SHORTENING IN THE CENTRAL ANDES AT THE TRANSITION TO FLAT-SLAB SUBDUCTION

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STATEMENT BY THE AUTHOR

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TABLE OF CONTENTS

Statement by the author ........................................................................................................1
Abstract ..................................................................................................................................4
Introduction .................................................................................................................................5
Background Information .............................................................................................................11
  Causes of flat-slab subduction ............................................................................................11
  Timing of slab flattening .......................................................................................................12
  Tectonic effects of the flat slab ............................................................................................15
  Timing of crustal deformation ...............................................................................................16
  Comparison of the central Andes and the Laramide orogeny ...........................................20
  Other possible factors influencing deformation style in the Andes ..................................21
Geologic Setting .........................................................................................................................26
  Regional structure ................................................................................................................29
  Rock units .............................................................................................................................32
  Cross-section .......................................................................................................................34
Analytical Methods ..................................................................................................................39
  Mineral separation ...............................................................................................................39
  Apatite fission track analysis ..............................................................................................40
  Apatite helium analysis ........................................................................................................43
  Zircon U-Pb analysis ...........................................................................................................45
Analytical Results ....................................................................................................................50
  Low-temperature thermochronology ..................................................................................50
  Geochronology ...................................................................................................................54
ABSTRACT

Shortening in the central Andes is considered to decrease north and south of the apex of the Bolivian orocline, partly owing to differences in the pre-existing stratigraphic architecture of the continental margin. Estimates of shortening in the central Andes of northern Argentina are scarce, but are required for assessment of the regional kinematic history of the orogenic system. The problem is acute at the north to south transition from the high elevation Puna plateau to the lower elevation region of the Sierras Pampeanas intra-foreland block uplifts, which corresponds with a transition to a flat segment of the subducting Nazca plate. Although deformation in the Eastern Cordillera appears to have propagated forelandward from west to east, the trend in the Principal Cordillera and the Sierras Pampeanas is not clear from existing data. Structures were mapped along a roughly E-W transect at latitude 28°S in the Sierra de Las Planchadas of the Principal Cordillera and a regional restorable cross-section was constructed for measuring total shortening. Low-temperature thermochronology was employed to help constrain the timing of exhumation and thrust propagation. A minimum of 60 km of shortening in the Sierra de Las Planchadas is estimated from the restored cross-section. When added to the 20 km shortening documented in ranges to the east by previous studies, this brings the total minimum shortening at this latitude to 80 km. Apatite fission track (AFT) ages from the hanging walls of thrusts are ~20 Ma, and apatite helium (AHe) ages range from 10.0 Ma west of the range to 2.3 Ma in the Fiambalá Basin which borders the range to the east. These ages are consistent with previously published AFT ages in the Fiambalá region and AHe ages in the Eastern Cordillera to the northeast and suggest synchronous deformation in the Principal Cordillera at 28°S and the Eastern Cordillera at 26°S. This implies that a single continuous thrust front connects the Eastern Cordillera and the Principal Cordillera in the southern central Andes.
INTRODUCTION

Subduction of an oceanic plate beneath a continental plate at an ocean-continent convergent margin is one of the most common processes responsible for the formation of large orogenic belts. Both the North and South American Cordilleras formed by this process. Although an oceanic plate typically subducts steeply into the mantle, it has been observed that in some cases sections of oceanic lithosphere flatten out and subduct very shallowly or sub-horizontally for an inboard distance of up to a thousand kilometers beneath a continent before steepening once again (Dickinson and Snyder, 1978; Gutscher et al., 2000; Jones et al., 2011). This process, called flat-slab subduction, appears to be an important first-order control on a number of upper-plate processes, including magmatism, shortening, and the structural style of crustal deformation (Dickinson and Snyder, 1978; Gutscher et al., 2000; Jones et al., 2011). Several modern flat slabs exist beneath the Andean orogenic belt, where the oceanic Nazca plate is subducting beneath the South American plate. The Andes thus provide an ideal laboratory for studying the tectonic causes and effects of flat-slab subduction.

The Central Andean Plateau is the second largest high-elevation plateau in the world after the Tibetan Plateau in central Asia and the first in a non-collisional setting. Often defined as the region above 3 km elevation, it spans 1800 km between Southern Peru and Northern Argentina and is up to 400 km wide (Allmendinger et al., 1997; Isacks, 1988).

The Central Andes were built by processes related to the convergence of the Nazca and South American plates, beginning during the Late Cretaceous. Analysis of basin deposits suggests that a regional foreland basin system first began to develop adjacent to the Bolivian Andes during the latest Cretaceous or early Cenozoic time (DeCelles and Horton, 2003; Horton, 2005; Sempere et al., 1997). The topography of the Central Andes is that of a classic ocean-
continent convergent margin. Nine tectonomorphic zones are present (Figure 1): (1) the offshore Peru-Chile trench of the subduction zone, (2) a remnant Mesozoic arc called the Coastal Cordillera, (3) a forearc basin, (4) the Precordillera, an inactive Paleogene arc, (5) the Western Cordillera, the modern-day arc, (6) the Altiplano Plateau, a high-elevation internally drained basin, (7) the Eastern Cordillera, which is part of the fold-thrust belt, (8) the Sub-Andean zone, the frontal part of the fold-thrust belt where most of the modern deformation is taking place, and (9) the Chaco Plain, a foreland basin above the Brazilian craton (McQuarrie et al., 2005). South of 24ºS, the Altiplano is replaced by the Puna, also a high-elevation, internally drained plateau broken up by an abundance of smaller basins and ranges, and the Sub-Andean zone is replaced by the Santa Barbara System, a thick-skinned fold-thrust belt (Allmendinger et al., 1997). South of 28ºS latitude, the fold-thrust belt geometry of the Altiplano and Puna is replaced by a narrow cordillera and the intra-foreland block uplifts of the Sierras Pampeanas. The magmatic arc is also extinguished at these latitudes (Cahill and Isacks, 1992).
The Central Andes are bordered to the east by a retroarc foreland basin system containing wedgetop, foredeep, forebulge, and backbulge depozones (Chase et al., 2009; Horton and DeCelles, 1997). The oldest foreland basin deposits are latest Cretaceous or early Paleocene in age, indicating the beginning of contractional deformation at that time (DeCelles and Horton, 2003; Sempere et al., 1997). The modern flexural forebulge of the Central Andes can also be observed in geoid data (Chase et al., 2009).

Figure 1 – Tectonomorphic zones in the Central Andes, after Carrapa et al. (2011b).
The shape and position of the subducting Nazca slab can be determined by observing the positions of mantle earthquake hypocenters in the “Wadati-Benioff Zone” in the slab. From 15°S to 27°S, the slab subducts at an angle of ~30 degrees. The slab then shallows out to be nearly horizontal from 28°S to 33°S, extending inland several hundred km with a dip of only 5 degrees. The slab steepens once again south of 33°S. A similar “flat slab” region is observed in the Northern Andes beneath Peru (Cahill and Isacks, 1992). Whereas the steep slab region of the Central Andes is associated with volcanism, a wide plateau, and thin-skinned deformation, the flat slab region lacks any modern volcanism and is characterized by a narrow or extinguished magmatic arc and basement-involved block uplifts which form the Sierras Pampeanas (Cahill and Isacks, 1992). The geometry of the Nazca plate beneath South America is displayed in Figure 2.

Figure 2: The geometry of the Nazca Plate beneath South America, as determined by earthquake hypocenters. Flat-slab segments are outlined with dashed lines.
The Central Andes exhibit large-scale along-strike variation in both the magnitude of crustal shortening and the style of deformation (Allmendinger and Gubbels, 1996; Kley and Monaldi, 1998). Kley and Monaldi (1998) compiled cross-sections from various studies in the Central Andes and constructed a curve of total shortening versus latitude from 5 °S to 40 °S. They then used this shortening curve to estimate crustal cross-sectional area along strike and compared these area estimates to estimates of crustal thickness based on topographic elevation compensated by airy isostacy (Figure 3). The crustal thickness predicted by airy isostacy is significantly greater than the thickness predicted by observed shortening, and the discrepancy is largest at the northern and southern edges of the Altiplano-Puna Plateau. This leads to speculation that other processes besides shortening have contributed to crustal thickening in the Central Andes. Proposed explanations (Kley and Monaldi, 1998) include tectonic underplating, lower-crustal transport of material, or pre-Neogene shortening. However, it is important to note that many of the cross-sections used in Kley and Monaldi (1998) are partial cross-sections that do not extend from the arc to the undeformed foreland, and the total shortening estimated from these cross-sections may therefore be an underestimate. Later work by McQuarrie et al. (2005) has found that the total shortening in the Bolivian Altiplano is greater than Kley and Monaldi (1998) estimate. McQuarrie et al. (2005) estimate that total shortening in the Bolivian Andes may exceed 500 km, which is enough to explain the observed crustal thickness in the Altiplano. The discrepancy persists in the Puna. However, it is possible that some shortening remains undocumented in the Puna Plateau. South of 28°S, in the region of the flat slab, documented shortening does appear to match the crustal thickness predicted by airy isostacy (Kley and Monaldi, 1998).
Figure 2: Shortening needed to produce crustal thickness estimated by assuming airy isostacy (black line) is compared to observed shortening from restored cross-sections at various latitudes in the Central Andes.

The purpose of this project is to examine the extent, style, and timing of crustal shortening along the southern flank of the Puna at a latitude of ~28° S, corresponding to the transition to flat-slab subduction. A balanced and restored cross-section and new thermochronologic and geochronologic data are presented to support an analysis of the kinematic history at this latitude.
BACKGROUND INFORMATION

Causes of flat-slab subduction

Flat subduction is a relatively common phenomenon, occurring along 10% of the world’s convergent margins (Gutscher et al., 2000). The causes of the flat slab segments are still debated. One theory posits that the changing curvature of the boundary between the two plates controls the angle of descent of the slab (Cahill and Isacks, 1992). Others have proposed that the subduction of oceanic lithosphere with thick oceanic crust causes the slab to become buoyant and flatten (Gutscher et al., 1999). The buoyant crust hypothesis is bolstered by the observation that in nearly all cases of flat subduction, the flat slab is correlated with a subducting oceanic ridge or plateau. In the Central Andes, this hypothesis is supported by earthquake hypocenter data (Anderson et al., 2007) which show that the shallowest location of the slab is at the inferred location of the subducted Juan Fernandez Ridge (Yanez et al., 2001). However, subduction of some large oceanic ridges does not result in flat subduction, such as the Iquique ridge subducting beneath the Bolivian orocline (Tassara et al., 2006). Still others have suggested that a high convergence rate with the overriding plate is the most important factor (Jarrard, 1986). However, it is unlikely that the overriding plate hypothesis alone can account for flat subduction in the Andes, as this hypothesis does not explain the along-strike segmentation in subducting slab angle which is observed in the Nazca plate regardless of the more or less constant westward velocity of the overriding South American plate.

A numerical model created by Espurt et al. (2008) predicts that flat slabs will occur only if a large segment of buoyant slab is forced to subduct by a fast overriding plate. Both a buoyant oceanic ridge or plateau and a high convergence rate are thus necessary preconditions for flat subduction. The Espurt model predicts that the flat slab will sink again at 600-700 km from the
trench, which agrees well with the known geometry of the Nazca slab (Gutscher, 2002). Another numerical model by van Hunen et al. (2002b) compares the importance of the effect of the buoyant ridge and the effect of the overriding plate velocity in producing a flat slab and concludes that the effect of the overriding plate velocity is 1-2 times greater than the effect of the buoyant ridge. Furthermore, while a buoyant oceanic ridge alone can produce a flat slab extending up to ~300 km from the trench, a fast overriding plate is needed to produce flat slab segments with the up to 500 km extent observed beneath Peru and northern Chile (van Hunen et al., 2002a). Thus it seems that a high convergence rate is a necessary precondition for producing a flat slab from buoyant, over-thickened oceanic crust.

**Timing of slab flattening**

The cessation of volcanism is characteristic of a flattening slab. As the asthenospheric wedge in front of the subducting slab is driven cratonward, the expression of volcanism at the surface also migrates cratonward, and in some cases eventually shuts off entirely. A similar pattern of inland-migrating volcanism is observed in the Rocky Mountains of North America from ca. 85 Ma to 40 Ma, which is attributed to a flat slab episode that also caused the Laramide Orogeny (Coney and Reynolds, 1977; Constenius, 1996).

The best evidence for the timing of the flattening of the slab beneath northwestern Argentina thus comes from an analysis of the migration and shutoff of arc volcanism at these latitudes, as outlined by Kay and Mpodozis (2002). The frontal arc at the latitudes of the flat slab transitioned from a continuous chain of andesitic stratovolcanoes to isolated dacite domes from 12-10 Ma. This is coeval with an increase in backarc volcanism. The frontal arc began to migrate eastward and broaden at 9-8 Ma and had completely shut off by 5 Ma. Backarc volcanism had
ceased by ~2 Ma, with the last known volcanic eruption at the latitudes of the flat slab occurring in the Sierra del Morro of the eastern Sierras Pampeanas at 1.9 Ma (Ramos et al., 2002). At latitudes north and south of the flat slab, arc volcanism migrated ~50 km eastward from 7-3 Ma but did not shut off.

The effects of the northeast trending arm of the Juan Fernandez ridge first arrived at the latitude of the modern flat slab region at ~14 Ma, and the east-west trending arm of the ridge began to subduct at ~10 Ma (Yanez et al., 2001). This timing is consistent with the beginning of eastward arc migration at 9 Ma. Figure 4 shows the reconstructed path of the Juan Fernandez ridge and the evolution of arc volcanism throughout the Neogene. It is important to note that reconstructions of the motion of the Juan Fernandez ridge are dependent on the reference frame used, although reconstructions from different reference frames seem to generally agree as far back as 20 Ma (Martinod et al., 2010).
Figure 4: Location of magmatic centers plotted with the position of the Juan Fernandez Ridge as determined by Yanez et al. (2001). The ridge position is indicated by the solid black line, and the predicted extent of the ridge’s influence on slab buoyancy is indicated by the dashed lines. Open circles are frontal arc eruptions and solid circles are backarc eruptions.
Tectonic effects of the flat slab

Both the deformation style and the total amount of crustal shortening in the Andes vary greatly along strike (Allmendinger and Gubbels, 1996; Kley and Monaldi, 1998). In the northern Central Andes, north of 24° S, the deformation is characterized by a thin-skinned fold-thrust belt where up to 340 km of shortening has been documented (McQuarrie and DeCelles, 2001). From 24° S to 28° S, the total crustal shortening decreases, and deformation is accommodated by the thick-skinned fold-thrust belts of the Santa Barbara System. South of 28° S, in the flat slab region, shortening is expressed much farther inland in the Sierras Pampeanas. It appears that the flattening of the slab both caused crustal deformation to migrate inland and altered the style of deformation.

The structural style of deformation in the Central Andes can be classified into three segments from north to south. In the northern Central Andes, the Altiplano is bounded to the east by a thin-skinned fold-thrust belt. South of 23° S, the Puna Plateau is bounded by the Eastern Cordillera, which consists of both thin- and thick-skinned structures. Active deformation is taking place in the Santa Barbara system. South of 28°S, in the region of the flat slab, thick-skinned deformation is characterized by block uplifting of the Sierras Pampeanas ranges along high-angle reverse faults. This change in structural style at the boundary of the flat slab region suggests that the slab angle is an important control on the structural style. However, other factors may be important as well, including the shape of the continental margin and along-strike variation in the thickness of the Phanerozoic sedimentary package (Allmendinger et al., 1997; McQuarrie, 2002a), or pre-existing basement structures (Grier et al., 1991).
Table 1 summarizes the above discussion. Magnitude of shortening, style of deformation, topographic characteristics, magmatic history, and slab angle are compared for three regions: the northern Central Andes of Bolivia, the southern Central Andes of Argentina, and the Sierras Pampeanas.

<table>
<thead>
<tr>
<th>Region</th>
<th>Shortening</th>
<th>Deformation Style</th>
<th>Topography</th>
<th>Active Arc?</th>
<th>Slab angle</th>
</tr>
</thead>
<tbody>
<tr>
<td>Northern Central Andes</td>
<td>&gt;500 km</td>
<td>Thin-skinned fold-thrust belt</td>
<td>Wide plateau</td>
<td>Yes</td>
<td>30 °</td>
</tr>
<tr>
<td>Southern Central Andes</td>
<td>50-150 km</td>
<td>Thin-/thick-skinned fold-thrust belt</td>
<td>Wide plateau</td>
<td>Yes</td>
<td>30 °</td>
</tr>
<tr>
<td>Sierras Pampeanas</td>
<td>50-150 km</td>
<td>Block uplifts</td>
<td>Narrow cordillera</td>
<td>No</td>
<td>5 °</td>
</tr>
</tbody>
</table>

Table 1: A comparison of three regions of the Andean orogen (Allmendinger et al., 1997; Kley and Monaldi, 1998; McQuarrie et al., 2005).

**Timing of crustal deformation**

North of the flat slab region, there is ample evidence to suggest that exhumation and uplift propagated mostly in sequence from west to east in the Eastern Cordillera. In northernmost Argentina, deformation propagated from the Eastern Cordillera eastward to the Subandean zone beginning at ~10-8 Ma. Deformation propagated in sequence from west to east across the Subandean zone until ~4.5 Ma, when some out-of-sequence thrusting began to occur (Echavarria et al., 2003). Apatite (U-Th)/He data suggest an eastward propagation of deformation across the Eastern Cordillera between 25 °S and 26°S, first reaching the Santa Barbara system at ~4 Ma (Carrapa et al., 2011b).
However, thermochronologic studies suggest non-uniform exhumation and uplift in the Northern Sierras Pampeanas. Some ranges may have been uplifted out of sequence. In the eastern Puna Plateau at ~27°S, the Sierra de Quilmes, which separates the El Cajón-Campo del Arenal basin from the Santa María basin, was uplifted at 6 Ma (Mortimer et al., 2007). This is coincident with, or possibly even subsequent to the uplift of the basin-bounding Sierra de Aconquija range to the east. Further evidence for possible out-of-sequence propagation of deformation comes from a comparison of AFT cooling ages of ranges in the southern Eastern Cordillera and the northern Sierras Pampeanas. Cooling ages in the Sierra de Aconquija are Late Cretaceous (Coughlin et al., 1998), whereas in the Chango Real range to the west, AFT cooling ages are Eocene-Oligocene (Coutand et al., 2001). If rock uplift is the cause of exhumation, this suggests that significant rock uplift and unroofing began in the Chango Real before the Aconquija. Still farther to the west, in the southern margin of the Puna Plateau, AFT cooling ages are early Miocene (Carrapa et al., 2006) AFT cooling ages from ranges throughout the northern Sierras Pampeanas actually suggest a trend of westward propagation of exhumation (Coughlin et al., 1998). Sedimentological proxies indicate a pulse of exhumation at ~ 6 Ma (Carrapa et al., 2008). Thus there is no clear directional trend for propagation of exhumation at latitudes south of 26°S.

Just south of the Puna Plateau, in the northern Sierras Pampeanas, two conflicting models exist to explain the timing of exhumation. There is evidence to suggest that the uplifts of the Sierras Pampeanas progressed from north to south from 7.6 to 2.6 Ma, based on the age of synorogenic deposits, AFT data, and hydrogen isotopes in alunite (Ramos et al., 2002). This suggests that the Nazca slab did not flatten coevally along strike, but rather the shallowing of the slab progressed from north to south. Davila and Astini (2007) studied the clast composition of
syn-orogenic conglomerates in the Sierra de Famatina of the northern Sierras Pampeanas (see Fig. 5 for the location of this range). They concluded that the foreland south of 27°S first became broken in the early Miocene, and that uplift of the Sierras Pampeanas progressed from north to south from the early Miocene to late Miocene-Pliocene time. This would suggest that uplift in the northern Sierras Pampeanas began before the collision of the Juan Fernandez ridge at this latitude (Yanez et al., 2001). Davila and Astini therefore propose that the slab may have begun to shallow before the collision of the ridge, and that shallowing of the slab initiated first in the north and propagated southward. However, the timing of uplift for many of the Sierras Pampeanas ranges is not well constrained. In particular, Ramos’ (2002) interpretation that the uplift of the ranges progressed from north to south depends on the apparent 2.6 Ma uplift of the Sierra de San Luis in the southern Sierras Pampeanas. The timing of uplift of the Sierra de San Luis is only poorly constrained by synorogenic deposits (Ramos et al., 2002).

Work by Carrapa et al. (2008) contradicts the Davila and Astini (2007) model. Sedimentologic data and U-Pb ages of ashes interbedded in Cenozoic deposits within several basins in the Sierras Pampeanas suggest a basin reorganization at ~6 Ma. The authors infer that broken foreland conditions initiated during the late Miocene, and that intra-foreland deformation occurred coevally along-strike (Carrapa et al., 2008). This is consistent with the timing of the collision of the Juan Fernandez Ridge and also implies that the slab shallowed coevally along strike.

Further thermochronologic analysis of both the Eastern Cordillera and the Northern Sierras Pampeanas ranges could shed more light on the timing of exhumation of the Plateau and the transition from a continuous to a broken foreland in the Sierras Pampeanas. Figure 5 is a topographic map with AFT cooling ages plotted from the three thermochronologic studies.
discussed above: Coughlin et al. (1998), Coutand et al. (2001), and Carrapa et al. (2006), as well as a vertical profile from Carrapa et al. (2011a). Although these studies cover a number of major ranges, systematic sampling of the northern Sierras Pampeanas and southern Puna Plateau has not been carried out and significant gaps in the data remain. These gaps must be filled in order to further our understanding of the timing of uplift in the region.

Figure 5: A summary of AFT ages (Ma) in the northern Sierras Pampeanas and southern Eastern Cordillera. Ages are from Coughlin et al., 1998 (green); Coutand et al., 2001 (red); Mortimer et al., 2007 (purple); Sobel and Strecker, 2003 (brown); Carrapa et al., 2006 (yellow) and Carrapa et al., 2011a (blue). Major mountain ranges that are mentioned in the discussion are also labeled.
**Comparison of the central Andes and the Laramide orogeny**

The Rockies were formed during subduction of the Farallon Plate beneath North America, which began during the late Jurassic (Coney and Reynolds, 1977). Tomography images indicate that the Farallon slab continues to sink into the mantle beneath North America and has currently reached a longitude of ~85 °W (Sigloch et al., 2008). The western margin of California is now a strike-slip boundary between the Pacific and North American Plates up to the Mendocino Triple Junction, where the Juan de Fuca plate meets the Pacific and North American plates. Although the Juan de Fuca slab currently subducts steeply, there is ample evidence to suggest that the Farallon slab experienced a period of flat subduction that is temporally correlated with the Sevier and Laramide orogenies and the formation of the Rocky Mountains (Coney and Reynolds, 1977; Constenius, 1996; Dickinson and Snyder, 1978; Jones et al., 2011). Evidence for a Farallon flat slab can be seen from the migration pattern of volcanism in western North America in the time interval ca. 85-40 Ma. Volcanism had migrated to a position nearly 1,000 km inland from the modern-day margin by 40 Ma (Figure 6) (Coney and Reynolds, 1977; Constenius, 1996). The eastward migration of volcanism could be explained by a shallowing of the subducting slab angle. Volcanism in western North America migrated back to the west from 40 Ma to ~15 Ma (Coney and Reynolds, 1977; Constenius, 1996), suggesting that the Farallon slab began to steepen again and the hinge “rolled back” to the west. This steepening of the slab could explain the collapse of the North American Cordillera at ~30 Ma and the initiation of extensional tectonics in the Western United States (Constenius, 1996). In the North American Rockies, the basement-cored Laramide uplifts extend more than 1000 km from the former trench of the subducting Farallon plate and more than 500 km from the easternmost extent of the Sevier fold-thrust belt (Burchfiel and Davis, 1975). The Laramide uplifts are thus very similar in style.
to the Sierras Pampeanas uplifts of the Central Andes. If the Farallon slab did indeed experience a period of flat-slab subduction, then the basement-cored uplifts of the Laramide orogeny may be an older counterpart of the more recent basement-cored uplifts in the Sierras Pampeanas of northwestern Argentina produced by the Argentine flat slab (Jordan et al., 1983).

![Figure 6](image)

*Figure 6: Location of the volcanic arc in the Southwestern United States from 85 Ma to present.*

**Other possible factors influencing deformation style in the Andes**

The width of the Andean orogen appears to be related to the angle of the subducting slab. Where the slab subducts steeply the orogen is characterized by a wide plateau bounded by the Western and Eastern Cordilleras. Conversely, where the slab is shallow, the orogen is characterized by a single narrow cordillera. In order to test how strongly the orogen width correlates with the angle of the slab, the width of the orogen above 1000 m elevation was measured at 5 degree latitude increments from 10 °S to 35 °S (Figure 7). These measurements were then compared to the slab angle at each latitude, which was estimated from slab depth
contours published by Cahill and Isacks (1992). Slab angle was determined trigonometrically using the slab depth and the down-dip length of the slab between 50 km and 150 km depths (Figure 8). Figure 9 is the resulting plot of orogen width versus slab angle. A trend line fit to the data by linear regression has a positive slope, indicating a positive correlation between orogen width and slab angle. However, the $R^2$ value of the linear regression is 0.07, so the trend is not strongly expressed. Thus it can be concluded that slab angle and orogen width are only loosely positively correlated. It is likely that other factors, such as the lithospheric architecture of the upper plate and pre-existing basement structures, are also important controls on orogen width.

**Figure 7:** The width of the Andean orogen, measured at 5° latitude intervals.
Figure 8: The distance from a slab depth of 50 km to 150 km. This distance is then used to trigonometrically calculate the average slab angle at each latitude.
Figure 9: Slab angle vs. orogen width. A linear regression trend line (red) has a positive slope, indicating that orogen width and slab angle are positively correlated. However, it is important to note that the $R^2$ value of this trend line is only 0.07, so it is not a statistically strong trend.

In addition to the slab angle, pre-existing basement fabrics may be an important control on both the timing and style of deformation in the Central Andes. The Cretaceous Salta Rift zone extends from ~23°S to ~26°S, and many of the high-angle normal faults associated with the rift have been reactivated as thrusts during Andean deformation (Grier et al., 1991; Monaldi et al., 2008). Reactivation of these rifts results in the block-uplift style of deformation which characterizes the Sierras Pampeanas. Thermochronologic evidence also suggests that pre-existing crustal architecture is an important control on the style of deformation. Cretaceous cooling ages in the southernmost Eastern Cordillera are interpreted to indicate paleotopographic highs on the flanks of Salta Rift basins which escaped burial and resetting of the cooling ages during the Cenozoic (Carrapa et al., 2011a). It is also important to note that geographically the Sierras Pampeanas are not entirely correlated with the flat slab. The southernmost extent of the Sierras
Pampeanas is ~35.5 °S, whereas the flat slab extends only to ~33 °S. Furthermore, any attempt to correlate the slab angle with the style of deformation involves comparing present day slab geometry with a structural style that was acquired in the past. This suggests that pre-existing crustal weaknesses may be a more important control than the present-day underlying flat slab on the style of crustal deformation. In addition to pre-existing structures, it is important to consider the architecture of the crust itself. In areas where a thick package of Paleozoic strata exists, thin-skinned fold-thrust belts can form. However, if very little sedimentary strata exist above the igneous basement rocks, thick-skinned deformation is likely to dominate (Allmendinger et al., 1997; McQuarrie, 2002b).
GEOLOGIC SETTING

Overview of regional geology

Figure 10 – Regional geologic map of southwestern Catamarca province published by SEGEMAR (Nullo et al., 1995). The field area for this study is indicated by the black outlined box.

Mapping for this project was focused on the Sierra de Las Planchadas and the adjacent Fiambalá Basin in the southern Catamarca province. Figure 10 outlines the field area on a regional geologic map. The Sierra de Las Planchadas bounds the Fiambalá Basin to the west and is composed mostly of a Paleozoic sedimentary package consisting of Ordovician siltstone, an ~200 m layer of Carboniferous sandstones and shales, and Permian eolian sandstone. The Sierra de Las Planchadas is part of the Principal Cordillera of the Andes, where thin-skinned deformation prevails. The Fiambalá basin is bounded by the Sierra de Fiambalá to the east,
which is composed entirely of Cambrian and Precambrian igneous basement intruded by Carboniferous granites. The Sierra de Fiambalá is one of the northernmost ranges of the Sierras Pampeanas. In the Sierra de Fiambalá, and in all major ranges to the east of the Fiambalá basin, thick-skinned block-uplift style deformation prevails. The Fiambalá basin contains an ~5 km thick section of upper Miocene to lower Pliocene synorogenic deposits which have been faulted and folded (Carrapa et al., 2008).

West of the field area is the Principal Cordillera of the Central Andes. The Paleozoic stratigraphic section here is quite distinct from the stratigraphy of the Sierra de las Planchadas. It consists of Devonian quartzite, a thick section of Carboniferous siltstones and shales, and Permian sandstone. Farther to the west, near the Laguna de los Aparejos, Ordovician sandstones and siltstones are also present in the section. The fact that the stratigraphy of the Principal Cordillera differs so markedly from stratigraphy in the Sierra de Las Planchadas suggests the existence of an intervening terrane boundary. Indeed it has been proposed that four different terranes all meet in the region of Los Aparejos: Chilenia, Cuyania, Famatina, and Antofalla. The Sierra de Las Planchadas is located on the Pampia terrane (Figure 11) (Ramos, 2009).
Figure 11 – Terrane boundaries in Northwestern Argentina. The black rectangle indicates the field area for this study.
Regional structure

The Fiambalá basin is bounded to the west by an east-verging high angle reverse fault that places Permian sandstone atop the Neogene synorogenic basin fill. The eastern edge of the basin is bounded by a buried west-verging high angle reverse fault with Cambrian and Precambrian granite and gneiss in the hanging wall. To the north, the basin has a depositional contact with the Cambrian and Precambrian igneous basement. To the south the basin is open and drains into the many smaller basins between intra-foreland uplifts of the Sierras Pampeanas system.

The Sierra de Las Planchadas exhibits several thrust faults and folds. One large thrust sheet has emplaced Ordovician siltstone atop Permian sandstone. A window eroded down through this thrust reveals folded Paleozoic section beneath. Although this structure was previously interpreted as two separate faults on a regional map published by SEGEMAR (Rubiolo et al., n.d.), I have reinterpreted it as a window. This interpretation is supported by satellite imagery that suggests that the fault is a single continuous structure and by chattermarks observed on the eastern exposure of the fault which indicate top-to-the-east sense of motion on the fault. In turn this suggests that the two previously mapped thrust faults with Ordovician rocks in their hanging walls are a single, eastward-verging major thrust fault (Fig. 13, note #9).

The Principal Cordillera contains numerous thrust faults. One major thrust places Devonian quartzite atop Permian sandstone. West of this fault the fold-thrust geometries are progressively covered by extensive Cenozoic volcanic rocks.

Maps of this area at the 1:250,000 scale had been published by SEGEMAR (Rubiolo et al., n.d.). However, these maps lack sufficient structural data to produce a cross-section. For this
study I mapped the Sierra de Las Planchadas and Fiambalá Basin at a more detailed 1:100,000 scale and collected structural data. Figure 12 is the geologic map produced in this study.
Figure 12 – Geologic map of the field area, with sample locations and line of cross-section
Rock units

Ordovician sedimentary rocks – The Ordovician section consists of fine-grained sandstone and siltstone. It is dark green with some tan beds. It is usually planar bedded or rippled and some beds contain burrows. It often cleaves parallel to bedding, but pencil cleavage is also observed in some localities. In some regions it is slightly metamorphosed. The base of the Ordovician is not exposed in the Sierra de Las Planchadas, so its thickness is not constrained, but is at minimum 1000 m.

Ordovician volcanic rocks – The Ordovician volcanic rocks are basaltic and are black or reddish colored. They are fine grained with plagioclase phenocrysts. These volcanic rocks are extensive in the southern part of the Sierra de Las Planchadas, where they blanket nearly the entire range.

Ordovician igneous intrusions – There are two major Ordovician-age plutonic units which intrude the Ordovician sedimentary rocks: a dark green, coarse-to-medium-grained granodiorite, and a coarse-to-medium-grained granite containing pink K-spar and chlorite. The granite is white but often altered to a rusty orange color. The granodiorite is observed in the central eastern part of the Sierra de Las Planchadas. Two major intrusions of the granite are observed: one in the western central part of the range and the other in the southeastern part of the range. A smaller intrusion of granite is also observed in the southern tip of the range.

Devonian – The Devonian section is not present in the Sierra de Las Planchadas, but it is present in the Principal Cordillera to the West of the field area, where it has been extensively mapped and dated by SEGEMAR (Rubíolo et al., n.d.). It consists almost entirely of fine-grained quartzite and contains occasional matrix-supported conglomerate beds. The clasts are fine gravel
sized and are mostly quartzite with a few granites. Clasts are rounded to subrounded and well sorted. The thickness of the Devonian is at minimum 3500 m.

Carboniferous – The Carboniferous section is ~200 m thick in the Sierra de Las Planchadas. It consists of shales, quartzites, and coarse-grained sandstones. The sandstones consist of quartz and feldspar grains. Some matrix-supported beds of pebbly conglomerate are present, with well-rounded poorly-sorted quartzite clasts. The shales are usually light grey, bluish grey, pastel green, or tan. It is usually planar bedded or rippled and contains burrows. The quartzite is fine-grained and white.

Permian – The Permian section is ~1000 m thick in the Sierra de Las Planchadas. It consists mostly of eolian sandstone capped by ~100 m of lacustrine facies. The sandstone is dark red and medium-grained. It contains quartz and feldspar grains and exhibits eolian trough cross strata. The lacustrine deposits at the top of the section are lower medium to upper fine grained sandstone interbedded with mudstones. Some mudstone layers have mudcracks. The sandstones are finely laminated with shaly cleavage. Some layers have a calcareous component and others are purely siliciclastic. The calcareous layers contain chert nodules. West of the field area in the Principal Cordillera the Permian sandstones contain some conglomerates. These conglomerates are poorly sorted, clast supported, and tabular bedded.

Upper Miocene and Pliocene syntectonic sedimentary rocks:

Tambería Formation – The Tambería Formation is a fluvial unit which consists of lower and upper members. Both members are ~1000 m thick. The lower member consists of tan medium-grained, tabular-bedded sandstones and mudstones. The sandstones are well sorted. The mudstones contain mudcracks and trace fossils. The upper member is a tan sandstone with matrix-supported channel conglomerates interbedded with grey lacustrine beds. Clasts in the
conglomerate are derived from the Permian sandstones, Devonian quartzites, and Ordovician granites. The sandstone exhibits both planar bedding and trough cross-strata. The Tambería Formation has been dated at 9.0-8.2 Ma based on magnetostratigraphic data (Reynolds, 1987) and zircon U-Pb dating of interbedded volcanic tuffs (Carrapa et al., 2008).

Guanchín Formation – The Guanchín Formation is an ~1500 m thick grey and pink colored fluvial sandstone containing channel conglomerates and interbedded with volcanic tuffs. It is planar bedded, trough cross stratified, and poorly consolidated. The Guanchín Formation has been dated at 8.2-5.5 Ma by zircon U-Pb dating of tuffs (Carrapa et al., 2008).

Punaschotter Formation – The Punaschotter Formation unconformably overlies the Guanchín and Tambería Formations. It is a poorly sorted, matrix supported cobble conglomerate. Some sections are horizontally stratified and some sections are massive. It is very poorly consolidated. The Punaschotter Formation has been dated at ~3.7 Ma by zircon U-Pb dating of tuffs (Carrapa et al., 2008).

Cross-section

I produced a cross-section (Fig. 13) based on the map data gathered in this study and the map published by SEGEMAR (Rubiolo et al., n.d.). The cross-section is line-length balanced and is a minimum-shortening interpretation. Subsurface data were not available at this latitude, so the deeper geometries of the cross-section are poorly constrained.

The following is a summary of assumptions made in constructing the cross-section. The locations where assumptions were made are indicated on the cross-section by number.

(1) The angle of the eastern basin-bounding fault is not constrained but is chosen at 45° to match the angle of reverse faults northeast of the field area documented by Grier et al. (1991).
(2) Bedding thickness variations are observed in the Miocene units in the Fiambalá Basin, indicating syntectonic growth strata. Thicknesses of these units in the western part of the basin are constrained by the map pattern. Thicknesses in the eastern part of the basin are chosen to be consistent with the measured section in Carrapa et al. (2008).

(4) The thickness of the Guanchín Formation outcropping to the east requires some erosion of the Guanchín in the footwall block of this fault. The Punaschotter Formation was then deposited post-erosion. A late-stage episode of slip on the fault then cut the Punaschotter Formation.

(6) A flat is inferred at the base of the Upper Tambería Formation in order to explain the absence of the Lower Tambería Formation in the easternmost thrust sheets.

(7) The depth to the basement is explained by the thicknesses of the Miocene and Paleozoic units. The thicknesses of Miocene, Permian, and Carboniferous units are all constrained by their map patterns and bedding orientations. The thickness of the Ordovician was chosen to fill the space in the hanging wall of the western basin-bounding fault (7a). The space that needs to be filled in the hanging wall of this fault is determined by the displacement on the fault, which is constrained by the thicknesses of the Miocene units in the footwall.

(8) The geometry at this branch line is exposed at the surface ~3 km north of the line of cross-section.

(9) The interpretation of this east-dipping fault as part of a continuous thrust sheet with the fault to the west of the anticline is based on a top-to-the-east sense of motion on this fault indicated by chattermarks.

(10) The location of the footwall ramp is constrained by the restoration of the bed length of the Lower Tambería Formation.
(11) This inferred backthrust explains the contact of Ordovician granite above Permian sandstone ~5 km north of the line of cross-section.

(12) The angle of the footwall ramp is constrained by the dip of the overlying syntectonic Upper Tambería Formation (12a).

(14) The presence of a thick layer of Carboniferous at depth is inferred from thick Carboniferous outcrops south of the field area that have been mapped by SEGEMAR (Rubiolo et al., n.d.).

Figure 14 is a stepwise restoration of the cross-section. Timing constraints are indicated based on the cross-cutting relationships between faults and the syntectonic sediments in the Fiambalá Basin. When restored, this cross-section shows a minimum of ~60 km of shortening. More shortening may be present if a duplex exists at depth, but subsurface data are needed to confirm this. Significantly less shortening is present in the ranges to the east of Fiambalá where the block-uplift style of deformation dominates. When our cross-section is combined with shortening estimates northeast of the field area (Allmendinger, 1986), this brings the total documented shortening at latitude 28° to ~80 km.
Figure 13 – Geologic cross-section
Figure 14 – Stepwise restoration of the cross-section. Timing constraints are based on crosscutting relationships between faults and syntectonic sediments in the Fiambalá Basin.
ANALYTICAL METHODS

Mineral separation

Zircon (ZrSiO$_4$) and apatite (Ca$_5$(PO$_4$)$_3$(OH,F,Cl)) are the two minerals used in this study for thermochronologic and geochronologic analysis. In order to be analyzed, these minerals must be separated from the other minerals in the rock sample. Both of these minerals are dense: zircon has a density of 4.65 g/cm$^3$, and apatite has a density of 3.19 g/cm$^3$. Unlike most dense minerals, zircon and apatite are also non-magnetic. Density and magnetism are thus the two characteristics that are traditionally used to separate zircons and apatites from a rock sample.

Samples collected in the field were broken with a sledgehammer into cobbles which were then crushed into gravel in a jaw crusher. The gravel was then pulverized in a roller mill. The resulting sand was run through a Wilfley table, which separates the sand grains by density. The densest fraction of the sand was run through a 500 micron sieve, and grains larger than 500 microns were discarded.

Sand grains then underwent two further steps of mineral separation. The sand was processed through a Franz magnetic separator device, which removes magnetic grains. The non-magnetic grains obtained from the Franz were then placed in liquid Methylene Iodide (MI), which has a density of 3.2 g/cm$^3$. Zircons sink in MI, while most other minerals float. A second round of density separation using MI diluted with acetone to 2.8 g/cm$^3$ was used to separate out the apatites. Both the apatite and zircon separates were passed through the Franz magnetic separator one final time to remove any remaining magnetic minerals.

All mineral separation for this study was carried out in the lab managed by Prof. George Gehrels at the University of Arizona.
Apatite fission track analysis

Apatite fission track analysis is a low-temperature thermochronometric method (e.g. Gallagher, 1998). When an apatite crystal forms, the crystal lattice contains some impurities of uranium. These uranium atoms decay to lead over time. One process by which uranium decays is nuclear fission, in which the $^{238}\text{U}$ splits into two particles. The two positively charged particles then repel each other in opposite directions, producing a trail of damage in the crystal known as a fission track. Due to the high relative abundance of $^{238}\text{U}$ to other isotopes of uranium, $^{238}\text{U}$ is effectively the only isotope which contributes to spontaneous fission track formation in the crystal (Tagami and O'Sullivan, 2005). If the concentration of $^{238}\text{U}$ and the density of tracks are known, an age can be calculated for the crystal.

At temperatures above ~110°C, the crystal lattice of apatite anneals and tracks are quickly erased from the crystal after they form. At temperatures below ~60°C the tracks are fully retained in the crystal (Gallagher et al., 1998). The temperature range 110-60 °C is known as the partial annealing zone (PAZ) (Gallagher et al., 1998). In this temperature zone, tracks get shorter over time and are eventually erased. The exact temperature limits of the PAZ are dependent on the cooling rate of the sample, with quickly cooled samples experiencing a smaller PAZ than slowly cooled samples (Gallagher et al., 1998). The fission track age calculated represents a time when the crystal was within the 110-60 °C temperature range. Since such temperatures are generally present at ~4.5 - 2.5 km depth (assuming a 25 °C/km geothermal gradient), the fission track age represents the time when the sample was exhumed through this depth on its way to the Earth’s surface. An apatite fission track age thus typically represents the timing of exhumation of the sample, as opposed to higher-temperature thermochronometers which typically yield a crystallization age.
Apatite grains are mounted in a chip of epoxy, which is polished down until the middles of the grains are exposed. The grains are then chemically etched according to the protocol of Murrell et al. (2009) until the tracks have widened enough to be viewed and counted under a microscope. The etching process widens the tracks without damaging the crystal because the damaged zones in the crystal lattice around the tracks dissolve much faster than the rest of the grain.

There are several methods for measuring the uranium concentration in the crystal. The method used in this study is known as the external detector method. A thin sheet of muscovite is placed over the etched grains, and the sample is then bombarded with thermal-neutrons in a nuclear reactor. Thermal neutrons induce fission in $^{235}$U (Donelick et al., 2005), and as particles are ejected from the apatite grains, they produce tracks in the muscovite. Each individual apatite crystal forms a print of tracks on the adjacent muscovite. These tracks are known as induced tracks, as opposed to the spontaneous tracks that accumulated naturally in the apatite over time.

Multiple samples interbedded with standards are stacked vertically in a tube and loaded into the reactor for bombardment. Glass dosimeters of known uranium concentration are bombarded with the samples in order to calculate the neutron flux. In some reactors, the neutron flux is constant throughout the sample tube, and only a single dosimeter glass is needed. However, in many reactors there is a gradient in neutron flux from the top of the tube to the bottom of the tube. In this case, a dosimeter glass is placed at both the top and bottom of the tube, and the gradient of neutron flux is calculated. From this gradient, a dosimeter track density is calculated for each sample based on its position in the tube.

After neutron bombardment, the muscovite is then etched in 5.5 M nitric acid at 21 °C for 20 seconds. The epoxy chip containing the apatites and the muscovite are placed side by side
under a microscope, and ~20 grains are selected for analysis. The spontaneous tracks are counted for each grain, and the induced tracks are counted in the corresponding print in the muscovite. The $^{235}\text{U}$ concentration of each grain can be calculated from the density of the induced tracks in the muscovite and the neutron flux. Since the relative abundance of uranium isotopes is a known constant, the $^{238}\text{U}$ concentration can then be calculated for the grain.

The age of an apatite is calculated from the equation:

$$t = \frac{1}{\lambda} \log \left( 1 + \frac{1}{2} \lambda \xi \rho_d \frac{\rho_s}{\rho_i} \right)$$

Where $\lambda$ is the decay constant for $^{238}\text{U}$, $\rho_s$ is the density of spontaneous tracks, $\rho_i$ is the density of induced tracks, and $\rho_d$ is the density of tracks in the dosimeter glass, or in this case, the calculated density for the position of the sample in the tube. The constant $\xi$ is a calibration factor to account for analyst bias in choosing which tracks to count. Each analyst must count a large number of standards to determine their personal $\xi$ factor before analyzing fission track samples. The above equation is derived in detail in Galbraith (2005).

For any basement sample with a simple cooling history, an average of twenty individual grain ages are usually obtained and should pass a $\chi^2$ statistical test (Galbraith, 2005). If the sample does not pass the $\chi^2$ test, then there is too much dispersion in the grain ages, and the sample has likely been partially reset through burial to a depth within the PAZ. Certain grains anneal faster than others, mostly according to the concentration of various elements in the grain, and thus an age dispersion develops. For detrital samples, the $\chi^2$ test indicates whether the sample has been reset to a homogenous age through burial below the PAZ.

Neutron bombardment for this study was carried out in the reactor at Oregon State University, and fission track analysis was carried out at the University of Arizona in the Fission Track Lab managed by Prof. Barbara Carrapa and Dr. Stuart Thomson.
**Apatite helium analysis**

A second thermochronometer used in this study is the (U-Th/He) system (Farley et al., 1996 for a review). When applied to apatites, this is referred to as apatite helium (AHe) dating. Helium accumulates in apatites over time as the result of radioactive decay of $^{238}\text{U}$, $^{235}\text{U}$, and $^{232}\text{Th}$ to various isotopes of lead. All three of these systems produce α particles (helium nuclei) in the steps of their decay chains. $^{147}\text{Sm}$ also makes a contribution to radiogenic helium production in apatites, but its contribution is insignificantly small. The equation for helium accumulation in an apatite grain is:

$$^4\text{He} = 8*^{238}\text{U}(e^{\lambda_{238}t}-1) + 7*^{235}\text{U}(e^{\lambda_{235}t}-1) + 8*^{232}\text{Th}(e^{\lambda_{232}t}-1)$$

Where $^4\text{He}$, $^{238}\text{U}$, $^{235}\text{U}$, and $^{232}\text{Th}$ are the respective concentrations of those elements, $\lambda$ is the decay constant for each system, and $t$ is the cooling age of the grain in years (Harrison and Zeitler, 2005).

Like AFT dating, the AHe system is sensitive to low temperatures, because at temperatures above ~85 °C helium diffuses out of the apatite grain (Wolf et al., 1998). The AHe system experiences a partial retention zone (PRZ), which is the equivalent of the PAZ in the fission track system. The PRZ for helium in apatites is 85-40 °C, and as in the AFT system the exact range of temperatures for partial retention depends on the cooling rate (Wolf et al., 1998). Assuming a geothermal gradient of 25 °C/km, AHe analysis thus gives an age for exhumation through depths of ~3.5-1.5 km. Because AHe is sensitive to lower temperatures than AFT, AHe ages are expected to be younger than AFT ages for any given sample.

One major complication with the (U-Th)/He method is the implantation or loss of helium from the outer rim of the grain due to “α ejection” (Reiners, 2002). When a decay event occurs, the α particles produced have a large kinetic energy and are able travel up to 20 μm within the
crystal lattice. Near the surface of the grain, this causes some α particles to be ejected from the grain and lost. This loss of helium from the grain can cause an error of 17-40% in the calculated cooling age (Reiners, 2002). For euhedral crystals, if the crystal dimensions are known, it is possible to calculate a correction to account for ejected α particles (Farley et al., 1996; Hourigan et al., 2005). Thus each grain must be imaged and measured before analysis. The correction is more significant for smaller grains, so the larger grains in a sample are preferred for AHe analysis.

In this study apatite grains were selected for analysis based on large size and a lack of inclusions. In the case of samples that had already undergone AFT analysis, apatite grains were plucked from the epoxy mounts. The grains were imaged with an optical microscope, and the three axial dimensions of each grain were measured from the images. Grains were then wrapped in Nb foil for analysis. Standard grains of sample 07WFS, which has an age of 65 ± 9 Ma (Reiners et al., 2007), were also prepared and analyzed along with the unknown samples.

Grains were analyzed according to the methods described in Reiners (2002). The grains were heated with a laser to extract the $^4\text{He}$ gas. The $^4\text{He}$ gas was then cryogenically purified and spiked with $^3\text{He}$. The $^4\text{He}/^3\text{He}$ ratio was measured with a quadrupole mass spectrometer. The grains were then retrieved and the U and Th concentrations were measured by isotope dilution ICPMS. An age was calculated for each grain and corrected for α ejection based on the grain’s dimensions.

All AHe analysis was carried out at the University of Arizona (U-Th)/He laboratory managed by Prof. Peter Reiners. The analytical procedures carried out in Prof. Reiners’ lab are described in further detail on the web page of the Arizona Radiogenic Helium Dating Laboratory: http://www.geo.arizona.edu/~reiners/arhdl/procs.htm (Reiners, 2007)
Zircon U-Pb analysis

Uranium-lead analysis was carried out on zircons (ZrSiO₄). The zircon U-Pb system has a closure temperature of >800°C (Gehrels, 2010), so unlike the AFT and AHe methods described above, zircon U-Pb analysis yields the crystallization age of the zircon and the igneous host rock from which it was derived.

Zircon is an ideal mineral for U-Pb analysis because the crystal lattice contains substitutions of uranium atoms, whereas Pb is mostly excluded during crystal formation. Thus there is very little initial daughter product in a zircon grain (Gehrels et al., 2008). Over geologic timescales the uranium decays to lead in the crystal: ²³⁸U decays to ²⁰⁶Pb with a half-life of 4.5 billion years, and ²³⁵U decays to ²⁰⁷Pb with a half-life of 700 million years. Both of these systems can independently be used to calculate an age. The age equations are:

\[ t = \frac{\ln\left(\frac{²⁰⁶Pb}{²³⁸U}\right) + 1}{\lambda_{²³⁸}} \]

\[ t = \frac{\ln\left(\frac{²⁰⁷Pb}{²³⁵U}\right) + 1}{\lambda_{²³⁵}} \]

Where \( t \) is the age in years, \( U \) is the concentration of uranium, and \( Pb^* \) is the concentration of radiogenically produced lead, as opposed to common lead (Gehrels, 2010).

Due to the low abundance of ²³⁵U, it is analytically simpler to measure the ratios of ²⁰⁶Pb/²⁰⁷Pb and ²⁰⁶Pb/²³⁸U. The ²⁰⁷Pb/²³⁵U ratio can be calculated from these measurements because the relative abundance of ²³⁸U/²³⁵U is constant at 137.88 for crustal rocks (Steiger and Jäger, 1977).

Two sources of error exist in the raw measurements which must be corrected. The first is the presence of a small concentration of initial common lead in the zircon crystal. We wish only to measure the radiogenically produced lead. Since the isotope ²⁰⁴Pb is not produced...
radiogenically, the concentration of common lead can be determined by measuring the ratios of \( \frac{^{206}\text{Pb}}{^{204}\text{Pb}} \) and \( \frac{^{207}\text{Pb}}{^{204}\text{Pb}} \) (Gehrels, 2010).

A correction must also be applied for fractionation between uranium and lead during the measurement process. In the LA-ICPMS method, this fractionation occurs because as the depth of the laser pit increases in the zircon grain, uranium atoms preferentially condense on the pit wall and are lost from the analysis (Hanchar and Hoskin, 2003). Fractionation is corrected by periodically taking measurements on standards of a known age throughout the analysis of unknown samples.

Several methods exist for U-Pb analysis, including isotope dilution thermal ionization mass spectrometry (ID-TIMS), secondary ion mass spectrometry (SIMS), and laser ablation inductively coupled plasma mass spectrometry (LA-ICPMS). LA-ICPMS is the method used in this study. Zircon grains are mounted in epoxy and polished down until the centers of the grains are exposed. The zircons are then shot with a laser beam that ablates a small portion of the grain. The spot diameter of the laser can range from 10-35 \( \mu \text{m} \) according to the spatial resolution required for the analysis. The atoms released by ablation are then carried by helium gas through a plasma torch, which ionizes the atoms. The ions are then analyzed with a mass-spectrometer. This method is described in detail in Gehrels et al., 2008.

In this study, U-Pb data are plotted on Wetherill concordia diagrams (Wetherill, 1956) (Figure 15a). These diagrams plot the \( \frac{^{206}\text{Pb}}{^{238}\text{U}} \) ratio against the \( \frac{^{207}\text{Pb}}{^{235}\text{U}} \) ratio. The concordia curve represents all points at which the 206/238 age agrees with the 207/235 age. An age is said to be “concordant” if it plots on the concordia curve and “discordant” if it plots off of the curve. Discordant ages plot below the concordia and reverse discordant ages plot above the concordia. Reverse discordant ages are typically the result of analytical error. There are two common causes
of discordant ages: loss of lead from the crystal or inheritance of older cores within the crystal grain.

Lead loss causes the calculated age to be younger than the real age of the sample, and $^{207}\text{Pb}$ and $^{206}\text{Pb}$ are lost in proportion to their relative composition in the grain. Thus as a grain loses lead over time its measured ratios move diagonally away from the concordia towards the origin. In samples which experienced a lead loss event at some time in the geologic past, the samples plot along a tie-line beneath the concordia curve which connects the age of the sample and the age at which the lead loss occurred. This tie-line is known as the “discordia” (Figure 15b). The discordia can be used to project analyses back to their original crystallization age.

Discordant ages can also result from grains that contain older xenocrystic cores. In this case the calculated age is actually a mixed age between the age of the outer domain of the zircon and the age of the core, and data will plot on a tie-line beneath the concordia curve which connects the ages of the two domains. Since zircon is a resistant mineral with a high melting point, it is not unusual for zircon to survive reincorporation into a magma and become preserved as xenocrystic cores within younger zircons. Thus it is important when dating igneous samples to take CL images of the grains before analysis. The CL image can allow the analyst to identify possible xenocrystic cores within the zircons, and the older and younger domains can be analyzed separately.
Figure 15: (a) A Wetherill concordia diagram. (b) Lead loss causes analyses to plot below the concordia on a tie-line between the original crystallization age and the age of the lead-loss event.
In this study zircons from igneous samples were isolated by standard mineral separation techniques. CL images were obtained to identify possible inherited domains within the grains. The zircons were then analyzed by LA-ICPMS. During analysis, older inherited cores were analyzed separately from the younger domains. Results are reported as Concordia plots for each sample.

Detrital samples were collected from the Fiambalá Basin for provenance analysis. After standard mineral separation to concentrate the zircons, the zircons from detrital samples were poured onto mounts in order to prevent any bias in the sampling. These mounts were then polished and analyzed by LA-ICPMS. The results are reported as both concordia plots and probability distribution plots which show the relative distribution of ages in the detrital samples.

Sample preparation and analysis were carried out in Prof. George Gehrels’ lab at the University of Arizona. Further details on analytical procedures in Prof. Gehrels’ lab can be viewed on the web page of the Arizona LaserChron Center at:

https://sites.google.com/a/laserchron.org/laserchron/home/ (Gehrels, 2010)
ANALYTICAL RESULTS

**Low-temperature thermochronology**

Samples were collected from the hanging walls of major thrust faults in the field area. The geologic map (Figure 12) indicates the locations of samples collected for thermochronologic analysis. Samples from the Sierra de Las Planchadas were analyzed by AFT, and some samples were also analyzed by AHe. Samples from the Miocene strata in the Fiambalá basin were analyzed only by AHe because prior work by Carrapa et al. (2006) showed that these strata were never buried deeply enough for AFT ages to be reset.

Table 2 summarizes AFT data. Table 3 summarizes AHe data. It is clear that some of these samples have been partially reset by burial. Two AFT samples do not pass the $\chi^2$ test, but the ages obtained are younger than the stratigraphic age, indicating partial resetting. There is significant dispersion in all of the AHe ages except for sample 10.02. In the case of partially reset AHe ages, the youngest or most reset age will be the closest to the true exhumation age.

Proper interpretation of thermochronologic ages requires an understanding of the structural context of the samples. Figure 16 displays the sample locations and ages obtained on one version of the cross-section constructed in this study, providing the necessary structural context.
<table>
<thead>
<tr>
<th>Sample</th>
<th>Longitude (°)</th>
<th>Latitude (°)</th>
<th>Central Age (Ma)</th>
<th>Age Dispersion (Ma)</th>
<th>Pχ²%</th>
<th>N_d</th>
<th>ρ_d (tracks)</th>
<th>ρ_i (tracks)</th>
<th>N_s (crystals)</th>
</tr>
</thead>
<tbody>
<tr>
<td>RS10.06</td>
<td>-68.30111</td>
<td>-27.613889</td>
<td>20.62</td>
<td>±1.81</td>
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<td>1</td>
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Table 3: AHe data
Figure 16 – Cross-section with thermochronologic data included. The youngest AHe ages for each sample are emphasized in bold because in partially reset samples the youngest age is the closest to the true exhumation age.
Geochronology

Zircon U-Pb analysis was completed on five igneous samples from the ranges bounding the Fiambalá Basin to the west, north, and east. U-Pb analysis was also completed on four detrital samples from the upper Miocene strata in the Fiambalá Basin. Provenance of the basin strata can thus be compared to the sources in the surrounding ranges.

Figure 17a is a satellite image with both igneous and detrital sample locations marked, and Figure 17b is a stratigraphic column with detrital sample locations marked. Sample locations that are within the mapped field area are also indicated in the geologic map of the field area (Figure 12). Data from the igneous samples are plotted on concordia diagrams in Figures 18-20 and 22-24. Data from the detrital samples are plotted on concordia diagrams (Figures 25, 27, 29 and 31) and probability distribution diagrams (Figures 26, 28, 30 and 32).
Figure 17a – A topographic map showing the locations of samples for zircon U-Pb analysis.

Igneous samples are blue and detrital samples are yellow. Samples that are within the mapped field area are also indicated on the geologic map (Fig. 12).
Figure 17b – A stratigraphic column of upper Miocene strata in the Fiambalá Basin with the locations of detrital zircon U-Pb samples indicated, modified from Carrapa et al. (2006).
Igneous Samples

RS10.01

Sample RS10.01 was collected from Ordovician granite in the Sierra de Las Planchadas. 23 analyses were performed on this sample. The sample has an age of 491.8 ± 5.9 Ma (Fig. 18). Ages are concordant, indicating a simple magmatic history for these zircons.

Figure 18 – Zircon U-Pb data for sample RS10.01
RS10.02

Sample RS10.02 was collected from a Miocene intrusion in the Valle de Chaschuil, west of the Sierra de Las Planchadas. 17 analyses were performed on this sample. The sample consists of young zircons with a concordant age of 14.0 ± 1.3 Ma (Fig. 19). A few of the zircons analyzed also contained inherited cores of Paleozoic and Proterozoic age.

Figure 19 – zircon U-Pb data for sample RS10.02
Sample RS10.10

Sample RS10.10 was collected from Ordovician granite in the Sierra de Las Planchadas. 22 analyses were performed on this sample. This sample has an age of $484.6 \pm 5.3$ Ma (Fig. 20). The ages are concordant, indicating a simple magmatic history.

Figure 20 – Zircon U-Pb data for sample RS10.10
Sample CN046 was collected from Cambrian-Precambrian granites near Cerro Negro on the northeastern flank of the Fiambalá basin. 39 analyses were performed on this sample. The sample consists of zircons with two domains: older cores surrounded by younger rims. Figure 21 is a CL image of several zircons from this sample in which the two domains are visible. The rims have a concordant age of 524.1 ± 5.8 Ma and the cores are older, with ages ranging from 644 – 1885 Ma (Fig. 22). Many of these ages are discordant, suggesting some lead loss and/or mixed ages. A cluster of ages at ~1000 Ma are likely inherited cores derived from the 1250-980 Ma Sunsas (Grenville) orogeny.

Figure 23 is a plot of the rim ages and the most discordant core ages. A trend line fitted to these data has a lower intercept of 534 ± 15 Ma, which is very close to the age of the rims, suggesting that the discordant trend is a result of mixed ages between the core and rim domains. The upper intercept of the trend line occurs at 1861 ± 30 Ma, which is interpreted as the age of the oldest cores.

Figure 21 – Zircons from sample CN046. Many of these zircons have clearly defined core and rim domains.
Figure 22 – zircon U-Pb data for sample CN046
Figure 23 – Plot of the rim ages and the most discordant core ages. A line fitted to these data has a lower intercept of 534 ± 15 Ma and an upper intercept of 1861 ± 30 Ma.
Sample AG161 was collected from Carboniferous leucogranites near Alto Grande, east of the Fiambalá Basin. 30 analyses were performed on this sample. With a few exceptions, ages are concordant, and the sample age is 322.9 ± 3.2 Ma (Fig. 24). This sample appears to have had a simple magmatic history.

Figure 24 – zircon U-Pb data for sample AG161
Detrital Samples

DZ003 – Lower Tambería Formation

Sample DZ003 is from the Lower Tambería Formation in the Fiambalá Basin. Significant peaks in the age distribution occur at ~1000 Ma, ~500 Ma and ~20 Ma (Fig 26). This distribution of ages suggests provenance from the Proterozoic basement to the east of the basin, the Ordovician intrusions to the west of the basin, and a component of early Miocene zircons associated with Andean volcanism. A small ~300 Ma component derived from the Carboniferous intrusions east of the basin is also present.

![Detrital zircon U-Pb data from the Lower Tambería Formation](image)

*Figure 25 – detrital zircon U-Pb data from the Lower Tambería Formation*
Figure 26 – Probability distribution plot of detrital zircon U-Pb data from the Lower Tambería Formation
DZ050 – Upper Tambería Formation

Sample DZ050 was collected from the Upper Tambería Formation in the Fiambalá Basin. The distribution of ages (Fig. 28) is similar to sample DZ003, with provenance from the Proterozoic basement east and northeast of the basin, the Ordovician intrusions to the west of the basin, and a component of early Miocene zircons associated with Andean volcanism. In addition, however, there is a significant component derived from the Carboniferous intrusions east of the basin. The Miocene component for this sample is not as large as in the Lower Tambería sample.

*Figure 27 – detrital zircon U-Pb data from the Upper Tambería Formation*
Figure 28 – probability distribution plot of detrital zircon U-Pb data from the Upper Tambería Formation.
DZ053 – Guanchín Formation

Sample DZ053 was collected from the Guanchín Formation in the Fiambalá Basin. The distribution of ages (Fig. 30) is similar to that of the Tambería Formation and suggests provenance from the Proterozoic basement to the east and northeast of the basin, the Carboniferous intrusions east of the basin, the Ordovician intrusions west of the basin, and late Miocene zircons associated with Andean volcanism. The Miocene component for this sample is younger than in the Tambería, with the youngest zircons at ~5-8 Ma. The Tambería Formation contained only early Miocene zircons.

Figure 29 – detrital zircon U-Pb data from the Guanchín Formation
Figure 30 – probability distribution plot of detrital zircon data from the Guanchín Formation
DZ177 – Punaschotter Formation

Sample DZ177 was collected from the Punaschotter Formation in the Fiambalá Basin. Peaks in the age distribution (Fig. 32) occur at similar ages to those observed in the Tambería and Guanchín Formations. However, the provenance for this sample is distinct from the Tambería and Guanchín Formations in that there is a much larger component of Miocene zircons associated with Andean volcanism. There is even a Pliocene component, with the youngest zircons at ~3 Ma. Other sources present are similar to the previous samples: the Proterozoic basement east and northeast of the basin, the Ordovician intrusions west of the basin, and the Carboniferous intrusions east of the basin.

Figure 31 – detrital zircon U-Pb data from the Punaschotter Formation
Trends in the Cenozoic populations

The youngest populations of Cenozoic Andean-derived zircons show an up-section younging trend. In the Lower Tambería the youngest zircon population is early Miocene. In the Upper Tambería the youngest population ranges from early to mid-Miocene. In the Guanchín Formation the youngest population is late Miocene, and in the Punaschotter the youngest grains range in age from late Miocene to Pliocene. Figure 33 is a comparison of the Cenozoic portion of the probability distribution plots for each sample. This younging trend indicates that local volcanism was occurring coevally with deposition in the Fiambalá Basin. The 14 Ma age obtained for sample RS10.02 confirms that magmatism was active during the Miocene.
Figure 33 – Cenozoic component of the detrital zircon samples
DISCUSSION

Structural data

The cross-section constructed in this study (Figure 13) shows a minimum of ~60 km of shortening in the Sierra de Las Planchadas. When combined with a cross-section completed through the ranges to the northeast (Allmendinger, 1986) this brings the total documented shortening at the latitude 27-28°S to ~80 km.

The Sierras Pampeanas ranges consist primarily of Laramide-style block uplifts, and thus there is generally little shortening in these ranges. All of the ranges to the east of the Sierra de Las Planchadas are block uplifts. The presence of thin-skinned style deformation in the Sierra de Las Planchadas is therefore significant in that it implies that undocumented thin-skinned shortening may also exist in the Principal Cordillera to the west. Unfortunately, much of the Principal Cordillera has been covered by Neogene volcanics, obscuring structural relationships.

The faulting and folding of sedimentary rocks within the Fiambalá Basin, involving rocks as young as the Pliocene Punaschotter Formation, indicates that active deformation persisted in the basin until recent geologic time. The young exhumation ages obtained in this study (~20 Ma AFT, <10 Ma AHe) also suggest that deformation within the Sierra de Las Planchadas is recent.

In addition, the syntectonic deposits in the Fiambalá Basin are all young, with the base of the Lower Tambería Formation dated at ~9 Ma by interbedded tuff beds (Carrapa et al., 2008). This is in stark contrast to the basins north of Fiambalá, where an entire >5 km sequence of foreland basin deposits dating back to the Paleocene is typically present (Coutand et al., 2001; DeCelles et al., 2011; Siks and Horton, 2011). South of Fiambalá, in the basins adjacent to the Sierra de Famatina, lower Miocene sedimentary rocks rest unconformably on Triassic sedimentary rocks (Davila and Astini, 2007). The lack of any sedimentary record prior to the late
Miocene in the Fiambalá Basin suggests that either earlier foreland sedimentary rocks were never deposited, or that the entire Paleogene section has been removed from the basin. The partially reset AFT and AHe ages suggest that the Paleozoic section was never buried deeper than 2.5-4.5 km. This supports the first hypothesis, that a full foreland basin sequence was never deposited in the Fiambalá Basin prior to the late Miocene. If such sedimentary rocks were deposited, they must have been less than 2.5-4.5 km thick and must have been removed before the Miocene sedimentary rocks were deposited.

**Thermochronologic data**

Some of the AHe ages appear to be partially reset, with significant age dispersion among grains within a sample. In the case of partial resetting, the youngest grains are the most reset and provide an upper constraint on the cooling age; thus only the youngest grain ages are considered in this discussion. In most samples the youngest grain ages fall into the 2-10 Ma range. The AHe ages suggest that samples were being actively exhumed during the Miocene. Exhumation is here interpreted to be a response to active tectonic deformation within the Sierra de Las Planchadas during the Miocene. This correlates well with the presence of Miocene syntectonic deposits in the Fiambalá basin. Sample RS10.18 from the Upper Tambería Formation has an AHe age of ~2.3 Ma, suggesting active deformation within the Fiambalá Basin as recently as the Pliocene; this observation is consistent with evidence cited above for post-Punaschotter Formation faulting.

The AHe ages show a general younging trend of 10-2 Ma moving from west to east across the Sierra de Las Planchadas and Fiambalá Basin. This may be an expression of the local eastward migration of exhumation and uplift. The one exception to this trend is the 4.2 ± 0.2 Ma
age obtained for sample RS10.10, which is east of the 7.0 ± 0.3 Ma age obtained for sample RS10.15. The 4.2 ± 0.2 Ma age may be an expression of out of sequence thrusting, perhaps in response to subcritical taper conditions in the thrust belt. It is also possible that the 7.0 ± 0.3 Ma age is not a fully reset exhumation age, as sample RS10.15 shows significant age dispersion among the grains. The AHe ages obtained in this study are consistent with ages in the Eastern Cordillera at ~26°S, where Carrapa et al. (2011b) found a west-to-east younging trend in AHe ages ranging from 14-3 Ma. If the two regions are correlated along strike, this indicates that crustal deformation in the Principal Cordillera at 28°S occurred synchronously with deformation in the Eastern Cordillera at 26°S and that the Eastern Cordillera and Principal Cordillera may be a single continuous thrust front (Figure 34).
Figure 34 - A summary of estimates of the timing and rate of the forelandward propagation of the Andean thrust front along strike, as compiled by Carrapa et al. (2011b). AHe data from this study (light blue) show a 10-2 Ma younging trend across the Sierra de Las Planchadas. This is consistent with a 14-3 Ma younging trend in AHe ages across the Eastern Cordillera found by Carrapa et al., (2011b) (yellow), suggesting that deformation in the Principal Cordillera at 28 °S is synchronous with deformation in the Eastern Cordillera at 26 °S. This implies a single continuous thrust front between the Eastern Cordillera and the Principal Cordillera (dashed red lines).
No similar younging trend in exhumation is observed in the AFT ages. Three samples have fully reset AFT ages that pass the $\chi^2$ test: samples RS10.06 and RS10.10 both have cooling ages of ~20 Ma, and sample RS10.14 has an age of ~160 Ma. The other three samples do not pass $\chi^2$ and are assumed to be partially reset. The reset AFT ages are older than the AHe ages, which is expected since AFT is a slightly higher temperature thermochronometer. The two fully reset ~20 Ma ages corroborate the interpretation from the AHe ages that exhumation was actively occurring during the Miocene.

When compared to regional AFT ages compiled by other studies (Fig. 35), the Miocene AFT ages determined in this study are younger than expected. AFT ages to the east of the Fiambalá Basin in the Sierra de Chango Real are Eocene-Oligocene in age (Coutand et al., 2001). Farther to the east in the Sierra de Aconquija AFT ages are late Cretaceous (Coughlin et al., 1998) and late Miocene-Pliocene (Sobel and Strecker, 2003). Thus there is no apparent west to east younging trend recorded by the AFT ages at this latitude. Coughlin et al. (1998) suggested that in the Sierras Pampeanas exhumation and deformation did not progress smoothly from west to east, but rather occurred out of sequence or possibly even backwards from east to west. Since the Sierra de Las Planchadas is east of the Sierras Pampeanas ranges, the young AFT ages reported in this study support the hypothesis that the AFT system does not have a clear younging trend from west to east at this latitude.
Figure 35 – A summary of AFT ages (Ma) in the northern Sierras Pampeanas and southern Eastern Cordillera. Ages are from Coughlin et al., 1998 (green); Coutand et al., 2001 (red); Mortimer et al., 2007 (purple); Sobel and Strecker, 2003 (brown); Carrapa et al., 2006 (yellow), Carrapa et al., 2011a (dark blue); and this study (light blue).

However, it is important to note that since the AFT system is sensitive to higher temperatures than the AHe system, it is possible that not enough erosion has occurred in this region to exhume fully reset AFT ages. Furthermore, paleotopography can have a significant effect on exhumation ages; samples exhumed from inverted rift basins may have experienced
burial to a great enough depth to reset the AFT age, whereas samples from the flanks of these rift basins may not be reset (Carrapa et al., 2011a, in revision). Sedimentologic evidence suggests that the Sierras Pampeanas were uplifted at ~6 Ma (Carrapa et al., 2008). This is not consistent with the old AFT ages found by Coutand et al. (2001) and Coughlin et al. (1998) and suggests that the AFT system does not record the uplift of the Sierras Pampeanas. The presence of a younging trend in the AHe ages and lack of any clear trend in the AFT ages suggests that AHe is a better suited thermochronometer for recording Cenozoic exhumation in this part of the southern central Andes.

**Geochronologic data**

The four detrital zircon samples from the Miocene strata show little variation in provenance. All four samples contain peaks of Grenville, Cambrian, Ordovician, and Carboniferous ages, as well as Neogene zircons associated with Andean volcanism. The Cambrian, Ordovician, and Carboniferous samples match well with the igneous samples that we dated from the ranges surrounding the Fiambalá Basin, indicating local sources. The Grenville signal likely comes from older inherited cores within the igneous zircons as well as zircons reworked from Paleozoic and Proterozoic strata. The youngest populations of Cenozoic Andean-derived zircons show an up-section younging trend, indicating that local volcanism was occurring coevally with deposition in the Fiambalá Basin.
CONCLUSIONS

The purpose of this study was to investigate the extent, style, and timing of crustal deformation at the transition between normal steep subduction and flat subduction at latitude 28°S. The cross-section constructed across the Sierra de Las Planchadas and Fiambalá Basin predicts a minimum of ~60 km of shortening, bringing the total documented shortening at this latitude to ~ 80 km. This is significantly less shortening than is observed in the Eastern Cordillera to the north. Although the documented shortening at this latitude is still insufficient to explain the crustal thickness, the presence of thin-skinned style deformation suggests that further undocumented shortening may exist in the Principal Cordillera to the west. East of the Fiambalá Basin, shortening is dominated by Laramide-style block uplifts.

AHe ages show a younging trend in exhumation from west to east, presumably indicative of the forelandward progression of uplift and deformation. The propagation of exhumation in the Principal Cordillera at this latitude appears to be synchronous with the Eastern Cordillera. No clear trend is present in the AFT ages at this latitude, and indeed the AFT ages obtained in this study are younger than those observed in the Sierras Pampeanas ranges to the east. This suggests that uplift and deformation may not have progressed simply from the Principal Cordillera eastward to the Sierras Pampeanas, but rather uplift of the Sierras Pampeanas ranges occurred synchronously with deformation in the Principal Cordillera during the late Miocene. However, the absence of a clear younging trend in the AFT ages may also indicate that the AFT system is not sensitive to low enough temperatures to systematically record Cenozoic exhumation in the central Andes.

This study provides further evidence for similarities and contrasts in the extent, style, and timing of crustal deformation between the regions of normal and flat-slab subduction. Although
contrasts in deformation style between the two regions may suggest that the slab angle is the primary control on the style of crustal deformation, it is important to consider that pre-existing crustal structures and architecture may play a key role. In particular, thin skinned deformation is most often prevalent where a thick sedimentary package exists which can be deformed, whereas block uplift often occurs where rifts can be reactivated as high-angle reverse faults. The existence of thin-skinned deformation in the Sierra de Las Planchadas juxtaposed with block-uplift deformation in the ranges to the east at the same latitude suggests that pre-existing structures played a more important role than the angle of the slab.
APPENDIX

A - Regional geologic map

I created a regional scale geologic map at the 1:1000000 scale by stitching together provincial maps published by SEGEMAR. A JPG file of this map is included on the attached CD-ROM.

B - Regional topographic maps

I created nine topographic maps of northwestern Argentina by generating 50 m contours from SRTM data and overlaying them on satellite images. The maps cover the Argentine Andes between latitudes 24-30 °S. These maps are numbered 1-9 and are included on the attached CD-ROM as jpeg files. Each map image has dimensions 40” x 40” and a resolution of 200 dpi. The full maps can be printed on a plotter, or sections of the maps can be enlarged and printed for field mapping. Figure 36 is an index map of the locations of the topographic maps.

Figure 36 – An index map of the topographic maps included on the attached CD-ROM.
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REFERENCES


Interpretations, and Applications, Volume 58: Reviews in Mineralogy and Geochemistry, Mineralogical Society of America.


Siks, B., and Horton, B.K., 2011, Growth and fragmentation of the Andean foreland basin during eastward advance of fold-thrust deformation, Puna plateau and Eastern Cordillera, northern Argentina: Tectonics, v. in press.


