Using basin thermal history to evaluate the role of Miocene–Pliocene flat-slab subduction in the southern Central Andes (27° S–30° S)

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ABSTRACT

Studies in both modern and ancient Cordilleran-type orogenic systems suggest that processes associated with flat-slab subduction control the geological and thermal history of the upper plate; however, these effects prove difficult to deconvolve from processes associated with normal subduction in an active orogenic system. We present new geochronological and thermochronological data from four depositional areas in the western Sierras Pampeanas above the Central Andean flat-slab subduction zone between 27° S and 30° S evaluating the spatial and temporal thermal conditions of the Miocene–Pliocene foreland basin. Our results show that a relatively high late Miocene–early Pliocene geothermal gradient of 25–35 °C km⁻¹ was typical of this region. The absence of along-strike geothermal heterogeneities, as would be expected in the case of migrating flat-slab subduction, suggests that either the response of the upper plate to refrigeration may be delayed by several millions of years or that subduction occurred normally throughout this region through the late Miocene. Exhumation of the foreland basin occurred nearly synchronously along strike from 27 to 30° S between ca. 7 Ma and 4 Ma. We propose that coincident flat-slab subduction facilitated this widespread exhumation event. Flexural modelling coupled with geohistory analysis show that dynamic subsidence and/or uplift associated with flat-slab subduction is not required to explain the unique deep and narrow geometry of the foreland basin in the region implying that dynamic processes were a minor component in the creation of accommodation space during Miocene–Pliocene deposition.

INTRODUCTION

In Cordilleran-type orogenic systems, the interaction between shortening, crustal thickening and magmatism during subduction processes produces predictable patterns in the geology of the overriding plate (Decelles et al., 2009, 2015). During periods of flat-slab subduction coupling between the lower and upper plate change these patterns (Coney & Reynolds, 1977; Jordan & Allmendinger, 1986; Dumitru et al., 1991; Ramos & Folguera, 2009); however, the effects of flat-slab subduction on the organization, thermal regime and subsidence patterns of the sedimentary basins overlying flat-slab subduction zones are poorly understood. In particular, it is difficult to distinguish between changes in surface geology related to normal – albeit commonly episodic – orogenic processes (Decelles & Graham, 2015) vs. dynamic effects associated with flat-slab subduction (Dávila et al., 2010; Liu et al., 2010; Painter & Carrapa, 2013; Dávila & Lithgow-Bertelloni, 2015; Heller & Liu, 2016; Hu et al., 2016).

The central Andean flat-slab subduction zone of the Sierras Pampeanas is located in a region of well-preserved Cenozoic sedimentary basins (Fig. 1). Present-day flat-slab subduction can be observed in the Sierras Pampeanas region using geophysical techniques (Fig. 1a) (Cahill & Isacks, 1992; Anderson et al., 2007; Gans et al., 2011; Ammirati et al., 2016); however, constraining the timing of slab flattening proves more challenging and temporal estimates of flat-slab initiation range from early Miocene to Pliocene based on the timing of arc magmatism, basin subsidence and reorganization and exhumation (Reynolds et al., 1990; Jordan et al., 2001; Ramos et al., 2002; Dávila & Astini, 2007; Carrapa et al., 2008; Kay & Coira, 2009; Dávila & Carter, 2013; Fosdick et al., 2015). The availability of geophysical data coupled with well-preserved surface geology provides an opportunity to understand the relationship between flat-slab subduction and sedimentary basin evolution not only in the Central Andes but also in other flat-slab subduction zones around the world. Understanding the spatial and temporal thermal regime in sedimentary basins in the Sierras
Pampeanas region – and specifically understanding if it is controlled by dynamic processes associated with subduction geometry – may be the lynchpin for more sufficiently distinguishing which changes in upper plate geology can be attributed to flat-slab subduction.

In the Sierras Pampeanas (Fig. 1), changes in thermal conditions and subsidence patterns have been tied to increase coupling between the upper and lower plate during flat-slab subduction (Gutscher et al., 2000) which is hypothesized to affect deformation in the fold and thrust belt (Johnson et al., 1986), drive dynamic subsidence and/or uplift in the basin (Dávila et al., 2010; Dávila & Lithgow-Bertelloni, 2015) and refrigerate the overriding plate (Collo et al., 2011, 2017; Dávila & Carter, 2013; Dávila & Lithgow-Bertelloni, 2015).

In particular, slab refrigeration is based on the assumption that removal of a hot asthenospheric wedge during periods of flat-slab subduction produces a uniquely low geothermal gradient (ca. 10–20 °C km⁻¹) above flat-slab subduction zones (Dumitru, 1990; Dumitru et al., 1991). This effect may be enhanced by the cold, flat subducting slab during the coupling between the upper and lower plates during flat-slab subduction (Collo et al., 2011, 2017; Dávila & Carter, 2013; Dávila & Lithgow-Bertelloni, 2015).

Low-temperature thermochronology is an effective tool to constrain the thermal history of sedimentary basins which in turn provides important insight into the tectonic history of a region. When applied to sedimentary rocks, low-temperature thermochronometers can reflect the cooling history of the sediment source regions or cooling and basin exhumation following sediment burial heating. An elevated geothermal gradient coupled with significant
burial can be sufficient to reset low-temperature thermochronometers such as apatite (U–Th–Sm)/He and fission track thermochronometers which have closure temperatures of 55–80 °C and 80–120 °C respectively (Green et al., 1989; Farley, 2000; Reiners & Brandon, 2006). Post-burial cooling can thus be used to constrain the magnitude of burial, the palaeogeothermal gradient and the timing of basin exhumation.

The goal of this study was to constrain the spatial and temporal thermal history of the Mio–Pliocene foreland basin (Fig. 1) located above the flat-slab subduction zone using the (U–Th–Sm)/He (AHe) and fission track systems (AFT) on apatite and to evaluate if a low geothermal gradient can be used as evidence for close coupling between a flat subducting slab and the overriding plate (Dumitru, 1990; Collo et al., 2011, 2017; Dávila & Carter, 2013; Dávila & Lithgow-Bertelloni, 2015). We combine these results with new and existing flexural modelling to evaluate if the proposed dynamic effects of flat-slab subduction (Dávila & Lithgow-Bertelloni, 2015; Hu & Liu, 2016) are required during basin development and/or exhumation.

**GEOLOGIC BACKGROUND**

The Central Andes between 27°S and 30°S are divided into discrete tectonomorphic zones based on unique morphological characteristics and deformation styles (Fig. 1a) (Ramos, 2009). The Mio–Pliocene depositional areas in this study straddle the boundary between two of these tectonomorphic zones, the Precordillera fold and thrust belt (Zapata & Allmendinger, 1996a,b; Allmendinger & Judge, 2014; Fosdick et al., 2015) to the west and the bivergent, basement-cored uplifts of the Sierras Pampeanas (Jordan & Allmendinger, 1986) to the east (Fig. 1a). These depositional areas are interpreted to have once been part of a regional foreland basin, extending over 400 km along strike, before being compartmentalized during flat-slab subduction in the Miocene–Pliocene (Ramos, 1970; Beer & Jordan, 1989; Jordan et al., 2001; Reynolds et al., 2001; Carrapa et al., 2006, 2008; Ciccioli et al., 2011).

The depositional areas in this study are located along the boundary of an Ordovician suture zone where the Famatinian arc terrane defines the eastern boundary and the Laurentian-affinity Cuyanian terrane defines the western boundary (Thomas & Astini, 1996; Ramos, 2010) (Fig. 1). The complex tectonic history of the region has affected the flexural rigidity of the crust (Cardozo & Jordan, 2001; Pérez-Gussinyé et al., 2007, 2008; Tassara et al., 2007) and flexural modelling by Cardozo & Jordan (2001) finds that variable effective elastic thickness best explains the geometry of Miocene sedimentary basins between 32° S and 29° S.

The initiation of foreland basin sedimentation in the Eastern Cordillera to the north (ca. 27°S) of the study region started in the Palaeogene (Decelles et al., 2011; Carrapa et al., 2012) and to the south near the Huaco area in the Bermejo foreland ca 30°S (Fig. 1) as early as the Eocene (Fosdick et al., 2017). This study focuses on six separate depositional areas (Figs 1 and 2) located along a 350-km stretch above the flat-slab subduction zone in the San Juan, La Rioja and Catamarca Provinces. From north to south these depositional areas are the Guanchin River and Rio La Troya depositional areas in the Fiambalá basin, the Valle Hermoso and Vinchina depositional areas in the Vinchina basin, and the La Flecha, and Huaco depositional areas in the northern Bermejo basin (Fig. 1). The stratigraphic nomenclature of these successions varies geographically and a chronometric summary across localities is detailed in Fig. 2. Although local heterogeneities are present, all of the depositional areas in this study are dominated by fluviol-fluvial coarsening upward sequences with limited lacustrine strata (Fig. 2). Constraints on the depositional age of fluviol-fluvial strata in each depositional area vary along strike; however, these variations also reflect disparities in geochronometric sampling (Fig. 2). Strata in the Huaco area are constrained to depositional ages between ca. 14 Ma and 2 Ma (Johnson et al., 1986; Beer, 1990), in the La Flecha depositional area between ca. 18 Ma and 12 Ma (Reynolds et al., 1990), in the Vinchina area between ca. 19 Ma and 2 Ma (Dávila et al., 2008; Ciccioli et al., 2014; Amidon et al., 2016; Collo et al., 2017), and in Fiambalá depositional area between ca. 9 Ma and 3 Ma (Carrapa et al., 2006, 2008) (Figs 1 and 2). The Valle Hermoso depositional area currently contains no geochronologic constraints (Fig. 2). Aeolian deposits of the ca. 33–23 Ma Vallecito Formation (Tripaldi & Limarino, 2005; Ciccioli et al., 2014; Fosdick et al., 2017) have been identified throughout the study area below the fluviol-fluvial succession (Fig. 2) and are identified here in the Valle Hermoso depositional area (Fig. 1).

The onset of Neogene flat-slab subduction in the central Andes (Fig. 1) is attributed to subduction of the Juan Fernandez Ridge (Gutscher et al., 2000), a ca. 40- to 60-km-wide hot spot trace with buoyant, over-thickened oceanic crust. Although presently subducting east-northeast (von Huene et al., 1997), plate reconstructions and palaeomagnetic records indicate that Juan Fernandez Ridge has a northeast-oriented dog leg geometry that caused the locus of ridge subduction to propagate from north to south in the Miocene (Pilger, 1981, 1984; Yáñez et al., 2001). Assuming coupling between lower and upper plate, the corollary of this reconstruction requires that the manifestation of flat-slab processes preserved in the surface geology should also migrate north to south.

In this study, we use the term flat-slab subduction to refer to areas where the subducting slab is horizontal. We
differentiate a horizontal flat slab from subduction angles that are subhorizontal, but more shallow than 30°, the average dip of the Nazca plate along the Andean subduction zone. Subhorizontal subduction is typically observed along the margins of a horizontal segment. In geologic time, subhorizontal subduction may occur during periods of incipient or waning flat-slab subduction. Temporal estimates of the initiation of flat-slab subduction in the southern Central Andes range from ca. 17 to 15 Ma (Dávila et al., 2004; Dávila, 2010) to ca. 12–10 Ma (Reynolds et al., 1990; Kay & Abbruzzi, 1996; Kay & Mpodozis, 2002), and ca. 6–5 Ma (Carrapa et al., 2008; Fosdick et al., 2015). Today, the subducting slab is flat between ca. 33° S and 32° S (Fig. 1); however, the depth to slab remains anomalously shallow (subhorizontal) between latitudes of approximately 33–27° S (Fig. 1) (Cahill & Isacks, 1992; Anderson et al., 2007; Gans et al., 2011).

Constraining the timing of flat-slab subduction is critical for understanding mechanisms driving basin subsidence in the mid Miocene–early Pliocene (Dávila et al., 2010; Dávila & Lithgow-Bertelloni, 2015; Hu & Liu, 2016; Hu et al., 2016) as well as for reconstructing the mechanisms driving flat-slab subduction (Schepers et al., 2017). Hu et al. (2016) suggested that dynamic suction...
forces drive coupling between the upper and lower plate during flat-slab subduction. Dávila & Lithgow-Bertelloni (2015) suggested that flat-slab subduction may cause dynamic uplift. Evidence for the onset of dynamic mechanisms is based on an extremely low (ca. 15 °C km⁻¹) geothermal gradient, attributed to refrigeration of the upper plate by the cold, shallowly subducting oceanic slab (Collo et al., 2011, 2017; Dávila & Carter, 2013; Dávila & Lithgow-Bertelloni, 2015).

Previous work on basin thermal history

The modern geothermal gradient can be estimated using temperatures measured in boreholes. However, this method, when applied to wells drilled using water and/or mud typically underestimates true rock temperatures by ca. 10–15 °C, thus geothermal gradients calculated using borehole temperatures are only minimum estimates (Deming, 1989). The modern geothermal gradient measured near the Guandacol borehole just southeast of the La Flecha depositional area, drilled using water, (Fig. 1b) yields an uncorrected (minimum) geothermal gradient of ca. 24 ± 5 °C km⁻¹ (F. Fuentes, personal communication). Slightly lower geothermal gradients have been calculated by Collo et al. (2017) from borehole temperatures farther south in the Pozuelos and Mataguasenos boreholes (Fig. 1b) which have uncorrected (minimum) geothermal gradients of ca. 21 and 18 °C km⁻¹. Drilling conditions are not described for these boreholes by Collo et al. (2017), but results from other studies suggest errors are likely ca. ± 4 °C for these boreholes (Deming, 1989).

The transformation from smectite to illite, a kinetically controlled process that occurs at temperatures between 70 and 250 °C (Perry, 1970; Pytte & Reynolds, 1989; Huang, 1993) has been used to constrain maximum burial temperatures in the Vinchina, La Flecha and Huaco depositional areas (Collo et al., 2009, 2011, 2017). The degree of thermal maturity of strata as deep as 7 km leads Collo et al. (2011, 2017) to suggest a geothermal gradient of ca. 15°C km⁻¹. Only one thermochronometer, AHe, was conducted in the Vinchina depositional area to test such a hypothesis (Fig. 2) (Collo et al., 2017). Samples taken at depths of ca. 5500 m contain single grain cooling ages (6.9 ± 0.4 Ma, 3.4 ± 0.2 Ma) younger than the depositional age of the host strata (ca. 12 Ma) consistent with significant post-depositional thermal resetting. AHe has also been applied to Miocene strata in the Huaco and La Flecha depositional areas (Fig. 2) (Fosdick et al., 2015; Collo et al., 2017). In the Huaco area, strata buried to depths of 4–5 km yield cooling ages younger than the depositional age, whereas strata with burial depths of ca. 2 km produce partially reset cooling ages suggesting that the sample spent significant time in the partial retention zone (PRZ) (see methods for explanation of PRZ) (Fosdick et al., 2015). Collo et al. (2017) reported that in the La Flecha depositional area, a sample buried to a depth of ca. 4.5 km has been fully reset (Farley, 2000).

One tuff sample in the Vinchina Formation was analysed by Tabbutt et al. (1989) using zircon fission track (ZFT) thermochronology, a system that is reset at temperatures between 190 and 220 °C (Zaun & Wagner, 1985; Bernet, 2009). The ZFT sample, taken from a burial depth of ca. 8 km, yields a cooling age of 7.3 ± 1.3 Ma. The original purpose of this study was to constrain the maximum depositional age of the strata in the Vinchina depositional area (Tabbutt et al., 1989); however, subsequent work indicates that the depositional age of this strata is likely between 14 and 12 Ma (Ciccioli et al., 2014; Collo et al., 2017) introducing the possibility that instead of a depositional age, the ZFT sample actually records a partially reset age or cooling post deposition.

In the Fiambalá basin (ca. 28° S), the northermmost extent of the Bermejo basin (Carrapa et al., 2008; Safipour et al., 2015), a stratigraphic section logged along the Guanchín River (Figs 1 and 2) contains ca. 4.5 km of Miocene–Pliocene fluviolacustrine and alluvial strata. At this location, samples at the top of the ca. 9–5.5 Ma Tambería Formation (covered by ca. 2–2.8 km of section) were fully reset for AHe but not for AFT thermochronology (Carrapa et al., 2006; Safipour et al., 2015) suggesting that burial temperatures were greater than 80 °C, but less than 120 °C (Green et al., 1989; Farley, 2000). Tabbutt et al. (1989) also analysed a sample from near the base of the Tambería Formation for ZFT which yielded a cooling age of 5.7 ± 0.8 Ma, consistent with thermal resetting.

METHODS

New samples of medium-grained sandstone were collected from four Miocene depositional areas above the flat-slab subduction zone in the La Flecha, Vinchina, Valle Hermoso and Rio La Troya areas (Figs 1 and 2) for detrital zircon geochronology and low-temperature thermochronology. In the La Flecha and Rio La Troya depositional areas, the stratigraphic age of samples was constrained by existing geochronologic constraints from measured sections in the area (Fig. 2). Chronostratigraphy in the La Flecha depositional area is constrained by palaeomagnetic data from the Las Juntas section from Reynolds et al. (1990), and in the Rio La Troya depositional area by geochronological and thermochronological analyses in the Fiambalá basin from Carrapa et al. (2006, 2008) and Safipour et al. (2015). In the Vinchina depositional area, we augment existing geochronological constraints (Ciccioli et al., 2014; Amidon et al., 2016; Collo et al., 2017) to better determine the depositional age of the strata. In the Valle Hermoso depositional area, we use
detrital zircon U–Pb geochronology to provide the first stratigraphic controls on the age of this section.

**U–Pb geochronology**

Where the depositional age of the stratigraphic section was poorly constrained, samples were collected for U–Pb geochronology on detrital zircons. This method relies on the presence of volanoogenic material within sandstone derived in part from the active Andean arc, to provide a constraint on the maximum depositional age of sedimentary strata (Decelles et al., 2007; Dickinson & Gehrels, 2009).

We analysed five stratigraphic intervals (Fig. 2) for detrital zircon geochronology in the Vinchina depositional area (Fig. 2). Four samples are from the Vinchina Formation (VN15-1-01, VN8-01, VN3-2020 and VN3-4032) and one sample is from the Toro Negro Formation (VN6-287). We analysed one sample from the Vallecito Formation in the Valle Hermoso depositional area (VH-01) (Fig. 2). Between 100 and 300 zircon grains were randomly analysed from each sample at the Arizona Laser-Chron Center using laser ablation on an Element2 HR ICPMS and a Nu HR ICPMS. Analytical methods are described in the supplementary information. Maximum depositional ages were calculated using the youngest age group in the detrital zircon samples using a refined methodology from Dickinson & Gehrels (2009) described in the supplementary information.

**Low-temperature thermochronology**

Low-temperature thermochronology is an effective tool to evaluate the thermal history of sedimentary basins (Naeser et al., 1989; Armstrong & Chapman, 1998; Armstrong, 2005). Thermochronometers reset after deposition in the basin should record a cooling age younger than the depositional age of the hosting strata. Alternatively, thermochronometers with cooling ages older than the depositional age of the sample reflect the thermal history of the mineral’s provenance source. Resetting thermochronometers is a function of both maximum temperature and the length of exposure time to temperatures higher than the closure temperature of a given thermochronometer (Dodson, 1973), and thus in sedimentary systems, constraining basin thermal history is critical to interpreting thermochronological data. The primary interest of this study was to determine the heating (burial) and cooling (exhumation) history of basin strata and not to derive the thermal history of detrital minerals source region; consequently, this study does not attempt to interpret the thermal history of minerals unreset by basin heating.

Our sampling strategy targeted sandstone originally buried to a minimum of 2,000 m in intervals between 500 and 2,000 m to evaluate thermal conditions at different stratigraphic depths (Fig. 2). In the La Flecha depositional area, we applied AHe thermochronology to two samples including one from the La Flecha Member (RLF-25) and one from the La Cay6 Member (RLF2-610) of the Vinchina Formation (Fig. 2). In the Vinchina depositional area, we analysed two samples from the Vinchina Formation for AHe thermochronology (VN3-2020 and VN3-4032) and three samples for AFT thermochronology (VN3-2020, VN3-4032 and VN4-704) (Fig. 2). In the Valle Hermoso depositional area, two samples were analysed for AHe thermochronology (Fig. 2) including one sample from the Vallecito Formation (VH-01) and one sample from the Vinchina Formation (VH2-365). Three samples were analysed in the Valle Hermoso depositional area for AFT thermochronology including VH-01 from the Vallecito Formation and VH-612 and VH2-365 from the Vinchina Formation (Fig. 2). In the Rio La Troya depositional area, samples were collected from the Tambelia Formation (Fig. 2). Sample TR1-10 was analysed for AHe thermochronology and samples TR1-10 and TR1-813 were analysed for AFT thermochronology (Fig. 2).

**Apatite fission track**

Fission track thermochronology utilizes the crystallographic imperfections created in a mineral during the spontaneous fission of $^{238}\text{U}$ (Price & Walker, 1963). In an apatite grain, spontaneous fission tracks are preserved at temperatures below 80–120 °C (Green et al., 1989); however, in the temperature range between 80 and 120 °C, known as the partial annealing zone (PAZ), fission tracks begin to anneal (Fleischer et al., 1975; Wagner et al., 1989). At temperatures above the PAZ, fission tracks are fully annealed. The number of spontaneous tracks relative to the original concentration of $^{238}\text{U}$ in the grain can be used to calculate the time when the mineral passed through the PAZ (Price & Walker, 1963). Samples were analysed for AFT at the University of Arizona using the external detector method (Hurford & Green, 1983). Where possible we counted between 25 and 50 grains per sample (Donelick, 2005). Detailed methodology can be found in the supplementary information.

**Apatite (U–Th–Sm)/He**

AHe thermochronology is based on the $\alpha$- particles emitted during the radioactive decay of the elements $^{238}\text{U}$, $^{235}\text{U}$, $^{232}\text{Th}$ and $^{147}\text{Sm}$ that occur naturally in minerals such as apatite and zircon (Farley et al., 1996; Ehlers & Farley, 2003). At high temperatures, $\alpha$- particles escape the crystal lattice of the mineral, whereas at low temperatures the $\alpha$- particles are retained. For apatite, the transition from $\alpha$- particle retention to loss occurs at temperatures between 55° and 80 °C (Farley et al., 1996; Farley, 2000), a temperature range known as the partial
retention zone (PRZ). We analysed five single apatite grains per sample from seven sandstones (locations described in Fig. 2). Analytical methods are described in the supplementary information.

**Thermal modelling**

Both the maximum temperature and duration of burial-induced heating control if a detrital thermochronometer is reset (Dodson, 1973). We use geological constraints including burial depth and depositional age of strata from this study (obtained by U–Pb geochronology) as well as chronostratigraphic constraints from other studies (Johnson et al., 1986; Reynolds et al., 1990; Carrapa et al., 2008; Ciccioli et al., 2014; Collo et al., 2017) to produce time-temperature histories that resolve both the magnitude and the duration of burial and subsequent cooling throughout the study area. These time-temperature histories can be evaluated using the forward modelling software, HeFTy (Ketcham et al., 2007), to test if they match the analytical AHe and AFT data. In forward modelling mode, HeFTy can predict the AHe and AFT cooling age of a sample for a given time-temperature history using the experimentally derived kinetic model for the used thermochronometric system. Three geothermal gradients, 15, 25 and 35 °C km⁻¹, were used to create forward models for a generic stratigraphic section that is similar to the stratigraphic sections in the depositional areas throughout the basin. The stratigraphic section is 8 km in depth and the sedimentation rate matches those observed in the depositional areas in this study (Jordan et al., 2001). Detailed forward model inputs are described in the supplementary information. Results from these models provide a quantitative evaluation of the stratigraphic levels where AFT and AHe systems should be reset given the unique time and temperature conditions of the depositional areas in this study.

**Flexural modelling**

We use a 2D flexural model to evaluate if flexural loading can account for the observed accommodation space of as high as 8.5 km, the thickest sedimentary package observed in the Mio-Pliocene depositional areas in this study. We apply the general equation for elastic flexure of the lithosphere from Turcotte & Schubert (2014) with a Young’s Modulus of 70 GPa and Poisson’s ratio of 0.25 (Turcotte & Schubert, 2014). We use a mantle density of 3300 kg m⁻³ with the infilling sediment density of 2700 kg m⁻³. Studies indicate that the topographic load in the Andean Precordillera by the end of the late Miocene was similar to the modern geometry (Val et al., 2016). For our model, we use a Precordilleran topographic load with an elevation of 2–2.5 km, and density of 2700 kg m⁻³, modelled using rectangular loads along a line-load width of 125 km adjacent to the basin. We argue that this represents a minimum load considering that modern elevations in the Precordillera are up to >4 km. We use the results from Cardozo & Jordan (2001) to anticipate variable effective elastic thickness which corresponds to variations in crustal strength associated with the Cuyania-Famatina terrane boundary (Thomas & Astini, 1996; Pérez-Gussinyé et al., 2007, 2008). Values of effective elastic thickness used in this study range from 20 to 5 km consistent with geophysical estimates (Tassara et al., 2007). We prescribe a 60-km-wide zone of relatively lower effective elastic thickness and evaluate the effect of this variable at four separate locations beneath both the depositional area and the topographic load. We use three model runs with combinations of effective elastic thickness values of 15 km and 5 km, 20 km and 10 km, and 20 and 5 km.

**Geohistory analysis**

Geohistory analysis, or 1D backstripping, quantifies the respective effects of tectonic forces and sedimentary and water loading on basin subsidence (van Hinte, 1978; Allen & Allen, 2009). The shape and magnitude of subsidence curves provide insight into the mechanism driving basin subsidence (e.g. Xie & Heller, 2009; Heller & Liu, 2016). We are specifically interested in distinguishing if subsidence in the investigated Miocene–Pliocene depositional areas reflects flexural subsidence or dynamic processes. This study follows the procedure described in Allen & Allen (2009) to conduct backstripping analysis on four stratigraphic sections including from the La Flecha and Vinchina depositional areas measured in this study. Input parameters and stratigraphic constraints are reported in the Supplemental Information.

**RESULTS**

**U–Pb geochronology**

In the Vinchina depositional area, the maximum depositional age of strata determined by zircon U–Pb geochronology consistently youngs upsection (Fig. 3). Sample VN15-1-01 located at a covered contact at the base of the section in the Vinchina Formation has a maximum depositional age constrained by three grains of 18.6 ± 0.4 Ma (Fig. 3). Intervals sampled from the Vinchina Formation at 1000 m (VN8-01), 3220 m (VN3-2020) and 5220 m (VN3-4032) above the base of the stratigraphic section yield maximum depositional ages of 13.8 ± 1.7 (n = 4), 9.2 ± 0.5 (n = 18) and 9.3 ± 0.4 (n = 4) respectively (Fig. 3). These results combined with existing geochronometers (Ciccioli et al., 2014; Collo et al., 2017) suggest that sedimentation rates in the Vinchina Formation reached ca. 1.0 mm yr⁻¹. Sample VN6-
287 sampled from the Toro Negro Formation has a maximum depositional age of $8.9 \pm 2.5$ Ma constrained by five grains (Fig. 3).

The distribution of detrital zircon ages from the Vinchina and Torro Negro formations (Fig. 4) is largely similar to distributions previously published by Ciccioli et al. (2014), Collo et al. (2017) and Amidon et al. (2016) with a range of Precambrain grains between ca. 1400 and 1000 Ma, major populations between 600 and 420 Ma, 380 and 250 Ma, and a Neogene population that decreased in age upsection.

In the Valle Hermoso depositional area, the maximum depositional age obtained from the Vallecito Formation at the base of the stratigraphic section constrained by 14 grains is $16.0 \pm 0.3$ Ma (Fig. 3). This is the youngest depositional age yet obtained from the Vallecito Formation (Jordan et al., 1993; Tripaldi & Limarino, 2005; Fosdick et al., 2017). The distribution of detrital zircon grains in VH-01 has significant populations between 1250 and 1000 Ma, 600 and 420 Ma, and 350 and 230 Ma (Fig. 4).

**Apatite fission track and (U-Th-Sm)/He thermochronology**

**La Flecha depositional area**

Two samples were analysed from the Vinchina Formation in the La Flecha depositional area (Fig. 2) for AHe thermochronology. Sample RLF-25 located 25 m from the base of the section (estimated depth, 5, 00 m) yielded four Mio-Pliocene cooling dates between $5.7 \pm 0.6$ Ma and $3.4 \pm 0.1$ Ma associated with eU values ranging from 6.5 to 156.2 ppm (Fig. 5, Table 1). One Oligocene cooling date of $27.0 \pm 0.53$ Ma has an eU value of 17.8 ppm (Fig. 5, Table 1). In contrast, sample RLF2-610 located 1570 m from the base of (estimated depth 3000 m) the stratigraphic section produced five dissimilar grain ages with cooling dates of $114.2 \pm 1.9$ Ma, $33.6 \pm 0.6$ Ma, $17.0 \pm 0.3$ Ma, $14.0 \pm 0.3$ Ma and $7.0 \pm 0.2$ Ma, with eU values ranging from 19.5 to 34.5 ppm (Fig. 5, Table 1).

**Vinchina depositional area**

Two intervals in the Vinchina Formation from the Vinchina section were sampled for AHe thermochronology (Fig. 2). Sample VN3-2020 located 3220 m from the base of the section (estimated depth of 5500 m) yields five cooling ages between $6.5 \pm 0.4$ Ma and $3.5 \pm 0.2$ Ma with eU values between 3.3 ppm and 2.1 ppm (Fig. 5, Table 1). Sample VN3-4032 yields four cooling dates between $6.3 \pm 0.2$ and $1.6 \pm 0.1$ Ma, with eU values between 13.5 and 4.8 ppm (Fig. 5, Table 1). VN3-4032 also yields one slightly older AHe of $11.6 \pm 0.3$ Ma with 69.6 ppm eU (Fig. 5, Table 1).
Three intervals from the Vinchina Formation were analysed for AFT thermochronology (Fig. 2). Sample VN3-2020 located 3,220 m from the base of the section (estimated depth of 5,500 m) (Fig. 2) yields an AFT cooling age of $8.9^{\pm}3.5$ Ma (Fig. 5, Table 2). Sample VN3-4032 located 5,220 m from the base of the section

![Fig. 4. Detrital zircon distributions for the six sandstone samples analysed for this study. The inset shows Phanerozoic detrital zircon distributions. Black bars are histograms and grey lines are kernel density estimates. Complete samples are plotted using bandwidth of 5 Ma and a bin width of 20 Ma and Phanerozoic samples are plotted using a bandwidth of 5 Ma and a bin width of 5 Ma.](image-url)

Three intervals from the Vinchina Formation were analysed for AFT thermochronology (Fig. 2). Sample VN3-2020 located 3,220 m from the base of the section
(estimated depth of 3,500 m) produces an AFT cooling age of 6.9 ± 1.5 Ma (Fig. 5, Table 2). The AFT cooling age from the shallowest sample located 5,750 m from the base of the section (estimated depth of 3,030 m), VN4-704, is 20.6 ± 2.6 Ma (Fig. 5, Table 2).

**Valle Hermoso depositional area**

Two intervals in the Valle Hermoso stratigraphic section from the Vallecito and the Vinchina formations were sampled for AHe thermochronology (Fig. 2). Sample VH-01 is located at the base of the stratigraphic section in the Vallecito Formation and sample VH2-365 is located at the top of the stratigraphic section, 1000 m above sample VH-01 in the Vinchina Formation (Fig. 2). The total burial depth at this location is unknown. Five apatite grains from sample VH-01 have AHe cooling dates ranging from 8.1 ± 1.0 Ma to 3.2 ± 0.1 Ma with eU values between 3.0 ppm and 9.0 ppm (Fig. 5, Table 1). In sample VH2-365, AHe cooling dates range in age from 9.6 ± 0.2 Ma to 3.5 ± 0.2 Ma (Fig. 5, Table 1). The eU of grains in sample VH2-365 range from 6.7 to 2.3 ppm.
Three samples were analysed in the Valle Hermoso locality for AFT thermochronology (Fig. 5, Table 2). In the Vallecito Formation, sample VH-01 produced an AFT cooling age of 6.8 ± 2.3 Ma (Fig. 5, Table 2). In the Vinchina Formation, sample VH-612 (located 612 m from the base of the stratigraphic interval) yielded a cooling age of 6.1 ± 1.4 Ma and VH2-365 (located at the top of the section) yielded a cooling age of 6.5 ± 1.5 Ma (Fig. 5, Table 2).

### Table 1. AHe Results

<table>
<thead>
<tr>
<th>Sample</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Diameter (µm)</th>
<th>U (ppm)</th>
<th>Th (ppm)</th>
<th>Sm (ppm)</th>
<th>eU (ppm)</th>
<th>Raw Age (Ma)</th>
<th>Corrected Age (Ma)</th>
<th>Error (±σ)</th>
</tr>
</thead>
<tbody>
<tr>
<td>RLF-25</td>
<td>29.37556,68.64417</td>
<td>36.02</td>
<td>121.60</td>
<td>147.44</td>
<td>537.90</td>
<td>156.24</td>
<td>3.5</td>
<td>5.7</td>
<td>0.1</td>
<td></td>
</tr>
<tr>
<td>VH-01</td>
<td>28.37583,68.05222</td>
<td>58.64</td>
<td>7.12</td>
<td>14.07</td>
<td>19.97</td>
<td>480.29</td>
<td>23.71</td>
<td>2.1</td>
<td>3.4</td>
<td></td>
</tr>
<tr>
<td>VH2-365</td>
<td>28.37583,68.05222</td>
<td>36.64</td>
<td>1.66</td>
<td>5.61</td>
<td>13.80</td>
<td>15.80</td>
<td>4.9</td>
<td>8.1</td>
<td>1.0</td>
<td></td>
</tr>
<tr>
<td>TR1-10</td>
<td>27.70104,67.70104</td>
<td>46.01</td>
<td>48.18</td>
<td>3.64</td>
<td>131.52</td>
<td>49.04</td>
<td>166.7</td>
<td>239.0</td>
<td>3.5</td>
<td></td>
</tr>
</tbody>
</table>

### Rio La Troya depositional area

One interval from the Tamberia Formation in the La Troya depositional area in the southern Fiambalá basin (south of the Guanchin section of Carrapa et al., 2006, 2008 and Safipour et al., 2015) was sampled for AHe thermochronology. Sample TR1-10 is located 10 m from the base of the section (estimated burial depth of 3,300 m calculated by projecting the stratigraphy along the
Guanchin river along strike) (Fig. 2). Four grains record cooling ages between 1.9 ± 0.1 and 0.6 ± 0.1 Ma which are younger than the depositional age of the strata with eU values between 14 ppm and 33 ppm (Fig. 5, Table 1). A fifth grain records a cooling age much older than the age of the strata, 239.0 ± 3.5 Ma with an eU value of 49 ppm. AFT thermochronology of two samples at located 10 m (TR1-10) and 800 m from the base of the section (TR1-813) indicate AFT cooling ages of 3.5 ± 0.8 Ma and 29.7 ± 2.7 Ma respectively (Fig. 5, Table 2). Estimated burial depths of 3,300 m (TR1-10) and 2,500 m (TR1-813) are calculated by projecting stratigraphy along the Guanchin river along strike.

### Flexural Modelling

The results from flexural modelling of a Precordilleran load on crust with variable effective elastic thickness produce a series of solutions with sufficient accommodation space to explain the observed basins thickness of up to 8–9 km (Fig. 7). A weak zone with an effective elastic thickness (Te) of 5 km produces greater total deflection than a weak zone Te of 10 km (Fig. 7). However, the location of the weak zone is the strongest control on the total magnitude of deflection. A weak zone located partially or completely under the load of the Precordillera produces the greatest magnitude of deflection both under the load and in the adjacent basin in all three simulations (Fig. 7).

### Geohistory Analysis

Geohistory analysis of four depositional areas in this study has tectonic subsidence curves (Fig. 8) that produce the magnitude of subsidence observed in the study depocentres. Maximum tectonic subsidence accounts for ca. 3.5 km of accommodation space in the Vinchina depositional area and ca. 1.5 km of accommodation space in the Huaco, La Flecha and Guanchin River depositional areas (Fig. 8). The slightly convex up shape of the subsidence curves for the La Flecha and Huaco depositional areas is consistent with a foreland basin (Decelles & Giles, 1996; Xie & Heller, 2009; Allen & Allen, 2009). In the Vinchina area, the subsidence curve is roughly linear, whereas in the Guanchin River depositional area, the subsidence curve is slightly concave upward. Deviations from the typical convex upward shape of foreland basins in these locations may reflect the short sampling period (10–4 Myr) of these data sets relative to classical models.
Fig. 6. (a) The time-temperature histories of three samples (black shapes) taken from different locations in a stratigraphic section. Each time-temperature history includes burial at a rate of 1 mm yr\(^{-1}\) and exhumation at 8 Ma. The sample is completely exhumed in this model by 2 Ma. The time-temperature history of the sample controlled by the geothermal gradient. Three geothermal gradients illustrated by coloured lines show variations in a sample’s time-temperature history with geothermal gradients of 15 °C km\(^{-1}\), 25 °C km\(^{-1}\), and 35 °C km\(^{-1}\). (b) The predicted distribution of AFT and AHe cooling ages given the time-temperature history in (a). Polygons indicated the predicted cooling ages given the assumed range of geothermal gradient (labelled at the top of each diagram). Vertical lines highlight the highest stratigraphic level of thermally reset sandstone for a given geothermal gradient. Details of this output are discussed in the text.
which sample from 50 to 100 Myr (Xie & Heller, 2009). Slow subsidence typical of early foreland sedimentary records is not documented in our study, but new work in the Bermejo basin at ca. 30°S identifies an earlier, Eocene period of foreland sedimentation (Fosdick et al., 2017). This record, if present in the Vinchina and Guanchin River depositional areas, would add a convex component to the subsidence curves in this study making the curve more consistent with traditional foreland basin subsidence curves.

Dynamic processes are predicted to produce a maximum subsidence of < 1 km over 10 Myrs and the shape of the subsidence curve is predicted to be concave up (Heller & Liu, 2016). Although we recognize that a dynamic control on a flexurally driven basin would be difficult to recognize, our data suggest that such component is not necessary to explain basin subsidence.

MIOCENE GEOTHERMAL GRADIENT

The thermochronological data produced in this study combined with previous work by Fosdick et al. (2015) in the Huaco depositional area, Collo et al. (2017) in the La Flecha and Vinchina depositional areas, as well as Carrapa et al. (2006, 2008) and Safipour et al. (2015) in the Fiambalá basin constrain the spatial and temporal thermal regime of the Mio-Pliocene strata above the flat-slab subduction zone (Figs 5 and 6). At five depositional areas with complete stratigraphic sections, Huaco, La Flecha, Vinchina, La Troya and Guanchin River, we directly calculate the minimum geothermal gradient using AFT and AHe cooling ages at varying depths. Estimates of maximum palaeodepth for each sample assume that strata have not been removed above the preserved stratigraphic sections. The young (ca. 4 – 6 Ma) depositional ages from the top of all but the Rio La Flecha section match the youngest (ca. 4 – 6 Ma) dates from thermochronometers buried in the section suggesting if additional strata was deposited, it was concurrent with basin exhumation and did not further bury the sample. The consistency among thermochronology cooling ages in the La Flecha section and other sections suggests the preserved section here was not significantly eroded. To calculate a geothermal gradient, we assume that AFT and AHe samples with cooling ages younger than the depositional age of the strata have been exposed to temperatures above the PAZ and PRZ, respectively, whereas samples with cooling ages older than the depositional age of the strata have not been exposed to temperatures in the PAZ and PRZ. For the AHe system, we assume that samples with cooling ages both above and below the depositional age of the strata have been exposed to temperatures in the PRZ (Farley, 2000; Flowers et al., 2009).

Our calculations described below are summarized in Table 3. In the Huaco depositional area, sample HC-07 buried to a depth of 2.2 km is partially reset by the AHe system suggesting that it was exposed to temperatures in the PRZ between 80° to 55 °C (Fosdick et al., 2015). These conditions imply a maximum geothermal gradient of 36.4 °C km⁻¹ and a minimum geothermal gradient of 25.0 °C km⁻¹ (Table 3). Similarly, in the La Flecha
depositional area, sample RLF2-610 buried to depths of 3 km yields AHe dates consistent with exposure to the PRZ. These constraints produce a maximum geothermal gradient between 26.7 and 18.3 °C km⁻¹ (Table 3). In the Vinchina depositional area, the AFT system in sample VN3-4032 (3.3 km depth) is completely reset, whereas sample VN4-704 is unreset. Assuming, conservatively, that 3.3 km depth is the last stratigraphic interval reset by the AFT system and knowing that this interval was exposed to temperatures between 120° and 80 °C allows us to calculate a geothermal gradient between 36.4 and 24.2 °C km⁻¹ (Table 3). In the La Troya depositional area, the AFT system is completely reset in sample TR1-01 (3.3 km depth) but not in TR1-813 (2.5 km depth). Again assuming conservatively that 3.3 km is the uppermost stratigraphic interval reset by the AFT system in the La Troya depositional area, results in a maximum geothermal gradient of 36.4 °C km⁻¹ and a minimum geothermal gradient of 24.2 °C km⁻¹ (Table 3). In the Guanchin River transect, the AHe system is completely reset in samples buried to 3.0 km depth (Safipour et al., 2015). We note that possible growth strata in this transect (Carrapa et al., 2008) indicate that sedimentation may have been coeval with exhumation making total burial depth difficult to constrain. Assuming that temperatures must have exceeded temperatures of the PRZ (55–80 °C) produces a minimum geothermal gradient of 26.7 °C km⁻¹. AFT samples from the Tamberia Formation (buried by ca. 4 km; Carrapa et al., 2006) are not reset producing a maximum geothermal gradient of 30 °C km⁻¹. In summary, the geothermal gradients measured across Miocene–Pliocene depositional areas along the Central Andes and within the Sierras Pampeanas are remarkably consistent despite the fact that these
Depositional areas are located over 350 km along strike from each other (Table 3). These results are also consistent with forward thermal models (Fig. 6) and together suggest that the geothermal gradient in the late Miocene–early Pliocene was not lower than 25 \( ^\circ \text{C} \cdot \text{km}^{-1} \), and likely was between 25 and 35 \( ^\circ \text{C} \cdot \text{km}^{-1} \). These results are consistent with the modern geothermal gradient whose minima range from ca. 24 \( ^\circ \text{C} \cdot \text{km}^{-1} \) near Guandacol (F. Fuentes, personal communication) to ca. 18 \( ^\circ \text{C} \cdot \text{km}^{-1} \) and ca. 18 \( ^\circ \text{C} \cdot \text{km}^{-1} \), in the Pozuelos and Matagusanos boreholes respectively (Collo et al., 2017).

### DISCUSSION

#### Late Miocene–Pliocene geothermal gradient

The thermal history of Miocene–Pliocene basins above the Central Andean flat-slab subduction zone is a subject of debate. A mid-late Miocene geothermal gradient between 25 and 35 \( ^\circ \text{C} \cdot \text{km}^{-1} \) derived from thermochronometers contradict proposals by Collo et al. (2011, 2017) that the base of the Vinchina depositional area could not have reached temperatures over 100 \( ^\circ \text{C} \), and suggesting a geothermal gradient between 12 and 18 \( ^\circ \text{C} \cdot \text{km}^{-1} \). The discrepancy between the thermochronometric and the illite/smectite geothermometer data used by Collo et al. (2011, 2017) may be explained by the presence of undetected sand-sized grains in the clay matrix which can more readily host fluids, a condition that can inhibit the transformation from smectite to illite (Pytte & Reynolds, 1989) following existing kinetic models (Huang, 1993). The consistency of our results from both AFT and AHe thermochronometers across stratigraphic sections supports a normal geothermal gradient between 25 and 35 \( ^\circ \text{C} \cdot \text{km}^{-1} \), as opposed to refrigerated and anomalously cool.

The late Miocene–Pliocene geothermal gradient is higher than expected for a retroarc flat-slab subduction zone where the removal of the asthenospheric wedge is hypothesized to reduce the geothermal gradient to ca. 10–20 \( ^\circ \text{C} \cdot \text{km}^{-1} \) (Dumitru, 1990). Minimum modern geothermal gradients range from 24 \( ^\circ \text{C} \cdot \text{km}^{-1} \) to 18 \( ^\circ \text{C} \cdot \text{km}^{-1} \) (Collo et al., 2017; Astini et al., 2005; F. Fuentes, personal communication) match those recorded by our study in the late Miocene. Dumitru et al. (1991) suggested that a reduction in the regional geothermal gradient during flat-slab-induced refrigeration has a lag time of 5–10 Myr following the inception of shallow subduction. This study, however, does not provide any evidence for a reduction in the geothermal gradient since the Miocene. This may indicate that regional refrigeration does not accompany flat-slab subduction or that the lag time for refrigeration is longer than previously supposed.

### Table 3. Estimation of geothermal gradient using low-temperature thermochronology

<table>
<thead>
<tr>
<th>Location</th>
<th>Total Section Thickness (km)</th>
<th>System</th>
<th>Assumption</th>
<th>Maximum Closure T (°C)</th>
<th>Minimum Closure T (°C)</th>
<th>Maximum Geothermal Gradient (°C km⁻¹)</th>
<th>Minimum Geothermal Gradient (°C km⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Guanchin River</td>
<td>4.5</td>
<td>AHe</td>
<td>(1) Sample RS.10.15 (3.0 km depth fully reset by AHe) cooled at temperatures greater than PRZ</td>
<td>NA</td>
<td>80</td>
<td>30</td>
<td>26.7</td>
</tr>
<tr>
<td>La Troya</td>
<td>4.5</td>
<td>AFT</td>
<td>(1) Projection of Fiambalá basin stratigraphy from Guanchin River places TR1-01 at 3.3 km burial depth. (2) Sample TR1-01 (3.3 km depth, fully reset by AFT) is the top of the PAZ</td>
<td>120</td>
<td>80</td>
<td>36.4</td>
<td>24.2</td>
</tr>
<tr>
<td>Vinchina</td>
<td>8.5</td>
<td>AFT</td>
<td>(1) Sample VN3-4032 (3.3 km depth, fully reset by AFT) is the top of the PAZ</td>
<td>120</td>
<td>80</td>
<td>36.4</td>
<td>24.2</td>
</tr>
<tr>
<td>La Flecha</td>
<td>4.9</td>
<td>AHe</td>
<td>(1) Sample RLF2-610 (3 km depth, partially reset by AHe) is in the PRZ</td>
<td>80</td>
<td>55</td>
<td>26.7</td>
<td>18.3</td>
</tr>
<tr>
<td>Huaco</td>
<td>5.2</td>
<td>AHe</td>
<td>(1) Sample HC-07 (2.2 km depth, partially reset by AHe) is in the PRZ</td>
<td>80</td>
<td>55</td>
<td>36.4</td>
<td>25.0</td>
</tr>
</tbody>
</table>
Spatio-temporal basin evolution in the Precordillera and Sierras Pampeanas regions

Cooling ages from low-temperature thermochronometers that have been reset by sediment burial reflect the timing of basin exhumation (Fig. 5). Within a given stratigraphic section from this study, thermochronometers young with depth (within error) as expected by increasing burial temperatures (Fig. 5). The difference in cooling ages for samples in the same depositional area, but at different stratigraphic levels is less than ca. 4 Myr for both AFT and AHe samples (Fig. 5) indicating rapid cooling and exhumation of individual depositional areas.

The distribution of cooling ages across depositional areas can be used to reconstruct the spatio-temporal evolution of deformation-related exhumation. Thermochronometers from all stratigraphic levels in the Huaco, La Flecha, Vinchina, Valle Hermoso and Guanchin River depositional areas display remarkable similarity despite the lateral separation of these depositional areas by ca. 250 km along strike (Fig. 5b). For these stratigraphic sections, reset AFT cooling ages are all between 9 and 6 Ma; reset AHe cooling ages are between 7 and 2 Ma (Fig. 5b). Cooling ages do not young systematically either north or south within these basins. AFT and AHe thermochronometers from the Rio La Troya depositional area are slightly younger than in other depositional areas. Here the single reset AFT sample is 3.5 ± 0.8 Ma and AHe ages are between 2 and 0.5 Ma.

The striking homogeneity in AFT and AHe cooling ages combined with similar late Miocene–early Pliocene geothermal gradients in the depositional areas provides constraints on the timing and location of the flat-slab during this time. Specifically, the north to southward migration of Juan Fernandez Ridge from ca. 27°S to 32°S suggested by Pilger (1981, 1984) and Yáñez et al. (2001) between ca. 14 Ma and 9 Ma would predict a younging of basin exhumation to the south which is not observed (Fig. 5). Nor does the predicted north to southward sweep of flat-slab subduction manifest itself in spatially differentiated geothermal gradients (Table 3). Instead, exhumation is uniform and concentrated between ca. 8 and 4 Ma in depositional areas between 30°S and 28°S. The geothermal gradient across depositional areas between 30°S and 27°S is within error.

Assuming that the basin history record is at least partially a response to flat-slab subduction, several alternatives could explain the uniform spatio-temporal distribution of thermal regimes and late Miocene–Pliocene basin exhumation: 1) The southward migration of the Juan Fernandez Ridge, and concurrent flat-slab subduction occurs too rapidly to be resolved by thermochronology data (<1 Myr). 2) The north-to-southward sweep occurred more recently (5–0 Ma) than proposed and the record of ridge migration is not yet manifested in the surface geology. 3) Subduction of the Juan Fernandez Ridge instigated flat-slab subduction that synchronously controlled basin exhumation along a wide range of the Andean margin, between 30°S and 27°S; in this case, the recent (<10 Myr) flat-slab event has not had sufficient time to reduce the regional geothermal gradient (Dumitru et al., 1991).

A final scenario (4) may also explain the temporal basin exhumation pattern identified in this study. Late Miocene to early Pliocene basin exhumation and cannibalization due to eastward migration of the retroarc fold and thrust belt has been observed throughout the Central Andes of western Argentina at latitudes of flat-slab subduction, ca. 30°S to 28°S (Zapata & Allmendinger, 1996a,b; Jordan et al., 2001; Carrapa et al., 2011, 2014; Allmendinger & Judge, 2014; Fosdick et al., 2015) as well as latitudes north of the flat-slab subduction zone ca. 27°S to 22°S (Echavarria et al., 2003; Decelles et al., 2011). The lateral extent of thrust belt migration beyond the Miocene–present flat-slab subduction zone introduces the possibility that thrust belt migration is not related to slab angle.

The advance of the deformation front between 30°S and 22°S may be linked by a shared mechanism; however, we propose that the propagation of foreland basin exhumation may have been additionally facilitated between ca. 28°S and 30°S by flat-slab subduction. For example, Fosdick et al. (2015) propose that out-of-sequence thrusting in the southern Precordillera (ca. 30°S) at ca. 5 Ma is driven by flat-slab subduction. Jordan et al. (2001) document a change in the drainage system in the same area between 7 Ma and 4 Ma that may be accompanied by uplift during the transition from conventional to broken foreland basin system. Carrapa et al. (2010) also suggest that a regional drainage reorganization observed in the Guanchin River depositional area (part of the Fiambalá basin) at ca. 6 Ma is related to flat-slab subduction. This evidence suggests that although an advancing deformation front may have controlled basin exhumation as early as 10 Ma, exhumation was further enabled by the onset of flat-slab subduction by ca. 5–6 Ma. Our basin analysis shows that the dynamic effects of the slab subduction on subsidence are not necessary during deposition between ca. 18 Ma and 5 Ma; however, dynamic uplift proposed by Jordan et al. (2001) and modelled by Dávila & Lithgow-Bertelloni (2015) could have facilitated basin exhumation. The timing of exhumation constrained by this study between 7 Ma and 4 Ma may indicate that flat-slab subduction did not occur until this time.
Implications for flat-slab subduction processes

In the Sierras Pampeanas, a prevailing low Miocene–Pliocene geothermal gradient (10–20 °C km\(^{-1}\)) interpreted to be associated with flat-slab subduction has been used to explain the abundant Mesozoic and Palaeozoic cooling ages from low-temperature thermochronometers in the region (Dávila & Carter, 2013). These basement-cored ranges bounded by steep reverse faults (Allmendinger et al., 1990) are hypothesized to have been exhumed and at least partially uplifted in the mid-late Miocene coincident with flat-slab subduction (Allmendinger et al., 1983; Jordan & Allmendinger, 1986; Ramos et al., 2002; Carrapa et al., 2008; Kay & Coira, 2009; Ramos & Folguera, 2009; Fosdick et al., 2015). In the absence of a low geothermal gradient, it will be important to revisit ranges with abundant pre-Neogene thermochronometric cooling ages to document the magnitude of Miocene exhumation relative to pre-Cenozoic exhumation and uplift events using a geothermal gradient between 25 and 35 °C km\(^{-1}\). If so, this revised palaeogeography will be an important consideration for reconstructing the organization of Cenozoic and even Mesozoic basins. Furthermore, decoupling the cause-and-effect relationship between flat-slab subduction and thick-skinned style deformation is an important advance in our understanding of the geologic manifestations of flat-slab subduction.

CONCLUSIONS

The Sierras Pampeanas has been used to investigate flat-slab subduction processes, and the geology of the region has been used to interpret processes associated with the slab angle in modern and ancient subduction zones around the world. Our work in the region has important implications for understanding the relationships between the lower and upper plate during flat-slab subduction. Although eastward migration of the deformation front through the region may have facilitated basin exhumation, the timing of rapid exhumation at 27°S to 30°S between 7 Ma and 4 Ma supports the role of flat-slab subduction on upper plate deformation. However, modelling results from this study indicate that the basin subsidence observed in Miocene–Pliocene basins within the region can be explained by flexure and that dynamic mechanisms are not necessary to explain basin accommodation space.

The normal (25–35 °C km\(^{-1}\)) late Miocene–Pliocene geothermal gradient throughout the region supports the initiation of flat-slab subduction no earlier than 10–5 My ago. The absence of both a north to south younging of basin exhumation and a change in the regional palaeo-geothermal gradient as predicted by a southward migration of flat-slab subduction indicates that either this sweep occurred more rapidly and/or recently than previously proposed (Yáñez et al., 2001) or that the geometry of the Juan Fernandez Ridge does not strongly control the spatial distribution of the flat-slab through time. Further work is required to distinguish between these different alternatives.

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SUPPORTING INFORMATION

Additional Supporting Information may be found in the online version of this article:

Data S1. Backstrip inputs.
Data S2. Detrital zircon data.
Data S3. Geochronology and thermochronology extended methods.

REFERENCES


TASSARA, A., SWAIN, C., HACKNEY, R. & KIRBY, J. (2007) Elastic thickness structure of South America estimated using...


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