Basin evolution in response to flat-slab subduction in the Altiplano

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Abstract: This paper assesses models for basin formation in the Altiplano. New magnetostratigraphy, palynology, and 40Ar/39Ar and U-Pb geochronology from the central Corque Syncline show that the 7.4 km thick section was deposited between 36.7 and 18.7 Ma. The base of the section post-dates exhumation in both the Western and Eastern cordilleras, precluding deposition in a classic retroarc foreland basin setting. Rotated palaeomagnetic vectors indicate counterclockwise rotation of 0.8° (myr)−1 since the early Oligocene. Detrital zircon provenance data confirm previous interpretations of Eocene–early Oligocene derivation from the Western Cordillera and a subsequent switch to an Eastern Cordilleran source. Flexural modelling indicates that loads consistent with palaeoelevation estimates cannot account for all the subsidence. Rather, the timing and magnitude of subsidence is consistent with Eocene emplacement and Oligocene–early Miocene re-steepening of a flat slab. Integration of the magmatic, basin and deformation history provides a coherent model of the effects of flat-slab subduction on the overriding plate. In this model, basin formation in the upper plate was controlled by flat-slab subduction, with subsidence enhanced in the zone of flat-slab subduction, but reduced over the crest of the flat slab. We conclude that the Altiplano was conditioned for plateau formation by Eocene–Oligocene flat-slab subduction.

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The locus, timing and pace of vertical movement of the Earth’s crust is a sensitive recorder of geodynamic processes. Downward movement of the crust (subsidence) is documented by the history of sediment accumulation. The pace of sediment accumulation therefore aids the development of geodynamic models. The Andes are the archetype example of a classic foreland and hinterland basin evolution in response to ocean–continent subduction (Ramos 2010; Horton 2018). Models developed in the Andes guide our understanding of subduction-driven orogenic systems worldwide and throughout time (e.g. Jordan and Allmendinger 1986; Leier et al. 2007; Ding et al. 2014).

Recent models for the Late Cretaceous–Holocene evolution of the central Andes (13–20°S) are dominated by multiple phases of contractual deformation, crustal thickening and associated basin formation (Horton et al. 2001, 2015; Perez and Horton 2014; Sundell et al. 2018; Buford Parks and McQuarrie 2019). However, further south there are well-documented Paleogene episodes of extension and basin formation (Jordan et al. 2001; Horton et al. 2016; Horton and Fuentes 2016; Horton 2018). Models invoking extensional basin formation have also been applied to the southern and central Altiplano (Fig. 1a) (Rochat et al. 1996, 1998, 1999, 2000; Baby et al. 1997; Elger et al. 2005). In addition, although Eocene–Oligocene flat-slab subduction has long been proposed, it has not been incorporated into a systematic model that integrates sedimentation, magmatism and deformation.

Models of both contractual and extensional deformation have sought to explain the great thickness of Cenozoic strata, most notably exposed in the eastern limb of the Corque Syncline (Fig. 1b). Sediment accumulation in this region is usually attributed to flexural subsidence in a retroarc foreland basin (e.g. Horton et al. 2001), sediment ponding in a piggy-back basin (Horton et al. 2002; DeCelles and Horton 2003; Buford Parks and McQuarrie 2019) and/or accumulation in the hanging wall of a west-dipping normal fault (Rochat et al. 1998). Potential thrust duplication of the Cenozoic strata during Neogene shortening (e.g. Lamb and Hoke 1997) has further complicated interpretation of this thick sedimentary sequence.

We present detrital zircon U-Pb geochronology, palynology, magnetostratigraphy and 40Ar/39Ar ages that document the depositional chronology and sediment sources of the Paleogene Potoco Formation and Neogene Totora Formation exposed in the eastern limb of the Corque Syncline (Fig. 2). Our results indicate that c. 7.4 km of strata were deposited between 36.7 and 18.7 Ma. Flexural modelling indicates that a subsidence mechanism beyond flexurally accommodated orogenic loading is needed to account for the accumulated thickness. We conclude that subsidence was driven by a combination of flexure associated with surface loads and dynamic loading associated with flat-slab subduction from the Eocene–Oligocene.

Geological setting

Convergence and subduction of the Nazca plate beneath the South American plate began in the Paleozoic (Mamani et al. 2010) and the margin became a well-organized compressional system in the Late
The Cretaceous (Horton 2018). The central Andes are divided into seven roughly longitudinal tectonomorphic domains that developed in response to shortening and are described from west to east (Fig. 1a). Along the western coast is the forearc, with its constituent Coastal Cordillera, Longitudinal Valley, Salar de Atacama and Precordillera. The steep monocline of the Precordillera marks the western slopes of the Western Cordillera continental magmatic arc. The high-elevation Altiplano separates the Western Cordillera from the bivergent fold–thrust belt of the Eastern Cordillera. East of the Eastern Cordillera are the Interandean and Subandean zones, marked by east-vergent thrust systems, and the undeformed Chaco and Beni plains (Fig. 1a).

The Western Cordillera of the central Andes from c. 14 to 22° is built on metamorphic and igneous rocks of the Arequipa–Antofalla Massif (Shackleton et al. 1979; Wasteney et al. 1995; Wörner et al. 2000; Loewy et al. 2004; Casquet et al. 2010). The Arequipa–Antofalla Massif has igneous and metamorphic rocks as old as 2.1–1.8 and 1.0–0.9 Ga (Sunsás-Grenville), respectively, with an Ordovician–Silurian subduction-related magmatic overprinting (Loewy et al. 2004; Casquet et al. 2010). The Phanerozoic cover of the Western Cordillera of southern Peru, Bolivia and northern Chile consists of scattered Paleozoic and widespread Mesozoic strata deposited primarily in a marine setting (Breitkreuz et al. 1988; Gonzalez 2004; Reimann et al. 2010). Ordovician–Devonian strata in the northern Altiplano were deposited in the Peru–Bolivia trough situated between the Arequipa–Antofalla Massif to the west and the Amazonian craton to the east (Cobbing et al. 1977; Bahlburg et al. 1987, 2011).
Late Paleozoic–Mesozoic back-arc extension produced the regionally extensive ≤3 km Mitu Group, as well as the emplacement of multiple Triassic intrusive complexes in the location of the modern Peruvian and Bolivian Eastern Cordillera (Kontak et al. 1990; Gillis et al. 2006; Perez et al. 2016; Spikings et al. 2016) and northern Chile (Munizaga et al. 2008; Coloma et al. 2017). Jurassic–Cretaceous post-rift subsidence allowed the deposition of thick blankets of clastic and carbonate strata across the Western and Eastern cordilleras. The Late Cretaceous–Cenozoic witnessed a transition to a well-organized contractional subduction regime (Jaillard and Soler 1996; Ramos 2010; Horton 2018).

The onset of orogenesis is marked by Late Cretaceous magmatism and exhumation in the Western Cordillera and is followed by Paleogene rotation of the vertical axis and flat-slab subduction. Evidence for the Late Cretaceous onset of orogenesis in the Western Cordillera is limited as a result of the extensive Neogene volcanic cover. Nevertheless, at c. 16° S, where deep canyons have cut through the volcanic cover, mapping reveals Late Cretaceous shortening (Vicente 1990). North of the Altiplano and in northern Chile (22° S), Eocene shortening in the Western Cordillera (the Incaic orogenic event) is well documented (Mégard 1984; Günther et al. 1998). In the Paleogene, the region north of c. 18–20° S experienced counterclockwise rotation of the vertical axis; a clockwise rotation defines the zone to the south (Roperch et al. 2000, 2006, 2011; Rousse et al. 2005; Arriagada et al. 2006b, 2008; Eichelberger and McQuarrie 2015). The central Andes from c. 16 to 24° S has experienced greater shortening than regions to the north or south (Gotberg et al. 2010; Eichelberger and McQuarrie 2015). Overlapping in age with the largest magnitude of rotation and shortening (Arriagada et al. 2008; Buford Parks and McQuarrie 2019), there is evidence for an Eocene–Oligocene period of shallow or flat-slab subduction (Sandeman et al. 1995; James and Sacks 1999; Perelló et al. 2003; O’Driscoll et al. 2012; Ramos 2018). A similar episode of flat-slab subduction has been proposed for northern Chile and Bolivia in the Oligocene–Miocene (Jiménez et al. 2009; Kay and Coira 2009; Ramos 2018).
Coeval with the onset of Andean orogenesis, deposition in the retroarc Altiplano region transitioned from marine to non-marine in the Late Cretaceous–Paleocene and deposition was locally continuous through at least the late Miocene (MacFadden et al. 1985; McArae 1990; Kay et al. 1998; Ropercher et al. 1999; Horton et al. 2001; Garzione et al. 2006). The youngest marine unit is the Maastrichtian–Danian (73–60 Ma) El Molino Formation, which was deposited in a restricted marine setting (Gayet et al. 1991, 1993; Sempere et al. 1997). Above the El Molino Formation, the regionally extensive, upwards-coarsening Paleocene (60.0–58.2 Ma) Santa Lucia Formation was deposited in alluvial and lacustrine settings (Gayet et al. 1991; Sempere et al. 1997; Horton et al. 2001). The overlying upper Paleocene (58.2–57.7 Ma) Cayara Formation consists of cross-bedded sandstones with minor mudstones, conglomerates and root traces (Sempere et al. 1997). The significance of the Cayara Formation is disputed, with some researchers interpreting it as overlying a regionally significant unconformity and some merging it with the underlying Santa Lucia Formation (Sempere et al. 1997; Horton et al. 2001).

We focus on the Potoco and lower Totoro formations exposed adjacent to the village of Chuquichambi on the east limb of the Corque Syncline at c. 18° (Figs 2 and 3) (Horton et al. 2001). Regionally, the Potoco Formation overlies a 20 to 100 m thick zone of palaesoals that overlies the Cayara or Santa Lucia formations and underlies the Miocene Totoro Formation along a transitional contact (Horton et al. 2001). The overlying upper Paleocene (58.2–57.7 Ma) Cayara Formation consists of cross-bedded sandstones with minor mudstones, conglomerates and root traces (Sempere et al. 1997). The significance of the Cayara Formation is disputed, with some researchers interpreting it as overlying a regionally significant unconformity and some merging it with the underlying Santa Lucia Formation (Sempere et al. 1997; Horton et al. 2001).

Methods

The methods are summarized in the following sections; a full description is given in the Supplemental Material.

Palaecology methods

Seven organic-rich siltstone samples from the lower 4500 m of the Chuquichambi succession were collected for palynology analysis at the University of California, Los Angeles (Fig. 3a). Sample preparation involved the digestion of 5–10 g of material from each sample in hydrochloric and hydrofluoric acids, followed by floating in a solution of zinc bromide (specific gravity 2.0). The materials were then mounted on glass slides and examined using a standard light microscope.

\(^{40}\)Ar/\(^{39}\)Ar geochronology methods

Five samples containing biotite crystals from ashfall tuff horizons near the top of the Chuquichambi section (Fig. 3a) were analysed using \(^{40}\)Ar/\(^{39}\)Ar geochronology at the University of California, Los Angeles, following the analytical procedures described by Quideleur et al. (1997), McDougall et al. (1999) and Horton (2005). The ages were calculated using conventional decay constants and isotopic abundances. Unless otherwise noted, the reported ages are weighted-mean ages calculated for the complete (ten-step) release spectra for each sample. All errors are reported at the 1σ level.

Magnetostatigraphy methods

Palaeomagnetic samples were collected from two stratigraphic sections, Chuquichambi I and Chuquichambi II, which are separated by the fault zone described earlier (Fig. 3c, d). A minimum of three oriented cores were collected with a Pomeroy drill from 335 stations, which were drawn into the field notes and recorded in the corresponding stratigraphic level (Fig. 3b, c).

All 335 samples were analysed for natural remanent magnetization (NRM) on the three-component 2G superconducting rock magnetometer at the University of Pittsburgh (Fig. 3d). The ‘A’ samples from each site were treated with 13 additional thermal demagnetization steps (200, 300, 400, 450, 500, 550, 570, 600, 620, 640, 650, 670 and 680°C) to determine their magnetic mineralogy and define appropriate treatments for the remaining samples at each site. An additional 38 sites had sufficient sample remaining for a fourth specimen to be subjected to stepwise alternating field demagnetization experiments using the following magnetic intensity steps: NRM, 2.5, 5, 10, 12.5, 15, 17.5, 20, 25, 30, 40, 45, 50, 60, 70, 80 and 90 mT. This treatment was successful with some samples, but did not separate the components of the NRM in other samples.

Subsidence methods

We evaluated tectonic subsidence to remove the effect of sediment loading and to compare tectonic subsidence at this location with other locations in the central Andes. We first calculated the decompacted sediment thickness through time and then calculated the tectonic subsidence assuming Airy (i.e. non-flexurally supported) isostatic compensation. Tectonic subsidence was calculated assuming a density for the mantle of 3.3 g cm\(^{-3}\) and no fill in the basin (a density of fill in the basin of 0 g cm\(^{-3}\)). The input parameters for the sediment accumulation, back-stripping and tectonic subsidence plots are presented in Supplemental Table S1.

Detrital zircon U–Pb geochronology methods

Eleven sandstone samples were collected for detrital zircon U–Pb geochronology after measurement of the stratigraphic section and correlated into the measured stratigraphic section by tracing well-exposed sandstone ridges from the collection points to the stratigraphic section (Figs 2 and 3; Supplemental Table). Stratigraphic uncertainties are estimated to be less than ±50 m. Zircons were extracted using standard procedures, including crushing to <400 μm, water table, heavy liquid (3.28 g cm\(^{-3}\)) density separation and magnetic separation using a Frantz magnetic separator. Eight of the samples were analysed at the University of Houston using methods defined by Shaulis et al. (2010) and Sundell (2017). The remaining three samples were analysed at the LaserChron Center at the University of Arizona following methods outlined by Gehrels et al. (2008). Full analytical data are reported in Supplemental Tables S2 and S4 and the ages are summarized in Supplemental Table S3.

To determine which regions and stratigraphic intervals contributed detritus to the Eocene–Miocene samples from Corque, we ran a Monte Carlo mixture model using DZmix (Saylor and Sundell 2016). Potential source samples from six sediment provenance studies were merged into eight stratigraphic and regional groups to determine both the location and structural level of exhumation.
Fig. 3. Stratigraphic section with sample locations, magnetostratigraphy and correlation to the Global Polarity Time Scale. (a) Sample names and data types are indicated with coloured diamonds; blue, zircon samples; red, tuff samples; green, palynology data. Palaeocurrent data from Horton et al. (2001). (b) Stratigraphic section reported by Horton et al. (2001). (c) Plot of the virtual geomagnetic pole latitudes of the palaeomagnetic sites in the Chuquichambi I and II sections v. their stratigraphic level in the lithostratigraphy. Black circles represent Class I data and open circles are Class II sites. (d) Polarity abstracted to a standard black and white polarity scale for correlation to the Global Polarity Time Scale. Black signifies normal polarity and white is reversed polarity. Reversals are placed halfway between adjacent sites with opposite polarities. There are two different interpretations in the lower Chuquichambi I section; the darker of the two is our preferred correlation. VGP, virtual geomagnetic pole.
Table 1. Location data for U–Pb detrital zircon and \(^{40}\)Ar/\(^{39}\)Ar tuff biotite samples

| Sample name | Stratigraphic level (m) | Latitude (° S) | Longitude (° W) |
|-------------|------------------------|----------------|----------------|
| Detrital zircon U–Pb samples |
| 300618-07   | 6690                   | 17.9999        | 67.83933       |
| 300618-06   | 4125                   | 17.96842       | 67.81998       |
| 300618-05   | 2775                   | 17.94960       | 67.81707       |
| CQT-2       | 2439                   | 17.94287       | 67.82296       |
| CQT-5       | 2092                   | 17.94085       | 67.81725       |
| 300618-04   | 727                    | 17.92167       | 67.80939       |
| 300618-03   | 530                    | 17.91787       | 67.80824       |
| 300618-02   | 175                    | 17.90814       | 67.81114       |
| CQT-1       | 163                    | 17.91343       | 67.80394       |
| 300618-01   | −50                    | 17.90596       | 67.81043       |
| 290618-02   | −700                   | 17.91679       | 67.77167       |
| Tuff biotite \(^{40}\)Ar/\(^{39}\)Ar samples |
| CBR2        | 6445                   | 17.99822       | 67.829713      |
| 03BE2       | 6927                   | 17.99422       | 67.851573      |
| CQT13       | 6949                   | 17.99491       | 67.852006      |
| CBR5        | 7127                   | 17.99761       | 67.853615      |
| CQT16       | 7336                   | 17.99795       | 67.858242      |

Samples below 0 m are tectonically deformed and the thickness is approximated.

Results

through time (Bahlburg et al. 2009; Leier et al. 2010; Reimann et al. 2010; Wotzlaw et al. 2011; Perez and Horton 2014; Sundell et al. 2015). Zircon ages >200 Ma were truncated from the Eocene–Miocene samples and potential sources to avoid incorporating the effects of Andean volcanism. All samples and sources were modelled as kernel density estimates with 15 myr bandwidths. Results are reported from the top 1% of 10,000 trials ranked by cross-correlation coefficient (Supplemental Table S5).

We calculated the maximum depositional age (MDA) for all samples using three methods: the youngest statistical population method (Coutts et al. 2019), the youngest graphical peak (Coutts et al. 2019) and the maximum likelihood age (Vermeech 2021) (Supplemental Table S6). All methods yielded comparable ages and are reported in the Supplemental Text.

Flexural modelling methods

Flexural modelling was conducted along a 2D cross-section assuming distributed loading of an infinite elastic lithosphere using equations outlined by Wangen (2010) and Allen and Allen (2013). Modelling simultaneously varied the topographic load of the Western and Eastern cordilleras and the effective elastic thickness (EET) to minimize the misfit between the modelled and observed subsidence at the Chuquichambi location. The eastern end of the Western Cordillera load is located 75 km from the eastern limb of the Corque Syncline, whereas the western end of the Eastern Cordillera load is located 25 km east of the Corque Syncline based on the distance to crustal loads at 24 Ma in the balanced cross-sections (McQuarrie 2002; Buford Parks and McQuarrie 2019). For comparison, the modern topographic load of the Eastern Cordillera is c. 45 km east of the Chuquichambi section and the topographic load of the Western Cordillera is at least 50 km west of the section. The EET was allowed to vary between 5 and 20 km based on regional EET maps (Stewart and Watts 1997; Tassara 2005; Pérez-Gussinyé et al. 2007; Tassara et al. 2007; Sacek and Ussami 2009). All the parameters are given in Supplemental Table S7. We conducted three sets of flexural models following the rationales laid out in the following text (Fig. 4). To take advantage of palaeoelevation constraints, all three models were run with two time intervals: >25 and >18 Ma.

In the first sets of models, we determined the optimum model result assuming a maximum elevation of 1.75 and 2.25 km for the Eastern and Western cordilleras, respectively, at >25 Ma (Fig. 4a, b). For the >18 Ma interval, we conservatively take the maximum palaeoelevation calculated by Leier et al. (2013) of 3.7 km as the maximum elevation of the Eastern Cordillera and a maximum of 6 km for the Western Cordillera (Saylor and Horton 2014; Sundell et al. 2019b). In the second set of models, we determined the loads needed to match the observed sedimentary thickness with only flexural subsidence mechanisms (Fig. 4c, d). We set a maximum load height for both the Western and Eastern cordilleras of 6 km. In the third set of models, we explored the effect of changing the distance between the depocentre and the tectonic loads (Fig. 4e, f). We retained the palaeoelevation constraints of the first set of models, but decreased the distance to the Western Cordillera load to 25 km. We maintained the distance to the Eastern Cordillera load at 25 km.

Results

Palynology results

Five of the seven samples yielded useable palynomorphs (Supplemental Table S8). Two of the five were previously reported by Horton et al. (2001). Specific species within the observed palynomorph assemblages provide age restrictions that range from late Eocene through to the Oligocene. The presence of Margocpolisites vanwijhei and the questionable presence of Retibreviriculites triangularis, which do not range above the late Eocene, suggest that the lower stratigraphic levels are of late Eocene age. For other samples, less precise late Eocene to early Oligocene age assignments can be made on the basis of occurrences of Echiperiportites estelae, Zonocostites ramonae and Podocarpidites sp. 1 and 2, which do not range below the late Eocene, and Scabraperiportites nativensis, which does not range above the early Oligocene.

\(^{40}\)Ar/\(^{39}\)Ar geochronology results

Weighted-mean \(^{40}\)Ar/\(^{39}\)Ar ages of these tuffs provide an independent anchor for magnetostratigraphic correlation. Step-heating results are presented in the Supplemental Text and Supplemental Table S9.

Magnetostatigraphic results

Laboratory results

The NRM magnetic dipole moment/unit volume (M) values vary between 1.16 × 10\(^{-3}\) and 6.36 × 10\(^{-2}\) A m\(^{-1}\). The M values were much stronger at the base and top of the Chuquichambi I section (below Q-70 and above Q-186). Sample NRMs between Q-70 and Q-186 approximate an even distribution between 1.16 × 10\(^{-3}\) and 6.06 × 10\(^{-2}\) Am\(^{-1}\).

A statistical mean direction of the palaeomagnetic vector was determined for each site (Fisher 1953) using a minimum of three samples per site. Statistically significant results using the R statistic (Watson 1956) suggest that the stable component of magnetization was isolated. As used here, Class I sites are those whose mean direction satisfies Watson criteria. When the number of samples, N, is three, then R > 2.62 is considered to be statistically significant (Watson 1956). Class II sites designate stratigraphic levels for which only two samples from the site survived to be analysed. In our analysis, Class II sites are only used to support single-site Class I polarity zones or to fill a long stratigraphic gap between two Class I sites of the same polarity.

Of the 335 palaeomagnetic sites collected, 281 (83%) are presented in the magnetic polarity columns (Fig. 3c, d). A total of
275 sites provided Class I data and six were Class II. Fifty-five sites produced results that either failed the criteria of Watson (1956) or were destroyed during transit or in the laboratory. Of these, 14 (4%) were Class II sites that were not included because their inclusion neither affected nor enhanced our interpretation of the results. Sixteen sites did not fit into a Class I or II category. Thermal demagnetization plots (Hatakeyama 2018) indicate that a minor component of magnetite is present in most samples, but that hematite is the primary carrier of NRM (see Supplemental Text).

Ages of the strata

Our interpretation of the age of the strata is based on magnetostratigraphic correlation with the Global Polarity Time Scale (Cande and Kent 1992, 1995; Ogg 2012), supported by isotopic ages obtained in this work and palynological samples (Horton et al. 2001). These data demonstrate that the Chuquichambi I and II sections were deposited between 36.85 ± 0.15 and 18.7 Ma (Fig. 3d). Vector average orientations are available in Supplemental Table S10. We place the base of the Chuquichambi I section between 37.0 and 36.7 Ma, c. 40 m below the base of Chron C16n.2. We place the top of the section between 27.9 and 27.4 Ma at the level of site Q195, which is situated at the approximate base of the fault zone within the upper portion of Chron C9r. The primary carrier of NRM at the top of Chuquichambi I is not readily obvious. This is possibly due to an influx of hot water into the fault zone, altering the original minerals and depositing a chemical remanent magnetization. The fault zone completely obliterates the palaeomagnetic interpretation of any strata deposited between the base of Chron C9n and the base of Chron C8n.2n. The base of the Chuquichambi II section occurs near the base of Chron C8n.2n between 26.4 and 26.0 Ma. The top of our

![Diagram](http://jgs.lyellcollection.org/Downloaded from http://jgs.lyellcollection.org/ by guest on July 14, 2022)
The palaeomagnetic section is assigned an age of c. 18.7 Ma at the approximate top of Chron C6n (Fig. 5).

Our Chuquichambi I correlation presents a normal polarity zone and a reversed polarity zone at the top of Chron C13n that do not appear on the Global Polarity Time Scale (Cande and Kent 1995; Ogg 2012). These are likely to be either cryptochron C13n-1 or C13r-1, respectively, identified in Cande and Kent (1992), but not recalibrated in later papers. Cryptochron C11r-1 (Cande and Kent 1992) appears to be present at the base of Chron C11n.2n. The extra normal polarity zone above Chron C10n.1n does not correspond to any cryptochron reported in Cande and Kent (1992). We suggest that this is either (1) a new cryptochron or (2) an unrecognized minor fault at the base of the fault zone repeating a thin slice of the section. The Chuquichambi II section is missing either Chron C7n.1n or C7n.1r and Chron C6Cn.2n, but is otherwise complete. Two explanations exist for the absence of these polarity zones: (1) deposition may not have occurred during these missing intervals (Cande and Kent 1995); or (2) our sample spacing was not detailed enough to register all the reversals present in the section.

Stochastic model

An independent evaluation of our interpretation is possible using the statistical models of Johnson and McGee (1983), summarized in Table 2. These models evaluate the probability of detecting the discovery of number of polarity reversals given the sampling density and the average amount of time covered by polarity zones for the given time period – in this case, the late Paleogene. Because of the high number of sample sites and lower number of reversals in the Chuquichambi I section, there is a lower probability of encountering a polarity reversal between palaeomagnetic sites. This lower probability results in similar time estimates for uniform, uniformly random and exponentially random sampling models with respect to time. For the Chuquichambi II section, with fewer sites, but more reversals, the uniformly random distribution model provides the best fit. The results suggest that our sample spacing was tighter than it needed to be to achieve similar results.

Sediment accumulation rate and tectonic subsidence

A plot of the stratigraphic level of reversals v. their age (Fig. 5) reveals a low initial sediment accumulation rate of c. 0.2 km (myr)^{-1} at the base of the Chuquichambi I section between 36 and 34 Ma. From 34 to 32 Ma, the rate gradually increases to c. 0.4 km (myr)^{-1} and roughly retains this linear rate to the top of the section. The Chuquichambi II section continues the linear sedimentation rate of c. 0.4 km (myr)^{-1} throughout the section. At the risk of overinterpreting the results, however, a slight acceleration is suggested between 25 and 24 Ma, followed by a return to the lower rate. The plot suggests a second short acceleration between 23 and 22 Ma. The inflections noted in both magnetostratigraphic sections create subdued sigmoidal curves similar to those observed in response to uplift in Neogene foreland strata from northern Argentina (Johnson et al. 1986; Reynolds et al. 1990, 2000, 2001; Malizia et al. 1995; Galli et al. 1996; Hernández et al. 1999; Echavarria et al. 2003).

The tectonic subsidence curve for the Chuquichambi I section shows a relatively slow subsidence of 0.5 km between 36.7 and 33 Ma (i.e. c. 0.16 km (myr)^{-1}; Fig. 6a). Subsidence accelerates to c. 0.18 km (myr)^{-1} from 33 to 27.9 Ma, before decreasing to 0.1 km (myr)^{-1} between 26.0 and 18.7 Ma. Subsidence rates increase again from 14.2 to 9 Ma before decreasing to their lowest rate between 9 and 5.3 Ma, based on the record at Callapa (Roperch et al. 1999; Garzonne et al. 2006).

Reversal test and rotation

Using the plotting protocol of Allmendinger et al. (2011), reversal tests (Fig. 7a, b) of the unambiguous Class I data for the

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Table 2. Estimates for the time intervals (Δt) in the Chuquichambi I and Chuquichambi II stratigraphic sections

|                  | No. of palaeomagnetic sites (N) | No. of reversals found (R) | Probability P = R(N - 1) | Mean spacing of palaeomagnetic sites* S = f(P) | Expected time interval\(\Delta t = S/n\) (myr) |
|------------------|---------------------------------|---------------------------|--------------------------|--------------------------------------------|---------------------------------|
| Chuquichambi I   | 166                             | 24                        | 0.1454                   | 0.205                                      | 12.3 ± 4.5 (1σ)                |
| Chuquichambi II  | 115                             | 27                        | 0.2368                   | 0.450                                      | 18.6 ± 4.5 (1σ)                |

*From Johnson and McGee (1983; equation 7 (\(P = S/(2S + 1)\)). Assuming that samples are distributed with a uniform randomness over the stratigraphic interval. The statistical terms used their paper are S-bar and \(\tau\)-bar.

*An estimate of the median length of polarity intervals (\(\tau\)) for the late Paleogene is 3.6 \(\times 10^2\) kyr.
Chuquichambi I and II sections show that, in both cases, the data fall into two distinct antipodal fields, suggesting that the magnetic component isolated by the demagnetization treatments represents the detrital remanent magnetization and not a thermal or chemical post-depositional overprint. Mean normal and reverse directions are within $5^\circ$ ($\gamma_c$) of antipodal for both the Chuquichambi I and II sections, providing positive reversal tests for each (Class A; McFadden and McElhinny 1990). Figure 7c suggests a (26–27) ± $6^\circ$ counterclockwise rotation of the Chuquichambi I section, whereas the Chuquichambi II section exhibits a counterclockwise rotation of $22 \pm 7^\circ$. This magnitude of rotation agrees with that reported by Roperch et al. (2000) for strata located c. 26 km to the NW of the Chuquichambi section.

Our results show mean inclinations between 32 and 34° in the Chuquichambi I section and 29 and 38° in the Chuquichambi II section. Expected inclinations for Amazonia craton rocks in the 20–30 Ma range are $-40 \text{ to } -43^\circ$. We assume that the difference is due to inclination shallowing during compaction of the strata.

**Detrital zircon U–Pb geochronology results**

The detrital zircon U–Pb age spectra show the characteristic modes identified in Cenozoic, Mesozoic and Paleozoic samples from the central Andes. As a result of the extensive post-Paleozoic recycling identified by Sundell et al. (2019a), most sources share similar age modes, albeit in slightly differing proportions. These differences are best approached using quantitative or modelling approaches, which take advantage of the relative magnitude of age modes for sample intercomparison or source identification. These include age modes <200 Ma that are associated with subduction-related volcanism. Age modes from 200 to 300 Ma are associated with Permo-Triassic extension-related volcanism and basin formation in Peru (e.g. the Mitu Formation; Spikings et al. 2016) and related intrusions in Bolivia (Jiménez and López-Velásquez 2003, 2008) as well as Permo-Triassic plutons in northern Chile (Munizaga et al. 2008; Coloma et al. 2017).

Zircons of these ages were widely dispersed in the Mesozoic and are present in the Jurassic–Lower Cretaceous strata of the Western Cordillera (Fig. 8). The merged sample also includes a broad age mode at 400–700 Ma that likely includes contributions of 400–500 Ma ‘Famatinian’ zircons and 500–700 Ma ‘Pampean-Braziliano’ zircons. These orogenic events are observed as separate age modes in some of the potential sources, such as Ordovician (the Sandia and Amatura formations) and Silurian–Carboniferous strata in the Eastern Cordillera (Fig. 8). The 1200–900 Ma (‘Sunsas/Grenville’) age mode is subdued in most samples, except for the stratigraphically lowest sample (290618-02) and one potential source: Ordovician strata (the Calapuja and San Gabon formations) from the Altiplano. There are few zircons older than 1500 Ma and these zircons likely reflect ultimate derivation from
Paleoproterozoic cratonic sources recycled from Paleozoic and/or Mesozoic strata (Sundell et al. 2019a).

Mixture modelling indicates that the lowermost sample is consistent with derivation from Paleozoic (Ordovician) strata of the Altiplano (Fig. 8c). The majority of the next eight samples in stratigraphic order can be accounted for by Paleozoic and Mesozoic sources from the Western Cordillera. There is a marked increase in the proportion of zircon ages consistent with Paleozoic sources from the Western Cordillera between c. 35 and 29 Ma (Fig. 8c). After 28 Ma, there is an abrupt increase in the proportion of zircon ages consistent with derivation from Paleozoic strata in the Eastern Cordillera. Samples with <100 ages show a wide range in model results and often significantly differ from stratigraphically adjacent samples.

The MDAs exhibit a generally upward younging trend, but there are both significant and minor deviations from this trend (Fig. 5). Samples with <100 grain analyses per sample deviate significantly from the true depositional age (TDA), which is based on magnetostratigraphy. No sample with fewer than 300 grain analyses per sample yields Cenozoic MDAs. We therefore focus on samples with >300 grain analyses per sample. The MDA of the lowermost sample (35.4 ± 0.5 Ma, 290619-02, below the measured section, all uncertainties reported at 2σ) is close to the age of the base of the section (36.7 Ma). However, MDAs from the next three samples increase upwards from 38.0 ± 1.8 to 39.5 ± 1.7 Ma. Even without consideration of the magnetostratigraphically determined TDA, these MDAs can be discounted as approaching the TDA due to the increase in age up-section. The next Cenozoic MDA (35.1 ± 1.4 Ma, 300618-04) is slightly greater than the TDA (33.1 Ma). However, the following Cenozoic MDA (27.2 ± 1.6 Ma, 300618-05) is slightly younger than the TDA (29.4 Ma) and the following two MDAs (29.0 ± 0.6 and 23.0 ± 0.8 Ma) are slightly older than their TDAs (27.5 and 20.4 Ma).

**Flexural modelling results**

Flexural modelling of both the >25 and >18 Ma intervals constrained by palaeoelevation estimates underestimate the stratigraphic thickness by 1.7–2.2 km (Fig. 4a, b). Relaxing the palaeoelevation constraints yields models that match the sediment accumulation at 25 Ma at Chuquichambi with loads of 3.0 and 3.6 km in the Western and Eastern cordilleras, respectively (Fig. 4c). This model can also match the observed sedimentary thickness at 18 Ma with loads of 6 and 5.3 km in the Western and Eastern cordilleras, respectively (Fig. 4d). Decreasing the distance to the Western and Eastern cordilleran loads to 25 km increases the subsidence at the Corque Syncline location, but the model still underestimates sediment thickness at 25 Ma by c. 0.6 km (Fig. 4c). However, the 18 Ma model with a 25 km spacing yields sediment accumulation that matches the observed sediment accumulation with load heights of 4.2 and 2.9 km for the Western and Eastern cordilleras, respectively.
Fig. 8. Detrital zircon U–Pb geochronology results shown as kernel density estimate plots. All the kernel density estimates were calculated with a fixed 15 myr bandwidth. (a) Detrital zircon U–Pb results from sandstone outcrops collected along the Chuquichambi measured section; \( n \) = number of ages <200 Ma; \( n_{\text{total}} \) = number of ages for the entire sample. (b) Detrital zircon U–Pb data from potential source samples. (c) Cumulative contributions from the potential sources shown in part (b) to the samples shown in part (a) based on Monte Carlo mixture modelling using DZmix. The ages of samples are indicated by Roman numerals below the graph, which correspond to the samples in part (a). Data for this figure are available in Supplemental Table S5.
Discussion

Maximum depositional ages

The observation that samples with <100 grain analyses did not yield any Cenozoic MDAs indicates that the syn-depositional ages comprise a relatively small component of the overall zircon population for Chuquichambi strata (Vermeesch 2004; Sharman and Malkowski 2020). The MDAs that approach the TDA in samples with >100 grain analyses provide a confirming case study of research that advocates increasing sample size to identify small age components and to fully characterize the relative proportions of age modes (Vermeesch 2004; Pullen et al. 2014; Saylor and Sundell 2016; Coutts et al. 2019; Sundell et al. 2020). Nevertheless, the observation that only 50% of the high-number samples yielded ages that are compatible with TDAs provides a cautionary note and indicates the need for multiple high-number samples to characterize the depositional history, even in this retroarc setting.

Subsidence and sediment accumulation

Detrital zircon MDAs generally young up-section, confirming that there is no kilometre-scale thrust repetition of the stratigraphic section. This conclusion is confirmed by the magnetostratigraphic results, which further refine the age model for this thick stratigraphic section. Sedimentation in the combined Chuquichambi palaeomagnetic section occurred from the late Eocene to early Miocene (36.5 to c. 19 Ma). At the latitude of our sections, an additional c. 5 km of overlying strata remains undated and we estimate the age at the axis of the syncline at this latitude to be c. 6.2 Ma based on the sedimentation rates at the top of the Chuquichambi II section (see Supplemental Text for calculations). We suggest that this is a minimum age because of the increased sedimentation rate during uplift of the Eastern Cordillera (Fig. 6) (Roperch et al. 1999; Garzione et al. 2006).

We therefore refine the Miocene portion of the record by incorporating the sediment accumulation record from the Callapa stratigraphic section near the centre of the Corque Syncline (Roperch et al. 1999; Garzione et al. 2006). A c. 5 km thick composite mid-Miocene magnetostratigraphic section, with an age range of 14–9 Ma, is reported from the Totora Formation (Roperch et al. 1999), c. 78 km to the NW of our sites, along the axis of the Corque Syncline. Magnetostratigraphy and isotopic ages of tuffs provide age constraints for an additional c. 1.2 km between 9 and 5.3 Ma (Marshall et al. 1992; Roperch et al. 1999; Garzione et al. 2006). The 0.4 km (myr)^{−1} calculated for the Chuquichambi II section is less than the sediment accumulation rate of 0.97 km (myr)^{−1} reported by Roperch et al. (1999). To incorporate the Miocene record, we project the 0.4 km (myr)^{−1} sediment accumulation rate from the top of the Chuquichambi section (18.7 Ma) to the base of the section reported by Roperch et al. (1999) at 14.2 Ma. We then added the stratigraphic thickness reported by Roperch et al. (1999) and Garzione et al. (2006) (Fig. 6a).

The subsidence record shown by incorporating both Paleogene and Neogene sediment accumulation records is inconsistent with the typical convex-upward subsidence history associated with advancing foreland basins (Nie and Heller 2009; Sundell et al. 2018). Nevertheless, it shares characteristics with other hinterland basin subsidence histories from across the Altiplano Basin. For example, sediment accumulation at both Corque Syncline and Ayaviri (Fig. 9) began in the late Eocene (39–36 Ma; Fig. 6). They both show an initially linear subsidence pattern, with Ayaviri showing some minor accelerations and decelerations in subsidence. By contrast, these two locations share relatively few characteristics with the sediment accumulation record at Cusco (Sundell et al. 2018). At Cusco, sediment accumulation begins in the Paleocene and does show the convex-upward subsidence pattern typical of an advancing foreland basin.

Like the subsidence record, the chronology of deposition indicates that the deposition of strata in the Corque Syncline did not occur in a classic retroarc foreland basin setting. Whereas sediment accumulation at Chuquichambi section begins at 36 Ma, exhumation in the Eastern Cordillera at 18° S was ongoing by 40–36 Ma (Barnes et al. 2012; Eichelberger et al. 2013; Buford Parks and McQuarrie 2019). This indicates that the Corque region had orogenic loads both to its east and west, a scenario that is not consistent with a classic foreland basin (cf. DeCelles and Giles 1996). Rather, the Corque and Ayaviri depocentres were hinterland basins throughout their Eocene–Modern history.

Rotations

Palaeomagnetic rotations occurred prior to the onset of exhumation of the Corque Syncline, implicating a larger scale tectonic process in the rotations. Figure 7 suggests a 26–27° ± 6° counterclockwise rotation of the Chuquichambi I section, whereas the Chuquichambi II section shows a counterclockwise rotation of 22 ± 7°, suggesting that rotation was in progress during deposition. Counterclockwise rotation is also reported in the Totora Formation (Roperch et al. 1999) further to the north, but still within the Corque Syncline. They report a 10.8 ± 2.9° counterclockwise rotation, concluding that the c. 8–14° rotation occurred in the last 9 myr. MacFadden et al. (1985) reported a c. 540 m thick upper Oligocene magnetostratigraphic section from the Salla–Luribay Basin located c. 95 km to the NNE of Chuquichambi. They date the Salla strata between 28.5 and 24 Ma and report no apparent rotation of the vertical axis.

Calculating the rotation for the reversed sample at 2 myr intervals reveals a linear decrease in rotation towards the top of the section. The reversed sample is used because, in all cases: (1) it contains more sites than the normal sample; and (2) the α95 of the reversed sample is less than that of the normal sample. Our data show a considerable counterclockwise rotation that we suggest began between 32 and 30 Ma. This coincides with a time frame during which the rate of sediment accumulation was increasing in the Chuquichambi section. Assuming that rotation occurred at the same rate for the section reported by Roperch et al. (1999) and the Chuquichambi I and II sections allows us to plot the regional vertical axis rotational history (Fig. 7c). Projecting a line to the upper boundary α95 value at 32 Ma and another through the lower boundary α95 value at 9 Ma creates a wedge of statistically significant data for all points. The area is defined by the slope of the line that bisects the wedge at 0.8° (myr)^{−1} ± 10%. This rotation history is consistent with previous research (Roussé et al. 2005; Arrigada et al. 2006b, 2008; Roperch et al. 2006, 2011; Eichelberger and McQuarrie 2015). Roperch et al. (1999, 2000) also conclude that the magnitude of counterclockwise rotation decreases with the distance from the axis of the bend in the Corque Syncline with counterclockwise rotation to the NW and clockwise rotation to the south. Based on the trend observed in Figure 7c, it is likely that the degree of rotation since 9 Ma in our area is consistent with that reported by Roperch et al. (1999). The majority of the deformation of the Corque Syncline occurred after 10 Ma (Lamb 2011). Hence rotation cannot be accounted for by local deformation prior to the late Miocene. We therefore follow previous studies in attributing at least the Paleogene–early Neogene portion of the record to the development of the Bolivian orocline.

Sediment provenance

New detrital zircon-based sediment provenance data are compatible with, but refine, previous interpretations of palaeoflow and sandstone petrography (Horton et al. 2001, 2002). Recognizing
that samples with <100 analyses yield highly variable results, we focus the following discussion on samples with >300 analyses per sample. The new data suggest that (1) the upper Eocene strata were locally derived from Proterozoic basement or Paleozoic strata in the Altiplano and (2) the uppermost Eocene–Oligocene strata document an unroofing sequence in the Western Cordillera. Samples from the base of the section to c. 4 km are consistent with a mixture of Proterozoic basement or Paleozoic strata and Mesozoic sources present in the Western Cordillera (Fig. 8c). This is compatible with eastwards-directed palaeocurrent orientations and sandstone compositions that include volcanic, sedimentary and basement (granite/gneiss) fragments (Horton et al. 2002) potentially derived from the Berenguela area of Peru (Tosdal 1996) or the Arequipa–Antofalla Massif sampled in the San Andres drillcore in Bolivia (Lehmann 1978).

The lowermost mid-Eocene sample (290618-02) is dominated by a c. 1 Ga (Grenville) age mode, which is consistent with derivation either from the basement of the Arequipa–Antofalla Massif or Ordovician strata, which were themselves derived from basement sources. The 1 Ga ages become diluted up-section, but are present in Oligocene strata on the western side of the Corque Syncline (Wotzlaw et al. 2011), suggesting that by the Oligocene the basin was receiving sediment from multiple sources and that a variety of structural levels was exposed in the Western Cordillera. Upper Eocene strata at Chuqichambi are dominated by sources consistent with Jurassic–Cretaceous strata of the Western Cordillera (Fig. 8c) and the relative proportion of zircons associated with Devonian strata (Bahlburg et al. 2009) increases up-section, consistent with an unroofing sequence in the Western Cordillera.

U–Pb age spectra from samples above 4 km stratigraphic height are entirely consistent with recycling of Paleozoic strata from the Eastern Cordillera (Fig. 8c). This is also consistent with a switch in palaeocurrent orientation from eastwards-directed to westwards-directed and an increase in quartzose, sedimentary and low-grade...
metamorphic sand grains in the sandstones (see Horton et al. 2001, 2002 for detailed palaeocurrent and sandstone petrology data). The new U–Pb data are therefore consistent with previous interpretations indicating an Eastern Cordilleran source for Oligocene–lower Miocene strata, including the upper Potoco, Totora, Luribay and Bolívar formations (Horton et al. 2002; Leier et al. 2010).

**Tectonic drivers of subsidence**

We consider five possible mechanisms for basin subsidence given the evidence for pre-existing and ongoing exhumation to both the east and west of the Altiplano Basin at the onset of sediment accumulation in the Eocene. These include: (1) flexural subsidence due to loading by the Western and Eastern cordilleras; (2) normal faulting in the Altiplano; (3) sediment ponding behind topographic barriers; (4) dynamic subsidence related to the emplacement of the flat slab beneath South America; and (5) dynamic subsidence related to the removal of a subcrustal flat slab or the removal of dense lithospheric material in response to a Rayleigh–Taylor instability.

**Flexural subsidence**

Flexural modelling suggests that orogenic loading can explain some, but not all, of the sediment accumulation in the Altiplano. When considering all available independent constraints, flexural subsidence underestimates sediment accumulation from 36 to 25 Ma by 2.2 km and from 36 to 18 Ma by 1.7 km (Fig. 4a, b). If palaeoelevation constraints are not included, orogenic loading by the Western and Eastern cordilleras alone can account for all the subsidence observed at Corque (Fig. 4c, d). Nevertheless, we consider the palaeoelevation-constrained subsidence to be minimum underestimates for the following four reasons.

First, maximum load heights of 2.25 and 1.75 km were set slightly above the maximum palaeoelevation estimates of 2 and 1.5 km for the Western and Eastern cordilleras at 15.5–17 and 17° S, respectively (Leier et al. 2013; Saylor and Horton 2014; Sundell et al. 2019b). Second, the palaeoelevation estimates are absolute estimates, whereas topographic loading is based on the difference in elevation between the basin and the orogenic load. Palaeoelevation estimates suggest that the Altiplano Basin was at c. 1 ± 1 km prior to the late Miocene (Gregory-Wodzicki et al. 1998; Gregory-Wodzicki 2000; Garzione et al. 2006; Ghosh et al. 2006; Garzione et al. 2008; Bershaw et al. 2010), meaning that the palaeoelevation estimates from the Western and Eastern cordilleras may significantly overestimate the effective orogenic load. Third, the applied EET (20 km) is at the upper end of the range of estimated EETs (Stewart and Watts 1997; Tassara 2005; Pérez-Gussinyé et al. 2007; Tassara et al. 2007; Sacek and Ussami 2009). Fourth, the loads are as close as possible based on palinspastically restored distances from balanced cross-sections (McQuarrie 2002; Buford Parks and McQuarrie 2019). These loads are based on deformation that is not apparent (e.g. as growth strata) in the stratigraphic record and we conclude that the loads were probably further from the depocentre than indicated in our model.

In all these elements we have taken a conservative approach that maximizes the subsidence at Corque due to flexure. Nevertheless, even with this conservative approach we are unable to match the observed stratigraphic thicknesses.

Our approach differs from several previous studies that have concluded that flexure alone can account for the extreme thicknesses of Paleogene subsidence observed in the Corque Basin (Hampton 2002; Buford Parks and McQuarrie 2019). We adopt a maximum EET of 20 km, consistent with those observed in the Andes as well as cordilleran-type orogens globally (Stewart and Watts 1997; Tassara 2005; Pérez-Gussinyé et al. 2007; Tassara et al. 2007; Sacek and Ussami 2009). Increasing the EET allows the transmission of topographic loading over a greater lateral distance, thereby increasing the constructive interference between the Western and Eastern cordilleran loads and increasing subsidence in the intervening basin. Buford Parks and McQuarrie (2019) achieved thicknesses approaching those observed at Chuquicamati, but doing so required an EET of 100 km for the Altiplano. For comparison, the EET for the Paleoproterozoic–Archean Amazonian craton ranges from 70 to 100 km (Pérez-Gussinyé et al. 2007; Tassara et al. 2007). Whereas Hampton (2002) considered subsidence directly adjacent to the modelled topographic load, we model the load as displaced from our locus of observation (Chuquicamati) by several tens of kilometres. This approach is consistent with the observed and modelled loads and most accurately represents the flexurally induced subsidence at the Chuquicamati location.

**Normal faulting**

Normal faulting in the Altiplano is not a plausible mechanism for the observed excess subsidence. Some researchers have attributed the variations in stratigraphic thickness or deformation patterns in the central Andes to strike-slip or normal faulting (Hénal et al. 1996; Rochat et al. 1996, 1998, 1999, 2000; Baby et al. 1997; Elger et al. 2005). However, these differences in stratigraphic thickness can be explained by the large-scale telescoping of a basin characterized by gradual changes in stratigraphic thickness, rather than the low shortening inversion of a basin characterized by rapid changes in stratigraphic thickness across a normal fault (e.g. Lamb and Hoke 1997; Horton et al. 2001; McQuarrie 2002; Buford Parks and McQuarrie 2019).

The stratigraphic stacking patterns are inconsistent with standard models of rift-related sedimentation. Idealized rift-fill sequences are characterized by upwards-fining followed by upwards-coarsening stratigraphic trends. By contrast, the interpreted seismic reflection data and regional correlations suggest that the evaporites at the base of the Chuquicamati section represent the base of the Cenozoic section (Rochat et al. 1996; Horton et al. 2001) and the section monotonically coarsens upwards (Fig. 3b). The fine-grained lithofacies that characterize the base of the stratigraphic section and the east-directed palaeocurrents of the lower 4 km are inconsistent with deposition in the hanging wall proximal to an active normal fault to the east. Seismic reflection data do not show fanning dips in the hanging wall of the Corque Thrust as would be expected if it had experienced a history of normal motion and sediment accumulation in a half-graben (Lamb and Hoke 1997, their fig. 8a). Inversion of a half-graben is incompatible with the geometry of the Corque Syncline (McQuarrie and DeCelles 2001).

We conclude that normal faulting is unlikely to account for the excess stratigraphic thickness observed in the Chuquicamati section.

**Sediment accumulation due to topographic ponding**

The sedimentary lithofacies are also inconsistent with deposition in an internally drained closed basin. The upper Eocene–lower Miocene sequence lacks significant thicknesses of lacustrine or evaporite deposits, as seen in other hydrologically closed basins, such as the Eocene Green River Basin in the North American Laramide province (Carroll and Bohacs 1999; Smith et al. 2014). Lacustrine and evaporitic conditions are documented elsewhere in the Altiplano–Puna plateau (Saylor and Horton 2014; Quade et al. 2015; Kat et al. 2016; Eden et al. 2020). However, those stratigraphic intervals are typically <1 km thick and so cannot account for the >1.7 km of excess sediment accumulation observed at Corque.
Dynamic subsidence due to flat-slab subduction

The excess subsidence may be the result of flat-slab subduction beneath the Peruvian and Bolivian Altiplano. Flat-slab subduction can produce dynamic subsidence in front of the downgoing portion of the slab and uplift over its crest (e.g. Dávila et al. 2010; Jadamec et al. 2013; Eakin et al. 2014). As the flat slab migrates beneath the overriding plate, the zone of dynamic subsidence will migrate with the forward-migrating hinge (Mitrovica et al. 1989; Liu and Gurnis 2010; Heller and Liu 2016). In numerical models, the magnitude of surface deflection depends on the model parameters, including the relative viscosity contrast between the layers, the rheology of the lithosphere, the density and geometry of the slab, and coupling between the subducting and overriding plates (Billen et al. 2003; Eakin et al. 2014). As a result, the magnitude of the predicted subsidence can range from <1 to 3 km (Gurnis 1993; Wheeler and White 2000; Jadamec et al. 2013; Molnar et al. 2015; Heller and Liu 2016). Nevertheless, a consensus is emerging that 0.5–1 km of dynamic subsidence is consistent with observations from nature (Braun 2010; Flamant et al. 2013; Liu 2015; Heller and Liu 2016; Eakin and Lithgow-Bertelloni 2018). In this scenario, flat-slab subduction-related dynamic subsidence and flexural subsidence due to orogenic loading by the Eastern and Western Cordilleras would constructively interfere to produce the observed total subsidence.

Subsidence due to removal of upper plate lithosphere

Numerical and analogue modelling indicates that the formation and culmination of a dense lithospheric root from the upper plate can also drive subsidence (Molnar and Houseman 2004; Göğüş and Pysklywec 2008a; Wang et al. 2015). The rate of tectonic subsidence in the Arizaro Basin on the Puna Plateau (northwesternmost Argentina) is similar to that observed in both Ayaviri and Corque (Fig. 6b) (DeCelles et al. 2015). Furthermore, the concave upwards subsidence curve at Arizaro is similar to that at Ayaviri (Fig. 6b). Nevertheless, we refrain from attributing late Eocene–early Miocene subsidence at Ayaviri and Corque to lithospheric ‘dripping’ because of the abundant evidence for a flat slab beneath the Altiplano during the late Eocene–Oligocene and the northern Puna plateau in the early Miocene.

A flat slab would have prevented the detachment of a lithospheric root. However, tearing and southwards migration of the flat slab in the late Oligocene–early Miocene may have stranded a portion of the flat slab in the early Miocene, or the associated magmatism may have formed an eclogitic or restitic root in the mid- to late Miocene (e.g. Currie et al. 2015). A dense stranded or newly formed root may account for the continued and accelerated subsidence through the Neogene observed at Corque and Arizaro (Wang et al. 2015; DeCelles et al. 2015; Currie et al. 2015) and may have set the stage for foundering of the lithospheric root and associated surface uplift between 10 and 6 Ma (Garrison et al. 2006, 2008; Ghosth et al. 2006; Pingel et al. 2020).

Paleogene flat-slab subduction in the central Andes

We integrate new and published data to outline a scenario in which a southwards migrating flat slab drove basin evolution, magmatism and deformation in southern Peru–northern Argentina (c. 14–25°S). Flat-slab subduction migrated from north to south, starting in southern Peru in the early Eocene (50–42 Ma) and migrating to northern Bolivia by the Oligocene (c. 30 Ma) (Sandeman et al. 1995; James and Sacks 1999; Perelló et al. 2003; Jiménez et al. 2009; Ramos 2018). In southern Peru, slab flattening is recorded as an inboard migration of magmatism between the development of the Toquepala Arc (91–45 Ma) in the Western Cordillera and the Andahuaylas–Anta Arc (45–30 Ma) in the Altiplano and Eastern Cordillera (Bissig et al. 2008; Mamani et al. 2010).

Prior to the initiation of the Andahuaylas–Anta Arc, there was a magmatic lull in the Western Cordillera extending from c. 14 to 19°S from c. 55 to 45 Ma (Central Andean Geochronology Database, including data compiled by Mamani et al. 2008, 2010; Pilger 2020). The zone of flat-slab subduction widened to the south and arrived in central Bolivia by the late Eocene–early Oligocene based on the record of magmatism and magmatic lulls in the central Andes. This was followed by a magmatic lull in the Chilean magmatic arc, which has been linked to an episode of flat-slab subduction from c. 37 to 26 Ma (Haschke et al. 2002, 2006). The record of magmatism associated with flat-slab subduction in Bolivia includes batholiths emplaced in the Eastern Cordillera in the latest Eocene–Oligocene (c. 34–25 Ma) (McBride et al. 1983; Kennan et al. 1995; Gillis et al. 2006; Jiménez and López-Velasquez 2008).

Slab re-steepening began in Peru at c. 31 Ma and migrated southwards to northern Chile and Bolivia by c. 25 Ma. Initial re-steepening in southern Peru at c. 31 Ma is documented by mantle-derived melts of the Tacaza Group in the Western Cordillera (Clark et al. 1990). The lower Tacaza Group is dominated by mafic shoshonitic volcanism, whereas the upper Tacaza Group (<26 Ma) is characterized by a mixture of shoshonitic and calc-alkaline magmatism (Wasteneys 1990). Continued re-steepening produced a mixture of mantle-derived and peraluminous melts of the Picotani Group (c. 25–22 Ma) in the Eastern Cordillera (Sandeman et al. 1995). Later Miocene magmatism is dominated by rhyolitic crustal melts of the Quenamari Group (Pichavant et al. 1988a, b).

Bimodal magmatism with both mantle- and crustal-derived components in the Eastern Cordillera has long been attributed to flat-slab subduction (Sandeman et al. 1995; Mlynarczyk and Williams-Jones 2005; Hoke and Lamb 2007; Mamani et al. 2010). Continued re-steepening of the slab is documented by expansion of the zone of bimodal volcanism across the Peruvian and Bolivian Altiplano by 25 Ma, followed by persistent widespread volcanism through the Neogene (Haschke et al. 2006; Trumbull et al. 2006; Hoke and Lamb 2007).

The initiation of flat-slab subduction is marked by exhumation in the Eastern Cordillera and, like the record of magmatism, documents a southwards migration in exhumation. Exhumation initiated in the southern Peruvian Eastern Cordillera in the mid- to late Eocene and migrated to the southern Bolivian Eastern Cordillera by the late Eocene–early Oligocene. K–Ar and 40Ar/39Ar data point to cooling at 49–37 Ma in the Eastern Cordillera of southern Peru between 13.5 and 14.5°S (Clark et al. 1990; Kontak et al. 1990).

Apatite fission track data from northern Bolivia at 15–17°S, interpreted along with He/He modelling and balanced structural cross-sections, yields a range of ages from 50 to 30 Ma (McQuarrie et al. 2008; Barnes et al. 2012). Thermokinematic modelling of the balanced cross-section yields better fits for the initiation of exhumation between 50 and 40 Ma (Bufford Parks and McQuarrie 2019). Zircon fission track, K–Ar and 40Ar/39Ar data point to exhumation in the Eastern Cordillera of northern Bolivia at 16–17°S between 45 and 40 Ma (Benjamin et al. 1987; McBride et al. 1987; Farrar et al. 1988; Gillis et al. 2006). In central–southern Bolivia (18–21°S), apatite fission track data, in some cases coupled with He/He modelling, yield an onset of exhumation between 41 and 30 Ma, with a concentration of ages between 36 and 30 Ma (Scheuber et al. 2006; Ege et al. 2007; Barnes et al. 2008; Eichelerberger et al. 2013).

Provenance, palaeocurrent and growth strata relationships in the Oligocene Luribay Formation (Bolivia; Leier et al. 2010) and Puno Group (Peru; Perez and Horton 2014) confirm that deformation in the Eastern Cordillera adjacent to each of these locations persisted into the Oligocene. Deformation migrated from the Peruvian Eastern Cordillera to the Interandean zone and Altiplano in the
late Oligocene–early Miocene following expansion of the zone of volcanism (Sempere et al. 1990; McQuarrie 2002; Perez and Horton 2014). Deformation migrated eastwards and westwards from the Bolivian Eastern Cordillera later, with deformation in the Bolivian Altiplano between the deposition of the Oligocene Salla Beds (MacFadden et al. 1985; McRae 1990; Kay et al. 1998) and the Pliocene Umal Formation (Marshall et al. 1992).

All of this suggests that the region of the Altiplano should have experienced dynamic uplift during flattening of the slab in the early Oligocene, followed by dynamic subsidence at the frontal edge of the flat slab in the late Oligocene–Oligocene (Fig. 9a, b) (Dávila and Lithgow-Bertelloni 2015). The timing and magnitude of dynamic subsidence due to the emplacement and removal of a flat slab beneath the Altiplano is consistent with the observed subsidence at Corque Syncline and Ayaviri. Initial flattening of the slab is predicted to be associated with uplift of the overriding plate (Eakin et al. 2014; Dávila and Lithgow-Bertelloni 2015; Axen et al. 2018). This is consistent with the extended deformation variability accompanied by palaeosol development or deposition in evaporite-rich, low-energy, non-marine environments in the late Paleocene to mid-Eocene sometimes associated with forebulge development (Horton et al. 2001, 2015; Perez and Horton 2014).

When subsidence began at Ayaviri and Corque, it was synchronous along-strike for at least 500 km from Ayaviri Basin (14.8° S; Perez and Levine 2020) to Chuquichambi (18° S) (Fig. 6b), suggesting a long-wavelength process consistent with subsidence in front of the flat slab (Dávila and Lithgow-Bertelloni 2015). Neither Ayaviri or Corque show the convex-upwards tectonic subsidence pattern observed at Cusco that is typical of flexural basins (Fig. 6b) (Xie and Heller 2009). The 9 myr interval in which excess sediment accumulation in the Corque depozone is first observed (36–25 Ma) overlaps with the migration of a flat slab into the Bolivian central Andes (Ramos and Folguera 2009; Ramos et al. 2018). Dynamic subsidence produces a deflection that tapers both into the Bolivian central Andes (Ramos and Folguera 2009; Ramos and Pysklywec 2008) and into the Eastern Cordillera of NW Argentina (Clark et al. 1990; Sandeman et al. 1995; Hoke and Lamb 2007; Mamani et al. 2010). After c. 25 Ma, magmatism is widespread across the Altiplano and consists of initially mafic mantle-derived melts that give way to more evolved and crustal-derived melts (Soler and Jiménez 1993; Sandeman et al. 1995; Mamani et al. 2010). This has been linked to the removal of a flat slab followed by either detachment of the mantle from the upper plate (Hoke and Lamb 2007) or melting of the now hydrated lithospheric mantle of the upper plate (e.g. Kay and Coira 2009). Nevertheless, there is a paucity of magmatism in the Bolivian Altiplano between c. 25 and 10 Ma (Fig. 9d). We suggest that this may reflect the stranded portion of the slab, which both limited magmatism in this region and drove continued subsidence in the Corque Syncline.

Full re-steepening of the slab may not have occurred until the mid-Miocene at latitudes south of Corque (e.g. Kay and Coira 2009) and hence the slab may have continued to drive subsidence at these southern latitudes during the early Miocene, as observed at Corque and Arizaro. Minor late Eocene exhumation across the Puna plateau and into the Eastern Cordillera of NW Argentina (c. 23–25° S; Kraemer et al. 1999; Mpodozis et al. 2005; Arriagada et al. 2006a; Hong et al. 2007; Carrapa and DeCelles 2008) is synchronous with the initiation of flat-slab subduction to the north and may be driven by the initiation of plate-wide contractional stresses, as indicated by the numerical models of Axen et al. (2018). Nevertheless, deformation waned and the region was integrated into a regional retroarc foreland basin (Carrapa et al. 2011; DeCelles et al. 2011).

When exhumation recommenced in the late Oligocene–early Miocene, it was focused in the eastern Puna plateau and Eastern Cordillera (Carrapa et al. 2005, 2011; Coutand et al. 2006; DeCelles et al. 2011) and is synchronous with a magmatic lull (Kay and Coira 2009). This late Oligocene–early Miocene exhumation and magmatic lull has been attributed to flat-slab subduction and is consistent with the southwards migration of flat-slab subduction described here and is consistent with the model proposed by Kay and Coira (2009) and Ramos (2018). The model proposed here refines these earlier models by integrating the basin, deformation and magmatic history from southern Peru to NW Argentina.

Summary

Subsidence at both Ayaviri and Corque was controlled by the emplacement and removal of the flat slab in addition to orogenic loading. Uplift associated with initial flattening of the slab produced a regional disconformity observed at the base of both sections. Both Ayaviri and Corque experienced dynamic subsidence associated with mantle convection in front of the flat portion of the oceanic plate. Passage of the hinge between the shallowly and steeply subducting portions of the slab at Ayaviri may have resulted in brief uplift over the crest of the flat slab. However, Corque remained above the downgoing hinge and so experienced continued subsidence. Initial tearing and rollback of the flat slab may have
driven a brief pulse of subsidence at Ayaviri between 30 and 25 Ma, but ultimately upwelling warm asthenosphere provided support and minimized further subsidence at Ayaviri and Cusco. The weight of the stranded slab drove continued subsidence at Corque into the Neogene. Mid- to late Miocene subsidence at Corque may have been driven by the formation of a dense lithospheric root, the detachment of which in the late Miocene led to the decrease in subsidence rate and dramatic surface uplift observed at Corque. The range of responses of the upper plate reflects the range of geodynamic forcing mechanisms and the complex setting produced by flat-slab subduction.

Conclusions

We present here new magnetostratigraphy, bio- 

otic 40Ar/39Ar ages, detrital zircon U–Pb geochronology and palynology from the 7.4 km thick Chuquicamati stratigraphic section in the central Corque Syncline, which consists of the Potocoto and lower Totora formations. The new chronology demonstrates that the section is continuous and that sediment accumulation extended from 36.7 to 18.7 Ma. Deposition is unbroken by significant hiatuses. Integration of this dataset with nearby datasets indicates that sedimentation was probably continuous from the Eocene to late Miocene.

The onset of depositional events the onset of exhumation in both the Western and Eastern cordilleras to the west and east, respectively, of the study location. This rules out a classic retroareal foreland basin depositional setting for the Potocoto or Totora formation. Flexural modelling indicates that loads consistent with palaeoelevation estimates and constrained by retrodeformed balanced structural cross-sections are insufficient to account for the magnitude of subsidence at Chuquicamati. Alternative mechanisms, such as normal faulting or the ponding of sediments behind the growing topography, are inconsistent with lithofacies, stratigraphic stacking patterns and the thickness of the accumulated sediment.

Rotation data from the Chuquicamati section indicate a linear counterclockwise rotation of 0.8° (myr)^{-1}. This has produced c. 30° of counterclockwise rotation since c. 32 Ma. We attribute the Paleogene–early Neogene rotation to the development of the Bolivian orocline. The mid Eocene–Pliocene portion of the record may be due to either the continued rotation of the Bolivian orocline or local exhumation of the Corque Syncline.

Detrital zircon provenance data are consistent with an unroofing sequence in which Mesozoic clastic sources yield to Paleozoic sources between 36 and 28 Ma. After c. 28 Ma, detrital zircon U–Pb age spectra are entirely consistent with derivation from Paleozoic sources. Integration with previously published palaeocurrent data indicates that the initial unroofing occurred in the Western Cordillera and the transition to exclusively Paleozoic sources after 28 Ma coincided with a transition to Eastern Cordillera sources. Integration of the magmatic, basin and deformation history of the central Andes provides a coherent model of the effects of flat-slab subduction.

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