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Cenozoic uplift of the Central Andes in northern Chile and Bolivia – reconciling paleoaltimetry with the geological evolution.

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Abstract.

The Cenozoic geological evolution of the Central Andes, along two transects between ~17.5°S and 21°S, is compared with paleo-topography, determined from published paleo-altimetry studies. Surface and rock uplift are quantified using simple 2-D models of crustal shortening and thickening, together with estimates of sedimentation, erosion and magmatic addition. Prior to ~25 Ma, during a phase of amagmatic flat-slab subduction, thick skinned crustal shortening and thickening (nominal age of initiation ~40 Ma) was focused in the Eastern and Western Cordilleras, separated by a broad basin up to 300 km wide and close to sea level, which today comprises the high Altiplano. Surface topography at this time in the Altiplano and the western margin of the Eastern Cordillera appears to be ~1 km lower than anticipated from crustal thickening, which may be due to the pull-down effect of the subducted slab, coupled to the overlying lithosphere by a cold mantle wedge. Oligocene steepening of the subducted slab is indicated by the initiation of the volcanic arc at ~27 - 25 Ma, and widespread mafic volcanism in the Altiplano between 25 and 20 Ma. This may have resulted in detachment of mantle lithosphere and possibly dense lower crust, triggering 1 – 1.5 km of rapid uplift (over << 5 Myrs) of the Altiplano and western margin of the Eastern Cordillera and establishing the present day lithospheric structure beneath the high Andes. Since ~25 Ma, surface uplift has been the direct result of crustal shortening and thickening, locally modified by the effects of erosion, sedimentation and magmatic addition from the mantle. The rate of crustal shortening and thickening varies with location and time, with two episodes of rapid shortening in the Altiplano, lasting < 5 Myrs, that are superimposed on a long term history of ductile shortening in the lower crust, driven by underthrusting of the Brazilian Shield on the eastern margin.

Keywords: Central Andes, Cenozoic uplift, Tectonics, Paleoaltimetry, Convergent plate margin.

1. Introduction

The uplift of large and high mountain ranges has a profound influence on our planet. With the advent of paleoaltimetry techniques, using the geomorphology of drainage systems, plant physiognomy, and geochemistry of surface deposits (see Quade et al. 2007), there have been a wealth of studies tracking the evolution of topography during the growth of Tibet, Andes, and parts of western North America. In some cases, these studies have yielded unexpected results, revealing high topography much earlier than was anticipated – for example, the Colorado Plateau (Wolf et al. 1995), Tibet (Garzzone et al. 2000, Spicer et al. 2003, Wang et al. 2008), Argentinian Northern Puna (Canavan et al. 2014). Or very young km-scale uplift – for example, the Bolivian Altiplano (Garzzone et al. 2006, 2008, Ghosh et al. 2006). In any case, the timing of uplift provides important constraints on the dynamical processes that cause it.

In general terms, uplift of the Earth's surface is a result of changes in the vertical density structure of the lithosphere. The largest density contrast here is between the crust and mantle, and so crustal shortening and thickening, driven by stresses that come from the interaction of the tectonic plates, is a powerful mechanism of surface uplift - the uplift will correlate closely with the amount and timing of crustal shortening, on time scales of millions to tens of millions of years. In marked contrast, numerical and theoretical models show that in certain cases an instability in the higher density and fluid-like lower part of the lithosphere can grow slowly during long term lithospheric shortening, then reach a critical stage where it rapidly detaches and sinks into the underlying hotter and lower density asthenosphere, leaving behind a thinned lithospheric mantle; isostatic rebound in the overlying crust could result in several kilometers of surface uplift, depending on the thickness of the mantle root – here, uplift post-dates shortening, and may occur on very short time scales ($<<5$ Ma) (Houseman et al. 1981, Molnar et al. 1993, Molnar and Garzzone 2007, Stern et al. 2013).

Lithospheric detachment has been developed by DeCelles et al. (2009) and Ducea et al. (2011) into a general model for orogenesis, with mantle and mafic lower crustal detachment as an inevitable late stage consequence of lithospheric shortening, thereby maintaining over geological time a relatively thin lithosphere, with an intermediate to felsic crust (Rudnick 1995). However, the role of lithospheric detachment in the uplift of Tibet has become less clear, now that paleoaltimetry shows that southern Tibet has been high since at least 20 Ma, and possibly 40 Ma (Garzione et al. 2000, Spicer et al. 2003, Wang et al. 2008). For this reason, the Central Andes is widely regarded as the best documented example of this mechanism in an actively deforming mountain belt (Garzione et al. 2008).

Following on from Lamb (2011), I test these ideas by integrating the documented Cenozoic geological evolution of the Bolivian Andes, between $\sim 17^{\circ}\text{S}$ and 22°S with published paleo-elevation estimates, in order to determine what processes are responsible for the observed uplift. This analysis also provides insight into the large scale forces in the Earth that create large mountain ranges.

1.1 Central Andes in northern Chile and Bolivia

1.1.1 Tectonic Setting

The Central Andes in northern Chile and Bolivia form a mountain belt up to 700 km wide, reaching elevations greater than 6500 m on the western margin of South America (Fig 1). Offshore, the ~ 50 Ma oceanic Nazca (or Farallon) plate is being subducted beneath the western margin of South America (Pardo-Casas and Molnar, 1987) along the Peru-Chile trench. The present day convergence is ~ 85 mm/yr in an ENE direction, relative to the South American plate (DeMets et al., 1994).

1.1.2 Topography

Figure 1a shows the topography of the Andes between 13°S and 23°S . The general topographic trend of the mountainous region varies from approximately NW in the north to approximately N farther south, defining the Bolivian bend. Two topographic profiles, north and south of the bend, show the main topographic features (Figs. 1b, c). The western margin of the high Andes is overall steeper ($\sim 2^{\circ}$) than the eastern margin ($\sim 1^{\circ}$). There is also a distinct plateau in the central part, at essentially the same

altitude north and south of the Arica bend, so that the topographic profiles have a general flat-topped hat shape.

The central plateau region is referred to as the Altiplano in Bolivia and forms a ~200 km wide region of subdued relief, ~600 km long in Bolivia and at an average altitude of ~3800 m, which has been essentially a sedimentary basin throughout much of the Cenozoic (Fig. 3). However, in detail, there appear to be distinct topographic highs that bound the western and eastern margins of the central plateau. The western high (referred to as the western Cordillera) is about 700 m higher than the Altiplano (at length scales greater than 40 km, Figs 1b, c). The eastern high is less than 500 m higher than the Altiplano at length scales greater than 40 km, though individual peaks reach over 6000 m (Figs 1b and c). This forms the high spine of the Eastern Cordillera. The eastern flank of the Eastern Cordillera drops toward the foothills of the Andes. Here there is a marked inflection in the topographic profile, where the topography begins to flatten out (Figs. 1b, c). The eastern margin of the Andes is at altitudes of 200 to 400 m.

This study focuses on the tectonic and magmatic evolution, and paleo-elevation history, of two transects, defined by the topographic profiles in Fig. 1a. These transects, referred to as the northern and southern profiles respectively, encapsulate the main geological features of this part of the Andes, immediately north and south of the Bolivian bend (Fig. 1). Figure 2 shows lithospheric-scale cross-sections along the northern and southern profiles.

2. Lithospheric Profiles

The lithospheric structure of the Central Andes is constrained by seismic receiver function analyses, combined with seismic tomography and gravity and flexural modeling (Fig. 2, Beck et al. 1996, 2002, Lyon-Caens and Molnar 1985, Watts et al. 1995, Stewart and Watts 1997, Myers et al. 1997, Yuan et al. 2000, 2002, Heit et al. 2007, Perez-Gussinye et al. 2009, Tassara and Echaurren 2012). These show that the high Andes form a broad region, ~500 km wide, where the thick crust is up to 75 km thick, but the lithosphere is less than 100 km thick. The mantle signature of helium emissions in geothermal systems in the high Andes provides an additional constraint

on the lithospheric structure here, because it defines the present day extent of active mantle melting, which is inferred to correspond to thin lithosphere <100 km thick (Fig. 6, Hoke and Lamb 2007). The Andean lithosphere is bounded in the west by the subducted slab of the Nazca plate, and in the east by the thick lithosphere (> 150 km), but normal crust (~35 km thick) of the Brazilian Shield (Fig. 2).

The topographic and structural trend in the northern profile is approximately NW – SE, and the high Andes are slightly narrower (~450 km width for elevations >2 km) than in the southern profile (~550 km width), where the trend is nearly N-S. The structural and stratigraphic evolution of the profiles is summarized in Figs. 2 - 5.

2.1 Forearc and Western Cordillera (Volcanic Arc)

The onshore part of the forearc rises abruptly from the coast to elevations 1 - 2 km. Farther inland, a relatively smooth monoclinical ramp of Miocene (10 – 20 Ma) ignimbrites and alluvial sequences slope up from elevations of ~1 km to 4 – 4.5 km over a distance of ~50 km, with minor faulting that accommodates shortening <5 km (Figs. 1, 2a, Isacks 1988, Victor et al. 2004, Lamb et al. 1997, Jordan et al. 2010). These overly with marked angular unconformity Mesozoic and Paleozoic sequences that have undergone significant Eocene shortening (Fig. 3a, 30 – 40 Ma, Armijo et al. 2015). The active volcanic arc (Fig. 6, Western Cordillera) is about ~50 km eastward of the western edge of the high plateau, at a base level of 4 - 4.5 km - individual arc volcanoes rise up to 2 km above this base level. There has been a ~50 km eastward migration of the arc since the early Cenozoic (Campusano 1990).

2.2 Altiplano

The Altiplano formed a broad sedimentary basin >200 km wide (Altiplano basin) during the Cenozoic, bounded by reverse faults on both margins, with up to 8 km of terrestrial red-bed accumulation in sub-basins (Figs. 2, 3, for example, Corque-Corocoro basin in northern profile, and Tambo Tambillo basin in southern profile). Paleocurrents show the Altiplano fluvial sequences were derived from both the Western and Eastern Cordilleras (Lamb et al. 1997, Horton and Hampton 2001, Leier 2010). Today, much of the region is covered by large salt pans, such as the Salar de

Uyuni and Salar de Coipasa, or shallow lakes (Lago Poopo).

In the central part of the Altiplano basin, marine latest Cretaceous (Maastrichtian) limestones and shales pass up conformably into Eocene to Late Oligocene fluvial sequences, with rare interbedded volcanic tuffs (Marshall et al. 1991, Gayet et al. 1991, DeCelles and Horton 2003). Sequences become markedly conglomeratic at ~25 Ma (Azurita Formation, Lamb and Hoke 1997), synchronous with the emplacement of mafic lavas and volcanic conglomerates (Kennan et al. 1995, Lamb and Hoke 1997, Hoke and Lamb 2007). On the margins of the basin, imaged in oil company seismic lines, these rest with marked angular unconformity on Mesozoic and older sequences, including Precambrian basement near the volcanic arc (Fig. 4, Troeng et al. 1994, Lamb and Hoke 1997, Lamb 2011). Large cobbles of this basement are present in the ~25 Ma Azurita Formation, outcropping in the centre of the Altiplano basin between ~18° and 20°S.

The presence of frequent interbedded volcanic tuffs, together with local angular unconformities, allow tight constraints on the timing of Neogene deformation in the Altiplano, with a total of 30 ± 10 km of shortening along the northern profile, and 75 ± 25 km of shortening along the southern profile (Figs. 3-5, Table 1, Bolivian Geological Survey 1978, Lavenue and Mercier 1991, Herail et al. 1993, McFadden et al. 1995, Kennan et al. 1995, Lamb and Hoke 1997, Elger et al. 2005, Lamb 2011). Altiplano tectonic shortening was markedly diachronous (Figs. 3-5), occurring ~7 Ma earlier along the southern profile (mainly between 11 and 16 Ma), compared to the northern profile (mainly between 5.4 and 9.5 Ma).

2.3 Eastern Cordillera

The Eastern Cordillera forms a rugged region east of the Altiplano that is mainly underlain by lower Paleozoic (Cambrian to Ordovician) marine sandstone and shale, up to 10 km thick, overlain unconformably by Cretaceous and Cenozoic marine and terrestrial deposits, up to 4 km thick sequences (Bolivian Geological Survey 1978, Marshall et al. 1991, Gayet et al. 1991, Rouchy et al. 1993). These have been deformed in a Paleogene fold and thrust belt since ~40 Ma, with a westward vergence in the west, on the margins of the Altiplano, and eastward vergence in the east,

towards the foreland with a total Paleogene shortening of 65 – 100 km, decreasing away from the axis of the Bolivian oroclinal (Fig. 2, Table 1, Bolivian Geological Survey 1978, Key 1996, 1999, Kley and Monaldi 1998, Lamb et al. 1997, Muller et al. 2002, McQuarrie 2002, Eichelberger et al. 2013). Along the northern profile, the tightly folded and thrust Oligocene Luribay Conglomerate, together with underlying Maastrichtian and older sequences, are overlain with angular unconformity by the 28 – 22 Ma Lower Salla Formation (Fig. 3a) and intruded by the ~26 Ma Illimani granitoids (Fig. 3a).

Since ~27 Ma, fluvial and lacustrine sequences (with well dated volcanic tuff layers) accumulated in local compressional basins, resting with angular unconformity on deformed Late Cretaceous and Paleogene sequences (Fig. 3b, Bolivian Geological Survey 1978, Lamb et al. 1997, Horton 2005). Folding and reverse faulting of these sequences indicate that the Eastern Cordillera has undergone more-or-less continual contractional deformation during the early to mid Miocene, accommodating 30 ± 10 km of shortening, possibly with a component of dextral strike-slip, although the precise timing of shortening varies from basin to basin (Fig. 3b, Kennan et al. 1995, Lamb et al. 1997, Herail et al. 1996, Muller et al. 2002, Horton 2005, Eichelberger et al. 2013).

An essentially undeformed, although uplifted, Late Cenozoic paleodrainage system, that fed via a small number of outlets into the foreland basin farther east (Gubbels et al. 1993, Kennan et al. 1997, Barke and Lamb 2006) are preserved throughout the Eastern Cordillera at elevations ~3 to 3.7 km as regional peneplain remnants with thin sedimentary cover (Fig. 4, San Juan del Oro surfaces). Their existence shows that internal folding and thrusting within the Eastern Cordillera had ceased ~12 Ma, although conjugate strike-slip continued and is active today (Lamb 2000, Funning et al. 2005).

2.4 Interandean and Subandean zones

A two-fold morpho-tectonic division is generally recognized in the Andes east of the Eastern Cordillera, comprising the Interandean zone and Subandean zone (see Kley 1996), generally at elevations <2 km. The Subandean zone comprises the eastern

foothills of the Andes, generally 100 – 150 km wide at elevations <1.5 km, forming a NE- to E-verging fold and thrust belt that deforms late Oligocene to recent fluvial sequences. The Interandean zone lies in the transition between the mainly lower Paleozoic outcrops in the Eastern Cordillera and deformed Cenozoic sequences of the Subandean zone, coinciding with a marked reduction in average elevation in southern Bolivia. It forms a zone 50 – 100 km wide where predominantly upper Paleozoic sequences (Devonian - Carboniferous) outcrop (Bolivian Geological Survey, 1978). The structural style in the Interandean zone is similar to that in the Subandean zone, and together they were the main locus of shortening on the eastern margin of the Andes since ~25 Ma, with deformation focused in the Subandean zone since ~10 Ma, accommodating underthrusting of the Brazilian Shield (Fig. 1b, Kley 1996, Lamb et al. 1997, Eichelberger et al. 2013).

Total shortening in the Interandean and Subandean zone since ~25 Ma is estimated to be 120 ± 25 km, with 85 ± 15 km since ~10 Ma (Fig. 2a, 3a, Table 1, Roeder, 1988; Herail et al., 1990; Baby et al., 1992, 1993, 1995; Wigger et al., 1993; Dunn et al., 1995, Kley and Monaldi 1998, Kley 1996, 1999, McQuarrie 2002, Uba et al. 2006, Barke and Lamb 2006, Eichelberger et al. 2013). The Brazilian Shield is overlain to the NE by a Cenozoic foreland basin up to 5 km thick, which thins towards the outcrop of the Brazilian shield (data held by oil companies, 1995). The latter forms the stable nucleus of this part of South America since the Proterozoic (Litherland et al., 1986).

2.5 Magmatism

Figure 6 shows the timing and distribution of magmatism in the Central Andes between 16°S and 22°S. Volcanic activity between ~26 and ~37 Ma is rare in this part of the Andes, with the absence of a volcanic arc, interpreted to be the result of flat slab subduction in this period (James and Sacks et al. 1999, Haschke et al. 2006, Hoke and Lamb 2007). Since ~26 Ma, andesitic - felsic volcanic activity has been widespread across the Central Andes, extending up to 300 km behind the arc (Fig. 6). Early - mid Miocene (14 – 26 Ma) Sn and Ag bearing subvolcanic stocks and granitoid intrusions (Fig. 6, Haschke et al. 2006, Oncken et al. 2006, Hoke and Lamb 2007), together with major Late Miocene ignimbrite fields (Figs. 4, 6, Los Frailes

ignimbrites, 2 – 12 Ma, and Morocoalla ignimbrites 6 – 9 Ma) are preserved on the western margin of the Eastern Cordillera (Schneider 1985, Baker and Francis 1978, Barke et al. 2007).

There has been mafic volcanism in the Altiplano, with the best documented periods 20 – 25 Ma, 11 – 13 Ma, and 0 – 5 Ma (Redwood and MacIntyre 1989, Davidson et al. 1992, Hoke et al. 1993, Lamb and Hoke 1997, Elger et al. 2005, Hoke and Lamb 2007). Today, active mantle melting is occurring throughout the high Andes, resulting in degassing of mantle helium in geothermal systems and active volcanoes (Fig. 6, Hoke et al. 1994, Hoke and Lamb 2007). The early mafic volcanism (20 – 25 Ma) coincides with the onset of widespread volcanism in this part of the Andes – presumably caused by slab steepening and the re-establishment of an asthenospheric corner flow beneath the high Andes - and is the most voluminous. It comprises lava flows, sills and shallow intrusive bodies that mainly outcrop on the northeastern and southeastern margins of the Salar de Uyuni (Fig. 4, 6, Tambo Tambillo, Rondal and Juliaca Formations), although sills and lavas are found in the Corque region (Azurita Formation). These are tightly folded, but originally represent ‘flood basalt’ type eruptions in a region up to 200 km long and ~100 km wide (Fig. 5a).

3. Paleo-altimetry

A change in elevation of a rock outcrop is described by rock uplift U_r . However, the average change in elevation of the Earth’s surface, referred to as surface uplift U_s , also depends on the average amount of vertical erosion E :

$$U_s = U_r - E \quad (1)$$

Surface uplift is the response to changes in the isostatic equilibrium with the underlying mantle, whereas paleo-altimetry techniques, which apply to specific rock units, document rock uplift U_r . The following is a summary of the three main techniques used in the Central Andes for estimating paleo-elevation. The locations, both in space and time, of published paleoaltimetry studies are shown in Figures 2 to 4.

3.1 Paleofloras

Fossil flora, preserved in volcanic or sedimentary deposits, can be used to estimate climatic conditions at the time of growth. There are broad correlations between the morphology of these flora and elevation, determined by studying living floras at different elevations and comparable climatic zones (Wolf 1995, Stranks and England 1994, Gregory 2000a, 2000b, Spicer et al. 2000). Estimates of paleo-elevation made this way have large uncertainties, generally > 1 km, but can be used to discriminate between high (~ 4 km) and low (< 2 km) elevations. In Bolivia, there are four sites with well-studied fossil floras, currently at elevations > 3.7 km and ages between 6 and 20 Ma, all of which suggest elevations > 2 km lower when they grew (Berry 1923, Gregory 1998, Gregory 2000a, 2000b, Graham 2001).

3.2 Paleodrainage systems

Well preserved paleodrainage systems are indicators of regional topographic gradients at the time they were active. By comparing these gradients with present day drainage gradients, it is possible to estimate subsequent elevation changes.

Barke and Lamb (2006) quantified drainage gradients in the Late Miocene San Juan del Oro peneplain surfaces (Fig. 4), preserved at elevations of 3 – 3.7 km in the Eastern Cordillera between 17°S and 22°S . By a combination of projecting these downstream to the foreland, projecting present day gradients upstream, and quantifying knickpoints in the drainage system, they showed that there had been a regional change in elevation (U_r) since ~ 8 Ma of 1.7 ± 0.7 km. Taking account of ~ 0.3 km of average erosion E in the same time period, this indicates 1.5 ± 0.7 km of surface uplift U_s (see equation 1).

Jordan et al. (2010) compared the present-day slopes along the western margin of the Andes, in northern Chile between 17°S and 23°S , with estimates of drainage slopes in alluvial fan systems, to determine Neogene paleo-relief of the Western Cordillera relative to the forearc. The cosmogenic flux measured in boulders preserved on ~ 13 Ma and older erosional surfaces indicates that the forearc has not changed significantly in elevation (< 500 m), certainly since ~ 13 Ma and most likely since ~ 25

Ma (Dunai et al. 2005, Evenstar 2014 and pers. com.). So the base level used in the geomorphological analysis of Jordan et al. (2010) can be treated as essentially fixed since ~25 Ma, and their relief development can be interpreted in terms of uplift relative to sea level. Thus, 2 – 3 km of the uplift of the western margin of the high Andes is likely to have occurred since ~25 Ma, with ~1 km in the last 10 Ma.

3.3 Geochemistry of paleosols

The ratios of carbon and oxygen isotopes in paleosol carbonates are a measure of temperature and isotopic composition of meteoric water from which the carbonates precipitated. These systematically change with altitude, and also regional and/or global climate. So any calibration for altitude needs to take account of the effects of climate change, with the assumption that the paleosols have not gone any significant alteration since they formed (Sempere et al. 2007, Quade et al. 2007). In addition, isotopic studies of modern paleosols show that there are unexplained systematic errors up to 1.5 km in predicted altitude, and the isotopic method can be ‘blind’ to elevations <1.5 km (Hoke et al. 2009). This suggests that the true error in isotopic paleo-altimetry could be significantly greater than the formal errors, and most likely > 1 km.

Figure 7 shows a $\delta^{18}\text{O}$ and temperature plot for paleosols from the Bolivian Andes, using data from Garzzone et al. 2006, Ghosh et al. 2006, Garzzone et al. 2008, Leier et al. 2013, Garzzone et al. 2014. The temperature is determined from the carbon isotopic signature (clumped $\Delta^{13}\text{C}$) and reflects that of the soil pore waters from which the paleosol carbonates precipitated. The $\delta^{18}\text{O}$ (as per mil with respect to SMOW) is that for soil waters in equilibrium with the carbonate, given the soil water temperature and the oxygen isotopic ratio in the carbonate. As noted by Quade et al. (2007), the data cluster about a curve (defined by the heavy black line in Fig. 7) that follows the general anticipated trend for $\delta^{18}\text{O}$ and temperature with altitude. Altitude calibrations used by Quade et al. (2007), Ehlers and Poulsen 2009, and Garzzone et al. 2014 are shown in Figure 7 by the short solid or dashed marks along the main black curve (the orientation of the marks reflects the ‘trend’ of the calibration elevation on the plot).

The Quade et al. (2007) calibration is based on present day variations in temperature

and $\delta^{18}\text{O}$ with elevation for the northern Bolivian Andes. The Ehlers and Poulsen (2009) calibration – referred to as the northern calibration in Figure 7 – is based on regional climate models (RCMs) which consider the effect of the Andes themselves on trajectories and precipitation of air parcels, referred to as the orogenic effect (see Leier et al. 2013). Garzzone et al.'s (2014) calibration is based on a comparison with low elevation (~0.4 km) Late Miocene (6 – 12 Ma) paleosols in the eastern foothills (Subandes) of the Central Andes at ~23°S – referred to as the southern calibration in Figure 7. There are clearly strong latitudinal effects, and the southern calibration is most applicable closest to the calibration samples at ~23°S – it seems to progressively overpredict paleo-elevations northwards, where soil temperatures are likely to be cooler at any given elevation, with wetter conditions, as noted by Garzzone et al. (2014).

In this study, the southern calibration is used for paleosols south of 20°S (Figs. 4, 8c, Quebrada Honda, Cerdas), and the northern calibration for paleosols farther north (Callapa, Lower and Upper Salla). For samples in the vicinity of 20°S (Figs. 4, 7, 8b, Quehua, Tambo Tambillo), the best estimate of paleo-elevation is taken to be somewhere between the two calibrations.

4. Quantifying processes of surface and rock uplift

Mechanisms for rock and surface uplift of the Bolivian Andes are: erosion, sedimentation, crustal shortening and thickening, magmatic addition to the crust, and changes in lithospheric structure such as mantle detachment (Lamb and Hoke 1997). The amount of uplift is also determined by the isostatic compensation factor k . A comparison between observed crustal thickness and elevation in the Bolivian Andes (Beck et al. 1996, 2002, Yuan et al. 2000, 2002, Barke and Lamb 2006) indicates that k is in the range 5.5 – 7.5, with higher values in the Eastern Cordillera, and lower values in the Altiplano (Lamb 2011).

4.1 Uplift due to erosion, sedimentation, and magmatic addition

Erosion will act to reduce surface uplift, but if it is not a uniform planation, it can also

be a powerful mechanism of rock uplift. Thus, erosion E will reduce surface uplift by an amount E/k , but will generate rock uplift $E \cdot (k-1)/k$ of relict areas; rock uplift >1 km is difficult to generate by this process as it requires very deep erosion which still leaves relict land surfaces. Geochronological data (Benjamin et al. 1987, Barnes et al. 2009) combined with direct observations of the depth of erosion from dissection and unroofing of deeper parts of the stratigraphy (Barke and Lamb 2006, Jordan et al. 2010), indicate that Neogene erosion plays only a minor role in the Western Cordillera and Altiplano. Erosion has caused 5 – 10 km of total unroofing in the Eastern Cordillera, with roughly equal amounts between ~40 and 25 Ma, and 25 and 0 Ma.

Sedimentation has played a role in the uplift of the Altiplano. Here, an average thickness of 3 – 5 km of Cenozoic red-beds accumulated across most of the width of the Altiplano, with roughly equal amounts between ~40 and 25 Ma, and 25 and 0 Ma - surface uplift will be a factor $1/k$ of the sediment thickness. Magmatic addition is only likely to be significant on timescales of several 10s of millions of years (section 4.4, Lamb and Hoke 1997). In addition, the added material from the mantle will have a mafic composition (basalt or basaltic andesite), with a greater density ($2.9\text{--}3 \text{ g/cm}^3$) compared to average crust ($2.7\text{--}2.85 \text{ g/cm}^3$), and so the isostatic compensation factor $k \sim 8\text{--}11$, or 1 km surface uplift for 8 – 11 km of added crust.

4.2 Uplift due to crustal shortening and thickening

In the case of pure shear crustal shortening and thickening, in 2-D vertical plane strain, uplift U_s can be expressed:

$$U_s = \Delta C/k - E = S \cdot C_f/((W_f + S) \cdot k) - E = S \cdot C_i/(k \cdot W_f) - E \quad (2)$$

where S is the observed horizontal tectonic shortening, C_i , W_i and C_f , W_f are the initial and final crustal thickness and widths respectively ($S = W_i - W_f$), ΔC is the change in crustal thickness ($C_f - C_i$), and, as before, k is the isostatic compensation factor, and E is the erosion (Lamb 2011).

4.2.1 Underthrusting and lower crustal shortening

Heat flow in the high Andes (Western Cordillera, Altiplano and western margin of the Eastern Cordillera) has mean values in the range $75 - 100 \text{ mWm}^{-2}$, indicating surface geothermal gradients $\sim 35^\circ\text{C/km}$ (Springer and Forster 1998). Such a high surface geothermal gradient must inevitably result in a very weak mid to lower crust, at depths $>20 \text{ km}$, even leading to lower crustal convection (Babeyko et al. 2002), especially if it has a mainly felsic composition, as indicated by the relatively low P wave seismic velocities (Beck et al. 2002).

Weak lower crust will be capable of undergoing significant ductile strain that is decoupled from that in the overlying brittle crust (Fig. 2), promoted by underthrusting of 'strong' lithosphere on the eastern margin, with a high $T_e > 50 \text{ km}$ (Watts et al. 1995, Stewart and Watts 1997, Perez-Gussinye et al. 2008). Thus, surface shortening on the eastern margin of the Andes will result in uplift of the high Andes through underthrusting of the rigid foreland (Brazilian Shield) along an inclined detachment, or by ductile lower crustal shortening and thickening that is decoupled from the overlying brittle crust (Fig. 2, Lamb 2011). The latter may be described by a simple 'piston' model, in which the lower crust is squeezed and thickened by the rigid 'piston' of the underthrust lithosphere (Fig. 2, Isacks 1988, Gubbels et al. 1993, Allmendinger and Gubbels 1996, Lamb 2000, Hindle et al. 2005, Barke and Lamb 2006, Sobolev et al. 2006, Lamb 2011). In this case, for a piston of thickness C_p , underthrust a distance S , lower crustal shortening and thickening across a final width W_f will result in surface uplift U_s :

$$U_s = \Delta C/k = C_p \cdot S/(W_f \cdot k) \quad (3)$$

where, as in equation (2), ΔC is the change in crustal thickness, k is a measure of the degree of isostatic compensation and deformation is 2-D vertical plane strain.

Where there is also brittle crustal shortening – for example, in the Altiplano and Eastern Cordillera during the Miocene - the total uplift is calculated by adding together equations (2) and (3); here equation (2) describes the contribution to uplift due to crustal thickening and erosion, and the shortening S in equation (3) includes all horizontal displacements in the crust. In addition, equation (3) may be modified if ductile crustal strains are not uniform along the length of the cross-section (see

equation (4) below).

4.2.2 Non-uniform distribution of lower crustal shortening

The distribution of lower crustal ductile strain is unlikely to be uniform, but will depend on the rheology and physical conditions (e.g. the effects of lateral variations in temperature, pressure, stresses and composition) in the ductile zone. The sensitivity of strain rate ($\dot{\epsilon}$) to these can be easily demonstrated with a typical flow law such as the experimental Dorn Equation (Ranalli 1995):

$$\dot{\epsilon} = A \tau^n \exp[-(E+pV)/RT] \quad (4)$$

where A is a constant, τ is the deviatoric stress, n is the power law exponent, E is the activation energy of flow, p is the pressure, V is the molar volume, R is the universal gas constant, and T is the temperature in Kelvin. For example, with a wet granite rheology (for ~40 km depth and 630°C reference temperature, $E = 137$ kJ/mol, $n = 1.9$, $V = 10^{-6}$ m³ mol⁻¹) indicated by the low crustal P velocities in the thick Altiplano crust (Beck and Zandt 2002), and the behind arc subduction setting (Ranalli 1995, Sobolev et al. 2006, Babeyko et al. 2002), increasing the mean temperature by ~50°C would increase the strain rate by a factor of ~3 for the same deviatoric stresses, whereas a change from a wet to dry felsic rheology, or to an intermediate composition, would reduce the strain rate by at least an order of magnitude (see also discussion in Lamb 2011).

Lamb (2011) introduced an empirical constant γ , referred to as the focusing factor, which takes account of horizontal variations in ductile strain in the plane of the cross-section, when it is partitioned between several regions:

$$\Delta C(l) = \gamma \Delta C \quad (5)$$

where $\Delta C(l)$ is the local change in crustal thickness as a consequence of ductile strain, and ΔC is the average change, defined by equation (3). In order that ductile shortening balances in 2-D – i.e. to preserve cross-sectional area – γ in one region places constraints on γ in adjacent regions. For example, if ductile shortening is partitioned

in 2-D into regions 1 and 2, with a focusing factor γ for region 1, then for region 2, $\gamma_2 = (1-\gamma\alpha)/(1-\alpha)$, where α is the proportion of the total final region underlain by ductile strain that is within region 1. Note that if $\gamma > 1/\alpha$, then region 2 will undergo crustal thinning and subsidence. The range of models is increased if flow of ductile lower crust out of the plane of the cross-section is also considered.

As discussed above, quite small lateral variations in composition, temperature and topographic contrasts can give rise locally to a large change in strain rate for the same driving deviatoric stresses, resulting in a local focusing factor γ of two or more (note that elsewhere in the plane of the cross-section $\gamma < 1$ to preserve cross-sectional area), although, in fact, values fairly close to 1 ($0.5 < \gamma \leq 1.6$) are sufficient to fit the uplift data in the Central Andes (Tables 1 and 2, and see analysis in Lamb 2011), and much more restricted than the extreme permitted values ($0 \leq \gamma < 3$) if all lower crustal shortening is focused within any particular region (assuming there is no crustal thinning).

A full analysis requires computer models of lithospheric deformation (see Sobolev et al. (2006) for a numerical simulation of the Cenozoic evolution of the Central Andes), where temperature, viscosity and stresses are calculated through time from specified boundary conditions, rheologies, rock properties and physical processes. This way, the plausibility of the values of γ required to fit the uplift data can be directly tested.

4.3 Uplift due to lower lithospheric detachment

The mantle lithosphere has a higher density than the underlying asthenosphere, with a typical density contrast in the range $0.04 - 0.06 \text{ g/cm}^3$ (Stern et al. 2006). For this reason, it is gravitationally unstable, and so, when weak enough, it will detach itself and sink into the underlying asthenosphere, essentially removing a vertical load from the overlying lithosphere, which will bob up in response. Simple isostasy requires a detachment of a 50 – 100 km thick layer of mantle lithosphere in order to generate an uplift of ~1 km. In addition, if the lower crust is mafic, then for crustal thicknesses $> 35 \text{ km}$ it is assumed to transform to eclogite (Babeyko et al. 2006), with a density $\sim 3.5 \text{ g/cm}^3$. This will also be gravitationally unstable and could detach along with the

underlying lithosphere. In this case, detachment of ~17 km thick layer of pure mafic eclogite would also result in uplift of ~1 km.

In the usual geological model for mantle detachment, uplift is preceded by a slow phase of crustal and lithospheric thickening, over tens of millions of years (Houseman et al. 1981, Molnar et al. 1993, Platt and England 1993, Garzione et al. 2006). This is followed by rapid detachment of the gravitationally unstable lithospheric root, although recent dynamical models suggest that this is only likely to occur if there is an abrupt lateral change in lithospheric thickness (Stern et al. 2013). If detachment does occur, the overlying lithosphere could undergo uplift of several kilometres, and this could be on a timescale of only a few million years or less (Molnar and Garzione 2007, Stern et al. 2013).

Uplift may be enough to raise the gravitational potential energy of the lithosphere to the state where it is in deviatoric extension, and so it could be accompanied by a phase normal faulting. Also, if hot asthenosphere flows into the region left behind by the detached lithospheric root, then this would result in mantle melting through heating and adiabatic decompression, triggering a brief phase of mafic volcanism (Kay et al. 1994). Thus, an association of uplift, mafic volcanism and extension that all immediately *post-dates* lithospheric shortening is taken as diagnostic of this mechanism of uplift (Platt and England 1993).

4.4 Changes in geometry of the subducted slab

A change in the geometry of the underlying subducted slab can also generate elevation changes in the overlying lithosphere. Thus, in flat-slab or more general amagmatic subduction, the subducted plate is coupled to the overlying plate by a cold mantle wedge, and so the negative buoyancy of the subducted plate can act as a load which will pull the overlying lithosphere down 0.5 – 1.5 km (Zhong et al. 1992, Stern and Holt 1992, Kudo and Yamaoko 2003, Davila et al. 2010, Manea et al. 2012). This is in contrast to subduction above a well developed volcanic arc, where the mantle wedge consists of ‘hot’ and low viscosity asthenosphere.

When the load from the subducted slab is removed, the overlying plate must bob up.

The simplest way that this can happen is by the slab sinking, creating space for an inflow of hot and low viscosity asthenospheric mantle into the overlying mantle wedge. Hydration and adiabatic decompression could result in prolific mantle melting, with the rejuvenation of the volcanic arc and widespread behind arc mafic – felsic magmatism (James and Sacks 1999, Haschke et al. 2006, Hoke and Lamb 2007).

5. Evolution of the Andean Lithosphere

Figure 8 shows the topographic evolution of the Altiplano and Western and Eastern Cordilleras along the northern ($\sim 17.5^\circ\text{S}$) and southern ($\sim 20^\circ\text{S}$) profiles (Fig. 1a, 2), and also slightly farther south at $\sim 21^\circ\text{S}$, based on the paleoaltimetry techniques discussed in section 3. Also shown are the major phases of shortening, as described in sections 2.1 and 2.2, the slab steepening and delamination event, inferred from the widespread and voluminous volcanism between 20 and 25 Ma (see sections 2.5 and 5.2.2). The different methods of paleo-altimetry are clearly consistent with each other, and changes in elevation through time since ~ 25 Ma correlate with phases of shortening and uplift.

The initiation of Cenozoic deformation and uplift is given a nominal age of ~ 40 Ma, when regional elevations are assumed to be close to sea level, although both the exact timing of initiation and uplift, as well initial elevation, may vary slightly from region to region. Calculated histories (within their 2σ uncertainties) of surface and rock uplift along the northern and southern profiles for various time periods are also shown in Figure 8 and Tables 1 and 2, based on section 4 and the simple 2-D models described by equations (1) – (4), together with published estimates of the timing and amount of crustal shortening, sedimentation, erosional unroofing and magmatic addition (Benjamin et al. 1987, Lamb and Hoke 1997, Muller et al. 2002, Elger et al. 2005, Barke and Lamb 2006, Eichelberger et al. 2013, Lamb 2015).

Errors for uplift due to crustal shortening and thickening are from Monte Carlo simulations, where the statistics of multiple realisations of the data are calculated, given the range in input values (Table 1). Errors for calculated paleo-elevations are

determined using a similar method. The values of the focusing factor γ are chosen to provide a good fit with paleo-elevation data, in the sense that the calculated uplift trajectories lie within the formal errors of the paleo-elevation observations. As noted in section 4.2.2, the required values of γ ($0.5 \leq \gamma \leq 1.6$) are within a significantly more restricted range than the extreme permitted values ($0 \leq \gamma < 3$) that would focus all lower crustal shortening in one particular region. In other words, it is not necessary to ‘crank up’ the effects of shortening with end-member crustal thickening scenarios in order to fit the paleo-elevation data within error.

Three-dimensional effects, such as the lateral flow of ductile lower crust along the length of the Andes, are not required to explain the paleo-elevation data. However, some form of large-scale change in lithospheric geometry is the most plausible explanation for the evidence of ~1 km subsidence and abrupt uplift prior to 25 Ma in the Altiplano and western margin of the Eastern Cordillera (section 5.2, Table 2, Fig. 8).

5.1 Uplift of the Altiplano and Eastern Cordillera since ~25 Ma

Since ~25 Ma, the timing and amount of crustal shortening and thickening beneath the Altiplano and Eastern Cordillera, with a smaller contribution from erosion and sedimentation, can easily explain the paleo-elevation data, within error. Accelerated episodes of uplift in the Altiplano correspond to intense phases of crustal shortening (Fig. 8a and b).

Crustal shortening in the brittle crust is generally decoupled from that in the lower crust, with lower crustal ductile shortening right across the high Andes, driven by underthrusting of the Brazilian Shield since ~25 Ma, providing a long wavelength mechanism of crustal thickening and uplift (Table 1, Fig. 2). Note that the focusing factor $\gamma > 0$ everywhere (see section 4.2.2), so that no areas are undergoing lower crustal thinning and extension. Thus, the paleo-elevation data is well matched along the southern profile with lower crustal flow being more focused beneath the Altiplano and Western Cordillera between ~20 and 10 Ma ($\gamma \sim 1.3$, Fig. 8b, Table 1) compared to beneath the Eastern Cordillera ($\gamma \sim 0.7$), when Altiplano elevations here were

initially < 1 km, and there was intense brittle crustal shortening in the Altiplano. Since ~10 Ma, the paleo-elevation data indicate that lower crustal flow was more focused beneath the Eastern Cordillera ($\gamma \sim 1.5$, Fig. 8b, Table 1) compared to beneath the Altiplano and Western Cordillera ($\gamma = 0.5 - 1$, Table 1). The situation for the northern profile is similar, except that brittle crustal shortening in the Altiplano was ~7 Ma younger, occurring between ~10 and ~5 Ma (Table 1, Fig. 8a).

5.2 Paleogene uplift of the Altiplano and Eastern Cordillera

The Eastern Cordillera and Altiplano were at or below sea-level in the latest Cretaceous, when marine limestones were deposited (Fig. 3a and b, Rouchy et al. 1993, Lamb et al. 1997), and so the crustal thickness was most likely similar to that in the foreland, and ~35 km thick. Thus, Paleogene thick-skinned shortening in Eastern Cordillera, would give rise to elevations of ~1.6 km at ~25 Ma (Tables 1 and 2). Erosion during or after this shortening has stripped off a thickness comparable to the Paleogene sedimentary sequences deposited in the Altiplano region (~ 3 km), reducing elevations by ~0.5 km to ~1.1 km (~3/k for $k \sim 6.5$, Table 2). Likewise, sedimentation and shortening in the Altiplano region should cause 0.5 – 1 km of surface uplift. All these estimates are about ~ 1 km higher than those indicated by paleo-altimetry studies for the Altiplano and western margin of the Eastern Cordillera (Fig. 8, Tables 1 and 2), indicating some process that is reducing elevations in these regions by this amount.

5.2.1 Subsidence during flat slab subduction

The virtual absence of both arc and behind-arc magmatism in the period ~27 to 37 Ma suggests that subduction at this time took place at a low-angle, with a ‘flat slab’ geometry, so that there was no asthenospheric corner flow beneath the Andes – this is similar to the situation today at 28°S – 33°S, where there has been no arc and behind arc volcanism since ~15 Ma (Kay and Abruzzi 1996). As discussed in section 4.4, flat-slab subduction could provide an additional vertical load which would be enough to pull down the lithosphere beneath the Altiplano and Eastern Cordillera about 1 km in the Paleogene, as predicted by dynamical models of flat-slab subduction beneath relatively weak lithosphere (Fig. 10, Davila et al. 2010).

5.2.2 Rapid uplift with the onset of steeper subduction

The abrupt onset of volcanism in the Bolivian Andes at ~26 Ma, extending up to 300 km behind the arc, suggests that the asthenospheric corner flow was re-established at this time, with rapid steepening of the subducted slab (James and Sacks 1999, Haschke et al. 2006, Hoke and Lamb 2007). This would have decoupled the overlying lithosphere from the effects of the negative buoyancy of the subducted slab, resulting in ~1 km of rapid uplift (Figs. 8, 9). The paleo-elevation estimates along the northern profile for the Upper and Lower Salla Formation, between 28 Ma and ~17 Ma, appear to have captured this uplift (Fig. 8a), whereas the estimate for Tambo Tambillo Formation, at ~25 Ma, along the southern profile, may be the lower 'flat slab' elevation (Fig. 8b). Subsequent uplift along the northern profile on the eastern margin of the Altiplano, between 25 Ma and 11 Ma, can be related to crustal thickening and > 4 km reverse motion on the Eucalyptus Fault, which decoupled the uplift history here from the interior of the northern Altiplano, in the Corque-Corocoro basin (Fig. 8a).

Seismic, gravity and geochemical evidence point to a thin lithosphere, 90 – 120 km thick, beneath the Altiplano and western margin of the Eastern Cordillera (Fig. 2, Beck et al. 1996, 2002, Yuan et al. 2000, 2002, Myers et al. 1997, Hoke and Lamb 2007, Heit et al. 2009, Tassara and Echurran 2012). This is also likely to be the case in the early Miocene, because REE element modeling of the ~22 – 25 Ma mafic volcanics in the Altiplano indicate that they are the result of mantle melting at depths between 40 and 90 km, most plausibly in the asthenosphere (Hoke and Lamb 2007). Thus, detachment of lithospheric mantle beneath the Altiplano probably occurred during initial slab steepening, between 27 – 26 Ma, or during flat slab subduction between 37 and 27 Ma. In either case, detachment is likely to have been caused by the change in geometry of the subducted slab, generating the stresses that strip away the lower part of the lithosphere.

It has been suggested that mantle detachment also resulted in the loss of an eclogitic lower crust, because P-wave velocities (<6.2 km/s) indicate a felsic crust beneath the high Andes (Beck et al. 2002). However, the felsic composition may be a much older feature of this part of the Andes, because even on the extreme eastern margin of the Eastern Cordillera, which has had no history of Cenozoic magmatism, but where the

crust is ~60 km thick and underthrust by the Brazilian Shield, the lower crust has P wave velocities $<6.2 \text{ km s}^{-1}$ and cannot be eclogite (Christensen and Mooney 1995, Beck et al. 2002).

5.3 Uplift of the Western Cordillera

The long-term evolution of topography of the Western Cordillera (volcanic arc) is most easily explained in terms of the build-up of volcanic edifices, superimposed on the effects of lower crustal shortening and thickening. The latter has pushed up or inflated the overlying brittle crust of the high Andes, driven by underthrusting farther east (Fig. 8b, Table 1). Thus, the monoclinical ramp or shoulder in the forearc, which is only broken by faults with a cumulative displacement $< 2 \text{ km}$ (Victor et al. 2004), marks the western edge of this ductile shortening in the lower crust (Fig. 2).

Presumably this is ultimately controlled by the thermal structure of the crust, and in particular, the geometry of the $\sim 350^\circ\text{C}$ isotherm, defining the boundaries of lower crust that is warm enough for ductile flow (Fig. 2, Springer and Forster 1998, Lamb and Hoke 1997, Lamb 2011). A similar shoulder, marking the eastern limit of the high Andes (Whipple and Gasparini 2014), may mark the eastern edge of the region with warm ductile lower crust (Fig. 2).

The crustal thickening calculations indicate that plausible lower crustal ductile shortening beneath the Western Cordillera will result in uplift that is still $\sim 1 \text{ km}$ less than required by the paleo-elevation data (Tables 1 and 2). The obvious source of additional uplift is magmatic addition to the crust beneath the arc, which has had more-or-less continual volcanism since $\sim 25 \text{ Ma}$ (Fig. 8b, Table 2, Francis and Hawkesworth 1994). Lamb and Hoke (1997) used the relation between topography and the thickness of the crustal root, determined from receiver function data (Beck et al. 1996), to infer that $\sim 40\%$ of the crustal root beneath the volcanic arc ($\sim 15 \text{ km}$ crustal thickness) is higher density mafic crust with a density of $3.0 - 3.1 \text{ g/cm}^3$. If this has accumulated since $\sim 25 \text{ Ma}$, when the arc was in its present location (after migrating $\sim 50 \text{ km}$ eastwards since the Eocene, Campusano 1990) then it indicates magmatic addition beneath the arc at $\sim 0.6 \text{ km/Ma}$, contributing a total surface uplift of $1 \pm 0.5 \text{ km}$ for a root density of $3.0 - 3.1 \text{ g/cm}^3$ ($15/k \text{ km}$ for $k \sim 13$). These simple calculations are within a factor of two of Francis and Hawkesworth's (1994)

conservative estimate of magmatic addition beneath the arc, based on the inferred volume of intrusive volcanic rock.

Intense crustal shortening in the Western Cordillera occurred prior to 25 Ma, with folding and faulting of Paleogene and older basement sequences. In addition, paleomagnetic data suggest significant rotation of the forearc at this time, most plausibly accommodated by along strike gradients of shortening in Western Cordillera. Estimates of crustal shortening at this time are poorly constrained, but must be at least 10 km across a width of 100 – 150 km. In this case, equation (2) predicts >0.5 km of surface uplift prior to ~25 Ma (Tables 1 and 2, Fig. 9b).

6. Discussion

The geological and paleo-elevation data indicate that the Central Andes in Bolivia and northern Chile formed by uplift, and then merging, of two distinct mountain ranges, in the Western and Eastern Cordilleras. These were initially separated by a sedimentary basin up to 300 km wide and close to sea level, in what is the high Altiplano today (Lamb et al. 1997, Lamb and Hoke 1997, Armijo et al. 2015) – this also accounts for the structural style in the Eastern Cordillera, with west verging thrusts on the western margin and east verging thrusts on the eastern margin (Fig 2).

The two mountain ranges nucleated in the Paleogene (~40 Ma), during a period of ‘flat’ subduction, within zones of weaker lithosphere, along the ‘hot’ volcanic arc (Western Cordillera) and much farther east, in the centre of a Late Cretaceous basin (Eastern Cordillera, Lamb et al. 1997, Lamb and Hoke 1997). They each had their own distinct geological evolution until ~16 Ma, when the initiation of intense shortening within the intervening proto-Altiplano region eventually resulted in a single wide and high mountain range (Lamb et al. 1997, Lamb and Hoke 1997).

A crucial stage in this evolution was the development in the late Oligocene (28 to 25 Ma) of thin lithosphere beneath what is today the high Andes, when it is likely that the subducted slab steepened, allowing both mantle detachment and the influx of ‘hot’ asthenosphere into the overlying mantle wedge. This effectively created a thermal boundary condition at the Moho, raising temperatures to close to those in the

asthenosphere (Springer and Forster 1998). The subsequent behavior of the lithosphere can be understood in terms of its rheology and the thermal re-equilibration, on a time scale of 5 – 15 Ma (Turcotte and Schubert 1982), dictating the timing and mode of uplift, depending on the viscosity of the lower crust combined with the topographic evolution itself.

6.1 Weak ductile lower crust

It is clear from the analysis in this study that shortening in weak ductile lower crust has had a profound effect on the uplift of the Central Andes during the last 25 Ma. Uplift will tend to be comparable over large areas - leading to the formation of a central high plateau as the Western and Eastern Cordilleras merged - because a weak lower crust will also act as a sort of spirit level, smoothing out topographic contrasts (Fig. 2, Lamb and Hoke 1997). This mode of deformation is markedly different to thrust sheet models (for example McQuarrie 2002), in which very large rigid basement thrust sheets are postulated in the mid-lower crust beneath the Altiplano and Eastern Cordillera, transmitting thrust displacement right across the high Andes. It is hard to reconcile thrust sheet models with the plausible thermal and rheological conditions of the Andean crust (Babeyko et al. 2002, Sobolev et al. 2006).

Isostasy in an underlying weak ductile lower crust will also amplify the thickness of local sedimentary basins. This may explain why Altiplano basins are often quite narrow (~50 km wide), but with very thick sediment fill (> 5 km). Thus, the isostatic response to a sediment load depends on the density contrast between the sediment and underlying layer it is displacing. If the lower crust is weak, large and local vertical displacement of lower crust may occur, compensated by more regional lower crustal flow with a much smaller isostatic mantle response. In this case, local basin subsidence will be mainly controlled by the density contrast between the sediment load and underlying ductile lower crust, which will be much less ($\sim 0.2 - 0.3 \text{ g/cm}^3$) than that with the mantle ($\sim 0.7 \text{ g/cm}^3$). For fluvial sedimentation, sediment is replacing air, and so the amplification factor for basin depth above weak lower crust could easily be >10, compared to ~5 for mantle isostasy. An initial ~1 km deep accommodation space could be rapidly filled by >10 km of sediment.

6.2 Diachronous uplift of the Altiplano

An interesting question is why Miocene shortening and uplift in the Bolivian Altiplano were diachronous, occurring ~7 Ma later in the vicinity of the northern profile, compared to the southern profile. This may just reflect the thermal histories of these regions. For example, if rheology of the lithosphere beneath the Bolivian Altiplano is uniform, then the central Bolivian Altiplano lithosphere heated slightly faster than that beneath northern Altiplano, so that it was weak enough by ~16 Ma to yield to the ambient stresses and undergo rapid internal strain. The northern Altiplano did not reach these conditions until ~7 Ma later. These differences in thermal history could be due to differences in the thickness of mantle lithosphere that was retained since ~25 Ma in each region.

Alternatively, the thermal history of the two regions may have been similar, but the rheologies were not uniform – for example, the northern Altiplano may have a slightly more mafic rheology compared to the central Bolivian Altiplano, so that it maintained strength for longer (Lamb 2011). In any case, the focusing factors for ductile deformation ($0.5 < \gamma < 1.6$), required to explain the crustal thickening and uplift, indicate that the mechanism driving the intense Miocene shortening in the Altiplano is localized weakening of the lower crust.

6.3 Rate of Uplift

Advocates of the model of lithospheric detachment as a mechanism for post 25 Ma uplift in the Central Andes tend to favour both the upper bound estimates of uplift and the lower bound estimates of the time period in which the uplift occurred, determined from isotopic paleo-elevation proxies (i.e. >3 km in <<4 Ma, Garzzone et al., 2006, 2008, 2014). Thus, these authors argue that the rate of uplift is too high, when compared with observed strain rates, for crustal shortening and thickening to be the main mechanism of post 25 Ma uplift. Instead, they have proposed separate brief events of lithospheric mantle detachment as the cause of the two phases of uplift in the Bolivian Altiplano, between 16 and 10 Ma and 9.5 and 5.4 Ma. However, the upper bound estimates of uplift rate (e.g. Garzzone et al. 2014) are no more probable

than lower bound estimates, and the coincidence between the timing of uplift and tectonic shortening remains. Previous workers (Leier et al. 2013, Whipple and Gasparini 2014, Saylor and Norton 2014) have also argued that there is a lack of correlation between km-scale uplift and phases of crustal shortening at the same location, again indicating that mantle detachment is likely to be the main mechanism of uplift.

In this study I show that since ~25 Ma, the *mean* values, and certainly lower bound estimates, of observed uplift can easily be accounted for both local and regional (in the lower crust) horizontal finite strains in the same time period, and so detachment is not required given the available data. The strain rates are typical of wide active plate boundary zones, with 7 – 15 mm/yr across 150 – 200 km, or $\sim 5 \times 10^{-8}$ /yr, (Lamb and Vella 1987, Lamb 2000, 2015, Flesch et al. 2007). At this strain rate, the rate of surface uplift through crustal thickening, for a final crustal thickness of ~60 km, would be ~0.5 mm/yr or ~0.5 km/Myr. Lower crustal ductile strain rates can also give rise to a regional km-scale uplift in < 5 Myr. Thus, even if rapid uplift occurs in regions without surface crustal shortening, crustal thickening at depth can still drive it.

However, given the uncertainties in both our understanding of paleo-elevations and tectonic deformation, minor episodes of detachment of small amounts of mantle lithosphere (<25 km) since ~25 Ma, driving <0.5 km of surface uplift, cannot be ruled out for the Altiplano or Eastern Cordillera (e.g. Sobolev et al. 2006). Note that some form of lithospheric detachment may have occurred farther south, beneath the Argentinian Puna, in the Late Cenozoic (Kay et al. 1994).

7. Conclusions

In this study, the Cenozoic geological evolution of the Central Andes, along two transects between ~17.5°S and 21°S, is compared with paleo-topography, determined from published paleo-altimetry studies. Surface and rock uplift are quantified using simple 2-D models of crustal shortening and thickening, together with estimates of sedimentation, erosion and magmatic addition. The principle conclusions are as follows:

(1) Prior to ~25 Ma, during a phase of amagmatic flat-slab subduction, thick skinned crustal shortening and thickening was focused in the Eastern and Western Cordilleras, separated by a broad basin up to 300 km wide and close to sea level, which today comprises the high Altiplano. Surface topography in the Eastern Cordillera at ~25 Ma is estimated from crustal shortening and erosion to be ~1 km, whereas paleo-elevations on the western margin of the Eastern Cordillera are close to sea level. This discrepancy can be explained in terms of a local 'pull-down' effect of the subducted slab, coupled to the overlying lithosphere by a cold mantle wedge.

(2) Oligocene steepening of the subducted slab is indicated by the initiation of the volcanic arc at ~27 - 25 Ma, and widespread mafic volcanism in the Altiplano between 25 and 20 Ma. This may have resulted in detachment of mantle lithosphere and possibly dense lower crust, triggering 1 – 1.5 km of rapid uplift (over << 5 Myrs) of the Altiplano and western margin of the Eastern Cordillera and establishing the present day lithospheric structure beneath the high Andes.

(3) Since ~25 Ma, surface uplift has been the direct result of crustal shortening and thickening, locally modified by the effects of erosion, sedimentation and magmatic addition from the mantle. The rate of crustal shortening and thickening varies with location and time, with two episodes of rapid shortening in the Altiplano, lasting < 5 Myrs, that are superimposed on a long term history of ductile shortening in the lower crust, driven by underthrusting of the Brazilian Shield on the eastern margin.

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Table 1. Calculated and observed crustal thickening and resultant surface uplift (U_s) in 2-D shortening models (see section 4.2) for Central Andes between 18°S and 21°S, using value of focusing factor γ that gives a good fit to observed paleo-elevations (see Table 2 and Fig. 8).

Time (mode) Ma	Final width ¹ km	Shortening ² km	Focusing factor γ ³	Final (initial) thickness ⁴ km	Crustal Thickening ⁵ km	Surface uplift ⁶ km
<i>Western Cordillera</i>						
40 - 25	120	>10	-	(35)	>3.5	>0.5
25-10 (ductile)	520	30+30+75±30	1.4	32.5*	14 ± 4	1.6 ± 0.6
10-0 (ductile)	520	85 ± 15	0.5	32.5*	3 ± 1	0.4 ± 0.1
<i>Northern Profile</i>						
<i>Altiplano</i>						
40 - 25	180	10 ± 5	-	(35)	1.8 ± 0.9	0.3 ± 0.2
25 - 10 (ductile)	450	30+30 ± 10	0.5	32.5*	2 ± 0.7	0.3 ± 0.1
10-5 (brittle)	140	28 ± 7	-	20	3.3 ± 1	0.5 ± 0.2
10-5 (ductile)	420	43+28 ± 15	1.6	32.5*	9.1 ± 3.3	1.3 ± 0.3
10-5 (total)					12.4 ± 4	1.9 ± 0.4
5-0 (ductile)	420	43 ± 7	1	32.5*	3.3 ± 0.6	0.5 ± 0.1
<i>Eastern Cordillera</i>						
40 - 28	180	65 ± 10	-	32.5*	10.4 ± 2	1.6 ± 0.3
25-15 (brittle)	150	30 ± 10	1	20	2.6 ± 2	0.5 ± 0.1
25-15 (ductile)	450	30+ (0.7 x 30) ± 5	1.5	32.5*	6.5 ± 3	0.7 ± 0.2
25-15 (total)					8.5 ± 3	1.2 ± 0.4
15-10 (ductile)	450	0.3 x 30 ± 5	1.5	32.5*	2.6 ± 1	0.4 ± 0.2
10-5 (ductile)	420	43+28±15	0.6	32.5*	4 ± 1	0.5 ± 0.1
5-0 (ductile)	420	43 ± 7	1	32.5*	3.3 ± 0.6	0.5 ± 0.1
<i>Southern Profile</i>						
<i>Altiplano</i>						
40 - 25	255	20 ± 10	-	(35)	2.6 ± 1.3	0.4 ± 0.2
25 - 16 (brittle)	240	15 ± 8		15	1 ± 0.5	0.14 ± 0.1
25 - 16 (ductile)	520	0.6 x (30+30)+15±30	1.3	32.5	4 ± 2	0.6 ± 0.2
25 - 16 (total)						0.7 ± 0.2
16-10 (brittle)	180	60±25	-	15	4.0 ± 2	0.6 ± 0.3
16-10 (ductile)	520	0.4 x (30+30)+60±30	1.3	32.5*	7.2 ± 3	1.1 ± 0.4
16-10 (total)					11.1 ± 4	1.7 ± 0.4
10 - 0	520	85 ± 15	0.7	35*	3.3 ± 1	0.6 ± 0.1
<i>Eastern Cordillera</i>						
40 - 28	250	100 ± 20	-	32.5	10.4 ± 2	1.6 ± 0.3
25-10 (brittle)	220	30 ± 10	-	(20)	2 ± 1	0.3 ± 0.1
25-10 (ductile)	520	30+30+75±30	0.5	32.5*	4.6 ± 1	0.7 ± 0.2
25-10 (total)					6.5 ± 2	1.0 ± 0.3
10 - 0	520	85 ± 15	1.5	32.5*	9 ± 2	1.2 ± 0.4

Notes: (1) Width over which shortening has taken place - final width after deformation of overlying brittle crust shown in brackets; (2) published shortening estimates for time period (after Muller et al. 2002, Elger et al. 2005, Barke and Lamb 2006, McQuarrie 2002, Lamb 2011, Eichelberger 2013); (3) best-fit focusing factor γ for ductile strain (see section 4.2.2); (4) star denotes 'piston' thickness in lower crust, otherwise crustal thickness (section 4.2.1); (5) crustal thickening using simple 2-D crustal shortening models (equations 2 – 4 in section 4.2); (6) surface uplift from shortening and thickening with 2 σ uncertainties based on Monte Carlo simulation, using isostatic factor $k = 6.5 \pm 1$ (see section 4.2); values used in Table 2 shown in bold.

Table 2. Calculated (this study, see section 4) and observed uplift for Central Andes between 18°S and 21°S. Calculated paleo-elevations for representative localities shown in bold.

Time Ma	Deposition ¹	Calculated uplift during time period in km				Paleo- elevation km ⁵	Observed rock uplift in time period (ref) km ⁷
		Erosion ¹	Crustal shortening ²	Magmatic addition ³	Subduction/ detachment ⁴		
<i>Western Cordillera</i>							
~40						0.0	
40 - ~25			> 0.5	Inactive arc	?		
~25					~?	1.3 ± 0.7	
25 - 10			1.6 ± 0.6	~0.6		3.5 ± 0.2	2 ± 0.5 (8)
10 - 0			0.4 ± 0.1	~0.4		4.3	1.2 ± 0.7 (8)
<i>Northern Profile</i>							
<i>Altiplano</i>							
~40						0.0	
40 - ~25	0.3 ± 0.2		0.3 ± 0.2			0.65 ± 0.4	
~28					-1.0 ± 0.5	-0.3 ± 0.5	
~25					1.0 ± 0.5	0.8 ± 0.4	
25 – 10	0.3 ± 0.2		0.3 ± 0.1			1.4 ± 0.4	1 ± 0.5
10 - 5			1.9 ± 0.4			3.4 ± 0.14	2.5 ± 1 (1-3)
5 - 0			0.5 ± 0.1			3.9	0.6 ± 1
<i>Eastern margin of Altiplano and western margin of Eastern Cordillera</i>							
~40						0.0	
40 - 28		-0.5	1.6 ± 0.3			1.1 ± 0.5	
-28					-1.0 ± 0.5	0.1 ± 0.5	
~25					1.0 ± 0.5	1.1 ± 0.5	1.5 (5)
25 - 15	0.15±0.05	-0.12	1.2 ± 0.4			2.35 ± 0.4	~0.85 (5)
15 - 10		-0.12	0.4 ± 0.2			2.6 ± 0.3	
10 - 0		+0.25	1.0 ± 0.2			3.9*	1.7 ± 0.7 (7)
<i>Southern Profile</i>							
<i>Altiplano</i>							
~40						0.0	
40 - ~25	0.3 ± 0.25		0.4 ± 0.2			0.7 ± 0.5	
~28					-1.0 ± 0.5	-0.3 ± 0.6	
~25					1.0 ± 0.5	0.6 ± 0.5	
25 – 16	0.3 ± 0.25		0.7 ± 0.1			1.6 ± 0.5	3 ± 0.7 (4,6)
16 - 10			1.7 ± 0.6			3.3 ± 0.2	
10 - 0			0.6 ± 0.1			3.9	0.9 ± 1 (6)
<i>Eastern Cordillera</i>							
~40						0.0	
40 - ~25		-0.5	1.6 ± 0.3			1.1 ± 0.5	
~28					(-1.0 ± 0.5) ⁶	(0.0 ± 0.5)⁶	
~25					(1.0 ± 0.5) ⁶	(1.0 ± 0.5)⁶	
25 - 10		-0.25	1.0 ± 0.3			1.8 ± 0.5	
10 - 0		+0.25	1.2 ± 0.4			3.2*	1.7 ± 0.7 (7)

Notes: (1) Average sedimentation or erosion (sign indicates whether erosion used to calculate rock (+ve) or surface (-ve) uplift (see section 4.1); (2) see preferred 2-D shortening models in Table 1; (3) magmatic addition beneath arc based on inferred high density crustal root (see section 4.1); (4) inferred effect of mantle detachment and/or subduction pull-down, consistent with dynamical subduction models (see text); (5) Paleo-elevation of representative locality with 2σ uncertainty at end of time period (also plotted in Fig. 8), based on data in columns to the left and assuming initial elevation at 40 Ma was sea-level (* indicates this is the elevation of an outcrop); (6) western margin of Eastern Cordillera only?; (7) estimates of rock uplift for time period from published paleo-altimetry studies using calibrations discussed in this study – see Figs. 4 and 7 -with references in brackets (1 = Garzzone et al. 2006, 2 = Ghosh et al. 2006, 3 = Quade et al. 2007, 4 = Garzzone et al. 2008, 5 = Leier et al. 2013, 6 = Garzzone et al. 2014, 7 = Barke and Lamb 2006, 8 = Jordan et al. 2010).

Tables

Table 1. Calculated and observed crustal thickening and resultant surface uplift (U_s) in 2-D shortening models (see section 4.2) for Central Andes between 18°S and 21°S, using value of focusing factor γ that gives a good fit to observed paleo-elevations (see Table 2 and Fig. 8).

Table 2. Calculated (this study, see section 4) and observed uplift for Central Andes between 18°S and 21°S. Calculated paleo-elevations for representative localities shown in bold.

Figures

Figure 1. (a) Topography of the Central Andes between 13°S and 22°S, on the western margin of South America. Here, the Andes form a broad region, up to 700 km wide and reaching elevations of 6500 m asl. Virtually all this topography is the result of deformation since ~40 Ma. Red lines show location of lithospheric scale profiles, analysed in this study to determine the relation between changes in paleo-topography and geological processes within the Andes. Box shows region covered by Fig. 4. (b) Topography for 50 km wide swath along the northern profile, showing the main physiographic provinces. Note the relatively abrupt flanks of the high Andes, where elevations >3 km. (c) Topography for 50 km wide swath along the southern profile.

Figure 2. Lithospheric-scale cross-sections along (a) the northern and (b) southern profiles (see Fig. 1 for lines of profiles). The subducted Nazca plate, dipping at ~30°, underlies the western part of profiles, with a ‘hot’ overlying asthenospheric wedge where arc volcanism is generated. Also shown are main physiographic provinces, referred to in the text, the timings of crustal shortening across the width of the Andes, and the location of published paleoaltimetry studies (see Figs. 3 and 4). Note the structural vergence of major thrust structures, and the broad region beneath the Western Cordillera, Altiplano and western margin of the Eastern Cordillera that is inferred from high heat flow in thick crust to be underlain by ductile lower crust (ALVZ = Altiplano Low Velocity Zone, Yuan et al. 2000). Shortening in this ductile

crust accommodates underthrusting of the Brazilian Shield on the eastern margin, which acts as a sort of ‘piston’ pushing into the lower crust of the Andes farther west; crustal thickening creates the long-wavelength uplift of the high Andes since ~25 Ma (see text) – the western margin forms a monoclinical flexure, accommodating differential uplift between the Western Cordillera and Forearc. Major brittle faults in the Altiplano most plausible nucleate close to the brittle ductile transition. Note the location of published Cenozoic paleo-elevation studies discussed in the text (see Fig. 4 for explanation of reference numbers).

Figure 3. Space-time diagrams illustrating the Cenozoic geological evolution along (a) the northern and (b) southern profiles. Also shown are the stratigraphic positions of published paleo-altimetry studies. Note the presence of numerous unconformities, discrete episodes of crustal shortening in the Altiplano and Eastern Cordillera, as well as the accumulation of sedimentary sequences 5 – 8 km thick in both the Altiplano, eastern margin of the Eastern Cordillera, and Subandes, with local sedimentary basins since ~25 Ma within the Eastern Cordillera. Since ~25 Ma, the Brazilian Shield has been underthrust beneath the eastern margin of the high Andes. Note the stratigraphic context of published Cenozoic paleo-elevation studies discussed in the text (see Fig. 4 for explanation of reference numbers).

Figure 4. Map showing the main geological features of the Altiplano and parts of the Eastern Cordillera between ~16°S and ~22°S (see box in Fig. 1a for location). The lines of the northern and southern profiles are also shown. A major fold and thrust zone, referred to as the Corque-Corocoro structure, deforms Eocene or Oligocene to Pliocene red-bed sequences, several kilometers thick, deposited in a series of sub-basins (see cross-section in Fig. 5a). Deformation between ~10 and ~2.7 Ma, combined with erosion and sedimentation, has uplifted and smoothed out a Middle - Late Miocene paleo-depression, 0.5 – 1.5 km deep (with salar/lacustrine deposits in its centre) – the southern and northern margins of this paleo-depression are not well defined, but significant deformation prior to ~10 Ma, and presumably uplift, occurred farther south at ~20°S. Farther south, Cenozoic sequences as young as ~16 Ma are tightly folded near Tambo Tambillo, and along the southeastern part of the Salar de

Uyuni, but are unconformably overlain by essentially flat-lying volcanics and fluvial sequences deposited ~10 Ma (see cross-section in Fig. 5b). Note also the location of published Cenozoic paleo-elevation studies discussed in the text (1 = Garzione et al. 2006, 2 = Ghosh et al. 2006, 3 = Quade et al. 2007, 4 = Garzione et al. 2008, 5 = Leier et al. 2013, 6 = Garzione et al. 2014, 7 = Barke and Lamb 2006, 8 = Jordan et al. 2010, 9 = Gregory 2000, 10 = Graham et al. 2001).

Figure 5. (a) Cross-section through the Corque-Corocoro basin (see line AA' in Fig. 4 for location), based on surface mapping and dating studies, and an interpretation of oil company seismic data (with kind permission of Yacimientos Petroliferos Fiscales Bolivianos, and see also Lamb and Hoke 1997, Lamb 2011). The structure forms a regional syncline, bounded by NE and SW verging thrusts that cut red-bed sequences as young as ~5 Ma. The regional dips outside the structure point to an underlying SW-verging thrust, possibly sowing in the brittle-ductile transition at 15 - 20 km depth. See Lamb (2011) for seismic data used to constrain parts of the cross-section. Note the part of the syncline, projected along strike from the north (see Fig. 5a), sampled for isotope paleo-altimetry studies (Garzione et al. 2006, Ghosh et al. 2006). (b) Cross-section through the Tambo Tambillo region (see line BB' in Fig. 4, and also Lamb and Hoke 1997). Note the tight concentric fold structure, with subvertical, or overturned limbs. A regional extensive ~23 Ma basaltic sill and mafic to felsic volcanics, with overlying red beds as young as ~16 Ma, are tightly folded in the core of the Tambo Tambillo syncline.

Figure 6. Shaded topographic map of the Central Andes, between 13°S and 23°S, showing the location of Cenozoic volcanic provinces, as well as the region, covering most of the high Andes, with detectable mantle helium emissions in geothermal systems ($>0.5 R/R_a$, Hoke and Lamb 2007). The latter indicates the extent of active mantle melting, most plausibly marking the region with thin lithosphere today (Hoke and Lamb 2007). Note the location of the profiles analysed in this study (thick grey lines), and the active volcanic arc, following the western margin of the high Andes. Felsic to intermediate shallow level intrusions, volcanic stocks, and ignimbrites, emplaced or erupted between 26 and 2 Ma, are found on the western margin of the Eastern Cordillera, more-or-less within the region of active mantle melting indicated

by helium emissions - magmatic activity here prior to ~26 Ma is rare. (b) Plot of age against distance perpendicular to the active arc of Neogene volcanic and plutonic rocks in the Andes between 16° and 22°S (compiled from Everden et al. 1977, Grant et al. 1978, Schneider 1985, Redwood and MacIntyre 1989, Kennan et al. 1995, Barreiro 1998, Woerner et al. 2000, Barke 2004, Hoke and Lamb 2007). EC = Eastern Cordillera. Arc and widespread behind arc volcanism started abruptly at ~26 Ma, after a ~10 Ma period of arc quiescence. Since then, there has been more-or-less continual felsic to mafic volcanism extending up to ~300 km behind the arc. Note the ~50 km westward migration of the eastern limit of magmatism. Also plotted is normalised $^3\text{He}/^4\text{He}$ in geothermal emissions along the southern profile, showing the pronounced mantle helium signature right across the high Andes, in the same region where there is behind arc dacitic-rhyolitic volcanism. This strongly suggests that the heat source for this volcanism is mantle melts at depth in the crust, which only occasionally erupt at the surface as basalts.

Figure 7. Published oxygen isotope and growth temperature data for water in equilibrium with Cenozoic (28 – 5 Ma) paleosol carbonates in the Bolivian Andes, at sites with elevations ranging between 3.7 and 3.9 km (see Figs. 2, 3, 4 for locations, Ghosh et al. 2006, Garzzone et al. 2006, 2008, Leier et al. 2013, Garzzone et al. 2014). Data show a marked trend, indicated by bold line, from ‘high’ temperature and ‘heavy’ oxygen isotopes to ‘low’ temperatures and ‘light’ oxygen, relative to SMOW. This trend also reflects a general increase in paleo-elevation, although the precise paleo-elevation depends on the calibration used. Short lines (heavy, dashed, thin dashed) show the inferred paleo-elevations along the trend line, using the calibrations of Quade et al. (2007), Ehlers and Poulsen (2009), and Garzzone et al. (2014). Garzzone et al.’s (2014) calibration is referred to as the southern calibration (Fig. 8), as it depends on a comparison with low elevation (~0.4 km) Late Miocene paleosols at ~23°S, whereas Ehlers and Poulsen’s (2009) is referred to as the northern calibration.

Figure 8. Plots showing the elevation evolution for parts of the high Andes along (a) the northern profile at ~17.5°S, (b) southern profile at ~20°S, and (c) slightly farther south at ~21°S. Paleo-elevations from published paleo-altimetry studies are plotted, using either the northern or southern calibrations shown in Fig. 7. Coloured lines

show uplift trajectories (with 2σ uncertainties), calculated in this study for different regions (green = Western Cordillera, red = Altiplano, blue = Eastern Cordillera), based on published estimates of crustal shortening (using ‘best-fit’ values of the focusing factor γ in Tables 1 and 2), erosion, sedimentation and magmatic addition, together with inferred subsidence/uplift between 28 and 25 Ma related to changes in lithospheric geometry (see Tables 1 and 2, and section 4). Note that the calculated trajectory at 21°S for the Eastern Cordillera since ~25 Ma has but shifted upward 300m to match the modern day elevation; the one for the Altiplano is the same as that for 20°S .

Figure 9. A series of cartoons illustrating the inferred lithospheric evolution for the Central Andes between 16° and 21°S . (a) In the Paleocene (~60 Ma), volcanic activity was mainly confined to the arc, with a broad region behind (east of) the arc, >400 km wide, of subsidence. Deformation and uplift in the behind arc region commenced at ~40 Ma. (b) Between ~37 – ~26 Ma, arc volcanism shut off, most likely caused by the flattening of the subducted slab. This way, the negative buoyancy of the subducted slab ‘pulled down’ the overlying lithosphere ~1 km, counteracting much of the elevation increase due to crustal thickening. The Altiplano formed a wide region, close to sea level, between the arc (Western Cordillera) and ‘low’ Eastern Cordillera. (c) Significant Andean lithospheric shortening in the Eastern Cordillera and Altiplano ceased at ~28 Ma. Arc volcanism, together with widespread behind arc mafic - felsic volcanism, resumed abruptly at ~26 Ma, most plausibly triggered by creation of an asthenosphere wedge as the subducted slab steepened (possibly with detachment of the lower part of the lithosphere). This decoupled the Andean lithosphere from the subducted plate, resulting in ~1 km of rapid uplift and the creation of thin lithosphere beneath the high Andes, which subsequently has remained largely unchanged. (d) Intense shortening was focused beneath the Altiplano, with rapid episodes, lasting < 5 Myrs, in the northern Bolivian Altiplano (along the northern profile) between ~9.5 and ~5.4 Ma, and in the central Altiplano (along the southern profile) between ~16 and 10 Ma. This, together with lower crustal ductile shortening beneath most of the width of the high Andes, driven by underthrusting of the Bolivian Shield on the eastern margin, resulted in km-scale uplift of the high Andes so that the Western

(volcanic arc) and Eastern Cordilleras merged. Widespread behind arc mafic – felsic volcanism was more-or-less continuous throughout this period (see Fig. 6b).

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