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Paleosurfaces, paleoelevation, and the mechanisms for the late Miocene topographic development of the Altiplano plateau

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ABSTRACT

We construct a topographic datum across the central Andes between 18°S and 22°S using extensive, well-preserved paleosurfaces that span the entire range. This datum, combined with estimates of topographic change from paleoaltimetry in the Altiplano and geomorphic relief generation estimates from the plateau flanks allow us to reconstruct the paleo-topography of the range at ~10 Ma. Using the 10 Ma topography we explore which geophysical processes can create the observed topographic change over the <4 Ma period between 10 and 6 Ma. Our results indicate that loss of a dense lithospheric root, and possibly contemporaneous crustal flow, are the most likely processes that produced the observed plateau-wide topographic change. Raising the plateau purely by late Miocene crustal thickening would require shortening rates four times greater than those reported for the Cenozoic.

1. Introduction

Elevation is a key parameter in geophysical observations of modern and ancient earth systems. Accurate geodetic observations on the modern earth surface are straightforward given a well-defined datum. Paleoelevation and extensive paleo-datums, despite being critical parameters, are much more difficult to constrain (Wolfe and Schorn, 1989; Molnar and England, 1990; England and Molnar, 1990). Much of our knowledge of how mountains build spatially over time is derived from studies of the deformation and sedimentation within and surrounding any given mountain range. The timing of upper crustal shortening constrains deformation history, and the proliferation of geometrically balanced cross sections (Suppe, 1983) provides a means for estimating the amount of displacement (i.e., crustal shortening) occurring in fold-and-thrust belts over time. Upper crustal shortening and crustal thickening histories have been suggested to record the topographic evolution of mountain ranges (e.g. McQuarrie, 2002a), which in turn are used to place constraints on results of models of lithospheric evolution. However, it remains unclear whether this is actually the case as lateral flow of a ductile lower crust (Royden et al., 1997; Husson and Sempere, 2003; Clark and Royden, 2000) and removal of a dense lower crust and mantle lithosphere (Kay and Kay, 1993; Kay et al., 1994) are also viable candidates for generating substantial topographic change in mountain ranges and particularly continental plateaus (Fig. 1).

The central Andean plateau is the type-example of a high continental plateau formed in a non-collision tectonic setting. Subduction of oceanic lithosphere along the western margin of South American began in the Jurassic, and the Andean phase of deformation began in the late Eocene (Mpodozis and Ramos, 1989; Elger et al., 2005; McQuarrie et al., 2005; Benavides-Caceres, 1999). The Cenozoic dynamics of subduction along the western margin of South America imparts the largest control on the style and tempo of continental deformation (Sobolev and Babeyko, 2005; Jordan et al., 2001; Iaffaldano et al., 2006). Changes in convergence direction and velocity have also been linked to major changes in the tectonics of the central Andes. Maxima in plateau width and cumulative shortening (Isacks, 1988) in the central Andes occurs about Gephart's (1994) axis of symmetry for the topography and the shape of the subducting Nazca plate. These maxima are contained within the area considered in this study, between 18°S and 22°S.

In this paper we use paleosurfaces to unite several previously published estimates of paleoelevation, and relief generation into a single common topographic datum. These combined observations allow for a plateau-wide examination of the geodynamic factors responsible for the observed elevation history of the central Andean plateau (Fig. 2) (the area of high topography comprising the Western Cordillera, Altiplano,

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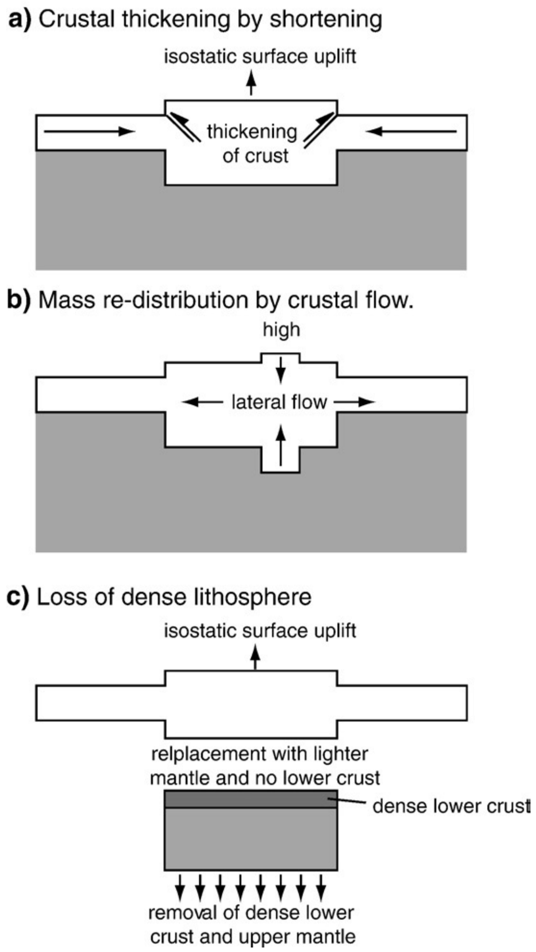

Fig. 1. Conceptual diagrams showing possible mechanisms for large magnitude surface uplift of the central Andean plateau described in the text. a) Shortening as a means for crustal thickening and b) lower crustal flow would result in isostatically induced surface uplift related to the thickening of the felsic (upper) continental crust. c) Loss of a dense lithospheric root would result in positive (upward) isostatic rebound.

and Eastern Cordillera) and its flanks between 18°S and 22°S. We combine previously published, quantitative and temporally similar observations of surface uplift based on stable isotope estimates of paleoelevation (Garzione et al., 2006, 2007) for the center of the plateau and quantitative geomorphic estimates of relief change on the flanks (Barke and Lamb, 2006; Hoke et al., 2007). Advances in our ability to measure paleoelevations (Garzione et al., 2006; Poage and Chamberlain, 2001; Quade et al., 2007; Rowley and Garzione, 2007; Rowley et al., 2001; Rowley and Currie, 2006; Ghosh et al., 2006) allow for a more comprehensive look at the factors that generate elevation change in mountain ranges. These data provide the most robust determinations of regional land surface paleoelevations to date. While, absolute paleoelevation determinations are typically single point observations, made in areas containing rocks suitable for this type of analysis, river incision based geomorphic measures of tectonically generated relief creation provide a regional-scale picture of relative elevation change that is not tied to any well-defined datum such as sea level. The ability to unify both types of data within a common datum, both in time and space, provides a powerful tool for reconstructing the paleotopography of entire mountain ranges that goes beyond extrapolating from disparate point sources of information.

Rapid surface uplift of the central Andean plateau, on the scale of kilometers, is constrained by radiometric and paleomagnetic ages to have occurred between ~10 Ma and present across the entire region (Garzione et al., 2006, in press; Barke and Lamb, 2006; Hoke et al., 2007, Ghosh et al., 2006; Gubbels et al., 1993; Schildgen et al., 2007; Kennan et al., 1997). Here we also combine and use well-preserved, range-wide late Miocene paleosurfaces as a datum in creating a reconstruction of the topography of the central Andes at 10 Ma. With this 10 Ma paleo-topography, which spans the entire central Andes, and basic lithospheric-scale flexural modeling, we explore which geodynamic processes (Fig. 1) are most likely to have resulted in the observed surface uplift of the Altiplano between 10 and 6 Ma.

2. Crustal shortening between 18–22°S

Crustal shortening across the entire Western Cordillera, Altiplano and Eastern Cordillera (central Andean plateau) between 18°S and 22°S, has been documented by numerous studies (Elger et al., 2005; Farías et al., 2005; Victor et al., 2004; McQuarrie, 2002b; Muñoz and Charrier, 1996; Kley, 1996; Herail et al., 1994; Lamb and Hoke, 1997). Estimates of the amount of crustal shortening across the entire orogen vary between 300 km (Allmendinger et al., 1997) and 500 km (McQuarrie, 2002a). The most comprehensive compilation of shortening rate data versus age are from Oncken et al. (2006) and McQuarrie et al. (2005). Elger et al.'s (2005) data show that the plateau and the flanking cordilleras have largely asynchronous shortening histories, and that shortening rate is in decline within the Altiplano, Eastern and Western Cordilleras by 10 Ma dropping to rates of <1 mm/yr (Fig. 3). As deformation across the plateau and especially in the Eastern Cordillera ceased at 10 Ma, it begins in the Sub-Andean belt (Allmendinger et al., 1997; Echavarria et al., 2003) (Fig. 2). The maximum shortening rate reported for the Altiplano and cordilleras occurs in the Eastern Cordillera domain and is 8 mm/yr at ~28 Ma. The maximum shortening rate reported for the Sub-Andean range is 11 to 12 mm/yr (McQuarrie et al., 2005; Echavarria et al., 2003) (Fig. 3). Cumulative shortening rates across the entire plateau and Sub-Andean ranges have maxima at 28 Ma (8 mm/yr), 12 Ma (8 mm/yr), and 8 Ma and 3 Ma (11 mm/yr) (Elger et al., 2005; Echavarria et al., 2003). McQuarrie et al. (2005) present a less episodic shortening history with long-term shortening rates of 8–10 mm/yr pre-Miocene and 5–7 mm/yr post Miocene. If the topographic evolution of the central Andean plateau is linked solely to the spatial and temporal distribution of crustal shortening, it should have attained nearly all of its modern elevation before 10 Ma (Elger et al., 2005; Victor et al., 2004; Muñoz and Charrier, 1996), a conclusion at odds with fossil plant physiognomic studies (Gregory-Wodzicki, 2000), transient landscape formation in the eastern and western flanks (Barke and Lamb, 2006; Hoke et al., 2007; Gubbels et al., 1993; Mortimer, 1980), and stable isotope paleoaltimetry (Garzione et al., 2006, Ghosh et al., 2006). Geophysical studies (Oncken et al., 2003; Beck et al., 1996) reveal that crustal thicknesses beneath the central Andean plateau are on the order of 60–70 km, consistent with significant crustal thickening and in good agreement with results derived from crustal-scale balanced cross section results (McQuarrie, 2002b; Hindle et al., 2005).

3. Range-wide late Miocene paleosurface

Prominent paleosurfaces are present throughout the central Andes, especially on the western flank of the orogen in Northern Chile and Peru. Bowman (Bowman, 1916) was the first to identify the Puno Surface of SW Peru. Later studies identified several other surfaces in Chile, Peru and Bolivia (Gubbels et al., 1993; Kennan et al., 1997; Clark et al., 1967; Mortimer and Saric, 1972; Mortimer et al., 1974; Lavenu, 1986) Here, our use of paleosurface applies to a landform or landscape, which formed under different geomorphic conditions than those of its current state. The paleosurface reflects different relief and climate conditions than those operating in the modern, resulting in a temporary or 'transient' condition as rivers re-adjust the landscape to reflect these conditions (Whipple et al., 1999). In theory, the "old" landscape operates under its former relief conditions until the wave of incision related to the new

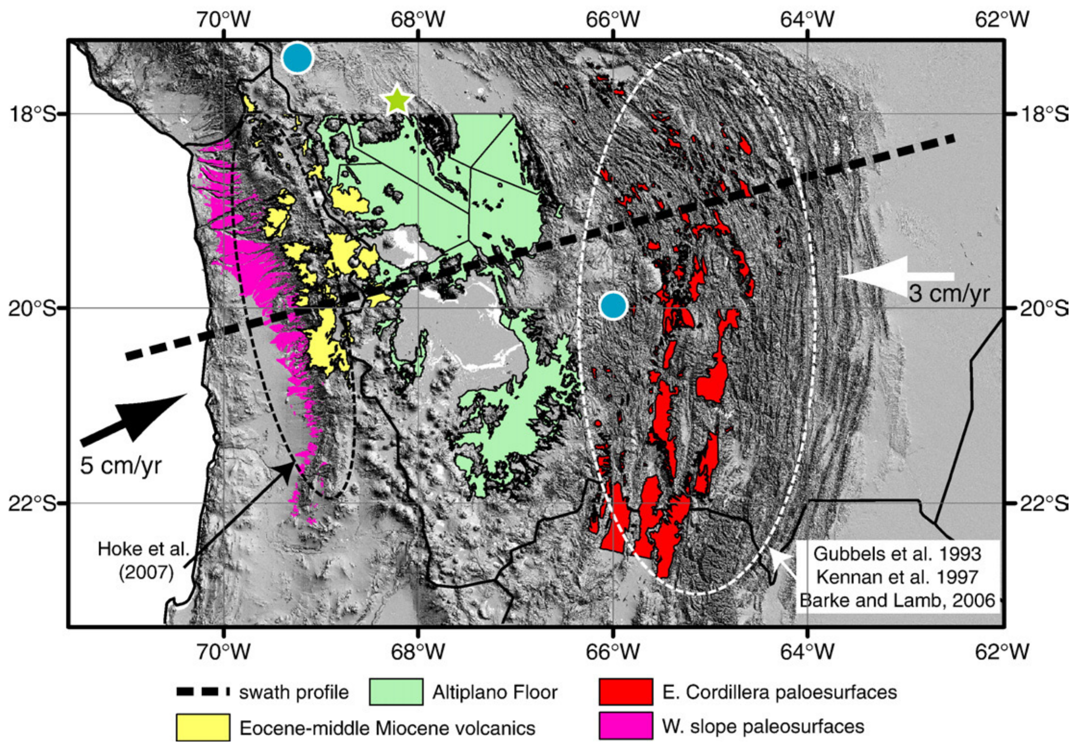

Fig. 2. Shaded relief map of the central Andes of the study area and locations of the surface uplift estimates used in this study. Studies by Hoke et al. (2007) and Barke and Lamb (2006) are shown by the dashed line circled areas, Garzoine et al. (2006, 2007) are shown with a green star and Gregory-Wodzicki (2000) and Gregory-Wodzicki et al. (1998) sample sites are marked with filled blue circles. Black lines show political boundaries. Thick arrows show the relative velocities and direction of the Nazca and South American plates. Mapped paleosurfaces from the Eastern Cordillera (red) (Gubbels et al., 1993; Kennan et al., 1997) and Western slope (magenta) (Hoke, 2006). Pre-late Miocene constructional volcanic features, which constituted part of the 10 Ma landscape, selected from geologic maps (Makepeace et al., 2002; Gubbels, 1993) and used in this study are shown in yellow. The modern floor of the Altiplano basin, used here as a proxy for the 10 Ma surface of the Altiplano is indicated on the map by the light green areas. The dashed black line marks the location of a 100 km wide swath profile taken in 5 km increments that is used in Figs. 4–6. The location of the profile line was chosen to bisect the range near the location of the Gephart (1994) symmetry plane and maximize sampling of the paloesurfaces.

relief conditions migrates through the entire landscape. The dominance of semi-arid to hyper-arid climate conditions across our study area is largely responsible for the high degree of paleosurface or transient landscape preservation. This extensive preservation also implies that, that post-10 Ma erosion and mass transfer out of the central Andean plateau, toward the foreland and forearc, is relatively small (see below).

We combined several previously published data sources to create a map of paloesurfaces that were part of the late Miocene landscape across the cenral Andean orogen between 18 and 22°S (Fig. 2). Here we add two additional paleosurfaces not considered in prior studies, constructional volcanic features >10 Ma and the floor of the Altiplano Basin. The range of ages of these paleosurfaces are based on stratigraphic age constraints with dated volcanic horizons, above and below the regionally extensive transient landscapes of the western Andean mountain front of northern Chile (Hoke et al., 2007; Mortimer et al., 1974; Hoke, 2006; Sillitoe et al., 1968) and the Eastern Cordillera of central and southern Bolivia (Gubbels et al., 1993; Kennan et al., 1997) are formed at the same time, between ~12 and 6 Ma (Fig. 2). The criterion for inclusion in the paleosurface map is an approximate surface age of ≥10 Ma or a region of minor deposition or tectonism, as is the case for the modern Altiplano basin.

The western flank of the central Andean plateau contains large tracts of an approximately 10 Ma relict depositional surface of the El Diablo and Altos de Pica formations (Hoke et al., 2007; Farías et al., 2005; García and Hérail, 2005) (magenta area in Fig. 1). This depositional surface formed under much lower average slope conditions (<1°) than those of today (2°–3°). Other studies demonstrate that locally, this surface is slightly younger (Hoke et al., 2007; von Rotz et al., 2005; Kober et al., 2006]. Nearby Ne terrestrial cosmogenic nuclide dating by Evenstar (Evenstar et al., 2005) yield stratigraphically impossible surface ages of 14–16 Ma because the cosmogenic age is older than the dated tuffs that flank the paleosurface. This surface was mapped in detail by Hoke (2006) using a combination of Landsat TM images, 90 m and 20 m digital elevation data, and geologic maps (SERNAGEOMIN, 2002). The El Diablo-Altos de Pica surface is spatially continuous and ranges in elevation from 500 m to 4000 m.

The next paleosurfaces to the east are Eocene to middle Miocene constructional volcanic features of the Western Cordillera identified in geologic maps (SERNAGEOMIN, 2002; Makepeace et al., 2002) which are not substantially eroded (yellow areas in Fig. 1). Because of their age, these volcanic rocks formed part of the 10 Ma landscape and therefore should be included with the paleosurfaces to the east and west. These constructional volcanic features are so well-preserved in some cases that, for example, Mortimer et al. (1974) determined Miocene ages for a volcano they expected to be Quaternary in age.

We chose the modern floor of the Altiplano (green areas in Fig. 1; mapped as Qal in Makepeace et al., 2002) as a proxy for the late Miocene paleosurface. This choice is made on the basis that there has been little sedimentation (<500 m) throughout most of the Altiplano after 10 Ma and that no significant post-10 Ma regional angular unconformity exists in seismic data (Elger et al., 2005; Allmendinger et al., 1997). Therefore, we deduce that the modern surface serves as a reasonable approximation for the surface at 10 Ma.

The paleosurfaces of the Eastern Cordillera (red areas in Fig. 2) are a spectacular example of a transient landscape. The southern, and most spatially extensive part of this surface is referred to as the San Juan del Oro surface (Gubbels et al., 1993). The San Juan del Oro surface comprises a complete landscape with highlands (hills), lowlands (valleys) and small sedimentary basins, which today are being actively incised by the modern Pilcomayo and San Juan del Oro Rivers (Kennan et al., 1997; Gubbels, 1993). The surface is only slightly affected by faults related to the Eastern Cordillera. It is likely that formation of the landscape

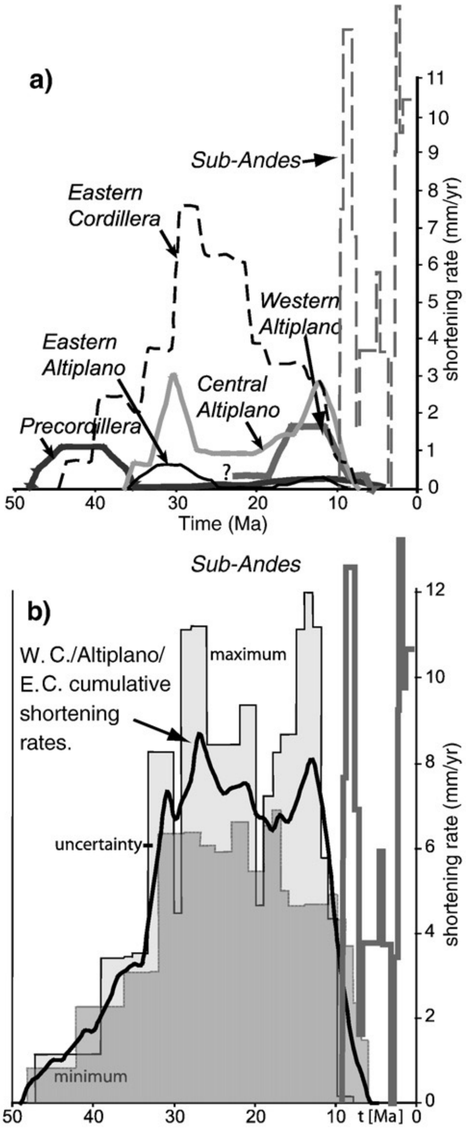

Fig. 3. a) Compilation of upper crustal shortening rates versus time across the Altiplano and its flanking [modified from Elger et al., 2005] and the minimum duration shortening interpretation in the Sub-Andean range (Echavarria et al., 2003). As shortening decreases in the plateau area, it initiates in the Sub-Andean range. b) Cumulative shortening rates for the Western Cordillera, Altiplano and Eastern Cordillera from Elger et al. (2005). The same curve for the sub-Andes from A is plotted for reference, but not combined with cumulative rates for the rest of the range Elger et al. (2005).

occurred over the middle to late Miocene, with the initiation of uplift related incision beginning between 10 and 6 Ma (Kennan et al., 1997). Geomorphic analysis by Barke and Lamb (2006) suggest that the San Juan del Oro paleolandscape developed with a maximum basin relief of no more than 2000 m. We remapped the paleosurfaces of the Eastern Cordillera utilizing the original maps of Gubbels (1993) and Kennan et al. (1997) and slightly expanded upon their mapping using the classification criteria of the original studies combined with transparent overlays of local slope, digital topography and Landsat TM (images from the University of Maryland's Global Land Cover Facility).

Using a mask of the areas mapped as paleosurfaces, the corresponding elevations were extracted from the 90 m Shuttle Radar Topography Mission (SRTM) DEM. The heavy gray line in Fig. 4 is a profile showing the average topography of the paleosurfaces in a 100 km × 5 km windows centered along the dashed line in Fig. 3. The prominent data gap between 500–550 km corresponds to the ~7 Ma Los Frailes ignimbrite shield area. Given the high percentage of paleosurfaces that exist in this region, the average topography across the Western Cordillera and Altiplano is very similar to the average modern topography (heavy gray line in Fig. 4). In the Eastern Cordillera the average paleosurface topography is about 500–200 m higher than the average modern topography, and supplies a rough estimate of the average landscape lowering since 10 Ma, which is in good agreement with the estimate of Barke and Lamb (2006). This small amount of landscape lowering across the entire region indicates that the isostatic response to the removal of eroded material post-10 Ma is only on the order of tens of meters. The amount of material removed below the paleosurfaces across the rest of the study area is negligible.

4. Paleoelevation

Our current understanding of elevation change across the central Andean plateau between 18°S and 22°S comes from determination of paleoelevations from fossil leaf margin analysis and stable isotope studies. Relief generation gleaned from the analysis of river profiles is an indirect measure of elevation change. Geomorphic indicators on the flanks of the central Andes, mostly river incision and tilting of large depositional surfaces, have long been cited as evidence for post 10 Ma surface uplift of the Altiplano (Gubbels et al., 1993; Mortimer, 1980; Lavenu, 1986; Tosdal et al., 1984) and were used as primary evidence by Isacks (1988) for a second stage of surface uplift of the central Andean plateau (Table 1). However, because these studies lacked the ability to quantify the amount of topographic change, they have been overlooked as indicators of surface uplift in favor of more easily quantified crustal shortening estimates. Gregory-Wodzicki et al. (1998) and Gregory-Wodzicki (2000) used fossil leaf physiognomy to provide the first quantitative estimates of paleoelevations for the central Andes (Table 1). Despite the large uncertainties in the estimates, they supported the notion that the plateau was at approximately half of its present elevation by 10 Ma.

Stable isotope paleoaltimetry in the Corque syncline of the Bolivian Altiplano suggests ~2500 m ± 1000 m of surface uplift in the period between 10 and 6.4 Ma (Garzione et al., 2006, in press; Quade et al., 2007; Ghosh et al., 2006). In these studies the $\delta^{18}O$ values of soil carbonates and palustrine carbonates are taken to reflect the isotopic composition of local meteoric waters at the time of their formation and the Δ_{47} values of paleosol carbonates record the temperature of carbonate precipitation. Elevations are derived by comparing the carbonate $\delta^{18}O$ and Δ_{47} values to those of modern meteoric river waters and surface temperature at different elevations. These calculations assume the same $\delta^{18}O$ versus altitude gradient and land surface temperature lapse rate as today. The low elevation starting temperature was shifted to a higher temperature to reflect the warmer global climate in middle to late Miocene time (Quade et al., 2007; Ghosh et al., 2006). The uncertainties in determining paleoelevations from stable isotopes are beyond the scope of this paper, but for a detailed treatment of this see Rowley and Garzione (2007) and Quade et al. (2007).

Analysis of rivers that drain the western flank of the plateau in northernmost Chile show the consistent presence of knickpoint bounded transient profiles. Using stream power based river modeling, these transient profile segments are projected to their former downstream outlet elevations. This modeling demonstrated that a minimum of 1080 ± 230 m of relief was generated on the western flank of the Altiplano between 10 Ma and present (Hoke et al., 2007). In the Eastern Cordillera of Bolivia, Barke and Lamb (2006) used several types of geomorphic analysis to constrain the original elevation (relief conditions) of the San Juan del Oro paleosurfaces arriving at an estimate of 1705 ± 695 m of rock uplift between 10 and 2.7 Ma.

4.1. Late Miocene paleosurface and paleotopography of the central Andes

The average topography of the 10 Ma paloesurfaces across the range can be combined with the relief estimates from the plateau flanks (Barke

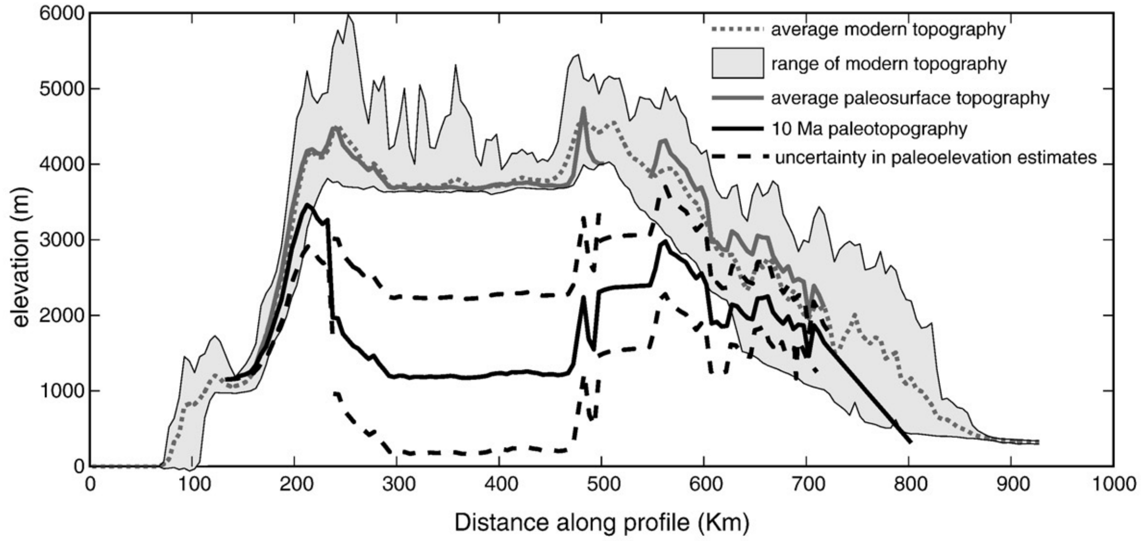

Fig. 4. Swath topographic profiles derived from 90 m SRTM data for a 100 km wide swath in 5 km steps centered on the dashed black line in Fig. 2. Thin black lines and filled light gray area represent the range of elevations between the maximum and minimum values for the modern topography. The gray lines show the average modern topography for the entire dataset within the swath (dashed) and the topography that only represents the 10 Ma paleosurface (solid). The topography of the central Andes at 10 Ma (heavy black line) and based on lowering the average topography of the 10 Ma paleosurfaces according to the estimates given by studies in the Altiplano (Garzione et al., 2006), Eastern Cordillera (Barke and Lamb, 2006) and western Andean mountain front (Hoke et al., 2007) (see text). The uncertainties given here (dashed heavy black line) are based on the uncertainties reported by each author.

and Lamb, 2006; Hoke et al., 2007) and the paleoaltimetry from the center of the plateau (Garzione et al., 2006, in press; Quade et al., 2007; Ghosh et al., 2006) to create a paleo-topographic profile across the range at 10 Ma. In this reconstruction of the 10 Ma topography from relief and paleoelevation estimates, we treat the Coastal Cordillera in the Chilean forearc and the foreland east of the modern Sub-Andean range elevation as pin-points, or sites of no elevation change, relative to the plateau and its flanks. The gaps in the paleosurface data were filled by connecting the Altiplano and Eastern Cordillera and the Eastern Cordillera foreland with linear interpolations of elevations from the endpoints of the profiles (Fig. 4). We divide the range into 3 domains, Western Cordillera, Altiplano and Eastern Cordillera. We use the more conservative minimum estimates of elevation change from Garzione et al. (2007) and Quade et al. (2007) to lower the entire Altiplano domain by 2500 m. The eastern and western flanks of the plateau were lowered by their corresponding amounts along decreasing linear ramps from their crests to the pin-points in the forearc and foreland. The resulting topography is shown in Fig. 4 with the nominal uncertainties (thick dashed lines in Fig. 4) assigned by each set of authors.

5. Cause of rapid surface uplift?

The nearly synchronous observations of surface uplift across the entire width of the range suggest the effect of some regional scale geodynamic process(es). Various studies, mostly on the flanks of the Altiplano, have ascribed elevation change to local effects. However, with our more complete view of the topography of the central Andes at 10 Ma, a more comprehensive treatment of plateau evolution is possible. There are three viable processes that could result in widespread, rapid surface uplift: crustal shortening, lower crustal flow, lower lithospheric loss, or some combination of the three processes. Our modeling results, presented below, are in accord with the conclusions of several studies that require the removal of dense lithosphere to explain current elevation of the Andes (Sobolev and Babeyko, 2005; Garzione et al., 2006; Ghosh et al., 2006; Beck and Zandt, 2002; Hoke et al., 1994; Molnar and Garzione, 2007).

5.1. Role of crustal shortening

If 2.5 km ± 1 km of regional surface uplift occur in the central Andes between 10 and 6 Ma as suggested by the combined paleoelevation estimates, then surface uplift resulting only from upper crustal shortening is an unlikely single candidate. As discussed above, the only place where significant upper crustal shortening took place in the Andes at this time is in the Sub-Andean fold-and-thrust belt. Taking a purely crustal shortening approach, we estimate the amount of crustal shortening necessary to arrive at today's average elevations from the paleotopography of the range at 10 Ma (Fig. 4). Using the 10 Ma topographic profile, we estimate the thickness of an isostatically compensated lithospheric root (Turcotte and Schubert, 2002)

Table 1
Data sources used in this study

Type	Method	Location	Value	Error	Units	Reference
Paleoelevation	O stable isotopes	S. Bolivia	2500	±1000	m	Garzione et al. (2006, 2007), Quade et al. (2007), Rowley and Garzione (2007)
Paleoelevation	Fossil plant physignomy	S. Bolivia	2000	±2000	m	Gregory-Wodzicki (2000)
Relief creation	River profiles	N. Chile	1180	±250	m	Hoke et al. (2007)
Relief creation	River profiles	E. Cordillera	1700	±600	m	Poage and Chamberlain (2001)
Paleosurface	Stratigraphic	E. Cordillera	10–6	Varies	Ma	Schildgen et al. (2007), Farías et al. (2005)
Paleosurface	Stratigraphic	N. Chile	12–5	Varies	Ma	Hoke et al. (2007), Lavenu (1986), Sillitoe et al. (1968), Kober et al. (2006)
Paleosurface	Mapping/stratigraphic	N. Chile/S. Bolivia	>10	Varies	Ma	This study,
Shortening	Balanced cross section	N. Chile to E. Cordillera	40–7		Ma	Elger et al. (2005)
Shortening	Balanced cross section	Sub-Andean range	9–0		Ma	Gregory-Wodzicki (2000)
Shortening	Balanced cross section	E. C. and Sub-Andes	40–0		Ma	McQuarrie et al. (2005)

assuming an initial zero elevation crustal thickeness of 35 km. We then sum the thickness of the root with the 10 Ma topography and the assumed initial continental crust thickness of 35 km to obtain a total crustal thickness at 10 Ma. Taking the same approach with the modern topography and subtracting it from the 10 Ma topography, we calculate the difference in area between the profiles of 5.57×10^9 m^2. Assuming that the principal post-10 Ma input of crustal material into the orogen is from the thrusting of the foreland and the Brazilian Shield of thickness = 35 km, over a period of 4 Ma (the maximum period over which surface uplift occurred in the Altiplano), a shortening rate of 40 mm/yr is necessary to arrive at today's observed elevations. This rate is ~4 times greater than the maximum long-term shortening rates observed at any given time since the Oligocene (McQuarrie et al., 2005). This estimate does not consider the potentially important effect of thickening eclogitic lower crust and/or mantle lithosphere and therefore should be considered the minimum required shortening rate. Calculating the necessary shortening rates to generate 2.5 km of surface uplift over a 10 Ma time period yields rates near the maximum observed rates, but still substantially higher than the average shortening rate in the Sub-Andean range over the last 10 Ma. In either case, crustal shortening alone cannot account for raising the plateau.

5.2. Crustal flow

Crustal flow, particularly lower crustal flow has been proposed as an important mechanism in plateau formation for the Andes (Husson and Sempere, 2003; Isacks, 1988) and Tibet (Clark and Royden, 2000). The flow of crustal material in the Altiplano is most likely not restricted to the lower crust based on geophysical evidence for partially molten material beneath the plateau (Oncken et al., 2003; Beck and Zandt, 2002; Chmielowski et al., 1999) and the extremely low effective elastic thickness across the plateau of 5–8 km (Tassara et al. 2005). Isacks (1988) first suggested the idea of a 'hydraulic ram' related to the underthrusting of the foreland and shortening of the ductile lower crust. This type of crustal thickening can account for some of the elevation change observed in the Andes. Using the more recent estimates of shortening in the Sub-Andes (Echavarria et al., 2003), Garzione et al., (2006) calculate that this process could account for ~1 km of surface uplift. Hindle et al. (2005) invoked extensive east-west and north-south mass re-distribution within the plateau to explain discrepancies in the results of structural modeling with observed crustal thickness. Husson and Sempere (2003) proposed the flow of ductile crustal material from the topographic highs of the western and eastern cordilleras towards the center of the plateau as an means for growing the plateau interior with minor local crustal shortening. In principle these two models for crustal flow present attractive solutions. However if flow of ductile material occurs over relatively short timescales (10^4–10^5 a), it becomes necessary to link the timing of crustal flow to times of crustal thickening. This would result in a much older Altiplano, since the maximums in crustal shortening for both of the cordilleras is Oligocene in age (Fig. 2). In addition, lower crustal flow cannot account for the simultaneous surface uplift of the cordilleras as is indicated by the rotation and incision of the flanks of the Andes. Rather, by lower crustal flow, the cordilleras should have remained at roughly constant elevations, or subsided as the Altiplano rose.

5.3. Removal of lower lithosphere

The loss of a preferentially dense part of the lower lithosphere (i.e. an eclogitized lower crust and the underlying lithospheric mantle; 'delamination' of Kay and Kay (1993)) is another process that should have a regionally extensive and possibly large topographic effect (Platt and England, 1993). The transition from granulitic lower crust to a denser eclogitic lower crust requires crustal thicknesses >50 km and the presence of water. Hetenyi et al. (2007) explored the necessary T-P and water contents conditions necessary to form eclogite in the lower crust at the southern end of the Tibetan Plateau. Aside from the presence of a density-based instability located in the lower crust, there is no consensus on the physical processes in which the lower crust and mantle lithosphere are removed. The different theories on the physical processes behind lithospheric loss range from ablative subduction (Pope and Willett, 1998) to rapid detachment by delamination (Kay and Kay, 1993; Kay et al., 1994), or the Rayleigh–Taylor instability driven drip model (convective removal) (Molnar and Houseman, 2004). Ablative subduction implies a steady loss of material over time and is unable to produce rapid elevation change. In the flexural modeling scenarios described below, we evaluate the effects of rapid removal of the lower lithosphere by rapid delamination or by convective removal below the Altiplano-Eastern Cordillera. The modeling results, assuming either a convective removal or delamination process, are the same. Thus, through the modeling alone we cannot ascribe a specific mechanism to the removal process. The only difference is that the drip model may result in remnants of lithospheric mantle attached to the crust (Molnar and Houseman, 2004). We propose that the drip model best reconciles the available observations with the current topography and geophysical state of the study area in that lower lying regions, such as the Salar de Uyuni, appear to have remnant high density mantle lithosphere and/or eclogite beneath them (Beck and Zandt, 2002). This convective removal process is also consistent with empirically derived flow laws for eclogite and mantle minerals based on required temperature of the Moho and mean stress within the mantle lithosphere (Molnar and Garzione, 2007).

6. Flexural modeling of lithospheric loss

Simple 1-D flexural modeling was employed to model the topographic effects of removal of dense lower lithosphere. We also considered the variations in crustal strength by using the variations in the effective elastic thickness (EET) of the crust according to the profiles of Tassara et al. (2005) and Tassara and Yañez (2003) (Fig. 5). Across the central Andes between 18 and 22°S the plateau is extremely weak (EET between 8 and 5 km) and responds in a manner similar to simple Airy isostasy, while the margins are strong (EET between 30 and 75 km) (Tassara et al., 2005; Watts et al., 1995). We performed flexural modeling using the program tAO which allows for variations in the effective elastic thickness parameter (García-Castellanos et al., 1997). We implemented the simplest flexural modeling algorithm in tAO, a 1-D model that calculates the total flexure based on the prescribed loads. In our modeling tAO, like all flexural modeling, assumes a thin elastic plate. We do not consider the effects of mass transfer outside to the foreland over geologic time, because as mentioned above, a total landscape lowering in the Eastern Cordillera >500 m would only account for tens of meters of surface uplift. We use average density values and do not consider depth dependent changes in rheology or the thermal structure of the lithosphere. We assume a density contrast of 100 kg/m^3 between a dense lower crust/lithospheric mantle and a 'normal' lithospheric mantle/lower crust with an underlying asthenosphere. The density contrast is the parameter to which topographic deflection is most sensitive, the greater the contrast the greater the topographic change. The value we use reflects the upper end of density contrasts for a dense lower crust in hot arc environments as modeled by Jull and Keleman (2001). This higher value is the one that respects the geometry of the suducting Nazca Plate while at the same achieving the observed rapid surface uplift seen in paleoelevation data. We assume that the mafic lower crust beneath the central Andes was >10 km thick because the modern geophysical properties of the central Andean crust (Oncken et al., 2003) suggest that the entire 60–70 km crustal thickness is felsic in composition and that no mafic lower crust remains under the plateau.

We show the results of two models. The first case shows a total removal of a dense lithospheric root 80 km thick and 390 km wide beneath the entire plateau (Fig. 6). The second scenario shows the

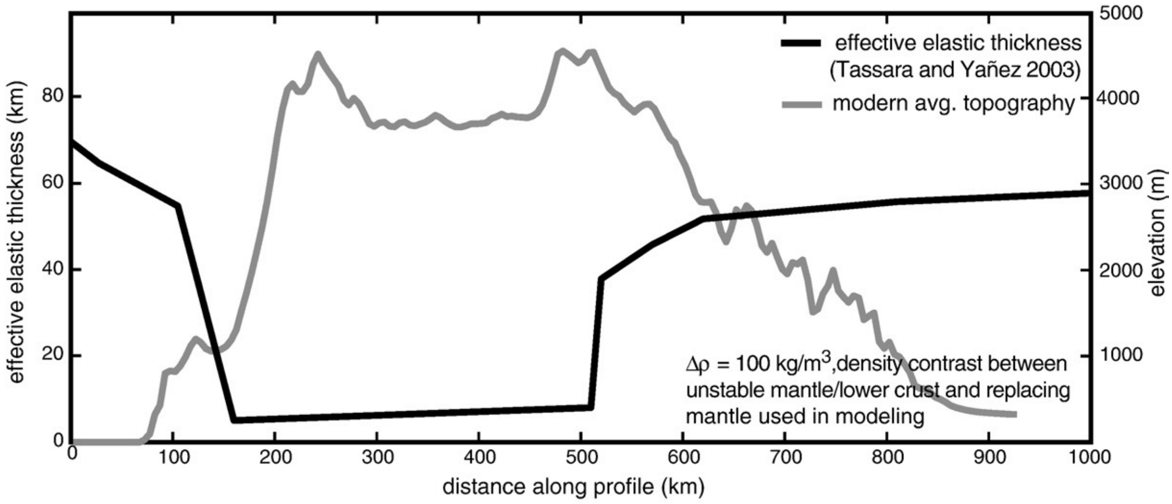

Fig. 5. Profile of the effective elastic thickness of the Andean lithosphere (black line) along the dashed black line in Fig. 1 based on data presented in Tassara et al. (2005). The average topography of the central Andes (gray line) from the same dashed black line in Fig. 2 and the topographic profiles in Fig. 4. The other parameters used in the flexural modeling are given in the lower right hand corner of the graph.

removal of a large volume of thickened and dense lithospheric mantle and lower crust of 140 km thick and 180 km wide beneath the Eastern Cordillera and eastern edge of the Altiplano, along with a 45 km thick and 110 km wide block beneath the western edge of the Altiplano according to the summary of geophysical data presented in Beck and Zandt (2002). The cross sectional area of lithosphere removed in each of the modeled scenarios is approximately equal. To evaluate the ability of the removal of lower lithosphere to produce the observed modern topography, we summed the modeled topographic deflections with the 10 Ma paleotopography of the range (Fig. 6). The

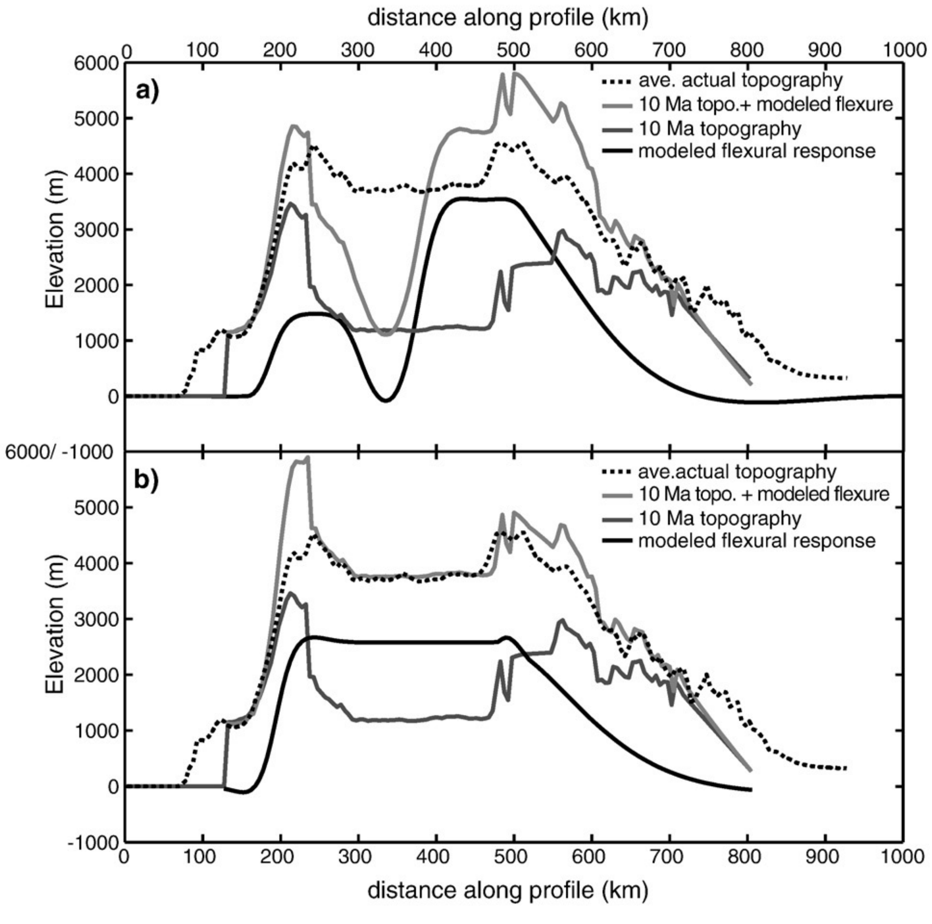

Fig. 6. Modeled topographic profiles (light gray line) of the central Andes based on summing the topographic deflection (black line) related to the loss of the lower lithosphere with the topography at 10 Ma (dark grey line). The average actual topography (dashed gray line) is shown for comparison. Two lithospheric loss scenarios are presented here, a.) loss concentrated beneath the Eastern Cordillera as suggested by broadband seismic and b.) loss of a uniform lower lithosphere 390 km×80 km thick beneath the entire Altiplano. All profiles are along the dashed black line in Fig. 2.

average actual topography of the range is also plotted in Fig. 6 a and b for comparison purposes. Both models suggest that lithospheric loss could account for the majority, if not all of the topographic change generated over the 4 Myr period between 10 and 6 Ma. Despite the simplifications inherent to themodeling, the results provide valuable insight intowhich processes are responsible for topographic change in the central Andes.

7. Discussion and Conclusions

Topography is a key parameter that affects the earth system in many different ways, from plate boundary interactions (Iaffaldano et al., 2006) to atmospheric circulation (Prell and Kutzbach, 1992; Lenters and Cook, 1995). Our reconstruction of central Andean paleo-topography between 18 and 22°S using range-wide paleosurfaces to create a datum and relief/paleoelevation estimates to restore that datum to its initial state at 10 Ma provides a tool for assessing the role of different geodynamic mechanisms for generating surface uplift in the Andes. The 10 Ma topographic profile across the range shown in Fig. 4 is very similar to topography generated by the thermomechanical numerical modeling studies of Sobolev and Babyeko (2005, Fig. 5, 25 Ma timeslice). Our study represents the most complete combination of quantitative surface uplift estimates for the central Andes using both regional geomorphology and absolute paleoelevations at specific sites in the Altiplano and Eastern Cordillera. Strong evidence exists for contemporaneous, large magnitude surface uplift to the north (Schildgen et al., 2007; Thouret et al., 2007) and south (Kay et al., 1994). Shortening rates derived from area balancing and the timing constraints on elevation change in the central Andes suggest that crustal shortening alone cannot generate the observed change in elevation. Modeling the regional isostatic effect of lower lithosphere loss allows for a first-order approximation of the impact of this process on topography across the entire range. We find either of the scenarios shown in Fig. 6 as plausible explanations based on different types of geophysical data. Of the two modeled scenarios, the best match of modeled to actual topography results from the total removal case represented in Fig. 6b.

Densification of the mafic lower crust by transformation to eclogite during crustal thickening is thought to result in the creation of a gravitational instability in the lower lithosphere which may result in its foundering. 'Delamination' of the lower crust and mantle lithosphere was first proposed for the southern extension of the central Andean plateau, the Puna, by Kay et al. (1994) on the basis of geochemical evidence from Plio-Quaternary mafic rocks with varying degrees of mantle affinity and crustal thickness. To the north in the Altiplano, there is far less mafic volcanism, but there is voluminous 10–6 Ma ignimbrite volcanism, in particular the Los Frailes and Morococala ignimbrite shields (Baker, 1981) in the western part of the Eastern Cordillera. Despite the lack of mafic volcanism related with the proposed delamination event in the south, L. Hoke et al. (1994) explain a pronounced anomaly in ^{3}He isotopes in surface thermal waters that occurs across the entire southern Altiplano as compelling evidence for mantle asthenosphere in contact with the base of the crust. These He isotope data lend support to the wholesale removal scenario of Fig. 6b. In addition, geophysical evidence for a largely felsic central Andean crust (Beck and Zandt, 2002) reveals that there is little or no eclogitic crust at these latitudes.

At first glance, the convective removal of lower lithosphere scenario represented by Fig. 6a, seems like a poor fit to the modern topography of the Eastern and Western cordilleras. However, if we take into account that excess crustal mass in one area can be redistributed rapidly via crustal flow in the hot and weak Andean crust beneath the Altiplano (Oncken et al., 2003; Tassara et al., 2005) as suggested by Husson and Sempere (2003), then this solution is a more viable. Furthermore, this solution fits best with what is know about the current geophysical state of the modern Andean lithosphere between 18 and 22°S. Seismic attenuation studies along the length of the central Andes identified the presence of a mantle lid beneath parts of the Altiplano, but not the Puna (Whitman et al., 1996). Later studies across the Plateau at 20°S confirmed and defined the outline of a high seismic velocity zone located below the center of the Altiplano (Myers et al., 1998). This piece of lithospheric mantle could be the remnants of a drip event, or possibly the initial re-growth of what was removed in the wholesale removal scenario. To the north, thermochronology results from the Ocoña–Cotahuasi canyon in SW Peru demonstrate near-synchronous surface uplift of the western flank of plateau as that observed in Northern Chile (Schildgen et al., 2007; Thouret et al., 2007). Based on the lack of contemporaneous regional volcanic events in the area, these authors suggest lower crustal flow as the dominant mechanism driving surface uplift (Schildgen et al., 2007).

Our modeling results demonstrate that lower lithosphere removal must be considered as an important, if not the principal, mechanism in the late Miocene surface uplift of the central Andean plateau and its flanks. While our results do not support crustal shortening or lower crustal flow as the sole means for generating the observed rapid late Miocene surface uplift of the central Andean plateau, we suggest that all of the pre-10 Ma crustal shortening within the plateau resulted in gradual surface uplift of up to several kilometers and provided the crustal thickness necessary to precondition the lower lithosphere for later removal. Ultimately lithospheric thickening generated a thickened welt of dense eclogite and mantle lithosphere that became unstable and detached by delamination or convective removal. Thickened, low-density upper crust set the height limit to which the plateau could rise during lower lithosphere removal.

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