Simulating foreland basin response to mountain belt kinematics and climate change in the Eastern Cordillera and Subandes: An analysis of the Chaco foreland basin in southern Bolivia

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ABSTRACT

The relative importance of crustal thickening, lithospheric delamination, and climate change in driving surface uplift and the associated changes in accommodation space and depositional facies in the adjacent foreland basin in the central Andes has been a topic of vigorous debate over the past decade. Interpretation of structural, geochemical, geomorphic, and geobiologic field data has led to two proposed end-member Tertiary surface uplift scenarios for the Eastern Cordillera and Subandes in the vicinity of the Bolivian orocline. A “gradual uplift” model proposes that the rate of surface uplift has been relatively steady since deformation propagated into the Eastern Cordillera during the late Eocene. In this scenario, the mean elevation of the region was >2 km above mean sea level (msl) by the late Miocene or earlier. Alternatively, a “rapid uplift” model suggests that the mean elevation of the Altiplano was <1 km above msl, and the peaks of the Eastern Cordillera were more than 2 km below their modern elevations until rapid uplift began ca. 10 Ma. Determining which of these uplift scenarios is most consistent with the stratigraphic record is complicated by the potentially confounding effects of global climate changes and lithospheric delamination in the stratigraphic record. In this study, we use a coupled mountain-belt–sediment-transport model to predict the foreland basin stratigraphic response to these end-member surface uplift scenarios. Our model results indicate that the location and height of the migrating deformation front play the dominant roles in controlling changes in accommodation space and grain size within the foreland basin. Changes in accommodation space and rates of sediment supply related to climate change and lithospheric delamination play secondary roles. Our results support the conclusion that the Eastern Cordillera likely gained most of its modern elevation prior to 10 Ma, in contrast with recent proposals that most of the modern elevation was obtained during the late Miocene. This conclusion is consistent with the most comprehensive paleoaltimetric analysis of the region to date.

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INTRODUCTION

Sediments eroded from mountain belts are primarily deposited in foreland basins, which are elongate troughs created by flexural loading of the lithosphere adjacent to mountain belts (DeCelles and Giles, 1996). Foreland basin stratigraphy can provide useful data for reconstructing climate and crustal deformation histories within ancient mountain belts (e.g., Heller et al., 1988; DeCelles et al., 1998; Marzo and Steel, 2000; Uba et al., 2007). The fluvial systems that transport sediments from eroding mountain belts are sensitive to changes in sediment supply, discharge, and sediment accommodation rates, and thus, the foreland basin stratigraphy in these systems is a partial record of changes in climate and the geometry of the mountain belt load through time. The flexural profile of foreland basin accommodation space depends on the rigidity of the lithosphere that is underthrust beneath the mountain belt as well as the width and elevation of the mountain belt load. If the foreland basin geometry and rigidity of the underthrust lithosphere can be constrained, then it is possible to infer changes in the mountain belt load that occurred during the development of the foreland basin. As such, numerous studies have inferred tectonic histories for ancient mountain belts by fitting foreland basin geometries predicted by a coupled mountain belt and foreland basin numerical model to observed foreland basin isopach data (e.g., Toth et al., 1996; Ford et al., 1999; Garcia-Castellanos et al., 2002; Prezzi et al., 2009).

The eastern margin of the central Andes (i.e., the portion of the mountain belt that is to the east of the Altiplano Plateau) in southern Bolivia is an ideal region for constraining late Cenozoic changes in mountain belt geometry and climate through numerical modeling of foreland basin stratigraphy. The sediments preserved in the foreland basin have been well documented, and field and laboratory constraints have been reported in the literature for shortening and exhumation rates in the hinterland for the period of time over which the eastern margin of the central Andes has developed. A W-E transect of the Cenozoic foreland basin stratigraphy is exposed within the retroarc fold-and-thrust belt. Detailed isopach maps and stratigraphic sections for the Cenozoic foreland basin stratigraphy have been developed based on a combination of measured sections from the fold-and-thrust belt and correlations between well-log data and two-dimensional (2-D) seismic data (Sempere et al., 1997; DeCelles and Horton, 2003; Echavarria et al., 2003; Uba et al., 2005, 2006). In addition to data from the foreland basin, the timing and magnitude of shortening and exhumation within the mountain belt have been constrained by field mapping and thermochronology (McQuarrie, 2002; Muller et al., 2002; Oncken et al., 2006; Ege et al., 2007; Barnes et al., 2008, 2012). While most of the papers in this volume focus on the NW Argentina transect through the central Andes, we focus here on the Bolivian transect due to these unusually rich data sets. We further assume that the two transects (Bolivia and NW Argentina) behave in a broadly similar way. This assumption is supported by basin evolution and thermochronologic studies (e.g., DeCelles et al., 2011; Carrapa et al., 2011a, 2011b) as described further in the Discussion section. At a larger scale, the Andes as a whole have been an invaluable natural laboratory for exploring feedbacks between climate and rock uplift (Montgomery et al., 2001; Strecker et al., 2007).

An unresolved issue for the central Andes is: When did the topography of the central Andes rise to its modern elevation? One model for the topographic development of the central Andes near the latitude of the Bolivian orocline posits that the mean topography reached near-modern elevation following Neogene crustal thickening within the Subandean zone; thus, earlier crustal thickening within the Eastern Cordillera was not sufficient for the eastern margin of the central Andes to rise to near-modern elevations (Isacks, 1988; Gubbels et al., 1993). In addition to crustal thickening, continental lithospheric foundering has been proposed as a mechanism for the generating at least some of the rapid Neogene surface uplift posited by this model (Kay and Kay, 1993; Garzione et al., 2008; DeCelles et al., 2009). Continental lithospheric foundering involves removal of negatively buoyant lower crust (i.e., eclogite root) and mantle lithosphere by either delamination or Rayleigh-Taylor instability. Initial paleoelevation and geomorphologic studies did support late Miocene rapid surface uplift of the region (Ghosh et al., 2006; Hoke et al., 2007; Garzione et al., 2008). Pedogenic carbonate and carbonate cement samples collected between 17°S and 18°S within the Altiplano and Eastern Cordillera Neogene stratigraphy contain decreasing δ18O values in progressively younger units (Garzione et al., 2008). Garzione et al. (2008) interpreted this oxygen-isotope trend as evidence for an increase in elevation of the Eastern Cordillera of 2.5 ± 1 km from 10 to 7 Ma. Although these results are in agreement with fossil leaf and clumped isotope data collected from the Altiplano and Eastern Cordillera (e.g., Gregory-Wodzicki et al., 1998; Ghosh et al., 2006), the elevation gain predicted by all three methods can be complicated (i.e., biased toward larger values) by climate change due to the uplift of the Andes (Ehlers and Poulsen, 2009). Also, recent studies have interpreted the change in Miocene oxygen isotopes to be the result of regional climate change caused by uplift of an already significantly elevated (>2 km) central Andes above an orographic threshold (Poulsen et al., 2010; Insel et al., 2012). Most recently, Quade et al. (this volume) showed that, while rapid uplift may have occurred locally in the Eastern Cordillera as documented by Garzione et al. (2008), the easternmost Puna and Eastern Cordillera rose gradually to >3 km by no later than 15 Ma. Evidence for late Miocene rock uplift is recorded by paleosurfaces that exist on both the eastern and western margins of the central Andes. Barke and Lamb (2006) estimated 1.7 ± 0.7 km of localized rock uplift for the San Juan Del Oro surface of the Eastern Cordillera and Interandean tectono-morphic regions since 12–9 Ma, when the surface was abandoned and incised. The Barke and Lamb (2006) results do not constrain the magnitude of surface uplift, however. The difference between rock and surface uplift is particularly significant because localized rock uplift can be much higher than mean regional surface uplift due to isostatic effects (England and Molnar, 1990). We will refer to this conceptual model for surface uplift as the rapid...
uplift end-member model. Although Quade et al. (this volume) appear to have disproven this model, it is still important to test this model using other lines of evidence (besides paleoaltimetry).

A second end-member model for the topographic evolution of the central Andes invokes gradual surface uplift since the late Eocene, when deformation propagated from the Western Cordillera into the Eastern Cordillera. Evidence for pre-Neogene deformation comes from Eocene exhumation ages within the Eastern Cordillera and changes in paleocurrent directions within Paleogene stratigraphy of the Altiplano and Eastern Cordillera (Horton et al., 2002; McQuarrie et al., 2005; Ege et al., 2007; Barnes et al., 2008). Although pre-Neogene shortening is considered low (<100–150 km) for building a high-elevation plateau, long-term shortening should have led to crustal thickening and isostatic rebound in the Eastern Cordillera. Therefore, unless erosion rates exceeded or were equal to rock uplift rates, the Eastern Cordillera should have been rising (i.e., surface uplift as defined by England and Molnar, 1990) since the late Eocene, barring some mechanism for keeping the elevation of the Andes low during an extended period of deformation. The gradual uplift end-member model posits that the Andes gained the majority of its modern elevation prior to ca. 10 Ma.

The Cenozoic stratigraphy of the Chaco foreland basin in Bolivia shows an increase in both grain size and sedimentation rates during the late Miocene (Uba et al., 2006, 2007). This stratigraphic change could have been the result of rapid surface uplift. However, previous studies have inferred that this depositional trend might instead be primarily controlled by distance from the approaching fold-and-thrust belt (DeCelles and Horton, 2003; Uba et al., 2006), which provides the topographic load to drive flexural subsidence and accommodation space creation. Thus, the primary mechanism that caused these changes in late Miocene stratigraphy is still uncertain. In addition to thrust belt migration, climate change and continental lithosphere foundering have also been emphasized as potential controls on the foreland basin stratigraphy of the central Andes in recent years. For example, Kleinert and Strecker (2001) documented a change from previously dry to wetter conditions in the Santa Maria Basin of northern Argentina between 9 and 7 Ma. Based on climate model studies, Poulsen et al. (2010) and Insel et al. (2012) have shown that surface uplift of a significantly elevated central Andes in the late Miocene would have caused an increase in precipitation along the eastern front of the central Andes. Uba et al. (2007) found a significant (i.e., factor of 5) increase in time-averaged depositional rates within the stratigraphy of the Subandean thrust belt in Bolivia at ca. 8–7 Ma based on U-Pb dating of tuffs within Miocene volcanic rocks. Coeval with changes in depositional rates and climate conditions within the foreland depositional facies of the Cenozoic fluvial units, a shift occurs from single-thread sinuous channels to more amalgamated alluvial megafan facies (Uba et al., 2006). Uba et al. (2007) interpreted these changes in depositional rates and facies to be a consequence of a change from semiarid to more humid climate conditions during the intensification of the South American monsoon. An increase in mean annual precipitation should increase sediment supply through enhanced erosion rates and increase the transport capacity of foreland basin fluvial systems via greater mean annual discharge.

Lower-crustal delamination is another mechanism that has been proposed for controlling the late Cenozoic foreland basin stratigraphy in the central Andes. The cordilleran cycle, a conceptual model for mountain belt development and cyclicity proposed by DeCelles et al. (2009), posits that as lower crust and mantle lithosphere are underthrust beneath a growing mountain belt, magmatic and petrologic processes lead to the formation of a dense eclogite root that acts as a subsurface negatively buoyant load, lowering the elevation of the mountain belt relative to a state of isostatic equilibrium (DeCelles et al., 2009; Pelletier et al., 2010). If sufficient shortening takes place, the eclogite root reaches a critical thickness or volume and is delaminated or removed via a Rayleigh-Taylor instability. This cordilleran cycle, as posited by DeCelles et al. (2009), includes episodic periods of modest (i.e., 0.5–1 km) increases in surface uplift and shortening rates, both of which could have had a significant effect on the adjacent foreland basin through the modification of both rates of sediment supply and creation of accommodation space. Presently, the Puna Plateau is thought to be in a postdelamination state in this conceptual model (Schurr et al., 2006; DeCelles et al., 2009) and, as a consequence, has a mean elevation that is significantly higher than the Altiplano Plateau. The Altiplano, in turn, is thought to have had a delamination event at 10 Ma (Kay et al., 1994). DeCelles et al. (2009) proposed that the Altiplano may already be in an early stage of a new cycle, and its lower elevation is a consequence of newly forming eclogite loads. Evidence for delamination beneath the Altiplano comes from seismic velocity analysis of the eastern Altiplano lithosphere at 20°S (Beck and Zandt, 2002). A low-velocity zone occurs within the upper mantle beneath the thickened crust of the eastern Altiplano and western edge of the Eastern Cordillera and is interpreted as a location where the cold-fast upper mantle has been removed by a delamination event. Might stratigraphic trends in the late Miocene Chaco foreland basin deposits in Bolivia be a signature of lithospheric foundering instead of climate change or fold-and-thrust belt propagation?

Previous studies have numerically modeled the evolution of the central Andes as a coupled mountain-belt–foreland-basin system. Flemings and Jordan (1989) first simulated the rapid uplift end-member model for the last 5 m.y. of deformation in the central Andes with a two-dimensional model. They concluded that the foreland basin adjacent to the central Andes should have shifted from narrow and underfilled to broad and overfilled as sediment supply outpaced sediment accommodation. However, the results of their model are potentially limited by the fact the mountain belt component of their model was simplified by assuming a constant topographic slope and sediment supply through time. Prezzi et al. (2009) recently applied a more rigorous deformation and erosion model to test the effect of mountain load geometry and elastic thickness of the South American lithosphere on the sediment accommodation rates within the foreland basin of...
the central Andes since the middle Miocene. They found that by decreasing the elastic thicknesses beneath the Eastern Cordillera and Interandean zones between 14 and 6 Ma and by deforming a detailed structural cross section developed by McQuarrie (2002), their model results adequately fit modern gravity anomalies and isopach distributions for the late Cenozoic stratigraphy recently described by Uba et al. (2006) in the Subandean zone of Bolivia.

In this study, we explore the linkages among thrust-belt kinematics, climate change, continental lithosphere delamination, and the foreland basin stratigraphy in the central Andes using a coupled, two-dimensional numerical model. Flemings and Jordan (1989) simulated the foreland response to a fixed topographic slope and constant sediment supply and, thus, did not capture the effects of feedbacks among mountain belt erosion, sediment supply to the foreland basin, and sediment accommodation within the foreland basin. The Prezzi et al. (2009) study focused on the changes in foreland basin accommodation based on surface uplift due to a very specific history of crustal deformation and changes in elastic thickness under constant climate conditions. Both studies focused on crustal thickening as the dominant mechanism for driving foreland basin development. In contrast, in this paper, we aim to constrain the relationship of the Cenozoic foreland basin stratigraphy to end-member surface uplift models, climate change, and lower-crustal delamination in order to place firmer constraints on the paleoelevation history of the central Andes. First, we determine which end-member surface uplift model is most consistent with the available stratigraphic and tectonic data for the development of the eastern margin of the central Andes. Second, we determine whether there is some signal (e.g., unconformity, grain size change) of climate change or continental lithospheric delamination recorded in the upper Miocene Chaco foreland basin stratigraphy.

**GEOLOGIC BACKGROUND**

The central Andes of southern Bolivia contain a high-elevation, internally draining, low-relief plateau. This portion of the Andes also has the highest magnitude of total shortening (i.e., ~285 km of minimum crustal shortening within the Eastern Cordillera, Interandean, and Subandean zones; Isacks, 1988; McQuarrie, 2002; Oncken et al., 2006). The orogenic belt is generally broken up into tectonomorphic regions, which include (from west to east): the Western Cordillera, Altiplano, Eastern Cordillera, Interandean, Subandean, and Chaco foreland basin zones (Fig. 1). The modern topographic divide between westward internal drainage into the Altiplano basin and eastward drainage into the Chaco foreland basin resides within the Eastern Cordillera, a bivergent fold-and-thrust belt. Thrusts within the Eastern Cordillera detach in Ordovician-aged horizons and predominantly exhume Paleozoic through Mesozoic units (McQuarrie, 2002). Further east, the Interandean and Subandean zones are where deformation is currently active. These zones contain dominantly eastward-verging imbricate thrusts that generally exhume younger stratigraphic units compared to the Eastern Cordillera. The modern deformation front is located between 0.5 and 1 km above mean sea level (msl). Beyond the Subandean zone, the Chaco foreland basin extends an additional 250–600 km east into Bolivia, Paraguay, and the Pantanal wetlands of Brazil until it onlaps Precambrian

![Figure 1. Digital elevation model for the central Andes displaying the tectonomorphic zones after Barnes et al. (2008). Topography is from the Shuttle Radar Topography Mission (SRTM) 90 m data set. The tectonomorphic regions are the following: WC—Western Cordillera; AL—Altiplano; EC—Eastern Cordillera; IA—Interandean zone; SA—Subandean zone. The thick white lines are thrust faults located on the boundary between major divisions, and the thin white lines mark political boundaries.](https://pubs.geoscienceworld.org/books/chapter-pdf/963891/mwr212-17.pdf)
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basement highs in eastern Bolivia and southwestern Brazil (Horton and DeCelles, 1997). An exception occurs between 19°S and 20°S, where the foredeep basin deposits pinch out onto the Alto de Izozog basement high over a distance that is less than 200 km from the deformation front (Uba et al., 2006).

The Chaco foreland basin stratigraphy has been documented within the exposed thrust sheets of the Eastern Cordillera, Interandean, and Subandean zones by DeCelles and Horton (2003), Echavarria et al. (2003), Horton (2005), and Uba et al. (2005). Isopach trends have also been developed for the buried modern foreland basin units based on correlations between well data and seismic cross sections (Uba et al., 2006). Starting at the base of the Cenozoic stratigraphy in the Subandean zone (Fig. 2), the Petaca and Yecua Formations contain fluvial deposits with well-developed paleosols, paleocurrent trends that dominantly show transport toward or along strike with the approaching mountain belt, and thicknesses that are low, considering the amount of time contained within the units. The Yecua Formation is unique to the Subandean zone Cenozoic units because it also contains shallow-marine facies in addition to fluvial and lacustrine facies in outcrops located north of 20°S. Moving up through the stratigraphy into the Tariquía, Guandacay, and Emborozú Formations, overall grain size increases from that of the Yecua Formation, paleocurrent data indicate a shift to transport from the west, and depositional rates increase. Based on dating volcanic ash layers in the Cenozoic foreland basin stratigraphy, there is a factor of five increase in depositional rates between the Yecua and Tariquía Formations (Uba et al., 2007). Higher in the stratigraphic section, depositional rates decrease by half upward into the Guandacay Formation. Although deposition is generally conformable within the Cenozoic stratigraphy, unconformities bound the Petaca Formation, and an angular unconformity separates the Guandacay

![Figure 2. Generalized stratigraphic columns for the Cenozoic stratigraphy of the Chaco foreland basin after DeCelles and Horton (2003) and Uba et al. (2006, 2007). Formation thicknesses are displayed beneath the formation names. The question marks represent uncertainty in the boundaries between formations due to a lack of absolute age data. A detailed discussion of the age uncertainties for the Cenozoic stratigraphy of the Eastern Cordillera, Interandean, and Subandean zones can be found in DeCelles and Horton (2003) and Uba et al. (2006). m—mudstone; s—sandstone; c—conglomerate.](https://pubs.geoscienceworld.org/books/chapter-pdf/963891/mwr212-17.pdf)
and Emborozú Formations. The upper unconformity of the Peta- 
caca Formation has been interpreted to record the uplift and pas-
sage of the forebulge as the Subandean zone passed from the 
back bulge into the foredeep basin (Uba et al., 2006). In contrast, 
the angular unconformity located between the Guandacay and 
Embomozú Formations has been interpreted as the initiation of 
 wedge-top deposition in the Subandean zone. Low-angle uncon-
formities have been documented at the base of coarsening-upward cycles within the age-equivalent stratigraphy of the Tariquía and 
Guandacay Formations in northern Argentina (Echavarria et al., 
2003). However, cycles and significant unconformities were not 
documented in the Tariquía and Guandacay Formations of the 
southern Bolivia Subandean zone (Uba et al., 2006). Farther 
west and older in age, similar stratigraphic trends occur within 
the remnant foreland basin stratigraphy of the Eastern Cordillera 
and Interandean zones (Fig. 2). The transition from the Cayara 
into the Camargo Formation involves an overall increase in grain 
size and changes in paleocurrent direction and apparent deposi-
tional rates. One stratigraphic difference between this remnant 
foreland basin and the Neogene foreland basin of the Subandean 
zone is that an unconformity does not bound the lowest Cenozoic 
unit in the section. Overall, the most significant trends within the 
Cenozoic foreland basin stratigraphy of the eastern margin of the 
central Andes are the apparent increase in depositional rates and 
grain size with decreasing age, and thus, any numerical models 
for the foreland basin evolution of the eastern margin of the cen-
tral Andes must honor these trends.

MODEL DESCRIPTION

Summary

The numerical model of this paper is a 2-D kinematic 
model that couples an actively deforming and eroding moun-
tain belt with sediment transport and flexure within a depo-
sitional foreland basin. A 2-D model is valid for this region 
of the Andes because the mountain front is nearly linear, and 
paleoflow directions for the majority of the Cenozoic foreland 
basin stratigraphy are perpendicular to the mountain front. The 
topography of the numerical model represents the mean topog-
raphy of the eastern margin of the central Andes (i.e., western 
edge of the model begins in the Altiplano) located between 
18°S and 21°S. In our kinematic model, shortening and rock 
uplift rates are prescribed. Although time-averaged shortening 
rates (Table 1) can be directly calculated from published val-
ues and used as preliminary input into the model, obtaining 
valid rock uplift rates is an iterative process of running simu-
lations, checking the results against independent published 
data (e.g., modern topography and exhumation magnitudes), 
and adjusting the rock uplift rates within each tectonomorphic 
zone (Fig. 3). In this paper, we report model outcomes that are 
consistent with published data for the central Andes, including 
the modern topography, shortening rates from balance cross 
sections, exhumation magnitudes, Cenozoic isopach data, and 
modern basin geometries.

All model simulations begin at 43 Ma. Evidence from field 
mapping and thermochronology suggests that deformation did

Figure 3. The distribution of prescribed zonal rock uplift rates within the eastern margin of the central Andes hinterland 
through time for both the gradual (A) and rapid (B) end-member uplift models. Symbols for each zone are the following: 
triangles—Altiplano (AP); dashed line—Eastern Cordillera back thrust (BT); dashed dotted-line—Eastern Cordillera fore-
thrust (FT); dotted line—Interandean (IA); asterisk—western Subandean (WSA); solid line—eastern Subandean (ESA).
not propagate into the Eastern Cordillera until the late Eocene, and, therefore, starting the model before then is unnecessary and would have been poorly constrained with available data. The older foreland basin section in the region was interpreted by DeCelles and Horton (2003) to represent distal foredeep to forebulge deposits and only represents a partial foreland basin profile. Having a more complete foreland basin profile that includes the foredeep is necessary to constrain the topography of the adjacent mountain belt.

**Rock Deformation**

Dynamic models for crustal deformation (e.g., Simpson, 2004) that directly solve for visco-plastic deformation within a fold-and-thrust belt could be applied to the central Andes. This approach, however, comes with the drawbacks of longer computational times and the difficulty of calibrating the model to multiple types of calibration data. We therefore take a simpler approach in this paper. In the model, the mountain belt is partitioned into individual blocks for each tectonomorphic region (i.e., Altiplano, Eastern Cordillera, etc.). Exceptions are made for the Eastern Cordillera and Subandean zones. The Eastern Cordillera, a bivergent fold-and-thrust belt, was divided into separate back-thrust and fore-thrust blocks because these two regions were rapidly exhumed at different periods of time (Barnes et al., 2008). The Subandean zone was divided into a western and eastern block to allow the central and eastern Subandean zone to continue to subside until deformation propagated into the region during the late Miocene. Deformation is simulated by uniform rock uplift and shortening applied to each actively deforming block. Time-averaged shortening rates were determined for input to the model by dividing the total shortening determined from field-based balanced cross sections for each tectonomorphic zone by the duration of fault activity (McQuarrie et al., 2005; Muller et al., 2002). The positions of the model nodes located to the east of the deformation front are fixed, while the model nodes to the west of the deformation front are allowed to translate toward the foreland basin as if riding over a décollement. The total propagation of the deformation front (and hence the forebulge) is ~536 km, which is close to the ~600 km of total propagation of the forebulge since the late Eocene estimated by DeCelles and Horton (2003).

**Bedrock Channel Erosion**

A stream-power model is used to determine bedrock incision rates within the mountain belt (Howard and Kerby, 1983; Whipple and Tucker, 1999): 

\[
\frac{\partial h}{\partial t} = -k_e A^{n} \left[ \frac{\partial h}{\partial x} \right] = -k_e A^{1/2} S = -k_e IS ,
\]

(1)

where \( h \) is elevation above msl (m), \( t \) is time (yr), \( k_e \) is the bedrock erodibility (yr\(^{-1}\) for \( n = 1.0 \) and \( m = 0.5 \)), \( A \) is the contributing drainage basin area (m\(^2\)), \( x \) is the lateral distance from the model origin (m), \( l \) is the distance along the principal channel from the headwaters of the drainage basin (m), \( S \) is the local channel slope (unitless), and \( m \) and \( n \) are constants that determine the dependence of local erosion on discharge and channel slope. In this standard 2-D implementation of the stream-power model, \( h \) represents the longitudinal profile of the main channel of a three-dimensional (3-D) river basin. The usual approach in such cases (e.g., Whipple and Tucker, 1999) is to assume that contributing area goes as the square of distance from the drainage divide. This assumption implicitly adds tributary area to the main channel in a manner that is statistically consistent with the way in which tributaries add area in real 3-D river basins. In Equation 1, we assume that the length of the principal channel is proportional to the square root of the contributing drainage area (Hack, 1957). We also assume that the area and slope exponents \( m \) and \( n \) have values of 0.5 and 1, respectively. Evidence from theory and field studies predicts that the ratio of \( m/n \) is near 0.5 (Whipple and Tucker, 1999). Although the slope exponent \( n \) can range between 0.66 and 2.0 depending on the relationship between slope and stream power (shear stress), we chose to make the stream power linearly proportional to the slope (\( n = 1.0 \) and \( m = 0.5 \)), consistent with the assumption of many other studies (e.g., Kirby and Whipple, 2001; Snyder et al., 2000). Bedrock incision rates for active mountain belts are on the order of 0.1–1.0 mm/yr (Montgomery and Brandon, 2002). When \( m/n \) is equal to 1/2, we found that \( k_e \) must be on the order of \( 10^{-6} \) (m/yr) to appropriately reproduce the range of calculated exhumation rates (i.e., 0.1–0.6 mm/yr) for time scales of \( 10^5 \) to \( 10^7 \) yr for the central Andes (Safran et al., 2005; Ege et al., 2007; Barnes et al., 2008). The boundary between the bedrock portion of the model (where stream-power-driven erosion of bedrock is modeled) and the alluvial portion of the model (where grain-size–dependent sediment transport is modeled) is allowed to fluctuate in the model. If a given model node contains sediment or sediment is deposited (i.e., upstream sediment supply exceeds the transport capacity), then the node is treated as an alluvial channel, and no bedrock incision occurs. Otherwise, bedrock incision occurs, and the amount of newly created sediment transported downstream is controlled by the transport capacity.

**Sediment Transport**

Sediment transport of a bimodal grain-size distribution (i.e., gravel and sand) is simulated with a modified version of the diffusion model approach, which states that sediment flux is proportional to local channel slope. Linear slope-dependent sediment flux, when combined with conservation of mass, gives a diffusion equation for the evolution of the longitudinal profile of the foreland basin (Paola et al., 1992). In the model of this paper, we added a threshold slope term to the diffusion equation model because transport of gravel requires...
a threshold slope to initiate transport. The equation for alluvial deposition and erosion along a channel profile in our model is a combination of mass balance (i.e., the Exner equation) and the slope-dependent sediment-transport equation:

\[ q_s = k_g (S - \phi), \quad S > \phi \]
\[ q_s = 0, \quad S \leq \phi \]  

(2)

\[ \frac{\partial h}{\partial t} = -\frac{\partial q_s}{\partial x} - \frac{q_{s,\text{upstream}} - q_{s,\text{downstream}}}{\partial x} \]  

(3)

where \( k_g \) is the transport coefficient (m²/yr), \( \phi \) is the threshold slope, which is a function of grain size, and \( q_s \) is the local sediment flux (m²/yr). The transport coefficients for gravel and sand are determined from a relationship that depends on discharge and river type (Paola et al., 1992). Paola et al. (1992) calculated the values for braided and meandering rivers to range between 1.0 and 7.0 \( \times 10^4 \) m²/yr for a drainage basin with a length of 100 km and a mean annual precipitation of 1 m. We chose a value of 1.0 \( \times 10^4 \) m²/yr for the gravel transport coefficient, which is a reasonable value for braided, gravel-dominated streams with relatively small drainage areas. The transport coefficient for sand was chosen to be \( \sim 6.5 \times 10^4 \) m²/yr, which is on the order of the transport coefficient used by Flemings and Jordan (1989) for their simulation of the central Andes. The threshold slope for gravel entrainment is \( \sim 0.001 \), which represents effective channel parameters of a median grain size of 0.02 m and bankfull flow depth of 1.85 m. The chosen median grain size is on order with the median grain size of bed-load material observed within the modern Rio Pilcomayo located in the eastern Subandean zone (Mugnier et al., 2006).

Channel armoring is an emergent property of the model obtained by incorporating a threshold slope for gravel transport. If discharge is insufficient to entrain gravel present on the channel bed during a time step, then the finer-grained sand deposited beneath the gravel is prevented from being transported during that time step. This process should reduce both alluvial and bedrock erosion in regions of lower regional slope where gravel is present. Active bed material layers in natural channels are on the order of 1–2 grains thick (Hassan et al., 2006). Although our active layer is on the order of a meter, it is valid to have a larger active layer for the purposes of this study because we are focusing on long-term (>10³ yr) grain-size trends that average over many individual flooding events.

**Flexural Isostasy**

The bedrock and alluvial surface dynamics models are coupled to a flexural foreland basin in order to quantitatively assess accommodation space creation and the migration of the forebulge through time in the model. The flexural model solves for the displacement of a thin elastic beam subjected to a spatially distributed vertical load (Turcotte and Schubert, 2002):

\[ D \frac{d^4 w(x)}{dx^4} + (\rho_m - \rho_s) g w(x) = L(x), \]

(4)

where \( w \) is the deflection of Earth’s crust (m), \( D \) is the flexural rigidity (Nm), \( \rho_m \) is the density of the mantle (kg/m³), \( \rho_s \) is the density of the mountain crust or foreland sediment (kg/m³), \( g \) is the acceleration due to gravity (m/s²), and \( L(x) \) is the topographic load (kg/m²). For each simulation, except for those involving eclogite root delamination (discussed later), we solve for the deflection due to a topographic load every time step using a Fourier transform method (Watts, 2001). Viscous relaxation effects were not considered in the model because we interpret stratigraphic patterns over geologic time scales that are greater than relaxation time scales.

Several studies have calculated the flexural rigidity of the central Andes using 2-D methods (Horton and DeCelles, 1997; Stewart and Watts, 1997; Tassara, 2005). Their results suggested that flexural rigidities along the eastern margin of the central Andes range between 1.5 \( \times 10^{23} \) and 4.0 \( \times 10^{24} \) Nm. We chose to use a flexural parameter of 150 km because this value best fits the observed modern basin geometry between 18°S and 20°S. This value is consistent with the results of Chase et al. (2009), who found that the flexural parameter should be less than 220 km for the central Andes. The flexural parameter is defined as the following (Turcotte and Schubert, 2002):

\[ \alpha = \left[ \frac{4D}{(\rho_m - \rho_s) g} \right]^{1/4}, \]

(5)

where \( \alpha \) is the flexural parameter (km). Rearranging Equation 5 and applying values from Table 2 yields a flexural rigidity of \( \sim 6.8 \times 10^{23} \) Nm, which is well within the range of calculated effective flexural rigidities for the central Andes. Although previous studies have suggested that the elastic thickness varies in space and time (e.g., Toth et al., 1996; Prezzi et al., 2009), such variations are difficult to constrain. Therefore, our model implements a uniform and constant elastic thickness.

**TABLE 2. PARAMETERS USED IN END-MEMBER MODEL SIMULATIONS**

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>( k_g ) (m²/yr)</td>
<td>( 1.0 \times 10^6 )</td>
</tr>
<tr>
<td>( k_b ) (m²/yr)</td>
<td>( 1.0 \times 10^4 )</td>
</tr>
<tr>
<td>( k_e ) (m²/yr)</td>
<td>( 6.5 \times 10^4 )</td>
</tr>
<tr>
<td>( \phi )</td>
<td>0.001</td>
</tr>
<tr>
<td>( \alpha ) (km)</td>
<td>150</td>
</tr>
<tr>
<td>( \rho_m ) (kg/m³)</td>
<td>2750</td>
</tr>
<tr>
<td>( \rho_s ) (kg/m³)</td>
<td>3300</td>
</tr>
<tr>
<td>( \rho_e ) (kg/m³)</td>
<td>3600</td>
</tr>
<tr>
<td>( g ) (m/s²)</td>
<td>9.8</td>
</tr>
</tbody>
</table>
Numerical Methods

We ran three types of experiments using our numerical model: (1) an end-member surface uplift model experiment, (2) an eclogite root foundering experiment, and (3) a climate change experiment. The end-member surface uplift model experiment simulates the last 43 m.y. when deformation is concentrated in the eastern margin of the central Andes. The purpose of this experiment is to contrast the foreland basin response to rapid surface uplift caused by Neogene crustal thickening. The parameters used in this experiment are found in Table 2. During the simulations of this experiment, mean annual precipitation rates were held constant, and time steps were ~700 yr. Topographic profiles were sampled at 22, 10, and 0 Ma for visual comparison.

Following the end-member surface uplift model experiment, we ran an eclogite root foundering experiment. The purpose of this experiment is to contrast the foreland basin response to rapid surface uplift in the mountain belt caused by crustal thickening alone against rapid surface uplift caused by a combination of foundering and crustal thickening. The model duration, time steps, and interval of topographic profile sampling in the foundering experiment were the same as in the end-member surface uplift experiment. We also applied the parameters in Table 2 and the same kinematic histories in the mountain belt (e.g., shortening rates and propagation of deformation) as in the rapid uplift model. An eclogite root grows in our model between 25 and 10 Ma beneath the Altiplano and Eastern Cordillera back-thrust zones, where significant crustal thickening has taken place such that lower-crustal rocks are subjected to pressures sufficient enough to produce eclogite. We assume that eclogite foundering is caused by a Rayleigh-Taylor instability (defined as the diapiric drip of a dense layer overlying a less dense layer) and that the time scales over which the growth of the eclogite drip occur can be calculated to within first order by the results of the linear-stability analysis of Turcotte and Schubert (2002):

\[ \tau_a = \frac{13.04 \mu}{(\rho_e - \rho_m)b}, \] (6)

where \( \tau_a \) is the amount of time required for an instability to grow by a factor of \( e \), \( \mu \) is the viscosity of the upper and lower layers, and \( b \) is the original thickness of the eclogite layer. The Rayleigh-Taylor instability is not modeled explicitly, but instead we prescribe a rate of eclogite root growth consistent with the observed spacing between foundering events in the central Andes and the thickness of the root required toinitiate foundering.

Equation 6 is rearranged to solve for \( b \) in order to prescribe the thickness of the eclogite root required to initiate foundering in our model. Based on geophysical models and paleoelevation proxies, the length of time over which the foundering event occurred beneath the Altiplano was ~3 m.y. (i.e., between 10 and 7 Ma; Molnar and Garzione, 2007; Garzione et al., 2008). We assume that the majority of the foundering time is spent growing the instability in the lower crust–mantle interface by the initial factor of \( e \) due to the high resistance to flow of the mantle when minimal perturbations in the lower crust–mantle interface exist. Later, when a significant pressure gradient exists at the base of the eclogite layer due to the formation of a drip, the pinching and foundering portion of the drip event occurs much more rapidly. Based on this argument, we inferred that \( \tau_a \) is approximately less than or equal to 3 m.y. Using densities of 3300 and 3600 kg/m\(^3\) for upper mantle and eclogite and a root growth period of 3 m.y., the range in maximum eclogite layer thickness is calculated to be between 1.87 and 46.88 km as the effective viscosity of the underlying mantle layer varies between \( 4.0 \times 10^{19} \) and \( 1.0 \times 10^{21} \) Pa s. This range in eclogite layer thickness is consistent with the 10–20 km eclogite root thickness that occurred prior to removal during a numerical simulation of mantle drip for the Puna Plateau (Quinetros et al., 2008). We used a similar scale (i.e., 12.5 km) for the thickness of the eclogite root required to initiate delamination.

Between 10 and 7 Ma, the root is removed at a linear rate until it is completely removed at 7 Ma. During the period of root foundering, we allowed our flexure algorithm to specify the magnitude of rock uplift due to foundering and superimposed that result on the results of the rapid uplift model. Our model is kinematic, so the viscous coupling between the sinking root and the overlying lithosphere is not simulated in our model. However, there is a possibility that viscous coupling may result in significant (i.e., on the order of hundreds of meters) subsidence and rebound (Göğüş and Pysklywec, 2008).

An experiment with climate change during the late Mio-
cene was also conducted in order to determine its impact on the foreland basin stratigraphy. At 9 Ma, we simulated an increase in mean annual precipitation by doubling both the bedrock erodibility and sediment transport coefficients. Such increases could be triggered by intensification of the South American monsoon system and/or by an increase in orographic activity associated with the mountain belt attaining a threshold elevation. In this experiment, it is also necessary to increase bedrock uplift rates along with the bedrock erodibility and sediment transport coefficients because exhumation must keep pace with erosion in order for the model to reproduce the modern topography of the central Andes at the end of the simulation.

The bedrock erodibility (\( k_b \)) of the stream power model and the transport coefficients (\( k_i \) and \( k_g \)) of the diffusion model have both been shown theoretically to be proportional to discharge (Paola et al., 1992; Whipple and Tucker, 1999). If we assume that mean annual discharge scales proportionally with mean annual precipitation rate, then erosion rates and sediment transport rates scale proportionally with mean annual precipitation rate. This is reasonable assumption for erosion rates because long-term erosion rates have been shown to correlate with mean annual precipitation rates in the Andes. Sediment transport rates in alluvial rivers have also been proposed to correlate with mean annual...
Figure 4. The evolution of the eastern margin of the central Andes as a series of cross sections in time for both the gradual (left column) and rapid (right column) end-member uplift models. Sedimentary deposits are shaded gray, and the location of the deformation front at the time represented by the snapshot is marked by the gray vertical line.
precipitation rates (Molnar et al., 2006). Although the magnitude of increase in mean annual precipitation rate at the onset of the South American monsoon is unknown, long-term erosion rates since the Eocene have not varied by more than a factor of two compared with modern rates (Safran et al., 2005; Ege et al., 2007; Barnes et al., 2008). Thus, both erosion rates and sediment transport rates increase by a factor of two at 10 Ma. The model duration, time steps, and interval of topographic profile sampling are the same as in the end-member uplift models.

RESULTS

Summary of Model Outputs

Topographic-profile and sediment-flux time-series plots show the development of the eastern margin of the central Andes in response to the gradual uplift end-member model (Figs. 4 and 5). Between 43 and 22 Ma, deformation is concentrated within the Eastern Cordillera. During this period of time, topographic slopes are low, and thus the sediment fluxes from the mountain belt into the Altiplano and foreland basin are also low. Sediment eroded off of the Eastern Cordillera is not transported beyond the foredeep, which is underfilled to completely filled. The sediment flux leaving the back-bulge basin toward the east is zero at this time. The Subandean zone in its predeformed state is located more than 400 km from the deformation front near the forebulge crest (Fig. 5A). Low sediment flux from the mountain belt is consistent with observed paleocurrent data from the Petaca Formation, located in the Subandean zone, showing westward paleotransport from the South American craton. Between 30 and 22 Ma, there is a period of higher-than-average sediment flux (Fig. 5B). Sediments with low erodibility contained within wedge-top basins are exhumed as the eastern portion of the Eastern Cordillera begins to uplift at 30 Ma. Exhumation of the wedge-top basin deposits causes a period of high sediment delivery to the foredeep.

By 22 Ma, rock uplift rates between 0.1 and 0.2 mm/yr cause the Eastern Cordillera to rise to a peak elevation of 3 km above msl. The increase in topographic relief causes the mean sediment flux from the mountain belt into the foreland basin to increase, such that sediment from the mountain belt outpaces foreland accommodation rates and is transported out of the foredeep. Subsidence rates in the back-bulge basin are an order of magnitude lower than in the foredeep. Consequently, the back-bulge basin is rapidly filled, and sediment from the mountain belt begins to bypass the filled back-bulge basin by 27 Ma (Fig. 5B). At this time, the deformation front propagates east into the Interandean zone and creates a step in topography located 500 km from the left end of the model domain in Figure 5A. Again, there is a period of higher-than-average sediment flux between 22 and 15 Ma caused by exhumation of wedge-top basins within the Interandean zone. The 22 Ma snapshot represents the time period when the Camargo Formation is transitioning into the overlying Mondragon Formation of the Eastern Cordillera and when the Petaca Formation is being deposited in the Subandean zone. These results are also consistent with available data from NW Argentina showing that the Eastern Cordillera was actively deforming and eroding at ca. 18–22 Ma (Deeken et al., 2006; Coutand et al., 2006). Similarly, thick sedimentary depocenters indicate high sedimentation rates in NW Argentina (Carrapa et al., 2011a).

By 10 Ma, the core of the Eastern Cordillera is within 1 km of its modern peak elevation, and the deformation front and

![Figure 5. (A) Topographic evolution of the central Andes and (B) time series of sediment supply rates into and out of the foreland basin for the gradual uplift model. The thick black lines in A represent snapshots of topography throughout the simulation, and the thick gray lines represent the maximum and mean topography for a north-south sweep of modern topography between 18°S and 21°S. The black rectangles represent the locations of the forebulge crest during each snapshot in time, with the oldest forebulge location on the left. In B, the thick line represents the sediment flux leaving the back-bulge basin or right edge of the model, and the thin lines represent the sediment flux at the deformation front into the foredeep of the foreland basin.](https://pubs.geoscienceworld.org/books/chapter-pdf/963891/mwr212-17.pdf)
forebulge migrate another 100 km toward the South American craton (Fig. 5A). As a result of the increase in relief and mountain front slopes, the average sediment flux into the foredeep increases from 20 to 30 m$^2$/yr and overfills the foredeep. A plot of sediment flux leaving the back-bulge basin shows that at least half of the mean sediment supply to the foreland is leaving the back-bulge basin at this time (Fig. 5B). A significant decrease in the amount of sediment leaving the back-bulge basin and a significant increase in the amount of sediment entering the foredeep occur between 5 and 2 Ma. This is the period in time when the eastern Subandean zone is actively uplifting in both southern Bolivia and northwestern Argentina. Thermochronologic data and growth structures (Carrapa et al., 2011b; DeCelles et al., 2011) indicate that the deformation front was in the Eastern Cordillera until ca. 4 Ma and migrated out into the Subandes after that. This is broadly similar to what observed in Bolivia (Barnes et al., 2008, 2012). High rock uplift rates in the eastern Subandean zone lead to exhumation of wedge-top basins and the development of a topographic step. Although sediment flux from the mountain belt is high during this time, rapid creation of topographic relief in the front of the thrust belt causes high subsidence rates in the foredeep. A significant portion of the sediment supply to the foreland basin is deposited and stored in the foredeep. The final topography at 0 Ma fits the average and maximum topography for a north-south sweep of the modern central Andes well. However, the alluvial slopes of the depositional foreland basin are a few hundred meters higher than the average channel elevations today. It is unclear if this error is due to an overprediction of the sediment supply, an underprediction of transport coefficient, or if sediment fluxes at the downstream end of the model domain are too low.

Although rates of shortening and the propagation of the deformation front are identical for both the gradual and rapid uplift models, the surface uplift and sediment flux responses to the rapid uplift model are notably different (Figs. 6A and 6B). Between 43 and 22 Ma, low rock uplift rates (i.e., 0.1–0.2 mm/yr) in the Eastern Cordillera lead to only 1.7 km of peak surface uplift over this period of time. As such, mean sediment supply rates from the mountain belt remain low (i.e., <10 m$^2$/yr) over the first 21 m.y. of simulation. Despite the fact that the mean sediment flux into the foreland basin is half the rate of the gradual uplift model, the foredeep and back-bulge basins in the rapid uplift model are able to fill and bypass sediment from the mountain belt by 25 Ma. Following an additional 12 m.y. of active uplift between 22 and 10 Ma, peak elevation of the Eastern Cordillera in the rapid uplift model rises to 2.3 km above msl, and deformation propagates east into the western Subandean zone. Mean sediment flux into the foreland basin and sediment fluxes leaving the back-bulge basin in the rapid uplift model remain low compared to the values for the gradual uplift model at this time. At 10 Ma, rock uplift rates across the mountain belt increase from 0.1 to 0.3–0.5 mm/yr as the central Andes rapidly uplift in response to deformation in the Subandean zone. As a result, mean sediment flux from the mountain belt increases from 20 to 60 m$^2$/yr. However, the sediment flux leaving the back-bulge basin decreases shortly after the initiation of rapid uplift across the mountain belt because subsidence rates in the foredeep quickly respond to surface uplift within the mountain belt, trapping sediment within the foredeep near the edge of the mountain belt. The final topography in the rapid uplift model at 0 Ma equally reproduces the modern mean and maximum topographies of the central Andes mountain

![Figure 6](https://pubs.geoscienceworld.org/books/chapter-pdf/963891/mwr212-17.pdf)

**Figure 6.** (A) Topographic evolution of the central Andes and (B) time series of sediment supply rates into and out of the foreland basin for the rapid uplift model. The thick black lines in A represent snapshots of topography throughout the simulation, and the thick gray lines represent the maximum and mean topography for a north-south sweep of modern topography between 18°S and 21°S. The black boxes represent the locations of the forebulge crest during each snapshot in time, with the oldest forebulge location on the left. In B, the thick line represents the sediment flux leaving the back-bulge basin or right edge of the model, and the thin lines represent the sediment flux at the deformation front into the foredeep of the foreland basin.
An analysis of the Chaco foreland basin in southern Bolivia

Constraints on Foreland Basin Depositional Rates

In both the gradual and rapid uplift models, surface uplift of the eastern margin of the central Andes leads to an increase in sediment supply and subsidence within the foreland basin. Figures 7A and 7B show the sediment thickness curves for four depozones (i.e., the forethrust region of the Eastern Cordillera, western Interandean zone, western Subandean zone, and eastern Subandean zone, respectively) within the foreland basin for both the gradual and rapid uplift models. The solid lines represent the results for the simulations described in the previous section. Between 43 and 22 Ma, deformation and surface uplift are limited to the Eastern Cordillera. At this time, both the eastern region of the Eastern Cordillera and western Interandean zones are located within the foredeep basin of the gradual uplift model (Fig. 7A). Between 40 and 30 Ma, the deformation front remains in the core of the Eastern Cordillera. A spatially fixed load leads to constant subsidence rates and thus constant depositional rates in the Eastern Cordillera depozones over this period of time. The Interandean depozone, however, is located in the distal foredeep, and therefore it develops an acceleration in depositional rates between 31 and 25 Ma that is characteristic of an increase in subsidence rates due to an approaching mountain belt. Further to the east, the western and eastern Subandean depozones are located within the back-bulge basin. Low subsidence rates lead to low depositional rates between 43 and 36 Ma. Between 36 and 25 Ma, the forebulge crest migrates through the western Subandean depozone and leads to erosion and development of an unconformity. Sediment deposited while the eastern Subandean depozone was located in the back bulge is completely eroded away during this time period. The shaded region between 36 and 25 Ma represents the approximate depositional age for the Camargo Formation. Predicted depositional thicknesses for the Camargo Formation range between 1 and 2 km in the Eastern Cordillera and Interandean depozones, but they underpredict the >2 km thickness of the Camargo Formation observed in the Camargo syncline (DeCelles and Horton, 2003). By 22 Ma, deformation propagates into the Interandean zone. Therefore, the Interandean depozone is uplifted and exhumed at this time. Between 22 and 10 Ma, depositional rates within the western Subandean zone accelerate as it reaches the foredeep. At 9 Ma, a sharp inflection point occurs that reflects a significant increase in depositional rates in the eastern Subandean zone. These high depositional rates are inferred to be the result of high rock uplift rates in the western Subandean zone. Gradual uplift model simulations were also conducted for constant and lower rock uplift rates for the western Subandean zone. However, a slow and gradual kinematic model for the deformation front could not reproduce the extreme increase in isopach thickness for the Tariquía Formation given the constraints on deformation propagation rates during the late Miocene. Reproducing accurate accommodation magnitudes for the foreland basin stratigraphy while also fitting the modern topography motivated us to apply an increase in rock uplift rates at the front of the mountain belt relative to the gradual uplift model. Rock uplift rates within the back thrust and forethrust of the Eastern Cordillera, however, remained gradual and constant. Increased uplift rates between 9 and 4 Ma caused an increase in sediment accommodation rates. As a result, a larger portion of the incoming sediment flux is stored in the foredeep depozone.

Figure 7. Uncompacted Tertiary sediment thickness for depozones located in the eastern Subandean zone (ESA), western Subandean zone (WSA), western Interandean zone (IA), and within the Eastern Cordillera forethrust region (EC) for the (A) gradual uplift and (B) rapid uplift models. The dotted, solid, and dashed lines represent simulations with flexural parameters of 100, 150, and 235 km. Shaded regions represent the time periods during which the Tariquía Formation of the Subandean zone and Camargo Formation of the Eastern Cordillera and Interandean zone were deposited. The subsidence curve from Uba et al. (2006) for the location in the Subandes that we sampled is represented by black triangles in both plots.
 Extremely high depositional rates continue until ca. 4 Ma, when the basin begins to uplift and exhume. This period of time represents the deposition of the Guandacay Formation and the transition into the Emborozú Formation.

Although shortening rates and thrust belt propagation rates in the rapid uplift model are the same as in the gradual uplift model, the total thicknesses of basins for the rapid uplift model are significantly different than those for the gradual uplift model (Fig. 7B). Early in the model experiment, between 43 and 22 Ma, lower rock uplift rates within the Eastern Cordillerá cause depositional thicknesses within the eastern forethrust region of the Eastern Cordillerá and Interandean zones to be 500 m less compared to thicknesses predicted in the gradual uplift model. Both locations develop Camargo Formation thicknesses of ~500 m, which is much less than the observed >1 km thickness for the Camargo Formation in the Camargo syncline. By 12 Ma, low rock uplift rates in the Eastern Cordillerá and Interandean zones lead to a factor of 2 decrease in the total sediment thickness of the western Subandean depozone compared to the thickness predicted for this depozone in the gradual uplift model. At 9 Ma, the entire mountain belt between the western Subandean zone and Altiplano rapidly uplifts. This rapid uplift causes an order of magnitude increase in depositional rates within the eastern Subandean depozone. Approximately 2 km of sediment are deposited between 8 and 6 Ma. However, this is still 1–1.8 km less than the maximum isopach values reported by Uba et al. (2006) for the Tariquía Formation.

Additional simulations were conducted for both surface uplift models to test the sensitivity of the sediment thickness curves to the rigidity of the South American plate. The dotted lines represent the results for simulations with the lowest rigidities within the eastern forethrust region of the Eastern Cordillerá and Interandean zones. The dashed line to the rigidity of the South American plate (i.e., 4.0 × 10^24 Nm) causes the maximum depositional thickness of the three westernmost depozones to increase with respect to the initial results. Changing the rigidities of the South American plate in the rapid uplift model has similar effects as in the gradual uplift model (Fig. 7B). Although the rigidity of the South American plate can range over an order of magnitude, the predicted thickness of the Camargo Formation within Eastern Cordillerá and Interandean depozones of the rapid uplift model is less than a kilometer.

Role of Eclogite Root Foundering on Surface Uplift and Foreland Development

The eclogite root foundering simulation involves the growth and removal of a subsurface load that modifies the flexural response of the foreland basin and rock uplift distribution within the mountain belt. Prior to 10 Ma, rock uplift rates and shortening rates are prescribed to be identical to the rapid uplift model for the region of the mountain belt that is deforming. Between 25 and 10 Ma, an eclogite load is allowed to uniformly grow beneath the Altiplano and back-thrust region of the Eastern Cordillerá, where the deformed crust is sufficiently thick for lower-crustal rocks to undergo a phase transition to eclogite (Fig. 8A). The subsurface eclogite load, the location of which is represented by the black bar in Figure 8, sets up a flexural profile (identified by the thick black line) that superimposes onto the flexural profile caused by the topographic load. Although the total deflection due to the presence of the eclogitic root is small (i.e., ~1 m) at 25 Ma, by 10 Ma...
the total deflection is on the order of 1 km at the center of the eclogite root. At 25 Ma, the edge of the eclogite load is located ~150 km from the current deformation front. As a result of the distance between the load and deformation front, subsidence caused by the eclogite load is small in the foredeep, and nearby to the east, subsidence changes to rock uplift in the distal foredeep. As time progresses, the Interandean and Subandean depozones are subject to an additional component of rock uplift because they are located on the flexural forebulge related to the eclogite root load. Following 10 Ma, the polarity of lithospheric deflection reverses as the eclogite root is removed (Fig. 8B). The solid black curve is the total amount of deflection that occurs over 3 m.y. as the subsurface load is removed. A significant amount of rock uplift occurs directly over the center of the load in the eastern Altiplano and western back-thrust region of the Eastern Cordillera and rapidly decays into the core of the Eastern Cordillera. Further east, the Interandean and western Subandean zones experience subsidence on the order of meters to tens of meters. Beyond the western Subandean zone, the deflection due the eclogite foundering is less than a meter. Rock uplift rates in the Altiplano and the core of the Eastern Cordillera due to the eclogite root foundering range between 3.0 and 6.0 × 10⁻⁴ m/yr. Maximum deflection rates in the eastern Interandean and western Subandean zones are on the order of 10⁻⁷ and 10⁻⁵ m/yr. Rock uplift rates near the mountain front and within the foreland basin (i.e., Interandean and Subandean zones) prescribed in the rapid uplift model at 10 Ma are on the order of 6.0–5.0 × 10⁻⁴ m/yr, and, therefore, the flexural signal from a delamination event is small when compared to the subsidence or rock uplift due to crustal thickening. Conversely, rock uplift rates due to the delamination are slightly greater in the Altiplano and the back-thrust region of the Eastern Cordillera than the prescribed uplift rates due to crustal thickening.

Sediment supply into the foredeep predicted by the eclogite root foundering model shows a similar behavior to that of the rapid uplift model. Sediment supply rates are low until ca. 10 Ma, when the eclogite drip event accompanied by the rapid uplift of the eastern Interandean and western Subandean zones lead to increased channel slopes in the front of the mountain belt (Fig. 9A). The amount of sediment bypassing the foreland basin in the eclogite root foundering model is also similar to the rates predicted by the rapid uplift model, with the exception of a greater magnitude of sediment flux between 15 and 5 Ma. This increase in sediment leaving the foreland per time period is either the result of increased erosion rates in the mountain belt or decreased sediment accommodation rates in the foreland basin.

A comparison of sediment thicknesses predicted by the rapid uplift and eclogite root foundering models through time for the Interandean, western Subandean, and eastern Subandean zones shows that the drip event modified sediment accommodation by less than 10% of the maximum basin thicknesses before foundering (Fig. 9B). The western Subandean zone basin deposits are ~100 m thicker at 22 Ma for the eclogite root foundering model (solid line) compared to the rapid uplift model (dashed line). Between 25 and 22 Ma, the western Interandean zone is located close enough to the edge of the eclogite load to experience subsidence, which causes the change in sediment thickness between models. Conversely, the Subandean basins are located far enough from the eclogite load at that time to undergo rock uplift. The western Subandean basin is the least affected basin by the drip event because of its proximity to the inflection point between subsidence and rock uplift caused by accumulation of eclogite. The eastern Subandean zone also experiences rock uplift due to foundering, and, thus, lesser sediment thicknesses between 20 and 10 Ma. Decreased accommodation in the Subandean zones leads to an increase in sediment leaving the basin prior to 10 Ma. Following 10 Ma, the drip event leads to minor subsidence in the eastern Subandean zone, and the sediment thickness difference between the two models becomes negligible.

Figure 9. (A) Time-series data for sediment fluxes into and out of the foreland basin and (B) uncompacted sediment thickness for depozones located in the eastern Subandean (ESA), western Subandean (WSA), Interandean (IA), and Eastern Cordillera (EC) tectonomorphic zones for the eclogite delamination model. The black symbol at the top of A and the bottom of B represents the growth and delamination of the eclogite root between 25 and 10 Ma. In A and B, the thick lines represent the eclogite delamination model, and the thin dashed lines represent the rapid uplift model.
Role of Climate Change in Foreland Basin Development

The results for the climate change experiment (i.e., a doubling of the bedrock erodibility and sediment transport coefficients in late Miocene time) are reported here. Bedrock uplift rates during the late Miocene within the actively deforming mountain belt in this simulation are increased with respect to those of the rapid uplift model to maintain similar surface uplift histories, and, thus, the basin accommodation histories between the rapid uplift model and the climate change model are the same. We only report observations for sediment flux and grain size. Although erosion rates and sediment transport rates are a factor of two lower for the first 35 m.y., the first-order behavior of sediment flux entering the basin appears to be unchanged from that of the rapid uplift model (Fig. 10). The effect of lowering both erosion and sediment transport coefficients by a factor of two early on in the simulation is a few-million-year delay in the time when sediments begin to exit the back-bulge basin. Decreasing both erosion rates and sediment transport rates also has a cumulative effect of decreasing sediment bypass rates (solid black line) by almost half the value of sediment bypass predicted by the rapid uplift model (dashed line). Following 9 Ma, the sediment bypass rates suddenly increase by a factor of 1.5 times the previous flux.

An abrupt change in erosion rates and transport efficiency due to climate change should be expected to have an effect on grain size within the depositional basin, and, therefore, we tracked the gravel-sand interface for each of the experiments (Fig. 11). A comparison of the gradual uplift, rapid uplift, and climate change model results (Fig. 11) reveals that in general the gravel-sand boundary closely tracks the location of the deformation front (shown as the dashed line). An exception to this observation occurs immediately following times when the deformation front propagates basinward, and the gravel-sand interface lags behind the deformation front for ~1–2 m.y. until the regional slope of the newly formed wedge-top basin increases above the threshold slope for gravel transport. Once gravel reaches the deformation front, gravel progradation into the foredeep appears to be limited within a narrow zone of ~50 km from the deformation front. Figure 11A compares the gravel-sand boundaries for the gradual and rapid uplift models. Prior to 10 Ma, the mean location of the gravel-sand boundaries for both models generally remain in front of the deformation front following the 2 m.y. lag periods. However, the initiation of rapid uplift in the Interandean and western Subandean zones at ca. 10 Ma causes the gravel-sand boundary to retrograde back into the wedge-top zone as accommodation rates exceed sediment supply rates. Both models appear to closely overlap each other, with two exceptions. Between 22 and 15 Ma, the gravel front in the gradual uplift model progrades...
more rapidly than the gravel front of the rapid uplift model due to a combination of greater uplift rates in the Interandean zone wedge-top basin and greater sediment supply rates. Between 8 and 5 Ma, the mean location of the gravel-sand interface of the rapid uplift model retrogrades more than the gravel-sand interface of the gradual uplift model due to greater subsidence rates. Figure 11B compares the results for the climate change model to the rapid uplift model and, thus, should highlight the effects of changes in precipitation on grain size. Although both the bedrock erodibilities and transport coefficients are a factor of two less in the climate change model compared to the rapid uplift model between 25 and 9 Ma, the gravel-sand boundaries generally track each other well through time. However, there are a few brief periods of time when the gravel-sand interface of the climate change model lags behind that of the rapid uplift mode. When the bedrock erodibilities and transport coefficients increase by a factor of root 2, the results of the climate change model do not vary significantly from the trends of the rapid uplift model.

DISCUSSION

Two end-member surface uplift models have been proposed for the central Andes (Garzione et al., 2008). One model involves a long history of constant deformation since the late Eocene, accompanied by a gradual increase in mean surface elevation prior to 10 Ma. The opposing end-member model suggests that the central Andes did not gain significant elevation until after 10 Ma. Our results show that if the elastic properties of the underthrust South American plate remain effectively constant, then rapid rock uplift is likely to have occurred within the Interandean and western Subandean zones during the late Miocene in order to generate sufficient accommodation space for the Tariquía Formation observed in outcrops within the central to eastern Subandean zone (Fig. 7). Constraints on the initiation of rapid rock uplift in the Interandean and Subandean zones are based on the isopach data for the Yecua and Petaca Formations. If rock uplift rates were greater in the Interandean and Subandean zones prior to 9–8 Ma, when the Tariquía Formation began to be deposited, then the isopach trends for the Yecua and Petaca Formations observed within the central Subandean zone would show maximum thicknesses greater than 825 m. Rock uplift rates prescribed for the Interandean and western Subandean zones in the gradual uplift model lead to maximum thicknesses for the Petaca-Yecua deposits that are on the order of 1 km. Although other studies have proposed that decreasing the flexural rigidity of the South American lithosphere can explain an increase in accommodation for the Tariquía Formation (Prezzi et al., 2009), we show that a constant flexural rigidity model can also fit the late Miocene isopachs well.

At 10 Ma in the models, both the Altiplano and Eastern Cordillera are sufficiently far from the central and eastern Subandean zones such that their contribution to the accommodation space is low compared to that of the Interandean and western Subandean zones. The deflection of an elastic beam in response to a point load decreases exponentially with distance from the center of the load. As such, a comparison of the sediment thickness time-series results for the gradual and rapid uplift models shows little change in accommodation rates following 10 Ma in the eastern Subandean zone, even though the peak elevations of the Eastern Cordillera differ by 2 km (Fig. 7). Although the difference in peak elevation within the Eastern Cordillera does not strongly affect late Miocene foreland basin sediment accommodation, we can constrain the early surface uplift history of the Eastern Cordillera and Altiplano with the middle Cenozoic stratigraphy of the Eastern Cordillera and western Interandean zones because these depozones were located at least 100 km closer to the Eastern Cordillera than the central Subandean zone. DeCelles and Horton (2003) measured a thickness of over 2 km for the Camargo Formation within the Camargo syncline of the Eastern Cordillera. A comparison of sediment thicknesses deposited between 20 and 22 Ma predicted by the gradual and rapid uplift models shows that a sediment accommodation thickness of 2 km within the early foreland basin is not achievable unless rock uplift rates in the core of the Eastern Cordillera more closely resembled the rock uplift rates specified within the gradual uplift model (Fig. 7). Although there is uncertainty in the depositional age of the Camargo Formation, the rapid uplift model could not create sufficient sediment accommodation space if deposition began as early as 40 Ma. If the erosion rates in our model are close to the effective erosion rates during the early Miocene and late Oligocene, then the thick deposits of the Camargo Formation would imply that the peak elevation of the Eastern Cordillera was between 2 and 3 km by 22 Ma.

An Eastern Cordillera peak elevation of 2–3 km might have acted as an effective topographic barrier to moisture that was being transported west across the central Andes. Today, the topography of the eastern flank of the central Andean Plateau prevents a significant amount of moisture that originates from the Atlantic Ocean and Amazon Basin from being transported west into the Altiplano and Western Cordillera by the South American summer monsoon (Strecker et al., 2007). A similar behavior occurs in the southern Andes, where the moisture from the Pacific Ocean carried by the Southern Hemisphere westerlies rains out on the west coast of South America and the Patagonian Andes, leading to semiarid conditions on the eastern flank of the Andes. The southern Andes are an effective moisture barrier even though their mean elevation is around 2 km lower than the central Andes, with peak elevations ranging from 2 to 3 km between 38°S and 50°S. We envision the topography of the central Andes during the early to middle Miocene as being similar to the southern Andes today, which suggests that the orogen may have been an effective barrier to moisture being transported to the southwest from the Atlantic. If this is the case, then basins within the Altiplano and Western Cordillera regions of the central Andes should become increasingly arid around or soon after 22 Ma. Based on abrupt changes in lacustrine facies within basins, the onset of hyperaridity has been documented to have occurred between 10 and 6 Ma for basins within the Western Cordillera between 18°S and 22°S (Gaupp et al., 1999; Sáez et al., 1999). However, based...
on oxygen isotopes, soil morphological characteristics, and salt chemistry. Rech et al. (2006) and Rech et al. (2010) proposed that the onset of hyperaridity could have occurred in the Atacama Desert earlier, between 19 and 13 Ma. In northern Argentina between 25°S and 26°S, recent thermochronologic data from the Puna Plateau show that the Eastern Cordillera was deformed and exhumed at the same time as the Eastern Cordillera further to the north in southern Bolivia (Carrapa et al., 2011a, 2011b). Interestingly, Vandervoort et al. (1995) also documented an earlier shift from nonevaporitic to evaporitic sedimentary deposits within Puna Plateau basins located between 24°S and 26°S at ca. 15 Ma. A middle Miocene onset of aridity-hyperaridity within the Andean Plateau (perhaps as far south as the Puna) and basins on the western margin of the central Andes would be consistent with the peak elevations within the Eastern Cordillera that were necessary to generate sufficient accommodation space for the Camargo Formation.

Another output of our numerical model during the end-member uplift model experiment was a time series for sediment leaving the foreland basin (Figs. 5 and 6). Changes in the sediment bypass rates would not be directly recorded in the foreland basin stratigraphy, and, therefore, one must look at the stratigraphy of adjacent intracratonic basins or the continental shelf to directly sample this signal. Presently, a topographic divide exists in southern Bolivia that splits the flow of major rivers draining off the central Andes north into the Amazon and southeast into the Rio del La Plata cratonic basins. Based on the sediment bypass time series for each of the simulations, the sediment flux component from the central Andes should steadily increase from the late Oligocene into the Quaternary, with a local minimum between 8 and 3 Ma. Analysis of drill-core sediments from the Amazon fan and Ceara Rise show that Andean-derived sediments reached the continental margin between 16.5 and 11.3 Ma (Dobson et al., 2001; Figueiredo et al., 2009). Between 9 and 6.8 Ma, sedimentation rates increased for both the Amazon fan and Ceara Rise. This increase in sedimentation rates was interpreted as the establishment of a transcontinental Amazon drainage system that fully linked the Andean forelands to the Amazon fan. Sedimentation rates continued to increase into the Pliocene-Pleistocene during the period of most rapid uplift within the northern Andes (Hoorn et al., 1995). An overall increase in sedimentation rates on the Atlantic shelf is predicted by our model. The higher the mean elevation of the mountain belt, the more the sediment supply may exceed available accommodation space. Our model also predicts a factor of two increase in sediment flux leaving the foreland basin during the late Pliocene to Pleistocene as foredeep accommodation rates decrease in response to slower rock uplift rates in the Interandean and Subandean zones. One aspect of our model results that is not recorded in the Amazon fan is a late Miocene to early Pliocene decrease in sediment supply rate. The absence of a drop in sedimentation rates may be expected because a small portion of the Amazon Basin drains the central Andes, and, therefore, sediment supply to the Amazon fan is more predominantly influenced by the northern Andes. A stratigraphic data set that would be more predominately influenced by the central Andes would be the deposits of deep-water fans offshore of the Rio del la Plata estuary, which samples the Andes between 18°S and 34°S. To our knowledge, no analysis has been conducted for the late Cenozoic sediments of the Rio del la Plata fans.

A hypothetical eclogite foundering event in the eastern Altiplano region during middle Miocene to Pliocene time was synchronous with increased grain size and depositional rate in the foreland basin stratigraphy. Our results show that the distance from the center of mass of the load is the most important parameter for determining how the growth and removal of a lower-crustal load would affect rock uplift and subsidence within the mountain belt and foreland basin (Fig. 8). An inflection point between subsidence and uplift occurs at a distance of \((\pi/2)\alpha\) (i.e., ~235 km for this study) from the growing eclogite root load. During the Oligocene to early Miocene, the foredeep depozone was located closer to the eclogite root than it would be for the remainder of the simulation (i.e., less than \((\pi/2)\alpha\)). As a result, the accumulation of eclogite drove an additional 100 m of subsidence, which is more than a factor of 2 greater than subsidence added to the western and eastern Subandean depositional basins during the late Miocene. The greatest deflection occurred directly above the center of the eclogite root, which was located in the eastern Altiplano and westernmost Eastern Cordillera. This deflection led to ~0.5–1 km of rock uplift in the Altiplano. Therefore, an additional 2 km of surface uplift due to crustal thickening and sediment deposition are required to achieve the modern mean elevation of the Altiplano if it was located at 1 km around 10 Ma. Similar amounts of rock uplift due to crustal thickening (i.e., 1–2 km) are required in the Eastern Cordillera to reach its modern mean and spatially averaged maximum elevations. However, less rock uplift due to crustal thickening is necessary if the average thickness of the eclogite root was significantly larger than the value applied in this study. We infer that this is less likely because a thicker eclogite root would grow a significantly sized instability in less than a period of 3 m.y., which is the proposed length for the delamination period. Our results also show that <0.5 km of rock uplift is contributed to the eastern edge of the Eastern Cordillera and western Interandean zones by eclogite removal. Barke and Lamb (2006) calculated that the San Juan del Oro paleosurface, which overlies the eastern part of the Eastern Cordillera, was uniformly uplifted by ~1 km. Therefore, the eclogite root must be located closer to or beneath the forethrust of the Eastern Cordillera to uniformly uplift it by 1 km. However, tomography data suggest that foundering is more likely beneath the Altiplano and western part of the Eastern Cordillera (Beck and Zandt, 2002). Thus, part of the 1 km of rock uplift of the San Juan del Oro surface is likely the result of crustal thickening.

Climate change was the final process that we tested for the foreland basin of the central Andes of southern Bolivia. Erosion rates and transport coefficients between 43 and 9 Ma were a factor of 2 less than the values between 9 and 0 Ma. As a direct result, peak sediment flux into the foreland basin and sediment bypass rates were significantly less than the values resulting from
the rapid uplift model (Fig. 10). At the onset of the South American monsoon, both sediment bypass rates and sediment flux rates into the foredeep increased significantly. The increase in sediment supply to the foredeep is predominantly controlled by the rapid uplift of the mountain belt instead of by the increase in mean annual precipitation. Increasing erosion rates and decreasing transport rates should lead to steeper slopes in the proximal foreland basin. Both sediment supply and transport capability increase as mean annual precipitation increases, which leads to minimal change in the regional slopes for the foreland basin, as these effects cancel each other. Regardless, regional slopes in the foreland basin component of the model appear to be predominately affected by rapid subsidence due to crustal thickening near the deformation front instead of by increasing precipitation. Another way to gauge the effect of climate change is to track the boundary between gravel and sand deposition. Paola et al. (1992) demonstrated that sinusoidal variations in both sediment supply and transport coefficients lead to migration of the gravel-sand interface, especially if the forcing period is small compared to the basin equilibrium time. Our results show that the first-order migration of the gravel-sand boundary appears to be unaffected by the factor of 2 increase in both of these parameters over the time scales that we are sampling (Fig. 11). Based on our results, the threshold slope term appears to be a more primary control on gravel progradation than the transport coefficient term. Rock uplift in the hinterland leads to steeper slopes that are capable of transporting gravel. As a result, gravel trapped near the mountain front can rapidly prograde toward the new deformation front when it is located in a wedge-top basin. Eventually, rock uplift leads to the exhumation of pre-Cenozoic units near the deformation front, which can produce gravel during bedrock incision. Conversely, the regional slope within the foreland is insufficient to transport gravels far into the foredeep. Therefore, a combination of processes causes progradation of the gravel-sand interface during forward propagation of the deformation front; climate change appears to have played a secondary role in controlling gravel progradation for the central Andes.

CONCLUSIONS

Based on our modeling results, we propose that the early surface uplift history (i.e., prior to 22 Ma) of the eastern margin of the central Andes more closely resembled the gradual uplift model. When the rigidity of the South American plate ranged between $1.5 \times 10^{23}$ and $4.0 \times 10^{23}$ Nm, surface uplift of the core of the Eastern Cordillera in the rapid uplift model produced grossly undermatched sediment accommodation during the deposition of the >2-km-thick Camargo Formation. The gradual uplift model more closely fits the observed sediment thicknesses for the Camargo Formation located in the Eastern Cordillera, and, thus, the core of the Eastern Cordillera experienced rock uplift rates that would have led to significant topography (i.e., peak elevations near 2–3 km as shown in Fig. 5A) by the early Miocene if basin-averaged erosion rates were on the order of $10^{-4}$ m/yr. Further crustal thickening during the middle Miocene may have increased mean topography of the Eastern Cordillera to the point where it became an effective barrier to easterly moisture derived from the Atlantic (as shown in Fig. 5A); in turn, this initiated aridity in the Western Cordillera and Atacama Desert regions at that time. Our results suggest that peak topography of the Eastern Cordillera was well above 3 km prior to 10 Ma and, therefore, do not support the rapid uplift model, which predicts ~2 km of late Miocene rapid surface uplift within that region. However, our results do support rapid rock and surface uplift within the Interandean and western Subandean zones to produce the amount of accommodation required to store the thick Tariquia Formation, which was deposited within a period of 2 m.y.

Our results also support the hypothesis that the first-order trends in the Cenozoic foreland basin stratigraphy of the Subandean zone were predominantly influenced by its distance from the approaching mountain belt. An increase in topographic loads located less than $(3/4)\pi \alpha$ km from a depozone can cause deflections up to the order of a kilometer. Beyond this distance, deflection ranges up to hundreds of meters near the forebulge. During the late Miocene, the eclogite root was located more than a distance of $(3/4)\pi \alpha$ km from the Subandean depozone. As such, the accommodation rates within the central-eastern Subandean zone foredeep were positive, but only on the order of tens of meters. Therefore, the deflection caused by an approaching mountain belt exceeded the deflection caused by eclogite delamination. Also based on our results, the factor-of-five increase in depositional rates was not likely the result of an increase in erosion rates and sediment transport rates caused by climate change. Increased erosion rates would lead to isostatic rebound of the mountain front, and thus decreased sediment accommodation within the foredeep, if crustal thickening does not increase as well. If crustal thickening does increase with erosion rates, as in our model, then the additional sediment supplied bypasses the foreland basin because it is already overfilled. In addition to accommodation rates, the location of the deformation front in time appears to exert a primary control on the gravel-sand interface. Increasing erosion rates and sediment transport rates by a factor of 2 due to the onset of the South American monsoon at 9 Ma did not generate a long-term progradation in the gravel-sand interface. Instead, the gravel front retreated toward the deformation front due to the rapid subsidence rates within the foredeep basin. Actively uplifting fold-and-thrust belts are able to achieve channel slopes that are above the threshold for gravel entrainment and produce gravel by bedrock incision. As a direct result, the gravel-sand interface rapidly progrades when the deformation front propagates into the foreland basin.

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