Limit of channel flow in orogenic plateaux

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ABSTRACT

The eastward growth of the Tibetan Plateau has been attributed to the flow of the plateau's weak lower crust into its foreland over a distance of 1500 km in 15 m.y. This channel-flow extrusion requires a very low-viscosity deep crust prior to thickening. Here, we show through triaxial thin sheet models that Tibet's uplift rate and plateau elevation demand a prethickening Moho temperature of 500–600 °C. Such temperatures are incompatible with the viscosity necessary for channel-flow extrusion >1000 km. Using two-dimensional (2-D) coupled thermomechanical numerical experiments and prethickening temperatures compatible with Tibet's uplift history, we show that mass redistribution processes are dynamically coupled and that coupling is sensitive to rheology, channel buoyancy, and boundary conditions. Channel-flow extrusion velocities are limited to less than 1 cm yr^{-1} by cooling in the foreland and by any upward deviation of the weak channel by extension in the plateau or by erosion at the plateau margin.

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PLATEAU-FORELAND TRANSITIONS AND CHANNEL-FLOW EXTRUSION

Fast and deep channel-flow extrusion (Bird, 1991) east of Tibet is based on two-dimensional (2-D) thin sheet models and uses, as observables, Tibet's uplift history and the topography across the transition between the plateau and the foreland to the east. In these models, there is no horizontal shortening as a >5000-m-high plateau is inflated by pumping material at a constant rate into one end of a 10–15-km-thick weak channel embedded within a stronger crust. The pressure gradient from the plateau forces the channel material 1500 km into the foreland (Royden, 1996; Royden et al., 1997; Clark and Royden, 2000; Royden et al., 2008; Medvedev and Beaumont, 2006). The input flux rate is chosen so a plateau elevation of 5000 m is reached in 20 m.y., compatible with the uplift history of Tibet, the elevation of which was established sometime after 50 Ma and before 10 Ma (Clark et al., 2005; Dupont-Nivet et al., 2008). The topographic profile at the transition between the plateau and the foreland is a function of the thickness of the channel region and its viscosity (Clark and Royden, 2000; Vanderhaeghe et al., 2003; Medvedev and Beaumont, 2006), and its surface width is a measure of the distance traveled by the material in the channel. The >1500-km-wide topographic profile along Tibet's eastern margin, on either side of the Sichuan Basin, requires a viscosity in the channel between 10^{18} and 10^{17} Pa s (Clark and Royden, 2000). Continental crust may attain such viscosities for temperatures >800 °C, or even >700 °C when melt is present (Medvedev and Beaumont, 2006).

This model poses a number of problems. First, convergence and the gravitational push of the plateau onto its foreland are not considered as a mode of plateau formation, yet viscous thickening of the foreland region can be critical to the topographic profile at the plateau-foreland transition (Medvedev and Beaumont, 2006). Second, although temperatures >800 °C can be explained in the plateau by transient radiogenic heating, it is not clear how such a high temperature can be achieved in the foreland, and how it would not make the foreland more prone to viscous thickening under the gravitational push from the plateau. Third, in 2-D plane strain models, high plateaux can develop from weak crust because the latter cannot flow in a direction perpendicular to convergence. This may not be possible in three dimensions (3-D).

Recently, channel-flow extrusion was investigated using a quasi-3D indenter model (Cook and Royden, 2008). In this model, the viscosity of the reference crust is uniform with depth and laterally homogeneous. Upon convergence, a weak lower crust starts to develop when a critical crustal thickness (50 km) is reached. On the side of the indenter, there is a dormant "weak crust" region with the ability to weaken faster than the surrounding crust. With such a design, a 5 km plateau can develop with little channel-flow extrusion until the dormant "weak crust" thickens and differentiates a weak lower crust, at which stage the plateau lower crust "spills" into the weak foreland lower crust. In this model, the foreland region must become a plateau before channel-flow extrusion can develop, and therefore channel-flow extrusion is a consequence of crustal thickening and not its cause. In the model by Cook and Royden (2008), transport in the weak lower crust foreland may not exceed a few tens to a few hundreds of kilometers, in contrast to the Clark and Royden (2000) model, in which channel-flow extrusion is the only mode of foreland thickening.

Hence, it is not clear whether a high plateau region can develop at a rate compatible with Tibet's uplift history while developing a >1000 km transitional region via channel-flow extrusion in less than 15 m.y. To attempt a more realistic understanding of crustal flow, plateau uplift, and plateau-foreland dynamics, we consider first a generic model of plateau formation in a triaxial stress setting, which suggests that, for average continental rheologies, Tibet could only have developed at the expense of a continental crust with a Moho temperature of 500–600 °C, a range incompatible with channel-flow extrusion >1000 km. Using this temperature range, we turn to 2-D numerical modeling and show that under such a condition, the length scale of channel-flow extrusion is an order of magnitude smaller than previously thought.

PLATEAU UPLIFT, ELEVATION, AND MOHO TEMPERATURE

In a 3-D setting, orogenic plateaux express the balance between the rate of thickening related to convergence and the rate of thinning related to lateral gravitational flow. Triaxial thin viscous sheet models (Fig. 1A)

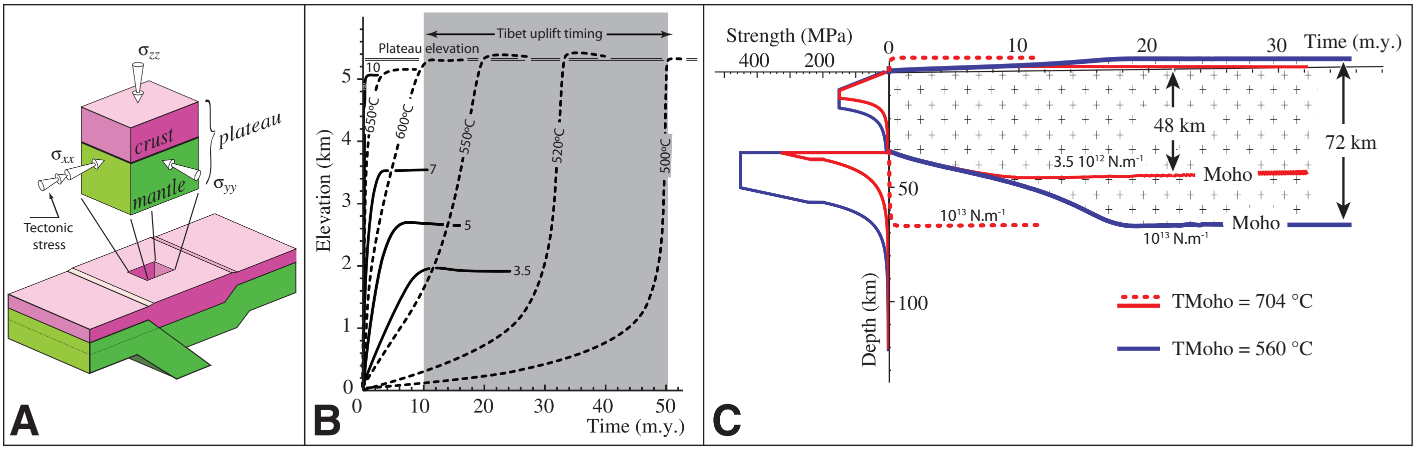

Figure 1. Triaxial thin sheet model of an orogenic plateau under ongoing convergence. (A) Conceptual model of an orogenic plateau submitted to a triaxial state of stress. (B) Uplift evolution for two series of experiments. Dashed lines—a constant driving force of 10^{13} N m^{-1} is applied to lithospheres of decreasing Moho temperatures. Continuous lines—a lithosphere with a Moho at 700 °C is submitted to various tectonic forces (with numbers ×10^{12} N m^{-1}). (C) The initial strength profiles and the position of the density interfaces through time for a hot and colder model.

have shown that both the uplift rate and the plateau elevation depend on the geotherm (Rey and Houseman, 2006; Rey and Coltice, 2008). Given Tibet's average elevation (~5000 m) and assuming that the plateau developed over the past 10–50 m.y., one can reconstruct the temperature at the Moho prior to thickening. Figure 1B shows the uplift history of a continental lithosphere for various prethickening Moho temperatures when this lithosphere is submitted to a triaxial state of stress, isostasy, and radiogenic heating. Using standard crustal and mantle rheologies (see GSA Data Repository[1]), our models require a tectonic force of 10^{13} N m^{-1} to reach an elevation compatible with that of Tibet. Our triaxial experiments also reveal that, in order to match both the timing of Tibetan uplift and its elevation, the continental lithosphere requires a prethickening Moho temperature in the range of 500–600 °C, well below the 700–800 °C necessary for channel-flow extrusion >1000 km. A plateau developing at the expense of a hot lithosphere (Figs. 1B and 1C) cannot match both the timing of Tibetan uplift and Tibet's elevation. Therefore, the proposition that channel-flow extrusion is >1000 km in east Tibet is incompatible with Tibet's elevation and uplift history.

MODES OF OROGENIC MASS TRANSFER AND PLATEAU COLLAPSE

We use fully coupled thermo-mechanical numerical experiments to examine the modes of orogenic mass transfer between plateau and foreland in a context in which crustal rheologies allow building of an orogenic plateau, and where the foreland is fixed or slowly retreats. The model setup (Figs. $2A_0$ and $2B_0$) describes a plateau-foreland transition after convergence has stopped, and also approximates a plateau "neutral margin" (Medvedev and Beaumont, 2006), such as the eastern margin of the Tibetan Plateau, on which convergent plate motion has a limited effect. We use frictional plastic and laboratory-derived viscous rheologies dependent on temperature, stress, strain rate, and melt fraction, along with melt-dependent viscosities and densities (see GSA Data Repository item [see footnote 1]). The role of erosion, already studied in detail (Beaumont et al., 2001, 2004), is not included in our experiments. Instead, we introduce weak fault-shaped anomalies and use a strain weakening function to induce heterogeneous deformation in an otherwise strong upper crust; this weakening promotes the localization of upward flow in the plateau and thrusting in the foreland.

The foreland region is in thermal equilibrium and has a Moho temperature of 560 °C, and the plateau region has accommodated some degree of thermal relaxation. Two cases are considered: (1) the Moho in the plateau region has reached 790 °C, and there is no melt at the onset of collapse (Fig. $2A_{0-4}$); and (2) the Moho in the plateau region has reached 870 °C, and the temperature in the lowest 14 km of the plateau crust is above the solidus, with a maximum melt fraction of 24% (Fig. $2B_{0-4}$), compatible with melt fraction estimates beneath Tibet (Schilling and Partzsch, 2001). Although a very weak foreland would not allow the development of a plateau, we tested the unlikely configuration in which an orogenic plateau is adjacent to a foreland for which viscosity is reduced by two orders of magnitude (Figs. $2A_4$ and $2B_4$) with a minimum of 10^{18} Pa s at the Moho.

Our results show that under fixed boundary conditions (Figs. $2A_{1-2}$ and $2B_{1-2}$), plateau collapse is accommodated by a combination of lateral channel-flow extrusion and upward mass transfer of the weak plateau channel that dynamically couples extension and the formation of a core complex in the plateau with shortening via thrusting in the foreland and bulk gravitational sliding of the plateau margin toward the foreland. As the weak plateau lower crust is partitioned into the foreland and the core complex, horizontal channel-flow extrusion competes with vertical flow in the plateau. Large melt fraction, large buoyancy of the channel, and weak foreland upper crust all favor the development of core complexes and coupled foreland shortening, which has the effect of impeding channel-flow extrusion (Fig. $2B_1$). In contrast, small melt fraction, low buoyancy, and a strong foreland upper crust favor channel-flow extrusion (Fig. $2A_2$). Isostasy alone is able to exhume the plateau ductile crust into a metamorphic core complex, even when the lower crust is not buoyant (Fig. $2A_{1-2}$; Wdowinski and Axen, 1992; Rey et al., 2009).

Divergent boundary conditions contribute to provide the necessary space for gravitational collapse (Rey et al., 2001) and the development of core complexes. Hence, one can expect that, under divergent boundary conditions, plateau-foreland coupling would be less important and surface extension in the plateau would be less dependent on the shortening of the foreland.

[1]GSA Data Repository Item 2010241, details on the density and thermal and rheological structures of the numerical models, is available at www.geosociety.org/pubs/ft2010.htm, or on request from editing@geosociety.org, Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301-9140, USA.

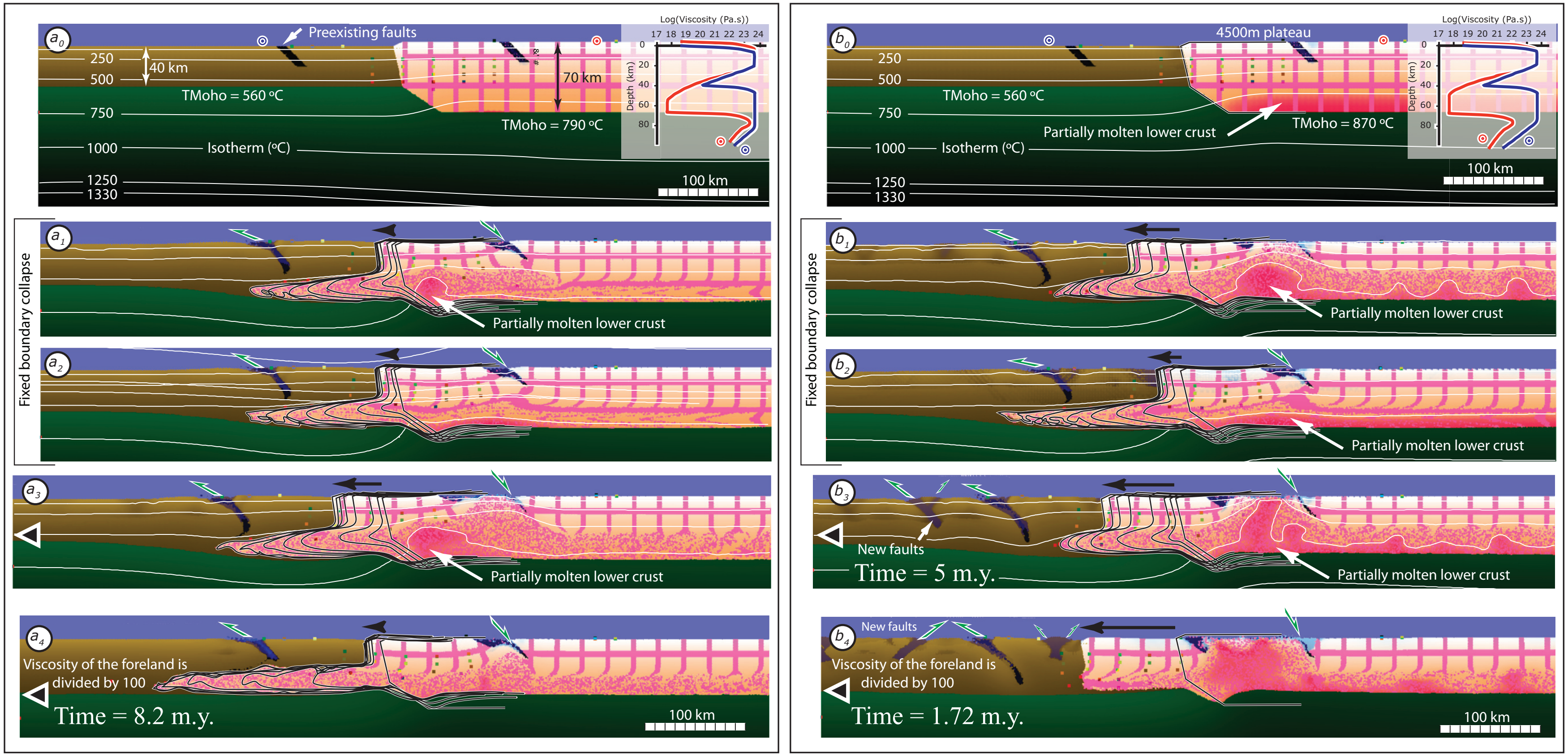

Figure 2. Numerical experiments of plateau collapse. Black lines show the plateau-foreland interface at 0, 1, 2, 4, 6, 8, 10, and 12 m.y. (A_{0-4}) Experiments with a initial plateau Moho temperature of 790 °C. (B_{0-4}) Experiments with a plateau Moho temperature of 870 °C. (A_1, B_1) Fixed boundary collapse. (A_2, B_2) Same as A_1–B_1 but with no decrease in density due to the presence of melt. (A_{3-4}, B_{3-4}) Models with slow foreland retreat to the left at 2.3 mm yr^{-1}. (A_4, B_4) Models with very weak forelands. The black arrow on top of each model shows the displacement of the plateau edge.

Under slow divergent boundary conditions (left side moving to the left at 0.23 cm yr^{-1}), the plateau flank travels 74 km in 5 m.y. at an average velocity of 1.5 cm yr^{-1}, approximately five times as fast as the imposed retreat of the foreland (Figs. $2A_3$ and $2B_3$). Consequently, stronger foreland shortening develops, accommodated by preexisting and newly formed reverse faults. A combination of gravity sliding and rear push from the developing gneiss (migmatite) dome in the plateau drives the motion of the plateau margin toward the foreland. There is a positive feedback between localized stretching in the plateau upper crust, induced by slow divergence, and the formation of a metamorphic core complex, which becomes the dominant driver of extension in the plateau and shortening in the foreland. Under fast divergence (2.3 cm yr^{-1}, not shown here), shortening of the foreland and channel-flow extrusion are reduced, while extension in the plateau is enhanced.

In the case in which the foreland is very weak but the plateau lacks melt-enhanced buoyancy (Fig. $2A_4$), channel-flow extrusion is the dominant mode of collapse, even though a gneiss dome has developed in the plateau. In sharp contrast, where plateau buoyancy is enhanced by partial melting (Fig. $2B_4$), the plateau gravitational push is sufficiently strong to shorten the whole foreland, displacing the flank of the plateau 100 km toward the foreland in less than 2 m.y. at an average velocity of 5 cm yr^{-1}. As in the model in Figure $2B_3$, shortening is coupled to an equivalent amount of extension in the plateau, which accommodates the formation of a large metamorphic core complex, in which the weak channel is exhumed.

In experiments most favorable to channel-flow extrusion, where foreland rheology allows for the development of an orogenic plateau, a wedge-shaped channel of the plateau lower crust travels at most 150 km into the foreland after 12 m.y. of gravitational collapse (Fig. $2B_2$). After only a few million years, the velocity of the channel front drops sharply from a few centimeters per year to 1 cm yr^{-1} or less when buoyant melt is present. This velocity decrease occurs as the pressure gradient at the foreland-plateau transition decreases and the hot channel cools to 500–650 °C as it travels into the foreland. Channel-flow extrusion is further inhibited by upward flow of the weak channel into metamorphic core complexes. Overall, channel flow velocities 10 cm yr^{-1}, which are necessary for channel-flow extrusion to travel 1500 km in 15 m.y., are not tenable.

APPLICATION TO SOUTHEAST TIBET

Despite ongoing shortening on Tibet's southern, northern, and eastern margins (Zhang et al., 2004; England and Molnar, 2005; Shen et al., 2005), east-west extension has been the dominant tectonic regime in the plateau since 15 Ma (Armijo et al., 1989; Molnar and Lyon-Caen, 1989). Conjugate strike-slip faults and N-S–trending normal faults with up to 7 km of vertical offset accommodate E-W extension in central and southern Tibet (Blisniuk et al., 2001). In southern Tibet, N-S–trending shear zones exhumed deep crust into metamorphic core complexes during orogen-parallel extension (Murphy et al., 2002; Jessup et al., 2008). In addition, channel flow related to aggressive erosion along the Himalayan front has been proposed to explain the exhumation of North Himalayan high-grade gneiss and leucogranite (Beaumont et al., 2001; Klemperer, 2006). Our experiments confirm that coeval extension in the plateau and contraction in the adjacent foreland is compatible with gravitational collapse under fixed or slowly retreating boundary conditions. In SE Tibet, the SE gradient in both crustal thickness and topography combined with limited upper-crustal shortening are the basis for the proposal of at least 400 km of SE-directed channel flow (Schoenbohm et al., 2006). Assuming that channel flow started 13 m.y. ago, following the rise of eastern Tibet (Schoenbohm et al., 2006; Clark et al., 2005), channel-flow extrusion can hardly have exceeded 150 km. The strong correlation between seismic anisotropy and surficial geology, including geodetic data, suggests a strong coupling between the crust and the mantle under E and SE Tibet (Holt, 2000; Sol et al., 2007). If the upper-crustal shortening has not been underestimated, then one can envision that crustal thickening is partitioned into the lower crust, possibly via bulk transpression against the Indian plate and around the eastern syntaxis, while strike-slip faulting affects the upper crust.

CONCLUSIONS

Tibet's uplift history and plateau elevation limit prethickening Moho temperatures to between 500 and 600 °C, a temperature range incompatible with channel-flow extrusion >1000 km. For rheologies and temperatures compatible with Tibet, our numerical experiments show that the rate of channel-flow extrusion is an order of magnitude smaller than previously proposed for eastern Tibet (i.e., 1 cm yr^{-1} rather than 10 cm yr^{-1}). Over 15 m.y., the length scale of channel-flow extrusion is 150 km rather than 1500 km. Our results show that during the plateau-growth stage of orogenic evolution, when extension associated with lateral gravitational spreading balances convergence-driven thickening, and when convergence slows or stops, mass is redistributed from the plateau to the foreland via a number of strongly coupled processes (Fig. 3), including: (1) gravity

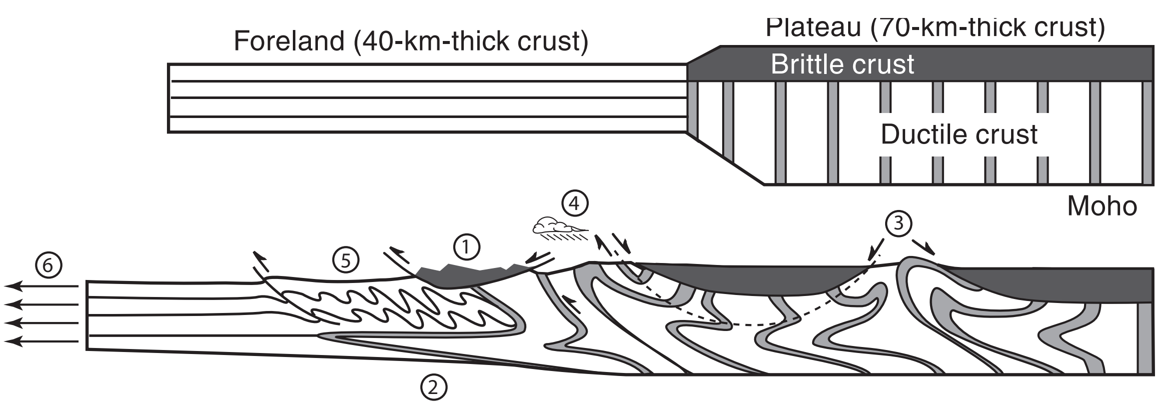

Figure 3. Conceptual model of gravity-driven plateau growth and orogenic collapse processes. Mass transfer is accommodated by (1) gravitational sliding of the edge of the plateau onto the foreland, (2) channel-flow extrusion of the plateau lower crust into the foreland, (3) upper-crustal extension and upward flow of the plateau lower crust into metamorphic core complexes, (4) upward flow of the plateau lower crust into regions of aggressive erosion, (5) shortening of the foreland immediately adjacent to the plateau, and (6) retreat of the foreland, promoting bulk extension and thinning of the plateau region.

sliding of the plateau margin toward the foreland; (2) channel-flow extrusion of plateau lower crust into the foreland lower crust; (3) upward flow of lower crust channel toward a domain of upper-crust extension to form a metamorphic core complex or (4) a site of focused erosion as shown by Beaumont et al. (2001); and (5) coeval contraction of the foreland, which balances extension in the plateau.

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