Topographic ooze: Building the eastern margin of Tibet by lower crustal flow

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ABSTRACT

Topography extracted from swath profiles along the northern, southern, and eastern margins of the Tibetan Plateau show two end-member morphologies: steep, abrupt margins and long-wavelength, low-gradient margins. Because the lack of significant upper crustal shortening across much of the eastern plateau margin implies that the crustal thickening occurs mainly in the deep crust, we compare regional topographic gradients surrounding the plateau to model results for flux of a Newtonian fluid through a lower crustal channel of uniform thickness. For an assumed 15-km-thick channel, we estimate a viscosity for the lower crust of 10^{18} Pa·s beneath the low-gradient margins, 10^{21} Pa·s beneath the steep margins, and an upper bound of 10^{16} Pa·s beneath the plateau. These results indicate that the large-scale morphology of the eastern plateau reflects fluid flow within the underlying crust; crustal material flows around the strong crust of the Sichuan and Tarim Basins, creating broad, gentle margins, and "piles up" behind the basins creating narrow, steep margins. These results imply that this portion of the Eurasian crust was heterogeneous, but largely weak, even prior to construction of the Tibetan Plateau.

Keywords: Tibet, topography, regional tectonics, lower crustal flow, viscosity.

INTRODUCTION

The behavior of the lower continental crust is an integral element of many recent tectonic models of intracontinental deformation, but the basic physical parameters that govern the mechanics of the lower crust remain poorly understood. Many authors have proposed that in regions where the continental crust is hot, the middle or lower crust acts as a weak viscous layer capable of flow on geologic time scales (e.g., Block and Royden, 1990; Wernicke, 1990; Kruse et al., 1991; Bird, 1991; Wdowinski and Axen, 1992). Thus, lower-crustal flow has been proposed as a mechanism by which lateral pressure gradients within the crust are equilibrated, reducing variations in topography and crustal thickness (e.g., Bird, 1991).

Lateral variations in topography, crustal thickness, and style of deformation suggest differences in strain distribution within active orogens and underscore the important role of the rheology of the lower continental crust in governing crustal deformation. Most studies aimed at constraining physical parameters of the lower crust have concentrated on the Basin and Range extensional province of the western United States and have yielded estimates of effective viscosity of the lower crust of 10^{17}–10^{19} Pa·s for assumed channel thicknesses of 10–15 km (e.g., Kruse et al., 1991; Wdowinski and Axen, 1992; Kaufman and Royden, 1994). However, estimates of lower-crustal viscosity beneath active convergent orogens are largely lacking. In this regard, the Tibetan Plateau and its associated mountain belts offer an excellent natural laboratory in which to explore the relationship of tectonics and topography in order to estimate rheologic parameters for the lower crust.

TIBET

The Tibetan Plateau has been created by the continent-continent collision of India with Eurasia since 45 Ma, and is an example of extreme regional topography over scales of hundreds to thousands of kilometers (e.g., Molnar and Tapponnier, 1975; Le Fort, 1975) (Fig. 1). The low-relief, but topographically high, central plateau that has persisted despite continuing Cenozoic shortening has provoked explanations ranging from continental underplating (Barazangi and Ni, 1982) and thermal uplift by delamination of the mantle lithosphere (e.g., England and Houseman, 1989) to deformation within a weak, fluid layer in the middle to lower crust (e.g., Bird, 1991; Zhao and Morgan, 1987).

Rock-mechanics laboratory experiments predict that for moderately high geothermal gradients or for thick crust, the middle or lower crust could contain a weak ductile zone, separating the more competent layers of the brittle upper crust from the rheologically strong upper mantle (e.g., Goetze and Evans, 1979; Brace and Kohlstedt, 1980; Kirby, 1983). Many geophysical observations from Tibet suggest the presence of a weak crustal zone; e.g., short-wavelength gravity anomalies imply compensation of topography within the crust (Jin et al., 1994), and the Project INDEPTH observation of coincident mid-crustal low velocities, high electrical conductivity, and reflection bright spots in Tibet indicate that the middle crust in Tibet contains fluids and may be partially molten (Nelson et al., 1996; Makovsky et al., 1996; Chen et al., 1996).

In contrast to the flat central plateau, most of the marginal mountain belts bordering the plateau are characterized by steep topographic gradients, such as along the southern margin of the Himalaya (e.g., Le Fort, 1975) (Fig. 2A). Previous topographic studies have concentrated on these steep margins of the plateau but have ignored low-gradient margins of the plateau (e.g. Fielding et al., 1994). For example, excluding the areas adjacent to the Sichuan Basin (Fig. 2B), the eastern margin exhibits low topographic gradients with a gradual change in topographic elevation and crustal thickness from the plateau to the outer foreland (Li and Mooney, 1998) (Fig. 2C and 2D).

Unlike the southern plateau margin, where crustal thickening can be explained by shortening of the upper crust through faulting and folding, most of the eastern margin has been uplifted without significant shortening of the upper crust along large-magnitude thrust faults (Burchfiel et al., 1995; Wang et al., 1998). The presence of such long-wavelength, regional topographic gradients in the absence of upper-crustal shortening suggests deformation within the lower crust by ductile flow. Therefore the elevated topography along the eastern plateau margin may be the direct result of thickening of the deep crust in concert with evacuation of the lower crust from beneath the central plateau (Royden, 1996; Royden et al., 1997).

TOPOGRAPHIC PROFILES

Several topographic swath profiles were taken from the northern, eastern, and southern margins of the Tibetan Plateau (Fig. 1). Topographic data are from the publicly available GTOPO30 digital elevation model (DEM) data set, which has ~1 km horizontal resolution (U.S. Geological Survey,

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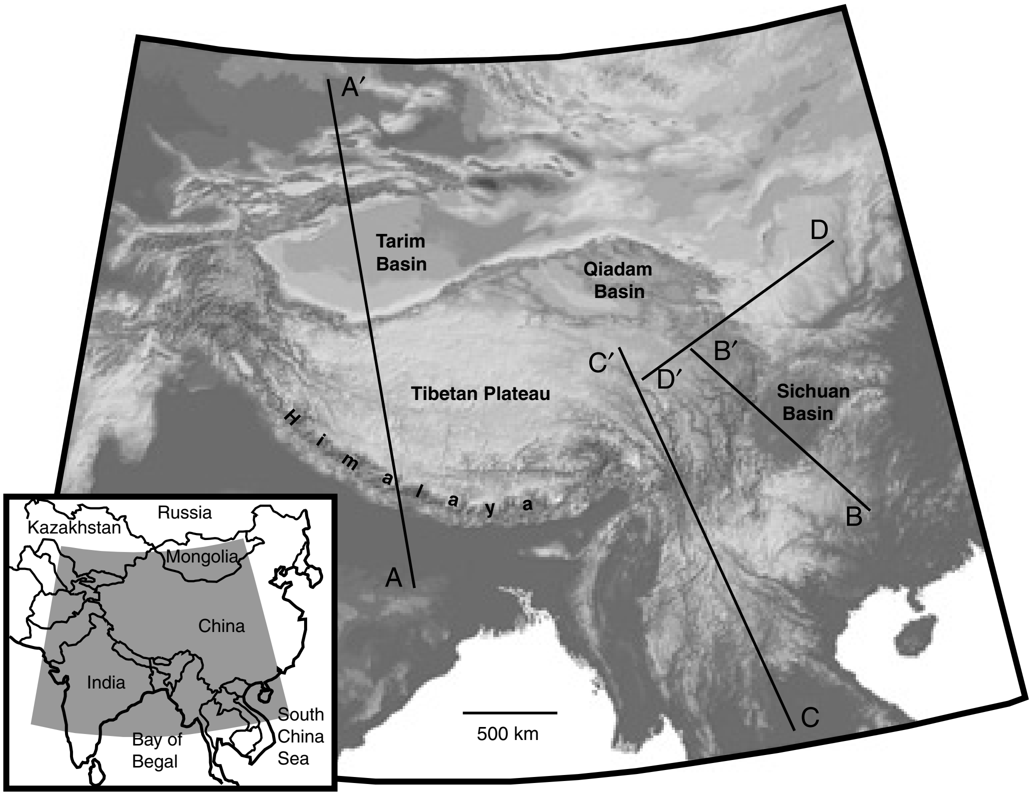

Figure 1. Topographic relief of Tibetan Plateau and surrounding regions. Lines represent axes of swath profiles for different margins of plateau. A–A′, Tibetan Plateau–Tarim Basin profile; B–B′, Sichuan Basin profile; C–C′, southeast margin profile; D–D′, northeast profile.

1993). Swath topography was selected by extracting a narrow rectangular patch of topography with the long axis perpendicular to the plateau margin. The topographic values of the swath are projected onto a vertical plane parallel to the long axis of the swath rectangle, and maximum, mean, and minimum topographic curves are calculated. The width of our swath profiles was 100–300 km, a range that was narrow enough to avoid averaging in the large-scale along-strike variations in geologic structure or crustal rheology. The envelope of maximum topography can be considered a proxy for the current elevated position of the pre-uplift surface, on the basis of previous studies by Wang et al. (1998), DEM analysis, and our own field observations along the eastern plateau margin, where large remnants of an old erosion surface are preserved between major drainage systems.

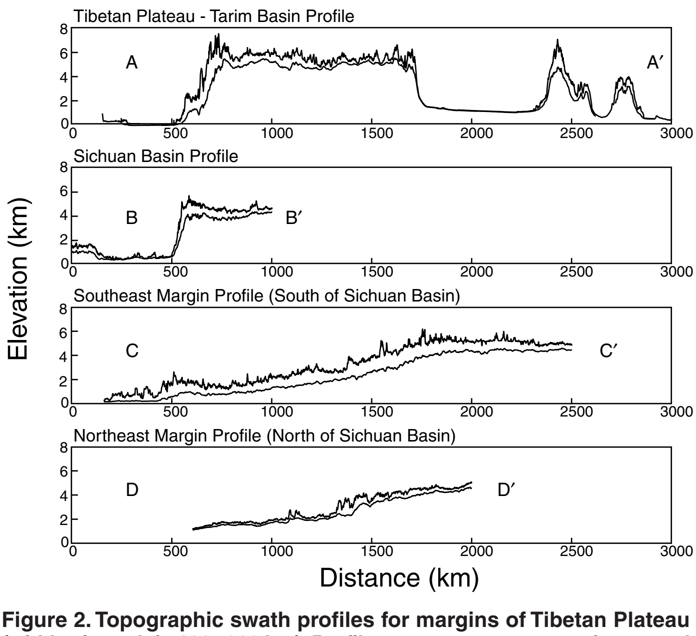

Figure 2. Topographic swath profiles for margins of Tibetan Plateau (width of swath is 100–300 km). Profile curves represent maximum and mean profiles projected onto axis of swath profile. Locations of profiles are shown in Figure 1.

The topographic profiles fall into two end-member categories. The first group, located along the eastern margin but excluding the area adjacent to the Sichuan Basin, shows similar low topographic gradients despite very different climates and erosional histories (Fig. 2C and 2D). Topography along the southeastern margin decreases from 5.2 km to sea level over a distance of ~2500 km (Fig. 2C). Profiles north of the Sichuan Basin decrease from 4.5 km to 1 km over nearly the same distance (Fig. 2D). Steep regional topographic gradients characterize the second group of profiles along the northern (Tarim Basin) and southern (Himalaya) borders of the plateau and along portions of the eastern border (Sichuan Basin) (Fig. 2A and 2B). The drop from plateau height (4.5–5 km) to <1 km elevation over only 50–200 km produces an average regional slope for these profiles that is more than an order of magnitude greater than along the low-gradient margins.

MODEL FOR DUCTILE FLOW IN THE LOWER CRUST

In a simplified model of ductile flow in the lower crust, we consider the region of the lower crust to be a channel of uniform thickness in which crustal material is allowed to flow in response to lateral pressure gradients. In a two-dimensional model we calculate the flux of a Newtonian fluid crust through a channel of thickness (h). For Poiseuille flow with zero velocity at the top and bottom of the channel, the velocity (u) of crustal material in the channel as a function of viscosity (μ), lateral pressure gradient (dp/dx), and depth (z) is

$$u = \frac{1}{2\mu}\frac{dp}{dx}\left(z^2 - hz\right) \tag{1}$$

(e.g., see Turcotte and Schubert, 1982). The flux (U) of material in the channel can be expressed by integrating the velocity of the material over the channel height (h),

$$U = \int_0^h u\left(z\right)dz, \tag{2}$$

and can be related to changes in crustal thickness (c) over time by

$$\frac{dc}{dt} = \frac{-dU}{dx} = \frac{1}{12} h^3 \frac{d}{dx}\left(\frac{1}{\mu}\frac{dp}{dx}\right). \tag{3}$$

We assume that the lateral pressure gradient in the channel is a function of topography only, with pressure given by $p = \rho_c\, g\, T(x)$, where ρ_c is the density of the crust (2600 kg/m^3), g is the acceleration due to gravity, and $T(x)$ is the topographic elevation. By considering only topographic wavelengths that are long compared to the flexural wavelength of the crust (Airy isostatic equilibrium), changes in topographic relief can be linearly related to changes in crustal thickness. Thus, the change in topographic elevations over time as a result of flux of crustal material in the lower crust can be expressed by

$$\frac{dT}{dt} = \frac{(\rho_m - \rho_c)}{\rho_m}\left[\frac{1}{12} h^3 \rho_c\, g \frac{d}{dx}\left(\frac{1}{\mu}\frac{dT}{dx}\right)\right], \tag{4}$$

where ρ_m is mantle density (3300 kg/m^3).

Topography is built by specifying a constant flux of material into the lower crustal channel from beneath the thick part of the plateau (Fig. 3). For the sake of simplicity, we require the channel thickness to be uniform and constant. Excess crustal material that does not participate in flow is accreted to the top and bottom of the channel; this material thickens the crust (Fig. 3). The flux rate into the channel from beneath the central plateau was chosen to allow a plateau margin of ~5 km elevation to develop over 20 m.y. (We chose 20 m.y. as an average time for plateau evolution on the basis of estimates of plateau uplift from geologic data; e.g., Harrison et al., 1992. Although the timing of the uplift is not precisely known, varying the model run time between 10 and 30 m.y. did not significantly alter results.) Crustal channel thicknesses have been estimated to be between 5 and 25 km (e.g., Wernicke, 1990). We chose a channel thickness of 15 km, although this approach provides nonunique results for viscosity because results scale with the variable (h^3/μ) (equation 4).

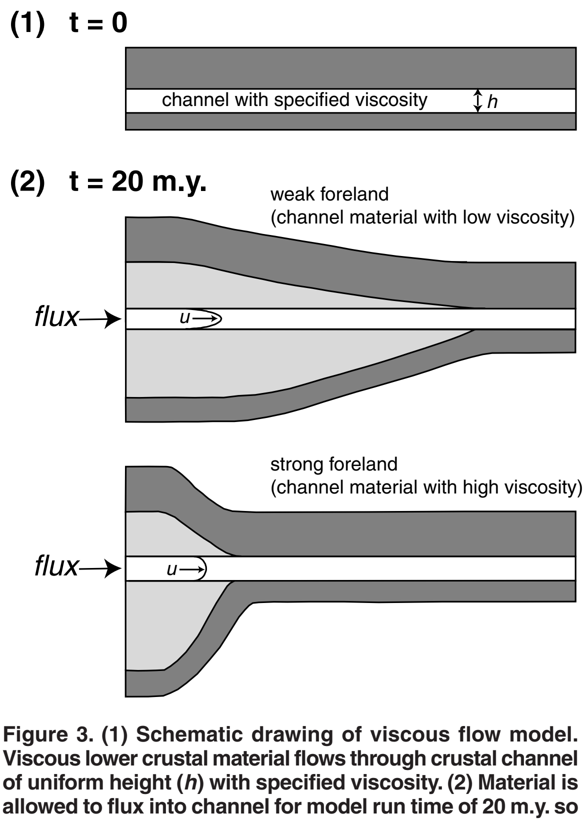

Figure 3. (1) Schematic drawing of viscous flow model. Viscous lower crustal material flows through crustal channel of uniform height (*h*) with specified viscosity. (2) Material is allowed to flux into channel for model run time of 20 m.y. so that elevation of 5 km is reached at edge of plateau. Parabolic channel flow is represented by velocity (*u*), and material that does not participate in channel flow is accreted to top and bottom of channel and thereby thickens crust (light gray areas). Viscosity of channel material in foreland determines regional topographic slope.

RESULTS

Topographic profiles were computed for a range of spatially uniform lower crustal viscosities within a 15-km-thick channel, and were compared with observed topographic swath profiles. Model profile results for the southeastern margin of the plateau yield an excellent fit to the observed topography for a lower crustal viscosity of 10^{18} Pa·s (Fig. 4A); for the northeast profile an acceptable fit is also obtained for a viscosity of 10^{18} Pa·s (Fig. 4B). The topography across the steep margins of the Sichuan Basin (Fig. 4C) and the southern Tarim Basin (Fig. 4D) are fit with a much higher channel viscosity of 10^{21} Pa·s. However, there must be a viscosity contrast between the flat central plateau and the margins of the plateau, because the central plateau must have sufficiently low viscosity so that no significant regional topographic slope can be maintained. Because viscosity estimates vary approximately linearly with topographic slope, we estimate an upper bound on the viscosity beneath the flat central plateau of 10^{16} Pa·s.

If channel viscosity is varied with crustal thickness (i.e., is less under higher parts of the margin), then compared to the uniform case described here the model curves would be steeper at low elevations and flatter at high elevations. This yields a poor fit to the topography of the southeast and northeast margins. The very long wavelength relief that is observed along these margins requires a very low channel viscosity beneath all parts of the margin. Model results yielded the best agreement with observations when viscosity in the channel was uniformly weak at the initiation of flux into the channel, suggesting that the low-gradient margins of Tibet were already weak before the onset of crustal thickening. (Qualitatively similar results would be obtained if a power-law rheology for the channel material were assumed, such as n = 3, although the upward concavity of the profiles would be less pronounced.)

DISCUSSION

The margins of the Tibetan Plateau fall naturally into two end-member categories (Fig. 2). Where the lower crust beneath the margin and adjacent foreland is weak, the lower crust of the central plateau escapes and flows over distances of 1000–2000 km; the result is the lack of a distinct edge to the plateau margin and little shortening deformation in the upper crust. By contrast, where the margins are strong, regional flow of lower-crustal material from the weak central plateau is inhibited, and a steep topographic margin develops. These end-member cases are exemplified by changes in

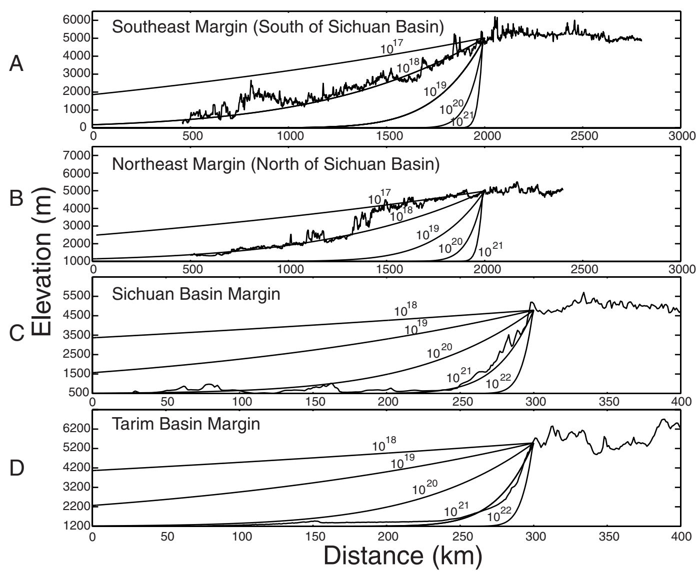

Figure 4. Model results vs. maximum topographic profiles. Model profiles represent runs for uniform lower crustal channel of thickness *h* = 15 km and variable viscosity (labeled in Pa·s).

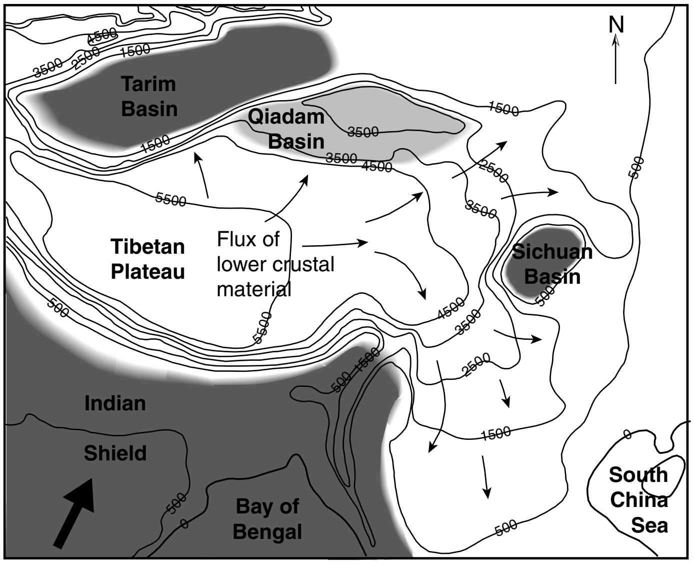

Figure 5. Contour plot of smoothed elevations of Tibetan Plateau and surrounding regions. Contour interval is 1000 m. Areas shaded in dark gray represent regions of cold, strong, continental material; light gray area represents intermediate strength; and white areas represent weak crustal regions. Thus, lower crust escapes from beneath thickened, elevated central plateau through regions where crust is weak.

along-strike morphology of the eastern plateau margin. A smoothed-contour elevation map shows topography flowing or oozing around the Sichuan Basin (Fig. 5), reflecting the strength heterogeneity of the crust along the eastern plateau margin.

The Sichuan Basin is an old, intact craton that has remained relatively undeformed despite orogenic events at its margins in Mesozoic and Cenozoic time (see also England and Houseman [1985] for discussion of crustal heterogeneities). We propose that the strength of the crust beneath the Sichuan Basin inhibits flow of lower crustal material from central Tibet and therefore builds a steep topographic margin that is steeper than that of the Himalayan front, despite the lack of large-magnitude thrust faulting.

Similarly, the Tarim Basin is a relatively strong crustal region that has undergone little internal deformation during Mesozoic and Cenozoic orogenies (e.g., Molnar and Tapponnier, 1981). Crustal material also appears to be flowing eastward around the Tarim Basin through areas of intermediate crustal strength in the Qaidam region and low crustal strength in the northeast corner region of the plateau. This analysis suggests that the greater part of the lower crust of Eurasia surrounding the eastern margin of the plateau was weak prior to building of the Tibetan Plateau.

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