Potential Influences of Middle and Lower Crustal Flow on Landscape Evolution: Insights from the Himalayan-Tibetan Orogen

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1 Introduction

Large exposures of lower crustal rocks in deeply eroded orogenic systems – such as the Scandinavian and East Greenland Caledonides – testify to a complex rheology as greater crustal depths during orogeny. Some tracts seem to have behaved more or less rigidly during mountain building, whereas others – particularly those containing large proportions of anatectic melt rocks (anatexites) – are penetratively deformed, with fabrics and mineral textures indicative of high strains. In regions of the lower crust dominated by such materials, it is easy to image large-scale flow during orogenesis. Several lines of geophysical evidence suggest that the crust beneath orogenic plateaus is sufficiently hydrated to be relatively weak and at high enough temperatures to be partially molten. If the viscosity of this crust is low enough, and if zones of low viscosity are sufficiently interconnected, it is theoretically possible that channels may form that enable this weak crust to flow laterally as a consequence of lateral pressure gradients (Bird, 1991).

Middle or lower crustal flow has been invoked in many models of the tectonic evolution of the Tibetan Plateau; see reviews by Hodges (2006), Klemperer (2006), and Hodges (2016). The concept has been exported to the Andean Plateau (Husson and Sempere, 2003; Gerbault et al., 2005; Ouimet and Cook, 2010) as well as many Precambrian to Cenozoic orogenic systems (e.g., Brown and Gibson, 2006; Brown and Juhlin, 2006; Cagnard et al., 2006; Hatcher and Merschat, 2006; Kuiper et al., 2006; Xypolias and Kokkalas, 2006; Jamieson et al., 2007; Rivers, 2008; Barreiro et al., 2010; Dorr and Zulauf, 2010; Dumond et al., 2010; Gee et al., 2010; Li et al., 2010; Raimondo et al., 2010; Hodges, 2016). Although such models remain controversial, this chapter explores how lateral flow of a fluid middle or lower crust might be manifested in the surface geomorphology of plateau interiors and margins. We focus on the Tibetan Plateau, where most of these ideas were developed, as an analog for other present-day plateaus like the Andean Plateau and plausible older plateaus in the geologic record.

2 Development and geophysical characteristics of the Tibetan Plateau

Developed as a consequence of the Cenozoic collision between India and Eurasia, the Tibetan Plateau is one of Earth's most remarkable physiographic features. It has roughly half the area of the continental USA and a mean elevation above sea level of approximately 5000 m (Figures 1 and 2). Rain gauge reanalysis data (Figure 2b) illustrates the profound precipitation gradients across plateau margins.

The transition is abrupt across the high mountain ranges that mark the southern rim of the plateau and form one of the most dramatic rain shadows on our planet as they block powerful storms coming from the south. The precipitation gradient is far more gradual across the gentler topographic slope of the southeastern margin of the plateau, indicating that summer storms of the Indian monsoon sweep around the eastern end of the Himalayan ranges and are funneled into this region through deep gorges formed as major river systems drain the southeastern plateau (shown as the Three Rivers region in Figure 2c). The unusual arcuate configuration of these gorges, and the fact that several of their traces extend northwestward as topographic lineaments within the plateau, attest to the fact that they follow major transcurrent fault systems (Molnar and Tapponnier, 1975; Tapponnier and Molnar, 1976; Tapponnier et al., 1982).

Continent–continent collision leads to crustal thickening within the collision zone and, if boundary conditions permit, the lateral expulsion of material between the two colliding plates. Both processes feature prominently in models that have been developed to describe Tibetan Plateau evolution. Escape models invoke a special kind of crustal flow, one in which rigid blocks of Tibetan crust slip past one another along a network of transcurrent/strike-slip faults, and the ensemble of blocks experiences macroscopic granular flow (Tapponnier et al., 1982; Peltzer and Tapponnier, 1988; Replumaz and Tapponnier, 2003). Although the net amount of lateral expulsion since collision is highly controversial (Houseman and England, 1993;

Tapponnier et al., 1982), and geodetic data demonstrate that the regions between major transcurrent faults within Tibet do not behave entirely rigidly (Chen et al., 2004; Zhang et al., 2004; Gan et al., 2007), many incarnations of the escape model provide a reasonable explanation of active surface deformation throughout much of Tibet (e.g., Meade, 2007; Thatcher, 2007).

Such models do not, however, fully describe lithospheric deformation in Tibet, particularly the development of an unusually thick crust beneath the plateau (60–80 km; Tseng et al., 2009; Li et al., 2021). England and Houseman (1986, 1988), Houseman and England (1993), and England and Molnar (1997) have demonstrated that continuum mechanical models treating the entire Tibetan lithosphere as a thin viscous sheet explain lithospheric thickening as well as limited expulsion of upper crust. These continuum models presume, for simplicity, a vertically averaged rheology of the lithosphere. However, the crust beneath the Tibetan Plateau is unusually hot compared to most continental regions (Francheteau et al., 1984) and may be unusually felsic (Wang et al., 2021). Moreover, that crust is a collage of compositionally and thus mechanically variable materials accreted to Eurasia during pre-Cenozoic orogenic events (e.g., Bally et al., 1980; Allegre et al., 1984; Dewey et al., 1988; Burtman, 2010). Together these characteristics suggest that the actual rheology of the Tibetan crust is likely to be considerably more complex than vertically averaged models would imply.

Classical models of the strength of the continental lithosphere (Goetze and Evans, 1979; Brace and Kohlstedt, 1980; Kirby, 1983) emphasize the changes in deformation mechanisms that occur with depth as a consequence of increasing temperature and changing lithospheric composition. They feature a brittle upper crust where rock strength is limited by friction and increases with depth with little regard for compositional variation (Byerlee, 1978; Marone, 1998). At deeper levels and higher temperatures, quartz-rich rocks exhibit ductile behavior and become progressively weaker. For a given strain rate, a variety of factors can influence the strength of rocks in this regime. Among the most important are composition (less felsic compositions are stronger: Burgmann and Dresen, 2008), the degree of hydration (wetter rocks are weaker: Kohlstedt, 2007), and the presence and volumetric proportion of anatectic melt (partially molten rocks are weaker: Renner et al., 2000; Rosenberg and Handy, 2005).

Figure 3 illustrates qualitative strength-depth profiles for the crust and uppermost mantle lithosphere beneath southern, and central and northern Tibet. They are based on available geological, geophysical, and geochemical data regarding compositions and thermal structure (cf., Klemperer, 2006). Seismic velocity data from southern Tibet (Wittlinger et al., 2009; Mechie et al., 2011) suggest the existence of a dry, mafic, and thus relatively strong lower crust. The Tibetan middle crust is less mafic and more hydrated and thus should be substantially weaker than the seismically active upper crust or the drier, more mafic lowermost crust. In central and northern Tibet, various seismic data suggest that the middle and lower crust and at least the uppermost mantle are hot and weak (Ni and Barazangi, 1983; McNamara et al., 1995; Wei et al., 2010).

A three-part, vertically stratified rheology of Tibet with a laterally continuous weak zone in the middle or lower crust is a venerable concept in tectonics, dating back at least as far as the work of Bird and Toksoz (1977), who deduced the existence of the weak layer from Rayleigh-wave attenuation observations. After detailed geophysical surveys of southern Tibet began as part of the US National Science Foundation INDEPTH project in the 1990s, more direct evidence for a weak middle crust became available (Nelson et al., 1996). Since then, much seismic and magnetotelluric data have confirmed the presence of a fluid-rich or partially molten middle crustal level beneath the southern half of the plateau, extending in some areas as far south as the crest of the Himalaya (Li et al., 2003; Unsworth et al., 2005; Klemperer, 2006; Ashish et al., 2009; Caldwell et al., 2009; Guo et al., 2009; Huang et al., 2009a; Zhang and Klemperer, 2010). Knowing the effects of fluids and melts on crustal rheology as we do, these observations imply that the middle crust may be sufficiently fluid to sustain channelized lateral flow (Bird, 1991; Nelson et al., 1996). The available geophysical data for northern and central Tibet (e.g., Klemperer, 2006) seem to indicate that much of the middle and lower crust there may have sufficiently low viscosity to support wholesale lateral flow rather than narrowly channelized flow in the middle crust (e.g., Bendick and Flesch, 2007). However, in southern Tibet, seismic low-velocity zones in the middle and lower crust – thought to be indicative of partial melts – are discontinuous, such that flow may be more localized approaching the Himalaya (Hetényi et al., 2011).

3 Gravitational potential energy gradients and the dynamics of middle crustal flow

One useful context for thinking about the tectonic and geomorphic implications of such flow derives from an analysis of lateral gradients in gravitational potential energy within and around the Tibetan Plateau (Hodges et al., 2001). The gravitational potential energy per unit area of a column of the lithosphere (*GPE*) can be defined as the integrated vertical stress due to the body force:

$$GPE = -\int_{-h}^{L}(L-z)\rho(z)g\,dz \qquad [1]$$

where *L* is lithosphere thickness, *h* is surface elevation, *z* is depth, $\rho(z)$ is density as a function of depth, and *g* is gravitational acceleration (e.g., Jones et al., 1996; Flesch et al., 2007). Lithospheric thickening in orogenic systems results in a change in gravitational potential energy of the system (GPE_0) relative to the surrounding, thinner lithosphere (Artyushkov, 1973). Thickening of the crust alone leads to an increase in gravitational potential energy, although concomitant thickening of the denser mantle lithosphere counterbalances this effect (Fleitout and Froidevaux, 1982). The convective removal (Molnar et al., 1993) – or delamination (Bird, 1991) – of a dense lithospheric root at some point in the system's evolution would result in a dramatic overall increase in the difference between the gravitational potential energy of the system and its surroundings. We can define this difference (ΔGPE) as:

$$\Delta GPE = GPE_0 - GPE_S \qquad [2]$$

where GPE_S represents the gravitational potential energy of the adjacent lithosphere. Large values of ΔGPE result in large differential stresses that are entirely due to lithospheric thickness variations; these stresses contribute significantly to the dynamics of continental deformation in orogenic systems (e.g., Molnar and Lyon-Caen, 1988; England and Jackson, 1989; Jones et al., 1996; Humphreys and Coblentz, 2007). The *GPE* of continental plateaus is characteristically high, and the *GPE* of Tibet is the highest on Earth. As a consequence, Tibet exerts a very large deviatoric stress on its surroundings, estimated at ~ 5–6 TN m^{-1} (6 x 10^{12} N m^{-1}) by Molnar and Lyon-Caen (1988) and Copley et al. (2010), or at a more modest ~2.5 TN m^{-1} by Ghosh et al. (2006). Either way, these values are of comparable magnitude to – or larger than – stresses exerted by processes such as ridge push, slab pull, or tractional flow in the asthenosphere per unit length along plate boundaries (Thatcher, 2009).

Within major orogenic plateaus, deviatoric stresses induced by ΔGPE are tensional (Artyushkov, 1973; Fleitout and Froidevaux, 1982), providing an explanation for why the active surface deformation of plateau interiors is frequently dominated by extensional and transcurrent faults rather than thrust faults (Dalmayrac and Molnar, 1981; Dewey, 1988; Molnar and Lyon-Caen, 1989). However, another phenomenon may also help smooth out gravitational potential energy anomalies beneath plateaus: the lateral flow of crust from areas with high *GPE* to areas with low *GPE*. One can think of the Himalayan–Tibetan orogenic system as a vast reservoir of excess *GPE*, built by India–Eurasia collision, that is, dissipating by the redistribution of mass outward from the plateau through a combination of brittle faulting in the Tibetan upper crust and lateral flow in the Tibetan middle crust (Westaway, 1995; Royden et al., 1997; Hodges et al., 2001; Shen et al., 2001; Royden et al., 2008). The flux of material outward from the plateau (*v*) can be represented as:

$$v = -\kappa(\Delta GPE) \qquad [3]$$

where κ is a proportionality constant. From this equation – analogous to Fourier's, Darcy's, and Fick's laws for heat flow, fluid flow, and chemical diffusion, respectively – it follows that the directionality of flow is influenced strongly by the gradient in *GPE* in the absence of resistance. At present, the highest *GPE* gradients in the Tibetan realm are along the southern margin of the plateau, which may explain why *GPE*-related deviatoric stresses within the plateau are N–S tensional (Ghosh et al., 2006). Hodges et al. (2001) noted that although continued N–S convergence across the Himalaya exerts a resistance against wholesale southward spreading of the plateau, a coupling of middle crustal extrusion with aggressive erosion along the Himalayan range front could provide an extremely effective outlet for energy dissipation.

Within the Tibetan Plateau, flow of material in a fluid middle crustal channel could redistribute mass laterally as a consequence of horizontal pressure gradients imposed by topographic gradients above the channel (Bird, 1991). Figure 4(a) illustrates the essential characteristics of such a channel that tunnels laterally through the crust. The major geomorphic expression of the laterally propagating tip of such a channel should be a propagating topographic front. For more detailed discussions of this process, readers are directed to Clark and Royden (2000), Clark et al. (2005a), Grujic (2006), and Medvedev and Beaumont (2006). Some workers have suggested that laterally propagating channels may emerge – or extrude (Figure 4(b)) – at plateau margins under specific tectonic and climatic conditions (e.g., Hodges et al., 2001; Beaumont et al., 2001; Hodges, 2006). In Section 5.15.4, we concentrate on how the marginal and interior landscapes of the Tibetan Plateau might relate to middle crustal flow, brittle deformation in the upper crust, and climatic processes.

4 Geomorphology and tectonics of the Tibetan Plateau

Even before the Cenozoic collision of India and Eurasia that built the Himalayan–Tibetan orogenic system, the tectonic history of Tibet had been long and complex. The bedrock geology of the plateau (Figure 5) tells the story of successive accretion of tectonostratigraphic blocks to a growing Eurasia over the Triassic–Eocene interval (e.g., Bally et al., 1980; Allegre et al., 1984; Dewey et al., 1988; Burtman, 2010; Kapp and DeCelles, 2019). However, the geomorphology of the plateau is largely reflective of Miocene-Recent tectonics after India–Eurasia collision. To best appreciate these relationships, it is useful to think of the physiography of Tibet in terms of long-wavelength (>10^2-km scale) and short-wavelength (<10^2-km scale) topographic variations. In Sections 4.1–4.3, we use a series of topographic profiles across the plateau (Figure 6) and more detailed profiles across select plateau margins (Figure 7) to describe salient features. The cross-plateau profiles in Figure 6 represent averages and ranges of 100-km-wide topographic swaths (Figure 2a) extracted from the Global 30 arcsec digital elevation dataset (GTOPO30), which has an effective resolution slightly better than 1 km px^{-1}. The more detailed plateau margin profiles in Figure 7 are based on Shuttle Radar Topography Mission (SRTM) data, with a resolution of slightly better than 90 m px^{-1}.

4.1 Long-wavelength topographic variations, south to north

High-resolution satellite images and DEMs of Tibet (Figures 1 and 2) reveal a land of low relief (10s to a few 100 m over a 5-km distance) surrounded by margins of extreme relief (1000s of meters over a 5-km distance) (Fielding et al., 1994). The Himalayan ranges, which correspond geographically to the southern margin of the plateau, harbor our planet's highest mountains and most dramatic relief. The foothills of the Himalaya (physiographic Lower Himalaya) rise from near sea level in the Indo-Gangetic plains in the south to typical maximum elevations of ~2000 m across a physiographic transition (left margin of Transect A–A') that marks the approximate surface trace of the Himalayan Sole thrust (HST), the active decollement along which India is subducted northward beneath Tibet (e.g., Seeber et al., 1981; Yeats and Lillie, 1991). However, an even more profound physiographic transition, marked by major knickpoints on rivers that flow southward off the plateau, commonly separates the foothills region from the physiographic Higher Himalaya (Hodges et al., 2001; Wobus et al., 2006). Although some researchers regard it as marking the trace of an N-dipping, surface-breaking, out-of-sequence thrust fault (Wobus et al., 2003, 2005; Hodges et al., 2004; Whipple et al., 2016), others regard it as the surface expression of the leading edge of a blind duplex system developed above a structural ramp in the HST (Cattin and Avouac, 2000; Bollinger et al., 2006; Herman et al., 2010; Elliott et al., 2016). These researchers nonetheless agree that deformation at this physiographic transition, whether at the surface or slightly deeper in the upper crust, has resulted in a major discontinuity in the regional rock and surface uplift patterns in Plio-Pleistocene times. To the north, the steep front of the Higher Himalaya spatially coincides with: (1) rapid denudation on the million-year and millennial timescales as evidenced by thermochronologic and terrestrial cosmogenic radionuclide data (e.g., Burbank et al., 2003; Wobus et al., 2003, 2005; Bojar et al., 2005; Blythe et al., 2007; Adams et al., 2009; Dortch et al., 2011; Godard et al., 2014; Morell et al., 2017); as well as (2) rapid surface uplift as measured by geodesy (Jackson and Bilham, 1994). Importantly, this front also bears the brunt of the South Asian monsoon in the Himalaya (Figure 2b; Barros et al., 2000; Barros and Lang, 2003; Burbank et al., 2003; Putkonen, 2004; Bookhagen and Burbank, 2006) and consequently supplies much of the eroded material carried southward toward the plains by major river systems draining off the southern plateau (e.g., Garzanti et al., 2007; Gabet et al., 2008).

A third important physiographic transition marks the crest of the steep Higher Himalayan front and the literal edge of the plateau. The position of this transition, which we refer to as PT1 (following Hodges et al., 2001), is commonly marked by a prominent knickpoint on rivers flowing off the plateau and across the Himalayan range front. Hodges et al. (2001) noted that the PT1 – as well as the physiographic transition at the foot of the Higher Himalaya (PT2) and at the foot of the Lower Himalaya (PT3) – correlates with fault traces in at least some well-studied transects through the central part of the range (e.g., Transect E–E’, Figure 7). We will return to the possible significance of this correlation later on. It is noteworthy that the topographic profile across the Himalayan front changes significantly along strike. In Bhutan (e.g., Transect C–C’, Figure 7), the physiographic Lesser Himalaya are missing, and instead of a well-developed PT2 in the hinterland, there is a pronounced topographic bench. In the NW Himalaya (e.g., Transect F–F’, Figure 7), PT2 is almost indistinguishable. Such variations likely represent along-strike variations in the topology of blind and surface-breaking fault systems (Duncan et al., 2003; Baillie and Norbu, 2004; Adams et al., 2013; 2015; 2016).

Despite the impressive height of the major Himalayan peaks (including most of Earth’s landscape above 8000 m above sea level [asl]), the mean elevation of the Himalayan crest is only slightly higher than the mean elevation of the Tibetan Plateau as a whole. Transect A–A’ shows a mean elevation of just under 5000 m asl, with a total relief of only approximately 1000 m between the Himalayan crest and the Qaidam Basin. Closer inspection of the profile and the digital elevation model in Figure 1 reveals two broad, E–W–trending topographic highs within the plateau between ~80° and 95° N, in addition to the Himalayan crest at the plateau’s southern margin and the Altyn Tagh–Qilian mountain complex along the northern margin.

From the Himalayan crest, the mean elevation decreases northward from ~5000 to 4000 m over roughly 200 km. A narrow topographic low occurs in the E–W-trending valley of the upper Yarlung Tsangpo River, which eventually flows eastward, then southward, off the plateau to empty into the Bay of Bengal as the Brahmaputra River. In southern Tibet, the river course generally follows the trace of the Eocene Indus–Tsangpo suture (ITS), a feature that marks the surface boundary between geology of the Indian realm to the south and the Eurasian realm to the north (Bally et al., 1980; Shackelton, 1981; Tapponnier et al., 1981). The high region immediately north of the Yarlung Tsangpo valley corresponds to the Transhimalayan ranges, which are largely underlain by magmatic rocks of a Cretaceous-Paleocene continental arc that marked the southern margin of Eurasia to India–Eurasia collision (Dewey et al., 1988). This Transhimalayan region – or what we will refer to in this chapter as the southern highlands of Tibet – broadly corresponds to the Lhasa tectonostratigraphic terrane (Figure 5). It has average elevations of 5000–5500 m asl and isolated peak elevations that rise to more than 7000 m asl (not evident on Transect A–A'). Farther north, is another subtle, E–W-trending topographic low that broadly corresponds to the surface trace of the Late Jurassic–Early Cretaceous Banggong-Nujiang suture (Girardeau et al., 1984; Dewey et al., 1988).

North of the Banggong-Nujiang suture lies a northern highlands region underlain by predominately Paleozoic– Mesozoic metamorphic and sedimentary rocks of the Qiangtang tectonostratigraphic block (e.g., Allegre et al., 1984; Dewey et al., 1988; Yin and Harrison, 2000; Kapp et al., 2005). The southern and northern highlands and the intervening topographic lowland collectively represent what is normally regarded as the central Tibetan Plateau. The vast area north of the drainage divide of the southern highland is internally drained (Figure 2c) and remarkably flat, with local relief typically measured in the hundreds of meters or less (Liu-Zeng et al., 2008; Dong et al., 2010).

This broad region transitions northward to elevations of ~3000 m asl across a narrow zone with high local relief corresponding to the Eastern Kunlun Shan. A second zone of high relief, the Altyn Tagh, separates the Qaidam Basin from lowlands to the north. The Altyn Tagh, Western Kunlun Shan, and Qilian Shan collectively mark the structurally complex northern margin region of the Tibetan Plateau, where active major left-lateral strike-slip structures like the Altyn Tagh and Kunlun faults are kinematically linked with numerous N-W striking and both SW- and NE-vergent thrust faults (Burchfiel et al., 1989; Duvall and Clark, 2010; Harkins et al., 2010). Like the Himalayan realm along the southern plateau margin, the northern margin of Tibet apparently became the locus of contractional deformation at the time of India–Eurasia collision in Eocene time (Clark et al., 2010; Yin, 2010); as noted in the first of those papers, the structural reactivation of the area in Miocene–Recent times implies a relatively stationary position of the northern plateau margin for the past 45 Ma.

4.1.1 Are E–W corrugations related to middle-lower crustal flow?

Although subtle, the broad E–W corrugation in the Tibetan Plateau – comprising the southern and northern highlands and the depression between them – has been interpreted in different ways. Jin et al. (1994) suggested that these might represent large-scale buckling of the Tibetan lithosphere. Bendick et al. (2008) presented a model in which the corrugations represent long-wavelength flexures of an elastic upper crust above a fluid lower crust that is being forced northward by a rigid Indian indenter. They based this hypothesis on a model of northern Tibet in which there is no single, middle crustal channel of low viscosity crust but instead a middle and lower crust that is uniformly fluid (Bendick and Flesch, 2007). In contrast, Pullen et al. (2011) interpreted the N–S, long-wavelength topographic variation as inherited from older tectonic patterns. They noted spatial correlations among the northern physiographic highlands, a Mesozoic structural antiform in the Qiangtang tectonostratigraphic block, and crustal-scale geophysical features. However, the extremely weak lower crust beneath northern Tibet seems inconsistent with the explicit suggestion by Pullen et al. (2011) that the Mesozoic surface structural architecture persists downward into the deep crust. Taylor et al. (2021) recently suggested that active thrust duplexing plays an important role in the uplift of the southern highlands of Tibet. Alternatively, tunneling of weak lower crust toward the south from central Tibet, driven by ΔGPE, could also drive active uplift of the southern highlands.

4.1.2 Has Tibetan middle crust extruded at the Himalayan front?

High erosion rates along the southern flank of the plateau, coupled with the kinematic characteristics of major north-dipping fault systems in the Himalaya, led Nelson et al. (1996) to propose a tectonic model for the Miocene evolution of this part of the orogenic system that involves the viscous flow of middle crust southward from beneath southern Tibet and removal of that material from the orogenic system by monsoon-driven erosion at the Himalayan front (Figure 4(b)). The viability of this channel extrusion hypothesis hinges on the correctness of two assumptions. The first is that there was a fluid middle crust in southern Tibet during the Miocene, similar to the one inferred to exist at present based on geophysical data. Proponents of the southward extrusion model of Nelson et al. (1996) point to the metamorphic core of the Himalaya – commonly referred to as the Greater Himalayan Sequence (Hodges, 2000) – as an excellent analogue for what the modern Tibetan middle crust might look like: high-grade (amphibolite to granulite facies) metamorphic rocks that contain abundant Miocene granites resulting from widespread anatectic melting. The outcrop belt for the north-dipping Greater Himalayan Sequence conforms to the Himalayan range front between PT1 and PT2 (Figures 6 and 7). The second key assumption has to do with the nature of the upper and lower contacts of the Greater Himalayan Sequence. If the sequence represents an emergent channel, it should be bound at the top and bottom by coeval, N-dipping fault zones with opposite vergence: S-directed thrusting at the base and N-directed extension (from a kinematic perspective) at the top. Evidence for Miocene, large-displacement (order 10^2 km) thrusting at the appropriate structural level has existed for decades (e.g., Le Fort, 1975), but the recognition of normal faulting near the Himalayan crest has come substantially more recently and not without some controversy.

Chinese and French geoscientists can be credited with the first recognition of N-dipping, low-angle, normal-sense detachments in the physiographic High Himalaya (Caby et al., 1983; Burg et al., 1984), but the first suggestion that these structures may be part of a major fault system all along the crest of the Himalayan arc can be traced to the work of Burchfiel and Royden (1985). Speculating that this system was broadly coeval with major south-directed thrust systems at deeper structural levels – the Miocene Main Central Thrust system or MCTS and Pliocene Main Boundary Thrust system or MBTS (Hodges, 2000) – Burchfiel and Royden hypothesized that steep Miocene–Pliocene topographic gradients resulted in a stress field consistent with synchronous slip on shallowly dipping thrust and normal faults in a continent-continent collisional setting, effectively accommodating southward expulsion of material between the contractional and extensional fault systems toward the orogenic foreland.

Subsequent research by an international community of independent teams confirmed the existence of a South Tibetan fault system (STFS) – comprising shallowly and steeply N-dipping normal faults and oblique transcurrent faults – near the range crest from as far east as Bhutan and as far west as NW India (as reviewed by Burchfiel et al., 1992; Hodges, 2000; and Kellett et al., 2018). Although the STFS appears to have initiated slip in the Middle–Late Miocene in many areas (e.g., Hodges et al., 1992, 1996; Edwards and Harrison, 1997; Searle et al., 1997; Wu et al., 1998; Walker et al., 1999), an ever-increasing body of evidence points to the existence of multiple detachments of multiple ages at the structural level of the STFS (Burchfiel et al., 1992; Hodges et al., 1996; Kellett et al., 2009; Searle, 2010; Cooper et al., 2012; 2013; 2015), some of them as young as Pleistocene (Hurtado et al., 2001; McDermont et al., 2013; 2015). Inasmuch as southward thrust faulting has been continuous along the Himalayan front since at least the Miocene (see Hodges, 2000 for a review), evidence for normal faulting near the Himalayan crest over most – if not all – of the same interval led Hodges et al. (2001) and Hodges (2006) to speculate that the southward extrusion of south Tibetan middle crust has been continual in the Himalaya for the past 20 Ma. Beaumont et al. (2001) presented elegant thermomechanical models for how the surfacing of a middle crustal channel could be a natural consequence of rapid, monsoon-driven erosion along the Himalayan front, and Hodges et al. (2001) noted how this coupling of tectonics and erosion provides an extremely efficient mechanism for dissipation of the excess gravitational potential energy from the orogenic system.

The channel extrusion hypothesis requires that both the basal thrust fault system and the capping normal fault system binding the Greater Himalayan channel have displacements measured in the tens or hundreds of kilometers. Although virtually all Himalayan researchers agree that this is true for major thrust structures of the MCT and MBT systems, not everyone agrees that the South Tibetan fault system is this significant. Geologic relationships described by Herren (1987), Burchfiel et al. (1992), Dezes et al. (1999), Long and McQuarrie (2010), and Searle (2010) from Bhutan to NW India seem to require at least 20 km of slip along the South Tibetan fault system detachments and, in some cases, much more (Cooper et al,. 2012). The notion of very large-scale displacements is supported by a growing body of petrologic, geochronologic, and thermochronologic data demonstrating dramatic differences in the late Cenozoic thermal history of the Greater Himalayan sequence and strata in the hanging wall of the STFS (e.g., Searle et al., 2003; Jessup et al., 2008; Cottle et al., 2011; Cooper et al., 2012; Schultz et al., 2017). However, a few authors (e.g., Yin, 2006; Sachan et al., 2010) regard the extensional South Tibetan fault system as a relatively minor feature of limited significance.

Additional arguments have been marshaled against the channel extrusion hypothesis by several researchers (e.g., Harrison, 2006; Kohn, 2008; Yin et al., 2010). Some reflect a concern that Greater Himalayan Sequence rocks are of Indian, rather than Eurasian, affinity. If the Greater Himalayan sequence is indeed the fossil remains of a Miocene channel flowing from southern Tibet, why are the rocks not similar to those exposed at the surface in the Transhimalayan ranges? Channel extrusion advocates counter that the channel in southern Tibet – as opposed to central and northern Tibet, north of the Indus–Tsangpo suture – may be Indian plate materials accreted across the HST during continental subduction at the time of India–Eurasia collision (Hodges, 2006; Harris, 2007). Another criticism (Kohn, 2008) is that the pressure–temperature–time history of the Greater Himalayan rocks is different from that predicted by published thermomechanical models of channel extrusion (e.g., Jamieson et al., 2004), but such models include so many explicit input parameters that the inconsistency of any specific model result with geologic observations is just as likely a reflection of parameter choice and simplifying assumptions as it is an indication of an invalid conceptual model. Still another expressed concern is that the model requires coupled deformation and monsoon-driven erosion, but erosion is focused on the Himalayan front, whereas the Greater Himalayan Sequence in some places occurs in klippen far south of the area of highest rainfall (Harrison, 2006). However, recent analysis of precipitation patterns shows a broader N–S distribution of heavy rainfall than previously inferred (e.g., Bookhagen and Burbank, 2010). Besides, there is no reason to assume that the latitude of heavy precipitation relative to the range front has not migrated substantially since the Miocene, and much reason to believe that some of the klippen are related to complex duplexing or out-of-sequence thrusting after initial emplacement of the Greater Himalayan Sequence in the MCT system hanging wall (Robinson et al., 2003, 2006; Pearson and DeCelles, 2005). Although southward channel extrusion is far from demonstrated unequivocally, none of the alternative models thus far proposed to explain that the Miocene–Recent structural architecture of the Himalaya are as consistent with the preponderance of geologic evidence.

4.2 Long-wavelength topographic variations, West to East

A second topographic swath shown in Figure 7 (Transect B–B') begins in the Hindu Kush, traverses eastward through the eastern Himalayan syntaxis and the northern highlands region of Tibet, and ends just east of the Longmen Shan. Unlike Tibet's southern and northern margins, the western margin is diffuse, as the high plateau dies out gradually into the high relief mountainous landscape of the Western Kunlun Shan and Karakoram. The prominent low in the mean elevation of this swath, ~200 km from its western edge, corresponds to the dramatic valley of the Indus River as it cuts through the Karakoram. The high dynamic range of the profile between 200 and 1000 km east of its western edge represents the rugged topography of the western plateau in the structurally complex region where the active Karakoram, Altyn Tagh, and Karakax fault systems converge (Burtman and Molnar, 1993; Robinson, 2009; Burtman, 2010). High-frequency relief diminishes substantially over the northern highlands of the plateau but then increases again as the mean elevation begins a gradual decrease from ~ 5000 m asl to slightly less than 4000 m asl before dropping precipitously at the Longmen Shan escarpment.

4.2.1 Does the gradual West-to-East decrease in mean elevation indicate eastward flow of the middle crust?

One of the most important geomorphic observations in the eastern Tibetan Plateau is that the region displays an elevated, low-relief landscape (Figure 2 and 7) that has been heavily dissected by river systems (e.g., Kirby et al., 2002; Clark et al., 2004; Schoenbohm et al., 2004; Clark et al., 2006). This landscape, with a relief of no more than a few hundred meters, is generally marked by bedrock erosional surfaces which are sometimes covered by a thin veneer of Tertiary or Quaternary sediments (Clark et al., 2006). On Transect B-B', the elevations of this landscape range from near 5000 m asl in the west to about 4500 m asl in the east, but regions to the north and south of the Longmen Shan show a greater variation in elevation. This is demonstrated by Transect J-J' (Figure 7), in which the low-relief landscape elevation is comparable to the profile mean and drops from ~4000 to ~ 2000 m asl over a distance of about 150 km. In fact, the southeast plateau margin is everywhere marked by a very gradual decline in mean elevation, although the local relief is high because the plateau gradient is parallel to the flow directions of deeply incised river systems (e.g., the Mekong, Salween, and Yangtze) which follow Cenozoic strike-slip faults as they drain off the plateau to the Bay of Bengal and South China Sea (Clark et al., 2004; Ouimet et al., 2010).

Clark and Royden (2000), Clark et al. (2004), Schoenbohm et al. (2006), and Royden et al. (2008) employed this gradual slope in the low-relief surface to argue in favor of a channel tunneling model (Figure 4a) in which the eastward flow of a fluid Tibetan middle crust – driven by pressure gradients imposed by overlying crust of varying thickness – helped accommodate lateral growth of the plateau. The channel tunneling model is generally supported by the teleseismic receiver function analyses of Xu et al. (2007) for SE Tibet that imply a mid-crustal low velocity zone comparable to that in southern Tibet. In the original model, which diverges markedly from classical escape models for Tibet (e.g., Tapponnier et al., 2001; Replumaz and Tapponnier, 2003), the major surface-breaking strike-slip fault systems in this region are inferred to be restricted to the elastic upper crust, although one could imagine a kinematic linkage (cf., Schoenbohm et al., 2006); some of the larger fault systems may continue as steep shear zones to depth and segment the mid-crustal fluid channel without disrupting its conductivity. Bai et al. (2010) presented magnetotelluric data indicative of high conductivity zones in the middle crust of eastern Tibet, south of the Longmen Shan and adjacent Sichuan Basin. They were interpreted as regions of crust weakened by high fluid contents or partial melts. Suppose these zones represent the lower crustal flow channels as envisioned by Clark and Royden (2000). In that case, such features appear to be laterally discontinuous and possibly bound by the down-dip projections of major transcurrent structures such as the Red River and Xiaojiang faults. A consistent result was obtained by Huang et al. (2009b) based on the seismic velocity structure of the region.

Clark and Royden (2000) also suggested that the steep Longmen Shan escarpment, west of the Sichuan Basin (Transect I-I', Figure 7), as well as the steep northern boundaries of the plateau against the Tarim and Qaidam basins (Transects G-G' and H-H', Figure 7) are evidence for thick, rigid crust beneath these basins serving as blockades that divert the lateral flow of fluid Tibetan crust to the region between the Qaidam and Sichuan basins and the SE between the eastern Himalayan syntaxis and the Sichuan Basin (Figure 1). This model was further elaborated by Clark et al. (2005a) and Cook and Royden (2008). Stüwe et al. (2008) and Dayem et al. (2009a; 2009b) explored how strength discontinuities between Tibet and the regions to the north could have helped concentrate strain on major thrust and transcurrent fault systems.

Just as the channel expulsion model is not accepted by all researchers working in Tibet, neither is the channel tunneling model first proposed by Clark and Royden (2000). For example, while Liu-Zeng et al. (2008) recognized the relict low-relief landscape within the eastern plateau, they questioned whether or not it predated plateau uplift and could be used as a passive marker to construct an argument of the eastward flow of the mid-crustal channel. Yang et al. (2015) also presented an alternative formation mechanism for the low-relief surfaces that do not require changes in vertical rock uplift rates that would be created by tunneling. They suggest that a change in lateral rock advection perpendicular to the direction of proposed tunneling could perturb river channel topology, which could, in turn, cause river capture events resulting in the high-elevation, low-relief surfaces similar to what is observed at the south-eastern border of the Tibetan Plateau. The stream-capture hypothesis was challenged by Whipple et al. (2017), who agreed that the stream-capture mechanism could produce high-elevation, low-relief surfaces in principle, but in practice, these surfaces would not look like those at the south-eastern edge of the Tibetan Plateau. Whipple et al. (2017) state that the stream-capture hypothesis cannot explain the co-planar nature of the dissected surfaces with a relatively consistent dip and dip-direction or the specific geometries of the headwaters of these landscapes. Fox and Carter (2020) revisited these hypotheses using a model of river channel inversion. They found that a smooth function of rock uplift and surface uplift rates can only predict ~55% of the elevation variability observed over the region defined by low-relief surfaces. These findings do not invalidate the tunneling hypothesis but suggest that a more complicated model that includes other complicating factors such as faulting, climate change over time, spatial changes in erosional efficiency, or stream piracy to explain the temporal evolution of elevations if tunneling had occurred.

A variety of authors (Holt, 2000; Flesch et al., 2005; Sol et al., 2007) have argued for strong coupling between the upper crustal and mantle strain fields beneath Tibet, which can be interpreted to be inconsistent with upper crustal decoupling from a weak middle crust, but may instead simply imply a similarity of farfield stresses that produce compatible deformation fields in the upper crust and mantle (cf., Klemperer, 2006). Copley et al. (2011) presented numerical experiments that assume a strong coupling between the down going Indian plate and the Tibetan lower crust and can reproduce the observed deformation field in southern and southeastern Tibet without appealing to channel flow. Of course, one viable model does not preclude the validity of another, in this case, either channel tunneling or channel extrusion. Thermomechanical modeling by Cook and Royden (2008) demonstrated how a fluid lower crust can develop in front of an on-coming rigid indenter and how lateral flow can be induced. However, using a different approach to thermomechanical modeling, Godard et al. (2009) showed how escarpments such as that of the Longmen Shan could be produced by wholesale gravitational flow of overthickened crust, enabled by focused erosion at the plateau margin, without the existence of a discrete channel of low-viscosity middle crust. Rey et al. (2010) argued that the extent of eastward channel flow invoked by Clark and Royden (2000) and others – more than 1000 km – seemed improbable, if not impossible. At the same time, Rey et al. (2010) seemed to accept the likelihood of southward extrusion. Suffice it to say that the mode and extent of eastward middle crustal channeling beneath southern Tibet shall remain controversial for some time.

4.3 Short-wavelength topographic variations on the Tibetan Plateau

Superimposed on the long-wavelength topographic variations across the plateau are short-wavelength variations related to roughly N–S-striking normal faults and kinematically linked, conjugate transcurrent faults (striking NW–SE or NE–SW) that collectively accommodate E–W extension of the plateau interior (Armijo et al., 1986; Taylor and Yin, 2009). The result is several small and discontinuous rift basins scattered across the central and southern plateau that is flanked by local uplifts with isolated peak elevations up to more than 7000 m (Figures 1 and 2). The highest concentration of these rift basins is in plateau regions with mean elevations of more than 5000 m asl. Although most of the related structures are presently active, their initiation ages are largely unknown. An increasing body of evidence suggests, however, that the extensional regime that now characterizes most of the central and southern plateau interior was established by at least 14–13.5 Ma (e.g., Coleman and Hodges, 1995; Blisniuk et al., 2001) and may be even older (Larson et al., 2020). Other than in some rift-flank uplifts, low-temperature thermochronology provides little evidence of significant erosional or tectonic rock exhumation in the central Tibetan interior after the Eocene (e.g., Wang et al., 2007).

The depth to which the extensional regime of the upper crust continues down into the lower crust is limited; most earthquakes associated with E–W rifts in the plateau interior have shallow (~18 km) hypocenters, and the middle and lower crust are aseismic. Masek et al. (1994) demonstrated how the morphology of rift-flank uplifts in Tibet implies a low flexural rigidity for the elastic upper crust, with isostatic compensation taking place deeper in the crust. Modeling by Jimenez-Munt and Platt (2006) supported the argument that only a relatively weak middle-lower crust could explain the observed E–W extension in the Tibetan upper crust given overall convergence between India and Eurasia.

In the 1980s, researchers such as Armijo et al. (1986), Mercier et al. (1987), and Peltzer and Tapponnier (1988) interpreted the E–W extension of the Tibetan interior as a consequence of escape tectonics: Material in the plateau extended and was displaced as rigid blocks eastward toward China. Subsequent authors (e.g., McCaffrey and Nabelek, 1998; Yin et al., 1999; Kapp and Guynn, 2004) appealed to longitudinal variations in the collisional boundary conditions to relate rifting in Tibet with the impingement of India. Another study of active extension in the plateau (Elliott et al., 2010) supported another long-standing hypothesis: that most of the extension in the high plateau has been driven by *GPE* variations (e.g., Molnar et al., 1993). A reasonable extension of this hypothesis is that the eastward flow of a fluid middle or lower crust beneath an elastic upper crust would result in tractional forces compatible with E–W upper crustal extension.

5. A self-consistent model of the Cenozoic topographic evolution of the Tibetan Plateau, assuming lower and middle crustal flow

Although the geoscience community has not yet reached a consensus regarding whether or not a fluid middle or lower crust exists beneath the Tibetan Plateau – and whether or not large-scale lateral flow of that fluid crust occurs – we present here a model for crustal deformation in Tibet that incorporates both tunneling and extrusion to illustrate its consistency with geological and geophysical observations.

It seems likely that the region that is now southern Tibet had already achieved significant elevation before the collision of India and Eurasia ~60–45 Ma (Kapp and DeCelles, 2019). Shortly after collision, however, stresses were quickly transmitted throughout the plateau region, such that contractional deformation along the southern (Himalayan) and northern plateau margins began at roughly the same time (as reviewed by Hodges, 2000; Dayem et al., 2009b; and Kapp and DeCelles, 2019).

By the Late Oligocene or Early Miocene, crustal thickening in southern Tibet had proceeded to the point that thermal conditions were right for the development of a fluid middle crustal channel. By that time, something like the modern monsoon rainfall pattern had been established, and the gravitational potential energy contrast between southern Tibet and India was sufficient to drive the development of the South Tibetan fault system, thereby enabling gravitational potential energy dissipation through southward channel expulsion and monsoon-driven erosion.

It has been widely assumed that the onset of E–W rifting coincided with the plateau reaching uniformly high elevations everywhere, that the onset of rifting was ~8–11 Ma everywhere on the plateau, and that only a catastrophic event – such as the abrupt convective removal of a dense lithospheric root – would have caused large-scale uplift over such a broad area (e.g., Molnar et al., 1993). However, we know so little about the initiation ages of most of the Tibetan rifts that it is not possible to evaluate whether or not they formed over a brief time interval, though the available data suggest a range of perhaps as much as 10 million years (~18–8 Ma). (Many of the available constraints are reviewed by Lee et al. (2011), though those authors reject age estimates older than ~14 Ma and younger than ~10 Ma). As reviewed by Spicer et al. (2021), there is no direct evidence to support an argument that Tibet as a whole abruptly achieved its high elevation. Plate reconstruction exercises performed by Molnar and Stock (2009) provided evidence that India–Eurasia convergence slowed significantly at ~20 Ma and has stayed roughly constant for the past ~11 million years. They suggested that the sudden exertion of an outward force by Tibet could provide a reasonable explanation of the change in convergence and implied that such an event could correspond to the removal of a lithospheric root and be temporally associated with the large-scale uplift of Tibet. However, a different plate reconstruction exercise by Copley et al. (2010) showed no evidence for Miocene slowing of India–Asia convergence that might be related to an abrupt change in the evolution of Tibet. Regardless, initiation of E–W extension in the upper crust of Tibet by the middle Miocene probably marked the beginning of lateral E–W flow of material within a mid-crustal channel of at least limited areal extent. It seems likely that a fluid middle crust was widespread beneath Tibet by late Miocene time, and the fluid layer may have extended to the base of the crust in northern Tibet, north of the northern edge of the subducted Indian Plate (Owens and Zandt, 1997; Nabelek et al., 2009). Once these fluid zones were established, the redistribution of mass under the influence of gravitational potential energy gradients began in earnest. This redistribution is manifested today in two long-wavelength topographic characteristics of the plateau interior: (1) enormous E–W corrugations (Transect A–A', Figure 6); and (2) a general west to east lowering of the mean elevation of the plateau related to the evacuation of the middle crustal channel to the east and southeast off the plateau margin (Transect B–B', Figure 6). This eastward flow was kinematically linked to strike-slip and normal faulting in the upper elastic crust. The flow was bound to the north by rigid crust beneath the Tarim and Qaidam basins, where strain was localized on major transcurrent fault systems and transpressional mountain ranges marking the plateau margin. Similarly, rigid crust beneath the Sichuan Basin served to divert crustal flow to the north and south, but an upstream localization of strain at the western basin margin produced the Longmen Shan (Kirby and Ouimet, 2011). Considerable attention has been focused on the strain field in the Longmen Shan – especially how late Cenozoic strain has been partitioned between the upper and middle crust – in light of the devastating effects of the 2008 Wenchuan (M_w7.9) earthquake (Burchfiel et al., 2008; Kirby et al., 2008; Hubbard and Shaw, 2009). If lateral middle-lower crustal flow remains an important tectonic process in Eastern Tibet, the Neogene and Quaternary upper crustal deformation in the Longmen Shan can be explained as an upper crustal manifestation of localized uplift and concomitant shortening related to the impeded eastward flow of Tibetan crust (cf., Burchfiel et al., 2008).

Questions remain regarding when or if southward channel extrusion ceased after eastward evacuation of the south Tibetan middle crust began. Most researchers presently favor models in which channel extrusion is an early to middle Miocene phenomenon. However, all of the tectonic and climatic ingredients necessary for southward extrusion still exist today, and the available geologic and geodetic data do not preclude it (Hodges et al., 2001; Hodges, 2006). If southward extrusion has been continual since the Middle Miocene, the erosion rate at the Himalayan front has roughly balanced the time-integrated rate of flux of material toward the front, so that the position of the plateau margin has remained roughly stationary over time.

6 Feedbacks among middle-lower crustal flow, landscape evolution, and climate

Orogenic plateaus develop due to crustal thickening and, in the eyes of many, the delamination or convective removal of a dense lithospheric root. These processes raise temperatures in the middle and lower crust by concentrating radioactive heat-producing elements and increasing mantle heat input to the base of the plateau crust. A consequent, dramatic thermal weakening of the crust sets the stage for middle or lower crustal flow, and gravitational potential energy gradients related to plateau uplift provide a motive for flow (eqn [3]) and strongly influence the directionality of that flow. In effect, although lithospheric plate interactions establish orogenic systems, the dynamics of orogenesis are strongly influenced by local variations in gravitational potential energy when an orogenic system becomes fully formed. That influence is especially important when the middle and lower crust are capable of lateral flow. In particular, depending on the magnitude of forces related to plate convergence, ΔGPE may largely drive the ways in which orogenic plateaus grow.

The principal impact of middle or lower crustal flow on the interior topography of plateaus is to limit local relief as the existence of a weak channel at depth reduces the effective elastic thickness of the upper crust (Bird, 1991; Masek et al., 1994). However, a closer analysis of the Tibetan Plateau interior reveals subtle, long-wavelength topographic variations that could be related to the flow field. Although the notion that the E–W corrugations in Tibet are related to buckling above a fluid substrate might be debatable – compare Bendick et al. (2008) and Pullen et al. (2011) – the long-wavelength slope in the mean elevation of the plateau from west to east can be interpreted with some confidence as a robust signature of the eastward evacuation of a crustal channel. Rugged, high-relief topography develops along plateau margins that are bound by rigid crust that impede flow, as evidenced by the Longmen Shan, Kunlun Shan, and Qilian Shan. It is now well-established that the existence of the Tibetan Plateau exerts considerable control on the weather patterns of South and East Asia and the strength of the Indian and perhaps South Asian monsoons (Hahn and Manabe, 1975; Kutzbach et al., 1993; An et al., 2001; Molnar, 2005; Molnar et al., 2010). Although, only recently, has the connection between climate and erosion in tectonically active landscapes, such as the Himalaya, been demonstrated from direct observations (Adams et al., 2020). However, long before then, thermomechanical models of the Himalaya (Beaumont et al., 2001; Beaumont et al., 2004) indicated how focused erosion along the southern Tibetan Plateau margin could 'attract' a mid-crustal fluid channel toward the south, a process that may conspire with shortening across the mountain front to produce the highest mountains on Earth. In addition, Westaway (2009) has suggested that the East Asian monsoon acts as a similar attractor for southeastward flow of the channel, where mass is removed from the plateau in part by the monsoon-driven incision of the major rivers draining the southeastern plateau.

7 Conclusions

Although questions remain regarding the rheology and dynamics of the middle and lower crust beneath the Tibetan Plateau, we are impressed by the overall consistency of the emerging geologic and geophysical database for the late Cenozoic Himalayan–Tibetan orogenic system with tectonic models that include crustal flow. The authors favor models in which the upper crust is largely decoupled from the middle and lower crust below (e.g., Royden et al., 2008; Hodges, 2016). Although lower- and middle-crustal dynamics may be influenced by N–S shortening related to collision, much of the late Cenozoic strain at that level reflects a response to gravitational forces. We are persuaded by the available evidence that fluid middle crust of southern Tibet – representing recycled, accreted Indian crust – flowed southward and extruded at the Himalayan front, attracted there by a steep gravitational potential energy gradient and focused erosion. This energy-dissipative mechanism appears to have been most active in the Early to Middle Miocene, but its persistence into the more recent past remains controversial. Eastward flow of material in the channel toward regions of lower GPE began soon after that, and abundant mineral cooling ages from the northeastern plateau that are likely related to this process record the arrival of the flow front between 13 and 8 Ma (e.g., Kirby et al., 2002; Clark et al., 2005b; Ouimet et al., 2010; Zheng et al., 2010; Lease et al., 2011).

Although deeper levels deformed by ductile flow, the upper elastic crust responded by a complex combination of strike-slip and extensional faulting that collectively resulted in: (1) east–west extension in the western, central, and southern plateau; (2) shortening in the northeastern plateau; and (3) the translation of narrow-upper crustal blocks around the eastern Himalayan syntaxis on arcuate strike-slip fault systems. This pattern of faulting, established in the Middle Miocene, resulted in upper crustal kinematics that are indistinguishable from the modern strain field as measured by geodesy (Zhang et al., 2004). Thus, the eastward displacement of Tibetan crust is manifested by kinematically consistent upper crustal faulting and lower and middle crustal flow.

The low-relief surfaces mapped by Clark et al. (2004) place an important constraint on the relationship between tectonics and landscape evolution across the plateau: In general, the rise of the Tibetan Plateau from its early Cenozoic elevation to its current mean elevation of 5000 m asl was largely passive, and we argue that topographic variations within the plateau are largely manifestations of deformational processes related to viscous flow of the middle crust. Along the plateau margins, landscapes have been shaped by three processes acting in concert and amplifying one another: (1) convergent tectonics related to India–Eurasia collision; (2) lateral crustal flow driven by local pressure gradients; and (3) a climate that is dominated by spatially variable monsoon rainfall.

Acknowledgments

This work has been influenced by many discussions over the years with the likes of Doug Burbank, Clark Burchfiel, Marin Clark, Arjun Heimsath, Jose Hurtado, Eric Kirby, Simon Klemperer, Peter Molnar, Wiki Royden, Lindsay Schoenbohm, Mike Searle, and Kelin Whipple. KVH would like to thank his current and former students and postdoctoral advisees at MIT and Arizona State for keeping him sharp. This work was been funded by National Science Foundation grants from the Tectonics and Continental Dynamics programs (most recently EAR-0711140, EAR-0838112, EAR-1007929, EAR-0708714 to KVH), and the Royal Society (R1\180068 to BAA) . We thank Eric Kirby, William Ouimet, and Lewis Owen for their insightful reviews of previous versions of this manuscript.

Figures and Captions

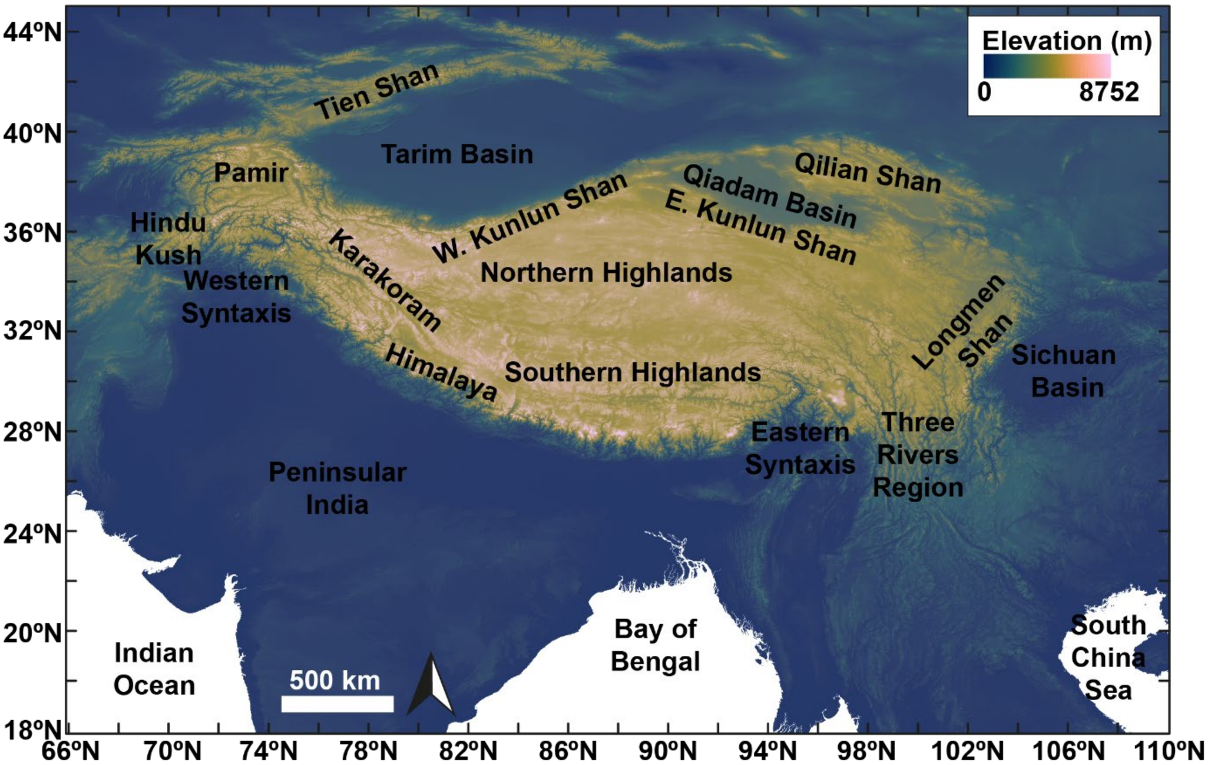
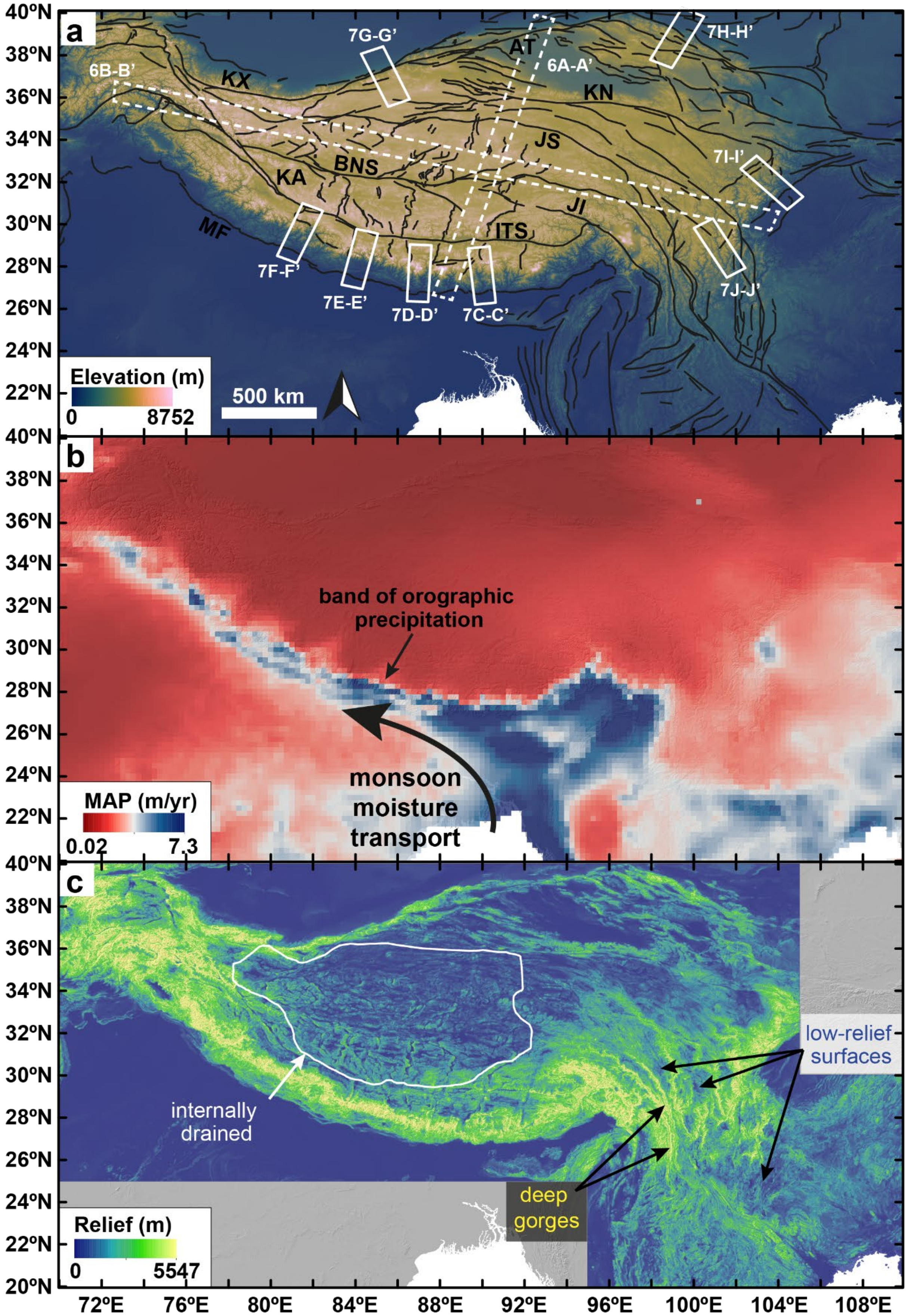


Figure 1 Digital elevation model (GTOPO30 1 km px^{-1}) of south Asia centered on the Tibetan Plateau. Major mountain ranges, basins, and other geographic elements are shown for reference. The plateau is defined in this paper as the high elevation region bordered by the

Himalayan ranges to the south, the Karakoram to the west, the W Kunlun Shan–Altyn Tagh–Qilian Shan mountain systems to the north, and the Longmen Shan and Three Rivers region to the east.



a
7G-G'
7H-H'
AT
6A-A'
6B-B'
KX
KN
JS
7I-I'
BNS
KA
JI
MF
ITS
7F-F'
7E-E'
7D-D'
7C-C'
7J-J'
Elevation (m)
0
8752
500 km
b
band of orographic precipitation
monsoon moisture transport
MAP (m/yr)
0.02
7.3
c
low-relief surfaces
internally drained
deep gorges
Relief (m)
0
5547
40ºN
38ºN
36ºN
34ºN
32ºN
30ºN
28ºN
26ºN
24ºN
22ºN
20ºN
72ºE
76ºE
80ºE
84ºE
88ºE
92ºE
96ºE
100ºE
104ºE
108ºE

Figure 2 a) Major fault systems of the Himalayan–Tibetan orogenic system from the compilation of Taylor and Yin (2009). Major named faults (black lettering) include: AT – Altyn Tagh; JI – Jiali; KA – Karakoram; KN – Kunlun; KX – Karakax; and MF – Main Frontal. Yellow acronyms refer to suture zones: BNS – Banggong-Nujiang; ITS – Indus Tsangpo; and JS – Jinsha. White boxes indicate the positions of topographic swath profiles (100 km wide). Profiles A–A' and B–B' are shown in Figure 6. Profiles C–C' through J–J' are shown in Figure 7. b) Map of mean annual precipitation based on the APHRODITE dataset (Yatagai et al., 2012). Note the prominent precipitation band created as the South Asian Monsoon interacts with the steep range front of the Himalaya. c) Local relief map showing the maximum relief within a 5-km diameter moving window. This map highlights the region of internal drainage on the Tibetan Plateau (outlined in white) and the distribution of remnants of a low-relief surfaces separated by deep gorges in the eastern plateau and off-plateau regions to the east.

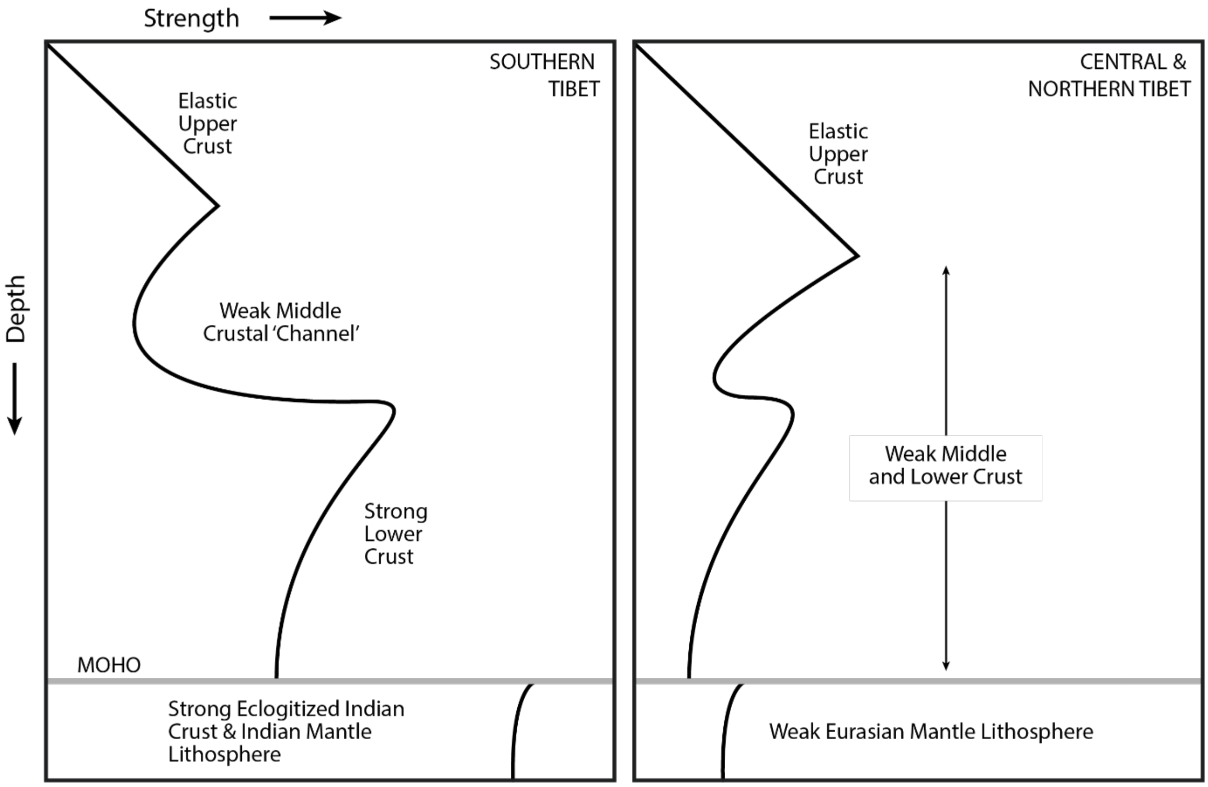


Figure 3 Generalized strength profiles for southern Tibet (left) and central and northern Tibet (right) based on a variety of sources, especially Klemperer (2006) and Wei et al. (2010). Gray horizontal line indicates the position of the Moho. Note that an elastic upper crust and a strong lower crust bound a relatively thin (perhaps 10–25 km?), weak channel in the middle crust of southern Tibet. Farther north, the elastic upper crust is inferred to extend deeper based on earthquake seismology in that region (Wei et al., 2010). An ensemble of geophysical data suggests that the middle crust, lower crust, and upper mantle are relatively weak beneath central and northern Tibet, north of the Banggong-Nujiang suture (Figures 2 and 5).

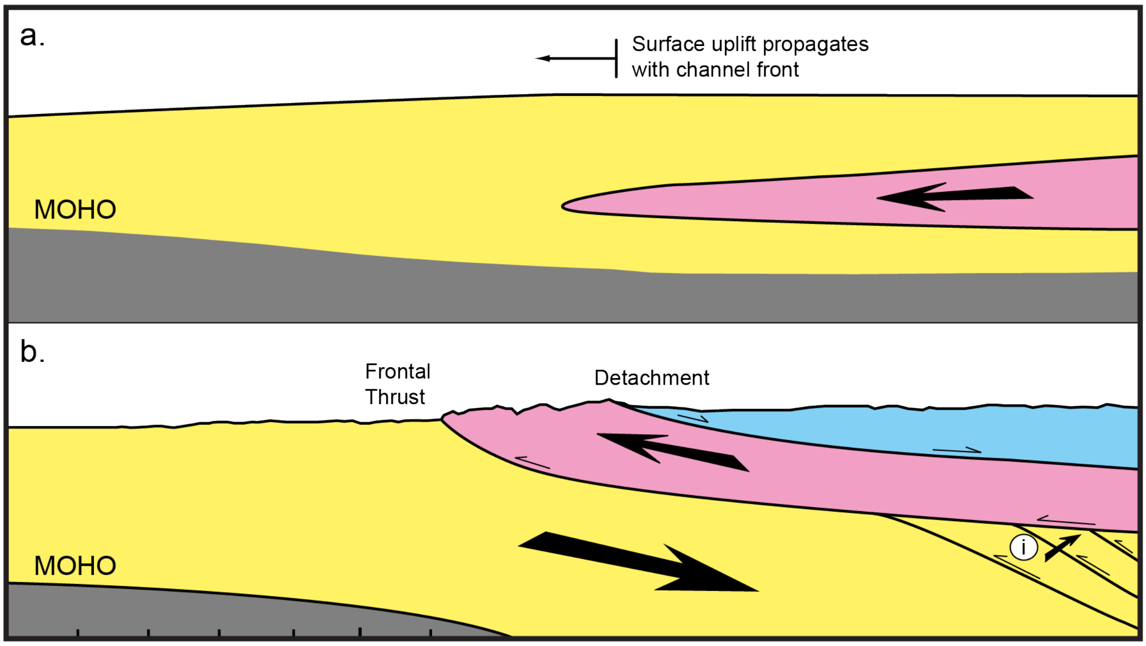


Figure 4 Schematic cross sections illustrating two modes of channel flow (after Hodges, 2006). (a) A weak channel (pink) is shown tunneling through stronger crust (yellow) under the influence of gravity. Arrow shows direction of flow. Note the propagating wave of surface uplift associated with the channel front. (b) Channel extrusion to the surface. This model can illustrate current ideas for Miocene (and possibly later) channel extrusion toward the Himalayan front as driven by the GPE contrast between the Tibetan Plateau to the north and the Indian foreland to the south. The channel is bound by a normal fault system on top (the South Tibetan fault system) and a thrust fault system on the bottom (Main Central thrust system in the Miocene Himalaya). Erosion along the range front removed mass (and therefore excess GPE) from the system. Structures labeled 'i' represent the accretion of down-going Indian Plate into the channel for recycling by channel extrusion toward the range front.

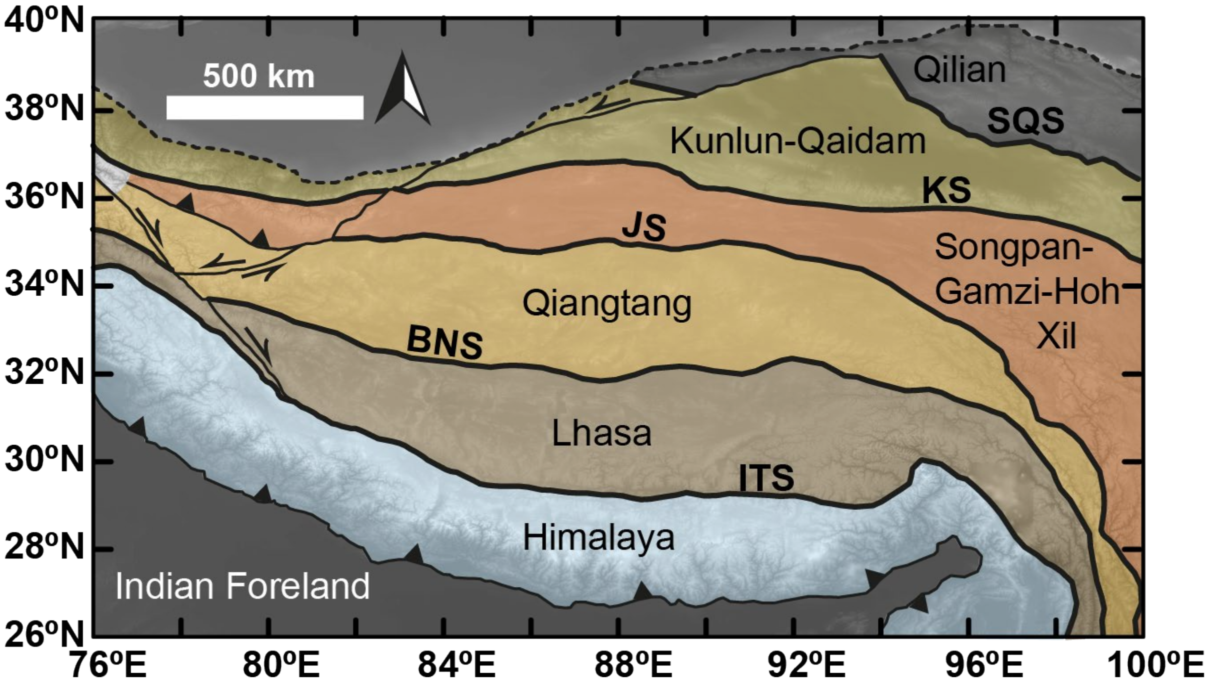


Figure 5 Generalized terrane map of Tibet, based on structures from Taylor and Yin (2009), overlain on elevation data. Major sutures, shown in thick black lines and identified with bold abbreviations, separate named tectonostratigraphic units. The sutures (from north to south) include the late Proterozoic (Devonian?) southern Qilian suture (SQS), Late Triassic Kunlun suture (KS), Early Jurassic (?) Jinsha suture (JS), Late Jurassic Bangong-Nujiang suture (BNS), and Eocene Indus–Tsangpo suture (ITS).

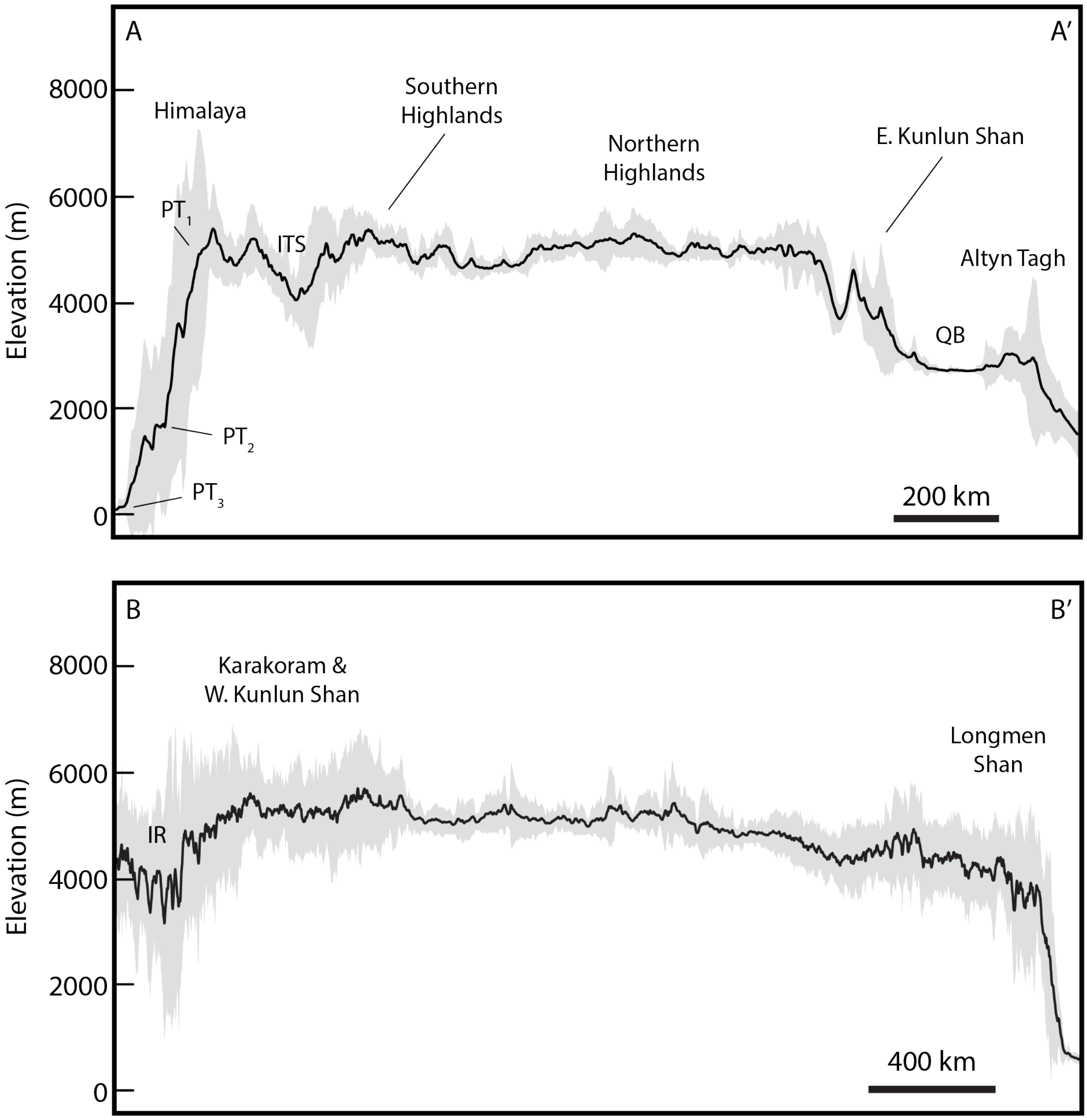


Figure 6 Topographic profiles A–A' (south to north) and B–B' (west to east) across the Tibetan Plateau. Black line indicates mean elevation for a 100 km-wide swath centered on the profile lines shown in Figure 2a. Gray-shaded areas indicate two standard deviations of the mean elevations within the swath. Note scale change between the two profiles. Important geographic and geologic features are indicated, some by acronyms: IR – Indus River gorge; ITS – Indus–Tsangpo suture; and QB – Qaidam Basin. On the Himalayan range front, major physiographic transitions PT1, PT2, and PT3 (Hodges et al., 2001) are shown for reference.

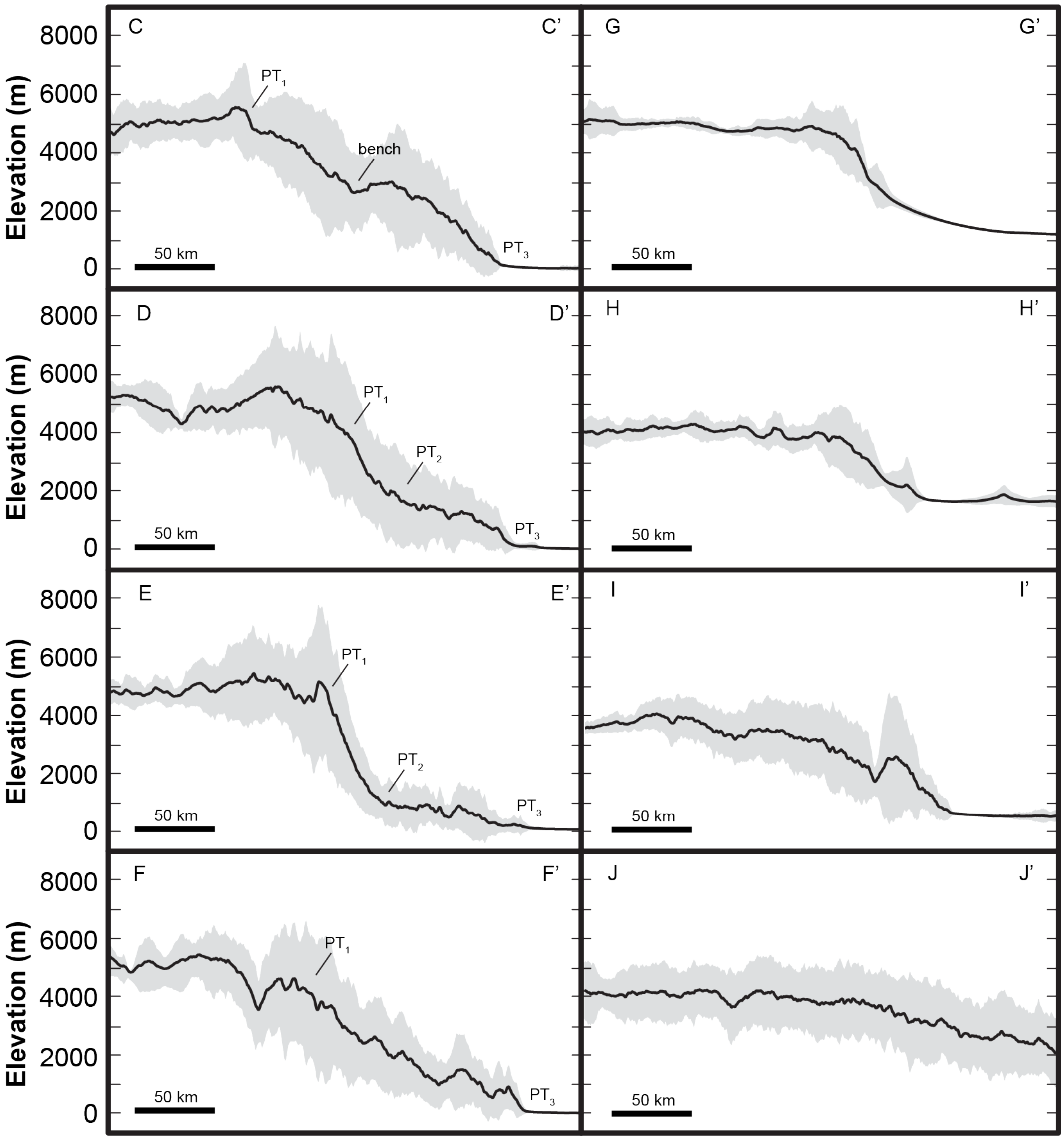


Figure 7 Topographic profiles C–C' through and F–F' across the Himalayan front, G–G' across the W Kunlun Shan–Tarim Basin front, H–H' across the north flank of the Qilian Shan, I–I' across the Longmen Shan-Sichuan Basin escarpment, and J–J' across the diffuse SE Tibet margin. Black line indicates the mean elevation for a 100 km-wide swath centered on the profile lines shown in Figure 2b. Gray-shaded areas indicate two standard deviations of the mean elevations within the swath. Important physiographic transitions discussed in the text are indicated with acronyms PT1. PT2, and PT3 (Hodges et al., 2001).

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