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The growth of northeastern Tibet and its relevance to large-scale continental geodynamics: A review of recent studies

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[1] Recent studies of the northeastern part of the Tibetan Plateau have called attention to two emerging views of how the Tibetan Plateau has grown. First, deformation in northern Tibet began essentially at the time of collision with India, not 10–20 Myr later as might be expected if the locus of activity migrated northward as India penetrated the rest of Eurasia. Thus, the north-south dimensions of the Tibetan Plateau were set mainly by differences in lithospheric strength, with strong lithosphere beneath India and the Tarim and Qaidam basins steadily encroaching on one another as the region between them, the present-day Tibetan Plateau, deformed, and its north-south dimension became narrower. Second, abundant evidence calls for acceleration of deformation, including the formation of new faults, in northeastern Tibet since ~15 Ma and a less precisely dated change in orientation of crustal shortening since ~20 Ma. This reorientation of crustal shortening and roughly concurrent outward growth of high terrain, which swings from NNE-SSW in northern Tibet to more NE-SW and even ENE-WSW in the easternmost part of northeastern Tibet, are likely to be, in part, a consequence of crustal thickening within the high Tibetan Plateau reaching a limit, and the locus of continued shortening then migrating to the northeastern and eastern flanks. These changes in rates and orientation also could result from removal of some or all mantle lithosphere and increased gravitational potential energy per unit area and from a weakening of crustal material so that it could flow in response to pressure gradients set by evolving differences in elevation.

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1. Introduction

[2] The height and lateral extent of the Tibetan Plateau make it a popular field laboratory for understanding the processes of large-scale continental deformation (Figure 1), because what is understood about these processes, as revealed by deformation either within Tibet or on its margins, should be transferrable to more modest belts. Accordingly, as a prototype for large-scale continental deformation, studies of Tibet have elicited a spectrum of different, in some cases mutually inconsistent, interpretations. Perhaps not surprisingly,

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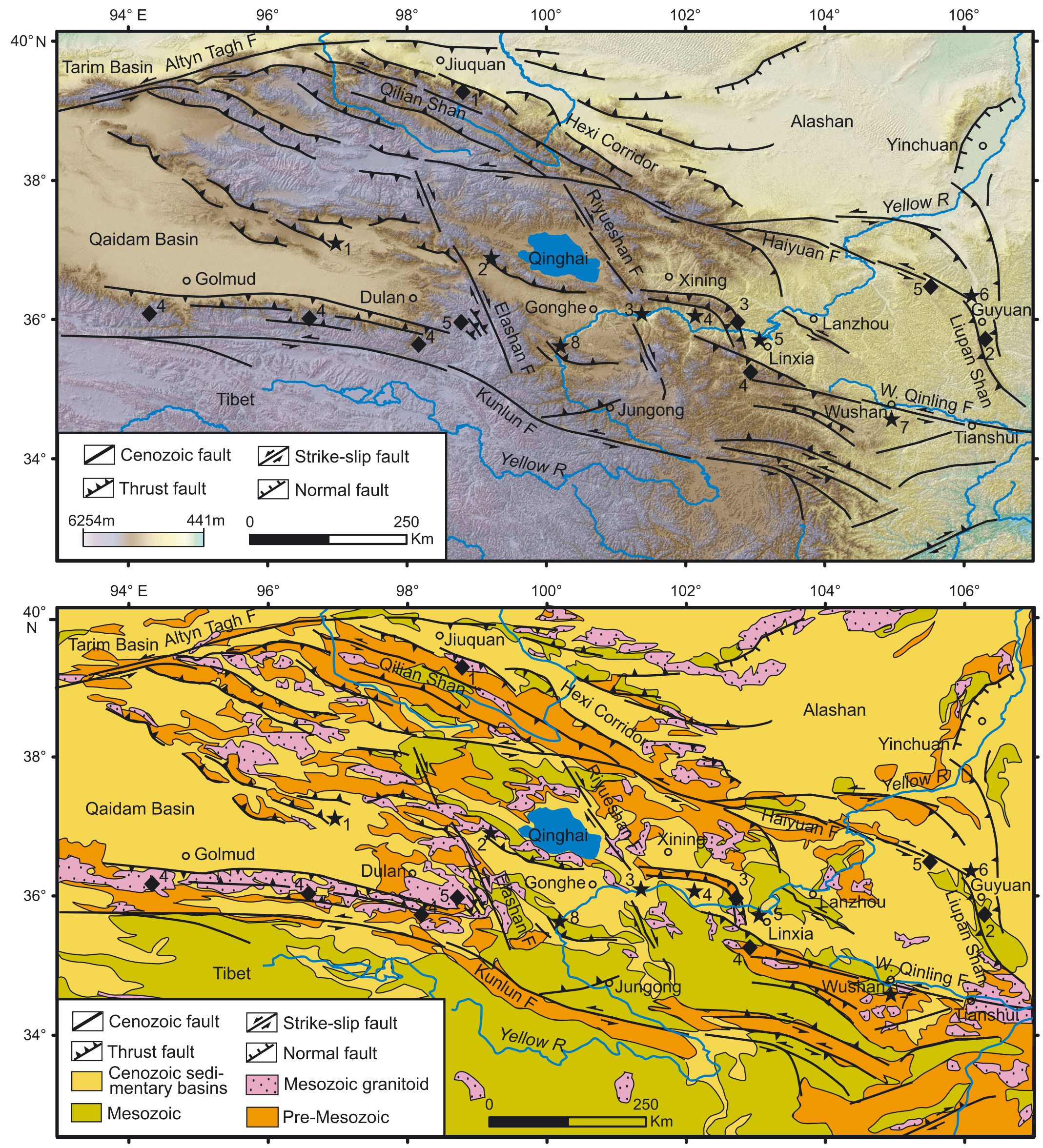

Figure 1. (top) Topographic map and (bottom) simplified geologic map of northeastern Tibet showing basic rock units, major faults, and sites where thermochronology, detrital zircon, and magnetostratigraphic studies were carried out. Diamonds show sites of thermochronology sampling: 1: *Zheng et al.* [2010]; 2: *Zheng et al.* [2006]; 3: *Lease et al.* [2011]; 4: *Clark et al.* [2010]; 5: *Duvall et al.* [2013]. Stars show sites of magnetostratigraphic profiles: 1: *Fang et al.* [2007]; 2: *Zhang et al.* [2012]; 3: *Fang et al.* [2005]; *Yan et al.* [2006]; 4: *Lease et al.* [2012a]; 5: *Fang et al.* [2003]; 6: *Wang et al.* [2011]; 7: *Wang et al.* [2012]; 8: *Craddock et al.* [2011].

too, given the remoteness of the region, new data can cause marked shifts in our perception of how Tibet has grown and in what these new perceptions mean for past ideas.

[3] One common view that seems doomed to modification, if not imminent dismissal, is that crustal deformation of rock that is now part of the Tibetan Plateau has monotonically migrated outward, largely northward, since collision with India, when an Andes-like margin may have lain along the southern edge of Eurasia. Although decades ago this idea was treated as common sense, tests of it were nevertheless proposed. Such northward growth followed logically from similarities of continental deformation north of India with that in a thin viscous sheet penetrated by a rigid indenter [*England and Houseman*, 1986]. *Tapponnier et al.* [2001] formalized this further by postulating that outward growth occurred in steps. As we discuss below, much evidence has accumulated in the 21st century to suggest that the northern and southern edges of Tibet were already defined when India and Eurasia collided at ~50 (or maybe at 40) Ma, and these edges have been moving closer to one another over time. By contrast, crustal deformation in northeastern and eastern Tibet seems to have developed since the collision, apparently with some outward growth, though among us, we do not all share the same view.

[4] A second, controversial, view is that deformation within Asia, not just Tibet, underwent a change some time

since ~15 Ma. The occurrence of a change is undeniable, for today, normal faulting on planes striking approximately north-south is widespread, and thrust faulting is limited to the margins of the plateau [e.g., *Elliott et al*., 2010; *Molnar et al*., 1993]. Such normal faulting implies not only east-west extension of the Tibetan Plateau but also thinning of at least the upper crust (where most earthquakes occur and active faults are apparent). As some subcrustal earthquakes also show normal faulting, albeit beneath only portions of Tibet [e.g., *Chen et al*., 1981; *de la Torre et al*., 2007; *Molnar and Chen*, 1983; *Zhu and Helmberger*, 1996], some of us presume that the entire crust is thinning. In any case, unlike the thrust faulting and folding that occurred within Tibet since collision and thickened the crust [e.g., *Dewey et al*., 1988; *Horton et al*., 2002; *Kidd et al*., 1988; *Rohrmann et al*., 2012; *Spurlin et al*., 2005; *Studnicki-Gizbert et al*., 2008; *Wang et al*., 2008], normal faulting is not a process by which crust thickens or one that builds a high plateau; hence, its prevalence today requires a change in geodynamics since the Tibetan Plateau began to grow [e.g., *England and Houseman*, 1989].

[5] Controversy surrounds the idea of a change since ~15 Ma also because dating still constrains timing only loosely. More importantly, not only the reasons given for such a change but also the precise mechanisms responsible for a change require additional tests to convince skeptics. *England and Houseman* [1989] showed that removal of some or all of Tibet's mantle lithosphere could lead to an increased mean elevation of the Tibetan Plateau, a change from contraction to extension and normal faulting across it, and a resulting change in the balance of forces (per unit length) on the margins of the plateau that would induce or accelerate deformation there. *Harrison et al*. [1992] suggested that this change occurred at ~8 Ma. *Molnar et al*. [1993] then synthesized data and exploited quantitative assessments of various processes to support removal of mantle lithosphere. Further tests of this idea, however, seem to have convinced only those predisposed to finding the tests convincing; many see the removal of mantle lithosphere as a bankrupt idea.

[6] An alternative possibility is that the slow heating of the Tibetan crust, either due to increased radiogenic heat production in thickened crust or aided by removal of mantle lithosphere, gradually weakened the middle and/or lower crust so that it could flow more easily than either the overlying brittle upper crust or the underlying, presumably strong, mantle lithosphere [e.g., *Clark and Royden*, 2000; *Cook and Royden*, 2008; *Clark et al*., 2005b; *Royden*, 1996, *Royden et al*., 1997]. *England and Houseman* [1989] argued that weakening of the entire lithosphere would lead to enhanced crustal shortening, not to normal faulting and crustal thinning, but if weakening were confined to the midcrust, flow there could result in thinning of the overlying upper crust. Depending on the rate of outward flow from high to lower terrain, channel flow of middle and/or lower crust could both induce the observed normal faulting and crustal thinning in the highest parts of the Tibetan Plateau [*Royden et al*., 2008] and inflate crust on the flanks of the plateau, causing its surface to rise.

[7] We synthesize recent work in northeastern Tibet (Figure 1), both undertaken together by the authors and carried out by others, that has been aimed at these questions: when did deformation begin, how has it changed since the collision between India and Eurasia occurred, and what processes might have led to such changes? In most cases, the data that we present do not directly address crustal deformation in the strict sense but surrogates of such deformation, such as changes either in exhumation rates of exposed high terrain or in sedimentation rates in basins adjacent to high terrain. Because tectonics is not the only class of processes that can affect exhumation or sedimentation rates, there is a risk that what we attribute to tectonics might reflect some other process (climate change, stream capture, changes in erodibility of material no longer present, etc.). We contend that because several observations imply tectonic changes in the region, the others are also likely to reflect initiation or acceleration of tectonic activity.

2. Deformation of Northern Tibet Shortly After Collision

[8] Evidence from many localities suggests a date for collision between 55 and 45 Ma [e.g., *Dupont-Nivet et al*., 2010a, 2010b; *Garzanti and van Haver*, 1988; *Green et al*., 2008; *Najman et al*., 2010; *Rowley*, 1996, 1998; *Zhu et al*., 2005]. In the western Himalaya, however, the initial collision seems to have been between India and a small fragment or an island arc, i.e., the Kohistan-Ladakh Batholith, not yet part of southern Eurasia [*Khan et al*., 2009]. Final closure between the combined Kohistan-Ladakh Batholith and Indian subcontinent with southern Eurasia occurred more recently, at ~40 Ma [*Bouilhol et al*., 2013]. Whether final closure of the sea between India and Eurasia farther east also occurred as recently as 40 Ma or earlier remains an open question [see also *Aitchison et al*., 2007, 2008; *Garzanti*, 2008].

[9] Abundant evidence demonstrates precollision crustal shortening in southern Tibet, which suggests that that region was high, like the present-day central Andes [e.g., *Burg and Chen*, 1984; *Burg et al*., 1983; *Dürr*, 1996; *England and Searle*, 1986; *Kapp et al*., 2007; *Murphy et al*., 1997; *van der Beek et al*., 2009; *Volkmer et al*., 2007]. Deformation occurred somewhat farther north or northeast in more central Tibet not long after collision, whether at ~50 or 40 Ma [e.g., *Horton et al*., 2002; *Rohrmann et al*., 2012; *Spurlin et al*., 2005; *Wang et al*., 2008]. More surprising, it is now clear that deformation throughout most of Tibet was underway not long after the collision. The accumulation of Eocene (if commonly late Eocene) or the earliest Oligocene sediment within basins now within Tibet [e.g., *Horton et al*., 2004; *Ritts et al*., 2004], in one case in a basin created by flexure [*Fang et al*., 2003] and in another with pollen from "high-altitude vegetation" [*Dupont-Nivet et al*., 2008a], implies newly created relief with elevations of ~1000 m or more. Offsets of early Cenozoic sedimentary rock by thrust faults surrounding the Qaidam basin in northern Tibet (Figures 1 and 2b) also imply deformation not long after collision [e.g., *Yin et al*., 2002, 2007, 2008a, 2008b; *Zhuang et al*., 2011], if perhaps not until 10 Myr after it. Declination anomalies measured on paleomagnetic samples from numerous basins in northeastern Tibet call for 20–40° of clockwise rotation since some time between 45 and 29 Ma; samples younger than ~25 Ma from most regions show declination anomalies of only ~10° or less [*Dupont-Nivet et al*., 2004, 2008b]. Conceivably, all of this rotation occurred in a short period after collision, but because timing is not precise and different basins and ranges

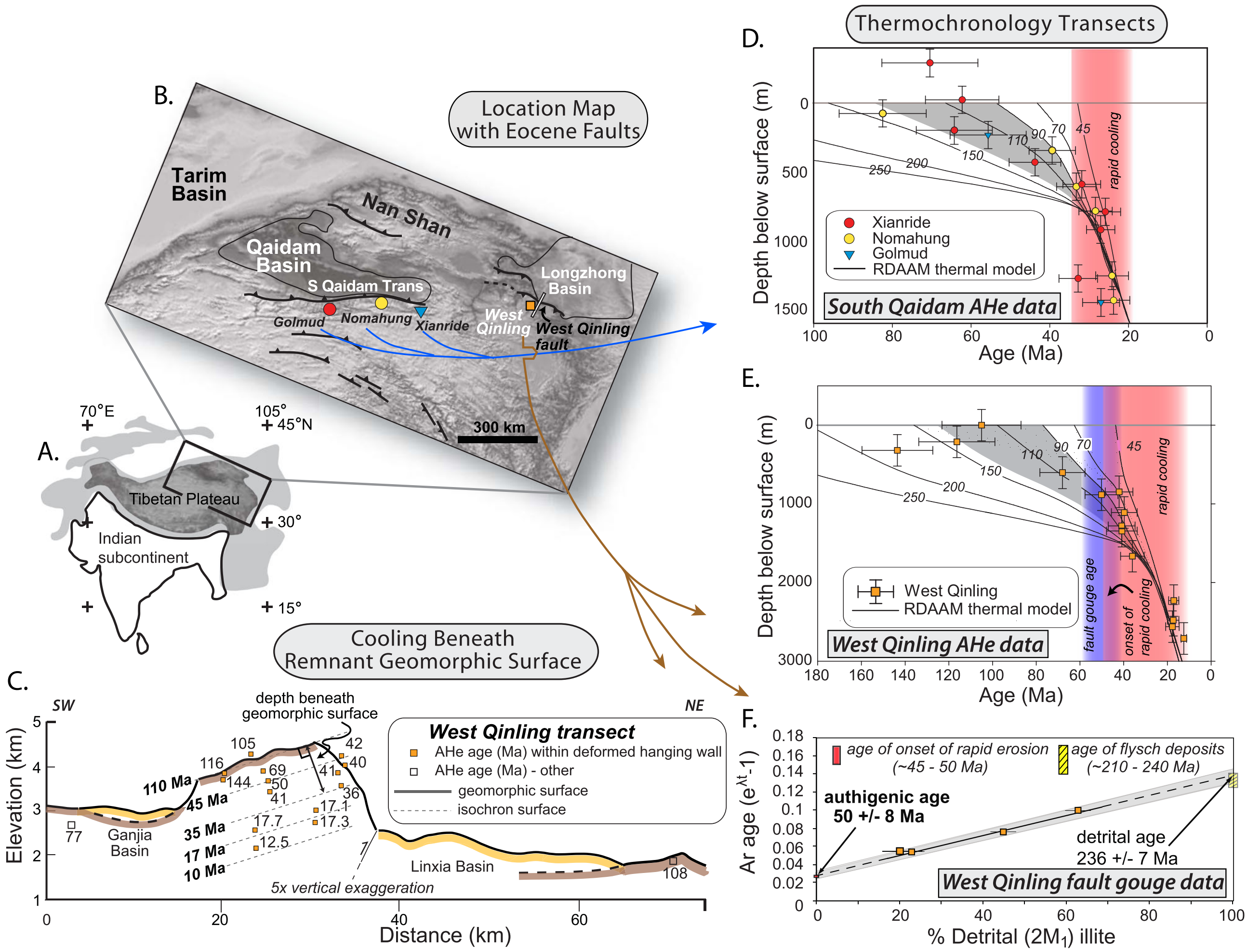

Figure 2. Regional map showing locations of (a) Paleocene-Eocene age faulting and (b) blowup of northeastern Tibet, showing evidence of Eocene deformation in the Linxia–West Qinling region. (c) Thermochronological sampling on an elevation transect shows an (e) abrupt increase in cooling rate near 45–50 Ma [*Clark et al.*, 2010]. Apatite helium ages are reported as averages of three or four single-grain analyses with 2σ standard error. Uncertainties on the depth measurement were determined from the amount of relief on the geomorphic surface. One-dimensional forward thermal models, lines with numbers, simulate initial heating (burial) and isothermal holding followed by accelerated erosion of the hanging wall of the West Qinling fault. Numbers on the lines refer to the initiation of isothermal holding (Ma), which continues until erosion at 0.1 mm/yr begins at 45 Ma, assuming a layered thermal structure and using the radiation damage model kinetics of *Flowers et al.* [2009]. The isothermal holding that begins some time between 140 and 90 Ma and continues until 45 Ma best fits the data (grey shading); the 50 Ma models are not shown but fit the data similarly. (d) Similar thermochronological sampling on elevation transects farther west also shows abrupt increases in cooling rate near 35 Ma, equivalent to erosion at 0.2 mm/yr, that follows burial and isothermal holding since 90–140 Ma [*Clark et al.*, 2010]. Finally, (f) fault gouge dating from the West Qinling fault (Figures 2b and 2c) demonstrates that the fault was active at 50 ± 8 Ma [*Duvall et al.*, 2011] (1σ standard error based on a York regression of four individual age/population aliquots as indicated by grey shading). Horizontal error bars represent 2M1% (detrital illite) uncertainties with 3–5% precision. The age function $(e^{\lambda t} - 1)$ is linearly proportional to the percentage of detrital mica, and the vertical error bars are smaller than the symbol size. λ is the decay constant of ^{39}K to ^{39}Ar.

could undergo rotations at different times, much of the clockwise rotation could have occurred ~20 Myr after collision. Thus, these data leave open the question of whether the northern margin of Tibet was active at the time of collision or not until another 10–20 Myr had elapsed.

[10] Recently, some of us have shown that the margin of high terrain in northeastern Tibet, now ~1000 km north of the Himalaya, not only underwent rapid exhumation at the time of the collision, but that the West Qinling fault (Figures 1 and 2b), the main thrust fault that dips beneath the plateau, was active at that time. Just south of the Linxia basin, which *Fang et al.* [2003] inferred had been flexed down by at least 29 Ma, a high, relatively flat terrain has been deeply incised. Roughly 2000 m of steep relief allowed *Clark et al.* [2010] to sample an age-elevation transect (Figure 2c) whose dates demonstrate slow cooling of rock

from 140–90 Ma to 45–50 Ma, when the rate of cooling accelerated by approximately 10 times (Figure 2e). *Clark et al.* [2010] also reported accelerated exhumation farther west along the Kunlun beginning at ~35 Ma (Figure 2d). Such rapid cooling calls for the removal of 2000 m of rock from above the Triassic granite now exposed in this region. The simplest, but not unique, explanation is that at 45–50 Ma, this region began to rise rapidly relative to the local base level, which presumably was the surface of the Linxia basin at that time, and erosion stripped the rock presently exposed on the Tibetan Plateau of its cover.

[11] To test this interpretation, *Duvall et al.* [2011] exploited fault gouge dating: the $^{40}Ar/^{39}Ar$ dating of mixtures of authigenic and detrital clay minerals in fault gouge [*van der Pluijm et al.*, 2001]. Clay minerals in gouge consist of particles ground up from the surrounding rock and of new minerals that grow in the gouge as slip occurs. Although individual particles cannot be isolated, the fraction of each type (detrital or authigenic) can be measured. Resulting $^{40}Ar/^{39}Ar$ ratios from such mixtures then give weighted averages of those derived from the two sources (Figure 2f). Extrapolations to pure detrital and pure authigenic material should give the ages of the wall rock and of a date when the fault was active. *Duvall et al.* [2011] obtained a sensible age of 236 ± 7 Ma for the surrounding Middle Triassic wall rock and an age of 50 ± 8 Ma for when the fault was active.

[12] Although uncertainties not only in these ages but also in the timing of collision allow for a delay, conceivably of 10 Myr, between collision and slip on this thrust fault, clearly, northeastern Tibet began to undergo horizontal shortening shortly after collision or, conceivably, before it [e.g., *Craddock et al.*, 2012].

[13] The concurrence of collision with deformation far from the locus of collision might seem surprising, particularly to those burdened by misconceptions about stress “propagating” so far so fast. Stress does not propagate instantaneously, but at speeds of elastic waves, and for time scales of geologic processes, the time for seismic waves to travel thousands of kilometers is obviously short. When tractions are applied to portions of boundaries of deformable media, related components of stress decrease away from those boundaries. The distances to which components of stress reach depend on how, as expressed by the equation of equilibrium, their gradients balance both one another and the body force due to gravity acting on lateral differences in density. For likely tractions that India might apply to the southern edge of Eurasia, the distance scale for the decrease of most components of stress from the collision zone into Eurasia should be proportional to the width of the boundary [*England et al.*, 1985], which, for the India-Eurasia collision, is the length of the Himalaya, i.e., ~2500 km. Simple calculations that exploit a thin viscous sheet show that at the time of collision, deformation should be expected 2000–3000 km from a boundary that is 2500 km long [*Dayem et al.*, 2009b]. Moreover, the presence of a relatively strong object, like the Tarim basin, immersed in weaker lithosphere, which surely characterizes Tibet today, will concentrate deformation near that boundary [e.g., *Dayem et al.*, 2009a; *England and Houseman*, 1985; *Vilotte et al.*, 1984, 1986]. In light of these considerations, an onset of deformation on the northern margin of the Tibetan Plateau at the time of the collision ought not to be a surprise.

[14] In addition, some recent paleomagnetic evidence suggests that Tibet has been shortened by only ~1000 km or less since collision. *Chen et al.* [1993] had measured inclinations of magnetization in sedimentary rock in southern Tibet suggesting 1700 ± 600 km of postcollisional northward displacement of southern Tibet with respect to southern Eurasia. Indeed, some recent work has suggested comparable amounts: 2140 ± 590 km [*Chen et al.*, 2010] and 1847 ± 763 km [*Liebke et al.*, 2010]. Others, however, have challenged earlier results, arguing that compaction of magnetized sediment has led to shallowing of inclinations and a bias toward low paleolatitudes; using volcanic rock that should not have undergone such flattening, they suggested less northward displacement of southern Tibet: 1100 ± 500 km [*Dupont-Nivet et al.*, 2010a] and, possibly, only 200 ± 900 km if allowance for errors in the virtual geomagnetic pole for Eurasia is included [*Dupont-Nivet et al.*, 2010b] or between 370 ± 280 km and 810 ± 600 km [*Tan et al.*, 2010]. Consistent with these small amounts, *Lippert et al.* [2011] measured only 290 ± 530 km of northward displacement of central Tibet relative to Eurasia. Thus, at the time of collision, northern and southern Tibet may have been less than 2000 km apart, which makes synchrony of collision and deformation of northern Tibet yet more easily understood.

[15] A stationary northern perimeter to the plateau also has a potential geodynamic consequence when viewed with plate convergence rates since collision [*Clark*, 2012]. Dividing the convergence rate by the distance across which greater India and Eurasia deform produces an average bulk strain rate for Tibet. Convergence rates have declined since collision [e.g., *Molnar and Stock*, 2009] as must the distance across the deforming region measured from the northern margin of the plateau to the northernmost point on the undeformed Indian plate. Both have decreased such that the average compressional strain rate across the plateau has remained constant since collision at a rate of 7.03×10^{-16} s^{-1}, which is equal to the modern rate determined by GPS. For both linear and nonlinear constitutive relationships between stress and strain rates, an average strain rate most simply implies constant average stress, or constant force per unit length along the margins of the belt. If this were so, then constant forcing would imply that the work done against gravity, as thickening occurred, would be small compared to the viscous resistance associated with the deformation of the lithosphere or that the thickening of the plateau occurs under neutrally buoyant conditions (i.e., thickening of the mantle lithosphere offsets the increase in buoyancy due to crustal thickening). A viscous resistance of the continental lithosphere has not been previously considered as a type of plate forcing, and eastern Asia may offer one extreme example of such.

3. Oligocene to Middle Miocene Deformation of Northern Tibet

[16] Approximately north-south (or NNE-SSW) shortening of northern Tibet seems to have continued until at least ~22 Ma, when rapid exhumation of rock in the Laji Shan began (Figures 3b and 3c) [*Lease et al.*, 2011] and when sedimentation in the adjacent Xunhua basin accelerated (Figure 3d) [*Lease et al.*, 2012a]. Although erosion has been so deep that no remaining rock contains a (U-Th)/He record of an earlier period of slow cooling and slow exhumation within the Laji Shan (Figure 3c), *Lease et al.* [2011] showed

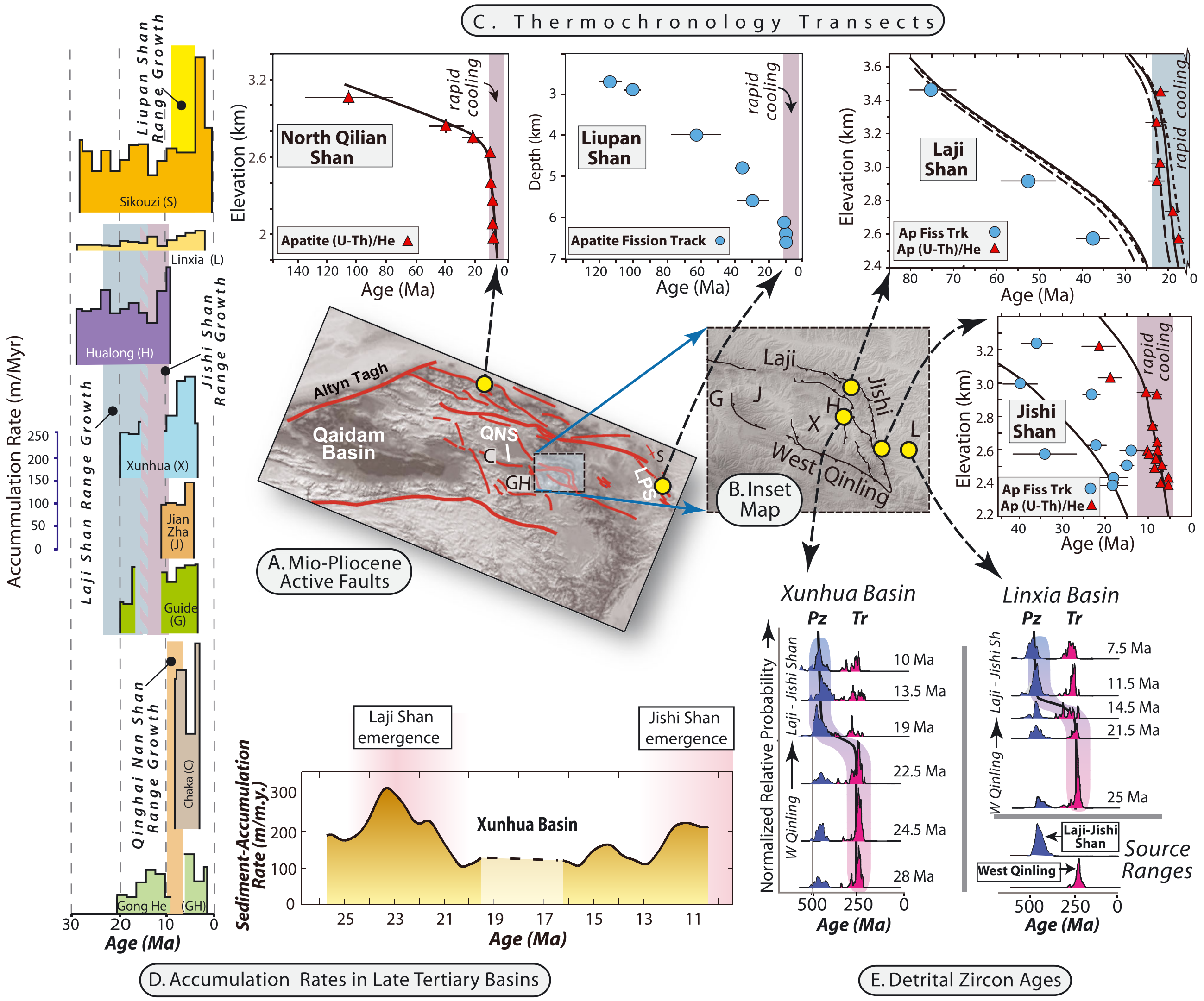

Figure 3. (a) Regional map with active Miocene and/or Pliocene faults (in red). C: Chaka section; GH: Gong He section; LPS: Liupan Shan; QNS: Qinghai Nan Shan; S: Sikouzi section. Locations of age-elevation transects are shown with yellow circles. (b) Sites bounding the Laji–Jishi Shan and the West Qinling Shan. Magnetostratigraphic sections used to calculate sediment-accumulation rates are as follows: G: Guide; H: Hualong; J: Jian Zha; L: Linxia; and X: Xunhua. (c) Age-elevation transects show 2σ errors on (U-Th)/He replicated single-grain and/or pooled fission track ages on apatite [*Lease et al.*, 2011; *Zheng et al.*, 2006, 2010]. Accelerated exhumation is inferred where sites spanning a significant altitudinal range show very similar ages. (d) Accumulation rates derived from magnetostratigraphically dated sections, not corrected for compaction (located in Figures 3a and 3b) [*Craddock et al.*, 2001; *Fang et al.*, 2003, 2005; *Hough et al.*, 2011; *Lease et al*. 2012a; *Zhang et al*., 2012]. Rates in the Jian Zha are from R. O. Lease et al. (unpublished data, 2011). Rates from the Xunhua basin, derived from the Hualong section (H) show accelerated accumulation attributed to the growth of the Laji and Jishi Shan at ~22 and 10 Ma, respectively. (e) Detrital zircons in the Xunhua basin reveal source-area changes from the West Qinling Shan to the Laji–Jishi Shan between 22 and 19 Ma, whereas zircons from the Linxia basin depict the rise of the Jishi Shan by 11 Ma [*Lease et al*., 2012a]. These data argue for a change in exhumation and sedimentation sometime since 15 Ma and, presumably, for the emergence of mountain ranges at that time.

that fission track ages imply a period of slow cooling, and presumably slow exhumation, until ~22 Ma, when (U-Th)/He ages show rapid cooling (Figure 3c). Moreover, doubling of the accumulation rate of sediment and the introduction of new detrital zircons in proximal portions of the adjacent Xunhua basin at the same time supports the inference that the Laji Shan emerged topographically and as a new detrital source area at ~22 Ma (Figure 3e) [*Lease et al*., 2012a]. Thus, given the east-west trend of the range from which rock was removed, *Lease et al.* [2012a] inferred that approximately north-south shortening, at least south of the southern end of the Qilian Shan, continued until the earliest Miocene time.

[17] Following the initiation of rapid exhumation of rock along the northern margin of the Kunlun at ~35 Ma

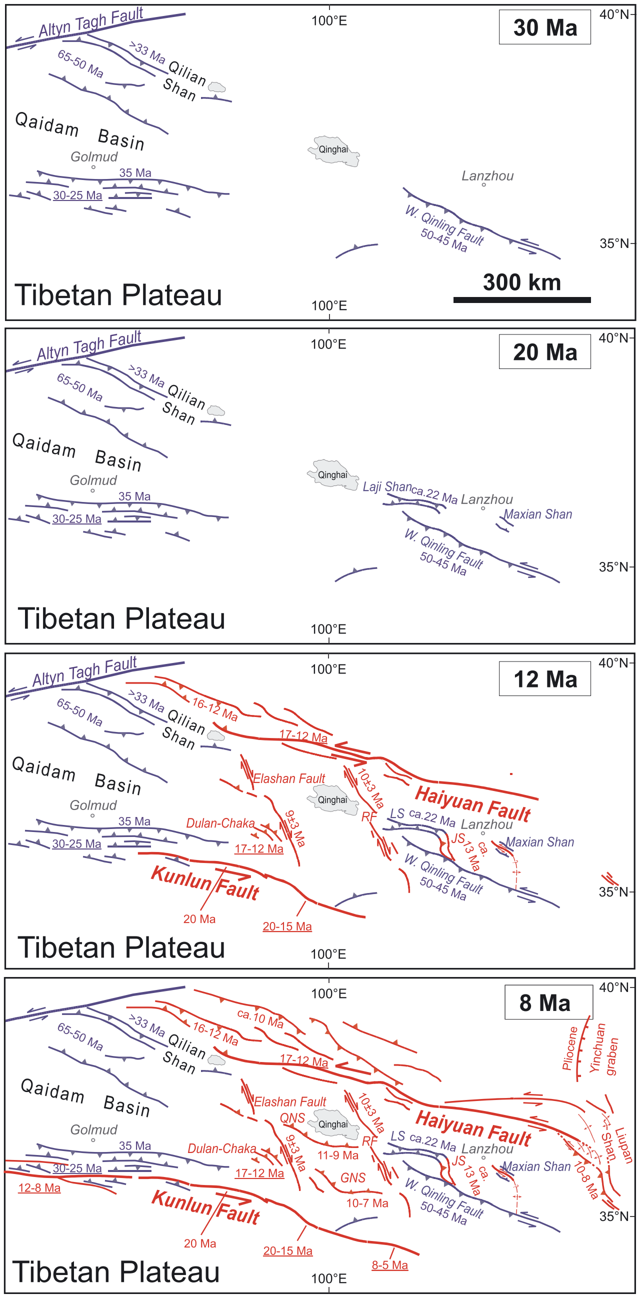

Figure 4. Sequence of maps showing locations, ages, and styles of fault deformation in northeastern Tibet, during Cenozoic time. Ages on faults indicate initiation of intervals of known or inferred activity. Faults that were initiated in the Paleogene are shown in blue, and newly initiated Neogene faults are in red; these different colors delineate fault styles that developed before and after the proposed ~15 Ma kinematic change. Note the roughly 150-km northeastward expansion of deformation, the coupled and growing strike-slip fault systems, and the switch from east trending toward NNW trending thrust faults in the middle to late Miocene time.

(Figure 2d) [*Clark et al.*, 2010], exhumation within ranges surrounding the Kunlun fault itself occurred between 30 and 20 Ma and more recently [*Duvall et al.*, 2013]. We interpret these dates for rapid exhumation as evidence that crustal thickening due to crustal shortening occurred during the period from ~35 to ~20–15 Ma (Figure 2d). Moreover, (U-Th)/He age-elevation transects in the Dulan-Chaka highland (Figure 4) suggest acceleration of exhumation near 15 Ma [*Duvall et al.*, 2013], a result consistent with fission track ages from this region [*Lu et al.*, 2012]. The prevalence of NW striking thrust faults in the Dulan-Chaka highland suggests that NE-SW crustal shortening occurred in this region, although only a few kilometers of exhumation occurred over steep to vertical reverse faults. Rather than accommodating much crustal shortening, the thrust faults of the Dulan-Chaka highland appear to be kinematically linked with the Elashan right-lateral fault. *Duvall et al.* [2013] associated rapid cooling of rock beginning at ~20–15 Ma to ~8 Ma within the Kunlun fault zone as evidence for left-lateral slip on the east-west striking Kunlun fault, which was established at least by middle to late Miocene time. In any case, the Oligocene and early Miocene exhumation of the Kunlun is consistent with crustal shortening along the northern edge of the Tibetan Plateau and south of the Qaidam basin.

[18] Deformation northeast of these regions of relatively rapid exhumation between ~50 and 15 Ma seems to show a different pattern. *Wang et al.* [2013] inferred from seismic reflection profiles across the Sikouzi basin and exposed sediment within the basin (Figures 1 and 3a) that normal faulting and northeast-southwest extension occurred in this region, when exhumation occurred in the terrain to the southwest of the basin. Thus, they questioned whether the Linxia and adjacent basins developed in a foreland setting, as *Fang et al.* [2003] inferred. Regardless of the nature of faulting along the West Qinling fault, it appears that north-south or northeast-southwest shortening did not extend far beyond that fault.

[19] As noted above, paleomagnetic declination anomalies call for 20–40° of clockwise rotation of parts of northeastern Tibet since 45–29 Ma [*Dupont-Nivet et al.*, 2004, 2008b] and, for some regions, as recently as ~15–10 Ma, although subsequent work by *Yan et al.* [2006] indicates 25 ± 5° of rotation of the Guide basin in middle Miocene time. Clockwise rotation is opposite to the sense today as revealed, for example, by GPS measurements in this region [e.g., *Duvall and Clark*, 2010; *Gan et al.*, 2007; *Zhang et al.*, 2004]. Insofar as clockwise rotation is a manifestation of right-lateral simple shear, the operative right-lateral shear would have been on approximately north-south (or NNE-SSE) trending planes [e.g., *Cobbold and Davy*, 1988; *England and Molnar*, 1990; *Kirby and Harkins*, 2013]. Hence, it would reflect more rapid northward displacement of the region to the west than that to the east relative to stable eastern China. Such right-lateral shear might continue today, for it is consistent with the widespread left-lateral strike-slip faulting on east-west (and NW-SE) planes. The consistency of declination anomalies of samples older than 45 Ma and their marked difference from samples younger than 29 Ma, however, imply that a kinematic change occurred since ~29 Ma. Moreover, in the case of the Guide basin (G in Figure 3b), the demonstration of 25 ± 5° of clockwise rotation between 17 and 11 Ma by *Yan et al.* [2006] suggests that such a change occurred since ~15 Ma.

[20] Collectively, the studies summarized above suggest that approximately north-south, or NNE-SSW, shortening occurred in northeastern Tibet from the time of collision

through ~22 Ma and until ~18 Ma, when exhumation slowed [*Lease et al*., 2011]. Insofar as clockwise rotation reflects right-lateral shear on roughly north-south planes, the amount of northward displacement of rock in northeastern Tibet, relative to the relatively stable region to its east, must decrease eastward. The present-day left-lateral shear of the region, demonstrated by GPS measurements [*Duvall and Clark*, 2010; *Gan et al*., 2007; *Zhang et al*., 2004], however, requires a subsequent change in the style of deformation.

4. Changes in Rates and Orientations of Deformation in Northeastern Tibet Since ~15 Ma

[21] As discussed in this section, rapid exhumation of rock within many of the mountain ranges in northeastern Tibet suggests that they emerged within a wider basin to create smaller adjacent, now intermontane, basins [e.g. *Horton et al*., 2002], since ~20 Ma in the case of the Laji Shan (Figure 3e) [*Hough et al*., 2011, 2013; *Lease et al*., 2012a] and since ~15 Ma for most of the others. Moreover, in the eastern part of the region, where orientations of crustal shortening can be inferred, such orientations tend to be closer to northeast-southwest, if not east-west, than to north-south (Figure 4). Collectively, these results suggest that the rate of deformation accelerated and that the orientation of crustal shortening has changed since ~20–15 Ma (Figure 4), though such acceleration and changes in orientation need not be simultaneous across the region, or even across a single mountain range.

[22] Cooling ages from the western end of the northern edge of the WNW trending Qilian Shan [*Zheng et al*., 2010], from the NNW trending Jishi Shan [*Lease et al*., 2011], and from the NNW trending Liupan Shan [*Zheng et al*., 2006] (Figures 3c and 4) suggest rapid exhumation since 8–13 Ma of rock that lies in the hanging walls of major thrust faults. Hence, the logical inference is that these dates define when the slip of these faults became rapid enough to create relief that erosive processes could attack. Moreover, roughly concurrently, sedimentation accelerated, provenance changed, and coarse deposits became common in proximal parts of basins adjacent to these ranges (Figures 3d and 4) [*Lease*, 2013]: in the Hexi corridor near the Qilian Shan [*Bovet et al*., 2009], in the Xunhua basin adjacent to the Jishi Shan [*Hough et al*., 2013; *Lease et al*., 2012a], and in the Sikouzi basin near the Liupan Shan [*Wang et al*., 2011, 2013]. In fact, increases in rates and/or grain size and changes in provenance occurred in most basins in northeastern Tibet since ~15 Ma and since ~10 Ma for many (Figures 3d and 3e): perhaps beginning earlier at ~16 Ma in the Wushan basin [*Wang et al*., 2012], at ~13 Ma in the Linxia basin [*Lease et al*., 2012a], and perhaps again near 6 Ma [*Fang et al*., 2003], nearer 8–9 Ma in the Guide basin [*Fang et al*., 2005; *Lease et al*., 2007], and in different parts of the Gonghe basin, between 10 and 7 Ma in the Tongde subbasin [*Craddock et al*., 2011], and near 12 Ma in the Chaka subbasin [*Lu et al*., 2012; *Zhang et al*., 2012]. On the flanks on the Gonghe Nan Shan, the timing of local range growth is revealed by strata that lie in angular unconformable contact with lower stratal packages. Absolute chronology from these strata brackets range growth to 7–10 Ma [*Craddock et al*., 2011]. A change in provenance of sediment in the Chaka subbasin indicates that the Qinghai-Nanshan, along the northeastern margin of the basin (Figures 3d and 4), emerged a bit later, near 6 Ma [*Zhang et al*., 2012]. A divergence of stable isotopes in pedogenic carbonates in the Linxia and Xunhua basins (Figure 4) between 16 and 11 Ma suggests that the Jishi Shan, which lies between them, rose in this interval and separated a larger basin into smaller ones [*Hough et al*., 2011]. Finally, an extrapolation of late Quaternary slip rates on two NNW-SSE trending right-lateral faults, i.e., the Elashan and Riyueshan faults (Figure 4), suggest that slip at a constant rate since 6–12 Ma would account for total offsets on these faults [*Yuan et al*., 2011].

[23] Although most of these studies did not quantify amounts of deformation since ~15 Ma, those that did commonly show shortening of 10–20 km [e.g., *Craddock et al*., 2011; *Lease et al*., 2012b], and similar amounts crudely apply to the regions studied by *Zheng et al*. [2006, 2010]. Thus, average rates of slip since 15 Ma compare with those measured over late Quaternary time [e.g., *Champagnac et al*., 2010; *Hetzel et al*., 2002, 2004; *Tapponnier et al*., 1990; *Zheng et al*., 2013a, 2013b]. When applied to a duration of ~15 Myr, these rates account for estimates of total crustal shortening across the region [e.g., *Lease et al*., 2012b; *Métivier et al*., 1998; *Meyer et al*., 1998], which is consistent with the acceleration and reorientation of deformation in northeastern Tibet since ~15 Ma.

[24] Most of these observations call attention to changes in exhumation rates or in sediment-accumulation rates, which, in turn, are most simply, if not uniquely, treated as evidence of accelerated tectonic activity and the growth of mountain ranges between intermontane basins. Although dating is imprecise for many, these new ages do not allow all of these changes to have occurred (or initiated) at the same time. For example, the emergence of the Qinghai-Nanshan (QNS in Figure 4) did not shed sediment into the Chaka subbasin of the Gonghe basin (Figure 4) until ~6 Ma [*Zhang et al*., 2012], and cooling of rock in the Liupan Shan and the western Qilian Shan did not seem to accelerate until after 10 Ma [*Zheng et al*., 2006, 2010]. On the other hand, the growth of the Jishi Shan seems to predate 10 Ma [*Hough et al*., 2011, 2013; *Lease et al*., 2011, 2012a]. Similarly, *Duvall et al*. [2013] showed that cooling, and presumably exhumation, of the Dulan-Chaka highland (Figure 4) was slow before 15 Ma and then sped up. Moreover, *Yan et al*. [2006] showed that $25 \pm 5°$ of clockwise rotation of the Guide basin occurred between 17 and 11 Ma. Thus, whether one sees abrupt acceleration of deformation since ~15 Ma or continuous evolution of deformation, which is biased by younger events being more clearly recorded, depends in part on prejudices brought by readers, as well as by some of us.

[25] This suggestion of accelerated deformation since ~15 Ma applies to much of Tibet's surroundings, not just northeastern Tibet, including evidence of deep incision and rapid erosion on the eastern margin of Tibet [e.g., *Clark et al*., 2005a; *Godard et al*., 2009; *Kirby et al*., 2002; *Ouimet et al*., 2010], the onset of normal faulting within the high part of Tibet [e.g., *Blisniuk et al*., 2001; *Harrison et al*., 1995; *Lee et al*., 2011; *Pan and Kidd*, 1992; *Woodruff et al*., 2013], accelerated folding and north-south shortening of the Indian plate [*Cochran*, 1990], deformation elsewhere surrounding Tibet, as reviewed recently by *Molnar and Stock* [2009], and even extension near the southern end of the Shanxi graben farther east [*Liu et al*., 2013]. In addition, *Duvall et al*. [2012] recently showed that erosion rates increased by

nearly an order of magnitude across the eastern part of the high Tibetan Plateau between approximately 11 and 4 Ma. They studied detrital material deposited along trunk streams that drain eight watersheds in the high part of eastern Tibet. These watersheds lie along a NNE-SSW transect that is southwest of nearly all of the ranges and basin in northeastern Tibet discussed above. *Duvall et al*. [2012] argued that the increased erosion between 11 and 4 Ma most likely reflects a tilting of the Tibetan Plateau down toward the east and with an increase in height of the drainage basins that they studied. The timing of this apparent tilting is approximately concurrent with the change in orientation of deformation in northeastern Tibet discussed above and demonstrated by *Lease et al*. [2011, 2012a] and *Zheng et al*. [2006].

5. Processes Responsible for a Change in Rates and Orientation of Deformation in Northeastern Tibet

[26] At least three processes during the overall development of the Tibetan Plateau might be responsible for the accelerated and reoriented deformation since ~15 Ma that we describe above. First, these changes might be part of the natural outward growth of the highest part of the plateau. Second, they may reflect a change in the balance between gradients in stress and the body force resulting from removal of some or all the mantle lithosphere beneath the Tibetan Plateau [e.g., *England and Houseman*, 1989; *Harrison et al*., 1992; *Molnar et al*., 1993]. Third, they might result from changes in the rheology of the Tibetan lithosphere, and particularly of the crust, such that outward flow of middle to lower crust from beneath the high plateau accelerated some time near 15 Ma [e.g., *Clark and Royden*, 2000; *Royden*, 1996]. We recognize that these processes are not mutually exclusive and may work together to produce the growth history of Tibet.

[27] As India has penetrated Eurasia, the Eurasian crust has thickened, and for most plausible lithospheric structures, such thickening would increase gravitational potential energy per unit area stored within high terrain. Hence, as such areas rise higher, an increasing amount of work (per unit area) must be done to raise surfaces yet higher [e.g., *Molnar and Lyon-Caen*, 1988; *Molnar and Tapponnier*, 1978], whereas surrounding lowland areas must face rising compressive deviatoric stresses for convergence to continue unabated. Consequently, surrounding areas deform, and the extent of the highest terrain expands outward. Although northern and perhaps also northeastern Tibet began to undergo crustal shortening shortly after collision, we imagine that the areal extent of the highest part of the plateau may have been both relatively small shortly after collision and confined to its southern margin. As noted above, before collision, southern Tibet may have been as high as it is today, for it had already undergone substantial north-south crustal shortening. The areal extent of the high plateau would subsequently have grown larger as time elapsed. High rates of present-day deformation in the Hexi corridor along the northeastern margin of Tibet (Figure 4) [*Champagnac et al*., 2010; *Hetzel et al*., 2002, 2004; *Tapponnier et al*., 1990] and relatively recent onsets of deformation [*Zheng et al*., 2013b] imply that northeastward growth of deforming terrain persists today [*Métivier et al*., 1998; *Zheng et al*., 2013a]. Thus, the deformation in the past 15 Ma has been, at least partly, the inevitable result of an outward growth of high terrain in Asia, both on the margins of the Tibetan Plateau and exterior to it. At the same time, this postcollision northeastward expansion of deformation has been relatively minor (~150 km) and largely since ~15 Ma (Figure 4).

[28] Removal of some or all of the mantle lithosphere beneath part or all of the plateau offers a second process that could lead not only to accelerated deformation on its flanks and a change in the orientation of deformation on the margins but also to crustal thinning within the high Tibetan Plateau [e.g., *England and Houseman*, 1989; *Harrison et al*., 1992; *Molnar et al*., 1993]. Such removal adds potential energy to the plateau, which can then power deformation of the surroundings. If mantle lithosphere were removed from only part of the plateau, it would be logical to infer that part to be northern Tibet. Relatively young basaltic volcanism is abundant in northern Tibet, and its composition suggests removal of mantle lithosphere [e.g., *Holbig and Grove*, 2008; *Turner et al*., 1993, 1996]. Crust is thinner in northern than southern Tibet [*Tseng et al*., 2009], and seismological studies of various kinds suggest lower speeds and higher attenuation of *P* and *S* waves in the upper mantle there than in southern Tibet [e.g., *Brandon and Romanowicz*, 1986; *McNamara et al*., 1997; *Molnar and Chen*, 1984; *Ni and Barazangi*, 1983; *Shapiro and Ritzwoller*, 2002], as if a hot, low-density uppermost mantle isostatically compensates for the thinner, if still thick crust, of northern Tibet. In particular, removal of mantle lithosphere, if it occurred, offers a sensible explanation for the change in orientation of deformation in northeastern Tibet. Thus, one might say that the history of crustal deformation in northeastern Tibet provides a test that removal of mantle lithosphere passes. Among us, some favor this explanation, but what it needs is an independent test, such as a demonstration that the high part of northern Tibet rose 1000 m or more since 15 Ma.

[29] A third process that might be responsible for the eastward growth of high topography, in the absence of significant crustal shortening, is the outward flow of weakened middle to lower crust. Weakening of the Tibetan middle or lower crustal layer could allow for about 10% of the Tibetan Plateau's modern crustal mass to flow to the east [*Clark and Royden*, 2000; *Royden et al*., 1997, 2008]. Enhanced radiogenic heat production in thickened crust could warm and weaken it. In addition, replacement of cold dense mantle lithosphere by warm asthenosphere could increase the basal heat flux, which could also contribute to thermal weakening of the crust. Eastward flow of midcrustal material may be conceptualized by a flow in a low-viscosity channel in the middle to lower crust [e.g., *Clark and Royden*, 2000]. Pressure gradients associated with laterally varying mean elevations would force lateral flow such that the flux of material would be proportional to the pressure gradient and to the cube of the channel thickness and inversely proportional to the viscosity of material in the channel. Such channel flow provides an explanation for the apparent absence of obvious, large amounts of crustal shortening in much of eastern Tibet, coupled with seismological observations that the crust is thick beneath that part of the plateau [e.g., *Yang et al*., 2012; *Yue et al*., 2012]. A demonstration of such channel flow awaits a definitive seismological test, such as a predictable alignment of crustal minerals revealed by seismic anisotropy.

[30] Among us, no one doubts the occurrence of lateral flow of material in the middle to lower crust, for large displacements on normal faults require such flow [e.g., *Block and Royden*, 1990; *Buck*, 1988; *McKenzie et al.*, 2000]. We all agree that in some of northeast Tibet, such as beneath the Qilian Shan where thrust faulting is widespread in the upper crust [e.g., *Meyer et al.*, 1998; *Tapponnier et al.*, 1990], lower crustal rock need not have moved tens of kilometers with respect to rock in the upper crust. We differ, however, on how extensive, both in areal extent and in magnitudes of displacement [e.g., *Lease et al.*, 2012b], such flow may have been across Tibet and on what the role of such flow has been in the reorientation of deformation that seems to have occurred in northeastern Tibet since ~15 Ma. In any case, both the acceleration of deformation and its change in orientation since 15 Ma provide a test that channel flow passes.

6. Conclusions

[31] The observations summarized here call attention to two aspects of Tibet's development that surely bear on how continental lithosphere deforms and mountain belts grow.

[32] First, virtually the entire plateau became active and started to grow near the time of collision. Although the loci of the highest rates of crustal shortening and of the highest terrain may have migrated across the plateau since collision, there seems little doubt that an outwardly expanding, grand-scale fold-and-thrust belt provides a poor analogy for Tibet's growth. Rather, the dimensions of the Tibetan Plateau seem to have been set by lithospheric strength; relatively weak regions deformed, while stronger surroundings did not. As the plateau narrowed, plate convergence slowed such that the average strain rate across Tibet has remained constant and equal to the modern geodetic value. *Clark* [2012] interpreted this constant average strain rate to indicate constant forcing of Tibet's margins toward one another, possibly implying that the viscous resistance of the continental lithosphere contributes to plate motion.

[33] Second, results obtained over the past decade augment evidence for a tectonic event at or since ~15 Ma and call for a less tightly dated change in orientation of deformation in northeastern Tibet since ~20 Ma. Although the majority of dates indicate an acceleration, if not an initiation, of tectonic activity close to this time, the span of individual estimates ranges from ~15 to 5 Ma, and large-scale simultaneity at any time in this interval can be ruled out. The breadth of this time range may be also due to the preservation of relevant datable rock or to uncertainties among different proxies for faulting or regional elevation gain. For now, many of us favor the idea of an event near or since ~15 Ma, but we leave open the possibility of earlier, and more gradual, changes in both style and orientation as future studies contribute new ages. The broadening of age estimates calls into question whether this shift was abrupt (spanning only a few Myr) or protracted (possibly including the latter quarter if not half of the Tibetan Plateau history). Changes in style and orientations of deformation, however, signal a shift in the geodynamics (including both boundary conditions and rheology) responsible for deformation within the Tibetan Plateau and on its margins. Some of us think that this change could result from gradual thickening and heating of the Tibetan crust, and others favor a more abrupt change in elevation due to removal of some, if not all, of the mantle lithosphere. In either case, the Tibetan Plateau would have reached some limiting elevation since ~15 Ma. The stresses applied to the margins of the plateau by converging plates obviously must do work against dissipative processes to deform rock. In general, they must also do work against gravity, given that raising the surface increases gravitational potential energy. Insofar as thickened mantle lithosphere does not absorb that increased potential energy, as *Clark* [2012] posited, when the surface becomes sufficiently high, those stresses can raise the surface no higher [e.g., *England and Houseman*, 1989; *Molnar and Lyon-Caen*, 1988].

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