

# Orographic Controls on Climate and Paleoclimate of Asia: Thermal and Mechanical Roles for the Tibetan Plateau

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## Key Words

geodynamics, East Asian monsoon, South Asian monsoon, atmospheric dynamics, climate-tectonics interaction

## Abstract

Prevailing opinion assigns the Tibetan Plateau a crucial role in shaping Asian climate, primarily by heating of the atmosphere over Tibet during spring and summer. Accordingly, the growth of the plateau in geologic time should have written a signature on Asian paleoclimate. Recent work on Asian climate, however, challenges some of these views. The high Tibetan Plateau may affect the South Asian monsoon less by heating the overlying atmosphere than by simply acting as an obstacle to southward flow of cool, dry air. The East Asian “monsoon” seems to share little in common with most monsoons, and its dynamics may be affected most by Tibet’s lying in the path of the subtropical jet stream. Although the growing plateau surely altered Asian climate during Cenozoic time, the emerging view of its role in present-day climate opens new challenges for interpreting observations of both paleoclimate and modern climate.

## 1. INTRODUCTION

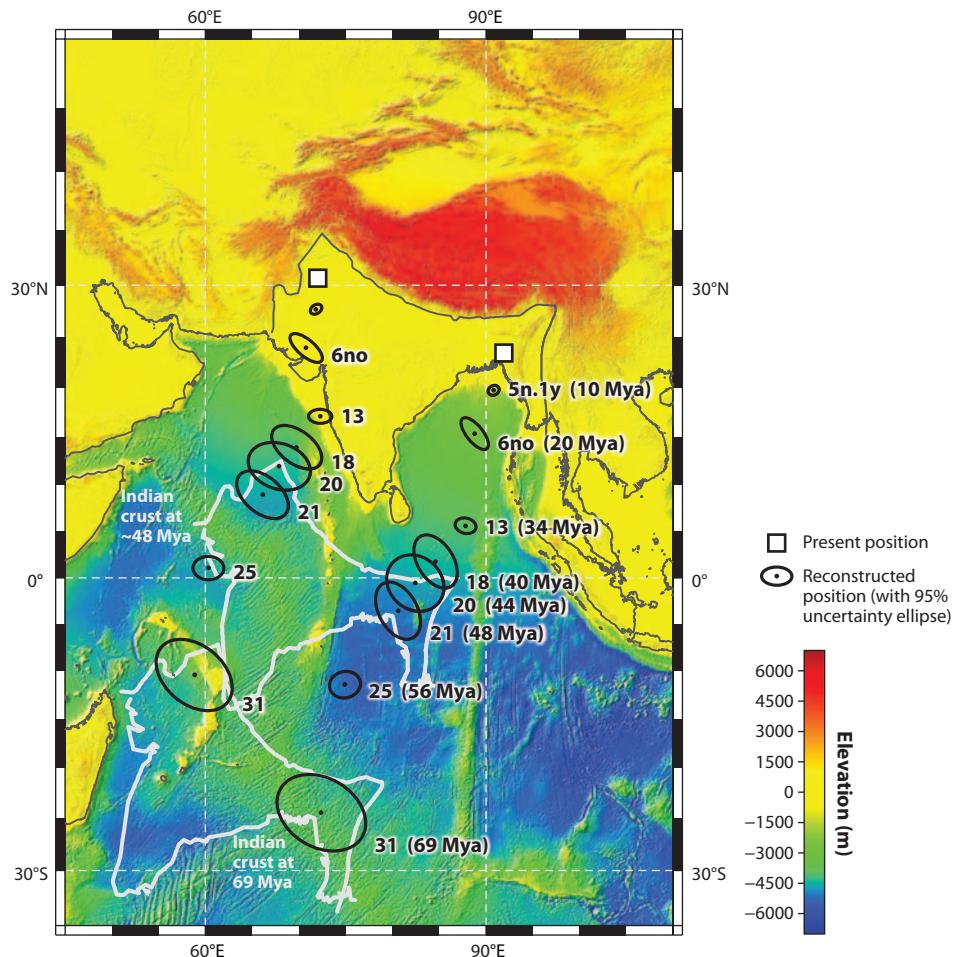
Since ~50 Mya, when India collided with southern Eurasia, the area to the immediate north has grown into the high, wide Tibetan Plateau. The obvious impact of this high, wide plateau on northern-hemisphere atmospheric circulation (e.g., Held 1983, Manabe & Terpstra 1974) and the close association of Tibet with the South Asian monsoon—the prototype for Earth’s monsoons—make the growth of Tibet a prominent evolving boundary condition for the geologic history of eastern Asia, if not global, climate. A high, laterally extensive terrain such as Tibet can interact with the atmosphere in two simple ways: It poses a physical obstacle to flow, and the heating of its surface alters the temperature structure and hence the pressure field of the atmosphere immediately above it. Moreover, these two interactions can work together or separately. Not surprisingly, interpretations of time series of proxies for paleoclimate in Asia typically implicate changes in monsoons, and with them a link to the growth of Tibet. Recent research, however, has cast Tibet’s role in Asian climate somewhat differently from prevailing views. Therefore, the time seems ripe to review how Tibet has grown, how its height and extent might affect the strength and duration of monsoons, and how paleoclimatic proxies might be sensitive to a changing monsoon and a growing Tibetan Plateau. First, despite much progress, the history of Tibet’s upward and outward growth remains controversial. Second, the emerging view of the South Asian monsoon places more emphasis on Tibet as a mechanical obstruction to circulation and less on Tibet as a heat source that drives a nonlocal circulation. Third, Tibet seems to affect precipitation in eastern China through atmospheric dynamics associated with its perturbation to the mid-latitude jet (jet stream). Fourth, the relationship of paleoclimate proxies to Tibet’s growth in turn may require a new perspective. These points are discussed below.

## 2. THE TOPOGRAPHIC HISTORY OF THE TIBETAN PLATEAU

The current high, wide Tibetan Plateau owes its existence to the north-northeastward penetration of the Indian subcontinent into the Eurasian continent (**Figure 1**) and to the buoyancy of continental crust. Underlain by relatively strong lithosphere, India has behaved as part of a nearly rigid object that indents the weaker deformable lithosphere of most of southern Eurasia. As India penetrated into Eurasia, and as crust of both southern Eurasia and the northern edge of India thickened, high terrain developed.

We can calculate the relative positions of India and Eurasia at times in the past using the known histories of separation between India and Africa, Africa and North America, and North America and Eurasia (**Figure 1**). Opinions concerning the timing of the India-Eurasia collision vary, but most researchers agree that the last oceanic crust vanished between 55 and 45 Mya (e.g., Garzanti & Van Haver 1988, Jain et al. 2009, Zhu et al. 2005). India’s rate of convergence with Eurasia decreased markedly near 45 Mya, consistent with the buoyancy of Indian crust resisting continued subduction of Indian lithosphere but not stopping India’s convergence with Eurasia.

A collision ~45–55 Mya calls attention to two aspects of Tibet’s history that are relevant to its effect on climate. First, before India collided with Eurasia, widespread folding and thrust faulting in southern Tibet suggest that Eurasia’s southern margin was high (perhaps ~4000 m), as in the present-day Central Andes (e.g., Burg & Chen 1984, England & Searle 1986, Kapp et al. 2003, Murphy et al. 1997, Volkmer et al. 2007), although defining the width and height of that precollision (“Andean”) margin remains a challenge. Second, India lay 2500–3000 km south of its present position when it collided with Eurasia (**Figure 1**). India’s northward movement into Eurasia has been absorbed by the north-south contraction of both (a) Indian crust to build the Himalaya and (b) Eurasian crust to build Tibet and high terrain farther north. Intact Indian crust



**Figure 1**

Map showing present and reconstructed positions of two points on the India plate with respect to the Eurasia plate at different times in the past. Numbers next to points identify chronos and corresponding ages. An outline of the Indian coastline and the northern edge of present-day, intact Indian crust is also reconstructed to its positions ~48 Mya, close to the time of collision, and 69 Mya, before the collision. (Modified from Molnar & Stock 2009.)

currently underthrusts the Himalaya and southern Tibet at ~15–20 km per Ma (Feldl & Bilham 2006, Lavé & Avouac 2000); if that rate were to apply to the entire period since collision, say, 50 Mya, then north-south contraction of Indian crust would account for 750–1000 km of that 2500–3000 km. Hence, 50 Mya southern Tibet would have lain ~1500–2250 km south of its present position relative to Siberia. Paleomagnetic measurements from southern Tibet corroborate this arithmetic and suggest  $1700 \pm 610$  km of convergence between southern Tibet and Siberia (Chen et al. 1993). Thus, 45–55 Mya, high terrain (~4000 m) like that of the present-day central Andes, thousands of kilometers long and perhaps 200 to 400 km wide, lay at 10–20°N ( $\pm 5^\circ$ ) along the southern margin of Eurasia, when northern India made contact with that margin.

Shortly after collision, deformation occurred throughout much of the region that is now part of the Tibetan Plateau, as shown by numerous studies from different parts of northern Tibet (see

Dayem et al. 2009 for a review). We cannot quantify the elevation history, but it seems likely that ~30 Mya a huge area north of Asia's precollisional southern margin extended from 20–25°N to nearly 40°N with a mean elevation perhaps as high as 1000 m. This initial widespread deformation can be explained by two factors: indentation by a wide indenter (India) whose penetration affects a region far from its edge (e.g., England et al. 1985), and the presence of strong objects such as the Tarim Basin north of the plateau that localize deformation adjacent to them (Dayem et al. 2009).

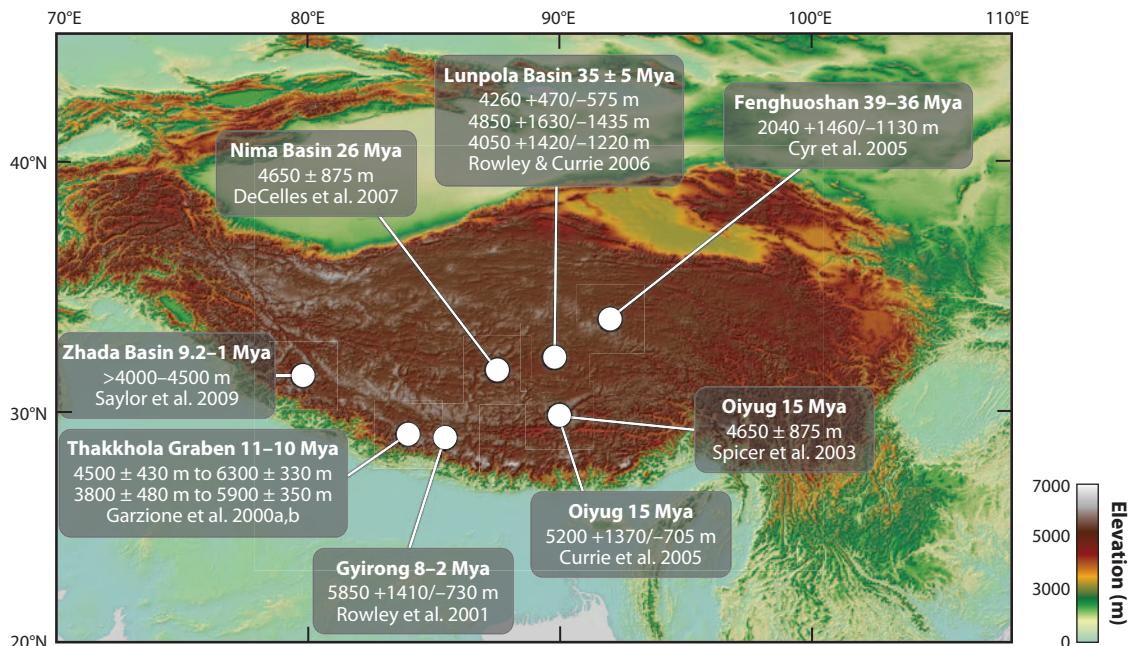
Most imagine a subsequent, progressive northward expansion of high terrain (e.g., England & Houseman 1986, Tapponnier et al. 2001), but despite enormous progress in the past decade, investigators have found few constraints on the growth of the plateau. Estimates of north-south contraction differ (e.g., Coward et al. 1988; Horton et al. 2002; Spurlin et al. 2005; C-S Wang et al. 2002, 2008), and most apply to a part of Tibet too small to provide quantitative bounds on the total north-south crustal contraction. Nevertheless, most agree that such contraction and thickening of crust accounts for its large thickness beneath the plateau, and by isostasy, for most of Tibet's current high elevation. Moreover, several studies suggest that the northeastern and eastern margins of the plateau have risen to their present-day heights since ~15–10 Mya (see Molnar & Stock 2009 for a review). This outward growth includes not just the high terrain within the plateau *sensu stricto* but also the high mountain belts farther north, the Tien Shan, Mongolia Altay, Gobi-Altay, and surrounding regions.

None of the work cited above quantifies past elevations of Tibet and surroundings. Cretaceous marine limestone deposited in southern but not northern Tibet (Hennig 1915) requires that southern Tibet lay at sea level ~80 Mya. Although few workers, if any, seem likely to suggest that northern Tibet stood high at that time, we have no evidence to constrain its elevation.

One of the major recent advances in the Earth sciences has been the quantification of paleoelevations. Among seven locations (eight studies) at which quantitative assessments of Tibet's paleoaltimetry have been made (**Figure 2**), most call for paleoelevations comparable with present-day elevations, if not higher by as much as 1000 m (Rowley et al. 2001, Saylor et al. 2009). Uncertainties for all are ~1000 m.

With one exception (Spicer et al. 2003), all estimates are based on concentrations of  $^{18}\text{O}$  ( $\delta^{18}\text{O}$ ) that accumulated in carbonate sediment when it was deposited. When water vapor condenses,  $\text{H}_2^{18}\text{O}$  enters the condensate more readily than  $\text{H}_2^{16}\text{O}$ , and the rate at which it does so depends on temperature. Both observations (e.g., Garzione et al. 2000b, Gonfiantini et al. 2001, Ramesh & Sarin 1992) and theory (e.g., Rowley et al. 2001) show that  $\delta^{18}\text{O}$  should decrease with elevation of terrain above which condensation and precipitation occurs. In regions where vapor travels short distances, condenses, and precipitates with no reevaporation during precipitation or later, Rayleigh distillation provides a model for isotopic fractionation as vapor condenses (e.g., Rowley et al. 2001). Studies of large-scale circulation, however, show Rayleigh distillation to describe isotopic concentrations inaccurately (e.g., Hendricks et al. 2000, Lee et al. 2007, Noone & Simmonds 2002).

The difficulties of predicting  $\delta^{18}\text{O}$  in precipitation, and hence in carbonate sediment, manifest themselves with data from northern Tibet (**Figure 2**). Cyr et al. (2005) reported values of  $\delta^{18}\text{O}$  that, when interpreted using the Rayleigh distillation model of Rowley et al. (2001), yielded a paleoelevation of only ~2000 m. Yet,  $\delta^{18}\text{O}$  in present-day precipitation decreases markedly across Tibet (Quade et al. 2007; Tian et al. 2001, 2003), so that  $\delta^{18}\text{O}$  values inferred by Cyr et al. (2005) for precipitation in northern Tibet 39–36 Mya differ little from those today. DeCelles et al. (2007), in fact, suggested that northern Tibet may have been ~5000 m high at that time. Using modern isotopes, Tian et al. (2001) inferred that precipitation on southern Tibet is derived from the Bay of Bengal and the Brahmaputra River, but water reaching northern Tibet not only originates elsewhere but also is recycled from the land surface. Hence, a basic assumption of Rayleigh distillation is violated.



**Figure 2**

Topographic map of Tibet showing estimates of paleoaltimetry. For each site shown, the authors cited in the figure offered estimates of paleoelevations using samples with approximate ages shown. Where more than one estimate is shown, more than one horizon of sedimentary rock was used. Spicer et al. (2003) based their estimates on sizes and shapes of fossil leaves; all other measurements exploit values of  $\delta^{18}\text{O}$  in carbonate sediment.

The various paleoaltimetric studies suggest that southern Tibet has remained approximately at its present height for as long as we can measure paleoelevations—26 Mya (DeCelles et al. 2007) and  $35 \pm 5$  Mya (Rowley & Currie 2006)—but the elevation history of northern Tibet is hardly constrained at all. If southern Tibet resembled the present-day central Andes before India collided with Eurasia, precollision elevations may have been as high as today's (e.g., Molnar et al. 2006).

The sensible interpretation of the current tectonics of Tibet is that its surface is subsiding slowly, and some studies yield paleoelevations greater than present-day elevations (e.g., Rowley et al. 2001, Saylor et al. 2009). Whereas north-south crustal contraction and thrust faulting thicken crust, which leads to high surface elevations because of isostasy, the present tectonics of Tibet include a combination of normal and strike-slip faulting. Global positioning system (GPS) velocities call for east-west extension of the plateau at a rate roughly twice that of north-south shortening (Zhang et al. 2004), and seismic moment tensors of earthquakes suggest that half of the extension results in crustal thinning (e.g., Molnar & Lyon-Caen 1989). Such normal faulting has been occurring since at least 8 Mya (Harrison et al. 1992, 1995; Pan & Kidd 1992) and perhaps since  $\sim 13$  Mya (Blisniuk et al. 2001) or earlier. Assuming crustal thinning at the current rate since 10 Mya and in a state of isostatic equilibrium leads to a  $\sim 500$ -m lowering of Tibet's surface since that time (Molnar et al. 1993, 2006).

In terms of geodynamics, the switch from crustal thickening to crustal thinning requires a change in the processes that built the plateau. Thick crust with high elevations stores more gravitational potential energy per unit area than thin low crust with a surface at sea level. Thus, to account for the switch from a regime of crustal thickening that creates potential energy per unit

**GPS:** global positioning system

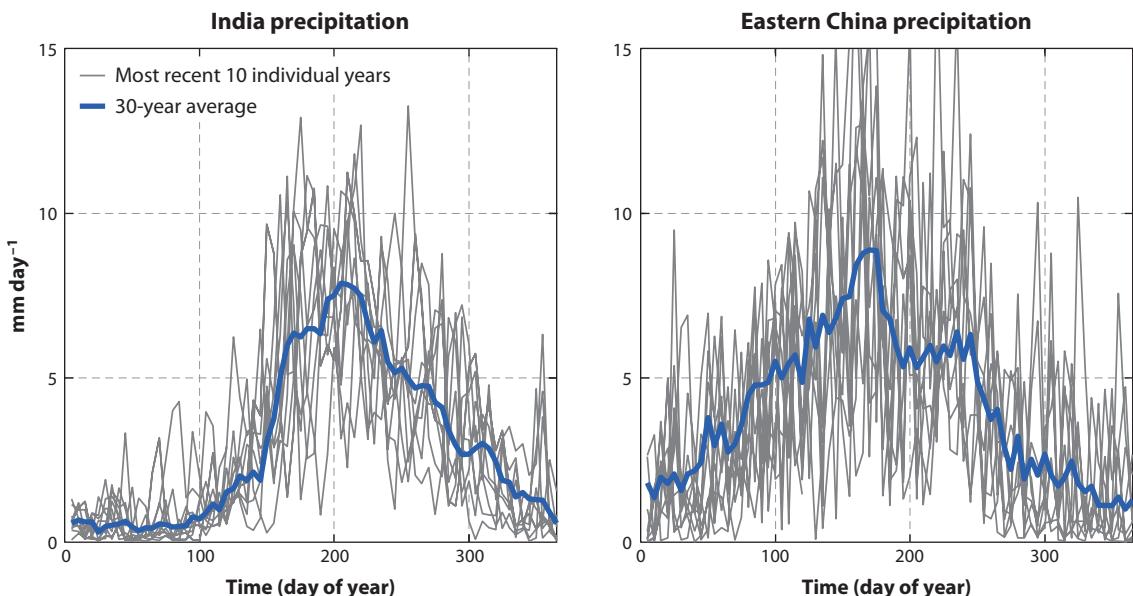
area, equivalent to available potential energy in meteorology (Lorenz 1955), to a regime of crustal thinning that loses potential energy per unit area, a change in the balance of forces must have occurred. England & Houseman (1989) showed that the simplest process causing such a switch is the removal of Tibet's cold, dense mantle lithosphere and its replacement by hotter asthenosphere, which by isostasy leads to an increase in surface elevation of ~1000 m. Although other observations, such as the composition and timing of volcanic rock that erupted in northern Tibet (Turner et al. 1993, 1996) and the change in convergence rate between India and Eurasia since 20 Mya (Molnar & Stock 2009), have been used as arguments in support of removal of mantle lithosphere, many doubt that this process has occurred, and conclusive tests remain to be carried out. Furthermore, given the paleoaltimetric estimates from southern Tibet, if mantle lithosphere were removed, such removal more likely occurred beneath northern Tibet than southern Tibet (if uncertainties in paleoaltimetry permit removal of some mantle lithosphere from southern Tibet).

One region that may have been particularly important in the climate history of eastern Asia is the high terrain in southeastern Tibet (**Figure 2**). Many are convinced that before India collided with Eurasia, the rock in this region lay to its west, south of or within what is now southern Tibet (e.g., Royden et al. 2008; Tapponnier et al. 1986, 1990, 2001). As usual, however, the elevation history is essentially unconstrained. Recent incision of deep river valleys in this area, where relatively flat surfaces characterize interfluves, suggests that this area rose thousands of meters since incision began 13–9 Mya (Clark et al. 2005, 2006). Moreover, the present-day velocity field measured with GPS shows that material moves southward relative to both Siberia and South China, as India penetrates northward into Eurasia (e.g., Zhang et al. 2004). The common image includes material of southern Tibet being extruded around the northeast corner of India as it penetrates into Eurasia. Hence, this flow of material in eastern Tibet suggests—but does not require—a southward growth of high terrain, relative to both Eurasia and South China.

In summary, the simple history of Tibet includes the likelihood of terrain near sea level ~80 Mya, but before ~50 Mya southern Tibet growing into belt of mountains, resembling the central Andes, that was 200–400 km wide and perhaps ~4000 m high, stretching from roughly 10°N ( $\pm 5^\circ$ ) at ~90°E to 20°N ( $\pm 5^\circ$ ) at ~70°E. The construction of the Himalaya, which is underlain by slices of India's northern margin that have been stacked atop one another, attests to tens of millions of years of such underthrusting. Conceivably, this slicing and north-south contraction of northern India's crust has accommodated as much as 1000 km of India's convergence with Eurasia. Roughly 1500–2000 km of that convergence has been accommodated by southern Tibet's moving northward relative to Siberia. Apparently soon after collision, most if not all of the territory between southern Eurasia's Andean margin and the present-day northern edge of Tibet underwent north-south contraction, so that the region reached a low but not insignificant altitude of ~1000 m. As India penetrated deeper into Eurasia, presumably the high part of Tibet (as high as 3000–4000 m) increased in width, while the region to its north (but south of ~40°N) decreased in width. Finally, during the past ~15 Ma, the plateau has grown outward on its northeastern and eastern sides, northern Tibet may have risen ~1000 m higher, and the high terrain to the north in the Tien Shan and Mongolia has developed largely in this period.

### 3. AN OVERVIEW OF THE SOUTH ASIAN AND EAST ASIAN MONSOONS

Climate in eastern and southern Asia is dominated by the marked seasonal cycle with relatively dry conditions in winter and heavy rain in late spring/early summer (the East Asian monsoon) and summer (the South Asian monsoon) (**Figure 3**). The South Asian monsoon includes the regional monsoons of India, the Indochina Peninsula, and the South China Sea, which serve as classical



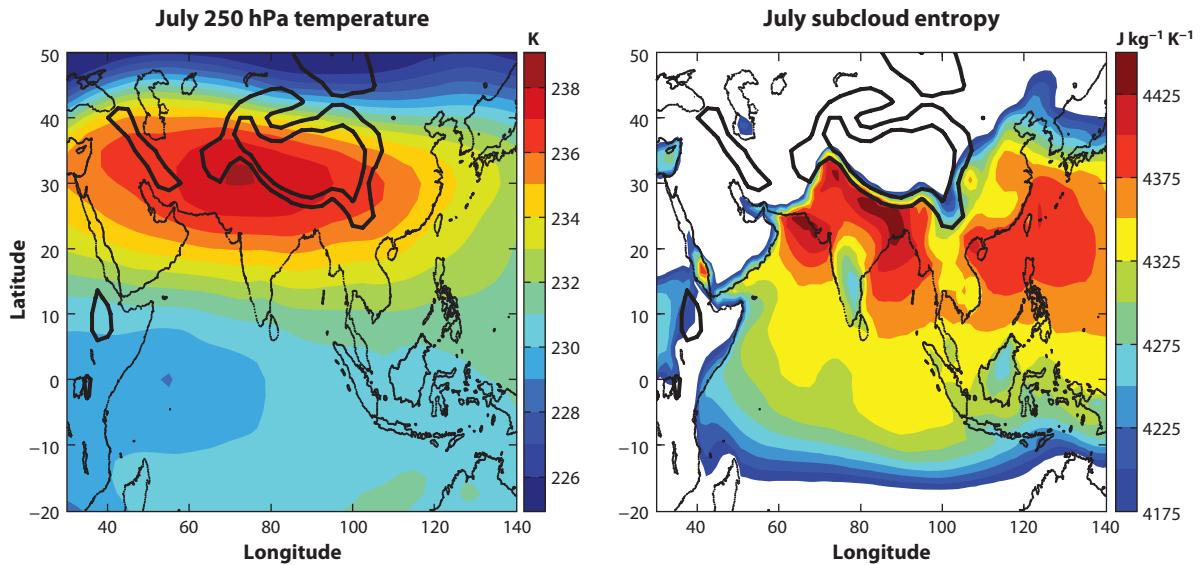
**Figure 3**

Annual cycles of precipitation for India,  $7.5^{\circ}$ – $30^{\circ}$ N,  $70^{\circ}$ – $87.5^{\circ}$ E (*left*), and eastern China,  $22.5^{\circ}$ – $30^{\circ}$ N,  $105^{\circ}$ – $122.5^{\circ}$ E (*right*), using pentad (5-day average) data for 1979–2008 from the Global Precipitation Climatology Project (GPCP). The gray lines show data for the most recent 10 individual years, and the blue lines show the 30-year average climatology. The India time series exhibits an abrupt increase near day 150, and the eastern China time series shows a near-linear increase up to day 175, followed by two step-wise decreases.

tropical monsoon circulations in which rain falls almost entirely in an intertropical convergence zone (ITCZ) displaced from the equator (e.g., Gadgil 2003). The East Asian “monsoon,” however, which encompasses the climates of China, the Korean Peninsula, and Japan, is distinctly extratropical in nature; precipitation and winds are associated with frontal systems and the jet stream. Thus “monsoon” is somewhat of a misnomer, but we use the term below and distinguish it with quotation marks.

The South Asian monsoon displays the following anatomy. Near the spring equinox, peak precipitation and large-scale ascent is centered over the equatorial islands of Indonesia. As the annual cycle progresses, this peak precipitation migrates northward to the Indochina Peninsula and its adjacent waters, the Bay of Bengal and the South China Sea. Then, by early June the region of large-scale ascent intensifies and expands westward across India. Deep convection throughout the northern Bay of Bengal and inland over India along the southern flank of the Himalaya characterize the peak of the South Asian summer monsoon, with temperature and pressure maxima in the upper troposphere centered in these regions (Figure 4). When the gradient in upper troposphere temperature reverses so that the air is colder on the equator than in the subtropics, balanced easterlies develop aloft over the region between  $\sim 10^{\circ}$ N and  $\sim 20^{\circ}$ N (Figure 5), and the potential for a strong, off-equatorial meridional overturning circulation develops (Plumb & Hou 1992). Thus, the onset of the monsoon correlates with a marked change in the wind shear to a regime with surface westerlies and easterlies aloft. Accordingly, Webster & Yang (1992) suggested that the strength of that wind shear provides a good measure of the strength of the South Asian summer monsoon. Goswami et al. (1999) modified Webster & Yang’s index (1992) to span the region where rain falls most on India and the adjacent Bay of Bengal and found that this modified monsoon index correlated better with monsoon rainfall on India than did the original. Many associate that

**ITCZ:** intertropical convergence zone



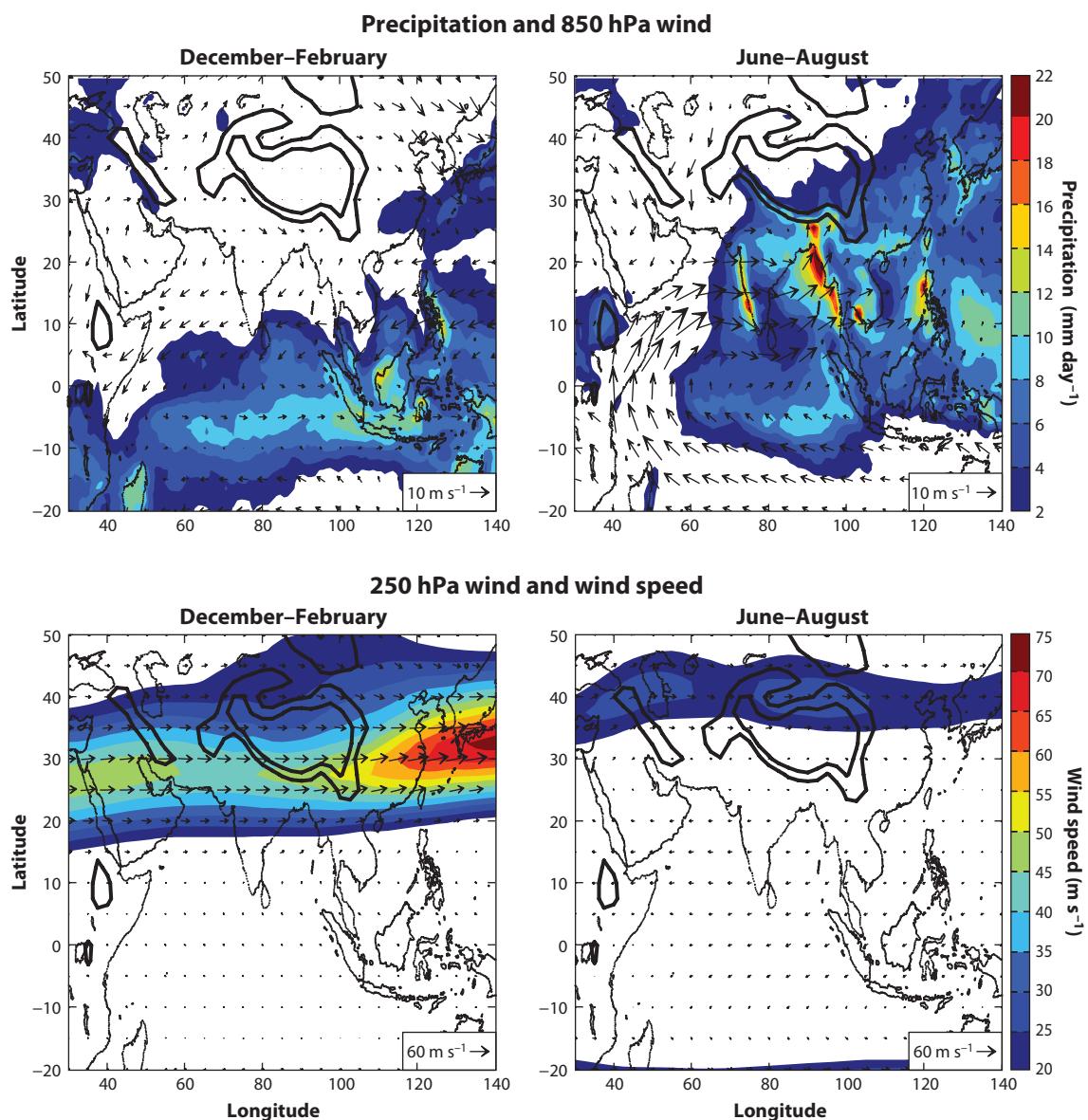
**Figure 4**

(left) Upper-tropospheric temperature (250 hPa) over Asia in July; the maximum overlies northern India and Pakistan, not the Tibetan Plateau. (right) The moist entropy in July on a terrain-following model level within 50 hPa ( $\sim 500$  m) of the surface. All quantities are means for 1979–2002 from the ERA-40 data set (Uppala et al. 2005).

temperature reversal aloft with the onset of the Indian monsoon (He et al. 1987, 2003; Li & Yanai 1996; Wu & Zhang 1998; Yanai & Wu 2006; Yanai et al. 1992).

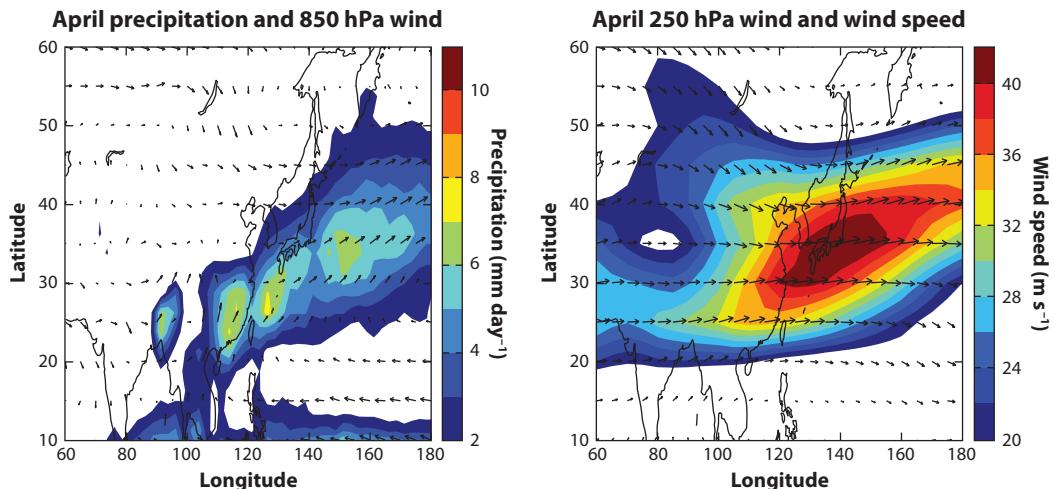
When seen in terms of meridional circulation, the onset of the monsoon circulation is associated with a rapid shift from a classical Hadley circulation, in which marked ascent at the equator is closed by descent in the subtropics, to an asymmetrical flow with ascent in the summer hemisphere, most of the descent in the winter hemisphere, and marked cross-equatorial flow (e.g., Webster et al. 1998). Bordoni & Schneider (2008) further showed that the onset of the Indian monsoon is marked by a transition between regimes in which large-scale atmospheric eddies play markedly different roles in transferring angular momentum.

The South Asian monsoon circulation includes the establishment of the Somali jet, which Findlater (1969) recognized and described. The core of the jet, at an elevation of  $\sim 1500$  m, follows the east coast of Africa (Figure 5), and then swings eastward at  $\sim 10$ – $15^\circ$ N, crosses the west coast of India where it dumps 2–4 m of summer rain, and continues into the Bay of Bengal. This jet results from northward cross-equatorial flow induced by heating in the summer hemisphere, and its dynamics are analogous to those of western ocean boundary currents (e.g., Anderson 1976; Hart 1977). Air crossing the equator carries negative vorticity (clockwise spin to geologists), whose conservation in the absence of dissipation would cause that air to curve back southward and return to the southern hemisphere. The combination of the high terrain of East Africa and greater friction over the East African land mass than over adjacent ocean imparts a torque (left-lateral simple shear to geologists) that cancels the negative planetary vorticity carried across the equator and allows flow to concentrate in a jet and remain in the northern hemisphere (Rodwell & Hoskins 1995). Like most of the elements of the South Asian monsoon, the wind speed in the Somali jet increases rapidly with monsoon onset (e.g., Fieux & Stommel 1977), and the kinetic energy of zonal flow across the Indian Ocean increases by an order of magnitude commonly within a two-week span in the spring (Boos & Emanuel 2009, Halpern & Woiceshyn 1999, Krishnamurti et al. 1981).



**Figure 5**

Top panels show horizontal wind on the 850 hPa pressure level (arrows) and precipitation for 1998–2006 from the TRMM 3B43V6 data set (shading, with an interval of  $2 \text{ mm day}^{-1}$ ). Bottom panels show horizontal winds on the 250 hPa pressure level (arrows) and wind speeds on the same level (shading, with an interval of  $5 \text{ m s}^{-1}$ ). All winds are from the ERA-40 data set for 1979–2002 (Uppala et al. 2005).



**Figure 6**

(left) The climatological April horizontal wind on the 850 hPa pressure level (arrows) and precipitation from CPC Merged Analysis of Precipitation (CMAP) (shading). (right) The climatological April horizontal wind on the 250 hPa pressure level (arrows) and wind speeds on the same level (shading).

The anatomy of the East Asian “monsoon” is fundamentally different from that of the South Asian monsoon. In the heart of northern-hemisphere winter, persistent, albeit weak, rainfall occurs in southern China (**Figures 3** and **5**). This precipitation is associated with and organized by a nearly stationary, convergent frontal circulation (e.g., Y-S Zhou et al. 2004). As winter gives way to spring, precipitation remains organized in a band known as the Meiyu Front that extends east-northeast over central China and the western Pacific (**Figure 6**). The front migrates northward and precipitation reaches a maximum in mid-June (**Figure 3**) over the Yangtze River basin of central China and eastward to Japan. Then the front breaks down during a relatively rapid transition to summer circulation, and precipitation becomes highly irregular—if intense—when it happens. The migration and evolution of these fronts are closely associated with seasonal changes in the East Asian jet stream (Liang & Wang 1998). Hence, despite the moniker “monsoon,” the anatomy of the spring and summer rains over East Asia do not conform to the common definition of a summer monsoon.

#### 4. DYNAMICS OF THE SOUTH ASIAN MONSOON

Heating of the atmosphere over the Tibetan Plateau has long been held to drive the Indian summer monsoon and strongly influence the broader-scale South Asian summer monsoon (e.g., Flohn 1968, Yanai & Wu 2006). A high-amplitude seasonal cycle in the flux initially of sensible and then of latent heat from the plateau surface into the mid-troposphere evolves concurrently with the South Asian monsoon circulation (e.g., Li & Yanai 1996, Luo & Yanai 1984). Recently, however, some observational and theoretical work has raised questions about the importance of the plateau’s heating for South Asian climate.

The Tibetan Plateau does not directly provide the dominant heat source for the South Asian summer monsoon, as the warmest upper-tropospheric air lies over northern India just southwest of, rather than over, the plateau (**Figure 4**). The largest diabatic heat source, inferred from reanalyzed atmospheric data, lies in the region of intense cumulus convection just south of the plateau

(e.g., Yanai & Wu 2006). A forcing of the summer monsoon by the plateau is often reconciled with these facts through the assumption that the plateau acts as a sensible heat pump, driving local ascent that causes low-level moist air south of the plateau to converge and to produce latent heating, which in turn drives the large-scale monsoon flow (Yanai & Wu 2006). Thus, the question of how Tibet alters South Asian climate becomes bound up with the question of how moist cumulus convection interacts with the large-scale flow. It is worth noting that there is an alternative, now widely used in both theoretical work and climate models (e.g., Arakawa 2004; Clift & Plumb 2008, Chapter 1; Emanuel 2007; Neelin 2007), to the idea of low-level moisture convergence causing convection that in turn drives large-scale atmospheric flow.

In this alternate view, moist convection does not act as a heat source for the large-scale circulation; instead, the circulation is correlated with and includes low-level moisture convergence but is not caused by it. Moist convection tightly couples free-tropospheric temperatures to the energy content of air below the base of cumulus clouds, similar to the way that dry convection maintains a dry adiabatic lapse rate with the surface temperature as its lower boundary condition (Emanuel et al. 1994). Free-tropospheric temperatures are said to be in a state of quasi-equilibrium (*a*) with the subcloud entropy  $s_b$ , or (*b*) almost equivalently with the low-level moist static energy  $b_b$ <sup>1</sup> (e.g., Clift & Plumb 2008, Chapter 1; Emanuel 2007; Neelin 2007; Plumb 2007).

Theoretical work has shown that the peak free-tropospheric temperature is expected to lie at the poleward boundary of a thermally direct monsoon circulation (Lindzen & Hou 1988). If cumulus convection simply couples free-tropospheric temperatures to the subcloud entropy, Lindzen & Hou's (1988) condition implies that the peak subcloud entropy (or low-level moist static energy) lies at the same latitude as the poleward boundary of the monsoon circulation, with the maximum ascent rate and precipitation lying slightly equatorward (Emanuel 1995, 2007; Neelin 2007; Privé & Plumb 2007a). In the South Asian summer monsoon (**Figure 4b**) and in idealized atmospheric model runs (e.g., Bordoni & Schneider 2008; Privé & Plumb 2007a,b), maximum free-tropospheric temperatures indeed lie directly over the peak subcloud entropy, which in turn lies over low terrain of northern India, just south of the Himalaya. The relation between the location of maximum  $s_b$ , or  $b_b$ , and precipitation is diagnostic, leaving open the question of how surface heat fluxes, radiation, and convective downdrafts interact to set the  $s_b$  distribution. Nevertheless, the concurrent locations of maximum  $s_b$  or  $b_b$  and maximum free-tropospheric temperatures suggest that the dominant thermal forcing for the summer monsoon occurs over the low regions of northern India, rather than the Tibetan Plateau (**Figure 4**), despite calculations for idealized states of radiative convective equilibrium showing that upper-tropospheric temperatures,  $s_b$ , and  $b_b$  do increase with the elevation of an underlying land surface (Molnar & Emanuel 1999). Those same calculations, however, show that the thermodynamic effect of surface elevation depends also on the albedo and moisture availability of the land surface, and these quantities on the Tibetan Plateau seem to have values different from those of lower regions of South and East Asia (Boos & Emanuel 2009).

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<sup>1</sup>Subcloud moist entropy is given by  $s_b = C_p \ln \theta_e$ , where  $C_p$  is the specific heat at constant pressure and  $\theta_e$  is the equivalent potential temperature of air below the base of cumulus clouds. Potential temperature,  $\theta = T(P_0/P)^{R/C_p}$ , is the temperature (in Kelvins) that a substance would have if it were compressed or decompressed adiabatically to some reference level,  $P_0$ , usually sea level, where  $T$  is the temperature at pressure  $P$ , and  $R$  is the gas constant. The atmosphere in general is stably stratified so that  $\theta$  increases with height. The equivalent potential temperature,  $\theta_e \cong T(P_0/P)^{R/C_p} e^{L_v q / C_p T}$ , allows for the effect of moisture. ( $L_v$  is the latent heat of vaporization, and  $q$  is the water vapor mixing ratio.)  $\theta_e$  is approximately the temperature that a parcel of air containing water vapor would have if it were lifted adiabatically high enough that all vapor condensed and precipitated out, if it warmed in response to the resulting latent heating, and if it were then lowered adiabatically to the reference pressure. Moist static energy, given by  $b = C_p T + L_v q + gZ$ , is the total energy per unit mass in a parcel of air, ignoring negligibly small kinetic energy; the last term is potential energy, with  $g$  gravity and  $Z$  height.

**GCM:** general circulation model  
**SST:** sea-surface temperature

This is not to say that topography has a negligible effect on the South Asian summer monsoon, for the peak precipitation has been shown to weaken dramatically and shift toward the equator when Asia's elevated topography is removed in numerical simulations (e.g., Hahn & Manabe 1975, Kutzbach et al. 1989, Yasunari et al. 2006). One might posit that the Himalaya could produce moisture convergence even in the absence of direct heating by Tibet, but it is difficult to imagine how a mountain range oriented parallel to the zonal flow and positioned poleward of likely meridional flow would produce such convergence. An alternate hypothesis is that the Himalaya, Tibet, and adjacent high terrain prevent cold and dry extratropical air from ventilating the warm and moist tropics (Chou et al. 2001, Privé & Plumb 2007b). Thus elevated topography would alter monsoon thermodynamics indirectly by blocking flow instead of serving as a direct heat source. This idea is consistent with the fact that sharp horizontal gradients in subcloud entropy are coincident with the Himalaya and Hindu Kush (**Figure 4**) (Boos & Emanuel 2009). Furthermore, Chakraborty et al. (2006) found that eliminating mountainous terrain west of the Tibetan Plateau in a general circulation model (GCM) run produced a larger reduction in the intensity of Indian summer rainfall than eliminating the Tibetan Plateau farther east, and they noted that in the absence of all topography, advection of cold air from the extratropics seemed largest in this region west of Tibet. Perhaps more conclusively, Boos & Kuang (2010) showed that winds, precipitation, and the thermodynamic structure of the large-scale South Asian monsoon in a GCM were almost unchanged when the plateau was eliminated, but a narrow Himalaya and the adjacent mountain ranges were preserved. The topographic configurations of Boos & Kuang (2010) and Chakraborty et al. (2006) were chosen not to mimic reality but to shed light on physical mechanisms; the high terrain west of Tibet that is hypothesized to prevent intrusion of cold, dry extratropical air seems to have grown concurrently with the uplift of Tibet's surface, and so could have coupled South Asian climate to the growth of the plateau.

A strong monsoon circulation requires only a strong off-equatorial thermal forcing, which can be obtained even in an atmospheric GCM with slab oceans and no land (often termed an aquaplanet) given a sufficiently high subtropical sea-surface temperature (SST) (e.g., Bordoni & Schneider 2008). Until recently it was thought that surface heat fluxes from the Tibetan Plateau provided this strong off-equatorial thermal forcing, either directly or via a positive feedback with low-level moisture convergence. As discussed above, an emerging alternative view holds that strong northward gradients of  $s_b$  or  $b_b$  and free-tropospheric temperature are maintained by the land and ocean surfaces south of Tibet, with the Himalaya and adjacent mountain ranges protecting this thermal maximum from the cooler and drier extratropics.

Our discussion of the monsoonal response to thermal forcing, regardless of how topography might influence that forcing, has thus far been limited mostly to the spatial structure of the circulation. The intensity and duration of the summer circulation are of comparable importance, as they set the annual amount of precipitation in South Asia. Theoretical work has shown that when the thermal forcing in an idealized monsoon climate becomes sufficiently strong, the circulation will enter a regime where its strength increases nonlinearly with that forcing (e.g., Boos & Emanuel 2008, Bordoni & Schneider 2008, Plumb & Hou 1992, Walker & Schneider 2006). Such theories are typically used to explain the observed threshold response to the insolation forcing, and specifically the abrupt onset of the summer monsoon, as documented using different criteria (e.g., Boos & Emanuel 2009, Fasullo & Webster 2003, He et al. 1987, Krishnamurti et al. 1981, Webster et al. 1998). Molnar et al. (1993) exploited the idea of a transition to a nonlinear circulation regime to suggest that a large increase in the intensity of the South Asian summer monsoon may have occurred on geological timescales when heating over the Tibetan Plateau exceeded a threshold as it grew higher. This hypothesis is predicated on the idea that the plateau provides the dominant thermal forcing for the monsoon circulation, but even if that idea proves to be false, monsoon

intensity could still have changed dramatically on geological timescales as evolving topography altered land-surface albedo, ocean circulation, or the intrusion of cold, dry air from the extratropics. For instance, some studies have reported an inverse relationship between Eurasian snowpack and Indian summer rainfall, but the robustness of the statistical relationship has been debated and the mechanism is not well understood (e.g., Shaman & Tziperman 2005, Yanai & Wu 2006).

The duration of the summer monsoon can also influence annual accumulations of precipitation. Except for Chakraborty et al. (2006), who found that removal of high terrain west of Tibet delayed the onset of the monsoon in their GCM runs, we are unaware of work that addresses how topography might alter duration. Fasullo & Webster (2003) showed a high correlation between rainfall in India and the duration of the monsoon, and both they and Goswami & Xavier (2005) showed that El Niño reduces South Asian rainfall primarily by shortening the length of the summer monsoon rather than by reducing its intensity. Although numerous theories have been proposed to explain why monsoon onset is abrupt, including the aforementioned ones that involve transitions to a nonlinear circulation regime, most of these theories leave the timing of onset open.

Regardless of how intensity and duration interact to produce various measures of summer-integrated monsoon activity, the accumulated precipitation in strong and weak monsoons differs from that in normal ones by only  $\sim$ 10% (e.g., Gadgil 2003, Webster & Fasullo 2003). Moreover, this remarkably regular summer mean rainfall results from a series of active and weak episodes within each summer season that have a high variance relative to the mean (**Figure 3**), which prompted Palmer (1994) to suggest that the small variations in total summer precipitation may result from changes in the number and duration of the active episodes. Although the intraseasonal variance of large-scale dynamical indices seems lower than that of precipitation (e.g., Fasullo & Webster 2003, Goswami & Xavier 2005, Webster & Yang 1992), there is still no commonly accepted theory for the physics of active and weak monsoon episodes (see the review by Goswami 2005), including how their character might depend on topography.

This review focuses mainly on the extreme elevations of Tibet and the Himalaya, but the lower mountains of the Western Ghats on the west coast of India and the Arakan Yoma and Bilauktang ranges on the west coast of the Indochina Peninsula also clearly play a role in the monsoon (Xie et al. 2006). The bulk of precipitation in the South Asian summer monsoon falls in narrow, intense bands on the windward sides of these ranges (**Figure 5**). These mountains are oriented perpendicular to the low-level monsoon westerly winds and so might naively be thought to play a role analogous to that of the Sierra Nevada or Andes in the extratropical westerlies, but there are important differences. First, the zonal monsoon flow is strongly baroclinic with easterlies aloft, rather than westerlies. Second, the latent heating on the windward side of these ranges might be strong enough to alter the mean circulation. It thus seems an open question whether the orographic lifting associated with these modest, north-south oriented ranges merely localizes precipitation within the monsoon domain, or whether it somehow alters the intensity or structure of the mean monsoon flow.

## 5. DYNAMICS OF THE EAST ASIAN “MONSOON”

The anatomy of the East Asian “monsoon,” summarized above, points to the central role that jet-stream dynamics plays in its development and suggests that Tibet also plays an important role by altering the jet stream. The subtropical jet (or jet stream) passes south of Tibet in winter, when wind speeds are highest (**Figure 5**). Flow accelerates east of Tibet, and the jet exits eastern China as a narrow, especially high-speed jet (**Figure 5**; e.g., Harnik & Chang 2004). Speeds reach a maximum between  $\sim$ 120°E and  $\sim$ 160°E (**Figure 6**), where the jet is flanked by attendant high and low pressure south and north of it. The acceleration of flow into this jet maximum is associated

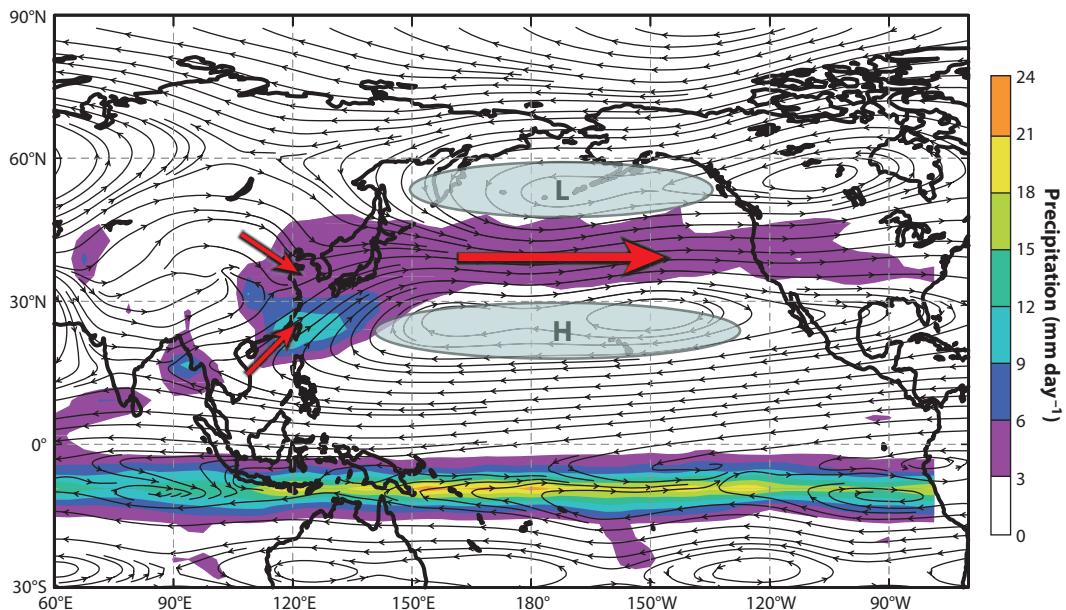
with the Coriolis torque on a meridional circulation transverse to the jet axis, with ascent and precipitation south of the jet-entrance region over eastern China (Liang & Wang 1998).

What role does orography play in the dynamics of the jet stream and its transverse circulations? Idealized models of the boreal winter troposphere have been used to show that the strong jet maximum just east of Asia during winter primarily results from large zonal asymmetries in extratropical heating; orographic forcing plays a secondary role (Chang 2009, Held et al. 2002). These studies, however, focused on the nondivergent circulations of the jet, whereas we are primarily concerned with the divergent (and convergent) flow that is associated with rainfall. Chang (2009) and Held et al. (2002) conducted experiments in which they exploited models forced by prescribed fixed heating, but they emphasized that in the real atmosphere this heating depends on the circulation, which is shaped in part by orography (as well as on latent heating in extratropical eddies and the convergence of heat transported by those eddies). Seager et al. (2002) showed that orography strongly affects the diabatic heating in storm tracks; if orography is removed in a coupled model, the climatological sea-surface temperatures and diabatic heating differ markedly. In the case of Tibet, Held et al. (2002) further noted a significant nonlinear relationship between the diabatic heating and orographic forcing.

To circumvent the problem of disassociating heating forced by orography and by land-sea temperature contrasts, K. Takahashi and one of us (D.S.B.) ran an atmospheric GCM coupled to a slab ocean but with the ocean surface elevated in the region of Tibet to the same height and breadth of the observed plateau. Thus, this model run lacks the strong zonal gradients in heating associated with a cold Asian continent and a warm Pacific Ocean during boreal winter. Nevertheless, it produces a localized storm track east of the plateau (hence, zonally asymmetric heating) and enhanced precipitation southeast of the water-covered plateau; the amplitude and location of this enhanced precipitation resemble those of the precipitation over South China in winter (**Figure 7**). This suggests that Tibet interacts with the jet to produce the zonal heating gradients that Held et al. (2002) and Chang (2009) found to be responsible for the intensity of the nondivergent flow, as well as the divergent circulations that were found to be responsible for precipitation over China. Wu et al. (2007) reported consistent findings from a GCM in which the height of Eurasian topography was varied; they found that spring rainfall in East China vanished when the topographic height was zero but that this rainfall resembled observations for a run with realistic topographic heights. Finally, Rodwell & Hoskins (2001) found that, in a model in which heating and orography were separately specified, orography alone could produce closed subtropical circulations. This model involves a region of ascent sandwiched meridionally between two regions of subsidence on the east side of mountains.

In summary, the winter rain over southeastern China can be seen to result from the jet interacting with the high orography of the Tibetan Plateau (**Figures 6** and **7**). This orography generates a stationary wave that features a local maximum in flow downstream of the plateau and a local, zonally asymmetric region of diabatic heating that is coincident with the local maximum jet speed. In turn, the zonally symmetric diabatic heating shapes the stationary wave generated directly by the orography (Held et al. 2002). The localized jet is maintained in part by eddy-driven transverse circulations. The net effect of the orography—both mechanical effects and the resulting downstream diabatic effects—creates low-level flow in the subtropics that sweeps moisture from the South China Sea into southern China, where low-level convergence and precipitation occur (**Figure 7**).

As time marches from winter to spring, rainfall over southeast China increases in intensity (**Figure 3**), perhaps because the humidity of the converging air that is imported from the warming ocean to the south increases. With the seasonal decrease in the equator-to-pole temperature gradient, the jet moves northward from its winter position south of Tibet to pass directly over the plateau and then north of it (**Figure 5**). In turn, the locus of convergence of moisture and



**Figure 7**

Calculated precipitation rates (*in color*) and lower tropospheric (850 hPa) streamlines superimposed on a map of Earth. A general circulation model (GCM) was coupled to a slab ocean in calculations (an aquaplanet), and included only the high terrain equivalent to that of Tibet. Perpetual equinox insolation was used; all forcings except orography are also symmetric about the equator. East of the high terrain that mimics Tibet, focusing (*converging red arrows*) and intensification of the upper-level jet (*large red arrow across the central Pacific*) set up circulation with lower-level convergence and heavy precipitation colocated where southeastern China is located. (From unpublished work of K. Takahashi and D.S.B.)

precipitation downstream of the plateau, the Meiyu Front, shifts northward into central China. In this view, the intensification and northward movement of the Meiyu Front from late winter to late spring can be seen as a result of (*a*) the jet interacting with the plateau and (*b*) the increasing humidity of air that is swept in from the south over a warming ocean. With the onset of the South Asian monsoon in early June, humid air is also transported over the Indo-Burman ranges and southeastern Tibet into southeast China (Y-S Zhou et al. 2004), contributing to the increase in precipitation at this time. Then approximately when the core of the jet stream moves northward to pass north of Tibet (Figure 5), the Meiyu Front disintegrates, and precipitation over China decreases (Figure 3).

The demise of the Meiyu Front as the locus of convection and intense precipitation over China in late June is inconsistent with the simple traditional model that equates monsoonal flow with increasing land-sea temperature gradients. Viewed in a more global context, the strong meridional temperature gradient in the extratropics weakens as spring moves into summer, causing the jet to weaken and shift to higher latitudes where the Tibetan Plateau affects it less. This might explain the rather sudden demise of the Meiyu Front in late June and the transition from nearly daily to more sporadic (yet intense) precipitation and overall drying despite the continuing increase in the land-ocean temperature contrast.

The view that we express here treats Tibet as an obstacle to flow aloft, and might suggest that heating over the plateau is secondary. Indeed, we hold that Tibet's principal effect results simply from its height, but some studies demonstrate that heating over the plateau directly affects

East Asian climate. For example, B. Wang et al. (2008) inferred that marked twentieth-century warming over Tibet ( $1.8^{\circ}\text{C}$  in 50 years) can account for enhanced rainfall in the Meiyu Front during this interval.

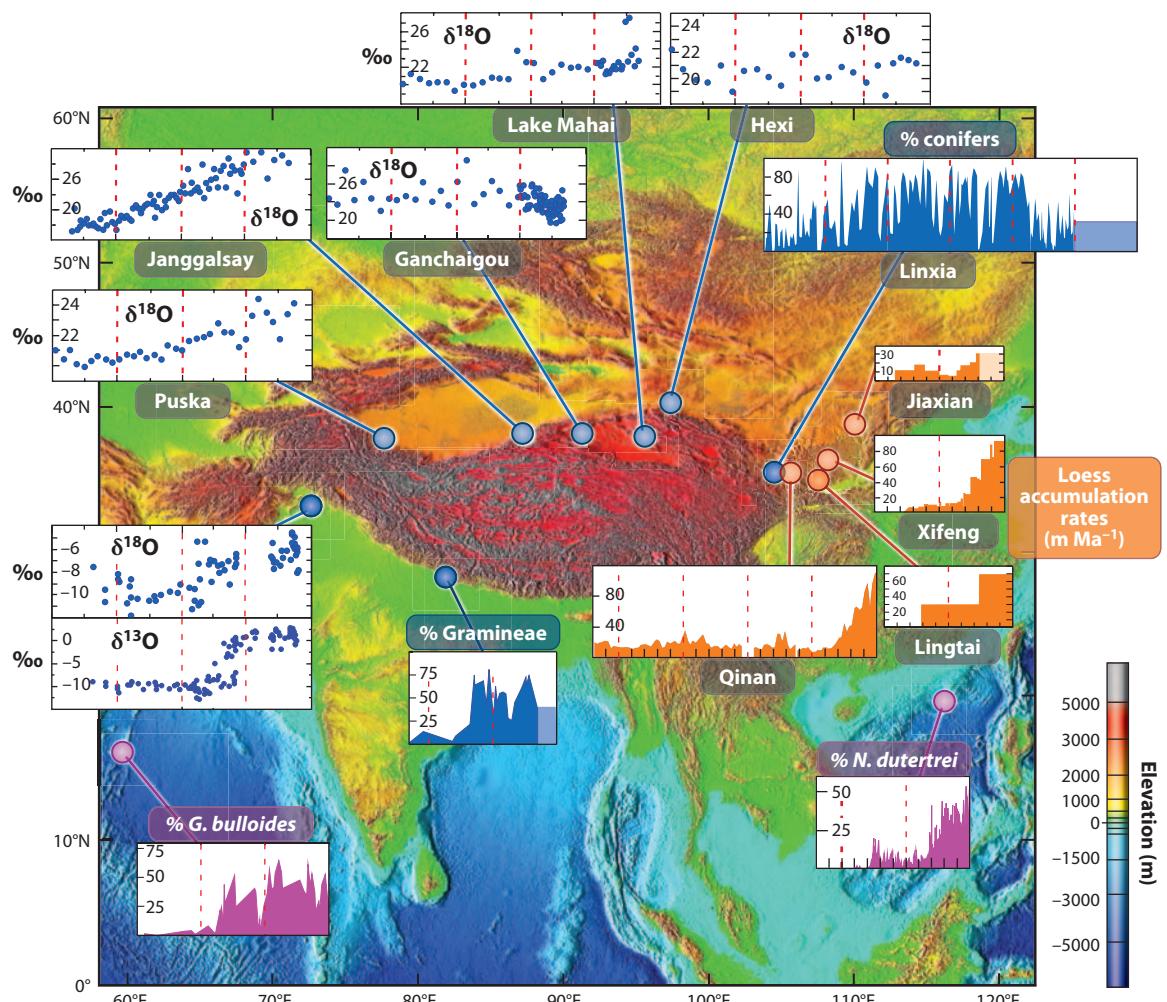
## 6. PALEOCLIMATES OF SOUTHERN AND EASTERN ASIA

Because of unusually good paleoclimate records from a continental region, eastern Asia has become a focus for paleoclimate study. Much recent work concerns the past few hundred thousand years, during which the topography of Asia has undergone negligible change. Thus, these records offer little information about how growing topography might have affected regional climate, and despite their tantalizing signals we ignore them here. On a million-year timescale, two times of climate change stand out. As is the case with paleoclimate records throughout Earth, nearly all proxies show a change between  $\sim 4$  and  $2.5$  Mya, when northern-hemisphere cooling accelerated and the growth and decay of ice sheets became recurring events in Canada and Fennoscandia. Numerous studies have ascribed paleoclimatic changes in Asia near 3 Mya to an alleged rise of Tibet at that time. We consider this association to have no foundation (e.g., Molnar 2005), for it would require either crustal thickening at rates disallowed by geologic observations or processes (*deus ex machina*) for which we have no evidence. Consequently, we refrain from discussing evidence for climate change at that time. Many paleoclimate records also show changes near 10 Mya, which we do discuss, because they may be associated with changes in topography.

Quade et al. (1989) reported abrupt changes in  $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$  values in pedogenic carbonates from northern Pakistan between  $\sim 10$  and  $\sim 6$  Mya, and they suggested that these changes marked a strengthening of the South Asian monsoon (Figure 8). Much subsequent work (see Molnar 2005 for a summary) has corroborated these isotopic changes. A roughly concurrent decline of large browsing mammals and an increase in grazers (Barry et al. 1985, 1995; Flynn & Jacobs 1982) support palynological and other observations that suggest a thinning of forests and an expansion of grasslands along the Himalaya (e.g., Garzione et al. 2003, Hoorn et al. 2000, Komomatsu 1997, Prasad 1993). In addition, changes in the architecture of sediment deposited on the plains of northern India call attention to increased flooding near 10 Mya, which DeCelles et al. (1998) and Nakayama & Ulak (1999) associate with increased frequency or magnitudes of strong monsoon rains. Although the interpretation of the isotopic shifts has become more complicated (e.g., Cerling et al. 1997, Dettman et al. 2001, Nelson 2005), the combination of isotopic shifts, expansion of grassland at the expense of forests, and suggestions of increased flooding concur with a shift near 10 Mya in South Asian climate toward seasonal aridity, like today's monsoonal climate.

One of the strongest arguments for a strengthening of the Indian monsoon near 10 Mya came from deep-ocean drilling in the western Arabian Sea. During both summer and winter monsoons, steady winds induce upwelling of cold, deep, nutrient-rich water. Modern studies of plankton from this region show that *Globigerina bulloides*, a planktonic foraminifer that thrives in cold water where nutrients upwell, becomes abundant during monsoons, particularly in summer, and dominates organic sediment accumulation (Curry et al. 1992, Prell & Curry 1981). *G. bulloides* evolved near 14 Mya, and for a few million years, it comprised only few percent of the annual mass accumulation of organic sediment in the western Arabian Sea. Then beginning near  $\sim 10$  Mya (using modern timescales), it suddenly comprised tens of percent of organic sediment deposition (Figure 8), suggesting a strengthening of monsoon winds at that time (An et al. 2001, Kroon et al. 1991, Prell et al. 1992).

Recently, however, some of the same authors have called this inference into question. First, by using only foraminifera of large sizes, Huang et al. (2007) found only a modest change in the fraction of *G. bulloides* near 10 Mya; second, they reported little change in estimates of sea-surface



**Figure 8**

Time series of paleoclimate from the region surrounding the Tibetan Plateau superimposed on a map of eastern Asia. In all cases, time series are up to 20 Ma in length with the present on the right, and the red dashed lines show 5-Ma intervals. The *G. bulloides* record from the Arabian Sea is from Kroon et al. (1991); the  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  values, in ‰, from Pakistan are from Quade & Cerling (1995) and Quade et al. (1989, 1992, 1995); percentage of Gramineae pollen is from Hoorn et al. (2000); and percentage of *N. duterrei* is from P. Wang et al. (2003). Loess accumulation rates are as follows: Lingtai from Ding et al. (1999), Xifeng from Sun et al. (1998), Jiaxian from Qiang et al. (2001), and Qinan from Guo et al. (2002). The percentage of conifers from Linxia is from Ma et al. (1998), and the five  $\delta^{18}\text{O}$  records from northern Tibet are from Kent-Corson et al. (2009).

temperatures based on the  $\text{U}_{37}^{\text{K}'}$  alkenone index in the western Arabian Sea. Peterson et al. (1992) showed that the accumulation rate of carbonate sediment at shallow depths and at low latitudes in the Indian Ocean increased abruptly near ~10 Mya; thus relatively small foraminifera should be better preserved since that time than before, possibly making them a biased record of productivity. Moreover, if monsoon winds did strengthen ~10 Mya, we might expect sea-surface temperatures to drop at that time, but the results of Huang et al. (2007) show no such change. If, however, the thermocline were deeper ~10 Mya than it is today, as Philander & Fedorov (2003) suggested, cold water might not have upwelled during monsoons. Thus, the results of Huang et al. (2007) do not

rule out a strengthening of the monsoon near 10 Mya, but they require that excuses be found for what appear to be inconsistencies.

Paleoceanographic data obtained from ocean drilling sites in the South China Sea suggest a shoaling of the thermocline near ~10–8 Mya and a more marked shoaling near 3 Mya (**Figure 8**) presumably because of stronger winds and deeper mixing (Li et al. 2004, P. Wang et al. 2003). As winds are strongest in winter in the modern climate, P. Wang et al. (2003) inferred that winter winds strengthened ~8 Mya.

Thick sequences of loess, wind-blown sediment, in northern China provide unusually complete time series of climatic variability within a continental setting (**Figure 8**). Although loess deposition clearly has occurred since 22 Mya (Guo et al. 2002), and perhaps since before 29 Mya (Garzione et al. 2005), several studies from widely separated places in the eastern part of China's Loess Plateau demonstrate initial accumulation near 10–8 Mya (Ding et al. 1999, Qiang et al. 2001, Sun et al. 1998). As loess deposition began earlier at sites farther west, it is easy to infer that the outward growth of the area affected by loess has responded in some way to the outward growth of Tibet.

On the northeastern margin of Tibet, Ma et al. (1998) reported a marked change in pollen near 9 Mya, with grass pollen becoming dominant at that time (**Figure 8**). They inferred that the region became more arid.

From measurements of  $\delta^{18}\text{O}$  from sediment deposited at numerous localities in and near northern Tibet, Kent-Corson et al. (2009) found a variety of time series. Some of these include no obvious change during the past 20 Ma, some show a continuous increase in  $\delta^{18}\text{O}$  since 15 or 20 Mya, and others hint at an increase in  $\delta^{18}\text{O}$  beginning near 10 Mya (**Figure 8**). The lack of a consistent pattern and ignorance of how climate change might have affected  $\delta^{18}\text{O}$  makes interpreting these different time series difficult, but the data do imply changing conditions as Tibet was growing.

The summary of paleoclimate data given here can be taken to suggest that climate changes of various kinds occurred near 10 Mya throughout the region surrounding Tibet.

## 7. TIBET AND PALEOCLIMATE VIEWED THROUGH THE LENS OF MODERN CLIMATE DYNAMICS

When Harrison et al. (1992), Prell et al. (1992), and Molnar et al. (1993) suggested that a rise of Tibet of ~1000–2000 m near 8 Mya led to a strengthening of the South Asian monsoon at that time, data were fewer and more consistent with that association than today. Climate on the Indian subcontinent does seem to have changed near 10–8 Mya, but it now appears that if Tibet did abruptly rise 1000 m, it began to do so closer to 15 Mya if not slightly earlier. Moreover, northern Tibet could have risen more than 1000 m, but southern Tibet apparently did not. Given that Tibet may affect the South Asian monsoon more by serving as a barrier to cool, dry air from the north than by serving as a source of heat to the atmosphere above it, assigning climate change in India to a change in Tibet's height requires a more subtle understanding of how the two are related than the simple assertion that Tibet acts like "a heat engine with an enormous convective chimney" (Flohn 1968, p. 37). Bansod et al. (2003) did find a negative correlation of Indian monsoon rainfall with temperature over northeastern Tibet, which they associated with interannual variations in Eurasian atmospheric circulation. Thus, a link between the growth of Tibet and Indian climate may exist, but connecting them convincingly poses a challenge—we must first understand how the flux of energy into Tibet's surface affects South Asian climate today, and then how an evolving Tibetan landscape would lead to variations in South Asian climate.

As discussed above, loess deposition in China accelerated near 10–8 Mya, and a connection with the evolving extent and height of Tibet might seem direct (An et al. 2001). Modern dust, however, is deposited in spring when outbreaks of cold arctic air pass over the high terrain in

and near Mongolia, ~1000 km north of Tibet, and storms grow in the lee of these mountains (Penny et al. 2010, Roe 2009). Although that high terrain seems to have grown largely since 10 Mya—concurrently with the outward expansion of Tibet—it is not obvious how the presence of a high Tibetan Plateau affects spring dust storms. In fact, given that ice seems to have expanded on Greenland near ~7 Mya (Larsen et al. 1994), cold outbreaks might have increased at that time, and the growth of Tibet may have played no role at all in increased loess deposition. Again, an inadequate understanding of modern climate dynamics challenges us.

As indicated in the discussion above, the region most sensitive to Tibet's growth might be eastern China because of the apparent dependency of precipitation on the path of the jet, south of Tibet in winter and north of it in summer. As the southern margin of Tibet steadily moves northward at ~20–25 km per Ma, at some time in the past the jet cannot have passed south of the plateau, and rainfall over China should have been different. Similarly, if northern Tibet has grown substantially, both higher by ~1000 m and outward by hundreds of kilometers, since 10–15 Mya and perhaps before that time, the jet spent longer summers north of Tibet than it does today. Thus, we might expect a paleoclimatic record from eastern China to record such topographic changes.

Eastern China contains one of the richest collections of paleoclimatic data from within a continent, and indeed proxies of precipitation are measured by the principal recorders: loess and speleothems in caves (e.g., An et al. 1991, Maher & Thompson 1995, Zhang et al. 2008). Ironically, however, the time spanned by most of that record is too short to record changes in the topography of eastern Asia. On the positive side, if methods like that of W-J Zhou et al. (2007) for inferring annual precipitation amounts from loess sequences can be extended to reach back to millions of years, they might constrain Tibet's elevation history. Also, because both loess (e.g., An et al. 1991) and especially speleothems (e.g., Kelly et al. 2006, Y-J Wang et al. 2008) record climatic variability on the timescales of ice ages (19–23 ka, 41 ka, and 100 ka), which are associated with variations in mid- to high-latitude insolation, this variability might be tied closely to small variations in the position and strength of the subtropical jet stream that are amplified in China by Tibet's obstruction of the jet. Thus, these data could provide constraints necessary to understand how Tibet affects the jet on scales that modern climate variability cannot sample.

In summary, both our current knowledge of how Tibet has grown on geologic timescales and our understanding of how its present-day orography affects modern climate are in a state a flux. A pessimist might think that we have regressed, but an optimist can see opportunities for rapid understanding, especially if all three communities—geodynamicists, atmospheric scientists, and paleoclimatologists—can work together.

## DISCLOSURE STATEMENT

The authors are not aware of any affiliations, memberships, funding, or financial holdings that might be perceived as affecting the objectivity of this review.

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## Errata

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