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Correspondence and requests for materials should be addressed to W.R.B. (billboos@alum.mit.edu)

Sensitivity of the South Asian monsoon to elevated and non-elevated heating

William R. Boos[1] & Zhiming Kuang[2]

[1]Department of Geology and Geophysics, Yale University, New Haven, Connecticut, USA, [2]Department of Earth and Planetary Sciences, and School of Engineering and Applied Sciences, Harvard University, Cambridge, Massachusetts, USA.

Elevated heating by the Tibetan Plateau was long thought to drive the South Asian summer monsoon, but recent work showed this monsoon was largely unaffected by removal of the plateau in a climate model, provided the narrow orography of adjacent mountain ranges was preserved. There is debate about whether those mountain ranges generate a strong monsoon by insulating the thermal maximum from cold and dry extratropical air or by providing a source of elevated heating. Here we show that the strength of the monsoon in a climate model is more sensitive to changes in surface heat fluxes from non-elevated parts of India than it is to changes in heat fluxes from adjacent elevated terrain. This result is consistent with the hypothesis that orography creates a strong monsoon by serving as a thermal insulator, and suggests that monsoons respond most strongly to heat sources coincident with the thermal maximum.

Elevated heating by the Tibetan Plateau was long thought to drive the South Asian summer monsoon[1–3]. However, we recently showed that eliminating the Tibetan Plateau in a climate model produced little change in this monsoon, provided the comparatively thin wall of the Himalayas and adjacent mountain ranges was preserved (Boos and Kuang[4], hereafter BK10). This model result is consistent with the hypothesis that topography creates a strong South Asian monsoon by preventing the intrusion of cold and dry extratropical air into the warm and moist thermal maximum of the monsoon, a hypothesis motivated by the fact that observed near-surface entropies (or equivalent potential temperatures) exhibit sharp gradients coincident with the mountain ranges just south and west of the Tibetan Plateau (see also refs. 5 and 6).

Wu et al.[7] (hereafter Wu12) argued that surface heat fluxes on the slopes of the Himalayas instead provide a dominant forcing for the South Asian monsoon via a "sensible-heat-driven air-pump". In their proposed mechanism, sensible heat fluxes from mountain slopes produce rising motion that draws surrounding air toward the mountains, converging moisture which then condenses and heats the atmosphere in a positive feedback[8]. We show here that surface heat fluxes from mountain slopes are no more important for the South Asian summer monsoon than heat fluxes from nearby non-elevated surfaces. We explain why this is so through use of a convective quasi-equilibrium (QE) framework for monsoon dynamics that has been used in theoretical studies[9–12] and has been shown to be consistent with multiple observed monsoons[13]. In this framework, deep precipitating convection maintains the off-equatorial, free-tropospheric temperature maximum of the thermally direct monsoon circulation in approximate equilibrium with the maximum equivalent potential temperature of air below the base of cumulus clouds, θ_{eb}. Surface fluxes of both sensible and latent heat that are distributed vertically by moist convection thus become important in altering θ_{eb}, free-tropospheric temperature, and the associated monsoon flow.

Results

A climate model with the same topographic modification used by BK10 and Wu12 is integrated as a control. This model retains the relatively narrow topography of the Himalayas and adjacent mountain ranges over South Asia, but the Tibetan Plateau and other elevated topography north of those ranges has been removed (Fig. 1). South Asian summer precipitation and winds in this model differ little from a model with full, standard topography that includes the Tibetan Plateau, a result of BK10 that was confirmed by Wu12.

We conducted two integrations in which surface sensible heat fluxes in select regions were not allowed to warm the atmosphere, both using the same topography as the control run. When surface sensible heat fluxes are suppressed over South Asian topography higher than 500 m, precipitation over the elevated terrain decreases and low-level monsoon westerlies weaken, showing that these surface heat fluxes do contribute to the intensity of the South Asian summer monsoon (Fig. 2a, b). When surface sensible heat fluxes are suppressed over a similarly

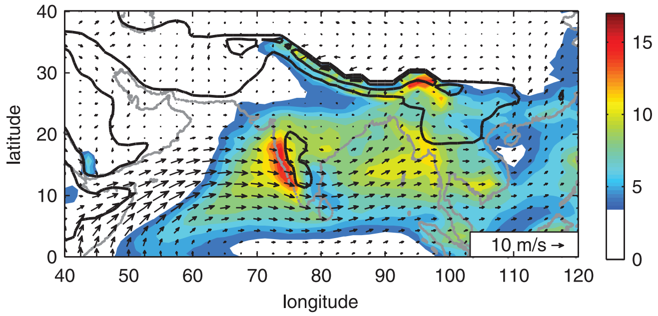

Figure 1 | Summer precipitation in control model. May-Sept. precipitation rate (shading, mm day^{-1}), horizontal wind on the 0.867 sigma level (arrows), and surface elevation (black contours, interval 2 km starting at 0.5 km).

sized, non-elevated region just south of the Himalayas, a larger decrease occurs in precipitation and in the strength of the low-level monsoon westerlies (Fig. 2c, d), even though the horizontally integrated reduction in surface sensible heat flux is smaller (Table 1). This shows that surface heat fluxes on sloping terrain are less important for the large-scale South Asian monsoon than heat fluxes from the non-elevated region of northern India.

This result can be understood by examining distributions of θ_{eb} and upper-tropospheric temperature. In the control integration, maxima of both of these quantities lie over the non-elevated part of northern India (Fig. 3a). The upper-tropospheric temperature maximum in this control run is displaced slightly south of its position in an integration with full topography (not shown), but even in that full-topography integration it lies over non-elevated parts of northern India, as it does in observations (BK10). Suppressing sensible heat fluxes over the Himalayas and adjacent elevated terrain decreases θ_{eb} and upper-tropospheric temperature poleward of the monsoon thermal maximum (Fig. 3b) and thus has a relatively weak effect on both the amplitude of that maximum and the meridional temperature gradient between that maximum and the equator. In contrast, suppressing surface heat fluxes in the non-elevated region directly beneath the maxima of θ_{eb} and upper-tropospheric temperature reduces the amplitude of these maxima and the associated meridional temperature gradient (Fig. 3c).

We have thus far emphasized changes in surface sensible heat fluxes, for consistency with Wu12, but surface latent heat fluxes are equally important in the QE description of monsoons discussed above because they also directly influence θ_{eb}. Surface latent heat fluxes fell in the regions in which surface sensible heat fluxes were suppressed because, like Wu12, we prevented sensible heat fluxes from warming the atmosphere while retaining those heat fluxes in the land surface energy budget. This amounts to prescribing a heat sink just above the land surface, which reduces surface temperature and thus surface evaporation. The reduction in horizontally integrated surface enthalpy (i.e. sensible plus latent heat) flux was 4% larger in amplitude when sensible heating was suppressed over non-elevated terrain than when it was suppressed over elevated terrain. Although it would be surprising if this small difference in surface enthalpy flux could explain the more than factor of two response in precipitation (e.g. Table 1), we conducted two additional integrations in which the surface enthalpy flux was perturbed by increasing land surface albedo in the same regions of elevated and non-elevated terrain described above. Qualitatively similar results were obtained, with the reduction in precipitation and low-level westerlies being stronger when albedo was increased over the non-elevated region, even though the change in surface enthalpy flux was slightly smaller than when albedo was perturbed over elevated terrain (not shown).

The amplitude of the θ_{eb} and upper-tropospheric temperature response shown in Fig. 3 is larger for the sensible heat flux perturbation over non-elevated terrain, despite the fact that the change in surface enthalpy flux was nearly the same as when sensible heat flux was suppressed over elevated terrain. A larger upper-tropospheric temperature response is expected when surface heat fluxes are perturbed in moist convecting regions near the thermal maximum, because surface flux perturbations in regions of large-scale subsidence will not be directly coupled to upper-tropospheric temperature due to the absence of deep convection. Furthermore, the net θ_{eb} and

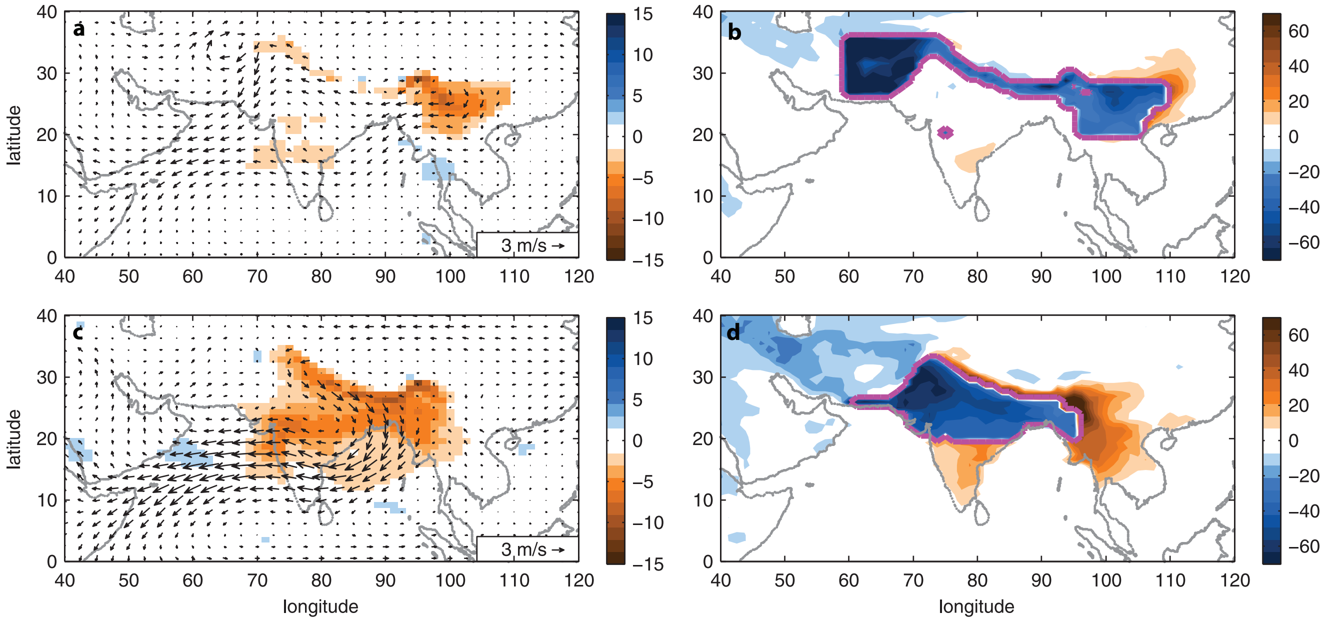

Figure 2 | Response of monsoon to suppression of surface sensible heat flux. Difference between the control model and the model with surface sensible heat fluxes suppressed over elevated terrain in (a), May-Sept. precipitation rate (shading, mm day^{-1}) and horizontal wind on the 0.867 sigma level (arrows) and (b), May-Sept. surface sensible heat flux (shading, W m^{-2}) with the magenta contour surrounding the region in which those heat fluxes were set to zero. (c, d) Show the same quantities but for the model with surface sensible heat fluxes set to zero over non-elevated terrain in northern India.

Table 1 | Properties of model runs with suppressed surface sensible heat fluxes

Region in which sensible heat flux is suppressed	Average surface elevation in region	Surface sensible heat flux change in region	Surface enthalpy flux change in region	Average low-level zonal wind change, 50–110°E, 10–35°N	Average precipitation change, 50–110°E, 10–35°N
Elevated terrain	1.5 km	−130 TW	−225 TW	$-0.89\ m\ s^{-1}$	$-0.47\ mm\ day^{-1}$
Flat northern India	0.26 km	−116 TW	−234 TW	$-1.1\ m\ s^{-1}$	$-1.2\ mm\ day^{-1}$

upper-tropospheric temperature response to a surface heat flux perturbation could be influenced by feedbacks with other tendencies on moist static energy, such as radiation and advection.

The Himalayas make up a relatively small fraction of the area in which surface heat fluxes were modified. Wu12 actually suppressed sensible heat fluxes over an even broader region of elevated terrain that stretched across the Iranian Plateau to 30E. We conducted another integration with surface sensible heat flux suppressed over that same region, effectively extending the region outlined in Fig. 2b westward over elevated terrain to 30E. This produced a decrease in horizontally integrated sensible heat flux roughly twice that shown in Fig. 2b, but the resulting decreases in precipitation and low-level zonal wind were only about 20 percent larger with a highly similar pattern to that displayed in Fig. 2a (not shown).

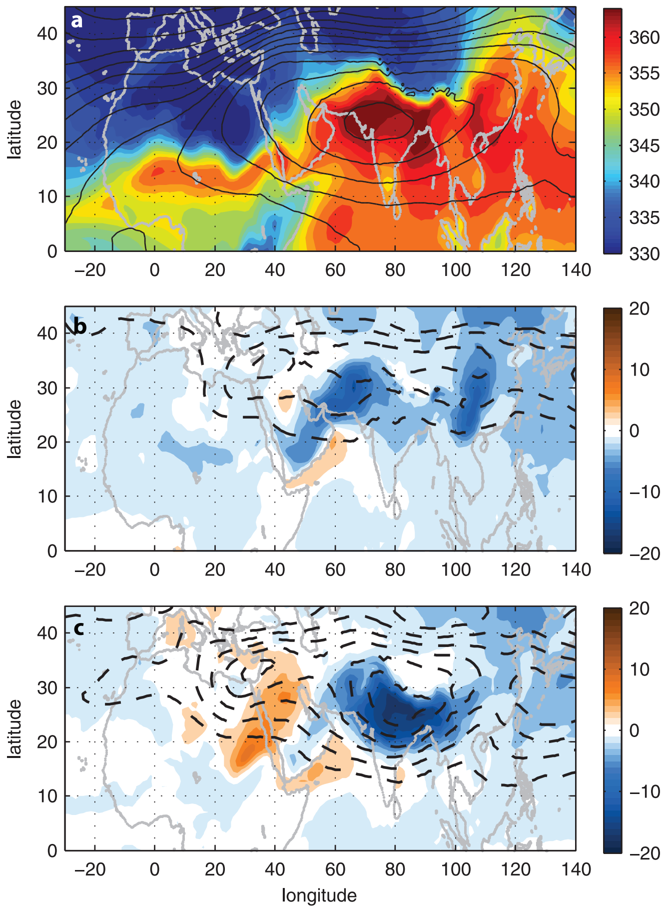

Figure 3 | Thermodynamic response to suppression of surface heat flux. (a), July mean equivalent potential temperature 30 hPa above the surface (θ_{eb}, color shading, contour interval 2K) and upper-tropospheric temperature (black contours, interval 1 K), both for the control integration. Difference in θ_{eb} (color shading, contour interval 2K) and upper-tropospheric temperature (contours, negative dashed, interval 0.5 K) between the control integration and (b), the integration with sensible heat flux suppressed over elevated terrain; (c), the integration with sensible heat flux suppressed over non-elevated northern India.

Discussion

The monsoon thermal maximum is maintained by surface fluxes of sensible and latent heat, so it is not surprising that a reduction in proximal surface heat fluxes reduces the strength of this thermal maximum and the intensity of the associated thermally direct circulation. In an idealized system in which a thermal maximum is perfectly isolated from a cold region by an insulating barrier, only changes in heating on the warm side of the barrier will alter the strength of the thermal maximum. This is analogous to reducing surface heat fluxes on the southern slopes of the Himalayas or over northern India: both are on the warm side of the insulating topographic barrier. The fact that the monsoon is somewhat more sensitive to surface heat fluxes over the non-elevated parts of India in the model used here is expected, since those fluxes are located closer to the thermal maximum (BK10) and thus provide the strongest correlation between heating and temperature. Nonzero sensitivity to heat sources on the cold side of the barrier (e.g. poleward of the Himalayas and Hindu Kush) is expected since the topographic barrier is not a perfect insulator.

The concept of a "sensible-heat-driven air-pump"[8] invoked by Wu12 relies on a positive feedback between low-level moisture convergence and precipitation. This sort of positive feedback is often called Conditional Instability of the Second Kind (CISK) and was once a central element of theories for tropical atmospheric dynamics[14,15]. CISK has since been shown to be energetically ill-founded: moist convection results from a local instability that occurs when θ_{eb} becomes sufficiently large, relative to the overlying free-tropospheric temperature, and convergence of low-level air toward that θ_{eb} maximum can only decrease its amplitude through advection[16,17].

In summary, Wu12 is correct that BK10 did not test the possibility that sensible heat fluxes from the slopes of the Himalayas provide a dominant forcing for the South Asian monsoon. The observed subcloud entropy distributions shown in BK10 provided no reason to suppose that heating from topographic slopes would be so important, and a QE theory that is consistent with observed monsoons does not assign surface sensible heat fluxes any greater importance than surface latent heat fluxes. We have shown here that, in a climate model, the strength of the South Asian monsoon is more sensitive to surface heat fluxes from non-elevated terrain directly beneath the thermal maximum than it is to fluxes from sloping terrain poleward of this maximum. These results are all consistent with off-equatorial maxima of θ_{eb} and free-tropospheric temperature being generated by local surface enthalpy fluxes and insulated from cold and dry extratropical air by topography.

Methods

The Community Earth System Model (CESM) version 1.0.4 of the National Center for Atmospheric Research (NCAR, http://www.cesm.ucar.edu) was integrated with fully active and coupled atmosphere, ocean, sea ice, and land components[18,19]. All integrations were performed using the finite volume dynamical core at $0.9 \times 1.25°$ horizontal resolution with 26 vertical levels. Although Wu12 and BK10 both used climate models with prescribed sea surface temperature (SST), we found that CESM produced a more realistic climatology of θ_{eb} and upper tropospheric temperature when coupled with an interactive ocean model. We repeated all integrations presented here using a model with prescribed climatological SST and obtained qualitatively similar results. Time averages were taken for the last 10 years of 25-year integrations, and any drift during those 25 years due to ocean equilibration did not qualitatively alter our results. The control model used topography modified to eliminate the bulk of the Tibetan Plateau while preserving the mountain ranges west and south of that plateau, using the exact procedure described in BK10 and repeated in Wu12. This sets surface elevations at each South Asian longitude to zero north of the point at which elevations reached two-thirds of their maximum at that particular longitude.

The influence of surface heat fluxes on the monsoon was assessed by setting the surface sensible heat flux into the atmosphere to zero at each time step in select regions. The land surface energy budget was not altered (meaning the surface sensible heat flux was still computed at each time step and allowed to influence land surface temperature), to match the methodology followed by Wu12. As in Wu12, one integration was conducted with surface sensible heat fluxes suppressed over sloping South Asian topography: between 60–110°E and 20–35°N in grid cells with surface elevations higher than 500 m. Another integration was conducted by suppressing surface sensible heat fluxes over an adjacent non-elevated region, defined as land surfaces south of the aforementioned slopes between 60–100°E, 20–35°N in grid cells with surface elevations less than or equal to 500 m. These two regions are marked in Figs. 2b and d, are roughly equal in horizontal area, and have roughly the same horizontally integrated surface sensible heat flux in the control run (Table 1). Additional integrations were conducted for which detailed results are not illustrated here, although the outcome is mentioned in the text of the results section. In one integration, surface sensible heat fluxes were suppressed between 30–110°E and 20–40°N in grid cells with surface elevations higher than 500 m, to more exactly match the model configuration of Wu12. In another two integrations, land surface albedo was set to 0.9 in the same regions in which surface sensible heat fluxes were previously suppressed.

Equivalent potential temperatures were computed on a terrain-following model level 30 hPa above the surface, using a standard definition of equivalent potential temperature[20]. Upper-tropospheric temperatures were calculated as a mass-weighted vertical average from 175–425 hPa. July climatologies of these thermodynamic variables are shown for easier comparison with previous studies[4,13], while May-Sept. climatologies of precipitation are shown to capture the effects of changes in the length of the rainy season.

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Author contributions

Both authors designed the research and interpreted results. W.R.B. performed and analyzed the model integrations and wrote the manuscript.

Additional information

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