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Tibetan Plateau climate dynamics: recent research progress and outlook

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ABSTRACT

This paper reviews progress in the study of Tibetan Plateau (TP) climate dynamics over the past decade. Several theoretical frameworks, including thermal adaptation and the TP sensible heat (SH) driving air-pump, have been developed to identify the mechanisms responsible for the circulation anomaly produced by thermal forcing of the TP. Numerical simulations demonstrate that the thermal effects of large-scale orography, including the Tibetan and Iranian Plateaus (TIP), are crucial for the formation of the East Asian and South Asian summer monsoons (SASM) because the surface SH of the TIP is the major driver of the water vapor transport required for the genesis of the north branch of the SASM. The large-scale orography of the TP affects the Asian climate through thermal forcing in spring and summer, and mechanical forcing in winter. The TP forcing can also influence the Asian summer monsoon (ASM) onset over the Bay of Bengal (BOB) by enhancing the BOB warm pool at the surface and by modulating the South Asian High (SAH) in the upper troposphere. On intra-seasonal timescales, the TP thermal forcing significantly modulates spring rainfall in southern China and generates the biweekly oscillation of the SAH in summer. Despite climate warming, the atmospheric heat source over the TP, particularly the spring SH, exhibits a clear weakening trend from the 1980s to 2000s. This weakening of the spring SH contributed to the anomalous 'dry in the north' and 'wet in the south' rainfall pattern observed over East China. Also discussed are challenges to further understanding the mechanism of TP forcing on the multi-scale variability of the ASM.

Keywords: Tibetan Plateau, climate dynamics, Asian monsoon, land–air–sea interaction, numerical simulation

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INTRODUCTION

Large-scale mountain ranges have significant impacts on atmospheric circulation through mechanical and thermal dynamical effects. The formation and variability of the Asian summer monsoon (ASM) are influenced by the spatial extent, elevation, steep topography, and the geographic position in subtropical eastern Eurasia of the Tibetan Plateau (TP).

Pioneering observational studies such as those undertaken by Yin [1], Yeh [2], Flohn [3], Yeh *et al.* [4], Murakami [5], Koteswaram [6], Staff Members of Academia Sinica [7], Yanai and Tomita [8], Tanaka *et al.* [9] and Ueda *et al.* [10] have examined the gross features of circulation and heating in different seasons over and around the TP, and discussed their possible connection with mechanical and thermal aspects of TP forcing. Many early theoretical and numerical modeling studies [11–14] focused on the influence of mechanical forcing on the formation of stationary waves in winter, while others found that planetary scale circulation can also be excited or maintained by thermal forcing alone [15–20]. With the seasonal evolution from winter to summer, strong surface sensible heat (SH) dominates the total diabatic heating over the TP in spring and regulates both the timing of the establishment of the ASM [21–23] and the inter-annual variability of the East Asian summer monsoon (EASM) [24–26]. As a large, intense, and elevated heat source, the TP plays a unique role not only in the formation of summer circulation [3,4,27–30], but also in the development of weather systems over East China

[31,32] and boreal climate patterns [33,34]. Moisture transport modulated by land surface processes in the TP is also an important source of water vapor for the EASM and western Meiyu rainfall over the Yangtze River valley [35,36]. Along with El Niño-Southern Oscillation events, the thermal forcing of the TP also contributes to the long-term changes in rainfall over eastern China and the large-scale atmospheric circulation associated with recent global warming [28,37,39].

The influence of the TP on climate variability can be assessed with sensitivity experiments using atmospheric general circulation (AGCM) or climate system (CSM) models. Kasahara and Washington [40], and Hahn and Manabe [34] examined the impact of topography on climate using AGCMs and found that the South Asian summer monsoon (SASM) cannot extend across the entire Indian subcontinent without the TP. Abe *et al.* [41] argued that the air–sea interaction in the North Indian Ocean could be modified by the influence of the TP, and further modulated the onset process of ASM. Kitoh *et al.* [42,43] also suggested that thermal processes associated with the TP impact the Asian monsoon by modulating the air–sea interaction. While Okajima and Xie [44] found that the uplifted mountains play a crucial role in the formation of the northwestern Pacific monsoon by modulating the air–sea interaction.

Influences of snow cover/depth on surface heat fluxes and atmospheric heat source/sink over the TP and the subsequent anomaly of ASM are still open questions. Barnett *et al.* [45] argued that there is no clear relationship between albedo effect caused by snow anomaly and the subsequent monsoon development without considering the snowmelt and evaporation effect, whereas Chou [46] suggested that the TP surface albedo becomes smaller due to less snow cover, leading to more intensified ASM and rain belt moves northeastward, but the tropospheric temperature shows small differences due to TP albedo changes.

Recent studies [47–49] show that the aerosols can significantly influence the surface albedo and hydrologic cycle on TP. The dust and black carbon [50] are transported into the northern and southern slope of TP in summertime. Therefore, the absorption of solar radiation by the aerosols heats the elevated surface over the TP and significantly influences the thermal effects of TP on the ASM.

Our understanding of TP climate dynamics has significantly advanced in the recent years as observational data and numerical models have improved. This has led to a better understanding of how the mechanical and thermal forcing of the TP can affect regional and global climate variability. This paper presents a brief review of this topic mainly based on the studies of this research group. We will review the basic theory of TP climate dynamics and discuss the influence of the TP on Asian climate and the global circulation in different seasons, the impacts of the TP on the formation of persistent rainfall in early spring (PRES) over southern China, the ASM, and the subtropical desert and monsoon, the ASM onset and evolution associated with the TP forcing, the atmospheric low-frequency oscillation related to the TP forcing and its impact on surrounding regions, as well as the decadal changes in atmospheric heat sources/sinks under a background of climate warming and possible feedbacks to the EASM.

INFLUENCE OF THE TP ON ASIAN CLIMATE PATTERNS AND GLOBAL CIRCULATION

Background and theoretical framework

During the 1930s and 1940s, Rossby [51] and Obukhov [52] proposed the theory of geostrophic adjustment to explain the interaction and adaptation between the wind and the mass fields within the atmosphere. To investigate the responses of the circulation to external thermal forcing, Wu and Liu [53] proposed the thermal adaptation theory to explain the circulation response to diabatic heating. The concept of 'overshooting' was also introduced to the study of thermal forcing by the TP. Above the level of diabatic heating, the 'overshooting' air parcel induced by the inertial ascent from below keeps a constant potential temperature, forming a cold anticyclonic circulation aloft. This concept was used to explain how a thick mixed potential temperature layer forms over the TP before monsoon onset, as reported by Yanai and Li [54], and why the surface SH is able to influence circulation in the upper troposphere (Fig. 1a).

Based on the thermal adaptation theory, a characteristic pattern of vertical motion induced by subtropical diabatic heating was proposed [53]. In summer, the zonal flow and vorticity advection across the subtropics are weak. By combining the continuity equation and the quasi-steady state geostrophic vorticity equation, the vertical velocity can be simplified as

$$w \approx -\frac{\beta \partial v}{f \partial z}. \tag{1}$$

Here, w, v, f, and β are the vertical motion, meridional wind, the Coriolis parameter, and its meridional gradient, respectively. The atmospheric response to an external heating (cooling) generates

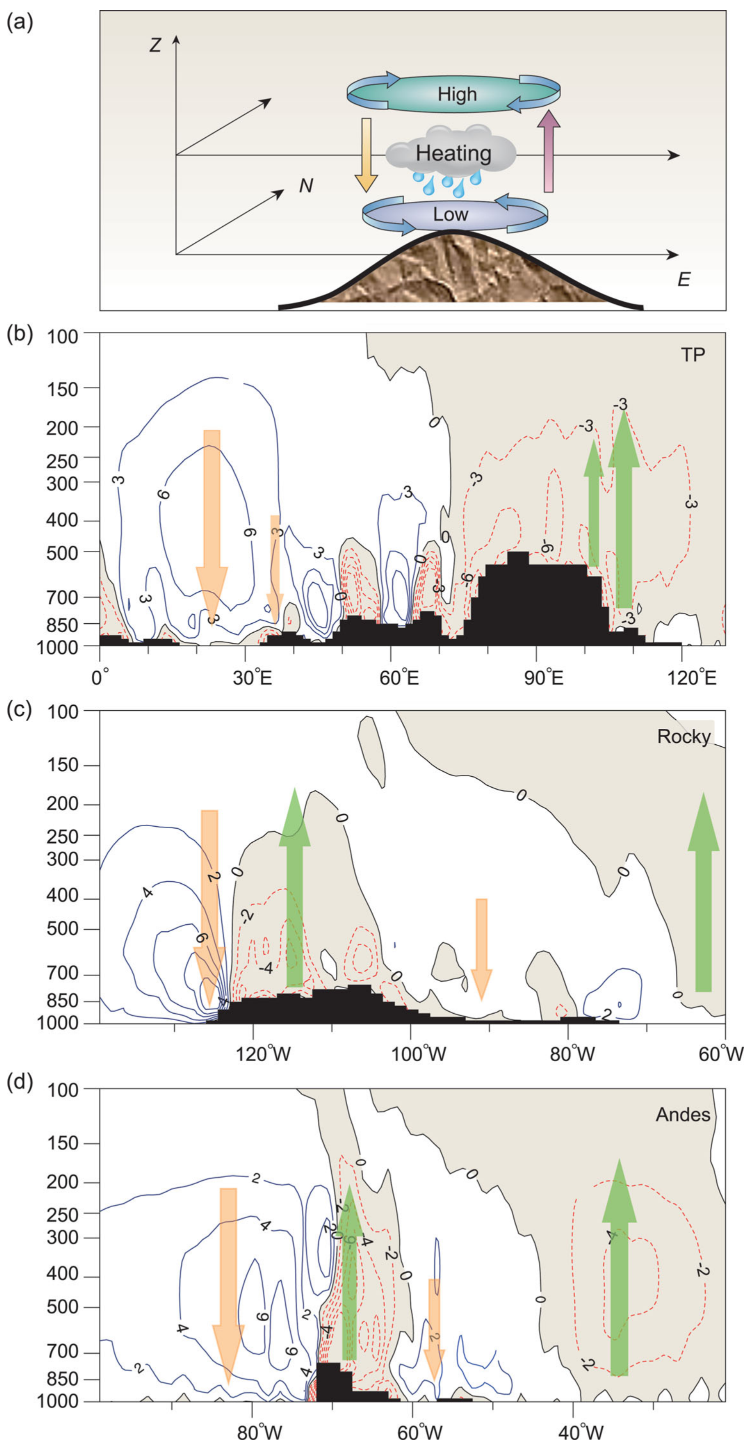

Figure 1. (a) Schematic diagram showing atmospheric response on the rotating earth to external heating in the summertime subtropics: cyclonic circulation near the surface and anticyclone circulation in the upper troposphere are generated with air ascent/descent developing over the eastern/western part of the heating. Also shown are 1979–2010 monthly mean longitude–pressure cross-sections of vertical velocity (unit: Pa s^{-1}) along 32.5°N (b) and 40.0°N (c) in July, and 30.0°S (d) in January, from ERA–Interim reanalysis (reproduced from [56]).

cyclonic (anticyclonic) circulation in the lower atmosphere and anticyclonic (cyclonic) circulation in the upper troposphere. According to (1), ascent should develop on the eastern (western) side of the heating (cooling) region and descent should develop on the western (eastern) side of the heating (cooling) region (Fig. 1a). This idealized vertical motion pattern is observed in summer over the TP, the Iranian Plateau, the Rocky Mountains, and the Andes Mountains (Fig. 1b–d) because of the uplift-related heating generated by these mountains [55]. Because the TP is located in the eastern Eurasian continent, the ascent induced by the TP is in phase with the ascent produced by the continental-scale thermal forcing; consequently, the strongest monsoon and largest deserts form over the Afro-Eurasian continent (Fig. 1b). In contrast, the Rockies and the Andes are located in the western parts of their respective continents; thus, the orography-induced ascent is geographically distinct from the ascent due to continental-scale forcing, and the deserts and monsoon climate are not as strongly developed as those over the Afro-Eurasian continent (Fig. 1c and d).

Yeh *et al.* [4] calculated each term of the thermodynamic equation and reported that the TP is a weak heat sink in winter but a large heat source in summer. Then the air column over the TP loses energy to create a forcing for descent in winter, but it is heated to create a forcing for ascent in summer. The corresponding surface circulation diverges away from the TP in winter and converges toward the TP in summer, acting like a large air pump. Based on a series of numerical experiments (Fig. 2), Wu *et al.* [56] demonstrated that SH on the sloping lateral surfaces of the TP is the major driver of the air pump. According to the thermodynamic equation,

$$\vec{V} \cdot \nabla\theta = Q, \tag{2}$$

where $\vec{V}$, θ, and Q are the horizontal wind, potential temperature, and diabatic heating, respectively. In the presence of surface SH on the sloping surface of a mountain (Fig. 2a and b), an air parcel moving along an isentropic surface impinges on the topography is heated as it encounters the SH along sloping surface of the mountain, and is able to penetrate the isentropic surface and ascend. In the area surrounding the topography, air from lower levels is lifted to the plateau region, producing strong ascent and heavy rainfall over the TP. In the top-heating-only case (Fig. 2c), the heating can produce convergence above the plateau, but cannot pull up air from below. This is because air parcels travelling in the lower layer and impinging on the TP must remain on the same isentropic surface without diabatic heating on the sloping lateral surface of the TP. The air parcel moves

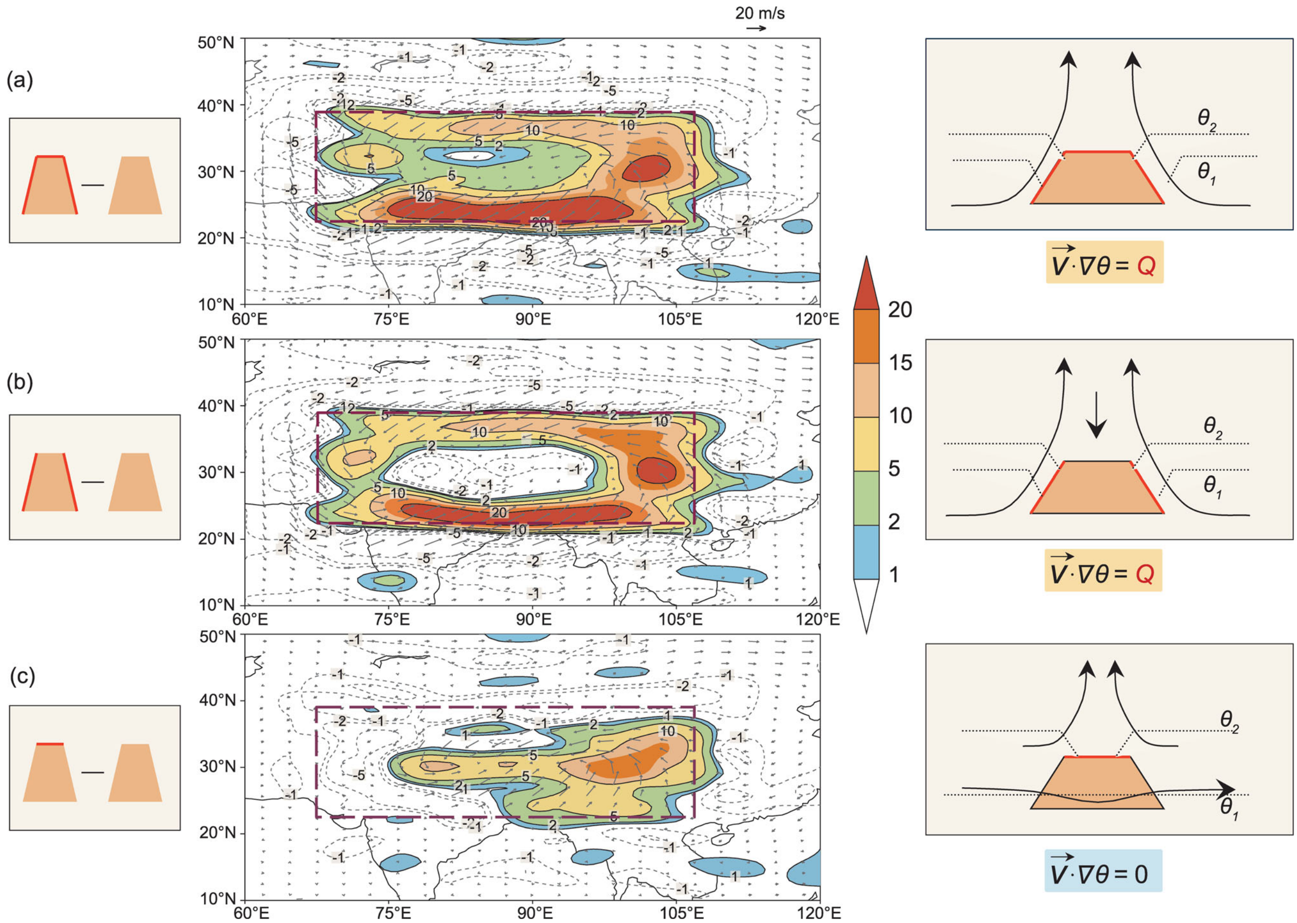

Figure 2. Distributions of difference in wind (vectors, m s^{-1}) and vertical velocity ($-\omega$, shading, 10^{-2} Pa s^{-1}) at the $\sigma = 0.991$ surface between the two perpetual July experiments: (a) ALLSH–NOSH, (b) SLPSH–NOSH, and (c) TOPSH–NOSH, with the dashed rectangle indicating the prescribed mountain base. Left panels indicate the experiment designs and right panels are interpretations of the relevant mechanisms, with orange shading representing mountain and heavy red bar denoting the imposed surface sensible heating. See the text for details [56].

around the TP on an approximately horizontal isentropic surface without ascending and has no significant impact on monsoon rainfall. It appears that air pumping is driven by the surface SH on the TP slope, and is therefore referred to as TP-SHAP [57]. Results from numerical experiments [58] demonstrate that such a thermal pumping also exists in the Iranian Plateau forcing; the pumping effect of Tibetan and Iranian Plateaus (TIP) forms a TIP-SHAP that drives water vapor from oceans and contributes to the formation of the ASM.

Recently, Boos and Kuang (BK hereafter) [59,60], citing results from the National Center for Atmospheric Research (NCAR) CSM model, claimed that the impact of TP thermal forcing is less important than the thermal insulation of the Himalayas in the formation of the SASM. BK proposed that the Himalayas block cold and dry air from north so that the warm–moist high surface entropy over North India is maintained, which is coupled with the upper-level temperature maximum through moist convection, and can drive the SASM. Their conclusions differ from those of Kasahara and Washington [40], and Hahn and Manabe [34], who employed AGCMs to study topographic impacts on climate and demonstrated that there was no cold and dry advection from the subtropics to North India in summer even after removing the mountains over the world. BK's conclusion is questionable. First, in summer the boreal subtropics receive more insolation than North India, and over India there does not exist such a cold and dry advection from north. Second, although high surface temperature in North India is necessary for local convection to develop, it is insufficient since high subcloud entropy requires not only high temperature, but also plenty of moisture. Strong surface hot depressions exist over many tropical land areas such as the Arabian Peninsula, but there is no monsoon over there due to the lack of moisture. Actually, in BK's experiment, the thermal pumping of the Himalayas is remained. As we will see later, it is such a thermal

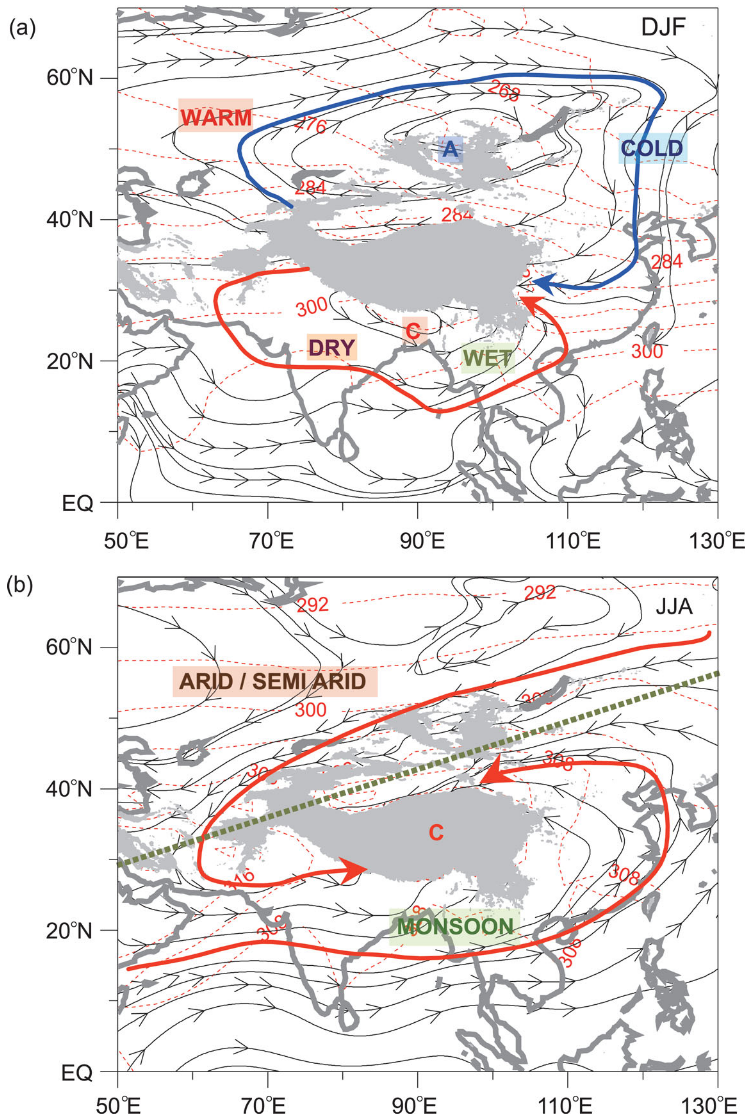

Figure 3. Distributions at 850 hPa of potential temperature (K) and stream fields composed of wind deviations from the corresponding zonal means based on ERA–Interim reanalysis for 1979–2010. (a) Winter average (December to February). (b) Summer average (June to August). Shading indicates where the elevation exceeds 1800 m (reproduced from [56]).

pumping effect that contributes to the monsoon occurrence over North India. Once the thermal effect is removed, monsoon over North India disappears even the mountain mechanical blocking exists.

Stationary waves forced by the TP

In winter, the TP reacts to impinging mid-latitude westerly winds by exerting a negative mountain torque on the atmosphere. As a result, the westerly winds are deflected by the topography and the zonal deviation of streamlines produces an asymmetric dipole with an anticyclone to the north of the TP and a cyclone to the south (Fig. 3a). The anticyclonic gyre in high latitudes transports warm air northward to its west and cold air southward to its east. As a result, the isotherms in the high latitudes of Asia tilt from the northwest to the southeast such that the temperature at 40°N near 130°E is more than 10 K colder than at 50°E. The cyclonic gyre in low latitudes transports dry air southward to the South Asia subcontinent and moist air northward to the Indochina Peninsula and South China, triggering the dry season in South Asia and the persistent rainy season in Southeast Asia and South China, which precedes the Asian monsoon onset.

In summer, the strong pumping of the TP-SHAP causes the surrounding flow to converge into the TP area. Consequently, the summer 850 hPa stream flow pattern resembles a cyclonic spiral and the TP-SHAP looks like a spiral pump (Fig. 3b) accompanied with air ascent/descent on its east/west (Fig. 1a). The TP is therefore an important location for the genesis of vortices that can propagate eastward and produce torrential rain along the Yangtze River in summer [36,61]. Although thermal forcing of the TP is dominant in summer, as an obstacle in the subtropics it can also influence the surrounding circulation. Numerical experiment results [62,63] demonstrate that the TP can deflect the impinging higher latitude westerly on its west and lower latitude easterly on its east, forming anticyclonic circulations on both sides. Since the isentropic surfaces along subtropics tilt down southward sharply, the mechanically forced anticyclonic circulations slide upward on the eastern TP and downward on its west. Consequently, the mechanically forced vertical motion is in phase with that due to its thermal forcing as shown in Fig. 1a.

TP forcing and its impacts on global circulation

In summer, the TP is not only a heating source but also a negative vorticity source for the atmosphere. Numerical experiment results showed that imposing a surface SH of 100 W m^{-2} in a region can produce a surface cyclonic vorticity of 10^{-5} s^{-1} within 2–3 h if no vorticity was exchanged with the underlying surface. Subsequently, a remarkable warm surface cyclonic circulation and a cold anticyclone circulation in a deep column aloft are formed in few days over the heating region [56]. This is because air converges in the lower layers and penetrates isentropic surfaces over the heating region, and diverges in the upper layers where there is no heating. Since continuity requires the air to rise in these layers above the heating region—a process which is referred to as overshooting, and since in this region

$$\frac{d\theta}{dt} \approx w\frac{\partial\theta}{\partial z} = Q = 0, \qquad (3)$$

ascent must occur in the nearly constant isentropic layer. In other words, a rising air parcel brings low potential temperature from the lower layer to the upper layer and forms a cold center aloft. According to the alternative form of the vertical vorticity equation [53],

$$\frac{\partial\zeta}{\partial t} + \nabla\cdot\vec{V}\zeta \propto \theta_z^{-1}(f+\zeta)\frac{\partial Q}{\partial z} + F_z, \quad \theta_z \neq 0 \qquad (4)$$

where $\vec{V}$, ζ, θ_z, and F_z are the horizontal wind, relative vorticity, static stability, and surface friction, respectively, and cyclonic circulation is generated on the underlying surface. Thus, the atmosphere receives a negative vorticity forcing due to surface friction, F_z. Furthermore, because surface SH decreases quickly with increasing height, this heating also generates negative vorticity in the atmospheric column. The generated negative vorticity is transported by the heating-induced ascent to the upper troposphere, forming an anticyclonic circulation in a thick layer with upper-level divergence that transports negative vorticity to the surrounding region. Therefore, the surface SH becomes a negative vorticity source for atmospheric motion.

To illustrate how the surface SH of the TP can affect the general circulation of the atmosphere, two numerical experiments were performed using the spectral atmospheric model of LASG/IAP (State Key Laboratory of Numerical Modeling for Atmospheric Sciences and Geophysical Fluid Dynamics/ Institute of Atmospheric Physics) [64], including a control experiment (CON) and another that removed the surface SH of the TP for elevations over 1.5 km (NSH). The difference in the July 200 hPa stream function between the two experiments (i.e. CON–NSH) has shown a strong difference in anticyclonic circulation developed aloft, indicating that the surface SH of the TP is an efficient producer of negative vorticity in the upper troposphere. Because the negative vorticity source is located in the mid-latitude westerlies, it also generates a Rossby wave train that influences the general circulation anomaly across the Northern Hemisphere.

Although the results were obtained through numerical experiments, they are also supported by observational evidence. Liu *et al.* [65] analyzed the correlation between July heating over the TP area and the corresponding 200 hPa geopotential height using monthly NCEP–NCAR Reanalysis data [66], and observed a similar Rossby wave pattern. This suggests that the TP-SHAP affects the circulation over the Northern Hemisphere as well as the local climate.

FORMATION OF SPRING PERSISTENT RAINFALL, DESERT, AND THE ASM

The ASM can be divided into the tropical and subtropical summer monsoons [67–70]. The tropical monsoon includes the Bay of Bengal (BOB) summer monsoon, the South China Sea (SCS) monsoon, the western North Pacific summer monsoon, and the Indian summer monsoon, while the subtropical monsoon typically refers to the summer monsoon over subtropical regions of China, Japan, and Korea. Before the ASM onset, an equally significant climate event is the persistent rainfall over South China in spring. We found that the TP forcing plays a vital role in the formation of both the spring persistent rainfall and the ASM.

PRES over southern China

PRES over southern China are a unique feature of Asian rainfall. They occur over southern China from late February to early May before the onset of the ASM. Tian and Yasunari [71] proposed the lag in the temperature contrast between land and sea in explaining the formation of PRES. Wan and Wu [72] showed that a similar time-lagged pattern of heating is also observed between Mexico and the western North Atlantic in spring, although there is no PRES in North America. Their results suggested that the time-lag mechanism is a necessary but not sufficient condition for the occurrence of PRES.

During spring, the TP-dipole-type stationary circulation pattern (Fig. 3a) still exists. The deflected flow on the eastern side of the TP brings cold air from the north and moist air from south, which converge over eastern China, contributing to the formation of PRES. This hypothesis has been verified by numerical experiments [57,73].

Multi-scale forcing and coexistence of subtropical desert and monsoon

There are three types of atmospheric forcing across the summertime subtropics that contribute to the occurrence of drought and wet climates in the subtropics. Wu *et al.* [74] proposed a new mechanism for the formation of desert over the western parts of continents and of monsoon over the eastern parts, and showed that a monsoon climate and deserts coexist as twin features of multi-scale forcing.

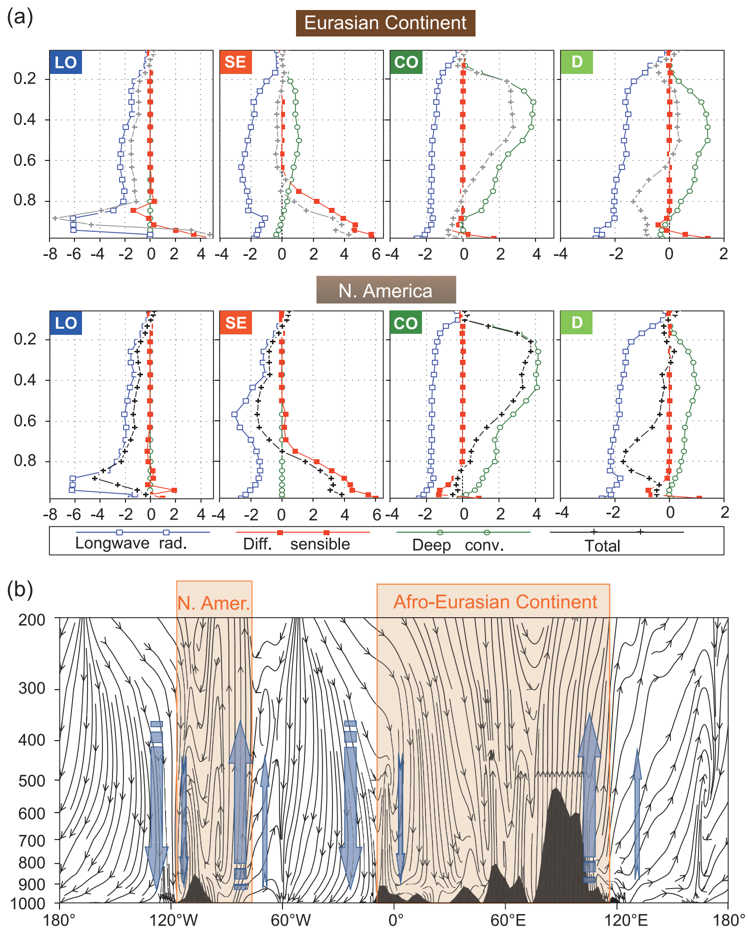

Figure 4. (a) July mean vertical heating profiles in sigma-coordinates across the Afro-Eurasian continent and adjacent oceans (upper panel), and North America and adjacent oceans (lower panel). The lines with open squares, filled squares, open circles, and 'plus' signs represent long-wave radiative cooling (LO), diffusive sensible heating (SE), deep convective condensation heating (CO), and total heating, respectively. Unit is K day^{-1}. (b) July mean vertical cross-section of subtropical circulation along 30°N (vertical velocity has been amplified by a factor of 100). The boxes indicate the locations of the North American and Afro-Eurasian continents. Black shading indicates orography [73 and 75].

First, continent-scale heating over land and cooling over ocean induces ascent over the eastern parts of continents and western oceans, and descent over eastern oceans and the western parts of continents. Second, local-scale sea breeze forcing in coastal regions enhances descent over the eastern oceans and ascent over the eastern parts of continents (Fig. 4). This leads to the formation of the well-defined summertime subtropical LOSECOD quadruplet-heating pattern across the continents and adjacent oceans in the summer subtropics, in both the Northern Hemisphere and Southern Hemisphere [74,75]. LOSECOD refers to long-wave radiative cooling (LO) over eastern oceans, sensible heating (SE), and condensation heating (CO), respectively, over the western and eastern continents, and double heating (D: LO+CO) over western oceans (Fig. 4). This heating pattern leads to a dry climate over the western regions and a wet climate over the eastern part of the continents. Third, regional-scale heating associated with orographic uplift generates ascent to the east of the orography and descent to the west. Accordingly, the deserts and monsoon climate develop over the western and eastern parts of subtropical continents, respectively [73].

Formation of the ASM

The CSM is an ideal numerical platform for studying the TP. The spectral version of the Flexible Global Ocean–Atmosphere–Land System (FGOALS-s), one of the IPCC CSMs, was developed mainly at LASG/IAP [63,76,77]. FGOALS-s has been widely used in studies of the thermal and dynamic effects of the TP [57,63,73,76,78,79].

In reality, there is another plateau, the Iranian Plateau, which has lower elevation but the comparable size with the TP and is located to the west of it. Numerical experiment [78] demonstrates that its presence generates a strong southerly to its east and southeast against the cold and dry northerly from the north that exists only if the TP forcing is imposed, thus contributing to the formation of the South Asian monsoon, at least to its northern part. The results have also been verified by a two-layer quasi-geostrophic model [80]. Therefore, it is more reasonable to think of the TP and Iranian Plateau as one unity TIP and investigate their joint impacts on the ASM.

Wu *et al.* [78] carried out a series of sensitivity experiments using FGOALS-s and demonstrated that the thermal effects of the TIP-SHAP are crucial for the formation of the SASM, primarily through the role of the surface SH. Fig. 5a shows the simulated boreal summer mean precipitation and near-surface stream function. Due to the pumping effect of the TIP driven by surface SH, the lower-level water vapor along the tropical water vapor conveyer belt is deflected northward over North India and the BOB region, and is then lifted on to the south slope of the TP, causing heavy precipitation and forming the northern branch of the SASM. However, without the SH of the TIP (Fig. 5b), the air flows along quasi-horizontal isentropic surfaces, and the streamlines are parallel to the topography near the TP. Therefore, water vapor cannot be transported up to the free atmosphere, and the northern branch of the SASM disappears. Fig. 5c shows a schematic diagram of the thermal control of the ASM. The water vapor conveyer belt over South Asia can be separated into three major branches. The first branch,

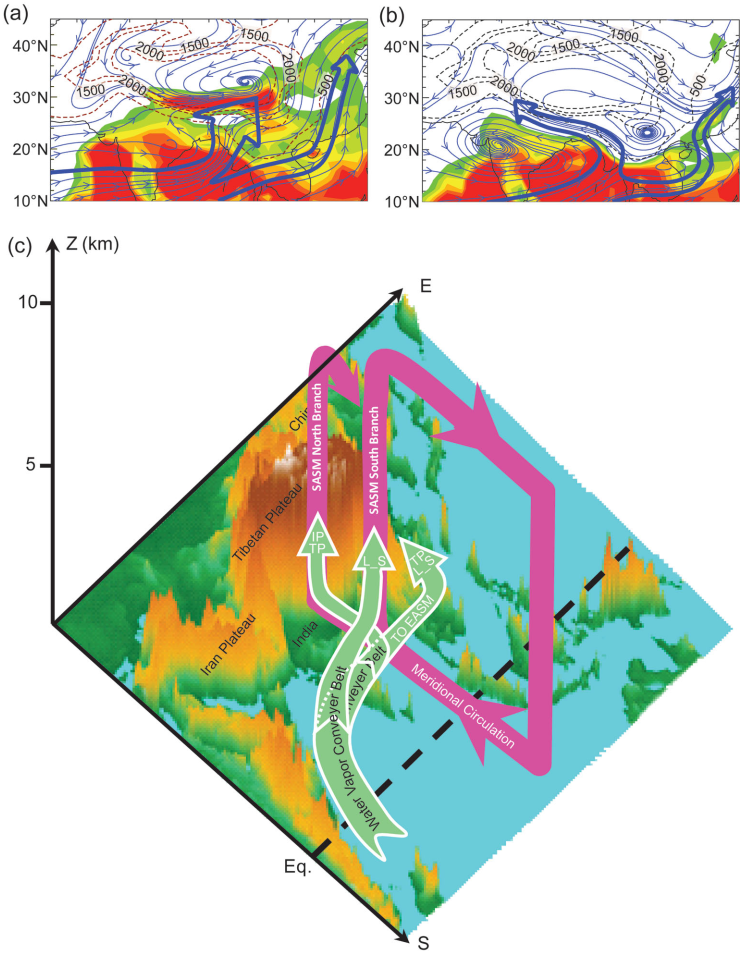

Figure 5. Relative contributions of the climbing and deflecting effects of mountains, showing the summer precipitation rate (shading, mm day^{-1}) and streamlines at the $\sigma = 0.89$ level (a) with surface SH on the mountain range and (b) without surface SH on the mountain range. Dashed contours indicate elevations higher than 1500 and 3000 m, with red and black colors, respectively, indicating with and without surface sensible heating of the mountains. Dark blue open arrows denote the main atmospheric flows impinging on the TP, either climbing up the plateau (a) or moving around the plateau, parallel to orographic contours (b). (c) Schematic diagram showing the gross structure of the ASM. For the southern branch, water vapor along the conveyer belt is lifted up mainly due to the land–sea (L–S) thermal contrast in the tropics; for the northern branch, the water vapor is drawn away from the conveyer belt northward toward the foothills and slopes of the TP and is uplifted to produce heavy precipitation that is controlled mainly by TIP-SHAP; the rest of the water vapor is transported northeastward to sustain the EASM, which is controlled by the land–sea thermal contrast as well as thermal forcing of the TP [78].

in the tropics across the Arabian Sea, the BOB, and the SCS, induces heavy precipitation and forms the southern branch of the SASM, mainly due to land–sea thermal contrast. The second branch flows northward across northern India and the BOB toward the south slope of the TP because of the TIP-SHAP and forms the northern branch of the SASM. The third branch transports water vapor to sustain the EASM, owing to the thermal effects of both land–sea contrast and the TP.

ONSET AND PROPAGATION OF THE ASM

The ASM onset process is characterized by three successive phases, with the earliest onset starting over the southeastern BOB, followed by onset over the SCS, and finally over India [23,82–84]. The elevated heating over the TP is important for the ASM onset process [54,85–87]. In spring, the TP-SHAP contributes to the seasonally abrupt change in the Asian circulation and anchors the earliest ASM onset over the eastern BOB [23,57,88]. Bao *et al.* [79] showed that realistic treatment of soil moisture over the TP leads to the accurate prediction of TP-SHAP and improves simulations of the ASM onset, especially for the EASM.

Phase locking of intra-seasonal oscillation during the ASM onset

Wu and Zhang [23] investigated thermal impact of the TP on the onset of the ASM in 1989. They found three periods with abrupt increases in temperature over the TP that occurred in late April, middle May, and early June (red in Fig. 6a), respectively, concurrent with the monsoon onsets over the eastern coast of the BOB in early May, over the SCS on 20th May, and over India on 10th June. Surges of enhanced surface SH over the TP were identified, leading the first and second temperature increases by about 10 days (blue in Fig. 6a). The ASM onset area shown in Fig. 6b (green) includes the northern parts of the SCS, the BOB and the eastern part of the TP, and can be regarded as the most sensitive region for the ASM onset. From April to June of 1989, there appeared four warm surges of the two-to-three-week oscillation (TTO) mode of upper tropospheric temperature, each of which propagated eastward (along A–B in Fig. 6b) from Europe into the TP area (Fig. 6c and d, left panel). The first warm surge, which reached the plateau region in the first half of April, did not meet with the northward propagating (along C–D in Fig. 6b) Madden–Julian oscillation (MJO) (Fig. 6c, right panel) and the westward propagating (along E–F in Fig. 6b) TTO of tropical divergence (Fig. 6d, right panel). Therefore, monsoon onset did not occur until the arrival of the second, third, and fourth warm surges, all of which were phase-locked with the arrival of both the northward propagating MJO and the westward propagating TTO. Together, the MJO and TTO led to the vigorous development of strong convection and torrential rain over a large area. Therefore, it was concluded that the timing of the monsoon onset was largely determined by the phase locking within the ASM onset area among different kinds of low-frequency oscillations from different directions.

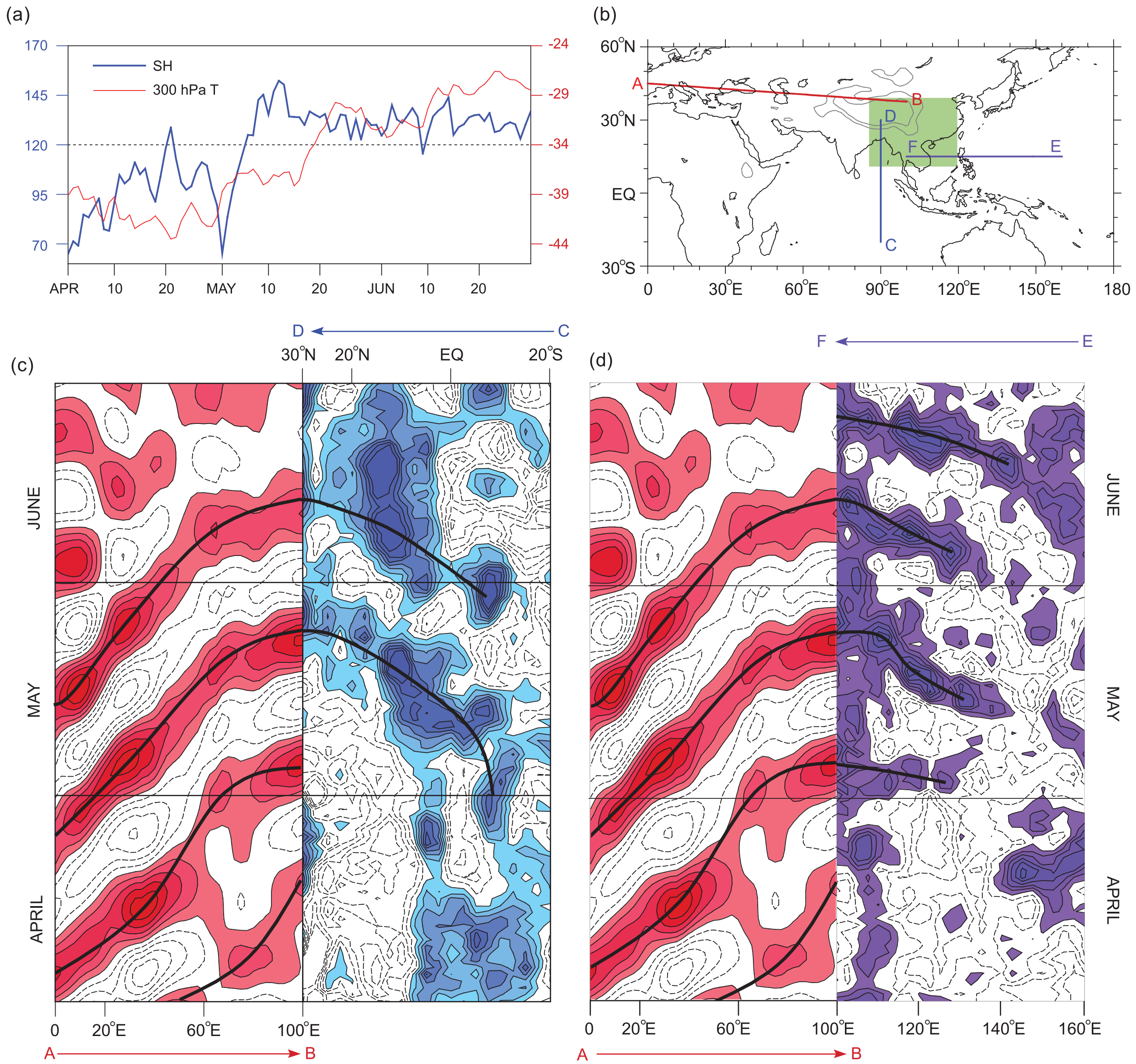

Figure 6. (a) Evolution during April–June 1989 of the surface sensible heat flux SH (left scale in W m^{-2}) and the 300 hPa temperature *T* (right scale in °C), averaged over the TP (27.5°–37.5°N, 80°–100°E). (b) Locations of the East Asian monsoon area as defined by the green rectangle, and cross-sections A–B, C–D, and E–F showing the propagation of different low-frequency oscillations. (c) Evolution along the A–B cross-section of the 200 hPa temperature, processed using a 15–25 day bandpass filter (left panel, interval of 1.0°C), and along the C–D cross-section of the 200 hPa divergence (right panel, interval of 2.0×10^{-6} s^{-1}). (d) (Left) Same as the left panel in (c). (Right) Evolution along the E–F cross-section of the 200 hPa divergence averaged from 10° to 20°N. Interval is 2.0×10^{-6} s^{-1}. Stippling indicated divergence (reproduced from [23]).

Onset process of the BOB summer monsoon

Most of the BOB summer monsoon onsets are characterized by the formation and development of the monsoon onset vortex (MOV). The MOV occurs over the eastern BOB because of the local surface air–sea interaction [88,89] and its close coupling with the upper divergence pumping, attributed to the development of the South Asian High (SAH) [90]. Both of the factors are closely associated with the TP forcing.

The influence of the TP on the formation of the short-lived BOB warm pool in spring and its contribution to the BOB MOV are illustrated in Fig. 7. The cold and dry northwesterly flow over India induced by TP forcing generates strong surface SH and cyclone circulation (Fig. 7a). The resultant southwesterly flow along the western offshore region of BOB, together with the equatorial westerly winds, forces

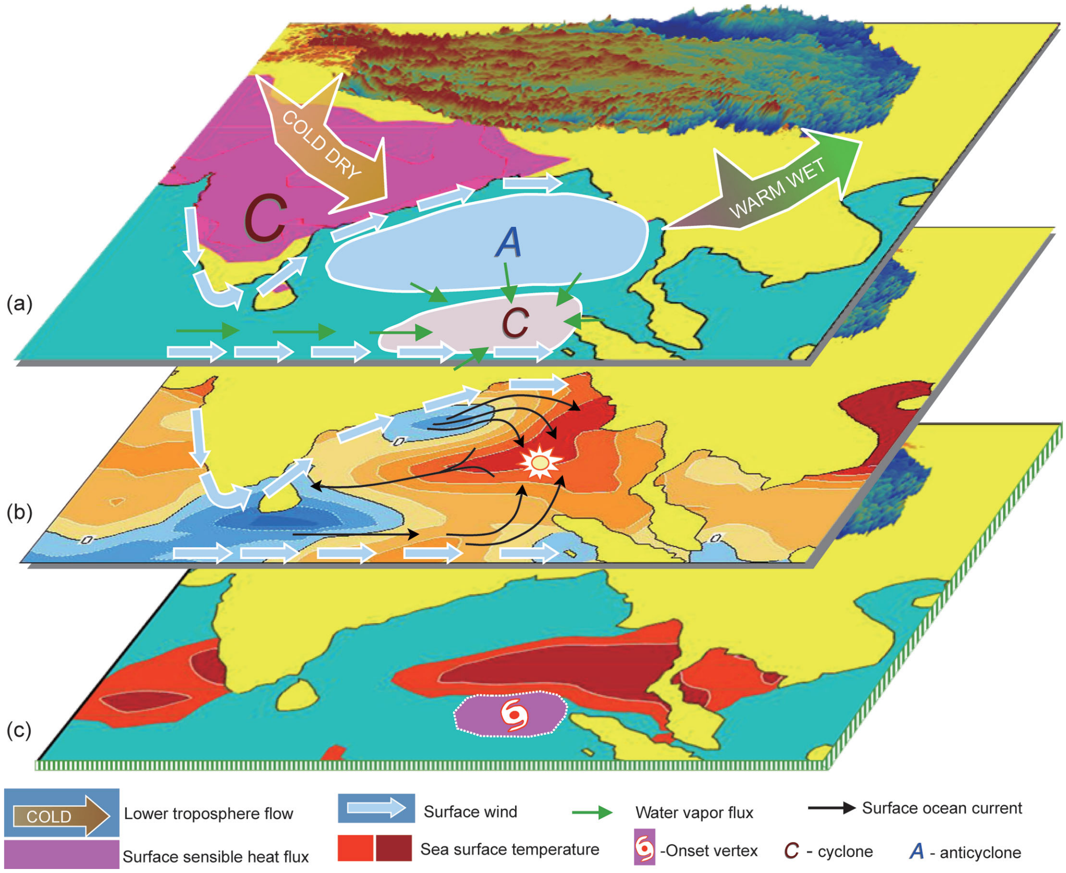

Figure 7. Schematic diagram showing the formation of the BOB MOV as a consequence of *in situ* air–sea interaction modulated by the land–sea thermal contrast in South Asia and TP forcing in spring [89]. See the text for details.

an anticyclonic circulation with descending air over the northern BOB, cyclonic circulation with ascending air over the southern BOB, and strong water vapor transport from the northern BOB and other areas toward the southeastern BOB (Fig. 7a). The strong southwesterly flow along the western offshore region of the BOB also results in an offshore ocean current and upwelling that produces cold sea surface temperatures (SSTs) in the western BOB. Warm surface water converges in the eastern part of the central BOB, which is under the influence of an anticyclone with clear skies, weak surface winds, and a shallow mixed layer in the ocean. As a result, strong solar radiation and weak energy loss act on a very thin mixed layer (Fig. 7b). Consequently, SSTs decrease in the western BOB and increase rapidly in the eastern BOB, forming a springtime warm pool in the eastern part of the central BOB prior to the formation of the MOV (Fig. 7c) near the region with strong surface SH. Finally, local convection develops because of the positive correlation between temperature and heating, leading to the generation of available potential energy that is transformed to kinetic energy, generating the BOB MOV.

The evolution of the SAH, which is intensified by the TP forcing, can also influence the BOB summer monsoon onset. As mentioned above, before the BOB summer monsoon onset the TP forcing induces a surface anticyclone over the northern BOB (Fig. 7a), accompanied by significant local descent. As the descending air reaches the sea surface, westerly winds and strong divergence develop over the ocean with convergence over the Indochina Peninsula [89]. The moisture from the BOB is then transported northeastward by the strong low-level southwesterly winds induced by the TP forcing, where it converges near the northern Indochina Peninsula, leading to the local rainfall maximum. Condensational heating from the rainfall thus contributes to the northwestward migration and strengthening of the SAH in the upper troposphere during the premonsoon period. Owing to the development of the SAH, the upper-layer divergence on the southwestern side of the SAH increases over the southern BOB and is then coupled with stronger ascending motion in the middle troposphere (Fig. 8a). Such a pumping effect triggers the genesis of the MOV near the surface and intensifies the surface southwesterly flow

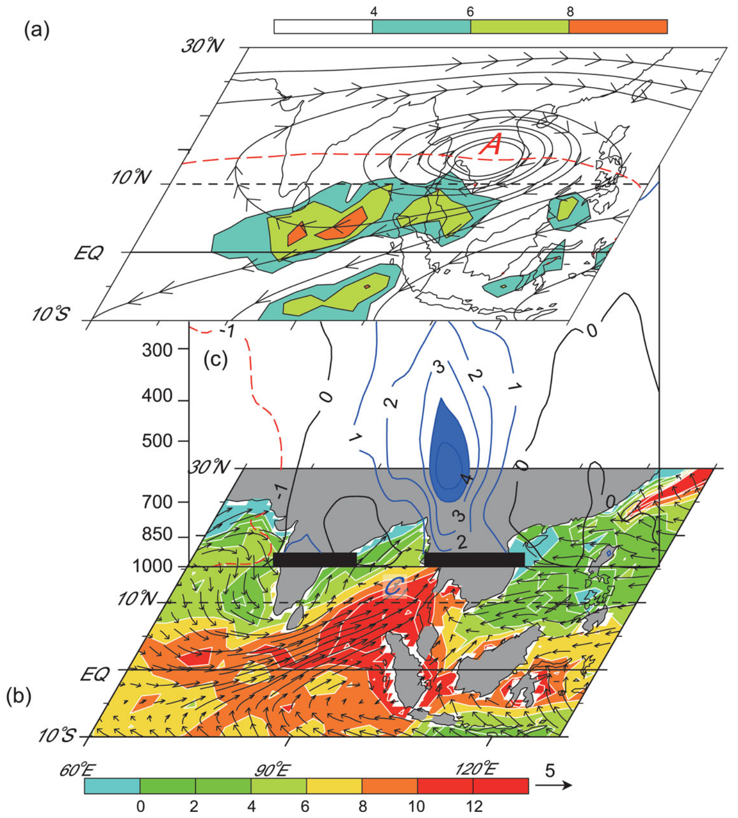

Figure 8. Schematic diagram of the vertical coupling between the upper and lower tropospheric circulations. (a) Due to the TP forcing in spring, the SAH in the upper troposphere is strengthened and migrates northwestward so that the upper-layer divergence pumping is located over the southeastern BOB. (b) In the lower layer, a strong SST warm pool is generated in the eastern BOB, favoring the development of convection. (c) Their coupling leads to vigorous baroclinic development and the BOB monsoon onset.

over the northeastern BOB. Subsequently, the development of onshore ocean currents and downwelling near the coast maintain the local warm SST and enhance the surface SH over the eastern BOB to further develop the MOV (Fig. 8b). Torrential rainfall occurs over the eastern BOB and western Indochina Peninsula with the vertical coupling of upper and lower circulation over the eastern BOB (Fig. 8c). As a result, the ASM commences over the BOB in early May. After the BOB summer monsoon onset, the SAH migrates onto the southern flank of TP [91].

INTRA-SEASONAL VARIABILITY IN THE ASM

10–20 day oscillation of PRES

The spring PRES over southern China undergoes a pronounced intra-seasonal oscillation (ISO) with a period of 10–20 days in most years [92]. Phase composite analyses demonstrate that the dry (wet) phase of the 10–20 day oscillation of PRES is characterized by the alternating occurrence of a large anomalous anticyclone (cyclone) encircling the TP in the lower troposphere, and anomalous low-level northeasterly (southwesterly) flow prevailing over southern China that produces lower tropospheric divergence (convergence). The structure and strength of the lower tropospheric circulation around the TP depend to a great extent on the TP thermal forcing associated with the surface SH [91]. During the driest phase (Fig. 9a), negative surface SH anomalies over the TP correspond to anomalous divergent winds flowing downslope toward the surrounding areas. An anticyclonic circulation encircling the TP in the lower troposphere, anomalous southwesterly flow over most of eastern China, and an anomalous northeasterly over the southern coastal region and SCS accompany the downslope flow. An east–west orientated band of surface divergence is located in southern China in association with the local anticyclonic circulation in the lower troposphere. Before the arrival of the wet phase, the positive heating anomaly extends southward from the northern TP and the surrounding air converges toward the TP due to its SHAP effect so that an anomalous cyclonic circulation is formed to encircle the entire TP in the lower troposphere. When positive SH anomalies completely cover the TP (Fig. 9b), the enhanced surface convergence over southern China results in the wettest conditions. Therefore, the generation and propagation of the 10–20 day ISO of PRES is modulated by the thermal state of the TP in spring.

Biweekly oscillation of the SAH

ISO (7–25 days) of convection also exists over the Yangtze and Huaihe River basins during the Meiyu season [93] in association with the significant influence of the TP and SAH.

In the upper troposphere, the SAH, as one of the important synoptic systems in the ASM region, exhibits evident biweekly oscillation, which is ascribed to the TP thermal forcing. Tao and Zhu [93], and Krishnamurti [94] identified a quasi-periodic oscillation of the SAH with a period of 10–16 days from spectral analysis. Previous studies have pointed out that such ISO in the upper troposphere over the monsoon region was associated with the forcing of diabatic heating [96]. Specifically, Liu *et al.* [62,63] found that if heating over the plateau led to a potential vorticity (PV) minimum that exceeded the observed mean by 1.5 or 2.0 times, an unstable flow would be developed to produce a quasi-biweekly oscillation. During this oscillation, the SAH changed from a single center over the southwestern side of

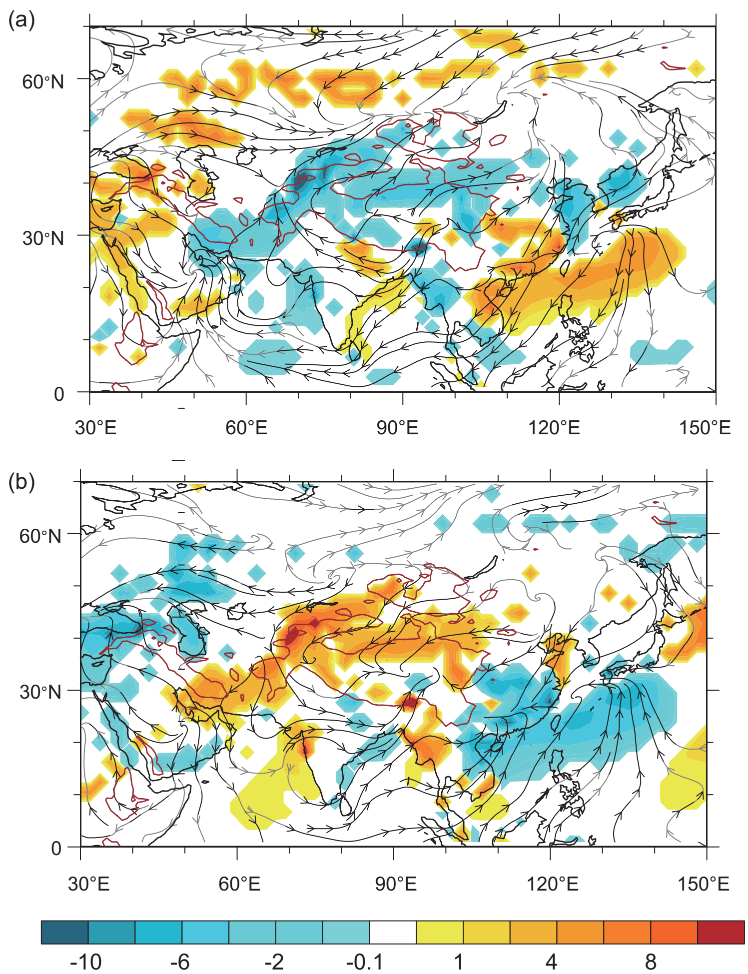

Figure 9. Composite of the 10–20 day filtered 10 m streamline and surface sensible heating flux (W m^{-2}, with shading denoting statistical significance at the 95% confidence level) at the driest phase (a) and wettest phase (b) of the 10–20 day ISO of the PRES over Southern China. The TP with terrain above 1500 m is outlined by a solid curve [92].

the plateau to a split (dipole) structure with centers over China and the Middle East, similar to the observed circulation variability in the same region [97]. Liu *et al.* [62] showed that the origin of such variability is due to the zonally extended PV minimum on a θ-surface [98], to a tendency for the PV above the heating over the TP to be reduced, and to the advection by the anticyclonic flow formed by the high-PV air from the east. On the other hand, the deep convection to the south and southeast of the plateau tends to suppress the quasi-biweekly oscillation because the low PV produced above it acts to reduce the reversal of the meridional PV gradient. The occurrence of the oscillation depends on the relative magnitude of the heating in the two regions [62,63].

DECADAL CHANGE IN HEAT SOURSE OVER TP AND ITS CLIMATE IMPACT

Many studies have reported climate warming over the TP during the second half of the 20th century [99–102]. In the vertical, there appears to be a coherent warming trend near the tropopause and a cooling trend in the lower stratosphere [26]. These features were reproduced in two coupled climate models forced by the observed 20th century CO_2 concentration, but not in model runs forced by CO_2 concentrations from Pre-Industrial Revolution. These results suggest that the recent climate warming over the TP may result from increasing emissions of anthropogenic greenhouse gas. The impacts of such gas emissions on climate change over the TP are probably stronger than elsewhere in the world [103].

From the 1980s to the end of the 20th century, the atmospheric heat source over the TP showed a significant trend characterized by weakened SH and enhanced cooling by radiation. Considering the small change in latent heating, the total atmospheric heat source over the TP, therefore, exhibited a decreasing trend, especially in spring [104–109]. The mechanism for the weakening SH trend in spring over the TP can be regarded as a regional response to the much larger warming amplitude in the mid and high latitudes over eastern Eurasia compared with that in the subtropics during the same period. The differential heating pattern induces a decrease in the meridional temperature and pressure gradients across the subtropics and a decelerated subtropical westerly jet stream over the region, as required by the geostrophic balance [110].

How does such a decadal change in the TP thermal forcing affect the EASM? Liu *et al.* [106] suggested that a weakened TP forcing in spring and summer is associated with a decreasing summer precipitation trend over Pakistan, North India, Bangladesh, and North China, and an increasing trend over northwestern China, South China, and Korea. Numerical experiments using an AGCM indicate that the reduction in upward surface SH over the TP leads to upper tropospheric cooling over the Eurasian continent, particularly in tropical and subtropical areas, and a weakened near-surface cyclonic circulation. The SH weakening scenario results in a weakened ASM circulation, characterized by an anomalous divergent anticyclonic circulation in the lower troposphere and an anomalous cyclonic circulation in the upper troposphere with a weakened western Pacific subtropical high. These changes in the lower tropospheric circulation lead to a weakened southerly flow over southern China, a convergence of anomalous water vapor transport in South

China, and a divergence of anomalous water vapor transport in North China. The circulation changes lead to an increase in precipitation in the south and reduced precipitation in the north within the East Asian monsoon region, consistent with previous observation results [109,111].

Snow cover or snow depth over the TP has been regarded as an early signal of the EASM anomaly. Zhang *et al.* [111] and Ding *et al.* [112] proposed that the TP winter snow has a significant positive correlation with the subsequent summer precipitation along the Yangtze River valley in China and in southern Japan after the late 1970s. Si and Ding [39] further argued that the shift in correlation patterns is closely associated with the decadal decrease in snow over the TP since the late 1990s. The increase in TP heating owing mainly to the decrease in winter snow, together with the decrease in SST in the tropical central and eastern Pacific, can possibly enhance the land–sea thermal contrast over East Asia in the subsequent spring and summer. However, the analysis of station observation data by Duan *et al.* [108] suggested that the SH and snow depth over the TP both decreased during 1980–2008, although only the trend in SH was significant. The similarity between the pattern of precipitation trends over China and the corresponding patterns of regression coefficients on the leading mode of the spring SH change over the TP demonstrates the distinct contribution of changes in spring SH over the TP. Analyses using observational data and numerical simulations have revealed that the reductions in SH over the TP have weakened the monsoon circulation and postponed the seasonal reversal of the land–sea thermal contrast in East Asia, consistent with the findings of Liu *et al.* [105]. While disagreements persist regarding the cause of variations in thermal forcing over the TP and the mechanism for the multi-scale variability in EASM, there is no doubt about the important influence of the TP on climate variability.

SUMMARY AND OUTLOOK

This paper reviewed the effects of mechanical and thermal forcing of the TP on both regional and global climate. In particular, recent studies have emphasized the unique role of the TP in driving the general circulation, large-scale climate patterns, and multi-timescale variability in the ASM.

The TP is a weak heat sink in winter but a strong, large, and elevated heat source in summer, which is characterized by the surface SH over the western TP and latent heating in the mid and upper troposphere over the central and eastern TP. In the presence of surface SH on the sloping surfaces, the heated air parcel at the sloping surface penetrate the isentropic surfaces and slide upward. The air in the lower elevation layers in the surrounding areas is therefore pulled into the plateau region, forming strong rising motion and even heavy rainfall over the TP.

In spring, the TP-SHAP contributes to the seasonal abrupt change of the Asian circulation and anchors the earliest ASM onset over the eastern BOB to the south of TP. In summer, the ascending induced by the TP heating is in phase with that produced by continental-scale forcing; the strongest monsoon and largest deserts are formed over the Afro-Eurasian continent. Moreover, the SH-induced vorticity source over the plateau forces a series of stream field anomalies in the form of a ray of Rossby waves downstream; hence, the SHAP over the TP affects not only the climate anomaly in the surrounding area but also the circulation over the Northern Hemisphere. Besides the unique role of the TP thermal forcing in the formation of ASM and large-scale climate regime in Eurasian continent, the intra-seasonal variability of SAH is also determined by the TP forcing to a considerable degree.

During the last two decades of the 20th century, the atmospheric heat source, particularly the spring SH over TP, shows a pronounced weakening trend which is induced mainly by the reduced surface wind speed. Such a decadal change in thermal forcing over the TP is responsible at least partly for the decreased summer precipitation over Pakistan, North India, Bangladesh, and North China, and the increasing trend over northwestern China, South China, and Korea.

In this review paper, we focus on the thermal forcing of the TP on the modern climate and its multi-variability within inter-decadal timescale. However, the involved dynamics and results are also applicable in understanding the paleo-climate.

Despite great progresses, many questions are still unresolved and challenges remain in the following aspects.

Observations are important for the study of climate, but there is currently a lack of adequate data for quantitatively understanding the role of diabatic heating over the TP, the relevant 3D structure, and the long-term temporal variation of circulation. Owing to their broad coverage in areas of complicated topography, the precision, content, and coverage of satellite datasets and *in situ* observations still do not satisfy the requirements of current studies. Obvious errors also exist in surface fluxes, energy budgets, cloud, and precipitation in the reanalysis data. To fully understand the effects of the TP on climate variations, it is necessary to establish more routine and automatic meteorological stations over the TP, particularly for data-scarce regions in the northwestern TP and at high elevations.

Regional and global CSMs are becoming increasingly important for understanding the nature of the climate system and its variability. However, the reliability of numerical simulations is limited by model uncertainties due to coarse resolution and the parameterization of physical processes such as cloud-radiation feedback, convection, and land processes.

The thermal effects of large-scale mountains are closely connected with their dynamic counterparts. For example, the elevated land surface of the TP obviously enhances the effect of the atmospheric heating source in summer. The different contributions of dynamic and thermodynamic effects of the TP to climate, and the intrinsic dynamical processes require further investigation. The questions of how human activity and natural climate variability together influence the TP thermal state and relevant forcing, and how changes in TP thermal forcing and oceanic forcing jointly affect regional and global climate are still unclear and should be further addressed.

In the future, international cooperation is necessary for studies regarding TP climate dynamics within the framework of the coupled land–ocean–atmosphere system. In 2013, 'The Third TP Observational Experiment' and a Key Program entitled 'Variation of the TP Land-air Coupled System and its Global Climate Effect' were launched by the China Meteorological Administration and the National Natural Science Foundation of China, respectively. These national programs will improve our understanding on the TP climate dynamics and weather forecast and climate prediction in the coming decade.

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