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Land–Climate Interaction Over the Tibetan Plateau

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Summary and Keywords

The Tibetan Plateau (TP) is the largest and highest plateau on Earth. Due to its elevation, it receives much more downward shortwave radiation than other areas, which results in very strong diurnal and seasonal changes of the surface energy components and other meteorological variables, such as surface temperature and the convective atmospheric boundary layer. With such unique land process conditions on a distinct geomorphic unit, the TP has been identified as having the strongest land/atmosphere interactions in the mid-latitudes.

Three major TP land/atmosphere interaction issues are presented in this article: (1) Scientists have long been aware of the role of the TP in atmospheric circulation. The view that the TP's thermal and dynamic forcing drives the Asian monsoon has been prevalent in the literature for decades. In addition to the TP's topographic effect, diagnostic and modeling studies have shown that the TP provides a huge, elevated heat source to the middle troposphere, and that the sensible heat pump plays a major role in the regional climate and in the formation of the Asian monsoon. Recent modeling studies, however, suggest that the south and west slopes of the Himalayas produce a strong monsoon by insulating warm and moist tropical air from the cold and dry extratropics, so the TP heat source cannot be considered as a factor for driving the Indian monsoon. The climate models' shortcomings have been speculated to cause the discrepancies/controversies in the modeling results in this aspect. (2) The TP snow cover and Asian monsoon relationship is considered as another hot topic in TP land/atmosphere interaction studies and was proposed as early as 1884. Using ground measurements and remote sensing data available since the 1970s, a number of studies have confirmed the empirical relationship between TP snow cover and the Asian monsoon, albeit sometimes with different signs. Sensitivity studies using numerical modeling have also demonstrated the effects of snow on the monsoon but were normally tested with specified extreme snow cover conditions. There are also controversies regarding the possible mechanisms through which snow affects the monsoon. Currently, snow is no longer a factor in the statistic prediction model for the Indian monsoon prediction in the Indian Meteorological Department. These controversial issues indicate the necessity of having measurements that are more comprehensive over the TP to better understand the nature of the TP land/atmosphere interactions and evaluate the model-produced results. (3) The TP is one of the major areas in China greatly affected by land degradation due to both natural processes and anthropogenic activities. Preliminary modeling studies have been conducted to assess its possible impact on climate and regional hydrology. Assessments using global and regional models with more realistic TP land degradation data are imperative.

Due to high elevation and harsh climate conditions, measurements over the TP used to be sparse. Fortunately, since the 1990s, state-of-the-art observational long-term station networks in the TP and neighboring regions have been established. Four large field experiments since 1996, among many observational activities, are presented in this article. These experiments should greatly help further research on TP land/atmosphere interactions.

Keywords: Tibetan Plateau, land/atmospheric interactions, Tibetan surface characteristics, Tibetan field measurements, Tibetan thermal forcing vs. orographic forcing, Tibetan snow effects, Tibetan land degradation

Introduction

The Tibetan Plateau (TP), the largest and highest plateau in the world, is located in Central and East Asia, mainly covering the Tibet Autonomous Region and the Qinghai Province in western China. With an average elevation exceeding 4,500 meters, the TP (also known as "the Roof of The World") stretches approximately 1,000 km north to south and 2,500 km west to east. The TP has the most complicated and prominent terrain on the globe and is the birthplace of ten large rivers in Asia—the Yangtze, the Yellow, the Mekong, the Irrawaddy, the Salween, the Ganges, the Brahmaputra, the Indus, the Amu Darya, and the Tarim—several of which are Asia's largest Rivers. They provide nourishment to the dry lands of Pakistan all the way to northern China, which is about 5,000 km away. Around two billion people and a third of the world's population depend on water from the TP. Because of this, the TP has been recognized as the "Asian water tower."

The land conditions in this region show strong spatial variability. In the central and eastern parts of the TP, the surface is typically characterized by alpine prairie and meadows, which occupy about 50% of the TP; its northern and western areas are covered by alpine and temperate deserts, while in the southeast TP, alpine boreal, evergreen, and deciduous forests are dominant (Figure 1).

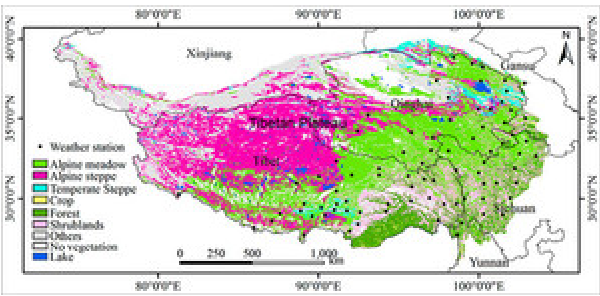

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Figure 1. Tibetan Plateau vegetation classification map (Shen, Fu, Yu, Sun, & Zhang, 2014).

An important geographical feature of the TP is thousands of lakes across the region. Many of the lakes are salt water lakes. The total lake area over the TP is about 15,000 km^2, whose water supply is mainly in the form of precipitation, surface runoff, and glacier melt (Shao et al., 2008). Another unique geographical feature is permafrost, snow, and glaciers that cover large areas on the high plateau. The TP is sometimes called the "Third Pole" because its high altitude allow it to contain some of the largest volumes of ice outside the Polar Regions. It contains 36,800 glaciers, amounting to a total glacial area of 49,873 km^2 and a total glacial volume of 4,561 km^3 (Yao, Pu, Lu, Wang, & Yu, 2007). The frozen ground of the TP is about 1,740,000 km^2 and accounts for 55–60% of the total permafrost area of the Northern Hemisphere (Cheng, 1990; Zhang, Barry, Knowles, Heginbottom, & Brown, 1999).

The complicated land conditions on this unique geomorphic unit produce complex interactions between the lithosphere, hydrosphere, cryosphere, biosphere, atmosphere, and anthrosphere (Yao et al., 2015). This article mainly presents the TP land processes and land/atmosphere interactions.

Characteristics of Land Surface Processes Over the Tibetan Plateau (TP)

Substantial Diurnal Variations and Distinct Seasonal March of Surface Radiation and Heat Fluxes in the Central and Eastern TP

Due to higher elevation and a relatively clean atmosphere, downward shortwave radiation is much larger over the TP than in other areas. For example, the solar irradiance over the Plateau is often observed to exceed 1,200 W m^{-2} near noon (Ma et al., 2005), which results in a very strong diurnal change of the surface energy components: upward shortwave and upward long-wave radiation, net radiation flux, soil heat flux, sensible and latent heat fluxes, and other near-surface meteorological variables. For instance, the diurnal range of the surface skin temperature can exceed 60 K (Yang, Chen, & Qin, 2009).

Figure 2 shows typical diurnal cycles in an automatic weather station, NPAM, (91º 42′1″E, 31º 54′15″N; Elevation: 5,063 m), which were measured during the GEWEX (Global Energy and Water cycle Experiment) Asian Monsoon Experiment on the TP (GAME/Tibet) (Ma et al., 2005).

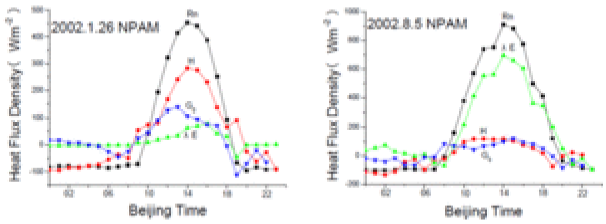

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Figure 2. The typical diurnal variations of heat fluxes in dry season with no snow cover (January 26, 2003) and wet season (August 5, 2002) in the Automatic Weather Station NPAM (91º 42′1″E, 31º 54′15″N, Elevation: 5063 m), which is listed as MS3478 in Figure 7. Note: Rn: net radiation; H: sensible heat flux; λE: latent heat flux; G_0: surface soil heat flux. Unit: w m^{-2}. The figure is based on Ma et al. (2005).

During the summer monsoon season (Figure 2B), net radiation at the surface could reach 1,000 W m^{-2} at noon. Most energy goes to latent heat flux, i.e., evapotranspiration. Different from many other regions, the ground heat flux is comparable to the sensible heat flux, leading to a very high surface temperature in the daytime. At night, both sensible heat and ground heat fluxes turn negative. During the winter dry condition (Figure 2A), net radiation at noon is only about half of that observed in the summer; sensible heat flux becomes dominant with the ground heat flux being the second largest component. Latent heat flux is rather small. At night, the negative sensible heat flux is quite large, around 100 W m^{-2}.

A unique feature of the TP surface eddy process is the high surface resistance for the heat flux and the momentum flux and its diurnal variation. The observed eddy covariance flux data collected over different TP land surface covers (bare soil, short grass, and forest) all show larger values of average heat bulk transfer coefficients of resistance than those of average momentum bulk transfer resistance (Wang & Ma, 2011), different from traditional turbulence transfer processes, in which momentum exchange is more efficient than heat transfer. The excess resistance to heat transfer also has apparent diurnal variation over the land surfaces of some in situ sites. It shows lower values at night and higher values in the daytime, especially in the afternoon. In the nighttime for relatively smooth surfaces, heat transfer efficiency may exceed that of momentum transfer.

The intraseasonal and seasonal variation of the downward shortwave radiation and net radiation at the surface are shown in Figure 3, which shows the monthly mean diurnal variation from 2002 to 2003.

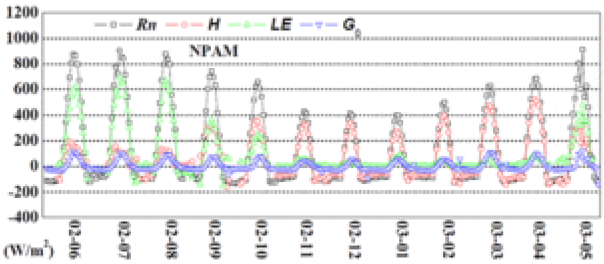

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Figure 3. Monthly diurnal variations of land surface heat fluxes at the NPSM station. Rn: net radiation; H: sensible heat flux; λE: latent heat flux; G_0: surface soil heat flux. Unit: w m^{-2}. The figure is based on Ma et al. (2005).

Before the onset of the monsoon in May, the surface is relatively dry and the sensible heat flux dominates the surface energy budget. After the onset, the land surface becomes wet due to frequent monsoonal rainfall events; the latent heat flux dominates the energy budget until the withdrawal of the monsoon in September. The sensible heat flux and latent heat flux are dominant in different seasons in the partitioning of net radiation flux on the TP.

The Contrast Between the Dry Western Region and the Wet Eastern Region

The various landscapes described in the INTRODUCTION correspond to the unique plateau climate. In large parts of the central and eastern TP, the annual precipitation amount is more than 200 mm, with a sharp northwest-southeast gradient (Figure 4D); the southeastern TP has the highest precipitation.

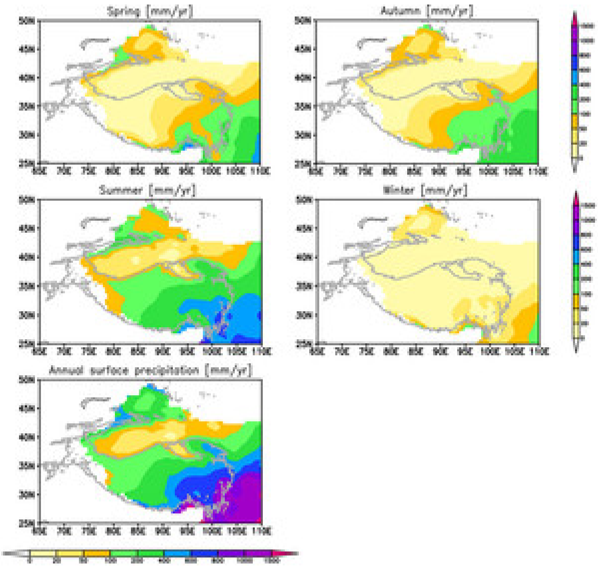

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Figure 4. Annual and seasonal precipitation (mm yr^{-1}). Based on Chinese Meteorological Administration Precipitation Data (Li, Guo, Xue, Fu, & Qiu, 2016). Spring: March–April–May; Summer: June–July–August; Autumn: September–October–November; Winter: December–January–February.

In the western and most northern TP, where alpine steppe and desert are dominant, the annual precipitation is less than 200 mm; over some areas, it is even less than 50 mm (Figure 4).

The TP precipitation also shows a strong seasonality (Figure 4). Summer rainfalls account for most of the total annual. The high precipitation in the southeast TP corresponds to the strong moisture transfer by the monsoon flow. The summer precipitation over the eastern TP has a strong diurnal variation with its local peak in the mid-afternoon to early evening (Bao, Zhang, & Sun, 2011; Liu, Feng, Chu, Zhou, & Ueno, 2002). The winter season is uniformly dry. However, the precipitation that occurs over western TP as the result of the orographic uplift of the westerly flow (Maussion et al., 2014) was not presented in Figure 4 due to sparce data for that area. During the spring and autumn, except the southeast TP, most parts of the TP are dry. The “precipitation barrier” induced by the Himalayan range is apparent (Maussion et al., 2014). The average values for the TP precipitation are 211.9 mm yr^{-1}, 55.8 mm yr^{-1}, and 399.6 mm yr^{-1} for a year, December–January–February (DJF), and June–July–August (JJA), respectively. The precipitation shows larger interannual variability and decadal trends, but there are large differences over different areas. Precipitation has increased in most parts of the TP since the 1970s, especially in the eastern and central TP, while the southern TP along the Himalaya Mountains has been drying during the same period (Xu, Gong, & Li, 2008B; Yao et al., 2013).

Figure 5 shows the spatial distribution of latent heat flux for four seasons, which was derived from Advanced Very High Resolution Radiometer (AVHRR) satellite-derived data (Ma et al., 2015).

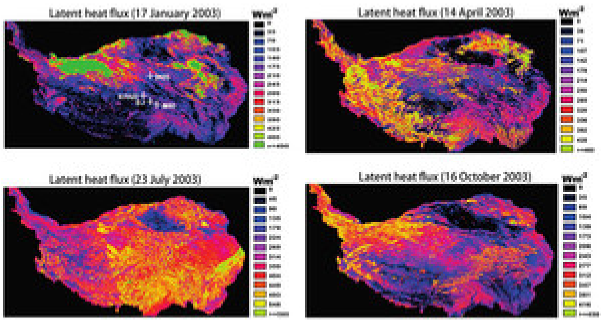

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Figure 5. Spatial latent heat flux (w m^{-2}) distribution for four seasons (Ma et al., 2015).

The pattern and seasonal variation are consistent with precipitation. The average latent heat fluxes over the TP were 170.9, 243.6, 395.0, and 225.8 W m^{-2} for January, April, July, and October, respectively, in 2003. The summer season had higher latent heat flux than other seasons, mainly in the central and eastern TP, where the land surface was wet and was covered by green and growing vegetation. During spring and winter, the northern and western TP had high latent heat flux, where there are glaciers, ice, snow, permafrost and seasonal frozen ground, etc., which may contribute to the complicated spatial distribution of latent heat flux during the spring.

Development of the Convective Atmospheric Boundary Layer (CABL)

The rapid increase in net radiation after sunrise results in surface heating and the development of the atmospheric boundary layer. In the plateau, the profiles of temperature and humidity are strongly affected by intensive convective activities. The composite temperature profile and water vapor profile are used to investigate the ABL evolution. Figure 6 shows the composite sounding profiles during June 4–14, 1998 (a dry period) for a CABL measurement at the Tibet Amdo site (32°15′50″N, 91°40′50″E; Elevation: 4,710 m), which is located in the central Tibetan Plateau between the Tanghla and the Nyainqentanghla Mountains.

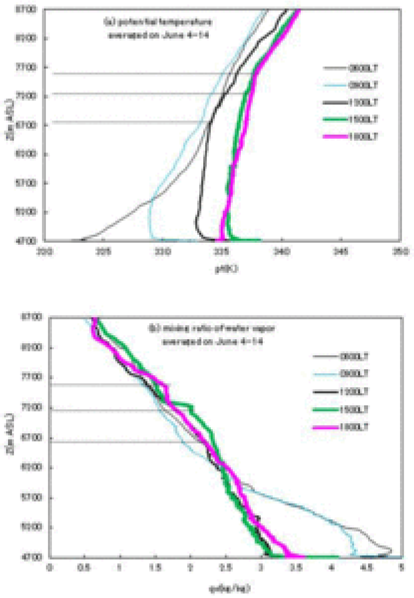

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Figure 6. CABL development: (a) Potential temperature (K); (b) mixing ratio of water vapor g kg^{-1}. Figure is based on Yang et al. (2004).

The maximum CABL height there is about 3 km above ground level, which is higher than previously estimated in the first Tibetan Plateau Atmospheric Scientific Experiment (TIPEX) for other Tibet sites (e.g., Ye & Gao, 1979) due to the Amdo station's high elevation. Over the lowlands, the CABL height is usually no more than 1.5 km. Three distinct periods in CABL development have been roughly identified in Figure 6A. (1) Period I (sunrise to mid-morning): the near-surface layer becomes warm quickly due to surface heating, but the upper layer becomes slightly cool due to atmospheric longwave emissions. The CABL extends to around 500 m above ground. (2) Period II (mid-morning to mid-afternoon): The CABL heats quickly and the CABL height extends rapidly. The height extends up to 2.0 km at 1200 local time and reaches about 3 km at 1500 local time. (3) Period III (mid-afternoon to late afternoon): the height of the CABL is stabilized. In the process, the specific humidity under 1 km was rapidly mixed during 9 a.m.–12 noon; there was not much diurnal variation above 1 km (Figure 6B). The ABL is characterized as a well-mixed layer of potential temperature and a nonuniform layer of specific humidity (Figure 6, Yang et al., 2004).

Other radiosonde data from a western TP station on a typical sunny day showed that after morning heating, CABL had grown to 3,600 m above ground by noon, eroding both the surface and residual layers, and after six hours this growing layer was 4,780 m deep (Ma et al., 2015). During the fast development of the CABL, the water vapor content and wind speed were well mixed by turbulence. These results suggest buoyant turbulence dominating development of the CABL.

Large Scale Field Measurements Over the TP and Corresponding Land Model Development

The IPCC AR4 reported that all the global climate models underestimated surface air temperature in the TP, and most models overestimated its precipitation (Hao, Ju, Jiang, & Zhu, 2013). Yang et al. (2009) compared four numerical weather prediction models over the TP and found that they had quite large discrepancies for simulating surface energy budgets. In addition, Takayabu et al. (2001) found that large uncertainties existed in the Bowen Ratio in an intercomparison of multiple offline land surface models, which were driven by observed meteorological variables to produce the surface fluxes. The energy transfer processes of the TP differ significantly from those in regions of lower altitude, as discussed in CHARACTERISTICS OF LAND SURFACE PROCESSES OVER THE TIBETAN PLATEAU (Smith & Shi, 1992; Zhang, Wu, Zhao, & Gao, 2014). To better understand interactions between TP land surfaces processes with atmosphere/monsoon, it is imperative to have adequate information regarding the land-surface energy and water balance as well as boundary layer and free-atmosphere dynamics over the TP.

Due to high elevation and harsh climate conditions, measurements over the TP used to be sparse. The TP is notorious for its lack of climate observations, which cripples the predictive power of land surface and global climate models for the region (Qiu, 2013). The First TIPEX was launched from May to August in 1979 to investigate the role of the TP in global atmospheric circulation. This experiment, for the first time, provided a qualitative understanding of general aspects of TP meteorology and its role in the Asian Monsoon (Ye & Gao, 1979; Xu, Zhang, et al., 2008). Since the 1990s, state-of-the-art climate observatories and long-term research stations in the TP and neighboring regions have been established. A large amount of data sets for energy and water cycles over the TP have been collected from in situ observations, which provide useful information about TP surface hydro-meteorology and are essential to elucidate climate processes and validate satellite data and models. With the vast TP landscape and relatively less human impact, the TP provides a valuable natural field laboratory for land/atmosphere interaction studies.

Four major large-scale field measurements, among many observational activities, have been carried out since 1996. They are the GAME/Tibet (1996–2000, Ma et al., 2005, 2008); the Coordinated Enhanced Observing Period (CEOP) Asia–Australia Monsoon Project on the Tibetan Plateau (CAMP/Tibet, 2001–2006, Ma et al., 2008); the New Integrated Observational System over the Tibetan Plateau (NIOST, since 2005, Xu, Zhang, et al., 2008); and the Third Pole Environment (TPE) Tibetan Observation and Research Platform (TORP, since 2009; Ma et al., 2008, 2009). Figure 7 illustrates the spatial distribution of major stations that are presented in this section and photos for some of these stations.

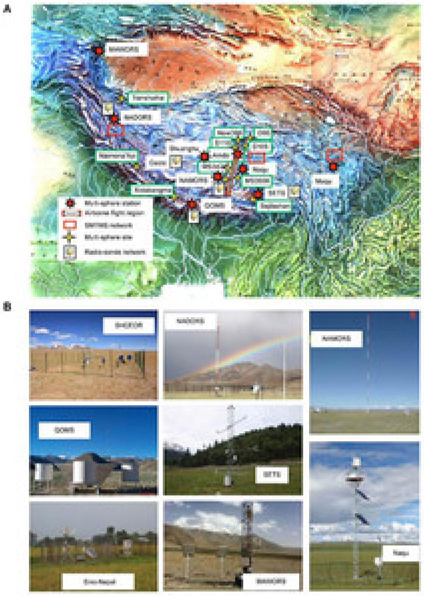

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Figure 7. (a) The spatial distribution of major stations of water-cryosphere-atmosphere-biology interactions over the TP; (b) the typical observation instruments for different land covers.

The TP field data set can be found at http://en.tpedatabase.cn/.

GAME/Tibet (1996–2000, Ma et al., 2005, 2008)

The GAME/Tibet program was carried out by scientists from Japan, China, and other Asian countries. The goal of the program was to clarify the interactions between the land surface and the atmosphere over the TP in the context of the Asian monsoon system. To achieve the scientific objectives of GAME-Tibet, a meso-scale observation network (150 x 250 km, 91°–92.5°E, 30.5°–33°N) was implemented in the central plateau where short grass was dominant. The field measurement network was mainly comprised of: (1) Amdo planetary boundary layer (PBL) station and NaquFx PBL station with a radiation observational system, turbulent flux measurement, soil temperature and moisture measurement, and radiosonde observation system; (2) a network of five Automatic Weather Stations (AWS. D66, TTH, Naqu, D110, and MS3608 stations) and two sets of flux-PAM (Portable Automated Meso-net) stations (MS3478 and MS3637 stations); (3) a soil temperature and soil moisture network (SMTMS), which consists of the Amdo PBL site, D66, TTH, D110, WADD, NODA, MS3478, MS3608, and MS3637 stations; and (4), a precipitation gauge net, which consists of the D66, TTH, D110, WADD, NODA, AQB, MS3478, MS3543, Zuri, NaquFx, MS3608, Naquhy, NaquRS, and MS3637 stations.[1]

CAMP/Tibet (2001–2006, Ma et al., 2008)

CAMP/Tibet was a continuation of the GAME/Tibet with more instruments. The objective of the project was to understand the land–atmosphere interaction process in the following four specific aspects: (1) to obtain the characteristics of the land–atmosphere interaction process systematically; (2) to understand the meteorological and hydrological process quantitatively; (3) to develop large-scale adaptive land surface process models; and (4) to develop and validate the methods for deriving land surface parameters. In addition to the instruments from GAME/Tibet, the CAMP/Tibet further included four AWSs (the D105, MS3478, BJ, and ANNI stations), one Airborne microwave radar, two microwave radiometers, one wind profiler plus Radio Acoustic Sounding System (RASS) in BJ, two turbulence towers in BJ and ANNI, and two deep soil temperature observation systems.

NIOST (Since 2005, Xu, Zhang, et al., 2008)

As one of the key initiatives of China–Japan intergovernmental cooperation, the New Integrated Observational System over the TP (NIOST) has been in development since 2005 by scientists from the Chinese Meteorological Administration (CMA), as well as the University of Tokyo, Japan. NIOST focuses on China and East Asian countries' weather/climate monitoring and forecasting needs and on energy and water cycle studies with key issues addressed on the predictability of disastrous weather, the role of geographic structure of the plateau on energy and water circulations, and climate change related issues. The NOIST datasets include traditional meteorological elements (wind, temperature, pressure, and moisture), energy budget components (net radiation, sensible and latent heat fluxes), total precipitable water, and various satellite products. The NIOST mainly contains 66 AWS stations, 6 PBL towers, 16 GPS sounding sites, 24 GPS stations, and two regional wind profiler stations.

Tibetan Observation and Research Platform (TORP, since 2009, Ma et al., 2008, 2009)

The TORP was set up under the Third Pole Environment (TPE) Program and the Chinese Academy of Sciences (CAS) in 2009. TORP focuses on the TP's land surface, land–atmosphere, and environmental processes. All the instruments available in GAME/Tibet and CAMP/Tibet are continuing for long-term observations in TORP. Moreover, more observation stations are included for land–atmosphere processes on different types of landscapes and over all parts of the TP.

TORP contains 27 comprehensive observation and research stations and 11 additional observational sites to monitor atmospheric processes from surface layer to the stratosphere and ground-surface processes. The additional sites include Mt. Everest (gravel), Nam Co (plateau meadow), Linzhi (high grass in a valley), Ali (desertification grassland), Mt. Mushitageta, Mt. Tanggula, Shuanghu (sparse meadow), and the Haibei, Golmud, Lhasa, and Mt. Gongga stations. Each comprehensive observation and research station includes 20-m ABL towers (wind speed, wind direction, air temperature, and humidity at five levels), a four-component radiation measurement system, a five-level SMTMS, a GPS radiosonde system, a RASS, a sonic turbulent measurement system, a CO_2/H_2O flux measurement system, a precipitation monitoring system, and a soil heat flux measurement system. Each additional observational site includes a 10-m AWS (wind speed, wind direction, air temperature, and humidity at three levels), a radiation measurement system, a SMTMS, a precipitation and snow depth system, and a soil heat flux system. One of the important parts of TORP is the meso-scale monitoring network (see Figure 7A), which had been used in GAME/Tibet (1996–2000) and CAMP/Tibet (2001–2006) and is continued in TORP. Furthermore, several comprehensive observation and research stations have also been set up in surrounding regions, such as Nepal, Pakistan, and Tajikistan since 2010.

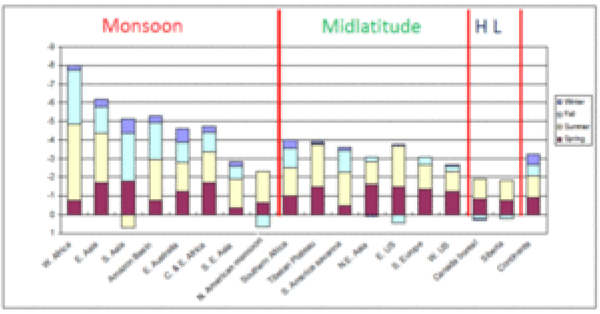
TP Land Model Development

Using field data, substantial research progress has been made in TP land and atmosphere process modeling. These field observations have played a fundamental role in providing large amounts of data for plateau scientific research and improving the quantitative understanding of land–atmosphere interactions over the TP. These data helped to develop land surface process models and data assimilation systems over large spatial scales, as well as to develop and validate satellite-based retrieval of land surface parameters (e.g., Qin et al., 2013; Zhao et al., 2014). Among them, the diurnal cycle of surface skin temperature, surface energy budget, and soil moisture are focuses (e.g., Chen, Yang, Zhou, Qin, & Guo, 2010; Yang et al., 2008; Zeng, Wang, & Wang, 2012). Since many land models have neglected the difference between thermal and aerodynamic roughness lengths and their diurnal variation, which has led to inadequate diurnal variation of TP surface temperature and improper net radiation, soil heat flux, and sensible heat flux, a new scheme based on observational data was proposed to reproduce the diurnal variation of surface resistances (Yang et al., 2008).

The cryosphere is a crucial part of the TP and directly interacts with the overlying atmosphere. The large area of snow and frozen ground in winter over the TP greatly influences hydrological and thermal processes (Niu & Yang, 2006; Yamazaki, Kubota, Ohata, Vuglinsky, & Mizuyama, 2006). A comprehensive snow scheme that includes snow layering and compaction processes (Sun, Jin, & Xue, 1999) has been introduced to the water and energy budget-based distributed hydrological model to evaluate the seasonal variation of spatial distribution of snow cover and land surface temperature over a Himalayan region at 1-km spatial resolution (Shrestha, Wang, Koike, Xue, & Hirabayashi, 2010, 2012; Wang, Koike, Yang, Jackson, et al., 2009). Furthermore, since permafrost ground covers large parts of the TP, it is necessary to include a frozen soil scheme for the simulation of soil temperature and moisture and the surface energy and water budget. Ice crystals in the frozen ground lessen soil infiltration capacity. As such, a higher proportion of snowmelt and spring rainfall is partitioned into surface runoff (Stähli, Jansson, & Lundin, 1999). Moreover, the phase transitions between the ice and liquid water alter the energy distribution in the soil and have an important influence on the snow layer above (e.g., Niu & Yang, 2006; Wang et al., 2009A). Therefore, the physical processes of snow and frozen soil should be explicitly addressed in land surface hydrological modeling. Zhang, Sun, and Xue (2007) and Li, Sun, and Xue (2010) have developed a comprehensive multilayer soil scheme for climate study, including water flow and heat transfer in permafrost soil with water phase change and validation using TP observational data. Zhang et al. (2013) further combined snow and frozen soil processes with hydrological processes by coupling a simultaneous heat and water model with a geomorphologically based distributed hydrological model. TP field measurement data have provided valuable information for these modeling efforts. Furthermore, due to the important contribution of glacial melt runoff in surface hydrology, a multilayer energy balance scheme for clean glacier and a single layer energy balance scheme for debris-covered glacier has included in the distributed hydrological model to study the surface water balance in the river basin (Shrestha et al., 2015). These land surface parameterization developments and their application have advanced understanding of land processes in this region (Hu, Ye, Zhou, & Tian, 2006; Ma, Tsukamoto, Wang, Ishikawa, & Tamagawa, 2002; Tanaka, Tamagawa, Ishikawa, Ma, & Hu, 2003; Yang, Koike, Ye, & Bastidas, 2005). However, these parameterizations have not been well integrated into state-of-the-art global Earth System models and regional climate models. There is still a big gap between these experimental studies and predictive Earth system model development over the TP region.

Development of TP Land–Air Interaction Studies and Controversies

Analyses of satellite-derived land surface products and modeling results have been applied to evaluate the role of land surface processes in the climate system. Xue et al. (2004, 2010) applied General Circulation Models (GCMs) coupled to different land surface parameterizations with varying degrees of physically based complexity in the representation of land surface processes to assess the global and temporal characteristics of seasonal climate/land surface processes interactions. The importance of land effects on climate was assessed based on the accuracy of simulations of observed precipitation by the GCMs with different land parameterizations. Since a more realistic representation of the land surface in a GCM should improve precipitation simulations if land has effects on the real climate system, the statistically significant reductions of absolute bias and root-mean-square errors (RMSE) between simulated precipitation and the observation were adopted as criteria to identify land effects. Figure 8 shows the reduced absolute annual mean bias of precipitation simulations (or improved prediction accuracy) over different regions over the world due to more realistic land representations in the GCMs.



Red: spring; Yellow: Summer; Cyan: Fall; Light blue: Winter

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Figure 8. The reduced absolute annual mean bias of precipitation (mm day^{-1}) simulations (or improved prediction accuracy) due to land processes in the GCM. Based on Xue et al. (2010).

The TP has been identified as the region with the largest impact of land surface processes on the climate system in the mid-latitude regions. Further analyses reveal that in the TP, land surface processes have the most impact during the summer and the spring, with some impact during fall. Therefore, the TP is a hot spot in the world for land/atmosphere interaction. In the following subsections, the effects of TP land heating, snow cover, and land degradation on the Asian monsoon are discussed, and some controversies regarding the land effects are also presented.

Plateau Heating vs. Himalayan Orographic Insulation

Scientists have long been aware of the role of the TP in atmospheric circulation. Earlier, thermal influences of the TP were inferred from the presence of a huge upper tropospheric anticyclone above the Plateau in summer (Flohn, 1957), heat balance in the lower troposphere (Yeh, Luo, & Chu, 1957), and from the pronounced diurnal variations in the surface meteorological elements along the periphery of the Plateau (Ye & Gao, 1979). More diagnostic and modeling studies have also confirmed that the TP provides a huge, elevated heat source to the middle troposphere, and the sensible heat pump plays a major role in the regional climate and in the formation of the Asian monsoon (Wu et al., 2007). The thermal effect of the TP is evident from positive temperature anomalies over the Plateau. The air above the Plateau is always warmer than the air over the surrounding areas when compared at the same level and at the same latitude. There is a large-scale vertical circulation thermally induced by the Plateau. The rising motion is seen near the surface of the western Plateau in December and spreads over the entire Plateau as the season progresses. An “explosive” expansion of warm air over the Plateau occurs from late spring to early summer over the areas extending from the eastern Plateau to southern China, leading to reversal of the meridional temperature gradient south of the Plateau. This warming causes the South Asian anticyclone to shift northward and the low-level south-westerlies to develop in the Indian Ocean. The area of ascent over the Plateau and that with the migrating rainfall belt merge together, extending the monsoon rains to the north. Analyses of the warming processes of the upper troposphere show that diabatic heating is the primary cause of the temperature increase over the eastern TP during the monsoon onset. The sensible heat flux from the ground surface is identified as the dominant factor in the heat budget of the TP. There is also a large diurnal change of temperature over the Plateau. A deep, nearly mixed boundary layer with high potential temperature is observed in the evening, suggesting the role of thermal convection in the upward transport of heat (e.g., Luo & Yanai, 1984; Seto, Koike, & Rasmy, 2013; Tamura & Koike, 2010; Tamura, Taniguchi, & Koike, 2010; Webster et al., 1998; Wu & Zhang, 1998; Yanai, Li, & Song, 1992; Yanai, Wu, & Wang, 2006).

Although this view that the TP's thermal and dynamic forcing drives the Asian monsoon has been prevalent in the literature for decades, a new spate of studies has been challenging it in recent years. Studies suggest that the south and west slopes of the plateau produce a strong monsoon by insulating warm and moist tropical air from the cold and dry extratropics, while a direct heat source cannot be considered as a factor for driving the Indian monsoon (e.g., Boos & Kuang, 2010; Chou, Neelin, & Su, 2001; Privé & Plumb, 2007). In fact, there have been a number of studies focusing on the TP's orographic effect which show a stronger South Asian summer monsoon owing to the presence of the TP (e.g., Hahn & Manabe, 1975; Kutzbach, Guetter, Ruddiman, & Prell, 1989; Yasunari, Saito, & Takata, 2006). However, in these studies, thermal effect was also considered as a major factor producing the monsoon.

Different from these early TP orographic effects, Boos and Kuang (2010, 2013) argue that the Himalayas' mechanical forcing exerts a blocking effect on the colder and drier air advection from the north. They further suggest that the high surface entropy over North India plays a more important role than the TP does in producing the local monsoon rainfall and upper warm center over North India. Figure 9C shows that after surface elevations to the north of the Himalayas are set to zero, June–August upper level temperature and surface precipitation can still be properly simulated by their GCM.

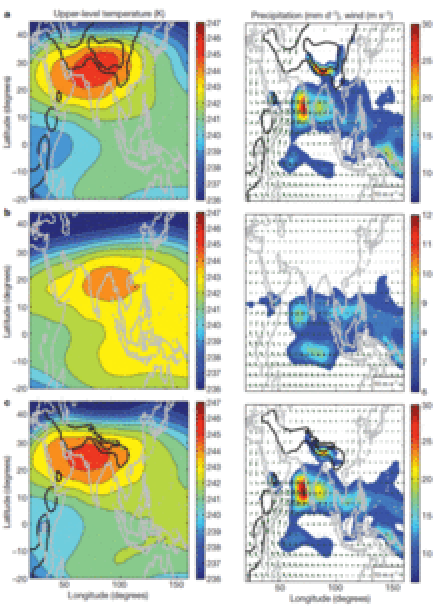

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Figure 9. Simulated June–August upper level temperature, precipitation, and wind from the atmospheric model: (a) standard topography; (b) no elevated topography; and (c) surface elevations to the north of the Himalayas set to zero, which demonstrates the effect of orographic forcing of the Himalayan south slope. Figures are adapted from Boos and Kuang (2010).

Its effect is similar to the one in which the full TP exists, including its thermal forcing (Figure 9A). The new, alternative hypothesis on Tibet has produced a challenge to understanding the South Asian monsoon comprehensively and encouraged more investigation. In responding to this challenge, Wu et al. (2012) emphasize that the sensible heat-driven pump effect of large mountains drives low-level moisture advection from the Indian Ocean to the southern TP to support the high surface entropy over North India and maintain the north branch of the Asian summer monsoon. Because high surface entropy requires high surface potential temperature as well as high specific humidity, both local surface heating and water vapor transport from ocean to land are required. Figure 10, which is based on Wu et al.'s study (2012), demonstrates that the Asian summer monsoon is thermally controlled.

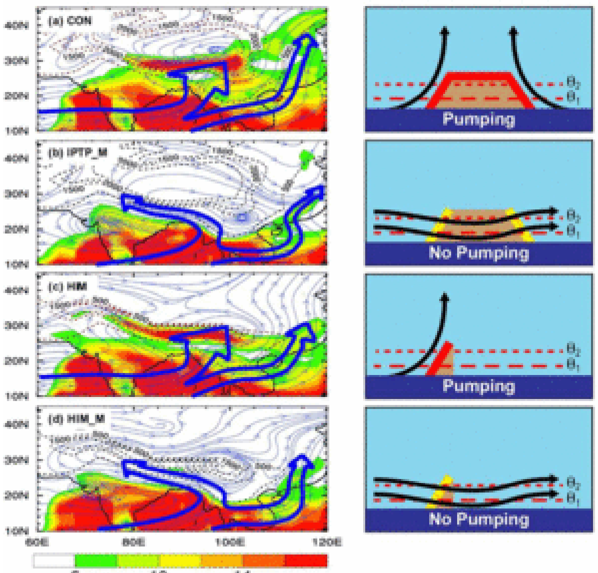

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Figure 10. Simulated June–August precipitation (color shading mm·d^{-1}) and wind flow at σ = 0.89 (left column) and corresponding mechanisms (right column): (a) the control experiment; the bold red lines in the right column on the Iranian-Tibetan Plateau surface represent heating; (b) the experiment with the mechanical forcing but without the surface sensible heating above the Plateau; (c) the experiment with the Himalaya heating and topography but without the TP plateau; and (d) the Himalaya topography only (the same as c but without the surface sensible heating above the Himalayas). Dark blue open arrows in the right column denote the main atmospheric flows impinging on the TP, either climbing up the plateau (a and c) or moving around the plateau, parallel to orographic contours (b and d). Figures are adapted and modified from Wu et al. (2012) and Wu et al. (2007).

Figure 10D shows that in their GCM simulation, when only the Himalayas' slope mechanical effect exists, the Asian monsoon flow is not comparable with the one that was produced by the thermal heating (Figures 10A and 10C). Based on the remote sensing-derived radiation analysis and numerical experiments, He, Wu, Liu, and Bao (2015) further argue that in the South Asian summer monsoon sector the mid-latitudes and subtropics receive more solar energy and possess higher surface potential temperature than North India during the summer monsoon season, and there is no cold/dry advection from the north into the Indian monsoon area. That said, the TP shielding does not exit. Others suggest that the fundamental problem in understanding the TP thermal and orographic effects is severe limitation of climate modeling in mountain regions like the TP and Himalayas due to complex topography and complicated ways it affects climate (Qiu, 2013). In fact, both Boos and Wu acknowledge that coarse resolutions in their models may be a major constraint to simulating TP effects adequately.

The Effects of TP Snow Cover

Snow, one of four components in the cryosphere, has long been considered an important factor in influencing variability of terrestrial hydrologic systems over a variety of time scales. Through change of surface albedo and regulation of turbulent heat and momentum fluxes at the surface, snow modifies the exchange of energy between the land surface and the atmosphere and significantly affects the distribution of diabatic heating in the atmosphere. Snow cover is also an effective insulator of the soil thermal column. In addition, snow melting represents an effective heat sink for the atmosphere and an important source of surface runoff and moisture for the soil.

Snow effects are a major subject of Asian land/atmosphere studies. As early as in 1884, Blanford suggested a possible negative influence of winter snowfall in the Himalayas on the drought season in Northwest India. Walker (1910) extended Blanford's study and addressed an out-of-phase relationship between the Himalayan snow depth at the end of May and the Indian summer monsoon rainfall. The identified Himalayan snow–Indian monsoon relationship they identified was further supported by other studies, for instance, Dey and Bhanu Kumar (1983) and Liu and Yanai (2002), in which the former used the December to March snow cover and the latter used April to May snow cover. Shukla and

Mooley (1987) proposed a physical mechanism for the snow–monsoon relationship, which suggests that the memory of a winter snow anomaly in the climate system resides in the wetness of the underlying soil as snow melts during spring and summer, which influences the ground temperature in the subsequent summer and modifies the regional monsoon.

However, Bamzai and Shukla (1999) found no significant relationship between the Himalayan seasonal snow cover and subsequent Indian monsoon rainfall using 22 years of satellite data. Some studies even show positive correlation between snow cover over the TP and the Indian monsoon (Lau & Li, 1984; Robock, Mu, Vinnikov, & Robinson, 2003), opposite to that of Blanford (1884). Furthermore, it has been found that the relationship between snow cover over the western Himalayas and the Indian monsoon has recently undergone a change in sign (Kripalani, Kulkarni, & Sabade, 2003). Zhao and Moore (2004) found that the correlations between snow cover over the eastern and western regions of the TP and Indian monsoon are of opposite signs. The different variability in eastern and western snow cover may contribute to the different TP snow and Indian monsoon relationship. Furthermore, studies (Robock et al., 2003; Liu & Yanai, 2002) indicate that while such correlations were high during some decades, they were rather low during others.

The snow cover's effect on the East Asian summer monsoon rainfall has also been studied. Using snow depth data at 60 stations over the TP for the period 1960–1998, Wu and Qian (2003) documented that the cold season (November–March) average snow depth over the TP has a positive correlation with June–July–August (JJA) rainfall in central China. Meanwhile, using station data from 1973–2001, it was found that an increase of April–May snow cover depth over the TP is associated with a decrease of June rainfall in the Yangtze and Hwai Rivers and an increase of rainfall in southeastern China (Zhao, Zhou, & Liu, 2007). The simultaneous relationship between summer snow cover anomaly over the TP and precipitation over the Mei-Yu–Baiu region has also been identified (Liu et al., 2014). In addition, Zhang et al. (2004) found a close relationship between the interdecadal increase of snow depth over the TP during March–April and a wetter summer rainfall over the Yangtze River valley and a dryer one on the southeast coast.

The hypothesis of Eurasian snow/Indian monsoon empirical relations was tested in a few AGCM sensitivity studies (e.g., Barnett, Dumenil, Schlese, Roeckner, & Latif, 1989; Vernekar, Zhou, & Shukla, 1995; Yasunari, Kitoh, & Tokioka, 1991; Yeh, Wetherald, & Manabe, 1983). Recently regional climate models have also been introduced to study the relationship between TP snow cover and the East Asian summer monsoon (Kyunghee & Hong, 2009; Xiao & Duan, 2016). All experiments have demonstrated that anomalous snow cover over Eurasia or the TP had impacts on the Indian or East Asian monsoon, but the impacts were highly variable among these studies. Moreover, these were sensitivity studies with specified extreme snow covers (such as no snow cover or double snow cover in the experimental design) or forced specified snow cover in the entire simulation that would violate the surface water and energy balances and affect their impact on atmospheric circulations. Furthermore, although snow over the Himalayas was included as one of the predictors in regression equations to predict the interannual variability in the Indian monsoon (Walker & Bliss, 1932), during the 1960s, the Indian Meteorological Department removed snow accumulation from the list of predictors routinely used for the long-range forecasting of summer monsoon rainfall (Bamzai & Shukla, 1999). Although it was later included for a time period (Guhathakurta, Rajeevan, & Thapliyal, 1999), after 1999, it is again no longer used as a predictor of Indian monsoon by the Indian Meteorological Department (Rajeevan, Pai, Kumar, & Lal, 2007; Thapliyal, 2001).

Robock et al. (2003) have noted that although snow-albedo feedback is always operating, the anomalous snow cover impacts on temperature were not prolonged by soil moisture feedbacks. This is because the soil moisture memory is very short, usually less than 2 months, and cannot be used as a bridge to link the winter snow cover and the subsequent summer monsoon. There was no obvious relationship between soil moisture and the Indian monsoon in their study. Moreover, there is poor persistence of the snow cover anomaly from winter through spring (Bamzai & Shukla, 1999). These imply that the argument of the winter snow anomaly residing in the climate system through soil moisture and its subsequent impacts on the Asian summer monsoon needs to be investigated further. In most of these modeling studies, simple land-surface models and snow parameterizations were used. Inadequate snow parameterization may affect the timing of snow melting for a couple of months, a matter of substantial implication for summer soil moisture and precipitation (Sud & Mocko, 1999). More studies with better snow modeling are necessary to more comprehensively understand these snow/monsoon relationships.

Assessment of the Effects of Land Degradation on the TP Climate System

The TP is one of the major areas in China greatly affected by land degradation (Figure 11A), which is caused by both natural processes and anthropogenic activities.

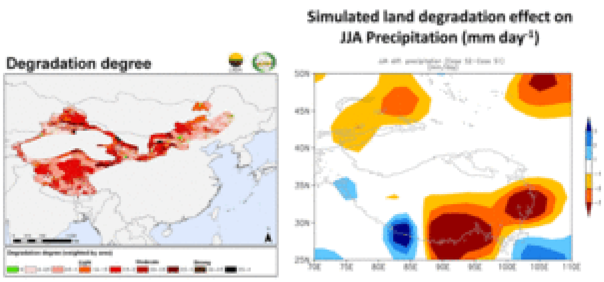

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Figure 11. (a) China land degradation map (adapted from the website http://www.fao.org/nr/lada/index.php?option=com:content&view=article&id=167&Itemid=200030&lang=en/). (b) Simulated June–August precipitation difference (mm day^{-1}) due to Tibet land degradation (Li & Xue, 2010).

The fragile TP environment, climate desiccation, and warming since the 1950s make the land vulnerable and contribute to different types of degradation (Zou et al., 2002; Cui & Graf, 2009; Li et al., 2010). Aeolian degradation, which is primarily driven by wind, is the primary form of land degradation on the TP

(Zhu & Liu, 1984; Wang, Wei, Zhao, Mengchun, & Zhao, 2004; Yang, 2006). Aeolian processes are very effective in areas with high available unconsolidated sediments and sparse vegetation conditions. Li, Zhang, Shen, Jia, and Li (2016) found that land degraded by aeolian degradation expanded rapidly from 1977 to 1990, then slowed down from 1990 to 2000. The anthropogenic cause of land degradation is related to overpopulation and excessive exploitation of natural resources in some local areas (Li, Yang, Gao, Chen, & Yao, 2004). For instance, the upper reaches of the Yellow River, which consists of extensive grasslands with intrinsic ecological value, has mostly suffered from overgrazing. Meanwhile, livestock numbers across the TP have also increased dramatically due to inappropriate land management practices (Du, Kawashima, Yonemura, Zhang, & Chen, 2004). In addition, the alpine grassland ecosystem has degraded significantly in association with global warming (IPCC, 1996). Recently available data for Phase 6 of the Coupled Model Intercomparison Project (CMIP6) LULCC experiment (Hurtt et al., 2011) show that 40–50% of the TP land suffered land degradation from 1948 to 1990.

Interaction between land cover change and regional climate is one of the major research subjects in land/atmosphere interaction studies (e.g., Fu, Yasunari, & Lütkemeier, 2004; Xue, 1996). For example, Xue et al. (2004) found that land degradation could cause delayed monsoon onset. However, the TP land degradation effect on the regional climate is just recently under investigation in a number of studies. The major challenge for conducting these types of studies is lack of reliable data of past land condition change as well as vegetation phenology and soil properties (Suh & Lee, 2004). With the development of remote sensing and retrieval methodology, a large number of satellite-derived land cover products have provided valuable information for climate models to assess TP land-degradation effects (Li & Xue, 2010; Cui, Graf, Langmann, Chen, & Huang, 2004).

In a GCM sensitivity experiment (Li & Xue, 2010), an AGCM was integrated using a “potential vegetation” map for the control simulation and a vegetation map with land degradation, in which grasslands in the central plateau and needle leaf evergreen trees in the southeast plateau were replaced by bare soil, for the degraded simulation. Different vegetation and soil properties, including surface albedo, leaf area index, vegetation root distribution, stomatal resistance, soil hydraulic conductivity, and surface roughness length, which are associated with the vegetation and soil types, were also changed according to the vegetation cover changes. This sensitivity study revealed that land cover change from vegetated land to bare ground over the TP increased surface albedo by 7%–9%, which results in decreased radiation absorbed by the surface and leads to weaker surface thermal effects and weaker vertical ascending motion. This and associated circulation change cause a decrease in the precipitation in the southeastern TP and weaker monsoon circulation (Figure 11B). A statistically significant relationship between observed precipitation in the southeastern TP and satellite-derived vegetation conditions has also been found (Zuo, Zhang, & Zhao, 2011), consistent with Li and Xue’s study (2010). Recent study (Yao et al., 2012) has identified a decreasing trend of precipitation in the southeastern TP, which seems consistent with Li and Xue’s study (2010). Further studies with proper land degradation settings based on observational data and models with high resolutions are warranted to realistically investigate land degradation effects on the regional climate system and surface hydrology.

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Notes:

(1.) The following are the latitude and longitude of some stations mentioned in the text: TTH: 34.22N, 92.43E; MS3637: 31.02N, 91.66E; WADD: 32.57N, 91.84E; NODA: 32.46N, 91.80E; MS3543: 31.58N, 91.90E; Zuri: 31.55N, 91.52E; BJ: 31.37N, 91.90E; ANNI: 31.25N, 92.17E; NaquFX: 31.36N, 91.54E; Naquhy: 31.42N, 91.98E; NaquRS: 31.42N, 91.98E.

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