Recent climate changes over the Tibetan Plateau and their impacts on energy and water cycle: A review

Kun Yang, Hui Wu, Jun Qin, Changgui Lin, Wenjun Tang, Yingying Chen

Abstract

The Tibetan Plateau (TP) exerts strong thermal forcing on the atmosphere over Asian monsoon region and supplies water resources to adjacent river basins. Recently, the Plateau experienced evident climate changes, which have changed atmospheric and hydrological cycles and thus reshaped the local environment. This study reviewed recent research progress in the climate changes and explored their impacts on the Plateau energy and water cycle, based on which a conceptual model to synthesize these changes was proposed and urgent issues to be explored were summarized.

1. Introduction

The Tibetan Plateau (TP) is an elevated region in the central Asia and stretches about 1000 km along latitude and 2500 km along longitude (Fig. 1). It is the highest plateau in the world, with an average elevation exceeding 4000 m ASL (above sea level) and an area of about $2.5 \times 10^6$ km$^2$. Its climate and environment are influenced by both westerlies and Asian monsoon, and, in turn, the Plateau modifies the climate of the adjacent and remote regions (Zhou et al., 2009). The TP is often called “Third Pole” and is subjected to strong interactions among the atmosphere, hydrosphere, cryosphere, and biosphere. In terms of regional energy cycles, the Plateau exerts a huge thermal forcing on the mid-troposphere over the middle-latitude of the Northern Hemisphere during spring and summer (Flohn, 1957; Ye and Gao, 1979). The thermal forcing effectively enhances the Asian summer monsoon and modulates its variability (Yanai et al., 1992; Wu et al., 2012). The surface heating also triggers vigorous deep convections over the TP (Yang et al., 2004) that greatly enhance the troposphere-stratosphere exchanges of water vapor and air pollutants (Fu et al., 2006). In addition, this region is the headwater areas of major rivers in Asia; particularly, it provides a large portion of water resources for economical activities in the oases of Northwest China (Yao et al., 2004).

Over the past three decades, the Plateau experienced evident climate changes (Kang et al., 2010), which have changed atmospheric and hydrological cycles and thus reshaped the local environment. For instance, river discharge and lake levels have responded to the climate changes (Cao et al., 2006; Ye et al., 2007). On the central Plateau, lakes expanded rapidly since the middle of the 1990s, which flooded surrounding grasslands and threatened the local economy and living. Glacier retreat due to warming was suggested to be the major cause of the expansion of glacier-fed lakes (Zhu et al., 2010). The warming also caused permafrost...
degradation (Cheng and Wu, 2007) and shortened the soil frozen period by approximately half a month per decade over the period of 1988–2007 (Li et al., 2012). Along with the climate changes, surface pressure over the Plateau increased significantly (Moore, 2012), and surface heating and atmospheric heating became weakened (Zhu et al., 2007; Duan and Wu, 2008; Yang et al., 2011a, 2011b). This warming and thermal weakening in spring and summer may affect summer precipitation downstream (Wang et al., 2008; Ding et al., 2009; Liu et al., 2012; Duan et al., 2013). In addition, Immerzeel et al. (2010) projected that the warming may lead to less water resources for the downstream regions in the future. Therefore, the TP climate changes have become the concerns of both the local and surrounding people, and the “Third Pole Environment” program was initiated to pool international efforts to understand climate and environment changes on the Plateau (Yao et al., 2012a).

In order to understand the processes of the regional water and energy cycle, several field campaigns have been conducted in the TP since the 1990s (Koike et al., 1999) and hydro-meteorological observations were further enhanced in recent years (Ma et al., 2008; Xu et al., 2008; Su et al., 2011; Yang et al., 2013). These field activities have advanced our understanding to the land–atmosphere interactions (e.g. Yamada and Uyeda, 2006; Ma et al., 2009; Zhang et al., 2012; Ueno et al., 2012), and supported the development of land surface models (e.g. Yang et al., 2009a; van der Velde et al., 2009; Chen et al., 2010; Gerken et al., 2012) and satellite remote sensing (e.g. Ma et al., 2011; Chen et al., 2013). These model improvements are crucial steps for understanding the response of water and energy budgets to the climate changes.

In this paper, we first reviewed recent findings in climate changes over the TP. We then focused on the processes and mechanisms how the water and energy budgets responded to the climate changes, following which a conceptual model is proposed to synthesize these climate changes and their impacts, in terms of the relationship between local warming and regional warming and in terms of the water and energy exchanges. Finally, relevant urgent issues to be clarified were recommended.

2. Observed climatic changes

The China Meteorological Administration (CMA) provided long-term station data for climate change studies in the TP region. These stations are sparsely distributed on this region and their operations started in different years and some stations were not regularly operated for some early years. In this review, we mainly addressed the climate changes over 1984–2006, because this period has not only accumulated more surface data for climate studies but also experienced outstanding climate changes over the TP. In addition, radiation budget data from satellites, as a complement to station data, became available over this period. Nevertheless, the decadal changes in wind speed and solar radiation are complex, so their investigations were extended to the 1960s and the results were compared with the case in Mainland China.

Fig. 2 shows an overview of the climate changes over the TP, including the observed linear trends in annual mean air temperature, air humidity, wind speed, and pan evaporation at the individual Plateau stations over 1984–2006. The air temperature and specific humidity increased significantly, with more warming over the North TP while more moistening over the South TP (Fig. 2a-b). The overall rapid warming since the 1980s was reported in many previous studies (e.g. Liu and Chen, 2000; Guo and Wang, 2012) and was also qualitatively produced with satellite data (Salama et al., 2012). The wind speed significantly declined at almost all stations over this period (Fig. 2c), which is consistent with the prevailing wind stilling occurring over Mainland China since the beginning of the 1970s (Xu et al., 2006; Jiang et al., 2010). Meanwhile, the pan evaporation dropped, as reported in previous studies (Zhang et al., 2007; Zhang et al., 2009), except at the Northeast TP. Details of these trends have been investigated by combining satellite data, reanalysis data, and model simulations in recent studies, and their major results are summarized below.

2.1. Spatial variability of surface temperature change

The environmental changes over the Plateau are mostly associated with the rapid surface warming. Understanding the elevation-dependence is crucial for the assessment of glacier/snow dynamics. According to station data, Liu and Chen (2000) showed that the warming was elevation-dependent. Qin et al. (2009) confirmed the differential warming rates between stations above 3000 m ASL (above sea level) and stations below 3000 m ASL and found that their difference was amplified since 2000 (0.1 K year$^{-1}$ in the region between 1000 and 3000 m and 0.21 K year$^{-1}$ in the region above 3000 m). However, little was known about the warming status for regions above 5000 m ASL,

Fig. 1. The terrain of the Tibetan Plateau and its surroundings. Major atmospheric circulation systems around the Plateau are shown, with yellow arrows for summer and blue arrows for winter. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
where most glaciers and snow surfaces are located but no CMA station is available. Also, the vast western Plateau has very few stations. Therefore, Qin et al. (2009) utilized satellite data to investigate the recent temperature change over the Plateau.

Qin et al. (2009) evaluated several MODIS monthly products of nighttime land surface temperature (LST); they found that version 4 of MODIS LST data is able to reproduce the station-observed trend, whereas the latest version does not. The observed warming at the individual CMA stations is not always comparable to the MODIS-derived one at the collocated pixel, but the latter approaches the former if averaged over multi-sites (~10 stations). This indicates that their differences in spatial representativeness may be effectively eliminated through the spatial average.

Based on this evaluation, version 4 of the MODIS LST product was applied to investigate the elevation-dependence of the warming over the Plateau. Fig. 3a shows the warming rate averaged over all pixels within each 200 m-elevation interval (the value given over each bar represents the number of MODIS pixels used for averaging the warming rate). The warming rate is clearly elevation-dependent: increased from 3000 m to 4800 m ASL, then became relatively stable between 5000 m and 6200 m ASL, and finally diminished near the highest elevations. Simultaneously, Liu et al. (2009b) projected the warming over the Tibetan Plateau with a general circulation model and found a very similar elevation-dependence pattern. The increasing pattern below 4800 m has been reported before (Liu and Chen, 2000). The decreasing trend above 6200 m ASL is a new finding but it is understandable. The land surfaces in this elevation range protrude into the middle-high troposphere and only occupy a small fraction of the space (as seen by the pixel number in Fig. 3a), so the surface warming over the very high elevations actually denotes the warming status of the mid-troposphere, which decreases with respect to the increase of elevation. However, the relatively stable warming rate between 5000 m and 6200 m ASL is unexpected in terms of the extrapolation of the station-derived elevation-dependence. Possibly, the presence of snow or a glacier that is typical above 5000 m ASL resists further warming through both reflecting solar radiation by the surface and exhausting energy by ice-melting. Another possibility is that this elevation range is a transitional zone between the increasing trend in the lower elevation and the decreasing trend in the higher elevation and thus the warming rate becomes stable.

The horizontal distribution of surface temperature change derived from the MODIS LST data was shown in Fig. 3b. It is clear that warming is a dominant phenomenon on the Plateau over recent years. The most warming areas are the southeastern TP and East Himalayas, a major glaciated area. Surprisingly, the west of 80°E, another major glaciated area, exhibited a strong cooling. In response, the glaciers in the southeastern TP and East Himalayas retreat the most rapidly (Yao et al., 2012b), whereas the glaciers in Karakorum and Western Kunlun Mountains were found fairly stable and even advancing (Scherler et al., 2011; Yao et al., 2012b).

In short, this study proves the feasibility to detect surface warming by remote sensing, although the time series used here is too short to detect the decadal change of the surface temperature. Along with the improvement of remote sensing technique and the extension of the observing period in the future, satellite observations can play an irreplaceable role in climate change studies.

2.2. Wind speed change

Since the beginning of the 1970s, wind speed over China declined (e.g., Xu et al., 2006; Jiang et al., 2010) and it declined more over the TP (Yang et al., 2011c). Several hypotheses have been proposed to explain the wind stilling over China, such as effects of urbanization and air pollution. However, these factors cannot play an important role in weakening wind speed over the Plateau, due to negligible urbanization and low-level aerosol load over there. Vautard et al. (2010) found that the wind stilling actually occurred widely over the Northern Hemisphere, and they concluded that the wind stilling can be largely explained by an increase in surface roughness over the past decades. This may explain the contrast trends between surface wind speed and upper-air wind speed occurring over Europe and North America, but it is not applicable to the case over the Plateau, where bare soils and short grasses are the dominant land types and the surface aerodynamic roughness lengths are generally small (mm–cm). Furthermore, the

Fig. 2. Trends in air temperature, specific humidity, wind speed and pan evaporation over 1984–2006 at the individual CMA stations on the Tibetan Plateau. The solid triangle symbol indicates a trend passes the t-test (p < 0.05) and its size indicates the magnitude of the trend. Panel (a–c): modified from Fig. 4 in Yang et al. (2011c).
major part of the Plateau is vegetation-free in wintertime, when wind speed decelerated, too. So, it is unlikely that the TP wind stilling was caused by an increase in the surface roughness.

Instead, Zhang et al. (2009) showed that in NCEP reanalysis the summertime pressure gradient force over the Plateau declined over the wind stilling period, and they attributed the wind decline to the adjustment of atmospheric circulation. Duan and Wu (2009) based on numerical modeling, further suggested that more warming over the high-latitude of Asia may be responsible for this weakening. To further understand characteristics of the wind speed change over this region, Lin et al. (2013) analyzed the longer-term variations of wind speed using both CMA data and IGRA (Integrated Global Radiosonde Archive; Durre et al., 2006) data, in which they not only provided direct evidence to support the finding by Duan and Wu (2009) but also revealed some new features, as explained below.

Since 1960, wind speed over China experienced three stages: jump of wind speed around 1970, a follow-on long-term decline until 2002, and a stable and even recovery trend afterwards. Fig. 4a shows the case for the TP. The jump of wind speed around 1970 is so outstanding that one may suppose that this was a spurious one. Lin et al. (2013) presented a cross-check to justify the jump. A wind speed and ground–air temperature gradient \( (T_g - T_a) \) together determine the sensible heat flux. If sensible heat flux dominates the surface energy budget, a high wind speed usually results in a low value of \( T_g - T_a \), and vice versa. This negative correlation between wind speed and \( T_g - T_a \) may be used to judge the reasonability of the wind jump. As in Fig. 4a, during the spring (MAM), a season with strong sensible heat flux, there is indeed a strong negative correlation between the wind speed and \( T_g - T_a \). Around 1970, the jump of the wind speed corresponds to a significant drop of \( T_g - T_a \); afterwards, the steady decrease in wind speed
is in contrast to the increase in $T_g - T_a$. Because wind speed, air temperature and ground temperature were measured independently, the opposite changes prove that the jump of the wind speed around 1970 and the follow-on wind stilling are very likely.

The jump and decline exhibit neither a diurnal cycle nor a seasonal cycle (not shown), indicating that the wind speed change was controlled by processes that have a scale beyond the atmospheric boundary layer. This can be further substantiated by the coherent changes between the surface wind speed and upper-air wind speed (see details in Lin et al.). In addition, another important finding is the elevation-dependence of the wind speed variability over China: the higher the elevation, the more the wind speed changed. In other words, the decadal variability of the surface wind speed over the Plateau is larger than the one over the rest of Mainland China. This fact implies that the signal of wind decline is transferred from the upper-air into the boundary layer. The wind speed over highlands may earlier respond to atmospheric adjustment than the one over lowlands; therefore, the recent recovery of wind speed over the TP might be a precursor of the overall change in atmospheric circulation over East Asia.

Lin et al. (2013) presented observed evidence that the wind speed change over China (including the Plateau) is due to the latitudinal gradient of surface warming over Central and East Asia. As shown in Fig. 4b, the difference in 500-hPa geopotential height between a low latitude belt (20°N to 25°N) and a high latitude belt (45°N to 50°N) in this region (70°E to 140°E) is highly correlated with the difference in surface temperature between the two belts, with correlation coefficient $R = 0.86$. As the anomaly of the temperature difference reflects the differential warming between the two latitude belts, the high correlation in Fig. 4b indicates that the latitudinal gradient of surface warming is an essential factor that first changed the upper-air pressure gradient force through lifting the geopotential height, then changed the upper-air wind through geostrophic adjustment, and finally changed the surface wind speed through downward momentum transport into the boundary layer. This substantiates the modeling results in Duan and Wu (2009). Specifically, during the period of wind stilling, the Siberian region got more warming than the south of China; during the recent period of wind speed recovery, the Siberian region received more snowfall (Ghatak et al., 2012) that may have suppressed its rapid warming.

### 2.3. Solar radiation change

Solar radiation is an important indicator of global climatic changes. A number of studies (e.g. Stanhill and Cohen, 2001; Liepert, 2002; Wild et al., 2005) suggested that there was a transition from global dimming to brightening around the end of the 1980s or beginning of the 1990s. Several radiation studies also stated that China had a similar transition (e.g., Che et al., 2005; Liang and Xia, 2005). Herein, we introduce a recent revisit to this issue.

A trend in solar radiation is so small and its estimate is sensitive to measurement errors, so quality control to station data is a prerequisite for the trend analysis (Shi et al., 2008). Tang et al. (2010) presented a method to control data quality of radiation measurement in China. After excluding spurious data and inaccurate measurements, there remain only ten radiation stations with long-term reliable records for the years from 1979 to 2006. Therefore, the radiation stations valid for detecting the radiation trend over China are very limited.

Alternatively, Tang et al. (2011) combined data at all CMA routine stations and radiation stations to revisit the radiation trend over the whole China, with a focus on the Plateau region. A widely-validated hybrid model was used to estimate solar radiation from sunshine duration data and other meteorological measurements at all CMA stations. The model was validated in terms of radiation trends against the observed
Fig. 5. (a) Variations of observed annual mean daily sunshine duration averaged over China and over the Plateau; (b) Variations of estimated annual mean solar radiation averaged over China and over the Plateau (modified from Fig. 6 in Tang et al., 2011). The star symbol (*) denotes a trend that passes the significance test ($p < 0.05$). The unit of the trend is $W \text{ m}^{-2} \text{ year}^{-1}$; (c) Linear trend slope of annual mean solar radiation at the individual Plateau stations over 1984–2006. The solid triangle symbol indicates a trend passes the $t$-test ($p < 0.05$) and its size indicates the magnitude of the trend.
ones from the ten quality-verified stations and the derived ones by an artificial neural network (ANN)-based model at all CMA radiation stations. Finally, the radiation data simulated by the hybrid model was used to extend the radiation trend analysis to all CMA stations and to a longer period.

Fig. 5a shows the changes of annual mean daily sunshine duration averaged over all CMA stations in both the TP and Mainland China. The sunshine duration averaged over the Mainland China decreased since the 1960s, but it became stable since 1990. The sunshine duration averaged over the Plateau is greater than the one over Mainland China, due to lower values of air mass, aerosol loads, and humidity over the Plateau. More importantly, the sunshine duration over the TP exhibits an increasing trend up to the end of the 1970s and a persistent decreasing trend afterwards. Clearly, the decadal change in sunshine duration over the TP is different from the one averaged over the whole China.

The variation of solar radiation exhibits a decadal change similar to the sunshine duration. As shown in Fig. 5b, the solar radiation averaged over all CMA stations has a decreasing trend before the 1990s, but it does not show an overall recovery trend since 1990. This result is somehow different from previous studies that stated solar radiation in China was recovering since 1990. In addition, the solar dimming rate before 1990 is about 2.5 W m\(^{-2}\) decade\(^{-1}\), which is about half of the values (4–5 W m\(^{-2}\) decade\(^{-1}\)) in previous reports. On the Plateau, the averaged solar radiation had an increase trend until the end of the 1970s and went down afterwards, indicating a transition from brightening to dimming around the end of the 1970s. Therefore, the solar radiation changes over the last 50 years are quite different between the TP and Mainland China. Fig. 5c shows the solar radiation changes at the individual Plateau stations for the period of 1984–2006. It is evident that the dimming occurred at most of the stations over this period. A further analysis shows that the dimming mainly occurred in summer.

Conventionally, a solar radiation trend is explained by aerosol loads and/or cloud changes. The Plateau region is one of the cleanest regions in the world and a considerable aerosol loading is not expected. Meanwhile, the total cloud cover decreased over the region, which is neither able to explain the solar dimming. To understand the mechanism of the solar dimming, Yang et al. (2012) analyzed the effects of aerosols, water vapor, and cloud cover. First, the inverse of visibility, a proxy of aerosol optical depth (AOD) (Wang et al., 2009), observed at CMA stations has a downward trend since 1980. The major aerosol over the TP is dust (Xia et al., 2011). Along with the wind stilling, dust amount and frequency should have decreased and thus AOD decreased, too. This is consistent with the observed trend in Fig. 6a. Therefore, it is unlikely that aerosol loading can explain the solar dimming. Second, an increase in water vapor amount is highly correlated with the solar dimming (Fig. 6b), but the direct absorption of solar radiation by the vapor (-1 W m\(^{-2}\)) is much lower than the observed dimming (-6 W m\(^{-2}\)). Third, deep cloud cover (DCC) is highly correlated with the solar radiation at both annual and decadal scales. DCC can be given by the International Satellite Cloud Climatology Project (ISCCP) data or approximately represented by the station-observed low cloud cover (LCC), and Fig. 6c shows that LCC has an increasing trend, opposite to the decreasing change in solar radiation. Such an opposite correlation can be seen in satellite observations, too. The increase in DCC is caused by the increase in convective available potential energy (CAPE), which is in turn due to the rapid warming and moistening over the TP. In summary, it is important for the radiation trend analysis to discriminate cloud types, in addition to the often mentioned cloud cover and aerosol changes.

3. Response of energy and water cycles

3.1. Land and atmospheric energy budget

Because of the importance of the Plateau thermal forcing in the context of the Asia monsoon, there have been a number of studies since the end of the 1970s to quantify the heat source over the TP and its effect on various weather and climate events. The total heat source (TH) comprises of three components, i.e., surface sensible heat (SH), latent heat release of condensation (LH), and radiative convergence (RC). Their estimation was first presented in Ye and Gao (1979) and followed by many studies (e.g. Chen et al., 1985; Zhao and Chen, 2000).

The surface sensible heat flux is usually calculated according to the bulk heat transfer equation, which needs input of observed wind speed and ground–air temperature difference (\(T_g - T_a\)) as well as parameterized heat transfer coefficient \(C_h\). Conventionally, \(C_h\) is either given a constant value (Chen et al., 1985) or simply correlated with meteorological variables (Chen and Wong, 1984). These parameterizations are widely applied in the Plateau heating studies. For instance, Duan and Wu (2008) used the Chen Scheme and found a weakening trend in sensible heat flux. However, near-surface atmospheric stability depends on wind speed and \(T_g - T_a\); with the wind stilling and increase in \(T_g - T_a\), the atmosphere became more unstable and thus \(C_h\) should have increased. So, a \(C_h\) scheme that does not properly account for the stability would cause vital errors in estimating the trend in sensible heat flux.

Accordingly, Yang et al. (2009b) developed a new scheme to estimate the heat flux from CMA station data. The new one disaggregates CMA station data to hourly to capture the diurnal variations of
atmospheric stability, and then, uses the Monin–Obukhov similarity theory to account for the stability effect. Yang et al. (2011a) applied this parameterization to the Plateau, in comparison with one conventional scheme given by Chen and Wong (1984) and another by Chen et al. (1985). As shown in Fig. 7, the estimated trend over the period of 1984–2006 is sensitive to the selected scheme. The Chen and Wong scheme assumes a linear relation between $f_D$ and the inverse of wind speed and thus $f_D$ increases too quickly with the wind stilling; as a result, it yields a positive trend in the sensible heat flux ($0.3 \text{ W m}^{-2} \text{ decade}^{-1}$). By contrast, the Chen scheme assumes a constant value of $C_D$ and thus ignores the effects of the atmospheric stability trend on $f_D$, so it gives a strong negative trend ($-2.5 \text{ W m}^{-2} \text{ decade}^{-1}$). The new scheme designed by Yang et al. (2009b) considers both diurnal variations of heat transfer and the atmospheric stability, and it yields a moderate decadal decreasing trend ($-1.2 \text{ W m}^{-2} \text{ decade}^{-1}$). Guo et al. (2011) used observed 6-hourly data and shows a trend similar to Yang et al. (2011a), after considering the diurnal variations of the land–atmosphere coupling intensity.

The trend in the sensible heat flux was also investigated with land surface modeling driven by CMA data (Yang et al., 2011c). The land surface model (LSM) was the Simple Biosphere Scheme (SiB2; Sellers et al., 1996), with modifications to accommodate the TP environment (Yang et al., 2009a). The simulation results were validated with surface temperature and a surface energy budget on the western Plateau. Again, Fig. 8 shows that the simulated sensible heat flux (SH) had a negative trend, which is comparable to the parameterized result by Yang et al. (2011a). Zhu et al. (2012) compared several datasets of sensible heat flux, including reanalysis products and LSM simulations, and they confirmed that the weakening trend in TP sensible heat flux occurred in most of the datasets. Therefore, all these studies reach a consensus on the weakening trend in the TP sensible heat flux over the warming period.

The radiative convergence (RC) is the difference between the net radiation at the top of atmosphere (TOA) and the one at the surface. Usually, this term has negative values, which thus represent a cooling effect. Based on Tibetan experimental data, Yang et al. (2011b) evaluated the surface net radiation from the LSM simulation and two satellite products (the Global Energy and Water Cycle Experiment–Surface Radiation Budget (GEWEX-SRB) and the ISCCP-Flux Data (ISCCP-FD)) and found that the modeled one has smaller errors. Thus, the surface net radiation from the LSM simulation and the TOA net radiation averaged on the two satellite products were used to estimate RC. Fig. 8 shows the trends in the estimated RC. The trends are negative for both seasonal mean and annual mean, indicating an enhanced cooling. The trend slope is about 6–8 $\text{ W m}^{-2} \text{ decade}^{-1}$ and much less than the one presented by Duan and Wu (2008), who calculated this term according to ISCCP-FD data only. A further investigation shows that the cooling trend was mainly caused by the increase in outgoing longwave radiation (OLR) (by 5–6 $\text{ W m}^{-2} \text{ decade}^{-1}$), and the weakening of the air-absorbed solar radiation played a secondary role (by 0–2 $\text{ W m}^{-2} \text{ decade}^{-1}$). The increase in OLR is due to both the warming and the decreasing in total cloud cover (particularly high cloud cover).

The latent heat release due to vapor condensation is proportional to surface precipitation and has a large inter-annual variability. We detected no significant temporal trend in the Plateau-averaged latent heat release.

In summary, the $LH$ trend is not significant, $SH$ declined slightly, and $RC$ declined much more. As a result, the total heat source ($SH + LH + RC$) has a significant negative trend ($-7 \text{ W m}^{-2} \text{ decade}^{-1}$) and is dominated by the $RC$ trend (Fig. 8). This trend slope is comparable to the one derived from the NCEP/NCAR reanalysis data by Ding et al. (2009).

3.2. Land surface water budget

This section introduces how climate changes influenced the surface hydrological cycle on the TP. As there are limited data of river discharge and evaporation, the surface water budget was usually investigated by land surface hydrological modeling. Yang et al. (2011c) validated the capability of the improved SiB2 in simulating the discharge trend for four large rivers sourced from the central and eastern Plateau. Zhou and Huang (2012) also showed that this model could reproduce the observed anomaly of the surface water budget in the upper Yellow River basin. Therefore, the simulated surface water budget presented in Yang et al. (2011c) was introduced below.

Fig. 9 shows that the surface water balance was changed over the recent decades. In general, evaporation has an overall increasing trend, which is consistent with a recent independent modeling study (Yin et al., 2013). Other water budget components exhibit a clear spatial pattern, as indicated by two curved regions. On the central Plateau, we can see increasing trends in both precipitation and evaporation. The trade-off of the two trends leads to an increasing trend in runoff, as observed by Liu et al. (2009a). On the South and the East Plateau, however, we can see decreasing trends in precipitation while increasing trends in evaporation; both contribute to lowering the runoff. This implies a reduction of water resources from the Plateau, if the enhancement of the glacier melts was excluded from the total discharge. The western Plateau has
a small amount of precipitation while strong solar radiation; therefore, the precipitation is almost evaporated and the runoff is negligible.

This spatial pattern of surface water budget corresponds to lake area changes on the TP. Since the middle of the 1990s, major lakes in the central Plateau have expanded, whereas many lakes along the marginal region of the South and East TP have shrunk (personal communication with Dr. Yanbin Lei). Another analysis (Lei et al., 2013) found that the increasing runoff simulated by the model could account for the major part of the lake expansion in the central Plateau, although enhanced glacier melt could be another major contributor to the expansion of some small lakes.

In addition to the change in precipitation amount, the change in precipitation spatial distribution may also significantly change the surface water budget, as is the case for the upper Yellow River basin on the TP. This basin experienced wet conditions in the 1980s and dry conditions in the 1990s. In response, the river discharge had similar change during the two periods. Since 2002, the precipitation amount increased drastically, but the discharge still kept a low level. This incoherent change is attributed to the change in the precipitation spatial distribution (Zhou and Huang, 2012). According to CMA observations, the increase in precipitation occurred in the arid area of this basin, whereas the humid area received less precipitation than normal. In the arid area where water availability controls evaporation, most of the precipitation turned into evaporation and the runoff did not increase much when precipitation increased. In the humid area where energy availability determines the evaporation, evaporation significantly increased due to the warming; the overlap of the increase in evaporation and the decrease in precipitation led to significant reduction in runoff. The increase in basin-averaged evaporation almost canceled out the one in precipitation, and thus the discharge did not increase evidently. Therefore, the change in precipitation spatial distribution changed the relationship between the precipitation, evaporation and runoff.

3.3. Response of heat and evaporation to climate change

The decrease in the pan evaporation on the Plateau is a prevailing phenomenon over the past decades (Fig. 2d). This may be attributed to both solar dimming and wind stilling. Zhang et al. (2009) showed that the solar dimming played a major role in the pan evaporation reduction on the South Plateau while the wind stilling was more important on the North Plateau.

The opposite trends between the simulated land evaporation (Fig. 9b) and the observed pan evaporation (Fig. 2d) may be attributed to a complementary effect (Bouchet, 1963; Brutsaert and Parlange, 1998). As shown in Fig. 10, the sum of the simulated actual evaporation (E) and the potential evaporation (Ep) is roughly equal to 2Ey (Ey is the evaporation in a wet environment in the absence of advection, i.e. equilibrium evaporation) in wet zones, which is consistent with Bouchet’s hypothesis. Nevertheless, the sum is lower than 2Ey in dry zones, indicating a deviation from Bouchet’s complementary theory. The deviation was attributed by Zhang et al. (2007) and Yu et al. (2009) to low vapor pressure deficit that may suppress the power of vapor transfer. Note that the decrease in pan evaporation implies that lake evaporation may have weakened, due to both solar dimming and wind stilling. This trend is another factor responsible for the lake expansion on the central Plateau, in addition to the contribution of increasing runoff and enhanced glacier melts. Nevertheless, further studies are needed to quantify the lake evaporation and its trend.
It is concerned with what caused the decrease in the sensible heat flux and the increase in the land evaporation (i.e., lower Bowen ratio) under the warming environment. Herein, we present a simple theory to explain the response of Bowen ratio to warming and wind stilling. According to the bulk transfer equation:

$$SH = \rho C_p C_H u (-T_g - T_a)$$  \hspace{1cm} (1)

$$IE = \frac{\rho C_p}{\gamma} \alpha(\theta) C_H u\left[ e_s(T_g) - rh \cdot e_s(T_a) \right]$$  \hspace{1cm} (2)$$

where $SH$ is the sensible heat flux, $IE$ is the latent heat flux, $\rho$ is the air density, $C_p$ is the specific heat capacity at a constant pressure, $\gamma$ is the psychometric constant, $C_H$ is the bulk coefficient for heat transfer, $T_g$ is the ground temperature, $T_a$ is the air temperature, $e_s$ is the vapor pressure at saturation, and $rh$ is the air relative humidity. $\alpha(\theta)$ is a function of soil moisture ($\theta$) and it has a value near zero for dry soils and approaches unity for wet soils.

The Bowen ratio is given by

$$B = \frac{SH}{IE} = \frac{\gamma}{\alpha(\theta)} \left[ \frac{(T_g - T_a)}{e_s(T_g) - rh \cdot e_s(T_a)} \right]$$  \hspace{1cm} (3)$$

As a first-order approximation, we assume that that $\alpha(\theta)$ is time-invariant at a decadal scale. Then, the relative change in Bowen ratio depends on the change in the ground temperature and the air temperature, as follows

$$\frac{\Delta B}{B} = \left[ \frac{\Delta T_g - \Delta T_a}{(T_g - T_a)} \right] \left[ \frac{\partial e_s}{\partial T_g} \frac{\Delta T_g - \Delta T_a}{e_s(T_g) - rh \cdot e_s(T_a)} \right] \left[ \frac{(T_g - T_a)}{(T_g - T_a)} + 1 \right]$$  \hspace{1cm} (4)$$

where $\Delta T_g$ is the increase in the ground temperature, $\Delta T_a$ is the increase in the air temperature, and $\Delta B$ is the change of the Bowen ratio after warming.

The climatology and trend of meteorological parameters over the Plateau are derived from CMA station data, which gives $T_g = 280.9$ K, $T_a = 276.5$ K, $rh = 56$ %, $\Delta T_g = 1.8$ °C and $\Delta T_a = 1.4$ °C. Given these conditions, it can be shown with Eq. (4) that the Bowen ratio decreases by 6% with respect to warming (indicated by the solid circle in Fig. 11), which indicates that a warming can result in a less sensible heat flux and more latent heat flux (or evaporation). This qualitatively explains the sensible heat flux weakening and evaporation enhancement. If in Eq. (4) we consider the contribution of soil moistening (Fig. 9d) to the lowering of the Bowen ratio, then the theory can explain most of the decrease in the LSM-simulated Bowen ratio.

The Bowen ratio response to the warming may be modulated by wind speed change. If the wind speed slows down, the efficiency of the energy exchange would decrease; accordingly, the ground would get warmer (i.e., $T_g$ increases), leading to the decline of the Bowen ratio. Over the period between 1970 and 2000, both the warming and wind stilling weakened the Bowen ratio, leading to the decrease in sensible heat flux. Since 2000, however, the warming and wind speeding up play contrasting roles in determining the trend in the Bowen ratio, leading to a stable sensible heat flux, as indicated by the parameterization result with the Yang scheme in Fig. 7.

In addition, other changes on the TP surface may have contributed to the decadal change of the surface energy partition. The first is the abrupt increase of snow cover in winter and spring since around 1977 (Ding et al., 2009). The snow cover cools the surface through albedo enhancement to reduce surface net radiation and through snow melting to exhaust surface net radiation, leading to lower sensible heat flux. The second is the increase in vegetation density (Zhong et al., 2010). This increase may result in more evaporation and less sensible heat flux and thus lower Bowen ratios, as shown in Zuo et al. (2011). The third one is the increase in soil moisture in the Plateau excluding the monsoon-impacted southern and eastern TP (Fig. 9d). This change also enhances evaporation and weakens sensible heat flux (Zhang and Zuo, 2011).

4. A conceptual model for the Plateau climatic changes

There are a number of studies on the climatic changes over the Plateau, but a general and consistent framework has not been established for the interpretation of all the changes in the Plateau climate system. Herein, we proposed a conceptual model to synthesize these changes and to link these changes with the global warming. This model is shown in Fig. 12.

Wind stilling is a response of the regional circulation to the changes in the latitudinal temperature gradient over the Central and East Asia. Under the background of the global warming, there was more warming over the high-latitude belt than over the low-latitude belt during the past three decades. As a result of the adjustment of geopotential height to the change in thermal contrast, the middle-level pressure gradient force between the low-latitude and the high-latitude declined (Fig. 4b), and therefore, the middle-troposphere wind speed became weakened (Section 2.2). Due to less momentum transport from free atmosphere down to the boundary layer, the surface wind speed declined, too. This situation over the TP and East Asia is different from the case over Europe and North America, where the surface wind slowed down while upper-air wind speeded up and Vautard et al. (2010) attributed their contrast changes to the increase in surface roughness lengths.

The wind stilling weakened the energy exchange between the Plateau and its surroundings. As the Plateau provides a huge thermal forcing to the middle-troposphere during the summer half-year, a weaker energy exchange would lead to more remaining energy to warm the local air over the Plateau. Indeed, we found a correlation coefficient of 0.6 between wind speed and air temperature during summers. So we speculate that the TP warming during the summer half-year is partially attributed to the wind speed decline. Other warming causes were also proposed in the literature. For instance, Chen et al. (2003) suggested with modeling studies that the TP warming is due to the CO$_2$ effects, but Rangwala et al. (2009) argued that the enrichment of surface water vapor enhanced the downward longwave radiation and is thus responsible for the prominent winter warming trend over the TP. Meanwhile, Zhang and Zhou (2009) found that the total ozone amount over the TP declined more than the one over other regions along the same latitudes and thus suggested that the ozone depletion is a possible reason for the more warming over the TP. Similarly, Guo and Wang (2012) suggested that the most significant warming occurring over the North TP may be related to radiative and dynamical heating that are results

![Fig. 11. Response of Bowen ratio to surface air warming (ΔT_a) according to Eq. (4), given T_g = 280.9 K, T_a = 276.5 K and ΔT_g = 1.28ΔT_a. The solid circle corresponds to the Plateau warming condition (ΔT_g = 1.4 °C, ΔT_a = 1.8 °C and rh = 56 %).](image-url)
of pronounced stratospheric ozone depletion. In a word, it is far from reaching a consensus to the understanding of the TP rapid warming and its seasonality, as the energy budget responsible for a regional warming is small while influenced by a lot of factors.

The warming and air moistening over the central Plateau upraised CAPE, which is favorable to triggering stronger clouds and thus causing the solar dimming (Section 2.3). The rapid planetary warming led to enhanced radiative cooling through outgoing longwave emission. Meanwhile, the warming and wind stilling lowered the Bowen ratio and thus reduced surface sensible heat flux. Both processes contributed to the thermal forcing weakening over the Plateau (Section 3.1), which in turn might contribute to the Asia monsoon weakening.

The warming enhanced land evaporation overall on the Plateau, but the solar dimming and wind stilling may result in weaker lake evaporation. The monsoon weakening may result in less rainfall over the monsoon regions (South and East Plateau), whereas the warmer and moister atmosphere on the central Plateau may trigger more convective precipitation and enhance water cycle. Accordingly, the climatic change may reduce runoff on the Plateau source areas of Asian large rivers but enhance runoff on the central Plateau (Section 3.2). The latter have contributed to the lake expansion on the central Plateau.

5. Recommendations for future studies

The TP surfaces experienced an overall rapid warming and moistening as well as wind stilling and solar dimming over the past three decades. These changes further influenced the water and energy cycle over the Plateau: weakened sensible heat flux and lake evaporation, enhanced radiative cooling and land evaporation, and changes precipitation amount and spatial distribution. All the changes may be deemed as different but inter-dependent aspects of a changing regional climate system. We proposed a conceptual model that can interpret major changes observed in the TP region. Meanwhile, some key processes in the conceptual model need to be strengthened in future studies, as current reanalysis datasets are not able to capture some crucial facts (Wang and Zeng, 2012). For example, the warming trend over the Plateau is not observed in NCEP reanalysis data (You et al., 2010), the wind decline is not observed in ERA-40 and ERA-interim data, the negative trend in the heat source is not seen in the JRA-25. A conceptual model is helpful for a consistent and integrated interpretation of observed climatic changes and for the prediction of the Plateau water and energy cycles from precursor signals of global and regional changes. To reach this goal, we recommend the following outstanding issues to be explored in future studies.

First, what determines the spatial pattern of the warming? The warming is the most striking phenomenon on the Plateau, but it has not been understood why the warming rate over the Plateau is higher than over the eastern part of China along the same latitude (Zhang and Zhou, 2009). Moreover, the spatial distribution of the warming rate within the Plateau is important for interpreting the spatial pattern of the glacier changes, but what determines the spatially distributed warming is little known.

Second, is a decadal climatic change occurring over the Plateau and East Asia? Since the beginning of the 1970s, wind speed declined very much over both the Plateau and East Asia. However, Lin et al. (2013) found that the wind speed started to recover over the Plateau since 2002. Simultaneously, the sensible heat flux on the Plateau stopped dropping. These signals are probably the precursor of a decadal climatic change over the Plateau and East Asia.

Third, what is the trend in the surface radiation budget? Solar dimming over the Plateau is evident; both downward and upward longwave radiations should have increased due to the warming. However, the trend in net radiation is far from understanding, as it is sensitive to the change in surface albedo and surface emissivity in addition to the dimming, warming and moistening. Future studies need to find a reliable way to quantify net radiation and its trend, in order to understand the warming. Considering that the nighttime warming rate is greater than the daytime one, we speculate that the trend in net radiation may have a diurnal variation, large in the nighttime and small in the daytime.
Last but not least, where is the source of water vapor over central Tibet? The lake expansion in this region indicates that the water vapor was transported from outside. In order to understand the hydrological cycle and its trend in this region, it is crucial to identify the source region and the transport processes of the water vapor.

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