INTERNATIONAL JOURNAL OF CLIMATOLOGY
Int. J. Climatol. **38** (Suppl.1): e1–e17 (2018)
Published online 15 January 2018 in Wiley Online Library
(wileyonlinelibrary.com) DOI: 10.1002/joc.5411

Review

Climatic and associated cryospheric, biospheric, and hydrological changes on the Tibetan Plateau: a review

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ABSTRACT: We review recent climate changes over the Tibetan Plateau (TP) and associated responses of cryospheric, biospheric, and hydrological variables. We focused on surface air temperature, precipitation, seasonal snow cover, mountain glaciers, permafrost, freshwater ice cover, lakes, streamflow, and biological system changes. TP is getting warmer and wetter, and air temperature has increased significantly, particularly since the 1980s. Most significant warming trends have occurred in the northern TP. Slight increases in precipitation have occurred over the entire TP with clear spatial variability. Intensification of surface air temperature is associated with variation in precipitation and decreases in snow cover depth, spatial extent, and persistence. Rising surface temperatures have caused recession of glaciers, permafrost thawing, and thickening of the active layers over the permafrost. Changing temperatures, precipitation, and other climate system components have also affected the TP biological system. In addition, elevation-dependent changes in air temperature, wind speed, and summer precipitation have occurred in the TP and its surroundings in the past three decades. Before projecting multifaceted interactions and process responses to future climate change, further quantitative analysis and understanding of the change mechanisms is required.

KEY WORDS climate change; Tibetan Plateau; cryosphere; biosphere; hydrosphere; streamflow

Received 27 June 2017; Revised 14 November 2017; Accepted 15 December 2017

1. Introduction

Tibetan Plateau (TP) is the source of all the major Asian rivers and serves as the main source of water for nearly 1.4 billion people. TP has an average elevation exceeding 4000 m and an area approximately 2.5×10^6 km^2. TP is covered by snow, glaciers, permafrost, and seasonally frozen ground. It exerts a profound climatic influence on adjacent and distant world regions due to the environmental and climatic influences of the Indian monsoon, East Asian monsoon, and westerlies on the TP (Immerzeel and Bierkens, 2010b; Duan *et al.*, 2012; Yang *et al.*, 2014). In contrast to the Arctic and Antarctic, TP experiences strong interactions among the atmosphere, cryosphere, hydrosphere, and biosphere. As a result, hydrologic and cryosphere changes on the TP caused by climate change directly affect the lives of billions of people.

Climate over the TP is categorized by low air temperatures, high daily temperature differences, low annual temperature difference, and strong solar radiation. A warming rate of 0.46 °C decade^{-1} was observed during 1984–2009. This is greater than the rate in Northern Hemisphere (0.38 °C decade^{-1}) and almost 1.5 times the rate of global warming (0.32 °C decade^{-1}) (Zhang *et al.*, 2013a). Recent studies (Liu *et al.*, 2009b; Qin *et al.*, 2006; Pepin, 2015; Guo *et al.*, 2016a, 2016b; Li *et al.*, 2017) have reported that the changing trends in air temperature, wind speed, and summer precipitation are elevation-dependent in the TP and its surroundings. Because of its high sensitivity to climate change, TP changes are considered as an indicator of global warming effects (Yao *et al.*, 2012b). Substantial glacial recession, snow melt, and permafrost degradation have occurred on the TP due to temperature increases which have led to modifications in the hydrological cycle and changes in water resources throughout the region (Jiao *et al.*, 2015).

The period of frozen soil has decreased approximately 15 day decade^{-1} as a result of permafrost degradation from 1988 to 2007 (Liu *et al.*, 2012). Long-term observations and reanalysis data have been used to evaluate spatio-temporal variations in temperature, precipitation, evapotranspiration (ET), and stream flow and to study TP

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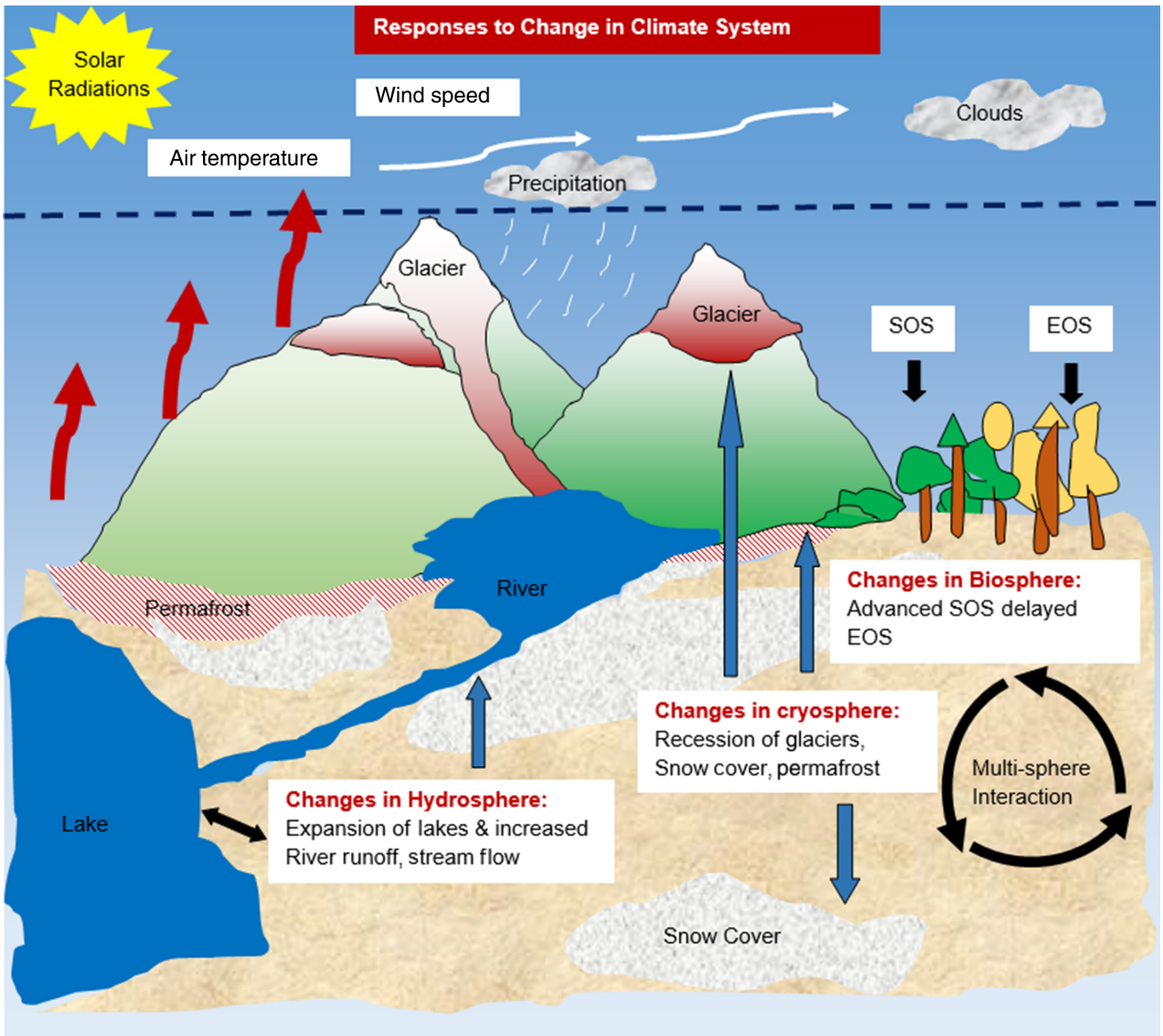

Figure 1. Responses of hydrosphere, cryosphere, and biosphere to changing climatic system on the TP. SOS means the start of vegetation growing season and EOS represents end of vegetation growing season.

energy and water cycles. Recent studies have focused on different climate variables (Yang *et al.*, 2014; Gao *et al.*, 2015). To develop a regional strategy for climate change and water resource management over TP, it is vital to study hydrological, cryospheric, and biological responses (Chen *et al.*, 2015). Many other reviews have discussed climate change and its impacts on various components of the TP climate system. Kang *et al.* (2010) reviewed the climate elements that were indicators for cryospheric changes on the TP. Cuo *et al.* (2014) reviewed the characteristics of surface hydrology and changes in streamflow hydrology over the TP. Yang *et al.* (2014) studied the impacts of climate change on the TP energy and water cycles. Shen *et al.* (2015b) studied the link between plant phenology and climate change and quantified the influence of phenological changes on TP ecosystems. Kuang and Jiao (2016) reviewed long-term trends in climate variables during last half century responsible for climate changes on the TP.

In contrast to previous reviews, we provide a comprehensive overview of climate change on TP and its associated hydrological, cryospheric, and biological changes in a systematic and integrated manner. This study evaluates current information and focuses on overall climate change characteristics and responses of the hydrosphere, biosphere, and cryosphere. Most intensive multi-spherical interactions take place on TP, impacting the life of about 20% of the world population. Figure 1 shows how changes in different climatic variables affect biosphere, hydrosphere, and cryosphere and how multi-spheric interactions take place.

2. Changes in the climatic indicators over the TP

2.1. Air temperature

Yearly mean temperature across TP is <0 °C and decreases from east to west (−10 °C in the western TP) (Sun *et al.*,

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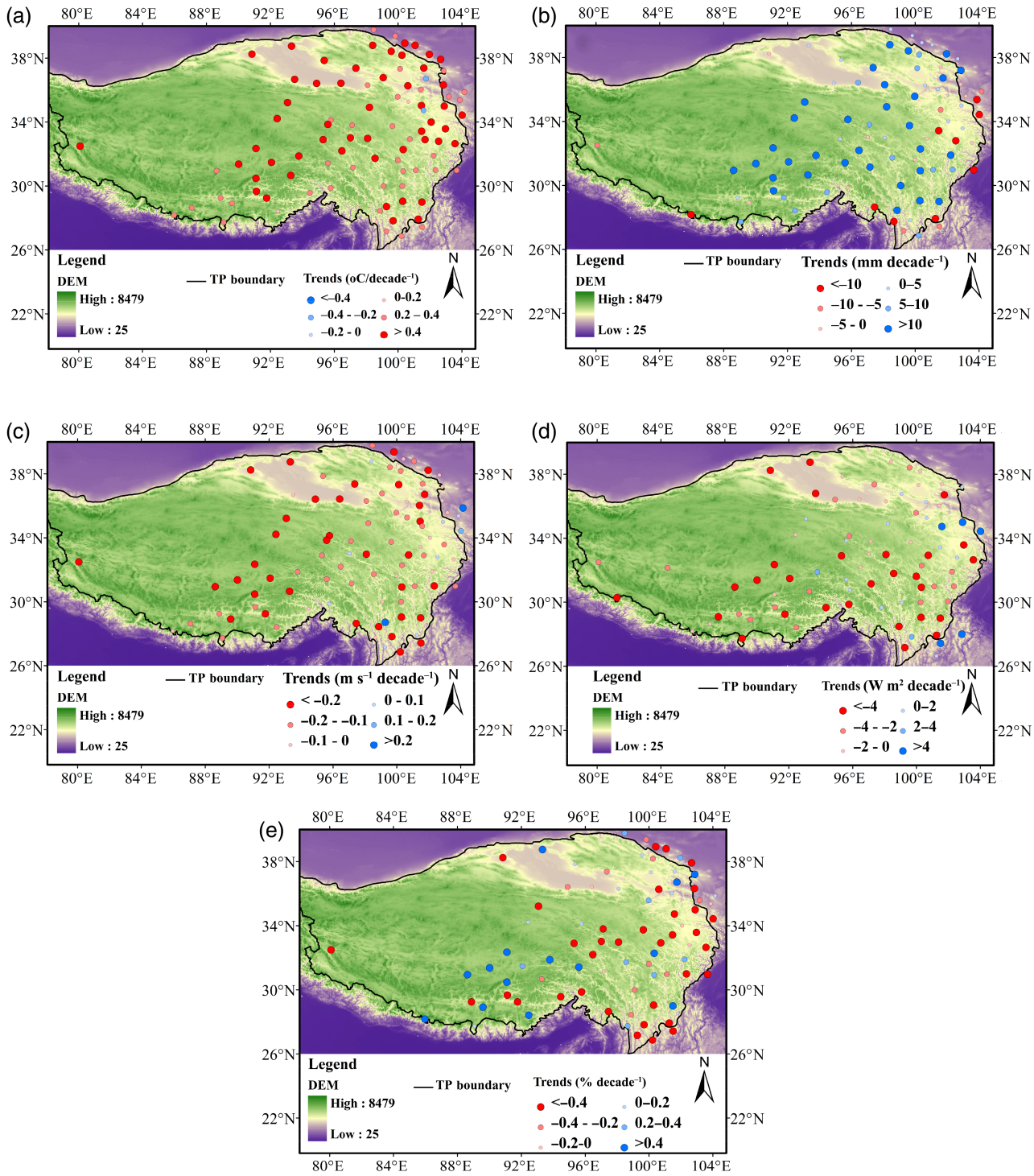

Figure 2. Trends in annual mean air temperature (1970–2012, °C decade^{-1}) (a), precipitation (1970–2012, mm decade^{-1}) (b), wind speed (1970–2012, m s^{-1} decade^{-1}) (c), solar radiation (1984–2006, W m^{-1} decade^{-1}) (modified from Yang *et al.*, 2011) (d), and relative humidity (1970–2012, % decade^{-1}) (e) over the TP.

2015). Surface air temperature from 1970 to 2012 has shown rapid warming (Figure 2(a)). After 1997, the warming trend accelerated (0.25 °C decade^{-1}) compared with an increase during 1980–1997 (0.21 °C decade^{-1}) (Duan and Xiao, 2015). The warming trend of the minimum temperature (0.41 °C decade^{-1}) is around twice the rate of the maximum temperature (0.18 °C decade^{-1}) (You *et al.*, 2016).

Most studies demonstrate that TP air temperature was high in the beginning of 1960s, from mid-1960s to the 1980s it decreased, and from the late 1980s onwards there is sustained increase in air temperature. Model projections have suggested more pronounced increases in minimum temperature than maximum temperatures, and greater winter warming than summer (Liu *et al.*, 2009b). Surface water vapour was identified as one of the main contributors

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to winter warming on the TP (Rangwala *et al.*, 2009). These will not only accelerate the melting of glaciers that affect the water supply to rivers but downstream precipitation might also be altered (Wang *et al.*, 2008). Nevertheless, the driving mechanisms for future temperature rise have not been revealed completely.

2.2. Precipitation

Similar to air temperature, precipitation is a vital climatic factor in the TP that impacts the cryospheric, biospheric, and hydrological regimes. For instance, the intensity, amount, duration, and phase of precipitation, alongside other characteristics such as its spatial, temporal, and elevational distribution and seasonality, control the development and formation of seasonal snowpack, glacier accretion, watershed storage conditions, soil moisture, and eventually runoff, as hydrological response. This is particularly important for the vulnerable region in the TP under the background of climate warming, which illustrates strong connections between the cryosphere, hydrosphere, and biosphere (Wu *et al.*, 2015; Dong *et al.*, 2016). TP climate is categorized by wet and humid summers with cool and dry winters. Gradual decrease in summer precipitation from southeast to northwest is approximately 700–50 mm. Since the 1960s, the TP has experienced a slight increase in precipitation (Xie *et al.*, 2010; Yang *et al.*, 2011; Gao *et al.*, 2014). This phenomenon is not as pronounced as temperature (Xie *et al.*, 2010). Variation in annual mean precipitation from 1970 to 2012 has been shown in Figure 2(b). The majority of the stations in semi-arid and humid-zones of central TP shows increasing precipitation trends, while at the periphery, decreasing trends have been observed. Summer precipitation increased with elevation (1000–4500 m) over TP from 1970 to 2014 at a rate of 0.83% decade^{-1} km^{-1} (Li *et al.*, 2017). The moisture transports from the west by westerlies and from the southwest by the Indian summer monsoon likely contributed the most to the precipitation over the TP (Zhang *et al.*, 2017a). However, complex topography and remoteness of most parts of the TP make it difficult to acquire ground-based meteorological observations.

The increase in atmospheric moisture over TP may be affected by a variety of mechanisms. Xu *et al.* (2008) proposed that water vapour transport increased over TP due to escalated monsoonal circulation and TP warming. Zhang *et al.* (2013a) suggested that melting of glaciers and permafrost due to a warming climate and slight growth in vapour transport from low altitude and Arabian Sea increased the moisture budget over the TP. Other possible mechanisms could involve the polewards shift of the westerly jet and an increase in summer monsoon circulations (Gao *et al.*, 2014).

2.3. Wind speed

Wind speed, as a function of horizontal pressure or temperature gradients, has direct or indirect effects on climatic variations (e.g. sensible heat flux and evaporation) (Guo *et al.*, 2016b). Wind speed has direct or indirect effects on climatic variations, e.g. density of sensible heat flux. Large-scale spatio-temporal variations in global temperature trends, especially the Elevation Dependent Warming (EDW), could be responsible for near-surface wind speed changes (Klink, 1999). The weakening of wind speed is a common phenomenon in China, particularly over TP (Lin *et al.*, 2013) (Figure 2(c)). The annual mean wind speed increases from southeast to northwest. During 1980–2005, differences in annual mean wind speeds were greater in the southeastern TP (1.5 m s^{-1}) than northwestern TP (3.5 m s^{-1}) and the annual entire TP mean value of 2.3 m s^{-1} was observed (You *et al.*, 2014). Generally, wind speed has decreased on the TP since the 1960s (Kuang and Jiao, 2016) or 1970s (Yang *et al.*, 2011) (Figure 2(c)). Guo *et al.* (2016b) proposed that the rate of wind speed reduction intensified with elevation with higher-elevation environments having greater changes in wind speed than lower-elevation areas. They suggested that EDW and increased surface roughness may have contributed to the elevation-dependent wind speed reduction in the TP.

2.4. Solar radiation

Worldwide declining trends in solar radiation have occurred at many locations since the 1960s, while pervasive decrease has been seen since the 1990s (Wild *et al.*, 2005). Over the TP, brightening was not observed (Figure 2(d)). The daily mean sunshine duration ranged from 6.8 to 7.4 h over TP (Liang *et al.*, 2013). Since the 1960s, a general decrease in sunshine duration was observed over TP (Yang *et al.*, 2014). From 1961 to 2005 decrease in seasonal mean sunshine durations of −5.7, −7.4, −3.4, and −3.1 h decade^{-1} was observed during spring, summer, autumn, and winter, respectively, with a maximum decrease in spring and summer (You *et al.*, 2010a, 2010b, 2010c). Since the 1980s, a significant trend of decreasing seasonal sunshine duration has been seen in summer and spring (Kuang and Jiao, 2016).

Changes in solar radiation on the TP have been influenced by anthropogenic aerosol emissions. Increased aerosol emissions have resulted in decreased solar radiation (You *et al.*, 2010a, 2010b, 2010c; Kuang and Jiao, 2016). During the past few decades, the summer monsoon has moved brown clouds from their origin (particularly the southern part of TP) to TP which may have also caused reduction in the amount of solar radiation (Ramanathan *et al.*, 2007; Zhang *et al.*, 2009). You *et al.*, 2010a, 2010b, 2010c found that solar radiation variation is associated with fluctuations of cloud cover over the TP. Yang *et al.* (2012) proposed that increased amounts of water vapour due to rapid warming and heavy cloud cover are key factors responsible for solar reduction on the TP. Further studies on the effects of decreased solar radiation on the TP are needed.

2.5. Relative humidity

During 1961–2013, mean annual relative humidity has decreased on TP at a rate of −0.23% decade^{-1} (Yin *et al.*, 2013). However, variability in annual relative humidity

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has been observed from the 1960s to 1990s, which has relatively decreased from the 1990s to the present (Kuang and Jiao, 2016).

From the 1960s to the early 21th century the northern TP showed an increasing trend in relative humidity (Zhang, 2007; Yin *et al.*, 2010). Linear trends of relative humidity during 1970–2012 showed large spatial heterogeneity (Figure 2(e)). Spatial distributional trends in relative humidity are complex, and both decreasing trends and increasing trends have occurred throughout the eastern and central part of TP (Kuang and Jiao, 2016). Over the entire region from 1961 to 2013, the seasonal mean relative humidity showed a decreasing trend (apart from winter). The trends were −0.11, −0.60, −0.39, and 0.04% decade^{-1} in spring, summer, autumn, and winter, respectively (Kuang and Jiao, 2016).

2.6. Causes of climate warming on the TP

The climate warming on the TP may be a result of the interaction of many different factors.

1. The recent climate warming on TP is primarily caused by the increased in anthropogenic greenhouse gas emissions (Guo and Wang, 2011; Duan *et al.*, 2012) due to change in local radiative forcing (Duan *et al.*, 2006). Big altitudinal difference and unique topography make the impacts of Green House Gases (GHGs) emissions on TP climate more devastating than any other region (Duan *et al.*, 2006, 2012).
2. Snow/ice-albedo and the cloud-radiation feedback are the main contributing factors to climate warming on TP (Rangwala *et al.*, 2010; Duan and Xiao, 2015), where the warming rate in high mountains can be more than in lower-elevation regions. Decreased snow cover due to increased absorbed solar radiation could be part the reasons for the substantial warming over TP.
3. Increased ozone concentration could be the possible reasons for decrease in total cloud cover over TP; however, it has been difficult to identify the mechanism (Zhang, 2007). On the other hand, the low-level cloud volumes display antagonistic trend, which decreases during daytime but increases at night (Duan and Wu, 2006). Increased night-time cloud cover could be one of the reasons for increased nocturnal warming in the TP.
4. Atmospheric aerosols can affect the atmospheric circulation over TP. Aerosols could increase lower atmospheric solar heating by nearly 50% (Ramanathan *et al.*, 2007; Kang *et al.*, 2010). Aerosols may be transported to the TP by atmospheric circulation (You *et al.*, 2010c) but it is uncertain how these influence the climate warming in the TP. Atmospheric aerosols could modify the regional climate, cryosphere, and hydrology by increased climate warming (Liu *et al.*, 2017). Aerosol particles impact the radiative forcing directly through absorption and reflection of solar and infrared radiation in the atmosphere. Aerosols appear to be more easily transported to the TP across the northern edge rather than the southern edge (Xu *et al.*, 2015).
5. Land use change may also account for climate warming over the TP. Most pronounced land use changes that not only impact the local climate but also have important hydrological implications on TP include grassland and permafrost degradation, deforestation, urbanization, and desertification (Cui and Graf, 2009). Anthropogenic land use change makes the climate warming on TP spatially intricate.

The mechanisms of climate warming on the TP are complex, so at present it is difficult to fully understand. Beside these, surface water vapour and changes in atmospheric circulation and atmospheric aerosol concentration are also considered as important factors that contributed to enhanced climate warming on the TP.

3. Responses of cryosphere to climate change over the TP

3.1. Snow cover

For Asian rivers originating at high altitudes, runoff from snow melt accounts for a greater fraction of the total runoff compared to glacier runoff in Yangtze, Yellow, Mekong, Brahmaputra, and Salween (Zhang *et al.*, 2013b). Precise mapping of the snow cover area is vital for agricultural production and mitigation of snow caused disasters such as floods (Liang *et al.*, 2008). Seasonal snow cover is a key environmental component on the TP and Himalayas regions. Because of cold climatic conditions and high elevation, precipitation usually occurs in solid form. In addition, as a result of low thermal conductivity and high albedo, the snow provides the water and energy interface between the land surface and atmosphere (Shrestha *et al.*, 2012). Snow cover in the southern TP and western edges is most persistent due to Indian monsoon spills onto the plateau (Pu *et al.*, 2007). The monsoonal precipitation (particularly inland) is predicted by spring snow cover on the TP (Immerzeel and Bierkens, 2010a). In contrast to extra tropical areas, snow cover thickness has increased in the TP from 1957 to 1992 (Kang *et al.*, 2010), which coincides with the recent increase in Antarctic and Greenland snow accumulation (Li, 1996).

With greater seasonal variation in spatial extent, snow cover over TP is a distinctive feature, which is a vital source of freshwater for western China and other Himalayan regions. General circulation and monsoon systems over eastern and southern Asia during spring and summer have a strong link with winter snow cover over the TP (Dickson, 1984; Yang, 1996). In regional climate studies, response of snow cover variability in TP to global warming has created an alarming situation (Qin *et al.*, 2006). Most persistent snow cover is situated in western and southeastern edges of TP and it persistency increases with increase in terrain elevation (Pu *et al.*, 2007).

An investigation carried out on snow cover dynamics of four typical lake basins in TP by using MODIS snow cover products (moderate-resolution imaging spectrometer) showed no evident trend of snow cover

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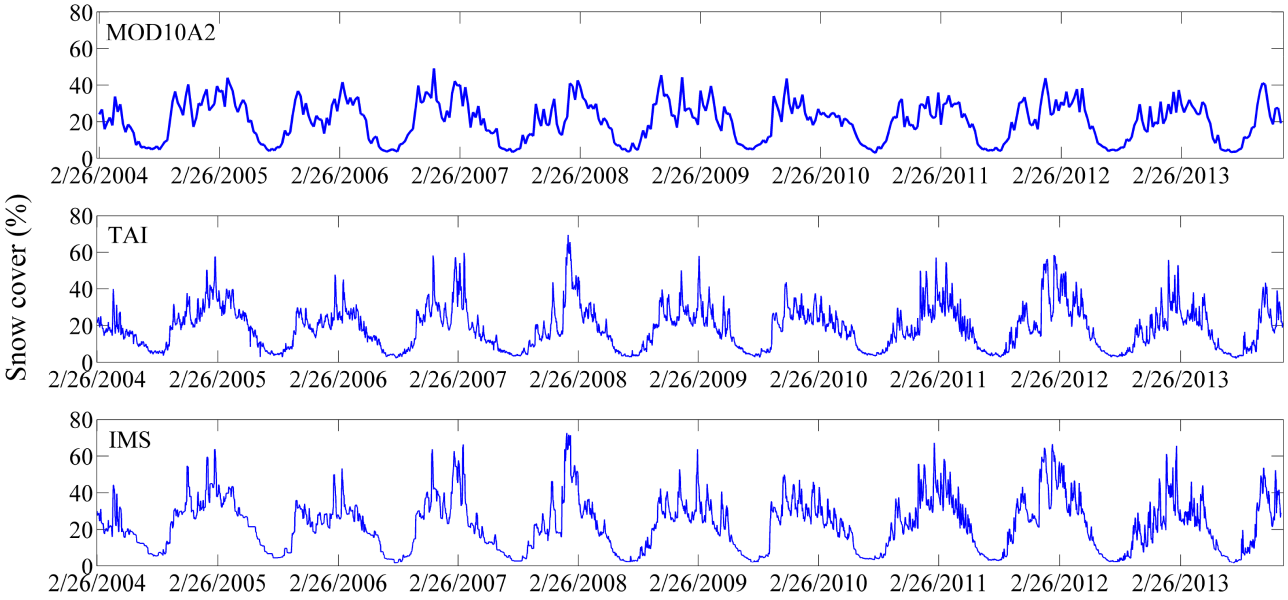

Figure 3. Time series of snow cover variations on the entire TP observed from MOD10A2, IMS, and TAI between 2004 and 2013 (modified from Yu *et al.*, 2016).

change during 2001–2010 (Zhang *et al.*, 2012). Because of biennial variations of the Indian monsoon, in western part of Nam Co Basin (central TP) monsoon and autumn snow cover indicated altering patterns. While snow-free areas and areas with permanent snow cover, snow cover trend was relatively stable (Kropacek *et al.*, 2010). From 1975 to 2000, increased snow depth over the TP was concurrent with a strong Indian–Burma trough, subtropical westerly jet intensification, and increased ascending air mass over the plateau during most of the year (Zhang *et al.*, 2004; Kang *et al.*, 2010). This phenomenon could influence the Asian summer monsoon due to strong snow and monsoon interactions (Ding *et al.*, 2014). As for other climatic factors, it is difficult to understand the duration of snowfall and seasonal snow cover changes on the TP due to limited availability of data on surface snow cover and depth (Li *et al.*, 2008). Yu *et al.* (2016) compared a time series of snow cover variations crosswise in the TP using MOD10A2, the Interactive Multi-sensor Snow and Ice Mapping System (IMS), and Terra–Aqua–IMS (TAI) between 2001 and 2013. Some MOD10A2 snow cover values (Figure 3) are >40%, while the daily TAI revealed more fine-grained snow cover variations, predominantly higher peaks of snow cover >50%.

3.2. Permafrost

Permafrost covers about 75% of the TP (Cheng and Jin, 2013). Permafrost is a chief component of the cryosphere and susceptible to climate change on different spatial and temporal scales (Cheng and Wu, 2007). In cold regions, thawing depth of frozen soil influences the runoff generation on hillslopes which substantially differs in temperate regions (Zhang *et al.*, 2013c). During the 1970–1990s, an increase in surface temperature over TP from 0.2 to 0.5 °C has caused permafrost reduction and permafrost loss is expected to continue in the future (Zhao *et al.*, 2004; Lemke *et al.*, 2007). Figure 4 shows the spatial distribution of trends in seasonally frozen ground during 1981–2010 for maximum freezing depths, freeze starting and ending dates, and the freeze duration at a depth of 1 m (Guo and Wang, 2013). Decreasing trends in maximum freezing depth have been observed in TP except for some southern regions. Similarly, freeze start, freeze end, and freeze duration have experienced delaying, advancing, and shortening trends except for some southern parts.

From the 1970s to 1990s, a preliminary assessment was made about of 100 000 km^2 areal reduction of permafrost in TP (Li *et al.*, 2008). Since the 1970s, permafrost areal extent has decreased about 36% along the Qinghai–Tibetan highway (from Amdo to Liangdaohe) (Lemke *et al.*, 2007). Based on long-term measurements of temperature, the permafrost altitudinal limit has increased 25 m in the northern TP during past 30 years, and increased 50–80 m in southern TP over the past 20 years (Cheng and Wu, 2007). Taliks (i.e. patches of unfrozen ground in an area of permafrost) areal extent stretched approximately 1.2 km alongside the highway. Areal extent of permafrost could shrink more rapidly over the next 50–100 years as predicted by a permafrost numerical model (Nan *et al.*, 2005).

Since the early 1980s, active layer thickness in TP (alongside the Qinghai–Tibetan highway) has increased about 1.0 m from its mean value (Zhao *et al.*, 2004), which is similar in magnitude (0.15–0.50 m) to other places over TP between 1996 and 2001 (Cheng and Wu, 2007). Model results indicate that the active layer thickness in TP has increased about 0.3 m in the north and 0.9 m in locally selected parts during 1980–2001 (Oelke and Zhang, 2007). From 1967 to 1997 thickness of seasonally frozen ground has been reduced by 0.05–0.22 m, and as a result of the earlier onset of spring thaw, duration of seasonally frozen ground has decreased by more than

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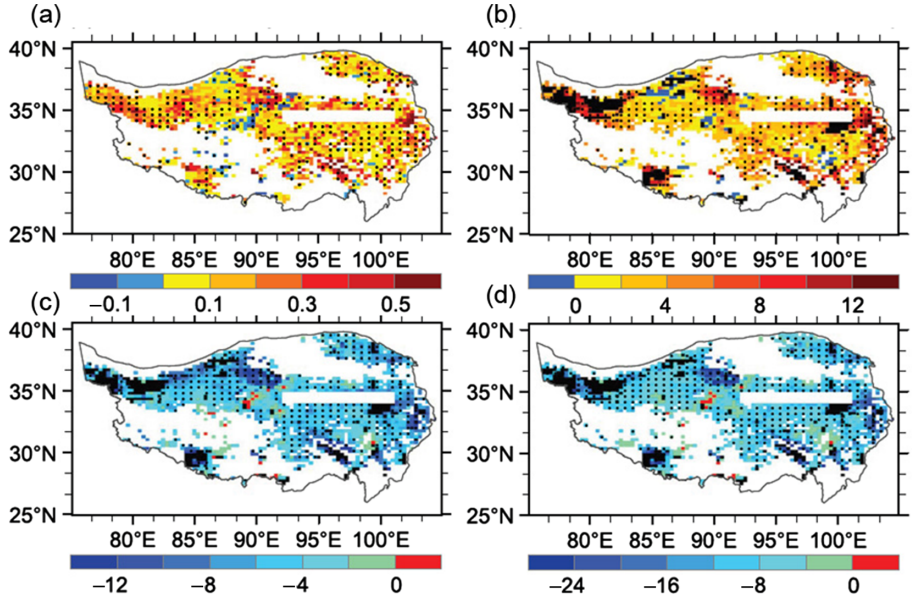

Figure 4. Trends in the simulated active layer thickness (m decade^{-1}) (a), date of freeze start at 1 m depth (day decade^{-1}) (b), date of freeze end at 1 m depth (day decade^{-1}) (c), and freeze duration at 1 m depth (day decade^{-1}) (d) for permafrost from 1981 to 2010 (modified from Guo and Wang, 2013).

20 days in the TP (Zhao *et al.*, 2004). Along with a climate change indicator, ground freeze–thaw cycles increased the heat-exchange intensity between the ground surface and the atmosphere, which expectedly increased the monsoon on TP (Chang *et al.*, 2007). Thawing can also result in earlier release of trapped carbon dioxide into the atmosphere (Zhang, 2007). Studies on thawing of seasonally frozen ground should be a priority in order to understand possible climatic feedbacks.

3.3. Glaciers

Recent changes in regional climatic conditions have had a negative impact on mass balance due to acceleration of glacier retreat (Kang *et al.*, 2007). By losing 4.5% of combined areal coverage, about 80% of glaciers have receded in western China (Ding *et al.*, 2006). This phenomenon is very evident in the Himalayas (Jin *et al.*, 2005; Bolch *et al.*, 2012; Immerzeel *et al.*, 2012, 2014), Qilian Mountains (Liu *et al.*, 2000; Liu *et al.*, 2002), and Tianshan Mountains (Liu *et al.*, 2006), with a shrinkage rate of between 5 and 10% during last 30 years. In contrast, glaciers in the interior TP are moderately stable (Liu *et al.*, 2000; Liu and Chen, 2000). Nevertheless, accelerated rate of glacier shrinkage has been reported in recent years (Shi, 2001; Li *et al.*, 2008). To present a comprehensive and systematic assessment of glacier status in and around TP, the whole region was divided into seven subregions shown clockwise as I–VII (Figure 5). The Himalaya shows maximum glacial reduction in both length and glacial areal extent (Figures 5(a) and (b)). Like glacial length and areal changes, the most negative mass balances also occur along the Himalayas ranging from −1100 to −760 mm year^{-1} with an average of −930 mm year^{-1} (Figure 5(c)) (Yao *et al.*, 2012a).

Since 1990, most TP glaciers have been rapidly shrinking due to global warming (Yao *et al.*, 2013). Satellite imagery and satellite altimetry have shown great variations in glacial mass and extent over the TP (Guo *et al.*, 2016a). From 2003 to 2009, the rate of glacial shrinkage over the Himalayas was measured at 10–30 mm year^{-1} (Gardner *et al.*, 2013), while the volume loss rate of 22 and 45 mm year^{-1} during 1999–2008 was recorded by a digital elevation model in eastern and western Himalayas, respectively (Gardner *et al.*, 2013). In Khumbu Himal, a 5% decrease in ice coverage was observed during 1962–2005, and a highest volume loss during 1992–2005 (Bolch *et al.*, 2008). In comparison to the mass loss over Himalayas, according to GRACE mission observation during 2003–2010, water mass over TP increased by 7 Gt year^{-1} as a result increase in water level of many lakes in the central TP (Jacob *et al.*, 2012; Song *et al.*, 2013; Zhang *et al.*, 2013a).

Precipitation change is a key factor contributing to the changes of glacial mass balance and this phenomenon is predominantly important in TP and surroundings. The greatest glacial shrinkage in the Himalayas corresponds to the declining precipitation rate accompanied by the weakening Indian monsoon (Molg *et al.*, 2012, 2014; Yao *et al.*, 2012a; Maussion *et al.*, 2014). Glacial retreat has resulted in hydrological changes, including greater river discharges, elevated lake levels, recurrent glacial lake upsurges, flooding, and increased debris flows from glacial, and water resources. These changes have been the focus of many studies (Lemke *et al.*, 2007; Yao *et al.*, 2007; Li *et al.*, 2008; Zhang *et al.*, 2013a; Cuo *et al.*, 2014; Yang *et al.*, 2014; Jiao *et al.*, 2015). In the 1990s, an increased river runoff (>5.5%) from TP caused by glacier retreat was observed, which is even greater in northwestern China (Yao *et al.*, 2007). Remote sensing data for many

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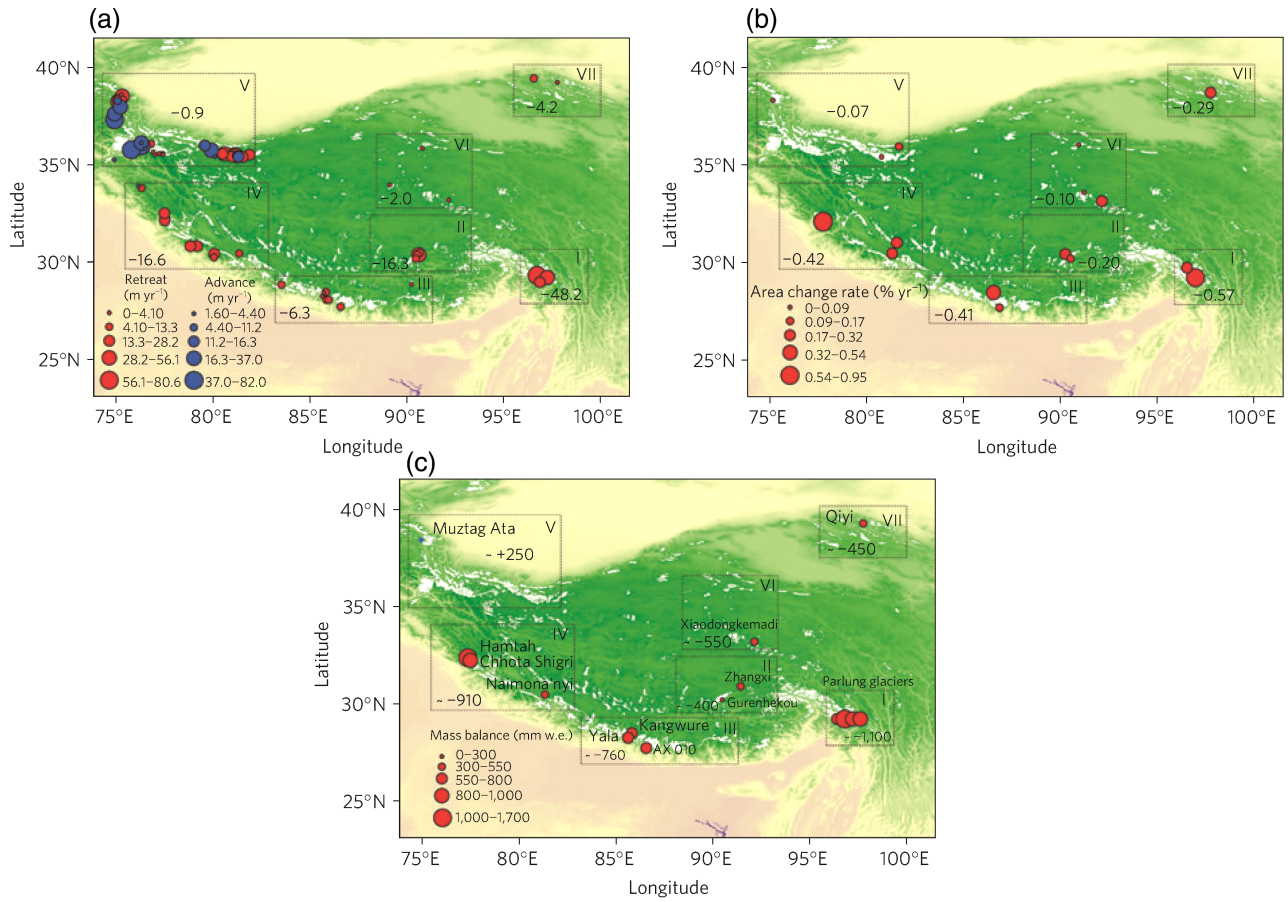

Figure 5. Spatial and temporal patterns of glacier status in the TP and surroundings for glacier retreat in 82 glaciers (a), area reduction in 7090 glaciers (b), and mass balance in 15 glaciers (c) (modified from Yao *et al.*, 2012a, 2012b).

lakes were used to quantify the contribution of glacier melt to rising lake level (Yao *et al.*, 2007). Zhang *et al.* (2017c) investigated that glacier mass loss contributes 13% and ground ice melt due to permafrost degradation contributed 12% in annual changes in lake area, level, and volume during 1970s–2015 in TP. Nevertheless, glacial volume changes and the synchronized environmental implications of glacial recession need additional study.

Despite the data limitations noted, observations have shown clear and systematic patterns of change in climatic regime and cryospheric response over TP. The various lines of evidence are consistent and mutually supportive. Warming has been pervasive, especially during winter and spring and at higher latitudes, while changes in precipitation have been more varied, both regionally and seasonally. Slight increase in precipitation has been observed, especially in the southeastern and southwestern parts of the region, and a decline in the fraction of precipitation falling as snow is very likely associated with rising winter/spring temperatures.

Cryosphere is sensitive to increasing temperature, particularly when precipitation occurs at temperatures near 0 °C. These changes are key drivers for extensive reductions in snow depth, duration, snow cover extent, and freshwater ice cover, predominantly in spring. This has led to an earlier occurrence of the spring freshet across the region. Warmer air temperatures are associated with rising permafrost temperatures, thawing and degradation of permafrost, and increasing glacier melt. Recent declines in winter and annual glacier mass balance at eastern and northeastern TP have been attributed to reduced snow accumulation (Deng *et al.*, 2017), but they may also be due to a shift in precipitation phase from snow to rain in some parts. Anomalously warm conditions led to reduced snow packs and early disappearance of snow cover, also leading to end of summer snow lines retreating off the tops of most glaciers and net negative glacier mass balance.

4. Responses of the biosphere to climate changes over the TP

Vegetation phenology is sensitive to climate changes. Changes in temperature and precipitation and other climate variables could alter the length of growing season. In cold regions, the start of the growth season (SOS) requires adequate temperatures (period preceding SOS that initiates vegetative growth) as a threshold to end ecodormancy (suspension of growth caused by factors outside the meristem but within the plant) during the spring season (Chuine *et al.*, 2003). Warming on TP resulted in a marked advance of SOS during 1982–1999 (Yu *et al.*, 2010; Piao *et al.*, 2011). Because of decreasing photoperiod (daylight)

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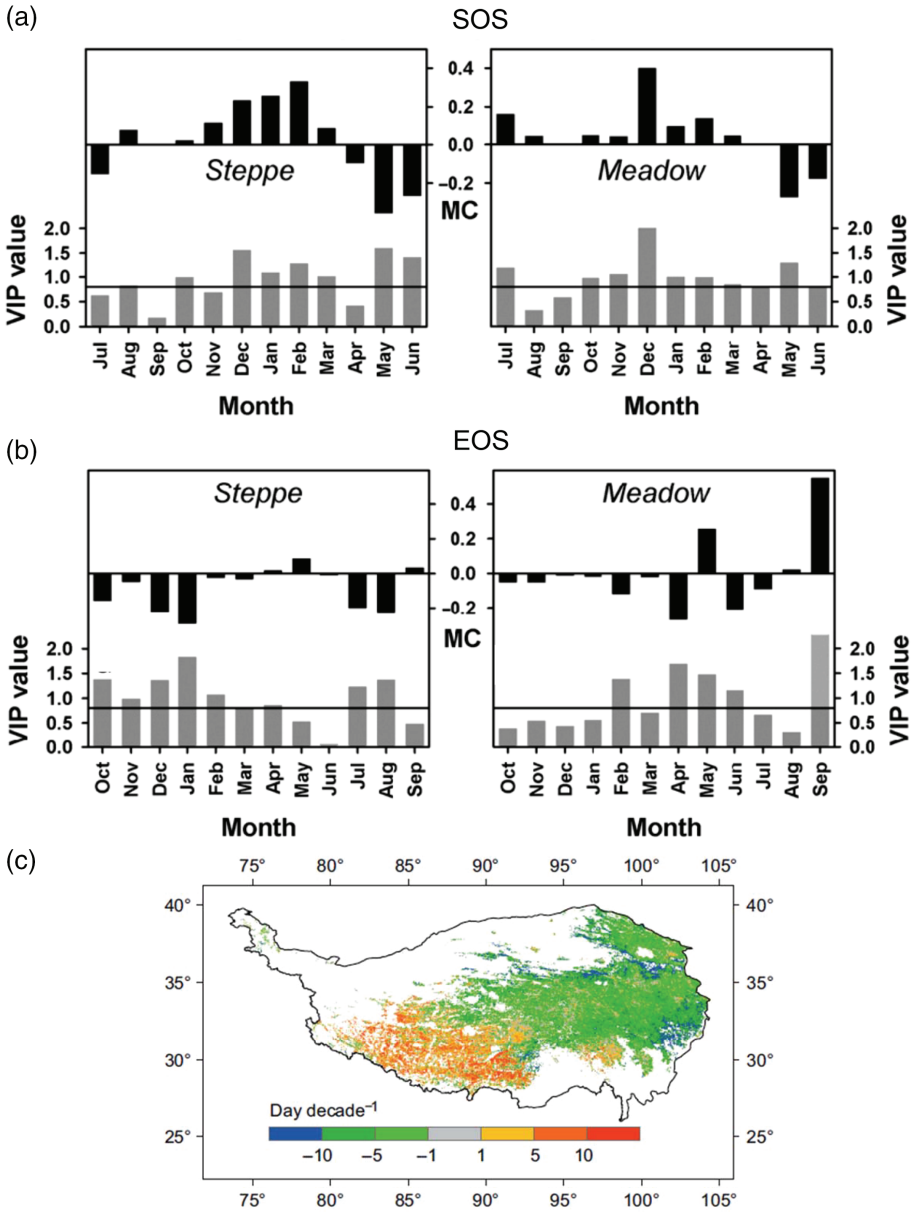

Figure 6. SOS (a) and EOS (b) in steppe and meadow vegetation of the TP in 1982–2006 (modified from Yu *et al.*, 2010) and spatial distribution of temporal trends (day decade^{-1}) of satellite-derived SOS over 1982–2011 (c) (modified from Shen *et al.*, 2015b). SOS is the start of vegetation growing season, and EOS is the end of vegetation growing season. VIP is variable importance plots and MC is model coefficients. [Correction added on 23 January 2018, after first online publication: This figure has been replaced and its caption has been updated in this current version.]

in the fall, endodormancy (non-growing season) follows ecodormancy and it breaks during dormant period typically at temperature lower than 0 °C during the dormant season (Chuine *et al.*, 2003). Hence, insufficient chilling due to climate warming in late autumn and winter could potentially result in delaying the start of the SOS (Shen *et al.*, 2015a). Delay of SOS in herbaceous plants due to climate warming in autumn and winter has not yet been observed because this temperature is adequate to meet chilling requirements (Shen *et al.*, 2014; Chen *et al.*, 2015). Thus, climate warming delayed the end of vegetation growing season (EOS) on TP due to partial positive inter-annual correlation between growing season temperature and EOS for the alpine steppe and meadow (Liu *et al.*, 2016). Figure 6 demonstrates the spatial distribution of satellite derived SOS temporal trends over TP during 1982–2011 (Shen *et al.*, 2015b) and response of the SOS (a–d) and EOS (e–h) in steppe and meadow vegetation to monthly temperature on TP during 1982–2006 (Yu *et al.*, 2010).

An increase in spring temperatures was observed in TP from 2000 to 2010, and significant decrease in SOS was seen in the western part of the plateau (Shen *et al.*, 2015b). Evidence of advancement in SOS due to pre-season precipitation has been found over the TP, and this phenomenon was more evident in arid parts of the region (Shen *et al.*, 2015a). During 1982–1999 extensive advancement in SOS was caused by the increase in pre-season precipitation

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and temperature, while a weakly delayed SOS in northwestern parts of TP might have been due to reduced precipitation (Shen *et al.*, 2015b). The EOS was also delayed on the TP by increases in precipitation during the growing season (Liu *et al.*, 2016). *In situ* observations revealed some positive effects of increasing temperature and precipitation on leaf phenology (Chen *et al.*, 2015). The results suggested that projected increase in temperature and precipitation on the TP would expand the growing season by advancing SOS and postponing EOS (Su *et al.*, 2013). There are several factors that could affect TP phenology. These are listed below:

1. Greater increase of warming at night on TP than during the day and the associated smaller diurnal temperature range (Piao *et al.*, 2015).
2. Intensity and timing of precipitation, freezing and thawing of soil, evaporation, and snowmelt, degradation of permafrost, increasing evaporation and soil water loss due to climate warming, and extended active layer may have reduced water availability to plants on TP due to the diverse soil properties (Hu *et al.*, 2009; Yang *et al.*, 2009; Zhao *et al.*, 2010; Peng *et al.*, 2013).
3. Interactions among different climatic factors also affect the phenology of herbaceous plants (Zhao *et al.*, 2010).
4. Photoperiod impacts on phenology over TP have not been adequately assessed (Vitasse and Basler, 2013; Laube *et al.*, 2014).
5. No obvious connection has been found between winter warming and spring phenology on TP (Shen *et al.*, 2015b).
6. Species composition can be altered either by climatic warming or by anthropogenic activities like overgrazing which ultimately affect the vegetation phenology (Harris, 2010; Niu *et al.*, 2010).

In mutualistic relationships, phenological changes also affect the tropic levels in contagious ecosystems with trophic mismatches in consumer resource dynamics and pollinator host mismatches (Kerby and Post, 2013). TP is a habitat of numerous Tibetan animal species (e.g. *Grusnigricollis*, *Ursusarctospruinosus*, and *Pantholopshodgsonii*), some of which are endangered, so preservation of TP vegetation needs particular attention (Shen *et al.*, 2015b) These changes could ultimately lead to changes in the entire climatic system (Richardson *et al.*, 2013). Both SOS and EOS timing substantially influence the annual carbon budget in the Northern Hemisphere which ultimately causes fluctuations in atmospheric CO_2 concentration (Keenan *et al.*, 2014). Some studies suggest that net carbon uptake will be increasing in spring due to warming driven advancement in leaf unfolding, whereas carbon uptake will either increase or decrease due to delayed leaf fall in autumn (Piao *et al.*, 2008; Keenan *et al.*, 2014). However, more studies are needed to understand the phenomenon of phenological changes on TP and how they affect regional carbon balance and energy cycles (Shen *et al.*, 2015b). Land surface biophysical parameters (sensible heat flux, albedo and trans-boundary layer turbulence) are directly modified by phenological changes which lead to changes in energy and water budgets and thus transform the local climate system (Jeong *et al.*, 2009a, 2009b).

Rapid climate change in the past half century has significantly modified vegetation phenology over TP. Lengthening of growing season is mainly caused by the advance in SOS, and delayed EOS. Plant phenology responds stronger to changes in precipitation patterns also; winter precipitation in the form of rainfall rather than snow requires less heat energy to melt and to warm soil and thus may advance the SOS. Changes in clouds cover and light intensity and other climatic factors also affect SOS and EOS. Some species may green up earlier, and others may senesce later, resulting in stretched growing season at both ends (early SOS and delayed EOS). Along with changes in climate variables, changes in permafrost can also have substantial effects on plateau vegetation phenology and it can also be used as an indirect indicator to monitor the change in permafrost. Further investigations are needed to understand biodiversity change in the alpine region under climate change scenario.

5. Responses of hydrosphere to climate change over TP

5.1. Evapotranspiration

Regional ET change patterns can also vary. Northwestern desert basin has lower ET compared to the southeastern forest cover basin (Kang *et al.*, 2010). Annual mean ET values are 359.7 and 306.6 mm in the upper Yellow and upper Yangtze River basins, respectively (Xue *et al.*, 2013). Linear trends of seasonal mean ET for the upper Yellow River basin, upper Yangtze River basin, Qiangtang Plateau, and Qaidam Basin during 1983–2006 are shown in Figure 7. ET for the four large basins showed increasing trends for all four seasons during past three decades consistent with regional warming (Li *et al.*, 2014). Overall, the inter-annual ET decreased from 2000 to 2010 over 42% of the region, mainly in the northwest of the TP. The dominant climatic factor that controls the long-standing variations of ET in the northwest TP was relative humidity. However, temperature drove ET under moist conditions (Song *et al.*, 2017).

5.2. Water storage changes

Because of its high elevation and glaciers (approximately 1.0×10^5 km^2, Yao *et al.*, 2012a), snow (approximately 41.9×10^9 m^3 year^{-1} water equivalent, Li *et al.*, 2008), and permafrost (approximately 1.5×10^6 km^2, Guo and Wang, 2013), TP is very important in global climatic and geodynamical studies (Guo *et al.*, 2016a). In the TP and the Himalayas, direct quantification of mass changes of glaciers, surface water, and groundwater is challenging due to data limitations (Bolch *et al.*, 2011; Kaab *et al.*, 2012). Studies conducted over the TP, Himalayas, and neighbouring areas indicate that water mass changes in the whole region are spatially heterogeneous (Guo *et al.*, 2016a). Water mass in the entire region, particularly in

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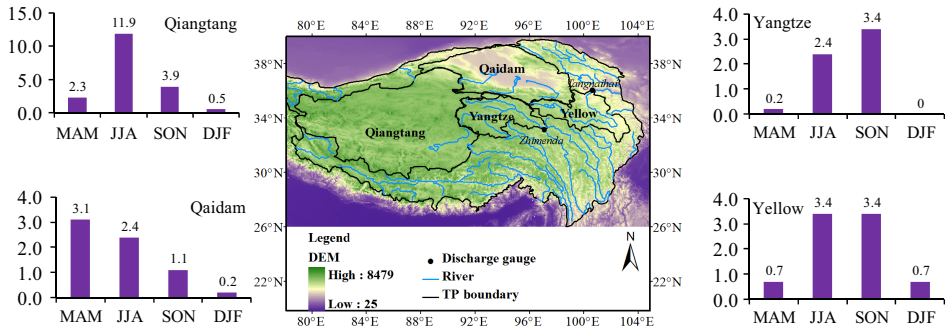

Figure 7. Linear trends of seasonal mean ET (mm decade^{-1}) for the upper Yellow River basin, upper Yangtze River basin, Qiangtang Plateau, and Qaidam Basin under the period of 1983–2006 (modified from Li *et al.*, 2014). The discharge gauges for the upper Yellow and Yangtze Rivers (Tangnaihai and Zhimenda) are given as solid points.

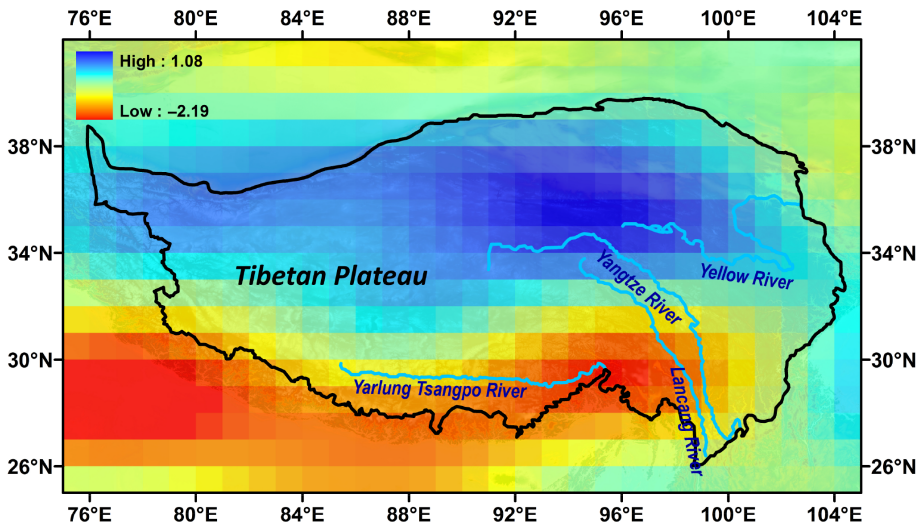

Figure 8. Spatial distribution of changing rates of total water storage (mm month^{-1}) in the TP from 2003 to 2012 (modified from Jiao *et al.*, 2015).

central and northern parts of the TP, increased between 2003 and 2012, while decreasing in the northern parts of India and Himalayas (Rodell *et al.*, 2009; Guo *et al.*, 2016a). Regularization solution Groupe de Recherches en GeodesieSpatiale (GRGS) investigated the mass changes rate (>10 mm year^{-1}) in the central TP and Qaidam Basin (Guo *et al.*, 2016a). Adjacent to the southern boundary of TP, obvious decrease in total water storage was observed during 2002–2012 (Figure 8). Melting of glacial ice and snow caused the decrease in total water storage along the Yarlung Tsangpo River (Jiao *et al.*, 2015).

Massive tectonic uplifting in the region could recompense the mass loss over the TP and ultimately it makes no substantial contribution to GRACE results (Jacob *et al.*, 2012). Satellite altimetry revealed gradual expansion of water levels of many lakes located in the central parts of the TP (Zhang *et al.*, 2011a; Song *et al.*, 2013). Satellite imagery also discovered many new glacial lakes in the plateau and the number of the new lakes tends to be increasing (Wang *et al.*, 2013). It has provided a positive indication about water mass storage in the form of lakes and increasing soil moisture in the TP (Zhang *et al.*, 2013a).

However, where did the lake water or the soil moisture originate? In previous sections, we noted that the entire plateau has experienced climatic warming since the 1980s which has altered the entire TP climatic system. Increases in surface air temperature, decreased solar radiation and humidity, increases in precipitation and melting of glaciers, and decreases in snow cover have all contributed to increased water mass over the TP.

5.3. Lakes

Anthropogenic activities have had minimal impacts on the alpine hydrological environment, so these regions are reasonable indicators of climate change. Understanding the response of alpine lakes to changing climate is crucial for accurate modelling of the hydrological and ecological process, and realistic prediction of seasonal and inter-annual water storage changes (Deniz and Yildiz, 2007; Medina *et al.*, 2008). The TP boasts the greatest concentration of high-altitude inland lakes in the world. With a total area of 41 800 km^2, it has been estimated that more than 1000 lakes distributed on TP are larger than 1 km^2 (Ma *et al.*, 2010; Witt *et al.*, 2016). Glacier melt has greatly supported the development of many inland TP lakes because these are fed by glaciers located in their

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catchments (Yao *et al.*, 2007; Zhu *et al.*, 2010; Günther *et al.*, 2016). Conversely, many other climatic factors such as increased precipitation and decreased evaporation from lake surfaces may have contributed significantly to the expansion of lakes which are not fed by glaciers in their catchment (Liu *et al.*, 2009a).

Rapidly rising temperature is the main driving force behind accelerated glacial retraction and thawing of permafrost which has been the chief source for inland lake expansion on the TP (Yao *et al.*, 2007; Zhu *et al.*, 2010). Since 1969, optical satellite imagery obtained from Landsat and other satellites has been used to study surface extent changes for a small number of lakes (Wang *et al.*, 2007). However, most studies have focused on quantitative detection and analysis of variation in either lake surface area or water level changes which do not provide sufficient information about response of lake water balance to changing climate (Wang *et al.*, 2007; Liu *et al.*, 2009a). Furthermore, due to inaccessibility and remoteness of the areas, the few observation stations are located on Lake Qinghai (Li *et al.*, 2005), Namco Lake (Zhang *et al.*, 2011a), Yamdrok Lake (Li and Wang, 2009), and Zabuye Salt Lake (Qi and Zheng, 2006) which can provide direct hydrological observations such as water level change data. The recent study on TP lakes exhibits an increasing trend of lake expansion from south to north and from west to east (Figure 9) (Lei *et al.*, 2014).

Complex climatic and hydrological environments for specific lakes exhibited variation in the different lake basins (Song *et al.*, 2013). For instance, Lake Qinghai located in northeastern TP, primarily supplemented by precipitation runoff, was shrinking until the 2000s despite of slight increase in precipitation; increase in evaporation rate in the basin was three times higher than precipitation (Zhang *et al.*, 2011b). Thus, increase in water storage cannot be simply explained by a large negative budget caused by 'precipitation–evaporation' in Lake Qinghai (Song *et al.*, 2013). Snow cover and glaciers melt have become chief sources of water for some of the lakes (Zhang *et al.*, 2011b). Similarly, levels of many lakes located in the upper Yellow River basin have shown greater variability in response to instabilities in climate features and water releases into the Yellow River (Song *et al.*, 2013).

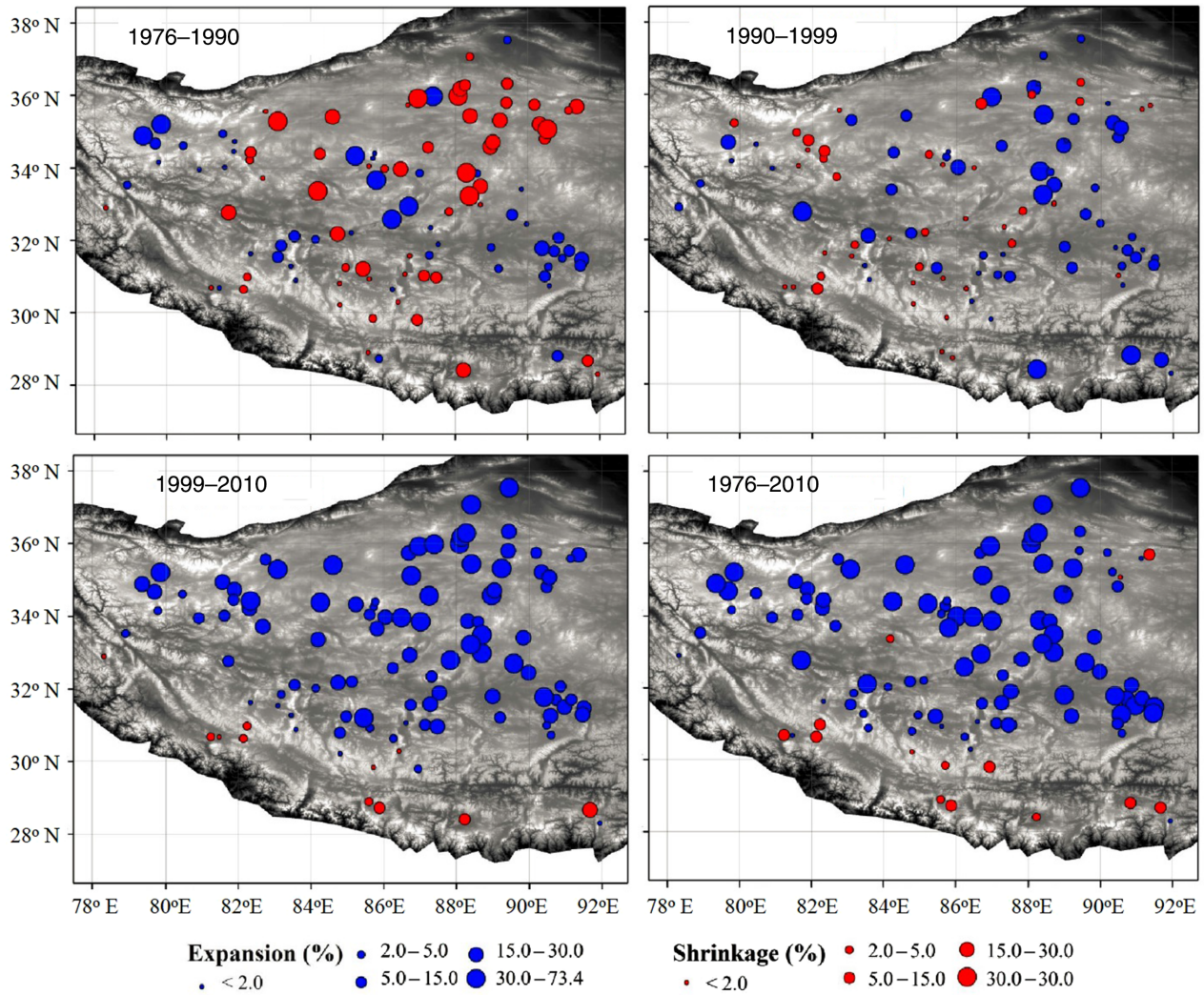

Figure 9. Changes in lake area on the interior of TP during the period of 1976–2010 (Lei *et al.*, 2014).

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Likewise, in the early 1970s and mid-1980s climate warming in the TP triggered glacier retraction, which resulted in increased water storage in these lakes (Bolch *et al.*, 2010), but water replenishment forces including rigorous evaporation and infrequent precipitation caused many lakes to shrink or even disappear (Song *et al.*, 2013). Distinct change in lake level during 1997/1998 is thought to be driven by large-scale atmospheric circulation changes in response to climate warming in the region (Zhang *et al.*, 2017b).

In addition to use of satellite images, numerical simulations from a basin-wide hydrological modelling can be effective to quantitatively investigate lake expansion/shrinkage over the TP in detail. This is especially true if *in situ* observations can be used to calibrate and validate the hydrological model. Through the process-based modelling strategy for a specific lake basin (including both water area and its surrounding land area), we are able to determine the key factors that contribute to the historical lake expansion/shrinkage (e.g. Jin *et al.*, 2015). Furthermore, this methodology is promising for prediction of lake changes, with help of the projections of future climatic variables (e.g. precipitation, air temperature, and wind speed).

5.4. Streamflow

Climatic variables such as temperature and precipitation regulate streamflow over the TP. During summer, precipitation is a dominant factor contributing to streamflow over the eastern and southeastern TP due to the strong influence of the South Asian monsoons (Yao *et al.*, 2013). However, spatial pattern understanding is required to quantify the long-term streamflow changes for all the basins on the TP. Streamflow peaks with increases in temperature and precipitation over TP (Yao *et al.*, 2012b). Meltwater is the key contributor to streamflow in the western part and melt increases with increased temperatures (Cuo *et al.*, 2014). In the westerly dominated central part of the TP, both meltwater and groundwater are significant contributors to streamflow during summer. The entire TP displays non-homogeneous spatial and temporal streamflow patterns (Cuo *et al.*, 2014). Variation in components and contributors occurs even for the same river system and streamflow changes from sub-basin to basin and from headwaters to downstream areas (Figure 10). Other factors include prevalent climatic system, environmental and watershed setting, and anthropogenic activities.

Temperature also strongly influences streamflow, and increased temperature positively affects the streamflow if the basin is extensively covered with glaciers. Streamflow response to temperature changes also depends on the forms and spatial distributions of precipitation. In the western TP that is controlled by westerly influences, annual precipitation increases in the range of 20–700 mm from the lowlands to the mountains (Mao *et al.*, 2006; Gao *et al.*, 2010). In some areas, low precipitation generates insufficient streamflow, while in mountainous regions precipitation primarily occurs as snow and ice and contributes to streamflow as meltwater.

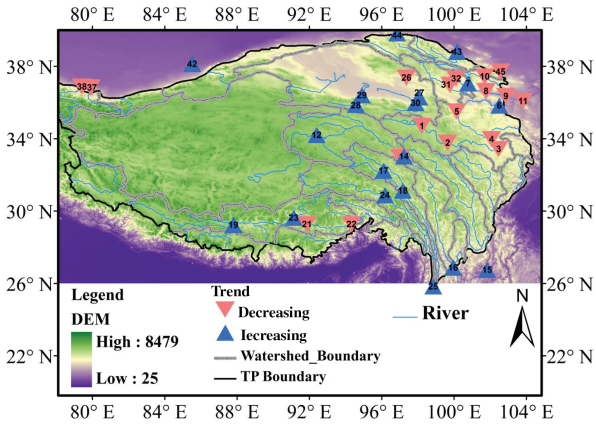

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Figure 10. Trends of streamflow in the major river basins on the TP (modified from Cuo *et al.*, 2014).

Similar to precipitation and temperature, another important factor that significantly affects the streamflow is actual ET. Studies about actual ET on TP are predominantly based on water balance and potential ET adjusted by available moisture content (Zhang, 2007; Cuo *et al.*, 2013b). Long-term observational actual ET on TP are not available but limited data suggest a general increase in actual ET along with spatial variation which would result in less availability of water for streamflow (Yang *et al.*, 2007; Zhang, 2007; Kang *et al.*, 2010). Larger effects of actual ET on streamflow occur during May–October compared to other months and it is the second most important factor that causes seasonal and annual variation in streamflow besides precipitation (Cuo *et al.*, 2013a).

Large-scale atmospheric systems such as the mid-latitude westerlies, East Asia and Indian monsoons, North Atlantic Oscillation, Arctic Oscillation, El Niño-Southern Oscillation (ENSO), and local circulations all play roles in affecting the weather and climate of TP (Yao *et al.*, 2012b; Cuo *et al.*, 2013b; Gao *et al.*, 2014; Lu *et al.*, 2015). Relating climate system indices and streamflow could possibly reveal the influences of the climate systems on streamflow. This would be helpful for understanding the spatial and temporal changes of streamflow over the TP.

Streamflow is affected by all cryospheric components (Kang *et al.*, 2010). On TP, relationship between seasonal distribution of streamflow and frozen soil is positive (Niu *et al.*, 2011). The degree-day modelling approach has been used to study the contribution of glaciers to seasonal streamflow in various TP basins (Su *et al.*, 2016), but the results of these studies evaluating the quantitative contribution of glaciers are inconsistent (Liu *et al.*, 2009a; Immerzeel *et al.*, 2010; Zhang *et al.*, 2013a). Numerous climate factors affect the future water availability in Asian river basins (Immerzeel and Bierkens, 2012; Lutz *et al.*, 2014). In general, glacier contributions are important

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primarily for headwaters or basins with large glacier coverage. The relative importance of snow to stream flow in this region is also controversial. Climate warming affected arid regions more severely than humid part of TP. For the assessment of climate change in the Third Pole and surrounding regions, it is particularly important to quantify the response of hydrosphere as water balance largely depends on climate regimes.

6. Conclusions

The TP has experienced rapid warming and increased precipitation over the last three decades, while solar radiation and wind speed have generally declined. In particular, dry areas have had stronger climate fluctuations than wet regions. The observations described here provided extensive and long-term data for examining climate system change and variability on the TP.

Increase in temperature drives changes in the cryosphere, biosphere, and hydrological cycles of the TP, resulting in seasonal streamflow variation. These changes will lead to changes in biodiversity and water shortages in regions with inadequate water storage capability. Spatial distribution of precipitation trends in the TP is more heterogeneous than temperature and there is significant inter-annual variability. Since the last decade of the 1900s, the rate of glacier retreat has accelerated, so more consideration should be given to subsequent hydrological processes, including greater discharges which can result in glacial lake outburst floods, rising lake levels, increased debris flows, and impacts on water resources in the TP.

Pervasive warming during winter and spring and both temporal and spatial variability in precipitation patterns have been observed over the TP. Decreased winter precipitation and the proportion of precipitation falling as snow may be associated with rising temperatures during winter and spring. These changes have resulted in an extensive reduction in snow cover, snow depth, and duration. Warmer air is the driving force behind rising permafrost temperatures which leads to permafrost thawing and increased glacier melting. Greatest reduction in glacial length and area and the most negative mass balance on the TP have been attributed to reduced snow accumulation (Yao *et al.*, 2012a). Because of the few meteorological stations in the western TP, limited solid precipitation data are available for this region. The responses of permafrost thawing and the reduction of seasonally frozen ground in response to changing climate on TP and its hydrological response across watersheds and gradients in different physiographic and thermal environments need to be studied. The influence of declining snowfall and duration on all facets of the hydrological cycle and streamflow will also be useful study topics.

Plant phenology has a broad influence on the structure and function of ecosystems but no model has been applied to study phenological responses on the TP. Because of our limited understanding of biosphere responses to climate change, additional modelling should be carried out to address phenological dynamics and responses to a changing climate such as future water availability. Ongoing research should continue evaluating the interactions between climate, hydrology, ecology, cryosphere, and other components of TP environments.

Acknowledgements

This work was funded by the "Strategic Priority Research Program" of the Chinese Academy of Sciences (XDA19070301) and the National Natural Science Foundation of China (Grant 91747201, 41571033). It was also supported by the International Partnership Program of the Chinese Academy of Sciences (131C11KYSB20160061) and the Top-Notch Young Talents Program of China (Dr. Lei Wang). The authors appreciate Dr. Xiaotao Zhang for providing figure 8.

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