Earth-Science Reviews 190 (2019) 33–57

Contents lists available at ScienceDirect

Earth-Science Reviews

journal homepage: www.elsevier.com/locate/earscirev

Climate background, fact and hydrological effect of multiphase water transformation in cold regions of the Western China: A review

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ARTICLE INFO

Keywords:
Multiphase water transformation
Climate warming
Hydrologic effect
Cold regions
Western China

ABSTRACT

The cold regions of western China are referred to as the “Asian Water Tower” and mainly include the Tibetan plateau and surrounding mountains. Its prominent hydrological feature is multiphase water transformation, which accelerates the water cycle and affects spatial and temporal patterns of water resources. Under the effect of lengthening ablation periods and increased annual precipitation, multiphase water transformation is accelerating. There are three main manifestations characterizing the transformation from solid to liquid water in the period since 1990: (i) the melting of glaciers has accelerated; (ii) the depth of permafrost active layers is increasing and their maximum freezing depth is decreasing; and (iii) a marked decrease in snowfall and increase in rainfall has been observed. The transformation from liquid to gaseous water was mainly concentrated on accelerating evapotranspiration. The transformation from gaseous to liquid water was observed as enhanced moisture recycling. The final hydrological effect of these transformations was observed in the change of the runoff components, increase in runoff, and lake expansion. A theory of multiphase water transformation is proposed, which is expected to contribute to the understanding of cold region hydrology in the future.

1. Introduction

Multiphase water transformation (MWT) refers to the frequent conversions of bodies of water between the solid, liquid, and gaseous states, which are the crucial links in the water cycle and have made great effect on spatial and temporal patterns of water resources. In this study, we consider the transformation from: solid to liquid water characterized by glacier ablation, permafrost degradation, and decreased snowfall; liquid to gaseous water by evapotranspiration; and gaseous to liquid water by moisture recycling. In cold regions, the prominent hydrological feature is the coexistence of multiphase water and its transformation. Fig. 1 shows a schematic diagram of the transformation process. Solid water reserves include glaciers, snow cover, and permafrost ground ice; liquid water mainly includes rivers, lakes, marshes, soil water, plant water, and groundwater; while gaseous water includes local evapotranspiration vapor and advection vapor. Under the effect of a warming climate, the MWT process is accelerating, which is affecting the hydrology, environment, water resources, ecology, and occurrence of natural disasters, resulting in subsequent social effects. Investigation of MWT will provide a new theoretical basis for explaining the balance of material and energy in cold regions, and for understanding the relationship between MWT and global warming. Quantifying the shrinkage of the cryosphere and the subsequent hydrological effects is expected to provide a new approach and direction for the development of hydrology in cold regions. In addition, the water balance and water resources are closely linked to changes in the MWT process, which play a crucial role in determining the frequency, intensity, and duration of water cycles in cold regions.

Based on data from the Intergovernmental Panel on Climate Change (IPCC, 2013), the global combined land and ocean temperature showed an increase of ~0.8 °C over the period 1901–2010 and about 0.5 °C over the period 1979–2010 when described by a linear trend. The mid−/high-latitudes and higher altitudes of the Northern Hemisphere showed an overall increase in precipitation during 1900–2010. Considering this, solid water is being transformed into liquid water at an increasing rate. It is likely that a decreasing number of snowfall events are occurring in most regions where increasing winter temperatures have been observed (Marty, 2008; Choi et al., 2010). In addition, almost all glaciers worldwide continue to shrink, as indicated by the measured changes in glacier length, area, volume, and mass over time. Glaciers have melted at an increasing rate over the last 35 years; the mean glacier loss was 221 mm/a in 1980–1989, 726 mm/a in 2000–2009,

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https://doi.org/10.1016/j.earscirev.2018.12.004
Received 11 July 2018; Received in revised form 26 November 2018; Accepted 6 December 2018
Available online 08 December 2018
0012-8252/

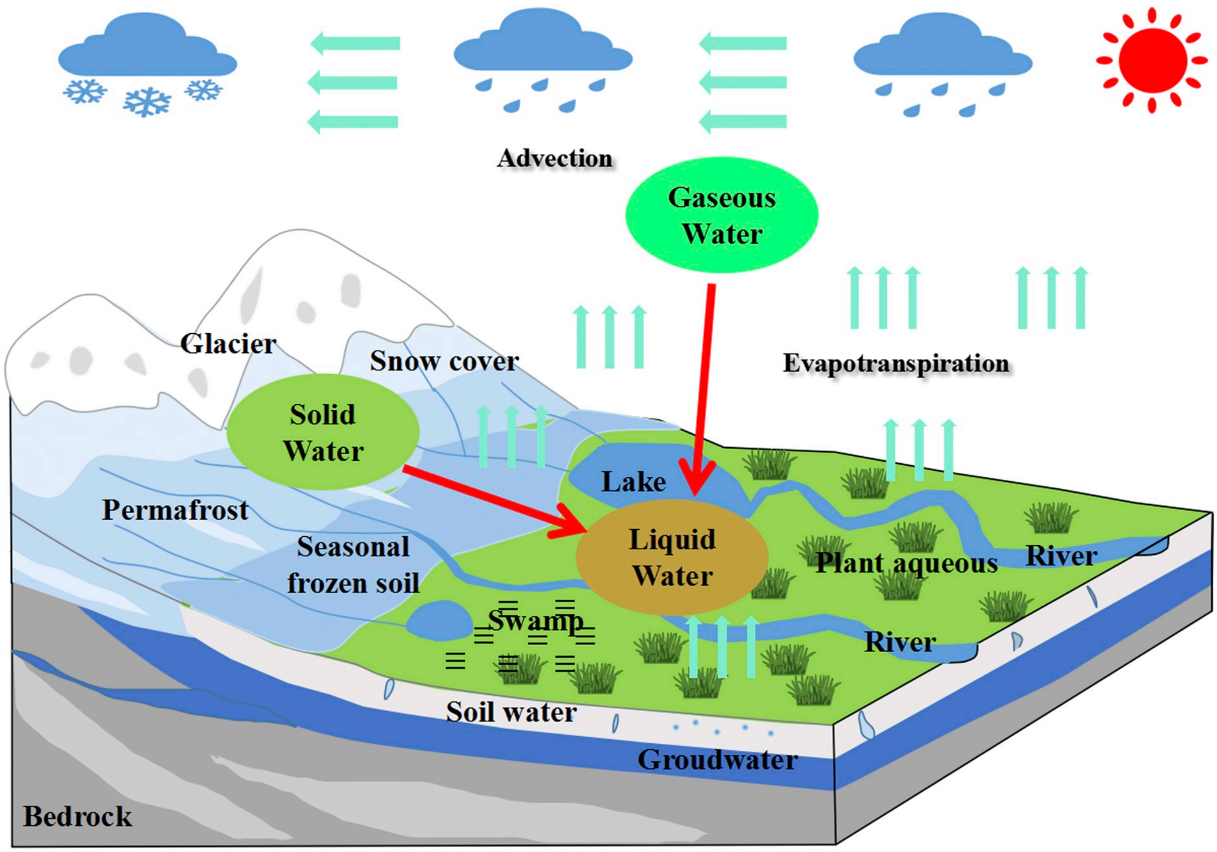

Fig. 1. The sketch of multiphase water transformation in cold regions of the western China.

and 836 mm/a from 2010 to 2014 (Arendt et al., 2002; Gardner et al., 2011; Andresen et al., 2012; Yao et al., 2012a; IPCC, 2013; Zhao et al., 2017). Furthermore, the extent of snow cover has decreased in the Northern Hemisphere, especially in spring, and the extent of annual mean snow cover decreased with statistical significance over the period 1967–2012, while no months had statistically significant increases (Robinson et al., 1993; Valt and Cianfarra, 2010). In addition, permafrost temperatures have increased in most regions since the early 1980s, although the rate of increase varied regionally; the temperature increase for colder permafrost was generally greater than for warmer permafrost, where the active layer thickness increased by a few centimeters to tens of centimeters since the 1990s (Jorgenson et al., 2006; Cheng and Wu, 2007; Li et al., 2012; IPCC, 2013).

In the case of transformation from liquid to gaseous water, evapotranspiration over land increased from the early 1980s up to the late 1990s (Jung et al., 2010; Wang et al., 2010; Wild et al., 2008); Wang et al. (2010) found that global evapotranspiration increased by a rate of 0.6 W/m^2 per decade for the period 1982–2002, and it has acted as a constraint to further increases in global evapotranspiration after 1998 (Jung et al., 2010). The hydrological effect is also significant; while the average runoff has not changed for the majority of rivers, year-to-year variability has increased (IPCC, 2013). Currently, the warming of the climate system is significant in China, where the annual mean temperature has increased by 0.9–1.5 °C in the past century; the rate of warming has increased since 1990 in the Tibetan Plateau (Ding and Wang, 2016). Considering this background, MWT can be characterized by the rapidly shrinking cryosphere (Kääb et al., 2007; Kabel et al., 2012; Malatinszky et al., 2013; Liu et al., 2015a; Zhao et al., 2017). In particular, the glacier and permafrost areas have been reduced by 18.6% and 10.1%, respectively, from the 1960s to 2010s in western China (Qin et al., 2005; Ren et al., 2010; Liu et al., 2015b; Zhao et al., 2017).

Multiphase water transformation has had a direct influence on hydrological processes and the water cycle in the studied region (Qin et al., 2005; Yao et al., 2012b; Li et al., 2014). For example, meltwater has been one of the main water sources and controls the seasonal pattern of runoff, which affects use of water resources and ecological safety patterns (Li et al., 2010a). Meanwhile, permafrost degradation has led to an extensive recession of the alpine ecosystem and variations in hydrological processes (Li et al., 2016a; Ma et al., 2016). MWT accelerates the water cycle and affects spatial and temporal distribution of precipitation and evapotranspiration, soil water content, and river runoff (Ganopolski and Rahmstorf, 2001; Qin et al., 2005; Li et al., 2008; Kundzewicz, 2008; Sorg et al., 2012; Yao et al., 2012a; Li et al., 2016b). These transformations also change the regional response of the water cycle to global climate change, such as variations in lake and runoff conditions (Chen et al., 2005, Chen and Han, 2010; Wang et al., 2013; Wei et al., 2017; Wang et al., 2017). Thus, a thorough review of MWT and its hydrological effect in cold regions is expected to contribute to a comprehensive understanding of global warming and regional climate responses.

The cold regions of western China mainly cover the Tibetan Plateau and surrounding regions (Fig. 2), which are very important for hydrological processes and water resources in the vast area of Central, Eastern, and Southern Asia. This cold region is called the “Asian Water Tower” and is the source of major Asian rivers, including the Yellow River, Yangtze River, Salween River (Nujiang River in China), Mekong River (Lancangjiang River in China), Bulamaputelahe River (Yarlung Zangbo River in China), Ganges River, Indus River, Ili River, Tarim River, Irtysh River, and Yeinisei River, which supply freshwater for the survival of about 2 billion people. The newest report from the second scientific expedition to Tibetan Plateau (2018) stated that the Asian Water Tower is becoming unstable, and the water reserves are decreasing in the eastern and southern Tibetan Plateau, but increasing in

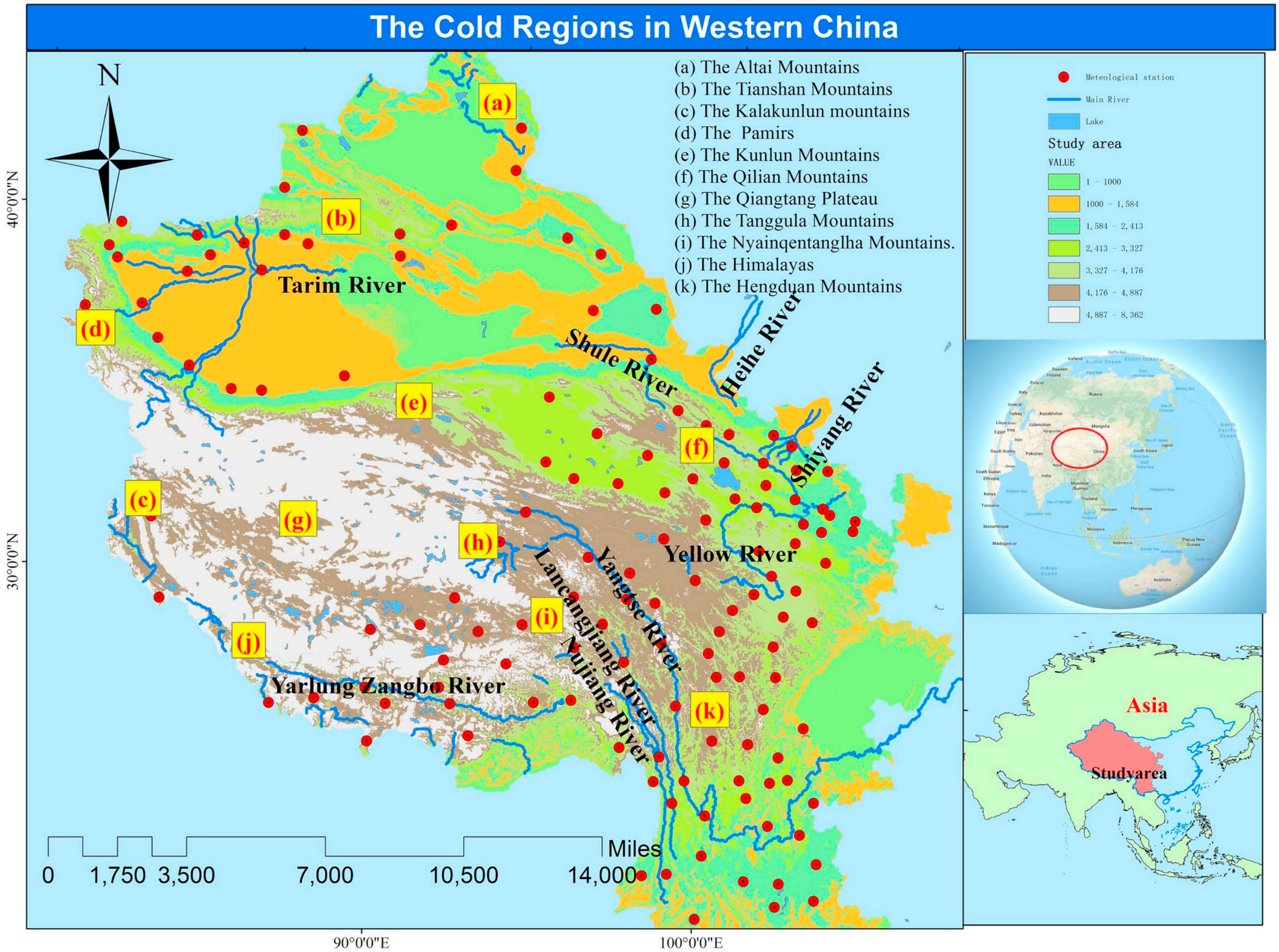

Fig. 2. Study region and its location in the Asia and the earth.

the western and northern regions. The structure and ratio of solid-to-liquid water is becoming imbalanced owing to an increasing amount of liquid water due to the accelerating melting of solid water (e.g., glaciers, permafrost, and snow cover). The accelerating transformation from solid to liquid water is increasing the risk of water shortages in some areas, with a corresponding risk of natural disasters resulting from e.g., glacial lake bursts, floods, and mud rock flows. Thus, a comprehensive study of MWT and its hydrological effects in the Asian Water Tower region is urgent in order to provide a scientific foundation for predicting future water resources and risk management.

Here, the climate background for MWT was analyzed for cold regions of western China based on meteorological data from 135 stations. MWT was explored using data for changes in glacier and permafrost properties, snowfall, evapotranspiration, and moisture recycling. Finally, the hydrological effects of MWT in study region are discussed. This analysis is expected to provide a broad understanding of the frequency, intensity, and duration of MWT and its hydrological effect on the Asian Water Tower. The study also develops a new theoretical basis for cold region hydrology.

2. Study region and data

In western China, the cold regions are mainly distributed in the Tibetan Plateau and its adjacent mountains (Fig. 2), which have a wide distribution of permafrost, glaciers, and snow cover, with a total area of about $5.1 \times 10^4\,km^2$, stretching from the Pamir and HinduKushin in the west to the Hengduan mountains in the east, and from the Altai mountains in the north to the Himalayas in the south, with an average elevation over 4000 m. This area is also referred to by scientists as the Third Pole (Qiu, 2008), which is geomorphologically the largest and highest mountain region on Earth. All peaks in the world over 7000 m a.s.l are within the study region, including fourteen globally acknowledged mountains over 8000 m a.s.l. (Yao et al., 2012b). Due to its high altitude and large area, the region plays a significant role in the Earth's climate system (Jin et al., 2005). The unique and complex interactions of atmospheric, cryospheric, hydrological, geological, and environmental processes have a large effect on the Earth's biodiversity, climate, and water cycles. Furthermore, the region borders more than ten countries and impacts > 2.0 billion people (Yao et al., 2012a), including the Tibet autonomous region, Qinghai province, and some regions of Gansu province, Xinjiang Uygur autonomous region, Sichuan province, and Yunnan province of China. This cold region provides resources, including water, pasture, and timber, as well as recreational and tourism opportunities to the billions of people inhabiting the plateau and surrounding regions (Yao et al., 2012a). Cryospheric processes in the study region are reacting sensitively to global changes, including glacier retreat, decrease in snow cover area, and permafrost degradation (Qiu, 2008).

To date, many studies have investigated the response of the cryosphere to climate warming in western China; however, these studies focused on a single river basin or single theme, such as glacier variations (Wu and Zhang, 2008; Liu et al., 2009; Yao et al., 2012b; Wan et al., 2014; Liu et al., 2015a; Li et al., 2014, 2016; Zhao et al., 2017). In addition, no systematic analysis of MWT in cold regions of western China has been performed. Therefore, in this study, we used meteorological observations, snowfall data, and GLEAM actual evapotranspiration data, in conjunction with the findings of previous studies of glaciers, permafrost, moisture recycling, runoff component, runoff variation, and lake changes to review the climate background, fact, and hydrologic effect of MWT in the study region. Details regarding the data sources are shown in Appendix A.

3. Results and discussion

3.1. Climate background

3.1.1. Lengthening ablation period

Climate warming has been significant in the cold regions of western China, where the annual average temperature has increased by 0.28 °C/10a during 1961–2016. The daily temperature range (DTR) significantly reduced by 0.18 °C/10a. This warming resulted in the gradual extension of the ablation period and continuous reduction of the freezing period; the number of ice days (ID) decreased by 2.35 d/10a during 1961–2016, while the number of frost days (FD) reduced by 4.09 d/10a in the study region (Fig. 3). Meanwhile, the growing season length (GSL) increased by 3.09d/10a at significant level (Fig. 3). The annual mean temperature also showed a general warming trend over the period of 1961–2016; in the study area, 134 stations exhibited a significant increase, while a higher degree of warming occurred at higher altitudes (Fig. 4). The DTR measured by 110 stations decreased, where the amplitudes of variation of the data from 104 stations were significant at the 0.05 level of significance (Fig. 4). The 26 stations showing increasing DTR were mainly distributed at lower altitudes. There were 130 stations for ID and 132 stations for FD that showed decreased values, where the larger reductions mainly occurred at higher altitudes (Fig. 4). About 131 stations for GSL presented an increase during 1961–2016, where the larger increase also occurred at higher altitudes (Fig. 4).

The increase in annual mean temperature in regions with an altitude above 3000 m a.s.l. was 0.05 °C/10a higher than that for altitudes of 1000–3000 m a.s.l. Correspondingly, the ID, FD, GSL, and DTR values for this high altitude region were 1.1 d/10a, 0.3 d/10a, 0.7 d/10a, and 0.02 °C/10a higher than in the lower altitude region, respectively. In general, the degree of warming increased with increasing altitude. The largest changes observed in the high-altitude regions (> 3000 m a.s.l.) that contain a large volume of water trapped in glaciers, permafrost, and snow cover were reflected by the clear extension of the ablation period.

Furthermore, the warming has accelerated since 1990; the temperature change was 0.39 °C/10a higher over the period of 1991–2016 than 1961–1990, while the corresponding decrease in DTR was 0.32 °C/10a (Fig. 5). The decrease in the number of ID and FD during

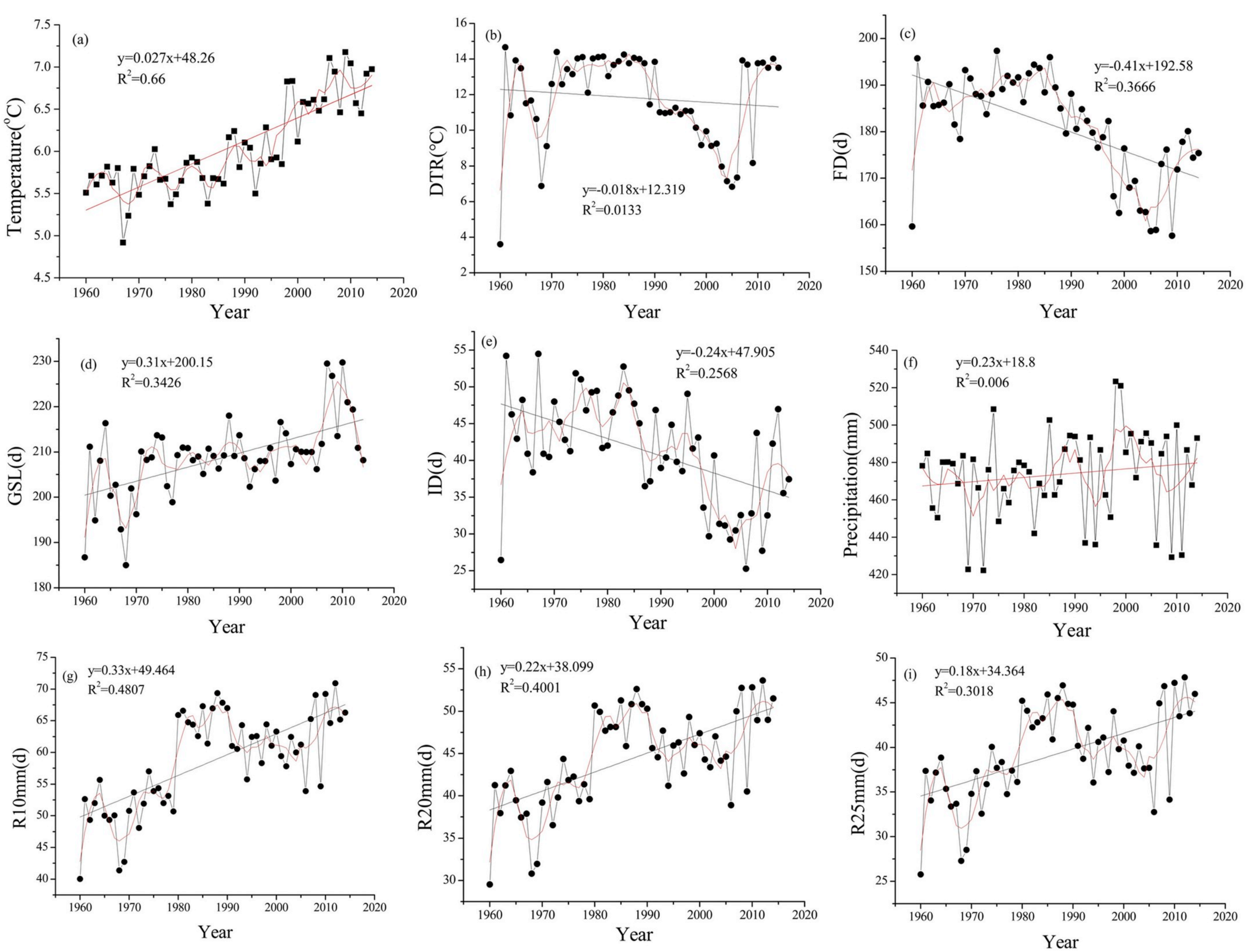

Fig. 3. Temporal variation of climate index in study region.

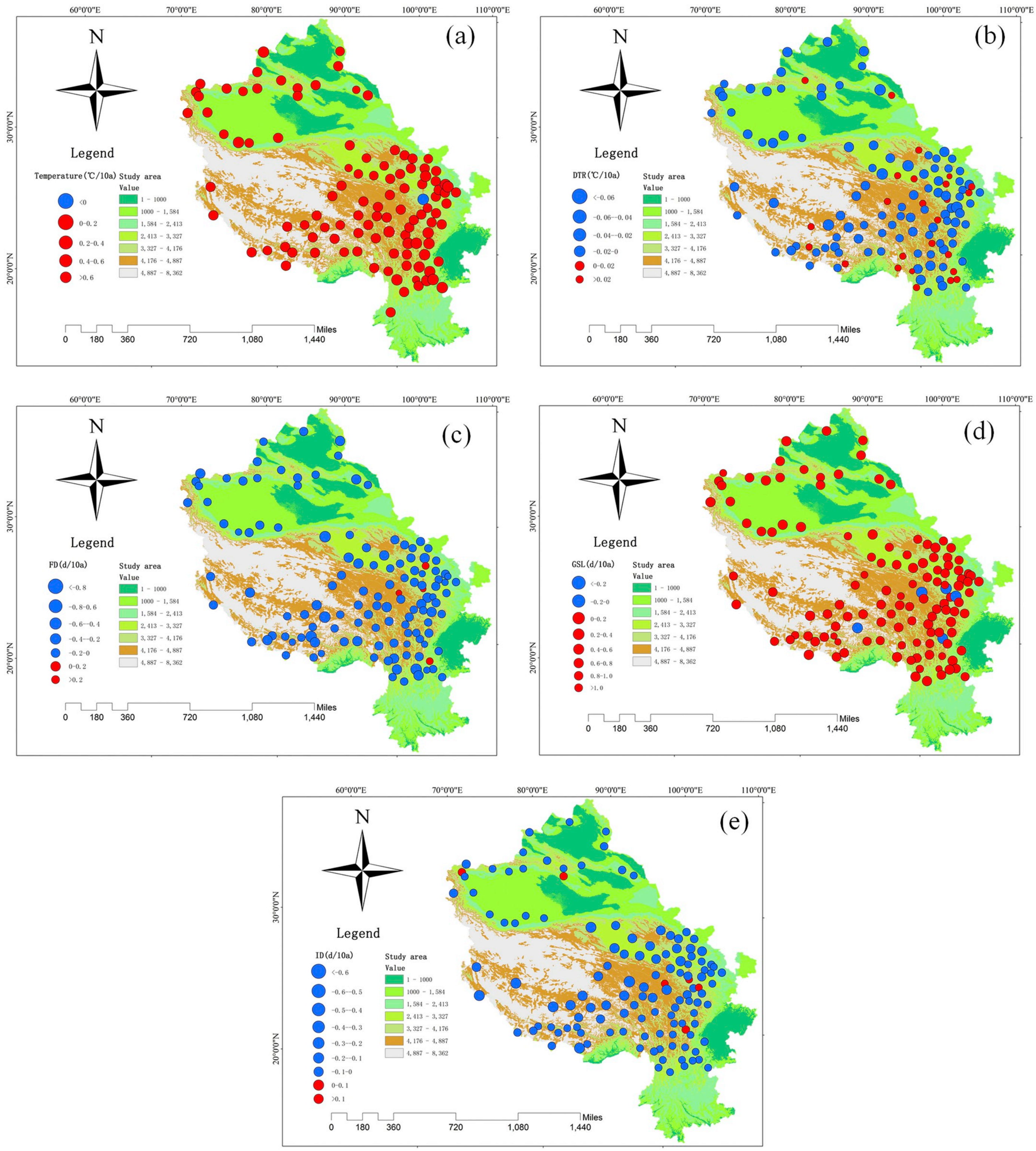

Fig. 4. Spatial changes of variational amplitudes for temperature index in study region.

1991–2016 was 1.58 d/10a and 4.47 d/10a higher than those during 1961–1990, while the increase in GSL was higher by 2.32 d/10a during 1991–2016 compared to 1961–1990 (Fig.5). These data indicate that accelerating warming is a main driver for MWT in the studied region.

3.1.2. Precipitation increase with altitude rise

In cold regions of western China, the increase in annual precipitation was also significant, with a rate of 2.52 mm/10a during 1961–2016 (Fig. 3). The R10mm, R20mm, and R25mm (number of days in a year where daily precipitation exceeds 10 mm, 20 mm, and 25 mm, respectively) increased by 3.29 d/10a, 2.24 d/10a, and 1.76 d/10a during 1961–2016 (Fig. 3), respectively. In the study area, 115 stations showed an increase in annual precipitation, while only 32 stations showed statistically significant increases at the 0.05 level over this time period (Fig. 6). Stations showing reduced precipitation were mainly distributed in the Hengduan mountains. However, the changes in R10mm,

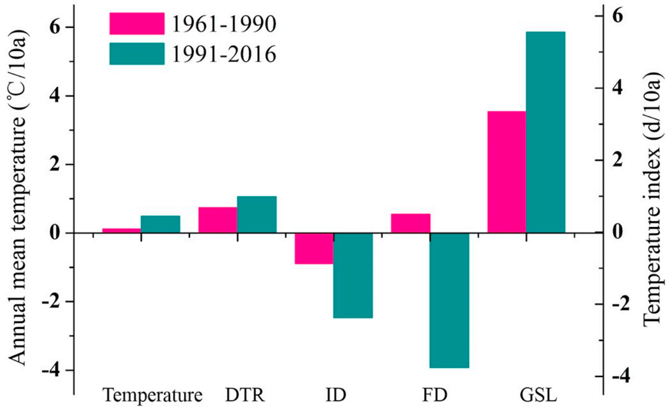

Fig. 5. The variational amplitudes of temperature index between 1961 and 1990 and 1991–2016.

R20mm and R25mm were widespread, with 124, 121, and 119 stations, respectively, showing a significant increase (Fig. 6), indicting the increasing trend of extreme precipitation events. Furthermore, the precipitation, R10mm, R20mm, and R25mm values in region with altitude above 3000 m a.s.l. were respectively 0.8 mm/10a, 4 d/10a, 2.9 d/10a, and 2.5 d/10a larger than those in the region with altitude of 1000–3000 m a.s.l. The increasing precipitation and number of rainy days indicates an increasingly humid climate in the cold regions of

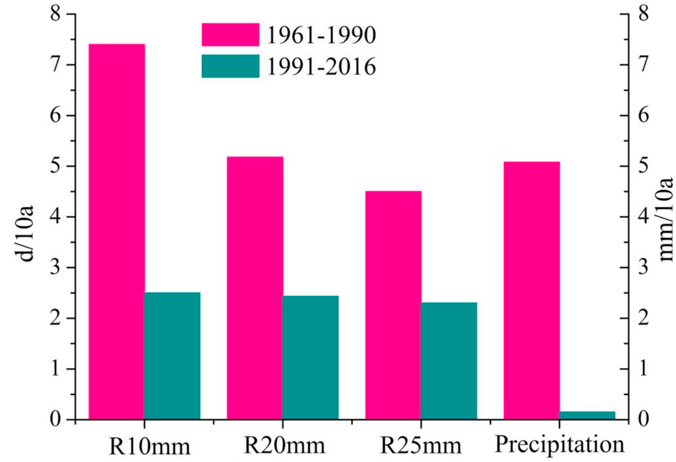

Fig. 7. The variational amplitudes of precipitation index between 1961 and 1990 and 1991–2016.

western China.

However, the rate of precipitation increase reduced since 1990; the increase was 4.88 mm/10a higher in 1961–1990 (5.08 mm/10a) than in 1991–2016 (0.2 mm/10a), while the precipitation fluctuated without a clear trend after 1990 (Fig. 7). The rate of increase in R10mm, R20mm, and R25mm also decreased after 1990; there was a decline in R10mm

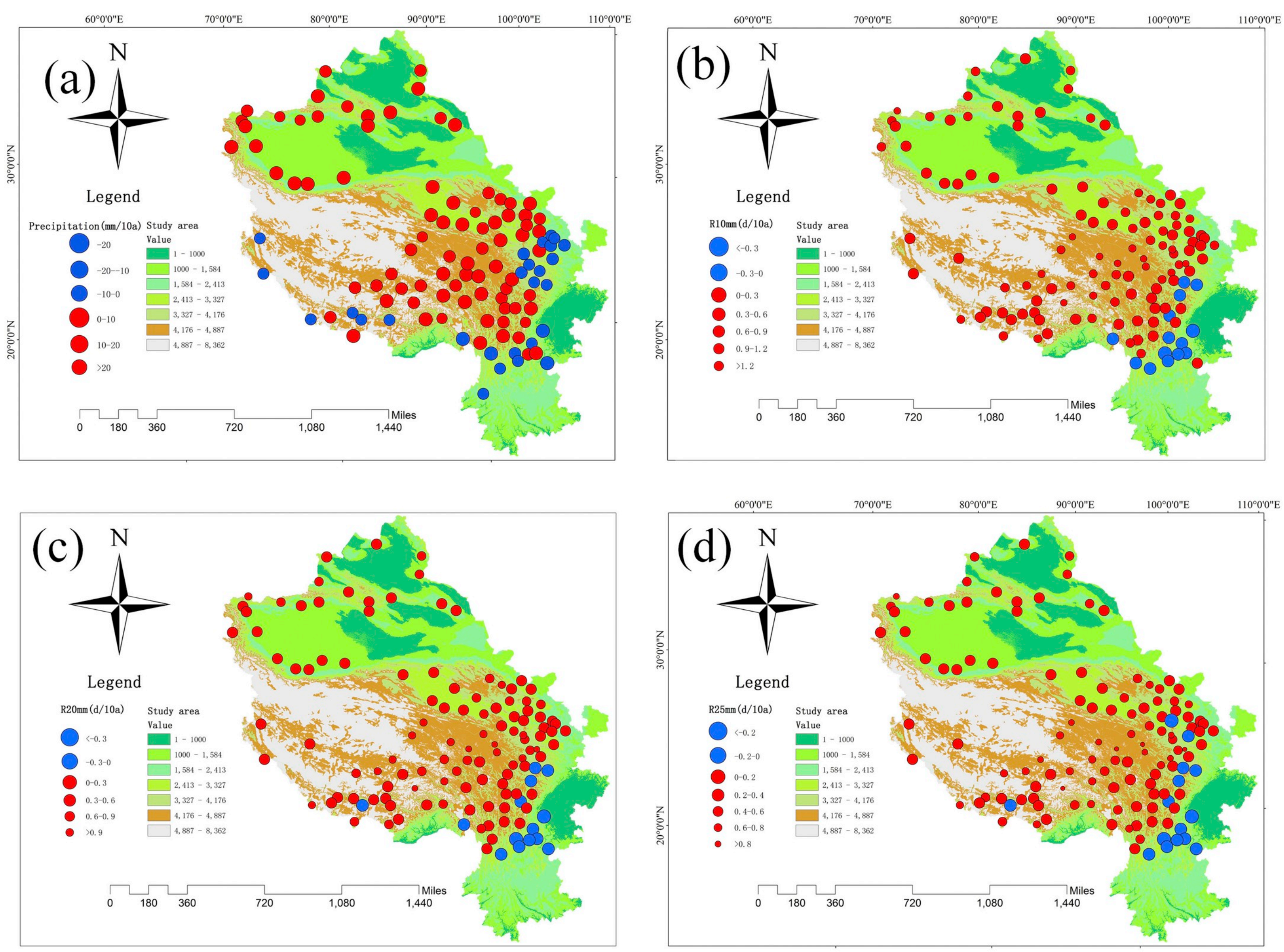

Fig. 6. Spatial changes of variational amplitudes for precipitation index in study region.

from 7.4 d/10a to 2.5 d/10a, in R20mm from 5.18 d/10a to 2.43 d/10a, and in R25mm from 4.5 d/10a to 2.3 d/10a over the period of 1961–1990 and 1991–2016, respectively (Fig. 7). These data confirmed that MWT was mainly driven by accelerating warming in the study region.

3.2. Facts of multiphase water transformation

3.2.1. Solid-liquid transformation

3.2.1.1. Glacier retreat. Glacier melting is the main transformation from solid to liquid water in cold regions of western China. Based on the first glacier inventory (measurements of glacier status between the 1950s to 1980s), there were 46,377 glaciers with the total glacier area of 59,425 km^2 and an approximate ice volume of 5600 km^3 in western China (Shangguan et al., 2004, 2007, 2009; Qin et al., 2004, 2005, 2016). The second glacier inventory (glacier status during 2004–2010) confirmed that there were 48,571 glaciers with an area of 51,800 km^2 and volume of 4300–4700 km^3 (Liu et al., 2015b). Around 72% of the total number of glaciers was within the Kunlun, Tianshan, and Kalakunlun mountains, Nyainqntanglha mountains, and the Himalayas; however, over 55% of the total glacier area and 60% of the total ice reserves were concentrated in the Kunlun mountains, Nyainqntanglha mountains, and Tianshan mountains, whiles around 82% of the glaciers in the study region have been shrinking over the past fifty years (Liu et al., 2015a). Many factors have caused glacier ablation, but precipitation and temperature were the most important, where the summer temperature mainly controlled glacier melting and winter precipitation mainly controlled the accumulation of new ice (Wang and Su, 2003; Liu et al., 2003; Liu et al., 2006; Liu et al., 2015a).

Different regional climates, environment, and topography result in local characteristics of glacier melting. The glacial areas measured for various mountain ranges over the period of 1960s–2010s are shown in Fig. 8, which are discussed here. In the northern Altai mountains, the glacier area was 666.00 km^2 in 1980, 614.30 km^2 in 2000, and 584.00 km^2 in 2010; however, in the southern Altai Mountains, the areas were 633.90 km^2 in 1972, 452.50 km^2 in 1989, 386.10 km^2 in 2000, and 329.00 km^2 in 2011 (Wang et al., 2011a; Bai et al., 2012; Yao et al., 2012c; Lv et al., 2012). In the Tianshan mountains, the glacier area on the northern slope was 2214.80 km^2 in 1990, 2078.70 km^2 in 2000, and 1884.20 km^2 in 2011, while the corresponding values were 4017.50 km^2 in 1990, 3794.50 km^2 in 2000, and 3490.9 km^2 in 2011 for the southern slope (Li et al., 2004, 2006; Shangguan et al., 2009; Xu et al., 2011; Wang et al., 2011b; He et al., 2014; Zhao et al., 2014; Xing et al., 2017). In the Kalakunlun mountains, glaciers area has reduced by 237.55km^2 during 1978–2015 (Xu, 2017). The glaciers area in the western Kunlun mountains was 2986.70 km^2 in 1990, 2984.20 km^2 in 2000, and 2979.00 km^2 in 2011, while the values for the eastern Kunlun mountains were 2197.40 km^2 in 1990, 2047 km^2 in 2000, and 1933.1 km^2 in 2010 (Li et al., 1998; Xu et al., 2006; Shangguan et al., 2007, 2009; Zhang et al., 2010a, 2010b; Jiang, 2012; Li, 2014).

The glacier area in the Tanggula mountains decreased from 2062.19 km^2 in 1990 to 1725.47 km^2 in 2015 (Zhang et al., 2010a, 2010b; Wang, 2017a). In the western Nyainqntanglha mountains, the glacier area was 931.50 km^2 in 1979, 878.40 km^2 in 1991, 852.00 km^2 in 2000, and 737.60 km^2 in 2011 (Bolch et al., 2010; Zhang et al., 2010a, 2010b; Wang et al., 2012; Ji et al., 2014, 2015). In the Qilian mountains, the glacier area was 2017.81 km^2 in 1956, 1761.3 km^2 in 1990, and 1597.1 km^2 in 2010 (Liu et al., 2003; Wang et al., 2007; Cao et al., 2010; Zhang et al., 2010a, 2010b; Wang et al., 2011c; Zhang et al., 2012; Pan et al., 2012; Sun et al., 2015). In the Hengduan mountains, the glacier area decreased from 252.1 km^2 in 1974 to 227.2 km^2 in 2010 in Gongga mountains (Liu et al., 2010; Pan et al., 2011), and it decreased from 15.40 km^2 in 1974 to 13.00 km^2 in 2009 in the Yulong mountains (Wang et al., 2011d; Du, 2011). In the northern Himalayas, the glacier area decreased from 8878.02 km^2 in 1990 to 7594.03 km^2 in 2010 (Ji, 2018). The glacier area in the Longbasahu lake basin of the middle Himalayas decreased from 167.31 km^2 to 157.43 km^2 during 1980–2010 (Jiang, 2015), and it decreased from 491.64 km^2 to 410.87 km^2 in the Luozha basin of the eastern Himalayas during 1980–2007 (Li et al., 2011a). It decreased from 2075km^2 to 1985km^2 during from 1970 to 2000 in Qiangtang plateau (Wang et al., 2011e).The glacier area in the eastern Pamirs showed an obvious decrease from 1780 km^2 in 1972 to 1670 km^2 in 2011 (Shangguan et al., 2004; Shangguan et al., 2007; Shangguan et al., 2009; Zhang et al., 2010a, 2010b; Zeng et al., 2013). Fig. 8 shows that more glaciers were retreating in the Qiangtang plateau and the Tinashan, Kunlun, Qilian, and Altai mountains with the largest glacier cover, while fewer were retreating in the Yulong and Gongga mountains with the smallest glacier area.

As shown in Fig. 9, the rate of glacier retreat in the studied region increased after 1990. For the southern Altai mountains, the rate of retreat increased from 5.61 km^2/a during 1972–1989 to 10.67 km^2/a during 1989–2011, while for the Tianshan mountains it increased from 35.9 km^2/a during 1990–2000 to 45.3 km^2/a during 2000–2010. In the Qilian mountains, the retreat rate was 4.33 km^2/a during 1956–1990, whereas it increased to 8.17 km^2/a during 1990–2010. The retreat rate increased from 2.49 km^2/a during 1970–1990 to 4.03 km^2/a during 1990–2000 in the Qiangtang plateau. In the Tanggula mountains and northern Himalayas, the retreat rate was 5.7 km^2/a and 10 km^2/a higher in 1990–2000 compared to 2000–2010, respectively. In the western Nyainqentanglha mountains, the retreat rate increased from 4.43km^2/a during 1979–1991 to 7.04 km^2/a during 1991–2011. In addition, the retreat rate increased from 14.76 km^2/a during 1974–1990 to 19.13 km^2/a during 1994–2013. However, the retreat rate decreased in the Kalakunlun, eastern Pamirs, and Kunlun mountains by 2.85 km^2/a, 2.42 km^2/a, and 3.38 km^2/a after 1990, respectively.

3.2.1.2. Permafrost degradation. In cold regions, increases in the air temperature can thermally degrade permafrost. Such changes have widespread impacts on construction of infrastructure, resource development, and ecosystem resilience and can increase the risks of flooding, positive climate feedback effects, and the resulting damage to infrastructure and the environment. Hence, the state of the permafrost has the potential to affect the wellbeing of millions of people and the sustainable development of the Tibetan plateau (Ran et al., 2018).

Permafrost degradation due to warming of the climate is another type of transformation from solid to liquid water in cold regions of western China. This degradation is characterized by a decreasing maximum freezing depth and increasing permafrost active layer depth. Fig. 10 shows these values for the studied region, which confirms widespread permafrost degradation. Fig. 10a shows the average permafrost active layer depth, Fig. 10b and c shows the maximum freezing depthin in Tibet autonomous region and Qinghai Province over the period of 1960s–2000s.

The permafrost active layer depth increased significantly over the studied time period in the region. Based on continuous observations from 8 stations along the Qinghai-Tibet road, the average permafrost active layer depth increased from 252 cm in 2006 to 276 cm in 2011 with an average rate of increase of 4.7 cm/a (Fig.10a). Meanwhile, the average rates of increase in permafrost temperatures at a depth of 15 m and at the permafrost table were 0.018 °C/a and 0.015 °C/a, respectively, during 2006–2011. The increase in the permafrost temperature at these two depths in cold permafrost regions was higher than that in warm permafrost regions (Liu et al., 2014). At some measurement sites on the Tibetan plateau, the active layer depth increased at 7.8 cm/a over the period of 1995–2010 (Wu and Zhang, 2010). The area-mean active layer depth increased by 15 cm/10a on the Tibetan plateau during 1981–2010 (Guo and Wang, 2013). In the headwaters of the Urůmqi river in the Tianshan mountains, the permafrost active layer depth has been increasing since 1991, where the annual mean ground temperature also increased from −1.6 °C in 1993 to −1.0 °C in 2008,

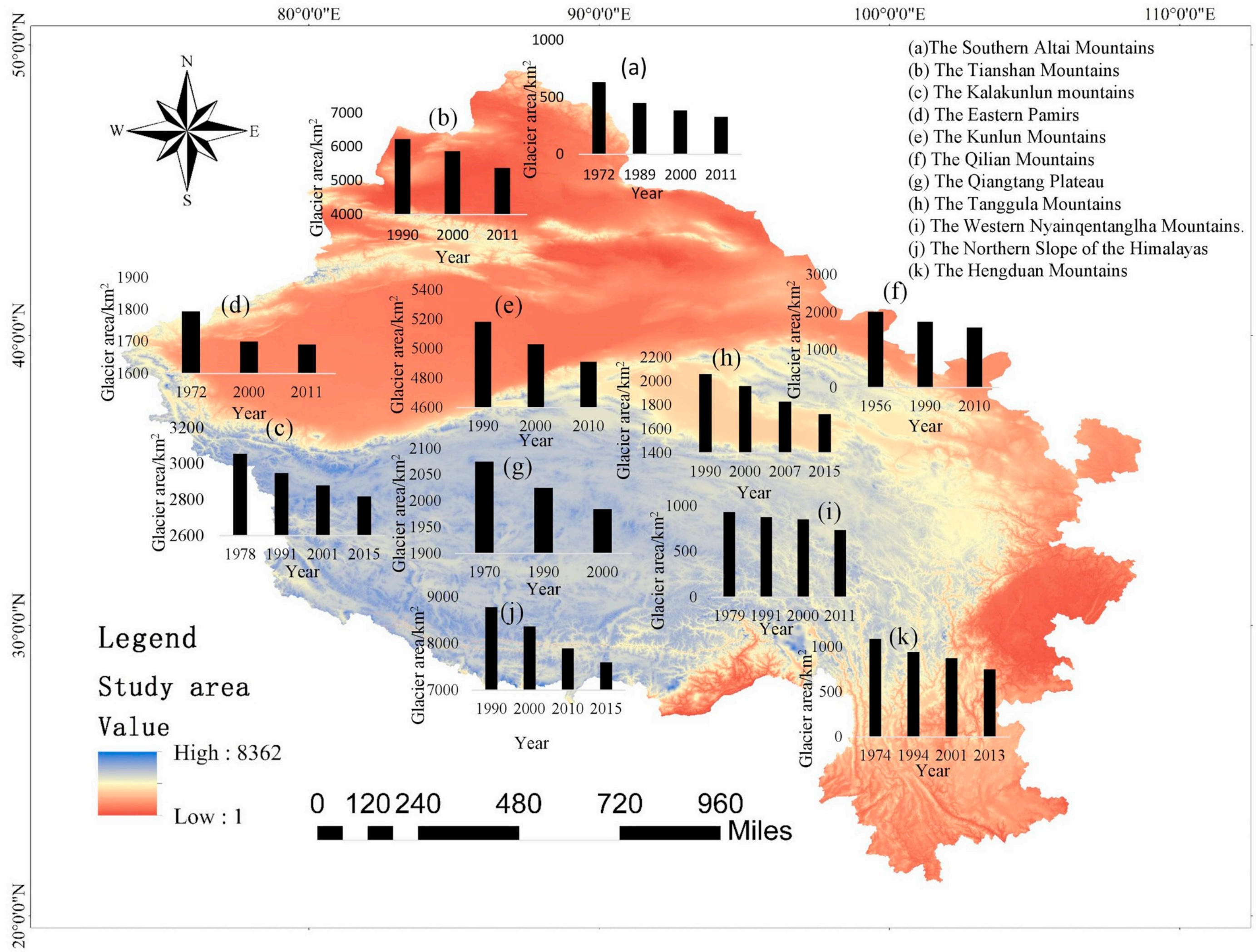

Fig. 8. Spatial distribution of glacier area change in study region (x-coordinate is year, and y-coordinate is glacier area).

and the estimated permafrost depth was 7.7 m less in 2008 than in 1992. Hence, it is concluded that the permafrost has recently been degrading rapidly from bottom to top, especially since the start of the 21st century (Zhao et al., 2010). Satellite data indicates that the onset dates of the spring thaw have advanced by 14 d, whereas the autumn freeze date was delayed by 10 d in the Tibetan Plateau over the period of 1988–2007 (Li et al., 2012b). The high rate of increase in the active layer depth may have been the result of local disturbances as more recent studies indicated rates of 1.33 cm/a for the period 1981–2010 and 3.6 cm/a for the period of 1998–2010 (Li et al., 2012a).

The maximum freezing depth also decreased in the studied region, especially after 1990. The data from 16 stations in Qinghai province showed that the maximum freezing depth continuously decreased by 4.8 cm/10a during 1961–2001, while the average depth decreased from 144 cm in 1961–1970 to 124 cm in 1990–2001 (Fig.10b). In the Tibet autonomous region, the average maximum freezing depth measured at 17 stations also decreased with a rate of 5.5 cm/10a during 1961–2010. The rate of decrease increased after 1990; it decreased by 14 cm from 1961 to 1990 to 1991–2010 (Fig. 10c), while the start date of thawing was advanced with a rate of 2.1–5.2 d/10a during 1971–2010. These variations were directly related to the significant increase in air and soil temperature (Du et al., 2012). The maximum freezing depth in the Tianshan mountains also decreased, but without a clear trend (Cheng and Wu, 2007; Zhao et al., 2010; Du et al., 2014). In addition, the maximum freezing depth decreased by 12 cm in the source region of the Yellow river over the past three decades (Jin et al., 2010). The depth of seasonally frozen ground in western China decreased by 20–40 cm since the early 1960s (Li et al., 2008), while it decreased by up to 33 cm since the middle of 1980s on the Tibetan plateau (Li et al., 2009). The permafrost depth in the Qinghaihu lake basin also clearly decreased (Yuan, 2016). The area-mean maximum freezing depth of seasonally frozen ground decreased by 34 cm/10a on the Tibetan plateau during 1981–2010, while the start dates for freezing of permafrost and seasonally frozen ground were linearly delayed by 3.8 and 4.0 d/10a, respectively, while the end dates of freezing advanced linearly by 5.9 and 4.6 d/10a, respectively, resulting in freeze durations that were shortened by 9.7 and 8.6 d/10a, respectively (Guo and Wang, 2013).

Permafrost degradation was also confirmed by the decreasing permafrost area on the Tibetan plateau, which can be divided into three periods: a stable period (1980s); a period of rapid degradation (1990s); and a period of slow degradation (in recent years) (Li, 2013; Feng et al., 2016). The degradation was concentrated around rivers, lakes, and valleys of the southern Tibetan plateau, especially the island permafrost regions, while there was no obvious variation in the northern Tibetan Plateau (Lu et al., 2017; Gao et al., 2017; Yang et al., 2017a; Wu et al., 2018). During 1981–2010, the near-surface permafrost area decreased by a rate of $9.20 \times 10^4\,km^2$/10a on the Tibetan Plateau (Guo and Wang, 2013). Under climate warming, the total area of thermally degraded permafrost was $\sim 153.76 \times 10^4\,km^2$ in 2010, corresponding to 88% of the permafrost area in the 1960s (Ran et al., 2018). The mean

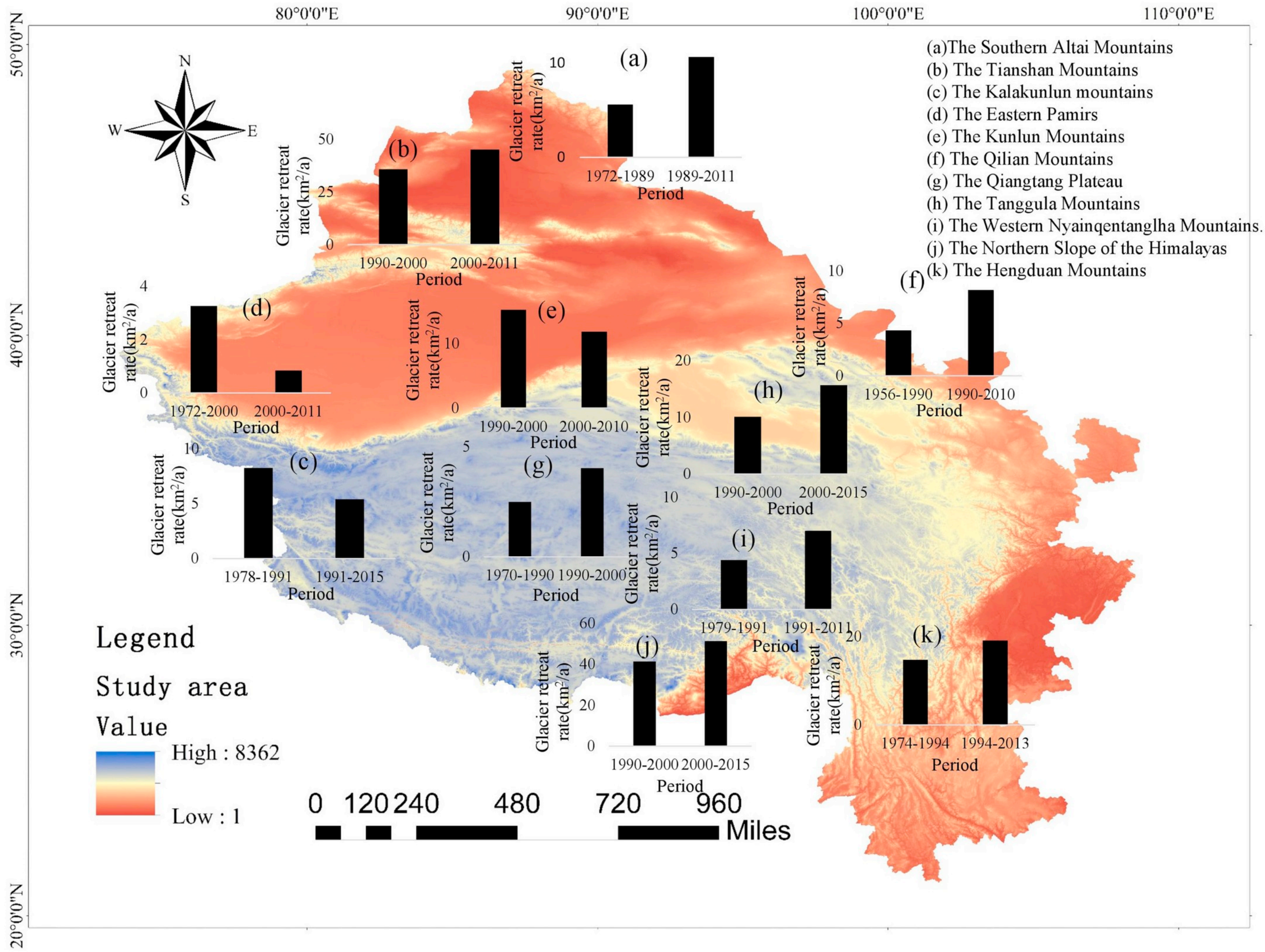

Fig. 9. Spatial distribution of glacier area retreat rate before 1990 and after 1990 in study region (x-coordinate is year, and y-coordinate is glacier area retreat rate).

lower boundary elevations of the very cold, cold, cool, warm, very warm and likely thawing permafrost areas have increased by 88, 97, 155, 185, 161, and 250 m, respectively. Over the past 40 years, the permafrost area in the Qilian mountains has reduced by 2.09×10^4 km^2, where most permafrost loss (1.19×10^4 km^2) occurred during the 1990s and 2000s (Zhang et al., 2014a). During 1972–2012, 833 km^2 of permafrost was lost in the source region of the Yellow river (Ma et al., 2017).

Simulations showed that the permafrost area continuously decreased on the Tibetan plateau over the past 50 years, with areas of 1.60×10^6, 1.49×10^6, 1.45×10^6, 1.36×10^6, and 1.27×10^6 km^2 lost in the 1960s, 1970s, 1980s, 1990s, and 2000s, respectively. Permafrost degradation has accelerated since the 1980s, with a total area of $\sim 3.3 \times 10^5$ km^2 lost; this is equivalent to ~20% of the total permafrost area in the 1960s (Cheng et al., 2012). Economic development on the Tibetan Plateau is mainly concentrated around livestock breeding and there is a very low level of industrialization (Hu, 2006). The main influence of human activities on the environment has been overgrazing, although localized and weak. Hence, it can be concluded that the permafrost degradation is due to climate warming; the annual mean temperature increased by 0.38 °C/10a since the 1980s, while the ground temperature of the permafrost active layer also increased by 0.1–0.3 °C (Cheng and Wu, 2007). The permafrost degradation is continuing; the most recent report indicated that the permafrost underground ice reserves over the Tibetan plateau was 12.7×10^{12} m^3, where $\sim 2.2 \times 10^{12}$ m^3 is deposited in > 10 m thickness of permafrost. It has been predicted that permafrost degradation could result in 300–600 $\times 10^8$ m^3/a of underground ice being converted into meltwater in the next 50 years (Zhao et al., 2017). These variations describe the persistent transformation from solid to liquid water by permafrost degradation in the studied region.

3.2.1.3. Decreasing snowfall. The transformation from solid to liquid water can be also characterized by decreasing snowfall and increasing rainfall. Fig. 11 shows the average snowfall over the period of 1961–2016. In the studied region, snowfall increased and then decreased during 1961–2016 (Fig. 11a). Specifically, the snowfall showed a statistically increasing trend from 1961 to 1990 with 3.6 mm/10a (Fig. 11b) and decreasing trend during 1991–2016 with 1.9 mm/10a (Fig. 11c), with a maximum around 1990. Meanwhile, the rainfall consistently increased by 2.52 mm/10a during 1961–2016 (Fig. 3).

The spatial distribution of the snowfall over the studied region is shown in Fig. 12. Fig. 12a shows that 127 stations measured increasing snowfall during 1961–1990, while the 31 stations that measured decreasing trends were mainly distributed in the Hendguan, eastern Qilian, and western Kunlun mountains. During 1991–2016, 99 stations showed a decreasing trend, while the 36 stations that showed an increasing trend were mainly located in the eastern Tibetan plateau and the Pamirs (Fig. 12b). Over the period of 1967–2012, little snowfall was

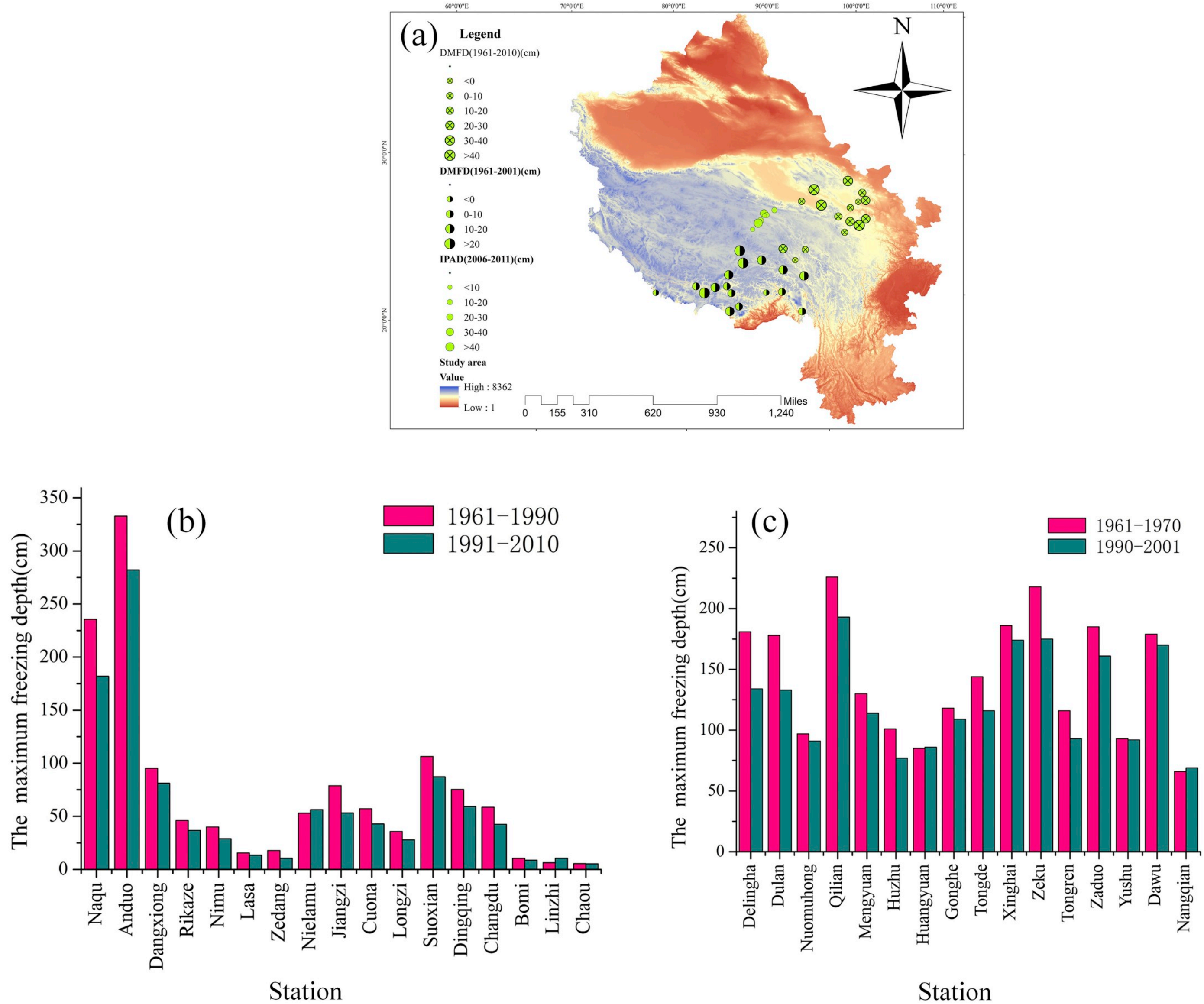

Fig. 10. Spatial distribution of the decreasing maximum freezing depth (the abbreviation is DMFD) and the increasing permafrost active layer depth (the abbreviation is IPAD) in study region (a); comparasion of the maximum freezing depth between 1961 and 1990 and 1991–2010 in Tibet autonomous region (b); comparasion of the maximum freezing depth between 1961 and 1970 and 1990–2001 in Qinghai province (c).

measured before the mid-1980s after the end of the 20th century, while there was a period of high snowfall from the late 1980s to the late 1990s. The Tibetan Plateau showed a significant decrease in annual mean snowfall days from 1981 to 2010, with a rate of 10.5 d/10a, where an abrupt change occurred in 1997 (Chu et al., 2017), accompanied by a decrease in winter and spring snowfall (Hu and Liang, 2014). A highly significant correlation existed between the decrease in snowfall days and the increase in air temperature. In addition, climate warming resulted in significant snowfall reduction in Tianshan mountains, especially at lower altitudes (Li et al., 2016). These data reflect the transformation from solid to liquid precipitation, especially after 1990.

Snow cover is the most important component of the cryosphere, with the largest seasonal and spatial variations over the Tibetan Plateau, which plays an important role in the hydrology and energy cycle for many Asian river basins. However, the snow cover depth in cold regions of western China clearly reduced from the end of the 1990s to 2005, attributed mainly to a decrease in snowfall and accelerated ablation (Ma, 2008). Previous studies showed that spring snow depth on the Tibetan plateau increased after the mid-1970s, but then decreased after 2002, while atmospheric heating also experienced interdecadal variations from cold to warm phases after 2002, which could be linked to the decreasing snow depth (Zhu et al., 2015). There was also a significant decrease in annual mean maximum snow depth on the Tibetan Plateau with a rate of 0.55 cm/10a during 1981–2010, while an abrupt decrease in snow depth also occurred around 1997 (Chu et al., 2018).

A decrease in snow cover area was also significant in the studied region. With increasing yearly average temperature, the snow cover area decreased on Mount Everest during recent years (Bengtsson and Berndtsson, 2003); some studies confirmed a positive correlation between the decrease in snow cover area and increase of cumulative average temperature over the Tibetan Plateau (Li, 1996; Yao et al., 2004; Wang et al., 2007; Chu et al., 2011). Between the 1960s and 1990s, the snow cover area on the Tibetan Plateau increased slowly, and has been decreasing slightly over recent years (Wang, 2017b). During 2001–2011, the seasonal snow cover area showed a significant negative correlation with temperature and precipitation in spring and summer, while the snow cover area showed a significant positive correlation with precipitation in winter, highlighting the influence of

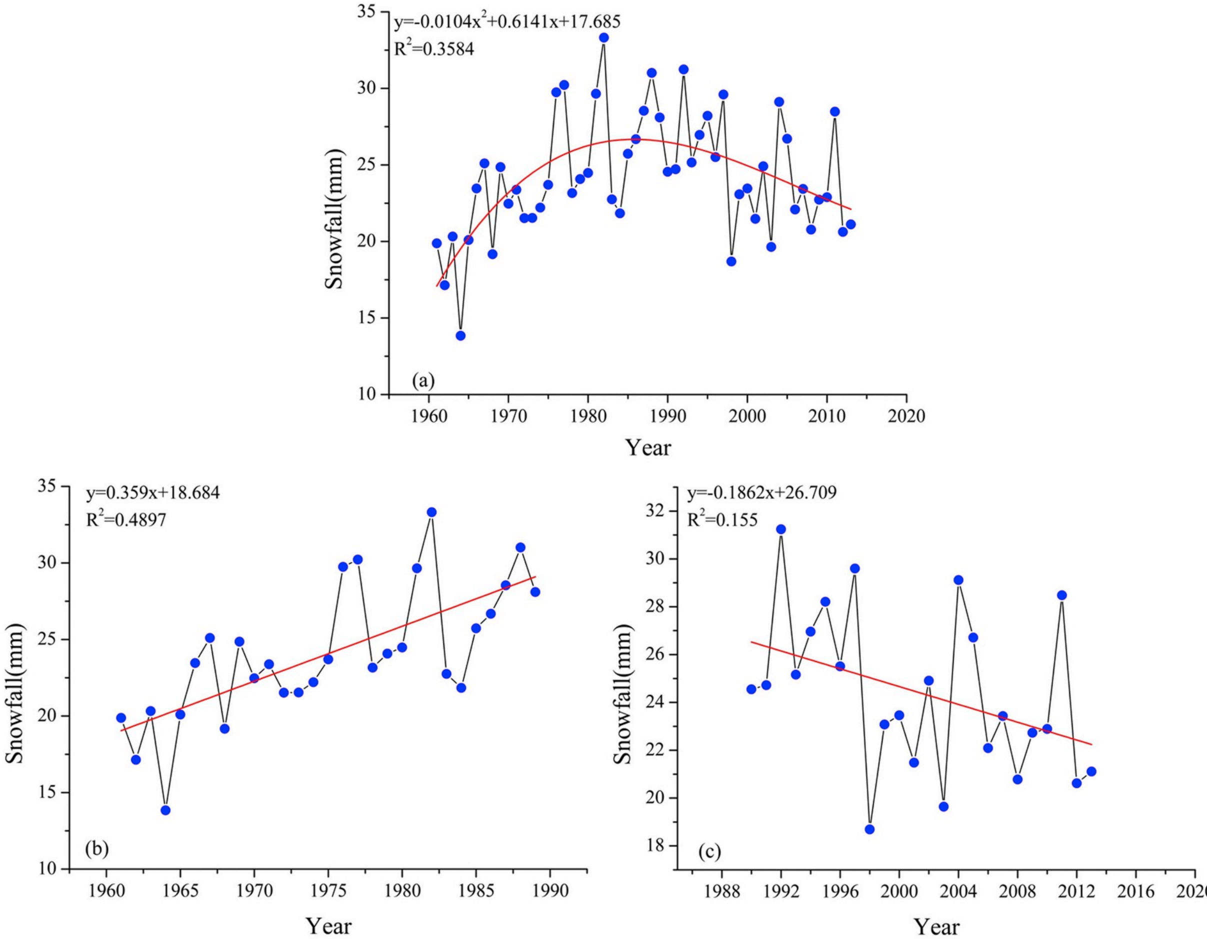

Fig. 11. Temporal variation of snowfall during 1961–2016 (a), 1961–1990 (b) and 1991–2016 (c) in study region.

climate warming and confirming decreased snowfall (Li, 2007; Wang et al., 2013). There was a very significant decrease in annual mean snow cover days on the Tibetan Plateau with a rate of 4.81 d/10a during 1981–2010, where 91.5% of the stations measured a decrease in snow cover days (Chu et al., 2015). Similarly, the snow cover area decreased in the Altai mountains during 2001–2014 (Chen et al., 2017), and also in the Tibet autonomous region (Chu, 2016) and Qilian mountains (Jiang and Ming, 2016). Both the decreasing snow depth and snow cover confirmed the accelerating transformation from solid to liquid precipitation due to climate warming.

3.2.2. Liquid-gaseous transformation

Under climate warming, enhanced evapotranspiration reflects the accelerated transformation from liquid to gaseous water. Actual evapotranspiration (AE) is the amount of water evaporated from water and soil surfaces and transpired by plants into the atmosphere. AE is an important component of the hydrologic cycle and global energy balance, and plays a significant role in the biosphere, hydrosphere, and atmosphere (Wang et al., 2012). Based on remote sensing data of the studied area, AE showed a significant increase (R^2 = 0.51) from 1980 to 2016 with a rate of 7.09 mm/10a (Fig. 13a), where the average evapotranspiration was 317.88 mm. This increasing trend was confirmed by the spatial distribution of AE (Fig. 13b and Fig. 13c), which showed the strongest effect in the Tianshan, Altai, and Qilian mountains and the southern Tibetan Plateau. When the average AE value for 1980–2016 was subtracted from the 2016 value, a positive value was determined, further confirming the increasing trend in the study region (Fig. 13c).

Global AE also increased by 7.1 ± 1.0 mm/10a from 1982 to 2007 (Jung et al., 2010). Over the last 32 years, the average AE significantly increased with a rate of 12.3 mm/10a in China, with an average value of 397.5 mm (Yang et al., 2015). In the case of potential evapotranspiration (PE), the highest increase of 15–25 mm/10a was observed for the northern region of Qinghai province, while an increase of 5–10 mm/10a was determined for the southeastern region (Liu et al., 2016). The temperature, precipitation, and AE in the source region of the Yellow river significantly increased by 0.34 °C, 11.4 mm, and 7.6 mm/10a, respectively, during 1970–2013 (Du et al., 2017). During 1981–2010, although PE decreased, AE increased over most areas of the Tibetan Plateau (Yin et al., 2013). The PE in the Tianshan mountains increased from 1960 to 2010 with a rate of 1.29 mm/10a, while this rate increased after 1990 (Liu et al., 2015b). During 1961–2010, the annual mean AE over the Tibetan Plateau was 543 mm, with a range of 147 to 687 mm; higher values were observed for the southern region, while lower values were measured for the northern Tibetan Plateau. In addition, the annual and seasonal mean AE showed a statistically significant increase for most stations. The annual area-averaged AE increased by 10 mm/10a, which tended to be affected by increasing soil water supply associated with the retreat of permafrost, increase in precipitation, and decrease in PE due to global warming (Zhang et al., 2018). These variations highlight that enhanced liquid-to-gaseous transformation is normal in the studied region under climate warming.

3.2.3. Gaseous-liquid transformation

The transformation from gaseous to liquid water is accelerating. This can be confirmed by moisture recycling, which includes contributions from terrestrial evaporation from the surface of soil and water, and plant transpiration to precipitation. Previous studies showed

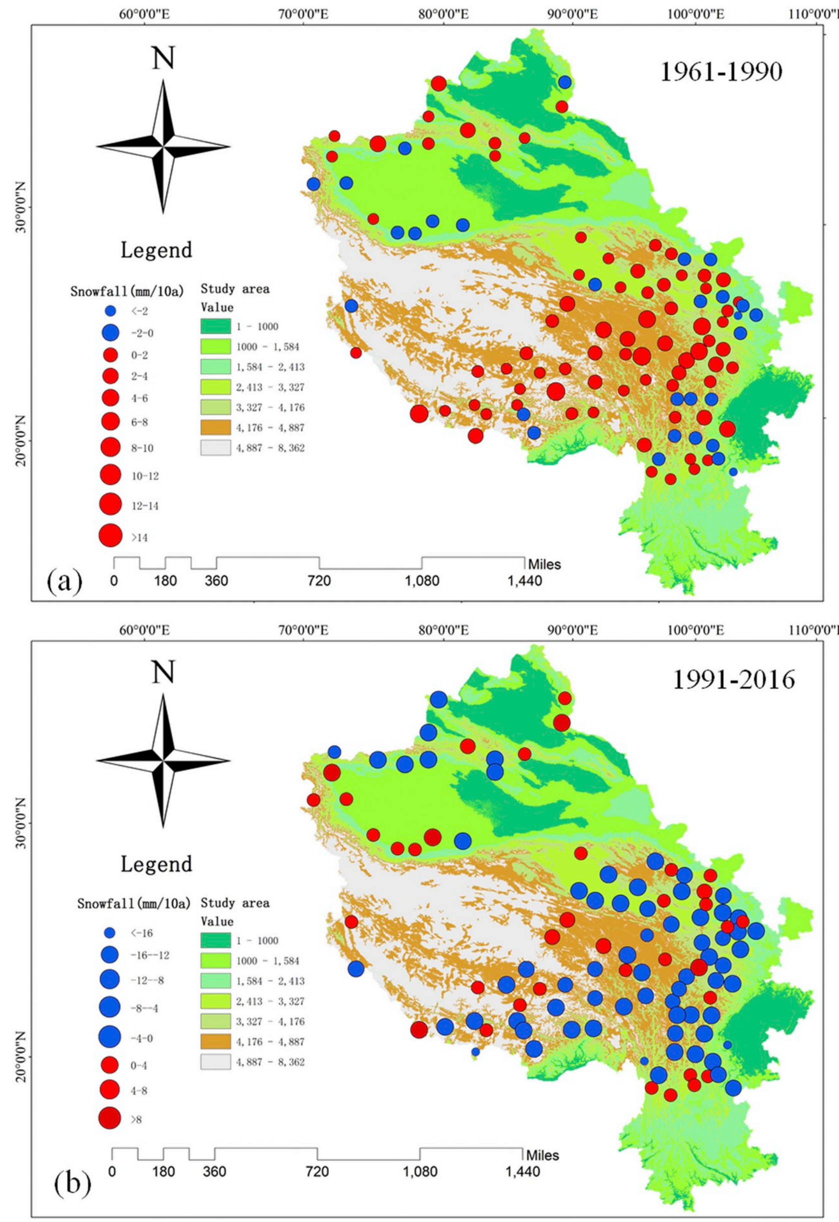

Fig. 12. Spatial distribution of snowfall trend during 1961–1990 (a) and 1991–2016 (b) in study region.

that moisture recycling had been a crucial part of local precipitation under climate warming, resulting in enhanced evapotranspiration, which has greatly affected the water cycle (Kong et al., 2013; Schlesinger and Jasechko, 2014; Li et al., 2016a, 2016b). Local moisture recycling played an essential role in maintaining an active hydrological cycle in the cold regions of western China.

Local soil evaporation contributed to 9.32% of the total precipitation in the Tianshan mountains, which was equivalent to 41.8 mm (Yao et al., 2016). Zhang et al. (2014a) found that moisture recycling in the Qilian mountains accounted for 20.76% of annual precipitation. Guo and Wang (2014) proposed that moisture recycling accounted for 40% of annual precipitation in the Tibetan Plateau; this contribution increased with a rate of 3.1%/10a during 1979–2008 (not including the western Tibetan plateau with an arid environment). Based on the newest research (Li et al., 2016b), moisture recycling contributed 87 mm (accounting for 24%) to total annual precipitation during May–September in the northern slope of the Qilian mountains; this contribution increased with increasing altitude (Fig. 14). The contribution from lake evaporation to local precipitation was 28.4–31.1% in the Nam Co basin of the central Tibetan Plateau during summer (Xu et al., 2011). In the Qinghaihu lake basin, the monthly contribution of lake evaporation to basin precipitation was around 3–38%, with an annual contribution of 23.42% (90.54 mm), the majority of which was concentrated over summer (Cui and Li, 2015). In the oasis stations of arid central Asia during the summer months, the contribution of recycling moisture to local precipitation was ~16% at Urumqi, but < 5% at small oases like Shihezi and Caijiahu (Wang et al., 2017b). Rudi et al. (2010) calculated that moisture evaporating from the Eurasian continent was responsible for 80% of China's water resource; furthermore, they showed that, due to the terrain, local moisture recycling was a key process near the Tibetan Plateau.

Deuterium excess records in two ice cores from the northwestern Tibetan Plateau showed that, on average, almost half of the precipitation was provided by local moisture recycling over the past decades, where the local moisture recycling ratio clearly increased, suggesting an enhanced hydrological cycle. This could be due to the rapid increases in temperature and precipitation, and changes in the land surface (An et al., 2017). The data discussed in this section are evidence of

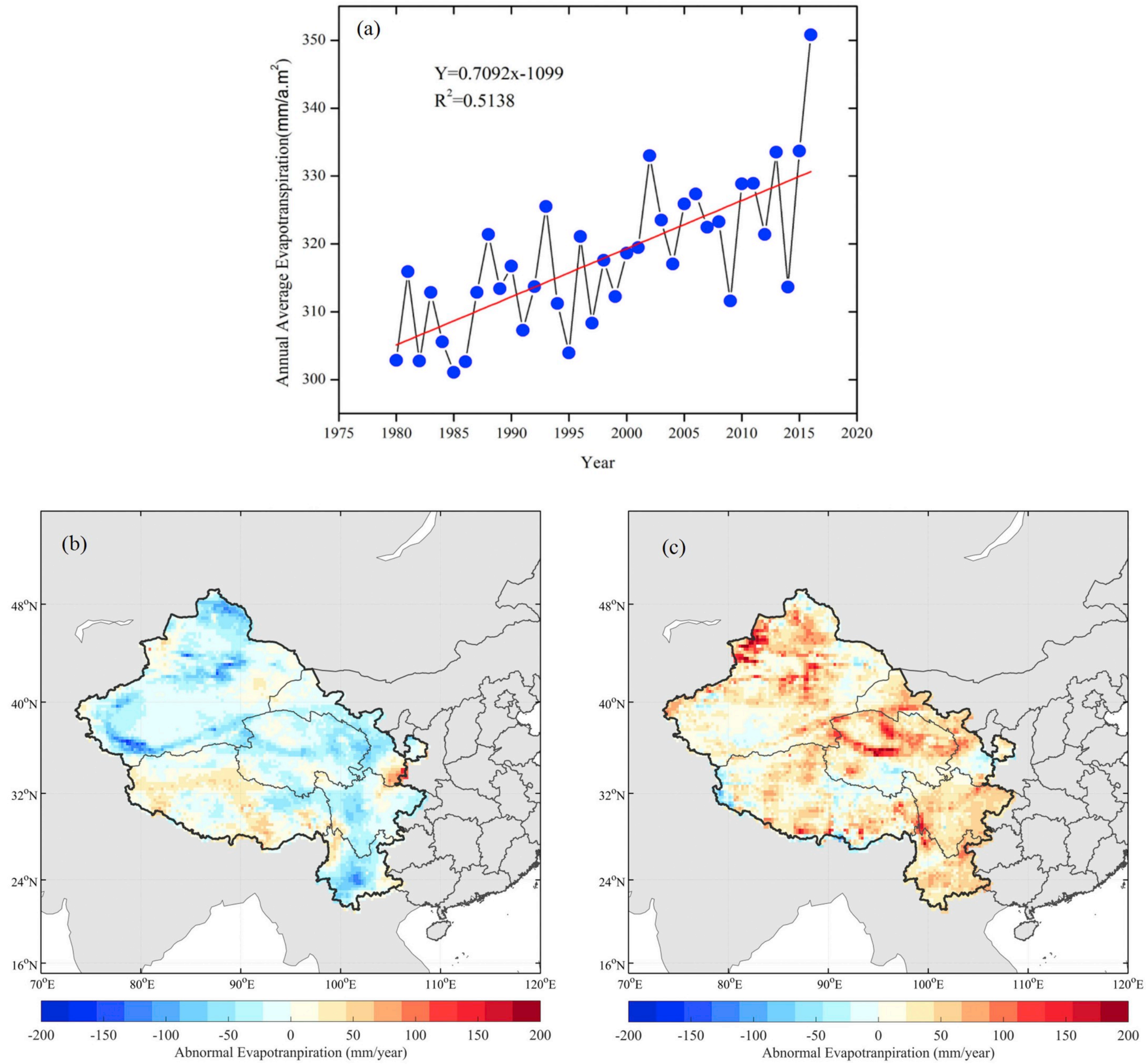

Fig. 13. Temporal variation of actual evapotranspiration during 1980–2016 (a); spatial distribution of actual evapotranspiration that the value in 1980 minus the average value during 1980–2016 (b); spatial distribution of actual evapotranspiration that the value in 2016 minus the average value during 1980–2016 (c).

the accelerating transformation from gaseous to liquid water under climate warming in the studied region, which is expected to play an important role in the future water cycle.

3.3. Hydrological effect of multiphase water transformation

3.3.1. The changed runoff component

Under MWT, glacier snow meltwater has been the main component of runoff in cold regions of western China, as shown in Fig. 15. The contribution of glacier snow meltwater to runoff was 13.4% in the headwaters of the Shulehe river basin in the western Qilian mountains with relatively large glacier covers (Zhou et al., 2015). Glacier snow meltwater accounted for 6% of the outlet river water in the Taolaihe river basin, while the contribution from supra-permafrost water was 15% (Li et al., 2016b). Supra-permafrost water and glacier snow meltwater contributed an average of 28% and 7%, respectively, to the outlet river water in Heihe river basin (Li et al., 2014, 2016c). Glacier snow meltwater only accounted for 3% of the outlet river water in Shiyanghe river basin, while the contribution from supra-permafrost water was 20% (Li et al., 2016a). Glacier snow meltwater from Qiyi glacier in the Qilian mountains increased by $0.41 \times 10^6\,m^3$ from 1960 to 1995 to 1996–2004, accompanied by a 0.41 °C temperature increase in the glacial region (Song et al., 2010). Gao et al. (2011a, 2011b) showed that the average glacier mass balance was −49.5 mm/a in the Qilian mountains during 1961–2006; the average contribution of meltwater to runoff was 14.1%, with an annual mean of $10.2 \times 10^8\,m^3$.

In the Tianshan mountains (Fig. 15), glacier snow meltwater accounted for 57–64% of the flow of the Kumalak river at 14 km, and > 57% in the Xiehela hydrology station (Kong and Pang, 2012). Throughout 2012, the proportions of precipitation and glacier snow meltwater were 17.6% and 14.7%, respectively, in the Urumqi river (Sun et al., 2015a). The contributions of glacier snow meltwater and

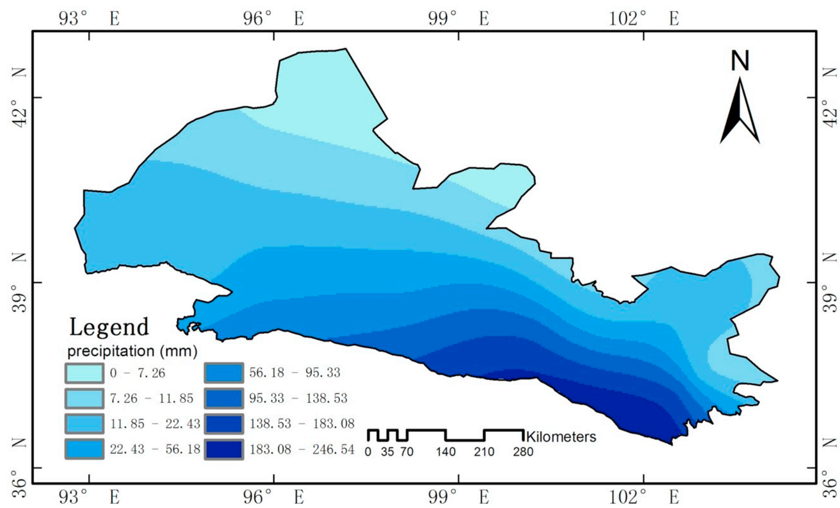

Fig. 14. Spatial pattern of the precipitation sourced from moisture recycling in Qilian mountains.

groundwater to the Yushugou river were 63% and 37%, respectively, in the Tianshan mountains (Wang et al., 2015). In the Aksu river, 36–44.4% of the runoff was derived from glacier snow meltwater (Sun et al., 2015b). Fan et al. (2015) confirmed that the contribution of glacier snow meltwater to runoff was 43% in the Tizinafu river. Meltwater increased by 563 mm water equivalent from 1959 to 1993 to 1994–2008 from Urumqi glacier No. 1, where the cumulative mass balance reached up to $-13{,}693$ mm over the past 50 years, equivalent to $3135.7 \times 10^4\,m^3$ (Sun et al., 2012). In the Tarim river basin, the glacier mass balance was -139.2 mm/a (1961–2006), where the contribution of glacier snow meltwater to river runoff was an average of ~41.5%; this contribution significantly increased after 1990 (Gao et al.,

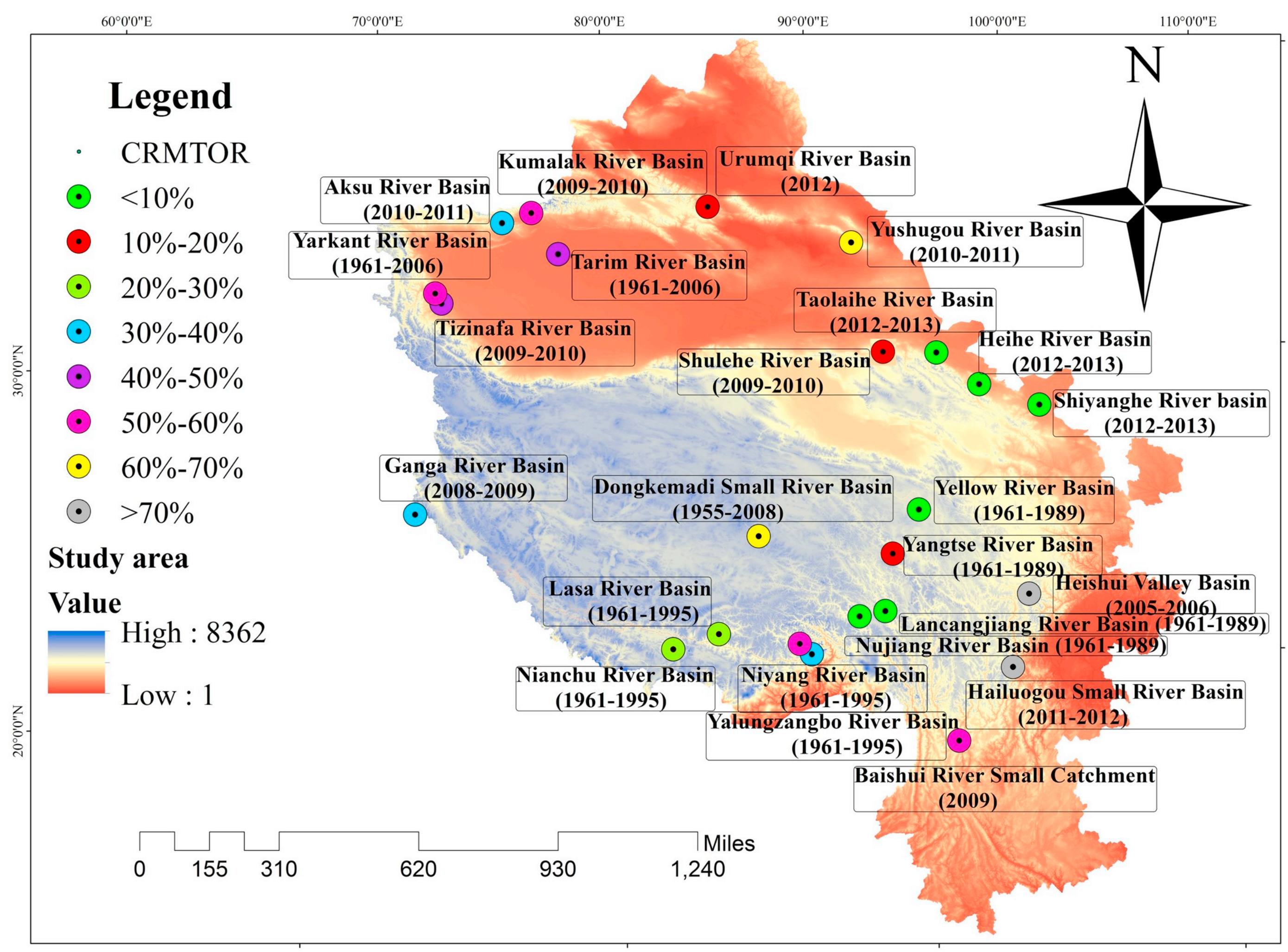

Fig. 15. Spatial pattern of the contribution rate from glacier snow meltwater to runoff (the abbreviation is CRMTOR) in study region.

2010a). The glacier melting depth in the upper reaches of the Yarkant river was 807.7 mm/a (1961–2006), while the contribution of glacier snow meltwater to river runoff was an average of ~51.3% over this period, which increased to 63.3% after 2000 (Gao et al., 2010b). Han et al. (2012) proposed that an increase of 2 °C in temperature would increase meltwater discharge by 45.0% for Tuomuer-type glacier basins in the Tianshan mountains. The contribution from glacier snow meltwater to runoff was clearly higher in the Tianshan mountains than in the Qilian mountains due to the larger glacier covers and ice volume.

In the Hengduan mountains (Fig. 15), hydrograph separation analysis showed that the contribution to runoff of glacier snow meltwater varied from 63.8% to 92.6%, and that of precipitation varied from 7.4% to 36.2% in the Heishui valley basin (Liu et al., 2008). In the small catchment area of the Baishui river in the Yulong mountains, a two component mixing model showed that and average of 53.4% of runoff came from glacier snow meltwater during the wet season, while the remainder was from precipitation (Pu et al., 2013). In addition, in the small Hailuogou river basin, glacier snow meltwater accounted for 72.84 ± 8.03% of the outlet runoff (Xing et al., 2015). Li et al. (2010a) showed that meltwater accounted for 54.6% of the annual mean runoff from the Hailuogou glacier basin in the Gongga mountains during 1994–2004, while the contribution of meltwater from higher altitudes to runoff increased in recent years. Within the Yanggong river basin in the Yulong mountains, an increase of 90.9% in the average glacier snow meltwater from 1979 to 1988 to 1994–2003 far exceeded the increased precipitation (1.1%) and river discharge (78.7%) (Li et al., 2010b). These results indicated the substantial contribution of meltwater to water resources.

Glacier retreat in the Himalayas has also greatly contributed to increased runoff in recent decades (Zhang et al., 2009; Li et al., 2011b). Maurya et al. (2011) showed that glacier snow meltwater, precipitation, and groundwater contributed to 32%, 53%, and 15% of runoff in the Ganga river basin. Glacier snow meltwater accounted for 38%, 21%, 26%, 50%, and 53% of runoff in the catchment areas of the Yarlung Zangbo river and its tributaries of Nianchu river basin, Lasa river basin, Niyang river basin, and Yigongzangbu river basin, respectively (Fig. 15) (Liu, 1999). In the Tibetan Plateau (Fig. 15), glacier snow meltwater accounted for 18.5%, 1.3%, 6.6%, and 8.8% of runoff in catchment areas of the Yangtze river, Yellow river, Lancangjiang river, and Nujiang river during 1961–1989, respectively (Yang et al., 2000). The runoff depth of the Dongkemadi river basin increased by 5.61 mm/a during 1955–2008, where ~66% of the increased runoff was contributed by glacier snow meltwater due to increased air temperature (Gao et al., 2011a, 2011b). By 2009, the glacier area decreased to 20.83% and 34.81% of the 1970 values in the Tuotuohe river and Baqu river basins, respectively, where the total meltwater supply in each basin was 2.56×10^9 and $1.24 \times 10^9\, m^3/a$, respectively. This was due to the increasing annual and summer stream flow during 1970–2009 (Wu et al., 2013).

The studies discussed here demonstrate that in cold regions of western China, basins with glaciers, snow cover, and permafrost have played a crucial role in regional water resources under global warming. The Tibetan Plateau, including the sources of Asian rivers, and the Altai and Tianshan mountains, are considered particularly significant for water resource management in the future.

3.3.2. Runoff increase

Under MWT, the outlet runoff also changed in cold regions of western China. With increasing precipitation, glacier ablation, and permafrost degradation, the annual average outlet runoff for major rivers in the Altai, Tianshan, Kunlun, and Qilian mountains, and the Tibetan Plateau increased, as shown in Fig. 16. In the Altai mountains, the runoff into the Barqin and Ulungar rivers increased with a rate of $1.87 \times 10^8\, m^3/10a$ during 1957–2008 and $0.27 \times 10^8\, m^3/10a$ during 1960–2010. As shown in Fig. 16, the 14 rivers sourced from the Tianshan mountains showed an increase in runoff during 1960s–2000s, while only the Tekes river showed a decrease during 1956–2000. Based on hydrological records of 8 rivers in the Kunlun mountains, the runoff for 7 rivers increased during 1960s–2000s, while that of the Golmud river decreased. In the Qilian mountains, the increase in runoff into the Heihe and Shulehe rivers was significant. The runoff increased from $12.73 \times 10^8\, m^3$ (1990–2002) to $15.56 \times 10^8\, m^3$ (2003–2016) in the Shiyanghe river basin (Li et al., 2017). During 1945–2014, annual average outlet runoff consistently increased, from $14.16 \times 10^8\, m^3$ (1945–1949) to $20 \times 10^8\, m^3$ (2011–2014) in the Heihe river basin (Cheng et al., 2017). In the western branch of the Heihe river basin, annual average outlet runoff increased by $1.37\, m^3/s$ (1990s–2000s) in the Taolaihe river basin (Xu et al., 2014). In the Shulehe river basin, annual average outlet runoff increased by $0.91 \times 10^8\, m^3/10a$ during 1958–2015, while it increased from $12.95 \times 10^8\, m^3$ in 1990 to $16.97 \times 10^8\, m^3$ in 2010 (Yang et al., 2017b). However, the runoffs into the Datong, Huangshui, and Buha rivers decreased during 1960s–2010s (Fig. 16).

In the Tibetan Plateau (Fig. 16), the runoff in the sources region of the Yellow river increased during 1956–1972, decreased during 1973–2000, then increased again during 2001–2012 (Kong and Pang, 2012). In the source region of the Yangtze River, the runoff increased with a rate of $6.67 \times 10^8\, m^3/10a$ during 1956–2012, where the rate of increase accelerated after 2000. The runoff increased with a rate of $0.5 \times 10^8\, m^3/10a$ during 1956–2012 in sources region of the Lancangjiang river basin (Kong et al., 2016). The runoff depth increased with a rate of 3.3 mm/10a during 1956–2013 in the sources region of the Yangtze river, which occurred in combination with increasing precipitation leading to relatively stable water storage (Du et al., 2017).

Changes in precipitation and glacier snow meltwater caused by climate warming may be major drivers of variations in streamflow in sources region of the Yangtze and Lancangjiang rivers, while the variations in the river flow of the Yellow river may be affected by precipitation, increased evaporation due to increased temperatures, and anthropogenic effects (Chang et al., 2018). The results indicated that precipitation, which mainly occurs during June–October (but varies in some monsoon-affected basins), was the major contributor to increased runoff in Tibetan Plateau basins (Liu et al., 2018a). In Yarlung Tsangpo river basin (Nuxia hydrological station), no distinct increase in annual streamflow during 1956–2013 was measured (Li et al., 2015). Decreasing runoff in the upper Yellow river basin and some sub-basins of Yalong river was attributed to the weakening east Asian monsoon (Liu et al., 2018b).

These findings suggested that MWT has accelerated the water cycle in cold regions of western China, indicated by the consistent increase in streamflow, especially in the Altai, Tianshan, Kunlun, and Qilian mountains, and in the sources region of the Yangtze and Lancangjiang rivers. The impact of MWT on streamflow is complicated. On one hand, annual evaporation could increase under warmer air conditions, which would result in decreasing streamflow. On the other hand, rainfall and meltwater can increase under MWT, which can increase streamflow (Li et al., 2015). Therefore, further hydrological modelling studies should be conducted to quantify streamflow changes and their uncertainty across the cold regions of western China.

3.3.3. Lake expansion

Lakes carry important information regarding global climate change and regional responses (Li et al., 2008; Wan et al., 2014) and are widely distributed in cold regions of western China (Yao et al., 2012a). Lakes affect the regional climate via the exchange of water and energy between the land and atmosphere, and also record past climate changes in their sediments. This study analyzed 78 big lakes (with areas > $10\, km^2$) in cold regions of western China to explore lake changes during recent decades. As shown in Fig. 17, most lake areas increased, other than those of Aibihu (Tianshan mountains), Zhiguicuo, Yangzhuoyongcuo, Xurucuo, Wurucuo, Renqingxiubucuo, Paikucuo, Lamucuo, Mapangyongcuo, Laangcuo, Gerencuo, Dangreyongcuo, Cuona, and

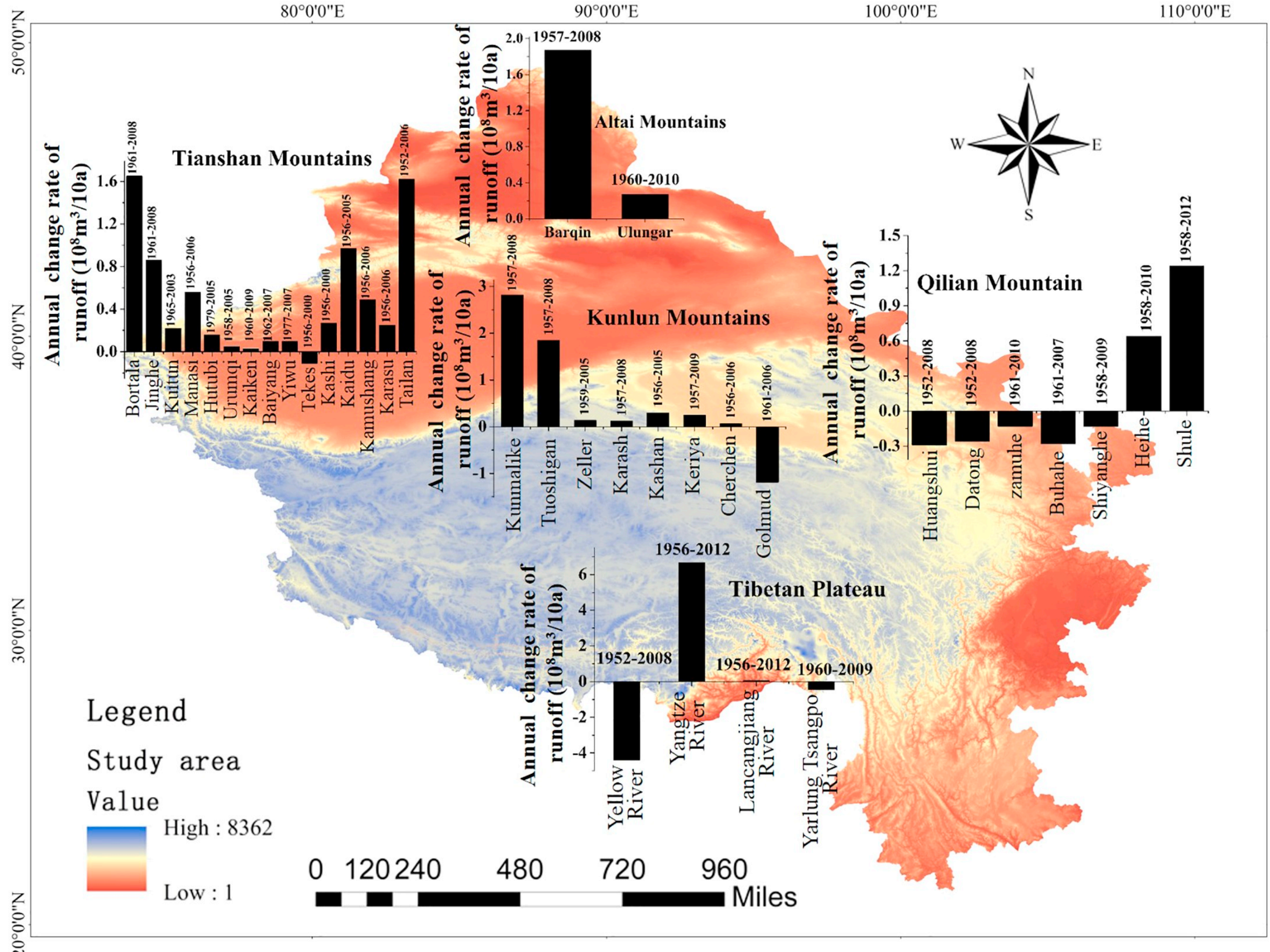

Fig. 16. Annual change rate of runoff for major rivers in study region.

Anglarencuo (southern Tibetan Plateau), Guozhuhu (western Tibetan Plateau), and Dongtaijiwaner (northern Tibetan Plateau). The overall lake area increased, with a mean value of 1.37 km^2/a (and a range of −3.92 to 14.82 km^2/a); Overall, 63 of the 78 lakes were in expansion, where these lakes are mainly located in the northern Tibetan Plateau, Hoh Xil region, Qilian mountains, Qaidam basin, and Kunlun mountains, while the shrinking lakes are mainly located in the southern Tibetan Plateau (Fig. 17).

The lake water level also increased in the studied region, with the exception of Yangzhuoyongcuo, Paikucuo, Mapangyongcuo, Laangcuo, Gerencuo, and Angzhicuo (southern Tibetan Plateau), and Maerguochaka, Jingyuhu, Cuorendejia, and Cuodarima (northern Tibetan Plateau). The lake water level change was between −1.480 and 1.038 m/a, with a mean value of 0.14 m/a (Fig. 17). The lakes with reducing water levels were mainly located in the southern Tibetan Plateau. From the 1970s to 2010s, the total lake area increased by 7240 km^2 in the Tibetan Plateau (18% of the 1970 area), while lake expansion mainly occurred during 2000–2010 (Zhang et al., 2017a). Over the past 25 years, 261 new lakes were observed in the Tibet autonomous region, and the total lake area increased from 24,161.1 to 30,549.2 km^2. A severe decline in lake area in the period of 1990–1995 was observed, followed by a rapid increase during 1996–2006, then stabilization during 2007–2013 (Liu et al., 2018a).

Under climate warming, most lakes in cold regions of western China clearly expanded, while only a few lakes in the southern Tibetan Plateau decreases in size (Song et al., 2013; Zhang et al., 2011; Zhang et al., 2014b). Lake changes are mainly affected by the geology, climate factors, and human activities (Zhang et al., 2017b), while climate factors showed the largest effect on lake changes in the studied region (Zhang et al., 2011). Under MWT, increased rainfall results in lake expansion, along with increased volumes of glacier snow meltwater and supra-permafrost water under climate warming (Ding et al., 2006; Wan et al., 2014; Chen et al., 2015). The main reason for the decreasing lake volume in the catchments of the Yellow river, Junggar basin, Turpan basin, Northern slope of the Karakorum mountains, and the Himalayas was that water loss from evaporation (due to the increased temperature) exceeded the increase in runoff from precipitation (Kang et al., 2010; Wan et al., 2014; Lin et al., 2017; Liu et al., 2018b). In addition, glacier-fed lakes showed a much more rapid expansion than non-glacier-fed lakes, which indicated that increasing glacier degradation was one of the main factors contributing to the expansion of Tibetan Plateau lakes (Yang et al., 2017a). In the Tibetan Plateau, lake variations agreed well with the spatial pattern of precipitation changes during the 2000s; however, glacier meltwater was shown to augment precipitation-driven lake expansion in this region (Song et al., 2014).

From the 1970s to 2010s, precipitation, glacier snow meltwater, and permafrost degradation have contributed about 74%, 13%, and 12%, respectively, to lake expansion in the Tibetan Plateau (Zhang et al., 2017a). Based on the analysis of Fang et al. (2016), increases in lake areas were attributed to: glacier melting and increased precipitation in

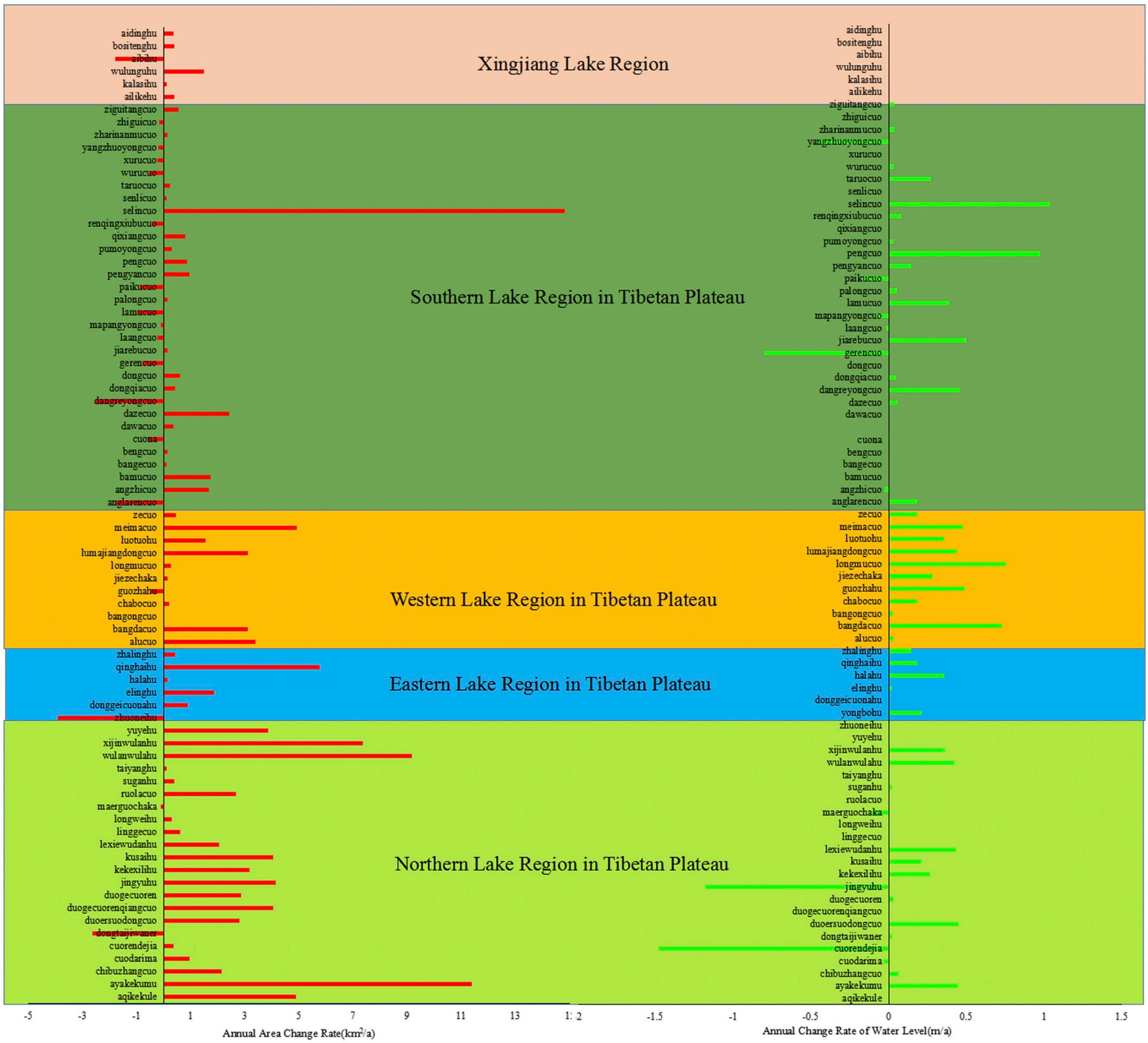

Fig. 17. The variation of lake area and water level in study region.

the Naqu region; increased precipitation, higher temperatures, and a decrease in evaporation in the Hoh Xil region; and rapid glacier melting and increasing precipitation in the Qilian mountains. Although evaporation throughout the Tibetan Plateau has decreased in the past 40 years, changes in precipitation played the dominant role in lake area changes (Yang et al., 2017b). The annual mean precipitation and temperature in this region increased considerably and evaporation decreased in recent years, resulting in increased lake areas since the 2000s in the Kunlun mountains (Song et al., 2014). Over the past 25 years, minimum temperatures, evapotranspiration, and high precipitation resulted in the rapid expansion of lake areas in the central Tibet autonomous region. However, high temperatures, low precipitation, a large amount of evapotranspiration, and melting of glaciers and permafrost are possible drivers of lake expansion in the northern region (Liu et al., 2018a). Hence, we conclude that MWT greatly contributes to lake expansion under climate warming, especially increased rainfall, glacier snow melting, and permafrost degradation.

4. Conclusions

As shown in Fig. 18, accelerating warming and a moist climate are the main drivers of MWT in cold regions of western China. The annual average temperature increased by 0.28 °C/10a during 1961–2016, which gradually lengthened the ablation period and continuously shortened the freezing period. The number of ID and FD significantly decreased by 2.35 d/10a and 4.09 d/10a, respectively, while GSL increased by 3.09 d/10a. Annual precipitation increased by 2.52 mm/10a (with larger increases at higher altitudes), while R10mm, R20mm, and R25mm also increased by 3.29, 2.24, and 1.76 d/10a, respectively during 1961–2016. After 1990, warming sped up, while the increase in precipitation slowed down.

All analyzed data showed that MWT processes are accelerating. Solid water is rapidly transforming to liquid water via three processes: (1) a clear retreat of glaciers which accelerated after 1990 in the southern Altai, Tianshan, Qilian, and Tanggula mountains, the

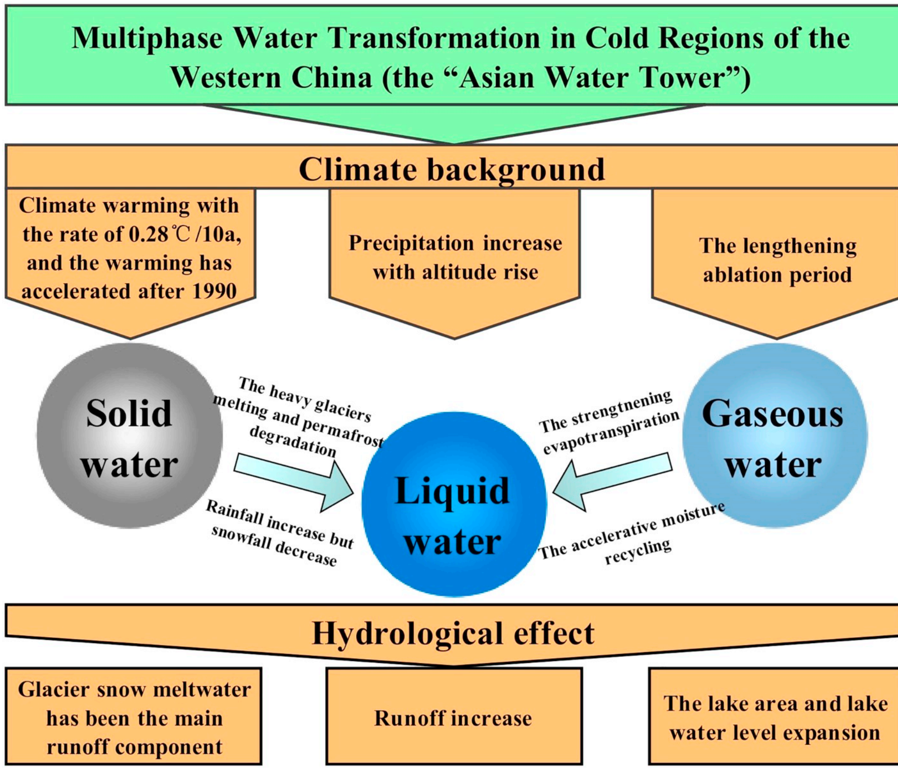

Fig. 18. The conceptual model of multiphase water transformation in cold regions of western China.

Qiangtang plateau, and the northern Himalayas; (2) permafrost degradation, where the average permafrost active layer depth increased by 24 cm (2006–2011) along the Qinghai-Tibet road, and the maximum freezing depth decreased by 24 cm (1961–2001) in Qinghai province and 14 cm (1961–1990 to 1991–2010) in the Tibet autonomous region; (3) solid precipitation reduced gradually and liquid precipitation increased. The snowfall decreased significantly by 1.9 mm/10a during 1991–2016. Considering the liquid to gaseous water transformation, evapotranspiration significantly increased during 1980–2016 (7.09 mm/10a). Moisture recycling greatly contributed to local precipitation. The final hydrological effect of MWT is characterized by three factors: (1) glacier snow meltwater was the main runoff component under enhanced glacier and snow melting; (2) Runoff in the 39 rivers clearly increased, with an average rate of $0.31 \times 10^8\ m^3/10a$. (3) The lake area and lake water level for the 78 big lakes also increased (Fig. 18).

This review provides a theoretical basis for explaining the water cycle, and enhances our scientific understanding of the mechanism and hydrological effects of MWT in the Asian Water Tower. The cold regions of western China profoundly affect Asian water resources and the local ecosystem. Under accelerating MWT, runoff is expected to continually increase, while lakes will expand for a certain period until they reach an inflection point where the decreasing cryosphere meltwater cannot provide for the demand in water resources, which would seriously threaten the livelihoods of the population and the sustainable development of the region. Therefore, future research should devote more attention to the environmental effect of MWT in cold regions of western China, which will elucidate the effect of an unbalanced and unstable Asian Water Tower. In addition, it is critical to develop scientific strategies for controlling or reducing potential threats to water security and ecological health in southern, central, and eastern Asia.

Acknowledgments

This study was supported by the Youth Innovation Promotion Association, CAS (2013274), National Nature Science Foundation of China (91547102), National Key R&D Program of China (2017YFC0404305), Longyuan Innovation and Entrepreneurship Project in Gansu Province, the open funding from State Key Laboratory of Urban and Regional Ecology (SKLURE2018-2-6). We greatly appreciate suggestions from anonymous referees for the improvement of our paper. Thanks also to the editorial staff.

Appendix A. Appendix

A.1. Meteorological data

The snowfall, daily precipitation, maximum temperature, and minimum temperature data were provided by the National Climate Center, China Meteorological Administration (CMA) (available from http://www.nmic.gov.cn/). The modern nationwide network of weather observing stations in China began operation in the 1950s. A total of 135 of the 158 stations in the original data set have maintained daily data since 1961. Of these, 23 were excluded because of data quality problems based on a quality control method previously reported (Li et al., 2012). The time span of the meteorological data was mainly from January 1st 1961 to December 31st 2016 in order to ensure that the time period analyzed for all data sets was the same. The distribution of the stations is uneven and very sparse in the western Tibetan plateau owing to the harsh observation environment. Finally, 135 meteorological stations were selected, which were considered to have data of sufficient quality, provide a relatively even spatial distribution, continuous records, and were built earlier than 1961. Detailed information about the stations, which are located at altitudes between 1012 m and 4900 m, is provided in Table.1. These stations belong to the WMO climate data exchange network and each has been allocated a WMO number. In addition, the index for FD, ID, GSL, DTR, R10mm, R20mm, and R25mm were calculated based on a previous study (Li et al., 2012).

Table 1
The selected weather stations in the study region.

WMO number	Name	Latitude	Longitude	Altitude(m)
52,323	Mazongshan	41.80	97.00	1770.4
52,436	Yumenzheng	40.27	97.00	1526
52,674	Yongchang	38.23	101.97	1976.9
52,679	Wuwei	37.92	102.67	1531.5
52,787	Wushaoling	37.20	102.87	3045.1
52,797	Jingtai	37.18	104.05	1630.9
52,889	Lanzhou	36.05	103.88	1517.2
52,983	Yuzhong	35.87	104.15	1874.4
52,984	Linxia	35.58	103.18	1917.2
52,986	Lintao	35.35	103.85	1893.8
52,993	Huining	35.68	105.08	2012.2
52,996	Huajialing	35.38	105.00	2450.6
56,074	Maqu	34.00	102.08	3471.4
56,080	Hezuo	35.00	102.90	2910
56,093	Minxian	34.43	104.02	2315
51,886	Mangya	38.25	90.59	2944.8
52,602	Lenghu	38.75	93.23	2770
52,633	Tuole	38.80	98.24	3367
52,645	Yeniugou	38.42	99.36	8320
52,657	Qilian	38.18	100.25	2787.4
52,707	Xiaozaohuo	36.80	93.47	2767
52,713	Dachaidan	37.85	95.24	3173.2
52,737	Delinha	37.37	97.24	2981.5
52,754	Gangcha	37.33	100.13	4301.5
52,765	Menyuan	37.38	101.62	7850
52,818	Geermu	36.42	94.59	2807.6
52,825	Nuomuhong	36.43	96.24	2790.4
52,836	Dulan	36.30	98.01	3191.1
52,842	Chaka	36.78	99.01	3087.6
52,866	Xining	36.72	101.75	2295.2
52,868	Guide	36.03	101.43	2237.1
52,876	Minhe	36.32	102.85	1813.9
52,908	Wudaoliang	35.22	93.01	4612.2
52,943	Xinghai	35.58	99.60	3323.2
56,004	Tuotuohe	34.22	92.24	4533.1
56,018	Zaduo	32.90	95.13	4066.4
56,021	Qumalai	34.13	95.48	4175
56,029	Yushu	33.02	97.00	3681.2
56,033	Maduo	34.92	98.12	4272.3
56,046	Dari	33.75	99.37	3967.5
56,065	Henan	34.73	101.60	4500
56,067	Jiuzhi	33.43	101.48	3628.5
56,125	Nangqian	32.20	96.25	3643.7
56,151	Banma	32.93	100.75	4530
56,038	Shiqu	32.98	98.01	4200
56,079	Ruoergai	33.58	102.97	3439.6
56,144	Dege	31.80	98.36	4184
56,146	Ganzi	31.62	100.00	3393.5
56,152	Seda	32.28	100.33	3893.9
56,167	Daofu	30.98	101.12	2957.2
56,172	Maerkang	31.90	102.23	2664.4
56,173	Hongyuan	32.80	102.55	3491.6
56,178	Xiaojin	31.00	102.35	2369.2
56,182	Songpan	32.65	103.57	2850.7
56,247	Batang	30.00	99.01	2589.2
56,251	Xinlong	30.93	100.32	4000
56,257	Litang	30.00	100.27	3948.9
56,357	Daocheng	29.05	100.30	3727.7
56,374	Kangding	30.05	101.97	2615.7
56,385	Emeishan	29.52	103.33	3047.4
56,459	Muli	27.93	101.27	2426.5
56,462	Jiulong	29.00	101.50	2987.3
56,475	Yuexi	28.65	102.52	1659.5
56,479	Zhaojue	28.00	102.85	2132.4
56,565	Yanyuan	27.43	101.52	2545
56,571	Xichang	27.90	102.27	1590.9
56,671	Huili	26.65	102.25	1787.3
55,228	Shiquanhe	32.50	80.01	4278.6
55,279	Banyi	31.38	90.00	4700
55,294	Anduo	32.35	91.01	4800
55,299	Naqu	31.48	92.01	4507
55,437	Pulan	30.28	81.13	4900
55,472	Shenzha	30.95	88.36	4672
55,493	Dangxiong	30.48	91.01	4200

(continued on next page)

Table 1 (*continued*)

WMO number	Name	Latitude	Longitude	Altitude(m)
55,578	Shigatse	29.25	88.59	3836
55,591	Lhasa	29.67	91.01	3648.9
55,598	Zedang	29.25	91.48	3551.7
55,655	Nielaer	28.18	85.60	3810
55,664	Dingri	28.63	87.01	4300
55,696	Longzi	28.42	92.25	3860
55,773	Pali	27.73	89.01	4300
56,106	Suoxian	31.88	93.48	4022.8
56,116	Dingqing	31.42	95.36	3873.1
56,137	Changdu	31.15	97.12	3306
56,202	Jiali	30.67	93.13	4488.8
56,227	Bomi	29.87	95.48	2736
56,312	Nyingchi	29.67	94.23	2991.8
56,434	Chaou	28.65	97.25	2327.6
51,186	Qinghe	46.67	90.24	1218.2
51,288	Beitashan	45.37	90.35	1653.7
51,437	Zhaosu	43.15	81.01	1851
51,467	Baluntai	42.73	86.13	1739
51,477	Dabancheng	43.35	88.13	1103.5
51,542	Bayingbuluke	43.03	84.02	2458
51,567	Yanqi	42.08	86.36	1055.3
51,628	Akesu	41.17	80.12	1103.8
51,633	Baicheng	41.78	81.59	1229.2
51,644	Kuche	41.72	82.60	1081.9
51,701	Tuergate	40.52	75.24	3504.4
51,705	Wuqia	39.72	75.13	2175.7
51,709	Kashen	39.47	75.60	1289.4
51,711	Aheqi	40.93	78.25	1984.9
51,716	Bachu	39.80	78.36	1116.5
51,720	Keping	40.50	79.01	1161.8
51,730	Alaer	40.55	81.13	1012.2
51,804	Tashenkuergan	37.77	75.12	3090.1
51,811	Shache	38.43	77.13	1231.2
51,818	Pishan	37.62	78.13	1375.4
51,828	Hetian	37.13	79.59	1375
51,839	Minfeng	37.07	82.47	1409.5
51,855	Qiemo	38.15	85.36	1247.2
51,931	Yutian	36.85	81.37	1422
52,101	Balitang	43.60	93.01	1677.2
52,313	Hongliuhe	41.53	94.47	1573.8
56,533	Gongshan	27.75	98.47	1583.3
56,548	Weixi	27.17	99.13	2326.1
56,586	Zhaotong	27.35	103.72	1949.5
56,651	Lijiang	26.87	100.22	2392.4
56,684	Huize	26.42	103.28	2110.5
56,739	Tengchong	25.02	98.35	1654.6
56,748	Baoshan	25.12	99.12	1652.2
56,751	Dali	25.70	100.18	1990.5
56,768	Chuxiong	25.03	101.55	1824.1
56,778	Kunming	25.00	102.65	1886.5
56,786	Zhanyi	25.58	103.83	1898.7
56,875	Yuxi	24.33	102.55	1716.9
56,543	Zhongyi	27.83	99.47	3276.7
56,886	Luxi	24.53	103.77	1704.3
56,951	Lingcang	23.88	100.08	1502.4
56,444	Deqin	28.48	98.59	3319
52,856	Qiabuqia	36.27	100.62	2835
56,880	Yiliang	24.92	103.17	1532.1
55,680	Jiangzhi	28.92	89.36	4040
51,330	Wenquan	44.97	81.00	1357.8
52,118	Yiwu	43.27	94.47	1728.6

A.2. Glacier data

This study reviewed change in the glaciers in western China based on data sourced from previous studies, as shown in Table 2.

Table 2
Data source of glacier area change from published literature.

Mountains	Times	Data sources
The north of Altai mountains	1980–2010	Bai et al., 2012; Wang et al., 2011a; Yao et al., 2012a; Lv et al., 2012;
The south Altai mountains	1972–2011	Bai et al., 2012; Wang et al., 2011b; Yao et al., 2012b;
The north of Tianshan mountains	1989–2011	Li et al., 2004, 2006; Shangguan et al., 2009; Xu et al., 2011; Wang et al., 2011a;He et al., 2014
The south of Tianshan mountains	1990–2011	Li et al., 2004, 2006; Shangguan et al., 2009; Xu et al., 2011;Wang et al., 2011b; Zhao et al., 2014; Xing et al., 2017;
Tanggula mountains	1973–2010	Zhang et al., 2010a, 2010b; Zhu, 2012;Wang, 2017a; Wang et al., 2017a;
The western Kunlun mountains	1990–2011	Li et al., 1998; Xu et al., 2006;Shangguan et al., 2007, 2009; Zhang et al., 2010a, 2010b; Li, 2014;
The eastern Kunlun mountains	1990–2010	Li et al., 1998; Xu et al., 2006; Shangguan et al., 2007; Shangguan et al., 2009; Zhang et al., 2010a, 2010b; Jiang, 2012;
Nyainqntanglha mountains	1979–2011	Bolch et al., 2010; Zhang et al., 2010a; Zhang et al., 2010b; Wang et al., 2012; Ji et al., 2014, 2015;
Qilian mountains	1990–2010	Liu et al., 2003;Wang et al., 2009;Cao et al., 2010; Zhang et al., 2010a, 2010b; Wang et al., 2011c; Zhang et al., 2012; Pan et al., 2012;
Gongga mountains	1974–2010	Liu et al., 2010; Pan et al., 2011; Zhang et al., 2010a, 2010b; Li, 2015;
Yulong mountains	1974–2009	Wang et al., 2011a; Du, 2011
Middle Himalaya Range	1980–2010	Jiang, 2015
Eastern Himalaya Range	1980–2007	Li et al., 2011a
Pamirs	1972–2011	Shangguan et al., 2004; Shangguan et al., 2007; Shangguan et al., 2009; Zhang et al., 2010a, 2010b; Zeng et al., 2013;
The northern Himalayas	1990–2015	Ji, 2018
Qiangtang plateau	1970–2000	Wang et al., 2011a
Kalakunlun Mountains	1978–2015	Xu, 2017;

A.3. Permafrost data

Based on the previous studies (Wang et al., 2005; Du et al., 2012; Liu et al., 2014), we calculated variations in the maximum freezing depth at 33 sites and the permafrost active layer depth at 8 sites in the studied region in order to explore permafrost degradation.

A.4. Actual evapotranspiration data

In order to analyze variations in AE, we used global land-surface actual evapotranspiration (GLEAM) data from 1980 to 2016 (downloaded from https://www.gleam.eu). GLEAM evapotranspiration data considers the influence of precipitation, surface soil moisture, and vegetation. This product has high precision and resolution, and is now used in most similar studies.

A.5. Moisture recycling data

Moisture recycling refers to the contributions from terrestrial evaporation and transpiration to precipitation (Jasechko et al., 2013; Rios-Entenza et al., 2014). It includes moisture evaporated from the surface of soil and water and moisture from plant transpiration (Davie, 2008), which provides precipitation in local regions, forming part of the scarce water resources in arid regions (Kong et al., 2013). This study reviewed the contribution from moisture recycling to precipitation in cold regions of western China based on data from previous research (Rudi et al., 2010; Xu et al., 2011; Zhang et al., 2014a; Guo and Wang, 2014; Li, 2015;Yao et al., 2016; Wang et al., 2017a; An et al., 2017; Li et al., 2017).

A.6. Runoff data

The data of runoff for 39 rivers in the cold regions of western China were sourced from published studies, as listed in Table 3.

Table 3
Data source of annual runoff variation rate from published literature.

Rivers	Times	Data sources
Kensiwate	1956–2006	Tang et al., 2013
Barqin	1957–2008	Qin et al., 2016
Ulungar	1960–2010	Qin et al., 2016
Bortala	1961–2008	Qiao, 2010
Jinghe	1961–2008	Qin et al., 2016
Kuitun	1965–2003	Ma et al., 2008
Manasi	1956–2006	Tang et al., 2013
Hutubi	1979–2005	Pu et al., 2004
Urumqi	1958–2005	Qin et al., 2016
Kaiken	1960–2009	Tian, 2013
Baiyang	1962–2007	Ren et al., 2010
Yiwu	1977–2007	Gao and Luo, 2009
Tekes	1956–2000	Qin et al., 2016
Kashi	1956–2000	Qin et al., 2016
Kaidu	1956–2005	Qin et al., 2016
Kamushang	1956–2006	Qin et al., 2016
Karasu	1956–2006	Qin et al., 2016
Tailan	1952–2006	Qin et al., 2016
Kunmalike	1957–2008	Qin et al., 2016
Tuoshigan	1957–2008	Qin et al., 2016
Yarkant	1957–2008	Mtalip et al., 2012
Zeller	1959–2005	Qin et al., 2016
Karash	1957–2008	Mtalip et al., 2012
kushan	1956–2005	Qin et al., 2016
Keriya	1957–2009	Ling et al., 2012
Cherchen	1956–2006	Aynur et al., 2010
Golmud	1961–2006	Wang et al., 2007
Yellow	1952–2008	Wang and Hu, 2011
Daxiahe	1953–2008	Qin et al., 2016
Taohe	1952–2008	Qin et al., 2016
Huangshui	1952–2008	Qin et al., 2016
Datong	1952–2008	Qin et al., 2016
Shulehe	1961–2010	Zhang et al., 2014a
Heihe	1961–2010	Qin et al., 2016
zamuhe	1961–2010	Qin et al., 2016
Buhahe	1961–2007	Cui et al., 2011
Yangtze river	1956–2012	Kong et al., 2016
Langcang	1956–2012	Kong et al., 2016
Yarlung Tsangpo River	1960–2009	Li et al., 2015

A.7. Lake data

We reviewed changes in the areas and water levels for 78 big lakes (area $> 10\,km^2$) in the study region. The data was mainly sourced from the research by Wang (2017a), Zhu et al. (2015), and Wu et al. (2017).

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