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Review

Permafrost dynamic change on the Tibetan Plateau under climatic warming since 1950s

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Abstract: The Tibetan Plateau is situated in one of Earth's extreme continental climate settings and is influenced by numerous climatic regimes such as the East Asian and Indian monsoons and westerlies. It is the largest region of the world where permafrost is present at mid-latitudes and its relative warmth and shallow thickness makes it more sensitive to climatic warming than the arctic region. Changes in permafrost temperature, extent and active layer thickness have been observed for decades. In this paper we review these changes in this dynamic permafrost region, especially noting the changes in permafrost distribution and area loss of about 99,000 km^2 per decade, increases in active layer thickness of 4-8 cm/year, and increases in thawing days of about 10-16 days/decade. As soil temperature and soil water content have changed, river run-off has decreased over most of the year. Warmer temperatures may have transformed the Tibetan Plateau from a net carbon source in the 20th century to a net sink as the increase rate of net primary production is faster than that of soil respiration.

Keywords: permafrost, Tibetan Plateau, climate change

INTRODUCTION

Permafrost, or perennially frozen soil, is soil that has a temperature lower than 0^{o}C continuously for at least 24 consecutive months [1-3]. Permafrost thickness, occurrence and geographic extent react sensitively to changes in surface temperature. Recent climatic warming has resulted in an almost ubiquitous increase in ground temperature globally and the resulting permafrost degradation has led to significant changes, at least locally, in soil moisture content, soil nutrient availability, ecology, and soil carbon cycle [4].

The Tibetan Plateau (TP) is the world's highest elevated region, averaging 4500 m above sea level. It is located between 74–105°E and 25–40°N and covers an area of approximately 2,545,000 km^2. The TP plays a crucial role in the provision of water resources to most of the Asian continent, both directly via the large rivers that originate there and also via the temperature contrast between TP and the Indian Ocean, which is an important controlling factor for both the Asian monsoon and global atmospheric circulation patterns [5]. Permafrost in the TP mainly occurs in the plateau's interior between the Kunlun Mountains and the Tanggula Mountains, with Xidatan to the north of the Kunlun as its northern limit and Anduo to the south of the Tanggula as its southern limit. At present, the permafrost area is about 1.27×10^6 km^2 in the TP, accounting for about 50% of its total area [6]. TP permafrost can be found both as discontinuous in the northern part and patch in the southern part of the Plateau (Figure 1). Direct mapping of the permafrost has been very limited to date due to the remote and harsh environment, especially in the central and western regions of the TP. Geographic information system and remote sensing technologies have been the main tools for monitoring the spatial distribution of permafrost [7], and maps are made using statistical fits to the sparse observation network. However, classification and mapping of the TP permafrost has not been consistent across the different studies [8]. Considerable uncertainty exists in permafrost maps that have been made at different times over the last 50 years. Currently, the 1 : 4,000,000 Map of the Glaciers, Frozen Ground and Deserts in China [9] is the most reliable. This map is based on multiple linear regression between latitude, altitude and mean ground temperature using a network of 76 borehole data stations across the TP [10]. Nan et al. [11] suggested that the permafrost was overestimated in western TP in Wang et al.'s map [9], although this is a much better map than the International Permafrost Association's map [12].

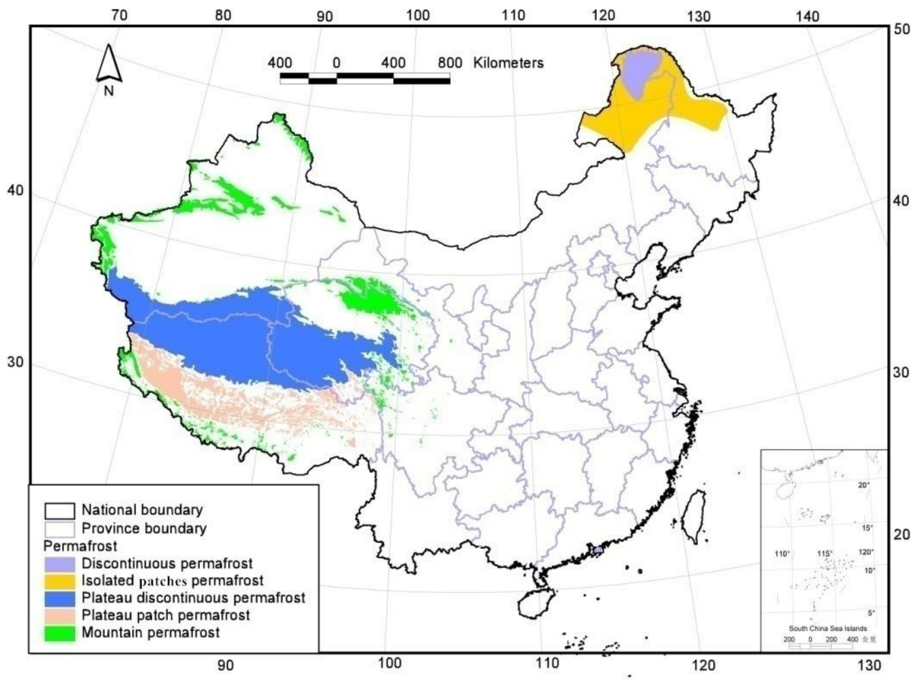


Figure 1. Permafrost distribution in China [9]

The permafrost on the TP is considered to be more sensitive to climatic warming than that in the arctic region: Tibetan permafrost is relatively warm and thin, mostly warmer than -2.0°C and less than 100 m in thickness [13]. As a result, it is highly sensitive to temperature changes, and significant warming, thawing, thinning and retreat of the permafrost have been reported throughout the plateau in recent decades [7, 14, 15]. This sensitivity has significant influences on regional water balance, biological diversity, carbon cycle and engineering construction in the TP [16]. In this paper we assess several key aspects in the changes of permafrost regions of the TP, including permafrost distribution and degradation, permafrost temperature, changes in active layer thickness, soil water content and surface run-off, and carbon emissions and sink.

PERMAFROST DEGRADATION

Significant permafrost degradation has occurred on the TP over recent decades [7]. The process of permafrost degradation can be divided into five stages, viz. starting stage, rising temperature stage, zero geothermal gradient stage, talik layers stage and disappearing stage. Permafrost in the high and middle mountains is in the rising temperature stage while that in the middle range and lower mountains is in the transition from the late rising temperature stage to zero geothermal gradient stage. The zero geothermal gradient stage is seen on high plateaus and valleys. The transition from zero geothermal gradient stage to the talik layers stage occurs in patchy permafrost area, and permafrost is disappearing from the margins [17].

Increasing ground temperatures since the 1970s have led to thinning of the permafrost and complete thawing of thin permafrost [18]. Li et al. [19] used a one-dimensional heat equation to simulate permafrost degradation near Maduo in eastern TP. Using a prescribed increase rate of the mean annual air temperature of 4 ^{o}C per century, the simulation results indicated that after 110 years, the base of permafrost rises from the initial value of 55.0 m to 15.2 m. The projected rate of basal permafrost degradation in low moisture content bedrock reaches a peak of 0.58 m/year after 60 years, assuming a constant thermal gradient at the base of 0.018 ^{o}C/m. This implies that the permafrost has a high degree of persistence under the applied warming scenario and thins mostly from the base and much less from the surface due to the high latent heat content of the near-surface ice-rich permafrost layers [19]. In addition to the temperature impact, changes in radiation balance due to recent wintertime increases in cloud amount at low elevation locations and decreases at higher elevation locations also contribute to permafrost degradation [20]. Li et al. [21] investigated permafrost distribution along a transportation corridor near Xining city (Figure 2). They found that elevation, potential direct solar radiation in June and topographic wetness index are important factors affecting the presence and distribution of the permafrost.

Since permafrost on the TP is characterised by high ground temperature and high ice content, its degradation may result in the formation of thermokarst lakes. Niu et al. [22] investigated thermokarst lakes between the Kunlun Mountain Pass and the Fenghuo Mountain Pass along the Qing–Tibet Railway (Figure 2). Over 250 lakes are distributed within a 200-m-wide band along both sides of the railway, covering an area of 1,390,000 m^2. The largest lake is about 60,000 m^2, while the smallest is smaller than 200 m^2, and the mean area is 5,580 m^2. The distribution of the thermokarst lakes is closely related to the ice content and the permafrost temperature, 83.8% of the lakes being in ice-rich permafrost regions and 54.9% in high-temperature permafrost regions.

Rise of Lower-Altitude Limit of Permafrost

Increasing ground temperatures since the 1970s have resulted in an increase in the 'island permafrost' boundary [18, 23]. Cheng [24] gave an empirical formula for the statistical relation between the lower limit of permafrost (H) and latitude (φ), expressed as a Gaussian distribution function:

$$H=3650 \exp[-0.003(\varphi-25.37)^2]+1428.$$

Maejo Int. J. Sci. Technol. **2016**, *10*(03), 242-255

Figure 2. Location of main permafrost observation stations on the TP

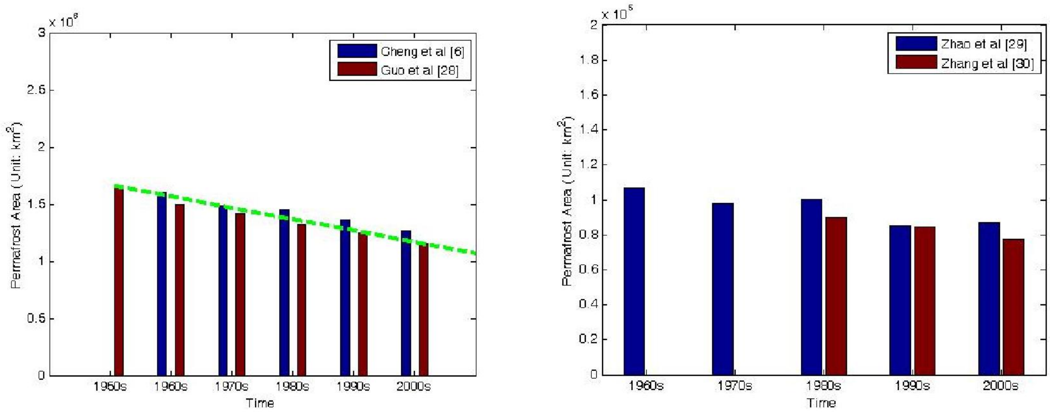
Based on it, Li et al. [25] further evaluated the integrated effect of latitude and altitude on permafrost distribution on the TP. Global warming has led to a rise of 40-80 m in the lower-altitude limit of permafrost on the TP since 1970s [14]. Cheng and Wu [13] further showed that the lower-altitude limit of the permafrost has moved up by 25 m in the north during the last 30 years and between 50-80 m in the south over the last 20 years. In addition, Fukui et al. [26] investigated changes in the lower limit of permafrost in the Khumbu Himal of the Nepal Himalayas, which is located in the south-west TP. In 1973 the permafrost lower limit was estimated to be 5200–5300 m above sea level on the southern slopes in this region while it was 5400–5500 m in 2004. So the permafrost lower limit has risen 100–300 m in the Khumbu Himal of the Nepal Himalayas during 1973-2004.

Shrinkage of Permafrost Area

Over the whole TP, the areal extent of permafrost has been reduced significantly during the past few decades [14]. Along the Qing-Tibet Highway (Figure 2), the permafrost boundary has moved 12 km northwards at the southern lower limit, whereas it has moved 3 km southwards at the northern lower limit. In Maduo county the horizontal change of permafrost zones is 15 km [14]. In Xidatan region Wu et al. [27] used 50-MHz ground-penetrating radar to detect the change of the northern boundary of discontinuous permafrost of the TP. Compared with observation data from 1975, Xidatan permafrost was reduced in area by 12%. For the whole TP, using elevation data, mean annual air temperature and the vertical lapse rate of temperature, Cheng et al. [6] estimated that the permafrost area decreased over the decades from the 1960s to 2000s (Figure 3a). The total area of degraded permafrost was about 3.3×10^5 km^2, which accounts for about one-fifth of the total area of permafrost in the 1960s. Guo et al.'s simulated results [28] (Figure 3a) showed that the permafrost area of the TP exhibits a significant decreasing linear trend, with a rate of decrease of

99000 km^2 per decade. Nan et al. [11] suggested that for an air temperature rise of 0.02°C/year, permafrost area on the TP would shrink 8.8-13.5% after 50 years and about 46% after 100 years. Based on outputs from Community Land Model version 4, Guo et al. [28] estimated that the permafrost area would decrease by approximately 39% by the mid-21st century and approximately 81% by the end of the 21st century.

Qilian Mountains (Figure 2) are located in the north-eastern part of the TP and comprise one of the most important alpine permafrost regions of the TP. Zhao et al. [29] used a logistic regression model to simulate the decadal change in the distributions of alpine permafrost from 1960 to 2009 in the Qilian Mountains (Figure 3b). The area of alpine permafrost shows an overall decrease. Later, based on the Map of Snow, Ice and Frozen Ground, topographic and meteorological factors, Zhang et al. [30] used a multi-criteria approach to simulate permafrost distributions in the same region (Figure 3b), and also presented an overall decreasing trend of about 17% (i.e. a decrease of about 15600 km^2) over the past 30 years. Extrapolating the climate change trend into the future suggests that the Qilian permafrost will continue to degrade by about 5100 km^2 over the next 15 years.



(a) (b)

Figure 3. The change in permafrost area of (a) the TP and (b) the Qilian Mountains (north-east TP) from 1960s to 2000s [6, 28-30] with a rate of decrease of 99,000 km^2 per decade [28] for the TP

INCREASE IN PERMAFROST TEMPERATURE

On the TP the mean annual ground surface temperatures are highly variable and are related to the mean annual air temperature, but are also influenced by land type, vegetation, snow cover and summer monsoons. During 1967-1997 the warmest ground surface temperatures occurred in north-west TP, with annual, cold- and warm-season temperatures of about 14.3, 1.7 and 26.9 ^{o}C respectively [31]. In Maduo County (Figure 2) Xue et al. [32] showed that the annual ground surface temperature increased at an average rate of 0.6°C per decade between 1980 and 2005. Increasing soil temperature also caused a 60-day increase in the number of surface thawing days between 1983 and 2001, when most exchange of energy, moisture and gases occur. Wu et al. [33] analysed the ground surface temperature records from 16 meteorological stations which were located in or adjacent to permafrost regions on the central TP. They also revealed a statistically significant warming. The mean annual ground surface temperature increased there at an average rate of 0.60°C per decade over the period of 1980–2007. The surface freezing index has decreased at a rate of 111.2 °C d/decade and the surface thawing index has increased at a rate of 125.0 °C d/decade on the central TP, which means a significant weakening of seasonal frost penetration and a great thickening of active layer, especially prominent in the source regions of Yangtze River.

Permafrost temperatures are closely related to ground temperatures and air temperatures [33]. The mean annual permafrost temperatures at the depth of 1 m have increased by about 0.12–1.65°C with average 0.91°C from 1996 to 2006 in permafrost regions of the TP [34]. Wu and Liu [35] used seven sites along the Qing–Tibet Highway between Golmu and Tuotuo River Pass (Figure 2, Table 1). Their data indicated that the maximal trend of annual permafrost temperature (more than 0.05°C/year at the depth of 6 or 8 m) occurred in the high mountain permafrost areas of the TP and the minimal trend occurred in the high plain permafrost. Wu and Zhang [34] further showed that long-term mean annual permafrost temperatures at 6-m depth varied from -0.19°C at the Touerjiu Mountains site near Ando to -3.43°C at Fenghuo Mountain Pass, with an average of about -1.55°C. Mean annual permafrost temperatures at 6-m depth increased 0.12-0.67°C with an average increase of about 0.43°C between 1996 and 2006.

Table 1. Trend of annual ground surface and permafrost temperature along Qing-Tibet Railway/Highway [33-36]

Location	Time	Soil depth	Trend (°C/year)
Tuotuohe Pass	1980-2007	Ground surface	0.079
Ando	1980-2007	Ground surface	0.062
Lhasa	1980-2007	Ground surface	0.062
Fenghuoshan Pass	1996-2006	1 m	0.075
Wudaoliang	1996-2006	1 m	0.091
Kunlun Mountain Pass	1995-2002	1 m	0.140
Tanggula Mountain Pass	1998-2006	1 m	-0.016
Wudaoliang	1996-2001	6 m	0.021
Wudaoliang	1996-2006	6 m	0.061
Fenghuoshan Pass	1996-2001	6 m	0.031
Fenghuoshan Pass	1996-2006	6 m	0.059
Kunlun Mountain Pass	1996-2001	6 m	0.048
Kunlun Mountain Pass	1995-2002	6 m	0.059
Tanggula Mountain Pass	1998-2006	6 m	-0.017
Beiluhe	2002-2012	6 m	0.0013 (desert grasslands) 0.0181 (alpine ecosystems)
66 Daoban	1996-2001	8 m	0.022
Beiluhe	2002-2012	10 m	0.0085 (desert grasslands) 0.0161 (alpine ecosystems)

Different land types can affect permafrost temperature. Xie et al. [37] studied the effects of desertification on the permafrost temperature. The annual ground temperature below the permafrost table overlain by a thick sand layer was lower than that of the natural ground surface, and the temperature drop was roughly constant at -0.2°C. Changes in permafrost temperature below a sand layer can be attributed to differences in albedo and heat conductivity between the sand layer and natural ground surface of the TP. When desertification occurs on the TP, the reflectivity of the sand surface can reach 0.25 to 0.4 while the reflectivity of a bare surface is only 0.1 to 0.25. Later on, in the Beiluhe region of the TP, Wu et al. [36] showed that from 2002 to 2012, the rate of increase in permafrost temperature at a depth of 6 m varied from 0.013°C/decade for desert grasslands to 0.181°C/decade for alpine ecosystems, and at a depth of 10 m, the rate varied from 0.085°C/decade for desert grasslands to 0.161°C/decade for alpine ecosystems. Moreover, permafrost temperature also has a significant variation for different amounts of vegetation cover. For instance, Wang et al. [38] investigated the source region of Yangtze River (near Tuotuo River Pass) between 2005-2009 and found that the decline in vegetation cover in alpine meadows resulted in an increase in the soil-thawing temperature and moisture, a decrease in the soil-freezing temperature and moisture, and an advance in the onset of seasonal changes in the soil temperature.

Over the whole TP, permafrost temperatures at 15-m depth are strongly controlled by elevation and latitude. Based on data from 190 boreholes along the Qing-Tibet Highway/Railway since the early 1960s, Wu et al. [39] gave an empirical formula by linear regression:

$$T_{15} = 52.78 - 0.79\varphi - 0.57H_{100}$$

where T_{15} is permafrost temperature (°C) at 15-m depth, φ is latitude (°) and H_{100} is altitude (in hectometres). This means that permafrost temperature at 15-m depth decrease at a rate of 0.57°C per 100-m altitude increase and 0.79°C per degree latitude moving north along the Qing-Tibet Railway.

Gao et al. [40] used wavelet methods to analyse the permafrost temperature data along the Qing-Tibet Highway over the period 1996-2012. The permafrost temperature varied on a longer time scale as depth increased. The wavelet power spectra of permafrost temperature displayed preferred time scales of about 6, 12, 22-29, 34-45, 60-81 and 102-108 month periods. Gao et al. [40] did not suggest causes for these dominant variations, but the obvious 5-7 years coincide with the variation of the North Atlantic Oscillation and variability in Westerlies.

DEEPENING OF ACTIVE LAYER

The active layer is the layer above the permafrost that thaws during the summer and freezes in winter. Active layer thickness (ALT) in the northern and western regions of the TP is larger than that in the eastern and southern regions [41]. In particular, along the Qing-Tibet Highway, between 1995 and 2007, the long-term and spatially averaged ALT was about 2.41 m with a range of 1.32–4.57 m [17]. The thickness of the active layer depends on summer (ground surface) temperature, soil texture (i.e. thermal properties of the ground), vegetation cover, soil moisture content, snow cover, etc. [42]. The mean increase rate of ALT was approximately 7.5 cm/year on the TP over the period 1995–2007 [17].

Soil Freeze–Thaw Cycle

Freezing and thawing processes at the soil surface play an important role in determining the ALT and its interactions with the atmosphere. The trend in the soil freeze–thaw cycle is positively correlated with climatic warming over the TP. Guo et al. [43-44] investigated land surface water and heat exchanges under different freezing and thawing conditions and showed that the diurnal cycle of unfrozen soil water, resulting from diurnal freeze/thaw cycles at the surface, has a significant impact on latent heat flux. As already noted, Xue et al. [32] found a 60-day lengthening of surface thawing days from 1983 to 2001. Later on, Li et al. [45] found a trend toward earlier onset date of soil thaw by approximately 14 days and later onset date of soil freeze by approximately 10 days over the period 1988–2007, and that the number of frozen days decreased over the TP by 16.8 days per decade. The most significant changes occurred in the north-eastern and south-western TP, while almost no change occurred in the north-western TP. Guo et al. [46] estimated that from 1981 to 2010, the freeze durations of the TP were shortened by 9.7 and 8.6 days/decade for permafrost and seasonally frozen ground respectively. In addition, Zhou et al. [47] investigated the impact of snow cover on soil freeze–thaw cycle. The soil freezes deeper for snow depths less than 20 cm, compared with the situation without snow cover. Shallow snow cover can increase albedo and then reduce the absorption of solar radiation and decrease soil temperature. However, this positive effect is lost with increasing precipitation and snow depth in winter.

Past Change of ALT

The depth of the active layer on the TP has increased progressively over the past several decades, which has directly affected the terrestrial hydrological cycle, ecosystems and climate feedback. The increase in ALT is primarily due to increasing summer air temperature, whereas changes in the winter air temperature and snow cover condition play no, or only a very limited, role [17]. Wu and Liu [35] set up seven observation sites along the Qing–Tibet Highway between Golmu City and Tuotuo River Pass (Figure 2). The data from all sites indicated that the mean annual deepening of the ALT was about 4.0–8.4 cm/year for high-mountain areas, 0.8–6.5 cm/year for high-plain areas, and 3–5 cm/year for mid-lower mountain areas from 1996–2001. In Beiluhe region of the TP, Wu et al. [36] showed that from 2002 to 2012, the average increase in ALT was about 4.26 cm/year. In addition, for Maduo County, the depth of the seasonal thawing layer increased by 54 cm, 102 cm and 77 cm in April, May and June respectively, from 1983 to 2003 [32].

Projections of Future ALT

Many researchers have explored various methods to simulate and predict the change in ALT. Pang et al. [31] calculated the seasonal freezing and thawing depth in the frozen ground regions over the TP, and found that ALT in the northern and western regions were greater than that in the eastern and southern regions. ALT in the permafrost regions along the Qing–Tibet Highway were generally more than 2.0 m, and the regions with ALT less than 2.0 m were mainly in the high mountain areas such as the Kunlun Mountains, the Fenghuo Mountains and the Tanggula Mountains. Li et al. [19] used the one-dimensional heat equation to simulate permafrost degradation near Maduo in eastern TP. Using a prescribed rate of increase in the mean annual air temperature of 4^{o}C per century, the simulation results indicated that after 110 years, the ALT would increase from a depth of 1.20 m to 1.48 m. When forcing with the HADCM3 model, Pang et al. [48] found that ALT would increase with rising air temperature over the 21st century. The average increase in ALT expected under modest temperature forcing from the B1 scenario was 0.4 m and an increase of 1.2 m was obtained under the more extreme A1F1 scenario during the 21st century. ALT change is very significant in sporadic permafrost regions while it is relatively small in the colder permafrost regions of the inland plateau. Similarly, Guo et al. [28] showed that under the A1B scenario the ALT of 0.5–1.5 m found at present would increase to approximately 1.5–2.0 m by 2030–2050. This increase would continue and reach a level of 2.0–3.5 m by 2080–2100. In addition, using the CoupModel, Zhou et al. [47] calculated an increase in maximum thawing depth from 1.5 m to about 3.5 m as a result of a 4°C warming and predicted a talik formation at the top of the permafrost as a result of a 6°C warming. Projections using a 25-km regional climate model forced by an Earth System Model under the A1B scenario suggested an annual warming of 4°C for the TP by 2100 [49].

Maejo Int. J. Sci. Technol. **2016**, *10*(03), 242-255

SOIL WATER CONTENT AND SURFACE RUN-OFF

Permafrost of the TP influences hydrology by providing an impermeable barrier to the movement of liquid water. In the TP the soil water regime tends to be greatly controlled by the active layer's thawing and freezing processes. Thawing can remove barriers to groundwater flow, which may decrease near-surface soil water storage [50-52]. In the past decades the near-surface soil layer in the TP has been found to become drier. These changes, in turn, have inhibited the growth of alpine meadow vegetation that has shallow root systems [32], although seasonal precipitation also controls water storage in the surface soil layer [16, 53].

Permafrost dynamic changes in the TP are one of the main causes of a lowering groundwater table at river-source regions, which in turn results in lowering lake water levels, drying swamps and shrinking grasslands [13]. In general this process is driven by soil temperature, soil moisture, air temperature, precipitation, etc., whereas soil moisture and soil temperature are the main contributors to run-off processes. Wang et al. [52] focused on the Yangtze River source regions (Figure 2) and investigated the influence of freeze–thaw cycles of active layer on surface run-off. Regardless of the level of flood run-off in spring or summer, the active layer thawing is the most important factor influencing the run-off while the precipitation has no significant impact. Niu et al. [22] investigated the Yellow River source regions (Figure 2), which had 43% permafrost coverage. During the period 1956–2007, the monthly run-off of the Yellow River had a negative trend except in June, which increased by 2.9% per decade. The run-off decreased by about 4% per decade from July to November and 1.0-4.0% per decade for December to May. In contrast to the Yellow River, Liu et al. [54] found a significant upward trend of run-off in winter (since 1985) but not in summer for the upper Lhasa River located in Lhasa region (Figure 2).

CARBON EMISSIONS AND SINKS

Permafrost on the TP is a huge carbon pool, but soil organic carbon stored in the permafrost is vulnerable in an increasingly warmer climate. In the TP, permafrost degradation has resulted in decreases in the soil organic carbon content of alpine meadows and alpine steppes [15, 55]. Based on changes of grassland vegetative cover between 1986 and 2000, soil organic carbon loss from the top 0.3-m soil layer of the permafrost region of the TP is about 120 gigagrams C/year [56]. Mu et al. [57] investigated Eboling Mountain in the central Qilian Mountains and indicated that the potential methane emission from soils near the permafrost table and deep permafrost (4.14–4.48 m) is closely related to climatic warming. In addition, carbon emissions from thermokarst lakes in the permafrost region were investigated in six lakes along the Qing-Tibet Highway [58]. The concentration of CO_2 and CH_4 varied greatly, with bubbles having CO_2 concentrations of 0.22-77.62% and CH_4 concentrations of 0.001-1.96%. Isotopic analysis of CO_2 and CH_4 suggested that CO_2 was primarily from decomposed lacustrine carbonates and oxidised organic mass, while CH_4 primarily originated from old sedimentary organic matter.

One important offset of carbon emissions from permafrost is that increased soil temperature and improved nutrient availability during permafrost degradation can increase rates of photosynthesis and also prolong the growing season, both of which would lead to greater sequestration of carbon from the atmosphere [15, 55, 59]. This provides a negative feedback, to some extent, on climate warming. Using the Terrestrial Ecosystem Model, Zhuang et al. [60] suggested that during the 20^{t}h century, the TP had changed from a neutral or small carbon source during the early part of the century to a sink later, with large inter-annual and spatial variability due to climate change and permafrost conditions. The increase rate of net primary production was faster than that of soil respiration, and the carbon sink in the whole TP was 36 teragrams C/year by the end of the 20th century.

DISCUSSION AND CONCLUSIONS

The TP, where permafrost is present at mid-latitudes, is more sensitive to climatic warming than the arctic region at high latitudes. Increasing air temperatures over the TP have resulted in a rise in soil temperatures, deeper active layer and degradation of the permafrost. The increasing ALT is an immediate indicator of warming while the retreat of the permafrost requires a much longer time. The extensive permafrost degradation on the TP exerts profound influences on the land surface energy and moisture balance, hydrological cycle, carbon exchange between the atmosphere and the earth surface, as well as the ecosystem dynamics [13]. Permafrost degradation increases the seasonal thawing depth and ground temperature, and decreases water content and floristic number [61-62]. A potential approach to mitigating permafrost degradation is the application of geoengineering schemes which involves intentional large-scale manipulation of the environment to reduce undesired anthropogenic climate change. We are carrying out related research in this direction.

Permafrost on the TP receives much less attention than that in the Arctic. To date, most permafrost research on the TP deals with applied aspects in connection with major infrastructure projects. As a consequence, permafrost monitoring sites are located mainly along the Qing-Tibet Railway/Highway except for some sites near the Qilian Mountains. Monitoring of borehole temperatures has demonstrated the warming of permafrost, deepening of the active layer and an extensive permafrost degradation on the TP during the last several decades. However, the current distribution of permafrost monitoring sites certainly does not provide a fully representative coverage in the scientific context. Since the relationships between climatic change, permafrost and environment are complicated, more and better monitoring sites and more long-term monitoring are urgently needed for a better understanding of these interrelations. Meanwhile, when observational data are limited, using an integrated approach of regional modelling, high-resolution remote sensing and geographic information system to simulate and project permafrost dynamic changes in the TP is suggested.

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