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Past, present, and future geo-biosphere interactions on the Tibetan Plateau and implications for permafrost

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ABSTRACT

Interactions between the atmosphere, biosphere, cryosphere, hydrosphere, and geosphere are most active in the critical zone, a region extending from the tops of trees to the top of unweathered bedrock. Changes in one or more of these spheres can result in a cascade of changes throughout the system in ways that are often poorly

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understood. Here we investigate how past and present climate change have impacted permafrost, hydrology, and ecosystems on the Tibetan Plateau. We do this by compiling existing climate, hydrologic, cryosphere, biosphere, and geologic studies documenting change over decadal to glacial-interglacial timescales and longer. Our emphasis is on showing present-day trends in environmental change and how plateau ecosystems have largely flourished under warmer and wetter periods in the geologic past. We identify two future pathways that could lead to either a favorable greening or unfavorable degradation and desiccation of plateau ecosystems. Both paths are plausible given the available evidence. We contend that the key to which pathway future generations experience lies in what, if any, human intervention measures are implemented. We conclude with suggested management strategies that can be implemented to facilitate a future greening of the Tibetan Plateau.

1. Introduction

The high mountains surrounding the Tibetan Plateau (TP) provide water resources for >1.4 billion people living on and downstream (Immerzeel et al., 2010; Immerzeel et al, 2020; Cheng and Jin, 2013; Wester et al., 2019). The TP encompasses an area of ~2.5 million km^2 (Zhang et al., 2014) and is often referred to as the Third Pole because of its extensive cryosphere (e.g., Yao et al., 2012, 2019; Thompson et al., 2018). The cryosphere of the TP includes abundant glaciers and the largest area of mid-to low-latitude permafrost, including ~1 million km^2 of permafrost and ~ 1.46 million km^2 of seasonally frozen ground (Fig. 1) (Cheng and Wu, 2007; Zou et al., 2017; Gruber et al., 2017; Bolch et al., 2019; Wester et al., 2019). Previous studies have documented the sensitivity of permafrost on the TP to climate change (Kang et al., 2010; Yang et al., 2010, Wester et al., 2019; Biskaborn et al., 2019; Devoie et al., 2019). Due to rising temperatures, TP permafrost is currently thawing (e.g., Lu et al., 2017; Bolch et al., 2019). This thawing results in its degradation, as evidenced by an increase in the thickness of the active layer, retrogressive thaw slumps, gullies, and active layer detachments (Mu et al., 2020). Thawing of permafrost and melting of glaciers in the Himalayas and Tibet has been referred to as one of the most significant environmental impacts of our time (Ali et al., 2015; Gruber et al., 2017; Bolch et al., 2019) due to the potential for massive degassing of CO_2 and CH_4 to the atmosphere, loss of soil organic carbon (Bosch et al., 2017), and landscape degradation (Wang et al., 2011; Li et al., 2015a). Indeed, permafrost on the TP (Fig. 1) and elsewhere is one component of the Earth system facing a potential tipping point (e.g., Lenton et al., 2008, 2019; Slater and Lawrence, 2013; MacDougall et al., 2012; Schaefer et al., 2014; McGuire et al., 2018; Rosier et al., 2021; Boettner et al., 2021; Brovkin et al., 2021). However, the previous adverse outcomes of global warming are also offset by more favorable developments, such as the enhanced biomass and biosphere expansion across thawing permafrost, including increased biodiversity (Zhu et al., 2016; McGuire et al., 2018; Zhong et al., 2019). Thus, the future of environmental change on the TP has both potential negative (permafrost thawing and degradation, glacier retreat) and positive (biosphere

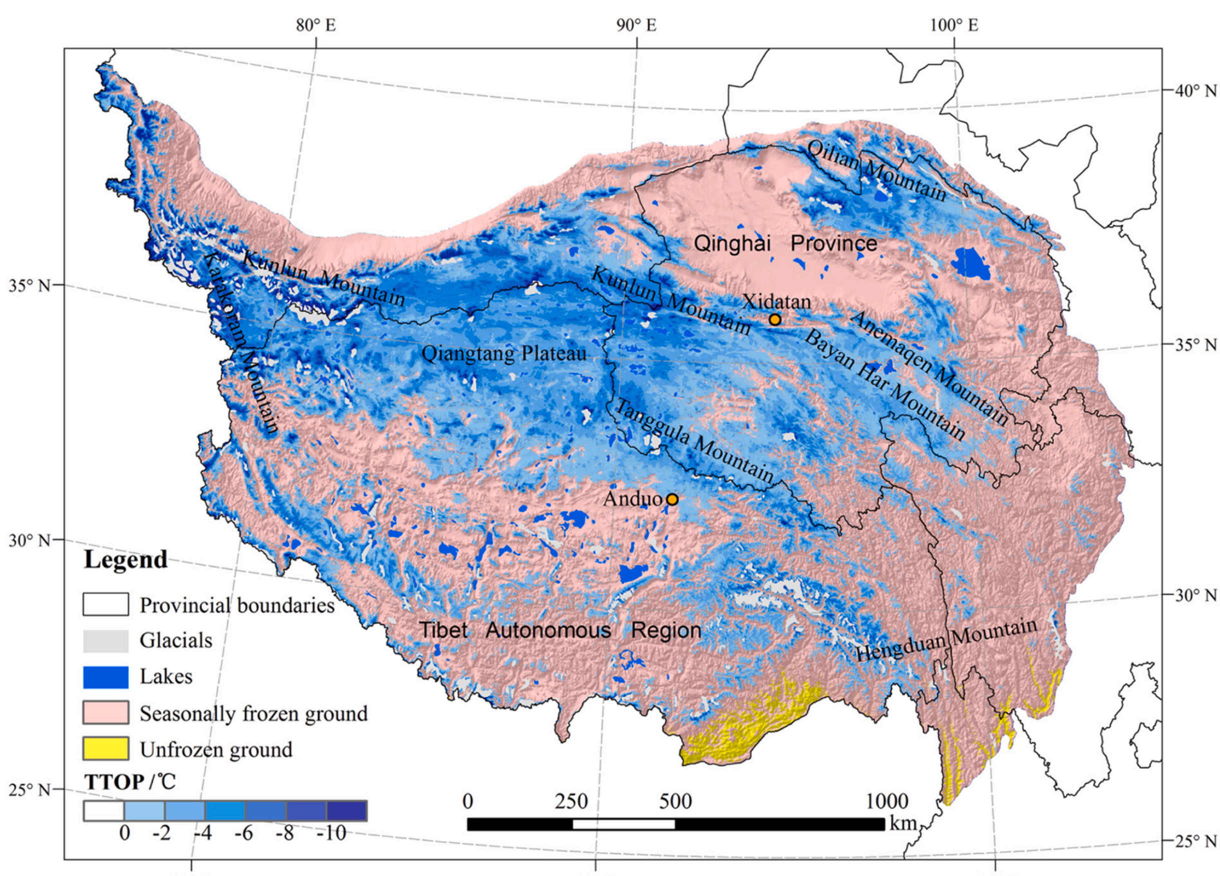

Fig. 1. Map of present-day (as of 2017) permafrost and seasonally frozen ground on the Tibetan Plateau. Blue shaded regions are permafrost, where TTOP represents the simulated mean annual temperature at the top of the permafrost. Other colors are indicated in the legend. Reproduced from Zou et al., 2017 (reproduced under Creative Commons Attribution 3 license). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

expansion) outcomes. The key to understanding the future outcomes of these paths lies in knowing how specific regions of the TP will respond to the combined effects of regional warming, changes in the water cycle, and possible landscape management strategies for sustainable use of natural resources.

Previous studies have documented how climate change has caused unfavorable and favorable impacts on the TP cryosphere and biosphere. For example, previous work indicates adverse effects to TP permafrost thawing over decadal and longer timescales. More specifically, a decrease in the area of permanent permafrost has been observed starting from at least the 1960s and is predicted to continue until at least 2100 (Nan et al., 2005; Lu et al., 2017; Bolch et al., 2019; Wester et al., 2019). Decreasing permafrost has been associated with an increased thaw depth of soil (i.e., the water saturated active layer) above the permanently frozen ground during summer (Wu and Zhang, 2010; Zhang et al., 2012). Of particular note are future projections for an enhanced rate of active layer thickening from 2000 to 2100 (Zhao and Wu, 2019; Cheng and Wu, 2007; Chen et al., 2015). These changes are driven by a warming of 0.3–0.4 °C per decade over the last 50 years (Chen et al., 2015) and are expected to continue with projected warming of 1.8 °C to 4.1 °C by 2100 (e.g., RCP2.6 to 8.5 of IPCC AR5, Su et al., 2013).

The previous changes in permafrost and active layer thickening are in contrast to some favorable impacts of global change on the TP biosphere. Present-day TP ecosystems (Fig. 2) are characterized mainly by Alpine Steppe and Kobresia Pastures (Zhong et al., 2019). These Kobresia pastures are pervasive (~450,000 km^2) across the flats of the TP and have biogeochemical linkages to the Earth system with large organic carbon and nutrient (N, P) stocks in the subsurface (Miehe et al., 2019). A demonstrated (positive) trend in greening across the plateau has been identified from 1982 to 2014 using remote sensing (NDVI) observations (Shen et al., 2015; Zhu et al., 2016; Zhong et al., 2019). However, increases in aboveground biomass are not necessarily linked with an increase in below-ground C allocation but may also be offset by a reduction in belowground C stocks (Hafner et al., 2012). Finally, increases in atmospheric CO_2 have also facilitated a greening of the plateau (Piao et al., 2012). Thus, the warming, moistening, and permafrost degradation across the TP over the last decades have led to an increase in total above-ground biomass. In particular, thawed permafrost and warmer and wetter conditions in a high CO_2 world provide a supportive environment for the continued greening of the TP.

However, two critical components of TP geo-biosphere interactions are currently missing from our understanding. First, integrated scenarios for the future of the entire TP are needed that encompass both the biosphere dynamics and the regional human impact. Second, if different trajectories can be identified, management recommendations for mitigating adverse outcomes are required. For example, although recent work suggests that favorable biosphere conditions (greening) are emerging across the TP today, these conditions may be short-lived. Due

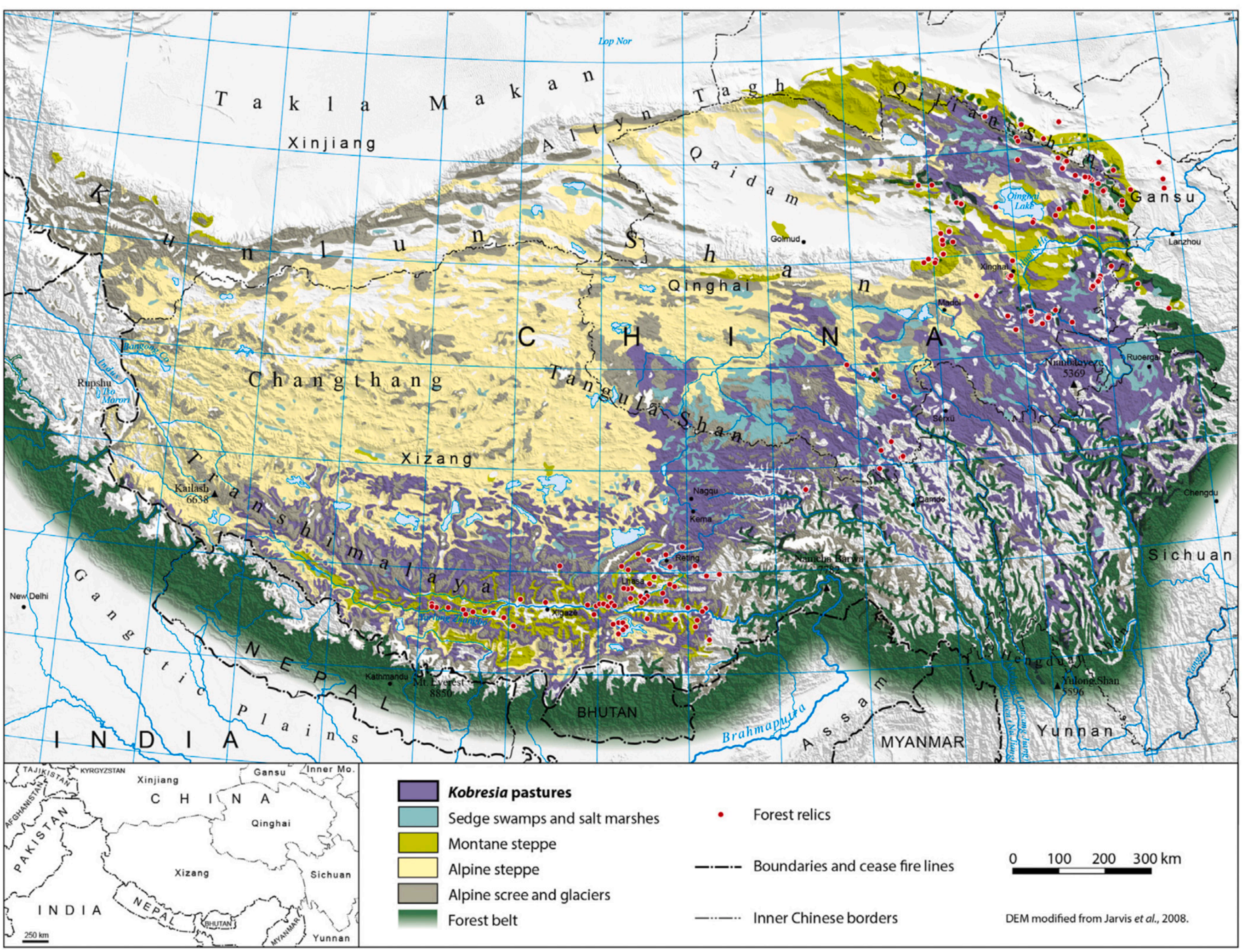

Fig. 2. Land cover of the Tibetan Plateau, including Kobresia pastures (purple) and forest relics. From Miehe et al. (2019) (reproduced under creative commons license 4.0). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

to altered geo-biosphere interactions, they may not persist over longer (centennial and longer) timescales. Reasons for this concern stem from the impact of anthropogenic land-use changes. One example of this is through changes induced by grazing that develop feedback loops that affect the entire ecosystem's carbon and nutrient cycling (Liu et al., 2018). This highlights the need for land-use management recommendations to mitigate longer-term adverse outcomes.

In this review, we investigate the response of the TP biosphere to atmospheric and hydrologic changes and the resulting impacts on the cryosphere. Our focus is on past, present, and future geo-biosphere interactions related to permafrost degradation. We evaluate the hypothesis that if the biosphere is currently benefiting from global warming, these benefits may be short-lived if tipping points in preserving degraded permafrost are crossed. To evaluate this, we present: 1) a summary of observed environmental changes on the TP (section 2.0); 2) scenarios for past and present global climate change impacts on the TP using geologic archives and RCP 2.6 and 8.5 of IPCC AR5 and AR6 (Su et al., 2013; IPCC, 2021 sections 3.0 and 4.0); 3) potential future biosphere response pathways to the previous scenarios and human ecosystem management strategies (section 5.0); and 4) proposed future management recommendations (section 6.0).

2. Observed environmental changes on the Tibetan Plateau

In the following, we document recent atmospheric and hydrologic changes that have impacted the TP cryosphere. We start with an emphasis on past decadal-scale changes in the atmosphere, hydrosphere, and cryosphere, and finally, the biosphere's response to these changes (Sections 2.1 to 2.4). We summarize geologic (millennial to million-year) observations of past warm periods and the Earth system response on the TP (Section 4.0). The latter consideration (past geologic observations) is not often considered in modern environmental change

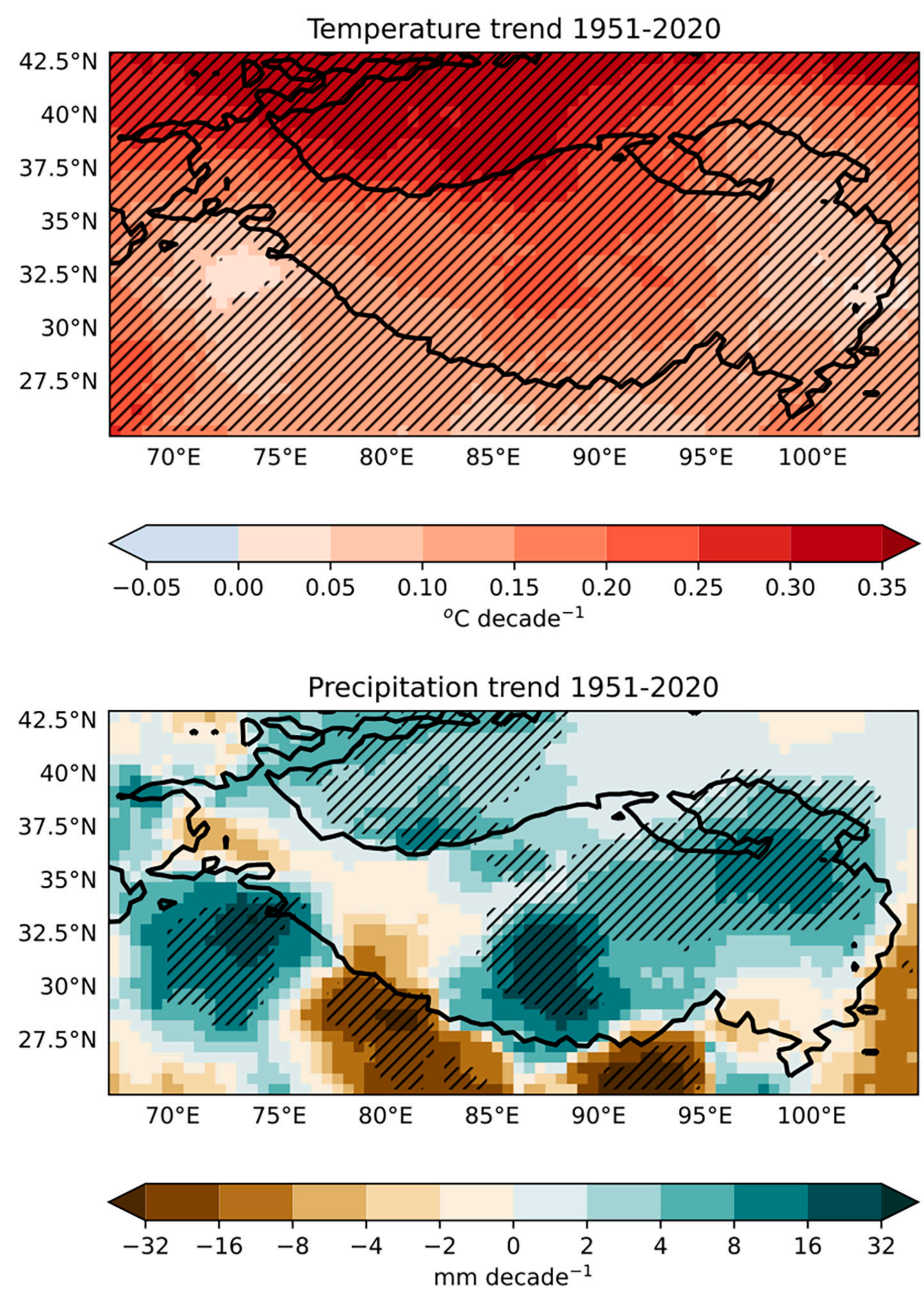

Fig. 3. Decadal trends in mean annual surface temperature and precipitation between 1951 and 2020 based on CRU gridded observation of surface stations. Hatched area indicates a significant trend at the $p < 0.05$ level. The black lines indicate the Tibetan Plateau region with a terrain height of 3000 m.

studies. Still, we consider it here because it provides not only an analog for what the future of the TP may look like but also the biosphere's response to warm periods in the absence of human impacts.

2.1. Observed climate change over the TP

Previous work related to TP precipitation changes found that annual mean precipitation increased at 2.54 mm per decade (from 1982 to 2000, ERA-40) across the plateau, indicating that the TP is overall becoming wetter (Zhong et al., 2011). However, trends in annual and seasonal precipitation rates across the TP from 1979 to 2018 document spatial heterogeneity (e.g., Fig. 3). One clear result from previous studies is that climate change impacts vary between the southern (Himalaya) and central to northern parts of the TP (Lei et al., 2014; Zhong et al., 2019). For example, the increasing aridity in the Himalayas and southern TP is most apparent, with the most significant rate of precipitation decrease occurring during the summer monsoon season. Finally, summer precipitation has also varied with elevation on the plateau and includes an increase with elevation at 1000–4500 m above the TP from 1970 to 2014, at 0.83% per decade km^{-1} (Li et al., 2017b). Thus, previous work has documented recent differences in TP climate change between the northern and southern parts of the plateau and different magnitudes of change with elevation.

In addition, land surface temperatures (LST) have risen on the plateau in the 20th and 21st centuries (Li et al., 2010; Liu and Chen, 2000; Bibi et al., 2018). Remote sensing observations show that annual and monthly LST anomalies increased the most over the central part of the TP and at high elevations (Salama et al., 2012). The LST anomaly trends on the Plateau were seasonally dependent and increased with elevation. Furthermore, Zhong et al. (2011) concluded that from 1982 to 2000, both the LST and the surface air temperature rose on the TP with a rate of LST increase of 0.26 ± 0.16 K per decade and surface air temperature increases of 0.29 ± 0.16 K per decade. The warming started in the early 1950s, much earlier than in the rest of the Northern Hemisphere (NH), which began in the mid-1970s (Liu and Chen, 2000). The TP experienced more substantial warming compared with the Northern Hemisphere and northern midlatitude (e.g., Bibi et al., 2018). It also experienced Elevation Dependent Warming (EDW; Pepin et al, 2015; Yan et al., 2016) with an amplified warming rate at elevations up to ~5000 m (Li et al., 2017b). Thus, the observed EDW has played a critical role in cryosphere change. Fig. 3 shows the regional temperature and precipitation trends over the TP from 1951 to 2020, confirming the warming and over wetting trend in the last decades.

The previous temperature and precipitation changes across the TP have been linked to changes in albedo, atmospheric water vapor, and sensible heat, amongst other things (Gao et al., 2019). Lin et al. (2013) and Zhong et al. (2019) analyzed atmospheric data collected between 1960 and 2014 and explained the physical mechanism of TP warming by a decrease in the plateau surface albedo, a weakening of wind speeds (Guo et al., 2017), decrease in the low cloud fraction and geometrical depth, and an increase in the nighttime low cloud coverage. Recent work shows the dominant role of snow depth change as a controlling factor in EDW (Guo et al., 2021). The significant increase in TP temperature, especially in high-altitude areas, caused accelerated changes in plateau glaciers and snow cover (Yao et al., 2019). Increases in precipitation over the TP (Yang et al., 2011) have been analyzed between 1979 and 2013 and are attributed to the enhanced influx from the southwest boundaries of the TP and enhanced local ET (Zhang et al., 2017).

Finally, changes inferred from ice core isotopic records (e.g., Yao and Thompson, 2017) capture the previous climate observations over the last 50 years and thus provide confidence in the reconstruction of longer-term trends over hundreds to thousands of years across the plateau (Thompson et al., 2006a, 2006b; Yao et al., 2002; Yao et al., 2007; Thompson et al., 2017; Thompson et al., 2018; Yao et al., 2019). For example, the opposing trends in annual precipitation rates between the Himalayas and the TP regions to the north previously observed by rain gauges are also reflected in longer time scale ice core records that record changes since 1600 CE (Chen et al., 2015). These records indicate opposite decadal-scale trends between Dasuopu in the monsoon-dominated south and Guliya, Puruogangri, and Dunde in the westerlies-dominated and transition regions (Thompson et al., 2006b). Furthermore, both ice core and tree ring (Liu et al., 2009b) records demonstrate a warming trend such that the last decade is the warmest in the past 2000 years. In situ observations of 2-m air temperature data from meteorological stations below 5000 m a.s.l (Gao et al., 2018) show temperatures are rising faster at higher elevations due to EDW, especially during the winter and fall seasons. Ice core isotopic data over the last 1000 years document similar enhanced warming at higher elevations across the Plateau (Thompson et al., 2018). Such tendencies have been manifested mainly since the 1990s and may continue in the future, particularly during the winter and spring seasons (Liu et al., 2009b). A future continuation of EDW is expected because the snow line elevation is rising under a warming climate, thus further enhancing warming at higher elevations.

2.2. Observed hydrosphere changes

The combined effects of spatiotemporal variations in precipitation, evaporation, and meltwater from permafrost and glaciers have led to variations in TP river discharge and lake levels. Su et al. (2019) reported large interannual variations in streamflow, but no significant trends over the entire TP were observed from the 1950s to the beginning of the 21st century. However, more localized spatial differences in streamflow were observed within different regions of the TP (Xie et al., 2010; Su et al., 2019; Zhang et al., 2019; Yang et al., 2014), which might be related to variations in the dominant hydrological process active within individual river basins (Cuo et al., 2014; Lutz et al., 2014). For example, previous work has suggested that precipitation dominates streamflow in the northern, eastern, and southern basins. In contrast, in the central and western basins, meltwater and groundwater contributions are more significant (Cuo et al., 2014). Annual streamflow in the source region of the Yellow River decreased while the Yangtze Mekong, Salween, and Brahmaputra River source regions experienced a moderate increase. These changes have been partly attributed to climate change over the TP (Yang et al., 2014; Lutz et al., 2014).

Spatial variations in the hydrologic cycle of lakes across the TP have also been documented. Estimates of the multiyear (2003–2016) average evaporation of the TP lakes from meteorological and satellite data (Wang et al., 2020a) are 29.4 ± 1.2 km^3 $year^{-1}$ for the 75 big dimictic lakes (lake area larger than 88 km^2) and 51.7 ± 2.1 km^3 $year^{-1}$ for all plateau lakes. Lakes at a higher elevation, with a small area, and higher latitude were identified as having a shorter ice-free season and lower evaporation. Han et al. (2021) estimated multiyear (2001–2018) land evapotranspiration across the TP using meteorological and satellite data. They found that annual evapotranspiration increased significantly at a rate of 2.62 $mm\,yr^{-1}$ ($p < 0.05$) in the eastern TP (longitude > 90° E) but decreased significantly at a rate of -5.52 $mm\,yr^{-1}$ ($p < 0.05$) in the western TP (longitude < 90° E). Other studies (Yang et al., 2011: Liu et al., 2009a) of the hydrologic cycle across the TP (from 1984-to 2006) document changes in the surface water balance whereby precipitation increases were insignificant in the central TP and decreased along the TP periphery. In contrast, evaporation showed an overall increasing trend across the TP. These factors lead to decreasing discharge at major TP water source areas in the semi-humid and humid zone of the eastern and southern TP. Spatial variations in lake levels between the TP interior and the Himalayas have also been documented (Lei et al., 2014; Zhang et al, 2020, ESR). The major lakes in the central TP have expanded significantly in area and volume since the 1990s, whereas many lakes along the marginal region of the south and east TP have shrunk. Recent work showed that enhanced precipitation was the main driver of the lake level increase followed by glacier melt and permafrost thaw (Zhang et al., 2020). Li et al. (2014) found that 79% of the increased lake levels observed between 1970-and 2010 on the central-northern TP are correlated with areas of thawing continuous permafrost, suggesting a connection between permafrost conditions and surface water budgets.

Hydrosphere changes across the TP have occurred laterally (as described above) and with increasing elevation in response to EDW. For example, glacier mass loss on the TP and its surrounding regions has been well documented (Bolch et al., 2019; Brun et al., 2017; Yao et al., 2019) and has contributed to increased surface water runoff. However, the changes are heterogenous, with the highest rates of glacial mass loss occurring in the SE Tibetan Mountains and western Himalayas and balanced conditions, to slightly positive mass budgets, in the west and northwest of the plateau, such as the Central Karakorum, Eastern Pamir and Western Kunlun (Bolch et al., 2019; Brun et al., 2017; Neckel et al., 2014; Kääb et al., 2015). This general pattern has been similar since at least the 1970s (Bolch et al., 2017; Zhou et al., 2017). However, the rate of mass loss strongly increased in most regions since the mid to late 1990s (Bolch et al., 2012, 2019; Maurer et al., 2019; Zhou et al., 2018), with the highest increase in parts of the western Himalayas (Mukherjee et al., 2018). Recent studies show that glacier mass loss prevails now even in areas with formerly balanced conditions (Hugonnet et al., 2021; Bhattacharya et al., 2021). Furthermore, the previous changes have been accompanied by snow-covered days and snow depth changes. Both snow-covered days and snow depth decreased in the north and northwest TP regions and increased in the southwest and southeast parts of the TP from 2000 to 2014 (Huang et al., 2016).

2.3. *Observed and projected estimates of permafrost change*

At the turn of the century, >50% of the area of the TP was influenced by permafrost (Cheng, 2005). Long-term temperature measurements indicate that mean permafrost temperature at 10–20 m depth increased by approximately 0.08–0.24 °C yr^{-1} since the 2000s (Hock et al., 2019). Other studies have documented an increase in permafrost temperatures at 6 m depth of 0.1 to 0.3 °C between 1996 and 2001 (Fig. 4) (Cheng and Wu, 2007). The previous increases in permafrost temperature changes are driven by increases in mean annual ground surface temperatures of 0.3–0.5 °C (documented between 1970 and 1990; Li et al., 2008). In addition, satellite monitoring shows that the number of frozen days decreased by 16.8 days per decade between 1988 and 2007 (Li et al., 2012). In the past half-century, the thermal state of 88% of the permafrost region of the TP has degraded to unstable levels (Ran et al., 2018; Zhao et al., 2021). Previous studies have documented a robust spatial heterogeneity in where warming occurred on the TP.

The positive feedback between warming increased snowmelt and rising snow-lines is causing a thickening of the permafrost active layer (Cheng and Wu, 2007; Piao et al., 2010; Yao et al., 2012; Guo and Wang, 2013; Yang et al., 2014; Gao et al., 2015; Xu and Liu, 2007). Both observational data and linked permafrost and general circulation model (GCM) simulations indicate that the lower altitudinal limit of permafrost has risen by >25 m between 1980 and 2000 in the north of the TP and > 50 m in the southern TP from between 1990 and 2000 (Cheng and Wu, 2007). These increases in the lower altitudinal limit of permafrost are coeval with increases in the permafrost active layer thickness since the start of measurements in 1995 (Bolch et al., 2019) by around 0.15–0.50 m (between 1996 and 2001; Cheng and Wu, 2007) and 0.15–0.67 m since the 2000s (Cheng and Jin, 2013, Hock et al., 2019). Furthermore, recent work by Xu and Wu (2021) conducted a multimodel comparison for permafrost thawing across the plateau and suggested that between the years 2015 to 2100, the thickness of the active layer will increase between 2.5 and 17.5 cm/decade, with the previous range of numbers reflecting spatial variability thickness increase across the plateau.

As a consequence of the previous changes, the southern boundary

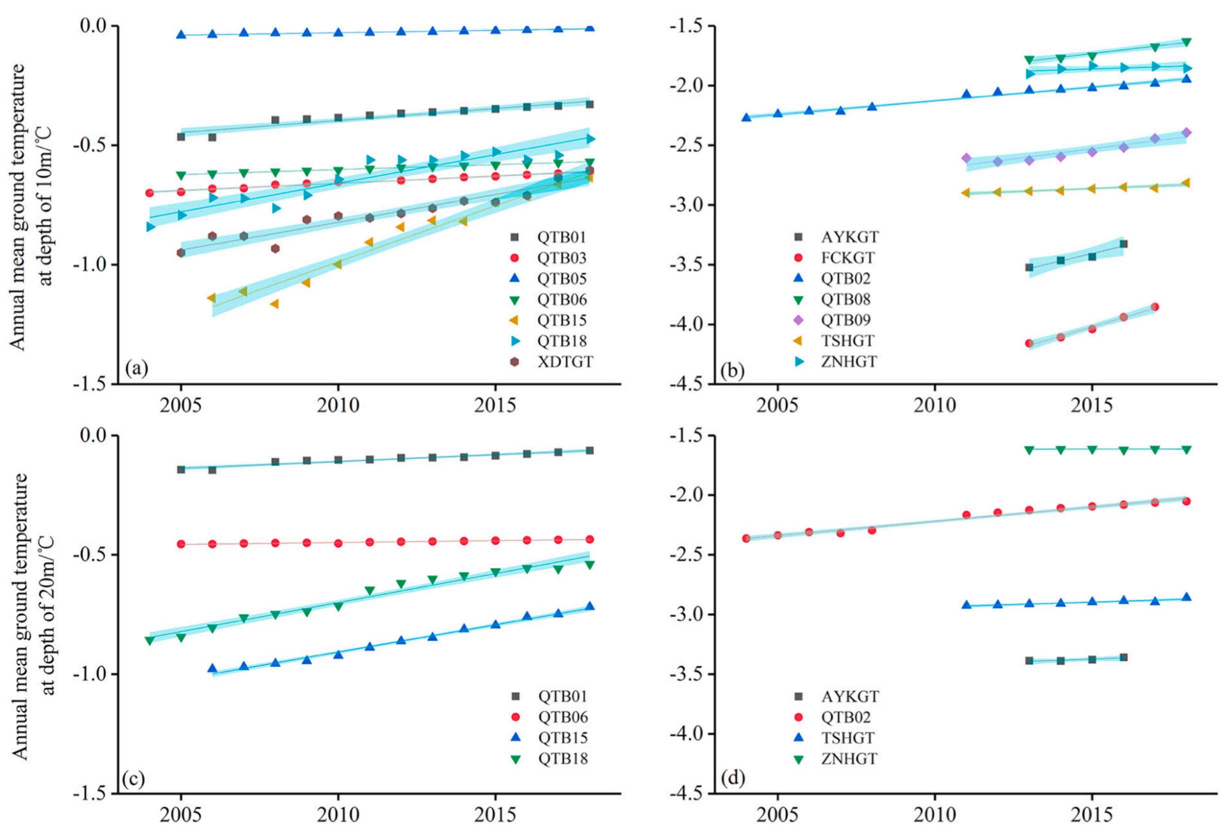

Fig. 4. Example of the permafrost ground temperatures warming between ~2005 to 2021 from a permafrost monitoring network across the TP. The different stations are located along a roughly north-south transect in the central TP at ~93° East Longitude (see Zhao et al., 2021 for exact locations). All stations, regardless of latitude, show decadal timescale warming of permafrost at depths of 10 and 20 m below the surface and increase the active layer thickness (not shown). Data and figure from Zhao et al. (2021) (reproduced with permission of Creative Commons Attribution 4.0).

between extremely stable and sub-stable permafrost (temperatures <-1.5 °C) has moved northward, as has the southern boundary of the transitional, unstable and likely thawing permafrost (temperatures of -1.5 °C to >0.5 °C). The northward migration of these boundaries results from an increased permafrost loss rate. The permafrost loss rate has been estimated as $9.52 \times 10^4 km^2$ per decade between 1960 and 2000, with a total area of thermally degraded permafrost of $153.76 \times 10^4 km^2$ (Ran et al., 2018). Wu et al. (2018) reported that about 59% of the total permafrost could be categorized as transitional and unstable permafrost.

The recent trends in permafrost degradation are predicted to continue into the near future. Recent work suggests that permafrost degradation will continue until at least 2090 (Fig. 5). Work by Wang et al. (2019b) applied machine learning and statistical analysis techniques trained on the observed permafrost extent across the plateau during a baseline period between 2003 and 2010. The trained algorithms were then applied to five GCM simulations to predict future

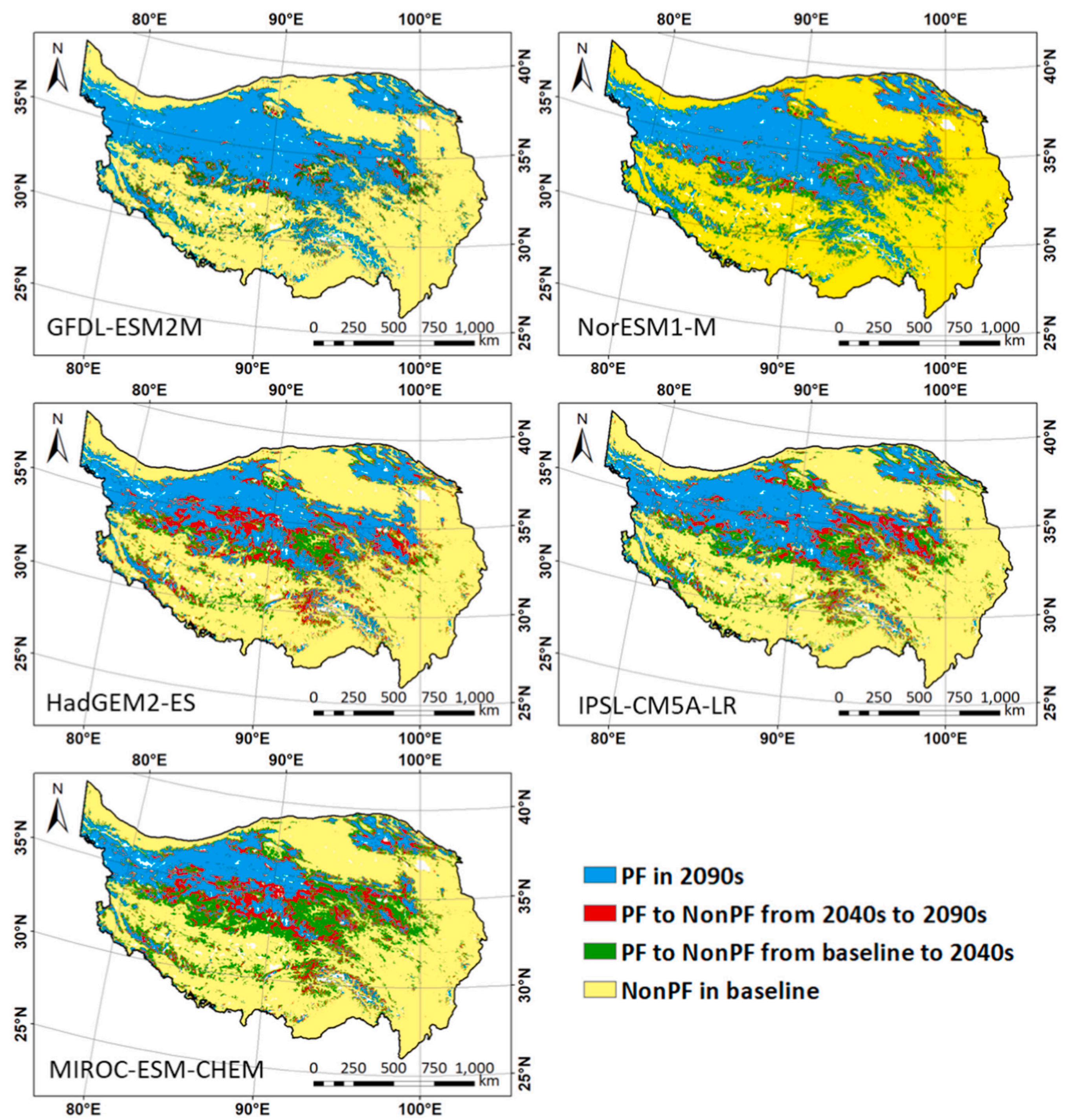

Fig. 5. Predicted future (until the year 2090) permafrost (PF) degradation across the Tibetan Plateau from five different GCM simulations. Results are shown relative to a baseline period between 2003 and 2010. Colored areas represent results from a majority voting approach using a combination of different machine learning and statistical techniques (i.e., logistic regression, support vector machine, and random forest techniques). Figure from Wang et al., 2019b (© Elsevier B.V., reproduced with permission).

permafrost extent when driven by RCP 4.5. They found that 25.9% and 43.9% of the current permafrost will disappear by 2040 and 2090, respectively. Thus, despite differences in parameterizations between the GCMs, or the analysis technique used (machine learning or statistical), a clear picture of continued future degradation of permafrost area is expected. Finally, the previous changes in active layer thickness and permafrost degradation have recently been identified as closely linked to ecosystem changes in permafrost areas across the Northern Hemisphere. More specifically, ecosystem changes (e.g., as identified by the leaf area index, Bao et al., 2014) are associated with enhanced degradation of permafrost. In contrast, more stable ecosystems are typically associated with slower rates of permafrost degradation (Ran et al., 2021a, 2021b).

2.4. Observed evidence of biosphere changes

Observations from long-term in situ ecosystem monitoring and satellite remote sensing have made advances in understanding the responses of the TP ecosystem to climate change (e.g., Fig. 6). Here, we provide evidence of the impacts of current rapid climate change and CO_2-fertilization on the biota's structure and functioning, including vegetation phenology, tree line position, ecosystem productivity, and carbon sinks.

Time series analysis of Normalized Difference Vegetation Index (NDVI) imagery of 315 SPOT vegetation scenes between 1998 and 2006 revealed that the spatial distribution of NDVI values is in agreement with the general climate pattern across the TP (Zhong et al., 2010), with warmer and wetter climate favoring higher NDVI (Figs. 6, 7). The Asian monsoon greatly influences seasonal variations in NDVI but vegetation density (inferred from NDVI values) on the entire TP has generally increased (e.g., Zhong et al., 2010; Li et al., 2015b). A time lag response has also been found between NDVI and climate variables using 1 km resolution land cover maps from the Global Land Cover 2000 database and monthly near-surface air temperature and precipitation at seven meteorological stations. Except in desert grassland settings of the western highlands, the NDVI of all selected sites correlated with air temperature and precipitation (Fig. 7). Variations in these correlations occurred due to the different land cover types at other locations. The strongest correlation was found in alpine and subalpine grassland, the weakest in desert grassland (Zhong et al., 2010; Shen et al., 2015). Fig. 6 shows changes in the seasonal Normalized Difference Vegetation Index (NDVI) over the TP from 1982 to 2016, which shows a dominant greening in the past decades.

Global warming during the past decades has significantly advanced spring vegetation phenology in the TP. The spring phenology of TP grasslands shows a general advance since the 1980s, increasing by 0.31–0.88 days year^{-1} during the 1980s and 1990s. Since ~2000, there has been no evidence of continuously advanced spring phenology over the whole plateau (Wang et al., 2017). Specifically, the spring phenology is delayed in the southwestern region but advances in the northeastern region. Such spatial variations in green-up dates are consistent with the changes in precipitation, whereby spring precipitation declined in the southwestern TP but increased in the northeastern TP (Zhong et al., 2019). These observations highlight the critical role of precipitation in regulating spring phenology on the TP. In addition, warming across the TP has shifted the tree line upwards, with spatially varying amplitudes. Over the past 100 years, tree lines in the eastern slope have migrated upward, with the highest rate reaching 80 m /century (Zhong et al., 2019). Besides warming, species competition and precipitation also regulate the migration rate. For example, species competition could explain over 70% of the variations in the migration rates of treeline over the TP.

Warming and CO_2-fertilization have also significantly increased vegetation productivity on the TP. Vegetation productivity shows a significant upward trend, which leads to an enhanced ecosystem carbon uptake and a carbon sink of 23.4–34.3 Tg C/yr (e.g., Piao et al., 2012). Warming-induced vegetation carbon accumulation is higher in Kobresia pasture settings than in the alpine steppe. Kobresia pastures show extreme values in root-to-shoot ratio, and it is yet unknown how increased aboveground biomass will translate into belowground C stocks especially considering feedback reactions via nutrient cycling (Schleuss et al., 2015). Ganjurjav et al. (2016) reported that experimental warming led to an increase in the net primary production of alpine meadows, while it decreased in alpine steppes due to warming-induced drought. Yang et al. (2008) used a repeated soil survey approach to quantify changes in topsoil (30 cm) carbon stocks in Tibetan grasslands. They found that soil organic carbon stocks remained relatively stable between the 1980s and 2004, with changes ranging from -36.5 to 35.8 g C m^{-2} yr^{-1}. With a contribution of at least 40% to total soil organic C, microbial necromass' role in the Tibetan grassland's belowground C stocks is substantial. Yet, its controlling factors (temperature, soil moisture, mineralogy) are not sufficiently understood for implementing this critical C pool in Earth system models (He et al., 2021). Therefore, estimates of changes in soil carbon stocks under climate change are still highly uncertain. This is mainly due to the high spatial heterogeneity in soil properties and the lack of information on deep-layer soil processes and their influence on water availability for vegetation. As warming continues, the permafrost soil carbon dynamics are crucial in the future carbon balance over the TP. Another factor is the interrelationship between warming and increases in CO_2 concentration. Han et al. (2019) report that elevated atmospheric CO_2 may contribute more to plant growth than organic carbon decomposition on the TP and may thus outbalance a likely carbon loss from soil due to warming.

3. Interactions between permafrost changes and the biosphere

Permafrost degradation on the TP has accelerated as the climate warms. This degradation affects regional hydrology and terrestrial ecosystems (Cheng and Wu, 2007; Ran et al., 2018; Cheng and Jin, 2013). Degradation of permafrost has changed groundwater recharge and exchanges between soil water and surface water such that available water for plants has decreased on the plateau. The decrease in water available to plants could lead to harsh conditions for vegetation with short roots, a reduction in species diversity, and the degradation of the stability of vegetation and terrestrial ecosystems (Jin et al., 2009; Cheng and Wu, 2007). These ecosystem changes also influence the storage of soil organic carbon and vegetation dynamics in the permafrost. Climate warming has increased the vegetation greenness and productivity since the 1980s (Piao et al., 2011; Chen et al., 2014) and will increase the content of active carbon and carbon fluxes in both the growing and non-growing seasons (Gao et al., 2016; Mu et al., 2017). While the soil organic carbon in the upper 2 m of the TP permafrost region is approximately 12 to 28 Pg (Yang et al., 2008; Mu et al., 2015; Ding et al., 2016, 2017a; Zhao et al., 2018), there is also around 35 Pg soil organic carbon stored below 3 m with an uncertainty range of 20 to 54 Pg (Wang et al., 2020c). The Cyperaceae wetlands covering permafrost are presently a carbon sink. Still, it is possible that it will shift to a carbon source due to the degradation of vegetation and loss of organic carbon along with permafrost degradation (Wang et al., 2017; Wang et al., 2020c).

In addition to the previous climate change effects on carbon storage in TP permafrost, anthropogenic impacts have also figured prominently for ecosystem change. For example, human activities (e.g., grazing, agriculture) influence the vegetation and ecosystem productivity of the TP (e.g., Li et al., 2016; Huang et al., 2022). Moderate grazing by cattle and underground dwelling rodents and other herbivores have played a crucial role in maintaining alpine ecosystems' sustainable high productivity and biodiversity (Ingrisch et al., 2015; Miehe et al., 2019; Wang et al., 2018; Stover and Henry, 2019). However, in the last decades, more sedentary grazing regimes have been developed, and the prevailing practice is the contracting of grasslands to individual households, including fencing and grazing control (Cao et al., 2017). In summary, climate warming and human activities have greatly influenced the vegetation dynamics and carbon storage on the TP. However,

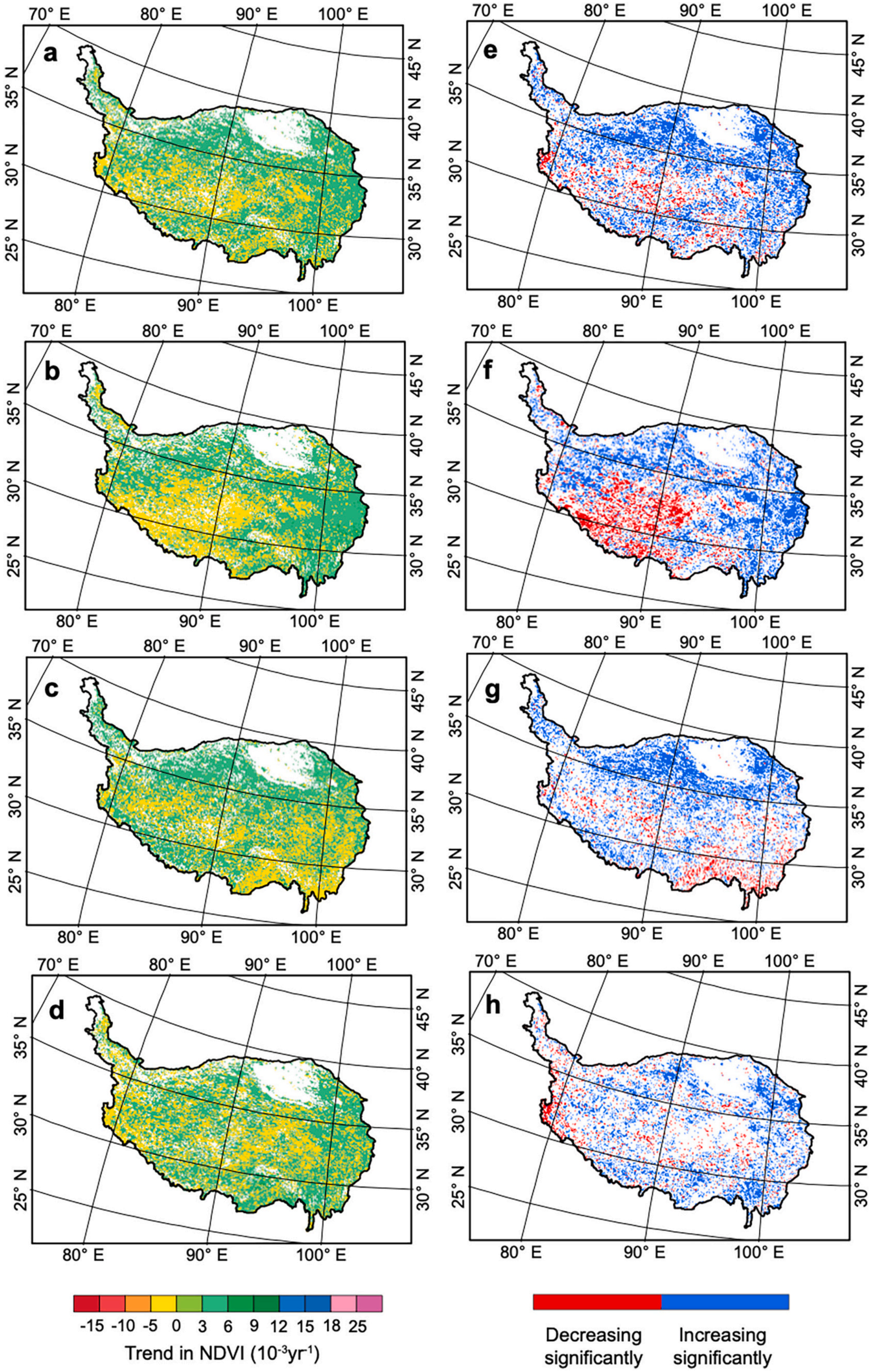

Fig. 6. Evidence from GIMMS NDVI3g data (Pinzon and Tucker, 2014) for the TP over the past decades (1982–2016) in response to warming and changing precipitation patterns. Trends of the mean Normalized Difference Vegetation Index (NDVI) are shown, along with areas showing significant trends at the $p < 0.05$ confidence level during the growing season (March–November) (a and e), spring (March, April, May) (b and f), summer (June, July, August) (c and g), and autumn (September, October, November) (d and h). An overall greening is clear, although most prominent in the summer.

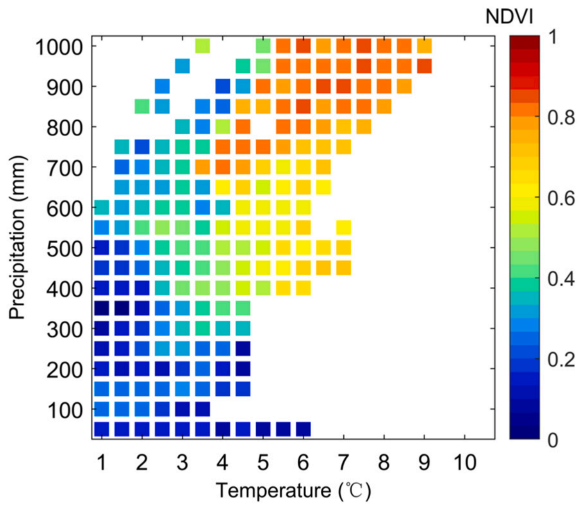

Fig. 7. The mean growing season NDVI as a function of climate. Colors represent (March–November) NDVI from 1982 to 2016 for intervals of 0.5 °C in mean growing season temperature and 50 mm intervals of mean growing season precipitation. The satellite-derived NDVI was retrieved from GIMMS NDVI3g data (Pinzon and Tucker, 2014). The climate space is constructed using data from the Climatic Research Unit, University of East Anglia (CRU TS 4.02). The NDVI data used here are the same as those shown in Fig. 6.

more work is needed to quantify the relative importance of climate change and human activities on permafrost ecology and the developing trends of the TP biosphere under global warming. Stepwise ecological restoration needs to explicitly consider the impacts of climate change (Liu et al., 2018).

4. Tibetan Plateau geo-biosphere interactions in the geologic past

Earth's history provides a treasure of climate and biosphere-related observations with insight into the response of the TP to environmental change (Fig. 8). Lessons learned from the geologic record document system changes and possible tipping elements for biosphere adaptation to climate change. Previous studies provide compilations of paleotemperature reconstructions across the TP (Burke et al., 2018). Here we focus on the Holocene Climate Optimum (HCO), Marine Isotope Stage 11 (MIS-11) in the Late Pleistocene, and the Middle Miocene Climate Optimum (MMCO). More specifically, during the HCO, temperatures in the northern hemisphere were warmer than today (Bartlein et al., 2011), and a slight temperature anomaly caused marked changes in the environment. The HCO, therefore, provides hints for understanding the effects of mild global warming on the TP. Northern hemisphere high latitude summer temperatures during MIS-11 peaked with temperatures between 2.1 and 3.4 °C above present (de Vernal and Hillaire-Marcel, 2008; Robinson et al., 2017) and match projections of intermediate RCP scenarios for the year 2100. Our current understanding of the rates and magnitudes of terrestrial temperature change is still limited. The MMCO between 17 and 15 Ma (Fig. 8) was a globally warm period (e.g., Zachos et al., 2001, Steinthorsdottir et al., 2021) with a reduced pole-to-equator temperature gradient, North Atlantic sea surface temperatures of >10 °C (Super et al., 2020) and mid-latitude temperatures about 6 °C above present (Flower and Kennett, 1994). Middle Miocene temperatures in Central Europe were relatively high, with rapid and extensive magnitude changes at the end of the MMCO (Methner et al., 2020). Previous work documents that northern hemisphere westerlies were instrumental in setting mid-latitude temperature and precipitation conditions (Methner et al., 2021). Although only moderately high atmospheric CO_2 concentrations of 300–600 ppmv prevailed during most of the mid-Miocene (Kurschner et al., 2008), the higher temperatures suggested from the geologic record make the MMCO a suitable analog for extreme RCP scenarios. Regardless of where, when, and what magnitude of climate change has occurred in the past, the endemic plant species in the Alpine Steppe of the western and central plateau testify to a high resilience to climate change due to their wide thermal and hygric range (Miehe et al., 2011). We detail the TP response to environmental change during these different time periods in the following.

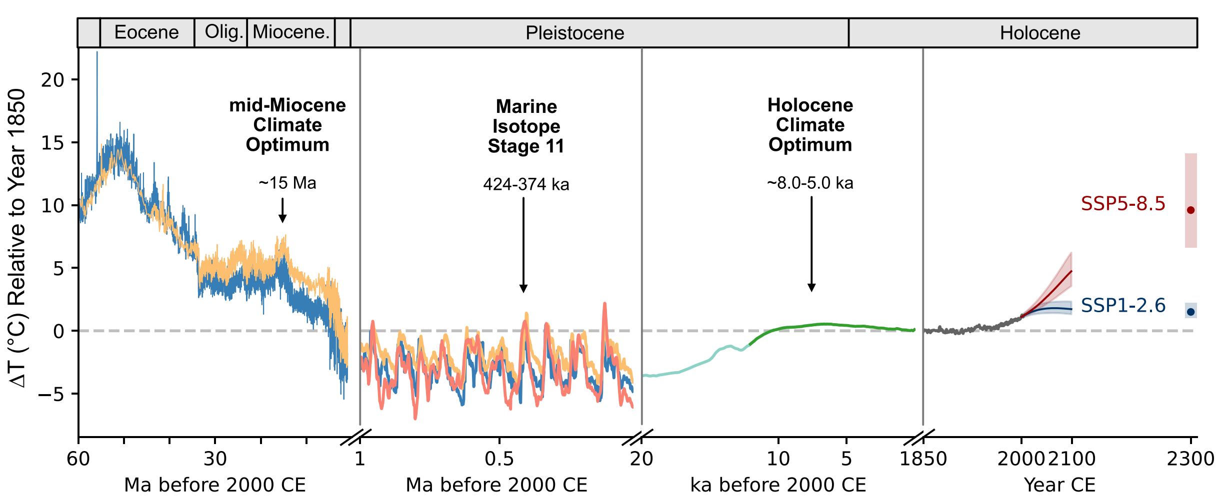

Fig. 8. Global mean temperature trends over the geologic past (last 60 Ma) and future projections (right side) based on AR6 scenarios SSP1–2.6 (blue) and SSP5–8.5 (red). Temperature changes are shown relative to Pre-Industrial conditions (year ~1850). Paleo temperature estimates are a composite of five proxy-based reconstructions (see Burke et al., 2018 for description). Note that the horizontal axis scale changes and is non-linear to highlight the high projected rate of future warming. The primary patterns evident include the long-term cooling trend from 65 Ma to the present, with periodic, higher frequency, variations resulting from changes in Earth's orbit over Milankovitch timescales. The rate of future warming is faster than experienced in the geologic past for the predicted magnitudes of warming under different scenarios. Data sources: dark blue line: Westerhold (2020); yellow line: Hansen et al. (2013); Pink line: Snyder (2016); light blue line (20–10 ka): Shakun et al. (2012); light green line (10 ka to present): Kaufman et al. (2020); Data between 2015 and 2100 are from IPCC WGI AR6: IPCC (2021) and Gulev et al. (2021); Data at 2300: Lee et al. (2021). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

4.1. Lessons from the Holocene Climate Optimum (HCO), ~8.0–5.0 ka BP

The present interglacial (Holocene) started ~11 kyr ago and was characterized by a long-term decline in summer insolation that influenced the temporal-spatial configuration of climate across the TP (Herzschuh, 2006; Ramisch et al., 2016; Herzschuh et al., 2019; Chen et al., 2020). Palaeoclimatic records and modeling results show (Fig. 8) that the Holocene started with postglacial climatic warming leading to a climate optimum between roughly 8.0 and 5.0 ka BP (Dallmeyer et al., 2015). According to compilations of pollen data (Bartlein et al., 2011) and PMIP2/PMIP3 climate model simulations shown in the IPCC AR5, mid-latitude northern hemisphere temperatures of the coldest month were not elevated, but those of the warmest month were ~ 1–2 °C higher than today. The Indian Monsoon strongly influenced the southern part of the plateau with teleconnections far to the north. At the same time, the SE Monsoon was prominent until the mid-Holocene and mainly affected the northeastern region and the northern TP, which was also heavily influenced by the Westerlies (Li et al., 2017a; Mutz et al., 2018; Mutz and Ehlers, 2019; Herzschuh et al., 2019; Chen et al., 2020).

The subsystems of the TP showed different responses to Holocene climate variability (Li et al., 2017a; Mutz et al., 2018; Mutz and Ehlers, 2019). While vegetation dynamics and the hydrological cycle seem to align with long-term natural climate deterioration towards the late Holocene, the cryosphere revealed complex responses. Long-term muting of the summer monsoon after the HCO resulted in climate cooling and a stronger invasion of dry air masses from inner Asia, as demonstrated by widespread lake-level decreases across the northern parts of the plateau during the late Holocene (Liu et al., 2013; Yu et al., 2019). However, spatial variability in past lake levels have been documented, and some lake levels increased. The highest lake levels, up to >200 m above present, are recorded on the southern plateau due to increased moisture transport from the Indian summer monsoon during the late glacial to mid-Holocene (Miehe et al., 2021). In addition to summer precipitation, glacial runoff contributed to a positive water budget. Mountain glaciers reacted to these changes and reached a threshold around 4.5 ka BP, at which point they tended to readvance in the valleys across the plateau. This advance is referred to as the Neoglacial stage (Xu et al., 2019).

Although there are only a few studies available, existing studies document that permafrost landscapes of the plateau varied in their distribution during the Holocene. Former thermokarst activity has been reported for the northeastern TP during the early Holocene high-lake-level stage, precisely in the same region that today shows rejuvenation of thermokarst subsidence (Opitz et al., 2013). The evaluation of cryogenic structures in drill cores and field outcrops suggests a 30–40% reduction in permafrost at the surface during the climate optimum. This permafrost reduction resulted from a deepening of unfrozen ground to 15–25 m below the surface and a rise of the permafrost table about several hundreds of meters in altitude. Neoglacial cooling reversed this trend to even more widespread permafrost conditions than today (Jin et al., 2009). From this high variability of permafrost in the past, it can be expected that permafrost will decline strongly in response to future warming.

4.2. Marine Isotope Stage 11 (MIS-11), 424–374 ka BP

Although the MIS-19 is often considered to represent the best analog of the present-day interglacial (MIS-1), we select MIS-11 for discussion here because it has the most extensive proxy records available for the TP and Central Asian region. The exceptionally long MIS-11 (425–395 ka, Yin and Berger, 2015) provides a reasonable analog for the impact of solar insolation increases on the TP. MIS-11 is considered warmer than MIS-1 when the effect of CO_2 is included (Yin and Berger, 2015) (Fig. 8), and it is characterized by the strongest amplitude deglaciation of the past several million years (Droxler et al., 2003). In their review of the role of MIS 11 as an analog for the current interglacial, Candy et al. (2014) provide evidence from EPICA Dome C, Antarctica, and a wide range of marine and lacustrine sequences that show a relatively consistent picture of a long interglacial with multiple and discrete climate peaks as temperature"plateaus." Furthermore, there is a substantial degree of consistency particularly in the relative timing of temperature maxima showing that maximum warmth occurred later in the interglacial (Candy et al., 2014). A range of records show that the end of MIS 11 was punctuated by multiple stadial and interstadial events (Candy et al., 2014). For the TP and adjacent areas, a warmer but drier MIS-11 is revealed by the thickest loess deposits on the western Chinese Loess Plateau. Generally, warm-humid climate dominated the paleosol development, but the climate was volatile throughout MIS-11 (Shi et al., 2016). Evidence from sediments in a paleolake in the Heqing Basin on the southeastern fringe of the TP suggests that climate quickly became drier after the peak of MIS-11, with relatively low climate variability afterward until the last glacial (Hu et al., 2015). In addition, the substantial decline of Tsuga (hemlock spruce) pollen at the end of MIS-11 indicates rapid cooling (Hu et al., 2015). Results from another sediment core in the same basin underline the importance of interhemispheric forcing driving Indian Summer Monsoon (ISM) variability at the glacial-interglacial time scale as today. Interglacial ISM maxima were dominated by an enhanced Indian low associated with global ice volume minima. During strong interglacials, ISM maxima coincide with Northern Hemisphere ice volume minima and Antarctic temperature maxima (An et al., 2011).

Zhao et al. (2019) used a high-resolution speleothem record of MIS-11 as a natural analog for Holocene Asian Summer Monsoon (ASM) variations. They found that orbital-centennial patterns of the ASM were remarkably similar during both interglacials, including their pacing and structure. A multi-millennial more vigorous monsoon late in MIS-11 is similar to the late Holocene strengthening of the ASM about 2000 years ago. From this, Zhao et al. (2019) speculate that the current century-long ASM warming trend may persist into the future if only natural forcings are considered.

Candy et al. (2014) proposed many of the critical attributes of MIS 11 including: the possible loss of the Greenland and Western Antarctic ice sheets, abrupt events late in the interglacial, and the occurrence of a long and warm interglacial. They suggest these events may be analogous to the future of the current interglacial, even if the exact parameters and precise structure of MIS 11 are not. More specifically, Hu et al. (2015) speculate that warming could have triggered a long-term persistent change of the land surface cover (vegetation, snow, and ice cover) on the TP, strong enough to cause a moderate shift in the ISM and keep its influence low on the southeastern plateau area for several hundred thousand years. New but yet to discover MIS-11 evidence from Gulija ice cores and lacustrine records (Haberzettl et al., 2019), combined with marine records, will be of significant interest in unraveling the effects of persistent warmth as well as feedbacks and teleconnections in the climate system. Thus, despite high uncertainty in the conditions of the MIS-11 interglacial period on the TP, what is clear from existing observations is that humid and warm conditions prevailed, well-developed soils formed, and vegetative land cover expanded. We can deduce from these findings that future warming of the plateau will result in permafrost loss and a greening of the plateau.

4.3. Lessons from the mid-Miocene Climate Optimum (MMCO), ~15 Ma

Paleoclimate and paleoelevation reconstructions suggest the long-term patterns of atmospheric circulation (Caves et al., 2015) and overall TP topography (e.g., Gébelin et al., 2013; Spicer et al., 2021) shared significant similarities with today's conditions (e.g., Fig. 8). Major adaptations towards modern biota and a peak of diversification characterized the middle Miocene (e.g., Favre et al., 2015, Mobrugger et al., 2018). For example, the 15 Ma lacustrine deposits in the Namling-Oiyug basin (situated at an altitude of about 4500 m) in the southern Tibetan

Plateau provide insight into MMCO conditions on the TP. The Namling-Oiyug fossil flora (e.g., Spicer et al., 2003; Khan et al., 2014) provides insight into the southern plateau region in a warmer climate similar to the extreme RCP 8.5 scenario. The Namling-Oiyug fossil flora indicates a temperate broad-leaved deciduous woodland was present on the southern TP, as exists today in high elevations of the Himalayas (Khan et al., 2014). Reconstruction of climate conditions using CLAMP analysis (Khan et al., 2014) indicates a mean annual temperature of the southern TP of around 8 °C (similar to Lhasa today). These floras also suggest mean temperatures of ~24 °C and ~ −6 °C for the warmest and coldest months, respectively, and a growing season of about 6 months.

There is considerable uncertainty in the reconstructed paleoclimate conditions during the MMCO, especially for the monthly growing season precipitation, because of unknown vapor transport over the Himalayas (Khan et al., 2014). There is increasing evidence that at ca. 15 Ma, the Himalaya had already formed an efficient orographic barrier for atmospheric circulation (e.g., Gébelin et al., 2013; Thiede and Ehlers, 2013; Spicer, 2017; Ding et al., 2017b) and potentially rose above the elevation of the TP in the mid-Miocene. Moreover, the lateral extent of the paleolake in the Namling-Oiyug basin is unknown, but the lake likely supported moisture in its surroundings (Khan et al., 2014). Despite uncertainties in paleoclimate conditions, the existence of forests in the southern TP during the MMCO is a robust result and a bold statement in the context of a greening of the TP under warmer conditions. However, the impact of the rate at which climate change occurs on the formation of forests or the potential failure of forest development under rapid anthropogenic driven climate change is unknown.

5. Trajectories for future change on the Tibetan Plateau

5.1. Future RCP scenarios for climate change

As part of the Coupled Model Intercomparison Project (CMIP5), 24 Global Climate Models (GCMs) were analyzed to project future changes in precipitation and temperature over the TP (Su et al., 2013). Here we focus on twenty-first-century trends of precipitation and temperature over the TP based on these GCM projections for scenarios RCP2.6 and RCP8.5. Collectively, all models suggest a warming trend until 2100 under RCP8.5, while RCP2.6 permits a slight cooling trend. Taking 1961 to 2005 as a mean annual temperature (MAT) baseline, the annual MAT at the end of the 21st century would be +1.8 °C and + 4.1 °C under the RCP2.6 and RCP8.5 scenarios, respectively. The projected long-term precipitation change amounts to +6.0% and 12.0% for RCP2.6 and RCP8.5, respectively (Su et al., 2013). Although changes in precipitation over the past 50 years show a contrasting spatial pattern with wetting in the northern TP and drying in the south, future precipitation projections by CMIP5 GCMs show a wetting trend over all of the TP. The previous scenarios for climate change have been suggested to cause an extension of the vegetation growing period by at least 6 weeks (Su et al., 2013). These results are supported by dynamic downscaling and a regional climate model that shows similar changes as coarser GCM simulations but provides more spatial details in the warming and overall wetting. Of particular interest is the appearance of a drying trend along the Himalaya (e.g., Ji and Kang, 2013) predicted by regional climate models. This prediction is in agreement with the observed changes in the last decades.

5.2. Future trajectories for permafrost

Permafrost over the TP is expected to undergo increased thawing and degradation during the 21st century due to increasing air temperature (Guo and Wang, 2016; Lu et al., 2017). An empirical-statistical model-based sensitivity experiment predicted that permafrost will reduce by about 18% by 2049 along with a temperature increase of 1.1 °C. A permafrost reduction of 58% is expected if air temperature increases by 2.9 °C by 2099 (Li and Cheng, 1999). Lu et al. (2017) indicated that the permafrost area will be strongly reduced by 22–64% by 2100 for RCP2.6 and RCP8.5. Wang et al. (2019a) projected a 44% decline in permafrost area by 2100 under RCP4.5. Furthermore, Yang et al. (2007) found that the active layer thickness will deepen by 0.1 to 0.4 m by 2050 and 0.2 to 0.7 m by 2100 along the Qinghai-Tibet Railway. Furthermore, according to the soil temperature predictions derived from CMIP5, by 2099 the ensemble mean soil temperature at 1 m (15 m) depth will increase along the TP engineering corridor by about 1.0 °C (1.0 °C) for RCP2.6, and 4.5 °C (−3.6 °C) for RCP8.5. These changes may result in a subsidence hazard by 2050, even under the best-case scenario of RCP2.6 (Guo and Sun, 2015). However, it should be noted that such coarse-scale simulations (Guo and Wang, 2016; Lu et al., 2017) are of limited use in quantifying permafrost changes and assessing related influences over the TP due to inadequate representation of complex terrain. Additional finer scale studies are needed, such as the one by Sheng et al. (2020).

One of the most dramatic and damaging consequences of permafrost degradation is the loss of soil organic carbon due to thawing. In frozen soil, microbial activity is strongly inhibited, leading to organic carbon accumulation (Zimov et al., 2006). According to Mu et al. (2015), permafrost soils of the TP store about 160 +/− 87 Pg C, representing approximately 8.7% of that in the northern circumpolar permafrost region. Thawing of permafrost will release large parts of the carbon into the atmosphere as CO_2 and CH_4 (Schuur et al., 2015). This is likely the main reason for the projected soil organic carbon losses on the TP (Crowther et al., 2016). The gradual permafrost thawing due to active layer deepening is projected to make 1.07 to 2.60 Pg (RCP4.5) or 2.87 to 5.30 Pg (RCP8.5) soil organic carbon vulnerable to microbial decomposition by the end of the century, which could largely offset the regional carbon sink (Wang et al., 2020b). Soil organic carbon losses due to permafrost thaw are projected to be in the range of 15 to 69 Pg until 2020 and up to 3 m soil depth (Mu et al., 2015; Ding et al., 2016; Bosch et al., 2017; Jiang et al., 2019), amounting up to more than one-third of the total soil organic carbon stocks. However, there is significant spatial heterogeneity in the predicted soil organic carbon losses until 2070. Estimates range from <6.5 kg C m^{-2} to >24 kg C m^{-2} according to RCP2.6 projections (Bosch et al., 2017). This calls for a detailed analysis of possible soil organic carbon losses for different landscape units of the TP and how the feedback mechanisms with vegetation and land use will change.

5.3. Future pathways for biosphere change

The previous work summarized in section 2 and 3 identifies rapid bio- and geosphere changes over the TP in the past decades (e.g., Fig. 3). Looking into the future (2100), nearly all projections using GCMs show an overall warming and wetting; exceptions to this are noted in Su et al. (2013). Furthermore, although there are regional and seasonal differences, warming at high elevations is generally larger than at lower elevations (e.g., Hu et al., 2015). The above climate change scenarios (including those summarized in section 5.1) over the TP will impact, and interact with, biosphere changes. Biosphere changes on the TP have poorly understood climate feedbacks through interactions with the albedo, ground surface temperatures, cloud formation, transpiration, precipitation, aerosol production, and wind patterns (Babel et al., 2014). For example, changes in Kobresia (bog sedges) grassland pastures on the TP from pre-industrial conditions to the present lead to a degradation (thawing) of permafrost. Permafrost thawing leads to an increased thickness of the active permafrost layer, resulting in lower soil water availability relative to pre-industrial conditions when the permafrost table was shallower and characterized by a thinner active layer. A thicker active layer results in a deeper depth and less accessible available water for plants even though the overall amount of soil moisture is higher. However, some of the disadvantages of a deeper soil water reservoir are offset by CO_2 fertilization, enabling Kobresia to develop deeper roots to access water. Furthermore, a thicker active layer allows plants to exploit more nutrients released from the thawing permafrost (Schuur et al., 2007).

The high organic carbon and nutrient storage in the felty horizon of Kobresia grasslands (Kaiser et al., 2008) is a physiological by-product of Kobresia in response to grazing (Miehe et al., 2008; Hafner et al., 2012). This response allows for an extreme root-to-shoot ratio of these grasslands and a tight nutrient cycle to avoid losses of essential elements from the system (Schleuss et al., 2015). Soils under Kobresia pasture (and grassland in general) thus store much larger amounts of carbon and nutrients in the subsurface compared to many other (e.g., forest) ecosystems (Guo and Gifford, 2002). The result of the warmer, wetter, and nutrient-rich present-day setting supports the previously described (section 2.4) observations of a greening of the TP. However, recent work has also documented an increase in evapotranspiration (ET) over the plateau and the larger water losses from soil associated with permafrost thawing (e.g., Shen et al., 2015; Han et al., 2021). Increased ET results from increased vegetation on the plateau and evaporation of near-surface water available in the active layer. It is uncertain if these two components of the hydrologic cycle will eventually result in a future drying out of available near-surface waters that lead to unfavorable conditions for future greening.

In the following, we explain two potential pathways for the future of the TP geo-biosphere, both of which are supported by recent studies and result from interactions of the Earth's surface with the climate system (Fig. 9). Which path will be followed in the future largely depends on land-use adaptation strategies humans choose over the next decade.

5.3.1. Biosphere Pathway 1: continued greening

In this pathway, continued warming and wetting of the plateau will lead to a continuation of processes that we observed over the past decades (Fig. 9). This pathway is based on the assumptions that: (1) precipitation and humidity in (at least) the northern and southeastern parts of the TP will increase; (2) plant biomass will increase in response to a warmer climate, CO_2 fertilization, and a longer growing season. These changes will compensate for enhanced soil respiration; (3) the decay of the alpine vegetation due to permafrost degradation does not lead to increased soil desiccation and erosion; and (4) humans do not intensively interfere with the natural succession of vegetation.

In this case, alpine vegetation in the northern and southeastern parts of the TP will be replaced by forests (e.g., Bastin et al., 2019). If, as during previous wet periods in the LGM to mid-Holocene, the interior basins of the plateau experience lake expansion, then forest migration there may be impeded (e.g., Miehe et al., 2010). This greening and forest expansion pathway could occur if continued warming and wetting of the plateau leads to a succession of shrublands and forests that replace Kobresia pastures (Fig. 9). For this ecological transition to occur, the soil organic carbon and nutrient pools and high subsurface water content of degraded permafrost need to remain intact to provide a favorable

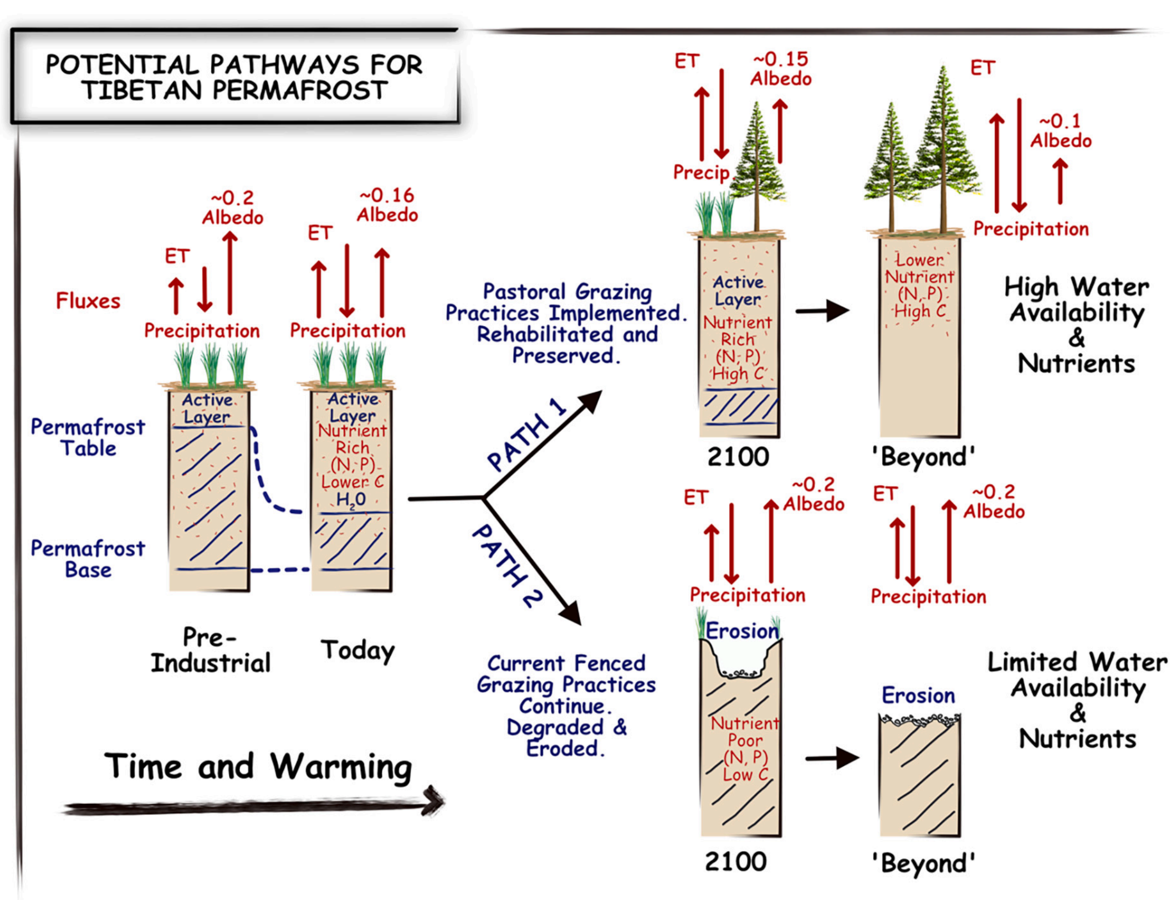

Fig. 9. Proposed mechanism and possible pathways for future biosphere response to permafrost thawing on the Tibetan Plateau. With future increases in temperature, the thickness of the active layer will increase as permafrost degrades. Pathway 1 leads to continued greening of the plateau if pastoral grazing practices are re-implemented. In contrast, pathway 2 leads to the desiccation and erosion of thawed permafrost and the loss of nutrients. Pathway 2 is likely to occur if existing grazing practices that rely on privately fenced pastures and reduced grazing radius are not changed. Current practices lead to overgrazing near permanent settlements and leave grazing resources in the remote parts unused. Which pathway occurs in the future will be determined by which human management strategies are implemented in the coming years. Red arrows and text indicate changes in the magnitude of fluxes in each scenario. ET denotes evapotranspiration. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

environment for continued greening. Continued CO_2 fertilization is also required to support the development of shrubs and forests with deeper root bases that can access water and new nutrient reservoirs supporting the increased biomass. The result of this pathway would be a TP with diverse (forest and shrub rich) ecosystems and Kobresia pastures that extend up to higher elevations that are currently bare. This pathway would transition the 'Third Pole Environment' characterized by glacial refugia to a 'Green Pole Environment.' An outcome of this pathway and ecosystem changes would be a continued increase in ET, an overall decrease in the albedo of the TP over forested areas, and a near-surface hydrologic scenario where forests access to water and nutrients from greater depths than Kobresia pastures. This pathway of TP greening is consistent with modeling studies (Piao et al., 2011, 2012) and with observations from the distant past (see section 4.).

5.3.2. Biosphere Pathway 2: degraded and eroded Kobresia pastures

This pathway is markedly different from the previous one. It suggests that although a 'greening' of the plateau is currently in progress, it is a temporary phenomenon, and a switch in primary processes and geo-biosphere interactions will occur sometime in the future (Fig. 9). This pathway is based on the assumptions that the biosphere will (in contrast to Pathway 1) experience large-scale degradation of the alpine vegetation if: (1) under a warmer climate and a higher atmospheric CO_2 concentration the primary productivity of plants increases but will not be able to compensate for enhanced soil mineralization (Chen et al., 2013; Han et al., 2019); (2) the gradual decay of the alpine vegetation will lead to increased soil desiccation and erosion (Wang et al., 2011; Li et al., 2015a) and (3) human activities such as private fenced grazing practices and water use further intensify the degradation and desertification (Hopping et al., 2018; Liu et al., 2018). More specifically, current grazing practices are problematic because they rely on privately fenced pastures with a reduced grazing radius than pastoral practices previously used on the plateau. The current practices lead to overgrazing near-permanent settlements and leave grazing resources in more remote parts unused. While the degradation suggested by pathway 2 occurs initially near settlements with fenced pastures, it is anticipated that it would spread to other areas with time as grazing would need to be relocated to more pristine settings, which would subsequently become degraded if the use of fenced pastures continues.

Thus, this pathway will occur if permafrost degradation and warming-induced soil moisture loss (Yang et al., 2018) and unsustainable grazing practices (Cao et al., 2017) lead to grassland degradation. Recent observations have documented that erosion of the nutrient and carbon-rich soil surface layer occurs on the TP, with the driving mechanism often being overgrazing of Kobresia pastures (Schleuss, 2017; Hopping et al., 2018; Liu et al., 2018). In this pathway, the removal of Kobresia by overgrazing or excessive water extraction results in lower vegetation cover and enhanced exposure of the fertile layer that is subsequently eroded (Fig. 9). Vegetation cover has been shown in recent observational and modeling studies to heavily impact catchment erosion rates through both stabilizing and destabilizing effects on surface processes (e.g., Schaller et al., 2018; Schmid et al., 2018; Werner et al., 2018; Starke et al., 2020; Sharma et al., 2021). In the case of TP Kobresia pastures, erosion of the surface layer would remove a large proportion of the ecosystems' nutrient pool (Liu et al., 2018) that would otherwise facilitate colonization of larger plants a reduction of the plant available soil water. This pathway would have a lower ET than the previous one because of lower biomass and the reduction of available soil moisture. These changes in ET would (amongst other factors) impact local meteorology and likely result in a modified precipitation pattern and a less active hydrologic cycle (Babel et al., 2014). However, whatever precipitation does occur, an unsustainable grazing regime would continue to have a high potential to facilitate carbon and nitrogen loss and erosion of the remnant Kobresia mats (He and Richards, 2015, 2017; Liu et al., 2018). Although Kobresia pastures might extend to higher elevations, their fate would similarly depend on land-use practices such as grazing (Miehe et al., 2019). The result of the previous processes would be a nutrient-poor and drier subsoil exposed at the surface with highly challenging conditions for colonization by vegetation. Thus, a starkly different TP would emerge in this scenario, and instead of rich shrublands and forests covering the plateau, a barren landscape would appear.

6. Management options and recommendations

The two pathways for the TP biosphere described in the previous section predict strongly contrasting outcomes (Fig. 9). Pathway 1 suggests a more positive effect for the TP, including a continued greening of the Third Pole into a Green Pole. In contrast, Pathway 2 would lead to a degraded TP environment with diverse socio-economic outcomes. One key element in determining one or the other pathway depends on human activity and associated land use.

If scenario 2 occurs, then slow vegetation recovery and continuously high anthropogenic pressure through grazing may lead to the disappearance of Tibetan pastures in some regions (Liu et al., 2018). Hence, proper and efficient pasture protection and restoration strategies must be implemented. However, the tendency in current practices is away from pasture protection. For example, in the last decades, mobile pastoralism has been considered inefficient in using rangeland ecosystems (Zhuang et al., 2019). Consequently, mobile pastoralism has mainly been replaced by individual contracted rangelands, which are divided and fenced. This approach has led to problems with overgrazing around villages and increased grassland degradation (He and Richards, 2015; Wang et al., 2018; Schleuss, 2017; Miehe et al., 2019). At the same time, grazing intensity is decreasing in remote areas. This leads to changing vegetation with a decrease of the grazing-adopted *Kobresia pygmaea* at the cost of grasses and shrubs (Miehe et al., 2019; Hopping et al., 2018). As a result, the below-ground carbon allocation has decreased (Hafner et al., 2012), likely leading to smaller soil organic carbon content and the destruction of the erosion-resistant root mat under Kobresia. If this situation continues, this will also reduce the foliage production for livestock. Continued warming will lead to more pressure on ungrazed grasslands than on grasslands under moderate grazing (Hopping et al., 2018). Zhuang et al. (2019) showed that the transition from community-based seasonal grazing to household-based continuous grazing increased the greenhouse gas emissions from -0.62 kg CO_2-eq kg^{-1} meat to 10.51 kg CO_2-eq kg^{-1}meat, thus highlighting the benefits of mobile pastoralism.

In summary, it is important to note that moderate grazing of the Kobresia grasslands is a prerequisite for its preservation. Overgrazing and too little grazing are detrimental for the survival of these critical ecosystems on the TP. To minimize environmental degradation and use the grasslands as a carbon sink, a community-based mobile pastoralism system with winter pastures should be advocated.

The potential future pathways for Kobresia pastures (section 4) highlight the multitude of interactions associated with geo-biosphere interactions in a sensitive landscape such as the TP. We presented the scientific basis for each pathway (sections 2 to 4) and suggest that we are currently on a threshold for influencing the future of the TP biosphere. Here we propose that which pathway occurs in the future can be controlled mainly by human behavior and that adaptation and management strategies can be implemented to determine the future of the TP biosphere. A continuation of current global warming trends and policy will result in a passive adaptation of the TP (i.e., 'business as usual') in terms of land and water use on the TP will lead us towards ecosystem degradation suggested by pathway 2.

However, passive adaptation can be changed to active adaptation, which means that intervention into the system through changes in environmental policy can be used to influence future change to maintain or re-establish a resilient TP ecosystem. If active adaptation is pursued, the ecosystem will be the winner through continued greening (pathway 1). But, active adaptation can have potential risks in the long term for the biosphere and society. Therefore, it is important to identify critical intervention points that stabilize the shift towards pathway 1, avoid potentially negative consequences, and increase the TP ecosystem system resilience. Based on experiences from current countermeasures such as setting up national parks, active interventions can be realized through four management approaches. First, a political and social reaffirmation is needed to make the paradigm of degradation and change towards pathway 2 possible. This requires a stable supply of ecosystem services to the TP and surrounding areas. Second, fragile TP ecosystems can be stabilized by reducing disturbances to them and stabilizing green infrastructure (Palmer et al., 2015; Huang et al., 2020). Third, for the already degraded ecosystems, stepwise ecological restoration is an important step to recovering a resilient ecosystem, and such restoration needs to consider future climate change (Liu et al., 2018). Finally (fourth), timely information on environmental change must be provided to decision-makers and stakeholders. Up-to-date information is needed to characterize better change that is presently occurring and for the evaluation of newly implemented adaptation strategies. Taken together, these four management approaches can promote pathway 1.

We recommend two 'initial step' adaptation strategies that are urgently needed to prevent future degradation of TP ecosystems (pathway 2) and facilitate continued greening (pathway 1, Fig. 9). First, improved land management strategies are needed. Land management strategies should consider different grazing practices, encourage traditional grazing practices that protect Kobresia mats and carbon stocks, and foster future expansion of Kobresia onto barren soil and rocks. Overgrazing has been shown (He and Richards, 2017; Liu et al., 2018) to foster the degradation of Kobresia mats, thereby leading to a transformation of the TP environment from a carbon sink to a carbon source. Implementing moderate grazing strategies would inhibit overgrazing and foster the eventual development of shrubs, trees, and forests in lower and more humid areas. Additional land management strategies should also include consideration of infrastructure development on the TP. Infrastructure such as roads, dams, and houses leads to increased desiccation and erosion, promoting pathway 2 (Fig. 9). Finally, land management strategies should also consider the promotion of nature protection through the designation of additional national parks or wilderness areas that promote pathway 1.

Second, adaptation strategies are needed to promote better water management on the TP. Human interventions that lower the water table or lead to desiccation will foster pathway 2. Practices such as water extraction via irrational unsustainable irrigation and forest plantations (Cao et al., 2017; Cao and Zhang, 2015) and construction of mega infrastructure (Wang et al., 2022; Li et al., 2022) should be avoided. In contrast, interventions that maintain a high-water table and promote precipitation and humidity will foster pathway 1. These interventions could include (for example) re−/afforestation and promotion of water protection areas.

7. Summary and conclusions

In this study, we demonstrate that recent temperature change across the TP, to the best of our knowledge, is occurring at a geologically unprecedented rate. Although evidence for past warm and wet conditions across the plateau is documented in geologic archives, the pace at which global warming occurred in the geologic past was significantly slower (e.g., over timescales of at least 10^3 yrs. when entering interglacials) compared to what is happening now. It is uncertain if the anthropogenically accelerated rate of warming leading up to 2100 and recent and ongoing anthropogenic practices for land and water management will lead the TP environment along the same pathway as past warming events or into new uncharted, territory.

We presented (section 6) two active adaptation strategies that are essential initial steps towards protecting the TP biosphere and fostering pathway 1 (Fig. 9) and a future greening of the TP. These adaptation strategies involve changes in land use and water use practices on the TP. What is not well known and warrants continued investigation is to what degree (i.e., over what spatial and temporal scales) these recommendations need to be implemented to assure a positive outcome. Although the concept of 'human engineering' of geo- and environmental systems is often seen with disdain, the issues highlighted in this study suggest that immediate human intervention on the TP will promote a favorable outcome for the biosphere and minimize undesirable consequences from current global change and human practices.

Author contributions

Conceptualization: T. Ehlers, D. Chen, V. Mosbrugger.

Writing draft: All authors contributed to writing individual sections in the original draft; T. Ehlers produced the complete draft within the framework of the manuscript concept.

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Writing review: T. Ehlers, D. Chen.

Funding Acquisition: T. Yao, V. Mosbrugger.

Data availability

All data used in this study are published, and the sources of the data are cited throughout the text.

Declaration of Competing Interest

The authors are aware of no financial or other conflicts of interest.

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