Permafrost Degradation Diminishes Terrestrial Ecosystem Carbon Sequestration Capacity on the Qinghai-Tibetan Plateau

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Abstract

Effects of permafrost degradation on carbon (C) and nitrogen (N) cycling on the Qinghai-Tibetan Plateau (QTP) have rarely been analyzed. This study used a revised process-based biogeochemical model to quantify the effects in the region during the 21st century. We found that permafrost degradation would expose 0.98±0.49 (mean±SD) and 2.17±0.38 Pg C of soil organic carbon under the representative concentration pathway (RCP) 4.5 and the RCP 8.5, respectively. Among them about 60% will be decomposed, enhancing heterotrophic respiration by 9.54±5.20 (RCP 4.5) and 38.72±17.49 (RCP 8.5) Tg C/yr in 2099. Deep soil N supply due to thawing permafrost is not accessible to plants, providing limited benefits to plant growth and only stimulating net primary production by 6.95±5.28 (RCP 4.5) and 27.97±12.82 (RCP 8.5) Tg C/yr in 2099. As a result, permafrost degradation would weaken the regional C sink (net ecosystem production) by 303.55±254.80 (RCP 4.5) and 518.43±234.04 (RCP 8.5) Tg C cumulatively during 2020–2099. Permafrost degradation has a higher influence on C balance of alpine meadow than alpine steppe ecosystems on the QTP. The shallower active layer, higher soil C and N stocks, and wetter environment in alpine meadow are responsible for its stronger response of C balance to permafrost thaw. This study highlights that permafrost degradation could continue to release large amounts of C to the atmosphere irrespective of potentially more nitrogen available from deep soils.

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Key Points:

- Soil carbon release due to permafrost degradation overwhelms net primary production on the Qinghai-Tibetan Plateau.
- Deep soil nitrogen addition from thawing permafrost has a limited benefit to plant carbon uptake.
- Permafrost degradation has more pronounced influence on carbon cycling of alpine meadow than alpine steppe.

Abstract

Effects of permafrost degradation on carbon and nitrogen cycling on the Qinghai-Tibetan Plateau (QTP) have rarely been analyzed. This study used a revised process-based biogeochemical model to quantify the effects in the region during the 21st century. We found that permafrost degradation would expose 0.98±0.49 (mean±SD) and 2.17±0.38 Pg C of soil organic carbon under the representative concentration pathway (RCP) 4.5 and the RCP 8.5, respectively. Among them about 60% will be decomposed, enhancing heterotrophic respiration by 9.54±5.20 (RCP 4.5) and 38.72±17.49 (RCP 8.5) Tg C/yr in 2099. Deep soil nitrogen (N) supply due to thawing permafrost is not accessible to plants, providing limited benefits to plant growth and only stimulating net primary production by 6.95±5.28 (RCP 4.5) and 27.97±12.82 (RCP 8.5) Tg C/yr in 2099. As a result, permafrost degradation would weaken the regional carbon sink (net ecosystem production) by 303.55±254.80 (RCP 4.5) and 518.43±234.04 (RCP 8.5) Tg C cumulatively during 2020–2099. Permafrost degradation has a higher influence on C balance of alpine meadow than alpine steppe ecosystems on the QTP. The shallower active layer, higher soil C and N stocks, and wetter environment in alpine meadow are responsible for its stronger response of C balance to permafrost thaw. This study highlights that permafrost degradation could continue to release large amounts of carbon to the atmosphere irrespective of potentially more nitrogen available from deep soils.

1 Introduction

Permafrost degradation due to climate warming makes large amounts of frozen soil organic matter (SOM) available for decomposition, leading to a positive feedback to the climate system by releasing greenhouse gases (GHG) (Koven et al., 2011; Koven, Schuur et al., 2015; Schaefer et al., 2014; Schuur et al., 2015). Enhanced decomposition in permafrost regions also increases nutrient availability however (Keuper et al., 2012; Salmon et al., 2016, 2018), stimulating plant production together with the effects of increasing air temperature, expanding growing season, shifting plant community composition, and rising atmospheric CO_2 concentration, resulting in a negative feedback to climate warming (Finger et al., 2016; Liang et al., 2018; Peng, X. et al., 2020; Zhuang et al., 2010). Considerable uncertainty of these positive and negative feedbacks limits our ability to model C balance of the permafrost region (Abbott et al., 2016).

Coupling the impact of permafrost degradation into land surface models has projected that the northern high latitudes would shift from a net C sink (Qian et al., 2010) to a net source to the atmosphere (Burke et al., 2013; Koven et al., 2011; MacDougall et al., 2016; Schaefer et al., 2014; Schaphoff et al., 2013). However, large uncertainties still exist in the magnitude of the estimated changes in soil, vegetation, and ecosystem C stocks (McGuire et al., 2016). Model structural differences, particularly in the soil carbon decomposition processes, have been attributed to be the largest uncertainty in quantifying permafrost carbon-climate feedbacks (Burke et al., 2017).

A critical model structural feature to represent the permafrost carbon-climate feedback is to consider soil C exposure with thaw depth (McGuire et al., 2018). In permafrost regions, part of soil organic carbon (SOC) is protected from decomposition by frozen soil but may become more susceptible to decomposition with warming-induced degradation of permafrost. To better account for the influence of permafrost dynamics on terrestrial carbon balance, soil C exposure due to thawing permafrost should be incorporated into land surface models. Hayes et al. (2014)

assessed the net effect of active layer dynamics on permafrost carbon feedback during 1970–2006 over the circumpolar permafrost region and found that permafrost degradation exposed 11.6 Pg C of thawed SOM to decomposition, resulting in 4.03 Pg C releasing from the decomposition of newly exposed SOM, but only 0.3 Pg C was compensated by net primary production (NPP). Using a carbon-nitrogen model that includes permafrost processes and depth-dependent changes in SOM, Koven et al. (2015a) found similar results and projected that permafrost carbon-climate feedback is sensitive to deep soil carbon decomposability, and the impact of deep soil nitrogen (N) mineralization on C budget is small. However, projections of five biogeochemical models that represent soil C explicitly with depth show that the northern permafrost region would likely act as a net C sink before 2100 due to stronger vegetation C uptake (McGuire et al., 2018). The different result found by McGuire et al. (2018) maybe because they only considered the active layer thickness (ALT) less than three meters. The decomposition of large quantities of deep carbon deposits (Schuur et al., 2015) may reverse this estimation.

The Qinghai-Tibetan Plateau (QTP) is characterized by a deep ALT (2.34±0.70 m), and about 15% of the QTP permafrost region has ALT greater than 3 m (Wang, T. et al., 2020). SOC storage on the QTP is estimated at 50.43 Pg, of which 35.10 Pg stores below 3 m soils and 37.21 Pg frozen in permafrost currently (Wang, T. et al., 2020). Based on the vertical distribution of SOC and the change of ALT, Wang, T. et al. (2020) estimated that permafrost thaw on the QTP would expose 1.86±0.49 and 3.80±0.76 Pg frozen C to decomposition, which could potentially turn the region from a net C sink to a net source. The lack of the simulation of SOM decomposition, and the use of net biome production (NBP) from CMIP5 models (which did not consider permafrost carbon (Jones et al., 2016) and hence the stimulation of N supply to vegetation production due to permafrost thaw), may bias the estimation of net C budget of Wang, T. et al. (2020). Through repeated soil carbon measurements on the QTP, Ding et al. (2017) suggested that the upper active layer of the QTP permafrost currently represents a substantial regional soil C sink, probably owing to enhanced vegetation growth. However, this study only examined SOC stocks in uppermost 30 cm. Carbon changes associated with deeper and older permafrost remain largely unknown (Mu et al., 2020). Inadequate studies that focus on representing soil C explicitly with depth, particularly representing soil C in deep soils, and simultaneously consider the feedback of vegetation to permafrost thaw impedes our understanding of permafrost carbon-climate feedbacks on the QTP.

Apart from deep ALT, the environment of the permafrost regions on the QTP is also different from the pan-Arctic in water condition. Specifically, annual precipitation decreases from southeastern to northwestern on the QTP. As a result, the eastern QTP receives more precipitation and dominated by alpine meadow ecosystem, while the widespread inner and western parts are controlled by alpine dry climate and dominated by alpine steppe ecosystems. Many studies have found that the alpine steppe and alpine meadow ecosystems respond differently to climate change (Hao et al., 2021; Li, S. et al., 2019; Liu et al., 2020; Peng, F. et al., 2020). For example, warming experiment revealed that warming increased plant productivity in alpine meadow but decreased productivity in alpine steppe (Ganjurjav et al., 2016). Temperature sensitivity of ecosystem respiration (Q10) in alpine meadow (3.4) can be twice the number in alpine steppe (1.7) (Wang, L. et al., 2018). The distinct environment and the different responses to climate change of these two ecosystems could result in different permafrost C feedbacks. However, currently the effects of active layer deepening on C balance of these two ecosystems are still not certain.

In this study, a process-based biogeochemical model was revised by coupling thaw depth with soil C exposure to analyze the following issues: (*i*) the quantity of frozen SOC in deep soils on the QTP that will be exposed to decomposition due to permafrost thaw in the 21st century, and the quantity of C that will be released into the atmosphere; (*ii*) the stimulation of deep permafrost thaw on vegetation productivity due to extra N supply and the net effect of permafrost thaw on ecosystem C balance (enhanced heterotrophic respiration (RH) versus stimulated net primary production (NPP)); and (*iii*) the different responses of C balance of the distinct alpine meadow and alpine steppe to permafrost thaw.

2 Methods and Data

2.1 The Terrestrial Ecosystem Model

The Terrestrial Ecosystem Model (TEM) is a process-based, global scale biogeochemical model, using spatially explicit soil, vegetation, and elevation data and climate forcing of radiation, precipitation, and air temperature data to simulate ecosystem C and N cycling (McGuire et al., 1992; Zhuang et al., 2002). A soil thermal model (STM) has been coupled into TEM to represent vertical soil thermal profile in permafrost- and non-permafrost-dominated ecosystems and gives TEM the ability to describe freeze-thaw cycles in cold regions (Jin et al., 2015; Zhuang et al., 2001, 2010). The effects of freeze-thaw dynamics on gross primary production (GPP) have also been considered in TEM (Zhuang et al., 2011). TEM has been extensively used to evaluate C dynamics in northern high latitudes and Tibet plateau (e.g., Hayes et al., 2014; Jin et al., 2013, 2015; Kicklighter et al., 2019; McGuire et al., 2018).

In previous versions of TEM, soil organic carbon (SOC) in the entire root zone is assumed to be available for decomposition, and thus take part in heterotrophic respiration. However, since active layer thicknesses (ALT) in permafrost regions are often shallower than root depth defined in TEM (Hayes et al., 2011, 2014), part of SOC in root zone is frozen in soil and protected from microbial decomposition. Therefore, SOC considered in heterotrophic respiration quantification is smaller than the entire SOC stock in root zone. TEM6 (Hayes et al., 2011, 2014) has considered the amount of SOC available for decomposition and treated it as a proportion the entire root zone SOC, depending upon the ratio of ALT to rooting depth. By calculating ALT variation over time through the STM module in TEM, this approach has been used to estimate the potential influence of permafrost thaw on the availability of soil organic matter to decomposition (Hayes et al., 2014; Kicklighter et al., 2019). However, this approach did not consider SOC below the rooting zone and assumed that they do not contribute to C flux. When warming continues in the future and ALT deepens below rooting depth, this approach may underestimate the influence of permafrost degradation on the availability of carbon to decomposition (Hayes et al., 2011).

On the QTP, ALT (2.34 m; Wang, T. et al., 2020) is usually much deeper than rooting depth (over 90% of plant roots are distributed within the top 30 cm soil layer; Yang, Y. et al., 2009). Future permafrost degradation would probably happen far below rooting zone. Therefore, we modified the approach by adjusting SOC according to the ratio of ALT of current time step to previous time step, rather than the ratio of ALT to rooting zone depth:

$$SOC_i = SOC_{i-1} * \left(1 + \frac{prop_i - prop_{i-1}}{prop_{i-1}}\right) \quad (1)$$

$$prop_i = \frac{a * ALT_i}{b + ALT_i} \quad (2)$$

where i and i-1 denote the current and previous year, respectively; ALT_i is the maximum ALT of the current year; $prop_i$ denotes the proportion of SOC in the active layer of current year to the total SOC in the entire soil profile. The hyperbolic function of $prop_i$ describes the vertical distribution of SOC in soil profile, with high density of SOC in shallow soil layers but decreases towards deep soil, as documented by Hayes et al. (2014). a and b are parameters controlling the shape of the curve that how SOC density changes with soil depth. We estimated parameter b for alpine meadow, alpine steppe, and alpine desert using observation data (Figure S1). For the remaining ecosystems, we set parameters b according to Kicklighter et al. (2019). Parameter values of b for various ecosystems are listed in Table S1. Parameter a will be divided when put eq. (2) into eq. (1).

Using this method, SOC within the soil layers that permafrost degraded was factored into total available SOC pools for soil decomposition. Although we considered the fact that SOC density decreases with soil depth, the vertical variations in SOC turnover time are still missing in this study, which may overestimate SOC decomposition rate to some extent (Shu et al., 2020).

Permafrost degradation influences not only the amount of SOC, but also soil organic N (SON), available inorganic N (AVALN), and soil water content. These variables were also adjusted according to the variation of ALT, in the same method as that for SOC. Although permafrost degradation can increase AVALN pools, if ALT deepens far below rooting depth, we assume that the additional AVALN cannot be accessed by vegetation. Therefore, only AVALN in rooting zone was used for vegetation N uptake (VegNup) calculation when ALT is greater than rooting zone depth.

Two freezing fronts are modeled in TEM, including freezing down due to cold air temperature and freezing up due to permafrost underneath the soils (Zhuang et al., 2001). ALT in permafrost regions is calculated as the total depth of unfrozen soil layers above the two freezing fronts. TEM does not explicitly ‘define’ permafrost occurrence in a grid cell, but infer its depth based on the 0 °C isotherm in the soil thermal profile. The permafrost region of this study was derived from permafrost map of China (Wang, 2019).

2.2 Model calibration and validation

TEM has been calibrated for terrestrial ecosystems on the QTP by Zhuang et al. (2010) and Jin et al. (2013, 2015). In this study, some key rate-limiting parameters (such as the maximum rate of photosynthesis, heterotrophic respiration, and plant N uptake and respiration rate) were recalibrated for three major ecosystems (i.e., alpine shrubland, alpine meadow, and alpine steppe), which occupy 82.6% of the permafrost regions on the plateau according to the 1:1000000 vegetation map of China (Figure 1; Chinese Academy of Sciences, 2001). The parameters were optimized by minimizing the differences between observations and simulations through altering parameters and iterating model simulations. The rest of the model parameters were set as previous studies. Site information of these three ecosystems is provided in Table S2. Observations of soil temperature (DST), net ecosystem production (NEP), and soil C and N stocks for each site are derived from references in Table S2 accordingly. The model was validated with 289 observations of soil C and N pools (data derived from Ding et al. (2016), Kou et al. (2019), and Wang et al. (2020)) on the plateau. These observation sites spread over alpine desert, alpine steppe, alpine meadow, and alpine cushion ecosystems. However, for the analysis of the effects of permafrost degradation, only the dominant ecosystems of alpine steppe and alpine meadow were focused on.

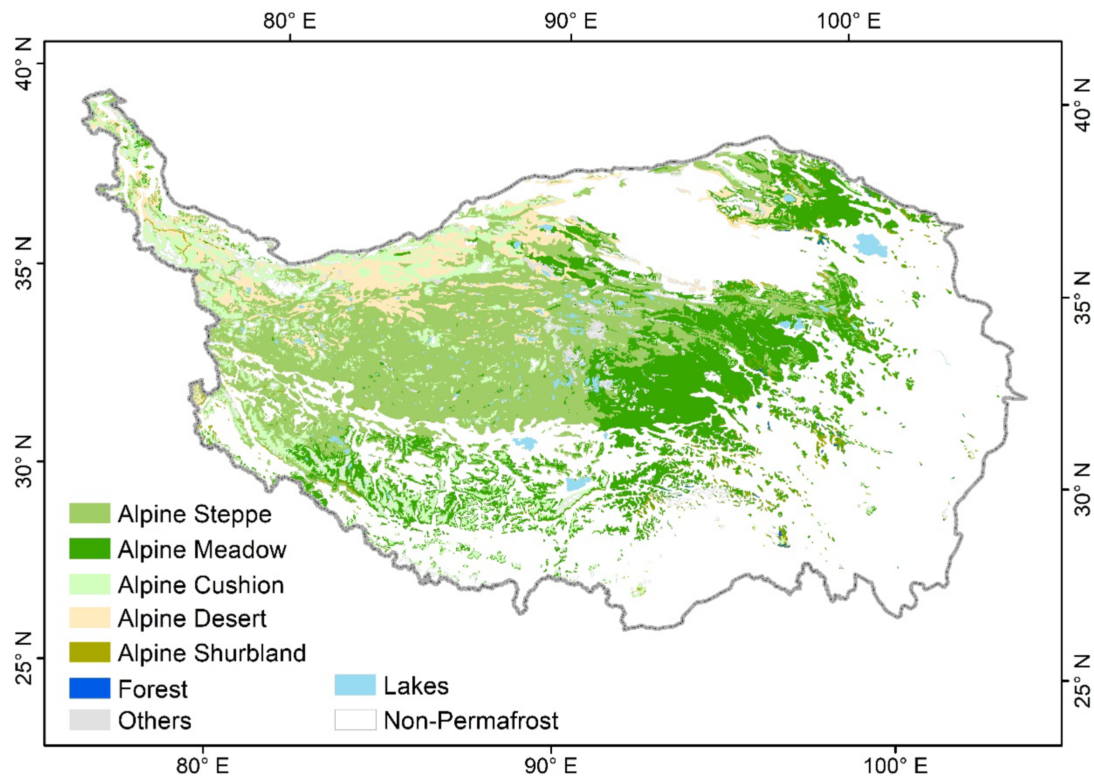

Figure 1. Vegetation distribution in the permafrost regions of the QTP. The vegetation distribution was derived from a 1:1,000,000 vegetation map of China (Editorial Committee of Vegetation Map of China, Chinese Academy of Sciences, 2001) provided by the Resources and Environment Science Data Center, Chinese Academy of Sciences. The permafrost map (all the colored patches exclude lakes) was obtained from frozen ground map of China (Wang, T., 2019).

2.3 Simulation protocol

To determine the direct effects of ALT deepening on ecosystem C dynamics, we carried out two simulations: the referenced simulation (S_R) and the transient simulation (S_T). The two simulations share the same model inputs but have differences in model structure. They are identical until 2020. After that, ALT changes with climate in S_T but remained no change in S_R. Therefore, C and N pools do not change with ALT in S_R after 2020 but does influenced by ALT in S_T. Through comparing the results of these two simulations, the direct effects of ALT changes on ecosystem C dynamics can be estimated.

TEM was spun up for 200 years in both simulations. During the spin-up procedure, climate data from 2006 to 2046 were used to force the model run and repeated continuously until dynamic equilibrium was achieved. The state variables of the dynamic equilibrium were then used as the initial value for simulations from 2006 to 2099. The spin-up procedures were identical for the S_R and S_T.

2.4 Data

TEM was driven with spatially referenced information on climate, soils, vegetation, and elevation for spatial extrapolation.

Climate data from 2006 to 2099 include air temperature (°C), precipitation (mm), and incident short-wave solar radiation (W/m^2), deriving from four global circulation models (GCMs: IPSL-CM5A-LR, GFDL-ESM2M, MIROC5, and HadGEM2-ES) in the second simulation round of the Inter-Sectoral Impact Model Intercomparison Project (ISI-MIP 2b). Known issues of previous round of ISIMIP have been solved for the ISIMIP2b through a series of adjustments, and atmospheric data provided by the ISIMIP2b have also been bias-adjusted to a new reference dataset of EWEMBI (Frieler et al., 2017). Regional simulation of this century was conducted under two climate scenarios of the Representative Concentration Pathway 4.5 (RCP 4.5) and RCP 8.5, which correspond to radiative forcing levels of 4.5 and 8.5 W/m^2 by 2100, respectively (Moss et al., 2010).

Soil texture data over the QTP were based on the Regridded Harmonized World Soil Database v 1.2 (Wieder, 2014). Vegetation and elevation data were derived from the 1:1000000 vegetation map of China (Chinese Academy of Sciences, 2001) and the Shuttle Radar Topography Mission (SRTM; Farr et al., 2007), respectively.

The spatial resolutions of soil texture and elevation data are 0.05 degree and 90 m, respectively. Compared with these land surface datasets, the climate data are much coarser, at the spatial resolution of 0.5 degree. On the QTP, topography, vegetation and soil may change greatly within 0.5-degree spatial resolution. To consider the effects of heterogeneities of these data, we resampled them into a spatial resolution of 0.1 degree. During resampling, we make every 25 grids of the resampled climate data share the same value as the grid of the same location of the original climate data. We did not adjust climate data according to elevation, which may introduce uncertainty. TEM is run on monthly time scale. We averaged the daily climate data to monthly before resampling.

3. Results

3.1 Model validation

We validated TEM with monthly observation data of DST and NEP for three dominant ecosystems on the QTP (Figure 1). TEM well reproduced the seasonal variation of DST and NEP at all the three sites, with R^2 greater than 0.9 and 0.8 for DST and NEP, respectively. Simulated SOC and SON at these three sites are closed to observations (Table S1).

Apart from these three sites, we also compared the simulated SOC and SON with 289 observations on the Plateau (Figure 2 and Figure S2). The discrepancies between modeled and observed data (Figure S2) for SOC and SON may be due to the following reasons. First, the calculated SOC and SON in this study are C and N stocks in active layers, while the observations represent C and N stocks in top two meters (top three meters for 173 sites for SON; Kou et al., 2019) of soil profile. The mean ALT on the QTP is estimated at about 2.3 m during 2006–2015. By comparison, ALT is shallower than 2 m in the northwestern, eastern, and southern QTP (Ni et al., 2021; Wang, T. et al., 2020). According to Ding et al. (2016), Kou et al. (2019), and Wang et al. (2020), about half of the observation sites are distributed in the regions where ALT is estimated less than 2 m. As a result, the simulated SOC and SON in active layer is smaller than observations (Figure 2). Second, although the heterogeneities in topography, vegetation, and soils were considered by running the model on 0.1 × 0.1 degree of spatial resolution, the great spatial heterogeneities in SOC (Mishra et al., 2021) and SON make it difficult to predict site level observations based on large grid simulations.

In general, the simulated SOC and SON are comparable to observations based on the means across vegetation types (Figure 2). For the entire permafrost regions on the QTP, the estimated SOC stocks in the QTP active layers during 2006–2015 under the RCP 4.5 scenario is 15.64±0.78 Pg C (mean ± SD), a little higher than 13.22 Pg C estimated by Wang, T. et al. (2020), but lower than 19.0±6.6 Pg C (in 0–2 m soils) estimated by Mu et al. (2020). The simulated ALT during 2006–2015 (2.72±0.07 m) is close to but a little higher than 2.34±0.70 m estimated by Wang, T. et al. (2020). The climate forcing data of RCP 4.5 scenario other than meteorological observations might cause this difference. SON stocks in alpine grassland active layers during 2013–2014 are estimated at 1.47±0.04 kg N/m^2, which is within the range (1.40–1.76 kg N/m^2) estimated based on observation data within 0–3 m soil depth (Kou et al., 2019). Estimated NPP in the whole QTP is 240.99±14.27 g C /m^2/yr, which is close to the upper bound of the MODIS normalized difference vegetation index (NDVI) based estimate of 219.8–242.5 g C /m^2/yr during 2002–2012 (Wang, S. et al., 2017). The use of RCP 4.5 data and the different time spans may be responsible for the higher estimate in this study.

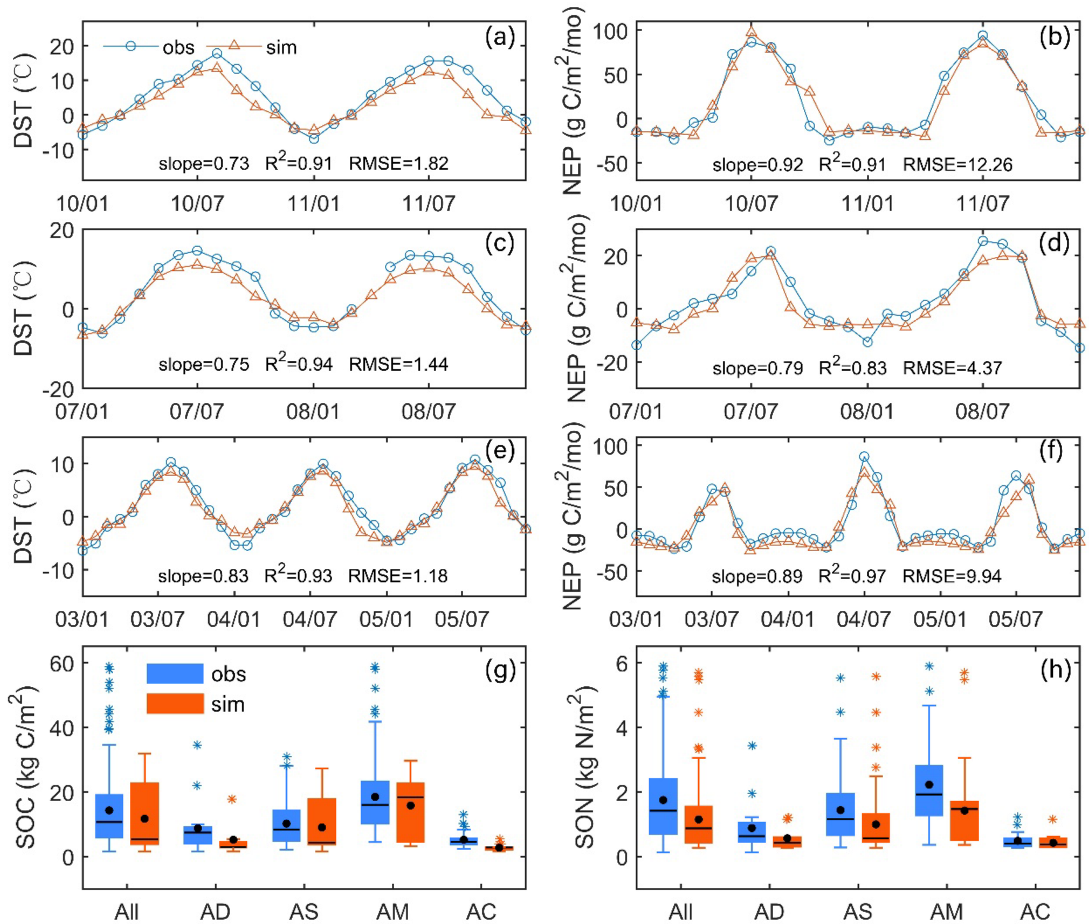

Figure 2. Comparison of model simulation (sim) with observations (obs) for soil temperature (DST) and NEP at alpine meadow (**a** and **b**), alpine steppe (**c** and **d**), and alpine shrub (**e** and **f**) sites, respectively. See Table S1 for detailed information of the observation sites. **g** and **h** represent the comparison of the simulated SOC and SON with 289 observations on the Plateau. AD, AS, AM, and AC in **g** and **h** denote alpine desert, alpine steppe, alpine meadow, and alpine cushion, respectively. The number of observation sites for AD, AS, AM, and AC are 16, 134, 106, and 33, respectively.

3.2 Changes in ecosystem C balance and the effects of ALT deepening

Permafrost regions on the QTP are projected to be a stronger C sink during the remaining of this century. VegC and SOC both increase under the RCP 4.5 and RCP 8.5 (Figure 3, S_T). As a result, by the end of this century, ecosystem C stocks (total VegC and SOC) would increase by 2.68±0.22 Pg C (0.17±0.02 Pg C gains from VegC and 2.51±0.26 Pg C gains from SOC) and 5.61±0.36 Pg C (0.44±0.08 Pg C gains from VegC and 5.18±0.35 Pg C gains from SOC), respectively, under the RCP 4.5 and RCP 8.5 scenarios. Under the RCP 4.5, the increases in NPP and RH are close, which are 53.99±24.58 Tg C/yr and 48.96±4.19 Tg C/yr, respectively. However, under the RCP 8.5, the increase in NPP (157.93±21.08 Tg C/yr) is much higher than RH (104.37±22.30 Tg C/yr). The permafrost regions on the QTP would sequestrate more C under the RCP 8.5.

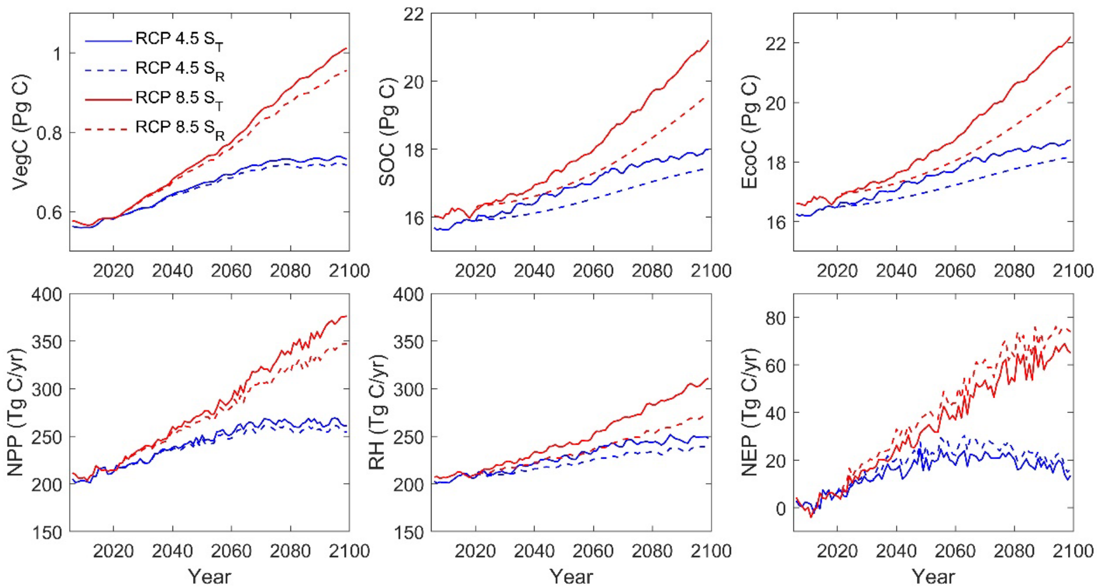

Figure 3. Changes in C pools (VegC, SOC, and EcoC) and fluxes (NPP, RH, and NEP) of the permafrost regions on the QTP in the 21st century. Differences between the results of the transient simulation (S_T) and referenced simulation (S_R) represent the direct effects of ALT. Standard deviation for each simulation did not show for clarity.

The effects of ALT deepening on C balance are represented by the difference between S_T and S_R. Figure 3 shows that ALT deepening have evident effects on C balance. We calculated the contribution of ALT deepening effects to C balance change by dividing the difference in S_T and S_R from the overall C balance changes (Figure 4b). Under the RCP 4.5, ALT deepening can contribute 9–18% of the C balance changes, and the contributions increase to 12%–37% under the RCP 8.5. The radical change in ALT (Figure 4a) under the RCP 8.5 results in a greater contribution. Under both RCPs, the deepening of ALT has more influences on SOC and RH than VegC and NPP (Figure 3 and Figure 4b).

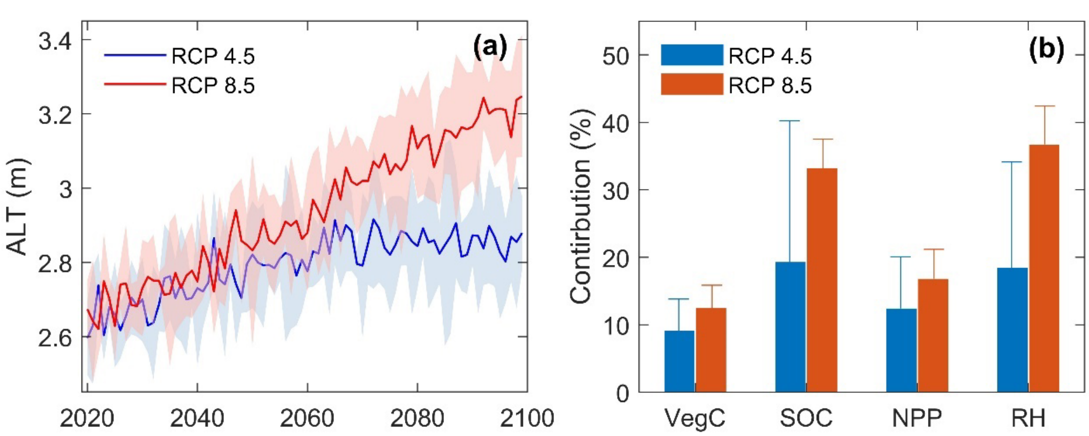

Figure 4. (a) Changes in mean ALT of the QTP under the RCP 4.5 and RCP 8.5 scenarios; **(b)** Contributions of ALT deepening effects to the overall C balance changes caused by climate change in permafrost regions of QTP during the 21st century. The shade in **(a)** and error bars in **(b)** denotes the standard deviation among all GCMs.

The greater values of S_T than S_R (Figure 3) under both RCPs suggest that ALT deepening increases ecosystem C storage in the region. However, it should be noted that the increasing in RH (9.54±5.20 (RCP 4.5) and 38.72±17.49 (RCP 8.5) Tg C/yr) is greater than NPP (6.95±5.28 (RCP 4.5) and 27.97±12.82 (RCP 8.5) Tg C/yr) due to ALT deepening (Figure 3 and Figure S3), resulting in a smaller NEP in S_T than S_R. This result illustrates that, although NPP increases more than RH in the S_T simulation (Figure 3), resulting in a stronger C sequestration over time, permafrost degradation itself strengthens RH more than NPP and cumulatively decrease NEP by 303.55±254.80 (RCP 4.5) and 518.43±234.04 (RCP 8.5) Tg C during 2020–2099.

3.3 Responses of C balance to ALT deepening in different ecosystems

The effects of ALT deepening on C balance of the QTP show an obvious spatial heterogeneity (Figure 5). The east and south parts of the QTP would increase more in both C pools and fluxes than the northwestern part due to ALT deepening, and the spatial heterogeneity is more noticeable under the RCP 8.5 than RCP 4.5. Given that east and south parts of the QTP are dominated by alpine meadow ecosystems and the northwestern part is dominated by alpine steppe ecosystems (Figure 1), these results suggest that ALT deepening has greater influence on C balance of alpine meadow than alpine steppe, particularly under the RCP 8.5. Moreover, changes in regional average ALT of these two ecosystems (Figure 6) show that increase in ALT is smaller in alpine meadow. However, C storage and fluxes of alpine meadow respond more significantly to ALT deepening.

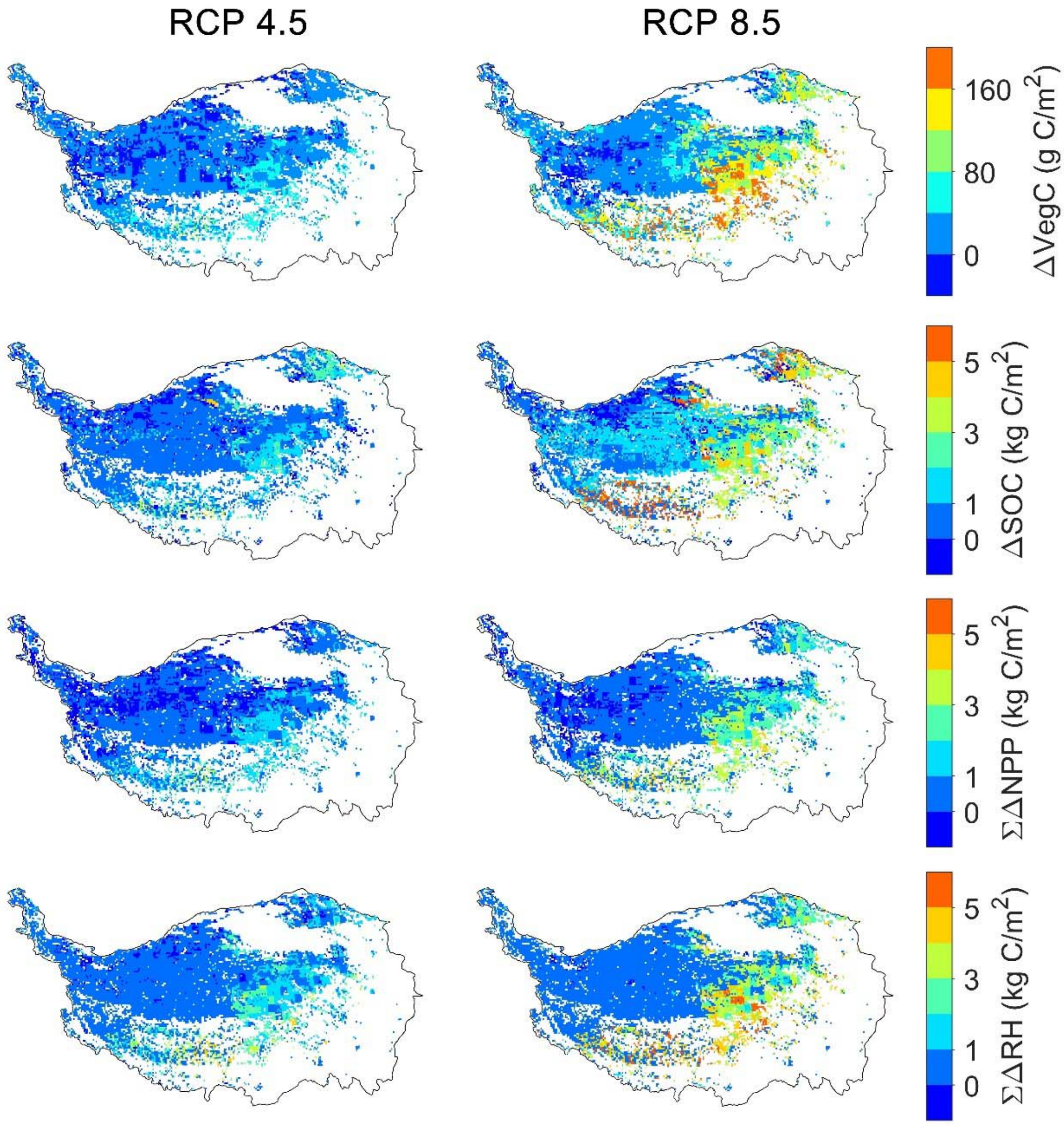

Figure 5. Changes in C balance of the QTP due to permafrost degradation under the RCP 4.5 and RCP 8.5. Δ before VegC and SOC mean the difference between the results of the transient simulation and referenced simulation (S_T-S_R) in 2099, and ∑ denotes the accumulation of each C flux from 2020 to 2099.

Under both RCPs and both ecosystems, the increasing in RH is greater than NPP (Figure 6), which is consistent with the results of the whole permafrost region (Figure 3), suggesting again that the direct effects of ALT deepening would weaken C sequestration. When comparing the two types of ecosystems, the C sequestration weakening tends to be stronger in alpine steppe ecosystems. The cumulative difference in NEP between S_T and S_R from 2020 to 2099 are -122.56±112.45 and -236.94±139.55 Tg C for alpine meadow under the RCP 4.5 and RCP 8.5, respectively, and -172.65±168.75 and -247.32±97.27 Tg C for alpine steppe under the RCP 4.5 and RCP 8.5, respectively. Given that the increasing in ALT in alpine steppe is about twice as much as that in alpine meadow (Figure 6), the C sequestration could reduce more in alpine meadow than alpine steppe with the same degree of ALT deepening.

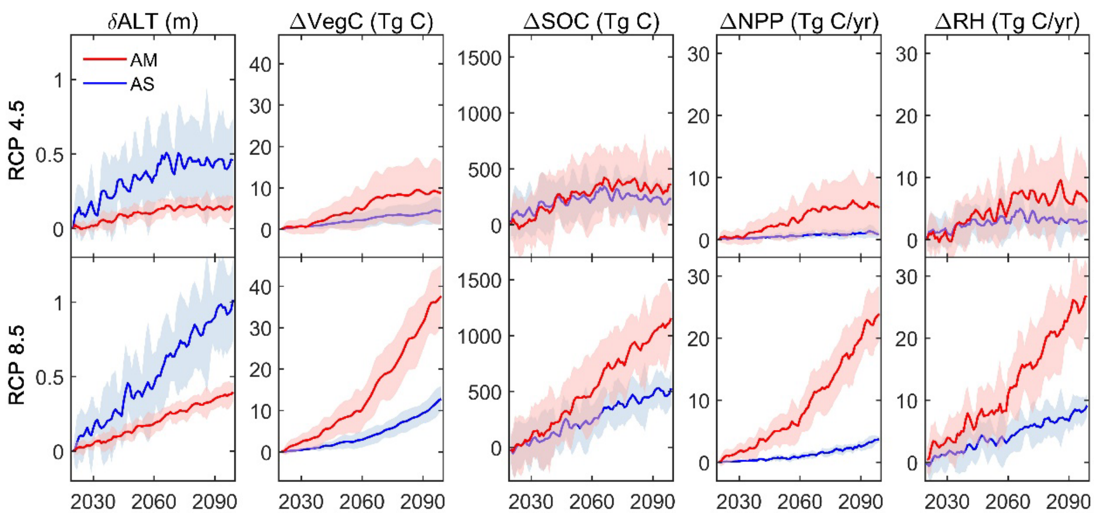

Figure 6. Anomalies of ALT from 2020 baseline and the effects of ALT deepening on ecosystem C balance of alpine meadow (AM) and alpine steppe (AS) ecosystems under the RCP 4.5 and RCP 8.5 scenarios. δ before ALT means the anomalies of ALT from 2020 baseline, and Δ before C balance components mean the difference between the results of the transient simulation and referenced simulation (S_T-S_R). Red and blue shade denotes the standard deviation among all GCMs for alpine meadow and alpine steppe ecosystems, respectively.

4. Discussion

4.1 Comparison with other studies

The estimated NPP and RH of the permafrost region on the QTP are 202.5 and 201.1 g C/m^2/yr during 2006–2011, both within the range of other estimates (120.8–329.6 g C/m^2/yr for NPP and 149.8–303.5 g C/m^2/yr for RH; Table 1). However, our NEP estimates are smaller than others. Particularly, our estimates with this version of TEM tends to be in low end of other TEM estimates, which is most likely due to more respiration from deep active layer in this study. In addition, this study focused on the permafrost region on the QTP, while Zhuang et al. (2010) and Jin et al. (2015) studied the entire QTP, and Yan et al. (2015) focused on grasslands on the QTP. Different study areas and ecosystem types considered in these studies may contribute to the discrepancies.

Existing studies show large differences in future carbon dynamics in the region (Table 1). Our estimation of flux density is higher than Jin et al. (2015) but smaller than Bosch et al. (2017). RH estimation of Bosch et al. (2017) depends only on annual precipitation, which may induce large uncertainties in their future projection because important factors like soil temperature and SOC quality and quantity were not considered. Thawing-induced CO_2 emissions are also considered in the study of Bosch et al. (2017). However, the potential C loss from thawing permafrost is estimated from incubation experiments with soil samples from the arctic region. SOC density on the QTP (14.4–17.5 kg/m^2 for alpine meadow and 6.6–7.7 kg/m^2 for alpine steppe in 0–2 m; Mu et al., 2020) can be substantially lower than that in the arctic (55.1±18.9 kg/m^2 for lowlands and 40.6±22.7 kg/m^2 for uplands in upper 1 m layer of the North American Arctic; Ping et al., 2008), which may cause an overestimate of RH in Bosch et al. (2017). The projections of Jin et al (2015) are lower for NPP and NEP in the 2090s under the RCP 8.5. TEM used in Jin et al (2015) was mainly calibrated for wetland ecosystems. Under the high warming scenario of the RCP 8.5, warming induced high evapotranspiration may result in water stress and suppress NPP for wetland ecosystems. In this study, TEM has been calibrated with the main ecosystems on the QTP, including alpine meadow, alpine steppe, alpine shrubland, and alpine desert. Different calibrations should partially contribute to the differences in those projections.

Table 1. Estimated regional C fluxes on the QTP from different studies. Units for regional total C fluxes are Tg C/yr, and units for regional mean flux density are g C/m^2/yr.

Model	Area (10^6 km^2)	Period	NPP		RH		NEP		Source
			Total	Density	Total	Density	Total	Density	
CENTURY	1.41	1901–2010	344.0	244.7	336.5	239.0	7.5	5.3	Lin et al. (2017)
TEM	1.01	1961–2010	193.0	193.7	182.9	183.7	10.1	10.0	Yan et al. (2015)
ORCHIDEE	1.39	1961–2009	219.8	158.1	208.2	149.8	11.6	8.3	Piao et al. (2012)
CASA	1.47	1982–2009	177.2	120.8					Zhang et al. (2014)
ORCHIDEE	1.39	1980–2010	323.9	233.0					Tan et al. (2010)
DOS-TEM		1981–2012		199.0		195.0		4.0	Yi et al. (2014)
TEM	1.37	1990s	451.6	329.6	415.8	303.5	35.8	26.1	Zhuang et al. (2010)
TEM	1.38	2006–2011	235.8	170.9	221.5	160.5	14.4	10.4	Jin et al. (2015)
		2090s[2]	319.9	231.8	247.1	198.7	45.8	33.1	
		2090s[3]	348.4	252.5	321.5	233.0	27.0	19.5	
RG[1]		2050[2]				388.6			Bosch et al. (2017)
		2050[3]				391.2			
		2070[2]				385.9			
		2070[3]				389.2			
TEM (S_T)	1.19	2006–2011[2]	242.1	202.5	240.5	201.1	1.6	1.3	This study
		2090s[2]	315.1	263.6	296.2	247.8	18.9	15.8	
		2090s[3]	441.2	369.0	363.2	303.8	78.0	65.2	

Notes: 1. RG denotes regression model; 2 and 3 represent estimations derived from the RCP 4.5 and RCP 8.5, respectively.

4.2 Impact of climate change on ecosystem C balance

Although permafrost degradation contributes a fraction to the overall changes in C balance (Figure 4), the differences between S_T and S_R shows that climate change is still a main factor affecting various ecosystem C processes, controlling the carbon balance (Figure 3). By

2099, ecosystem C storage and NEP of the permafrost region on the QTP would increase by 1.67±0.32 Pg C and 15.64±3.52 Tg C/yr under the RCP 4.5 and 3.65±0.49 Pg C and 70.56±9.43 Tg C/yr under the RCP 8.5, respectively. Six earth system models from phase five of the Coupled Model Inter-comparison Project (CMIP5) that did not consider the effects of permafrost degradation on SOC stocks show that NPP and RH increase from 1850 to 2100, and the carbon sink of the QTP would be higher during 2006–2100 in comparison with 1850–2005 under RCP4.5 scenario (Li et al., 2015). Increasing net nitrogen mineralization rate enhanced N availability, together with warming air temperature and rising CO_2 concentrations, promote the regional C sink on the QTP (Zhuang et al., 2010).

In the northern high latitudes, warming and elevated atmospheric CO_2 concentration have been found as key drivers to NPP and RH dynamics, but precipitation changes are less important (McGuire et al., 2018). In contrast, on the QTP, air temperature, precipitation, and atmospheric CO_2 concentration in permafrost regions would all increase under both the RCPs (Figure S4). Evidence from experimental warming (Ganjurjav et al., 2016; Zhao, J. et al., 2019), model simulation (Zheng et al., 2020), and meta-analysis (Wang, G. et al., 2019) all suggest that C exchange in alpine meadow is mainly regulated by air temperature, but C exchange in dry alpine steppe is dominated by water conditions. Given that the cover of alpine steppe (7.5×10^5 km^2) is approximately three times higher than alpine meadow (2.5×10^5 km^2) (Ni, 2000), increment in precipitation should also be associated with C exchanges in the permafrost regions on the QTP as warming and elevated atmospheric CO_2 concentration (Chen, B. et al., 2014; Piao et al., 2012;) do. Carbon dynamics responses on the QTP can be different from the conclusion of McGuire et al. (2018) for northern permafrost regions.

4.3 Nitrogen subsidies from permafrost degradation

Permafrost degradation will not only release old carbon but also supply additional nitrogen that can be utilized to boost plant productivity. The direct effects of permafrost thaw would increase ecosystem C storage and enhance C cycling on the QTP in the remaining of this century (Fig 2 and 4). Our analysis suggests that N subsidies from permafrost degradation should be the main reason for these increases in C storages and fluxes. When considering N addition, soil organic N (SON) experiences a remarkable surge in S_T (73.58±37.79 (RCP 4.5) and 156.92±21.48 (RCP 8.5) Tg N) and changes from decrease in S_R (-2.13±1.63 (RCP 4.5) and -4.12±1.80 (RCP 8.5) Tg N) to a large amount of increase in S_T (Figure 7). The increase of N availability enhances plant N uptake (VegNup; S_T-S_R: 12.41±4.72 (RCP 4.5) and 26.56±13.03 (RCP 8.5) Tg N/yr) and makes plant gain more N (VegN; S_T-S_R: 0.62±0.46 (RCP 4.5) and 1.97±0.90 (RCP 8.5) Tg N), which therefore, stimulate plant growth and increase NPP.

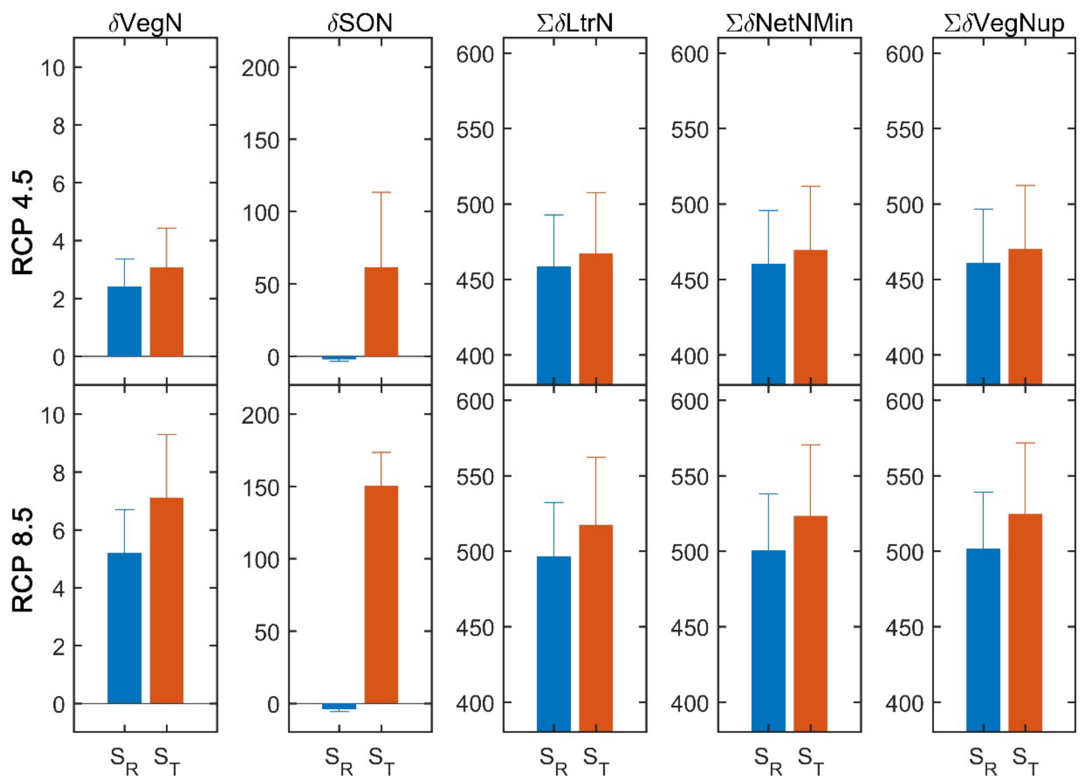

Figure 7. Changes in ecosystem N balance from 2020 baseline for the referenced simulation (S_R) and transient simulation (S_T) under the RCP 4.5 and RCP 8.5 scenarios. δ means the anomalies of N pools of 2099 from 2020 baseline; ∑ denotes the accumulation of each N flux from 2020 to 2099. The error bars represent the standard deviation among all GCMs. Units for N pools (vegetation N (VegN) and soil organic N (SON)) are Tg N, and units for N fluxes (litter N (LtrN), net N mineralization (NetNMin), and vegetation N uptake (VegNup)) are Tg N/yr.

However, although both VegN and SON benefit from permafrost thaw, the increase in VegN is much smaller than in SON (Figure 7). Based on soil C and N balance, this study estimated that 0.98±0.49 and 2.17±0.38 Pg C of SOC, and 76.14±39.45 and 162.90±23.44 Tg N of SON will be added to active layer due to permafrost thaw (Table 2), under the RCP 4.5 and RCP 8.5, respectively. Among them, 57–63% of these additional C would be consumed by microbes and discharged through RH. However, only 10–13% of these addition N would be absorbed by plant.

Figure S5 shows that N subsidies from permafrost thaw only alleviates plant N limit slightly. We find that although there is a large amount of N added to active layer due to permafrost thaw, only a small fraction of the amount can be accessed by plant. The average maximum rooting depth in permafrost regions on the QTP is estimated as 1.2 m, while the average ALT would increase to 2.9±0.1 and 3.3±0.1 m under the RCP 4.5 and RCP 8.5, respectively (Figure 4a), which is much greater than rooting zone depth. Therefore, only a small fraction of N released from thawed permafrost is absorbed by plant. The benefit of permafrost thaw is limited, but the consequence of permafrost thaw to C emission is large. As a result, the net effect of permafrost thaw on the QTP is to weaken C sequestration. Similar results have also been found over the circumpolar permafrost regions. Permafrost thaw from 1970 to 2006 exposed 11.6 Pg C of SOC to decomposition, resulting in 4.03 Pg C emission but only 0.3 Pg C compensated by the stimulated plant C uptake (Hayes et al., 2014). In the forests of Northern Eurasia, the gain of vegetation C that is benefited from N supply due to permafrost degradation is only 29.8% (RCP4.5) to 49.2% (RCP8.5) of the loss of soil organic C. Permafrost degradation overall diminishes C sequestration in these forests (Kicklighter et al., 2019). Although N released from new thawed permafrost can be used to stimulate plant productivity, plant may not be N limited or may not access to them (Koven, Lawrence et al., 2015). However, new released SOC will be decomposed, although part of them take a longer time. Consequently, the stimulated RH may overwhelm the NPP increase (due to more available N to plant from thawing permafrost), decreasing the regional C sequestration.

Table 2. Changes in soil C and N balance due to permafrost thaw.

C	ΔSOC	∑ΔLtrC	∑ΔRH	SOC_{PF}	∑ΔRH/ SOC_{PF}
RCP 4.5	0.65±0.26	0.33±0.27	0.66±0.53	0.98±0.49	0.57±0.41
RCP 8.5	1.59±0.21	0.84±0.49	1.42±0.74	2.17±0.38	0.63±0.24
N	ΔSON	∑ΔLtrN	∑ΔNetNMin	SON_{PF}	∑ΔVegNup/ SON_{PF}
RCP 4.5	75.57±39.08	8.46±7.18	9.04±7.69	76.14±39.45	0.10±0.09
RCP 8.5	160.93±22.55	20.84±12.12	22.81±13.05	162.90±23.44	0.13±0.06

Notes: Δ means the differences between the transient simulation and referenced simulation (S_T-S_R); ∑Δ means the cumulative of S_T-S_R from 2020 to 2099; SOC_{PF} and SON_{PF} mean SOC, and SON added to active layers from permafrost thaw, respectively. LtrC, LtrN, NetNMin, and VegNup denote litter C, litter N, net N mineralization, and vegetation N uptake, respectively. Units for C and N variables are Pg C and Tg N, respectively.

There are several reasons in the limited plant access to N released from permafrost degradation on the QTP. First, ALT (greater than 2 m; Ni et al., 2021; Wang, T. et al., 2020; Wu, X.B. et al., 2018) on the QTP is already much deeper than rooting depth (generally smaller than 30 cm; Yang et al., 2009) currently. N released from further deep permafrost in the future, therefore, could be far beyond plant accessibility if there are no substantial changes in plant species. Second, N mineralization in permafrost soils on the QTP has been found largely regulated by microbial traits (Mao et al., 2020; Zhang et al., 2020). Microbial biomass in permafrost on the QTP is 91.3% lower than that in active layer soils (Mao et al., 2020). The low abundance of microbes can become a key restriction on N mineralization after permafrost thaw, because of the direct role that microbes play in N transformation (Schimel and Bennett, 2004) and the positive effects of the extracellular enzymes (Ali et al., 2021; Luo et al., 2017) and microbial biomass (Li, Z. et al., 2019; Wu, H. et al., 2021) on soil N mineralization. Apart from microbial biomass, the higher abundance of bacteria than fungi in QTP permafrost may also has negative impacts on net N mineralization processes (Mao et al., 2020), because of the significantly higher metabolic N-demand and immobilization in bacteria than fungi (Kooijman et al., 2016). Further, the phase lag of heat transfer into deep soils can shift the deep soil organic matter mineralization later into fall and winter, producing a seasonal offset from the peak period of N demand during spring and summer (Koven, Lawrence et al., 2015), and reducing the access of plant to the additional N released from deep permafrost thaw. The remaining large fraction of N that cannot be used by plant could leach into aquatic ecosystems, causing far-reaching consequences on their functions and structure (Guo et al., 2019; Wickland et al., 2018), or could be lost via gaseous form of N_2O through nitrification and denitrification processes (Elberling et al., 2010; Voigt et al., 2017, 2020; Wilkerson et al., 2019), intensifying non-carbon feedback to climate warming (IPCC, 2013; Xu et al., 2012; Yang, G. et al., 2018). Considering the serious consequences and the large amount of these additional N, sufficient attention should be paid to their fate on the QTP.

4.4 Nitrogen subsidy regimes from ALT deepening in different ecosystems

The more pronounced influences of ALT deepening on C balance in alpine meadow than alpine steppe on the QTP can be due to different C and N subsidies from permafrost thaw. Although the increase in ALT is smaller in alpine meadow comparing to alpine steppe, alpine meadow gains more C and N (Figure 6 and Figure 8). The maximum rooting depth of alpine meadow and alpine steppe ranges from 0.5 to 1.9 m, but the average ALT in alpine meadow (2.05±0.01 m) is shallower than alpine steppe (2.96±0.03 m) currently. Although ALT increases more in alpine steppe, it mainly occurs in deep soils far from rooting zone, therefore, only very limited N released from thawed permafrost can be accessed by alpine steppe. On the contrary, owing to the shallower ALT in alpine meadow, more N released from thawed permafrost is within its rooting zone, thus can be used. The addition of SOC and SON released from permafrost degradation in these two ecosystems (Table 3) support such explanations.

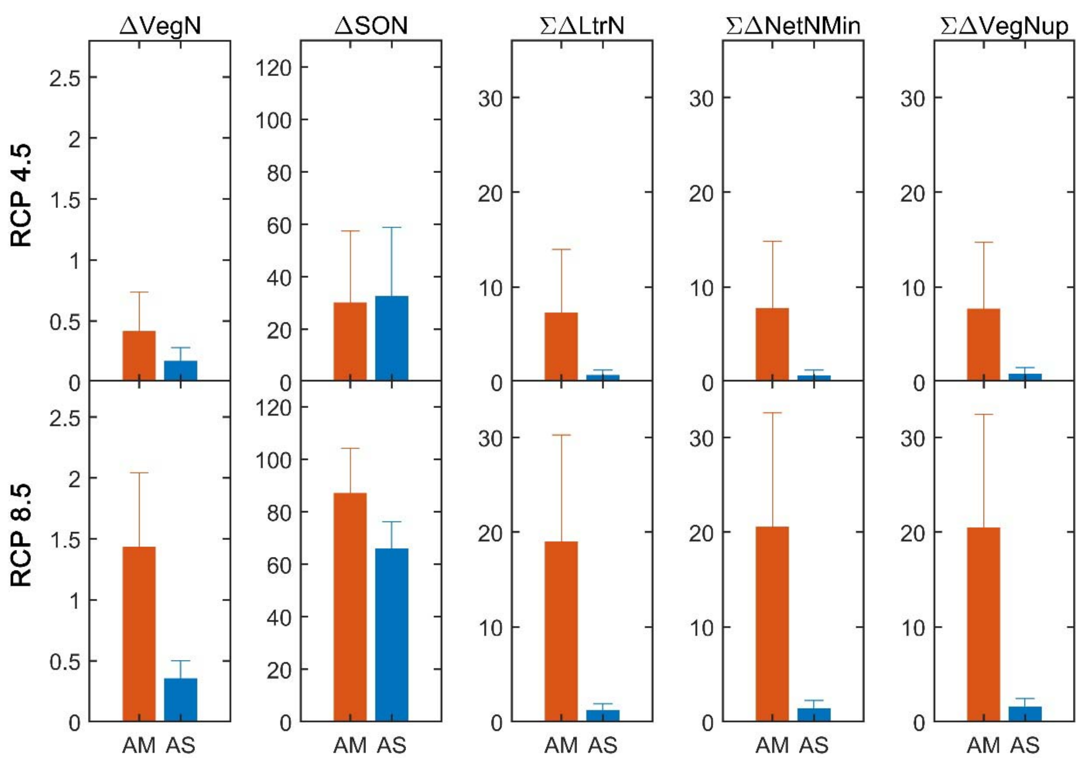

Figure 8. Differences in N balance between alpine meadow (AM) and alpine steppe (AS) ecosystems due to ALT deepening. Δ means the simulation differences between the transient

simulation and referenced simulation (S_T-S_R) in 2099; ∑ denotes the accumulative differences in N fluxes from 2020 to 2099. The error bars represent standard deviations among all GCMs. Units for N pools (vegetation N (VegN) and soil organic N (SON)) are Tg N, and units for N fluxes (litter N (LtrN), net N mineralization (NetNMin), and vegetation N uptake (VegNup)) are Tg N/yr.

In addition, SOC and total nitrogen stocks in alpine meadow is about twice of that in alpine steppe on the QTP (Ding et al., 2016; Zhao, L. et al., 2018), which may result in the significant increasing in ecosystem C and N pools in alpine meadow ecosystem even with less increases in ALT. The higher density of SOC underneath alpine meadow can provide more substrates for microbe than that underneath alpine steppe, triggering a stronger RH response (Crowther et al., 2016). More SOC can also stimulate microbial activities and increase microbial biomass in alpine ecosystems (Li, Y. et al., 2019), strengthening RH. Apart from quantity, incubation experiments found that SOM under alpine meadow is highly decomposable (Chen, L. et al., 2016), while SOM under steppe is largely composed of stable organic compounds (Wu, X. et al., 2014). These differences may also contribute to a more sensitive RH in alpine meadow ecosystems (Eberwein et al., 2015). The consumption of SOC that released from permafrost thaw (Table 3) also shows that most of the additional SOC (79–85%) in alpine meadow would be consumed by microbe through RH, while only nearly a half (44–51%) of them would be consumed in alpine steppe.

The stronger response of ΔNPP to ALT in alpine meadow can partly be explained by the significantly higher usage of N supplied by permafrost thaw (Table 3). More than 25% of additional N is utilized in alpine meadow, while only less than 3% is absorbed by alpine steppe, therefore, contributing to distinct vegetation responses in ΔNPP. Alpine meadow ecosystem is mainly distributed in the eastern QTP. Relatively adequate precipitation, but also cold climate makes alpine meadow has higher soil water content than alpine steppe, which is distributed in arid climate. High soil water content in alpine meadow can facilitate N utilize by plant as water availability influences the diffusion of soil N to plant (Humbert et al., 2016; Simkin et al., 2016). Conversely, water limiting in alpine steppe constrains its nutrient availability and make it respond differently to climate warming and nitrogen addition from alpine meadow (Ganjurjav et al., 2016; Li, S. et al., 2019; Zheng et al., 2020). Different water conditions in these two ecosystems may result in distinct differences in N usage (Table 3) and fluxes (Figure 8).

Table 3. Changes in SOC and SON in alpine meadow and alpine steppe ecosystems due to permafrost thaw.

	Alpine meadow		Alpine steppe	
C	SOC_{PF}	$\sum\Delta RH/SOC_{PF}$	SOC_{PF}	$\sum\Delta RH/SOC_{PF}$
RCP 4.5	0.57±0.31	0.85±0.22	0.42±0.25	0.51±0.29
RCP 8.5	1.40±0.37	0.79±0.30	0.79±0.14	0.44±0.13
N	SON_{PF}	$\sum\Delta VegNup/SON_{PF}$	SON_{PF}	$\sum\Delta VegNup/SON_{PF}$
RCP 4.5	30.70±29.23	0.31±0.13	32.60±27.48	0.02±0.00

RCP 8.5	88.73±18.62	0.25±0.09	66.15±7.41	0.03±0.01

Notes: ∑Δ means the cumulative difference between the transient simulation and referenced simulation (S_T-S_R) from 2020 to 2099; SOC_{PF} and SON_{PF} mean SOC, and SON added to active layers from permafrost thaw, respectively. Units for soil C and N are Pg C and Tg N, respectively.

5. Conclusions

We modified a process-based biogeochemistry model by considering the effects of extra carbon and nitrogen from thawing permafrost to quantify ecosystem C balance on the Qinghai-Tibetan plateau in the 21st century. Permafrost regions on the plateau would sequestrate more C and become a stronger C sink by the end of this century under both RCP 4.5 and RCP 8.5 scenarios. However, permafrost degradation on the QTP is projected to diminish the ecosystem C sequestration capacity, because the increase of NPP stimulated by N supply from thawing permafrost is smaller than the enhanced soil carbon respiration. Due to the deep active layer on the QTP, N released from thawing permafrost is mainly distributed in deep soils, which is not accessible by plant, thus having a limited benefit to plant carbon uptake. As a result, the C sink would decrease by 303.55±254.80 (RCP 4.5) and 518.43±234.04 (RCP 8.5) Tg C during 2020–2099.

Permafrost degradation on the QTP has different effects on C balance of the distinct ecosystems of alpine meadow and alpine steppe. C storages and fluxes respond more significant to ALT deepening in alpine meadow than alpine steppe, which could be due to the shallower ALT, the higher C and N stocks, and the wetter environment in alpine meadow.

Although more than half of the C released from thawing permafrost would be decomposed and emitted into the atmosphere, only 10% of the released N is utilized by plant. The remaining large amount of N could have significant impacts on aquatic ecosystems and may intensify climate warming through another potent greenhouse gas N_2O release. Therefore, the study on this gas release shall draw sufficient attention on the plateau.

This study highlights the important effects of permafrost thaw on C and N cycling in permafrost dominated ecosystems and underscore the critical role of factoring deep permafrost carbon into land surface modeling in order to make more reliable projections of future permafrost carbon-climate feedbacks.

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Global Biogeochemical Cycles

Supporting Information for

Permafrost Degradation Diminishes Terrestrial Ecosystem Carbon Sequestration Capacity on the Qinghai-Tibetan Plateau

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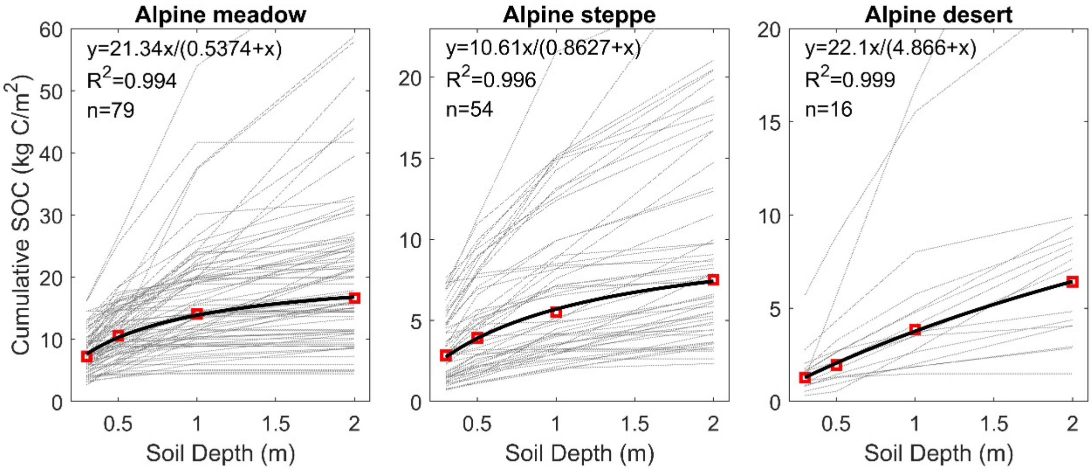

Figure S1. Changes of cumulative SOC with soil depth in alpine meadow, alpine steppe and alpine desert ecosystems in the permafrost region of the QTP. The dotted gray lines denote observation results; the red r rectangles represent the median value of every soil depths; and the thick black lines are regression curves. 'n' in each panel is the number of observation sites. Observation data derived from Zhao et al. (2018).

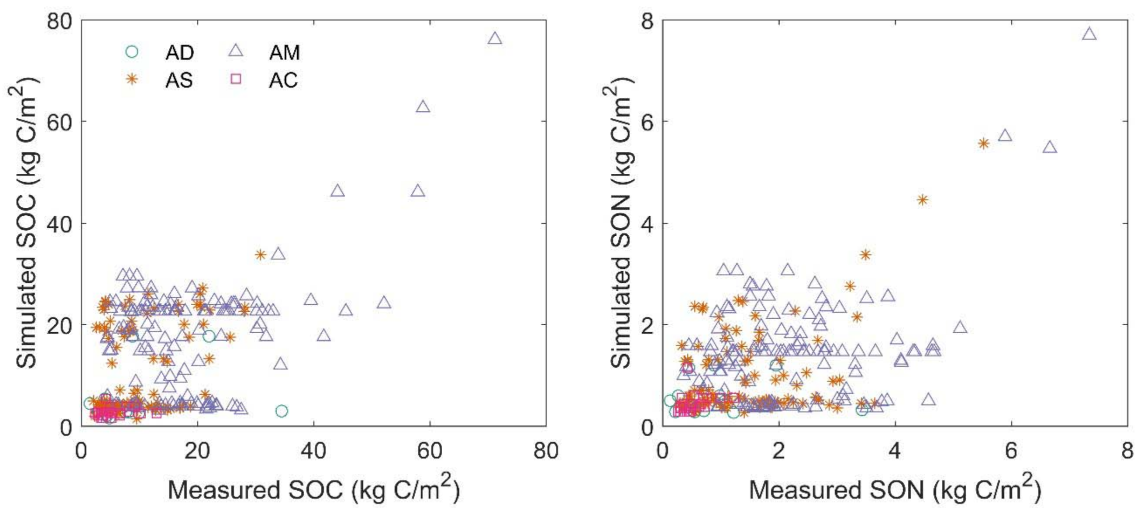

Figure S2. Validation of simulated SOC and SON vs. 289 observations on the QTP. AD, AS, AM, and AC represent alpine desert, alpine steppe, alpine meadow, and alpine cushion, respectively.

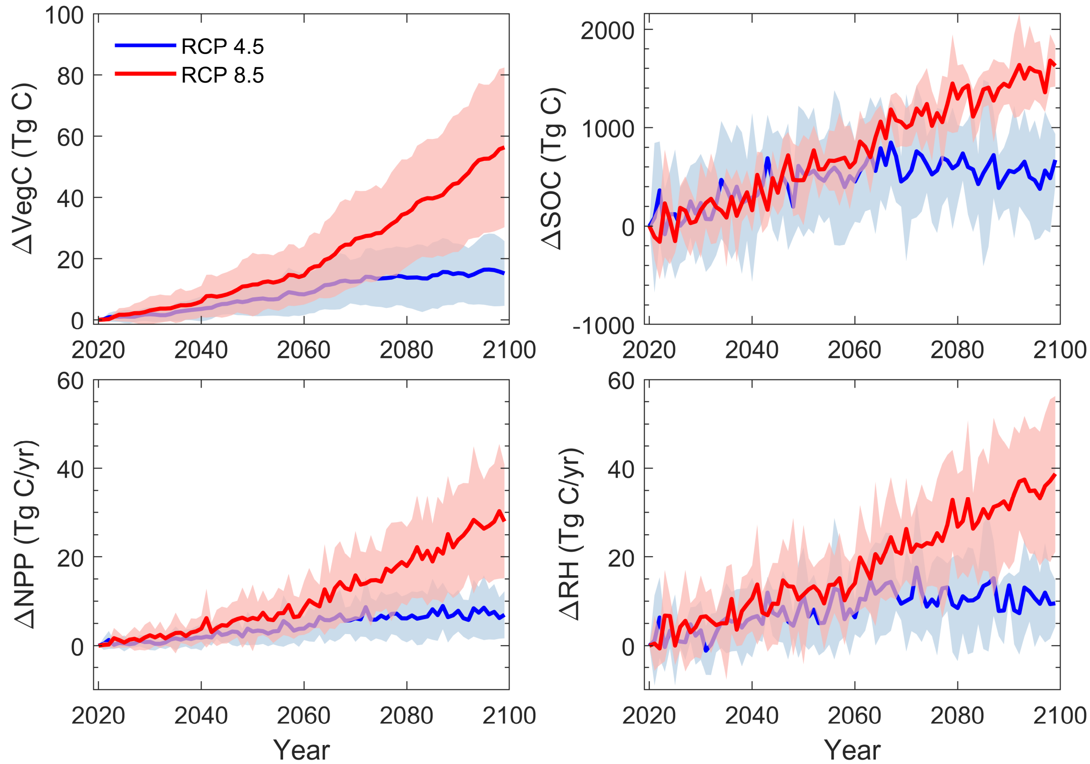

Figure S3. Effects of ALT deepening on ecosystem C balance of permafrost regions on the QTP under the RCP 4.5 and RCP 8.5 scenarios. Δ means the difference between the results of ST and SR (ST-SR).

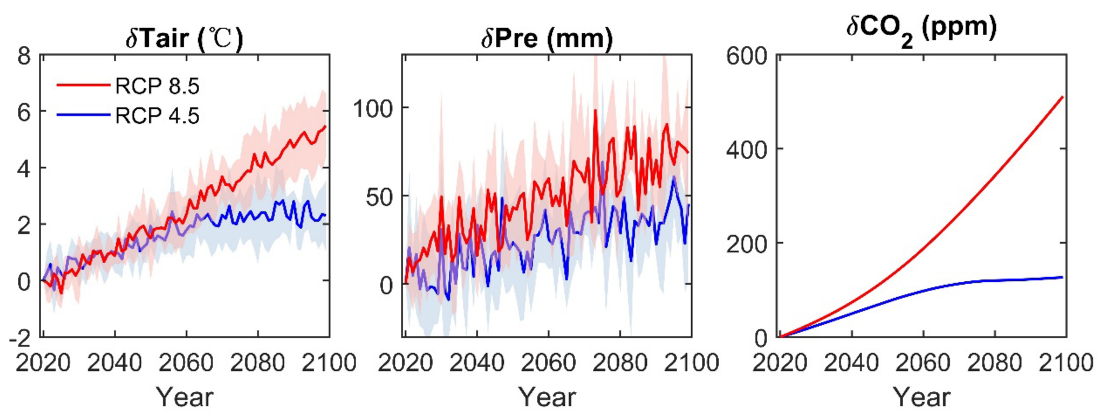

Figure S4. Changes in air temperature (Tair), precipitation (Pre), and atmospheric CO2 concentration of the permafrost region on the QTP based on 2020.

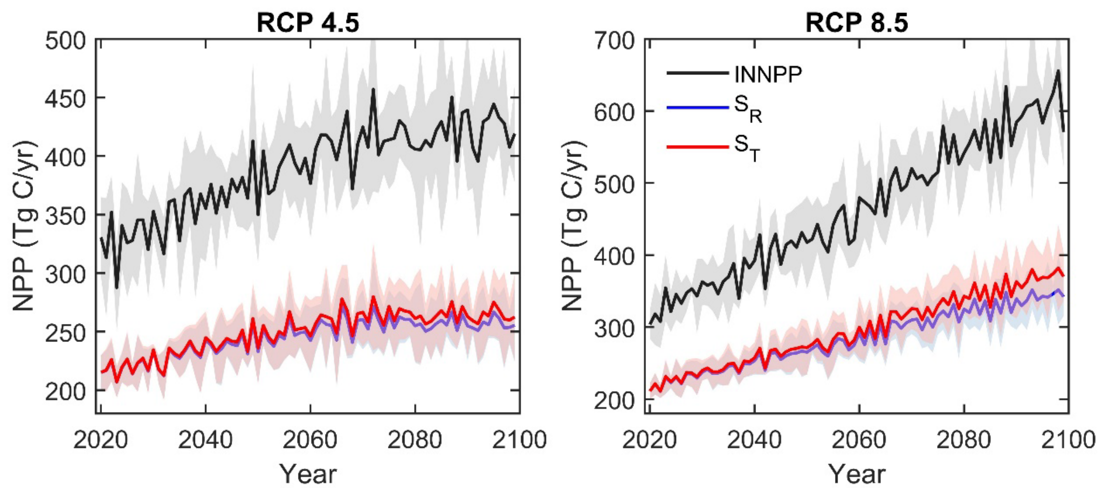

Figure S5. Changes in NPP in ST and SR simulations and changes in INNPP during 2020-2100 under the RCP 4.5 and RCP 8.5. INNPP is the NPP estimated by TEM that is not constrained by N supply but only regulated by climate.

Steppe	Meadow	Sparse	Desert	Shrubland	Forest	Others
0.8627*	0.5374*	0.5937**	4.866*	1.2553**	0.4746**	0.6920**

Table S1. The value of parameter b in eq. (2) for each ecosystem in the permafrost region of QTP.

Notes: * denotes the data calculated from observation data (Fig. S1); ** denotes data derived from Kicklighter et al. (2019).

Ecosystem		Alpine meadow	Alpine steppe	Alpine shrubland
Site		Zoige	NamCo	Haibei
Latitude		33°53′ N	30°46′ N	37°37′ N
Longitude		102°8′ E	90°57′ E	101°19′ E
Elevation (m)		3423	4730	3190
SOC	obs*	30895.51	4698.55	23440.00
(g C/m^2)	sim*	30329.27±37.4	4723.02±5.57	22581.90±14.97
SON	obs*	1984.78	399.04	2616.20
(g N/m^2)	sim*	2032.76±1.83	398.78±0.25	2793.79±0.40
Reference		Huo et al., 2013; Shang et al., 2016	Ding et al., 2016; Kou et al., 2019; Nieberding et al., 2020	Tao et al., 2007; Fu et al., 2018; ChinaFLUX**

Table S2. Information of observation sites for model validation.

Notes: *: obs and sim denote observation and simulation, respectively; **: http://www.chinaflux.org/gczd/index.aspx?nodeid=1003