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Vegetation can strongly regulate permafrost degradation at its southern edge through changing surface freeze-thaw processes

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A R T I C L E I N F O

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A B S T R A C T

Permafrost contains twice as much carbon as the atmosphere and the degradation of permafrost due to climatic warming, which could potentially change the global carbon cycle and could also enhance global climate change. It is well studied that permafrost degradation could result in vegetation transition. Aboveground vegetation can act as a buffer for climatic warming, however its role in regulating permafrost degradation remains unclear. In this study we examined how different vegetation types regulated the amplitude and duration of diurnal soil freeze/thaw (FT) cycles and the timing of seasonal soil FT. Soil temperature data (hourly and half hourly) was collected from paired forest-steppe sampling plots spanning a large spatial gradient from northern China to southern Siberia, Russia from 2008 to 2015. FT cycles were found to be larger in amplitude and longer in duration in steppe sites in comparison to forest sites. Soils in the forest sites and steppe sites freeze almost simultaneously, but experience a delay in thawing of approximately 14, 19 and 25 days for deciduous broadleaf forest, evergreen coniferous forest, and deciduous coniferous forest, respectively. Variations in snow accumulation due to differences in vegetation structure as opposed to solar radiation were responsible for the disparity in thaw timing. These findings imply that deciduous conifer forest in east Eurasia could reduce carbon emissions more effectively than evergreen conifer forest in west Eurasia by slowing down warming-induced permafrost degradation during spring thaw.

1. Introduction

Permafrost (permanently frozen ground) comprises a region underlying 23.9% of the land area of the Northern Hemisphere. Currently permafrost contains about twice as much carbon as the atmosphere (Zhang et al., 1999; Zimov et al., 2006). These large quantities of carbon stored in frozen soils can be released into the atmosphere due to warming-induced permafrost degradation, which is further enhanced by a warming climate (Hodgkins et al., 2014; Hollesen et al., 2015; Schuur et al., 2015). This degradation could cause permafrost regions to shift from being a sink to a source of CO_2 by the end of the 21st century (Koven et al., 2011). Permafrost degradation mainly occurs at the southern edge of the permafrost area due to the significant poleward movement of permafrost with climate warming (Guo and Wang, 2016). Soil freezing–thawing (FT) processes are measured by the amplitude and duration of the diurnal soil FT cycle as well as seasonal FT timing and can effectively mediate permafrost degradation. Increasing soil surface thawing days can enhance permafrost degradation by deepening active layer thickness (Peng et al., 2017; Wu et al., 2015; Zhang, 2005). The soil suffers remarkable FT cycles during the transition period from cold/warm to warm/cold seasons. The transition between frozen soil to thawing is a continuous process lasting from several days to a few weeks (Cheng et al., 2014; Wang et al., 2013), and is susceptible to rapid climate warming, in particular the higher warming rates that occur in winter and spring (IPCC, 2013). Stronger and longer diurnal FT cycles and earlier seasonal permafrost thawing may increase carbon loss from the soil.

There has been significant research examining spatio-temporal changes in soil FT processes in response to a warming climate at regional and continental scales, mainly using land surface temperature data from satellites (Kim et al., 2014; Li et al., 2012; Zhang, 2003) and land surface model simulations (Guo and Wang, 2014). Despite this, uncertainties still exist in the conclusions of these studies due to the coarse resolution of the available data and poor representation and

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parameterization of soil FT processes. To comprehensively understand soil FT processes at a regional scale, long-term direct monitoring of these processes in the field over a large spatial and climactic gradient is needed (Cheng et al., 2014; Sun et al., 2012). However, this is currently rare, preventing a detailed understanding of patterns and drivers of soil FT processes.

Long-term climate warming could disequilibrium the relationship between vegetation and permafrost (Camill and Clark, 1998). Most previous studies have focused on how permafrost degradation affect biomass and aboveground net primary production (Camill et al., 2001), vegetation recruitment (Camill et al., 2010), and vegetation fragmentation (Baltzer et al., 2014). Also, there are some researches concerning the vegetation impact on permafrost degradation. Field measurements in interior Alaska (Jorgenson et al., 2010), Jubany (King George Island) and Signy Island in the Maritime Antarctic (Cannone et al., 2006) found that vegetation could exert negative effect on permafrost degradation. However, a significant knowledge gap remains in terms of the regulative role of vegetation in permafrost degradation over a regional scale and long-term direct monitoring of soil temperature as proxy, especially by soil FT processes underground. Different vegetation types have distinct community structures, and consequently differ in solar radiation income, litter cover and snow accumulation patterns. Increasing LAI (leaf area index) could increase vegetation canopy and then raise the interception of snow which could affect snow accumulation in the interior of vegetation during the winter period (Pomeroy et al., 2002; Rasmus et al., 2011). Different snow accumulations could produce various influences on soil temperature variations (Iwata et al., 2010; Park et al., 2014b). These differences can substantially affect the energy balance within the soil-vegetation-atmosphere system and the thermal dynamics of the soil. We thus hypothesize that vegetation type can significantly influence diurnal soil FT cycles and the timing of seasonal soil FT.

Eleven paired forest-steppe measurement sites were located along a vast latitudinal (> 10°) and longitudinal (> 30°) gradient on the edge of the permafrost region of southern Siberia, Russia and northeastern China where permafrost degradation is more sensitive than in Canada and the USA (Guo and Wang, 2016). Soil FT processes were recorded directly and continuously at these sites from 2008 to 2015 (Fig. 1). Using the soil temperature observations collected during this eight year period, this study investigated the effects of vegetation type on the duration and amplitude of diurnal FT cycles and the timing of seasonal soil FT (see Materials and Methods).

2. Materials and methods

2.1. Study region

Our study region is located in southern edge of a permafrost region in Inner Mongolia, China and southern Siberia, Russia (Fig. 1). This region is predominantly covered by forest-steppe transitional vegetation, with forests distributed mainly on shady slopes and steppes found on sunny slopes and flatlands (Gutiérrez-Jurado et al., 2013; Hu et al., 2013; Namzalov et al., 2012). The mean annual temperature ranges from −4.5 °C in the north to 3.5 °C in the south of this region, and mean annual precipitation ranges from 350-550 mm (Table S1). We established four sampling sites in the Altai Mountains, three sampling sites in the Trans-Baikal region of Russia and four sampling sites on the eastern Inner Mongolian Plateau in China (Fig. 1).

2.2. Vegetation types

In each site, we selected two typical aspect-dependent vegetation plots. One was located in forest and the other was in the adjacent steppe. These paired plots were several hundred meters apart. Forest sample plots were either dominated by *Larix sibirica* (larch), *Pinus sylvestris* (pine) or *Betula platyphylla* (birch). These species belong to different vegetation types, deciduous conifer, evergreen conifer and deciduous broadleaf forest, respectively and their morphological and ecophysiological features differ from each other. While *L. sibirica* and *P. sylvestris* are both conifers, *P. sylvestris* is evergreen and can intercept some snow in the canopy layer, decreasing its penetration to the forest interior, however the deciduous *L. sibirica* cannot.

2.3. Soil temperature measurements

We placed temperature data loggers (HOBO, Onset Computer Corporation, MA, USA) into about 10 cm in depth below ground surface, and logged soil temperature hourly or half hourly. Data loggers were set in the interior of the sample plots, without depression and root around. Loggers were installed in the soil in 2008, with the exception of the Altai Mountains, where loggers were not installed until 2010. The data was downloaded from the loggers during the transition from summer to autumn of each year and the battery was also checked, and if necessary replaced, before replacing the logger into the soil.

Total potential solar radiation (TPSR). TPSR were calculated by using Solar Calculator tool (http://www.meteoexploration.com/products/SolarCalculator.html). While giving the specific location (latitude and longitude), timing, altitude, temperature, etc (Table S1), we calculated the TPSR for each forest-steppe sample plots.

2.4. Soil temperature, precipitation and solar radiation of USCRN (the U.S. climate reference network)

USCRN is a network of meteorological stations maintained by the National Oceanic and Atmospheric Administration which have been measuring soil temperatures since 2011 (Bell et al., 2013). In order to explore the impact of snow depth and solar radiation on soil thaw timing with different vegetation types, USCRN is prior to this research which contains soil temperature, precipitation and solar radiation and could help us to build connection between vegetation types and soil thaw timing. It appears that precipitation during freezing time is a proxy for snow accumulation.

2.5. Estimation of soil freeze/thaw parameters

Difference in Julian day of year (DOY) was used to compare the difference in soil thawing and freezing dates between paired forest-steppe samples for different vegetation types. New Year's Day is defined as DOY 1, and the last day of the year is DOY 365 or 366 in a leap year. Soil temperature was used to determine periods of soil thawing and freezing. When daily minimum soil temperature exceeded 0 °C, it was categorized as a soil thawing day and when daily maximum soil temperature was below 0 °C, it was categorized as a soil freezing day.

When the daily maximum soil temperature exceeded 0 °C and the minimum soil temperature was below 0 °C in one day, this implies that the soil thawed during the day and froze again during the night, producing a FT cycle. We then calculated both the duration and amplitude of diurnal soil FT cycles during the soil freezing–thawing transitional periods. Duration of diurnal soil FT cycles was calculated using the number of days with diurnal soil FT cycles (Bartsch et al., 2007; Guo and Wang, 2014; Han et al., 2010). Duration was calculated for both freeze-to-thaw and thaw-to-freeze stages for each of the paired sample plots. Amplitude of diurnal soil FT cycles was evaluated using diurnal temperature ranges (difference between maximum and minimum soil temperature) for periods when diurnal soil FT cycles occurred (Henry, 2007).

3. Results

3.1. Diurnal soil FT cycles of different vegetation types

Comparisons of temperature variations within diurnal soil FT cycles