Environmental Research Letters

LETTER

Permafrost response to vegetation greenness variation in the Arctic tundra through positive feedback in surface air temperature and snow cover

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Keywords: arctic vegetation, permafrost, snow cover

Abstract

The permafrost response to variations in Arctic vegetation remains controversial. We investigated the consequences of Arctic vegetation greenness variation over the past three decades using a coupled land-atmosphere model and found that it induces air temperature perturbation, which is further amplified by snow cover variation and eventually leaves a footprint on soil temperature. Compared to the atmospheric impacts of vegetation, local shading of vegetation canopy has relatively minor effects on soil temperature. Significant soil warming was observed along the summer snowline between the Low and High Arctic, indicating the direct impact of snow cover variation led by vegetation changes. In the Low Arctic, the winter snowpack insulates the soil from colder air, resulting in less permafrost. In the High Arctic, snow persists for more than 330 d per year and has a strong protection effect on the permafrost as it insulates soil from warmer summer air and reflects solar radiation.

1. Introduction

Recently, the Arctic has experienced an increase in surface air temperature almost twice as large as the global average. Known as the Arctic temperature amplification (Serreze and Francis 2006), this phenomenon leads to thawing of the permafrost which can result in carbon release from frozen organic soil as carbon dioxide or methane (Tarnocai et al. 2009, Chen et al. 2015). In addition, this process reduces surface water through vertical groundwater movement (Liao and Zhuang 2017). In other words, temperature amplification can significantly change the ecological, hydrological, and climate processes in the Arctic. Simultaneously, permafrost is affected by many environmental factors, including air temperature, wildfires, snowpack, and vegetation growth (Grosse et al. 2011).

Increasing vegetation has been observed at multiple sites across the Arctic over the past few decades in response to rising air temperatures (Myers-Smith et al. 2011). The normalized difference vegetation index (NDVI) is often used to detect the vegetation greenness in the Arctic (e.g. Bhatt et al. 2010, Beck and Goetz 2011). Beck and Goetz (2011) analyzed the NDVI product (MOD13A3) of the moderate resolution imaging spectroradiometer (MODIS) between 1992 and 2008 and noted an increasing NDVI trend in the Arctic tundra. Bhatt et al. (2017) studied the NDVI records of the advanced very high-resolution radiometer (AVHRR). The authors found that the annual maximum NDVI in the pan Arctic tundra showed positive trends from 1982 to 2015, but the annual time-integrated NDVI, the sum of the biweekly NDVI time-integrated NDVI, the sum of the biweekly NDVI, showed a log-linear relationship from 1982 to 2001 and increased from 2001 to 2015. The time-integrated NDVI better represents gross primary production than the maximum LAI (Tucker and Sellers 1986) and is better correlated with climate parameters (Bhatt et al. 2010).

Vegetation changes influence the soil temperature via multiple mechanisms. The field experiment of Blok et al. (2011) in the north-eastern Siberian tundra suggested that a larger canopy could shade the ground
and mitigate deep permafrost thawing during the growing season. In contrast, simulations using the coupled land-atmosphere global climate model suggested that a larger leaf area index (LAI) could accelerate permafrost thawing by reducing albedo and increasing surface air temperature in spring (Lawrence and Swenson 2011). However, simulations based on global LAI increases over the past three decades show that parts of the Arctic surface are being cooled due to increased evapotranspiration and large-scale interannual atmospheric circulation (Zeng et al 2017). Snow cover is another factor that influences permafrost thawing. The snowpack insulates soil from colder air in the winter, prevents deep freezing of the near-surface ground, and results in less permafrost in the following summer (Bulygina et al 2011). Stieglitz et al (2003) found that snow cover influenced the variation in soil temperature in addition to near-surface air temperature in the Arctic. Johansson et al (2013) increased the amount of winter snow in a manipulative experiment in northern Sweden and reported increased permafrost thawing with ground subsidence. Park et al (2015) modified winter precipitation in simulations involving the Pan-Arctic land surface and found that increased winter precipitation resulted in less permafrost in northern Siberia. However, Zhang et al (2018) compared in situ observation records of boreal soil and air temperatures and argued that snow insulation weakens the warming trend in boreal soil more significantly than that of the near-surface air.

Previous studies have shown multiple consequences of Arctic vegetation changes, involving variations in air temperature, snow cover, and solar radiation on the ground. The significances of the multiple pathways of environmental impacts on permafrost can be investigated with numerical simulations. To show accelerated permafrost thawing by reducing albedo and increasing surface air temperature in the spring, Lawrence and Swenson (2011) simulated the permafrost response to Pan-Arctic vegetation changes using the coupled land-atmosphere global climate model while considering LAI, which persistently increased from 2000 to 2100. However, this Arctic greening trend is not consistent with spaceborne observations from 2000 to 2015 (Bhatt et al 2017). Integrating long-term observations regarding vegetation greenness coupled with land-atmosphere simulations could improve our understanding of the consequences of tundra greening and browning in recent decades (Bhatt et al 2017). Therefore, this study investigates the permafrost response, specifically soil temperature, to LAI changes in the Arctic tundra over the last three decades. In addition, snow cover and surface air temperature involved in this process are explored.

2. Methods

We used Community Land Model 4.5 (CLM 4.5, Oleson et al 2013) and Community Atmosphere Model 5 (CAM, Neale et al 2012), which are coupled in the framework Community Earth System Model 1.2.1 (CESM, Hurrell et al 2013) with a 1.9° × 2.5° spatial resolution. The sea surface conditions referred to in this paper are comprised of sea ice data, based on spaceborne observations by the special sensor microwave imager (Wentz 1997) and sea surface temperature (SST) data. This is a merged product based on the monthly mean Hadley Center SST and the national oceanic and atmospheric administration (NOAA) weekly optimum interpolation SST (Hurrell et al 2008). Furthermore, the simulation of fractional snow cover in CLM4.5 was improved from CLM4.0 based on observations by calculating the snow-free and snow-covered fractions of a grid cell separately (Swenson and Lawrence 2012).

Changes in the Arctic vegetation were noted with the monthly LAI for each plant functional type (PFT) in each grid cell. In this study, we use PFTs derived from Global Land Cover (GLC2000, Bartholomé and Belward 2005), a static land cover product with a resolution of ∼1/112°. We choose this method to prevent the generation of false trends and errors in recent multiple-year PFT products such as MODIS (MOD12, Friedl et al 2010) and the climate change initiative (CCI) land cover of the European Space Agency (Poulter et al 2015). The Arctic shrub coverage in the GLC2000 product is ∼24%, largely consistent with the corresponding value of 26% in AVHRR (Walker et al 2005), while the 500 m annual PFTs of MOD12 record a shrub cover of 63% (Lawrence and Swenson 2011). The CCI land cover agrees better with the results of Walker et al (2005), but shows a persistent declining trend in shrub cover since 1992 which contradicts the abundant surface observations of shrub expansion from the 1980s to 2010s (Myers-Smith et al 2011). Hence, current multiple-year PFT products are limited in terms of distinguishing Arctic shrub and grass. Field observations indicate that Arctic shrubs have expanded by ∼1% per decade since the 1950s (Sturm et al 2001a, 2001b). In our study spanning 34 years, these shrubs expanded by ∼3%, which is one order of magnitude less than the LAI changes. Thus, this study focused on the LAI, rather than land cover variations.

The monthly LAI in this study was generated from biweekly Arctic NDVI between 1982 and 2015 (Bhatt et al 2017) using an empirical equation (Shaver et al 2007). It was derived from field experiments on Arctic shrub-grass tundra (van Wijk and Williams 2005, Williams et al 2006 and Street et al 2007) and was validated in the Siberian tundra (Juutinen et al 2017). Loranty et al (2011) also applied this method to a simple ecosystem model for the tundra. In contrast, MODIS LAI products (MOD15, Myneni et al 2002)
were found to overestimate the LAI for the Alaskan tundra (Verbyla 2005). MOD15 data were generated using a lookup table method from the reflectance and NDVI data for each PFT. In this process, the inaccuracy of the MODIS PFTs was inherited by its LAI products. Moreover, a single shrub type in MOD12 spans the area from the tropics to the Arctic. Therefore, an empirical relationship between the LAI and NDVI for global shrublands is not ideal for the Arctic region. Although the equation developed by Shaver et al. (2007) may be limited by the spatial heterogeneity of the vast Arctic region, it remains the best estimation available to date. Furthermore, we follow the canopy height used in Lawrence and Swenson (2011), by setting maximum canopy height of Arctic grass and shrubs at 0.1 m and 1.0 m, respectively. For the regions except for Arctic, the LAI and the canopy height remain the same for the default settings of CLM4.5 (Lawrence and Chase 2007).

CLM is first spun up for 50 years to reach a stable snow depth in the High Arctic while avoiding any biases of simulated Arctic snow depth due to its initial value (Oleson et al. 2013). Then, CLM was coupled with CAM and spun up for an additional 5 years. During the last stage, two 34 year simulations were performed from the spin-up: one with the realistic surface data changed annually from 1982 to 2015 (SCLIM), and the other with the fixed land surface data of 2000 (S2000), the year with the maximum time-integrated LAI.

3. Results

The Arctic tundra LAI (i.e. LAI of grass and shrub) in 2000 was significantly higher than the mean from 1982 to 2015 (figures 1 and 2(a)). As shown in the averages of the two runs (S2000 and SCLIM) in figure 2, such a difference in LAI does not lead to any significant differences in land surface albedo, latent heat, air temperature, and soil temperature for the whole Arctic. A decrease in summer snow depth in S2000 relative to SCLIM was observed. In SCLIM, snow exists in a period of August in northern Canada and the Siberian coast (figure 3(a)). These regions are consistent with the regions with 330–360 snow cover days in the satellite dataset of NOAA (Brown and Mote 2009). Although the LAI difference between S2000 and SCLIM induced weak environmental consequences for the mean of the entire Arctic region, the impacts on snow cover and soil temperature in specific regions are significant (figure 3). Comparing S2000 to SCLIM, the summer snow line was shown to retreat in northern Canada with a higher annual maximum soil temperature at a depth of 1 m. In this region, the active layer thickness (ALT) significantly increased with the soil temperature, while ALT decreased along the southernmost belt of Arctic, but this change was not significant (figure 3(f)). Besides soil temperature, albedo and latent heat are also significantly influenced by snow cover. With regard to the location of the snow fraction reduction, a decline in albedo and increase in latent heat were observed (figure 4). In July, the strongest albedo reduction occurs in the southern Arctic, with the largest LAI increase (figure 1(b)) and snow layer less than 0.4 m (figure 4(f)). In this condition, shrubs can protrude above the snow surface, reduce the surface albedo of snow and result in an increase in air temperature and melting (Sturm et al. 2001a).
In this study, we define the High Arctic as a region with an August mean snow area fraction >80% and the Low Arctic as a region with a snow fraction of <20% (figure 3(a)). The remaining area of the Arctic was defined as the buffer zone, where the snow cover fraction changes dramatically every year. By estimating the correlation coefficients between the variables related to the Arctic surface conditions in the different regions (i.e. High Arctic, buffer zone, Low Arctic, and whole Arctic; table 1), the major controlling factors of the soil temperature differences between the two runs were identified. The air temperature showed a significant positive relationship with soil temperature in all regions. In contrast, the canopy shade, which is represented by solar radiation absorbed by the ground in the Low Arctic, did not have a significant impact on soil temperature. Significant negative correlation coefficients between soil temperature at a depth of 1 m and snow fraction in the growing season suggest that snow fraction is a key factor affecting permafrost thawing in the buffer zone and High Arctic. Furthermore, deeper snow in the previous winter significantly warmed the soil only in the Low Arctic, suggesting a spatial limitation of the winter snow insulation effect.

To identify the major factors controlling the snow cover changes, correlation coefficients among the variables related to surface conditions were estimated (table 2). We inferred that less snow cover in the growing season reduced albedo and increased air temperature. Snow reduction was related to more melting in warmer weather, not less precipitation. Hence, snow melting creates a positive feedback loop for air temperature, where increases are amplified, as summarized in figure 5. In addition, snow melting is significantly accelerated by latent heat increase through water vapor-induced long-wave radiation changes (Wang et al 2015). Furthermore, the correlation coefficient between air temperature and LAI can be as low as 0.24, indicating that the annual perturbations in the meteorological variables are mainly caused by the unstable nature of the atmosphere rather than LAI. We eliminated these perturbations in the 34 year means (e.g. figures 3 and 4), which only contain the footprints for LAI forcing. However, these perturbations were meaningful for indicating the sensitivity of...
Figure 3. (a) The snow area fraction in August in SCLIM (34 year mean), (b) differences between the snow area fraction in August in S2000 and SCLIM (34 year mean), (c) annual maximum soil temperature at a depth of 1 m in SCLIM, (d) differences between the annual maximum soil temperature at a depth of 1 m in S2000 and SCLIM, (e) active layer thickness in SCLIM, and (f) differences between the active layer thickness in S2000 and SCLIM. The color shading indicates significance in a paired t-test.
Figure 4. (a) Albedo in July in SCLIM (34 year mean), (b) differences in July albedo between S2000 and SCLIM, (c) August latent heat in SCLIM, and (d) differences in August latent heat between S2000 and SCLIM, (e) July snow depth in SCLIM, (f) differences in August snow depth between S2000 and SCLIM. The color shading indicates significance in a paired $t$-test.
soil temperature in response to different variables, especially the impacts of snow fraction and air temperature. This analysis also suggests that increased vegetation does not induce significant increases in evapotranspiration, as indicated by the relationship between latent heat and LAI in table 2.

Comparing S2000 to SCLIM, the surface air is significantly warmer in central and northern Siberia, the Far East, Alaska, and northern Canada, while Greenland is the only region with cooler air (figure 6). The time-integrated LAI for S2000 is significantly higher than that for SCLIM in most regions around the Arctic circle, except for northern Alaska and sparsely vegetated land in northern Canada (figure 1b). The maximum LAI increase occurred in Siberia near 90°E, resulting in a vast area experiencing surface air

![Figure 6](image.png)

**Figure 6.** Snow fraction changes between S2000 and SCLIM in central and northern Siberia, the Far East, Alaska, and northern Canada.

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**Table 1.** Correlation coefficients of the soil temperature at a 1 m depth with multiple variables for three regions of the Arctic. All variables are the differences between S2000 and SCLIM.

| Variable                              | High arctic | Buffer zone | Low arctic | Whole arctic |
|---------------------------------------|-------------|-------------|------------|--------------|
| Snow fraction in the growing season   | −0.65\(^a\) | −0.77\(^a\) | −0.45      | −0.71\(^b\)  |
| Solar radiation absorbed by the ground| 0.76\(^a\)  | 0.68        | 0.54       | 0.62         |
| \(T_{air}\)                           | 0.78\(^a\)  | 0.54\(^a\)  | 0.56\(^b\) | 0.77\(^a\)   |
| Snow depth in the previous winter     | −0.01       | −0.12       | 0.30\(^a\) | −0.09        |
| LAI                                   | −0.24       | −0.03       | −0.24      | −0.18\(^a\)  |

\(^a\) Indicates significance at the \(p = 0.01\) level in the \(t\)-test of multivariable linear regression and.

\(^b\) Indicates significance at \(p = 0.05\).

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**Table 2.** Correlation coefficients for Arctic parameters. All variables are the differences between S2000 and SCLIM.

| Variable   | Precipitation | Snow melting | Albedo | Latent heat | \(T_{air}\) |
|------------|---------------|--------------|--------|-------------|-------------|
| Snow fraction | 0.05          | −0.64\(^a\) | 0.99\(^a\) | −0.87\(^a\) | −0.70\(^a\) |
| LAI        | 0.03          | −0.07        | −0.12  | 0.30        | 0.24        |
| \(T_{air}\) | 0.04          | 0.59\(^a\)   | −0.70\(^a\) | 0.57\(^a\)  |

\(^a\) Indicates significance at the \(p = 0.01\) level in the paired \(t\)-test.
heating. The Arctic surface is largely warmer in S2000 than SCLIM, as indicated by the higher LAI. This relationship is consistent with the results reported by Lawrence and Swenson (2011). However, the locations of LAI increase and air heating do not totally match, suggesting that other factors (i.e. atmospheric circulation and snow cover) affect air temperature. The deep soil is only warmer in limited regions of the Arctic, which matches the pattern of snow fraction reduction in the summer (figure 3(b)) rather than those of the LAI (figure 1(b)) or air temperature changes (figure 6(b)). These results suggest that deep soil temperature is directly affected by snow, and snow is influenced by changes in LAI, air temperature and melting.

4. Discussion and conclusions

From 1982 to 2015, the annual integrated LAI in the Arctic tundra increased over the first two decades and subsequently declined in the latter 14 years (Bhatt et al 2010). This study showed that vegetation greenness variations in the Arctic led to changes in surface air temperature, which were amplified by snow cover, eventually leaving a footprint on the soil temperature at a depth of 1 m, but these variations had a minimal effect on Arctic climate. Such mechanism is robust regardless of the time period we consider: only the period with increasing LAI, i.e. the first two decades, or longer periods until the recent time (not shown). Wang et al (2015) used satellite-based snow albedo in the coupled land-atmosphere simulations and found that surface air temperature in the Arctic is sensitive to albedo change. We found that snow cover is also sensitive to the change in surface air temperature, rather than precipitation. Hence, this positive feedback loop in snow cover and surface air temperature can amplify climate change in the Arctic.

Annual perturbations in surface air temperature strongly affected the annual maximum soil temperature in most Arctic regions. However, on the decadal scale, neither LAI-induced air warming nor canopy shading significantly impacted the local soil temperature at a depth of 1 m. In contrast, soil warming occurred along the summer snowline between the Low and High Arctic regions, highlighting the impact of snow cover amplification. The winter snow insulation effect is significant only in the Low Arctic. Summer snow cover in the High Arctic had a stronger but opposite impact on soil temperature by insulating the soil from warmer summer air and reflecting solar radiation.

Previous studies have shown that winter snow insulation can result in warmer soil and reduced permafrost in the following summer (Bulygina et al 2011). The results presented herein are in agreement with these observations in the Low Arctic (table 1), but the winter snow insulation effect is weaker than the effects of air temperature perturbations on an annual scale and does not leave a significant footprint on soil temperature at a depth of 1 m on the decadal time scale.

The winter snow insulation effect in the High Arctic observed in this study, was also reported in previous studies. Park et al (2015) simulated the Arctic soil temperature using a land surface model (a coupled hydrological and biogeochemical model), with climate

![Figure 6.](image)
forcing from 1901 to 2009. The authors showed that a 30% increase in winter precipitation resulted in significantly higher soil temperatures in the summer. However, we found that snow exerted the opposite effects in High Arctic summers. The summer snowpack insulates soil from the warmer summer air, reflects solar radiation, and results in a cooler ground. Although variation in summer snow cover showed a larger effect on summer soil temperature in the High Arctic, it was not considered by Park et al. (2015). Moreover, our study suggests that the fraction of summer snow is controlled by the melting and air temperatures during the growing season rather than winter precipitation.

This study shows that the canopy shading effect in the growing season does not have a significant effect on the soil temperature of vegetated Arctic lands. Blok et al. (2011) found that a larger canopy can shade and cool the ground in controlled field experiments in north-eastern Siberian tundra. Unlike Blok et al. (2011), this study was performed over vast sparsely vegetated regions and bare ground, where snow and air temperature play more significant roles.

Lawrence and Swenson (2011) investigated the permafrost response to shrub expansion using a coupled land-atmosphere global climate model and compared two opposite consequences of shrub expansion. The surface air heating effect was stronger in permafrost at a larger scale than that of the canopy shading effect. In this study, we reveal a more specific mechanism for air warming: Arctic air temperature perturbations are amplified through a positive feedback of linked variations in snow cover and albedo. This positive feedback loop amplifies the air temperature change at a larger spatial scale, while ground shading plays a smaller role and is not involved in this amplification process.

Recently, Zeng et al. (2017) evaluated the surface air temperature response to the global LAI increase over the past 30 years using a coupled land-atmosphere model. The authors showed that the LAI increase caused more evapotranspiration and earth surface cooling in several regions in the Arctic. However, our study shows that evapotranspiration over the Arctic is dominated by surface temperature, not LAI. This difference can be attributed to vegetation changes in the temperate and tropic regions included in model of Zeng et al. (2017), which affect the Arctic through large-scale atmospheric circulation. In addition, the LAI product used by Zeng et al. (2017) was derived from the AVHRR global inventory modeling and mapping studies (Zhu et al. 2013) based on the MODIS PFT product (Friedl et al. 2010), which overestimates LAI for the Alaska tundra (Verbyla 2007). Furthermore, our results show that latent heat has a strongly negative relationship with snow cover fraction, which is consistent with the findings of Westermann et al. (2009) in Svalbard which show that most latent heat fluxes occur during the snow melt period (particularly during the snow-free period in the summer and fall), while latent heat flux occurrences through snow sublimation are negligible.

This study characterized the Arctic snow cover amplification of a weak air-temperature change. This amplification may occur due to factors other than small vegetation changes, inducing large variations in the Arctic air temperature. We used historical sea ice and SST data to force the coupled land-atmosphere model. However, the sea ice extent is correlated with air temperature and snow (Blanchard-Wrigglesworth et al. 2015). Hence, including other aspects of natural systems will improve our understanding of the complex Arctic ecosystem.

Acknowledgments
This work was supported by the Korea Polar Research Institute (KOPRI and PE18900) and a National Research Foundation of Korea Grant from the Korean Government (MSIT; the Ministry of Science and ICT) (NRF-2016M1A5A1901769) (KOPRI-PN19081). Z Wang is also supported by Yonsei University Research Fund (Postdoctoral Research Supporting Program of 2017, 2017-12-0038). J Mao is supported by the Reducing Uncertainties in Biogeochemical Interactions through Synthesis and Computing Scientific Focus Area (RUBISCO SFA), which is sponsored by the Regional and Global Climate Modeling (RGCM) Program in the Climate and Environmental Sciences Division (CESD) of the Biological and Environmental Research (BER) Program in the US Department of Energy Office of Science. Oak Ridge National Laboratory is managed by UT-BATTELLE for the DOE under contract DE-AC05-00OR22725. The Global Land Cover (GLC2000) dataset used in this study was downloaded from http://forobs.jrc.ec.europa.eu/products/glcc2000/data_access.php.

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