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Snowmelt Events in Autumn Can Reduce or Cancel the Soil Warming Effect of Snow–Vegetation Interactions in the Arctic

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ABSTRACT

The warming-induced growth of vegetation in the Arctic is responsible for various climate feedbacks. Snow–vegetation interactions are currently thought to increase the snow-insulating capacity in the Arctic and thus to limit soil winter cooling. Here, we focus on autumn and early winter processes to evaluate the impact of the presence of erect shrubs and small trees on soil temperature and freezing. We use snow height and thermal conductivity data monitored near Umiujaq, a low-Arctic site in northern Quebec, Canada (56°N, 76°W), to estimate the snow thermal insulance in different vegetation covers. We furthermore conducted a field campaign in autumn 2015. Results show that the occurrence of melting at the beginning of the snow season counteracted the soil warming effect of snow–vegetation interactions. Refrozen layers on the surface prevented wind drift and the preferential accumulation of snow in shrubs or trees. Snowmelt was more intense in high vegetation covers, where the formation of refrozen layers of high thermal conductivity at the base of the snowpack facilitated the release of soil heat, accelerating its cooling. Consequently, the soil was not necessarily the warmest under high vegetation covers as long as melting events occurred. We conclude that under conditions where melting events become more frequent in autumn, as expected under climate warming, conditions become more favorable to maintain a negative feedback among the growth of erect vegetation, snow, and soil temperature in the Arctic, rather than a positive feedback as described under colder climates.

1. Introduction

In response to global warming, vegetation covers are changing in the Arctic (Hinzman et al. 2005; Post et al. 2009; Serreze et al. 2000). Shifts in species and abundance

are being observed in many Arctic and subarctic regions, and the boundaries between the various vegetation communities are moving. Woody plants benefit the most from these changes, with the forest–tundra ecotone (also called tree line) moving northward (Danby and Hik 2007; Harsch et al. 2009) and shrubs expanding on the tundra (Myers-Smith et al. 2011; Ropars and Boudreau 2012; Tape et al. 2006). The invasion of trees and shrubs (i.e., Arctic greening) is expected to feed back on climate either directly by decreasing surface albedo and increasing atmospheric water vapor concentrations, or indirectly by

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impacting soil temperature (Chae et al. 2015; Chapin et al. 2005; Lawrence and Swenson 2011; Pearson et al. 2013) and therefore carbon recycling.

Permafrost covers 25% of Northern Hemisphere lands and stores huge amounts of organic carbon (Hugelius et al. 2014). With permafrost thawing, conditions are becoming favorable for the microbial degradation and release of this carbon to the atmosphere as CO₂ and CH₄, constituting one of the strongest positive climate feedbacks (Schuur et al. 2015). However, the relationship between vegetation and soil temperature under climate change is still not quantitatively understood (Chapin et al. 2005; Pearson et al. 2013). The thermal regime of permafrost is correlated to atmospheric temperatures, but is also highly dependent on surface properties such as the presence of vegetation and snow. Both vegetation and snow act as buffer layers between the soil and the atmosphere by insulating the soil from atmospheric temperatures (Bertheaux et al. 2017). The insulating properties of the snowpack depend on its physical properties (i.e., height, density, thermal conductivity, and albedo; Domine et al. 2007; Zhang 2005). Snow physical properties evolve naturally over time through a process called snow metamorphism, which is highly affected by the presence of vegetation (Sturm et al. 2005a; Domine et al. 2016). All these components evolve together and are driven by and feed back on climate. It is thus crucial to study them simultaneously to understand their interactions within the climate–vegetation–snow–permafrost feedback loop and to estimate the net feedback on global climate.

Erect shrubs considerably impact snow physical properties, as shown by studies conducted in the Arctic. In Arctic Alaska, Sturm et al. (2001b) found that shrubs favor the accumulation of wind-drifted snow and the formation of low thermal conductivity snow layers, increasing the insulating capacity of the snowpack. In the Canadian high Arctic, Domine et al. (2016) reported that by trapping drifting snow, shrubs increase snow height, but only up to their own height. They measured snow density and thermal conductivity on herb tundra and in shrubs and found that shrubs reduced the density and thermal conductivity of the snow by limiting its compaction and enhancing metamorphism and depth hoar formation. All these effects significantly increased the insulating effect of the snowpack. These studies therefore suggest that shrubs limit soil winter cooling because they enhance the accumulation of snow with lower thermal conductivity. Warmer soil temperatures favor nutrient recycling (Saccone et al. 2013), which enhances vegetation growth; this therefore constitutes a positive feedback, as reported by Sturm et al. (2005a) and illustrated in the red feedback loop of Fig. 1.

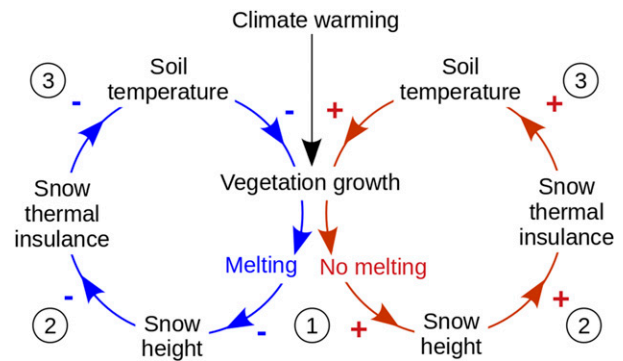


FIG. 1. Arctic vegetation–snow–soil interactions constituting negative and positive feedback loops depending on whether or not melting happens in autumn. The positive feedback loop (red arrows) has been suggested by Sturm et al. (2005a). The negative feedback loop (blue arrows) is proposed in this study. We focus on the three first steps influencing the soil temperature.

Other processes could further enhance this feedback. Sturm et al. (2005b) found that shrub branches protruding above the snow can reduce the surface albedo by up to 30%, further contributing to surface warming. The effect of shrubs on snowmelt in spring has also been studied by Pomeroy et al. (2006) and Marsh et al. (2010), and the general conclusion was that protruding shrubs highly affect the surface radiative budget. By absorbing more solar radiation in spring, shrub branches become warmer than the surrounding snow, resulting in both conductive heat transfer and longwave radiative transfer to snow. Both effects contribute to surface warming and thus accelerate snowmelt.

These previous studies, however, focused mostly on spring and, to a lesser extent, on winter processes, but there are hardly any observations of snow–vegetation interactions in autumn (mid-September–November), when the snowpack starts forming in the Arctic. This period is of major importance, as it is the time when soil freezing and intense snow metamorphism take place. These are fundamental Arctic processes that strongly impact average annual soil temperatures, and they therefore need to be understood to project permafrost thawing. Studying autumn processes may be critical because, for example, the feedback proposed by Sturm et al. (2005a) does not account for melting events that are likely to take place in autumn, when decreasing temperatures hover around 0°C for a while. Melting affects wind drift because refrozen layers on the surface prevent the erosion of snow by wind and therefore the preferential accumulation of snow in high vegetation covers. Shrubs or trees may also affect autumn melting. In fact, Domine et al. (2016) found more signs of autumn snow melting in shrubs in the high Arctic, suggesting that lower surface albedo caused by protruding branches

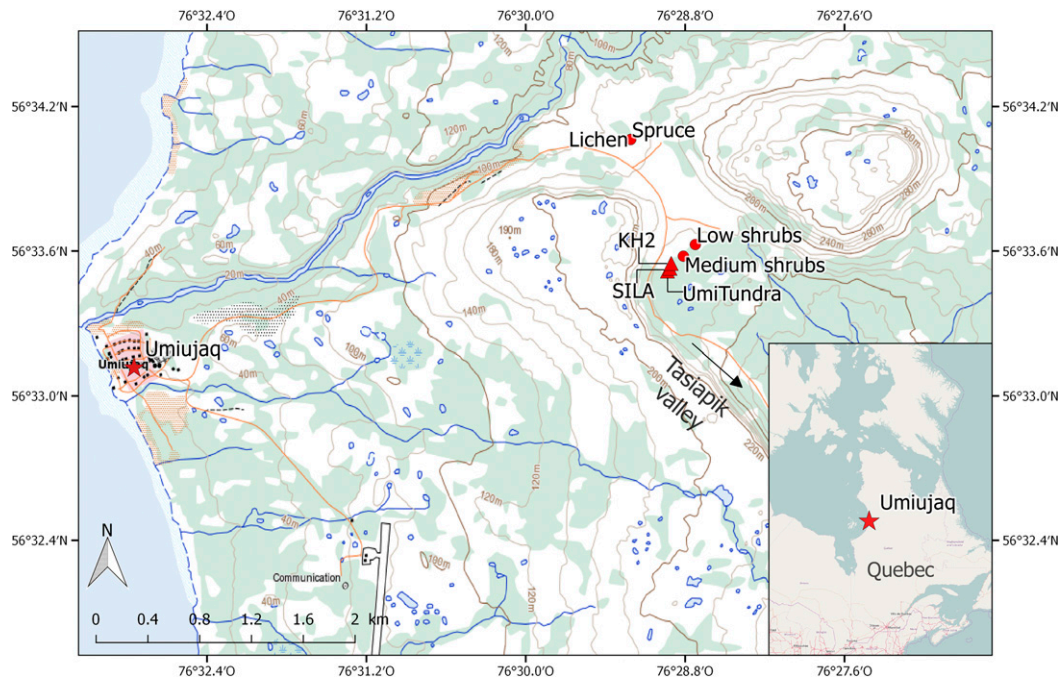


FIG. 2. Location of the general study area and of specific sites near Umiujaq in northern Quebec, Canada. The automatic weather stations SILA, KH2, and UmiTundra are indicated with triangles, and snow pits are indicated with circles. Map source: Natural Resources Canada (<http://atlas.gc.ca/toporama/en/index.html>).

can facilitate snowmelt and the formation of refrozen layers with high thermal conductivity. More melting in erect vegetation also means thinner snow. If the snow in shrubs or trees is thinner and more conductive than that on lichen tundra, then the red feedback loop of Fig. 1 may be reduced or even canceled.

The purpose of this study is to test the hypothesis that the presence of erect shrubs or trees can, in some cases, lead to a negative feedback where snow accumulation in these high vegetation covers is limited because drifting is impeded by refrozen layers on the surface. Also, more intense melting leads to thinner snow with higher thermal conductivity, which in turn results in faster soil cooling under erect vegetation. This feedback is featured with blue arrows in Fig. 1. Of course, most observations will probably witness a combination of both feedbacks, complicating the identification of the blue component. Here, we use automated measurements of meteorological, snow, and soil conditions near Umiujaq ($56^{\circ}33'31''\text{N}$, $76^{\circ}28'56''\text{W}$), a low-Arctic community on the eastern Hudson Bay shore in Nunavik, as well as manual observations and measurements performed during an extended campaign in autumn 2015 to attempt to detect the actions of the negative feedback proposed in Fig. 1. With continued warming, there is the possibility that this negative feedback may grow in importance, which motivated this investigation.

2. Methods

a. Study area

The study site near Umiujaq, northern Quebec (Fig. 2), is situated in the forest–tundra ecotone and features most vegetation types found in the Arctic (lichen tundra, low- and high-shrub tundra, forest tundra), providing a unique opportunity to study the effects of the lichen–shrub–forest transition on snow properties and the resulting soil temperature. Starting in 2012, we began deploying instruments to measure snow and soil variables including thermal conductivity (Domine et al. 2015), thus complementing atmospheric data recorded since 1997. In addition, we conducted a field study during 6 weeks, from 24 October to 7 December 2015. The variables studied then were snow height and stratigraphy and vertical profiles of snow density, thermal conductivity, and temperature. To our knowledge, this is the first study measuring snow physical properties of an Arctic snowpack in autumn that considers the effects of different vegetation types.

Near Umiujaq (Fig. 2), the mean annual temperature measured between 1997 and 2015 is -3°C , and the annual precipitation ranges from 700 to 1000 mm for the 2012–15 period (CEN 2016). The particularity of this region comes from the cooling influence of Hudson Bay, repelling the tree line farther inland and leading to an

abrupt transition between lichen and shrub tundra near the coast and forest tundra a few kilometers inland (Gregoire and Begin 1993). Wind-sheltered areas host forest patches mostly comprising black spruce (*Picea mariana*), while uplands are colonized by lichens (*Cladonia stellaris*) and shrubs (*Betula glandulosa*, *Salix* sp. -mostly *planifolia*- and *Alnus crispa*). Dwarf birch (*B. glandulosa*) is the dominant shrub species, whose height ranges from 20 to 200 cm. Situated in the discontinuous permafrost zone (Allard and Seguin 1987), the region is undergoing widespread permafrost degradation. Numerous thermokarst hollows and ponds are visible at the surface, and only a few permafrost mounds remain (Payette et al. 2004). Shifts in vegetation have also been observed during the last decades, with shrubs expanding onto open tundra at the expense of lichens (Provencher-Nolet et al. 2014; Ropars and Boudreau 2012).

Close to the village, the Tasiapik Valley goes through these different ecosystems to the Guillaume-Delisle Lake to the southeast. The upper part of the valley gives the opportunity to study sites dominated by various vegetation types within 1 km, so that meteorological conditions show very little spatial variation (Fig. 3). Low shrubs (dwarf birch; 20–30 cm high) grow in isolated patches on lichen areas. Medium shrubs are mostly 60–80-cm-high dwarf birches regrouped in bushes of larger extent. Isolated spruce stands (200–250 cm high) are also observed. These are relicts of a warmer past (Payette et al. 1994; Laberge et al. 2000) and appear as thickets of trees about 10 m long in the middle of shrub tundra, looking like krummholz. We focus here on the plant size, as it is probably the primary factor influencing the snow-insulating properties (Domine et al. 2016; Sturm et al. 2001b). Other plant characteristics, such as the branch density, may of course have some impact, but evaluating these will require further study. We therefore here use plant height as the metric to characterize vegetation, while being aware that more complex metrics may be needed in the future.

Three automatic weather stations have monitored the meteorological conditions of our study area, as well as snow and soil properties (Fig. 3), since at least 2015. Installed in a lichen and low-shrubs area (dwarf birch; 30 cm high), the UmiTundra station records fairly complete meteorological data (air temperature and humidity and wind speed measured at 2.3-m height, upwelling and downwelling shortwave and longwave radiation), along with snow height with an SR50A acoustic gauge (Domine et al. 2015). Note that this SR50A snow gauge is situated above a lichen area surrounded by patches of low shrubs. A few meters away, in the same vegetation cover, a 10-m tower (hereafter called SILA) records air temperature and humidity,

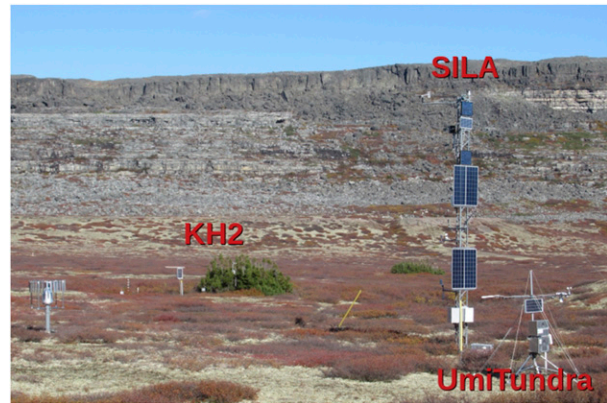


FIG. 3. Weather stations KH2, SILA, and UmiTundra located in the Tasiapik Valley on 26 Sep 2017. The general area is covered with lichens and low (30 cm) to medium (70 cm) dwarf birches. Spruce stands like KH2 are isolated thickets of trees on the shrub tundra.

wind speed and direction at 10 m, snow height above lichen, and precipitation rate (CEN 2016). Finally, instruments were deployed in the middle of the spruce stand KH2 and next to it in medium shrubs (dwarf birches 60–70 cm high) in autumn 2014. In particular, SR50A gauges measure snow height at both sites. Variables are measured every minute, and hourly averages are recorded.

Snow thermal conductivity and temperature are measured with Hukseflux TP02 or TP08 heated needle probes (NPs) installed at heights above the lichen or moss surface of 5, 15, 30, 47, and 64 cm in low shrubs (UmiTundra station); at 2, 10, and 30 cm in medium shrubs; and at 4, 12, and 32 cm in spruce (KH2 station). Briefly, the method consists of monitoring the temperature rise during a 100-s heating cycle, whose rate depends on the thermal conductivity of the snow (Domine et al. 2015; Morin et al. 2010). When the snow temperature is close to the melting point, the heated needle can trigger snowmelt and alter both the measurement and the snow microstructure. To avoid that, measurements are taken once a day at 0500 LT (UTC – 5 h) and only if the snow temperature is below -2°C . About 50 cm from the NPs, Decagon 5TM sensors record soil temperature and liquid water content at the same three sites (spruce, medium, and low shrubs) at depths of 5, 10, 15, 30, and 50 cm. Additional 5TM sensors were installed in soil under lichen about 3 m away from the low-shrubs sensors. Finally, time-lapse cameras on the SILA tower provide several pictures a day of the KH2 and UmiTundra sites.

b. Experimental methods

Our study investigated snow physical properties in relation to vegetation cover, including lichen, shrubs (low, medium, and high), and spruce. We have particular

interest in the snow thermal properties (i.e., height and thermal conductivity) because they directly govern the heat flux F between the soil and the atmosphere following Fourier's heat equation:

$$F = -k_{\text{eff}} \frac{dT}{dz}, \quad (1)$$

where dT/dz (K m^{-1}) is the vertical temperature gradient through the snow, and k_{eff} is the snow effective thermal conductivity ($\text{W m}^{-1} \text{K}^{-1}$), so named because it is an effective parameter that includes the effects of several processes, such as conductive heat transfer and latent heat transport by sublimation–condensation cycles during snow metamorphism (Sturm et al. 1997). As the snowpack is a layered medium, its insulating capacity can be characterized by its thermal insulance R_T ($\text{m}^2 \text{K W}^{-1}$), given as:

$$R_T = \sum_i \frac{h_i}{k_{\text{eff},i}}, \quad (2)$$

with h_i and $k_{\text{eff},i}$ are the thickness and thermal conductivity of snow layer i , respectively.

At each study site, areas of about 20 m^2 were dedicated to digging several successive snow pits during autumn where we measured the vertical profiles of snow properties. Stratigraphies were described, and the density profiles were measured with 3-cm-resolution using a 100-cm^3 box cutter. Snow thermal conductivities were measured with a TP02 heated needle probe. It can be difficult or impossible to measure snow thermal conductivity with an NP when the snow is too warm, so that some expected k_{eff} data could not be obtained when conditions were too warm. The time-lapse pictures were also used to assess the snow cover evolution.

The accuracy of the NP method to measure snow thermal conductivity has been discussed in detail by Domine et al. (2015). Sources of error include random errors and systematic errors. Systematic errors can reach 20%; however, here we compare thermal conductivity values obtained with the same method, and relative values are more important than absolute ones. Domine et al. (2015) mention a random error of the NP method of 5% and a 3% error due to the algorithm used to analyze the heating curve, so that the total random error of interest for a comparison of values is 6%.

3. Results and discussion

a. Wind events and occurrence of drifting

Local meteorological conditions govern snow accumulation through air temperature for the precipitation

phase and wind speed for snow redistribution (Fig. 4). The 2015/16 winter was the warmest winter on record in the Arctic (Cullather et al. 2016; WMO 2016). At our study site, measurements from the UmiTundra station indicate that air temperature during the beginning of the winter (September–January) was, on average, 2°C warmer in 2015/16 than the two previous winters and 1°C warmer than in 2016/17. Snow began to accumulate on 17 October 2015, 10 days earlier than the two previous winters. When we arrived on site on 24 October, the snow cover was already between 0 and 50 cm thick and was wet because of positive temperatures and rain (Fig. 4). The thermal gradient between the unfrozen sea and the snow-covered land was intense during the day, triggering winds and blizzards strong enough to prevent any field measurement some days. An intense warm spell occurred between 18 and 20 November 2015, with positive temperatures, rain on snow, and strong winds triggering intense melting and the disappearance of the snow cover in many places. Immediately afterward, a significant snowfall restored the snow cover under windy conditions, with important snow redistribution. The weather remained wet and windy until late December, when the sea finally froze. The air temperature remained definitely below 0°C after 26 December 2015, about 1 month later than usual. In autumn 2016, the snowpack formed on 12 October, earlier than in autumn 2015, and went through numerous warm spells, but only until mid-November (Fig. 4). Rain-on-snow events occurred each time the temperature exceeded 1°C . Solar radiation was very low at that time of year, about 10 times lower than measured by Marsh et al. (2010) in spring, while the longwave radiation was similar. Hence, the radiative budget was very different to spring because the shortwave contribution was much reduced, while the longwave radiation had a dominant effect. Given that protruding shrubs are known to increase longwave radiative transfers to snow (Pomeroy et al. 2006), we expect snow to be warmer in high vegetation covers.

Large heterogeneities in snow accumulation have a strong influence on the local environment and in particular on the soil temperature (Sturm et al. 2001a). The driving mechanism is the wind-induced snow redistribution, which is affected by the vegetation trapping effect (step 1 of Fig. 1). Usually, snowdrift occurs for wind speeds $\geq 6 \text{ m s}^{-1}$, depending on snow properties (Vionnet et al. 2013).

Many events in both years illustrate the various steps in both the red and blue feedbacks of Fig. 1. Here, we focus on the inhibition of the first step of the positive feedback, where the presence of shrubs or spruce does not increase snow height during wind events when surface melting takes place. We describe three episodes:

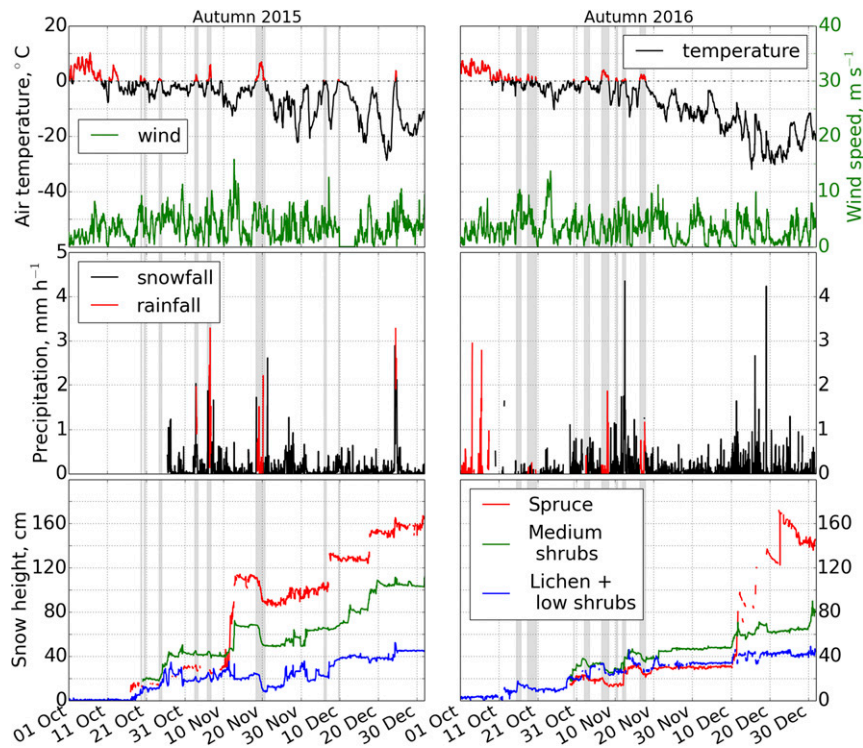


FIG. 4. Evolution of meteorological conditions at 2.3-m height and snow heights during the beginning of the (left) 2015/16 and (right) 2016/17 snow seasons. Snow heights were automatically measured at UmiTundra (lichens with patches of low shrubs) and KH2 (medium shrubs and spruce) stations. A connectivity issue affected the precipitation gauge until we fixed it on 26 Oct 2015. Gaps in snow height were caused by faulty ultrasonic sensors. They were replaced on 28 Oct 2016 at KH2. The shaded areas indicate the occurrence of warm spells (i.e., episodes with positive air temperature while snow was covering the ground).

6 November 2015, 25 December 2015, and 16 November 2016 (and following days each time). During these periods, most branches of the low shrubs were buried in snow, so their trapping effect was suppressed. On 6 November 2015, it rained under air temperatures reaching 6°C , leading to a slight decrease in snow height everywhere (Fig. 5a). At night, it froze, and a refrozen crust formed while wind speed reached 10 m s^{-1} . Despite these strong winds, no drifting took place, and the three snow gauges indicate unchanged snow heights. Interestingly, the small snowfall of 7 November under negative temperatures led to preferential accumulation on lichen and low shrubs, but this was transitory and only illustrates the wind-induced motion of drifting snow. The main conclusion from the 6 November event was that it led to similar decrease in snow height everywhere and completely suppressed snow drift and therefore the preferential accumulation in shrubs.

On 24 December 2015 (Fig. 5b), a day-long mixed snow-rain event took place. Winds were moderate during precipitation (5 m s^{-1}) but increased to 8 m s^{-1} in the evening after precipitation had ended and the

atmosphere had cooled down. The snow at the surface was wet because of positive temperatures during precipitation, after which it rapidly froze because of a 10.6°C temperature drop in 6 h, and no drifting was observed. The event nevertheless resulted in accumulation, and it is noteworthy that the most important accumulation was on lichen (+15 cm) and the lowest in medium shrubs (+7 cm). The following night, winds reached 10 m s^{-1} , but no drifting at all was observed.

On 16–17 November 2016, it rained, and snow height decreased similarly everywhere (Fig. 5c). On 19 November, winds reached 10 m s^{-1} , but no drifting took place. The subsequent small changes in snow height were due to small snowfalls at the end of that day, as confirmed by time-lapse photographs.

In summary, the observations and data presented here demonstrate that melting can prevent drifting under wind speeds up to 10 m s^{-1} , and we even observed enhanced accumulation on lichen in one instance (Fig. 5b), possibly because the mixed-phase accumulation included more water in medium shrubs than on lichen, as it was probably slightly warmer there. This illustrates that

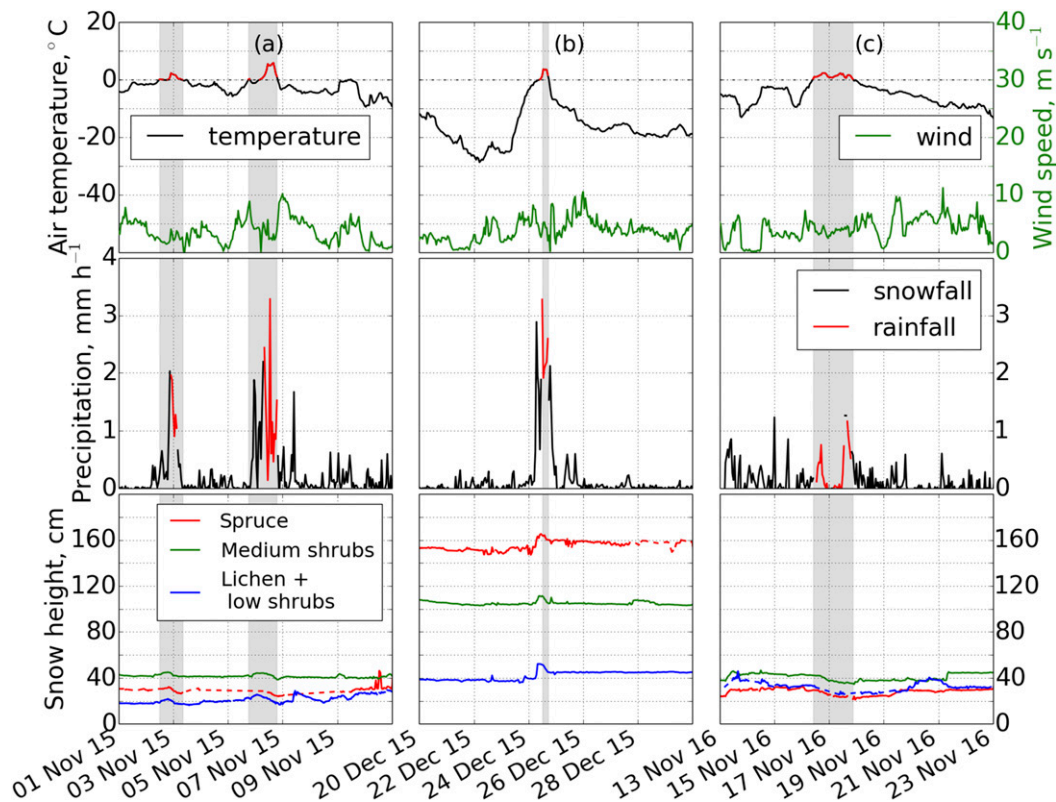


FIG. 5. Evolutions of atmospheric conditions and snow heights during three wind events when snowmelt occurred: (a) 6 Nov 2015, (b) 25 Dec 2015, and (c) 17 Nov 2016. Missing snow height data are filled with dashed lines.

under warm conditions, the first stage of the positive feedback of Fig. 1 can be stopped.

b. Snow thermal insulance

To investigate step 2 of Fig. 1, we examined the evolution of snow thermal insulance R_T , measured automatically at the different stations to assess the consequences of vegetation growth on the snow thermal properties in autumn. The R_T values were calculated from Eq. (2) using snow height and thermal conductivities monitored simultaneously at low (30 cm) and medium (60–70 cm) shrubs and spruce. Several k_{eff} values are missing because snow temperatures were too warm for measurements. In these cases, we partially filled the data gaps with available values for the same layer in the previous or next days. We also made use of our manual measurements from the autumn 2015 campaign. This makes the assumption that k_{eff} varied little in a given layer within a few days, which is usually true for layers at the bottom of the snowpack if they are not affected by melting. Otherwise, we extrapolated values measured at other heights. Also, the absence of k_{eff} sensors above 30 cm at KH2 sites limits the accuracy of the R_T estimation. When the snowpack was thicker than 30 cm,

upper fresh snow layers of low k_{eff} were not considered anymore. This could cause a negative bias (underestimation) in the R_T estimation in spruce and medium shrubs. However, with the settling of the snowpack and the increase in k_{eff} , the negative bias decreases over time. Nevertheless, the snowpack is relatively thin at this period, and the comparison of R_T between sites and of their respective evolutions shows interesting behaviors.

Figure 6a shows that the snow R_T in spruce was lower than in low and medium shrubs in late October and early November until the occurrence of a major drift event on 12 November 2015. The same applies to Fig. 6b during most of autumn until drifting occurred on 11 December 2016. This is explained by a fast increase in snow thermal conductivity in spruce, as a result of a higher compaction rate and probably stronger melting than in shrubs. Indeed, the canopy structure is different in spruce and birch. Birches have a network of small branches dense enough to hold up the snow and prevent compaction, while spruce just have a few larger branches whose network is not dense so that snow compacts easily under its own weight. The presence of needles and the greater height of the trees optimize radiation absorption and therefore melting. These effects, in combination with

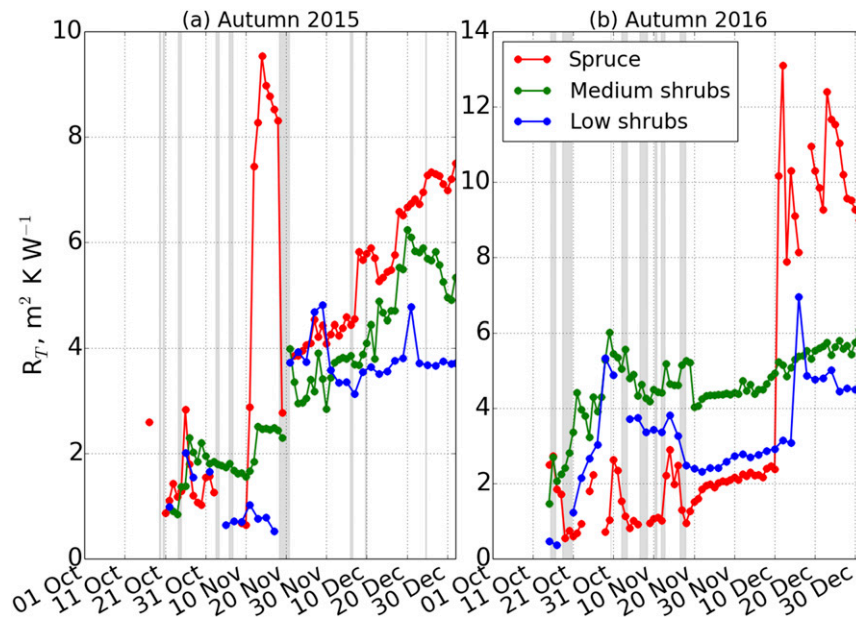


FIG. 6. Evolution of snow thermal insulances R_T at the different sites during autumn and early winter (a) 2015/16 and (b) 2016/17. Missing R_T values are due to the lack of k_{eff} measurements when snow layers were too warm. The shaded areas indicate the occurrence of warm spells (i.e., positive air temperatures).

the lack of preferential accumulation in spruce (inhibition of step 1 of Fig. 1), leads to the inhibition, or even reversal, of step 2.

In shrubs, the low k_{eff} values produce a higher R_T value. Two periods are, however, noteworthy when R_T was almost as high as or higher in low than in medium shrubs: 20–30 November 2015 and 21–31 October 2016. The intense warm spell of 18–20 November 2015 triggered snowmelt throughout the whole snowpack, which strongly reduced R_T everywhere. Areas where snow accumulated the least (i.e., where the snow water equivalent was the lowest) totally melted out. On 21 November, fresh snow fell on bare ground or on a thin (≤ 10 cm) old snow layer on lichen and low-shrubs sites, whereas about 50 and 90 cm of refrozen snow of high k_{eff} remained at medium-shrub and spruce sites, respectively (Fig. 4). The thinner snow cover favored the establishment of a large thermal gradient ($>100^\circ\text{C m}^{-1}$) that rapidly formed low thermal conductivity faceted grains. Consequently, the snowpack and especially its bottom layer became more insulating in lower vegetation covers, and R_T was the highest in low shrubs until the end of November 2015 (Fig. 6a).

Neither snow gauge of the KH2 station was functioning in October 2016, so we used snow height data from the UmiTundra station to calculate R_T between 12 and 29 October 2016. It should not affect the results, given the homogeneous accumulation observed on

29 October, when the gauges were repaired. The R_T value in spruce shows rapid increases caused by snow accumulation, followed by rapid decreases caused by the partial melting of the whole snowpack and the resulting increase in k_{eff} on 19 October and on 3 and 16 November (Fig. 7). Snowmelt was less intense in shrubs where basal layers were not affected and thus kept very low k_{eff} values ($0.03\text{--}0.05 \text{ W m}^{-1} \text{ K}^{-1}$). Upper layers affected by

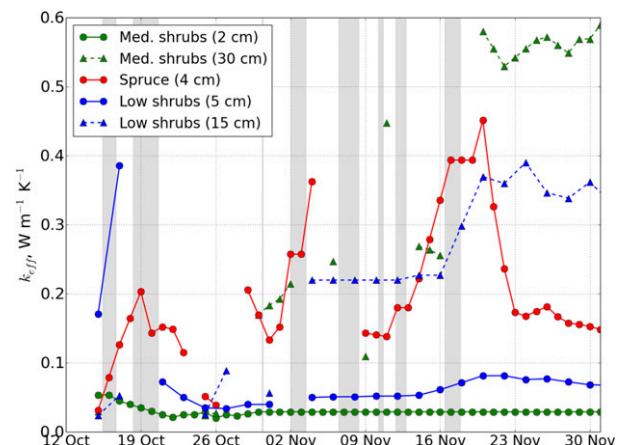


FIG. 7. Evolution of snow thermal conductivities k_{eff} automatically measured at the station sites during autumn 2016. Values are missing when snow layers were too warm to perform k_{eff} measurement.

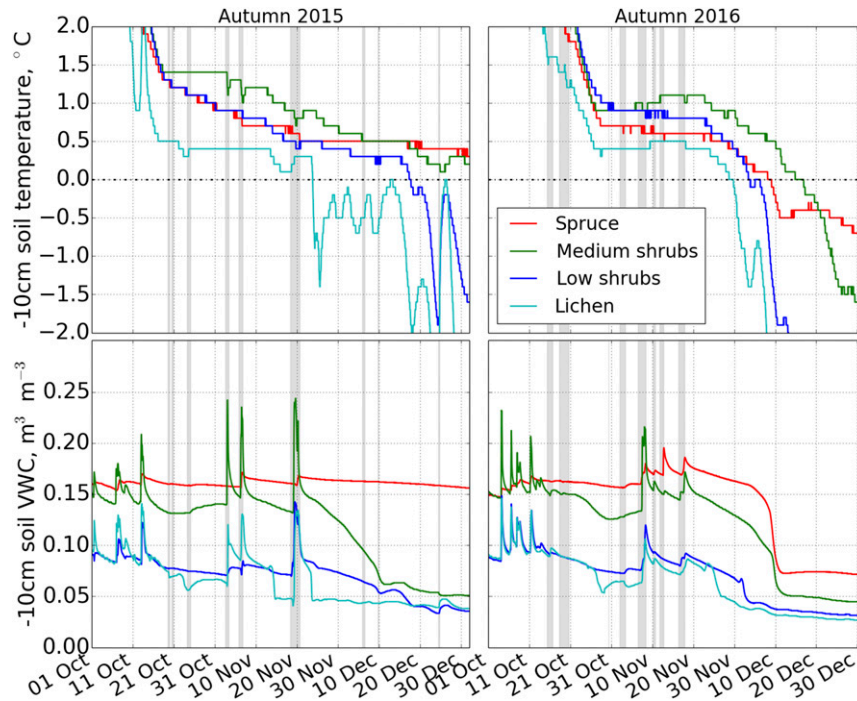


FIG. 8. Evolution of soil temperatures T and volumetric water contents (VWC) at 10 cm below the surface.

snowmelt in shrubs showed, on the contrary, very high k_{eff} values, especially in medium shrubs (Fig. 7). It indicates, expectedly, that snowmelt affected the snow in high vegetation covers more. Consequently, the occurrence of melting impeded the second step of the feedback of Fig. 1, consistent with the blue loop.

c. Soil temperature

The third step of the loops of Fig. 1 refers to the impact of snow properties on soil temperature. Figure 8 shows the soil temperature monitored at 10 cm below the surface during autumns 2015 and 2016. The volumetric water content (VWC) is also shown, as both soil VWC and snow R_T highly influence soil cooling and freezing processes. In both years, the coldest soil temperatures were recorded under lichen, while the water content there was similar to that of low shrubs. The impact of air temperature fluctuations is the most visible under lichen, attesting to low snow thermal insulance at this site. The soil under spruce was always colder than or at least as cold as soil under low and medium shrubs until the formation of thick drifts in spruce. This is remarkable, given that the soil water content was the highest under spruce and thus delayed freezing. The fast cooling of the soil under spruce in autumn is related to the high k_{eff} values, which conferred a low insulating capacity to the snow and thus facilitated soil heat loss (Fig. 6).

Soil temperature was the warmest under medium shrubs, but Fig. 8 shows periods when the soil under low shrubs cooled slower. For example, between 22 November and 7 December 2015, the temperature change at -10 cm was $-0.02^\circ\text{C day}^{-1}$ under low shrubs and $-0.03^\circ\text{C day}^{-1}$ under medium shrubs. Between 22 and 31 October 2016, these rates were -0.11° and $-0.18^\circ\text{C day}^{-1}$, respectively. It corresponds to the two periods described in the previous paragraph, when melting appeared more intense in high vegetation covers, thus accelerating soil heat loss through the snow.

These results demonstrate that the soil warming effect of snow–vegetation interactions is not always predominant. Both aspects of red (soil warming) and blue (soil cooling) loops of Fig. 1 are visible on Fig. 8. This figure also attests that the positive feedback between vegetation growth and soil temperature predominates after melting events cease in December. Hence, the sign of the feedback among snow, vegetation, and soil depends on the occurrence of melting. There are many cases where the feedback switched due to the changing conditions. Overall, our observations indicate that each time snow melted in autumn, the negative snow–vegetation–soil feedback occurred. Because the negative feedback kept manifesting itself after melting, it is difficult to determine when exactly the feedback became positive again. Our data of Figs. 4, 6, and 8 show that in

2015, between snow onset and 31 December, melting took place for 11 days. If we estimate that the effect of melting continued 2 days after it ceased, then the negative feedback took place 43% of the time until 31 December. For 2016, it was 44% of the time. This shows that for a fairly cold site such as Umiujaq, the negative feedback, although very important, does not predominate but can be expected to gain importance with continued warming. In addition, the investigations of the decreases in autumn soil temperature intuitively focus on sensible heat, while latent heat effects are also important. Because more vegetation means wetter soil, this latent heat effect will apparently favor the red loop and mask episodes where the blue loop may be predominant. In summary, we found that snowmelt events in autumn can reduce or even cancel the soil warming effect of vegetation growth in the Arctic.

4. Conclusions

The vegetation–snow–soil interactions at the beginning of the snow season are difficult to predict because they are highly sensitive to meteorological conditions (i.e., air temperature, wind, and radiation). We found that the occurrence of melting events in autumn reduced the insulating effect of snow–vegetation interactions described in earlier work, where melting was not considered (Domine et al. 2016; Sturm et al. 2001b). The appearance of liquid water and its refreezing increased the cohesion of the surface snow and therefore prevented drifting. These conditions were frequent in autumn because of numerous warm spells in October and November, which therefore delayed the redistribution of snow by wind and the formation of thick snowpacks over erect shrubs and trees. In addition, snowmelt appeared to be more intense in high vegetation covers, which we explain by greater energy transfers, both radiative and conductive, from the branches to the snow. The formation of refrozen layers of high thermal conductivity facilitated the release of soil heat and accelerated its cooling. This effect was particularly strong in spruce, while in shrubs, the canopy structure, by limiting compaction, helped maintain a high insulation capacity at the base of the snowpack.

The December data of Fig. 8 indicate that at this low-Arctic site, the positive feedback between vegetation growth and soil temperature predominated after December, when melting events ceased. The exceptionally warm autumn 2015 triggered a longer and stronger negative feedback compared to colder years. This situation was favorable to observe the soil cooling effects of snow–vegetation interactions (i.e., the negative feedback of Fig. 1). While we document this negative

feedback with detailed observations at just one site, it is probably quite common since melting events occur throughout the Arctic in autumn (Forbes et al. 2016; Langlois et al. 2017; Sokolov et al. 2016). We have also observed the presence of melt–freeze crusts and other signs of melting at other times and places in Arctic shrub patches [e.g., at Umiujaq (Domine et al. 2015) and Bylot Island (Domine et al. 2016)] but also on Arctic herb tundra (e.g., at Barrow, Alaska; now known as Utqiagvik; Domine et al. 2012).

Umiujaq is still a cold site, as shown in the detailed meteorological data of Domine et al. (2015). With continued warming, rain-on-snow and melting events can be expected to become more frequent (Hansen et al. 2014; Liston and Hiemstra 2011). In addition to the growth of vegetation, these conditions may increase the probability of snowmelt to occur in autumn and early winter and thus amplify the negative feedback represented by the blue loop of Fig. 1. We therefore expect the positive feedback among vegetation growth, snow thermal insulation, and soil temperature to decrease in strength or perhaps even reverse under warmer climates because of the increase in both vegetation height and the frequency of snowmelt events.

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