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Impact of snow reduction on decomposition and soil respiration

Vliv úbytku sněhu na dekompozici a respiraci půdy

Bachelor´s thesis

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Prohlášení

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V Praze dne 30.4.2025

Anežka Klárová

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Abstract

The global decline in snow cover affects biochemical processes in soil, potentially leading to changes in the carbon cycle. This thesis describes global and regional snow cover trends and summarizes current knowledge and methods used in decomposition and soil respiration research. Soil biochemical processes during winter depend on the frequency of freeze-thaw cycles, litter input, and the activity of soil biota. However, the results of individual studies are inconsistent due to varying experimental conditions across ecosystems and the lack of standardized methodical approaches. Decomposition and soil respiration contribute to the annual carbon exchange even under reduced snow conditions and may alter the annual carbon balance between soil and the atmosphere.

The objective of the practical part of this thesis was to study the effect of snow reduction on winter soil temperature, decomposition, and respiration at Králický Sněžník. An observation of plots with different snow depth combined with snow manipulation experiment was conducted in two winters, 2023 and 2024. Significantly lower decomposition was found with lower snow depth in plots measured in spring 2024. Significantly lower soil respiration was found with lower snow depth in plots measured in winter 2024. Decomposition was influenced by long-term temperature conditions, whereas soil respiration responded to recent soil temperatures.

Focusing on the impact of snow reduction on decomposition and soil respiration is essential for understanding potential changes in the carbon cycle and for predicting future climate change.

Keywords: global warming, albedo, carbon cycle, CO_2 flux, snow manipulation, litterbags

Abstrakt

Globální úbytek sněhové pokrývky ovlivňuje biochemické procesy v půdě, které vedou k potenciálním změnám uhlíkového cyklu. V této práci jsou zhodnoceny globální i regionální trendy sněhové pokrývky a shrnuty poznatky a metody ve výzkumu dekompozice a respirace půdy. Půdní biochemické procesy během zimy závisí na četnosti cyklů zamrzání a tání, přísunu opadu a aktivitě půdní bioty. Výsledky jednotlivých studií však nejsou konzistentní vlivem odlišných experimentálních podmínek napříč ekosystémy a nejednotným metodickým přístupem. Dekompozice a respirace půdy přispívají k jejich celoroční míře i za redukovaných sněhových podmínek a potenciálně mění celoroční bilanci uhlíku mezi půdou a atmosférou.

Cílem praktické části práce je odpovědět na hypotézy týkající se vlivu úbytku sněhu na teplotu půdy, dekompozici a respiraci půdy na Králickém Sněžníku. Zkoumání ploch s různými výškami sněhové pokrývky v kombinaci s experimentem manipulace sněhu bylo prováděno během dvou zim, roku 2023 a 2024. Signifikantně nižší respirace půdy byla za nižší sněhové pokrývky na plochách měřených na jaře 2024. Signifikantně nižší dekompozice byla za nižší sněhové pokrývky na plochách měřených v zimě 2024. Dekompozice byla ovlivněna dlouhodobými teplotami, zatímco respirace půdy reagovala na aktuální stav teploty půdy.

Zaměřit se na vliv úbytku sněhu na dekompozici a respiraci půdy je zásadní k porozumění potenciálních změn cyklu uhlíku a předpovídání budoucí klimatické změny.

Klíčová slova: globální oteplování, albedo, cyklus uhlíku, tok CO_2, manipulace sněhem, opadové sáčky

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1 Introduction

Around one third of the Northern Hemisphere land surface is seasonally snow-covered (Robinson, 2023). Due to its direct snowmelt response, snow cover is highly susceptible to global warming. Areas most vulnerable to snow cover changes are high latitude and altitude areas, but the snow cover is projected to decrease globally (Pongracz et al., 2024). Climate is the main factor in changing the snow cover.

In turn, snow cover reduction causes positive feedback to global warming. With reduced snow cover extent (SCE), land surface albedo decreases, amplifying ground heating and subsequently contributing to climate change (Edwards et al., 2007).

One of the key aspects of this phenomenon is the impact of snow reduction on soil. Snow is an insulating layer, protecting the soil from extreme temperature fluctuations and preventing it from freezing during winter (Maurer and Bowling, 2014). When snow cover decreases, the soil is more frequently exposed to temperatures fluctuating around the freezing point, which leads to freeze-thaw cycles (FTC) (Hobbie and Chapin, 1996). These freezing conditions can cause mechanical damage to soil structure and the mortality of soil organisms, affecting the soil in two opposing ways. While the soil may become enriched with new nutrients due to the mechanical disturbance, the abundance of microbial communities tends to decrease. During thaw periods, soil biota can become reactivated (Hobbie and Chapin, 1996). The overall impact of FTC on soil biota is highly variable, which results in an ambiguous effect on soil biochemical processes.

In this context, decomposition and soil respiration processes are critical, as they play a key role in transforming and releasing carbon from the soil into the atmosphere. Soil is a significant carbon reservoir, and these processes substantially influence the global carbon cycle. The decomposition of organic matter leads to the release of carbon through heterotrophic respiration by microorganisms, while autotrophic respiration occurs through the metabolic activity of plant roots (Šimek et al., 2019). Soil conditions altered by snow cover reduction can influence these processes. At lower temperatures, a general slowdown in soil biochemical processes is expected. However, temperature fluctuations and moisture availability can cause variability in these processes. The presence of active microorganisms and the amount of available organic matter are also critical factors. With increased soil biota activity, decomposition and soil respiration are enhanced. Importantly, their intensity throughout the winter plays an appreciable role in annual rates of decomposition and soil respiration (Bond-Lamberty et al., 2024; Gavazov, 2010).

Previous research on the effects of snow cover reduction on soil processes has often focused on specific biomes, targeted only the vegetation season, or examined partial aspects such as microbial abundance or the occurrence of various carbon forms (Blankinship and Hart, 2012;

Bokhorst et al., 2013; Edwards et al., 2007; Kosolapova and Altshuler, 2024; Li et al., 2016). The outcomes of these studies also vary depending on the characteristics of the site, the length of the study period, and the data collection methodology. Finally, a comprehensive literature review of decomposition and soil respiration under reduced snowpack has been missing.

This thesis aims to describe trends in snow cover, summarize current knowledge on the effects of snowpack reduction—particularly on soil decomposition and respiration—and review the methods used in research on this topic. The literature review is complemented by the results of an observation study combined with snow manipulation experiment conducted at Králický Sněžník, which tests the effect of snow reduction on soil temperature, decomposition and respiration. Finally, the thesis aims to highlight existing research gaps and the potential for further investigation in this field.

2 Theoretical part

2.1 Global distribution of snow

Snow cover extent (SCE) is defined as the amount of land covered by snow at any given time (Mudryk et al., 2024). The average annual SCE across Northern Hemisphere (NH) land areas was 24.3 million km^2 in 2023 (Robinson, 2023). Snow cover occurs mainly in the winter season and mostly prevails in areas of higher latitudes and altitudes.

Snow cover distribution can be altered by snow sublimation, wind redistribution, specific topography, forest interception or water movement through the snowpack (ČHMÚ, 2011). Biomes that typically experience extended snow cover include boreal forests and tundra. Those cover the Northern Hemisphere above 60°N latitude (Mudryk et al., 2024). In mountainous regions of high altitudes, the distribution of snow cover is highly variable due to the diverse terrain (Björk and Molau, 2007; Schinner, 1983). In these areas, freezing temperatures prevail, and the vegetation season is commonly shorter than in low latitude and altitude areas. Due to the low temperatures, soil processes are slowed down, leading to the accumulation of organic matter. Consequently, soils of these typically anoxic ecosystems such as wetlands and polar regions, tend to sequester the most carbon of all principal biomes (Šimek et al., 2019).

2.1.1 Trends in snow cover and depth

Snow cover extent (SCE) across Northern Hemisphere is continually calculated by the Global Snow Lab from maps created by the US National Ice Center (Estilow et al., 2015). A satellite SCE climate data record was progressively developed to improve meteorological and climate forecasting (Robinson et al., 2012). Snow-covered areas of the Northern Hemisphere declined by 3,2 % in 2023 compared to the 1966 – 2023 period mean (Robinson, 2023).

It is evident that the SCE long-term trends differ depending on the season (Fig. 1). For the years 1967 – 2024, fall SCE increased. However, in the last decade, the minimum SCE was in 2024.The winter SCE remains stable. In contrast, spring experiences a decrease in SCE which indicates earlier snowmelt and beginning of growing season. The earlier snowmelt is the most determinative factor for the snowpack impact on ecosystems since the length of the following vegetation season is altered.

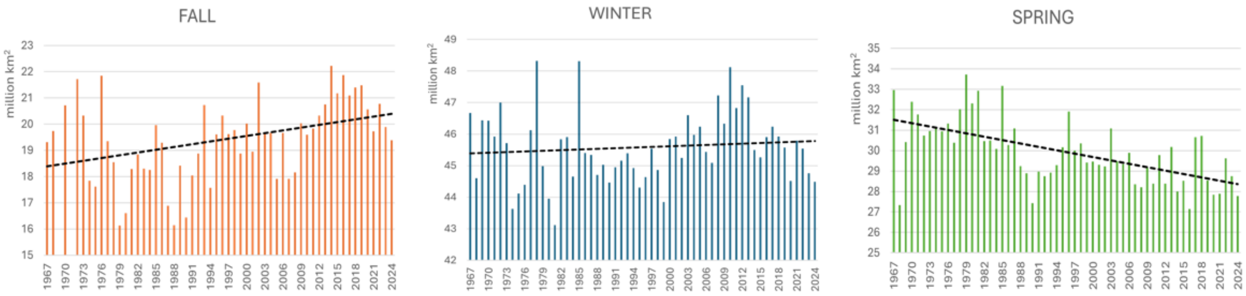

Figure 1: Snow cover extent (SCE) for the Northern Hemisphere (NH) for different seasons in years 1967–2024. Annual SCE over NH mean 25,1 million km^2. Monthly SCE values have been averaged for each year for fall (September, October, November; data from 1969 and 1971 missing), winter (December, January, February), and spring (March, April, May). Source: Graphs are of own making based on Rutgers University GSL (Robinson et al., 2012).

In addition to the snow cover extent, snow depth is also important. Estimating global changes in future snow depth is complicated due to regional, especially spatial and temporal, differences. Long-term decline in snow depth is expected due to global warming (Mudryk et al., 2024). Pongracz et al. (2024) assessed a decrease in annual median of snow depth by 3-5 cm (20%) by 2100 in arctic-boreal regions combining all future scenarios (Fig. 2). However, under the same scenarios, snow depth is projected to increase in the coldest regions such as Siberia and Northern Canada. The decline holds true within Europe, western Russia, southern Alaska, and southern Canada.

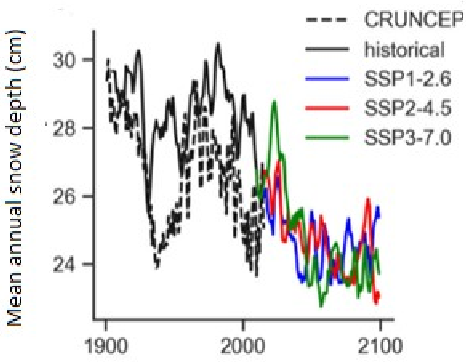

Figure 2: Mean annual snow depth projection combining climatic models. Black dashed line represents snow depth simulated by the applied model, black full line represents real historical data, three coloured lines are each for one climatic model. Source: Pongracz et al. 2024

Trends in the Czech Republic

The Czech Republic has a decent network of snow measuring stations, both manual and automatic, providing data of snow water equivalent (SWE) and snow depth (SD). Linear trends of SD show continual decrease over the years 1961–2022. Snow cover duration declined by 16 days over the years 1991–2020 compared to 1961–1990 (Šustková et al., 2023). The network is gradually densified to provide the most accurate water storage values, which are calculated weekly from the beginning of November until the end of April (Šustková et al., 2023).

Winter temperatures between 1961 and 2019 show significant deviations, particularly in the first decade 1961–1970, which was considerably colder, and in the last decade 2011–2019, when temperatures rose by 1.4 °C compared to the years 1961–1990. The largest increase is expected during winter seasons. According to the more severe scenario, an increase by 3.4°C to 5°C is expected in individual seasons by the end of the 21st century compared to the period 1981–2010 (Marek et al., 2022). Reduction in snow cover will be most likely influenced by a decline in days in which the temperatures do not rise above freezing temperatures (Marek et al., 2022) and decreased snow/precipitation ratio, despite any observed trend in annual precipitation (Šustková et al., 2023). The snow is projected to decrease until the end of the 21st century at all elevations above 600 m a.s.l. Furthermore, snow-covered season will be shortened by 40-60 days due to earlier spring melt in reference to 1980–2005 period (Jeníček et al., 2021).

2.2 Climate change impact on snowpack

Climate warming, primarily induced by increased greenhouse gas presence in the atmosphere, greatly impacts snow cover. Along with the precipitation, air temperature is the main factor changing snow cover trends. Nevertheless, changes in affected areas vary depending on the location of different topographies, primarily latitudes, altitudes, and biogeography. The response of albedo effect to snow cover changes in turn causes positive feedback to climate change impact.

2.2.1 Air temperature

Temperatures increased by 1.1 °C in 2015 compared to 1950, and this increase is predicted to exceed 2 °C by the end of the 21st century (IPCC, 2023). The period 2014–2023 was the warmest decade worldwide, and 2024 was the warmest year (NOAA, 2025). Temperatures are projected to rise globally. Higher temperatures accelerate snowmelt and decrease snow-covered areas, which has a direct impact on the albedo effect. Furthermore, air temperature affects whether precipitation falls as rain or snow.

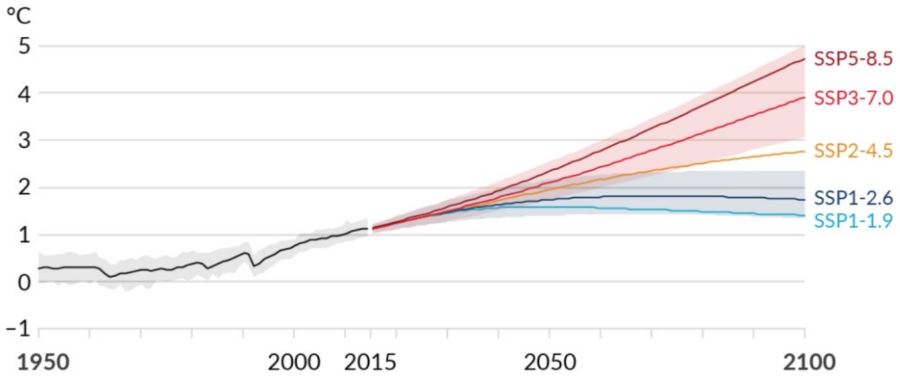

Figure 3: Global temperature change based on 20-year averaging periods from 1850 – 1900. Coloured lines each for different climate model. Source: IPCC, (2023)

2.2.2 Precipitation

Despite the projected rise of global temperatures, precipitation is inconsistent in long-term trends (Sanders-DeMott et al., 2019) but extreme and heavy precipitation is predicted (IPCC, 2023). Thus, air temperatures shape the precipitation aspect of climate impact on snow. With more frequent temperatures above the freezing point, rainfall will prevail and affect the snowpack depth and snow cover extent and duration by acceleration of snowmelt (IPCC, 2023). However, some sites could experience more snowfall when warming but remaining below the freezing point (Pongracz et al., 2024). Warming increases evaporation from snow; thus, the precipitation is more frequent (Barnett et al., 2005). However, in lower altitudes, the snowfall turns into rainfall sooner, and possible warm air advection events occur, which causes a decrease in snow cover. In higher altitudes, particularly in the mountains, the temperatures stay below or at the freezing point even under warming, accelerating the snowfall and preserving the mountainous snow cover longer (Barnett et al., 2005; Jenicek et al., 2021).

2.2.3 Albedo effect

Contrary to the land surface, snow greatly induces the albedo effect. Snow can reflect up to 90 % of solar energy, while the potential of soil reflection may decline to 20 %. The albedo effect is the most pronounced at high latitudes and altitudes, where the duration of snow cover is longer (IPCC, 2023). Snowpack reduction increases land surface solar radiation absorption due to decreasing albedo (Edwards et al., 2007). Higher annual absorption increases the ground heating. That can lead to a positive feedback loop (Fig. 4).

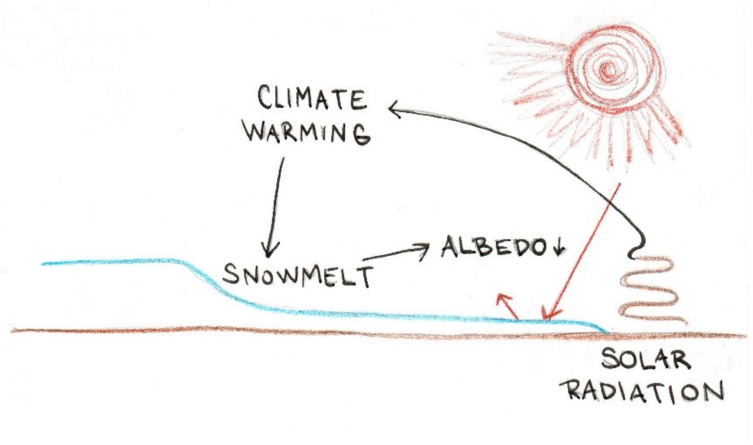

Figure 4: Positive feedback loop of climate warming and albedo. Warming causes snowmelt, which decrease albedo. Thus, more solar radiation is captured by soil and heated ground increases air temperatures. Climate warming is induced by decreased albedo.

2.3 The impact of snowpack on soil

Snowpack has a direct effect on soil temperature and water content which further influence vegetation, microbial communities and key biogeochemical processes (Edwards et al., 2007; Maurer and Bowling, 2014) as decomposition and soil respiration (Fig. 5). The essential snow properties which is the snow extent, depth and duration are given by physical properties of snow such as density, structure and thermal conductivity (Saccone et al., 2013). Those are impacted by meteorological and climate conditions as mentioned in 2.2.

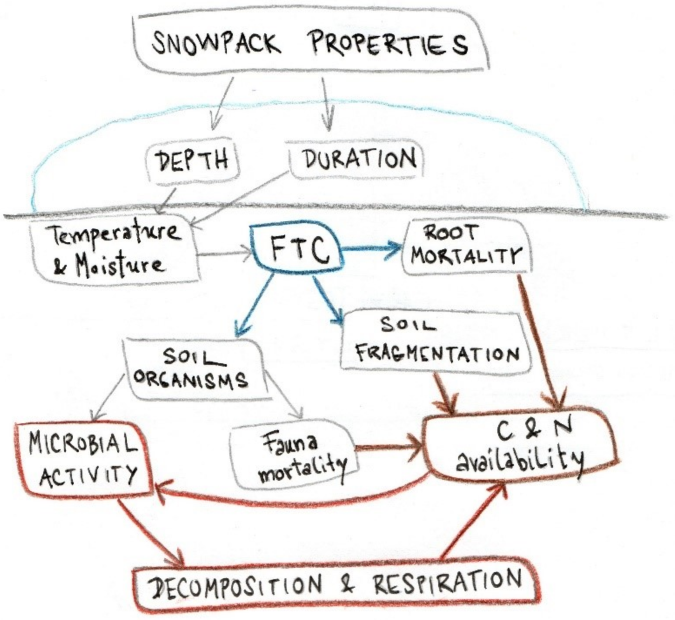

Figure 5: The influence of snowpack on soil properties leading to changes in decomposition and soil respiration. FTC: freeze-thaw cycle, C: carbon, N: nitrogen

2.3.1 Soil temperature

Snowpack is an insulation layer separating the cold air from the soil due to its low thermal conductivity (Maurer and Bowling, 2014). Just as cold air does not get through the snowpack, heat accumulated in the ground does not escape into the air. Thus, snow depth and onset of first snow have an essential effect on soil temperature (Brooks et al., 1998; Maurer and Bowling, 2014). If a sufficient snow layer covers the soil before the air cools down, warmer and stable soil temperatures are maintained (Maurer and Bowling, 2014; Xu et al., 2023). Brooks et al., (1998) suggest that the minimal sufficient layer to moderate the soil temperature is at least 30 cm. (Bartlett et al., 2004) claim that a snow depth of 1 m is needed to protect soil from air temperature fluctuations over the whole winter period.

The trend of earlier snowpack onset during fall, as described in chapter 2.1.1, could positively affect maintaining warmer soil temperatures over the winter period. On the other hand, early snowpacks could be prone to melt due to possibly warmer fall. That depends on the different biomes and their monthly temperatures. Furthermore, the intensity of the first snowfall depends on precipitation fluctuations. Thus, the overall future impact of snow on soil temperatures seems complex to predict.

2.3.2 Soil moisture

Snow cover provides moisture, and soils under deeper snow layers remain wet during winter. Soil water content is affected mainly by pre-snowpack conditions such as evapotranspiration from soil and its moisture(Maurer and Bowling, 2014). The water supply of continuous snowpack is measured as Snow Water Equivalent (SWE), which is defined as the amount of water contained in the snow layer after melting. This water storage is ordinarily kept until early spring and is important for the vegetation period. However, snowmelt can be induced during winter and supply soil with additional liquid water, which enhances soil activity (Brooks et al., 1998; Maurer and Bowling, 2014).

2.3.3 Freeze-thaw cycle

Warmer temperatures often lead to insufficient snow layer, which causes more frequent freeze-thaw cycles. The freeze-thaw cycle (FTC) describes the transition of water between liquid and ice occurring mainly during winter on exposed soils due to fluctuating temperatures (Henry, 2008). It occurs more frequently due to reduced snowpack. The frost phase severely affects soil physical properties by its mechanical disruption, litter fragmentation, or root damage. Thawing may increase the flush and leaching of elements and nutrients such as carbon and nitrogen and enhance microbial activity (Hobbie and Chapin,1996).

On a global scale, the area of seasonally frozen soil will be reduced, and the depth and duration of penetration will decrease (IPCC, 2023) mainly due to earlier onset of spring snowmelt, which may alter the effects on soil in an annual context (Campbell et al., 2010). Furthermore, early spring snowmelt or snow reduction may result in warmer soils than would remain under snow cover (Matzner and Borken, 2008), which could also be important considering the soil processes.

Soil frost

Sanders-DeMott et al. (2019) reviewed several snow removal experiments across different landscapes, elevations and soil types, confirming consistent results. Snow removal increases the depth and duration of soil frost. As mentioned in the freeze-thaw cycle chapter, soil structure can be mechanically damaged by frost. Steinweg et al., (2008) hypothesized that soil physical disruption may induce macroaggregates turnover. Macroaggregate turnover refers to formation or breakdown of soil aggregates, which is typically done by soil organisms. Aggregates determine soil structure, health, aeration or water retention potential (Šimek et al., 2019). Bigger particles (250–2000μm) disrupted by ice expansion would disintegrate into microaggregates which would increase the organic matter release and enhanced its bioavailability (and potential for humification) (Steinweg et al., 2008). However, they measured higher concentrations of organic matter even without changes in aggregates which indicates organic matter release due to litter fragmentation, root damage or mortality.

Litter fragmentation provides organic matter input into soil. Soil fauna is responsible for disrupting litter during vegetation season (Šimek et al., 2019). During winter, however, this impact is replaced by an ice expansion through the soil profile, similar to the aggregation (Steinweg et al., 2008). Released fresh matter allows microbial colonization and further activity while freezing (Gavazov, 2010).

Root mortality affects especially smaller roots of low order (fine roots) which are essential to water and nutrient uptake. Damaged fine roots hinder the nutrient uptake which enhances nitrogen availability for microorganisms and thus alters the balance of mineralized and immobilized nitrate (Tierney et al., 2001).

Consequently, root mortality effect on further soil activity may vary. Damaged fine roots are more susceptible to decomposition (Tierney et al., 2001), which would intensify the decomposition rate. In contrast, increased nitrogen availability possibly results in its deficit which may slow down decomposition. Root mortality has also a possible effect on autotrophic respiration. Severe damage to root structure may decline the respiration efficiency.

2.3.4 Litter and its properties

Litter as upper layer of soil organic matter is composed of plant residues of vegetation, typically controlled by climate and nutrient availability at the site. Litter is also colonized by some organisms and their excrements (Šimek et al., 2019). Litter susceptibility to decomposition is determined by its quality given by the content of C, N, C/N ratio, cellulose, lignin and the content of other elements such as P and K (Wu et al., 2010). The litter C/N ratio is its proportion of carbon and nitrogen. The higher the ratio is, the more carbon relative to nitrogen is included which slows down the decomposition.

Vegetation of snow-covered regions is mostly coniferous and deciduous trees of temperate forests, moss, lichens, and evergreen small shrubs. The litter of evergreen shrubs, dominant in ecosystems experiencing very short vegetation season, is the most recalcitrant (Gavazov, 2010). Overall, sites with extremely short growing seasons suffer from insufficient amounts of decomposable material (Schinner, 1983).

2.3.5 Soil organisms

Soil organisms can be classified into main groups by their size. Microorganisms, including bacteria, archaea, and fungi, are the most abundant group of soil biota, representing 80 % of soil biomass (Šimek et al., 2019). Those are believed to contribute most to the soil processes, such as decomposition and respiration. In literature, this group is often described as microbial biomass. The rest (mesofauna, macrofauna), including arthropods, contribute to decomposition primarily through litter fragmentation during the initial stage (Šimek et al., 2019).

The insulating effect of snow significantly influences soil organisms by protecting them from freezing temperatures. Snow depth plays a crucial role in determining the vertical distribution of organisms, as they may migrate deeper into the soil in search of warmer conditions while the upper layer is frozen. Consequently, community structures may shift since the less cold-adaptive, primarily macrofauna, are absent in the upper horizon (Bokhorst et al., 2013). Moreover, frost-induced mortality decreases the abundance of arthropods in subsequent summer (Templer et al., 2011). Additionally, organic matter may be redistributed throughout a wider soil profile due to migrating fauna. The resulting changes in the abundance of soil organisms and organic matter can potentially affect the decomposition rate.

A deeper snowpack enhances microbial activity (Yin et al., 2024) and may even induce its annual maximum, especially among bacteria and fungi (Isobe et al., 2018). In contrast, freeze-thaw cycles caused by snowpack reduction attenuate soil organisms' activity (Bokhorst et al., 2013; Sorensen et al., 2018). However, the freezing temperatures do not cease the activity equally among all groups. It is assumed that fungi are more susceptible to frost since their abundance declines more than that of bacteria (Schmitt et al., 2008). Nevertheless, of all soil organisms, fungi appear to be the most plentiful in biomass (Šimek et al., 2019), which could explain the decline among fungal communities. The alteration of the fungi/bacteria ratio could have consequences for the decay of various compounds of organic matter. While bacteria aim for the easily accessible carbon, fungi decompose very effectively more complex compounds. In a simplified way, bacteria induce cellulose mass loss while fungi reduce lignin content in the soil (Yin et al., 2024).

Finally, the shift of communities among the seasons is probable. Cold-adapted microbes start to decline with higher temperatures, and soil is enriched with their necromass, which provides food for summer communities emerging by the early spring (Gavazov, 2010), most likely contributing to the early spring peak of microbial activity (Fig. 6).

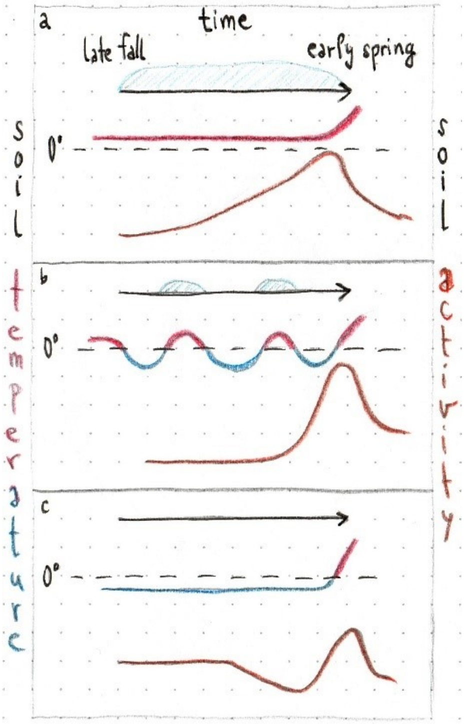

Figure 6: The hypothetical effect of different snow cover conditions on soil temperature and soil biota activity from late fall to early spring. The arrow illustrates the snowpack duration over winter. The red curve represents soil temperature above zero, while the blue curve shows soil temperature below zero. The brown curve indicates the likely soil biota activity under the given temperature conditions.
a) consistent snowpack – gradual increase in soil biota activity due to consistently above-zero temperatures. b) varying snowpack - The accumulation of organic matter due to fluctuating temperatures significantly enhances soil biota activity during the spring thaw. c) absent snowpack - Extended periods of soil freezing reduce soil biota activity and diminish the intensity of the spring activity peak. Based on Liptzin et al. 2009.

Soil freezing has two opposing effects. While the activity of soil organisms is suppressed, the soil is simultaneously enriched with necromass and newly available nutrients. During the subsequent thawing events as part of the freeze–thaw cycles (FTCs), frost-resistant organisms may take advantage of these newly released compounds and increase their activity again. Therefore, it remains unclear whether biogeochemical processes in the soil throughout the winter are overall suppressed or enhanced by snow reduction compared to deeper snowpack conditions.

2.4 Impact of snow reduction on soil carbon cycling

Soil is a significant carbon sink on Earth (Šimek et al., 2019). Changes in terrestrial carbon storage are fundamental to understanding the carbon cycle and future climate change projections (Bond-Lamberty et al., 2024). Biological processes in winter contribute significantly to annual fluxes of carbon and should not be underestimated (Gavazov, 2010; Pongracz et al., 2024).

Carbon inputs to the soil include litter and fine root biomass. Decomposition processes transform plant biomass inputs into residues in different phases of decay, which, together with microbial necromass residues, form soil organic matter. As decomposition continues, the organic carbon is converted into terminal inorganic product carbon dioxide, CO_2 (Šimek et al., 2019). The carbon exchange between the terrestrial sink and the atmosphere is mainly through the respiration of CO_2 and partially other carbon compounds (Fig. 7). For instance, methane (CH_4) should not be neglected when studying soil C losses. It is produced by anaerobic microorganisms under anoxic conditions in soil or sediments (Šimek et al., 2019). However, following review is focusing only on aerobic respiration which produces only CO_2 as its final product.

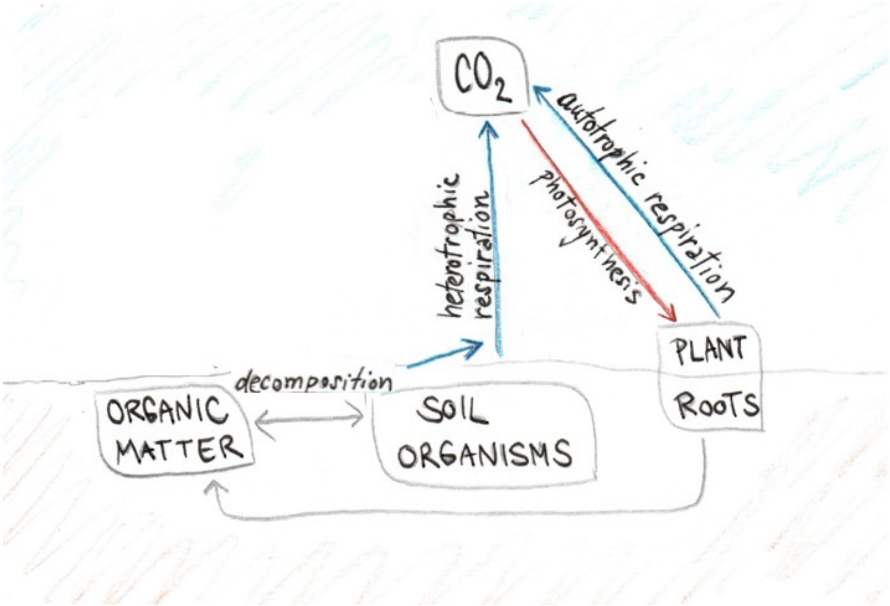

Figure 7: Soil processes of decomposition and soil respiration contributing to exchange of CO_2 between soil and atmosphere. Recreated from Šimek et al., 2019

2.4.1 Impact of snowpack reduction on litter decomposition

Litter decomposition is the process of breaking down organic matter into diffusible gases. The key contributors are soil organisms responsible for fragmentation and further decay facilitated by metabolic processes, excreted enzymes, and diffusion of single chemical structures into cells (Šimek et al., 2019). The decomposition rate is also enhanced by the leaching of soluble material enabled by soil moisture (Gavazov, 2010).

In general, decomposition is slowed down by lower soil temperatures, and its rate during the winter period is significantly lower than during the growing season. However, the decomposition rate also varies throughout the winter, as the soil is susceptible to temperature fluctuations caused by varying snow depth (Bokhorst et al., 2013; Kreyling et al., 2013).

The decomposition rate during winter may be enhanced by the warmer soil temperatures kept under the snow cover (Gavazov, 2010; Yin et al., 2024). According to meta-analysis conducted by Yin et al. (2024), decomposition increased by 17% in average compared to sites with reduced snow depths. Therefore, it is reasonable to assume that the decomposition rate would decrease with snowpack reduction due to soil temperatures declining. However, the assumption of a lower decomposition rate under lower snow may not be straightforward due to factors such as climate, frequency of freeze-thaw cycles, litter quality and soil biota activity.

Climate effect

Spatial variability is often a significant factor influencing winter decomposition rate. This is particularly common in high mountains where snow cover persists for longer periods. In certain depressions known as snowbeds, the duration of the snowpack can be further extended which increases the decomposition rate (Schinner, 1983; Björk and Molau, 2007). In contrast, exposed areas of the mountains experience significant wind, creating extreme sites uncovered with snow, dried out by wind, and having low temperatures. In such sites, decomposition rates remain low due to the poor organic substrate and stressed microbial activity (Schinner, 1983).

Furthermore, interesting shifts in decomposition rates could be visible in mountainous areas due to the impact of climate on the plant community composition. Since the snow cover will be decreasing in higher altitudes, alpine vegetation colonizes the upper zones of mountains, which can alter the microecosystems and increase the decomposition rate (Gavazov, 2010).

Freeze-thaw cycle effect

Some studies conducted in boreal forests have proven that soils under reduced snow cover experience slower decomposition. The difference compared to soils with consistent snow cover throughout the winter can be up to double (Kreyling et al., 2013). The difference in decomposition values increases with the difference in snow depths (Christenson et al., 2010). Therefore, the decomposition rate may depend on soil frost duration and snow depth (Kurka et al., 2000). Soil frost duration is substantially affected by frequent freeze-thaw cycles.

Given the increasing impact of FTC on organic matter release and microbial activity, the assumption that reduced snowpack leads to lower decomposition could be disproved. Wu et al. (2010) demonstrated that winter decomposition influenced by freeze-thaw cycles (FTC) significantly contributes to the annual decomposition rate. However, the primary factors driving this winter decomposition remain unclear. Decomposer activity may be more pronounced at the onset of winter, as freshly fallen litter releases large amounts of dissolved organic carbon (DOC), making it highly susceptible to rapid breakdown (Bokhorst et al., 2013; Wu et al., 2010). The quality of decomposed organic matter might play a more crucial role than the depth of the snowpack in determining the winter decomposition rate (Baptist et al., 2010). Elevated concentrations of DOC can also be observed during the spring thaw. Throughout the winter, FTC continuously releases organic matter that increases microbial activity as spring thaws (Schinner, 1983). Thus, while detecting high decomposition rates immediately after winter is noteworthy, it does not necessarily confirm that these increased values occurred during winter.

Litter quality dependence

Additionally, the predisposition of bacteria and fungi microbes to variable chemical structures emphasizes the key role of vegetation and litter composition. Decomposition rates differ according to litter type and its various chemical composition (Cornwell et al., 2008; Schinner, 1983; Walker et al., 1999). Typically, during growing season, lowest decomposition rate is associated with highest litter C:N ratio (lowest N concentrations). Surprisingly, some findings show the opposite. Litters lower in N had 5-15 times higher winter litter decomposition rates (Bokhorst et al., 2013). Thus, greater nutrient concentrations do not always mean faster decomposition. Therefore, decomposition processes during winter cannot be directly inferred from those occurring in the growing season.

Summary - processes influenced by decomposition under reduced snowpack

Decomposition under reduced snowpack is influenced by spatial conditions, freeze-thaw cycles, microbial communities, and litter type. Due to lower winter decomposition rates, litter layer remains largely intact over the winter, which can lead to heightened microbial activity in early spring. It is also implied that lower decomposition under reduced snow cover may promote carbon sequestration (Ni et al., 2014). Understanding the effect of decreased winter decomposition is important to evaluate growing season decomposition and annual carbon cycle changes (Kreyling et al., 2013). Finally, lower winter decomposition can initially suggest reduced soil-to-atmosphere CO_2 flux during winter.

2.4.2 Impact of snowpack reduction on soil respiration

The main final product of decomposition in aerobic conditions is carbon dioxide (Šimek et al., 2019). It diffuses from soil to the atmosphere by its respiration. The importance of total soil respiration to the carbon cycle is emphasized by its contribution to the earth's carbon flux. It is a large carbon release source, exceeding anthropogenic carbon emissions (Bond-Lamberty and Thomson, 2010). Total soil respiration comprises heterotrophic and autotrophic respiration. The heterotrophic covers microorganisms and fauna. Through heterotrophic respiration, soil microbial activity is measured. Plant roots perform autotrophic respiration.

Climate and biome effects

Soil respiration is expected to decrease due to reduced snowpack. However, results of decreased soil respiration differ among various biotopes (Kosolapova and Altshuler, 2024). Whereas consistent results of reduction in CO_2 flux under lower snowpack are for a subalpine spruce forest (Monson et al., 2006; Yang et al., 2019) and temperate grassland (Kurganova et al., 2017), such effect is not visible in subalpine grassland (Gavazov et al., 2017) or boreal forest/tundra (Groffman et al., 2006), neither confirmed by a meta-analysis of snow manipulation on carbon dynamics (Li et al., 2016). Notably, forests have generally lower respiration rates under freezing temperatures compared to open spaces (Monson et al., 2006). Thus, climatic conditions of various ecosystems affect the results of soil respiration.

Generally, the snow cover effect on CO_2 efflux in the mid-elevation temperate forests is marginal compared to high-latitude and altitude areas with much longer winter seasons and snowpack duration (Schindlbacher et al., 2014). The elevation factor in respiration measurements significantly affects the amount of respired carbon, with CO_2 flux generally being higher in low-elevation areas (Groffman et al., 2009). In mid-elevation areas, the effect of climate change on CO_2 efflux is crucial during the growing season (Schindlbacher et al., 2014).

Additionally, organic and mineral soils react to snow removal differently. Based on laboratory results, snow reduction's effect on microbial respiration response is more sensitive in organic soils (Yang et al., 2021). According to Gavazov et al. (2017), soil respiration rate in deeper mineral horizons remains stable and lower than in the organic horizon. Deeper layers remain unfrozen and do not contribute to winter fluxes. Therefore, it is important not to overlook the soil profile while estimating winter CO_2 flux.

Freeze-thaw cycle effect

Under freeze-thaw cycles, the respiration rate fluctuates according to changing temperature and moisture conditions. Thus, soil respiration is unevenly distributed throughout the winter (Liptzin et al., 2009). For instance, in thawing periods, significant flux occurred, representing up to 64% of the total heterotrophic respiration over 5 months of measurements (Khoroshaev et al., 2023). Values also differ depending on the length and frequency of measurements. It is suggested that longer measurements could decrease the measured fluxes. Due to gas accumulation in the gas analysing chamber, the CO_2 flux may be inhibited (Groffman et al., 2006). It is also suggested that accurate measurement requires multiple values collected periodically throughout the winter. Otherwise, the estimation of overall CO_2 flux is highly imprecise (Khoroshaev et al., 2023; Bond-Lamberty et al., 2024).

Considering the soil frost effect, the respiration rate also depends on the duration of FTC. Prolonged frost periods decrease microbial biomass and soil respiration (Goldberg et al., 2008). This assumption is underlined by (Aanderud et al., 2013), who estimated soil respiration sensitivity to temperature and moisture. In the low snow plots, CO_2 flux was the most sensitive to moisture fluctuations. Soils exposed to ambient air temperatures undergo the transition between liquid and ice most frequently. This occurs within a temperature range of approximately -1°C and 1°C (Aanderud et al., 2013). Under these conditions, soil respiration is likely determined by moisture availability. Outside this temperature range, soil respiration is likely to be more sensitive to soil temperature.

Impact among different seasons

Reduced snow cover can significantly alter the annual carbon balance by modifying soil respiration dynamics across seasons. The accumulation of available organic matter due to numerous FTC and impeded decomposition leads to a larger CO_2 burst during the early spring snowmelt. CO_2 flux during the early spring thawing contributes the most to the annual flux. The contribution is significantly larger from frost-influenced soils (Kurganova et al., 2017). Therefore, snowpack reduction could increase annual soil respiration considerably despite the prevailing claims of current literature emphasising the decrease in respiration due to reduced snowpack.

Reinmann and Templer (2018) estimated freezing-induced annual total soil respiration increase by 27,6 %. They attribute the rise of total soil respiration to enhanced root mortality. Root mortality increases the decomposition of root necromass, the positive priming effect of soil organic matter decomposition, and the compensatory growth of roots in the following growing season (Gavazov et al., 2017; Reinmann and Templer, 2018).

However, further respiration rate in the subsequent summer season will likely decline when snow is reduced (Haei et al., 2013; Muhr et al., 2009). According to Haei et al. 2013), summer heterotrophic respiration has a higher rate due to increased soil frost. Conversely, autotrophic respiration in summer is decreased due to damaged roots, which overrides the effect of higher heterotrophic respiration. Consequently, the overall annual respiration rate may decline. Consequences of reduced respiration during winter are most likely important for the subsequent summer. Nonetheless, these consequences are difficult to describe. Reduced carbon output during summer may result from arid conditions, regardless of the effects of soil and root freezing (Muhr et al., 2009).

2.4.3 Possible carbon sequestration under reduced snowpack

Lower soil respiration due to reduced snowpack may lead to greater carbon retention in the soil. Enhanced C sequestration due to decreased CO_2 flux should not be dismissed (Gavazov et al., 2017; Monson et al., 2006). However, lower soil respiration is caused by decreased decomposition. With a low decomposition rate, soil carbon is inefficiently transformed into organic matter and incorporated into microbial biomass, leading to its immobilisation. Carbon sequestration is more likely to increase when decomposition is sufficient, but respiration remains low. However, the balance of the soil processes is complex and highly unpredictable in winter conditions. Wang et al. (2022) observed a decrease in immobilisation within three years of reduced snowpack, followed by an increase after four years. This suggests that the short-term effects of snowpack reduction may not explain long-term soil carbon dynamics. It remains unclear whether frost conditions will enhance the priming effect–due to the release of soil organic matter–or whether microorganisms will instead immobilize carbon, thus promoting its sequestration (Gavazov et al., 2017). Therefore, further long-term research is required to understand these mechanisms better.

2.5 Methods for studying the effects of snowpack changes on decomposition and soil respiration

Several methods are used to study the effect of different snow conditions on decomposition and soil respiration. This can be done by field observation of sites with different snow depth, but often also other factors (such as parent material, vegetation, climate) can vary between these sites. The best approach is to experimentally alter the depth of snow cover in a manipulation experiment in which the other confounding factors can be controlled.

2.5.1 Snow manipulation method

Snow cover can be altered by different methods. Commonly used is a “snow removal” method where the snow cover is manually redistributed with shovels. At one plot, snow is being constantly removed to keep the soil exposed. Removed snow may be added to a second plot to deepen the snow layer, and a third plot serves as a control plot (Aanderud et al., 2013; Muhr et al., 2009).

Another possibility to alter the snowpack is the “snowfence” method (Fig. 8). A snow fence is placed to disrupt the wind stream, which leads to snowpack accumulation just behind the fence. The layer subsequently recedes, which redistributes the snow to variable depths.

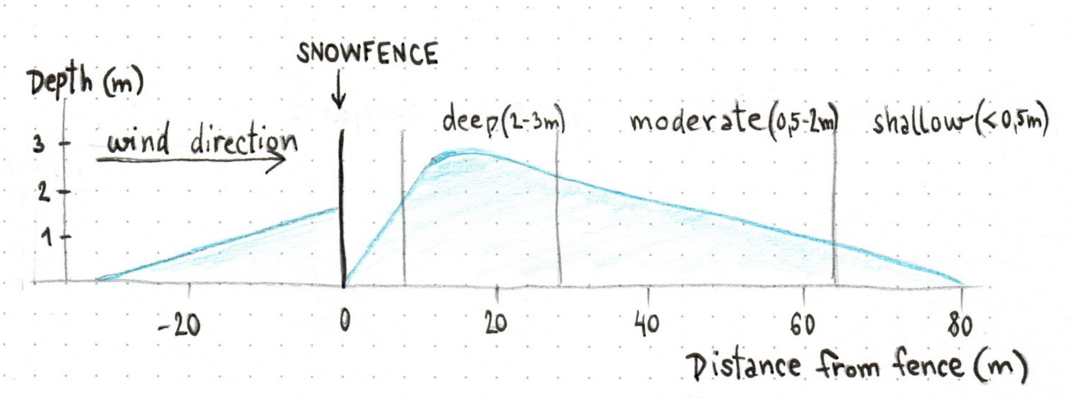

Figure 8: Snowfence method design. Recreated from Walker et al., (1999).

Another possible way to alter the snowpack is by adding dust to the snow surface, as they did in (Conner et al., 2017). The dust reduces the albedo effect, and snow reflects less solar radiation, which induces the snowmelt and reduces snow. Additionally, the forest gap method uses the surrounding vegetation canopy to distinguish plots with deeper or shallower snow. More snow tends to accumulate in forest gaps, resulting in longer snowpack duration (Wang et al., 2022; Yin et al., 2024). Finally, snow shelters can be built to prevent the snow from reaching the ground (Li et al., 2016; Yang et al., 2021).

2.5.2 Soil respiration

Total soil respiration is measured as a CO_2 flux from soil to the atmosphere. Chamber methods, using both manual and automated chambers, are commonly used. First, collars are inserted into the soil, to which a gas-measuring chamber is attached. It is important to prevent any flux leakage from the chamber. Subsequently, the concentration of captured CO_2 is measured using an infrared gas analyser or gas chromatography. Then, soil respiration flux is quantified (Bond-Lamberty et al., 2024).

Soil respiration is often measured manually at various intervals, providing insight into CO_2 fluxes across heterogeneous terrain. However, temporal conditions, such as precipitation, influence soil respiration, which can cause significant fluctuations. Variability in fluxes during temporal changes can only be captured through continuous measurements, and results from single-time measurements may be highly misleading. Thus, automated long-term measurements have the potential to reveal the relationship between soil respiration and yearly climate conditions. However, standardized methods and protocols are being developed, and technical issues must be addressed (Bond-Lamberty et al., 2024).

2.5.3 Soil decomposition

The litterbag method is widely used to examine litter decomposition. A litterbag is a nylon-made meshed bag filled with decomposable material, usually the litter samples from the examined site or some standard material (e.g. cellulose or tea). The preferential mesh size is 1mm, which allows microbes and micro and meso-fauna to pass through but excludes macro-fauna colonization. It also enables the litter fragments loss. Litterbags are best laid on soil surface to assess results closest to the natural decomposition processes. Selected weight and sampling duration affect the results substantially (Xie, 2020). The decomposition rate is calculated based on the mass loss over a certain period.

3 Practical part

3.1 Aim and hypotheses

Aim of the practical part was to assess the impact of snowpack depth on soil temperature, decomposition rate and soil respiration in Králický Sněžník. Based on previous findings, there were three hypotheses:

Hypothesis 1: Soil temperature decreases due to snow reduction.

Hypothesis 2: Decomposition decreases with lower soil temperature due to snow reduction.

Hypothesis 3: Soil respiration decreases with lower decomposition rate due to snow reduction.

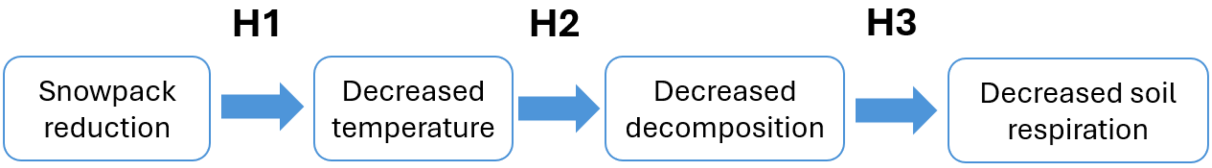

Figure 9: Diagram, where each H(1,2,3) represent one of the hypotheses. H1: Soil temperature decreases due to snow reduction. H2: Decomposition decreases with lower soil temperature due to snow reduction. H3: Soil respiration decreases with lower decomposition rate due to snow reduction.

3.2 Materials and methods

3.2.1 Study site

The study was carried out at the top of the mountain Králický Sněžník, 1423 m a.s.l., (50.2075000N, 16.8475581E), the highest peak of the mountain range Králický Sněžník situated on the border of northeastern Czechia and Poland. Mountain top is covered with subalpine vegetation with grass and heath.

Data were collected in two subsequent winter seasons, 2022/2023 and 2023/2024 and the experimental design was the same for both winters (Table 1).

Table 1: Individual phases of the experiment at Králický Sněžník. My participation marked with *.

	Site preparation	Winter sample collection	Spring sample collection	In-situ period (days)	Lab processing		Data analysis
					winter	spring	
Winter 2023	10th Nov (2022)	07 – 10 Feb *	29th Apr	170	*		*
Winter 2024	9th Nov (2023)	23 – 25 Jan	4th Apr	147		*	*

The field preparations were on the 10th of November in 2022 and on the 9th of November for the subsequent season in 2023. For the experiment, two types of plots, six in total, were set up (fig. 10). The first three plots were selected on the summit, with less snow due to stronger wind conditions. The second three plots were located on the hill's lee side where the snow accumulated. These two types of plots differing in their snow depth were marked as low snow (LS) and high snow (HS), respectively. The dominant vegetation in LS plots was heath (*Calluna vulgaris*), while the dominant vegetation in HS was blueberry (*Vaccinium myrtillus*).

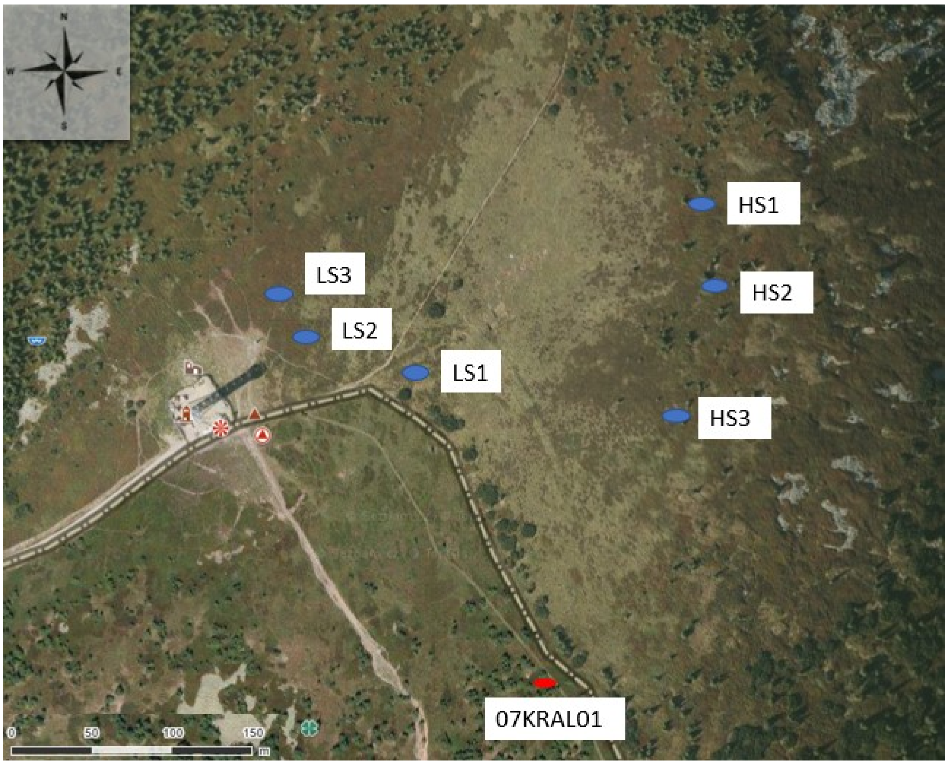

Figure 10: Map of LS and HS plots distribution on the study site and meteorological station of Králický Sněžník (07KRAL01).

Each of the six plots was divided into three subplots, determined as “Removal,” “Control,” and “Add” (R, C, A) according to their future treatment. The centre position of the plot was marked by a bamboo stick and entered into the GPS. Strings were then stretched from the bamboo stick to the corners of R and A subplots. In the case of the R subplot, the string continued along the diagonal (fig.11). Furthermore, in each R subplot were placed dataloggers (HOBO), for measuring temperature, and already prepared litterbags. Every litterbag was made of nylon mesh (mesh size 42 µm) and filled with filter paper 10x10 cm of equal weight 1 g (±0.001). Moreover, litterbags were marked with the snow thickness plot, subplot number, and treatment. They were fastened to the ground by nails. Three litterbags were added to all the "control" and "add" subplots. Six litterbags were placed on each "removal" subplot so that three samples could be collected in both winter and spring. For every plot, a scheme of subplots position and distribution of litterbags and data loggers was prepared, and a preferential approach in the winter season was also included (Fig. 11). In total, 72 litterbags were distributed for each season. The plots were left unimpaired until the first sample collection while covered by the snow.

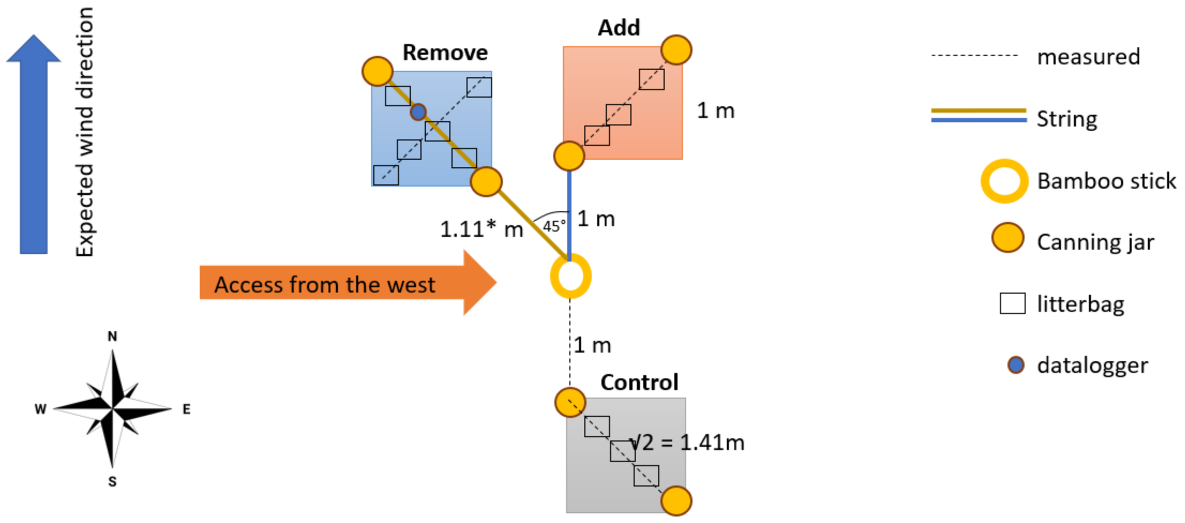

Figure 11: Distribution of the three subplots, (Remove, Add, Control) at the site with arrows of preferential wind and access direction to avoid compaction of snow on experimental plots in the winter.

3.2.2 Snowpack manipulation

The first samples collection took place from 7. – 10.02.2023 for the first season and 23. – 25.1.2024 for the second season. At each of the six plots, the subplots were distinguished by the snowpack manipulation method (described in 3.7) according to the prepared scheme (fig. 11). The plot was approached from the west to prevent the subplots from trample. Afterward, the snow was removed from the R subplot and added to the A subplot to increase the snow depth. The last subplot, treated as a C, was left unimpaired. After the snow removal, snowpack depth at the R subplot was measured using a probe. Thereafter, three litterbags were removed from the uncovered R subplots. Each winter, 18 litterbags were collected.

The spring samples collection was performed after the spring snow melt on 29.4.2023 and 4.4.2024. This time, all the remaining litterbags were collected, together with the temperature data loggers. In total, 54 litterbags were collected each spring.

During the sample collection in winter 2024, litterbags from the subplot LS3C were accidentally collected instead of LS3R. Therefore, six litterbags from LS3R collected in spring 2024 were considered as treatment C because they remained unimpaired (with the 2 sets of litterbags). This caused the dataset from spring 2024 to miss the treatment R completely, because there were no litterbags left at the LS3C plot until the spring.

3.2.3 Sample processing

After removing litterbags from the plots, each was dried until constant weight at 40°C overnight and paper was removed from the litter bag and weighed. The cellular mass loss is calculated as *initial weight - reduced weight = cellulose mass loss.* This was then divided by original weight of paper (1 g) and multiplied by 100 to obtain cellulose mass loss in %. In the advanced stage of decomposition, the smaller pieces firmly attached to the nylon-made litterbag had to be scraped off with the edge of tweezers or scissors.

Figure 12: Reduced filter paper removed from the LS and HS plots, respectively in spring 2024. Filter paper is firmly attached to the nylon material from the HS plot.

3.2.4 Soil respiration

The soil respiration was measured with the EGM-5 CO_2 gas analyser. It was measured in both winter seasons at every snow removal (R) subplot after the snowpack manipulation. In 2023, respiration was measured only once per subplot. In 2024, it was measured three times in each R subplot which were then averaged. In the analysis I work with the mean value of the three measures at each plot.

3.3 Data and analysis

I worked with four datasets - two from winter and two from spring. I took the mean values of the three litterbags collected from each subplot. For analysing winter data, I used a t-test to examine the effect of plot type (independent factor, two levels: high snow/ low snow) on measured snow depth, soil temperature, decomposition, and respiration. For analysing spring data, I used two-way ANOVA because I had two independent factors to examine: plot type (high snow/low snow) and snow treatment (three levels: removal/addition/control). To control the variance homogeneity, I used the Shapiro-Wilk normality test. Therefore, I used one non-parametric, the Scheirer-Ray-Hare test, for spring 2023.

From the temperatures measured every hour, four different temperature indicators were derived: *mean temperature*, *mean temperature over the last three days*, *ratio temperature higher than 0°*C, to estimate which part of the measured temperatures were above freezing point, and *time with temperature higher than 0°C*, which shows how many values were above freezing point. Finally, I made a correlation matrix testing which temperature indicators correlate the most with decomposition and soil respiration.

3.4 Results and discussion

3.4.1 Snow depth

It was confirmed (t-test, Table 2), that the low snow (LS) and high snow (HS) plots were selected correctly to gain diverse snow depths. On high snow (HS) plots was accumulated deeper snow layer than on low snow (LS) plots (Fig. 13). It emphasizes the spatial effect on snow distribution. On the mountain top, snow is reduced by strong wind, whereas it is accumulated on the lee side of the hill.

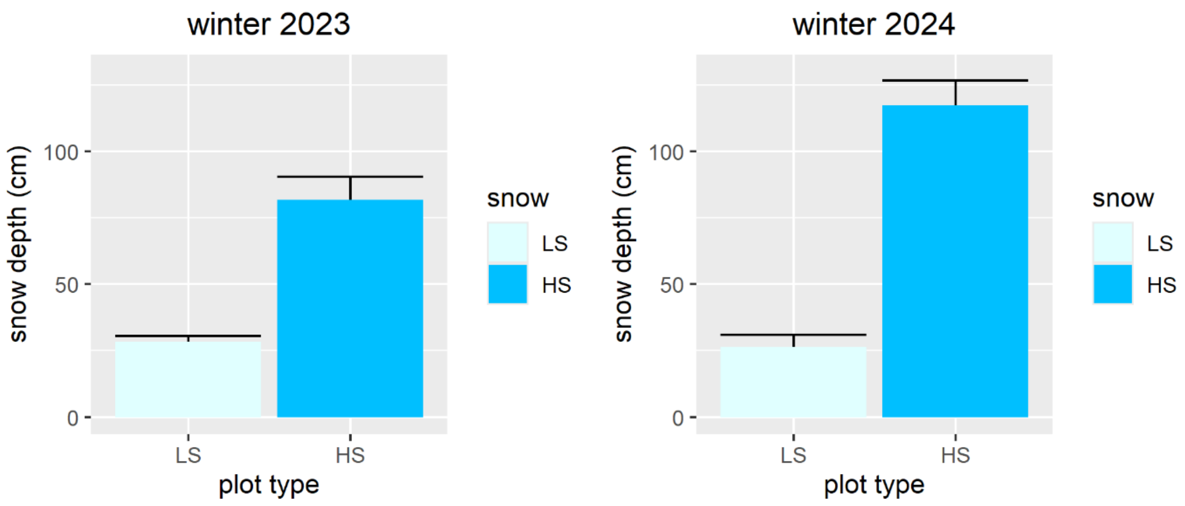

Figure 13: Snow depth (±SEM, n=3) at low snow (LS) and high snow (HS) plots in winters 2023 and 2024, respectively.

Table 2: T-test results for snow depth at LS/HS plots for winters 2023 and 2024. df: degrees of freedom, t: t-value, Significant results correspond with $p < 0.05$.

year	2023			2024		
	df	t	p-value	df	t	p-value
Plot type	4	-5.99	0.0039	4	-8.72	0.000952

When interpreting the results, the development of snow cover across individual years should not be neglected. The snowpack onset for the first season was in mid-November and there was a rapid melt from the end of the December reversed in mid-January by intensive snow increase. The following season experienced earlier onset and more consistently increasing snowpack than the previous winter.

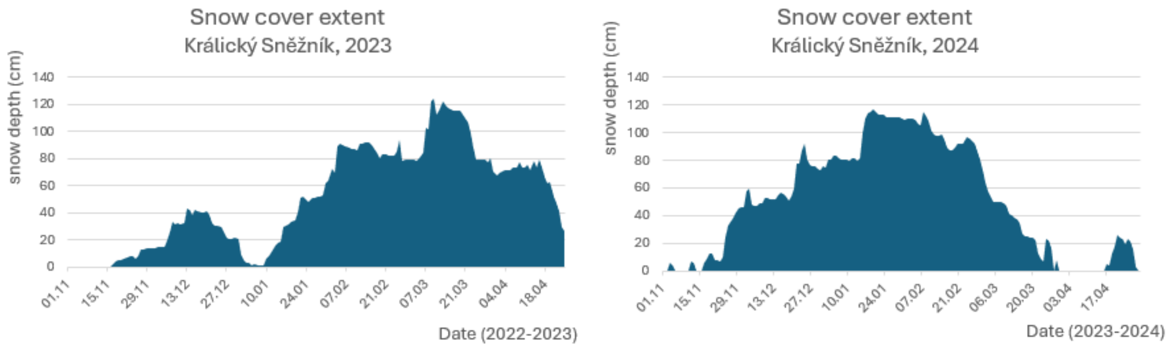

Figure 14: Snow cover extent (SCE) from November to April at the Králický Sněžník for the two winter seasons 2022/2023 and 2023/2024. Data from meteorological station 07KRAL01, 1402 m a.s.l. (provided by ČHMÚ Ostrava, Ing. Pavel Lipina, Martin Balík)

3.4.2 Soil temperature

There was a tendency for the soil temperature to have higher mean values on high snow plots than on low snow plots (fig.15). However, this was not statistically significant (t-test, Table 3). Soil temperature increases with consistent sufficient snow depth, which could be the case in winter 2024 (fig. 14). Despite the intact depths of snow between the plots, the insulation effect of snowpack could have been insufficient even among the HS plots. As Bartlett et al. (2004) claims, the snow depth should be above 1 m to prevent the temperature fluctuations, which is surpassed only in HS plot in winter 2024. During the winter 2023, average temperature was below the freezing point. The insignificance of the results may have been influenced by deficiency in measured data at LS plots since for each winter, measurement from one of the three plots is missing. Given the insignificant results, the first hypothesis of the experiment cannot be confirmed.

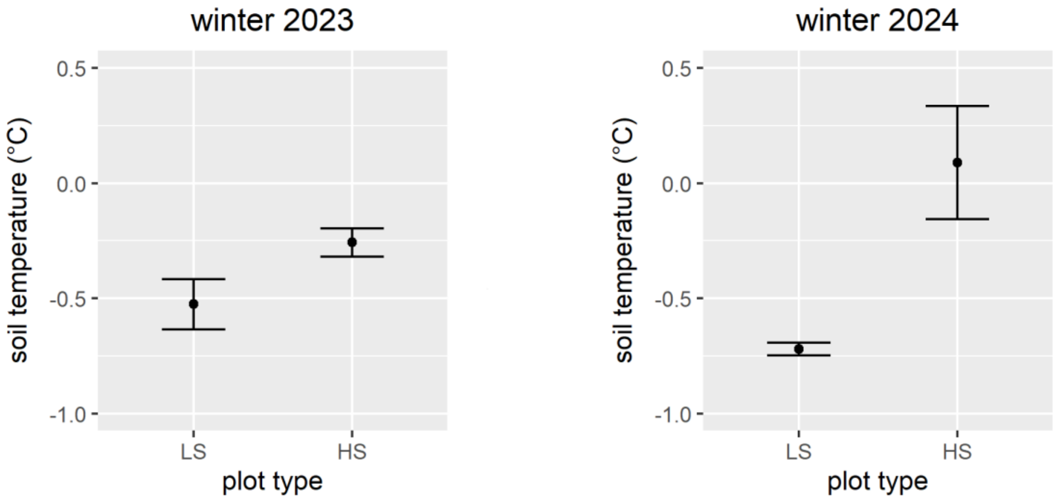

Figure 15: Soil temperature (±SEM, n=3) measured at low snow (LS) and high snow (HS) plots in the winter 2023 and 2024, respectively.

Table 3: T-test results for soil temperature at LS/HS plots. df: degrees of freedom, t: t-value, Significant results correspond with $p < 0.05$.

year	2023			2024		
	df	t	p-value	df	t	p-value
Plot type	3	-2.36	0.0997	3	-2.54	0.0846

At each plot were measured fluctuating temperatures from -1°C to 1°C. In this temperature range frequently occurs FTC as described in Aanderud et al., (2013). More frequent FTC can enhance soil biota activity. Thus, frequency of FTC is determinative for both decomposition and soil respiration.

3.4.3 Decomposition

There was a tendency for the decomposition as measured by cellulose mass loss to have higher mean values on high snow plots than on low snow plots (Fig.16). However, this was not statistically significant in both winters (t-test, Table 3). Nevertheless, this difference became significant for litterbags sampled in spring (two-way ANOVA, Fig.17, Table. 5). The second hypothesis may be therefore confirmed only based on spring results. These findings are consistent with the meta-analysis conducted by Yin et al. (2024). It can be explained by the different duration of soil processes acting on the samples, as the spring litterbags were placed on the soil for a longer period. The tendency for higher values of decomposition at the HS plots in 2024 than in 2023 for both winter and spring (Fig. 16,17) are probably given by the deeper and consistent snowpack during the second season (Fig. 14) but this could be a question for a follow-up analysis. This would potentially stress the importance of snowpack duration as discussed in Kreyling et al. (2012), Christenson et al. (2010) and Kurka et al. (2000).

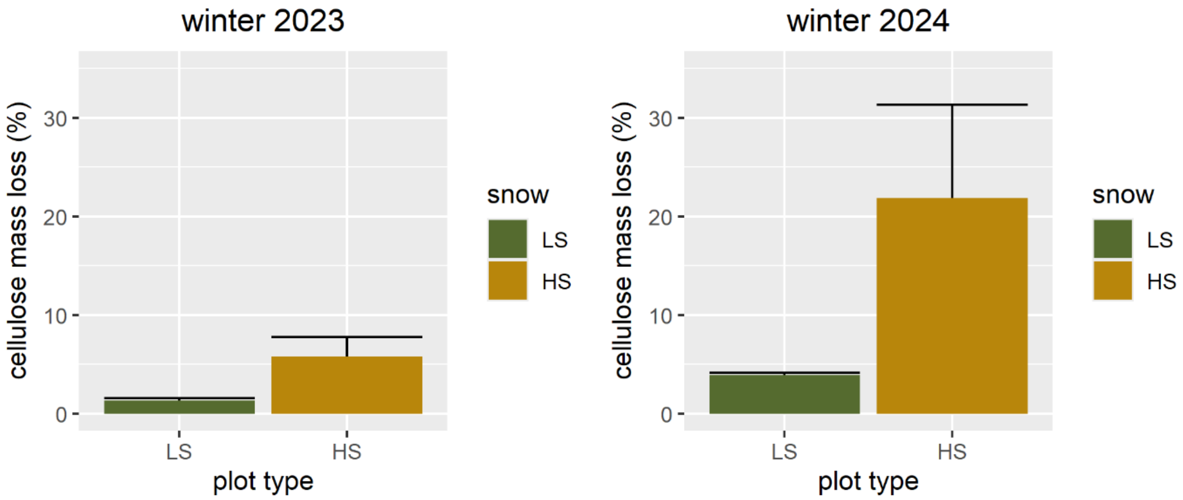

Figure 16: Decomposition (±SEM, n=3) at low snow (LS) and high snow (HS) plots in winters 2023 and 2024, respectively.

Table 4: T-test results for decomposition at LS/HS plots for winters 2023 and 2024. df: degrees of freedom, t: t-value, Significant results correspond with $p < 0.05$.

year	2023			2024		
	df	t	p-value	df	t	p-value
Plot type	4	-2.24	0.0887	4	-1.89	0.131

Decomposition was not influenced by snow manipulation treatment as assessed in spring (Fig. 17). The expected trend at the treatment subplots was from the lowest to highest decomposition rate at R, C, A plots, respectively. In line with the expectation, there was a tendency for the snow removal treatment to reduce decomposition compared to control and snow addition, especially in the high snow plots, but this was not statistically significant (Table 5). Insignificant results could have been caused by poor manipulation. Snow removal can be considered only a temporary disturbance, given that the manipulation was carried out just once. Subsequently, the area could have been covered by drifting snow or subjected to snowfall. On the other hand, after the snow manipulation, the soil could have rapidly frozen. In that case, the subsequent snow cover would have had little effect on the results.

Furthermore, the unexpected highest decomposition rate within the control plot could suggest that the insulation effect was moderated due to altered snowpack density. Nonetheless, O'Lear and Seastedt, (1994) provide a different explanation as they gained similar results in snow manipulation plots. Lower mass loss at add plots could have been caused by prolonged snowpack duration, thus later snowmelt prevents soils to be heated by warmer spring temperatures and decomposition is attenuated.

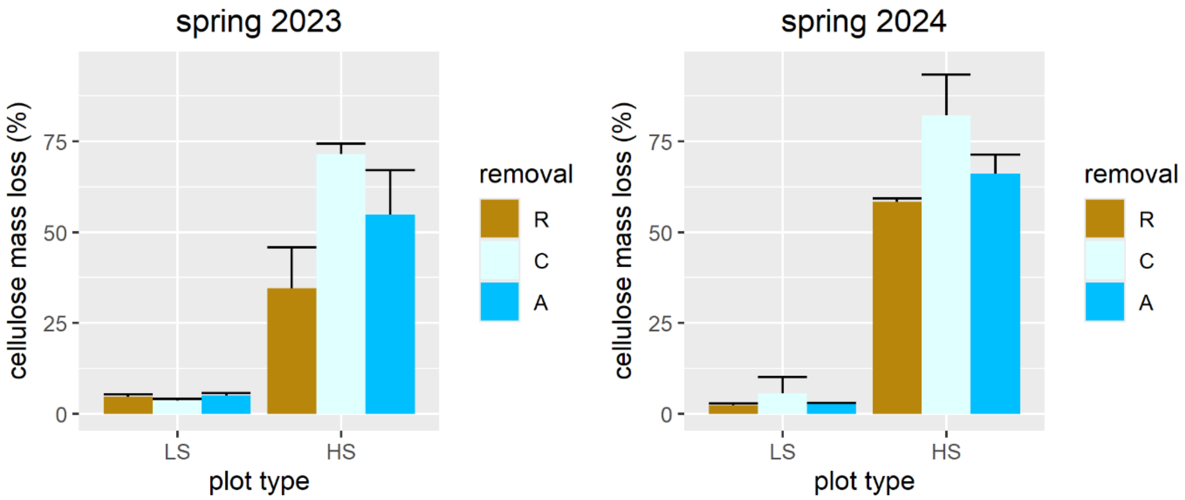

Figure 17: Cellulose mass loss (±SEM, n=?) at low snow (LS) and high snow (HS) plots with subplot treatment R (removal), C (control), A (add) in springs 2023 and 2024.

Table 5: Scheirer-Ray-Hare test results for cellulose mass loss for spring 2023 and Two-way ANOVA results for cellulose mass loss for spring 2024. df: degrees of freedom, F: F-value, Significant results correspond with p < 0.05. Significance codes: 0 '***' 0.001 '**' 0.01 '*' 0.05 '.' 0.1 ' ' 1.

year	2023				2024			
	df	H	p-value	Sig.	df	F	p-value	Sig.
Plot type	1	12.79	0.00035	***	1	192.11	2.611e-08	***
treatment	2	0.25	0.884		2	3.25	0.0779	.
Plot type:treatment	2	1.20	0.548		2	1.57	0.250	

residuals	12				11			

3.4.4 Soil Respiration

CO_2 flux was not significantly influenced by plot type in 2023 but was in 2024 (Table 4). Measurements taken in 2023 were conducted only once per plot, which likely introduced greater variability into the dataset and could have influenced the significance of the results. Therefore, the third hypothesis may be confirmed only based on the results from 2024.

Overall, there were higher rates in the 2023 season. The temperatures were below the freezing point and did not differ significantly (fig. 18), which can explain the similar rates of respiration among the plot types. Nevertheless, Goldberg et al. (2008) suggest that the longer frost duration decreases soil respiration. The 2023 plots had additional moisture due to snowmelt at the beginning of January (Fig. 14), which could have enhanced microbial activity and led to a higher respiration rate.

As shown in the soil temperature graph for 2024 (fig. 15), temperatures fluctuated around the freezing point at HS plots, while LS plots exhibited even lower average temperatures. LS plots may be insufficiently supplied by moisture, which results in a lower rate of respiration. At this temperature range, between -1°C and 1°C, respiration is sensitive to moisture (Aanderud et al., 2013), scarce at freezing temperatures.

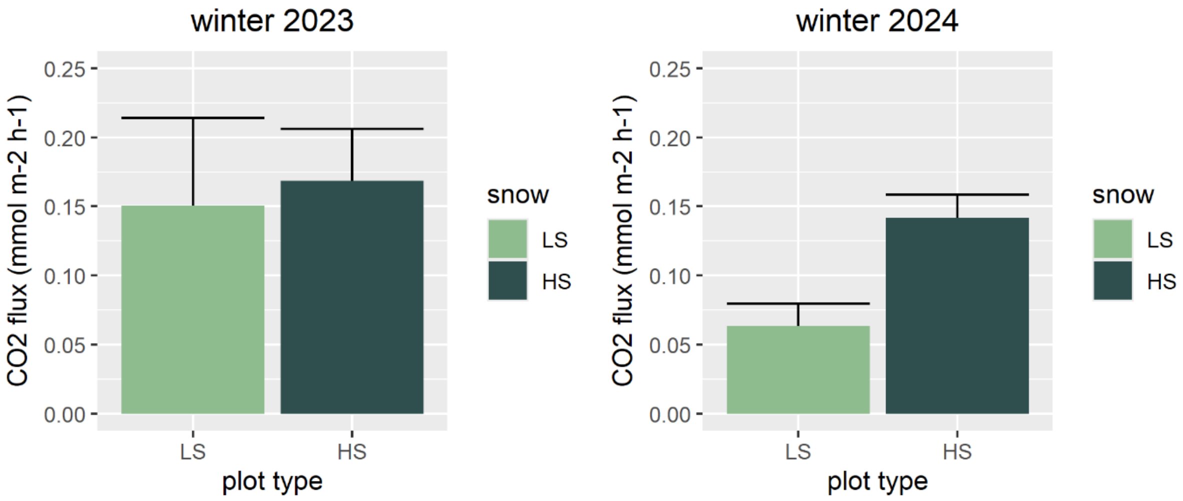

Figure 18: CO_2 flux (±SEM, n=3) at low snow (LS) and high snow (HS) plots in winters 2023 and 2024, respectively.

Table 6: T-test results for CO_2 flux at LS/HS plots for winters 2023 and 2024. df: degrees of freedom, t: t-value, significant results correspond with $p < 0.05$.

year	2023			2024		
	df	t	p-value	df	t	p-value
Plot type	4	-0.24	0.820	4	-3.37	0.0281

3.4.5 Potential predictors of winter decomposition and soil respiration

Decomposition and respiration were not correlated and responded differently to snow depth and different temperature indicators (Fig. 19).

Decomposition in 2023 positively correlated with *snow depth* and *ratio temperature higher than 0°C*. In 2024, *decomposition* was positively correlated with *ratio temperature higher than 0°C, mean temperature* and *time with temperature higher than 0°C*, but not with *snow depth*. *Therefore,* common significance for both winters is only for *ratio temperature higher than 0°C.* Conclusively, the number of days with positive temperatures seems to be the best predictor of the decomposition rate out of the tested potential drivers.

Soil respiration was not correlated with any of the potential drivers in 2023. In 2024, soil respiration was positively correlated with *snow depth*, *mean temperature,* and *mean temperature over the last 3 days*. The most significant correlation was with *mean temperature over the last 3* days. Conclusively, the temperature of recent days could be the best predictor of soil respiration out of the tested potential drivers.

Significant correlations between decomposition and soil respiration support the assumption that decomposition is more dependent on the long-term temperature conditions of the soil, whereas soil respiration responds more directly to the current state of the soil.

Additionally, snow depth was not correlated with any of potential drivers in 2023. However, in 2024, snow depth was positively correlated with *mean temperature* and *mean temperature in the last three days*. The latter is close to significance in 2023 too. Conclusively, the positive correlation between *snow depth* and *mean temperature* partially confirms the first hypothesis, that soil temperature decreases due to snow reduction.

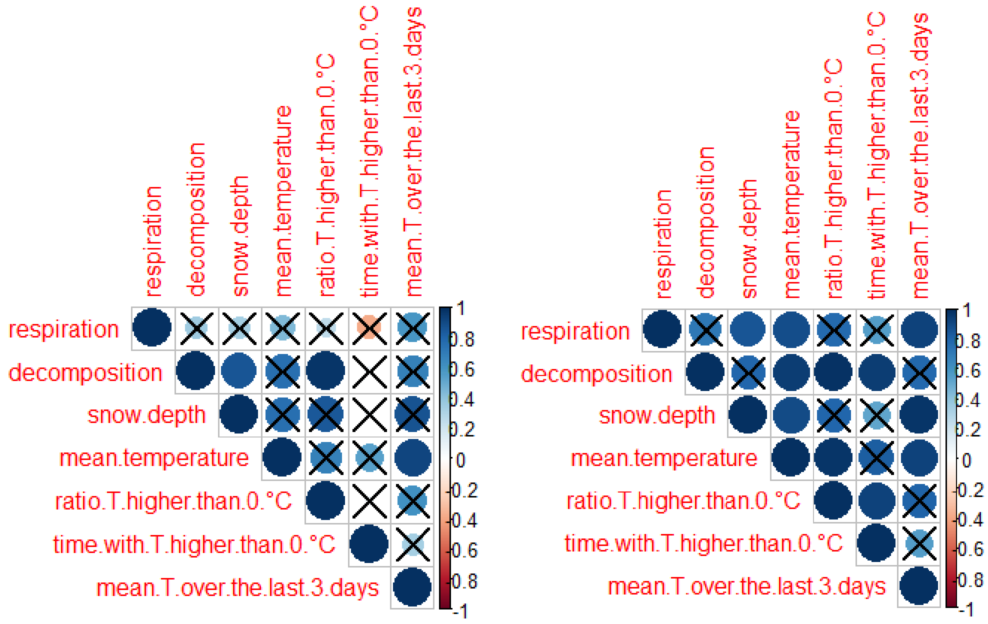

Figure 19: Correlation matrix for all factors measured in winters 2023 and 2024, respectively. Significant results correspond with $p < 0.05$. The crossed-out points are not significant.

3.5 Conclusions

Rising temperatures and earlier spring melt drive the global decline in snow cover. The duration and depth of snow cover are expected to decrease globally in the future, including in the Czech Republic. Snow cover duration is primarily influenced by earlier spring melt and is expected to decrease, even if the onset of snowfall in autumn occurs earlier. However, in some of the coldest areas, an increase in snow depth is projected.

As a result of the reduced insulating snow layer, soils will be increasingly exposed to extremes of freeze-thaw cycles (FTC). During freeze periods, the soil is enriched with new nutrients and organic matter due to increased mortality and enhanced litter fragmentation, but the activity of soil biota is halted. During thaw periods, the necromass is utilized by soil biota which becomes active again. The contrasting effects of these conditions on soil biota activity have variable impacts on decomposition and soil respiration throughout the winter. However, the prevailing effect is a decrease in both decomposition and soil respiration. The impact of reduced snow cover on decomposition and respiration is dependent on litter input, fluctuating temperatures, moisture availability, variable spatial and temporal conditions. Despite the non-negligible decomposition and soil respiration during winter, their rates are at their highest during spring thaw. Preceding winter conditions partly determine the intensity of these processes throughout the following growing season. Therefore, assessing the overall carbon cycle balance requires a year-round perspective on the interactions between these processes.

To estimate the effect of reduced snowpack on soil, snowpack manipulation methods are used, such as shoveling, snow dusting, constructing snow fences or shelters, and utilizing natural gaps in the forest canopy. To ensure consistent results in decomposition and soil respiration studies, it is essential to standardize the number, duration, and timing of measurements.

The results of the snow reduction experiment on soil temperature, decomposition, and soil respiration at Králický Sněžník partially confirmed that the rates of decomposition and soil respiration decrease with reduced snow cover, in line with the second and third hypotheses. A positive correlation between snow depth and mean temperature indicated support for the first hypothesis—that soil temperature decreases with lower snow cover. From the results of potential predictors, it could be inferred that decomposition depends on long-term temperature trends while soil respiration reacts to temperatures of recent days.

3.5.1 Gaps in knowledge and recommendations for further research

Understanding the impact of snow cover reduction on soil processes is particularly important for gaining a clearer picture of the carbon balance between the soil and the atmosphere. This impact varies significantly under different spatial and temporal conditions, highlighting the need for research across various ecosystems and their subsequent comparison.

Further research coupling decomposition and soil respiration is essential to evaluate the potential of carbon retention in soil while the snow cover is reduced. In addition, snowpack reduction may potentially alter both the quality and quantity of litter, which can also influence the carbon balance. It is therefore crucial to monitor carbon fluxes and decomposition not only during the winter season but also throughout the year and over longer time scales, as only then can the overall impact on the carbon balance be properly assessed.

The snowpack properties strongly affect all the results, as snow depth, duration, onset, and snowmelt fluctuate throughout winter. Long-term and continuous data collection is needed to prevent uneven results. Single data collection within the examined winter period cannot demonstrably ascertain the impacts and influences of measured variables.

As for the next steps in the analysis of data from Králický Sněžník, results from the different seasons could be statistically compared and key drivers of variable results within separate years could be explored. Abiotic site conditions such as air temperature and the temporal trend of snowpack should be involved in such data analysis.

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