Key Points:
- Deeper snow increased winter soil temperatures and CO_2 emissions of Arctic tundra
- Late growing season emissions were driven by year- to decade-old C ($\Delta^{14}C = 62 \pm 8‰$) and winter emissions by Holocene soil and fossil shale ($\Delta^{14}C = -64 \pm 14‰$)
- In gelisols, weathering of C-rich parent material may contribute to C emissions

Supporting Information:
- Supporting Information S1

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Winter Ecosystem Respiration and Sources of CO₂ From the High Arctic Tundra of Svalbard: Response to a Deeper Snow Experiment

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Abstract Currently, there is a lack of understanding on how the magnitude and sources of carbon (C) emissions from High Arctic tundra are impacted by changing snow cover duration and depth during winter. Here we investigated this issue in a graminoid tundra snow fence experiment on shale-derived gelisols in Svalbard from the end of the growing season and throughout the winter. To characterize emissions, we measured ecosystem respiration (R_{eco}) along with its radiocarbon (^{14}C) content. We assessed the composition of soil organic matter (SOM) by measuring its bulk-C and nitrogen (N), ^{14}C content, and n-alkane composition. Our findings reveal that greater snow depth increased soil temperatures and winter R_{eco} (25 mg C m^{-2} d^{-1} under deeper snow compared to 13 mg C m^{-2} d^{-1} in ambient conditions). At the end of the growing season, R_{eco} was dominated by plant respiration and microbial decomposition of C fixed within the past 60 years ($\Delta^{14}C = 62 \pm 8‰$). During winter, emissions were significantly older ($\Delta^{14}C = -64 \pm 14‰$), and likely sourced from microorganisms decomposing aged SOM formed during the Holocene mixed with biotic or abiotic mineralization of the carbonaceous, fossil parent material. Our findings imply that snow cover duration and depth is a key control on soil temperatures and thus the magnitude of R_{eco} in winter. We also show that in shallow Arctic soils, mineralization of carbonaceous parent materials can contribute significant proportions of fossil C to R_{eco}. Therefore, permafrost-C inventories informing C emission projections must carefully distinguish between more vulnerable SOM from recently fixed biomass and more recalcitrant ancient sedimentary C sources.

1. Introduction

Permafrost-dominated landscapes in the Northern Hemisphere contain in excess of 50% of the organic carbon (C; 930–1,690 Pg C) found in soils globally (Hugelius et al., 2014; Tarnocai et al., 2009). A large fraction of this soil C, 50–90% (Hugelius et al., 2014), is frozen in permafrost and accumulated over millennia (Horwath et al., 2008). With ongoing permafrost thaw (Romanovsky et al., 2010), increasing amounts of ancient permafrost C will enter the atmosphere as the greenhouse gas carbon dioxide (CO_2) or methane (Burke et al., 2013; Elberling et al., 2013; MacDougall et al., 2012; Schaefer et al., 2011; Schneider von Deimling et al., 2012; Schuur et al., 2015). Yet the permafrost C feedback to climate change has not been fully integrated in global climate models (Arora et al., 2013).

High latitudes have warmed at almost twice the rate of the northern hemisphere mean over the past decades (Bekryaev et al., 2010). The most intense warming has been observed during winter, with up to 4 °C of warming in 30 years (Bintanja et al., 2011). Projections for snowfall for the Arctic, with the exception of Greenland, show a pronounced reduction during summer and autumn with an overall decrease in snow cover duration (Derksen & Brown, 2012). Conversely, winter precipitation is predicted to intensify considerably (Bintanja & Andry, 2017), increasing overall snowpack depth. Along with these projections, expansion of shrub populations (Epstein et al., 2013; Pearson et al., 2013) is modifying regional snowpack depth distribution (Myers-Smith et al., 2011). Changes in depth and timing of snow cover can have significant implications on regional biogeochemistry (i.e., C and nitrogen (N) cycling), albedo, and hydrology that,

however, remain highly uncertain (Callaghan et al., 2011; Hinzman et al., 2005; Steffen, 2006; Welker et al., 2005) and could have implications for constraining the fate of thawing permafrost C.

In the High Arctic (areas above 70°N), winter is the dominant season that lasts from September through May. Thus, any changes in the biogeochemical processes during winter, such as CO_2 efflux (Björkman et al., 2010; Fahnestock et al., 1998, 1999; Jones et al., 1999; Morgner et al., 2010; Nobrega & Grogan, 2013; Oechel et al., 2014, 1997; Semenchuk et al., 2016) and N and phosphorus mineralization (Hobbie & Chapin, 1996; Schimel et al., 2004, 2006; Semenchuk et al., 2015), can represent a large portion of the annual budgets. Winter C fluxes remain a key unknown in estimating the annual C balance of the tundra (Belshe et al., 2013; Euskirchen et al., 2012; Euskirchen et al., 2016; McGuire et al., 2012).

Snow's impact on wintertime biogeochemistry of Arctic ecosystems remains one of the greatest uncertainties to tundra C processes. Deeper snowfall, especially in early winter, prevents the soils from cooling rapidly, and thus, soil temperatures remain warmer compared to soil temperatures under ambient temperatures (Hinkel & Hurd, 2006; Lafrenière et al., 2013; Morgner et al., 2010; Webb et al., 2016; Welker et al., 2004). Deeper snow can stimulate microbial activity in winter and plant N supply (Leffler & Welker, 2013) and leaf-level C fixation and ecosystem C sequestration (Lupascu, Welker, Seibt, Maseyk, et al., 2014; Pattison & Welker, 2014; Rogers et al., 2011) during the subsequent growing season. When these deep snow conditions are experimentally put in place for many years, the accumulated consequence is a deepening of the active layer (Park et al., 2015) and, depending on ground ice content, subsidence and changes in surface hydrology (Hinkel & Hurd, 2006), plant phenology, growth, and fungal attack (Cooper et al., 2011; Olofsson et al., 2011; Rumpf et al., 2014; Semenchuk et al., 2013). Increase in soil temperature may result in greater wintertime ecosystem respiration (R_{eco}; Brooks & Williams, 1997; Morgner et al., 2010; Nobrega & Grogan, 2013; Semenchuk et al., 2016; Walker et al., 1999; Webb et al., 2016; Welker et al., 2000). Furthermore, this might lead to ancient permafrost C, fixed by photosynthesis during the early Holocene or even the Pleistocene, becoming vulnerable to decomposition by soil microbes (Lupascu, Welker, Seibt, Maseyk, et al., 2014; Nowinski et al., 2010). Similar to emissions of ^{14}C-free C from the combustion of fossil fuels, emissions of ancient C that has accumulated in permafrost soils and disconnected from the modern C cycle represent a net flux to the atmosphere. However, it is still uncertain to which degree ancient permafrost C contributes to R_{eco} under current and future winter conditions.

Our understanding of the sources and ages of respired CO_2 from landscapes underlain by permafrost (tundra and taiga) has been greatly enhanced by the use of the natural abundance of radiocarbon (^{14}C) isotopes in the environment (Czimczik et al., 2006; Hartley et al., 2012; Hicks Pries et al., 2013; Lupascu, Welker, Seibt, Maseyk, et al., 2014; Nowinski et al., 2010; Schuur et al., 2009; Schuur & Trumbore, 2006).

Because ^{14}C decays at a constant rate, C fixed by ancient ecosystems is isotopically distinct from the modern C cycle. Specifically, we can distinguish (a) "fossil (^{14}C-free) C," which is present in soils and sediments formed during or before the Pleistocene; (b) "ancient C," which accumulated in permafrost during the late Pleistocene and early Holocene; c) "older C," which is residing in soils for centuries; and (d) "modern" C, which was fixed from the atmosphere after 1950s (e.g., Elder et al., 2018).

Up until now, our estimates of permafrost C emissions from tundra have, however, been confined to the growing season (Czimczik & Welker, 2010; Hicks Pries et al., 2013, 2015; Lupascu, Welker, Seibt, Maseyk, et al., 2014; Lupascu, Welker, Seibt, Xu, et al., 2014; Lupascu, Welker, Xu, et al., 2014; Natali et al., 2015, 2011; Nowinski et al., 2010), with the only winter losses estimated in a single alpine tundra in interior Alaska (Schuur et al., 2009; Trucco et al., 2012).

Our study addresses two questions in the High Arctic of Svalbard: (a) what are the magnitudes and sources of R_{eco} and in the soil pore CO_2 of graminoid tundra at the end of the growing season and throughout the winter and (b) what are the effects of deeper snow on R_{eco} fluxes and sources?

2. Material and Methods

2.1. Site Description

This study was conducted in the Adventdalen valley (78°10′N, 16°06′E) on the island of Spitsbergen, Svalbard, Norway, in the continuous permafrost zone (Figure 1). Adventdalen is situated in one of the warmest fjords
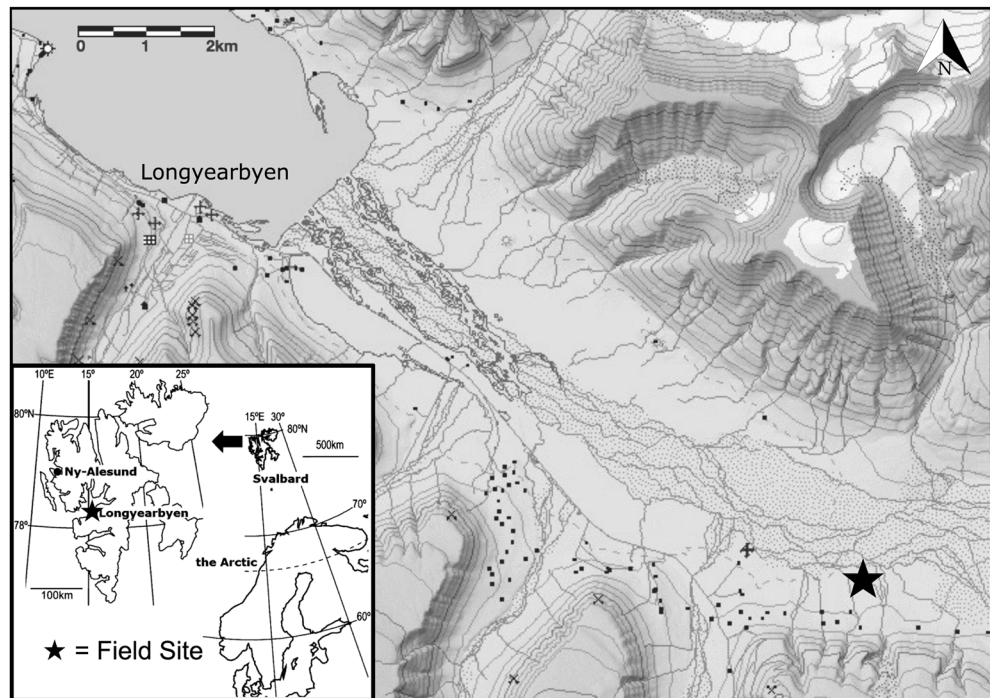
Figure 1. Maps showing the location of the snow fence study site in Adventdalen Valley.

(Isfjorden) in Svalbard. The mean annual air temperature at Svalbard airport is below −10 °C, with a minimum of −14.0 °C in February and a maximum of 6.3 C in July (1976–2010; Bednorz & Kolendowicz, 2013). The study area is also in one of the driest parts of Svalbard, with a mean annual precipitation of 191 mm (1981–2010), deposited predominantly as snow between November and March (Eckerstorfer & Christiansen, 2011; Humlum, 2002). The region has undergone recent warming and wetting, with an increase in temperature of 0.25 °C and in precipitation of 1.9% per decade (1912–2011, Svalbard airport; Førland et al., 2011).

Adventdalen is a SE-NW oriented glacial valley, covered by unconsolidated glacial, colluvial, alluvial, marine, and eolian deposits, and drained by the braided Adventelva River. Adventelva tributaries have built large alluvial fans along the valley sides. Valley sediments were deposited, permafrost aggraded, and overlying soils formed during the Holocene (6,800 to 3,000 years ago). Previously, Adventdalen was glaciated until enhanced orbital forcing resulted in upvalley glacial retreat and 60–70-m rebound of the valley bottom (Gilbert, 2014; Lønne & Nemec, 2004; Mangerud et al., 1992). Soil parent materials are complex, and include Paleozoic and Mesozoic siliciclastic rocks, including sandstones and siltstones, coal, and shale (Lønne & Nemec, 2004; Major et al., 2001).

Our experiment was conducted within a long-term snowpack manipulation experiment using three (of the 12 established) 1.5-m-tall and 6.2-m-long snow fences in an area of approximately 0.04 km^2 on a fluvial terrace (Cooper et al., 2011; Morgner et al., 2010; Figure 1). Fences were erected in fall 2006, perpendicular to the dominant winter southeastern wind direction and created a continuously tapering snowdrift on their leeward side.

The vegetation at our study site was Arctic graminoid tundra (meadow), dominated by *Alopecurus magellanicus* LAM., *Luzula arcuata* (WAHLENB.) SW. subsp. *confusa*, *Bistorta vivipara* (L.) GRAY, *Salix polaris* WAHLENB., and *Dryas octopetala* L. (Cooper et al., 2011), while *Cassiope tetragona* was present in patches. Soils were Histels overlying alluvial gravel derived from shale.

2.2. Experimental Setup and Data Collection in the Field

At each snow fence (*n* = 3), we investigated two snow depth regimes: natural snow cover ("ambient," 10–35 cm) and experimentally increased snow cover ("deep," approximately 150 cm) both measured in a
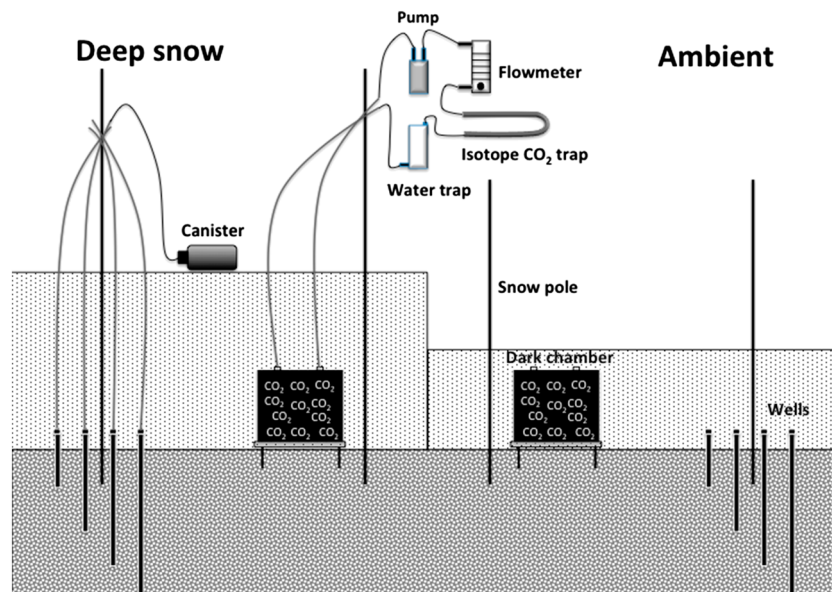
Figure 2. Sampling design for collecting CO_2 produced in soils with gas wells and CO_2 emitted from the soil surface with chambers during winter in areas with experimentally increased or ambient snow. Wells were inserted in sets to 30-, 50-, 70-, and 90-cm depth, and accessed from common sampling points. Soil gas was collected in preevacuated canisters via flow-restricting capillaries. Chambers were placed on chamber bases, inserted into the soil to 5 cm, and CO_2 was collected on molecular sieve traps. Chamber tops were only placed onto the bases during sampling prior to snowfall, and then permanently left on the bases, but flushed with ambient air prior to sampling (dotted white boxes represent snowpack).

snow survey in March 2013. At each snow regime, we monitored the rate and isotopic composition of R_{eco} and the concentration and isotopic composition of CO_2 within the soil pore space ($n = 3$/snow regime) 3 times between September 2012 and April 2013 (Figure 2).

Before the first snow, R_{eco} was measured in early September and October 2012 using opaque, dynamic chambers (30-cm i.d., 30-L volume; Lupascu, Welker, Seibt, Maseyk, et al., 2014). Chamber bases had been inserted in mid-August 2012 to about 5-cm depth, sealed with soil material on the outside, and left in place during the entire study period. To calculate R_{eco}, air was circulated between the chamber's headspace connected to an infrared gas analyzer and a data logger (LI-840, LI-1400, LI-COR Biosciences, Lincoln, NE, USA) at a rate of 0.5 L/min. Flux rates were estimated from the slope of time versus CO_2 concentration curves using linear regression.

In April 2013, when there was an established snowpack, winter R_{eco} was measured using a gradient approach, where CO_2 concentrations were recorded at the snow-soil interface and in the ambient air (Fahnestock et al., 1998, 1999). In combination with snow density profiles through the snowpack, CO_2 fluxes were calculated using Fick's law [1]:

$$J_g = D_g(d[g]/dz)\, f \qquad (1)$$

where J_g is the gas flux, D_g is the diffusion coefficient of CO_2 in air (1.39×10^{-5} m^2/s), g is the difference between $[CO_2]_{atm}$ and $[CO_2]_{measured}$, z is the snow depth, and f is the snow porosity ($1 - $ (density/973) measured in March 2013 (ambient 355 g/m^3 and deep snow 380 g/m^3; $n = 5$). Winter R_{eco} measurements were combined as "winter" period.

We collected R_{eco} for ^{14}C analysis on 7 September and 26 October 2012, and during 3–5 February and 10–12 March 2013. Before snowfall, chambers were left closed until the CO_2 concentration inside the chamber was about twice that in ambient air (up to 48 hr at the end of the summer and up to 72 hr in winter). After measuring the CO_2 concentration inside the chamber headspace, CO_2 was collected by circulating the air inside the chamber through drierite (W.A. Hammond Drierite Co. Ltd., Xenia, OH, USA) followed by a preconditioned, activated molecular sieve (powder-free 133 8/12 beads, Grace) trap at a rate of 0.5 L/min for 15 min (Gaudinski et al., 2000).

Before the beginning of winter, long metal extensions (up to 150-cm length) were connected to the chamber ports for sampling above the snowpack. At the end of each capillary, stopcocks were attached for connecting the sampling system (Figure 2). During each sampling event, we also collected CO_2 in ambient air on molecular sieve traps.

Soil CO_2 for ^{14}C analysis were collected via stainless steel gas wells (0.35-cm i.d., 0.6-cm o.d.) from 25-, 45-, 65-, or 85-cm soil depth (Figure 2). Wells were inserted at the end of August 2012, extended to a common sampling point above the snowpack with Bev-A-line tubing, capped with luer lock stopcocks, and left in the ground during the entire study period. Soil CO_2 was sampled in late summer (7 and 21 September), fall/early winter (19–21 October 2012), and late winter (3–5 February and 10–12 March 2013) in 1-L evacuated stainless steel canisters via flow-restricting stainless steel capillaries (0.010 × 0.063 × 30 cm; Fisher Scientific, Pittsburgh, PA, USA) to minimize disturbing the soil CO_2 concentration gradient and sampling air from other than the sampling depth (Gaudinski et al., 2000).

In each treatment (n = 1/treatment), we also continuously monitored soil temperature (S-TMB-M002, Onset, Bourne, MA, USA; accuracy ±0.3 °C) and soil moisture contents (Onset EC5; ±1%). Soil data were collected every 30 min at 30- and 50-cm soil depth under ambient snow and at 10-, 30-, and 50-cm depth under deep snow. Air temperature data at 2 m above ground was obtained from the University Centre in Svalbard (UNIS) Adventdalen micrometerological station that was about 3 km from the study site (http://www.unis.no/20_RESEARCH/2060_Online_Env_Data/weatherstations.html).

We collected soil samples and characterized the composition of soil organic matter (SOM) via analyses of the elemental content and isotopic ratios of C and N, and of n-alkanes. Soil samples were collected with a trowel in August 2012 from the ambient and deep snow experimental plots (to about 40-cm depth) and from a pit dug into a nearby river terrace under ambient snow (to 160-cm depth), stored frozen, and dried at 60 °C at UNIS.

2.3. Laboratory Analyses

Sample CO_2 was released from molecular sieve traps (R_{eco}) by baking at 650 °C for 45 min or extracted from canisters (gas wells) using a vacuum line, purified cryogenically, and reduced to graphite via Zn reduction (Xu et al., 2007). For samples exceeding 0.3 mg C, an aliquot of the CO_2 was analyzed for its $\delta^{13}C$ ratio (GasBench II, DeltaPlus XL, Thermo, Waltham, MA, USA). The ^{14}C content of the graphite was measured with accelerator mass spectrometry (NEC 0.5MV 1.5SDH-2 AMS) alongside processing standards and blanks at the KCCAMS laboratory of UC Irvine (Beverly et al., 2010). The measurement uncertainty for $\Delta^{14}C$ was <2‰, and years are expressed as ^{14}C ages (years BP).

The ^{14}C content of R_{eco} was corrected for the amount of CO_2 from ambient air present in each chamber's headspace:

$$\Delta^{14}C_{cor} = \frac{\Delta^{14}C_{obs} - \left(f_{air} \times \Delta^{14}C_{air}\right)}{\left(1 - f_{air}\right)} \qquad (2)$$

with $\Delta^{14}C_{cor}$ being the actual ^{14}C content of R_{eco}, $\Delta^{14}C_{obs}$ the measured ^{14}C content of a given sample, and f_{air} the fraction of CO_2 derived from ambient air, calculated as the difference of the CO_2 concentrations inside the chamber immediately before trapping and in ambient air.

Soil pH was measured along the soil profile from subsamples collected from the ambient pit core with an Orion 250A pH meter (resolution 0.01/0.1 pH; accuracy ±0.02). For elemental and isotopic analyses, soil samples were sieved (<2 mm), and a subsample was ground to powder and acid-washed with 2 M HCl to remove inorganic C. The C and N elemental and stable isotopic composition was measured alongside processing standards and blanks with EA-IRMS (Fisons NA-1500NC, DeltaPlus XL, Thermo). The measurement uncertainty was 0.1‰ for $\delta^{13}C$ and 0.2‰ for $\delta^{15}N$ (1σ, from long-term measurements of secondary standards). For ^{14}C analysis, soil samples were combusted to CO_2 in precombusted, evacuated quartz tubes with cupric oxide for 2 hr at 900 °C at the University of California, Irvine, and then processed like gaseous CO_2 samples (see above).

Alkane analyses were performed at the University of South Carolina, School of the Earth, Ocean, and Environment, USA. The lipid fraction from SOM was extracted with dichloromethane/methanol/
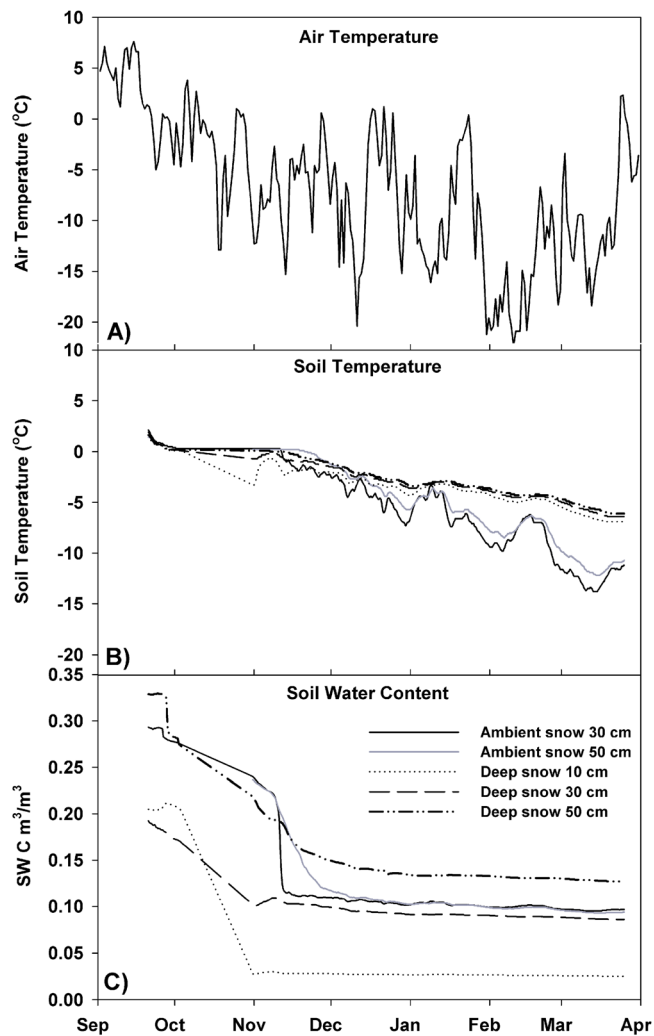
Figure 3. Weather and soil conditions during the study period from September 2012 to May 2013. (a) Air temperature at the UNIS Adventdalen micrometerological station. (b) Soil temperatures at the study site at 10-, 30-, and 50-cm soil depth under ambient and deep snow. (c) Volumetric soil water content at the study site at 10-, 30-, and 50-cm soil depth under ambient and deep snow.

phosphate buffer (1:2:0.8) and sonicated at room temperature before being left to extract at room temperature overnight. Depending on the amount of organic C in the sample, between 2 and 160 gr of dried sample material was used for extraction. The total extract was separated on a silica gel column into three fractions: aliphatic hydrocarbons and aromatic hydrocarbons (dichloromethane), neutral lipids (acetone), and polar lipids (methanol). Only compounds in the hydrocarbon fraction were used for this study, while the other fractions were stored for future analyses. Hydrocarbon compounds were identified and quantified using a gas chromatograph with mass selective detector (GC-MSD; Agilent 7890B/5977A) and compared with an external n-alkane standard mixture (Sigma-Aldrich). Scanning ion monitoring (SIM) using ion m/z of 57 was used for the identification and quantification of n-alkanes. Laboratory blanks were analyzed with each set of samples to assess contamination.

Lipid analyses were used to calculated proxies and indices that relate to the quality of the soil organic C, for example, the C preference index (CPI) and average chain length (ACL). The CPI was introduced by Bray and Evans (1961) as the ratio of odd- to neighboring even-numbered alkanes; this is a measure of the alteration of organic matter. The CPI is used as a degradation and alteration proxy by quantifying the odd over even (n-alkanes, employed here) or even over odd (n-fatty acids) predominance of the C chains (Bray & Evans, 1961; Glombitza et al., 2009). A low CPI is indicative of mature or degraded organic matter (e.g., CPI of shale ≈1). The n-alkane ACL value is the concentration-weighted mean of different C chain lengths in a geological/soil sample (Poynter, 1989), and is proxy for organic C sources. We used the C23–C33 interval, since vascular C3 plants are expected to have an ACL of about 28.5.

2.4. Statistical Analyses

Statistical analyses were performed using SPSS Statistical Software. Group means (ambient versus deep) were compared using t tests. Differences were considered significant when $p < 0.01$. The sample size of ^{14}C measurements of R_{eco} is 20, while that of soil CO_2 is 43 ($n = 3$). Standard deviations reported for ^{14}C content of R_{eco} reflect spatial variability.

3. Results

3.1. Soil Temperature, Moisture, and pH

Air temperatures in Adventdalen exhibited a typical seasonal decline from summer into winter, with a minimum in mid-March of −21 °C (Figure 3a). In the fall, soil temperatures at the onset of freezing (0.6 ± 0.0 °C) were similar between the snow treatments and soil depths (Figure 3b). Soil temperatures in each snow treatment cooled throughout the winter, and cooled more near the surface than at depth. Under deep snow, soil temperatures remained higher than under ambient snow, and were up to 8 °C warmer in later winter.

Soil water contents (SWC) were higher at the end of the growing season and rapidly decreased during the onset of winter (Figure 3c). At 30-cm depth, SWC was higher under ambient than deep snow at the end of the growing season, but similar during the winter (~0.10 m³/m³). Winter SWC was similar at 30- and 50-cm depth under ambient snow, but SWC increased with depth under deep snow, likely due to the warmer temperatures. The driest soils were found at 10 cm. Surface soil data from ambient snow conditions were lost due to sensor failure.

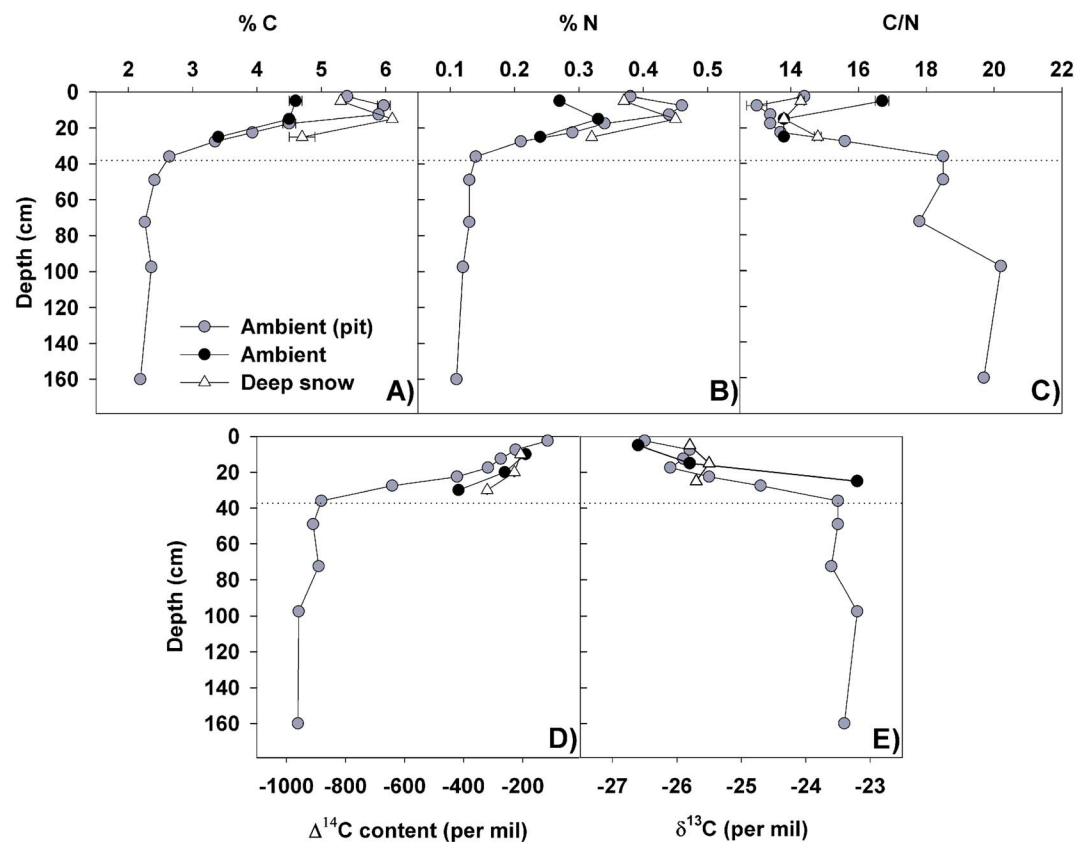Soil in situ pH values ranged from 5.0 near the surface to 6.5 at depth.

Figure 4. Composition of soil organic matter as a function of depth under ambient and deep snow at the study site and nearby riverbank ("pit"). (a) Bulk soil carbon content, (b) soil nitrogen content, (c) soil C/N ratio, (d) radiocarbon content, and (e) stable carbon isotope composition (dotted line indicates top of gravel (shale)).

3.2. Soil Organic Matter

The percentage of bulk organic C and N was high in the topsoil with a maximum of 5.4% C and 0.5% N at about 10-cm depth in ambient conditions, and then gradually decreased to a constant value at about 40-cm depth (~2.1 and 0.1%, respectively; Figures 4a and 4b and Table 1). In addition, the soil texture was distinctly different between the surface and deeper soil. This indicates that the soil contains at least two parent materials (clay and silt overlaying gravel), which is likely a consequence of this area being an abandoned river terrace. Consequentially, C/N ratios also sharply increased to a maximum of 20 at the soil-regolith boundary (Figure 4c). There was no clear difference in the C content in the soil under deep snow compared to under ambient snow.

All bulk soil C below the vegetation and litter layer (below 0-cm soil depth) was dominated by C with a mean age of >60 years (fixed from the atmosphere before thermonuclear bomb testing in the 1950s), as indicated by negative $\Delta^{14}C$ values (Figure 4d and Table 1). The mean age of C in the top ~40 cm of soil was $2,282 \pm 950$ years BP compared to the $21,054 \pm 4,270$ years BP in the deeper profile (>40 cm). The latter value likely indicates that this was a mixture of fossil (^{14}C-free, $\gg$50,000 years BP $= -1,000$‰ $\Delta^{14}C$), shale-derived C and SOM-C that may vary in age from current to Pleistocene. Bulk stable isotope data for $\delta^{13}C$ also showed a clear distinction between top (<40 cm) and deeper soil that was dominated by shale, with -26.0 ± 0.4 and -23.4 ± 0.1‰, respectively (Table 1 and Figure 4e).

Analysis of bulk C in the alkanes (Figure 5) showed that most of the SOM is highly degraded organic material mixed with fossil C from weathered shale, as indicated by low ACL (19.3 ± 0.1; Figure 5a) and low CPI (1.3 ± 0.1; Figure 5b) values. However, we also observed more plant-like material in the surface soil, with a maximum at 20-cm depth.

Table 1
Carbon (C) and Nitrogen (N) Elemental and Isotopic Composition of Soil Organic Matter Under Ambient and Deep Snow at the Study Site and Nearby Riverbank ("Pit"; Average ± 1σ, n = 1–2)

	Average depth (cm)	Total C (%mass)	Total N[a] (%mass)	C/N	$\delta^{15}N$	$\delta^{13}C$ (‰)	UCI-AMS no.[b]	Age (years BP)	$\Delta^{14}C$ (‰)
Ambient snow (pit)	2.5	5.40	0.38	14.39	1.17	26.50	124,988	925 ± 20	−115.5 ± 1.7
	7.5	5.97 ± 012	0.46	13.05 ± 0.26	1.97 ± 0.01	25.77 ± 0.03	124,989	1,975 ± 20	−223.8 ± 1.7
	12.5	5.89	0.44	13.37	1.99	25.91	124,990	2,510 ± 20	−273.8 ± 1.5
	17.5	4.50 ± 0.07	0.34	13.40 ± 0.04	1.29 ± 0.07	26.09 ± 0.03	124,991	3,010 ± 20	−317.6 ± 1.4
	22.5	3.93	0.29	13.75	1.96	25.54	124,992	4,345 ± 20	−422.3 ± 1.2
	27.5	3.35 ± 0.03	0.21	15.65 ± 0.05	2.15 ± 0.07	24.71 ± 0.13	124,993	8,175 ± 25	−641.4 ± 0.9
	36.0	2.64	0.14	18.54	2.94	23.53	124,994	17,065 ± 45	−881.4 ± 0.6
	49.0	2.41	0.13	18.55 ± 0.09	2.97 ± 0.11	23.47 ± 0.00	124,995	19,210 ± 60	−909.2 ± 0.6
	72.5	2.26	0.13	17.77	3.23	23.60	124,997	17,695 ± 50	−890.4 ± 0.7
	97.5	2.36 ± 0.01	0.12	20.22 ± 0.05	3.18 ± 0.14	23.21 ± 0.05	124,998	25,410 ± 110	−958.0 ± 0.6
	160.0	2.19	0.11	19.74	3.06	23.42	124,999	25,890 ± 120	−960.5 ± 0.6
Ambient snow	5	4.57 ± 0.08	0.27	16.66 ± 0.17	1.17 ± 0.07	26.58 ± 0.10	125,003	1,630 ± 20	−189.9 ± 1.7
	15	4.50 ± 0.04	0.33	13.79 ± 0.03	1.16 ± 0.09	25.80 ± 0.07	125,004	2,365 ± 20	−260.7 ± 1.4
	25	3.37 ± 0.01	0.24	13.83 ± 0.13	1.90 ± 0.17	25.14 ± 0.01	125,005	4,275 ± 20	−417.2 ± 1.2
Deep snow	5	5.31 ± 0.03	0.37	14.34 ± 0.13	0.09 ± 0.29	25.77 ± 0.06	125,000	1,800 ± 20	−206.8 ± 1.6
	15	6.15 ± 0.02	0.45	13.75 ± 0.07	0.23 ± 0.01	25.53 ± 0.06	125,001	2,020 ± 20	−228.1 ± 1.5
	25	4.70 ± 0.02	0.32	14.81 ± 0.06	0.83 ± 0.03	25.73 ± 0.06	125,002	3,040 ± 20	−320.2 ± 1.3

[a]1σ < 0.01. [b]Sample code of W.M. Keck carbon cycle accelerator mass spectrometer facility.

No difference between ambient and deep snow was detected in the alkane data. The alkane analyses also suggest that the shale becomes the major component of the soil at about ~40 cm (Figure 5), below which depth there is little evidence of roots.

We were unable to confidently estimate the mean age of C in SOM. If the meadow soil developed on top of the gravel bed, the mean age of bulk SOM in the surface soil was old, but within the estimated time span of pedogenesis in the region (Table 1). It is more likely, however, that the meadow soil is a mixture of more recent SOM and fossil gravel—as suggested by short ACL and low CPI values (Figure 5) and the soil's texture and C content (Table 1). A simple mass balance assuming a constant fraction of fossil shale-C in each depth interval resulted in unrealistically young SOM mean ages and several age reversals with depth (data not shown).
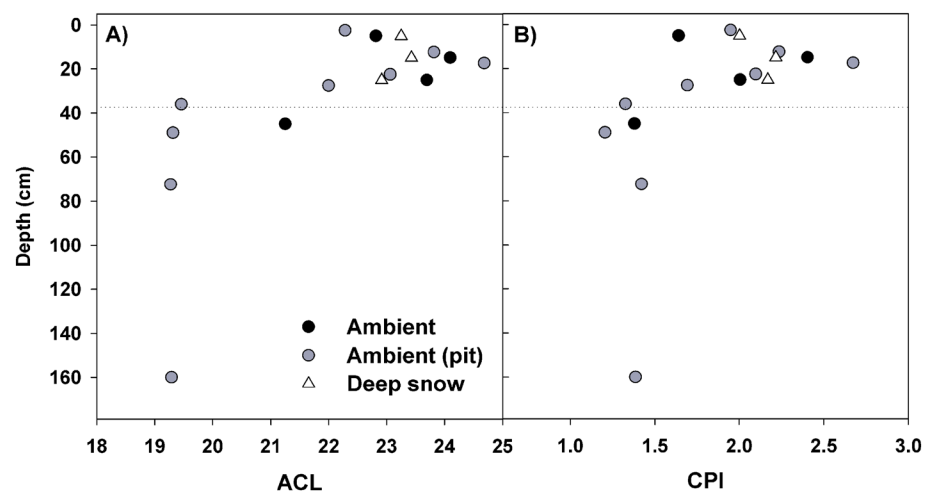

Figure 5. Composition of n-alkanes in soil organic matter under ambient and deep snow. (a) Average chain length (ACL) and (b) odd-to-even ratio (CPI) of n-alkanes as a function of depth (dotted line indicates top of gravel (shale)).
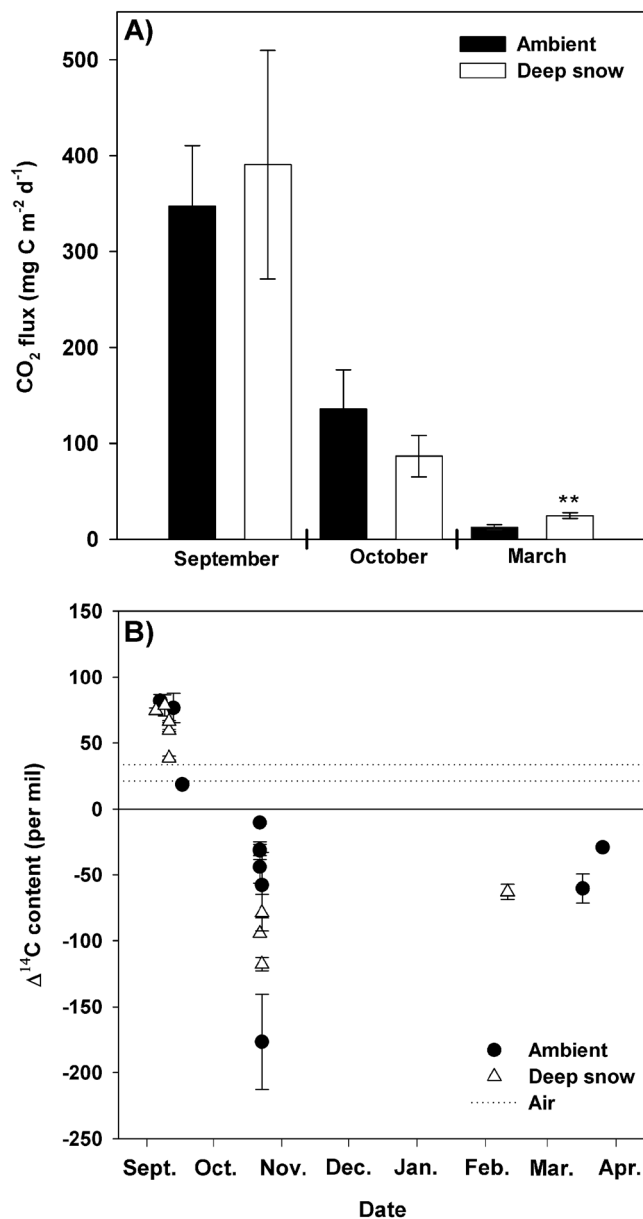
Figure 6. (a) Soil carbon efflux (mg C m^{-2} d^{-1}, average ± SE, $n = 6$; see also Table S1; $**p \leq 0.01$) and (b) ^{14}C content of emitted CO_2 at the end of the growing season and winter under ambient and experimentally increased snow in Svalbard's Adventdalen Valley (dotted lines indicate the range of $^{14}CO_2$ in the atmosphere for 2012–2013).

3.3. Ecosystem Respiration

Rates of R_{eco} declined from about 400 mg C m^{-2} d^{-1} at the end of the growing season to about 20 mg C m^{-2} d^{-1} in the winter (Figure 6a) with no significant difference ($p > 0.15$) in late summer or fall between ambient and higher snowpack plots. The snowpack initially developed in early November as soil and air temperatures fell to below 0 °C. Winter flux measurements were taken when the snowpack was fully developed. In winter, rates of R_{eco} were twice as high, ($p < 0.01$) where the snowpack was deep, averaging about 25 compared to 13 mg m^{-2} d^{-1} (Figure 6a and Table S1) in ambient conditions.

3.4. Radiocarbon Content of Ecosystem Respiration and Soil CO_2

The ^{14}C content of R_{eco} under both ambient and deep snow gradually declined from late summer to winter, with no significant differences between ambient and deep snow areas (Figure 6b and Table S2). At the end of the growing season, Δ^{14}C values of R_{eco} were modern, and slightly enriched (18 to 82‰) compared to atmosphere values, but after plant senescence, with the starting of winter, they declined dramatically to approximately -55 ± 52‰ in ambient areas, and to -88 ± 23‰ in deep snow areas. However, no significant statistical difference in the ^{14}C of R_{eco} was found between ambient and deep snow treatment.

Overall the ^{14}C content of soil CO_2 (Figure 7) exhibited more depleted values at under increased snowpack (mean -102 ± 19‰) than in ambient areas (mean -50 ± 14‰; unpaired t test, $p < 0.05$). No statistical difference between the sampling periods was recorded.

4. Discussion

4.1. Cold-Season CO_2 Emissions

During our study period, rates of R_{eco} from this High Arctic ecosystem exhibit a strong seasonal decline. Rates were highest at the end of the growing season, when air temperatures were about 10 °C, and declined by an order of magnitude in winter. This seasonal decrease in the rates of R_{eco} is closely associated with that in soil temperature from 0.9 ± 0.1 °C in September to -12.5 °C in March (Figure 3b). Seasonal reductions in R_{eco} rates of this magnitude have been reported from this study site in prior years (Björkman et al., 2010; Morgner et al., 2010; Semenchuk et al., 2016), and in the High Arctic of Northern Elsmere Island (Welker et al., 2004). When we compare R_{eco} rates below ambient snow to those below deep snow, we see a difference in late winter, but not during the fall (Figure 6a). In our late winter sampling, R_{eco} rates were twofold higher where snow was deeper. These higher R_{eco} rates are probably driven by a diversion in soil temperature between treatments as the winter progresses. In the fall, we see no difference between the two treatments as soil temperatures were very similar (Figure 3b). On the other hand, after January, soil temperatures were 4.9 ± 0.1 and 4.1 ± 0.1 °C warmer under deep snow at 30 and 50 cm, respectively, than under ambient snow (Figure 3b). Our soil temperature results confirm previous studies at this same site (Morgner et al., 2010; Semenchuk et al., 2016). Higher rates of winter R_{eco} from areas with experimentally deeper snow have also been reported in Arctic Alaska (Fahnestock et al., 1998, 1999; Jones et al., 1999; Welker et al., 2000) and Canada (Nobrega & Grogan, 2013).

Our results suggest that snowpack-mediated increases in soil temperature can stimulate C emissions from permafrost-affected soils. However, the timing and duration of this increase might differ at different locations as at our site soil temperature in the deeper-snow site increased many months (January) after the first snow compared to ambient conditions.
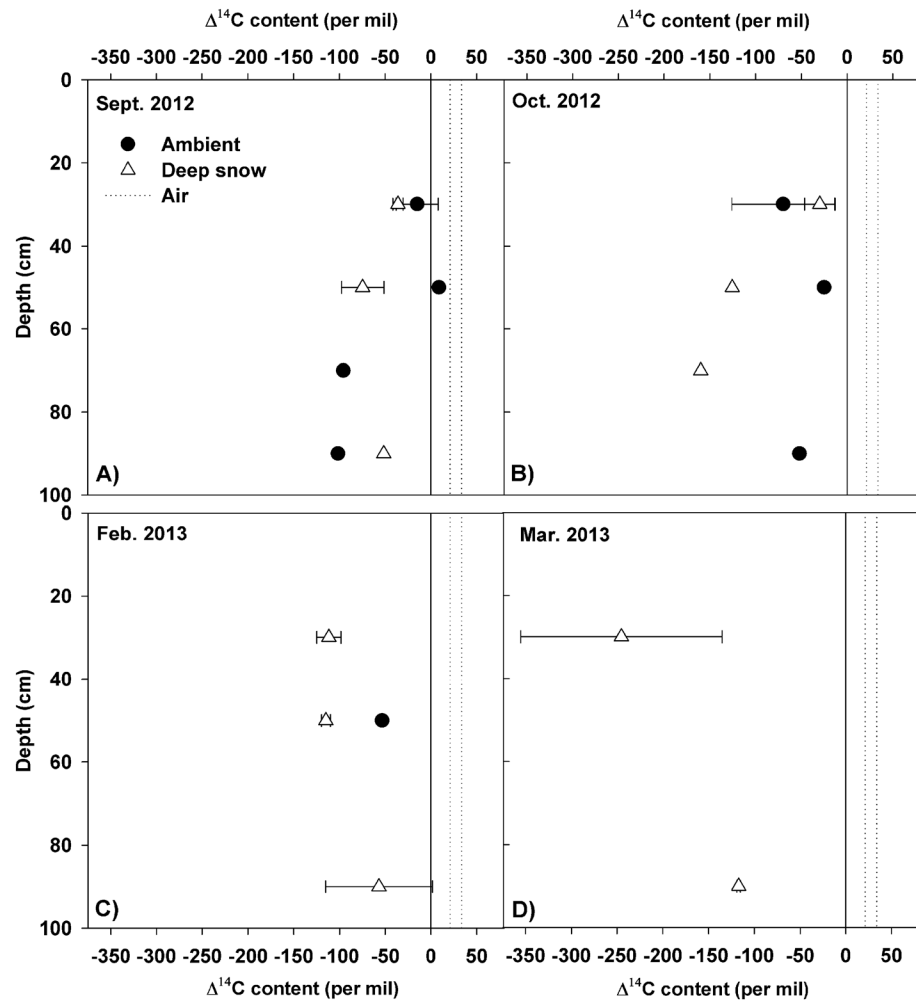
Figure 7. Soil ^{14}C-CO_2 (‰) content throughout the soil profile in areas with ambient and deep snow depth in Adventdalen in (a) 7 and 21 September 2012, (b) 19–21 October 2012, (c) 3–5 February 2013, and (d) 10–12 March 2013 (dotted lines indicate the range of $^{14}CO_2$ in the atmosphere for 2012–2013; many data points in February and march are missing due to the low CO_2 concentration and hence impossibility to run ^{14}C analysis and/or samples were air-contaminated).

At our site, deeper snow did not affect key soil parameters (organic C, N, C/N ratio (Figure 4), ACL, CPI (Figure 5)) but instead has caused a change in vegetation composition with less *Dryas* and *Cassiope*, more *Bistorta*, moss and dead material along with more bare ground in the deeper snow plots (data not published). A decline in aboveground biomass and an increase in root biomass have also been previously recorded (Semenchuk et al., 2016). However, these shifts in C allocation and in vegetation composition were not apparent in R_{eco} rates during the growing season, although plants are very active at this time. Consequently, we do not think that these factors could have led to this late winter R_{eco} increase under deeper snow especially considering that plants are dormant during this period (Edwards & Jefferies, 2013; Koller & Phoenix, 2017). Nevertheless, further studies are needed to seasonally apportion R_{eco} into plant and microbial emissions and to quantify how snowpack affects C allocation and respiration in plants year-round.

It is more likely that the differences in late winter R_{eco} between snow treatments were driven by soil microbial activity. Even small temperature changes can drastically increase liquid water availability and thus microbial activity (Öquist et al., 2009) and R_{eco}. This is particularly true for soil temperatures between 0 and −5 °C, below which there is a sharp decrease in microbial activity (Elberling & Brandt, 2003). A study by Mikan et al. (2002) further demonstrated that temperature controls are more important than substrate quality in controlling CO_2 production in laboratory incubation between 0 and −10 °C.

By the end of the century, climate projections see a reduction in Arctic snowfall during summer and autumn, when temperatures are close to the melting point, but a considerable intensification during winter (Bintanja & Andry, 2017). This, in combination with warmer temperature (Collins et al., 2013), could lead to warmer winter soils coupled with an increase in cold-season R_{eco} and hence results in an increased annual C loss (Morgner et al., 2010; Nobrega & Grogan, 2013; Schimel et al., 2004). Nonetheless, it is important to mention that this potential change in the annual C budget due to enhanced deeper snow may be not as significant, especially considering the magnitude of C loss during the winter time compared to the growing (Morgner et al., 2010; Semenchuk et al., 2016) and shoulder seasons (28 and 16 times lower in March compared to September, in ambient and deep snow plots respectively; this study). A potential increase in winter C loss could be partially offset by a reduced growing-season R_{eco} due to prolonged depletion in the labile soil C pool following multiple years of enhanced wintertime CO_2 release under deepened snow (Christiansen et al., 2018; Semenchuk et al., 2016).

4.2. Cold-Season Carbon Sources

Our data show that during fall, when plants are still active, sources of R_{eco} are modern ($>$1950) and become older once plants become dormant and temperature drops (Figure 3b).

During fall senescence, more modern C-rich plant litter accumulates on the soil surface and roots are actively acquiring and translocating nutrients below ground (Chapin & Bloom, 1976; Iversen et al., 2015) which are preferentially decomposed by microbes compared to more recalcitrant old C sources (Sinsabaugh et al., 2013), probably leading to our enriched ^{14}C-R_{eco} signature (Figure 6b). Sources of available SOM change throughout the early winter as the soil freezes from the top-down and upward from the permafrost table and throughout the winter as the soils continues to cool and dry. With colder temperature and dormant plants, a lack of a continuous supply of labile modern C from the roots (Edwards & Jefferies, 2013; Koller & Phoenix, 2017) can drive microorganisms to consume more of the recalcitrant old C compounds, as shown in a previous incubation study (Loya et al., 2004). In Loya et al. reductions of acid insoluble compounds—C (AIS; lignin, suberin, cutin, humic acids) over a simulated winter ($<$0 °C incubation) coupled with a decrease in the AIS ^{14}C signature, suggested that these relatively recalcitrant compounds can contribute to winter respiration in Arctic tundra soils.

Within this context, a study from Wild et al. (2016) showed how increased availability of plant-derived organic C could stimulate decomposition in permafrost subsoil horizons through priming (defined as an increase in the decomposition of native SOM stocks). On the other hand, a more recent study from Lynch et al. (2018) found no evidence of priming by low molecular weight carbon (aka root exudates) in Arctic soils underlying *Eriophorum vaginatum* and *Betula nana*. None of these studies, however, measured ^{14}C-R_{eco} and were not carried out in winter. We cannot exclude that priming could play a role in the release of old C during winter in both our treatments. However, priming is unlikely to come from the dormant rooting zone, especially considering that our soil ^{14}C-CO_2 pore-space data show old C coming mainly below 25 cm (Figure 7). Consequently, due to the lack of data and of a total scientific consensus, we refrain from confirming that this could be a plausible explanation for our ^{14}C-R_{eco} data.

At our site during winter time, the source of this older C is clearly from deeper in the soil profile. Currently, the wintertime R_{eco} has an average age of 879 ± 116 years BP (Table 1). This age is much older compared to previous R_{eco} measurements (postbomb C; $>$1950) from a birch tundra site in Abisko, Sweden (~110 pM; Hartley et al., 2013), a spruce forest in Norway (~86–89‰; Muhr et al., 2009), or from another graminoid tundra snow-pack manipulation site in Alaska (~38‰; Natali et al., 2011).

This R_{eco} flux and underlying bulk soil C are old (preindustrial), but formed during the Holocene (6,800 to 3,000 years ago; Gilbert, 2014). We cannot exclude that preaged C deposited on top of the regolith or the underlying shale regolith which contains fossil C and can be remineralized by microbes (Petsch et al., 2001) and therefore may contribute to the old mean ages of soil CO_2 and R_{eco}. This flux may contribute a significant proportion to total ecosystem emissions in winter, when soil C decomposition fluxes are small. In the warmer soil underlying the deeper snow, C from deeper parts of the soil profile (which are older; Figure 7) are likely to contribute a larger proportion to R_{eco} than under ambient snow. This trend has been shown also during the growing season in another experimental snow-depth site in the High Arctic of Greenland (Lupascu, Welker, Xu, et al., 2014).

However, our pore-space ^{14}C-CO_2 results are not reflected in the ^{14}C-R_{eco} data, which shows no statistical difference between ambient and deep snow areas. This mismatch between soil CO_2 with the ^{14}C-R_{eco} could be due to different factors. One could be a result of the low number of ^{14}C samples analyzed in February and March, as we had to discard many due to the low CO_2 concentrations (which disallowed reliable ^{14}C analysis). Furthermore, measurements of R_{eco}, integrate the entire soil profile, including older C at depth, but are strongly biased toward the younger CO_2 produced by higher microbial activity in the rhizosphere (Lupascu, Welker, Seibt, Maseyk, et al., 2014). Consequently, while the belowground CO_2 might be older under the deep snow area (Figure 3), a stronger modern signal from the rooting/litter zone is able to overall dilute the ^{14}C signal of the R_{eco} masking the older CO_2 coming from deeper soils as shown in other studies (Hicks Pries et al., 2015; Lupascu, Welker, Seibt, Maseyk, et al., 2014). This could be the case especially for our late summer samples when plants are still active.

Our findings depict the complexity of how initial changes in winter climates, such as snowfall, may affect biosphere-atmosphere interactions in the High Arctic. While the proportional contribution of old C to wintertime R_{eco} is very high (close to 100%), the emission rates are where the initial sensitivities lie. As snow depth increases, it is possible that not the source but amount (mg m^{-2} season) of old C that is transferred from the soil to the atmosphere will dramatically change. While at our study site, microbes are decomposing a mixture of old highly degraded soil C and fossil C coming from shale deposits (Table 1), this process might be true also in areas where organic deposits are deeper. Furthermore, ages of respired CO_2 in winter might become progressively older as active layers deepen with current warming trends, as reported for Svalbard (Christiansen et al., 2010), other regions of the Scandinavia Arctic and sub-Arctic, and in Arctic Alaska, Canada, and Greenland (Schädel et al., 2014).

Another caveat in the interpretation of our results for the annual C budget of tundra systems lies in the duration of these high old C emissions periods. Areas with deeper snow in winter will have 2–3-week-longer winter season emissions (Borner et al., 2008; Lupascu, Welker, Xu, et al., 2014; Morgner et al., 2010; Pattison & Welker, 2014; Welker et al., 2000) before snowmelt compared to ambient conditions, and thus smaller rates of C uptake during the shorter growing season (Lupascu, Welker, Xu, et al., 2014).

4.3. Sedimentary Deposits and Carbon Sources

Our data raise further questions on how organic and mixed sedimentary deposits might behave with a changing climate. While microbes most often consume the most labile and therefore youngest C (Trumbore, 2009), previous studies have illustrated that in some settings microbes may use ancient C. For example, ancient sedimentary organic matter that is exposed through erosion can become vulnerable to microbial oxidation (Horan et al., 2017). Radiocarbon analyses of fatty acids in microbes cultured from newly exposed shale found that over 75% of the C the microbes incorporated into their cells was from the ^{14}C-free shale (Petsch et al., 2001). Additionally, microbes have been found to incorporate ancient depleted C in the phospholipid fatty acids of their cell membranes after petroleum hydrocarbons were introduced to rock pools (Ahad et al., 2010; Bostic et al., 2018; Mahmoudi et al., 2013). In depositional environments such as Svalbard and in regions of the Canadian High Arctic, it is possible that soil microbes are consuming a combination of modern and ancient organic matter. Simple C inventories that do not differentiate SOM and (very low quality) organic C-rich sedimentary rocks pools are likely to significantly overestimate the permafrost C feedback to climate change from the High Arctic. For example, our soil C data suggest a mixture of organic and shale C (Table 1 and Figure 5). Furthermore, it is important to consider that in addition to the pool of permafrost C that will be vulnerable in a warmer Arctic, there may be additional pools of unaccounted C, such as new deglaciated soils exposed by glacial retreat (Schmidt et al., 2008). The degradation of this sedimentary C pool could be offset if plant succession and soil organic aggradation was faster than sediment C mineralization. Weathering is typically assumed to be much slower than SOM aggradation; however, this might not be the case in marginal Arctic soils (Kabala & Zapart, 2012). Therefore, this study highlights the importance of considering the heterogenous nature of Arctic soils, and the source of old carbon that may be respired in a warmer Arctic.

5. Conclusions

Our results show that snow cover duration and depth play an important role in affecting soil temperatures and consequently the magnitude of R_{eco} in winter with higher values under deeper snow cover. Radiocarbon analysis of R_{eco} shows that late growing season emissions were driven by year- to decade-old

C (Δ^{14}C = 62 ± 8‰), while winter emissions, on the other hand, were dominated by Holocene soil and fossil shale (Δ^{14}C = −64 ± 14‰) with no significant differences between ambient and deeper snow cover. This study further highlights how weathering of C-rich parent material in Arctic gelisols soils may contribute to R_{eco} with significant proportions of fossil C. Consequently, future permafrost-C inventory and emission budgets should distinguish between modern and ancient sedimentary C sources.

Our work emphasizes that the shallow nature of the organic layer in High Arctic soils adds an additional level of complexity to permafrost C pool and emission estimates.

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