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RESEARCH ARTICLE

Key Points:
- Cumulative summertime R_{eco} was positively correlated to water-year precipitation
- Under increased snow, summertime Reco was reduced by up to 40%
- Soil CO_2 was older under increased snow

Rates and radiocarbon content of summer ecosystem respiration in response to long-term deeper snow in the High Arctic of NW Greenland

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Abstract The amount and timing of snow cover control the cycling of carbon (C), water, and energy in arctic ecosystems. The implications of changing snow cover for regional C budgets, biogeochemistry, hydrology, and albedo due to climate change are rudimentary, especially for the High Arctic. In a polar semidesert of NW Greenland, we used a ~10 year old snow manipulation experiment to quantify how deeper snow affects magnitude, seasonality, and ^{14}C content of summer C emissions. We monitored ecosystem respiration (R_{eco}), soil CO_2, and their ^{14}C contents over three summers in vegetated and bare areas. Additional snowpack, elevated soil water content (SWC), and temperature throughout the growing season in vegetated, but not in bare, areas. Daily R_{eco} was positively correlated to temperature, but negatively correlated to SWC; consequently, we found no effect of increased snow on daily flux. Cumulative summertime R_{eco} was not related to annual snowfall, but to water year precipitation (winter snow plus summer rain). Experimentally increased snowpack shortened the growing season length and reduced summertime R_{eco} up to 40%. Soil CO_2 was older under increased snow. However, we found no effect of snow depth on the R_{eco} age because older C emissions were masked by younger CO_2 produced from the litter layer or plant respiration. In the High Arctic, anticipated changes in precipitation regime associated with warming are a key uncertainty for understanding future C cycling. In polar semideserts, water year precipitation is an important driver of summertime R_{eco}. Permafrost C is vulnerable to changes in snowpack, with a deeper snowpack-promoting decomposition of older soil C.

1. Introduction

The High Arctic (>70°N) is undergoing rapid warming [*Overpeck et al.*, 1997] and "greening" [*Bhatt et al.*, 2010; *Epstein et al.*, 2013]. Here polar semideserts cover approximately 1×10^6 km^2 and, together with deserts and mires, hold an estimated 12 pg of organic carbon (C) in soils affected by near-surface permafrost [*Burnham and Sletten*, 2010]. These soil C pools are exceedingly heterogeneous in their vertical and surface traits [*Horwath et al.*, 2008], as differential frost heave and patterned ground result in extensive vertical mixing and a surface that is comprised of vegetated and bare areas [*Kessler and Werner*, 2003; *Walker et al.*, 2004]. Field-based studies investigating what controls C losses and sequestration in the complex soils of the High Arctic are important to fully understanding C dynamics in permafrost soils globally [*Welker et al.*, 2004].

A key driver of C cycling in arctic ecosystems is the onset, duration, and amount of snow cover. Arctic tundra is covered in snow for 8–10 months of the year, although spatial and interannual variability of snow cover is high [*Fahnestock et al.*, 1999; *Jones et al.*, 1999; *Callaghan et al.*, 2011]. Anticipated changes in precipitation, including the amount and timing of snow cover as a consequence of climate warming, and their implications for regional biogeochemistry (i.e., C cycling), albedo, and hydrology are highly uncertain [*Hinzman et al.*, 2005; *Steffen*, 2006]. Global climate models (GCM) projection in snow cover for 2050 indicates increases in maximum soil water equivalent (SWE) by up to 15% for most of the panarctic, with the greatest increase (15–30%) over Siberia. On the other hand, GCM project a decrease in snow cover duration (SCD) by ~10–20% for most of the panarctic by 2050, with the smallest decreases over Siberia (<10%) and the greatest decreases over northern Scandinavia and Alaska (30–40%) [*Callaghan et al.*, 2011]. Along with these projections, continuing expansion of erect shrub populations [*Epstein et al.*, 2013; *Pearson et al.*, 2013] will affect regional snowpack depth distribution, with higher snowpack where shrub density is greater [*Sturm et al.*, 2005]. Thus, understanding the consequences of changes in the depth and timing of snow cover is an important part of quantifying the functional traits of these landscapes in the future.

Due to its high albedo and low thermal conductivity, the depth and timing of snow cover control tundra water and energy budgets [*Pomeroy et al.*, 2006; *Adam et al.*, 2009], active layer thermal regime [*Lafrenière et al.*, 2013], and permafrost stability [*Lawrence et al.*, 2008]. Snow cover also influences ecosystem biogeochemistry, such as the cycling of C and nutrients. These effects are mediated via changes in (1) growing season length that in turn affects plant productivity, phenology, reproductive success, and tissue composition [*Galen and Stanton*, 1995; *Rieley et al.*, 1995; *Molau and Edlund*, 1996; *Welker et al.*, 2000, 2005; *Groendahl et al.*, 2007; *Borner et al.*, 2008; *Bhatt et al.*, 2010; *Cooper et al.*, 2011; *Olofsson et al.*, 2011; *Leffler and Welker*, 2013; *Epstein et al.*, 2013] and (2) soil microbial activity [*Schimel et al.*, 2004; *Sturm et al.*, 2005; *Borner et al.*, 2008]. Associated changes in winter nutrient cycling can have subsequent effects on annual C budgets [*Welker et al.*, 2000], culminating in shifts in tundra vegetation composition and canopy structure [*Sturm et al.*, 2005; *Wahren et al.*, 2005; *Tape et al.*, 2006; *Forbes et al.*, 2010], which, in turn, feedback to how snow accumulation is distributed across the landscape [*Sturm et al.*, 2005].

Several studies have explored the effects of increased winter snow depth on soil and ecosystem C cycling. Deeper snowpack typically results in greater wintertime soil temperatures [*Jones et al.*, 1998; *Walker et al.*, 1999; *Schimel et al.*, 2004; *Welker et al.*, 2004; *Hinkel and Hurd*, 2006] and ecosystem respiration (R_{eco}) [*Brooks et al.*, 1997, 1998; *Walker et al.*, 1999; *Welker et al.*, 2000; *Schimel et al.*, 2004; *Morgner et al.*, 2010; *Nobrega and Grogan*, 2013]. However, in both the High and Low Arctic effects on summertime C cycling are unclear, with some studies exhibiting an increase in R_{eco} under experimentally increased snowpack [*Natali et al.*, 2011; *Rogers et al.*, 2011], a decrease [*Jones et al.*, 1998; *Welker et al.*, 2000; *Björkman et al.*, 2010] or no effect [*Björkman et al.*, 2010; *Morgner et al.*, 2010]. Furthermore, only two studies, both carried out in Alaska [*Nowinski et al.*, 2010; *Natali et al.*, 2011], have investigated the effects of winter snow manipulation on soil C storage and turnover.

Here we investigated the effects of long-term (10 years) increases in winter snowpack on the magnitude, seasonality, and sources of summertime R_{eco}. We monitored the rate of R_{eco} and the concentration of CO_2 in the soil pore space along with their radiocarbon (^{14}C) content; nondestructive techniques used to infer changes in ecosystem and soil C sequestration or loss [*Trumbore*, 2006, 2009]. All living organisms are labeled with ^{14}C, which is produced in the atmosphere, oxidized to CO_2, and enters the food chain via photosynthesis. Testing of thermonuclear bombs aboveground in the midtwentieth century increased the amount of atmospheric $^{14}CO_2$ above natural production levels. The amount of this bomb ^{14}C in the atmosphere is declining due to mixing with terrestrial and ocean C pools and dilution by fossil (^{14}C-free) CO_2 [*Levin et al.*, 1980]. Consequentially, the ^{14}C content of recent photosynthetic products as well as that of plant and rhizosphere respiration is similar to the ^{14}C content of current atmospheric CO_2. The respiration of soil microbes decomposing organic matter made from photosynthetic products years to decades ago is enriched in ^{14}C. Furthermore, microbial respiration decomposing hundreds to thousands or years old organic matter is depleted in ^{14}C due to radioactive decay. Any increases in decomposition of old permafrost C would have important consequences for atmospheric CO_2. Decomposition of old C that was not part of the active C cycle for millennia results in an increase flux of C to the atmosphere while rapid cycling of young C between plants and microbes has a near-zero effect on the atmospheric C pool [*Trumbore*, 2009]. Specifically, with this study we asked what are the effects of increasing winter snowpack on (1) the concentration and ^{14}C content of CO_2 within the active layer and (2) on the magnitude, seasonality, and ^{14}C content of R_{eco} during the summer.

2. Material and Methods

2.1. Site Description

The study was carried out in prostrate dwarf-shrub tundra in the High Arctic of northwest Greenland near Thule Air Base (76°32′N, 68°50′W; 200–350 m above sea level) [*Sullivan et al.*, 2008; *Sharp et al.*, 2013]. Mean annual air temperature is −11.3 ± 1.3°C, with 122.6 ± 45.4 mm of mean annual precipitation (1952–2012). During the study period (2010, 2011, and 2012) mean summer temperature was 5.6 ± 2.0°C, precipitation was 24.0 ± 18 mm, and mean winter snowfall 1306 ± 200 mm (data from Thule (THU) airport). During the last climate normal period (1983–2012), air temperature shows an increasing trend of 1.03°C/decade with the strongest warming during the winter months (December, January, and February) of 1.84°C/decade [*Lupascu et al.*, 2014b].

The vascular plant community at our site is dominated by the deciduous dwarf-shrub *Salix arctica* PALL., the graminoid *Carex rupestris* ALL., and the wintergreen dwarf-shrub *Dryas integrifolia* VAHL. The live biomass and

litter of these three species account for approximately 70% of vascular plant cover. The soil is a Typic Haploturbel [*U.S. Department of Agriculture*, 1999] with a thin, patchy litter layer, high-gravel content, and, based on soil pits near the experimental area, a maximum thaw depth of about 1.0 ± 0.2 m [*Sullivan et al.*, 2008, *Horwath*, 2007]. Soils are well drained and subject to cryoturbation. In vegetated areas, the topsoil (0–12 cm) texture is 67–74% sand, 20–34% silt, and 5–8% clay, while in bare areas the texture is 54–64% sand, 33–38% silt, and 3–7% clay [*Sullivan et al.,* 2008]. For the top 40 cm, soil bulk density is approximately $1.10 \, \mathrm{g \, cm^{-3}}$ and the organic soil C content is 0.2–1.6% for vegetated areas and 0.1–0.2% for bare areas.

2.2. Experimental Set Up

Measurements were conducted at a long-term snowpack manipulation experiment consisting of two 1.2 tall and 6.0 m long snow fences erected in the summer of 2003 [*Rogers et al.*, 2011]. Fences were aligned perpendicular to the dominant winter wind direction, and snow naturally accumulates on the leeward side in a continuously tapering drift. At each fence, $2.0 \times 0.8 \, \mathrm{m^2}$ plots were delimited within three larger, preestablished main plots at each of three levels of winter snowpack: ambient (control, ~0.25 m), intermediate ((Snowfence medium) SFM, ~0.55 m), and deep ((Snowfence high) SFH, ~1.1 m) [*Rogers et al.*, 2011]. Ambient plots used in this study are the same control plots used in the factorial warming × irrigation experiment described by *Lupascu et al.* [2014a, 2014b] and *Sullivan et al.* [2008]. Plots were oriented to span the transition between vascular plants and bare soil/cryptogamic crust, such that each comprised approximately 50%, to facilitate scaling from the plot to ecosystem level. Thus, each plot had two subplots representing the vegetated versus bare areas (Figure S1 in the supporting information). Direct measurements of snow depth were not made as part of this study, as the sites could not be accessed during winter. Measurements were taken during the snow-free period, from when it was possible to insert chamber bases and conduct R_{eco} measurements without compacting snow (from May/ early June under ambient conditions to 20 August in 2010, 2011, and 2012).

2.3. Weather Data

Weather data were calculated based on temperature, precipitation, and snowfall data from the Thule airport (THU) weather station for the period 2009–2012. Daily mean temperatures are calculated as the mean of the daily minimum and maximum temperatures. Mean summer temperatures represent the average of the daily mean temperature for the June–August period. Cumulative precipitation represents the sum of the daily precipitation (rain, mm) for the summer (June, July, and August). Cumulative snowfall is the sum of the daily snowfall from September to May.

Water year precipitation, a concept commonly used in hydrology to describe bioavailable precipitation in (semi-) arid regions, is the sum of precipitation and snowfall (expressed as SWE). We calculated SWE as the average of the cumulative snowfall times its minimum ($252 \, \mathrm{kg \, m^{-3}}$) or maximum ($354 \, \mathrm{kg \, m^{-3}}$) density, using density values reported for high flat areas in the High Arctic [*Woo et al.*, 1983].

2.4. Ecosystem Respiration and Soil Pore Space CO_2 Concentrations

Ecosystem respiration and pore space CO_2 concentrations were measured once per day during 9 A.M. to 12 noon, 2 to 3 times per week. Ecosystem respiration was determined through the use of opaque chambers (30 cm internal diameter; 8 L volume). In June, collars were inserted to ~ 2 cm depth, sealed with soil material on the outside, and left in place for the sampling season. To calculate $R_{\mathrm{eco,}}$ air was circulated between the chamber's headspace connected to an infrared gas analyzer and a data logger (LI-840, LI-1400, and LI-COR Biosciences, Lincoln, NE, USA) at a rate of $0.5 \, \mathrm{L \, min^{-1}}$. Flux rates were estimated from the slope of time versus CO_2 concentration curves using linear regression. In parallel, we measured soil temperature (15–077, Fisher Scientific, resolution ±0.1°C) at 5 and 10 cm depth below the soil surface and soil water content (SWC, Hydrosense, Campbell Scientific, Logan, UT, USA, resolution ±0.1%) at 5 cm depth. Direct measurements of active layer depth could not be made due to the small plot size and high-gravel content. However, we estimated thaw progression using our gas wells (see below), as soil gas cannot be retrieved from frozen soils. Rarely, soil gas could not be retrieved from wells, because soils were temporarily water logged during snowmelt or following rain storms.

Carbon dioxide concentrations in the soil pore space were analyzed via stainless steel gas wells (0.35 cm ID, 0.6 cm OD) from 20, 30, 60, or 90 cm soil depth. Wells were inserted in 2010 and 2012, capped with rubber septa (Blue septa, Grace, Deerfield, IL, USA), and left in the ground during the entire study period including

the winter ($n = 1-2$). Gas samples were taken with a 60 mL syringe (BD, Franklin Lakes, NJ, USA) and injected (first sample extraction was always discarded) into an infrared gas analyzer connected to a data logger (LI-800, LI-1400, and LI-COR). Concentrations were recorded manually.

2.5. Estimating Summer Ecosystem Respiration Fluxes

To estimate summer R_{eco}, we used the relationship between respiration and temperature [*Lloyd and Taylor*, 1994], represented by an Arrhenius type equation (1), where the effective activation energy for respiration varies inversely with temperature.

$$R_{eco} = Ae^{\frac{-E_o}{(T-T_o)}} \qquad (1)$$

with (E_o) = 308.56 K and (T_o) = 227.13 K and where (A) is a data set-dependent variable. (A) was first obtained using the data collected in situ, and then daily R_{eco} for the missing days was calculated using the average daily air temperature (T). Daily R_{eco} for the growing season (snow-free period) was then summed up to yield cumulative summertime R_{eco}.

2.6. Gas Sampling for Isotope Analyses

To sample R_{eco} for ^{14}C analysis, chambers were left closed until the CO_2 concentration inside the chamber was about twice that in ambient air (up to 48 h). 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^{-1} for 15 min [*Gaudinski et al.*, 2000]. During each sampling event, we also collected CO_2 in ambient air on molecular sieve traps.

Pore space CO_2 was collected in 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]. During each sampling event, we also collected CO_2 in ambient air on molecular sieve traps.

Carbon dioxide respired from roots in the control plot were sampled by manually extracting all roots from a core of 30 × 5 cm the same day the soil was sampled. Roots were rinsed with water and placed into a 1 L mason jar, which was flushed with CO_2-free air. After 24 h, the CO_2 produced was collected on a molecular sieve trap. To conserve the integrity of this long-term experiment, roots could only be retrieved from control areas outside the experimental plots.

2.7. Isotope Analyses of CO_2

Carbon dioxide was released from molecular sieve traps by baking at 650°C for 45 min or extracted from canisters using a vacuum line, purified cryogenically, and reduced to graphite via Zn reduction [*Xu et al.*, 2007]. A split of the CO_2 was analyzed for its $\delta^{13}C$ ratio (GasBench II, DeltaPlus, Thermo Scientific). The ^{14}C content of the graphite was measured with accelerator mass spectrometry (NEC 0.5MV 1.5SDH-2 accelerator mass spectrometry (AMS)) at the Keck Carbon Cycle (KCC) AMS laboratory of the University of California (UC) Irvine [*Southon and Santos*, 2007]. Data are reported relative to National Institute of Standards and Technology OX-I (SRM 4990a) and OX-II (SRM 4990c) [*Stuiver and Polach*, 1977]. The measurement uncertainty for $\Delta^{14}C$ was < 2 per mil.

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

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

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 from the CO_2 concentrations inside the chamber immediately before trapping and in ambient air.

2.8. Statistical Analyses

Control plots had a sampling size of $n = 2$, while we increased the sampling size of treatment plots from one in 2010 to two in 2011 to four in 2012. Our sample for vegetated areas consisted of 662 observations for R_{eco},
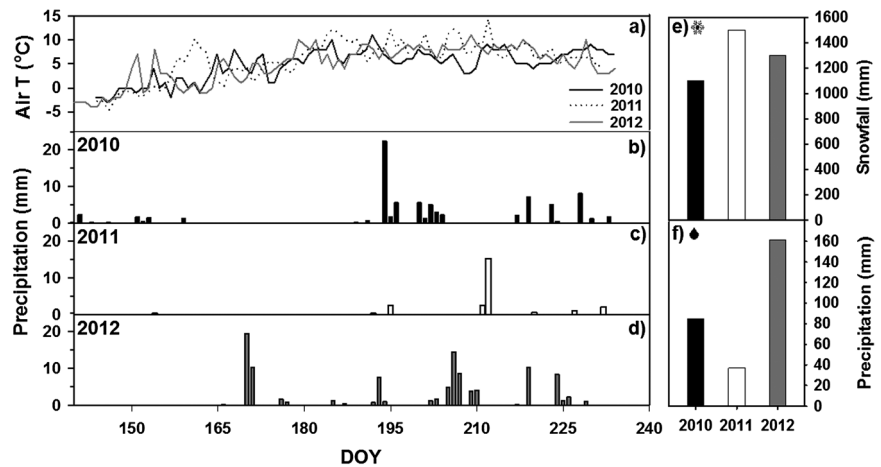
Figure 1. (a) Summer daily mean air temperature and (b–d) daily precipitation (rain) patterns for Thule, Greenland, (Thule airport, THU) for 2010–2012. (e) Cumulative snowfall for the preceding winter (2010 = winter 2009/2010) and (f) cumulative summer precipitation for 2010–2012.

with the control plot as experimental unit. We conducted a two-way analysis of variance (ANOVA) to examine the fixed effects of measurement year (2010, 2011, and 2012), treatment type (control, SFM, and SFH) and the interaction of the two on R_{eco}. The same analysis was applied for SWC (427 observations), soil temperature (393 observations). Compared to R_{eco} fluxes, in the ^{14}C analyses, the extra variable "plot type" (vegetated versus bare areas) was taken into account and 96 observations were used for the ANOVA.

3. Results

3.1. Weather Conditions

The three sampling years exhibited contrasting weather. Average summer air temperature was highest in 2011 (~6.1 ± 3.1°C), with the maximum difference observed in July (+1–2°C) (Figure 1a). Summertime temperature was similar in 2010 and 2012 (~5.3 ± 3.0°C; Figure 1a). In contrast, cumulative summer precipitation was highest in 2012 (161.3 mm) compared to 2010 (84.8 mm) or 2011 (36.7 mm) (Figure 1f). We also observed a striking difference in the timing of summer precipitation. In 2010 and 2011 (Figures 1b and 1c), more than half of summer rain fell during the month of July (61–73%) while June was very dry (1.1–6.5%). In 2012 (Figure 1d), precipitation was more evenly distributed, with 31, 46, and 22% of the total summer precipitation falling in June, July, and August, respectively. Cumulative snowfall was highest in winter 2010/2011, with +36% more snow compared to 2009/2010 and +15% compared to 2011/2012 (Figure 1e). If we consider the water year precipitation, 2012 was still the wettest year with 557 ± 67 mm, followed by 2011 (492 ± 77 mm) and 2010 (410 ± 56 mm).

3.2. Soil Water Content and Temperature

Surface SWC in vegetated areas (Figure 2, middle) was strongly influenced by interannual difference in snow cover, summer precipitation patterns, and experimental treatment. The water year precipitation affected SWC throughout the growing season in all years with SWC being greater in 2012 than in 2011 or 2010. Soil water content increased during spring due to snowmelt and again in July and August due to rainfall. We observed the highest average summer SWC in 2012 (31.8 ± 3.7 vol %), compared to 2010 and 2011 (15.1 ± 4.4 and 17.2 ± 4.4 vol %, respectively) (year difference, $F = 36.88$, $p < 0.001$). Higher experimental snowpack level increased average summer SWC by +4.1 ± 0.5 (SFM) and +5.2 ± 1.9 vol % (SFH) relative to the ambient snow conditions but did not alter the seasonal pattern ($F = 9.90$, $p < 0.002$). Soil water content in the bare areas did not show any statistical difference between deep and ambient snow areas (data not shown).

Surface soil temperature (Figure 2, top) reflected the seasonal change in air temperature, with maximum values in July, as previously observed [*Czimczik and Welker*, 2010]. In vegetated areas, experimentally increased snowpack affected soil temperatures during all summers. Compared to the control (mean 8.6 ± 0.3°C), experimental plots (SFM and SFH) were +0.9 ± 0.2°C warmer ($F = 7.055$, $p < 0.024$). No statistical difference
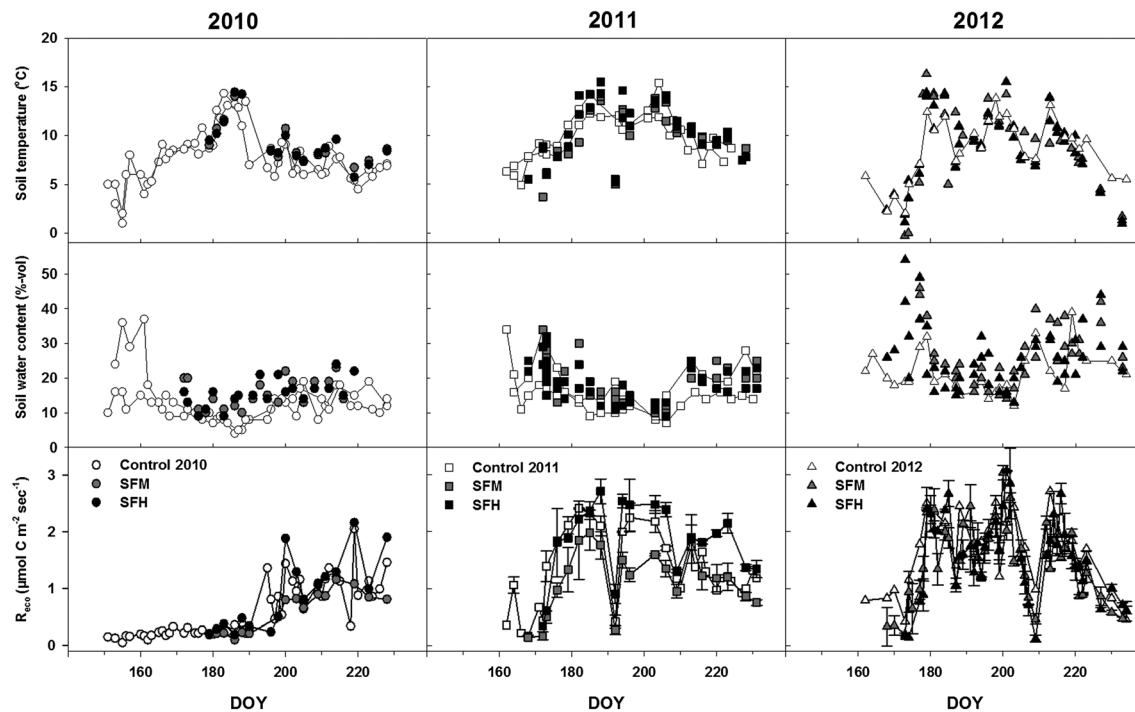
Figure 2. (top) Soil temperature and (middle) water content at 5 cm and (bottom) average daily ecosystem respiration (R_{eco}) for ambient (control), medium (SFM), and high (SFH) snowpack conditions in vegetated areas for the 2010–2012 summer period.

between the two snowpack treatments was found. On the other hand, a distinct difference between the years ($F = 6.112$, $p < 0.002$) was observed, with soil temperatures being the warmest in 2011 (mean $10.0 \pm 0.3\,°C$), the year with the most snow.

3.3. Growing Season Length

The length of the growing season (i.e., snow-free period) varied between years due to the different snowpack level, with more snowfall resulting in a shorter growing season. The growing season was longest in 2010 with 81 days in the control compared to 71 and 72 days in 2011 and 2012, respectively. In the treatments, growing season length was shorter than in the control, starting about 6 days later in SFM (medium snowpack plots) and about 11 days later in SHM (high snowpack plots) (Table 1).

3.4. Ecosystem Respiration

Summer ecosystem respiration was noticeably influenced by the seasonal variability in precipitation and temperature, experimental snowpack level, and water year precipitation. On a daily scale, all plots and treatments showed a weak negative correlation between daily R_{eco} and SWC ($R^2 = 0.22 \pm 0.19$), but a robust positive correlation with soil temperature ($R^2 = 0.56 \pm 0.19$). Furthermore, R_{eco} diminished considerably during cold spells, which typically concurred with rainfall events.

We observed two different seasonal patterns of R_{eco} fluxes. In 2011 and 2012, R_{eco} fluxes in the vegetated areas showed a sharp peak during snowmelt, and then followed the seasonal changes in air temperature, with a maximum in July (Figures 2 (bottom) and 1a). Midsummer fluxes were of similar magnitude in 2011 and 2012 (control, mean ± standard deviation: $2.36 \pm 0.18\,\mu mol\,C\,m^{-2}\,s^{-1}$). In 2010, we did not observe a snowmelt peak as our measurements commenced a couple of weeks after snowmelt (DOY 151), and R_{eco} fluxes could be divided into two distinct periods. Ecosystem respiration was low and stable (control: $0.28 \pm 0.11\,\mu mol\,C\,m^{-2}\,s^{-1}$) from early June (DOY 152) to mid-July (DOY 195), and greater from mid-July to the end of August (control: $0.60 \pm 0.29\,\mu mol\,C\,m^{-2}\,s^{-1}$; Figure 1b). In bare areas, R_{eco} fluxes were lower than in vegetated areas ($<1\,\mu mol\,C\,m^{-2}\,s^{-1}$), but seasonal trends were similar (Figure S2).

Table 1. Ecosystem Respiration (R_{eco}) Fluxes From Vegetated and Bare Areas and From the Landscape (Vegetated + Bare) During Three Sampling Years (2010–2012)

Months	Daily Flux (ave. ± SE, $n = 1$–4)			Days[a]	Monthly Flux		
	Vegetated	Bare	Landscape		Vegetated	Bare	Landscape
	($g\,C\,m^{-2}\,d^{-1}$)				($g\,C\,m^{-2}\,d^{-1}$)		
	June 2010						
Control	0.24 ± 0.00	0.04 ± 0.03	0.28 ± 0.00	30	7.3 ± 0.1	1.1 ± 0.7	8.4 ± 0.7
SFM	0.23 ± na	0.05 ± na	0.29 ± na	22	5.3 ± na	1.4 ± na	6.7 ± na
SFH	0.12 ± na	0.07 ± na	0.27 ± na	17	2.1 ± na	1.2 ± na	3.3 ± na
	July 2010						
Control	0.53 ± 0.01	0.07 ± 0.05	0.60 ± 0.00	31	16.5 ± 0.2	2.0 ± 1.4	18.5 ± 1.4
SFM	0.40 ± na	0.10 ± na	0.50 ± na	31	12.5 ± na	3.0 ± na	15.5 ± na
SFH	0.36 ± na	0.11 ± na	0.48 ± na	31	11.3 ± na	3.6 ± na	14.9 ± na
	August 2010						
Control	0.61 ± 0.02	0.09 ± 0.06	0.69 ± 0.00	20	12.1 ± 0.5	1.8 ± 1.2	13.9 ± 1.3
SFM	0.47 ± na	0.12 ± na	0.59 ± na	20	9.3 ± na	2.5 ± na	11.8 ± na
SFH	0.42 ± na	0.13 ± na	0.54 ± na	20	8.3 ± na	2.5 ± na	10.8 ± na
	June 2011						
Control	0.86 ± 0.12	0.12 ± 0.07	0.98 ± 0.14	20	17.2 ± 2.3	2.4 ± 1.4	19.6 ± 2.7
SFM	0.61 ± 0.07	0.14 ± 0.01	0.76 ± 0.07	14	8.6 ± 1.0	2.0 ± 0.1	10.6 ± 1.0
SFH	1.14 ± 0.03	0.15 ± 0.01	1.07 ± 0.04	9	10.2 ± 0.5	1.4 ± 0.2	11.6 ± 0.5
	July 2011						
Control	1.43 ± 0.09	0.22 ± 0.13	1.65 ± 0.16	31	44.4 ± 2.7	6.9 ± 4.1	51.3 ± 4.9
SFM	1.11 ± 0.17	0.26 ± 0.00	1.36 ± 0.17	31	34.4 ± 5.3	7.9 ± 0.1	42.3 ± 5.3
SFH	1.71 ± 0.04	0.22 ± 0.02	1.93 ± 0.04	31	53.0 ± 1.2	6.8 ± 0.5	59.8 ± 1.7
	August 2011						
Control	1.42 ± 0.11	0.22 ± 0.13	1.64 ± 0.17	20	28.4 ± 2.2	4.4 ± 2.5	32.8 ± 3.3
SFM	0.85 ± 0.10	0.20 ± 0.01	1.04 ± 0.07	20	16.9 ± 2.1	3.9 ± 0.2	20.8 ± 2.1
SFH	1.29 ± 0.01	0.16 ± 0.01	1.46 ± 0.01	20	25.9 ± 0.2	3.3 ± 0.1	29.2 ± 0.2
	June 2012						
Control	1.43 ± na	0.33 ± na	1.76 ± na	21	30.1 ± na	7.0 ± na	37.1 ± na
SFM	0.97 ± 0.06	0.23 ± 0.11	1.20 ± 0.06	16	15.5 ± 0.9	3.7 ± 1.8	19.2 ± 2.0
SFH	1.08 ± 0.05	0.18 ± 0.05	1.26 ± 0.04	11	11.8 ± 0.8	2.0 ± 0.7	13.8 ± 1.1
	July 2012						
Control	1.99 ± na	0.48 ± na	2.46 ± na	31	61.6 ± na	14.8 ± na	76.4 ± na
SFM	1.36 ± 0.08	0.34 ± 0.17	1.70 ± 0.18	31	42.2 ± 2.4	10.4 ± 5.2	52.6 ± 5.7
SFH	1.41 ± 0.09	0.20 ± 0.07	1.61 ± 0.12	31	43.6 ± 2.9	6.3 ± 2.1	49.9 ± 3.6
	August 2012						
Control	1.74 ± na	0.41 ± na	2.15 ± na	20	34.7 ± na	8.2 ± na	42.9 ± na
SFM	1.08 ± 0.04	0.26 ± 0.12	1.34 ± 0.08	20	21.6 ± 0.8	5.2 ± 2.4	26.8 ± 2.5
SFH	1.20 ± 0.09	0.16 ± 0.04	1.37 ± 0.07	20	24.1 ± 1.9	3.3 ± 0.9	27.4 ± 2.1

[a]Actual days of study period representing the growing season length; na = data not available ($n = 1$).

In all treatments, cumulative summertime R_{eco} showed a positive trend with total water year precipitation, but not with winter snowfall. Summertime R_{eco} was lowest in 2010 (Figure 3)—the year with the lowest water year precipitation. Additional water year precipitation in 2011 and 2012 stimulated summer R_{eco} (Figure 3; year difference in control, $F = 74.17$, and $p < 0.001$). However, this increase in summer R_{eco} was muted in the snowpack manipulations.

In the snowpack manipulations, daily R_{eco} fluxes in vegetated areas were similar to those in the control (Figure 2, bottom). However, due to a shorter snow-free period, cumulative summertime R_{eco} fluxes in the SFM and SFH treatments were reduced by −29.8 ± 6.1 and −22.6 ± 27.2%, respectively, compared to the control ($F = 16.42$, $p < 0.001$; Figure 3). In bare areas, however, daily R_{eco} fluxes in the SFM and SFH treatments were greater or similar to (Table 1) the control R_{eco} in 2010 and 2011, but lower in 2012. We found no statistical differences in the cumulative summertime R_{eco} fluxes between the two snowpack treatments.

3.5. Soil Pore Space CO_2

Pore space CO_2 concentrations reflect both production at the measured depth and diffusion within the soil. Concentrations were strongly affected by water year precipitation, seasonal changes in active layer depth, and
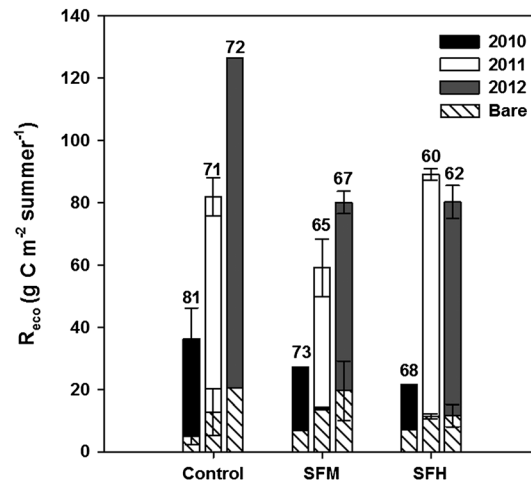
Figure 3. Cumulative summertime ecosystem respiration (R_{eco}) from High Arctic tundra during 2010–2012 under ambient (control), medium (SFM), and high (SFH) snowpack conditions (average ± SE, $n = 1–4$) (The number on the top of each histogram represents the growing season length calculated from the span of the actual measurement period.).

snowpack manipulation (Figure 4). Vegetated areas showed overall greater CO_2 concentrations compared to the bare areas (unpaired t test, $p < 0.005$), especially in the surface soil. Typically, CO_2 concentrations peaked between 30 and 60 cm, followed by a decrease toward the permafrost table (Figure S3).

During the summer, CO_2 concentrations in the surface soils exhibited a bimodal trend: A first peak after snowmelt (up to 18000 ppm in vegetated areas), and a second peak usually coupled with maximum air temperatures and active layer depth (Figure 4). The magnitude of the snowmelt peak between years was positively related with snowpack, with 2010 being the lowest (Figure 1e). However, interannual variability in the concentration of pore space CO_2 was smaller than that of R_{eco}. The timing of the snowmelt and the associated CO_2 peak were delayed in years with greater snowfall and in the snowpack treatments (Figure 4).

Snowpack manipulation significantly increased CO_2 concentrations at all depths compared to the control (Figure 4). For example, in 2012 seasonally averaged CO_2 concentrations in the SFM treatment were 4037 ± 903, 4660 ± 628, and 4418 ± 732 ppm at 20, 60, and 90 cm versus 2150 ± 187, 3500 ± 299, and 3481 ± 432 ppm in the control (seasonal average ± SE; $p < 0.002$). In 2010 and 2011, we found no difference in
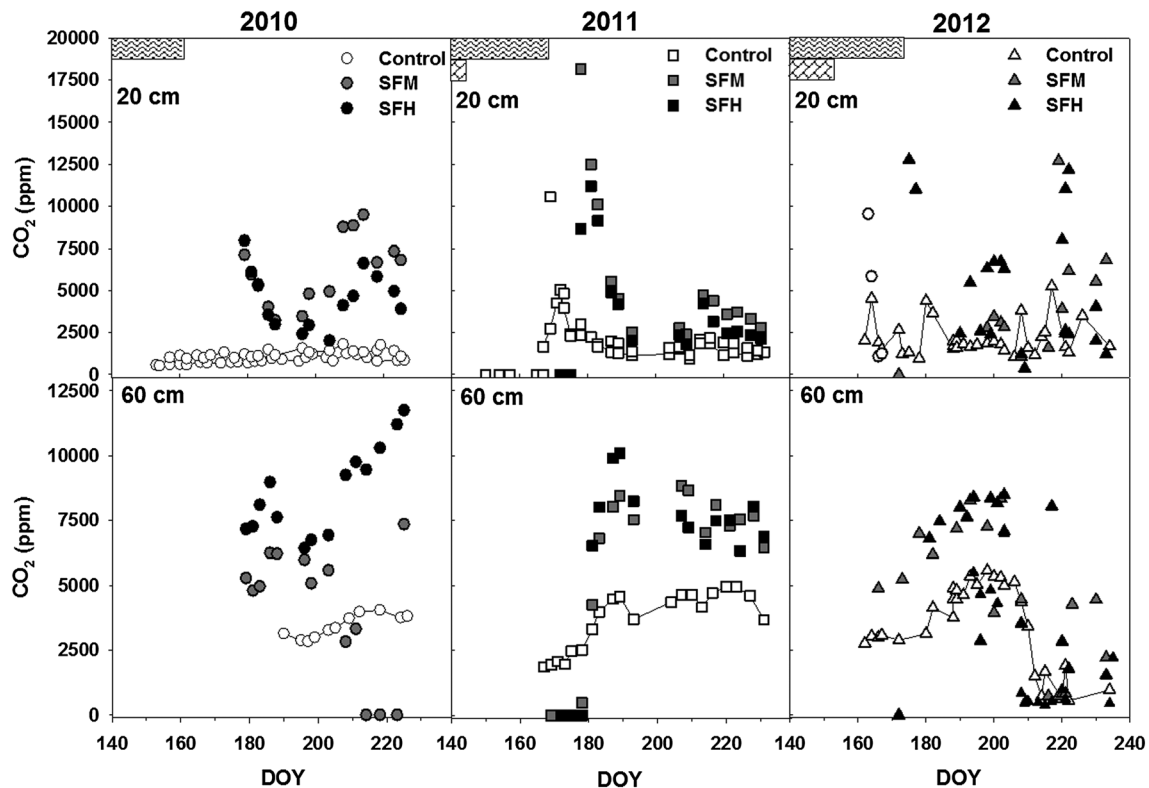

Figure 4. Pore space CO_2 concentration at 20 and 60 cm depth in vegetated areas under ambient (control), medium (SFM), and high (SFH) snowpack conditions during the summer of 2010, 2011, and 2012 ($n = 1–2$ per treatment). Brick- and zigzag-patterned histograms represent the presence of snow at the time of measurements in ambient and snow fence plots, respectively. (Note difference in scale between concentrations at different depths.)
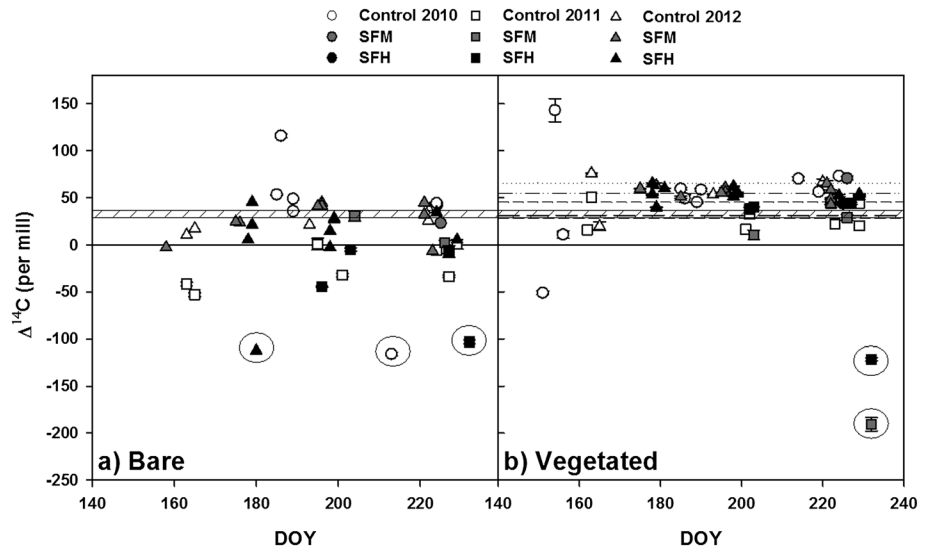
Figure 5. Radiocarbon content of ecosystem respiration ($R_{eco} \pm$ analytical error, $n = 1$) from (a) bare and (b) vegetated areas under ambient (control), medium (SFM), and high (SFH) snowpack conditions during the summers of 2010–2012 (Coarse area indicate the range of $^{14}CO_2$ in the atmosphere for 2010–2012; dotted line represent the litter layer; short-dash line, dash-dot-dot line, and long-dash line represent root incubation from the control plot in 2010, 2011, and 2012, respectively; circled symbols represent old C pulses concurring with small rain event (<4 mm)).

the CO_2 concentration between the two snow depth manipulations. In 2012, CO_2 concentrations were highest in the SFH treatment.

3.6. Radiocarbon Content of Ecosystem Respiration

The ^{14}C content of R_{eco} was significantly different between vegetated and bare areas and interannually but showed only negligible differences seasonally or between control and experimental snow depth manipulations. In vegetated areas, R_{eco} was significantly enriched (younger) in ^{14}C compared to R_{eco} in bare areas in both the control and snowpack manipulations ($F = 1.16$, $p < 0.007$; Figure 5). For example, the average summertime ^{14}C values of R_{eco} in the SFH treatment in 2010–2012 were 43.8 ± 2.0, 30.7 ± 1.7, and 54.2 ± 2.2 per mil, respectively, in vegetated areas, and 23.2 ± 0.7, -18.8 ± 12.8, and $18.8.6 \pm 5.4$ per mil, respectively, in bare areas.

Variability in weather conditions between the three sampling years significantly affected the mean age of R_{eco} ($F = 16.63$, $p < 0.001$), with the most ^{14}C depleted (oldest) R_{eco} observed in the warmest year 2011 for both vegetated and bare areas (Figure 5). For example, the average summertime ^{14}C value of R_{eco} in vegetated areas of the SFM treatment was 19.9 ± 9.2 per mil in 2011, compared to 7.6 ± 2.0 per mil in 2010 and 55.9 ± 2.2 per mil in 2012.

There was no seasonal trend in the ^{14}C content of R_{eco} in any year (Figure 5). Sporadically, however, we detected very depleted ^{14}C values (up to -190.8 ± 7.1 per mil) indicative of old R_{eco}, which appeared to be coupled to rain pulses in July and August [*Lupascu et al.*, 2014b]. We found no statistical difference between the ^{14}C values of R_{eco} across the three snow treatment conditions ($F = 2.03$, $p < 0.158$).

3.7. Radiocarbon Content of Soil Pore Space CO_2

The ^{14}C content of CO_2 in soil gas, sampled below the rooting zone, provides insight to changes in microbial activity and soil physical properties [*Phillips et al.*, 2013; *Torn et al.*, 2013]. The ^{14}C values of pore space CO_2 exhibited two different depth trends (Figure 6): (1) An increase in age with depth or (2) an increase to about 60 cm followed by a decrease toward the permafrost table. Overall, the ^{14}C values of pore space CO_2 were older in bare than in vegetated areas (unpaired t test, $p < 0.007$).

Experimental increases in snow affected the ^{14}C content of pore space CO_2 throughout the soil profile, with deeper snow resulting in older C at all depths. For example, the summertime average ^{14}C content of soil CO_2 at 60 cm in bare areas was -14.0 ± 11.4 and -67.7 ± 11.5 per mil for SFM and SFH, respectively, compared to
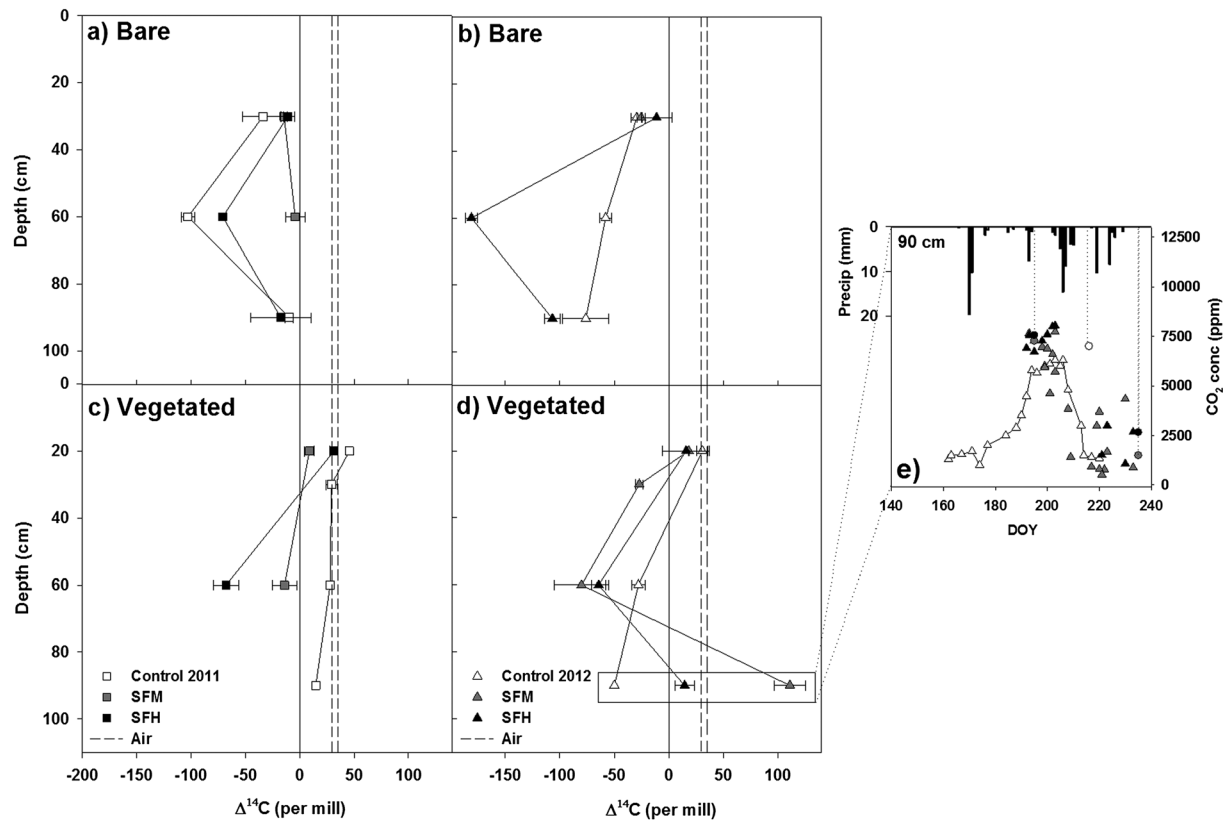
Figure 6. Radiocarbon content of pore space CO_2 along the soil profile under ambient (control), medium (SFM), and high (SFH) snowpack conditions in (a, b) bare and (c, d) vegetated areas during 2011 and 2012 (average ± SE, $n = 1–3$) (Dotted lines indicate the range of $^{14}CO_2$ in ambient air.). (e) Concentration of pore space CO_2 at 90 cm and precipitation during the summer of 2012. (Circle symbols with dotted lines represent the timing of ^{14}C samples for each treatment.)

27.9 per mil in the control. Two exceptions to this pattern were observed: pore space CO_2 in the control was older than in the treatments in bare areas in 2011 (Figure 6a), and in vegetated areas at 90 cm in 2012 (Figures 6d and 6e). For the latter event, sampling of CO_2 from the snowpack manipulations, but not from the control, coincided with intense rain events (>8 mm). Within a given snow depth treatment, we found no significant seasonal or interannual differences in the ^{14}C values (ages) of pore space CO_2.

4. Discussion

4.1. Effects of Snowpack on SWC and Soil Temperature

In vegetated areas we found that greater winter snowfall and experimentally increased snow depth resulted in greater summer SWC and soil temperatures throughout the growing season. While some studies have reported similar results to ours [*Chimner and Welker*, 2005; *Morgner et al.*, 2010; *Rogers et al.*, 2011], others did not find any increase in SWC over the summer [*Buckeridge and Grogan*, 2010]. This is not surprising as soil (porosity and water-holding capacity) and vegetation type can play a crucial role in affecting SWC [*De Michele et al.*, 2008]. Furthermore, depending on the photosynthesis rates, plants can use more or less water [*Sullivan and Welker*, 2007] thus affecting SWC. In our study area, the prolonged higher SWC is likely due to greater water retention in the litter layer and rhizosphere as bare areas drained much faster.

It was previously shown elsewhere [*Welker et al.*, 2000; *Schimel et al.*, 2004; *Morgner et al.*, 2010] that deeper snowpack increases soil temperature during the winter, but summertime effects are more complex. While many studies showed lower summertime temperatures in areas with higher snowpack [*Walker et al.*, 1999; *Schimel et al.*, 2004; *Hinkel and Hurd*, 2006], others found no difference or warmer soil temperatures [*Jones et al.*, 1998; *Rogers et al.*, 2011]. These contrasting observations can be reconciled by differences in water content, vegetation cover, and increased litter fall [*Fahnestock et al.*, 2000]. As water has a higher heat capacity than air, soil with a greater SWC gain and loose heat more slowly than drier soils [*Al-Kayssi et al.*, 1990].
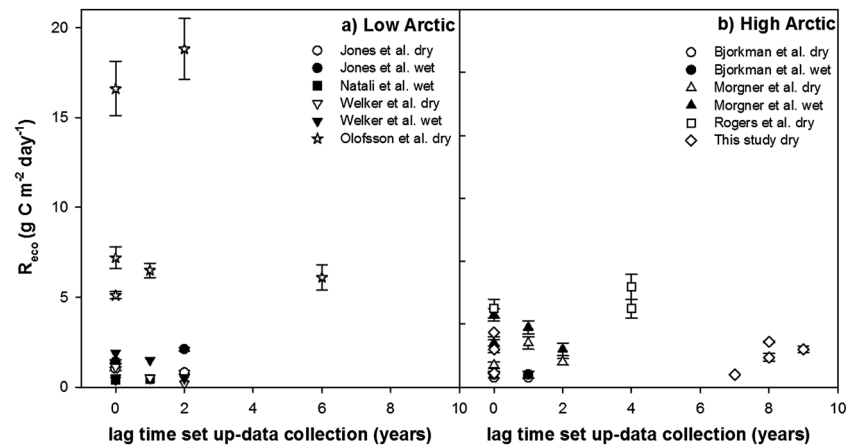
Figure 7. Synthesis of mean daily summer ecosystem respiration (R_{eco}) fluxes from wet (full symbols) and dry (open symbols) tundra in the (a) Low and (b) High Arctic under increased snowpack as a function of time since the experiment was established. Measurements from the same studies under ambient snowpack conditions are plotted at time zero. (Data from this study and *Björkman et al.* [2010], *Jones et al.* [1999], *Morgner et al.* [2010], *Natali et al.* [2011], *Olofsson et al.* [2011], *Rogers et al.* [2011], and *Welker et al.* [2000]. See supporting information for details (Table S1).)

In addition, soils and vegetation types with a better water retention capacity also minimize energy dissipation by latent heat flux, further increasing soil warming. Greater leaf litter deposition associated with deeper snow may further insulate the soil surface from the atmosphere [*Fahnestock et al.*, 2000].

4.2. Seasonality and Magnitude of Ecosystem Respiration

Interactions of air (and tightly coupled soil) temperature and precipitation (and hence SWC) are key drivers of summertime R_{eco} fluxes in polar semideserts, as recently demonstrated in a summertime precipitation and temperature manipulation experiment adjacent to the study reported here [*Lupascu et al.,* 2014a]. On a daily timescale, R_{eco} fluxes were driven by temperature and showed a peak in mid-July that coincided with maximum air temperatures and active layer depth. Mean daily summertime R_{eco} fluxes observed in this study under ambient snow are similar to those seen in other studies of High Arctic prostrate dwarf-shrub and herb tundra ecosystems [*Jones et al.*, 2000; *Lloyd*, 2001; *Welker et al.*, 2004; *Björkman et al.*, 2010; *Morgner et al.*, 2010; *Rogers et al.*, 2011]. On a seasonal timescale, the magnitude of total summertime R_{eco} was controlled by the amount of available water year precipitation. Thus, changes in the amount of both summertime rainfall and snow cover are critical uncertainties in understanding future C cycling in rapidly warming polar semideserts.

We found no effect of experimental increases in snow depth on daily R_{eco} fluxes. Similar responses were found in other dry tundra summer studies [*Welker et al.*, 2000; *Björkman et al.*, 2010] (Figure 7 and Table S1). However, our findings are in contrast to a previous study from our site [*Rogers et al.*, 2011], where higher snow depth enhanced R_{eco} fluxes in the SFH, but not the SFM treatment. The difference between the two studies may be related to differences in weather conditions or possibly to changes in the substrate availability for microbial degradation between 2007 and our measurement period. In 2007, the measurement period of *Rogers et al.* [2011], mean summer air temperature was 5.5 ± 3.8°C and thus similar to our study (5.3–6.1°C). However, water year precipitation was only 279 ± 196 mm and thus much less than in our study (410–557 mm). Therefore, plants or microbes in the control and SFM plots might have been affected by drought stress that was alleviated in the SFH treatment. In addition, since weather conditions are highly variable on a daily basis, difference between studies might derive from disparities in the number of measurements as *Rogers et al.* [2011] calculated R_{eco} from about 8 to 10 points compared to our almost 50 measurements.

The dominant effect of an experimentally increased snowpack was a shorter growing season. As a consequence, cumulative summer R_{eco} was smaller in both experimental snowpack manipulations (SFM and SFH) compared to control levels (with the exception of SFH in 2011; Figure 3). This confirms earlier summer studies in dry tundra systems [*Björkman et al.*, 2010, *Welker et al.*, 2000] where shorter plant growing seasons were associated with deeper snowpack [*Rieley et al.*, 1995; *Cooper et al.*, 2011; *Mallik et al.*, 2011; *Semenchuk et al.*, 2013].

The timing and amount of future snow is uncertain and greatly varies between regions [*Callaghan et al.,* 2011; *Liston and Hiemstra,* 2011]. Increases in erect shrub populations [*Epstein et al.,* 2013; *Pearson et al.,* 2013] will locally create areas with shorter growing season, since shrubs act as natural snow barriers that cause snow accumulation and delayed melt [*Sturm et al.,* 2005]. Whether changes in SCD and vegetation cover will have net positive or negative climate feedbacks, effects remain uncertain today [*Welker et al.,* 1997; *Anderson-Smith et al.,* 2014].

Our study also demonstrates the effectiveness of long-term monitoring studies for understanding C cycling in ecosystems with extreme interannual variability in weather such as polar semideserts. Compiling literature data from long-term snowpack manipulation studies in the High and Low Arctic showed supports our hypothesis: Interannual variability of mean daily R_{eco} within a given treatment was typically much larger than the treatment effect (Figure 7). Sustained, strong snow additions, however, significantly reduce the growing season length and may result in water logging and disease and vegetation collapse [*Hinkel and Hurd,* 2006; *Olofsson et al.,* 2011].

4.3. Soil CO$_2$ Concentrations at Different Depths

Greater CO_2 concentrations in the topsoil during the snowmelt period as opposed to later in the summer suggests that microbial activity was stimulated by the input of new water and/or C and nutrients from melting snow leaching through the litter layer and surface soil [*Hirano,* 2005; *Scott-Denton et al.,* 2006]. In addition, some of this CO_2 may originate from the release of previously produced CO_2 that had been trapped in the frozen active layer during the winter [*Albert and Perron,* 2000; *Schimel et al.,* 2006]. These short-term effects during the shoulder seasons may be especially important to quantify, while measurements during midsummer only would fail to unravel the complexity of these High Arctic systems [*Walker et al.,* 2008].

Long-term snowpack increases clearly affected belowground CO_2 concentrations. Increased snowpack resulted in greater CO_2 concentrations due to the additional melt water and greater SWC that persisted throughout the growing seasons compared to ambient conditions in all years. These results are consistent with a parallel study that found that weekly summer water additions increased belowground CO_2 concentrations [*Lupascu et al.,* 2014a].

4.4. Radiocarbon Content of Ecosystem and Soil Respiration

Analyses of the ^{14}C content of R_{eco} and pore space CO_2 provide a means to determine the sources of ecosystem C loss [e.g., *Czimczik and Welker,* 2010; *Nowinski et al.,* 2010]. Ecosystem respiration is a combination of CO_2 derived from three sources: (a) current plant metabolism (^{14}C ~30 per mil), (b) decomposition of recently fixed plant tissue (fixed since 1950: ^{14}C >30 per mil), and (c) decomposition of old soil organic matter (^{14}C <0 per mil). As a result of radioactive decay coupled with changes in the atmospheric ^{14}C abundance due to nuclear testing, each of these sources has a distinct ^{14}C value [*Trumbore,* 2006].

We found that both soil CO_2 and R_{eco} in vegetated areas were enriched (younger) in ^{14}C compared to those in bare areas, consistent with results from nearby concurrent or previous studies [*Czimczik and Welker,* 2010; *Lupascu et al.,* 2014a, 2014b]. This is likely due to proportionally higher plant and topsoil microbial respiration of CO_2 with a ^{14}C content similar or more enriched than current atmospheric CO_2 [*Czimczik et al.,* 2006; *Schuur and Trumbore,* 2006] that can mask older CO_2 diffusing from depth. Furthermore the bulk C pool in areas where plants are present is overall younger than where absent, due to the continuous input of fresh exudates and litter [*Horwath et al.,* 2008]. Differential frost heaving, which results in a ridge (bare)-trough (vegetated) patterning in the form of polar polygons or stripes [*Kessler and Werner,* 2003], thus dictates microtopographic variance in the ages of R_{eco} polar semideserts.

Interannual changes in air temperature and summer precipitation strongly affected the mean age of R_{eco}. We observed a significant release of older C from depth during the warmest summer (2011) compared to the colder and wetter years (2010–2012). This suggests a loss of deeper and older C pools by microbial decomposition, as shown in a related summer warming and wetting study from the same area [*Lupascu et al.,* 2014a]. During wetter summers (2010 and 2012), in contrast, R_{eco} in vegetated areas was enriched in ^{14}C compared to atmospheric CO_2, implying that R_{eco} originated predominantly from the respiration of microorganisms decomposing aboveground or belowground litter.

Soil CO_2 was depleted (older) in ^{14}C under experimentally increased snowpack in both vegetated and bare areas. Other studies from moist acidic tussock tundra in the Low Arctic found older C being emitted during summertime under increased snowpack manipulation in both R_{eco} [*Nowinski et al.*, 2010; *Natali et al.*, 2011] and soil pore space [*Natali et al.*, 2011]. Unfortunately, we were limited in our ability to collect samples to measure the ^{14}C content of root respiration and microbial respiration in litter and bulk soil C data from the SFM and SFH treatments. More destructive sampling is required to predict the contributions of old C decomposition to R_{eco} from soil CO_2 measurements.

Together, older CO_2 and higher CO_2 concentrations suggest greater decomposition of soil C under deeper snow and contribute to increasing atmospheric CO_2 concentrations and global climate change [*Czimczik and Welker*, 2010; *Schaefer et al.*, 2011; *Hicks-Pries et al.*, 2013; *Lupascu et al.*, 2014a, 2014b]. In our monthly observations, however, we found no effect of snowpack on the mean age of R_{eco}. This occurred in part, because R_{eco} integrates C dynamics within the entire active layer as well as the aboveground plant community. Emissions of old, deep C were likely masked by variable contributions from young, near-surface C that quickly respond to weather fluctuations. In addition, our inability to detect older C emissions from the experimental treatments may be due in part to our discontinuous sampling scheme.

We observed episodic release of older CO_2, irrespective of treatment. In June, some of the older values of R_{eco} are likely due to the release of CO_2 trapped over winter [*Schimel et al.*, 2006]. In July and August, these ^{14}C data further support episodic flushing of the active layer. The oldest values (most depleted in ^{14}C) we measured all coincided with small intensity ($<4\,mm$) precipitation events. Similarly, intense rain events ($>8\,mm$) coincided with observations of younger CO_2 at depth (Figure 6d), suggesting that larger precipitation events translocate modern C from the litter layer and/or rhizosphere to depth where it is decomposed. The decomposition of this younger C is likely masking potential decomposition of older, in situ C at these depths. These phenomena have been previously reported in an adjacent summertime climate manipulation study [*Lupascu et al.*, 2014a, 2014b].

5. Conclusions

Anticipated changes in the timing and amount of winter snowfall are uncertain, but our study revealed the complexity of how changes in snowfall regime can significantly affect C cycling in the High Arctic. Under experimentally increased snowpack, we found a decrease in cumulative summertime R_{eco} due to a reduction in growing season length. Interannually, the magnitude of summertime R_{eco} showed a positive trend with water year precipitation and thus SWC. In addition, our data suggested greater decomposition of older soil C under deeper snowpack. Additional, multiyear measurements of plant productivity and net ecosystem exchange along with more continuous ^{14}C measurements of R_{eco} and of soil C pools are needed to investigate the effects of snow depth on the total ecosystem C balance and land-atmosphere CO_2 exchange. To quantify the annual C balance of High Arctic tundra, these studies should not cover only the growing season but also the winter, as recent work demonstrated that microorganisms function well below formerly assumed temperature and moisture limits.

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