Environmental factors regulating winter CO_2 flux in snow-covered boreal forest soil, interior Alaska

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Abstract

Winter CO_2 flux is an important element to assess when estimating the annual carbon budget on regional and global scales. However, winter observation frequency is limited due to the extreme cold weather in sub-Arctic and Arctic ecosystems. In this study, the continuous monitoring of winter CO_2 flux in black spruce forest soil of interior Alaska, over a permafrost regime, was performed using NDIR CO_2 sensors at 10, 20, and 30 cm above the soil surface during the snow-covered period (DOY 357 to 466) of 2006/7. The atmospheric pressure was divided into four phases: >1000 hPa (HP: high pressure); 985<P<1000 (IP: intermediate pressure); <986 hPa (LP: low pressure); and a snow-melting period (MP); for the quantification of the effect of the environmental factors determining winter CO_2 flux. The winter CO_2 fluxes were 0.22±0.02, 0.23±0.02, 0.25±0.03, and 0.17±0.02 gCO_2-C/m^2/d for the HP, IP, LP, and MP phases, respectively. Wintertime CO_2 emission represents 20% of the annual CO_2 emissions in this boreal black spruce forest soil. Atmospheric temperature and soil temperature explained 56 and 31% of winter CO_2 flux, respectively, during the snow-covered period of 2006/7, when snow depth experienced one of its lowest totals of the past 80 years. Atmospheric temperature and soil temperature at 5 cm depth, modulated by atmospheric pressure, were found to be significant factors in determining winter CO_2 emission and fluctuation in snowpack. Regional/global process-based carbon cycle models should be reassessed to account for the effect of winter CO_2 emissions, regulated by temperature and soil latent-heat flux, in the snow-covered soils of Arctic and sub-Arctic terrestrial ecosystems of the Northern Hemisphere.

1 Introduction

While winter CO_2 flux is an important carbon source in snow-covered sub-Arctic and Arctic ecosystems for the estimation of the annual carbon budget (Zimov et al., 1993, 1996; Oechel et al., 1997; Winston et al., 1997; Fahnestock et al., 1998; Kim et al., 2007), there are few reports on continuous winter CO_2-flux measurements in high latitudinal regions. Most studies, rather, have intermittently measured winter CO_2 flux with a static chamber built on the snow surface. These flux measurements are limited due to the extreme cold weather from December to February and issues with static and/or continuous chamber operation at identical sampling points, made difficult by newly accumulated snow in high latitudes. Winter CO_2 emissions, though, correspond to 10-30% of the annual soil respiration rate in alpine, sub-Arctic, and Arctic regions during the long (>200-day) snow-covered period (Sommerfeld et al., 1993; Zimov et al., 1993, 1996; Brooks et al., 1996; Oechel et al., 1997; Mast et al., 1998; Wickland et al., 2001; Kim et al., 2007; Liptzin et al., 2009; Seok et al., 2009), suggesting that the winter carbon contribution should not be overlooked when evaluating the annual carbon budget on regional and global scales. In this study, the monitoring of continuous winter CO_2-flux measurements was conducted using non-destructive infrared (NDIR) CO_2 sensors, installed before snowfall in black spruce forest soils during the seasonally snow-covered period of 2006/7. These sensors have been used in temperate forests during winter before (Hirano, 2005; Takagi et al., 2005); however, this study is the first to use these sensors to report on continuous winter CO_2-flux measurement in the boreal black forest of interior Alaska, under an environment of extreme cold.

The environmental factors influencing winter CO_2 flux are atmospheric pressure and wind speed (Massman et al., 1997; Takagi et al., 2005; Massman and Frank, 2006; Bowling et al., 2009: Seok et al., 2009), atmospheric temperature (Takagi et al., 2005), soil temperature (Zimov et al., 1993, 1996; Oechel et al., 1997; Winston et al., 1997; Hirano, 2005; Monson et al., 2006), soil moisture (Hirano, 2005: Liptzin et al., 2009), and snow depth (Fahnestock et al., 1998; Takagi et al., 2005). Liptzin et al., (2009) demonstrated the conceptual model of the seasonal pattern of CO_2 flux under four distinct zones, divided by changes in the environmental factors (e.g., freeze-thaw cycles, soil temperature, soil moisture, and carbon availability) and based on variability in snow coverage in the subalpine forest. Moreover, atmospheric pressure affects wind speed and atmospheric temperature, subsequent wind speed influences CO_2 fluctuation within the snowpack, and the ambient temperature modulates snow/soil temperatures. The soil temperature, depending on snow depth and atmospheric temperature, also governs the strength of microbial activity that terminally establishes the magnitude of CO_2 production in soils. We investigated each of these environmental factors affecting continuous winter CO_2-flux measurement through the snowpack in this study.

Several process-based ecosystem carbon models (e.g., Biome-BGC, TEM, and Sim-CYCLE) have used atmospheric temperature data as one of the key parameters for the assessment of the cycle and budget of terrestrial carbon on regional and global scales (e.g., Running and Coughlan, 1988; Kimball et al., 1997; McGuire et al., 2000; Ito and Oikawa, 2002; Lagergren et al., 2006). However, the implication of winter carbon emissions in the snow-covered Arctic and sub-Arctic terrain of the Northern Hemisphere upon the regional/global carbon budget is poorly accounted for in these models. Because vegetative photosynthesis and respiration does not occur in environments of extreme cold, soil-originated CO_2 emission through the snowpack represents the only ecosystem respiration during the winter. Therefore, continuous winter CO_2 emission, dependent on environmental factors, is a significant key in the winter carbon contribution to process-based terrestrial ecosystem carbon models, as well as to the assessment of the terrestrial carbon cycle/budget on regional and global scales.

2 Materials and Methods

2.1 Sampling Locations and Methods

The study site is a typical boreal forest in Fairbanks, in the Alaska interior (64°52'N, 147°51'W; 155 masl). The average monthly temperature in Fairbanks between 1971 and 2005 was lowest in January at –23.2°C, and highest in July at 16.9°C, with an annual average of –2.9°C (Shulski and Wendler, 2007). The average annual precipitation was 263 mm, of which approximately 37% fell as snow, and the rest as rain. The minimum temperatures at 80 cm above the soil surface and in soil 5 cm below the surface were -45.4°C (DOY 418 to 421) and -11.2°C (DOY 425 to 430), respectively, during the winter of 2006/7. The average snow depth during the winter of 2006/7 was 25 cm; this average was the third lowest since 1929 (Alaska Climate Research Center, 2008).

Black spruce (*Picea mariana*) is the dominant overstory tree species, with ages from 45 to 120 years (Vogel et al., 2005). The black spruce canopy is sparse. The average canopy height is about 3.5 m, but there are taller trees of up to 6 m, sporadically. Understory vegetation includes typical boreal forest shrubs, such as *Rhododendron groenlandisum*, *Vaccinium uliginosum*, *Vaccinium vitis-idaea*, and *Betula glandulosa*, as well as some *Carex* species. The forest floor is almost completely covered by mosses, such as *Sphagnum capillifolium*, *Sphagnum magellanicum*, *Sphagnum riparium*, *Calliergon stramineum*, *Aulacomnium palustre*, and patchy lichen, such as *Cladonia* species. Discontinuous permafrost is widely distributed 40 cm below the surface, and a thin, silty clay layer exists on the upper-most permafrost (Kim et al., 2007).

The sensor system was built on sphagnum and feather moss layers and was in operation from October 6, 2006 (DOY 280) to April 30, 2007 (DOY 485) for the monitoring of continuous CO_2 concentration in snowpack during the winter of 2006/7 (Figure 1). The non-dispersive infrared sensor (NDIR; Vaisala GMD 20; Helsinki, Finland) was set on a length of wooden stick (3 cm diameter, 100 cm long) at four directional levels (10, 20, 30, and 50 cm above the moss surface) for prevention of disturbance (Figure 1). This sensor is the same type used for prior winter CO_2-flux measurements (Hirano et al., 2003; Takagi et al., 2005). The installed sensor was covered with a PVC pipe (48 mm OD; 40 mm ID; 170 mm long), open on one end, for water and sensor-window protection. The *in-situ* sensor head (155 mm long and 15 mm in diameter) has an NDIR source, optical filter, and detector, and a 50-mm long and 4-mm wide slit on the head that allows CO_2 from the soil to diffuse through membranes into the small sample cell (ca. 2.6 cm^3), as used by Hirano et al. (2003) and Takagi et al. (2005). The sensor detects CO_2 concentration through molecular diffusion from the soil to the snowpack, assuming that soil-originated CO_2 emission within the diameter of the sensor (e.g., 20 cm) is constant. The cable from each sensor was connected to a datalogger (CR 1000, Campbell Scientific Inc., USA) within an ice cooler for the storage of CO_2 data. A commercial heating pad was used for operation of the logger during winter. CO_2 concentration measured at the 50 cm level above the surface is not discussed here due to unexpected failure of the sensor in the extremely cold weather.

The calibration of each sensor was conducted using certified EPA protocol for a CO_2 standard cylinder (1000.0 ppm; Airgas Inc., USA) and zero gas (pure N_2 cylinder; Airgas Inc., USA), before and after the monitoring of CO_2 concentration in the laboratory. The sensor responded to the standard cylinder within 10 seconds, and repeatedly measured standard CO_2 concentration for 60 minutes. The precision of each sensor was determined using zero gas and 1000.0 ppm standard cylinders, ranging from 978±6 ppm (0.61%) to 1020±47 ppm (4.30%) before the observation, and from 967±7 ppm (0.72%) to 1031±47 ppm (4.57%) after the observation, for the calibration of 1000.0 ppm standard CO_2 cylinder over an hour. The precision on each sensor ranged from 0.6 to 4.6%, and the CO_2 concentration of each sensor was corrected. The CO_2 concentration in the snowpack was calculated at 30-min intervals for each corrected sensor.

Temperatures in snow and soil were measured at 10, 20, 30, 40, 50, and 80 cm above, and at 5, 10, 15, 20, 30, 40, and 50 cm below the soil surface, and were monitored at a 1.5-h interval, with sensors (TMC6-HC, Onset Computer Corporation, USA) and 4 external channel-loggers (U-12 HOBO, Onset Computer Corporation, USA). Soil moisture was monitored at 5 and 20 cm below the surface at a 1.5-h interval using sensors (ML2x, Dynamax Inc, USA) and a 2-channel logger (THLOG-2, Dynamax Inc, USA). The monitoring of temperature and soil moisture was conducted from September 12, 2006 to September 6, 2007. Atmospheric pressure was recorded by barometer (CS100, Campbell Scientific Inc., USA) every 30 min at 8 m at the eddy-covariance tower site. The daily snow-depth data was taken from the Alaska Climate Research Center of the Geophysical Institute (GI) at the University of Alaska Fairbanks (UAF) during the winter of 2006/7 (Alaska Climate Research Center, 2008: Figure 2). Because the snow depth was much lower than in normal years, winter CO_2 flux was estimated between DOY 357 (December 23, 2006) to 466 (April 11, 2007), when the snow depth was higher than 25 cm. While the snow depth was less than 20 cm before DOY 257, the winter flux could not be estimated. The accumulated snowpack began to melt on DOY 446 (March 21, 2007). The snow survey was also conducted at a two-week interval. Two to five snow samples were collected using a snow density sampler (4 cm H × 5 cm W × 5 cm D) and a snow cutter for the estimation of snow porosity (Kim et al., 2007).

2.2 Estimation of Winter CO_2 Flux

The winter CO_2 flux through snowpack to the atmosphere was obtained by applying the following equation under a steady-state condition: $F_{CO2} = D \cdot (\partial C/\partial z) \cdot \tau \cdot \theta$ (Kim et al., 2007), where D is CO_2 diffusivity corrected only for the in-situ temperature within the snowpack measured in cm^2/sec (Sommerfeld et al., 1993; Fahnestock et al, 1999); $\partial C/\partial z$ is the vertical CO_2 concentration gradient observed within the snowpack in ppmv/cm; τ is tortuosity; and θ is the snow porosity. The CO_2 concentration gradients from 10 to 20 cm and from 20 to 30 cm were similar, indicating that the gradient is almost linear; the gradient ratios for the 10-20 cm and 20-30 cm ranges varied from 0.87 to 1.22 and showed no difference under the 95% confidence level. Porosity was calculated from the density of ice (ρ_{ice}=0.91) and the water contents of the snowpack over the gradient interval. Tortuosity is difficult to measure and is usually described as a function of porosity, with values ranging from $\theta^{1/3}$ to $\theta^{2/3}$ (Striegl, 1993). In this study, the tortuosity of the snowpack was estimated by the theoretical relation $\tau = \theta^{1/3}$ (Millington, 1959), which yielded values ranging from 0.74 to 0.92. These values are similar to the range of 0.70 to 0.91 for the whole observation period for boreal forest snowpack in interior Alaska. Sommerfeld et al. (1993), Mast et al. (1998), and Kim et al. (2007) reported similar data (0.68 to 0.90) in subalpine snowpack in Wyoming and Colorado, and in boreal forest snowpack in Alaska. The snowpack at high-latitude boreal black spruce forest sites has always been under dry conditions, except for the snow-melting period. At a snow density of 150 kg/m^3 and assuming all other variables unchanged, the diffusion rate was 79% faster than at a snow density of 300 kg/m^3, indicating that errors in the estimate of CO_2 flux through the snowpack caused by incorrect measurements of density varied as snow density changed (Seok et al., 2009). In our case, measured snow density and snow depth were much smaller than Seock et al. (2009)'s values. Nevertheless, we used the sensitivity of calculated CO_2 fluxes to estimate snow density as suggested by Seok et al. (2009) (see Figure S1). They demonstrated that the propagated errors from porosity and tortuosity estimation resulted in snow density uncertainties estimates of ±10, 20, and 30%, shown as a function of absolute snow density value. For example, a 10% error in the measurement of snow density resulted in an error in the estimated CO_2 flux on the orders of 3 and 5% for a snow density of 150 and 300 kg/m^3, respectively. We estimated that the error in calculating CO_2 flux ranged from 1 to 11%, compared to the 2-9% error evaluated by Seok et al. (2009). Crust was formed by the sublimation; however, we did not consider the effect of the ice layer upon estimating CO_2 flux, because freeze-thaw events did not occur under the cold environment before the onset of snow thaw.

2.3 Analysis of Soil Heat Flux

We correlated winter CO_2 flux with the non-conductive heat flux component of the active layer. The non-conductive heat component, r_h, is expressed in terms of volumetric heat production in W/m^2 and is estimated by considering one-dimensional energy conservation as formulated:

$$r_h = c_h \frac{\partial T}{\partial t} - k_h \frac{\partial^2 T}{\partial z^2} \quad (1)$$

where c_h is the volumetric bulk heat capacity, k_h is bulk thermal conductivity, T is temperature, t is time, and z is depth. Neglecting energy exchange below the lowest measurement, the total amount of non-conductive heat components, R_h, is the result of r_h multiplied by the thickness of the soil layer, d:

$$R_h = \sum_i r_h^i d^i \quad (2)$$

where the subscript i represents the i-th layer from surface to bottom. We set the mid-depth of the i-th layer to be at the i-th measurement depth from the surface. Accordingly, the soil column was divided into three layers, the thicknesses of which were 5, 7.5, and 10 cm from the surface to bottom (25 cm). Finite element formulations to solve equations (1) and (2) are described in Ishikawa et al. (2006).

We assumed k_h to range from 5.5×10^{-7} to 8.0×10^{-7} J/kg/K, referring to the thermal diffusivity for frozen silty clay shown by Yershov (1998; $d_h = 5.5 \times 10^{-7}$- 8×10^{-7} m^2/s) and to its heat capacity shown by Roth and Boike (2001; c_h = 2.2±0.2 × 10^6 J/m/K). These calculations neglected the contribution of soil air because of its very low mass density.

3. Results and Discussion

During the winter, wind speed and direction have been important factors affecting winter CO_2 fluxes in temperate and subalpine regions (Massman et al., 1997; Takagi et al., 2005; Massman and Frank, 2006; Bowling et al., 2009; Liptzin et al., 2009; Seok et al., 2009). The wind speed measured at 2 m from the eddy covariance tower was less than 2 m/s in the black spruce forest of interior Alaska during the observed winter period, compared to the 0-6 m/s measured in subalpine regions (Massman et al., 1997; Filippa et al., 2009; Liptzin et al., 2009; Seok et al., 2009), and 0-3 m/s in a temperate-climate region (Takagi et al., 2005), which were affected by wind-pumping when estimating CO_2 flux through the snowpack. Relationships here between air pressure and wind speed at 2 m, and between CO_2 concentration gradient and wind speed at 2 m, had low correlation, showing correlation coefficients of 0.017 and 0.069, respectively. This suggests wind speed in the black spruce forest of interior Alaska during the winter may not have played a significant role when estimating CO_2 flux in response to changes in wind speed, contrary to strong wind speed in subalpine and temperate regions. However, most (>96%) of the wind speed at 2 m during winter at our location was less than 1.0 m/sec. In order to validate the effect of wind pumping on CO_2 concentration variability at each level in the snowpack, we used the 2nd order polynomial fit ($y = cx^2 + bx + a$) as estimated by Seok et al. (2009), illustrating the relationship between CO_2 concentration at each level and wind speed at 2 m (not shown in the relationship). The finding is quite different than Seok et al. 's (2009) result. The c in this study, characterized by the curvature of the best-fit equation, tends to decrease with increased depth, indicating little sensitivity toward wind speed under shallow snowpack and much weaker wind speed environment, than the findings of Seok et al. (2009) under deeper snowpack and strong wind speed during winter. The trend in b is similar to term c. The regression term a, denoting the zero-wind speed snowpack CO_2 concentration at each height, increases linearly moving from the bottom of the snowpack, indicating that the CO_2 source is from the soil. When wind speed is zero in this study, the average CO_2 concentrations at each height over the whole winter are 627, 532, and 474 ppm at 10, 20, and 30 cm in the snowpack, respectively, suggesting that wind speed is much weaker at this study site. This demonstrates that there is no wind-pumping effect on the black spruce forest soil of interior Alaska during the seasonal snow-covered period of 2006/7. Thus, contrary to temperate (Takagi et al., 2005) and subalpine regions (Seok et al., 2009; Liptzin et al., 2009), winter CO_2 flux as an effect of wind pumping is not considered in our study and was estimated through the application of Fick's law. We used 6-h averages of CO_2 concentration, winter CO_2 flux, atmospheric pressure, temperatures in air and soil, and soil moisture, during the snow-covered period of 2006/7.

3.1. Environment Factors and CO_2 Concentration

Soil moisture and the temperatures at 80 cm above the soil surface and at 5 cm below the surface were monitored from DOY 255 to 614 (Figure 3). Atmospheric temperature showed a higher daily variation, and the temporal fluctuation of soil temperature was lower. Soil moisture at 5 cm below the surface was affected by low (<0°C) atmospheric temperature and soil freezing, and the freezing rate from 5 to 20 cm was 0.75 cm/d, suggesting that the time it took the freezing front to reach 20 cm was 20 days. Kim et al. (2007) reported a freezing rate of 4 cm/d for 10 to 30 cm below the surface during the winter of 2004/5 in the same observation area. In our study, the frozen soils began to thaw at 5 cm by DOY 489 (May 4, 2007), and at 20 cm by DOY 508 (May 22, 2007); the melting rate over these 19 days was 0.78 cm/d, similar to the freezing rate in early winter.

Ambient pressure and temperature ranged from 943 to 1020 hPa and from -45 to 17°C, respectively, during the period of DOY 350 to 466. The temporal variation in pressure showed an inverse tendency to the change in temperature. Thus, in order to quantify the effects of pressure and temperature for winter CO_2 flux, the magnitude of pressure during the snow-covered period was divided into four phases: high pressure (HP: >1000 hPa); intermediate pressure (IP: 985 hPa < P < 1000 hPa); low pressure (LP: <985 hPa); and a snow-melting period (MP, after DOY 466); all shown in Figure 2. Atmospheric temperature was -31.9±11.0°C (Coefficient of Variance [CV]: 35%) for HP; -22.1±8.6°C (CV: 39%) for IP; -21.5±6.8°C for LP; and -8.4±12.4°C (CV: 146%) for MP. These air pressure phases, then, correspond to the magnitude of air temperature.

Figure 4 shows the relationship between ambient temperature and temperatures in snow (10, 20, and 40 cm above the soil surface), and soil (5 cm below), in order to demonstrate additional influence on ambient temperature. The ambient temperature indicates correlation coefficients (R^2) of 0.995, 0.99, and 0.79 for snow at 40, 20, and 10 cm above the surface, respectively, and 0.08 for soil 5 cm beneath the surface, suggesting that the extent of atmospheric temperature influence reached to 20 cm within the snowpack when the snow depth was less than 40 cm.

CO_2 concentrations at 10, 20, and 30 cm above the soil surface are shown with temporal variations in pressure in Figure 5. The 6-h average CO_2 concentrations in the snowpack were 627±19 ppm (CV: 3.0%) for 10 cm, 532±18 ppm (CV: 3.3%) for 20 cm, and 473±32 ppm (CV: 6.7%) for 30 cm. The concentration range of 365 to 692 ppm in sphagnum/feather moss regimes is comparable to the 400 to 740 ppm measured in tussock tundra/sphagnum moss regimes of boreal forest soils (Kim et al., 2007), during which tussock tundra was also found to be one of the carbon sources in boreal forest and Arctic terrestrial ecosystems of the Northern Hemisphere (Oechel et al., 1997; Kim et al., 2007). The temporal variations in CO_2 concentration showed a similar trend at 10, 20, and 30 cm levels, and may be affected by ambient pressure, as is the case in the relationship between pressure and ambient temperature.

3.2 Estimation of Winter CO_2 Flux

Winter CO_2 flux varied from 0.19 to 0.26 gCO_2-$C/m^2/d$ for the HP phase (>1000 hPa), from 0.19 to 0.27 gCO_2-$C/m^2/d$ for IP (985<P<1000), from 0.20 to 0.32 gCO_2-$C/m^2/d$ for LP (<985 hPa), and from 0.14 to 0.24 gCO_2-$C/m^2/d$ for MP. The average winter CO_2 flux and atmospheric temperature for the four pressure phases are shown in Table 1. Average winter CO_2 flux among the three pressure phases, excluding the snow-melting period, was not significantly different based on a one-way ANOVA with a 95% confidence level. During the snow-covered period of 109 days, the average CO_2 flux was 0.22±0.02 gCO_2-$C/m^2/d$ (CV: 10%), indicating a value corresponding to those measured by concentration profile (0.21±0.06 gCO_2-$C/m^2/d$) and chamber (0.26±0.06 gCO_2-$C/m^2/d$) methods during the winter of 2004/5 in the same black spruce forest soils of interior Alaska (Kim et al., 2007). Furthermore, the snow depth in the winter of 2004/5 was much deeper (>20 cm) than 2006/7. Although the snow depth was greater, the minimum soil temperature at 5 cm below the surface was -17°C, due to an extremely cold ambient temperature of -55°C (January 12, 2005). This suggests that the greater snow depth (68 cm) plays little role in insulating the soil below -50°C. In the temperate forests and grassland soils of northern Japan, greater snow depth (>80 cm) has kept soil at 5 cm beneath the surface warmer than zero (Takagi et al., 2005), and an increase in snow depth (35 to 70 cm) caused a temperature jump from -0.42 to 0.15°C at a 5-cm soil depth (Kim and Tanaka, 2002). This kind of change in soil temperature modulates the magnitude of soil CO_2 production by affecting soil microbial activity in tundra soils during the winter (Oechel et al., 1997; Panikov et al., 2006).

Temporal variations in pressure and ambient temperature for winter CO_2 flux are shown in Figure 6. The temporal variation of winter CO_2 flux shows a tendency that is qualitatively inverse to that of pressure (Figure 6a) but is similar to that of ambient temperature (Figure 6b). The winter CO_2 flux abruptly decreased from 0.28 to 0.17 gCO_2-C/m^2/d by DOY 446 (March 21, 2007), which was the first day of snow melting—when ambient temperature increased to above zero, as shown in Figure 3. Also, the temperature dropped from 1.23 to -13.8°C, and the pressure increased from 959 to 980 hPa. Therefore, the atmospheric temperature, modulated by the pressure, is a significant factor in determining winter CO_2 flux in the seasonally snow-covered boreal forest soil of interior Alaska.

3.3 Environmental Factors Regulating Winter CO_2 Flux

Winter CO_2 flux has a direct relationship to atmospheric pressure for HP (>1000 hPa), LP (<985 hPa), and MP (snow-melting) days of the snow-covered period (Figure 7a), indicating an inverse correlation for each pressure phase. The data for the IP (985<P<1000) phase is virtually excluded in Figure 7a—the temperatures in air and soil during IP are discussed below. The correlation coefficients (R^2) were 0.25 for HP, 0.31 for LP, and 0.18 for MP. Ambient pressure has a lesser effect in determining winter CO_2 flux through the snowpack to the atmosphere during the winter season.

Winter CO_2 flux shows a strong exponential relationship to ambient temperature, though, for three pressure phases: the correlation coefficients were 0.80 at low temperature for HP, 0.26 at high temperature for LP, and 0.58 for MP (Figure 7b). The regression curves in Figure 7b are $Y = 0.27e^{(0.069T)}$ for HP, $Y = 0.29e^{(0.007T)}$ for LP, and $Y = 0.18e^{(0.065T)}$ for MP. Figure 7c shows the relationship between CO_2 flux and ambient temperature for IP (985<P<1000), which also has strong correlation, suggesting that the air temperature accounted for 58% of the variability of winter CO_2 emission during IP, with a regression curve of $Y = 0.27e^{(0.064T)}$. During the LP phase, the coefficient was less than half the coefficient in either HP, MP, or IP. Winter CO_2 flux during the early days of the snow-covered period was much higher than during the remainder of this period. This may be due to a higher concentration difference between the 10- and 20-cm levels before DOY 368, when the snow depth was less than 27 cm. As a result, we calculated the 6-h average CO_2 concentration gradient before and after DOY 368; the difference in CO_2 flux is likely due in part to warm soil temperature before DOY 368. The soil temperature is dependent on the snow depth and affects the soil microbial physiology and the community composition (Brooks et al., 1996; Oechel et al., 1997; Kim and Tanaka, 2002; Takagi et al., 2005; Monson et al., 2006).

The Q_{10} is the temperature coefficient of the reaction and is defined as the ratio of reaction rate at an interval of 10°C. Our Q_{10} values were 1.22 for HP, 1.25 for LP, 1.26 for MP, and 1.37 for IP. These values are much lower than those of previous studies during the winter (Oechel et al., 1997; Monson et al., 2006). Monson et al. (2005) reported the R_T (a first-order exponential coefficient analogous to the Q_{10} coefficient used in biochemical studies) was 105 near trees and $1.25x10^6$ in the open space of the LTER Niwot Ridge Ameriflux site. These values are several orders of magnitude higher than the range of Q_{10} values found in previous studies of terrestrial ecosystem soils, demonstrating that higher temperature sensitivity invokes a physical limitation to substrate diffusion—as liquid water disappears below zero (between 0 and -1°C). Panikov et al. (2006) proved that soil CO_2 production occurred even under the extremely cold soil temperature of -39°C, with soil core samples (0-30 cm) from Barrow, Siberia, and Sweden; their Q_{10} values ranged from 2.1 to 8.5.

Estimated soil non-conductive heat flux evolved negatively through the period from DOY 357 to 460, for both upper and lower d_h and c_h (Figure 8a). Assuming that this heat arises from a single phase transition of water, we compared soil moisture observed, using R_h divided by the enthalpy of the transition from phases α to β, $L^{\alpha\beta}$ ($\{L^{sl}, L^{lv}\} = \{0.333, 2.45\}$ MJ/kg, where the superscript s, l and v represent solid, liquid, and vapor, respectively), and found that this heat was mostly from vaporization. As shown in Figure 8a, winter CO_2 flux showed a decreasing trend until the end of the snow-covered period, while latent heat flux showed an increasing trend. Also, both fluxes showed significant correlations (R^2 = 0.49 and 0.52, with $p < 0.001$ in both) before the onset of snow melting (Figure 9). These findings suggest that the higher upward vapor movement in the soil column occurred in accordance with the smaller, soil-originated CO_2 flux. We postulate that winter soil-originated CO_2 is hampered by the reduction of snow pores linked to the atmosphere—due to compaction of the snowpack, vapor condensation in the snow column, and subsequent snow metamorphism. This consideration is supported by the comparison between CO_2 flux and the snow temperature gradient (Figure 8b). Winter CO_2 flux was occasionally greater when the snow temperature profile approached the isotherm and the condensation rate was reduced (e.g., DOY 364, 368, 392-399, 406, 411, 423, 434; Figure 8a and b). The temperature gradient governed the vapor pressure gradient through the snow and soil column, modulating evaporation, condensation, and vapor movements. This modified the passage of winter CO_2.

Figure 10 shows the correlation coefficient percentages (%) for atmospheric pressure, temperature, soil temperature, and winter CO_2 flux during each pressure phase. During HP, IP, and MP, the strongest environmental factors determining winter CO_2 flux were atmospheric temperature, soil temperature, and atmospheric pressure, respectively. Takagi et al. (2005) implied that winter CO_2 flux responded directly to ambient temperature, and not to soil temperature, even beneath a 1-m snowpack of temperate forest soils in Japan. They inferred that the atmospheric temperature affected the root activity of trees through their trunks and that the variation in root respiration strongly affected fluctuation in CO_2 concentration in soil under the snowpack. Vogel et al. (2005) suggested that the contribution of root respiration in mature black spruce forest soils varied from 81-85% of total soil respiration during the winter. Moreover, because their site is similar to the study site here, Kim et al. (2007) demonstrated that the $\delta^{13}CO_2$ of -22.5‰ originated from root respiration rather than heterotrophic respiration in black spruce forest soils of interior Alaska during the winter. Atmospheric temperature and soil temperature at 5 cm, depending on ambient pressure, therefore, play significant and key factors in regulating winter CO_2 emission through the snowpack in these forest soils during the snow-covered period.

3.4 Implication for Regional Winter Carbon Budget

Average wintertime CO_2 emission was 24.3±1.3 gCO_2-C/m^2/season (CV: 5%) during the experimental period of 109 days. For our four pressure phases, average emissions are shown in Table 2. The cumulative snow depth in 2006/7 was one of the lowest years in Fairbanks over the past 80 years—the snow depth during the 212-day winter period of 2006/7 corresponds to merely half of a normal season. However, winter CO_2 emission has always occurred before DOY 357 and after DOY 466. Thus, winter CO_2 emission was reevaluated as a half of the average flux (0.23±0.02 gCO_2-C/m^2/d) measured before DOY 357 (December 23, 2006) and as a half of the flux (0.17±0.02 gCO_2-C/m^2/d) measured after DOY 466 (March 21, 2007), based on additional CO_2 flux-measurement (0.07±0.03 gCO_2-C/m^2/d; n=24) with the dynamic chamber on the melting snow surface during the snow-melting period (DOY 100) of 2007/8. The wintertime CO_2 emission was 36±1.7 gCO_2-C/m^2/season (CV: 5%) during the winter of 2006/7. This emission corresponds to 20% of the annual CO_2 emitted from boreal black forest soils in interior Alaska; the CO_2 emission was 142±57 gCO_2-C/m^2/season (CV: 40%) in the same study site during the growing period of 2006. Kim et al. (2007) reported that the wintertime CO_2 emission was 49±13 gCO_2-C/m^2/season in the same boreal forest during the winter of 2004/5, a difference due to a longer snow-covered period and greater snow depth than in this study. Also in the boreal forest of interior Alaska, Vogel et al. (2005) measured a winter respiration of 36-54 gCO_2-C/m^2/season in three different tree ages (75, 110, and 120 years) of black spruce forests of Bonanza Creek, interior Alaska, representing 8-18% of the annual CO_2 emission. Further, the winter respiration was 40-55 gCO_2-C/m^2/season in black spruce and jack pine forests of the BOREAS study area (Winston et al., 1997) and later 25-35 gCO_2-C/m^2/season in the black spruce forest of the BOREAS area (Wang et al., 2003), in which the winter carbon contributions accounted for 5-19% of the annual respiration. The BOREAS study area contained one south-facing and one north-facing vegetation distribution, resulting in a difference of soil drainage that greatly affects the species composition and functions of the boreal forest ecosystem (Wang et al., 2003). The magnitude of soil drainage regulates the decomposition rate of soil organic carbon and the vegetation biophysical conditions.

In snow-covered Arctic tundra ecosystems of the North Slope of Alaska, the winter CO_2 emissions from moist tussock tundra and coastal wet sedge were 70 and 20 gCO_2-C/m^2/season, respectively (Oechel et al., 1997). That is the main share of the total annual net carbon emission in Arctic tundra ecosystems. In the same Arctic tundra ecosystems, winter CO_2 emission ranged from 1.3 to 11 gCO_2-C/m^2/season (Fahnestock et al., 1998), depending on the vegetation community types, flux that represents up to 17% of the annual carbon flux of Arctic tundra ecosystems.

Considering all the snow-covered tussock and moss regimes in the Northern Hemisphere (6.5 x 10^{12} m^2; Whalen and Reeburgh, 1998), winter carbon emission should not be overlooked when estimating regional and global carbon budgets. Furthermore, regional/global process-based CO_2 cycle models should be sufficiently discussed and modified to include winter CO_2 contribution, considering atmospheric temperature as a key regulating factor and depending on atmospheric pressure, in snow-covered soils of Arctic and sub-Arctic terrestrial ecosystems in the Northern Hemisphere.

4 Conclusions

The continuous monitoring of winter CO_2 flux in snowpack was conducted in sphagnum and feather moss regimes of black spruce forest soils of interior Alaska during the winter of 2006/7. Measurements were taken of key environmental factors that regulate winter CO_2 flux, such as atmospheric pressure and temperatures in air and soil, during the snow-covered period from DOY 357 (December 23, 2006) to 466 (March 11, 2007). Atmospheric pressure was divided into four phases: >1000 hPa (HP: high pressure), <986 hPa (LP: low pressure), 985<P<1000 (IP: intermediate pressure), and the snow-melting period (MP), for the quantification of the effect of atmospheric pressure on temperature-modulated winter CO_2 flux. Winter flux greatly depends on atmospheric temperature, which is governed by these four pressure phases. Pressure is an important factor in indirectly and directly influencing atmospheric temperature and winter CO_2 flux. The transport of CO_2 emissions through soil and snow columns is modified by snow compaction and metamorphism and is modulated by evaporation, condensation, and vapor movements through the columns. Moreover, atmospheric temperature and soil temperature play significant roles in determining winter CO_2 flux, demonstrated by the fact that atmospheric temperature accounted for an average of 56% of the variability of winter CO_2 emission during the snow-covered period. Because snow-covered tussock and moss regimes are widely distributed in Northern Hemisphere, wintertime carbon emission is of considerable significance when estimating seasonal, regional and global carbon budgets, as this emission represented 20% of the annual soil carbon emissions in black spruce forest soils in interior Alaska during the cold winter of 2006/7. In order to evaluate the effect of wind pumping when estimating CO_2 flux, we need additional study on wind-pumping effect using installation of pressure sensors and build-up of NDIR in the snowpack in other, relatively sparse black spruce forest of interior Alaska. Regional/global process-based CO_2 cycle models should be reassessed to consider the effect that atmospheric temperature and soil latent-heat flux have in regulating winter CO_2 emissions in snow-covered soils of Arctic and sub-Arctic terrestrial ecosystems in the Northern Hemisphere.

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Figure Legends

Figure 1. Sampling scheme of observation system in sphagnum and feather moss regimes of black spruce forest, interior Alaska during the winter of 2006/7.

Figure 2. Daily snowfall (black line) and cumulative snowpack (grey line) during the snow-covered period of DOY 357 (December 23, 2006) to 466 (April 11, 2007). Winter CO_2 flux is estimated when the snow depth is over 30 cm, due to the sensor levels.

Figure 3. Seasonal variations of 6-h average temperatures at air 80 cm above (grey line) and 5 cm below (black line) soil surface, and soil moistures at 5 cm (grey circles) and 20 cm (black circles) below surface from August 12, 2006 (DOY 254) to September 10, 2007 (DOY 618).

Figure 4. Relationship between ambient temperature and temperatures of soil (5 cm below surface) and snow (10, 20, and 40 cm above surface) from DOY 254 to 618. When the snow depth was over 20 cm, the snow temperatures at 20 and 40 cm above the soil surface depend on the ambient temperature. The symbols indicate solid circles for soil at 5 cm below the surface, open grey circles for snow at 10 cm above, grey squares for snow at 20 cm above, and crossed squares for snow at 40 cm above, respectively. The dotted line is a 1:1 line.

Figure 5. Time series of CO_2 concentrations at 10, 20, and 30 cm above the soil surface within the snowpack, with seasonal change in atmospheric pressure. The concentration data was comparable with measurements from 392 to 742 ppm during the winter of 2004/5 (Kim et al., 2007).

Figure 6. Temporal variations of winter CO_2 flux along with 6a) atmospheric pressure and 6b) ambient temperature. Atmospheric pressure affects temperature, which regulates the magnitude of winter CO_2 flux. Thus, the pressure is divided into four phases: high pressure (HP: >1000 hPa), low pressure (LP: <985 hPa), intermediate pressure (IP: 985<P<1000), and a snow-melting period (MP: since DOY 446), all shown in Table 1. The temperature was much higher for LP than for HP—a difference of over 10°C on average.

Figure 7. Relationships between winter CO_2 flux and: 7a) atmospheric pressure during HP, LP, and MP; 7b) ambient temperature during HP, LP, and MP; 7c) ambient temperature during IP. The empty area for IP in 7a and 7b denotes exclusion between 985 and 1000 hPa. Winter CO_2 fluxes show good exponential relations with ambient temperature for four pressure phases. The symbols are open circles for HP, stars for LP, solid circles for MP, and solid triangles for IP.

Figure 8. 8a) Temporal variations in winter CO_2 flux (g CO_2-C/m^2/d) and soil non-conductive heat flux, R_h (W/m^2), estimated from equation (1), as written in the text. Both fluxes were averaged on a daily basis. CO_2 flux is represented by a solid grey line. R_h was estimated for upper bounds (solid black line, $\{d_h, c_h\} = \{8.0 \times 10^{-7}$ m^2/s and 2.4×10^6 J/m/K$\}$) and lower bounds (dashed black line, $\{d_h, c_h\} = \{5.5 \times 10^{-7}$ m^2/s, 2.0×10^6 J/m/K$\}$). 8b) Snow temperature gradients between the soil surface and 10 cm above, from DOY 357 to 460. Ts (0 cm) and Ts (10 cm) denote temperatures at soil surface and at 10 cm above the surface.

Figure 9. Correlations between non-conductive heat and winter CO_2 flux until DOY 445. Solid and grey circles are maximum and minimum soil heat flux, respectively, as described in Figure 10. Solid and dashed lines show the relationships between winter CO_2 flux and soil non-conductive heat fluxes at maximum and minimum, respectively.

Figure 10. The percentage (%) correlation, measured using the correlation coefficient (R^2), between environmental factors and winter CO_2 flux during 10a) HP, 10b) IP, 10c) LP, and 10d) MP. The numbers on the lines are the percentages of correlation between both parameters. The dotted, thin solid, and thick solid lines denote $R^2<0.20$, $0.20<R^2<0.40$, and $R^2>0.40$, respectively.