

Winter fluxes of CO₂ and CH₄ from subalpine soils in Rocky Mountain National Park, Colorado

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Abstract. Fluxes of CO₂ and CH₄ through a seasonal snowpack were measured in and adjacent to a subalpine wetland in Rocky Mountain National Park, Colorado. Gas diffusion through the snow was controlled by gas production or consumption in the soil and by physical snowpack properties. The snowpack insulated soils from cold midwinter air temperatures allowing microbial activity to continue through the winter. All soil types studied were net sources of CO₂ to the atmosphere through the winter, whereas saturated soils in the wetland center were net emitters of CH₄ and soils adjacent to the wetland were net CH₄ consumers. Most sites showed similar temporal patterns in winter gas fluxes; the lowest fluxes occurred in early winter, and maximum fluxes occurred at the onset of snowmelt. Temporal changes in fluxes probably were related to changes in soil-moisture conditions and hydrology because soil temperatures were relatively constant under the snowpack. Average winter CO₂ fluxes were 42.3, 31.2, and 14.6 mmol m⁻² d⁻¹ over dry, moist, and saturated soils, respectively, which accounted for 8 to 23% of the gross annual CO₂ emissions from these soils. Average winter CH₄ fluxes were -0.016, 0.274, and 2.87 mmol m⁻² d⁻¹ over dry, moist, and saturated soils, respectively. Microbial activity under snow cover accounted for 12% of the annual CH₄ consumption in dry soils and 58 and 12% of the annual CH₄ emitted from moist and saturated soils, respectively. The observed ranges in CO₂ and CH₄ flux through snow indicated that winter fluxes are an important part of the annual carbon budget in seasonally snow-covered terrains.

1. Introduction

Landscapes at northern latitudes and high elevations account for a significant percentage of the Earth's land surface and include arctic tundra, boreal forest, temperate forest, subalpine forest, and alpine tundra [Bouwman, 1990]. Soils in many of these landscapes have been identified as important sources and sinks of atmospheric CO₂ that may be particularly sensitive to climate change [Oechel *et al.*, 1997]. For example, soils in arctic ecosystems contain large pools of stored carbon that may now be acting as net sources of CO₂ to the atmosphere because of increases in surface soil temperatures over the past 3 decades [Oechel *et al.*, 1997]. Ciais *et al.* [1995] recently identified a strong terrestrial biospheric sink of CO₂ in the northern hemisphere, although the process of uptake has not yet been identified. Wetlands are the largest natural source of CH₄ to the atmosphere, and northern and high-elevation wetlands may account for a third of this natural source [Moosavi and Crill, 1996].

As much as half of northern latitude and high-elevation landscapes are snow covered for most of the year [Sommerfeld *et al.*, 1993]; yet, most attention has been focused on soil gas emissions in these environments during the growing season. Soils under

seasonal snowpacks commonly are warm enough to support microbial activity through the winter, but relatively little is known about the processes and rates of gas flux from snow-covered soils [Melloh and Crill, 1996; Sommerfeld *et al.*, 1993]. In addition, few studies have quantified the importance of winter gas emissions from different soil types in terms of the annual carbon budgets for these ecosystems [Brooks *et al.*, 1995].

CO₂ fluxes from snow-covered soils are largely due to microbial respiration, although root respiration may be important in some environments, and rates appear to be related to the effectiveness of the snowpack in insulating soils from winter air temperatures. Sommerfeld *et al.* [1993] observed positive CO₂ fluxes from subalpine soils in Wyoming throughout the winter where the snowpack was deep enough to prevent soils from freezing. In the alpine zone in Colorado, CO₂ fluxes were low in early winter because a shallow snowpack caused surface soils to freeze, then fluxes increased after additional snow accumulation allowed frozen soils to thaw [Brooks *et al.*, 1997]. Winter CO₂ emissions from soils in a boreal forest ecosystem in Canada appeared to be derived from a deep soil source because surface soils froze owing to the combination of shallow snowpacks and low air temperatures [Winston *et al.*, 1995]. Zimov *et al.* [1993] and Oechel *et al.* [1997] measured winter CO₂ emissions from soils in the arctic zone. In these environments, extremely cold air temperatures caused soil temperatures to drop well below the threshold for microbial respiration, and CO₂ emissions probably were related to physical processes that released CO₂ produced during the early part of the cold season [Oechel *et al.*, 1997].

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Winter CH₄ production from microbial activity in anoxic environments may comprise a significant part of the annual CH₄ budget for wetlands. *Dise* [1992] measured CH₄ fluxes from several moisture regimes in a Minnesota peat bog and found that winter fluxes accounted for as much as 21% of the annual flux in this environment. Winter CH₄ fluxes accounted for 4% of the annual flux from a temperate wetland in New Hampshire [Melloh and Crill, 1996] and between 4 and 43% of the annual flux from different plant communities in a subarctic muskeg [Whalen and Reeburgh, 1988]. Snow-covered soils also may serve as sinks of atmospheric CH₄ through the winter. CH₄ consumption by forest soils was observed to occur through winter in the subalpine zone of the Rocky Mountains [Mast et al., 1994; Sommerfeld et al., 1993] and in a boreal ecosystem in Canada [Wickland and Striegl, 1997].

This study investigates soil CO₂ and CH₄ emissions during winter in the subalpine zone of the Loch Vale watershed in Rocky Mountain National Park, Colorado. Gas concentrations in the snowpack were measured during the snow-covered seasons of 1994, 1995, and 1996 as a means of estimating gas flux through snow. Physical properties of the snowpack and the soils were measured during the winter. Direct measurements of flux also were made at the snow surface using a static chamber technique. These data allowed us to (1) investigate the effect of the snowpack on the exchange of gas between the soil and atmosphere, (2) determine the magnitude of and temporal change in CO₂ and CH₄ fluxes from different soil types during the snow-covered season, (3) calculate the percentage of the annual flux occurring in winter, and (4) evaluate two methods of estimating gas flux through deep snowpacks. Because the study site has topographic, vegetative, and climatic characteristics typical of many high-elevation and northern ecosystems, the results obtained from this study should have broad applicability to snow-covered terrains. *Mosier et al.* [1993] noted that the global extent of subalpine ecosystems is not well defined but suggested that forest-meadow systems with similar characteristics may account for as much as 12×10^6 km² of the Earth's land surface.

2. Study Site Description

The study was conducted in the Loch Vale watershed (latitude 40°18', longitude 105°40') located in Rocky Mountain National Park, approximately 80 km northwest of Denver, Colorado. The site is designated as a United Nations Biosphere Reserve [Baron, 1992] and currently is one of five sites funded by the U.S. Geological Survey Water, Energy, and Biogeochemical Budgets Program to conduct field-oriented research on the movement of water and carbon between the land and atmosphere [Lins, 1994]. Loch Vale is a subalpine/alpine basin that drains the eastern side of the Continental Divide and ranges in elevation from 3000 to 4000 m. Land cover types in the basin include exposed bedrock surfaces, talus slopes, alpine tundra, and subalpine forests and wetlands. Climate is characterized by long, cold winters and a short, 2- to 3-month growing season. Average daily air temperatures range from -6.0°C in winter to 13.7°C in summer [Baron, 1992]. Average annual precipitation is 100 cm, and as much as 70% accumulates in a seasonal snowpack between November and April. Snowmelt generally begins in early May, and snow cover may persist at higher elevations into July.

Gas flux measurements were made at an elevation of 3200 m in a small (0.4 ha) subalpine wetland and in a forested area near the

wetland. Vegetation at the wetland site grades from drier grassy areas or subalpine forest around the perimeter to sedge (*Carex utriculata* and *Carex aquatilis*) and arctic rush (*Juncus articus*) toward the wetter center. Transitional zones along the edge of the wetland are covered by moss (*Sphagnum* spp.), tufted hairgrass (*Deschampsia cespitosa*), and alpine timothy (*Phleum commutatum*). Soils in the center of the wetland are anoxic near the surface and consist of a thick (at least 175 cm) peat layer that is covered with 5 to 10 cm of water during the growing season. Forest vegetation is old-growth Englemann spruce (*Picea engelmannii* Pargy) and subalpine fir (*Abies lasiocarpa* Hook) with a sparse ground cover of myrtleleaf blueberry (*Vaccinium myrtillus*) [Arthur, 1992]. Forest soils are well-drained Cryoborals typically less than 50 cm in depth [Baron, 1992].

3. Methods

Winter gas fluxes were estimated using snowpack properties and concentration gradients of CO₂ and CH₄ in the snowpack in 1994, 1995, and 1996. In 1994, gas concentrations were measured monthly to bimonthly at one wetland site and one forested site from January through May. In 1995, sampling began just prior to snowmelt in early May and continued biweekly through the end of June. In addition to the two sites established in 1994, measurements were made at four additional sites in the wetland. In 1996, measurement efforts focused on the wetland. Ten sites were established along a 120-m transect that crossed a range of soil moisture conditions from drier areas along the wetland perimeter to saturated areas in the center of the wetland. Measurements were made approximately monthly from January through April and biweekly through the snowmelt period.

In 1994, snowpack gases were sampled through fixed-interval samplers installed before significant snow accumulation had occurred. The sampler consisted of 25-mm-diameter sampling ports, covered with stainless steel screens, that were suspended above the ground on metal arms that extended out 35 cm from a central PVC pole. Pairs of sampling ports were placed at the soil surface and at 45-cm intervals in the snowpack. The sampling ports were accessed by nylon tubing (2-mm outer diameter) that ran vertically along the central PVC pole to the snow surface. Snowpack gases also were sampled just below the snow surface through a thin metal tube inserted into the snow. In spring 1994 and in 1995 and 1996, snowpack gases were sampled through a 10-mm-diameter stainless steel probe 4 m in length that was inserted to different depths in the snowpack. The probe had an inner bore diameter of 3 mm and was slotted over a 5-cm interval near the tip. The probe and fixed-interval sampler produced similar results, but the probe had the added advantage of not being restricted to a fixed site location and sampling interval.

CO₂ was measured in the field using a portable infrared gas analyzer (IRGA) that was calibrated with 350 and 2010 ppmv standards carried into the field in mylar balloons. The precision of the field IRGA measurements decreased with increasing concentration from 5% at 350 ppmv to less than 1% for concentrations over 2000 ppmv. The IRGA was equipped with an internal pump that pulled gas from a sampling line attached to the probe at a maximum rate of 0.3 L min⁻¹. During the measurements, gas from the snowpack was pulled through the probe by the IRGA pump until the concentration stabilized (usually 15 to 30 s). After taking

the CO₂ reading, the sampling line was attached to a 20-cm³ nylon syringe in which a sample was collected after purging the syringe three times with air from the sampling line. The syringes were returned to the laboratory for CH₄ analysis by gas chromatography, which was done within 48 hours of collection. Gas standards collected in the syringes showed no significant change in concentration during a week of storage in the laboratory or during transport from the field site to the laboratory, which was 1370 m lower in elevation. The gas chromatograph had a detection limit of 0.1 ppmv CH₄ and a precision of 2.5% for concentrations below 2.0 ppmv and 1% for concentrations above this level.

Physical snowpack measurements were made at the wetland site within 2 days of each gas collection date. The water content of the snowpack was measured from a vertical snowpit face using a 1-L stainless steel cutter, and snow temperatures were measured every 10 cm in the pit wall using a dial stem thermometer (precision ±0.5°C). Temperatures were measured hourly at depths of 1, 5, and 10 cm below the soil surface at one site in the forest from March 1994 through August 1995 and at two sites in the wetland (one at the center and one at the perimeter) from October 1995 through the end of the study, using permanently installed thermometers (thermocouples were used at the forest site) connected to data loggers. Soil water content was measured hourly at the perimeter of the wetland from October 1995 through June 1997, using a 15-cm-long water content reflectometer (probe CS615, Campbell Scientific, Inc.) inserted vertically into the soil. Air temperature and snow depth were measured continuously at a weather station located in the wetland, and barometric pressure was measured at a weather station located 300 m from the wetland at an elevation of 3150 m.

Fluxes of CO₂ and CH₄ from the snow were calculated from gas concentrations in the snowpack using a one-dimensional form of Fick's law describing ordinary gaseous diffusion in porous unsaturated media as discussed by *Striegel* [1993],

$$q_A = -D_{eff} \frac{dC_A}{dz} \quad (1)$$

where q_A is the mass flux of gas A (mol m⁻² d⁻¹), D_{eff} is the effective diffusion constant of gas A through the snowpack (m² d⁻¹); and dC_A/dz is the measured concentration gradient of gas A in the snowpack (mol m⁻³ m⁻¹).

The effective diffusion constants D_{eff} for CO₂ and CH₄ were estimated on each sampling date using the relation,

$$D_{eff} = \theta \tau D_A \left(\frac{1}{p} \right) \left(\frac{T + 273.15}{273.15} \right)^{1.823} \quad (2)$$

where θ is the snowpack porosity, τ is the tortuosity coefficient, and D_A is the diffusion constant for gas A through air at STP, corrected for the average daily barometric pressure (atmospheres) and snow temperature (degrees Celsius). Porosity was calculated from the density of ice ($\rho_{ice}=0.91$) and the water content of the snowpack over the gradient interval. Tortuosity is difficult to measure and, in soils, usually is described as a function of porosity with values ranging from $\theta^{1/3}$ to $\theta^{2/3}$ [*Striegel*, 1993]. In this study, the tortuosity of the snowpack was estimated from 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 0.74 to 0.90 reported by *Sommerfeld et al.* [1996] for a subalpine snowpack in Wyoming. A recently developed method for measuring the

tortuosity of snow from images of cut, polished snowpack sections yielded tortuosity values of about 0.9 for snowpacks having porosities of about 80% (*R. Davis*, Cold Regions Research and Engineering Laboratory, written communication, 1997). Table 1 lists the effective diffusion coefficients used to calculate CO₂ and CH₄ fluxes on each sampling date. Calculated diffusivities for CO₂ decreased from 1.25 m² d⁻¹ in midwinter to 0.53 m² d⁻¹ in late spring, which primarily resulted from a reduction in porosity of the snowpack over the winter. The midwinter values are slightly higher than the range of 0.70 to 0.86 m² d⁻¹ reported by *Winston et al.* [1995] for CO₂ diffusion through a midwinter snowpack in a boreal forest. The higher values calculated for Loch Vale are mostly due to the lower atmospheric pressure at this site, which increases the diffusivity of CO₂ through air by 30% compared to sea level. *Zimov et al.* [1993] reported experimentally determined diffusivities ranging from 1.90 to 2.86 m² d⁻¹ for wind-packed snow in Siberia. The values reported by *Zimov et al.* [1993] are considerably higher than the diffusivity of CO₂ through air, suggesting that their experimental design measured advective as well as diffusive movement of CO₂ through the snow.

CO₂ flux measurements also were made at the snowpack surface using a static chamber technique on selected dates. The 55 × 65 cm rectangular chamber having a height of 15 cm was inserted 5 cm into the snow surface during measurements. Air inside the chamber was mixed with a small pump, and CO₂ concentrations were measured by circulating the chamber air through the IRGA. CO₂ concentrations were monitored every 60 s for up to 30 min. The maximum change in concentration observed in the chamber headspace was 152 ppmv over a 16-min interval. Flux was estimated from the linear portion of the concentration versus time curve, which usually included the first 8 min of concentration data. For all chamber measurements, the r^2 of the linear regression was greater than 0.98. Fitting the data with a nonlinear, diffusion-based model, as suggested by *Anthony et al.* [1995], did not significantly change the flux estimates.

4. Results

4.1. Snowpack and Soil Characteristics

During the 1993-1994 winter, a seasonal snowpack began to accumulate at the study site in early November that remained fairly shallow (< 50 cm) through the end of December. Owing to cold air temperatures and shallow snow depths in early winter, a layer of temperature gradient (TG) snow developed at the base of the snowpack. The snowpack, which reached a maximum depth of 190 cm at the weather station, became isothermal between the April 20 and May 5 sampling dates, and snow covered the study site until early June. Snowpack characteristics in 1994-1995 were substantially different than in 1993-1994 owing to an unusually warm period in early spring. Average daily air temperature at the wetland was above 0°C at least one half of the days between March 6 and April 7, 1995, and snow accumulation during this period was minimal. Cool, wet weather followed in April and May, and the snowpack nearly doubled in depth during this period. Owing to the warm weather in early spring, a laterally continuous, 2- to 5-cm thick ice layer developed at the snow surface that was subsequently covered by nearly 1 m of late-season snow. Maximum snow depth was

Table 1. Average Daily Barometric Pressure, Air Temperature, Snow Depth, Snowpack Temperature, Snowpack Porosity, Calculated Tortuosities and Effective Diffusivities D_{eff} of CO₂ and CH₄ for Each Sampling Date

Date	Atmospheric Pressure, atm	Air Temperature, °C	Snow Depth, cm	Snow Temperature, °C	θ	τ	$D_{\text{eff}} \text{ CO}_2$, m ² d ⁻¹	$D_{\text{eff}} \text{ CH}_4$, m ² d ⁻¹
Jan. 10, 1994	0.672	-11.8	95	-3.4	0.78	0.92	1.25	1.76
Feb. 15, 1994	0.681	-4.8	135	-3.2	0.71	0.89	1.10	1.55
March 10, 1994	0.679	-4.3	155	-2.7	0.68	0.88	1.04	1.46
April 1, 1994	0.681	-0.7	170	-1.1	0.63	0.86	0.95	1.33
April 20, 1994	0.689	5.0	160	-0.3	0.54	0.82	0.77	1.09
May 5, 1994	0.689	7.3	160	-0.3	0.52	0.81	0.74	1.04
May 25, 1994	0.689	2.9	65	-0.3	0.53	0.81	0.74	1.05
May 10, 1995	--	-0.1	205	-1.0	0.75	0.91	1.20	1.69
June 1, 1995	0.648	4.4	210	-0.4	0.49	0.79	0.71	1.01
June 18, 1995	0.689	11.5	160	0.0	0.47	0.78	.64	.90
June 22, 1995*	0.689	7.8	105	0.0	0.47	0.78	.64	.90
Jan. 9, 1996	0.683	-2.2	130	-5.1	0.67	0.88	1.00	1.40
Feb. 5, 1996	0.676	-7.9	225	-3.8	0.64	0.86	.95	1.34
March 20, 1996*	0.663	-10.2	310	-2.0	0.55	0.82	.80	1.13
April, 8, 1996	0.681	0.8	290	-0.7	0.53	.81	.75	1.06
May 7, 1996	0.684	6.6	280	-0.2	0.49	0.79	.68	.95
May 15, 1996	0.686	9.5	240	0.0	0.41	0.74	.53	.75
May 21, 1996	0.683	0.8	205	0.0	0.41	0.74	.54	.75
May 31, 1996	0.684	7.7	200	0.0	0.42	0.75	.53	.78
June 4, 1996	0.692	8.0	170	0.0	0.42	0.75	.53	.78
June 11, 1996*	0.691	11.4	115	0.0	0.42	0.75	.53	.78
June 18, 1996*	0.692	11.4	75	0.0	0.42	0.75	.53	.78

*Snowpack temperature and porosity estimated for these dates.

210 cm on May 15, 1995; cool air temperatures delayed the onset of snowmelt until the end of May, and snow covered the wetland until early July. During the 1995-1996 winter, a seasonal snowpack began to accumulate the first week of November and increased slowly until the middle of January when several storm systems deposited more than 50 cm of snow during the following 2-week period. This rapid accumulation of snow triggered a large avalanche on January 30, 1996, that destroyed the forest on the edge of the wetland and covered the field site with avalanche debris. The snowpack, which reached a maximum depth of 300 cm on April 23, 1995, was isothermal on the May 7 sampling date, and snow persisted at the wetland until the third week of June.

Soils in the center of the wetland remained saturated and just above 0°C through the snow-covered season (Figure 1). In all 3 years, a thick ice layer (10 cm) developed at the surface of the wetland in early November before significant snow accumulation. The ice layer remained intact until the onset of spring snowmelt when it began to melt under the snowpack. Soil temperatures at sites in the forest and along the wetland perimeter also remained close to 0°C through the snow-covered season. The record of soil water content at the wetland perimeter is discontinuous owing to equipment failure and destruction of the instrument installation by the January 1996 avalanche. The most continuous part of the

record begins in April 1996 at the beginning of the snowmelt period (Figure 1). Melting of the overlying snowpack caused soil water content to increase until the site was free of snow. The soil dried over the course of the summer until several heavy rain events rewet the soil in early September. Seasonal snow cover was re-established at the beginning of November 1996, after which soil water content increased gradually through the early winter months. As in the previous year, soil water content increased rapidly once snowmelt began.

4.2. Snowpack CO₂ and CH₄ Concentrations

Snowpack CO₂ and CH₄ concentrations varied temporally and spatially during the period of study. Figure 2 shows snowpack CO₂ profiles at a location near the perimeter of the wetland through the 1996 snow-covered season. These profiles are similar to those observed at the other sampling sites, although the slopes and shapes of the profiles varied from site to site depending on snowpack properties and underlying soil characteristics. Concentrations of CO₂ at all sites increased toward the base of the snowpack, indicating that soils were a source of CO₂ to the atmosphere throughout the snow-covered period. Concentrations at the base of the snowpack generally increased through the winter until the

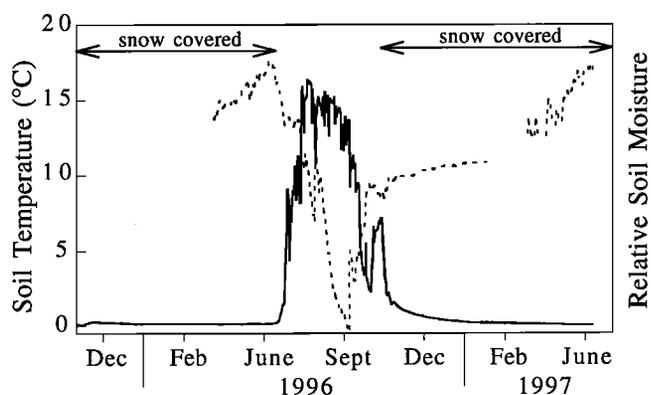


Figure 1. Daily soil temperature at 5 cm depth (solid line) in the center of the wetland and soil moisture content of the upper 15 cm (dashed line) at a site along the wetland perimeter.

onset of snowmelt. Maximum CO₂ concentrations at the snowpack base in 1996 ranged from 2900 ppmv in the center of the wetland to 11,550 ppmv in a grassy area adjacent to the wetland. These concentrations are comparable to those reported for snowpacks covering subalpine and montane soils [Solomon and Cerling, 1987; Sommerfeld *et al.*, 1996] but higher than those reported for snowpacks in alpine and arctic environments [Brooks *et al.*, 1997; Zimov *et al.*, 1993]. Figure 2 also shows a seasonal pattern in the CO₂ concentration gradients (dC/dz) that increased through the midwinter months (solid lines), reaching a maximum on May 7 at the onset of snowmelt. These changes in gradient reflect a combination of changes in snowpack density and in gas fluxes from the underlying soils.

Snowpack CH₄ concentrations also increased toward the bottom of the snowpack at sampling locations in the wetland, indicating that wetland soils were a net source of CH₄ to the atmosphere through the winter (Figure 3). As observed for CO₂, concentrations of CH₄ at the base of the snowpack generally increased through the winter until the onset of spring snowmelt. Maximum CH₄ concentrations at the soil/snow interface at sites within the wetland ranged from 34 to 360 ppmv, which are comparable to concentrations reported in a shallow snowpack overlying a temperate peatland in New Hampshire [Melloh and Crill, 1996]. The change in CH₄ gradient over the season was similar to CO₂, with the gradients showing a pronounced increase prior to the May 7 sampling date. CH₄ concentrations over drier sites in the forest and along the wetland perimeter typically decreased toward the snowpack base during the midwinter sampling dates, indicating that these soils were a sink of atmospheric CH₄. Two sites at the wetland perimeter in 1995 switched from net CH₄ consumption to net production after soils became saturated by snowmelt, as evidenced by the reversal in gradient between the June 1 and 22 sampling dates (Figure 4). The reason for the apparent midpack gradient reversal on June 18 is unclear. It may have been caused by horizontal diffusion of CH₄ along ice layers from high concentration zones in the center of the wetland or, alternatively, to disturbance of the concentration gradient by a transient process such as wind pumping [Massman *et al.*, 1997].

Large spatial variations in snowpack CO₂ and CH₄ concentrations were observed across the wetland transect as illustrated by a

contour plot of concentrations in the snowpack on March 30, 1996, just prior to peak snow accumulation for the year (Figure 5). Snowpack depths ranged from 230 to 370 cm across the wetland owing to redistribution of snow by wind and the January avalanche. CO₂ and CH₄ concentrations in the snowpack varied significantly across the wetland. Although some of the variation may be related to snow depth and density, the pattern appears to correlate most strongly with soil moisture. Soil moisture data across the transect are available only during the summer months [Wickland, 1997], but the relative differences between sites are not expected to vary significantly among seasons. The highest CO₂ and lowest CH₄ concentrations were measured from dry soils on a grass-covered rise at the edge of the wetland (left side of Figure 5). Soils at this site have an organic matter content of 25% and an average volumetric water content during snow-free periods of 35% in the upper 10 cm [Wickland, 1997]. The highest CH₄ concentrations and lowest CO₂ concentrations were measured in the center of the wetland where soils are covered by standing water during snow-free months. Vegetation at these locations is dominated by sedge and arctic rush, and soils have an average organic matter content of 77%. CO₂ and CH₄ concentrations at the margins of the wetland were intermediate between the dry and saturated soils. Soils in these areas are moist, having an average volumetric water content of 72%, and organic matter contents ranging from 71 to 77%. Vegetation at these locations is predominantly moss, tufted hairgrass, and alpine timothy.

4.3. Winter CO₂ and CH₄ Fluxes

Gas fluxes through the snowpack were calculated for each site using (1), where the concentration gradient (dC/dz) was determined from a linear fit of the snowpack concentration profiles. Examples of the most common profile shapes observed during the study period are shown in Figure 6. Profiles similar to the curve shown in Figure 6a were most commonly observed during

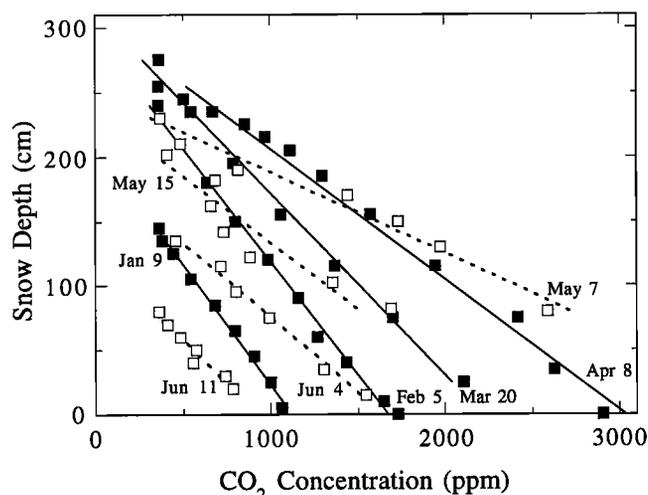


Figure 2. CO₂ snowpack concentrations during the 1996 snow-covered season at a site along the wetland perimeter. Solid and dashed lines represent profiles measured before and after the onset of spring snowmelt, respectively.

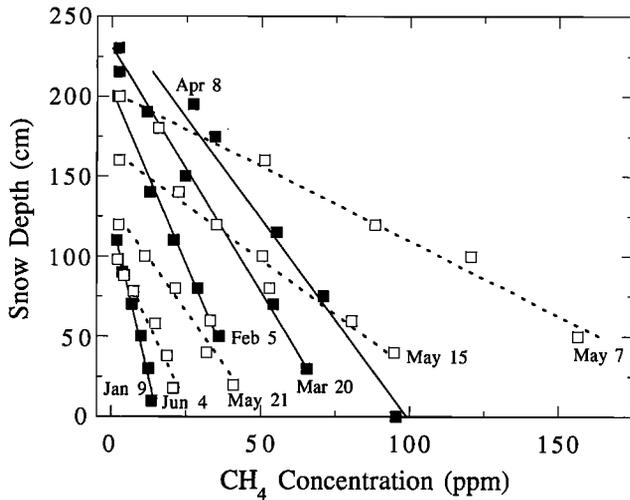


Figure 3. CH₄ snowpack concentrations during the 1996 snow-covered season at a site in the center of the wetland. Solid and dashed lines represent profiles measured before and after the onset of spring snowmelt, respectively.

midwinter conditions when the snowpack density and stratigraphy were relatively homogeneous. In these cases, the concentration gradient was calculated from a linear fit of the entire concentration profile. Curvilinear profiles as shown in Figure 6b were observed during the 1996 snow-covered period as a result of increasing snow density with depth. In these cases, the concentration gradients were calculated from the upper half of the snowpack where the profiles were close to linear. Although not used in this study, a second approach would be to correct the snowpack concentrations for changes in snow porosity with depth using a finite difference model [Ishii *et al.*, 1989] before fitting the linear regression. A midpack ice layer, which developed in spring 1995, restricted gas diffusion through the snow, causing the stepped profile shown in Figure 6c. For these profiles, the concentration gradient was estimated from the portion of the gradient above the ice layer. For the nonlinear profiles shown in Figures 6b and 6c, inclusion of all points in the gradient calculation or a two-point calculation using base of snow and snow surface concentrations would have resulted in an overestimation of gas fluxes by as much as 25% because concentrations toward the base of the snowpack were elevated by variations in snowpack structure and density. Gradient estimates of CO₂ and CH₄ fluxes during the snow-covered periods of 1994, 1995, and 1996 are summarized in Table 2. Sample locations were grouped according to relative soil moisture content [Wickland, 1997]. Saturated soils are in areas of the wetland covered by standing water; dry soils include forested and unforested areas outside the wetland, and moist soils are transitional areas along the wetland perimeter. Fluxes on each date represent one to four measurements for each moisture grouping.

Average winter CO₂ emissions from dry, moist, and saturated soils were 42.3 (standard deviation (s.d.) = 17.7), 31.2 (s.d. = 11.5), and 14.6 (s.d. = 10.3) mmol m⁻² d⁻¹, respectively. Fluxes from individual sites over the study period ranged from 16.0 to 76.0 mmol m⁻² d⁻¹ for dry soils, 3.0 to 101 mmol m⁻² d⁻¹ for moist soils, and 1.2 to 36.5 mmol m⁻² d⁻¹ for saturated soils. Fluxes from dry and moist sites were similar to average winter fluxes from subalpine soils (27.5 to 73.0 mmol m⁻² d⁻¹) in the Snowy Range,

Wyoming [Sommerfeld *et al.*, 1996], but slightly greater than fluxes from alpine soils (2.5 to 26.4 mmol m⁻² d⁻¹) on Niwot Ridge, approximately 20 km south of Loch Vale [Brooks *et al.*, 1997]. The flux estimates reported by Sommerfeld *et al.* [1996] may be low because they did not correct the diffusion coefficient for the lower atmospheric pressure at their study site (3200 m). Winter fluxes measured in this study generally were greater than those observed in high-latitude ecosystems. For example, winter fluxes ranged from 5.8 to 21.7 mmol m⁻² d⁻¹ from arctic tundra soils in Siberia [Zimov *et al.*, 1993] and 0.8 to 16.6 mmol m⁻² d⁻¹ from arctic soils on the North Slope, Alaska [Oechel *et al.*, 1997]. Winston *et al.* [1995] reported midwinter CO₂ fluxes ranging from 0.0 to 36.0 mmol m⁻² d⁻¹ from snow-covered soils in the boreal forest in Canada. Temporal patterns in winter CO₂ fluxes in 1994 and 1996 are presented in Figure 7. Fluxes at the moist and saturated sites were lowest in midwinter, peaked at the onset of spring snowmelt, then decreased to near premelt levels through the remainder of the snowmelt period. The largest increase in flux occurred at the moist sites in 1996, which nearly tripled between the February 5 and May 7 sampling dates. The seasonal pattern at dry sites was similar except that the late spring fluxes were significantly lower than the early winter rates. The magnitude of and changes in fluxes at the dry and saturated sites were similar between 1994 and 1996 despite differences in site locations and in the timing and depth of the snowpack.

Average winter CH₄ fluxes from dry, moist, and saturated soils were -0.016 (s.d. = 0.009), 0.274 (s.d. = 0.288), and 2.87 (s.d. = 2.99) mmol m⁻² d⁻¹, respectively. Fluxes from individual sites over the study period ranged from -0.035 to 0.016 mmol m⁻² d⁻¹ for dry soils, -0.014 to 2.08 mmol m⁻² d⁻¹ for moist soils, and 0.046 to 13.9 mmol m⁻² d⁻¹ for saturated soils. CH₄ consumption rates measured over dry soils were similar to the range of -0.004 to -0.016 mmol m⁻² d⁻¹ reported by Sommerfeld *et al.* [1993] for subalpine soils in Wyoming during winter. Fluxes from saturated soils were similar to average winter fluxes from a temperate peatland in New Hampshire, which ranged from 1.3 to 3.5 mmol m⁻² d⁻¹ [Melloh and Crill, 1996], and from an open bog site in Minnesota, which

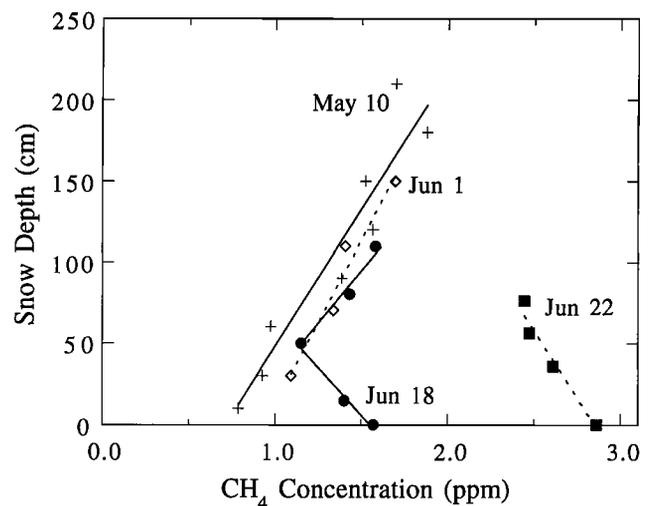


Figure 4. CH₄ snowpack concentration profiles on four dates in 1995 at a site along the wetland perimeter.

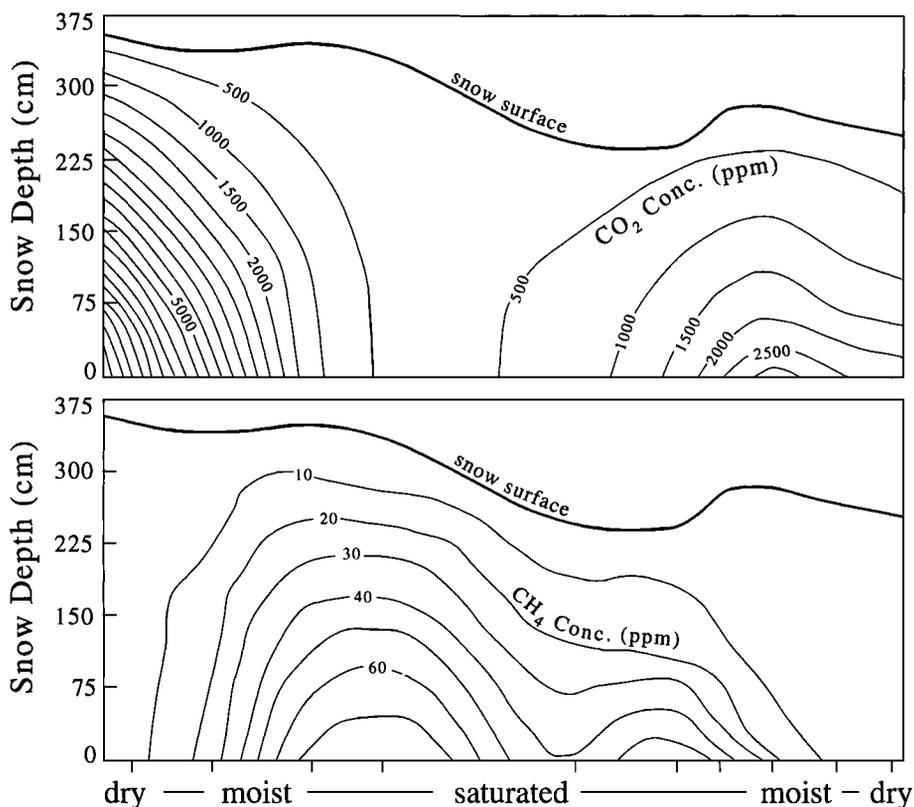


Figure 5. Contour plots of CO₂ and CH₄ snowpack concentrations along a 120 m transect through the wetland on March 30, 1996. Tick marks on the *x* axis represent relative positions of the 10 sampling sites along the transect.

ranged from 0.3 to 3.7 mmol m⁻² d⁻¹ [Dise, 1992]. Temporal patterns in winter CH₄ flux in Loch Vale differed among soil types. At the dry site in 1994 (Figure 8b), CH₄ uptake was greatest during midwinter then declined to near zero during the snowmelt period. The seasonal variation at the dry site in 1996 was less pronounced, although a slight decline in flux was observed at the onset of snowmelt (Figure 8d). Moist and saturated soils showed a slight increase in CH₄ flux through the midwinter months, followed by a dramatic increase in flux at the onset of snowmelt (Figures 8a and 8c). The largest change in flux occurred at the saturated site in 1994, which increased by an order of magnitude between the February 15 and May 25 sampling dates. In 1996, the flux from saturated soils decreased through the remainder of the snowmelt period, aside from a second increase in early June. At the moist sites in 1996, the flux returned to premelt levels following the initial release and then increased slightly through the remainder of the snow-covered period.

On several sampling dates in 1995, gradient and chamber measurements were made at the same locations in order to compare these two methods of estimating gas flux through snow. The results of eight pairs of measurements made over different soil types are presented in Table 3. CO₂ gradients in the snowpack at the eight sites ranged from -27 to -105 mmol m⁻², and the corresponding gradient flux estimates ranged from 17.0 to 62.7 mmol m⁻² d⁻¹. Chamber flux estimates made at the same sites ranged from 5.5 to 21.1 mmol m⁻² d⁻¹ and were consistently lower than the gradient flux estimates by nearly a factor of 3.

5. Discussion

5.1. Effect of Snow Cover on Winter Gas Fluxes

Recent studies have shown that gas transport through seasonal snowpack occurs largely by diffusion and that fluxes at the snowpack surface primarily are controlled by biological activity in the underlying soils [Sommerfeld *et al.*, 1996; Brooks *et al.*, 1997; Winston *et al.*, 1995]. Within-snow processes, such as biological production or consumption, adsorption on ice, and dissolution in water, are thought to have a minor effect on gas exchange between the soil and atmosphere during winter [Sommerfeld *et al.*, 1996; Brooks *et al.*, 1993]. Pressure fluctuations at the snowpack surface caused by wind may enhance gas transport through snow; however, the importance of this process in terms of the total winter flux is poorly understood [Albert and Hardy, 1995; Massman *et al.*, 1995; Massman *et al.*, 1997]. Comparison of snowpack CO₂ profiles and snowpack physical characteristics in Loch Vale showed variations one would expect if gas exchange through the snowpack was controlled primarily by porosity and tortuosity variations within the snowpack (Figure 6). The nearly linear concentration gradient in Figure 6a is typical of most sampling dates and reflects the lack of structure and relatively narrow range of densities in midwinter snowpacks. A more detailed analysis of this profile, however, showed a subtle break in slope at the 60-cm level.

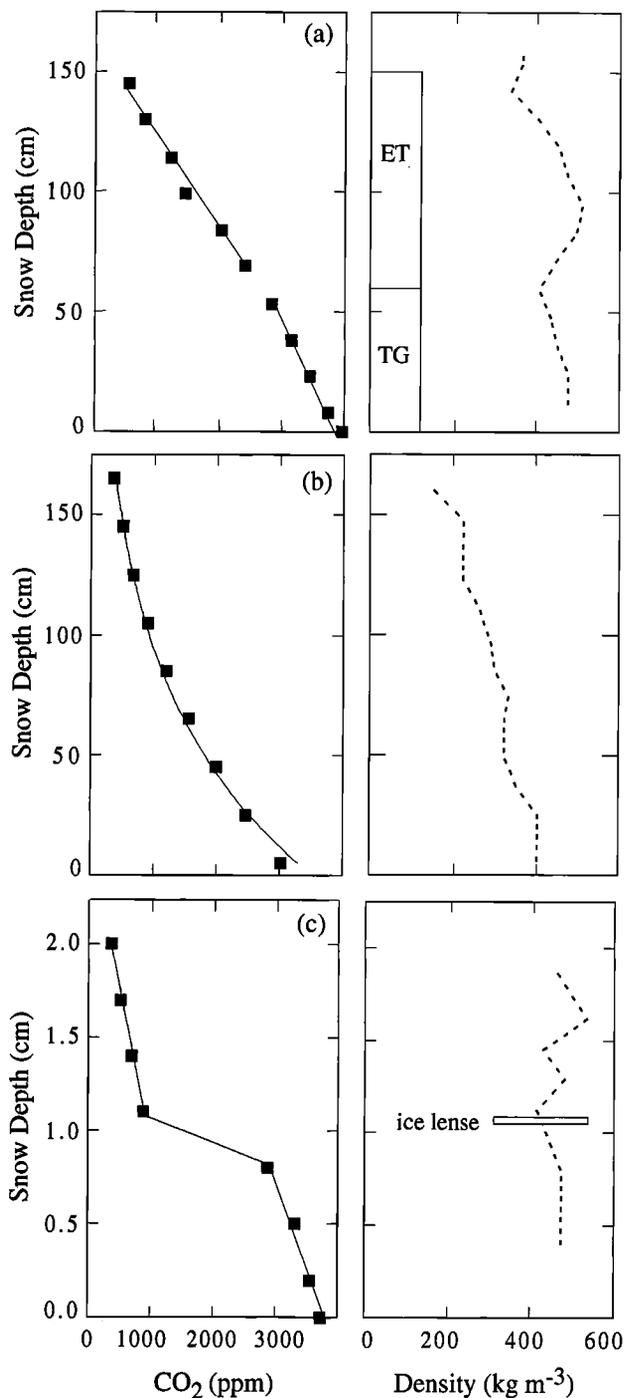


Figure 6. CO₂ snowpack concentration profiles in (a) 1994, (b) 1996, and (c) 1995 with measured snow-density profiles. ET represents a layer of equitemperature snow, and TG a layer of temperature gradient snow.

If separate regression lines are fit to the two regions, the concentration gradient above 60 cm ($-25.5 \text{ ppmv cm}^{-1}$, $p < 0.001$) is slightly greater than the gradient below 60 cm ($-21.0 \text{ ppmv cm}^{-1}$, $p < 0.001$). The change in slope appears to reflect a change in snow metamorphic type rather than snow density. The snowpack stratigraphy indicated a layer of a coarse-grained temperature gradient (TG) snow overlain by a fine-grained equitemperature

(ET) layer. ET metamorphism occurs in response to a vapor-pressure difference on the individual snow grains and results in a strengthening of the snowpack due to the formation of necks between individual grains [Sommerfeld and LaChapelle, 1979]. In contrast, TG metamorphism occurs in response to a temperature gradient across the snowpack that results in a loss in snowpack cohesion as the individual grains enlarge at the expense of the necks [Sommerfeld and LaChapelle, 1979]. This loss of cohesion between grains may have reduced the tortuosity of this layer, which caused the gradient to be slightly shallower in the lower half of the snowpack. Conway and Abrahamson [1984] found a similar relation between air permeability and snow metamorphic type; increasing permeability appeared to be associated with TG metamorphism and decreasing permeability with ET metamorphism. On several sampling dates in 1996, profiles in the wetland had a distinct curvilinear shape as shown by Figure 6b. The snowpack characteristic that appeared to influence the shape of the gradient was a distinct increase in snow density with depth, which probably was caused by the unusually rapid accumulation of snow during the month of January. For a few profiles, the concentrations were adjusted for changes in porosity and tortuosity with depth which caused the profiles to straighten, indicating that snow density was the primary factor controlling the profile shape.

Several studies have suggested that ice layers and melt features within the snowpack may have a significant effect on gas transport from snow-covered soils [Hardy et al., 1995; Winston et al., 1995; Melloh and Crill, 1995, 1996]. An ice layer was observed to have a dramatic effect on gas concentration profiles in Loch Vale during the spring of 1995. The profile in Figure 6c shows a large increase in CO₂ concentration at the same level as a thick ice layer located midway through the snowpack. Concentration gradients above and below the ice layer were linear, although the gradient was slightly steeper below than above the ice layer. This profile demonstrates that the ice layer had a considerably lower diffusivity than the surrounding snow but did not block gas flux from the snowpack surface. The basal ice layer that developed at the wetland surface also was sufficiently permeable to allow continuous gas exchange between the soil and atmosphere as evidenced by the CO₂ and CH₄ gradients measured in the snowpack over the wetland throughout the snow-covered period. Melloh and Crill [1996] also observed the buildup of CH₄ below a midpack ice layer; however, near-ambient snowpack CH₄ concentrations above the layer indicated that the ice inhibited gas flux at the snowpack surface. At a few sites along the wetland perimeter, high CH₄ concentration layers were observed between ice layers that developed in the spring. Because a similar pattern was not observed for CO₂, we believe that these high-concentration layers may have resulted from horizontal diffusion of CH₄ between the ice layers from high concentration zones in the center of the wetland. These results show that snowpack structures can, at least temporarily, affect the exchange of gas between the soil and atmosphere; however, once the snowpack begins to melt, gas trapped beneath or between ice layers will eventually be released [Melloh and Crill, 1996].

Perhaps the most important role the snowpack plays in controlling winter gas production or consumption is providing insulation for soils from cold midwinter air temperatures. Despite average air temperatures in Loch Vale of -9°C from November through February, soil temperatures remained slightly above 0°C and well above the minimum temperature range of -5° to -7°C reported for

Table 2. Average Winter CO₂ and CH₄ Fluxes From Dry, Moist, and Saturated Soils

Date	Average Daily CO ₂ Flux, mmol m ⁻² d ⁻¹					Average Daily CH ₄ Flux, mmol m ⁻² d ⁻¹						
	Dry	<i>n</i>	Moist	<i>n</i>	Saturated	<i>n</i>	Dry	<i>n</i>	Moist	<i>n</i>	Saturated	<i>n</i>
Jan. 10, 1994	34.8	1	--		0.9	1	--		--			
Feb. 15, 1994	29.0	1	--		4.1	1	-0.035	1	--		0.60	1
March 10, 1994	45.6	1	--		4.6	1	-0.029	1	--		1.64	1
April 1, 1994	53.3	1	--		10.4	1	-0.026	1	--		1.39	1
April 20, 1994	56.7	1	--		14.8	1	0.000	1	--		2.92	1
May 5, 1994	71.8	1	--		38.2	1	0.000	1	--		3.63	1
May 25, 1994	16.0	1	--		33.6	1	-0.006	1	--		13.9	1
May 10, 1995	57.3	3	--		20.4	2	-0.026	2			2.84	1
June 1, 1995	47.1	2	29.1	1	12.2	3	-0.011	2	0.53	2	4.56	2
June 18, 1995	47.0	3	38.6	1	32.4	2	-0.015	3	0.16	2	6.10	2
June 22, 1995	52.8	3	26.7	1	17.0	1	-0.021	2	--		1.77	2
Jan. 9, 1996	50.9	1	28.4	4	10.1	4	-0.019	2	.060	4	0.64	4
Feb. 5, 1996	--		12.4	4	13.4	4	--		0.18	4	0.92	4
March 20, 1996	62.2	1	16.5	4	8.7	4	-0.012	2	0.22	4	0.68	4
April 8, 1996	76.0	1	33.7	4	9.6	4	-0.011	1	0.17	4	1.55	4
May 7, 1996	37.0	1	63.0	4	26.0	3	-0.013	1	1.2	4	5.52	3
May 15, 1996	17.7	1	33.5	4	23.0	4	-0.021	1	0.18	4	1.88	4
May 21, 1996	30.2	1	29.8	4	10.4	4	-0.014	1	0.13	4	1.68	4
May 31, 1996	39.1	1	35.5	4	10.6	4	-0.018	1	0.23	4	1.45	4
June 4, 1996	17.4	1	33.5	4	9.6	3	-0.010	1	0.15	4	3.53	3
June 11, 1996	21.0	1	29.2	4	5.3	4	-0.023	1	0.15	4	0.73	4
June 18, 1996	25.8	1	27.5	1	5.1	2	-0.019	1	0.25	1	2.27	2

n denotes number of measurements.

microbial respiration [Flanagan and Bunnell, 1980; Coxson and Parkinson, 1987; Brooks *et al.*, 1996]. Without a snow cover, most surface soils in the study area certainly would freeze and microbial activity would be greatly diminished in winter. The effectiveness of the snowpack in insulating soils in different environments appears to be controlled by snowpack depth and average winter air temperature. In the subalpine zone in Loch Vale, snowpacks generally are deep enough to prevent soils from freezing during the early winter when average daily air temperatures are as low as -9.0°C. Similar results were reported by Sommerfeld *et al.* [1996] for subalpine soils in Wyoming, where soil temperatures remained close to 0°C once a consistent snowpack had been established in early winter. In contrast, Brooks *et al.* [1996] found that alpine soils on Niwot Ridge froze early in the winter owing to a shallow snow cover, and surface soil temperatures fell as low as -14°C. As snow began to accumulate later in winter, the underlying soils warmed to above -5°C and microbial activity was observed. These three Rocky Mountain sites are within 500 m of elevation and probably receive comparable amounts of snowfall. However, the Niwot site is on an alpine ridge that is subject to wind scouring, whereas the Loch Vale and Wyoming sites are in areas that are more conducive to snow accumulation.

At lower elevations, the period of snow cover tends to be less consistent through the winter and the effectiveness of the snowpack as an insulator may vary from year to year. For example, in a temperate peatland in New Hampshire, winters having lower

average air temperatures were found to have higher CH₄ emissions than those having higher air temperatures [Melloh and Crill, 1996]. The authors suggested that a deeper and more continuous snow cover, maintained by lower air temperatures, limited frost penetration, which resulted in a longer winter period of microbial activity in years having lower average air temperatures. Dise [1992] observed a similar relation between frost depth and snow cover in a Minnesota peatland. In a year with heavy snowfall, the peat surface never fully froze, but in a subsequent dry year, the peat froze to a depth of 30 cm at some locations. In high-latitude ecosystems, winter soil temperatures generally are below the -5° to -7°C threshold for biological activity owing to the combination of relatively shallow snowpacks and extremely low air temperatures. In the boreal region of Canada, Winston *et al.* [1995] observed surface soil temperatures as low as -14°C in midwinter, which caused CO₂ production to cease in the shallow soil zone. The frozen surface soils, however, remained gas permeable because of low soil moisture [Wickland and Striegl, 1997], which allowed CO₂ from deeper soil sources to diffuse to the soil surface [Winston *et al.*, 1995]. In arctic ecosystems in Siberia and the North Slope, Alaska, soil temperatures as low as -20°C were observed at soil depths as deep as 100 cm during the midwinter months [Zimov *et al.*, 1993; Oechel *et al.*, 1997]. During this period, soil temperatures were too low for biological activity and CO₂ flux was attributed to physical processes that released CO₂ produced during the earlier part of the cold season [Oechel *et al.*, 1997]. On the North

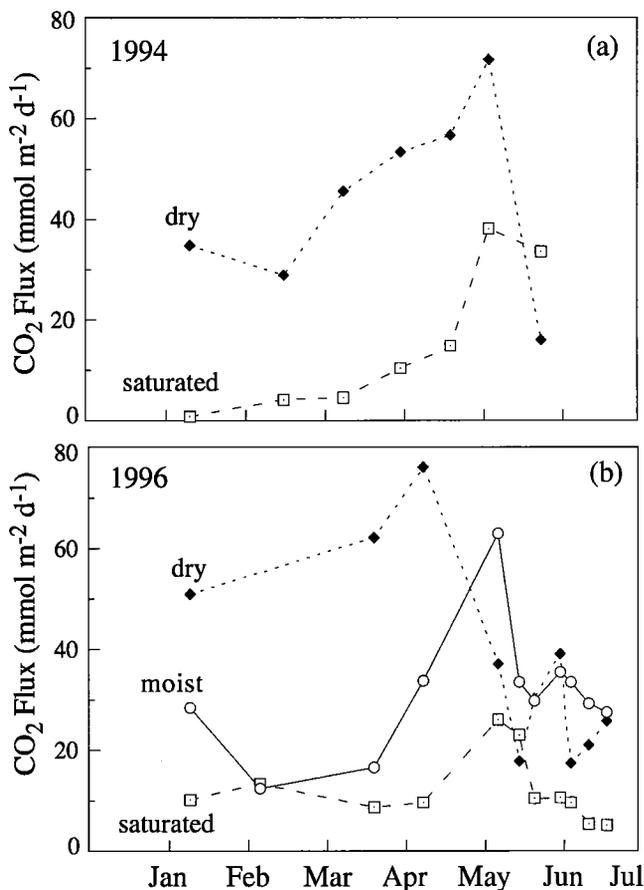


Figure 7. Temporal patterns in winter CO₂ fluxes from dry, moist, and saturated soils in (a) 1994 and (b) 1996.

Slope, Alaska, there is also a deeper geologic source of CO₂ that reaches the surface through fractures in the bedrock [R. G. Striegel and R. W. Healy, U.S. Geological Survey, written communication, 1997].

5.2. Controls on Spatial and Temporal Variations in Winter CO₂ and CH₄ Fluxes

The maintenance of above-freezing soil temperatures by the insulating snowpack is the primary reason that microbial activity persisted through the winter in Loch Vale soils. However, spatial patterns were observed in winter gas fluxes that could not be attributed to the effects of soil temperature, which remained nearly constant through the snow-covered season. The spatial patterns in winter CO₂ and CH₄ fluxes in Loch Vale were similar to those observed at this site during snow-free months and appear to be controlled primarily by soil-moisture conditions [Wickland *et al.*, 1996; Wickland, 1997]. The highest CH₄ flux was found in saturated areas of the wetland where conditions were favorable for methanogenesis. At dry sites outside the wetland, CH₄ was consumed through most of the winter, indicating soils were aerobic and favorable for methanotrophs. The pattern of CO₂ flux across the wetland transect was opposite the pattern of CH₄, with the lowest fluxes in the wetland center and the highest fluxes at the

perimeter. The dominant source of CO₂ in winter from aerobic soils at high elevations is the decomposition of organic matter by heterotrophic microbial communities [Brooks *et al.*, 1997]. Heterotrophic respiration probably is also the source of CO₂ in the center of the wetland, but the fluxes may be lower because less oxygen is available to surface soils in the wetland environment. CO₂ from root respiration in the wetland was assumed to be negligible in the winter.

In addition to spatial trends, CH₄ and CO₂ fluxes in Loch Vale varied temporally through the winter. Most sites had similar patterns for CO₂ and CH₄; fluxes were lowest in early winter, increased to maximum values at the onset of spring snowmelt, then dropped again through the remainder of the snowmelt period. The exception was CH₄ consumption at the dry sites, which was greatest in early winter and declined during snowmelt. Because soil temperatures remained within a very narrow range over the measurement period, other processes are probably controlling the observed temporal trends in emissions. Brooks *et al.* [1997] proposed that increases in winter CO₂ fluxes from alpine soils on Niwot Ridge were driven largely by substrate availability. In this alpine environment, soils froze early in the winter then subsequently thawed with continued snowpack accumulation. Brooks *et al.* [1997] hypothesized that, following the thawing event, labile carbon substrates were released into the soil, which stimulated the growth of microbial populations. This mechanism probably does not apply to the soils in this study because soil temperatures were observed to stay above 0°C through the winter and there was no thawing event to release substrates into the soil.

An early winter minimum in CO₂ fluxes was also observed by Sommerfeld *et al.* [1996], who suggested that temporal trends in winter gas fluxes might be driven by changes in soil moisture. Data from the soil moisture probe at the edge of the wetland indicate that soil moisture during the winter was at a minimum at the time a permanent snow cover was established in early November, then increased gradually through the midwinter until the onset of snowmelt in early May (Figure 1). This is consistent with the early-winter minimum and gradual increase in CO₂ flux prior to snowmelt at the dry and moist sites (Figure 7). The mechanism for the midwinter increase in soil moisture is unclear since free water typically is not present in the snowpack until late April or early May. Sommerfeld *et al.* [1996] hypothesized that changes in soil moisture in early winter might be driven by temperature gradients in the snowpack that cause vapor to be transported from the soil toward the snowpack. Zimov *et al.* [1993] observed an increase in soil moisture in the upper layer of arctic soils in early winter that was attributed to the upward migration of water from deeper soil horizons toward the frozen surface.

Once the snowpack becomes isothermal in spring, there is a large influx of meltwater into the soil environment. At dry sites, CO₂ fluxes decreased significantly after the onset of snowmelt, perhaps because diffusion of oxygen and CO₂ into and out of the saturated soils was limited. The onset of snowmelt also appeared to cause CH₄ consumption to slow in aerobic soils (Figure 8), and, in some cases, soils became net CH₄ producers after being saturated by snowmelt (Figure 4). Mosier *et al.* [1993] noted the opposite pattern in subalpine meadow soils during snow-free months where drying after snowmelt caused soils to switch from net CH₄ production to CH₄ consumption. In the center of the wetland, we observed large releases of CH₄ that were coincident with the onset of snowmelt (Figures 8a and 8c). We believe this episodic release

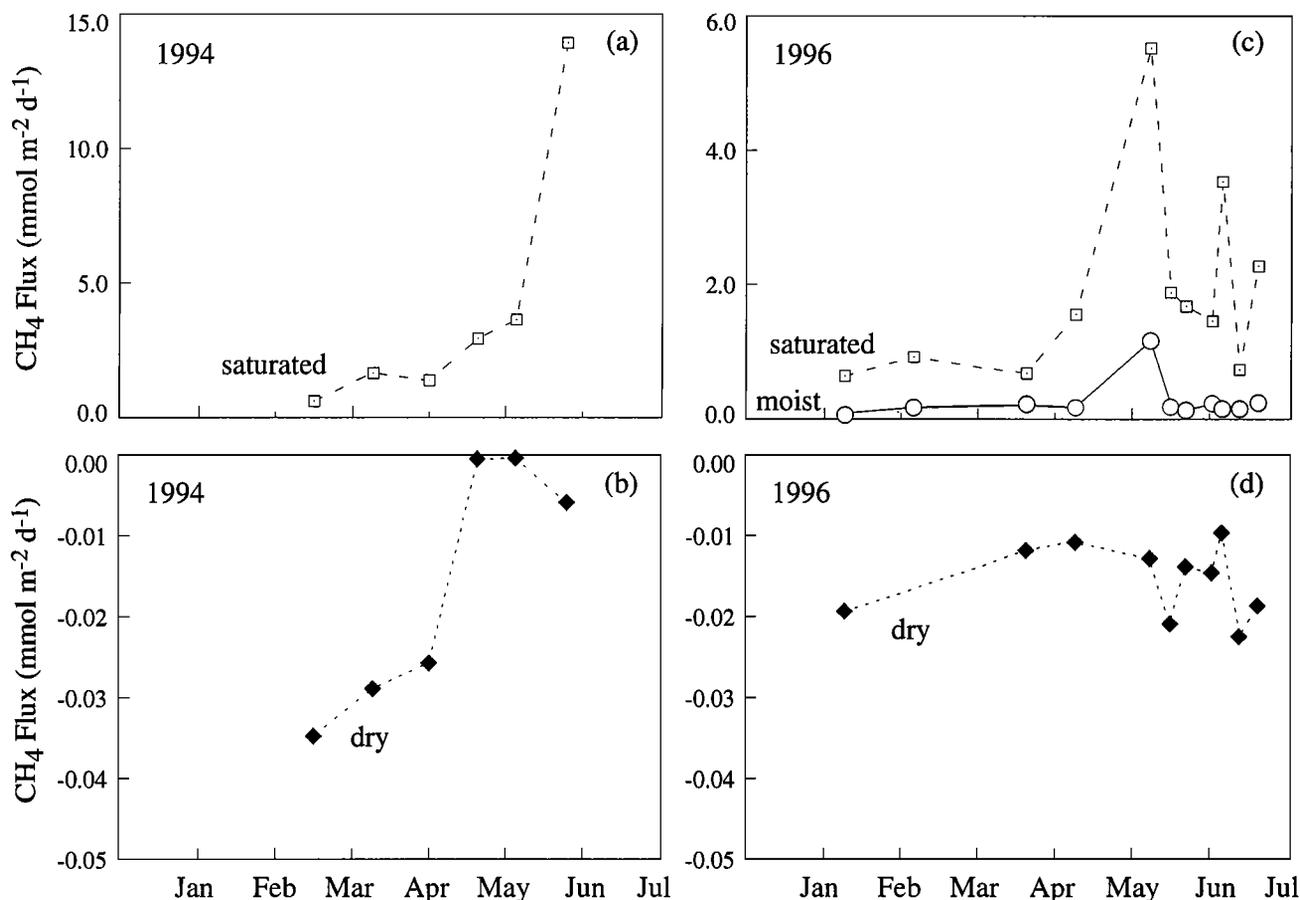


Figure 8. Temporal patterns in winter CH₄ fluxes from (a) saturated and (b) dry soils in 1994, and from (c) moist and saturated, and (d) dry soils in 1996.

of CH₄ may be related to hydrologic changes within the wetland rather than increases in microbial activity. One possible mechanism for episodic release is melting of the ice layer on the wetland surface and subsequent release of gas trapped beneath it. This is consistent with our snowpit observations that the basal ice layer began melting shortly after the snowpack became isothermal. Another hydrologic process that might cause episodic release is the influx of meltwater into the shallow groundwater system. Infiltrating snowmelt from talus slopes above the wetland might push older, CH₄-rich waters to the wetland surface, causing subsequent degassing and release of CH₄ into the overlying snowpack. This hypothesis is supported by isotopic and geochemical studies in Loch Vale, which show that streamflow during the initial stages of snowmelt is derived primarily from premelt groundwaters pushed into the stream channel by infiltrating snowmelt [Mast et al., 1995; Campbell et al., 1995]. Following the episodic release of CH₄, fluxes declined at both the moist and saturated sites through the remainder of the snowmelt period. This decline may be the result of the influx of oxygenated snowmelt into the anaerobic wetland soils that should cause CH₄ production to decrease and CH₄ consumption to increase.

5.3. Winter Gas Fluxes as a Percentage of Annual Budgets

Several recent studies have determined that winter fluxes of CH₄ and CO₂ from snow-covered soils comprise an important part

of the annual budgets of these gases [Dise, 1992; Sommerfeld et al., 1993; Brooks et al., 1997; Oechel et al., 1997]. In Loch Vale, total winter CO₂ emissions in 1995-1996 (November 1 to June 18) for dry, moist, and saturated soils were 11.4, 6.5, and 2.7 mol m⁻², which accounted for 23, 12, and 8%, respectively, of the gross annual CO₂ emissions from these soils [Striegl et al., 1996; K. P. Wickland, unpublished data, 1997]. Total winter CO₂

Table 3. Comparison of CO₂ Flux Estimates Based on Concentration Gradients in the Snowpack (F_g) and Chamber Measurements (F_c) Made on the Snowpack Surface

Date	F_g	F_c	F_g/F_c
May 10	59.8	19.6	3.1
May 10	55.6	18.2	3.1
June 1	20.6	8.6	2.4
June 22	31.3	10.0	3.1
June 22	62.7	18.8	3.3
June 22	57.2	21.1	2.7
June 22	17.0	5.5	3.1
June 22	26.7	10.2	2.6

F_c and F_g in units of $\text{mmoles m}^{-2} \text{d}^{-1}$.

emissions at other high-elevation sites in the Rocky Mountains range from 1.1 mol m⁻² for alpine soils [Brooks *et al.*, 1997] to 10.9 mol m⁻² for subalpine soils [Sommerfeld *et al.*, 1993], which accounts for as much as 25% of the carbon fixed by primary producers during the growing season [Brooks *et al.*, 1997; Sommerfeld *et al.*, 1993]. Total winter CH₄ fluxes in Loch Vale from dry, moist, and saturated soils were -3.7, 50.3, and 315 mmol m⁻², respectively. Using mean daily flux values from Wickland *et al.* [1998] for snow-free months, winter fluxes were calculated to account for 12% of the annual CH₄ consumption by dry soils, and 58 and 12% of the annual CH₄ produced by moist and saturated soils, respectively. The differences in the percentage of winter CH₄ flux among these three soil types appear to be related to the response of microbial activity to soil temperature during snow-free months [Wickland, 1997]. The results for saturated soils in Loch Vale are similar to those at a Minnesota peatland, where winter fluxes accounted for as much as 21% of the annual CH₄ flux [Dise, 1992], but slightly higher than at a temperate peatland in New Hampshire, where only 4% of annual CH₄ emissions occurred under the snowpack [Melloh and Crill, 1996].

5.4. Comparison of Gradient and Chamber Flux Estimates

There is no widely accepted method for estimating gas flux through seasonal snowpacks, and most previous studies have utilized one of the following two techniques: (1) measurement of snowpack concentration gradients and application of a one-dimensional flux model [Solomon and Cerling, 1987; Sommerfeld *et al.*, 1993] or (2) direct measurement of flux at the snow surface using a closed-chamber technique [Whalen and Reeburgh, 1988; Winston *et al.*, 1995]. Fluxes in this study were determined by measuring snowpack concentration gradients; however, several closed-chamber measurements were made in 1995 in an effort to compare these two methods of measurement. The data presented in Table 3, which compare gradient and chamber flux estimates at eight locations, show that the chamber fluxes were consistently lower than the gradient fluxes by nearly a factor of 3. Because uncertainties in θ , τ , and gas concentrations generally do not introduce more than a 15% error in gradient flux estimates [Sommerfeld *et al.*, 1996], the discrepancy between these two methods indicates that chamber measurements at the snow surface probably underestimate gas flux through snow. One possible explanation for lower chamber fluxes is that an adequate seal was not achieved between the chamber and the snow surface [Livingston and Hutchinson, 1994]. On a few occasions, we observed that the concentration in the chamber headspace was reset to ambient levels by strong gusts of wind, which indicated that the seal was not adequate on windy days. Dise [1992] reported that an adequate seal could not be made between the chamber and snow surface during winter CH₄ flux measurements at a Minnesota peatland. Whalen and Reeburgh [1988] used chambers to measure winter CH₄ fluxes from subarctic muskeg communities. To establish an airtight seal, permanent chamber bases were installed in the soil to which vertical sections were added through the winter to stay even with the snow surface. While this technique appears to work well in areas with shallow snowpacks, it would be impractical in the Rocky Mountains, where seasonal snowpacks typically reach several meters in depth.

Several studies have reported that chamber measurements made at the soil surface systematically underestimate true fluxes because increasing concentrations in the chamber headspace distort the concentration gradient near the soil surface [Hutchinson and Mosier, 1984; Jury *et al.*, 1982; Mosier, 1989; Livingston and Hutchinson, 1994]. Using a 3-D simulation, Healy *et al.* [1996] demonstrated that chamber measurements made at the surface of porous media consistently underestimated true fluxes and that the bias increased with increasing air-filled porosity. For example, their model predicted ratios of field-based chamber fluxes to true fluxes ranging from 0.955 for a porosity of 0.1 to 0.435 for a porosity of 0.5. The average chamber flux F_c to gradient flux F_g ratio of 0.372 (s.d. = 0.095) calculated from our field data agrees fairly well with the ratio of 0.435 predicted by Healy *et al.* [1996] for a porosity of 0.5. Because the snowpack porosity averaged 0.52 during our measurements, the model appears to provide a reasonable explanation for the discrepancy between the two methods of measurement. Winston *et al.* [1995] also compared chamber and gradient estimates of CO₂ flux through snow but observed a wide range of flux ratios ($F_c/F_g = 0.2$ to 5.5), some of which indicated the chamber fluxes were significantly higher than the gradient fluxes. The higher chamber fluxes were attributed to the presence of melt channels and tree wells that may have preferentially channeled gas from the soil to the snow surface [Winston *et al.*, 1995; Hardy *et al.*, 1995]. These results suggest that gradient measurements may be inaccurate under some conditions, particularly in areas with shallow or heterogeneous snowpacks.

6. Summary and Conclusions

CO₂ and CH₄ fluxes were measured from snow-covered soils in and adjacent to a subalpine wetland to investigate the processes controlling the exchange of gas between the soil and atmosphere during winter. Gas concentration profiles in the snowpack varied with the physical properties of the snow. Ice layers appeared to slow the diffusion of gas through the snowpack but did not inhibit gas flux at the snow surface. The most important role the snowpack played in controlling winter CO₂ and CH₄ fluxes was in providing insulation for soils from cold midwinter air temperatures, which allowed microbial activity to continue through the snow-covered season. Soil moisture and hydrology appeared to be the principal factors controlling CO₂ and CH₄ fluxes during the winter in Loch Vale. Spatial patterns in CO₂ and CH₄ emissions along the wetland transect were consistent with the soil-moisture gradient across the wetland. Temporal changes in CO₂ flux from aerobic soils may have been driven by gradual increases in soil moisture prior to snowmelt. Episodic release of CH₄ from the wetland was probably related to hydrologic changes in the wetland rather than an increase in microbial activity. Winter fluxes of CO₂ and CH₄ were found to account for a significant percentage of the annual budgets of these gases. Winter CO₂ emissions in 1995-1996 accounted for 23, 12, and 8% of the gross annual CO₂ emissions from dry, moist, and wet soils, respectively. Winter microbial activity accounted for 12% of the annual CH₄ consumption in dry soils and 58 and 12% of the annual CH₄ emitted from moist and saturated soils, respectively. Owing to the inherent high porosities of snow, estimates of gas flux from snowpacks are likely to be sensitive to the method of measurement. Snowpack concentration gradients appear to provide a more accurate technique for

estimating gas flux than static chamber measurements made at the snow surface, particularly in areas where seasonal snowpacks reach meters in depth.

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