CO_2 in Arctic Snow Covers: Landscape Form, In-Pack Gas Concentration Gradients, and the Implications for the Estimation of Gaseous Fluxes

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ABSTRACT

The physical characteristics and CO_2 concentrations of snow cover in the western Canadian arctic were examined at sites with different landscape forms (valley floor, hillslope, plateau). The greater exposure of plateau snow cover to blowing snow results in differences in the structure of the snow cover and different snow strata compared with snow covers on the other landscape forms. Both higher in-pack concentrations of CO_2 and the largest vertical CO_2 concentration gradients were found in plateau snow cover, the smallest in the deeper hillslope and valley snows. CO_2 gradients in all landscape snow covers followed two patterns i.e. where concentrations at the soil-snow interface are higher than those just below (5 cm) the snow-atmosphere interface and vice-versa. The latter pattern is due to the transport of the gas from the lower levels to the upper levels of the snowpack by wind-induced advection (windpumping) and is indicative of non steady-state, non-diffusive processes. These latter processes should thus be considered in any studies on CO_2 fluxes from Arctic soil where snow cover topography and winds are conducive to windpumping and where concentration gradients resulting from diffusive processes have not been clearly identified.

INTRODUCTION

Snow cover plays an important role in governing gaseous exchange between the soil and the atmosphere by forming a multi-phase porous structure that is sensitive to fluxes of heat and mass. Gaseous exchange between the soil and the atmosphere is a well known phenomenon in Arctic ecosystems throughout the winter season (Coyne and Kelley, 1971, 1974; Zimov et al, 1993; Fahnestock et al, 1998; Jones et al, in press). Gas fluxes through snow are a significant part of the global carbon budget and can be the largest component of the annual flux in the Arctic because of the long snow-covered season and possible microbial respiration or release of gas trapped by frozen soils (Oechel et al., 1997; Jones et al., 1998; Scott and Larson, 1985; Kappen, 1989; Winston et al, 1997). Fluxes of gases such as CO_2 and N_2O from snow cover to the atmosphere have been measured directly using chambers placed on snow (CO_2, Winston et al, 1995; Oechel et al., 1997; Mast et al, 1998) or calculated from concentration gradients and assumed transfer coefficients of CO_2 or N_2O within the snowpack (Solomon and Cerling, 1987; Sommerfeld et al, 1991, 1993; van Bochove et al, 1996; Fahnestock et al., 1998; Mast et al,1998; Jones et al., in press). If the gradients are linear, and mass transport is due to diffusion, then gas fluxes can be calculated using Fick's first law of diffusion as,

$$F = D_s(d[g]/dz)/ \quad (1)$$

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where F is the steady state flux of the gas (e.g. mg CO_2 m^{-2} h^{-1}), D_s is the diffusion coefficient for the gas in snow, (m^2 h^{-1}), d[g]/dz is the concentration gradient (ppmv m^{-1}) and / the conversion factor for the concentration unit of volume/volume (ppmv) to that of mass/volume (mg m^{-3}).

D_s may be estimated directly from experiments in the field (Zimov et al, 1993; Jones et al, 1999) or calculated as:

$$D_s = D_a\theta/\delta \quad (2)$$

where D_a is the diffusion coefficient of the gas in air, θ is the porosity of the snow (dimensionless) and δ is the tortuosity of the snow (dimensionless).

The porosity of snow can be obtained from the density of snow as:

$$\theta = 1-\rho_s/\rho_i \quad (3)$$

where ρ_s and ρ_i are the density of snow and ice respectively. Typical values of snow density in various environments and seasons are listed by Pomeroy and Gray (1995).

The tortuosity of snow is, however, very difficult to measure (Winston et al, 1995) as it can vary with the degree of metamorphism of snow grains, the presence of ice crusts and lenses, and water content (Perla, 1982). Consequently there are few values available in the literature. In some studies on Alpine snow cover where the snow is relatively deep and homogeneous with little structural differentiation, the value of δ has been assumed to be either equal to 1 (Sommerfeld et al, 1993) or a function of the observed porosity ($\delta = \theta^{1/3}$, Mast et al, 1998). In the case of shallower Arctic snow cover, some studies have taken tortuosity into account by reducing the calculated fluxes by 35% of their original values (Fahnestock et al, 1998; Jones et al, in press).

When concentration gradients within snowpacks cannot be determined due to a restricted number of measurements, then some studies have assumed the applicability of Fick=s law to calculate gas fluxes from concentrations measured at the base of the snowpack and in the air above the snow. For instance, Zimov et al (1993) and Jones et al (in press) assumed that steady-state diffusion dominated CO_2 exchange and that therefore concentrations measured at the soil surface gave rise to linear vertical gradients in the over-lying snow cover. The assumption of steady-state diffusion in such calculations may lead to erroneous CO_2 fluxes during sampling periods if the snowpack has horizontal impermeable ice lenses which impede diffusion (Hardy et al, 1995) or if exchange processes other than diffusion are significant.

In particular, the assumption that steady-state diffusion is the only process for CO_2 transport through snow must be carefully evaluated when applied to the Arctic environment, where snow covers can be highly irregular (Pomeroy et al., 1993; Marsh and Pomeroy, 1996; Essery et al., 1999), episodic CO_2 emissions occur (Coyne and Kelly, 1971; Zimov et al., 1993) and irregular soil and snow microtopography, high wind exposure, and intense temperature gradients can cause mass transport by convection (Benson and Sturm, 1993) and/or advection (Colbeck, 1989; Clark and Waddington, 1991; Albert et al., 1996).

If transmission of gas does take place by air flow, then the permeability of the snow must be taken into account (Albert and Hardy, 1995). The permeability influences fluid transport through pore space according to Darcy=s law.

Permeability, K, m^2, is defined as:

$$K = V\mu[\, dx/dP] \quad (4)$$

where V is the volumetric flux of air, m s^{-1}, μ is the viscosity of air, kg m^{-1} s^{-1}, and dx/dP the reciprocal of the pressure gradient in m^2 s^2 kg^{-1}. Permeability is affected by porosity and tortuosity but there is no direct relationship between the three parameters (Hardy et al. 1995).

The focus of this paper is to demonstrate the influence of snowpack structure and depth on the mode of gaseous transport within the snow covers of different Arctic landscape types. CO_2 was chosen as the gas due to ease of measurement and the availability of data from prior studies.

METHODOLOGY

Experimental sites

Sampling of the snowpack, ambient air and air in snow for CO_2 was carried out at two sites in the continuous permafrost zone of the western Canadian Arctic (Figure 1). The site climate, vegetation and topography is fully described by Pomeroy and Marsh (1997). The first, Havikpak Creek, a small valley 10 km southeast of Inuvik, Northwest Territories (68°N 134°W) is dominated by a sparsely-wooded, open black spruce forest. Vegetation at the snow-soil interface consists of lichens and mosses. A thick, organic layer (0.3 - 1.0 m) overlies poorly drained mineral soils. The seasonally-thawing soil layer is shallow, between 0.3 and 0.5 m deep. Snow air samples for CO_2 analyses were taken at six points, spaced 0.5 - 1 m apart, along a transect through the woodland (Transect A, 6 m; Figure 1) on April 16 1992.

The second site, Trail Valley Creek, is an upland tundra basin, situated 50 km north of Inuvik. Vegetation consists of grasses, mosses, lichens and shrubs. Soils consist of a organic mat (0.1 to 0.5 m thick) overlaying mineral soil. The organic mat is thicker in poorly-drained lowland sites. The seasonally-thawing soil layer has a variable depth, between 0.3 and 2 m, depending on snow insulation (snow depth) and drainage. Snow atmosphere samples were taken along a transect (Transect B, 100 m; Figure 1) over three landscape forms - a tundra valley floor (30 m), a snow-drift covered hillslope (50 m; along a rise of 30 m in height) and a tundra upland plateau (20 m) which was well-exposed to wind. There were six sampling points in the valley, three points on the hillslope, and four points on the plateau. The distance between each sampling point was 5 m on the plateau and in the valley and 15 m along the hillslope. The samples were taken on April 24 (plateau), April 25 (valley) and April 26 (hillslope and plateau).

Snow cover structure and temperature profiles

Temperature was taken by thermocouples (Havipak Creek) or dial-type thermometers (Trail valley Creek). Wind speed was measured adjacent to the Havipak sites in the valley bottom and nearby to the Trail Valley Creek site on an exposed plateau; each site using 4 calibrated NGR cup-anemometers in a vertical array. Snow grain size and snow density were recorded using a standard snow measurement kit from the National Hydrology Research Institute of Environment Canada. Snow structure was classified using the International Classification for seasonal snow on the ground (Colbeck et al., 1990).

CO_2 sampling and analyses

Snow atmosphere and air samples for CO_2 analyses were taken with plastic syringes (10 ml); the sample size was 4 ml. Samples were taken from the top of the pack by inserting the needle 4-5 cm into the snow and drawing air into the syringe. Samples from the bottom of the pack were taken by first carefully removing some part of the upper strata of the snow cover (particularly wind slab on the plateau) to provide access and then thrusting the syringe to the ground surface through the depth hoar. The method generally disturbs the snow cover more than the use of long tubes for grab samples of snow gases. However, observations in the field lead us to conclude that relatively little disturbance occurs at the soil-snow interface at the point of sampling by the syringe needles.

Samples were withdrawn from the syringes by suction into previously evacuated tubes (vacutainer, Fisher). The sealed tubes were then transported to the laboratory and CO_2 analyses carried out by the use of a gas chromatograph (Perkin-Elmer Sigma 300) with a Porapak Q 80/100 mesh column. Helium was used as the carrier gas. Injector and column temperatures were both 40°C and the CO_2 was detected by a thermal-conductivity detector. Blanks were run on evacuated tubes carried into the

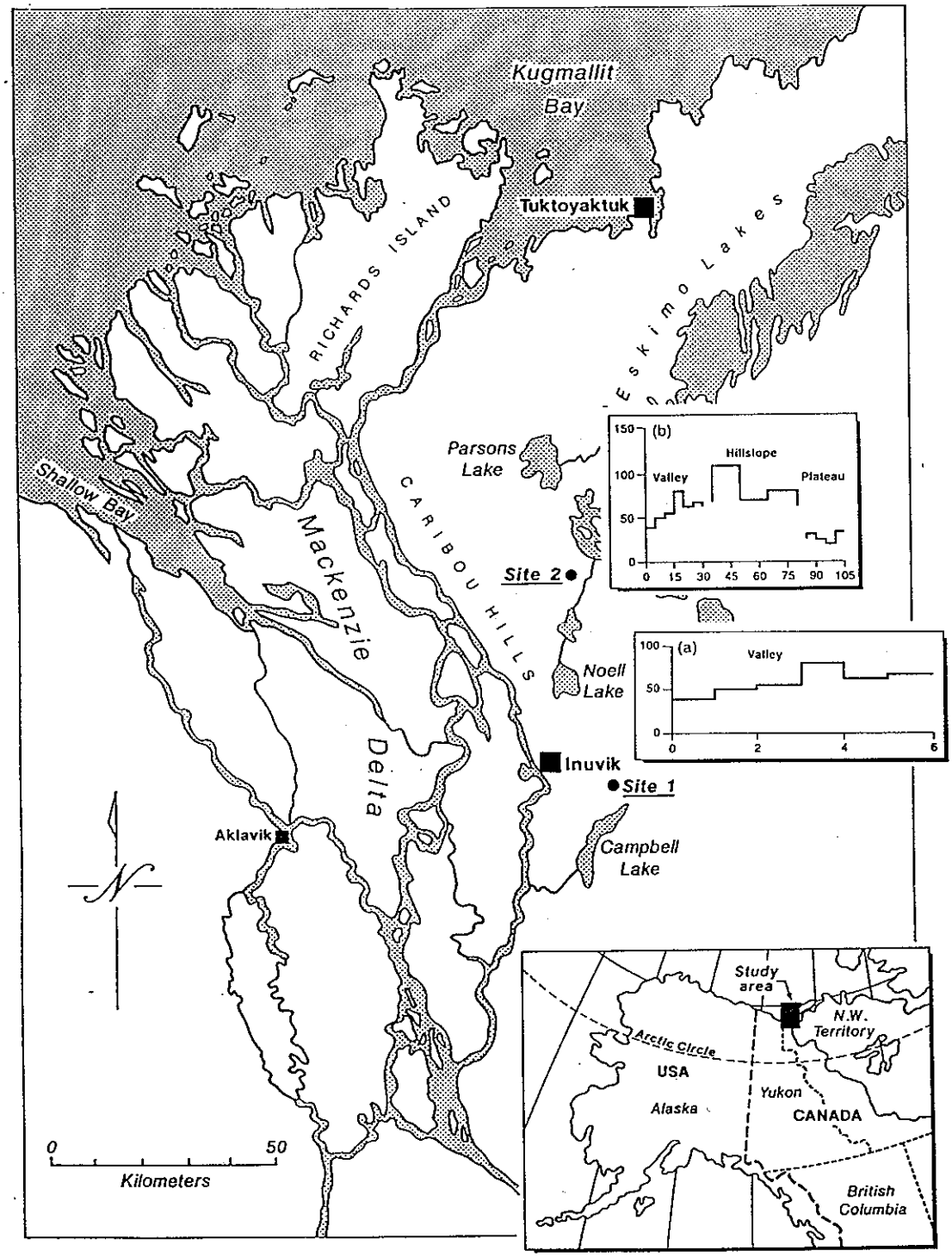

Figure 1: Study sites, and sample transects, Inuvik, Northwest Territories, Canada, April 1992. Insets, a) snow depth (cm) and sample transect (m) Site 1, Havikpak Creek, b) snow depth (cm) and sample transect (m) Site 2, Trail Valley Creek

field and filled with helium prior to the analyses. Control samples were those of ambient air sampled from above the snow cover. CO_2 concentrations were corrected for the temperature at the time of sampling. The detection limit for CO_2 was approximately 5 ppmv.

RESULTS AND DISCUSSION

Snow structure

Snow depth, density, layering and classification for representative snow pits in each of the four sampling areas are shown in Figure 2. Classification terminology follows that for seasonal snow cover properties as outlined by Colbeck et al. (1990). The snow profiles show substantial differences in snow stratigraphy as snow crystal/grain type, density and depth vary according to terrain. The Trail Valley Creek "valley" sites contained low density, metamorphosed snow that was quite similar to that found in the valley at Havikpak (Fig 2). The valley snow profiles are typical of cold subarctic snow covers with depths of 0.7-0.8 m, a deep depth hoar layer and no hard crusts. The Trail Valley Creek plateau sites have a shallow depth (25 cm), very dense, fine-grained snow in the upper crust with interior layer of ice lenses overlying a low-density depth hoar layer. These snows are typical of wind-scoured snowpacks and were extremely Ahard≡, supporting the investigators weight without any boot imprint. The Trail Valley Creek hillside sites are typical of tundra snow drifts with much deeper snow than elsewhere, less evidence of metamorphosis and greater density. Pomeroy et al. (1997) and Essery et al. (1999) have shown that the varying snow depths in this region are due to wind redistribution of snow; better-exposed snowpacks such as that on the plateau being the source of snow for that redeposited in valleys and hillsides.

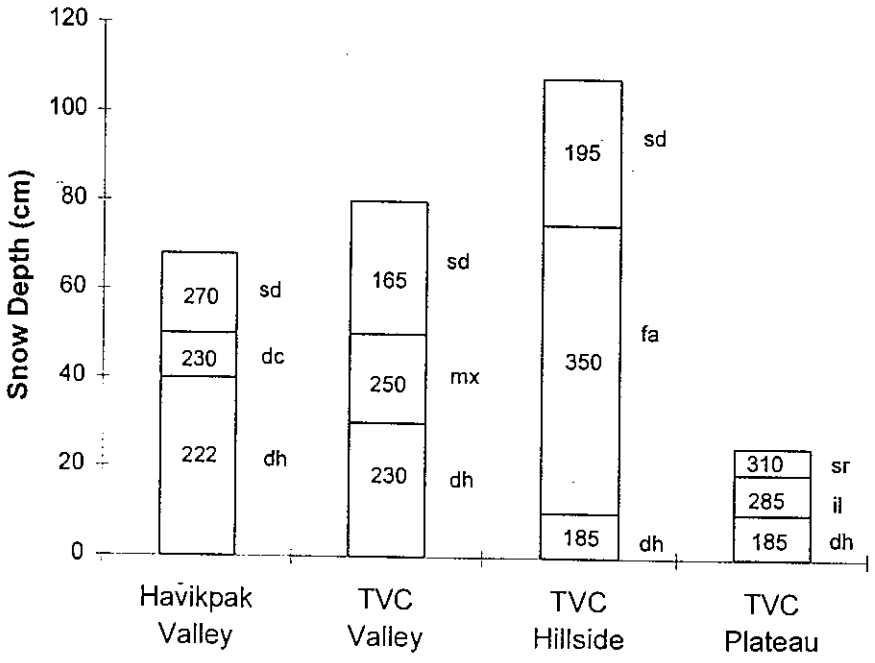

Figure 2: Snow stratigraphy of samples on the plateau, hillside and in the valleys, Havipak Creek and Trail Valley Creek, Inuvik, Northwest Territories, Canada, April 1992. Snow depth, snow density, layers and crystal type from representative snow pits for the sampling transects. Densities are shown within each snow layer in kg m^{-3}. Crystal type is shown next to each snow layer, sd = stellar dendrites (fresh particles), sr = small rounded grains, dc = partially decomposed particles, dh = columns of depth hoar, mx = mixed rounded grains, fa = solid faceted crystals, il = horizontal ice layer/depth hoar.

The variation in snow density results in a range of porosity, tortuosity and hence diffusion coefficients of gas in snow. The range of porosity in individual snow layers varies from 0.73-0.84 for valley snow, 0.65-0.82 for hillside snow and 0.69-0.82 for plateau snow. The hillside and plateau snowpacks are dominated by high density, low porosity snow. The ice and wind crust layers in the plateau snow, by definition, must have sections with extremely low porosity. However, the variation in bulk porosity is relatively small amongst sites. Tortuosity cannot be directly calculated from density and crystal size; values of 1.0 to 1.42 have been assumed by other studies (Massman et al., 1995; Hardy et al., 1995). Here it is suggested that extremely high tortuosity values could develop in ice and wind crust layers such as found on the plateau. Low values of permeability would also be expected for the same features.

CO_2 concentrations in Snow

Within-pack CO_2 concentrations measured near the top and bottom of the snowpack for all sites are shown in Figure 3 along with snow depth. For snow deeper than about 0.35 m, snow-air CO_2 concentrations were within 75 ppmv of ambient atmospheric concentrations (356"2 ppmv). However for shallow snow, CO_2 concentrations were up to 404 ppmv greater (760 ppm) than the ambient concentrations and large values occurred in both the top and bottom of the snowpack (Fig. 3). Regrouping sampling sites by topography showed that the concentrations of CO_2 in the upper strata of snow cover in the valley transects (mean depth = 0.57 m) were similar to those in the ambient air (mean difference = 12, standard deviation of difference = 9 ppmv). The CO_2 concentrations in the upper strata of the hillslope snow (mean depth = 0.86 m), showed a mean difference with ambient air of 25 (standard deviation of difference = 3) ppmv. In the case of the plateau sites (mean depth = 0.28 m) the mean concentration difference with ambient air was 91 (standard deviation of difference = 65) ppmv.

The generally low values for CO_2 in valley and hillslope and the low soil temperatures showed that emissions of CO_2 were low. Probably, very little respiration was taking place in horizons close to the surface and the CO_2 production in the deeper layers were confined by the low permeability of the frozen surface soil (Zimov et al., 1993). The high CO_2 measurements all occurred in snow on the

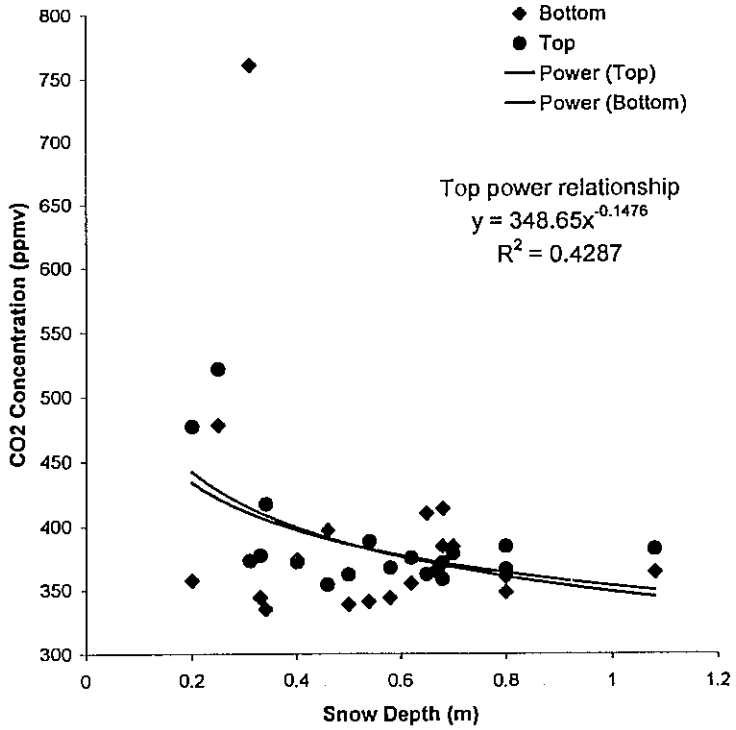

Figure 3. CO_2 concentrations (ppmv) found at the top and bottom of snowpacks as a function of total snow depth (m) for all sites.

plateau. The temperature of the soil would indicate that the production of CO_2 by the soil on the plateau should be similar to the valley and hillslope sites. One explanation for the greater concentrations of CO_2 in the plateau snow cover maybe due to the greater permeability of the organic soil surrounding the mineral soil which constitute distinct hummocks (Quinton and Marsh, 1998). Alternately, it may be due to the greater prevalence of the plateau soils to release episodic emissions of CO_2 under wind-induced thermal stresses of the soil structure (Zimov et al., 1993). If an episodic emission does occurs, the presence of crusts or windslab could reduce or prevent CO_2 exchange by diffusion and trap relatively high concentrations of the gas within the snowpack.

CO_2 Profiles

Examples of CO_2, snow strata and temperature profiles for different sites in the snow covers of valley and plateau are shown in Figure 4. The differences between within-pack CO_2 concentrations

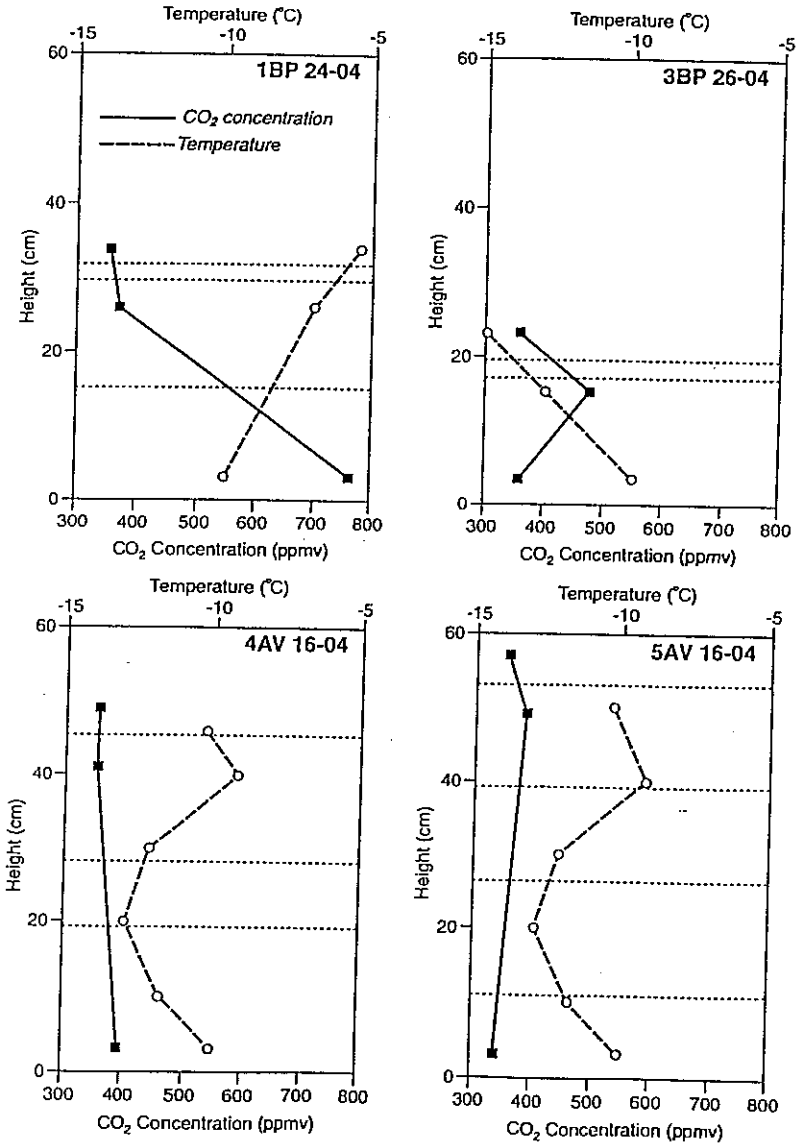

Figure 4. Snow structure, CO_2 concentrations and temperature in and just above Arctic snow covers on the plateau and in the valley, Inuvik, April 1992. The CO_2 profiles show the presence of positive and negative concentrations gradients of the gas in two shallow, wind-blown plateau sites (1BP and 3BP) and two sheltered valley sites (4AV and 5AV). Snow strata are shown along with varying temperature gradients.

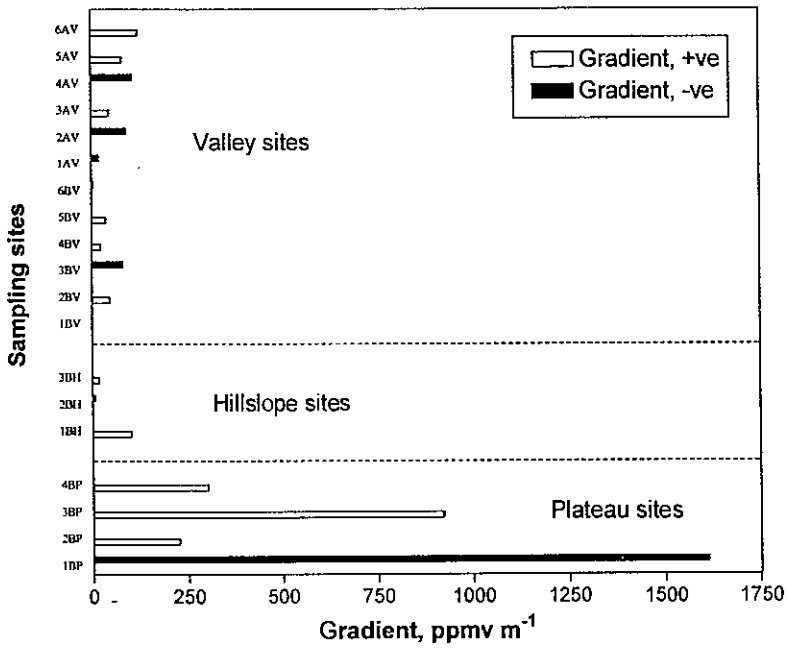

Figure 5. CO_2 concentration gradients (ppmv m^{-1}) in Arctic snow cover regrouped by landscape type (P, plateau; H, hillside; V, valley bottom).

were either negative (higher concentrations in the lower strata) or positive (higher concentrations in the upper strata) depending on site. Both examples are shown. It is clear from these examples that errors in flux calculations could occur from applying Fick=s law by presuming a linear gradient through snow and measuring concentrations only at the bottom of the snowpack and in air (Zimov et al., 1993; Jones et al., in press).

Figure 5 shows the distribution of the calculated CO_2 gradients (both positive and negative) in all terrain types. In contrast to valley and hillslope snows there were much stronger gradients of both types in the plateau sites. In the case of the snow cover sites with decreasing bottom-to-top concentrations, we conclude that the negative gradients (Figure 5) are the result of diffusion of CO_2 from the soil through the snow. The production of negative gradients can be the result of both steady-state CO_2 production in soil (Sommerfeld et al., 1993) or non steady-state CO_2 production (Sommerfeld et al., 1996). However, in the situation where the pattern of the CO_2 is reversed i.e. positive CO_2 gradients, we can conclude that diffusion of CO_2 from the soil through snow was not responsible for the observed CO_2 pattern.

The possibility that the CO_2 originated elsewhere than the soil is small, as no CO_2 is produced within such dry, cold snow. Chemical interaction between calcareous dust and acids (Delmas et al, 1996) and microbiological activity (Jones, 1991) by heterotrophic populations in melting snow (Stein and Admunsen, 1967) can generate CO_2 in snow, but require liquid water. The temperature of the snow cover was too low (-7° to -13°C) for appreciable amounts of liquid water to be present in the hillslope and plateau snowpacks. The valley snowpacks in Trail Valley Creek did reach mid-afternoon temperatures (-5 to -2°C) where a small amount of interstitial liquid water may have been present. However, the short time period spent at this temperature, negative daily net radiation and the internal energy deficit necessary to overcome before phase change could begin (Marsh and Pomeroy, 1996; Brandt and Warren, 1993) would not have permitted sufficient liquid water for significant chemical or microbial activity.

Positive gradients could be generated if the soil/ground cover became a sink for CO_2. Then the drawdown of CO_2 from the pack would lead to positive gradients. However, the upper limit for concentrations of CO_2 in the top of the pack would be the atmospheric concentration (356 ppm).

Sinks could be result of photosynthetic activity by higher plants and algae at the soil surface although photoautotrophic and chemoautotrophic bacteria may also contribute in a minor fashion. However, this would be extremely unlikely in the low mid-winter temperatures of a desiccated Arctic soil.

Transport Mechanisms

Given no within-pack source of CO_2 and no soil sink, the higher CO_2 concentrations in the upper snow strata can only have resulted from convective transport of CO_2 from the soil to the snow-atmosphere interface. Convective transport requires airflow; airflow in snow can be forced by wind-pumping (Colbeck, 1989; Massman et al, 1995; Albert and Hardy, 1995) and/or thermal convection (Powers et al., 1985; Sturm, 1991; Sturm and Johnson, 1991).

Thermal convection in snow gives rise to upwelling of relatively warm air from the soil and the compensating downwelling of cold air from the surface (Sturm and Johnson, 1991). The snow covers at Havipak Creek and the valley site at Trail valley Creek are similar in porosity and grainsize to the snow cover chosen by Sturm (1991) for his study of thermal convection. However, no thermal convection could have taken place in the spring period we examined as temperature gradients were very weak (<6 $^{o}C\ m^{-1}$) and most sites had negative (stable) gradients, showing colder air within, or at the base of, the pack than in the upper strata and ambient air.

Windpumping is the result of three processes.

1) *barometric pumping* is driven by changes in air pressure associated with synoptic weather systems and causes air to move very slowly in and out of the snow cover as it expands and contracts (Colbeck, 1989; Cunningham and Waddington, 1993). Barometric pumping could have occurred equally at all sites as it is not sensitive to wind speed or permeability of snow. However, barometric pressure fluctuations were less than 5% over the measurement period and therefore the gas exchange due to barometric pumping would have had a negligible impact for such shallow snow.

2) *turbulent pumping* is caused by fast pressure fluctuations at the surface associated with turbulent eddies in wind flow over snow (Clarke et al., 1987; Clarke and Waddington, 1991). The pressure fluctuations that cause turbulent pumping increase exponentially with wind speed and therefore snowpack exposure to wind (Colbeck, 1989). There are notable differences (40 fold) in the estimate of air flow due to turbulent pumping by the analyses of Colbeck and Clarke and Waddington (1991). Waddington et al. (1996) suggest that large numbers of holes in relatively impermeable surface layers such as found on wind-crusted snow could induce Arelatively rigourous air motion at depth≡.

3) *topographic pumping* is occurs when pressure differences in wind travelling over an undulating snow-cover induce quasi-static pressure fields that are strong enough to initiate air currents within the snow structure (Colbeck, 1989). The velocity of the airflow depends on the snow depth, wavelength of surficial topography surface roughness, wind velocity, porosity and air permeability. Waddington et al. (1996) showed that the air velocity due to topographic pumping over an undulating surface can be calculated as (Equation 5):

$$Q_z = \frac{K}{\mu}\frac{6\rho_{air}U_{10}^2}{\pi}\frac{h}{\lambda}\frac{1}{\lambda}\frac{\sqrt{\alpha^2+1}}{\alpha} \qquad (5)$$

where Q_z is vertical air flux (m/s), K is permeability (m^2), μ is air viscosity, ρ_{air} is air density (kg/m^3), U_{10} is wind speed (m/s), h is undulation height (m), λ is undulation wavelength (m) and α is dune orientation with respect to wind direction.

Turbulent and topographic pumping may influence gaseous exchange within snow cover. Massman et al. (1997) studied a model of pumping in snow due to short-term turbulent atmospheric pressure fluctuations. They concluded that this type of pumping would not significantly influence

the rate of gaseous diffusion of CO_2 through snow unless the pumping effect reached down to the snow-soil interface and impacted on CO_2 concentrations at that level. Thus shallow sites on the plateau could be more sensitive to pumping of this nature than valley and hillslope sites.

The presence of air currents due to topographic pumping can play a more significant role in air exchange through snow. The movement of air through the snow will dominate over the diffusive process by physical transport of CO_2 within the pack in both vertical and horizontal directions. We would expect windpumping to have little effect on the advection of CO_2 in the valley and hillside snow covers compared to the plateau sites. At the valley and hillslope sites the snow surface is relatively smooth (valley coefficient of variation of snow water equivalent = 0.16 {Pomeroy et al., 1998}) and wind speeds are significantly lower (approximately 50%) than over exposed plateaux because of deceleration of air flow over concave topography (Essery et al., 1999).

In contrast, the snow cover on the plateau at Trail Valley Creek is subject to strong winds, has a distinctive wind crust (Fig. 2) and is covered with snow dunes (coefficient of variation of snow water equivalent = 0.31 {Pomeroy et al., 1998}). Such conditions would promote forced convection via turbulent and topographic pumping. The surface is heterogeneous and the permeability therefore varies; this will create a system in which air may flow into the snow cover in relatively permeable areas (turbulent pumping), but be restricted from leaving the snow cover in other less permeable areas. Unfortunately there are presently no methods to calculate the contribution of turbulent pumping through a heterogeneous surface crust. Topographic wind pumping, however, may be estimated from measurements of snow characteristics and wind speed. If we assume a mean snow permeability of $100@10^{-10}$ m^2, a snow surface characteristic wavelength (λ) of 1 m and undulation height (h) of 0.03, and a porosity (θ) of 0.8, then the calculated airflow velocities in the shallow plateau snow cover using Eq. 5 are of the order of $4@10^{-3}$ m s^{-1} at a depth of 0.2 m for a wind speed of 5 m s^{-1}. At this velocity, wind pumping would have displaced any CO_2-rich air from the base of the snow cover into the upper strata within a matter of minutes.

Thus during to the sampling period, airflow carrying CO_2 may have been occurring in the upper strata of the plateau snows. Wind speed during the sampling dates varied from 2 to 8 m s^{-1}. CO_2-rich pockets in the upper strata could also have due to the trapping of CO_2 under the relatively impermeable wind slab if airflow had decreased or stopped prior to sampling. The wind slab/ice layers have a lower porosity and lower permeability than underlying snow strata. If pockets of CO_2 were to be trapped in the upper snow strata in regions of low air flow as wind speed diminished, then the subsequent redistribution of the trapped CO_2 within the snowpack would be by diffusion. We can estimate the persistence of CO_2 pockets by calculating the half-life for the trapped gas. Using a simple dynamic equilibrium calculation for CO_2 diffusion (D_s, 0.025 m^2 h^{-1}) between two "impermeable" boundaries (wind crust and soil), presuming no advection and imposing mass continuity, the calculated decay of CO_2 concentration in an entrapped CO_2-rich (500 ppmv) pocket of air (0.05 m^3) under the crust was 425 ppmv in 1 hour and to 390 ppmv in 2 hours (i.e. a half life of 57 mins).

In the above calculation the value for D_s is estimated from a porosity of snow equal to 0.6; it assumes a snow tortuosity of 1. However, with ice lenses the effective tortuosity may be higher and the half life for the decrease of CO_2 will be greater. In the case of any wind pumping that may have occurred in hillslope or valley sites, the increased porosity of upper snow strata and the absence of ice layers would not have lead to any significant trapping of air currents.

The existence of positive gradients of CO_2 which can be explained by snow structure and prevailing meteorological conditions on landscape forms show that non steady-state non-diffusive processes can govern the in-pack transport of gas. The results show that the estimation of CO_2 fluxes through snow cover should consider such processes where temperature, snow topography, wind speed and snow depth are conducive either to thermal convection or windpumping and where in-pack gas concentration gradients cannot be unequivocally be assigned to transport by diffusion. Controls on the transport of snow can thus vary spatially and temporally. When this variability is superimposed on the changes in CO_2 production in the soil during the cold season (Sommerfeld et al., 1996) and the

episodic release of CO_2 at the soil-snow interface by physical processes (Zimov et al., 1993), the determination of CO_2 emissions becomes extremely difficult. Consequently, we cannot estimate at this particular time the extent to which the phenomenon of transport of CO_2 through snow by forced convection could affect global CO_2 emissions in cold regions.

CONCLUSION

The presence of CO_2-rich pockets of air in the upper strata of snow relative to the snow-soil interface and atmosphere constitutes evidence for the unsteady, non-diffusive transport of CO_2 through exposed Arctic snow covers. In the absence of thermal convection, the concentration patterns are indicative of transport of CO_2 by forced convection (topographic windpumping) within the snow cover. These results show that one cannot assume that only diffusive processes are responsible for gaseous concentrations in snow cover and flux to the atmosphere. Attempts to calculate gaseous fluxes between the soil and the atmosphere using Fick=s law in the absence of concentration gradients that can not be unequivocally be assigned to diffusion processes, will give rise to errors in the compilation of CO_2 budgets for arctic winter conditions.

ACKNOWLEDGEMENTS

The authors acknowledge the logistical and financial assistance of the Aurora Institute, Inuvik Research Laboratory; Polar Continental Shelf Project, Natural Resources Canada; Natural Sciences and Engineering Research Council of Canada; NATO Collaborative Grants Programme; Canadian GEWEX Programme and the National Hydrology Research Institute, Environment Canada. Superb field and scientific support was provided by Mr. Cuyler Onclin in extremely adverse conditions.

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