Soil carbon dioxide fluxes and profile concentrations in two boreal forests

S.A. Billings, D.D. Richter, and J. Yarie

Abstract: Because a large fraction of the world’s carbon exists in the soil of boreal forests, understanding how soil temperature and moisture affect soil respiration is vital for predicting soil response to climate change. We measured soil respiration and CO₂ concentrations within soils of floodplain and upland forests in interior Alaska from 1996 to 1997. At each site, a 0.10-ha-area shelter was constructed that prevents summer precipitation from infiltrating into the soil. Measurements of soil profile CO₂, soil respiration, soil temperature, and soil moisture were made inside (treatment) and outside (control) the sheltered areas through two growing seasons and the winter of 1996–1997. Sheltered soils had decreased profile concentrations and surface flux of CO₂. At the upland control site, individual flux rates ranged from 0.10 to 0.95 g·m⁻²·h⁻¹ in the summer and at sites under the shelter from 0.10 to 0.53 g·m⁻²·h⁻¹. Rates at the floodplain control site ranged from 0.11 to 1.45 g·m⁻²·h⁻¹ and under the shelter from 0.11 to 0.55 g·m⁻²·h⁻¹. Fick’s Law could predict surface CO₂ flux when the CO₂ concentration gradient within the profile accurately represented the soil surface gradient and biological sources and sinks of the gas did not overwhelm flux calculations.

Résumé: Étant donné qu’une fraction importante du carbone dans le monde se trouve dans le sol des forêts boréales, il est vital de comprendre de quelle façon la température et l’humidité du sol affectent la respiration du sol pour prédire la réaction du sol à un changement climatique. Nous avons mesuré la respiration du sol et les concentrations de CO₂ dans le sol de forêts situées sur une plaine alluviale et sur un plateau de l’intérieur de l’Alaska de 1996 à 1997. À chaque endroit, un abri d’une superficie de 0,10 ha a été construit pour empêcher les précipitations estivales de s’infiltrer dans le sol. Les mesures du CO₂ dans le profil de sol, de la respiration, de la température et de l’humidité du sol ont été prises à l’intérieur (traitement) et à l’extérieur (témoin) des sites couverts pendant deux saisons de croissance et pendant l’hiver 1996–1997. Les sols protégés avaient des concentrations de CO₂ dans le profil et un flux de CO₂ en surface plus faibles. Pendant l’été sur le plateau, les taux individuels de flux variaient de 0,10 à 0,95 g·m⁻²·h⁻¹ dans les zones témoin et de 0,10 à 0,53 g·m⁻²·h⁻¹ dans les zones sous abri. Sur la plaine alluviale, les taux variaient de 0,11 à 1,45 g·m⁻²·h⁻¹ dans les zones témoin et de 0,11 à 0,55 g·m⁻²·h⁻¹ dans les zones sous abri. La loi de Fick pouvait prédire le flux de CO₂ en surface lorsque le gradient de concentration de CO₂ dans le profil représentait fidèlement le gradient à la surface du sol et que les sources et les puits biologiques de CO₂ ne venaient pas fausser les calculs de flux.

Introduction

Between 16 and 24% of the world’s soil carbon is in the vast boreal forest (Gates 1993), which comprises 1.2 x 10¹³ ha (Walter and Breckle 1986). This estimate is conservative because of significant, unquantified reservoirs of carbon immobilized in permafrost (Billings 1987). Any future changes in climate could increase depth of thaw in these soils, thus influencing soil water availability in this region. Increased depth of thaw and altered soil moisture will affect rooting depth, root respiration, and soil microbial community composition. Under such a scenario, soil respiration could be altered dramatically. This could have a significant effect on the global carbon cycle (Billings 1987, 1995). Because soil respiration is an indicator of the biological activity occurring in a soil, monitoring soil respiration in conjunction with related soil parameters can help us evaluate how microbial and root activity in active-layer soils may respond to a changed climate (Schlentner and Van Cleve 1985; Thierron and Lau delout 1996).

Soil CO₂ efflux has been studied in boreal (Gordon et al. 1987; Schlentner and Van Cleve 1985; Weber 1985), arctic (Billings et al. 1982, 1984), temperate (Wiant 1967; Wintz 1969; Raich and Schlensinger 1992; Solomon and Cerling 1987; Landsberg 1986), and tropical systems (Trumbore et al. 1995). However, few of these studies have examined the effects of experimentally altering water availability on both soil profile CO₂ dynamics and surface efflux. Measurements of this kind are especially important in the boreal zone, where depth of soil thaw affects water availability and both gas profile and surface fluxes.

Johnson (1993) discussed the need for studying factors controlling forest soil CO₂ fluxes such as soil porosity. This
requires incorporating in soil CO$_2$ flux models the effects of water availability and soil air-filled porosity. Soil temperature must also be taken into account to determine when water is in a frozen state and not biologically available. In order to better determine how carbon gas fluxes may be altered by changing water availability, we explored how altering the precipitation regime at boreal forest sites affects soil respiration and soil profile CO$_2$ concentrations. With this profile information, Fick’s Law was used to calculate soil CO$_2$ fluxes near the soil surface. Soil surface fluxes were also examined in relation to soil moisture and temperature data.

**Materials and methods**

Soil atmosphere samples, soil respiration, and soil temperature and moisture data were collected at two sites with contrasting moisture regimes in the Bonanza Creek Experimental Forest, 20 km southwest of Fairbanks, Alaska (64°4′N, 148°2′W). The mean annual temperature is −3.5°C; the growing season is 90–100 days (Van Cleve and Yarie 1986). About 37% of mean annual precipitation, 269 mm, is snow (Viereck et al. 1993). Potential evapotranspiration was 466 mm (Patric and Black 1968). Snow cover generally is present from the end of September through mid-April.

One site is a well-drained upland site (64°41′28″N, 148°21′45″W) on a ridge 308 m above mean sea level. The site is on a 25° slope facing east-southeast. The soils are loessial Alfis Cryaquepts. Approximately 7 cm of decomposing litter overlies a rock-free, silt loam – silt subsoil, which supports (2767 stems/ha) paper birch (*Betula papyrifera* Marsh.), white spruce (*Picea glauca* (Moench) Voss), and balsam poplar (*Populus balsamifera* L.). Dominant tree height is about 18 m. It has been approximately 80 years since the last forest fire.

The other site (64°43′24″N, 148°08′58″W) is on poorly drained, alluvial Entisols, classified as Typic Cryaustolls. The top 25–30 cm of the soil profile is silt loam, and from approximately 30 to 100 cm the profile is coarse sand. Several sampling sites were re-located 10–15 m horizontally on this surface to the top 100 cm of the soil profile. Litter depth on the soil surface is about 9 cm. This floodplain site is 10.7 km from the upland site and is on a 0° slope. The site supports 780 stems/ha, comprising 40–

50-year-old white spruce reaching into an upper canopy of 3–3.5 m above river water, and the site is 139 m above mean sea level. When the river level rises, the water table can reach to 1 m below the soil surface. Permafrost is not present at either site. We collected data at both sites throughout the 1996 and 1997 growing seasons and during the winter of 1996–1997.

At each site, a 0.10-ha shelter, approximately at breast height level, was constructed to prevent summer precipitation from reaching the soil. These shelters have been installed since 1989 and are removed each fall to allow a snowpack to accumulate. The shelters are made of translucent, corrugated fiberglass. Litter accumulation on the shelter tops was spread over the treatment area at the end of the growing season. A moisture barrier of plastic sheeting was buried vertically down to 0.6 m around the perimeter of each shelter to limit lateral moisture flow into the treatment area. Sampling took place at three locations beneath each shelter and three locations outside these shelters at both the upland and floodplain sites.

Soil atmosphere CO$_2$ concentrations were sampled from tubing inserted into the ground at 20-, 40-, and 100-cm depths at locations picked randomly inside and outside the shelters. Nalgene plastic tubing, perforated at the ends to allow air flow, and with flexible rubber tubes sealed on the tops with silicon, was used at the 20- and 40-cm depths. The 100-cm depths have copper tubing installed down to a 10-cm-diameter, 30-cm-long PVC pipe chamber, perforated at the bottom. The copper tubes are sealed at the top with flexible rubber tubes and silicon sealant. Syringe samples were taken weekly at each 20-, 40-, and 100-cm depth at three locations inside and three outside the shelters at each site. At the upland site, 200-cm copper tubing was installed at the end of the 1996 growing season for monitoring over the following winter in a similar fashion. Syringes were 10-mL glass BD Pack syringes with stopcocks epoxied to the tips. Sampling locations were randomly placed in and out of the rainout shelters at each site.

CO$_2$ surface efflux was measured next to the sampling locations for depth concentrations using chamber techniques. White plastic 5.5-L containers with airtight lids were set into the soil approximately 10 cm. Locations were chosen with no visible photosynthesizing plants. We allowed them to equilibrate for several weeks to let the initial flush of CO$_2$ from dying fine roots dissipate. Syringe samples were taken from septa installed in the lids using the same type of syringe as for below-ground atmosphere sampling. A sample of above-ground atmosphere was taken immediately adjacent to the chamber to compare with the initial, time zero sample inside the chamber. Subsequent samples were taken every 15 min over a period of 45 min.

Samples were immediately transported to a laboratory at the University of Alaska in Fairbanks and were analyzed within 24–48 hours. No difference was found between samples immediately analyzed and those allowed to sit for up to 4 days. Samples were analyzed with a thermal conductivity detector in a Shimadzu GC-14A gas chromatograph (Shimadzu, Columbia, Md.) with a 2-m stainless steel column packed with Porapak Q. Standards were Scotty gas standards (Scotty Gases, San Bernardino, California). We determined the rate of CO$_2$ efflux by fitting an equation to the curve describing the increase in CO$_2$ over time within the chamber and taking the derivative of this equation. The value of this derivative at time zero, used in conjunction with data for chamber volume and area, was considered the soil respiration rate.

Throughout each growing season, soil temperature at 20, 40, and 100 cm was continuously recorded with thermistors (MK820, Siemens, Erlangen, Germany) connected to a Delta-T datalogger (DL-2, Delta-T, Burwell, U.K.). Thermistors were placed at three depths (20, 40, and 100 cm) at three locations inside and three outside the shelters at each site. Soil moisture was measured weekly using time domain reflectometry (TDR, Tektronix, Beaverton, Oregon) at two locations inside and two outside the shelters at 5, 10, 20, and 50 cm.

Gas flux out of the soil profile was calculated according to Fick’s Law. This assumes that the main direction of soil gas flow is vertical and that it is primarily driven by diffusion gradients. Fick’s Law states that gas flux $J$ is a function of the gas concentration gradient and the gas diffusion coefficient in free aboveground air:

$$J = -D_g \frac{\partial C}{\partial z}$$

where $J$ is flux rate (grams CO$_2$ per square centimetre per second), $D_g$ is the diffusion coefficient of CO$_2$ in air (square centimetres per second), $C$ is gas concentration (parts per million), and $z$ is depth (centimetres).

The value of $D$ in a soil medium is an estimated fraction of the known diffusion coefficient in air, due to the longer path length and any moisture through which soil gas must travel. Estimates of flux calculated from Fick’s Law are limited in accuracy by this estimated value of $D$ (Rolston et al. 1978), as well as by how $\frac{\partial C}{\partial z}$ is calculated. Several investigators have estimated $D$ in various soil

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types (Penman 1940; Marshall 1959; Millington and Quirk 1961; Currie 1965). We chose Marshall’s (1959) form of Fick’s Law to predict upland soil CO$_2$ fluxes, which estimates $D$ by multiplying by an air-filled porosity term, $\phi$. This exponent was changed to 1.1 to find the equation that best predicted floodplain fluxes. Air-filled porosity, $\phi$, was calculated from TDR data at 20 cm according to Vomocil (1965).

Statistical tests on the effects of the shelters were performed using $t$ tests for equal means. When the data were nonnormally distributed, the nonparametric Wilcoxon rank-sum test was used. Relationships between CO$_2$ efflux and moisture and temperature were tested using linear regression. Significance was determined at $\alpha = 0.05$.

**Results**

**Soil moisture**

TDR measurements both inside and outside the rainout shelters show the effect of preventing precipitation infiltration into the soil (Fig. 1). Under the shelters at each site, volumetric soil moisture content (VMC) decreased with availability of water from snowmelt. At the upland site, moisture content inside the shelters at 20 cm ranged from 8 to 24% in 1996 and from 8 to 39% in 1997, with the higher values only occurring immediately after snowmelt. Outside the shelters, the range was approximately the same, but higher values were not limited to immediately after snowmelt.

At the floodplain site, volumetric water content was much higher than at the upland site. The water table at this site, sometimes as shallow as 1 m below the forest floor, maintained generally wetter soil conditions than at the upland site. Soil moisture at 20 cm ranged from 32 to 61% inside and from 61 to 65% outside the shelter in 1996, values often close to saturation for these soils. Values ranged from 28 to 46% inside and from 46 to 50% outside the shelter in 1997.

**Soil temperature**

Soil temperature increased throughout each growing season at each depth measured (Fig. 2, 20-cm data shown). At each site, soil temperature differences were found between locations inside and outside the rainout shelters. At the upland site, temperatures inside the rainout shelter were consistently higher than those outside in 1996 at all depths measured. The difference was generally less than 1.5°C. The same trend was noted at the beginning of the 1997 season.
Temperatures under the shelter were not recorded for the latter half of the 1997 season. Soil temperatures at 20 cm in 1996 ranged from 4 to 12°C and in 1997 from 2 to 20°C.

At the floodplain site, soil temperatures under the shelter (Fig. 2) were consistently lower at 20 cm in 1996, for much of 1997, and throughout both growing seasons at 40 cm. The maximum temperature difference was 3.0°C. At the floodplain, 20-cm temperatures in 1996 ranged from 0.5 to 7°C; in 1997, they ranged from 1 to 7.8°C. The lower temperatures in 1996 at both sites were the result of an unusually shallow snowpack in the winter of 1995–1996.

Soil profile CO₂ concentrations

Soil profile CO₂ concentrations generally increased throughout the growing seasons at the upland site (Fig. 3). Significantly greater CO₂ concentrations were measured outside the rainout shelter compared with inside at the 20-cm depth in 1996 and 1997 (P < 0.05). At 40 cm, CO₂ concentrations were not significantly greater outside the shelter in 1996 or in 1997. There were no differences in concentration inside the shelter at the 100-cm depth in 1996, but in 1997 the drier sites had higher CO₂ levels than the wetter ones (P < 0.05), and the same pattern existed at 200 cm that year (P < 0.05). Winter profile CO₂ concentrations at the upland site show a decrease at all depths to a steady minimum (Fig. 4). At the floodplain site, it was much more difficult to discern the effects of the rainout shelter on profile CO₂ concentrations (Fig. 5). Significant differences were found only at 100 cm in both 1996 and 1997 (P < 0.05). Concentrations at the floodplain were more variable, especially during spring both years, reflecting sudden pulses of CO₂ released from the thawing of buried decomposing material and (or) water.

Soil surface CO₂ efflux

Soil surface respiration inside the rainout shelters was significantly lower than that from the unaltered soil moisture regime at both sites in 1996 (Figs. 6A and 6C, P < 0.05). A late summer maximum was reached, coinciding with the warmest summer air temperatures (data not shown). In 1997, a significant negative effect of the rainout shelter on soil respiration was found only at the floodplain site (Fig. 6D, P < 0.05). At the upland site, although the mean of soil respiration data under the shelter was lower than that outside the shelter, the difference was not significant at the 0.05 level (Fig. 6B).

Soil temperature at 20 cm affected soil respiration both in and out of the rainout shelters at the upland site in 1996 (Table 1A, P < 0.05) and out of the shelter at the floodplain site in 1996 (Table 1A, P < 0.05). In spite of the positive
relationship between soil temperature and respiration at both sites in 1997, at neither site was the relationship significant at the 0.05 level. A significant positive relationship between CO₂ efflux and soil moisture was found in 1996 under the shelter at the floodplain and upland sites and in 1997 both outside and under the shelter at the upland site (Table 1B, \( P < 0.05 \)). Soil temperature was a more influential parameter on CO₂ efflux than soil moisture in multiple regressions of CO₂ efflux on soil temperature and soil moisture (Table 1C) in 1996. In 1997, neither parameter was a significant influence on soil CO₂ efflux.

**Discussion**

Predicting the effect of potential climatic changes on soil respiration requires an understanding of the processes governing both soil surface and soil profile fluxes. This study examined seasonal trends of these parameters and examined their relationship, if any, to soil temperature and moisture.

**Soil surface CO₂ fluxes**

Most of the soil respiration rates outside the shelters at both sites (0.42 ± 0.06 (SE) and 0.42 ± 0.03 g·m⁻²·h⁻¹ for upland and floodplain, season averages) were within the range of rates from other boreal forest soils, 0.16–0.46 g·m⁻²·h⁻¹ (Gordon et al. 1987; Schlentner and Van Cleve 1985; Weber 1985). The highest rate recorded at the floodplain site, 1.45 g·m⁻²·h⁻¹, was most likely the result of a sudden, short-term release of CO₂. At the upland site, however, rates outside the shelter during 1996 were higher than previously published ranges for this area. These rates are similar to and even higher than some rates found in temperate forests (Witkamp 1969; Schlesinger 1991; Thierron and Laudelout 1996). Our flux measurements represent an integrated response of many biological and physical processes in the soil, all of which respond differently to various parameters. Thus, it is difficult to obtain a strong correlation between flux measurements and a single variable, although clearly the rainout shelters had an effect on soil surface CO₂ efflux (Fig. 6).

There are several factors that could have reduced soil surface CO₂ efflux underneath the shelters. Litterfall from the shelter tops is returned to the sheltered soil surface each fall, providing a carbon source for microbial activity similar to litterfall outside the shelter; however, decreased throughfall inside the shelter may result in decreased quantities of...
dissolved organic carbon (DOC) at these drier locations. DOC may be a significant source of respired carbon. Altered soil temperature under the shelters also may have affected CO₂ efflux. At the floodplain site, soil temperatures at 20 cm were up to 3°C cooler inside the shelter in 1996, possibly limiting soil respiration compared with soil outside the shelter. Multiple regressions of CO₂ efflux on soil moisture and soil temperature (Table 1C) indicate that soil temperature can affect soil respiration more than soil moisture, as was the case at both sites in 1996. However, we suspect that temperature differences inside and outside the shelters were not the sole factor affecting CO₂ flux rates for three reasons.

First, the upland and floodplain shelters had opposite effects on soil temperatures at each site. At the floodplain site, soil temperatures at 20 cm were cooler outside the shelter in 1996 (P < 0.05), suggesting that decreased radiation under the shelter maintained cooler soils. At the upland site, soil temperatures at 20 cm were warmer outside the shelter (Fig. 1, P < 0.05), suggesting that the upland shelter served as a greenhouse on this southeast-facing slope.

Second, the maximum observed difference in soil temperature between soils inside and outside the shelter was 3°C at the floodplain site in 1996. At several points during this summer, an equal if not greater difference in CO₂ efflux occurred when the temperature difference was about 1–2°C.

Third, regressions indicate that parameters other than soil temperature also have influence on CO₂ efflux (Table 1A). Only on three of seven occasions tested did temperature have a significant effect on soil CO₂ efflux (P < 0.05), and when the relationships were significant, from 56 to 24% of the variation in CO₂ efflux was unexplained by soil temperature. The lower flux values under the shelter suggest that river water alone, although plentiful, does not provide the same hydrologic environment as does precipitation.

Although decreases in soil temperature may have been partially responsible for the decreases in soil respiration observed under the shelter at the floodplain site, soil temperature does not entirely explain the soil respiration responses to the shelter. At the upland site, elevated soil temperatures underneath the shelter should have resulted in increased soil respiration at those plots, if temperature alone was governing soil respiration rates. The opposite effect was seen. The data suggest that other parameters such as reductions in soil moisture may have, in part, governed the decreased rates of soil CO₂ efflux underneath the shelters. Only in 1997 at the upland site was a nonsignificant difference in CO₂ flux rates observed inside and outside the shelter, when the summer was naturally dry and moisture differences inside and outside the shelter were limited (Fig. 6B).

**Table 1.** Summary of regressions of soil CO₂ efflux on (A) soil temperature, (B) volumetric water content, and (C) soil profile concentrations of CO₂ at 20 cm, upland and floodplain sites, sheltered and unsheltered soils (dashes indicate missing data).

<table>
<thead>
<tr>
<th></th>
<th>Upland Sheltered</th>
<th>Upland Unsheltered</th>
<th>Floodplain Sheltered</th>
<th>Floodplain Unsheltered</th>
</tr>
</thead>
<tbody>
<tr>
<td>(A) Soil temperature</td>
<td>0.007</td>
<td>0.76</td>
<td>0.014</td>
<td>0.44</td>
</tr>
<tr>
<td>(B) Volumetric water content</td>
<td>0.021</td>
<td>0.37</td>
<td>0.012</td>
<td>0.42</td>
</tr>
<tr>
<td>(C) Soil profile concentrations of CO₂</td>
<td>0.0001</td>
<td>0.87</td>
<td>0.001</td>
<td>0.87</td>
</tr>
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</table>

**Fig. 4.** CO₂ profile concentrations at all depths measured, upland site, winter 1996–1997. The gap indicates new year.

CO₂ concentration, winter 1996-1997, upland site

CO₂ in these soil profiles results mainly from microbial and root respiration, but also from carbonate dissolution (Burton and Beauchamp 1994). Thus, any decrease in concentration of this gas in the profile indicates a limitation on one or more of these processes. The lower CO₂ concentrations at the upland site at 20 cm inside the shelter in both
1996 and 1997 \((P < 0.05)\) could suggest a decrease in microbial activity and (or) root respiration due to moisture limitations. Because sheltered soils at the upland site were warmer than unsheltered soil, it is doubtful that temperature alterations induced by the shelter caused this decrease in CO2 concentrations at 20 cm. At the floodplain site, the same pattern of lower CO2 concentrations at 20 cm in the profile is visible in 1996 but is not significant at that site either year. Lower CO2 concentrations within the soil profile at both sites may have occurred as a result of decreased throughfall, which may contain significant quantities of DOC.

The significantly lower concentrations at the floodplain at 100 cm under the shelter in 1996 are particularly surprising given the lack of an effect at the more shallow depths that year (Fig. 5A). The low soil temperatures at 100 cm in the sheltered area probably affected these concentrations. The 100-cm layer inside the shelter at this site remained frozen for the first half of the summer both years. At the end of July, when this soil thawed, a pulse of CO2 was observed. This may result from a flush of carbon available from microbes killed by freezing (Skogland et al. 1988; Burton and Beauchamp 1994), stimulation of microbial activity with warmer temperatures, or the release of CO2 previously trapped in soil ice (Coxson and Parkinson 1987).

Dramatically higher floodplain CO2 concentrations at 40 and 100 cm compared with the upland site probably result from the buried organic layers in the floodplain soil profile from past flooding events or from carbonate dissolution as river water percolates up into the profile (Marion et al. 1993a, 1993b). By limiting diffusion to the aboveground atmosphere, higher soil moisture contents partially inhibit flux of this CO2 to the soil surface, resulting in higher CO2 concentrations.

High soil water contents at the floodplain can affect soil profile CO2 concentrations via soil chemistry as freeze–thaw events occur. As ice in the profile thaws, solubility decreases as water temperature increases, and CO2 is released (Dormaar and Sauerbeck 1983). Pulses of CO2 concentrations in the spring likely reflect this process. We expect that profile CO2 concentrations at the floodplain are mediated by the ability of cold soil water to absorb aqueous CO2 (Stumm and Morgan 1981). This effect is largely absent at the upland site, where soil moisture content at 10 and 20 cm was commonly less than 10% VMC, and even less deeper in the profile.

The wintertime profile concentrations at the upland site show how the effects of the shelter are diminished with increasingly colder temperatures and a deep snowpack (Fig. 4). By mid-October, approximately 60 cm of snow covered the site. Air temperatures were frequently \(-30^\circ\text{C}\), but the coldest soil temperatures at 20 cm were about \(-4^\circ\text{C}\) because of the insulating snowpack. The CO2 concentrations at 20 cm still were much higher than the aboveground atmosphere well into November and remained at over twice atmospheric concentrations through the winter. This CO2 was either still producing surface efflux or was being dissolved into the remaining soil water. Given the water limitations at this site, we can assume that the CO2 eventually reached the surface and either escaped directly through the snowpack or accumulated underneath it, to be degassed in
pulses with any snow disturbance. Fluxes from soil surface also may be dissolved into melting snow (Solomon and Cerling 1987). Burton and Beauchamp’s (1994) findings of high CO₂ concentrations within the profile during winter months support our results. They reported high accumulations of CO₂ (up to 4000 µL·L⁻¹) in the early winter during one year of their study, which corresponds to our findings.

There is a strong positive relationship between soil CO₂ efflux and the 20-cm profile CO₂ concentration at the upland site in 1996 (Table 2). If the linearity of this relationship between upland soil respiration and CO₂ concentration at 20 cm holds true through winter months, the wintertime profile data indicate that up to 33% of the summer upland flux could occur from October to April. This corroborates well the study of Coxson and Parkinson (1987) where winter soil respiration rates were sufficiently high to account for the amount of winter litter decomposition in a Canadian aspen forest.

The production of CO₂ is linked to water movement through these soil systems. Increases in profile CO₂ concentrations at various times throughout both summers and at both sites occurred when soil moisture increased.

Fig. 6. CO₂ efflux rates from sheltered and unsheltered soils, upland and floodplain sites, 1996 and 1997. *P* values are from a *t* test for equal means; error bars are standard errors of the mean. When data were non-normally distributed, the Wilcoxon rank-sum test was used. Open circles indicate unsheltered soils; solid circles indicate sheltered treatment soils.

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indicate soil-forming events in addition to biogenic processes (Burton and Beauchamp 1994). Precipitation leaching through the soil profile can serve as an agent of soil formation, dissolving CaCO₃ and producing CO₂ in the process.

At the floodplain site, the dissolution of CaCO₃ as a result of leaching has been documented (Marion et al. 1993a, 1993b). The upland site soils are loessal silt, derived from the same parent material in the Alaska range as the soils in the Tanana River floodplain, and thus can contain significant quantities of CaCO₃. The lower CO₂ concentrations inside the shelters could reflect limited CaCO₃ dissolution due to limited precipitation as well as limited microbial and (or) root respiration. This is supported by Burton and Beauchamp (1994), who suggested that seasonally variable pedogenesis via CaCO₃ dissolution may be responsible for a significant part of total soil CO₂ efflux in Canadian agricultural soils.

**Fick’s Law predictions**

Because of the inherent difficulties in determining the relationships between soil water limitations, soil temperature, and soil respiration, we attempted to predict soil CO₂ efflux using Fick’s Law. Estimating the coefficients for D, the diffusion of CO₂ in air, is a major constraint on the accuracy of Fick’s Law. We altered the exponent on φ from 1.5 (Marshall 1959) to 1.1 to empirically fit the calculated flux to measured flux. However, this ignores other significant factors affecting the ability of Fick’s Law to predict soil surface gas fluxes.

The ability of Fick’s Law to predict soil surface CO₂ efflux depends also on how closely values of ∂C/∂z represented the concentration gradient across the soil surface. When soil moisture was below 15% VMC at the upland site, the measured gradient between 20 cm in the profile and immediately adjacent to the soil surface more closely reflected the soil surface gradient; when soil moisture was above 15% VMC, the measured 20-cm concentration gradient poorly represented the soil surface gradient, resulting in flux estimates that did not match measured values (Figs. 7A and 7B). At the floodplain site, soil moisture was consistently high enough to create discrepancies between the measured concentration gradient and the gradient near the soil surface needed to compute the CO₂ flux at z = 0 (Figs. 7C and 7D).

**Fig. 7.** Calculated versus measured soil CO₂ efflux, upland and floodplain sites, 1996 and 1997. Circles indicate unsheltered sites; squares indicate sheltered treatment soils. Close correspondence would be indicated by the points falling on a one-to-one line, similar to Fig. 7A.
Table 2. Summary of multiple regressions of soil CO₂ efflux on soil temperature at 20 cm and volumetric water content at 20 cm, upland and floodplain sites, sheltered and unsheltered soils.

<table>
<thead>
<tr>
<th></th>
<th>Sheltered</th>
<th>Unsheltered</th>
</tr>
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<tbody>
<tr>
<td></td>
<td>(P_{\text{temp}})</td>
<td>(R_{\text{H,O}})</td>
</tr>
<tr>
<td>Upland</td>
<td>0.633</td>
<td>0.410</td>
</tr>
<tr>
<td>1997</td>
<td>0.633</td>
<td>0.410</td>
</tr>
<tr>
<td>Floodplain</td>
<td>0.002</td>
<td>0.437</td>
</tr>
<tr>
<td>1996</td>
<td>0.002</td>
<td>0.437</td>
</tr>
<tr>
<td>1997</td>
<td>0.647</td>
<td>0.298</td>
</tr>
</tbody>
</table>

If we assume that the concentration of CO₂ increases linearly with depth from the soil surface, the gradient from 20 cm to the aboveground atmosphere more accurately represents the gradient at the midpoint of the 20-cm soil compartment, at 10 cm, rather than the gradient across the soil surface. Thus the values used for \(\partial C/\partial z\) in Fick’s calculations result in flux estimates that more aptly reflect fluxes within the soil profile at 10 cm. Given the generally shallow rooting zone of the boreal forest, significant biological activity occurs in the top 10 cm of the soil profile, creating discrepancies between measured and calculated fluxes. This is best described by the one-dimensional continuity equation, which states that the change in gas concentration over a time increment equals the change in gas flux with depth, plus any locally produced or consumed gas within that depth (de Jong and Schappert 1972, Striegl 1993). At any given point in the soil:

\[
\frac{\partial C}{\partial t} = \frac{\partial q}{\partial z} + S
\]

If we integrate over depth \(z\), we obtain

\[
z \cdot \frac{\partial C}{\partial t} = q_0 - q_z + z \cdot S
\]

where \(q_0\) is the flux at the soil surface plane and \(q_z\) is the flux across the plane at depth \(z\), the overbar symbol indicates depth-averaging over \([0, z]\), and \(S\) is local sources and sinks within depth \(z\). Hence, \(q_0\) (measured surface CO₂ efflux) will equal \(q_z\) (calculated Fickian flux) only if the source of CO₂ in the soil profile compartment exactly equals the change in concentration with time. This is not likely, since over the time interval that \(q_0\) was measured (45 min), \(\partial C/\partial t = 0\). Differences between flux calculations \((q_z)\) and measurements \((q_0)\) (Figs. 7A and 7B) thus likely reflect the effect on \(S\) of the proliferation of roots and, we assume, microbes in the top 10 cm of the soil profile.

Several studies have used Fick’s Law to estimate soil surface gas fluxes with \(\partial C/\partial z\) calculated in various ways (Potter et al. 1996; Dunfield et al. 1995; Yavitt et al. 1990, 1995; Burton and Beauchamp 1994; Sommerfeld et al. 1993; Whalen et al. 1992). Fick’s Law is best applied when \(\partial C/\partial z\) estimates most accurately reflect the concentration gradient across the soil surface plane. When concentration gradients across the top layers of the soil profile are used as a replacement for the soil surface gradient, estimates of surface flux from Fick’s Law must take conservation of mass into consideration, particularly when the gas in question has prominent sources or sinks in the top layers of the soil profile. This is the case in boreal forests such as these, where the rooting zone can be as shallow as 20 cm.

Because the boreal forest contains such large amounts of soil carbon, it is important to learn the effects of altered moisture on root and soil microbial activity. If global climate model predictions prove true, high-latitude regions will experience both shifts in precipitation patterns and increases in temperature (Manabe and Wetherald 1986; Wilson and Mitchell 1987; Schlesinger and Zhao 1989). If this results in a reduction in water availability for these interior Alaskan forests, soil respiration from this boreal forest could be significantly reduced. However, the net effect of increasing temperature and soil thaw season over longer time periods may increase forests’ productivity and result in higher net soil CO₂ efflux. Further studies are needed to quantify the area of upland versus floodplain forest types, and the predicted future water supply in these regions, in order to more accurately quantify the changes in carbon gas fluxes across boreal landscapes such as these.

References


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