SIBERIAN CO_2 EFFLUX IN WINTER AS A CO_2 SOURCE AND CAUSE OF SEASONALITY IN ATMOSPHERIC CO_2

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Abstract. Over three years, we found a consistent CO_2 efflux from forest tundra of the Russian North throughout the year, including a large (89 g C m^{-2} yr^{-1}) efflux during winter. Our results provide one explanation for the observations that the highest atmospheric CO_2 concentration and greatest seasonal amplitude occur at high latitudes rather than over the mid-latitudes, where fossil fuel sources are large, and where high summer productivity offset by winter respiration should give large seasonal oscillations in atmospheric CO_2. Winter respiration probably contributed substantially to the boreal winter CO_2 efflux. Respiration is an exothermic process that produces enough heat to warm soils and promote further decomposition. We suggest that, as a result of this positive feedback, small changes in surface heat flux, associated with human activities in the North or with regional or global warming, could release large quantities of organic carbon that are presently stored in permafrost.

1. Introduction

High northern latitudes (53-83° N) are unique in exhibiting the greatest seasonal amplitude in atmospheric CO_2 concentration (Fung et al., 1987; Tans et al., 1990; Denning et al., 1995). This is surprising because, at high latitudes, photosynthetic carbon gain in summer (depleting atmospheric CO_2) is generally low (Wielgolaski et al., 1981), and frozen soils should prevent respiration from causing a large CO_2 efflux to the atmosphere in winter. Data-based global carbon models must assume twice the net primary production (NPP) measured in northern ecosystems in order to generate observed seasonal patterns of atmospheric CO_2 at high latitudes (Fung et al., 1987). This discrepancy could result from large (2-fold) errors in measurements of NPP, greater winter respiration than generally assumed (previously assumed to be negligible), violation of the steady-state assumptions of the model (if northern ecosystems have a non-zero carbon balance), or there may be important processes not captured by the model. Recent measurements in temperate alpine ecosystems indicate substantial winter respiration beneath the snow (Sommerfeld et al., 1993), but only preliminary measurements of respiration during one winter have been

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reported from the permafrost-dominated Arctic (Zimov et al., 1993a; Zimov et al., 1993c).

High concentrations of atmospheric CO_2 occur over the tundra and boreal forest above 53° N latitude (Tucker et al., 1986; D'Arrigo et al., 1987; Fung et al., 1987; Zavarzin and Clark, 1987; Tans et al., 1990; Denning et al., 1995), suggesting that northern ecosystems today could be large *net* sources of CO_2 per unit area. Indeed, a recent study suggested that, although there is a large terrestrial carbon sink from 30-60° N, latitudes north of 60° N may be a net CO_2 source (Ciais et al., 1995). Although 95% of fossil fuels are burned in the belt between 20° and 60° N latitude (Houghton et al., 1990), this region may not be a strong *net* source, if CO_2 inputs from fossil fuel combustion are balanced by increased productivity in response to elevated CO_2, nitrogen deposition (Tans et al., 1990; Kauppi et al., 1992; Ciais et al., 1995), and reforestation following agricultural abandonment (Wofsy et al., 1993). Some CO_2 measured at high latitudes is delivered from mid-latitudes by northward-moving winds (D'Arrigo et al., 1987), but this cannot explain why *maximum* atmospheric concentrations occur at high latitudes. Other factors contributing to the high atmospheric CO_2 concentrations at high latitudes include the high ratio of land to ocean surface and latitudinal variation in vertical atmospheric mixing. Northern ecosystems could be a large net source of CO_2, if their large pools of soil carbon (25% of the world's soil organic carbon or 350 GT; Post et al., 1982; Billings, 1987) are being released by soil respiration, as suggested by recent measurements in Alaska (Oechel et al., 1993) and Russia (Zimov et al., 1993c). If northern ecosystems exhibit a large net efflux of CO_2 to the atmosphere, this would serve as a positive feedback to augment the rate of global warming.

In this article we address three questions: Is there sufficient summer photosynthesis or winter respiration in northern ecosystems to explain the large seasonal fluctuations observed in atmospheric CO_2 at high latitudes? What processes account for winter CO_2 flux in the North? Are northern ecosystems a net source of atmospheric CO_2? To address these questions we began daily monitoring of CO_2 flux in September 1989 in an area of forest tundra in northeast Russia. Here we summarize results from the first three years of that study.

2. Methods

The study was conducted at the North-East Scientific Station of the Pacific Institute for Geography near Cherskii, Yakutia, Russia (69° N, 161° E) 10 km east of the Kolyma River and 100 km south of the Arctic Ocean. The region is characterized by a mosaic of wet tundra, shrub tundra, forest tundra, and boreal larch forest. After an initial survey of net CO_2 flux from ground-layer vegetation and underlying soil in 135 sites, spanning the regional variation in vegetation (Zimov et al., 1993b), we selected an area of forest tundra adjacent to the Cherskii station whose net CO_2 flux was close to the median value of the region as a whole, providing a basis for regional extrapolation. Choice of a site close to the station also made it logistically feasible to measure CO_2 flux frequently. The study area is on upland loess deposits capped by a 5–15 cm thick organic mat typical of the region. The vegetation consists of scattered larch (*Larix dahurica*) trees 3–8 m tall with an understory of 0.5-m tall shrub birch (*Betula middendorffii*) and a carpet of ericaceous dwarf shrubs (*Vaccinium uliginosum*, *Vaccinium vitis-idaea*, *Ledum palustre*, *Empetrum nigrum*, *Arctous alpina*), grasses (*Calamagrostis* spp.), mosses (*Hylocomium splendens* and *Sphagnum* spp.), and lichens (*Cladonia* spp., *Cetraria* spp.) (Zimov et al., 1993b).

In the first year we measured CO_2 flux daily in winter (September 1989 to May 1990, $n = 80$ plots) and summer (June to August 1990, $n = 15$ plots). Measurements in this first year showed a surprisingly high winter CO_2 efflux to the atmosphere (30–200 g C m^{-2} yr^{-1}) (Zimov et al., 1993a), so we continued the measurements for an additional two years in nine representative vegetated plots and one plot from which vegetation and litter were removed. These 10 permanent plots were selected in winter beneath the snow at random points along a 200-m transect. The vegetated plots included lichens, mosses, grasses, and dwarf shrubs but excluded scattered large shrubs and trees. Here we summarize the results of 11,000 CO_2 flux measurements from these 10 plots for 1990–92.

From May through October we measured CO_2 flux using the chamber technique (Whalen and Reeburgh, 1988). Each permanent plot was surrounded by a 26 cm-diameter metal cylinder that extended 5 cm below and 5-20 cm above the soil surface. During summer measurements, each cylinder was covered with a transparent plastic chamber. Air samples were collected with a 300 ml syringe from each chamber 5–15 minutes after initiation of the measurement and transferred to a valved rubber sample bag. Samples from each chamber and of ambient air were immediately analyzed with a GIAM-SM infrared gas analyzer with an accuracy of ± 1 ppm, calibrated against standards from the Scripps Institute of Oceanography (Zimov et al., 1993c). CO_2 flux was calculated, taking into account the length of the sample interval, the chamber volume, and the change in CO_2 concentration.

Measurements of each chamber were made each day at noon throughout the 3-yr study (except May 1990). To estimate daily flux we measured the daily time course of CO_2 flux at 3-hr intervals once a week during the summer in the three plots shown in the figure. Daily fluxes on intermediate days were estimated for each chamber by (1) calculating the difference (p) between the noon flux and the average daily flux for each date in which diurnal measurements were made, (2) linearly interpolating the value of p between consecutive dates with diurnal flux measurements, (3) adding this interpolated value of p to the noon flux measured on each day, and (4) expressing this flux on a 24-h basis. The average noon flux for all 10 chambers was calculated for each day. This mean flux at noon was converted into a daily flux using the average value of p from the chambers and the interpolation procedure described above.

In the winter, when we could not use the chambers without disturbing the snow profile, we estimated CO_2 flux from the CO_2 concentration gradient between the top and bottom of the snow pack. CO_2 concentration at the bottom of the soil profile was measured from a rubber tube attached to 300 ml reservoirs (inverted plastic bottles) placed at the soil surface in autumn. Knowing the concentration in air and at the base of the snow pack and the effective diffusion coefficient (Zimov et al., 1993b, 1993c), we calculated CO_2 flux. We calibrated this technique against the chamber technique through simultaneous measurements of the CO_2 gradient and CO_2 flux using one tall chamber in November–April. This calibration gave us a direct measurement of the effective diffusion coefficient (d) through snow ($d = F/(C_b - C_a)$, where F is the flux measured with a chamber, C_b is the CO_2 concentration below the snow, and C_a is ambient CO_2 concentration above the snow. Details of the measurements and validation of the procedure are presented elsewhere (Zimov et al., 1993b, 1993c). The procedure is similar to that used to measure winter respiration in alpine meadows (Sommerfeld et al., 1993).

To determine whether sampling artifacts might be caused by the frequent revisitation of the same sites over the 3-yr study or, by inadequate mixing within these chambers, or by shading from the chamber base, we made 21 measurements with a chamber 2.5 × 2.5 × 0.6 m that contained large *Betula* shrubs but lacked trees. Five simultaneous measurements were made in different parts of the chamber 2, 4, 8, 16, and 32 minutes after installing the chamber. Fluxes estimated from this large chamber did not change with time and did not differ from those measured in our smaller chambers. Thus, it is not likely that disturbance, shading by the chamber collars, or lack of mixing created artifacts in our measurements.

As a further check for bias in our measurements, we measured the vertical CO_2 concentration gradient in the afternoon on a clear day in June, 1995, when photosynthesis should have been maximal, between the surface of tall shrubs (50 cm height – approximate zero plane displacement) and the top of the tree canopy (5 m) to determine whether tree photosynthesis might offset the CO_2 efflux we measured systematically in the ground layer. At four points along the transect we measured CO_2 concentration at 1-min measurements first at the top of the shrub canopy and then above the tree canopy with a LICOR-6250 portable infrared gas analyzer. This measurement sequence was repeated five times at each of the four points along the transect.

Our measurements and any other non-continuous measurements are subject to errors of extrapolation due to diurnal and seasonal variations in weather. Our diurnal measurements, which were made every 7 days, are a larger proportion (14%) of the growing-season days than in any previously published study of CO_2 flux. By making additional noon measurements *every* day, we could account for day-to-day variations in weather, which has not been attempted in any previous study.

To determine the possible source of winter CO_2 efflux, we excavated two soil columns 90 cm in diameter to the maximum depth of thawed soil (80 cm) in early September. In the center of each profile we placed thermocouples and gas-sampling reservoirs at 10, 20, 50, and 80 cm depth, and in one of the profiles we placed 4 kg

Table I

Concentrations of CO_2 measured at the bottom (50 cm) and top (5 m) of the forest canopy at the Cherskii study site on a clear afternoon. Positive values for the CO_2 gradient indicate highest concentration near the ground surface (flux to the atmosphere). Data are means $\pm$ SE, $n = 5$

Profile	CO_2 concentration (ppm)		CO_2 gradient
Number	50 cm	5 m	(ppm)
1	362.70 + 0.02	362.20 + 0.05	0.50
2	362.35 + 0.02	362.03 + 0.04	0.32
3	362.78 + 0.07	362.58 + 0.02	0.20
4	363.37 + 0.06	362.90 + 0.63	0.47

of wheat flour as a labile carbon source at 50–80 cm depth. Soil was replaced in the two profiles, preserving the original soil profile and bulk density as closely as possible. We monitored temperature and CO_2 flux weekly from September through November. Because these profiles were unreplicated, we interpret these data only qualitatively.

3. Results

When all plots were averaged, there was a net CO_2 efflux from ground-layer vegetation to the atmosphere in each month of the three-year study (Fig. 1a). In only one plot, where grasses dominated, was there any net monthly carbon gain. Net efflux was generally highest in summer, when temperatures were above freezing. Within the summer period, efflux was highest in the wettest year (1990), indicating that soil drying did not promote summer respiration in this arid Kolyma climate. This contrasts with the wetter Alaskan tundra, where dry conditions may enhance CO_2 efflux (Oechel et al., 1993). Total net efflux during the summer averaged 150 ± 49 g C m^{-2} yr^{-1} ($n = 3$ yr), excluding the net aboveground carbon gain of trees and large shrubs, which in this ecosystem was about 25 g C m^{-2} yr^{-1} (estimated from harvest of current year's aboveground growth of the scattered trees and shrubs). Thus, during summer, there was probably a net efflux of 75–175 g C m^{-2} yr^{-1}, similar to rates of 35–160 g C m^{-2} yr^{-1} measured in Alaskan tundra (Oechel et al., 1993).

In June, when conditions were favorable for photosynthesis, atmospheric CO_2 concentration was consistently lower at the top than at the bottom of the forest canopy (Table 1), consistent with our conclusion that this ecosystem was, at the time of measurement, a net CO_2 source to the atmosphere.

Winter CO_2 flux was highest in September–April and in May and minimal in December–April, when temperatures were lowest, suggesting a strong temperature dependence of winter CO_2 flux (Figure 1). Winter flux showed no consistent diurnal

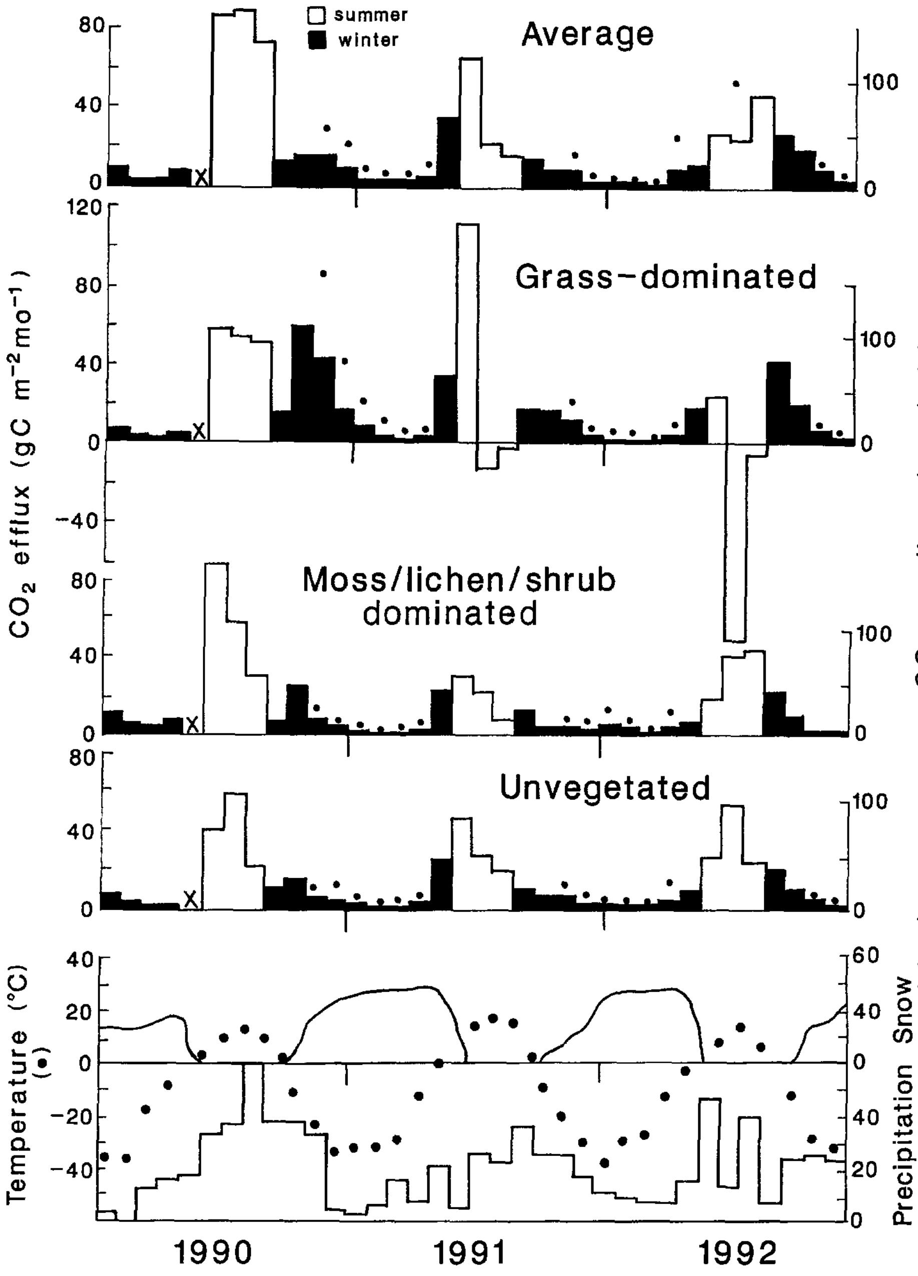

Figure 1. Three-year record of monthly CO_2 efflux from the ground layer of forest tundra, measured in the Kolyma Lowlands of northern Russia. Each bar is the total monthly CO_2 flux during winter (filled bars) and summer (open bars). *X* indicates no data collected. Positive values indicate flux from land to the atmosphere. Also shown is the average concentration gradient of CO_2 through the snowpack in winter, average monthly air temperature and precipitation (measured every 3 hr), and snow depth (measured daily). CO_2 flux is shown for the average of all plots ($n = 10$), a representative moss/lichen/dwarf-shrub-dominated plot, a productive grass-dominated plot, and a moss/lichen plot cleared of vegetation and litter.

pattern (data not shown). Assuming negligible aboveground respiration by large shrubs and trees (which we did not measure), winter CO_2 efflux averaged 89 (83–94) g C m^{-2} yr^{-1} across the three years of measurement. Thus, total efflux from this ecosystem during the 9-month winter was probably similar in magnitude to that which occurred during the 3-month summer.

The winter flux of CO_2 could derive from either (1) gases produced during the summer, which gradually leak out during the winter or (2) winter soil respiration. If we assume the *maximum* soil CO_2 concentration (0.4% at 50 cm depth) observed in the soil atmosphere during the three-month period of weekly monitoring (Sept. to Nov.) and maximum observed thaw depth (80 cm), the total CO_2 stored in the soil profile would be 0.43 g C m^{-2} in the gas phase and 147 g C m^{-2} dissolved in the soil solution (assuming CO_2 saturation at 0 °C of the estimated 160 L water m^{-2}). Together these sources of CO_2 stored in the soil profile could account for the observed winter flux. Addition of wheat flour (a labile carbon source) to soils in September caused a 4-fold increase in surface CO_2 flux (from 16 to 70 g CO_2-C mo^{-1}) during September to November, when soil temperatures at 50–80 cm were 0.2–0.4 °C, indicating that microbial respiration can occur at substantial rates in winter. Moreover, the positive correlation between CO_2 efflux and temperature in both early and late winter (Figure 1) are more readily explained by respiration than by a gradual outgasing of CO_2 during winter. Laboratory incubation of these soils showed substantial respiration at temperatures exceeding –4 °C (Zimov et al., 1993c), indicating a very low temperature threshold for respiration in these northern soils. We conclude that winter respiration contributed to winter CO_2 efflux at our site.

4. Discussion

The results presented here and the preliminary results from the first year (Zimov et al., 1993a) are the only complete winter budgets of CO_2 efflux from northern ecosystems of which we are aware, so it is difficult to know how broadly they can be extrapolated. If winter CO_2 efflux to the atmosphere is extrapolated to progressively larger areas (but with progressively less certainty), northern loess-covered, permafrost-dominated landscapes of northern Siberia (1×10^6 km^2; Tomirdiaro, 1980) would contribute 90 Tg C yr^{-1}; larch-dominated Siberian forests (4.7×10^6 km^2; Chapin and Matthews, 1993) would yield 400 Tg C yr^{-1}; and boreal needle-leafed forests (12.3×10^6 km^2; Chapin and Matthews, 1993) would give 1100 Tg C yr^{-1}. These winter CO_2 effluxes are equivalent to 3–42% of the observed annual fluctuation in atmospheric CO_2, north of 66° N. [This calculation assumes that 8% of the earth's surface and the atmospheric CO_2 pool is north of 66° N and that the annual amplitude in atmospheric CO_2 for this region is 16 ppm (Fung et al., 1987).] For comparison, Oechel et al. (1993) estimate that arctic tundra contributes 190

Tg C yr^{-1} during summer, based on fluxes similar to those that we measured in summer.

Global calculations also suggest that the winter effluxes we measured in Siberia are globally significant. Global NPP and respiration are about 60 Gt C yr^{-1}, of which 85% occurs simultaneously, much of this in the tropics (I. Fung, personal communication). Thus, the seasonal cycle of atmospheric CO_2 reflects the 15% of NPP and respiration (i.e., 9 Gt C yr^{-1}) that are seasonally asynchronous. Clearly, our estimated winter efflux from boreal forests (0.09–1.1 Gt C yr^{-1}) could be a globally significant cause of fluctuation in atmospheric CO_2.

The CO_2 concentration that we measured beneath the snow in winter was similar to that measured in arctic tundra (Kelley et al., 1968; Coyne and Kelley, 1974) and temperate alpine (Sommerfeld et al., 1993), suggesting that many snow-covered ecosystems could exhibit large winter CO_2 effluxes, such as we measured. Thus, winter CO_2 efflux is a major component of annual CO_2 flux in boreal Russia, where we measured it and, if general, could contribute to the high atmospheric CO_2 concentration observed in winter in the Far North. If this phenomenon is widespread, it suggests that the large seasonal amplitude of atmospheric CO_2 at high latitudes is due more to a substantial CO_2 efflux in winter rather than to high productivity in summer. Consequently, smaller values of NPP at high latitudes would be required to generate the observed seasonal fluctuation in atmospheric CO_2 at high latitudes than was previously assumed (Fung et al., 1987).

Winter respiration has important implications for the soil thermal regime. Biological oxidation of organic matter to CO_2 is an exothermic reaction releasing approximately 8 kcal g^{-1} C (3–5 kcal g^{-1} organic matter or 6–10 kcal g^{-1} organic C) (Hodgman, 1959). Thus, the winter CO_2 efflux of 89 g C m^{-2} must have released 700 kcal m^{-2} (3 MJ m^{-2}) of heat. This heat derived from respiration could allow soils to remain unfrozen and biologically active longer during the winter, creating a positive feedback that promotes further CO_2 release. The impact of respiratory heat input to soil thermal regime depends on the timing of respiration, the depth of winter snow, and the thermal conductivity of surface soil and snow (Outcalt et al., 1990, 1992).

Obviously, a net annual CO_2 efflux from an ecosystem of 150–300 g C m^{-2} yr^{-1} (our estimated annual flux) cannot be sustained indefinitely. Since the Pleistocene, northern soils have accumulated carbon because of the combined effects of low temperature and waterlogged soils characteristic of permafrost terrain (Billings, 1987). What has reversed this trend? We suggest that some combination of regional or global warming (Hansen and Lebedeff, 1987; Chapman and Walsh, 1993; Oechel et al., 1993) or a decline in the thermally insulating cover of lichens and mosses (Van Cleve et al., 1991) may have increased heat flux into northern soils. Any decline in mosses and lichens due to their sensitivity to pollutants, which have increased substantially in the Arctic in recent decades (Jaffe et al., 1991), could increase the summer energy influx to arctic permafrost soils, perhaps contributing to the unexplained warming of Alaskan permafrost in the past 40 years (Lachenbruch and

Marshall, 1986). Because of the positive feedback caused by the heat input from respiration, small changes in surface thermal regime could be amplified through time, melting the permafrost, exposing additional soil carbon to decomposition, and increasing atmospheric CO_2 concentration and its greenhouse effect on global temperature.

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