Sensitivity of Boreal Forest Carbon Balance to Soil Thaw

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We used eddy covariance; gas-exchange chambers; radiocarbon analysis; wood, moss, and soil inventories; and laboratory incubations to measure the carbon balance of a 120-year-old black spruce forest in Manitoba, Canada. The site lost 0.3 ± 0.5 metric ton of carbon per hectare per year (ton C ha⁻¹ year⁻¹) from 1994 to 1997, with a gain of 0.6 ± 0.2 ton C ha⁻¹ year⁻¹ in moss and wood offset by a loss of 0.8 ± 0.5 ton C ha⁻¹ year⁻¹ from the soil. The soil remained frozen most of the year, and the decomposition of organic matter in the soil increased 10-fold upon thawing. The stability of the soil carbon pool (~150 tons C ha⁻¹) appears sensitive to the depth and duration of thaw, and climatic changes that promote thaw are likely to cause a net efflux of carbon dioxide from the site.

The climate of the boreal zone has warmed significantly in this century and is predicted to warm further in the next century (1). The seasonally and perennially frozen soils of boreal forests contain one of the largest pools of carbon in the terrestrial biosphere, 200 to 500 gigatons of carbon (1 Gt = 10¹⁰ metric tons) (2), an amount that could increase the concentration of CO₂ in the atmosphere by as much as 50% if it were released by climatic warming (3, 4). However, the sensitivity of boreal forest carbon balance to temperature, and hence the magnitude and even direction of such a feedback, is unknown.

We used the eddy covariance technique (5), soil gas-exchange chambers (6), natural abundance radiocarbon analysis (7), and wood, moss, and soil inventories (8, 9) to investigate the carbon balance of a 120-year-old black spruce forest in Manitoba, Canada, from 1994 to 1997 (10). The site was typical of much of the North American boreal forest (11). Natural vegetation extended for several kilometers in all directions across a flat landscape. Dense black spruce (10 m tall) and feather moss grew in slightly higher, better drained areas, and sparse spruce (1 to 6 m tall) and sphagnum moss grew in lower, wetter locations. About 45% of the area within 500 m of the eddy flux tower was covered by feather moss, 45% was covered by sphagnum moss, and the remaining 10% was covered by fen (8).

The winters were harsh. The minimum temperature observed was −52°C and 3 or 4 months of each year had daily highs of −10°C to −25°C (Fig. 1C). The growing season began and ended with abrupt temperature transitions, usually in May and October. Daytime temperatures in the growing season were 15°C to 25°C, and July and August were nearly frost free. The soil and lower moss layers warmed gradually in summer and stayed below air temperature until October (Fig. 1C) (12). Patches of soil remained frozen in late summer beneath sphagnum hummocks and beneath feather moss at depths of 75 to 100 cm, a pattern of discontinuous permafrost that is common in central Canada (13).

The site contained 40 ± 20 tons C ha⁻¹ in aboveground live and dead spruce (9) and 45 ± 13 tons C ha⁻¹ in live and dead moss (7, 8). The largest pool of carbon was in the soil—200 ± 50 tons C ha⁻¹ in the sphagnum areas and 90 ± 20 tons C ha⁻¹ in the feather moss areas (8). These stocks are typical of northern forests (2). Most of the soil carbon was in a humified layer we refer to as the deep pool, which was ~40 to 80 cm beneath the surface and just below the depth penetrated by fires. The deep carbon had bulk ages of several hundred to 7000 ¹⁴C years before present (7) and is considered to be nearly inert in most biogeochemical models.

We made eddy covariance measurements of whole forest CO₂ exchange for 22,000 hours from March 1994 to October 1997 (5). Daily net uptake started in early May, peaked at 10 to 15 kg C ha⁻¹ day⁻¹ from late May to early July, and declined to near zero in August and September (Fig. 1A). The forest lost 6 to 8 kg C ha⁻¹ day⁻¹ in October and 2 to 3 kg C ha⁻¹ day⁻¹ from December to April. The absolute rates and seasonal course of exchange were markedly different from those ob-

Fig. 1. (A) Daily carbon balance measured by eddy correlation (NEE), (B) Accumulated CO₂ exchange (ΣNEE) starting at zero on 1 October 1994. (C) Daily mean temperatures at a height of 30 m (——), 10 cm beneath the sphagnum moss surface (– – –), 20 cm beneath the surface (—), 50 cm beneath the surface (T_{soil} (— – – – –), and 100 cm beneath the surface (— – – –). Lines connect means for 7 days with complete observations; unconnected points in (A) indicate a gap in the observations of at least 13 days. Vertical bands begin 1 July and end 30 September. Positive fluxes in (A) indicate a net efflux of CO₂ from the forest (5, 14, 15).

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served above a deciduous oak and maple stand at the Harvard Forest in central Massachusetts (14). Carbon accumulation at Harvard Forest began in late May and continued at a comparatively high rate into September. Net uptake at Harvard Forest lagged that at the evergreen boreal site by a month or more even though Harvard Forest is 13° farther south.

The integrated eddy flux measurements at the boreal site showed a loss of 0.7 ± 0.5 ton C ha\(^{-1}\) from October 1994 to October 1995, a loss of 0.2 ± 0.5 ton C ha\(^{-1}\) from October 1995 to October 1996, and a gain of 0.1 ± 0.5 ton C ha\(^{-1}\) from October 1996 to October 1997 (Fig. 1B) (15, 16). We found no evidence of high rates of carbon accumulation, possibly associated with nitrogen deposition, elevated CO\(_2\), or climatic warming, as has been suggested is occurring in the boreal zone on the basis of analyses of atmospheric CO\(_2\) (17). We emphasize, however, that the forest is relatively old, and stands in the area that are younger or dominated by deciduous vegetation may be accumulating carbon at high rates. The annual gross production remained consistent from year to year at ~8.0 tons C ha\(^{-1}\) year\(^{-1}\), whereas annual respiration ranged from 8.9 tons C ha\(^{-1}\) year\(^{-1}\) in 1995 to 7.9 tons C ha\(^{-1}\) year\(^{-1}\) in 1997 (18). The carbon balance was the small residual of two much larger fluxes, a pattern that may make it extremely sensitive to climate.

Analyses of forest recovery after fire indicated that, on average, 120-year-old black spruce stands in the area accumulate 0.2 to 0.4 ton C ha\(^{-1}\) year\(^{-1}\) in the moss layer (8). Measurements of stem increment at the site showed that 0.2 to 0.4 ton C ha\(^{-1}\) year\(^{-1}\) has been stored in woody biomass during the past 10 years (9). This combined gain (0.6 ± 0.2 ton C ha\(^{-1}\) year\(^{-1}\)) was lost from the deep pool.

A direct measurement of the loss of soil carbon was not feasible because of the heterogeneity, although several lines of indirect evidence support this conclusion. The midsummer decline in net uptake (Fig. 1A) was associated with an increase in respiration. Respiration increased by ~10 kg C ha\(^{-1}\) day\(^{-1}\) from July to August, whereas photosynthesis remained relatively constant except for a decline of ~5 kg C ha\(^{-1}\) day\(^{-1}\) caused by decreasing day length. The midsummer increase in respiration was not caused by warmer air temperatures (Fig. 1C); whole-forest respiration, after controlling for air temperature by considering only 8° to 12°C windy nights, increased from 41 kg C ha\(^{-1}\) day\(^{-1}\) in July to 51 kg C ha\(^{-1}\) day\(^{-1}\) in August (Fig. 2). Rather, the midsummer increase in respiration coincided with warming at depths of 20 to 100 cm (Fig. 1C).

The nocturnal CO\(_2\) efflux at the moss surface measured with automated chambers (6) showed a midsummer increase comparable to that observed by eddy covariance. From June to August moss surface respiration increased by 8 to 15 kg C ha\(^{-1}\) day\(^{-1}\) at all three sphagnum sites measured independent of moss temperature and by 10 kg C ha\(^{-1}\) day\(^{-1}\) at one of two feather sites (Fig. 3A). This seasonal increase in respiration (\(R_{\text{deep}}\)) (19) was correlated with the soil temperature at a depth of 50 cm (\(T_{50}\)) (Fig. 3B). These observations eliminate the possibilities that aboveground or moss-layer sources were responsible for the midsummer increase in respiration because the increase was insensitive to diurnal and synoptic variations in air temperature. The measurements are also inconsistent with an increase in root growth being responsible because the increase was smallest beneath feather moss, where root respiration appears to be greatest (20). Instead, the observations imply that the decomposition of deep carbon increased in midsummer by ~10 kg C ha\(^{-1}\) day\(^{-1}\).

Analyses of \(^{14}\)CO\(_2\) in the pore space of the moss layer provided further evidence of the decomposition of deep carbon (7). The \(^{14}\)C of moss-layer CO\(_2\) in late fall and winter, after correction for atmospheric dilution, was ~80 to ~150‰, corresponding to a bulk source age of 600 to 1300 \(^{14}\)C years. This depletion of \(^{14}\)CO\(_2\) indicates that autotrophic respiration was minor and that deep decomposition was the main source of the winter efflux observed by eddy covariance (2 to 3 kg C ha\(^{-1}\) day\(^{-1}\)) and the similar efflux observed from the snowpack with chambers (7).

The sensitivity of \(R_{\text{deep}}\) to \(T_{50}\) (Fig. 3B) was much greater than is considered typical for soil respiration (21), which suggests that deep decomposition was controlled by a mechanism other than the direct effect of temperature on enzyme activity. Laboratory and field studies have shown that anoxia due to flooding controls decomposition of organic matter in tundra soils (3, 4), and the apparent high sensitivity of \(R_{\text{deep}}\) to \(T_{50}\) could have resulted from changes in water-table depth that correlated with \(T_{50}\). However, soil at the site did not dry out in midsummer (22), and \(R_{\text{deep}}\) did not vary with rainfall (Fig. 3B).

An alternative explanation is that \(T_{50}\) was related to the depth of the active layer and that the midsummer increase in respiration was a result of the increasing volume of unfrozen soil. The CO\(_2\) efflux from deep soil incubated in the laboratory increased from 0.016 kg C per ton C day\(^{-1}\) at ~2°C to 0.15 kg C per ton C day\(^{-1}\) at 5°C (23). These observations correspond to an increase at the site from 1.5 to 3 kg C ha\(^{-1}\) day\(^{-1}\) in completely frozen soils to 15 to 30 kg C ha\(^{-1}\) day\(^{-1}\) in thawed soils. These rates are consistent with the field observations (Figs. 1 to 3) and provide evidence that soil thaw directly controlled the seasonal cycle of deep decomposition. In contrast, chamber measurements in nearby wetlands do not show a seasonal increase in deep decomposition (\(R_{\text{deep}}\)), apparently because anoxia at those sites limits decomposition even after thaw (24).

Fig. 1. Studies of CO\(_2\) efflux (A) and respiration (B) at the Harvard Forest (open squares) and at the nearby boreal site (open circles), the latter indicates the forest is relatively old, and stands in the area that are younger or dominated by deciduous vegetation may be accumulating carbon at high rates. The annual gross production remained consistent from year to year at ~8.0 tons C ha\(^{-1}\) year\(^{-1}\), whereas annual respiration ranged from 8.9 tons C ha\(^{-1}\) year\(^{-1}\) in 1995 to 7.9 tons C ha\(^{-1}\) year\(^{-1}\) in 1997 (18). The carbon balance was the small residual of two much larger fluxes, a pattern that may make it extremely sensitive to climate.

Fig. 2. Nocturnal CO\(_2\) efflux (whole forest respiration) on windy nights with air temperatures of 8° to 12°C. Points are means for individual nights with at least four valid 30-min observations (5, 14). Both respiration and photosynthesis were reduced during periods with freezing nights (5), typically before mid-June and after early September, apparently because of changes in leaf activity.

Fig. 3. (A) Nocturnal respiration at a sphagnum moss surface as a function of moss temperature (\(T_{\text{mass}}\)) before 18 July 1996 (open circles), and after 19 July 1996 (filled triangles). The efflux of CO\(_2\), which reflects the combined rate of moss, root, and microbial metabolism, was measured every 3 hours with an automated chamber (5). (B) Seasonal increase in respiration (\(R_{\text{deep}}\) or open circles) [see (19)], soil temperature 50 cm beneath a sphagnum surface (\(T_{50}\)) (solid line), and rainfall (millimeters divided by 4) (vertical lines). The least squares fit in (B) was \(R_{\text{deep}}\) (kg C ha\(^{-1}\) day\(^{-1}\)) = 0.76 + 1.09 \(T_{50}\) (°C) \(n = 207, r^2 = 0.69\). The slopes were 1.7 and 1.3 at the other sphagnum chambers and 1.5 and ~0.4 at the feather moss chambers.
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The chamber measurements integrated from May to October 1994 show a significant CO2 flux at the moss sites and no net exchange at the sphagnum sites. This pattern apparently reflects a large difference in respiration and litter input.

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The soil at a depth of 50 to 100 cm thawed for 100 to 50 days each year (Fig. 1C), implying the decomposition of 1.0 ton CO2 ha−1 year−1 for the northern study area Old Black Spruce site (W. C. Oechel, Science 271, 1576 (1996)). We calculated the daily and annual carbon balances as described in (14) but with a 7-day averaging interval and made the same correction for the second largest on record and implies that solar radiation was 8% above average (12). The annual mean temperatures during the study were close to the 1969–1996 average. See K. J. van Cleve, C. W. Oechel, M. E. McArthur, and J. T. McDaniel, New York, ed. 2, 1975). 22. P. Ciais and P. Tans, in press. M. L. Goulden, J. W. Munger, S.-M. Fan, B. C. Daube, S. C. Wofsy, Global Change Biol. 2, 169 (1996).
Evidence of Shallow Fault Zone Strengthening After the 1992 M7.5 Landers, California, Earthquake

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Repeated seismic surveys of the Landers, California, fault zone that ruptured in the magnitude (M) 7.5 earthquake of 1992 reveal an increase in seismic velocity with time. P, S, and fault zone trapped waves were excited by near-surface explosions in two locations in 1994 and 1996, and were recorded on two linear, three-component seismic arrays deployed across the Johnson Valley fault trace. The travel times of P and S waves for identical shot-receiver pairs decreased by 0.5 to 1.5 percent from 1994 to 1996, with the larger changes at stations located within the fault zone. These observations indicate that the shallow Johnson Valley fault is strengthening after the main shock, most likely because of closure of cracks that were opened by the 1992 earthquake. The increase in velocity is consistent with the prevalence of dry over wet cracks and with a reduction in the apparent crack density near the fault zone by approximately 1.0 percent from 1994 to 1996.

A fault plane undergoes sudden stresses, shaking, and cracking during an earthquake. Extensive research has been directed toward understanding this phenomenon (1, 2), but many facets remain obscure.

We focus on the rate at which a fault regains its strength following a large earthquake. This rate is needed to understand how fault zones strengthen or “heal” after an earthquake, but so far, only simple laws have been assumed based on laboratory experiments rather than direct observations in the field. In addition, the healing rate may affect the probability of another earthquake in a fault zone. Experimental studies (3) indicate that a longer interval since the previous episode of faulting correlates with higher stress drop in the subsequent rupture. Studies of repeated earthquakes along a fault (4) show trends that are consistent with state- and rate-dependent healing models (5).

We had the particularly favorable situation of probing the evolution of a shallow fault that had recently undergone large displacements. Earlier efforts to identify dilatancy that might be detectable near fault zones failed because of low sensitivity coupled with what we now know to be subtle precursory changes due to dilatancy (2, 6). Repeated surveys near Parkfield (7) is showing small changes in velocity over time, but in the absence of a large earthquake and with uncertainty about the precise location of the velocity change, its significance is hard to assess. A comparison of earthquakes before and after a specific large event showed a small coseismic reduction in wave velocity at stations with unconsolidated sedimentary rocks that were strongly shaken (8), suggesting the temporal change was a shallow effect of shaking rather than a physical change in the bedrock. One study (9) suggests changes in scattering of P waves with a 1-s period around the time of large earthquakes, but the pattern is not yet well established.

We conducted two identical seismic experiments on 2 November 1994 and 6 August 1996 (Fig. 1) to monitor the change of fault zone physical properties after the 1992 M7.5 Landers earthquake. A pair of explosions, or “shots,” in 30-m boreholes along the Johnson Valley fault segment of the Landers fault zone were detonated in each experiment, using 400 to 700 pounds of chemical emulsions for each. A pair of linear three-component seismic arrays recorded the arrivals of seismic waves for each explosion. The arrays were 3-km in length and aligned perpendicular to the fault. The two arrays were separated by 13 km, and the explosions were located between the arrays (Fig. 1). The array along line 1 had 36 stations and the array along line 3 had 21 stations.

Line 1 is centered at the region that experienced the maximum amount of slip—about 3 m—on the Johnson Valley fault during the Landers earthquake (Fig. 1). Slip is smaller near line 3, and also diminished to the north until the surface rupture connected with the Homestead Valley fault. Fault slip at depth is more difficult to determine, but seems to resemble the slip at the surface (10). The recurrence of faulting on the Johnson Valley fault is estimated to exceed 1000 years (11, 12).

The data from line 3, collected in 1996, had P waves visible on all traces near 1 s. The S waves had a longer period and were more prominent on the horizontal components near 2 s, and the fault zone trapped modes appeared from 3 to 8 s (Fig. 2). The trapped waves showed larger amplitudes,

Fig. 1. Map of the study region showing locations of two seismic arrays at line 1 and line 3 and two explosions SP4 and SP5 in the fault zone of the 1992 Landers, California, earthquake. Only the southern half of the Landers rupture lies within this map, and the centroid surface-fault slip profile is shown (inset) to the left of the map (12). JVF, Johnson Valley fault; KF, Kickapoo fault; HVF, Homestead Valley fault; and PMF, Pinto Mountain fault.