

Quantifying Carbon Dioxide and Methane Emissions and Carbon Dynamics from Flooded Boreal Forest Soil

Maren Oelbermann* and Sherry L. Schiff University of Waterloo

The boreal forest is subject to natural and anthropogenic disturbances, but the production of greenhouse gases as a result of flooding for hydroelectric power generation has received little attention. It was hypothesized that flooded soil would result in greater CO₂ and CH₄ emissions and carbon (C) fractionation compared with non-flooded soil. To evaluate this hypothesis, soil C and nitrogen (N) dynamics, CO₂ and CH₄ mean production rates, and ¹³C fractionation in laboratory incubations at 14 and 21°C under non-flooded and flooded conditions and its effect on labile and recalcitrant C sources were determined. A ferro-humic Podzol was collected at three different sites at the Experimental Lakes Area, Canada, with a high (19,834 g C m⁻²), medium (18,066 g C m⁻²), and low (11,060 g C m⁻²) soil organic C (SOC) stock. Soil organic C and total N stocks (g m⁻²) and concentrations (g kg⁻¹) were significantly different (*p* < 0.05) among soil horizons within each of the three sites. Stable isotope analysis showed a significant enrichment in δ¹³C and δ¹⁵N with depth and an enrichment in δ¹³C and δ¹⁵N with decreasing SOC and N concentration. The mean CO₂ and CH₄ production rates were greatest in soil horizons with the highest SOC stock and were significantly higher at 21°C and in flooded treatments. The δ¹³C of the evolved CO₂ (δ¹³C-CO₂) became significantly enriched with time during decomposition, and the greatest degree of fractionation occurred in the organic Litter, Fungal, and Humic forest soil horizons and in soil with a high SOC stock compared with the mineral horizon and soil with a lower SOC stock. The δ¹³C-CO₂ was significantly depleted in flooded treatments compared with non-flooded treatments.

THE boreal forest occupies nearly 32% of the total global forest area. Of this proportion, more than 85% of the carbon (C) is stored in the soil organic matter (SOM) pool (Prentice, 2001). Slower decomposition rates, because of limited microbial activity, poor drainage, and a shorter growing season in the cold and humid boreal region, make these soils a C sink (Krishnan et al., 2006). Until recently, it was believed that hydroelectric energy was a clean source of energy, with little contribution to greenhouse gas (GHG) emissions, compared with fossil fuel burning (Delmas et al., 2005).

Some researchers (Gagnon and Chamberland, 1993; Svensson and Ericson, 1993; Rosa and Schaeffer, 1994) have suggested that the creation of artificial reservoirs for the production of hydroelectric power constitutes an anthropogenic source of GHG. This was confirmed by Kelly et al. (1997), who showed that a wetland became a net C source instead of a sink once it was flooded. Jugnia et al. (2006) evaluated how much additional methane (CH₄) emissions from flooding could be re-oxidized in a hydroelectric reservoir in Quebec, Canada. Blais et al. (2005) suggested that decomposition of organic matter from aboveground vegetation and that stored in the soil may be rapid, depending on local physicochemical conditions, leading to the formation and release of GHGs from the reservoir's surface.

For example, St. Louis et al. (2000) estimated that 1.5 million km² of the global surface area is covered by reservoirs, which are used for hydroelectric power generation or for irrigation. It is expected that GHG production rates will increase globally as a result of further reservoir expansion because currently only 17% of the potential hydroelectric sites have been developed (Pircher, 1993). The sustained production rate due to post-initial disturbance from these reservoirs yields an annual global CO₂ production rate of 10 × 10¹⁴ g CO₂ yr⁻¹ and a global CH₄ production rate of 0.7 × 10¹⁴ g CH₄ yr⁻¹ (St. Louis et al., 2000). Nitrous oxide (N₂O) is one of the major GHG, with a global warming potential 310 times greater than that of CO₂ (IPCC, 2007). However, N₂O emissions from reservoirs from the boreal forest region and their contribution to global warming appear to be minor (Hendzel et al., 2005).

Canada has flooded 73,000 ha for hydroelectric power generation (St. Louis et al., 2000), and this area is expected to be expand-

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*Corresponding author (moelberm@fes.uwaterloo.ca).

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677 S. Segoe Rd., Madison, WI 53711 USA

M. Oelbermann, Dep. of Environment and Resource Studies, Univ. of Waterloo, Waterloo, ON Canada N2L 3G1. S.L. Schiff, Dep. of Earth and Environmental Sciences, Univ. of Waterloo, Waterloo, ON Canada N2L 3G1.

Abbreviations: CHR, charred layer; ELA, Experimental Lakes Area; GHG, greenhouse gases; JP-B, Jack pine-*Betula*; JP-L, Jack pine-*Ledum*; JP-P, Jack pine-*Polytrichum*; OE, organic matter enriched; SOC, soil organic carbon; SOM, soil organic matter.

ed because of increasing power demand. Future hydroelectric sites will be located in Canada's north, including the boreal forest, because southern sites have been developed and are urbanized. Given the large stores of C in boreal forest soil and vegetation biomass, flooding vast areas of this landscape could contribute a substantial amount of GHG to the atmosphere. St. Louis et al. (2000) estimated that global production rates of GHG from reservoirs may contribute up to 7% of current anthropogenic emissions on a CO₂ equivalent basis.

In this study, it was hypothesized that flooded soil would result in greater CO₂ and CH₄ emissions and C fractionation compared with non-flooded soil. The aim of this paper is to provide information on C, N, CO₂, and CH₄ dynamics from flooded and non-flooded soil with different soil organic carbon (SOC) stocks collected in the boreal forest of north-western Ontario, Canada. Soil organic C and N stocks and the relationship between δ¹³C and δ¹⁵N stable isotope ratios to SOC and N concentration within the soil profile were also characterized for each site. Additionally, this study evaluated soil CO₂ and CH₄ production rates and C fractionation during decomposition by assessing the δ¹³C of CO₂ under spring (14°C) and summer (21°C) ambient temperatures using laboratory incubations.

Carbon isotope dynamics during the biological transformation of plant tissues have been well studied, but data on the mechanisms driving changes in the natural abundance of δ¹³C in SOM are limited (Boström et al., 2007). Another objective was to elucidate the ¹³C dynamics (δ¹³C-CO₂) during decomposition of each horizon, from each site, in addition to evaluating the effects of flooding and temperature on labile and recalcitrant, including charred material, C sources. Information presented in this paper helps to evaluate how soil with different C stocks and vegetative communities will respond to flooding and thereby provides further insight into the dynamics of GHG production rates and the C budget as a result of reservoir creation. This will also aid in the planning and management of new and existing hydroelectric reservoirs, especially if a payment for environmental services is to be implemented to promote the conservation of forest resources and to cap C emissions.

Materials and Methods

Study Site

The study took place at the Experimental Lakes Area (ELA) (49°34' N, 93°36' W). Soils of the study site are classified as a ferro-humic Podzol (Canadian System of Soil Clas-

sification, 1998), and vegetation is characteristic of a boreal subclimax forest dominated by jack pine (*Pinus banksiana* Lamb.), black spruce [*Picea mariana* (Mill.) BSP], trembling aspen (*Populus tremuloides* Michx.), and white birch (*Betula papyrifera* Marsh.). In this region of Ontario, forest fires have a 50- to 100-yr cycle, and the last forest fire in the study site occurred in 1980 (Schindler et al., 1996).

This study focused on three different sites: jack pine-*Betula* (JP-B), jack pine-*Ledum* (JP-L), and jack pine-*Polytrichum* (JP-P). These sites were located within 1 km from each other and were approximately 1 ha in size, with contrasting plant communities and soil C stocks (Table 1). At the time of this study, trees at all three sites were approximately 20 yr of age. Tree diameter measured in 1998, at a tree age of approximately 18 yr since the last forest fire in 1980, at 15 cm above ground level varied between the three sites. For example, at the JP-B site, *P. banksiana* had a mean diameter of 7.6 cm, and *B. papyrifera* had a mean diameter of 4.1 cm. Trees at the JP-L site ranged from 5.4 cm for *P. banksiana* to 3.9 cm for *B. papyrifera*, and those at the JP-P site ranged from 5.8 cm for *P. banksiana* to 2.2 cm for *B. papyrifera*.

The JP-B site is dominated by jack pine and white birch, with *Lycopodium* spp. and *Polytrichum* spp. as understory vegetation. The underlying soil profile at JP-B is of mineral origin, consisting of an organic L (0–2 cm) and FH horizon (2–7 cm), followed by a charred (CHR) layer (7–10 cm), which overlay the mineral Ah horizon (10–20 cm). The JP-L site is dominated by jack pine, with Labrador tea (*Ledum groenlandicum* Oeder.) and *Sphagnum* spp. in the understory. The underlying soil profile at JP-L is also of mineral origin, consisting of an organic L (0–4 cm) and FH horizons (4–7 cm), followed by a CHR layer (7–10 cm), which overlay the mineral C horizon (10–20 cm). The JP-P site is similar to the JP-B site, but the JP-P site has 30% of its area covered with extensive bedrock outcrops with no parent material. Some of the outcrops are coated with patches of moss (*Pleurozium schreberi*, *Polytrichum* spp., *Lycopodium obscurum*) and lichen (*Cladina rangiferina* and other *Cladina* spp.). Underneath the moss-lichen patch is a 6- to 10-cm-thick deposit of material intermediate between mineral soil and organic soil, with a SOC concentration ranging from 130 to 180 g kg⁻¹ (Canadian System of Soil Classification, 1998). In this study, results from the JP-P site focus on the soil beneath the moss-lichen patch, herein referred to as organic matter-enriched (OE) mineral soil. This study also incorporates results from the CHR layer, which is part of the soil profile, because forest fires are a natural occurrence in this ecosystem.

Table 1. Biophysical characteristics of three different upland Boreal forest sites at the Experimental Lakes Area in northwestern Ontario, Canada.

Site	Location above		Soil organic C stock	Total soil N stock	Plant communities
	sea level	Mean soil depth			
	m	cm	g C m ⁻²		
Jack Pine-Birch	425	20	19,834 (high)	646	<i>Pinus banksiana</i> - <i>Betula papyrifera</i> - <i>Lycopodium</i> spp.- <i>Polytrichum</i> spp.
Jack Pine-Ledum	430	20	18,066 (medium)	534	<i>Pinus banksiana</i> - <i>Ledum groenlandicum</i> - <i>sphagnum</i> spp.
Jack Pine-Moss	430	20	11,060 (low)	787	<i>Pinus banksiana</i> - <i>Betula papyrifera</i> - <i>Vaccinium</i> sp.- <i>Polytrichum</i> sp., <i>Pleurozium schreberi</i> , <i>Lycopodium obscurum</i> , <i>Cladina rangiferina</i>

In the boreal forest, hydroelectric power generation occurs throughout the year. However, soil in upland areas is flooded only from May to September; for the remainder of the year, water levels are lowered (drawdown), leaving previously flooded upland soil exposed to the atmosphere. Drawdown in hydroelectric reservoirs helps to increase the efficiency of power production during the winter months in northern climates. As such, the incubation temperatures used in this study represent the mean annual temperature in May and September (14°C) and in July (21°C).

Soil Sampling and Elemental Analysis

Each of the three sites was represented by a plot approximately 1 ha in size. Within each plot, three subplots measuring 2 m by 2 m in size were randomly located. Within each 4-m² subplot, three randomly selected soil samples were taken with a 5-cm inner diameter soil corer. The soil samples were bulked according to their vertical stratification. Soils were air-dried and not sieved to preserve the soil's structure; however, coarse roots (>1 cm diameter) and large pieces of wood were removed. Each soil sample was divided into three subsamples, which were used for the analysis of (i) soil bulk density, (ii) SOC and total soil N concentration (g kg⁻¹) and δ¹³C and δ¹⁵N stable isotopes, and (iii) CO₂ and CH₄ mean production rates and δ¹³C-CO₂. The first set of subsamples was oven-dried at 105°C for 48 h to determine soil bulk density. Bulk density was calculated using the inner diameter of the core sampler, horizon depth, and the oven-dry weight of the soil. Bulk density was not adjusted for rock volume (i.e., mineral particles >2 mm diameter) because of low soil rock content in the collected samples.

For the second subsample, 30 g of the air-dried soil was ground to a fine powder (<80 μm) using a ball mill (Model ZM1; Retsch, Haan, Germany) for C and total soil N concentration (g kg⁻¹). To evaluate the C and total soil N concentration in the CHR layer, charred particles were picked out separately, rinsed with deionized water, air-dried, and ground for elemental analysis. Such processing of the charred particles prevented any organic material or mineral soil from adhering to them, yielding more accurate results of its C and total N concentration. No charred particles were observed in the mineral and Litter, Fungal, and Humic (LFH) horizons. The finely ground soil and charred particles were weighed into tin capsules (D1008; Elemental Microanalysis Ltd., Okehampton, UK) and analyzed for SOC and total N concentration (g kg⁻¹) on an Elemental Analyzer (Carlo Erba, Cernusco, Italy). Soil C and N stocks were quantified by multiplying SOC and N concentration (g kg⁻¹) by the mass of soil per hectare, using soil bulk density and horizon depths.

Aboveground leaf samples from the predominant plant species (*P. banksiana*, *B. papyrifera*, *L. groenlandicum*, *Cladina rangiferina* (L.) Nyl., and *Polytrichum* spp.) at each site were collected in July for elemental analysis. Plant material was sampled using a pruning pole, and five samples were randomly taken and bulked from the canopy of each tree. Three trees per species and per plot were sampled. Plant samples were oven-dried at 65°C for 48 h and ground into a fine powder (<80 μm) using

a Retsch (Model ZM1) ball mill and weighed into tin capsules (D1008; Elemental Microanalysis Ltd.) and analyzed for C and N concentration (%) on a Carlo Erba Elemental Analyzer.

Soil Incubation and Greenhouse Gas Mean Production Rates

For the third subsample, 10 g of the L horizon, 20 g of the FH horizon, 25 g of the CHR layer, and 60 g of the Ah or C horizons were placed in 1000-mL jars. Before incubation, soil water content was adjusted to 50% (w/w) and was kept constant during the incubation by adding deionized water as necessary. There were three replications per soil horizon, including the CHR layer, for the non-flooded treatments. In the flooded treatment, using the same weight for each of the soil horizons as in the control treatment, deionized water was added to the soil, resulting in a water level that was maintained at 2 cm above the soil surface throughout the experiment. This was achieved by replacing evaporative losses with deionized water every 3 d. There were three replications per soil horizon, including the CHR layer, for the flooded treatment.

Three blank jars, containing no soil, were used to represent ambient CO₂ and CH₄ concentrations. Each jar (blanks and those containing samples) were closed with a rubber stopper. The rubber stopper contained a glass tube sealed with a septum that was used to extract gas samples from the jar's headspace. A set of jars and their replications, containing L, FH, CHR, and Ah or C horizons (flooded and non-flooded) and blank jars were incubated at 14°C for 96 d, and a second set of jars and their replications, containing L, FH, CHR, and Ah or C horizons (flooded and non-flooded) and blank jars were incubated at 21°C for 96 d. For non-flooded and flooded treatments, the CHR layer was not pre-treated by picking-out and washing charred pieces to obtain the most accurate representation of this section of the soil profile. However, this may have resulted in the inclusion of organic and mineral soil particles within the CHR layer.

Before each weekly sampling, the jars were fully aerated by removing the rubber stopper for 20 min. The jars were then sealed tightly with the rubber stopper, which was secured by electrical tape and left in place for 24 h. The CO₂ and CH₄ concentration in the headspace of the jar was measured on a Shimadzu GC-8A Gas Chromatograph (Shimadzu, Columbia, MD) with a Peak-Simple Interface Data System 202 (SRI Instruments, Torrance, CA) on a weekly basis by extracting 250 μL at each sampling. External CO₂ and CH₄ standards of 500, 1000, and 15,000 μmol mol⁻¹ in helium were used for calibration. The daily CO₂ and CH₄ production rates were determined using the following equation adapted from Hogg et al. (1992):

$$R = (C_s - C_A) VD / (M/t) \quad [1]$$

where R (μg g⁻¹ d⁻¹) is the amount of CO₂ and CH₄ (μg) evolved per gram of (dry) soil per day; C_s is the concentration of CO₂ and CH₄ evolved from the soil (μL L⁻¹); C_A is the concentration of CO₂ and CH₄ (μL L⁻¹) from the blank jars with no soil; V is the volume of effective headspace (liters); D is

the density of CO₂ and CH₄ adjusted for temperature (14 and 21°C), pressure, and humidity (g L⁻¹); M is the dry mass of the soil sample (g); and t is the sampling time interval in days.

Before the headspace of the flooded treatment was sampled, the incubation jars were shaken for 60 min to displace the accumulated CO₂ in the water to the headspace. Because of the large volume of headspace relative to the volume occupied by the water and soil, according to Henry's Law, the CO₂ accumulated in the water would be driven into the jar's headspace, where it would reach equilibrium. To determine the amount of time needed to reach headspace equilibrium, 700 mL of deionized water (pH 7) was supersaturated with 15,000 ppm CO₂ for 25 min. Three incubation jars were filled with 300 mL of the supersaturated water and tightly sealed. The jars were left to equilibrate for 1.5 h before their CO₂ concentration was sampled, and a second sampling took place after 24 h. At this time, the jars were placed on a platform shaker (120 rpm), and the CO₂ concentration in the headspace was sampled at 15-min intervals for a total duration of 135 min. It was determined that after 60 min of shaking and intermittent headspace sampling, the CO₂ evolution from the water to the headspace reached equilibrium. Equilibrium is defined as no further change in the amount of CO₂ evolving into the headspace from the supersaturated water. It is speculated that shaking the flooded jars may be similar to the actual conditions during flooding and drawdown in the field, which has resulted in the translocation of organic matter from the LFH soil horizons to that of lower mineral soil horizons (unpublished data).

Stable Isotope Analysis

The finely ground soil, charred particles, and plant samples were weighed into tin capsules (D1008; Elemental Micro-analysis Ltd.) and analyzed for stable C and N isotope ratios using a Carlo Erba Elemental Analyzer interfaced with an IsoChrom Isotope Ratio Mass Spectrometer (Micromass, Manchester, UK). The stable C isotope ratio was expressed as δ¹³C ‰ units relative to the Pee Dee Belemnite standard, and the stable N isotope ratio was expressed as δ¹⁵N ‰ relative to air.

At the same time as the CO₂ and CH₄ production rates were sampled, an additional sample from the headspace of each incubation jar was injected and stored in a 12-mL extainer (839W; Labco Ltd., Buckinghamshire, UK). To characterize the change in the δ¹³C of CO₂ (δ¹³C-CO₂) with time, the stored samples were injected into a Micromass IsoChrome Continuous Flow Isotope Mass Spectrometer interfaced with a gas chromatograph at the end of the experiment (HP6890; Hewlett-Packard, Wilmington, DE). As such, samples for δ¹³C-CO₂ analysis were evaluated 24 h after the initiation of the incubation (Day 1) and on Days 11, 25, 35, and 96. The stable isotope ratio was expressed as δ¹³C-CO₂ ‰ units relative to a gas standard with a δ¹³C-CO₂ value of -44.1 ‰. The fractionation (enrichment/depletion) factor (Δδ¹³C), the δ¹³C-CO₂ during decomposition, was quantified by evaluating the difference in stable isotope composition of the soil before its incubation (δ¹³C_{soil}) and the cumulative respired CO₂

between the solid and gas phases (δ¹³C-CO_{2soil-headspace}) and was determined according to the following equation adapted from Wynn et al. (2006):

$$\Delta\delta^{13}\text{C} (\text{‰}) = \delta^{13}\text{C-CO}_{2\text{soil-headspace}} - \delta^{13}\text{C}_{\text{soil}} \quad [2]$$

Statistical Analysis

All data were examined for homogeneity of variance and were found to have normal distributions. Data for soil characteristics (SOC and N stocks [g m⁻²], C and N concentrations [g kg⁻¹], C/N, δ¹³C, and δ¹⁵N) were tested using a one-way ANOVA in SPSS (SPSS Science Inc., 1989). From the ANOVA, significant differences were further tested using Tukey's least significant difference multiple comparison test (Steel et al., 1997). Within each site, differences in the CO₂ and CH₄ mean production rates in flooded and non-flooded treatments and for temperature (14 and 21°C) of each soil horizon were analyzed using a multiple factor ANOVA in SPSS. Within each site, differences in the δ¹³C-CO₂ for flooded and non-flooded treatments and for temperature (14 and 21°C) of each soil horizon were analyzed using a multiple-factor ANOVA in SPSS. Significantly different main effects were further tested using the least significant differences multiple comparison test (Steel et al., 1997). Significant simple effects were tested using the estimated marginal means function in SPSS. For all statistical analyses, the threshold probability level for determining significant differences was *p* < 0.05.

Results

Soil Organic Carbon and Nitrogen

Soil bulk density values at the JP-B and JP-L sites increased with depth and were significantly different among horizons at each site (Table 2). Soil organic C and total N concentrations (g kg⁻¹) at the JP-B and JP-L sites decreased with soil depth (Table 2). Concentration of SOC and total N were significantly different among horizons at JP-B and JP-L, with the exception that the total N concentrations in the L and CHR layers were not significantly different. The organic matter-enriched mineral soil (OE) beneath the moss-lichen patch had a SOC and total N concentration similar to that of the CHR layers at the JP-B and JP-L sites. The soil C/N ratio at the JP-B and JP-L sites decreased with depth and was significantly different between soil horizons. The soil C/N ratio for the OE horizon beneath the moss-lichen patch at the JP-P site was 14.1 (Table 2).

The SOC and total N stocks (g C m⁻²) were significantly different among the soil horizons within the JP-B and JP-L sites (Table 2). The greatest SOC and total N stocks were located in the top 10 cm of the soil, incorporating the organic LFH horizons and the CHR layer at each of these sites. Although the mineral soil horizons Ah and C had the greatest depth (10 cm) as one single layer, its SOC and total N stocks were significantly lower compared with the LFH horizons and the CHR layer at the JP-B and JP-L sites. The OE horizon underneath the moss-lichen patch (JP-P site) was relatively

Table 2. Soil characteristics and stable isotope ratios of soil from three different sites at the Experimental Lakes Area in northwestern Ontario, Canada. Standard errors are shown in parentheses ($n = 3$).

Site	Soil horizon	Horizon depth cm	Bulk density g cm ⁻³	SOC† g kg ⁻¹	Total soil N	C/N ratio	Total soil N stock		$\delta^{13}\text{C}$ ‰	$\delta^{15}\text{N}$ ‰
							SOC stock g C m ⁻²	stock		
JP-B	L	0–2	0.77a‡	534 (11)a	8.1 (1)a	66.0a	4117 (264)a	62 (12)a	-30.47 (0.04)a	-3.65 (0.14)a
	FH	2–7	0.96b	509 (18)b	16.2 (1)b	31.4b	9780 (289)b	311 (34)b	-29.37 (0.18)b	-1.49 (0.29)b
	CHR§	7–10	1.04c	168 (15)c	6.2 (1)a	26.6c	5126 (269)c	193 (23)c	-27.40 (0.17)c	1.16 (0.08)c
	Ah	10–20	1.64d	21 (2)d	1.1 (3)c	10.1d	811 (65)d	80 (8)d	-26.37 (0.03)d	3.24 (0.17)d
JP-L	L	0–4	0.79a	472 (12)a	8.9 (1)a	53.0a	11,186 (453)a	211 (23)a	-29.35 (0.21)a	-1.90 (0.40)a
	FH	4–7	0.89b	175 (13)b	6.6 (1)b	26.5b	3902 (145)b	147 (18)b	-27.68 (0.12)b	0.52 (0.19)b
	CHR§	7–10	1.01c	156 (16)c	7.7 (1)a	20.3c	2371 (134)c	117 (15)c	-26.20 (0.12)c	1.96 (0.01)c
	C	10–20	1.48d	4 (1)d	0.4 (0.1)c	10.3d	607 (43)d	59 (6)d	-26.31 (0.12)d	3.22 (0.05)d
JP-P	OE¶	0–6	1.64	145 (13)	8.0 (1)	14.1	11,060 (432)	787 (45)	-24.49 (0.14)	2.75 (0.13)

† Soil organic carbon.

‡ Values within a column followed by the same lowercase letter are not statistically significantly different at $p < 0.05$ (ANOVA) for JP-B and JP-L when comparing soil horizons.

§ CHR describes the charcoal layer that has accumulated within the soil profile as a result of forest fires.

¶ OE describes organic matter-enriched mineral soil.

shallow (6 cm) but showed a SOC stock comparable to that of the LFH horizon at the other sites. The total soil N stock in the OE horizon (787 g m⁻²) to a 6-cm depth was greater than that of the soil profile to a 20-cm depth at the JP-B (646 g m⁻²) and JP-L (534 g m⁻²) sites. Values of plant C and N concentrations (g kg⁻¹) are presented in Table 3.

Soil Incubation and Greenhouse Gas Mean Production Rates

The mean production rate of CO₂ and CH₄ in the headspace of the jars, expressed as dry mass of soil, over the 96-d incubation was strongly influenced by temperature, flooding, and soil type (organic vs. mineral). Soil with a high SOC and total N stock, such as the LFH horizons, had a greater mean production rate of CO₂ and CH₄ compared with the mineral and OE horizons and the CHR layer. The overall mean CO₂ production rate was 50% greater, and that of CH₄ was 10% greater, at 21°C compared with that at 14°C. Flooding increased the mean production rate by more than 50% for CO₂ and by 60% for CH₄.

When the CO₂ production rate for all soil horizon at the JP-B and JP-L sites was expressed as a mean value for each temperature treatment over the 96-d incubation, the mean production rate was significantly greater at 21°C compared with that at 14°C (Table 4). However, no such difference was observed for the JP-P site. When the CO₂ production rate for all soil horizons within each site was expressed as a mean value for each flooding treatment over the 96-d incubation, the mean production rate was significantly greater in the flooded treatments. The accumulation of CO₂ in the incubation jar headspace over the 96-d incubation period was strongly influenced by the type of soil. For example, the organic LFH horizons showed a greater mean CO₂ production rate compared with the CHR layer and the mineral soil. When expressing the CO₂ production rates as a mean over the 96 d of incubation for each individual soil horizon at the JP-B and JP-L sites, the sequence was L > FH > CHR > Ah ≈ C. Interaction effects of the mean CO₂ production rate with respect to time-by-flooding and time-by-temperature were not significant.

When the CH₄ production rate for all soil horizons at the JP-B site was expressed as a mean value for each temperature treatment over the 96 d of incubation, the mean production rate showed no significant difference (Table 4). However, expressing the CH₄ production rate for all soil horizons as a mean value for each flooding treatment reveals that the mean CH₄ production rate for the flooded treatment was significantly greater. At the JP-L site, when expressing the CH₄ production rate for all soil horizon as a mean value for each temperature treatment over the 96 d of incubation, the mean production rate was significantly greater at 21°C compared with 14°C. When the CH₄ production rate is expressed for all soil horizons as a mean value for each flooding treatment, the mean CH₄ production rate shows no significant difference. The OE horizon at the JP-P site showed no significant difference in the mean CH₄ production rate for flooding and temperature. The accumulation of CH₄ in the incubation jar headspace also was strongly influenced by the type of soil. When expressing the CH₄ production rate as a mean over the 96 d of incubation for each individual soil horizon at the JP-B and JP-L sites, the sequence was L > FH > CHR > Ah ≈ C. Interaction effects of the mean CH₄ production rate with respect to time-by-flooding and time-by-temperature were not significant for all three sites.

Stable Isotope Dynamics

Values of $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ from plant material are typical of a C₃-dominant ecosystem (Table 3). Soil at the JP-B and JP-L sites became significantly enriched in $\delta^{13}\text{C}$ with depth and a

Table 3. Plant leaf C and N concentrations (g kg⁻¹) and stable isotope ratios of common plant species collected in July at the Experimental Lakes Area in northwestern Ontario, Canada. Standard errors are shown in parentheses ($n = 3$).

Plant species	C and N concentrations		$\delta^{13}\text{C}$ ‰	$\delta^{15}\text{N}$ ‰
	C g kg ⁻¹	N g kg ⁻¹		
<i>Pinus banksiana</i>	536 (18)	13.3 (2)	-28.73 (0.03)	-3.96 (0.03)
<i>Betula papyrifera</i>	526 (17)	19.4 (5)	-29.14 (0.04)	-4.84 (0.08)
<i>Cladina rangiferina</i>	469 (19)	5.7 (1)	-24.74 (0.03)	-1.66 (0.03)
<i>Ledum groenlandicum</i>	563 (15)	19.2 (4)	-30.16 (0.01)	-3.28 (0.06)
<i>Polytrichum</i> spp.	495 (19)	10.3 (1)	-27.42 (0.03)	-3.00 (0.22)

Table 4. Production rates of CO₂ and CH₄ (μg g⁻¹ d⁻¹) expressed as a mean value over the 96-d incubation from three sites with high, medium, and low C stocks under non-flooded and flooded conditions at the Experimental Lakes Area, northwestern Ontario, Canada at 21 and 14°C. Standard errors are given in parentheses (n = 3).

Temperature	Horizon†	JP-B‡		JP-L		JP-P	
		Non-flooded§	Flooded§	Non-flooded§	Flooded§	Non-flooded§	Flooded§
CO ₂ (14°C)¶	L	23.7 (2.1)	65.0 (5.4)	16.5 (2.1)	18.5 (2.4)		
	FH	16.5 (4.5)	37.8 (2.3)	4.0 (1.2)	7.8 (1.4)		
	CHR	3.5 (0.7)	9.1 (0.9)	2.4 (0.8)	4.7 (1.2)		
	Ah or C or OE	1.2 (0.4)	1.3 (0.5)	0.2 (0.1)	0.1 (0.1)	3.5 (1.1)	6.6 (1.5)
CO ₂ (21°C)¶	L	47.9 (4.6)	75.0 (6.4)	18.3 (3.1)	30.8 (4.3)		
	FH	17.9 (1.2)	31.1 (3.2)	7.8 (1.1)	4.2 (0.9)		
	CHR	6.0 (1.0)	9.4 (2.1)	3.4 (0.6)	2.2 (0.9)		
	Ah or C or OE	1.2 (0.4)	1.4 (1.1)	0.8 (0.1)	0.9 (0.1)	6.1 (2.2)	7.9 (1.3)
CH ₄ (14°C)#	L	0.81 (0.01)	1.13 (0.01)	0.48 (0.01)	1.02 (0.01)		
	FH	0.25 (0.01)	0.76 (0.01)	0.25 (0.01)	0.54 (0.01)		
	CHR	0.09 (0.01)	0.21 (0.01)	0.11 (0.01)	0.13 (0.01)		
	Ah or C or OE	0.01 (0.01)	0.04 (0.01)	0.02 (0.01)	0.04 (0.01)	0.05 (0.01)	0.02 (0.01)
CH ₄ (21°C)#	L	0.91 (0.1)	1.27 (0.04)	1.54 (0.06)	2.30 (0.1)		
	FH	0.13 (0.03)	0.86 (0.02)	0.10 (0.02)	0.46 (0.03)		
	CHR	0.08 (0.01)	0.26 (0.01)	0.02 (0.01)	0.32 (0.01)		
	Ah or C or OE	0.01 (0.01)	0.04 (0.01)	0.01 (0.01)	0.05 (0.01)	0.12 (0.01)	0.21 (0.03)

† CO₂ and CH₄ fluxes are significantly different (p < 0.05) between soil horizons.

‡ JP-B, Jack Pine-Birch; JP-L, Jack Pine-Ledum; JP-P, Jack Pine-Polytrichium.

§ CO₂ and CH₄ fluxes are significantly greater (p < 0.05) in flooded treatments.

¶ CO₂ flux is significantly greater (p < 0.05) at 21°C than at 14°C at the JP-B and JP-L sites; there was no significant difference (p < 0.05) at the JP-P site.

CH₄ flux is not significantly different (p < 0.05) between 21°C and 14°C at all three sites.

lower SOC concentration (Table 2). The OE soil horizon at the JP-P site was the most enriched in δ¹³C compared with the other sites. The same pattern was observed for the δ¹⁵N values, showing enrichment with depth and a lower soil N concentration at the JP-B and JP-L sites. The OE horizon beneath the moss-lichen patch had a δ¹⁵N value intermediate between that of the other two sites. The isotopic distribution of δ¹³C and δ¹⁵N changed at the JP-B and JP-L sites according to the SOC and N concentration (g kg⁻¹). For example, JP-B and JP-L showed depletion in δ¹³C and δ¹⁵N with increasing SOC and N concentration, respectively (Fig. 1 and 2).

At all three sites, the C fractionation (enrichment/depletion) factor (Δδ¹³C) of the evolved δ¹³C-CO₂ during decomposition was significantly different between each of the five sampling times (Days 1, 11, 25, 35, and 96) at 14 and 21°C in flooded and non-flooded treatments (Table 5). The greatest degree of C fractionation (enrichment) occurred on Days 11 and 35.

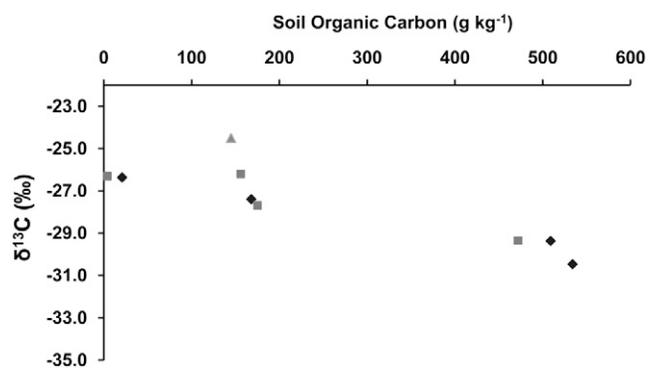


Fig. 1. Isotopic composition of δ¹³C (‰) and its relationship to soil organic C concentration (g kg⁻¹) of each soil horizon at the JP-B (squares), JP-L (diamonds), and JP-P (triangles) sites at the Experimental Lakes Area, northwestern Ontario, Canada.

Carbon fractionation was also significantly different among soil horizons within each site at each of the five sampling times. The greatest degree of C fractionation occurred in soil with the highest SOC stock (LFH horizons), followed by the OE soil, the CHR layer, and the mineral soil. The Δδ¹³C-CO₂ was enriched at 21°C for the JP-B site but depleted at this temperature for the JP-L and JP-P sites. The Δδ¹³C-CO₂ was depleted in the flooded treatments compared with the non-flooded treatments. Interaction effects of the Δδ¹³C-CO₂ with respect to time-by-flooding and flooding-by-temperature were not significant.

Discussion

Soil Carbon and Nitrogen

Plant C and N concentrations (g kg⁻¹) at all sites were similar to those observed by Preston et al. (2006a). Preston et al. (2006a) reported a C concentration of 540 to 573 g kg⁻¹ and an N concentration of 3.4 to 7.7 g kg⁻¹ for needles of *P. banksiana* in a Canadian boreal forest transect study spanning from Saskatchewan to Manitoba, Canada. Soil C and N

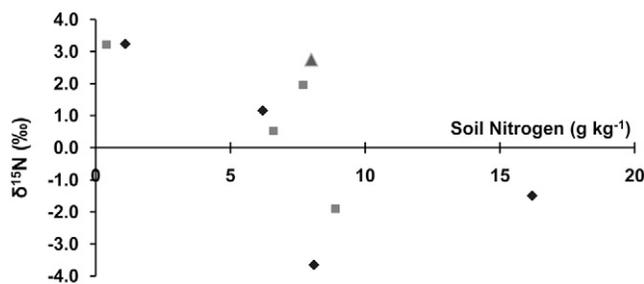


Fig. 2. Isotopic composition of δ¹⁵N (‰) and its relationship to soil organic C concentration (g kg⁻¹) of each soil horizon at the JP-B (squares), JP-L (diamonds), and JP-P (triangles) sites at the Experimental Lakes Area, northwestern Ontario, Canada.

Table 5. Carbon fractionation [$\Delta\delta^{13}\text{C}$ (‰)] of non-flooded soil and flooded soil during decomposition at 14 and 21°C at three sites with high, medium, and low soil organic carbon content at the Experimental Lakes Area, northwestern Ontario, Canada.

Site†	Soil horizon	Temperature °C	$\Delta\delta^{13}\text{C}$				
			Day 1	Day 11	Day 25	Day 35	Day 96
			‰				
Non-flooded soil							
JP-B	L	14	4.92A,a‡§	4.08B,a	3.74C,a	4.40D,a	4.10E,a
	FH		3.67A,b	3.27B,b	3.26B,b	3.83C,b	3.25E,b
	CHR		1.09A,c	1.37B,c	1.50C,c	1.62C,c	1.41E,c
	Ah		0.18A,d	-0.16B,d	-0.54C,d	0.06C,d	-0.08E,d
JP-B	L	21	3.45A,a	3.42A,a	3.00B,a	4.07C,a	3.85D,a
	FH		2.44A,b	2.84B,b	2.48C,b	3.53D,b	3.84B,b
	CHR		0.60A,c	0.77B,c	0.45C,b	0.98D,c	0.89E,c
	Ah		-0.03A,d	-0.24B,d	-0.83C,c	-1.08D,d	-1.16E,d
JP-L	L	14	1.65A,a	1.09B,a	0.45C,a	1.20D,a	0.37E,a
	FH		-0.09A,b	-0.64B,b	-1.02C,b	-0.55D,b	-0.55D,b
	CHR		-1.63A,c	-2.85B,c	-2.61C,c	-1.99D,c	-2.27E,c
	C		-2.23A,d	-2.52B,d	-1.79C,d	-1.81D,d	-2.81E,d
JP-L	L	21	2.07A,a	2.05B,a	2.36C,a	2.71D,a	2.20E,a
	FH		0.43A,b	0.65B,b	0.06C,b	0.92D,b	0.89D,b
	CHR		-1.74A,c	-0.62B,c	-1.37C,c	-1.17D,c	-0.36E,c
	C		-0.34A,d	-0.93B,d	-0.54C,d	-0.11C,d	-1.09D,d
JP-P	OE	14	-2.38A	-3.60B	-2.91C	-2.34D	-1.53E
JP-P	OE	21	-5.25A	-1.79B	-1.98C	-2.05D	-1.87E
Flooded soil							
JP-B	L	14	3.32A,a	4.66B,a	4.43C,a	4.93D,a	4.54E,a
	FH		3.11A,b	3.45B,b	3.59C,b	3.66D,b	3.60C,b
	CHR		1.36A,c	1.18B,c	1.17B,c	1.21B,c	0.96C,c
	Ah		0.24A,d	-0.28B,d	0.75C,d	0.66D,d	-0.06E,d
JP-B	L	21	2.96A,a	4.12B,a	4.16C,a	4.44D,a	3.64E,a
	FH		1.84A,b	2.15B,b	3.76C,b	4.10D,b	3.50E,b
	CHR		-1.16A,c	0.77B,c	0.48C,c	0.63D,c	0.18E,c
	Ah		-0.29A,d	0.35B,d	-0.49C,d	-0.83D,d	-0.88E,d
JP-L	L	14	1.03A,a	0.29B,a	0.23C,a	1.28D,a	-0.18E,a
	FH		-0.59A,b	-1.35B,b	-1.29C,b	-0.95D,b	-1.87E,b
	CHR		-2.18A,c	-2.81B,c	-2.85B,c	-1.85C,c	-3.55D,c
	C		-3.39A,d	-2.57B,d	-2.06C,d	-1.66D,d	-3.62E,d
JP-L	L	21	1.54A,a	1.96B,a	2.16C,a	2.16D,a	2.31E,a
	FH		-0.12A,b	0.36B,b	0.19C,b	0.97D,b	1.41E,b
	CHR		2.33A,c	-0.89B,c	-0.68B,c	-1.10C,c	-1.96D,c
	C		-0.29A,d	-2.03B,d	-1.61B,d	-1.60D,d	-0.60E,d
JP-P	OE	14	-1.80A	-3.01B	-1.99C	-1.72D	-0.96E
JP-P	OE	21	-5.12A	-1.91B	-0.81D	-0.36D	0.49E

† JP-B, Jack Pine-Birch; JP-L, Jack Pine-Ledum; JP-P, Jack Pine-Polytrichium.

‡ Values followed by the same lowercase letter are not statistically significantly different at $p < 0.05$ when comparing between soil horizons and within each site for JP-B and JP-L.

§ Values followed by the same uppercase letter are not statistically significantly different at $p < 0.05$ when comparing between days and within each soil horizon for each site.

concentration (g kg^{-1}), C/N ratio, and bulk density values at all sites were also similar to those reported in other Canadian studies (Hazlett et al., 2005; Jerabkova et al., 2006; Preston et al., 2006a; Hazlett et al., 2007) and were typical for soil of the boreal forest region.

At the JP-B and JP-L sites, the accumulation of organic material for the combined depth of the L and FH horizons led to the formation of a forest floor with a greater SOC and N stock, compared with the CHR layer and the mineral (Ah or C) soil horizons (Table 2). For example, at the JP-B site, the SOC stock for the combined depth of the L and FH horizons was $13,879 \text{ g C m}^{-2}$, compared with 5126 g C m^{-2} in the CHR layer and 811 g C m^{-2} in the Ah horizon. At the JP-L site, the SOC stock for the combined depth of the

L and FH horizons was $15,088 \text{ g C m}^{-2}$, compared with 2371 g C m^{-2} in the CHR layer and 607 g C m^{-2} in the C horizon. These results are similar to those determined by Liu et al. (2002), who found that the combined soil litter and mineral layers, not including a charred layer, had a SOC stock of $15,600 \text{ g C m}^{-2}$ in Ontario's Boreal forest ecosystem.

Wildfire is a major disturbance factor in boreal forests; it plays a key role in rejuvenating soil properties and encourages new tree seedling generation and growth (Preston and Schmidt, 2006). Czimczik et al. (2003) suggested that forest fires play a major role in the formation of burned layers in the forest soil profile. They noted that forest fires resulted in the formation of 0.005 to 0.013 Pg yr^{-1} of charred material in northern hemisphere forests, of which 80% remained in the

soil and ultimately entered the SOC pool, while the remaining 20% was emitted as aerosols. The concentration of C in the charred material observed by Czimczik et al. (2003) was 311 g kg⁻¹, which is slightly higher compared with the results observed in this study (Table 2). These differences may be due to the ash concentration of the material in addition to some mixing with the mineral material in this study. Results from this study showed that the CHR layer contributed 13 to 26% of the total soil C pool to a 20-cm depth in the forests of the ELA. These results were similar to those found by Schmidt et al. (1999) and Glaser et al. (2001), who determined that the amount of black C in the mineral soil contributed 30 to 60% of the total SOC pool. Results from Zachrisson et al. (1998) showed a significantly smaller charred particle stock (98–207 g m⁻²) compared with that of this study, which may be related to a low conversion rate of organic C to charred material in addition to surface erosion and the reallocation of charred particles to much greater depths within the soil profile.

Carbon Dioxide and Methane Mean Production Rate

The cumulative amount of CO₂ evolved at 21°C was higher from soil at all three sites because of a greater rate of C mineralization at this temperature. The higher temperature likely created a more favorable biophysical environment and thereby increased soil microbial activity, which led to a greater amount of available nutrients in the soil solution, resulting in a higher rate of decomposition and microbial respiration. Smith et al. (2003) noted that under laboratory and field conditions, the release of CO₂ from SOM as a result of heterotrophic respiration increased exponentially with temperature. Williams et al. (2006) used a model to evaluate soil response under increasing temperature and found that the rate of CO₂ respiration was over two times greater at 21°C compared with 14°C. The effect of temperature, however, on the mean CH₄ production rate in this study was limited and was similar to results observed by Savage et al. (1997), Smith et al. (2003), and Thérien and Morrison (2005). For example, Yang and Chang (1998) determined that at temperatures of 4°C and 15°C, there was little variation in CH₄ production. However, Yang and Chang (1998) and Cheng et al. (2000) noted that at a temperature greater than 25°C, the production rate of CH₄ increased, which they attributed to the faster rate of decomposition of organic matter at this temperature. In this study, the maximum incubation temperature was 21°C and therefore was likely not high enough to detect any significant changes in the CH₄ mean production rate.

Soil moisture was likely the main factor that led to the greater mean CO₂ production rate observed in our study. Savage et al. (1997) noted that the CO₂ production rate increased under extreme wet conditions, whereas the CO₂ production rate was not greatly affected when the soil was moist. Rey et al. (2002) observed a linear increase in soil respiration with increasing soil water content, and Kirk (2004) found that the rate of organic matter decomposition was greater than expected in a soil that was continuously flooded for 4 yr. He suggested that, in the field, other factors (e.g., soil pH, soil nutrient levels, and micro-

bial communities) may influence the rate of decomposition in submerged soils apart from temperature, aeration, and the quality and quantity of organic matter in the soil (Kirk, 2004). In the field, the CO₂ production rate is also influenced by the addition of organic matter from aboveground vegetation that has died as a result of inundation. This leads to a greater availability of organic matter and nutrients for decomposition compared with a non-flooded forest. For example, a study by Kelly et al. (1997) showed that a wetland became a C source instead of a sink once it was flooded. Before flooding the wetland, it sequestered 6.6 g C m⁻² yr⁻¹; after flooding, it became a C source, releasing 120 g CO₂ m⁻² yr⁻¹ and 9 g CH₄ m⁻² yr⁻¹. Kelly et al. (1997) noted that the majority of the organic C decomposed within 3 yr of flooding, but after 7 yr of flooding, CO₂ and CH₄ production rates were recorded to be the highest over the course of the experiment.

In this study, the greater mean CH₄ production rate in the flooded treatment was likely due to water-logging, leading to temporary anoxic conditions in the incubation jar. The mean CH₄ production rate observed in this study may also have been affected by soil disturbance, such as mixing, which occurs as a result of flooding. Ambus et al. (2001) noted that the disturbance of upland forest soil via mixing resulted in CH₄ emission compared with soil left intact and undisturbed. Therefore, the flooding of forest soil as a result of reservoir creation may modify biochemical parameters, thereby influencing CH₄ and CO₂ dynamics for some period of time in this newly created environment.

The SOC stock in the soil horizons also affected the mean production rate of CO₂ and CH₄, showing a greater production of these GHGs in the L and FH horizons compared with the CHR layer and the mineral horizons. This was likely due to the greater availability, quality, and lability of C compounds in the litter layer. Van der Nat and Middelburg (1998) noted that plant litter removal, where the majority of labile C is located, decreased the CH₄ production rate by a factor of two, and Ambus et al. (2001) observed that the application of materials rich in organic matter fostered CH₄ effluxes in a Danish upland soil. Briones et al. (2007) suggested that the response of decomposition sensitivity to temperature is likely related to SOM age, recalcitrance, and turnover rate. These findings are similar to those of Michaelson and Ping (2003), who showed that organic soil horizons from northern Alaska were 1.7 times more sensitive to an increase in temperature than mineral horizons. Michaelson and Ping (2003) also observed that soil physical characteristics influenced soil biological activity.

A debate remains on the sensitivity of different C substrates to decomposition (Knorr et al., 2005). This sensitivity is dictated by the chemical recalcitrance of the material being decomposed, in addition to its interactions with the mineral soil and its accessibility to microorganisms and their enzymes (von Lütow et al., 2006). Results from our study showed that the CHR layer (non-flooded treatment) is more resistant to decomposition, contributing to a more stable C pool, compared with the labile organic material in the L and FH

horizons. However, when flooded, the mean CO₂ and CH₄ production rate in the CHR layer increased. A similar increase in the mean CO₂ and CH₄ production rate from the CHR layer was observed when the incubation temperature increased from 14 to 21°C. Such an increase in the mean production rate of CO₂ and CH₄ from the CHR layer may be due to the composition of the charred material, which is made up of more stable C compounds that are resistant to decomposition and labile C that is decomposed at a similar rate to that of pre-burned material (Harden et al., 2000).

Stable Isotopes

The δ¹³C values of vegetation and SOM at ELA are within a typical range of a C₃-dominated ecosystem (Table 3) and are similar to results reported by Flanagan et al. (1999) and Preston et al. (2006a). The enrichment in the δ¹³C with soil depth found in this study and in other studies (Preston et al., 2006a) was typical of a biologically active soil and could be attributed to a decrease in the rate of SOC and N transformation (Nadelhoffer and Fry, 1988) and the age of SOC (Wynn et al., 2006). The observed enrichment with soil horizon depth at the JP-B and JP-L sites may also be influenced by the δ¹³C value of the organic material that enters the soil. For example, the L and FH horizons at the JP-B and JP-L sites had a δ¹³C values similar to that of the surrounding vegetation, whereas the mineral soil was enriched in δ¹³C because the SOM had undergone chemical transformations as a result of microbial activity. Gleixner (2005) also proposed that the enrichment in δ¹³C with soil depth may be due to a shift in the trophic level, suggesting that C in deeper horizons is derived mostly from microbes. However, Gioacchini et al. (2006) proposed that several mechanisms are responsible for isotopic enrichment with soil depth, including the preferential uptake of isotopically depleted C and N compounds. These compounds originate from litter decomposition, leaching from the soil, or consumption by soil microorganisms.

Soil at JP-B and JP-L was enriched in δ¹⁵N with soil depth. Such enrichment may be due to the δ¹⁵N value of the organic matter that enters the soil, by N₂ fixation, by atmospheric N deposition, or by N loss (leaching) from the soil. Nadelhoffer and Fry (1988) determined that a soil horizon-enriched δ¹⁵N could be correlated to low soil N concentrations, as was observed in this study. The plant and soil δ¹⁵N values measured in this study fell within the same range as that observed by Amundson et al. (2003).

Incubation temperature strongly affected the δ¹³C-CO₂ respired from the soil collected at all three sites. At higher temperatures, the CO₂ was more depleted in δ¹³C, whereas it was more enriched at lower temperatures at the JP-B site, but this trend was reversed at the JP-L and JP-P sites. The incubation temperature likely affected the use of C sources that were more recalcitrant and enriched in δ¹³C by soil microbes. This suggests that the microbes preferentially consumed compounds enriched in δ¹³C at lower temperatures or that they consumed more compounds enriched in δ¹³C due to changes that occurred in the enrichment of organic matter with time.

As such, the pathway of decomposition was likely not identical at 14 and 21°C, which resulted in more fractionation of the same C sources at the lower temperature. Similar results were observed by Biasi et al. (2005), who suggested that temperature, rather than substrate limitation, was responsible for selective substrate use because temperature may increase the solubility of recalcitrant substrates. Andrews et al. (2000) also observed a temperature-dependent fractionation of C in soil heterotroph-respired CO₂ that was associated with a shift in microbial community structure. They noted that samples incubated at 22 and 40°C showed no difference in the δ¹³C of respired CO₂ and were similar to the δ¹³C value of the soil before incubation. However, soil incubated at 4°C was enriched in δ¹³C-CO₂ compared with that of non-incubated soil. Andrews et al. (2000) suggested that there may be a change in the kinetics of C mineralization or that there may be a shift in the C pool that is mineralized at lower temperatures.

Flooding strongly affected the δ¹³C-CO₂ respired from the soil at all three sites, resulting in a depletion of the flooded soil compared with that of the non-flooded soil. Similar results were observed by Bilek et al. (1999), who evaluated the response of different rice cultivars to soil fertilization, plant biomass, and water management. They observed that the δ¹³C-CO₂ of the soil was depleted during flooding but became increasingly enriched with time when flooding ceased. Bilek et al. (1999) suggested that a dramatic change in soil moisture may affect the availability of organic substrates for decomposition, influencing soil microbial dynamics and the fractionation of ¹³C during decomposition.

An evaluation of the δ¹³C-CO₂ evolved into the headspace of the incubation jars, throughout the course of incubation on Days 1, 11, 25, 35, and 96, showed a variation in depletion and enrichment throughout this time (Table 5). This suggests that the most labile C sources were consumed during the first 11 d of incubation, followed by another bout of consumption on Day 35. Similar results were also reported by Crow et al. (2006), Plante and McGill (2002), and Pendall and King (2007), suggesting that the use of C sources changed with time, leading to the decomposition of more recalcitrant sources reflecting a depletion in the δ¹³C. The mineral horizons, the OE soil, and the CHR layer did not show distinct phases of depletion-enrichment-depletion; instead, the respired CO₂ became depleted in δ¹³C with time. This may be due to the more recalcitrant C sources, especially for the charred material, which is the most recalcitrant form of C in SOM (Czimczik et al., 2003), limiting microbial activity and thereby soil respiration. Gleixner (2005) indicated that recalcitrant C sources (e.g., lignin) are depleted in δ¹³C by up to 6‰ compared with cellulose. Preston et al. (2006b) indicated that other recalcitrant materials, such as tannins and lipids derived from cutin and suberin, are depleted in δ¹³C, which would also cause a shift toward more depleted δ¹³C values. Therefore, an increase in these materials of decomposition content could cause an isotopic shift to more depleted δ¹³C values, such as those observed in this study. This change in the δ¹³C-CO₂ with time indicates ongoing processes, including

microbial degradation and microbial synthesis, or differences in the utilization of C source materials (Pendall and King, 2007; Gleixner, 2005). As such, with depth, microbes are consuming more recalcitrant materials because these are the only substrates available.

Conclusions

Results from this study indicated that the greatest mean production rate of CO₂ and CH₄ is derived from soil horizons with the most labile sources of C. Soil organic C in the surface soil was readily decomposable and vulnerable to increased mineralization when flooded and incubated at a higher temperature. Flooding strongly affected the δ¹³C of respired CO₂, resulting in an isotopic depletion. A dramatic change in soil moisture, such as that observed in this study between flooded and non-flooded treatments, likely affected the availability of organic substrates for decomposition, thereby influencing soil microbial dynamics and the fractionation of ¹³C during decomposition. Recently, the potential of GHG emissions from hydroelectric power reservoirs has gained attention because the number of dams and reservoirs is expected to increase with a rising demand for power on a global scale. One of the major challenges is to gain further insight into the short- and long-term effects of flooding on the soil microbial community structure, soil C turnover, and soil C stabilization, including stable forms of C (e.g., charred material). This will aid in future recommendations of the location of hydroelectric reservoirs to help minimize GHG emissions, especially if a payment for environmental services is to be implemented to promote the conservation of forest resources and to limit C emissions.

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