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THE IMPACT OF EXPERIMENTAL RESERVOIR CREATION ON GREENHOUSE GAS FLUXES FROM FORESTED UPLANDS.

by

Elizabeth M. Joyce ©

A thesis submitted to the Faculty of Graduate Studies and Research in partial fulfillment of the requirements for the degree of Master of Science

Department of Biological Sciences

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Dr. David W. Schindler

Robert Grant

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Date: 11 JULY 2001

I would like to dedicate this work to my husband, Stephen Page, and our family of five.

ABSTRACT

The FLooded Uplands Dynamics EXperiment (FLUDEX) examined the effect of reservoir creation on greenhouse gas (GHG) fluxes from three boreal upland subcatchments that varied in the amount of organic carbon stored in their soils and vegetation. Prior to flooding, soils in all subcatchments respired on average (± standard error) 2300±350 mg CO_2-C m^{-2} day^{-1} and consumed via oxidation 0.9±0.2 mg CH_4-C m^{-2} day^{-1}. Following flooding, the subcatchments became large sources of CO_2 and CH_4 due to decomposition of flooded organic carbon. Mean CO_2 fluxes from flooded soils were three times higher (6500±2700 mg C m^{-2} day^{-1}) in all reservoirs than soil fluxes before flooding. Once reservoir sediments became anoxic, they emitted CH_4 at rates of 41±24 mg C m^{-2} day^{-1}. In the first year of flooding, total fluxes of GHGs from reservoirs were not related to the total amount of organic carbon flooded, but to the amount of labile organic carbon readily available for decomposition.

ACKNOWLEDGEMENTS

This work would not have been possible without the overwhelming support of many. I would like to thank my advisors, Vince St. Louis, Dale Vitt, David Schindler, and Robert Grant for guidance, advice and support. I would especially like to thank Vince for the many hours of editing. Sherry Schiff has been a continual source of advice, support, and scientific spunk. To Ken Beaty, I owe many thanks for explanations of all things hydrological. I would like to thank Ray Hesslein for being so brilliant and patient with my endless barrage of questions over the last 2 years. I would like to thank Natalie Boudreau, Jason Venkiteswaran, James Crichton, and Cory Mathews for good conversation and assistance in the field and lab. Finally, a great big thank-you to all the folks at ELA who strive to keep the station top notch.

For, the rest of you, as there are too many to list as above, I would like to use word associations to thank you:

Cheeko- warm, sweet fur
Brad Park- fresh outdoor frogs and pollywogs. Dig em'
Kurt Hangle- now it won't be so easy to push my buttons
Marnie Potter- ease of mind, stomach, and hands
Susan Crites- Momma Bird, glad to be one of your nestlings. Thank you
Marc Song- recipe cards make for a good thesis. Ajax the conqueror!
Scott Higgins- Arr be darr, Scart! Your confidence warms the cockles of my heart
Candace Fisher- glad I didn't have to get pregnant to finish this! Very glad you did.
James Crichton- you're one entertaining circus freak. I'm forever changed!
Stephen Hanus- hey is that a nose-hair caught in your willow furniture?
Joy Kennedy- you sing like an angel and break wind like the devil. The ultimate woman
Selene Winchester- fast hands, faster mouth, and a heart the size of Vegas
Neil Fisher- we're going to need to whip up one serious flambé when this is done!
Lyn Marlow- you're my clairvoyant connection to everything indulgent.
Britt Hall- thanks for teaching me to drink like a Manitoban
Roger Mollot- are you downtown Rahj or a cop?

Lisa Dueck- you got funk, you got style!
Andrew Majewski- Andy Betty, you may be dirty, but you're one hell of a nifty nurse
Allyson Jung- you're the most refreshingly familiar of all
Paul Humenchuk- here's hoping to have some time floating down a lake with you.
Suzanne Bayley- thank you for inspiring me to sink my teeth into botany and science
Kory Pennebaker- Dallas, you're the salt of Texas, pretty as a flower, and hey, what ya got glowin' under water?
Carolyn Geekie- may your aura blast all the colors of the rainbow!
Katherine Peach- the name says it all… gal with sweet fruit and a stony pit
Karen Scott- hope I don't get you kicked out of your office
Pauline Gerrard- Noy, you inspire me on many levels and are so dear to me

The strength and cohesion of my family helped tremendously through this process. To my mother, I'd like to thank you for giving me colors, and being a friend (and a mother when you needed to be). To Michael, you've always inspired me to reach my potential through your own bright example. I thank Agnes for keeping my feet on the ground. To Pops, I'd like to thank you for taking such a keen interest in all this gas business. To Louise and Roger, thank you for love, support and perspective. Above all else, I would like to thank Stephen Page for making me laugh, wiping my tears, tolerating me, feeding me, marrying me, and supporting me through every step of this process. I would not have been able to do it without you. You put the ding in rammalamma ding dong!

TABLE OF CONTENTS

LIST OF TABLES

LIST OF FIGURES

INTRODUCTION

Greenhouse gases and climate change

Anthropogenic emissions of gases such as carbon dioxide (CO_2), methane (CH_4), carbon monoxide (CO), nitrous oxide (N_2O), and chlorofluorocarbons (CFCs) to the troposphere are of concern because of their global warming potential (GWP) (Lashof and Ahuja, 1990; Prather, 1994). These greenhouse gases (GHGs) absorb infrared radiation emitted from the Earth's surface and re-emit this energy back towards the earth (Houghton *et al.*, 1996). The potential effects of climatic warming are controversial and far-reaching, ranging from reduction of the cryosphere (global snow, ice, and permafrost) to loss of late succession forest species (Houghton *et al.*, 1996). Post-industrial revolution increases in atmospheric concentrations of CO_2 and CH_4, and increases in mean annual temperatures have been well documented (Harvey, 2000). The combustion of fossil fuels, cement production, and changes in tropical land use account for most of the anthropogenic CO_2 emissions to the atmosphere (Houghton *et al.*, 1996). Of the other GHGs, CH_4 is the most important because of its relatively high global emissions and high GWP (24.5 times that of CO_2 over a 100-year period; Houghton *et al.*, 1996). Major human sources of CH_4 emission include animal waste, animal husbandry, rice cultivation, landfills, oil and natural gas, coal mining, and burning of forest biomass (Graedel *et al.*, 1993; Hogan *et al.*, 1991; Mooney *et al.*, 1987).

Extent of reservoir creation

Although hydroelectric reservoirs are still considered by some to be a carbon-free method of generating power (Hoffert *et al.*, 1998; Victor, 1998), Rudd *et al.* (1993) discovered that they were sources of CO_2 and CH_4 to the atmosphere. Whether reservoirs are created for hydroelectric generation, water supply, recreation, irrigation, aquaculture, or flood control (ICOLD, 1998), all reservoirs studied to date have been found to emit CO_2 and CH_4 to the atmosphere (St. Louis *et al.*, 2000). Mean fluxes from the surfaces of temperate reservoirs range from 220-3450 mg CO_2 m^{-2} day^{-1} and 3-79 mg CH_4 m^{-2} day^{-1} (St. Louis *et al.*, 2000). Fluxes from tropical reservoirs are much higher ranging from 2900-4460 mg CO_2 m^{-2} day^{-1} and 65-1140 mg CH_4 m^{-2} day^{-1} (St. Louis *et al.*, 2000). In Canada alone there are approximately 73,000 km^2 of existing reservoirs in four of 10 provinces, and approximately 1.5 million km^2 the world over (St. Louis *et al.*, 2000). Presently the global GHG flux from reservoirs may account for 7% of the GWP of all anthropogenic emissions and yet they are not included in any national GHG inventories (St. Louis *et al.*, 2000).

Reservoir creation and GHG flux

When landscapes are flooded to create reservoirs, the impact on atmospheric GHGs is two-fold. First, all of the existing vegetation that previously assimilated carbon through photosynthesis dies. Second, all the flooded organic matter in the dead vegetation and soils decomposes to CO_2 and CH_4, which are then released to the atmosphere. As a result of flooding, the landscape is therefore transformed from a carbon sink to a significant carbon source (Campo and Sancholuz, 1998; Duchemin *et al.*, 1995; Galy-Lacaux *et al.*, 1997; Kelly *et al.*, 1997; Kelly *et al.*, 1994; Scott *et al.*, 1999; Rosenberg *et al.*, 1997; Rudd *et al.*, 1993). In the presence of oxygen (O_2), flooded chemoheterotrophic bacteria use O_2 as an electron acceptor in the oxidation of electron-rich organic carbon, capturing energy and producing water (H_2O) and CO_2 (Schlesinger, 1997). Once O_2 has been depleted, bacteria use other electron acceptors such as manganese (Mn^{4+}), nitrate (NO_3^-), iron (Fe^{3+}), sulfate (SO_4^{2-}), and CO_2 in order of most electronegative to least in the process of decomposition (Kelly *et al.*, 1988). In the absence of O_2, methanogenesis involves a partnership between eubacteria that hydrolyze and ferment organic carbon to acetate (CH_3COOH), H_2, and CO_2, and archaeobacteria that convert these end products to CH_4 (Mah, 1982; Schlesinger, 1997; Williams and Crawford, 1985). Archaeobacteria have two metabolic pathways; they either split acetate producing CO_2 and CH_4 or reduce CO_2 to produce H_2O and CH_4 (Cicerone and Oremland, 1988). In reservoirs, the tendency towards anoxia in the stagnant waters encourages the anoxic production of CH_4.

Ecosystem carbon storage

The amount of GHG emitted from a reservoir in the long term after flooding theoretically depends on the quantity of decomposable organic carbon stored in the ecosystem prior to flooding. The amount of stored organic carbon varies between ecosystems with differences in watershed hydrology and disturbance periodicity (e.g., fire, harvest). At one extreme are peatlands, where long-term storage of sequestered atmospheric CO_2 in peat carbon generates an average store of approximately 130 kg C m^{-2} (Gorham, 1991). At the other extreme are upland forests with thin soils storing an average 15 kg C m^{-2} in northern regions and approximately 10 kg C m^{-2} in tropical climates (Schlesinger, 1997). Since the last glaciation, boreal forests have been accumulating soil carbon, particularly in organic peat (Sellers *et al.*, 1997). Overall, carbon sequestration in upland boreal forests is low, amounting to the small difference between photosynthetic gains and losses through plant, root and soil respiration (Sellers *et al.*, 1997). The small remaining difference, often greatest in early succession forests, is defined as net ecosystem production (Schlesinger, 1997). From an atmospheric point of view the difference between the amount of CO_2 removed by photosynthesis and the amount emitted through respiration is termed *net ecosystem exchange*. Over an 80-100 year period, frequent intense fires in the boreal ecoregion may burn upland vegetation and soils back to bedrock, resulting in a zero net ecosystem exchange of CO_2 with the atmosphere. Boreal forest soils consume 0 to 5 mg CH_4 m^{-2} d^{-1}, however, through oxidation by methanotrophic bacteria (Adamsen and King, 1993; Castro, 1994; Crill, 1991; Houghton, 1992; Lessard, 1994; Savage, 1997; Whalen, 1992). In assessing the *net* impact of reservoir creation, one must consider not only the GHG flux from the reservoir surface, but also the loss of CO_2 uptake by photosynthetic plants and CH_4 oxidation in forest soils.

Net effects of reservoir creation: ELARP

Although GHG emissions from existing reservoirs have been measured (St. Louis *et al.*, 2000), the Experimental Lakes Area Reservoir Project (ELARP) was the first experiment designed to determine the *net* effect of reservoir creation (Kelly *et al.*, 1997). Two years of preflood and many years of postflood gas flux measurements were taken to determine the net effect of reservoir creation on GHG emissions. A boreal wetland was chosen for this experiment for two reasons. The first is that boreal wetlands are often the areas initially inundated because of their low-lying position on the landscape. Second, a wetland was chosen as a worse case scenario for potential long-term decomposition of organic carbon and flux of GHG to the atmosphere because they store more organic carbon per m^2 (as peat) than any other ecosystem in the world (Post *et al.*, 1982). Flooding changed the ELARP wetland from a carbon sink (7 g C m^{-2} yr^{-1}) to a large carbon source (130 g C m^{-2} yr^{-1}) in the form of CO_2 and CH_4 emitted to the atmosphere. Measurements of GHG flux from the surface of this reservoir are still being measured to determine long-term trends. Recommendations made following three years of flooding the ELARP wetland were to minimise both the total area of land and amount of organic carbon flooded to potentially reduce GHG flux to the atmosphere.

FLUDEX

The FLooded Uplands Dynamics EXperiment (FLUDEX) was designed to address the hypothesis that reservoirs created on boreal upland forest, with relatively low stores of organic carbon in thin soils, will emit less GHG to the atmosphere in the long term than reservoirs created over peatlands. Three types of upland boreal forest were chosen for inundation along a moisture and carbon storage gradient. The objectives of the GHG portion of this experiment were:

1. To quantify both vegetative and soil organic carbon stores in each forest subcatchment prior to flooding;
2. To assess fluxes of CO_2 and CH_4 from the soils of three upland forests in their natural state (prior to flooding);
3. To assess GHG flux from the three reservoirs after flooding and to gain an understanding of processes associated with this flux;
4. To compare both preflood and postflood GHG flux data to other studies including the ELARP experiment.

MATERIALS AND METHODS

Site description

We experimentally flooded three upland boreal forest subcatchments located at the Experimental Lakes Area in NW Ontario (Figure 1). The three subcatchments differed in hydrology and water storage, and as a result, in vegetation community types and carbon storage in soils. The three subcatchments included a low-lying moist/high carbon forest (high C subcatchment; Figure 2), a dry/intermediate carbon forest (intermediate C subcatchment; Figure 3), and a very dry/low carbon ridgetop forest (low C subcatchment; Figure 4). The high C subcatchment was 0.69 ha in area, and was dominated by two plant communities (Table 1). Approximately 53% of the vegetation community consisted of water-loving plants such as *Sphagnum spp.* and Labrador tea *(Ledum groenlandicum)*, and jack pine (*Pinus banksiana)*, while the remaining vegetation community was primarily a drier jack pine forest with a feather moss (*Polytrichum spp.*) ground cover (Heubert, 1999). This subcatchment had an average soil depth of 41 cm and the highest amount of carbon stored in soils and above ground vegetation (6.6 kg C m^{-2}) of all three subcatchments (Boudreau, 2000). The intermediate C subcatchment was 0.50 ha in area with a homogeneous dense jack pine and birch *(Betula papyrifera)* community (Table 1). There was very little ground vegetation due to a dense canopy. The average soil depth was 39 cm, and approximately 5.8 kg C m^{-2} was stored in soils and above ground vegetation prior to flooding (Boudreau, 2000). The low C subcatchment was 0.66 ha in area and approximately 73% of this subcatchment consisted of a jack pine forest with a blueberry shrub *(Vaccinium spp.)* understory, while the remaining vegetation community consisted of *Polytrichum spp.* and lichens (*Cladina spp.*) on exposed bedrock (Heubert, 1999). Average soil depth was 15 cm and carbon stores in above ground vegetation and soils were approximately 4.3 kg C m^{-2}, the lowest carbon stores of the three subcatchments prior to flooding (Boudreau, 2000).

Experimental flooding and reservoir water budgets

During the preflood summer of 1998, dikes were constructed along the low-lying contours of the three forested subcatchments (Figures 2-4). Dikes were constructed

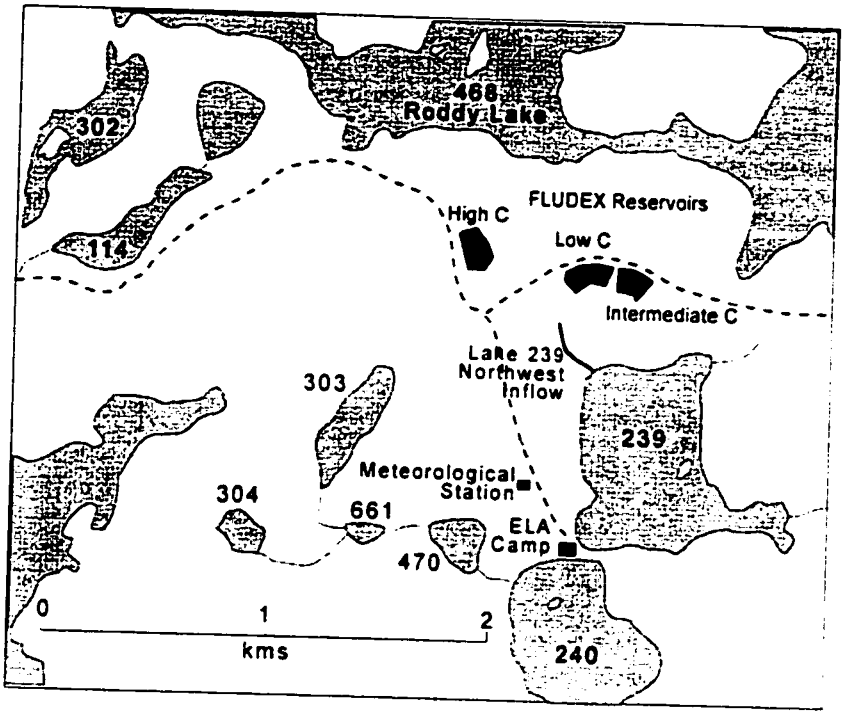
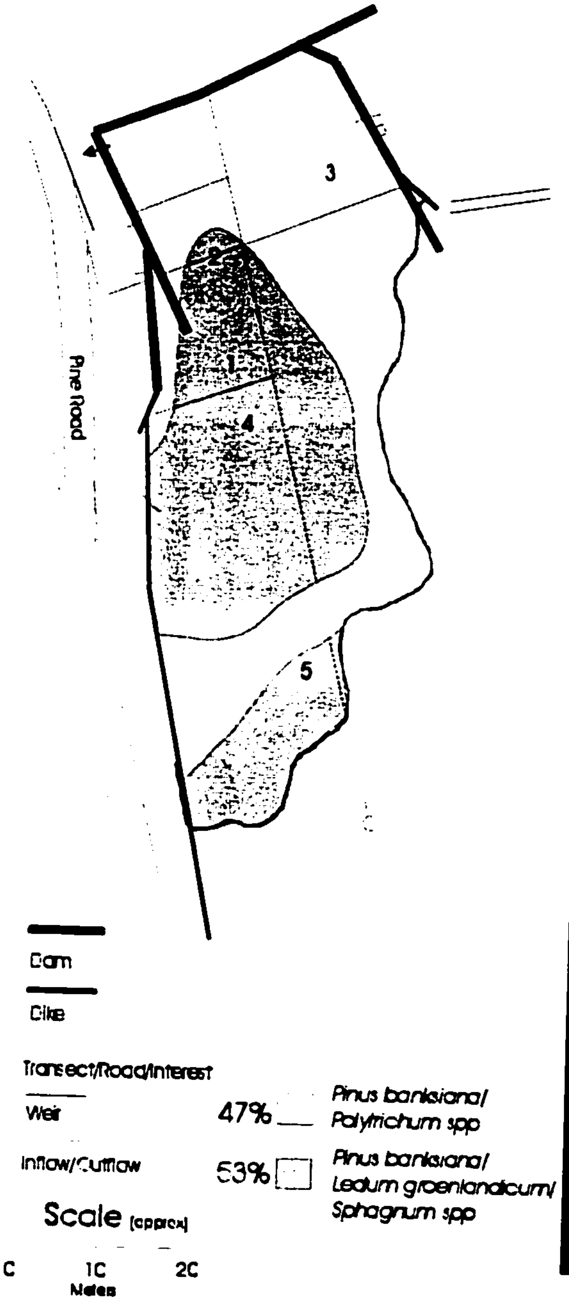


Figure 1. Map of the Experimental Lakes Area, showing the FLUDEX reservoirs, Lake 468, Lake 239 NWIF and a photograph of dike construction.

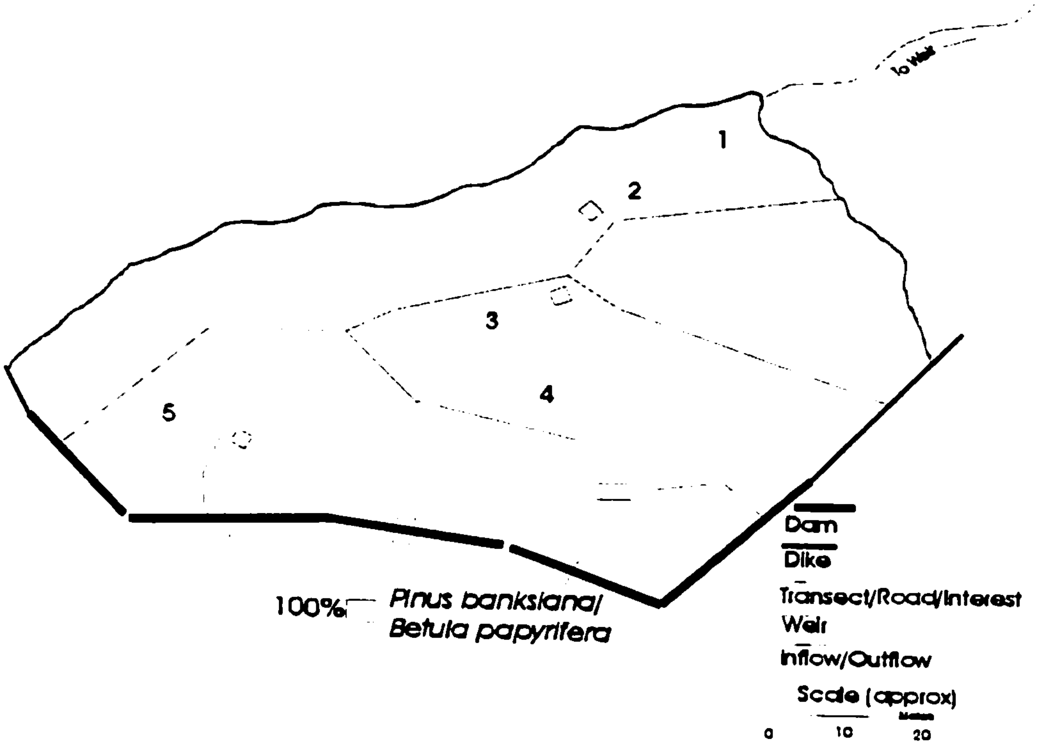


***Pinus/Ledum/Sphagnum* community**

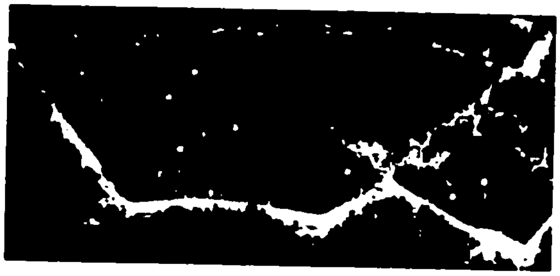
Aerial view preflood

Aerial view postflood

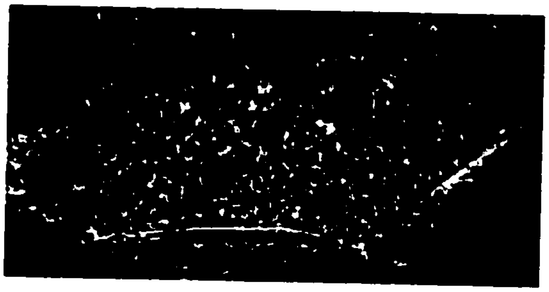
Figure 2. Diagram of the high C subcatchment showing sampling sites 1-5, cut transects, and the two vegetative communities.



Preflood *Pinus/Betula* community

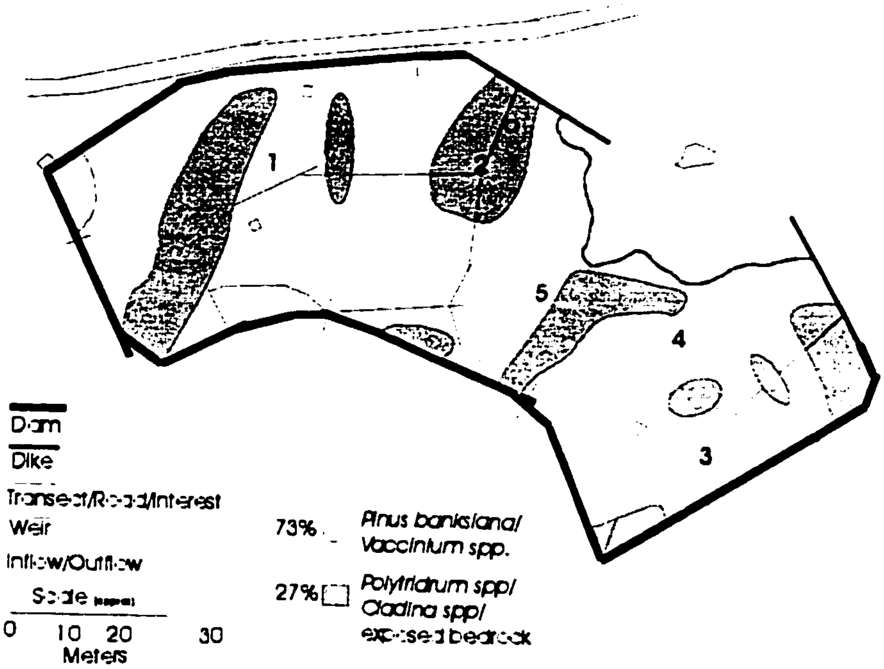


Aerial view preflood



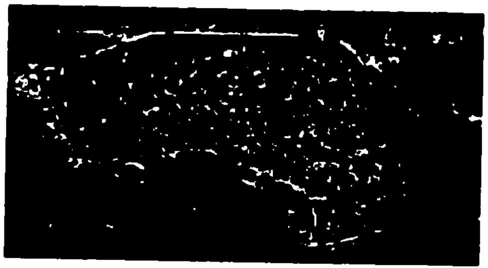
Aerial view postflood

Figure 3. Diagram of the intermediate C subcatchment showing sampling sites 1-5, cut transects and vegetative community.



Preflood vegetative community structure

Aerial view preflood



Aerial view postflood

Figure 4. Diagram of the low C subcatchment showing sampling sites 1-5, cut transects and the two vegetative communities.

Table 1. Site descriptions and carbon stores in the three upland forest subcatchments flooded to create the experimental reservoirs. Vegetation data from Huebert (1999). Soil data from Boudreau (2000).

Subcatchment	High Carbon		Intermediate Carbon	Low Carbon	
Size (ha)	0.69		0.50	0.66	
Average soil depth (cm)	41.1		39.3	15.2	
Dominant vegetation	*Pinus/Ledum/ Sphagnum*	*Pinus/ Polytrichum*	*Pinus/Betula*	*Pinus/ Vaccinium*	*Polytrichum/ Cladina*
Above ground vegetation (Kg C m^{-2})	1.99±0.90	3.63±1.17	2.77±2.85	2.70±1.83	0.01±0.01
Soils: (Kg C m^{-2})					
Litter and FH range (mean)	1.07-5.72 (3.00)	1.24-5.84 (3.20)	1.16-4.41 (2.27)	1.00-2.39 (1.53)	1.44-5.67 (2.53)
% C composition	37		46	39	
Mineral soil range (mean)	0.39-1.87 (0.73)	0.43-1.20 (0.79)	0.49-1.24 (0.77)	0.18-0.96 (0.52)	0.26-1.06 (0.53)
% C composition	1		3	4	
Total soil (mean)	3.73	3.99	3.04	2.05	3.06
Soil + vegetative carbon (Kg C m^{-2})	5.72	7.62	5.81	4.75	3.07
Subcatchment total (Kg C m^{-2})	**6.61**		**5.81**	**4.30**	

such that the maximum flooding depth was two meters. Plastic-lined wooden dikes were attached directly to the bedrock where flooding depth was to be deeper than 1 meter. Soil cleared to expose the bedrock was backfilled both inside and outside the dikes. Plastic lined gravel dikes were constructed where flooding was less than 1 meter. No dikes were built when natural contours of the subcatchments were greater than the maximum height of water. Beginning on 22 June 1999, water was continuously pumped from the nearby oligotrophic Roddy Lake (L468, Figure 1), via a diesel pump situated on its shore, through irrigation pipes to fill the three reservoirs. Each reservoir had three inlets located above the maximum water level so that water would be oxygenated as it entered the reservoirs (to maintain fish populations). Water renewal time was maintained between 8-11 days for all reservoirs. The reservoirs were drawn down on 6 October 1999 and took approximately 4 days to empty.

Measurements of CO_2 and CH_4 fluxes from forest soils preflood

The true effect of reservoir creation on the atmosphere is the *net* difference between fluxes of GHGs prior to flooding and those after flooding. We were not able to measure net ecosystem exchange of CO_2 (gross photosynthesis minus total ecosystem respiration) prior to flooding. Instead we measured rates of net dark respiration by forest soils and ground vegetation using static chambers. We also measured rates of CH_4 oxidation by soils simultaneously with CO_2 fluxes. In May of the 1998 preflood year, two intersecting transects were cut through each subcatchment and five sampling sites were chosen randomly along these transects. Three permanent grooved chamber collars were dug into the soils of each sampling site, to give a total of 15 chambers per subcatchment. The high C subcatchment had four sampling sites (sites 1, 2, 4, and 5) located in the moist *Pinus/Ledum/Sphagnum* community and one site (site 3) in the *Pinus/Polytrichum* community (Figure 2). The intermediate C subcatchment had 5 sites located in the homogeneous *Pinus/Betula* community (Figure 3). The low C subcatchment had three sites (sites 1, 3, and 4) located in the *Pinus/Vaccinium*/litter community and two sites (sites 2 and 5) in the open *Polytrichum/Cladina*/bedrock community (Figure 4). Boardwalk platforms were installed at each site to avoid disturbing soils when sampling. 9-L polycarbonate water jugs that had their bottoms removed were used as static chambers. The chambers were covered with foil to omit light and prevent excessive heat build-up. They were gently placed into the water-filled collar grooves to create a sealed chamber. To avoid disturbing the chamber while measuring fluxes, gas samples were extracted through a 20-cm long 2 mm inner diameter tygon tubing attached via a 18-gauge needle to a septa inserted in the side of the chamber. Concentrations were measured initially and every 10 minutes over a 30-minute period. Chambers were sampled in each subcatchment approximately every second week from 17 June to 10 October 1998. Gas flux calculations were corrected for in situ barometric pressure and temperatures monitored inside the chambers.

CO_2 samples were collected in evacuated 60-ml Wheaton bottles and were analysed on a Shimadzu mini-2 gas chromatograph (GC) with a flame ion detector (FID). A haysep D column separated the CO_2 and CH_4, and a ruthenium methanizer reduced the CO_2 to CH_4. Detection limit was 12 ppm CO_2. A gas volume of 0.2 mls was injected into an injection port using a pressure loc syringe. Eight CO_2 standards ranging between 20 and 19900 ppm were used to calibrate the GC and as quality control throughout sample runs. Only calibration curves with an $R^2 > 0.98$ were accepted. An SP4270 Spectraphysics integrator integrated peak areas.

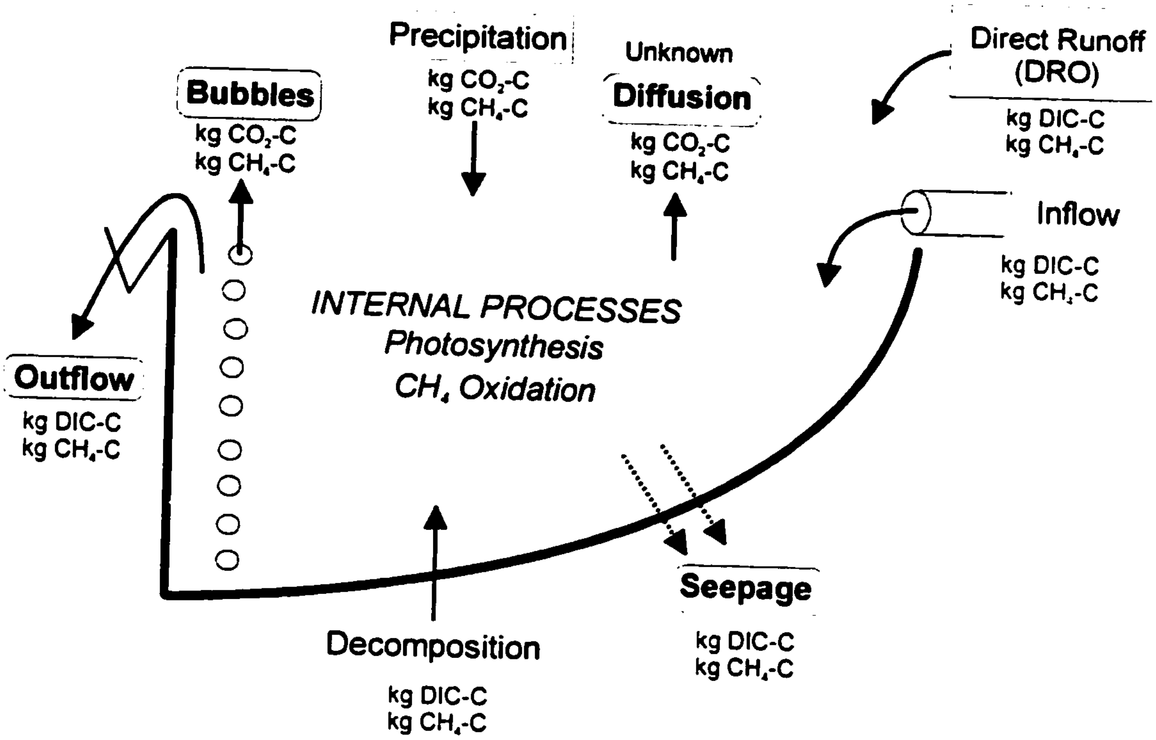
CH_4 samples were collected in 5-ml syringes with 3-way-valves and analysed the same day. Samples were injected into a 3 ml sample loop on a Varian series 3700 GC with a Haysep D column and FID. Detection limit was 0.2 ppm CH_4 (Scott *et al.*, 1999). Three CH_4 standards of 1.6, 4, and 6 ppm were used to calibrate the GC and as quality control throughout sample runs. Only calibration curves with an $R^2 > 0.98$ were accepted. An SP4270 Spectraphysics integrator integrated peak areas.

Gas concentrations (ppm) were corrected for temperature and barometric pressure and flux rates were calculated by linear regression of the four measurements taken over the 30-minute interval. For CO_2, regressions calculated on the first two sets of chamber runs consistently showed an $R^2 > 0.98$ indicating that fluxes were linear, and as a result, only initial and 30 minute final samples were taken in the latter part of the sampling season. Correlations for CH_4 were less consistent and so four measurements were always taken over the 30-minute time period. An $R^2 > 0.80$ was required for a rate to be accepted. Rates (mg GHG m^{-2} day^{-1}) were calculated using the quantity of GHG that accumulated in the chamber over time, the volume of the chamber and collar, and the soil surface area within the collar. A mean of the measured flux rates on a given day was calculated.

Each reservoir contained two sites where copper-constance thermocouple rods attached to a Campbell Scientific datalogger recorded soil temperatures every 30 seconds at 0, -5, -15, and -20 cm during the 30-minute sampling period. Regressions of soil temperature (mean of temperatures collected at 0 and -5 cm) to CO_2 and CH_4 flux were conducted for each reservoir. Ground water level was monitored with wells at the 5 sampling sites per reservoir, except site 2 in the low C subcatchment where only 5 cm of soil was present above the bedrock. Regressions of ground water level with CO_2 and CH_4 flux were conducted for each reservoir. An analysis of variance (ANOVA) was conducted to determine difference in CO_2 and CH_4 flux between subcatchments using within-subcatchment variation. All statistical analyses were conducted using SYSTAT 9, an SPSS program (Wilkinson, 1999).

Measurements of CO_2 and CH_4 emissions after flooding

GHG fluxes from reservoir surfaces are typically measured directly with static chambers, or indirectly with the thin boundary layer calculation (Duchemin *et al.*, 1999; Sellers *et al.*, 1995). However, the use of floating flux chambers was not practical because they are labor intensive when quantifying diurnal patterns of fluxes. The thin boundary layer method was not possible because the abundance of trees protruding from the reservoir surfaces made it impossible to estimate accurate windspeeds for the whole reservoir, and windspeeds lower than 3 m sec^{-1} tend to underestimate flux from the surface (Duchemin *et al.*, 1999). Therefore, a carbon mass balance input-output budget was attempted to determine the GHG flux from the surface of the reservoirs for the whole period that the subcatchments were flooded (Figure 5). Inputs of dissolved inorganic carbon (DIC) and CH_4 to the reservoirs consisted of inflow water pumped from Roddy Lake, runoff from the associated drainage area, atmospheric deposition, and internal microbial production of CH_4 and CO_2 from decomposition of flooded organic carbon. Outputs of DIC and CH_4 from the reservoirs include fluxes from the surface (unknown), the outflow from the weir, the seepage along the reservoir walls and through bedrock fissures, and carbon lost out the drainage pipe when the reservoirs were drawn down. Processes affecting the mass carbon balance were the loss of CO_2 via algal photosynthesis and the loss of CH_4



$$IN_{inflow} + IN_{decomposition} + IN_{precipitation} + IN_{runoff} = OUT_{outflow} + OUT_{seepage} + OUT_{flux}(unknown)$$

Processes: Photosynthesis consumes CO_2 in the water column and methane oxidation consumes CH_4 in the water column. Ebullition contributes to flux from the surface of the reservoir.

Figure 5. Diagram of carbon inputs and outputs in kilograms of DIC-C and CH_4-C and the processes involved in the calculations of a seasonal carbon mass balance budget. The total inputs of carbon should balance the total outputs of carbon.

via microbial methane oxidation in the water column (also unknown). Ebullition of bubbles high in CH_4 and CO_2 from the sediments also contributes to the flux of CH_4 and CO_2 from the surface of the reservoir.

Inputs and outputs of carbon to the reservoirs

Decomposition

Decomposition of flooded soils and ground vegetation was measured with submerged chambers placed on the same collar sites used for measuring GHG flux prior to flooding. Submerged chambers were weighted 9-L polycarbonate bottles fitted with 9-volt battery operated motors and propellers. The propeller rotated 1 rpm to slowly mix the water in the chambers to avoid gradient build up. Clear chambers were used to allow light penetration and account for CO_2 consumption via photosynthesis within the chamber. As a result, measured fluxes were considered conservative estimates of decomposition. One submerged chamber was placed on a collar at each of the five sites in each reservoir for 4-5 hours every other week. In the high and intermediate C reservoirs, the low water levels at sampling sites 5 and 1, respectively, made these sites inaccessible for sampling, and therefore two chambers were placed at one of the other sampling sites. On each sampling date, one chamber was sampled every hour to ensure linear accumulation of gases, while the other four chambers were sampled initially when chambers were lowered onto collars and at the end of incubations. Samples were collected from the chambers through a 2.4 mm inner diameter tygon tube into a 60-ml syringe with a three-way valve. The water was then transferred from the syringe into a 30 ml evacuated Wheaton bottles with a 5-ml headspace of UHP nitrogen. After collection, 0.05 mls of concentrated HCl was injected to convert all bicarbonate to CO_2 (Garrels and Christ, 1965). The bottles were agitated for 10 minutes on a wrist shaker to partition dissolved CO_2 and CH_4 into the headspace. 0.2 ml of headspace was injected into a Varian 3800 GC with a haysep D column, a ruthenium methanizer, and a FID. The detection limit for CO_2 was 6 ppm and 1.6 ppm for CH_4. A Star WorkStation program integrated peak areas. Seven CO_2 standards ranging from 20 to 19900 ppm and seven CH_4 standards ranging from 1.6 to 20100 ppm and were used to calibrate peak areas. Only calibration curves with an $R^2 > 0.98$ were accepted. A full set of standards was analysed before and after a sample run and several standards were analysed throughout a run as checks. All samples were run in duplicate.

Gas concentrations (ppm) were corrected for temperature and barometric pressure. Where several samples were taken over the 4-5-hour interval, a linear regression of CO_2 or CH_4 build-up with an $R^2 > 0.80$ was required for a rate to be calculated. Rates of accumulation in chambers (mg DIC-C and CH_4-C m^{-2} day^{-1}) were calculated from the difference between initial and final concentrations of CO_2 and CH_4 in the chambers, the volume of the chamber and collar, and the sediment surface area within the collar. To calculate the contribution of decomposition to the carbon budget, mean rates measured at 3-5 chamber sites were multiplied by the area of each reservoir and the number of days between the consecutive sampling times to get the mean amount of carbon produced per interval. All intervals over the flooded period were summed for each reservoir to get the total net amount of DIC-C and CH_4-C produced presumably by decomposition of flooded organic carbon.

Regression analyses of both sediment temperature (mean of temperatures collected at 0 and –5 cm) and water depth with sediment GHG fluxes were conducted. Each reservoir contained the same two thermocouple sites used for preflood measurements. Water level was recorded at the 4-5 sampling sites in each reservoir. An ANOVA of sediment DIC and CH_4 flux (mg C m^{-2} day^{-1}) between reservoirs was performed using within reservoir variance to determine differences between reservoirs (Wilkinson, 1999). ANOVA was also used to determine if there were differences between preflood and postflood fluxes using the reservoirs as replicates.

Inflows and outflows

Inflow and outflow samples were collected in evacuated Wheaton bottles of known volume (approximately 160 ml) and backfilled with 10 mls UHP nitrogen gas. After collection, 0.2 mls of concentrated HCl was injected to convert all bicarbonate to CO_2, and analysis of DIC and CH_4 followed the same procedure outlined for decomposition samples. Two injections were analyzed from each sample bottle.

Daily mean flow rate (L s^{-1}) and total cumulative water volume entering the reservoirs (m^3) were measured using in-line flow meters (Beaty, 2000). Gas samples were collected weekly from a sampling port in the inflow pipe close to the Roddy Lake pump.

Occasionally samples were taken at the inflow pipe to each reservoir to ensure that gas concentrations did not change between the pipe intake and the input of water to the reservoirs. Gauged V-notch weirs and water level recorders were used to quantify water volume exiting the reservoirs (Beaty, 2000). Outflow samples were collected every 3-6 days just above the V-notch weir of each reservoir. Initially, samples were collected in duplicate, but concentrations between samples were found to be similar, so only one sample was collected after the second week of flooding. Inflow or outflow DIC and CH_4 concentrations (μmol L^{-1}) were averaged over two consecutive sampling dates and multiplied by the total volume of water entering or exiting each reservoir between those dates. All intervals were summed to estimate the total amount of DIC and CH_4 entering or exiting the reservoirs over the flooded period. Carbon leaving the reservoirs at the end of the sampling season in water that flowed out the drainage pipe was estimated by applying concentrations measured in the drainage water to total volume of water that exited the reservoirs through the drainage pipe. The volume of water exiting the drainage pipe was calculated by subtracting the volume of the top 10 cm (water that flowed out the weir once the pump was turned off) from the total volume of the reservoir.

Direct runoff and precipitation

Each reservoir received direct runoff (DRO) from small undiked and ungauged areas of upland forest. Water inputs from these areas were estimated from the water yield at the nearby gauged Lake 239 northwest inflow (L239NWIF) (Figure 1) and corrected for the drainage area of each reservoir (Beaty, 2000). The high C reservoir had the largest drainage area of 4.76 ha, while the intermediate and low C reservoirs had very small drainage areas (0.59 and 0.04 ha, respectively). DIC and CH_4 concentrations measured in runoff from subcatchments prior to flooding (during spring melt or periods of high rainfall) were applied to these water yields because runoff from the area could rarely be measured directly after flooding. Mass of carbon entering the reservoirs via direct runoff was estimated by multiplying the adjusted water yield from L239NWIF with the average CH_4 and DIC concentrations measured in preflood runoff waters.

Precipitation inputs of DIC and CH_4 to the reservoirs were estimated by multiplying average concentrations measured in ELA precipitation by the total amount of rain that fell on the reservoirs (throughfall). Throughfall input was estimated from data collected at the ELA meteorological station and multiplied by 0.554 to account for loss due to evapotranspiration (St. Louis *et al.*, personal communication). Trees retained the majority of their foliage for the entire 1999 flooded period.

Seepage and Evaporation

The total amount of water that seeped out the high and intermediate C reservoirs along the dikes was calculated as a residual term in the water balance equation (Beaty, 2000). The calculation for seepage in the low C reservoir involved estimating high water losses through a large bedrock fracture as well as water seeping along the dike wall. As in the high and intermediate C reservoirs, seepage was calculated as a residual term; however, seepage along the dikes was estimated by direct measurement of pooled rivulets using flow meters. This dike seepage estimate was subtracted from the total seepage to estimate the volume of water loss via the bedrock fracture in the low C reservoir. Gas concentrations in seepage were not applied to the total amount of water seeping out of each reservoir because processes affecting water enroute through soils and dikes are not known. To estimate the maximum amount of carbon lost in seepage, the total seepage volume of each reservoir was multiplied by mean DIC and CH_4 concentrations measured in the sipper bottom waters (20 cm to 2 cm above sediments; see below). To estimate the minimum amount of carbon lost in seepage, mean concentrations of DIC and CH_4 measured in outflow waters was applied to the total volume that seeped out of that reservoir. For the high and intermediate C reservoirs, although the maximum and minimum range was calculated, we assumed most of the water lost through slow seepage originated from the bottom of the reservoirs and hence, we used bottom water concentrations in budget seepage calculations. In the low C reservoir, we assumed that the large volume of seepage water originated from the water column and hence, outflow concentrations were used in the budget seepage calculations. Evaporation pans placed in the shallows of the intermediate and low C reservoirs measured water losses via evaporation (Beaty, 2000).

Ebullition

Volume of gas ebullating directly from the reservoirs in bubbles was estimated using 15 cm diameter inverted funnels floating just below the surface of the reservoirs. Five traps were deployed in each reservoir and sampled weekly for bubble volumes. Bubbles from sediments had very high concentrations of CO_2 and CH_4. Once trapped, CO_2 and CH_4 diffused back into the water column. Therefore, fresh bubbles were collected in another funnel trap by probing the sediments. One ml of air was removed from aired out 13 ml Wheaton bottles and 1 ml of fresh bubble gas was injected into the bottle. Bubble gas was injected into air to oxidize any sulfide prior to injection into the GC methanizer. Mass of carbon exiting the reservoirs in bubbles was estimated by multiplying the weekly ebullition volume from each reservoir by measured CO_2 and CH_4 concentrations, and summing to estimate the total flux for the season.

Challenges to calculating the mass balance budget for CO_2 and CH_4

An unmeasured sink of CO_2 in the reservoirs was uptake via photosynthesis, and as a result, the estimate of *net* CO_2 flux out the surface of the reservoirs is high. Microbial CH_4 oxidation was also not quantified. However, $\delta^{13}CH_4$ isotope data showed that CH_4 outflow signatures were very depleted (-64 $^0/_{00}$ to -89 $^0/_{00}$) indicating a microbial production of CH_4, but no enrichment by microbial CH_4 oxidation (Boudreau, 2000).

Other measurements

Surface water samples were collected and analyzed weekly for dissolved CO_2 and CH_4 from 5 sites in each reservoir. Water was collected just below the surface in 160 ml evacuated Wheaton bottles with a 10 ml UHP nitrogen headspace and 9 grams of KCl for preservation.

Each reservoir had two "sippers" (interval samplers) that were sampled biweekly to determine how well mixed the water column was and where the highest concentrations of GHG occurred. Sippers were constructed by attaching tygon tubing (5 mm outer diameter) to PVC pipes. Samples were collected every 20 cm throughout the water column and approximately every 5 cm into the sediments to 20 cm. In the high C reservoir, one sipper was located at site 3 (*Pinus/Polytrichum* community) and the other sipper was located at site 1 (*Pinus/Ledum/Sphagnum* community). Sippers in the intermediate C reservoir were placed at sites 3 and 5 in the homogeneous *Pinus/Betula* community and both sippers in the low C reservoir were placed in the *Pinus/ Vaccinium* community (sites 1 and 3). CO_2 and CH_4 concentrations were measured in the water column every 20 cm and every 5-15 cm in the sediment pore water. Water was drawn into a 60 ml syringe with a three-way valve and transferred into evacuated 60 ml bottles with a 5 ml UHP nitrogen headspace. Samples were acidified and analyzed as previously described.

RESULTS

Temperatures prior to and following flooding

There was a strong seasonal and depth related pattern to soil temperatures prior to flooding (Figure 6). Soil temperatures prior to flooding were highest in the intermediate C subcatchment with a mean (± standard error) soil temperature of 17.2±1.5 °C, ranging from 10.2 °C to 30.0 °C (Figure 6). Mean temperature was second highest in the low C subcatchment at 16.3±1.1 °C, ranging from 9.5 °C to 27.8 °C. The high C subcatchment had the lowest soil temperatures, ranging from 8.0 °C to 27.7 °C with a mean temperature of 14.9±1.1 °C. The highest temperatures, and greatest differences between depths, were observed mid-season in all three subcatchments. Soil temperatures were generally highest at the soil surface and declined with depth. However, in October 1998, as the air cooled, the temperatures differed less between depths, and were actually higher in the lower depths than near the soil surface.

Sediment temperatures after flooding ranged much less between depths and sampling dates than they did prior to flooding (Figure 6). The highest mean sediment temperature of 17.9±0.6 °C was observed in the low C reservoir followed by the intermediate C reservoir at 17.5±0.6 °C. Mean sediment temperatures were lowest in the high C reservoir (14.5±0.4 °C). The highest sediment temperatures were observed at 0 cm and decreased with depth. In September 1999, however, as the overlying water cooled, the inverse was observed and sediment temperatures increased with depth. In general, sediment temperatures after flooding were highest at the beginning of the sampling season and decline toward October.

Preflood groundwater levels and postflood water column height

Water levels prior to flooding never approached the soil surface in any of the subcatchments (Figure 7), and most wells contained very little water with the exception of 2 sampling events (Table 2). In the high C subcatchment, water wells that were often dry showed large increases in water levels on 30 June and 10 October 1998 following major rain events (Table 2). On these two occasions, water levels rose to approximately

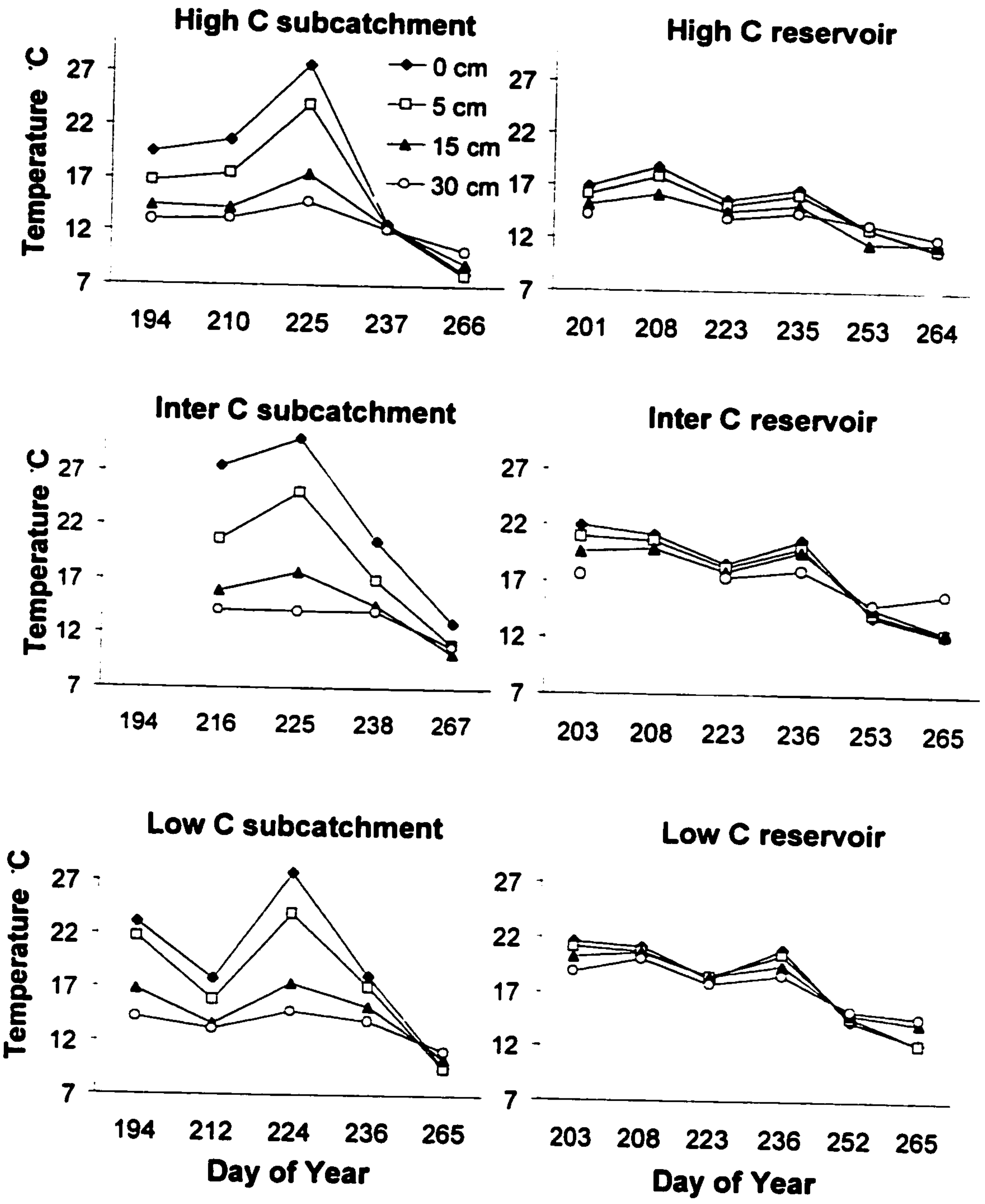


Figure 6. Soil temperatures (°C) in the three subcatchments before and after flooding. Measurements were collected at the soil surface (0 cm) and at 5, 15 and 30 cm below the soil surface.

28 cm below the soil surface. The overall mean (± std err) water table in the high C subcatchment was 36±2 cm below the soil surface. High water levels (water approximately 29 cm below the soil surface) were only seen on 10 October 1998 in the intermediate C subcatchment, which had a mean water table of 39±1 cm below the soil surface. The low C subcatchment had a low water table throughout the season and no increase on 30 June or 10 October 1998 (mean of 27±2 cm below the soil surface). After flooding, water column height at the five sampling sites in the high C reservoir ranged from 12 to 125 cm with a mean depth of 92±21 cm (Figure 7). In the intermediate C reservoir water column height ranged from 9 to 130 cm with a mean depth of 71±21cm above the collar sites. The water column heights in the low C reservoir ranged from 37 to 153 cm with a mean of 94±23 cm.

Overall comparison of preflood soil CO_2 fluxes to postflood sediment CO_2 fluxes

Postflood fluxes from the sediments to the bottom water were significantly higher than preflood fluxes of CO_2 from the soils to the atmosphere ($p<0.001$, Table 3) when all three reservoirs were treated as replicates. Mean carbon from CO_2 (CO_2-C) fluxes increased 2.5 times, from 2400±530 mg CO_2-C m^{-2} day^{-1} prior to flooding to 6300±3200 mg DIC-C m^{-2} day^{-1}, in the high C reservoir after flooding (Figure 8). The mean rate of C production in the intermediate C reservoir increased by a factor of 3.4 following flooding, from 2700±280 mg CO_2-C m^{-2} day^{-1} to 9200±3010mg DIC-C m^{-2} day^{-1}. In the low C reservoir, there was a 2.2-fold increase in C production from 1900±240 mg CO_2-C m^{-2} day^{-1} prior to flooding to 4200±1750 mg DIC-C m^{-2}post flooding (Figure 8).

A comparison between subcatchments of preflood CO_2-C fluxes from forest soils to the atmosphere

Prior to flooding, the soil and ground vegetation in all subcatchments were found to respire 680-4200 mg CO_2-C m^{-2} day^{-1} in the dark (Figure 8, Appendix 1). Average (± std err) rates of respiration were highest in the intermediate C subcatchment (2700±280 mg CO_2-C m^{-2} day^{-1}) followed by the high C (2400±530 mg CO_2-C m^{-2} day^{-1}) and low C subcatchments (1900±240 mg CO_2-C m^{-2} day^{-1}). Although CO_2 fluxes were variable at

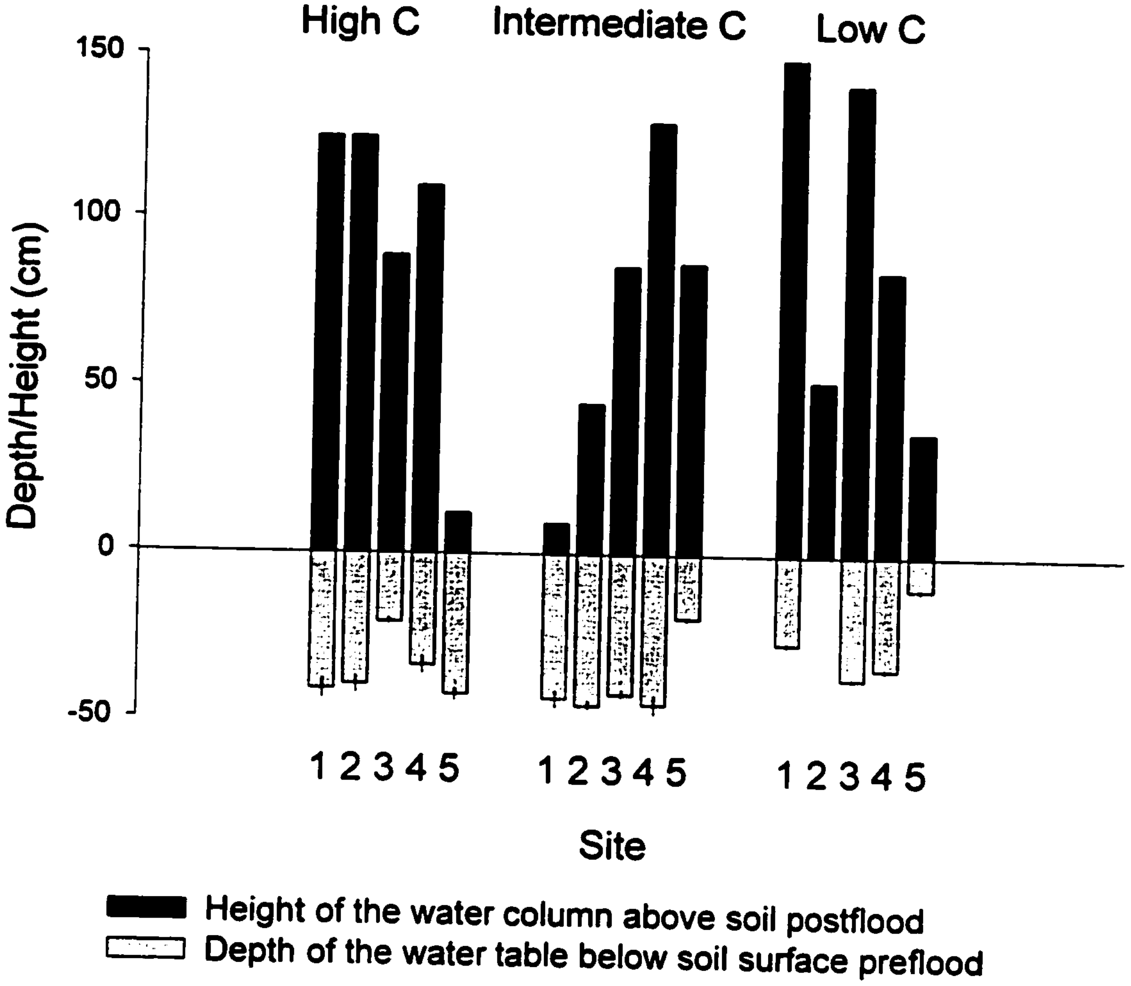


Figure 7. Mean depth of the water table (±std err) in cm below surface prior to flooding and height of the water column (cm) above the soil surface after flooding.

Table 2. Groundwater levels (cm) in the water wells of three boreal subcatchments.

Subcatchment	High C					Intermediate C					Low C			
Sampling site	1	2	3	4	5	1	2	3	4	5	1	3	4	5
Total well depth from soil surface	**-47.5**	**-43.5**	**-21.5**	**-38.0**	**-45.0**	**-46.5**	**-46.5**	**-46.5**	**-48.5**	**-21.0**	**-28.5**	**-38.0**	**-36.0**	**-12.0**
Date	Distance to groundwater from soil surface													
17 June	-43.0	-43.5	-21.0	-37.5	-45.0	-46.5	-46.5	-46.0	-48.5	-21.0				
30 June	-23.0	-26.0	-20.5	-20.5	-33.0	-46.5	-45.7	-43.0	-48.5	-18.5	-26.0	-36.5	-33.2	-8.5
14 July	-47.0	-41.5	-20.5	-38.0	-44.5	-46.5	-45.5	-42.5	-48.0	-18.5	-26.0	-36.5	-36.0	-10.0
31 July	-40.0	-42.5	-21.0	-38.0	-45.0	-46.5	-46.0	-44.0	-48.5	-19.5	-27.5	-37.0	-34.0	-9.5
13 August	-47.5	-41.5	-21.0	-37.5	-45.0	-46.5	-45.0	-44.0	-48.0	-19.5	-27.5	-36.5	-34.0	-11.0
25 August	-47.0	-41.0	-20.5	-37.5	-45.0	-46.5	-45.0	-44.0	-48.0	-19.5	-27.0	-37.0	-33.5	-10.5
23 September	-47.0	-40.5	-21.0	-38.0	-45.0	-46.0	-46.0	-44.0	-47.5	-19.0	-26.5	-37.5	-33.5	-9.5
10 October	-34.5	-40.5	-21.5	-23.5	-36.5	-24.0	-46.0	-31.0	-24.5	-19.0	-26.5	-38.0	-34.0	-11.0
Mean water depth	**-41.1**	**-39.6**	**-20.9**	**-33.8**	**-42.4**	**-43.6**	**-45.7**	**-42.3**	**-45.2**	**-19.3**	**-26.7**	**-37.0**	**-34.0**	**-10.0**

individual collars sites and between sites in each subcatchment on a given day (Appendix 1), significant differences amongst subcatchments (ANOVA, $p=0.011$, Table 3) were observed. Rates of dark respiration in the intermediate C subcatchment were significantly higher than those in the low C subcatchment ($p=0.010$, Table 3). Rates of dark respiration in the high C subcatchment did not differ significantly from rates measured in the intermediate or low C subcatchments ($p>0.05$). Regression analysis showed no significant relationship between pooled preflood CO_2 fluxes from all sites and subcatchments and mean surface (0 and -5 cm) soil temperatures ($p>0.05$, Table 4). No significant relationship was seen between water table height and rates of CO_2 dark respiration ($p>0.05$, Table 4). However, regressions of pooled subcatchment CO_2 flux with time show a significant seasonal decline ($p<0.001$, $r^2=0.175$, Table 4).

Table 3. Results from ANOVA tests comparing chamber fluxes of pooled subcatchments preflood vs. pooled reservoirs postflood and a comparison of preflood and postflood chamber fluxes amongst the three experimental boreal subcatchments. NS= not significant

Between Subcatchments/ Reservoirs	**Factor** (subcatchment /reservoir)	High C	Intermediate C	Low C
Reservoirs Pooled (Pre vs. Postflood)	**Factor** (Flooding)			
CO_2 flux n=202	$p<0.001$			
CH_4 flux n=198	$p=0.001$			
Preflood CO_2 flux	$p=0.011$, $n=120$			
High C		1		
Intermediate C		NS	1	
Low C		NS	$p=0.010$	1
Postflood CO_2 flux	$p=0.050$, $n=82$			
High C		1		
Intermediate C		NS	1	
Low C		NS	$p=0.049$	1
Preflood CH_4 flux	$p=0.001$, $n=116$			
High C		1		
Intermediate C		$p=0.001$	1	
Low C		NS	$p=0.013$	1
Postflood CH_4 flux	$p=0.010$, $n=82$			
High C		1		
Intermediate C		$p=0.035$	1	
Low C		$p=0.019$	NS	1

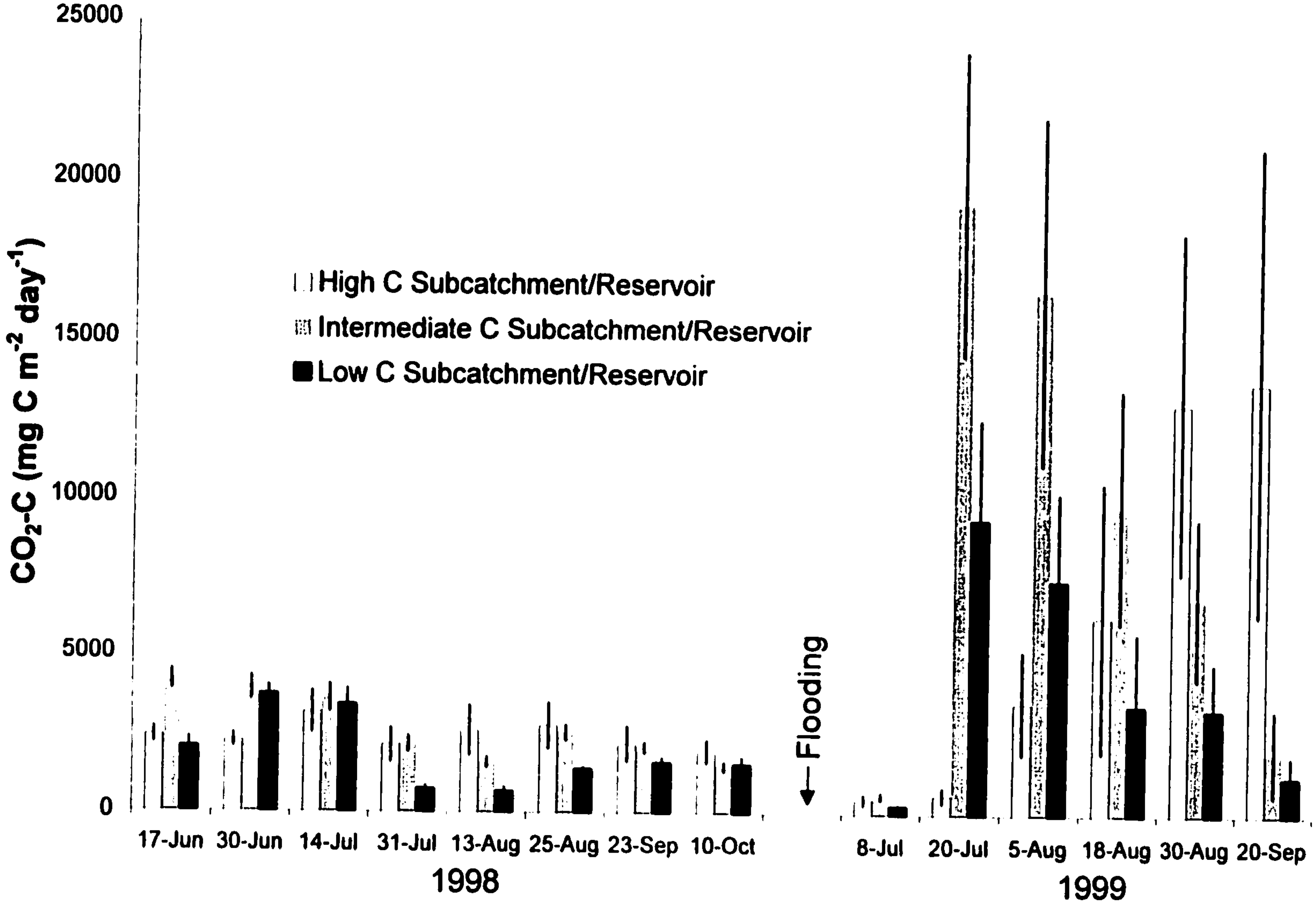


Figure 8. Mean (± std err) CO_2-C and DIC flux (mg C m^{-2} day^{-1}) from the soils in three subcatchments before and after flooding. Flooding occurred on June 22, 1999.

Table 4. Linear regressions of chamber flux rates to soil temperature, water depth, and sampling date in the three experimental boreal subcatchments before and after flooding. NS=not significant.

Subcatchments Pooled	Preflood Fluxes				Postflood Fluxes			
	CO_2		**CH_4 oxidation**		**DIC**		**CH_4 emission**	
Soil temperature	p	r^2	p	r^2	p	r^2	p	r^2
	NS	NS	<0.001	0.179	NS	NS	0.001	0.122
n =	85		84		82		82	
df =	83		82		80		80	
Water depth	p	r^2	p	r^2	p	r^2	p	r^2
	NS	NS	0.013	0.058	NS	NS	NS	NS
n =	108		105		82		82	
df =	106		103		80		80	
Date	p	r^2	p	r^2	p	r^2	p	r^2
	<0.001	0.175	0.043	0.035	NS	NS	0.008	0.084
n =	120		116		82		82	
df =	118		114		80		80	

A comparison between reservoirs of postflood DIC fluxes from the sediments to the bottom waters

Rates of sediment DIC production varied considerably within and between reservoirs over the flooded period (265-19300 mg DIC-C m^{-2} day^{-1}; Appendix 3), but significant differences were observed between reservoirs in DIC-C flux (ANOVA, p=0.050, Table 3). The highest mean rate of sediment DIC-C production occurred in the intermediate C reservoir (9166±3015 mg C m^{-2} day^{-1}) and was significantly different (p=0.049, Table 3) from the lowest rate occurring in the low C reservoir (4171±1749 mg C m^{-2} day^{-1}). Mean rates of DIC-C flux in the high C reservoir (6300±3200 mg C m^{-2} day^{-1}) were not significantly different than the low C or intermediate C reservoirs (p>0.050, Table 3). Within each reservoir, specific sampling sites generally produced similar rates of DIC flux between sampling dates (Appendix 3). For example, site 2 (located in the *Pinus/Ledum/Sphagnum* community) in the high C reservoir had high rates of flux on most sampling dates (>10000 mg DIC-C m^{-2} day^{-1}) while site 3 (located in the *Pinus/Polytrichum* community) had low fluxes (<2500 mg DIC-C m^{-2} day^{-1}). The intermediate C reservoir (homogenous *Pinus/Betula* community) had high fluxes at sites 2 and 4 and lower fluxes at sites 3 and 5. Sites in the low C reservoir had relatively less variation than the other two reservoirs, but low rates of flux were consistently observed at site 1 (located in the *Pinus/ Vaccinium* community). Regression analysis showed no significant relationship between sediment DIC-C flux and temperature (mean of temperatures collected at 0 and -5 cm) amongst all sites combined ($p>0.05$, Table 4). There was also no relationship between water column height and DIC-C sediment flux ($p>0.05$, Table 4) and no significant temporal trend ($p>0.050$, Table 4) was observed.

Overall comparison of preflood soil CH_4 fluxes to postflood sediment CH_4 fluxes

Postflood fluxes of CH_4 were significantly different than preflood soil fluxes of CH_4 ($p=0.001$, Table 3) when all three reservoirs were treated as replicates. Prior to flooding, the upland soils oxidized CH_4 at mean rates of 0.77 ±0.24, 1.08±0.13, 0.72±0.13 mg CH_4-C m^{-2} day^{-1} in the high, intermediate, and low C subcatchments, respectively (Figure 9). After flooding, the forest soils no longer oxidized CH_4, but began to emit CH_4. The most dramatic increase in mean sediment CH_4 flux rates to the bottom water was observed in the high C reservoir (110±65 mg CH_4-C m^{-2} day^{-1}). Smaller increases in sediment CH_4 flux rates was observed in the intermediate C reservoir (10±5.7 mg CH_4-C m^{-2} day^{-1}) and the low C reservoir (3.3±2.2 mg CH_4-C m^{-2} day^{-1}) following flooding.

A comparison between subcatchments of preflood CH_4-C fluxes from forest soils to the atmosphere

Forest soils oxidized CH_4 in all subcatchments at rates of 0.33 -1.31 mg CH_4-C m^{-2} day^{-1} (Figure 9, Appendix 2). Similar to patterns seen for rates of respiration, the intermediate C subcatchment had the highest mean (±std err) rate of CH_4 oxidation (1.08±0.13 mg CH_4-C m^{-2} day^{-1}). The high C subcatchment oxidized CH_4 at a mean rate of 0.77±0.24 mg CH_4-C m^{-2} day^{-1} and the lowest mean oxidation rate was observed in the low C subcatchment (0.72±0.13 mg CH_4-C m^{-2} day^{-1}). Although fluxes measured within collar sites and between collar sites are highly variable (Appendix 2), there was a significant difference amongst subcatchments in CH_4 oxidation (ANOVA, $p=0.001$, Table 3). Rates in the intermediate C subcatchment were significantly higher than rates in the high C and low C subcatchments ($p=0.001$ and $p=0.013$, respectively, Table 3). There was no significant difference in rates of CH_4 oxidation between the high and low C subcatchments ($p>0.05$). Rates of CH_4 oxidation were positively related to temperature ($p<0.001$, $r^2=0.179$, Table 4), but a significant negative relationship was observed

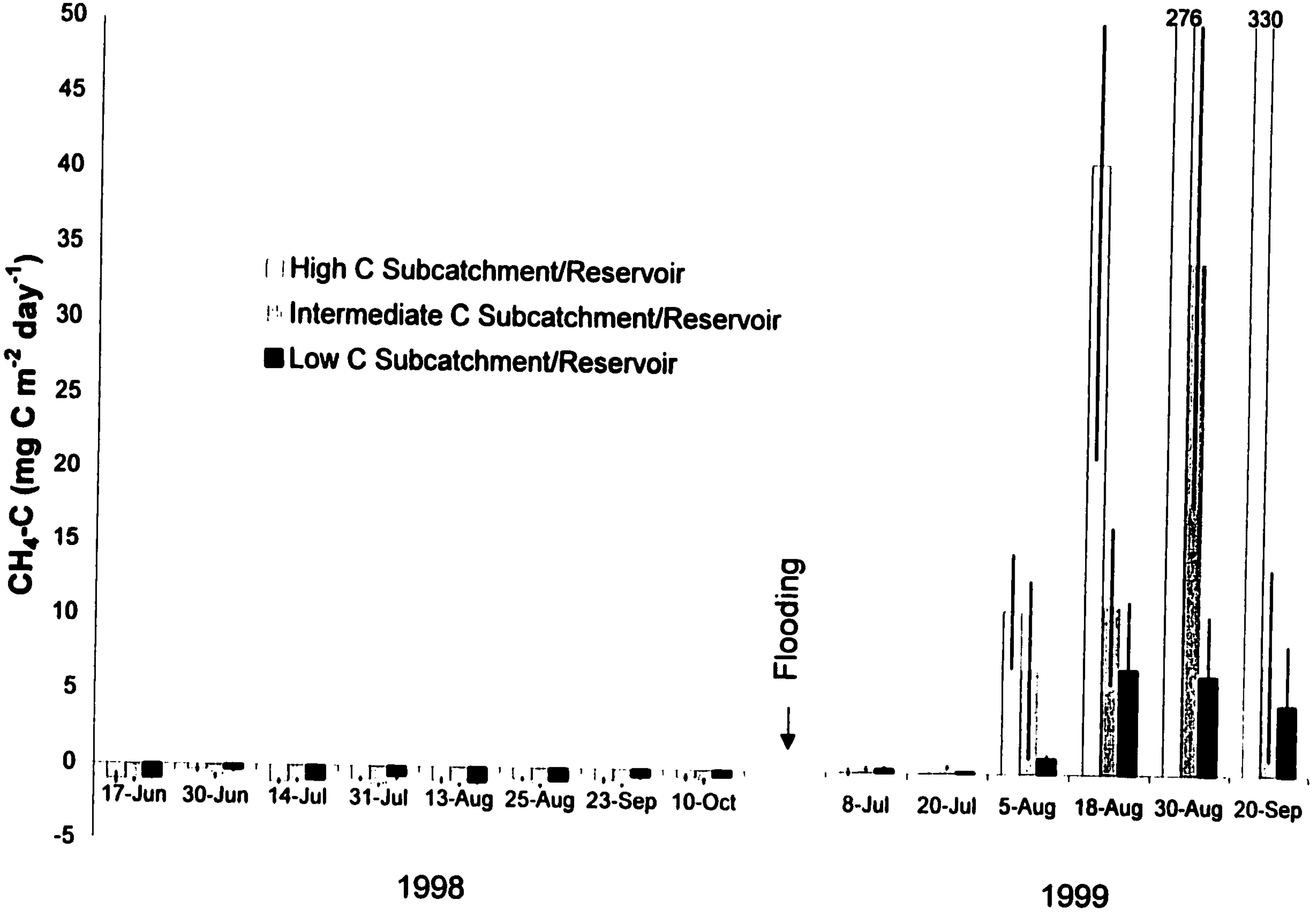


Figure 9. Mean (±std err) CH$_4$-C flux (mg C m^{-2} day^{-1}) from the soils in three subcatchments before and after flooding. Flooding occurred on June 22, 1999.

between rates of CH_4 oxidation and increases in water table height ($p=0.013$, $r^2=0.058$). Rates of CH_4 oxidation declined significantly over the season ($p=0.043$ and $r^2=0.035$).

A comparison between reservoirs of postflood CH_4-C fluxes from the sediments to the bottom waters

Mean sediment CH_4 fluxes to the bottom water were highest in the high C reservoir (110±65 mg CH_4-C m^{-2} day^{-1}) over the sampling season (Figure 9, Appendix 4). Mean rates of CH_4 production in the intermediate and low C reservoirs were at least ten fold lower (10.2 ±5.7 and 3.3±2.2 mg C m^{-2} day^{-1}, respectively) than in the high C reservoir. Rates of CH_4 flux from the sediments were significantly different amongst the reservoirs (ANOVA, $p=0.010$, Table 3). CH_4 flux in the high C reservoir were significantly higher than rates in the intermediate and low C reservoirs ($p=0.035$ and $p=0.019$, respectively). CH_4 bottom fluxes remained low until 18 August 1999, unlike DIC bottom flux that increased immediately following flooding. In the high C reservoir, high rates of CH_4 flux were observed at sites 1, 2, and 4 (the moist *Pinus/Ledum/Sphagnum* community) while rates remained low throughout the season at site 3 (the drier *Pinus/Polytrichum* community)(Appendix 4). Site 5 in the intermediate C reservoir had rates of CH_4 bottom flux four times higher than all other sites. The low C reservoir showed no discernible differences amongst sampling sites in CH_4 flux. Regressions show a significant negative relationship was observed between CH_4 sediment flux and temperature ($p=0.001$, $r^2=0.122$, Table 4). No relationship between water column height and CH_4 sediment flux ($p>0.05$) was observed, but CH_4 flux amongst all sites increased significantly over time ($p=0.008$, $r^2=0.035$).

Inflow/Outflow concentrations

Concentrations of DIC in inflow waters to all reservoirs fluctuated very little throughout the flooding period (Figure 10), averaging 131±0.3 μmol L^{-1}. There was no significant seasonal trend in inflow DIC concentrations ($r^2<0.5$, Table 5). Concentrations of DIC in reservoir outflow water were higher than in the inflow water immediately after flooding (Figure 10). The intermediate C reservoir had the highest mean DIC concentration in the outflow (431±18 μmol L^{-1}) followed by the high C reservoir (383±9 μmol L^{-1}). The lowest mean DIC outflow concentration was observed at the low C reservoir (323±12 µmol L^{-1}). Outflow DIC concentrations were consistently highest in the intermediate C reservoir until mid September 1999, when DIC concentrations in the outflow of the high C reservoir surpassed them. There was a significant decline in outflow water DIC concentrations with time in the intermediate C reservoir ($p<0.001$, $r^2=0.83$, Table 5). No temporal trend in outflow DIC concentration with time was observed in the high C and low C reservoirs ($p>0.05$).

Table 5. Linear regressions of concentrations in inflow and outflow water with sampling date in the three experimental reservoirs after flooding. NS= not significant

	DIC				CH_4			
	p	r^2	n	df	p	r^2	n	df
Inflow	NS	NS	14	12	NS	NS	14	12
Reservoir Outflow								
High C	NS	NS	19	17	<0.001	0.669	19	17
Intermediate C	<0.001	0.832	19	17	<0.001	0.604	19	17
Low C	NS	NS	19	17	<0.001	0.624	19	17

CH_4 concentrations in inflow waters (mean of 0.09±0.01 µmol L^{-1}, Figure 10) to all reservoirs showed no significant temporal trend ($p>0.05$, Table 5). In contrast to outflow DIC concentrations, which were elevated immediately after flooding, CH_4 concentrations remained low until 5 weeks after flooding when they began to increase, peaking between weeks 8 and 10 in all three reservoirs. Mean outflow CH_4 concentrations were greatest in the high C reservoir (1.1±0.3 µmol L^{-1}, Figure 10), followed by the intermediate C reservoir (0.9±0.2 µmol L^{-1}) and the low C reservoir (0.7±0.2 µmol L^{-1}). A significant positive temporal trend in CH_4 outflow concentrations was observed in the high C reservoir ($p< 0.001$, $r^2=0.669$, Table 5), the intermediate C reservoir ($p<0.001$, $r^2=0.604$), and the low C reservoir ($p<0.001$, $r^2=0.624$).

Carbon inputs in the high C reservoir

Decomposition of flooded organic matter on the bottom of the reservoirs, measured by submerged chambers, was by far the largest input of carbon to all reservoirs (94-97% of total inputs). The high C reservoir had the largest mean total input of DIC-C

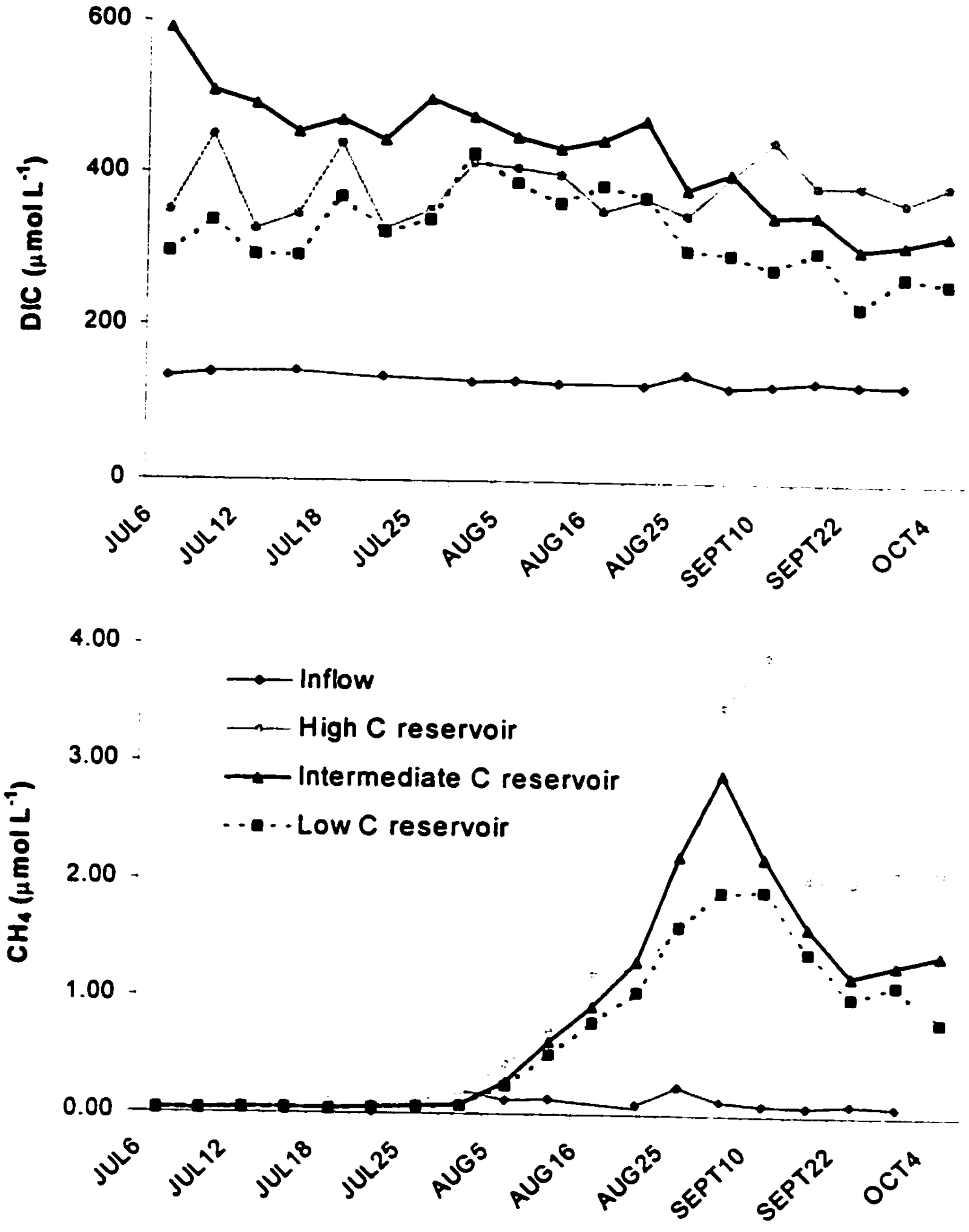


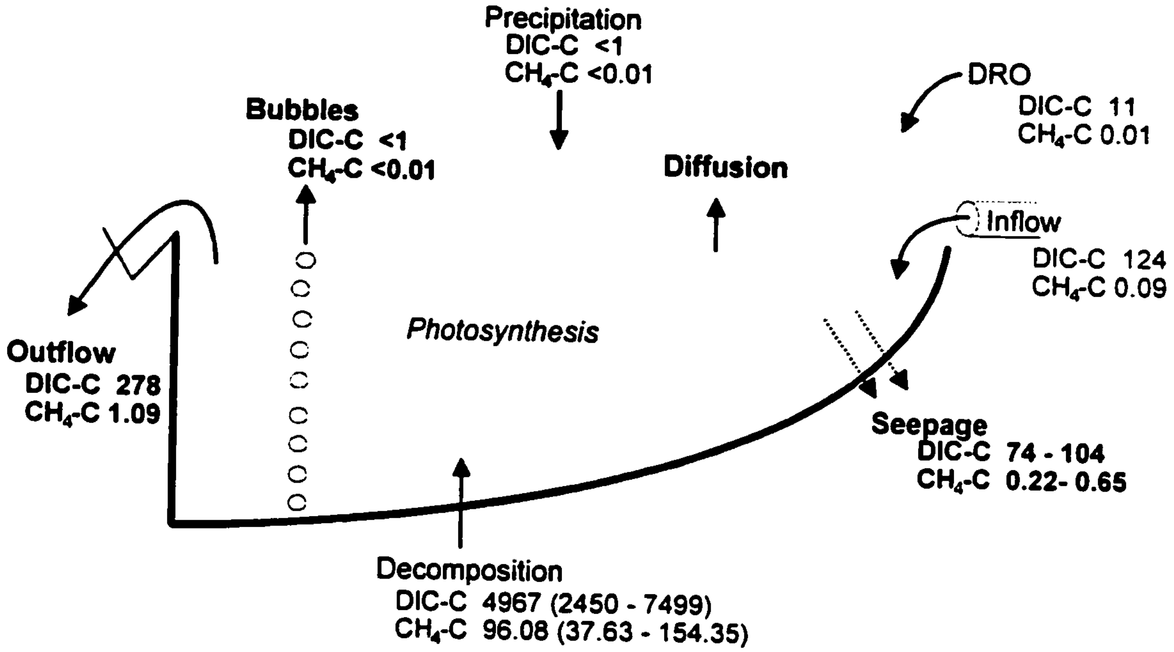
Figure 10. DIC and CH_4 concentrations (μmol L^{-1}) in the inflow and outflow waters of the FLUDEX reservoirs.

(5103 Kg, Table 6, Figure 11) and CH_4-C (96 kg). Both the total volume of inflow water (Beaty, 2000, Table 6) and mass of inflow carbon were similar in the high and intermediate C reservoirs (Table 6). The high C reservoir received 124 Kg of DIC-C and 0.09 Kg of CH_4-C via 78600 m^3 of inflow water. DIC-C and CH_4-C in direct runoff from the small uplands surrounding the reservoirs was greatest in the high C reservoir (11 Kg of DIC-C and 0.01 Kg of CH_4-C) where the area of the drainage area was largest (4.76 ha yielding 3954 m^3 of runoff water). Mean concentrations of DIC in precipitation were 34 μmol L^{-1} and 0.15 μmol L^{-1} of CH_4. Annual throughfall volume was small in the high C reservoir (1162 m^3, Table 6) and contributed very little carbon to the mass balance budget (under one Kg of C) in this reservoir. The contribution of mean decomposition on a mass basis was greatest in the high C reservoir, accounting for 4970 Kg of DIC-C and 96.1 Kg of CH_4-C (Table 6).

Carbon outputs in the high C reservoir

The high C reservoir had a total output of 455 Kg of DIC-C and 3 Kg of CH_4-C (based on assigning bottom water concentrations to seepages) in 83709 m^3 of outgoing water (Table 6, Figure 11). Total volume of weir discharge (Beaty, 2000) and mass of carbon exiting the weir was similar in the high and intermediate C reservoirs (Table 6). A volume of 60074 m^3 water discharged 278 Kg of DIC-C (60% of total DIC-C output) and 1.09 kg of CH_4-C (35% of total CH_4-C output) in the high C reservoir. Outputs of DIC-C and CH_4-C in seepages were much slower (approximately 20% of the total output) in the high and intermediate C reservoirs than the low C reservoir and hence, bottom water concentrations were used in the calculations. In the high C reservoir, dike seepage (15912 m^3 water) accounted for a loss of 104 Kg of DIC-C and 0.65 Kg of CH_4-C. The amount of DIC-C lost in water that flowed out the drainage pipe during drawdown was < 10% of the total DIC-C output in this and the other reservoirs. For CH_4-C, however, approximately 30% of the total output was lost in the drainage waters of all three reservoirs. The high C reservoir lost 56 Kg of DIC-C and 0.97 Kg of CH_4-C in 6250 m^3 of drawdown water (Table 6). An evaporation rate of 2.2 mm day^{-1} was estimated for all reservoirs by the evaporation pans (Beaty, 2000). The high C reservoir lost 1473 m^3 of water to evaporation (Table 6).

High C Reservoir



	CO_2	CH_4
Total Inputs (Kg C): mean (min-max)	5103 (2585-7634)	98 (38-154)
Total Outputs (Kg C): max (min-max)	**455 (408-455)**	**3 (2-3)**
Difference (Kg C): mean (min-max)	4608 (2177-7179)	93 (36-152)
g C m^{-2} $season^{-1}$	675	13.6

Figure 11. Inputs and outputs of GHGs (Kg C) from the 1999 flooded season in the high C reservoir.

Carbon inputs in the intermediate C reservoir

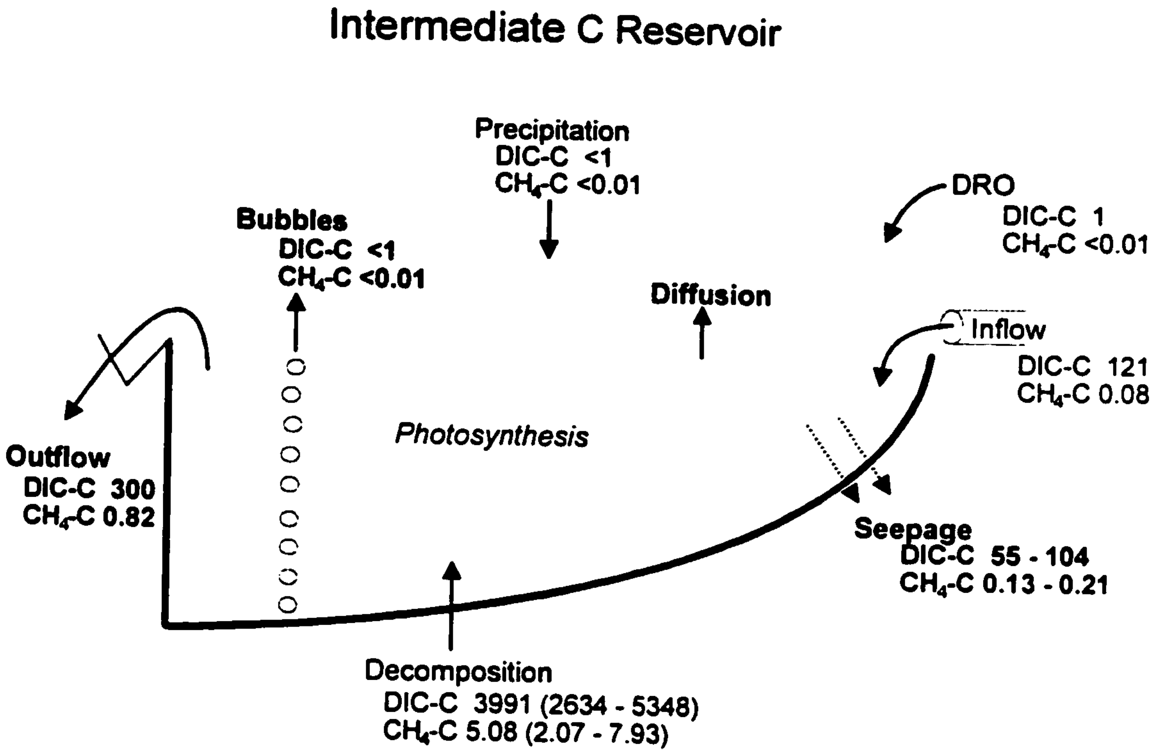
The intermediate C reservoir had the second highest mean DIC-C (4113 Kg) and CH_4-C (5 Kg) input (Table 6, Figure 12). Similar to the high C reservoir, the intermediate C reservoir received 121 Kg of DIC-C and 0.08 Kg of CH_4-C via 76748 m^3 of inflow water. This reservoir had a much smaller upland drainage area than the high C reservoir (0.59 ha yielding 476 m^3 water). The contribution of direct runoff to the mass balance from the intermediate C reservoir was less than one Kg of DIC-C and CH_4-C (Table 6, Figure 12). Annual throughfall in this reservoir (840 m^3) contributed very little carbon to the mass balance budget (under one Kg of C). The intermediate C reservoir had similar mean DIC-C decomposition inputs (3990 Kg) as the high C reservoir, but the CH_4-C input was 20 times lower (5.1 Kg).

Carbon outputs in the intermediate C reservoir

Total outputs of 437 Kg of DIC-C and 1 Kg of CH-C in 78064 m^3 (Table 6, Figure 12) were observed in the intermediate C reservoir (based on assigning bottom water concentrations to seepages). In the intermediate C reservoir, 60661 m^3 of outflow water transported 300 Kg DIC-C (approximately 70% of the total output) and 0.82 Kg CH_4-C (60% of the total). Outputs of DIC-C and CH_4-C in seepages of this reservoir (approximately 20% of the total output) accounted for a loss of 104 Kg of DIC-C and 0.21 Kg of CH_4-C in 12686 m^3 of seepage water. 28 Kg of DIC-C and 0.32 Kg of CH_4-C were exported in 3660 m^3 of drawdown water and 1057 m^3 evaporated from the intermediate C reservoir over the 1999 season.

Carbon inputs in the low C reservoir

The low C reservoir had the smallest DIC-C (2553 Kg) and CH_4-C (2 Kg) total mean input over the whole flooded period (Table 6, Figure 13). The highest amounts of DIC-C (153 Kg) and CH_4-C (0.10 Kg) in inflow water (96900 m^3) were delivered to the low carbon reservoir where high levels of seepage required that more inflow water be pumped into this reservoir than the others to keep it full. This reservoir had the smallest upland drainage areas (0.04 ha yielding 30 m^3 water) of all the reservoirs. The contribution of direct runoff to the mass balance of the low C reservoir was less than



	CO_2	CH_4
Total Inputs (Kg C): mean (min-max)	4113 (2756-5470)	5 (2-8)
Total Outputs (Kg C): max (min-max)	**437 (303-437)**	**1 (1-1)**
Difference (Kg C): mean (min-max)	3676 (3139-5033)	4 (1-7)
g C m^{-2} $season^{-1}$	735	1

Figure 12. Inputs and outputs of GHGs (Kg C) from the 1999 flooded season in the intermediate C reservoir.

Table 6. Mass balance budget of water and carbon (DIC-C or CH_4-C) (Kg) in the high, intermediate and low carbon reservoirs. Means, minimums, and maximums are given for the decomposition term. The seepage term is calculated using outflow concentrations (minimum) and bottom water concentrations (maximum). All water budget data are from Beaty (2000).

	Total water (m^3)			DIC (Kg of carbon)			CH_4 (Kg of carbon)		
Reservoir	High C	Intermediate C	Low C	High C	Intermediate C	Low C	High C	Intermediate C	Low C
Inputs									
Inflow	78593	76748	96898	124	121	153	0.09	0.08	0.10
Direct run-off	3954	476	30	11	1	<0	0.01	<0.00	<0.00
Throughfall	1162	840	1035	<0	<0	<0	<0.00	<0.00	<0.00
Decomposition mean				4967	3991	2399	96.08	5.08	2.39
(min-max)				(2450-7499)	(2634-5348)	(1384-3420)	(37.6-54.4)	(2.1-7.9)	(0.70-3.87)
Total inputs	**83709**	**78064**	**97963**	**5103**	**4113**	**2553**	**96**	**5**	**2**
(min-max)				**2585-7634**	**2756-5470**	**1537-3575**	**38-155**	**2-8**	**1-4**
Outputs									
Outflow weir	60074	60661	33463	278	300	125	1.09	0.82	0.37
Dike seepage	15912	12686	23052	74-104	55-104	89-454	0.22-0.65	0.13-0.21	0.19-0.55
Bedrock seepage	-	-	33125	-	-	123-653	-	-	0.27-0.80
Drawdown	6250	3660	6987	56	28	37	0.97	0.32	0.79
Evaporation	1473	1057	1336	-	-	-	-	-	-
Total outputs	**83709**	**78064**	**97963**	**408-455**	**383-437**	**374-1269**	**2-3**	**1**	**2-3**
Inputs-Outputs	**0**	**0**	**0**	**4658**	**3676**	**2179**	**93**	**4**	**0**
(min-max)				**2177-7179**	**3139-5033**	**1163-2306**	**36-152**	**1-7**	**-1-1**
Season summary for 92-98 flooded days[1] (g C m^{-2} $season^{-1}$)				**675**	**735**	**330**	**14**	**1**	**0**

[1] The season summary is calculated using mean values for inputs and outputs.

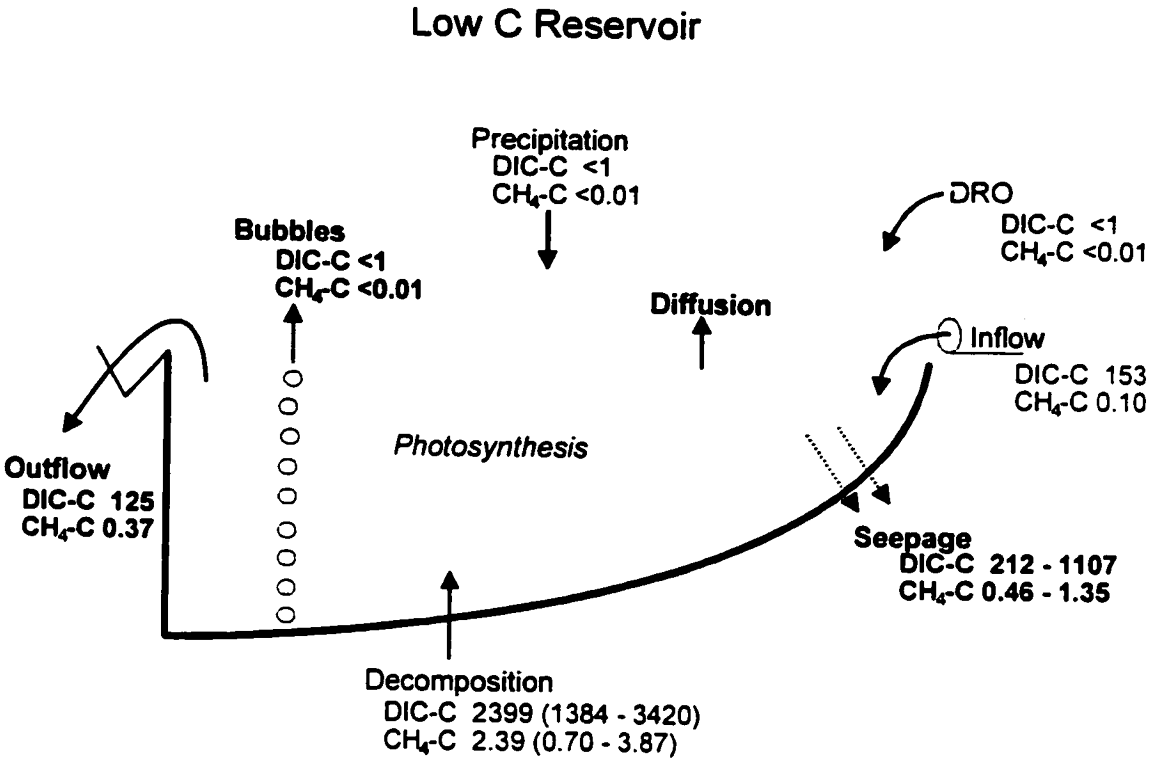
one Kg of DIC-C and CH_4-C. Annual throughfall volume was also small (30 m^3, Table 6) and contributed less than one Kg of C in this reservoir. The lowest mean inputs of carbon from decomposition occurred in the low C reservoir where DIC-C added 2400 Kg and CH_4-C added 2.4 Kg (Table 6, Figure 13).

Carbon outputs in the low C reservoir

Total outputs of water (97963 m^3, Beaty, 2000) and DIC-C (374-1269 Kg) would have been greatest in the low C reservoir due to the large volume of water that seeped out of this reservoir (approximately 60% of the total inflow, Table 6, Figure 13). However, because we used outflow concentrations in the seepage calculations, the amount of DIC-C lost in seepage (374 Kg) was similar to that seen in the other reservoirs. Total CH_4-C output from this reservoir was 3 Kg. Only 30% (125 Kg) of total DIC-C output and 20% (0.37 Kg) of total CH_4-C output was discharged in 33463 m^3 of outflow water through the weir in the low C reservoir. Large volumes of seepage (33125 m^3 water, Beaty, 2000) through bedrock fractures were observed only in the low C reservoir (Table 6). Total DIC-C seepage loss (212 Kg) included bedrock fractures (123 Kg) and dike leakage (89 Kg), and accounted for 60% of total output of DIC-C. Bedrock seepage exported 0.27 Kg CH_4-C and dike seepage exported 0.19 Kg CH_4-C in the low C reservoir. A total of 37 Kg of DIC-C and 0.79 Kg of CH_4-C were discharged in 6987 m^3 water from the low C reservoir during drawdown. The low C reservoir lost 1336 m^3 of water to evaporation, Table 6).

Overall Inputs - Outputs

The C budget for the 92-98 day flooded period in 1999 was largely driven by decomposition inputs of carbon. In the high and intermediate C reservoirs, mean inputs minus outputs yielded 4658 Kg and 3676 Kg of DIC-C, respectively (Table 6, Figures 11 and 12). Due to the high volume of seepage and lower rate of decomposition in the low C reservoir, inputs minus outputs yielded only 2179 Kg of carbon. An excess of 93 Kg CH_4-C was calculated in the high C reservoir, 20 times more than the excess in the intermediate C reservoir (4 Kg CH_4-C). We calculated more outputs than inputs of CH_4-C (-0.36 Kg) in the low C reservoir due to the large volume of seepage and



	CO_2	CH_4
Total Inputs (Kg C): mean (min-max)	2553 (1537-3420)	2 (1-4)
Total Outputs (Kg C): max (min-max)	**374 (374-1269)**	**2 (2-3)**
Difference (Kg C): mean (min-max)	179 (1163-2306)	0 (-1-1)
g C m^{-2} $season^{-1}$	330	1

Figure 13. Inputs and outputs of GHGs (Kg C) from the 1999 flooded season in the low C reservoir.

relatively low rates of anaerobic decomposition. When calculated on a unit area basis (g C m^{-2} $season^{-1}$) for the 92-98 day season, the intermediate C reservoir actually had the highest excess DIC-C (735 g C m^{-2} $season^{-1}$). The high C reservoir had the second largest excess of DIC-C (675 g C m^{-2} $season^{-1}$) and the low C reservoir had the lowest excess of DIC-C (330 g C m^{-2} $season^{-1}$).

Total input minus output of CH_4-C on a unit area basis were highest in the high C reservoir (14 g C m^{-2} $season^{-1}$), followed by the intermediate C reservoir (1 g C m^{-2} $season^{-1}$), and negative in the low C reservoir (-0.01 g C m^{-2} $season^{-1}$). In the calculation of inputs minus outputs, the remaining DIC-C in the reservoirs was either consumed by algal photosynthesis, or fluxed out the surface of the reservoirs. In the case of CH_4-C, the excess was either consumed by CH_4 oxidation (insignificant in 1999) or fluxed out the surface of the reservoirs.

Very little CO_2-C (<0.00 Kg to 0.03 Kg) and CH_4-C (<0.00 Kg) ebullated from the three reservoirs. When bubbles could be collected, high concentrations of CO_2 (~32000 ppm) and CH_4 (~22000 ppm) were observed, but so few bubbles accumulated in the bubble traps over the 1999 flooded season that bubbles contributed little to the flux from the surface of the reservoirs.

Surface water concentrations of dissolved CO_2 and CH_4

Concentrations of CO_2 in the surface waters (Figure 14) of the reservoirs closely reflected patterns seen in outflow DIC concentrations (Figure 10). The high C reservoir displayed the second highest mean surface water concentration of 258±11.4 μmol CO_2 L^{-1} (Figure 14). The highest mean concentration of dissolved CO_2 (294±9.5 μmol L^{-1}) was measured in the intermediate C reservoir and the lowest mean surface CO_2 concentration (194±9.2 μmol L^{-1}) was measured in the low C reservoir. Surface CO_2 concentrations declined significantly in the intermediate C ($p<0.001$, $r^2=0.55$, Table 7) and low C ($p<0.001$, $r^2=0.54$) reservoirs over the flooded sampling season, while dissolved CO_2 concentrations in the high C reservoir did not ($p>0.05$).

Mean surface CH_4 concentrations were highest in the high C reservoir (1.35±0.09 μmol L^{-1}), followed by the intermediate C reservoir (0.80±0.03 μmol L^{-1}) and the low C reservoir (0.72±0.04 μmol L^{-1}). Surface CH_4 concentrations increased significantly in the high C reservoir ($p<0.001$, $r^2=0.74$, Table 7), the intermediate C reservoir ($p<0.001$, $r^2=0.85$) and the low C reservoir ($p<0.001$, $r^2=0.84$) over the flooded season.

Mean surface dissolved CO_2 concentrations measured in the ELARP wetland reservoir in the first year of flooding were much lower (157±4.1 μmol CO_2 L^{-1}, Kelly *et al.*, 1997) than those we observed in our reservoirs (Figure 14). Mean surface CH_4 concentrations, however, were at least 5 times lower in the upland reservoirs than those measured in the ELARP wetland reservoir (5.6±0.37 μmol CH_4 L^{-1}).

Table 7. Linear regressions of concentrations in surface water to sampling date in the three experimental reservoirs after flooding. NS= not significant

	DIC				CH_4			
	p	r^2	n	df	p	r^2	n	df
Reservoir								
High C	NS	NS	43	41	<0.001	0.738	43	41
Intermediate C	<0.001	0.553	42	40	<0.001	0.846	42	40
Low C	<0.001	0.538	42	40	<0.001	0.838	42	40

Water column and porewater profiles

Concentrations of DIC-C and CH_4 in water at depths between 140 cm to 20 cm above the sediments were similar, indicating the water column was well mixed until 20 cm above the sediments, below which a dramatic increase in concentrations with depth was observed in all reservoirs (Figure 15 and 16). In the high C reservoir, the lowest DIC concentration (306 μmol L^{-1}) was observed at 120 cm above the sediment surface on 23 August 1999 in sipper 1 (Figure 15). The highest mean DIC-C concentration (12400±2160 μmol L^{-1}) was observed at 25 cm below the sediment surface two weeks later. The two sippers located in the high C reservoir yielded similar DIC numbers at all depths throughout the sampling season. DIC concentrations were highest at 25 cm below the sediment surface for both sippers on all sampling dates indicating the highest zone of microbial DIC production occurred deep in the sediments. The intermediate C reservoir had the highest mean DIC concentration (7450±1890 μmol L^{-1}) early in the season (27 July 1999) at 2 cm below the sediment surface, while the lowest mean DIC concentration (317±6 μmol L^{-1}) was measured just prior to drawdown at 80 cm above the sediment

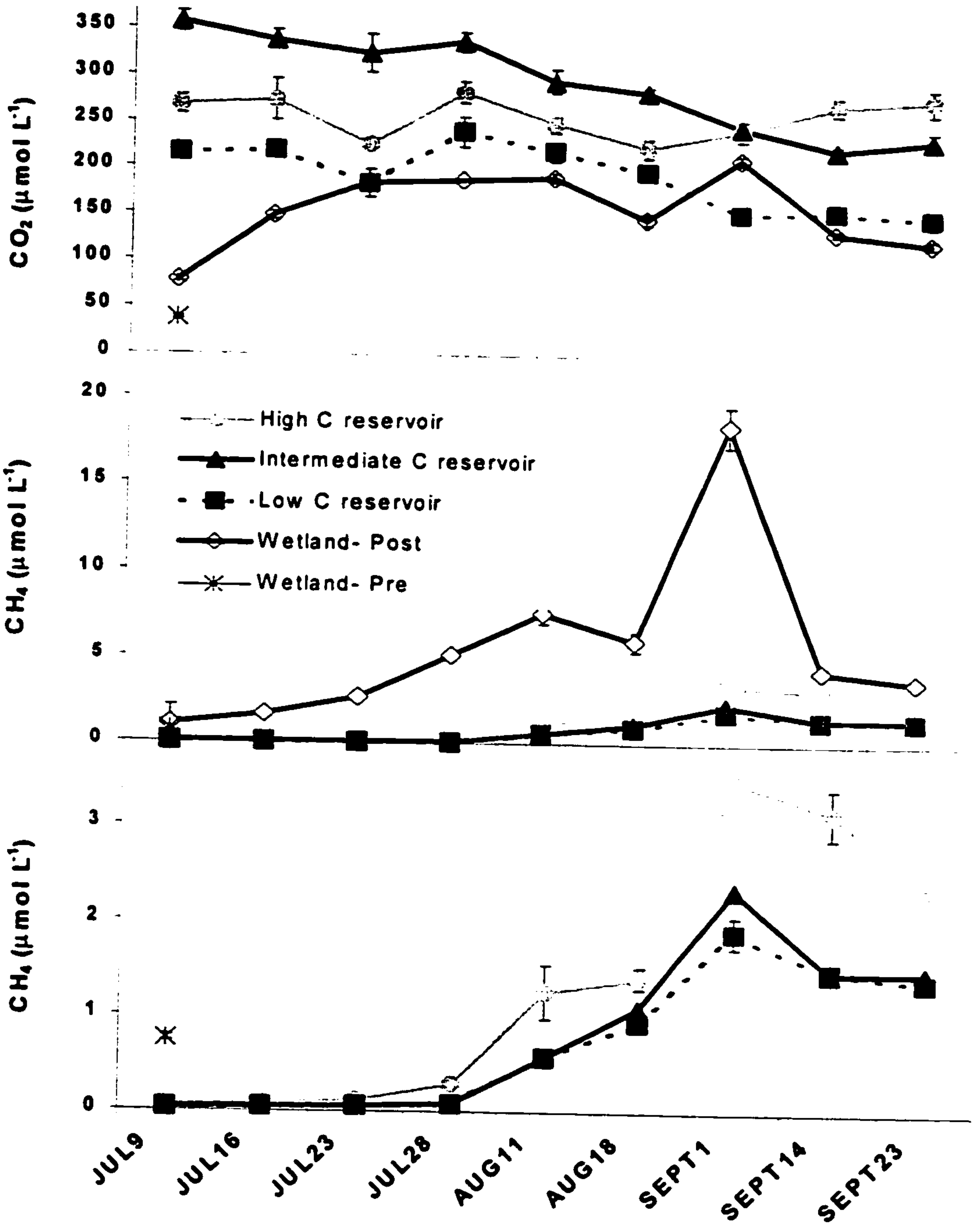


Figure 14. Mean (±std err) surface water CO_2 and CH_4 concentrations (μmol L^{-1}) in the upland and ELARP reservoirs. The bottom graph shows the surface water concentrations of CH_4 in the upland reservoirs only.

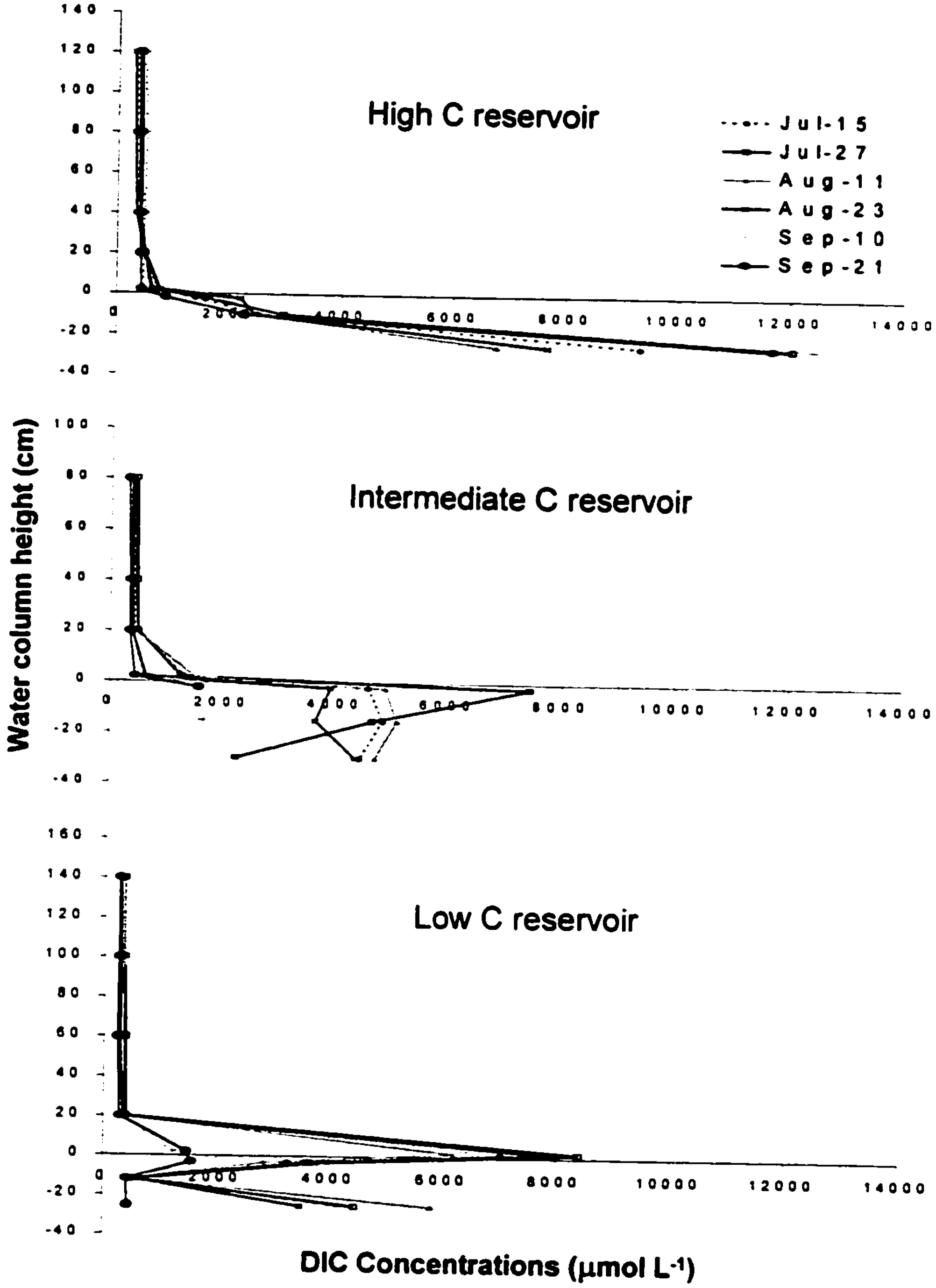


Figure 15. Profiles of mean DIC concentrations (μmol L^{-1}) in the water column and sediment pore water of the upland reservoirs. Standard error bars have not been included so individual profiles can be distinguished from one another.

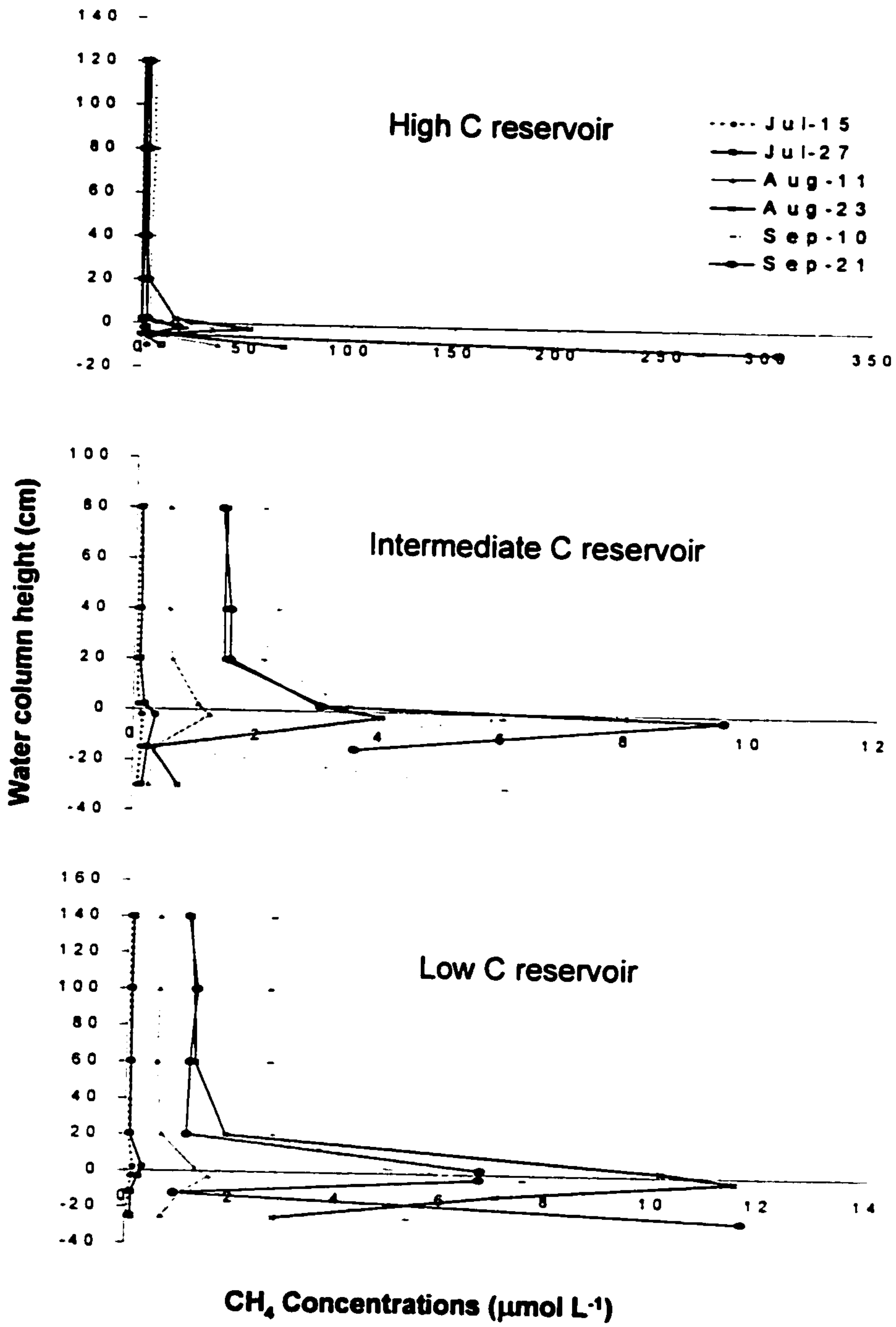


Figure 16. Profiles of mean CH_4 concentrations (μmol L^{-1}) in the water column and sediment pore water of the upland reservoirs. Standard error bars have not been included so individual profiles can be distinguished from one another.

surface. The two sippers located in this reservoir were variable throughout the sampling season, but consistently had the highest concentrations at 2 cm to 5 cm below the sediment surface indicating a more shallow zone of microbial DIC production than that of the high C reservoir. For the low C reservoir, the highest DIC-C concentration of 8390 μmol L^{-1} was observed at 25 cm below the sediment surface early in the season in sipper 3 while the lowest mean concentration of 260±20 μmol L^{-1} DIC-C was measured at 100 cm above the sediment surface just prior to drawdown. The two sippers in this reservoir differed greatly in DIC-C concentrations from one another throughout the season. The highest concentrations of DIC-C at the site 3 sipper were measured at 2 cm above and at 25 cm below the sediment surface indicating two zones of high production, while low DIC concentrations were measured at all depths throughout the season at the site 1 sipper. These two sippers were located in the same vegetative community and were expected have similar DIC profiles.

Mean CH_4 concentrations, like DIC concentrations, were highest in the high C reservoir where the maximum concentration (305 μmol L^{-1} CH_4) was observed at 25 cm below the sediment surface in sipper 1 just prior to drawdown (Figure 16). The lowest mean CH_4 concentration (0.07±0.01 μmol L^{-1}) was observed at 40 cm above the sediment surface on 15 July. Although the two sippers in the high C reservoir were similar in DIC concentrations, they differed dramatically in CH_4 concentrations. The sipper at site 1 (*Pinus/Ledum/Sphagnum* community) had concentrations up to 300 μmol L^{-1} CH_4 at the lowest sediment depths while site 3 (*Pinus/Polytrichum* community) concentrations never exceeded 20 μmol L^{-1} CH_4 at any depth. These sipper data confirm the trend seen in the submerged chamber data (Appendix 2) where all sampling sites located in the moist vegetation community (1, 2 and 4) displayed high rates of CH_4 flux while site 3, located in the drier vegetative community, had much lower fluxes. CH_4 concentrations from sippers in the intermediate and low C reservoirs never exceeded 20 μmol L^{-1} (Figure 13). In the intermediate C reservoir the highest CH_4 sipper concentration (9.6 μmol L^{-1}) was seen at 2 cm below the sediment surface in sipper 3 just prior to drawdown. The lowest mean CH_4 concentration was 0.05±0.00 μmol L^{-1} on July 15 at a depth of 20 cm. Both sippers in the intermediate C reservoir showed similar concentrations that, like the DIC concentrations, were highest at 2 cm to 5 cm below the sediment surface. Mean CH_4 concentrations were highest at 3 cm below the sediment surface (11.5±5.4 μmol L^{-1}) on 24 August 1999 in the low C reservoir and lowest at 25 cm below the sediment surface (0.05±0.01 μmol L^{-1}) on 15 July 1999. For mean CH_4 concentration, the two sippers in this reservoir did not differ as much as mean DIC concentrations. In addition, the highest CH_4 concentrations were observed at 3 cm below the sediment surface as opposed to 25 cm below the sediment surface for DIC concentrations.

DISCUSSION

Fluxes of CO_2 and CH_4 from forest soils prior to flooding

Rates of CO_2 respiration from the three subcatchments (0.6-21 g CO_2 m^{-2} day^{-1}, Appendix 1) prior to flooding were similar to those measured in other northern forest sites (0-27 g CO_2 m^{-2} day^{-1}; Table 9). Most of those studies found a strong positive relationship between soil temperature and rates of CO_2 respiration (Crill, 1991; Lessard *et al.*, 1994; MacDonald *et al.*, 1995; Oberbauer *et al.*, 1992; Raich and Potter, 1995; Ryan *et al.*, 1997; Savage *et al.*, 1997; Schlentner and Van Cleve, 1985; Weber, 1990). Although we observed a significant seasonal decline in rates of CO_2 respiration from June to October, soil temperature did not explain this trend. This may be due to a high degree of variability in rates of CO_2 respiration between sampling sites on a given day.

Table 8. Summary of northern studies on soil CO_2 respiration and CH_4 oxidation rates (g CO_2 or CH_4 m^{-2} day^{-1})

Forest type	CO_2 flux ($g\ m^{-2}\ day^{-1}$)	CH_4 flux ($mg\ m^{-2}\ day^{-1}$)	Reference
This study	1 to 21	0.6 to −2.6	
Boreal		-0.2 to −0.6	Whalen *et al., 1992*
Boreal	0 to 27	0.0 to −2.6	Savage *et al., 1997*
Subarctic		-1.0 to −3.0	Adamsen and King, *1993*
Subarctic	9 to 11		Schlentner and Van Cleve, *1985*
Temperate	2 to 7		Weber, *1990*
Temperate	2 to 15	0.0 to −1.1	Lessard *et al., 1994*
Temperate	0 to 25	0.2 to −4.9	Crill, *1991*
Temperate		-3.7 to −5.3	Castro *et al., 1994*

Soil moisture proved to be poor predictor of rates of CO_2 respiration in this study (Table 3) and others (Crill, 1991; Lessard *et al.*, 1994; Ryan *et al.*, 1997; Savage *et al.*, 1997). Some studies have shown a significant increase in rates of CO_2 respiration with increasing soil moisture (Schlentner and Van Cleve, 1985; Singh and Gupta, 1977) while others found a negative relationship (Douglas and Tedrow, 1959; Raich and Potter, 1995; Weber, 1990). Overall, mean rates of CO_2 respiration were lower in the moist, high C subcatchment than in the drier subcatchments. Others have observed similar trends in boreal and temperate forests (MacDonald *et al.*, 1995; Savage *et al.*, 1997; Weber, 1990). For example, Savage *et al.* (1997) studied a range of boreal forest sites with different moisture regimes and found their highest rates of CO_2 respiration in forest sites of intermediate moisture. Most likely these patterns were observed because high levels of soil moisture reduce soil temperatures as well as diffusion of CO_2 from the soil into the sampling chamber (Crill, 1991; MacDonald *et al.*, 1995).

Both MacDonald *et al.* (1995) and Savage *et al.*, (1997) found that microbial activity, and subsequent CO_2 flux to the atmosphere was significantly related to substrate quality. In general, soils with high organic C content had the highest rates of CO_2 flux. For example, Weber (1990) found that CO_2 flux was higher from soils under aspen (*Populus tremuloides*) stands where the soil litter layer consisted of high nutrient, readily decomposable labile leaves than in jack pine stands where the accumulated litter was lower in nutrients and more recalcitrant. Although the high C subcatchment had the largest amount of accumulated soil organic carbon, the percent organic carbon of litter and FH layers was highest in the intermediate C forest (Table 1) where the highest rates of CO_2 respiration were seen.

Rates of CH_4 oxidation in the subcatchments (0.6-2.6 mg m^{-2} day^{-1}, Appendix 2) also fell within the range measured in other northern forests (0.2-5.3 mg m^{-2} day^{-1}, Table 9). The significant decline in rates of CH_4 oxidation over the season may be explained by the significant effect of soil moisture or temperature. This study showed a significant decrease in CH_4 oxidation with decreasing temperatures. Most field studies show that rates of CH_4 oxidation increase with increasing temperatures up to approximately 10°C (Castro *et al.*, 1995; Crill, 1991; Whalen *et al.*, 1991), but then become independent of temperature (Castro *et al.*, 1995; Crill, 1991; Dunfield *et al.*, 1993; Lessard *et al.*, 1994; Savage *et al.*, 1997; Whalen *et al.*, 1991). Those studies would suggest that in the boreal region, temperatures in the fall, winter, and spring would limit microbial CH_4 oxidation. However, once temperatures increase in spring and methanotrophic communities become established, CH_4 oxidation is most likely driven by concentration gradients and diffusion more than temperature. Since soil temperatures in this study were rarely below 10 °C, the significant decrease in CH_4 oxidation with decreasing temperature suggests that at our sites temperature was not the only factor driving oxidation.

Increases in water table height had a significant negative affect on CH_4 oxidation in this study and others (Castro *et al.*, 1994; Castro *et al.*, 1995; Dunfield *et al.*, 1993; Lessard *et al.*, 1994; Whalen *et al.*, 1991). Crill (1991) and Savage *et al.* (1997) speculated that high water tables not only increase anoxia and favourable conditions for methanogenisis, but that increases in soil moisture fill soil pore spaces and slow diffusion of CH_4 to oxidizing sites. This may explain why the drier intermediate C subcatchment had significantly higher rates of oxidation than the other subcatchments where soils were either thinner or wetter. In the high C subcatchment, for example, CH_4 production was observed in *Sphagnum* communities on two occasions when the water table was high (Appendix 1). CH_4 production was also observed on two occasions in the low C reservoir during wet periods, but only at sampling sites where very thin layers of saturated *Polytrichum* moss grew above the bedrock.

CO_2 fluxes from forest soils postflood

The boreal region is considered a sink for atmospheric carbon (Sellers *et al.*, 1997 and Fan *et al.*, 1998) through the accumulation of organic soils and vegetative biomass on the long term, especially in peatlands. Although we were unable to evaluate net ecosystem exchange, it is likely that since the three boreal subcatchments burned in 1980, they were accumulating carbon prior to flooding. Upon flooding these forest sites quickly changed from being a sink for atmospheric carbon to being a considerable source. The dramatic increase in CO_2 flux from the forest floor immediately following flooding indicated an increase in oxic mineralization of plant, litter, and soil organic carbon. Although temperature and height of the water column may influence rates of carbon mineralization after flooding, the amount and decomposability of the organic carbon present in each reservoir most likely determined how much GHG was produced. We expected the high C reservoir with the largest stores of organic carbon to produce the most CO_2 in the presence of oxygen (Fudge, 2000). However, in the first year of flooding, the intermediate C reservoir had the highest rates of CO_2 production, followed by those in the high C and low C reservoirs. The litter and FH layer of the intermediate C reservoir had the highest percent carbon composition of all the reservoirs and was the only reservoir where labile birch leaves were a significant component of this layer

(Boudreau, 2000). These two factors likely explain why this reservoir had the highest CO_2 production.

Parameters used for predicting rates of decomposition for different types of vegetation are plentiful. For example, several studies have shown that the Klason lignin:nitrogen ratio is a good predictor of decomposition in the first several years of decomposition (Aerts, 1997; Bridgham *et al.*, 1998; Couteaux *et al.*, 1995; Moore, 1984; Moore *et al.*, 1999; Taylor *et al.*, 1989; Trofymow *et al.*, 1995). Jack pine and birch leaf lignin:N ratios were 26 and 33, respectively, and were found to decompose more quickly than species like Douglas fir (*Pseudotsuga menziesii*) that had a lignin:N ratio of 43 (Moore *et al.*, 1999; Trofymow *et al.*, 1995). No lignin:N ratios are available for other species in this study. However, Boudreau (2000) conducted a laboratory decomposition experiment with foliage from individual plant species found in the reservoirs. Boudreau (2000) determined that the leaves of birch, jack pine, and Labrador tea decomposed more rapidly and in that order, than blueberry, *Polytrichum*, and *Cladina* based on rate of DIC production in oxic incubation containers over 110 hours.

The presence of an abundant layer of birch and jack pine litter in the intermediate C reservoir likely explains why it not only had the highest sediment fluxes of CO_2, but was significantly different from the low C reservoir where the forest floor consisted of less labile organic carbon in blueberry, *Polytrichum*, and *Cladina* vegetation. The highest *amount* of total carbon stored in the litter and FH layers occurred in the high C reservoir, however, the highest *percent* carbon was found in the litter and FH layers of the intermediate C reservoir (Table 1). Quality of organic matter, however, does not explain within reservoir variation in sediment fluxes of CO_2 in the intermediate and low C reservoirs (Appendix 3). For example, the two vegetative communities in the low C reservoir did not have consistently different sediment CO_2 fluxes, even though we expected the *Polytrichum*/*Cladina* community to be more recalcitrant than the *Pinus*/ *Vaccinium* community. Site-specific parameters such as organic matter available for mineralization, the composition of the microbial community, and the availability of oxygen may explain within-reservoir variations in sediment CO_2 flux. In the high C reservoir, high CO_2 fluxes observed at the three *Pinus*/*Ledum*/*Sphagnum* sites suggest that the flooded organic carbon was more easily mineralized than at the *Pinus*/*Polytrichum* site where CO_2 fluxes where consistently lower. Unlike in the high C reservoir, sediment

CO_2 production in both the intermediate and low C reservoirs was high immediately following flooding and steadily declined over the season (Figure 8). This suggests that the most labile material decomposed very rapidly in these two reservoirs. The steady increase in CO_2 production over the season in the high C reservoir is most likely the result of more recalcitrant *Sphagnum* (Clymo and Hayward, 1982) and *Polytrichum* moss (Russell, 1990) being present there. Once these materials began to decompose, rates of CO_2 flux remained high until the reservoir was drawn down in the fall.

Postflood soil fluxes of CO_2 relative to other reservoir studies

CO_2 fluxes measured on the bottom of hydroelectric reservoirs in northern Quebec were approximately 3.5 times lower than the mean of 6500 mg CO_2-C m^{-2} day^{-1} seen in this study (Duchemin *et al.*, 1995). Maximum CO_2 fluxes from flooded peatland and forest soils were 1800 mg CO_2-C m^{-2} day^{-1} in the Quebec reservoirs (Duchemin *et al.*, 1995). Rates of CO_2 flux measured on the bottom of a 23 year-old hydroelectric reservoir did not differ from those measured on the bottom of a 2 year-old reservoir, but lower fluxes were measured over flooded forest soils than over flooded peat. Rates of CO_2 flux from inundated peat in the ELARP study (maximum rate in 1995 of 2200 mg CO_2-C m^{-2} day^{-1}; Karen Scott, unpublished data) were very similar to those measured in the Quebec reservoirs. Although rates of CO_2 flux were much higher in this study than others, there was ample labile organic carbon in the flooded vegetation and litter to support these rates of mineralization. For example, the 12000, 6600 and 3000 Kg C that was available in labile vegetation and litter in the high, intermediate and low carbon reservoirs, respectively (Boudreau, 2000) supported total sediment fluxes to the bottom water of 5000, 4000 and 2400 Kg C $season^{-1}$, respectively (Table 7). These fluxes suggest that labile carbon from the litter and FH layer available for mineralization may be low in the second year of flooding, leaving behind the more recalcitrant carbon stored in flooded soils and vegetation to decompose over the long term. However, the sediment fluxes of CO_2 measured in the first year of flooding may also be high because the flooded jack pines, with most of their canopies above water, remained alive, but stressed, during the first season of flooding. Most likely these stressed trees were respiring and emitting CO_2 through their submerged roots. Jack pine, an upland conifer without a well-developed root parenchyma layer, is not well adapted to flooding and the products of fermentation eventually kill the roots (Janusz Zwiazek and Gregory Taylor, personal communication). Toward the end of the 1999 flooded season the needles were beginning to yellow, but not fall off, possibly partially explaining the overall decline in sediment fluxes throughout the flooding period.

Measurements of CH_4 fluxes from forest soil postflood

The forest floor changed from a sink to a significant source of atmospheric CH_4 once the subcatchments were flooded. CH_4 production did not begin until early August, however, when concentrations of O_2 at the lower depths in the reservoirs declined below 1 mg/L (Fudge, 2000). Because methanogens are strict anaerobes (Williams and Crawford, 1985), both O_2 concentration and substrate quality are most likely more important predictors of CH_4 production than temperature and height of the overlying water column. In fact, the significant, counterintuitive, negative relationship we observed between CH_4 flux and temperature (Table 3) is most likely only a covariance between decreasing reservoir temperatures and decreasing O_2 concentrations, because decreasing temperatures should slow rates of decomposition and CH_4 flux (Cicerone and Oremland, 1988). Depleted CH_4 isotopic $\delta^{13}C$ signatures of -65 to -77 ‰ in the submerged chambers indicated the dominant methanogenic pathway is CO_2 reduction (Boudreau, 2000, Aravena, 1993; Oremland, 1987; Landsdown, 1992). Enriched $\delta^{13}C$ signatures between approximately -50 to -65 ‰ would have indicated that the dominant methanogenic pathway was acetate fermentation (Bellisario *et al.*, 1999; Blair and Carter, 1992; Burke *et al.*, 1988; Oremland *et al.*, 1987; Sugimoto and Wada, 1993).

Sediment CH_4 and CO_2 flux in the high C reservoir increased throughout the season. High rates of CH_4 production in this reservoir relative to the other two reservoirs may be the result of large concentrations of dissolved CO_2 available for reduction to CH_4 at the time the anoxic methanogenic population became established (Figure 8). In the high C reservoir, low rates of CH_4 flux from the *Pinus/Polytrichum* sampling site relative to the *Pinus/Ledum/Sphagnum* samplings sites may indicate a methanogenic community was already established at the *Pinus/Ledum/Sphagnum* site (Appendix 4). The sipper profiles in the high C reservoir also show much higher bottom and sediment pore-water CH_4 concentrations in the *Pinus/Ledum/Sphagnum* community than in the *Pinus/Polytrichum* community. In both the intermediate and low C reservoirs, it appears that anaerobic decomposition (and subsequent CH_4 flux from sediments) became established as aerobic decomposition declined (as measured by the CO_2 flux). CH_4 fluxes at specific sampling sites were variable within the high and low C reservoirs (Appendix 4) and were likely dependant on a combination of factors such as oxygen concentration, type of methanogenic community present, and type of substrate available.

Postflood soil fluxes of CH_4 in the experimental reservoirs relative to other reservoirs

The postflood sediment CH_4 fluxes measured in the FLUDEX reservoirs agreed very well with those measured in other studies. In a 2 year-old Quebec reservoir, 15 and 260 mg CH_4-C m^{-2} day^{-1} were measured fluxing from flooded forest and peatland soils, respectively (Duchemin *et al.*, 1995). Submerged peat surfaces in the ELARP study produced on average 6 to 164 mg CH_4-C m^{-2} day^{-1} (Scott *et al.*, 1999). Mean sediment CH_4 fluxes in the FLUDEX reservoirs ranged from 3 to 110 mg CH_4-C m^{-2} day^{-1} (Figure 8) with the highest flux occurring over the flooded *Sphagnum spp.* sites in the high C reservoir. These results indicate that factors controlling methanogenesis are similar in the three studies and that a inundated *Sphagnum spp.* peat community produces more CH_4 than do soils without this vegetation, possibly due to the presence of already established methanogenic communities in the saturated areas where *Sphagnum spp.* grew prior to flooding.

Carbon Mass Balance Budget: DIC

Dissolved concentrations of gases in surface waters and windspeeds measured above the water surface are often used to estimate a gas flux from a water body to the atmosphere. At low windspeeds, however, gas exchange across the air-water interface is largely independent of windspeed (Duchemin *et al.*, 1999; Liss and Slater, 1974; Sellers *et al.*, 1995; Wanninkhof, 1992; Wanninkhof *et al.*, 1987). Windspeeds above the water surface at the FLUDEX reservoirs were below 1 m s^{-1} (St. Louis *et al.*, unpublished data) and therefore could not be used to calculate a flux to the atmosphere. Without a known gas exchange coefficient for the reservoirs, flux out the surface of the reservoirs could not be calculated. We therefore originally attempted to estimate fluxes of CO_2 and CH_4 to the atmosphere using a mass balance approach. However, as described below, uncertainties in the carbon budgets of the reservoirs prevented this.

Inputs of C from reservoir sediments were much higher (minimum of 94 % of total inputs) than all other C inputs to the reservoirs, showing that mineralization of flooded organic material dominated the carbon budget of the reservoirs. This large sediment flux resulted in a surplus of DIC available for either uptake by photosynthesis or flux out the reservoir surfaces to the atmosphere. Unfortunately, neither net primary production nor surface flux of CO_2 was quantified in the first year of flooding. Large epipelic mats (communities of algae, bacteria and fungi growing on the submerged branches and trunks of vegetation) and epilithic (attached to rock) communities were observed growing throughout the intermediate and low C reservoirs. Less algal growth was observed in the high C reservoir, but enough was present to indicate that primary production could be an important component of the C budget. To estimate the maximum potential primary production in the reservoirs, we used rates of primary production measured from Lake 375 at the ELA (Turner *et al.*, 1994). Although Lake 375 had some of the highest rates of net primary production (maximum 1162 μmol C m^{-2} hr^{-1}) measured at the ELA, rates there most likely still underestimated that observed in the reservoirs in the first year after flooding due to the larger area available for community colonisation. The bottom of Lake 375 is a relatively planar surface, whereas surfaces available for epipelic and epilithic growth on the submerged vegetation and rocks in the reservoirs could be up to 10 times greater than that in a normal lake (Michael Turner, personal communication). Using the primary production rates from Lake 375, and an approximate surface area ten times greater for epipelic or epilithic production, we estimate that the attached algal community could consume up to 2400 mg C m^{-2} day^{-1} or 230 g C m^{-2} for the whole period the reservoirs were flooded. Most likely, however, all the C taken up by these communities was lost back to the atmosphere due to decomposition of this highly labile material following drawdown of the reservoirs in the autumn and before reflooding the following late spring.

All three FLUDEX reservoirs in the first year of flooding had higher surface water CO_2 concentrations than those seen in the first two years of flooding at the ELARP reservoir (Figure 11). In fact, surface water CO_2 concentrations in the intermediate C reservoir were twice as high as those in ELARP. Unfortunately we do not have actual surface flux measurements. However, if we account for the estimated net primary production that we calculated above in our budgets, the amount of carbon remaining to flux from the surface of the reservoirs to the atmosphere during the flooded period was 440 g C m^{-2} (4500 mg CO_2-C m^{-2} day^{-1}) in the high C reservoir, 503 g C m^{-2} (5200 mg CO_2-C m^{-2} day^{-1}) in the intermediate C reservoir and -67 g C m^{-2} $season^{-1}$ (-730 mg CO_2-C m^{-2} day^{-1}) in the low C reservoir (Table 10). Potential surface CO_2 fluxes from the FLUDEX reservoirs (excluding the low C reservoir) for a 92-98 day flooded period were up to 2.3 times higher than fluxes measured from the ELARP reservoir for a 220-day season (Table 10), possibly due to the large amount of easily decomposable leaf litter in the FLUDEX reservoirs compared with the more recalcitrant carbon in peat in the ELARP reservoir.

Table 9. A comparison of estimated FLUDEX surface GHG flux (for a 92-98 day season) to ELARP surface GHG fluxes (for a 220-day season).

Reservoir	**High C**		**Inter C**		**Low C**		**ELARP**	
g C m^{-2} $season^{-1}$	**CO_2**	**CH_4**	**CO_2**	**CH_4**	**CO_2**	**CH_4**	**CO_2**	**CH_4**
Budget summary	675	13.55	735	0.76	330	-0.05		
Primary production	235		232		220			
Est. surface flux	440	13.55	503	0.76	110	-0.05	105-220	7.9-15

Carbon Mass Balance Budget: CH_4-C

The excess CH_4-C to the atmosphere was relatively small compared to the excess DIC. As with DIC, sediment CH_4 flux was the largest input (minimum 96% of total inputs, Table 7) of CH_4 to the reservoirs. This sediment input overwhelmed the whole CH_4-C budget in the high and intermediate C reservoirs. However, seepage in the low C reservoir was so large relative to the other reservoirs (60% of incoming water seeped out) that no surplus CH_4 remained despite high inputs from decomposition. Excess CH_4 within the reservoirs was lost either to methane oxidation or out reservoir surfaces to the atmosphere. CH_4 isotope $\delta^{13}C$ data show that CH_4 signatures in outflow water (-64 $^{0}/_{00}$ to -89 $^{0}/_{00}$) and submerged chamber water (-65 $^{0}/_{00}$ to -77 $^{0}/_{00}$) were very depleted (Boudreau, 2000). This indicates a microbial production of CH_4 at the bottom, but little or no enrichment by microbial CH_4 oxidation as water leaves the reservoirs in outflow. This implies that all CH_4 produced on the bottom eventually made its way to the atmosphere, either through reservoir surface flux or out the outflow or in seepage. Calculated CH_4-C flux from the surface to the atmosphere in the high C reservoir (13.55

g C m^{-2} $season^{-1}$, Table 10) is close to the upper range seen in ELARP (7.9-15 g C m^{-2} $season^{-1}$) even though the surface water CH_4 concentrations in the high C reservoir were five times lower than observed in ELARP (Figure 11). Boudreau (2000) did see some enrichment of $\delta^{13}C$ in this reservoir so methane oxidation may explain the discrepancy between sediment CH_4 flux and CH_4 surface concentrations in the high C reservoir. In the intermediate C reservoir, the low calculated surface CH_4 flux (0.76 g C m^{-2} $season^{-1}$) corresponded well with the low concentrations of CH_4 measured in surface waters.

CONCLUSION

The FLUDEX and ELARP experiments are the only whole-ecosystem experiments that have measured the net impact of reservoir creation by examining GHG fluxes from ecosystems both prior to and following flooding. Although net ecosystem exchange measurements were not feasible for CO_2 in FLUDEX prior to flooding, rates of dark soil and root CO_2 respiration were found to be characteristic of rates measured at other boreal forest sites. Furthermore, because the forests flooded were approximately 19-year-old fire regenerated forests, they were most likely acting as sinks for CO_2 by accumulating vegetative biomass and organic soil at the time of flooding (Schlesinger, 1997). These forest sites were also sinks for CH_4 at rates similar to those seen in other boreal forest studies. After flooding a marked increase in rates of flux from flooded soils was observed. Although these rates were higher than those observed in two other studies (Duchemin *et al.*, 1995, Scott, 1999), the amount of organic carbon stored in labile vegetative biomass and litter in each reservoir easily supported the rates of flux observed in the first year of flooding. It appears that the organic carbon stored in the intermediate C reservoir, although less in quantity than that in the high C reservoir, may have been more labile, leading to higher rates of sediment CO_2 flux there. CH_4 fluxes were far higher in the *Pinus/Ledum/Sphagnum* community of the high C reservoir than in the communities of any other reservoir, suggesting that a methanogenic community already existed in these saturated *Sphagnum spp.* areas prior to flooding.

Although budget calculations allowed us to compare sources and sinks of GHGs in the reservoirs, direct comparison of FLUDEX GHG surface fluxes to those of other studies was difficult because we were unable to balance the budgets. For both DIC-C and CH_4-C budgets, losses of these GHGs in seepage made the estimate of surface flux difficult. Furthermore, losses of CO_2 to algal photosynthesis could only be estimated, making calculations of CO_2 flux from the surface uncertain. A gas exchange coefficient is currently being calculated using a sulfur hexafluoride (SF-6) tracer gas that will eventually allow back calculations of a surface flux from the FLUDEX reservoirs using surface concentrations measured in 1999. However, measurements taken on the FLUDEX reservoirs in the first year of flooding clearly show that mineralization of the labile organic litter in flooded soils and vegetation to CO_2, and to a lesser extent CH_4, is rapid and that, at least in the first year of flooding, it exceeds rates observed in ELARP.

In determining whether upland boreal forest should be considered an alternative to low-lying wetland ecosystems we must consider both the productivity of the FLUDEX sites and length of time they could remain productive. It is likely that most of the labile material in the FLUDEX reservoirs has decomposed within the first year of flooding and that rates of GHG production will decline in the following years to rates much lower than those seen in ELARP (which remain high to present). One must also consider the global warming potential of CH_4 (25 times that of CO_2) and that rates of CH_4 production were much lower in the FLUDEX sites than in the ELARP reservoir despite the availability of labile organic carbon for decomposition. In the long term, upland forest may be a better choice for reservoir development than wetland ecosystems, but more data must be collected on the FLUDEX reservoirs to be certain.

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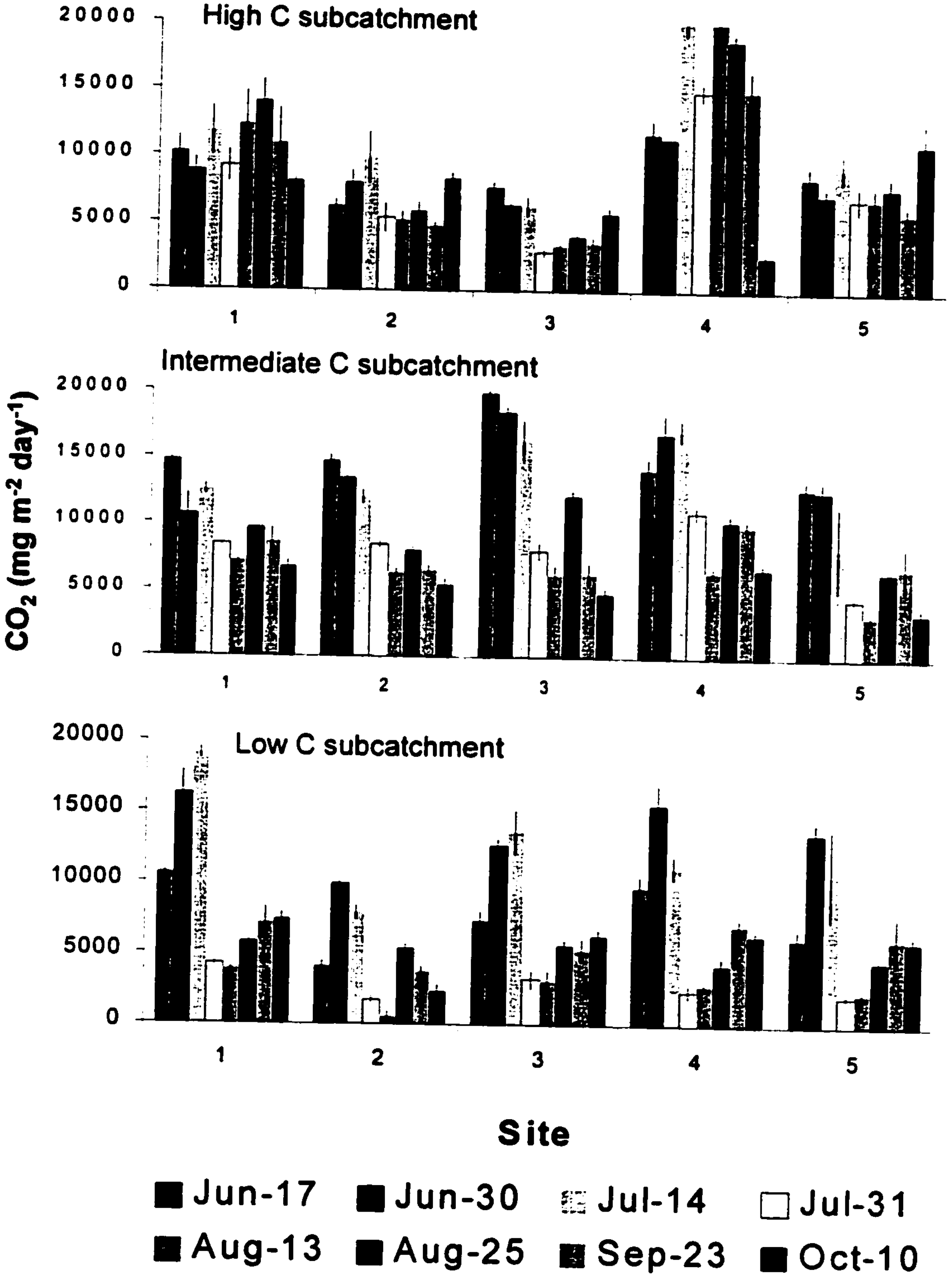
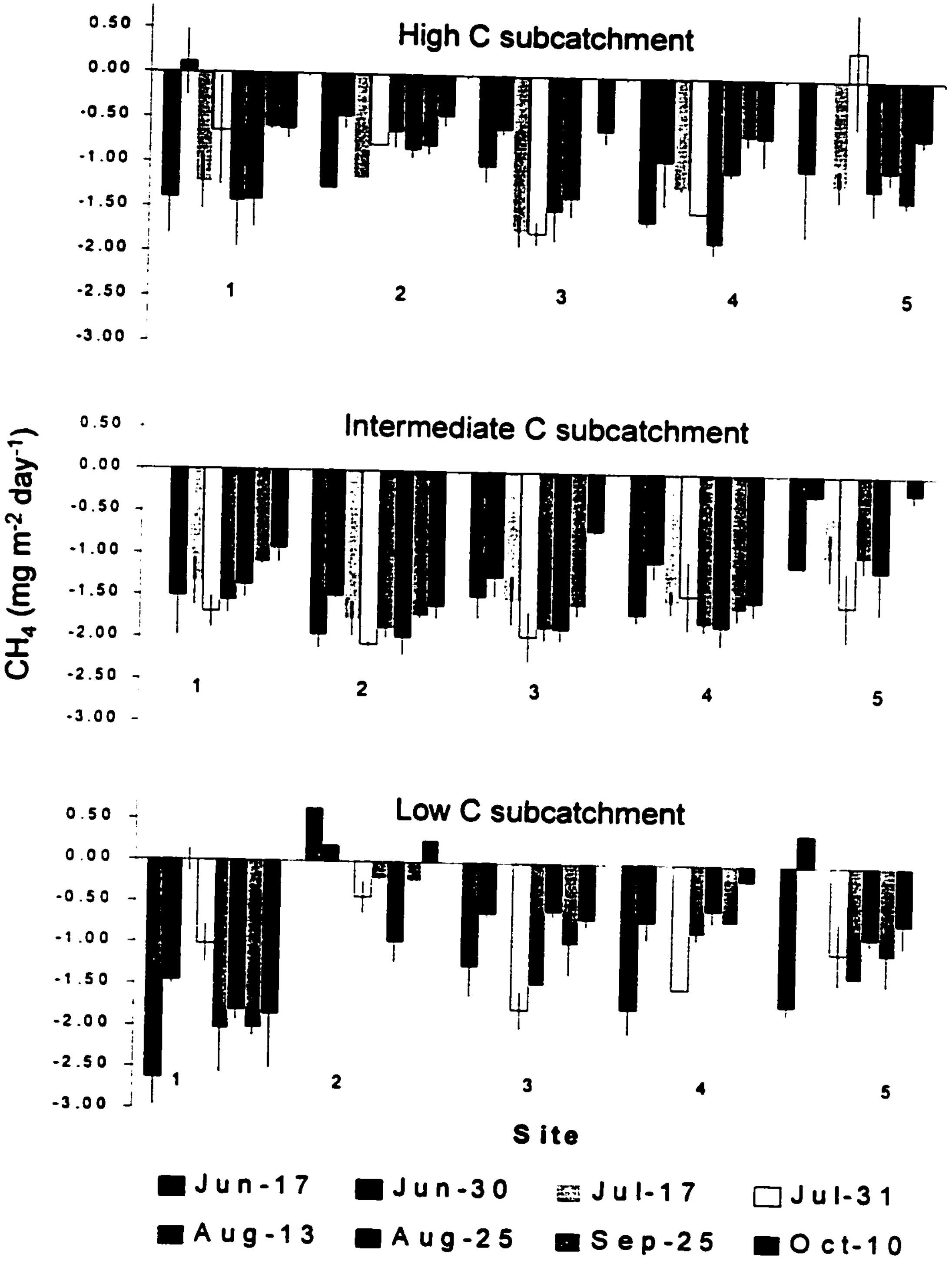
APPENDICES

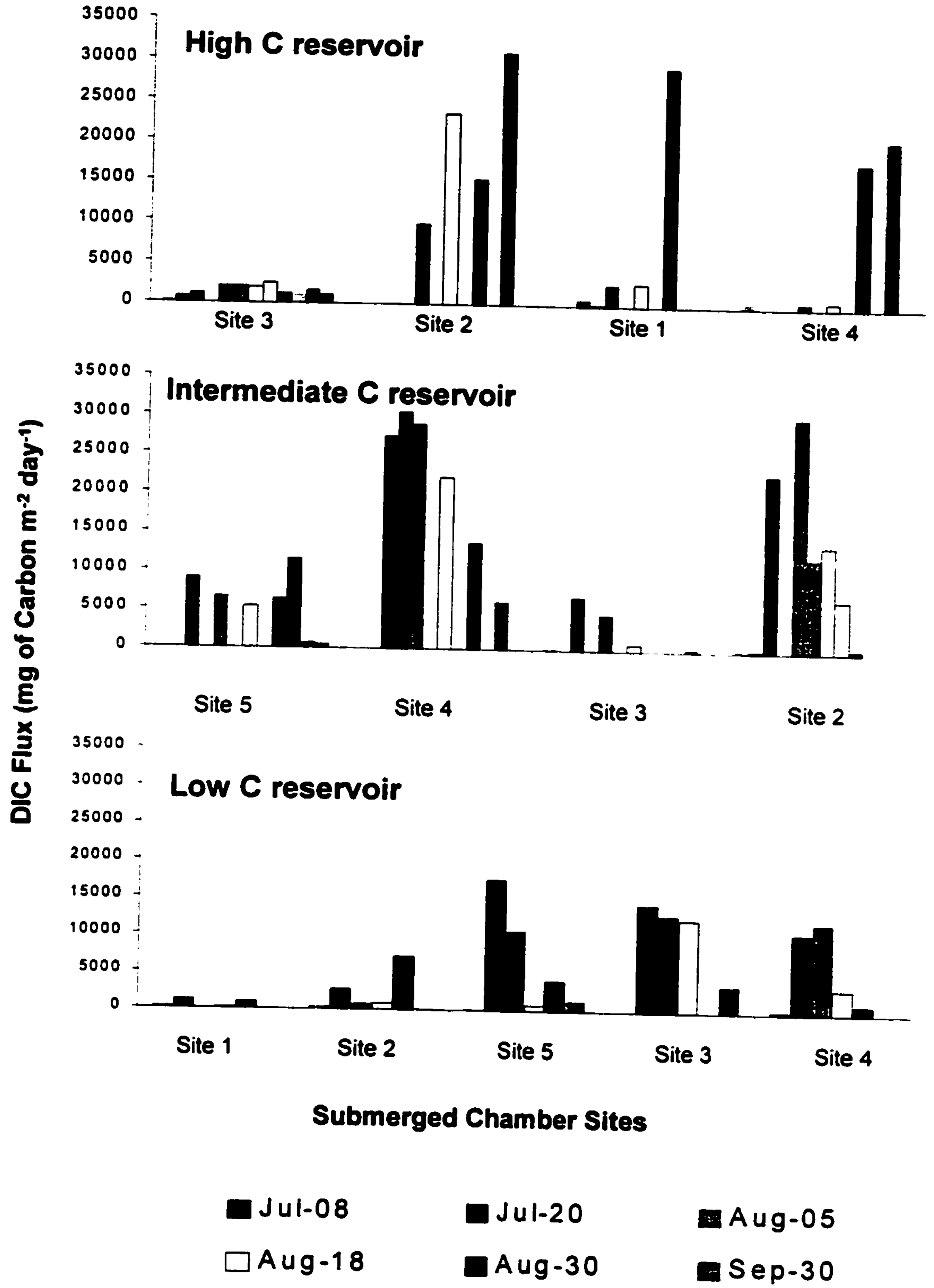
Appendix 1. Mean (±std err) CO_2 flux (mg m^{-2} day^{-1}) from the sampling sites of three boreal subcatchments before flooding.



Appendix 2. Mean (±std err) CH_4 flux (mg m^{-2} day^{-1}) from the sampling sites of three boreal subcatchments before flooding.



Appendix 3. Sediment DIC-C flux (mg m^{-2} day^{-1}) from the sampling sites of three FLUDEX reservoirs after flooding.



Appendix 4. Sediment CH_4-C flux (mg m^{-2} day^{-1}) from the sampling sites of three FLUDEX reservoirs after flooding.