2010

Carbon Dioxide Exchange in a Permafrost-Dominated Boreal Wetland in the Northwest Territories, Canada

Andrea J. Kenward
Wilfrid Laurier University

Your file Votre référence
ISBN: 978-0-494-68712-3
Our file Notre référence
ISBN: 978-0-494-68712-3

Carbon Dioxide Exchange in a Permafrost-Dominated Boreal Wetland in the Northwest Territories, Canada

By

Andrea J. Kenward
B.Sc. University of Victoria, 2005

THESIS
Submitted to the Department of Geography and Environmental Studies
in partial fulfillment of the requirements for
the Master of Science Degree

Wilfrid Laurier University, 2010

Abstract

Northern boreal wetland complexes are substantial reservoirs for carbon and play a crucial role in both regional and global carbon budgets but they are showing significant signs of impact by climate change. This study examined the carbon dioxide flux of a high boreal wetland during the snowmelt and growing season of 2008 in Scotty Creek Basin, located near Fort Simpson (61°18'N, 121° 18'W), Northwest Territories. This basin is not only responding to shifts in atmospheric temperatures, but it is also under additional pressure from increasing permafrost degradation. A dynamic closed-system chamber was used to monitor and quantify mid-day total respiration (R_{tot}), gross ecosystem production (GEP), and net ecosystem exchange (NEE) at nine sites, in order to characterize and compare the gas flux gradients for three landscape units typical of the lower Liard River valley (channel fens, ombrotrophic flat bogs and peat plateaus).

Each landscape unit exhibited increasing rates of R_{tot} and GEP for the duration of study. Instantaneous rates of R_{tot} and NEE were highest in the permafrost plateau and channel fen, while the flat bog remained consistently low throughout the season. While there was significant variation in magnitude, the results demonstrated relatively similar temporal variability between landscapes. Temporal and spatial variability in CO_2 exchange was further examined through the relationships with local environmental conditions: photosynthetically active radiation, air temperature, soil temperature, soil moisture, and frost table and water table depth. Light response curves derived using an exponential model showed GEP was primarily driven by photosynthetically active radiation, yet significant scatter suggested additional environmental influences. Differential development in R_{tot} appeared to be most influenced by temperature and moisture regimes. Ambient air temperature, and soil and water temperatures at 20 cm all showed strong positive correlations with R_{tot}, while decreasing frost and water table depth, and soil moisture enhanced R_{tot}.

These relationships for the 2008 season were used with assistance from meteorological stations to develop a continuous dataset for this region. In addition, remote sensing technology was used to scale the continuous dataset to the ecosystem level. Results showed that while the individual channel fen examined was the greatest emitter of CO_2 into the atmosphere, it was the permafrost plateau that had the greatest total flux over a larger area. The potential future regional flux for this region as a sink or source for CO_2 was also examined through site specific instantaneous gas flux and a simplified continuous model. This study highlights the need for long term measurement in order to develop an annual budget for CO_2 and capture a more complete carbon profile of permafrost-dominated boreal wetlands. Further study will also result in a more holistic understanding of how CO_2 gas flux gradients vary between the three distinct landscape units and periods of climatic variability. As the climate in northern ecosystems continues to alter, understanding the interactions between the physical, biochemical, and environmental conditions of different landscapes and the processes which define them can aid in the parameterization and interpretation of current and future climate and biogeochemical models.

Acknowledgements

During my time at Laurier I have accumulated a long list of people who have played an invaluable role in my graduate experience. First and foremost I must thank my supervisors, Bill Quinton and Rich Petrone, the gurus of Scotty Creek and CO_2. They not only showed great patience while enduring my many questions as I pieced it all together but they were always there with encouragement, guidance, and advice that kept me on track and taught me so much. I would also like to thank my other committee members, Laura Chasmer and Merrin Macrae, for their insightful comments and ideas.

This opportunity has allowed me to challenge myself in new ways and explore an area of the world that not many people get the chance to see. Through my experiences I have not only discovered my inability to float while traversing wetlands (much to my assistant's amusement), but a love for snowmobiling across frozen tundra (from which Eric may still habour bruises). I thank Mike English for this latter discovery, allowing me to tag along on another trip of a lifetime. In addition to all the adventures associated with my research, I have also discovered a love of teaching. As the daughter of two teachers, I never thought I would admit to such a thing, but I must confess I have enjoyed the challenge and pleasure of helping students.

Along the way my technical support, Trevor and Scott, always found a way to clear up my confusions and help me out. Tyler, my field assistant extraordinaire, managed to keep me sane and laughing and the whole Scotty Creek experience wouldn't have been the same without him. I only hope he's as lucky as I was to find a partner in crime as he embarks on his own graduate and Northern adventure.

I want to say a special thank you to Dan Smith. During my 1st year at UVIC he was a guest lecturer in my geography elective, and inspired a budding biologist to switch majors. As my undergraduate career veered into the world of physical geography he helped me discover the glaciers of Jasper National Park, took me on as a lab assistant, and helped me get my first job after university. Dan has played a huge role in my academic career and I sincerely thank him for everything he has taught me.

To my friends and fellow scholars, I am sure there would be much disagreement as to who procrastinated more, but an overall agreement to the fact that it was worth the memories. I have thoroughly enjoyed the last few years and have gained lifelong friends. During the tough times they were always handy for support and advice, while the many highs have been filled with potlucks, office chats, patios and most of all a lot of laughing.

Finally and most importantly, I thank my family for supporting me on this adventure and of course waiting until after I lived in a tent for three months to tell me I was crazy. They have taught me the meaning of hard work, perseverance, and to believe that I could do anything if I set my mind to it. They were the first to show me that I could love both science and art. They are my inspiration, my support, and my home.

Table of Contents

List of Figures

List of Tables

Chapter 1 Introduction

1.1 Boreal Wetland Ecosystems

As climate change continues to be the dominant environmental issue, the present and future condition of Northern environments is a growing concern with public, political and scientific communities. As one of the most sensitive environments in the world it is not only affected by, but also affects global climate change (ACIA, 2004). However, there is still much that is not understood as a result of remote locations and extreme climates that have historically limited the development of intensive long term study in the North (ACIA, 2004; Elberling, 2007; IPCC, 2007; Tarnocai *et al.*, 2007). High latitude regions are defined by considerable spatial and temporal variability in climate, resulting in an assortment of regional climates that exhibit different ecological and physical climatic characteristics and responses (ACIA, 2004). To develop quantitative models that accurately represent these ecosystems, it is important to understand the physical processes that define them.

A wetland is considered the transition between land and water that is neither firm land nor open water (NWWG, 1988). These areas are waterlogged either most or all of the time with a fluctuating water table (NWWG, 1988), dominated by "hydric soils, hydrophobic vegetation, and biological activity" (Tarnocai, 1980) that are adapted to the presence of excess water. Waterlogged conditions often promote greater rates of plant production than decomposition, resulting in peat accumulation (NWWG, 1988). Peat is the product of the incomplete decomposition of plant material in water-saturated conditions (NWWG, 1988). Peatlands are organic wetlands that have accumulations of peat greater than 40 cm in depth (NWWG, 1988). Approximately 97% of Canadian peatlands occur in the boreal wetland and subarctic wetland regions (Tarnocai, 2006).

In Canada wetlands are estimated to cover approximately 14% of the total land mass, the distribution of which is a result of local climate and surficial geology (NWWG, 1988). Wetland and peatland ecosystems extend from temperate and boreal into subarctic regions, distinguished by variations in canopy and vegetation coverage, hydrology, wetland features, and permafrost distribution (Kuhry and Turunen, 2006). The development of peatlands is also distinguished by latitude as temperate peatlands have experienced higher rates of peat accumulation in comparison to those in the north and subarctic (Kuhry and Turunen, 2006). Temperate regions experience warmer climates, while boreal and sub-arctic wetlands experience cold winters with wet and limited growing seasons (NWWG, 1988). As a result, the boreal and subarctic regions currently experience a greater sensitivity to shifts in temperature, precipitation and permafrost loss (Kuhry and Turunen, 2006).

Boreal wetland regions, in particular, extend across the Canadian landscape from coast to coast. The ecological diversity encompassed by this distribution has resulted in four regions: high, mid-, and low and Atlantic boreal wetlands (NWWG, 1988). The continental high boreal wetlands situated in Northwestern Canada lie within the southern boundaries of the Taiga Plains and in the centre of the Mackenzie River Basin, Canada's largest river basin (Natural Resources Canada, 2009). Saturated conditions in a flat, cold northern ecosystem result in poor drainage with substantial deposits of peat and a wetland composed of three landscape units: permafrost plateaus, flat bogs, and channel fens (NWWG, 1988; Quinton *et al.*, 2003) (Figure 1-1).

Permafrost plateaus rise ~ 1-2 m above the surrounding wetlands as forested islands underlain by permafrost (Quinton *et al.*, 2003; Robinson and Moore, 2000). It is believed that permafrost plateaus evolved from bogs (Robinson and Moore, 2000), their expansion and elevation the result of peat accumulation over time (Geological Survey of Canada, 1998) and uplift due to the expansion of water as it freezes (Harris and Schmidt, 1994). As the peat elevates above the bog water level it is exposed to colder temperatures, surface wind, and greater drainage. These conditions result in an increased penetration of winter frost initiating permafrost growth. Moreover, the relatively drier peat acts as an insulating barrier allowing further permafrost growth. The development of vegetation and tree species sustain the presence of permafrost by decreasing the amount of incident solar radiation reaching the ground surface. The water table of permafrost plateaus fall 0.5 m or more below the ground surface during the annual thaw and drainage of the active layer (Quinton *et al.*, 2003). As a result, they are not wetlands by definition (Quinton *et al.*, 2003).

Permafrost plateaus direct runoff into adjacent bogs and fens, maintaining the current hydrological drainage network (Quinton *et al.*, 2003). Plateaus vary in size, form, and stage of development (ranging from youthful, to mature, to old). Initially in the development of the permafrost plateau there is little diversity in the distribution of vegetation. However, with maturity the vegetative species common to plateau environments begin to establish themselves (Geological Survey of Canada, 1998). Typically, permafrost plateaus support a tree canopy of *Picea mariana* (black spruce), and a ground cover dominated by lichen, feather and *Sphagnum* mosses, Labrador tea, and ericaceous shrub species (NWWG, 1988). Plateaus with a dense canopy, often have a well decomposed surface peat with high nutrient concentrations, while an open-canopy is often more fibric at the surface with lower nutrient concentrations (NWWG, 1988). Permafrost plateaus will eventually reach a maximum stage of development as limited by the local conditions and at which point permafrost will begin to slowly degrade. Thawing of the permafrost causes ground surface subsidence and local flooding (Robinson and Moore, 2000). This degradation and disappearance of plateaus will result in canopy loss and a shift in vegetation dominance as the sub-surface becomes unstable for tree roots and saturated.

Flat bogs are ombrotrophic, low-lying features that appear as patches on the landscape. They are broad and poorly defined (NWWG, 1988; Quinton *et al.*, 2003). Most flat bogs are isolated in terms of surface and near-surface flow interactions, and are surrounded on all sides by raised plateau. Hydrologically-isolated by the surrounding elevated permafrost they receive water from precipitation and sub-surface runoff from the plateaus and experience loss through evaporation and possibly groundwater recharge (Quinton *et al.*, 2009). Hydrologically-connected bogs are those that are connected during periods of high water supply to channel fens by surface and near-surface flows (Quinton *et al.*, 2009). The water table remains close to the surface throughout the year in bogs. Stunted trees, typically *Picea mariana*, can be found near the edges with plateaus (NWWG, 1988); while the surface is often dominated by *Sphagnum* mosses (NWWG, 1988). The underlying peat is generally composed of fibric mosses and ericaceous leaves at the surface, which overlies a mesic layer of moderately decomposed material, and lastly a well-decomposed basal layer with residues of sedges and tree wood.

Channel fens are minerotrophic, low-lying features that appear as large linear features on the landscape along the drainage network of a basin (NWWG, 1988; Quinton *et al.*, 2003), with broad (50 to > 100 m wide) channels. The water table remains close to the surface throughout the year resulting in consistent saturation, similar to the flat bog (Quinton *et al.*, 2003). Water received from the surrounding permafrost plateau and bog landscapes is conveyed through the drainage network toward the basin outlet (Quinton *et al.*, 2003). As a result of this function, channel fens are often relatively nutrient-enriched (NWWG, 1988). Channel fens typically have a buoyant peat mat on the surface that responds to changes in the water table, resulting in an inconsistent surface elevation (Quinton *et al.*, 2003). The mat, which sits just below the water surface, supports the development of sedges, grass, herbs and shrubs above the water table, while trees are usually absent (NWWG, 1988).

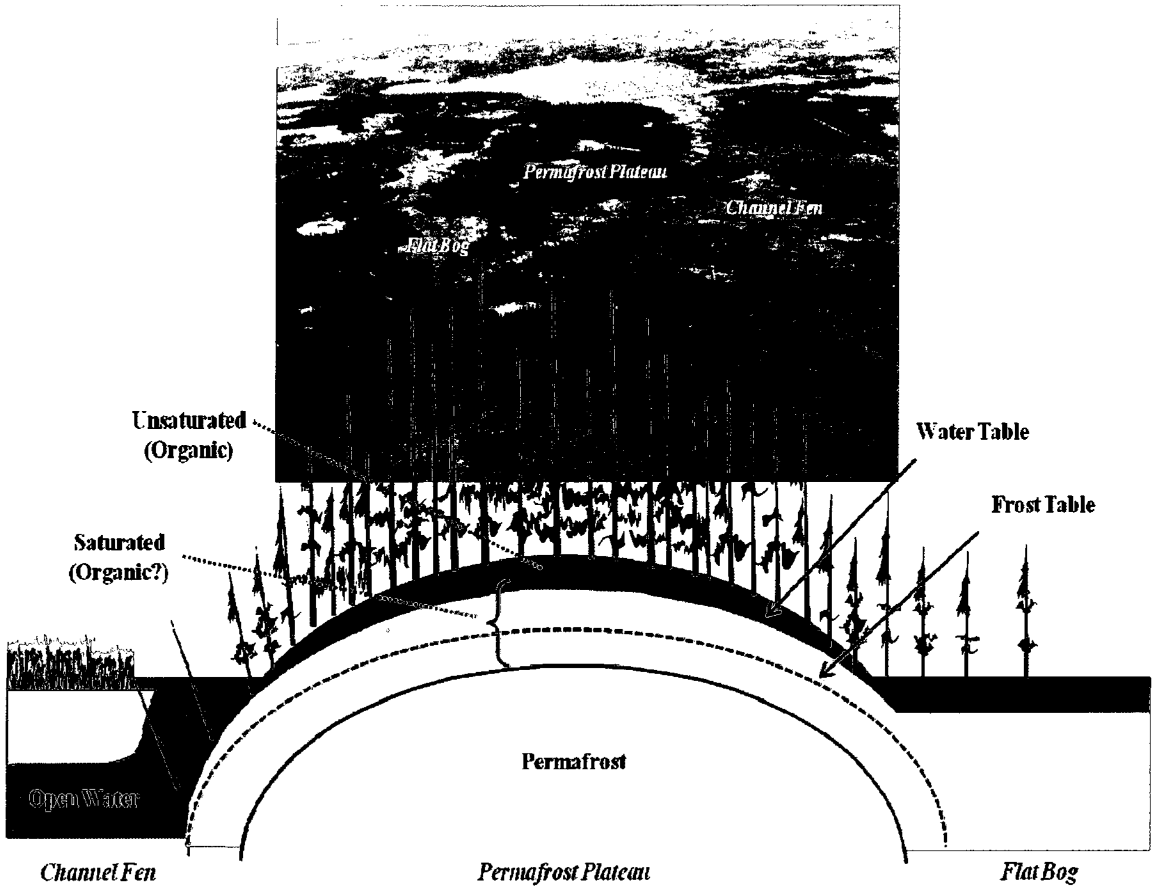

Figure 1-1 Image depicting a typical boreal wetland complex composed of peat plateaus, flat bogs, and channel fens, looking south (Goose Lake in the background), Northwest Territories, Canada (photo courtesy of William Quinton). Schematic cross-section of a boreal wetland complex detailing a permafrost plateau flanked by a flat bog and channel fen (Quinton *et al.*, 2009).

1.2 Permafrost Distribution and Landscape Evolution

Permafrost is ground that remains at or below 0°C for a minimum of 2 consecutive years (Brown and Kupsch, 1974). Approximately 50% of the Canadian land mass is underlain by permafrost (Geological Survey of Canada, 1998) and approximately 50% of the peatlands in Canada are found in permafrost-affected regions (Robinson *et al.*, 2003).

A significant portion of Northern boreal wetlands lie within the zone of discontinuous permafrost, which is both extensive and sporadic. Permafrost in the zone of discontinuous permafrost is particularly vulnerable to disturbance (Camill, 2005) and thaw can be initiated by disturbances such as fire, deforestation, erosion, flooding, and climate warming. In recent decades, increasing thaw depth and shallow ground temperatures have been documented around the globe and here in Canada (Wright *et al.*, 2008). Permafrost with a temperature close to 0°C is extremely sensitive to increases in the mean annual air temperature (Robinson, 2002). Halsey *et al.* (1995) documented a continuous northward shift of the southern boundary of permafrost throughout Alberta, Saskatchewan, and Manitoba as a result of over 200 years of relatively warm climate. Based on current global climate predictions permafrost retreat is expected to accelerate over the coming decades (e.g. Anisimov *et al.*, 2002; Camill and Clark, 1998). However, the increase in thaw depth could be even greater than what has been forecasted by modelling estimates.

While warming can enhance the productivity of an ecosystem, Zoltai (1993) and Vitt *et al.* (1994) argue the importance of permafrost to the survival of ecosystems in the wetland-dominated zone of discontinuous permafrost. Permafrost has three crucial roles in the climate system (Anisimov *et al.*, 2001; Nelson *et al.*, 1993; U.S. Arctic Research Commission Permafrost Task Force, 2003): first as a "geoindicator" of environmental change, acting as a temperature archive that does not experience seasonality in its warming or cooling below depths of 15 to 20 cm; second as a control for the hydrological and biological processes in its surrounding ecosystem, defining the seasonally thawed depth (active layer) to which they will be confined, altering surface and subsurface water fluxes and the functions of overlying vegetation; and third its control over the release of trace gases back into the atmosphere.

Rising air temperatures threaten the current physical characteristics of wetland landscapes as they evolve from permafrost to non-permafrost terrain. Further, with continued alteration to the growing season and its productivity, snow depth, and the timing of annual freeze-thaw cycles on a local scale, it is unknown how long permafrost will remain (Camill and Clark, 1998). What is known is that seasonal alterations further enhance the natural hydrological shift and subsidence of permafrost plateaus into collapse scar bogs (Camill and Clark, 1998). Permafrost distribution is also endangered by human disturbance. When the tree canopy is removed by human disturbance or natural processes, radiation received at the surface increases, resulting in an increased depth of thaw (Quinton *et al.*, 2009). By creating a localized depression in the frost table water from the surrounding area will drain towards it, increasing the water table depth and soil moisture in the unsaturated zone (Quinton *et al.*, 2009). Thermal conductivity of peat is enhanced with increasing soil moisture (Wright *et al.*, 2009), allowing thermal energy to transfer deeper into the sub-surface, resulting in a greater depth of thaw and further drainage (Quinton *et al.*, 2009). Unable to survive in water-logged conditions, the canopy is further thinned out resulting in even greater radiation loading at the surface over a larger area. Eventually an isolated flat bog will develop and over time it will grow in size with continued degradation, potentially leading to the elimination of the permafrost plateau.

1.3 Atmospheric Carbon and Global Change

General Circulation Models (GCMs) predict an increase in the average annual air temperature of 1.7-4.0 °C in northwestern Canada by 2050 (IPCC, 2007). The global mean annual air temperature has increased by approximately 0.74 °C ± 0.18 °C over the last 100 years (1906 - 2005) (IPCC, 2007). Atmospheric temperatures are strongly influenced by the concentrations of greenhouse gases (including carbon dioxide (CO_2), methane (CH_4), nitrous oxide (N_2O), and halocarbons), the concentrations of which have steadily increased over the industrial era, largely a result of human activities (IPCC, 2007). Anthropogenic emissions of CO_2 into the atmosphere since 1850 have been well documented (Watson *et al.*, 2000; Apps, 2002). Atmospheric CO_2 concentrations have increased as a result of fossil fuel combustion (coal, oil, and natural gas); however, during the last century a growing contributor to CO_2 increase has been the shift in ecosystem carbon responses due to increased land use change (IPCC, 2007). Approximately 43% of CO_2 emissions over the last century have been retained in the atmosphere (Apps, 2002). This accumulation could potentially contribute over 60% of the total increase in radiative forcing (Ruddiman, 2002). The growth rate for CO_2 in the atmosphere from 1995 to 2005 was 1.9 ppm yr^{-1} resulting in an increase in radiative forcing of 0.28 Wm^{-2} during the same time period (IPCC, 2007).

Northern peatlands are a substantial reservoir for greenhouse gases, accumulating carbon for thousands of years (Yu *et al.*, 2002; Rosenberry *et al.*, 2006). During the Holocene, post-glacial conditions resulted in a significant sink for atmospheric carbon in such environments (Makiranta *et al.*, 2009). The circumpolar boreal zone contains close to 30% of the global terrestrial carbon storage within its peatlands (Watson *et al.*, 2000;

Apps, 2002). The shifting balance between production and decomposition of peat has created a reservoir estimated at 455 Pg of C (Gorham, 1991). The cycle of CO_2 removal from, and emission into, the atmosphere has a large impact on global temperatures. However, the direction and magnitude of feedback to air temperatures resulting from CO_2 exchange between northern peatlands and the atmosphere is poorly understood and requires further research (IPCC, 2007; Schlesinger and Andrews, 2000; Yu *et al.*, 2002).

1.4 Ecosystem - Atmosphere Carbon Dioxide Exchange

The carbon cycle of peatlands includes carbon removal from the atmosphere through photosynthesis, storage in living and decomposing plant tissues in the active layer and permafrost, and its release to the atmosphere through autotrophic and heterotrophic respiration and peat oxidation. The general carbon (C) balance for terrestrial ecosystems is represented by,

$$\Delta C\ (NEE) = CO_2\ (GEP) - CO_2\ (R_{tot}) - CH_4 - DOC\ - DIC \tag{1.1}$$

where ΔC (NEE) is the net change in carbon storage within the ecosystem (g C m^{-2} day^{-1}), GEP is the gross ecosystem production, which represents total plant uptake or release of CO_2 from the system, R_{tot} is the loss of CO_2 from roots, microbial activity (decomposition) and vegetation respiration, CH_4 is methane, and DOC and DIC represent dissolved organic and inorganic carbon, respectively. However, in this study only gross ecosystem production and total respiration were examined to monitor the temporal and spatial variability and potential controlling variables on net ecosystem exchange and carbon accumulation. Thus, this balance was simplified for this study as,

$$NEE = GEP + R_{tot} \tag{1.2}$$

where net ecosystem exchange is the balance between photosynthesis (the process of CO_2 uptake from the atmosphere) and plant and soil respiration (the release of CO_2 back into the atmosphere) (Bubier *et al.*, 2003).

Over thousands of years, significant amounts of carbon dioxide (CO_2) have been removed from the atmosphere through photosynthesis and sequestered as un-decomposed organic matter (Camill and Clark, 1998). Photosynthesis is the process of converting light into chemical energy to sustain life in the biosphere (Schlesinger, 1977) and carbon is assimilated by plants during this process. CO_2 is diffused through the plant stomata into the leaf from the atmosphere, in exchange for water, which is available in excess in the leaf, and PAR is the photosynthetically active radiation from the sun (between 400 and 700 nm) (Campbell and Norman, 1998). Energy that is not utilized by the plant for metabolism and heterotrophic respiration is distributed through the plant to above-ground leaves and shoots, and below-ground to the roots (Raven *et al.*, 1999). Carbon dioxide, water, and light are fixed by the vegetation and converted into a carbohydrate (CH_2O) and oxygen (Campbell and Norman, 1998),

$$6CO_2 + 12H_2O + PAR = C_6H_{12}O_6 + 6O_2 + 6H_2O \tag{1.3}$$

Decomposition determines the rate of carbon respiration and carbon accumulation on the ground and varies with plant species, nutrient concentrations, and oxygen availability (Gorham 1991; Thormann *et al.*, 2002). The aerobic zone near the ground surface enables high rates of decay (Yu, 2002) as unsaturated conditions promote greater microbial activity. In the anaerobic zone, saturated conditions limit microbial activity; however, at greater depths climatic warming is no longer influential and the rate of decay is more constant (Yu, 2002). Older peat layers located at the bottom of the active layer are situated in direct contact with the underlying permafrost. At this depth, where the soil is frozen and saturated, organic material freezes, halting decomposition. As long as it exists in such a state, the carbon will remain stored within frozen residues (Figure 1-2). Slightly higher values of carbon are often recorded nearer the ground surface where new peat and fresh plant matter is initially less decomposed (Vardy *et al.*, 2000). However, as aerobic conditions persist the volume of carbon released will increase (Vardy *et al.*, 2000). Less mobile carbon is present at greater depths as the organic material will have undergone a greater degree of decomposition over time.

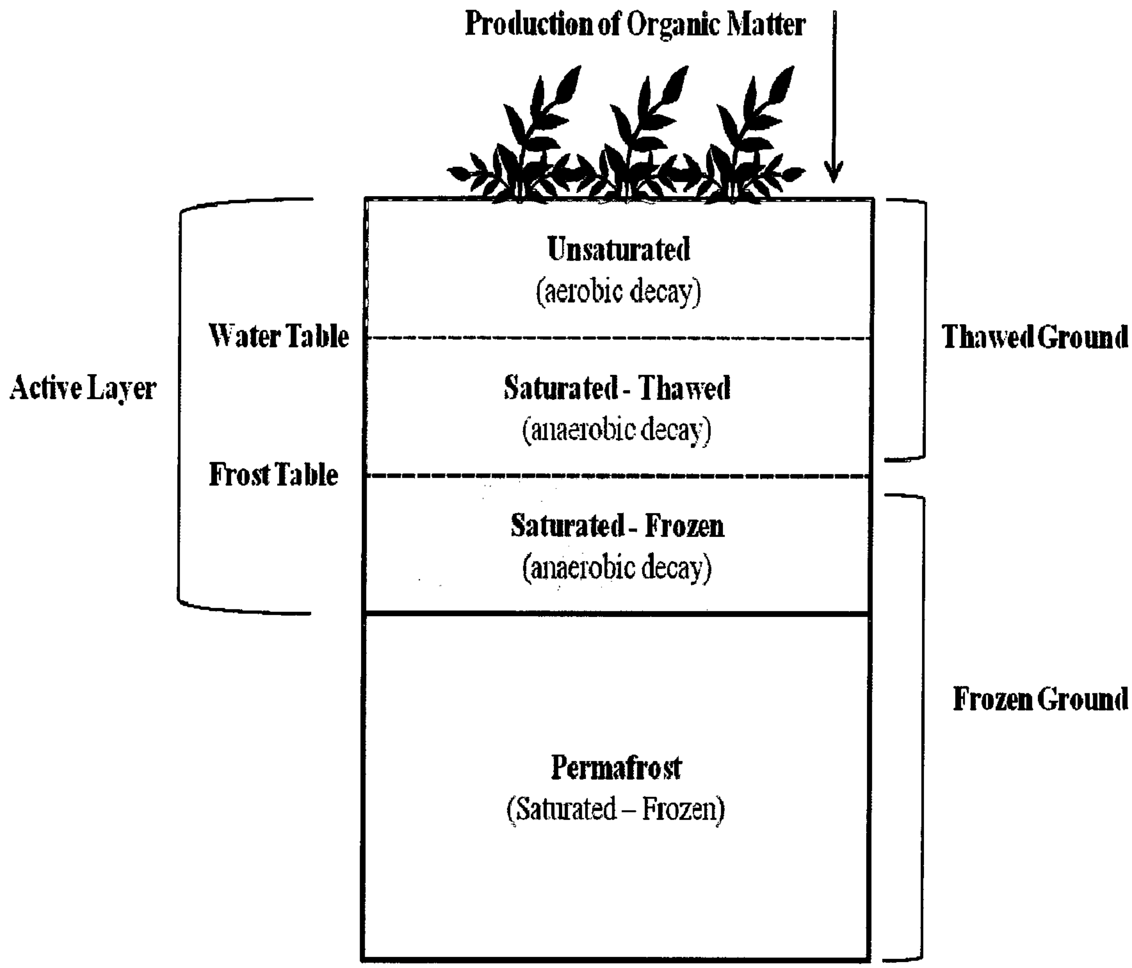

Figure 1-2 Accumulation and decay of organic matter in peatlands underlain by permafrost.

Carbon is released from peatlands through respiration and soil leaching. Respiration is the process of releasing carbon from plants, the soil, and organic compounds in the form of CO_2. Soil respiration involves a combination of processes: biotic (rhizosphere (root and root exudates)), heterotrophic (microbial and faunal), chemical (oxidation of soil carbonates), and physical (degassing) (Raich and Schlesinger, 1992; Suyker *et al.*, 2003).

1.5 Environmental Factors Controlling CO_2 Gas Flux

In addition to monitoring CO_2 flux, examination of site specific environmental variables, such as localized climate and unique physical characteristics of a landscape, provides a more comprehensive understanding of what drives the CO_2 flux within an ecosystem. As there is considerable spatial variability of sub-surface carbon, environmental measurements should be made at the location where CO_2 is being monitored and in the immediate surroundings. By determining the relationship between CO_2 and its localized environmental conditions, it is possible to relate temporal and spatial variability to changes in climatic, hydrological, vegetative, and soil characteristics.

The most influential climatic variables to CO_2 flux are air temperature and photosynthetically active radiation. Air temperature not only drives near-surface soil temperatures, creating a temperature gradient between the surface and atmosphere but also affects photosynthesis through stomatal control. Photosynthetically active radiation (between 400 and 700 nm) (Campbell and Norman, 1998) used in photosynthesis, helps vegetation fix CO_2 from the atmosphere, as discussed in the previous section.

Hydrological characteristics of a landscape are also important to CO_2 flux. During winter, CO_2 flux varies with depth of snow and the presence or absence of ice layers within the snow profile, directly above the surface (Bubier *et al.*, 1998). An overlying snowpack insulates the ground surface from cooler air temperatures, defining the depth of the frost table (Dingman, 1994; Sturm *et al.*, 2005). Below the ground surface, the frost table represents the lowest extent of the seasonally thawed sub-surface (Woo, 1986). As the relatively impermeable upper surface of the frozen and saturated soil beneath it, it also coincides with the zero-degree isotherm (Carey and Woo, 1998; Quinton *et al.*, 2000). The frost table depth is an indicator of the amount of available space for microbial activity to increase decomposition rates and the amount of thawed organic matter available to decompose and release CO_2. A snowpack also temporarily stores precipitation and reduces its contribution to the wetland surface. During snowmelt meltwater is unable to infiltrate past the frost table. Seasonal ice in the active layer decreases the available storage capacity and movement of water in the thawed and saturated portion of the soil profile (Wright *et al.*, 2008). As a result, water must drain laterally between the water table and frost table (Hayashi *et al.*, 2004; Wright *et al.*, 2008), downslope from the permafrost plateau into the adjacent wetlands. As meltwater infiltrates the ground surface and refreezes it convects heat into the sub-surface causing a rise in ground temperature and a lowering of the frost table (Woo, 1986). Thermal energy is conducted from the ground surface to the frost table at a higher rate in wet, rather than dry, peat (Hayashi *et al.*, 2007). The occurrence of overland flow on the typically dry plateau surface is rare; however, small ponds may develop on the surface (Wright *et al.*, 2009) during snowmelt and early spring when stored water is released, overwhelming an already heavily saturated environment. The large majority of water is stored in the saturated layer, with the remainder in the unsaturated zone. In relation to the amount of CO_2 within an ecosystem, the position of the water table controls the depth and size of the aerobic and anaerobic zones which are critical to productivity, decomposition, and peat accumulation (Rouse *et al.*, 1997). Lowering of the water table will result in a thickening of the aerobic zone and increase in oxygen availability needed to increase decomposition rates and thereby CO_2 loss (Bubier *et al.*, 2003). Water depth may also play a role in determining the number and type of microbial communities; thereby, indirectly influencing decomposition rates.

Vegetative and soil characteristics also play a role in determining site specific CO_2 flux, and can often be correlated with spatial variability within and between landscape units. Plant species in wetland environments are controlled by nutrient availability, soil characteristics, and hydrology. In turn they influence the soil microclimate structure through their root development, moisture and nutrient use, and the quantity and quality of detritus supplied to the soil profile (Raich and Tufekcioglu, 2000). Most importantly, plant species determine the rate of carbon assimilated from the atmosphere through photosynthesis and the amount contributed to the soil profile through the process of organic decomposition. Below the surface, characteristics of the soil profile, such as texture and porosity, can determine the ease and volume of water and gas movement in and out of the soil. The amount of thawed organic material and its total carbon and nitrogen composition can determine the volume of mobile carbon. Greater concentrations of nitrogen in the soil profile can stimulate plant production and therefore, result in a potential increase in storage of atmospheric CO_2 in soil organic matter (Holland *et al.*, 1997). Soil temperature is also a strong driver of CO_2 flux. Not only an indicator of frozen ground, soil temperature plays a role in controlling productivity and decomposition rates throughout the soil profile (Rouse *et al.*, 1997). Microbial activity is sensitive to temperature; however, recent studies (Minkkinen *et al.*, 2007; Makiranta *et al.*, 2009) have documented this sensitivity as having large spatial variation. Makiranta *et al.* (2009) suggest this variation may be a result of different soil microorganisms responding differently to temperature.

1.6 Chamber Measurement of Carbon Dioxide Flux

Although gas fluxes can be directly measured by eddy covariance at the ecosystem scale, continuous flux measurement requires extensive instrumentation, which can be difficult to maintain in harsh northern environments. In addition this method relies on atmospheric instability and sufficient wind velocities (Rouse *et al.*, 2002) and can be complicated by advection. Defining a footprint for eddy covariance can often be difficult to determine in ecosystems with heterogeneous landscapes. Chambers provide a much finer scale measurement from the soil and living vegetation at the spatial scale defined by the size of the chamber. While chambers do not provide an ecosystem scale assessment of fluxes or quantify the complete flux of an ecosystem (trees and shrubs are not included in monitoring), chambers do capture the diversity of each landscape through the characterization of temporal and spatial heterogeneity of CO_2 and its controls at the surface-atmosphere boundary. An infrared gas analyser attached to a portable chamber is used to measure CO_2 respiration. There are two chamber techniques that can be used: open-system and closed-system.

Open-system chambers allow interaction between the chamber environment and surrounding atmospheric conditions (Davidson *et al.*, 2002) and are sealed to the ground surface where CO_2 concentrations can be measured for hours, or days, at a time (Streever *et al.*, 1998). While chamber designs vary according to the objectives of a project there are some similar construction techniques. Ambient air is often pumped into the chamber, where it circulates thoroughly with the application of internal fans and then finally pumped out of the chamber through an exhaust feature (Streever *et al.*, 1998). As a result of the interaction with the surrounding atmosphere during measurement, consideration must be taken in regards to temperature, velocity, and flow of air entering and exiting the chamber (Alterio *et al.*, 2006; Balogh *et al.*, 2005). One of the main advantages to open-system chambers is their ability to measure continuously over longer periods, ranging from several hours to days at a time (Streever *et al.*, 1998). However, they are not the most conducive to some field studies as the equipment involved can be expensive and is not easy to transport if several sites are involved due to the complexity of its construction. In addition, the needed power supply is not always accessible in remote situations. In comparison, closed-system chambers do not incorporate surrounding atmospheric conditions. The chamber is sealed to the ground surface for the duration of measurement without any ventilation features. The use of an infrared gas analyzer (IRGA) is common for closed-system chambers and can be applied using two methods. The first, circulates air from the chamber to an external IRGA (Petrone *et al.*, 2008; Solondz *et al.*, 2008; Waddington *et al.*, 2003), through tubing connecting the two. The second technique measures CO_2 concentrations directly within the chamber through the insertion of a portable IRGA into the chamber structure.

Chambers are most commonly cylindrical structures that can be constructed out of materials such as vinyl chloride (Bekku *et al.*, 1995), transparent perspex plastic (Streever *et al.*, 1998; Strom and Christensen, 2007), or polycarbonate sheets (Grau, 1995; Bubier *et al.*, 2003). Clear chambers permit light penetration during measurement. Photosynthetically active radiation can have a direct relationship to the CO_2 concentrations being measured in the chamber and should be monitored simultaneously (Streever *et al.*, 1998). To construct a 'dark' chamber, covers can be placed over the chamber instantaneously (Streever *et al.*, 1998; Strom and Christensen, 2007) or in degrees of darkness using multiple shrouds of varying mesh sizes (Bubier *et al.*, 2003). The size and volume of a chamber should be small enough to capture minor flux measurements but large enough so as to minimize chamber disturbance effects.

To ensure data quality, attention must be given to the environmental factors that influence chamber performance: air temperature, relative humidity, pressure, and oxygen. A closed-system chamber sealed to the ground surface eliminates interaction with the surrounding atmosphere but as a result can develop its own microclimate, which may not accurately represent the current conditions of the system being measured. An enclosed chamber on the ground surface can result in potential flux disturbances related to air pressure differences between the chamber headspace and the ambient atmosphere, modification to wind speeds, influence on the soil-to-air CO_2 gradient, and alterations to temperature, relative humidity, and rates of evaporation within the chamber (Davidson *et al.*, 2002; Lund *et al.*, 1999). However, if the duration of measurement is short, this impact is minimal. Sampling time and duration has a direct influence on respiration rates. Bekku *et al.* (1995) found that longer sampling periods resulted in a significant decrease in CO_2 emission as the CO_2 gradient between the soil and air decreased within the chamber between 20-25 min after the start of measurement. Researchers that have utilized a closed-system chamber typically apply a sample period of 2-10 minutes in order to avoid gradient errors. A secondary consideration is the time of day when the measurements are made. Taking measurements at the same point and time every day creates an inaccurate representation of the site (Davidson *et al.*, 2002). CO_2 flux is not constant throughout a 24-hour period and measurements should reflect this by changing the order in which sites are sampled and at what time. Chamber measurements may also be affected by lateral diffusion of carbon in the soil profile. During simulations of both steady and non-steady-state chambers, Davidson *et al.* (2002) found that there was an alteration in the CO_2 gradient outside the perimeter of the collar placed on the ground surface. Therefore, a chamber may not be measuring total respiration at the surface of a soil column and losses may be occurring laterally beyond the chamber's capacity.

1.7 Temporal and Spatial Variability of CO_2 Flux

Northern ecosystems are experiencing changes in carbon cycling as a result of climate warming and land use change. In order to determine, or predict, the magnitude of response it is important to understand the sensitivity of the carbon cycle to local environmental changes and fluctuations. Studies examining the CO_2 flux of Arctic and Sub-Arctic regions have often focused on the fluxes of the mid-growing season to the exclusion of other seasons, eliminating several months that could be contributing to the net annual CO_2 balance. Elberling (2007) estimated that as much as 30-40 % of annual CO_2 loss can occur during the winter months. Therefore, this study focused on the transition between snowmelt conditions and the early stages of vegetation growth to capture the shoulder seasons that play a key role in the hydrology of northern wetlands and work towards a greater understanding for annual gas flux gradients.

Temporal variability of net CO_2 exchange during the shoulder seasons of spring and fall has often been attributed to plant species type, their growth cycle, nutrient conditions, and the variations in their interaction and response to soil moisture, photosynthetically active radiation and temperature (Bubier *et al.*, 1998; Frolking *et al.*, 1996; Griffis *et al.*, 2000; Kindermann *et al.*, 1996; Waddington *et al.*, 1998). The transition from cold winters to cool wet summers, results in slight shifts in climate, hydrology, vegetation, and soil characteristics that can impact the balance between a sink and source for CO_2. While snow cover in Northern Canada builds up over several months, snowmelt can occur in a matter of days. During snowmelt and spring, increases in air temperature and the amount of water alters the landscape. Wet and warm conditions promote soil thaw, plant growth and photosynthesis (Griffis *et al.*, 2000). In addition, the timing of leaf-out, can often correlate to an annual change from net loss to net gain of CO_2 (Griffis *et al.*, 2000) based on the assumption of increased plant activity (i.e. photosynthesis) sequestering CO_2. Furthermore, according to Frolking *et al.* (1996) the timing of spring could be one of the most important factors for inter-annual variability and should not be discounted. At high latitudes, snowmelt occurs when day length is near the annual maximum (Bubier *et al.*, 1998); therefore, reducing the delay in leaf-out and gas flux response once bare ground is snow-free.

During late summer and early fall, decreases in air temperature and the amount of water once again alter the landscape. Cool conditions promote vegetation senescence and thinning of the thaw depth as freeze-up begins and the winter season approaches. This cooling in addition to a decline in plant productivity can result in a change, once again, to the balance between net gain and net loss of CO_2. However, Goulden *et al.* (1998) demonstrated that soil respiration persists for an extended period at depth where soil cooling lags behind the cooling of shallower depths. Despite the initial lag in deep soil cooling, a significant portion of the soil profile will eventually freeze-up; limiting decomposition and the emission of CO_2 during winter months. The timing of surface thaw in the spring and freeze-up in the fall is therefore, important to the net exchange of CO_2 (Bubier *et al.*, 1998).

As previously discussed, in a northern boreal wetland there are very distinct differences between a permafrost plateau, flat bog and channel fen. Spatial variability of net CO_2 exchange can occur both within and between landscape units, thus, the significance of environmental controls may differ at various sites and scales (Maestre and Cortina, 2003). The relationship between CO_2 and site-specific conditions of microclimate, hydrology, vegetation and soil profiles are complex. Studies have attempted to clarify these intricate relationships through field observations and modeling predictions but they have yet to remain consistent between ecosystems. However, these studies have demonstrated that there is a significant temporal and spatial variation in gas flux based on changes in temperature, moisture, vegetation and sub-surface regimes (Bubier *et al.*, 1998; Davidson *et al.*, 2006; Elberling, 2007; Frolking *et al.,* 1998; Goulden *et al.*, 1998; Schreader *et al.*, 1998; Silvola *et al.*, 1996; Solondz *et al.*, 2008; Waddington and Roulet, 1996).

1.8 Study Rationale and Objectives

Northern boreal wetland ecosystems are large natural reservoirs for carbon, but are showing significant signs of impact by climate change. Changing climates have the ability to influence the balance of CO_2 by altering its relationships with permafrost distribution, local air temperature, and other environmental variables that play a role in the uptake, storage, and respiration of CO_2. Understanding the current relationships within northern permafrost-dominated peatlands and the drivers of CO_2 at the ecosystem scale can contribute to modelling and predicting how they will continue to respond to ongoing change (Lafleur, 2002).

Many previous and ongoing studies have focused on monitoring the carbon storage of peatlands in Norway, Sweden, and Greenland (Frolking *et al.*, 1998; Groendahl *et al.*, 2007; Lindroth *et al.*, 1998; Makiranta *et al.*, 2009; Silvola *et al.*, 1996; Waddington and Roulet, 1996; Waddington and Roulet 2000). However, approximately half of the global peatlands are distributed between Canada's boreal forest and arctic (taiga and low arctic tundra) zones. Studies and projects, such as the Boreal Ecosystem-Atmosphere Study (BOREAS), examined the boreal forests and peatlands stretching from Quebec, Ontario, Manitoba, Saskatchewan, into Northern Alberta (Bubier *et al.*, 1999; Griffis *et al.*, 2000; Lafleur *et al.*, 2001; Moore, 1989; Petrone *et al.*, 2001; Petrone *et al.*, 2003; Roulet, 1991; Solondz *et al.*, 2008; Trumbore *et al.*, 1999; Turetsky *et al.*, 2007). While these studies observed peatland and boreal ecosystems, few addressed the impacts and influence of permafrost. Moving north into Alaska (Fahnestock *et al.*, 1998; Jones *et al.*, 1998; Oechel *et al.*, 1995; Poole and Miller, 1982; Vourlitis *et al.*, 2003) and the Canadian Arctic tundra (Lafleur and Humphreys, 2007; Oberbauer *et al.*, 2007), permafrost environments have been studied under projects such as the International Tundra Experiment (ITEX). However, the examination of peatlands with underlain permafrost, in reference to CO_2 flux, has received comparatively little study. In the Fort Simpson, Northwest Territories, Canada region methane flux has been previously examined (Liblik *et al.*, 1997) but no studies have yet monitored CO_2 flux for the boreal wetlands of this area. While cold and remote locations in northern environments prove more difficult to monitor continuously they are the greatest indicators of potential CO_2 loss to the atmosphere with changing climate.

Historically, permafrost-dominated landscapes have acted as a net sink for CO_2 but the balance between uptake and emission is now tenuous enough that small changes in air and soil temperature, water table depth, timing and extent seasonal thaw and freeze-up could favour decomposition over plant production (Bubier *et al.*, 1998; Carroll and Crill, 1997; Chivers *et al.*, 2009; Shurpali *et al.*, 1995; Waddington and Roulet, 1996). Disturbances caused by changes in both climate and land use are resulting in landscape evolution both physically (permafrost degradation) and chemically (CO_2). Saturated and frozen conditions, maintained through the presence of permafrost and cooler air temperatures, limit decomposition thereby preserving larger quantities of carbon. If this accumulation exceeds respiration the sink function is maintained. However, climate warming has resulted in the thawing of permafrost. As frozen organic matter thaws and is available for decomposition, large quantities of carbon can then be lost in the form of CO_2, potentially shifting the ecosystem from a sink to a source of atmospheric CO_2. This loss of CO_2 back into the atmosphere from peat oxidation (Gorham, 1991) will exacerbate the warming climate trend and continue thawing. In addition, the landscape will experience increases in mean annual air temperature, increasing precipitation variability, and lengthening of the growing season (Price and Waddington, 2000). These changes will affect water availability, evapotranspiration and photosynthesis. The temporal and spatial variability of net CO_2 ecosystem exchange within a northern boreal wetland and between its representative landscapes is for the most part sparsely documented and unknown. Being unable to quantify these relationships makes it difficult to predict the future condition of boreal wetlands as a sink or source for CO_2 as they adapt to permafrost thaw and landscape evolution.

This study attempts to bridge the gap of limited research and understanding of CO_2 changes in northern boreal wetlands. To date there is no monitoring system for ecosystem or finer-scale CO_2 flux data collection at the Scotty Creek Basin, Northwest Territories, Canada. This study examines the interaction between the atmosphere and fluxes at the soil and ground surface interface. In order to relate CO_2 flux to its environmental controls measurements must be made at the same temporal and spatial scale (Davidson *et al.*, 2002). Therefore, the use of a closed-system static chamber to measure CO_2 flux at a finer scale provides the most direct way to measure net ecosystem exchange (NEE) and total respiration (R_{tot}). Moreover, by incorporating the surrounding environmental conditions of each site this research identifies significant relationships influential in controlling the CO_2 balance. Identification of these relationships can aid in the prediction of future shifts between sink and source of CO_2 and the greater understanding of the spatial variability of this physical process.

The objectives for this research project are to 1) characterize and compare the mid-day CO_2 flux for three landscape units: permafrost plateau, flat bog, and channel fen from the period of snowmelt through the growing season until late summer; 2) determine and compare the temporal patterns and ranges of total respiration, net ecosystem exchange, and gross ecosystem production between and within each landscape unit as they develop different light, vegetation, thermal and moisture regimes; 3) define the existing relationships between the gas flux gradients and their environmental controls to improve the understanding of boreal wetland CO_2 exchange; and 4) scale the relationships, as defined in objective #3, to the spatial variability of CO_2 for each landscape unit and use IKONOS satellite imagery and aerial photography to scale to the ecosystem level, in order to explore the potential changes in gas flux gradients as this ecosystem responds to landscape evolution.

Chapter 2 Study Site

2.1 Geographical Location

The study site chosen for this research is located within the Scotty Creek basin (61° 18'N, 121° 18'W). Covering an area of 152 km^2, this wetland-dominated drainage basin is situated approximately 50 km south of Fort Simpson, Northwest Territories, Canada in the lower Liard River valley. Located centrally within Canada's largest river basin, one of the most unique ecological and hydrological areas of Northern Canada, it also lies within the boundaries of Canada's continental high boreal wetland just slightly south of the transition into low subarctic wetland (NWWG, 1988), and in an area dominated by discontinuous permafrost (Heginbottom and Radburn, 1992) (Figure 2-1). The study site in Scotty Creek basin was located 1.1 km North of Goose Lake (Figure 2-2) with sites distributed across permafrost plateau, flat bog, and channel fen landscape units. Field measurements were limited to two trips, the first made between April and June 2008, to evaluate CO_2 exchange during the months capturing snowmelt and early growing season conditions. The second, only involved one day of measurements at the end of August 2008 to capture an example of late summer conditions.

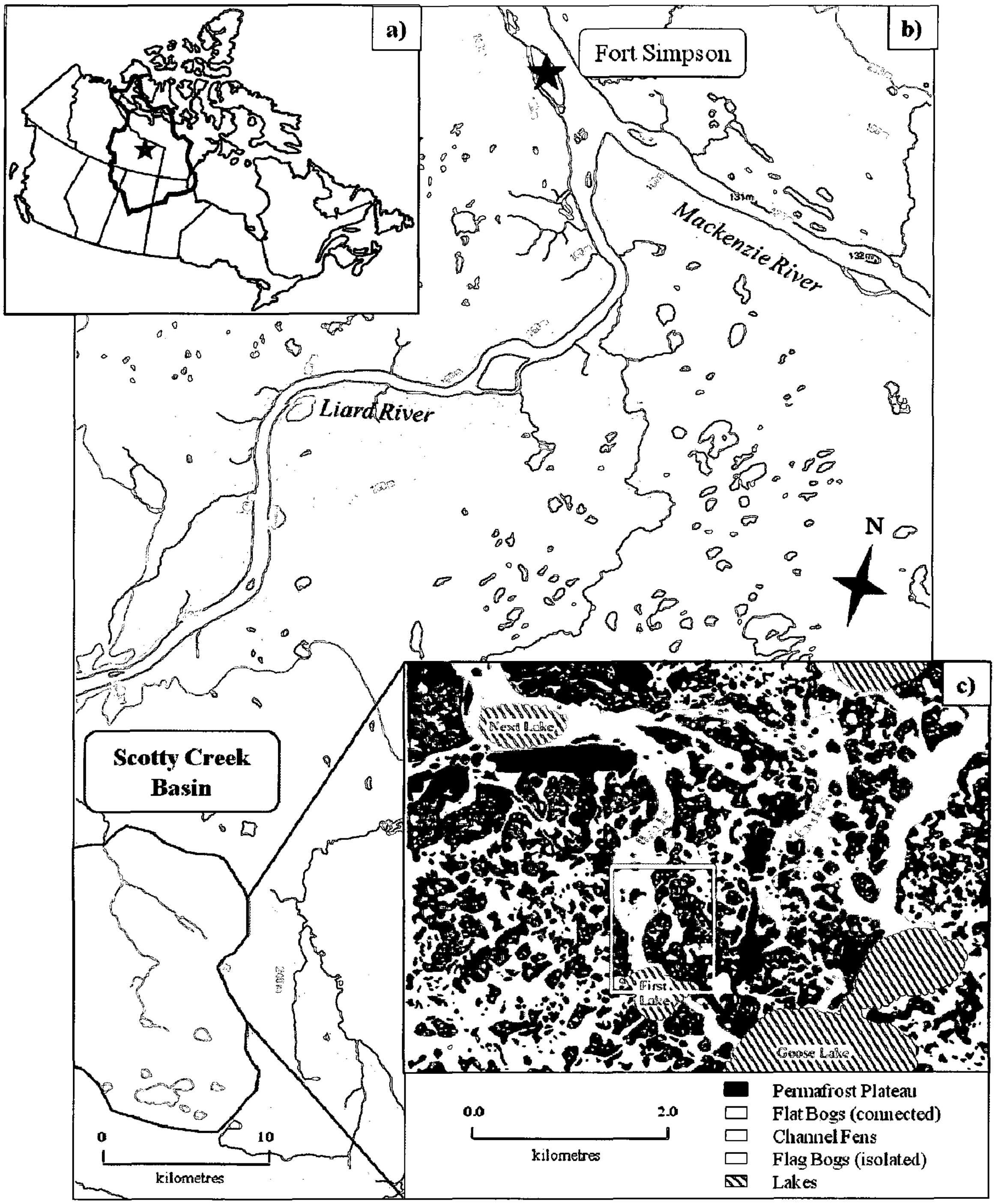

Figure 2-1 (a) Location of study site within the Mackenzie Basin (Cohen, 1997). (b) The lower Liard River valley near Fort Simpson, Northwest Territories, Canada (Natural Resources Canada, 2009). (c) Major ground-cover types in Scotty Creek within a 22 km^2 area of interest (Quinton *et al.*, 2009). The area presented in (**Figure 6-1**) is outlined in white in Figure 2-1(c).

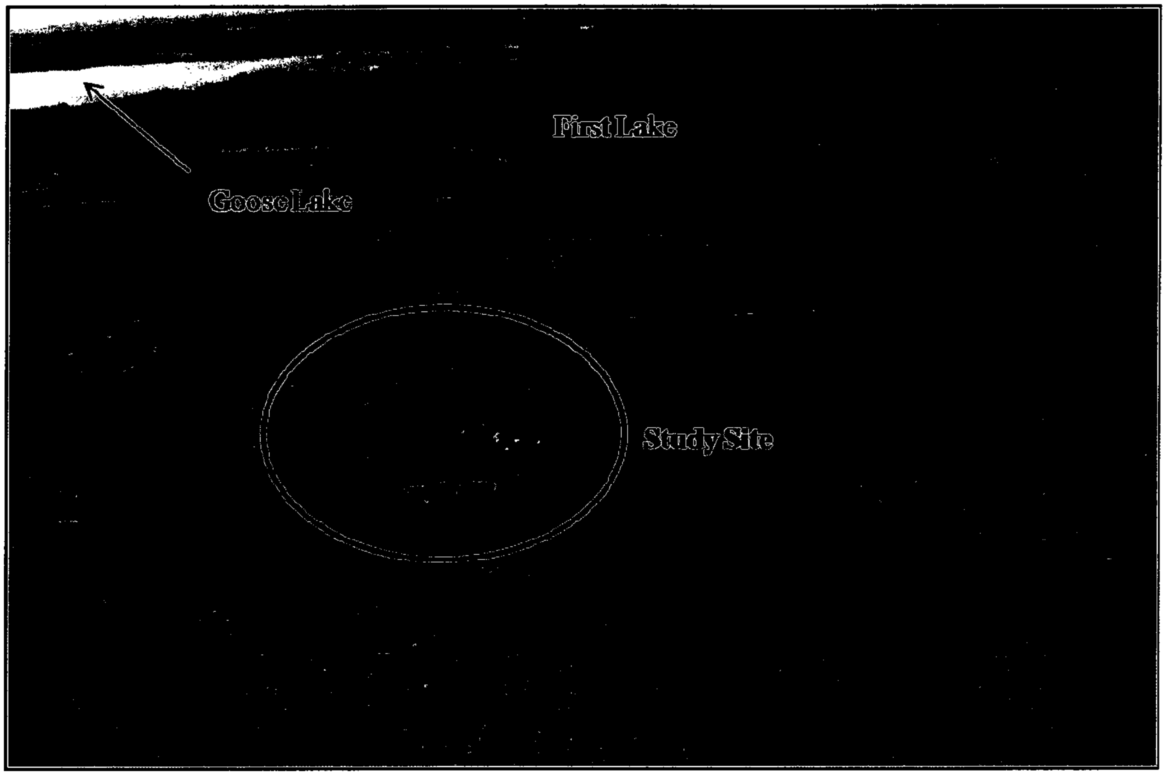

Figure 2-2 Location of study site, looking south with Goose Lake in the background (photo courtesy of William Quinton).

2.2 Physiography

Scotty Creek basin is composed of three main landscape units: permafrost plateaus, flat bogs, and channel fens. These landscape units are easily identified in aerial photography by abrupt changes in vegetation coverage and their distinguishable forms. Ground classification conducted by Quinton *et al.* (2009) on a sub-section of the Scotty Creek basin, covering an area of approximately 22 km^2, indicated the dominance or permafrost plateaus on the landscape. Permafrost plateaus occupied the greatest areal portion (43%), followed by isolated and connected bogs (26.7%), and channel fens (21%). As was previously discussed in the introductory chapter, each of these landscape units exhibits differences in topography, hydrology, vegetation and canopy coverage, and underlying organic and soil profiles. These unique characteristics are described in the following sub-sections.

2.2.1 Permafrost plateau

Permafrost plateaus are easily distinguished on the landscape as the only forested terrain above the water table. Situated ~ 1-2 m above the surrounding wetlands these elevated landscape units are also the only features underlain by permafrost (Quinton *et al.*, 2003; Robinson and Moore, 2000). Plateaus vary in size, form, and stage of development (youthful, mature, and old) and are dominant in the Fort Simpson region; however, increasing rates of maturity and permafrost degradation are threatening this landscape dominance. Permafrost thickness in the region has been reported to be between 5-10 m (Burgess and Smith, 2000). The plateau examined in this study rises 0.9 m above the surrounding wetlands and appears to be degrading. Along most of its edges it shows signs of collapse as well as additional subsidence in its central portions, which may be a result of isolated patches of surficial permafrost loss. The water table of permafrost plateaus fall 0.5 m or more below the ground surface during the annual thaw and drainage of the active layer (Quinton *et al.*, 2003). In terms of its role in the hydrological network, permafrost plateaus experience surface and sub-surface runoff as water drains from the plateau into the topographically lower wetland of permafrost-free isolated and connected bogs and channel fens (Quinton *et al.*, 2009). It also acts as a physical barrier to obstruct and redirect water flow owing to the high elevation of the permafrost table above the surrounding wetlands (Quinton *et al.*, 2003). However, additional release of water from permafrost thaw and the development of flow pathways between landscape units previously separated by permafrost plateaus are to be expected as basin storage and runoff response to hydrological inputs change.

The study plateau supports an open tree canopy, composed predominantly of black spruce (*Picea mariana*). Maximum measured tree height is 9.3 m with a mean tree height of 3.1 ± 2.2 m, and density of 1 stem m^{-2} (Wright *et al.*, 2008). Ground cover is dominated by lichen, feather and *Sphagnum* mosses, in addition to Labrador tea, and ericaceous shrub species (Table 2-1). Scotty Creek has an organic cover that can range up to 8 m in depth under which a silt-sand layer and thick clay to silt-clay layer (with low permeability) respectively lie (Aylesworth and Kettles, 2000). The organic cover contains an upper layer (the top 0.5-0.2 m) composed of living vegetation and lightly decomposed fibric peat under which lies a layer of dense sylvic peat in a more advanced state of decomposition with dark, woody material and the remains of lichen and moss, rootlets and needles (Quinton *et al.*, 2003).

2.2.2 Flat Bog

The flat bog is a low-lying feature in comparison to the permafrost plateau without the underlying presence of permafrost to elevate it above the water table. They appear as patches on the landscape (Quinton *et al.*, 2003) with no consistency in size or form. The water table remains close to the surface throughout the year, except during the snowmelt runoff period when it rises above its peat mat, resulting in a high level of consistent saturation (Quinton *et al.*, 2009).

The flat bog examined in this study is connected to the nearby channel fen, therefore during periods of high water there is potential for sub-surface lateral flow into the fen. The flat bog is dominated by *Sphagnum* mosses with thick underlying layers of *Sphagnum* peat (Table 2-1). There is little to no canopy coverage due to unstable conditions for trees to take root and remain upright. Therefore, only a few black spruce (*Picea mariana*) stand stunted and scattered throughout the bog. The ratio of above water vegetation to open water fluctuates according to changes in the water table and the growth of vegetation, altering the presence and distribution of mosses above the water surface.

2.2.3 Channel Fen

Like the flat bog, the channel fen is a low-lying feature in comparison to the permafrost plateau as it does not have the underlying presence of permafrost to raise it above the water table. They appear as large linear features on the landscape along the drainage network of a basin (Quinton *et al.*, 2003), conveying water received from the surrounding permafrost plateau and bog landscapes through the drainage network toward the basin outlet (Quinton *et al.*, 2003). Like the flat bog, the water table remains close to the surface throughout the year retaining a high level of consistent saturation (Quinton *et al.*, 2003).

The channel fen examined by this study runs the length of the study site, along the northwest side of the permafrost plateau and connected to the bog. The fen has a buoyant *Sphagnum riparium-dominated* peat mat on the surface, approximately 0.5 - 1.0 m thick (Hayashi *et al.*, 2004) that responds to changes in the water table, resulting in an inconsistent surface elevation. The mat, which sits just below the water surface, supports the development of sedges, feather mosses, grasses, and various herbs and shrubs above the water table (Table 2-1). There is no canopy coverage due to unstable conditions for trees to take root and remain upright. Therefore, the presence of trees does not extend far beyond the fringe of the channel fen. Beneath the peat mat, at a depth of 3 m below the water surface, is a dense organic layer with mineral soils (Hayashi *et al.*, 2004).

Table 2-1 Vegetation species for each landscape unit based on percent coverage in area captured by chamber collars for each site. Refer to **Figure 3-1** for site locations.

Site	Canopy Cover	Ground Cover	Above Ground Biomass	%
PPC-1	Closed	*Sphagnum capillifolium ; Sphagnum fuscum*		90%
	Picea mariana	*Sphagnum girgensohnii*		10%
			Rubus chamaemorus	2%
			Ledum groenlandicum	5%
			Oxyoccus microcarpus	5%
			Vaccinium vitis-idaea	5%
			Betula glandulosa	10%
PPC-2	Closed	*Sphagnum capillifolium*		90%
	Picea mariana	*Sphagnum girgensohnii*		10%
			Oxyoccus microcarpus	2%
			Rubus chamaemorus	5-10%
PPC-3	Closed	*Cladina mitis; Cladina rangiferina*		85%
	Picea mariana	bare ground		15%
			Ledum groenlandicum	2%
			Rubus chamaemorus	10%
BC-1	Open	*Sphagnum capillifolium ; Sphagnum fuscum*		50%
		Sphagnum riparium		50%
			Carex spp.	2-5%
			Andromeda polifolia	5%
			Chamaedaphne calyculata	5%
BC-2	Open	*Sphagnum capillifolium ; Sphagnum fuscum*		50%
		Sphagnum riparium		50%
			Carex spp.	1%
BC-3	Open	*Sphagnum capillifolium ; Sphagnum fuscum*		50%
		Sphagnum riparium		50%
			Chamaedaphne calyculata	1%
			Carex spp.	2%
FC-1	Open	*Helodium blandowii*		50%
		Sphagnum squarrosum; Brachythecium rivulare		40%
		Aulacomnium palustre		10%
			Oxyoccus microcarpus	5%
			Potentilla palustris	10%
			Salix pedicellaris	10%
			Carex spp.	15%
FC-2	Open	*Aulacomnium palustre*		100%
			Oxyoccus microcarpus	5%
			Galium trifidum	5%
			Carex spp.	40%
FC-3	Open	*Brachythecium rivulare*		90%
			Calla palustris	10%
			Carex spp.	20%

2.1 Climate Record

The region experiences a dry continental climate characterized by cold, long winters and short, dry summers limiting the growing season (NWWG, 1988; Robinson and Moore, 2000). Climate data is available from the Fort Simpson airport, which is the nearest Environment Canada weather station, 50 km north of the study site and 169 m above sea level. The mean annual air temperature for this region has increased by approximately 2.0 °C over the last 100 years (1906-2005) (Environment Canada, 2010). Based on the 30-year annual averages (1971-2000), the Fort Simpson region receives 369 mm of precipitation of which 170.3 cm is snowfall while the average annual temperature is -3.2 °C (Environment Canada, 2010) (Figure 2-3). During the year in which sampling occurred (2008) the average annual temperature was slightly colder at -4.1 °C, with 360.5 mm of precipitation and 259.8 cm of snowfall. Snowmelt typically begins in the later weeks of March and continues through the month of April, with little to no snow remaining by May (Hamlin *et al.,* 1998) at which point all waterways are open and flowing. It is these cold and wet conditions dominating in the low-lying flat terrain of a northern wetland that perpetuates the poorly drained landscape with a sub-surface profile of peat, organic soil and underlying permafrost.

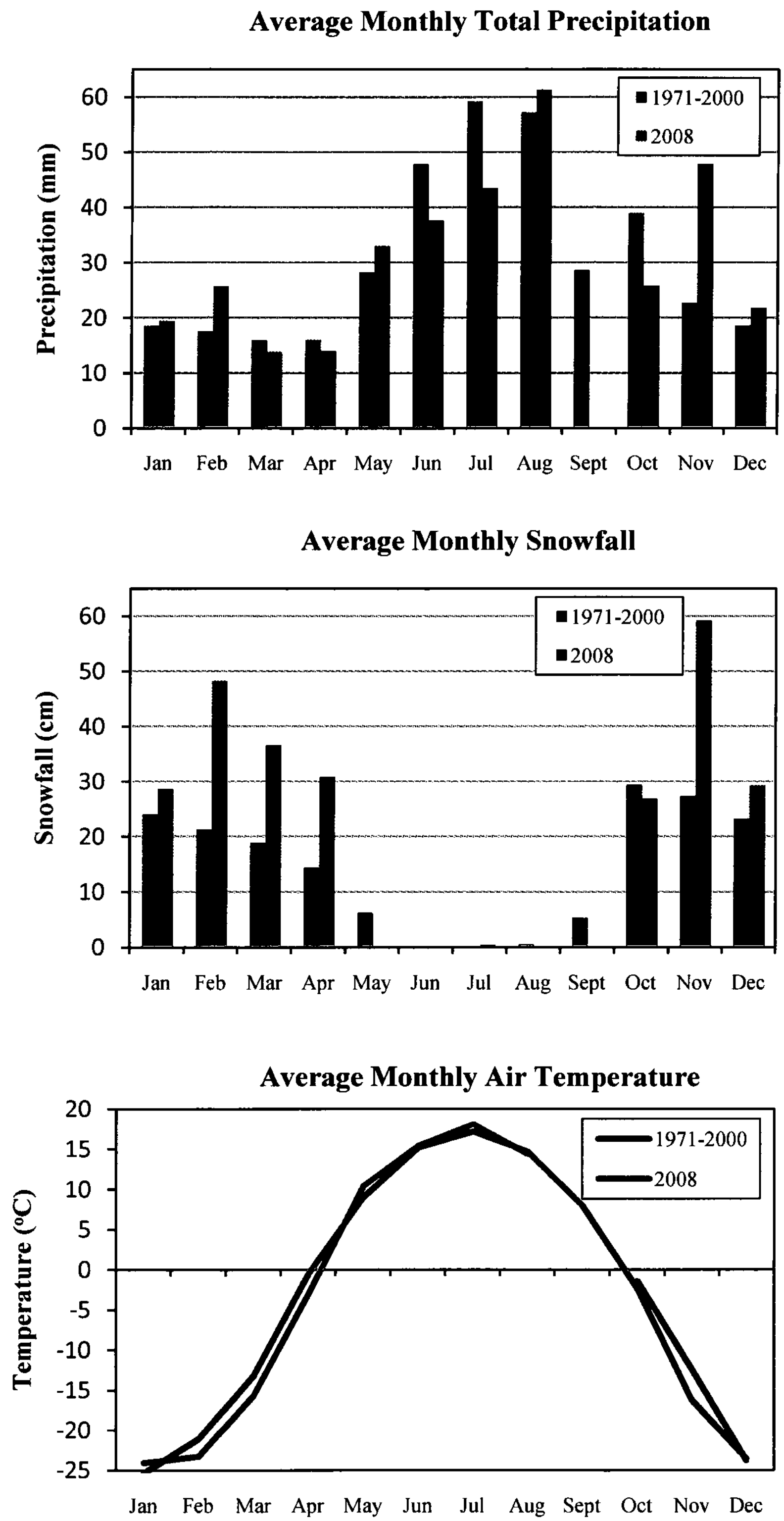

Figure 2-3 Fort Simpson climate normals for 1971-2000 and monthly averages for 2008 (Data Source: Environment Canada, 2010).

2.2 Permafrost Distribution and Landscape Evolution

Approximately half of the peatlands in the Fort Simpson region are underlain with permafrost (Robinson *et al.*, 2003). However, due to the disequilibrium of discontinuous permafrost with the current climate (Geological Survey of Canada, 2007) and increasing land use changes, this region is at risk of rapid change to its current permafrost distribution. Slight shifts in temperature can have significant impacts on the sensitive state of permafrost in the Fort Simpson region, which has experienced some of the greatest warming in Canada during the past century and where permafrost temperatures are generally just slightly below 0°C (Robinson, 2002). Scotty Creek basin is also situated within a General Use Zone in the Dehcho territory (Dehcho Land Use Planning Committee, 2006). Seismic cut lines and winter roads in this region have traversed bog, fen, and permafrost plateaus, removing canopy and vegetation in their path. This has resulted in permafrost degradation and alterations to the natural drainage pattern, habitat and migration patterns in the immediate surrounding area. The combined length of roads and seismic cut lines within the Scotty Creek basin is 133.2 km, and the density of linear features within the 152 km^2 area is 0.88 km^{-1} (Quinton *et al.*, 2009).

In the Mackenzie Valley air photo coverage extends back to the late 1940's. In a study conducted by the Geological Survey of Canada (Geological Survey of Canada, 2001), comparisons between these historical aerial photographs and more recent IKONOS satellite images were used to examine and quantify landscape evolution from permafrost degradation. For an area approximately 40 km west of Fort Simpson, the terrain encompasses within a 1947 aerial photograph and IKONOS satellite image taken in 2000 was delineated and classified as unfrozen or frozen. In 1947, the landscape was classified as 45% frozen and 55% unfrozen while in 2000 it was classified as 22% frozen and 78% unfrozen. A documented loss of approximately 23% of the landscape, once frozen ground, over a time span of 53 years identifies significant and rapid loss of permafrost for this region. It also documents the disappearance and degradation of permafrost plateaus, shown by the expansion and merging of bogs and fens. Similar analysis was conducted by Chasmer *et al.* (in press), looking more specifically at a 1 km^2 area within the Scotty Creek basin. Land cover classification was delineated from a series of aerial photographs taken between 1947 and 2008, in addition to IKONOS multispectral satellite imagery (4 m resolution) from 2000. In 1947, the 1 km^2 subset area was classified as 70% frozen, decreasing to 43% in 2008. The degradation and disappearance of plateaus have resulted in growing and merging wetlands (i.e. bogs and fens), corresponding to a 38% loss in permafrost for this area over a time span of 61 years (Chasmer *et al.*, (in press); Quinton *et al.,* (in review)).

Chapter 3 Methodology

3.1 Experimental Design

Three landscape units were selected based on their close proximity to one another and the existing presence of meteorological (MET) stations in each one: permafrost plateau, flat bog and channel fen (Figure 3-1). Sites within each landscape unit were chosen based on their representation of major site types. Nine sites in total were constructed over the entire site, with three placed in each landscape unit (Figure 3-1). Permafrost plateau collars (PPC) were placed across the width of the plateau. PPC-1 and PPC-3 were placed on opposite flanks, with PPC-2 placed on the plateau crest (Figure 3-1). Flat bog collars (BC) extended from the edge of the plateau to the center of the bog. BC-1 was placed in a bog near the edge of a permafrost plateau; BC-3 was in the centre of the bog, and BC-2, roughly mid-way between the two. Representative sampling across the width of the fen was not possible due to the persistence of standing water. Therefore, all fen collar (FC) sites (FC-1, FC-2, FC-3) were located along one side of the fen which was accessible for sampling.

The respective MET stations monitored meteorological variables automatically within each landscape unit during the study. Additional environmental variables and CO_2 flux were measured discretely at each site. Each site within the bog and the fen, in addition to PPC-1 and PPC-2 were developed as 4 m long linear transects. Environmental variables were measured every 0.5 m along the length of each transect, while chambers were located at the 2 m mark to monitor CO_2 flux. The ninth site, located at PPC-2, was developed with a 5 m by 5 m grid (6 transects, 9 points along each). Environmental variables were measured every 1 m along the length of the transect, while the chamber was situated in the center of the grid. While the bog and fen sites do not experience much variation in topography, the elevated peat plateau had some slight variations over the length of the study site. In order to capture the variation between elevated and depressed points, measurements were taken at the three sites (PPC-1, PPC-2, and PPC-3) along each transect from the surface at each point up to a given height. Measurement of environmental variables began on April 13th; however, chamber measurements were delayed until April 26th when the melting snowpack was able to support the weight of the chamber.

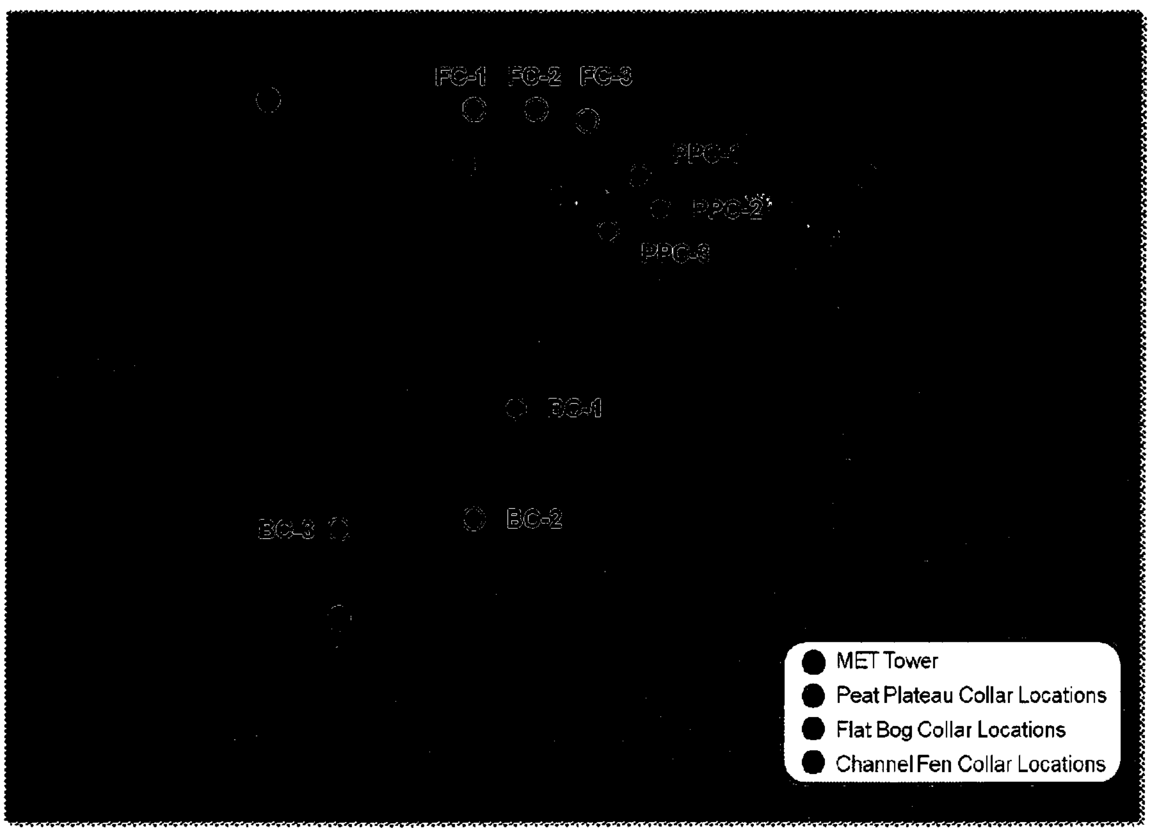

Figure 3-1 Sampling Design chosen for Scotty Creek outlining locations chosen for collar placements and chamber measurement (photo courtesy of William Quinton).

3.2 CO_2 Measurement

At each site, CO_2 respiration and net ecosystem exchange were measured with a closed-system chamber and a portable infrared gas analyzer (IRGA), the Vaisala CARBOCAP® GMP343 Carbon Dioxide Probe (Vaisala Oyj, Vantaa, Finland). The GMP343 is a single-beam, dual-wavelength, non-dispersive infrared sensor. Inside the chamber, pulsed light is emitted from a miniature filament lamp, reflecting and refocusing back to the infrared detector which is situated behind a silicon-based Fabry-Perot Interferometer. The CO_2 gas present in the chamber absorbs a portion of this light at a particular wavelength. The detector is designed to measure this wavelength and recognize it as a loss in the light transmission. The detector is also designed to measure another wavelength that acts as a reference signal, which has no absorption and records it as no loss in light transmission. The ratio between the two signals is the gas concentration value. All data is recorded and sent to the M170 indicator which simultaneously serves as a display, communicator (with the probe), and data-logger.

NEE was measured using a clear lexan chamber to capture both plant uptake simultaneously with soil and plant respiration. A dark shroud (two black plastic bags) was placed over the chamber for the measurement of R_{tot} for soil and plant respiration (Figure 3-2). Polyvinylchloride (PVC) collars (interior radius = 12.3 cm) were inserted at each location on April 12th, 2 weeks prior to sampling and remained permanently in place for the duration of the entire study. This was done to minimize plant and soil disturbance and allow for repeated measurement at a single location. Chamber measurements were then conducted from April 26th to June 6th, and then again on August 23rd. Measurement on the fen sites began on May 7th once the sites could be safely reached.

Figure 3-2 Dark and clear chambers measuring CO_2.

Grooves cut along the top of each collar were filled with water before the chamber was placed on top; this ensured an airtight seal to limit interaction between the chamber environment and the surrounding atmosphere. The chamber enclosed a surface area of 0.56 m^2, with a volume of 0.03 m^3. As seen in Figure 3-2, to avoid the development of a "greenhouse effect", from lack of circulation, the interior air temperature was cooled by a tube that circulated water pumped in to maintain an atmosphere within 2 °C of ambient conditions. In addition, a small fan was mounted to the inside of the chamber, circulating air to equilibrate the gas concentration and avoid an artificial decrease in gradient. The fan was mounted near the top of the chamber to also minimize aeration of the soil. The GMP343 probe inserted into the side of the chamber, recorded every minute for ten minute durations. To ensure the quality control of CO_2 flux for each measurement, a trial test was conducted at each site before the start of the study period. The regression slope (change in CO_2 over the duration of each sampling interval) for the trial tests was used as a reference for all sampling intervals conducted during the rest of the study. Samples that deviated significantly from this linear regression slope were removed. Dark and clear chambers were run at each location daily. Measurements were made during peak daylight hours between 10:00 and 17:00 daily. The measurement times at each site, and the order of measurement was varied day to day to minimise a temporal bias in the sampling regime.

The dark chamber measurements were made first, after which the chamber was removed from the collar and aired out for approximately two minutes. The water in the collar groove was replaced, and the clear (i.e. un-shrouded) chamber was set-up. The gas flux was then computed from the rate of increase in CO_2 concentration in the chamber. CO_2 concentration, in ppm, was used to estimate the average gas flux F and converted into mg CO_2 m^{-2} sec^{-1} using:

$$F = \frac{\left[\left(\frac{\Delta \times MM}{N} \times \frac{CV}{A}\right) \times CF\right] \times CF_2}{T} \tag{3.1}$$

where F is the gas flux (g CO_2 m^{-2} sec^{-1}), Δ is the linear change in CO_2 concentrations with time (ppm min^{-1}), MM is the molar mass of CO_2 (44.01 g mol^{-1}), N is the molar volume of an ideal gas (22.414 L mol^{-1}) at standard temperature and pressure (STP), CV is the temperature corrected volume within the chamber (m^3), A is the chamber area (m^2), CF is the conversion factor from ppm to mol (1ppm = 10^{-6} mol), CF_2 is the conversion factor from grams to milligrams (1 g = 1000 mg), and T is the conversion factor for time (60 sec min^{-1}).

Gas flux calculated from the dark chamber represents the gross R_{tot} for above and below ground respiration. The gas flux calculated from the clear chamber represents the NEE of both above and below ground respiration in addition to photosynthesis. Since the R_{tot} and NEE measurements were made within minutes of each other the difference between the two types of chamber measurement is considered the gross ecosystem production (GEP) at that particular light level. The computation of GEP determines the influence of photosynthesis by subtracting total respiration from the net ecosystem exchange:

$$GEP = NEE - Rtot \tag{3.2}$$

Positive fluxes indicate CO_2 emission from respiration, while uptake of CO_2 is indicated by negative fluxes. The field study was divided into four periods to better depict the temporal relationships: snowmelt (SM), pre-green (PG), green (G), and late-green (LG) (Table 3-1). Seasonal periods are based on the 2008 environmental conditions where changes in snow cover, temperature and vegetation growth clearly defined four stages of development within the duration of measurement (Table 3-2). The Snowmelt season extended until the sites were snow-free, showing visible signs of exposed bare ground. Pre-Green season extended from when the ground became snow-free until the emergence of immature vegetation. The Green season represents the early summer period following leaf-out. Late-Green represents only a single measurement day at the end of summer, a result of limited time at the study site. Fluxes were then averaged within these periods to get temporal flux values.

Table 3-1 Growing seasons for measurement during the 2008 field season transitioned from snowmelt (SM), visible melting snowpack exposing bare ground; pre-green (PG), emerging immature vegetation; green (G), maturing vegetation; to late-green (LG), the onset of dormancy.

Season	Dates
Snowmelt	April 26 – April 30
Pre-Green	May 1 – May 10
Green	May 11 – June 6
Late-Green	August 23

Table 3-2 Environmental conditions and stages of landscape unit development that defined the four growing seasons for the 2008 field season. Refer to **Table 2-1** for vegetation species composition and percent coverage in collar.

	Permafrost Plateau			Ombrotrophic Flat Bog			Channel Fen		
	PPC-1	**PPC-2**	**PPC-3**	**BC-1**	**BC-2**	**BC-3**	**FC-1**	**FC-2**	**FC-3**
Maximum Snow Depth (cm)	95	72	74	72.5	58	62	62	52	41
Date Measured	Apr. 13th	Apr. 14th	Apr. 13th	Apr. 14th	Apr. 13th	Apr. 13th	Apr. 13th	Apr. 13th	Apr. 13th
Bare Ground Exposure	May 8th	May 2nd	May 5th	May 3rd	May 1st	May 1st	May 2nd	May 1st	May 1st
Maximum Flood Depth (cm)		19							
Date Measured		May 22nd							
Duration of Flooding		May 2-28th							
Minimum Plant Height (cm)	35	3	15	2			5	6	24
Percent coverage in collar (%)	5	1	2	1			1	2	5
Date Measured	May 7th	May 21st	May 9th	May 17th			May 16th	May 14th	May 8th
Maximum Plant Height (cm)	36	14	12	15	5	4	60	50	85
Percent coverage in collar (%)	35	10	10	20	5	5	100	100	60
Date Measured	June 6th	June 6th	June 6th	Aug. 23th	Aug. 23th	Aug. 23th	Aug. 23th	Aug. 23th	Aug. 23th

3.3 Environmental Factors Controlling CO_2 Gas Flux

To define relationships between CO_2 flux and the environmental factors discussed in Chapter 1, the latter were monitored continuously at meteorological towers and discretely at each chamber site (Figure 3-1) during the CO_2 flux measurements. Cloud cover and general weather conditions were observed and noted daily at the time of the flux measurements.

The climatic conditions measured were: air temperature (T_a), relative humidity (RH), photosynthetically active radiation (PAR), soil (T_{soil}) and water temperature (T_{water}), and precipitation (P). Mounted on the bog and plateau MET towers at a height of 1.7 m and 2.4 m above the ground surface, T_{air} and RH, respectively, were measured every 60 seconds and then averaged every 30 minutes by a HMP45C Temperature and Relative Humidity Probe (Vaisala, Finland) and recorded by a CR10X and CR1000 Datalogger (Campbell Scientific Inc., Utah). Mounted on the interior wall of the chamber, a Veriteq Spectrum SP-2000-20R data logger (Veriteq Instruments Inc., Canada) monitored T_{air} and RH for the enclosed environment. Measurement within and outside the chamber monitored and ensured the chamber's enclosed environment remained representative of its surroundings. PAR was monitored outside of the chamber during flux measurements of the clear chamber using a Quantum Metre, Model QMSS (Apogee Instruments Inc., Utah). Placed unobstructed on top of the chamber, the quantum meter measured PAR every minute simultaneously with the IRGA. T_{soil} readings were taken at six points around the chamber collar (from which an average was calculated) at a depth of 20 cm using a Digi-Sense Thermocouple (Type T) Model 60010-20 (Eutech Instruments, Netherlands). As a result of high water tables at the bog and fen sites T_{water} readings were taken rather than T_{soil}. Readings were taken directly after chamber measurements were completed in order to prevent disturbance while CO_2 was being recorded. At the bog meteorological tower, a tipping bucket rain gauge (0.2 m diameter, 0.35 m height) (Jarek Manufacturing Ltd., Canada), calibrated to 0.25 mm per tip, recorded total half-hourly P.

Hydrological conditions: snow depth (SD), snow water equivalent (SWE), frost table depth (FT), water table depth (WT), precipitation (P), and soil moisture (VMC) were also measured at each site simultaneously with the CO_2 flux measurements. Snow depth and SWE were measured with an MSC fibreglass snow tube beside the collar. Once the ground surface at a measurement point became snow-free, the depth to the frost table was measured at the same point. The frost table depth was measured with a graduated 1.2 m steel rod that was driven into the ground to the depth of refusal. Frost table depths greater than 1.2 m were not measured. Water table depth was measured daily, approximately 0.25 m in front of the chamber, after flux measurements were completed. A shovel was used to cut into the ground surface and pull back the soil. The water was then given time to settle after initial disturbance before being measured as a depth below the surface. The point of measurement was not done at the same point as frost table depth in order to prevent disturbance to the frost table measurements, through physical disturbance to the soil profile or by increased exposure at the measurement point for radiation permeation. VMC was measured beside the chamber with a Hydrosense soil water measurement system, Model CD620 (Campbell Scientific, Canada). Inserted into the surface vertically between the frost table and water table measurement points, these measurements represent the integrated soil moisture in the 0-20 cm layer directly below the surface. Due to the limitation of the probe length, once frost table depths were greater than 20 cm all soil moisture measurements only account for the top 20 cm portion of the thawed profile. Vegetation surveys were conducted for each site on August 24^{th} and 25^{th}. Vegetation identification and percent coverage was documented within each collar. Soil cores were sampled on August 24^{th}: 6 in total, 2 for each collar. Cores were inserted into the ground on either side of the collar at a depth of 50 cm and removed for sampling. As a result of open water and saturated conditions at the bog and fen sites, soil core samples of similar depth were compacted and not viable for sampling. Therefore, samples recovered from the fen and bogs were much shallower in depth than those from the permafrost plateau, capturing only the first 20 cm below the surface. At the bog sites, the lack of unsaturated material at the surface within the collar resulted in core samples taken from beside the collar at the nearest representative ground vegetation that was located above the water table and easier to extract.

The cores were cut horizontally into 5 cm sub-samples, each with a volume of approximately 210 cm^3. The bulk density (ρ_b), porosity (Φ), specific yield (Φ_d), degree of decomposition, organic matter content (SOM), and C:N ratios of each sub-sample were measured or computed. Bulk density measured the mass of soil per unit volume in addition to available pore space, and was computed from:

$$Bulk\ Density = \frac{Weight\ of\ Oven\ Dried\ Sample\ (g)}{Volume\ of\ Sample\ (cm^3)} \tag{3.3}$$

The total porosity measured the amount of space in the soil occupied by air and water and was computed from:

$$\%\,Porosity = \frac{Saturated\ Mass\ (g) - \ Dry\ Mass\ (g)}{Volume\ (cm^3)} \times 100$$

(3.4)

Specific yield measured the drainable porosity under the force of gravity after saturation of a soil and was derived from:

$$Specific\ Yield = \frac{Saturated\ Mass\ (g) - \ Drained\ Mass\ (g)}{Saturated\ Mass\ (g)}$$

(3.5)

The VonPost humification scale is the qualitative measure of decomposition for a soil, which was used to classify each sub-sample with a degree of decomposition. The amount of organic matter in each sub-sample was measured by loss on ignition (Konen *et al.*, 2002) from:

$$\%\,Loss\ on\ Ignition \ = \frac{[(W_{cso} - W_c) - (W_{csi} - W_c)]}{(W_{cso} - W_c)} \times 100$$

(3.6)

where W_c is the weight of the crucible (g), W_{cso} is the weight of the oven dried soil in the crucible (g), and W_{csi} is the weight of the remaining (inorganic) soil after combustion (g). For C:N ratios, the samples were oven dried at 100°C for 24 hours and ground in a tumbling ball mill for 2-5 minutes until homogenized into a powder for sampling.

Percentage of total carbon (TC) and nitrogen (TN) was determined through combustion utilizing an Isochrom – elemental analysis, Carlo – Erba Isotope Ratio Mass Spectrometry, autocombustion carbon – nitrogen analyzer (Micromass UK, Ltd., Environmental Isotope Laboratory, Dept. of Earth Sciences, University of Waterloo, Waterloo, Ontario, Canada).

3.4 Relationship Between Gas Flux and Environmental Factors

The first relationship examined was derived for GEP, to determine its dependence on incident PAR. The relationship was fitted empirically using an exponential model (Gomes *et al.*, 2006; Goudriaan, 1979):

$$GEP = \{GP_{max}[1 - exp(-\alpha \times PAR)]\} \tag{3.7}$$

where α is the initial slope of GEP versus PAR, PAR is the measured PAR (μmol m^{-2} sec^{-1}), and GP_{max} is the empirically derived gross photosynthetic exchange of CO_2. GEP values were converted to positive values for the ease of computation, in order to use the exponential model.

To determine the environmental controls that play a role in total respiration, relationships between R_{tot} and T_a, T_{soil}, T_{water}, VMC, snow depth, frost table depth, and water table depth were also examined. Due to the water table conditions of the bog and fen sites soil temperature, VMC, and water table depth were only monitored at the permafrost plateau sites. Therefore, a comparison between landscape units, in regards to the relationship between R_{tot} and T_{soil} and VMC, cannot be made. However, these relationships are still examined for the permafrost plateau to understand the variability within an evolving plateau. In addition, the frost table depth at the fen remained immeasurable (>1.2 m) for the duration of the study so variability was not captured at any of the channel fen sites in relation to frost table depth.

Since this study collected instantaneous midday fluxes, the relationships that were determined were then applied to the continuous climate data for this site during the 2008 season of study. Using half hourly air temperature and PAR averages from the bog and plateau MET towers, GEP, R_{tot} and NEE were modeled for each landscape unit. The bog MET tower data was used for both the fen and bog landscapes due to the similarities of canopy conditions and their close proximity. This modeled gas flux was then summed to determine total seasonal flux for each landscape unit. These values were then scaled to a portion of the Scotty Creek basin based on the ground classification conducted by Quinton *et al.* (2009).

3.5 Statistical Analysis

The standard error was used to define the uncertainty and magnitude of error in the reported mean for this study. While standard deviation captures the dispersion of the data, standard error captures the potential of sampling error, therefore describing the confidence of the mean rather than the variability. To model the relationship between R_{tot} and air temperature, most studies implement an exponential model to best describe the relationship (Fang and Moncrieff, 2001). This also applies to the relationship between R_{tot} and soil temperature, based on the assumption that microbial activity increases at an accelerated rate as temperature increases. As a result, an exponential model depicts this reaction best rather than a linear model. For the relationship between R_{tot} and hydrological conditions of each landscape, the uses of a linear or quadratic model are most suited. Linear models are more accurate when measurements are made in either "wet" or "dry" conditions (Simek *et al.*, 2004). However, if the range of conditions during measurement occurs during both wet and dry conditions, a quadratic model is then best suited to depict the relationship (Davidson *et al.*, 1998). Since this study conducted measurement during a range of conditions from snowmelt to late-green season, the quadratic model was used to represent the relationship between R_{tot} and VMC as well as water table depth, snow depth and frost table depth.

Chapter 4 Results: Temporal Variability in CO_2 Flux and its Environmental Controls

4.1 Relative Variability in CO_2 Flux

Daily instantaneous flux measurements of CO_2 were taken at the nine locations with both clear and dark chambers. Time-series plots were used to show the average instantaneous gas flux of the three sites for each landscape unit over the duration of study (Figure 4-1). The permafrost plateau showed distinct trends for R_{tot}, NEE and GEP. R_{tot} fluctuated between 0 and 0.02 mg CO_2 m^{-2} sec^{-1} during the latter half of April and first week of May, increasing for the remaining duration of study. GEP remained low until late-May at which point it began to increase (negatively) and show greater variability. NEE was similar to R_{tot}, demonstrating that average instantaneous midday gas flux for the three sites on the plateau were sources of CO_2 to the atmosphere at the time of measurement. However, data collected on August 23rd responded as a sink for CO_2 at that particular time of measurement, suggesting a decline in NEE later in the summer. The channel fen exhibited a similar pattern to the permafrost plateau in its average instantaneous gas flux variability (Figure 4-1). R_{tot} steadily increased for the duration of the study; however, GEP increased (negatively) at a greater rate later in the growing season, which corresponded to a lower rate of NEE than that found at the plateau sites. These declining rates of NEE signified that the average instantaneous midday flux at the fen sites started as sources and became sinks for CO_2 at the time of measurement as the growing season progressed.

The average instantaneous gas flux for the flat bog did not show any distinct trends for R_{tot}, NEE and GEP (Figure 4-1). R_{tot} fluctuated below 0.02 mg CO_2 m^{-2} sec^{-1} during the latter half of April and first week of May, with only a slight increase for the remainder of the season. GEP remained almost negligible, resulting in an NEE flux similar to R_{tot}, which signifies the average instantaneous midday gas flux for the three sites on the bog were sources of CO_2 to the atmosphere at the time of measurement. Figure 4-1 clearly shows that there was some temporal and spatial variability between the three landscapes and their seasonal development.

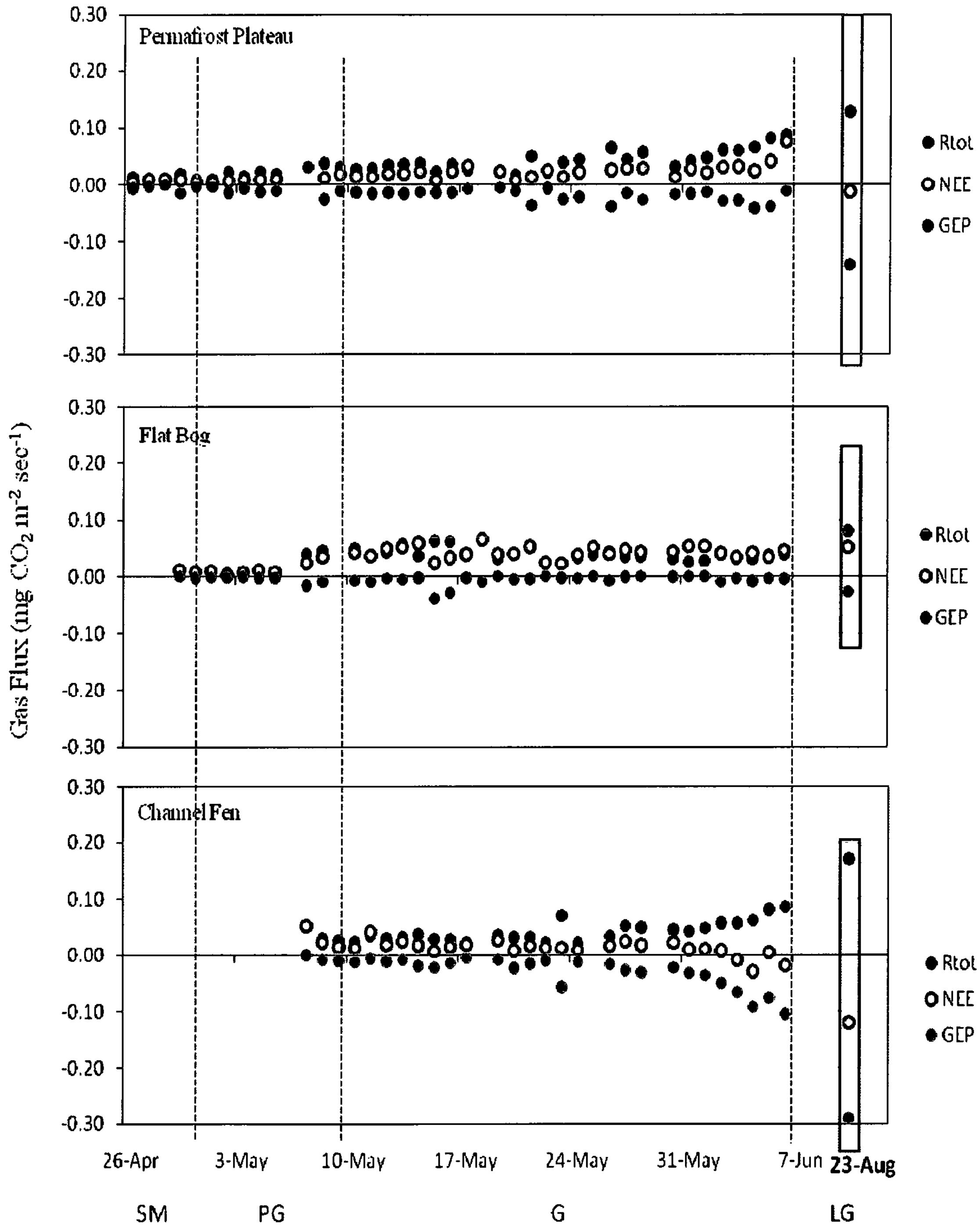

Figure 4-1 Instantaneous average gas flux of total respiration (R_{tot}), net ecosystem exchange (NEE), and gross ecosystem production (GEP) for each landscape unit during snowmelt (SM), pre-green (PG), green (G), and late-green (LG) (defined and separated by red lines) of the 2008 field season at Scotty Creek, Northwest Territories, Canada. Positive values indicate the emission of CO_2 and negative values indicate the ecosystem's uptake of CO_2.

4.2 Temporal Variability in CO_2 Flux

To depict the temporal variability within and between landscapes, the average gas flux was also computed for each of the four seasonal periods. For each season and landscape unit the average R_{tot}, NEE, and GEP was compared (Figure 4-2). The maximum, minimum, and average point measurements for each landscape unit during each season are presented in Table 4-1. These values represent an instantaneous CO_2 flux during midday and do not take into account fluctuations throughout the remainder of the day or night, and therefore cannot be extrapolated to average daily values. However, by averaging within periods a representative range of environmental conditions and controls were captured which can be used to determine functional relationships.

4.2.1 Permafrost Plateau

There was a steady increase in R_{tot} between snowmelt and late green for all three sites (Figure 4-2 and Table 4-1). Snowmelt had the lowest average R_{tot} of 0.011 ± 0.003 mg CO_2 m^{-2} sec^{-1} with fluxes ranging from 0.001 to 0.047 mg CO_2 m^{-2} sec^{-1}, while late-green had the highest average R_{tot} of 0.128 ± 0.019 mg CO_2 m^{-2} sec^{-1} with fluxes ranging from 0.103 to 0.165 mg CO_2 m^{-2} sec^{-1}. NEE fluctuated between snowmelt and late-green, due to a strong CO_2 uptake during pre-green and late-green periods for PPC-1, in comparison to the other two plateau sites. As a result, pre-green had an average uptake of -0.037 ± 0.047 mg CO_2 m^{-2} sec^{-1} with fluxes ranging from -1.201 to 0.034 mg CO_2 m^{-2} sec^{-1}, while the green period had the highest average emission of 0.023 ± 0.003 mg CO_2 m^{-2} sec^{-1} with fluxes ranging from -0.156 to 0.108 mg CO_2 m^{-2} sec^{-1}. There was a small increase in GEP between snowmelt and late-green. Again snowmelt had the lowest average GEP of -0.006 ± 0.004 mg CO_2 m^{-2} sec^{-1} with fluxes ranging from -0.002 to -0.045 mg CO_2 m^{-2} sec^{-1}, while late-green had the highest average GEP of -0.141 ± 0.067 mg CO_2 m^{-2} sec^{-1} with fluxes ranging from -0.065 to -0.275 mg CO_2 m^{-2} sec^{-1}.

4.2.2 Flat Bog

There was a steady increase in R_{tot} between snowmelt and late-green for all three sites (Figure 4-2 and Table 4-1). There was a small fluctuation during the pre-green period with BC-3 experiencing CO_2 uptake. As a result, pre-green had the lowest average R_{tot} of -0.010 ± 0.040 mg CO_2 m^{-2} sec^{-1} with fluxes ranging from -0.964 to 0.106 mg CO_2 m^{-2} sec^{-1}, while late-green had the highest average R_{tot} of 0.081 ± 0.014 mg CO_2 m^{-2} sec^{-1} with fluxes ranging from 0.056 to 0.106 mg CO_2 m^{-2} sec^{-1}. NEE steadily increased between snowmelt and late-green. Snowmelt had the lowest average emission of 0.009 ± 0.001 mg CO_2 m^{-2} sec^{-1} with fluxes ranging from 0.005 to 0.015 mg CO_2 m^{-2} sec^{-1}, while late-green had the highest average emission of 0.053 ± 0.008 mg CO_2 m^{-2} sec^{-1} with fluxes ranging from 0.041 to 0.068 mg CO_2 m^{-2} sec^{-1}. There was a slight increase in GEP between snowmelt and late-green. Snowmelt and the green period had the lowest average GEP. Snowmelt had an average GEP of -0.002 ± 0.001 mg CO_2 m^{-2} sec^{-1} with fluxes ranging from -0.001 to -0.006 mg CO_2 m^{-2} sec^{-1}, and the green period had an average GEP of -0.006 ± 0.001 mg CO_2 m^{-2} sec^{-1} with fluxes ranging from -0.001 to -0.057 mg CO_2 m^{-2} sec^{-1}. Late-green had the highest average GEP of -0.028 ± 0.014 mg CO_2 m^{-2} sec^{-1} with fluxes ranging from -0.011 to -0.056 mg CO_2 m^{-2} sec^{-1}.

4.2.3 Channel Fen

Unlike the flat bog and permafrost plateau sites there were no flux measurements recorded during snowmelt for the channel fen. There was a small increase in R_{tot} between pre-green and green with a sudden and rapid increase during late-green for all three sites (Figure 4-2 and Table 4-1). Pre-green and green had the lowest average R_{tot}. Pre-green had an average R_{tot} of 0.030 ± 0.003 mg CO_2 m^{-2} sec^{-1} with fluxes ranging from 0.022 to 0.052 mg CO_2 m^{-2} sec^{-1}, and the green period had an average R_{tot} of 0.044 ± 0.003 mg CO_2 m^{-2} sec^{-1} with fluxes ranging from 0.007 to 0.155 mg CO_2 m^{-2} sec^{-1}. Late-green had the highest average R_{tot} of 0.171 ± 0.027 mg CO_2 m^{-2} sec^{-1} with fluxes ranging from 0.120 to 0.214 mg CO_2 m^{-2} sec^{-1}. NEE slightly decreased between pre-green and green, and then rapidly increased in CO_2 uptake during late-green. In this case, pre-green had the highest average emission of 0.020 ± 0.005 mg CO_2 m^{-2} sec^{-1} with fluxes ranging from -0.002 to 0.053 mg CO_2 m^{-2} sec^{-1}. While the green period had the lowest average emission of 0.012 ± 0.002 mg CO_2 m^{-2} sec^{-1} with fluxes ranging from -0.037 to 0.083 mg CO_2 m^{-2} sec^{-1}. In comparison, late-green had the highest, and only, average CO_2 uptake of -0.120 ± 0.038 mg CO_2 m^{-2} sec^{-1} with fluxes ranging from -0.172 to -0.046 mg CO_2 m^{-2} sec^{-1}. There was a strong increase in average GEP between pre-green and late-green. Pre-green had the lowest average GEP of -0.010 ± 0.003 mg CO_2 m^{-2} sec^{-1} with fluxes ranging from -0.005 to -0.025 mg CO_2 m^{-2} sec^{-1}, while late-green had the highest average GEP of -0.291 ± 0.030 mg CO_2 m^{-2} sec^{-1} with fluxes ranging from -0.259 to -0.351 mg CO_2 m^{-2} sec^{-1}.

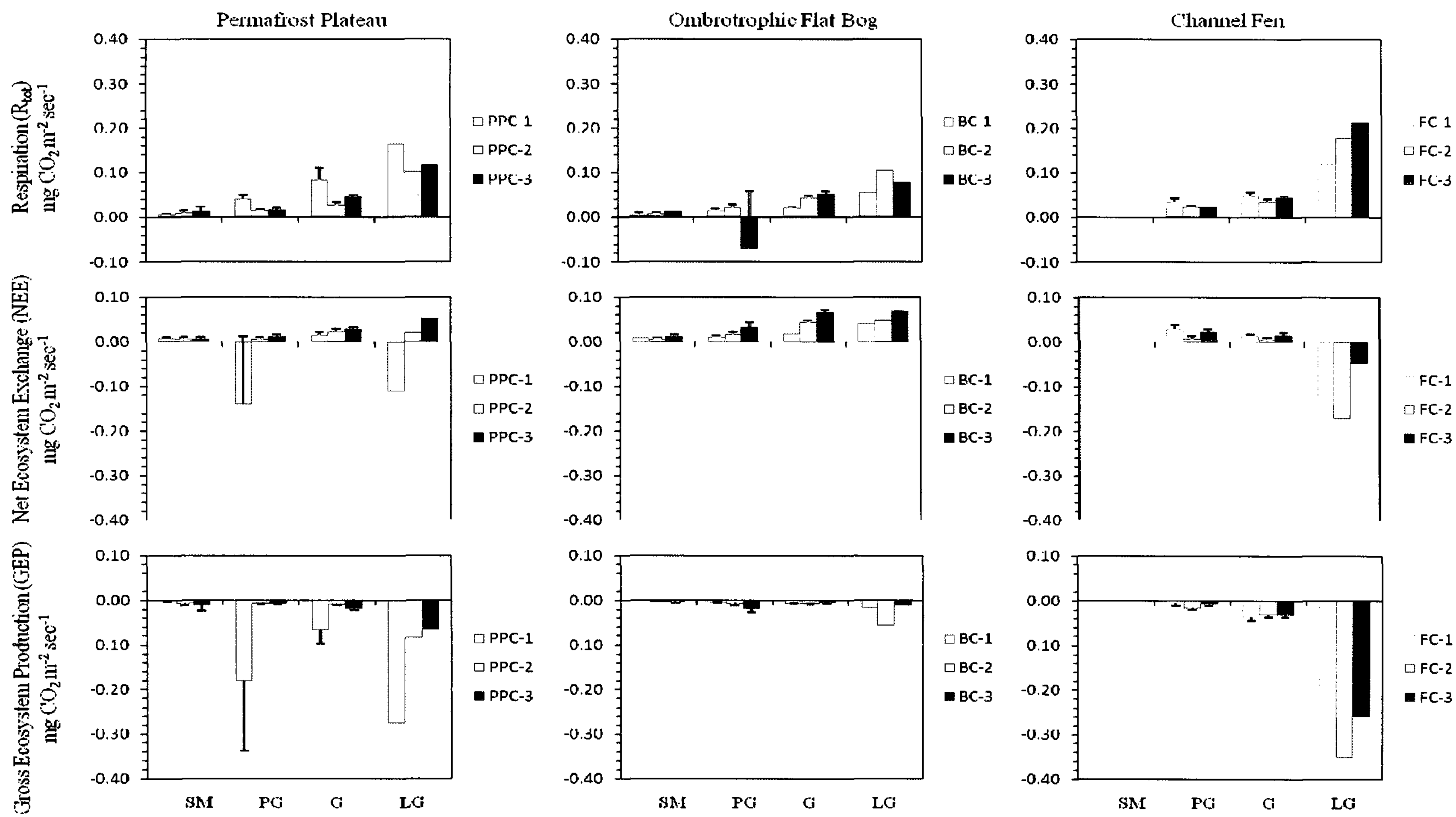

Figure 4-2 Average total respiration (R_{tot}), net ecosystem exchange (NEE), and gross ecosystem production (GEP) for the permafrost plateau, flat bog, and channel fen during snowmelt (SM), pre-green (PG), green (G), and late-green (LG) during 2008 at Scotty Creek, Northwest Territories, Canada. Positive values indicate the emission of CO_2 and negative flux values indicate the ecosystem's uptake of CO_2, error bars represent standard error.

Table 4-1 Minimum, maximum, and average measurements for total respiration (R_{tot}), net ecosystem exchange (NEE), and gross ecosystem production (GEP) in mg CO_2 m^{-2} sec^{-1} for the permafrost plateau, flat bog, and channel fen for the 2008 field season at Scotty Creek, Northwest Territories, Canada. Note: GEP is a negative value therefore minimum and maximum values are in opposition to the format used for R_{tot} and NEE.

Season	Landscape Unit	Total Respiration (R_{tot})			Net Ecosystem Exchange (NEE)			Gross Ecosystem Production (GEP)		
		Min.	Max.	Average	Min.	Max.	Average	Min.	Max.	Average
Snowmelt	Plateau	0.001	0.047	0.011±0.003	0.001	0.017	0.008±0.001	-0.002	-0.045	-0.006±0.004
	Bog	0.004	0.015	0.010±0.002	0.005	0.015	0.009±0.001	-0.001	-0.006	-0.002±0.001
	Fen	-	-	-	-	-	-	-	-	-
Pre-Green	Plateau	-0.001	0.072	0.024±0.004	-1.201	0.034	-0.037±0.047	-0.002	-1.266	-0.061±0.048
	Bog	-0.964	0.106	-0.010±0.040	0.001	0.072	0.020±0.004	-0.001	-0.089	-0.010±0.004
	Fen	0.022	0.052	0.030±0.003	-0.002	0.053	0.020±0.005	-0.005	-0.025	-0.010±0.003
Green	Plateau	0.002	0.628	0.053±0.009	-0.156	0.108	0.023±0.003	-0.001	-0.784	-0.031±0.011
	Bog	0.006	0.155	0.040±0.002	0.005	0.130	0.043±0.003	-0.001	-0.057	-0.006±0.001
	Fen	0.007	0.155	0.044±0.003	-0.037	0.083	0.012±0.002	-0.001	-0.146	-0.033±0.004
Late-Green	Plateau	0.103	0.165	0.128±0.019	-0.110	0.052	-0.013±0.050	-0.065	-0.275	-0.141±0.067
	Bog	0.056	0.106	0.081±0.014	0.041	0.068	0.053±0.008	-0.011	-0.056	-0.028±0.014
	Fen	0.120	0.214	0.171±0.027	-0.172	-0.046	-0.120±0.038	-0.259	-0.351	-0.291±0.030

4.3 Temporal Variability of Environmental Controls on CO_2 Flux

To compare the environmental factors with the CO_2 flux demonstrated in each landscape unit, they were divided into the same four seasonal periods (Table 3-1). Due to the proximity of the three landscape units examined and the lack of a substantial canopy on the plateau sites, there was little to no spatial variability among the landscape units in regards to T_{air}, RH and P (Figure 4-3). At the start of the snowmelt season the average daily T_{air} was 2 °C. However, as the winter season ended, the average daily T_{air} steadily increased, peaking in early to mid-July at 23.9 °C. As the growing season progressed temperatures remained warm, cooling off in mid-August during the late-green period. RH fluctuated throughout all the seasons but peaked in mid-May at 95 %. This was however, quickly followed by a sudden decrease in RH at the end of May to 38.5 %. Periods of low RH that occurred during early May, June, and July were dry with very little P. The other environmental factors examined during this study did vary among landscape units, and are therefore discussed independently in the following sections. In comparison to the climatic conditions described above these factors were measured discretely, for the duration and time at which chamber measurements occurred. They represent conditions at the time of measurement rather than a continuous record averaged for each day.

4.3.1 Permafrost plateau

PAR fluctuated throughout the seasons but there was an overall trend of increasing PAR measured at each chamber from snowmelt into the green season (Figure 4-4). PAR values peaked at the end of May at 1536 $\mu mol\ m^{-2}\ sec^{-1}$, and then decreased in late August to 685 $\mu mol\ m^{-2}\ sec^{-1}$ on August 23rd. Several of the fluctuations that represented large decreases in measured PAR appeared to coincide with large P events. Snow and soil temperatures were low during the snowmelt season; however, once the ground surface was exposed the temperature steadily increased, peaking in early May at 7 ºC (Figure 4-4). In late August soil temperatures were significantly higher with a value of 10.4 ºC on August 23rd.

As depicted in Figure 4-5, snowmelt measurement began in mid-April when the snowpack depth was 84.5 cm (SWE = 172 mm) and continued daily until the pack disappeared in the first week of May. The peak snow water equivalent of 205 mm was measured on April 20th. The frost table depth steadily increased as the pre-green and early green seasons progressed. By August 23rd the average frost table depth was 105 cm. As the sub-surface thawed, the frost table depth was shallow and the initial soil profile completely saturated, in some cases with ponded water as the water released during snowmelt could not infiltrate the soil. As the frost table deepened the average water table dropped steadily below the surface reaching 27 cm and on August 23rd dropping to 40 cm. During the green season soil moisture steadily increased, with occasional fluctuations, starting from 35 % and reaching 84 % on June 1st. By August 23rd soil moisture had decreased slightly coming down to 68 %.

The first 20 cm of the soil profile below the ground surface had an average organic content of 96.5%. The average percentage of carbon was 44.7 %, while nitrogen was low at 0.8 %, resulting in a C:N ratio of 69.6. When the entire 50 cm core was examined it showed an increasing state of decomposition, increasing bulk density, and lower specific yield with depth. Porosity on the other hand remained fairly consistent with very little change with depth (Table 4-2).

4.3.2 Flat Bog

PAR fluctuated throughout the seasons with a slight trend of increasing photosynthetic activity from snowmelt to the late-green season (Figure 4-4). PAR values peaked on June 1st at 1566 μmol m^{-2} sec^{-1}, and then decreased in late August to 936 μmol m^{-2} sec^{-1} on August 23rd. Similar to the plateau, several of the fluctuations that represented large decreases in measured PAR appeared to coincide with large P events. Open water conditions at the bog sites resulted in chamber collars submerged in water rather than solid ground; therefore, T_{water} measurements were more relevant to these conditions in comparison to the T_{soil} recorded on the plateau sites. Snow and water temperatures were low during the snowmelt season; however, once the snow cover diminished temperatures rapidly increased, peaking in early June at 17 °C (Figure 4-4). In late August water temperatures remained high with a value of 14.5 °C on August 23rd.

As depicted in Figure 4-5, snowmelt measurement began in mid-April when the snowpack depth was 64 cm (SWE = 120 mm) and continued daily until the pack disappeared in the first week of May. The peak snow water equivalent of 179 mm was measured on April 18th. The bog sites experienced an above ground water table once the snowpack disappeared. As a result, the monitoring of the frost table depth was interpreted at these sites as the depth to the ice layer below the water surface. Under this definition the frost table depth rapidly increased as the pre-green season progressed. By May 9th the ice had completely disappeared and the average frost table depth was over 120 cm.

The average soil profile for the first 20 cm below the water surface had an organic content of 97 %. The average percentage of carbon was 47.4 %, while nitrogen was low at 0.9 %, resulting in a C:N ratio of 52.4. When examined, the cores showed an increasing state of decomposition, increasing bulk density and a slight decrease in porosity and specific yield with depth (Table 4-2).

4.3.3 Channel Fen

The channel fen experienced similar conditions to the flat bog. PAR fluctuated throughout the seasons initially with a trend of decreasing PAR during snowmelt into the first few weeks of the green season, and increasing for the remainder of the study period (Figure 4-4). PAR values peaked on May 27th at 1621 $\mu mol\ m^{-2}\ sec^{-1}$, and then decreased in late August to 922 $\mu mol\ m^{-2}\ sec^{-1}$ on August 23rd. Similar to the bog, open water conditions at the fen sites resulted in chamber collars submerged in water rather than solid ground; therefore, T_{water} measurements were more relevant to these conditions. Snow and water temperatures were low during the pre-green season; however once the snow cover diminished temperatures rapidly increased, peaking in late May at 17 °C (Figure 4-4). In late August water temperatures remained high with a value of 13 °C on August 23rd.

As depicted in Figure 4-5, snowmelt measurement began in mid-April when the snowpack depth was 52 cm (SWE = 120 mm) and continued daily until the pack disappeared at the end of April. The peak snow water equivalent of 148 mm was measured on April 18th. Similar to the bog, the frost table represented the depth of ice below the water surface. Under this definition the frost table rapidly increased as the pre-green season progressed. By May 8th the ice had completely disappeared and the average frost table depth was over 120 cm.

The average soil profile for the first 20 cm below the water surface had an organic content of 94 %. The average percentage of carbon was 48.7 %, while nitrogen was low at 1.2 %, resulting in a C:N ratio of 43.4. When examined, the cores showed an increasing state of decomposition, a slight increase in bulk density, a slight decrease in specific yield, and a variable porosity with depth below the surface (Table 4-2).

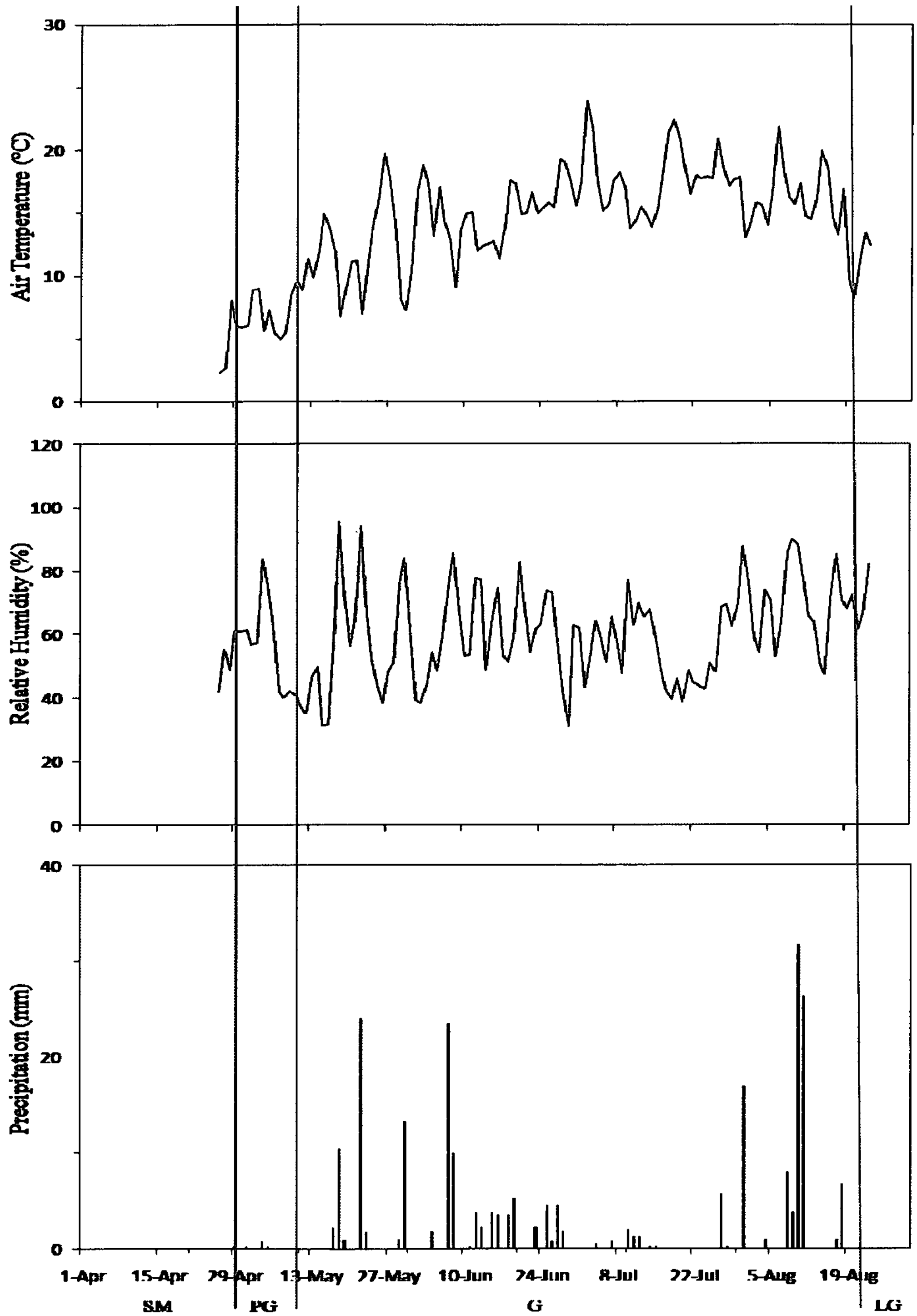

Figure 4-3 Average air temperature (°C), relative humidity (%), precipitation (mm) for all three landscape units (permafrost plateau, flat bog, and channel fen) during snowmelt (SM), pre-green (PG), green (G), and late-green (LG) (defined and separated by red lines) for the 2008 field season at Scotty Creek, Northwest Territories, Canada.

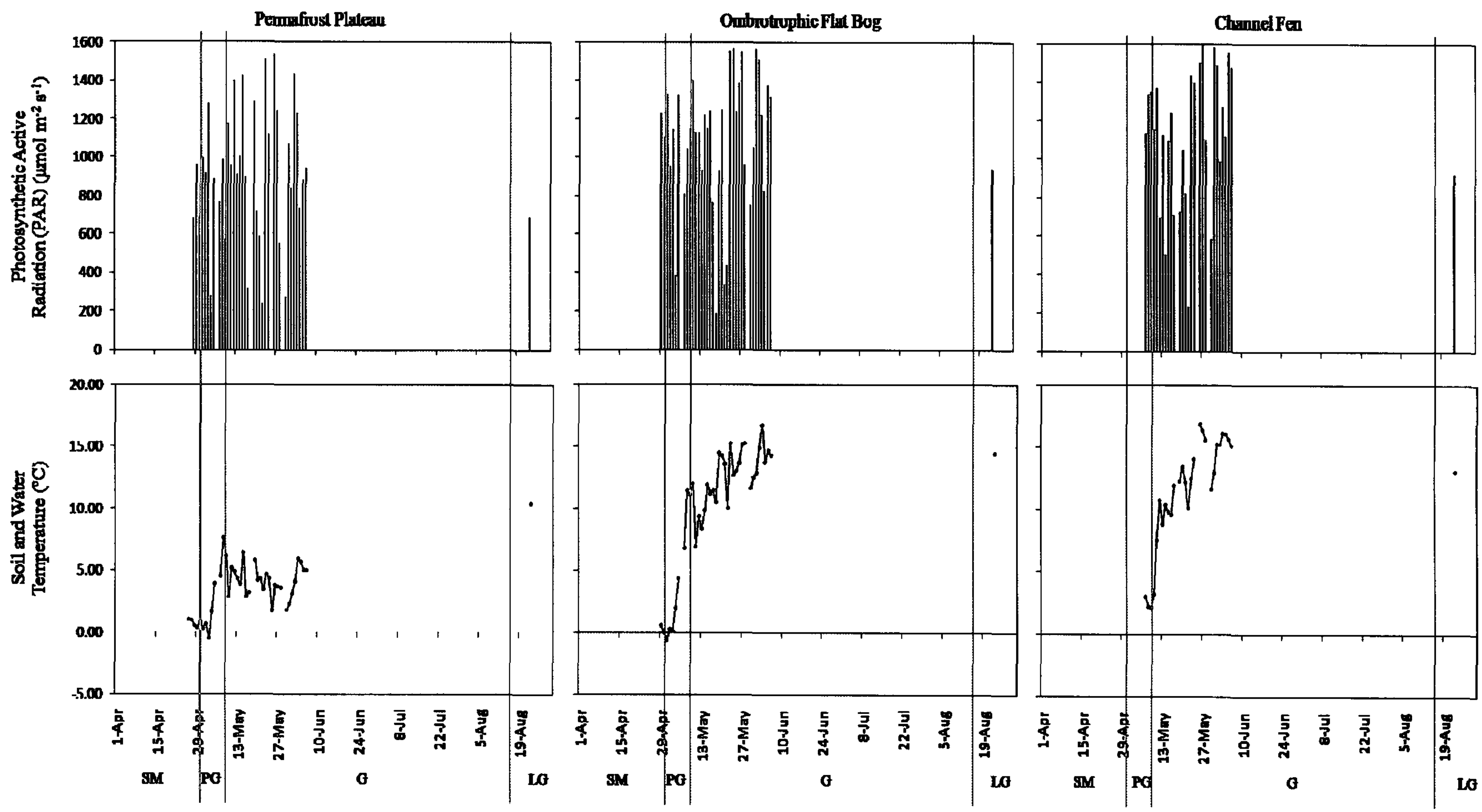

Figure 4-4 Average photosynthetically active radiation (PAR), and soil and water temperature (°C) for each landscape unit during snowmelt (SM), pre-green (PG), green (G), and late-green (LG) (defined and separated by red lines) for the 2008 field season at Scotty Creek, Northwest Territories, Canada. T_{soil} values are shown for the permafrost plateau, while T_{water} values are shown for the bog and fen.

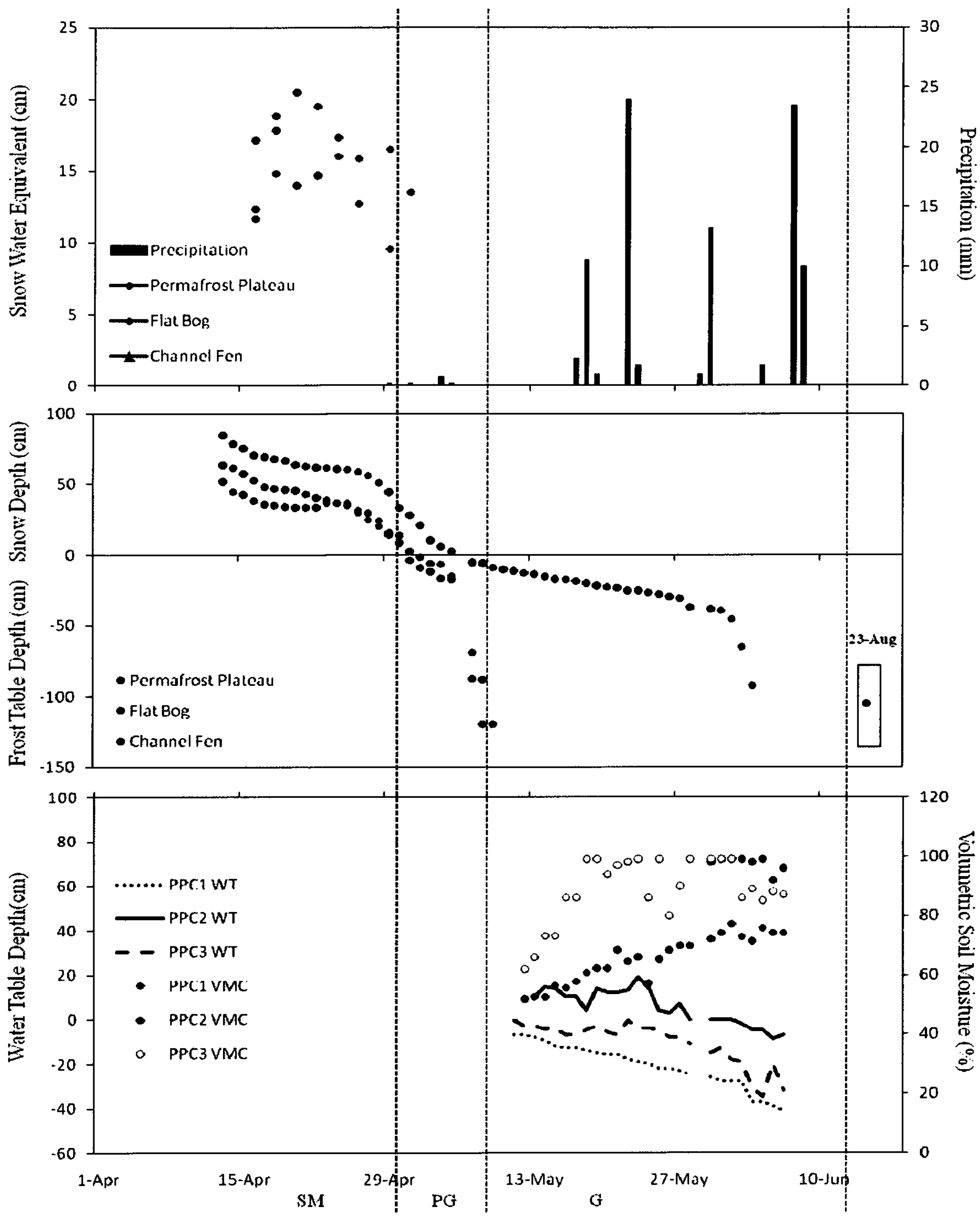

Figure 4-5 Average snow water equivalent (cm), precipitation (mm), snow depth (cm) (> 0 cm), frost table depth (cm) (< 0 cm), water table depth (cm), and soil moisture (%) for each landscape unit during snowmelt (SM), pre-green (PG), green (G), and late-green (LG) (defined and separated by red lines) for 2008 at Scotty Creek, Northwest Territories, Canada. VMC values and water table (WT) depths are only available for the permafrost plateau due to open water conditions in the bog and fen.

Table 4-2 Soil characteristics by depth for each landscape unit. All measurements are landscape unit averages based on individual cores taken at each site within each landscape unit. Measurements of LOI, TC, TN, and C:N were taken for only the first 20 cm below the surface at each site and averaged for each landscape unit. Refer to **Figure 3-1** for site locations.

Landscape unit	Depth (cm)	ρ_b (g/cm^3)	Φ (%)	Φ_d (%)	VonPost	LOI (%)	TC (%)	TN (%)	C:N
Permafrost Plateau	LFH	0.04	-	-	-	96.53	44.68	0.80	69.63
	5-10	0.08	78.68	0.24	H1-H2				
	10-15	0.15	81.87	0.17	H2				
	15-20	0.15	81.40	0.18	H2-H3				
	20-25	0.15	84.69	0.15	H3				
	25-30	0.15	83.10	0.17	H3				
	30-35	0.16	81.87	0.16	H3				
	35-40	0.21	82.36	0.09	H4				
	40-45	0.20	78.29	0.08	H5				
	45-50	0.20	81.05	0.09	H4				
Flat Bog	LFH	0.02	53.34	0.58	H1	97.19	47.36	0.91	52.42
	5-10	0.04	44.30	0.42	H2				
	10-15	0.04	29.51	0.23	H4				
	15-20	0.05	40.47	0.47	H4				
Channel Fen	LFH	0.03	32.81	0.46	H2	94.12	48.71	1.20	43.35
	5-10	0.04	30.70	0.36	H3				
	10-15	0.06	27.68	0.39	H4				
	15-20	0.03	38.68	0.36	H3				

Chapter 5 Results: Spatial Variability in CO_2 Flux and Environmental Controls

5.1 Spatial Variability in CO_2 Flux

Measuring CO_2 daily at nine locations with both clear and dark chambers, the average mid-day gas flux was computed for each of the four seasonal periods. Taking the average chamber R_{tot}, and GEP for each landscape unit, these values were plotted against PAR, T_a, T_{water}, T_{soil}, VMC, water table depth, snow depth, and frost table depth to establish functional relationships that can be used to extend the point flux measurements to a continuous dataset for each landscape in a given area. Understanding these relationships also aids in identifying the extent to which surrounding environmental conditions influence CO_2 gas flux. Taking the areal coverage of each landscape unit these results were then scaled to the ecosystem level. Ascertaining gas flux response at the larger scale is important in the prediction of future shifts between a net sink and net source of CO_2 for any ecosystem. For the Scotty Creek region, as permafrost plateaus shrink in size and number due to permafrost degradation and landscape subsidence, it is becoming increasingly crucial to classify and understand the difference between different landscape units as it relates to CO_2 gas flux.

5.1.1 Transect Topography

Variation in elevation between landscapes plays a significant role in the hydrology, and ultimately CO_2 flux, of a sub-arctic boreal wetland. It physically defines sub-surface and overland flow between and within each landscape. Topography of the ground surface influences the spatial distribution of soil moisture (Western *et al.*, 2001), therefore indirectly influencing thaw and rates of CO_2 exchange. While the fen and bog experienced fluctuating microtopography due to water level changes in much more unstable terrain environments, the plateau had a fixed ground surface. Vertical displacement between each plateau site was identified through measurements conducted along the length of each site's transect. The highest plateau site was PPC-1 with its collar located at 285 asl, followed by PPC-2 at 283 asl, and PPC-3 at 280 asl (Figure 5-1). Site PPC-1 was situated in the middle of a slight slope on the surrounding topography. It experienced the greatest drop in water table depth and driest soils of all three sites on the plateau. In addition to its sub-surface hydrology, it also had the most established vegetation cover and as a result the greatest rates of R_{tot} and was the only plateau site to experience NEE uptake during the study. The lowest plateau, PPC-3, was situated slightly lower than its surrounding topography. With moist soils and a relatively bare surface this site had a steady R_{tot} rate and NEE respiration throughout the study. Last of the three plateau sites, PPC-2 was situated in a depression. Lower than the surrounding topography this site experienced extended periods of flooding and high saturation in comparison to the other plateau sites. As a result, it had the lowest rates of R_{tot} and NEE respiration of all three sites.

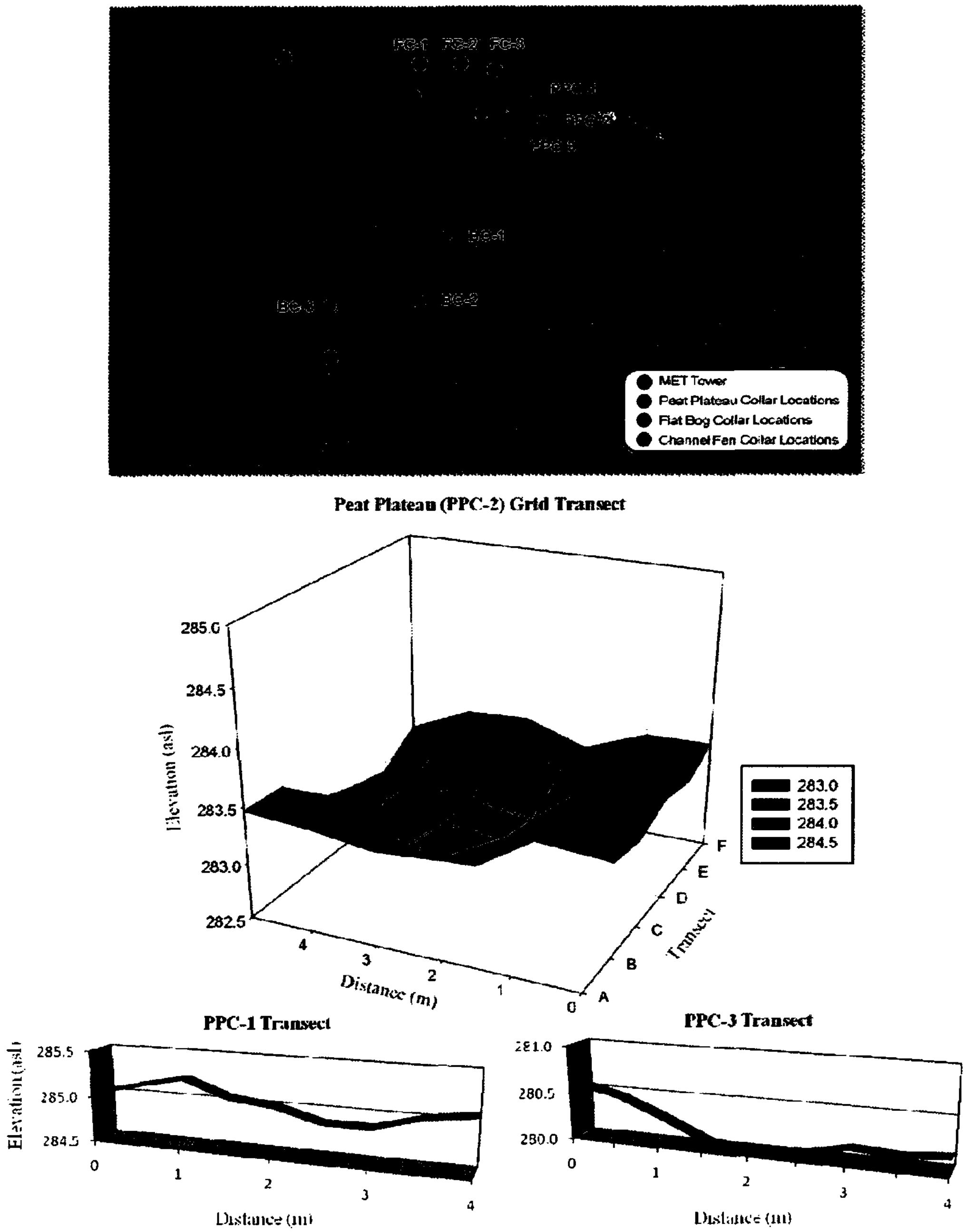

Figure 5-1 Elevation of peat plateau sites (m) above sea level during the 2008 field season at Scotty Creek, Northwest Territories, Canada. PPC-1 and PP-3 are 4 m long linear transects with measurement points at every 0.5 m and the chamber located at 2 m. PPC-2 was developed as a 5 m by 5 m grid (6 transects, 9 points along each) with measurement points at every 1 m and the chamber located in the center of the grid (Photo courtesy of William Quinton).

5.1.2 Relationship with Gross Ecosystem Production

The first relationship examined was derived for GEP, to determine its dependence on incident PAR. The relationship was fitted empirically using an exponential model (Gomes *et al.*, 2006; Goudriaan, 1979) (Figure 5-2). GEP values were converted to positive values for the ease of computation, in order to use the exponential model. In addition, GEP values were averaged based on PAR given that the chamber data was point values in time and not simultaneous among collars, making time averaging invalid. The permafrost plateau had a much greater slope of increasing productivity with increasing light intensity than the channel fen and flat bog. GP_{max} was reached between 0.02 and 0.03 mg CO_2 m^{-2} sec^{-1}, at PAR values recorded around 500 µmol m^{-2} sec^{-1} ($r^2 = 0.3$). The flat bog and channel fen experienced more gradual slopes for increasing productivity with increasing PAR. The flat bog does not reach a GP_{max} within the constraints of the dataset but is predicted to occur at approximately 0.06 mg CO_2 m^{-2} sec^{-1} ($r^2 = 0.4$). The channel fen also does not show GP_{max} but is predicted to occur at approximately 0.04 mg CO_2 m^{-2} sec^{-1} ($r^2 = 0.4$). Each landscape unit received similar light levels but displayed considerable scatter in GEP response, which indicates the possibility that other environmental factors may be contributing to the variation in flux response between landscape units as seen in Figure 5-2.

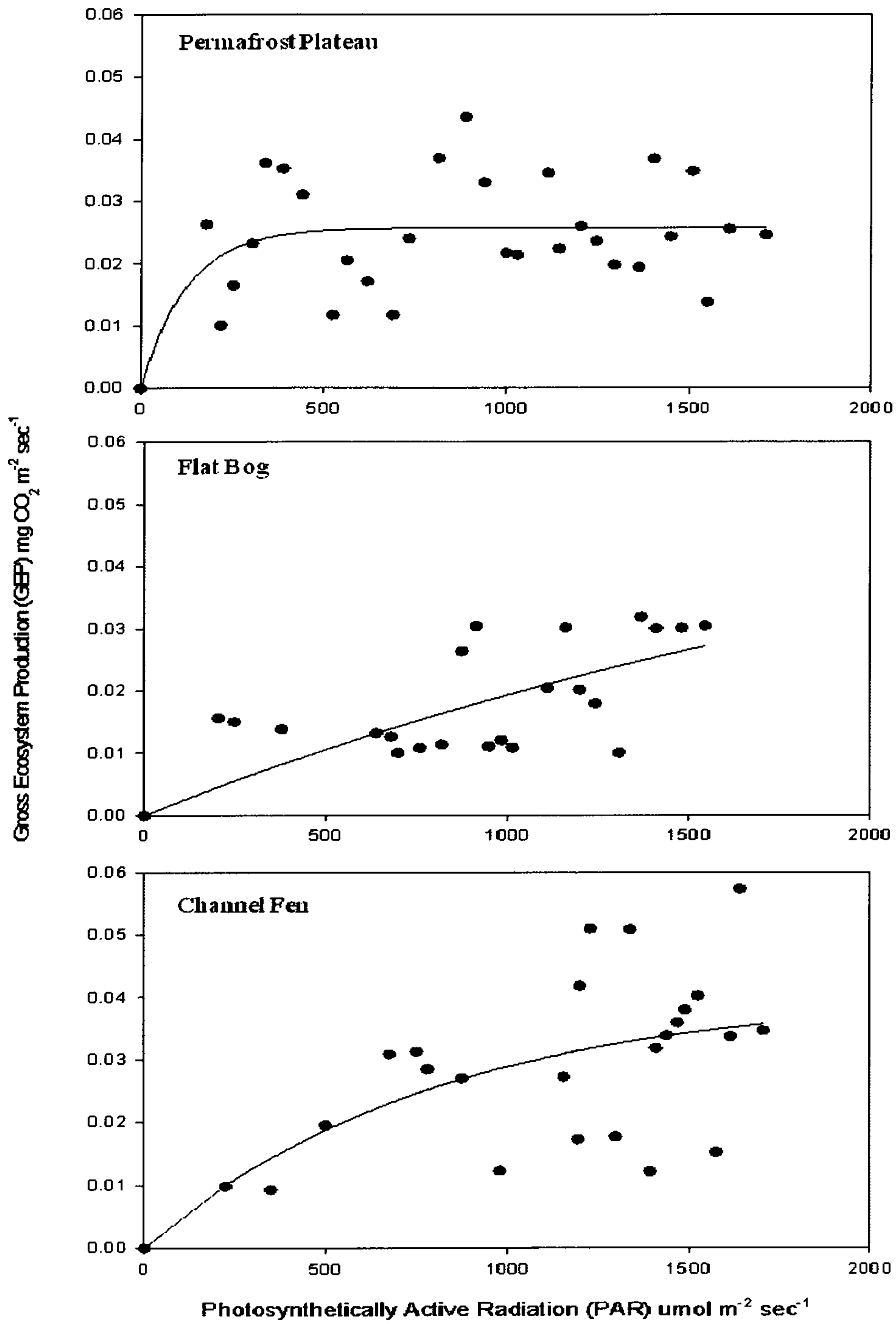

Figure 5-2 Relationship between gross ecosystem production (GEP) of CO_2 and photosynthetically active radiation (PAR) for 2008 at Scotty Creek, Northwest Territories, Canada. Showing the variation between landscape units: permafrost plateau (n=104, r^2=0.3), flat bog (n=110, r^2=0.4), and channel fen (n=84, r^2=0.4). The fitted curve for GEP versus PAR is an exponential model from Eq. (3.7). Symbols denote GEP bin-averages based on PAR intervals of 10 μmol m^{-2} sec^{-1}.

5.1.3 Relationships with Total Respiration

The relationship between R_{tot}, T_a, T_{soil}, and T_{water} examines the influence that temperature can have upon the surface and sub-surface processes of respiration (Figure 5-3). R_{tot} values were averaged based on changes in temperature given that the chamber data were point values in time and not simultaneous among collars, making time averaging invalid. The permafrost plateau had the strongest positive correlation between R_{tot} and T_{air} (r^2 = 0.5). While the flat bog and channel fen displayed a more moderate positive correlation in comparison (r^2 = 0.3, r^2 = 0.4, respectively), the fen appeared to have similar responses (and in some instances stronger) to changes in T_{air} between 10 – 20 °C. The variability between landscapes and the lower response by both the bog and fen to temperature changes suggested that there were also influenced by additional environmental variables.

Despite being unable to compare the relationship between R_{tot}, T_{soil}, and T_{water} among the three landscape units, these relationships depict other environmental variables that could be playing a role in controlling rates of total respiration within specific landscapes. Measured only on the permafrost plateau, R_{tot} showed a strong exponential correlation with T_{soil} (r^2 = 0.5). Surface water temperatures were measured in the flat bog and channel fen where high and fluctuating water tables dominated. Water controls heat diffusion in peat (Roulet *et al.*, 1992); therefore T_{water} was considered a reasonable a surrogate for T_{soil}. Both the bog and fen showed moderate to strong relationships between T_{water} and R_{tot} (r^2 = 0.3, r^2 = 0.3, respectively). The channel fen appeared to be more responsive with R_{tot} rates than the flat bog at lower temperatures, while the bog displayed more variability at greater temperatures.

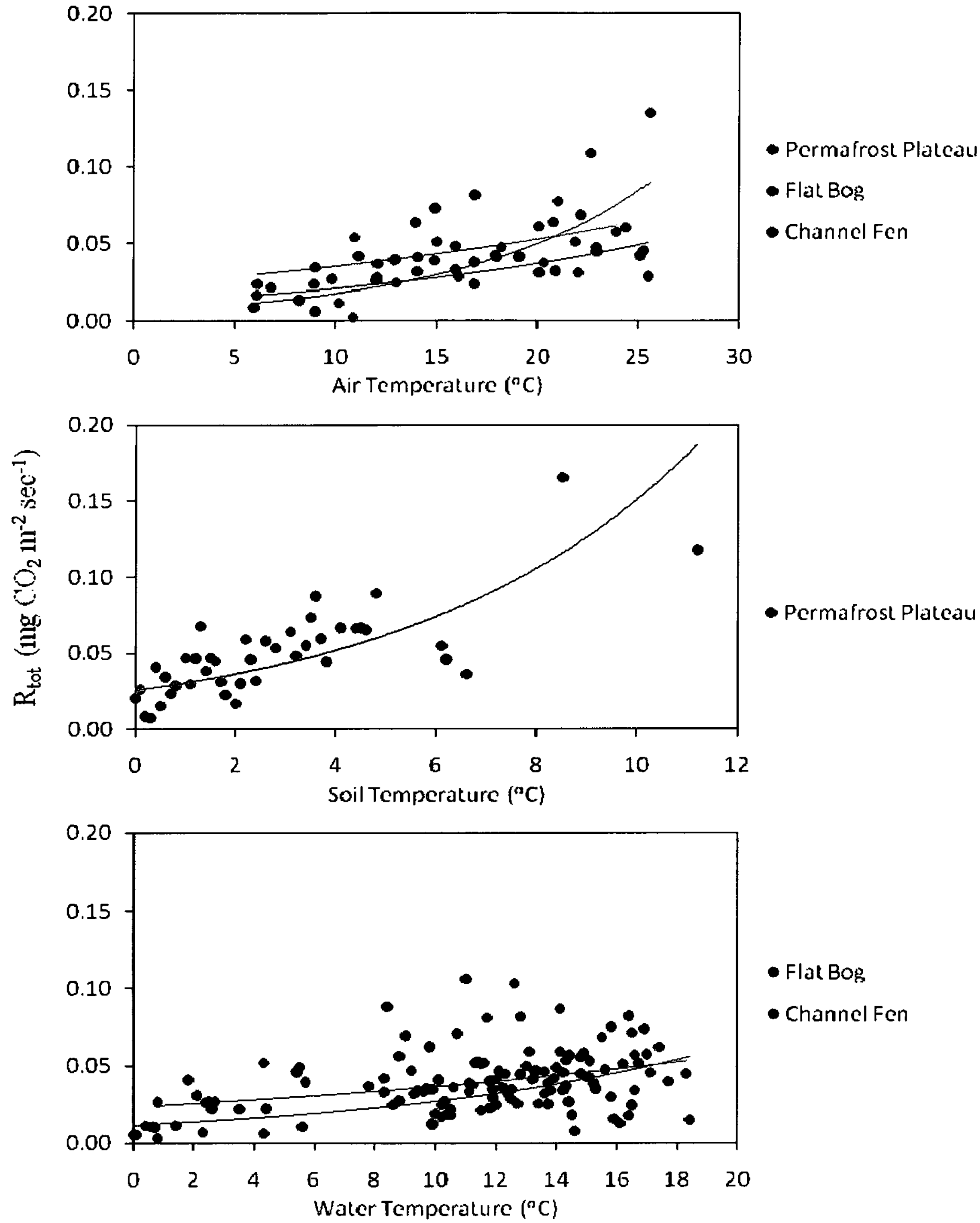

Figure 5-3 Relationship between total respiration (R_{tot}) and temperature for each landscape unit: permafrost plateau (n=104), flat bog (n=110), and channel fen (n=84). Symbols denote R_{tot} bin-averages based on 1 °C intervals for air temperature and 0.1 °C intervals for soil and water temperature. All data is from the 2008 field season at Scotty Creek, Northwest Territories, Canada.

The relationship between R_{tot}, snow depth, and frost table depth examines the influence that snow cover insulation and thaw depth can have upon the surface and sub-surface process of respiration (Figure 5-4). Measurement on the fen sites began on May 7th once the sites could be safely reached; however, by this time snow cover was gone and the frost table depth was greater than 120 cm below the surface and immeasurable. The permafrost plateau and flat bog on the other hand displayed moderate to strong relationships between R_{tot}, snow depth and frost table depth. Both landscapes showed consistently low rates (above zero) of R_{tot} for the duration of snow coverage; however, with increasing snowmelt and frost table depth R_{tot} began to steadily increase. In the flat bog frost table depths greater than 120 cm were reached early on in the season corresponding with a rapid increase in rates of R_{tot}.

The relationship between R_{tot}, soil moisture and water table depth examines the influence of hydrological conditions on the process of respiration (Figure 5-5). These conditions were only monitored at the permafrost plateau where the water table was below the surface for the majority of the study season. Water table depth showed a strong quadratic relationship with R_{tot} ($r^2 = 0.6$), which demonstrated that the dropping of the water table, allows soils and peat to aerate, increasing rates in R_{tot}. PPC-2 was the only plateau site to experience above surface flooding for a period of 26 days during May 2008. As can be seen in Figure 5-5 water table depths above the surface reflected in lower rates of R_{tot}. Soil moisture displayed no apparent relationship with R_{tot} on the permafrost plateau ($r^2 = 0.1$). There was a significant amount of variation between sites, making it difficult to discern any overall patterns for this landscape unit.

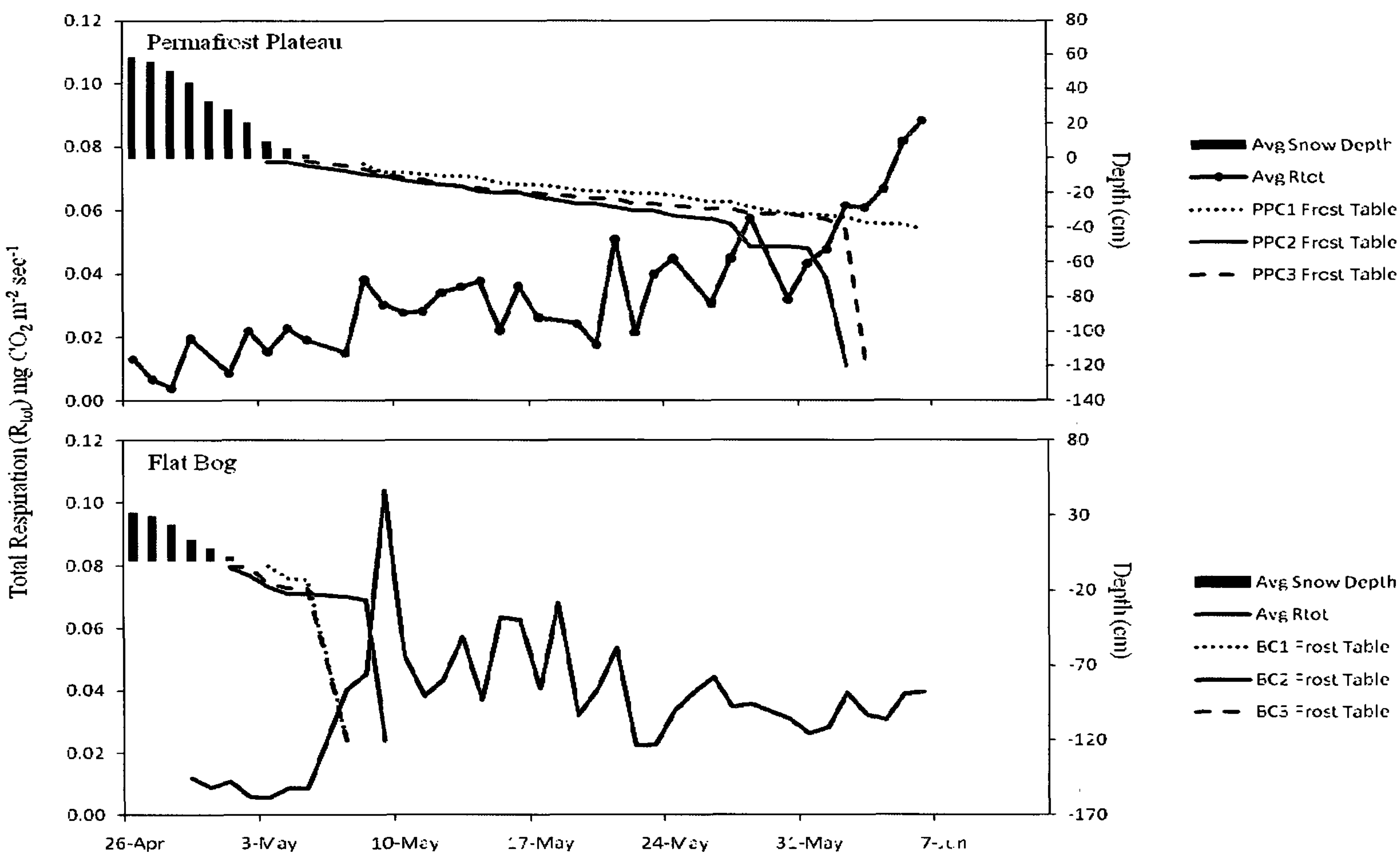

Figure 5-4 Relationship between total respiration (R_{tot}), snow depth, and frost table depth on the permafrost plateau (n=110), and the flat bog (n=114). All data is from the 2008 field season at Scotty Creek, Northwest Territories, Canada.

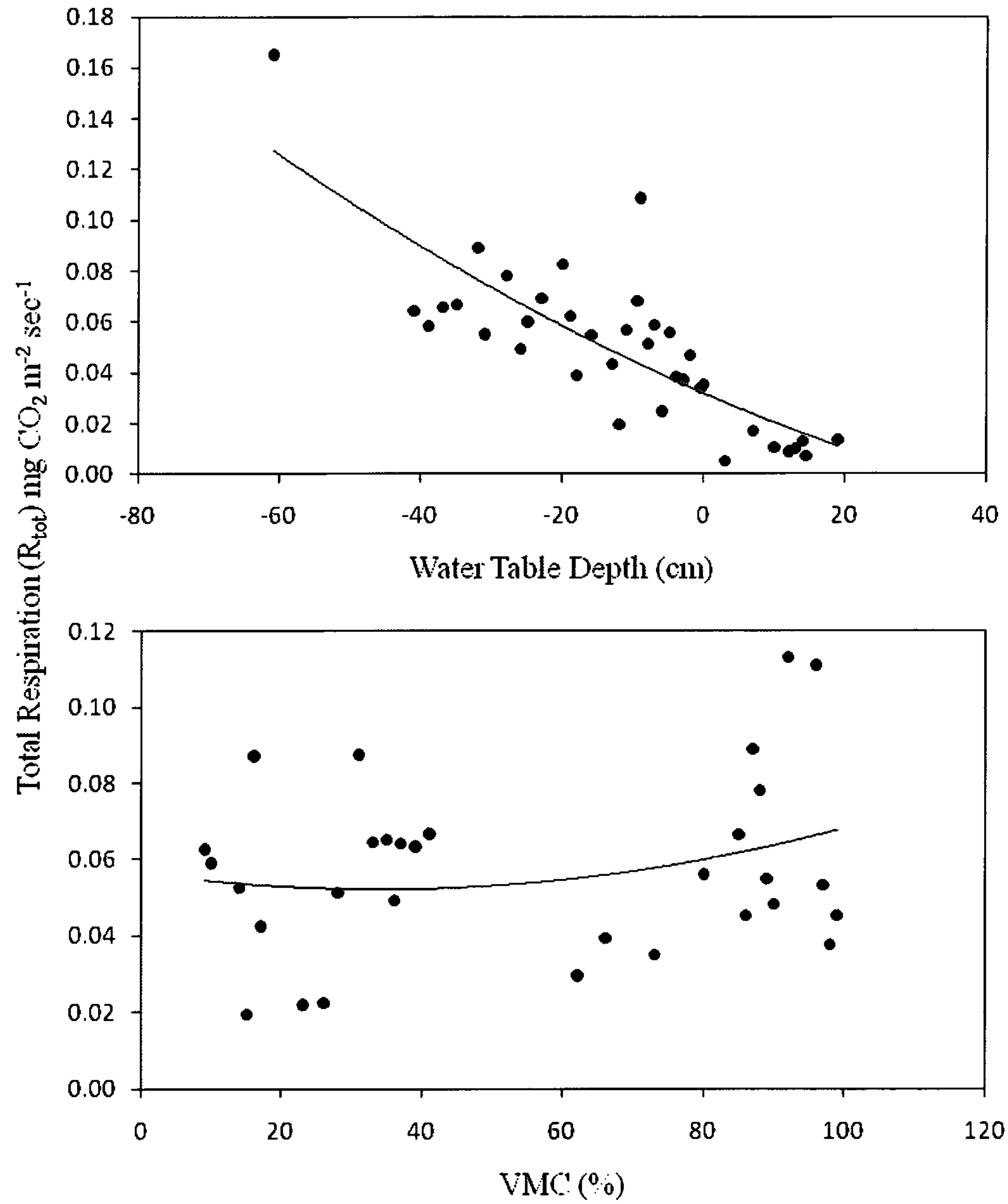

Figure 5-5 Relationship between total respiration (R_{tot}), water table depth (n=70, r^2=0.6), and volumetric soil moisture (n=54, r^2=0.1) on the permafrost plateau. All data is from the 2008 field season at Scotty Creek, Northwest Territories, Canada.

5.1.4 Modelling Continuous Seasonal and Ecosystem CO_2 Flux

Based on the relationships examined in this chapter, a model was developed to estimate continuous flux of GEP, R_{tot}, and NEE for each landscape unit between April 1st and August 23rd of the 2008 study season. GEP was modelled based on the light response curves. R_{tot} is strongly influenced by temperature and T_{air} was the only temperature variable continuously measured at all three landscapes during the 2008 season; therefore, R_{tot} was modelled based on T_{air}. The modelled total daily gas flux shown in Figure 5-6 clearly identifies the greatest variability in flux rates on the permafrost plateau. Despite differences in magnitude, the temporal responses are similar in all three landscapes. Sinks for CO_2 occurred pre-dominantly during April and early May, while positive rates of NEE were greatest in the months of July and August. Table 5-1 summarizes the modelled total seasonal flux of GEP, R_{tot}, and NEE for each landscape unit. From these results, each site showed fairly similar total seasonal flux of GEP; however, they varied significantly in R_{tot}. The channel fen experienced the greatest total flux for GEP, R_{tot}, and NEE, while the flat bog experienced the lowest fluxes of all three landscapes. Based on the total seasonal flux for NEE, the channel fen was the greatest source of CO_2, while the flat bog was the lowest.

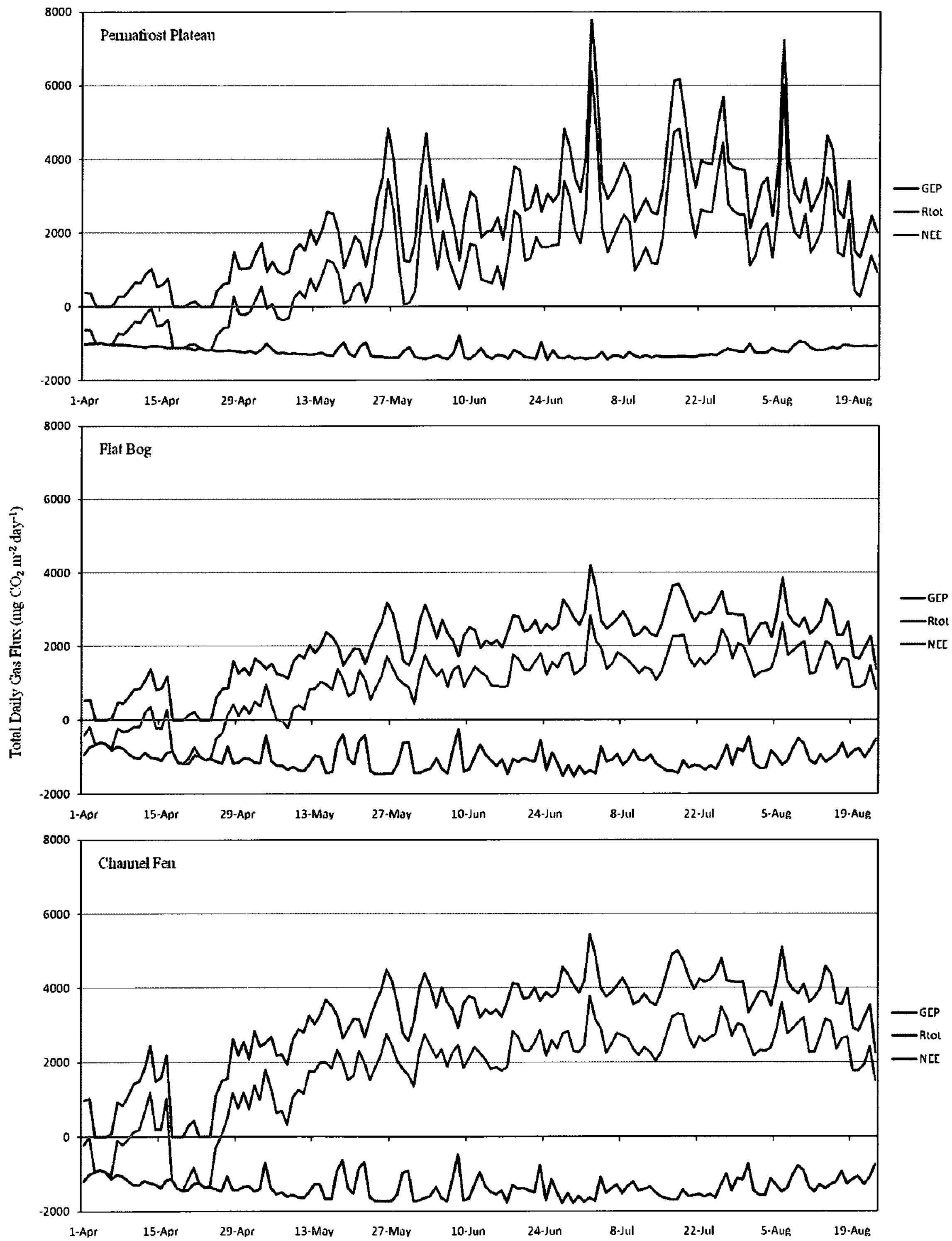

Figure 5-6 Modelled total daily gas flux for each landscape unit during the 2008 field season at Scotty Creek, Northwest Territories, Canada. Results based on the computed continuous flux of gross ecosystem production (GEP), total respiration (R_{tot}), and net ecosystem exchange (NEE) between April 1st and August 23rd, 2008.

Total Seasonal Flux	Permafrost Plateau	Flat Bog	Channel Fen
GEP (g CO_2 m^{-2})	178.0	153.5	196.5
Rtot (g CO_2 m^{-2})	353.7	295.7	451.8
NEE (g CO_2 m^{-2})	175.7	141.5	255.3

Table 5-1 Total seasonal flux for each landscape unit examined during the 2008 field season at Scotty Creek, Northwest Territories, Canada. Results are based on the modelled continuous flux of gross ecosystem production (GEP), total respiration (R_{tot}), and net ecosystem exchange (NEE) between April 1st and August 23rd, 2008.

Ground classification conducted by Quinton *et al.* (2009) on a sub-section of the Scotty Creek basin, covering an area of approximately 22 km^2 (Figure 2-1 (c)) shows the dominance of permafrost plateaus on the landscape (Table 5-2). Permafrost plateaus occupy the greatest areal portion (43%), followed by isolated and connected bogs (26.7%), and channel fens (21%).

Cover Type	N	Area (km^2)	Area (%)
Permafrost Plateaus	609	9.52	43.0
Flat Bogs (isolated)	999	0.89	4.0
Flat Bogs (connected)	-	5.03	22.7
Channel Fens	2	4.65	21.0
Lakes	4	2.06	9.3

Table 5-2 Selected results of detailed ground cover classification of sub-section of the IKONOS image of Scotty Creek, Northwest Territories, Canada, representing an area of ~ 22 km^2 on the ground. N is the number of samples of each cover type. Deriving the number of connected flat bogs was not attempted (Quinton *et al.*, 2009). Refer to **Figure 2-1** (c) for map detailing classification area.

The total seasonal NEE results (g CO_2 m^{-2}) computed from the model for continuous seasonal CO_2 flux (Table 5-1) were applied to the areal coverage of the three landscape units (Table 5-2) over an approximately 22 km^2 sub-section of the Scotty Creek basin. This resulted in an ecosystem-scaled gas flux (g CO_2) summarized in Table 5-3. The greatest source of CO_2 for this ecosystem is the permafrost plateaus which dominate the landscape, while the channel fens and flat bogs appear to be lower total contributors to the emission of CO_2 back into the atmosphere.

Ecosystem-Scaled Seasonal Flux	Permafrost Plateau	Flat Bog	Channel Fen
Area (km^2)	9.52	5.92	4.65
Weighted Flux (g CO_2)	**1,672,812**	**837,642**	**1,187,067**

Table 5-3 Ecosystem-scaled net ecosystem exchange (NEE) for each landscape unit between April 1st and August 23rd of 2008. Results based on modelled continuous flux for each landscape and ground cover classification of a sub-section of Scotty Creek, Northwest Territories, Canada, representing an area of ~ 22 km^2 on the ground. Refer to **Table 5-1** and **Table 5-2** for source values.

Lastly, the weighted fluxes calculated in Table 5-3 were multiplied by the known areal coverage of each landscape unit over the 22 km^2 sub-section of Scotty Creek basin. The sum of these calculations were equivalent to the total weighted average CO_2 flux (g CO_2) for this portion of the Scotty Creek Basin (Table 5-4), representing the current distribution of landscape units. Modifications made to the areal coverage of each landscape and re-calculation of the total CO_2 flux for this area would quantify the potential impact of landscape evolution on ecosystem-scaled CO_2 exchange. For example, if the plateau coverage declined by 10%, resulting in an increase in bog coverage by 10%, the weighted average of the basin based on the results of this study would decline by 83, 516 g CO_2.

Proportional Flux	Permafrost Plateau	Flat Bog	Channel Fen
Area (%)	43.0	26.7	21
Weighted Flux ($g\ CO_2$)	**719,309**	**223,650**	**249,284**
Weighted Average Flux of the Basin ($g\ CO_2$)		**1,192,243**	

Table 5-4 Proportional net ecosystem exchange (NEE) for each landscape unit between April 1st and August 23rd of 2008. Results are based on modelled continuous flux for each landscape and ground cover classification of a sub-section of Scotty Creek, Northwest Territories, Canada, representing an area of ~ 22 km^2 on the ground. Refer to **Table 5-2**, and **Table 5-3** for source values.

Chapter 6 Discussion

6.1 Temporal and Spatial Variability of CO_2 Exchange

The three landscapes examined in this study in the Scotty Creek basin are representative of a typical continental high boreal wetland. Based on the 30-year annual averages for 1971-2000, the Fort Simpson region experiences an average annual air temperature of -3.2 °C and receives approximately 369 mm of precipitation of which 170.3 cm is cumulative snowfall. Figure 2-3 however, illustrated that during 2008 average annual air temperatures were cooler (-4.1°C), and while total precipitation did not vary significantly (360.5 mm), there was greater cumulative snowfall (259.8 cm). Examining the seasonal periods as defined by this study, the winter of 2008 experienced a greater volume of snowfall, the snowmelt and pre-green period was wetter and colder, and the green period was drier and warmer than normal.

Temperature and moisture play a significant role in the balance between carbon acquisition and loss. Increasing temperatures are more favourable for vegetation growth, resulting in an increase in GEP; however increasing temperatures also favour thaw depth and decomposition, resulting in an increase in R_{tot} (Chimner *et al.*, 2010; Griffis and Rouse, 2001; Groendahl *et al.*, 2007; Zimov *et al.*, 1996). Dry conditions can often result in limited vegetation growth due to a lack of available water, reducing GEP, while wet conditions can suppress R_{tot} (Chimner *et al.*, 2010; Griffis and Rouse, 2001; Groendahl *et al.*, 2007).

Examining instantaneous fluxes during 2008 captured the temporal variability of CO_2 flux at Scotty Creek. Flux rates observed at each site reflected rates for CO_2 flux found in similar studies (i.e. Bubier *et al.*, 2003; Griffis *et al.*, 2000; Lafleur, 1999, Moore, 1989; Petrone *et al.*, 2001; Silvola *et al.*, 1996). There was a large accumulation of snow during this study and the months previous. Deep snowpacks act as insulators to enhance thaw depth and microbial activity below the ground surface with the potential result of high rates of R_{tot}. While the presence of ice layers within the snowpack can be an impediment to the flow of gas between the surface and the atmosphere, this was not the case on the plateau where snowpacks were monitored every few days with no sign of ice layers. However, low rates of CO_2 emission were still recorded during the snowmelt season at each site. Therefore, the effect of snow as an insulator was negligible as cooler air temperatures during the shoulder season initially maintained a shallow frost table. The amount of available moisture over-saturated the remaining available thawed space, resulting in low rates of R_{tot}. The majority of studies that have had the opportunity to examine CO_2 fluxes during snowmelt have also observed a net source of CO_2 to the atmosphere (i.e. Corradi *et al.*, 2005; Zimov *et al.*, 1996). The predictions and observations of Elberling (2007) and Zimov *et al.* (1996), among others, continue to suggest that while CO_2 flux rates during the winter are often relatively small (Aurela *et al.*, 2002), as seen in this study, the cumulative efflux of CO_2 during winter in many ecosystems may be near or equal to the amount lost during the growing season, playing a critical contribution to annual carbon budgets.

In northern environments the presence of snow cover later in the year corresponds to near maximum light levels and large amounts of available water at the time when the ground becomes snow and ice-free (Bubier *et al.,* 1998). In addition, pre-green season nutrient levels are often high initially as a result of snowmelt supplying the sub-surface with nutrients, labile root and detrital organic matter preserved from the previous year by freezing temperatures (Schlesinger, 1977). These types of conditions in addition to increasing air temperatures typically initiate a swift response by the ecosystem after snowmelt, resulting in the quick succession of vegetation and increasing R_{tot} and GEP (Bubier *et al.*, 1998). Sudden and rapid development of root structures and shoots will often give rise to the removal of CO_2 from the atmosphere (Rouse *et al.,* 2002) creating a net sink during early spring, captured in studies such as Griffis *et al.* (2000) and Lafleur and Humphreys (2007). This was the case for each of the landscapes studied at Scotty Creek. All three experienced late snow cover and an abundance of water and organic matter after snowmelt, resulting in a short pre-green season due to the quick succession of vegetation. On the plateau the ground surface was bare and exposed with a large amount of available water both above and below the surface. Available water on the plateau and its runoff into the surrounding wetlands can supply nutrients to areas where microbes and vegetation can utilize them, enhancing decomposition and R_{tot} (Bekku *et al.*, 2003). Both the daily and seasonal averages reflected these conditions with increases in R_{tot}. Rates remained low for both R_{tot} and GEP at PPC-2 and PPC-3 most likely due to the over-abundance of water which flooded and over-saturated the two sites, suppressing CO_2 flux. Meanwhile, GEP increased substantially at the vegetated plateau site, PPC-1, resulting in a sink for CO_2. The channel fen and flat bog had little to no vegetation growth and high water tables, which was reflected in small GEP flux rates; however, R_{tot} rates increased similar to the plateau.

During the transition between the pre-green and green season, when vegetation is in its early stages of development, the timing of leaf-out can signify a shift in the carbon budget (Bubier *et al.*, 2003). Through observation, the plateau began to show signs of vegetation growth by the end of the pre-green season (~ May 7th), while the bog and fen began developing in the first two weeks of the green season (~ May 14th). Throughout the entire growing season nutrients in the sub-surface are accessed by root growth and dissolution from precipitation and sub-surface water flow (Schlesinger, 1977). Photosynthetic activity and surface warming are controlled by the presence and seasonal development of an above ground canopy which limits the amount of light able to penetrate down to the ground surface. Frost table and water table depths define and allocate the amount of the sub-surface able to thaw and aerate, allowing microbial activity to operate at high rates and have access to a greater store of organic matter. All of these conditions further stimulated vegetation growth, nutrient enrichment, and decomposition at each site while maintaining and increasing the rates of R_{tot} and GEP. Whether a landscape becomes a sink or source for CO_2 during this time, under such conditions, is dependent upon temperature and moisture regimes in addition to the specific nature of the landscape. Studies in the sub-arctic tundra have documented the growing season as a consistent source for CO_2 (Lafleur, 1999; Lafleur and Humphreys, 2007), while other peatlands have found considerable sinks for CO_2 as productivity outweighed total respiration (Bubier *et al.*, 2003; Rouse *et al.*, 2002; Waddington and Roulet, 1996). Or in the case of Groendahl *et al.* (2007), the growing season began as a source but shifted to a sink during the month of August. At Scotty Creek, the greatest increases in both R_{tot} and GEP occurred in the productive vegetated landscapes of the plateau and fen, with a lower response in the bog, which retained only slight vegetation growth above a high water table. As seen in both the time-series and seasonal averages there is growing variability during this season both between and within landscapes. However, the green period remained an overall source of CO_2 for the majority of sites. Despite only a portion of the summer months being captured for this study, it is assumed that R_{tot} and GEP would continue to increase, the rates of which would depend upon the limitations or advantages provided by environmental conditions.

In the late-green season, at the end of summer and start of fall, GEP can show signs of decline as light levels decrease, soil temperatures cool, frost table depths become shallow, and vegetation nears senescence (Goulden *et al.*, 1998). R_{tot} however, may remain high if temperature and moisture deficits are not strong enough to break down microbes and slow down decomposition rates (Goulden *et al.*, 1998). The one day measured during the late-green season showed high rates of both R_{tot} and GEP for the plateau and fen, again most likely a result of significant vegetation growth and lower water tables than the flat bog. While only one day's worth of data was collected for the late-green period, it would be expected that these rates would begin to decrease over more time and this would be captured in a more extensive study, seen in studies such as Silvola *et al.* (1996).

The greatest rates of R_{tot} between landscapes occurred in the permafrost plateau and channel fen with the lowest rates in the flat bog. For both the channel fen and flat bog, the greatest rates of R_{tot} took place at the sites with the least vegetation (FC-3, BC-3 and BC-2). Meanwhile, the greatest rates of R_{tot} on the permafrost plateau took place at the most elevated and vegetated of the three sites, PPC-1, and the lowest rates at the flooded site, PPC-2. The greatest rates of GEP between landscapes occurred in the permafrost plateau, then the channel fen, with the lowest in the flat bog. Within the landscapes the bog showed little variation between sites, similarly in the fen with only slightly greater GEP in FC-2. The permafrost plateau however, showed more variability between sites. The vegetated PPC-1 had significantly greater rates of GEP compared to the bare and flooded sites of PPC-3 and PPC-2, respectively. For both R_{tot} and GEP, the flat bog displayed the least amount of variability and range in magnitude in comparison to the permafrost plateau and channel fen. This response in respiration is similar to other studies that have found low CO_2 flux in bog terrains (Bubier *et al.*, 1998).

6.2 Environmental Controls of CO_2 Exchange

The three landscape units have characteristic properties (presence or absence of permafrost, canopy cover, microtopography, vegetation and key species, soil profile, thermal and light regimes, and hydrology). Therefore, the differences in R_{tot}, NEE, and GEP among the landscape units are considered in terms of these contrasting site properties and their microclimate. The bog and fen sites were the first to become snow-free and the first to have frost table depths reach to an immeasurable depth resulting in saturated and open water conditions at each site. Both experienced high and variable water tables, high light levels, and increasing air and water temperatures; however, they had very different temporal responses in vegetation distribution and development, and CO_2 exchange. The bog sites had little vascular plant or shrub development above water and were dominated by *Sphagnum* moss species. When a major portion of the moss species is underwater, in a saturated and anoxic environment, it slowly dies with lack of light. The weight of overlying peat compresses the dead plant material as it weakens to decay (Johnson *et al.*, 1990). Anoxic conditions due to a high water table create very slow rates of decomposition, often resulting in greater rates of accumulation (Aerts *et al.*, 2001; Bubier *et al.*, 1998). This is further inhibited by the nutrient-limited conditions found in northern bogs that receive nutrients solely through atmospheric deposition and surrounding landscape runoff (Aerts *et al.*, 2001; Bubier *et al.*, 1998). In comparison, the fen sites had a buoyant peat mat at the water surface once the initial snowmelt water supply receded. Dominated by *Sphagnum squarrosum, Helodium blandowii, Brachythecium rivulare,* and *Aulacomnium palustre* moss species; in addition to ericaceous shrub and sedge species, the fen had a more competitive and diverse vegetative cover. As water conveyors, minerotrophic channel fens are also more nutrient-rich than ombrotrophic flat bogs and therefore more productive, as demonstrated by their abundance in vegetation at the sites monitored. In this case, the channel fen had a greater percent of total nitrogen than the bog.

The differential development between the two landscapes is reflected in the gas flux results. The fen had greater temporal variance, in both direction and magnitude, than the bog. A lack of canopy and high PAR values throughout the season resulted in increased photosynthetic activity for the fen, while the bog experienced little to no GEP due to its limited vegetation coverage above the water table. For NEE, the lowest emission occurred at the *Sphagnum*-dominated bog due to slow decomposition and low abundance of vascular plants, especially trees. Bogs typically have less primary production and slower decomposition (Silvola *et al.*, 1996; Thormann and Bayley, 1997).

The highest emission occurred at the photosynthetically active channel fen, which received fresh, nutrient-rich water from the adjacent plateau soils.

The permafrost plateau is a more complex landscape in comparison to the flat bog and channel fen. In addition to climatic, hydrological and biological variables, the presence of a canopy and variation in microtopography will influence CO_2 flux and respond differently than the surrounding wetlands (Bubier *et al.*, 1998; Petrone *et al.*, 2010; Solondz *et al.*, 2008; Waddington and Roulet, 1996). During the winter months the presence of a canopy decreased the potential decline in snow depth from wind-driven drift and sublimation. With the deepest snowpacks of the three landscapes, the plateau had a slower rate of loss and was the last to become snow-free. The presence of permafrost creates unique sub-surface conditions that heavily determine the survival of the over-lying plateau (Vitt *et al.*, 1994; Zoltai, 1993). Through its presence it elevates the surface and allows the development of tree roots that need relatively dry conditions to survive. While in the absence of permafrost this elevation does not occur and soils once dry become saturated, reducing potential growth and increasing mortality of forest stands. Permafrost thickness and the depth of its overlying active layer are dependent upon, and influence, the stability of climatic variables, soil characteristics, hydrology, and vegetative cover at the surface, all of which are controls on the CO_2 flux.

The frost table depth only reached an average depth of 105 cm for the duration of the study season, in comparison to the bog and fen, and the fluctuating water table was continuously re-defining the depth below the surface that was saturated and unsaturated. Lowering of the water table results in greater R_{tot} due to the increasing availability of oxygen and organic matter as the aerobic zone thickens, thereby enhancing microbial activity and decomposition rates (Billings *et al.*, 1983; Griffis *et al.*, 2000; Oechel *et al.*, 1995). Therefore, high water tables or flooding often reflect in lower rates of R_{tot}. While over-saturation can reduce decomposition rates, increases in soil moisture can positively influence thaw depth by increasing thermal conductivity (Hayashi *et al.*, 2004; Wright *et al.*, 2009). Standing water on the surface, unable to percolate into saturated soils, will transfer heat from the warmer surface into the cooler soil. This effect will strongly influence ground thaw in locations where lateral flow is converging into a depressed area (Guan *et al.*, 2010; Wright *et al.*, 2009). The strongest rates of R_{tot} were measured at PPC-1 with the lowest water table, while the lowest rates were measured at PPC-3 which was flooded for a majority of the study period.

Vegetation and ground cover varied on the plateau. For example, PPC-1 was predominantly covered with *Sphagnum* and lichen species while PPC-3 was bare. Moss can potentially have a strong effect on NEE of CO_2 (Petrone *et al.*, 2004). *Sphagnum* moss species are seasonally photosynthetic, usually peaking in the middle of the growing season (Botting and Fredeen, 2006; Swanson and Flannagan, 2001). The thin black spruce canopy cover on the plateau allowed high levels of PAR to reach the ground surface, stimulating plant production, and thereby, increasing rates of GEP during the growing season. Incident PAR at the ground surface also increases air and soil temperatures. Air temperature steadily increased; however, soil temperature remained low, which could be due to the insulating properties of *sphagnum* moss (Oechel and Van Cleve, 1986; Van Cleve *et al.*, 1983).

Further, the plateau had slightly lower average total nitrogen and total carbon percentages than the flat bog and fen, with the highest C:N ratio. Typically higher quality soils (low C:N) have greater nutrient availability for microbial activity (Moore *et al.*, 1998; Raich and Schlesinger, 1992) and thereby the possibility for greater soil respiration. Although the permafrost plateau had a lower quality soil profile in comparison to the bog and fen sites, other factors may have played a greater role in stimulating soil respiration.

When examining the dependence of the CO_2 flux to these environmental controls previous studies have attributed temporal variability in GEP to PAR, and R_{tot} to T_{soil} and T_{air} (Botting and Fredeen, 2006; Bubier *et al.*, 1998; Goulden *et al.*, 1998; Law *et al.*, 2002; Raich and Schlesinger, 1992; Waddington and Roulet, 1996). The exponential model used in this study has been proven as an adequate quantitative tool for fitting light response curves that realistically portrays the ecosystem being examined (Gomes *et al.*, 2006). The relationship between GEP and PAR showed significant scatter and variability for all three landscapes suggesting the influence of additional environmental variables on gas flux other than PAR. This type of scatter is also seen in other studies (Bubier *et al.*, 1998; Lafleur, 1999) and can be a reflection of limiting environmental conditions such as temperature and moisture. Persistence of winter conditions later in the year in northern environments, in combination with the climatic conditions at Scotty Creek during 2008 may have limited early spring growth with cool and wet conditions and growth during peak growing season with the warm and dry conditions during summer. These could have influenced the scatter between PAR and GEP.

Temperature appeared to be an influential factor on CO_2 exchange on all three landscapes with strong positive correlations between R_{tot}, T_{air}, T_{soil} (permafrost plateau) and T_{water} (flat bog and channel fen). Hydrological characteristics also appeared to significantly influence CO_2 exchange. As the bog (and most likely the fen based on similar development) became snow-, and ice-free, and water temperatures increased, R_{tot} increased. Similar strong negative trends were found with snow depth, frost table depth, and water table depth on the permafrost plateau. VMC had a significant amount of scatter and no trends were established as each plateau site responded differently to the same VMC. This could potentially be due to instrument error or the possibility that other environmental conditions measured were more influential to CO_2 exchange.

Microtopography and vegetation appeared to influence each landscape differently. In the flat bog the two sites (BC-2 and BC-3), located furthest away from the plateau in open water conditions with little above water vegetation, had stronger responses in R_{tot} than BC-1 to temperature variation and thaw depth. For the channel fen, all three sites were located along the edge of the fen and only varied in the amount of vegetation present, which did not appear to significantly influence their responses in R_{tot}. While the bog and fen sites differed little in topography, the three sites studied on the permafrost plateau did. PPC-1 was the highest site, situated on a slope and predominantly vegetated, PPC-2 was situated in a depression and flooded, while PPC-3 was the lowest site, situated at the bottom of a slight slope and bare. Typically higher topographical features have a larger NEE uptake than lower sites due to greater aerobic conditions (Waddington and Roulet, 1996). This was replicated at the plateau sites where PPC-1 had the greatest uptake of CO_2. PPC-1 had the strongest relationship with the environmental variables examined, PPC-3 had the strongest rates of R_{tot}, while and the magnitude of CO_2 exchange at PPC-2 appeared to be significantly impacted by its flooded conditions.

6.3 Modelled Seasonal Flux: Sink or Source for CO_2

Peatlands have been net sinks for CO_2 since deglaciation (Makiranta *et al.*, 2009). However, the balance between CO_2 uptake and loss is sensitive enough that a small change in water table depth, temperature, or timing of thaw, leaf-out and senescence can favour decomposition over plant production (Bubier *et al.*, 1998; Carroll and Crill, 1997; Chivers *et al.*, 2009; Shurpali *et al.*, 1995; Waddington and Roulet, 1996). When the data collected from the 2008 study season was scaled into a continuous dataset for each landscape, the total daily flux showed similar temporal variability but differed in magnitude. The flat bog displayed the least variability in CO_2 flux of the three landscapes and recorded the lowest total seasonal flux for R_{tot}, GEP, and NEE. Meanwhile, the permafrost plateau displayed the greatest variability in CO_2 flux; however, it was the channel fen that ultimately recorded the highest total seasonal flux for R_{tot}, GEP, and NEE. While in situ measurements showed sinks for CO_2 in both the plateau and fen landscapes during the pre-green and green periods, respectively, the continuous data alternated between sink and source trends during snowmelt and developed into a consistent source during the growing season, with signs of decline towards the end of the season (August 23rd, 2008).

Studies examining the seasonal patterns and controls on net ecosystem CO_2 exchange in peatland, wetland, and northern environments have found differing results. What they all appear to agree on is the sink and source variability during snowmelt and pre-green. Winter and snowmelt typically respond as small consistent sources of CO_2 to the atmosphere (i.e. Corradi *et al.*, 2005; Zimov *et al.*, 1996), while pre-green often switches to a sink of CO_2 (i.e. Griffis *et al.*, 2000; Lafleur and Humphreys, 2007; Rouse *et al.,* 2002). The greatest discrepancy occurs during the green (growing) season. The ecosystems examined by Botting and Fredeen (2006), Burton *et al.* (1996), Griffis *et al.* (2000), Griffis and Rouse (2001), Groendahl *et al.* (2007), Lafleur *et al.* (1997), Lafleur and Humphreys (2007), Petrone *et al.* (2001), Rouse *et al.* (2002), Schreader *et al.* (1998), Shurpali *et al.* (1995), Swanson and Flannagan (2001), and Waddington and Roulet (1996) had a net loss of CO_2 to the atmosphere during the growing season. As many of these studies are based on several years of data they have also monitored growing seasons with net uptake of CO_2 from the atmosphere, in addition to Bubier *et al.* (2003), and Lafleur (1999). For example, Griffis *et al.* (2000) examined a similar transition from pre-green to late-green and reported variability ranging from a net sink of – 235 g CO_2 m^{-2} in one year to a net source of 76 g CO_2 m^{-2} in the same peatland a few years earlier. For many of these studies, temperature and moisture played a large role in annual alteration between sink and source. Cool, wet conditions during the growing season often resulted in greater sinks for CO_2, while warm, and dry conditions resulted in greater sources for CO_2. Drier conditions can change a peatland from a sink to a source within a single season (Griffis *et al.*, 2000; Joiner *et al.*, 1999; Oechel *et al.*, 1995; Oechel *et al.*, 2000; Shurpali *et al.*, 1995). As previously mentioned, the growing season during 2008 at Scotty Creek was warmer and drier than normal which could explain the overall sources of CO_2 observed for each landscape unit.

While instantaneous rates of R_{tot}, GEP, and NEE were comparable between this study and previous studies, the continuous data ranged in magnitude (i.e. Griffis *et al.*, 2000; Petrone *et al.*, 2001). This could be due to differences in the methodology between studies. For example, many studies incorporated eddy covariance measurements, which include below and above canopy fluxes while this study was more site-specific at the atmosphere-ground surface interface and then scaled up. Not including canopy cover and shrub vegetation in this study eliminates a potentially significant source of CO_2 uptake from these predictions. Therefore, the quantitative comparison between studies is difficult to make. Chamber measurements were also taken daily at midday, which can result in higher R_{tot} and GEP rates than if the fluxes were continuously monitored throughout the day and night. Moreover, many studies also focused their measurements only on the growing season between mid-May and August, or into October, while this study focused on the months of April to June, and the end of August. This could potentially create variability in the relationships established between CO_2 flux and its environmental controls, from which this study developed its continuous dataset. In addition, respiration rates can be greater in northern boreal wetlands than lower latitudes because the soil does not experience the same degree of water deficit in the summer months (Law *et al.*, 2002), which could explain the high rates of R_{tot} captured during this study.

Applying remote sensing information to spatially extend in situ measurements allowed for a preliminary assessment of the relative importance of each landscape to the CO_2 functioning of this ecosystem. Raich and Schlesinger (1992) found respiration was more dominant in plateau features than fens and bogs. Computing the total seasonal flux for NEE (g CO_2 m^{-2}) at Scotty Creek, the channel fen was identified as the greatest source of CO_2 to the atmosphere (255.3 g CO_2 m^{-2}), then the permafrost plateau (175.7 g CO_2 m^{-2}), and lastly the flat bog (141.5 g CO_2 m^{-2}). This could reflect the unique characteristics of permafrost environments, or a particularly nutrient-rich fen. Once these cumulative fluxes were areally weighted to cover a larger area (~22 km^2), the permafrost plateau was identified as the greatest cumulative source of CO_2 (1, 672, 812 g CO_2) back into the atmosphere, a reflection of its dominance on the landscape, covering approximately 43 % of the region examined. It was followed by the channel fen (1, 187, 067 g CO_2) and lastly the flat bog (837, 642 g CO_2). Although the channel fen had the greatest seasonal flux its areal coverage on the landscape was much less than the permafrost plateaus.

As the climate continues to change and the landscape evolves, the current state of the Scotty Creek basin as an overall source of CO_2 is unstable and will continue to transform. Images covering a 1 km x 1 km subset area of Scotty Creek, dating from 1947-2008 clearly shows the threat of permafrost degradation (Figure 6-1). In 1947 approximately 70.4 % of this region was underlain by permafrost. In the 61 years since 1947 there has been a consistent and steadily increasing decline in plateau coverage. In 2008, the year of this study, the region was re-estimated at only 43.3 % underlain by permafrost. Since 1970 the rate of loss was occurring at approximately 0.2 % per year. However, from 2000 to 2008, this rate has increased to 0.8 % per year. Continued subsidence of permafrost plateaus will result in increasing bog landscape coverage. This evolution into a northern boreal wetland predominantly composed of bog landscapes, could potentially result in an initial increasing loss of CO_2 to the atmosphere as permafrost degrades and exposes more of the sub-surface to the processes of respiration. However, this trend will eventually reach a threshold and decline as the percentage of bogs on the landscape increases, the weakest emitters of CO_2.

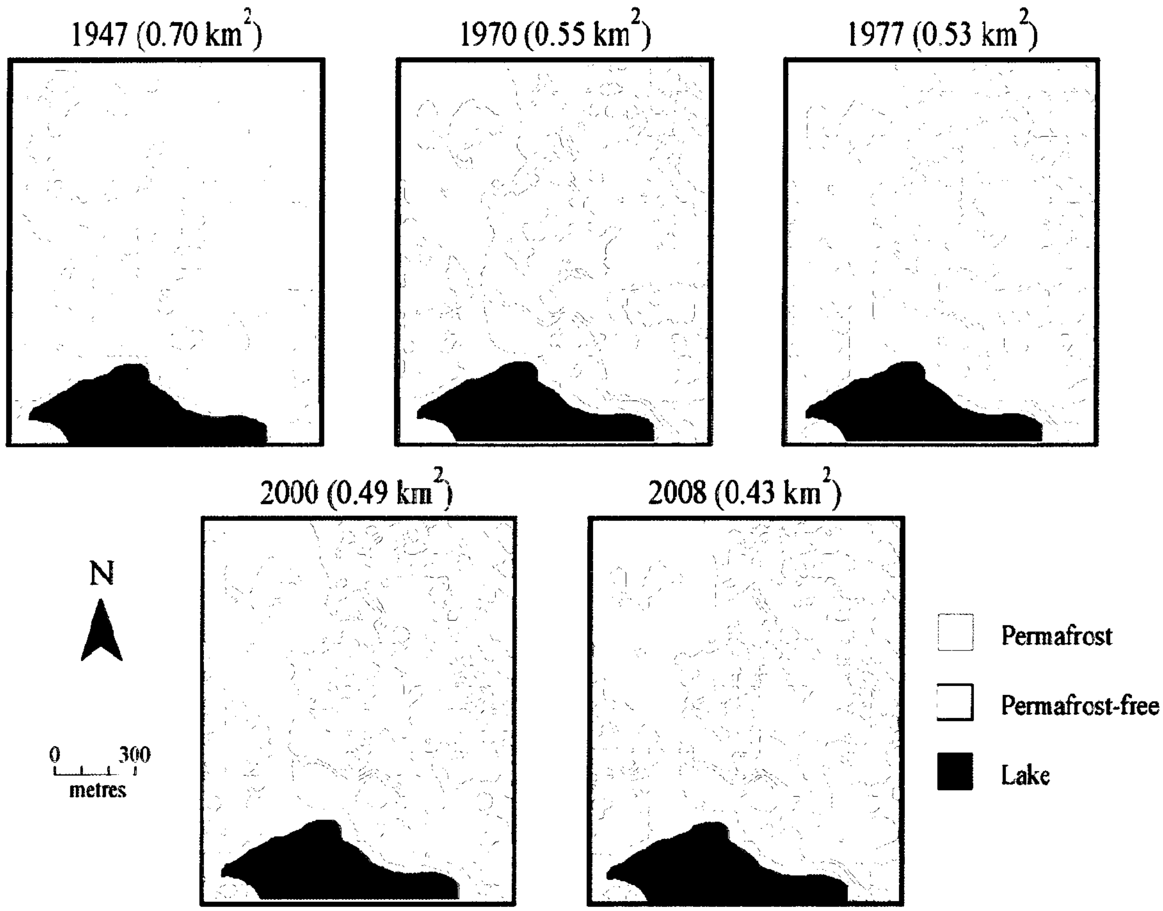

Figure 6-1 Evidence of permafrost loss, degradation and evolving permafrost plateau distribution in a 1 km^2 sub-set area of Scotty Creek, Northwest Territories, Canada in 1947, 1970, 1977, 2000, and 2008 (Quinton *et al.* (in review)).

Chapter 7 Conclusion

Northern boreal wetlands represent an ecosystem that has hydrological, biological, and greenhouse gas exchange significance. This study observed the region-specific temporal and spatial variability of net ecosystem CO_2 exchange and its driving environmental controls. By examining the different landscape units within a boreal wetland these physical processes were further defined by determining how differences in canopy cover, vegetation distribution, sub-surface profile, and other site specific environmental controls influenced CO_2 exchange. The mean annual air temperature for the Fort Simpson region has increased by approximately 2.0 °C over the last 100 years (Environment Canada, 2010). General Circulation Models (GCMs) predict a further increase in the average annual air temperature of 1.7 - 4.0 °C in northwestern Canada by 2050 (IPCC, 2007). Relationships identified in this study with temperature, thaw depth and water table depth reflect the possibility of further stress on CO_2 exchange in this region in response to such climate change. While potential limitations in methodology create some uncertainty in the strength of the spatial relationships captured, the primary interest of this study was to examine and attempt to understand the relative difference in CO_2 exchange within and between three unique landscape units in a boreal wetland.

Through the use of environmental relationships and remote sensing technology this study found that during the study season of 2008, while the individual channel fen examined was the greatest emitter of CO_2 into the atmosphere it was the permafrost plateau that had the greatest total seasonal flux over a larger area. Based on this assessment, the continued subsidence of permafrost plateaus will initially result in an increasing amount of CO_2 emitted into the atmosphere as plateaus decline in number and size. This trend however, will most likely reach a threshold and begin to decline as bog landscapes become more prominent on the landscape. Whether this ecosystem will ultimately shift from a source to sink of CO_2 over time is uncertain based on the results found.

This study has illustrated that although previous work has examined the spatial and temporal variability of CO_2 exchange in many ecosystems, it can't necessarily be extrapolated to the northern boreal wetlands of Scotty Creek basin. This basin is a representative example of northern boreal wetlands, their current landscape composition, and their ongoing adaption to landscape evolution. Their position on the landscape in relation to one another, maintained by the presence of permafrost, defines their hydrology, nutrient status, and vegetation composition, all of which control the current CO_2 flux of this landscape. As the climate in northern ecosystems continues to change, understanding the interactions between the physical, biochemical, and environmental conditions of different landscapes and the processes which define them can aid in the parameterization and interpretation of current and future climate and biogeochemical models. This study highlights the need for long term measurement in order to capture and examine a variety of climatic conditions, which can strengthen the relationships captured in 2008, and validate whether they are similar to those found in climatically different years.

Chapter 8 References

ACIA. *Impacts of a Warming Arctic: Arctic Climate Impact Assessment*. Cambridge University Press. 2004.

Aerts, R., Wallen, B., Malmer, N., and de Caluwe, H. 2001. Nutritional constraints on Sphagnum-growth and potential decay in northern peatlands. *Journal of Ecology,* 89(2): 292-299.

Alterio, G., Giorio, P., and Sorrentino, G. 2006. Open-system chamber for measurements of gas exchanges at plant level. *Environmental science & technology,* 40(6): 1950-1955.

Alyesworth, J.M. and Kettles, I.M. 2000. Distribution of peatlands. In *The Physical Environment of the Mackenzie Valley, Northwest Territories: a Base Line for the Assessment of Environmental Change,* (ed.) Dyke, L.D. and Brooks, G.R., Geological Survey of Canada, Bulletin 547, 49-55.

Anisimov, O.A., Fitzharris, J.O.H., Jeffries, R., Marchant, H., Nelson, F.E., Prowse, T. and Vaughan, D.G. 2001. Polar regions (Arctic and Antarctic). In *Climate Change: Impacts, Adaptation, and Vulnerability, the Contribution of Working Group II of the Intergovernmental Panel on Climate Change, Third Assessment Review.* Cambridge University Press, Cambridge, p.801-841.

Anisimov, O.A., Shiklomanov, N.I., and Nelson, F.E. 2002. Variability of seasonal thaw depth in permafrost regions: A stochastic modeling approach. *Ecological Modeling,* 153(3): 217-227.

Apps, M.J. 2002. Peatlands in Perspective: Circumpolar Boreal Carbon Cycles. In *Proceedings of the International Workshop on Carbon Dynamics of Forested Peatlands: Knowledge Gaps, Uncertainty, and Modeling Approaches; Long-Term Dynamics and Contemporary Carbon Budget of Northern Peatlands*, Natural Resources Canada: Edmonton, Alberta; 17-10.

Aurela, M., Laurila, T., and Tuovinen, J.P. 2002. Annual CO_2 balance of a subarctic fen in northern Europe: Importance of the wintertime efflux. *Journal of Geophysical Research*, 107(D21): 1-12.

Balogh, J., Foti, S., Nagy, Z., Czobel, S., Pinter, K., Peli, E., and Tuba, Z. 2005. Comparison of carbon dioxide fluxes over sandy grassland vegetation as measured by the eddy-covariance technique and by open system chamber. *Acta Biologica Szegediensis,* 49(1-2): 143-145.

Bekku, Y., Koizumi, H., Nakadai, T., and Iwaki, H. 1995. Measurement of Soil Respiration using Closed-Chamber Method: an IRGA Technique. *Ecological Research,* 10(3): 369-373.

Bekku, Y.S., Nakatsubo, T., Kume, A., Adachi, M., and Koizumi, H. 2003. Effect of warming on the temperature dependence of soil respiration rate in arctic, temperate and tropical soils. *Applied Soil Ecology*, 22(3): 205-210.

Billings, W.D., Luken, J.O., Mortensen, D.A., Peterson, K.M. 1983. Increasing Atmospheric Carbon-Dioxide: Possible Effects on Arctic Tundra. *Oecologia,* 58(3): 286-289.

Botting, R.S. and Fredeen, A.L. 2006. Net ecosystem CO_2 exchange for moss and lichen dominated forest floors of old-growth sub-boreal spruce forests in central British Columbia, Canada. *Forest Ecology and Management,* 235(1-3): 240-251.

Brown, R.J.E. and Kupsch, W.O. 1974. *Permafrost Terminology.* Technical Memorandum No.111 Associate Committee on Geotechnical Research, National Research Council of Canada, Ottawa, Canada.

Bubier, J.L., Crill, P.M., Moore, T.R., Savage, K., and Varner, R.K. 1998. Seasonal patterns and controls on net ecosystem CO_2 exchange in a boreal peatland complex. *Global Biogeochemical Cycles,* 12(4): 703-714.

Bubier, J.L., Frolking, S., Crill, P.M., and Linder, E. 1999. Net ecosystem productivity and its uncertainty in a diverse Boreal peatland. *Journal of Geophysical Research D: Atmospheres,* 104(22).

Bubier, J.L., Bhatia, G., Moore, T.R., Roulet, N.T., and Lafleur, P.M. 2003. Spatial and temporal variability in growing-season net ecosystem carbon dioxide exchange at a large peatland in Ontario, Canada. *Ecosystems,* 6(4): 353-367.

Burgess, M.M. and Smith, S.L. 2000. Shallow ground temperatures. In *The Physical Environment of the Mackenzie Valley, Northwest Territories: a Base Line for the Assessment of Environmental Change,* (ed.) Dyke, L.D. and Brooks, G.R., Geological Survey of Canada, Bulletin 547, 89-103.

Burton, K.L., Rouse, W.R. and Boudreau, D.L. 1996. Factors affecting the summer carbon dioxide budget of subarctic wetland tundra. *Climate Research,* 6: 203-213.

Camill, P. 2005. Permafrost thaw accelerates in boreal peatlands during late-20th century climate warming. *Climatic Change,* 68: 135-152.

Camill, P. and Clark, J.S. 1998. Climate change disequilibrium of boreal permafrost peatlands caused by local processes. *American Naturalist,* 151(3): 207-222.

Campbell, G.S. and Norman, J.M. 1998. *Introduction to environmental biophysics (*2nd Eds.), Springer, New York.

Carey, S.K. and Woo, M. 1998. Snowmelt hydrology of two subarctic slopes, southern Yukon, Canada. *Nordic Hydrology,* 29(4-5): 331-346.

Carroll, P. and Crill, P. 1997. Carbon balance of a temperate poor fen. *Global Biogeochemical Cycles,* 11(3): 349-356.

Chasmer, L., Hopkinson, C., and Quinton, W. (in press). Quantifying errors in discontinuous permafrost plateau change from optical data Northwest Territories, Canada: 1947 to 2008. *Canadian Journal of Remote Sensing.*

Chimner, R.A., Welker, J.M., Morgan, J., LeCain, D., and Reeder, J. 2010. Experimental manipulations of winter snow and summer rain influence ecosystem carbon cycling in a mixed-grass prairie, Wyoming, USA. *Ecohydrology.*

Chivers, M.R., Turetsky, M.R., Waddington, J.M., Harden, J.W., and McGuire, A.D. 2009. Effects of Experimental Water Table and Temperature Manipulations on Ecosystem CO_2 Fluxes in an Alaskan Rich Fen. *Ecosystems,* 12(8): 1329-1342.

Cohen, S.J. 1997. *Mackenzie Basin Impact Study (MBIS).* Final Report, Environment Canada, Environmental Adaptation Research Group.

Corradi, C., Kolle, O., Walter, K., Zimov, S.A, and Schulze, E.D. 2005. Carbon dioxide and methane exchange of a north-east Siberian tussock tundra. *Global Change Biology,* 11(11): 1910-1925.

Davidson, E.A., Belk, E., and Boone, R.D. 1998. Soil water content and temperature as independent or confounded factors controlling soil respiration in a temperate mixed hardwood forest. *Global Change Biology,* 4(2): 217-227.

Davidson, E.A., Savage, K., Verchot, L.V., and Navarro, R. 2002. Minimizing artifacts and biases in chamber-based measurements of soil respiration. *Agricultural and Forest Meteorology,* 113(1-4): 21-37.

Davidson, E.A., Richardson, A.D., Savage, K.E., and Hollinger, D.Y. 2006. A distinct seasonal pattern of the ratio of soil respiration to total ecosystem respiration in a spruce-dominated forest. *Global Change Biology,* 12(2): 230-239.

Dehcho Land Use Planning Committee. 2006. *Final Draft Dehcho Land Use Plan.* Dehcho Land Use Planning Committee, Northwest Territories, Canada.

Dingman, S.L. 1994. *Physical hydrology.* Prentice Hall, Englewood Cliffs, N.J.

Elberling, B. 2007. Annual soil CO_2 effluxes in the High Arctic: The role of snow thickness and vegetation type. *Soil Biology & Biochemistry,* 39(2): 646-654.

Environment Canada. 2010. *Canadian Climate Normals*. Meteorological Service of Canada, Ottawa, Ontario.

Fahnestock, J.T., Jones, M.H., Brooks, P.D., Walker, D.A., and Welker, J.M. 1998. Winter and early spring CO_2 efflux from tundra communities of northern Alaska. *Journal of Geophysical Research-Atmospheres,* 103(D22): 29023-29027.

Fang, C. and Moncrieff, J.B. 2001. The dependence of soil CO_2 efflux on temperature. *Soil Biology & Biochemistry,* 33(2): 155-165.

Frolking, S., Goulden, M.L., Wofsy, S.C., Fan, S.M., Sutton, D.J., Munger, J.W., Bazzaz, A.M., Daube, B.C., Crill, P.M., Aber, J.D., Band, L.E., Wang, X., Savage, K., Moore, T., and Harriss, R.C. 1996. Modeling temporal variability in the carbon balance of a spruce/moss boreal forest. *Global Change Biology,* 2(4): 343-366.

Frolking, S.E., Bubier, J.L., Moore, T.R., Ball, T., Bellisario, L.M., Bhardwaj, A., Carroll, P., Crill, P.M., Lafleur, P.M., McCaughey, J.H., Roulet, N.T., Suyker, A.E., Verma, S.B., Waddington, J.M., and Whiting, G.J. 1998. Relationship between ecosystem productivity and photosynthetically active radiation for northern peatlands. *Global Biogeochemical Cycles,* 12(1): 115-126.

Geological Survey of Canada. 1998. *Living with Frozen Ground: a field guide to permafrost in Yellowknife, Northwest Territories*. Natural Resources Canada, Ottawa, Ontario.

Geological Survey of Canada. 2001. *Northern Peatlands to Monitor Recent and Future Warming: Final Report of CCAF project A313.* Natural Resources Canada, Ottawa, Ontario.

Geological Survey of Canada. 2007. *Permafrost and Climate Change*. Natural Resources Canada, Ottawa, Ontario.

Gomes, F.P., Oliva, M.A., Mielke, M.S., de Almeida, A-A.F., and Leite, H.G. 2006. Photosynthetic irradiance-response in leaves of dwarf coconut palm (Cocos nucifera L. 'nana', Arecaceae): Comparison of three models. *Scientia Horticulturae,* 109(1): 101-105.

Gorham, E. 1991. Northern Peatlands: Role in the Carbon-Cycle and Probable Responses to Climatic Warming. *Ecological Applications,* 1(2): 182-195.

Goudriann, J. 1979. A family of saturation type curves, especially in relation to photosynthesis. *Annals of Botany*, 43: 783-785.

Goulden, M.L., Wofsy, S.C., Harden, J.W., Trumbore, S.E., Crill, P.M., Gower, S.T., Fries, T., Daube, B.C., Fan, S.M., Sutton, D.J., Bazzaz, A., and Munger, J.W. 1998 Sensitivity of boreal forest carbon balance to soil thaw. *Science,* 279(5348): 214-217.

Grau, A. 1995. A Closed-Chamber Technique for Field Measurement of Gas-Exchange of Forage Canopies. *New Zealand Journal of Agricultural Research,* 38(1): 71-77.

Griffis, T.J, Rouse, W.R., and Waddington, J.M. 2000. Interannual variability of net ecosystem CO_2 exchange at a subarctic fen. *Global Biogeochemical Cycles,* 14(4): 1109-1121.

Griffis, T.J. and Rouse, W. 2001. Modeling the interannual variability of net ecosystem CO_2 exchange at a subarctic sedge fen. *Global Change Biology,* 7(5): 511-530.

Groendahl, L., Friborg, T., and Soegaard, H. 2007. Temperature and snow-melt controls on interannual variability in carbon exchange in the high Arctic. *Theoretical and Applied Climatology,* 88(1-2): 111-125.

Guan, X.J., Spence, C., and Westbrook, C.J. 2010. Shallow soil moisture – ground thaw interactions and controls – Part 2: Influences of water and energy fluxes. *Hydrology and Earth System Sciences*, 14: 1387-1400.

Halsey, L.A., Vitt, D.H., and Zoltai, S.C. 1995. Disequilibrium Response of Permafrost in Boreal Continental Western Canada to Climate Change. *Climatic Change,* 30(1): 57.

Hamlin, L., Pietroniro, A., Prowse, T., Soulis, R. and Kouwen, N. 1998. Application of indexed snowmelt algorithms in a northern wetland regime. *Hydrological Processes*, 12: 1641-1657.

Harris, S.A. and Schmidt, I.H. 1994. Permafrost aggradation and peat accumulation since 1200 years B.P. in peat plateaus at Tuchitua, Yukon Territory (Canada). *Journal of Paleolimnology*, 12: 3-17.

Hayashi, M., Quinton, W.L., Pietroniro, A., and Gibson, J.J. 2004. Hydrologic functions of wetlands in a discontinuous permafrost basin indicated by isotopic and chemical signatures. *Journal of Hydrology,* 296(1-4): 81-97.

Hayashi, M., Goeller, N., Quinton, W.L., and Wright, N. 2007. A simple heat-conduction method for simulating the frost-table depth in hydrological models. *Hydrological Processes*, 21: 2610-2622.

Heginbottom, J.A. and Radburn, L.K. 1992. Permafrost and ground ice conditions of northwestern Canada. Geological Survey of Canada, Map 1691A.

Holland, E.A., Braswell, B.H., Lamarque, J.-F, Townsend, A., Sulzman, J., Muller, J.-F., Dentener, F., Brasseur, G., Levy H. II, G., Penner, J.E., and Roelofs, G.-J. 1997. Variations in the predicted spatial distribution of atmospheric nitrogen deposition and their impact on carbon uptake by terrestrial ecosystems. *Journal of Geophysical Research,* 102(D13): 15849-15866.

IPCC. 2007. Climate Change 2007: The Physical Science Basis. *Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change* (ed.) Solomon, S., Qin, D., Manning, M., Chen, Z., Marquis, M., Averyt, K.B., Tignor, M., and Miller H.L. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, 996pp.

Johnson, L.C., Damman, A.W.H., Malmer, N. 1990. Sphagnum macrostructure as an indicator of decay and compaction in peat cores from an ombrotrophic south Swedish peat-bog. *The Journal of Ecology,* 78(3): 633-648.

Joiner, D.W., McCaughey, J.H., Lafleur, P.M., and Bartlett, P.A. 1999. Water and carbon dioxide exchange at a boreal young jack pine forest in the BOREAS northern study area. *Journal of Geophysical Research D: Atmospheres,* 104(22).

Jones, M.H., Fahnestock, J.T., Walker, D.A., Walker, M.D., Welker, J.M. 1998. Carbon dioxide fluxes in moist and dry arctic tundra during season: Responses to increases in summer temperature and winter snow accumulation", *Arctic and Alpine Research,* 30(4): 373-380.

Kindermann, J., Wurth, G., Kohlmaier, G.H., Badeck, F. 1996. Interannual variation of carbon exchange fluxes in terrestrial ecosystems. *Global Biogeochemical Cycles,* 10(4): 737-755.

Konen, M.E., Jacobs, P.M., Burras, C.L., Talaga, B.J., and Mason, J.A. 2002. Equations for predicting soil organic carbon using loss-on-ignition for north central US soils. *Soil Science Society of America Journal,* 669(6): 1878-1881.

Kuhry, P. and Turunen, J. 2006. *The* Postglacial Development of Boreal and Subarctic Peatlands. *Boreal Peatland Ecosystems*, 188: 25-46.

Lafleur, P.M. 1999. Growing season energy and CO_2 exchange at a subarctic boreal woodland. *Journal of Geophysical Research,* 104(D8): 9571-9580.

Lafleur, P.M. 2002. Peatland Carbon Flux Measurements for Model Evaluation and Assessment of Contemporary Sink-Source Status: An Overview. In *Proceedings of the International Workshop on Carbon Dynamics of Forested Peatlands: Knowledge Gaps, Uncertainty, and Modeling Approaches; Long-Term Dynamics and Contemporary Carbon Budget of Northern Peatlands*, Natural Resources Canada: Edmonton, Alberta; 18-20.

Lafleur, P.M., McCaughey, J.H., Joiner, D.W., Bartlett, P.A. and Jelinski, D.E. 1997. Seasonal trends in energy, water, and carbon dioxide fluxes at a northern boreal wetland. *Journal of Geophysical Research,* 102(D24): 29,009-29,020.

Lafleur, P.M., Roulet, N.T., and Admiral, S.W. 2001. Annual cycle of CO_2 exchange at a bog peatland. *Journal of Geophysical Research*, 106(D3): 3071-3081.

Lafleur, P.M. and Humphreys, E.R. 2007. Spring warming and carbon dioxide exchange over low Arctic tundra in central Canada. *Global Change Biology*, 14: 740-756.

Law, B.E., Falge, E., Gu, L., Baldocchi, D.D., Bakwin, P., Berbigier, P., Davis, K., Dolman, A.J., Falk, M., Fuentes, J.D., Goldstein, A., Granier, A., Grelle, A., Hollinger, D., Janssens, I.A., Jarvis, P., Jensen, N.O., Katul, G., Mahli, Y., Matteucci, G., Meyers, T., Monson, R., Munger, W., Oechel, W., Olson, R., Pilegaard, K., Paw U, K.T., Torgeirsson, H., Valentini, R., Verma, S., Vesala, T., Wilson, K., and Wofsy, S. 2002. Environmental controls over carbon dioxide and water vapor exchange of terrestrial vegetation. *Agricultural and Forest Meteorology*, 113: 97-120.

Liblik, L.K., Moore, T.R., Bubier, J.L., and Robinson, S.D. 1997. Methane emissions from wetlands in the zone of discontinuous permafrost: Fort Simpson, Northwest Territories, Canada. *Global Biogeochemical Cycles,* 11(4): 485-494.

Lindroth, A., Grelle, A., and Moren, A. 1998. Long-term measurements of boreal forest carbon balance reveal large temperature sensitivity. *Global Change Biology,* 4(4): 443-450.

Lund, C.P., Riley, W.J., Pierce, L.L., and Field, C.B. 1999. The effects of chamber pressurization on soil-surface CO_2 flux and the implications for NEE measurements under elevated CO_2. *Global Change Biology,* 5(3): 269-281.

Maestre, F. and Cortina, J. 2003. Small-scale spatial variation in soil CO_2 efflux in a Mediterranean semiarid steppe. *Applied Soil Ecology,* 23(3): 199-209.

Makiranta, P., Laiho, R., Fritze, H., Hytonen, J., Laine, J., and Minkkinen, K. 2009. Indirect regulation of heterotrophic peat soil respiration by water level via microbial community structure and temperature sensitivity. *Soil Biology & Biochemistry,* 41(4): 695-703.

Minkkinen, K., Laine, J., Shurpali, N.J., Makiranta, P., Alm, J., and Penttila, T. 2007. Heterotrophic soil respiration in forestry-drained peatlands. *Boreal Environment Research* 12: 115-126.

Moore, T.R. 1989. Plant Production, Decomposition, and Carbon Efflux in a Subarctic Patterned Fen. *Arctic and Alpine Research,* 21(2): 156-162.

Moore, T.R., Roulet, N.T., and Waddington, J.M. 1998. Uncertainty in Predicting the Effect of Climatic Change on the Carbon Cycling of Canadian Peatlands. *Climatic Change,* 40(2): 229-245.

National Wetlands Working Group (NWWG). 1988. *Wetlands of Canada.* Ecological Land Classification Series, No.24. Sustainable Development Branch, Environment Canada, Ottawa, Ontario, and Poly Science Publications, Montreal.

Natural Resources Canada. 2009. The Atlas of Canada: Environment. Natural Resources Canada, Ottawa, Canada.

Nelson, F.E., Lachenbruch, A.H., Woo, M.K, Koster, E.A., Osterkamp, T.E., Gavrilova, M.K. and Cheng, G.D. 1993. Permafrost and Changing Climate. In *Proceedings of the Sixth International Conference on Permafrost: Vol. II,* South China University of Technology Press, Wushan, Guangzhou, China, pp 987-1005.

Oberbauer, S.F., Tweedie, C.E., Welker, J.M., Fahnestock, J.T., Henry, G.H.R., Webber, P.J., Hollister, R.D., Walker, M.D., Kuchy, A., Elmore, E., and Starr, G. 2007. Tundra CO_2 Fluxes in Response to Experimental Warming across Latitudinal and Moisture Gradients. *Ecological Monographs,* 77(2): 221-238.

Oechel, W.C., and Van Cleve, K. 1986. The role of bryophytes in nutrient cycling in the taiga. *In: Van Cleve, K., and Chapin, F.S III, Dryness, C.T., Viereck, L.A., and Flanagan, P.W. (eds.) Forest ecosystems in the Alaskan taiga. A synthesis of structure and function.* Springer Berlin Heidelberg, New York, NY.

Oechel, W.C., Vourlitis, G., Hastings, S., and Bochkarev, S. 1995. Change in Arctic CO_2 flux over two decades: Effects of climate change at Barrow, Alaska. *Ecological Applications,* 5(3): 846-855.

Oechel, W.C., Vourlitis, G.L., Hastings, S.J., Zulueta, R.C., Hinzman, L., and Kane, D. 2000. Acclimation of ecosystem CO_2 exchange in the Alaskan Arctic in response to decadal climate warming. *Nature,* 406(6799): 978-981.

Petrone, R.M., Waddington, J.M., and Price, J.S. 2001. Ecosystem scale evapotranspiration and net CO_2 exchange from a restored peatland. *Hydrological Processes,* 15: 2839-2845.

Petrone, R.M., Waddington, J.M., and Price, J.S. 2003. Ecosystem-scale flux of CO_2 from a restored vacuum harvested peatland. *Wetlands Ecology and Management,* 11(6): 419-432.

Petrone, R.M., Price, J.S., Waddington, J.M., and von Waldow, H. 2004. Surface moisture and energy exchange from a restored peatland, Québec, Canada. *Journal of Hydrology,* 295(1-4): 198-210.

Petrone, R.M., Devito, K.J., Silins, U., Mendoza, C., Brown, S.C., Kaufman, S.C., and Price, J.S. 2008. Transient peat properties in two pond-peatland complexes in the sub-humid Western Boreal Plain, Canada. *Mires and Peat,* 3(3).

Petrone, R.M., Solondz, D.S., Macrae, M.L., Gignac, D., and Devito, K.J. (in review). Microtopographical and canopy cover controls on moss carbon dioxide exchange in a western Boreal Plain peatland. *Ecohydrology.*

Poole, D.K. and Miller, P.C. 1982. Carbon-Dioxide Flux from Three Arctic Tundra Types in North-Central Alaska, USA. *Arctic and Alpine Research,* 14(1): 27-32.

Price, J.S. and Waddington, J.M. 2000. Advances in Canadian wetland hydrology and biogeochemistry. *Hydrological Processes,* 14: 1579-1589.

Quinton, W.L., Hayashi, M., and Chasmer, L.E. (in review). Permafrost thaw induced land-cover change in the Canadian subarctic: Implications for water resources. *Hydrological Processes.*

Quinton, W.L., Hayashi, M., and Chasmer, L.E. 2009. Peatland Hydrology of Discontinuous Permafrost in the Northwest Territories: Overview and Synthesis. *Canadian Water Resources Journal,* 34(4): 311-327.

Quinton, W.L., Gray, D.M., Marsh, P. 2000. Subsurface drainage from hummock-covered hillslopes in the Arctic tundra. *Journal of Hydrology*, 237: 113-125.

Quinton, W.L., Hayashi, M., and Pietroniro, A. 2003. Connectivity and storage functions of channel fens and flat bogs in northern basins. *Hydrological Processes,* 17(18): 3665-3684.

Raich, J.W. and Schlesinger, W.H. 1992. The global carbon dioxide flux in soil respiration and its relationship to vegetation and climate. *Tellus,* 44(B2): 81-99.

Raich, J.W. and Tufekcioglu, A. 2000. Vegetation and soil respiration: Correlations and controls. *Biogeochemistry,* 48(1): 71-90.

Raven, P.H., Evert, R.F., and Eichhorn, S.E. 1999. *Biology of Plants.* 7th Edition. W.H. Freeman and Company, New York, NY.

Robinson, S. 2002. Peatlands of the Mackenzie Valley: Permafrost, Fire, and Carbon Accumulation. In *Proceedings of the International Workshop on Carbon Dynamics of Forested Peatlands: Knowledge Gaps, Uncertainty, and Modeling Approaches; Long-Term Dynamics and Contemporary Carbon Budget of Northern Peatlands*, Natural Resources Canada: Edmonton, Alberta; 21-24.

Robinson, S.D. and Moore, T.R. 2000. The influence of permafrost and fire upon carbon accumulation in high boreal peatlands, Northwest Territories, Canada. *Arctic, Antarctic, and Alpine Research,* 32(2): 155-166.

Robinson, S.D., Turetsky, M.R., Kettles, I.M., and Wieder, R.K. 2003. Permafrost and peatland carbon sink capacity with increasing latitude. In *Proceedings of the 8th International Conference of Permafrost*, Zurich, Switzerland, Balkema Publishers, 2: 965-970.

Rosenberry, D.O., Glaser, P.H., and Siegel, D.I. 2006. The hydrology of northern peatlands as affected by biogenic gas: current developments and research needs. *Hydrological Processes,* 20(17): 3601-3610.

Roulet, N.T. 1991. Surface Level and Water Table Fluctuations in a Subarctic Fen. *Arctic and Alpine Research,* 23(3): 303-310.

Roulet, N.T., Ash, R., and Moore, T.R. 1992. Low boreal wetlands as a source of atmospheric methane. *Journal of Geophysical Research*, 97: 3739-3749.

Rouse, W.R., Douglas, M.S.V., Hecky, R.E., Hershey, A.E., Kling, G.W., Lesack, L., Marsh, P., McDonald, M., Nicholson, B.J., Roulet, N.T., and Smol, J.P. 1997. Effects of climate change on the freshwaters of arctic and subarctic North America. *Hydrological Processes,* 11(8): 873-902.

Rouse, W.R., Bello, R.L., D'Souza, A., Griffis, T.J., and Lafleur, P.M. 2002. The annual carbon budget for fen and forest in a wetland at Arctic treeline. *Arctic,* 55(3): 229-237.

Ruddiman, W.F. 2002. *Earth's Climate: Past and Future*. W.H. Freeman and Company, New York, NY.

Schlesinger, W.H. 1977. Carbon balance in terrestrial detritus. *Annual Review of Ecology and Systematics,* 8: 51-81.

Schlesinger, W.H. and Andrews, J.A. 2000. Soil respiration and the global carbon cycle. *Biogeochemistry,* 48(1): 7-20.

Schreader, C.P., Rouse, W.R., Griffis, T.J., Boudreau, L.D., and Blanken, P.D. 1998. Carbon dioxide fluxes in a northern fen during a hot, dry summer. *Global Biogeochemical Cycles,* 12(4): 729-740.

Shurpali, N.J., Verma, S.B., and Kim, J. 1995. Carbon dioxide exchange in a peatland ecosystem. *Journal of Geophysical Research*, 100(D7): 14,319-14,326.

Silvola, J., Alm, J., Ahlholm, U., Nykanen, H., and Martikainen, P.J. 1996. CO_2 fluxes from peat in boreal mires under varying temperature and moisture conditions. *Journal of Ecology,* 84(2): 219-228.

Simek, M., Elhottova, D., Klimes, F., and Hopkins, D.W. 2004. Emissions of N2O and CO_2, denitrification measurements and soil properties in red clover and ryegrass stands. *Soil Biology and Biochemistry,* 36(1): 9-21.

Solondz, D.S., Petrone, R.M., and Devito, K.J. 2008. Forest floor carbon dioxide fluxes within an upland-peatland complex in the Western Boreal Plain, Canada. *Ecohydrology,* 1: 361-376.

Streever, W.J., Genders, A.J., and Cole, M.A. 1998. A closed chamber CO_2 flux method for estimating marsh productivity. *Aquatic Botany,* 62(1): 33-44.

Strom, L. and Christensen, T. 2007. Below ground carbon turnover and greenhouse gas exchanges in a sub-arctic wetland. *Soil Biology and Biochemistry,* 39(7): 1689-1698.

Sturm, M., Schimel, J., Michaelson, G., Welker, J.M., Oberbauer, S.F., Liston, G.E., Fahnestock, J., and Romanovsky, V.E. 2005. Winter biological processes could help convert arctic tundra to shrubland. *Bioscience,* 55(1): 17-26.

Suyker, A.E., Verma, S.B., and Burba, G.G. 2003. Interannual variability in net CO_2 exchange of a native tallgrass prairie. *Global Change Biology,* 9(2): 255-265.

Swanson, R.V. and Flanagan, L.B. 2001. Environmental regulation of carbon dioxide exchange at the forest floor in a boreal black spruce ecosystem. *Agricultural and Forest Meteorology,* 108(3): 165-181.

Tarnocai, C. 1980. Canadian wetland registry. In *Proceedings, Workshop on Canadian Wetlands*. (Eds.) Rubec, C.D.A. and Pollett, F.C. Land Directorate, Environment Canada. Ecological Land Classification Series No.12. Ottawa, Ontario.

Tarnocai, C. 2006. The effect of climate change on carbon in Canadian peatlands. *Global and Planetary Change*, 53: 222-232.

Tarnocai, C., Ping, C.L. and Kimble, J. 2007. Carbon Cycles in the Permafrost Region of North America. In *The First State of the Carbon Cycle Report (SOCCR): The North American Carbon Budget and Implications for the Global Carbon Cycle*. A Report by the U.S. Climate Change Science Program and the Subcommittee on Global Change Research (eds.) King, A.W., Dilling, L., Zimmerman, G.P., Fairman, D.M., Houghton, R.A., Marland, G., Rose, A.Z. and Wilbanks, T.J. National Oceanic and Atmospheric Administration, National Climatic Data Center, Asheville, NC, USA, pp. 127-138.

Thormann, M.N. and Bayley, S.E. 1997. Aboveground net primary production along a bog-fen-marsh gradient in southern boreal Alberta, Canada. *Ecoscience*, 4(3): 374-384.

Thormann, M.N., Bayley, S.E., and Currah, R.S. 2002. The Microbial Wildcard in Peatland Carbon Storage: Implications for Global Warming. In *Proceedings of the International Workshop on Carbon Dynamics of Forested Peatlands: Knowledge Gaps, Uncertainty, and Modeling Approaches; Long-Term Dynamics and Contemporary Carbon Budget of Northern Peatlands*, Natural Resources Canada: Edmonton, Alberta; 25-28.

Trumbore, S.E., Bubier, J.L., Harden, J.W., and Crill, P.M. 1999. Carbon cycling in Boreal wetlands: A comparison of three approaches. *Journal of Geophysical Research,* 104(D22): 27,673-27,682.

Turetsky, M.R., Wieder, R.K., Vitt, D.H., Evans, R.J., and Scott, K.D. 2007. The disappearance of relict permafrost in boreal North America: effects on peatland carbon storage and fluxes. *Global Change Biology,* 13: 1922-1934.

U.S. Arctic Research Commission Permafrost Task Force. 2003. *Climate Change, Permafrost, and Impacts on Civil Infrastructure*. Special Report 01-03, U.S. Arctic Research Commission, Arlington, Virginia.

Van Cleve, K., Dyrness, C.T., Viereck, L.A., Fox, J., Chapin, F.S III, and Oechel, W. 1983. Taiga ecosystems in interior Alaska. *Bioscience,* 33(1): 39-44.

Vardy, S.R., Warner, B.G., Turunen, J., and Aravena, R. 2000. Carbon accumulation in permafrost peatlands in the Northwest Territories and Nunavut, Canada. *Holocene,* 10(2): 273-280.

Vitt, D.H., Halsey, L.A., and Zoltai, S.C. 1994. The bog landforms of continental western Canada in relation to climate and permafrost patterns. *Arctic & Alpine Research,* 26(1): 1-13.

Vourlitis, G.L., Verfaillie, J., Oechel, W.C., Hope, A., Stow, D., and Engstrom, R. 2003. Spatial variation in regional CO_2 exchange for the Kuparuk River Basin, Alaska over the summer growing season. *Global Change Biology,* 9(6): 930-941.

Waddington, J.M. and Roulet, N.T. 1996. Atmosphere-wetland carbon exchanges: Scale dependency of CO_2 and CH4 exchange on the developmental topography of a peatland. *Global Biogeochemical Cycles*, 10(2): 233-245.

Waddington, J.M. and Roulet, N.T. 2000. Carbon balance of a boreal patterned peatland. *Global Change Biology,* 6(1): 87-97.

Waddington, J.M., Griffis, T.J., and Rouse, W.R. 1998. Northern Canadian wetlands: Net ecosystem CO_2 exchange and climatic change. *Climatic Change,* 40(2): 267-275.

Waddington, J.M., Rochefort, L., and Campeau, S. 2003. Sphagnum production and decomposition in a restored cutover peatland. *Wetlands Ecology and Management,* 11: 85-95.

Watson, R., Noble, L., Brolin, B., Ravindranath, N.H., Verardo, D.J., and Dokken, D.J. 2000. Land use, land-use change, and forestry. *A special report to IPCC.* Cambridge University Press, Cambridge.

Western, A.W., Bloschl, G., and Grayson, R.B. 2001. Toward capturing hydrologically significant connectivity in spatial patterns. *Water Resources Research*, 37(1): 83-97.

Woo, M. 1986. Permafrost hydrology in North America. *Atmosphere-Ocean,* 24(3): 201-234.

Wright, N., Quinton, W.L., and Hayashi, M. 2008. Hillslope runoff from an ice-cored peat plateau in a discontinuous permafrost basin, Northwest Territories, Canada. *Hydrological Processes,* 22(15): 2816-2828.

Wright, N., Hayashi, M., and Quinton, W.L. 2009. Spatial and temporal variations in active layer thawing and their implication on runoff generation in peat-covered permafrost terrain. *Water Resources Research,* 45(5).

Yu, Z. 2002. Does the Convex Pattern of Peat Accumulation Matter in Understanding Peat Carbon Dynamics. In *Proceedings of the International Workshop on Carbon Dynamics of Forested Peatlands: Knowledge Gaps, Uncertainty, and Modeling Approaches; Long-Term Dynamics and Contemporary Carbon Budget of Northern Peatlands*, Natural Resources Canada: Edmonton, Alberta; 53-56.

Yu, Z., Apps, M.J., and Bhatti, J.S. 2002. Introduction: Challenges and Opportunities in Understanding and Modeling Peatland Dynamics. In *Proceedings of the International Workshop on Carbon Dynamics of Forested Peatlands: Knowledge Gaps, Uncertainty, and Modeling Approaches; Long-Term Dynamics and Contemporary Carbon Budget of Northern Peatlands*, Natural Resources Canada: Edmonton, Alberta; 1-6.

Zimov, S.A., Davidov, S.P., Voropaev, Y.V., and Prosiannikov, S.F. 1996. Siberian CO_2 efflux in winter as a CO_2 source and cause of seasonality in atmospheric CO_2. *Climatic Change*, 33: 111-120.

Zoltai, S.C. 1993. Cyclic development of permafrost in the peatlands of northwestern Alberta, Canada. *Arctic & Alpine Research,* 25(3): 240-246.

Appendix

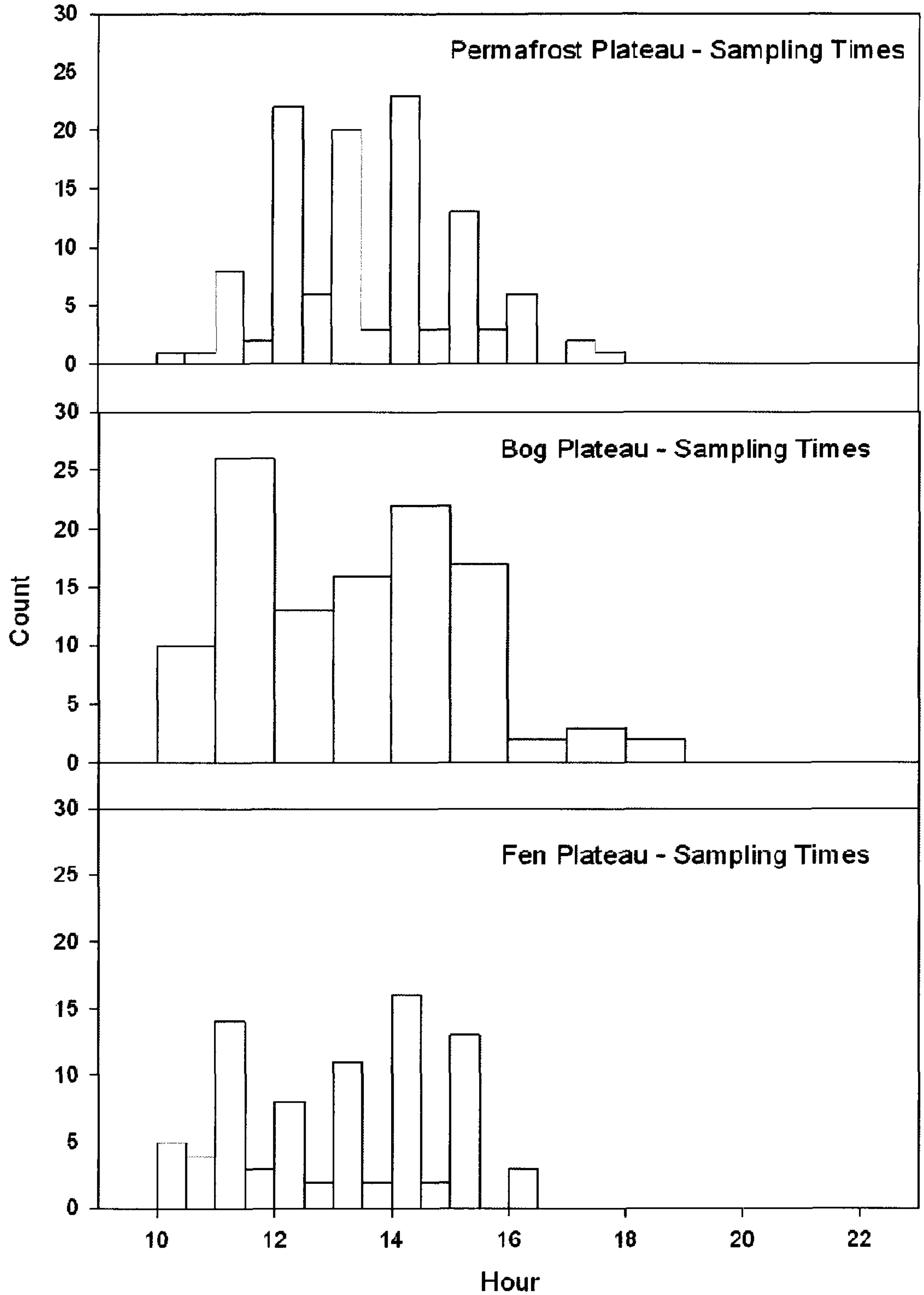

Figure A-1 Frequency distribution of sampling times during daylight hours: permafrost plateau (n=108), flat bog (n=106), and channel fen (n=83) for the 2008 field season at Scotty Creek, Northwest Territories, Canada.