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Authors: McLaughlin, Jim, and Webster, Kara

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Effects of Climate Change on Peatlands in the Far North of Ontario, Canada: a Synthesis

Jim McLaughlin and*
Kara Webster†

*Corresponding author: Ontario Forest Research Institute, Ontario Ministry of Natural Resources, 1235 Queen Street East, Sault Ste. Marie, Ontario, P6A 2E5, Canada,
jim.mclaughlin@ontario.ca
†Canadian Forest Service, Great Lakes Forestry Centre, 1219 Queen Street East, Sault Ste. Marie, Ontario, P6A 2E5, Canada

Abstract

The Hudson Bay Lowlands (HBL) is the largest peatland complex in North America. More than 75% of the HBL occurs in Ontario, where the provincial government mandates that ecosystem carbon storage and sequestration be considered in land-use planning. Accomplishing this task requires identifying carbon indicators and assessing their responses to changing ecosystem processes, such as succession, permafrost thaw, and evapotranspiration (ET). Therefore, we synthesized information on peat carbon indicators and ecosystem process from the literature. Findings indicate that the long-term carbon accumulation, carbon dioxide (CO_2) sequestration, peat depth, and peatland age were similar ($p > 0.10$) between dry and wet peatland features. Furthermore, CO_2 sequestration displayed the highest variability and ponds were net CO_2 emitters. Recent carbon accumulation, CH_4 emission, and ET were highest ($p < 0.01$) in wet features, with CH_4 emission displaying wide variation. Increased active layer thickness (105 ± 92 cm per 100 years) in permafrost was the most variable ecosystem process analyzed in this study, while variation in permafrost loss (53 ± 23% per 100 years) was similar to that of carbon accumulation and ET rates. Processes creating wet and pond conditions may increase landscape-scale CO_2 and CH_4 emissions to the atmosphere, weakening peatland carbon sinks. Dry conditions may reduce CH_4 emissions but potentially increase peatland susceptibility to fire. Knowledge of these changes should be useful for climate change vulnerability and adaptation assessments for large landscapes. However, better understanding of variability in CO_2 sequestration, CH_4 emission, and permafrost dynamics is required to design such assessments for small landscapes.

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Introduction

Northern peatlands account for approximately 25% of the global soil carbon (C) pool, with most of the peatlands occurring in Russia and Canada (Yu, 2012). Peatlands are categorized as ombrotrophic bogs (receiving water primarily from precipitation) and minerotrophic fens (receiving water from precipitation and groundwater inflow). Bogs and fens cycle C differently (Bauer et al., 2009), and their C sink strength may change as peat warms and becomes drier or wetter (Webster et al., 2013). However, vulnerability and adaptation of peatland C budgets in a changing climate are largely unknown (Frolking et al., 2011).

The Far North of Ontario (located between approximately 50 and 57°N, 79 and 94°W) is a highly variable landscape that consists of about 45 million hectares distributed relatively evenly between the Canadian Shield and Hudson Bay Lowlands (HBL). Here, peatlands are the dominant land class, accounting for nearly 50% of the area (Fig. 1). These peatlands store an estimated 36 Gt (1 Gt = 1 Pg = 10^{15} g) of C as peat (Far North Science Advisory Panel, 2010), which is approximately 25% of the C stored in all of Canada's peatlands. The HBL, which contains nearly 75% of the C stored in the Far North, is the largest peatland complex and has the southernmost extent of non-alpine permafrost (subsurface earth materials that remain below 0 °C for two or more consecutive years [Brown, 1967]; see Table 1 for permafrost features and their definitions) in North America (Martini, 2006).

Peatlands in the HBL have been important in Holocene climate cooling because since deglaciation they have sequestered large amounts of atmospheric carbon dioxide (CO_2) (van Bellen et al., 2011; Bunbury et al., 2012). However, when disturbed, these peatlands may release much of this C back to the atmosphere as CO_2 or methane (CH_4), which may speed global warming (Webster and McLaughlin, 2011). For example, although warmer and drier peat may increase tree and shrub biomass (Weltzin et al., 2003), peat decomposition and CO_2 loss may dominate over its gain through net primary production (NPP). In contrast, warmer and wetter peat enhances sedge and moss biomass, characteristic of fens that support rapid C accumulation in, and CH_4 emission from, peat (Webster et al., 2013). Therefore, although CO_2 is sequestered and accumulated as peat, CH_4 losses may offset C gain and enhance atmospheric global warming potential (GWP; sum CO_2 and CH_4 fluxes expressed as CO_2 equivalents].

Ontario House Bill 191, the Far North Act (http://www.ontla.on.ca/bills/bills-files/39_Parliament/Session2/b191ra.pdf), which achieved Royal Assent in October 2010, describes four land-use planning objectives. One of those objectives is "the maintenance of biological diversity, ecological processes and ecological functions, including the storage and sequestration of carbon in the Far North." Furthermore, Far North land-use planning teams may request the development of several policy statements, one of which concerns "ecological systems processes and functions, including considerations for cumulative effects and for climate change adaptation and mitigation."

Achieving the act's planning objective and developing an appropriate policy statement for land-use planning requires an understanding of how climate change affects peatland C storage and sequestration at the landscape scale. Spatial configurations of permafrost and peatland types, succession and plant communi-

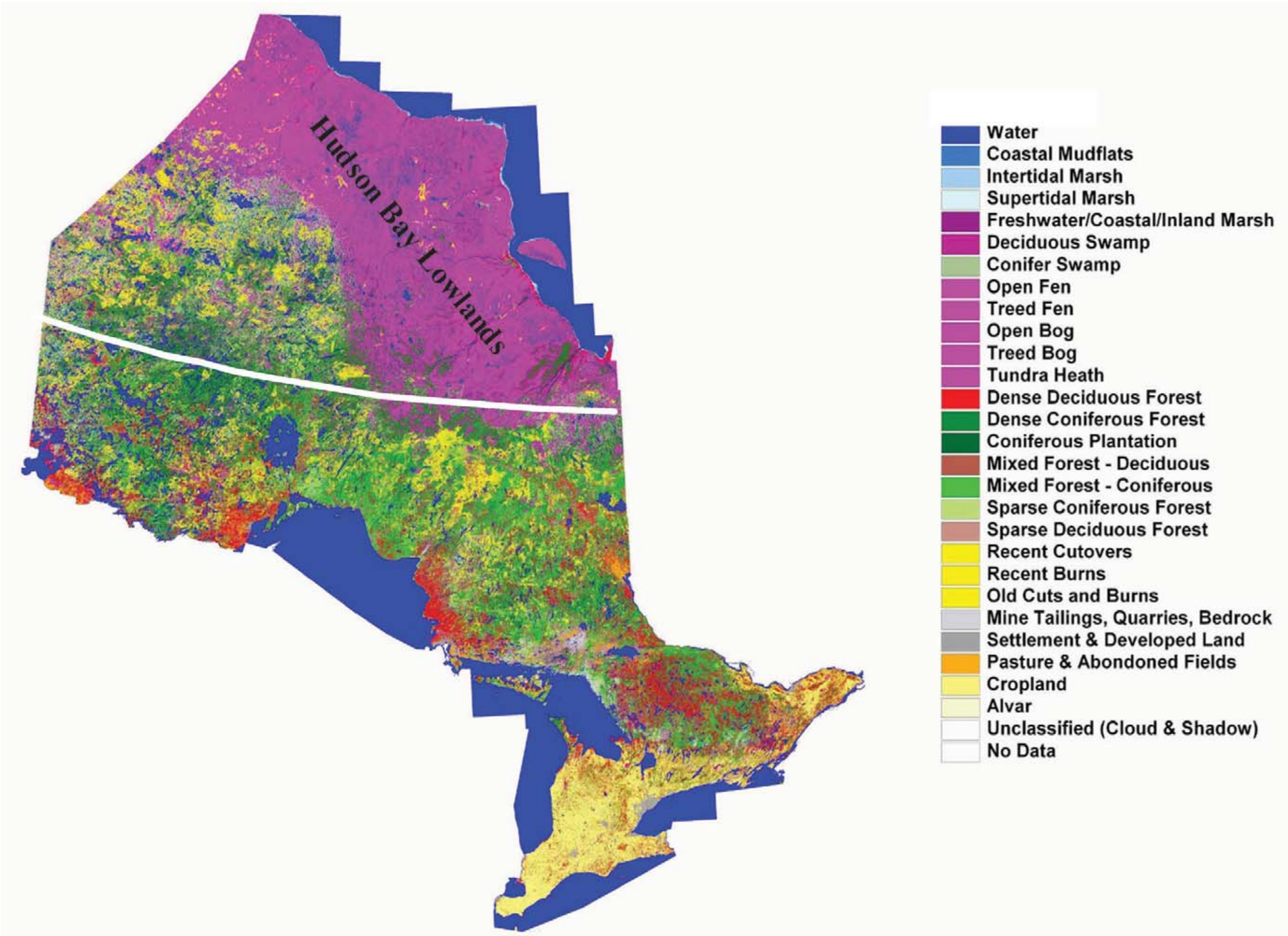


FIGURE 1. Ontario land cover map. Solid white line approximates the Far North of Ontario boundary (adapted from OMNR, 2000, with permission).

TABLE 1

Features common in the continuous (CPZ) and discontinuous (DPZ) permafrost zones in the Far North of Ontario.

Permafrost feature	Characteristics	Zone of occurrence
	Aggrading	
Pingo	Land elevated due to ice-wedge growth	CPZ
Polygonal peatland	Land elevated due to cracks in shrinking permafrost	CPZ
Palsa	Peat elevated due to ground ice expansion; ice penetration into mineral sediment	CPZ, DPZ
Peat plateau	Peat elevated due to ground ice expansion; ice core only in peat; coalesced palsas	CPZ, DPZ
Lithalsa	Mineral sediment elevated due to ground ice expansion	CPZ, DPZ
	Degrading	
Thermokarst		
Pond	Surface inundated due to ice melt	CPZ, DPZ
Thaw lake	Ground collapsed following thawing of ground ice in regions underlain by permafrost resulting in shallow body of water	CPZ, DPZ
Internal lawn	Area inundated outside peat plateau or palsa due to melting of ice	DPZ
Collapse scar	Internal regions of a peat plateau collapse where ice has melted	CPZ, DPZ
Talik	Patch of permanently unfrozen ground underneath thaw lakes	CPZ, DPZ

ties, along with subsequent evapotranspiration (ET) are important controls on landscape C storage and sequestration. For example, plant communities alter peatland C cycling through plant photosynthesis and the quantity and quality (i.e., decomposability) of litter added to peat (Laiho, 2006). Permafrost amount and thaw rates can also alter landscape-scale C budgets by creating drier or wetter conditions (Malmer et al., 2005). The latter inhibits organic matter decomposition relative to plant NPP, but enhances CH_4 emissions, whereas the opposite occur when peat dries (Johansson et al., 2006b). Furthermore, ET is positively correlated with CO_2 sequestration under wet conditions (Sonnentag et al., 2010). However, dry peat created by prolonged elevated ET relative to water input is also more susceptible to fires that burn deeper into peat (Turetsky et al., 2004) than occur in current HBL landscapes (Balshi et al., 2007), further contributing net peatland C losses to the atmosphere.

In this paper, we synthesize the processes controlling peatland C accumulation and greenhouse gas (GHG) exchange in the HBL. The approach was to compile, analyze, and discuss the data on peatland C fluxes, succession, permafrost, and ET from the literature. We conclude by presenting a climate change vulnerability and adaptation assessment framework for HBL peatlands, discussing challenges to its development and its application in land-use planning.

Peatland Carbon Indicators

Successful communication among researchers, policymakers, and land managers requires common terminology. For this paper, peat C content is defined as mass of C contained in a unit area (e.g., kg C m^{-2}) of dry peat. Carbon content is calculated

TABLE 2

Number of peatland studies reporting data on carbon accumulation and sequestration and methane release that were used in this synthesis by peatland type/feature.

Peatland type/feature	Number of studies	Total number of data points	Number of data points assessed in this study
Long-term carbon accumulation rate			
Overall	26	146	26
Bog	11	88	11
Fen	9	44	9
Thermokarst features	6	10	2
Palsa/peat plateau	3	4	4
Ponds	0	0	0
Recent carbon accumulation rate			
Overall	7	112	24
Bog	4	75	9
Fen	2	12	7
Thermokarst features	2	7	3
Palsa/peat plateau	1	18	5
Ponds	0	0	0
Carbon dioxide sequestration			
Overall	29	201	40
Bog	8	42	11
Fen	14	123	17
Thermokarst features	4	5	4
Palsa/peat plateau	4	12	4
Ponds	3	19	4
Methane emission			
Overall	22	265	58
Bog	5	36	8
Fen	9	76	23
Thermokarst features	5	29	5
Palsa/peat plateau	5	35	7
Ponds	11	89	15

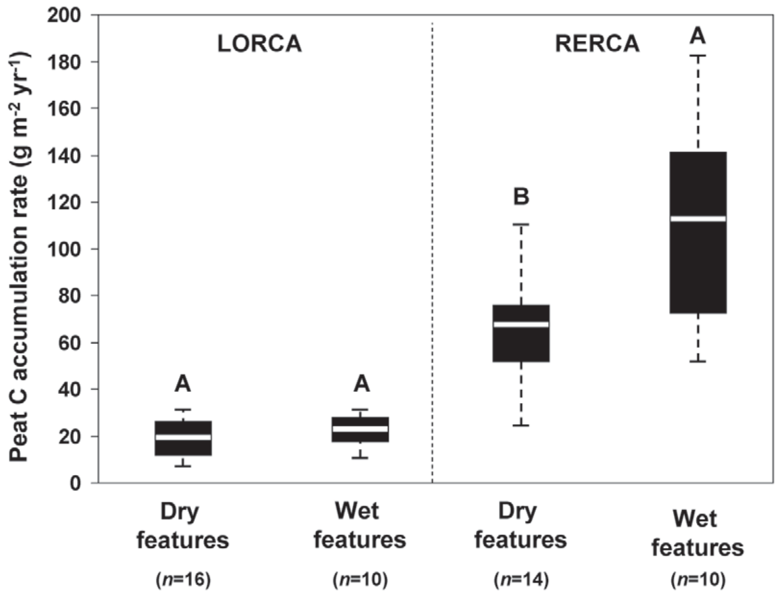


FIGURE 2. Box plots of long-term rates of carbon accumulation (LORCA) and recent rates of carbon accumulation (RERCA), where the box height indicates the interquartile range, the solid white line within the box represents the median value, and the dashed lines from the top and bottom extend to the highest and lowest values, respectively. Different letters for LORCA and RERCA, respectively, indicate statistical differences at p < 0.05.

as the product of C concentration, peat bulk density, and sample increment thickness. Peat C storage is defined as the product of C content, area of peatland in a landscape, and peat depth—the latter not well documented in subarctic peatlands. Carbon dioxide sequestration is defined as short-term (annual to decadal) balance of atmospheric CO_2 uptake by plants and its release through plant respiration and organic matter decomposition. Carbon dioxide sequestration is commonly measured using gas flux chambers and eddy covariance flux towers. Carbon accumulation is defined as net C gain in peat at century to millennial scales. Peat C accumulation is measured by (1) dating of surface peat with specified tracers, such as ^{210}Pb and ^{137}Cs, that provide decadal to century (e.g., 50 to 200 years) time scales of peat accumulation (i.e., recent rate of C accumulation [RERCA]; Oldfield et al., 1995) and (2) radiocarbon (^{14}C) dating of basal peat and peat profile age/depth relationships to estimate century to millennial C accumulation (i.e., long-term rates of C accumulation [LORCA]; Kuhry and Vitt, 1996). These rates are considered apparent because without intensively dating individual peat cores and identifying wetness using paleoecological indicators, timelines for and causes of LORCA and RERCA changes during peatland development cannot be inferred (Yu, 2012). However, differentiating sequestration from accumulation is critical in climate change vulnerability assessments because CO_2 sequestration in plant biomass and surface litter does not equate to long-term C accumulation as peat.

Data sets for peatland C indicators analyzed in this study were compiled from 79 published papers; the number of studies and data points available for LORCA (g C m^{-2} yr^{-1}), RERCA (g C m^{-2} yr^{-1}), CO_2 sequestration (g C m^{-2} d^{-1}), and CH_4 emission (mg C m^{-2} d^{-1}) are shown in Table 2. A total of 680 values were retrieved from the publications, with approximately 70% corresponding to GHG exchange and the remainder to C accumulation rates. Mean values were calculated for each study reporting multiple rates by peatland type; this resulted in 129 data points to conduct the statistical analyses described below. Because of small sample sizes and infrequent differences among bog and fen hummocks, permafrost palsas, peat plateaus, and polygonal peatlands, they were classified as “dry features.” Similarly, collapse scars, lawns, carpets, hollows, and pools in bogs and fens were classified as “wet features.” Although only four ponds were represented in the final data set for CO_2 sequestration, they were considered a separate category because all were net CO_2 and CH_4 sources to the atmosphere. A meta-analysis to assess the effects of experimental manipulations on C fluxes was beyond the scope of this paper, as the primary interest was an initial evaluation of potential C indicator variability. Meta-analysis is, however, recommended as the next step in peatland C indicator selection.

Differences in LORCA and RERCA between dry and wet features were assessed with the Mann-Whitney test, while a Kruskal-Wallis (Conover, 1980) test was used to assess differences in CO_2 sequestration and CH_4 emission from wet, dry, and pond features. Coefficient of variation (CV) and sample sizes were used to calculate the power (probability of failing to reject a false null hypothesis; Type II error) of the statistical tests. In addition, sample sizes required to detect a 20% difference in mean values of each indicator at the 0.05 level of significance (α) and a power of 0.70 were calculated following the methods of McLaughlin and Phillips (2006). Quartile analysis was used to define the bounds of natural variation for the indicators using the inter-quartile range between the first through third quartiles (25th through 75th percentiles), to which data compiled for HBL peatlands were compared.

Long-term C accumulation rate is summarized in Table 3; it was the most promising indicator (power = 0.78) analyzed in this study. No difference ($p = 0.281$) occurred between dry and wet features; therefore, data were combined into one peatland class. Natural variation of LORCA ranged from 18.0 to 24.5 g C m^{-2} yr^{-1} (Fig. 2), and its mean value in HBL peatlands (16.4 ± 4.9 g C m^{-2} yr^{-1}; $n = 11$) was below the 25th percentile (18.0 g C m^{-2} yr^{-1}) of the overall data set. Although the power of the Mann-Whitney test was 0.40 for RERCA, it was higher ($p = 0.009$) in wet than dry features (Fig. 2). Furthermore, mean RERCA in HBL peatlands (77 ± 42 g C m^{-2} yr^{-1}; $n = 9$) was similar to the 50th percentile (78 g C m^{-2} yr^{-1}) of the overall data set. The above results suggest RERCA may be a viable decade-to-century indicator of

TABLE 3

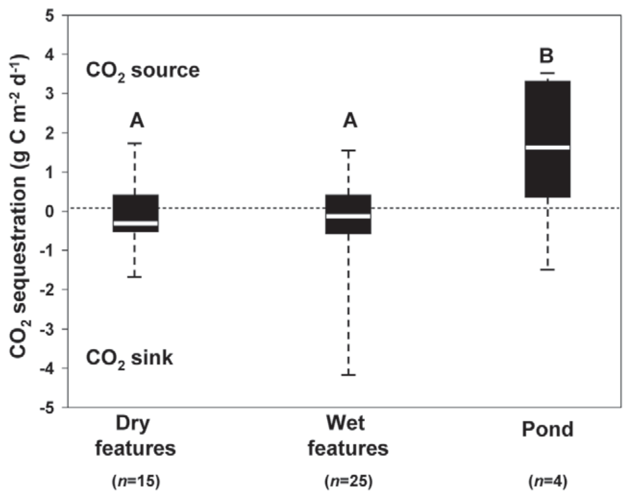
Recent and long-term carbon accumulation rates reported for northern peatlands.

Peatland type	Location	Long-term carbon accumulation ($g\ C\ m^{-2}\ yr^{-1}$)	Recent carbon accumulation ($g\ C\ m^{-2}\ yr^{-1}$)	References
Bogs	Ontario	10 to 25.5	55 to 178	Turunen et al. (2004); Bunbury et al. (2012)
	Québec	14.4 to 34.6	40 to 126	Turunen et al. (2004); Loisel and Garneau (2010); van Bellen et al. (2011)
	Western Canada (9 sites)	13.6 to 34.9		Kuhry and Vitt (1996)
	Saskatchewan	20.5 to 38.5	53 to 115	Kuhry et al. (1992); Kuhry (1994); Bauer et al. (2009)
	Alberta	19.7	53 to 125	Kuhry and Vitt (1996); Turetsky et al. (2000)
	Northwest Territories	19.1 to 32.3		Robinson and Moore (2000); Robinson (2006)
	Finland	16.0 to 35.3		Tolonen and Turunen (1996); Turunen et al. (2002)
	Siberia	5.4 to 34.9		Turunen et al. (2001); Borren et al. (2004); Beilman et al. (2009)
Fens	Québec		42 to 184	Pendea and Chmura (2012)
	Alberta	25.5 to 37.5		Kubiw et al. (1989); Nicholson and Vitt (1990); Yu et al. (2003); Yu (2006)
	Saskatchewan	13.2 to 33.6	74 to 78	Bauer et al. (2009)
	Northwest Territories	18.1 to 33.6		Robinson and Moore (1999)
	Alaska	29.0 to 38.0		Klein et al. (2013)
	Finland	15.6 to 24.9		Tolonen and Turunen (1999); Mäkilä et al. (2001); Turunen et al. (2002)
	Siberia	19.9 to 33.6		Borren et al. (2004)
Palsa/Peat plateaus	Québec	24.2	61 to 133	Ali et al. (2008); Lamarre et al. (2012)
	Manitoba		25 to 140	Camill et al. (2001)
	Saskatchewan	12.5 to 12.7	55 to 57	Turetsky et al. (2007)
	Alberta	19.7	55 to 95	Turetsky et al. (2000, 2007)
	Northwest Territories	10.8 to 18.4		Robinson and Moore (2000); Vardy et al. (2000); Robinson (2006)
	Sweden	29.5		Kokfelt et al. (2010)
Thermokarst	Québec		131	Ali et al. (2008)
	Manitoba		110 to 170	Camill et al. (2001)
	Northwest Territories	26.3		Robinson (2006)
	Alaska	13.1 to 35.2		Jones et al. (2012)

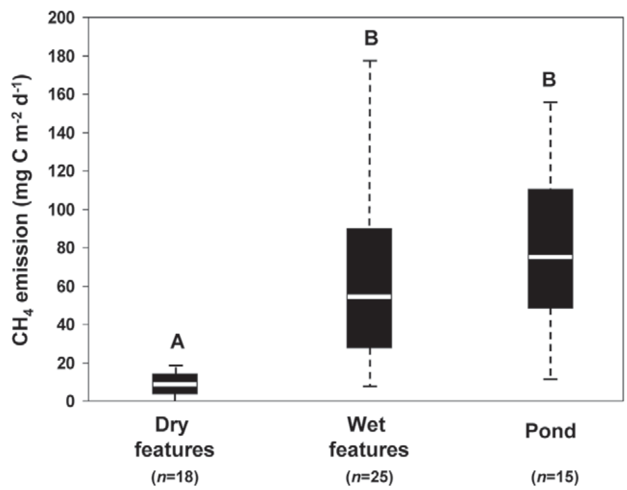
C accumulation. However, it would be beneficial to conduct additional RERCA measurements in boreal and subarctic peatlands, including the HBL, to increase statistical power.

Greenhouse gas fluxes are summarized in Table 4; CO_2 sequestration (power = 0.17) and CH_4 emission (power = 0.24) were characterized by high variation (Fig. 3, parts a and b) and low statistical power. Furthermore, CO_2 sequestration did not differ ($p = 0.766$) between dry and wet features, and they were weak CO_2 sinks (Fig. 3, part a). Ponds may be a CO_2 source; however, statistical inferences are limited by the sample size ($n = 4$) and caution is recommended when interpreting the pond CO_2 exchange data. For example, ponds covering mineral sediments are large CO_2 sources to the atmosphere, whereas those over organic sediments alternate between being CO_2 sinks and sources and tend to be smaller sources when assessed on an annual basis (Macrae et al., 2004). Differences in mineral and organic sediments should, therefore, be considered in scaling and modeling CO_2 sequestration in peatland ponds.

Mean CO_2 sequestration in dry features of the HBL (–0.62 ± 1.04 g CO_2-C m^{-2} d^{-1}) was similar to the 50th percentile (–0.64 g CO_2-C m^{-2} d^{-1}) of the overall data set (Fig. 3, part a). However, 7 of the 11 data points from the HBL analyzed here came from the 1990 Northern Wetland Studies in the southern HBL (Glooschenko et al., 1994) and may not be applicable to the entire HBL or its current landscape structure. Nearly all the peatlands (97%) analyzed in this study were net emitters of CH_4 to the atmosphere, with wet features and ponds releasing approximately 25 times more CH_4 than dry features ($p < 0.0001$; Fig. 3, part b). Variation in CH_4 emission was least in ponds (CV = 0.55), followed by wet (CV = 0.77) and dry (CV = 0.91) features. In the HBL, mean CH_4 emission from wet (41 ± 41 mg CH_4-C m^{-2} d^{-1}, $n = 10$) and dry (5.4 ± 4.4 mg CH_4-C m^{-2} d^{-1}, $n = 8$) features, as well as ponds (87 ± 41 mg CH_4-C m^{-2} d^{-1}, $n = 8$), fell within the bounds of natural variation established in this study (Fig. 3, part b).



(A)



(B)

FIGURE 3. Box plots of greenhouse gas fluxes from boreal and subarctic peatlands to the atmosphere calculated from the literature: (a) CO_2, where the box height indicates the interquartile range, the solid white line within the box represents the median value, and the dashed lines from the top and bottom extend to the highest and lowest values, respectively, and (b) CH_4, where the box height indicates the interquartile range, the solid white line within the box represents the median value, and the dashed lines from the top and bottom extend to the highest and lowest values, respectively. Different letters for CO_2 and CH_4, respectively, indicate statistical differences at $p < 0.05$.

Ecological Processes and Carbon Associations

PEATLAND SUCCESSION

Peatland development in the HBL was primarily controlled by recession of the Tyrell Sea and subsequent isostatic rebounding of the land (Glaser et al., 2004). These peatlands emerged more than 7000 years ago, although ^{14}C basal peat dates produced variable ages (Table 5). The results reported in Table 5 were summarized from 54 peat depth and 42 ^{14}C basal peat dates, respectively, collected in the Hudson Bay region. These data produced mean peat depth and age of 236 ± 92 cm and 4826 ± 1356 ^{14}C years, respectively, and neither depth ($p = 0.256$) nor age ($p = 0.523$) differed among peatland types; statistical power was near 0.70 for both depth and age. Natural variation for the combined data was 168 to 266 cm for depth and 4003 to 5810 for ^{14}C age. Variation in both depth (CV = 0.39) and age (CV = 0.28) was comparable to variation of LORCA.

In the HBL, peat and C accumulation rates were generally high, between 6500 and 4200 cal yr BP (calibrated years before present), then decreased as rich and intermediate fens transformed to poor fens, with *Larix laricina* (larch) and ericaceous plants present (Arlen-Pouliot and Bhiry, 2005; Bunbury et al., 2012). Since 2500 cal yr BP, poor fens generally transformed to bogs, and in the northern HBL bogs may have contained permafrost. In general, bogs may have accumulated the least amount of peat during development (Arlen-Pouliot and Bhiry, 2005). Further south, low peat and C accumulation rates were also reported during bog and poor fen successional stages (Bunbury et al., 2012). Here, however, the fastest accumulation rates occurred between 1000 and 600 cal yr BP and have since declined. At the southernmost portion of the HBL (Eastmain region of Québec), high C accumulation rates occurred in bogs during dry conditions because of high bulk density and C concentration in buried wood (Loisel and Garneau, 2010). Also, rapid accumulation rates during wet conditions were dependent on the presence of hummock-mosses because of their slow decomposition rates (van Bellen et al., 2011).

Peat temperature and wetness are important controls on peatland succession, and plant communities can evolve rapidly in response to modified peatland and permafrost conditions. For example, sampling in a permafrost peatland complex in northern Sweden over 30 to 50 years showed more trees, ericaceous shrubs, and hummock-mosses in drier areas after permafrost thaw (Malmer et al., 2005). However, sedges expanded at the landscape scale due to spreading of wet areas at the cost of hummocks as permafrost thawed, contributing to increased CO_2 sequestration and CH_4 emissions (Johansson et al., 2006b). Resampling of another Swedish peatland complex indicated that tree cover increased from less than 5% of plots initially sampled to approximately 30% of plots sampled 42 years later (Gunnarsson et al., 2002). Percentage of plots containing mosses also increased, whereas those containing sedges decreased. In a large-scale water table and temperature manipulation experiment in northern Minnesota (Weltzin et al., 2003), ericaceous shrub cover increased 50% four years after peat in a bog mesocosm was warmed from 1.6 to 4.1 °C; again, sedge cover decreased. Furthermore, above- (30%) and belowground (100%) biomass increased when water tables were low (Weltzin et al., 2003). Higher water table levels in a fen mesocosm resulted in increased sedge cover, whereas lower water table levels produced shrub results similar to those in the bog.

Remote sensing studies also indicated increased tree and shrub vegetation under warmer and drier conditions, or sedges, hollow-mosses, and tall shrubs under wetter conditions within the past 50 years (Cornelissen et al., 2001; Tape et al., 2006). Therefore, establishing plant function types (PFTs) should facilitate scaling and modeling C budget responses to landscape-level drying or wetting. Combining remote sensing of peatland PFTs and peat accumulation rates and GHG exchange measured at research sites and monitoring networks allows the use of calibrated and validated models to estimate C balances (St-Hilaire et al., 2010). The PFTs used depend on scale. For large landscapes, vascular (e.g., treed bogs and fens) and non-vascular (e.g., open bogs and fens) plant groups have been used (Dorrepaal, 2007). Results presented in this paper suggest dry, wet, and pond classifications may also be useful

TABLE 4

Carbon dioxide (CO_2) sequestration and methane (CH_4) emissions for northern peatlands.

Peatland type	Location	CO_2 sequestration (g C m^{-2} d^{-1})	CH_4 emission (mg C m^{-2} d^{-1})	References
Bogs	Ontario	–1.7 to 0.5	3.3 to 52	Roulet et al. (1992); Bubier et al. (1993); Klinger et al. (1994); Moore et al. (1994b); Neumann et al. (1994); Lafleur et al (2003); Strilesky and Humphreys (2012)
	Québec	–0.2	6.1 to 74	Moore et al. (1990); Roehm and Roulet (2003); Pelletier et al. (2007); McEnroe et al. (2009)
	Northwest Territories	0.4 to 1.3		Chasmer et al. (2012)
	Finland	–0.6 to 0.9		Alm et al. (1999)
	Sweden	–1.2 to 0.2	28	Sagerfors et al. (2008); Bäkstrand et al. (2010)
	Siberia	–0.7 to –0.1		Arneth et al. (2002); Friborg et al. (2003)
Fens	Ontario	–1.0 to –0.4	0.1 to 60	Roulet et al. (1992, 1994); Bubier et al. (1993); Klinger et al. (1994); Moore et al. (1994b); Bubier (1995)
	Québec	–0.6 to 0.2	0.6 to 65	Moore et al. (1990); Bubier (1995); Strack et al. (2006); Pelletier et al. (2007)
	Manitoba	–0.9 to 0.7	5.4 to 145	Bubier (1995); Bubier et al. (1995, 1999); Burton et al. (1996); Schreader et al. (1998); Griffis et al. (2000); Joiner et al. (1999)
	Saskatchewan	–4.3 to –0.7		Suyker et al. (1997); Griffis et al. (2000); Sonnentag et al. (2010)
	Alberta	–1.3 to –0.3		Glenn et al. (2006); Syed et al. (2006); Adkinson et al. (2011)
	Northwest Territories	0.4 to 1.6		Chasmer et al. (2012)
	Finland	–0.8 to –0.1	80 to 204	Heikkinen et al. (2002); Nykanen et al. (2003); Aurela et al. (2007); Leppälä et al. (2011); Maanavilja et al. (2011)
	Sweden	–0.7 to –0.4	87 to 137	Johansson et al. (2006b); Bäkstrand et al. (2010); Jackowicz-Korczy ski (2010)
Palsa/Peat plateaus	Saskatchewan		<0.1	Turetsky et al. (2002)
	Northwest Territories	0.4 to 0.5		Chasmer et al. (2012)
	Alaska	–0.8 to 1.0	–1.6 to 24	Wickland et al. (2006)
	Finland		1.0 to 2.1	Nykanen et al. (2003)
	Sweden	–0.7 to 0.7	0.5	Bäkstrand et al. (2010)
	Siberia		–0.1 to 15	Christensen et al. (1995); Flessa et al. (2008)
Thermokarst	Manitoba	–0.54		Bubier et al. (1999)
	Northwest Territories		29 to 160	Libilik et al. (1997)
	Alaska	–2.1 to –0.7	84 to 298	Wickland et al. (2006); Meyers-Smith et al. (2007)
	Finland	–1.1	80 to 234	Nykanen et al. (2003)
	Sweden	–0.5 to 0.3	87 to 137	Bäkstrand et al. (2010); Jackowicz-Korczy ski (2010); Christensen et al. (2012)
	Siberia		0 to 106	Christensen et al. (1995)
Ponds	Ontario	3.0 to 3.1	30 to 163	Roulet et al. (1992, 1994); Hamilton et al. (1994)
	Québec	0.22	32 to 190	Moore et al. (1990); Roulet et al. (1994); McEnroe et al. (2009)
	Manitoba	1.1 to 4.5		Macrae et al. (2004)
	Saskatchewan		88 to 120	Rask et al. (2002)
	Nunavuk	0.1 to 0.8		Laurion et al. (2010)
	Siberia		16 to 192	Flessa et al. (2008); Desyatkin et al. (2009)

TABLE 5

Basal peat ages in peatlands in permafrost zones of Canada.

Peatland type/province	Permafrost feature	Peat depth (cm)	Age (^{14}C yr BP)	References
Bogs				
Ontario	Non-permafrost	150	7280 + 70	Kettles et al. (2000)
Ontario	Non-permafrost	445	5920 + 90	Glaser et al. (2004)
Ontario	Non-permafrost	266	4810 + 70	Glaser et al. (2004)
Ontario	Non-permafrost	220	5220 + 80	Glaser et al. (2004)
Ontario	Non-permafrost	127	3700 + 70	Glaser et al. (2004)
Ontario	Non-permafrost	236	3960 + 60	Glaser et al. (2004)
Ontario	Non-permafrost	223	3730 + 50	Glaser et al. (2004)
Ontario	Non-permafrost	264	4000 + 80	Kettles et al. (2000)
Ontario	Non-permafrost	209	4500 + 70	Glaser et al. (2004)
Québec	Peat plateau	102 to 260	4400 + 90 to 6480 + 120	Bhiry et al. (2007)
Québec	Palsa	—	280 + 60 to 3410 + 100	Bhiry et al. (2007)
Manitoba	Palsa	169	5970 + 90	Kuhry (2008)
Manitoba	Peat plateau	166	5810 + 90	Kuhry (2008)
Saskatchewan	Peat plateau	186 to 197	5065 ± 70 to 5780 ± 90	Sannel and Kuhry (2008)
Fens				
Ontario	Non-permafrost	149 to 234	5200 + 60 to 5370 + 80	Glaser et al. (2004)
Ontario	Non-permafrost	104 to 127	3910 + 80 to 4550 + 70	Glaser et al. (2004)
Ontario	Non-permafrost	98 to 109	3840 + 70 to 4010 + 80	Glaser et al. (2004)
Québec	Palsa	225 to 252	4880 + 100 to 5100 + 100	Arlen-Pouliot and Bhiry (2005)
Québec	Filled thermokarst pond	37	350 + 60	Arlen-Pouliot and Bhiry (2005)
Québec	Palsa	200	4070 + 70	Bhiry and Robert (2006)

for scaling and modeling current and future C accumulation and GHG changes of HBL landscapes. Peatlands may also be more narrowly defined using vegetation classes, such as (1) mosses, sedges, shrubs, and trees or (2) genera- and species-level PFTs for smaller areas (Dorrepaal, 2007). For example, sedges, tall shrubs, and *Sphagnum* mosses have been used successfully to scale CH_4 emissions at site (Johansson et al., 2006b) and regional (Dorrepaal, 2007) scales.

PERMAFROST PATTERNS

Permafrost distribution based on mean annual air temperature (MAAT) is generally well mapped across Canada (Fig. 4, part a). However, its association with mean annual ground temperature (MAGT) is complex, and only sparse data exist that describe MAAT and MAGT associations (Zhang et al., 2008). Furthermore, aggrading and degrading permafrost features (Table 1) are present in various stages of maturation within a climatic zone (Payette et al., 2004). These patterns result from local variations in vegetation, topography, snow cover, and soil conditions (i.e., peat versus mineral soil) that largely control MAGT and subsequent permafrost presence (Johansson et al., 2006a). Variations in environmental conditions are important when MAAT is close to 0 °C, as in the HBL (Fig. 4, part b). This temperature range may be associated with the fastest peat C accumulation rates in northern peatlands; rates decrease northward because of low NPP and southward because of faster peat decomposition (Beilman et al., 2009).

In the northern hemisphere, during the 21st century permafrost is expected to disappear from 16 to 90% of the land base, and active layer (surface peat layer experiencing seasonal thaw that is the location of rapid biogeochemical cycling and water fluxes) thickness (ALT) may increase by 30 to 300 cm (Table 6). Mean (± standard deviation) percent permafrost loss calculated from data in Table 6 was 53% ± 25% per 100 years, with a CV of 0.47 and a statistical power of 0.65; increasing the sample size from the 37 compiled in this study to 43 would increase the power to 0.70. Furthermore, natural variation in permafrost loss ranged from 33 to 73% per 100 years. Mean increase in ALT was 105 ± 92 cm per 100 years, with a CV of 0.88, a power of 0.24, and natural variation between 50 and 100 cm; sample size would need to be 143 to attain a power of 0.70. However, models project that the HBL has already surpassed its temperature threshold for maintaining permafrost (Koven et al., 2011), or will do so by 2020 (Zhang et al., 2008). Permafrost thawing is, therefore, an important concern because MAAT in the HBL is projected to rise between 2.5 and 8.0 °C, with the most rapid warming occurring in the most permafrost-dense area in the region (Gagnon and Gough, 2005; McKenney et al., 2010). Yet, variable ALT may impede documenting current and future implications to landscape C balances.

Spatial configurations of permafrost features depend on a variety of internal (e.g., peat accumulation and acidification) and external (e.g., climate and fire behavior) factors (Johansson et al., 2006a). In the northernmost regions, palsa/peat plateaus and thermokarst features dominate landscape patterns, along with non-permafrost bogs and fens (Payette et al., 2004; Kuhry, 2008). In contrast, mostly non-permafrost bog and fen complexes occur in southern regions of the HBL, with forested bogs that are susceptible to burning (Turetsky et al., 2004), covering much of the area (Klinger and Short, 1996). Spatial distribution of permafrost extent implies that during the 21st century southern and northern portions of the HBL may respond differently to climate change. For example, the boreal zone is projected to warm and dry, increasing the likelihood of faster peat decomposition and more frequent and severe burning of forested palsas, peat plateaus, and bogs. Subsequently, elevated CO_2 losses may weaken the C sink strength of southern HBL peatlands (e.g., Turetsky et al., 2004). In contrast, permafrost thawing in the northern portion of the HBL may maintain current peat moisture levels, or increase the amount of saturated soils on the landscape (Zhang et al., 2008), with warmer temperatures enhancing peatland CO_2 sequestration and CH_4 emission (Johansson et al., 2006b). However, thermokarst pond expansion may partially offset peat CO_2 sequestration and further enhance CH_4 emission to the atmosphere (Macrae et al., 2004; Laurion et al., 2010).

At coarse landscape scales, permafrost thaw has been modeled for the Hudson Bay region using both equilibrium and disequilibrium assumptions between MAAT and MAGT, although small sample sizes preclude statistical analyses. Gagnon and Gough (2005) assumed MAAT and MAGT equilibrium would result in 35 to 67% losses in permafrost by 2100 in the Hudson Bay region. In contrast, Zhang et al. (2008) assumed non-equilibrium between MAAT and MAGT and reported only minor losses (about 16%) in permafrost occurrence in the HBL during the 21st century, primarily in the southern fringe of the discontinuous permafrost zone (DPZ). Gagnon and Gough's (2005) results were generally within the range of natural variability defined in this study, although some climate models produced results greater than that for the 75th percentile. In contrast, results from Zhang et al. (2008) were below the 25th percentile. That study simulated permafrost loss using the process-based Northern Ecosystem Soil Temperature (NEST; Zhang et al., 2003) model, which projects less permafrost loss by accounting for plant structure, snow depth, and soil organic layer thickness effects on ground thermal dynamics.

Aerial photographic and remote sensing studies provide evidence that permafrost thawing is changing HBL landscapes. For example, at Boniface River, Québec, permafrost decreased from approximately 80% of the land area in 1957 to 20% in 2003 (Payette et al., 2004). This decrease was accompanied by increases in (1) fen area from less than 5% in 1957 to approximately 50% in 2003 and (2) thermokarst pond area from 20 to 40% during the same time frame. Similar changes have occurred at the Stordalen mire complex in northern Sweden, also weakening landscape C and GWP sink strengths (Malmer et al., 2005; Johansson et al., 2006b). Those results were consistent with this current synthesis, where wet features and ponds emitted approximately 25 times more CH_4 than dry features. Combined with potentially more CO_2 release from the landscape, pond (and wet feature) expansion will likely play a vital role in C balances in northern HBL landscapes.

Carbon and GWP sink strengths over decadal to century time frames following thermokarst formation are relatively unknown. Some evidence supports less CH_4 emission from *Sphagnum*-dominated collapse scars as they age (Prater et al., 2007), but landscape CH_4 emissions may remain similar as ponds transform to sedge-dominated fens because of the latter's rapid CH_4 emission (Webster et al., 2013). Moreover, Frolking et al. (2011) suggested that by 2100, 30 to 100 Mt CO_2-C yr^{-1} may be sequestered and up to 35 Mt CH_4-C yr^{-1} released from permafrost regions (beyond current values), which was consistent with coupled carbon-climate-change biospheric terrestrial model projections (Koven et al., 2011; Schaefer et al., 2011).

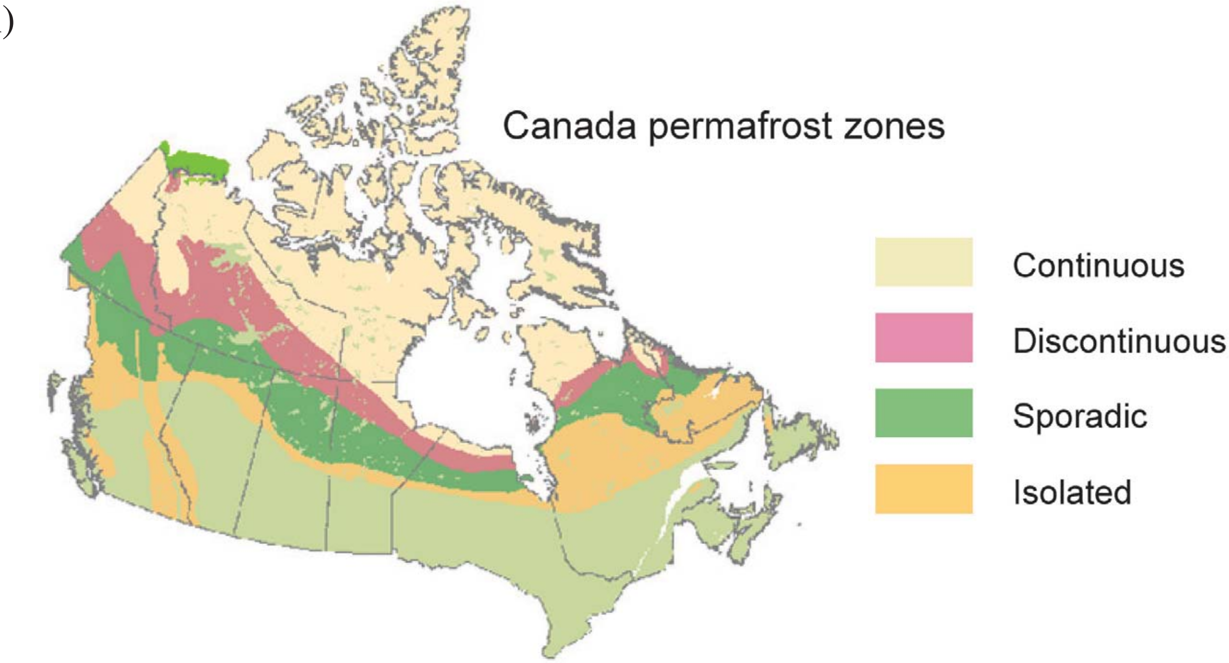
PEATLAND EVAPOTRANSPIRATION

Evapotranspiration is reported as actual (ET_a) and potential (ET_p), and northern peatlands frequently evaporate water at rates lower than their potential (Rouse, 2000); the rate of ET is strongly influenced by the surface energy balance (i.e., latent heat flux; Q_e). Evapotranspiration affects peatland C exchange at landscape scales primarily through its contribution to water balances. For example, when water was abundant, ET and gross primary production in bogs were positively correlated (Sonnentag et al., 2010), but under water-stressed conditions, photosynthetic depressions were linked to elevated ET in a variety of peatland plants (Euskirchen et al., 2012). Furthermore, elevated ET relative to water inputs may contribute to long-term peatland drying, leading to higher frequency and severity of fires, potentially contributing further to weakening peatland C sink strengths (Turetsky et al., 2004). Because of its relationship with C exchange, accurate estimates of ET are required to construct water balances and assess their potential responses to climate change. However, the amount of water withdrawn through ET varies between 20 and 80% of the water balance, depending on terrain unit (e.g., site type), permafrost conditions, and temperature and precipitation regimes (Rouse, 1998).

Since large areas of the northern HBL occur as palsa-bog-fen-pool/pond complexes and small lakes (Martini, 2006), ET estimates for dominant features are needed before a scaled value can be derived for a specific peatland complex (e.g., Petrone et al., 2000). Although ET rates vary (Table 7), tundra ponds, small lakes, and wet sites including fens, marshes, and tundra wetlands displayed the highest ET rates. Dry features, including spruce forests, non-forested bogs, lichen/heath, heath/shrub, tussock tundra, and peat plateaus/polygonal tundra had the lowest ET rates. Analysis of data compiled from 28 studies (Tables 7 and 8) indicated ET was higher ($p = 0.002$) in wet and pond than in dry features (power = 0.61). Evapotranspiration rates in wet features in the HBL (2.8 to 3.5 mm d^{-1}) were within the range of natural variability (2.5 to 3.5 mm d^{-1}) calculated from the overall data set. Wet features (0.71 ± 0.15) also partitioned more energy ($p = 0.003$) to ET than dry features (0.49 ± 0.08), as defined by the Q_e-to-net radiation (Q^*) ratio (Table 8). The statistical power was 0.58, but the small *p* value suggests Q_e/Q^* should be further pursued as an indicator of wet and dry feature energy partitioning and water budgets in developing methods to assess climate change vulnerability and adaptation.

Looking at estimates of how expected changes in climate might influence annual ET patterns in a sedge fen at Churchill, Manitoba, Rouse (1998) reported that a two times CO_2 climate warming scenario with a MAAT increase of 4 °C and no precipitation change increased summer water deficit, triggered mainly by greater ET during

A)



B)

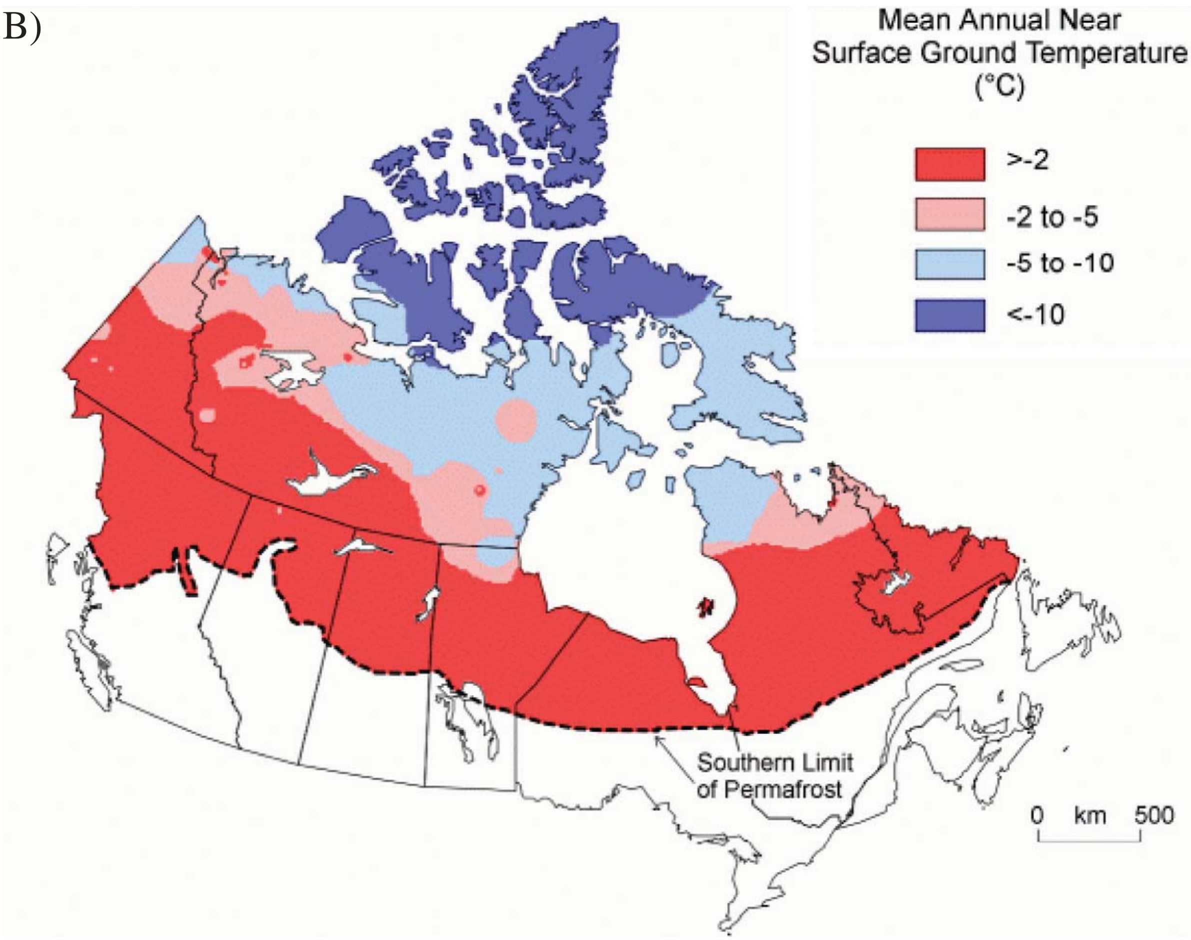


FIGURE 4. (a) Permafrost distribution across Canada illustrating permafrost zones, which are defined by the proportion of area underlain by permafrost. (b) Mean annual near-surface temperature distribution across Canada. Image (a) adapted from Department of Natural Resources Canada, <http://geogratis.gc.ca/api/en/nrcan-rncan/ess-sst/092b663d-198b-5c8d-9665-fa3f5970a14f.html>. Image (b) from <http://www.ccin.ca/home/ccw/permafrost/future>.

TABLE 6

Predicted change in permafrost area and active layer depth during the 21st century in Canada.

Domain	Climate change scenario	Model**	Percent permafrost loss	Active layer depth increase (cm)	References
Alaska	A1B [A*]	1	7	162	Marchenko et al. (2008)
Alaska	AIB [B]	2	22 to 61	69 to 105	Schaefer et al. (2011)
Canada	AIB [B]	2	22 to 63	55 to 90	Schaefer et al. (2011)
Canada	B2, A2 [C]	3	16 to 20	30 to 70	Zhang et al. (2008)
Canada	A1B [C]	3	21 to 24	30 to 80	Zhang et al. (2008)
Northern Hemisphere	A1B [D]	4	27	—	Euskirchen et al. (2006)
Northern Hemisphere	A2, B2 [E]	5	40 to 57	50 to 300	Saito et al. (2007)
Northern Hemisphere	B1, A1B, A2 [F]	6	60 to 90	50 to 300	Lawrence and Slater (2005)
Northern Hemisphere	AIB [F]	6	73 to 88	—	Lawrence and Slater (2010)
Northern Hemisphere	AIB [F]	6	80 to 85	50 to 300	Lawrence et al. (2008)
Northern Hemisphere	AIB [B]	2	20 to 39	56 to 92	Schaefer et al. (2011)

* Climate models [A] HADCM2; [B] CCSM3, HadCM3, and MIROC3.2; [C] CGCM, CSIROM, ECHAM, GFDL, HadCM, and NCAR; [D] NCAR; [E] MIROC3.2; [F] CCSM3.
** Permafrost models: [1] GIPL 2.0 - Spatially Distributed Model of Permafrost Dynamics in Alaska (Sazonova and Romanovsky, 2003); [2] Simple Biosphere/Carnagie-Ames-Stanford Approach Mode (Schaefer et al., 2008); [3] **Northern Ecosystem Soil Temperature (Zhang et al., 2008)**; [4] Terrestrial Ecosystem Model (Raich et al., 1991); [5] Minimal Advanced Treatments of Surface Interaction and Runoff (Takata et al., 2003); [6] Community Land Model (Dickenson et al., 2006).

May. The water deficit could be counterbalanced by a 23% increase in summer rainfall (Rouse, 1998). Similar results were recorded in the James Bay Lowland (Lafleur et al., 1993). Roulet et al. (1992) assumed a 3.5 °C increase in temperature and 1 mm day^{-1} increase in precipitation for several northern fens and reported an increase in the soil moisture deficit in excess of 100%. Other studies across Canadian peatlands have since shown that thresholds of about a 25% increase in water supplied by precipitation or permafrost thawing would be required to maintain soil moisture levels under warming-induced enhancement of ET (Quinton et al., 2011).

Effects of climate change on ET are more difficult to predict for dry than for wet features. This has been attributed to the phenology of vascular plants, predominantly leaf area index (LAI), and its control on energy partitioning in sparsely vegetated bogs (LAI < 1.0) (Admiral et al., 2006). Leaf area index is minimal in spring, when peat is generally saturated (Lafleur and Humphreys, 2008). This corresponds to the time of highest solar energy reaching bog surfaces and high Q_e/Q^* values due to *Sphagnum* mosses, which account for 50 to 70% of the total annual ET from bogs (Kellner, 2001). Increasing LAI decreases the amount of radiation reaching the surface (Admiral et al., 2006). At this time, ET switches from moss- to vascular plant–dominated, and subsequently dries the peat (Kellner, 2001).

Understanding moss and vascular plant interactions with projected earlier snowmelt is critical to comprehend future ET trends of dry features because of the earlier onset of ET by mosses. However, few studies have documented the competition between mosses and vascular plants following snowmelt. Therefore, it is uncertain if earlier snowmelt will only shift moss ET to earlier in the year or if soil warming during spring will enhance nutrient mineralization and subsequent vascular plant biomass and LAI, prolonging soil moisture deficits. Most model results project that during the 21st century vascular plant biomass and LAI will increase. Such increases have been positively correlated with aboveground biomass and soil C accretion (Frolking et al., 2011). However, LAI increases are also positively correlated with fire frequency and area burned because of dry surface peat (Turetsky et al., 2004). As such, we view (1) moss and vascular plant competition following earlier snowmelt and (2) extended growing season, energy partitioning, and peatland drying on C accumulation rates and GHG exchange rates as key research areas, where data are vital to understand and improve future bog C budget projections.

Permafrost is a strong heat sink and generally has a larger ground heat flux (Q_g) than non-permafrost peatlands, and changes in temperature and moisture may alter permafrost thaw rates (Quinton et al., 2011). However, few data exist to confidently describe and model the energy balance in permafrost features. Boike et al. (2008) reported that during a wet year in polygonal tundra of the continuous permafrost zone (CPZ), energy partitioning remained typical of wet features (Table 8). However, when soil was dry, a relatively even partitioning of energy occurred (Table 8). In addition, during dry years, energy partitioning was not affected by differences in air temperature. However, during wet years Q_e increased with increasing temperatures (Boike et al., 2008). Similar patterns were shown for wet and dry tussock tundra underlain by permafrost (Harazono et al., 1998) and mesic and heath/shrub tundra (Lafleur and Humphreys, 2008).

Climate Change Vulnerability and Adaptation Assessment Framework to Assess Peatland Carbon Indicators

The large contribution of the HBL to provincial and national C budgets implies that understanding peatland C accumulation and

TABLE 7

Evapotranspiration (ET) rates for common site types in high boreal and subarctic drainage basins north of 50° latitude.

Site type	Study period	Location	ET (mm d^{-1})	References
Tundra lake	June–August 1985	Manitoba	6.1	Bello and Smith (1990)
	June–August 1991	Manitoba	4.8	Boudreau and Rouse (1995)
	June–November 1997 to 1999	Northwest Territories	2.3	Rouse et al. (2003)
Open water	May–September 1977	Québec	2.8	Boudreau and Rouse (1995)
Tundra pond	June–September 2000	Northwest Territories	2.5	Spence et al. (2003)
Flat fen	June–August 1983	Northwest Territories	4.5	Roulet and Woo (1986)
Sedge fen	June–September 1985	Ontario	3.5	Rouse et al. (1987)
	June–August 1991	Manitoba	3.5	Boudreau and Rouse (1995)
	May–August 1990	Québec	2.8	Moore et al. (1994a)
	June–September 1987 to 1994	Manitoba	2.7	Rouse (1998)
	June–August 1989	Manitoba	2.5	Lafleur et al. (1993)
	June–August 1996 to 1997	Manitoba	2.5	Petrone et al. (2000)
Wooded fen	July–August 2004	Alberta	2.0	Humphreys et al. (2006)
	July–August 2004	Saskatchewan	2.0	Humphreys et al. (2006)
Open poor fen	July–August 2004	Alberta	2.5	Humphreys et al. (2006)
Pine fen	April–October 2006	Finland	2.2	Wu et al. (2010)
	April–October 2007	Finland	1.6	Wu et al. (2010)
Marsh	May–August 1984	Ontario	2.8	Price and Woo (1988)
Alder-willow covered ridge	May–August 1985	Ontario	3.0	Woo and diCenzo (1989)
Willow-birch wetland	June–August 1991	Manitoba	3.2	Boudreau and Rouse (1995)
	May–September 1977	Québec	2.7	Boudreau and Rouse (1995)
Open spruce, tamarack	June–September 1989	Manitoba	2.1	Lafleur (1992)
Shrub bog	June–September 1998 to 2000	Siberia	2.1	Kurbatova et al. (2002)
	June–September 1998 to 2000	Russia	1.9	Kurbatova et al. (2002)
Open bog	June–September 1996 to 1997	Sweden	2.0	Kellner (2001)
String bog	June–July 1989	Labrador	2.5	Price et al. (1991)
Blanket bog	May–July 1989	Newfoundland	1.7	Price (1991)
Spruce forest	June–August 1991	Manitoba	3.3	Boudreau and Rouse (1995)
	June–August 1989	Manitoba	2.2	Lafleur (1992)
Lichen-heath	June–August 1991	Manitoba	2.3	Boudreau and Rouse (1995)
	May–September 1977	Québec	1.7	Boudreau and Rouse (1995)
Mixed, heath and shrub tundra	May–August 2004 to 2006	Northwest Territories	1.4	Lafleur and Humphreys (2008)
Tussock tundra	July–October 2003 to 2004	Siberia	1.5	Corradi et al. (2005)
Coastal tundra, wet	June–August 2004	Alaska	1.3	Vourlitis and Oechel (1997)
Polygonal tundra	June–August 1999, 2003	Siberia	1.6	Boike et al. (2008)

GWP gains and losses is vital to sustaining C storage and sequestration in the ever-changing HBL landscape. Although scant, evidence is convincing that PFTs, permafrost, and ET are changing in the HBL. However, rates and direction of peatland changes are variable and depend on whether peat becomes drier or wetter as it warms (Fig. 5). Some changes, such as thermokarst formation (Arlen-Pouliot and Bhiry, 2005) and transformations within fen types as they convert to bogs (McLaughlin and Webster, 2010), occur rapidly (i.e., decades); those two processes have opposite effects on peatland C balances (Fig. 5). However, successional trajectories in the HBL are poorly documented (although recent summaries of data collected from the 1970s and early 1980s are available [Riley, 2011]). Also, the remoteness of the region hinders large-scale sampling endeavors due to high costs.

TABLE 8

Energy balance partitioning, Bowen ratio [latent heat flux (Q_e)/net radiation (Q*); sensible (Q_h)/Q*; ground heat flux (Q_g)/Q*; lake heat storage Q_w/Q*], Bowen ratio energy balance (β), and Priestly-Taylor alpha (α) for different site types in peatlands and permafrost features north of 50° latitude.

Site type	Location	Years	Q_e/Q*	Q_h/Q*	Q_g/Q* (Q_w/Q* for lakes)	References
Deep lake	Northwest Territories	1997–1998	0.34	–0.01	0.68	Eaton et al. (2001)
Shallow lake	Northwest Territories	1991–1995	0.75	0.23	0.03	Eaton et al. (2001)
	Ontario	1972	0.74	0.26	0.00	Stewart and Rouse (1976)
	Manitoba	1985	1.05	–0.14	0.04	Bello and Smith (1990)
Sedge wetland	Manitoba	1990–1995	0.64	0.28	0.11	Eaton et al. (2001)
	Québec	1990	0.63	0.25	0.10	Moore et al. (1994a)
	Ontario	1985	0.69	0.16	0.15	Rouse et al. (1987)
Willow-birch wetland	Manitoba	1991	0.64	0.21	0.14	Boudreau and Rouse (1995)
Wetland tundra	Northwest Territories	1996–1997	0.64	0.28	0.06	Eaton et al. (2001)
Open bog	Ontario	1990	0.46	0.34	0.10	den Hartog et al. (1994)
	Sweden	1996–1997	0.58	0.17	0.08	Kellner (2001)
	Siberia	1996	0.53	0.31	0.16	Eugster et al. (2000
Conifer forest	Northwest Territories	1996	0.46	0.45	0.09	Eaton et al. (2001)
	Manitoba	1993–1994	0.54	0.39	0.07	Eaton et al. (2001)
Shrub tundra	Northwest Territories	1996–1997	0.52	0.28	0.21	Eaton et al. (2001)
	Manitoba	1990–1991	0.56	0.30	0.14	Eaton et al. (2001)
Lichen-heath tundra	Manitoba	1991, 1996	0.50	0.38	0.14	Eaton et al. (2001)
	Ontario		0.57	0.41	0.04	Rouse et al. (1987)
Tussock tundra – wet	Alaska	1995	0.49	0.40	0.17	Harazono et al. (1998)
Tussock tundra – dry	Alaska	1995	0.37	0.48	0.15	Harazono et al. (1998)
Polygonal tundra – dry	Siberia	1999	0.31	0.40	0.29	Boike et al. (2008)
Polygonal tundra – wet	Siberia	2003	0.61	0.22	0.17	Boike et al. (2008)

Remote sensing using various platforms and sensors and modeling techniques have been employed to assess PFTs, permafrost thawing, ET, and C indicators at the landscape scale (Dorrepaal, 2007). Furthermore, the ecosystem processes presented here have consistently been correlated with various components of C budgets and successfully scaled across landscapes (e.g., Johansson et al., 2006b; Beilman et al., 2009). However, few researchers have attempted to develop correlations between climate change and peatland C indicators for the HBL. Significant efforts are required to sample peat and calibrate satellite and aerial photographs for PFTs, permafrost coverage, and ET to calibrate permafrost and peatland C models.

In terms of C indicators assessed in this study, low variation and high statistical power in LORCA comparisons point to it being a promising C indicator. Coupled with relatively small variability in peat depth and ^{14}C basal peat age, long-term peatland processes may be effective metrics in climate change assessments. However, LORCA in the HBL peatlands was on the low end of the natural variation calculated across boreal and subarctic peatlands. Furthermore, LORCA values were from a limited number of locations and peatland types in the HBL; thus, additional measurements in the HBL are warranted to further evaluate LORCA variability. Although statistical power was relatively low for RERCA and CH_4 emissions, rates were much higher in wet (and CH_4 for ponds) than dry features, suggesting their potential usefulness in evaluating landscape-scale peatland C balances in response to decade-to-century permafrost and PFT changes. However, much of the data from the HBL analyzed in this synthesis pre-dates the mid-1990s and may not be applicable across the region or to its current landscape structure. In addition, high variation and low statistical power in CH_4 emission (and CO_2 sequestration) hinders construction of viable GHG balances in HBL peatlands. Assigning probabilities to landscape GHG changes may be possible but requires further analyses.

Current and future peatland C accumulation and GHG emissions in northern peatlands may also be underestimated due to (1) poorly defined climate change rates and subsequent peatland adaptation and (2) whether new climate states will be within the range of Holocene variation (Frolking et al., 2011). As indicated by Frolking et al. (2011), the only approach available to address climate change and C budgets is rudimentary process-based modeling, the results of which are inconsistent with one another and field data. For example, warming may increase peatland C losses by 40 to 85% in coming centuries (Ise et al., 2008), or peatland C stocks may stabilize in response to drying (Frolking et al., 2010). However, none of the available permafrost (e.g., NEST) and C (e.g., McGill Wetland Model [St-Hilaire et al., 2010], Wetland-DNDC [Zhang et al., 2002]) models allow plant communities or spatial patterns in permafrost and peatland features to vary during simulations, significantly limiting their applicability to functional landscapes.

Information about the quality (decomposability) of peat exposed to aerated or flooded conditions with warming temperatures is also lacking. Deeply buried peat is primarily "recalcitrant" material due to prior decomposition in aerobic surface peat layers (Laiho, 2006). However, the effects of warmer and drier (or wetter) conditions on decomposition rates of "recalcitrant" peat are conflicting, with some studies reporting faster decomposition (Dorrepaal, 2007) and others showing no rate changes (Davidson and Janssens, 2006) under warmer temperatures. Furthermore, peat enclosed in permafrost tends to be less decomposable than that of enclosed mineral soil organic matter (Turetsky et al., 2007). However, data on short- and long-term decomposition of C released through ice melting in permafrost regions are scarce. Finally, dissolved organic C (DOC) may be an important regulator of microbial decomposition in peatlands (Laiho, 2006). The quality of DOC to support decomposition is known to differ among peatland types and between dry and wet conditions (Webster and McLaughlin, 2010). How those changes will be manifested in future organic matter decomposition in peatlands, however, is virtually unknown.

Other uncertainties and challenges also impede the use of climate and C indicators for assessing climate change vulnerability and adaptation. For climate change, lack of weather and permafrost monitoring networks in the HBL significantly limits understanding of past and current temperature, precipitation, and permafrost regimes, contributing to highly variable projections in future conditions. For example, mean monthly precipitation projections differed by more than 200% (Gagnon and Gough, 2005; McKenney et al., 2010). Precipitation amounts are important to landscape permafrost coverage within the same temperature zone (Rouse, 2000; Johansson et al., 2006a). Interactions between air temperature and precipitation with peat temperature and moisture dynamics are also some of the most poorly calibrated coefficients used to project permafrost and peatland C dynamics (Zhang et al., 2008; St-Hilaire et al., 2010).

Few valid estimates of permafrost amount and its thaw rate exist for the HBL. Payette et al. (2004) reported that between 1957 and 2003 percent annual permafrost loss increased from 2.5 to 5.3%, which is much higher than the natural range of variation (e.g., 0.33% yr^{-1} to 0.73% yr^{-1} assuming a 100-year time frame over the 21st century) calculated from coarse-scale climate and process-based models. Current models, such as NEST, may constrain variability by accounting for plant structure, snow depth, and soil organic layer thickness effects on energy and water transfer in soil-vegetation-atmosphere systems, and subsequent ground thermal dynamics. However, all models fail to address changes in landscape pattern, which is a critical modeling gap. For example, assuming a 50% increase in fen and 20% increase in pond areas in the northern HBL (Payette et al., 2004) during the 21st century, landscape-level CO_2 and CH_4 emissions to the atmosphere

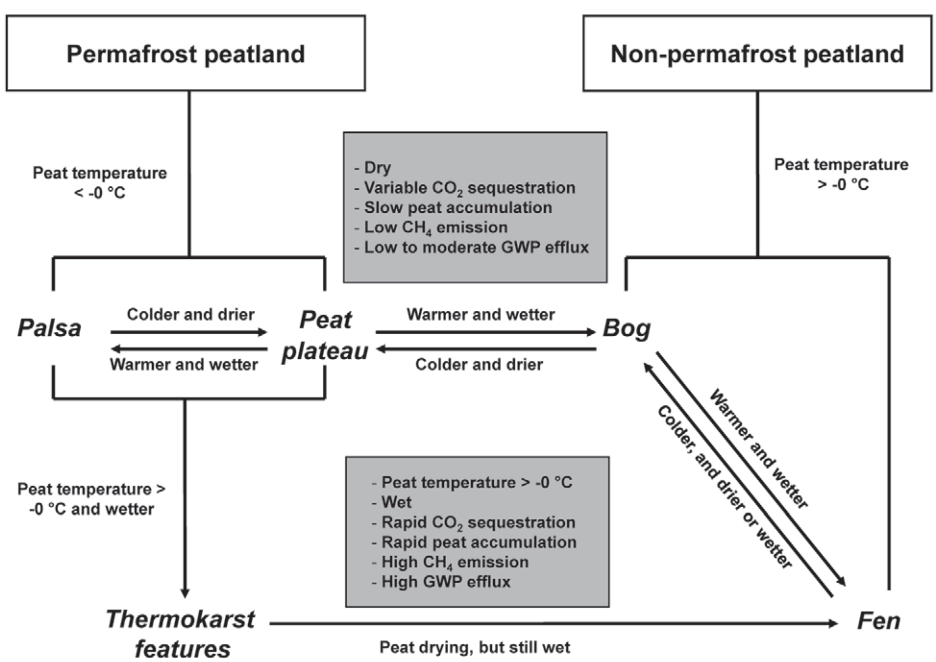


FIGURE 5. Conceptual model of permafrost and non-permafrost peatland succession (arrows) in response to changing peat conditions (gray boxes) in the Hudson Bay Lowlands of Ontario. C = carbon, CO_2 = carbon dioxide, CH_4 = methane, GWP = global warming potential.

may increase significantly and would not be detectable in coarse-scale projections. Such changes have been shown to significantly weaken C and GWP sinks (Malmer et al., 2005; Johansson et al., 2006b).

In conclusion, this synthesis indicates that changes in C accumulation rates and CH_4 emissions in response to altered ecosystem processes, such as succession, permafrost loss, and ET may be useful to assess peatland climate change vulnerability and adaptation for large HBL landscapes. Furthermore, C accumulation and CH_4 emission may be evaluated through remote sensing identification of changes in wet, dry, and pond landscape features. However, an improved understanding of the interactions between GHG fluxes and permafrost dynamics (loss and ALT) is needed before such assessments are possible for small landscapes. Therefore, enhanced research and monitoring studies are needed to estimate current and project future peatland C storage and sequestration with reasonable certainty to support land-use planning in the HBL.

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