UNIVERSITÉ DU QUÉBEC À MONTRÉAL

DYNAMIQUE ÉCOHYDROLOGIQUE ET SÉQUESTRATION DU CARBONE AU COURS DE L'HOLOCÈNE DANS LES TOURBIÈRES OMBROTROPHES MARITIMES DE L'ESTUAIRE ET DU GOLFE DU SAINT-LAURENT

THÈSE

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GABRIEL MAGNAN

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AVANT-PROPOS

La thèse est composée de trois articles scientifiques rédigés en anglais qui ont été soumis à des revues scientifiques internationales avec comité de lecture. Les autres sections de la thèse sont rédigées en français selon les exigences de l'Université du Québec à Montréal. J'ai agi à titre de premier auteur pour les trois articles (chapitres), ce qui signifie que j'ai été le principal responsable de la collecte des données, des analyses en laboratoire, de l'interprétation des données et de la rédaction. Ma directrice de thèse Michelle Garneau est deuxième auteure des trois articles composant la thèse. Mon co-directeur Serge Payette (Université Laval) est co-auteur du Chapitre 1 portant sur le développement à long terme des tourbières. Afin de faciliter la lecture individuelle de chaque chapitre, les articles ont été conservés dans leur intégralité ce qui entraine certaines répétitions notamment dans les sections décrivant les régions et sites d'étude et la méthodologie.

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AD	*Anno domini*, années après Jésus-Christ
AMS	Accelerator mass spectrometry
m. a.s.l.	Meters above sea level
BC	Baie-Comeau
BP	Before present (avant l'actuel, i.e. AD 1950)
BSW	Bog surface wetness
C	Carbone
^{14}C	Radiocarbone
Cal BP	calibrated years before present (AD 1950)
CAR	Carbon accumulation rates
CHAR	Charcoal accumulation rates
CO_2	Dioxide de carbone
CH_4	Méthane
DCA	Detrended correspondence analyses
DGPS	Differential Global Positioning System
étal. BP	Étalonnées avant AD 1950
Gt	Gigatonnes
HSP	Havre-Saint-Pierre
ka	kiloannum (milliers d'années BP)
LIA	Little Ice Age
LOI	Loss-on-ignition

LORCA	Long-term average apparent rate of carbon accumulation
LOWESS	Locally weighted regression smoothing
MAP	Mean Annual Precipitation
MAT	Mean Annual Temperature
MCA	Medieval Climate Anomaly
NPP	Net Primary Production
NLWIS	National Land and Water Information Service
OM	Organic Matter
RMSEP	Root-Mean-Square Error of Prediction
SE	Standard Error
SD	Standard deviation
sp.	Species (singulier)
spp.	Species (pluriel)
WTD	Water Table Depth

RÉSUMÉ

Les tourbières nordiques représentent l'un des plus importants puits terrestres de carbone (C) organique et sont aussi d'importantes sources de gaz à effet de serre. Afin de mieux comprendre entièrement le rôle des tourbières dans le cycle global du C, il est essentiel d'améliorer les connaissances sur les processus contrôlant l'accumulation à long terme du C dans ces écosystèmes à l'échelle régionale. Au cours des millénaires, l'accumulation du C dans les tourbières est influencée par une combinaison de facteurs internes (e.g. topographie du bassin) et externes (e.g. climat, feux). Cette thèse vise à évaluer l'influence de la variabilité climatique holocène sur la dynamique du C dans les tourbières maritimes de la région de la Côte-Nord dans l'est du Canada. Les tourbières ombrotrophes (bogs) sont répandues sur les deltas postglaciaires le long de l'estuaire et du golfe du St. Laurent où elles représentent d'importantes composantes hydrologiques et écologiques au sein des environnements côtiers. Dans ce projet, le rôle du climat sur le développement à long terme de ces tourbières a été évalué en utilisant une approche comparative multi-sites entre deux régions écoclimatiques (boréal et subarctique) près de Baie-Comeau (BC) et Havre-St-Pierre (HSP). La séquestration du C au cours de l'Holocène a été évaluée en calculant les taux apparents d'accumulation à long terme du C (LORCA) au sein de huit tourbières. Les variations temporelles de séquestration du C ont été comparées aux reconstitutions écohydrologiques basées sur l'analyse des macrorestes végétaux et des thécamoebiens fossiles. L'influence des feux a aussi été déterminée par l'analyse à haute résolution des charbons de bois macroscopiques.

Sur la Côte-Nord, la chronologie de formation des tourbières est étroitement liée aux fluctuations holocènes des niveaux marins. L'accumulation de la tourbe s'est amorcée sur les terrasses deltaïques sableuses suite au retrait de la Mer de Goldthwait à partir de 7500 ans étal. BP et au-dessus des dépôts silto-argileux associés à la transgression marine laurentienne après 4200 ans étal. BP.

Cette thèse a montré plusieurs évidences de l'influence climatique sur la dynamique écohydrologique des tourbières au cours de l'Holocène. Le rythme des successions végétales initiales semble avoir été indirectement influencé par des conditions climatiques régionales distinctes. La transition vers l'ombrotrophie est survenue plus rapidement dans la région de Baie-Comeau, favorisée par une accumulation rapide de la tourbe, une indication que le climat était plus propice au développement des tourbières dans cette région. De plus, les données paleohydrologiques montrent des périodes cohérentes de changements à l'échelle régionale correspondant à des phases climatiques documentées dans le nord-est du Canada. L'assèchement des surfaces des bogs et des conditions hydrologiques plus variables après 3000 ans étal. BP coïncident avec la transition entre l'optimum climatique holocène et le refroidissement Néoglaciaire.

Le climat de la Côte-Nord, caractérisé par une faible évapotranspiration et une humidité atmosphérique élevée, a favorisé une croissance soutenue des sphaignes au cours des phases ombrotrophes. Les feux locaux ont été rares (intervalles de feux : 1160 ans à Lebel et 1640 ans à Plaine) et dans l'ensemble ils ont eu peu d'impacts sur la végétation de surface tels que l'élimination des épinettes noires (*Picea mariana*) et l'introduction de mousses typiques des buttes de milieux ombrotrophes.

Au cours de l'Holocène, la différence des taux d'accumulation du C entre Havre-St-Pierre (LORCA : 16–46 g m^{-2} yr^{-1}) et Baie-Comeau (LORCA : 53–68 g m^{-2} yr^{-1}) suggère un contrôle climatique régional sur le bilan à long terme du C. Les taux d'accumulation du C plus élevés et plus stables dans les bogs de Baie-Comeau ont été favorisés par une stabilité des conditions hydrologiques. En contrepartie, les bogs de Havre-St-Pierre ont connu d'importantes variations d'accumulation du C associées notamment à des variations marquées des nappes phréatiques. Cette différence révèle une plus grande sensibilité des tourbières de Havre-St-Pierre aux changements climatiques passés, spécialement au cours du Néoglaciaire lorsque leur hydrologie et leur végétation semblent avoir été affectées par des saisons de croissance plus courtes et possiblement une influence accrue du gel saisonnier. La diminution graduelle de séquestration du C dans les tourbières étudiées résulte d'une influence combinée du développement vertical autogène et des changements climatiques régionaux survenus au cours de l'Holocène récent.

Les résultats suggèrent en effet que les trajectoires du développement holocène des tourbières ont été ultimement contrôlées par le climat mais plusieurs changements écohydrologiques à l'échelle décennale et centenaire semblent résulter principalement de processus internes. Les patrons distincts de variations des taux de séquestration du C entre les sites suggèrent que des facteurs internes de chaque tourbière ont affecté le bilan net d'accumulation du C. L'influence du climat sur le bilan à long terme du C dans les tourbières de la Côte-Nord s'est fait par l'intermédiaire d'interactions complexes entre les processus d'accumulation et de décomposition et les variations hydrologiques. Ces résultats sont conformes avec les modèles conceptuels et suggèrent que la dynamique des tourbières est aussi influencée par des processus autogènes et des mécanismes de rétroaction écohydrologiques.

Mots clés : Tourbières ombrotrophes, Holocène, carbone, paléoécologie, paléohydrologie

ABSTRACT

Northern peatlands are one of the largest terrestrial sink of organic carbon (C) and are important sources of greenhouse gases. In order to understand past and future roles of peatlands in the global C cycle, it is essential to improve knowledge on the processes controlling the long-term C sequestration in these ecosystems at a regional scale. Over millennial timescales, peatland C accumulation is driven by a combination of internal (e.g. basin topography) and external factors (e.g. climate, fires). This thesis aims to evaluate the influence of Holocene climate variability on the C dynamics within the maritime peatlands of the Québec North Shore (eastern Canada). Ombrotrophic peatlands (bogs) are widespread on the postglacial deltas along the Estuary and Gulf of St. Lawrence and are major hydrological and ecological components within the coastal environments. In this project, the role of climate over their long-term development was evaluated using a comparative multi-site approach within two peatland complexes from two maritime ecoclimatic regions (boreal and subarctic) near Baie-Comeau (BC) and Havre-St-Pierre (HSP). The Holocene C sequestration was evaluated by calculating the long-term average rates of C accumulation (LORCA) within eight peatlands. The temporal variations in C sequestration were compared with the ecohydrological reconstructions based on plant macrofossils and testate amoebae analyses. The influence of fire was also evaluated using high-resolution macroscopic charcoal analyses.

In the Québec North Shore, the timing of peat inception was closely associated with Holocene sea-level variations. Peat accumulation started over sandy deltaic terraces after the withdrawal of the Goldthwait Sea from 7500 cal yr BP and above silt-clay deposits associated to the Laurentian marine transgression after 4200 cal yr BP.

This thesis showed many evidences of climatic influences on the ecohydrological dynamics of the peatlands over the Holocene. The rates of early vegetation successions seem to have been indirectly influenced by distinct regional climatic conditions. The transitions to ombrotrophy have occurred much earlier in the peatlands of Baie-Comeau favored by faster initial rates of peat accumulation, an indication that climate conditions were more propitious to peat development in this region. Moreover, the paleohydrological data show coherent periods of changes at the regional scale which correspond to documented climate periods in northeastern Canada. The surface drying and increased hydrological variability in the bogs after 3000 cal yr BP coincide with the transition from the Holocene Climatic Optimum to the Neoglacial cooling.

The climate of the St. Lawrence North Shore characterised by low evapotranspiration and high moisture inputs was very propitious to *Sphagnum* growth during the ombrotrophic stages. The local fires were rare within the bog stages (fire-free intervals: 1160 yr in Lebel and 1640 yr in Plaine) and overall they had minor impacts on surface vegetation such as the removal of black spruce and the appearance of non-*Sphagnum* mosses typical of bog hummocks.

Over the Holocene, the difference in the C sequestration rates between Havre-St-Pierre (LORCA : 16–46 g m^{-2} yr^{-1}) and Baie-Comeau (LORCA : 53–68 g m^{-2} yr^{-1}) suggests a regional climatic control on the long-term C balance. The higher and steadier rates of C accumulation in the bogs of Baie-Comeau were promoted by stable hydrological conditions which contrast with the bogs of Havre-St-Pierre that experienced high fluctuations in C sequestration related to marked shifts in water table levels. This reveals an increased sensitivity of these bogs to past hydroclimatic changes especially during the Neoglacial period when peatland hydrology and vegetation were likely affected by shorter growing seasons and possibly longer periods of seasonal freezing. In the studied peatlands, the gradual slowdown in the rate of C production was driven by a combined influence of autogenic vertical development and regional climate change during the late-Holocene.

The results suggest that the long-term trajectories of peatland development were driven by climate but many decadal-to-centennial-scale changes in ecohydrological conditions seem to have resulted primarily from internal processes. The distinct patterns of C sequestration rates between sites suggest a complex response to regional climate variability. Our data suggest that the influence of climate on the long-term C balance was mediated by complex interactions between the accumulation/decay processes and the water table changes. These results are in line with conceptual models and suggest that peatland dynamics are also influenced by autogenic processes and ecohydrological feedback mechanisms.

Keywords : Ombrotrophic peatlands, Holocene, carbon dynamics, paleoecology, paleohydrology

INTRODUCTION

Mise en contexte

Les tourbières couvrent environ 3% de la surface terrestre et sont principalement concentrées dans les régions boréales et subarctiques de l'hémisphère nord (Charman, 2002). Au Canada, ces écosystèmes occupent environ 12% du territoire (~1,1 million km^2) et sont très abondants dans les Basse-terres de la Baie d'Hudson et de James et dans les régions maritimes de l'est du pays (Tarnocai, 2006). Les tourbières sont caractérisées par un bilan positif entre la production de biomasse et la décomposition de la matière organique qui en fait d'importants puits terrestres de carbone (C). Au cours de l'Holocène (derniers ~11 700 ans), les tourbières nordiques auraient entreposé environ 547 Gt de C (Yu *et al.*, 2010) ce qui correspond à plus de 50 % du C contenu dans l'atmosphère terrestre (Frolking et Roulet, 2007) et au tiers du C organique contenu dans les sols de la planète (Gorham, 1991). Les tourbières représentent aussi d'importantes sources naturelles de méthane (Lelieveld, Crutzen et Dentener, 1998) et sont une composante majeure du cycle global du C et de sa rétroaction au sein des cycles climatiques planétaires.

Les tourbières ombrotrophes (bogs) qui sont alimentées uniquement par les précipitations atmosphériques ont un grand potentiel de séquestration du C (Clymo, Turunen et Tolonen, 1998 ; Turunen *et al.*, 2002) notamment associé aux faibles taux de décomposition des sphaignes (Moore et Basiliko, 2006). Les bogs sont très répandus dans les régions maritimes côtières de l'est du Canada (Glaser and Janssens, 1986). Dans l'est du Québec, sur la Côte-Nord, de larges complexes de tourbières ombrotrophes se sont formées sur les dépôts fluvioglaciaires et deltaïques formés par les tributaires de la mer postglaciaire de Goldthwait (Bernatchez, 2003). Ces écosystèmes jouent un rôle fondamental sur les processus hydrologiques et les bilans de C au sein des environnements côtiers de l'estuaire et du golfe du Saint-Laurent. Malgré leur importance spatiale et écologique, le développement holocène de ces tourbières demeure peu connu et jusqu'à présent aucune étude n'a évalué la dynamique d'accumulation à long terme du C dans ces milieux. La compréhension des facteurs qui ont influencés la dynamique écohydrologique et l'accumulation du C dans ces tourbières dans le passé est essentielle pour évaluer leur rôle actuel et futur dans le cycle global du C.

Dans l'est du Canada, peu d'études paléoécologiques ont été effectuées dans les tourbières maritimes (Garneau, 1998 ; Payette *et al.*, 2013 ; Robichaud et Bégin, 2009 ; Tolonen, Huttunen et Jungner, 1985). Les travaux antérieurs portant sur le développement des tourbières ont été concentrés dans les Basses-terres de la Baie d'Hudson et de James (Arlen-Pouliot et Bhiry, 2005 ; Beaulieu-Audy *et al.*, 2009 ; Bhiry, Payette et Robert, 2007 ; Bunbury, Finkelstein et Bollmann, 2012 ; Glaser *et al.*, 2004 ; Lamarre, Garneau et Asnong, 2012 ; Magnan, Lavoie et Payette, 2012 ; van Bellen, Garneau et Booth, 2011) et dans le sud du Québec et l'est de l'Ontario (e.g. Elliott, Roe et Patterson, 2012 ; Lavoie et Richard, 2000 ; Muller, Richard et Larouche, 2003).

La dynamique holocène d'accumulation du C demeure en général peu connue dans les tourbières boréales et subarctiques de l'Amérique du Nord, en particulier dans les régions maritimes de l'est du Canada. Les taux apparents d'accumulation à long terme du C (LORCA) ont été évalués dans des tourbières de la Baie de James et varient entre 10 et 34 g C m^{-2} an^{-1} (Loisel et Garneau, 2010 ; van Bellen *et al.*, 2011) alors qu'un taux moyen de 19 g C m^{-2} an^{-1} a été calculé pour des tourbières ombrotrophes de l'est du Canada (Turunen *et al.*, 2004). Les valeurs de LORCA peuvent varier significativement entre les tourbières d'une même région en raison d'influences topographiques et géomorphologiques (Korhola *et al.*, 1996 ; Roulet *et al.*, 2007 ; van Bellen *et al.*, 2011). De plus, les LORCA sont des taux apparents et tendent ainsi à diminuer avec l'âge des dépôts de tourbe en raison de la décomposition continuelle dans la portion anoxique des tourbières (catotelme) (Clymo, Turunen et Tolonen, 1998 ; Turunen, Tomppo et Tolonen, 2002).

Influence du climat sur l'accumulation du C dans les tourbières

L'accumulation nette du C dans les tourbières est directement influencée par les conditions climatiques en particulier les précipitations et la température (Malmer *et al.*, 2005 ; Yu, Beilman and Jones, 2009) qui contrôlent le bilan entre la productivité primaire nette (PPN) et la décomposition dans la portion supérieure oxygénée de la tourbière (acrotelme). Cette influence climatique se reflète par une tendance nord-sud d'augmentation des taux d'accumulation du C qui est positivement corrélée avec la température moyenne annuelle (Beilman *et al.*, 2009), la durée de la saison de croissance (Clymo, Turunen et Tolonen, 1998) et le rayonnement photosynthétiquement actif (Charman *et al.*, 2013). Dans les tourbières nordiques, les taux d'accumulation du C les plus élevés sont répertoriés dans les régions avec des températures moyennes annuelles variant de 0 à 2.5°C et des précipitations moyennes annuelles de 450 à 550 mm (Yu, Beilman et Jones, 2009).

L'influence des variations climatiques sur la séquestration du C dans les tourbières se fait généralement par l'intermédiaire de l'hydrologie et de la végétation de surface (Malmer *et al.*, 2005 ; van Bellen, Garneau et Booth, 2011). La relation entre le climat et l'accumulation du C est complexifiée par l'existence de mécanismes de rétroaction entre la production de la tourbe et la profondeur de la nappe phréatique (Belyea et Clymo, 2001 ; Swindles *et al.*, 2012). Le bilan net d'accumulation du C est principalement lié aux conditions oxiques dans l'acrotelme et varie ainsi en fonction de la profondeur de la nappe phréatique. L'accumulation de la tourbe est généralement maximale avec des profondeurs intermédiaires de nappes phréatiques (~10 cm) favorisant une décomposition minimale et une productivité végétale optimale (Belyea et Clymo, 2001). Les changements dans les communautés végétales peuvent aussi affecter l'accumulation du C en modifiant la production de biomasse (PPN) et le taux de décomposition dans l'acrotelme (qualité de la litière) (Belyea et Malmer, 2004 ; Laiho et Vasander, 2003).

Les tourbières comme archives paléoclimatiques

Dans les tourbières ombrotrophes, les conditions hydrologiques en surface de sont directement influencées par le bilan entre les précipitations et l'évapotranspiration (Charman *et al.*, 2009). Dans les régions océaniques, les nappes phréatiques des bogs seraient principalement contrôlées par le déficit hydrologique au cours de la saison de croissance, lequel est essentiellement déterminé par les précipitations atmosphériques (Charman, 2007). En Europe, plusieurs travaux ont interprétés les variations d'humidité de surface des bogs déduites de proxies biotiques (e.g. thécamoebiens, macrorestes végétaux) et abiotique (e.g. humification de la tourbe) comme des changements climatiques régionaux (e.g. Barber, Chambers et Maddy, 2003 ; Borgmark, 2005 ; Charman et Hendon, 2000 ; Hendon, Charman et Kent, 2001 ; Mauquoy et Barber, 1999). Ces interprétations sont supportées par la cohérence des variations hydrologiques au sein d'une même tourbière et entre les régions (Charman, 2006 ; Charman *et al.*, 1999 ; Hendon, Charman et Kent, 2001 ; Langdon *et al.*, 2012) de même que les liens avec des reconstitutions climatiques indépendantes (Barber et Langdon, 2007).

Toutefois, le rôle du climat sur la variabilité hydrologique à long terme des tourbières demeure très débattu dans la communauté scientifique. Des modèles conceptuels suggèrent que les nappes phréatiques des tourbières seraient résistantes aux variations climatiques de courte durée (e.g. échelle décennale à centenaire) en raison de mécanismes de rétroaction écohydrologiques (Belyea et Clymo, 2001 ; Swindles *et al.*, 2012). En contrepartie, la corrélation de reconstructions paléohydrolologiques à haute résolution avec des données climatiques instrumentales suggère que l'hydrologie de surface des bogs est étroitement liée au bilan régional des précipitations et de l'évapotranspiration (Booth, 2010 ; Charman, 2007 ; Charman *et al.*, 2012).

L'influence des variations climatiques sur l'hydrologie des tourbières ombrotrophes demeure généralement peu connue particulièrement dans les régions maritimes boréales du nord-est américain. Très peu de reconstitutions paléohydrologiques ont été réalisées dans les tourbières de l'est du Canada et la variabilité climatique holocène demeure peu documentée dans cette région. Dans la région de la Côte-Nord, les dépôts de tourbe offrent un grand potentiel pour effectuer des reconstitutions paléoclimatiques car ils sont souvent entièrement ombrotrophes et dominées par les sphaignes (Glaser et Janssens, 1986 ; Payette *et al.*, 2013).

Rôle des facteurs internes et externes sur le développement holocène des tourbières boréales

Les modèles conceptuels (e.g. Belyea et Baird, 2006 ; Swindles *et al.*, 2012) et les études paléoécologiques (e.g. Payette, 1988 ; Loisel and Garneau, 2010 ; van Bellen, Garneau et Booth, 2011) suggèrent que le développement des tourbières est influencé par une combinaison de facteurs internes (autogènes) et externes (allogènes). Globalement, le climat a influencé l'expansion des tourbières au cours de l'Holocène (Korhola *et al.*, 2010 ; MacDonald *et al.*, 2006). Au Québec nordique, la formation des tourbières subarctiques a été très active entre ~6300 et ~4200 ans étal. BP favorisé par un climat plus chaud et possiblement plus humide à cette époque (Payette, 1984). Sur la côte nord du golfe du St. Laurent, la paludification et l'expansion des tourbières sur les sables deltaïques aurait été favorisée par une transition vers des conditions climatiques plus humides à partir de 5500 ans étal. BP (Payette *et al.*, 2013).

A l'échelle régionale, la formation des tourbières et les successions végétales initiales sont aussi influencées par les contextes géomorphologiques et topographiques (Bauer, Gignac et Vitt, 2003 ; Bhiry, Payette et Robert, 2007 ; Glaser *et al.*, 2004 ; Ireland *et al.*, 2013 ; van Bellen *et al.*, 2011). Dans les régions maritimes côtières, les variations du niveau marin et le relèvement isostatique ont directement influencés la formation des tourbières au cours de l'Holocène (Bhiry, Garneau et Filion, 2000 ; Robichaud et Bégin, 2009). La morphologie du bassin sédimentaire joue un rôle important dans le développement à long terme des tourbières. Lorsque l'expansion latérale des tourbières est contrainte par la topographie, le stockage de l'eau dans le catotelme tend à diminuer graduellement en lien avec la croissance verticale de la tourbe (Belyea et Clymo, 2001). Ainsi, à l'échelle des millénaires, la diminution graduelle de l'accumulation de la tourbe a été attribuée à l'assèchement graduel des surfaces associé au développement vertical autogène (Yu *et al.*, 2003b).

Dans les tourbières boréales et subarctiques, les successions végétales suivent généralement une transition partant de conditions initiales minérotrophes (fens) vers des conditions ombrotrophes (bogs). Cette transition du régime trophique est caractérisée par l'isolement des communautés biotiques en surface des tourbières par rapport aux eaux de la nappe phréatique enrichies en éléments nutritifs. L'ombrotrophication peut résulter essentiellement de l'accumulation verticale de la tourbe, toutefois des changements climatiques peuvent aussi influencer le rythme de cette transition trophique (Hughes et Barber, 2004 ; Hughes et Dumayne-Peaty, 2002 ; Payette, 1988 ; Robichaud et Bégin, 2009).

Dans le nord-est du Canada, des études ont montré que la dynamique écohydrologique des tourbières a été influencée par des variations climatiques significatives au cours de l'Holocène notamment lors du refroidissement Néoglaciaire et du Petit Âge glaciaire (Bunbury, Finkelstein et Bollmann, 2012 ; Lamarre, Garneau et Asnong, 2012 ; Loisel et Garneau, 2010; van Bellen *et al.*, 2013; van Bellen, Garneau et Booth, 2011). Dans l'est de Terre-Neuve, l'étude de Hughes *et al.*, (2006) a montré l'existence de liens persistants entre les variations d'humidité de surface d'un bog et le forçage océanique et solaire au cours de l'Holocène.

Dans les tourbières ombrotrophes boréales et subarctiques, les feux représentent une importante perturbation écologique pouvant affecter les successions végétales et l'hydrologie (Camill *et al.*, 2009 ; Turetsky *et al.*, 2002). Toutefois, la fréquence des feux dans les tourbières ombrotrophes continentales est généralement nettement inférieure à celles des forêts régionales (Kuhry, 1994 ; Magnan *et al.*, 2009 ; van Bellen *et al.*, 2012). La fréquence des feux dans les tourbières maritimes de l'est du Canada demeure très peu documentée et aucune étude n'a évalué l'impact de cette perturbation sur les successions végétales dans ces écosystèmes.

Cadre géographique du projet

Dans le cadre de cette thèse, les tourbières de deux secteurs de la Côte-Nord ont été étudiées suivant un gradient de nordicité et d'influence océanique entre la région de l'estuaire (Baie-Comeau) et du golfe du St. Laurent (Havre-St-Pierre). Les tourbières étudiées proviennent de deux des plus grandes formations deltaïques de la Côte-Nord (Manicouagan et La Romaine). Dans ces deux régions, les tourbières sont principalement retrouvées au sein de deux contextes géomorphologiques. Elles se sont formées d'une part sur les terrasses sableuses deltaïques (altitude >16 mètres a.s.l.) émergées suite au retrait de la Mer de Goldthwait. Plus près du littoral, dans les portions inférieures des deltas (altitude <16 mètres a.s.l.), les tourbières recouvrent des dépôts silto-argileux associés à la transgression marine laurentienne qui a érodée la côte jusqu'à une altitude de ~14-16 mètres au-dessus du niveau marin actuel vers 4.3 ka BP et qui aurait reculé à un niveau similaire à l'actuel vers 3 ka BP (Bernatchez, 2003).

Le climat régional de la Côte-Nord est de type maritime avec des précipitations annuelles totalisant approximativement 1000 mm dans les deux régions étudiées. La température moyenne annuelle est de 1.5 °C à Baie-Comeau et de 1.1°C près de Havre-St-Pierre (Environnement Canada, 2013). La durée moyenne des saisons de croissance diminue significativement entre Baie-Comeau et Havre-St-Pierre alors que le nombre de degré-jour de croissance (>0°C) passe de ~1924 à ~1779 (Hutchinson *et al.*, 2009).

L'influence croissante des conditions subarctiques vers le golfe du St. Laurent, notamment associée au courant froid du Labrador, se reflète entre autre par l'ouverture graduelle du couvert forestier dans les environnements côtiers. La région de Baie-Comeau appartient au domaine bioclimatique de la sapinière à bouleau blanc (*Betula papyrifera*) alors que Havre-St-Pierre se retrouve à la limite est de la forêt boréale fermée au sein de la pessière à mousse. La fréquence des feux dans les forêts régionales demeurent très peu documentée sur la Côte-Nord.

Dans les tourbières, le gradient climatique vers le nord-est s'exprime notamment par l'expansion graduelle des superficies occupées par les mares. Cette transition dans la morphologie de surface des bogs peut résulter d'une diminution du potentiel d'évapotranspiration et/ou d'une influence accrue des phénomènes de gel dans la région du golfe du St. Laurent tel que suggéré par Glaser et Janssens (1986). Près de Baie-Comeau, la surface des bogs est convexe et les mares sont généralement peu abondantes et de petites tailles. Dans la région de Havre-St-Pierre, les tourbières en forme de plateaux sont parsemées de mares de tailles variables et de petits lacs occupant jusqu'à environ 50 % des surfaces. Sur la Côte-Nord, l'épaisseur moyenne des dépôts de tourbe tend à diminuer entre l'estuaire et le golfe du St. Laurent, une indication que le climat régional aurait contrôlé l'accumulation de la tourbe et du C associé au cours de l'Holocène.

Objectifs et approche méthodologique

L'objectif principal de cette thèse est d'évaluer le rôle du climat sur l'accumulation du C au cours de l'Holocène dans les tourbières ombrotrophes de deux régions écoclimatiques de la Côte-Nord. La première hypothèse qui a été testée est que les contextes climatiques distincts de Baie-Comeau et de Havre-St-Pierre auraient influencés des dynamiques écohydrologiques différentes dans les tourbières au cours de l'Holocène. Ensuite, nous avons émis l'hypothèse que les tourbières de Havre-St-Pierre auraient des bilans d'accumulation à long terme du C plus faibles en raison de leur position écotonale à la limite climatique de distribution des tourbières nordiques (*sensu* Yu, Beilman et Jones, 2009).

Une approche comparative multi-sites a été utilisée afin d'évaluer l'influence respective des facteurs externes (e.g. climat, feux) et internes (e.g. accumulation de la tourbe, topographie) sur le développement des tourbières. Dans un premier temps, afin d'évaluer l'influence potentielle du climat sur l'accumulation du C et la dynamique des tourbières, deux régions écoclimatiques maritimes (boréal et subarctique) ont été comparées. Deuxièmement, au sein de chaque région, des tourbières d'âges variables ont été étudiées dans deux contextes géomorphologiques distincts sur les sables deltaïques (alt. 16-30 mètres a.s.l.) et les dépôts marins (alt. 10-14 mètres a.s.l.). Cette approche permet de déterminer l'influence du substrat minéral sur la formation initiale des tourbières et d'évaluer l'influence de l'âge des dépôts de tourbe sur les taux d'accumulation à long terme du C.

Les tourbières étudiées ont été sélectionnées en fonction de leur représentativité à l'échelle régionale suite à des analyses de photographies aériennes et des relevés sur le terrain (e.g. inventaires de végétation, mesures de profondeurs de tourbe et détermination du sédiment basal). Les profils de tourbe ont été prélevés dans la portion centrale, la plus profonde afin d'obtenir les plus longues reconstitutions temporelles possibles et de minimiser l'influence topographique (*sensu* Belyea et Clymo, 2001). Les carottes de tourbe ont été prélevées à partir de lanières dominées par les sphaignes car ces microformes sont plus sensibles aux changements environnementaux que les buttes (Nordbakken, 1996). Le cadre chronologique de l'ensemble des profils est basé sur la datation au radiocarbone (^{14}C) de 53 échantillons de pièces végétales fossiles par spectrométrie de masse par accélérateur (SMA).

Spécifiquement les objectifs de la thèse sont de :

1) Évaluer le rôle du climat et des feux sur les successions végétales à long terme

Dans le chapitre 1, le développement initial et les successions végétales à long terme ont été documentés par l'analyse des macrorestes végétaux. L'influence des feux sur la végétation a également été évaluée à l'aide d'analyses de charbons de bois de taille macroscopique sur une séquence de tourbe provenant de chaque région.

2) Détecter l'influence des variations climatiques holocènes sur la paléohydrologie des tourbières.

Dans le chapitre 2, des reconstitutions détaillées des variations hydrologiques basées sur l'analyse de thécamoebiens fossiles ont été réalisées à l'aide d'une fonction de transfert (Lamarre *et al.*, 2013). Les changements hydrologiques synchrones au sein de chaque région sont utilisés pour évaluer l'influence du climat sur l'hydrologie de surface des tourbières au cours de l'Holocène.

3) Évaluer la dynamique à long terme du C entre deux contextes écoclimatiques maritimes boréaux et subarctiques.

Dans le chapitre 3, l'influence du climat régional sur l'accumulation du C a été évaluée en comparant les valeurs de LORCA et les variations temporelles d'accumulation du C entre les tourbières. Les valeurs d'accumulation du C sont basées sur des mesures de densité volumique sèche de la tourbe et de pertes au feu. Les variations d'accumulation du C ont été comparées aux processus de décomposition aérobique (nappe phréatique et humification de la tourbe) sur un profil de tourbe au sein de chaque région.

CHAPITRE I

HOLOCENE DEVELOPMENT OF MARITIME OMBROTROPHIC PEATLANDS ALONG THE ESTUARY AND GULF OF ST. LAWRENCE IN EASTERN CANADA

Gabriel Magnan[1], Michelle, Garneau[1], Serge Payette[2]

[1]GEOTOP - Université du Québec à Montréal (UQAM), C.P. 8888 Succursale Centre-ville, Montréal, Québec, Canada H3C 3P8.

[2]Département de biologie, 1045 avenue de la Médecine, Université Laval, Québec, Canada, G1V 0A6.

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Résumé

L'analyse de plantes macrofossiles et les datations au radiocarbone ont été utilisées pour reconstituer les successions végétales à long terme au sein de tourbières ombrotrophes (bogs) qui se sont développées sur les deltas postglaciaires de l'estuaire (Manicouagan) et du golfe du St. Laurent (La Romaine). L'histoire des feux locaux a été reconstituée dans une tourbières au sein de chaque région par l'analyse à haute résolution de charbons de bois macroscopiques. Dans les deux régions, la chronologie de formation des tourbières est étroitement liée aux fluctuations holocènes des niveaux marins. L'accumulation de la tourbe s'est amorcée sur les terrasses deltaïques sableuses après le retrait de la Mer de Goldthwait à partir de 7500 ans étal. BP et au-dessus des dépôts silto-argileux après la transgression laurentienne vers 4200 ans étal. BP. Les successions végétales initiales ont été similaires au sein des deux contextes géomorphologiques où des fens pauvres dominés par les cypéracées et le mélèze laricin (*Larix laricina*) se sont développés rapidement suite à l'émersion des terres. L'ombrotrophication est survenue beaucoup plus rapidement dans les tourbières de Baie-Comeau. La différence régionale dans le rythme des transitions trophiques est associée à une dynamique distincte d'accumulation de la tourbe entre les deux régimes climatiques. En général, le climat maritime de la Côte-Nord a favorisé une croissance soutenue des sphaignes au cours des phases ombrotrophes. Les feux locaux ont été rares au sein des bogs (intervalles de feux : 1160 ans à Lebel et 1640 ans à Plaine) et ont eu des impacts limités sur la végétation tels que l'élimination des épinettes noires (*Picea mariana*) et l'introduction de mousses caractéristiques de milieux ombrotrophes secs.

Abstract

Plant macrofossil analyses and radiocarbon dating were used to reconstruct the long-term vegetation successions within ombrotrophic peatlands (bogs) distributed on postglacial deltas of the Estuary (Manicouagan) and of the Gulf of St. Lawrence (La Romaine). Local-scale fire history was also reconstructed within one peat sequence from each region using high-resolution macroscopic charcoal analysis. In both regions, the timing of peat inception was closely associated with Holocene sea-level fluctuations. Peat accumulation started over the sandy deltaic terraces after the withdrawal of the Goldthwait Sea from 7500 cal yr BP and above silt-clay deposits following a marine transgression after 4200 cal yr BP. The early vegetation successions were similar in both geomorphic settings where poor Cyperaceae-dominated fens with eastern larch (*Larix laricina*) established in most sites after land emergence. The ombrotrophication occurred significantly earlier in the peatlands of Baie-Comeau. The regional difference in the rate of the fen-bog transitions is related with different rates of initial peat accumulation between the two ecoclimatic regions. Overall, the maritime climate of the North Shore has favored a sustained growth of *Sphagnum* within the ombrotrophic stages. The incidence of local fires was rather low (fire-free intervals : 1160 yr in Lebel and 1640 yr in Plaine) and in most cases, had only minor impacts on peatland vegetation such as black spruce removal and introduction of non-*Sphagnum* mosses typical of dry ombrotrophic conditions.

1.1. Introduction

Peatlands are an important feature of the boreal landscape in North America. In Canada, they cover an estimated area of 1.1 million km^2 corresponding to ~12 % of the land surface (Tarnocai *et al.*, 2005). Ombrotrophic peatlands (bogs) are common ecosystems in the maritime regions of the Atlantic seaboard (Glaser and Janssens, 1986). Along the north coast of the Estuary and Gulf of St. Lawrence, most ombrotrophic peatlands have developed over the postglacial deltaic sands deposited by the tributaries of the Goldthwait Sea. Near the shoreline, some peatlands formed above silt-clay deposits associated with a marine transgression that reached an altitude of 14-16 meters above present-day sea-level around 4.3 ka BP and retreated to the present-day level around 3 ka BP (Bernatchez, 2003 ; Dionne, 2001).

Despite the large extent and ecological significance of peat bogs in the maritime region of eastern Canada, little is known on the context of peat initiation and long-term development of these ecosystems. Only few paleoecological studies have been conducted in some maritime bogs of eastern Canada (Garneau, 1998 ; Hughes *et al.*, 2006 ; Payette *et al.*, 2013 ; Robichaud and Bégin, 2009 ; Tolonen, Huttunen and Jungner, 1985). Most previous peatland paleoecological studies focussed on continental boreal regions of the James Bay and Hudson Bay lowlands (Arlen-Pouliot and Bhiry, 2005 ; Beaulieu-Audy *et al.*, 2009 ; Bunbury, Finkelstein and Bollmann, 2012 ; Glaser *et al.*, 2004 ; Lamarre, Garneau and Asnong, 2012 ; Magnan, Lavoie and Payette, 2012 ; van Bellen, Garneau and Booth, 2011) and in temperate peatlands within the St. Lawrence Lowlands (Elliott, Roe and Patterson, 2012 ; Lavoie and Richard, 2000 ; Muller, Richard and Larouche, 2003).

In northern peatlands, the long-term vegetation successions typically follow a hydrosere from shallow open water to poor minerotrophic conditions (fen) and most often to ombrotrophic conditions (bog). The rates and pathways of vegetation changes in peatlands are driven by a combination of internal processes (e.g. plant competition, peat build-up) and external factors (e.g. climate and fires). The isolation of peat-forming vegetation from nutrient-rich groundwater may result from water table drawdown or peat growth which can both be related with changes in the atmospheric moisture balance (Hughes and Barber, 2003). The geomorphic setting and morphology of the underlying mineral basin are important factors controlling the early vegetation pathways in peatlands (Bauer, Gignac and Vitt, 2003 ; Bhiry, Payette and Robert, 2007 ; Ireland *et al.*, 2013 ; van Bellen *et al.*, 2011). However, the Holocene climate variability has also been an important factor controlling peatland initiation and expansion at the regional scale (Korhola *et al.*, 2010 ; MacDonald *et al.*, 2006).

Natural disturbances such as fire play an important ecological role in ecosystems of the boreal biome (Payette, 1992) and although the incidence of fires in boreal peat bogs is usually much lower than in upland forests (Kuhry, 1994 ; Magnan, Lavoie and Payette, 2012), they can have significant impacts on long-term dynamics of peatlands (Camill *et al.*, 2009 ; Turetsky *et al.*, 2002). Previous paleoecological studies suggested that fires rarely occur in the maritime bogs of eastern Canada (Lavoie, Filion and Robert, 2009 ; Robichaud and Bégin, 2009), however the role of fire on the vegetation dynamics within these ecosystems remains largely unknown.

The main objective of the present study is to evaluate the factors controlling the long-term vegetation successions in the maritime ombrotrophic peatlands from the north shore of the Estuary and Gulf of St. Lawrence. More specifically we aim to 1) evaluate the timing and initial context of peat inception, 2) compare the long-term vegetation successions between two ecoclimatic and geomorphologic contexts and 3) evaluate the role of fire in the vegetation dynamics within the bogs.

1.2. Methods

1.2.1. Study area and sites

Two of the largest ombrotrophic peatland complexes in the North Shore of the Estuary and Gulf of St. Lawrence were investigated on the Manicouagan delta near Baie-Comeau (BC) and on the La Romaine delta near Havre-St-Pierre (HSP) (Fig. 1.1). In each region, peatlands were selected within two distinct geomorphic contexts below and above the Laurentian marine transgression limit (i.e. ~12 meters a.s.l.). Lebel, Plaine and Romaine peatlands developed over well-drained deltaic sands at altitudes ranging from 22 to 34 meters a.s.l. Baie, Manic and Morts peatlands are located on a lower terrace at 12-16 meters a.s.l. and are underlain by marine silt-clay deposits associated with the Laurentian transgression (Table 1.1).

In the Baie-Comeau area, the dominant types of peatland are raised bogs dominated by *Sphagnum* spp., ericaceous shrubs (mainly *Chamaedaphne calyculata*, *Rhododendron groenlandicum*, *Kalmia angustifolia*) and dwarf black spruce (*Picea mariana*). Near Havre-St-Pierre, the La Romaine delta is covered by large plateau bogs largely treeless and their surfaces are typically dominated by lichens (*Cladonia stellaris*, *C. mitis* and *C. stygia*) and ericaceous shrubs (*Rhododendron groenlandicum*, *Empetrum nigrum*, *K. angustifolia*) interspersed with patches of *Sphagnum* spp. (Fig. 1.2). Mean annual precipitation totals approximately 1000 mm in both regions and mean annual temperature is 1.5° C in Baie-Comeau and 1.1°C near Havre-St-Pierre (climate normals 1971-2000 ; Environment Canada). In the Baie-Comeau area, the regional forests are dominated by balsam fir (*Abies balsamea*) and black spruce along with paper birch (*Betula papyrifera*) and trembling aspen (*Populus tremuloides*). The forest cover is increasingly fragmented towards the Gulf of St. Lawrence as the climatic conditions become more oceanic and subarctic. Havre-St-Pierre is located at the eastern limit of the closed boreal forest within the spruce-moss domain characterised by dense stands of *Picea mariana* in protected sites and fragmented forest stands in more wind-exposed uplands.

1.2.2. Field work

At each site, measurements of peat thickness were performed with a metal probe at 25-m intervals along a series of transects distributed evenly throughout the peatland to evaluate the morphology of the underlying basin before selecting the coring sites. Peat cores were collected from the deepest section of the peatlands from *Sphagnum*-dominated lawn microforms using a Russian corer (diameter: 7.5 cm) (Jowsey, 1966). The uppermost 100 cm of the peat sequences was sampled with a Box corer (105×8×8 cm). The altitude (m.a.s.l) of the sampling location was determined using a differential global positioning system (DGPS) (Table 1.1). Sampled sediments were wrapped in plastic film, placed in PVC tubes and stored in the fridge at 4°C. In the laboratory, sub-samples were taken at every centimeter along the peat profiles for further analyses.

1.2.3. Plant macrofossil analyses

Vegetation assemblages were evaluated based on detailed plant macrofossil analyses (4-cm interval; 5-cm^3 peat samples) within two peatlands of Baie-Comeau (hereafter Lebel and Baie) and two peatlands of Havre-St-Pierre (hereafter Plaine and Morts) (Fig. 1.1). In each region, the two detailed plant macrofossil records have been compared with the stratigraphy of an adjacent peatland (Manic and Romaine) to evaluate the synchronicities of the trophic transitions at the scale of the delta. In these two peat sequences, the main vegetation types were evaluated within 2-cm^3 peat samples at 4-cm interval. Samples were processed according to the protocol of

Mauquoy, Hughes and van Geel (2010). Macrofossil remains were separated from the organic matrix by heating the material for about 30 minutes in a solution of potassium hydroxide (KOH 5%) to dissolve humic and fulvic acids. The material was wet-sieved through a 0.125-mm mesh screen and fossil plant remains were identified and counted in a petri dish using a stereomicroscope at 4 to 40x magnification. The volume percentages of the main vegetation types (*Sphagnum*, other bog mosses, brown mosses, Cyperaceae, wood, roots and leaves) were calculated using a gridded petri dish and dividing the area covered by each macrofossil by the total sample area. The other plant macrofossils (e.g. seeds, conifer needles, *Cenococcum* sclerotia) were quantified using a five-point scale of abundance (1= presence, 2= occasional, 3= frequent, 4= very frequent 5 = abundant). Hereafter in this study, we refer to bog mosses for non-*Sphagnum* bryophytes characteristic of dry ombrotrophic conditions and to brown mosses for species typical of wet minerotrophic environments.

References used for plant identification were Ireland (1982), Lévesque, Dinel and Larouche (1988), Mauquoy and van Geel (2007) and the plant macrofossil reference collection from the laboratory of continental paleoecology (Geotop, UQÀM, Canada). Botanical nomenclature follows Marie–Victorin (1995) for vascular plants and Crum and Anderson (1981) for mosses.

The degree of peat decomposition was evaluated in each sample based on the state of preservation of bryophytes from intact (1) to poorly preserved moss leaves (5). The plant macrofossil diagrams were plotted using software package C2 1.7.2 (Juggins, 2011). Macrofossil zones were determined visually according to the main changes in vegetation assemblages and peat decomposition.

1.2.4. Chronology

A total of 51 macrofossil plant remains (mostly *Sphagnum* stems and leaves) were submitted to Keck carbon cycle laboratory (University of Irvine, California) for

^{14}C dating with accelerator mass spectrometry (AMS) (Table 1.2). Radiocarbon dates were first conducted on the main stratigraphic transitions along the peat profiles and further dating was conducted to optimise the chronological control. Age-depth models were developed using Clam software (Blaauw, 2010) by applying linear interpolation between each dated level. Radiocarbon dates were calibrated using the IntCal09 calibration curve (Reimer *et al.*, 2009). The age of the peat surface was established at −60 yr BP (i.e. AD 2010 : year of coring). All dates were rounded to the nearest decade.

1.2.5. Charcoal analyses

Macroscopic charcoal analysis was conducted in the oldest peatland of each region that developed over the deltaic sands (Lebel, Plaine). Charcoal sampling was conducted at 1-cm interval from the fen-bog transition to the top of the cores, assuming that fires are very unlikely to occur under wet minerotrophic conditions. Peat samples (1-cm^{-3}) were soaked in a solution of sodium hydroxide (NaOH 10%) for 24 h to dissolve humic acids and bleach charcoal particles to help differentiating charred fragments from dark organic matter. The remaining material was wet sieved through a 0.5-mm mesh screen. The macroscopic charcoal fragments (long axis >0.5 mm) were counted under a Leica stereomicroscope at 16x magnification and recorded in two size-classes (0.5–2 mm) and (>2 mm). Macroscopic charcoal particles (>0.5 mm) are reliable indicators of local and/or nearby fire events whereas large charcoal pieces (>2 mm), which are unlikely transported over long distances, provide strong evidence of local fire events (Ohlson and Tryterud, 2000). Charcoal accumulation rates (CHAR; particles cm^{-2} yr^{-1}) were calculated using CHARSTER software 0.8.3 (Gavin, 2006). The charcoal series were resampled at equal time intervals defined as the median resolution of the record. The identification of local-scale fire events is based on distinct peaks in the CHAR record combined with the presence of large charcoal fragments >2 mm. The fire-free interval during the bog stage was calculated by dividing the age of the fen–bog transition by the number of fire events recorded.

1.3. Results

1.3.1. Vegetation successions and fire history in Lebel peatland (BC)

Peat inception on the sandy terrace at an altitude of 22 m.a.s.l. was dated at 5820 years cal BP. The basal peat consisted of highly decomposed *Sphagnum* with needles of *P. mariana* along with Cyperaceae (Zone L1) (Fig. 1.3; Table 1.3). *C. calyculata* shrubs rapidly colonized the site and the ectomycorrhizal fungi of *Cenococcum* (sclerotia) suggest relatively dry conditions in the early stage. This interpretation is supported by a water table depth (WTD) reconstruction based on testate amoebae analyses conducted on this peat core (chapitre 2 ; Table 1.3). Peat growth above the well-drained sediment was very rapid at this site (1.5 mm yr^{-1}; Fig. 1.3 and Fig. 1.4). Black spruce declined after 5500 cal BP and *Sphagnum* became dominant along with bog mosses (*Pohlia nutans* and *Pleurozium schreberi*) (Zone L2). From 5200 to 3500 cal BP, Ericaceae shrubs with sparse *Larix laricina* and sedges were present on the site (Zone L3). After 3500 cal BP, the abundance of Ericaceae decreased and spruce re-established shortly until approximately 3100 cal BP when slower peat accumulation was recorded (0.74 mm yr^{-1}; Zone L4).

In Lebel bog, the fire-free interval was evaluated at 1160 years, though all the fire events (n=5) occurred after 2400 cal BP. The most important charcoal peak around 2400 cal BP (Zone L5; Fig. 1.3 and Fig. 1.5) coincides with a drastic decline in *Sphagnum* at the expense of woody plant remains (Fig. 1.3). *Sphagnum* re-established after another fire dated at 2000 cal BP as spruce briefly recolonized the peatland surface. Over the last 1400 years, *Sphagnum* dominated the site with sedges and Ericaceae such as *C. calyculata* (Zone L6). Three local fires were recorded in

Zone L6 but no significant changes in surface vegetation are recorded by the macrofossil data except for an appearance of few *Dicranum* mosses. These three charcoal peaks contains few large particles >2 mm (Fig. 1.3) and may thus correspond with fires that occurred nearby without affecting the plant cover at the coring location.

1.3.2. Vegetation successions and fire history in Plaine peatland (HSP)

Plaine peatland developed over the Romaine river delta at an elevation of 30 m.a.s.l. The early peatland stage (7450-6560 cal BP) corresponds to a relatively poor treed fen colonised by *L. laricina*, Cyperaceae and brown mosses such as *Calliergon* spp. and *Warnstorfia* spp. (Zone P1; Fig. 1.6). *P. mariana* was present nearby on the site from 7200 cal BP. The fen-bog transition was dated around 6560 cal BP with the disappearance of brown mosses and the local establishment of *P. mariana* along with *Sphagnum* spp. and *P. nutans* mosses (Zone P2; 6560-5610 cal BP).

In Plaine, four local-scale fires occurred within the bog phase which represents a fire-free interval of 1640 years. *P. mariana* disappeared above the most important charred layer dated at 5760 cal BP (Fig. 1.5; Fig. 1.6) and S*phagnum* became dominant between 5520 and 4160 cal BP (Zone P3). A local fire that occurred *c.* 4200 cal BP coincides with a major slowdown in peat accumulation (from 0.98 to 0.24 mm yr^{-1}) and a rise in the water table as *Carex* spp. colonized the site along with *L. laricina* (Zone 4; 4160-2430 cal BP). After 2430 cal BP, *Sphagnum* remained dominant along with ericaceous shrubs and sporadic spruce cover (Zone P5). After 900 cal BP, sedges and Ericaceae became more abundant on the site (Zone P6). Two fires horizons with few large charcoal >2 mm were identified around 440 cal BP and 1950 AD (Fig. 1.5) but their impacts on the vegetation seem limited except for the disappearance of sedges after the most recent fire.

1.3.3. Vegetation successions in Baie peatland (BC)

Baie peatland has formed above a marine silt-clay deposit in the lowest section of the Manicouagan delta (12 m.a.s.l.) after the Laurentian marine transgression around 4200 cal BP. The marine influence at this site was confirmed by the identification of foraminifera tests at the peat/mineral contact (Fig. 1.7). The absence of brackish plants suggests that the influence of marine waters was short-lived. *Myrica gale* and *L. laricina* rapidly colonized the emerged silt-clay and a poor fen with brown mosses (*Calliergon* spp., *Warnstorfia* spp.) has developed (Zone B1; 4210-4080) as described by Garneau (1998). Peat accumulation rates were very high within the fen stage (2.8 mm yr^{-1}). Ombrotrophic conditions established after 4080 cal BP as suggested by a sharp decline in Cyperaceae and expansion of ericaceous shrubs (Zone B2). The disappearance of *L. laricina* after 3800 cal BP suggests that the peatland surface was isolated from the mineral-rich waters. The ericaceous shrub cover slightly declined after 3250 cal BP but *C. calyculata* remained abundant from 3250 to 2980 cal BP (Zone B3). After 3000 cal. BP, the abundance of woody plants and sclerotia of *Cenococcum* increased significantly when much lower peat accumulation rates are recorded (0.54–0.94 mm yr^{-1}; Zone B4) under particularly dry conditions (Table 1.3). *Dicranum* mosses were identified around 1850 cal BP (Zone B5). Between 980 and 280 cal BP (Zone B6), *Sphagnum* dominated the vegetation assemblages accompanied by bog mosses (*P. strictum*, *P. nutans* and *Dicranum* spp.) after 280 cal BP.

1.3.4. Vegetation successions in Morts peatland (HSP)

The influence of marine waters in the lower deltaic terrace (12 m.a.s.l) is shown by the foraminifera tests within the basal silt-clay sediment (Fig. 1.8). The initial conditions with abundant Cyperaceae remnants along with *Scheuchzeria* *palutris, Myrica gale, Menyanthes trifoliata* and sparse *Abies balsamea* correspond to the freshwater edge along the coastal topographic sequence as described by Garneau (1998). Peat accumulation started within a sedge-dominated environment (Zone M1; 3250-2980 cal BP) where the moss *Campylium stellatum* indicates minerotrophic conditions (Garneau, 1998 ; Vitt and Chee, 1990). *L. laricina* gradually colonized the peatland with brown mosses common to moderate-poor fens (*Tomenthypnum nitens*, *Calliergon* spp.) and ericaceous shrubs (*C. calyculata*, *Andromeda* spp.) (Zone M2; 2980-2520 cal BP). Ombrotrophic conditions developed from *c.* 2520 cal BP with the disappearance of brown mosses at the expense of *Sphagnum* mosses. *L. laricina* persisted through this transitional stage (Zone M3; 2520-2170 cal BP) but was replaced by *P. mariana* and bog mosses (*P. strictum*, *P. nutans* and *Dicranum* spp.) as the peatland surface was very dry (Zone M4; 2170-1470 cal BP; Table 1.3). Macrofossil data indicate a return to a *Sphagnum*-dominated cover with *C. calyculata* shrubs and sporadic presence of conifers after 1470 cal BP (Zone M5). After 500 cal BP, *L. laricina* and *P. mariana* were present locally but disappeared recently at the expense of heath shrubs and lichens (Zone M6).

1.4. Discussion

1.4.1. Role of internal and external factors on long-term peatland development

In the two studied regions, the initial context of peatland formation was similar and the timing of peat initiation was directly influenced by the Holocene sea-level variations. The three oldest peatlands (Plaine, Romaine and Lebel) developed above the emerged sandy deltaic terraces (22-34 m.a.s.l.) following the withdrawal of the Goldthwait Sea between 7500 and 5800 cal yr BP. Baie, Manic and Morts peatlands developed after 4200 cal yr BP above the silt-clay deposits (12-16 m.a.s.l.) associated with the Laurentian marine transgression.

Over the sandy terrace of La Romaine River (HSP), peat inception occurred around the same period in Plaine and Romaine peatland (Fig. 1.9). Peat accumulation in these sites coincides with an important period of fires recorded between 7360 and 6660 cal BP in a nearby forest (Payette *et al.*, 2013). Consequently, it is possible that peat inception in these sites result from a rise in the local water table associated with the removal of trees by fires (Payette *et al.*, 2013). In Lebel (BC), the ombrotrophic peat accumulated directly over the well-drained sands of the Manicouagan delta *c.* 5800 cal BP without an initial minerotrophic phase. Paludification process on the well-drained deltaic sands required major modification in the local hydrological balance which may be related to a change in the atmospheric moisture conditions. Payette *et al.*, (2013) showed that the paludification on the La Romaine delta coincided with the cessation of fire activity around 5500 cal BP which is an indication of a shift towards wetter conditions in this region. Paludification and extensive forest retreat in Northern Québec and Newfoundland after 6000 cal BP were also attributed to a change towards more oceanic climate conditions (Crawford, Jeffree and Rees, 2003).

In Baie and Morts peatlands, the macrofossil data suggest that the early coastal vegetation communities were rapidly isolated from the influence of marine waters by isotactic land uplift and change in relative sea-level. On the Manicouagan delta (BC), peat inception was synchronous in Baie and Manic peatlands which are located at the upper limit of the Laurentian transgression (Table 1.1; Fig. 1.9) suggesting that peat accumulation started rapidly after land emergence around 4200-4100 cal BP. In Morts (HSP) peat accumulation started approximately 1000 years later following the post-transgression conditions (*c.* 3240 cal BP). This could indicate that marine waters have persisted longer at this site or that the regional climate was not propitious to peat accumulation prior to that period.

1.4.2. Early vegetation successions and pathways to ombrotrophy

Previous studies showed that the geomorphic setting (substratum and basin topography), which controls the initial moisture and nutrient supply, influence the early developmental pathways in peatlands (Bauer, Gignac and Vitt, 2003 ; Bhiry, Payette and Robert, 2007 ; Ireland *et al.*, 2013 ; van Bellen *et al.*, 2011). In our study, the regional pattern towards ombrotrophy was similar in the peatlands over silt-clay and above deltaic sands which suggests that the geomorphic setting had a limited influence on the early vegetation successions. The six studied peatlands have relatively flat underlying topography which has likely favored a fast drainage after the withdrawal of marine waters and allowed a rapid development of poor minerotrophic plant communities.

Our macrofossil data show a regional difference in the rates of the trophic transitions (Fig. 1.9). The early development of ombrotrophic communities on the Manicouagan delta was promoted by very high rates of peat accumulation during the early peatland stage and suggests that the regional climate was very conducive to *Sphagnum* growth around 5800 and 4000 cal BP. In the peatlands of Havre-St-Pierre, the ombrotrophication seems to have been delayed by lower primary production within the initial fens, an indication that the regional climate was less propitious to *Sphagnum* growth. The succession from fen to bog in peatlands has previously been considered as a purely autogenic transition associated with peat accumulation (Payette, 1988 ; Yu *et al.*, 2003). However, the rate of this trophic transition may have been favored by a change in the atmospheric moisture balance as suggested by Hughes and Barber (2003). Along the Atlantic coast in New-Brunswick, Robichaud and Bégin (2009) attributed the ombrotrophication of large portions of a raised bog to a shift towards wetter/cooler climate after 5500 years BP. Further studies will be needed to evaluate the potential influence of regional climate change on this trophic transition at the scale of the peatland ecosystems.

1.4.3. Vegetation successions and fire dynamics within the bogs

Macrofossil data show that the bog systems were largely dominated by *Sphagnum* within the two studied regions (Fig. 1.9). The climate of the St. Lawrence North Shore characterised by low evapotranspiration and high moisture inputs seems to have been favorable to a sustained *Sphagnum* growth. The macroscopic charcoal data showed that the fire-free intervals were long within the bog stages in Lebel and Plaine (1160 and 1640 years respectively). These results support the observations from previous paleoecological studies and suggest that fires rarely occur in maritime ombrotrophic peatlands of eastern Canada (Lavoie, Filion and Robert, 2009 ; Robichaud and Bégin, 2009 ; Tolonen, Huttunen and Jungner, 1985). Low fire incidence in these bogs may be due to the scarcity of the woody cover and the prevailing wet conditions.

Overall, fires had few long-term impacts on the vegetation communities as previously observed in continental peatlands (Kuhry, 1994 ; Magnan, Lavoie and Payette, 2012). Although it is possible that some fire-induced shifts in vegetation have been too short-lived to be captured by the temporal resolution of the plant macrofossil analyses. Our data show that only the most important charcoal peaks with many large particles >2 mm coincided with significant changes in the vegetation cover. This suggests that most fires recorded by the charcoal data occurred nearby without directly affecting the surface vegetation at the coring location.

In Plaine bog, the fire event that occurred early after the ombrotrophication *c.* 5740 cal BP has eliminated a dense cover of *P. mariana*. This fire horizon containing many large charcoal fragments >2 mm coincides with an important period of fire recorded in a nearby paludified forest (Payette *et al.*, 2013). This suggests that the climate was drier and more propitious to fire during this period. The fire recorded around 4200 cal BP in Plaine bog is associated with an important slowdown in peat accumulation and a shift in vegetation which suggests that fire directly burnt the peat deposit. In Lebel, only the oldest fire dated around 2410 cal BP had a noticeable impact on the peat-forming vegetation. This fire which significantly affected the *Sphagnum* cover occurred during a dry period recorded in Baie and Lebel bogs in Baie-Comeau (Table 1.3), an indication that fire propagation on the peatland was promoted by drier climatic conditions.

1.5. Conclusion

Along the north shore of the St. Lawrence Estuary and Gulf, the timing and initial context of peat inception was closely linked with the Holocene variations in relative sea-level. The studied peatlands have developed over the deltaic terraces emerged following the withdrawal of the Goldthwait Sea from 7.5 ka BP and above silt-clay deposits after the Laurentian transgression from 4.3 ka BP. The long-term vegetation successions in these peatlands were influenced by a combination of internal processes and external climate forcing. The influence of the geomorphic setting on the early developmental pathways seems to have been relatively limited. The flat deltaic plains were rapidly drained after marine withdrawal and intermediate- to poor fens rapidly developed. The regional contrasts in the rates of vegetation successions are associated with distinct dynamics of peat accumulation. The faster transitions to ombrotrophy in the peatlands of Baie-Comeau seem to have been promoted by higher rates of peat accumulation, an indication that the regional climate was very conducive to *Sphagnum* growth in this region especially after the Laurentian marine transgression (i.e. ~4200-4000 cal BP). Also, the occurrence of fire was low and most events had limited impacts on the vegetation successions. Only the most important charcoal peaks coincided with significant changes in the vegetation cover.

This study provides further insights into the development of coastal maritime peatlands notably on potential influences of Holocene climate on long-term developmental pathways. However, the ecohydrological response to climate forcing and fires has likely varied spatially and temporally within the peatland basins and this dynamic is not fully covered by our approach using a single peat core per site. Further research is needed to evaluate the spatial dynamics of peatland development at the scale of the ecosystems to better understand the climatic influence on the contrasted development of maritime bogs from the boreal to the subarctic ecoclimatic domain.

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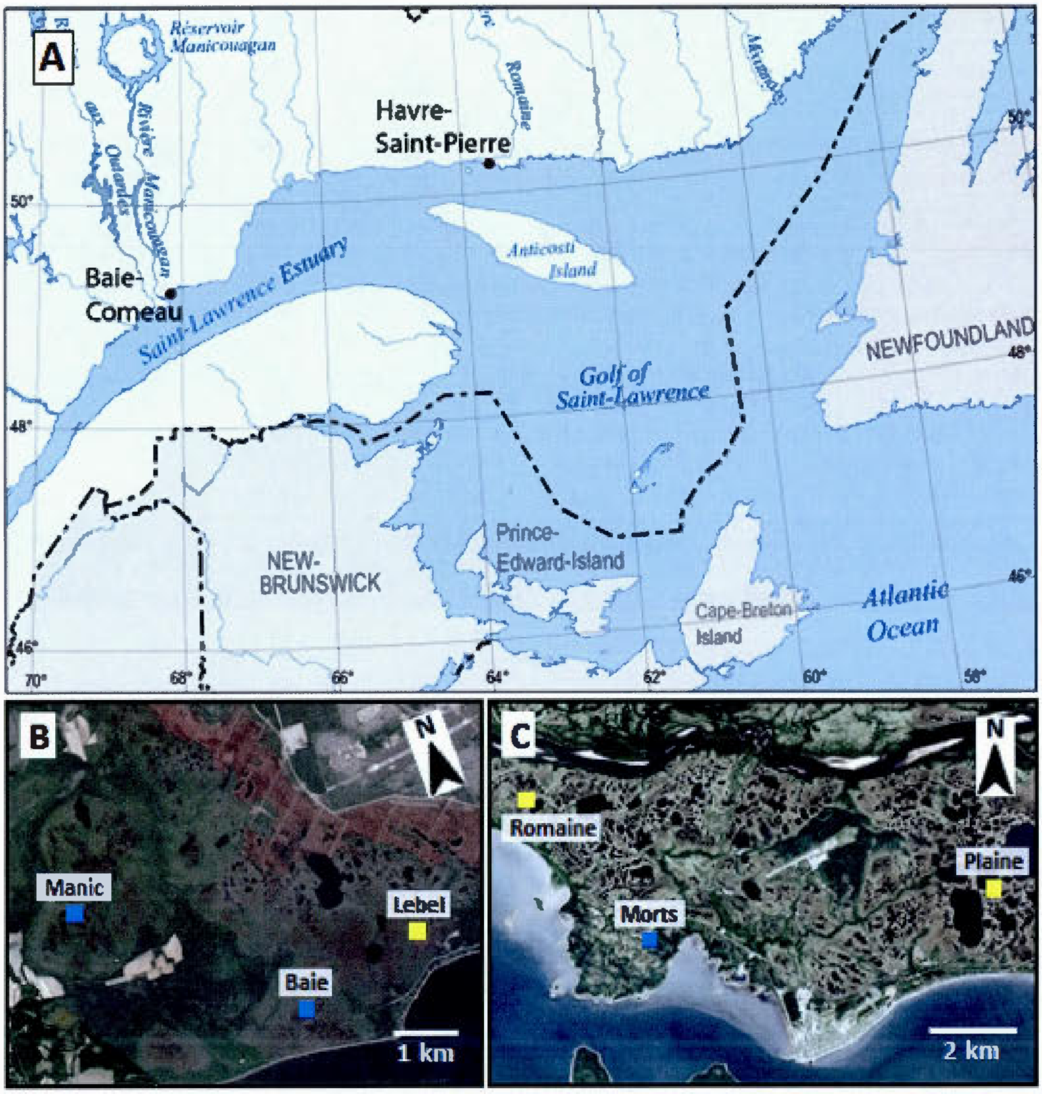

Figure 1.1. A) Location of the two studied regions in eastern Canada and aerial views of the peatlands investigated on B) the Manicouagan delta in the St. Lawrence Estuary and C) the La Romaine delta in the Gulf of St. Lawrence. Yellow squares : sites on deltaic sands, Blue squares : sites on marine silt-clay. Satellite images from Google Earth 2013.

Figure 1.2. Photographs of Lebel bog on the Manicouagan delta (left) and of Plaine bog on the La Romaine delta (right).

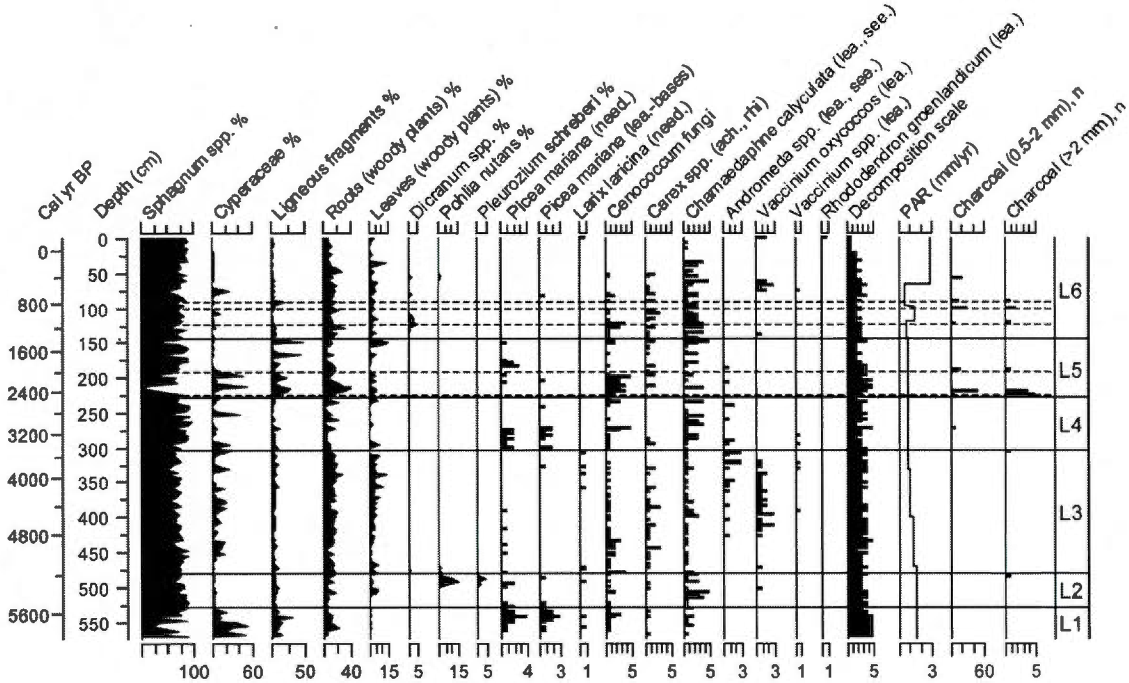

Figure 1.3. Plant macrofossil diagram of Lebel bog. Dashed lines show the fire events. The bars represent the scale of abundance (1= rare, 5= abundant). *need*–needles, *ach*–achenes, rhiz–rhizomes, lea–leaves, see–seeds, PAR–Peat accumulation rates.

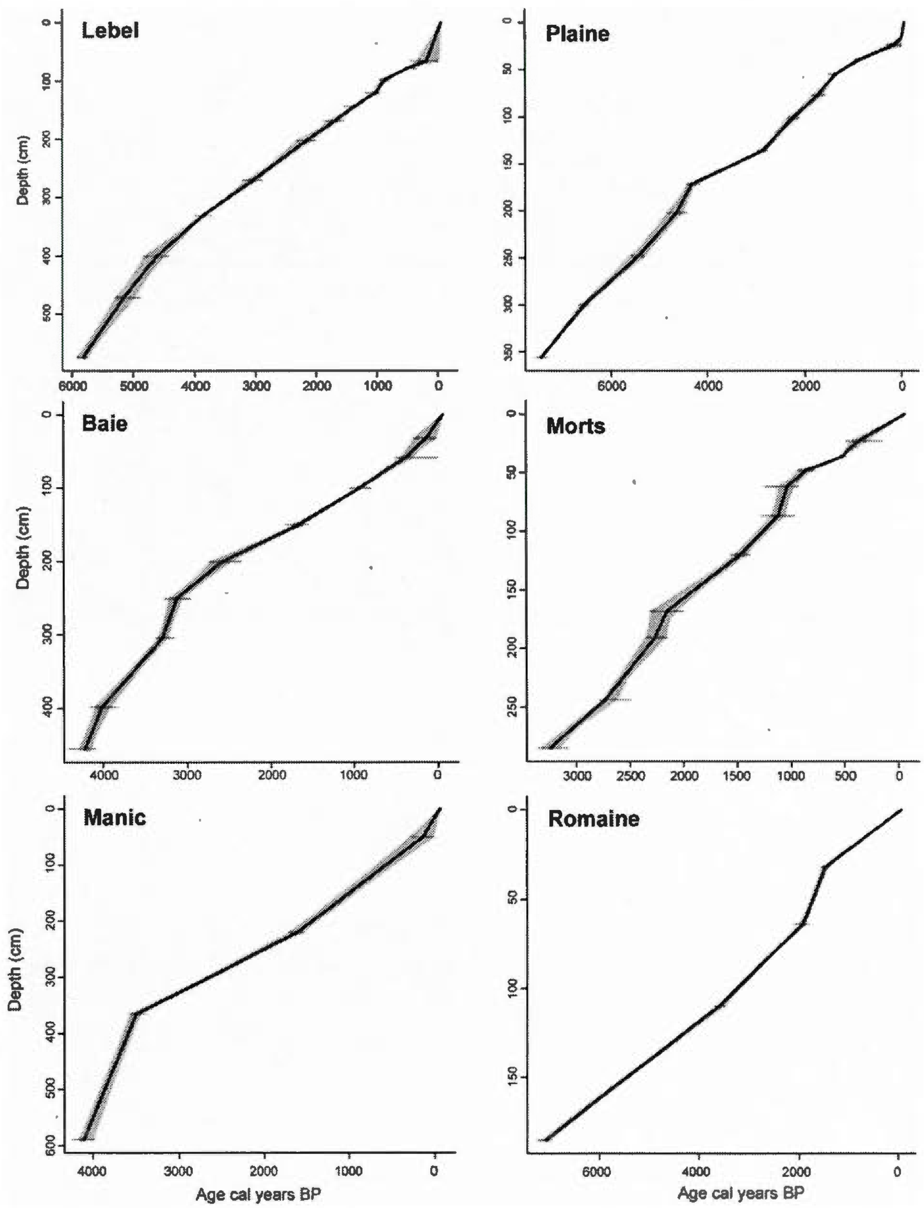

Figure 1.4. Age-depth models developed using Clam software with the 95% confidence intervals for the six studied peatlands.

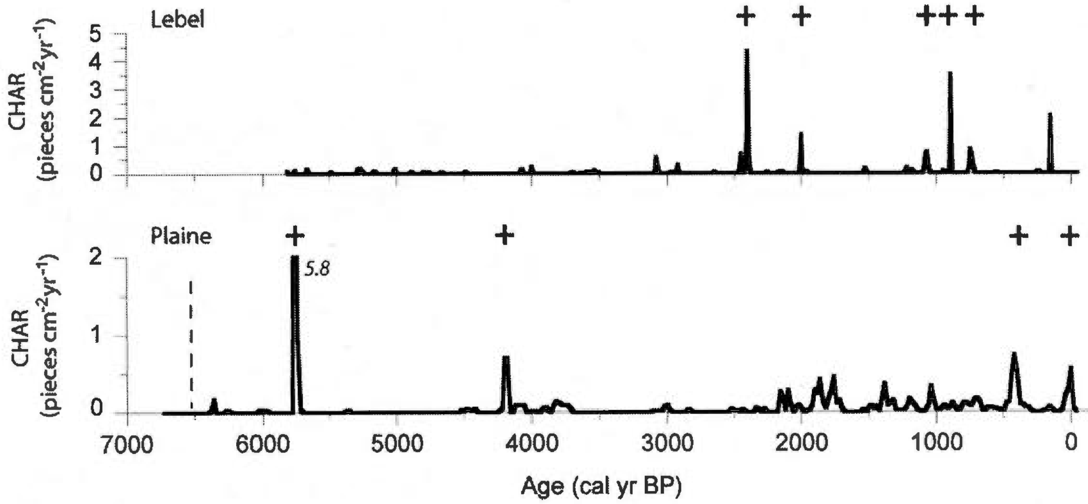

Figure 1.5. Charcoal accumulation rates (CHAR). The "+" symbol shows the identified fire horizons comprising large charcoal particles >2 mm. The vertical dashed line show the fen-bog transition in Plaine.

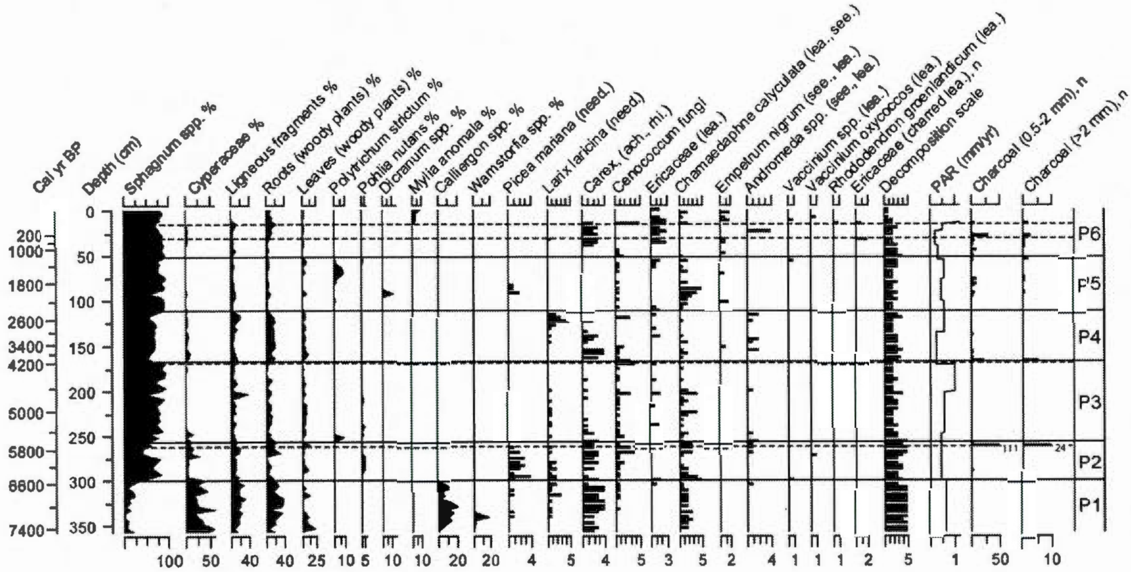

Figure 1.6. Plant macrofossil diagram of Plaine bog. Dashed lines show the fire events. The bars represent the scale of abundance (1= rare, 5= abundant). *need*–needles, *ach*–achenes, rhiz–rhizomes, lea–leaves, see–seeds, PAR–Peat accumulation rates.

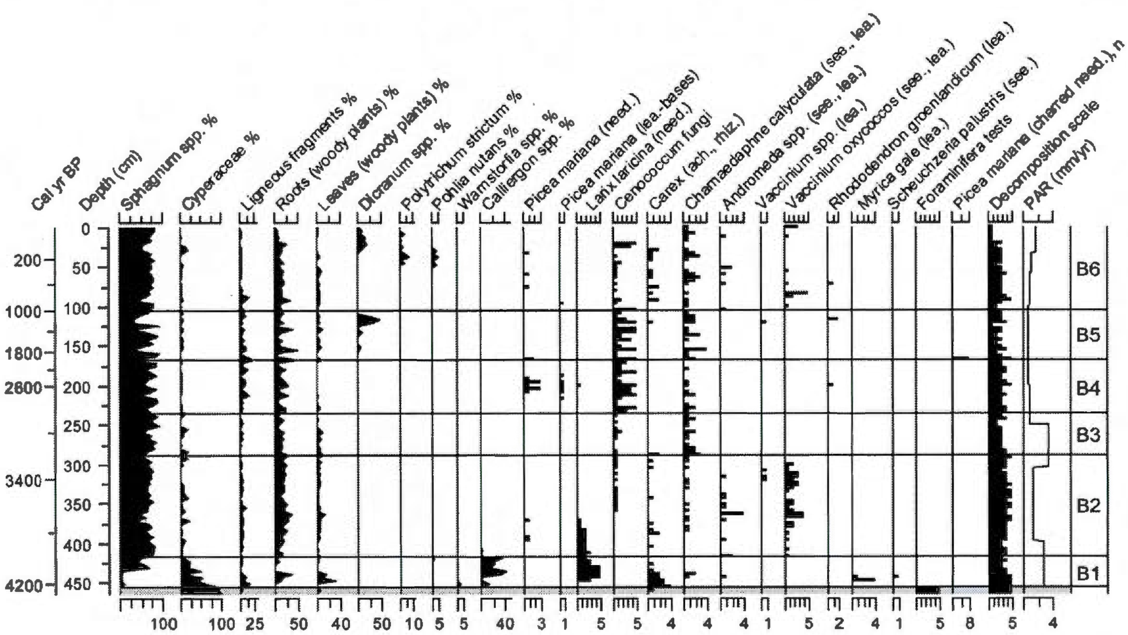

Figure 1.7. Plant macrofossil diagram of Baie bog. The bars represent the scale of abundance (1= rare, 5= abundant). *need*–needles, *ach*–achenes, rhiz–rhizomes, lea–leaves, see–seeds, PAR–Peat accumulation rates. The grey rectangle shows the lower peat section with high mineral content (>50%).

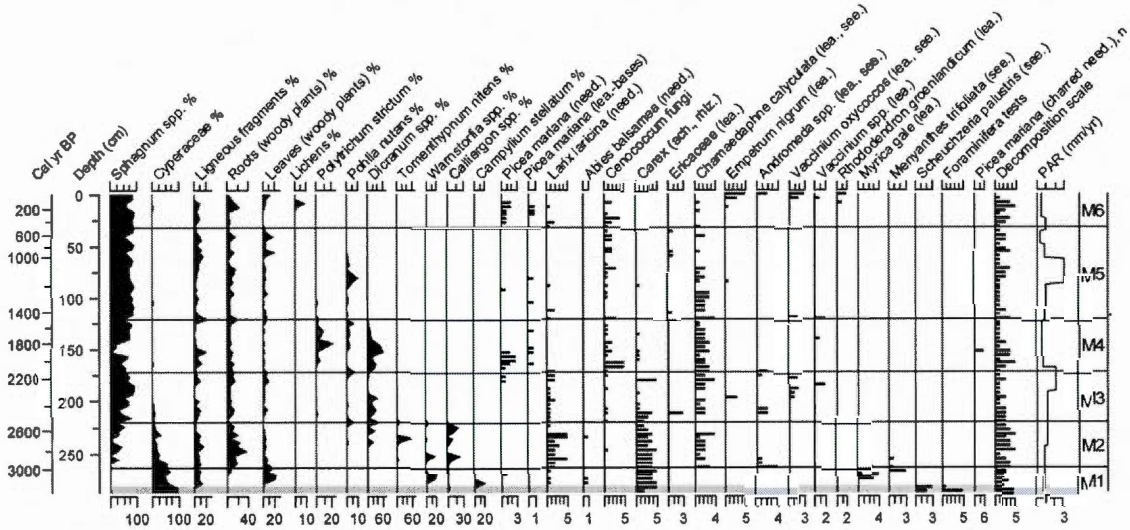

Figure 1.8. Plant macrofossil diagram of Morts peatland. The bars represent the scale of abundance (1= rare, 5= abundant). *need*–needles, *ach*–achenes, rhiz–rhizomes, lea–leaves, see–seeds, PAR–Peat accumulation rates. The grey rectangle shows the lower peat section with high mineral content (>50%).

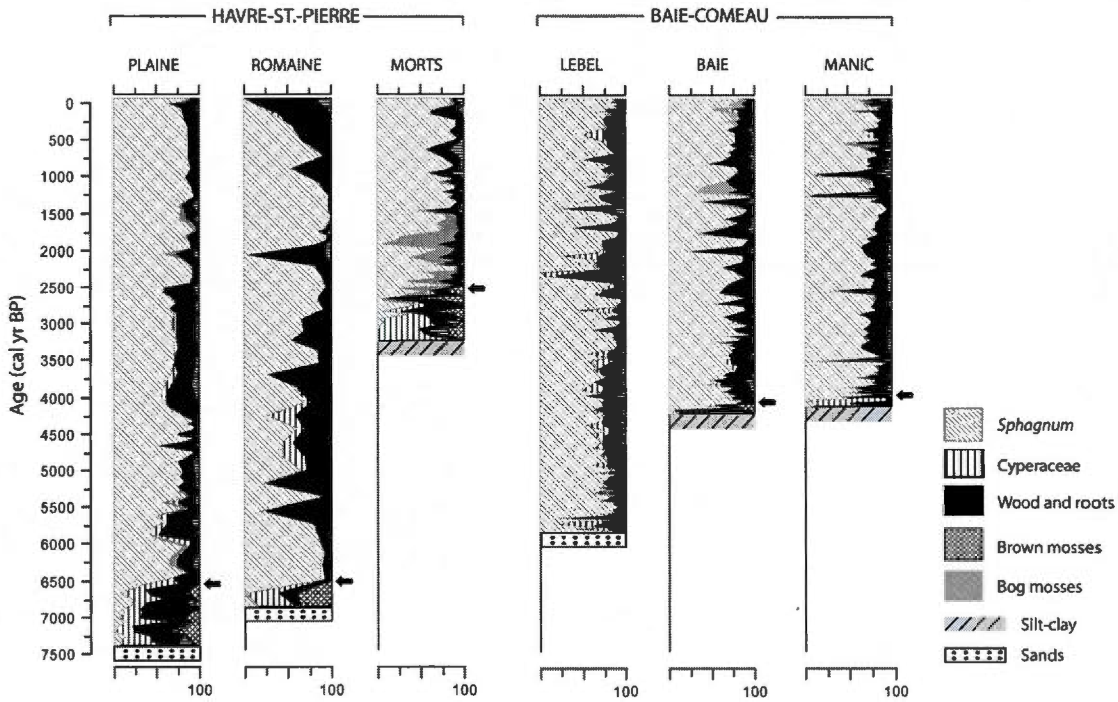

Figure 1.9. Synthesis of the peatland vegetation types for the six studied peat cores. The fen-bog transition is shown by the black arrows.

Study region	Sites	Lat. (N)	Long. (W)	Alt. peat surface (m.a.s.l.)	Alt.basal sediment (m.a.s.l.)	Basal sediment	Peat depth (cm)	Basal age (cal yr BP)
Baie-	Lebel	49° 5.9'	68° 13.3'	22.4	16.7	Sand	575	5820
Comeau	Baie	49° 5.8'	68° 15'	16.8	12.2	Silt-Clay	461	4210
	Manic	49° 7.1'	68° 18.2'	22.2	16.3	Silt-Clay	589	4100
Havre-	Plaine	50° 16.5'	63° 32.3'	34.2	30.6	Sand	356	7450
St-Pierre	Romaine	50° 17.7'	63° 42.9'	24.2	22.3	Sand	187	7080
	Morts	50° 15.8'	63° 40.1'	14.5	11.7	Silt-Clay	285	3240

Table 1.1. Characteristics of the studied peatlands.

Site	Sample depth (cm)	Lab. no. (UCI-AMS)	Material dated[1]	^{14}C age (yr BP)	2 σ range (cal. BP)	Cal. BP[2]
PLAINE	16-17	67838	Sph.	Modern	0	0
	24-25	79504	Sph.	150 ±15	6-281	150
	39-40	73854	Sph.	935 ±25	793-919	850
	55-56	98873	Sph.	1480 ±15	1328-1399	1360
	77-78	73855	Sph.	1800 ±25	1629-1818	1740
	101-102	73856	Sph.	2245 ±25	2157-2338	2240
	136-137	98874	Sph.	2765 ±15	2793-2921	2850
	172-173	80191	Sph.	3875 ±15	4244-4407	4320
	202-203	73857	Sph., Eric. seeds	4100 ±25	4523-4808	4630
	248-249	79505	Sph., Eric. seeds	4715 ±15	5327-5577	5430
	300-301	67839	Sph.	5755 ±15	6495-6632	6560
	356-357	67840	*Carex* ach.	6540 ±15	7426-7475	7450
MORTS	23-24	98870	Sph.	290 ±15	300-428	370
	36-37	79502	Sph.	480 ±15	506-531	520
	48-49	98871	Sph.	955 ±15	797-926	860
	62-63	73858	Sph.	1145 ±25	976-1168	1030
	87-88	73859	Sph.	1190 ±25	1014-1178	1120
	120-121	80188	Sph., Eric. lea.	1580 ±15	1415-1521	1470
	168-169	79503	Sph., Eric. seeds	2150 ±15	2066-2299	2150
	191-192	98872	Sph., Eric. lea.; *Picea* need.	2250 ±15	2162-2338	2260
	244-245	67844	Sph., Eric. lea.; *Larix* need.	2585 ±20	2625-2754	2730
	284-286	67843	*Carex* ach.	3025 ±15	3166-3327	3240
ROMAINE	32-33	79506	Sph.	1595 ±15	1416-1529	1470
	64-65	73860	Sph.	1975 ±25	1878-1988	1920
	110-111	79507	Sph.	3350 ±15	3490-3639	3590
	185-187	67845	*Carex* ach.	6200 ±15	7016-7168	7080
LEBEL	66-67	73861	Sph.	210 ±25	(-4)-303	180
	79-80	80190	Sph., Eric. lea.; *Carex* ach.	430 ±15	484-514	500
	98-99	79499	Sph., Eric. lea./seeds	965 ±15	798-929	870
	120-121	98875	Sph.	1125 ±15	979-1061	1020
	143-144	73862	Eric. lea.	1475 ±25	1311-1400	1360
	168-169	73863	Sph., Eric lea.	1775 ±25	1613-1810	1680
	202-204	98876	Sph	2145 ±15	2063-2297	2150
	270-271	79500	Sph. ; Eric. seeds	2905 ±15	2966-3139	3040
	331-332	79501	Sph.	3570 ±15	3832-3910	3870
	400-401	73864	Sph.; Eric lea.	4090 ±25	4452-4805	4620
	471-472	98877	Sph.	4465 ±15	4978-5277	5160
	574-575	67837	Sph.	5090 ±15	5753-5908	5820
BAIE	32-33	98867	Sph.	105 ±15	28-259	130
	58-59	73865	Sph. Eric. lea.	300 ±25	300-455	380
	100-101	98868	Eric. lea.	1000 ±15	840-958	920
	150-151	79496	Sph.	1755 ±15	1614-1712	1660
	200-202	98869	Sph. Eric. lea.	2495 ±15	2491-2715	2590

	250-251	79497	Sph.	2950 ±15	3065-3209	3130
	304-305	80196	Sph. Eric. seeds/lea.	3060 ±15	3219-3344	3290
	398-399	73866	Sph.	3670 ±25	3915-4086	4010
	455-456	79498	Sph., *Carex* ach.; Larix need.	3830 ±15	4153-4288	4210
MANIC	50-51	80192	Sph.	110 ±15	24-261	130
	220-221	80193	Sph.	1700 ±15	1548-1691	1610
	365-366	80194	Sph.	3265 ±15	3447-3556	3490
	588-590	80195	Sph.; *Carex* ach.	3745 ±15	4000-4153	4100

[1] Sph = *Sphagnum*, Eric = *Ericaceae*, lea.= leaves, ach. = achenes, need. = needles
[2] Calibrated ages inferred from Clam software (Blaauw, 2010).

Table 1.2. Radiocarbon dates.

Site	ZONE	Depth (cm)	Age (cal yr BP)	Main features of the zones	Mean WTD*
LEBEL	L6	144-0	1370-Present	*Sphagnum with Carex* spp. and Ericaceae	7.2 (3)
	L5	228-144	2500-1370	Woody plants with *P. mariana*	10.1(4)
	L4	304-228	3500-2500	*Sphagnum* with *P. mariana* and *C. calyculata*	8.4(2)
	L3	480-304	5220-3500	*Sphagnum* with Ericaceae	5.5(1)
	L2	528-480	5520-5220	*Sphagnum* with bog mosses and *C. calyculata*	14.4(5)
	L1	574-528	5820-5520	*Sphagnum* and *Cyperaceae* with *P. mariana*	24(5)
BAIE	B6	104-0	980-Present	*Sphagnum* with Ericaceae and bog mosses	6.5(3)
	B5	168-104	1990-980	*Sphagnum* with *Dicranum* spp.	9.3(5)
	B4	236-168	2980-1990	*Sphagnum* with *P. mariana*	10.5(7)
	B3	288-236	3240-2980	*Sphagnum* with *C. calyculata*	7.2(1)
	B2	416-288	4080-3240	*Sphagnum* with Ericaceae	6(4)
	B1	456-416	4215-4080	Poor fen with *Calliergon* spp. and *L. Laricina*	6.3(2)
PLAINE	P6	52-0	1270-Present	*Sphagnum* with *Carex* spp. and Ericaceae	11(6)
	P5	112-52	2430-1270	Well-preserved *Sphagnum*	17.7(7)
	P4	168-112	4160-2430	*Sphagnum* with *Carex* spp. and *L. Laricina*	6.0(5)
	P3	256-168	5610-4160	Well-preserved *Sphagnum*	17.6(8)
	P2	300-256	6560-5610	*Sphagnum* with *P. mariana*	7.6(4)
	P1	357-300	7450-6560	*Carex*-fen with *L. laricina* and *Calliergon* spp.	-1.5(4)
MORTS	M6	32-0	470-Present	*Sphagnum* with *P. mariana* and *L. laricina*	17.9(6)
	M5	120-32	1470-470	Well-preserved *Sphagnum*	12(6)
	M4	172-120	2170-1470	Bog mosses with *P. mariana*	20.7(8)
	M3	220-172	2520-2170	*Sphagnum* with *L. laricina* and Ericaceae	16.1(7)
	M2	264-220	2980-2520	Poor fen with brown mosses and *L. laricina*	-
	M1	284-264	3240-2980	Coastal *Carex*-fen with brown mosses	-

* Mean water table depths with standard deviation presented in Chapter 2.

Table 1.3. Macrofossil diagram zonation for the four main peat profiles.

CHAPITRE II

EVALUATING LONG-TERM REGIONAL CLIMATE VARIABILITY IN THE MARITIME REGION OF THE ST. LAWRENCE NORTH SHORE (EASTERN CANADA) USING A MULTI-SITE COMPARISON OF PEAT-BASED PALEOHYDROLOGICAL RECORDS

Gabriel Magnan[1], Michelle, Garneau[1]

[1]GEOTOP - Université du Québec à Montréal (UQAM), C.P. 8888 Succursale Centre-ville, Montréal, Québec, Canada H3C 3P8.

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Résumé

Cette étude présente les premières reconstitutions paléohydrologiques dans les tourbières ombrotrophes (bogs) de la côte nord de l'estuaire et du golfe du St. Laurent dans l'est du Canada. Les fluctuations à long terme des nappes phréatiques ont été reconstituées au sein de quatre tourbières ombrotrophes (bogs) provenant de deux régions écoclimatiques maritimes (boréal et subarctique) par l'analyse des thécamoebiens fossiles en utilisant une fonction de transfert adaptée aux tourbières nordiques de l'est du Canada. La comparaison des reconstitutions paléohydrologiques entre ces tourbières a été utilisée pour distinguer l'influence des changements climatiques par rapport aux variations liées aux processus internes. La cohérence des changements hydrologiques au sein de chaque région suggère une influence climatique commune sur la paléohydrologie des bogs. Toutefois, les différences marquées à l'échelle régionale impliquent aussi une influence d'autres facteurs non-climatiques (e.g. processus internes des tourbières et rétroactions écohydrologiques). L'assèchement des surfaces à partir de 3000 cal BP coïncide avec la transition de l'optimum climatique holocène vers le Néoglaciaire documentée par plusieurs proxies climatiques au nord-est du Canada. Les bogs de Havre-St-Pierre ont connus des abaissements majeurs des nappes phréatiques au cours des 2500 dernières années qui suggèrent d'importants déficits hydrologiques possiblement liés au refroidissement Néoglaciaire. Les différences dans l'amplitude des fluctuations hydrologiques entre les deux régions peuvent résulter de conditions climatiques distinctes ou peuvent indiquer une sensibilité accrue de l'hydrologie de surface des bogs de Havre-St-Pierre aux variations climatiques passées.

Abstract

This study presents the first paleohydrological reconstructions from peat bogs of the north shore of the Estuary and Gulf of St. Lawrence in eastern Canada. Past water table depths were reconstructed from analyses of fossil testate amoebae within four ombrotrophic peatlands (bogs) of two maritime ecoclimatic regions (boreal and subarctic) using a transfer function adapted to the northern peatlands of eastern Canada. The multi-site comparison of the water table records was used to distinguish the climate-driven changes from the variations related to site-specific internal peatland processes. The coherence between the water table reconstructions suggests a common climate forcing on the bog paleohydrology but the marked differences at the regional scale also imply an influence of non-climatic factors (e.g. internal peatland processes and ecohydrological feedbacks). The surface drying after 3000 cal BP coincides with the transition from the Holocene Climatic Optimum to the Neoglacial cooling documented by many proxy climate records in northeastern Canada. The bogs of Havre-St-Pierre have experienced major water table drawdown during the late-Holocene indicating important annual-to-centennial water deficits possibly linked with the Neoglacial cooling. The regional difference in the magnitude of the hydrological fluctuations may result from distinctive regional climate conditions or could indicate that the surface wetness in the bogs of Havre-St-Pierre was more sensitive to past climate changes.

2.1. Introduction

Ombrotrophic peatlands (bogs) which receive their moisture inputs solely from precipitation have been widely used as a source of proxy-climate data. Testate amoebae (single-celled protists) are robust indicators of hydrological conditions in peat bogs and are commonly used for quantitative reconstruction of water table depths in these ecosystems (Mitchell, Charman and Warner, 2008). The atmospheric water balance was often considered as the main driver of surface-wetness in ombrotrophic peatlands (Charman *et al.*, 2009). In northwestern Europe, long-term variations in bog surface wetness (BSW) have been interpreted as changing regional climate more specifically in terms of variations in effective precipitations (precipitation minus evapotranspiration) (Barber, Chambers and Maddy, 2003 ; Charman and Hendon, 2000). These interpretations are supported by the agreement in the patterns of water table changes within and between sites (Charman, 2006 ; Charman, Hendon and Packman, 1999 ; Hendon, Charman and Kent, 2001 ; Langdon *et al.*, 2012) and linkages with independent proxy climate data (Barber and Langdon, 2007). However, at the scale of the peatland, long-term changes in surface hydrology are also driven by internal processes and ecohydrological feedbacks which complicates the detection of external forcing (Belyea and Baird, 2006 ; Swindles *et al.*, 2012). This can be particularly problematic when using a single peat core.

Over the last decade, peatland paleohydrological research has been expanding in the boreal and subarctic regions of northeastern Canada (Bunbury, Finkelstein and Bollmann, 2012 ; Lamarre, Garneau and Asnong, 2012 ; Loisel and Garneau, 2010 ; van Bellen *et al.*, 2013 ; van Bellen, Garneau and Booth, 2011). These studies have linked long-term trends in bog surface wetness with documented climate changes such as the Neoglacial cooling and the Little Ice Age (LIA). A multi-proxy study conducted in a bog from eastern Newfoundland suggested the existence of persistent links between bog surface wetness variability and oceanic and solar forcing over the Holocene (Hughes *et al.*, 2006). Apart from this study, few peat-based paleohydrological reconstructions have been conducted in the maritime regions of eastern Canada. Little is known about the Holocene climate variability in these regions and its impact on the hydrology of peatland ecosystems. Ombrotrophic peatlands are widespread in the coastal plains along the north shore of the Estuary and Gulf of St. Lawrence. These peat sequences offer great potential for high-resolution paleohydroclimatic reconstructions since their stratigraphy is often entirely composed of fast-accumulating *Sphagnum* (Glaser and Janssens, 1986).

In the present study we reconstruct the Holocene paleohydrological variability from the coastal ombrotrophic peatlands of the St. Lawrence North Shore. The main objective is to detect regional paleohydroclimate changes using peat-based paleohydrological reconstructions. We compared the water table records from four distinct peat bogs as a means to disentangle the external climatic forcing from the site-specific variations associated with internal peatland processes.

2.2. Study region

We investigated four ombrotrophic peatlands from two ecoclimatic regions along the North Shore of the Estuary and Gulf of St. Lawrence in eastern Canada. Sites are located on the postglacial deltaic formations of Manicouagan and La Romaine. Peat cores were taken from the ombrotrophic peatlands Lebel (49°6'N, 68°13'W) and Baie (49°6'N, 68°15'W) near Baie-Comeau and from Plaine (50°16'N,63°3'W) and Morts (50°16'N, 63°40'W) near Havre-St-Pierre (HSP) (Fig. 2.1). Lebel and Baie are *Sphagnum*-dominated raised bogs with low abundance of pools whereas Plaine and Morts are extensive plateau bogs with ponds and lakes of all sizes.

In each region, we compared two peatlands that developed in different periods of the Holocene within two different geomorphic contexts. This approach aims to evaluate whether long term hydrological trends such as surface drying have resulted from gradual change in surface structure during peatland development. Lebel and Plaine peatlands formed on the sandy deltaic terrace that emerged after the withdrawal of the Goldthwait Sea during the early Holocene whereas Baie and Morts peatlands developed on silt-clay deposits associated with a mid-Holocene marine transgression (Bernatchez, 2003). The climate along the St. Lawrence North Shore is maritime with mean annual precipitations of approximately 1000 mm in both regions and mean annual temperature of 1.5° C in Baie-Comeau and 1.1°C near Havre-St-Pierre (Environment Canada, 2013). The length of the growing season decreases between the two regions. Mean value for growing degree-days (>0°C) is ~1924 in BC and ~1779 in HSP (climate record 1971-2003 from National Land and Water Information Service; Hutchinson *et al.*, 2009).

Between the Estuary and the Gulf of St. Lawrence, there is an important transition in peatland surface characteristics revealing distinct regional ecoclimatic conditions. In the peatlands, the density of shrubs and trees decreases gradually eastward from Baie-Comeau to Havre-St-Pierre and the area covered by pools is expanding significantly (Fig. 2.1). On the Manicouagan delta, bog surfaces are largely covered by *Sphagnum* spp., heath shrubs and dense krummholz of black spruces (*Picea mariana*). The plateau bogs of La Romaine delta are largely treeless and covered by lichen-dwarf shrub carpets interspersed with patches of *Sphagnum*.

2.3. Methods

2.3.1. Field sampling

Peatlands were selected according to their regional representativeness based on analyses of aerial photographs and field validations (e.g. vegetation inventories and peat thickness measurements). In order to minimise the topographic influence on peatland paleohydrology (*sensu* Belyea and Clymo, 2001), we selected sites with relatively flat basin identified with manual probing and peat cores were collected from the central domed sections. The upper first meter of peat was sampled with a Box corer (105×8×8 cm) and a Russian peat corer (diameter = 7.5 cm; length = 100 cm) was used for deeper peat sections. In the field, peat cores were wrapped in plastic and aluminum foil, transferred into PVC tubes and stored in the fridge at 4°C until analysis.

2.3.2. Chronology

A total of 43 samples of plant remains were radiocarbon (^{14}C) dated by accelerator mass spectrometry (AMS) at Keck laboratory (University of California, Irvine) (Table 2.1). The age-depth models were developed with Clam software version 1.7.2 (Blaauw, 2010) using linear interpolation between each ^{14}C date (Fig. 2.2). Radiocarbon dates were calibrated using the IntCal09 calibration curve (Reimer *et al.*, 2009). The age of the peatland surface was fixed at −60 cal yr BP (i.e. AD 2010 ; year of coring).

2.3.3. Testate amoebae analyses

Testate amoebae analyses were conducted within the four peat sequences at intervals varying between 2 and 4 cm in order to optimise the temporal resolution by using lower intervals in sections with higher deposition time (i.e. low rates of peat accumulation). The analyses were mainly restricted to the ombrotrophic peat sections which provide more sensitive archive of past hydroclimate changes (Charman *et al.*, 2009) but also cover the fen-bog transition. The identification of the fen-bog transition is based on the plant macrofossil data presented in Chapter 1. The main peat types (*Sphagnum*, bog mosses, ligneous remains, Cyperaceae and brown mosses) are displayed in the testate amoebae diagrams.

In the laboratory, 1-cm^{-3} peat samples were collected and treated following standard extraction procedures (Booth, Lamentowicz and Charman, 2010). Two tablets of *Lycopodium* were added as exotic markers to calculate test concentrations. The size fraction between 300 and 15 µm was analysed. Tests were identified and counted under the optical microscope at a 400x magnification using the taxonomy of Charman, Hendon and Woodland (2000) with the modifications from Booth and Sullivan (2007). A minimum of 100 tests was counted for most samples, however for some samples with very low concentrations, a minimum of 75 counts was considered reliable following Payne and Mitchell (2009) who stated that 50 specimens are adequate for some samples and 100 counts are sufficient in most cases. Fossil assemblages with at least 50 individuals are shown in the testate amoebae diagrams but were not included in the water table reconstructions. The relative abundance of each species was calculated as a percentage of the total. The rotifer *Habrotrocha angusticolis* was excluded from the total counts and is not included in the WTD reconstructions. Testate amoebae diagrams were constructed using C2 software version 1.7.2 (Juggins, 2011).

2.3.4. Past water table depth reconstructions

Paleohydrological reconstructions were carried out with C2 software using a new testate-amoebae transfer function based on 206 modern assemblages including 120 samples from the St. Lawrence North Shore (Lamarre *et al.*, 2013). A weighted average tolerance downweighted (WA-tol) model with classical deshrinking was used to infer the past water table depths (WTD). This model has the best performance statistics in cross-validation for the prediction of WTD (r^2 = 0.82; RMSEP = 5.65 cm). Also, the reconstructions based on the WA-tol model were shown to be statistically significant when compared with randomly generated WTD data (Telford and Birks, 2011). Sample-specific standard errors for the reconstructed WTD were calculated using bootstrapping (1000 cycles).

2.3.5. Ordination with DCA

Detrended correspondence analysis (DCA) was applied on the entire fossil data set (four records combined) to explore the testate amoebae associations by excluding species with less than three occurrences. Species with low occurrence have also been downweighted to minimise their impact on the ordination. The Axis 1 scores of the fossil samples derived from the DCA of the testate amoebae data were used along with the inferred WTD values to define the diagram zones.

2.4. Results

2.4.1. DCA ordination of the fossil assemblages

Detrended correspondence analysis (DCA) of the fossil testate amoebae assemblages (Fig. 2.3) shows a good separation of the species along the main axis of variability (Axis 1). The ordination along Axis 1 reveals a coherent surface-wetness gradient with dry-indicating species on the right (e.g. *Trigonopyxis minuta*, *Trigonopyxis arcula*) (higher values) and more hygrophilous species on the left (e.g. *Amphitrema wrightianum*, *Difflugia globulosa*) (lower values). *Arcella discoides* has an affinity with dry-indicating taxa in fossil data which contrasts with its wet optima in modern samples (Lamarre *et al.* 2013). This species is often found with dry-indicating species such as *Hyalosphenia subflava* in fossil samples corresponding to periods of high decadal-to-centennial hydroclimate variability (Sullivan and Booth, 2011). *Difflugia pulex* is positioned at the extreme right on Axis 1 indicating a clear affinity with the driest taxa in fossil assemblages.

The position of most species in the DCA ordination of the fossil samples (Fig. 2.2) is coherent with their known ecology in the modern samples (Lamarre *et al.*, 2013). This suggests that the fossil species assemblages are clearly related with past surface-wetness conditions. The DCA sample scores are well correlated with the inferred WTD values (Spearman's rank coefficient 0.81, $p < 0.001$). The similarity of the species associations in both fossil and modern assemblages suggests that the composition of the fossil assemblages is not significantly affected by differential decay (Mitchell, Payne and Lamentowicz, 2008 ; Swindles and Roe, 2007).

2.4.2. Paleohydrological reconstruction of Lebel bog (BC)

Particularly dry conditions are inferred during the early peatland stage (mean WTD : 18.5 cm) by high abundance of *H. subflava* and *T. arcula* (Zone L1; 5680-5240 cal BP) (Fig. 2.4; Table 2.2). The water table levels gradually increased and very stable and wet conditions installed between 5240 and 3180 cal BP (dominance of *Amphitrema flavum*)(Zone L2). The bog surface became drier and water table levels were more variable from 3180 to 2490 cal BP (Zone L3). The co-dominance of two species on opposite ends of the wetness gradient (*A. discoides* and *H. subflava*) between 2490 and 2070 cal BP (Zone L4) suggests an important seasonal-to-decadal hydrological variability with significant periods of water table drawdown. After 2070 cal BP, a return to wet conditions (Zone L5 and L6) is indicated by the presence of *A. wrightianum* and high levels of *A. flavum*. The recent dry shift (<100 cal BP) is mainly related to high abundance of *Nebela militaris* (Zone L7) although its exact timing is unknown due to poor chronological control in the upper peat layers.

2.4.3. Paleohydrological reconstruction of Baie bog (BC)

From 4180 to 3250 cal BP (Zone B1; Fig. 2.5), the testate amoebae (mainly *A. flavum*, *Hyalosphenia papilio*) show relatively wet surface conditions except for a single dry shift around 3900 cal BP (*H. subflava* and *T. arcula*). A slight shift to lower intermediate WTD is recorded between 3250 to 2980 cal BP (Zone B2) and this is followed by a period of low and fluctuating water tables (Zone B3; 2980-1990 cal BP) with a co-dominance of *H. subflava* (dry taxa) and *A. discoides* (wet taxa) and sustained presence of *D. pulex* and *Assulina muscorum*. An important dry event is recorded around 2290 cal BP but wetter intermediate conditions returned after 1990 cal BP (Zones B4, B5 and B6) as shown by high levels of *A. flavum*. Very wet conditions are inferred around 720 cal BP by significant presence of *A. wrightianum* but the water table decreased gradually thereafter.

2.4.4. Paleohydrological reconstruction of Plaine bog (HSP)

Very wet conditions are inferred prior to 6560 cal BP by high abundance of *A. wrightianum* and correspond to the early fen with *Cyperaceae* and brown mosses (Zone P1) (Fig. 2.6). A slight water table drawdown followed the transition to ombrotrophic *Sphagnum*-dominated peat but relatively wet conditions persisted until 4800 cal BP (Zone P2), except for a dry shift at *c.* 5700-5400 cal BP (high levels of *H. subflava*). Particularly dry surfaces are recorded between 4800 and 4240 cal BP (Zone P3) with the occurrence of *A. muscorum*, *D. pulex* and *T. arcula*. This is followed by a sharp shift to wet conditions around 4240 cal BP (*A. wrightianum* and *A. discoides*) that persisted until 2640 cal BP (mean WTD: 5 cm) (Zone P4). From 2640 cal BP, a gradual water table drawdown is recorded (*A. muscorum*, *D. pulex*, *H. subflava*) with the lowest levels around 1900 and 1700 cal BP (~32 cm). These dry conditions persisted locally until 950 cal BP (Zone P5) followed by a wet shift (Zone P6) until the recent surface drying (Zone P7).

2.4.5 Paleohydrological reconstruction of Morts bog (HSP)

The testate amoebae assemblages reveal relatively wet surface conditions (high levels of *A. flavum*) prior to 2310 years cal BP (Zone M1; Fig. 2.7) especially before the fen-bog transition dated at 2520 cal BP. Very dry conditions installed between 2310 and 1470 (Zone M2) (*D. pulex* and *H. subflava*) except for a single wet shift (*A. discoides*) around 2040 cal BP. A significant rise in the water table levels after 1470 cal BP (Zone M3) was interrupted by a temporary dry event around 1150 cal BP followed by a sharp wet shift. Much drier conditions (*D. pulex*) are recorded between 520 and 240 cal BP (Zone M4) followed by intermediate water table levels in recent peat deposits (Zone M5).

2.4.6. Multi-site comparison of the water table records

Our data show similar regional patterns of water table changes for the comparable time periods (Fig. 2.8). In Lebel and Baie there is a synchronous trend towards increased water table depth variability starting from ~3000 cal BP with a particularly dry event around 2600-2200 cal BP followed by generally wet conditions after 2000 cal BP. In Plaine, the dry shift that started from 2640 cal BP coincides with a drying related to the fen-bog transition in Morts and dated around 2520 cal BP. Water tables were very low and highly variable in Morts and Plaine between 2300 and 1500 cal BP but there are also many discrepancies between these two records. For example the dry phase in Morts between 520 and 240 cal BP contrasts with the near-surface water tables in Plaine. Also, some short-term shifts in the water table levels such as the wet event *c.* 2040 in Morts are clearly site-specific.

The data also show coherent periods of water table changes between the bogs of the two regions such as the dry phase in Lebel and Plaine from ~5700 to ~5200 cal BP and the wet conditions in Lebel, Baie and Plaine between 4200 and 3200 cal BP and after 1000 cal BP. The dry shift in Plaine and the fen-bog transition in Morts *c.* 2500 cal BP coincide with the onset of the driest period in the two bogs of Baie-Comeau. However, there are also clear regional differences such as the dry event in Plaine between 4800 and 4240 cal BP when surface conditions are very wet in Lebel. Also, the late-Holocene dry shift started earlier in Lebel and Morts (*c.* 3000 cal BP) than in Plaine and Morts (*c.* 2650-2300 cal BP).

2.5. Discussion

2.5.1. Detecting climate forcing using peat-based paleohydrological records

The multi-site comparison of WTD records was used to distinguish the climatic influence on long-term peatland hydrology. The differences in the timing and magnitude of the paleohydrological variations between Baie-Comeau and Havre-St-Pierre may result from differences in the climatic conditions between the Estuary and the Gulf of St. Lawrence. Some discrepancies within and between sites may also be explained by changes in the sensitivity of the bog surface-wetness to external climate forcing associated with the gradual vertical peatland development. In Morts bog, the late-Holocene surface drying may have been delayed by the persistence of minerotrophic conditions and possible influence of water runoff at the site until around 2500 cal BP. Although testate amoebae communities are not directly influenced by vegetation composition (Woodland *et al.*, 1998), the hydrological variability may be influenced by changes in vegetation structure and density at the bog surface (Sullivan *et al.*, 2011). This influence was likely limited in our study since the four studied peat profiles are largely dominated by bog *Sphagnum* mosses.

Our data support the idea that testate amoebae are reliable indicators of past hydrological conditions in ombrotrophic peatlands (Mitchell, Payne and Lamentowicz, 2008). The similarity of species associations in both fossil (Fig. 2.2) and modern samples (Lamarre *et al.*, 2012) suggests that testate amoebae shells were not significantly affected by decay or other taphonomic processes which supports the findings of previous studies (Mitchell, Charman and Warner, 2008 ; Swindles and Roe, 2007).

Overall, the coherent periods of hydrological changes within each region provide evidence for a common climate forcing. However, the results also suggest that peatland water tables have not responded consistently to climatic changes. The inconsistencies between the water table reconstructions suggest that the bog hydrology was also driven by internal peatland processes (e.g. peat build-up). These results are consistent with conceptual models and suggest that peatlands are complex adaptative systems (Belyea and Baird, 2006 ; Morris, Baird and Belyea, 2012). Peatland water tables are continuously responding to a combination of internal processes and external climate forcing. While some water table shifts may have been climatically-driven, some others may have resulted from changes in peat accumulation dynamics (Belyea and Clymo, 2001). Peatland water tables may be partially disconnected from external influences due to ecohydrological feedback mechanisms (Swindles *et al.*, 2012) and only significant climate changes may thus be recorded. Yu *et al.*, (2003) suggested that long-term surface drying in peatlands may results primarily from autogenic vertical development associated with peat growth. In our study, this potential effect was tested by comparing two peatlands of significantly different ages within each region. The synchronicity of the late-Holocene surface drying within each region suggested an external climate influence.

2.5.2. Comparison of the paleohydrological variability with mid-to late Holocene climate changes in northeastern Canada

The main periods of water table changes over the last 4200 cal BP are similar to those recorded by Hughes *et al.*, (2006) in a peat bog from Newfoundland (Fig. 2.9) and also correspond with documented climate periods in northeastern Canada. The stable and wet surface conditions from 4200 to 3000 cal BP coincide with the end of the warm and moist Holocene Climatic Optimum identified by proxy climate records in Northern Québec (Filion, 1984 ; Filion *et al.*, 1991 ; Viau *et al.*, 2006). The late-Holocene surface drying in the peatlands of the St. Lawrence North Shore corresponds to a widespread cooling trend (Neoglacial) mainly forced by declining insolation at northern latitudes (Renssen *et al.*, 2005) which has been recorded by many terrestrial proxies. In northwestern Québec, particularly cold and dry conditions were documented after ~3000 cal BP by increased aeolian activity (Filion, 1984 ; Filion *et al.*, 1991), opening of the boreal forest (Gajewski, Payette and Ritchie, 1993 ; Payette and Gagnon, 1985) and increased fire occurrence (Carcaillet and Richard, 2000 ; Desponts and Payette, 1993). In the peatlands of the James Bay region, particularly dry surface conditions were also recorded after 3000 cal BP (van Bellen, Garneau and Booth, 2011). The driest surface conditions in the peatlands of the St. Lawrence North Shore and in Nordan's Pond Bog between 2400 and 1800 cal BP (Fig. 2.9) coincide with an important sea-surface cooling in the Gulf of St. Lawrence (Dhahri, 2010). The higher water tables after 1600 cal BP coincide with a period of warmer sea-surface temperature in the Gulf of St. Lawrence (Dhahri, 2010). In northwestern Québec, a warmer and/or moister period was also documented after *c.* 1100 cal BP with improved tree growth (Arseneault and Payette, 1997) and lower aeolian activity (Filion, 1984 ; Filion *et al.*, 1991). In eastern Canada, the warming trend linked to the Medieval Climate Anomaly (MCA) was interrupted by a major climate deterioration after 600 cal BP (Little Ice Age; LIA) documented by permafrost growth in the peatlands of the lower North Shore (Dionne and Richard,

2006) and forest retreat in the Labrador plateau (Payette, 2007). During this period, our WTD records indicate relatively wet and stable conditions except in Morts bog where an important surface drying occurred from 470 to 240 cal BP. However, the limits inherent in our ^{14}C chronology and the relatively low resolution of our analyses in upper peat layers limit the interpretation of recent hydrological changes (last 400 years). Further high-resolution studies within the upper peat layers are needed to evaluate the impact of the LIA cooling and recent climate changes on the hydrology of the studied peatlands.

2.5.3. Paleoclimatic interpretation of water table variability in the bogs of the St. Lawrence North Shore

The agreement of the bog paleohydrology with documented mid-to late Holocene climate changes in northeastern Canada suggests that these ecosystems have responded to some external climate forcing. However, the lack of independent proxy climate records in this region limits the interpretation of the key climate variables driving the reconstructed water table changes. The Holocene climatic patterns may have slightly differ between the St. Lawrence North-Shore and northern Quebec where most proxy climate data are available. Further proxy climate data will be required to evaluate more accurately the relationship between Holocene climate variability and the bog surface wetness in the maritime regions of eastern Canada.

The shifts in bog surface wetness have commonly been interpreted as changes from cool and/or wet to warm and/or dry climate in mid-Latitude Europe (Barber, Chambers and Maddy, 2003 ; Barber and Langdon, 2007 ; Charman and Hendon, 2000). It was suggested that water table records in peat bogs primarily reflect changes in summer water deficit and that moisture deficit in oceanic peatlands is mainly determined by precipitations (Charman, 2007 ; Charman *et al.*, 2004). In oceanic peat bogs of mid-Latitude Europe it was usually assumed that the winter conditions have few impacts on the bog surface wetness variability (Charman *et al.*, 2009).

In the present study, the wettest and more stable hydrological conditions between 4200 and 3000 cal BP coincides with a warm period reported from other proxy climate records whereas the driest surface conditions are related to a climate cooling documented during the late-Holocene. The increased surface-wetness variability and generally drier conditions during the colder late-Holocene cannot simply be interpreted in terms of decreased summer precipitations. In these coastal bogs from low boreal and subarctic environments, surface-wetness deficits may also have resulted from increased duration of seasonal freezing, wind-exposure and related decrease in snow covers at the bog surfaces.

In boreal and subarctic regions of North America, the bog surfaces are exposed to harsh wind-exposed winter conditions. The snow cover plays an important role in the water budget of nordic peatlands by providing important water supply at the start of growing season and by controlling the intensity and length of the seasonal freezing. The influence of these winter processes on testate amoebae communities is poorly documented in spite of their important implications for the interpretation of paleohydrological records in northern peatlands.

The surface hydrology of the plateau bogs of Havre-St-Pierre seems to have been particularly affected by the Neoglacial cooling. The particularly dry conditions in these bogs especially around 2400-1500 cal BP suggest important decadal-to-centennial water deficits which may have been amplified by the multiannual persistence of frozen peat as suggested by van Bellen, Garneau and Booth (2011). The change in the testate amoebae communities after 3000 cal BP suggests increased subsurface hydrological variability in the studied peatlands. In Baie and Lebel, the co-dominance of *A. discoides* (wet optima) and *H. subflava* (dry optima) between 3000 and 2000 cal BP can be interpreted as increased seasonal-to-interannual hydrological variability (Barber and Langdon, 2007 ; Sullivan and Booth, 2011 ; van Bellen, Garneau and Booth, 2011). In Morts and Plaine, the dry phase from

2600 cal BP is mainly related to high abundance of *D. pulex*. This species remains problematic in peatland paleohydrological studies due to its low abundance in modern training set. Sullivan and Booth (2011) showed that this species is characteristic of highly variable environmental conditions. Lamarre, Garneau et Asnong (2012) have associated permafrost growth in a subarctic peatlands with high abundance of *D. pulex* supporting the idea that it is indicative of drier and more variable hydrological conditions. Further studies will be needed to improve knowledge on this species which is often abundant in fossil samples but poorly represented in modern surface samples.

The differences in the amplitude of the water table fluctuations and in the timing of the late-Holocene drying between Baie-Comeau and Havre-St-Pierre may reflect distinctive regional climate conditions. In the north shore of the Gulf of St. Lawrence, the climate is strongly influenced by the cold water from the Labrador Current. In this region, the sea-ice cover plays an important role in the regional climate notably by modifying wind strength; air temperature and cloud cover (Faucher *et al.*, 2004). It is possible that the increased sea-ice cover duration and decreased sea-surface temperature (SST) in the Gulf of St. Lawrence during the Neoglacial cooling have amplified the atmospheric thermal response causing greater climatic impacts in the Havre-St-Pierre area. The similar patterns of hydrological variability between our study and the one of Hughes *et al.*, (2006) over the last 4200 cal BP (Fig. 2.9) provides further evidence of large-scale climatic influences on the paleohydrology of maritime bogs in eastern Canada.

2.6. Conclusion

This study provides detailed reconstruction of past hydrological variability in the peatlands of the north shore of the Estuary and Gulf of St. Lawrence in eastern Canada. The comparison of four peat-based paleohydrological records was used to detect long-term trends in regional hydroclimate variations. The climatic influence on bog hydrology is supported by the synchronous late-Holocene drying in the bogs and linkages with documented climate variations in northeastern Canada. The coherent periods of water table changes in the two regions since 4200 cal BP suggest an external forcing in both the Estuary and Gulf of St. Lawrence area. However, further proxy climate data are needed from terrestrial and oceanic sites to strengthen the interpretation of the climate drivers of bog paleohydrology in maritime regions. Further research is needed to better understand the land, oceanic and atmospheric interactions in the St. Lawrence North Shore. Considering the role of sea-ice conditions in this region, quantitative reconstructions of past sea-surface conditions in the Estuary and Gulf of St. Lawrence would be very useful to evaluate the key climate factors driving the paleohydrology of these maritime bogs.

The bogs of Havre-St-Pierre have experienced much drier surface conditions than those of Baie-Comeau over the late-Holocene which suggests a contrasted hydrological response to the Neoglacial cooling. This may reflect distinct peatland sensitivities between the Estuary and the Gulf of St. Lawrence likely related to the increasing subarctic influence. The greater sensitivity of the plateau bogs of Havre-St-Pierre to past climate deteriorations such as the Neoglacial cooling may be due to their ecotonal position at the wet and cold limit of northern peatland distribution (*sensu* Yu, Beilman and Jones, 2009).

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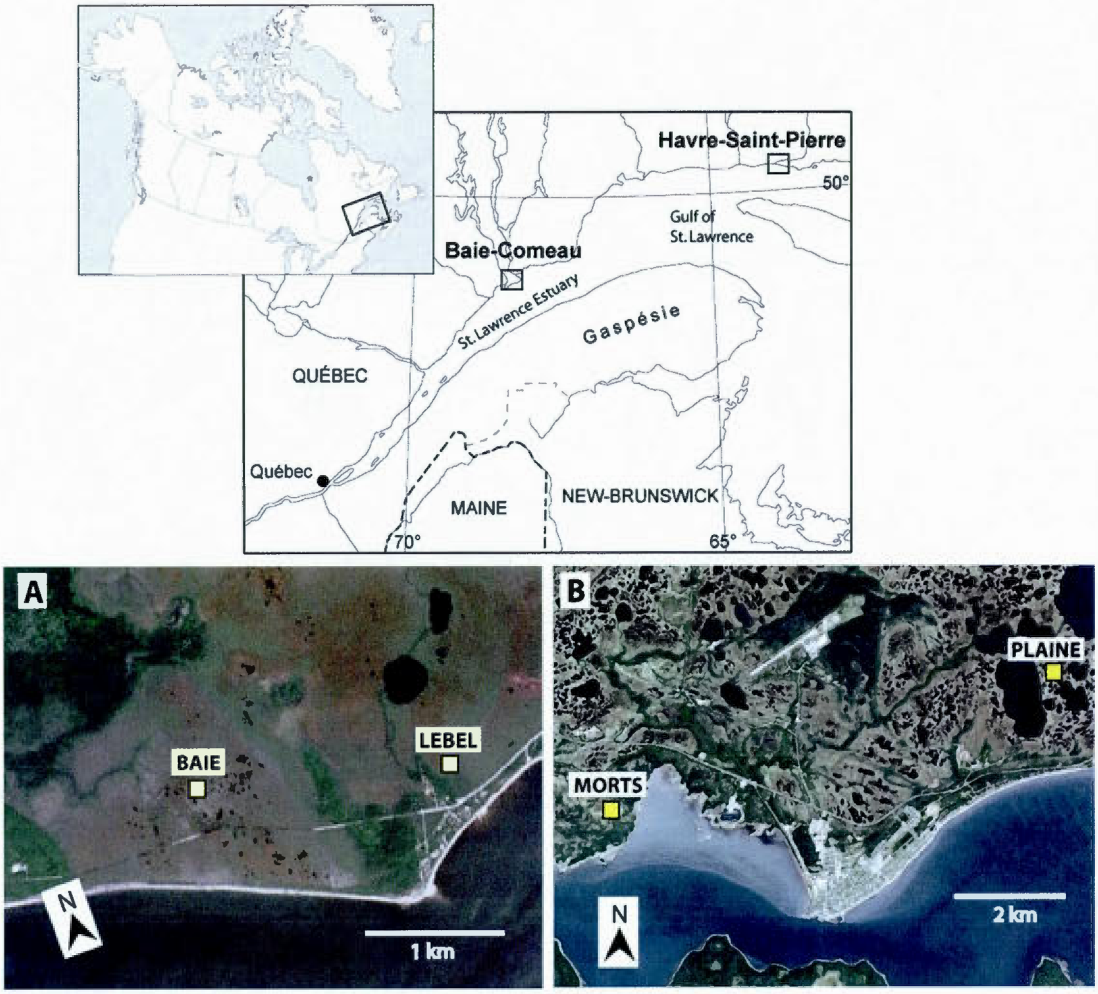

Figure 2.1. Location of the two study area and of the coring sites within the four studied peatlands along the St. Lawrence North Shore on A) the Manicouagan delta (BC) and B) the La Romaine delta (HSP). *Source*: Satellite images from Google Earth 2013.

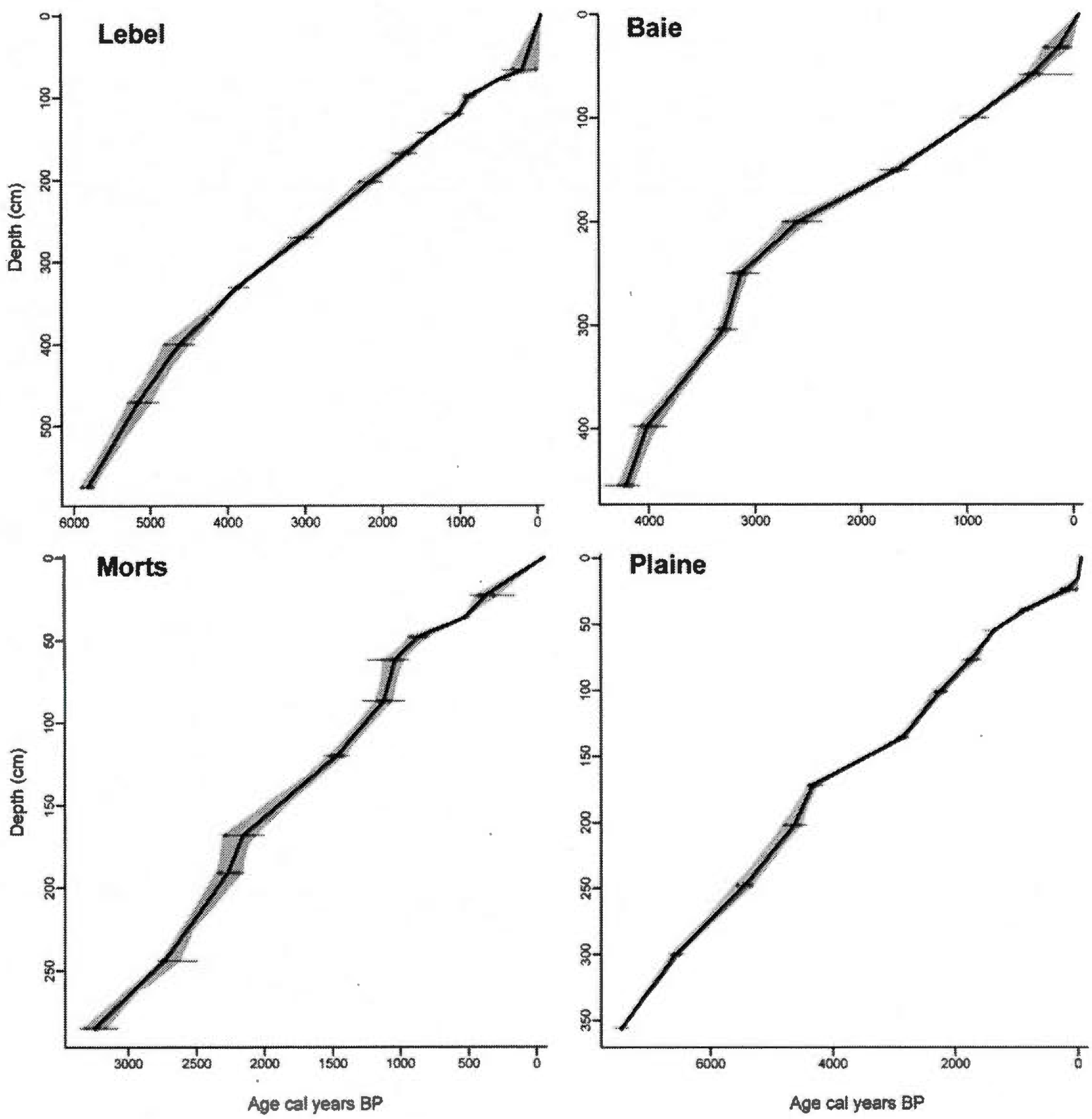

Figure 2.2. Age-depth models developed using Clam software by applying linear interpolation with the 95% confidence intervals for the four studied peat records.

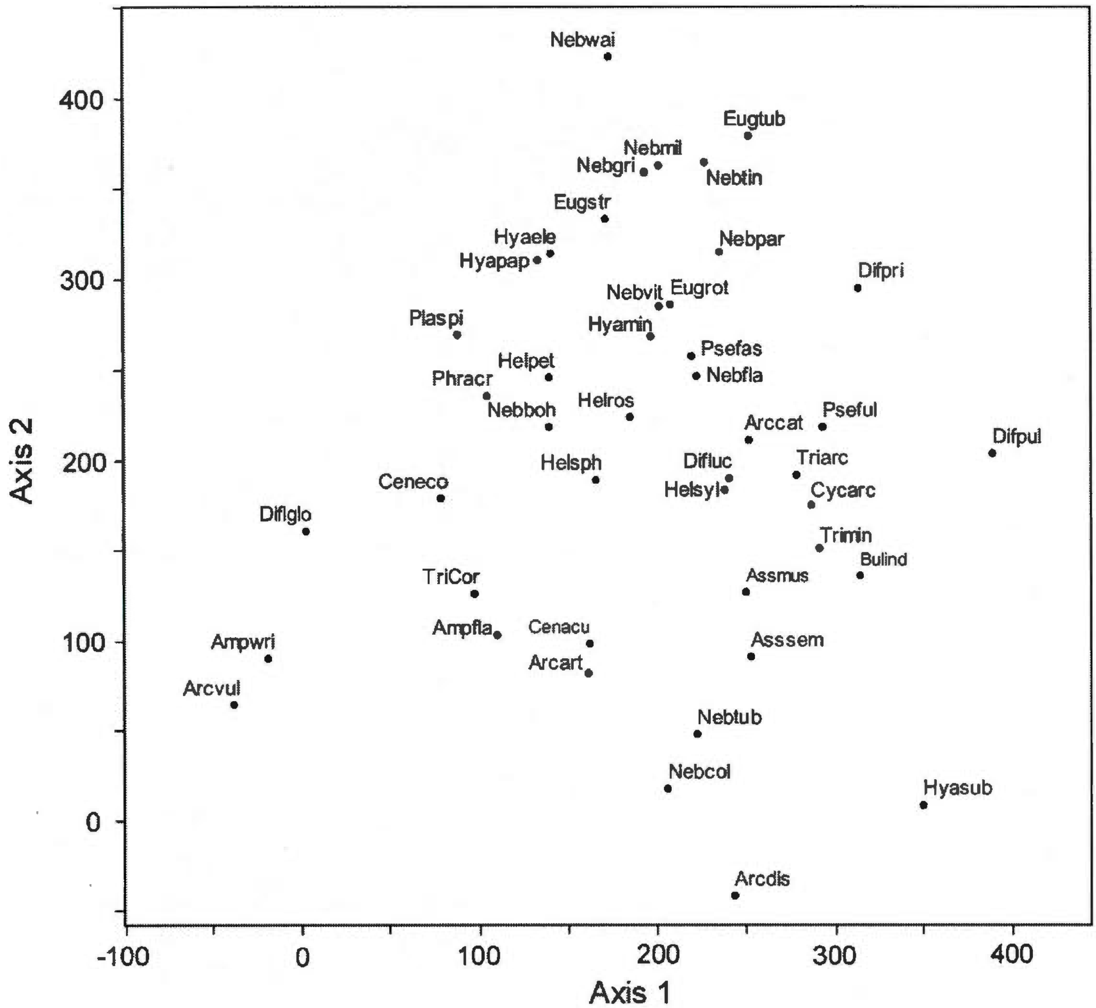

Figure 2.3. DCA ordination of the fossil testate amoebae data. Full taxa names : Ampfla–*Amphitrema (Archerella) flavum*, Ampwri–*Amphitrema wrightianum*, Arccat–*Arcella catinus*, Arcdis–*Arcella discoides*, Arcvul–*Arcella vulgaris*, Assmus–*Assulina muscorum*, Asssem–*Assulina seminulum*, Bulind–*Bullinularia indica*, Cenacu–*Centropyxis aculeata*, Cycarc–*Cyclopyxis arcelloides*, Ceneco–*Centropyxis ecornis*, Difglo–*Difflugia globulosa*, Diffluc–*Difflugia lucida*, Difpri–*Difflugia pristis*, Difpul–*Difflugia pulex*, Eugrot–*Euglypha rotunda*, Eugstr–*Euglypha strigosa*, Eugtub–*Euglypha tuberculata*, Helpet–*Heleopera petricola*, Helros–*Heleopera rosea*, Helsph–*Heleopera sphagni*, Helsyl–*Heleopera sylvatica*, Hyaele–*Hyalosphenia elegans*, Hyamin–*Hyalosphenia minuta*, Hyapap–*Hyalosphenia papilio*, Hyasub–*Hyalosphenia subflava*, Nebboh–*Nebela bohemica*, Nebcol–*Nebela collaris*, Nebfla–*Nebela flabellulum*, Nebgri–*Nebela griseola*, Nebmil–*Nebela militaris*, Nebpar–*Nebela parvula*, Nebtin–*Nebela tincta*, Nebtub–*Nebela tubulosa*, Nebvit–*Nebela vitrea*, Nebwai–*Nebela wailesi*, Nebpen–*Nebela penardiana*, Phracr–*Phryganella acropodia*, Plaspi–*Placocista spinosa*, Psefas–*Pseudodifflugia fascicularis*, Pseful–*Pseudodifflugia fulva*, Triarc–*Trigonopyxis arcula*, Trimin–*Trigonopyxis minuta*, TriCor–*Trinema/Corythion* types.

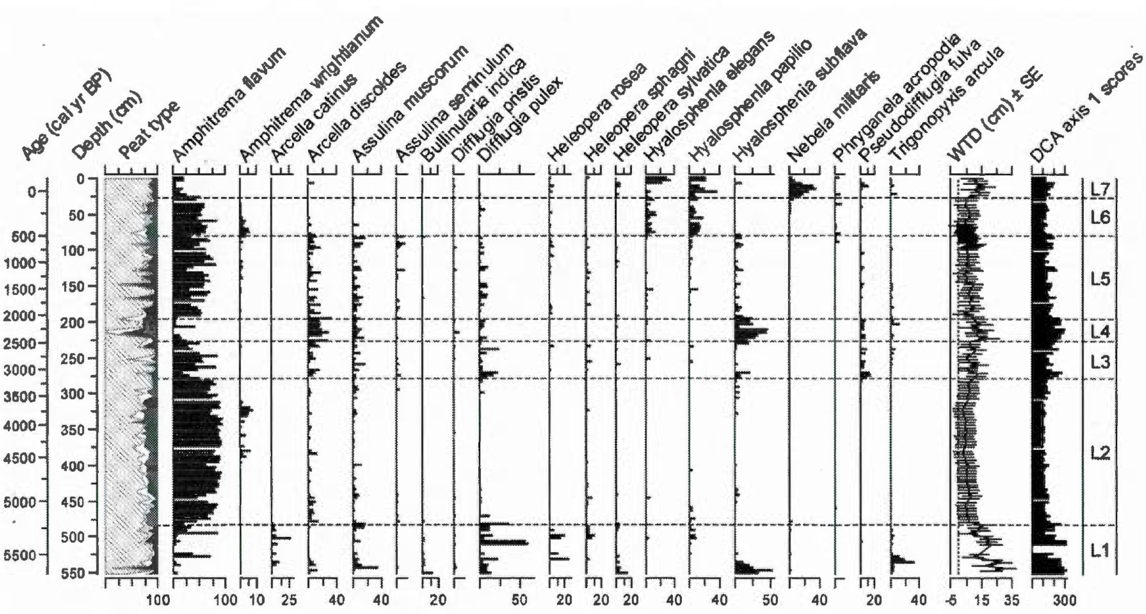

Figure 2.4. Testate amoebae diagram (%) of Lebel peatland (BC). Peat types : *Sphagnum* (diagonal lines), other bog mosses (light grey), Cyperaceae (white), ligneous (dark grey) and brown mosses (black).

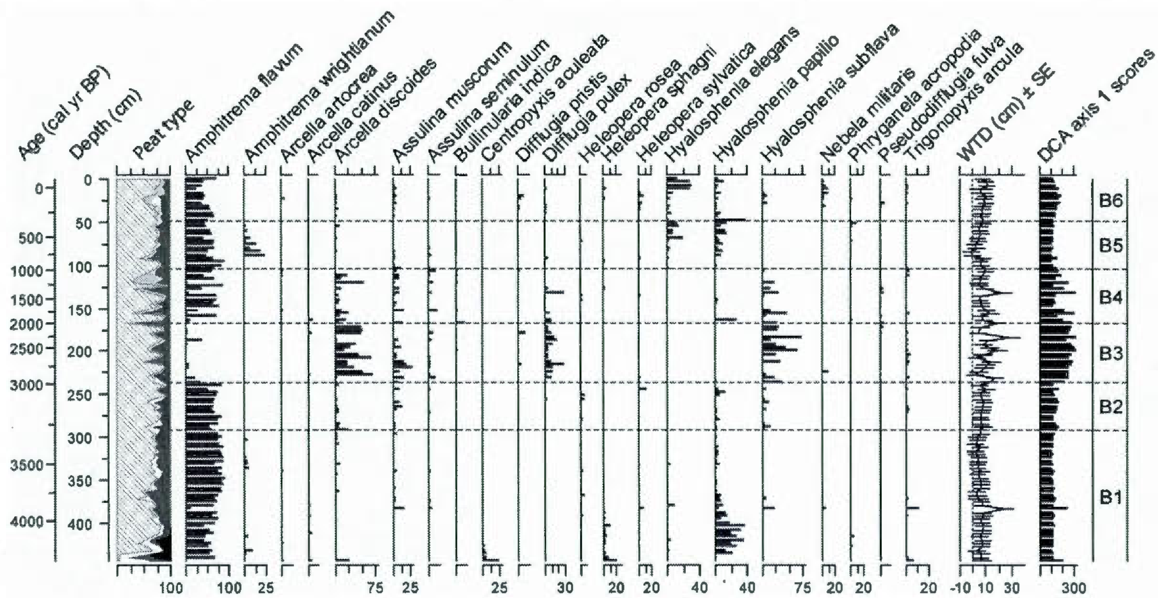

Figure 2.5. Testate amoebae diagram (%) of Baie peatland (BC). Peat types : *Sphagnum* (diagonal lines), other bog mosses (light grey), Cyperaceae (white), ligneous (dark grey) and brown mosses (black).

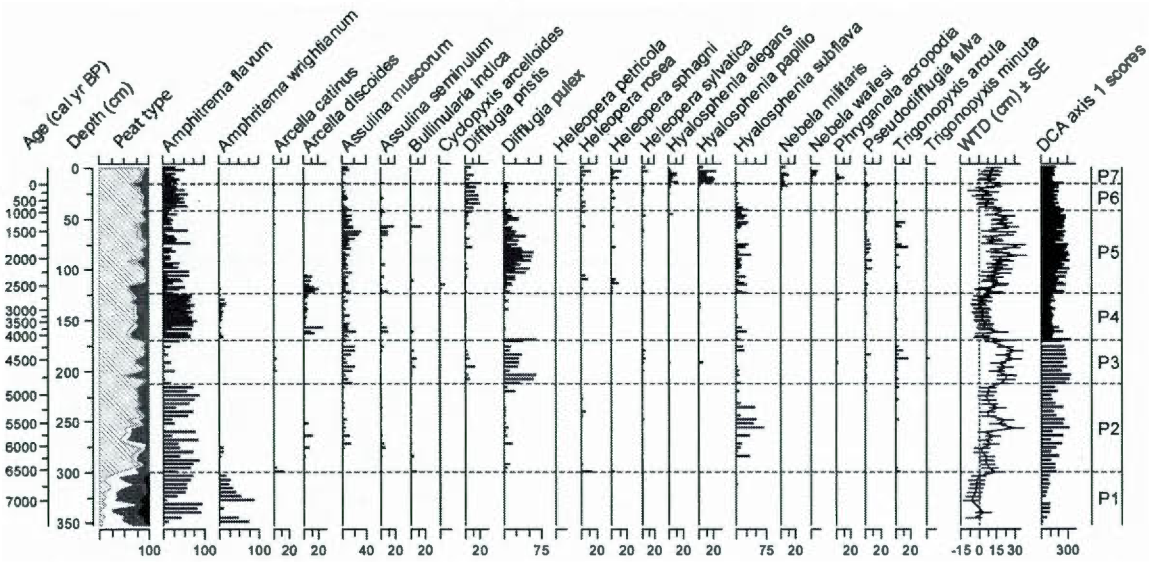

Figure 2.6. Testate amoebae diagram (%) of Plaine peatland (HSP). Peat types : *Sphagnum* (diagonal lines), other bog mosses (light grey), Cyperaceae (white), ligneous (dark grey) and brown mosses (black).

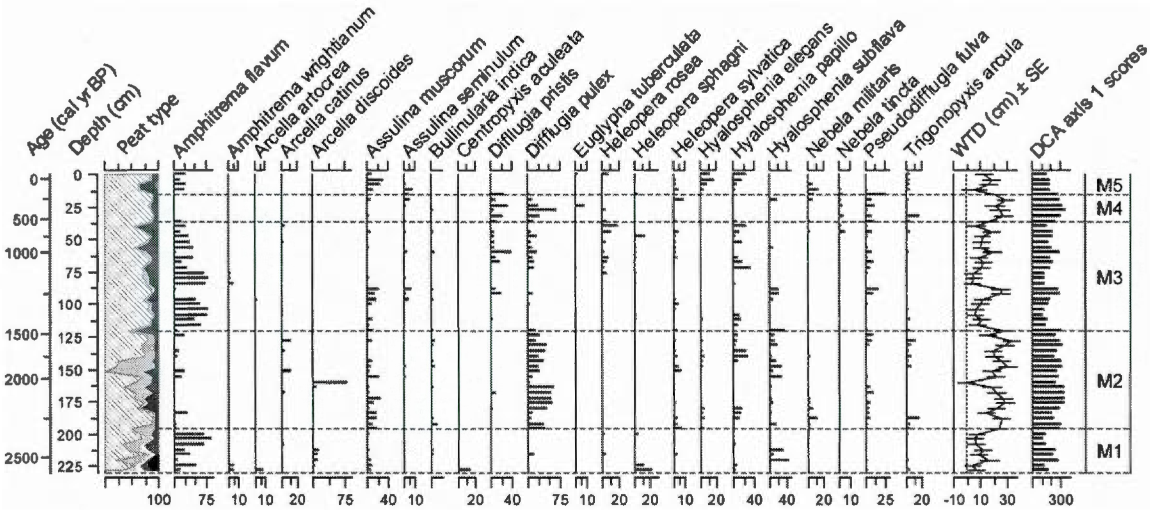

Figure 2.7. Testate amoebae diagram (%) of Morts peatland (HSP). Peat types : *Sphagnum* (diagonal lines), other bog mosses (light grey), Cyperaceae (white), ligneous (dark grey) and brown mosses (black).

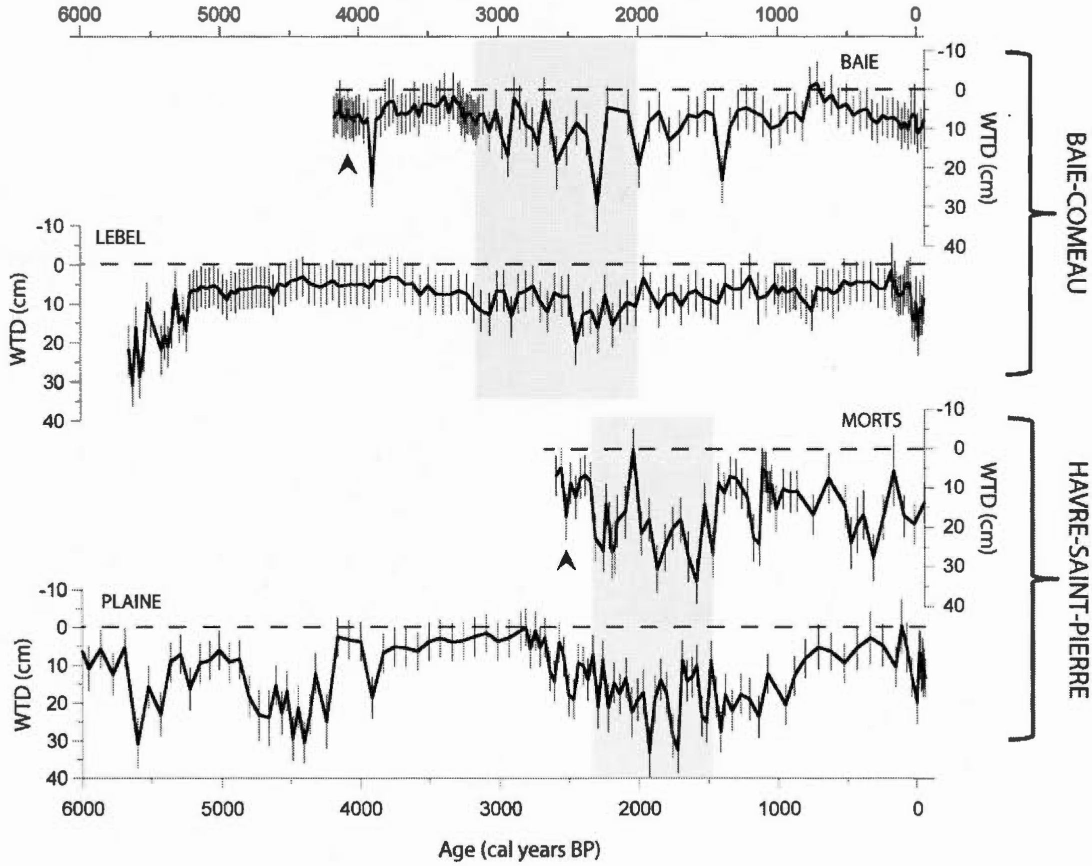

Figure 2.8. Comparison of the four water table depth records for the last 6000 cal years BP with sample-specific standard errors. The shaded boxes show the common late-Holocene period of increased hydrological variability and drier conditions in each region. The black arrows show the fen-bog transition in Baie and Morts.

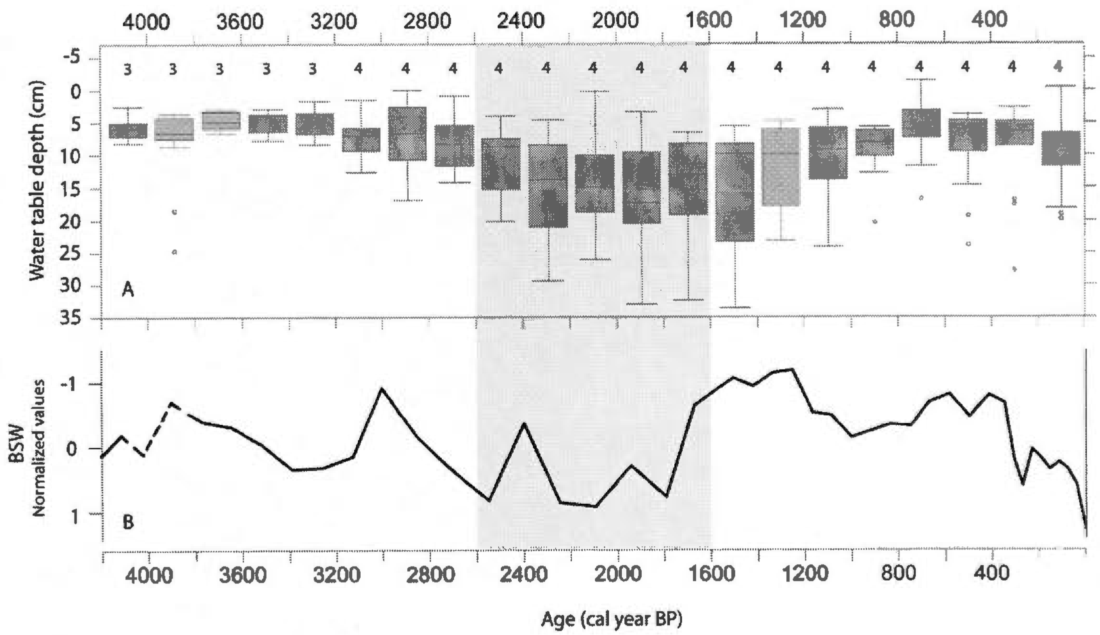

Figure 2.9. Synthesis of the WTD variations since 4200 cal BP from the four peat cores. Data were pooled into 200-year bins. The box plots show the lower and upper quartile, median and outliers. The combined record (A) is compared with the surface wetness record (BSW) from Nordan's pond bog in Newfoundland (B) (Hughes *et al.*, 2006). The shaded box shows the driest period from 2600 to 1600 cal BP.

Site	Sample depth (cm)	Lab. no (UCI-AMS)	Material dated*	^{14}C age (yr BP)	2 σ range (cal yr BP)	Cal. BP (*Clam* best estimate)
PLAINE	16-17	67838	Sph.	Modern	0	0
	24-25	79504	Sph.	150 ±15	6-281	150
	39-40	73854	Sph.	935 ±25	793-919	850
	55-56	98873	Sph.	1480 ±15	1328-1399	1360
	77-78	73855	Sph.	1800 ±25	1629-1818	1740
	101-102	73856	Sph.	2245 ±25	2157-2338	2240
	136-137	98874	Sph.	2765 ±15	2793-2921	2850
	172-173	80191	Sph.	3875 ±15	4244-4407	4320
	202-203	73857	Sph., Eric. seeds.	4100 ±25	4523-4808	4630
	248-249	79505	Sph., Eric. Seeds	4715 ±15	5327-5577	5430
	300-301	67839	Sph.	5755 ±15	6495-6632	6560
	356-357	67840	Carex ach.	6540 ±15	7426-7475	7450
MORTS	23-24	98870	Sph.	290 ±15	300-428	370
	36-37	79502	Sph.	480 ±15	506-531	520
	48-49	98871	Sph.	955 ±15	797-926	860
	62-63	73858	Sph.	1145 ±25	976-1168	1030
	87-88	73859	Sph.	1190 ±25	1014-1178	1120
	120-121	80188	Sph., Eric. lea.	1580 ±15	1415-1521	1470
	168-169	79503	Sph., Eric. seeds	2150 ±15	2066-2299	2150
	191-192	98872	Sph., Eric. lea.; Picea need.	2250 ±15	2162-2338	2260
	244-245	67844	Sph., Eric. lea.; Larix need.	2585 ±20	2625-2754	2730
	284-286	67843	Carex ach.	3025 ±15	3166-3327	3240
LEBEL	66-67	73861	Sph.	210 ±25	(-4)-303	180
	79-80	80190	Sph., Eric. lea.; Carex ach.	430 ±15	484-514	500
	98-99	79499	Sph., Eric. lea./seeds	965 ±15	798-929	870
	120-121	98875	Sph.	1125 ±15	979-1061	1020
	143-144	73862	Eric. lea.	1475 ±25	1311-1400	1360
	168-169	73863	Sph., Eric. lea.	1775 ±25	1613-1810	1680
	202-204	98876	Sph	2145 ±15	2063-2297	2150
	270-271	79500	Sph ; Eric. seeds	2905 ±15	2966-3139	3040
	331-332	79501	Sph stems/lea.	3570 ±15	3832-3910	3870
	400-401	73864	Sph stems/lea; Eric lea.	4090 ±25	4452-4805	4620
	471-472	98877	Sph.	4465 ±15	4978-5277	5160
	574-575	67837	Sph.	5090 ±15	5753-5908	5820
BAIE	32-33	98867	Sph.	105 ±15	28-259	130
	58-59	73865	Sph. Eric. lea.	300 ±25	300-455	380
	100-101	98868	Eric. lea.	1000 ±15	840-958	920
	150-151	79496	Sph.	1755 ±15	1614-1712	1660
	200-202	98869	Sph. Eric. Lea.	2495 ±15	2491-2715	2590
	250-251	79497	Sph.	2950 ±15	3065-3209	3130
	304-305	80196	Sph. Eric. seeds/lea.	3060 ±15	3219-3344	3290
	398-399	73866	Sph.	3670 ±25	3915-4086	4010
	455-456	79498	Sph., Carex ach.; Larix need.	3830 ±15	4153-4288	4210

* Sph = *Sphagnum*, Eric = *Ericaceae*, Lea.= leaves, ach. = achenes, need. = needles

Table 2.1. Radiocarbon data.

SITE	ZONE	Depth (cm)	Age (cal yr BP)	Mean WTD (cm) with SD	Main indicative species*
LEBEL	L7	28-0	40-Present	12.7(3.1)	HYAELE, HYAPAP, NEBMIL
	L6	80-28	520-40	5.4(1.6)	AMPFLA, AMPWRI, HYAELE, HYAPAP
	L5	196-80	2070-520	7.3 (2.1)	AMPFLA, ARCDIS, ASSMUS, DIFPUL
	L4	228-196	2490-2070	13.0 (4.2)	ARCDIS, HUASUB
	L3	280-228	3180-2490	9.0 (2.7)	AMPFLA,ARCDIS, DIFPUL, PSEFUL
	L2	484-280	5240-3180	5.7(1.3)	AMPFLA, AMPWRI
	L1	552-484	5680-5240	18.5 (6.5)	ARCCAT, ASSMUS, HYASUB, TRIARC
BAIE	B6	48-0	280-Present	8.4(1.6)	AMPFLA, HYAELE, NEBMIL
	B5	104-48	980-280	4.7(3.1)	AMPFLA, AMPWRI, HYAELE, HYAPAP
	B4	168-104	1990-980	8.7(4.5)	AMPFLA, ARCDIS, HYASUB,
	B3	236-168	2980-1990	11.0 (7.0)	ARCDIS, ASSMUS, DIFPUL, HYASUB
	B2	292-236	3250-2980	7.3(1.5)	AMPFLA
	B1	444-292	4180-3250	6.0(3.6)	AMPFLA, HYAPAP
PLAINE	P7	16-0	0-Present	11.5 (3.8)	HYAELE, HYAPAP, NEBMIL
	P6	42-16	950-0	8.0 (5.3)	AMPFLA, DIFPRI, HYAPAP, HYAELE
	P5	124-42	2640-950	17.5(6.5)	ASSMUS, DIFPUL, HYASUB
	P4	170-124	4240-2640	4.9(4.2)	AMPFLA, AMPWRI, ARCDIS
	P3	212-170	4800-4240	21.8(5.4)	ASSMUS, BULIND, DIFPUL
	P2	300-212	6560-4800	10.4(6.5)	AMPFLA, HYASUB
	P1	350-300	7350-6560	−1.5 (3.6)	AMPFLA, AMPWRI
MORTS	M5	16-0	240-Present	14.0(6.0)	AMPFLA, ASSMUS, HYAELE, HYAPAP
	M4	32-16	470-240	21.1(4.6)	DIFPRI, DIFPUL
	M3	120-32	1470-470	11.3(5.1)	AMPFLA
	M2	196-120	2310-1470	21.4(7.2)	DIFPUL, ASSMUS, TRIARC
	M1	228-196	2590-2310	9.0(3.9)	AMPFLA

* Full taxa names are presented in Fig. 2.3

Table 2.2. Details of the testate amoebae diagram zonation for the four peat records.

CHAPITRE III

A REGIONAL COMPARISON OF THE LONG-TERM CARBON DYNAMICS WITHIN MARITIME OMBROTROPHIC PEATLANDS ALONG THE NORTH SHORE OF THE ESTUARY AND GULF OF ST. LAWRENCE, EASTERN CANADA

Gabriel Magnan[1], Michelle, Garneau[1]

[1]GEOTOP - Université du Québec à Montréal (UQAM), C.P. 8888 Succursale Centre-ville, Montréal, Québec, Canada H3C 3P8.

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Résumé

Les tourbières ombrotrophes sont très répandues dans les régions maritimes côtières de l'est du Canada et représentent d'important puits terrestres de carbone (C). L'objectif de la présente étude est de comparer la dynamique à long terme d'accumulation du C entre les tourbières de deux régions écoclimatiques de la Côte-Nord dans l'est du Québec. L'influence climatique régionale sur la dynamique holocène du C a été évaluée en calculant les taux d'accumulation à long terme du C (LORCA) au sein de huit tourbières entre l'estuaire (Baie-Comeau) et le golfe du St. Laurent (Havre-St-Pierre). Aussi, afin d'évaluer l'influence des changements climatiques régionaux sur le bilan d'accumulation du C, les variations temporelles des taux de séquestration du C ont été comparées entre six tourbières. De plus, les processus de production et de décomposition de la tourbe ont été comparés à la végétation et aux nappes phréatiques. Les résultats montrent une relation négative entre les LORCA et l'âge des dépôts de tourbe qui reflète l'influence d'une décomposition anoxique constante. Toutefois, l'accumulation à long terme du C (LORCA) est significativement moins élevée dans les tourbières de Havre-St-Pierre (16–46 g m^{-2} an^{-1}) que de Baie-Comeau (53–68 g m^{-2} an^{-1}). Cette différence régionale suggère un contrôle climatique persistant sur la productivité primaire dans ces écosystèmes au cours de l'Holocène. Cependant, l'influence de la variabilité climatique sur l'accumulation du C s'est fait par l'intermédiaire d'interactions complexes entre la nappe phréatique et la production primaire et suggère l'influence de rétroactions écohydrologiques. Enfin, les données semblent indiquer que le bilan du C des tourbières de Havre-St-Pierre a été plus affecté par les changements climatiques passés en particulier au cours du refroidissement Néoglaciaire.

Abstract

Ombrotrophic peatlands are widespread in the coastal maritime regions of eastern Canada and these ecosystems are important terrestrial carbon (C) sinks. In this study, we aimed to compare the Holocene peatland C dynamics between two maritime ecoclimatic regions from the Estuary (Baie-Comeau) and the Gulf of St. Lawrence (Havre-St-Pierre). Long-term apparent rates of C accumulation (LORCA) were calculated for eight peatlands and the temporal variations in C accumulation rates were compared between six peat cores to evaluate potential influences of climate variability on their C balance. We also evaluated the role of water table variability and vegetation changes on the long-term balance of C input and decay. Our results show a negative relationship between the age of the peat deposits and the LORCA which reflects an influence of a constant anoxic decay. The Holocene rates of C accumulation (LORCA) are significantly lower in Havre-St-Pierre (16–46 g m^{-2} yr^{-1}) than in Baie-Comeau (53–68 g m^{-2} yr^{-1}). This regional difference suggests a pervasive climatic control on primary production in these ecosystems during the Holocene. However, the influence of climate variability on C accumulation was mediated by complex interactions between water table levels and the production/decay balance which point to the existence of ecohydrological feedbacks. Our data suggest that the C balance in the peatlands of Havre-St-Pierre was more affected by past climate changes especially during the Neoglacial cooling.

3.1. Introduction

Peatlands are widespread in boreal and subarctic regions of North-America. These ecosystems have sequestered large amounts of organic carbon (C) over the Holocene due to a positive balance between biomass production and decomposition of organic matter under waterlogged and nutrient-poor conditions. Globally, northern peatlands store about 547 Gt of C (Yu *et al.*, 2010) corresponding to approximately one third of world's soils C (Gorham, 1991). In Canada, peatlands cover about 12 % of the land area (~1.1 million km^2) and store ~147 Gt of C (Tarnocai, 2006). They are abundant in the maritime regions of eastern Canada especially in the coastal plains along the St. Lawrence North Shore where large complexes of ombrotrophic peatlands cover the postglacial deltaic formations. Along the North Shore, the average thickness of the peat deposits tends to decrease significantly between the Estuary (Baie-Comeau) and the Gulf of St. Lawrence (Havre-St-Pierre) (see Chapter 1). This regional difference suggests a climatic control on the accumulation of peat and related C over the Holocene. The C dynamics of these ombrotrophic maritime peatlands remains unknown in spite of their fundamental ecological and hydrological functions in the coastal environments.

In eastern Canada, previous studies on Holocene peatland C dynamics were mainly conducted in continental regions within the St. Lawrence Lowlands (Muller, Richard and Larouche, 2003 ; Roulet *et al.*, 2007) and in the James Bay Lowlands (Lamarre, Garneau and Asnong, 2012 ; Loisel and Garneau, 2010 ; van Bellen, Garneau and Booth, 2011). Further data on long-term C accumulation is needed from various peatland types under a wider range of climate contexts to improve global estimates of C sinks. Improving our understanding of the factors driving the long-term C dynamics in these peatlands is essential to evaluate their response to anticipated environmental and climate changes.

Net C sequestration in peatlands is directly influenced by precipitation and temperature (Malmer *et al.*, 2005) that control the net primary production (NPP) and the aerobic decay in the upper peat layers (acrotelm). This climatic influence is reflected in a north-south trend of decreasing C accumulation rates positively correlated with mean annual temperature (Beilman *et al.*, 2009), number of growing degree-days (Clymo, Turunen and Tolonen, 1998) and photosynthetically active radiation (Charman *et al.*, 2013). In northern peatlands, highest rates of C accumulation are registered in regions with intermediate mean annual temperatures (0–2.5°C) and mean annual precipitations (450–550 mm) (Yu, Beilman and Jones, 2009). At the regional scale, long-term rates of C accumulation may vary significantly within and between peatlands due to geomorphic and topographic influences (Korhola *et al.*, 1996 ; van Bellen *et al.*, 2011) and the variable ages of the peat deposits (Turunen, Tomppo and Tolonen, 2002).

The present study is the first to explore the Holocene C dynamics in maritime peatlands from the St. Lawrence North Shore in eastern Canada. The main objective is to compare the Holocene C dynamics between two maritime ecoclimatic contexts from the boreal to the subarctic domain. Specifically we aim to 1) evaluate the role of Holocene climate on the long-term patterns of C accumulation and 2) evaluate the interactions between the bog surface wetness and the C accumulation rates within these peatlands. We hypothesize that the long-term C accumulation is lower in the bogs near Havre-St-Pierre due to their ecotonal position at the wet/cold limit for peatland development (*sensu* Yu, Beilman and Jones, 2009).

3.2. Methods

3.2.1. Study area and sites

A total of eight ombrotrophic peatlands were studied from the maritime boreal (St. Lawrence estuary) and the subarctic regions (Gulf of St. Lawrence). We have evaluated the C dynamics in the raised bogs of the Manicouagan delta near Baie-Comeau and in the plateau bogs of La Romaine delta near Havre-St-Pierre (Fig. 3.1). In each region, we compared peatland ecosystems of different ages that developed in distinct geomorphic contexts to evaluate how these factors have potentially influenced the long-term C accumulation. In the two studied regions, peat accumulation started over sandy deltaic terraces after the withdrawal of the Goldthwait Sea from 7.5 ka BP and above silt-clay deposits associated with a marine transgression after 4.2 ka cal BP. More details on the context of peat inception are presented in Chapter 1.

Regional climate is maritime with mean annual precipitations of approximately 1000 mm in both regions and mean annual temperature of 1.5° C in Baie-Comeau and 1.1°C near Havre-St-Pierre (climate normals 1971-2000; Environment Canada). The growing seasons are longer in Baie-Comeau with degree-days >0°C of 1924 (±128) compared to 1779 (±138) in Havre-St-Pierre (Hutchinson *et al.*, 2009). Along the Gulf of St. Lawrence, the climate conditions are more subarctic due to the influence of the cold Labrador Current and the increased wind-exposure. In the landscape, this is reflected by a gradual opening of the regional forest cover and a change in the general morphology of the bogs from the Estuary to the Gulf area.

3.2.2. Field sampling

The sites were first selected from aerial photographs according to their representativeness of the regional peat-forming systems. We have chosen sites with relatively homogenous underlying basins, where lateral peat expansion is not strongly constrained, in order to limit potential topographic influences on long-term peatland development (Belyea and Clymo, 2001 ; van Bellen *et al.*, 2011). At each site, peat thickness was measured along a series of transects evenly distributed throughout the peatlands in order to find the deepest sections.

Four peat cores (Lebel, Plaine, Romaine, Romaine-C) were collected on the sandy deltaic terraces (alt. 16-30 m.a.s.l) and four others (Baie, Manic, Morts, Morts-Sud) were retrieved above marine deposits (alt. 11-16 m.a.s.l) (Fig. 3.1). Peat cores were taken from *Sphagnum*-dominated lawns in the central and deepest sections of the peatlands in order to minimise edge effects. Cores were retrieved using a Russian corer (diameter: 7.5 cm; Jowsey, 1966). The upper first meter of peat was sampled with a Box corer (105×8×8 cm). Sediments were wrapped with cellophane and placed in PVC tubes before being stored in the fridge at 4°C until analysis.

3.2.3. Radiocarbon dating and chronology

Fifty-three (53) samples of *Sphagnum* moss and vascular plant remains were submitted to Keck laboratory (University of California, Irvine) for AMS radiocarbon dating (Table 3.1). The age-depth models were developed using Clam software (version 1.0.2) (Blaauw, 2010) by applying linear interpolation between each ^{14}C date and attributing an age of −60 cal yr BP (i.e. AD 2010; year of coring) to the peatland surface. Radiocarbon dates were calibrated using the IntCal09 calibration curve (Reimer *et al.*, 2009). All dates are expressed in calendar years before present and rounded to the nearest decade (BP: before AD 1950).

3.2.4. Organic matter and C data

Analyses were conducted on contiguous peat samples (interval: 1 cm; volume: 3-cm^{-3}). Dry bulk densities (g cm^{-3}) were determined after overnight drying and organic matter (OM) contents (ash-free bulk densities) were evaluated with loss-on-ignition (LOI) at 550°C for 3 hours (Dean, 1974). C concentration was estimated at 50% of the OM content which is in line with measurements from previous studies (Charman *et al.*, 2013 ; Turunen *et al.*, 2002).

The average long-term apparent rates of C accumulation (LORCA ; g m^{-2} yr^{-1}) was calculated for eight peat profiles by dividing the total mass of C accumulated by the age of the basal peat (Turunen *et al.*, 2002). Long-term variations in C accumulation rates were also calculated by dividing the C content of every 1-cm increment by the deposition time (years cm^{-1}) and values are reported on an area basis (m^{-2}). In order to minimise the confounding influence of peat age on the apparent rates of C accumulation (Beilman *et al.*, 2009 ; Clymo, Turunen and Tolonen, 1998), the LORCA were compared for defined time periods since 2000 and 1000 cal years BP from six peat cores for which age-depth models were developed. These dates were estimated by linear interpolation between two adjacent ^{14}C-dated levels.

3.2.5. Peat humification and paleohydrological records

Peat humification was used as a high-resolution proxy to investigate the variations in the balance between production and decay within the two oldest peat cores of each region (i.e. Lebel and Plaine). Analyses were conducted on contiguous peat samples (1-cm interval) using the colorimetric method (Chambers, Beilman and Yu, 2010). The samples (0.2 mg) were boiled in a solution of sodium hydroxide (NaOH 8%) for 1 hour, filtered through Whatman Qualitative 1 filter paper and diluted. Readings on the extracted alkali were performed with a Thermo Spectronic

Unicam spectrophotometer using a wavelength of 540 nm and recorded as percentage light-transmission (% LT) (lower values indicating more humified peat). Values were averaged from two or three replicates in each sample. Humification data were corrected to take into account the mineral content based on the loss-on-ignition (LOI) values (Chambers, Beilman and Yu, 2010 ; Payne and Blackford, 2008) using the formula $LT_c = LT_r/(1/LOI)$ where LT_c is the corrected light-transmission, LT_r is the raw LT value (%) and LOI is expressed as a proportion. The LT record was linearly detrended to remove the effect of continuing anaerobic decay and data were smoothed with a LOWESS (2 % smoothing factor) using Kaleidagraph 3.6. The C accumulation rates for Plaine and Lebel were also compared to a water table depth (WTD) reconstruction based on testate amoebae analyses previously performed on these two peat records (see Chapter 2).

3.3. Results

3.3.1. Long-term apparent rates of C accumulation (LORCA)

LORCA vary significantly between the studied peatlands ranging from 16.0 to 67.8 g m^{-2} yr^{-1} and the average rates of peat accumulation varies between 0.26 and 1.44 mm yr^{-1} (Table 3.2). The rates of C accumulation are significantly higher in the peatlands of Baie-Comeau (Fig. 3.2). However, the data show a negative relationship between the age of peat inception and the LORCA values. In each region, the highest rates of C accumulation are recorded in the yougest peat deposits over the marine deposits whereas the lowest values are found in the older peatlands that formed over the sandy deltaic terraces (Fig. 3.2).

3.3.2. Long-term patterns of C sequestration

The age-depth relationships are represented as cumulative C mass curves in Fig. 3.3. In the six studied peat cores, the C mass versus age relationships are either linear or convex. The data show a clear regional difference in the patterns of C sequestration as the total mass of C accumulated is about twice in the peatlands of Baie-Comeau. In these sites most of the C was sequestered rapidly during the early peatland stages. In Baie and Manic (BC), about half of the total C mass was accumulated during the first 1000 years following peat inception (Fig. 3.3). The regional differences in the rates of C sequestration are particularly important over the last 2000 years. The LORCA values calculated since 2000 cal BP are about two to three times higher in the bogs of Baie-Comeau (Fig. 3.4). The regional differences are particularly important since 1000 cal BP, for example the LORCA is very low (13.7 g m^{-2} yr^{-1}) in Romaine (HSP) and high (66.5 g m^{-2} yr^{-1}) in Lebel (BC) during that period.

3.3.3. Temporal variations in C accumulation rates (CAR)

The patterns of the C accumulation rates diverge significantly from site to site even within each region (Fig. 3.5). The temporal variations in C sequestration are very marked in the peatlands of Havre-St-Pierre which contrasts with the gradual changes in the Baie-Comeau sites. The CAR were very high during the early fen stage in Manic and Baie (maximum: 188 and 149 g m^{-2} yr^{-1} respectively) while the accumulation rates were much lower within the fen stage in Plaine and Morts (35.9 and 48.3 g m^{-2} yr^{-1} respectively).

3.3.4. Relationship between C accumulation and paleohydrological variability

The processes controlling the long-term C balance were evaluated for Plaine and Lebel by linking the C accumulation rates with vegetation composition, peat humification and water table depths (Fig. 3.6). In Plaine, the C accumulation rates have been highly variable and relatively low through time (mean = 27.6 g m^{-2} yr^{-1}). The main shifts in C accumulation coincide with changes in WTD and peat humification. C sequestration was higher within the fen stage (mean = 35.9 g m^{-2} yr^{-1}) and declined slightly following the transition to ombrotrophy. Within the bog, the highest rates of C accumulation (mean = 40.8 g m^{-2} yr^{-1}) coincide with a dry phase around 4650-4300 cal BP in a section highly dominated by *Sphagnum*. This was followed by a period with much lower accumulation (mean = 12.5 g m^{-2} yr^{-1}) from 4300 to 2900 cal BP during a wet phase with abundant wood remains. A shift towards higher C sequestration rates between 2800 and 2250 cal BP (28.6 g m^{-2} yr^{-1}) corresponds to an important water table drawdown and highest peat humification. Over the last 2000 years, the slowdown in C accumulation coincides with low humification and a gradual rise in the water table depth.

In Lebel, the rates of C accumulation were very stable and high (mean = 65.9 g m^{-2} yr^{-1}) and were related to a persistence of relatively wet surface conditions (high WTD) and high dominance of *Sphagnum*. Prior to 5250 cal BP, high rates of C accumulation (mean = 83.6 g m^{-2} yr^{-1}) and minimal peat humification are associated with a gradual rise in water table levels associated with the paludification over well-drained sands (Chapitre 1). C sequestration rates were particularly stable over the last 4600 years in spite of the variations in peat humification and WTD, an indication that the balance between production and decay remained stable despite the changes in aerobic decay. The very high C accumulation rates (>150 g m^{-2} yr^{-1}) and low humification recorded in recent peat layer (<200 cal BP) likely correspond with the acrotelm in which peat is still subjected to aerobic decomposition.

3.4. Discussion

3.4.1. Holocene C sequestration between the two regions

Our study shows a clear difference in the Holocene C sequestration rates between the two studied regions which validate the main research hypothesis. The long-term apparent rates of C accumulation (LORCA) from the peatlands on the deltaic sands of Havre-St-Pierre (16–29.3 g m^{-2} y^{-1}) are similar to published data from eastern Canada (19 g m^{-2} yr^{-1}; Turunen *et al.*, 2004), continental North America (29 g m^{-2} yr^{-1}; Gorham, 1991) and James Bay lowlands (10 to 34 g m^{-2} yr^{-1}; Loisel and Garneau, 2010; van Bellen, Garneau and Booth, 2011). The LORCA in the peatlands of Baie-Comeau (52.8–67.8 g m^{-2} y^{-1}) are among the highest reported from boreal maritime peatlands (Yu, Beilman and Jones, 2009). Our data showed a negative relationship between the LORCA and the age of the peat deposits that reflects an influence of a constant anoxic decay on the apparent rates of C sequestration (Fig. 3.2). These results are in line with Clymo, Turunen and Tolonen (1998) and suggest that it can be misleading to compare apparent rates of C accumulation between peatlands with significantly different basal ages.

3.4.2. Long-term patterns of C accumulation

In the studied peat cores, the cumulative C mass versus age relationships were either linear or slightly convex (Fig. 3.3) as previously reported from many boreal peatlands (Kuhry and Vitt, 1996 ; Mäkilä, Saarnisto and Kankainen, 2011 ; Turunen and Turunen, 2003 ; van Bellen *et al.*, 2011 ; Yu, 2006). This contrasts with the theoretical bog growth model (BGM) which predicts a concave C mass-age relationship when C input is constant and decay is proportional to the remaining mass (Clymo, 1984 ; Clymo, Turunen and Tolonen, 1998). In the present study, we cannot assume that C input has been constant over time because the linear C mass-age relationship (Fig. 3.3) would imply that the rate of anoxic was nearly zero. Since we found evidence of continuing anoxic decay (Fig. 3.2) the data rather suggest that the rate of C input has gradually decreased over time.

3.4.3. Processes controlling the long-term C balance

The C accumulation curves show that the balance between C input and decay has varied significantly over time and more importantly in the bogs of Havre-St-Pierre (Fig. 3.5 and Fig. 3.6). Changes in long-term C accumulation were previously attributed to shifts in peat-forming vegetation affecting both the production of biomass (NPP) and the decomposition (litter quality) (Loisel and Garneau, 2010 ; van Bellen, Garneau and Booth, 2011). In our study, the changes in vegetation composition seem to have had few impacts on the C balance. This may be mainly due to the fact that the stratigraphy of two studied peat cores (Lebel and Plaine) was relatively homogenous and largely dominated by *Sphagnum* (Fig. 3.6).

In Manic, Baie and Plaine, C accumulation rates were maximal during the early fen stage as reported by earlier studies (Anderson, 2002 ; Bunbury, Finkelstein and Bollmann, 2012 ; Vardy *et al.*, 2000). The initial C accumulation rates within the fens were likely much higher than recorded considering that these older peat layers are the most affected by anoxic decay. The very high rates of C accumulation recorded within the fens of Manic and Baie (BC) around 4200-4000 cal BP likely resulted from high primary production promoted by a nutrient-enrichment from brackish waters after the withdrawal of marine waters at these sites. Overall, in Plaine bog, the rates of C accumulation tend to be higher in the sections highly dominated by *Sphagnum* peat.

Our results showed that the relationship between the water table depth and the C balance was complex (Fig. 3.6). In Plaine peatland, the highest C accumulation rates coincided with relatively dry conditions and high peat humification whereas the lowest rates occurred during wet periods with low humification. These seemingly contradictory results suggest that the increased decay under dry conditions was offsets by higher plant productivity while the low C accumulation rates under wet conditions resulted from low plant productivity rather than increased decay.

In Lebel, the high rates of C accumulation during the early peatland stage are related with low humification values and dry conditions, an indication that the surface was continuously isolated from the water table by fast peat accumulation. Overall, the higher rates of C accumulation in Lebel seem to have been promoted by the stability of the hydrological conditions as previously shown in the bogs of the James Bay area (van Bellen, Garneau and Booth, 2011). In Lebel bog, the peat-forming processes and water table levels may have interacted to maintain a stable acrotelm thickness with high rates of litter production as suggested by a conceptual model (Belyea and Clymo, 2001).

3.4.4. Role of climate on the C sequestration over the Holocene

The discrepancies in the patterns of C accumulation between sites even within each region (Fig. 3.5) suggest that the long-term balance of peat production and decay was predominantly controlled by internal peatland processes. The relation between the water table depth and the C balance appears to be non-linear and point to the existence of ecohydrological feedback mechanisms (Swindles *et al.*, 2012). The influence of long-term climate variability on C accumulation is more complex than previously suggested by peat-based paleoclimatic studies (Bunbury, Finkelstein and Bollmann, 2012 ; Lamarre, Garneau and Asnong, 2012 ; Mauquoy *et al.*, 2004 ; Yu *et al.*, 2003a). Although past climate changes may have directly influenced the hydrology of these bogs (Chapitre 2), the C balance has not responded consistently to changes in water table depths. These results agree with conceptual models and suggest that peatlands are complex adaptative systems (Belyea and Baird, 2006 ; Morris, Baird and Belyea, 2012). Changes in C sequestration may result from shifts in surface conditions with external climate forcing pushing the system over some thresholds (Belyea and Malmer, 2004). However, our study suggests that the long-term trajectories in peatland C accumulation were ultimately driven by the regional climate conditions through their persistent influence on the primary production.

In the peatlands of the St. Lawrence North Shore, the C input to the catotelm has significantly decreased over time. This contrasts with other maritime peatlands characterised by a constant input of C over the course of the development (Clymo, 1984). The paleohydrological reconstructions (Fig. 3.6) show that the slowdown in C accumulation is not simply resulting from a gradual surface drying related to autogenic development (Yu *et al.*, 2003b). We suggest that the gradual slowdown in C sequestration in the bogs of the Québec North Shore was driven by a combined influence of autogenic vertical development and climate change especially during the late-Holocene. As the bogs developed vertically, the plant productivity may have been affected by increased wind-exposure and this autogenic effect was likely amplified during the Neoglacial cooling. The particularly low C sequestration rates since 1000 cal yr BP in the peatlands of Havre-St-Pierre (Fig. 3.4) may be attributable to the LIA cooling as proposed by previous studies (Loisel and Garneau, 2010 ; Mauquoy *et al.*, 2002 ; van Bellen, Garneau and Booth, 2011). In northeastern Québec, a major climate deterioration after 600 cal BP was documented by permafrost development in the bogs of the Lower North Shore (Dionne and Richard, 2006), a cooling of the sea-surface temperatures in the Gulf of St. Lawrence (Dharhi, 2010) and an opening of the forest cover in the Labrador plateau (Payette, 2007). Further research is needed to evaluate how these recent climate changes have affected the C dynamics in the bogs of the St. Lawrence North Shore.

The contrasting long-term C sequestration patterns between the two regions suggest a climatic control on the long-term C balance over the Holocene. The marked shifts in the rates of C accumulation in the peatlands of Havre-St-Pierre (Fig. 3.5) seem to reveal that the production/decay balance in these ecosystems was more sensitive to past climate changes. Considering that these bogs are located at the wet/cold limit of northern peatland development (Yu, Beilman and Jones, 2009), it is possible that only slight changes in past climate conditions had significant impacts on their C dynamics. The bogs of Havre-St-Pierre seem to have been particularly affected by the Neoglacial cooling when the primary production was likely restricted by shorter and colder growing season and prolonged periods of seasonal freezing.

3.5. Conclusion

This study contributes to our understanding of the factors controlling the C sequestration in maritime peatlands Over the Holocene, the peatlands of Baie-Comeau and Havre-St-Pierre were characterised by distinct C dynamics and different sensitivities to climate changes. The long-term rates of C sequestration were much lower in the peatlands of Havre-St-Pierre particularly since 2000 cal years BP and their C balance was more sensitive to external forcing. The steadier and consistently higher C accumulation rates in Lebel bog (BC) was favored by stable hydrological conditions while in Plaine bog (HSP), the production/decay balance was affected by important shifts in water table levels. Our data suggest that the influence of past climate variability on the C balance of these peatlands was mediated by complex interactions between the water table depths and the peat-forming processes and point to an influence of ecohydrological feedbacks. However, our data showed that the long-term trajectories of C accumulation have been driven by a pervasive climatic influence at the regional scale especially from the late Holocene to the present.

Overall, the lower Holocene C accumulation in the bogs of Havre-St-Pierre may be related to shorter and colder growing seasons with increased influence of seasonal freezing that significantly affected the plant productivity. The coastal maritime peatlands of eastern Canada are particularly affected by harsh wind-exposed winter conditions that impact on frost depth and duration. Further research will be needed to evaluate the role of these winter processes on the C dynamics of the peatlands at the southern limit of subarctic environments. Also, more proxy climate data will be needed for this region to be able to evaluate the key climate variables that influenced the past C accumulation in peatlands.

Understanding the sensitivity of the peatlands C balance to past climatic and environmental changes is key to evaluate their future role in the global C cycle. The St. Lawrence North Shore has already experienced important warming over the last decades (Hutchison *et al.*, 2009). These ongoing climate changes may affect peatland hydrology and hence primary production, decomposition rates and fluxes of CO_2 and CH_4. In this context, the plateau bogs in the Gulf of St. Lawrence are particularly vulnerable due to their ecotonal position where slight climate changes may transform their C sequestration function.

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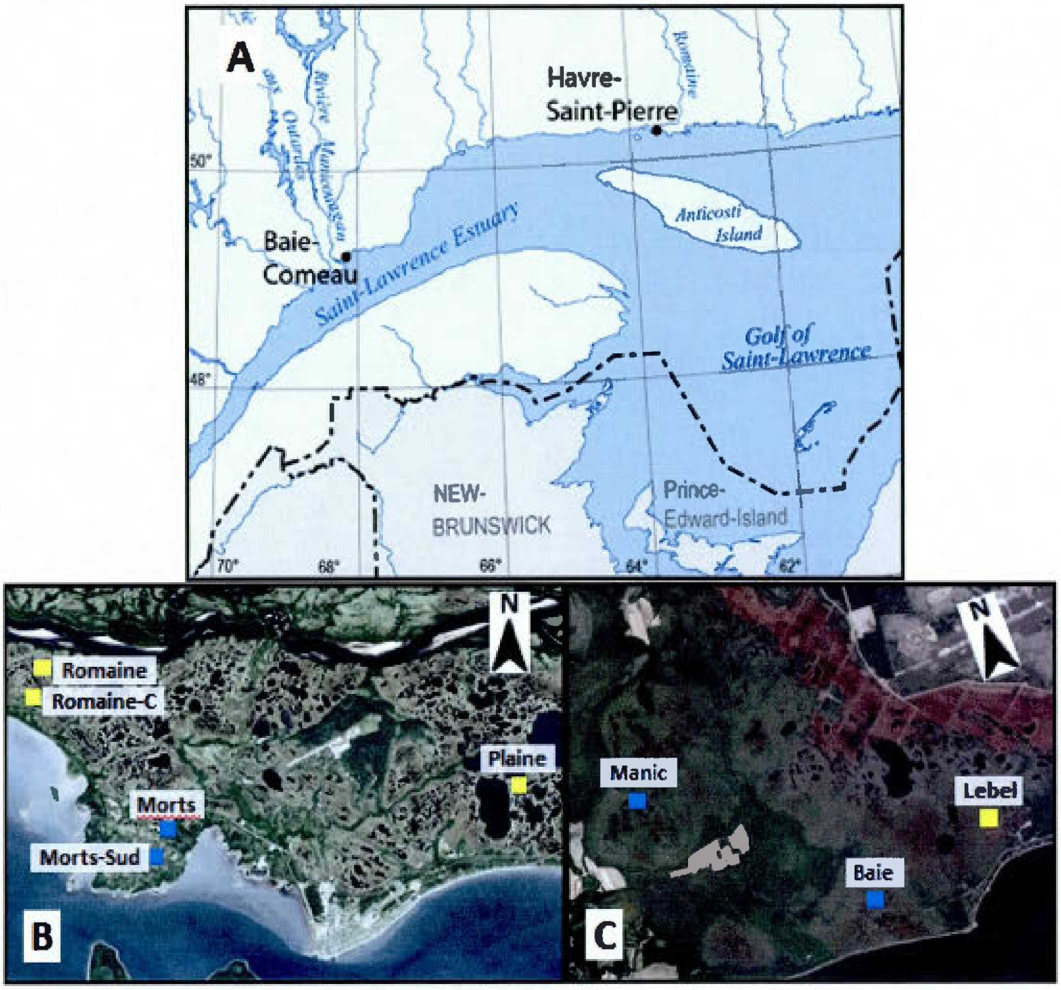

Figure 3.1. A) Location of the two studied regions along the St. Lawrence North Shore in eastern Canada and aerial views of the studied peatlands on B) the Manicouagan delta near Baie-Comeau and C) the La Romaine delta near Havre-St-Pierre. Yellow squares : sites on deltaic sands, Blue squares : sites on marine silt-clay. *Source*: aerial views from Google Earth 2013.

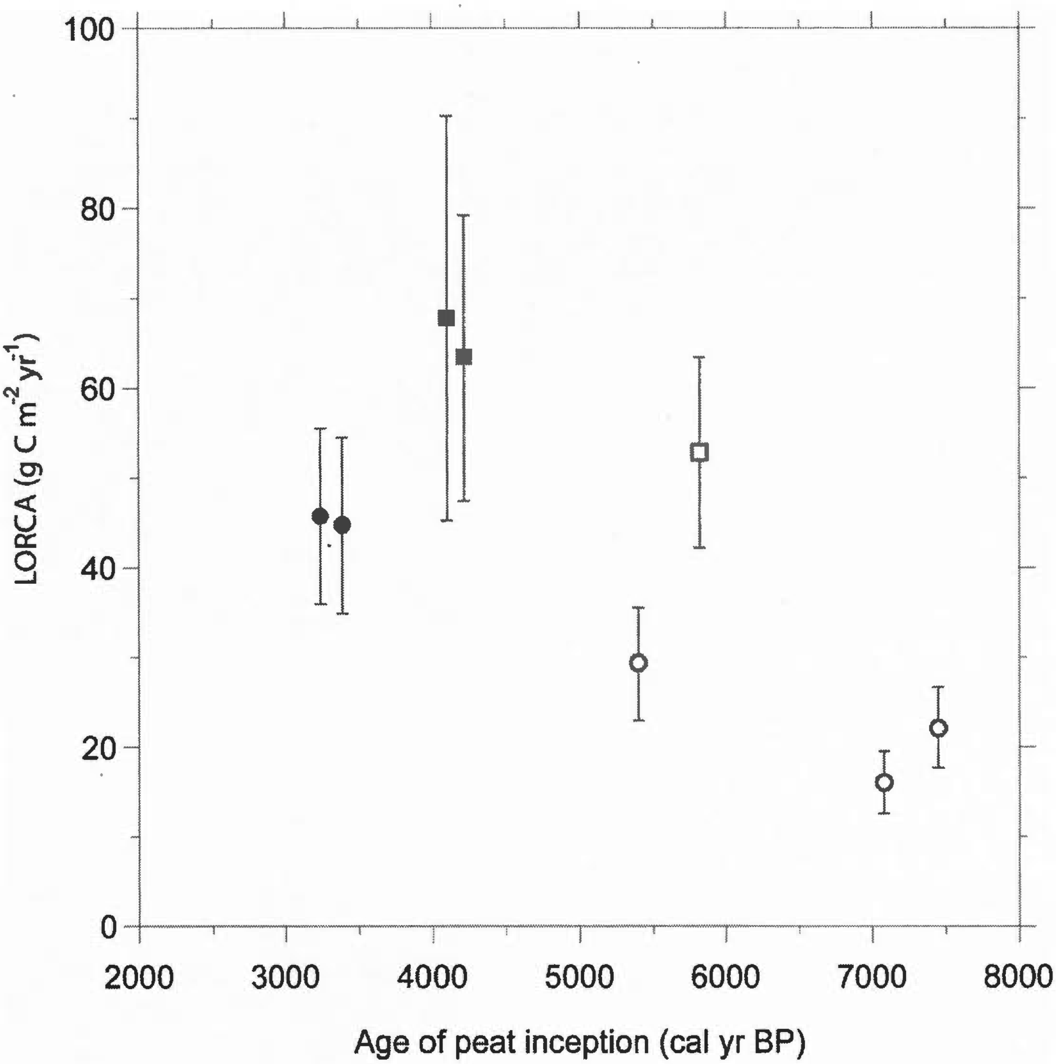

Figure 3.2. Relationship between the age of peat inception and the LORCA values with standard deviations for the peatlands of Havre-St-Pierre (circles) and Baie-Comeau (squares) that are underlain by deltaic sands (open symbols) or silt-clay deposits (filled symbols).

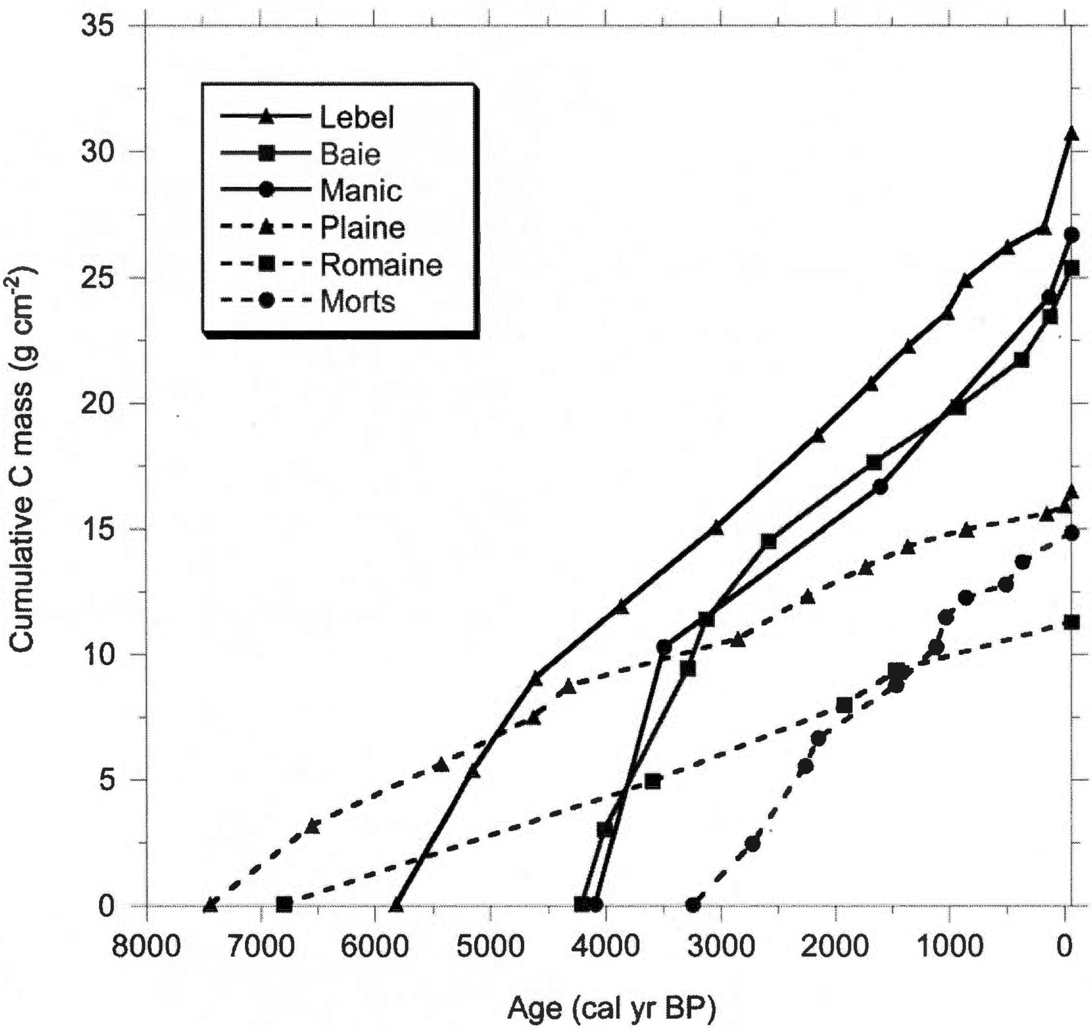

Figure 3.3. Cumulative C mass versus age curves for the six peatlands with age-depth models from Havre-St-Pierre (dashed line) and Baie-Comeau (continuous line).

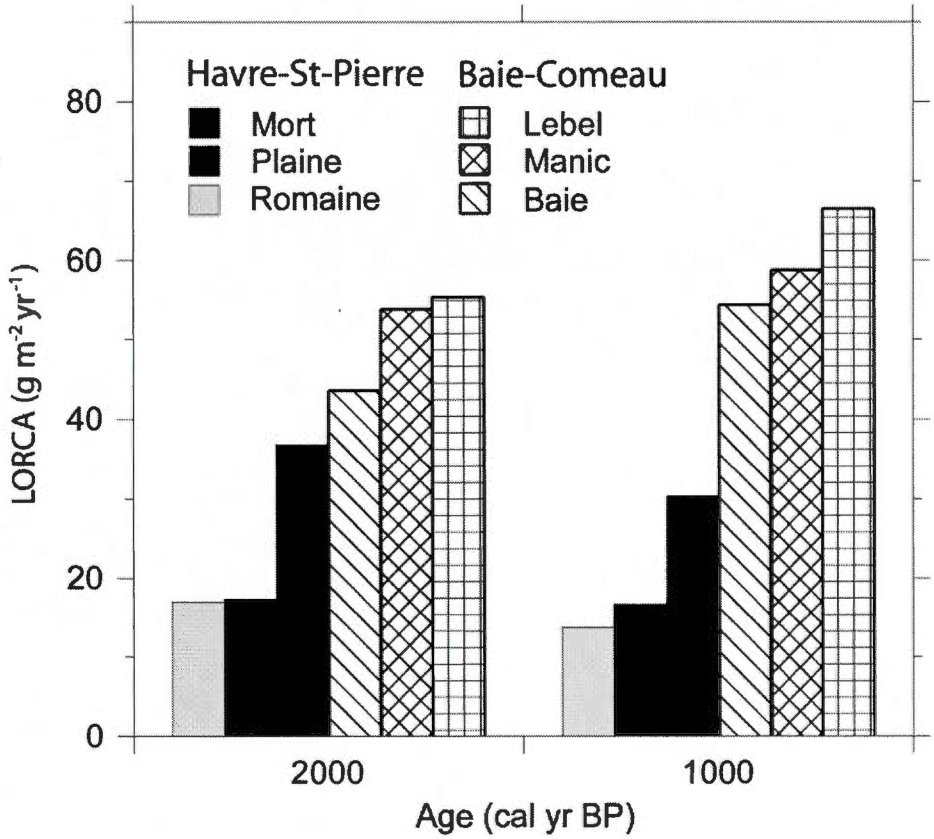

Figure 3.4. Long-term rates of C accumulation (LORCA) since 2000 and 1000 cal years BP for the six studied peatlands.

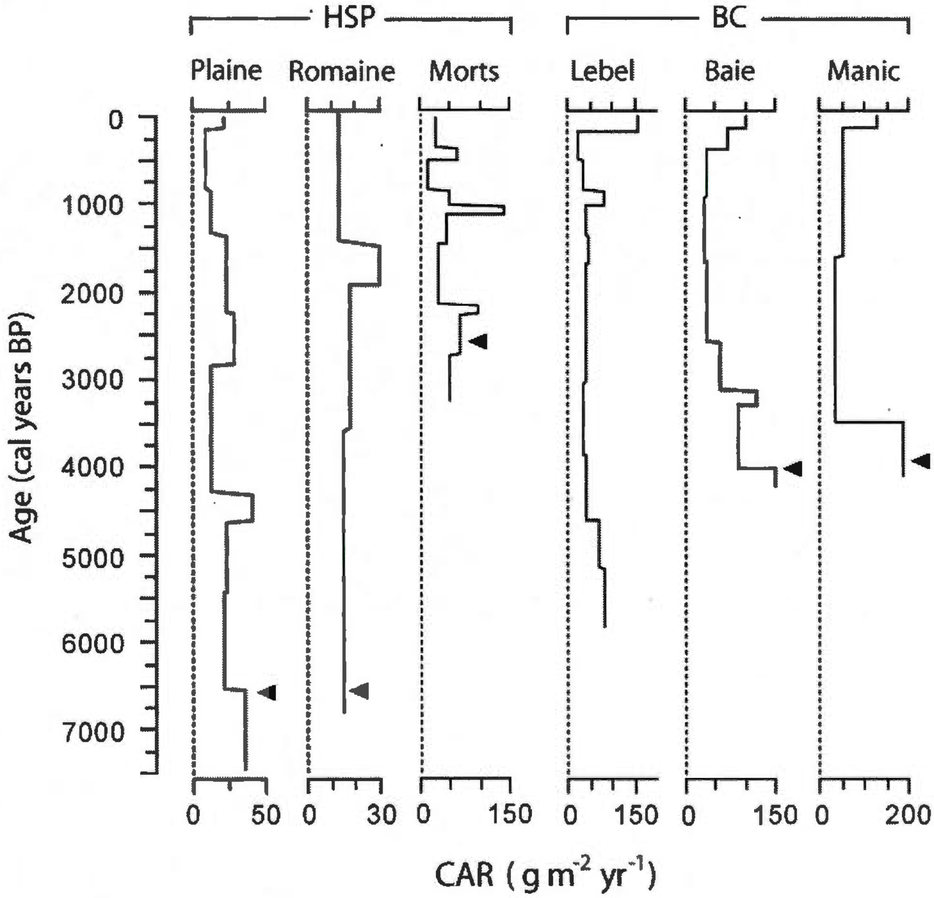

Figure 3.5. Temporal variations in the rates of C accumulation (CAR; mean value between each ^{14}C date). The age of the fen-bog transition is indicated by a triangle.

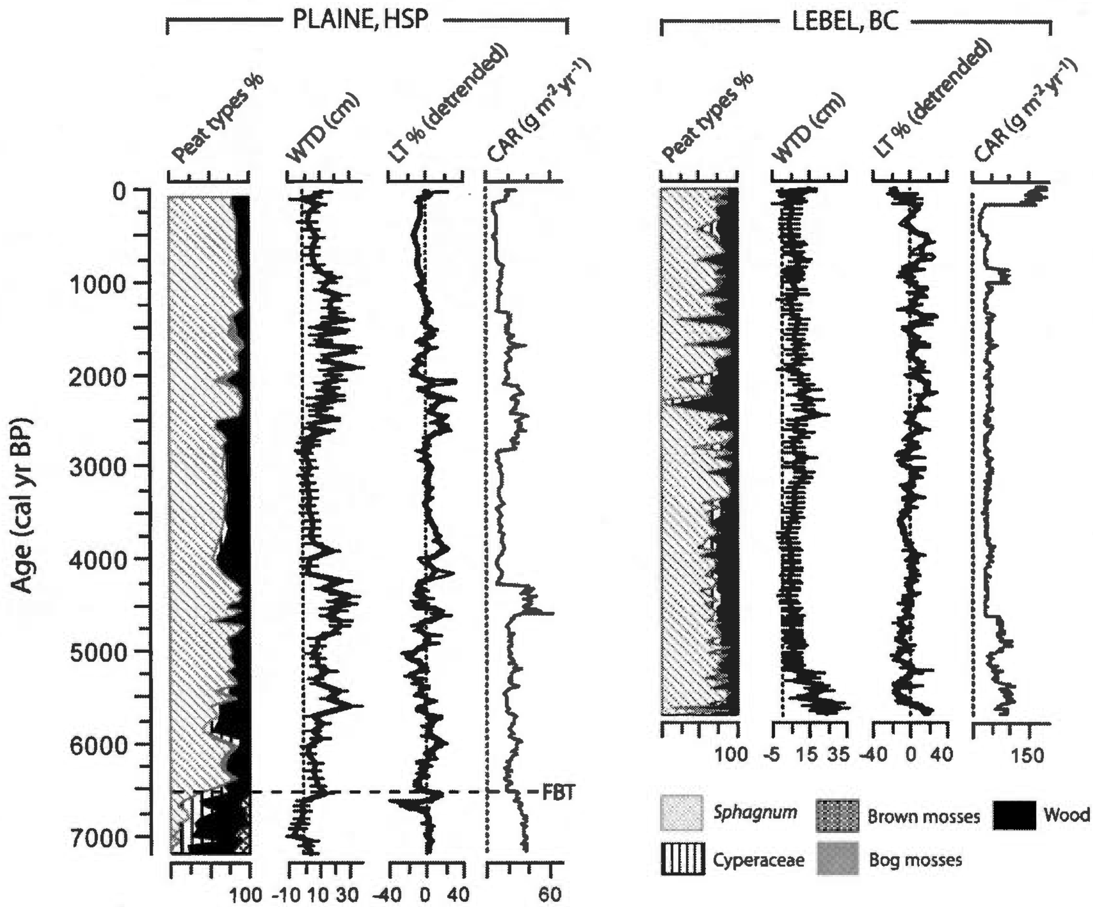

Figure 3.6. Comparison of the C accumulation rates (CAR) with peat vegetation types, water table depth (WTD) and peat humification (Light-transmission, LT) from Plaine and Lebel bog. The timing of the fen-bog transition (FBT) in Plaine is shown.

Site	Sample depth (cm)	Lab. no. UCI-AMS	Material dated[1]	^{14}C age (yr BP)	2 σ range (cal yr BP)	Cal yr BP[2]
PLAINE	16-17	67838	Sph.	Modern	0	0
	24-25	79504	Sph.	150 ±15	6-281	150
	39-40	73854	Sph.	935 ±25	793-919	850
	55-56	98873	Sph.	1480 ±15	1328-1399	1360
	77-78	73855	Sph.	1800 ±25	1629-1818	1740
	101-102	73856	Sph.	2245 ±25	2157-2338	2240
	136-137	98874	Sph.	2765 ±15	2793-2921	2850
	172-173	80191	Sph.	3875 ±15	4244-4407	4320
	202-203	73857	Sph., Eric. seeds	4100 ±25	4523-4808	4630
	248-249	79505	Sph., Eric. seeds	4715 ±15	5327-5577	5430
	300-301	67839	Sph.	5755 ±15	6495-6632	6560
	356-357	67840	*Carex* ach.	6540 ±15	7426-7475	7450
MORTS	23-24	98870	Sph.	290 ±15	300-428	370
	36-37	79502	Sph.	480 ±15	506-531	520
	48-49	98871	Sph.	955 ±15	797-926	860
	62-63	73858	Sph.	1145 ±25	976-1168	1030
	87-88	73859	Sph.	1190 ±25	1014-1178	1120
	120-121	80188	Sph., Eric. lea.	1580 ±15	1415-1521	1470
	168-169	79503	Sph., Eric. seeds	2150 ±15	2066-2299	2150
	191-192	98872	Sph., Eric. lea.; *Picea* need.	2250 ±15	2162-2338	2260
	244-245	67844	Sph., Eric. lea.; *Larix* need.	2585 ±20	2625-2754	2730
	284-286	67843	*Carex* ach.	3025 ±15	3166-3327	3240
MORTS-SUD	255-258	80189	*Carex* ach.	3170±15	3363-3443	3400
ROMAINE	32-33	79506	Sph.	1595 ±15	1416-1529	1470
	64-65	73860	Sph.	1975 ±25	1878-1988	1920
	110-111	79507	Sph.	3350 ±15	3490-3639	3590
	185-187	67845	*Carex* ach.	6200 ±15	7016-7168	7080
ROMAINE-C	259-262	80197	Sph	4665±15	5319-5465	5390
LEBEL	66-67	73861	Sph.	210 ±25	(-4)-303	180
	79-80	80190	Sph., Eric. lea.; *Carex* ach.	430 ±15	484-514	500
	98-99	79499	Sph., Eric. lea./seeds	965 ±15	798-929	870
	120-121	98875	Sph.	1125 ±15	979-1061	1020
	143-144	73862	Eric. lea.	1475 ±25	1311-1400	1360
	168-169	73863	Sph., Eric lea.	1775 ±25	1613-1810	1680
	202-204	98876	Sph	2145 ±15	2063-2297	2150
	270-271	79500	Sph. ; Eric. seeds	2905 ±15	2966-3139	3040
	331-332	79501	Sph.	3570 ±15	3832-3910	3870
	400-401	73864	Sph.; Eric lea.	4090 ±25	4452-4805	4620
	471-472	98877	Sph.	4465 ±15	4978-5277	5160
	574-575	67837	Sph.	5090 ±15	5753-5908	5820
BAIE	32-33	98867	Sph.	105 ±15	28-259	130
	58-59	73865	Sph. Eric. lea.	300 ±25	300-455	380
	100-101	98868	Eric. lea.	1000 ±15	840-958	920
	150-151	79496	Sph.	1755 ±15	1614-1712	1660

	200-202	98869	Sph. Eric. lea.	2495 ±15	2491-2715	2590
	250-251	79497	Sph.	2950 ±15	3065-3209	3130
	304-305	80196	Sph. Eric. seeds/lea.	3060 ±15	3219-3344	3290
	398-399	73866	Sph.	3670 ±25	3915-4086	4010
	455-456	79498	Sph., *Carex* ach.; Larix need.	3830 ±15	4153-4288	4210
MANIC	50-51	80192	Sph.	110 ±15	24-261	130
	220-221	80193	Sph.	1700 ±15	1548-1691	1610
	365-366	80194	Sph.	3265 ±15	3447-3556	3490
	588-590	80195	Sph.; *Carex* ach.	3745 ±15	4000-4153	4100

[1] Sph = *Sphagnum*, Eric.= *Ericaceae*, Lea.= leaves, ach. = achenes, need. = needles

[2] Calibrated ages inferred from Clam software (Blaauw, 2010).

Table 3.1. Radiocarbon dates.

STUDY REGION	Peatland	Lat. N	Long. W	Peat depth (cm)	Basal age (cal yr BP)	Mean peat accretion rate (mm yr^{-1})	Total C content (g cm^{-2})	LORCA (g m^{-2} yr^{-1}) (SD)
Baie-Comeau	Lebel	49° 5.9'	68° 13.3'	575	5820	0.99	30.7	52.8 (11)
	Baie	49° 5.8'	68° 15'	461	4215	1.09	26.8	63.5 (16)
	Manic	49° 7.1'	68° 18.3'	590	4100	1.44	27.8	67.8 (22)
Havre-St-Pierre	Plaine	50° 16.5'	63° 32.3'	356	7448	0.48	16.5	22.1 (5)
	Romaine	50° 17.7'	63° 42.9'	187	7080	0.26	11.3	16.0 (3)
	Rom-C	50° 17.3'	63° 43.3'	262	5400	0.49	15.8	29.3 (6)
	Morts	50° 15.8'	63° 40.1'	285	3245	0.89	15.3	45.7 (10)
	Morts-S	50° 15.6'	63° 39.7'	258	3390	0.76	15.2	44.7 (10)

Table 3.2. Description of the eight peat profiles with peat and C accumulation data.

CONCLUSION GÉNÉRALE

Cette thèse avait comme principal objectif d'évaluer l'influence du climat holocène sur la dynamique écohydrologique et l'accumulation du C dans les tourbières ombrotrophes maritimes de la région de la Côte-Nord. L'utilisation d'une approche comparative inter et intra-régionale a permis d'évaluer l'influence des facteurs externes dont le climat et les feux par rapport aux processus internes des tourbières. Dans un premier temps, l'analyse des macrorestes végétaux a permis de déterminer le contexte de formation des tourbières au cours de l'Holocène et d'évaluer la dynamique à long terme des communautés végétales en lien avec le climat et les feux survenus sur la tourbière. Ensuite, les reconstitutions paléohydrologiques basées sur l'analyse des thécamoebiens ont permis de détecter certaines influences climatiques régionales sur l'hydrologie des tourbières au cours de l'Holocène. Enfin, la dynamique à long terme du C a été évaluée et discutée en lien avec les changements écohydrologiques et les variations climatiques régionales.

Rôle du climat sur les changements écohydrologiques dans les tourbières de la Côte-Nord au cours de l'Holocène

Dans la région de la Côte-Nord, le contexte de formation des tourbières a été étroitement lié aux variations holocènes des niveaux marins relatifs. Les plus anciennes tourbières étudiées (~7500-5800 ans étal. BP) se sont formées sur les terrasses deltaïques sableuses (alt. 16-30 mètres a.s.l) édifiées par les rivières tributaires de la Mer de Goldthwait et qui furent émergées par le relèvement isostatique et la baisse du niveau marin relatif. Plus près du littoral, dans la portion inférieure des deltas (alt. 12-16 m.a.s.l.), certaines tourbières se sont formées après

4200 ans étal. BP sur des dépôts silto-argileux associés à la transgression marine laurentienne.

La paludification sur les sables deltaïques bien drainés suggère l'existence de conditions climatiques très favorables à l'accumulation de la tourbe bien que des facteurs édaphiques tels que l'induration des sols aient pu aussi jouer un certain rôle (Payette *et al.*, 2013). Le climat régional semble avoir affecté indirectement le rythme des transitions trophiques par son influence sur la productivité végétale (accumulation de la tourbe) au début de la formation des tourbières. La transition très rapide vers l'ombrotrophie dans les tourbières de Baie-Comeau a été favorisée par une accumulation rapide de la tourbe et indique que le climat régional était très propice à la croissance des sphaignes suite à l'émersion des terres. Dans les tourbières de Havre-St-Pierre, l'ombrotrophication a possiblement été retardée par la faible productivité végétale des communautés minérotrophes initiales, une indication que le climat était moins favorable à l'accumulation de la tourbe dans cette région.

Dans les deux régions étudiées, le climat maritime côtier caractérisé par un faible potentiel d'évapotranspiration et un apport élevé en humidité atmosphérique, a favorisé une croissance soutenue des sphaignes au cours du développement des tourbières. Dans ce contexte, les feux ont été peu fréquents et ont eu un impact limité sur la dynamique du couvert végétal dans les deux tourbières ombrotrophes étudiées (Plaine et Lebel).

Les reconstitutions paléohydrologiques ont permis de détecter certaines influences climatiques communes sur les tourbières des deux régions. La période cohérente de changement hydrologique à partir de ~3000 ans étal. BP coïncide avec la transition de l'optimum climatique holocène vers le Néoglaciaire qui a été documentée par plusieurs proxies terrestres dans le nord-est du Canada. Cependant, il y a aussi d'importantes différences entre les reconstitutions hydrologiques à l'échelle régionale qui suggèrent l'existence d'influences internes et de rétroactions écohydrologiques (Belyea et Clymo, 2001 ; Swindles *et al.*, 2012).

Rôle des variations climatiques sur la séquestration du carbone

Cette thèse a montré que les taux d'accumulation à long terme du C ont été beaucoup plus faibles dans les tourbières de Havre-St-Pierre, confirmant ainsi l'une des principales hypothèses de recherche. Les différences régionales des taux d'accumulation du C ont été particulièrement marquées au cours des 2000 dernières années. Ces résultats suggèrent que le climat régional a exercé une influence persistante sur la productivité primaire dans ces écosystèmes au cours de l'Holocène. Dans l'ensemble, l'accumulation plus faible du C dans les tourbières de Havre-St-Pierre peut résulter d'une productivité primaire moins élevée associée à des saisons de croissance plus courtes dans la région du Golfe du St. Laurent.

Les données présentées dans cette thèse suggèrent que le bilan à long terme d'accumulation du C a été affecté par une combinaison de facteurs autogènes associés à l'accumulation de la tourbe et de changements climatiques régionaux. Dans la région de la Côte-Nord, la séquestration du C a graduellement diminuée au cours du développement des tourbières. Les reconstitutions paléohydrologiques suggèrent que cette diminution n'est pas uniquement attribuable à un assèchement graduel des surfaces en lien avec le développement vertical autogène (Yu *et al.*, 2003b). Toutefois, il est possible que la croissance verticale des tourbières ait graduellement accrue la sensibilité des communautés biotiques aux conditions climatiques. L'effet du développement autogène sur la productivité végétale semble avoir été amplifié par le refroidissement du Néoglaciaire. Au cours des 3000 dernières années, il est possible que l'accumulation du C ait été affectée négativement par des saisons de croissance plus courtes et une influence accrue des phénomènes de gel.

Les divergences dans les variations d'accumulation du C entre les tourbières, même au sein de chaque région, indiquent que les processus internes ont exercé un important contrôle sur le bilan entre la production et la décomposition de la tourbe. Les données suggèrent que l'influence du climat sur l'accumulation du C s'est effectuée par l'intermédiaire d'interactions complexes entre la nappe phréatique, la végétation et les processus de production et de décomposition. L'analyse des processus contrôlant les variations d'accumulation du C dans les tourbières Lebel et Plaine suggère l'existence de mécanismes de rétroaction entre la productivité végétale et l'hydrologie de surface. Ces données appuient les modèles conceptuels et suggèrent que les tourbières sont des systèmes adaptatifs complexes (Belyea et Baird, 2006) continuellement influencés par une combinaison de facteurs internes et de forçages climatiques. Toutefois, les données de cette thèse indiquent que le bilan holocène d'accumulation de la tourbe et du C associé a été influencé par des conditions climatiques distinctes entre Baie-Comeau et Havre-St-Pierre. Afin d'arriver à dissocier les influences internes et externes, les travaux futurs devraient favoriser l'utilisation combinée de reconstitutions paleoenvironnementales à haute résolution et de modèles écohydrologiques du développement des tourbières.

Une sensibilité distincte des tourbières face aux changements climatiques entre l'estuaire et le golfe du St. Laurent

Les taux très élevés d'accumulation du C dans les tourbières de Baie-Comeau semblent avoir été favorisés par une grande stabilité des conditions hydrologiques au cours de leur développement. En contrepartie, la grande amplitude des variations hydrologiques dans les bogs de Havre-St-Pierre a affecté négativement le bilan d'accumulation du C. Ces résultats suggèrent que les tourbières de Havre-St-Pierre ont été plus sensibles aux variations climatiques passées. Ces tourbières ont subi d'important déficits hydrologiques au cours des 2500 dernières années qui ont possiblement été amplifiés par une influence accrue du gel saisonnier dans la tourbe lors du refroidissement Néoglaciaire tel que suggéré par l'étude de van Bellen, Garneau et Booth (2011).

Perspectives futures

Cette thèse permet d'accroitre les connaissances sur l'influence du climat holocène sur la dynamique des communautés végétales et la variabilité hydrologique des tourbières maritimes de l'est du Canada. Les données recueillies permettent d'améliorer la compréhension des processus contrôlant l'accumulation à long terme du C dans ces écosystèmes. Toutefois, il existe une complexité spatio-temporelle du développement à l'échelle de l'écosystème (Kohrola *et al.*, 1996 ; Ireland *et al.*, 2013 ; van Bellen *et al.*, 2011) qui n'est pas intégrée par notre approche basée sur l'utilisation d'une carotte par tourbière. De plus, le calcul des taux d'accumulation du C (LORCA) basé sur des carottes de tourbe centrales peut entrainer une surestimation des valeurs réelles pour l'ensemble de la tourbière (van Bellen *et al.*, 2011). Ainsi, des travaux futurs seront nécessaires pour comprendre l'influence des changements climatiques sur les processus écohydrologiques et l'accumulation du C à l'échelle du bassin des tourbières.

Dans l'est du Canada, la variabilité climatique holocène demeure peu connue, ainsi des données paléoclimatiques supplémentaires basées sur d'autres proxies biotiques terrestres et océaniques (e.g. grains de pollen, dynokystes, chironomides) seront nécessaires dans l'avenir. Ces données permettraient notamment de mieux évaluer quelles furent les principales variables climatiques influençant la dynamique à long terme des tourbières de la Côte-Nord. De plus, il serait essentiel d'évaluer le rôle de la variabilité climatique intra-annuelle en particulier l'influence des conditions hivernales sur le bilan net d'accumulation du C. Des travaux futurs sont nécessaires pour évaluer l'influence du gel saisonnier sur la dynamique écohydrologique et les processus de productivité végétale et de décomposition de la tourbe dans les tourbières boréales et subarctiques.

La compréhension des facteurs ayant contrôlé l'accumulation du C dans les tourbières maritimes de la Côte-Nord au cours de l'Holocène permet une meilleure évaluation du rôle actuel et futur de ces écosystèmes dans le cycle global du C. La région de la Côte-Nord a connu un important réchauffement climatique au cours des dernières décennies (Hutchison *et al.*, 2009). Les changements climatiques en cours dans cette région affecteront potentiellement les bilans hydriques, la productivité végétale, la décomposition aérobique et la séquestration du C dans les tourbières.

Les tourbières de Havre-St-Pierre sont particulièrement vulnérables aux changements climatiques en raison de leur position à la limite froide et humide de distribution des tourbières nordiques (Yu, Beilman et Jones, 2009). Dans ces régions, même de faibles variations des températures et des précipitations pourraient avoir des impacts significatifs sur la productivité végétale et la décomposition.

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