CONDITIONS DE L'ACCUMULATION DU CARBONE DANS UNE TOURBIERE DU QUEBEC MERIDIONAL : L'INFLUENCE DES FACTEURS AUTOGENES ET DES CONTROLES ALLOGENES

par

Serge D. MULLER

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Cette thèse intitulée

Conditions de l'accumulation du carbone dans une tourbière du Québec méridional : l'influence des facteurs autogènes et des contrôles allogènes

présentée et soutenue à l'Université de Montréal par

Serge D. Muller

a été évaluée par un jury composé des personnes suivantes :

Président du jury	Bhawan Singh, Professeur, Université de Montréal
Directeur de recherche	Pierre J.H. Richard, Professeur, Université de Montréal
Directeur de recherche	Jacques-Louis de Beaulieu, Directeur de recherches au CNRS, Université d'Aix-Marseille III
Examinateur externe	Serge Payette, Professeur, Université Laval
Membre du jury	Nigel T. Roulet, Professor, McGill University
Représentant du doyen de la FES	André Bouchard, Professeur, Université de Montréal

RÉSUMÉ

Conditions de l'accumulation du carbone dans une tourbière du Québec méridional : l'influence des facteurs autogènes et des contrôles allogènes

La tourbière de Mirabel (basses terres du St-Laurent, Québec méridional) a été étudiée dans le but de préciser l'influence exercée par les différents paramètres autogènes et allogènes sur l'accumulation postglaciaire du carbone. (1) Les dynamiques végétales régionales révèlent un contrôle climatique général, modulé localement par le contexte paléogéographique, la physiographie et les processus écologiques. (2) L'histoire postglaciaire du climat régional fut reconstituée par la méthode des analogues modernes, contrainte par les fluctuations du niveau du lac Hertel (Mont St-Hilaire). Les résultats obtenus témoignent d'un important réchauffement entre 13 000 et 11 000 cal. BP, d'un optimum thermique autour de 8000 cal. BP, et de deux périodes sèches (10 000-6500 et 5000-3000 cal. BP). (3) Les dynamiques postglaciaires internes de la tourbière furent indépendamment reconstituées en trois dimensions par l'analyse pluridisciplinaire de sept profils. Ces derniers montrent des successions végétales similaires, conformes au modèle de terrestrialisation, mais asynchrones. Ils témoignent en outre de conditions de surface très humides avant 6700 cal. BP, suivies de nombreuses fluctuations asynchrones sur l'ensemble du site. L'accumulation du carbone présente en revanche des tendances similaires dans tous les profils, caractérisées par de fortes accumulations avant 6200 cal. BP (moyennes de 7.1 à 19.1 $gC.m^{-2}.an^{-1}$) et de faibles accumulations depuis (moyennes de 2.7 à 6.3 $gC.m^{-2}.an^{-1}$). (4) Les différentes conditions postglaciaires du développement de la tourbière sont confrontées. L'accumulation du carbone paraît avoir été essentiellement contrôlée par les interrelations entre la physiographie (alimentation minérotrophe initiale) et les processus autogènes, qui ont progressivement isolé la végétation tourbigène de la nappe phréatique. Notamment, ni le climat, ni les feux ne semblent avoir joué de rôle significatif.

Mots-clés : Paléoécologie, tourbière, facteurs autogènes et allogènes, végétation tourbigène, hydrologie interne, accumulation du carbone, physiographie, climat, niveaux lacustres, Québec méridional.

ABSTRACT

Conditions of carbon accumulation in a southern Québec peat bog : the influence of autogenic factors and allogenic controls

Mirabel bog (St. Lawrence lowlands, southern Québec) was studied with the aim to specify the influences exerted by the different autogenic and allogenic parameters on postglacial carbon accumulation. (1) Regional vegetation dynamics reveal a general climatic control, locally modified by the palaeogeographical context, physiography and ecological processes. (2) The postglacial history of the regional climate was reconstructed by the modern analogues method, constrained by the water level fluctuations of Lac Hertel (Mont St-Hilaire). Obtained results provide evidence for an important climate improvement between 13 000 and 11 000 cal. BP, a thermal optimum around 8000 cal. BP and two dry periods (10 000-6500 and 5000-3000 cal. BP). (3) Postglacial internal dynamics were independently reconstructed in three dimensions by the multidisciplinary analysis of seven profiles. They show similar plant successions, conforming to the terrestrialisation model but were asynchronous. They moreover attest the occurrence of very humid conditions prior to 6700 cal. BP, followed by numerous asynchronous fluctuations throughout the site. By contrast, carbon accumulation presents similar trends in all profiles, characterised by high accumulation rates prior to 6200 cal. BP (averages from 7.1 to 19.1 $gC.m^{-2}.year^{-1}$) and low rates since this date (averages from 2.7 to 6.3 $gC.m^{-2}.year^{-1}$). (4) The different postglacial conditions of peatland development are confronted. Carbon accumulation appears to have been essentially controlled by the interplay between physiography (initial minerotrophic supply) and autogenic processes, which progressively isolated the peat-forming vegetation from groundwater. Notably, neither climate nor fire seem to have played a significant role.

Key-words: Palaeoecology, peatland, autogenic and allogenic factors, peat-forming vegetation, internal hydrology, carbon accumulation, physiography, climate, lake-levels, southern Québec.

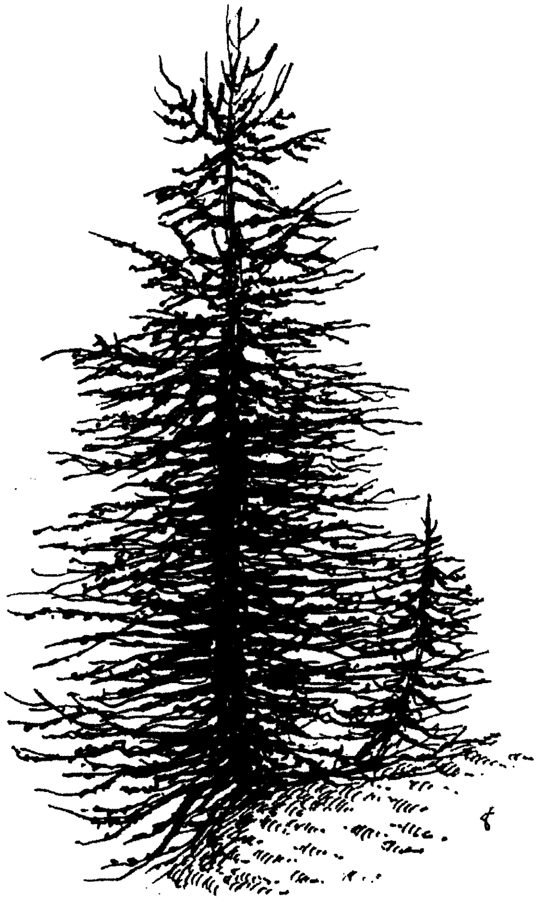
TABLE DES MATIÈRES

LISTE DES TABLES

LISTE DES FIGURES

LISTE DES SIGLES ET ABRÉVIATIONS

AMS :	Analyse Spectrométrique de Masse / *Mass Spectrometry Analysis*
BDPMQ :	Base de Données Pollinique et Macrofossile du Québec
cal. BP :	Années étalonnées avant aujourd'hui / *Calibrated years before present*
CH_4 :	Méthane / *Methane*
CO_2 :	Dioxyde de carbone / *Carbon dioxide*
cm, cm^3 :	Centimètre, centimètre carré / *Centimeter, square centimeter*
CMT :	Température du mois le plus froid / *Coldest month temperature*
E :	Evaporation réelle / *Actual evaporation*
et al. :	*et allii*
Fig., Figs. :	Figure, Figures / *Figure, Figures*
g :	Gramme / *Gram*
gC :	Gramme de carbone / *Carbon gram*
Gt :	Gigatonne / *Gigatonne*
GDD0 :	Degrés-jours de croissance au dessus de 0°C / *Growing degree-days above 0°C*
GDD5 :	Degrés-jours de croissance au dessus de 5°C / *Growing degree-days above 5°C*
km, km^2 :	Kilomètre, kilomètre carré / *Kilometer, square kilometer*
m, m^2, m^3 :	Mètre, mètre carré, mètre cube / *Meter, square meter, cubic meter*
mm, mm^3 :	Millimètre, millimètre cube / *Millimeter, cubic millimeter*
n.a. :	Non disponible / *Not available*
P :	Précipitations annuelles / *Annual precipitation*
PE :	Evaporation potentielle / *Potential evaporation*
PC :	Principal components / *Composantes principales*
r^2 :	Coefficient de corrélation / *Correlation coefficient*
°C :	Degré Celsius / *Degree Celsius*
% :	Pourcent / *Percent*



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AVANT-PROPOS

Ce mémoire est présenté sous forme de publications indépendantes et complémentaires, chacune représentant une étape de la thèse. La problématique générale de l'étude est définie en introduction, et les résultats obtenus sont synthétisés et discutés en conclusion.
La liste des publications, soumises ou en cours de soumission, est la suivante :

Muller, S.D. & Richard, P.J.H., accepté (2001). Postglacial plant migration in conterminous Montréal lowlands, southern Québec. *Journal of Biogeography.*

Muller, S.D., Guiot, J., Richard, P.J.H., Beaulieu, J.-L. de & Fortin, D., soumis-a (2001). Postglacial climate dynamics in the St. Lawrence lowlands, southern Québec: pollen and lake-level evidence. *Palaeogeography, Palaeoclimatology, Palaeoecology.*

Muller, S.D., Richard, P.J.H. & Larouche, A.C., soumis-b (2001). Postglacial development of a southern Québec bog: a spatio-temporal reconstruction based on pachymetry, sedimentology, micro- and macrofossils. *The Holocene.*

Muller, S.D., Richard, P.J.H. & Beaulieu, J.-L. de, soumis-c (2001). Postglacial carbon accumulation in a boreal bog (southern Québec). *Quaternary Research.*

Les contributions respectives des différents coauteurs sont les suivantes : Pierre J.H. Richard et Jacques-Louis de Beaulieu ont initié et supervisé l'ensemble du travail ; Joël Guiot m'a encadré pour la partie de la thèse qui concerne la reconstitution climatique ; David Fortin a réalisé l'analyse pollinique utilisée pour la reconstitution des niveaux lacustres ; enfin, Alayn C. Larouche m'a encadré et aidé lors de l'analyse macrofossile. Tous les coauteurs ont en outre contribué par de nombreuses relectures à l'amélioration des différents manuscrits.

Afin d'alléger le texte et d'éviter d'inutiles répétitions, les références bibliographiques ont été regroupées en fin de manuscrit.

INTRODUCTION

Dans l'hypothèse d'un futur réchauffement climatique associé à une augmentation de la concentration atmosphérique de CO_2, il apparaît important de mieux comprendre le cycle biogéochimique global du carbone (Houghton *et al.*, 1990, 1996 ; Sundquist, 1993). De nombreux travaux ont montré l'importance du rôle joué dans ce cycle par les écosystèmes terrestres en général (par exemple, Woodwell *et al.*, 1978 ; Peng *et al.*, 1995 ; Peng & Apps, 1997) et plus particulièrement par les tourbières (par exemple, Gorham, 1988, 1991 ; Malmer, 1992). Ces dernières, principalement concentrées dans les zones boréale et subarctique, renferment 455 Gt [1] de carbone (Gorham, 1991 ; Lappalainen, 1996), ce qui représente un tiers des quantités de carbone contenues dans l'ensemble des sols non tourbeux de la planète (1400 Gt [1] ; Post *et al.*, 1982) et l'équivalent du carbone atmosphérique (450 Gt [1] ; Houghton *et al.*, 1990). Leur énorme capacité de stockage à long terme du carbone confère ainsi aux tourbières une importance majeure dans le contrôle des échanges entre biosphère et atmosphère (Gorham, 1991 ; Malmer, 1992). L'influence du climat sur le développement des tourbières a été mise en évidence à plusieurs reprises (par exemple, Godwin, 1952 ; Damman, 1979 ; Barber, 1981 ; Frenzel, 1983). Leur comportement vis-à-vis d'un brutal réchauffement est cependant susceptible de dépendre de leur localisation géographique : les tourbières côtières et boréales tempérées devraient être affectées par la montée des eaux marines et l'assèchement du climat des moyennes latitudes, ce qui aurait pour conséquence le retour dans l'atmosphère de grandes quantités de carbone, tandis que les tourbières subarctiques et boréales nordiques devraient au contraire voir décuplée leur capacité d'accumulation, en particulier suite au recul du permafrost (Kettles & Tarnocai, 1999) . Toutefois, trop d'incertitudes concernent encore l'écologie des tourbières (Walker, 1961), et plus particulièrement la relation climat-tourbière (Houghton *et al.*, 1990 ; Moore *et al.*, 1998), pour permettre la prise en compte de ces écosystèmes dans les modélisations du stockage terrestre de carbone (Peng *et al.*, 1995, 1998).

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[1] Une gigatonne (Gt) correspond à 10^9 tonnes et à 10^{15} g.

Conditions de l'accumulation de la tourbe et du développement des tourbières

L'accumulation de la tourbe ou tourbigenèse, qui définit l'écosystème tourbière, est la résultante de trois processus: la production de matière organique par la végétation tourbigène (Clymo & Reddaway, 1971 ; Forrest & Smith, 1975), la respiration et la décomposition, extrêmement réduite du fait de la forte acidité et de la faible oxygénation du milieu (Svensson, 1980 ; Harriss *et al.*, 1985 ; Matthews & Fung, 1987 ; Clymo & Pierce, 1995 ; Bellisario *et al.*, 1998). L'accumulation traduit un déséquilibre entre ces processus, les quantités de carbone atmosphérique piégées sous forme de tourbe étant supérieures aux quantités relâchées sous forme de CH_4 et de CO_2 (Clymo, 1984). De nombreux paramètres influencent la tourbigenèse en contrôlant l'importance relative de la production et de la décomposition (Fig. 1). Ces paramètres sont le plus souvent regroupés en deux catégories : d'une part, les paramètres allogènes tels que le climat, le contexte géomorphologique, la morphométrie du site et les perturbations, et d'autre part, les paramètres autogènes tels que la succession végétale et l'hydrologie interne. Tandis que les paramètres autogènes sont par définition locaux, les paramètres allogènes peuvent exercer des influences locales, qui n'affectent qu'une partie du système – c'est le cas de la morphométrie du substrat et des perturbations – ou générales, affectant le système dans son entier – c'est le cas du climat et du contexte géomorphologique. La problématique de l'étude du développement des tourbières fut souvent posée en terme de dualité entre forces autogènes et contrôles allogènes (Frenzel, 1983 ; Payette, 1988 ; Foster & Jacobson, 1990 ; Foster & Wright, 1990 ; Korhola, 1992, 1996 ; Hu & Davis, 1995 ; Campbell *et al.*, 1997 ; Robichaud, 2000). L'influence des processus autogènes sur le développement des tourbières fut reconnue dès le début du siècle au travers de « l'hypothèse de la terrestrialisation [2] » (Weber, 1908) et de la « théorie de la régénération cyclique » (Post & Sernander, 1910 ; Osvald, 1923). Ces deux modèles ont été vérifiés dans certains cas : bon nombre de tourbières boréales se sont en effet initiées par terrestrialisation [2] et des successions cycliques autogènes furent observées à plusieurs reprises (Walker & Walker, 1961 ; Tolonen *et al.*, 1985 ; Payette,

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[2] Pour éviter les confusions avec le texte des publications, écrites en anglais, les termes *terrestrialisation* et *paludification* sont utilisés ici dans leur sens anglo-saxon. La *terrestrialisation* correspond au comblement d'un plan d'eau libre par le développement centripète de la végétation riveraine et la *paludification* correspond au développement de tourbières directement sur un sol minéral ou forestier (Tallis, 1983).

1988). Toutefois, d'autres travaux invalidèrent ces modèles en tant que processus directeur du développement des tourbières. Tallis (1983) décrit les nombreuses alternatives connues – dont la paludification [2] – à l'hypothèse de Weber (1908), qui montrent clairement que cette dernière ne représente qu'une possibilité parmi d'autres. L'observation de successions cycliques d'origine microclimatique (Aaby, 1976 ; Barber, 1981) ainsi que de phénomènes acycliques (Backéus, 1972 ; Aaby, 1976 ; Barber, 1981 ; Payette, 1988 ; Svensson, 1988) permirent de la même manière de réfuter le modèle des cycles de régénération.

Les études réalisées à l'échelle de l'Amérique du Nord illustrent bien la complexité des processus en œuvre dans le développement des tourbières. Damman (1979) mit en évidence l'influence du climat, en basant son analyse sur la correspondance entre les gradients de répartition des tourbières et les gradients climatiques (nord-sud, continental-maritime). Glaser & Janssens (1986) attribuèrent quant à eux les mêmes gradients de répartition des tourbières à l'action des processus autogènes, en considérant que les différences physionomiques entre tourbières reflétaient principalement des stades de développement différents. Leur analyse s'appuie sur de nombreuses datations basales, qui montrent que les tourbières méridionales et maritimes sont plus anciennes que, respectivement, les tourbières septentrionales et continentales.

Les facteurs autogènes

Les paramètres autogènes du développement des tourbières sont fortement liés au processus d'accumulation de la tourbe, dans lequel ils interviennent à la fois comme cause et comme conséquence (Fig. 1). Deux principales catégories de facteurs autogènes peuvent être distinguées : la succession végétale autogène (voir Walker, 1970 ; Tallis, 1983) et l'hydrologie interne (voir Ingram, 1983 ; Carter, 1986). Leurs fortes interrelations et les difficultés liées à leur mise en évidence ont souvent conduit les paléoécologues à ne pas les séparer et à les traiter comme une « boite noire ». Leur influence est alors généralement déterminée de manière indirecte et correspond à ce qui ne peut être expliqué par les contrôles allogènes (par exemple, Tolonen *et al.*, 1985 ; Hu & Davis, 1995 ; Lavoie *et al.*, 1995 ; Lavoie & Richard, 2000a, 2000b ; Robichaud, 2000).

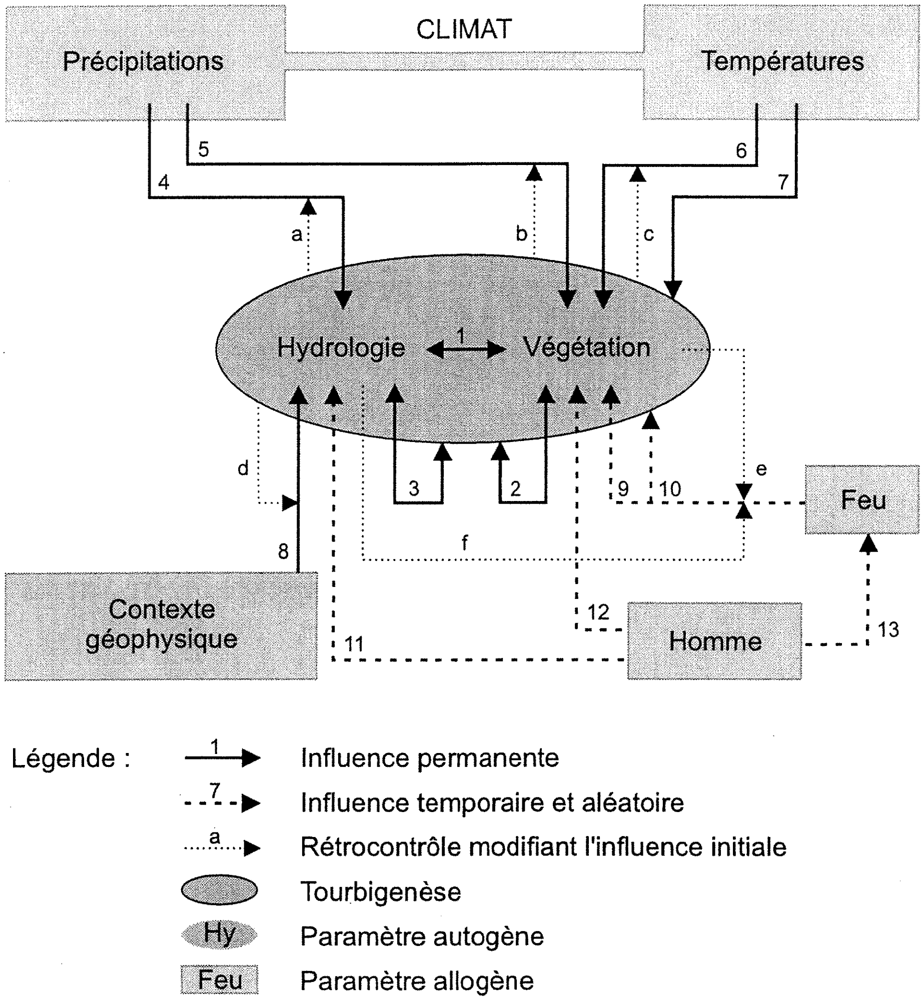

FIGURE 1. Modèle conceptuel des conditions de l'accumulation de la tourbe (tourbigenèse). Se référer au texte pour les détails.

La succession végétale autogène

La succession végétale autogène est définie comme un changement temporel cumulatif (directionnel) de la composition d'une communauté sous sa propre influence. Elle est opposée en ce dernier point à la succession végétale allogène, définie comme un changement non directionnel sous l'influence de facteurs externes au système (climat, physiographie, perturbations ...). Les effets de la succession autogène sur le développement des tourbières furent abordés à différentes échelles par Weber (1908), Post & Sernander (1910) et Osvald (1923). La « théorie de la régénération cyclique » (Post & Sernander, 1910 ; Osvald, 1923), qui concerne les tourbières ombrotrophes ou bogs [3], postule une succession à l'échelle microtopographique depuis les dépressions caractérisées par des espèces hydrophiles (par exemple, *Sphagnum cuspidatum*, *S. papillosum*, *Andromeda glaucophylla*) jusqu'aux buttes formées d'espèces plus xérophiles (par exemple, *Sphagnum fuscum*, *S. rubellum*, *Kalmia angustifolia*, *Eriophorum vaginatum*). Elle prévoit une alternance cyclique entre les deux microhabitats se traduisant par l'élévation en hauteur de la surface de la tourbière. Ce modèle fut vérifié dans quelques cas (Tolonen *et al.*, 1985 ; Foster & Wright, 1990 ; Payette, 1988). L' « hypothèse de la terrestrialisation [2] » (Weber, 1908) décrit quant à elle la succession végétale à l'échelle écosystémique sur la base des relations entre végétation et nappe phréatique : (1) stade limnique, en eau relativement profonde ; (2) stade telmatique, en eau peu profonde ; (3) stade semi-terrestre, inondé ou émergé selon un rythme saisonnier ; et (4) stade terrestre, toujours situé au dessus de la nappe phréatique. Cette succession reflète le concept de facilitation développé par Connell & Slatyer (1977) : chaque stade, de par l'accumulation de matière organique qu'il occasionne, modifie les conditions du milieu, en particulier l'hydrologie, et favorise ainsi l'installation et le développement du stade suivant (Fig. 1, voies 1 et 2). Cette succession idéale est toutefois souvent incomplète et des inversions peuvent avoir lieu sous l'emprise des facteurs externes (Tallis, 1983).

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[3] Les bogs sont définis comme des tourbières ombrotrophes, alimentées seulement par les eaux de pluies, et les fens comme des tourbières minérotrophes, alimentées à la fois par les précipitations, les eaux de ruissellement et la nappe phréatique (Gore, 1983).

L'hydrologie interne

L'hydrologie des tourbières se modifie de manière importante au cours de leur développement, et principalement au cours de la transition fen-bog. Le stade fen (minérotrophe) est caractérisé par une alimentation mixte par les eaux de ruissellement et les précipitations, et le stade bog (ombrotrophe) par une alimentation limitée aux eaux de pluies (Moore & Bellamy, 1974 ; Gore, 1983). Les fens présentent en outre une nappe phréatique affleurante au moins une partie de l'année (stade semi-terrestre), tandis que la surface des bogs, souvent constituée d'une mosaïque de buttes et de dépressions, est toujours située au dessus (stade terrestre). Comme la végétation, l'hydrologie des tourbières concerne des processus différents selon l'échelle à laquelle on se réfère. L'échelle microtopographique est caractérisée par un gradient hydrologique entre buttes et dépressions. Les buttes présentent un acrotelme (zone au moins temporairement oxique, située au-dessus de la position inférieure de la nappe phréatique ; Ingram, 1978) plus épais et une plus grande amplitude annuelle de la nappe phréatique que les dépressions (Sjörs, 1948 ; Malmer, 1962 ; Andrus *et al.*, 1983). Ce gradient est en partie responsable de la zonation végétale entre les deux microhabitats (Andrus *et al.*, 1983 ; Malmer, 1986 ; Fig. 1, voie 2).

A l'échelle de l'écosystème, les tourbières sont caractérisées par la persistance d'une nappe phréatique proche de la surface, et ce, malgré l'élévation en hauteur de cette dernière dans les bogs, sous l'effet de l'accumulation de tourbe. Ce phénomène fut d'abord attribué à l'action de forces de capillarité (Moore & Bellami, 1974), avant d'être relié à la très faible perméabilité de la tourbe (*groundwater mound hypothesis* ; Ingram, 1982). Les relations entre l'hydrologie interne et la tourbigenèse (Fig. 1, voie 3), abordées dans de nombreuses études (par exemple, Ingram, 1967 ; Boatman & Tomlinson, 1973 ; Damman, 1986, 1996), furent en outre explorées par différents modèles hydrologiques (Ingram, 1982 ; Winston, 1994 ; Hilbert *et al.*, 2000). La plupart de ces travaux ne distingue toutefois pas l'hydrologie interne de l'influence hydroclimatique. Il s'ensuit que l'influence exercée à l'échelle de l'écosystème par les processus hydrologiques autogènes, principalement liés à la perméabilité de la tourbe (dépendant de son degré de décomposition et de sa nature ; Ingram, 1983), est encore mal comprise.

Les contrôles allogènes

Le climat

La mobilisation scientifique mondiale sur le thème d'un futur réchauffement climatique (Houghton *et al.*, 1990, 1996 ; Laiho *et al.*, 1996) se reflète dans le nombre de travaux concernant l'influence du climat sur le développement des tourbières. A la suite des travaux pionniers de Weber (1908) et Granlund (1932), de nombreux chercheurs ont tenté de préciser le rôle joué par le climat, surtout par les précipitations, sur l'accumulation de la tourbe. L'influence des conditions climatiques sur la formation des tourbières fut récemment discutée par Svensson (1988), Korhola (1995a, 1996) et Halsey *et al.* (1998) : un assèchement climatique serait favorable à l'initiation par terrestrialisation[2], tandis qu'une augmentation du bilan hydrique favoriserait plutôt l'initiation par paludification[2]. L'influence du climat sur le développement postglaciaire des tourbières fut mise en évidence à de nombreuses reprises. Toutefois, si certaines d'entre elles semblent avoir répondu essentiellement aux variations hydroclimatiques passées (par exemple, Aaby, 1976 ; Barber, 1981 ; Winkler, 1988 ; Campbell *et al.*, 1997), d'autres paraissent avoir été également influencées par d'autres paramètres, tels que les facteurs autogènes (Payette, 1988), la topographie (Korhola, 1992 ; Mäkilä, 1997) ou les perturbations (Kuhry, 1994). Par ailleurs, quelques études suggèrent que les températures puissent avoir également joué un rôle important dans l'accumulation de la tourbe, principalement en contrôlant la productivité tourbigène (Almquist-Jacobson & Foster, 1995 ; Lavoie, 1998 ; Lavoie & Richard, 2000a, 2000b). Les principaux pôles d'influence du climat sur la tourbigenèse peuvent être résumés à trois voies principales : l'hydrologie interne (Fig. 1, voie 4), la végétation (Fig. 1, voies 5 et 6) et la décomposition (Fig. 1, voie 7).

La physiographie

L'influence de la physiographie n'a été, à ma connaissance, que peu abordée par les études portant sur le développement des tourbières. Heinselman (1970) a mis en évidence l'influence majeure de la géomorphologie sur la répartition et le développement des tourbières du lac pro-glaciaire Agassiz, dans le Minnesota. L'influence de la physiographie au travers du contrôle de l'hydrologie (Fig. 1, voie 8) fut également mise en évidence sur le développement des tourbières de l'ouest Canadien (Halsey *et al.*, 1998 ; Bauer, 2001). La topographie basale, qui correspond à la part du contexte géophysique en contact direct avec la tourbière, contrôle quant à elle l'expansion latérale des tourbières (Korhola, 1992 ; Almquist-Jacobson & Foster, 1995 ; Mäkilä, 1997). La nature et la topographie du substrat sont également susceptibles d'influencer leur mode d'initiation, notamment en permettant ou non l'existence d'un plan d'eau initial, et son entourbement éventuel (Lavoie, 1998).

Les perturbations

Les incendies naturels constituent un élément important du fonctionnement des écosystèmes boréaux (Wein & MacLean, 1983). S'ils n'exercent généralement que des influences locales et temporaires sur les dynamiques végétales des zones humides (Wein, 1983 ; Tolonen, 1985 ; Zoltai *et al.*, 1998 ; Fig. 1, voie 9), ils semblent influencer de manière beaucoup plus sensible l'accumulation de la tourbe (Fig. 1, voie 10), à court (Kuhry, 1994) et à long terme (Mäkilä, 1997 ; Robinson, 1998 ; Pitkänen *et al.*, 1999). Les feux ont également favorisé dans certains cas le processus de paludification [2] (Heinselman, 1975 ; Moore, 1975 ; Tallis, 1975 ; Wein, 1983 ; Tolonen, 1985). L'activité anthropique a souvent constitué une autre perturbation majeure pour les tourbières. Dans les zones cultivées, elle s'est généralement manifestée par la construction de canaux de drainage (Moore *et al.*, 1986 ; Fig. 1, voie 11), par la déforestation des zones entourbées (Warner, 1993 ; Fig. 1, voie 12) ou par l'augmentation de la fréquence des incendies (Caseldine & Hatton, 1993 ; Fig. 1, voie 13).

Vers un modèle synthétique du développement des tourbières

L'importance et la complexité des multiples interrelations entre les divers facteurs du développement des tourbières ont été notées par Heinselman (1963). Leur compréhension, prémisse indispensable à l'édification d'un modèle synthétique de l'accumulation de la tourbe, est toutefois encore loin d'être complète. Les relations entre paramètres autogènes (Fig. 1, voie 13) furent abordées par de nombreuses études écologiques, principalement sous l'angle des relations entre la végétation, la géochimie de l'eau et la nappe phréatique (par exemple, Jeglum, 1971 ; Vitt *et al.*, 1975a ; Glaser *et al.*, 1981 ; Andrus *et al.*, 1983 ; Rydin, 1985 ; Malmer, 1986 ; Vitt & Chee, 1990 ; Glaser, 1992a ; Anderson *et al.*, 1995 ; Belland & Vitt, 1995). Ces travaux ont démontré l'étroite interdépendance de ces différents paramètres. Les données recueillies ont en outre permis l'élaboration de deux modèles conceptuels appliqués au développement des bogs : celui de Clymo (1978, 1984), basé sur les propriétés de la tourbe, et celui d'Ingram (1982), qui développe l'aspect hydrologie. Ces deux modèles ont constitué la base théorique de plusieurs modèles de simulation (Winston, 1994 ; Korhola *et al.*, 1996 ; Hilbert *et al.*, 2000 ; Frolking *et al.*, 2001a, 2001b), dont le but est de tester des hypothèses par confrontation avec des données empiriques (voir Yu & Campbell, 1998). Les modèles de simulations constituent à ce titre des outils particulièrement intéressants pour étudier les modes d'action et les interrelations des différents paramètres de l'accumulation de la tourbe.

Toutefois, les dynamiques à long terme de ces interrelations sont encore relativement mal connues. Très peu d'études paléoécologiques se sont attachées à reconstituer l'ensemble des conditions postglaciaires du développement des tourbières. La plupart d'entre elles ont abordé dans le détail l'influence d'un paramètre particulier, comme le climat (Barber, 1981 ; Winkler, 1988 ; Kuhry *et al.*, 1992 ; Lavoie & Richard, 2000a, 2000b), la topographie basale (Korhola, 1992) ou les perturbations (Caseldine & Hatton, 1993 ; Kuhry, 1994 ; Pitkänen *et al.*, 1999). Le rôle joué par les processus autogènes, soit directement (Fig. 1, voies 12-14), soit sous forme de rétrocontrôles (définis comme des modifications de la sensibilité des tourbières aux contrôles allogènes sous l'effet de leur propre développement), a été beaucoup moins documenté. Les rétrocontrôles concernent en particulier les influences exercées par le processus d'accumulation de tourbe sur les contrôles climatiques de l'hydrologie (Fig. 1, voie a) et de la végétation (Fig. 1, voie b-c), et sur la sensibilité des tourbières au contexte physiographique (Fig. 1, voie d). Quelques travaux ont en outre mis en évidence des rétrocontrôles exercés par la végétation (Davis, 1979 ; Fig. 1, voie e) et l'hydrologie (Kuhry, 1994 ; Fig. 1, voie f) sur le régime et la sévérité des incendies.

Malgré les nombreuses incertitudes concernant les interactions entre facteurs allogènes et autogènes, quelques travaux de synthèse témoignent de la volonté d'intégration des données disponibles en un tout cohérent. Les modèles de Forrester (1961) et Wildi (1978) utilisent des paramètres non quantifiables ou trop nombreux qui rendent difficiles leur application et leur validation. Le modèle d'Almquist-Jacobson & Foster (1995) apparaît quant à lui comme le plus élaboré des modèles conceptuels, bien qu'il se limite à une représentation en deux dimensions. Ce modèle prend en compte les dynamiques internes des tourbières, déjà modélisées par Clymo (1978, 1984) et Ingram (1983), et les relie aux paramètres allogènes, tels que le climat et la physiographie.

Objectifs généraux et choix méthodologiques

La thèse présentée s'inscrit dans cette démarche synthétique, comme une tentative de reconstitution de l'ensemble des conditions du développement d'une tourbière du Québec méridional. La tourbière de Mirabel fut choisie pour cette étude, principalement en raison de sa taille moyenne (2,15 km^2) qui autorisait son étude spatiale détaillée et de sa proximité de la ville de Montréal qui permettait un important travail de terrain. En outre, les basses terres montréalaises et les zones collinéennes adjacentes, qui comprennent un grand nombre de sites ayant précédemment fait l'objet d'analyses polliniques, constituent une zone idéale pour réaliser de solides reconstitutions paléophytoécologiques et paléoclimatiques.

Les principaux objectifs de ce travail sont (1) de préciser l'influence respective de chaque paramètre (climat, physiographie, perturbations, succession végétale, hydrologie) sur le développement postglaciaire de la tourbière étudiée, (2) de dégager les interactions passées entre les différents paramètres, (3) de déterminer quels ont été les paramètres clés (facteurs contraignants et limitants) du développement de la tourbière et (4) de préciser les rétrocontrôles exercés par les paramètres autogènes.

Paléophytogéographie postglaciaire de la région de Montréal

L'histoire postglaciaire de la végétation est bien connue à l'échelle du Québec-Labrador (Richard, 1987, 1989, 1993, 1994a, 1995) et du Québec méridional (Richard, 1977a, 1994b ; Richard & Larouche, 1994 ; Richard *et al.*, 1992). D'importantes différences régionales ont été mises en évidence, et attribuées à l'influence des facteurs locaux, d'ordre physiographique et mésoclimatique (Richard, 1989, 1993). Le premier chapitre de la thèse précise l'influence de ces facteurs locaux sur la dynamique postglaciaire de la végétation dans la région étudiée (basses terres du St-Laurent), afin de définir le cadre phytogéographique du développement de la tourbière de Mirabel. Soixante-quatorze dates radiocarbone provenant de quinze sites (voir Muller & Richard, accepté) sont utilisées pour définir une chronologie postglaciaire homogène pour l'ensemble de la région. La corrélation de ces quinze analyses polliniques permet de préciser les patrons régionaux des migrations végétales postglaciaires. Ce travail permet en outre de corréler entre eux tous les sites utilisés dans la reconstitution climatique ultérieure et d'obtenir un cadre de référence fiable pour établir la chronologie du développement de la tourbière.

Climat postglaciaire des basses terres du St-Laurent

Aucune reconstitution climatique quantitative n'avait jusqu'à présent été réalisée dans le sud du Québec. Plusieurs travaux avaient permis d'y préciser les tendances climatiques générales (Richard *et al.*, 1992 ; Richard, 1994b ; Richard & Larouche, 1994 ; Lavoie & Richard, 2000c) et les patrons saisonniers (Carcaillet & Richard, 2000). Ces données qualitatives provenant de diverses sources (pollen, charbons, niveaux lacustres) sont toutefois délicates à utiliser dans l'étude des relations entre le climat et l'accumulation de la tourbe (Lavoie & Richard, 2000a, 2000b). Le deuxième chapitre de la thèse présente une reconstitution quantitative du climat à l'échelle des basses terres du St-Laurent. Ce travail est réalisé à l'aide de la méthode des analogues modernes, contrainte par une reconstitution de niveaux lacustres (Guiot, 1990 ; Guiot *et al.*, 1993 ; Magny *et al.*, sous presse). La méthode est appliquée à sept diagrammes polliniques et une moyenne est calculée pour chaque paramètre reconstitué afin d'établir un signal climatique régional et de réduire l'influence des paramètres locaux sur la reconstitution. Cette première application de la méthode des analogues modernes à une échelle régionale est finalement comparée aux résultats obtenus avec les données déjà disponibles pour le Québec, et avec les reconstitutions climatiques réalisées dans d'autres régions de l'est de l'Amérique du Nord.

Développement postglaciaire de la tourbière de Mirabel

L'influence et les interrelations des facteurs autogènes sont relativement peu abordées dans la littérature portant sur le développement à long terme des tourbières (Tolonen *et al.*, 1985 ; Foster & Wright, 1990 ; Payette, 1988 ; McCarthy *et al.*, 1995). Pourtant, l'analyse du contenu micro- et macrofossile de carottes de tourbe permet de reconstituer de façon indépendante les dynamiques végétales et les conditions hydrologiques qui ont prévalu tout au long de l'accumulation postglaciaire de la tourbe. Les données paléobotaniques (pollen, spores et macrorestes) sont ici utilisées pour reconstituer le développement passé de la végétation locale, tandis que les thécamoebiens (ou rhizopodes), les algues, les champignons et l'humification de la tourbe documentent les conditions hydriques locales (Moore, 1986). La teneur en matière organique de la tourbe est quant à elle évaluée par des mesures de perte-au-feu (Dean, 1974). Ces analyses sont réalisées sur sept profils sédimentaires prélevés dans la tourbière de Mirabel, basses terres du St-Laurent, Québec méridional. Le nombre et la localisation des profils sont déterminés en fonction de la végétation actuelle et de la topographie basale dans l'optique d'obtenir un enregistrement caractéristique des différentes unités fonctionnelles de la tourbière. La reconstitution en trois dimensions du développement de la tourbière et de l'accumulation de la tourbe offre une opportunité d'étudier les relations passées entre végétation, hydrologie et tourbigenèse à deux échelles spatiales, celle des communautés et celle du système. Les résultats obtenus seront discutés en relation avec les observations actuelles, qui mettent en évidence des liens importants entre les trois paramètres (par exemple, Malmer 1962, 1986 ; Jeglum, 1971 ; Vitt *et al.*, 1975a, 1975b ; Andrus *et al.*, 1983 ; Glaser, 1992a, 1992b ; Gorham & Janssens, 1992). En outre, l'approche multiscalaire doit permettre d'évaluer l'influence de l'échelle d'observation sur les reconstitutions paléoécologiques.

Conditions de l'accumulation du carbone dans la tourbière de Mirabel

Le rôle du climat sur le développement des tourbières au Québec méridional a fait l'objet des travaux récents de Lavoie (1998) et de Lavoie & Richard (2000a, 2000b). N'ayant mis en évidence aucune relation claire entre le régime des précipitations (déduit d'une reconstitution de niveaux lacustres ; Lavoie & Richard, 2000c) et l'accumulation de la tourbe, ces deux auteurs suggèrent un rôle majeur joué par les températures et les facteurs autogènes. Leurs études ne prennent toutefois pas en compte les autres facteurs susceptibles d'avoir joué un rôle, comme le feu ou la physiographie, ni les multiples interrelations entre les différents paramètres. Ce quatrième chapitre de la thèse complète sur ces points les résultats de Lavoie & Richard (2000a, 2000b) en examinant l'influence et les interrelations de l'ensemble des paramètres allogènes et autogènes : le contexte physiographique (Laverdière *et al.*, 1972), le climat régional (Muller *et al.*, soumis-a), la topographie basale (Muller *et al.*, soumis-b), les perturbations (feu, homme), la végétation locale, l'humidité de surface et l'accumulation du carbone (Muller *et al.*, soumis-b). Cette analyse sera réalisée en tenant compte des différents niveaux d'organisation (c'est-à-dire des gammes d'échelles) concernées par chacun des paramètres. Le climat ou la physiographie contrôlent en effet les processus à long terme, tandis que les paramètres locaux (perturbations, végétation locale, hydrologie interne) exercent des influences sur le court terme.

POSTGLACIAL VEGETATION MIGRATION IN CONTERMINOUS MONTRÉAL LOWLANDS, SOUTHERN QUÉBEC

Serge D. Muller & Pierre J.H. Richard

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Abstract

This work examines the postglacial plant migration patterns at the regional scale in the St. Lawrence lowlands and adjacent highlands, southern Québec. The postglacial plant dynamics in the St. Lawrence lowlands are reconstructed based on fifteen pollen diagrams and seventy-four radiocarbon dates. Migratory trends are shown for the first established plant formations (tundra, boreal forest and mixed forest), while the subsequent onset of deciduous forests shows no particular pattern. Several factors are involved in this contrasting behaviour. First, the palaeogeographic context, principally determined by the location of the ice sheet and Champlain Sea, played a major role during the initial stages. The Champlain Sea constituted a barrier to seed dissemination between the Appalachians and Laurentians, favouring the development of *Populus* instead of *Picea* in the latter region. Second, ecological processes partly determined differences in establishment between conifer-dominated forests and deciduous forests. The former resulted from the successive onset of dominant species (*Picea* and *Abies*), while the deciduous forest stages resulted from domination shifts between already established taxa. Third, physiographic traits explain local departures to the general migratory pattern. The major departure comprises the onset of tundra, boreal forest and mixed forest on the Monteregian hills (most of which constituted islands within Champlain Sea) before their establishment in Appalachians.

Key words: Plant migration, plant succession, regional scale, postglacial period, palaeophytogeography, pollen analysis, St. Lawrence lowlands, southern Québec.

Résumé

Migration végétale postglaciaire au travers des basses terres de Montréal, Québec méridional.

Ce travail examine les dynamiques de la migration végétale postglaciaire à une échelle régionale, dans les basses terres du St-Laurent et les hautes terres adjacentes, Québec méridional. Les dynamiques végétales postglaciaires sont reconstituées sur la base de quinze diagrammes polliniques et de soixante-quatorze dates radiocarbone. Les formations végétales pionnières (toundra, forêts boréale et mixte) sont caractérisées par des tendances migratoires claires, tandis que le développement des formations décidues ne présente aucun patron particulier. Plusieurs facteurs sont impliqués dans ce comportement contrasté. Premièrement, le contexte paléogéographique, principalement déterminé par la présence de l'Inlandsis laurentidien et de la Mer de Champlain, a joué un rôle majeur au cours des stades initiaux. La Mer de Champlain a constitué une barrière à la dissémination des graines entre les Appalaches et les Laurentides, favorisant le développement de *Populus* aux dépens de *Picea* sur ces dernières. Deuxièmement, les processus écologiques ont partiellement déterminé les différences dans le mode d'installation des forêts conifériennes et décidues. Le développement des premières a consisté en l'implantation successive des espèces dominantes (*Picea* et *Abies*), tandis que le développement des secondes a consisté en des changements de dominance entre taxons déjà présents. Troisièmement, les caractéristiques physiographiques expliquent certains écarts locaux au modèle migratoire général. Les principales exceptions sont constituées par l'installation précoce de la toundra, de la forêt boréale et de la forêt mixte sur les collines montérégiennes (qui, pour la plupart, constituaient des îles au sein de la mer de Champlain) avant leur implantation dans les Appalaches.

Mots clés : Migration végétale, succession végétale, échelle régionale, période postglaciaire, paléophytogéographie, analyse pollinique, Basses terres du St-Laurent, Québec méridional.

Introduction

Palaeophytogeographical studies undertaken at the scale of eastern North American (Bernabo & Webb, 1977; Davis, 1983; Delcourt & Delcourt, 1985; Gaudreau & Webb, 1985; Webb, 1988; Jackson *et al.*, 1997), at the scale of Canada (Ritchie, 1987) or Québec (Richard, 1994b) have demonstrated a general south-north postglacial plant migration pattern. This pattern corresponds roughly to the present-day north-south latitudinal plant zonation : tundra was followed by boreal forest, then by mixed forest, and deciduous forest finally succeeded in the southernmost glaciated area (Richard, 1977a, 1994b). However, Richard (1989, 1993) pointed out a number of local variations to this general model, especially during the Early Holocene. The deglaciation and subsequent marine and lacustrine episodes played a major role in vegetation dynamics by defining their physiographic context (Richard, 1989, 1994a). Moreover, pro-glacial waters influenced directly plant colonisation by creating geographical barriers to dispersion (Mott, 1978) and maintaining harsh climatic conditions just south of the ice-front (Gangloff *et al.*, 1971; Gangloff, 1981; Richard *et al.*, 1992; Richard, 1994a). In southwestern Québec, altitudinal variations related to the presence of the Appalachian and Laurentian highlands are also likely to have complicated the pattern of Holocene plant migration. These local climatic and physiographic influences can be studied by meso-scale approaches such as those of Liu (1990), Richard *et al.* (1992), Fuller (1997), Muller & David (2000) and Muller *et al.* (2000).

In the present study, we reconstruct Holocene plant migration at the regional scale, comprising the Montréal lowlands and adjacent highlands. The reconstruction is based on fifteen pollen diagrams from lakes and peatlands, chronological control being provided by seventy-four radiocarbon dates. Our objective is to answer the following questions : Is the postglacial migration pattern evident at continental and sub-continental scales, also present at the regional scale? What are the respective influences of the palaeogeographic context, physiography and ecological processes in vegetation dynamics?

Study area

Palaeogeography of deglaciation

Ice sheet retreat began to uncover south-western Québec (Fig. 2) around 15 000 cal. BP (Dyke & Prest, 1987). The first lands to emerge from the ice were the Appalachian highlands and the highest Monteregian hills (LaSalle, 1966). Following the ice regression northwards, the Montréal lowlands successively experienced the freshwater *Candona* Lake stage (before 13 800 cal. BP), the Champlain Sea episode (13 800 to 12 000 cal. BP) then the *Lampsilis* Lake stage (Fig. 2; Elson, 1969; Parent & Occhietti, 1988, 1999). The Champlain Sea period is characterised by deposition of widespread, albeit discontinuous, thick blue clays over the entire lowlands, a feature that deeply affected the drainage conditions. Outcropping bedrock is scant in the lowlands but sandy post-Champlain fluvial deposits are widespread along the ancient sea margins (Brown MacPherson, 1967). The Laurentian highlands were completely deglaciated after the St. Narcisse moraine episode, dated between ~13 200 and ~12 700 cal. BP and interpreted as a cooling event contemporary with the Younger Dryas (LaSalle & Elson, 1975). Due to isostatic rebound, the retreat of waters occurred in the lowlands between ~11 500 and ~6,500 cal. BP, shaping terraces in the lowlands until the emplacement of the present-day St. Lawrence River (Elson, 1982).

Environmental context

The Montréal lowlands (Fig. 3) mainly comprise a flat plain of marine clay, deposited above tills on a sandstone-dolomitic bedrock (Laverdière *et al.*, 1972). Several intrusive Monteregian hills emerge from the plain roughly along an east-west axis. Lakes are rare in the lowlands and, except for the now filled-up Lac Romer, they are located on the Monteregian hills. The regional vegetation constitutes the northern limit of deciduous forests dominated by *Acer saccharum* Marsh, in association with *Betula alleghaniensis* Britton on highlands, with *Tilia americana* L. in the north-eastern part of the plain and with *Carya cordiformis* (Wang.) K. Koch south-westwards of Sorel (Grandtner, 1966; Richard, 1987). The mature, mesic formations also comprise *Fagus grandifolia* Ehrh., *Pinus strobus* L., *Tsuga canadensis* (L.) Carr. and *Fraxinus pennsylvanica* Marsh (Grandtner, 1966), while

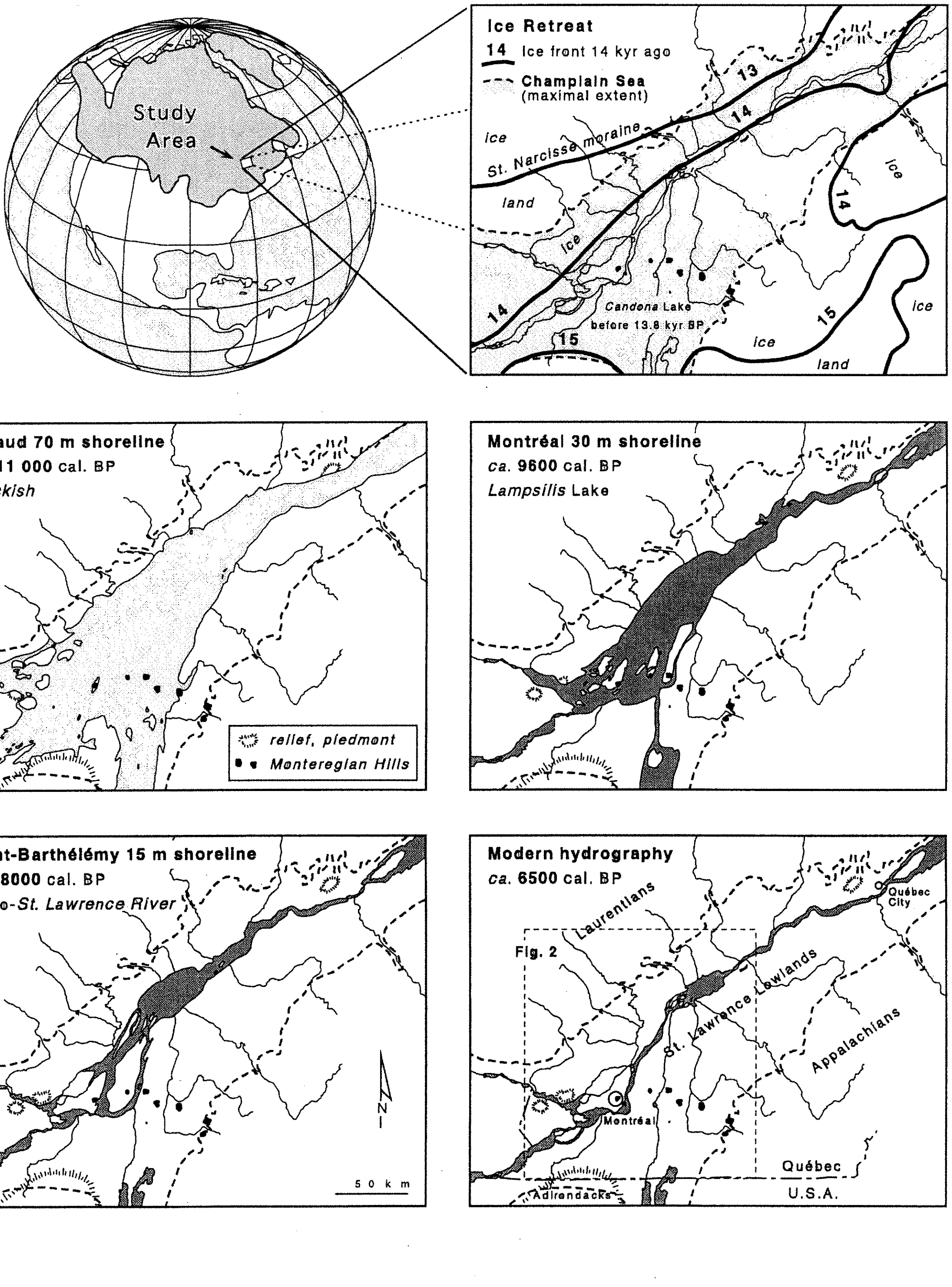
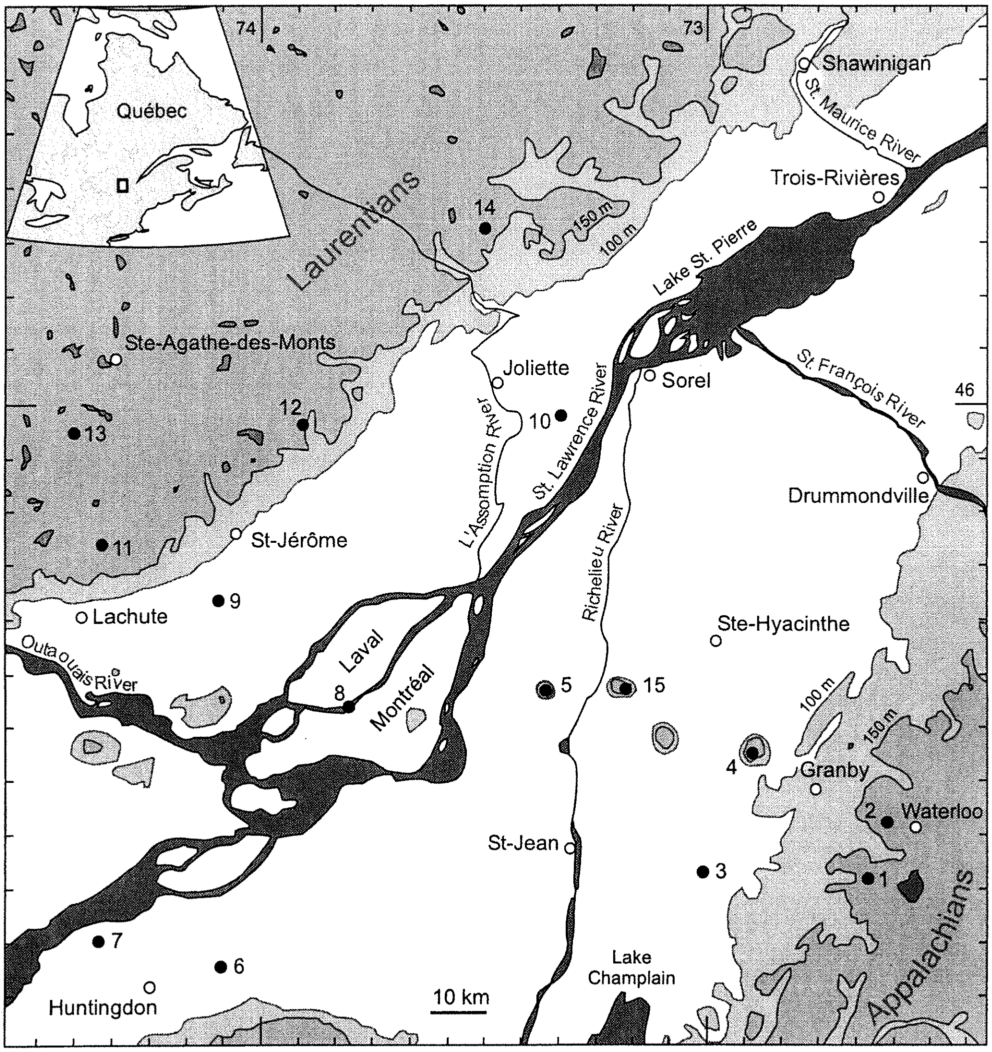


FIGURE 2. Location of the studied area on a full-glacial Globe, chronology of ice retreat, late-glacial marine invasion and postglacial lacustrine episodes (modified from Brown Macpherson, 1967).



1. Lac Bromont, BDPMQ
2. Shefford site, Richard, 1977, 1978
3. Farnham bog, Lavoie *et al.*, 1995
4. Lac Yamaska, BDPMQ
5. Lac-aux-Atocas, Gauthier, 1981
6. Ormstown bog, BDPMQ
7. Large Tea Field, Laframboise, 1987
8. Saraguay site, BDPMQ
9. Mirabel bog, this study
10. Lac Romer, Comtois, 1982
11. Lac Tania, Richard, 1977
12. Lac de St-Calixte, BDPMQ
13. Lac à St-Germain, Savoie & Richard, 1979
14. Lac de St-Gabriel, Richard, 1977
15. Lac Hertel, LaSalle, 1966

FIGURE 3. Study area and location of sites used in the vegetation reconstruction. The acronym BDPMQ points unpublished diagrams provided by the *Base de Données Polliniques et Macrofossiles du Québec.*

secondary facies are characterised by *Betula papyrifera* Marsh, *B. populifolia* Marsh, *Populus tremuloides* Michaux and *P. grandidentata* Michaux (Bergeron *et al.*, 1988; Brisson *et al.*, 1988). Depressions are occupied by peatlands which harbour boreal elements such as *Picea mariana* (Miller) BSP, *Larix laricina* (DuRoi) K. Koch and *Abies balsamea* (L.) Miller (Joyal, 1970; Jean & Bouchard, 1987). Finally, dry sites, located mainly on lowland sandy deposits and on highland summits, are dominated by *Pinus banksiana* Lamb., *P. strobus*, *Thuja occidentalis* L. and *Quercus rubra* L. (Bouchard & Maycock, 1978). The transitional nature of these forests was emphasised by Maycock (1961) and Bouchard & Maycock (1978), who noted the boreal and temperate elements of which they are constituted.

Material and methods

Pollen data and chronology

Fifteen pollen diagrams were used in this study (Table 1, Fig. 3). Fourteen of them were chosen for the quality of their chronological control (3-10 radiocarbon dates per site) and for their location within the Montréal lowlands region. Five diagrams are taken from previous publications (Richard, 1977a; P. Richard, 1978; Savoie & Richard, 1979; Comtois, 1982; Lavoie *et al.*, 1995), two from an unpublished M.Sc. thesis (Gauthier, 1981; Laframboise, 1987) and six from BDPMQ (*Base de Données Polliniques et Macrofossiles du Québec*). Moreover, the Lac Hertel diagram (LaSalle, 1966) presenting only a basal date was partially added to more precisely examine the early plant dynamic at Mont St-Hilaire. All previously published diagrams are available from the *North American Pollen Database* at the *National Geophysic Data Centre*, Boulder, Colorado. The eight unpublished diagrams and the partially published one from Lac Tania (Richard, 1977b) are presented in Figs. 4 to 12.

The seventy-nine available radiocarbon dates, among which five are rejected, are presented in Table 2. The majority of them were performed on bulk sediments (peat or gyttja). The potential for hard-water effect is relatively low in the area. However, sandstone-dolomitic rocks outcrop at several places within the lowlands, the marine clay may locally contain significant amounts of carbonates and some carbonatites occur on

TABLE 1. Site characteristics

Site name	Site code	Nature	Longitude	Latitude	Altitude (m)	Size (ha)	References
Bromont	Brom1	Lake	-72.67	45.26	133	46	BDPMQ
Shefford	She	Carr	-72.58	45.36	282	4.7	P. Richard, 1977a, 1978
Farnham	Farn1	Bog	-72.98	45.28	53	200	Lavoie *et al.*, 1995
Yamaska	Yam	Lake	-72.87	45.45	265	2	BDPMQ
Atocas	Ato	Lake	-73.31	45.54	114	1.2	Gauthier, 1981
Hertel	Hertel	Lake	-73.15	45.55	173	33.5	LaSalle, 1966
Ormstown	Orm1	Bog	-73.88	45.08	75	100	BDPMQ
Large Tea Field	Ltfl	Bog	-74.20	45.12	50	550	Laframboise, 1987
Saraguay	Sar83	Carr	-73.75	45.50	23	0.01	BDPMQ
Mirabel	Mir1	Bog	-74.07	45.68	75	215	This study
Romer	Rom	Carr	-73.33	45.96	18	6	Comtois, 1982
Tania	Tania	Lake	-74.30	45.77	305	2	Richard, 1977b; this study
St-Calixte	Cal	Lake	-73.87	45.96	261	1.7	BDPMQ
St-Germain	Ger	Lake	-74.37	45.94	473	5	Savoie & Richard, 1979
St-Gabriel	Gabriel	Lake	-73.47	46.27	250	3	Richard, 1977a

Note: Sites noted BDPMQ were studied by P.J.H. Richard and are provided by the *Base de Données Polliniques et Macrofossiles du Québec.*

Monteregian hills. These calcareous substrates could induce some hard-water effects, especially for basal dates in contact with inorganic deposits. Notably, the date UQ-749 of Lac Bromont is located on the top of a clayey layer deposited by the late-glacial discharge of an inlet. Its age, older than the deglaciation of the site, is consequently thought to reflect a redeposition or a contamination by old carbonates. The basal date of the Large Tea Field sequence (Beta-14373) is also obviously too old for post-Champlain Sea deposits, which could reveal another contamination by old carbonates. Two AMS dates, in contradiction with dates on bulk sediments, are rejected for Mirabel bog: the date TO-8168 presented abnormal recovery yield revealing a treatment problem (R. Beukens, personal communication) and the date TO-8169 was rejected on the basis of palynostratigraphic correlations with Mer Bleue bog, located in the St. Lawrence lowlands near Ottawa (P.J.H. Richard, unpublished data). Finally, the date I-8786 at St-Gabriel is clearly

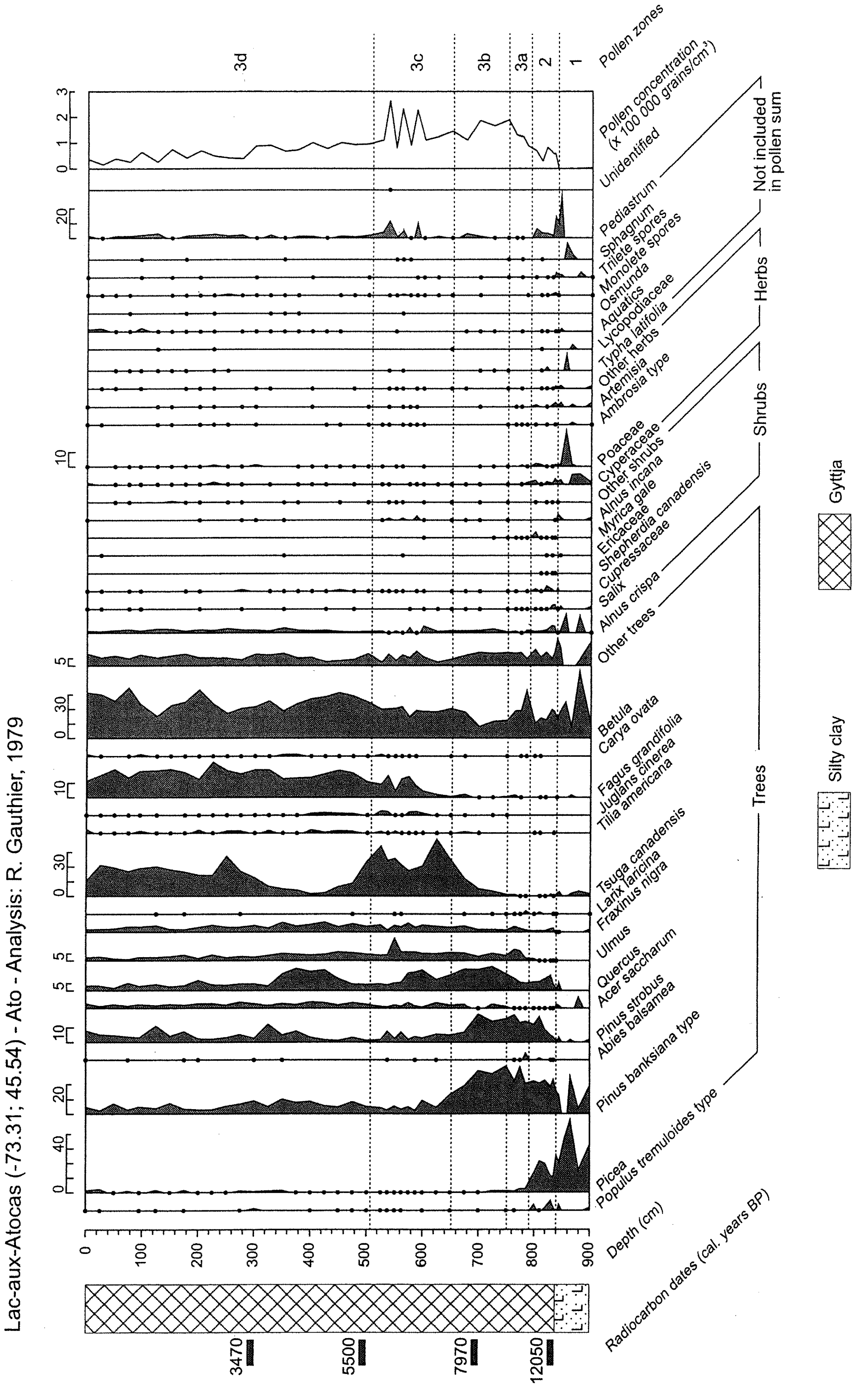


FIGURE 4. Lac-aux-Atocas. Simplified pollen percentage diagram. Dots represent less than 1% values.

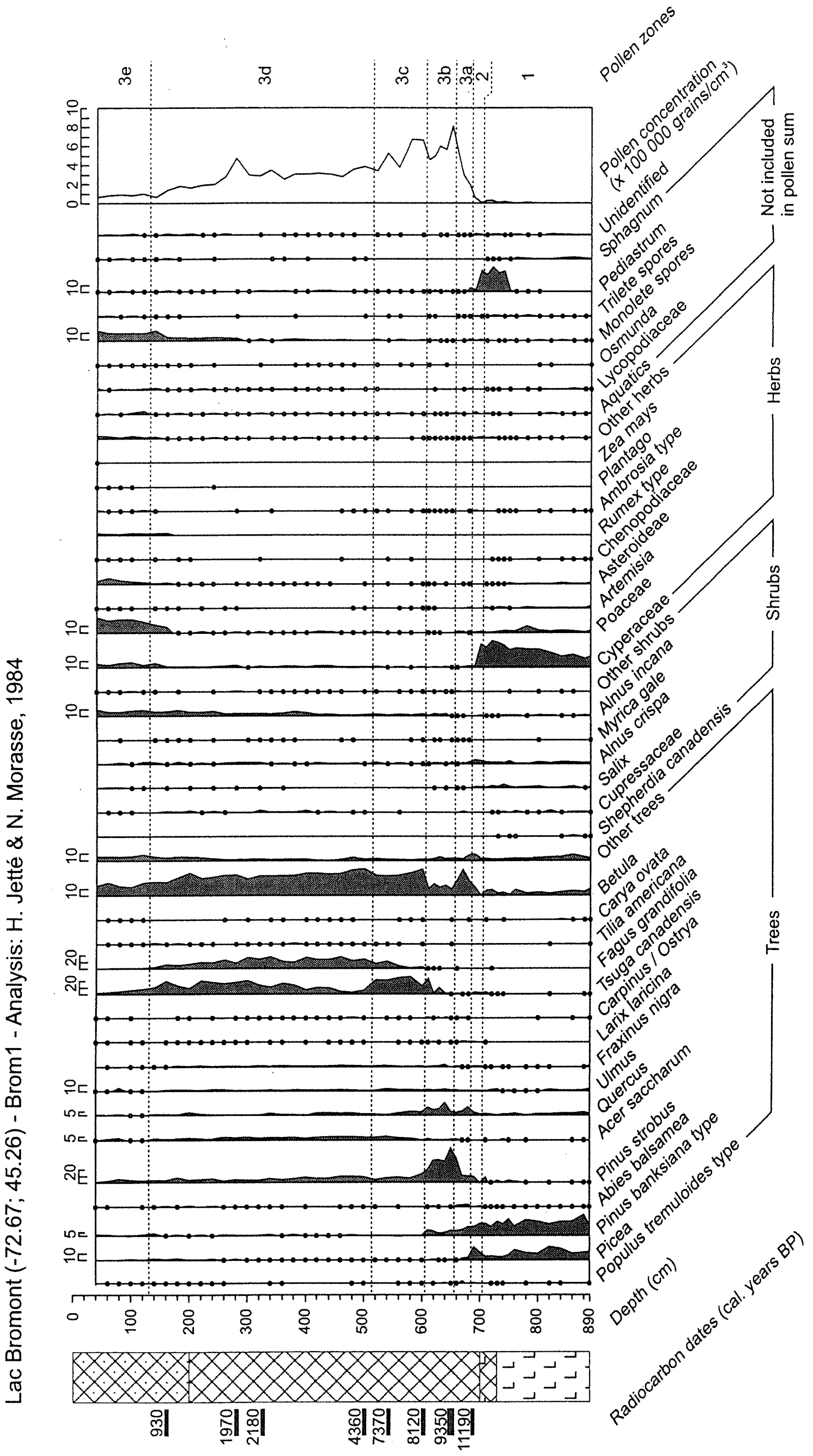


FIGURE 5. Lac Bromont. Simplified pollen percentage diagram. Dots represent less than 1% values.

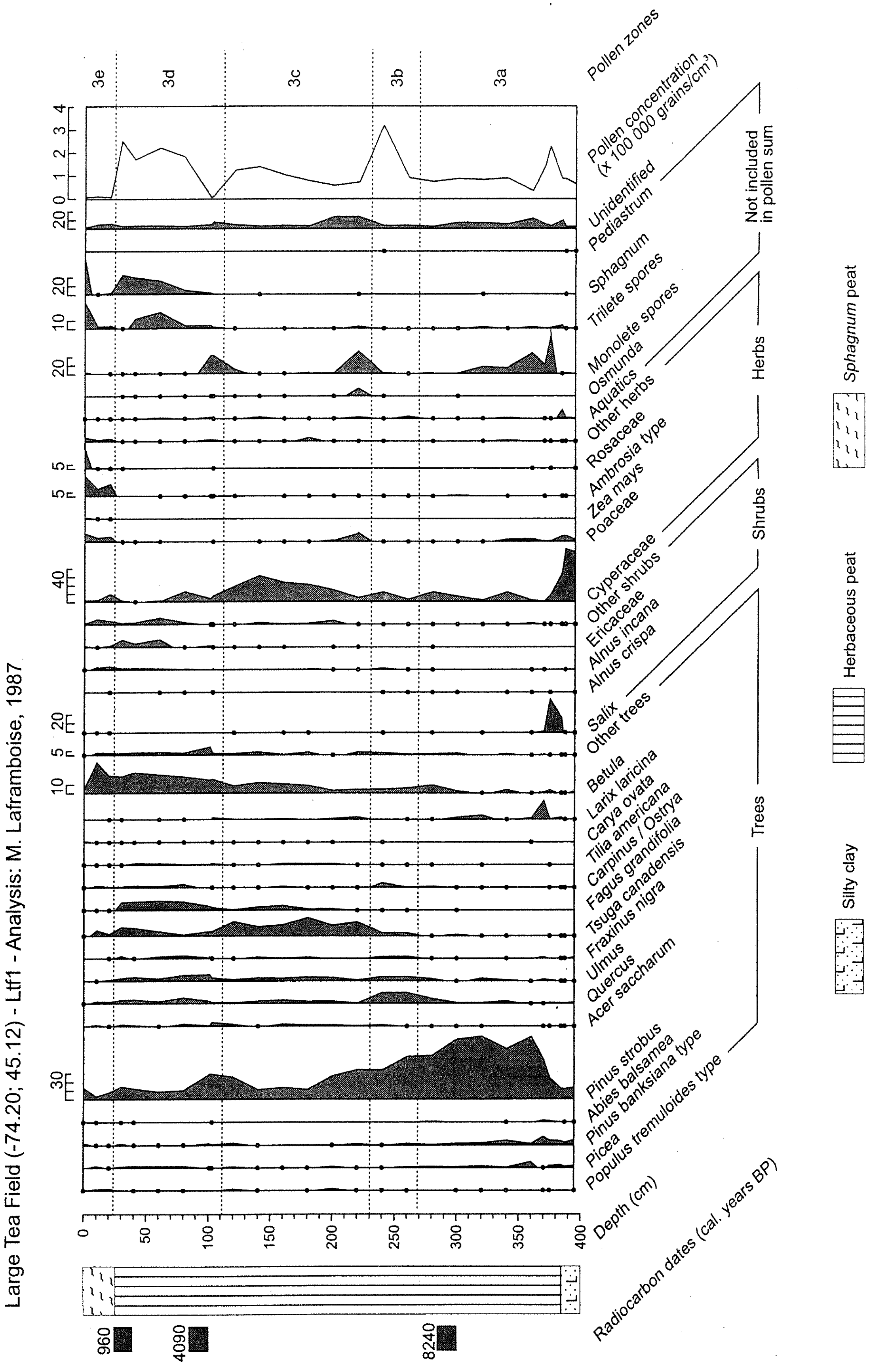


FIGURE 6. Large Tea Field. Simplified pollen percentage diagram. Dots represent less than 1% values.

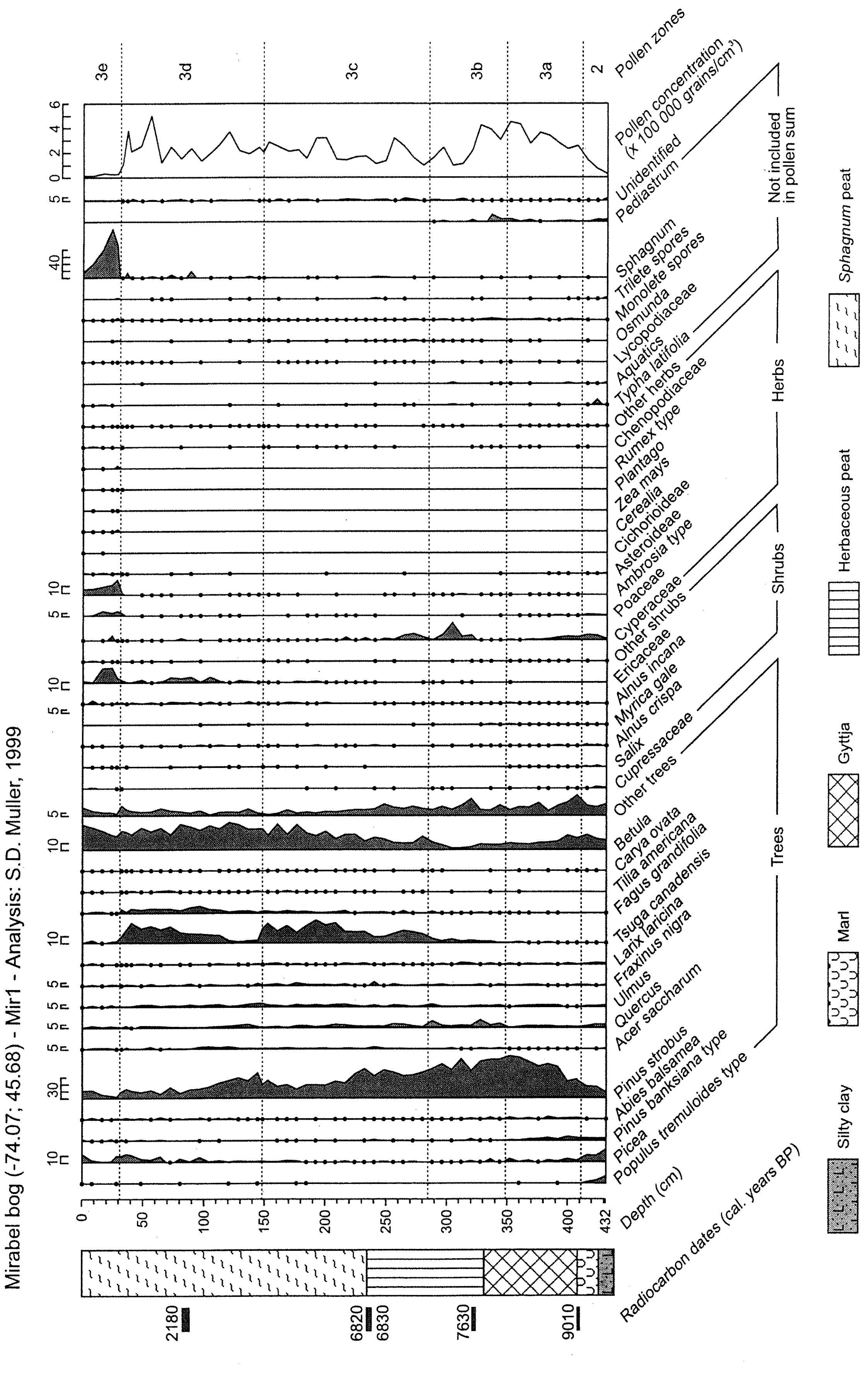


FIGURE 7. Mirabel bog. Simplified pollen percentage diagram. Dots represent less than 1% values.

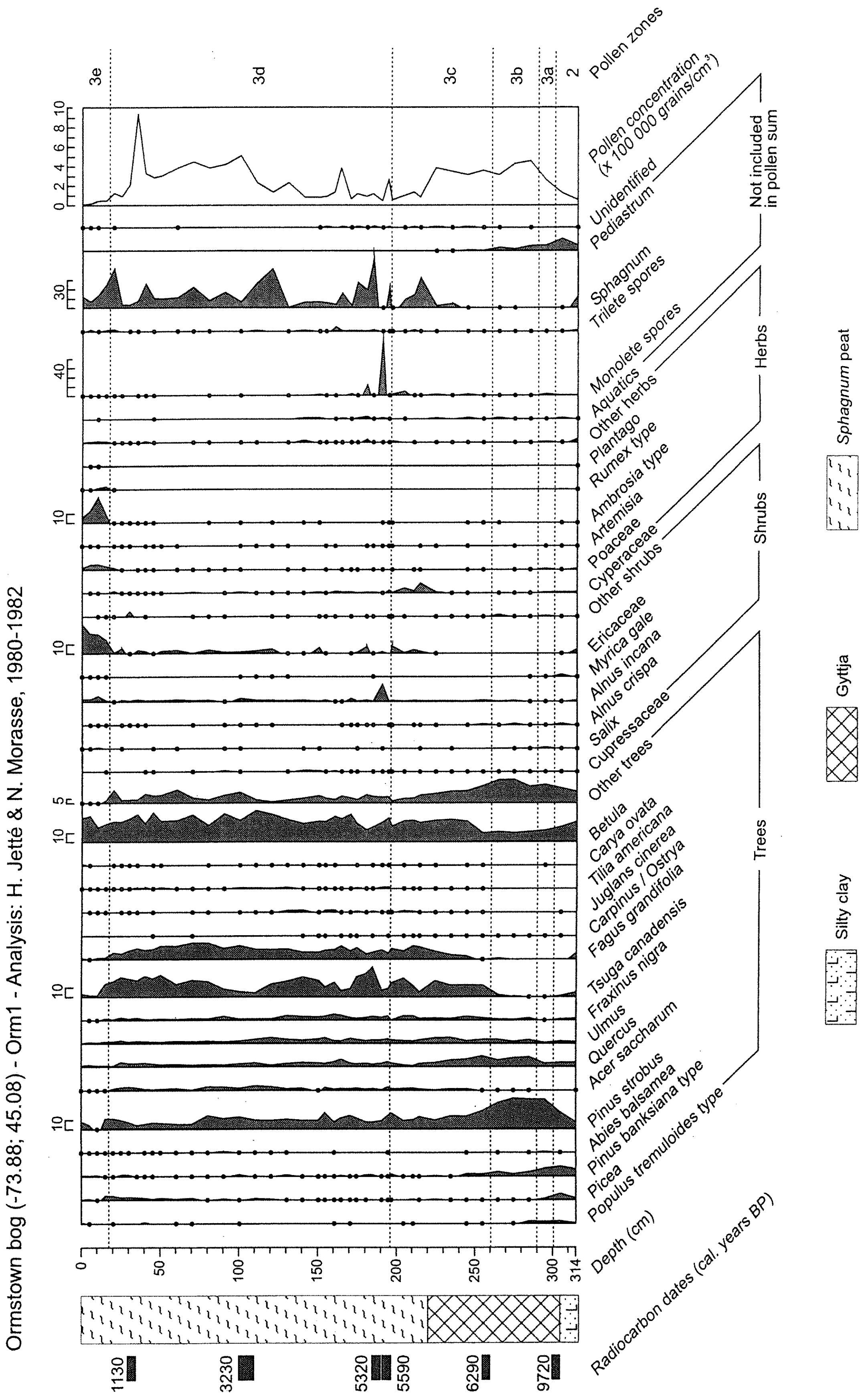


FIGURE 8. Ormstown bog. Simplified pollen percentage diagram. Dots represent less than 1% values.

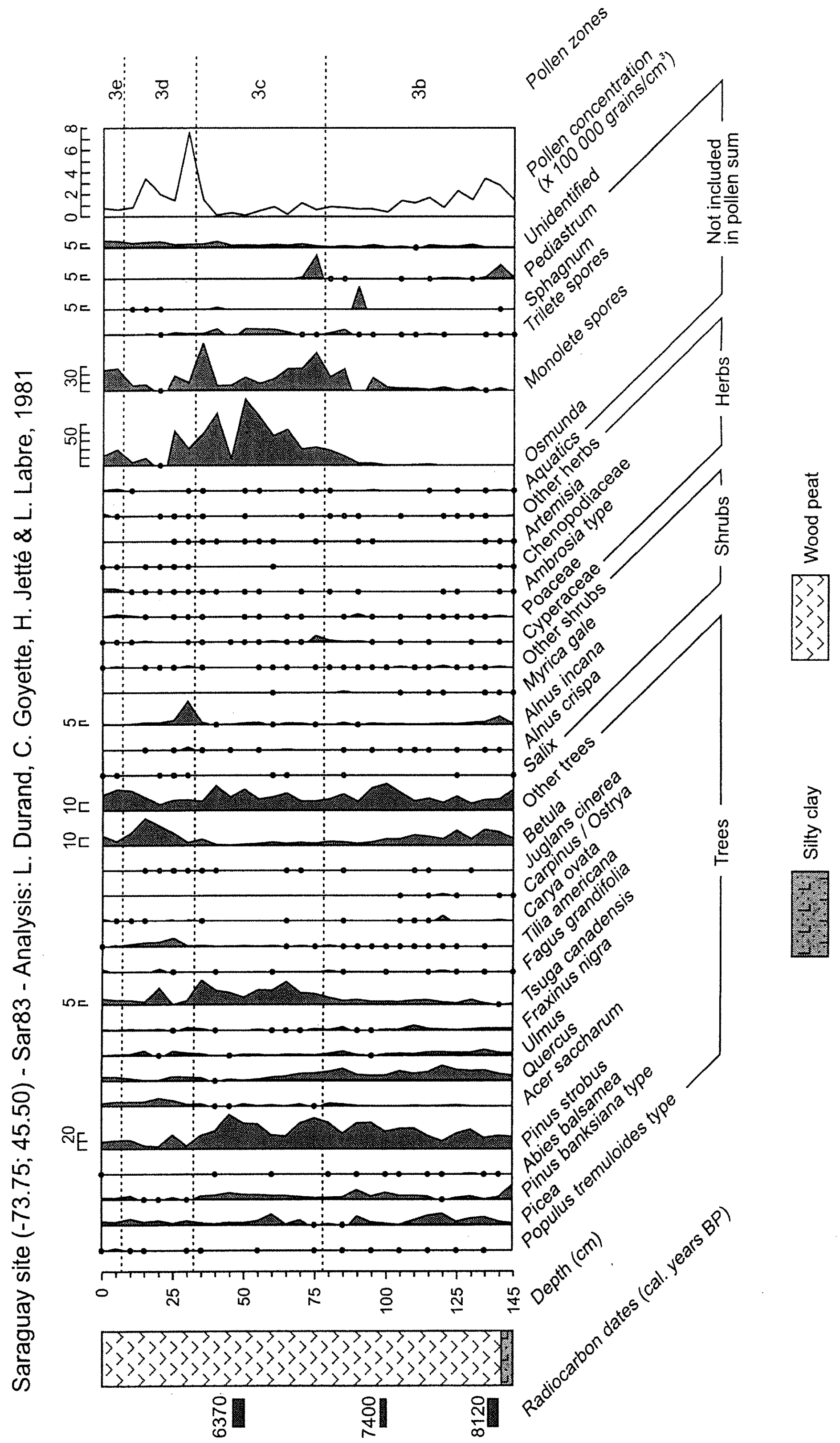


FIGURE 9. Saraguay site. Simplified pollen percentage diagram. Dots represent less than 1% values.

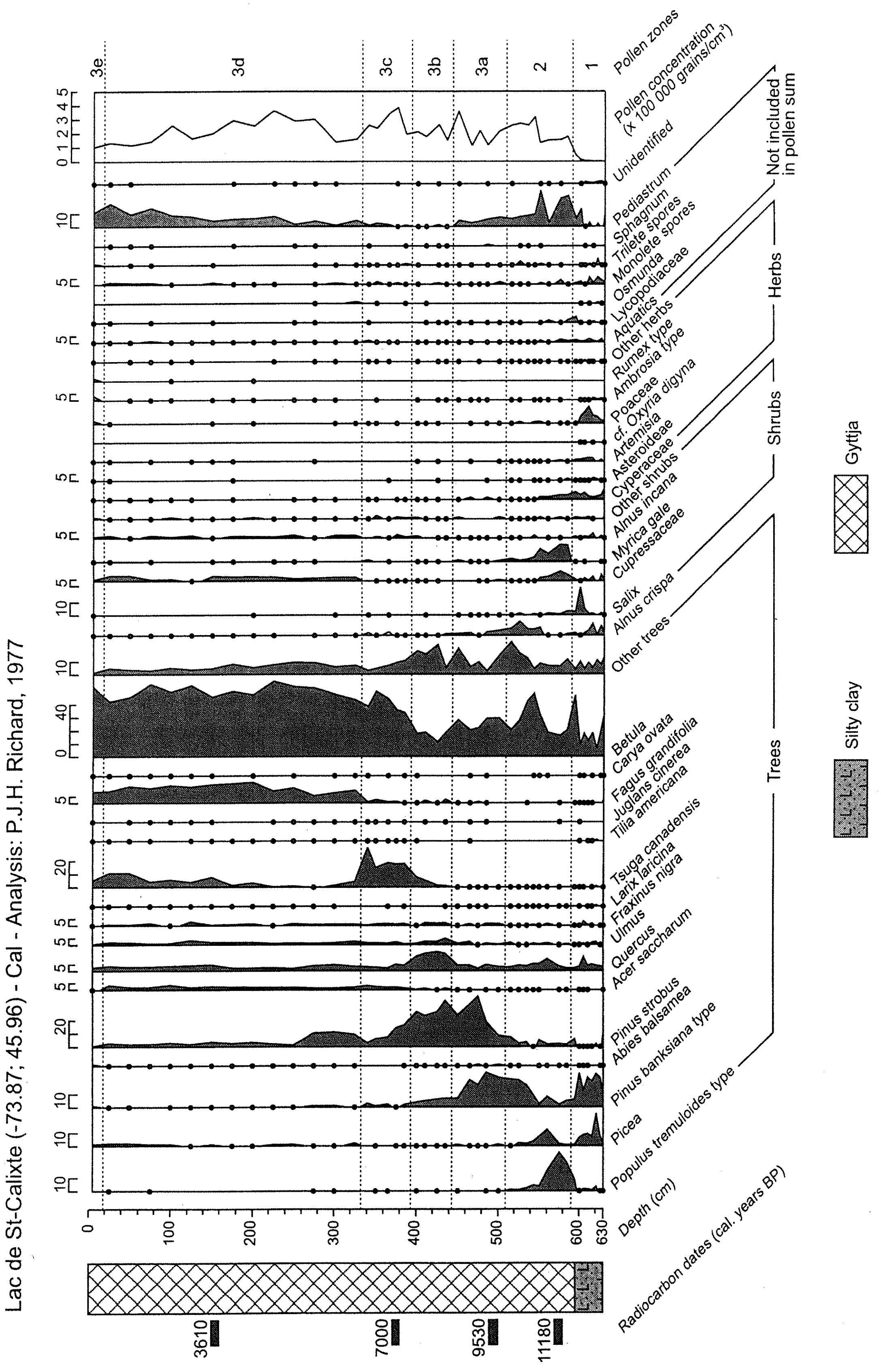


FIGURE 10. Lac de St-Calixte. Simplified pollen percentage diagram. Dots represent less than 1% values.

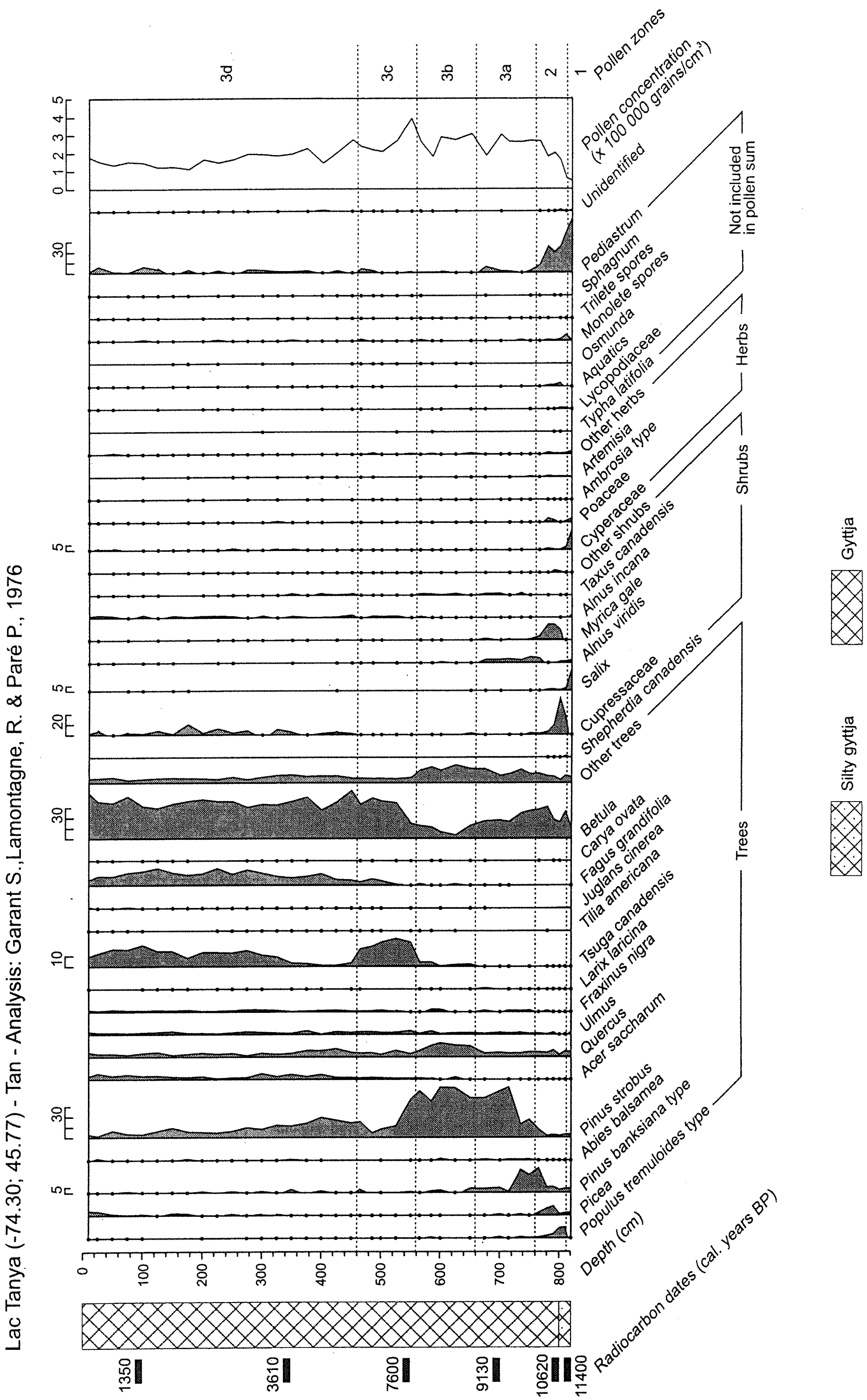


FIGURE 11. Lac Tania. Simplified pollen percentage diagram. Dots represent less than 1% values.

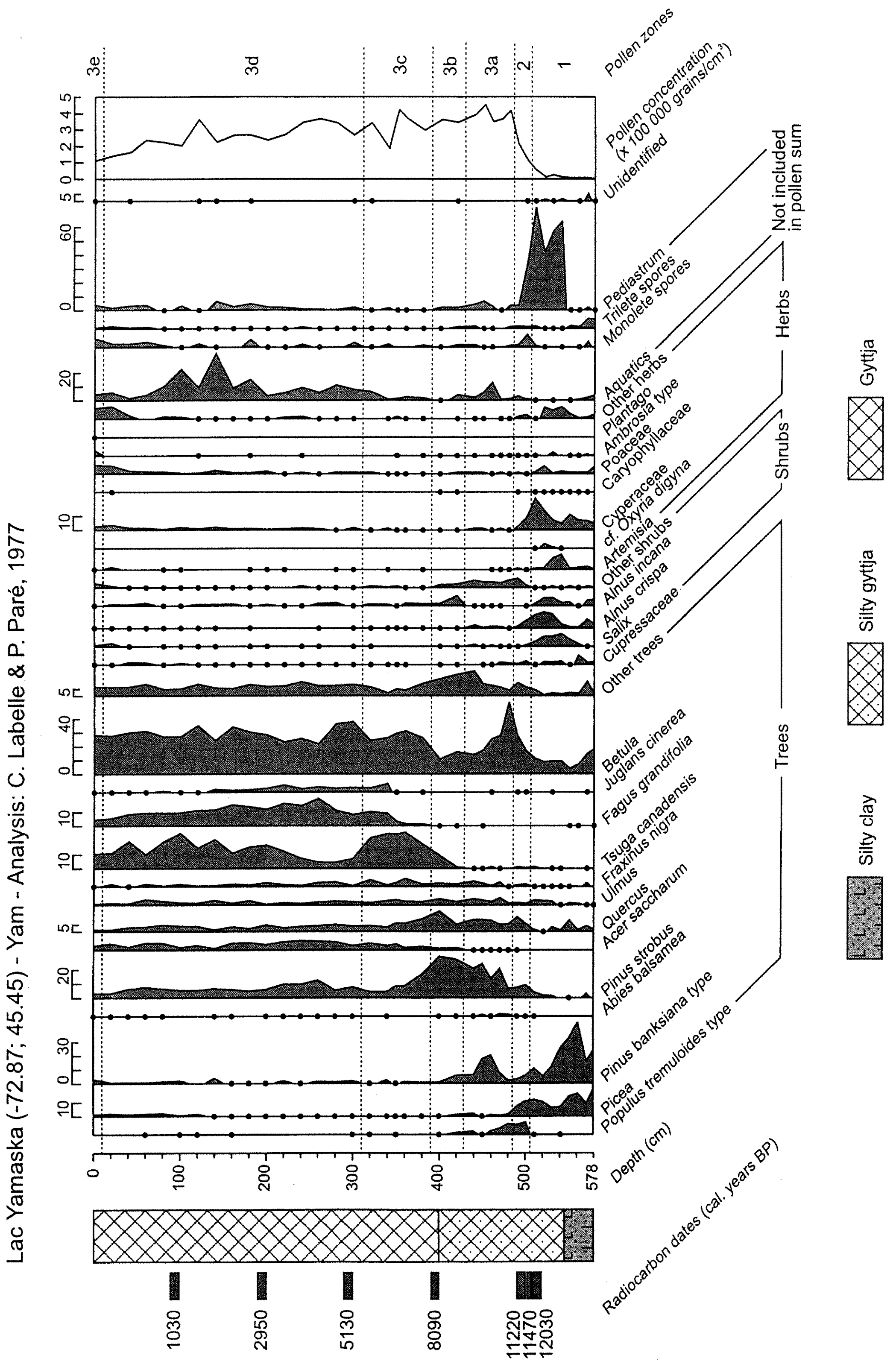


FIGURE 12. Lac Yamaska. Simplified pollen percentage diagram. Dots represent less than 1% values.

TABLE 2. Conventional and calibrated radiocarbon ages. Ages in italics are rejected (see age-depth models for details). All dates were performed on bulk sediments, either gyttja or peat, except the dates TO-8166, TO-8167, TO-8168, TO-8169 (Mirabel bog), which were performed by AMS on terrestrial plant macrofossils.

Site name	Depth (cm)	Conventional age	Calibrated age	Confidence interval 2σ	Laboratory code	Dated material
Atocas	290-300	3275 ± 95	3470	3720-3270	I-11190	Gyttja
	490-500	4790 ± 95	5500	5710-5310	I-11191	Gyttja
	690-700	7180 ± 120	7970	8270-7750	I-11192	Gyttja
	825-835	10250 ± 150	12050	12810-11260	I-11193	Gyttja
Bromont	160-165	1020 ± 200	930	1310-560	UQ-664	Gyttja
	280-285	2020 ± 70	1970	2150-1820	Beta-10529	Gyttja
	325-330	2210 ± 120	2180	2490-1900	UQ-662	Gyttja
	500-504	3910 ± 110	4360	4790-3990	Beta-10530	Gyttja
	540-545	6440 ± 140	7370	7580-7010	UQ-663	Gyttja
	600-605	7330 ± 240	8120	8590-7670	UQ-660	Gyttja
	645-655	8320 ± 180	9350	9630-8780	Beta-10531	Gyttja
	685-690	9770 ± 160	11190	11690-10640	Beta-8269	Gyttja
	690-700	*13300 ± 450*	*15980*	*17130-14310*	*UQ-749*	*Gyttja*
Farnham	77-83	5145 ± 200	5910	6310-5470	UQ-67	Peat
	102-108	6545 ± 110	7430	7610-7250	UQ-68	Peat
	125-130	7260 ± 160	8090	8390-7750	UQ-69	Peat
	173-178	8150 ± 150	9060	9490-8600	UQ-69	Peat
Hertel	760-790	10880 ± 260	12920	13430-12150	GSC-482	Gyttja
Large Tea Field	25-39	1060 ± 60	960	1160-800	Beta-15990	Peat
	86-100	3730 ± 70	4090	4350-2870	Beta-15991	Peat
	285-300	3450 ± 90	8240	8410-8030	Beta-15992	Peat
	374-384	*10570 ± 210*	*12740*	*13010-11700*	*Beta-14373*	*Peat*
Mirabel	82-88	2200 ± 50	2180	2340-2060	Beta-135343	Peat
	81-90	*3380 ± 120*	*3630*	*3960-3360*	*TO-8169*	*Macrofossils*
	235-237	6010 ± 160	6830	7250-6450	TO-8167	Macrofossils
	232-238	6050 ± 80	6820	7180-6680	Beta-135344	Peat
	321-323	*5550 ± 220*	*6310*	*6850-5910*	*TO-8168*	*Macrofossils*
	318-325	6800 ± 80	7630	7790-7510	Beta-135345	Peat
	408-410	8050 ± 170	9010	9470-8430	TO-8166	Macrofossils
Ormstown	30-35	1200 ± 60	1130	1270-970	UQ-586	Peat
	100-110	3030 ± 60	3230	3380-3000	UQ-591	Peat
	184-194	4620 ± 90	5320	5590-4990	Beta-8278	Peat
	190-196	4830 ± 90	5590	5740-5320	UQ-585	Peat
	254-259	5480 ± 130	6290	6530-5940	UQ-583	Gyttja
	299-304	8760 ± 180	9720	10240-9430	UQ-582	Gyttja
Romer	50-60	620 ± 70	600	670-510	Wis-1058	Gyttja
	240-250	3860 ± 70	4250	4500-4020	Wis-1057	Gyttja
	310-320	4520 ± 80	5140	5450-4870	Wis-1056	Gyttja
	430-440	6920 ± 80	7740	7930-7590	Wis-1055	Gyttja
Saraguay	47-50	5570 ± 50	6370	6450-6280	UQ-655	Peat
	98-100	6410 ± 110	7400	7560-7030	UQ-658	Peat
	136-139	7320 ± 120	8120	8380-7870	UQ-744	Peat

TABLE 2. Continued.

Site name	Depth (cm)	Conventional age	Calibrated age	Confidence interval 2σ	Laboratory Code	Dated material
Shefford	65-75	970 ± 110	920	1170-670	Qu-515	Peat
	195-205	2200 ± 90	2180	2350-1950	Qu-516	Peat
	244-250	4350 ± 95	4870	5290-4650	I-8834	Gyttja
	300-305	4445 ± 155	5040	5580-4620	I-8835	Gyttja
	350-355	5570 ± 145	6370	6720-5990	I-8836	Gyttja
	420-425	6345 ± 380	7270	7940-6320	I-8837	Gyttja
	460-465	8715 ± 215	9640	10240-9280	I-8838	Gyttja
	505-515	11100 ± 230	13130	13780-12640	I-8839	Silty gyttja
	543-550	11170 ± 230	13150	13800-12650	I-8840	Silty gyttja
	565-575	11400 ± 340	13410	14980-12660	I-8841	Silty gyttja
St-Calixte	150-160	3350 ± 60	3610	3810-3460	Wis-938	Gyttja
	370-380	6135 ± 75	7000	7250-6760	Wis-940	Gyttja
	490-500	8565 ± 90	9530	9710-9330	Wis-939	Gyttja
	570-580	9755 ± 100	11180	11340-10760	Wis-937	Gyttja
St-Gabriel	95-100	1860 ± 200	1820	2330-1330	I-8788	Gyttja
	375-380	3340 ± 175	3620	4070-3160	I-8787	Gyttja
	485-490	*3360 ± 125*	*3610*	*3900-3350*	*I-8786*	*Gyttja*
	645-650	7605 ± 380	8390	9460-7670	I-8785	Gyttja
	700-705	7965 ± 220	8920	9470-8350	I-8784	Gyttja
	725-733	9105 ± 175	10240	10690-9700	I-9038	Gyttja
St-Germain	270-280	3990 ± 145	4430	4840-3990	Gx-5229	Gyttja
	490-500	8060 ± 210	9010	9520-8410	Gx-5230	Gyttja
	658-668	10005 ± 280	11400	12080-10640	Gx-5232	Gyttja
	700-712	10420 ± 430	12520	13180-10760	Gx-5233	Gyttja
Tania	90-100	1490 ± 200	1350	1860-970	I-10188	Gyttja
	340-350	3360 ± 190	3610	4090-3080	I-10189	Gyttja
	540-550	6760 ± 300	7600	8170-7010	I-10190	Gyttja
	690-700	8180 ± 140	9130	9490-8650	I-10191	Gyttja
	790-800	9380 ± 130	10620	11110-10240	I-9278	Gyttja
	810-820	10000 ± 195	11400	12350-10870	I-9279	Silty gyttja
Yamaska	90-100	1135 ± 60	1030	1230-930	Wis-979	Gyttja
	190-200	2850 ± 65	2950	3210-2780	Wis-980	Gyttja
	290-300	4510 ± 75	5130	5450-4870	Wis-981	Gyttja
	390-400	7265 ± 85	8090	8280-7880	Wis-982	Gyttja
	490-500	9840 ± 100	11220	11550-11090	Wis-983	Gyttja
	500-505	10040 ± 95	11470	12260-11230	Wis-985	Gyttja
	505-515	10260 ± 100	12030	12760-11420	Wis-987	Gyttja

inconsistent with the presumed age of the *Tsuga* event (see below) and gives almost the same age as the date I-8787 located 90 cm above. The conventional radiocarbon dates are calibrated with the CALIB 4.0 program, by using the INTCAL98 dataset (Stuiver & Reimer, 1993). Calibrated dates, noted cal. BP (calendar years before present), are used to construct age-depth models (Fig. 13). The same method is used for all curves; it consists of interpolating the simplest curve connecting dates within the 2-sigma confidence intervals. Additionally, we used two dates for regional events considered to be synchronous throughout the study area, i.e. the pathogen-related *Tsuga* decline at ~5500 cal. BP (Davis, 1981) and the *Ambrosia* rise resulting from human deforestation and dated within the study area between 250 and 200 cal. BP.

Field and laboratory methods

All sites except Lac Hertel were analysed in the laboratoire Jacques-Rousseau (Université de Montréal) between 1975 and 1999: field and laboratory methods were similar for all. Lake profiles were collected with a modified Livingstone corer (Livingstone, 1955; Wright, 1967), and peatland profiles with a Russian sampler (Jowsey, 1966). Pollen extraction followed the conventional protocol according to Faegri & Iversen (1989). Pollen identifications were generally based on the pollen atlases of Richard (1970) and McAndrews *et al.* (1973). A total of 149 pollen taxa were identified. Pollen percentages are calculated on a sum excluding aquatic species and Pteridophyta spores (Berglund & Ralska-Jasiewiczowa, 1986). The minimum number of pollen grains included in this sum is 500 for all sites. The notation of pollen taxa follows Birks & Birks (1980) : the mention "type" groups several taxa not distinguishable by their pollen and the mention "cf." is used to point the most probable taxa. Diagrams were constructed with the GPalWin computer program (Goeury, 1997).

Pollen diagram zonation

Pollen zones are defined on sets of neighbouring pollen assemblages with similar frequency distribution (Moore *et al.*, 1991; Faegri & Iversen, 1989). They correspond to the following stages in postglacial vegetation dynamics (adapted from Richard, 1977a, 1987, 1994b):

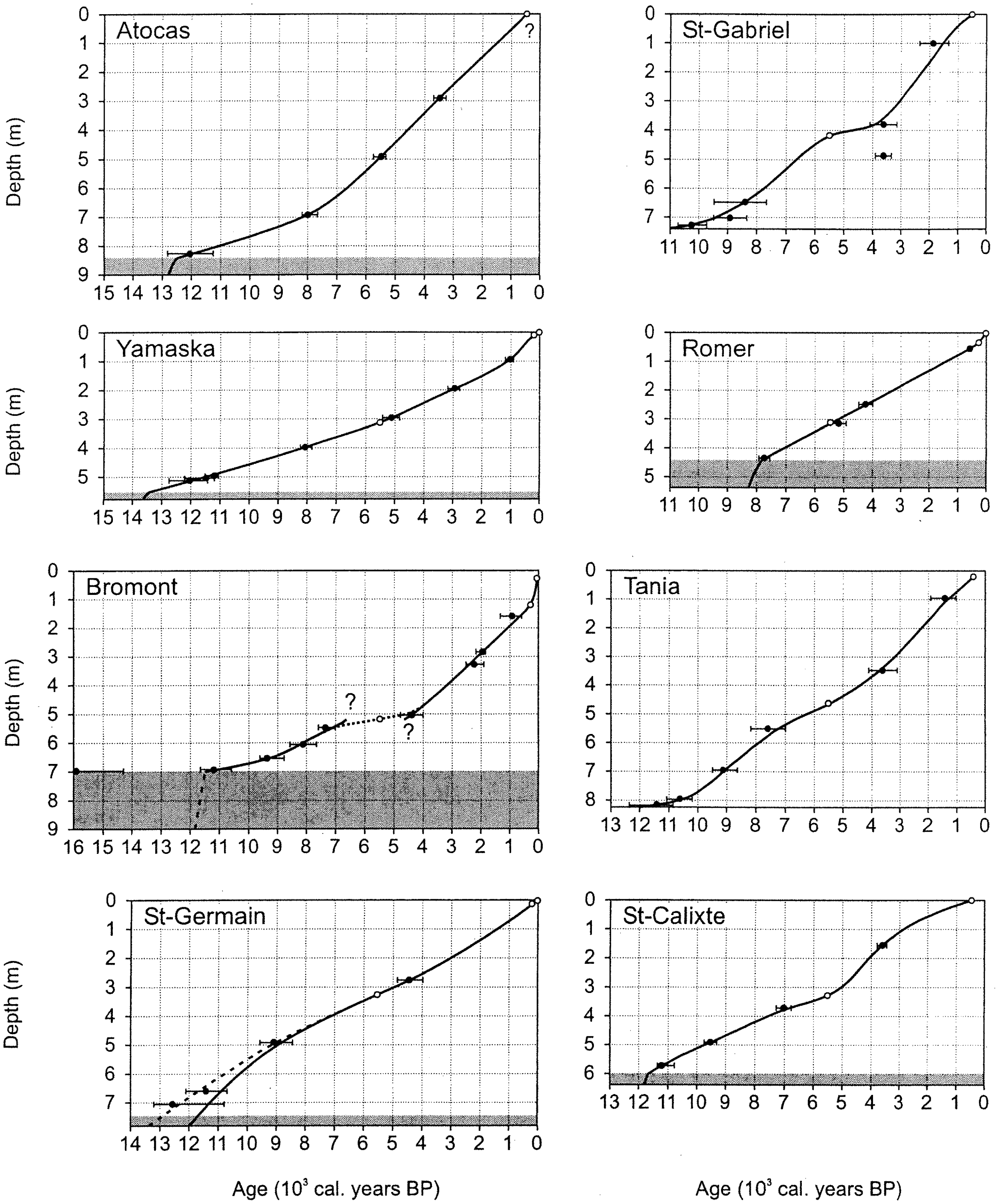


FIGURE 13. Age-depth models of regional diagrams. Error bars correspond to 2-sigma confidence intervals (95.4 %). White dots represent regional synchronous events (*Tsuga* decline and *Ambrosia* rise). Grey zones delimit the lower inorganic deposit (silty clay or sand) from the overlying organic layers (gyttja). Compared to the closest sites, the initial age of St-Germain (dashed line, Savoie & Richard, 1979) appears too old. To uniformise the regional pollen signal, we used a new model within the 2-sigma confidence intervals. The Lac Bromont presents a strong decrease in accumulation rate between ~7300 and ~4400 cal. BP. This feature, non confirmed by higher pollen concentration values (Fig. 4), is thought to reveal a hiatus of sedimentation for an undetermined period between these dates.

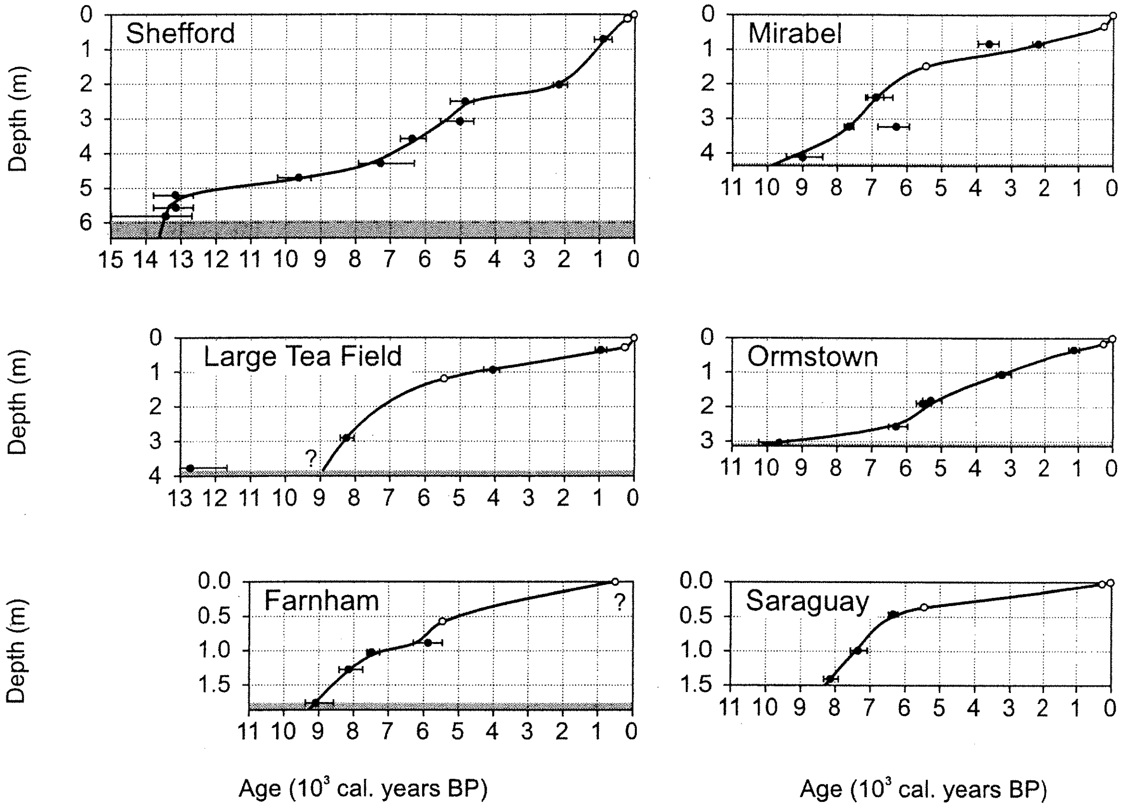


FIGURE 13. Continued

- Zone 1: Non-arboreal stage

  Periglacial desert: very low pollen concentration, dominance of long-distance transported arboreal pollen, occasional herb and shrub pollen representation, some arctic taxa.

  Herbaceous tundra: increased pollen concentration (2-10 x), Cyperaceae, Poaceae, arctic taxa.

  Shrubby tundra: *Salix*, *Betula* shrubs, arctic taxa, sustained pollen concentration.

- Zone 2: Afforestation stage (increasing pollen concentration)

  Boreal forest: *Picea*, *Populus*, associated or not with Cupressaceae (probably *Juniperus*), *Alnus crispa*, *Pinus banksiana* type and *Betula* trees.

- Zone 3: Forest stage (pollen concentration high and variable)

  3a. Mixed forest: *Abies*, *Pinus strobus*, *Betula* trees.

  3b. Maple-oak deciduous forest: *Acer saccharum*, *Quercus*, *Abies*.

  3c. Maple-hemlock deciduous forest: *Acer saccharum*, *Tsuga*.

  3d. Maple-beech deciduous forest: *Acer saccharum*, *Fagus*.

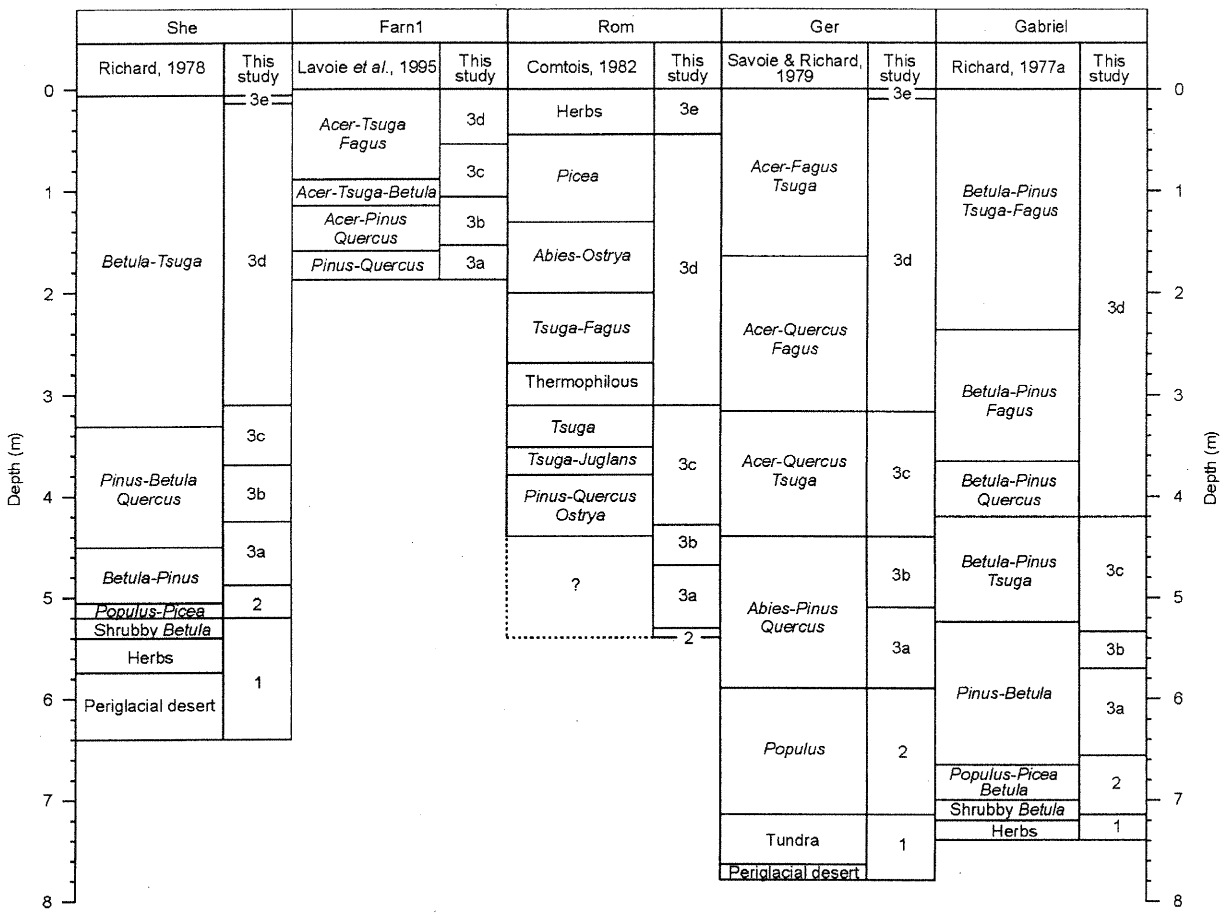
  3e. Human deforestation period: *Ambrosia*, *Plantago*, Poaceae, decrease in the arboreal pollen percentage.

This pollen zonation differs somewhat from the ones previously defined in published diagrams (Richard, 1977a; P. Richard, 1978; Savoie & Richard, 1979; Comtois, 1982; Lavoie *et al.*, 1995), notably concerning the deciduous forest stages. The new standardised zonation is compared to previous ones in Table 3.

Spatial heterogeneity of past plant formations within the study area

In order to evaluate the heterogeneity of pollen zones through the study area, a principal components analysis (Fig. 14) was carried out on pollen percentages (Ritchie & Yarranton, 1978). The analysis was performed from both original pollen spectra and mean pollen spectra calculated by averaging pollen spectra by zone – as defined above – and by site. The results obtained from the mean pollen spectra set were clearer and are consequently used to illustrate this paper.

TABLE 3. Comparison between pollen zonations defined in previously published diagrams and in this study, respectively.



The analysis contrasts homogenous forest zones with heterogeneous herbaceous-dominated ones: this opposition translates well the dominant taxa's influence. The most homogenous zones (mixed forest and deciduous forests) are characterised by the constant dominance of one or two taxa: *Abies*, *Acer-Quercus*, *Acer-Tsuga* and *Acer-Fagus*, respectively. The boreal forest zone appears more heterogeneous, probably because of the dominance of *Picea* in some sites and of *Populus* in others. Finally, the most heterogeneous zones correspond to the tundra and European period, which present a large variety of pollen assemblages through the area. In particular, the elongated shape of the tundra zone (Fig. 14) reveals taxinomic

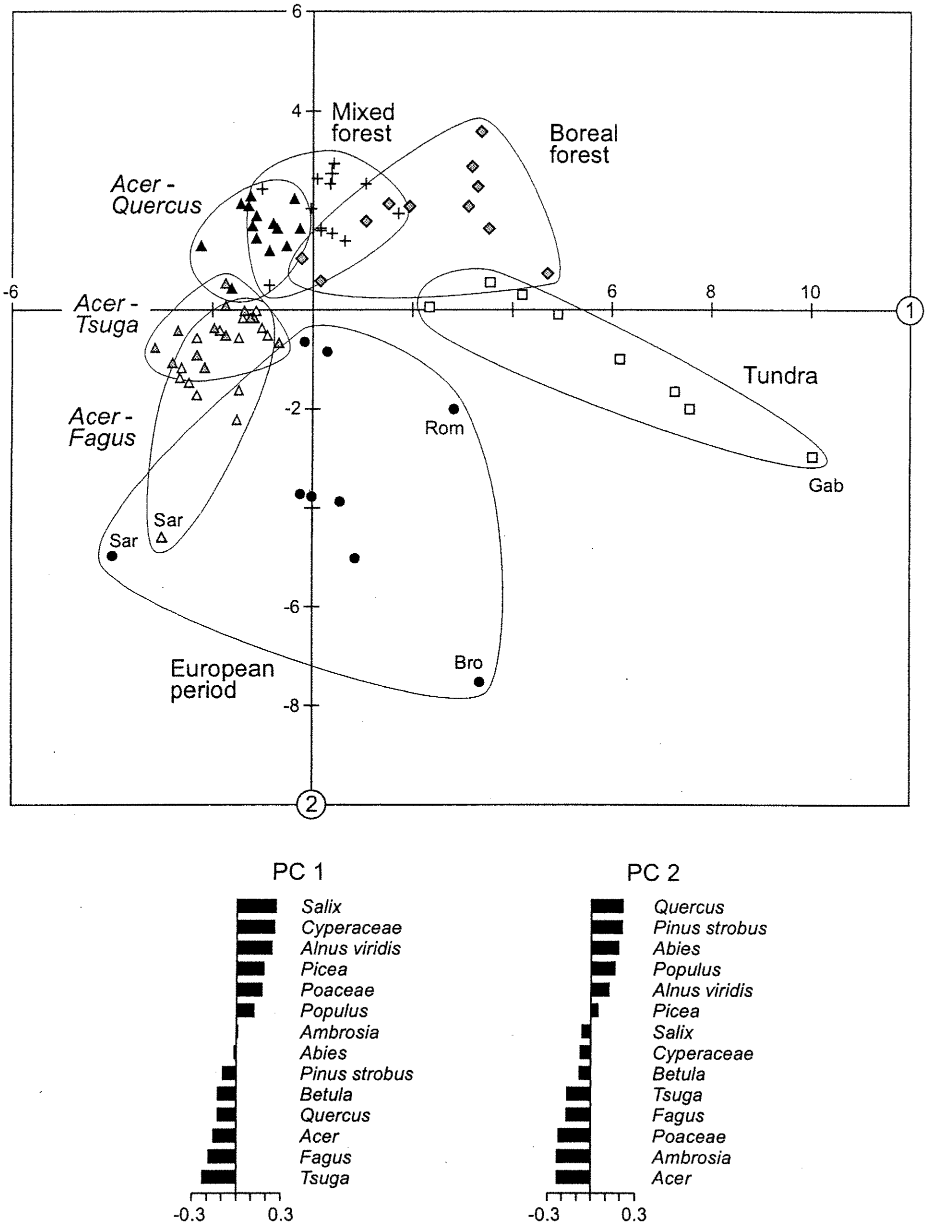


FIGURE 14. Ordination of mean pollen spectra on the first and second principal components (PC), which account for 18.2 % and 12.0 % of the total variance, respectively. Among the thirty-seven taxa used in the analysis, only the most important are classified by order of their respective contributions to principal components.

differences between sites, emphasised by the ambiguous influence of *Betula*. This taxon is principally represented in forests by arboreal species, but occurs also in tundra as shrubby species. Consequently, the abundance of *Betula* in tundra pollen spectra contributes to their displacement toward the forest zones. Based on this interpretation, the extreme position of the St-Gabriel tundra spectra (Fig. 14) is explained by the rarity of *Betula* and the dominance of other tundra taxa such as *Salix* and Cyperaceae. However, local characteristics also play an important role, as revealed by the isolated location of Saraguay, Bromont and Romer within the *Acer-Fagus* zone and the anthropic one (Fig. 14). First, Saraguay site (Fig. 9) is a forest hollow located on an island within the St. Lawrence river, with a very local pollen source area and very low net sediment accumulation in the upper layers. These characteristics are likely to separate it from larger open sites, characterised by a more regional pollen record. Second, the European period spectra of Bromont and Romer (Fig. 14) differ from others by more important percentages of Cyperaceae and Poaceae, which could indicate a greater deforestation in both regions.

Results

Comparison of the pollen assemblages (Fig. 15) leads to a palaeophytogeographic reconstruction throughout the Montréal lowlands during the entire postglacial period (Fig. 16). The age of the very first pollen assemblages in each diagram is difficult to pinpoint, due to the lack of dates in the clay deposits underlying organic sediments. Our basal age extrapolations are based on the assumption of a rapid accumulation rate in inorganic sediments (King, 1985). This feature moreover is suggested by variable mineral granulometry and very low pollen concentrations (Figs. 3 to 11) which indicate the absence of local vegetation likely to obstruct the erosion of surrounding surficial deposits (Richard, 1977a, 1977b).

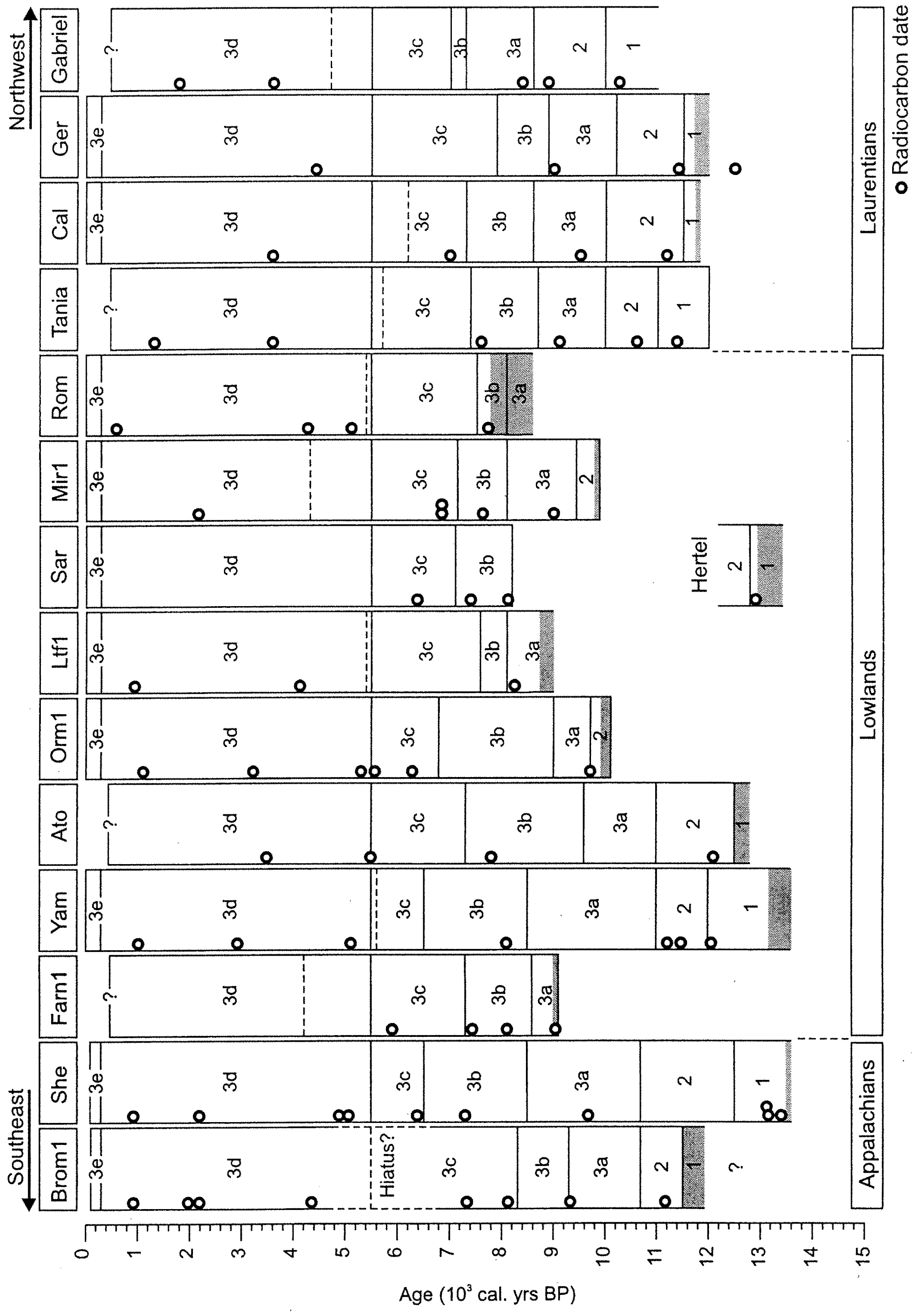


FIGURE 15. Comparison between regional pollen diagrams, based on radiocarbon dates and palynostratigraphy. Grey zones correspond to the basal mineral deposition.

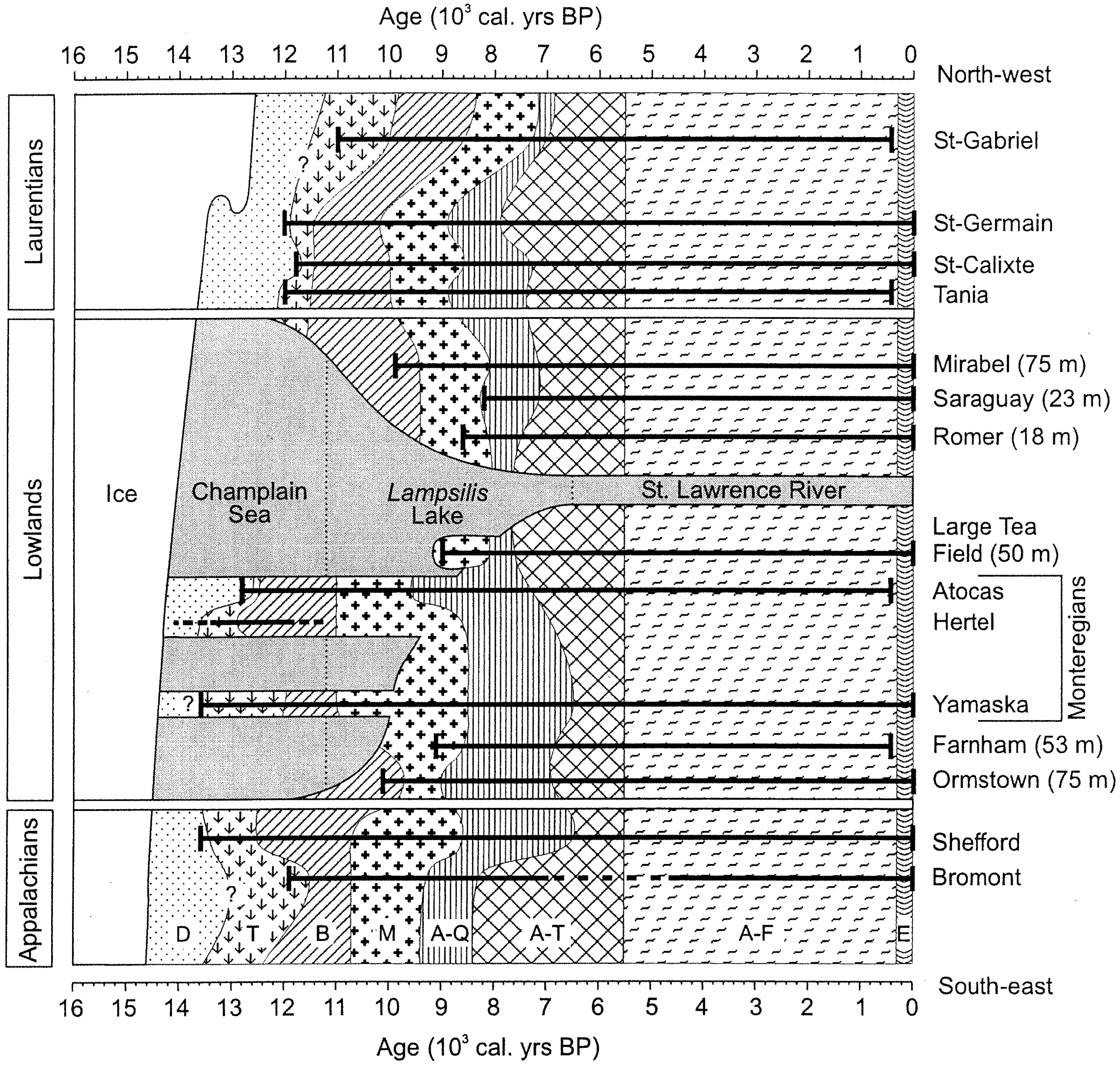


FIGURE 16. Space-time reconstruction of the vegetation history in Montréal region. D: Periglacial desert; T: Tundra; B: Boreal forest; M: Mixed forest; A-Q: *Acer-Quercus* forest; A-T: *Acer-Tsuga* forest; A-F: *Acer-Fagus* forest; H: Human period

Non-arboreal stages: periglacial desert and tundra

The non-arboreal stage was not recorded in the sediments of Farnham, Large Tea Field, Mirabel, Ormstown, Romer and Saraguay, located on the post-Champlain Sea deposits. The basal pollen assemblages of Large Tea Field sequence resembling zone 1 (Poaceae, Cyperaceae, *Salix*) and the rejected date Beta-14373 are believed to result from redeposition of older material. Yamaska and Lac-aux-Atocas, located on Monteregian hills (Figs. 2 and 3), probably emerged as nunataks within regressing ice just before isolation on islands within the Champlain Sea (Figs. 2 and 16). In all other sites except Bromont, a more or less complete vegetation succession occurred during this stage.

Periglacial desert. Despite the uncertainty concerning basal dates, the first pollen records on the Appalachian highlands and the Monteregian hills are roughly simultaneous around 13 600 cal. BP (Fig. 15). Pollen evidence of periglacial desert were also found on the Laurentian highlands, at St-Calixte (Fig. 10) and St-Germain (Savoie & Richard, 1979), dated around 12 000 cal. BP.

Herbaceous tundra. The onset of tundra was recorded at Shefford and Yamaska around 13 400 cal. BP (P. Richard, 1978), at Lac-aux-Atocas around 12 600 (Fig. 4), at St-Germain around 12 000 (Savoie & Richard, 1979) and at St-Calixte at ~11 700 (Fig. 10). These records indicate clearly a south-north migratory trend from the Appalachians to the Laurentian highlands (Fig. 16).

Shrubby tundra. Shrubby tundra succeeded herbaceous tundra around 13 000 cal. BP at Shefford and Yamaska, and around 12 500 at Lac-aux-Atocas. In those three sites, the first organic deposition is contemporaneous with the shrubby tundra development, showing the influence of surrounding vegetation on lake sediment type. The build-up of a humus layer under riparian formations and humate transport in lakes by seepage were responsible for the water eutrophication indicated by the sedimentary shift. This also occurred at St-Calixte where shrubby tundra formations appear at ~11 600 cal. BP (Fig. 10). The shrubby tundra stage is apparently absent at St-Germain (Savoie & Richard, 1979). The latest local shrubby tundra is recorded at St-Gabriel, the northernmost site around 10,200 cal. BP (Richard, 1977a).

Afforestation stage: boreal forest

The earliest pollen evidence for boreal forest in the study area is recorded at Lac Hertel, located on Mont St-Hilaire (Fig. 3). After a tundra phase, an open mixed forest of *Picea mariana* and *Populus tremuloides* developed at ~12 800 cal. BP (LaSalle, 1966). Similar formations replaced tundra around 12 500 cal. BP at Shefford and Lac-aux-Atocas, at ~12 000 at Yamaska, then around 11 500 at Bromont. The apparent establishment of a boreal forest on Mont St-Hilaire before its installation around the Appalachian sites could be related to dating problems or to particular local conditions. The bowl shape of the lake's catchment on the hill could effectively have provided protected conditions, favourable to the early development of local arboreal populations.

The afforestation stage in the Laurentian highlands began at ~11 500 cal. BP in the south (St-Calixte and St-Germain) and at ~10,000 in the north (St-Gabriel). Contrary to the Appalachians and Monteregians, the afforestation of the Laurentians shows a clear dominance of *Populus* associated with Cupressaceae (Fig. 10; Richard, 1977a; Savoie & Richard, 1979). *Populus* pollen was determined as *P. tremuloides* type (Richard, 1994b), but Comtois & Payette (1984) have envisaged that *P. balsamifera* also played a role during the Late-glacial in southern Québec. On the basis of ecological preferences, Cupressaceae pollen probably comes from *Juniperus* rather than *Thuja* (Richard, 1977a; Liu, 1990). The presence of juniper shrubs, associated with *Myrica*, marks the open character of the *Populus* forest, which could have been similar to the present-day Aspen Parkland of Saskatchewan (Mott, 1978; Savoie & Richard, 1979). In the four Laurentian diagrams, a slight maximum of *Picea* pollen follows the *Populus* phase, probably revealing the integration of scattered spruces into the aspen formation.

Forest stage: mixed and deciduous forests

The mixed forest was defined for Québec as the set of formations dominated by *Abies balsamea* in association with *Betula papyrifera* or *B. alleghaniensis* (Richard, 1993). Despite the presence of coniferous species (*Pinus strobus* and *Tsuga canadensis*), Grandtner (1966) considers the Laurentian maple forests as northern deciduous forests, because of the dominance of *Acer saccharum*, *Fagus grandifolia*, *Betula* spp., *Quercus* spp. and a wealth of deciduous companion species. Only two sites located in the lowlands (Mirabel and Ormstown) recorded the regional development of a fir forest; the others

(Farnham, Large Tea Field, Romer and Saraguay), in areas still submerged by the *Lampsilis* Lake, did not record this development. Their pollen record began during the mixed forest period (zone 3a, Figs. 15 and 16).

Fir – birch forest (mixed forest). Mixed forest development was first recorded at Yamaska and Lac-aux-Atocas around 11 000 cal. BP (Figs. 15 and 16) but occurred ~300 years later at Bromont and Shefford. This delay between sites on central Monteregian hills and those on the Appalachian piedmont could be related to dating inaccuracies. Nevertheless, these data point to a very early onset of mixed forest on Champlain Sea islands, at least synchronous, if not prior to the southernmost coastal areas. In all these southern sites, the balsam fir forest phase is accompanied by a maximum in *Betula* pollen percentages, generally occurring at the onset of the zone, with, except in Lac-aux-Atocas, an abrupt increase in *Pinus strobus* pollen percentages. Considering the important pollen production and widespread dispersion of these taxa, the increase in their pollen percentage probably reveals their regional development within the fir forest. Measurements of *Betula* pollen size point to *B. papyrifera* as the dominant species during maximum representation (Richard, 1993). Lac-aux-Atocas and Ormstown did record the *Pinus strobus* pollen increase, but earlier, during the boreal forest (Figs. 4 and 8). These observations could reveal an early local installation of *Pinus strobus*, in association with *P. banksiana*, around Lac-aux-Atocas on the Mont St. Bruno and highlight the particularly late development of mixed forest around Ormstown (Fig. 16).

Sugar maple – oak forest. As for the boreal and mixed forests, the sugar maple (*Acer saccharum*) forest appears firstly on a Monteregian hill. The *Acer-Quercus* assemblage is identified on Mont St-Bruno (Lac-aux-Atocas) around 9600 cal. BP, before its development at Bromont around 9300. Other Appalachian and Laurentian sites recorded the development of deciduous forest between 9000 and 8500 cal. BP, except St-Gabriel at ~7300 cal. BP (Figs. 15 and 16). In the lowlands, two sites (Farnham and Ormstown) recorded this stage at the same time as at the highland sites, while the others (Large Tea Field, Mirabel, Saraguay and Romer) recorded it ~500 years later. Throughout the area, the increase in *Tsuga canadensis* pollen percentages recorded during the *Acer-Quercus* forest stage (zone 3a) reveals the installation of hemlock in the vegetational cover.

Sugar maple – hemlock forest. The dominance shift from *Quercus* to *Tsuga* in the maple forest is first recorded at Bromont, in the Appalachians, around 8300 cal. BP and at St-Germain, in Laurentians, around 7900 cal. BP. The latest records of this event are those of Shefford and Yamaska, at ~6,500 cal. BP. As for the *Acer-Quercus* forest stage, the constitution of the *Acer-Tsuga* formation does not present any migratory trend in the study area (Fig. 16). The important delay between records at Bromont and Shefford implies a better recording of the extra-local pollen input (Janssen, 1973; Jacobson & Bradshaw, 1981) compared to the regional one.

Sugar maple – beech forest. Hemlock (*Tsuga canadensis*) populations declined sharply throughout eastern North America at ~5500 cal. BP (Davis, 1981). This rapid, widespread decline has been attributed to a phytophagous insect outbreak (Davis, 1981; Allison *et al.*, 1986; Bhiry & Filion, 1996) and, more recently, to a dry climate (Haas & McAndrews, 2000). As a consequence of the *Tsuga* decline, several taxa increased in abundance, including *Acer saccharum*, *Fagus grandifolia* and *Betula* (Davis, 1981; Fuller, 1998). These changes occurred at most sites in the study area, except at Shefford where the *Tsuga* decline was not clearly recorded (P. Richard, 1978), which corroborates the local character of the pollen record in this site. Based on interpolation between radiocarbon dates, the *Tsuga* decline is dated between 6200 and 4200 cal. BP within the study area (Fig. 15). However, it is directly dated around 5500 cal. BP at two sites (Lac-aux-Atocas, Ormstown; Fig. 15) and by interpolation at two others (Saraguay and St-Germain; Fig. 15), and only one such date (I-8786; Table 2) is inconsistent with this age. This points to the reliability of the commonly admitted mean age of ~5500 cal. BP for the event (Davis, 1981).

Human deforestation period. The European settlement of the St. Lawrence lowlands resulted in a massive deforestation starting between 250 and 200 cal. BP and culminating 150 years ago. Forest clearance and subsequent agricultural practices allowed the invasion of adventitious taxa (which became >25% of the total vascular flora), mainly represented in pollen spectra by *Ambrosia* type. This major anthropogenic shift was not recorded in four sites (Farnham, Lac-aux-Atocas, Tania and St-Gabriel), mostly because the uppermost sediments were not recovered during coring operations.

Discussion

The first postglacial plant formations (tundra, boreal forest and mixed forest) present clear migratory trends throughout the entire study area, while the development of deciduous forests shows no particular pattern. Several factors appear to have influenced this contrasting behaviour.

First, the palaeogeographic context played a major role during the first stages. The migration of tundra can be directly related to the presence of the ice sheet and related periglacial climate (Gangloff *et al.*, 1971; Gangloff, 1981). The Champlain Sea itself promoted a harsh coastal climate (Delage *et al.*, 1985), and iceberg scours were identified near Huntingdon (Fig. 3; Delage & Gangloff, 1993). Additionally, the Champlain Sea constituted an important barrier to seed dispersion during the early stages, and subsequently, the *Lampsilis* Lake controlled the Laurentian plant colonisation (Fig. 2). The best evidence for the barrier effect is given by the expansion of *Populus* and *Juniperus* on the southern margin of the Laurentian highlands, prior to that of *Picea*. The presence of the sea favoured *Populus*, that possesses easily wind-dispersed seeds (Mott, 1978), and *Juniperus*, an ornithochorous taxon commonly propagated by granivorous migrating birds. As the initial coloniser, *Populus* held *Picea* in check for a long time, probably helped by a harsh fire regime (Carcaillet & Richard, 2000). The reduction of the barrier effect during retreat of the Champlain Sea, then during the *Lampsilis* Lake episode, explains the rapid and irregular spread of deciduous forests over the lowlands.

Secondly, ecological processes also partly determined several differences between conifer-dominated forests and deciduous forests. Development of boreal and mixed forests was controlled by the migration of their dominant species (respectively *Picea mariana* and *Abies balsamea*) during the early Holocene climate warming. The arrival of each new tree species created new competitive relationships for the previous species. By contrast, the development of *Acer*-dominated forests resulted more from a dominance shift between already present, local taxa rather than from a migratory invasion. *Acer saccharum* and *Quercus* (cf. *rubra*) were already present in the mixed forest formation dominated by *Abies* before replacing it (Ritchie, 1987; Jackson *et al.*, 1997). In the same way, the installation of *Tsuga canadensis* was recorded at all sites during the *Acer-Quercus* stage, prior to its own dominance stage. The lack of spatial trends at the onset of

Acer-Quercus and *Acer-Tsuga* assemblages (Fig. 16) imply species replacement by competition instead of a migration process. This suggests a prominent control of these dynamics by local processes such as soil maturation. The most obvious evidence of the absence of a spatial pattern is given by the important delay between the records at Bromont and Shefford. Despite the proximity of the sites, the presence of the *Acer-Tsuga* forest is recorded ~1800 years later at Shefford than at Bromont (Fig. 15). Finally, the replacement of *Tsuga* by *Fagus* and *Betula* is believed to result mostly from the decline of the former and the subsequent modification of competitive equilibrium (Bennett, 1988; Fuller, 1998).

Thirdly, physiographic traits can explain some local departures from the general migratory pattern. Tundra, boreal forest and mixed forest developed on the central Monteregian hills before the Appalachian sites: tundra was first recorded at Lac Yamaska, Lac Hertel and Shefford, boreal forest at Lac Hertel, Shefford and Lac-aux-Atocas (Fig. 2), and mixed forest at Lac Yamaska and Lac-aux-Atocas. Early establishment on former periglacial islands – except the Shefford site – can be related to especially favourable conditions, such as protected slopes. The asynchrony between the deciduous stages of Shefford and Bromont also shows the major influence of local conditions on forest evolution. The number and the consistency of radiocarbon dates at both sites lead us reject the hypothesis of dating problems and to consider that the asynchrony points out the local character of pollen record at sites separated by only 10 km.

Conclusions

Two major patterns of postglacial plant migration are recognised at the regional scale in southern Québec. The vegetation changes during pioneer stages correspond to a succession of relatively discrete communities, as described by the holistic model (Clements, 1916; Braun-Blanquet, 1932), while the evolution of late-successional mature formations better conforms to the individualistic model, which defines plant succession as the result of independent demographic response of individual species (Gleason, 1926, 1939; Curtis, 1959). The first mode of plant migration by 'community replacement' reproduces, at the regional scale, the south-north migratory pattern evidenced at the continental and sub-continental scales. By contrast, the second mode by 'domination shifts between already established taxa' is characterised by non-patterned regional changes. This suggests a greater dependence of late-successional communities to local conditions, such as soil maturation, than for pioneer formations.

However, if ecological processes can be considered to have controlled the major trends of plant migration in the St. Lawrence lowlands, the palaeogeographic context and the physiography explain some particular features. The Champlain Sea is shown to have strongly influenced the early vegetation stages, notably by constituting a geographical barrier to seed dispersion. Moreover, the islands formed by the Monteregian hills within the sea recorded the onset of early plant formations prior to their development in coastal Appalachian sites. This could be due to dating inaccuracies; however, the consistency of the first records of pioneer formations (tundra, boreal forest and mixed forest) between several insular sites rather suggests that the islands provided particularly favourable conditions for plant installation.

Finally, our study highlights the spatial heterogeneity of non-arboreal stages (tundra and European period), which can essentially be related to various plant assemblages without any dominant taxa. The relative homogeneity of mixed forest and deciduous forests is, on the contrary, assigned to the constant dominance of one or two taxa (*Abies*, *Acer-Quercus*, *Acer-Tsuga* and *Acer-Fagus*, respectively).

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POSTGLACIAL CLIMATE DYNAMICS IN THE ST. LAWRENCE LOWLANDS, SOUTHERN QUÉBEC: POLLEN AND LAKE-LEVEL EVIDENCE

Serge D. Muller, Joël Guiot, Pierre J.H. Richard,
Jacques-Louis de Beaulieu & David Fortin

Palaeogeography, Palaeoclimatology, Palaeoecology – Submitted

Abstract

This study provides the first quantitative palaeoclimate reconstruction of southern Québec. Pollen and lake-level data are used to reconstruct past climate changes in the St. Lawrence lowlands, southern Québec. Past lake-level changes are assessed from sedimentological, pollen and macrofossil records from a single lateral core of Lac Hertel, which lies in the central part of the study area. Three low lake-level periods are recognised: some time before 8000, 7600-6600 and 4800-3400 cal. BP. The modern analogues method is applied on pollen data from seven well-dated sites located in the St. Lawrence lowlands and adjacent mountain areas, constrained and unconstrained by lake-level changes. At the regional scale, the results confirm the general trends known for eastern North America: a dry and cold late-glacial episode due to the presence of pro-glacial lakes and seas; a rapid climate improvement between 12 500 and 11 000 cal. BP possibly caused by increasing summer insolation; a dry period from 10 000 to 6500 cal. BP; a brief, slight cooling event between 9000 and 8000 cal. BP, possibly related to summer cooling of Arctic airmasses; a temperature optimum around 8000 cal. BP, which could represent a regional "Hypsithermal"; a dry climatic phase between 5000 and 3000 cal. BP; and finally, a progressive decrease in summer temperature and an increase in (winter ?) precipitation over the 4500 last years. These results point to the importance of taking into account seasonal patterns in climate reconstruction, by the way of merging pollen and lake-level data.

Key words: Lake-levels, climate reconstruction, regional scale, modern analogues method, St. Lawrence lowlands, southern Québec

Résumé

Dynamiques climatiques postglaciaires dans les basses terres du St-Laurent, Québec méridional : une reconstitution basée sur le pollen et les niveaux lacustres.

Cette étude constitue la première reconstitution paléoclimatique quantitative pour le sud du Québec. Les variations passées du climat dans les basses terres du St-Laurent sont reconstituées sur la base de données polliniques et de variations de niveaux lacustres. Ces dernières sont reconstituées à partir d'analyses sédimentologiques, polliniques et macrofossiles réalisées sur une carotte latérale du lac Hertel, lui-même situé au centre de la région étudiée. Trois bas niveaux lacustres sont mis en évidence : avant 8000, de 7600 à 6600 et de 4800 à 3400 cal. BP. La méthode des analogues modernes a été appliquée, sous la contrainte des niveaux lacustres, aux données polliniques de sept sites bien datés et localisés dans les basses terres du St-Laurent et les hautes terres adjacentes. A une échelle régionale, les résultats confirment les tendances générales connues pour l'est de l'Amérique du Nord : un tardiglaciaire froid et sec, en particulier à cause de la présence de lacs et de mers proglaciaires ; un rapide réchauffement climatique entre 12500 et 11000 cal. BP, peut-être dû à une augmentation de l'insolation estivale ; une période sèche entre 10000 et 6500 cal. BP ; un bref et léger refroidissement entre 9000 et 8000 cal. BP, possiblement lié à un refroidissement des masses d'air arctiques ; un optimum thermique autour de 8000 cal. BP, pouvant représenter un « hypsithermal » régional ; une phase sèche entre 5000 et 3000 cal. BP ; et finalement, une diminution progressive des températures estivales et une augmentation des précipitations (hivernales ?) au cours des 4500 dernières années. Ces résultats mettent en évidence l'importance de tenir compte des patrons saisonniers dans les reconstitutions climatiques, en combinant données polliniques et niveaux lacustres.

Mots clés : Niveaux lacustres, reconstitution climatique, échelle régionale, méthode des analogues modernes, basses terres du St-Laurent, Québec méridional.

Introduction

According to the assumption of a relationship between climate and plant distribution (Woodward, 1987), past climates can be reconstructed from past phytogeographies. However, our perception of this relationship depends greatly on scales (Ritchie, 1986; Webb, 1986, 1993). At the scale of continents, the present-day latitudinal zonation of vegetation matches well the north-south climate improvement (Greller, 1989; Huntley *et al.*, 1989). Based on that relationship, past climates have successfully been reconstructed from pollen data at a continental scale (e.g., Prentice *et al.*, 1992; Guiot *et al.*, 1993; T. Webb *et al.*, 1993). At larger scales, several factors such as disturbances (Ritchie, 1986), physiography or ecological processes (Davis *et al.*, 1986; Prentice, 1986) may create non-equilibrium states between vegetation and climate, and consequently complicate pollen-based climate reconstructions. Moreover, past plant formations without modern analogues may constitute another source of difficulties for these reconstructions.

The aims of this paper are (1) to attempt a regional pollen-based climate reconstruction taking in account these problems and (2) to discuss this reconstruction in regards to the qualitative knowledge of past climates (e.g., Richard *et al.*, 1992; Richard, 1994; Lavoie & Richard, 2000a). The reconstruction is made using the modern analogues method constrained by lake-level changes, shown to be linked to climate dynamics in eastern North America (Harrison, 1989). Previous studies in the St. Lawrence lowlands, south-western Québec (Fig. 1), have defined the past influence of local factors on plant geography (Muller & Richard, accepted) and discussed the analogies between past and modern plant communities (Richard, 1989, 1993). Moreover, the study area comprises several well-dated pollen diagrams (4-10 dates per site), which allows reducing the influence of local physiographic parameters by calculating regional climatic averages. Finally, Lac Hertel, a small lake characterised by a small catchment, lies right in the central part of the area and likely provides a reliable and independent record of proxy hydroclimatic data, usable to constrain the climate reconstruction. A previous study by LaSalle (1966) attests the a complete postglacial pollen record from the central zone of the lake.

Study area

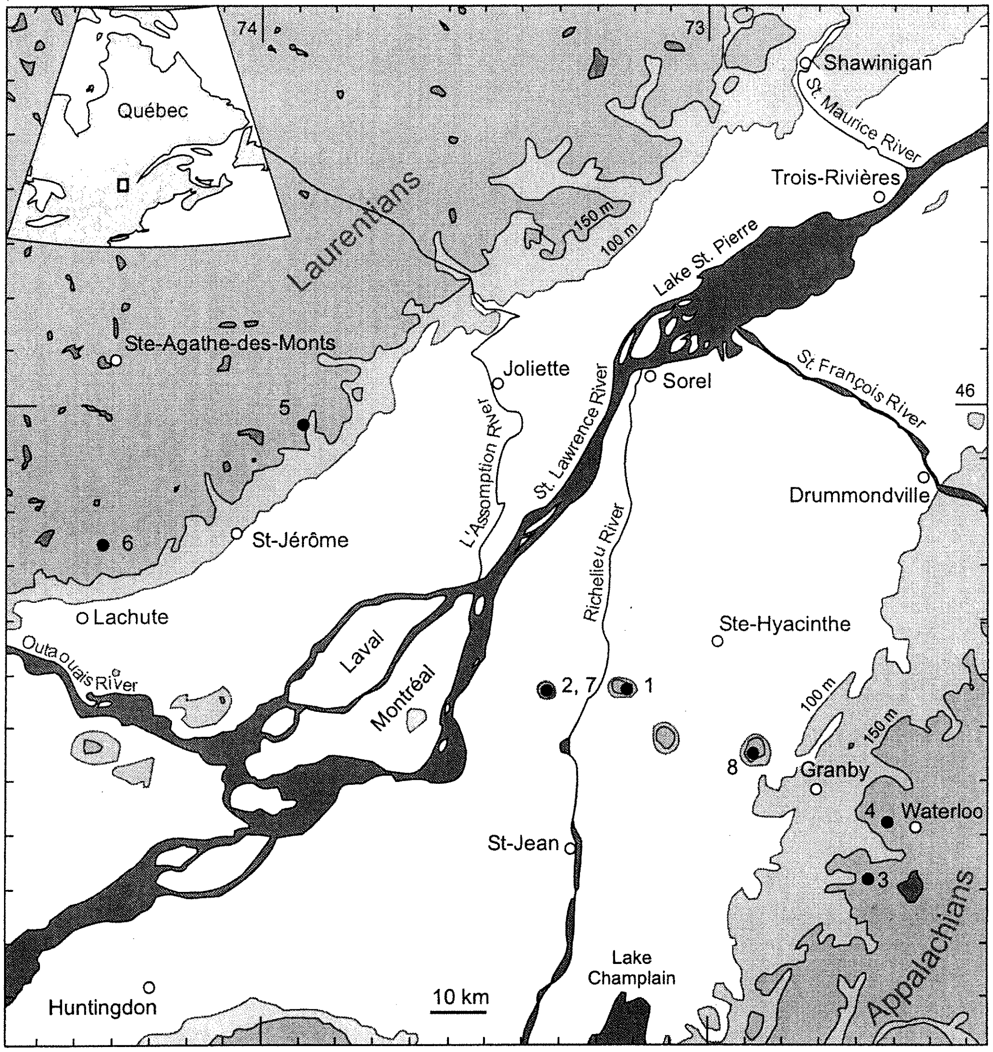
The Montréal lowlands (Fig. 17) consist mainly of a flat plain of marine clay deposited over tills on a sandstone-dolomitic bedrock (Laverdière *et al.*, 1972). Several intrusive hills (Monteregian hills) emerge from the plain roughly along an east-west axis. Lakes are rare in the lowlands and, except for the now filled-up Lac Romer, they are all located on the Monteregian hills. The natural vegetation of the region, at the northern limit of deciduous forests, is dominated by *Acer saccharum*, in association with *Betula alleghaniensis* on highlands, with *Tilia americana* in the north-eastern part of the plain and with *Carya cordiformis* south-westwards of Sorel (Grandtner, 1966; Richard, 1987). The transitional nature of these forests is well marked by the boreal and temperate elements of which they are constituted (Maycock, 1961; Bouchard & Maycock, 1978). However, cultivated lands today occupy the most of the plain, except moraines and bedrock outcrops.

Lake-level changes

Lac Hertel (Figs. 17 and 18) occupies the central depression of Mont St. Hilaire, one of eight so-called Monteregian hills. Located 50 km east of Montréal, this intrusive massif is mainly composed of nepheline syenite on the east zone and of essexite (gabbro) on the west zone (Horváth & Gault, 1990). The 28.4 ha lake, situated 173 m above sea level, drains a very small watershed (about 3.47 km^2) that allows considering it as a good recorder of past precipitation variations. The lake: watershed ratio is 13.

Methods

The lake-level changes are established by using a combination of evidences from sediment composition and vegetation distribution (Dearing & Foster, 1986; Digerfeldt, 1986), reconstructed from sedimentological, macrofossil and pollen analyses conducted on a single lateral core. The studied core was taken 220 m offshore, facing an inlet (Fig. 18) with a modified Livingstone piston sampler (Livingstone, 1955; Wright, 1967), in a present-day water-depth of 6 m. Loss-on-ignition (at 600 °C during ½ h; Dean, 1974) was measured on contiguous 1-cm thick samples.



1. Lac Hertel, this study
2. Lac-aux-Atocas, Gauthier, 1981; Muller & Richard, accepted
3. Lac Bromont, Muller & Richard, accepted
4. Shefford site, Richard, 1977, 1978
5. Lac de St-Calixte, Muller & Richard, accepted
6. Lac Tania, Muller & Richard, accepted
7. Lac-à-la-Tortue, Gauthier, 1981
8. Lac Yamaska, Muller & Richard, accepted

FIGURE 17. Study area and location of sites used in climate and lake-level reconstructions

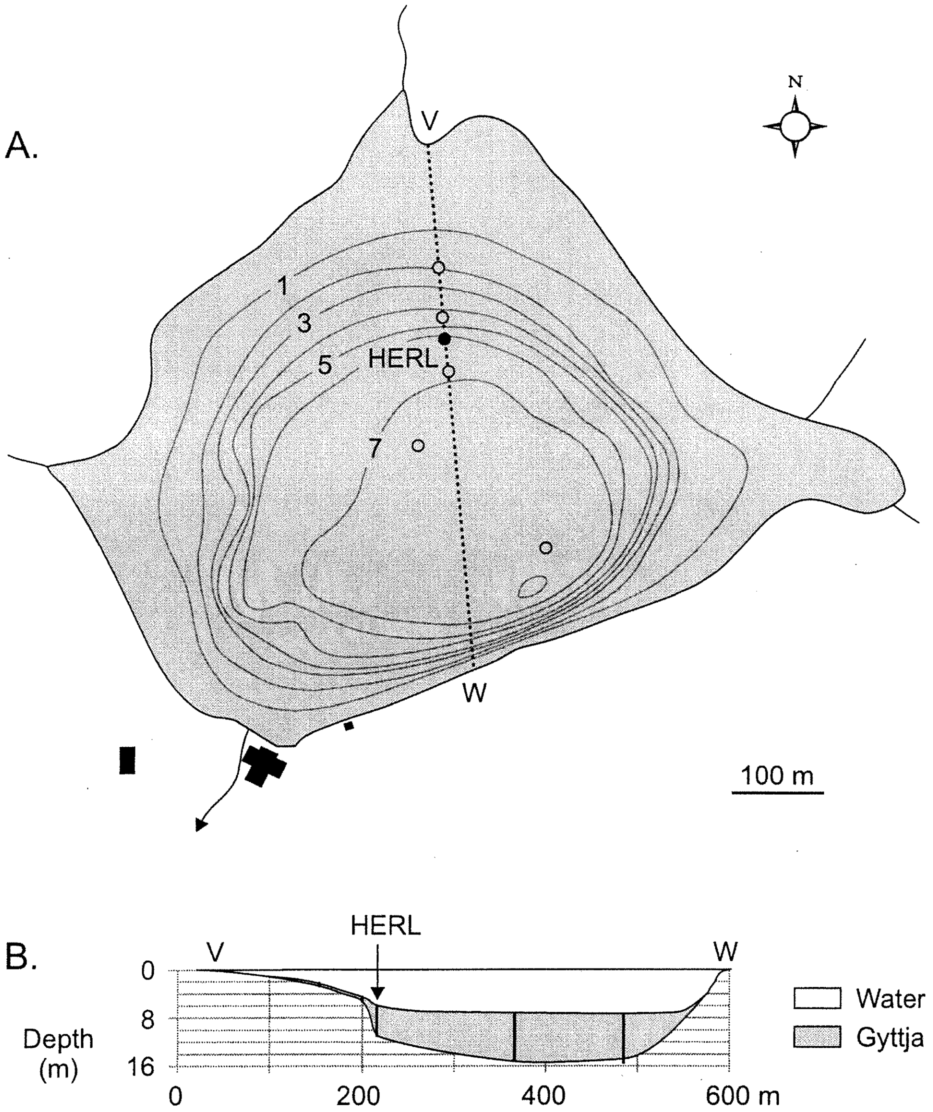


FIGURE 18. Bathymetry of Lac Hertel.
A. Map showing the location of the investigated profile (black dot). Other dots localise measurements of sediment thickness. Depth in meters.
B. Cross-section joining the inlet mouth to the coring site, along the transect V-W. For convenience, sediment thickness measurements are represented as a linear transect.

TABLE 4. Conventional and calibrated radiocarbon ages. Rejected ages are noted in italic (see text for details).

Core name	Laboratory code	Dating method	Calibrated age	Confidence interval 2σ	Conventional age	13C/12C ratio	Depth (cm)
Herl	TO-8736	AMS	3700	3830-3610	3460 ± 40	n.a.	153-155
	TO-8737	AMS	4700	4830-4530	4160 ± 50	n.a.	277-278
	TO-8738	AMS	6370	6490-6290	5600 ± 50	n.a.	371-375
	Beta-154759	*bulk*	*9440*	*9520-9250*	*8460 ± 70*	*- 29.6 ‰*	*456-459*
	TO-8739	*AMS*	*10 440*	*11 070-10 180*	*9280 ± 150*	*n.a.*	*461-466*

Note: AMS dates were performed on terrestrial macrofossils. n.a.: not available.

Chronological control is provided by three radiocarbon dates (Table 4) and by two well-dated regional events used as chronological pollen markers: the *Tsuga* decline dated around 5500 cal. BP and the *Ambrosia* rise resulting from European deforestation between 250 and 200 cal. BP. Conventional radiocarbon dates are calibrated with the CALIB 4.0 program (Stuiver & Reimer, 1993). Calibrated dates, noted cal. BP (calendar years before present), are used to construct the age-depth model (Fig. 19), by interpolating the simplest curve connecting dates within the 2-sigma confidence intervals. The basal dates (TO-8739 and Beta-154759; Table 4) were rejected on the basis of pollen assemblages. Regional correlations lead us to attribute an age of about 8000 cal. BP to the interface clay-gyttja. In addition, assuming a constant pollen input between 140 and 40 cm, we calculated that the silty gyttja between 120 and 80 cm (S8, Fig. 20) accumulated between 2600 and 2050 cal. BP, i.e. three times faster than in the adjacent gyttja layers.

Pollen extraction is based on the conventional protocol (Faegri & Iversen, 1989). Pollen percentages are calculated on a sum excluding hydrophilous species' pollen and Pteridophyta spores (Berglund & Ralska-Jasiewiczowa, 1986). *Alnus incana* is also excluded from the pollen sum, due to the local origin of its pollen in some sections of the core. The minimal number of pollen grains counted and included in the sum is 500.

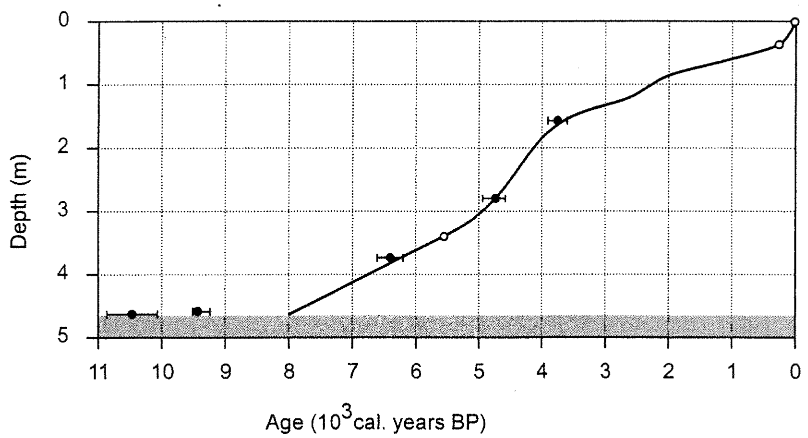


FIGURE 19. Age-depth curve of the profile Herl (Lac Hertel). Black dots represent calibrated radiocarbon dates and white dots represent well-dated regional events (*Tsuga* decline, European settlement) and the sediment surface. The curve is modified between 120 and 80 cm on the basis of pollen concentrations. See text for details.

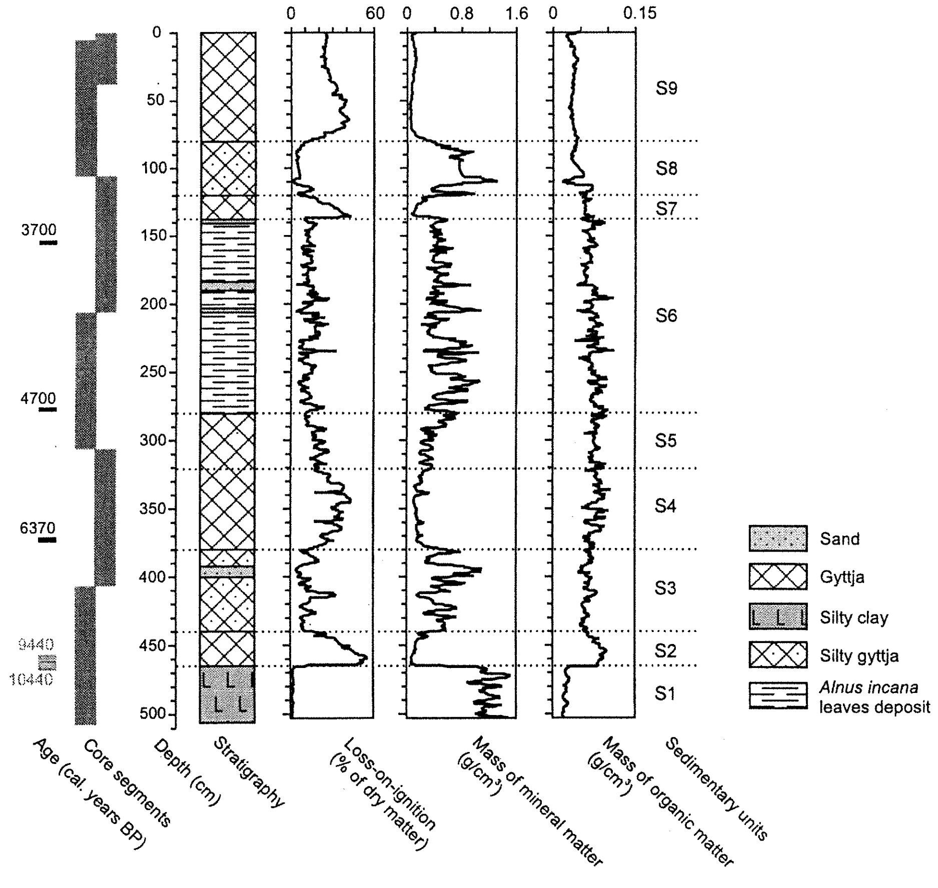


FIGURE 20. Stratigraphic data and loss-on-ignition of the profile Herl (Lac Hertel)

Macrofossil extraction was realised by sifting under a soft water jet (Grosse-Brauckmann, 1986). Volumes analysed, comprised between 10 and 60 cm^3, are standardised to a volume of 100 cm^3. The notation of pollen taxa follows Birks & Birks (1980): the mention "type" groups several taxa indistinguishable by their morphology and the mention "cf." points the most probable taxa. Diagrams were built with the GPalWin computer program (Goeury, 1997).

Results

Stratigraphic changes in the sediment, pollen and macrofossil content are presented in Figs. 20, 21 and 22, respectively. The pollen diagram zonation (Fig. 21) allowed us to correlate the lateral profile Herl with a previous central diagram (LaSalle, 1966) and with other regional studies (compiled in Muller & Richard, accepted).

The minerogenic content is here assumed to reveal past variations in water depth (the sediment limit), and consequently, in the distance between the shoreline and the coring site (Digerfeldt, 1988). It could also reflect variations in the inlet clastic input, but given the small size of the watershed drained by the brook (1.1 km^2), such variations are thought to be negligible. Macrofossil assemblages are also used to reconstruct the water depth and the distance to the shoreline. The taxinomic diversity change is summarised for each vegetation belt (Fig. 23) so as to take into account the origin of plant remains and to improve the interpretation. The aquatic macrofossil subset depends mainly on water depth (which determines the coverage area and the location of submerged plant communities), while the riparian subset depends rather on the distance from the shore and on the spatial development of riparian formations. Due to the basin morphometry (Fig. 18), the relation between water depth and the shoreline position appears to be non-linear: in relation to the lake filling-in, important shoreline variations can occur causing only slight modifications of the water depth at the coring site. Moreover, varying clastic inputs from the inlet could have disturbed macrofossil assemblages by carrying riparian and especially aquatic taxa from shallow waters to water lake depths.

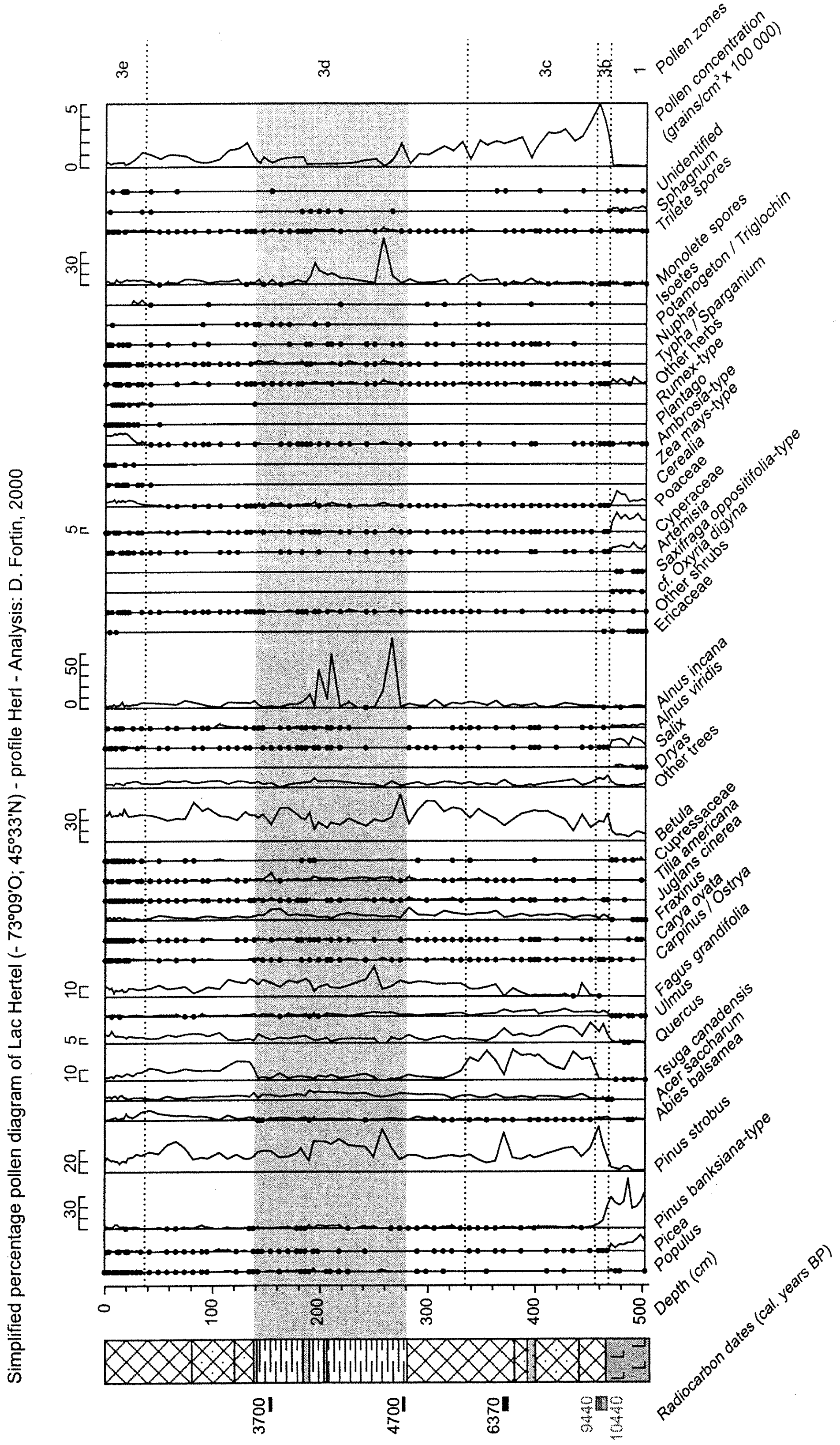


FIGURE 21. Simplified percentage pollen diagram of the profile Herl (Lac Hertel). Pollen zones correspond to the regional vegetation history (see text). The grey zone marks the sand and *Alnus* leaves deposit (unit S6, Fig.4).Stratigraphic legend in Fig. 4.

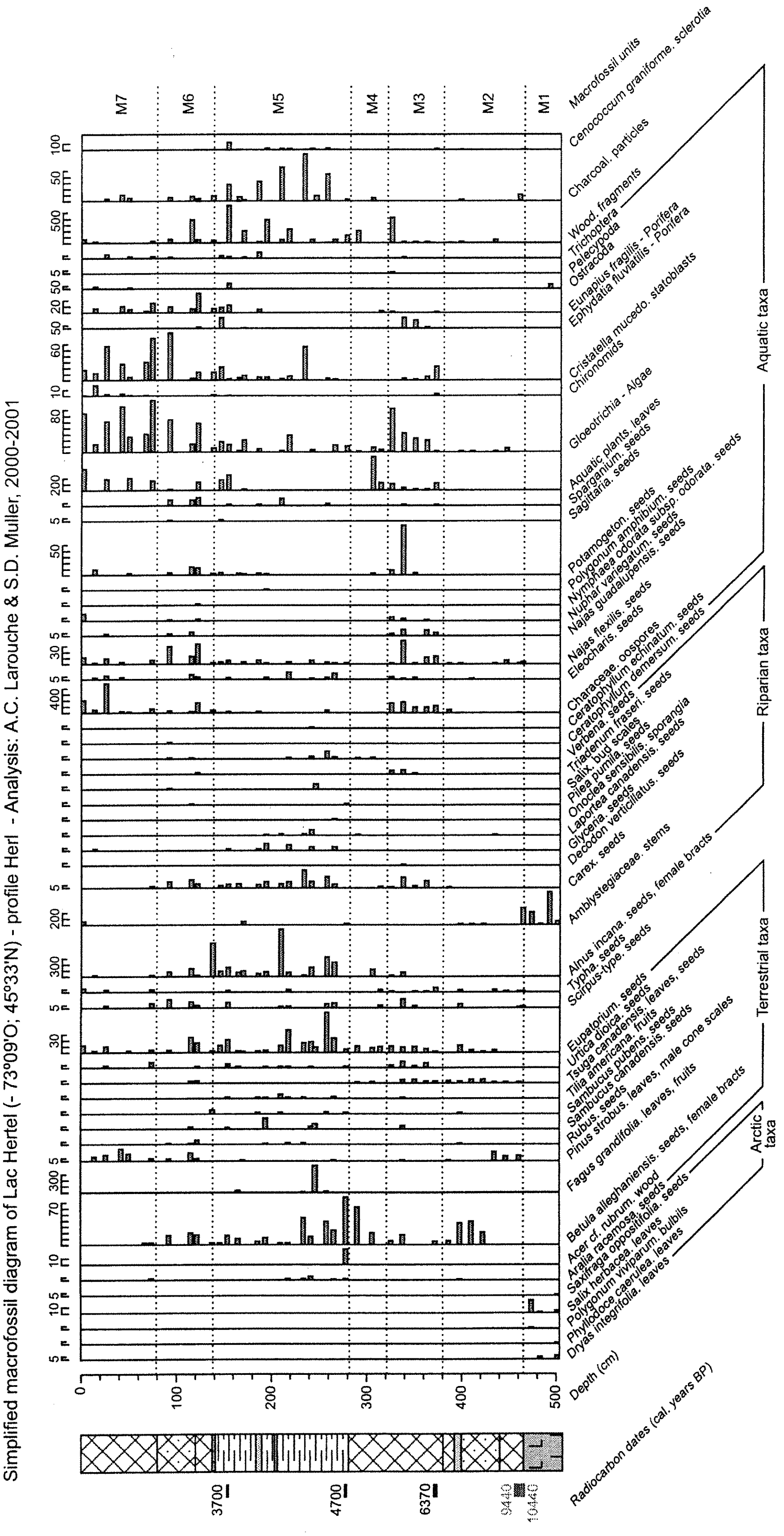


FIGURE 22. Macrofossil concentration diagram of the profile Herl (Lac Hertel). Concentrations are standardised to 100 cm^3. Stratigraphic legend in Figure 4.

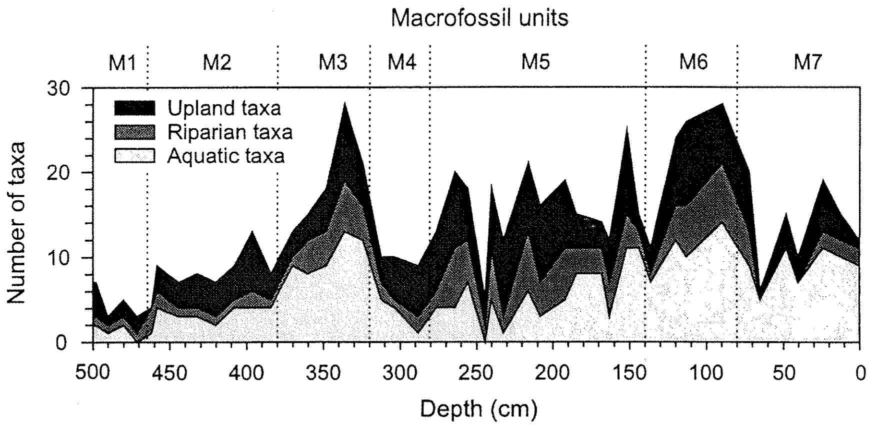


FIGURE 23. Taxinomic diversity of vegetation belts in the profile Herl (Lac Hertel)

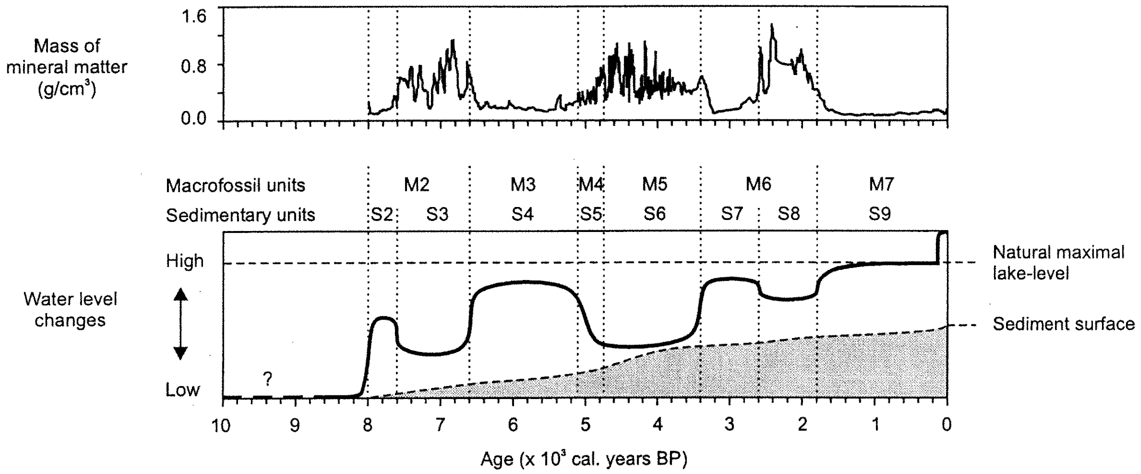


FIGURE 24. Lake-level changes at Lac Hertel

Nine sedimentary units (S1-S9) and seven macrofossil units (M1-M7) were identified from the sediment composition (Fig. 20) and the macrofossil record (Fig. 22), respectively. Comparisons between both records are used to infer past water level fluctuations of Lac Hertel (Fig. 24). Nine periods characterised by different lake-levels were recognised, among which some present contradictory sedimentary and macrofossil evidence. Alternative interpretations are discussed below. The present water level, inherited from the construction of a dam on the outlet (Fig. 18), is thought to be located between two and three meters above the natural lake-level.

Before 8000 cal. BP [hiatus]: low lake-level, >10.50 m lower than the present-day

The basal, silty clay (unit S1, Fig. 20) implies an initially high lake level just after Lac Hertel became isolated from the receding Champlain Sea waters (LaSalle, 1966). Pollen (Fig. 21) and macrofossil evidence clearly point to a very sparse tundra vegetation during this period. The sparseness of the vegetational cover is indicated by the very low pollen concentration and by the fact that silt and clay were then eroded from the watershed. The normal chronosequence should register the classification of the plant cover and the progressive change to a herb-rich, then a shrubby tundra, with a parallel increase in pollen concentration (Muller & Richard, accepted). Lack of these and of the subsequent pollen zones 2 and 3a, respectively representing boreal and mixed forest stages in the lateral core, is interpreted as being caused by an important lowering of the water level sometime during the tundra phase. The hiatus of sedimentation during this low lake-level is also evidenced by the sharp increase in loss-on-ignition (Fig. 20) at the limit between inorganic (unit S1, Fig. 20) and organic (unit S2, Fig. 20) sediments. The emersion at the coring site probably triggered erosion of previously deposited sediments and a subsequent redeposition of older organic material when the local zone was flooded. Redeposition could explain the old dates obtained at the base of the profile on upland macrofossils (TO-8739 and Beta-154759, Table 4). However, the presence of *Tsuga canadensis* seeds, which implies a younger age for the dated macrofossils (Muller & Richard, accepted), rather suggests a contamination due to old carbonates.

8000-7600 cal. BP [465-440 cm]: intermediate lake-level, 5-6 m lower than the present-day

The low minerogenic content of sediment (unit S2, Fig. 20) suggests a greater distance of the coring site from the shoreline than previously. This is corroborated by the low number of *Najas flexilis* seeds (unit M2, Fig. 22), which could indicate an important water depth at the coring site. Indeed, its seed production was shown to be optimal in light shallow waters (Haas, 1996a). Moreover, the regular occurrences of upland wind-dispersed macrofossils (e.g., *Betula*, *Pinus*, *Tsuga*; Fig. 22), which are commonly found in the centre of deep lakes (Birks, 1973), support the important water depth at the coring site. However, the constant input of organic material from surrounding forests, evidenced by the corresponding pollen and macrofossil records of *Betula alleghaniensis* and *Tsuga canadensis* (Figs. 21 and 22), may also be related to the inlet input. That would illustrate a more effective transport of macroremains from upland plants growing along the brooklet, than that of riparian taxa mainly located in quiet, shallow waters off the inlet mouth. Finally, the low macrofossil diversity of riparian taxa (unit M2, Fig. 23), which could be related to the distance between the shoreline and the coring site and also to a faint spatial development of riparian communities, leads us to propose a lake-level of 5-6 m lower than today. The water surface would then have been located just at the top of the steep slope (Fig. 18), at a position unfavourable for enhanced development of riparian formations.

7600-6600 cal. BP [440-380 cm]: slight lake-level lowering

In unit S3 (Fig. 20), the mineral matter mass is characterised by high values and important fluctuations. However, except for one sand layer near 400 cm, the mineral matter is mainly composed of very thin mica particles, likely to be easily transported away to lake depths. Although this sedimentary feature may indicate inlet divagations or variations in its erosive activity, it is more probably related to a lake-level lowering, sufficiently moderate not to have affected the representation of vegetation belts in the macrofossil assemblages (unit M2, Fig. 23).

6600-5100 cal. BP [380-320 cm]: high lake-level, around 3 m lower than the present-day

The low minerogenic content (unit S4, Fig. 20) corresponds to a gyttja deposition, generally related to an important water depth, while the very diversified aquatic plant remains (unit

M3, Figs. 22 and 23) seem to indicate a local water depth of around 2 meters. Submerged and floating-leaved species are well represented by shallow waters species like *Najas flexilis*, *Najas guadalupensis*, *Nuphar variegatum*, *Sparganium*, *Eleocharis*, *Potamogeton pusillus*-type and Characeae (Haas, 1996a; Dieffenbacher-Krall & Halteman, 2000). Moreover, the increase in diversity of riparian plant macroremains (Fig. 23), dominated by *Carex* sp. and *Triadenum fraseri* seeds (Fig. 22), could reveal the proximity of the coring site to the shoreline (Dieffenbacher-Krall & Halteman, 2000). However, compared to the previous zone, it seems more probable that the macrofossil record corresponds to a transport from widespread riparian and aquatic formations developed on the shallow marginal terrace (Fig. 18) rather than to a strictly local production. Consequently, water levels were probably about 3 meters lower than the present-day between 6600 and 5100 cal. BP.

5100-4800 cal. BP [320-280 cm]: progressive lake-level lowering

Although the gyttja deposited during this period contains no macroscopic mineral particles, loss-on-ignition reveals an increase in the minerogenic content (unit S5, Fig. 20) which is interpreted as an important but progressive water level lowering. The macrofossil record (unit M4, Fig. 22) is not well characterised, but corresponds to that of unit M2, recognised as a period of intermediate lake-levels. The end of the zone is marked by an abrupt increase in the inorganic matter mass (Fig. 20) and by an enhanced macrofossil representation of upland taxa (*Betula alleghaniensis*, wood fragments; Fig. 22), which likely indicate the proximity of the shoreline.

4800-3400 cal. BP [280-140 cm]: low lake-level, 6-7 m lower than the present-day

This period shows a straightforward increase in sedimentation rate (Fig. 19), corroborated by very low pollen concentrations (Fig. 21). The sediment is composed of fine sand particles with abundant *Alnus incana* subsp. *rugosa* leaves (unit M5, Fig. 22). Associated with the leaves deposit, *Alnus* pollen peaks (Fig. 21) correspond certainly to catkins falling in sediment, pointing to a very local origin of *Alnus incana*. Similarly, the macrofossil and pollen or spore records of *Glyceria* sp. and *Onoclea sensibilis* (Poaceae and Monolete spores, respectively; Fig. 21) attest the local development of both taxa. Moreover, the macrofossil record during this period shows a maximal diversity of riparian and upland plants, associated to a low diversity of aquatic taxa (unit M5, Fig. 23). The sediment (unit S6, Fig. 20) is also characterised by high values and important fluctuations in the inorganic matter mass, which signify the sediment heterogeneity, mainly composed of sand and *Alnus* leaves. This zone, interpreted as an almost *in situ* litter of an *Alnus incana* riparian thicket, indicates a very low lake-level, with a local water depth less than 1 m.

3400-2600 cal. BP [140-120 cm]: high lake-level, 3 m lower than the present-day

The important increase in aquatic taxa diversity (unit M6, Fig. 23), associated with the deposition of gyttja (unit S7, Fig. 20), indicates a water level rise around 3400 cal. BP. The macrofossil record (Fig. 22) is characterised by numerous submerged, floating-leaved and emergent aquatic taxa (Characeae, *Najas flexilis*, *N. guadalupensis*, *Nymphaea odorata* subsp. *odorata*, *Nuphar variegatum*, *Potamogeton* sp., *Sagittaria* sp., *Sparganium* sp. and *Gloeotrichia* sp.), among which several thrive commonly in shallow waters, and by the persistence of *Carex* seeds, which could indicate the proximity of the shoreline. However, similarly to the unit M3, this macrofossil record more probably results from a transport from widespread aquatic and riparian formations developed on the marginal terraces (Fig. 18). Consequently, it is also considered to correspond to a high lake-level, about 3 m lower than the present-day.

2600-1800 cal. BP [120-80 cm]: moderate lake-level lowering

The incorporation of sand particles in the gyttja between 120 and 80 cm (unit S8, Fig. 20) probably indicates a lake-level lowering. However, this lowering was sufficiently slight so as to leave the macrofossil assemblages unchanged (unit M6, Fig. 22), thus representing similar plant zonation as previously.

1800-0 cal. BP [80-0 cm]: high lake-level

Although all vegetation belts present a taxinomic diversity decrease, aquatic taxa dominate the macrofossil record (unit M7, Fig. 23). Along with the sediment composition (unit S9, Fig. 20), this indicates an important rise in water level around 1800 cal. BP, and a persistent high water level afterwards. This period however differs from previous high lake-level stages by the abundance of Characeae, unidentified aquatic plants (leaves), *Gloeotrichia* sp., Chironomids and *Cristatella mucedo* (unit M7, Fig. 22). A final, man-made rise occurred from A.D. 1775 to the present, due to successive constructions of dams over the outlet (Fig. 18). This resulted in the present-day maximal level, two or three meters above the natural lake-level.

Palaeoclimate reconstruction

Modern data

The modern pollen dataset is obtained from Sawada (2001) and from the *Base de Données Polliniques et Macrofossiles du Québec* (BDPMQ). It consists in 1674 modern pollen spectra from northeastern North America between 90°W and 50°W and between 35°N and 80°N (Fig. 25). Thirty-one taxa (16 trees, 4 shrubs, 11 herbs) were selected for their climatic significance, evaluated by principal components analysis. Pollen percentages are calculated relative to the sum of those thirty-one taxa. The modern climate dataset (Fig. 26) is provided by Environment Canada (1994) and the U.S. National Climatic Data Centre (1994). Monthly mean temperature (1702 stations), monthly minimum temperature (1889 stations), total precipitation (2519 stations) and sunshine (120 stations) were smoothly interpolated across the 1674 pollen sites. These climatic parameters are used to estimate several bioclimatic variables: growing degree-days above 0 and 5°C are calculated after a daily interpolation of the monthly temperature by summing the daily temperature above 0 and 5°C, respectively, and actual evapotranspiration is obtained from the hydrological model of BIOME1 (Prentice *et al.*, 1992).

Chronology of the pollen diagrams

Pollen diagrams from seven sites (Fig. 17, Table 5), located in a 80 km radius around Montréal City, were chosen for their good chronological control (4-10 radiocarbon dates). Two regional palynological events considered as synchronous throughout the study area and used in the Lac Hertel chronology were also used here as chronological pollen markers: the *Tsuga* decline (~5500 cal. BP) and the *Ambrosia* rise (250-200 cal. BP) (Muller & Richard, accepted).

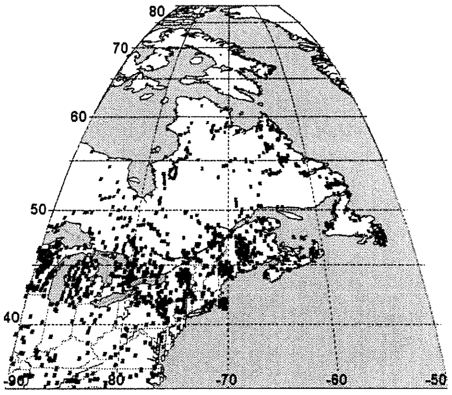
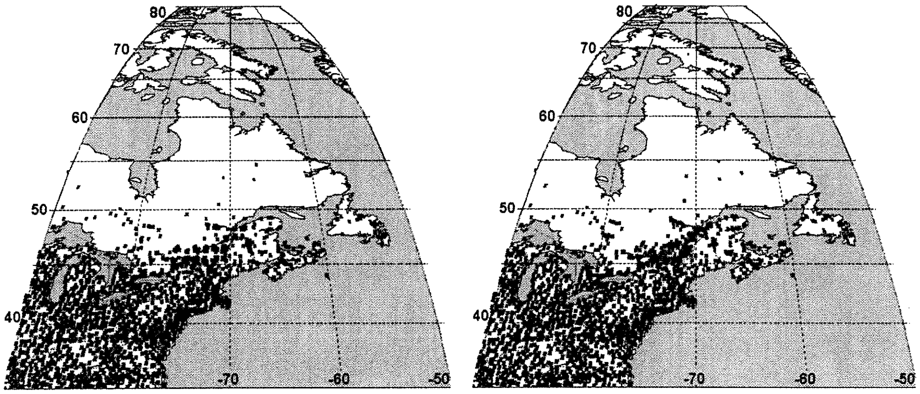
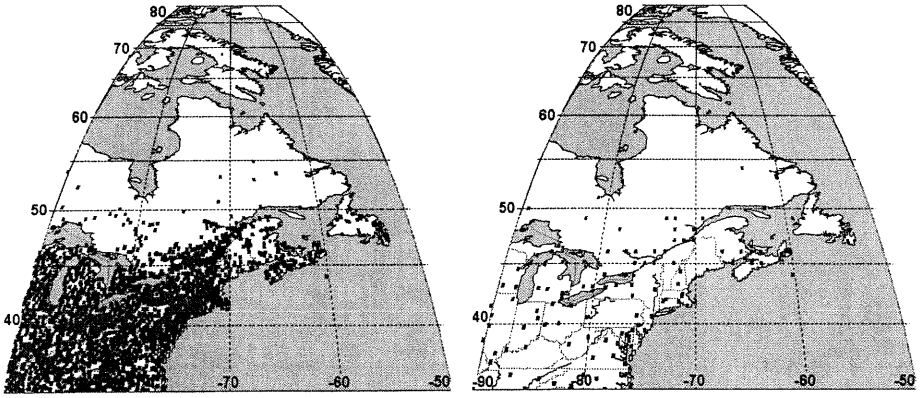


FIGURE 25. Distribution map of modern pollen sites



Mean monthly temperature

Minimal monthly temperature



Total precipitation

Sunshine percentage

FIGURE 26. Distribution map of modern climate data

TABLE 5. Site characteristics. Coordinates are noted in decimals.

Site name	Site code	Nature	Longitude	Latitude	Altitude (m)	Size (ha)	References
Atocas	Ato	Lake	-73.311	45.543	114	1.2	Gauthier, 1981; Muller & Richard, accepted
Bromont	Brom1	Lake	-72.670	45.265	135	50	Muller & Richard, accepted
Hertel	Herl	Lake	-73.153	45.547	173	28.4	This study
Shefford	She	Carr	-72.585	45.359	282	4.7	P. Richard, 1977a, 1978
St-Calixte	Cal	Lake	-73.868	45.961	261	1.7	Muller & Richard, accepted
Tania	Tania	Lake	-74.304	45.775	305	2	Richard, 1977b
Tortue	Tor	Lake	-73.317	45.546	137	2.5	Gauthier, 1981
Yamaska	Yam	Lake	-72.872	45.458	265	2	Muller & Richard, accepted

Note: Sites noted BDPMQ were studied by P.J.H. Richard and are provided by the *Base de Données Polliniques et Macrofossiles du Québec.*

Methods

The reconstruction of past climate in the St. Lawrence lowlands is done using the modern analogues method (Guiot *et al.*, 1989; Guiot, 1990), constrained by lake-level data (Guiot *et al.*, 1993; Cheddadi *et al.*, 1997; Magny *et al.*, 2001). The method consists of finding, in a dataset of modern pollen spectra, a few spectra similar to a fossil spectrum. The correlation between modern and fossil spectra is based on a chord distance (Overpeck *et al.*, 1985). The spectra, called analogues, are sorted out using the values of P-E (P: Annual precipitation; E: Annual actual evaporation): if the P-E of the analogue is higher than present and that the lake-level corresponding to the fossil spectra is lower, it is rejected and conversely. With this constraint, a lake-level change < 3 m (relative to the natural maximal level) is assumed to induce a change < 200 mm for P-E, and a lake-level change > 3 m is assumed to induce a change > 200 mm for P-E. The method is applied separately on all the pollen spectra of each site. The results for each pollen diagram were then averaged to time slices of 250 years to get a single regional reconstruction for our study area. Computer procedures were performed using the 3Pbase program (Guiot & Goeury, 1996).

Tsuga canadensis is likely to induce a bias in the climatic reconstruction, due to the drastic decline of its populations, believed to integrate a climate shift (Haas & McAndrews, 2000) and a phytophagous insect outbreak (Davis, 1981; Allison *et al.*, 1986; Bhiry & Filion, 1996). Davis *et al.* (2000) negate climate as a possible trigger to the *Tsuga* decline, but in order to evaluate the influence of this taxon on the climate reconstruction, the modern analogues method was applied by both including or excluding it from the pollen sum.

Climate-vegetation relationships

The ability of the modern analogues method to reconstruct climate is tested by comparing actual and estimated modern climate data (Fig. 27). Estimated values were generated for each modern spectra by sorting out the best analogues from the entire modern spectra dataset exclusive of the considered spectra. In this validation step, the lake-level constraint is evidently not used. The pollen spectra are more closely correlated with temperature variables (r^2 around 0.90), than with precipitation and the ones related to moisture-balance (r^2 around 0.72). Compared to previous similar works (Guiot *et al.*, 1993; Davis *et al.*, 2000), these r^2 point to a particularly close relationship between climate and vegetation in north-eastern North America. For instance, a very high relationship is evidenced for runoff (Fig. 27), which is theoretically considered to be a better predictor of lake-level changes than of vegetation changes (Guiot *et al.*, 1993; Harrison *et al.*, 1993).

Results

The modern analogues method was applied, successively unconstrained and constrained by lake-level changes, on a pollen dataset including *Tsuga canadensis* (Fig. 28). The results obtained by excluding lake-level changes show no significant differences for temperature and only minor departures for precipitation during the entire postglacial period. Paradoxically, the steep pollen decline of *Tsuga* recorded at 5500 cal. BP seems to reduce the amplitude of the reconstructed precipitation decrease (Fig. 28). The inclusion of *Tsuga* within the pollen sum consequently is considered to create no bias in the climate reconstruction. In comparison, the inclusion of lake-level changes appears to induce one major change between 5000 and 3500 cal. BP (Fig. 28). This period corresponds to the low lake-level inferred from the deposit of sand and *Alnus* leaves in profile Herl (Fig. 22). Our

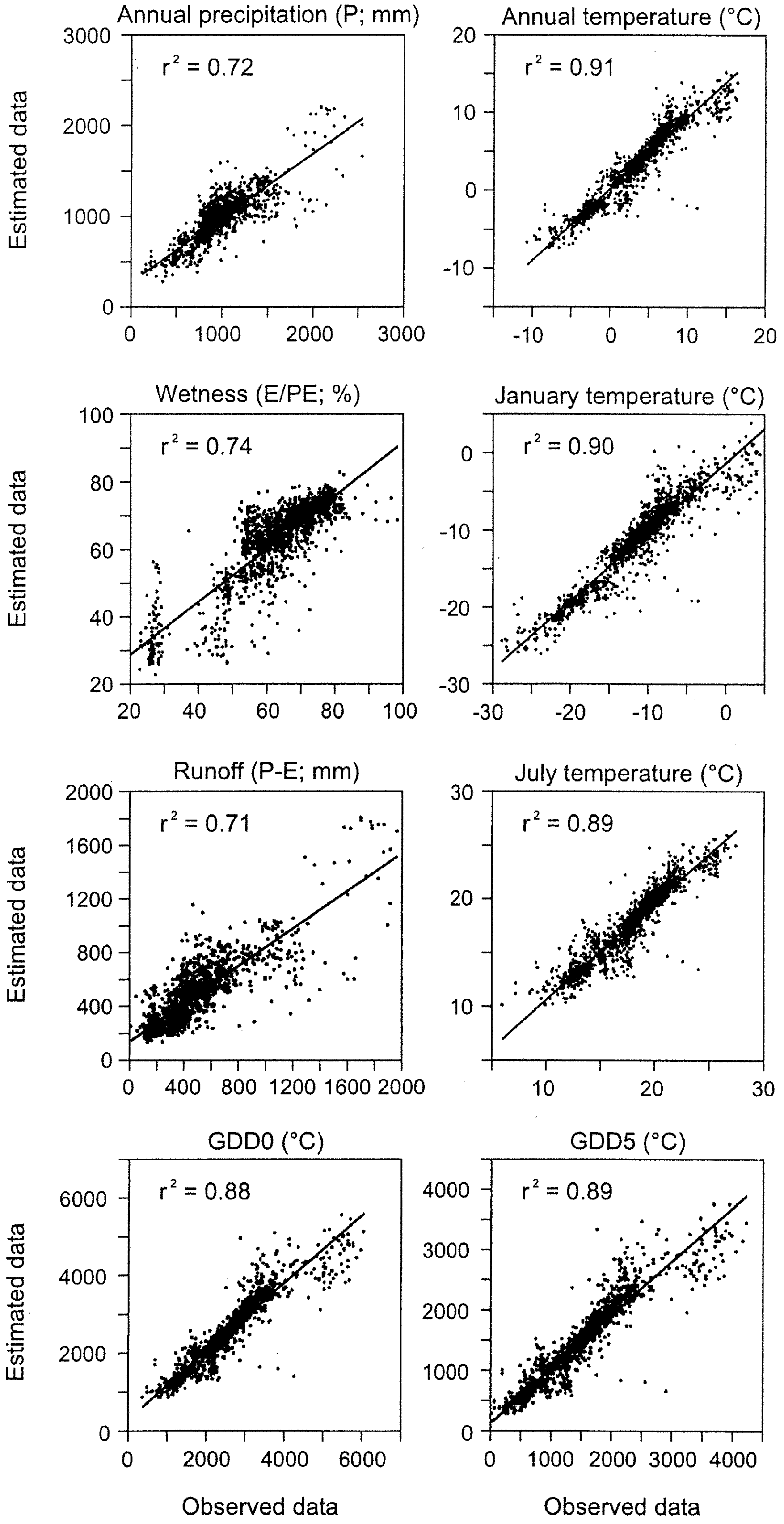


FIGURE 27. Modern climate-vegetation relationships. The wetness is approximated by E/PE and the runoff by P-E. E: Actual evaporation; PE: Potential evaporation; P: Annual precipitation; GDD0, GDD5: Growing degree-days above 0 and 5°C, respectively.

reconstruction associates a decrease in annual precipitation of about 150 mm and an increase in July temperature of about 1 °C (Fig. 28) to this event.

Discussion

Holocene climatic moisture patterns in the St. Lawrence lowlands

During the early Holocene, southern Québec was characterised by an important climatic dryness evidenced by very low annual precipitation (Fig. 28) and low lake-levels (Lavoie & Richard, 2000c; Fig. 24). Similar conditions were reported from lake-level status in northeastern United States (Harrison & Metcalfe, 1985; Harrison, 1989; T. Webb *et al.*, 1993) and from the eastward expansion of the prairie across the Midwest (Webb *et al.*, 1983; Winkler *et al.*, 1986). This widespread dry interval was believed to be the result of greater summer insolation (Kutzbach & Guetter, 1986). However, the constrained reconstruction (Fig. 28B) does not show the peak of aridity observed soon after 8000 cal. BP in northeastern United States (Webb *et al.*, 1983; Harrison & Metcalfe, 1985; Harrison, 1989) and in the western Great Lakes region (Davis *et al.*, 2000). During this, Lac Hertel exhibits a rise of the water level rapidly followed by a slight lowering (Fig. 24), which could reveal subtle climatic events not recorded by vegetation. Moreover, charcoal influx and pollen percentages of *Populus* and *Pinus banksiana* suggest shifts from dry to wet summers in southern Québec around 9000 cal. BP (Richard, 1994b) and between 8000 and 7000 cal. BP (Carcaillet & Richard, 2000, Carcaillet *et al.*, 2001). These shifts, not recorded by eastern North-American lake-levels, may be related to a greater sensitivity of lakes to winter rather than to summer precipitation (Vassiljev *et al.*, 1998; Carcaillet & Richard, 2000).

Between 7000 and 6000 cal. BP, annual precipitation increased in the St. Lawrence lowlands (Fig. 28). This feature is shown by both constrained and unconstrained reconstructions, pointing to a similar climate record by pollen and lake-levels. Southern Québec lakes have recorded a mid-Holocene water level rise: Lac Hertel between 6600 and 5000 cal. BP (Fig. 24) and Lake Albion between 6900 and 6100 cal. BP (Lavoie & Richard, 2000c). The different timing could be related to the different geographical

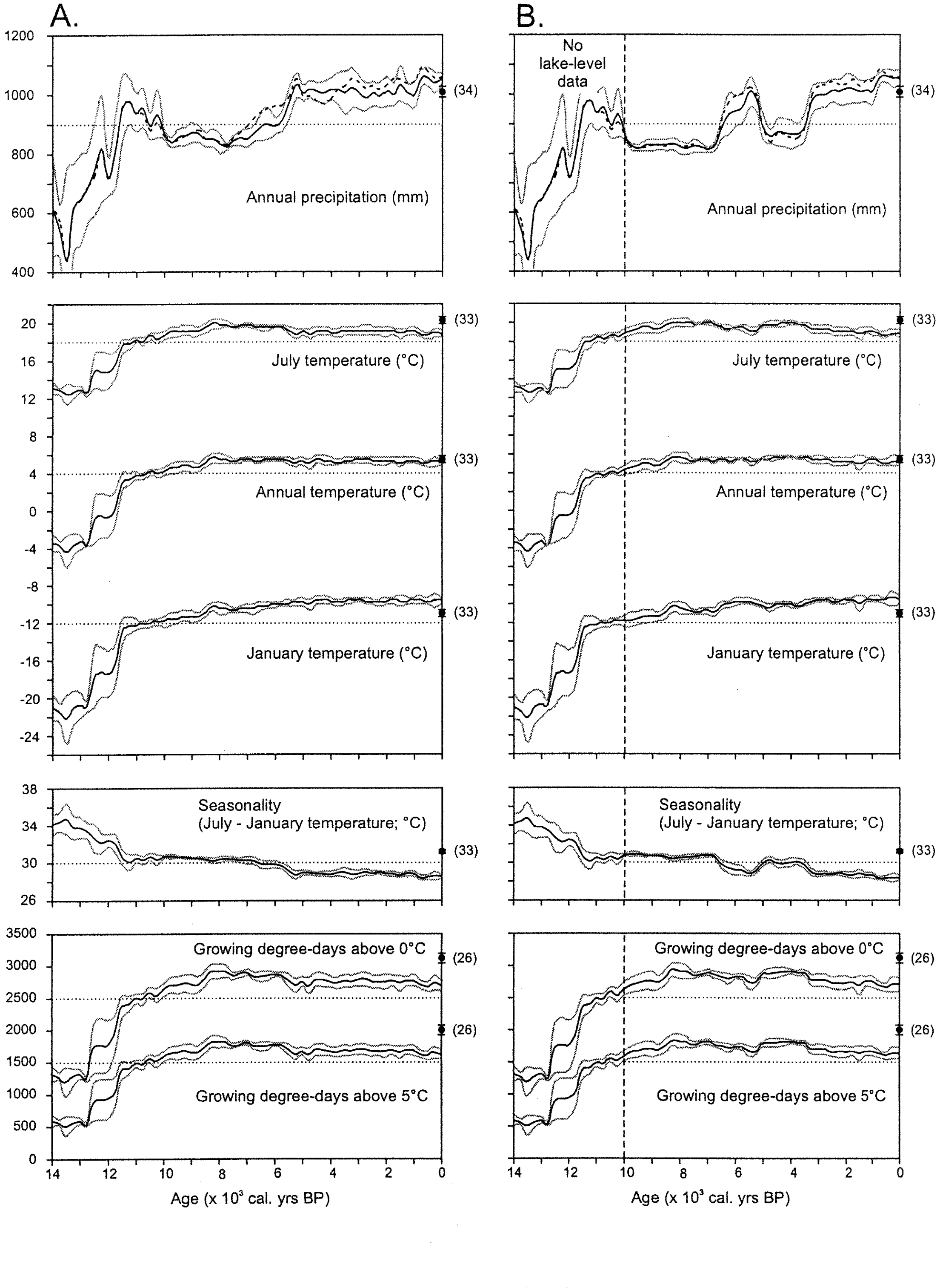


FIGURE 28. Climate reconstruction. A. Unconstrained. B. Constrained by lake-levels Full lines correspond to reconstructions including *Tsuga canadensis* (grey ones define the confidence intervals) and dashed lines to reconstructions excluding it (shown only for precipitation). Dots on the right axes of graphs represent actual modern values, calculated by averaging meteorological data from the Montréal lowlands area. The number of meteorological sites used in this calculation is noted in brackets.

situations or to dating problems, particularly at Lake Albion (Lavoie & Richard, 2000c). The increasing number of high lake-levels (Harrison, 1989; R.S. Webb *et al.*, 1993) and the increase in precipitation reconstructed from pollen data (Davis *et al.*, 2000) reveal similar trends in northeastern and midwestern USA, respectively. However, numerous low lake-levels were reported during the same period for Ontario (Szeicz & MacDonald, 1991; Yu & McAndrews, 1994; Bunting *et al.*, 1996; Yu *et al.*, 1996; Campbell *et al.*, 1997) and for Maine (Almquist *et al.*, 2001). These contradictory evidences denote the complex mid-Holocene climatic patterns at the scale of northeastern North America.

Although no major climatic change was recorded by pollen during this period (Fig. 28A), Lac Hertel sedimentary and macrofossil data (Figs. 20 and 22, respectively) attest to the occurrence of a dry climatic phase in southern Québec between 5000 and 3500 cal. BP (Fig. 28B). This regional dry interval was also recorded between 6100 and 4400 cal. BP in Lake Albion (Lavoie & Richard, 2000c). Similar conditions, documented between 7000 and 3000 cal. BP in Subarctic Québec (Payette & Filion, 1993), in Ontario (Sreenivasa & Duthie, 1973; Szeicz & MacDonald, 1991; Yu & McAndrews, 1994; Bunting *et al.*, 1996; Yu *et al.*, 1996, 1997; Campbell *et al.*, 1997), in Maine (Almquist *et al.*, 2001) and in Wisconsin (Winkler *et al.*, 1986; Baker *et al.*, 1992), point to a widespread climatic dryness. This discrepancy between pollen and lake-level records implies that the climatic change was not strong enough or did not influence the relevant parameters to significantly affect the structure of vegetation. Notably, we can note that the *Tsuga* decline dated at about 5500 cal. BP does precede and not follow the decrease in annual precipitation, as recorded by lake-levels. Although this timing could be due to the inertia in the lake's response to climate change, our results seem to indicate that the *Tsuga* decline was probably not related to a drought episode in southern Québec, as it may have been in southern Ontario (Haas & McAndrews, 2000).

The final rise of southern Québec lake-levels (Lavoie & Richard, 2000c; Fig. 24) is consistent with the trends observed in northeastern United States (Harrison & Metcalfe, 1985; Harrison, 1989; Almquist *et al.*, 2001) and in southern Ontario (Sreenivasa & Duthie, 1973; Szeicz & MacDonald, 1991; Yu & McAndrews, 1994; Bunting *et al.*, 1996; Yu *et al.*, 1996, 1997; Campbell *et al.*, 1997). This widespread rise of lake-levels suggests a general increase in annual precipitation throughout northeastern North America, which is supported at the regional scale by our climatic reconstruction (Fig. 28). Finally, lake-level modelling (Vassiljev *et al.*, 1998) and climatic multiproxy comparison including fire, lake-level and $\delta^{18}O$ (Carcaillet & Richard, 2000) suggest that high lake-level status result from increasing winter precipitation.

Holocene temperature patterns in the St. Lawrence lowlands

The Late-glacial in southern Québec was characterised by the occurrence of pro-glacial lakes and seas (Hillaire-Marcel, 1979; Parent & Occhietti 1988, 1999). Associated with the inlandsis proximity, these cold waters induced harsh climatic conditions throughout the St. Lawrence lowlands region (Gangloff *et al.*, 1971; Gangloff, 1981). According to our climatic reconstruction (Fig. 28), the mean annual temperature of the late-glacial period would have been around -4°C (about 9°C below present). The first increase in temperature was recorded around 12500 cal. BP (Fig. 28), simultaneously to both the afforestation and the beginning of organic sedimentation in lakes in southern Québec (Richard, 1994b). This temperature increase which was also reported at this date in southern Minnesota and Wisconsin (Webb & Bryson, 1972; Webb & Clark, 1977) and in sea surface temperature in the Gulf of St. Lawrence (Vernal *et al.*, 1993) points to a widespread climate improvement around 12 500 cal. BP probably driven by higher summer solar radiation (Kutzbach & Guetter, 1986).

Continental and oceanic temperatures increased abruptly until 11 500-10 500 cal. BP (Webb & Bryson, 1972; Webb & Clark, 1977; Vernal *et al.*, 1993; Fig. 28). At that time, the sea surface and the midwestern USA temperatures were shown to attain roughly their modern values (Webb & Clark, 1977; Vernal *et al.*, 1993), or even a slighty higher (Webb & Bryson, 1972). The St. Lawrence lowlands were characterised by a temperature 1.5°C below present at 11 000 cal. BP (Fig. 28). This difference could indicate the proximity of inland ice and of catabatic winds and the influence of cold water inflow from glacial lakes discharge, which was particularly important in the region located north of the Great Lakes (Lewis & Anderson, 1989; Anderson & Lewis, 1992).

Between 9000 and 4500 cal. BP, the spread of *Pinus strobus* and other thermophilous species was interpreted as indicating a warmer climate (Richard, 1994b). Annual temperature is effectively shown to have increased progressively between 11 500 and 8000 cal. BP (Fig. 28). A similar pattern is also clearly shown by other temperature-related parameters, such as growing degree-days (Fig. 28). A pronounced climate cooling however was reported between 8900 and 8300 cal. BP from an isotopic record in an annually-laminated Minnesota lake (Hu *et al.*, 1999). Our results seem to also show a slight temperature decrease from 9000 to 8250 cal. BP (Fig. 28). Mainly marked by July temperatures (Fig. 28), this climate change may confirm the influence of atmospheric circulation and especially of summer cooling of Arctic airmasses, as suggested by Hu *et al.* (1999). After this cool episode, annual temperature in the St. Lawrence lowlands reached their optimal values at about 0.5°C above the estimated modern value and 0.2°C above the actual value (Fig. 28). This thermal optimum was also reported to span 9000 to 7000 cal. BP in the western Great Lakes region (Davis *et al.*, 2000) and in the Gulf of St. Lawrence, where it was interpreted as the regional signature of the "Hypsithermal" (Vernal *et al.*, 1993). The St. Lawrence lowlands experienced from 5750 to 5000 cal. BP another summer cooling, which culminated at 5500 cal. BP (Fig. 28) simultaneous to the *Tsuga* decline.

After 4500 cal. BP, Richard (1994b) notes that the increase in boreal elements (*Picea*, *Abies*) within the regional vegetation could indicate a climate worsening, although it could also be related to autogenic development of peatlands, as reported for the Frontenac peatland (Lavoie & Richard, 2000a). This observed progressive general summer cooling trend is effectively attested over the last 4000 years in the Gulf of St. Lawrence (Vernal *et al.*, 1993) and in midwestern USA (Webb & Bryson, 1972; Webb & Clark, 1977; Davis *et al.*, 2000). Moreover, this pattern appears to be associated in southern Québec (Fig. 28) and in the western Great Lakes region (Davis *et al.*, 2000) with a continuously increasing winter temperature, which could explain the persistence of thermophilous species (Richard, 1994b).

Conclusions

This study provides the first quantitative climate reconstruction for southern Québec. It allows specifying regional values and ranges of climatic events. Most of these intervals was shown to affect simultaneously the Gulf of St. Lawrence, the southwestern Québec lowlands and the midwestern USA. They are: 1) a dry and cold late-glacial episode, especially harsh in southern Québec due to the proximity of inland ice, and to the influence of pro-glacial lakes and seas; 2) an abrupt climate improvement between 12 500 and 11 000 cal. BP probably catching up the increasing summer insolation during a regional ice retreat; 3) an important dry period from 10 000 to 6500 cal. BP; 4) a brief cooling event between 9000 and 8000 cal. BP, only observed from upland records and possibly related to a summer cooling of Arctic airmasses; 5) a temperature optimum around 8000 cal. BP, which could represent a regional "Hypsithermal"; 6) and finally, a progressive decrease in summer temperature and an increase in (winter ?) precipitation over the last 4500 years. By contrast, southern Québec experienced an increase in precipitation from 6500 to 5000 cal. BP, followed by a dry climatic interval between 5000 and 3000 cal. BP. This trend was not obviously evidenced throughout northeastern North America, pointing to a complex mid-Holocene climatic pattern at this scale.

The agreement of our reconstruction with previous results points to the reliability of the application of the modern analogues method at the regional scale. However, the south-north vegetation spread following the deglaciation of southern Québec was not in total equilibrium with climate: several other factors, such as seed dispersion characteristics or physiography (Muller & Richard, accepted), have played a role in this process. This implies carefull consideration of the minor fluctuations of reconstructed curves, especially during late-glacial and early postglacial periods. The reconstructed climate, though not representing exactly the actual past climate, corresponds at least to the minimal conditions necessary for the postglacial vegetation migration. Moreover, compared with modern values, our reconstruction appears to underestimate the seasonal contrasts of temperature (July-January) and the growing-degree days values (Fig. 28). Despite this bias, the annual temperature amplitude presents an obvious decreasing trend over the whole postglacial period, particularly well marked during the Late-glacial (this trend surprisingly presents a very strong inverse relationship with annual precipitation: $r^2 > -0.97$). Very similar patterns were observed for sites located at similar latitudes on the shore of Lake Michigan (Davis *et al.*, 2000). This enforces the hypothesis of a linkage between temperature seasonality and Earth orbital parameters suggested by these authors on the basis of a latitudinal transect. However, the seasonality is also known to represent a measure of continentality (Guiot *et al.*, 1993), and thus to depend to the proximity of water bodies. The influence of the Great

Lakes is clearly shown by comparing an inland to an island site, both located at similar latitudes (Davis *et al.*, 2000). The similarity between Holocene seasonality trends in the St. Lawrence lowlands and in Great Lakes shores suggests that the St. Lawrence River could have influenced the past temperature pattern within the southern Québec lowlands.

Moreover, our results shed light on two particular vegetational features of the St. Lawrence lowlands. First, the decline of *Tsuga canadensis* was attributed by numerous studies to a pathogen outbreak (e.g., Davis, 1981; Allison *et al.*, 1986; Bhiry & Filion, 1996) and, on the basis of a lake-level reconstruction in southern Ontario, to a drought episode (Haas & McAndrews, 2000). According to our climatic reconstruction, the *Tsuga* decline in southern Québec appears to have occurred previous a precipitation decrease inferred from lake-levels and simultaneously due to a pronounced summer cooling. These results suggest that the influence of climate is equivocal and occurred in various ways depending on location. Its role was probably limited in the *Tsuga* decline to favour or obstruct locally the spread of the hemlock looper (*Lamdina fiscellaria*, Lepidoptera). Second, macrofossil analyses conducted on the lateral core of Lac Hertel reveal past occurrences of several taxa unmentioned today on Mont St. Hilaire (Maycock, 1961). With the exception of Arctic species, these taxa are *Ceratophyllum echinatum*, *Decodon verticillatus*, *Najas guadalupensis* and *Nymphaea odorata* subsp. *odorata*. Seeds of *N. guadalupensis* were found sporadically between 370 and 24 cm (Fig. 22). The probable persistence of this species during more than six millennia suggests that it could be present albeit unknown today on Mont St. Hilaire. The other species were encountered only once in the whole profile: *D. verticillatus* (336 cm), *C. echinatum* (240 cm) and *N. odorata* subsp. *odorata* (120 cm). In regards to the present-day rarity of *C. echinatum* and *N. guadalupensis* in Québec (Bouchard *et al.*, 1983; Lavoie, 1992), their subfossil signatures in Lac Hertel may reveal past modifications of their northern distribution limit, likely induced by climate change.

Finally, this study highlights seasonal patterns in postglacial climate history. It corroborates the ideas that pollen and charcoal data would mainly indicate summer precipitation (Carcaillet & Richard, 2000) while lake-levels would rather depend on the winter precipitation balance (Vassiljev *et al.*, 1998). These relationships make sense, especially for northern zones characterised by harsh and long winters. The high amount of water liberated by the spring snow melt (50-60 % of the annual runoff in southeastern Québec lowlands; Plamondon *et al.*, 1996) largely exceeds the water-holding capacity of soils. It is believed to have contributed more to lake-levels than to groundwater recharge. In comparison, summer precipitation, which are regularly spread during the whole season, contributes mostly to the soil recharge and consequently to the vegetation development. These considerations point to the importance of merging summer and winter precipitation proxy-data by the way of pollen-based precipitation reconstructions constrained by lake-levels.

Acknowledgements

We thank Hans Asnong, Alayn Larouche, Isabelle Legeai and Nicole Morasse for field and laboratory assistance, and Martin Lechowicz, Marc-André Langlois and the World Biosphere Reserve of Mont St. Hilaire for the authorisation and the facilities to work. The research has been supported by the Natural Sciences and Engineering Research Council of Canada and the Fond pour la Formation de Chercheurs et l'Aide à la Recherche of Québec, grants to P.J.H. Richard. S.D. Muller benefited from a French fellowship (Allocation de recherche du Ministère de la Recherche et de la Technologie) and was supported by the Peatland Carbon Study Program (NSERC-Strategic). Thoughtful comments by Hans Asnong, Christopher Carcaillet, Gunnar Digerfeldt, Alayn Larouche and Martin Lavoie were greatly appreciated.

POSTGLACIAL DEVELOPMENT OF A BOREAL BOG (SOUTHERN QUÉBEC): A SPATIO-TEMPORAL RECONSTRUCTION BASED ON PACHYMETRY, SEDIMENTOLOGY, MICRO- AND MACROFOSSILS

Serge D. Muller, Pierre J.H. Richard & Alayn C. Larouche

The Holocene – Submitted

Abstract

This study examines the past relationships of autogenic processes in the development of a southern Québec bog. Local vegetation, surface moisture and peat accumulation were independently reconstructed from palaeoecological analyses (microfossils, macrofossils, sedimentology, loss-on-ignition and radiocarbon dating) conducted on seven profiles. The observed successional stages, from pond to marsh, rich fen, poor fen then bog, represents a typical hydrosere. The early wetland development provides an obvious illustration of the facilitation model: each community modified hydrological conditions by the way of peat accumulation which then favoured the onset of the following stage. Our results moreover highlight the difficulty for tackling past autogenic relationships from palaeoecological data. No direct link is evidenced among vegetation, moisture and peat accumulation, while all of them clearly reveal the initial influence of minerotrophic groundwater supply. Notably, the decrease in this allogenic influence is well translated by the increasing asynchrony of vegetation changes. Surface moisture was also highly asynchronous, especially since 7000 cal. BP. The integration of the seven profiles shows several moisture fluctuations at the peatland scale but none seems to be related to climate. In contrast with moisture and vegetation, net peat accumulation presents similar trends in all marginal profiles and at the peatland scale. As a consequence of the initial minerotrophic supply, peat accumulation rates were high during the early Holocene and low during the mid- and late Holocene. However, the last centuries also presented an apparent increase in peat accumulation rate, which is attributed to the lack of decomposition in the acrotelm layer.

Key words: Peatland development, autogenic factors, peat accumulation, observation scale, groundwater supply, southern Québec.

Résumé

Développement postglaciaire d'une tourbière boréale (Québec méridional) : une reconstitution spatio-temporelle basée sur des analyses pachymétriques, sédimentologiques, micro- et macrofossiles.

Cette étude examine les relations passées entre les processus autogènes du développement d'une tourbière du Québec méridional. La végétation locale, l'humidité de surface et l'accumulation de la tourbe furent indépendamment reconstitués à partir d'analyses paléoécologiques (microfossiles, macrofossiles, sédimentologie, perte-au-feu et datation radiocarbone) réalisées sur sept profils. La succession végétale observée – étang, marais, fen riche, fen pauvre et bog – représente une hydrosère typique. Le développement initial de la zone humide constitue une illustration du modèle de facilitation : chaque communauté a clairement modifié les conditions hydrologiques par le biais de l'accumulation de tourbe, et ainsi favorisé le développement du stade suivant. En outre, nos résultats mettent en évidence la difficulté d'appréhender les relations autogènes passées à partir de données paléoécologiques. Aucun lien direct n'a été mis en évidence entre végétation, humidité et accumulation de tourbe, bien que les trois paramètres révèlent clairement l'influence de la nappe phréatique minérotrophe. Notamment, la réduction de cette influence allogène est bien traduite par l'augmentation de l'asynchronisme des changements de végétation. L'humidité de surface fut également fortement asynchrone, en particulier depuis 7000 cal. BP. L'intégration des sept profils met en évidence plusieurs fluctuations de l'humidité à l'échelle de la tourbière, dont aucune n'apparaît reliée au climat. Contrairement à l'humidité et à la végétation, l'accumulation nette de la tourbe présente des tendances similaires dans tous les profils marginaux et à l'échelle de la tourbière. En réponse à l'enrichissement minérotrophe initial, les taux d'accumulation de tourbe furent élevés durant l'Holocène ancien et faibles durant l'Holocène moyen et l'Holocène récent. Toutefois, les derniers siècles présentent une apparente augmentation du taux d'accumulation, attribuée à la faible décomposition au sein de l'acrotelme.

Mots clés : Développement d'une tourbière, facteurs autogènes, accumulation de tourbe, échelle d'observation, apport de la nappe phréatique, Québec méridional.

Introduction

Peatlands or mires, defined in a general sense as peat-accumulating ecosystems, can be subdivided into bogs and fens, the latter being called swamps when treed (Moore & Bellamy, 1974; Gore, 1983). Both peatland types are differentiated by the nature of water supply: minerotrophic in the case of fens and ombrotrophic in the case of bogs. Fens and bogs were proposed to represent the ultimate stages of the succession from limnic to terrestrial communities: lake, pond, marsh, fen, then bog (Weber, 1908; Clements, 1916). This terrestrialisation process is believed to reflect the concept of facilitation developed by Connell & Slatyer (1977): each stage modifies local conditions, especially water table, and then favours the onset and spread of the following stage. However numerous departures to this ideal sequence were reported: the hydrosere is often only partially realised, successional inversions are not rare and peatlands can initiate by an alternate way, called paludification (Tallis, 1983). Terrestrialisation nevertheless appears to represent a relatively common pathway of wetland change in boreal zones (Korhola, 1995a).

It is notably the case of the peatland this paper focuses on in the aim to explore the past autogenic dynamics in regard to hydroseral and facilitation models. Autogenic dynamics were essentially stressed by ecological studies, which point to strong modern relationships between plants and surface moisture, at the microtopographical scale (e.g. Vitt *et al.*, 1975b; Andrus *et al.*, 1983; Malmer, 1986) and at the peatland scale (e.g. Malmer 1962, 1986; Jeglum, 1971; Glaser, 1992a, 1992b; Gorham & Janssens, 1992). Past relationships are assumed to have been similar to modern ones and to be able to be tackled by palaeoecological studies (Moore, 1986). However, these assumptions were not verified and, if the first one can be considered as a conceptually acceptable postulate, the second one, more methodological, needs to be discussed. With this aim, we independently reconstruct local vegetation (from pollen, spores and plant macrofossils), surface moisture (from testate amoebae, fungi, algae and humification) and organic accumulation (from loss-on-ignition, radiocarbon dates and pachymetry).

These data, obtained from several profiles, moreover allow stressing the influence of observation scale on palaeoecological reconstructions. The study of Belyea & Warner (1995) demonstrates important variations in peat accumulation reconstruction at a microtopographical scale. Our study provides an opportunity to compare reconstructions at the community or seral stage scale, assumed to be adequately translated by the analysis of a single profile, and at the ecosystem scale, tackled by integrating all studied profiles.

Geographic setting

The Montréal lowlands, southern Québec (Fig. 29) mainly comprise a flat plain of marine clay, deposited above tills on sandstone-dolomitic bedrock (Laverdière *et al.*, 1972). Several intrusive Monteregian hills emerge from the plain roughly along an east-west axis. Lakes are rare in the lowlands and, except for the now filled-up Lac Romer, they are located on Monteregian hills (Fig. 29). The regional vegetation constitutes the northern limit of deciduous forests dominated by *Acer saccharum*, in association with *Betula alleghaniensis* on highlands, *Tilia americana* in the north-eastern part of the plain and *Carya cordiformis* south-westwards of Sorel (Grandtner, 1966; Richard, 1987). The mature, mesic formations also comprise *Fagus grandifolia*, *Pinus strobus*, *Tsuga canadensis* and *Fraxinus pennsylvanica* (Grandtner, 1966), while secondary facies are characterised by *Betula papyrifera*, *B. populifolia*, *Populus tremuloides* and *P. grandidentata* (Bergeron *et al.*, 1988; Brisson *et al.*, 1988). Depressions are occupied by peatlands, which harbour boreal elements such as *Picea mariana*, *Larix laricina* and *Abies balsamea* (Joyal, 1970; Jean & Bouchard, 1987). Dry sites, mainly located on lowland sandy deposits and on highland summits, are dominated by *Pinus banksiana*, *P. strobus*, *Thuja occidentalis* and *Quercus rubra* (Bouchard & Maycock, 1978). The transitional nature of these forests was emphasised by Maycock (1961) and Bouchard & Maycock (1978), who noted the boreal and temperate elements of which they are constituted.

Site description

Mirabel bog, previously called St. Canut bog (MacPherson, 1966), is an ombrotrophic 2.15 km^2 raised bog located 35 km northwest of Montréal island (Fig. 29). It comprises an elongate excentric non-forested zone and several peripheral carrs of various extent

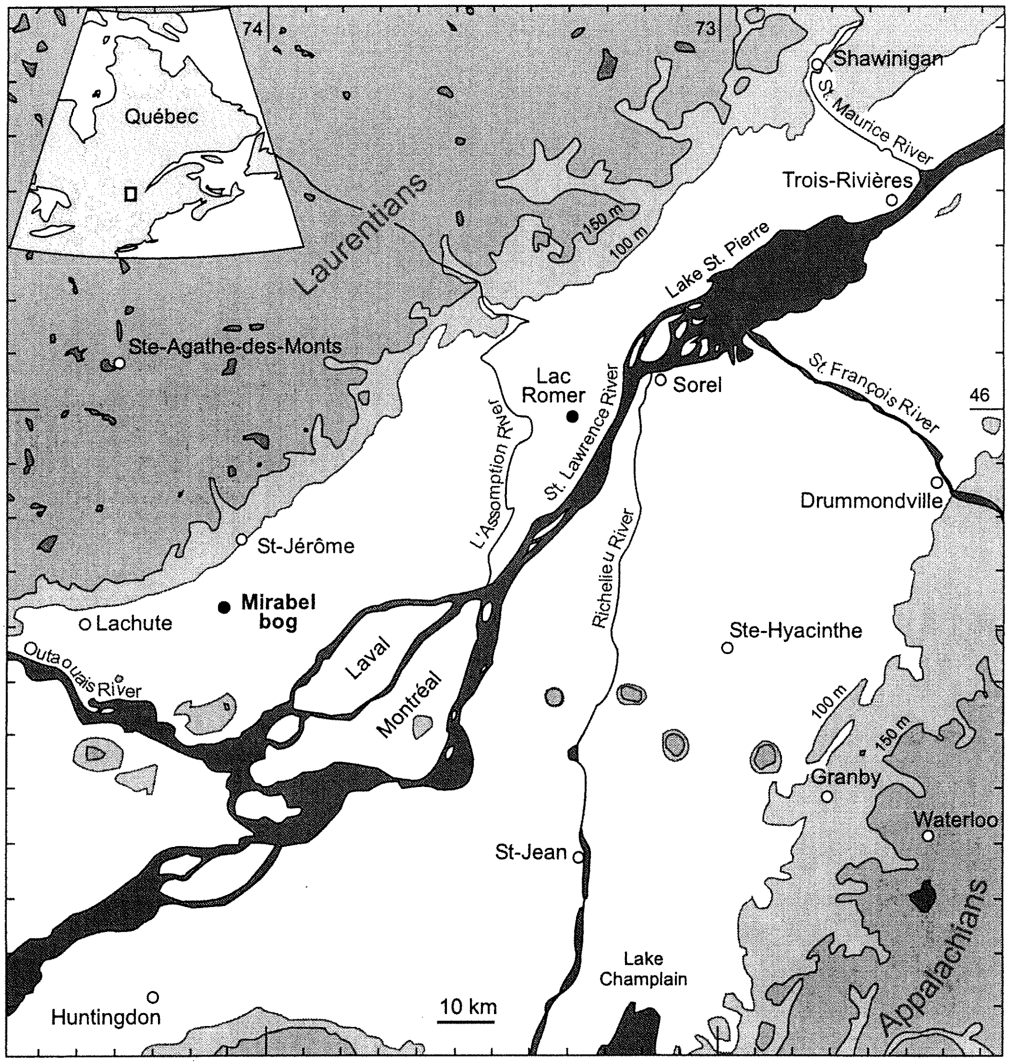


FIGURE 29. Location of Mirabel bog in the St. Lawrence lowlands, southern Québec

(Fig. 30). The non-forested zone is dominated by hummocks of *Sphagnum capillifolium* and *Chamaedaphne calyculata*, in association with *Sphagnum magellanicum*, *Polytrichum strictum*, *Kalmia angustifolia*, *K. polifolia*, *Rhododendron groenlandicum*, *Viburnum cassinoides* and *Eriophorum vaginatum* subsp. *spissum*. Moreover, scattered *Larix laricina*, *Picea mariana* and *Betula populifolia* trees are present in the open zone and in places, constitute some small treed islands.

A belt of *Aronia melanocarpa* and *Nemopanthus mucronatus* forms the ecotone between open bog expanse and forested margins. Located in the northern part of the peatland, the largest carr is mainly constituted by dense formations of *Picea mariana* and *Larix laricina* overlying a discontinuous bryophytic carpet of *Pleurozium schreberi*, *Sphagnum russowii*, *S. magellanicum* and *S. fimbriatum*. Other marginal carrs are dominated by *Acer rubrum* and *Abies balsamea*, with understories of *Osmunda cinnamomea*, *Dryopteris intermedia* and *Lycopodium annotinum*. Besides, the western carr harbours an important population of *Cypripedium acaule*, and *Woodwardia virginica*, an endangered species in southern Québec (Ministère de l'Environnement du Québec, 2000). Finally, an eroded, enriched zone, located on an ancient drainage ditch, is colonised by *Sphagnum cuspidatum*, *Eriophorum vaginatum* subsp. *spissum* and *Hamatocaulis palustre*.

Material and methods

Field methods

The basin morphometry was reconstructed from 140 peat depth (pachymetric) measurements and 120 surface topography surveys (Fig. 31). Fourteen basal peat monoliths were taken along five transects (Fig. 30) for radiocarbon dating. Seven profiles (Mir1 to Mir7; Figs. 30 and 32) were collected in 1999 and 2000 with a Russian sampler (Jowsey, 1966). For each coring, successive segments were alternately taken in two different holes, 30 cm apart. The seven cores were investigated for microfossils, macrofossils, sedimentology and loss-on-ignition. Number and location of profiles were determined from modern vegetation distribution and basal topography, with the aim to characterise functional units of the peatland. Modern vegetation and basal topography are effectively well

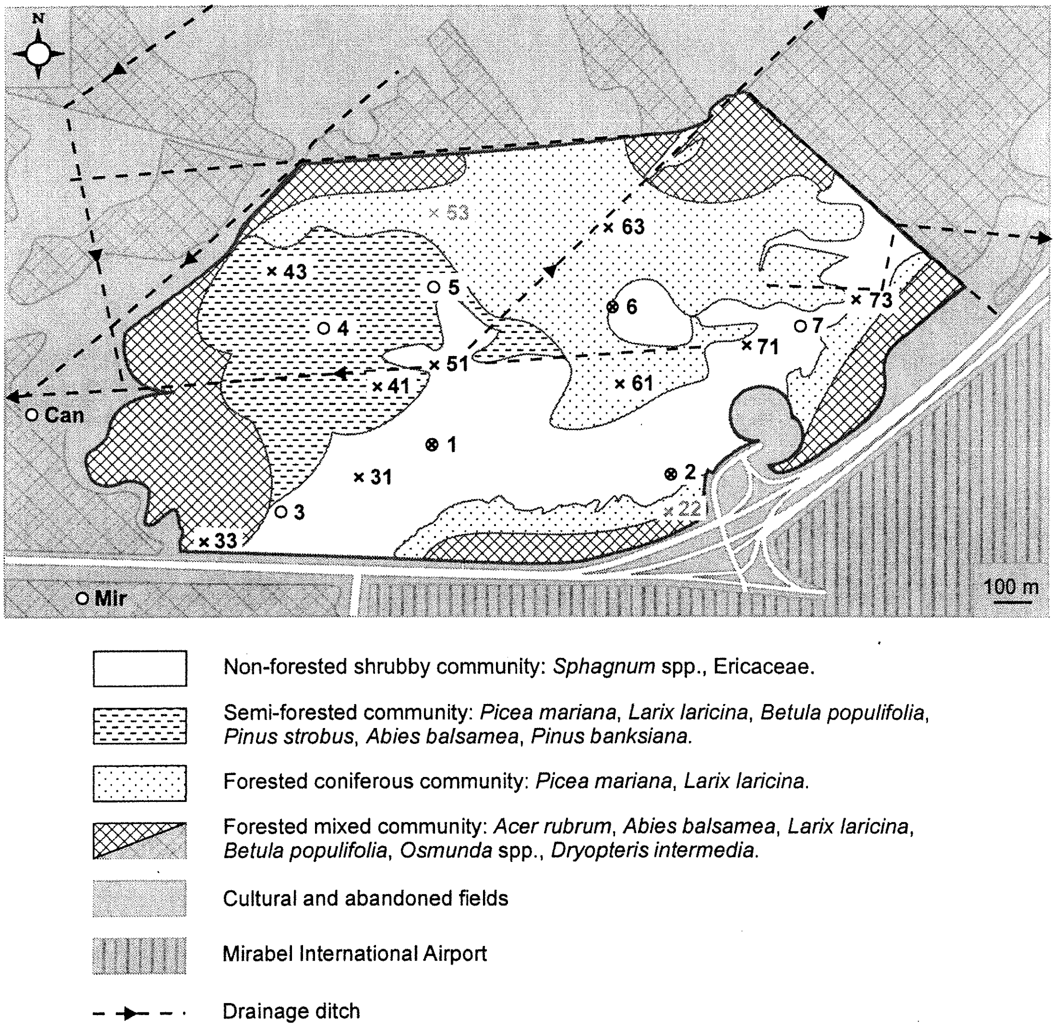


FIGURE 30. Vegetation of Mirabel bog and location of studied profiles (dots) and basal dates (crosses). Rejected basal dates are noted in grey.

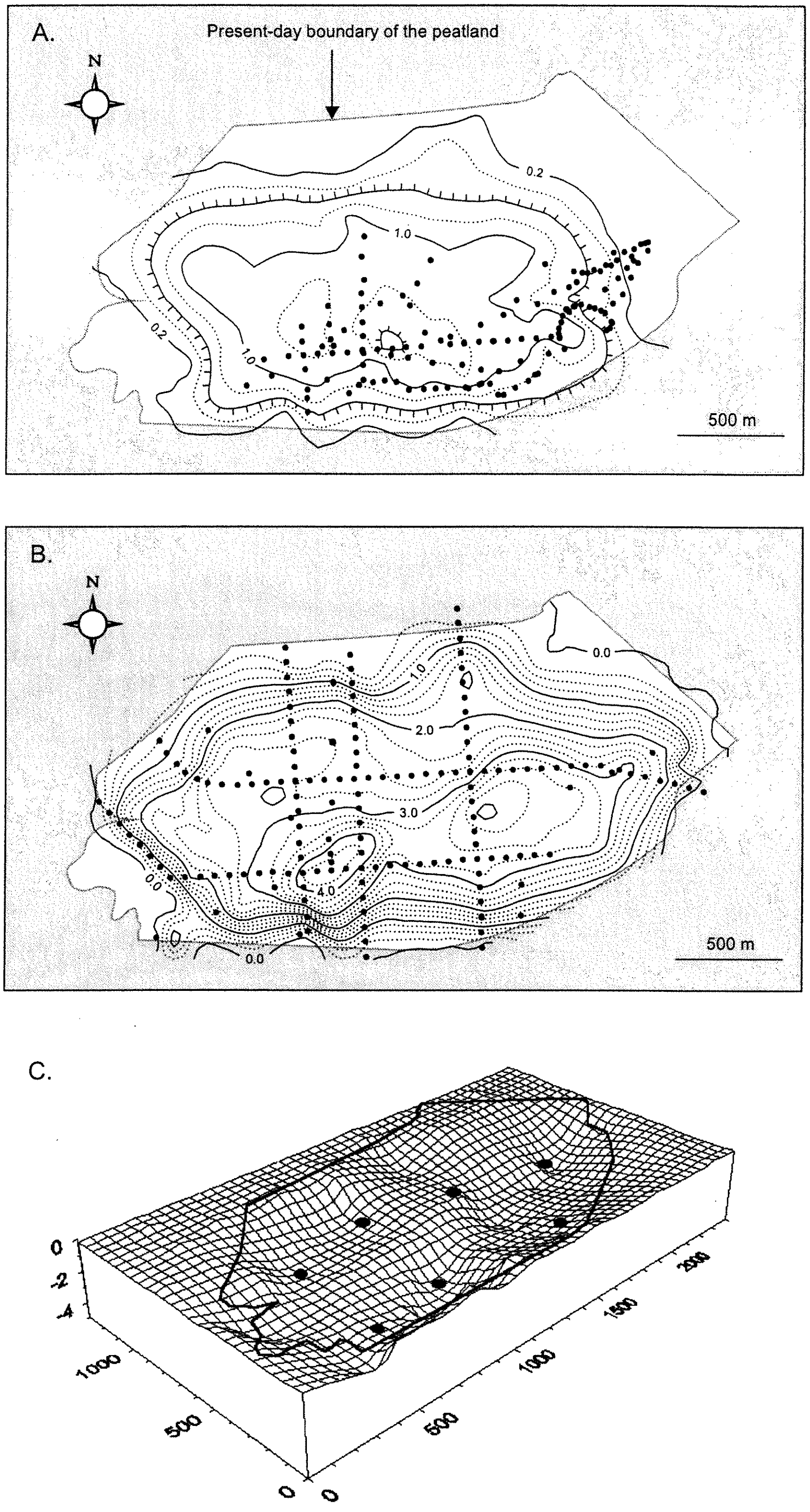


FIGURE 31. Surface topography (A), peat depths (B) and basin morphometry (C) of Mirabel bog. The solid lines represent the modern boundary of the peatland and dots localise planimetric measurements (A), pachymetric measurements (B) and studied profiles (C). Distances and depths are in m. Relative altitudes outside the peatland are assumed to equal 0 m.

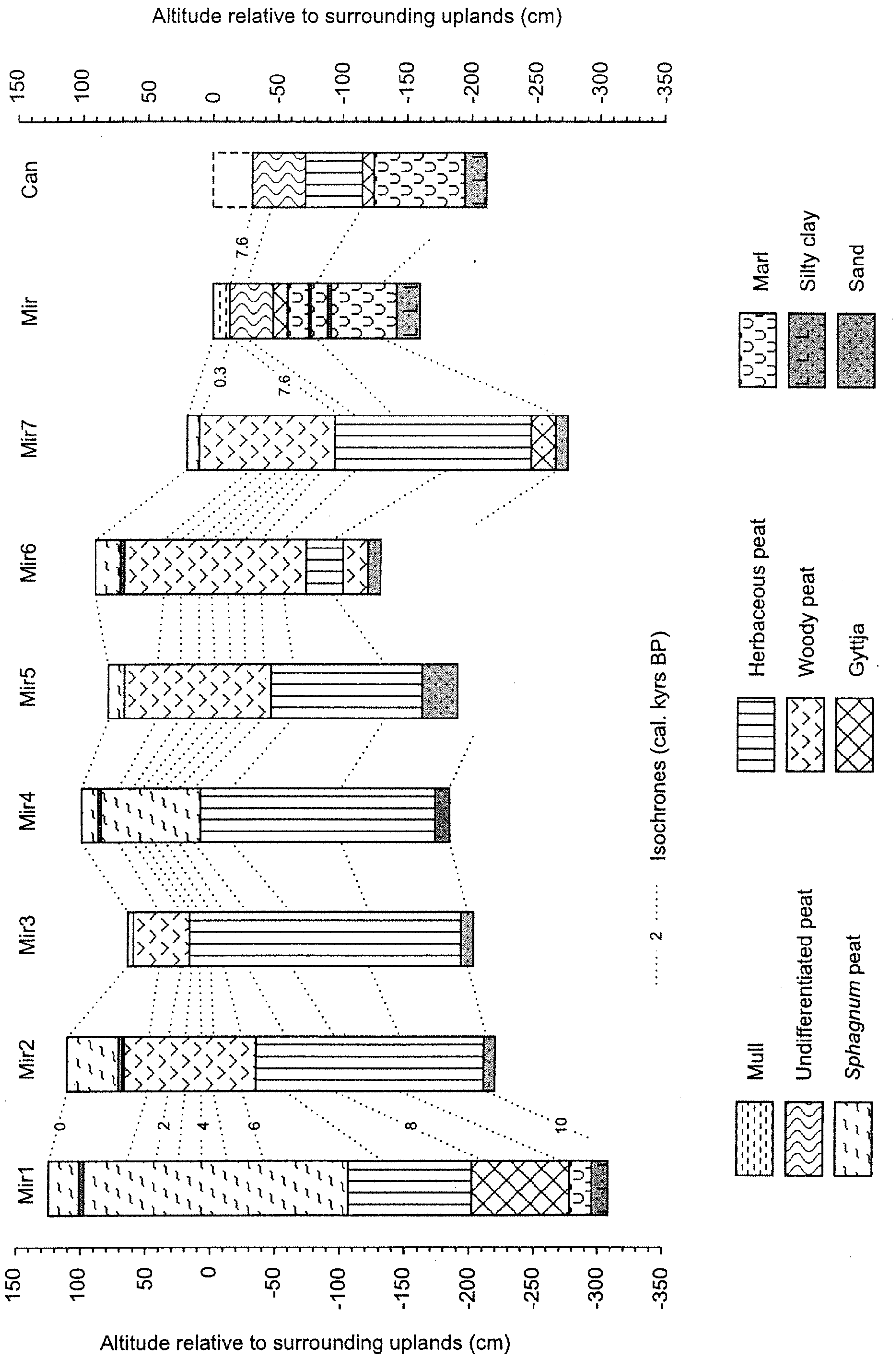


FIGURE 32. Stratigraphy and temporal correlation of Mirabel bog profiles, including the two peripheral profiles: Can (MacPherson, 1966) and Mir (Richard, P.J.H., unpublished data, 1974). The altitude of the surface of the profiles Mir and Can is assumed to be equal to surrounding uplands.

correlated (Figs. 30 and 31), which suggests that spatial patterns did not significantly change during the Holocene. Two previous unpublished pollen diagrams from surrounding zones were also used (Figs. 30 and 32). Profile Can was collected with a Hiller borer and analysed for pollen by MacPherson (1966). Profile Mir was directly taken on a section along a ditch in 1973, and also analysed for pollen (P.J.H. Richard, unpublished data, 1974).

Laboratory methods

Pollen extraction was based on the conventional protocol according to Faegri & Iversen (1989), but without the hydrofluoric acid treatment. Pollen identifications were generally based on the pollen atlases of Richard (1970) and McAndrews *et al.* (1973). Pollen percentages are calculated relative to a sum excluding hydrophilous species' pollen and Pteridophyta spores (Berglund & Ralska-Jasiewiczowa, 1986). Except some particularly poor samples, the minimum number of pollen grains counted and included in this sum is 500 for Mir and Mir1, 300 for Mir2 to Mir7, and 200 for Can. Pollen concentrations (grains.cm^{-3}) are based on volumetric measurements (Davis, 1965).

Palynomorphs (testate amoebae, fungi, algae…) were counted during pollen analysis. Chemical treatments were reported to destroy selectively a part of the testacean thanatocoenose (Tolonen, 1986; Hendon & Charman, 1997; Charman *et al.*, 2000). However, no hydrofluoric acid was used here during chemical treatments and previous studies have shown that the general ecological significance was relatively well preserved by the remaining assemblages (e.g., Aaby & Tauber, 1975; Kuhry, 1997; Lavoie & Richard, 2000a). Because of the generally low palynomorph sums, we chose to present results, not in percentages as it is usually made (Tolonen, 1986; Warner, 1990), but in volume-based concentrations standardised to 1 mm^3 to avoid big figures (Van Geel, 1986).

Macrofossil extraction was realised on 3 cm thick samples by sifting under soft water jet (Grosse-Brauckmann, 1986). Volumes analysed, comprised between 6 and 180 cm^3, were standardised to a volume of 100 cm^3.

The notation of taxa follows Birks & Birks (1980): the mention "type" groups several taxa indistinguishable by their morphology and the mention "cf." points the most probable taxa.

Pollen, palynomorph and macrofossil diagrams were constructed with the GPalWin computer program (Goeury, 1997).

The organic content of sediments was evaluated from loss-on-ignition measurements (at 600°C for 0.5 h ; Dean, 1974). The humification degree was estimated discontinuously on ca. 1-cm thick samples at every 10 cm from macroremains preservation (especially mosses) and decomposed material content, and was expressed in Von Post scale from 0 to 10. The obtained value was ascribed to the corresponding 10 cm interval (see Figs. 38 to 41).

Chronological control

Sixteen radiocarbon dates (Table 6), together with palynostratigraphic correlations, provide the chronological control of the Mirabel bog development. Four AMS dates are rejected: the dates TO-8168, TO-8876 and TO-8881 present abnormal recovery yields, which could have resulted in dating errors (R. Beukens, personal communication, 2000) and the date TO-8169 was rejected on the basis of palynostratigraphic correlations with Mer Bleue bog, located in the St. Lawrence lowlands near Ottawa (Richard, 2001; P.J.H. Richard, unpublished data, 2000).

The age-depth model of Mir1 was constructed using a smooth interpolation (splines; Guiot & Goeury, 1996) between radiocarbon dates including the standard deviations (Fig. 33). This procedure minimises abrupt changes in net peat accumulation rate, thus avoiding the bias induced by an insufficient or irregular radiocarbon dating. It is particularly justified in studies which, like ours, focus on the timing of the variations. Other age-depth models (Fig. 33) were similarly developed, but were based on chronological assessment from palynostratigraphic correlations with Mir1. The correlated events correspond to boundaries of the regional palynozones described in Muller & Richard (accepted). Conventional radiocarbon dates are calibrated with the CALIB 4.0 program (Stuiver & Reimer, 1993). Calibrated dates used in the text are noted cal. BP (calendar years before present, i.e. 1950).

TABLE 6. Conventional and calibrated radiocarbon ages. Rejected ages are noted in italic. AMS dates were performed on terrestrial plant macrofossils.

Core name	Depth (cm)	Laboratory code	Datation method	Calibrated age	Confidence interval 2σ	Conventional age	$^{13}C/^{12}C$ ratio
Mir1	82-88	Beta-135343	Conventional	2180	2340-2060	2200 ± 50	-28.4
	81-90	*TO-8169*	*AMS*	*3630*	*3960-3360*	*3380 ± 120*	*n.a.*
	232-238	Beta-135344	Conventional	6820	7180-6680	6050 ± 80	-26.8
	235-237	TO-8167	AMS	6830	7250-6450	6010 ± 160	n.a.
	318-325	Beta-135345	Conventional	7630	7790-7510	6800 ± 80	-19.8
	321-323	*TO-8168*	*AMS*	*6310*	*6850-5910*	*5550 ± 220*	*n.a.*
	408-410	TO-8166	AMS	9010	9470-8430	8050 ± 170	n.a.
Mir2	321-326	Beta-138673	Conventional	10030	10220-9700	8900 ± 80	-27.6
Mir22	*188-190*	*TO-8876*	*AMS*	*7670*	*7840-7490*	*6820 ± 100*	*n.a.*
Mir31	351-355	Beta-145847	Conventional	8580	8980-8370	7770 ± 110	-26.9
Mir33	210-212	TO-8877	AMS	9540	9710-9490	8600 ± 70	n.a.
Mir41	318-320	TO-8878	AMS	9910	10190-9630	8850 ± 70	n.a.
Mir43	254-257	TO-8879	AMS	9540	9710-9490	8600 ± 70	n.a.
Mir51	291-295	TO-8880	AMS	10030	10210-9700	8890 ± 70	n.a.
Mir53	*186-190*	*TO-8881*	*AMS*	*6820*	*7680-5920*	*6020 ± 430*	*n.a.*
Mir61	296-300	Beta-145846	Conventional	7450	7610-7320	6580 ± 80	-26.3
Mir6	214-217	TO-8882	AMS	9430	9530-9130	8370 ± 80	n.a.
Mir63	115-120	TO-8883	AMS	8410	8590-8350	7650 ± 60	n.a.
Mir71	286-290	Beta-145845	Conventional	8760	9230-8430	7930 ± 120	-25.9
Mir73	196-199	TO-8884	AMS	9570	9910-9530	8670 ± 70	n.a.

n.a.: not available

Results

Plant communities

The vegetation history of Mirabel bog is inferred from stratigraphy and palaeobotanical data (Figs. 34 to 37). The recognition of each environment (seral stages) and/or plant community is based on the following evidences:

- Pond. Marl and gyttja deposits. Aquatic taxa (*Brasenia schreberi*, *Chara*, *Ephydatia fluviatilis*, *Gloeotrichia*, *Isoetes*, *Nuphar*, *Nymphaea odorata* subsp. *odorata*, Mollusca, *Myriophyllum*, *Najas flexilis*, Ostracoda, *Potamogeton*).
- Marsh. Herbaceous peat deposit. Emergent taxa (*Carex*, Cyperaceae, *Eleocharis*, *Iris versicolor*, *Schoenoplectus tabernaemontani*, *Triadenum fraseri*, *Typha*).

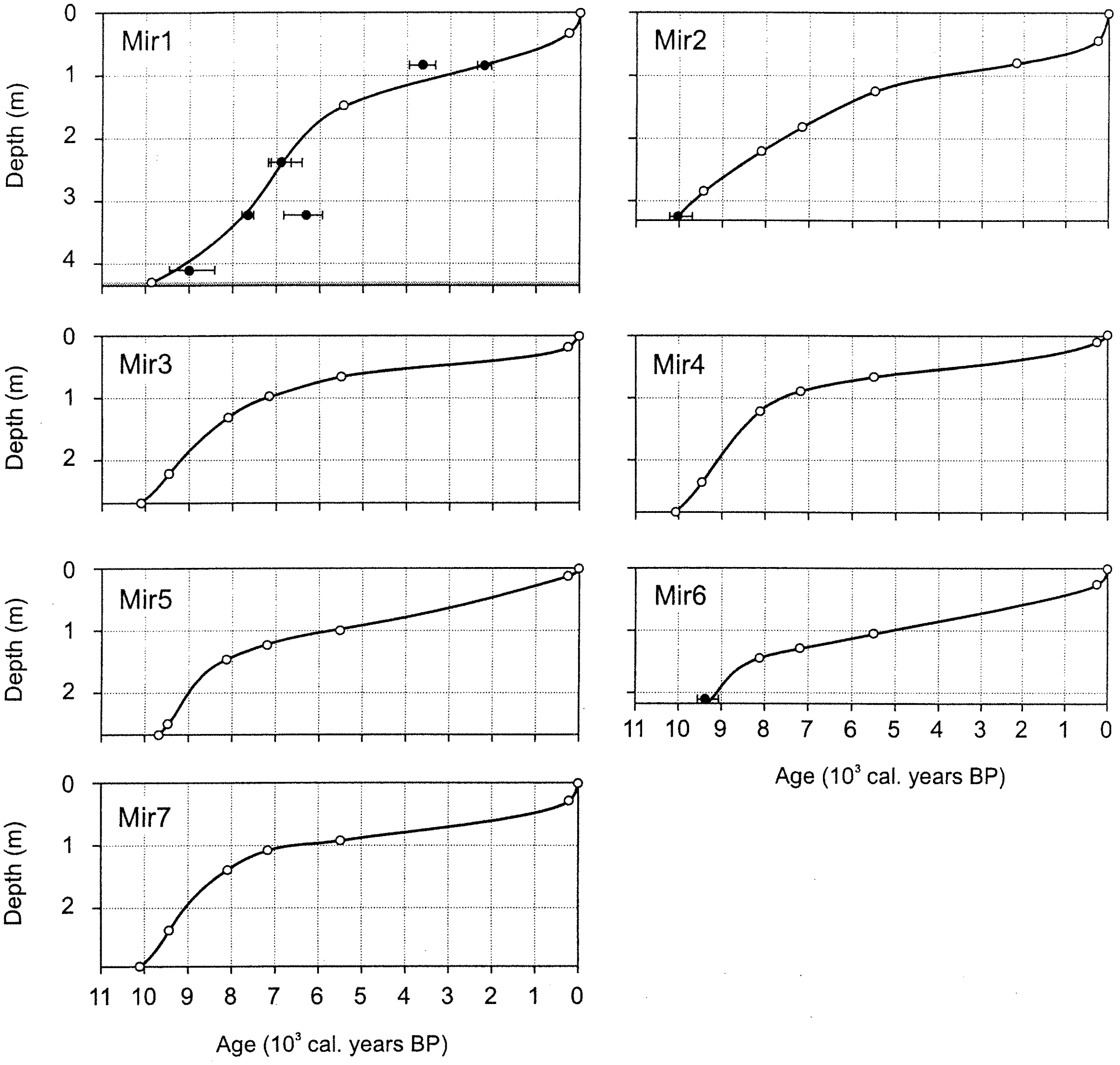


FIGURE 33. Age-depth models. Black dots represent radiocarbon ages and white dots palynostratigraphic correlations, including the two well-dated regional events (*Tsuga* decline at 5500 cal. BP and European settlement at 250 cal. BP).

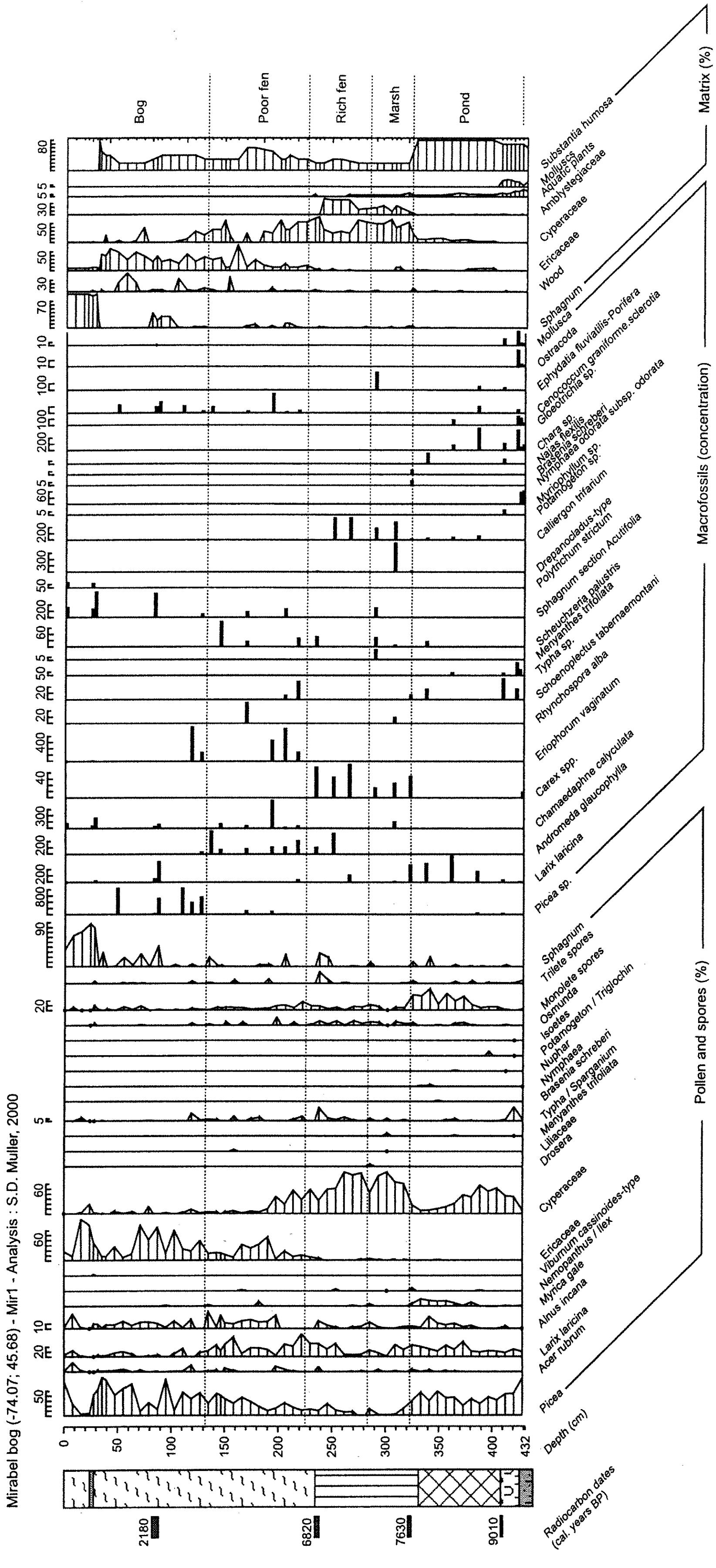


FIGURE 34. Pollen, spores (in percentages) and macrofossils (in concentration) of local taxa in profile Mir1. Dots represent less than 1 %. Macrofossils concentrations are standardised to 100 cm^3 of sediment Scales are not constant. Stratigraphic legend in Fig. 32.

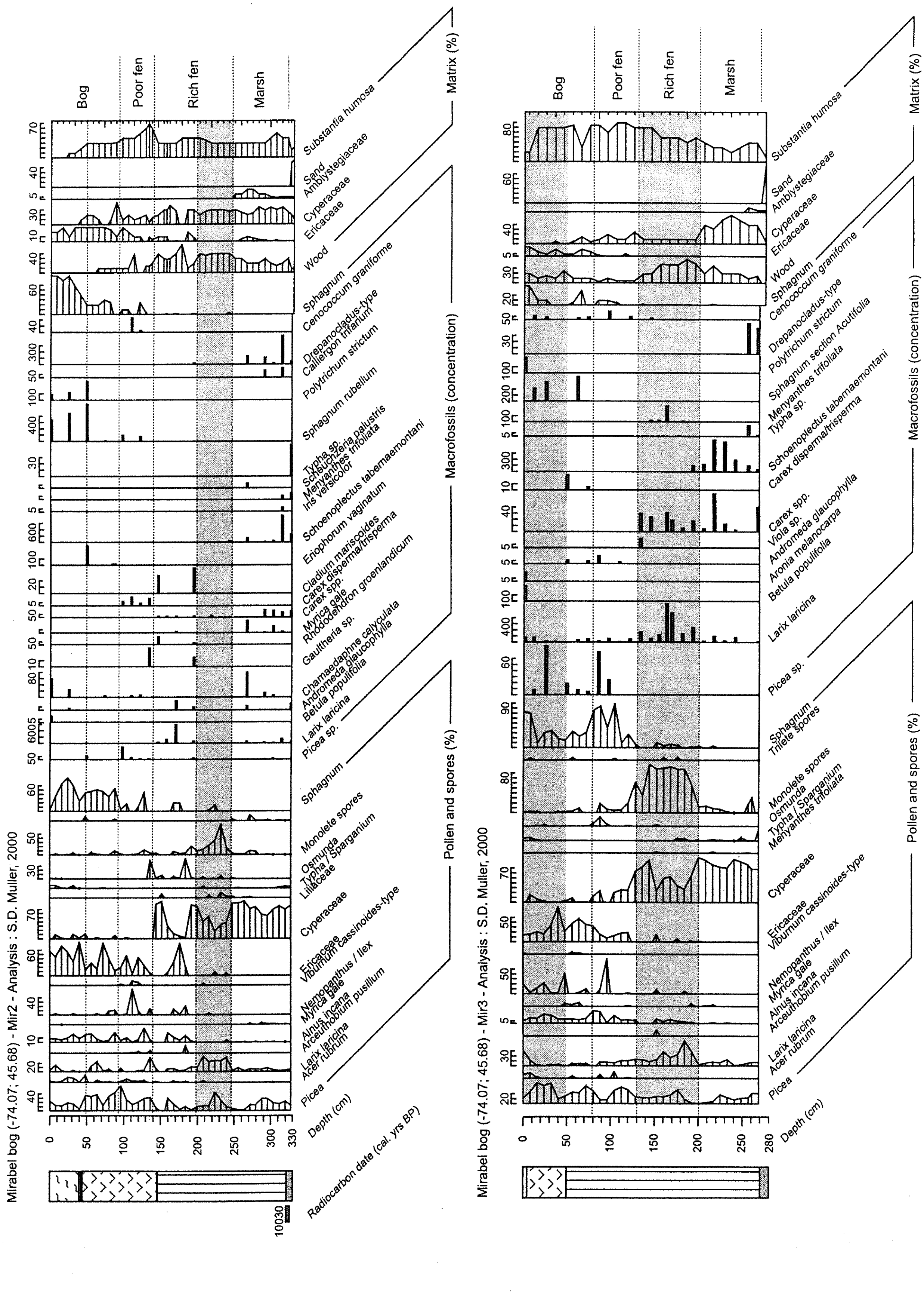


FIGURE 35. Pollen, spores and macrofossils of local taxa in profiles Mir2 and Mir3. Grey zones represent treed communities. See legend in Fig. 34.

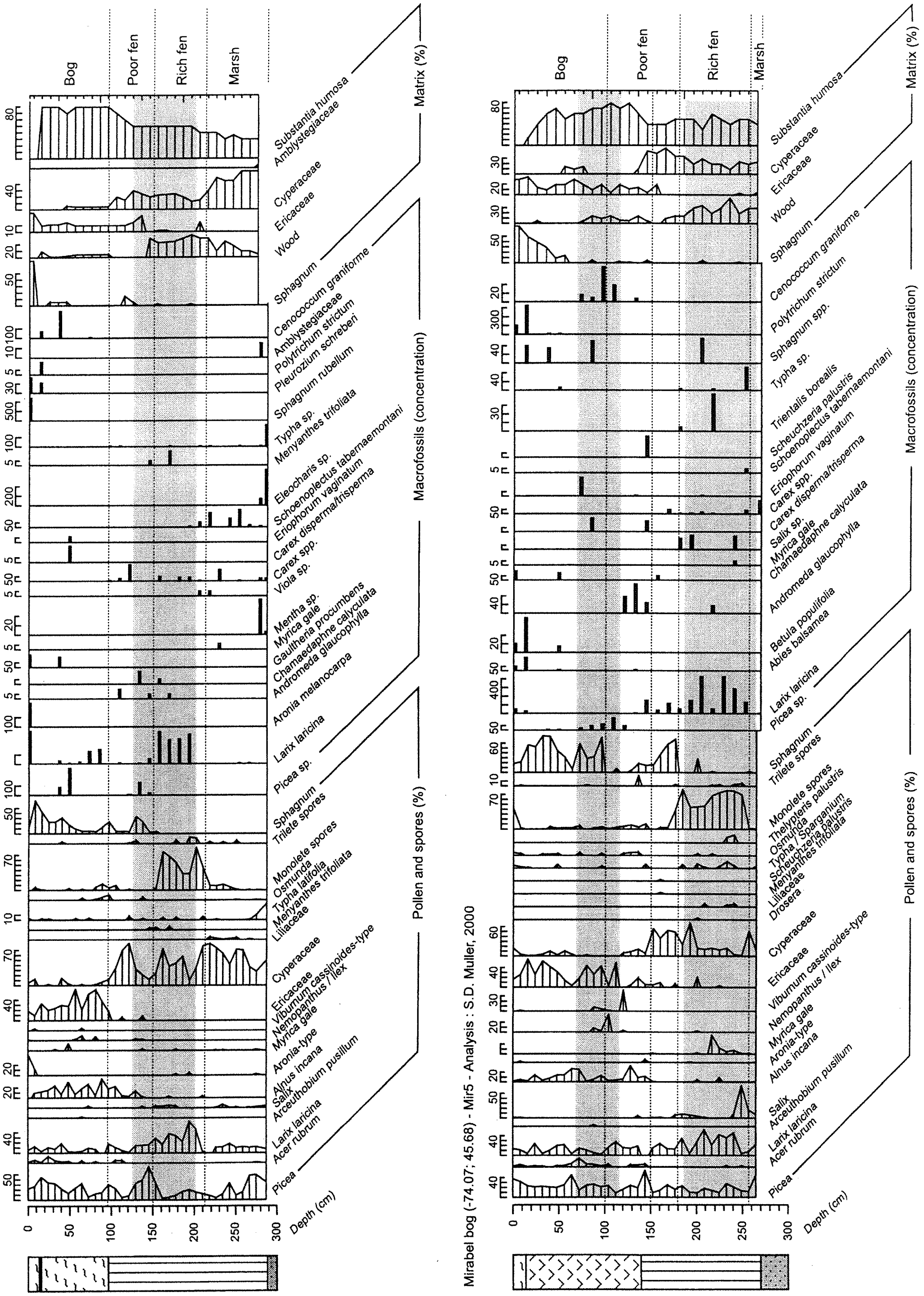


FIGURE 36. Pollen, spores and macrofossils of local taxa in profiles Mir4 and Mir5. Grey zones represent treed communities. See legend in Fig. 34

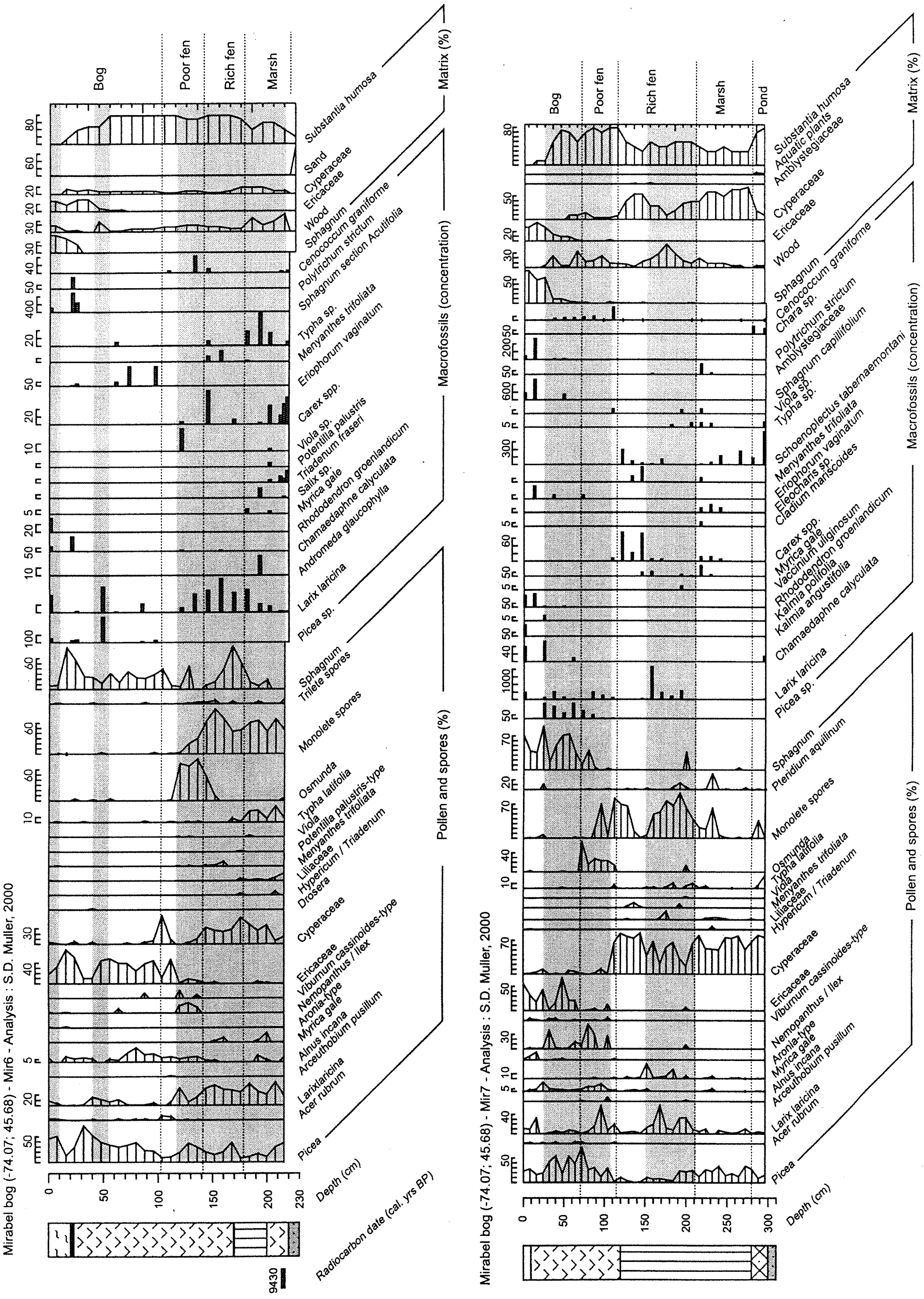


FIGURE 37. Pollen, spores and macrofossils of local taxa in profiles Mir6 and Mir7. Grey zones represent treed communities. See legend in Fig. 34.

- Rich fen. Herbaceous or woody peat deposit. Minerotrophic hydrophilous taxa (*Carex*, *Cladium mariscoides*, Cyperaceae, *Larix laricina*, *Menyanthes trifoliata*, *Myrica gale*, Pteridophyta, *Salix*).
- Poor fen. Herbaceous, woody or *Sphagnum* peat deposit, characterised by simultaneous occurrences of *Cyperaceae*, *Ericaceae* and *Sphagnum* remains. Ombrotrophic taxa (*Chamaedaphne calyculata*, *Sphagnum*), minerotrophic ones (*Osmunda*, *Nemopanthus/Ilex*, *Schoenoplectus tabernaemontani*, *Viola*) and transgressive ones (*Andromeda glaucophylla*, *Carex disperma/trisperma*, *Rhynchospora alba*, *Scheuchzeria palustris*). Simultaneous occurrences of ombrotrophic and minerotrophic taxa, and domination by transgressive species are considered to be the main characteristic of poor fen communities, which represent a transitional state between rich fens and bogs (Glaser *et al.*, 1981; Gorham & Janssens, 1992).
- Bog. *Sphagnum* or woody peat deposit. Ombrotrophic taxa (*Chamaedaphne calyculata*, Ericaceae, *Eriophorum vaginatum*, *Picea*, *Polytrichum strictum*, *Sphagnum*).

Surface moisture conditions

Past surface moisture conditions of Mirabel bog are inferred from microfossil data, exclusive of pollen and spores (Figs. 38 to 41). Interpretations are mainly based on testate amoebae, which were shown to be reliable indicators of water table depth (Tolonen *et al.*, 1992, 1994; Charman & Warner, 1997; Mitchell *et al.*, 2000). Other microfossils and humification data are used as complementary evidences (Moore, 1986).

- Pond (permanent open waters). Gyttja deposit, continuous records of Algae taxa and regular occurrences of Neorhadocoela oocytes.
- Pools (temporary open waters). Discontinuous records of Algae taxa and Neorhadocoela oocytes; highly humified herbaceous peat.
- Wet conditions. Peaks of *Habrotrocha angusticollis*, *Amphitrema flavum*, Copepoda spermatophores and Neorhadocoela oocytes, generally associated with low degrees of humification and sporadic occurrences of Algae taxa. Wet conditions are moreover often marked by peaks of cf. *Entophlyctis lobata*.

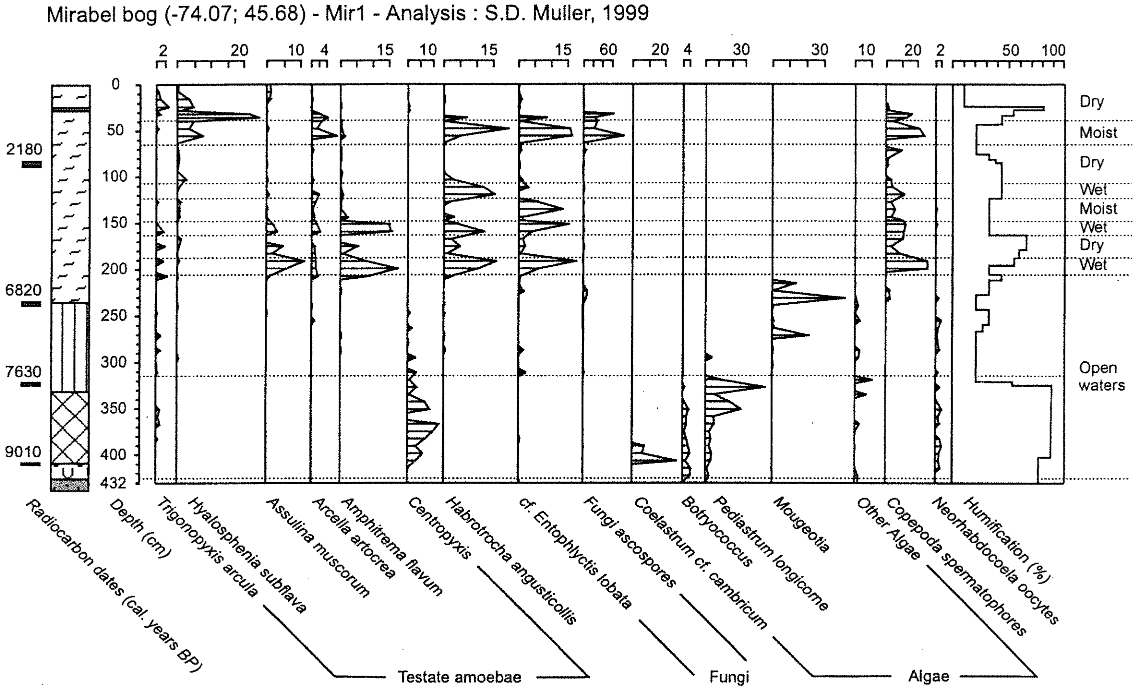
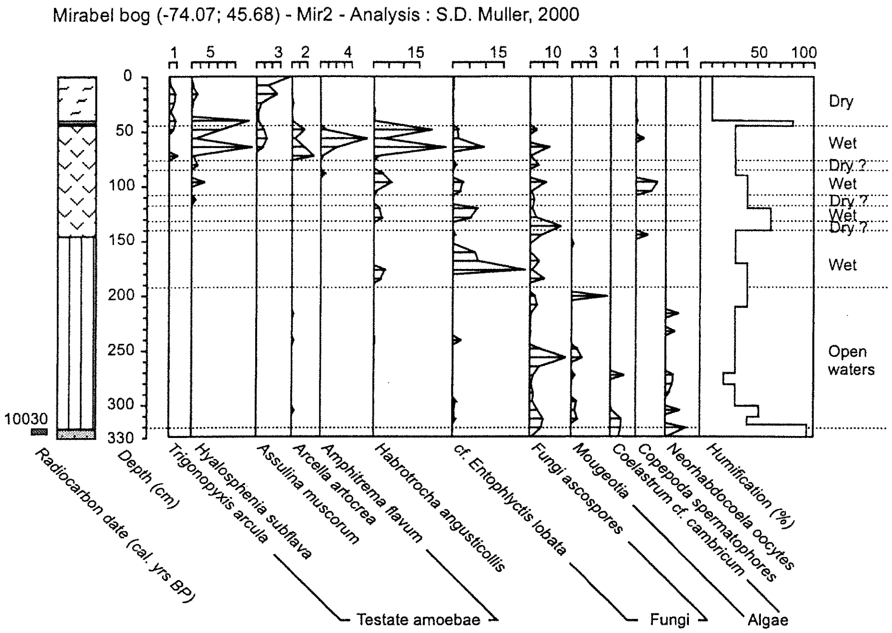


FIGURE 38. Selected microfossil taxa and humification degree of profile Mir1. Concentrations are standardised to 1 mm^3 of sediment. Scales are not constant. Stratigraphic legend in Fig. 32.



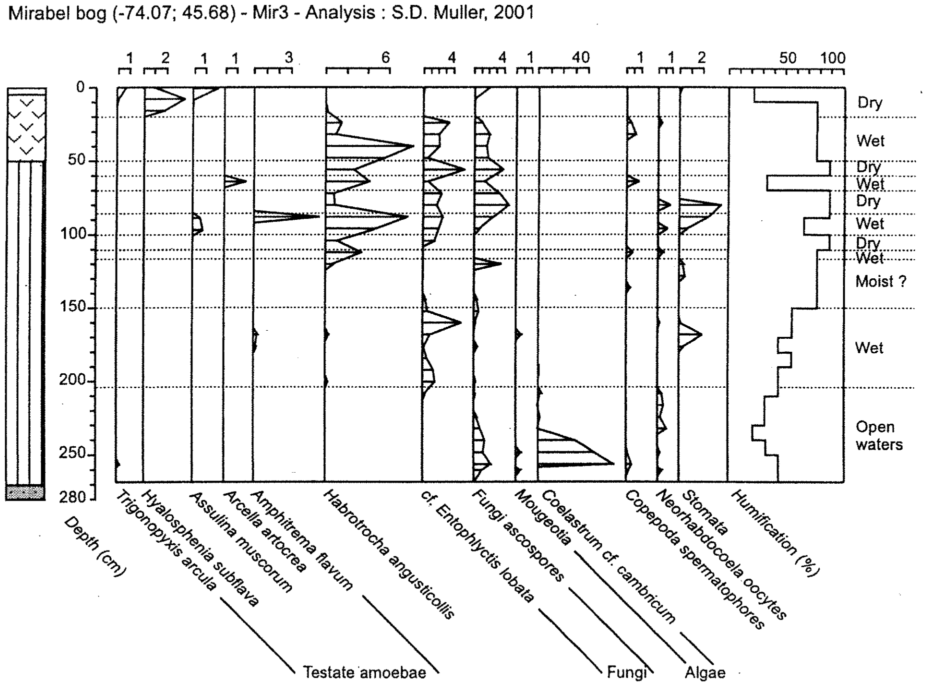
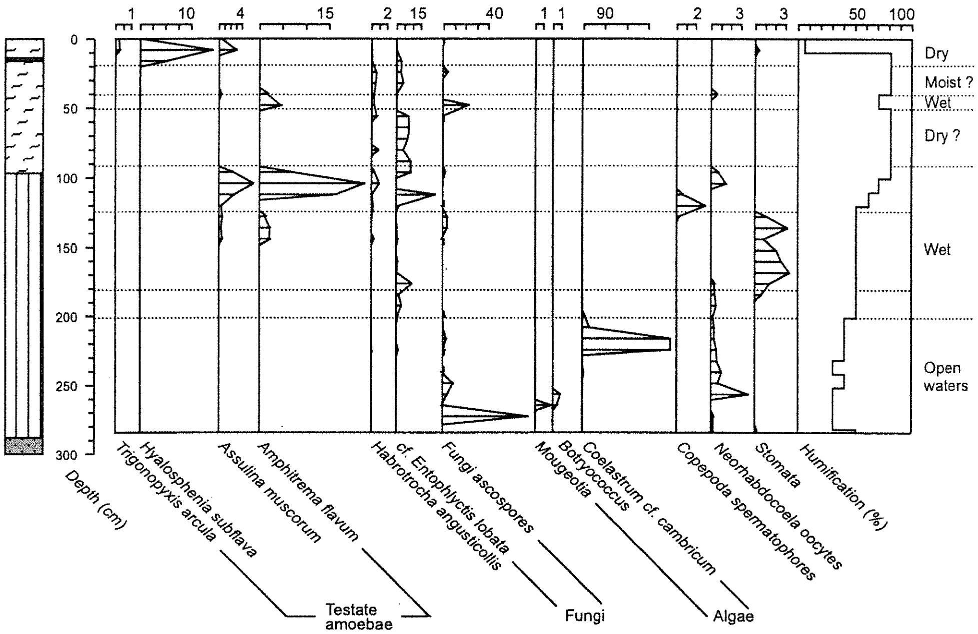


FIGURE 39. Selected microfossil taxa and humification degree of profiles Mir2 and Mir3. See legend in Fig. 38.

Mirabel bog (-74.07; 45.68) - Mir4 - Analysis : S.D. Muller, 2001



Mirabel bog (-74.07; 45.68) - Mir5 - Analysis : S.D. Muller, 2001

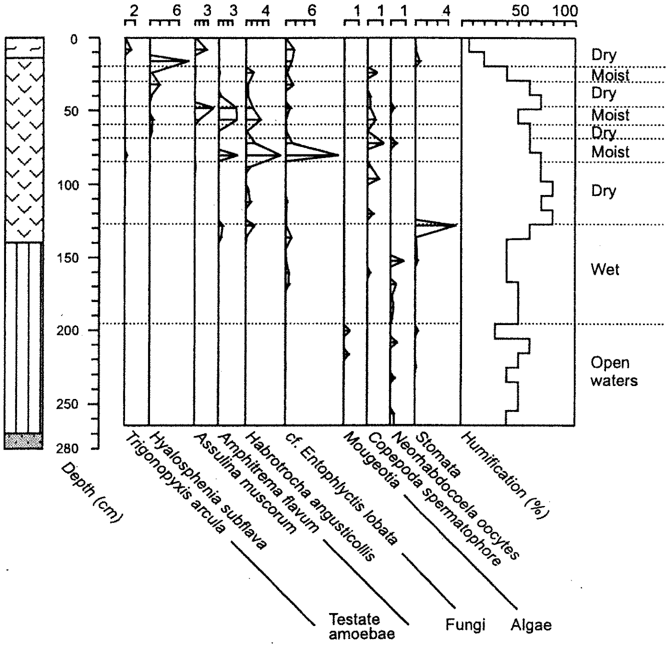
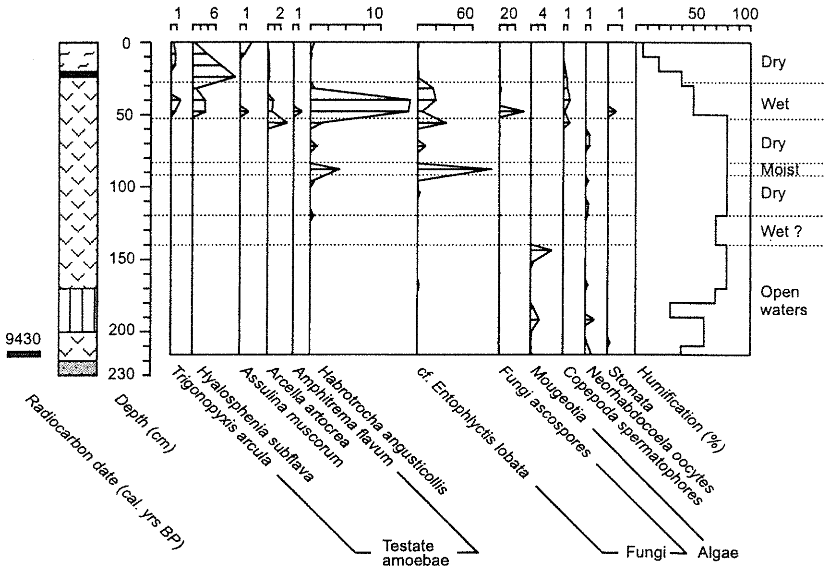


FIGURE 40. Selected microfossil taxa and humification degree of profiles Mir4 and Mir5. See legend in Fig. 38.

Mirabel bog (-74.07; 45.68) - Mir6 - Analysis : S.D. Muller, 2001



Mirabel bog (-74.07; 45.68) - Mir7 - Analysis : S.D. Muller, 2001

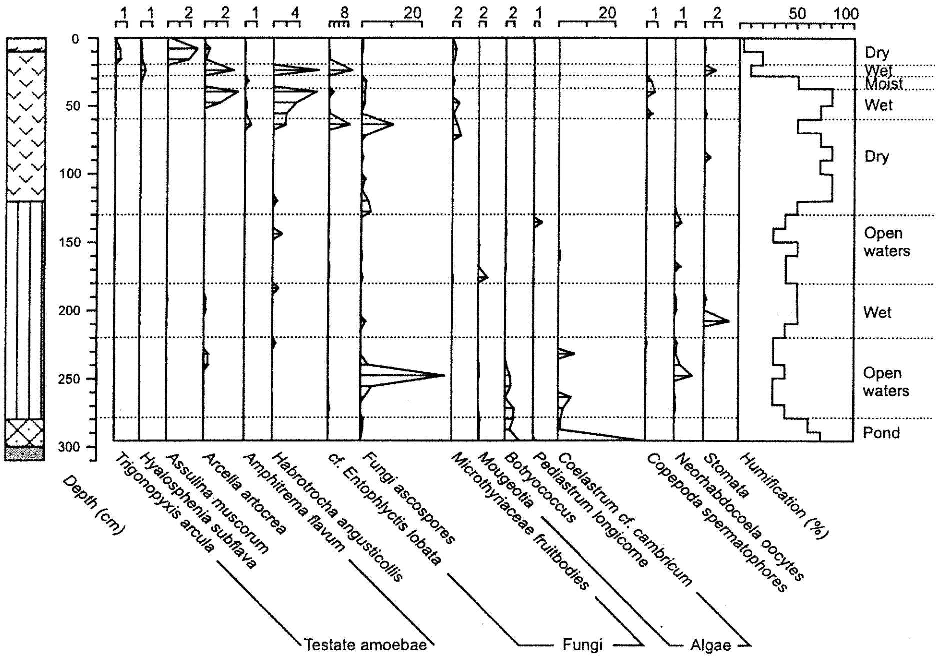


FIGURE 41. Selected microfossil taxa and humification degree of profiles Mir6 and Mir7. See legend in Fig. 38.

- Moist conditions. Simultaneous occurrences of taxa indicative of wet conditions (*Habrotrocha angusticollis*, *Amphitrema flavum*), and of taxa indicative of dry conditions (*Hyalosphenia subflava*). Variable degrees of humification. These conditions often represent a transition between wet and dry layers.
- Dry conditions. Peaks of *Hyalosphenia subflava* and *Trigonopyxis arcula*, generally associated with high degrees of humification and occurrences of Fungi ascospores. Dry conditions are also inferred from very low testate amoebae concentrations (Charman *et al.*, 2001).

Organic accumulation

The organic accumulation was evaluated for individual profiles from loss-on-ignition measurements and radiocarbon or palynostratigraphical dating (Fig. 42). All profiles present similar accumulation trends, except Mir1 characterised by a marked, delayed maximum around 6700 cal. BP. The maximal accumulation rates were recorded in Mir1 (198.5 $g.m^{-2}.yr^{-1}$) and Mir5 (151.0 $g.m^{-2}.yr^{-1}$). The recent (last 250 years) high values could be related to human activity (fertilisation), to climate change or more probably, to the incomplete decomposition in the acrotelm layer. The latter possibility is supported by the progressive increases observed over the last millennium, which suggests a non-anthropic cause.

Ecosystem-scale internal dynamics

Internal dynamics were estimated at the peatland scale by integrating results from individual profiles (Fig. 43). First, the correlation of the seven local vegetation reconstructions allows to evaluate the timing of hydroseral changes throughout the peatland. Notably, all the seral changes appear to have been asynchronous, especially the shift from fen to bog which occurred over more than 4000 years. Second, surface moisture reconstruction evidences widespread wet conditions during the early Holocene, followed by dryer phases around 6000, 4400, 2200 and 200 cal. BP. And third, in the aim to estimate the organic accumulation at the scale of the peatland, 250-years-steps averages calculated for each profile were smoothly interpolated by accounting for basal and surface topographies (Fig. 31). Highest integrated peat accumulation rates were recorded during

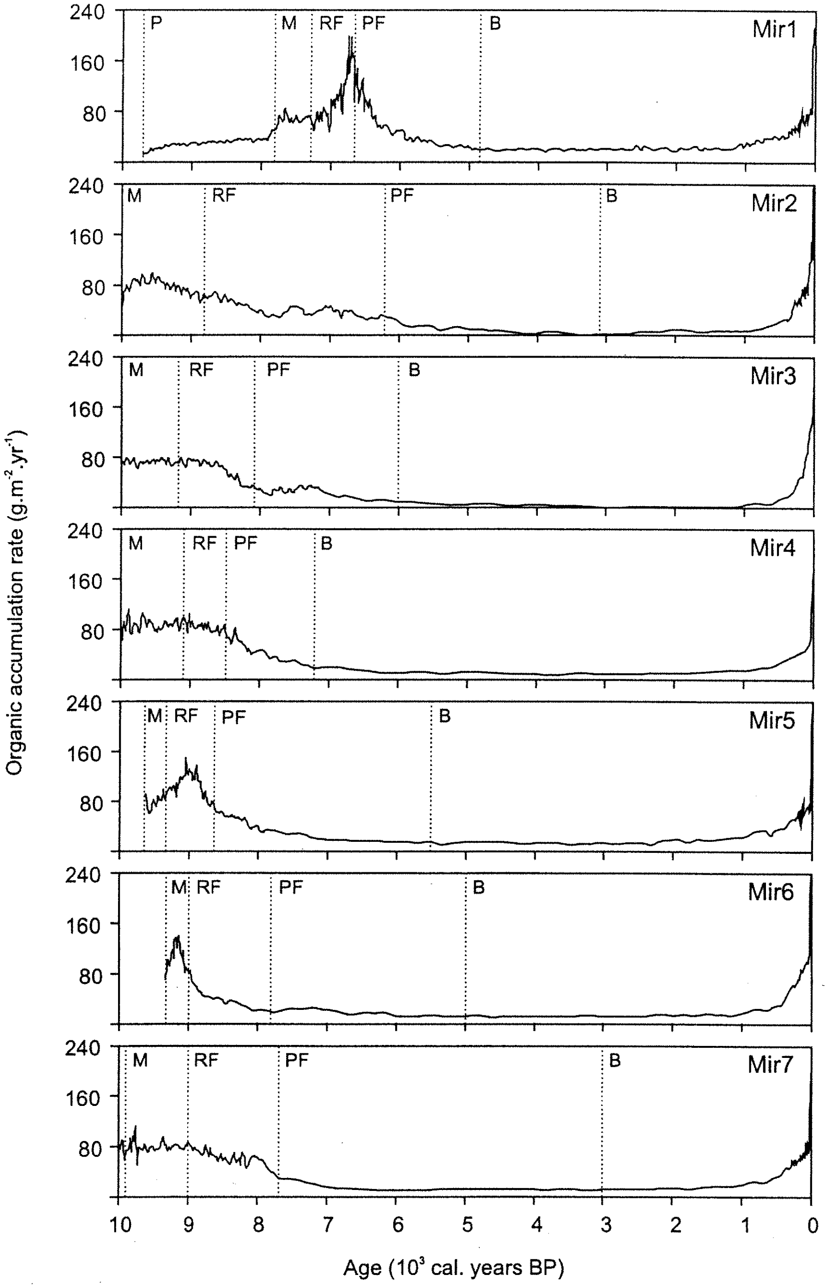


FIGURE 42. Organic accumulation rates at Mirabel bog. Letters refer to local vegetation phases. B: Bog; M: Marsh; P: Pond; PF: Poor fen; RF: Rich fen.

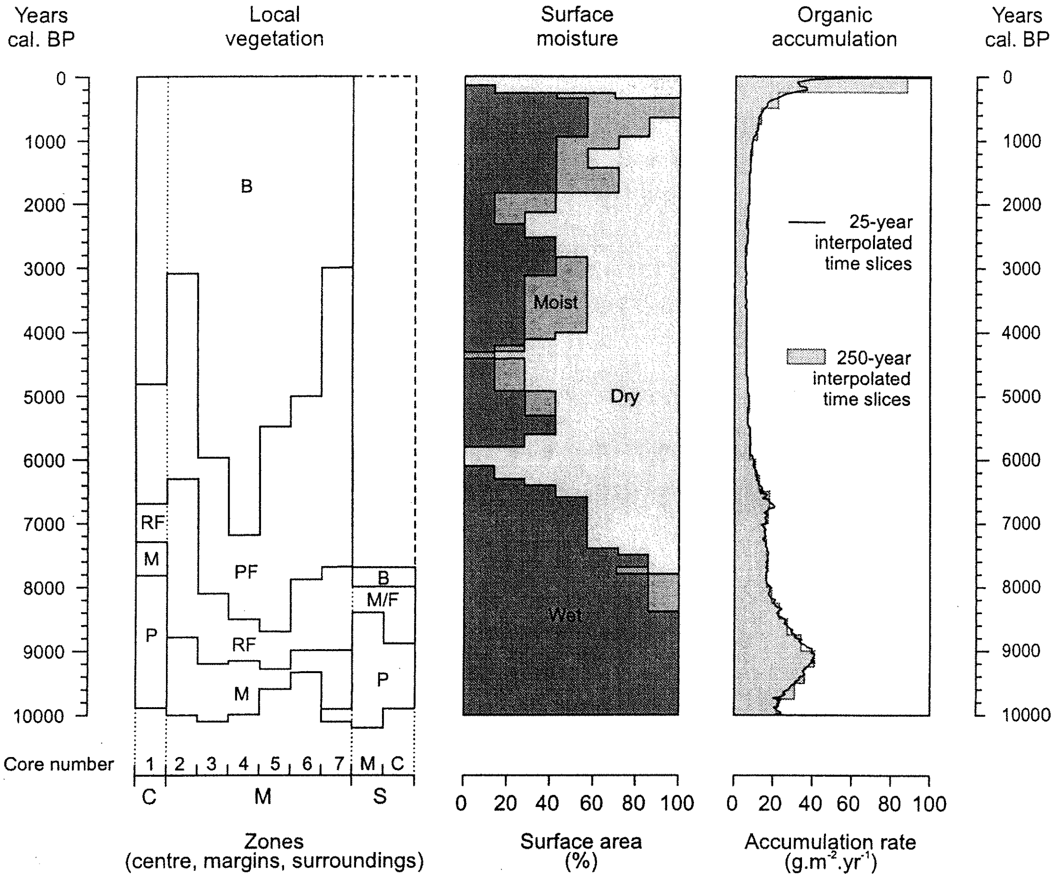


FIGURE 43. Ecosystem-scale dynamics of local vegetation, surface moisture and organic accumulation. Local vegetation phases are: P: Pond; M: Marsh; M/F: Marsh/fen; RF: Rich fen; PF: Poor fen; B: Bog.

early Holocene (40.5 $g.m^{-2}.yr^{-1}$) and the last two centuries (87.5 $g.m^{-2}.yr^{-1}$), while mid- and Late-Holocene were characterised by very low rates (5.8 - 12.4 $g.m^{-2}.yr^{-1}$). The first peak, recorded between 9250 and 9000 cal. BP, represents the dynamics of marginal zones that evolved from marshes to rich fens (Mir2-7; Fig. 42), while the second one, recorded between 7000 and 6500 cal. BP, translates essentially the behaviour of the non-forested central zone during the local transition from rich fen to poor fen (Mir1; Fig. 42).

Postglacial development of Mirabel bog

Pond stage

Dated by palynostratigraphic correlation before 10 300 cal. BP, the basal clay of Mir presents pollen occurrences of *Ruppia maritima* (P.J.H. Richard, unpublished data, 1974), a characteristic species of brackish waters, which indicates here a glacio-marine environment. The initial pond (Fig. 44) was rapidly colonised by a submerged vegetation composed of *Potamogeton*, *Myriophyllum*, *Najas flexilis*, *Chara*, *Gloeotrichia* (Figs. 34 and 37) and planktonic algae (Figs. 38 and 41). The great abundance of *Chara* oospores (Fig. 34) points to the important role played by these algae in marl formation. This calcareous environment allowed the development of a rich aquatic invertebrate community, among which numerous molluscs. Floating-leaved species were sporadically represented by *Nymphaea odorata* subsp. *odorata*, *Brasenia schreberi* and *Nuphar* macrofossils, suggesting the low abundance of these species in Mirabel ponds at the time.

The transition between marl and gyttja occurred around 9300 cal. BP at Mir1 and around 8700 at Mir (Fig. 32). It is characterised in profile Mir1 (Fig. 38) by some millimetric dark layers constituted almost exclusively by a globulate palynomorph, also observed in most of other profiles (Figs. 39 to 41). This microfossil (Fig. 45), which could correspond to the unidentified types 228 or 233 described by Van Geel *et al.* (1989), is thought to be the colonial planktonic algae *Coelastrum* type *cambricum* (Scenedesmaceae or Coelastraceae). *Coelastrum* has been widespread during the early developmental stages of Mirabel bog, and seems to be related to the pond eutrophication indicated by important vegetation changes. Notably, from 8900 to 8400 cal. BP, surroundings evolved toward marsh-fen communities

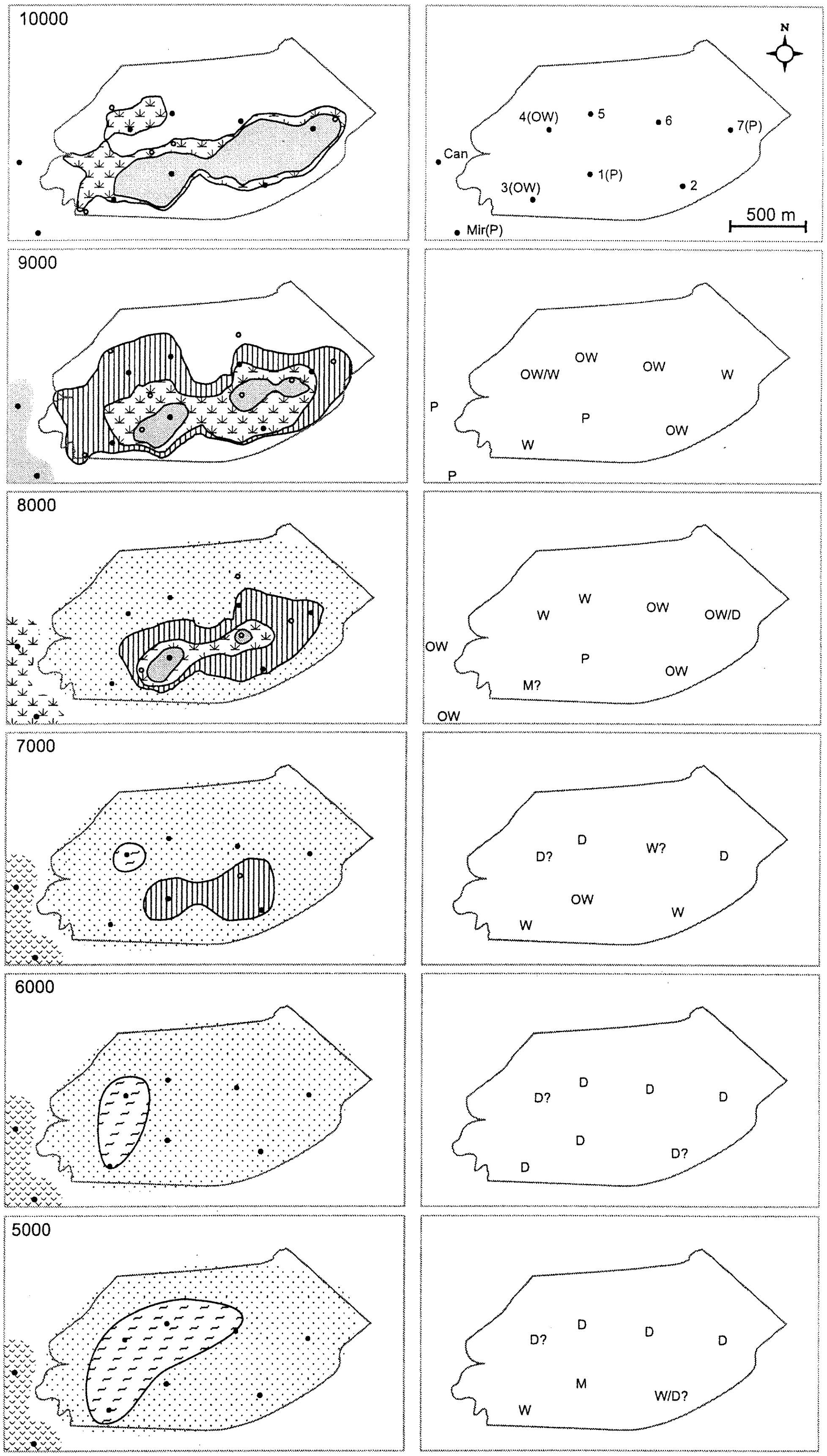


FIGURE 44. Spatio-temporal reconstruction of vegetation mosaic (left column) and surface moisture (right column) at Mirabel bog. Black dots represent the studied profiles. The shape of initial pond and marshes is inferred from the basin topography (Fig. 32) and basal radiocarbon dates (white dots).

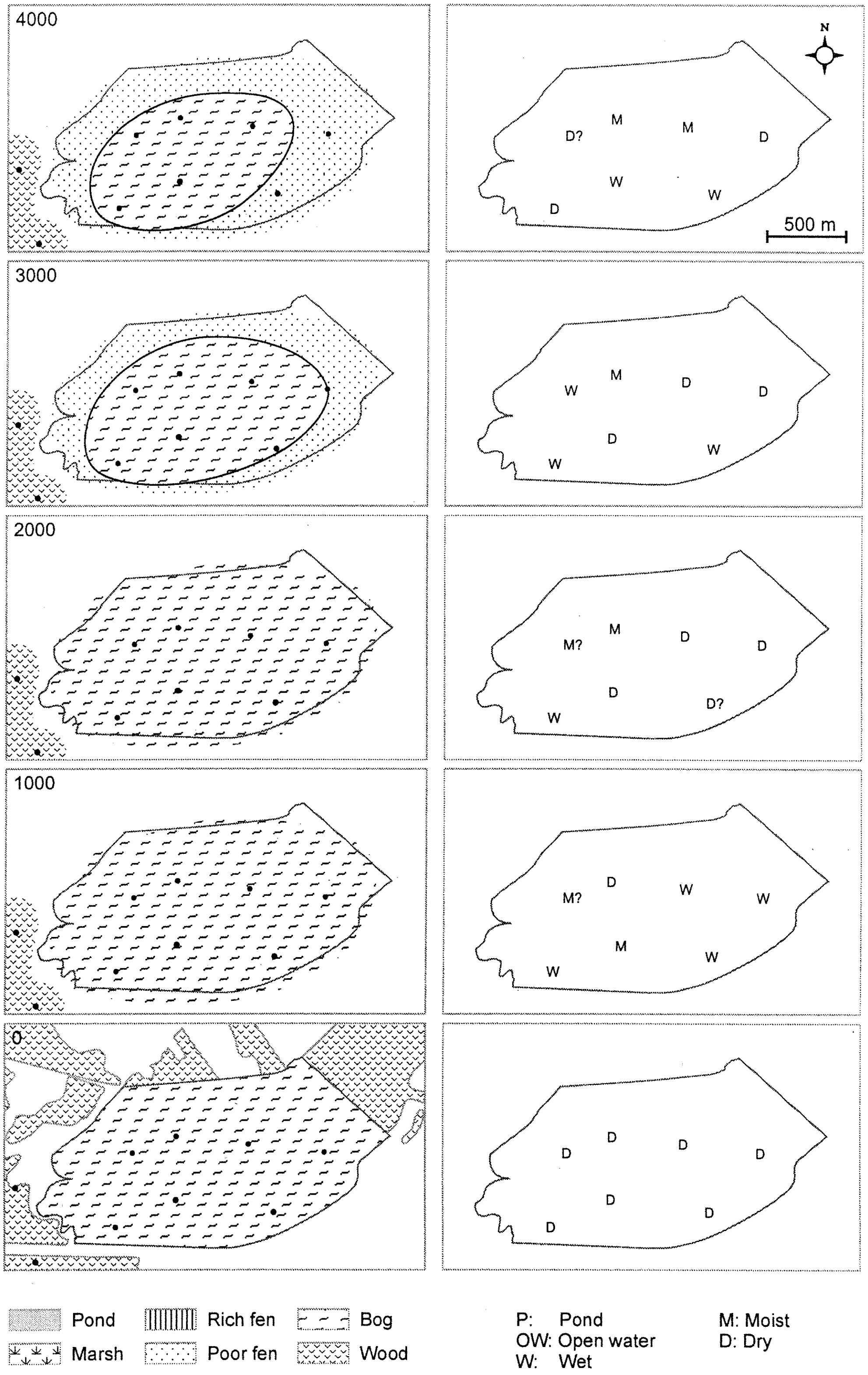


FIGURE 44. Continued.

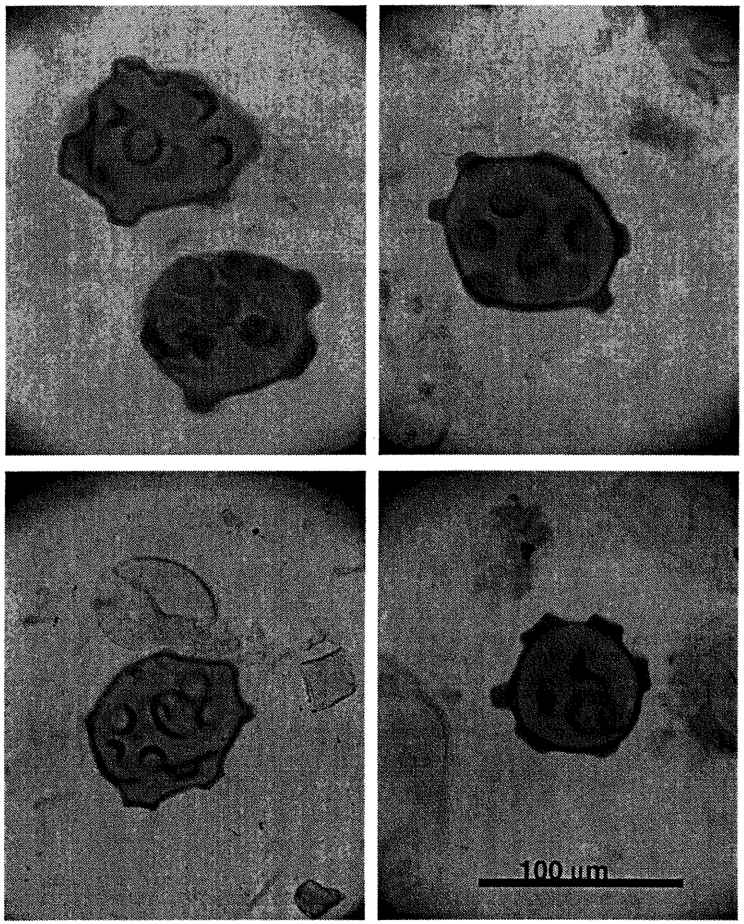


FIGURE 45. *Coelastrum* type *cambricum* (Chlorophyceae: Chlorococcales: Scenedesmaceae or Coelastraceae). Globose microfossil with large protuberances often terminated by a pore.

characterised by Cyperaceae and *Menyanthes trifoliata* (P.J.H. Richard, unpublished data, 1974) and from 9300 to 8800 cal. BP, marginal zones experienced the replacement of sedge meadows by rich, treed fens (mostly *Larix*; Figs. 35 to 37).

Marsh stage

The edges of ponds were covered by marshes dominated by the emergents *Schoenoplectus tabernaemontani*, *Typha*, *Eleocharis* and several species of *Carex* (Figs. 35 to 37). These communities accumulated thick herbaceous peat layers marked by numerous occurrences of the planktonic algae *Mougeotia* and *Botryococcus* (Figs. 38 to 41), which attest the existence of open water areas within the sedge meadows. Waterlogged marshes were also characterised by several species of Amblystegiaceae, including *Calliergon trifarium* in profiles Mir1 and Mir2 (Figs. 34 and 35). This characteristic species of very wet, highly calcareous fen habitats (Crum & Anderson, 1981) indicates the persisting influence of the calcareous clay on the water chemistry over centuries. Due to the centripetal spread of riparian marshes, the size of the initial pond decreased progressively until the total filling-up recorded around 7800 cal. BP at Mir1 (Figs. 34 and 42).

Rich fen stage

Around 9300 cal. BP, marginal marshes began to be replaced by swamps, inhabiting *Larix laricina*, *Myrica gale*, *Carex*, *Menyanthes trifoliata* and fern populations (Figs. 35 to 37). The fern spore morphology, excluding *Thelypteris palustris* and *Onoclea sensibilis*, points to three taxa whose ecological preferences correspond to the reconstructed community: *Woodwardia virginica*, *Dryopteris spinulosa* s.l. (which notably includes *D. intermedia*, present today in marginal maple woods) and *Dryopteris cristata*. The abundance of monolete spores in all profiles leads us to consider these minerotrophic formations as "fern fens" or "fern swamps".

While paludification proceeded outwards during the ninth millennium cal. BP, the centripetal overgrowing of marshes by rich fen communities occurred until 7300 cal. BP, along with the progressive drying of the peatland surface (Fig. 44). Between 8700 and 7800 cal. BP, marginal zones shifted from rich fens to poor fens, and surrounding zones (profiles Mir and Can) from marsh-fen formations to non-peat-forming communities (forested swamps, carrs; Fig. 44). The first occurrences of *Acer rubrum* pollen (during the regional palynozone 3b, recognised as *Acer-Quercus* deciduous forest; Muller & Richard, accepted) are then recorded in all marginal cores, which indicates the onset of peripheral red maple woods, still present today around the open, ericaceous bog (Fig. 30).

Poor fen stage

The disappearance of minerotrophic groundwater supply is marked in profile Mir1 by the disappearance of *Calliergon trifarium* around 6700 cal. BP (Fig. 29). That major event was also revealed by profound modifications of plant communities, such as the replacement of minerotrophic plants like *Carex*, *Menyanthes trifoliata*, *Myrica gale*, Amblystegiaceae and Pteridophyta by ombrotrophic taxa like Ericaceae, *Eriophorum vaginatum* subsp. *spissum* and *Sphagnum*. The corresponding disappearance of open water areas is characterised, especially in profile Mir1 (Fig. 38), by a faunal change from Neorhabdocoela to Copepoda communities. Neorhabdocoela, which are known to live in freshwater aquatic and semiaquatic ecosystems including bogs (Haas, 1996b), seem to characterise more humid environments than Copepoda.

At all sites, following the trophic change from rich to poor fen, communities still harboured some minerotrophic elements (*Osmunda*, *Schoenoplectus tabernaemontani*, *Viola*) and transgressive taxa (*Andromeda glaucophylla*, *Carex disperma/trisperma*, *Rhynchospora alba*, *Scheuchzeria palustris*) for some time before becoming strictly ombrotrophic. This heterogeneous floristic composition attests the transitional trophic status of these poor fen formations, but they dominated most of the peatland from 8000 to 3-4000 cal. BP

Bog stage

Ombrotrophic communities began to replace poor fens around 7200 cal. BP at Mir4 (Figs. 36 and 44). They spread first on the northeastern part of the peatland then overgrew the mire expanse around 4800 cal. BP, and the southeastern margin around 3000 cal. BP (Fig. 44). Since then, the seven coring sites present similar plant formations as today, dominated by ombrotrophic taxa, such as Ericaceous species, *Sphagnum* and *Eriophorum vaginatum* subsp. *spissum* (Figs. 34 to 37). Although no macrofossil remains of *Nemopanthus mucronatus* were found, pollen records attest the presence of this species around the peatland as early as 7000 cal. BP (Figs. 34 to 37).

Since 6000 cal. BP, the peatland surface was characterised by fluctuating moisture conditions (Figs. 43 and 44). Four dry intervals were recognised, around 6000, 4400, 2200 and 200 cal. BP. The two last millennia (Fig. 44) were relatively humid, as shown by the widespread development of the Rotifer *Habrotrocha angusticollis* and the testate amoebae *Arcella artocrea* and *Amphitrema flavum* (Figs. 38 to 41). The two last centuries were marked by the spread of *Betula populifolia* (revealed by the decrease in *Betula* pollen diameter), *Nemopanthus mucronatus* and *Aronia melanocarpa* (Figs. 34 to 37), and by the successive development of *Hyalosphenia subflava* and *Trigonopyxis arcula* in uppermost peat layers (Figs. 38 to 41). Both testacean species, reported to be commonly associated with artificially drained peatlands (Tolonen, 1986), attest that these recent changes were induced by the numerous man-made ditches still visible today throughout the peatland (Fig. 30).

Discussion

Past autogenic relationships of Mirabel bog development

Before 7700 cal. BP, Mirabel wetland experienced rapid successions from ponds and marshes to rich then poor fens and forests, and waterlogged conditions and high peat net accumulation rates (Fig. 43). The most productive communities (Fig. 42) differ somewhat between profiles: wet *Carex* meadows with *Andromeda glaucophylla* and *Scheuchzeria palustris* (Mir1, Fig. 34), marshes of *Schoenoplectus tabernaemontani*, *Carex* and Amblystegiaceae (Mir2-4-7, Figs. 35 to 37), *Larix* swamps characterised by underwoody communities of ferns and *Carex* (Mir3-4-5-7, Figs. 35 to 37) and treed *Typha* marshes harbouring *Larix laricina*, *Triadenum fraseri* and *Carex* (Mir6, Fig. 37). The variety of such productive communities points to the external influence of the minerotrophic groundwater supply, likely to have enhanced the plant productivity.

The delay between maximal peat accumulation phases on margins and on the focal expanse indicates from 10 000 to 6500 cal. BP a centripetal displacement of plant communities, hydrological conditions and associated high accumulation rates. The decrease in surface moisture between 8000 and 6000 cal. BP (Fig. 43) consequently reveals the progressive highering of the peat surface. These new local conditions explain the subsequent decrease in peat accumulation rates (Figs. 42 and 43), because dryer conditions enhanced the decomposition process, and the reduction in minerotrophic water supply limited the overall plant productivity.

Vegetation changes occurring since 6000 cal. BP were minor compared to previous ones: they only consisted in shifts from relatively diversified communities dominated by *Andromeda glaucophylla*, *Carex*, *Scheuchzeria palustris*, *Sphagnum*, *Picea*, *Viola* and *Osmunda* to ombrotrophic formations of Ericaceous shrubs, *Sphagnum* and *Picea* (Figs. 34 to 37). During the same period, the surface moisture presented a number of asynchronous local fluctuations notably marked by four dry phases evidenced at the peatland scale (Fig. 43). However, neither the vegetation changes nor the hydrological fluctuations appear to have influenced peat accumulation, which remained relatively constant until *ca.* 1000 cal. BP, both at the community and peatland scales (Figs. 42 and 43).

Finally, the important increase in peat accumulation rate that occurred during the last millennium could be an artefact due to the lack of decomposition within the acrotelm layer. In that case, the high recent values cannot be directly compared to previous ones. The impact of the human-related draining on the peatland productivity, which is likely to have occurred only within the last 200 years, is consequently difficult to evaluate.

Peatland development: hydrarch succession and facilitation model

On the basis of palaeobotanical evidences, we distinguished each of the main stages of the hydrarch succession described by Weber (1908) and Clements (1916). The seven studied profiles provide examples of both initiations by terrestrialisation (from pond, Mir1 and 7) and by paludification (from marshes, Mir2-6). Despite these different origins, all coring sites present similar plant successions with no inversion nor missing steps compared to the hydroseral model. Moreover, early successional changes, which resulted from autogenic peat accumulation through hydrological changes, well illustrate the facilitation model. Effectively, changes from ponds to marshes then rich fens resulted from water depth decrease and changes from rich fens to poor fens resulted from the decrease in groundwater supply. Besides, it is noteworthy that Mir1 and Mir2 sites record rich fens while the other sites around already attained the poor fen stage (Figs. 42 and 44). We suggest that the hydraulic gradient from the surficial deposits (clay) around and under Mirabel peatland was then stronger that the surficial drainage from the surroundings, thus sustaining elevated minerotrophy at Mir1 and Mir2.

Ability of palaeoecological data to tackle past autogenic relationships

Two main periods were recognised within the developmental history of Mirabel bog: from 10 000 to 7000 cal. BP, autogenic dynamics were linked by minerotrophic groundwater influence, while they seem to have been independent since 7000 cal. BP. These results show that our palaeoecological data allow to demonstrate the constraining external influence of groundwater supply on autogenic dynamics, but are insufficient to tackle their more subtle interrelations. This notably implies that past moisture patterns could not be reliably estimated from our palaeobotanical data. This could be partly related to the important decomposition rates of most peat deposits, which did not allow a good conservation of macroremains, especially of mosses.

Observation scale and autogenic dynamics

The integration of the seven individual profiles (Fig. 43) provides a way for comparing autogenic dynamics at the community (major seral stages) and ecosystem scales. First, the plant history appears to be greatly enriched by this spatial approach, notably concerning the timing of successional changes. The successive vegetation changes respectively spanned over 600 (from marshes to rich fens), 2400 (from rich fens to poor fens) and 4100 years (from poor fens to bogs) in marginal zones, revealing the decrease in allogenic groundwater influence before the onset of ombrotrophy. This suggests that the timing of vegetation changes can be used to evaluate the respective influences of autogenic and allogenic factors. Second, the surface of the whole peatland presented several moisture fluctuations (Fig. 43), which could be *a priori* attributed to climate forcing. However, the last dry period which occurred for two centuries is clearly related to human practices; the two prior ones respectively dated around 4400 and 2200 cal. BP were not equally recorded in all peatland zones (Fig. 44); and the first one, although simultaneously recorded in all profiles, occurred during a wet climatic phase evidenced from pollen and lake-level data (Muller *et al.*, submitted-a). Moreover, the asynchrony of moisture fluctuations between all profiles (Fig. 44) points to local controls of internal hydrology. Such a pattern suggests here a different situation than in several United Kingdom peatlands, shown to present a strong within-site synchrony of moisture changes (Hendon *et al.*, 2001).

And third, ecosystem-scale peat accumulation does not present significant differences with marginal profiles (Figs. 42 and 43). By contrast to vegetation and moisture dynamics, these synchronous patterns point to the external influence of groundwater supply as the main control of peat accumulation during marsh and fen stages. This moreover indicates that marginal profiles, in the case of Mirabel, are more representative of the ecosystem-scale peat accumulation dynamics than the central core, probably because of the complex bottom topography.

Conclusions

Two main periods were recognised within the developmental history of Mirabel bog. The first one (10 000-7000 cal. BP) is characterised by intricate, directional dynamics of local vegetation, surface moisture and peat accumulation under control of minerotrophic water supply. The second one (7000-0 cal. BP) presents independent changes in vegetation composition and surface moisture, without any obvious influence on peat accumulation rates. The trophic shift, from 8000 to 6000 cal. BP, resulted essentially from the peat accumulation process, which progressively isolated the peat-forming vegetation from the groundwater supply. The asynchrony of vegetation and moisture changes within the peatland, especially since 6000 cal. BP, reveals the lack of climatic control on its development.

These results were based on a thorough analysis of past plant communities and peat surface moisture. We distinguished from palaeobotanical evidences the main successional stages of the typical hydrarch succession (pond, marsh, rich fen, poor fen, bog). Notably, poor fens were evidenced in all studied profiles as transitional albeit widespread and long-lasting stages between rich fen and bog communities, by their heterogeneous floristic composition, comprising ombrotrophic, minerotrophic and transgressive taxa. Among the latter, the most characteristic species appears to have been *Andromeda glaucophylla*, known to normally represent the first stage of the ericaceous succession (Marie-Victorin, 1995). Poor fens differ from previous minerotrophic formations by the lack of several taxa, notably monolete ferns and *Larix laricina*, observed to have been associated in all marginal profiles, during a "fern fen" or "fern swamp" stage. Despite this palaeobotanical effort, our data do not evidence the autogenic relationships known from modern studies. This raises questions about the ecological significance of fossil assemblages and suggests that plant macroremains are not an accurate proxy-data for moisture conditions.

Finally, our study provides insights on the role of observation scale on palaeoecological reconstructions, by linking community-scale and peatland-scale approaches. The general asynchrony of vegetation and moisture changes between profiles points to the difficulties of assessing the system behaviour from single records, and the importance of accounting for the spatial heterogeneity of peatlands. However, the integration of several profiles may lead, notably in the case of moisture, to mistaken interpretations. It may be tempting to consider fluctuating trends in surface moisture as translating climatic forcings, but the results obtained at Mirabel bog suggest that the peat surface moisture was more a function of mesoscale topography. Obviously, this peatland as a whole did not register any climate-related changes. Moreover, the similar peat accumulation trends presented by all marginal cores suggest that, in concentrically-developed peatlands like Mirabel bog, a transect of cores from shallow margins to the deepest zone could be sufficient for tackling the essence of peat accumulation behaviour of the entire system. In the contrary, the general understanding of past vegetation and moisture patterns appears to require corings from all functional units (defined from modern plant communities and basal topography) constituting the peatland.

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CONDITIONS OF CARBON ACCUMULATION IN A BOREAL BOG

Serge D. Muller, Pierre J.H. Richard & Jacques-Louis de Beaulieu

Quaternary Research - Submitted

Abstract

This study examines the conditions of carbon accumulation in a southern Québec peatland, Mirabel bog, by confronting independent evidences relating to physiographic context, basal topography, climate, fire, local vegetation and surface moisture. Special attention is paid to scales and to their relationships with the different factors. Notably, carbon accumulation only presents long-term trends, which reflect the interplay between minerotrophic groundwater supply and autogenic processes: the initially alkaline conditions enhanced plant productivity and peat accumulation, which, in turn, progressively isolated peat-forming vegetation from the groundwater influence. Other physiographic parameters strongly influenced the peatland development: the geomorphological context determined the location and extent of peat deposits, and the basal topography controlled the rate of lateral expansion, both in centrifugal and centripetal directions. Climate appears to have not influenced in any way the internal dynamics. Notably, precipitation shows no obvious correlation with surface moisture, and carbon accumulation, which approximates the vegetation productivity, was not related to temperature. Finally, fire did not seem to influence the peat accumulation process, fire events occurring mostly during waterlogged early stages, unfavourable to peat burning.

Key words: Peatland development, allogenic controls, autogenic factors, physiographic context, climate, fire regime, vegetation succession, surface moisture, carbon accumulation, southern Québec.

Résumé

Conditions de l'accumulation du carbone dans une tourbière du Québec méridional

Cette étude examine les conditions de l'accumulation du carbone dans une tourbière du Québec méridional (tourbière de Mirabel), en confrontant entre eux le contexte physiographique, la topographie basale, le climat, le régime des incendies, la végétation locale et l'humidité de surface. Une attention particulière est accordée aux échelles concernées par les différents facteurs. Notamment, l'accumulation du carbone présente seulement des tendances à long terme reflétant les interactions entre l'alimentation en eau minérotrophe et les processus autogènes. Les conditions alcalines initiales augmentèrent la productivité végétale et l'accumulation de la tourbe, qui, en retour, isola progressivement la végétation tourbigène de la nappe phréatique. Les autres paramètres physiographiques influencèrent le développement de la tourbière : le contexte géomorphologique détermina la localisation et l'extension du dépôt tourbeux, et la topographie basale contrôla l'expansion latérale, centrifuge et centripète. Le climat n'a pas influencé les dynamiques internes. Notamment, les précipitations ne sont pas corrélées avec l'humidité de surface, et l'accumulation du carbone qui représente une approximation de la productivité tourbigène n'est pas liée aux températures. Finalement, le feu n'a pas non plus influencé l'accumulation de la tourbe, les incendies locaux ayant surtout eu lieu durant les stades initiaux, où les conditions humides n'étaient pas propices à la destruction de la tourbe.

Mots clés: Développement d'une tourbière, contrôles allogènes, facteurs autogènes, contexte physiographique, climat, régime d'incendies, succession végétale, humidité de surface, accumulation du carbone, Québec méridional.

Introduction

Peatlands were shown to represent a key compartment in the biogeochemical cycle of carbon, one major element implied in the greenhouse effect (Gorham, 1991). However, despite the number of related works, they are still counted among the functionally less-understood ecosystems. Difficulties of peatland studies are mainly related to the complex relationships between the different influences of their development (Frenzel, 1983). The respective influences of autogenic and allogenic factors have long been recognised, notably for autogenic succession (Weber, 1908; Clements, 1916; Tallis, 1983) and climate change (Blytt, 1876; Godwin, 1952; Heinselman, 1970; Moore & Bellamy, 1974; Aaby, 1976; Barber, 1981). Their interplay was more recently assessed (Frenzel, 1983; Payette, 1988; Almquist-Jacobson & Foster, 1995) and shown to greatly depend on spatio-temporal patterns. It is effectively thought to change during peatland evolution: allogenic factors would control the early stages, while autogenic processes would determine the late ones (Svensson, 1988; Korhola, 1995a; Vitt & Kuhry, 1992; Nicholson & Vitt, 1994). Moreover, it is likely to depend on the spatial heterogeneity of peatlands: for instance, mire margins would be less sensitive to climate than mire expanses (Korhola, 1992; Korhola *et al.*, 1996). Most studies concerning peatland development focuses on particular factors in the aim to examine their respective influences on peatland dynamics. Notably, this 'individualistic' approach greatly improved the understanding of the role played by climate (e.g., Barber, 1981; Winkler, 1988; Kuhry *et al.*, 1992; Korhola, 1995b; Lavoie & Richard, 2000a, 2000b), basal topography (e.g., Korhola *et al.*, 1996; Lavoie, 1998) or disturbances (e.g., Caseldine & Hatton, 1993; Kuhry, 1994; Pitkänen *et al.*, 1999). However, it is unlikely to allow the understanding of the entire peatland development, shown to result from the complex causal linkage among numerous parameters (Frenzel, 1983; Payette, 1988; Almquist-Jacobson & Foster, 1995).

The scale-related integrated approach

One way to account for these interrelations consists in confronting independent evidences of different factors. Almquist-Jacobson & Foster (1995) used such an 'integrated' approach to interpret the development of Swedish raised-bogs. They notably demonstrated the general control exerted by climate moisture and temperature on fen development and raised-bog distribution, respectively, and the local influence of substrate characteristics on the lateral expansion of fens. We develop here a similar method for assessing the past carbon dynamics of a southern Québec bog.

Our method consists first in collecting independent evidences of the past conditions of peatland development: geomorphological context, basal topography, regional climate, fire incidence, local vegetation, peat surface moisture and carbon accumulation. Second, these parameters are confronted between them with the aim to reconstruct their interplay, to reveal their linkages and antagonisms, and specify the eventual shifts in limiting/constraining factors. This step has to be made with respect to the spatio-temporal scales at which the different processes operate. Globally, climate and physiography may be considered to define the general framework within which local parameters would operate (Almquist-Jacobson & Foster, 1995). This implies a better synchrony between- (Winkler, 1988; Korhola, 1995b; Mauquoy & Barber, 1999a, 1999b) and within-sites (Barber, 1981; Hendon *et al.*, 2001) for climate- and physiography-induced changes than for local modifications, a feature likely to be revealed by spatial studies. However, if all factors can be assigned to specific spatial scale ranges, only the dynamical ones can be associated to temporal scales. Notably, structural elements such as underlying substrate characteristics are not likely to change during peatland development: their eventually changing influence has to be induced by other parameters.

Study area and site description

The Montréal lowlands (Fig. 46) consist mainly in a flat plain of marine clay deposited above tills on a sandstone-dolomitic bedrock (Laverdière *et al.*, 1972). Several intrusive hills (Monteregian hills) emerge from the plain roughly along an east-west axis. Lakes are rare in the lowlands and, except for the now filled-up Lac Romer in the Lanoraie peatland complex (Fig. 46), they are located on Monteregian hills. The Lanoraie complex and Mirabel bog presently represent the two last important natural peatlands in the Montréal lowlands. The natural vegetation of the region constitutes the northern limit of

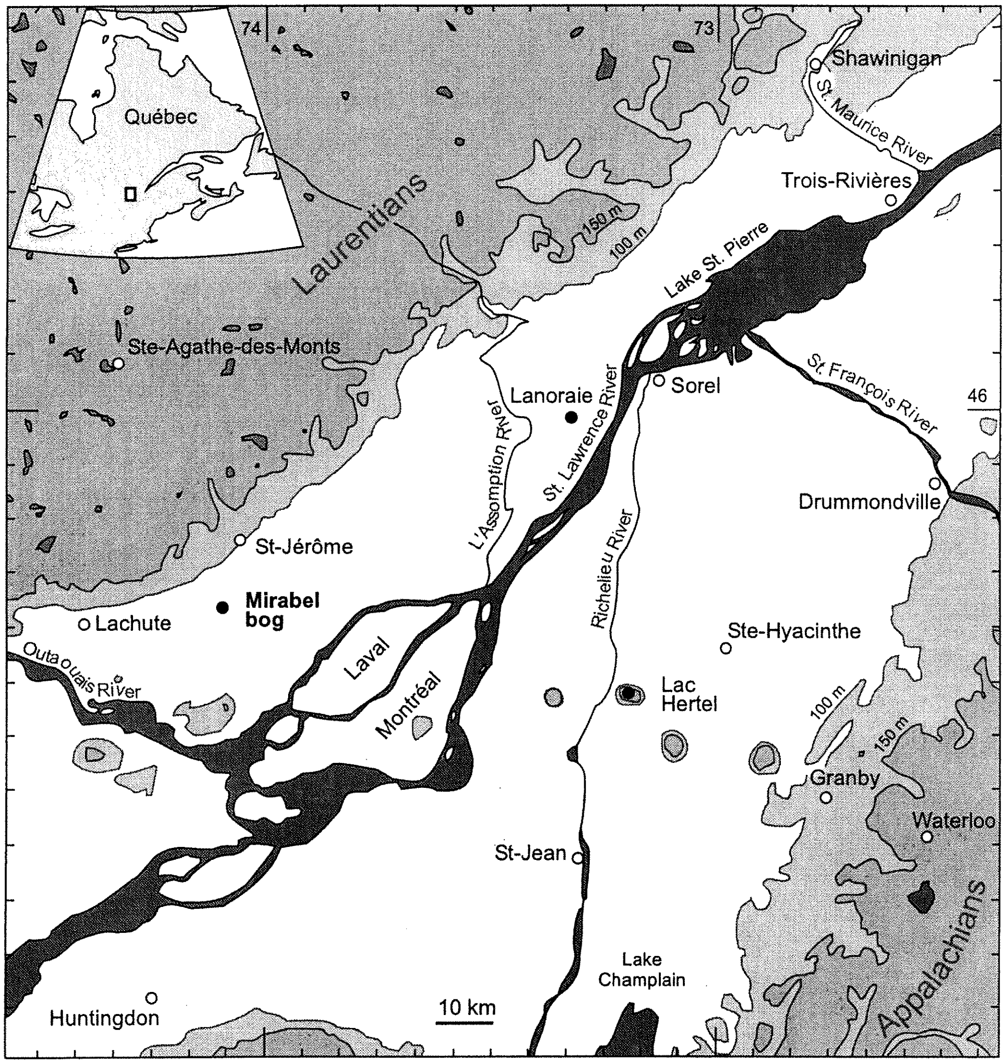


FIGURE 46. Location of Mirabel bog in the St. Lawrence lowlands, southern Québec

deciduous forests dominated by *Acer saccharum*, in association with *Betula alleghaniensis* on highlands, with *Tilia americana* in the north-eastern part of the plain and with *Carya cordiformis* south-westwards of Sorel (Grandtner, 1966; Richard, 1987). The transitional nature of these forests is well marked by the boreal and temperate elements of which they are constituted (Maycock, 1961; Bouchard & Maycock, 1978). However, cultivated lands today occupy the most of the plain, except moraines and bedrock outcrops.

Mirabel bog, previously called St. Canut bog (MacPherson, 1966), is an ombrotrophic, 2.15 km^2 raised bog located at an altitude of 75 m, 35 km north of Montréal island (Figs. 46). The peatland developed in a closed depression localised between the clayey plain of Ste-Monique and the sandstone-dolomitic massif of Belle-Rivière (Fig. 47). Ablation tills formed by the melting of residual ice blocks into the Champlain Sea waters, were generally recovered by marine clay, except in the Ste-Scholastique region where the clayey deposits do not exceed 60-70 m above sea-level. That low altitude is explained by the absence of sedimentation during most of the Champlain Sea period, associated to submarine turbulences (Laverdière *et al.*, 1972). Within the depressions of the Belle-Rivière massif, retreating marine waters deposited some siltous and sandy beaches, like the ones outcropping today in the north of Mirabel bog, below scattered formations of *Pteridium aquilinum*. The modern peatland vegetation is constituted by a mosaic of plant formations from open shrubby communities dominated by Ericaceae and *Sphagnum* spp. to densely forested communities of *Larix laricina* and *Picea mariana* (Fig. 48), also on a carpet of *Sphagnum* spp.

Material and methods

Field and laboratory

Mirabel bog was sampled in 1999 and 2000. Peat depth (pachymetry) was measured from 140 points at each 50 m along five transects and a surface topographic survey totalling 120 measurements was realised in the southern half of the peatland. Fourteen basal peat monoliths were taken for radiocarbon dating and seven complete profiles were collected with a Russian sampler (Jowsey, 1966). The seven cores were investigated for

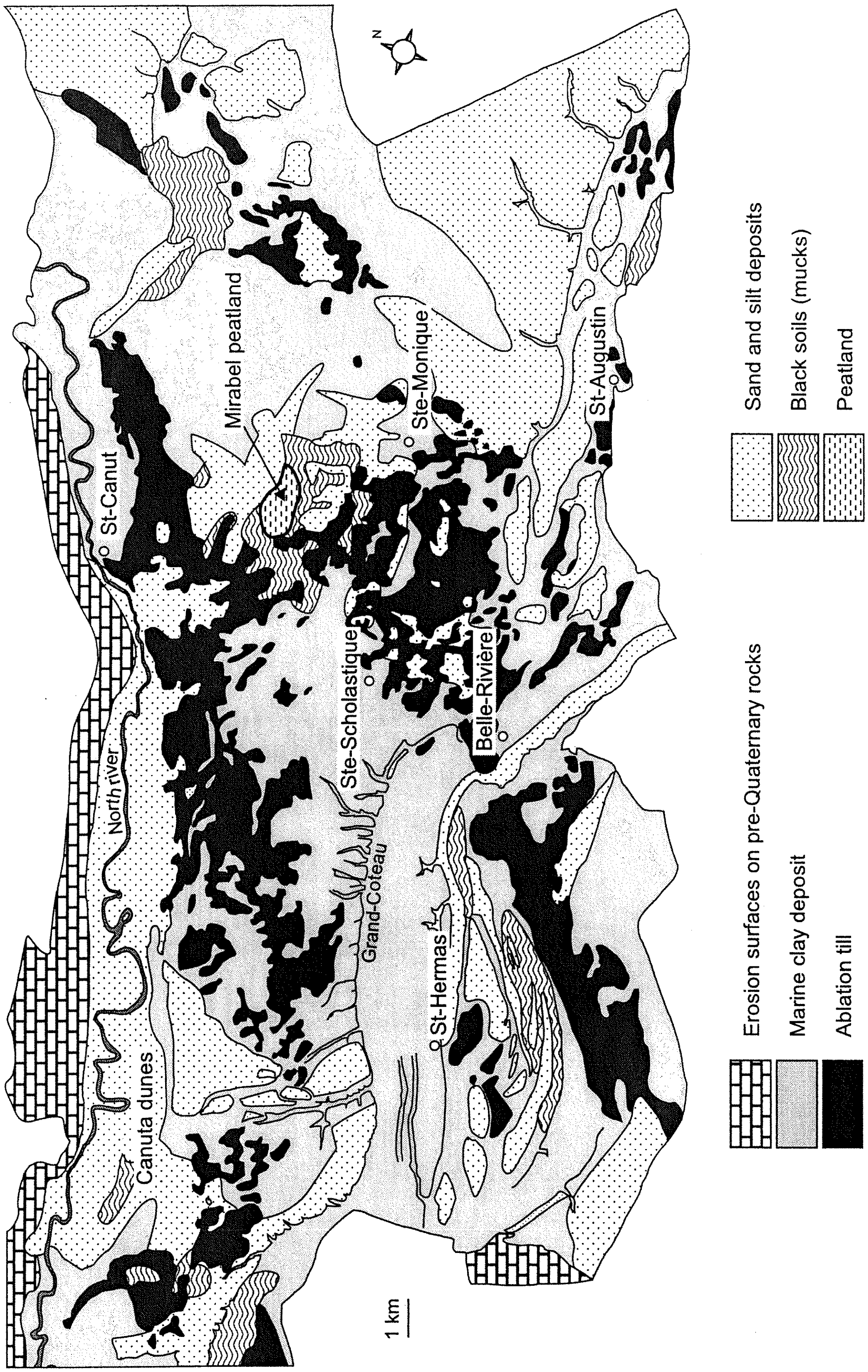


FIGURE 47. Surficial deposits of Mirabel area (following Laverdière *et al.*, 1972)

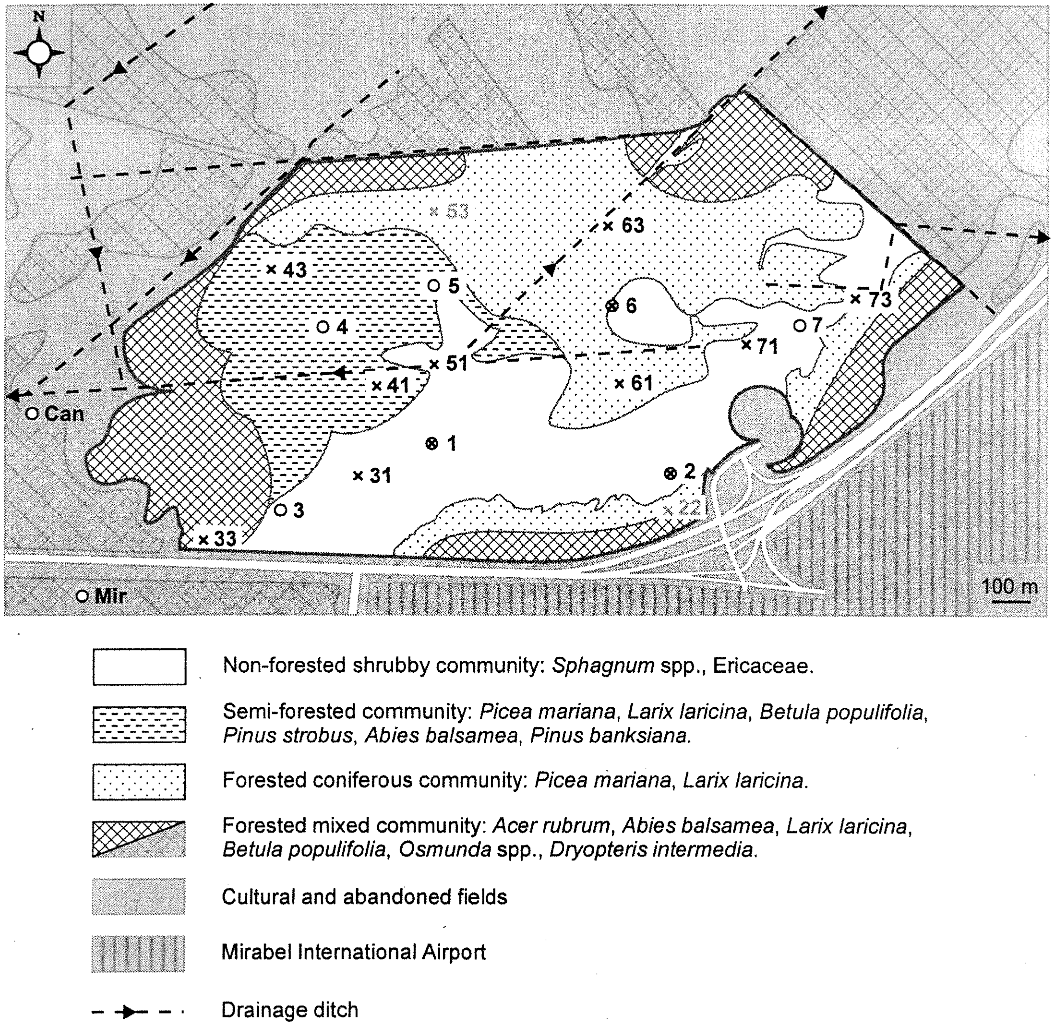


FIGURE 48. Vegetation of Mirabel bog and location of studied profiles (dots) and basal dates (crosses). Rejected dates are noted in grey.

sedimentology, loss-on-ignition, microfossils (pollen, spores, testate amoebae, algae, fungi, microcharcoals) and macrofossils. Two previous studies were used to characterise the vegetation history of the peatland surroundings. Profile Can was collected with a Hiller borer and analysed for pollen by MacPherson (1966). Profile Mir was directly taken on a section along a ditch in 1973, and also analysed for pollen (P.J.H. Richard, unpublished data, 1974).

Other microfossils (testate amoebae, algae, fungi, microcharcoals) were counted during pollen analysis. Chemical treatments followed the conventional method according to Faegri & Iversen (1989) without the hydrofluoric acid step. Macrofossil extraction was realised on 3 cm thick samples, representing volumes between 6 and 180 cm^3, by sifting under soft water jet (Grosse-Brauckmann, 1986). Charcoals were counted during both pollen and macrofossil analyses. Microcharcoals were counted by taking in account their surface areas on three lines per slide and macrocharcoals were identified in order to specify the burned plant communities (S.D. Muller and B. Talon, unpublished data, 2001). Carbon content of sediments was evaluated on 1 cm-thick samples every two cm from loss-on-ignition measurements (at 600°C), by dividing the dry organic matter mass by 2.13 (Dean, 1974).

Chronological control

The chronological control of the Mirabel bog development is provided by sixteen radiocarbon dates (Table 7) and two well-dated regional events (the *Tsuga* decline around 5500 cal. BP and the *Ambrosia* rise around 250 cal. BP). Four AMS dates are rejected: the dates TO-8168, TO-8876 and TO-8881 present abnormal recovery yields, which could have resulted in dating errors (R. Beukens, personal communication, 2000) and the date TO-8169 was rejected on the basis of palynostratigraphic correlations with Mer Bleue bog, located in the St. Lawrence lowlands near Ottawa (P.J.H. Richard, unpublished data, 2000). The age-depth model of Mir1 was constructed using a smooth interpolation between radiocarbon dates (Fig. 49), which allows, by minimising peat accumulation rate variations, to conserve only undoubted changes. This method, which allows avoiding the bias induced by an insufficient or irregular radiocarbon dating, is particularly justified in studies which, like ours, focus on the timing of variations. Other age-depth models (Fig. 49) were

TABLE 7. Conventional and calibrated radiocarbon ages. Rejected ages are noted in italic. AMS dates were performed on terrestrial plant macrofossils.

Core name	Depth (cm)	Laboratory code	Datation method	Calibrated age	Confidence interval 2σ	Conventional age	$^{13}C/^{12}C$ ratio
Mir1	82-88	Beta-135343	Conventional	2180	2340-2060	2200 ± 50	-28.4
	81-90	*TO-8169*	*AMS*	*3630*	*3960-3360*	*3380 ± 120*	*n.a.*
	232-238	Beta-135344	Conventional	6820	7180-6680	6050 ± 80	-26.8
	235-237	TO-8167	AMS	6830	7250-6450	6010 ± 160	n.a.
	318-325	Beta-135345	Conventional	7630	7790-7510	6800 ± 80	-19.8
	321-323	*TO-8168*	*AMS*	*6310*	*6850-5910*	*5550 ± 220*	*n.a.*
	408-410	TO-8166	AMS	9010	9470-8430	8050 ± 170	n.a.
Mir2	321-326	Beta-138673	Conventional	10030	10220-9700	8900 ± 80	-27.6
Mir22	*188-190*	*TO-8876*	*AMS*	*7670*	*7840-7490*	*6820 ± 100*	*n.a.*
Mir31	351-355	Beta-145847	Conventional	8580	8980-8370	7770 ± 110	-26.9
Mir33	210-212	TO-8877	AMS	9540	9710-9490	8600 ± 70	n.a.
Mir41	318-320	TO-8878	AMS	9910	10190-9630	8850 ± 70	n.a.
Mir43	254-257	TO-8879	AMS	9540	9710-9490	8600 ± 70	n.a.
Mir51	291-295	TO-8880	AMS	10030	10210-9700	8890 ± 70	n.a.
Mir53	*186-190*	*TO-8881*	*AMS*	*6820*	*7680-5920*	*6020 ± 430*	*n.a.*
Mir61	296-300	Beta-145846	Conventional	7450	7610-7320	6580 ± 80	-26.3
Mir6	214-217	TO-8882	AMS	9430	9530-9130	8370 ± 80	n.a.
Mir63	115-120	TO-8883	AMS	8410	8590-8350	7650 ± 60	n.a.
Mir71	286-290	Beta-145845	Conventional	8760	9230-8430	7930 ± 120	-25.9
Mir73	196-199	TO-8884	AMS	9570	9910-9530	8670 ± 70	n.a.

n.a.: not available

based on palynostratigraphic correlations with Mir1. The correlated events correspond to boundaries of the regional palynozones described in Muller & Richard (accepted). The inverse trends presented by upland pollen concentrations (Fig. 49) attest to the consistency of age-depth models. Departure from those trends indicate episodes of higher or lower net peat accumulation rates. However, due to the coarse sampling interval, the duration of these departures is unknown. Conventional radiocarbon dates are calibrated with the CALIB 4.0 program (Stuiver & Reimer, 1993). Calibrated dates used in the text are noted cal. BP (calendar years before present, i.e. 1950).

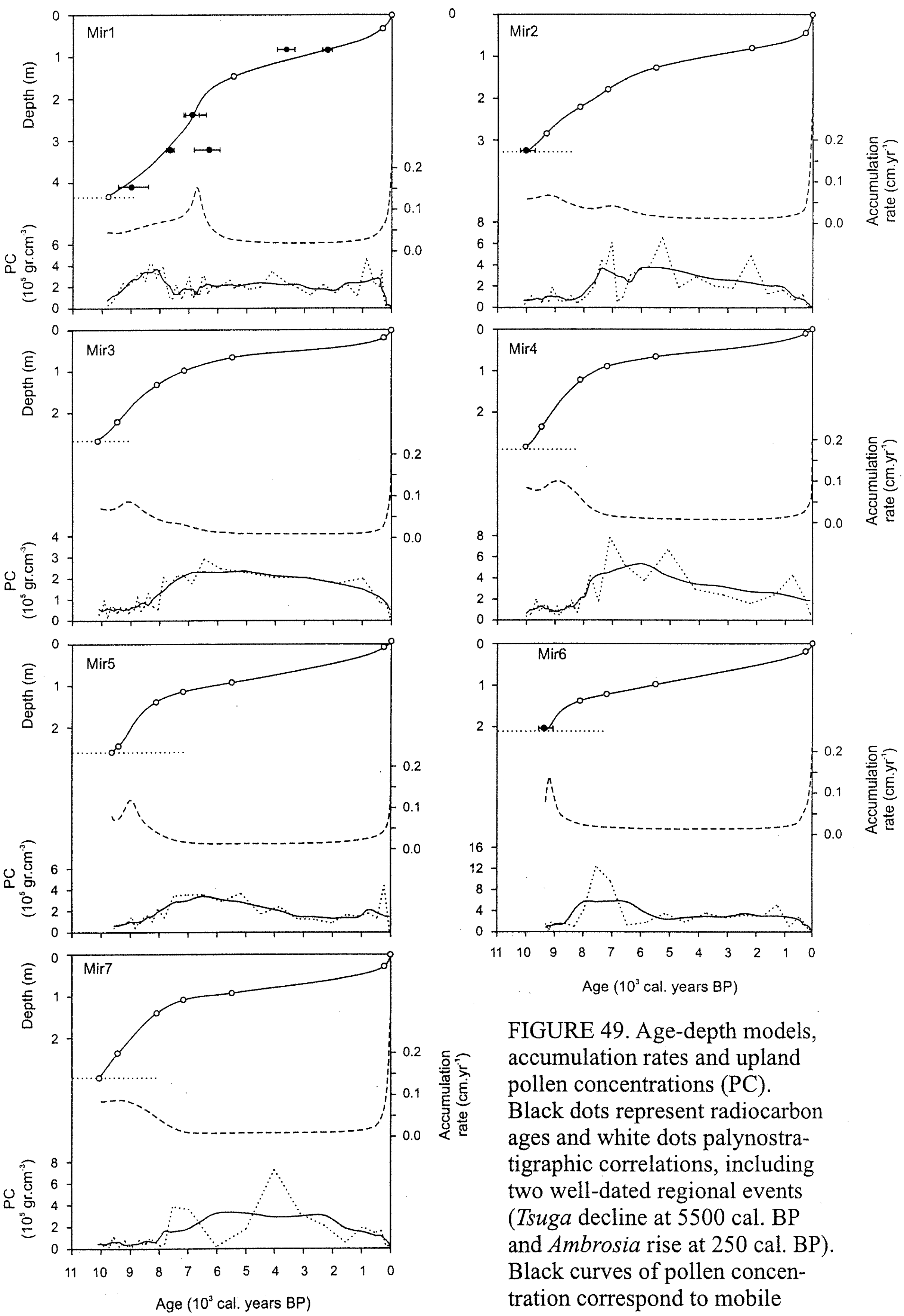


FIGURE 49. Age-depth models, accumulation rates and upland pollen concentrations (PC). Black dots represent radiocarbon ages and white dots palynostratigraphic correlations, including two well-dated regional events (*Tsuga* decline at 5500 cal. BP and *Ambrosia* rise at 250 cal. BP). Black curves of pollen concentration correspond to mobile averages calculated on five points.

Results

Carbon accumulation

Carbon accumulation at Mirabel bog presents consistent temporal trends at community and at peatland scale (Fig. 50). High accumulation rates were recorded during the early Holocene, from 10 000 to 6000 cal. BP, and during the two last centuries. The latter increase is considered to be an artefact due to the lack of decomposition within the acrotelm layer. The two peaks, evidenced at peatland scale around 9000 and 6700 cal. BP (Fig. 50) translate the delay of maximal peat accumulation between margins (Mir2-7) and the deepest zone (Mir1), respectively, which reflects the centripetal spread of associated plant communities and moisture conditions.

Local fire incidences

The charcoal influx, evaluated at the local and peatland scales (Fig. 51), was high during the early Holocene, low during the mid- and late Holocene, and finally high during the two last centuries. The charcoal record could be biased by the varying time resolution resulting from variable accumulation rates (Fig. 49). However, the observed pattern is consistent with the fire regimes previously reported for southern Québec (Richard, 1995; Carcaillet & Richard, 2000; Lavoie & Richard, 2000b).

Peat surface moisture

Past surface moisture conditions were inferred from local microfossils (testate amoebae, fungi, algae) and peat humification (Muller *et al.*, submitted-b). The Mirabel site presented very wet conditions during the early Holocene period followed by several fluctuations (Fig. 52). Notably, four dry periods were evidenced at the peatland scale: 6000, 4400, 2200 and 200 cal. BP. The first and last ones concerned the entire peatland surface area, while the two others were mostly recorded in the northeastern zone. The recent dry conditions can be directly related to man-made draining of the peatland.

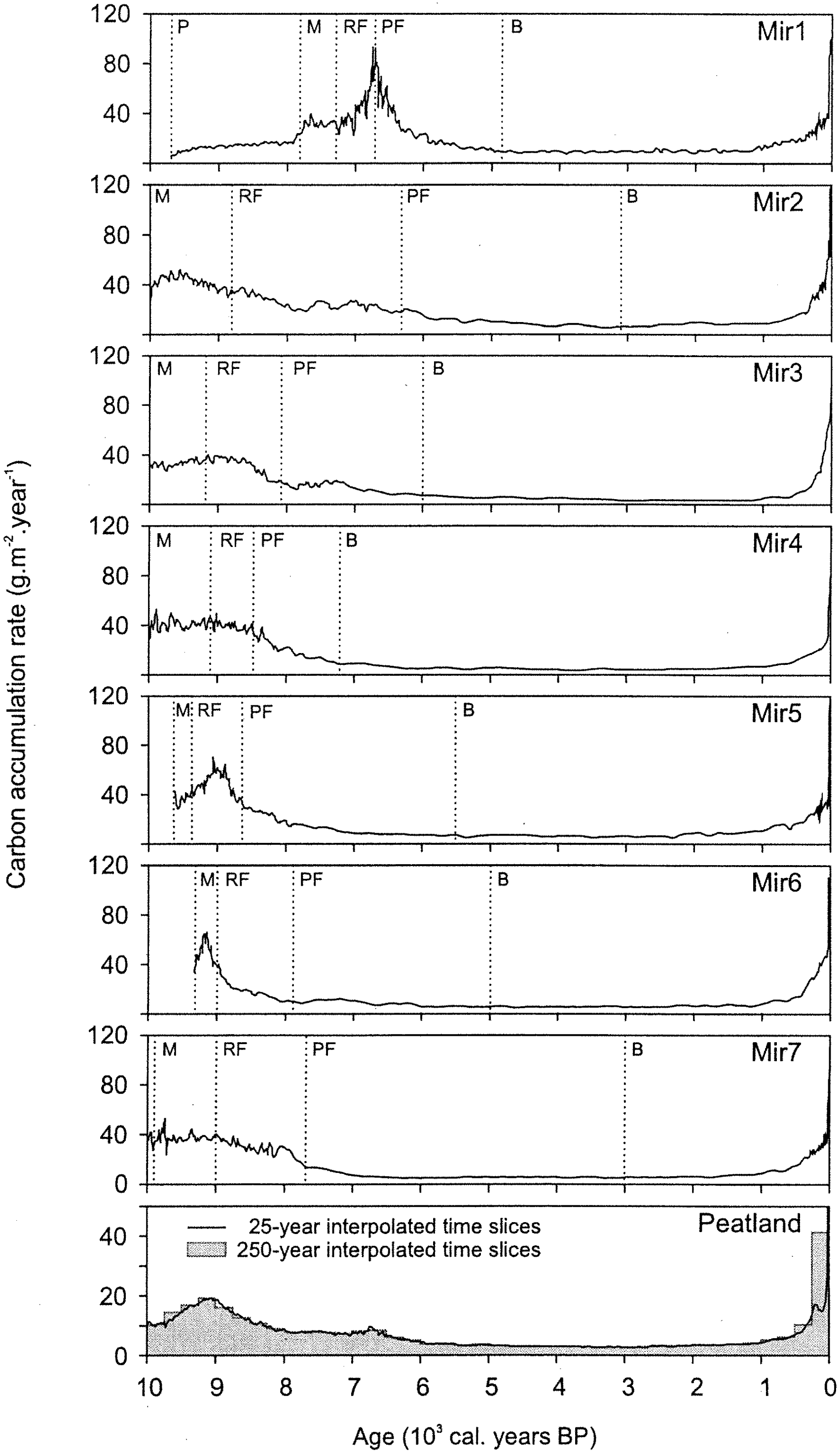


FIGURE 50. Carbon accumulation at Mirabel bog. Letters refer to local vegetation phases. B: Bog; M: Marsh; P: Pond; PF: Poor fen; RF: Rich fen.

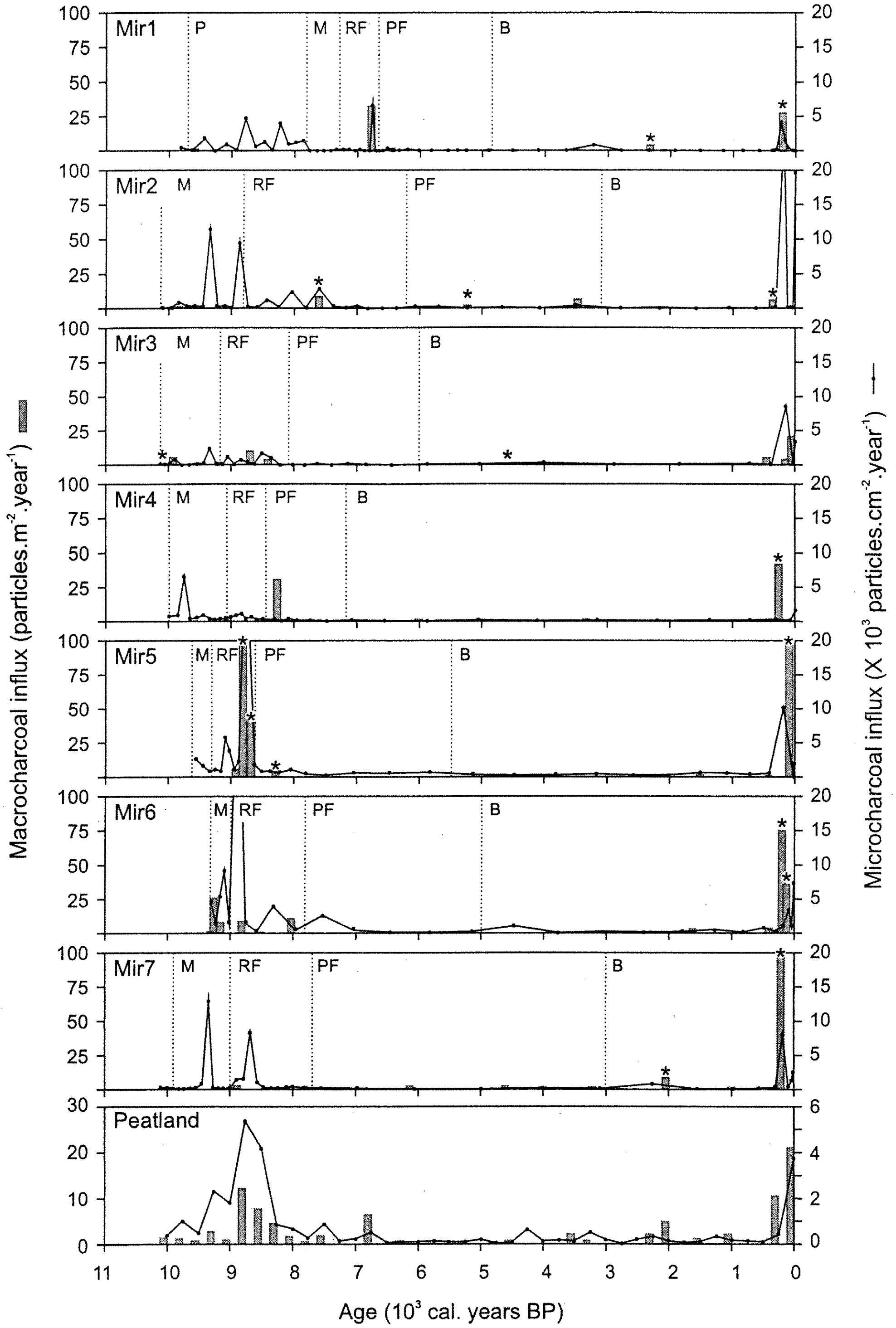


FIGURE 51. Charcoal influxes. Curves represent microcharcoals and histograms macrocharcoals. Asterisks point local fires, as shown by macrocharcoal determination (Muller, S.D., and Talon, B., unpublished data, 2001). The total charcoal influx was calculated by averaging the seven records by time slices of 250 years. Letters refer to local vegetation phases. B: Bog; M: Marsh; P: Pond; PF: Poor fen; RF: Rich fen.

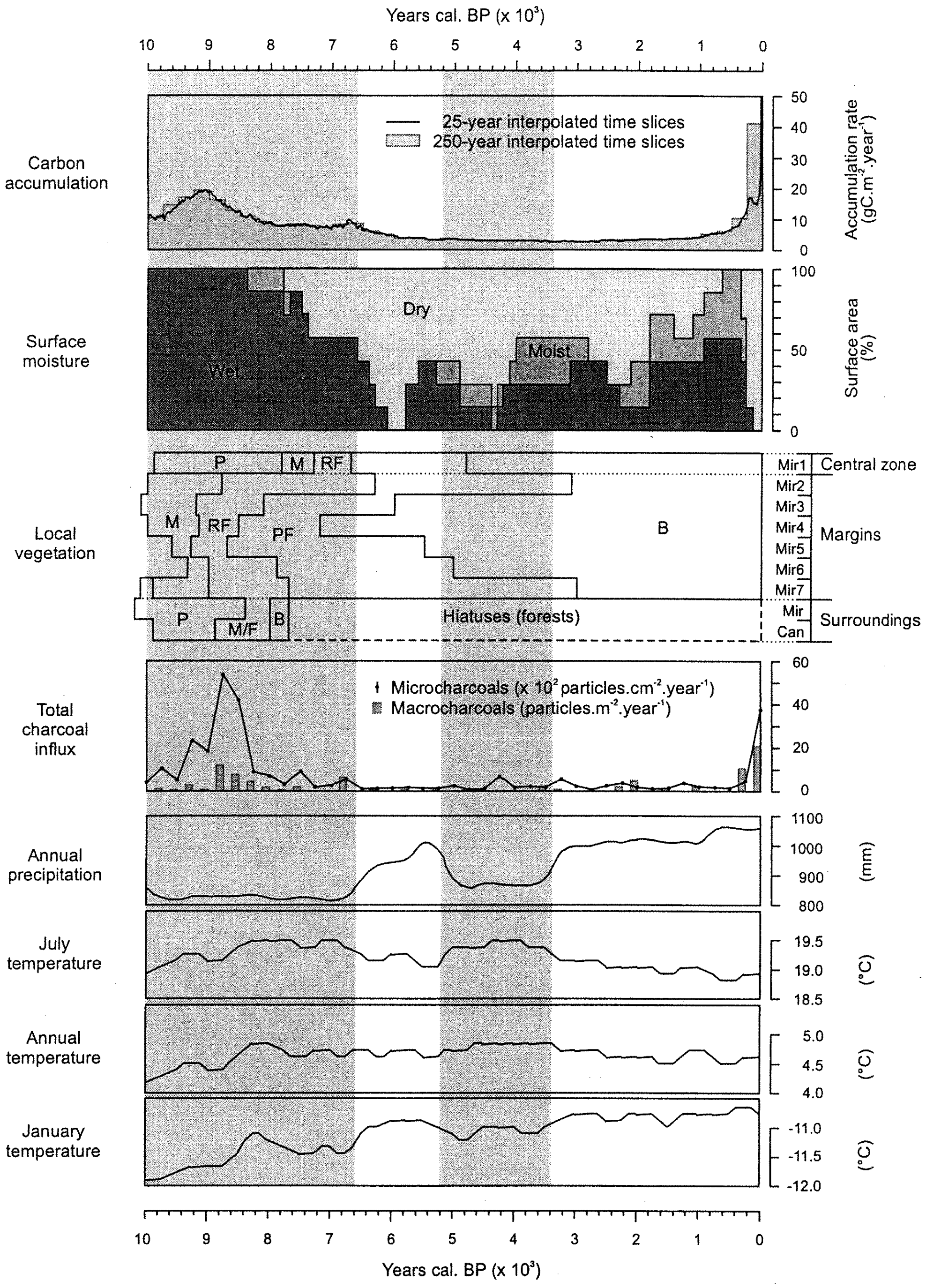


FIGURE 52. Postglacial conditions of Mirabel bog development: autogenic parameters (from Muller *et al.*, submitted-b), climate (from Muller *et al.*, submitted-a) and fire incidences. Grey zones point dryer periods inferred from precipitation reconstruction. Local vegetation phases are: P: Pond; M: Marsh; M/F: Marsh/fen; RF: Rich fen; PF: Poor fen; B: Bog.

Local vegetation

The local vegetation development was generally in conformity with the typical succession described as 'terrestrialisation process' (Weber, 1908; Tallis, 1983), and consists in all profiles in the following successive stages: pond, marsh, rich fen, poor fen and bog (Fig. 52; Muller *et al.*, submitted-b). The asynchrony of community shifts throughout the site increased over the Holocene period, from 2100 years for the transition from marsh to rich fen to 4200 for the transition from poor fen to bog.

Regional climate dynamics

Postglacial temperature and precipitation patterns were reconstructed at a regional scale in Muller *et al.* (submitted-a). The reconstruction was based on the modern analogues method (Guiot, 1990; Guiot *et al.*, 1993) constrained by the water level changes of Lac Hertel, on Mont St. Hilaire (Fig. 46). The reconstruction shows: (1) a dry period between 10 000 and 6500 cal. BP, characterised by a temperature increase until the postglacial optimum, attained around 8000 cal. BP; (2) an important increase in precipitation synchronous to a summer cooling from 6500 to 5000 cal. BP; (3) a second dry period well marked by a low water level in Lac Hertel from 5000 to 3500 cal. BP; and (4) a wet period characterised by relatively low summer and high winter temperatures since 3500 cal. BP.

Discussion

Influence of physiographic context on peatland initiation

The general physiographic context of Mirabel bog has been defined by the last deglaciation which began to uncover south-western Québec around 15 000 cal. BP (Dyke & Prest, 1987). Following ice regression northwards, the Montréal Lowlands experienced successively the *Candona* Lake stage, the Champlain Sea episode, then the *Lampsilis* Lake stage (Elson, 1969; S.H. Richard, 1978; Parent & Occhietti, 1988, 1999). The Champlain Sea period was characterised by the deposition of widespread, thick blue clays over the entire lowlands, a feature that deeply affected the drainage conditions (Brown MacPherson, 1967). These conditions prevented the formation of peatlands on the northern side of the St. Lawrence River, except in two areas respectively isolated from the regional hydrographic network by two successive deltas. Mirabel bog initiated within the Palaeo-

Outaouais River delta (Fig. 47) and Lanoraie peatlands within the Palaeo-St. Lawrence River delta (Comtois, 1979). However, local conditions were different in both areas and resulted in different initiation modes. The irregular clayey plateau of Mirabel favoured initiation by terrestrialisation within shallow depressions, while the permeable sandy depressions of Lanoraie were most favourable to paludification.

Influence of basin topography on peatland initiation and development

The good correlation between isohypsal curves and peat initiation ages (Fig. 53) suggests a strong influence of basin morphometry on the lateral expansion of early peat-forming communities. The basal topography controlled both the centrifugal peatland expansion (paludification) and the centripetal pond filling-in (terrestrialisation). The former influence was previously shown by several studies to be directly related to the substrate slope (Korhola, 1992; Almquist-Jacobson & Foster, 1995; Korhola *et al.*, 1996; Mäkilä, 1997). The control of the terrestrialisation process may be due to a relationship between the water depth and the converging spread of floating mats. The modern distribution of vegetation (Fig. 48) also fits relatively well with basal topography (Fig. 53): deepest zones are non-forested, intermediately deep zones are sparsely forested and shallow zones are densely forested. This implies an important control of vegetation dynamics by substrate topography over the entire Holocene period, probably mediated by internal hydrology.

Hydrological relationships

Peat surface moisture represents the local context of plant growth and partly governs peat accumulation by determining the rates of production and decomposition (Shurpali *et al.*, 1995; Silvola *et al.*, 1996; Waddington & Roulet, 1996). Although it is controlled by the multiple interrelations between external and internal factors (Moore, 1986), this parameter was considered to mainly reflect the interplay between climate, groundwater and internal hydrology (Hilbert *et al.*, 2000; Roulet, 2001). We here confront independent evidences of groundwater supply, climate precipitation and peat surface moisture with the aim to specify their past relationships (Fig. 54).

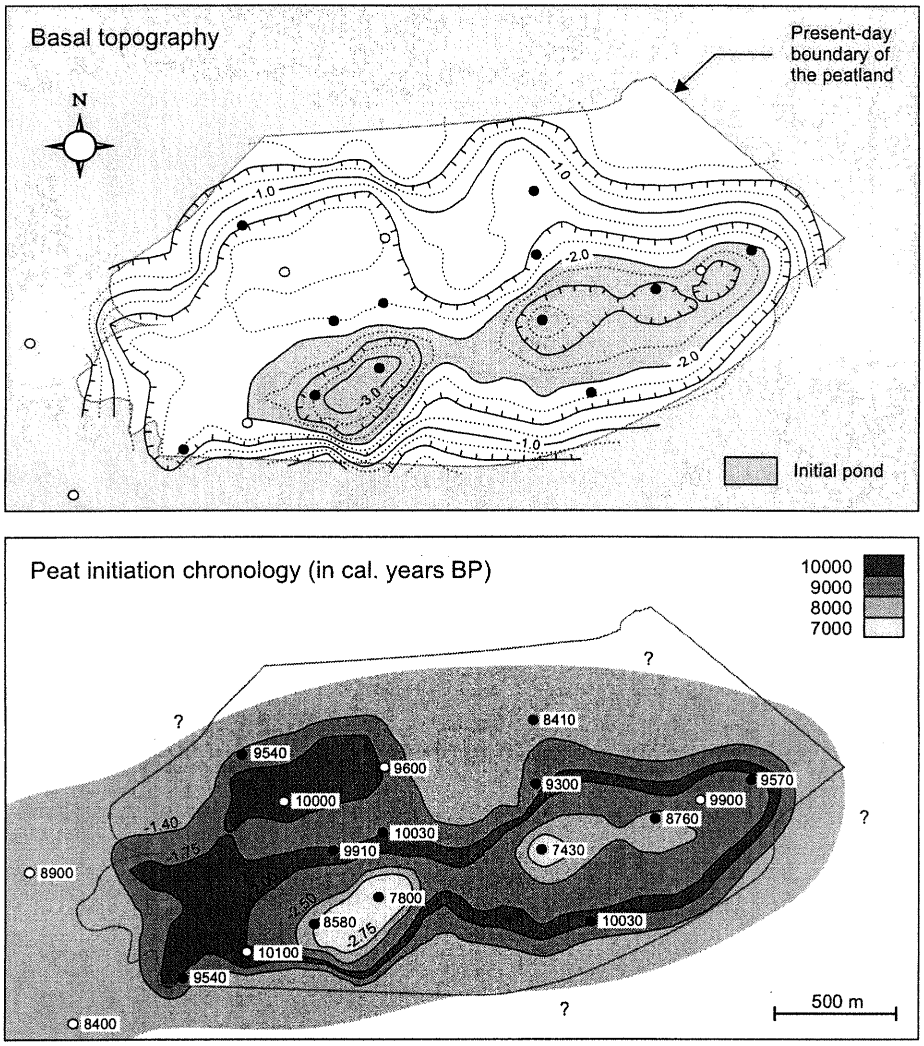


FIGURE 53. Basal topography and chronology of peat initiation at Mirabel bog. The age of peat deposited just above clay, sand or gyttja is obtained from radiocarbon dating (black dots) and palynostratigraphic correlations (white dots). These ages fit well with the isohypsal curves -1.40, -1.75, -2.00, -2.50 and -2.75 m (by reference to the mean altitude of surrounding fields).

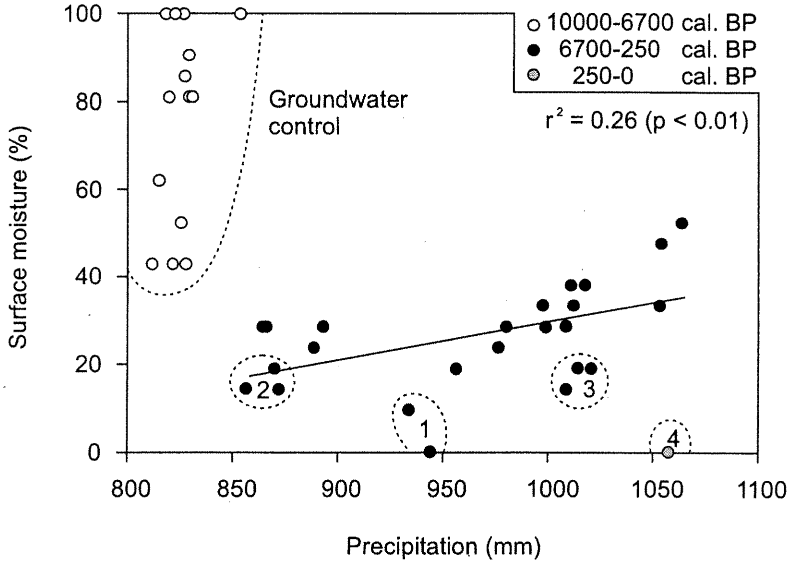


FIGURE 54. Relationships between precipitation and peatland-scale surface moisture. The regression curve is calculated for the period 6700-250 cal. BP (black dots). Numbers from 1 to 4 point the local dry phases recorded around 6000, 4400, 2200 and 200 cal. BP, respectively.

Occurrences of *Calliergon trifarium* remains reveal the persistence of highly alkaline conditions until 6700 cal. BP in the central part of Mirabel bog (Muller *et al.*, submitted-b). These conditions reflect the influence of the basal marine clay, rich in carbonates inherited from the abrasion of the dolomitic bedrock by glaciers (Laverdière *et al.*, 1972). The early Holocene groundwater supply was contemporaneous with very wet surface conditions and low precipitation values (Fig. 52), which points to its predominant influence on initial wetland stages (Fig. 54). The marked decrease in peat surface moisture occurring between 8000 and 6000 cal. BP (Fig. 52) resulted from the mire surface elevation, which progressively isolated the peat-forming vegetation from groundwater. Since 6700 cal. BP, changes in surface moisture were highly asynchronous throughout the peatland (Muller *et al.*, submitted-b), suggesting local controls rather than climatic ones. The four dry phases, evidenced at the peatland scale around 6000, 4400, 2200 and 200 cal. BP (Fig. 52), occurred over a large range of annual precipitation, from 850 to 1060 mm (Fig. 54). The last one, resulting from the anthropogenic cutting of drainage ditches, was clearly not related to climate. Among the three other dry intervals, only the first one was simultaneously recorded at the seven coring sites, but during an important increase in precipitation (Fig. 52). This enforces the idea that, despite a weakly positive correlation between precipitation and peat surface moisture from 6700 to 250 cal. BP (Fig. 54), the postglacial control exerted at Mirabel bog by local factors was predominant over hydroclimatic influence.

Role of disturbances

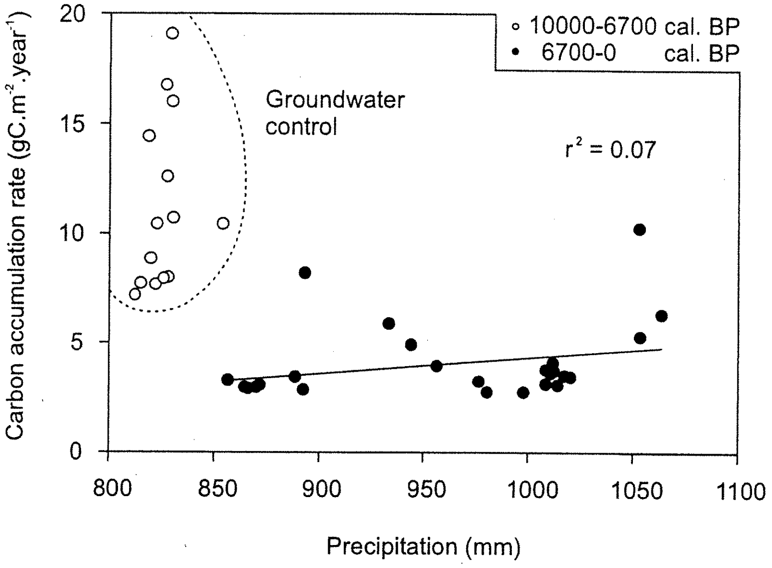
As shown by micro- and macrocharcoal records (Fig. 51), local fires occurred principally during the early- and the late Holocene periods, synchronously with important peat accumulation rates (Fig. 52). Early fire events did not destroy peat probably because of the waterlogged conditions of the site (Kuhry, 1994), but they could have maintained sedge meadows (marshes and fens) by inhibiting the growth of shrubs and trees (Davis, 1979). Also surprisingly, the charcoal record suggests that the relatively dry peat surface conditions occurring from roughly 6000 to 2000 cal. BP (Fig. 52) have not promoted peat surface fires. That feature could be related to the development of communities less flammable than sedge meadows (Davis, 1979) or, more probably, to a general mid-Holocene decrease in fire frequency (Carcaillet & Richard, 2000). Finally, burned macroremains of *Polytrichum strictum* and Ericaceae attest peat destruction during the late Holocene, and principally during the two last centuries (Fig. 51). Recent peat surface fires may have been favoured by the man-induced draining of Mirabel bog (Fig. 52), which is attested by the numerous ditches still visible throughout the whole peatland (Fig. 48). The dry conditions then prevailing on the peatland were also favourable to the spread of *Betula populifolia*, *Nemopanthus mucronatus* and *Aronia melanocarpa* (Muller *et al.*, submitted-b).

Controls of carbon accumulation

Carbon accumulation is characterised, both at community and at peatland scales, by high values during early Holocene and recent centuries (Fig. 50). This long-term trend translates mainly the groundwater influence, which enhanced plant productivity and peat accumulation prior to 6700 cal. BP, contemporaneously with a dry climatic interval. The lack of minerotrophic supply since this date resulted in regular, low carbon accumulation, which reflected neither peat surface moisture fluctuations nor climate changes (Fig. 52). Moreover, carbon accumulation appears to have been as much independent from precipitation as from temperature (Fig. 55, Table 8).

TABLE 8. Relationships between climate and carbon accumulation evaluated by a multiple correlation analysis. Only independent climate parameters were used: growing degree-days above 5°C (GDD5), coldest month temperature (CMT) and annual precipitation (P). Correlation coefficients and bootstrap verification evidence no significant linear correlation between climate and peat accumulation.

Peat accumulation	Multiple correlation coefficient	Verification correlation	Regression coefficients		
			GDD5	CMT	P
Mir1	0.269 ± 0.081	-0.192 ± 0.251	0.027	0.191	-0.480
Mir2	0.417 ± 0.284	-0.203 ± 0.286	-0.294	-0.085	-0.228
Mir3	0.458 ± 0.262	0.245 ± 0.410	-0.328	-0.165	-0.319
Mir4	0.680 ± 0.150	0.532 ± 0.302	-0.402	-0.371	-0.406
Mir5	0.619 ± 0.158	0.485 ± 0.322	-0.470	-0.298	-0.384
Mir6	0.474 ± 0.228	0.082 ± 0.313	-0.420	-0.121	-0.212
Mir7	0.522 ± 0.200	0.318 ± 0.380	-0.394	-0.225	-0.331
Peatland	0.390 ± 0.252	0.060 ± 0.341	-0.327	-0.064	-0.283



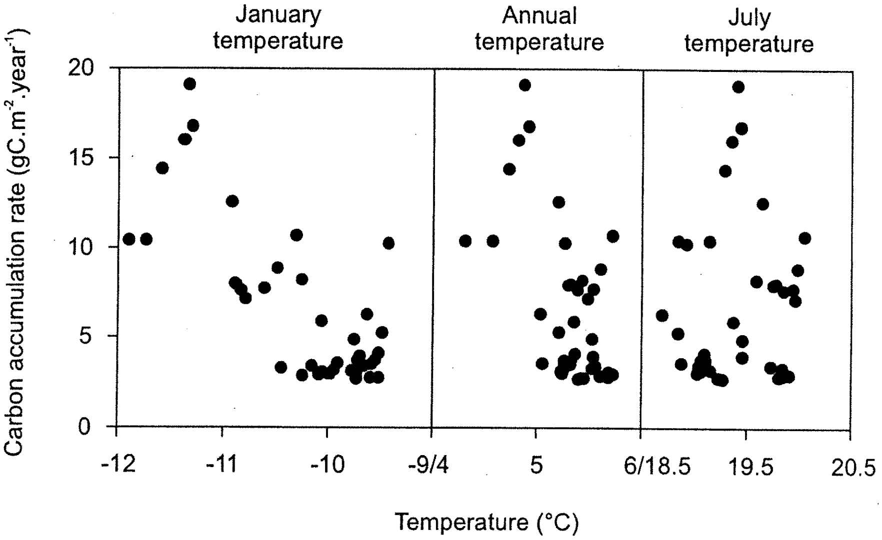


FIGURE 55. Relationships between climate and carbon accumulation. The regression curve of precipitation is calculated for the period 6700-0 cal. BP.

Short-term fluctuations, which are proportional to the carbon accumulation rate (Fig. 50), reproduce precisely the variations in loss-on-ignition measurements and cannot be interpreted as reflecting any past influence. Consequently, carbon accumulation curves do present only long-term trends, and no obvious medium- or short-term ones as reconstructed by previous works (e.g. Kilian *et al.*, 2000). This could be related to the smooth interpolation method used to construct age-depth models (Fig. 49), which minimises the inferred peat accumulation rates. However, pollen concentrations confirm the long-term trends and do not present any other general feature common to all profiles (Fig. 49). Mir2 is the only core to present a slight medium-term fluctuation, which is simultaneously evidenced between 7000 and 6000 cal. BP by radiocarbon dates and pollen concentrations. This observation suggests that our carbon accumulation reconstructions are, despite their simplicity, likely to translate medium-term fluctuations as well as long-term ones.

Conclusions

Two periods, based on autogenic relationships, were previously recognised in the postglacial development of Mirabel bog: local vegetation, surface moisture and carbon accumulation were strongly interrelated from 10 000 to 6700 cal. BP and independent since 6700 cal. BP (Muller *et al.*, submitted-b). The integration of allogenic factors (physiography, basal topography, climate and disturbances) provides insights into past controls of carbon accumulation and specifies the respective influences of each external parameter.

The postglacial carbon accumulation at Mirabel bog reflects almost exclusively the evolution of the trophic status of the ecosystem: high rates were related to the early minerotrophic groundwater supply, while low rates were associated with later ombrotrophic conditions. Carbon accumulation consequently appears to have resulted from the interplay between minerotrophic supply and autogenic processes, which progressively induced the isolation of peat-forming vegetation from groundwater. Similar initially alkaline conditions were reported from Finland (Korhola, 1992), Ontario (Campbell *et al.*, 1997) and western Canada (Kuhry *et al.*, 1993; Kuhry, 1997), suggesting that they could have been more important in early wetland dynamics than previously thought.

Other physiographic parameters have strongly influenced the developmental history of the peatland. Notably, the geomorphological context determined the location and extent of peat deposits and basal topography controlled the rate of lateral expansion, both in the centrifugal (paludification) and centripetal (terrestrialisation) directions. By contrast, the peatland development appears to have been independent of climate and fire during the entire Holocene period. The lack of linkage between peatland and regional hydroclimatic conditions was also reported from two other southern Québec sites (Lavoie & Richard, 2000a, 2000b), but these authors proposed that temperature could have played an important role in past peatland dynamics; this is definitively not observed at Mirabel bog. The independence of carbon accumulation from fire is evidenced by the synchrony between high charcoal influxes and high accumulation rates. Compared to previous studies emphasising the role of fire in peatland development (e.g., Kuhry, 1994; Mäkilä, 1997; Pitkänen *et al.*, 1999), this points to the influence of local moisture on the relationships between fire and peat accumulation.

Our results suggest that the linkage previously evidenced between peatland and climate (Aaby, 1976; Barber, 1981; Winkler, 1988; Campbell *et al.*, 1997; Hendon *et al.*, 2001) depends on the geographical location and cannot be generalised. Mirabel bog constitutes an example of peatland which development was controlled by physiography and autogenic processes, independently from climate.

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CONCLUSIONS ET PERSPECTIVES

Le but de cette thèse a été d'explorer les influences exercées par les diverses conditions de l'accumulation du carbone dans une tourbière du Québec méridional à la lumière de leurs interrelations. Il a pour cela été nécessaire de reconstituer de façon indépendante l'ensemble des paramètres susceptibles d'avoir joué un rôle et de les confronter entre eux, et avec l'accumulation du carbone.

Le contexte paléophytogéographique

Dans un premier temps, le contexte paléophytogéographique fut reconstitué à partir de quinze diagrammes polliniques régionaux (dont un partiel) et de soixante-quatorze dates radiocarbone. Les objectifs de ce travail furent d'étudier les patrons spatiaux de la succession végétale postglaciaire (Richard, 1977a) à l'échelle des basses terres montréalaises et d'en préciser les facteurs déterminants. Deux modes de migration ont été mis en évidence : les formations pionnières (toundra, pessières et sapinières), au contraire des formations matures (érablières), ont présenté des tendances migratoires nettes depuis les Appalaches, au sud, jusqu'aux Laurentides, au nord. Ces dynamiques contrastées paraissent pouvoir être essentiellement imputées à des facteurs non climatiques. Les lacs et mers proglaciaires recouvrant les basses terres ont constitué au tardiglaciaire et au début de l'Holocène des barrières géographiques à la progression de la végétation. La physiographie, notamment dans les îles formées par les collines montérégiennes au sein de la Mer de Champlain, a également influencé de manière forte les dynamiques végétales initiales. Enfin, les conditions locales et les phénomènes de compétitions ont joué un rôle majeur depuis l'Holocène moyen.

Ces résultats mettent en évidence la complexité de la relation entre végétation et climat à l'échelle régionale. Au vu de l'hétérogénéité spatiale de la végétation postglaciaire des basses terres montréalaises, il apparaît impossible d'en reconstituer les dynamiques climatiques sur la seule base d'études polliniques isolées comme l'ont fait Winkler (1988) et Campbell *et al.* (1997). Lavoie (1998) s'est d'ailleurs appuyé sur une reconstitution de niveaux lacustres (Lavoie & Richard, 2000c) pour appréhender l'influence des conditions hydroclimatiques sur le développement de deux tourbières des Appalaches et des Laurentides (Lavoie & Richard, 2000a, 2000b). L'absence de corrélation entre les fluctuations de niveaux lacustres et la dynamique de l'entourbement, ainsi que les différences d'accumulation entre les deux tourbières, ont conduit ces auteurs à proposer un contrôle plus important exercé par les températures et les conditions autogènes.

Le climat régional

La deuxième étape de la thèse fut de réaliser une reconstitution climatique régionale dans le but de compléter les résultats de Lavoie (1998) et Lavoie & Richard (2000a, 2000b, 2000c). La méthode des analogues modernes développée par Guiot (1990) a permis de reconstituer de manière quantitative les dynamiques climatiques passées, par la recherche d'analogues actuels aux spectres polliniques fossiles. Jusqu'à présent, cette méthode avait surtout été appliquée dans le but d'étudier les patrons spatiaux des dynamiques climatiques, en comparant entre elles des reconstitutions réalisées pour des sites individuels (par exemple, Overpeck *et al.*, 1992 ; Cheddadi *et al.*, 1998 ; Peyron *et al.*, 1998 ; Sawada *et al.*, 1999 ; Davis *et al.*, 2000). L'approche développée dans cette thèse consiste à reconstituer une dynamique climatique régionale en moyennant plusieurs reconstitutions individuelles. Ceci permet, en réduisant l'influence des paramètres locaux tels que la proximité des masses d'eau (Davis *et al.*, 2000), d'obtenir une reconstitution plus robuste. Sept sites bien datés ont été choisis au sein de la région étudiée et fait l'objet de reconstitutions indépendantes, qui ont ensuite permis le calcul de moyennes régionales. En outre, la reconstitution climatique a pu être améliorée, en particulier en ce qui a trait aux paramètres hydroclimatiques, par la prise en compte de données de niveaux lacustres (Guiot *et al.*, 1993 ; Magny *et al.*, 2001). Le lac Hertel, situé sur une colline montérégienne au centre des basses terres du St-Laurent, fut spécialement étudié afin de reconstituer les fluctuations passées de son niveau et de les utiliser pour contraindre la reconstitution climatique.

Les résultats obtenus ont permis de mettre en évidence trois phases sèches au cours de l'Holocène, dont la première (14 000-11 500 cal. BP) fut caractérisée par des températures froides et les deux suivantes (10 000-6500 ; 5000-3500 cal. BP) par des températures plus élevées (Muller *et al.*, soumis-a). La correspondance de la plupart des variations reconstituées pour les basses terres montréalaises avec des événements mis en évidence dans le golfe du St-Laurent et le nord-est des USA (Webb & Bryson, 1972 ; Webb & Clark, 1977 ; Webb *et al.*, 1983 ; Harrison & Metcalfe, 1985 ; Harrison, 1989 ; Vernal *et al.*, 1993 ; Hu *et al.*, 1999 ; Davis *et al.*, 2000) montre la validité de notre reconstitution.

Dynamiques autogènes du développement de la tourbière de Mirabel

Les contextes phytogéographique et climatique régionaux ainsi définis, la troisième étape de la thèse fut de préciser les dynamiques postglaciaires des facteurs autogènes : végétation locale, hydrologie de surface et accumulation du carbone (Muller *et al.*, soumis-b). Afin d'obtenir des reconstitutions générales pour l'ensemble du site, sept profils répartis sur l'ensemble de la tourbière firent l'objet d'analyses paléoécologiques détaillées qui furent, dans un second temps, interpolées à la surface totale. Les dynamiques passées des paramètres autogènes furent ainsi reconstituées en trois dimensions, mettant en évidence l'importante hétérogénéité spatiale qui a caractérisé la végétation locale et l'hydrologie de surface tout au long de l'Holocène. Ce résultat souligne une fois de plus la nécessité d'étudier plusieurs profils lors de l'étude du développement des tourbières. L'accumulation du carbone présente en revanche une dynamique relativement homogène sur l'ensemble du site, bien que traduisant la progression centripète de la végétation tourbigène lors du comblement de l'étang initial. Le début de l'Holocène fut caractérisé par d'importantes accumulations de carbone, culminant autour de 19.0 $gC.m^{-2}.an^{-1}$ pour l'ensemble de la tourbière, avec un maximum de 93.2 $gC.m^{-2}.an^{-1}$ enregistré dans la partie centrale. La productivité tourbigène s'est par la suite fortement réduite pour atteindre 2.7 $gC.m^{-2}.an^{-1}$ autour de 3000 cal. BP. La période récente a finalement enregistré de fortes valeurs d'accumulation, attribuées à la réduction progressive du degré de décomposition au sein de l'acrotelme et à la faible compaction de la tourbe de surface (Muller *et al.*, soumis-b). Les résultats obtenus ne montrent aucun lien direct entre les différents paramètres autogènes (Fig. 56, voies 1-3), excepté entre 10 000 et 7000 cal. BP où ils étaient tous fortement influencés par la nappe phréatique minérotrophe.

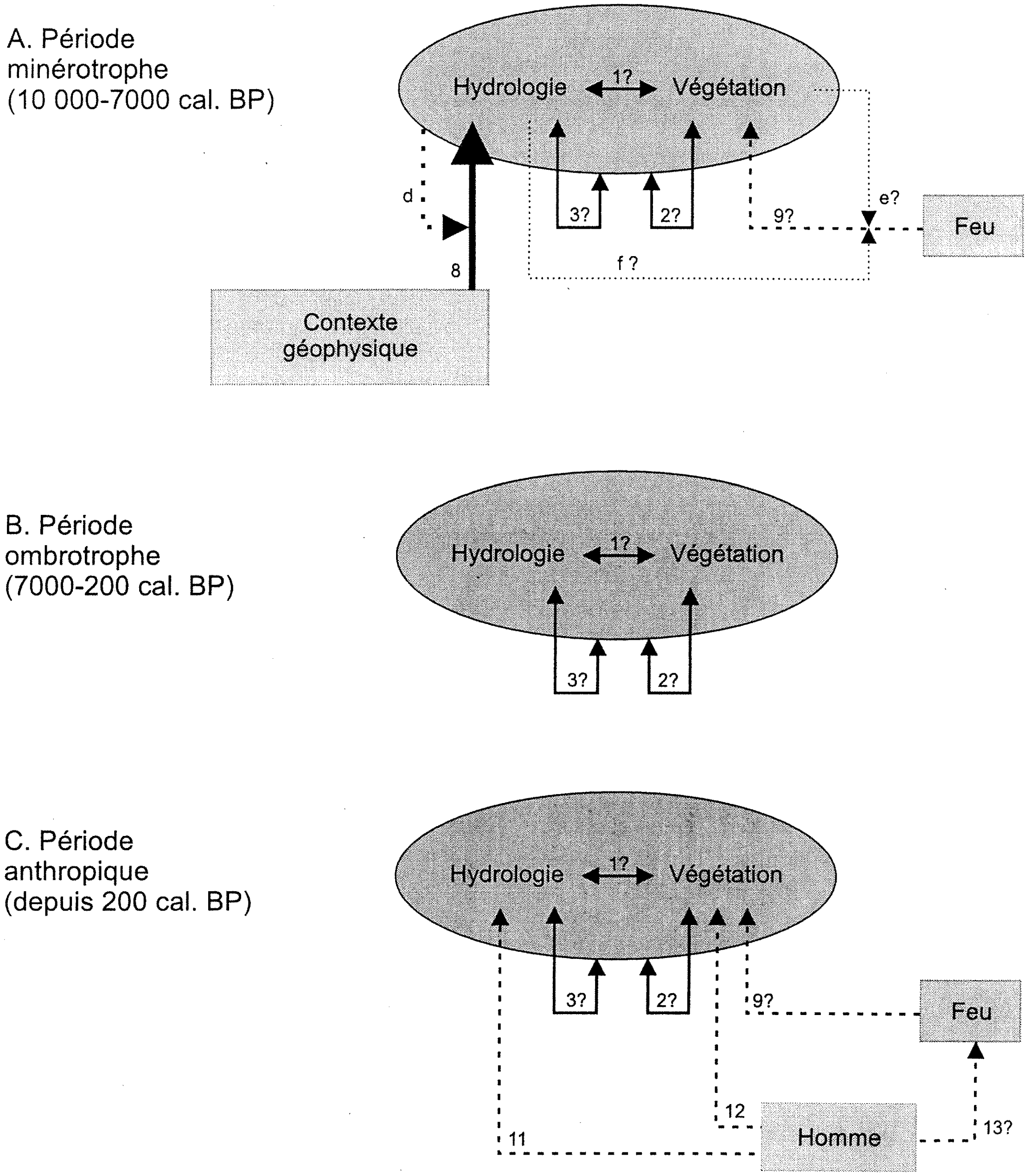

FIGURE 56. Reconstitution des conditions postglaciaires de l'accumulation du carbone dans la tourbière de Mirabel, d'après le modèle conceptuel présenté Fig. 5.
A. L'accumulation du carbone est contrôlée par les interrelations entre le contexte géophysique (nappe phréatique minérotrophe; voie 8) et les paramètres autogènes (voie d). Les importants influx de charbons attestent une situation favorable à des incendies locaux (voie 9), qui n'ont cependant pas influencé le processus d'accumulation de la tourbe.
B. L'accumulation du carbone n'a été influencée ni par le climat, ni par les perturbations, ni par le contexte géophysique. Elle semble donc avoir été essentiellement contrôlée par les paramètres autogènes (voies 1-3).
C. L'action de l'homme et la recrudescence d'incendies locaux ont fortement modifié la végétation et l'hydrologie de surface de la tourbière (voies 9, 11 et 12). Toutefois, l'influence de ces perturbations récentes sur l'accumulation du carbone est difficile à préciser, principalement en raison de la faible décomposition et de l'absence de compaction de la tourbe de surface.

Conditions de l'accumulation du carbone

A l'issue de cette reconstitution spatialisée du développement de la tourbière, l'ensemble des données concernant les conditions de l'accumulation du carbone était réuni : le contexte physiographique (Laverdière *et al.*, 1972), le climat régional (Muller *et al.*, soumis-a), le régime des feux, la topographie basale et les facteurs autogènes (Muller *et al.*, soumis-b). La confrontation de ces différents paramètres fit l'objet de la quatrième et dernière partie de cette thèse (Muller *et al.*, soumis-c). La physiographie apparaît avoir joué un rôle déterminant quant à la localisation, l'extension et la dynamique initiale de la tourbigenèse (Fig. 56A, voie 8). En particulier, la morphométrie du bassin a fortement contrôlé à la fois l'expansion latérale centrifuge (terrestrialisation) et le comblement centripète de l'étang originel (paludification). L'alimentation par les eaux minérotrophes provenant de la nappe phréatique paraît toutefois avoir constitué l'influence majeure des dynamiques internes initiales (Fig. 56A, voie 8). Sous l'effet de l'accumulation de la tourbe, la végétation tourbigène et les conditions hydriques de surface se sont progressivement affranchies des apports minérotrophes (Fig. 56, voie d).

L'absence de relation entre le climat et les dynamiques autogènes, et l'important métachronisme des changements de végétation et d'hydrologie au sein de la tourbière, attestent son indépendance vis-à-vis du climat. Ce résultat est cohérent avec ceux précédemment obtenus par Hu *et al.* (1995) dans le Maine et Lavoie & Richard (2000a, 2000b) au Québec méridional ; il s'oppose en revanche aux observations réalisées au Danemark et au Royaume-Uni, qui attestent un fort contrôle climatique sur les dynamiques tourbigènes (Aaby, 1976 ; Barber, 1981 ; Hendon *et al.*, 2001). Ces différences suggèrent que l'influence exercée par le climat n'est pas universelle, et qu'elle est susceptible de varier considérablement selon la zone géographique concernée. Les feux paraissent également n'avoir eu aucune influence significative sur l'accumulation postglaciaire du carbone à Mirabel (Fig. 56A et C) contrairement aux observations reportées dans des études antérieures (Kuhry, 1994 ; Pitkänen *et al*, 1999 ; Robichaud, 2000). L'absence de toute influence allogène sur le développement de la tourbière de Mirabel depuis 7000 ans implique son contrôle exclusif par les facteurs autogènes (Fig. 56B, voies 1-3). Les relations entre végétation, hydrologie et accumulation du carbone n'ont toutefois pas été directement mises en évidence par les données paléoécologiques.

Influence du climat : hypothèse d'un effet de seuil

Les résultats obtenus à Mirabel apportent un éclairage nouveau à la problématique du déterminisme de la tourbigenèse en général, et des relations climat-tourbière en particulier. Ils attestent pour la première fois l'indépendance d'une tourbière vis-à-vis du climat. Les études antérieures avaient mis en évidence, soit de forts contrôles climatiques (Aaby, 1976 ; Barber, 1981 ; Hendon *et al.*, 2001), soit une action conjointe du climat et des facteurs autogènes (Winkler, 1988 ; Kuhry *et al.*, 1992 ; Hu & Davis, 1995 ; Campbell *et al.*, 1997 ; Mäkilä, 1997 ; Lavoie & Richard, 2000a, 2000b ; Robichaud, 2000). Bien que la plupart de ces travaux ne permet pas d'évaluer précisément le rôle – ou l'absence de rôle – joué par le climat, l'ensemble des données recueillies révèle des influences climatiques extrêmement variables selon les tourbières. Cette variabilité du contrôle climatique constitue vraisemblablement un des éléments clé du processus de la tourbigenèse, mais les données disponibles à l'heure actuelle paraissent encore trop parcellaires pour permettre sa compréhension. Tout au plus est-il possible d'esquisser des hypothèses quant à ses causes et mécanismes, qui pourront servir de pistes pour de futures recherches.

Les tourbières, comme tous les écosystèmes, se développent sous le contrôle contraignant du climat. Cette influence climatique générale, qui détermine les conditions favorables ou défavorables à leur développement, se traduit notamment par la correspondance des gradients géographiques et des gradients climatiques (Damman, 1979 ; Foster & Jacobson, 1990 ; Almquist-Jacobson & Foster, 1995). Le climat est en outre susceptible d'agir à un second niveau, en tant que facteur limitant. La variabilité de l'influence climatique évoquée ci-dessus suggère l'existence d'un seuil, en deçà duquel le climat limiterait la tourbigenèse, et au-delà duquel ses variations resteraient sans effet (Fig. 57). Dans le cadre de cette hypothèse, les tourbières du Québec méridional (Lavoie & Richard, 2000a, 2000b ; Muller *et al.*, soumis-c) et du Maine (Hu *et al.*, 1995) pourraient être interprétées comme ayant toujours bénéficié d'un climat non limitant, au contraire de celles du Danemark (Aaby, 1976) et du Royaume-Uni (Barber, 1981 ; Hendon *et al.*, 2001).

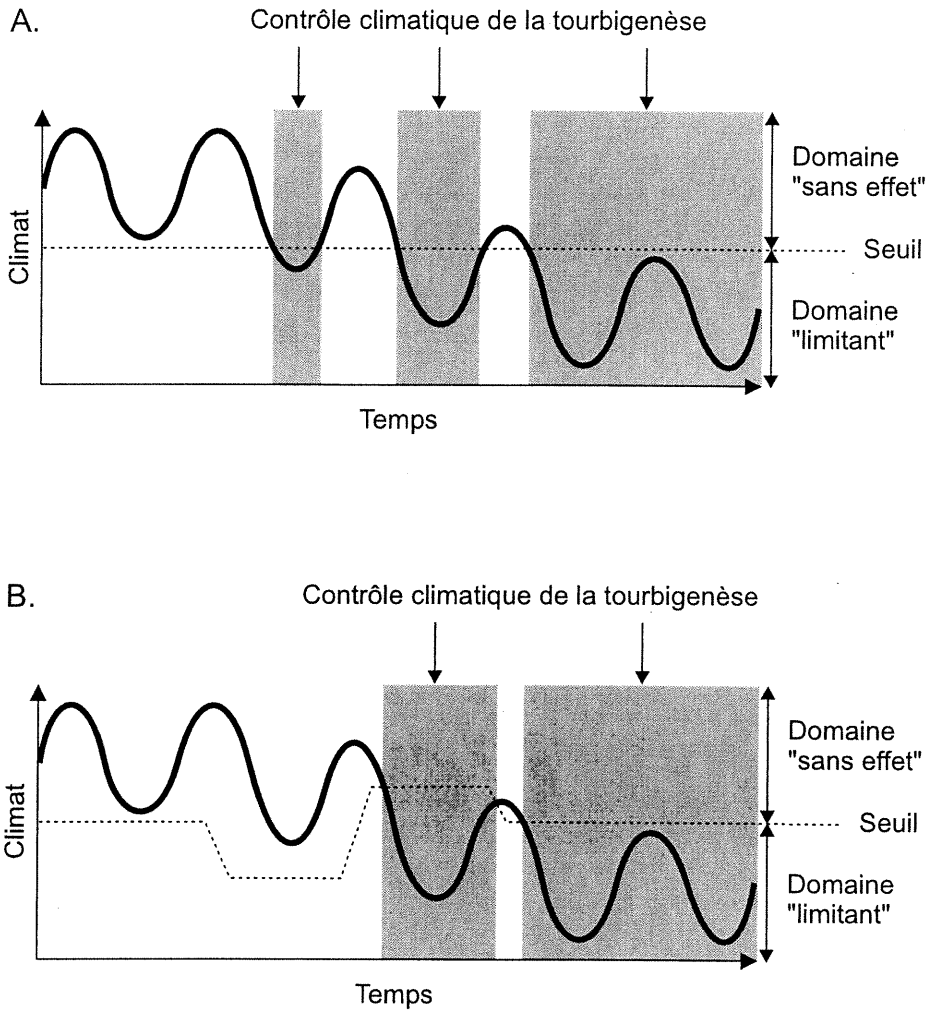

FIGURE 57. Phénomène de seuil dans l'influence du climat sur la tourbigenèse A. Seuil constant ; B. Seuil variable.

La difficulté pour vérifier cette hypothèse réside dans la mise en évidence de l'effet de seuil, et des mécanismes qui le détermine. On peut concevoir que des seuils indépendants existent pour chaque paramètre climatique, et que leurs effets puissent se succéder, voire de se cumuler. Il semble en outre probable que ces seuils varient dans le temps, en fonction des interrelations dynamiques des différents paramètres déterminants (Fig. 57B). Ils traduiraient alors une certaine sensibilité des tourbières vis-à-vis des variations climatiques, laquelle serait dépendante de l'ensemble des conditions de leur développement : végétation, statut trophique, hydrologie, nature du substrat ...

Perspectives

La reconstitution de l'accumulation du carbone dans la tourbière de Mirabel (Muller *et al.*, soumis-b, soumis-c) vient s'ajouter aux données obtenues pour les tourbières de Frontenac (Lavoie & Richard, 2000a), de Malbaie (Lavoie & Richard, 2000b) et de Mer Bleue (Richard, 2001 ; P.J.H. Richard, données non publiées). L'ensemble de ces résultats permet de dresser un bilan de l'accumulation du carbone à l'échelle du Québec méridional, qui devrait apporter des éléments de réponse aux questions laissées en suspens. Notamment, la prise en compte de la variabilité géographique au sein de la région concernée est susceptible de renseigner sur l'influence passée du climat sous un angle complémentaire de celui de cette thèse. Toutefois, la première analyse proposée par Richard (2001) suggère que quatre tourbières ne sont peut-être pas suffisantes pour mettre en évidence d'éventuels patrons spatiaux. La confrontations des données obtenues pour ces quatre sites avec les nombreuses autres tourbières étudiées dans un but paléophytogéographique par P.J.H. Richard pourrait peut-être permettre d'obtenir une idée plus générale et fiable de l'accumulation de la tourbe dans le sud du Québec. Ces études polliniques antérieures ne sont généralement basées que sur un seul profil, mais elles constituent cependant une base de données précieuse quant à la dynamique verticale de la tourbigenèse en de nombreuses localités.

Les résultats obtenus dans le cadre de cette thèse devraient en outre permettre de tester et, le cas échéant, de modifier des modèles récents d'accumulation de tourbe (Hilbert *et al.*, 2000 ; Frolking *et al.*, 2001a, 2001b). Il sera particulièrement intéressant de voir comment ces modèles parviendront à reconstituer une accumulation indépendante du climat. Un autre obstacle à l'application de ces modèles réside dans la non-conformité de l'accumulation de la tourbe à Mirabel vis-à-vis du modèle de développement autogène (Clymo, 1978, 1984).

Les courbes âge-profondeur de la tourbière de Mirabel présentent en effet des profils convexes, en contradiction avec les prédictions du modèle de Clymo (Belyea & Warner, 1995). Enfin, les modélisations pourront également être utilisées pour tester l'hypothèse d'un effet de seuil dans l'influence du climat, en comparant des simulations réalisées pour plusieurs tourbières dans des contextes climatiques différents.

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Annexe 1 : Inventaire floristique de la tourbière de Mirabel

La récolte des données floristiques et des spécimens d'herbier a couvert une période de deux ans (octobre 1998 - octobre 2000). Les spécimens, notés SP dans la liste ci-dessous, sont conservés dans mon herbier personnel. Les déterminations de plusieurs lichens et bryophytes furent vérifiées par Claude Roy et Robert Gauthier, herbier Louis-Marie, Université Laval, Sainte-Foy.

LICHENS

Cladoniaceae

Cladina mitis (Sandst.) Hale & Culb. – Carrs
Cladina rangiferina (L.) Harm. – Carrs et dépressions sèches de la zone centrale (SP)
Cladina stellaris (Opiz) Brodo – Carr ouest
Cladonia cenotea (L.) Fr. – Dépressions sèches de la zone centrale (SP)
Cladonia cristatella (Ach.) Schaer. – Carrs et dépressions sèches de la zone centrale (SP)
Cladonia fimbriata (Ach.) Schaer. – Dépressions sèches de la zone centrale (SP)
Cladonia phyllophora (Ehrh.) Hoffm. – Bordure sud de la zone centrale (SP)

BRYOPHYTES

Calopogejaceae

Calopogeja neesiana (Mass. & Carrest.) K. Müll. – Zone centrale et zone est non boisée (SP)

Lophocoleaceae

Lophocolea heterophylla (Schrad.) Dum. – Zone est non boisée (SP)

Cephaloziaceae

Cephalozia connivens (Dicks.) Lindb. – Zone centrale et zone est non boisée (SP)

Ptilidiaceae

Ptilidium pulcherrimum (G.Web.) Hampe – Zone centrale, une seule collecte (SP)

Sphagnaceae

Sphagnum angustifolium (C.Jens. ex Russ.) C.Jens. *in* Tolf. – Zone centrale et carr nord (SP)
Sphagnum capillifolium (Ehrh.) Hedw. – Carrs (SP)
    = *S. nemoreum* Scop.
Sphagnum cuspidatum Erhr. ex Hoffm. – Dépressions et anciens canaux de drainage (SP)
Sphagnum fallax (Klinggr.) Klinggr. – Dépressions du carr nord (SP)
Sphagnum fimbriatum Wils. – Carrs nord et ouest, en colonies éparses (SP)

Sphagnum fuscum (Schimp.) Klinggr. – Buttes isolées sur l'ensemble de la tourbière (SP)
Sphagnum magellanicum Brid. – Zone bombée et carr nord (SP)
Sphagnum papillosum Lindb. – Carr nord, en importantes colonies (SP)
Sphagnum rubellum Wils. – Espèce dominante dans la partie centrale (SP)
Sphagnum russowii Warnstorf – Carr nord (SP)

Dicranaceae
Dicranum polysetum Sw. – Carrs (SP)

Tetraphidaceae
Tetraphis pellucida Hedw. – Zone est non boisée (SP)

Bryaceae
Pohlia nutans (Hedw.) Lindb. – Sur l'ensemble de la tourbière (SP)

Aulacomniaceae
Aulacomnium palustre (Hedw.) Schwaegr. – Sur l'ensemble de la tourbière (SP)

Amblystegiaceae
Hamatocaulis palustre (Hedw.) Schwaegr. – Dépressions humides et canaux de drainage (SP)
= *Drepanocladus vernicosus* (Mitt.) Warnst.

Brachytheciaceae
Brachythecium salebrosum (Web. & Mohr) BSG – Zone est non boisée (SP)

Hylocomiaceae
Pleurozium schreberi (Brid.) Mitt. – En colonies importantes dans tous les carrs (SP)

Hypnaceae
Hypnum pallescens (Hedw.) P.Beauv. – Carr sud (SP)

Polytrichaceae
Polytrichum strictum Brid. – Très abondant sur l'ensemble de la tourbière (SP)

PTÉRIDOPHYTES

Lycopodiaceae
Lycopodium annotinum L. – Carrs
Lycopodium clavatum L. – Carrs (SP)
Lycopodium obscurum L. – Carrs (SP)

Osmundaceae
Osmunda cinnamomea L. – Érablières d'érables rouges périphériques
Osmunda claytoniana L. – Érablières d'érables rouges périphériques
Osmunda regalis L. – Érablières d'érables rouges périphériques

Blechnaceae

Woodwardia virginica (L.) Sm. – Une colonie (10 individus dénombrés) dans l'érablière d'érables rouges du carr ouest (SP)

Dennstaedtiaceae

Pteridium aquilinum (L.) Kuhn – Plusieurs colonies près du canal de drainage nord

Dryopteridaceae

Dryopteris intermedia (Muhl.) A.Gray – Carrs (SP)
Onoclea sensibilis L. – Érablières d'érables rouges périphériques, carrs sud et ouest

SPERMAPHYTES GYMNOSPERMES

Pinaceae

Abies balsamea (L.) Mill. – Très abondant en périphérie, individus dispersés dans les carrs (SP)
Larix laricina (Du Roi) Koch – Carrs et zone centrale
Picea glauca (Moench) Voss. – Surtout abondant dans le carr ouvert nord-ouest (SP)
Picea mariana (Mill.) BSP – Carrs et zone centrale
Pinus banksiana Lamb. – Quelques individus dans les carrs nord et nord-ouest
    = *P. divaricata* (Ait.) Dumont
Pinus strobus L. – Relativement abondant dans tous les carrs
Pinus sylvestris L. – Quelques individus dans les carrs nord et nord-ouest (SP)
Tsuga canadensis L. – Cinq individus rencontrés dans le carr nord (SP)

SPERMAPHYTES ANGIOSPERMES

MONOCOTYLÉDONES

Liliaceae

Clintonia borealis (Ait.) Raf. – Érablière d'érables rouges périphérique, limite ouest
Maianthemum canadense Desf. – Érablières d'érables rouges périphériques
Smilacina trifolia (L.) Desf. – Zone centrale et carrs
Streptopus roseus Michaux – Érablière d'érables rouges périphérique, limite ouest (SP)
Trillium erectum L., Liliaceae – Érablière d'érables rouges périphérique, zone ouest

Iridaceae

Iris versicolor L. – Érablière d'érables rouges périphérique, zone ouest

Cyperaceae

Carex disperma Dew. – Carr sud (SP)
Carex oligosperma Michaux – Dépressions dans les carrs sud et nord (SP)
Carex pauciflora Lightf. – Carr nord (SP)

Carex paupercula Michaux – Carr nord (SP)
Carex trisperma Dew. – Abondant dans la zone centrale et le carr nord (SP)
Eriophorum vaginatum L. subsp. *spissum* (Fern.) E. Hultén – Abondant dans la zone centrale et dominant dans les zones érodées (SP)
= *E. spissum* Fern.
Eriophorum virginicum L. – Zone centrale et dépressions humides du carr nord

Poaceae
Glyceria canadensis (Michaux) Trin. – Bordure des canaux de drainage, carrs nord et est
Leersia oryzoides (L.) Sw. – Bordure de canal de drainage, limite nord

Orchidaceae
Calopogon tuberosus (L.) BSP – Carr nord
= *Calopogon pulchellus* (Salisb.) R.Br.
Cypripedium acaule Ait. – Abondant dans les carrs nord et ouest
Platanthera blephariglottis (Willd.) Lindley – Abondant dans le carr nord
= *Habenaria blephariglottis* (Willd.) Hook.

Lemnaceae
Lemna minor L. – Canaux de drainage

Sparganiaceae
Sparganium emersum Rehmann – Canaux de drainage nord et ouest
= *S. chlorocarpum* Rydb.

Typhaceae
Typha latifolia L. – Partout en périphérie de la tourbière
Typha angustifolia L. – Partout en périphérie de la tourbière

DICOTYLÉDONES

Betulaceae
Alnus incana (L.) Moench subsp. *rugosa* (Du Roi) Clausen – Bordure des canaux de drainage, limites nord et est
Betula populifolia Marsh. – Très abondant : zone centrale et carrs (SP)
Corylus cornuta Marsh. – Érablière d'érables rouges périphérique, limite ouest (SP)

Salicaceae
Salix discolor Mühl. – Érablière d'érables rouges périphérique, limite sud
Salix petiolaris J.E. Smith. – Partout en périphérie de la tourbière (SP)
Salix rigida Mühl. – Érablière d'érables rouges périphérique, limite sud (SP)

Ulmaceae
Ulmus americana L. – Érablière d'érables rouges périphérique, limite est

Ranunculaceae
Coptis groenlandica (Oeder) Fern. – Carrs

Sarraceniaceae
Sarracenia purpurea L. – Individus isolés dans la zone centrale et le carr nord

Violaceae
Viola selkirkii Pursh – Érablière d'érables rouges périphérique (carr ouest)

Rosaceae
Aronia melanocarpa (Michaux) Ell. – Arbustaies denses dans les carrs et en ceinture de la zone centrale (SP)
Prunus serotina Ehrh. – Carr sud (SP)
Rubus idaeus L. – Érablières d'érables rouges périphériques, limites ouest et sud
Spiraea latifolia (Ait.) Borkh. – Érablières d'érables rouges périphériques
Spiraea tomentosa L. – Érablières d'érables rouges périphériques

Lythraceae
Lythrum salicaria L. – Bordures des canaux de drainage et plate-forme de remblais

Oxalidaceae
Oxalis stricta L. – Érablière d'érables rouges périphérique, limite ouest

Aceraceae
Acer rubrum L. – Dominant dans les Érablières d'érables rouges périphériques (SP)

Aquifoliaceae
Nemopanthus mucronatus (L.) Trel. – Arbustaies denses dans les carrs et en ceinture de la zone centrale

Cornaceae
Cornus alternifolia L.f. – Érablière d'érables rouges périphérique, limite est (SP)
Cornus canadensis L. – Érablière d'érables rouges périphérique, limite ouest (SP)
Cornus stolonifera Michaux – Érablière d'érables rouges périphérique, limite ouest

Araliaceae
Aralia nudicaulis L. – Érablière d'érables rouges périphérique, carr ouest

Primulaceae
Trientalis borealis Raf. – Carrs

Pyrolaceae
Monotropa uniflora L. – Relativement abondant dans les carrs, surtout en pessière (SP)

Ericaceae
Andromeda glaucophylla Link. – En petites colonies dans toute la tourbière (SP)
Chamaedaphne calyculata (L.) Moench – Dominant dans la zone centrale (SP)
= *Cassandra calyculata* (L.) D.Don
Gaultheria hispidula (L.) Mühl. – Quelques colonies sous pessières dans le carr nord (SP)
Gaultheria procumbens L. – En colonies importantes sur l'ensemble de la tourbière (SP)
Kalmia angustifolia L. – Abondant dans la zone centrale et dominant dans les carrs (SP)

Kalmia polifolia Wang. – Peu abondant mais présent dans toute la tourbière (SP)
Rhododendron canadense (L.) Torr. – Nombreuses colonies en ceinture de la zone centrale et dans les carrs (SP)
Rhododendron groenlandicum (Oeder) Kron & Judd – Abondant dans toute la tourbière (SP)
  = *Ledum groenlandicum* Retz.
Vaccinium angustifolium Ait. – Abondant dans toute la tourbière (SP)
Vaccinium myrtilloides Michaux – Abondant dans toute la tourbière (SP)
Vaccinium oxycoccos L. – En colonies importantes dans toute la tourbière (SP)
  = *Oxycoccus microcarpus* Turcz.

Scrofulariaceae
Chelone glabra L. – Bordure de canal de drainage, limite ouest

Lamiaceae
Lycopus uniflorus Michaux, Lamiaceae – Bordure de canal de drainage, limite nord

Rubiaceae
Galium palustre L. – Bordures des canaux de drainage et de la route, limite sud

Caprifoliaceae
Viburnum cassinoides L. – Abondant sur l'ensemble de la tourbière (SP)

Asteraceae
Bidens cernua L. – Bordure des canaux de drainage (SP)
Eupatorium perfoliatum L. – Bordure de canal de drainage, limite ouest, et bordure de la route, limite sud

Espèces trouvées dans les zones perturbées en bordure de la tourbière (liste non exhaustive)

Acer saccharum Marsh., Aceraceae – Jeunes plants, érablière d'érables rouges périphérique, limite ouest
Agropyron repens (L.) Beauv., Poaceae – Plate-forme de remblais
Agrostis scabra L., Poaceae – Plate-forme de remblais (SP)
Alisma triviale Pursh, Alismataceae – Plate-forme de remblais
Bromus inermis Leyss., Poaceae – Bordure de la route, limite sud
Carex blanda Dewey, Cyperaceae – Bordure de la route, limite sud
Carex crinita Lam., Cyperaceae – Plate-forme de remblais (SP)
Carex lurida Wahl., Cyperaceae – Bordure de la route, limite sud
Carex tenera Dewey, Cyperaceae – Bordure de la route, limite sud
Cinna arundinacea L., Poaceae – Plate-forme de remblais (SP)

Cerastium holesteoides Fries (= *C. vulgatum* L.), Caryophyllaceae – Bordure de la route, limite sud (SP)
Chenopodium album L., Chenopodiaceae – Plate-forme de remblais
Chrysanthemum leucanthemum L., Asteraceae – Bordure de la route, limite sud
Clematis virginiana L., Ranunculaceae – Bordure de la route, limite sud
Dactylis glomerata L., Poaceae – Bordure de la route, limite sud
Eleocharis smallii Britton (= *E. palustris*), Cyperaceae – Bordure de la route, limite sud
Equisetum arvense L., Equisetaceae – Plate-forme de remblais et bordure de la route, limite sud
Erigeron philadelphicus L., Asteraceae – Bordure de la route, limite sud
Fragaria vesca L. subsp. *americana* (Porter) Staudt (= *F. americana* (Porter) Britton), Rosaceae – Bordure de la route, limite sud
Hemerocallis fulva L., Liliaceae – Bordure de la route, limite sud
Hordeum jubatum L., Poaceae – Bordure de la route, limite sud
Hypericum perforatum L., Hypericaceae – Bordure de la route, limite sud
Juncus effusus L., Juncaceae – Bordure de la route, limite sud
Lotus corniculatus L., Fabaceae – Plate-forme de remblais
Medicago lupulina L., Fabaceae – Bordure de la route, limite sud
Melilotus albus Med., Fabaceae – Plate-forme de remblais
Œnothera biennis L. (= *Œ. pumila* L.), Onagraceae – Bordure de la route, limite sud (SP)
Pastinaca sativa L., Apiaceae – Plate-forme de remblais
Phleum pratense L., Poaceae – Plate-forme de remblais
Phragmites australis (Cav.) Trin. ex Steudel, Poaceae – Bordures de la route, limite sud et plate-forme de remblais
Plantago major L., Plantaginaceae – Bordure de la route, limite sud
Platanthera hyperborea (L.) Lindley, Orchidaceae – Bordure de la route, limite sud
Poa pratensis L., Poaceae – Bordure de la route, limite sud et plate-forme de remblais (SP)
Populus balsamifera L., Salicaceae – Bordure de la route, limite sud
Populus grandidentata Michaux, Salicaceae – Bordure de la route, limite sud
Populus tremuloides Michaux, Salicaceae – Bordures de la route, limite sud et plate-forme de remblais
Potentilla anserina L., Rosaceae – Bordure de la route, limite sud
Potentilla argentea L., Rosaceae – Bordure de la route, limite sud (SP)
Prunella vulgaris L., Lamiaceae – Bordure de la route, limite sud
Puccinellia distans (L.) Parl. – Plate-forme de remblais (SP)
Ranunculus acris L., Ranunculaceae – Bordure de la route, limite sud et plate-forme de remblais
Rudbeckia hirta L., Asteraceae – Friches, limite nord
Schoenoplectus tabernaemontani (C.G. Gmelin) Palla (= *S. validus* Vahl.), Cyperaceae – Plate-forme de remblais (SP)
Scirpus cyperinus (L.) Kunth., Cyperaceae – Bordure de la plate-forme de remblais
Sisyrinchium angustifolium Miller, Iridaceae – Bordure de la route, limite sud (SP)
Solidago canadensis L., Asteraceae – Nombreux endroits en friches en périphérie de la tourbière
Stellaria graminea L., Caryophyllaceae – Bordure de la route, limite sud (SP)
Spergularia rubra (L.) J. & C. Pesl., Caryophyllaceae – Plate-forme de remblais (SP)

Suaeda maritima (L.) Dumort., Chenopodiaceae – Plate-forme de remblais (SP)
Thlaspi arvense L., Brassicaceae – Plate-forme de remblais
Trifolium pratense L., Fabaceae – Bordure de la route, limite sud
Vicia cracca L., Fabaceae – Bordure de la route, limite sud

Annexe 2: Autorisation de l'éditeur pour l'article Muller & Richard (accepté)

Date / Date : 14 Août / August 2001

Nom de l'étudiant / Student name : Serge D. Muller

Grade postulé / Postulated grade : Ph.D.

Programme / Program : Géographie

Nom de la revue / Journal name : Journal of Biogeography

Adresse / Address : Blackwell Science Ltd, 25 John Street, London WC1N 2BS, UK

Nom de l'éditeur / editor's name : Philip Stott

Identification de l'article / Paper identification :

Muller, S.D. & Richard, P.J.H., accepted 2001. Postglacial plant migration in the St. Lawrence lowlands, southern Québec. *Journal of Biogeography.*

L'étudiant Serge D. Muller est autorisé à inclure l'article ci dessus dans sa thèse de doctorat qui a pour titre : « Conditions de l'accumulation du carbone dans une tourbière du Québec méridional : l'influence des facteurs autogènes et des contrôles allogènes ».

The student Serge D. Muller is authorised to include the article mentioned above in his Ph.D. thesis which name is : "Conditions of carbon accumulation in a southern Québec bog: the influence of autogenic factors and allogenic controls".

Professor Philip STOTT	[redacted]	08/14/01
Editor	Signature	Date

Annexe 3 : Autorisation des coauteurs des articles

Article 1 : Muller, S.D. & Richard, P.J.H., accepté. Postglacial plant migration in conterminous Montréal lowlands, southern Québec. *Journal of Biogeography.*

Nom de l'étudiant : Serge D. Muller
Grade postulé : Ph.D.
Programme : Géographie

A titre de coauteur de l'article identifié, je suis d'accord pour que Serge D. Muller inclut cet article dans sa thèse de doctorat qui a pour titre « Conditions de l'accumulation du carbone dans une tourbière du Québec méridional : l'influence des facteurs autogènes et des contrôles allogènes »,

Pierre J.H. Richard [redacted] 20/09/01
Date

Article 2 : Muller, S.D., Guiot, J., Richard, P.J.H., Beaulieu, J.-L. de & Fortin, D., soumis-a. Postglacial climate dynamics in the St. Lawrence lowlands, southern Québec: pollen and lake-level evidence. *Palaeogeography, Palaeoclimatology, Palaeoecology*.

Nom de l'étudiant : Serge D. Muller
Grade postulé : Ph.D.
Programme : Géographie

A titre de coauteur de l'article identifié, je suis d'accord pour que Serge D. Muller inclut cet article dans sa thèse de doctorat qui a pour titre « Conditions de l'accumulation du carbone dans une tourbière du Québec méridional : l'influence des facteurs autogènes et des contrôles allogènes ».

Joël Guiot [redacted] 13/09/01
Signature Date

Pierre J.H. Richard [redacted] ....20/09/01.....
Signature Date

Jacques-Louis de Beaulieu ..[redacted] 13/09/01
Signature Date

David Fortin ..[redacted] ....20/09/01....
Signature Date

Article 3 : Muller, S.D., Richard, P.J.H. & Larouche, A.C., soumis-b. Postglacial development of a southern Québec bog: a spatio-temporal reconstruction based on pachymetry, sedimentology, micro- and macrofossils. *The Holocene.*

Nom de l'étudiant : Serge D. Muller
Grade postulé : Ph.D.
Programme : Géographie

A titre de coauteur de l'article identifié, je suis d'accord pour que Serge D. Muller inclut cet article dans sa thèse de doctorat qui a pour titre « Conditions de l'accumulation du carbone dans une tourbière du Québec méridional : l'influence des facteurs autogènes et des contrôles allogènes »,

Pierre J.H. Richard [redacted] 20/09/01
Signature Date

Alayn C. Larouche [redacted] 20/9/01
Signature Date

Article 4 : Muller, S.D., Richard, P.J.H. & Beaulieu, J.-L. de, soumis-c. Postglacial carbon accumulation in a boreal bog (southern Québec). *Quaternary Research.*

Nom de l'étudiant : Serge D. Muller
Grade postulé : Ph.D.
Programme : Géographie

A titre de coauteur de l'article identifié, je suis d'accord pour que Serge D. Muller inclut cet article dans sa thèse de doctorat qui a pour titre « Conditions de l'accumulation du carbone dans une tourbière du Québec méridional : l'influence des facteurs autogènes et des contrôles allogènes »,

Pierre J.H. Richard [redacted] 20/09/01
Date

Jacques-Louis de Beaulieu .... [redacted] 13 09 01
Date