UNIVERSITÉ DU QUÉBEC À MONTRÉAL

BILAN HYDRIQUE ET APPORT D'EAU SOUTERRAINE AUX TOURBIÈRES DU MOYEN-NORD QUÉBECOIS (CANADA) DANS LES CONDITIONS ACTUELLES ET EN CONDITIONS DE CHANGEMENTS CLIMATIQUES.

MÉMOIRE

PRÉSENTÉ

COMME EXIGENCE PARTIELLE

DE LA MAÎTRISE EN SCIENCE DE LA TERRE

PAR

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AVRIL 2022

REMERCIEMENTS

Dans un premier temps, je voudrais saluer le dévouement de l'ensemble des membres de l'équipe d'hydrogéologie du département des sciences de l'UQAM ainsi que son réseau de soutien aux multiples ramifications où j'aurai trouvé camaraderie, support et réconfort.

Je tiens à remercier ma directrice de recherche Marie Larocque, véritable pilier du projet, qui a su m'accompagner et m'orienter avec énormément de générosité et d'humanité, et ce malgré un horaire pour le moins vertigineux. J'ai ainsi eu la chance de bénéficier grandement de ses critiques constructives, ce qui m'a permis de cheminer professionnellement et personnellement. J'aurai intégré ton équipe avec fierté et je la quitte avec d'autant plus d'admiration pour tes réalisations.

Merci également aux agents de recherche Sylvain Gagné et Marjolaine Roux pour leur constance et leur regard éclairant face aux défis qui se sont présentés sur ma route. Je salue particulièrement le courage de Marjolaine qui a su m'initier de main de maître au travail de terrain par son expertise et ses bons conseils, mais aussi par sa présence et son sens de l'humour bourgeonnant. Je ne peux aussi m'imaginer cette aventure nordique sans le travail acharné et le dévouement de mes assistants de terrain, merci à Tiffany Audet, Béatrice Lefebvre Wolff, Jérémie Maillé-Bizier et Samuel Bolduc, m'ayant tous épaulée dans ce périple. Votre présence, votre ténacité et votre vivacité d'esprit m'auront permis de respecter mes nombreux d'échéanciers, et ce malgré les conditions climatiques dont la clémence fût très timide. À Tiffany Audet, qui a généreusement décidé de revenir affronter la campagne de terrain du mois de juillet, qui fut pour le moins éprouvante, un merci particulier pour ta ferveur imperturbable et ton professionnalisme sans faille.

Merci à ma conjointe pour son soutien inconditionnel durant toutes ces années ainsi qu'à ma famille et mes amis qui ont su m'écouter, m'encourager et me donner la force d'avancer lors des moments plus difficiles.

Merci également aux membres de l'équipe de recherche élargie composée de Michelle Garneau, Éric Rosa, Marc-André Bourgault et Paul Morris, qui ont apporté un regard multidisciplinaire au projet. À Michelle, je tiens à dire un merci particulier pour son exceptionnelle générosité et sa confiance.

Je tiens finalement à saluer et souligner la présence de tous ceux qui ont contribué à mon obtention de différentes bourses scolaires et de recherche, dont celle du CRSNG, ce qui m’aura permis de me consacrer pleinement à mon travail avec une paix d’esprit certaine. Enfin, je tiens à remercier les nombreux partenaires du projet Nemaska Lithium ainsi que la Mine Renard de Stornoway, où nous avons été accueillis comme des membres de leur grande famille et pris en charge de façon remarquable par Yan Bhérer, Maryse Godin et Anissa Amri.

AVANT-PROPOS

Ce projet est rendu possible grâce au financement de Nemaska Lithium, Stornoway et du CRSNG (subvention RDC J 513542-17). Nous remercions également les professionnels de la mine Renard ayant participé activement à l'organisation logistique des campagnes de terrain.

Ces travaux s'inscrivent dans une démarche multidisciplinaire dont l'objectif est d'identifier les principaux processus ayant influencé le déséquilibre écohydrologique associé aux fens pauvres boréaux. L'approche fut subdivisée en quatre volets distincts (1) reconstruire les conditions paléohydrologiques et paléoécologiques ayant influencé l'accumulation de tourbe et la dynamique du carbone, (2) reconstruire la végétation régionale et les variations climatiques associées durant l'Holocène, (3) modéliser la dynamique hydrologique ainsi que les apports d'eau souterraine actuels et en situation de changement climatique dans les bassins versants respectifs (constituant l'objet du présent mémoire) et (4) simuler l'effet des différents forçages météorologiques des dernières années sur le fonctionnement écohydrologique des tourbières des dernières années.

Ce mémoire est présenté sous la forme d'un article qui sera soumis dans un numéro spécial sur les écosystèmes dépendants des eaux souterraines de la revue Frontiers in Earth Sciences.

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RÉSUMÉ

Les contextes géologiques, topographiques, hydrologiques et climatiques dans lesquels se produisent les apports d'eau souterraine dans les tourbières du Moyen-Nord québécois et l'importance de ces apports dans le bilan hydrique de ces tourbières sont encore très peu étudiés. Les deux tourbières étudiées dans ce projet (Misask et Cheinu) sont considérées représentatives des écosystèmes de cette région du Québec. Il s'agit de tourbières structurées de type minérotrophe pauvre montrant une récente ombrotrophisation dans leurs zones centrales profondes. Elles se sont développées dans les dépressions topographiques drumlinoïdes et morainiques du till sablo-silteux composant le substrat géologique régional. Les objectifs de cette recherche étaient de 1) quantifier le bilan hydrique actuel de deux tourbières du Moyen-Nord du Québec, 2) déterminer si l'afflux d'eau souterraine est une contribution importante à leur bilan hydrique et documenter leur sensibilité aux changements de recharge.

Les campagnes de terrain (essais hydrauliques *in situ*, carottage de tourbe, levés topographiques détaillés, suivis horaires des niveaux de nappe, mesures de débits) ont permis de caractériser la géomorphologie du bassin tourbeux, les propriétés hydriques des dépôts organiques et des sédiments composant l'aquifère superficiel environnant, ainsi que la dynamique hydrologique de la nappe phréatique leur étant associée. Un modèle numérique tridimensionnel (MODFLOW) en régime permanent simulant adéquatement les conditions hydrogéologiques actuelles a permis de quantifier le bilan hydrique de chaque tourbière. Une analyse de sensibilité des niveaux et des flux à la recharge a fourni des données permettant d'estimer la vulnérabilité des deux tourbières aux changements climatiques.

Au site de la tourbière Misask, les flux entrants sont dominés par la recharge (78 %) tandis que la contribution de l'eau souterraine n'est que de 22%. Les flux sortants de la tourbière sont évacués principalement par ruissellement de surface (85 %) et dans une moindre mesure par une contribution de la tourbière à l'aquifère superficiel (15 %). Le site de la tourbière Cheinu montre un bilan hydrique dont les flux entrants sont dominés par les apports en eau souterraine (56 %) et, dans une proportion presque égale, par la recharge (44 %). Ces débits sont exfiltrés principalement par le ruissellement superficiel (74 %), et par des flux souterrains de la tourbière vers l'aquifère (26 %). Enfin, les modèles numériques furent soumis à des changements de recharge (-50%, -20% et +20%) afin de simuler les effets hydrogéologiques potentiels des changements climatiques. Dans les deux sites, les variations de niveau de nappe moyenne sont plus faibles dans les dépôts organiques que dans l'aquifère environnant. Les charges hydrauliques moyennes ont varié de -1 à +0,3 cm pour la tourbière Misask et de -9 à +2 cm pour la tourbière Cheinu. De plus, sur le site Misask, les échanges de flux avec l'aquifère ont montré une sensibilité limitée à la recharge (-50, -20 et +20%) avec des variations de -26, -8 et +7% des flux provenant de l'aquifère et de -27, -9 et +7 % pour les flux exfiltrés de la tourbière. Le ruissellement de surface montre une plus grande sensibilité aux changements de recharge, avec une variation de -48, -19 et +19% respectivement pour les trois scénarios. Sur le site de Cheinu, les scénarios de recharge ont généré des changements de flux d'eau souterraine plus importants qu'au site Misask. Ceux-ci ont engendré des variations de -33, -12 % et +11 % des flux entrants et de -44, -17 et +17 % des flux sortants dans la tourbière. Le ruissellement de surface a montré des variations de -39, -15 et +15%.

Les scénarios étudiés ayant engendré des effets différents sur les deux sites, montrent que la sensibilité des tourbières aux changements climatiques est complexe et provient des facteurs intrinsèques distinguant les deux tourbières étudiées (morphologie du bassin tourbeux, contribution des échanges aquifère-tourbière au bilan hydrique, gradient topographique). Dans le cadre de futurs travaux, l'identification de ces facteurs pourrait permettre de faciliter la compréhension et l'estimation de leur vulnérabilité grâce à des méthodes de classification applicable à l'échelle régionale.

Mots clés : Eau souterraine, bilan hydrique, tourbière, Moyen-Nord québécois, modélisation hydrogéologique, MODFLOW, régime permanent, changement climatique

CHAPITRE I

INTRODUCTION GÉNÉRALE

1.1 Problématique générale

Les écosystèmes tourbeux couvrent approximativement 12 à 17% du territoire canadien (Letts *et al.*, 2000; Payette et Rochefort, 2001; Lafleur *et al.*, 2005) et de 9 à 12% du territoire québécois (Gorham *et al.*, 1998; Payette et Rochefort, 2001), avec une forte prévalence en zone boréale. Dans le cadre de l'étude des milieux humides, l'aqualyse se caractérise par l'expansion et la coalescence des mares causant la dégradation de la végétation qui compose leur structure microtopographique ainsi qu'une mortalité significative des espèces ligneuses (Payette, 2008; Dissanska *et al.*, 2009; Proulx-McInnis *et al.*, 2013). Ce phénomène de coalescence, affectant les tourbières nordiques (Glaser et Janssens, 1986), est documenté dans la région du Moyen-Nord du Québec (Robitaille *et al.*, 2021) définit comme la zone biogéographiquement subarctique. Puisque les tourbières nordiques constituent des puits significatifs de carbone (Charman, 2002; Holden, 2005a; Rydin et Jeglum, 2006; Daniels *et al.*, 2008), l'aqualyse pourrait être une manifestation de changements importants dans la dynamique de séquestration et d'émission de CO_2 et de CH_4 ainsi que dans les fonctions hydroécologiques des tourbières nordiques (Trudeau *et al.*, 2013).

Dans plusieurs régions de l'hémisphère Nord, l'écotone délimitant les biomes de la forêt boréale ouverte (taïga) et fermée, représente également la limite septentrionale

de la prévalence des tourbières ombrotrophes (Damman, 1979; Seppälä et Koutaniemi, 1985; Foster et Fritz, 1987; Payette et Rochefort, 2001; Rydin et Jeglum, 2013). Celles-ci sont ensuite progressivement remplacées par des fens réticulés oligotrophes dont l'abondance de formes aquatiques augmente progressivement vers le nord (Payette et Rochefort, 2001). Cette dominance bien documentée des tourbières minérotrophes en région nordique résulterait d'un gradient de diminution des précipitations et de l'écourtement de la période de production végétale qui ne seraient pas favorables au développement des conditions ombrotrophes supportées par une croissance importante des sphaignes (Batzer et Baldwin, 2012).
Les conditions hydrologiques et climatiques des derniers siècles reconstituées grâce aux indicateurs paléohydrologiques (Robitaille *et al.*, 2021) montrent que les tourbières du Moyen-Nord initialement minérotrophes ont évolué vers des conditions plus humides (aqualyse) et qu'elles sont depuis quelques décennies sujettes à une ombrotrophisation de leur zone centrale. La documentation de ce changement de régime écologique suggère une migration nordique de l'étendue géographique de la dominance ombrotrophique de ces écosystèmes avec le réchauffement du climat (Damman, 1979), notamment alimentée par des changements de régimes hydriques.

Il est reconnu que la profondeur de nappe dans les dépôts tourbeux et la température de la tourbe constituent les deux facteurs dominant la régulation des fonctions de stockage du carbone et de l'émission de gaz à effet de serre tel que le méthane (Moore et Knowles, 1989; Dise, 1993; Bubier, 1995; Alm, 1997). Des études récentes ont démontré qu'une augmentation du niveau de nappe peut entraîner une augmentation des émissions de méthane (CH_4) (Ojanen *et al.*, 2010; Ojanen *et al.*, 2013; Korkiakoski *et al.*, 2019) et de dioxyde de carbone (CO_2). Un assèchement augmente les émissions de CO_2 mais aussi d'oxyde nitreux (N_2O) induites par l'augmentation potentielle de la décomposition de la couche de tourbe profonde dans les tourbières drainées boréales (Ojanen *et al.*, 2010; Ojanen *et al.*, 2013; Ojanen et Minkkinen, 2019).

L'importance de comprendre le bilan hydrique des tourbières nordiques est le fondement de la capacité à anticiper les effets pondérés des changements climatiques sur chaque composante afin d'en évaluer l'impact sur la pérennité de ses écosystèmes. Les travaux se penchant sur la connectivité des tourbières à l'aquifère régional ont étudié les échanges aquifères-tourbières à l'aide d'indicateurs de végétation (Wassen *et al.*, 1988; Goslee *et al.*, 1997; Lewis, 2012; House *et al.*, 2015; Larocque *et al.*, 2016), des analyses de données piézométriques (Stein *et al.*, 2004; Ferlatte *et al.*, 2015), de l'hydrogéochimie (Adamus *et al.*, 1991; Kehew *et al.*, 1998; Auterives, 2006; Kerr *et al.*, 2008; Ferlatte, 2013), de différents isotopes (Corbett *et al.*, 1997; Hunt *et al.*, 1998; Hogan *et al.*, 2000; Cook *et al.*, 2008; Rodellas *et al.*, 2012; Berthot *et al.*, 2016; Hare *et al.*, 2017; Isokangas *et al.*, 2017) et de modèles numériques (Ronkanen et Kløve, 2008; Susilo *et al.*, 2013; Bourgault *et al.*, 2014; Levison *et al.*, 2014; Rossi *et al.*, 2014; Quillet *et al.*, 2017; Taufik *et al.*, 2019). Il a été démontré que les apports d'eau souterraine peuvent représenter une contribution représentant jusqu'à 50 % d'un bilan hydrique méridional des tourbières (Bourgault *et al.*, 2014; Levison *et al.*, 2014). De plus, van Bellen *et al.* (2018) ont suggéré, à l'aide d'un modèle conceptuel, une sensibilité de ces systèmes aux variations hydroécologiques. Des changements dans les flux d'eau souterraine peuvent alors constituer un des grands moteurs des changements dans l'équilibre hydrologique des tourbières. Malgré leur importance dans le biome boréal, le bilan hydrique des tourbières nordiques reste peu documenté, recevant moins d'attention scientifique, notamment en raison de la complexité d'accès de ces régions éloignées.

Les biomes boréaux seront vulnérables et sensibles aux conditions climatiques changeantes (Diffenbaugh et Field, 2013; Price *et al.*, 2013; Gauthier *et al.*, 2014; Gauthier *et al.*, 2015), notamment en raison de l'interaction étroite entre les précipitations (P) et l'évapotranspiration potentielle (ETP). En effet, les fluctuations climatiques entrainent plus facilement une modification du taux net de précipitation (P-ETP), c.-à-d. l'estimation de la recharge des eaux souterraines (Bergengren *et al.*, 2011; Flanagan et Syed, 2011; Ireson *et al.*, 2015; Schneider *et al.*, 2016). Ces variation de recharge sont difficilement quantifiable et varient spatialement sur le territoire canadien (Scibek et Allen, 2006; Jyrkama et Sykes, 2007; Rivard *et al.*, 2009; Croteau *et al.*, 2010; Levison *et al.*, 2014; Wang *et al.*, 2014; Dubois *et al.*, 2021). Les conditions de recharge affectant par la suite la proximité nappe-topographie (Devito *et al.*, 2005; Haitjema et Mitchell-Bruker, 2005; Hokanson *et al.*, 2020) et donc indirectement les flux d'eau échangés entre les aquifères et les tourbières.

Comme tous les écosystèmes de la région nordique, les tourbières sont sensibles à ces changements climatiques. Il est essentiel de documenter la dynamique hydrologique actuelle des tourbières nordiques afin de définir l'origine des changements de régime hydrologique documentés et ainsi d'anticiper plus efficacement les impacts des changements climatiques.

1.2 État des connaissances

1.2.1 Les tourbières

Les tourbières se caractérisent par une production végétale nette excédant leur taux de décomposition (Damman, 1979; Glaser, 1987; Payette et Rochefort, 2001). Il en résulte l'accumulation progressive d'un dépôt organique qui peut atteindre plusieurs mètres d'épaisseur. Le régime hydrique de ces écosystèmes dicte leur regroupement en deux principales classes, soit les tourbières ombrotrophes alimentées principalement par des apports provenant des précipitations (faiblement minéralisée) et les tourbières minérotrophes dépendantes des afflux d'eau souterraine (plus minéralisée). Généralement, le développement des tourbières s'amorce par un système minérotrophe qui évolue vers un état ombrotrophe (Gorham et Janssens, 1992; Kuhry *et al.*, 1993). Il est toutefois important de souligner que les tourbières ombrotrophes peuvent aussi recevoir de l'eau souterraine via les dépôts minéraux sous-jacents et être connectées latéralement à l'eau souterraine par les portions minérotrophes situées en périphérie.

Deux horizons principaux peuvent être distingués dans la stratigraphie des tourbières, soit l'acrotelme et le catotelme (Ingram, 1978). L'acrotelme se compose d'organismes végétaux vivants et de matière organique fibrique créant un médium de porosité et de conductivité hydraulique élevée à travers lequel sont circonscrites les fluctuations annuelles de la nappe phréatique. Cette couche est d'une épaisseur moyenne de 35-50 cm dans les tourbières ombrotrophes (Belyea et Clymo, 2001) et de 10-20 cm dans les tourbières minérotrophes (Price et Maloney, 1994). Le catotelme, pour sa part, est composé de matière organique plus ou moins bien décomposée et saturée dont les processus de décomposition anaérobique engendrent une réduction de la porosité effective limitant parallèlement sa conductivité hydraulique (Morris *et al.*, 2011). Les variations saisonnières de la profondeur de la nappe d'eau souterraine sous-tendent une interface acrotelme-catotelme à transition graduelle (Clymo, 1992).

Les tourbières de haute latitude c.-à-d. de subboréales à arctiques, sont dominées par les bryophytes et les cypéracées (Kuhry *et al.*, 1993; Glaser *et al.*, 2004). Elles constituent des écosystèmes complexes représentant des environnements intermédiaires entre les conditions terrestres et aquatiques (Mitsch et Gosselink, 2000) et dont le développement est étroitement associé aux contextes hydrologiques et hydrogéologiques favorisant le maintien de conditions de résurgence ou de saturation superficielle.

En surface, les tourbières sont caractérisées par une microtopographie caractérisée par des buttes, des creux et des mares. Plusieurs aspects du mécanisme de développement de mares restent encore méconnus, cependant les liens avec le contexte géologique de l'aquifère sous-jacent (Comas *et al.*, 2005; Comas *et al.*, 2011), avec la morphologie du bassin tourbeux (Lowry *et al.*, 2009) ainsi que leur perpendicularité par rapport au gradient de pente principal (Foster et Fritz, 1987; Belyea et Lancaster, 2002) ont été récemment démontrés. De plus, la biogéochimie caractéristique de l'eau qui y circule est liée à la morphologie de la tourbière ainsi qu'à des facteurs saisonniers (Arsenault *et al.*, 2018). La biogéochimie est également liée à l'importance de la succession de microformes dans la biodiversité faunique (Danks et Rosenberg, 1987; Murkin et Batt, 1987; Standen *et al.*, 1999; Mazerolle, 2005; Mazerolle *et al.*, 2006) et végétale (Glaser, 1992; Fontaine *et al.*, 2007). Cette microtopographie de buttes et de creux (*hollow-hummock*), commune dans les tourbières nordiques réticulées, serait provoquée par l'alternance des espèces végétales et leur taux de décomposition spécifiques (Johnson *et al.*, 1990; Johnson et Damman, 1991; Nungesser, 2003).

1.2.2 Propriétés physiques de la tourbe

La dynamique hydrique des tourbières assure leur équilibre écologique et biogéochimique en soutenant les processus de circulation et d'emmagasinement en eau ainsi que le renouvellement et la disponibilité des éléments dissous. Les propriétés physiques de la tourbe sont des facteurs dominants de ses échanges hydriques (Päivänen, 1973; Rycroft *et al.*, 1975a, 1975b; Price et Woo, 1986; Ours *et al.*, 1997; Rezanezhad *et al.*, 2016) et régulent la redistribution de l'eau dans le bassin tourbeux (Waddington *et al.*, 2015).

La porosité totale de la tourbe peut exercer 80% (Boelter, 1968) et se compose d'interstices et de pores aux morphologies complexes (Hayward et Clymo, 1982) dont la taille et la connectivité diminuent en profondeur (Rezanezhad *et al.*, 2010). Ce gradient vertical engendré par les mécanismes de décomposition organique (Price *et* *al.*, 2005) et de compaction de la matière organique conduit à l'augmentation de sa densité ainsi qu'à la réduction de la porosité effective. En contexte de toundra arctique, Quinton *et al.* (2000) ont obtenu des valeurs de porosité effective passant de 80% en surface à moins de 50% à 50 cm de profondeur. Plusieurs études ont démontré également que la décroissance verticale de la porosité effective en profondeur ne répond pas à une progression linéaire (Rosa et Larocque, 2008; Moore *et al.*, 2015; Dettmann et Bechtold, 2016; Bourgault *et al.*, 2017). Dans les dépôts tourbeux, Price (1996) estime que la porosité effective est comparable à la capacité d'emmagasinement.

Une partie de la teneur en eau de la tourbe est encapsulée à l'intérieur de pores fermés et isolés composant sa fraction dite immobile (Caron *et al.*, 2015). Le concept de double porosité repose sur la dualité démontrée entre cette fraction immobile et la fraction mobile attribuée à la porosité effective. Cette propriété de la tourbe fait l'objet d'études plus approfondies tentant de mettre en lumière ses implications notamment dans la régulation de régime d'écoulement (Rezanezhad *et al.*, 2016), le transport d'éléments dissous (Ours *et al.*, 1997; Reeve *et al.*, 2001; Rezanezhad *et al.*, 2012) et implicitement son rôle dans la production de gaz à effet de serre et du cycle du carbone.

Quinton *et al.* (2008) ont démontré que les données relatives à la caractérisation de la géométrie des pores de la tourbe, lorsqu'intégrées à des relations mathématiques s'appuyant sur les équations de Hagen–Poiseuille et Kozeny–Carman, peuvent prédire adéquatement la conductivité hydraulique saturée (K_{sat}) des dépôts tourbeux. Le profil vertical des propriétés physiques contribue activement au maintien des niveaux de nappe viables au développement des tourbières. En effet, la conductivité hydraulique élevée de l'acrotelme facilite l'évacuation rapide des surcharges épisodiques (évènements pluvieux, crue printanière) tandis qu'en condition d'étiage, le catotelme, qui génère un régime d'écoulement lent, diminue la vitesse des échanges horizontaux afin d'en inhiber les impacts (Boelter, 1968).

Puisque les régimes d'accumulation et de décomposition de la matière organique sont influencés par une multitude de facteurs qui fluctuent spatialement et temporellement, la tourbe est un médium hautement hétérogène (Letts *et al.*, 2000; Baird *et al.*, 2012; Baird *et al.*, 2016) et anisotrope (Beckwith *et al.*, 2003a; Surridge *et al.*, 2005). Les changements de propriétés hydrauliques verticales peuvent donc se produire à l'échelle centimétrique (Rosa et Larocque, 2008; Bourgault *et al.*, 2017). Pour cette raison, l'utilisation d'essais de perméabilité à charge variable (Baird et Gaffney, 1994; Holden et Burt, 2003; Baird *et al.*, 2004; Clymo, 2004) peut présenter des biais importants liés aux hypothèses d'homogénéité et d'isotropie qu'elle sous-tend (Butler Jr *et al.*, 1996). Des méthodes alternatives visant la quantification de la conductivité hydraulique saturée, telle que la méthode des cubes modifiée (MCM) (Bouma et Dekker, 1981; Beckwith *et al.*, 2003a; Surridge *et al.*, 2005), permettent d'estimer les profils de conductivité hydraulique verticale (K_v) et horizontale (K_h) en profondeur dans les dépôts tourbeux. Les résultats rapportés dans la littérature montre une dynamique d'écoulement de subsurface (< 1 m) généralement dominée principalement par les écoulements latéraux (Hoag et Price, 1995; Ours *et al.*, 1997; Beckwith *et al.*, 2003b; Whittington et Price, 2013), une décroissance marquée des K_{sat} en profondeur (Beckwith *et al.*, 2003a; Surridge *et al.*, 2005) et une présence sporadique de couches aux propriétés hydrauliques distinctes (Rosa et Larocque, 2008). Exceptionnellement, Chason et Siegel (1986) ont obtenu un profil vertical de conductivité hydraulique montrant une constance inhabituelle en profondeur lors d'une étude au nord du Minnesota. Lapen *et al.* (2005) ont par ailleurs montré que les conductivités hydrauliques faibles des dépôts organiques en marge des bogs peuvent contribuer au maintien des niveaux de nappe dans les portions centrales.

1.2.3 Les processus d'échange aquifère-tourbière

Les interactions hydrogéologiques liant les tourbières à l'aquifère superficiel adjacent ont initialement bénéficié de peu d'attention dans la littérature. Les faibles conductivités hydrauliques des couches basales des bassins tourbeux ont amené plusieurs auteurs à négliger les flux verticaux (Toth, 1963; Freeze et Cherry, 1979; Ivanov, 1981; Ingram, 1982, 1983) lors de la conceptualisation de la dynamique hydrique de ces écosystèmes, et ce, malgré la documentation de gradients hydraulique verticaux (Gafni et Brooks, 1990). Reeve *et al.* (2001) ont modélisé la nappe phréatique associée aux tourbières ombrotrophes du lac Agassiz. Leurs travaux suggèrent la présence d'une piézométrie locale de forme convexe associée au dôme de tourbe. L'étude des gradients verticaux et des profils géochimiques de ce site ont confirmé la présence de flux verticaux significatifs (Siegel et Glaser, 1987; Hill et Siegel, 1991). Enfin, des apports en eau souterraine ont également commencé à être identifiés dans des tourbières ombrotrophes généralement considérées comme étant alimentées exclusivement par les précipitations (Drexler *et al.*, 1999b; Fraser *et al.*, 2001).

À l'aide de simulation numérique (MODFLOW et MODPATH), Reeve *et al.* (2000) ont analysé la dynamique des flux aquifère-tourbière pour en évaluer la sensibilité à la topographie régionale, à la proéminence du dôme tourbeux et à la conductivité hydraulique de l'aquifère environnant. Leurs résultats suggère que les propriétés hydrauliques de l'aquifère sont le facteur principal contrôlant l'orientation et la magnitude des écoulements dans les dépôts organiques. En effet, un aquifère composé de dépôts sédimentaires de faibles conductivités hydrauliques favorise les flux horizontaux, tandis que les dépôts de conductivités hydrauliques élevées favorisent les échanges verticaux avec la tourbière. Reeve *et al.* (2000) estiment également que les gradients hydrauliques verticaux sont proportionnels au contraste de conductivité hydraulique qui existent à l'interface tourbière-aquifère. D'autres études ont toutefois décelé la présence de flux verticaux en l'absence de contraste significatif de conductivités hydrauliques entre les deux milieux (Siegel, 1983; Siegel et Glaser, 1987).

Alors que les flux locaux constituent des apports importants aux tourbières (Reeve *et al.*, 2001) et peuvent présenter des inversions (Devito *et al.*, 1997; Levison *et al.*, 2014) c.-à-d. une contribution de la tourbière à l'aquifère superficiel, l'étude détaillée de la piézométrie locale permet davantage de qualifier et/ou de quantifier (Fournier, 2008) les échanges aquifère-tourbière. Une méthode adaptée de l'application de la loi de Darcy (Rosenberry et LaBaugh, 2008) peut être utilisée afin de les estimer spatialement. Cette approche implique certaines hypothèses sur la géométrie du système, les gradients hydrauliques et les conductivités hydrauliques qui rendent très incertains les résultats obtenus. Hunt *et al.* (1996) comparent les résultats d'estimation des apports en eau souterraine dans des milieux humides du Wisconsin avec 1) des bilans massiques d'isotope stable, 2) la modélisation du profil de températures, 3) la modélisation numérique du bilan hydrique et 4) la méthode de Darcy. Ils concluent que les méthodes dépendantes d'une estimation de la conductivité hydraulique (3 et 4) montrent des incertitudes plus grandes.

Bourgault *et al.* (2019) suggèrent que la profondeur de nappe moyenne mesurée dans une tourbière ombrotrophe est liée avec la conductivité hydraulique de l'aquifère qui la borde en aval. En effet, le contact en aval avec des dépôts sédimentaires à faible K_{sat} contribue au maintien d'un niveau de nappe élevée au sein de la tourbière. Ces différents contrôles sur la topographie de nappe dans les tourbières influencent intrinsèquement les gradients dictant les échanges aquifère-tourbière (Price, 1992). Avec des coupes piézométriques transversales dans deux tourbières en environnement montagneux, Millar *et al.* (2018) ont montré que les tourbières de cuvette sont dominées par des gradients hydrauliques verticaux et se caractérisent par un plus faible apport d'eau souterraine. Les tourbières affectées d'un gradient de pente significatif dépendent quant à elles d'un afflux d'eau souterraine plus élevé qui génère principalement des échanges de flux horizontaux. Ferlatte *et al.* (2015) ont proposé des conditions types pour les connexions aquifère-tourbière. Les auteurs décrivent trois types de flux latéraux, soit 1) afflux d'eau souterraine suivant la topographie et se dirigeant vers le centre des dépôts tourbeux, 2) flux convergent entre l'afflux d'eau souterraine et les flux provenant du centre de la tourbière et 3) flux divergent composé d'un écoulement dans les dépôts tourbeux dirigé vers le centre, combiné en périphérie, à un flux de la tourbière vers l'aquifère peu profond.

Hare *et al.* (2017) utilisent des profils thermiques, traçage d'isotopes stables de l'eau ($\delta^{18}O$ et $\delta^{2}H$) et sondage géoradar, afin de cartographier les apports d'eau souterraine dans une tourbière de kettle où la production de canneberges a cessé deux ans avant l'étude. La morphologie du bassin tourbeux, plus précisément l'angle (gradient de pente) de l'interface tourbe-sédiment, semble être le facteur dictant le développement de deux dynamiques distinctes de résurgence. Les flux diffus semblent localisés principalement en marge de la limite de la tourbière où le bassin est peu profond et faiblement en pente, tandis que les flux préférentiels, de type *pipe flow* (Jones, 1981; Ingram, 1983; Woo et DiCenzo, 1988; Glaser, 1992; Holden, 2004, 2005b) se retrouvent dans les zones de changements topographiques abrupts du bassin tourbeux. Des traceurs isotopiques tels que le radon (^{222}Rn) (Berthot *et al.*, 2016), le ratio $^{87}Sr/^{86}Sr$ (Hunt *et al.*, 1998; Hogan *et al.*, 2000) et les isotopes stables de l'eau $\delta^{18}O$ et $\delta^{2}H$ (Hunt *et al.*, 1996; Hare *et al.*, 2017; Isokangas *et al.*, 2017) ont permis la caractérisation des interactions aquifère-tourbière dans différents contextes.

La température de l'eau souterraine profonde (Anderson *et al.*, 2005; Constantz, 2008) et peu profonde (Hare *et al.*, 2017) montre une stabilité distincte et contrastée par rapport aux variations diurnes et saisonnières de l'eau de surface, davantage influencée par la température de l'air. Cette dynamique particulière permet l'élaboration de méthodes de traçages thermiques de l'eau souterraine, notamment des apports aux tourbières (House *et al.*, 2015; Hare *et al.*, 2017) ou des exfiltrations des dépôts organiques vers des ruisseaux (Lowry *et al.*, 2007). Des approches s'appuyant sur la cartographie d'indicateur végétal (Munger *et al.*, 2014; Larocque *et al.*, 2016) ont également été développées. Comme cette méthode inclut plusieurs facteurs associés à la réactivité de la flore aux changements de la composition chimique de l'eau, elle pose plusieurs défis (Glaser *et al.*, 1990). Le traçage effectué à l'aide des concentrations en éléments totaux dissous (TDS) (Ferlatte, 2013; Berthot *et al.*, 2016) doit également tenir compte des échanges de cations provenant des interactions avec l'atmosphère (Drever, 1988) et de la remobilisation des éléments chimiques contenus dans la matière organique (Urban *et al.*, 1995). C'est pourquoi des approches combinant plusieurs indices sont proposées (Hunt *et al.*, 1998; Isokangas *et al.*, 2014; House *et al.*, 2015; Larocque *et al.*, 2016; Hare *et al.*, 2017).

Toutefois, les conditions géologiques, topographiques et climatiques dans lesquelles se produit l'afflux d'eau souterraine dans les tourbières du Moyen-Nord québécois, et plus précisément leur quantification et leur contribution spatiale, restent encore méconnues. La connaissance des interactions aquifère-tourbière est essentielle afin estimer leur vulnérabilité aux changements hydrogéologiques qu'entraîneront les fluctuations climatiques.

1.2.4 Bilan hydrique des tourbières

Le bilan hydrique d'une tourbière permet de déterminer l'importance relative de chacune des composantes de la dynamique hydrique de ces écosystèmes (Ingram, 1983; Price et Maloney, 1994). Selon les termes dominants, il est possible d'estimer leur vulnérabilité aux changements hydrogéologiques, hydrologiques et météorologiques. Un bilan hydrique peut également être utilisé pour estimer un processus non mesuré. C'est le cas notamment dans plusieurs études pour l'estimation des flux échangés avec l'aquifère (Owen, 1995; Rouse, 1998; Drexler *et al.*, 1999a; Fraser *et al.*, 2001; Van Seters et Price, 2001; Kane et Yang, 2004; Déry *et al.*, 2005; Fournier, 2008; Jutras *et al.*, 2009; Proulx-McInnis *et al.*, 2013).

L'équilibre hydrologique des tourbières peut être décrit par le bilan hydrique simplifié suivant :

$$P + G_E + R_E - (ET + G_S + R_S) = \Delta S \qquad (1)$$

Où P représente les précipitations [L], l'indice E les débits entrants [L], l'indice S les débits sortants [L], R les flux surfaciques [L], G les flux d'eau souterraine [L], ET l'évapotranspiration [L] et ΔS la variation de l'emmagasinement [L].

Une multitude de facteurs influencent le taux d'évapotranspiration (ET), tels que la disponibilité en eau, le vent, la topographie, le type de végétation et l'énergie disponible (Bailey *et al.*, 1997). L'ET est le flux sortant dominant le bilan hydrique des tourbières ombrotrophes (Lafleur *et al.*, 2005) et surpasse généralement celle des tourbières minérotrophes (Bridgham *et al.*, 1999). Comme sa quantification est très complexe, l'évapotranspiration potentielle (ETP) est souvent utilisée pour estimer sa valeur maximale (Penman, 1948; Oudin *et al.*, 2005). Dans les tourbières nordiques, les valeurs d'ET mesurées sont principalement comparables à l'ETP ou légèrement plus faibles (Lafleur, 1990; Lafleur et Roulet, 1992; Kim et Verma, 1996; Wu *et al.*, 2010). Exceptionnellement, Anderson *et al.* (2005) ont mesuré une ET représentant 110% de l'ETP dans des fens riverains peuplés d'espèces végétales vascularisées au Danemark. Des instruments tels que le lysimètre (Koerselman et Beltman, 1988) ainsi que des méthodes statistiques telles que la covariance des turbulences (Brümmer *et al.*, 2012; Helbig *et al.*, 2020) font partie des approches visant à mesurer l'ET. L'estimation de l'ET privilégiée par l'Organisation des Nations Unies pour l'alimentation et l'agriculture (FAO) est celle de Penman-Monteith (Monteith, 1965).

Selon le contexte étudié, la corrélation entre la profondeur de la nappe et l'ET est parfois importants (Ivanov, 1981; Lafleur et Roulet, 1992; Kim et Verma, 1996), faible (Roulet *et al.*, 1997; Thompson *et al.*, 1999; Lafleur *et al.*, 2005) ou inexistante (Parmentier *et al.*, 2009; Wu *et al.*, 2010). Selon Wu *et al.* (2010), ces différences peuvent s'expliquer par la faible variation des niveaux de nappe et leur proximité de la surface, même en période d'étiage.

Le ruissellement (R) au sein des tourbières est catégorisé selon trois mécanismes : emmagasinement, transmission et contribution (Black, 1997; Spence et Woo, 2006). Les facteurs contrôlant la transition entre ces trois fonctions distinctes sont 1) les propriétés hydriques des dépôts tourbeux (Waddington *et al.*, 2015), 2) la profondeur de la nappe (Verry *et al.*, 1988) et 3) le type et la position de l'écosystème dans le bassin versant (Devito *et al.*, 2012). L'emmagasinement se produit principalement en période humide et se poursuit jusqu'à la saturation, au-delà de laquelle la transmission s'établit et un écoulement de surface ou de sub-surface se produit (Spence *et al.*, 2011; Goodbrand *et al.*, 2019). Le mécanisme de contribution est associé aux périodes d'étiage où la tourbière libère un débit à un exutoire distinct (Ireson *et al.*, 2015). Au sud du Québec, la contribution annuelle des tourbières au cours d'eau internes et limitrophes de l'écosystème peut atteindre respectivement, de 47 à 100% (Bourgault *et al.*, 2014) et 4 à 7% (Levison *et al.*, 2014) de leur débit de base.

La topographie des dépôts tourbeux montre une oscillation réversible d'une magnitude centimétrique (Ingram, 1983) se produisant à l'échelle d'un évènement pluvieux et à l'échelle saisonnière (Price, 2003; Fritz *et al.*, 2008). Ce régime de contractions et d'expansions de la tourbe est modulé par la variation de l'emmagasinement en eau (ΔS) (McGarry et Malafant, 1987; Hogan *et al.*, 2006) et varie spatialement dans l'espace (Roulet, 1991; Gilman, 1994; Price et Schlotzhauer, 1999; Whittington et Price, 2006). L'emmagasinement en eau montre une proportionnalité avec degré de décomposition de la tourbe (Price et Schlotzhauer, 1999; Clerc, 2009). Ce mécanisme, au comportement hystérétique (Da Silva *et al.*, 1993; Naasz *et al.*, 2005; Fritz *et al.*, 2008), atténue l'amplitude des variations de la profondeur de nappe par rapport à la surface (Roulet, 1991; Van der Schaaf, 1999; Van Seters et Price, 2001; Whittington et Price, 2006). Le changement de volume des dépôts organiques montre une corrélation significative avec les espèces végétales qui la compose (Whittington *et al.*, 2007), mais ne semble pas lié avec la masse volumique apparente (« bulk density ») ou à la présence de textures fibriques dans la tourbe (Price *et al.*, 2005).

Enfin, les dernières composantes du bilan hydrique sont celles des échanges de flux entrants (G_E) et sortants (G_S) d'eau souterraine entre l'aquifère et la tourbière. Ces écoulements peuvent prendre plusieurs formes, tel que des échanges verticaux à la base des dépôts tourbeux qui suivent parfois des chemins préférentiels dans la tourbe (« pipe flow »). De plus, plusieurs types d'interactions latérales sont possible tel que les flux parallèles, convergents et divergents (Ferlatte *et al.*, 2015), décrivant la relation entre les directions d'écoulement dans la tourbe et dans l'aquifère superficiel.

1.2.5 Modélisation

La modélisation numérique des échanges de flux aquifère-tourbière se bute bien souvent à la complexité de la dynamique unique des interactions entre les composantes hydrologique et hydrogéologique de ces écosystèmes (Swain et Wexler, 1996). Néanmoins, elle permet d'étudier la répartition tridimensionnelle des flux, la sensibilité des modèles aux paramètres utilisés, et la réactivité des processus aux changements des données d'entrée.

Des efforts importants ont été consacrés dans les dernières années au développement de modèles pouvant reproduire cette dualité de la dynamique hydrique des tourbières. Restrepo *et al.* (1998) proposent un outil de simulation dédié à la représentation des milieux humides, le *Wetland Package*, pouvant être utilisé dans MODFLOW (Harbaugh, 2005). La complexité des données nécessaires à son implémentation représente un frein important à son utilisation. En effet, le progiciel requiert de définir les coefficients α et β de l'équation de Kadlec, le coefficient de rugosité de la sphaigne, les porosités effectives des différentes couches organiques ainsi que la profondeur de la zone capillaire. Wilsnack *et al.* (2001) ont simulé adéquatement, à l'aide de ce progiciel, la dynamique des écoulements en eaux souterraines des milieux humides hydrauliquement connectés à l'aquifère superficiel de la région de Miami-Dade. À l'échelle régionale, les milieux humides sont souvent conceptualisés sous forme de zone de charge constante, de zones de lac ou de zones de drainage (Halford, 1998). Ces techniques sont cependant mésadaptées aux études hydrogéologiques à l'échelle locale.

Feinstein *et al.* (2020) ont développé un outil de simulation de la zone non saturée, le *UZF Package « Unsaturated Zone Flow »*. Le progiciel UZF (Niswonger *et al.*, 2006) utilise l'élévation topographique de la nappe phréatique afin de départager la proportion du taux d'infiltration, imposé en donnée d'entrée, qui intègre l'aquifère (percolation) sous forme de recharge des débits excédentaires alors rejetés des zones sursaturées du modèle hydrogéologique. En surface, cet ajout au modèle conceptuel permet de simuler et cartographier les zones de résurgence dont les débits sortants associés sont modulés par les valeurs de conductivités hydrauliques verticales attribuées à la couche numérique supérieure. Cette approche fut initialement proposée par Feinstein *et al.* (2020) pour cartographier la distribution régionale des tourbières minérotrophes en se basant sur l'hypothèse que la viabilité de ces écosystèmes est étroitement associée à ses zones de résurgence.

Des travaux effectués sur le territoire québécois ont quantifié les échanges aquifère-tourbière avec des approches numériques (Bourgault *et al.*, 2014; Levison *et al.*, 2014; Quillet *et al.*, 2017). Jutras *et al.* (2009) proposent un sous-modèle (PHIM) au modèle hydrologique HYDROTEL (Fortin *et al.*, 2001a, 2001b; Rousseau *et al.*, 2003) afin d'intégrer l'impact hydrologique des milieux humides sur les débits simulés du bassin versant de la rivière Nécopastic, Baie-James, Québec.

Plusieurs modèles numériques adoptent une approche écosystémique afin de soutenir une représentation de la complexité des interactions hydriques, physiques, chimiques, thermiques et organiques entre la tourbière, l'eau souterraine et l'eau de surface (BIOME-BGC (Running et Coughlan, 1988); SHAW (Flerchinger et Saxton, 1989); Expert-N; (Engel et Priesack, 1993); COUP (Jansson, 2004); Wetland-DNDC (Zhang *et al.*, 2002); PEATLAND-VU (Van Huissteden *et al.*, 2006); McGill wetland model (St-Hilaire *et al.*, 2010), DigiBog (Baird *et al.*, 2012)).

1.2.6 Changements climatiques

Dans la région Moyen-Nord de la province de Québec (Canada), les projections climatiques utilisant les scénarios d'émission RCP4.5 (modéré) et RCP8.5 (élevé), c'est-à-dire adaptées des scénarios de forçage radiatif de Rogelj *et al.* (2012) pour 2041-2070, montrent des températures de l'air plus chaudes (respectivement +2,3 et +3,3°C) et une augmentation des précipitations annuelles (respectivement +63 et +100mm) (Ouranos, 2015). Selon ces projections climatiques, les tourbières nordiques connaîtront un accroissement du déficit de pression de vapeur (Yuan *et al.*, 2019), qui se traduira par une augmentation de l'évapotranspiration des tourbières par rapport à l'ensemble de la forêt boréale (Helbig *et al.*, 2020). Les tourbières nordiques seront vulnérables à ces changements climatiques (Roulet *et al.*, 1992), en particulier dans les plaines boréales où les proportions de précipitation et d'évapotranspiration potentielle sont similaires, dictant le taux de précipitations net c.-à-d. l'estimation de la recharge (Bergengren *et al.*, 2011; Flanagan et Syed, 2011; Ireson *et al.*, 2015; Schneider *et al.*, 2016) et affectant subséquemment la relation nappe-topographie (Devito *et al.*, 2005; Haitjema et Mitchell-Bruker, 2005; Hokanson *et al.*, 2020).

Des études réalisées au Canada ont été menées à l'aide de modèles hydrologiques (Jyrkama et Sykes, 2007; Croteau *et al.*, 2010) et de modèles hydrogéologiques (Scibek et Allen, 2006) afin de comprendre et d'anticiper les impacts des changements climatiques. Elles prédisent diverses fluctuations de recharge montrant une diminution (Wang *et al.*, 2014), une augmentation (Scibek et Allen, 2006; Jyrkama et Sykes, 2007; Dubois *et al.*, 2021) ou les deux (Croteau *et al.*, 2010; Levison *et al.*, 2014). Larocque *et al.* (2019) montrent cependant que ces différentes prédictions climatiques ne présentent pas de tendances spatiales claires dans l'est de Canada.

Les conditions transitoires attendues dans un climat changeant devraient également induire des températures hivernales plus chaudes (Ouranos, 2015). Ceci aurait pour effet d'étendre les périodes de recharge automnale et printanière, de réduire le stock de neige au sol l'hiver et donc l'amplitude de la fonte, et provoquer des événements de recharge sporadiques (Jyrkama et Sykes, 2007; Okkonen et Kløve, 2011). L'augmentation de l'intensité des précipitations (Sillmann *et al.*, 2013; Ouranos, 2015) devrait modifier différemment la recharge tout au long de l'année (Fu *et al.*, 2019; Dubois *et al.*, 2021). Ces changements intra-annuels à court terme du bilan hydrique des tourbières pourraient également entraîner des changements hydrologiques à long terme (annuels à décennaux) et causer des modifications fondamentales de l'accumulation de tourbe.

1.3 Objectifs et structure du mémoire

Les objectifs de cette recherche étaient de 1) quantifier le bilan hydrique actuel de deux tourbières du Moyen-Nord du Québec, 2) déterminer si l'afflux d'eau souterraine est une contribution importante à leur bilan hydrique et documenter leur sensibilité aux changements de recharge. Les hypothèses de recherche sont (1) que les deux tourbières étudiées reçoivent un apport en eau souterraine en provenance de l'aquifère superficiel environnant qui est localisé dans des dépôts perméables de till, contribuant ainsi à l'équilibre hydrologique et au régime nutritif de ces tourbières et (2) que des variations de recharge peuvent modifier l’équilibre des conditions hydrologiques de ses écosystèmes en influençant la dynamique des échanges aquifère-tourbière aux tourbières étudiées.

Les deux tourbières étudiées dans le cadre de ce projet sont considérées représentatives des écosystèmes de la région circumpolaire du Moyen-Nord du Québec, essentiellement caractérisées par une dominance de microformes aquatiques telles que les mares et les dépressions humides. Il s’agit de tourbières structurées de type minérotrophe pauvre montrant des signes de transition vers des conditions ombrotrophique dans leurs zones centrales. Elles se sont développées dans les dépressions topographiques drumlinoïdes et morainiques du till sablo-silteux composant le substrat géologique régional.

Des levés topographiques détaillés ont permis l'élaboration d'un modèle conceptuel 3D de la géométrie des tourbières et de leurs bassins versants. Des puits et piézomètres répartis sur les deux tourbières ont été instrumentés de sondes de suivi des niveaux de nappe échelonnées sur les années 2018 et 2019. Des essais hydrauliques ainsi qu'un carottage manuel de la tourbe ont été réalisés afin d'estimer la conductivité hydraulique des dépôts sédimentaires et organiques respectivement. Un calcul de flux de Darcy, selon une méthode par segment, a été utilisé afin d’estimer et de spatialiser les échanges aquifère-tourbière. Le développement d'un modèle hydrogéologique en régime permanent (MODFLOW) a permis la quantification du bilan hydrique des deux sites d’étude, tout en constituant un outil permettant d’évaluer leur dépendance aux conditions hydrogéologiques environnantes actuelles et futures. Ces travaux ont permis de développer les connaissances sur les tourbières du territoire nordique québécois et seront présentés en trois chapitres distincts.

Le premier chapitre, fait la revue des connaissances actuelles sur les tourbières. Le cœur du mémoire est ensuite présenté sous la forme d’un article qui sera soumis dans un numéro spécial de la revue Frontiers in Earth Sciences portant sur les écosystèmes dépendants des eaux souterraines. Cet article présente l’ensemble de la démarche scientifique utilisée ainsi que les résultats obtenus à la suite du traitement des données *in situ* et de la modélisation hydrogéologique (MODFLOW) ayant permis notamment, la quantification des échanges aquifère-tourbière. Enfin, le chapitre III conclu la démarche récapitulative, discutant notamment de la portée et de la reproductibilité de la recherche.

CHAPITRE II

AQUIFER-PEATLAND HYDROLOGIC CONNECTIVITY AND CONTROLLING FACTORS IN HIGH LATITUDE PEATLANDS

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ABSTRACT

The conditions in which groundwater inflow occurs in oligotrophic boreal peatlands of eastern Canada and its contribution to peatland water balance are still poorly understood. The objectives of this research were to quantify aquifer-peatland hydrological connectivity and to identify controlling factors in high-latitude peatlands of north-central Quebec (Canada). The peatlands were instrumented with piezometers and groundwater levels were monitored during two growing seasons. The peatland water budgets were simulated for the two peatlands with a steady-state groundwater flow model and a sensitivity analysis to recharge was performed. The two peatlands have contrasted water budgets, with recharge representing the largest inflow and subsurface runoff representing the largest outflows at the peatland with the smallest contributing area (Misask). The peatland with the largest contributing area (Cheinu) is also located downgradient within the regional hydrosystem. Its inflows are dominated by groundwater and its outflows are mostly towards subsurface runoff. The two peatlands are in conditions of precipitation excess and a recharge reduction would not affect their water table depths markedly. However, recharge changes could induce larger modifications in groundwater inflows and outflows for the peatland with a larger contributing area. The dominating peatland hydrological functions are thus contrasted at the two sites, and it is hypothesized that the water table depths thresholds triggering changes between storage, transmission and runoff functions are also different. Although work remains to be done to understand transient-state hydrological conditions and the long-term impacts of a changing climate and the resulting feedbacks, this work shows that understanding peatland hydrology requires assessing aquifer-peatland connectivity at the regional scale.
KEY WORDS
Peatland; aquifer; MODFLOW model; climate change; boreal (Québec, Canada)

1.4 Introduction

Peatlands cover 12 to 17% of the Canadian territory (Letts *et al.*, 2000; Tarnocai *et al.*, 2000; Payette et Rochefort, 2001; Lafleur *et al.*, 2005) and approximately 9 to 12% of the Quebec territory (Gorham *et al.*, 1998; Payette et Rochefort, 2001). Their occurrence in the boreal biome is estimated to account for half of the Earth's wetland cover (Matthews et Fung, 1987; Aselmann et Crutzen, 1989; Bridgham *et al.*, 1999). In the north-central region of the Province of Quebec (Canada), peatlands are mostly characterized by poor fens dominated by a succession of elongated strings and pools microforms parallel to the slope (Payette et Rochefort, 2001; Rydin *et al.*, 2013) that compares well with Fennoscandia and northwestern Russia boreal aapa mires (Botch et Masing, 1979; Seppälä et Koutaniemi, 1985; Foster et Fritz, 1987).

Aqualysis is described as the result of a hydrological imbalance in peatlands due to the increase in water table depths promoting more aquatic conditions to the detriment of terrestrial surfaces (Payette, 2008; Dissanska *et al.*, 2009; Proulx-McInnis *et al.*, 2013). It has been hypothesized that aqualysis could be linked to increased influxes of both surface water and groundwater inflow to northern peatlands (van Bellen *et al.*, 2018). However, hydrological conditions that lead to this disequilibrium are yet unknown. Since aquifer-peatland exchanges are also not well understood, it remains unclear if changes in groundwater inflows can play a role in peatland aqualysis.

It has been estimated that groundwater inflow can represent approximately 50% of the total inflow of water in peatlands from southern Quebec (Bourgault *et al.*, 2014; Levison *et al.*, 2014). Incoming fluxes from an aquifer are recognized to be highly variable according to geology, morphology and climate (Kløve *et al.*, 2014). Peatlands have received scientific attention especially for their hydrology (Whittington et Price, 2006; Spence *et al.*, 2011; Waddington *et al.*, 2015; Goodbrand *et al.*, 2019), their capacity to store carbon (Moore *et al.*, 1998; Blodau, 2002; Ju *et al.*, 2006; Strack et Waddington, 2007; Miquelajauregui *et al.*, 2019) and their resilience to anthropic pressures such as roads (Strack *et al.*, 2018; Saraswati *et al.*, 2020; Elmes *et al.*, 2021), fires (Morrissey *et al.*, 2000; Bourgeau-Chavez *et al.*, 2020; Nelson *et al.*, 2021) and drainage (Whittington et Price, 2013; Menberu *et al.*, 2016; Menberu *et al.*, 2018; Gyimah *et al.*, 2020; Harris *et al.*, 2020). However, peatlands located in remote northern conditions have been the object of relatively few studies, probably in part because access and harsh winter conditions are a challenge but also because they are subject to much less anthropogenic pressure.

According to climate change projections, northern peatlands will experience an increase in vapor pressure deficit (Yuan *et al.*, 2019) resulting in increased evapotranspiration that could exceed that of the surrounding boreal forest (Helbig *et al.*, 2020). In the central-north region of the province of Quebec (Canada), climate projections for 2041-2070 under RCP4.5 (representative concentration pathway) (moderate emission increase) and RCP8.5 (high emission increase) suggest that future temperatures will be warmer (respectively +2.3 and +3.3°C) and future precipitation will be higher (respectively +63 and +100 $mm.yr^{-1}$) (Ouranos, 2015). These climate conditions could lead to important changes in groundwater recharge (Bergengren *et al.*, 2011; Flanagan et Syed, 2011; Ireson *et al.*, 2015; Schneider *et al.*, 2016), affecting water table depth and the influx of groundwater to peatlands (Devito *et al.*, 2005; Haitjema et Mitchell-Bruker, 2005; Hokanson *et al.*, 2020). Recent studies for southern locations suggest a range of climate-driven changes in recharge (Levison *et al.*, 2014; Wang *et al.*, 2014; Mackay *et al.*, 2020; Pozdniakov *et al.*, 2020; Dubois *et al.*, 2021; Jannis *et al.*, 2021). Combined with the challenges of simulating aquifer-peatland connections (Devito *et al.*, 2005; Price *et al.*, 2013; Hokanson *et al.*, 2019), predicting the impact of hydrological conditions in peatlands under uncertain future climate is a challenge.

The objectives of this research were to 1) quantify the current water budget of two peatlands in north-central Quebec, 2) determine whether groundwater influx is an important contribution to their water budget and how it is sensitive to changes in recharge. Two peatlands located in north-central Quebec (Canada) were characterized and their hydrology was monitored during two growing seasons. A groundwater flow model developed for each peatland was used to quantify their water budgets and to provide insights into the effects of changing recharge.

1.5 Study sites

1.5.1 Geomorphology and hydrogeology

In the north-central region of the province of Quebec (Canada), peatlands developed mostly in depressions of the precambrian Shield. Dyke (2004) estimated the retreat of the Laurentide Ice Sheet to occur between 6500 and 7000 cal. BP using regional paleogeographic reconstruction based on ^{14}C dating (Bouchard, 1980).

Surface deposits are dominated by sandy to silty (including boulders) glacial till showing elongated erosional (drumlins) forms associated with subparallel eskers oriented at N220° (Lamarche et Hébert, 2020). Nearby drilling shows no major marine or lacustrine deposit and reveals that glacial till can reach 30 m (Ortega et O’Connor, 2005; Desbiens, 2008; Leblanc et Kendle, 2008) and glaciofluvial deposits can be more than 40 m in thickness (Lamarche et Hébert, 2020). Nevertheless, the present-day thickness of the unconfined aquifer and the hydraulic properties of both the Quaternary deposits and the underlying bedrock are unknown. However, the widespread and thick highly permeable deposits, and the cold and humid climate suggest potential important groundwater exchanges between aquifer and peatlands.

1.5.2 Misask and Cheinu peatlands

The Misask (52° 43' 27" N, 72° 12' 50" W) and Cheinu peatlands (52° 38' 48" N, 72° 11' 31" W) (unofficial names) are located at the ecotone between the open (taiga) and closed boreal forest biomes (Fig. 1). Peat accumulation initiated around *ca* 6430 and 6560 cal. BP (Robitaille *et al.*, 2021), suggesting an early post-glacial initial vegetation cover. This region represents the biogeographic limit of ombrotrophic peatlands, which dominate the boreal region and are gradually replaced by oligotrophic patterned fens, with abundant pools increasing northward (Payette et Rochefort, 2001). This biogeographical distribution, related to greater surface runoff and cooler climate limiting *Sphagnum* growth, is observed across the northern hemisphere (Seppälä et Koutaniemi, 1985; Foster et Fritz, 1987; Payette et Rochefort, 2001; Rydin *et al.*, 2013).

Both peatlands show a minerotrophic surface pattern with alternating elongated strings and pools parallel to the slope. Robitaille *et al.* (2021) suggested that these two previously minerotrophic peatlands have registered an increase in their hydrological conditions i.e. causing aqualysis, from the Neoglacial cooling to the end of the Little Ice Age followed by a recent ombrotrophication trend during the past decades (Robitaille *et al.*, 2021). Surface vegetation is dominated by *Sphagnum* moss species on both sites with local changes controlled by microtopography as herbaceous species (*Carex* spp and *Trichophorum cespitosum*) are mainly found in depressions and ericaceous species (*Chamaedaphne calyculata* and *Kalmia angustifolia*) in strings (Robitaille *et al.*, 2021). Peripheral minerotrophic areas are characterized by thinner peat (< 1 m), with black spruce (*Picea mariana*) and larch (*Larix laricina*) and show recent and ongoing paludification process.

The studied peatlands have developed in local topographic depressions created by the drumlin/moraine landscape, reaching an estimated average depth of 6.2 and 11 m at Misask and Cheinu site respectively (Fig. 1). These topographic depressions are underlain by boulders inherited from excessive leaching of till (Dionne, 1978), which represent an unusual permeable substratum for peatland development. The Misask peatland (0.102 km^2) has an average topographic slope of 1% oriented southwards. Shallow pools, covering about 10% of its total area, are found exclusively in the eastern portion of the peatland. The Cheinu peatland (0.147 km^2) developed in a deeper bowl-shaped depression located at the bottom of a significant slope (~6.7%). The peatland has an average southeast 0.5 % slope. The Cheinu peatland shows well developed (perpendicular to slope) pools, covering 60% of its total area.

Peat depths were surveyed using manual probing at > 300 observation points per site with an approximate 25 m spacing. The reconstructed peat basin shows a mean peat thickness of 0.71 and 0.87 m for Misask and Cheinu peatlands, respectively. Maximal peat thickness is reached in the central patterned area of the Misask site (2.08 m) and in southwest floating moss-mat area of the Cheinu site (3.28 m). This probing survey also confirmed that the beds of the studied peatlands are dominated by boulders and sand filled interstice.

Located in the upper part of the large watershed of the Eastmain River, the studied peatlands are also included in the watershed of the Misask River i.e. an important tributary of the Eastmain River. Their respective sub-watershed, covering areas of 0.226 km^2 (Misask) and 0.487 km^2 (Cheinu) show successions of small lakes (mostly kettle-lakes) linked by minor streams. No significant tributary streams pass within the studied peatlands. Permanent outflows are present at the two peatlands in the shallower paludified fen areas dominated by living mosses and black spruce. The eastern outlet of the Misask site feeds a small stream running south and a second outlet located southwest also running south (Fig. 1). Narrow paludified areas in the southern portion of the Cheinu peatland contribute to a 1.8 m wide outlet stream that flows into the downstream lake. Subwatersheds of studied sites site altitude ranged between 483 and 468 m at Misask (average slope 1.9%) and 482 and 455 m (average slope 3.1%).

1.5.3 Meteorological conditions

Meteorological data is obtained from the weather station located approximatively 15 km north, at the Renard mine. Between 2017 and 2019, annual precipitation (P) varied between 779 and 821 $mm.yr^{-1}$ (average 797 $mm.yr^{-1}$). The average annual temperature was -3.0°C, with snow occurring each year between mid-October and May. The potential evapotranspiration rate (PET) was calculated using the Oudin *et al.* (2005) equation based on the mean daily air temperature, and the Morton (1983) estimation of the extraterrestrial radiation. Annual PET for 2017 to 2019 ranged between 341 and 372 $mm.yr^{-1}$ (average of 360 $mm.yr^{-1}$). Documented actual evapotranspiration rates in northern peatlands were found to be mainly equivalent to or slightly lower than PET (Lafleur, 1990; Lafleur et Roulet, 1992; Kim et Verma, 1996; Wu *et al.*, 2010; Tardif *et al.*, 2014). The nearest weather station, located 376 km northeast of Renard Mine, provided climatic data between 1981 and 2010 with a mean annual temperature of -2.9°C and a mean annual precipitation of 647 mm (Environnement Canada, 2006).

1.6 Methodology

1.6.1 Peatland instrumentation and monitoring

Between June 2018 and October 2019, each peatland was instrumented with five piezometers ((P_m or P_c “m” stands for Misask and “c” stands for Cheinu; Fig. 1) and five observation wells (W_m or W_c; Fig. 1) equipped with hourly water level-recording loggers (*Solinst*). Each piezometer and well consisted of a 25 mm diameter PVC tube equipped with a 0.3 m screen in the mineral substrate or screened from top to bottom in the peat deposits. A barometer (*Solinst*) was installed at the Cheinu site for barometric compensation. In the Misask and Cheinu peatlands respectively, five and eight additional non-instrumented piezometers and wells were used for manual head measurements during each field visit (June, July, September of 2018 and June, September of 2019).

On each site, a metal rod was installed through the peat (lawn) and fixed into the underlying mineral to allow measurements of peat contraction and expansion between field campaigns. The variation of measured seasonal peat elevations reached a maximum of 1 cm (Misask) and 2.5 cm (Cheinu). A similar method was used at the Cheinu site outlet lake to quantify water level fluctuations compared to a metal bar referenced level. The maximum lake level variation of 19.2 cm was considered negligible in the numerical model.

1.6.2 Hydraulic conductivities

Two marginal and one central (1 m) peat cores were collected from the deepest portion of each peatland (Fig. 1) using a Box corer (Jeglum *et al.*, 1992). In the laboratory, the peat cores were divided in 8 cm x 8 cm x 8 cm cubes on which vertical and horizontal peat hydraulic conductivities were measured using the Modified Cube Method (Beckwith *et al.*, 2003a).

Slug tests were performed in piezometers located in the surrounding glacial till and in fluvioglacial deposit to estimate their respective hydraulic conductivity. Falling-head tests were conducted by adding volume of water i.e. from 10 to 90 ml depending on recovery time, or by using the logger as a slug, in the PVC tube with recording frequency set to a 2 second interval. For each aquifer piezometer, two or three hydraulic tests were carried out with an average duration of 5 minutes. Slug test results were interpreted using the Hvorslev (1951) method.

1.6.3 Piezometric map

At both sites, a local piezometric map was estimated as a subdued replica of topographic data with water bodies constrained to topography (Fig. 1). A detailed digital elevation model (DEM) was produced using photogrammetric drone data provided by Stornoway Renard Mine managers (unpublished data) and a differential global positioning system (DGPS Trimble) field survey. The digital surface model (DSM) produced by the photogrammetric survey with mean RMS error of 0.005 (Misask) and 0.004 m (Cheinu), was converted to a digital terrain model (DTM) by processing point clouds with 3DReshaper software (Technodigit, 2009). The resulting DTM was geo-referenced to match the DGPS survey with respective XY error and Z error of 0.004 and 0.005 m (Misask) and 0.005 and 0.006 (Cheinu). Elevations for all piezometers and wells, for the surrounding water bodies, as well as data from three existing observation wells (5 m deep in the granular aquifer; average water table depth 35 cm; UWP-01 to UWP-03) at the Renard Mine airport, were extracted from this DTM and used to build the piezometric map.

1.6.4 Groundwater inflows, recharge and surface outflow rates

Hydraulic gradients were quantified between selected close pairs of peat wells and mineral piezometers and were used to estimate aquifer-peatland groundwater exchanges ($q_{A\text{-}P}$) with the Darcy Law and the segment method (Rosenberry et LaBaugh, 2008):

$$q_{A-P} = \sum_{j=1}^{n} K_h l_j d_p i_h \Big/ A_p \qquad \text{eq. 1}$$

where $q_{A\text{-}P}$ is the unitary flow exchanged horizontally between the peat and the mineral deposits (mm/yr), K_h is the geometric average horizontal hydraulic conductivity (m/s) between mineral sediments (average value derived from the slug tests) and peat (average value for the top 30 cm of peat derived from MCM peat core measurement), l_j is the length of each segment around the peatland (m; colored peat contour sections on Fig. 1), d_p is either the average peat depth or the maximum peat depth (m) (the two values are used to set an interval of possible fluxes), i_h is the horizontal hydraulic gradient (m/m) estimated from the average heads available for the piezometer-well pair, and A_p is the peatland area (m^2).

Groundwater recharge was estimated using annual net precipitation values (P_{net} = P-PET) between 2017 and 2019, based on the hypothesis that there is limited runoff due to the high hydraulic conductivity of surface deposits. P_{net} varies between 414 and 450 $mm.yr^{-1}$ (average 434 $mm.yr^{-1}$). The flow rate of the main outlet of the Cheinu peatland was measured manually on each site visit with a current flow meter (*Swoffer*). Flow rates were calculated using the velocity-area method (Herschy, 1993). The outlets of the Misask peatland were too diffuse to allow for field measurement.

1.6.5 Groundwater flow model

A finite-difference steady-state groundwater flow model was developed with MODFLOW (Harbaugh, 2005) for each peatland. A grid with a maximum cell size of 128 m was locally refined to 4 m cells near the peatlands area. It was then vertically discretized into 14 layers reaching a total depth of 70 m, starting with a minimum thickness of 0.25 m within the first 2 m and gradually increases with depth to reach a maximum layer thickness of 30 m.

The Misask and Cheinu models cover respectively 15.37 and 7.14 km^2, i.e. approximatively 47 and 11 times their watershed area to ensure that the boundary conditions have no effect on the modeling results. Model limits were composed of rivers and lakes integrated as full-depth constant head boundaries, connected by no-flow segments identified from the piezometric map. The bottom of the 70 m deep numerical grid, that represents hypothetical bedrock contact, was also set as a no-flow boundary. Due to the lack of information on their connectivity to the surficial aquifer, the small lakes and wetlands located inside the models were represented with constant heads restrained to the top layer (Fig. 2). The peat bathymetry were defined based on bathymetric data from Robitaille *et al.* (2021).

Outside the peatlands, infiltration zones and hydraulic conductivity zones were established based on the Quaternary deposits map. Each model was discretized into five distinct zones including the peatland (Fig. 2). Sediment types are different for the two peatlands since the Misask watershed includes a more marked topography covered with thin glacial till (< 1 m) and the Cheinu model includes significant boulder fields. The budget zone of the peat deposits was divided in two zones, the upgradient and downgradient portions of the peatlands.

Within the peat deposits, hydraulic conductivities for the first meter were set to the mean measured K_h and K_v values estimated with the Modified Cube Method. Hydraulic conductivities for the mineral deposits (glacial till and fluvioglacial deposits) were estimated from slug test results and literature values. A vertical anisotropy (K_h/K_v) of 10 was imposed to the peat layer below 1 m and to all surrounding mineral deposit.

The unsaturated zone flow (UZF) package (Niswonger *et al.*, 2006) was used to redirect infiltration into groundwater recharge and leakage which was here attributed to subsurface runoff that is exported out of each cell. This package, used to map fen regional distribution by Feinstein *et al.* (2020), offers a means to better represent shallow subsurface flow within a peatland and thus provide a more reasonable water budget. The UZF package parameter SURFDEP (Niswonger *et al.*, 2006) which represents the topographic variation maximum amplitude within a grid cell, was fixed at 0.8 m. Feinstein *et al.* (2020) find this parameter to be not sensitive for surficial fen hydrology. Even if the UZF package allows controlling the top surficial layer K_v as a calibration parameter, this value was fixed to match the underlying K_v material. Feinstein *et al.* (2020) have shown that variation of this parameter may lead to unrealistic results.

1.6.6 Model calibration and sensitivity to recharge

Mineral K_h and infiltration rates were calibrated using PEST (Doherty, 2005) to reproduce average measured heads in the peatlands and in the aquifer. Parameters were afterward manually adjusted to ensure decreasing values of K with depth.

A sensitivity analysis was performed on recharge. Various studies have been carried out in southern regions of Canada to quantify recharge (Jyrkama et Sykes, 2007; Croteau *et al.*, 2010). They predict various recharge fluctuations with decreases reaching -50% (Bourgault *et al.*, 2014; Wang *et al.*, 2014), increases reaching +53% (Scibek et Allen, 2006; Jyrkama et Sykes, 2007), and a variety of changes between from -59% to +15% (Rivard *et al.*, 2009; Croteau *et al.*, 2010; Levison *et al.*, 2014; Dubois *et al.*, 2021). Here, model sensitivity to recharge was tested with scenarios of -50%, -20% and +20% with an emphasis on recharge reduces as Robitaille *et al.* (2021) revealed a recent increase in temperature and in growing degree-days combined with a decrease in precipitation. Similar range of recharge changes has been used by (Bourgault *et al.*, 2014) and (Levison *et al.*, 2014).

1.7 Results

1.7.1 Measured hydraulic conductivities

Slug tests in the glacial till (Misask, Fig. 3a) show K values ranging between 1.2×10^{-5} m.s^{-1} and 3.3×10^{-7} m.s^{-1} (average 4.1×10^{-6} m.s^{-1}). Slug tests in the fluvioglacial deposits (Cheinu, Fig. 3b) show slightly higher K values ranging between 1.2×10^{-4} m.s^{-1} and 7.6×10^{-6} m.s^{-1} (average 3.8×10^{-5} m.s^{-1}). No K measurement was conducted in washed out till nor thin till. It is assumed that they were, respectively, equal to K values in fluvioglacial deposits and in the same range as K values in glacial till.

At both sites, MCM results showed generally decreasing trends with depth of average K_h and K_v with maximum K values observed at the surface (Fig. 3). The average maximum K_h and K_v values (topmost cubes) were, respectively, 8.7×10^{-4} and 8.6×10^{-4} m.s^{-1} at Misask (Fig. 3a) and 5.3×10^{-4} and 5.5×10^{-4} m.s^{-1} at Cheinu (Fig. 3b). Top cubes (0-8 cm) of Cheinu peat cores were dismissed due to the inability to execute MCM tests properly (wax filling peat pores and friability of the living moss). Hydraulic conductivities rapidly decrease within the acrotelm reaching 0.28 m (Misask) and 0.44 m (Cheinu), followed below by less clear decreasing trends. Minimum K values were not directly associated with the deepest tested cubes, as minimum average K_h and K_v were respectively 1.6×10^{-6} $m.s^{-1}$ (0.84 m) and 4.7×10^{-7} $m.s^{-1}$ (0.68 m) at Misask and 5.5×10^{-7} $m.s^{-1}$ (0.92 m) and 3.6×10^{-6} $m.s^{-1}$ (0.76 m) at Cheinu. At both sites, hydraulic conductivity profiles were relatively similar for the three locations, but cores located downgradient had slightly lower hydraulic conductivities than the upstream and central cores.

The mean vertical anisotropy (K_h/K_v) ranged from 1.02 (0.04 m) to 7.72 (0.44 m) at the Misask peatland, with low values under 0.20 m followed by irregularly decreasing values deeper in the peat (Fig. 3a). A similar trend was observed at the Cheinu peatland with a top K_h/K_v of 1.69 followed by a maximum of 10 at 0.20 m followed by a constant decrease with depth and a minimum value of 0.24 at 0.84 m (Fig. 3b).

1.7.2 Water table depths and heads

Monitored wells within the organic deposits at Misask show water table depth (WTD) (Fig. 4) variations of between 57 and 6 cm below the surface. Minimum WTD occurred upgradient of the central peatland area immediately following snowmelt (June 2018: W_m05), while maximum WTD was observed in the paludified downgradient area during the growing season (July 2018: W_m08). The Cheinu peatland WTD showed a similar range, from 55 to 7 cm below the surface. Minimum WTD occurred within the central peatland area at the end of the growing season (October 2018: W_c28) and maximum WTD was observed near the woody main outlet during the growing season (August 2018: W_c22). The minimum recorded WTD of 6 cm (Misask) and 7 cm (Cheinu) show that no flooding occurred in the monitored areas between June and October 2018 and 2019. At both sites, the most central peatland wells (W_m04 and W_c28) were characterized by shallow water tables, with respective average WTD of 18 cm and 15 cm.

Except for the central peatland wells (W_m04, shallow WTD) and one peatland well located close to the northern boundary (W_m16, deeper than expected WTD), the Misask peatland generally shows increasing WTD from its northeastern section toward the southwestern area. A similar pattern of increasing WTD is visible at the Cheinu peatland. Except for one peatland well located close to the northern edge (W_c23, deeper than expected WTD) and for the central peatland well (W_c28, shallow WTD), the average WTD generally decreases from north (average WTD 19 cm) to south (average WTD 41 cm). These are indications of northeast-southeast and north-south flow directions within the Misask and Cheinu peatlands respectively. Exceptions to these trends at W_m16 and W_c23 showing deeper WTD, correspond to the forest border where peat is shallow, poorly decomposed with high porosity and associated with current lateral expansion.

In the mineral sediments, WTD (Fig. 4) varied from 78 to 6 cm at Misask and from 68 to 1 cm at Cheinu. In both sites, minimum WTD was measured at upstream piezometers following snowmelt in 2018 (P_m15) and the end of the growing season in September 2018 (P_c24). Maximum WTD occurred in downstream piezometers during the growing season, in August 2019 (P_m09) and in July 2019 (P_c30). Piezometers in mineral deposits surrounding the Misask peatland showed deeper average WTD compared to those installed at Cheinu (except for P_m03). Piezometers at Misask have higher average WTD northwest of the peatland than southwest of the peatland. Similarly, the piezometers north of the Cheinu peatland have higher average WTD than those southward.

At the two sites, heads reflect the general groundwater flow directions within the mineral deposits (Fig. 5). At the Misask peatland, heads in the mineral deposits indicate groundwater flows from P_m07 to P_m09, from northeast to southwest and in the direction of the diffuse southwest outlet, and from P_m07 to P_m06, i.e., towards the southern peatland outlet. At the Cheinu peatland, heads in the mineral deposits indicate flow from P_c13 to P_c26 and P_c30, i.e., from north to south. Heads follows a similar direction within the peat, from W_c17 to W_c28 and W_c22, towards the southern peatland outlet.

1.7.3 Calculated hydraulic gradients and flows

The piezometric map shows that local topography-driven flow system contributes to both upgradient and downgradient areas of the peatlands. At Misask site, general flow direction is oriented from northeast to south-west. Although, north-west border receive flow directed southeast.

The studied peatlands are primarily fed by precipitation and groundwater inflow (no tributary streams). Hydraulic gradients were calculated for all the sections used in the Darcy flow calculation. A negative hydraulic gradient (NHG) drives flow from the peatland to the aquifer and a positive hydraulic gradient (PHG) drives flow from the aquifer to the peatland. The average horizontal hydraulic gradients calculated between piezometers in the mineral deposits and peatland wells varied between -0.025 $m.m^{-1}$ (P_m06-W_m04) and 0.010 $m.m^{-1}$ (P_m13-W_m16) at Misask, and between -0.004 $m.m^{-1}$ (P_c25-W_c28) and 0.012 $m.m^{-1}$ (P_c13-P_c23) at Cheinu (Fig. 6). Both sites were characterized by similar PHG ranges. The Misask peatland mostly received inflow from northern (P_m15) and western (P_m13) piezometers, while at Cheinu, piezometers in the northern (P_c13 and P_c24) and western (P_c27) portions of the peatland indicated inflow to the ecosystem. At both sites, inflows driven by local topography were also found to occur in areas located close to downgradient peatland flow locations (e.g., P_m06-W_m08 and P_m09-W_m10 at Misask and P_c30-W_c22 at Cheinu). The only temporary hydraulic gradient inversions observed at the two sites were at P_m06-W_m08 and P_m15-W_m05 (Misask) and P_c30-W_c22 (Cheinu) and lasted a maximum duration of four consecutive days in August (2018 and 2019) and October (2018).

At Misask, the highest NHGs (absolute value) were observed in the southeastern (P_m06-W_m04) and eastern (P_m12-W_m05) portions of the peatland, close to the main diffuse outlet. Results from P_m09-W_m04 (south) and P_m12-W_m04 (east) also indicate that the peatland feeds the aquifer laterally in these areas. At Cheinu, a NHG was only found for P_c25-W_c28 (south) in the vicinity of the outlet. This NHG is markedly smaller than those observed at Misask. Considering that the peatlands developed in shallow depressions, it is interesting to note that topography is clearly not the only factor determining groundwater inflows to the Misask peatland. If that were the case, a PHG would be observed along its entire periphery. Topography is apparently a stronger driver of horizontal hydraulic gradients at Cheinu, with only one NHG of limited amplitude.

Three vertical hydraulic gradients (VHG) were calculated at the Misask peatland: P_m03-W_m16, P_m07-W_m05, and P_m11-W_m10. The P_m03-W_m16 pair probably provides the most reliable information, located only 6 m apart, whereas the other two pairs are located 25 m apart. The results showed VHG values of -0.17 m.m^{-1} (P_m11-W_m10), -0.10 m.m^{-1} (P_m03-W_m16), and 1.16 m.m^{-1} (P_m07-W_m05), indicating flow from the peatland to the underlying permeable mineral sediments in the northwestern and central-western parts of the ecosystem, and flow from the underling sediments to the peatland in its northeastern part. VHG values were relatively constant over time, with inversions observed over very short periods of only a few days.

Because almost no hydraulic gradient inversions were observed throughout the study period, horizontal Darcy flows were calculated for each section using average heads for all the well-piezometer pairs. Darcy inflows of 92 and 269 mm.yr^{-1} were calculated at Misask, and of 111 and 418 mm.yr^{-1} at Cheinu, with the minimum value estimated using average peat depth and the maximum value estimated using maximum peat depth at each site. Darcy outflows of 73 and 212 mm.yr^{-1} were calculated at Misask, and of 48 and 131 mm.yr^{-1} at Cheinu, again using average and maximum peat depths. At Misask, vertical outflows of 30 mm.yr^{-1} (P_m11-W_m10) and 18 mm. yr^{-1} (P_m03-W_m16) were estimated from the peatland to the underlying aquifer, and vertical inflows to the peatland of 116 mm.yr^{-1} were estimated at P_m07-W_m05 (no estimates of vertical flow were available at Cheinu). It is important to underline that due to some difficulties in installing piezometers in the mineral sediments and in the peat deposits, some well pairs might not be optimally located (e.g., P_c09-W_c10 or P_c15-W_c04).

Because flows in the lower southwestern portion of the peatland were mostly diffuse, outflows from the Misask peatland could not be measured. Outflows were able to be measured at the Cheinu peatland outlet. The average measured flow rate at this outlet was 730 m^3.d^{-1} (ranging between 371 m^3.d^{-1} in June 2019 and 1150 m^3.d^{-1} in September 2018), which is equivalent to 1812 mm.yr^{-1} (921 to 2855 mm yr^{-1}) when divided by the area of the peatland.

1.7.4 Model calibration

In both peatlands, the calibrated infiltration rate shows realistic values in term of the associated surface deposit and is similar to the range of P_{net} values of 414 to 450 mm.yr^{-1} mentioned above and assumed to correspond to infiltration rates. In the Misask model, the maximum calibrated infiltration (450 mm.yr^{-1}) is associated with the peatland where the top of the acrotelm, characterized by a high hydraulic conductivity, facilitates percolation through the porous upper layer. The lowest values (414 mm.yr^{-1}) were attributed to glacial till and thin till for which *in situ* observations have shown silt fractions that could induce slightly lower hydraulic conductivity. The washed till and fluvioglacial deposits are associated with intermediate values of 434 and 431 $mm.yr^{-1}$, respectively. For the Cheinu model, the maximum calibrated infiltration rate of 450 and 427 $mm.yr^{-1}$ were respectively estimated for the boulder field area and the peatland. Lower values were found in washed and glacial till (414 $mm.yr^{-1}$) and in fluvioglacial deposits (416 $mm.yr^{-1}$).

Calibrated K_h values for glacial till and fluvioglacial deposits are within field measured intervals (Fig. 7). At depths without measurements, calibrated K values correspond to the expected ranges for the different types of mineral deposits (Domenico et Schwartz, 1998) and show decreasing trends with depth, as expected in unconsolidated sediments. As mentioned above, K_h values for the top 1 m of peat were set to measured averages with a single value for each model layer (not calibrated). Below 1 m, peat K_h were set as gradually decreasing to reach 1×10^{-8} $m.s^{-1}$ at maximum peat depth. Using these K values, the models were calibrated satisfactorily with mean errors close to zero, mean absolute errors of 0.141 and 0.154 m for Misask and Cheinu respectively, and low RMSE values of 0.179 and 0.196 m respectively (Fig. 8).

1.7.5 Water budget

At the Misask peatland, the model suggests that 1% of the calibrated infiltration rate does not reach the saturated zone and is rejected from the cell as runoff. The resulting peatland recharge (450 $mm.yr^{-1}$) dominated the peatland inflows (78% of water input) while the groundwater contribution to the peatland of 130 $mm.yr^{-1}$ represented 22% of the peatland inflows. Groundwater inflows were distributed all around the peatland, with upgradient and downgradient portions receiving respectively 63 and 67 $mm.yr^{-1}$ (Table 1). Results show that subsurface runoff from the peatland

($495\ mm.yr^{-1}$) is the largest outflow from the peatland (85% of water output), while groundwater output through the saturated zone represents 15% of the water output. Groundwater outflow (total $87\ mm.yr^{-1}$) is equally exfiltrated around the peatland. Groundwater exchange flux resulted in a net inflow of $43\ mm.yr^{-1}$.

At the Cheinu peatland, the unsaturated zone rejects 0.2% of the applied infiltration rate through runoff leading to a recharge rate of $427\ mm.yr^{-1}$ (Table 1). The recharge represents 44% of water input to the Cheinu peatland while the proportion of inflow from the surrounding aquifer is 56% ($545\ mm.yr^{-1}$). The upgradient portion of the peatland received 86% of the total groundwater inflow, the rest of groundwater inflow being fed to the peatland by the aquifer through areas located in its downgradient portion. Water outputs were dominated by subsurface runoff which accounted for 74% of total outflows while groundwater flow from the peatland to the aquifer represented 26% of total outflows (Table 1). Outflows were relatively uniformly distributed all around the peatland. Compiled groundwater flow result in a net inflow of $291\ mm.yr^{-1}$.

Over the entire models, the unsaturated zone shows that 7.3% and 2% of the infiltration rate is rejected as runoff for the regional Misask area and for the regional Cheinu area, respectively. For both models, the spatial distribution of the rejected recharge and leakage (discharge) are located mainly near stream and wetland cells located in topographic depressions. The water budget input of Misask and Cheinu models shows a dominant inflow from constant heads, i.e. 63% and 59% respectively, and a recharge input of 37% and 41%. Groundwater subsurface runoff represents 69% (Misask) and 56% (Cheinu) of their water output while 31% and 44% is removed through constant head cells in the Misask and Cheinu models, respectively. Constant head cells, related to lakes and rivers, were divided in two categories 1) boundary conditions (BC) i.e. constant over the entire grid depth, and 2) lake cells (LC) located within the simulated area i.e. constant head on the first grid layer. The

Misask models showed BC representing 35% of inflow and 17% of outflow while LC represented 65% of inflow and 83% of outflow associated with constant heads flux. The Cheinu model showed 35% of inflow and 17% of outflow for BC and showed 65% of inflow and 83% of outflow.

1.7.6 Sensitivity of recharge

Simulated changes in recharge (-50%, -20% and +20% change in recharge) suggested that increased recharge leads to increased heads at the two sites, and vice-versa for decreasing recharge. In both cases, head changes in response to recharge changes are smaller in the organic deposits than in the mineral sediments (Fig. 9). However, recharge has a larger impact on heads at the Cheinu peatland than at the Misask peatland. Head changes vary from -1 to +0.3 cm for the Misask peatland and from -9 to +2 cm at the Cheinu peatland. Head changes vary from -4 to +0.9 cm for the Misask mineral deposits and from -20 to +5 cm for the Cheinu mineral deposits.

At the Misask site, exchanges with the aquifer showed limited sensitivity to recharge (-50, -20 and +20%) with variations of -26, -8 and +7% for fluxes from the aquifer to the peatland (initial value of 130 $mm.yr^{-1}$) and -27, -9 and +7 % for fluxes from the peatland to the aquifer (from initial value of 87 $mm.yr^{-1}$) (Fig. 9a). Subsurface runoff from the peatland to surface flow was more sensitive, with variation of -48, -19 and +19 % for the three recharge scenarios respectively (initial value of 495 $mm.yr^{-1}$). At the Cheinu site, the recharge change generated larger variation in fluxes from the aquifer to the peatland than at Misask, with variations of -33, -12% and +11 % (from initial value of 545 $mm.yr^{-1}$). Changes in fluxes from the peatland to the aquifer were also larger at Cheinu with variation of -44, -17 and +17 % (initial value of 255 $mm.yr^{-1}$) (Fig. 9b). Subsurface runoff from the peatland to surface flow was less sensitive to recharge changes than at Misask with -39, -15 and +15% (initial value of 719 mm.yr^{-1}).

At both sites, recharge change did not cause significant modifications in the dominating water budget component (Fig. 10). As recharge rate variedd from -50% to +20%, both peatland water budgets showed an increase of recharge contribution (70 to 80% at Misask; 37 to 46% at Cheinu), leading to a reduction of groundwater inflow to the peatland (30 to 20% at Misask; 63 to 54% at Cheinu). However, the proportion of outflow showed reversed trends between the two sites. Misask showed an increased proportion of subsurface runoff (from 80 to 85%) with a lowering contribution of groundwater outflow from the peatland (from 20 to 14%) (Fig. 10a). Cheinu had only a slightly decreasing proportion of subsurface runoff percentage (from 75 to 73%) and an increased proportion of groundwater outflow (from 25 to 27%) with the three recharge scenarios (Fig. 1b).

1.8 Discussion

1.8.1 Measured hydraulic conductivities

Hydraulic conductivity results from the peat cores showed that despite having similar time of peat initiation, regional climate exposure and geological forming context as well as geographical proximity, local factors influencing peat accumulation and decomposition rates can lead to different anisotropy profiles and therefore drive different flow dynamics. At the Misask site, the entire top 1 m of peat has $K_h > K_v$, indicating dominant lateral flow. At the Cheinu site, the vertical anisotropy resembles that of Rosa et Larocque (2008), with $K_h < K_v$ in deeper layers (> ~75 cm) showing the importance of vertical flows below the acrotelm.

The observed decreasing K trends with depth and the range of values compared well with those of other studies using the MCM method (Beckwith *et al.*, 2003a; Surridge *et al.*, 2005; Rosa et Larocque, 2008; Bourgault *et al.*, 2018). However, Rosa et Larocque (2008) and Bourgault *et al.* (2018) found higher range of K within the first 30 cm of peat in southern Quebec peatlands i.e. averaging 10^{-1} to 10^{-4} m/s. Also, fast decreasing rate of K reaching four orders of magnitude (Bourgault *et al.*, 2018) were distinct of result from Surridge *et al.* (2005) and Beckwith *et al.* (2003a), studying northern peatlands located at similar latitudes in the United Kingdom, who measured lower range of K in the acrotelm and compared well with result from the present study i.e. averaging 10^{-4} to 10^{-5} m/s. This could be linked to many local factors, like vegetation species, microtopography, WTD and weather conditions at the core sampled location. Nevertheless, this suggests a possible northern-southern trend in accumulation/decomposition processes leading to more decomposed acrotelm conditions in northern region.

The vertical profile of hydraulic conductivity measured at both sites probably enhances their capacity to buffer (Boelter, 1968; Quinton et Roulet, 1998; Price et Waddington, 2000) higher wetness with redistribution of excess water between flow within the peatland (Fraser *et al.*, 2001; Rezanezhad *et al.*, 2016; Quillet *et al.*, 2017) and subsurface runoff mechanisms (Kværner et Kløve, 2008) that represent the dominating outflows at the two sites (75%: Misask ; 85% Cheinu). Larger pores of near surface acrotelm decrease potential rise of water table and higher K provided greater runoff through living moss. Inversely, water table drawdown is limited by thinner pore and low flow rate of the catotelm.

1.8.2 Measured water table depths and heads

The average WTDs of central peat wells (W04: -18 cm and W28: -15 cm), recorded during the studied period, are shallower than the average value of -27 cm reported by Bourgault *et al.* (2019) for seven peatlands in southern Quebec (Canada) during 2015. In this study, the short and cool growing season could limit evapotranspiration and contribute to higher water tables within the peatlands. Multiple mechanisms have been shown to participate in the resilience of peatlands hydrology such as near-surface water table enhancing root water availability, reducing peatland vegetation dependence on precipitation, and providing greater resistance to dryer meteorological conditions (Farrick et Price, 2009).

For the two peatlands, the results showed more WTD variability in the thinner peat areas at the forested fen margins i.e. where connectivity to the surrounding aquifer is highest (Fig. 4). Stable near-surface WTD were obtained in the thicker peat areas close to the peatland center. Morris et Waddington (2011) have shown that groundwater-dependent peatlands present near-surface water tables and this was confirmed especially for ombrotrophic zones by (Lukenbach *et al.*, 2015). Waddington (2018) have shown that areas with limited organic matter (30-50 cm) experience two- or three-times larger water table fluctuations than areas with thicker peat and are thus more susceptible to drying conditions. This author also showed that lower peat hydraulic conductivity can induce shallower water table in upgradient aquifer. Margin zones are frequently characterized by a lagg transition margin with minerotrophic vegetation (Pellerin *et al.*, 2009; Howie et Tromp-van Meerveld, 2011; Langlois *et al.*, 2015; Paradis *et al.*, 2015) where groundwater inflow from the aquifer can be important (Glaser *et al.*, 1997). Ferlatte *et al.* (2015) found similar spatialization of connectivity which peaked at the peatland margin and was attenuated or delayed in the central zone.

1.8.3 Simulated flows

At the Cheinu peatland, the simulated subsurface runoff rate, which corresponds to surface outflow at the outlet, is much lower than the measured flow rates at the peatland outlet. It could be because of the sporadic flow rate measurements obtained during field campaigns when rain events or snowmelt occurred. However, the minimum measured outflow of 371.1 m^3/day is of the same order of magnitude as the simulated value of 246.2 m^3/day (719 $mm.yr^{-1}$). Outflow rates were not measured at the outlets of the Misask peatland because flows are mostly diffuse, and the simulated values could not be validated. Groundwater inflows and outflows from Darcy calculation showed a very large range of values. This approach can lead to various results, with a tendency to generate lower value (Hunt *et al.*, 1996) and it had seems to be challenged by the complexity of the peatland shape, despite the availability of observation points. However, the simulated groundwater flow remained between the Darcy's flows intervals obtained.

1.8.4 Contrasted water budgets

The results suggest that the complexity of peatland hydrology resides in local-scale factors. Despite the similarities between the two peatlands (mineral substratum, peat hydraulic conductivity, initiation age, climate exposure), water budget components and responses to recharge changes differ.

The Misask water budget was dominated by recharge and subsurface runoff and includes smaller flows from and to the aquifer than the Cheinu peatland. Other studies in natural (Fraser *et al.*, 2001; Tardif *et al.*, 2014) and restored peatlands (Van Seters et Price, 2001; Shantz et Price, 2006; Tiemeyer *et al.*, 2006) in north America and Europe showed similar fluxes. The Cheinu peatland water budget was characterized by groundwater inflow (56%) similar to the inflow from recharge. This contrasted from Misask results, but compared well with results from Bourgault *et al.* (2014) and Levison *et al.* (2014) who found groundwater input representing up to 50% of the water budget in partly ombrotrophic peatlands in southern Quebec (Canada). The contrast between the two peatlands could be due to their position within the watershed, with the Cheinu peatland being located close to the outlet and thus accumulating more flow from upgradient areas.

At Misask, the relatively small groundwater-peatland exchanged fluxes were uniformly distributed around the peatland, suggesting that the complex shape of the peat basin combined with changing local topography allows inflows to occur from downgradient areas as well as outflow to take place in upgradient areas. These results underline the critical influence of the local flow system on aquifer-peatland exchanges (Reeve *et al.* (2001) and the importance of converging flow in the peatland margin (Bourgault *et al.* (2019). At Cheinu, the simulated groundwater inflows enter mostly through the upgradient area, probably due to the higher slope in this part of the watershed. Nevertheless, groundwater outflow showed the same uniformed spatial distribution as that at Misask.

Subsurface runoff was highlighted as the dominant outflow component at both sites (85% at Misask and 74% at Cheinu) despite the high hydraulic conductivity in fluvioglacial deposits bordering downgradient margins at the Cheinu site. This reinforces the importance to include some representation of hydrological processes such as subsurface runoff when simulating peatland water budgets. Tardif *et al.* (2014) documented that runoff dominated the water budget term in the northern fen systems of La Grande River region, Quebec. The higher proportion of the outflow occurring as subsurface runoff at the Misask peatland (85% compared to 74% at Cheinu) is consistent with its steeper peat topographic gradients (1% at Misask compared to 0.5% at Cheinu).

The contrasted simulated groundwater inflows (130 and 545 $mm.yr^{-1}$ at Misask and Cheinu peatlands, respectively) and groundwater outflows (87 and 255 $mm.yr^{-1}$) compare well to other studies from northern peatlands in Quebec. Using a hydrological model in a northern James Bay watershed (northern Quebec), Jutras *et al.* (2009) simulated higher groundwater inflows (964 $mm.yr^{-1}$). Drexier *et al.* (1999) have calculated groundwater outflows of an Illinois fen as a water budget residual and obtained values between 2477 $mm.yr^{-1}$ (84% of total inflow) and 4685 $mm.yr^{-1}$ (88% of total inflow) during a two year study. In a southern Quebec peatland, Levison *et al.* (2014) found simulated average groundwater inflow of 300 $mm.yr^{-1}$ using MODFLOW while obtaining 491 $mm.yr^{-1}$ with a Darcy flow calculation. Those results support the quality of this type of Darcy flow cross-checking method. On the Villeroy ombrotrophic peatland (southern Quebec), Larocque *et al.* (2013) also used the Darcy approach and obtain a groundwater inflow of 283 $mm.yr^{-1}$ and an outflow of 40 $mm.yr^{-1}$ which compares well with Misask peatlands aquifer interaction. However, this work has shown that despite the number of piezometers available, they do not provide sufficient data to provide a complete understanding of lateral and vertical flows exchanged between aquifer and peatland.

1.8.5 Driving factors of peatland-aquifer exchanges

It remains a challenge to compare peatland water budgets since they are affected by many inherent characteristics. In this study, the contrasting result differencing the Misask and the Cheinu sites could be explained by different local factors. The Cheinu site has developed in a deeper topographic depression (6.2 m at Misask and 11 m at Cheinu) with steeper upgradient topographic slope (~6.7%). This is supported by results from Hare *et al.* (2017) who identified the morphology of the peat basin as a factor influencing discharge from the aquifer to the peatland. These authors showed that peatlands with steeper margins are more likely to develop additional preferential discharge zones (pipe flow) facilitating exchanges through peatland bottom.

The Cheinu peatland aquifer showed higher hydraulic conductivity than that of the Misask peatland, including the presence of distinctive fluvioglacial deposits in the downgradient area. The direct impact of hydrogeological setting surrounding peatlands has previously been linked to differences in peatland-aquifer connections water budget (van der Kamp et Hayashi, 2009), flow directions (Reeve *et al.*, 2001) and WTD within the peatland (Winter, 1999, 2001; Dimitrov *et al.*, 2014; Lukenbach *et al.*, 2015; Bourgault *et al.*, 2019). Vertical peat-aquifer exchanges are highly impacted by the contrast in peat and aquifer K values (Reeve *et al.*, 2000). In this study, vertical exchanges could be quantified only at the Misask peatland, because of the absence of piezometer nests at the Cheinu peatland.

1.8.6 Vulnerability to recharge changes

Results suggest that the Misask peatland heads are not very sensitive to changes in recharge and that aquifer heads are only slightly more sensitive. Subsurface runoff flows change at the same rate as recharge, as those changes in the peatland water appear to occur mostly in the surficial layers and are mainly reflected as runoff. At the Cheinu site, recharge changes led to marked peatland water table changes and even higher changes in aquifer heads. At this site, changes in recharge induce proportionately smaller changes in outflow to the aquifer than to surface leakage. Within the hydrogeological and topographic context of the Cheinu peatland, decreasing recharge causes greater water table lowering than at the Misask peatland which is located in a flatter landscape. These results suggest that the local aquifer system can react very differently to recharge changes. Modification of recharge rates due to climate change could have different impacts on peatlands with different

dominating water budget components. For example, peatlands with larger inputs from aquifers have been shown to have higher thermal regulation (Hare *et al.*, 2017) with a warming effect in March and a cooling effect in July. This is known to contribute to biodiversity (Hunt *et al.*, 2006; Parish *et al.*, 2008) and could increase resilience of peatlands in changing recharge conditions. Conversely, warmer temperature in vertical peat profiles, a condition which could be exacerbated in peatlands with limited groundwater inflow, could also lead to conditions favorable to methane production occurring deeper in the peat profile (McKenzie *et al.*, 2007).
Climate change is expected to induce more frequent extreme conditions as well as a long-term increase of annual temperature and precipitation (higher quantity and intensity) and an extended growing season influencing evapotranspiration rates. However, the scientific literature is still scarce on future trends of groundwater recharge in northern regions.

Those WTD changes can trigger vegetation changes (Waddington *et al.*, 2015) with succession shifts as different sphagnum species thrive in specific WTD conditions (Bobrov *et al.*, 2002) and influence their sustainability (Bubier, 1995; Silvola *et al.*, 1996). Although peatland ecosystems are highly resilient, it has been shown that northern peatlands have experienced marked ecohydrological changes in the past (van Bellen *et al.*, 2018; Robitaille *et al.*, 2021).

Studies on northern peatlands have shown how water table drawdown can lead to a decrease of methane (CH_4) emissions (Roulet *et al.*, 1992; Moore et Roulet, 1993; Roulet *et al.*, 1993; Nykänen *et al.*, 1995; Nykänen *et al.*, 1998) and an increase of carbon dioxide (CO_2) emission (Moore et Knowles, 1989; Moore et Roulet, 1993). Lowering water levels can modify the peatland carbon cycle (Holden, 2005a; Mitsch *et al.*, 2009), by enhancing decomposition and depleting accumulation rates (Mitsch *et al.*, 2009). Evapotranspiration, relying on water availability, can also decrease with water table drawdown, i.e. especially below average root depth, and fens are expected to be more impacted than bogs (Bridgham *et al.*, 1999). Many authors have also documented irreversible peat consolidation affecting peat hydraulic properties and peat hydrology as a result of lowering of the water table in drained peatlands (e.g., Whittington et Price (2006)).

In increasing water table conditions, major shifts in peatland ecology i.e. from peatland to wet-herbaceous environments, have been estimated to take place after 1500 to 1000 years of exposure to wetter condition (Belyea, 2009). However, the timescale in which aqualysis can occur can be much shorter. Dissanska *et al.* (2009) have shown with 50 years of recorded data that signs of aqualysis are evident in the James Bay region, Quebec while Robitaille *et al.* (2021) showed that it ceased in Misask and Cheinu peatlands. Those contradicting results show how changes in peatland wetness can differ regionally. This study provided insight of peatland resistance to increasing regional wetness with a scenario of +20% recharge rate, as a surrogate for potential aqualysis conditions. Results led to contrasting water table rise magnitude suggesting the importance of inflow source to the water budget as Misask mostly governed by exchange between recharge and seepage, showed low level rise while Cheinu, which mainly groundwater-fed, showed greater rise. Finally, seepage representing main output source suggest great capacity of those peatlands to redirect excess through surface flow mechanism and limit aqualysis.

1.8.7 Limitations of simulation results

This work has shown that steady-state groundwater flow modeling is a useful approach to quantify the water budget and estimate the sensitivity of peatland hydrosystems to changes in meteorological conditions. Although they represent frost-free conditions, the models developed here provide means to evaluate the possible impact a changing climate could have on the peatland water budget and average WTD.

However, to understand intra-year variations of storage and discharge through subsurface runoff and groundwater flow (Oswald *et al.*, 2011) linked to changes in WTD (Quinton et Roulet, 1998; Spence et Woo, 2003; Kværner et Kløve, 2008; Carrer *et al.*, 2015), and to identify temporary local flow reversals (Levison *et al.*, 2014)), a transient-state model would be needed. Transient conditions anticipated in a changing climate are also expected to induce warmer winter temperatures and shorter winter duration (Ouranos, 2015) inducing possible recharge conditions in late fall and early spring, as well as possible winter recharge events (Jyrkama et Sykes, 2007; Okkonen et Kløve, 2011). Increasing rainfall intensity (Sillmann *et al.*, 2013; Ouranos, 2015) could also modify recharge patterns throughout the year (Fu *et al.*, 2019; Dubois *et al.*, 2021). These short-term intra-annual changes in the peatland water budget could also lead to long-term (multi-year to multi-decade) hydrological changes that could lead to fundamental modifications in peat accumulation.

The models developed in this study include the UZF package previously used by Feinstein *et al.* (2020) to delineate fens distribution (discharge area) in the Mukwonago River basin located in southeastern Wisconsin. This package provides an alternative method to the modeler-controlled approach of blanketing a peatland with drains to simulate the near surface runoff dynamic (Halford, 1998). In this work, rejected recharge was considered negligible as it represents 1% (Misask) and 0.2% (Cheinu) of the imposed infiltration rate (thus recharge is almost equal to infiltration). The UZF package simulates leakage, considered here as subsurface runoff, information that is not otherwise in MODFLOW. It has been shown elsewhere that peat hydraulic conductivities at the peatland margins are lower than those in the central portion (Howie et Tromp-van Meerveld, 2011). Lapen *et al.* (2005) have shown that this can limit lateral drainage to the aquifer and contribute to raise the groundwater table at the center of a bog. The models developed in this study do not take into account this lateral heterogeneity of peat properties. Because lateral heterogeneity was not well defined at the two sites (only three peat cores), this heterogeneity could not be considered here. The resulting model simplification is considered to have relatively limited impacts on the results. Nevertheless, it could have influenced or modified peatland-aquifer exchanges, as well as flow direction and magnitude within peat deposit.

Finally, it must be underlined that although the calibrated models represent relatively well the available data, non-negligible uncertainty remains in the actual hydrostratigraphy of the two peatlands. It is yet unclear if deeper groundwater contributes to the peatlands inflow and in what proportion. Lack of available geological data is often a constraint in remote northern areas and in this study, the influence of deeper conditions could not be explored.

1.9 Conclusion

The objectives of this research were to quantify aquifer-peatland hydrologic connectivity and identify controlling factors in two pristine peatlands in north-central Quebec (Canada). Considering that the two peatlands have a similar history of peat initiation, are exposed to the same climatic conditions, and have developed on the same glacial deposit landscape, contrasting measured and simulated results emphasized the important role of local factors on their hydrogeological behavior. Measured hydraulic conductivities showed different trends of vertical anisotropy for the two peatlands, but WTD analysis showed stable near-surface levels within the central ombrotrophic areas and deeper water tables in marginal fens area at the two sites.

At the Misask peatland, inflows were dominated by recharge and the contribution of groundwater was relatively limited. Outflows were mostly exfiltrated by surface seepage and outflow to the aquifer was small. The Cheinu site had a distinct water budget, with inflows dominated by groundwater and most of outflow as seepage. The Misask peatland shows limited sensitivity to changes in recharge, while the Cheinu peatland reacts more markedly to the same changes. It is hypothesized that the greater groundwater dependence of the Cheinu site can be explained by the peatland geomorphological setting, by the size of its contributing area and by its position in regional flow. These factors necessarily influence the hydrological functions of the two peatlands and their sensitivity to long-term recharge changes. This work underlined the importance of the peatland watershed and groundwater contributing area to assess its resilience to hydrometeorological changes.

More work needs to be done to confirm results from this work, notably the influence of transient processes which could trigger feedback processes leading to either stable long-term conditions or in contrast drastic changes in hydrological conditions leading to changes in peat accumulation or degradation. To achieve this, long-term transient-state simulations including flow, vegetation and carbon dynamics need to be performed.

ACKNOWLEDGEMENTS

The authors wish to acknowledge the funding from National Science and Engineering Research Council (NSERC), and private funding partners, Nemaska Lithium and the Stornoway for the essential support that made this research project possible. We are also grateful for the participation of Michelle Garneau, Éric Rosa, Marc-André Bourgault and Paul Morris from the extended research team, for they multidisciplinary point of view from the field through the discussion process.

TABLES AND FIGURES

Table 1 : Simulated water budget flow rates ($mm.yr^{-1}$) for the Misask and Cheinu peatlands.

	Flow rate in $mm/.yr^{-1}$(% of water budget)			
	Misask		Cheinu	
Unsaturated zone flux	**Inflow**	**Outflow**	**Inflow**	**Outflow**
Infiltration through peatland surface	454.4		428.1	
Rejected recharge		-4.6		-1.0
Groundwater flux				
Recharge to water table	449.8 (78%)		427.1 (44%)	
Seepage		-494.6 (-85%)		-718.8 (-74%)
Aquifer exchanges				
Upgradient zone	62.6	-43.9	470.9	-133.3
Downgradient zone	67.1	-43.2	74.5	-121.6
Total	129.8 (22%)	-87.1 (-15%)	545.4 (56%)	-254.8 (-26%)
Total budget flow	579.6	-581.7	972.4	-973.6

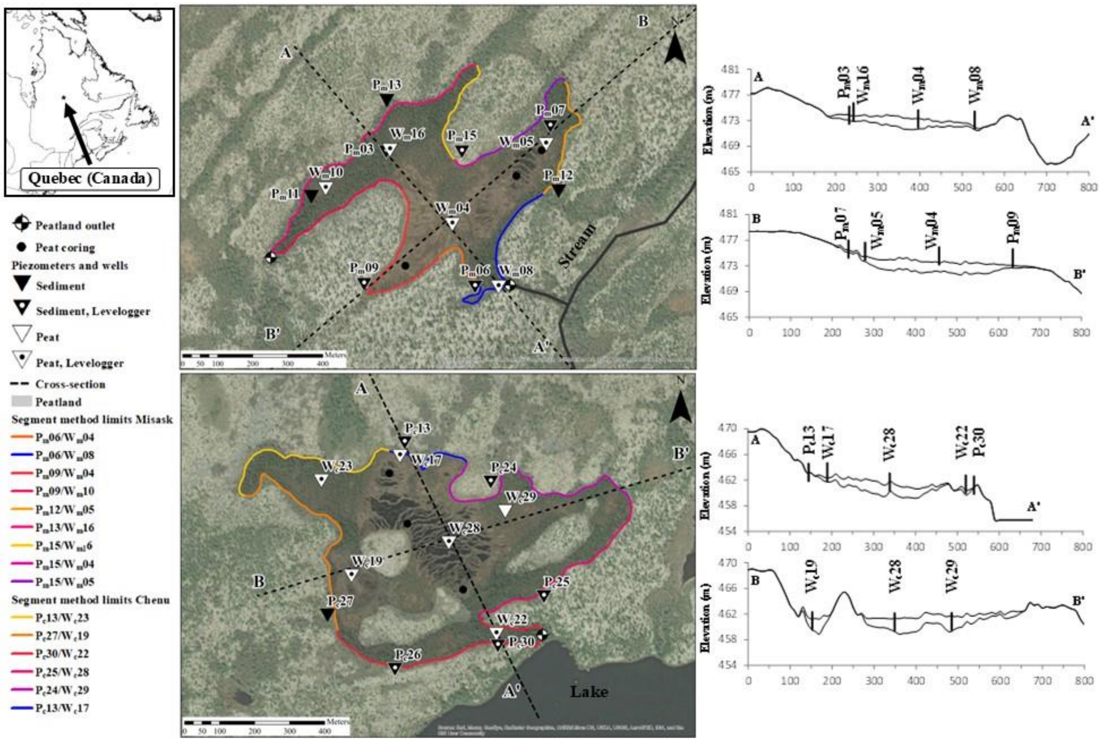

Figure 1 : Location of the studied peatlands with locations of all wells (W) and piezometers (P), and cross sections with only monitored observation point at a) the Misask peatland and b) the Cheinu peatland.

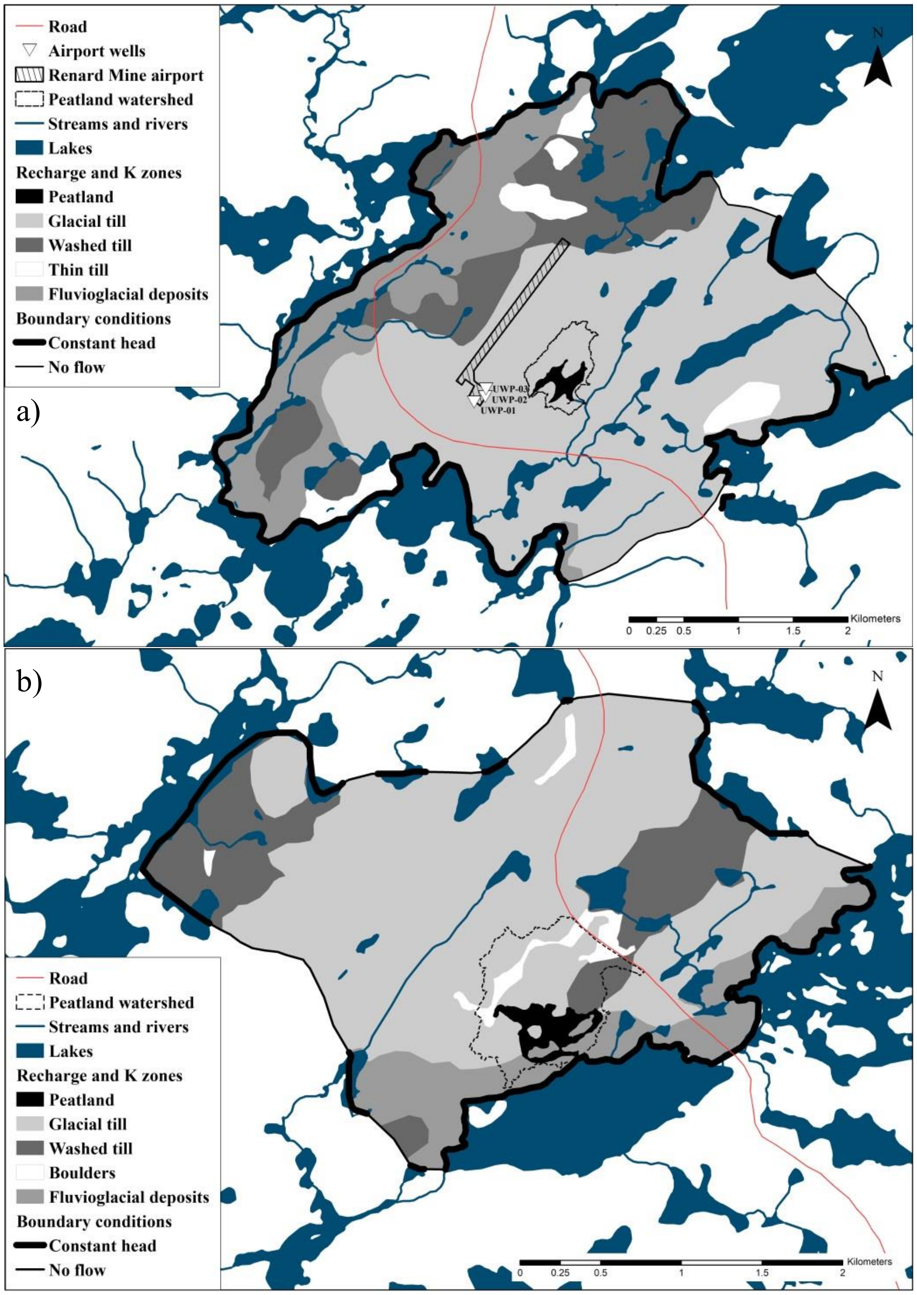

Figure 2 : The conceptual groundwater flow context, showing boundary conditions as well as recharge zones and hydraulic conductivity zones for a) the Misask peatland and b) the Cheinu peatland.

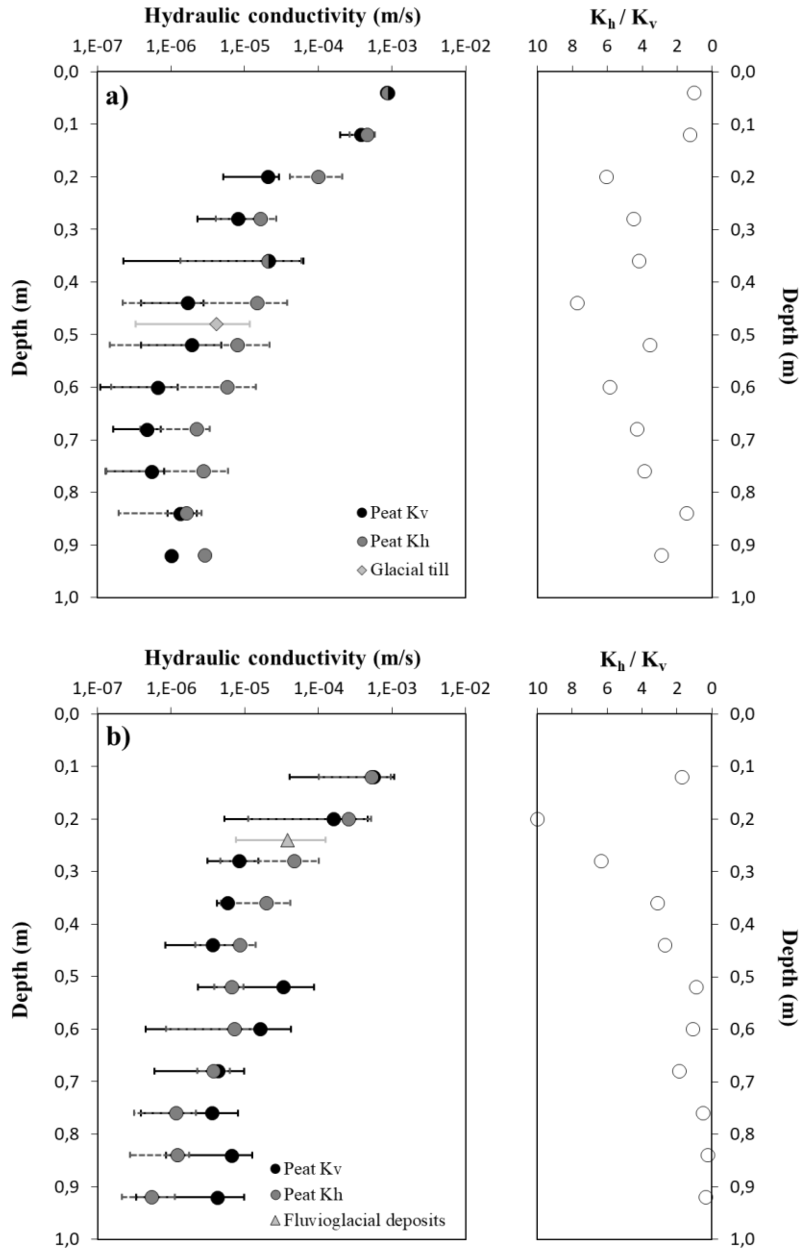

Figure 3 : Minimum, maximum and average measured hydraulic conductivities in the three cores for a) the Misask peatland and b) the Cheinu peatland. The left panels illustrate vertical hydraulic conductivities (K_v) and horizontal hydraulic conductivities (K_h) for the peat (estimated with the MCM method) and for mineral deposits (estimated with slug tests). The right panels illustrate K_h/K_v ratios.

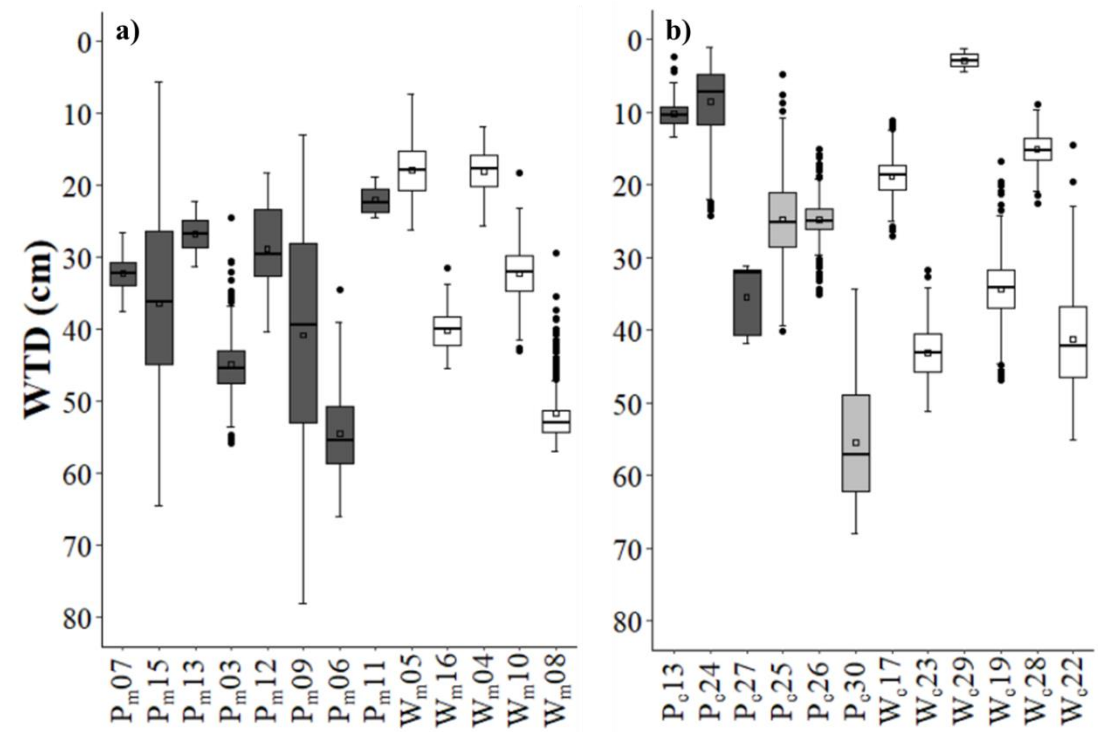

Figure 4 : Average daily water table depths (WTD) from June to October (2018 and 2019) for monitored wells (peat) and piezometers (mineral deposits), and for unmonitored stations (P_m13, P_m12, P_c27 and W_c29) used in the Darcy flow calculation for a) the Misask peatland and b) the Cheinu peatland. The squares indicate mean values.

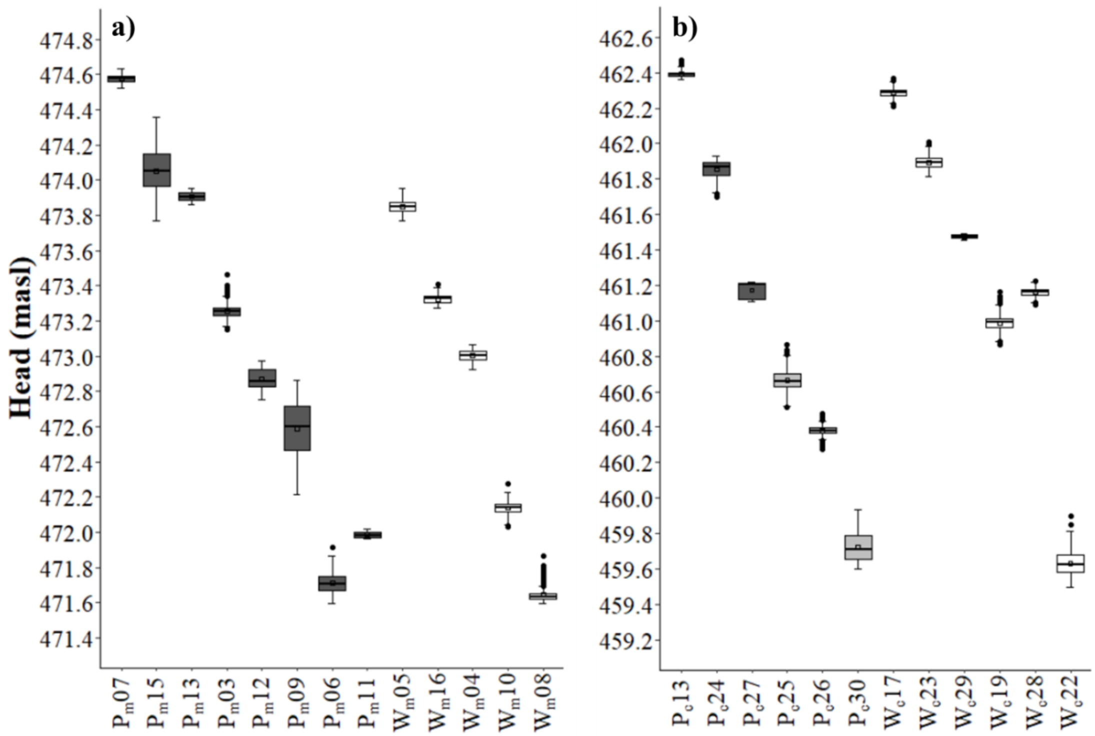

Figure 5 : Average daily heads from June to October (2018 and 2019) for monitored wells (peat) and piezometers (mineral deposits) and for unmonitored stations used in the Darcy flow calculation (P_m13, P_m12, P_c27 and W_c29) for a) the Misask peatland and b) the Cheinu peatland. The squares indicate mean values.

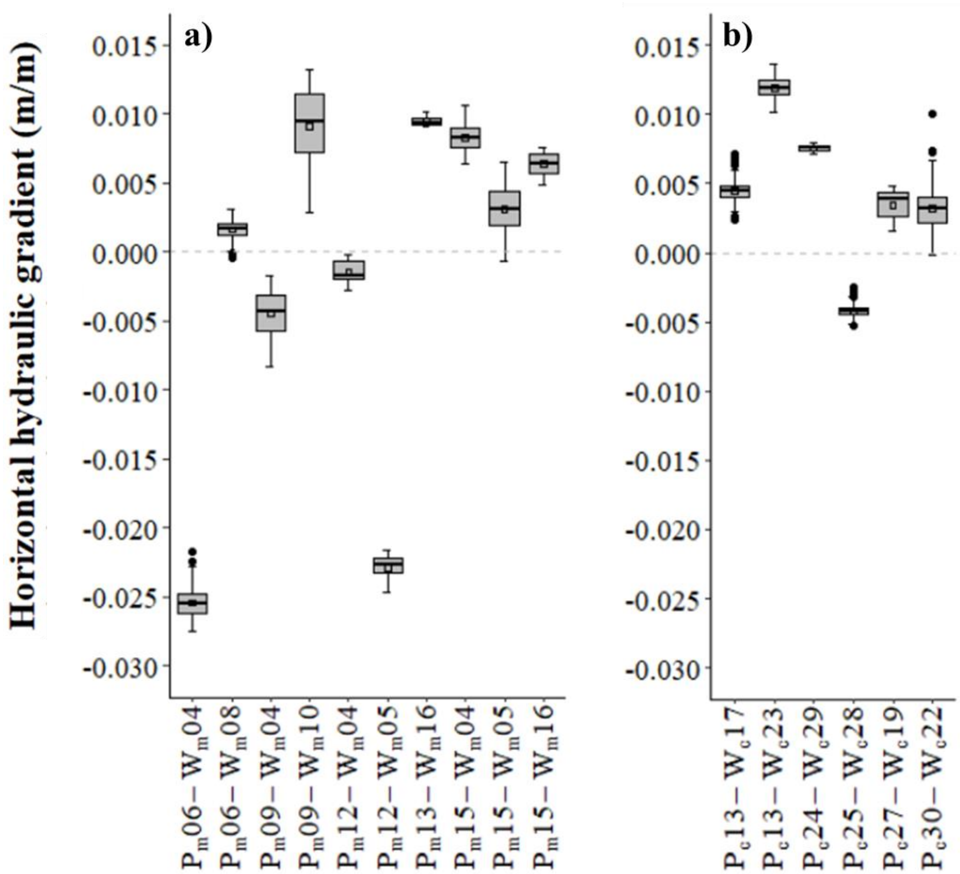

Figure 6 : Horizontal hydraulic gradient between stations used in the Darcy flow calculation for a) the Misask peatland and b) the Cheinu peatland. Positive values indicate flow from the aquifer to the peatland and negative values indicate flow from the peatland to the aquifer. The squares indicate mean values.

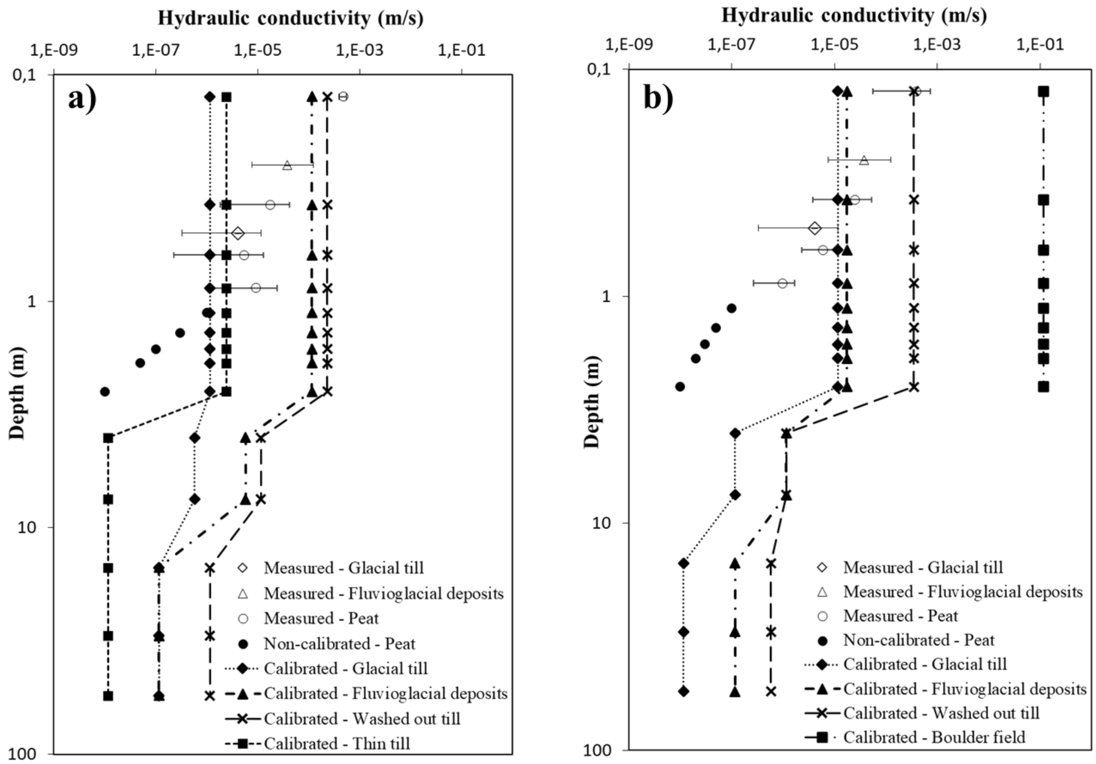

Figure 7 : Depth profiles of measured and calibrate horizontal hydraulic conductivities for a) the Misask peatland, b) the Cheinu peatland.

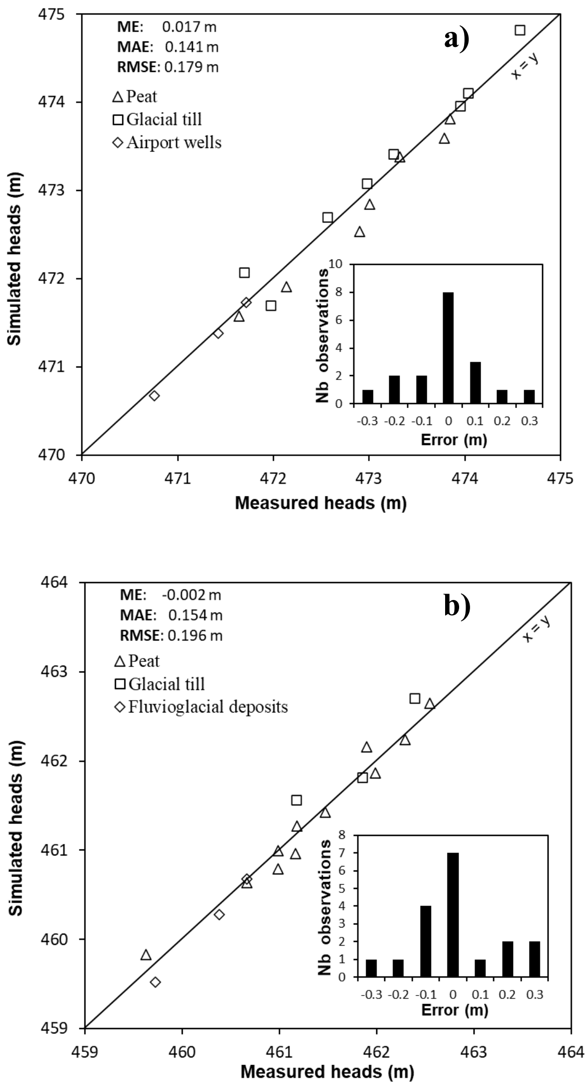

Figure 8 : Measured and simulated steady-state piezometric heads with associated error statistics for a) the Misask peatland and b) the Cheinu peatland.

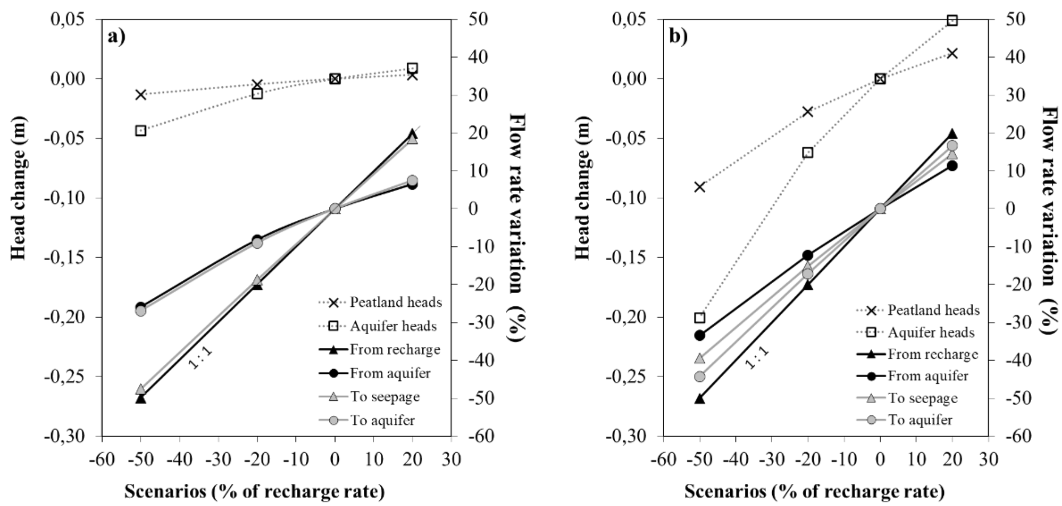

Figure 9 : Effect of climate change scenarios on average simulated peatland heads, mineral heads and water budget flow components at a) Misask peatland and b) Cheinu peatland.

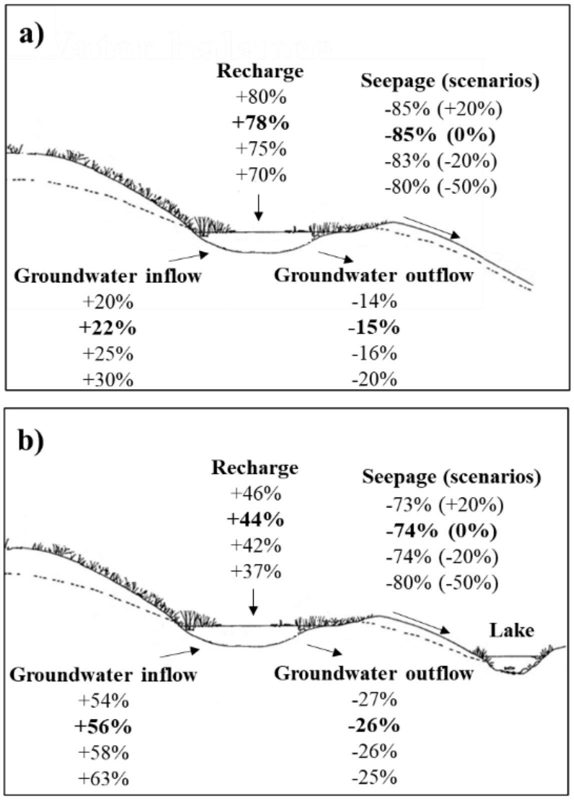

Figure 10 : Water budget components for the recharge rate scenarios i.e. +20%, 0%, -20% and -50%, for a) the Misask and b) the Cheinu peatlands.

CHAPITRE III

CONCLUSION GÉNÉRALE

Ce projet de maîtrise s'inscrit dans une démarche de développement des connaissances sur le régime hydrique des tourbières des régions nordiques circumpolaires québécoises. Les sites étudiés sont similaires à plusieurs égards, notamment en termes d'âge depuis l'initiation de l'accumulation de tourbe, ils sont situés à proximité l'un de l'autre (15 km) et ils sont donc exposés au même climat régional et ils se trouvent dans le même contexte géologique glaciaire. Cependant, les résultats mesurés et simulés contrastés soulignent l'importance des facteurs locaux sur la connectivité aquifère-tourbière.

Les conductivités hydrauliques mesurées ont montré une décroissance marquée dans le premier mètre de tourbe aux deux sites, mais des anisotropies verticales différentes. De plus, l'analyse des profondeurs de nappe entre juin 2018 et octobre 2019 (sans couvert neigeux) a montré un niveau proche de la surface et stable dans les zones ombrotrophes centrales et des niveaux significativement plus profonds dans les zones des minérotrophes périphériques caractérisées également par une plus grande connectivité à l'aquifère. Ceci confirme les propriétés tampons des tourbières observées par de nombreux autres auteurs et pouvant contribuer à leur résilience aux changements hydrogéologiques environnants.

La modélisation numérique a fourni des outils quantitatifs afin d'estimer la composante du bilan hydrique associée aux échanges aquifère-tourbière ainsi qu'une analyse de sensibilité aux changements de recharges anticipés par les fluctuations climatiques futures. Les modèles MODFLOW en régime permanent ont été calibrés de manière satisfaisante avec les conditions hydrogéologiques actuelles. L’approche utilisée pour définir le modèle conceptuel a permis une première application à l’échelle locale, du module UZF (MODFLOW) s’étant révélée très prometteuse pour simuler la dualité hydrologie-hydrogéologique de ces écosystèmes complexes.

Sur le site de Misask, les apports en eau sont dominés par la recharge (78%), tandis que la contribution de l’eau souterraine n’est que de 22%. Ces débits sont principalement exfiltrés à partir de résurgences alimentant le ruissellement de surface (85%), tandis que 15% de l’eau sort de manière souterraine de la tourbière pour alimenter l'aquifère superficiel. Au site de Cheinu, les apports d'eau souterraine (56%) dominent les flux entrants, mais sont quasiment égalés en proportion par la recharge (44%). Les débits sortants par le ruissellement de surface sont encore une fois dominants (74%), tandis que les flux sortant de la tourbière vers l’aquifère sont plus importants qu’à la tourbière Misask (26%).

Une analyse de sensibilité du taux de recharge a été menée comme proxy des effets hydrogéologiques possibles des changements climatiques (-50%, -20% et + 20% de la recharge imposés sur le modèle calibré). Sur le site de Misask, les scénarios de recharge étudiés ont conduit à des changements plus faibles des niveaux de nappe moyens que sur le site de Cheinu. Ceci pourrait être expliqué par les caractéristiques distinctes de ce site, incluant le contexte topographique, les paramètres hydrogéologiques de l’aquifère (conductivités hydrauliques) et la morphologie du bassin tourbeux. Pour les deux sites, les niveaux dans l’aquifère adjacent se sont révélés plus sensibles à la recharge.

Les travaux futurs pourraient inclure une analyse de sensibilité des facteurs locaux; le contexte géomorphologique, la superficie du sous-bassin-versant, la position au sein du bassin versant contrôlant de l’écoulement régional, dictant le comportement de la nappe phréatique et du bilan hydrique. Plusieurs tests complémentaires peuvent être appliqués aux modèles numériques construits et calibrés dans le cadre de ce projet de recherche. Les modèles développés pourraient également être calibrés en régime transitoire à l’aide de l’ensemble des données horaires disponibles (chronique piézométrique, précipitation, ETP). L’ampleur de l’instrumentation de suivi des niveaux d'eau permettrait également la documentation de la réactivité de chaque site aux évènements pluvieux à l'aide d'analyse comparative des puits et piézomètres c.-à-d. le décalage des maximums, leur amplitude et leur distribution spatiale. Cela permettrait de mieux documenter l'effet des variations de précipitations liées aux changements climatiques (fréquence, intensité et quantité) sur les différents secteurs de la tourbière.

Cette étude a montré comment un changement futur de la recharge pourrait conduire à des modifications hydrologiques importantes dans les tourbières nordiques du Québec et pourrait subséquemment engendrer des perturbations de l'emmagasinement et du relargage du carbone. Ces travaux s'inscrivaient dans une initiative de développement des connaissances sur les milieux humides nordiques du Québec. Cette recherche fournit des outils facilitant ce type de quantification du bilan hydrique dans des contextes géo-climatiques similaires et ainsi améliore la compréhension de ces écosystèmes complexes. La méthodologie appliquée lors de ces travaux s'appuie sur des principes généraux qui peuvent s'adapter à des contextes géographiques ou climatiques différents.

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