Hydrologic Functioning of Glacial Moraine Landscapes Within Alberta's Boreal Plains

by

Craig E. Thompson

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Department of Earth and Atmospheric Sciences
University of Alberta

Abstract

Within the Boreal Plains of north-central Alberta, catchments situated within low permeability glacial terrain are composed of a mosaic of landscape units including ponds, peatlands, and upland aspen forest ecosystems within a sub-humid climatic zone where water deficit conditions are frequent. These ecosystems host ecologically and commercially significant habitat and natural resources; however, they are threatened by expanding anthropogenic development and climate change. Within this framework, characterization of the processes governing water movement within and between landscape units is paramount for proper management of existing ecosystems and restoration of disturbed landscapes.

Hydrologic data were collected over eleven years to evaluate hydrologic interactions occurring between landscape units. Two-dimensional numerical models were developed using the fully-integrated groundwater-surface water model HydroGeoSphere to evaluate key landscape features and processes that allow these ecosystems to persist within the sub-humid climate. Results show that dynamic interactions between the pond and peatlands are driven by precipitation and evapotranspiration, with pond and peatland water levels reflecting recent climatic trends. Limited hillslope contributions to the peatlands occur, indicating they are not required within this climatic setting for long-term maintenance. Instead, the peatlands conserve water within the landscape and supply it to adjacent landscape units. By contrast, the pond and the aspen forested hillslopes are dominated by high rates of evapotranspiration, and represent net water sinks within the landscape.

A two-dimensional numerical model was also developed using MODFLOW-SURFACT to quantify the effects of seasonal peatland freezing on water distribution and water table position through investigation of changes due to variations in peatland hydraulic conductivity and storage properties. Results indicate that seasonal freezing is expected to maintain higher water table conditions by restricting infiltration of snowmelt and spring precipitation, thereby supporting higher rates of spring evapotranspiration, with discharge at the peat surface as surface ponding and overland flow. Subsurface hydrologic connectivity between the peatland and pond is also restricted due to the lower hydraulic conductivity of frozen peat. The degree of influence of the frozen peat is dependent on the relative timing of snowmelt and peatland ice recession. Where sufficient

ice remains to prevent infiltration of spring meltwater and rains, less water may be available to the peatlands. This decrease in available water may have negative implications for growing season productivity and fire susceptibility, as well as hydrologic interactions with neighboring ecosystems.

A two-dimensional numerical model was also developed using HydroGeoSphere to assess the hydrologic impact of aspen harvesting. Study results indicate that aspen harvesting has limited impact on groundwater levels and stream flows. This outcome is because of the sub-humid climate, with low-frequency of large storms, large soil-moisture storage capacity of heterogeneous glacial materials, and high evapotranspiration rates of regenerating aspen. Despite an estimated increase in hillslope groundwater levels of up to 3 m, pond and peatland water levels increased by less than 0.3 m and were accompanied by increased stream flows of less than 10 mm/yr. However, groundwater level and stream flow predictions were sensitive to regenerating aspen evapotranspiration rates, which can be enhanced by appropriate harvesting techniques but may be reduced by climate change. These results are consistent with previous results for the Boreal Plains, but they differ from aspen harvesting studies conducted in other settings where appreciable increases in stream flows have been reported. This disparity highlights the need to consider the integrated response of the hydrologic system when evaluating impacts from disturbance and making comparisons between settings.

Two-dimensional numerical simulations were also conducted using HydroGeoSphere to predict potential climate change impacts for a range of projected scenarios. Results indicate peatland water levels may decline by up to 1 m; however, sensitivity simulations indicate that the decline in water levels may be moderated by several feedback mechanisms that restrict evaporative losses and moderate water level changes. In contrast, higher evapotranspiration losses from the aspen hillslopes are predicted to result in near-surface soils becoming increasingly drier. Thus, the aspen may frequently be water-stressed and increasingly susceptible to secondary maladies such as pests and disease. Reduced pond water levels are also predicted with the development of frequent ephemeral conditions in warmer and drier scenarios. Concurrent decreases in stream flow may further impact downstream ecosystems. Further research into the regional health and sustainability of Boreal Plains ecosystems is warranted.

Preface

Craig Thompson is the primary author of this thesis. Hydrologic data from the Utikuma Region Study Area (URSA) used to support studies within this thesis were collected by Dr. Kevin Devito and Dr. Carl Mendoza, along with students and research assistants operating under their supervision. Chapters 2, 4, and 5 have been published. Co-authors for each publication provided intellectual supervision and editorial comment.

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CHAPTER 1

Introduction

1.1. Alberta's Boreal Plains

The Boreal Plains ecozone extends from northeastern British Columbia to south-central Manitoba (Figure 1-1A). Within north-central Alberta, the Boreal Plains are composed of a mosaic of ponds, peatlands, and upland forest ecosystems (Petrone *et al.*, 2007) situated within a sub-humid climatic zone (Marshall *et al.*, 1999) where water deficit conditions are frequent (Devito *et al.*, 2005a). These ecosystems are both regionally and globally significant due to their importance as a large carbon store (Kleinen *et al.*, 2012), spatially variable influence on climate (Krinner, 2003), source of seasonal habitat for migratory birds (Smith and Reid, 2013), and source of commercial lumber (David *et al.*, 2001). However, they are under significant development pressures from expanding resource extraction (Devito *et al.*, 2012) whose impacts are expected to be further compounded by climate change (Cerezke, 2009). Consequently, characterization of the hydrologic interactions and linkages between these ecosystems is of pressing importance for sustainable long-term management of existing ecosystems and restoration of those that have already been disturbed within the framework of a changing climate.

The Boreal Plains region is characterized by a sub-humid climate with synchronized growing season peaks in precipitation and evapotranspiration, low topographic relief, and thick heterogeneous glacial deposits (Devito *et al.*, 2017). Within this framework, evapotranspiration is the dominant growing season flux in open water and upland forest ecosystems (Petrone *et al.*, 2007; Zha *et al.*, 2010; Brown *et al.*, 2014); however, evapotranspiration is limited in peatlands by water table depth and internal feedback mechanisms (e.g., surface desiccation) that restrict water loss (Petrone *et al.*, 2007; Waddington *et al.*, 2015; Schneider *et al.*, 2015). As a result, runoff yields are typically low (Buttle *et al.*, 2009) and these low yields are correlated to the prevalence of water-conserving peatlands (Devito *et al.*, 2005a; Prepas *et al.*, 2006; Devito *et al.*, 2017). Due to the high evapotranspiration demand during the growing season, groundwater recharge is primarily dependent on spring snowmelt and late autumn rains (Smerdon *et al.*, 2008; Redding and Devito, 2011). However, infiltration may be impeded during these periods by seasonally frozen ground, particularly within peatlands (Petrone *et al.*, 2008;

Smerdon and Mendoza, 2010; Redding and Devito, 2011; Ireson *et al.*, 2015). Within the forested uplands, deep water tables that do not mimic topography are typical, along with thick unsaturated zones which have large soil moisture storage potential (Ferone and Devito, 2004; Smerdon *et al.*, 2005).

The Boreal Plains region has been the focus of recent hydrological studies (Buttle *et al.*, 2000) which have further developed the understanding of many processes occurring within individual ecosystems (i.e., peat, pond, or hillslope) across a range of glacial landforms and soil textures; however, few have directly investigated the hydrologic interactions and linkages that occur between ecosystems, particularly within fine-grained, glacial moraine settings that comprise roughly one third of the area of the Boreal Plains. Thus, additional study is warranted to further develop and refine existing conceptual models that incorporate the complex interaction between climate, vegetation, and geology within the region (Devito *et al.*, 2005b; Alberta Environment, 2008; Johnson and Miyanishi, 2008; Devito *et al.*, 2012; Ireson *et al.*, 2015). Such models are important because they serve to guide management, development, and reclamation[1] plans. One example is the oil-sands mining region, where complete reconstruction of the landscape will be required; the concepts that have been developed from natural analogs are crucial for design of plans that incorporate a higher probability of successful implementation. Within this thesis, hydrologic studies were undertaken at the Utikuma Region Study Area (URSA) to address these research gaps.

1.2. Utikuma Region Study Area

Initiated in 1999, the URSA is part of a long-term hydrological study that has examined the local and regional hydrology of Boreal Plains pond-peatland-upland ecosystems across a range of glacial substrates. The URSA is located approximately 350 km northwest of Edmonton and 150 km south of the discontinuous permafrost zone in north central Alberta (Woo and Winter, 1993; Figure 1-1A) with climate normals characterized by cold winters and warm summers with average annual temperature near 0°C (Marshall *et al.*, 1999).

[1] Within this thesis, usage of the term reclamation is intended to broadly encompass the R4 terms (i.e., remediation, reclamation, restoration and rehabilitation) as defined by Lima *et al.*, (2016). Specific definition of an R4 term will be dependent on the end goals for land use and will be problem-specific.

The URSA is situated within thick glacially-derived terrain that ranges in depth from 20 m to 200 m (Pawlowicz and Fenton, 2005; MacCormack *et al.*, 2015) and generally transitions from coarse-textured glaciofluvial sands and gravels in the northwest to finer-grained glacial moraine material within much of the central portion of the study area to fine-grained glaciolacustrine and organic material in the southeast (Figure 1-1B). Topographic relief is generally subtle with surface drainage that is typically poorly developed and regularly modified by beaver activity. Land cover across the URSA is comprised of shallow lakes and ponds in lowlands and surface depressions, peatlands with sparse black spruce and tamarack in flat-lying poorly drained areas, and forested hillslopes with primarily pine and white spruce in sandy coarse-grained glacial outwash areas and primarily aspen as well as poplar and white spruce in finer-textured moraine and glaciolacustrine landforms.

Previous studies at the URSA have investigated pond and wetland water budgets situated in a range of glacial landforms (Ferone and Devito, 2004; Smerdon *et al.*, 2005; Riddell, 2008); groundwater-surface water interactions in glacial outwash terrain (Smerdon *et al.*, 2007); groundwater recharge and runoff dynamics (Smerdon *et al.*, 2008; Redding and Devito, 2008, 2010, 2011; Devito *et al.*, 2017); the impacts of aspen clear-cutting on upland groundwater recharge (Hairabedian, 2011); the variability in ecosystem evapotranspiration (Petrone *et al.*, 2007; Brown *et al.*, 2010, 2014), the influence of subsurface ice formation (Petrone *et al.*, 2008; Smerdon and Mendoza, 2010; Redding and Devito, 2011), and hydrogeological controls on peatland wildfire vulnerability and burn severity (Hokanson *et al.*, 2016, Hokanson *et al.*, 2018).

1.3. Thesis Objectives and Format

The objective of this thesis was to advance the understanding of hydrologic interactions occurring within catchments situated in fine-grained glacial moraine settings within the Boreal Plains, including evaluation of the role seasonally frozen peatlands, and to evaluate potential ranges in hydrologic response caused by anthropogenic development (e.g., timber harvesting) and climate change. Specifically, this thesis explores the following research questions:

1. What are the dominant controls on water movement in ecosystems situated within fine-grained glacial terrain in the Boreal Plains of Alberta and what degree of interaction occurs between landscape units?

2. What degree of influence do seasonally frozen peatlands exert on shallow groundwater flow dynamics?
3. What changes to the hydrology of these systems can be expected due to development and climate change?
4. What are the key features of these ecosystems that need to be incorporated to improve the potential for successful reclamation in reconstructed landscapes?

These questions were addressed using studies conducted within the catchments of Pond 40 and Pond 43 (Figure 1-1), situated in glacial moraine deposits in the north-central portion of the URSA. Field data collected over an eleven-year period from 2000 to 2011 were used to evaluate the hydrologic interactions occurring between the different landscape units characteristic of Alberta's Boreal Plains, as well as to parameterize and calibrate numerical models that were used to simulate the key features of the existing landscape and explore the potential influence of a variety of climatic and development scenarios.

This thesis follows the paper format style and has been organized into six chapters. Following the first introductory chapter, Chapters 2, 3, 4, and 5 present the results of separate studies focused on different components of the research questions listed above. The outline of each study is as follows:

Chapter 2: Field observations from the pond, peatlands, and hillslopes located within the catchment of Pond 43 were used to evaluate the controls on water movement within the landscape and assess the degree of interaction between landscape units. The field data were used to develop and calibrate two-dimensional (2D) numerical models along two transects using the fully integrated groundwater-surface water code HydroGeoSphere (Aquanty, 2013). Sensitivity simulations were used to determine the key features governing hydrologic interactions within the fine-grained setting. A version of this chapter was published in 2015 (Thompson *et al.*, 2015).

Chapter 3: The influence of seasonal peatland freezing on shallow groundwater flow dynamics within the Pond 43 catchment was evaluated using field observations of peatland ice depth and simplified 2D numerical simulations conducted using MODFLOW-SURFACT (HGL, 2015). Simulations were used to quantify the influence of seasonal peatland freezing on peatland water table position and water distribution.

Chapter 4: The hydrologic influence of aspen harvesting was evaluated following clear-cutting of two stands in successive winters within the catchment of Pond 40 using the unharvested Pond 43 catchment as a reference. 2D numerical simulations were conducted using HydroGeoSphere to directly quantify the integrated hydrologic response of the system and predict system responses to varying aspen regeneration rates and atmospheric conditions. A version of this chapter was published in 2018 (Thompson *et al.*, 2018).

Chapter 5: The potential influence of climate change on ecosystems within Alberta's Boreal Plains was investigated using the calibrated 2D HydroGeoSphere model developed in Chapter 2. Thirteen climate change scenarios were simulated to examine the potential range in future hydrologic response over the 21st century. A version of this chapter was published in 2017 (Thompson *et al.*, 2017).

The final chapter comprises a summary of the findings from the preceding chapters and outlines suggestions for potential future research.

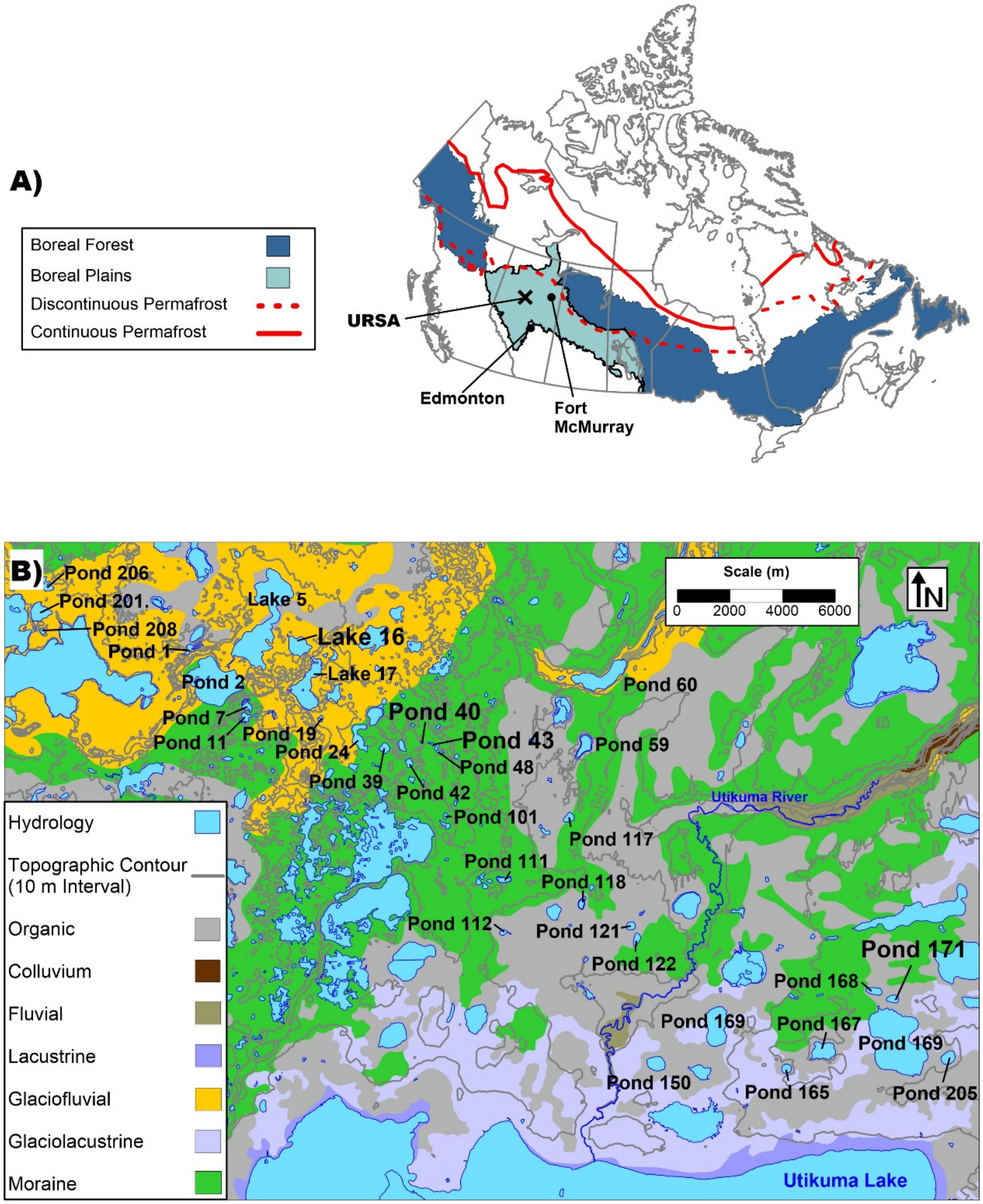

Figure 1-1. Location of the Utikuma Region Study Area (URSA): A) Location of the URSA within the Canadian Boreal Plains relative to the discontinuous permafrost zone. B) Surficial geology of the URSA based on Paulen *et al.* (2004, 2006) along with study locations including Pond 40 and Pond 43.

CHAPTER 2

Climatic Controls on Groundwater-Surface Water Interactions Within the Boreal Plains of Alberta: Field Observations and Numerical Simulations[2]

2.1. Introduction

The Boreal Plains of northern Alberta, Canada, are host to numerous wetlands, small lakes, and ponds interspersed within forested uplands. The wetlands provide important seasonal habitat for migratory birds and also represent a significant carbon pool (Smith and Reid, 2013; Gorham, 1991). However, they are under significant development pressures as a result of expanding petroleum developments and timber harvesting (Devito *et al.*, 2012). Furthermore, large areas have been disturbed by open-pit mining of oil sands which will require reclamation on an unprecedented scale over the next 30 to 50 years (Kelln *et al.*, 2008). Consequently, characterization of the processes governing the movement of water within these ecosystems is of pressing importance for both managing existing ecosystems and restoring those which have already been disturbed.

Ecosystems within the Boreal Plains are sustained by sub-humid climatic conditions, where annual precipitation is commonly less than potential evapotranspiration (Marshal *et al.*, 1999). Water deficit conditions occur frequently, making them highly vulnerable to developments that may alter their water budget. A large portion of annual precipitation falls during the summer months (Marshal *et al.*, 1999) when the potential evapotranspiration is greatest (Petrone *et al.*, 2007; Brown *et al.*, 2014). Therefore, in contrast to more humid regions such as the Boreal Shield, limited water is available to replenish the subsurface during the growing season in most years. Instead, the majority of groundwater recharge is derived from spring snowmelt and rain events occurring outside the growing season (Smerdon *et al.*, 2008; Redding and Devito, 2011), particularly in the forested uplands where evapotranspiration by species such as aspen may exceed growing season precipitation (Brown *et al.*, 2014).

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[2] A version of this chapter has been published:
Thompson, C, Mendoza, C.A., Devito, K.J., and Petrone, R.M., 2015. Climatic controls on groundwater-surface water interactions within the Boreal Plains of Alberta: Field observations and numerical simulations. Journal of Hydrology, 527: 734-746, doi: 10.1016/j.jhydrol.2015.05.027.

The elevated upland evapotranspiration, combined with the deep glacial soils within the region (Vogwill, 1978), result in deep upland water tables that do not follow topography and often decline away from adjacent ponds and peatlands (Ferone and Devito, 2004; Smerdon *et al.*, 2005). Consequently, following snowmelt and rain events the upland hydrologic response is dominated by fluctuations in storage, with little potential for generation of overland flow (Devito *et al.*, 2005b; Redding and Devito, 2008). Thus, pond and peatland water levels are dominated by the atmospheric fluxes of precipitation and evapotranspiration (Ferone and Devito, 2004; Smerdon *et al.*, 2005), with upland contributions occurring infrequently, particularly in lower permeability settings.

Despite their primary reliance on precipitation for maintenance within the dry sub-humid climate, pond and peatland ecosystems comprise a large and essential portion of the landscape within the Boreal Plains. Restoration of these ecosystems will represent an important challenge in the coming years, particularly in the oil sands mining region where complete reconstruction of the landscape will be required (Devito *et al.*, 2012). The re-establishment of peatland terrain has not been previously attempted (Price *et al.*, 2010), thus questions remain regarding the key features that may ultimately lead to successful reclamation, particularly within the context of the relatively dry sub-humid climate. Research aimed at addressing these questions has been primarily process-based, with most studies focusing on hydrological aspects of individual landscape units (i.e., pond, peatland, and hillslope; Devito *et al.*, 2012) for relatively short periods of time (i.e., 1 to 2 years) with variable climatic conditions.

This study evaluated the hydrologic linkages occurring between landscape units at a pond-peatland complex characteristic of Alberta's glaciated Boreal Plains region. A multi-year hydrologic dataset was collected over eleven years at a heterogeneous low permeability catchment within the region to assess the interactions occurring between landscape units and to evaluate how they vary over a range of meteorological conditions. The empirical dataset was used to develop numerical models capable of representing the dominant hydrological processes and to evaluate the key features of the landscape that allow these ecosystems to persist within the sub-humid climate. The goals of the study were to assess how the interactions vary between landscape units within the observed meteorological conditions, to evaluate the sensitivity of the system to a range of soil characteristics and catchment configurations commonly found within the Boreal Plains, and to address how catchment hydrology within the region is influenced by the configuration of landscape units to improve the design of (re)constructed landscapes.

2.2. Study Area

The study was conducted within the undisturbed catchment of Pond 43 (20 ha; Lat: 56.07 N, Long: 115.5 W; Figure 2-1), situated within the Utikuma Region Study Area (URSA). The URSA is located within the plains region of the western Boreal Forest approximately 350 km northwest of Edmonton, Alberta, Canada, and approximately 150 km south of the discontinuous permafrost region (Woo and Winter, 1993). Previous studies at the URSA have investigated pond water budgets situated in a range of glacial landforms (Ferone and Devito, 2004; Smerdon *et al.*, 2005); groundwater-surface water interactions in glacial outwash terrain (Smerdon *et al.*, 2007); groundwater recharge and runoff dynamics (Smerdon *et al.*, 2008; Redding and Devito, 2008, 2010, 2011); the variability in ecosystem ET (Petrone *et al.*, 2007; Brown *et al.*, 2010, 2014), and the influence of subsurface ice formation (Petrone *et al.*, 2008; Smerdon and Mendoza, 2010; Redding and Devito, 2011).

The URSA experiences cold winters and warm summers, with average January and July temperatures of -14.5°C and 15.6°C. The climate is sub-humid, with annual potential evapotranspiration (517 mm) exceeding annual precipitation (485 mm) in an average year (Marshal *et al.*, 1999). However, wetter years where precipitation exceeds potential evapotranspiration occur every 10 to 25 years (Mwale *et al.*, 2009). A large fraction of precipitation (i.e., 50-60%) falls during the summer months when the potential evapotranspiration is greatest (Petrone *et al.*, 2007; Brown *et al.*, 2014); therefore, the majority of groundwater recharge occurs outside of the growing season (Smerdon *et al.*, 2008; Redding and Devito, 2011).

The catchment of Pond 43 is situated on a regional topographic high and functions hydrologically as a regional recharge zone with predominantly vertical groundwater flow (Ferone and Devito, 2004). The catchment is characterized by deep, low permeability glacial disintegration moraine deposits with subtle topographic relief. The till is heterogeneous with sand and silt lenses interspersed throughout the soil column (Figure 2-2), and ranges in depth from 40 m to 50 m (Pawlowicz and Fenton, 2005). The till is underlain by marine shales of the Upper Cretaceous Smoky Group (Vogwill, 1978).

Mature forests consisting of predominately aspen cover the hillslopes (62% of area). In these areas, near-surface soils are classified as gray Luvisols and overlay oxidised clay till 5 m to 8 m in depth with unoxidised clay till beneath (Ferone and Devito, 2004). In flatter areas and depressions, peatlands with low density stunted black spruce (32%) are the dominant surface cover and directly overly unoxidised clay till. The organic materials within the peatlands range in depth from 2 to 5 m. Groundwater and surface water flow within the peatlands are influenced by seasonally frozen lenses, which can persist well into the summer months (Petrone *et al.*, 2008; Brown *et al.*, 2010). In the vicinity of the pond (6%), the peatlands transition to gyttja that extends to greater than 3 m depth. Surface drainage within the peatlands is generally poorly developed (Ferone and Devito, 2004) and frequently modified by beaver activity.

2.3. Field Data

2.3.1. Precipitation and Evaporation

Precipitation data were collected using 1 to 2 automated tipping bucket rain gauges and were checked using manual measurements from bulk rain gauges distributed across the site. Snow surveys were used to measure the snowpack each winter and spring. The surveys indicated similar snow depths accumulated within each landscape unit, with little drifting observed. Snow water equivalent (SWE) was determined from composite samples obtained during each survey.

Rainfall interception within aspen and spruce stands was estimated using integrated throughfall measurements. Throughfall was collected in 10 m long troughs placed below the shrub understory within both forest types. Bulk samples were collected at roughly two-week intervals following the snowmelt period before aspen leaf on through to mid-September.

Pond evaporation was measured using a Class A evaporation pan which was partially submerged to maintain thermal equilibrium between the pond and pan. Daily to continuous water level measurements within the pan were obtained throughout the pond's ice-free season. Additional evaporation data from Pond 43, peatlands, and the aspen forests have previously been reported (Petrone *et al.*, 2007; Brown *et al.*, 2010, Brown *et al.*, 2014).

2.3.2. Surface Water

Continuous pond stage measurements were obtained during the ice-free season using pressure transducers and augmented with manual measurements at established staff gauges. The bathymetry of Pond 43 was determined using depth to gyttja measurements from the water surface along two perpendicular transects. The bathymetry measurements were contoured and incorporated into a 1 m resolution LiDAR surface to provide a continuous surface for the study area.

Surface water flows and water levels in the pond and peatland channels were measured daily to weekly at V-notch weirs throughout the study period. The frequency of measurements was higher during the spring melt and following summer rain events to better capture increased flow rates during these periods. The resulting dataset was interpolated between measurements to estimate daily surface water inflows and outflows from the pond.

2.3.3. Groundwater

Groundwater levels were monitored throughout the duration of the study period in monitoring well and piezometer nests installed within each material. A total of 180 monitoring points were installed to a maximum depth of 23 m (Figures 2-1 and 2-2). Groundwater levels were typically measured weekly to biweekly from spring through autumn with less frequent measurements occurring during the winter. Continuous measurements were obtained at a number of locations using pressure transducers.

The hydraulic conductivity of the glacial substrate, peat, and gyttja has been previously reported (Ferone and Devito, 2004; Petrone *et al.*, 2008; Redding and Devito, 2008). The dataset was augmented with the results of slug tests conducted at 51 additional monitoring points distributed within each material. The dataset indicates that the hydraulic conductivity within the peat and gyttja displays a distinct decreasing trend with depth. Similar results have been reported by other researchers (e.g., Beckwith *et al.*, 2003; Quinton *et al.*, 2008). Within the glacial till, results of infiltration tests indicate a similar decline in hydraulic conductivity with depth in the near-surface materials, ranging from 10^{-3} m/s in the upper 0.1 m to 10^{-8} m/s at 0.75 m depth (Redding and Devito, 2010). However, no distinct depth trend was observed from slug tests conducted at greater depth. Instead, significant variation in hydraulic conductivity (i.e., 10^{-5} m/s to 10^{-10} m/s) is present throughout the material due to its highly heterogeneous nature.

2.4. Numerical Models

Two-dimensional (2D) models were developed to simulate the hydrologic interactions occurring between the different landscape units. The goal of the simulations was to reproduce observed hydrologic responses to climatic forcing and evaluate the sensitivity of the system to changes to specified parameters and boundary conditions, rather than to simulate hydraulic heads exactly at a specific well within the heterogeneous subsurface. Therefore, the models were simplified representations of the hydrologic system that retained sufficient complexity to simulate the dominant behavior, but may not capture finer-scale intricacies present at a given location.

Simulations were performed from January 1, 2000 to March 15, 2011. Models were developed for two transects to bracket the range in hydrologic conditions present at the site (Figure 2-1 and 2-2). The location of each transect was selected to follow interpreted surface water/groundwater flow paths while incorporating the largest number of data points available. Each transect features a pond near the center of the domain along with peatlands of varying depth and length on either side. Section A includes an aspen forested hillslope at its northern end that rises in elevation approximately 9 m above the pond. Section B differs by not having a hillslope in the vicinity of the pond and by terminating at an adjacent pond (i.e., Pond 48).

Simulations were performed using HydroGeoSphere (HGS; Aquanty, 2013). HGS is a physically-based, fully-integrated groundwater-surface water code that simultaneously solves the diffusion-wave approximation of the Saint Venant equations in the surface water domain and the variably-saturated Richard's equation in the subsurface domain. HGS was selected as the numerical simulator due to the integrated simulation capabilities of the code which allow the user to specify boundary conditions in the form of atmospheric fluxes, rather than specifying groundwater recharge rates and surface water levels that necessitate additional *a priori* assumptions. HGS has been applied to a wide range of problems, ranging in scale from catchment (Jones *et al.*, 2008) to continental (Lemieux *et al.*, 2008).

The finite-element mesh for each 2D model was discretized using uniform 1 m nodal spacing horizontally. Vertically, nodal spacing varied from 0.05 m to 0.1 m in the upper 0.5 m. Below this depth, finite-element layers increased to a maximum thickness of approximately 0.25 m. The surface of the mesh was set to ground surface using the 1 m resolution LiDAR surface with the pond's bathymetry incorporated. Glacial sediments within the modeled area reach depths of 50 m, thus no natural near-surface hydrologic boundary was present to define the base of the model. Instead, the base of the mesh was set to an arbitrary elevation of 633 m asl within each model resulting in a minimum domain thickness of 20 m.

2.4.1. Material Properties

Three hydrogeologic units were specified within the models including peat, gyttja, and glacial till. Parameter zones within each unit were defined based on observed depth trends in hydraulic conductivity. Within the peat and gyttja, the hydraulic conductivity was specified to decrease with depth (Table 2-1). An anisotropy ratio of 10:1 was assumed for both materials (Beckwith *et al.*, 2003; Nagare *et al.*, 2013). The remainder of the model was specified to be composed of glacial till. Highly decomposed organic materials at the edge of the peatland were also included within this unit, as their hydraulic properties were found to be similar to the till (i.e., low permeability and porosity, high bulk density). The glacial till was subdivided into three zones with hydraulic conductivity decreasing with depth. The depth of the upper two layers was specified to include near-surface sand and silt lenses observed in available boreholes as well as to simulate enhanced near-surface permeability resulting from fractures in the till, decayed roots, and animal burrows (Hayashi and van der Kamp, 2009). The remainder of the material was assigned a lower hydraulic conductivity of 1×10^{-8} m/s. An anisotropy ratio of 100:1 was assumed for each till layer and evaluated during model calibration. An additional 0.1 m thick forest floor was specified above the glacial till. Soil water characteristic and hydraulic conductivity curves were derived from the literature for the peat and gyttja (Silins and Rothwell, 1998; Price *et al.*, 2010) and glacial till using the properties of a clay loam (Carsel and Parrish, 1988). Limited data was available to characterize the porosity and specific storage of each material; therefore, measurements were augmented with literature values (Redding and Devito, 2006; Petrone *et al.*, 2008; Quinton *et al.*, 2008; Smith and Wheatcraft, 1993). Within the peat and gyttja, the porosity and specific storage were specified to follow a similar decreasing with depth trend as hydraulic conductivity; within the heterogeneous glacial materials uniform values were specified with depth (Table 2-1).

2.4.2. Boundary and Initial Conditions

Boundary conditions applied to the models consisted of a combination of specified fluxes and hydraulic heads. Specified fluxes consisting of daily rainfall and snowmelt were applied to the surface of each model (Table 2-2). Rainfall and snowmelt rates differed between landscape units to account for varying precipitation interception, snow sublimation, and frozen season length. Within the pond, all rainfall was assumed to reach the surface during the non-ice covered period. The rainfall rate within the forested areas was calculated based on a simple relationship used to relate measured interception, which was found to be similar in aspen and black spruce areas, to rainfall intensity and season. For the summer months when the evapotranspiration is highest, the daily canopy interception was assumed to be 2 mm, with the remaining quantity of rainfall reaching the ground. During the cooler month of May and the remaining winter months, the interception was reduced to 0.5 mm and 0 mm, respectively. The resultant calculated throughfall was found to provide a reasonable match to measured values (Figure 2-3) with an average interception of approximately 25% of annual precipitation.

Spring snowmelt was applied based on snow surveys and measured changes in water levels within the pond and observation wells. The depth of the winter snow pack was manually calculated in each landscape unit and applied as snowmelt over a 1 to 3 week period during the spring. Rain falling during the snow-covered period was assumed to be incorporated into the snowpack. Snow surveys indicated similar SWE volumes were present within each landscape unit in the spring, suggesting that varying snow interception was largely offset by higher mid-winter melt and evaporation rates in more open areas (Koivusalo and Kokkonen, 2002). Therefore, similar combined snowmelt evaporation and sublimation equal to 25% and 30% of the annual snowfall were assumed within the respective forested and pond areas.

Daily evapotranspiration was also applied to the models' surface from the spring through the fall. Within the pond, evapotranspiration was limited to ice-free periods, which were estimated from measured air temperatures and observed pond water levels. In the aspen-dominated areas, evapotranspiration was applied from May 15 to September 15 (Brown *et al.*, 2014); whereas, in the peatlands, evapotranspiration was specified on all non-snow-covered days (Brown *et al.*, 2010).

Measurements of evapotranspiration were not available for the entire study period; therefore, the temperature-based Hamon method (1963) was used to derive a continuous estimate. Calculated values were found to be similar to measured pond evaporation (Petrone *et al.*, 2007). The generated dataset was then scaled to account for differing evapotranspiration within the peatlands (Petrone *et al.*, 2007; Brown *et al.*, 2010) and aspen areas (Brown *et al.*, 2014). The resultant applied evapotranspiration ratios between the peatlands and pond and between the aspen and pond were 0.5 and 1.3, respectively, and were held constant for each year.

Evapotranspiration parameters were assigned to zones delineated by landscape unit to represent the different characteristics of the predominant vegetation type. Within the aspen hillslopes and black spruce peatlands, the evapotranspiration was distributed within the upper 3 m and 0.5 m, respectively, based on typical rooting depths (Debyle and Winokur, 1985; Lieffers and Rothwell, 1987). In the pond, evaporation was limited to the upper 0.1 m. In all zones, evapotranspiration was focused at near-surface nodes through use of a cubic function available within HGS. The actual evapotranspiration was limited by the available water, with the maximum rate occurring at saturations greater than field capacity (i.e., -33 kPa) and decreasing to zero at the permanent wilting point (i.e., -1500 kPa).

Within the pond, the specified precipitation and evapotranspiration fluxes also included measured stream flow into and out of the pond. The continuous stream flow dataset was converted to an areal flux using the estimated pond area for a given stage.

Additional boundary conditions applied to the models included specified heads at the base of the domain, a specified head at the eastern edge of Section B where it terminates at Pond 48, and no flow conditions along the remaining lateral edges of the models. At the base, a uniform hydraulic head was applied to simulate connection with the regional groundwater flow system. Measured hydraulic heads from deeper wells (i.e., >15 m depth) showed little fluctuation throughout the year; therefore, a constant head was assumed to provide a reasonable representation. At the eastern edge of Section B, daily specified hydraulic heads were applied within the peatland to simulate the adjacent Pond 48. Along the remaining lateral edges of the models, no flow boundaries were specified as groundwater flow has been observed to be predominately vertically downwards.

Initial conditions were generated by specifying the pressure head across the surface of the models and running them to steady state. Pressure heads were determined from measured hydraulic heads nearest to the start of the simulation period and interpolated between measurement points. Boundary conditions applied to the base (i.e., constant head) and lateral edges (i.e., no flow and specified head at Pond 48) of the models were unchanged from the transient simulations.

2.5. Results and Discussion

2.5.1. Meteorological Conditions

The weather was relatively dry throughout most of the study period. From 1999 to 2006, annual precipitation was less than average in all but 2005 (Table 2-3, Figure 2-5). The driest years occurred in 2001 and 2002, with both years receiving less than 60% of average precipitation. Wetter conditions occurred in consecutive years in 2007 and 2008 before dropping again in the following years. Despite the variability in annual meteorological conditions, rainfall patterns were relatively consistent with the majority of daily rainfall occurring in small events. On average, 90% of days with rain received less than 10 mm, with 51% of total annual rainfall occurring on days receiving less than 10 mm. Larger storms occurred infrequently, with daily rainfall totals exceeding 30 mm occurring on a total of 5 days over the duration of the study period, resulting in little potential for overland flow generation from the hillslopes (Redding and Devito, 2008).

2.5.2. Observed Groundwater Flow Patterns

Peatland water levels were found to mimic the pond stage throughout the study period (Figure 2-4). Water levels in both landscape units reflect recent trends in meteorological conditions (Figure 2-5), although the correlation is stronger in the pond (i.e., correlation coefficient (R) = 0.65) than the peatlands (R = 0.40). Seasonally, peatland water levels were generally higher than the pond following spring snowmelt. The resulting groundwater flow direction was from peatland to pond, with the annual discharge period varying in duration in response to the frequency and magnitude of early spring rains. Pond discharge from the peatlands continued through the growing season following drier periods when pond water levels were low (e.g., 2002 to 2003) as the deeper, lower hydraulic conductivity peat layers maintained peatland water levels above the pond. In contrast, lateral groundwater flow from the pond to the peatlands was observed from late summer through spring following the relatively dry 1999 growing season (Ferone and Devito, 2004); however, pond water levels were elevated during this period by comparison. During wetter periods when the pond was full (e.g., 2007 to 2008), pond-peatland interactions were much more dynamic, with hydraulic gradients reversing frequently in response to rain events. However, as the growing season progressed, peatland water levels generally declined below the pond as previously observed (Ferone and Devito, 2004), resulting in lateral groundwater flow from the pond to the peatlands throughout a large portion of the year. In contrast to horizontal hydraulic gradients, water level data collected within the peat, gyttja, and underlying shallow glacial materials indicated continuous downward groundwater flow occurred regardless of meteorological conditions, consistent with the interpretation that the area functions as a regional recharge zone (Ferone and Devito, 2004).

Within the hillslope, water level elevations were generally well below those observed in both the peatland and pond as previously observed (Ferone and Devito, 2004), indicating that the hillslope acted as a net water sink to the peatland. All wells remained dry throughout the study period except at the hillslope toe. At the toe, wells were dry prior to the spring of 2003 (Figure 2-4), indicating that the water table declined rapidly at the peatland edge to a depth of at least 2 to 3 m below the base of the pond. Annual snowfall during this period was well below average, suggesting limited water was available to recharge the glacial materials. As a result, the direction of groundwater flow was always from the peatland to the hillslope regardless of season. In 2003, spring water levels at the toe of the hillslope increased to within about 1 m of those in the peatland following an average snow fall year. However, a rapid water level decline was observed throughout the growing season despite near average rainfall, suggesting that most of the water was either removed through evapotranspiration or incorporated into storage. Consequently, by mid-July the water table dropped below the base of the well located at the hillslope toe.

Water levels at the hillslope toe remained low until the spring of 2005, when they rose above those in the adjacent peatland following snowmelt. The increase in water level followed another average snow fall year that was preceded by a relatively wet autumn, suggesting that the majority of groundwater recharge in the aspen hillslopes occurs outside of the growing season in the spring and late fall, consistent with the findings of other studies in the URSA (Smerdon *et al.*, 2008; Redding and Devito, 2011). Comparable responses were observed in the relatively wet years of 2006 to 2009, with water levels at the toe of the hillslope exceeding those in the peatland each spring. Thus, some groundwater discharge from the hillslope to the peatland likely occurred, although the combination of generally low hydraulic conductivity glacial till and seasonally frozen peat probably prevented significant water exchange.

2.5.3. Comparison of Simulated and Observed Hydraulic Heads

Simulated transient hydraulic heads were evaluated at 111 monitoring wells (i.e., 68 in Section A; 43 in Section B) along with the computed stage at Pond 43. Calculated calibration statistics (i.e., R, residual mean (r_m), root mean square error (RMS), and RMS normalized by the range in observed water levels (NRMS)) indicate the models were able to provide a reasonable representation of the hydrologic system (i.e., R = 0.65; RMS = 0.29 m; Figure 2-6d). Water levels within the pond were particularly well-represented (i.e., R = 0.97; RMS = 0.06 m; Figure 2-6a), although some bias is present at shallower pond depths where the predicted stage is somewhat over-predicted. Greater discrepancy is present at peatland locations (i.e., R = 0.77; RMS = 0.17 m; Figure 2-6b), with the RMS equal to about half the range of annual peatland water level fluctuations. The largest residuals were generally from deeper monitoring wells within the glacial till (i.e., R = 0.47; RMS = 0.50 m; Figure 2-6c), where considerable heterogeneity is present.

Temporally, simulated trends in water levels were found to be in good agreement with measured values (Figure 2-7). Computed residuals typically decreased from spring to fall when precipitation inputs were lowest, with similar error values obtained in both dry and wetter years. At the hillslope toe in Section A, the general trend of a rapidly rising water level following snowmelt and a continuous decline throughout the growing season is well-represented by the model (Figure 2-7e). Water levels at this location were over-predicted early in the simulated period, which likely occurred as a result of the initial conditions in the hillslope being too wet. In later years, the predicted peak of the spring water level rise differed from measured values in some years. This may be due to a number of factors not incorporated into the model. Such factors might include drifting of snow, higher hydraulic conductivity lenses or fractures that lead to more rapid transmission of water to this location, the presence of frozen peat that may act as a dam preventing water exchange, or flow contributions from outside of the assumed 2D flow system. Simulated hydraulic heads during the winter and early spring are not as well represented, as flow of water from the peatlands towards the hillslope leads to an increase in water level not observed in the field. This discrepancy occurs as a result of not including freezing and thawing within the simulations, as the peatlands freeze solid throughout the winter preventing water loss to the hillslope.

Above the hillslope toe, all wells within the hillslope remained dry throughout the duration of the simulation, with the simulated water table elevation declining with distance upslope. The high aspen evapotranspiration and thick unsaturated zone prevented significant recharge from reaching the water table, resulting in minimal fluctuations in groundwater levels and no overland flow generation. Thus, results of the simulations suggest that significantly wetter climate cycles would be required before noticeable rises in groundwater levels could occur. The simulated actual evapotranspiration integrated over the entire hillslope was limited by soil moisture availability, ranging from 56-76% of the potential rate (Table 2-3), and was found to be similar to measured rates at an adjacent catchment (Brown *et al.*, 2014). In contrast, simulated actual evapotranspiration within the pond and peatlands was predicted to equal the potential rate throughout the study period.

Despite neglecting the effects of peatland freezing and thawing, observed seasonal trends in hydraulic heads at peatland wells are well replicated (Figure 2-7a, b, d). The model is able to capture the early spring rise in water levels following snowmelt and the general decline throughout the growing season. Periodic smaller scale increases in water levels following larger rain events are reasonably well captured, although some deviations are apparent which may be a result of spatially varying rates of interception that were not incorporated.

The stage of Pond 43 was also well replicated (Figure 2-7c) and was found to mimic measured water levels throughout most of the simulation. Some discrepancy is apparent during the summers of 2002 to 2004 when the measured water level decline is under-predicted. This may be due to the relatively simple air temperature-based method that was used to calculate the daily pond evapotranspiration. Incorporation of a more rigorous approach that includes the effects of water depth and temperature may have provided a better representation of the process and could be considered for future simulations.

2.5.4. Sensitivity Simulations

Additional simulations were conducted to evaluate the sensitivity of the hydrologic system to changes in selected parameter values, boundary conditions, and catchment configurations. Each scenario did not necessarily provide a realistic representation of the system, as simulated water levels at some locations were significantly different than what has been observed. However, the simulated cases do provide insight into the controlling aspects of the system, which may be relevant to construction requirements in a reclamation setting. Results of the sensitivity simulations are described below and summarized in Table 2-4.

The first simulations investigated the sensitivity of the model results to the specified set of boundary conditions. Simulated water levels within the pond were found to be particularly sensitive to the timing and duration of snowmelt in both the pond and peatlands. Modification of the timing of the snowmelt period, particularly within the peatlands, led to simulated peaks in pond stage being out of phase with measured values. Annual water levels within the peatlands were not as sensitive to this boundary input provided that the snowmelt was added in a similar time frame as the observed increase in peatland water level. The overall response of the system was less sensitive to the timing of snowmelt in the hillslope, suggesting limited water was transferred to the adjacent peatland. Modification to the onset of the melt period primarily resulted in changes to the duration of the peak water level at the toe, which was predicted to persist until removal of water from aspen evapotranspiration commenced each spring.

To assess the uncertainty in the tabulated evapotranspiration boundary fluxes for each landscape unit, simulations were conducted with the specified evapotranspiration modified by 30%. Pond and peatland water levels were sensitive to this boundary flux. Increasing the evapotranspiration led to significantly reduced water levels (Figure 2-8). Conversely, where the evapotranspiration decreased, pond and peatland water levels were predicted to be elevated relative to the base case simulation. Similar responses in water levels at the toe of the hillslope were predicted for these scenarios, although the magnitude of the change was generally less than within the peatlands. Simulation results were less sensitive to increased aspen evapotranspiration, as the simulated actual evapotranspiration was already limited by the available soil moisture (Figure 2-8). Reducing the aspen evapotranspiration led to increased water levels at the toe of the hillslope, although this had little effect on water levels within the pond and peatland. Similar to the base case simulation, all wells located further up the hillslope remained dry in both scenarios.

The next sensitivity simulations evaluated the influence of the hydraulic conductivity distribution. Results were relatively insensitive to the hydraulic conductivity of the organic materials. Modification of the rate of decline in peatland hydraulic conductivity with depth resulted in minor deviations from the base case. Uniform order of magnitude increases or decreases to the hydraulic conductivity to either the peat or gyttja also led to relatively small changes to pond and peatland water levels. Similarly, small variations in water levels at the toe of the hillslope were predicted for changes in peat hydraulic conductivity.

Water levels in all materials were most sensitive to the hydraulic conductivity of the glacial till. Increasing the hydraulic conductivity of each till unit by an order of magnitude led to water levels in the pond and peatlands dropping by up to 0.5 m due to increased vertical groundwater flow. Consequently, the pond was predicted to dry out in 2010 for this scenario (Figure 2-9). Conversely, decreasing the hydraulic conductivity of each till unit by an order of magnitude led to increased pond and peatland water levels of about 0.3 m. Representation of the glacial materials as a uniform material of high (i.e., 1×10^{-7} m/s) or low (i.e., 1×10^{-8} m/s) hydraulic conductivity yielded similar results as the respective increased and decreased hydraulic conductivity simulations.

Groundwater levels at the toe of the hillslope were also sensitive to the hydraulic conductivity of the till. For the increased hydraulic conductivity scenario, water level peaks at the toe of the hillslope following snowmelt were generally lower and the magnitude of annual fluctuations was greater (Figure 2-9). In contrast, for the decreased hydraulic conductivity scenario, the simulated water level did not fluctuate in response to snowmelt inputs. Instead, the annual influx of water was held in storage prior to being removed by aspen evapotranspiration during the growing season. A similar response was predicted for simulations that represented the glacial materials as a uniform high and low permeability material.

The final sensitivity simulations investigated the individual influence that the pond and peatlands exerted on the hydrological behavior of the system. In the first simulation, the pond was removed from the modeled area to represent catchments that occur within the URSA that have extensive peatlands that do not terminate at a surface water body. For this scenario, the topography within the footprint of the pond was set to a similar elevation as the adjacent peatlands. The base of the organic materials was unchanged from the base case. Results indicated that the pond acts as a net water sink within the landscape, as water levels increased throughout the peatland (Figure 2-10). Similar conclusions were reached by Smerdon *et al.* (2005, 2007) at a lake situated within more permeable glacial outwash within the region. For this simulation, peatland water levels rose above ground surface by the spring of 2007, with ponded water predicted to remain to the end of the simulation. Thus, in the absence of a pond, increased overland flow would have been generated for the simulated meteorological conditions.

The final sensitivity simulation was used to investigate the influence the peatlands exert within the landscape. For this scenario, the peatlands were replaced with glacial till and aspen was assumed to be the vegetative cover. Although this scenario is unlikely to occur in an undisturbed setting, it may ensue following disturbance by timber harvesting, as roads and skid trails are commonly developed on flatter areas occupied by peatlands. Furthermore, it could be representative of a potential reclamation scenario within the oil sands mining region if difficulties are encountered in reestablishing peatlands within the landscape. Results indicated that the peatlands play a major role in maintaining the pond, as the predicted pond water level decreased significantly relative to the base case (Figure 2-10). Extension of the aspen stand led to increased removal of water through evapotranspiration. As a result, wells that were formerly located within the peatland were predicted to dry out seasonally, leading to increased seepage losses from the pond. Thus, the peatlands act to conserve water within the landscape, providing a counter-balance to the higher evapotranspiration demands associated with both the pond and hillslope.

2.5.5. Importance of Climate

The eleven-year study period permitted the evaluation of the hydrologic response of the system to a range of meteorological conditions. Results indicate dynamic interactions occur between Pond 43 and the peatlands that are driven by the fluxes of precipitation and evapotranspiration, with little to no additional input from the surrounding uplands. As a result, pond and peatland water levels reflect recent climatic variability (Figure 2-5), generally following trends in the cumulative difference between precipitation and evapotranspiration. Comparable responses to climate forcing have been observed in prairie pothole wetlands such as those found in the Cottonwood Lake area of North Dakota. Situated in a regional recharge zone within low permeability glacial terrain, water levels measured over 2 decades were found to mirror trends in precipitation (Winter *et al.*, 2001). In contrast to the pond and peatlands, water levels within the hillslope are more indicative of longer-term trends in climatic variability. Periods of groundwater recharge within the hillslope are generally limited to the late fall and to the spring following snowmelt as most precipitations inputs are lost to evapotranspiration over the growing season or taken up into storage. Therefore, a series of wetter years are required before soil moisture thresholds are exceeded and an increase in water levels is exhibited (Redding and Devito, 2010).

The rapid response to fluctuations in meteorological conditions at the pond and peatlands is a direct consequence of their position within the context of regional groundwater flow systems. Due to their location at elevated topographic positions, the pond and peatlands and their associated catchments function as regional recharge areas with little potential for discharge of groundwater to the system from outside sources. Thus, they are heavily reliant on precipitation inputs and have limited buffering capacity available to sustain water levels during periods of drought. Wetlands falling into this category are highly vulnerable to changes in climate at both short- and long-term time scales (Winter, 2000).

The long-term viability of wetlands within the Boreal Plains of Alberta may therefore be in question when considering the potential impact of climate change. Projections for the region indicate a gradual warming trend, with mean annual temperature increases of approximately 2 to 5°C by the 2050's (Barrow and Yu, 2005). It is probable that the length of the growing season will also increase as the climate warms, potentially leading to increased water stress resulting from greater evapotranspiration losses. This stress may be offset by predicted general increases in annual precipitation. Nevertheless, there is considerable uncertainty in these predictions, with some scenarios predicting reduced precipitation, particularly during the summer period (Barrow and Yu, 2005). However, the ability of peatlands to restrict water losses during periods of drought (Kettridge and Waddington, 2014; Waddington *et al.*, 2015) along with the resilience of the deeper peat to decomposition (Waddington *et al.*, 2015) may render them more resistant to climate change than otherwise expected. Future research will involve developing simulations of predicted climate change scenarios that may prove insightful to the future sustainability of these ecosystems.

2.5.6. Influence of Frozen Materials

Thick ice lenses that persist well into the growing season have been observed to develop within the peatlands (Petrone *et al.*, 2008; Smerdon and Mendoza, 2010; Brown *et al.*, 2010). However, despite neglecting their influence within the simulations, water levels within Pond 43 and the peatlands were generally well replicated. This may be due to a combination of the thermal behavior of the near-surface peat along with the low hydraulic conductivity of the deeper peat layers and underlying glacial substrate. The higher hydraulic conductivity near-surface peat that transmits most of the water has been observed to thaw relatively rapidly in the spring in response to rising air temperatures, thereby reducing the influence of the frozen material on water distribution within the peatlands. Furthermore, the presence of the deeper, lower permeability peat and glacial till restricts vertical groundwater flow in a similar fashion as the development of frozen peat, thereby maintaining peatland water levels during the frozen period.

Although the numerical simulations were able to capture the overall trends in the pond and peatlands, simulated peatland water levels were frequently under- and over-predicted during the spring and late autumn, respectively (Figure 2-6). This discrepancy may be due to temporal changes to the peat hydraulic conductivity resulting from the seasonal formation of ice. During the spring and early in the growing season the hydraulic conductivity of the peat may be lower than specified within the base case simulation due to the presence of ice. As a result, increased vertical groundwater flow is predicted, leading to peatland water levels being under-predicted. Likewise, during the latter part of the growing season, once the ice has melted, the specified hydraulic conductivity of the peat may be too low to allow adequate seepage and lowering of peatland water levels. Thus, the specified hydraulic conductivity distribution may be representative of an average thermal state of the peat allowing adequate simulation of the system.

In the context of a warming climate, the influence of the seasonally frozen peatlands may be diminished as the duration of the frozen season is reduced. As a result, the volume of water retained in ponds and peatlands within the Boreal Plains may be reduced in the early spring due to increased seepage losses. Future studies will focus on the dynamics of how seasonal ice formation influences water distribution within the peatlands, and how this may be altered by climate change.

2.5.7. Applicability to Mine Reclamation

The impetus for this study is driven by reclamation requirements within the Boreal Plains region associated with activities such as timber harvesting and petroleum developments (Devito *et al.*, 2012). In particular, expanding open-pit mining of oil sands in northern Alberta has created the need to reclaim large areas of land (Kelln *et al.*, 2008; Price *et al.*, 2010). Regulatory requirements specify that the disturbed land be returned to an “equivalent capability”, with the expectation that a vegetative regime similar to what was present prior to disturbance will be reestablished (OSWWG, 2000; Alberta Environment, 2008). The similarity in hydrologic characteristics (i.e., climate, geology, and vegetation) of the study area to many of the impacted areas allow it to be used as a proxy for future landscape reconstruction and reclamation.

This study indicates that peatlands within the Boreal Plains do not necessarily require regional groundwater discharge or significant flow contributions from adjacent uplands for long-term maintenance. However, in the context of a reconstructed landscape, this conclusion is predicated on the assumption that the peatlands be underlain by materials of sufficiently low permeability and be uninfluenced by drainage infrastructure (e.g., ditches or drained roads). Study results also suggest the peatlands conserve water within the landscape, supplying water to adjacent ponds and hillslopes within the sub-humid climatic setting. Thus, when considering the hydrologic impact of forestry, the largest effect to neighbouring wetland ecosystems may not be due to tree removal. The deep upland water table, large available water storage, and rapid aspen regeneration may limit the impacts of harvesting (Devito *et al.*, 2005b; Macrae *et al.*, 2005, Macrae *et al.*, 2006; Carrera-Hernandez *et al.*, 2011). Instead, the primary impacts of forestry may be due to placement of roads within flat-lying peatlands, which may disrupt shallow groundwater flows and potentially lead to isolated pond and peatland areas that are highly susceptible to drought (Lieffers and MacDonald, 1990; Smerdon *et al.*, 2009).

Although natural established peatlands in the region have continued to persist despite experiencing extended drought conditions with large water level declines, prolonged periods of drought may produce conditions that are hostile to newly placed peat that is less well established. Price *et al.* (2010) indicate that peatland water pressure should not drop below -1 m to encourage moss establishment and peat development, although the biological limit may be as low as -6 m (McCarter and Price, 2014). The upper bound of this threshold was exceeded frequently during the study period. Therefore, an external source of water may be needed to maintain wetland vegetation if initial reclamation is followed by drier climate cycles. However, contrary to some reclamation guidelines (e.g., OSWWG, 2000; Alberta Environment, 2008), the placement of small hills and hummocks within the landscape is unlikely to generate significant water. Aspen forested hillslopes in this climate have been shown to act as water sinks to the wetlands, storing and transpiring precipitation inputs and receiving lateral groundwater flow from the peatlands throughout much of the year. Thus, from a purely “health of peat” perspective, inclusion of small hills and hummocks within the reclaimed landscape may prove detrimental to peatland development, removing water from the system rather than supplying it to the wetlands as intended.

Table 2-1. Calibrated subsurface parameters within the numerical models.

Material	Depth Range (m)	Hydraulic Conductivity (m/s)		Porosity (-)	Specific Storage (m^{-1})
		Horizontal	Vertical		
Peat	0.0 - 0.1	3×10^{-3}	3×10^{-4}	0.90	1×10^{-4}
	0.1 - 0.3	3×10^{-4}	3×10^{-5}	0.82	5×10^{-5}
	0.3 - 0.5	8×10^{-5}	8×10^{-6}	0.72	8×10^{-6}
	0.5 - 1.0	4×10^{-5}	4×10^{-6}	0.60	2×10^{-6}
	1.0 - 1.5	2×10^{-6}	2×10^{-7}	0.50	5×10^{-7}
	1.5 - 2.0	3×10^{-8}	3×10^{-9}	0.45	2×10^{-7}
	2.0 - Base	1×10^{-8}	1×10^{-9}	0.40	1×10^{-7}
Gyttja	0.0 - 1.0	1×10^{-6}	1×10^{-7}	0.45	3×10^{-6}
	1.0 - 2.0	3×10^{-8}	3×10^{-9}	0.30	4×10^{-7}
	> 2.0	5×10^{-9}	5×10^{-10}	0.22	1×10^{-7}
Glacial Till	Upper	1×10^{-5}	1×10^{-7}	0.20	1×10^{-4}
	Mid	5×10^{-7}	5×10^{-9}	0.20	1×10^{-4}
	Lower	1×10^{-8}	1×10^{-10}	0.20	1×10^{-4}
Forest Floor	0.0 - 0.1	1×10^{-4}	1×10^{-4}	0.80	1×10^{-4}

Notes:

[a] Specified specific storage values for peat and gyttja are low; however, the influence on simulation results is negligible due to small fluctuations in water levels (i.e., less than 1 m).

Table 2-2. Summary of annual boundary fluxes applied in the base case models.

Year	Pond (mm)					Peat (mm)			Aspen (mm)		
	Rain [a]	Snow [a]	Stream Flow In	ET [b,c]	Stream Flow Out	Rain [a]	Snow [a]	ET [b,c]	Rain [a]	Snow [a]	ET [b,c]
2000	382	60	0	410	0	317	125	315	369	72	518
2001	187	71	0	440	0	155	102	293	159	98	508
2002	226	63	0	405	0	220	70	278	220	70	507
2003	335	107	0	421	0	293	150	315	330	113	527
2004	300	77	0	391	0	297	81	320	297	81	512
2005	342	149	47	430	134	319	173	329	323	168	532
2006	298	134	73	411	89	356	76	318	362	70	557
2007	318	212	365	463	520	312	218	359	315	214	578
2008	354	151	62	450	218	324	181	333	327	178	574
2009	239	152	81	437	201	209	182	310	212	179	528
2010	227	55	0	481	12	220	62	339	223	59	539

Notes:

[a] Proportions of rain and snow vary by landscape unit due to different frozen season lengths. Total precipitation (i.e., total precipitation = rain + snow) was assumed to be uniform.

[b] Maximum evapotranspiration (ET) applied within each landscape unit. The actual evapotranspiration removed within the simulations was limited by the available water.

[c] ET includes rain interception and snow sublimation that was removed externally from the model.

Table 2-3. Comparison of simulated aspen actual evapotranspiration (AET) to potential evapotranspiration (PET) and net precipitation (P) including snowmelt and rain throughfall.

Year	P (mm)	ET [a,b] (mm)	AET[a,b] (mm)	AET/ET (%)	P - AET (mm)
2000	442	518	395	76	47
2001	257	508	293	58	-35
2002	290	507	284	56	6
2003	443	527	370	70	72
2004	378	512	323	63	55
2005	491	532	392	74	99
2006	432	557	366	66	66
2007	530	578	403	70	127
2008	504	574	397	69	107
2009	391	528	352	67	39
2010	282	539	303	56	-21

Notes:

[a] Maximum ET applied within aspen areas. The AET removed within the simulations was limited by the available water.

[b] Includes rain interception and snow sublimation that was removed externally from the model.

Table 2-4. Summary of sensitivity simulation results.

Scenario	Influence on Landscape Unit			Overall Sensitivity
	Pond	Peat	Aspen	
Pond Snowmelt	Timing of spring water level peak shifted, minimal change throughout rest of year.	Modified timing and duration of spring water level peak in the vicinity of the pond.	Little to no change from base case.	Moderate to Low
Peat Snowmelt	Modified timing and duration of spring water level peak.	Timing of spring water level peak shifted, minimal change throughout rest of year.	Modified duration of spring water level peak.	Moderate
Aspen Snowmelt	Little to no change from base case.	Modified timing and duration of spring water level peak at the peatland edge.	Modified duration of spring water level peak.	Low
Pond PET	Large influence on water level.	Large influence on water levels.	Moderate influence on water levels.	High to Moderate
Peat PET	Large influence on water level.	Large influence on water levels.	Moderate influence on water levels.	High to Moderate
Aspen PET	Minor change to water level.	Small changes to water levels, decreasing with distance from the hillslope.	Moderate influence on water levels.	Low to Moderate
Peat K	Minor change to water level.	Low to moderate change to water levels.	Minor change to water levels.	Low
Glacial Till K	Large influence on water level.	Large influence on water levels.	Large influence on water levels.	High
No Pond	-	Large increase in water levels.	Large Increase in water levels.	High
No Peat	Large decrease in water level.	-	Small influence on overall response. No water level recovery during the winter.	High to Moderate

Notes:

[a] PET = potential evapotranspiration. K = hydraulic conductivity.

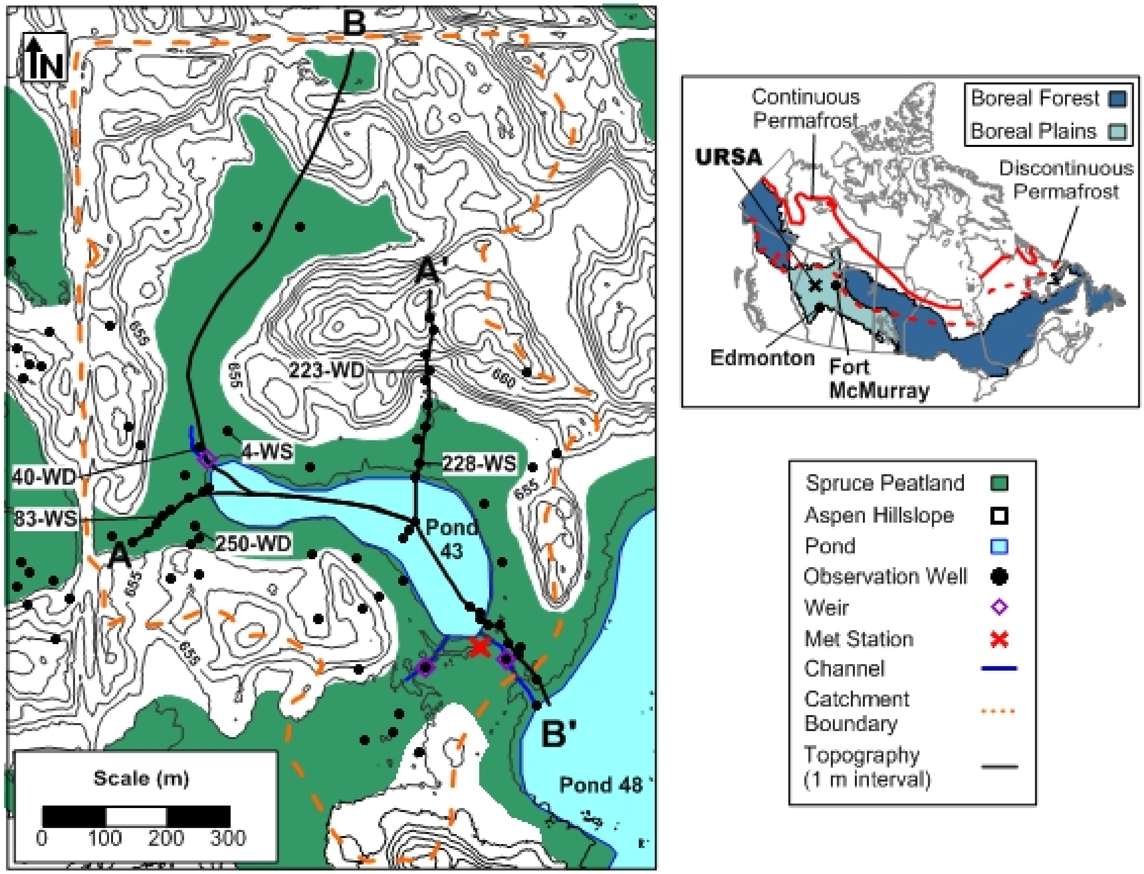

Figure 2-1. Location of the Utikuma Region Study Area (URSA). Right: Location of the URSA within the Canadian Boreal Plains relative to the discontinuous permafrost zone. Left: Enlarged map showing the Pond 43 study area including instrumentation, vegetative cover, selected monitoring well locations, surface water catchment boundaries, and the locations of the 2D numerical models.

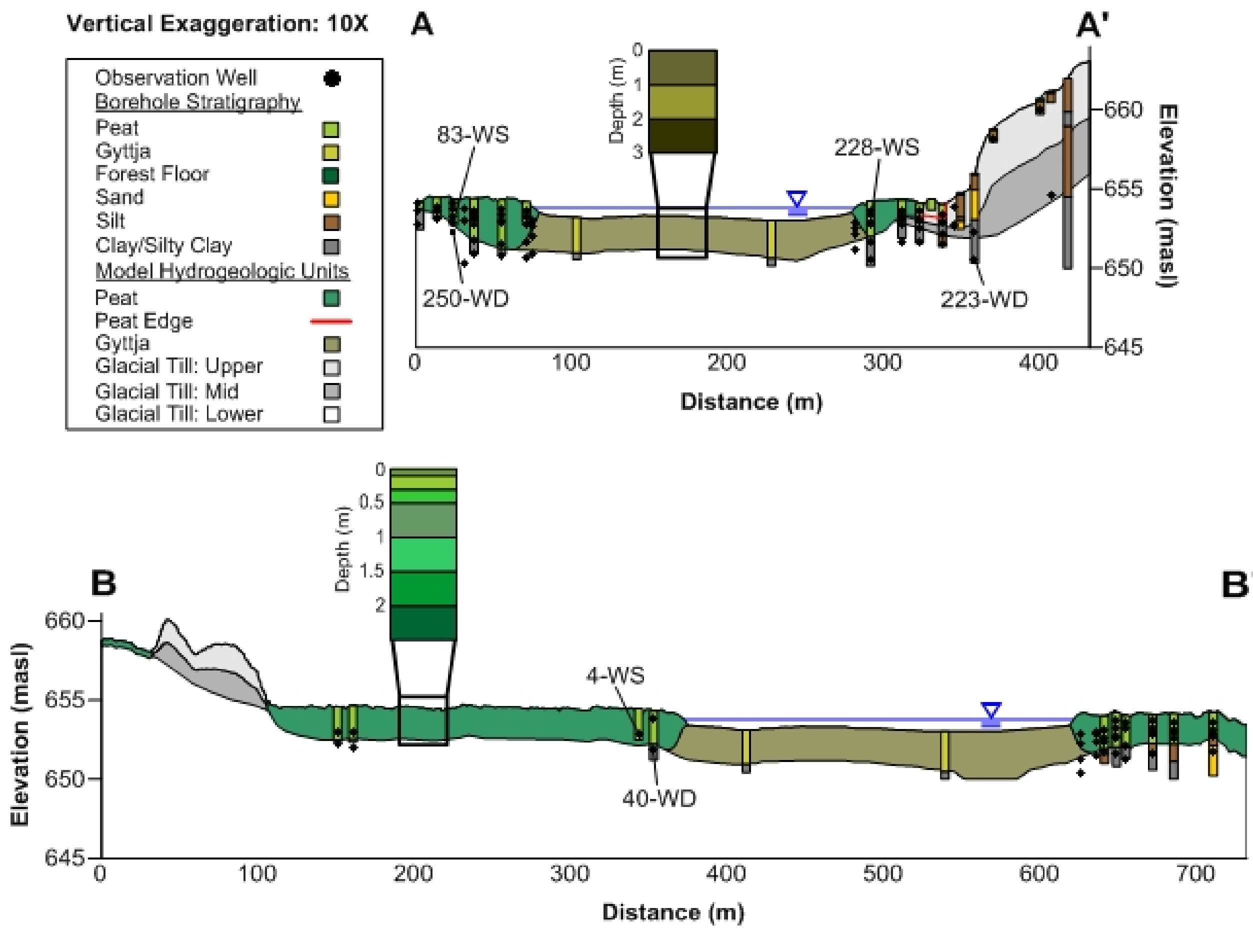

Figure 2-2. Model domains for cross-sections A and B, including borehole stratigraphy, hydrogeologic units, and observation points. Cross section locations are indicated in Figure 2-1. Note that the base of the models has been truncated at 645 m asl for illustration purposes.

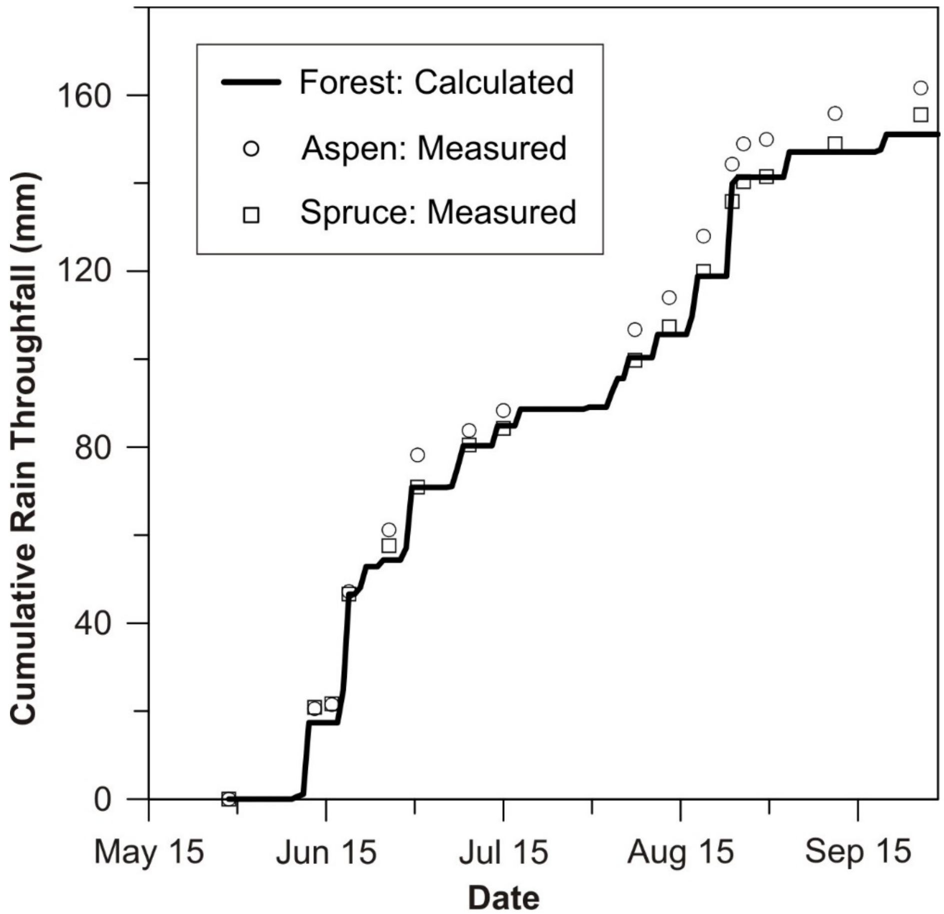

Figure 2-3. Comparison of measured and calculated cumulative throughfall to the forested landscape units.

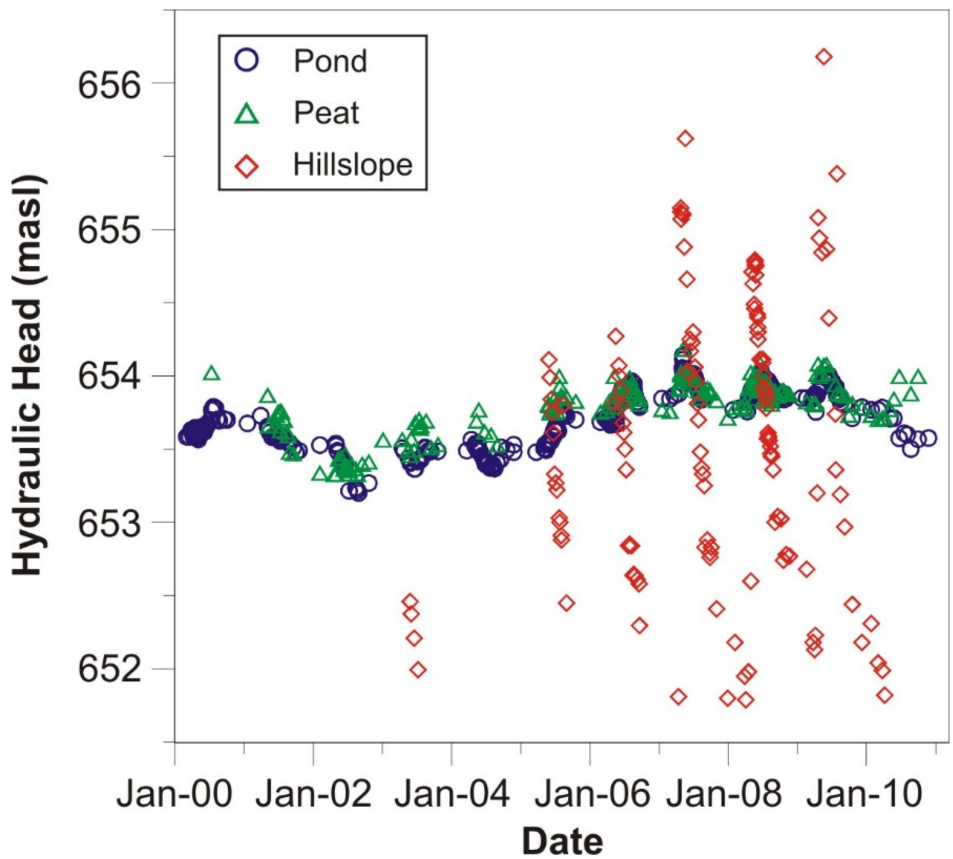

Figure 2-4. Observed water levels over the study period illustrating the similarity between pond and peatland (228-WS) water levels, as well as the large lateral hydraulic gradients between the peatland and hillslope (223-WD). Note that the hillslope well was dry prior to spring 2003, from mid-July 2003 through to spring 2005, and throughout most of 2010. Well locations are shown in Figures 2-1 and 2-2.

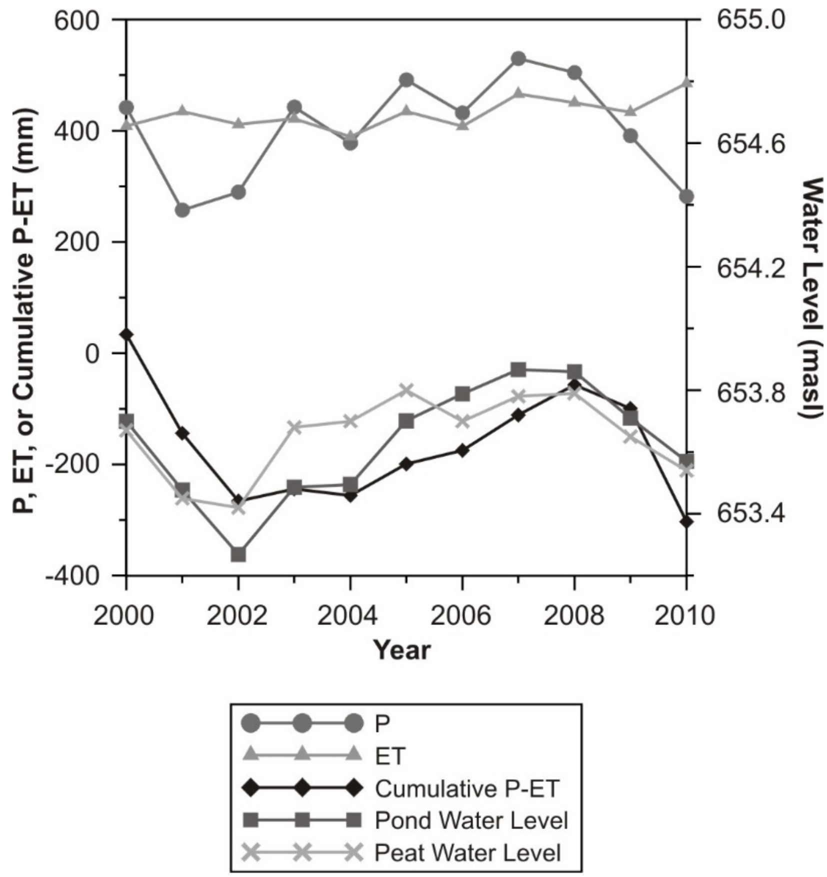

Figure 2-5. Variation in annual precipitation (P), pond evapotranspiration (ET), and cumulative pond P - ET over the study period. Observed average annual pond and peatland water levels were found to mimic trends in cumulative P - ET due to limited inputs from adjacent uplands.

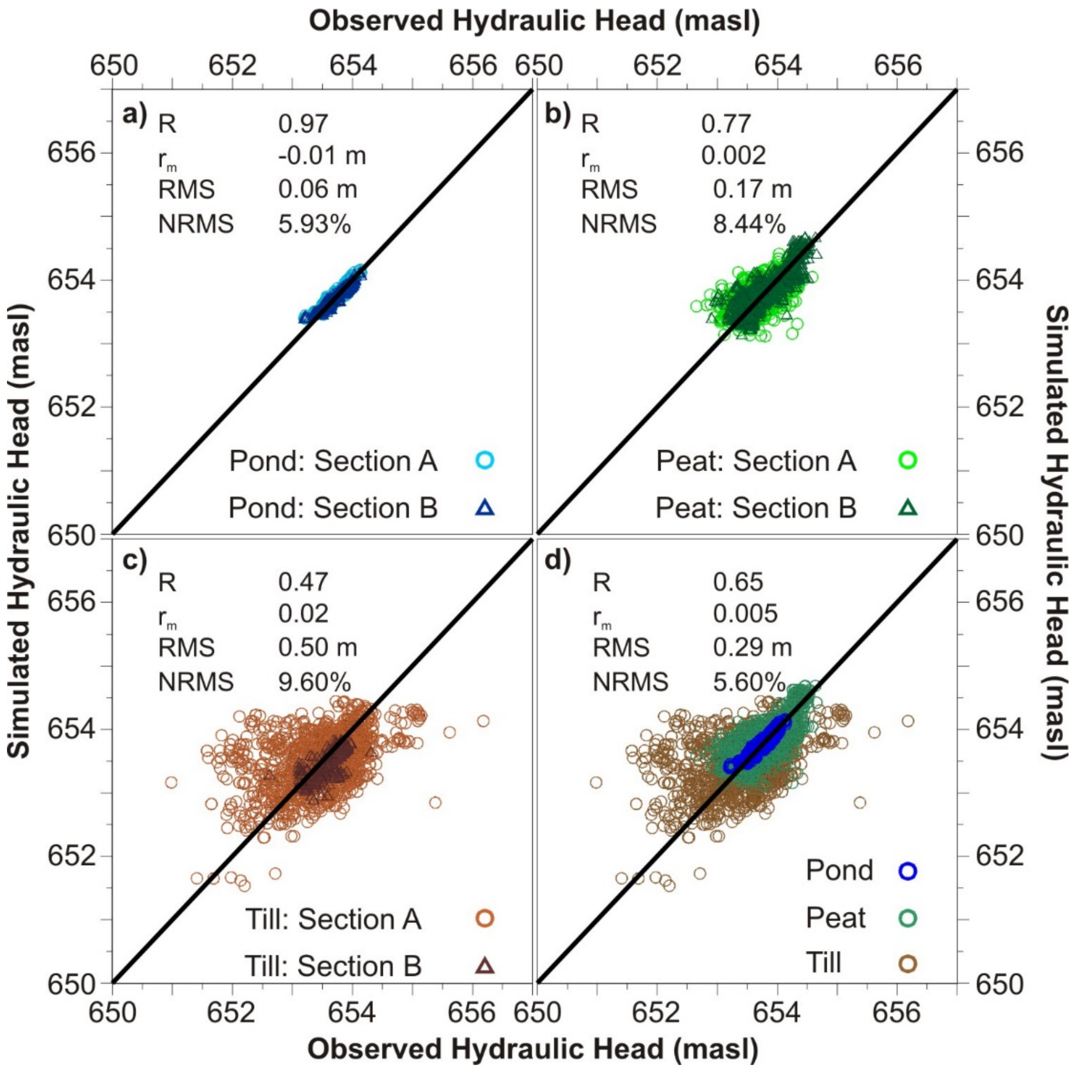

Figure 2-6. Comparison of simulated and observed water levels within a) the pond, b) peatlands, c) glacial till, and d) whole model.

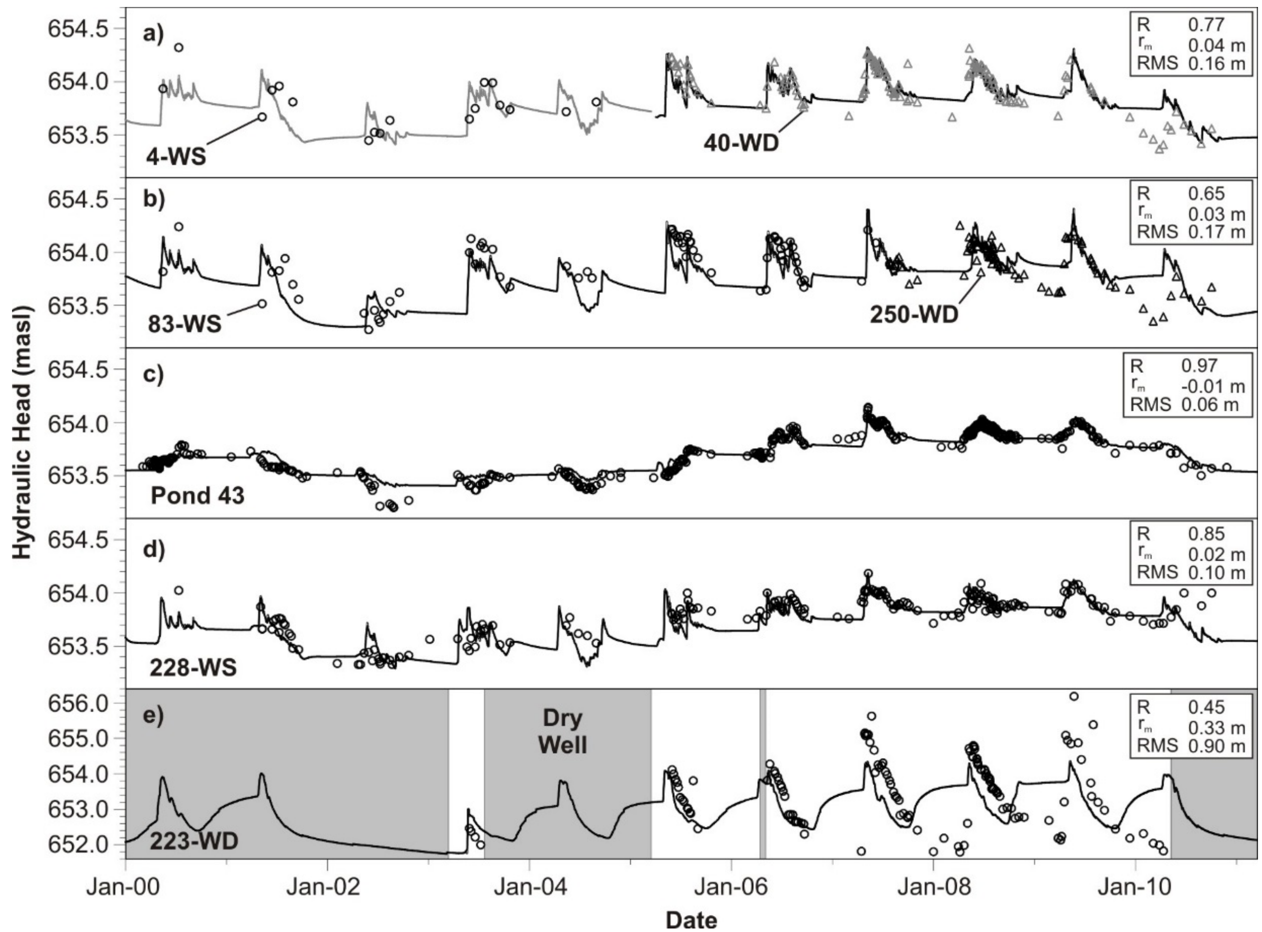

Figure 2-7. Comparison of simulated (solid lines) and observed (symbols) water levels for Pond 43 and selected monitoring wells. Dry wells are indicated by breaks in lines (simulated) and shaded periods (measured). Well locations are shown in Figures 2-1 and 2-2.

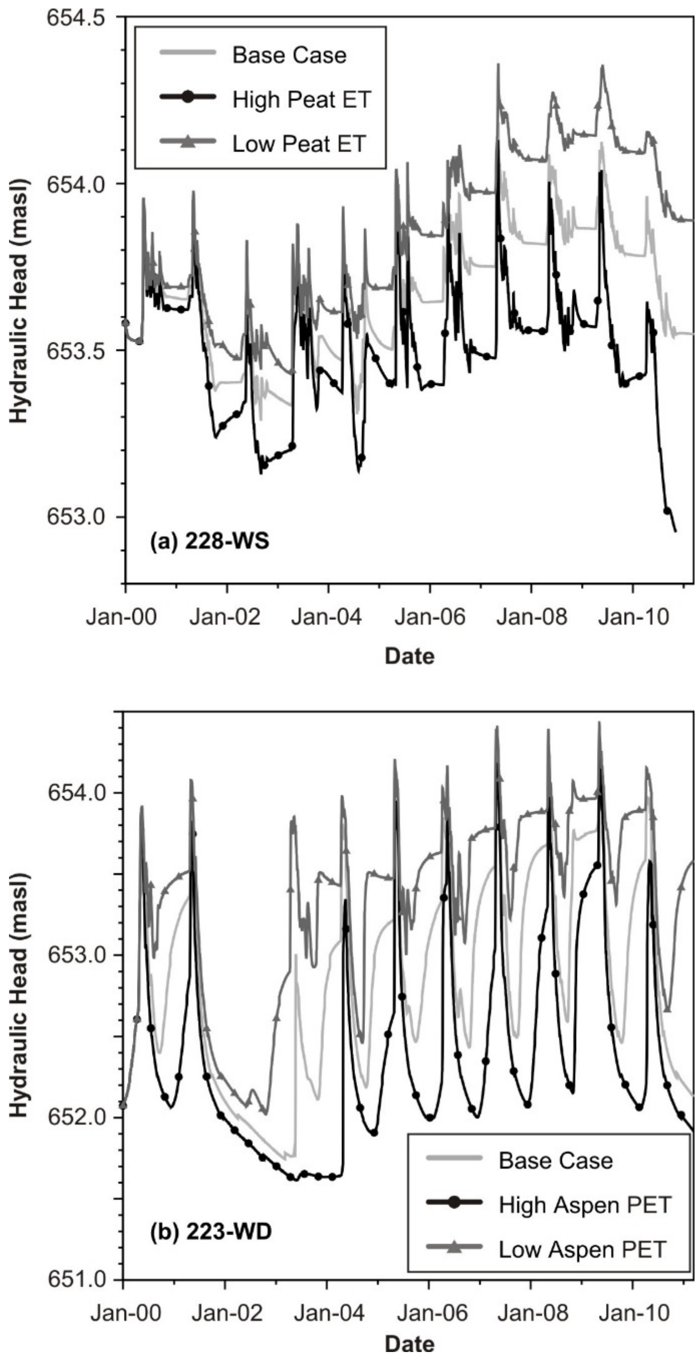

Figure 2-8. Predicted sensitivity to evapotranspiration (ET): comparison of simulated water levels for simulations with a) the ET of the peat modified by 30% and b) the ET of the aspen modified by 30%. Simulated dry wells are indicated by breaks in lines.

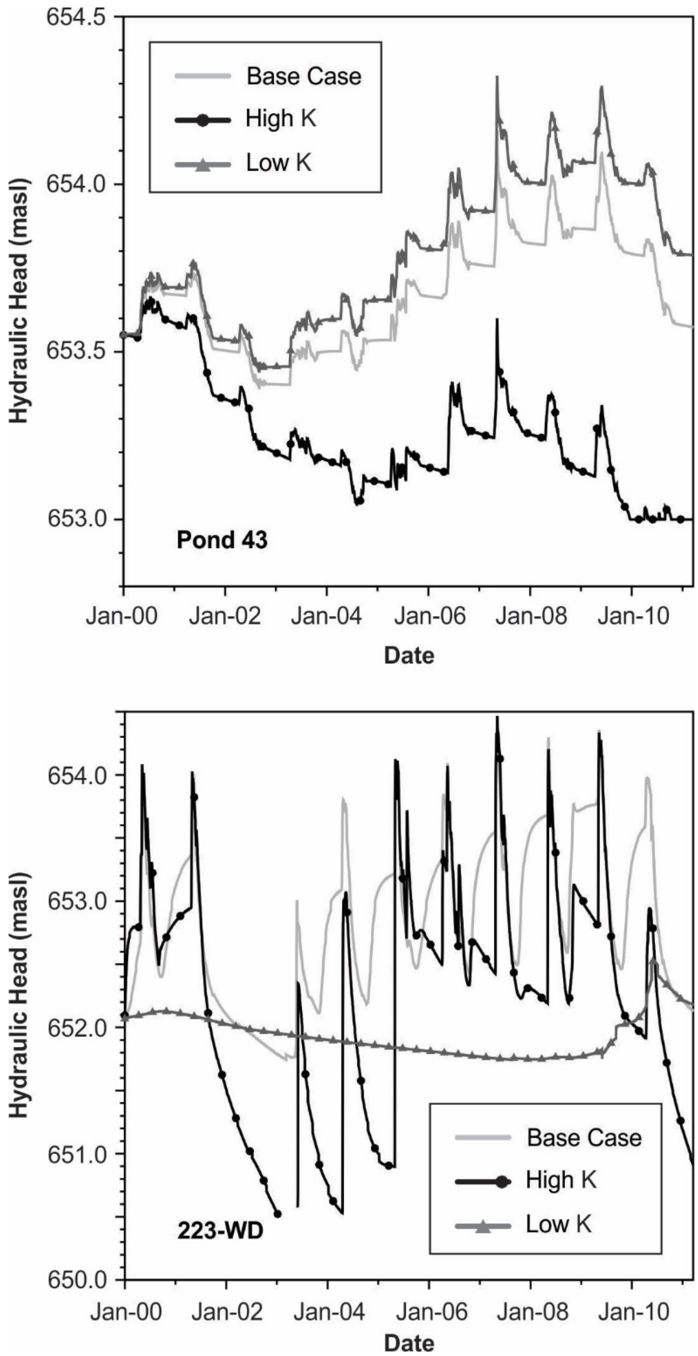

Figure 2-9. Predicted sensitivity to glacial till hydraulic conductivity (K): comparison of simulated water levels for simulations with the K of each glacial till unit increased and decreased by one order of magnitude. Simulated dry wells are indicated by breaks in lines.

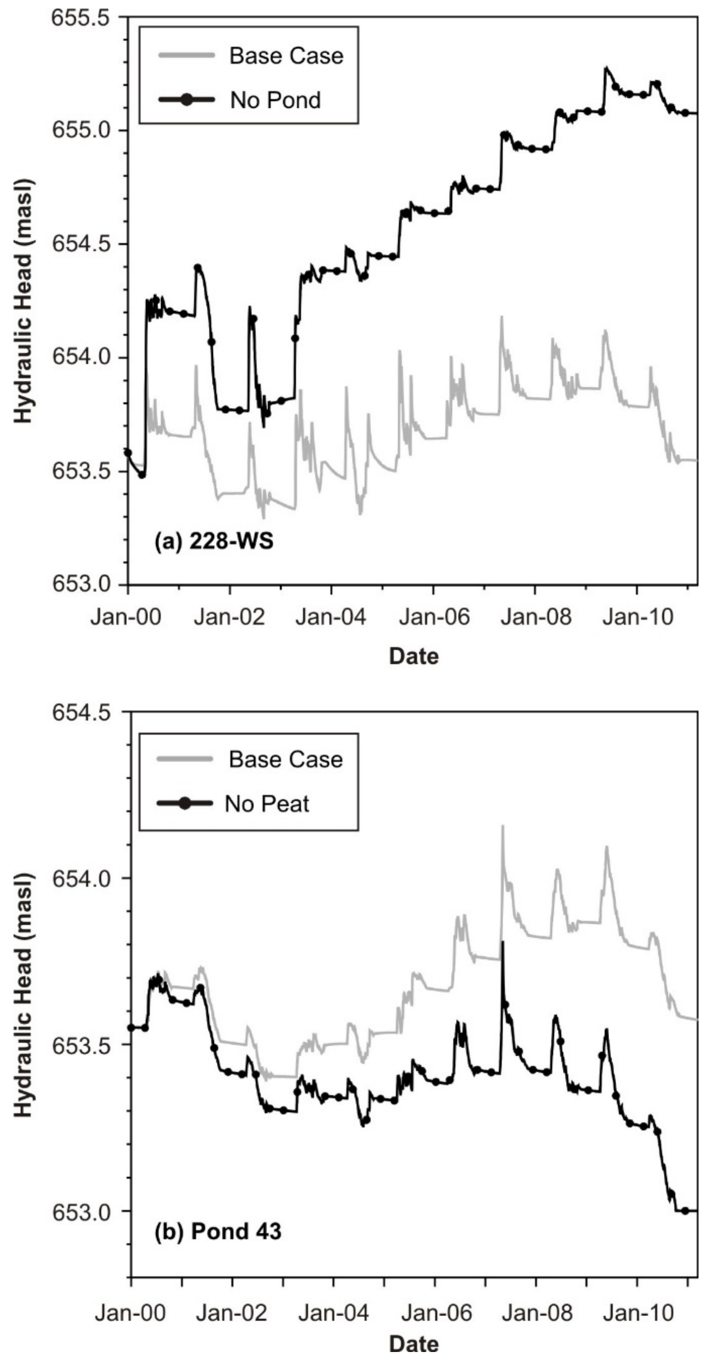

Figure 2-10. Predicted sensitivity to pond and peatland landscape units: comparison of simulated water levels for simulations with a) the pond replaced with a continuous peatland and b) the peatlands replaced by aspen forested glacial till.

CHAPTER 3

Influence of Seasonally Frozen Peatlands on Water Distribution in Alberta's Boreal Plains

3.1. Introduction

The Boreal Plains of north-central Alberta are composed of a mosaic of peatlands, ponds, and upland forests (NRC, 2006; Devito *et al.*, 2017) situated within a sub-humid climatic zone (Marshall *et al.*, 1999) where water deficit conditions are frequent (Devito *et al.*, 2005b). Located adjacent to the discontinuous permafrost region (Woo and Winter, 1993), seasonal freezing of the peatlands is an important process influencing ecosystem water availability, including rates of evapotranspiration, overland flow, and infiltration (Smerdon and Mendoza, 2010), which in turn has implications for overall ecosystem health and sustainability (Chapter 2). Peatlands play an important role in the hydrologic functioning of Boreal Plains catchments, supplying water to ponds, streams, and adjacent hillslopes following snowmelt and rain events, and conserving water within the landscape during periods of drought (Chapter 2; Gracz *et al.*, 2015; Devito *et al.*, 2017). Thus, understanding the role of seasonal peatland freezing on shallow Boreal Plains groundwater flow dynamics, both now and in a future warmer climate, is important to protect the region's vital habitat for a suite of organisms (Petrone *et al.*, 2007) and commercially valuable natural resources (Devito *et al.*, 2012).

Seasonal peatland freezing influences shallow groundwater flow dynamics by lowering both the effective hydraulic conductivity and the water storage potential of the peat as pores are blocked by ice (Woo and Marsh, 2005). At the land surface, the result is restricted infiltration with increased potential for surface ponding and overland flow (Hayashi, 2013; Ireson *et al.*, 2013). Seasonal peatland freezing also plays an important role in moderating groundwater-surface water interactions and shallow pond permanence (Smerdon and Mendoza, 2010), restricting lateral peatland seepage losses (Chapter 2), increasing early growing season moisture conditions and evapotranspiration through perched water and melting ice (Petrone *et al.*, 2007; Brown *et al.*, 2010), moderating wildfire burn depth and severity (Turetsky *et al.*, 2011), decreasing peatland compression (Petrone *et al.*, 2008), plant and soil fauna productivity (Hayashi, 2013), carbon sequestration and release (McKenzie *et al.*, 2007a), and solute dynamics (Hayashi, 2013; Ireson *et al.*, 2013).

Research conducted in northern climates has led to a thorough understanding of many hydrological, biological, and geochemical processes in permafrost settings (e.g., Quinton *et al.*, 2018); however, comparatively fewer published examples are available for non-permafrost areas with seasonally frozen peat. Furthermore, quantification of peatland hydrologic processes influenced by seasonal freezing using models remains a major gap (Ireson *et al.*, 2013), although there has been renewed interest in the literature in recent years (e.g., McKenzie *et al.*, 2007b; Atchley *et al.*, 2016; Painter *et al.*, 2016). Nevertheless, the significance of peatland freezing on water partitioning and evapotranspiration remains a knowledge gap in non-permafrost areas of the ecologically sensitive Boreal Plains region (Ireson *et al.*, 2015). Furthermore, development of a holistic understanding and quantification of the influence of peatland ice is vital not only for the proper management of ecosystems and natural resources within the Boreal Plains region, but it may also provide crucial insights for predicting future hydrologic dynamics in more northern locales as the prevalence of permafrost declines due to climate change.

In this study, field measurements collected from 2000 to 2011 were used in conjunction with loosely coupled numerical simulations to quantify the influence of seasonal peatland freezing on shallow groundwater flow dynamics at a catchment characteristic of fine-grained, low hydraulic conductivity, glacial moraine within Alberta's Boreal Plains. The goals of the study were to:

- Quantify the influence of seasonal peatland freezing on peatland water table position and water distribution, including evapotranspiration rates.
- Demonstrate how peatland hydrology may be influenced by variations in ice depth and continuity.

Numerical simulations were based on previous simulations results (Chapter 2), with seasonal peatland freezing represented by reduced peatland hydraulic conductivity and storage parameters based on observations of peatland ice depth and the results of coupled simulations of groundwater flow and heat transport with water-ice phase change (Appendix A).

3.2. Study Area

The study was conducted in a small Boreal Plains catchment (Lat: 56.07 N, Long: 115.5 W) within the Utikuma Region Study Area (URSA; Figure 3-1A) located approximately 350 km northwest of Edmonton, Canada and approximately 150 km south of the discontinuous permafrost region (Woo and Winter, 1993). The regional climate is sub-humid, with annual potential evapotranspiration (517 mm) greater than precipitation (485 mm) on average (i.e., 1961-1990; Marshall *et al.*, 1999). Monthly temperatures range from -15 °C (January) to 16 °C (July) with an annual average of about 1 °C (Marshall *et al.*, 1999).

The study catchment consists of a central shallow pond (i.e., Pond 43) underlain by about 3 m of gyttja that is surrounded by peatlands reaching depths up to 5 m and forested hillslopes of predominately aspen with about 10 m of topographic relief. Surface drainage is poorly developed, ephemeral, and frequently altered by beavers. The underlying subsurface consists of thick (i.e., 40 to 50 m), heterogeneous, generally low hydraulic conductivity, glacial disintegration moraine (Pawlowicz and Fenton, 2005) that is underlain by marine shales of the Upper Cretaceous Smoky Group (Vogwill, 1978).

Groundwater and surface water flow within the peatlands are influenced by seasonally frozen lenses, which can persist well into the summer months (Petrone *et al.*, 2008; Brown *et al.*, 2010; Chapter 2). In contrast, subsurface ice has been found to have negligible influence on catchment water distribution within the glacial materials, as the hydrologic response to snowmelt and spring rains is dominated by infiltration with low potential for overland flow (Redding and Devito, 2011).

3.3. Methods

3.3.1. Field Data

Hydrometric data collected within the study watershed, including measurements of precipitation, snowpack, rainfall interception, pond stage, and groundwater levels, including 75 peatland monitoring locations (Figure 3-1A), have previously been summarized (Ferone and Devito, 2004; Chapters 2 and 4). Additional data used in this study included measurements of air temperature and depth to ice within the peatlands. Air temperature was measured on a 30-minute interval and recorded using automated loggers for the duration of the study period. Depth to ice measurements were obtained from 2003 to 2009 by pushing a metal rod of known length through the peat to the ice surface. Using this method, the maximum depth of ice that could be detected was about 0.7 m. Weekly to monthly measurements were collected most years from early spring to late summer at peatland locations distributed across the study area.

3.3.2. Numerical Model

Two-dimensional (2D) numerical simulations were conducted using MODFLOW-SURFACT (HGL, 2015), a saturated/unsaturated finite-difference groundwater flow model based on the popular MODFLOW code (McDonald and Harbaugh, 1988). MODFLOW-SURFACT was selected as the numerical simulator for this study due to its ability to simulate both variably-saturated groundwater flow and time-varying material properties. The simulated model domain (Figure 3-1B) and horizontal discretization (i.e., 1 m) were the same as utilized in Chapter 2. The vertical discretization was modified to provide greater near-surface resolution for representing peatland ice, with grid blocks specified to be 0.1 m thick to a depth of 4.5 m. Below this depth, the thickness of grid blocks was gradually expanded to a maximum of 2.5 m at the base of the model. Simulations were performed using daily boundary conditions along with adaptive time-stepping that allowed for sub-daily time-steps.

Base case simulations were conducted from January 2000 to March 2011 for scenarios with (frozen scenario) and without (unfrozen scenario) peatland freezing and thawing to directly evaluate the influence of seasonal ice formation on peatland water table position and water distribution. Additional sensitivity simulations were performed to evaluate the influence of selected boundary conditions and peatland ice characteristics on the model predictions.

3.3.2.1. Material Properties

Simulated hydrogeologic units included peat, gyttja, and glacial till with 0.1 m of overlying forest floor on uplands. The extent of each hydrogeologic unit and the unfrozen parameter distribution were defined based on observed depth trends in hydraulic conductivity (Figure 3-1B and Table 3-1) and were unchanged from the simulations described in Chapter 2. Soil water characteristic and unsaturated hydraulic conductivity curves were derived from the literature for the peat, gyttja, and forest floor (Silins and Rothwell, 1998; Price *et al.*, 2010), and glacial till using the properties of a clay loam (Carsel and Parrish, 1988) consistent with the simulations in Chapter 2.

Within the frozen scenario, seasonal peatland freezing was simulated by transiently reducing peatland hydraulic conductivity, specific yield, and specific storage by a factor of 10^3 in grid blocks specified to be frozen. The duration and depth of frozen peat was specified based on the results of simulations conducted using SUTRA (Voss and Provost, 2010) that incorporated groundwater flow and heat transport with water-ice phase change (McKenzie *et al.*, 2007b; Kurylyk *et al.*, 2014) that was calibrated to measured peatland ice depth (Appendix A; Figure 3-2). These simulations were completed over the same time period and model domain as the current analysis, and thus provided the framework for a loosely-coupled representation of the physical system. Freezing and thawing were assumed to occur completely across each grid block (i.e., 0.1 m depth interval) over the course of one day and to be spatially uniform as lateral differences in simulated peatland ice prevalence were generally negligible (Appendix A).

3.3.2.2. Boundary Conditions

Daily boundary conditions were consistent between the frozen and unfrozen scenarios and were applied to the model domain based on the results Chapter 2. Within the peatlands, net precipitation (i.e., snowmelt and rainfall) that accounted for sublimation and interception was applied to the peat surface, while evapotranspiration constrained by water table depth was specified to be active during snow-free periods. To quantify surface ponding and overland flow, seepage at the peat surface was simulated using head-dependent boundary conditions restricted to only discharge (i.e., drain). Pond 43 was simulated using head-dependent boundaries that allowed exchange with the groundwater system below the specified pond stage and was restricted to outflow above. Along the surface of the aspen-covered hillslope, a no-flow boundary was assumed because observed and simulated hillslope water levels have generally remained well below the base of the peatlands (Chapter 2). No-flow boundaries were also specified along the lateral edges of the model where groundwater flow is predominately vertical, and a constant head boundary was specified along the model base, consistent with previous simulations (Chapter 2).

3.4. Results

3.4.1. Field Observations

3.4.1.1. Temperature

Temperature trends were relatively consistent between years, with average daily temperatures ranging from lows of -30 to -40 °C during the winter months to highs of 20 to 25 °C during the summer months (Table 3-2). Seasonally, the winter period commenced in late October to early November when temperatures dropped below 0 °C and extended until late March to April (Figure 3-2A). Despite persistent sub-zero temperatures over the winter period, appreciable variation in daily temperatures occurred between years, ranging from -40 °C to around 0 °C. From late March through May, temperatures gradually rose above 0 °C and consistently remained above freezing by mid-May. On an annual basis, average temperatures ranged from -2 °C in 2002 to almost 4 °C in 2006.

3.4.1.2. Water Levels

Water levels within the peatlands and pond mirrored recent trends in precipitation (Table 3-2; Chapter 2) and fluctuated through an approximately 1 m range during the study period that varied from just below ground surface to about 1 m depth (Figure 3-2B). Water levels were lowest in 2002, following two consecutive years with below average precipitation. In 2005, water levels rose in response to above average precipitation and remained elevated through 2009 as wetter conditions persisted. However, in 2010 water levels began to decline again following a low snowpack (Table 3-2). On a seasonal basis, water levels typically increased rapidly following spring snowmelt, before gradually declining throughout the growing season. These trends were generally consistent across the study period, except for 2002 when no snowmelt response was observed at monitoring locations across the study area.

3.4.1.3. Peatland Ice

During the study period, peatland ice was generally encountered at shallow depths of less than 0.05 m prior to snowmelt (Figure 3-2C). Following snowmelt, appreciable differences in peatland ice depth were observed between years, reflecting differences in both the onset (i.e., mid-April to mid-May) and rate (0.006 to 0.014 md^{-1}; Table 3-3) of ice recession. In all years, no ice was detected beyond early August; however, the maximum depth of ice that could be measured was limited to about 0.7 m.

3.4.2. Numerical Simulations

Simulated peatland water table position and water table trends were generally similar to available observations (e.g., Figure 3-2B) and were consistent with previous simulation results (Chapter 2). As shown for 2006 (Figure 3-3B), both the observed and the simulated peatland water table rose each spring in response to snowmelt and early spring rains before generally declining throughout the growing season (Figure 3-3B). However, in 2002 the observed peatland water table did not rise in response to snowmelt (Figure 3-2A), resulting in the simulated peatland water table position being overpredicted for both frozen and unfrozen scenarios (Figure 3-3A). Although consistent with previous simulations (Chapter 2), a potential cause for the discrepancy between observed and simulated peatland water table position in this year was investigated using a sensitivity simulation described later.

Simulation results indicate that seasonal peatland freezing supports higher spring evapotranspiration rates (Figure 3-4) and increased discharge at the peat surface as surface ponding and overland flow (Figure 3-5) by maintaining higher water table conditions. The influence of peatland ice is predicted to be greatest during dry years (e.g., 2002), where the frozen peat results in a higher early spring peatland water table (Figure 3-3A) and evapotranspiration rates that are almost doubled (Figure 3-4B) for the frozen scenario relative to the unfrozen scenario. Similarly, the frozen peat is also predicted to restrict infiltration during the early spring, leading to discharge at the peat surface (Figure 3-5B). During wetter periods (e.g., 2006), generally negligible difference in evapotranspiration rates are predicted between frozen and unfrozen scenarios (Figure 3.4C), as the peatland water table position is predicted to be similar between scenarios (Figure 3-3B). However, early spring peat surface discharge is still predicted to be greater for the frozen case (Figure 3-5C). On a seasonal basis, the frozen peat is predicted to exert the greatest influence on the water table position and water fluxes during the spring when the depth to ice is shallow. As the ice was simulated to recede from spring to mid-summer, negligible differences in peatland water table position, evapotranspiration rates, and surface discharge are predicted.

Seasonal peatland ice is also predicted to influence subsurface water exchange between the peatland and pond (Figure 3-6BC). Peak discharge rates from the peat to the pond are predicted to increase for the frozen scenario relative to the unfrozen scenario during spring snowmelt and rain events, similar to peatland surface discharge. However, during the frozen season the reduced hydraulic conductivity of the frozen peat is predicted to reduce discharge to the pond resulting in the cumulative discharge to the pond being generally lower for the frozen scenario on an annual basis (Figure 3-6A).

3.4.2.1. Sensitivity Simulations

3.4.2.1.1 2002 Snowmelt

The position of the simulated peatland water table was overpredicted in 2002 in both the frozen and unfrozen scenarios following spring snowmelt (Figure 3-3A), when the observed peatland water table did not rise in response to snowmelt. This response was consistently observed in peatland monitoring wells completed across the study area (Figure 3-1). Annually, 2002 was the coldest year during the 10-year study period (Table 3-2). In combination with the low spring water table that was present, it is possible that ice formation effectively isolated the saturated subsurface from infiltrating snowmelt during this year, thereby preventing an increase in spring peatland water levels. Anecdotal evidence of ponded water at the peat surface in 2002 support this hypothesis; however, detailed observations are not available to confirm.

To test this hypothesis, the base case frozen scenario was run with 2002 snowmelt removed from the simulation under the assumption that the meltwater was lost as overland flow. Alternatively, the meltwater could have collected in depressions in the peat surface and either infiltrated as the ice receded or was lost as evaporation. Unfortunately, field observations such as early spring stream flow measurements are not available to confirm or refute these alternatives. Nevertheless, simulation results (Figure 3-7) indicate that the hypothesized scenario may be viable, as the predicted peatland water table position in 2002 is improved relative to the base case frozen scenario (Figure 3-7C) providing a better representation of field observations. Predicted peatland water table position in subsequent years also remains consistent with available observations (Figure 3-7D). Because of the loss of meltwater, predicted evapotranspiration in 2002 is appreciably reduced (Figure 3-7A) and no further discharge is predicted at the peat surface throughout this year.

3.4.2.1.2 Peatland Ice Depth

The extent of peatland ice within the base case frozen scenario was specified based on the results of loosely-coupled simulations (Appendix A); however, only data for the depth to top of ice were collected and data are not available to confirm the estimated base of the frozen peat or frozen thickness. To evaluate the influence of peatland ice thickness, a sensitivity simulation was conducted with the depth of ice restricted to 0.5 m below ground surface relative to 0.8 m to 1.8 m below ground surface in the base case frozen scenario.

Results of this sensitivity simulation indicate that reduced peatland ice thickness has relatively minor influence on discharge at the peat surface (Figure 3-8A) and peatland evapotranspiration (results not shown), as negligible difference in the peatland water table position is generally predicted throughout the majority of each year (Figure 3-8C and D). However, the reduced ice depth is predicted to lead to greater subsurface connectivity between the peat and pond, with higher net discharge to the pond predicted for the sensitivity case in most years (Figure 3-8B).

3.4.2.1.3 Peatland Ice Continuity

Within the base case frozen scenario, seasonal peatland ice formation was assumed to be laterally continuous. However, ice formation within the peat may be influenced by local-scale differences in saturation due to heterogeneities in peat properties (e.g., hydraulic conductivity, degree of decomposition) and hummock-hollow microtopography (Waddington *et al.*, 2015), potentially resulting in the formation of discontinuous ice lenses.

To evaluate the influence of peatland ice continuity, two sensitivity simulations were conducted with peatland ice specified to form in either 2 m or 5 m long segments separated by 1 m segments of unfrozen material. Simulation results indicate that the degree of peatland ice continuity has limited influence on the predicted peatland water table (Figure 3-9CD) and peatland evapotranspiration (results not shown). Peatland ice continuity is predicted to influence discharge at the peatland surface, with generally reduced water discharged as the simulated ice segment length declines (Figure 3-9A). Peatland ice continuity is also predicted to influence subsurface peatland-pond exchange (Figure 3-9B); however, the difference in the magnitude of exchange between scenarios is small at less than 3 mm.

3.5. Discussion

3.5.1. Spring Snowmelt and Precipitation Partitioning

Simulation results indicate that seasonal freezing of the peatlands can support higher evapotranspiration rates in spring (Figure 3-4) by restricting vertical infiltration and maintaining higher water table positions (Figure 3-3) and consequently greater near-surface soil moisture. Brown *et al.* (2010) came to similar conclusions based on the results of field studies conducted within the URSA, where the spring peatland water table was found to be perched above frozen peat before steadily declining as the ice receded during the growing season. However, where sufficient near-surface ice remains to prevent snowmelt and early spring rains from infiltrating, as may have occurred in the spring of 2002, the loss of spring meltwater may cause the peatlands to become water-stressed in the early growing season with reduced evapotranspiration (Figure 3-7A) and increased susceptibility to fire (Waddington *et al.*, 2015). Moreover, associated decreases in peatland water table position may have further adverse effects on neighboring ecosystems through reduction of subsurface groundwater discharge to lakes and ponds (Chapter 5), modification of nutrient loading rates as greater depths of warmer peat are exposed to active groundwater flow and aeration (Kane *et al.*, 2010; Plach *et al.*, 2016), and reduced water availability for trees and shrubs at the peatland edge (Chapter 2).

Seasonal peatland freezing is also predicted to result in greater discharge of water at the peat surface as surface ponding and overland flow (Figure 3-5), which may be retained locally within peatland depressions and eventually recharge the groundwater system or be lost as evaporation (Ireson *et al.*, 2013), or may be rapidly transmitted as overland flow over the frozen zone (Ireson *et al.*, 2015) with implications for flood generation (Woo and Marsh, 2005). Predicted discharge rates at the peat surface are influenced by the degree of ice continuity (Figure 3-9A), with the rate of discharge generally decreasing as the proportion of unfrozen material increases. Breaks in the frozen material will further contribute to smaller-scale differences in moisture distribution between peatland hummocks and hollows that in turn affect susceptibility to fire (Waddington *et al.*, 2015).

The frozen peat is also predicted to reduce the subsurface hydrologic connectivity between the peatlands and pond (Figures 3-6 and 3-8B). Within the study catchment, where net groundwater discharge from the peatlands to the pond is prevalent, the result is reduced discharge of water from the peatlands to the pond throughout much of the year.

During most years, reductions in subsurface groundwater discharge from the peatlands are predicted to be accompanied by increased spring peatland surface discharge (Figure 3-5), which may still replenish the pond through overland flow or shallow groundwater flow through the peat over underlying ice lenses, although the timing and nutrient quantity of water reaching the pond would be impacted as flow path lengths and associated residence times are altered (Plach *et al.*, 2016). However, ponded spring meltwaters and precipitation will be exposed to greater surface water evaporation rates (Petrone *et al.*, 2007), which may ultimately lead to reduced water reaching the pond.

In different hydrologic settings where flow-through conditions (Townley and Trefry, 2000) are prevalent at pond and peatland ecosystems, reduced subsurface pond-peatland connectivity may be important for restricting pond seepage losses and maintaining pond water levels (Smerdon *et al.*, 2007). Thus, in these settings, anticipated reductions in peatland ice due to climate change may deleteriously affect pond permanence (Ireson *et al.*, 2013), with diminished habitat for aquatic organisms, mammals, and migratory birds and the potential for encroachment of neighboring terrestrial vegetative species (Chapter 5).

3.5.2. Climate Change Considerations

The hydrologic functioning of Boreal peatlands may be particularly sensitive to a future warmer climate, as small temperature differences determine the state of precipitation, magnitude and timing of snow accumulation and melt (Carey *et al.*, 2010), and consequently the onset and duration of frozen substrate. Based on climate change projections for the study area and surrounding region (IPCC, 2013; Chapter 5), annual temperature is likely to increase by 2 to 6 °C by the 2080s while precipitation is projected to increase by 0 to 25%. However, in most projection scenarios less precipitation is predicted to fall as snow during the increasingly shorter and warmer winters. Snow is an effective insulator due to its low thermal conductivity (Ireson *et al.*, 2013) and moderates changes in ground temperature at the snowpack base due to fluctuations in daily air temperature (Zhang, 2005). Thus, the declining winter snowpack could be accompanied by greater peatland ice development as less insulation is provided at ground surface from sub-zero winter air temperatures. Such effects have been inferred from historical data at locations across Russia (Frauenfeld *et al.*, 2004). Conversely, projections from Canadian locations suggest that the influence of lower winter snowpack could be more than offset by rising winter temperatures, resulting in an overall reduction in ice development (Henry, 2008). Simulations by Lawrence and Slater (2010) further suggest that the warming climate and changing snowpack could result in either relative soil warming or cooling, depending on the magnitude of change experienced at a particular location. While not restricted to peatland locations, these studies indicate that the influence of changing climate and snowpack dynamics on future trends in peatland ice development, and consequently impacts on peatland hydrology, remains uncertain.

3.5.3. Study Limitations and Future Research

Study results provide an indication of the hydrologic influence of seasonally frozen peat on peatland water table position and water distribution in a catchment characteristic of fine-grained, low hydraulic conductivity, glacial moraine within Alberta's Boreal Plains. This analysis utilized simplified numerical simulations with peatland ice represented using a fixed approach that did not allow feedback between the thermal state of the system and the simulated groundwater flow system. Future research could benefit from use of an integrated groundwater-surface water model that incorporates fully coupled flow and heat transport (e.g., Advanced Terrestrial Simulator; Painter *et al.*, 2016), which could allow representation of processes such as rain-on-snow/ice events and mid-winter snowmelt that may play important roles on the thermal state of the subsurface, peatland ice persistence, overland flow generation, and water levels that were not incorporated into this analysis. Future studies could also benefit from inclusion of a robust representation of snowpack dynamics, which may have both a cooling (e.g., high albedo, high emissivity) or warming (e.g., high absorptivity of long waver radiation) influence on temperature at the land surface (Zhang, 2005; Ireson *et al.*, 2013).

Prediction and management of the future consequences of climate change on northern water resources represents a formidable challenge; however, further study is warranted as the natural resource-rich Boreal Plains is expected to be an area of maximum ecological sensitivity in the 21st century (Carey *et al.*, 2010; Ireson *et al.*, 2015). Consideration of the interaction and feedback among climatic, hydrologic, geochemical, and biologic processes (Hayashi, 2013) with a focus on quantifying their relative magnitude (Chapter 5) is the next step to identify probable scenario trajectories.

3.6. Conclusions

Evaluation of a pond-peatland complex situated within glacial moraine deposits in the Boreal Plains region shows seasonal ice formation influences peatland water table position, water distribution, and hydrologic connectivity. Simulation results indicate seasonal freezing maintains higher water table positions and surface ponding within the peatlands that can increase the potential for overland flow. Seasonal freezing also supports higher spring evapotranspiration rates but reduces the subsurface hydrologic connectivity between the peatland and pond. The degree of influence of the frozen peat is dependent on the relative timing of snowmelt and peatland ice recession. Where sufficient ice remains to prevent infiltration of spring snowmelt and precipitation, the reduced water available to the peatlands may have negative implications for peatland productivity and fire susceptibility, as well as hydrologic interactions with neighboring ecosystems. Future regional trends in peatland ice development and persistence will be dependent on both changes in temperature and precipitation that remain uncertain. Nevertheless, further study of shallow groundwater flow dynamics and groundwater-surface water interactions within areas of seasonal peatland ice formation such as the Boreal Plains is warranted as they may provide a natural laboratory for understanding future hydrologic dynamics in northern locales as the prevalence of permafrost declines due to climate change.

Table 3-1. Unfrozen parameters for simulated hydrogeologic units.

Material	Depth Range (m)	Hydraulic Conductivity (m/s)		Porosity (-)	Specific Storage (m^{-1})
		Horizontal	Vertical		
Peat	0.0 - 0.1	3×10^{-3}	3×10^{-4}	0.90	1×10^{-4}
	0.1 - 0.3	3×10^{-4}	3×10^{-5}	0.82	5×10^{-5}
	0.3 - 0.5	8×10^{-5}	8×10^{-6}	0.72	8×10^{-6}
	0.5 - 1.0	4×10^{-5}	4×10^{-6}	0.60	2×10^{-6}
	1.0 - 1.5	2×10^{-6}	2×10^{-7}	0.50	5×10^{-7}
	1.5 - 2.0	3×10^{-8}	3×10^{-9}	0.45	2×10^{-7}
	2.0 - Base	1×10^{-8}	1×10^{-9}	0.40	1×10^{-7}
Gyttja	0.0 - 1.0	1×10^{-6}	1×10^{-7}	0.45	3×10^{-6}
	1.0 - 2.0	3×10^{-8}	3×10^{-9}	0.30	4×10^{-7}
	> 2.0	5×10^{-9}	5×10^{-10}	0.22	1×10^{-7}
Glacial Till	Upper	1×10^{-5}	1×10^{-7}	0.20	1×10^{-4}
	Mid	5×10^{-7}	5×10^{-9}	0.20	1×10^{-4}
	Lower	1×10^{-8}	1×10^{-10}	0.20	1×10^{-4}
Forest Floor	0.0 - 0.1	1×10^{-4}	1×10^{-4}	0.80	1×10^{-4}

Notes:

[a] For frozen scenarios, peat hydraulic conductivity, specific yield, and specific storage were decreased by a factor of 10^3.

[b] Specified specific storage values for peat and gyttja are low; however, the influence on simulation results is negligible due to small fluctuations in water levels (i.e., less than 1 m).

Table 3-2. Summary of atmospheric conditions over the study period.

Year	Temperature (°C)			Precipitation (mm)		
	Average	Minimum	Maximum	Rain	Snow	Total
2000	1.7	-29.1	20.1	283	88	371
2001	1.8	-29.1	20.1	382	60	442
2002	-1.6	-36.8	21.3	164	93	257
2003	-0.3	-41.5	20.9	207	82	290
2004	0.7	-36.6	22.7	310	133	443
2005	2.2	-34.8	19.1	291	87	378
2006	3.8	-30.5	23.9	348	143	491
2007	1.6	-29.0	24.9	384	48	432
2008	1.5	-37.0	23.0	343	186	530
2009	0.2	-36.7	20.2	326	179	504
2010	2.7	-36.6	21.4	214	177	391
2011	1.1	-29.3	20.2	358	131	489

Table 3-3. Summary of peatland ice depth observations.

Year	Rate of Ice Recession (md^{-1})	Ice Persistence
2003	0.006	Mid to Late July
2004	0.006	Late July to Early August
2005	0.014	Mid-June
2006	-	-
2007	-	-
2008	0.011	Early July
2009	0.013	Mid-June

Notes:

[a] "-" indicates insufficient data available to estimate.

b Average ice recession rate.

c Ice persistence restricted by 0.7 m depth detection limit.

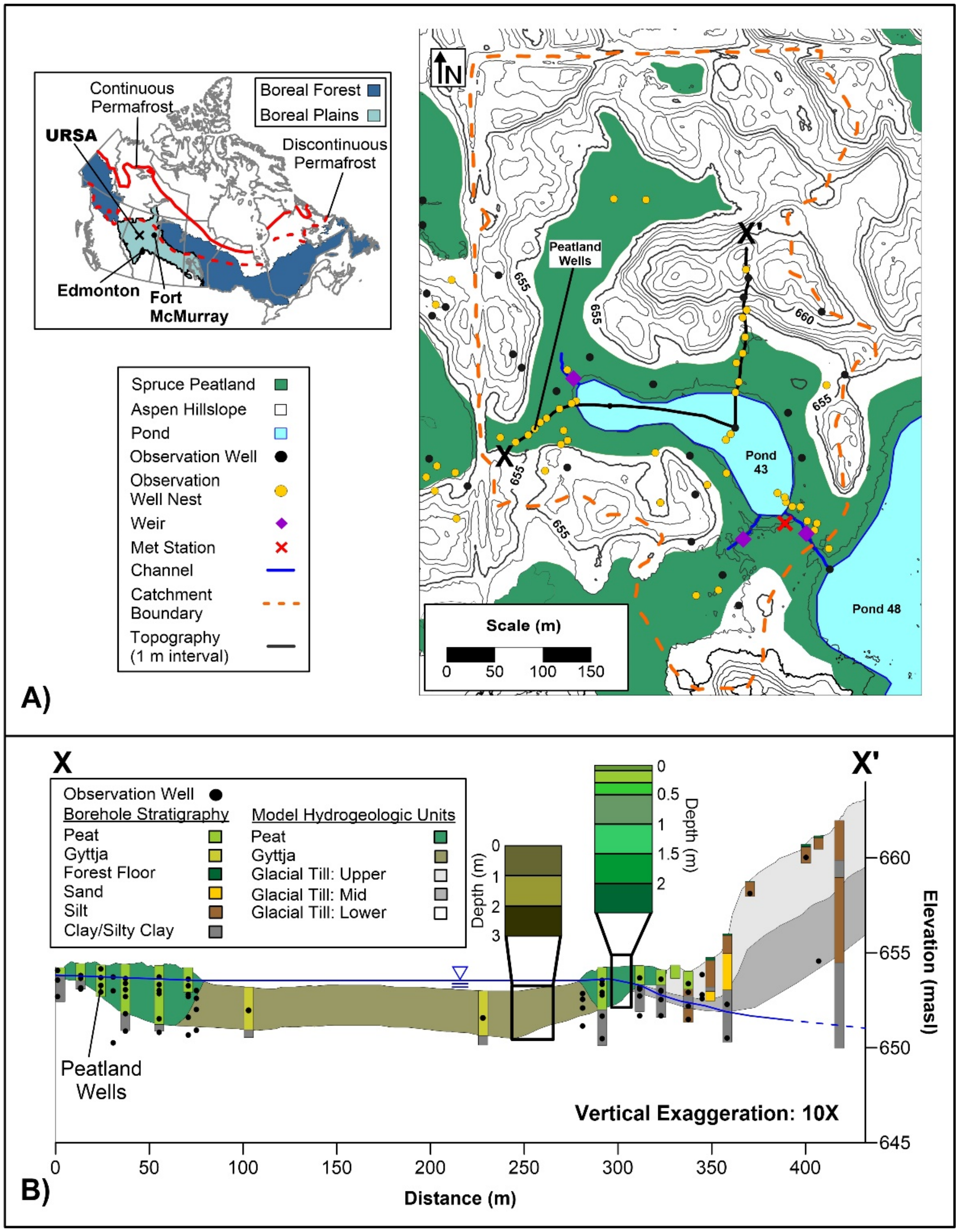

Figure 3-1. A) Location of the study area including instrumentation, vegetation classification, and location of the cross-sectional model. B) Model domain including hydrogeologic units, observation points, and average water table. Inset columns show layering specified within the peat and gyttja. Note that the base of the model has been truncated for illustration purposes.

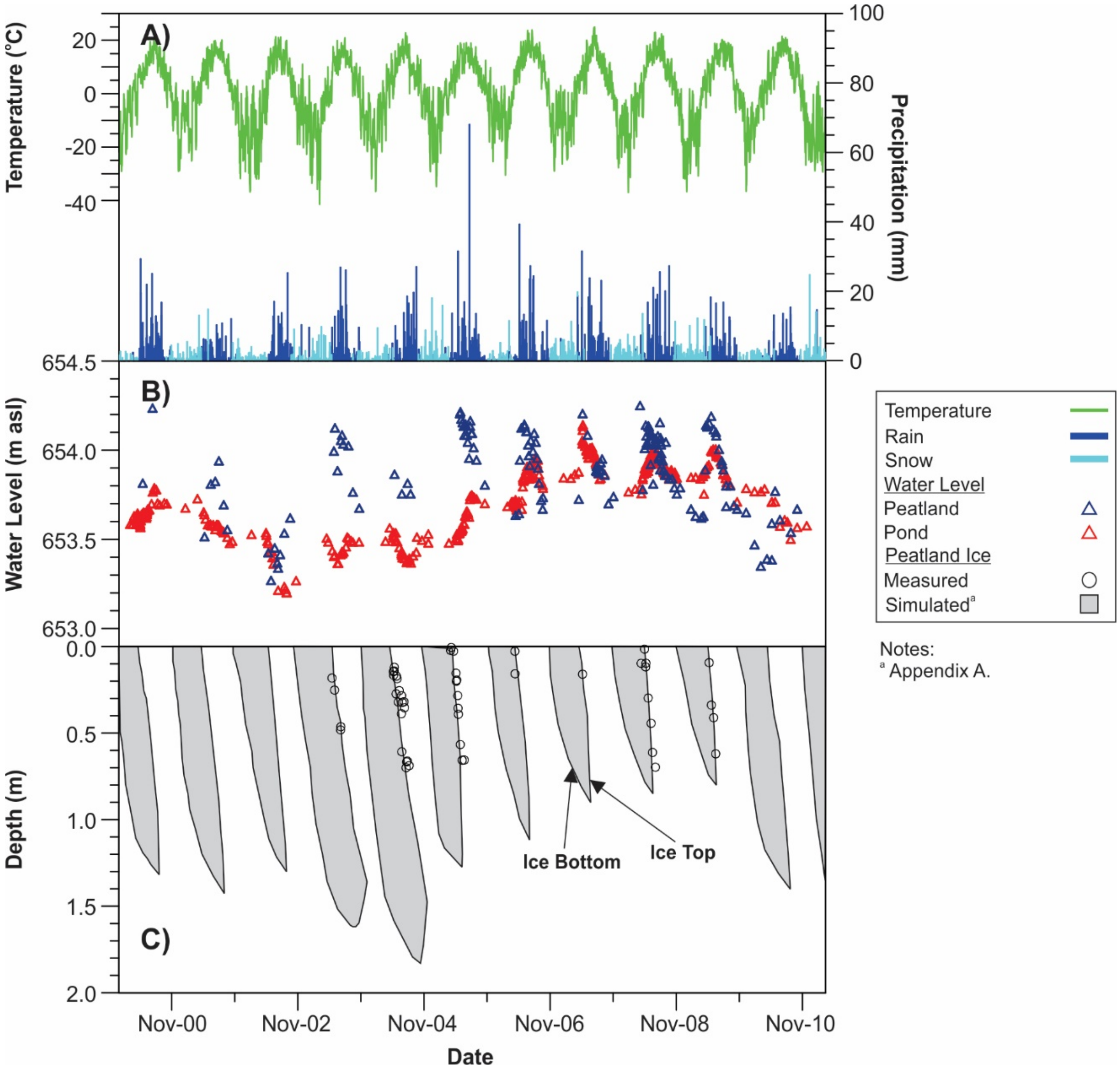

Figure 3-2. Summary of (A) daily temperature and precipitation, (B) peatland water table and pond stage, and C) distribution of peatland ice over the study period. The peatland measurement location is shown in Figure 3-1.

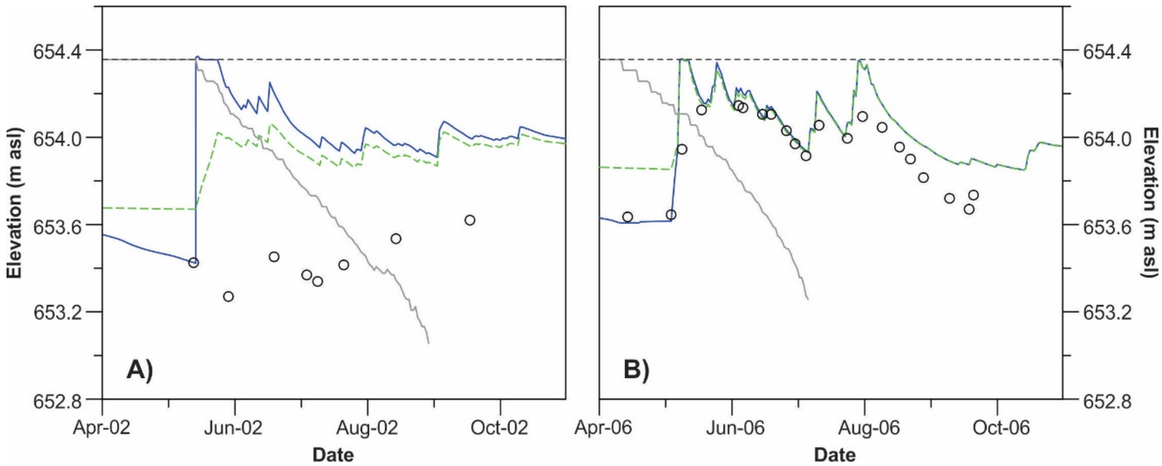

Figure 3-3. Simulated peatland water table for frozen (blue lines) and unfrozen (green lines) scenarios along with observed peatland water table (circles) and specified top of peatland ice (grey lines) for 2002 (A) and 2006 (B). Years 2002 and 2006 represent respective low and high peatland water table conditions (Figure 3-2). Peatland ground surface indicated by black dashed line and measurement location shown in Figure 3-1.

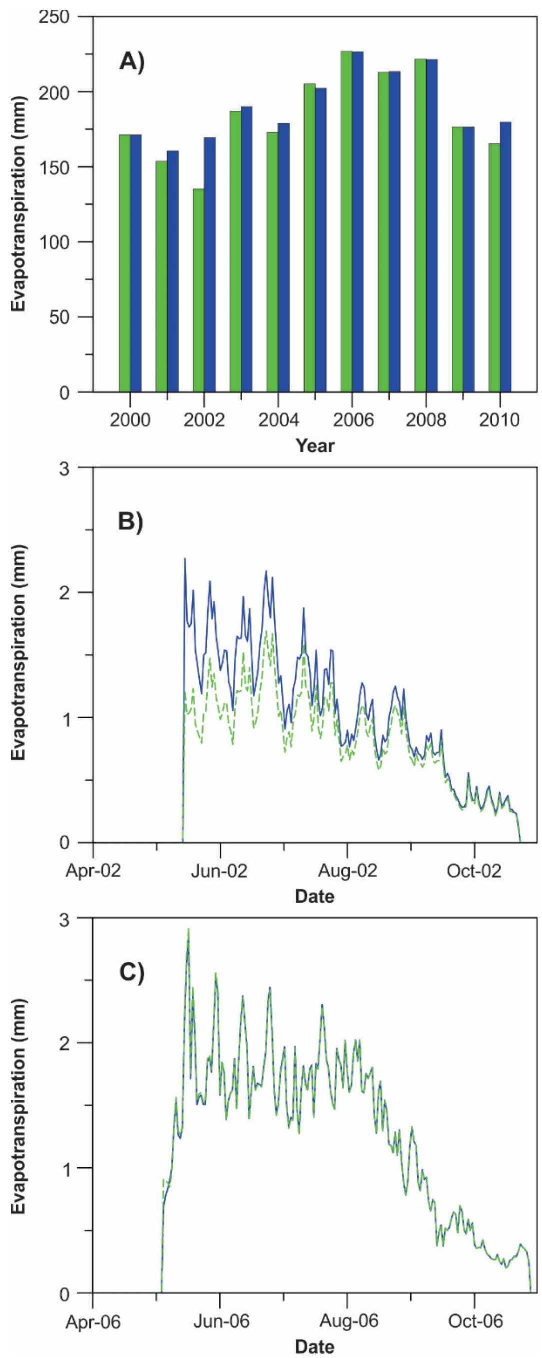

Figure 3-4. Annual peatland evapotranspiration (A) and daily peatland evapotranspiration for 2002 (B) and 2006 (C) simulated for frozen (blue) and unfrozen (green) scenarios. Years 2002 and 2006 represent respective low and high peatland water table conditions (Figure 3-2).

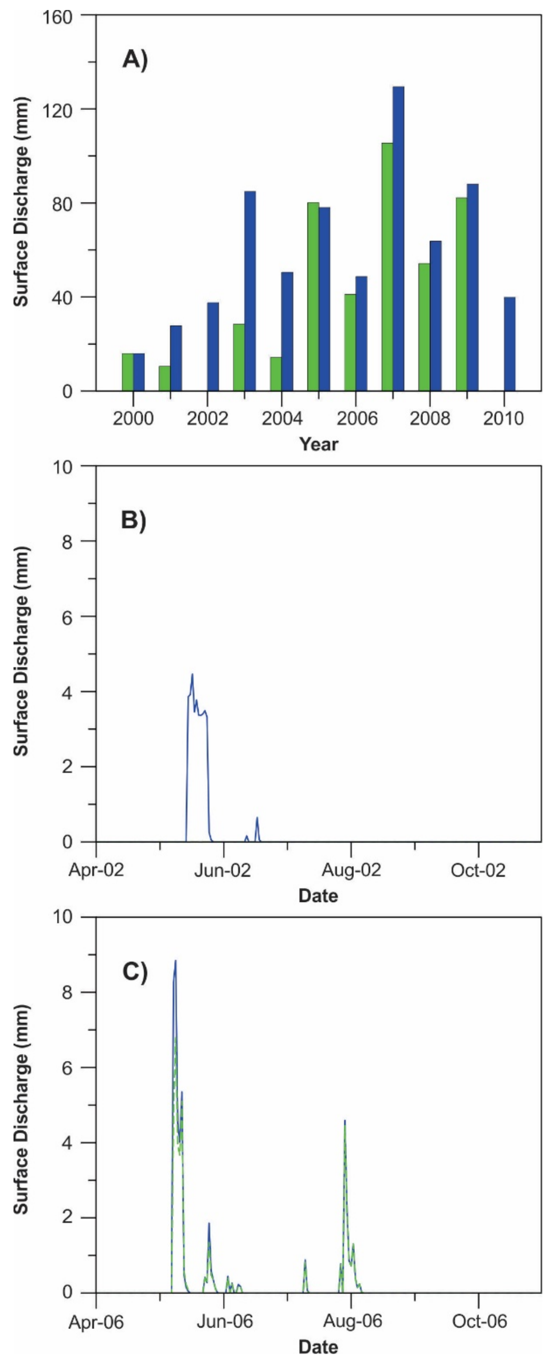

Figure 3-5. Annual peatland surface discharge (A) and daily peatland surface discharge for 2002 (B) and 2006 (C) simulated for frozen (blue) and unfrozen (green) scenarios. Years 2002 and 2006 represent respective low and high peatland water table conditions (Figure 3-2).

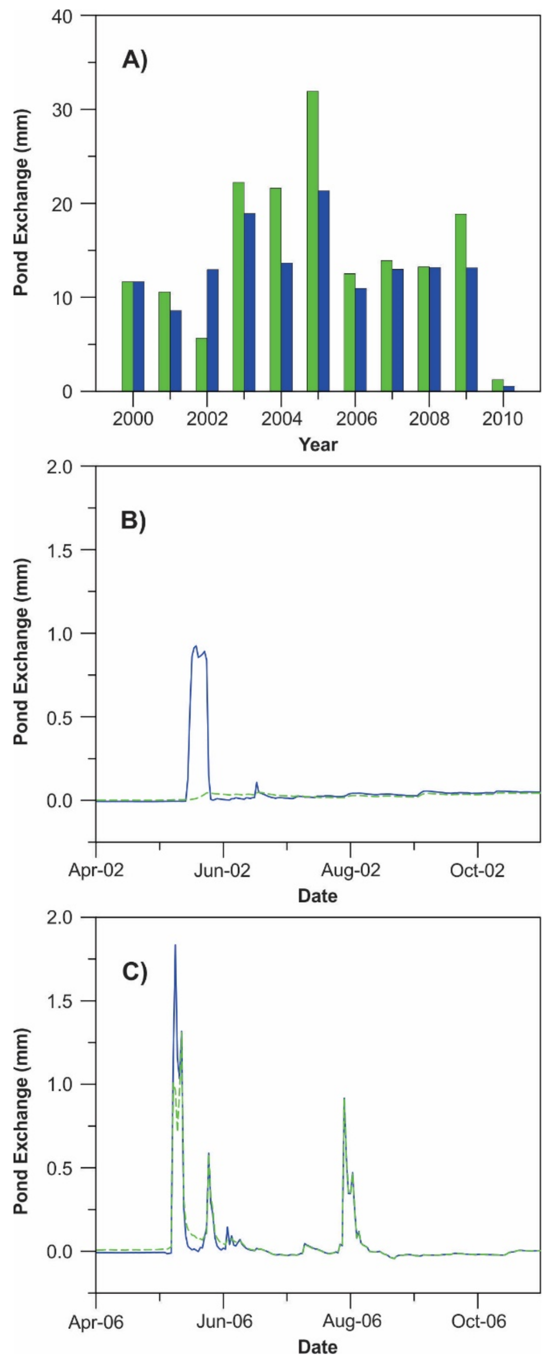

Figure 3-6. Annual net pond-peatland exchange (A) and daily net pond-peatland exchange for 2002 (B) and 2006 (C) simulated for frozen (blue) and unfrozen (green) scenarios. Years 2002 and 2006 represent respective low and high peatland water table conditions (Figure 3-2). Positive and negative values represent discharge to and seepage from the pond, respectively.

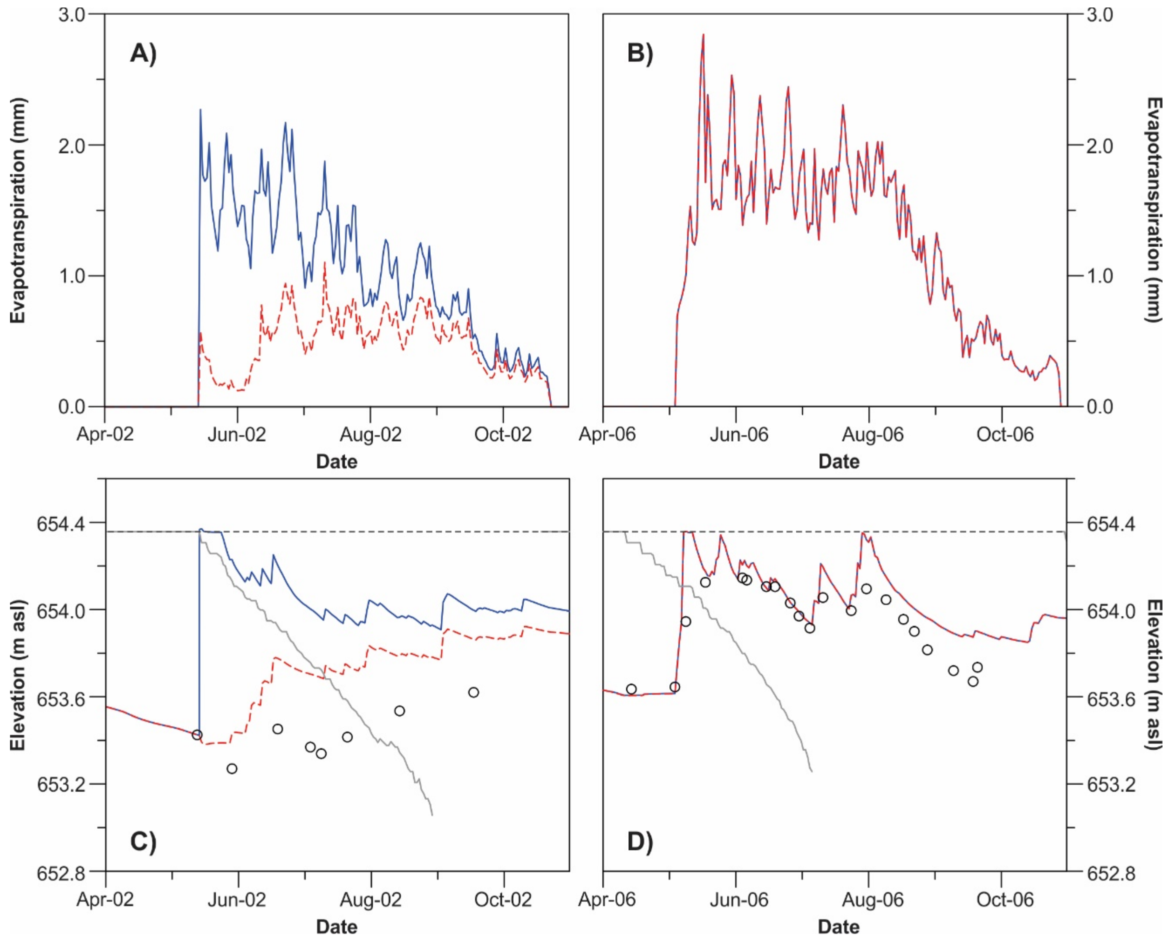

Figure 3-7. Predicted sensitivity to 2002 snowmelt: evapotranspiration (A and B) and peatland water table (C and D) for the frozen base case (blue lines) and sensitivity case (red lines) for 2002 (A and C) and 2006 (B and D). Observed peatland water table (circles) and specified top of peatland ice (grey lines) also shown. Years 2002 and 2006 represent respective low and high peatland water table conditions (Figure 3-2). Peatland ground surface indicated by black dashed line and measurement location shown in Figure 3-1.

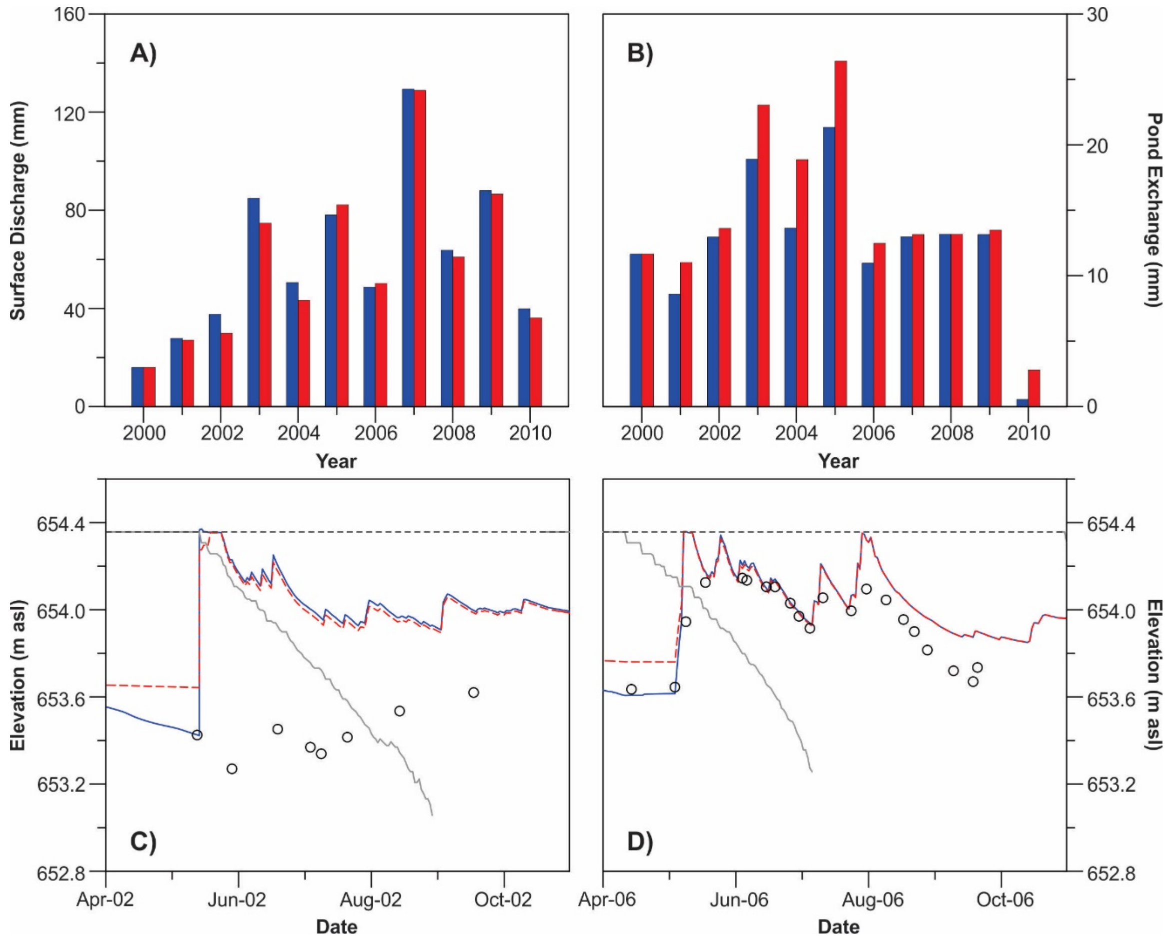

Figure 3-8. Predicted sensitivity to peatland ice depth: annual peatland surface discharge (A), annual net pond-peatland exchange (B), and peatland water table for 2002 (C) and 2006 (D) for the frozen base case (blue lines) and sensitivity case (red lines). Observed peatland water table (circles) and specified top of peatland ice (grey lines) also shown. Years 2002 and 2006 represent respective low and high peatland water table conditions (Figure 3-2). Peatland ground surface indicated by black dashed line and measurement location shown in Figure 3-1.

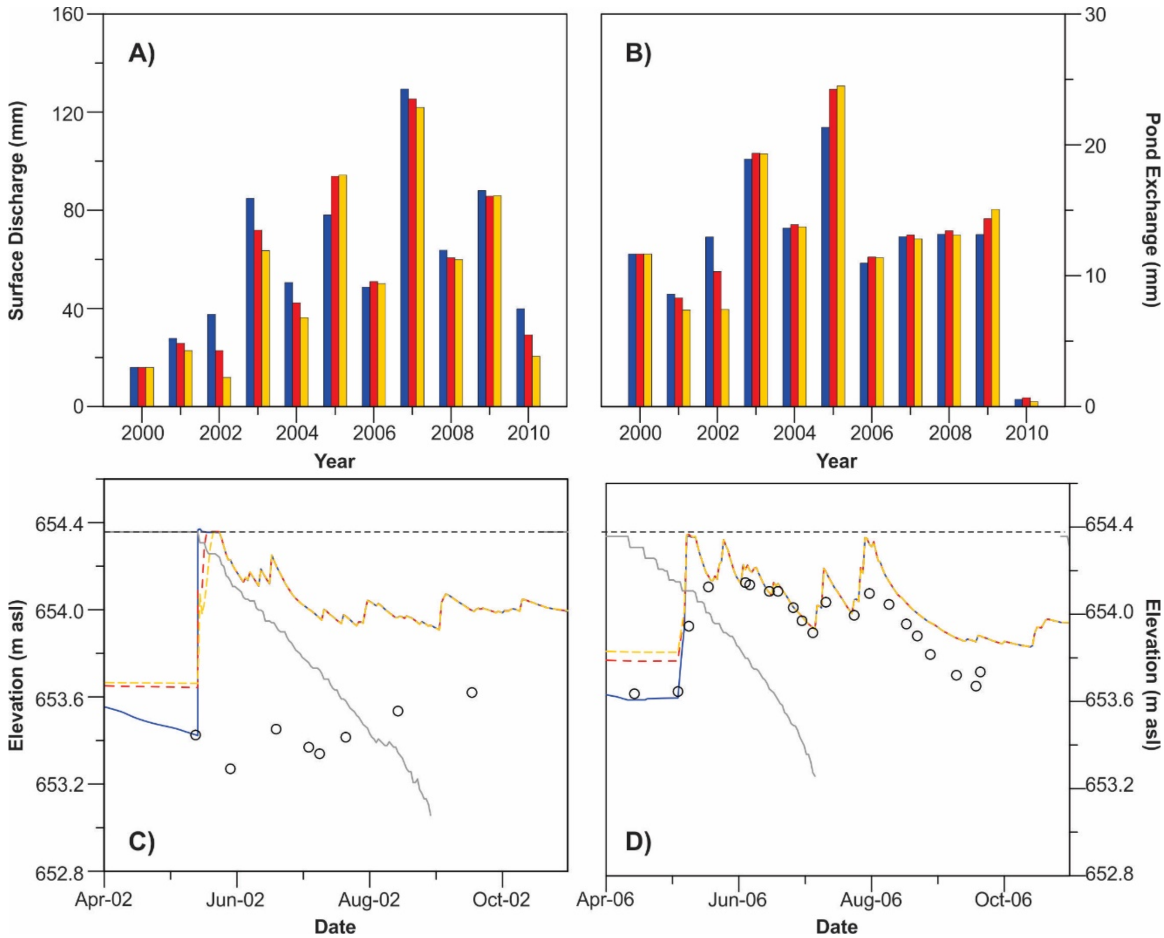

Figure 3-9. Predicted sensitivity to peatland ice continuity: annual peatland surface discharge (A), annual net pond-peatland exchange (B), and peatland water table for 2002 (C) and 2006 (D) for the frozen base case (blue lines) and sensitivity cases with 2 m (yellow lines) and 5 m (red lines) ice segments. Observed peatland water table (circles) and specified top of peatland ice (grey lines) also shown. Years 2002 and 2006 represent respective low and high peatland water table conditions (Figure 3-2). Peatland ground surface indicated by black dashed line and measurement location shown in Figure 3-1.

CHAPTER 4

Hydrologic Impact of Aspen Harvesting Within the Sub-Humid Boreal Plains of Alberta[3]

4.1. Introduction

Aspen (*Populus tremuloides*) forests occupy many uplands composed of fine-grained glacial deposits within the sub-humid Boreal Plains of Alberta, Canada. Commercial harvesting of these forests has accelerated since the 1980s (MacKenzie, 2010), with potentially detrimental effects on neighboring wetland ecosystems (Smerdon *et al.*, 2009). Furthermore, landscape disturbance may be enhanced by expansion of other resource developments (e.g., petroleum and mining; Devito *et al.*, 2012) and climate change (Chapter 5). Thus, the hydrologic impacts of aspen harvesting within the region, and how they vary relative to atmospheric variability, need to be understood to promote the long-term sustainability of the aspen resource and adjacent ecosystems.

Timber harvesting may directly alter ecosystem hydrology by lowering canopy interception and decreasing evapotranspiration (Buttle *et al.*, 2000). The resulting increase in available water can increase surface flow and groundwater recharge, and decrease soil-moisture storage potential. Within humid locales, where annual precipitation frequently exceeds potential evapotranspiration, the resulting response to harvesting is typically manifested by a rise in water table and an increase in stream flow (e.g., Dubé *et al.*, 1995; Jones, 2000; Moore and Wondzell, 2005) as thresholds required to exceed soil-moisture storage capacity are regularly surpassed. In drier climatic regions, such as the sub-humid Boreal Plains, results of harvesting studies have been variable, with increased stream flows and hillslope drainage observed at some locations (e.g., Swanson and Hillman, 1977; Kachanoski and De Jong, 1982) and little to no observable response reported at others (e.g., Devito *et al.*, 2005b; Whitson *et al.*, 2005).

The variability in results of studies within the Boreal Plains reflects interactions between climate, geology, and vegetation (Elliott *et al.*, 1998; Devito *et al.*, 2005a; 2005b).

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[3] A version of this chapter has been published:
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Within the sub-humid climate, the synchronicity of precipitation with peak potential evapotranspiration during the growing season leaves little excess water available to recharge the subsurface throughout most of the year. This effect leads to the persistence of deep water tables and thick unsaturated zones within forested hillslopes (Devito *et al.*, 2005b; Smerdon *et al.*, 2008; Chapter 2) that may buffer the system (Redding and Devito, 2008) and limit harvesting impacts (Carrera-Hernandez *et al.*, 2011). Furthermore, when combined with the low frequency of large storms and small runoff coefficients common to the region, harvesting may have limited potential to affect stream flows (Buttle *et al.*, 2009). Generalization of post-harvest response is further confounded by the heterogeneity of glacial sediments, which result in spatial differences in buffering capacity (Devito *et al.*, 2005b) and stream-flow generation mechanisms (Monteith *et al.*, 2006), along with differential growth characteristics of planted or regenerating tree species. Differential growth may result in rapid recovery of system evapotranspiration losses (e.g., several years for aspen; Elliott *et al.*, 1998; Devito *et al.*, 2005b; Whitson *et al.*, 2005) or more gradual recovery (e.g., upwards of 30 years for jack pine; Barr *et al.*, 2012).

While each of these climatic, geologic, and vegetative considerations also apply to other locales, their specific combination within the Boreal Plains make quantification of harvesting impacts difficult (Devito *et al.*, 2005b; Smerdon *et al.*, 2009), particularly when using traditional paired-catchment experiments (Buttle *et al.*, 2009). Nevertheless, distinguishing the signal of disturbance from that due to natural variability (i.e., atmospheric, geologic, vegetative effects) remains important for quantifying harvesting impacts on physical, chemical, and biological processes within these ecosystems (Buttle *et al.*, 2005), and may be improved through use of combined monitoring-modelling studies (Buttle *et al.*, 2005; 2009).

This study evaluated the hydrologic response to aspen harvesting within a catchment characteristic of Alberta's Boreal Plains. The goal was to test the following hypotheses:

1. Post-harvest reductions in interception and evapotranspiration result in increased groundwater levels and stream flow within this setting,
2. Post-harvest hydrologic impacts are generally negated by the combination of sub-humid climate, rapid aspen regeneration, and large-moisture storage potential of the thick unconsolidated substrate.

Hydrometric data were collected from 2004 to 2010 at a small 13 ha catchment within the Boreal Plains where two aspen stands were clear-cut during the winters of 2007 and 2008. The high degree of heterogeneity, both within and between an adjacent reference catchment, precluded rigorous statistical evaluation of impacts; consequently, numerical simulations were used in conjunction with the observational dataset to assess the effects of harvesting. The combined monitoring-modelling approach permitted direct evaluation of harvesting impacts and their causes through consideration of the integrated hydrologic response of the system. Further simulations were used to evaluate system responses to varying aspen regeneration rates and atmospheric conditions.

4.2. Study Area

The Utikuma Region Study Area (URSA; Lat: 56.07 N, Long: 115.5 W) is located within Alberta's Boreal Plains approximately 350 km northwest of Edmonton (Figure 4-1) and 150 km south of the discontinuous permafrost region (Woo and Winter, 1993). Average regional climate (i.e., 1961-1990) is sub-humid, with annual potential evapotranspiration (517 mm) slightly greater than precipitation (485 mm), and monthly average temperatures ranging from -14.5 °C (January) to 15.6 °C (July; Marshall *et al.*, 1999).

The harvested and reference catchments (Figure 4-1 and Table 4-1) are located central to the URSA where the landscape is characterized by gently rolling topography with low topographic relief of 10 to 20 m. Catchment hillslopes are covered predominately by aspen approaching maturity following regeneration from wildfire in 1962 (Petrone *et al.*, 2015). Flat-lying clay-rich areas contain shallow ~1 m deep ponds (i.e., Pond 40 and Pond 43; Figure 4-1) underlain by 2 to 5 m of organic-rich lake sediments referred to as gyttja. The ponds are surrounded by peatlands that are up to 4 m thick with low-density, stunted black spruce. A further peatland is located between the northern margin of the study catchments at the top of a hillslope that contributes water to both catchments during wetter periods. Peatland hydrological dynamics are affected by seasonal subsurface ice (Petrone *et al.*, 2008; Brown *et al.*, 2010). Ephemeral surface drainage within the peatlands is generally poorly developed and frequently altered by beavers.

The subsurface is composed of deep glacial disintegration moraine deposits that are 40 to 50 m thick (Pawlowicz and Fenton, 2005) underlain by marine shales of the Upper Cretaceous Smoky Group (Vogwill, 1978). The glacial deposits are highly heterogeneous, containing a complex arrangement of finer-grained clay-rich and coarser-grained silt and sand-rich materials along with discontinuous sand lenses. The hydraulic conductivity of the glacial till is generally low, although large variations are observed (i.e., range of 10^{-10} to 10^{-5} m/s; Chapter 2). However, measured glacial till hydraulic conductivity in the northern hillslope of the harvested catchment was generally lower than elsewhere as clay-rich materials were more frequently encountered. Available measurements at more than 10 profiles indicate the hydraulic conductivity of the peat and gyttja declines with depth (Chapter 2), similar to observations from other studies (e.g., Beckwith *et al.*, 2003; Quinton *et al.*, 2008). Hydraulic conductivities range from near-surface values of up to 10^{-3} m/s (peat) and 10^{-6} m/s (gyttja), to less than 10^{-8} m/s near the base (peat and gyttja).

4.3. Field Data

4.3.1. Forest Harvest

Aspen harvesting was completed during two consecutive winters. The northern hummock was clear-cut in March 2007; the southern hummock was clear-cut in February 2008 (Figure 4-1). Harvesting was completed using a feller-buncher and skidder with all trees removed. Slash was piled and burned the following winter other than small piles left for habitat development. Road reclamation was the only further post-harvest treatment. Observations during the harvests indicated the soils were frozen, with more than 0.6 m of snow accumulation.

4.3.2. Hydrometric Measurements

Study area precipitation was estimated using two automated tipping bucket gauges (Figure 4-1) and confirmed with bulk rain gauges distributed across the site. Snow depths were quantified using snow surveys conducted from the winter through spring each year, with the snow water equivalent (SWE) determined from composite samples. Air temperature in the study area was measured at half hour intervals and recorded using automated loggers.

Integrated throughfall measurements were used to estimate rainfall interception for aspen and black spruce stands. Throughfall was collected in 10 m long troughs placed below the shrub understory. Bulk volumes were measured approximately every 2 weeks from spring to mid-September. Following harvesting, throughfall troughs were reestablished at the same height to estimate changes in rainfall interception by recovering shrubs and regenerating aspen.

Pond water levels were monitored using submerged pressure transducers with integrated dataloggers and manual measurements from established staff gauges. Manual data were collected daily to bi-weekly during ice-free periods and monthly during the winter, while automated records were limited to the ice-free season. Stream flows were estimated using flow and water level measurements collected daily to weekly at V-notch weirs located at the outlet of each study catchment.

Manual groundwater levels were measured throughout the study period in monitoring wells and piezometer nests distributed across the study area to a maximum depth of 23 m (Figure 4-1). Measurements were generally collected weekly to biweekly from spring through autumn and monthly during the winter.

4.4. Numerical Model

A 2D model was used to quantify the impact of aspen harvesting. The model domain extended from the centre of the peatland located north of the northern hillslope to the crest of the southern hillslope, with Pond 40 situated near the centre, and was a minimum of 18 m deep (Figures 4-1 and 4-2). The model domain was discretized using uniform 1 m finite-elements horizontally; vertical node spacing varied from 0.05 m near the surface to 0.5 m at depth. Daily boundary conditions along the upper surface of the domain, described below, were implemented using an adaptive time-stepping scheme that allowed for refined temporal discretization using sub-daily time-steps.

The model was developed using HydroGeoSphere (HGS; Aquanty, 2013), a physically-based, fully-integrated, variably-saturated, groundwater-surface water code. The HGS code was previously applied within the reference catchment to assess hydrologic interactions occurring within these pond-peatland-aspen upland ecosystems (Chapter 2) and to evaluate how they may be impacted by the effects of climate change (Chapter 5).

4.4.1. Simulation Scenarios

The base case simulation was driven by measured atmospheric variables (e.g., precipitation and temperature) and performed from November 2003 through March 2011, encompassing an undisturbed period and the initial response to clear-cutting of both hillslopes. Model performance was evaluated at 59 monitoring points along with the computed stage of Pond 40. Calculated statistics used to evaluate model performance included the correlation coefficient (R) and normalized root mean square error (NRMS), given by:

$$R = \frac{\sum_{i=1}^{n}(h_o - h_{om})(h_s - h_{sm})}{\sqrt{\sum_{i=1}^{n}(h_o - h_{om})^2 \sum_{i=1}^{n}(h_s - h_{sm})^2}}$$

$$NRMS = \frac{\sqrt{\frac{1}{n}\sum_{i=1}^{n}(h_s - h_o)^2}}{max(h_o) - min(h_o)} \times 100\%$$

where h is the water level and subscripts o, s, and m indicate observed, simulated, and mean, respectively.

To directly evaluate impacts attributable to aspen harvesting, the base case model was run for both cut and synthetic uncut scenarios, with aspen hillslopes specified to retain boundary inputs associated with mature stands (Tables 4-2 and 4-3) in the uncut case. A further simulation was also performed to evaluate the influence of reduced post-harvest aspen evapotranspiration on model predictions.

Six additional hypothetical simulations (i.e., scenarios S1 to S6) were conducted to investigate the expected post-harvest range in hydrologic response due to atmospheric variability (Table 4-4). For these simulations, precipitation and evapotranspiration boundary conditions were derived from long-term climatic data from Fort McMurray, Alberta (Environment Canada, 2016) scaled to match study area climate normals from 1961 to 1990. Each simulation consisted of a 15-year period, with the first 5 years used to generate wet or dry pre-harvest conditions. Clear-cutting was simulated to occur on both hillslopes in the winter of the fifth year, with the remaining ten years used to evaluate the post-harvest response. Four post-harvest scenarios were considered. These scenarios included (1) the wettest and (2) driest periods on record, and (3) the wettest and (4) driest years on record simulated sequentially. For all scenarios, both cut and uncut simulations were conducted. For scenarios (1) and (2), simulations were performed for both wet and dry pre-harvest conditions; whereas, for scenarios (3) and (4), simulations were performed for either wet (scenario 3) or dry (scenario 4) pre-harvest conditions.

4.4.2. Boundary Conditions

Daily boundary conditions were applied to the model surface. Inputs to the model surface consisted of rain and spring snowmelt (Table 4-3), and varied by surface cover to account for interception that was manually calculated as outlined in Chapter 2. Mean SWE within the snowpack was calculated external to the model by summing precipitation falling during the simulated frozen period and included rainfall. Simulated snowmelt was applied over 1 to 3 weeks, as estimated from snow surveys, air temperature, and measured water levels.

Outputs from the model surface consisted of potential evapotranspiration estimated using the temperature-based Hamon (1963) method and scaled by surface cover (Chapter 2) based on available measurements of both potential and actual evapotranspiration within the study catchments (Petrone *et al.*, 2007; Brown *et al.*, 2010; Brown *et al.*, 2014). Simulated actual evapotranspiration was limited by simulated water availability, declining from the potential rate at field capacity (-33 kPa) to zero at the permanent wilting point (-1500 kPa). The simulated actual evapotranspiration was focused at shallow depths and distributed vertically across rooting zones within aspen hillslopes (3 m; Debyle and Winokur, 1985), peatlands (0.5 m; Lieffers and Rothwell, 1987), and the pond (0.1 m) using a cubic decay function available within HGS (Aquanty, 2013). Following aspen harvesting, the applied daily surface fluxes were modified (Tables 4-2 and 4-3) to account for reduced interception and evapotranspiration based on data collected within the study area and other locations within the Boreal Plains (Carrera-Hernandez *et al.*, 2011) and Boreal Forest (Murray and Buttle, 2003).

Stream flow from the pond was simulated using two different boundary conditions that depended upon the simulation scenario. For simulations with known stream flow (i.e., 2003 to 2011), measured outflow was converted to an areal flux using the estimated pond area and added to the specified pond evapotranspiration. For hypothetical scenarios, stream flow from the pond was simulated using a pumping boundary condition during the ice-free season, with water specified to be removed whenever the pond stage exceeded the elevation of the sill of the outflow channel.

Based on field measurements, and as in previous simulations (Chapters 2 and 5), a uniform constant hydraulic head was specified along the model base and no-flow boundaries were specified along the inferred divides at the lateral edges of the domain.

4.4.3. Material Properties

The model was divided into three hydrogeologic units including peat, gyttja, and glacial till. Parameter zones within each unit were defined based on observed depth trends in hydraulic conductivity (Figure 4-2 and Table 4-5) and previous simulation results within the reference catchment (Chapter 2). Soil-water retention and hydraulic conductivity curves were derived from the literature for the peat and gyttja (Silins and Rothwell, 1998; Price *et al.*, 2010), and glacial till using the properties of a clay loam (Carsel and Parrish, 1988).

4.5. Results

4.5.1. Precipitation and Interception

The study area was relatively dry prior to harvesting in the winter of 2007. From 1999 to 2006, annual precipitation (Table 4-6) was less than the long-term average of 485 mm in all years but 2005. The driest years occurred in 2001 and 2002, with both years receiving less than 60% of average precipitation. However, in 2007 and 2008, when the respective northern and southern hillslopes were harvested, wetter conditions occurred with annual precipitation exceeding the long-term average in both years. The wetter conditions did not persist beyond the harvest years, as annual precipitation returned to less than average in both of the following two years. In all years, rainfall patterns were relatively consistent, with daily rainfall exceeding 10 mm on less than 10% of days with rain.

Rainfall interception prior to harvesting was similar for both aspen and spruce stands in the study area, totaling approximately 25% on an annual basis (Chapter 2). Following aspen harvesting, estimated annual rainfall interception was reduced to about 14% and 21% in the respective first and second years following harvest. By the third post-harvest year, negligible differences in interception were observed between cut and uncut areas; rapid aspen regeneration occurred with dense stands reaching heights of 1 to 2 m height (Petrone *et al.*, 2015).

4.5.2. Hillslope Groundwater Levels

Prior to aspen harvesting, seasonal groundwater level trends were generally similar across both study catchments. Each spring groundwater levels increased to a peak in early to mid-May following snowmelt. During the growing season, peatland groundwater levels responded dynamically to rain events, but displayed an overall decreasing trend towards the fall. Within the hillslopes, groundwater levels generally exhibited a continuous decline during the growing season. At coarser-grained silt and sand-rich locations, hillslope groundwater levels showed little response to rain events and consistently dropped below peatland groundwater levels towards the end of the growing season (Figure 4-3). At fine-grained clay-rich locations, greater growing season response to rain events was apparent, as evidenced by short-term groundwater level increases. Moreover, fine-grained hillslope groundwater levels displayed an overall increasing trend prior to harvesting of the northern hillslope, and generally remained above pond and peatland water levels throughout the year (Figure 4-3).

4.5.2.1. Harvest Impact at Finer-Grained Locations

Post-harvest response within the finer-grained northern hillslope of the harvested catchment (i.e., wells HA-3 and HA-4) was compared to well RA-3 within the reference catchment (Figures 4-1 and 4-3CD). Following aspen harvesting of the northern slope during the winter of 2007, spring groundwater levels in the hillslope peaked above what was observed in the preceding years. This year received above average snowpack and similarly increased groundwater level peaks were observed at unharvested locations throughout both study catchments (Figure 4-3). During the growing season, groundwater levels in the harvested hillslope declined in a manner similar to preceding years, although the response to rain events may have increased. Groundwater levels in the hillslope also remained elevated in the fall relative to the preceding year at some locations; whereas, no change was apparent in the reference catchment. From 2008 to 2010, similar seasonal groundwater level trends were apparent in the hillslope relative to the reference catchment, although the magnitude of the spring groundwater level rise was larger in 2008 and 2009 in the harvested catchment. During the growing season, groundwater levels declined at similar rates in both catchments, with little response to rain events.

4.5.2.2. Harvest Impact at Coarser-Grained Locations

Post-harvest response within the coarser-grained southern hillslope was negligible based on observations from well HA-5 within the harvested catchment relative to wells RA-1 and RA-2 within the reference catchment (Figures 4-1 and 4-3AB). Following aspen clear-cutting of the southern hillslope in winter 2008, the magnitude of the spring peak in hillslope groundwater levels and rate of growing season decline was similar to that observed within the reference catchment. The rate of groundwater level decline did decrease in early July 2008 following a series of rain events; however, a similar response was also observed within the reference catchment at RA-1. In the second year following harvesting of the southern hillslope, groundwater levels within both catchments responded similarly to snowmelt and declined at a comparable rate throughout the growing season.

4.5.3. Pond and Peatlands

From 2000 to 2004, estimated stream flow from both catchments was negligible as pond and peatland water levels were low following a series of dry years (Table 4-6). In 2005 and 2006, annual pre-harvest stream flows from the harvested catchment were 19 mm and 9 mm greater, respectively, than stream flows from the reference catchment. Annual stream flows in 2007 and 2008 from the harvested catchment were 4 mm and 18 mm greater than the reference catchment following clear-cutting of the respective northern and southern hillslopes, with stream flows from both years generally within the range observed pre-harvest (Table 4-6). Annual stream flow from the harvested catchment did increase in 2009, with an estimated discharge about 37 mm greater than the reference catchment. However, peatland groundwater levels were apparently unaffected, as similar trends were exhibited within harvested and reference catchments (Figures 4-1 and 4-3EF).

4.5.4. Numerical Simulations

4.5.4.1. Evaluation of Model Performance

Computed statistics for 59 monitoring points and Pond 40 stage indicate the model provided a reasonable representation of the overall hydrologic system (i.e., R = 0.95; NRMS = 4.7%; Table 4-7). Water levels were particularly well-represented within Pond 40 (i.e., R = 0.95; NRMS = 6.3%) and the peatlands (i.e., R = 0.97; NRMS = 3.6%). Greater discrepancy between observed and simulated groundwater levels was present within the glacial till (i.e., R = 0.95; NRMS = 7.2%) where considerable heterogeneity is present. Temporally, the model captured the observed seasonality, with computed water levels generally well replicated before and after harvesting occurred (Figure 4-4).

4.5.4.2. Simulated Response to Harvesting

Comparison of simulation results between harvested and unharvested cases indicates that aspen harvesting led to increased hillslope water levels ranging from 3 m (northern hillslope, Figure 4-5A) to 1.5 m (southern hillslope, Figure 4-5B). The largest impact occurred during the growing season, where the lower evapotranspiration of the regenerating aspen (Figure 4-6) resulted in a slower decline in hillslope groundwater levels following snowmelt and increased response to rain events. Greater pre-melt groundwater levels also led to an increase in snowmelt response in spring 2010 in both hillslopes, although predicted differences between scenarios were generally diminished by the end of the year following a relatively dry year.

Outside the cut blocks, predicted pond and peatland water levels increased by 0.05 to 0.3 m by the end of 2009 relative to the unharvested scenario (Figure 4-5C), with the greatest impact predicted late in the growing season. Differences in predicted pond water levels between harvested and unharvested cases suggests annual stream flows were increased by up to 6 mm due to harvesting.

4.5.4.3. Influence of Regenerating Aspen Evapotranspiration

Simulation results were sensitive to reduced regenerating aspen evapotranspiration, with water levels in both hillslopes predicted to rise 1 to 3 m relative to the base case (Figure 4-7AB). At some locations within the northern hillslope (e.g., HA-2; Figure 4-1), observed groundwater levels were better replicated in this scenario relative to the base case (Figure 4-7A). Following harvesting, hillslope water levels were predicted to remain elevated following snowmelt, with decreased available storage and more frequent growing season water level fluctuations following rain events. Concurrent increases in pond and peatland water levels of up to 0.6 m were also predicted for this scenario (Figure 4-7C), with the increase in pond level equivalent to an increase in stream flow of almost 20 mm in 2008.

4.5.4.4. Influence of Atmospheric Variability

Simulation results indicate that aspen harvesting may lead to increased hillslope groundwater levels ranging from less than 0.5 m to 3 m (Figure 4-8AB). Peak increases were generally predicted during the growing season due to reduced aspen evapotranspiration following harvesting; however, similarly large differences between scenarios were also predicted in the spring when elevated groundwater levels persisted in the cut scenario prior to snowmelt.

For scenarios S1 to S4 with wet pre-harvest conditions and/or post-harvest conditions, peak hillslope groundwater level increases were predicted to occur in the initial

1 to 3 years after harvesting, as infiltrating water rapidly reached the water table. During this period, the water table at the toe of the hillslopes fluctuated within 0.5 m to 1 m of ground surface. However, the duration of harvest-induced increased groundwater levels was predicted to be diminished by the seventh year when simulated regenerating aspen evapotranspiration approached mature rates.

For scenarios S5 and S6 with both dry pre- and post-harvest conditions, smaller increases in hillslope groundwater levels of less than 0.2 m were predicted immediately after harvesting relative to wetter cases, as differences in aspen evapotranspiration between cut and uncut scenarios were generally negated by limited available water. For the driest post-harvest case S6, where dry conditions were simulated to persist, hillslope groundwater levels increased by less than 0.5 m throughout the simulation period as the water table remained near the base of the simulated aspen roots depth at 2 to 3 m depth. Conversely, for the dry case S5 increased groundwater levels comparable to the wetter cases occurred following wetter conditions; the increased water levels persisted to the end of the simulated 10-year period. However, as opposed to the wetter cases, hillslope water tables were predicted to remain at 1 m to 3 m depth.

Outside of the simulated cut blocks, aspen harvesting was predicted to have limited influence on peatland water levels (not shown) and stream flows from the pond (Figure 4-8C). Within the peatlands, groundwater levels were predicted to increase by less than 0.3 m in all scenarios, with similar temporal trends as predicted within the adjacent hillslopes. Similarly, stream flows were predicted to be increased by less than 10 mm in most years, with differences between cut and uncut cases generally predicted to become negligible after about 7 years.

4.6. Discussion

4.6.1. Hydrologic Impact of Boreal Plains Aspen Harvesting

Study results indicate that aspen harvesting has limited hydrological impact within the study catchment. The combination of sub-humid climate with low frequency of large storms (Buttle *et al.*, 2009) and large storage capacity of the glacial materials (Devito *et al.*, 2005b) provides substantial system buffering capacity (Redding and Devito, 2008), thereby limiting hydrologic impacts (Carrera-Hernandez *et al.*, 2011). Harvesting is predicted to result in increases of hillslope water levels of up to 3 m (Figures 4-5AB and 4-8AB); however, in general, peak increases do not persist beyond the initial post-harvest years. Rapid recovery of aspen evapotranspiration (Petrone *et al.*, 2015), although spatially variable (Figure 4-7A), is predicted to prevent water-logging of near-surface soils with low potential for runoff generation (Price *et al.*, 2005; Buttle *et al.*, 2009). Where harvesting does result in increased hillslope groundwater levels, impacts may be difficult to identify (e.g., Figure 4-3A) as observed and predicted post-harvest responses are similar in magnitude to observed fluctuations due to variability in atmospheric conditions (Figures 4-3AC, 4-5AB, and 4-8AB) with comparable seasonal trends (i.e., peak following snowmelt, general decline during growing season) for both cut and uncut locations.

Study results further indicate that harvesting disturbances are primarily restricted to the glacial substrate within the cut blocks. The relatively low hydraulic conductivity and generally deep water table within the glacial materials limits groundwater exchange between the hillslopes and adjacent peatlands (Chapter 2), resulting in modest impacts with increases of less than 0.3 m in peatland groundwater levels (Figure 4-5C) and consequently stream flows (Devito *et al.*, 2005b; Prepas *et al.*, 2006) even in the wettest of scenarios (Figure 4-8C). However, hydrologic impacts from aspen harvesting are not limited to the removal of trees, as greater disturbance may occur from the placement of roads and skid trails on the flat-lying peatlands that exert significant control on water retention within the landscape (Chapter 2). The peatlands would be susceptible to harvesting effects during most years (Buttle *et al.*, 2009), thus care should be taken to minimize their disturbance when planning and executing aspen management plans.

Although limited hydrologic impact from aspen harvesting was generally found in this study, the predicted hydrologic response was sensitive to the assumed evapotranspiration rate of the regenerating aspen, with stream flows predicted to increase by almost 20 mm in some years. Thus, timber harvesting practices that encourage aspen regeneration and minimize soil disturbance, such as clear cutting and winter harvesting (Frey *et al.*, 2003; Berger *et al.*, 2004) should be preferred if minimizing disturbance is an important objective. Furthermore, these results suggest that harvesting of other tree species that grow more slowly may lead to more pronounced and longer-lasting impacts.

4.6.2. Comparison to Other Locations

Results of this study are consistent with other aspen harvesting studies conducted within the sub-humid Boreal Plains region. Devito *et al.* (2005b) found there was low potential for harvesting to impact stream flows at a catchment in north-central Alberta, with no difference in phosphorous (Macrae *et al.*, 2005) or nitrogen (Macrae *et al.*, 2006) concentrations between forested and harvested areas. Similarly, Whitson *et al.* (2005) found little difference in soil moisture between harvested and forested sites 3 years after harvesting mixed aspen and white spruce stands. In a numerical modelling study, Carrera-Hernandez *et al.* (2011) predicted similar water table increases of 1 to 3.5 m following hypothetical aspen harvesting near Fort McMurray, Alberta, with the largest impacts following wet climate cycles.

At locations within the United States, greater impacts following aspen harvesting have been observed. Verry (1986) stated that aspen clearcutting will increase stream flow by 90 mm over the cut area. Similarly, Debyle (1976) and Debyle and Winokur (1985) noted stream flow increases of 100 to 150 mm following clearcutting based on studies across the United States, with aspen harvesting suggested as a method for increasing water yield for "economically higher purposes". In these studies, increases in stream flows were found to become negligible 7 to 15 years following harvesting. Comparable studies within the boreal forests of Scandinavia were not found in a literature search; however, in a review paper by Worrell (1995) similar regenerative growth rates for European aspen in Norway were noted, suggesting that the range in outcomes from harvesting may be comparable to those reported in the United States.

While outcomes from studies conducted outside the Boreal Plains appear contradictory to those conducted within, the results obtained are not directly comparable due to differences in the overall hydrologic system (e.g., climate and geology). For example, in Minnesota the humid climate (i.e., precipitation of almost 800 mm and potential evapotranspiration of about 550 mm; Nichols and Verry, 2001) results in more pronounced harvesting effects as water-limiting conditions are infrequent, in direct contrast to the sub-humid climate of the Boreal Plains where large water-deficits are prevalent (Devito *et al.*, 2005b). In settings such as the Boreal Plains, the importance of the integrated response of the overall hydrologic system (i.e., climate, geology, and vegetation) is magnified and should be considered when quantifying harvesting impacts and predicting the potential for negative impacts (Devito *et al.*, 2005a). However, study results from other climatic settings could prove increasingly informative for future aspen forest management in the Boreal Plains depending on the future manifestation of climate change.

4.6.3. Climate Change

Annual temperatures across the Boreal Plains are projected to increase by 2 to 5°C by the 2050s (Lemmen *et al.*, 2008), while over the same period precipitation is projected to change by -3% to +9% (Mbogga *et al.*, 2010; Wang *et al.*, 2014). As temperatures rise, greater evaporative water losses are likely to increase the disparity between annual precipitation and evapotranspiration, leading to increased frequency of water-stressed conditions (Chapter 5). Projected increases in fire frequency (Flannigan *et al.*, 2013) and northward migration of pests and diseases (Lemmen *et al.*, 2008) will further stress aspen resources, likely leading to declines in stand productivity (Worrall *et al.*, 2013). Thus, post-harvest aspen regeneration may be impeded, with the potential for greater hydrologic impact than predicted here, particularly if changing climate also includes an increase in the frequency or intensity of large storms (Sillmann *et al.*, 2013). However, widespread aspen decline may lag changes in climate (Schneider *et al.*, 2015) and could be mitigated through silvicultural practices (e.g., periodic thinning, removal of diseased trees; Cerezke, 2009). Effective forest management in the future may increasingly require the use of combined monitoring-modelling studies that integrate the overall hydrologic interactions occurring among ecosystem components and permit evaluation of system responses over a wider range of atmospheric conditions than would be possible strictly through observational approaches.

4.7. Summary and Conclusions

Changes to groundwater levels and stream flows resulting from aspen harvesting in the Boreal Plains were evaluated in a small catchment situated within glacial moraine deposits. Study results indicate harvesting had limited impact on the hydrologic system. Despite an estimated increase in hillslope groundwater levels of up to 3 m, pond and peatland water levels increased by less than 0.3 m and were accompanied by increased stream flows of less than 10 mm/yr. Sensitivity simulations indicate that the potential for substantial increases in stream flows and generation of soil water-logging conditions from aspen harvesting within these Boreal Plains' systems is low due to the combination of sub-humid climate, deep glacial substrate with large soil storage capacity, and rapid aspen regeneration. However, predicted increases in stream flows and groundwater levels were sensitive to regenerating aspen evapotranspiration, which can be enhanced by appropriate harvesting techniques but may be reduced by climate change.

These results highlight the need to consider the integrated response of hydrologic systems when evaluating impacts from disturbance and making comparisons to other settings. Furthermore, the use of combined monitoring-modelling approach of this study was an effective method for evaluating system responses over a wider range of atmospheric conditions than would be possible through a strictly observational approach and should be considered as a method to enhance effective forest management.

Table 4-1. Harvested and reference catchment summary.

Catchment Component	Catchment	
	Harvested	Reference
Area (ha)	13	20
Surface Cover (% of Area)		
Aspen	45	64
Peat	49	30
Pond	6	6

Table 4-2. Summary of parameters to quantify surface boundary conditions for mature and regenerating aspen.

Surface Cover	Aspen Regeneration Year	Rooting Zone PET_h/PET_u [a]	Rain Interception (mm)		Sublimation (% of Snowfall)
			Spring	Summer	
Pond	-	-	0.0	0.0	30
Peatlands	-	-	0.5	2.0	25
Aspen	Mature	1.00	0.5	2.0	25
	0	0.70	0.5	0.8	10
	1	0.74	0.5	1.5	20
	2	0.83	0.5	2.0	25
	3	0.88	0.5	2.0	25
	4	0.92	0.5	2.0	25
	5	0.96	0.5	2.0	25
	6	1.00	0.5	2.0	25

Notes:

[a] Potential evapotranspiration (PET) subscripts h and u denote harvested and unharvested, respectively.

Table 4-3. Summary of annual boundary fluxes applied in the base case model.

Year	Pond (mm) [e]			Peat (mm)			Aspen (mm)				
									PET [b,c]		
	Rain [a]	Snowmelt [a]	PET [b,c]	Rain [a]	Snowmelt [a]	PET [b,c]	Rain [a]	Snowmelt [a]	Mature [d]	Northern Hillslope [d]	Southern Hillslope [d]
2004	300	77	391	297	81	277	297	81	469	469	469
2005	338	154	421	319	173	329	323	168	532	532	532
2006	298	134	411	356	76	318	362	70	557	557	557
2007	318	212	463	312	218	359	315	214	578	375	578
2008	354	151	450	324	181	333	327	178	574	435	377
2009	241	150	439	209	182	310	212	179	528	456	400
2010	227	55	481	220	62	339	223	59	539	486	464

Notes:

[a] Proportions of rain and snow varied by surface cover due to different lengths of frozen seasons. Total precipitation (i.e., sum of rain and snow) was assumed to be uniform.

[b] Potential evapotranspiration (PET) applied to each surface cover. Simulated actual ET was restricted by the available water.

[c] PET includes rainfall interception and snow sublimation that were removed externally from the model.

[d] Harvested northern hillslope and southern hillslope. Mature PET used for simulations that assumed no aspen harvesting occurred.

[e] Stream flow from the pond (Table 4-6) was simulated as an areal flux from the pond and added to the specified PET.

Table 4-4. Summary of pre- and post-harvest precipitation for hypothetical sensitivity scenarios.

Pre- Or Post-Harvest Period	Case [a]	Scenarios	Years	Snow (mm)			Rain (mm)			Precipitation (mm)		
				Min	Avg	Max	Min	Avg	Max	Min	Avg	Max
Pre	Dry	3, 5, And 6	1945 - 1949	52	70	98	216	283	343	278	352	395
Pre	Wet	1, 2, And 4	1972 - 1976	90	159	282	301	435	559	497	594	715
Post	Dry*	6	Variable	55	83	130	194	233	285	249	316	356
Post	Dry	4 And 5	1942 - 1951	52	100	270	199	268	372	278	367	468
Post	Wet	2 And 3	1967 - 1976	90	152	282	235	393	559	349	546	715
Post	Wet*	1	Variable	90	157	201	399	473	559	597	631	715

Notes:

[a] Wet and dry cases incorporated years from the respective wettest and driest five-year (i.e., pre-harvest) or ten-year (i.e., post-harvest) period on record arranged by temporal sequence. Wet* case incorporated the ten wettest years arranged from highest to lowest total precipitation. Dry* case incorporated the ten driest years arranged from lowest to highest total precipitation.

Table 4-5. Summary of parameters to quantify surface boundary conditions for mature and regenerating aspen.

Material	Depth Range (m)	Hydraulic Conductivity [a] Horizontal (m/s)	Hydraulic Conductivity [a] Anisotropy (K_H:K_V) [b]	Porosity (-)	Specific Storage (m^{-1})
Peat	0.0 - 0.1	$5x10^{-4}$ - $3x10^{-3}$	10	0.90	$1x10^{-4}$
	0.1 - 0.3	$1x10^{-5}$ - $3x10^{-4}$	10	0.82	$5x10^{-5}$
	0.3 - 0.5	$1x10^{-6}$ - $8x10^{-5}$	10	0.72	$8x10^{-6}$
	0.5 - 1.0	$1x10^{-7}$ - $4x10^{-5}$	10	0.60	$2x10^{-6}$
	1.0 - 1.5	$3x10^{-8}$ - $2x10^{-6}$	10	0.50	$5x10^{-7}$
	1.5 - 2.0	$1x10^{-8}$ - $3x10^{-8}$	10	0.45	$2x10^{-7}$
	2.0 - Base	$1x10^{-8}$	10	0.40	$1x10^{-7}$
Gyttja	0.0 - 1.0	$1x10^{-6}$	10	0.45	$3x10^{-6}$
	1.0 - 2.0	$3x10^{-8}$	10	0.30	$4x10^{-7}$
	> 2.0	$5x10^{-9}$	10	0.22	$1x10^{-7}$
Glacial Till	Upper	$1x10^{-5}$	100	0.20	$1x10^{-4}$
	Middle	$5x10^{-7}$	100	0.20	$1x10^{-4}$
	Lower	$1x10^{-8}$	100	0.20	$1x10^{-4}$

Notes:

[a] Peat hydraulic conductivity (K) varied due to variations in peat accumulation and decomposition. The lowest values were assigned to peat at the toe of the northern hillslope.

[b] Subscripts denote horizontal (H) and vertical (V) hydraulic conductivity. Based on Beckwith *et al*. (2003) and Nagare *et al*. (2013) for organic materials, and Hendry (1988) for glacial till.

[c] Specified specific storage values for peat and gyttja are low; however, the influence on simulation results is negligible due to small fluctuations in water levels (i.e., less than 1 m).

Table 4-6. Summary of measured annual precipitation and stream flows over the study period.

Year	Precipitation (mm)	Stream Flow (mm)		
		Harvested	Reference	Difference
1999	371	-	-	-
2000	442	-	0	0
2001	257	-	0	0
2002	290	-	0	0
2003	443	0	0	0
2004	378	0	0	0
2005	491	53	34	19
2006	432	16	7	9
2007	530	49	45	4 [a]
2008	504	33	15	18 [b]
2009	391	52	15	37
2010	282	0	0	-

Notes:

[a] Northern hillslope harvested winter 2007.

[b] Southern hillslope harvested winter 2008.

Table 4-7. Model calibration statistics (hydraulic head).

Material	n (-)	R (-)	r_m (m)	RMS (m)	NRMS (%)
Pond	200	0.95	0.01	0.04	6.3
Peat	2,840	0.97	0.15	0.29	3.6
Glacial Till	1,640	0.95	-0.18	0.81	7.2
All	4,680	0.95	0.03	0.53	4.7

Notes:

[a] n = number of measurements, R = correlation coefficient, r_m = residual mean, RMS = root mean square error, NRMS = normalized (i.e., by elevation) RMS.

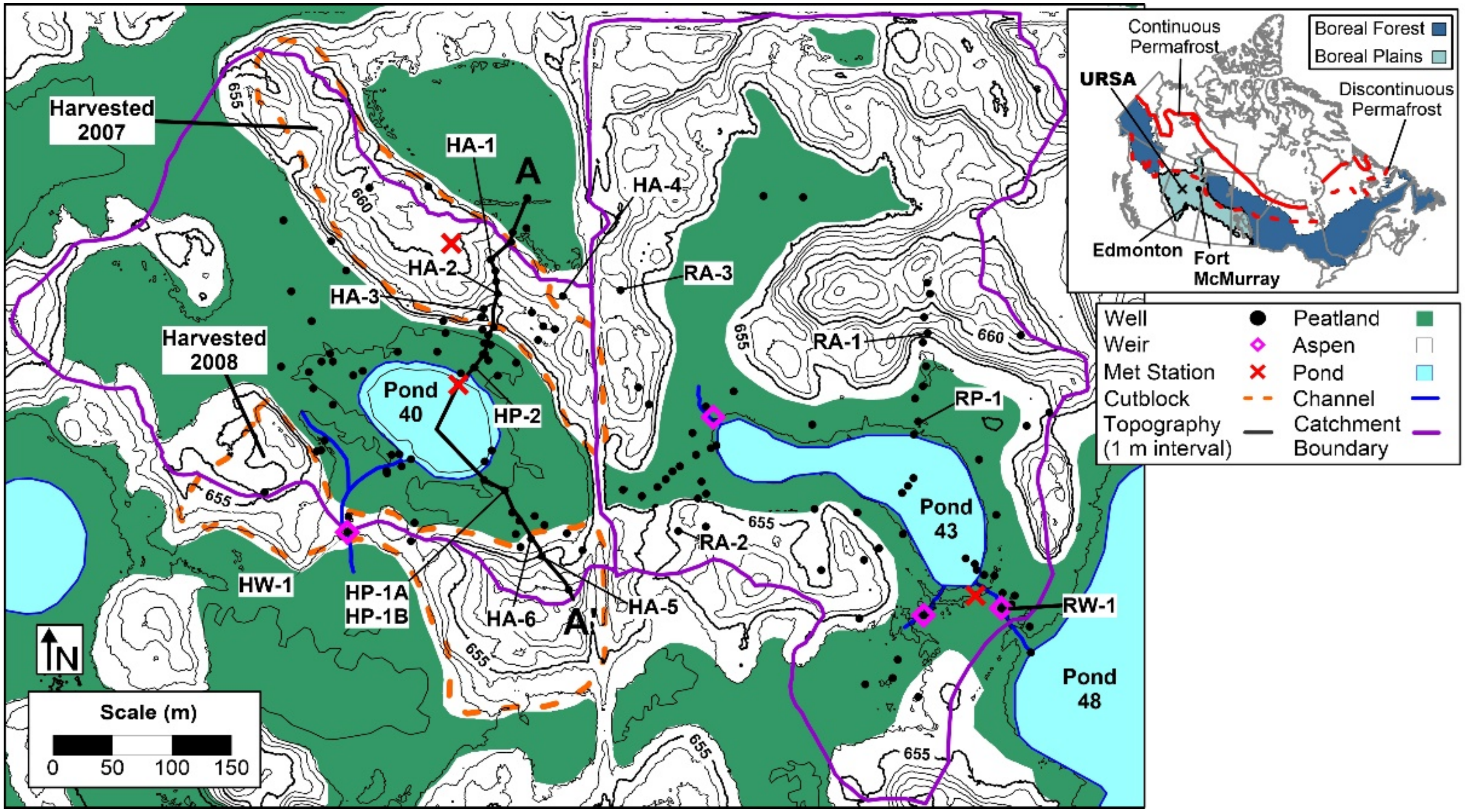

Figure 4-1. Study watershed delineation with topography, gross vegetation classifications, instrumentation locations, and the location of the numerical model transect (A - A'). The Pond 43 catchment serves as the uncut reference; sections of the Pond 40 and surrounding catchments were harvested in Winter 2007 and 2008, as shown. Data from labelled instrumentation (H = harvest; R = reference; P = peatland; A = aspen; W = weir) locations illustrated in the analyses.

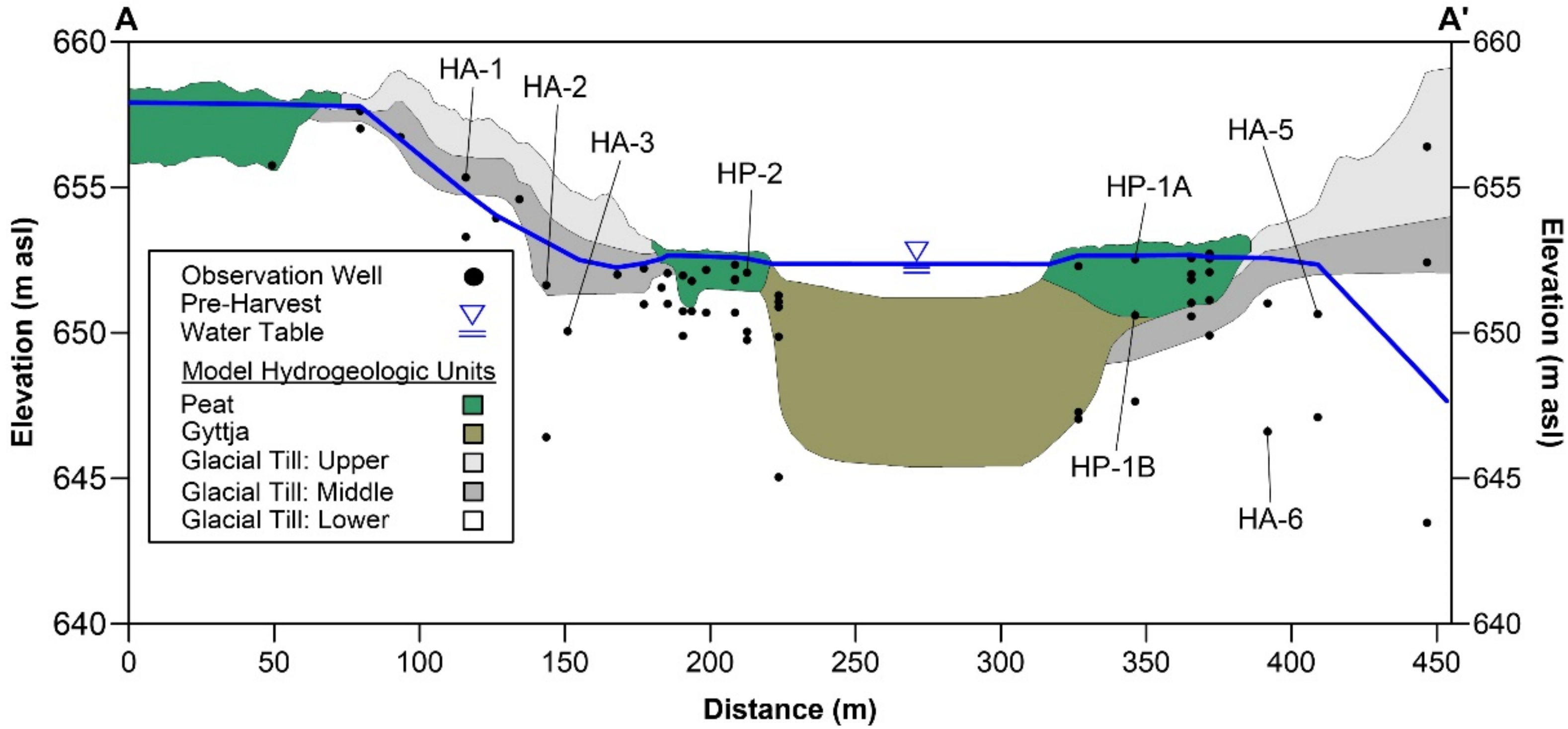

Figure 4-2. Interpreted cross-section specified within the 2D model domain with hydrogeologic units, observation wells (mid-screen elevation shown), and pre-harvest water table. The base of the numerical model extends to 633 m asl.

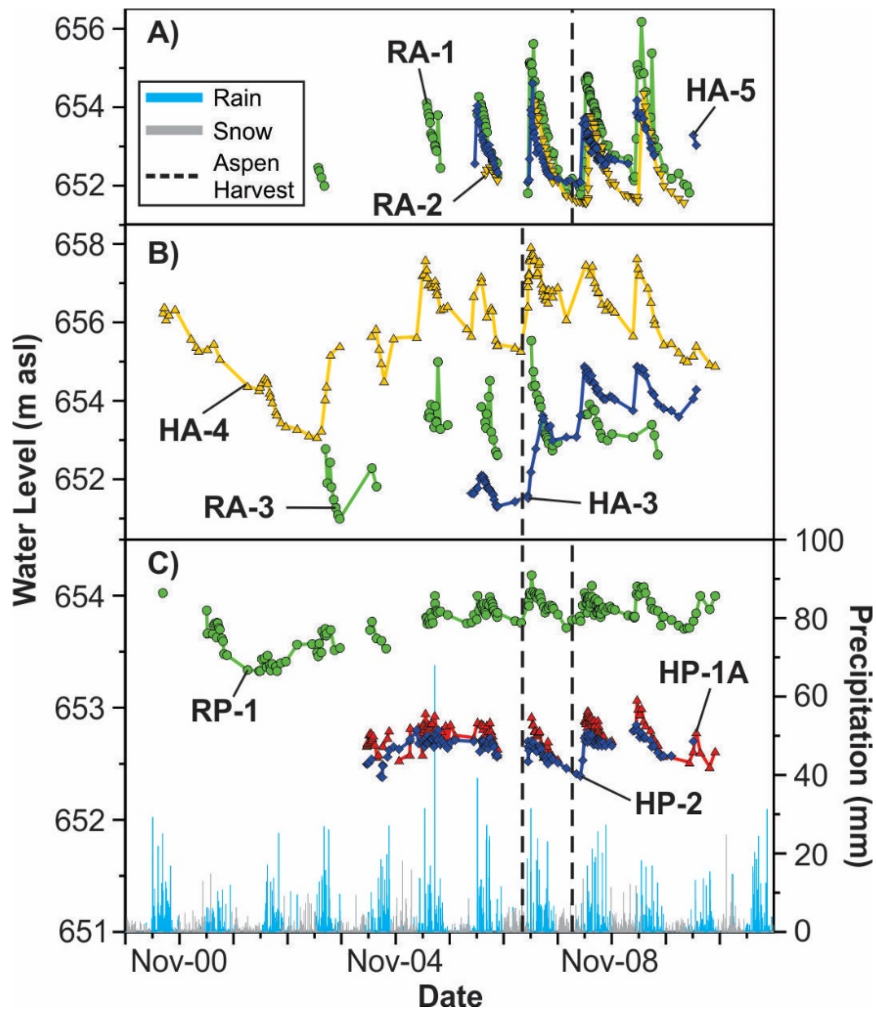

Figure 4-3. Comparison of groundwater levels within A) silt and sand-rich hillslopes, B) clay-rich hillslopes, and C) peatlands in the harvested and reference catchments along with daily rain and snowfall. Monitoring well locations are shown in Figure 4-1.

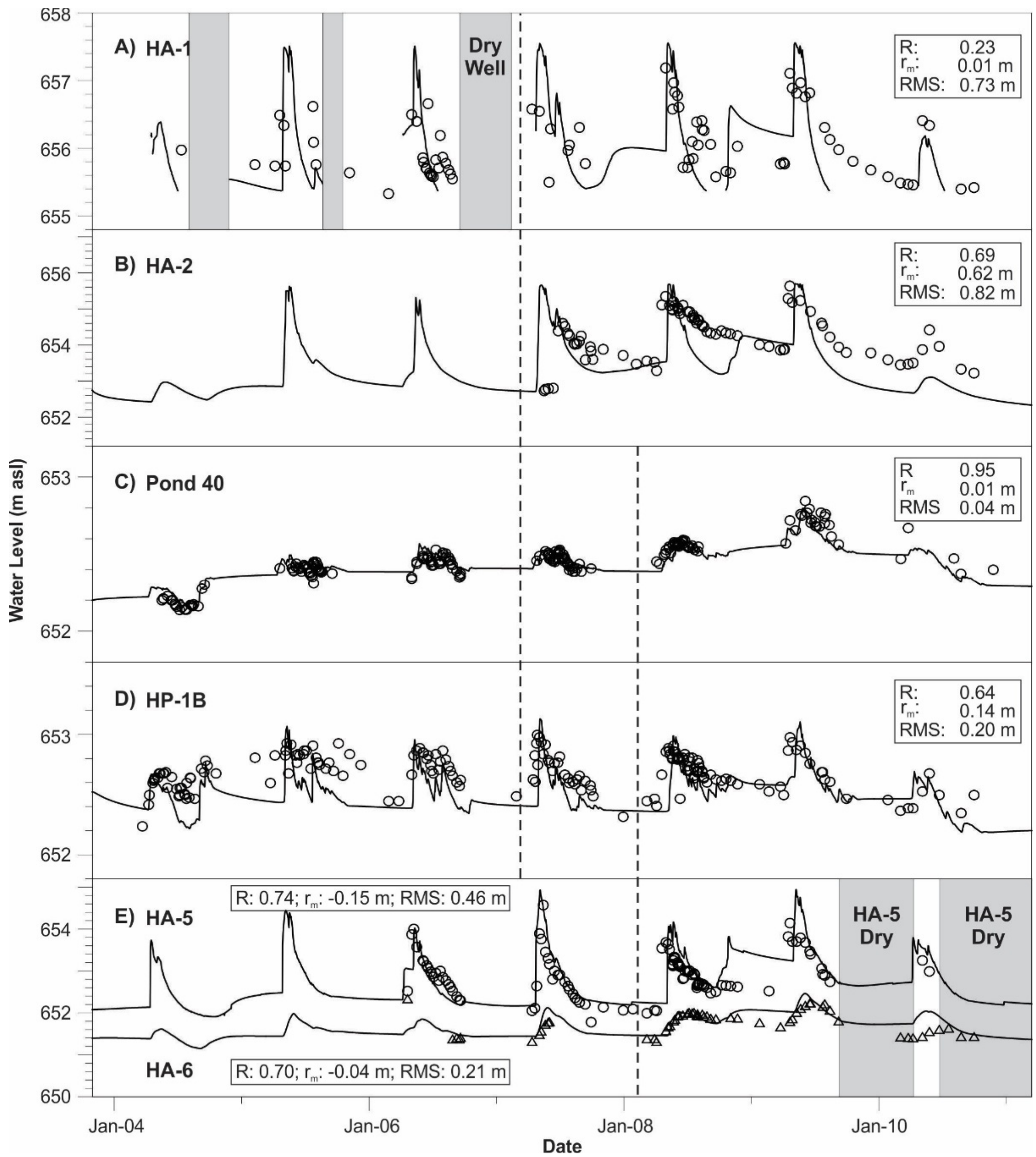

Figure 4-4. Observed (symbols) and simulated (lines) water levels at monitoring locations within the harvested catchment along with calibration statistics (R = correlation coefficient, r_m = residual mean, RMS = root mean square error). Dry wells are indicated by shaded periods (observed) and line breaks (simulated). Monitoring well locations are shown in Figure 4-1. Vertical dashed lines indicate the timing of aspen harvesting.

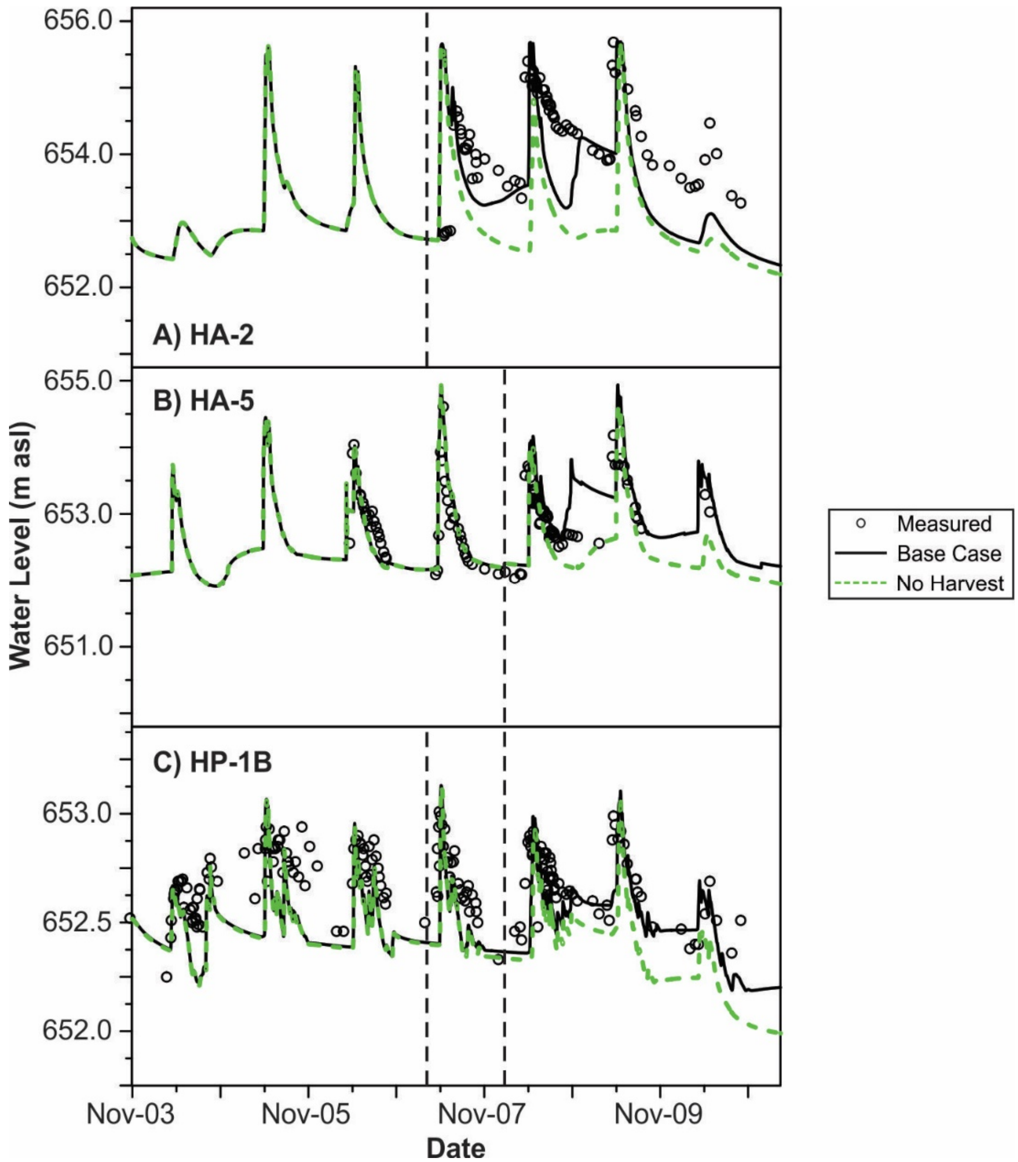

Figure 4-5. Predicted influence of harvesting on groundwater levels at A) the harvested northern hillslope, B) the harvested southern hillslope, and C) the peatlands. Monitoring well locations are shown in Figure 4-1. Vertical dashed lines indicate the timing of aspen harvesting which was simulated in the base case.

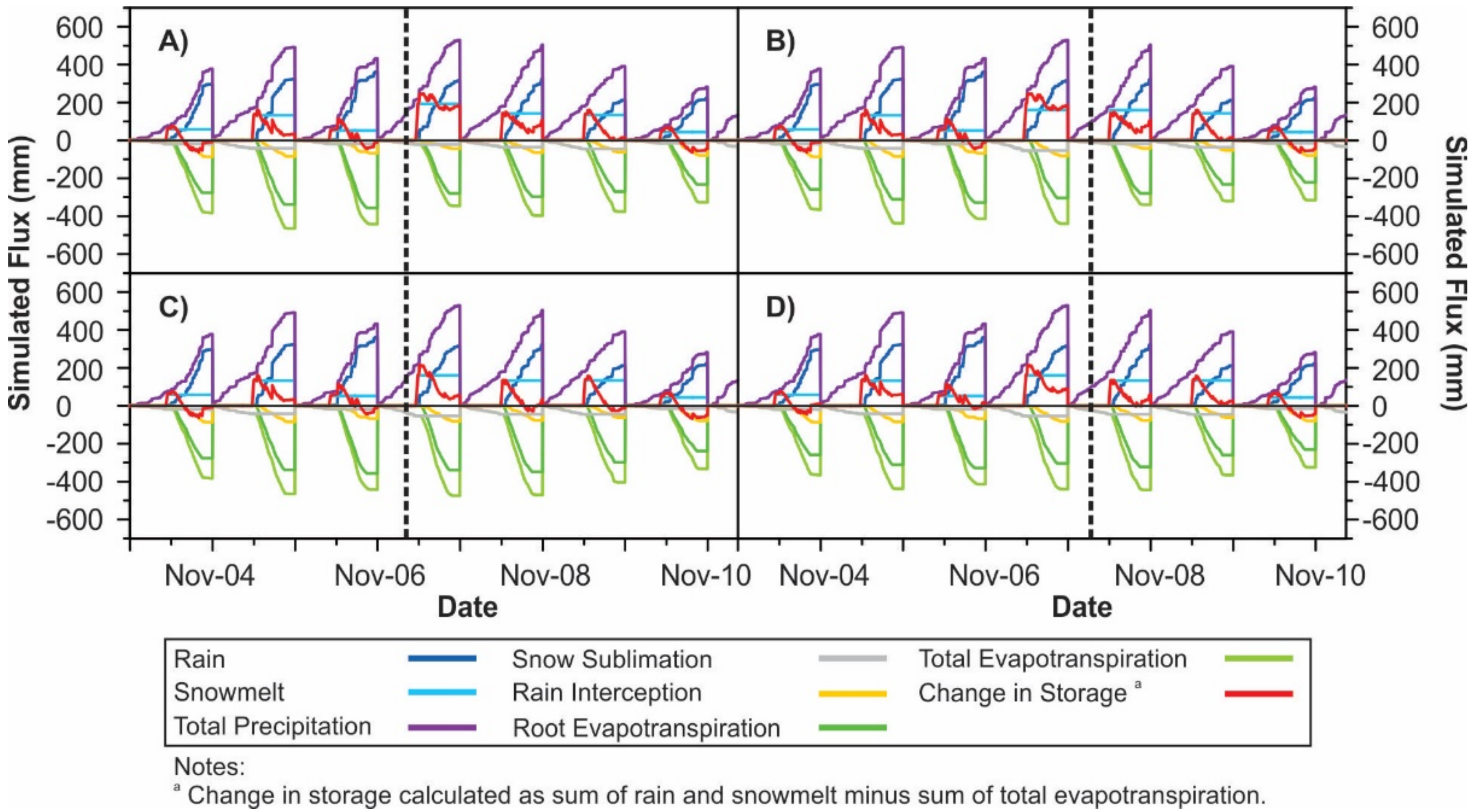

Figure 4-6. Comparison of cumulative annual simulated boundary fluxes for the base case harvested (A, B) and unharvested (C, D) cases in the northern hillslope (A, C) and southern hillslope (B, D). Vertical dashed lines indicate the timing of aspen harvesting.

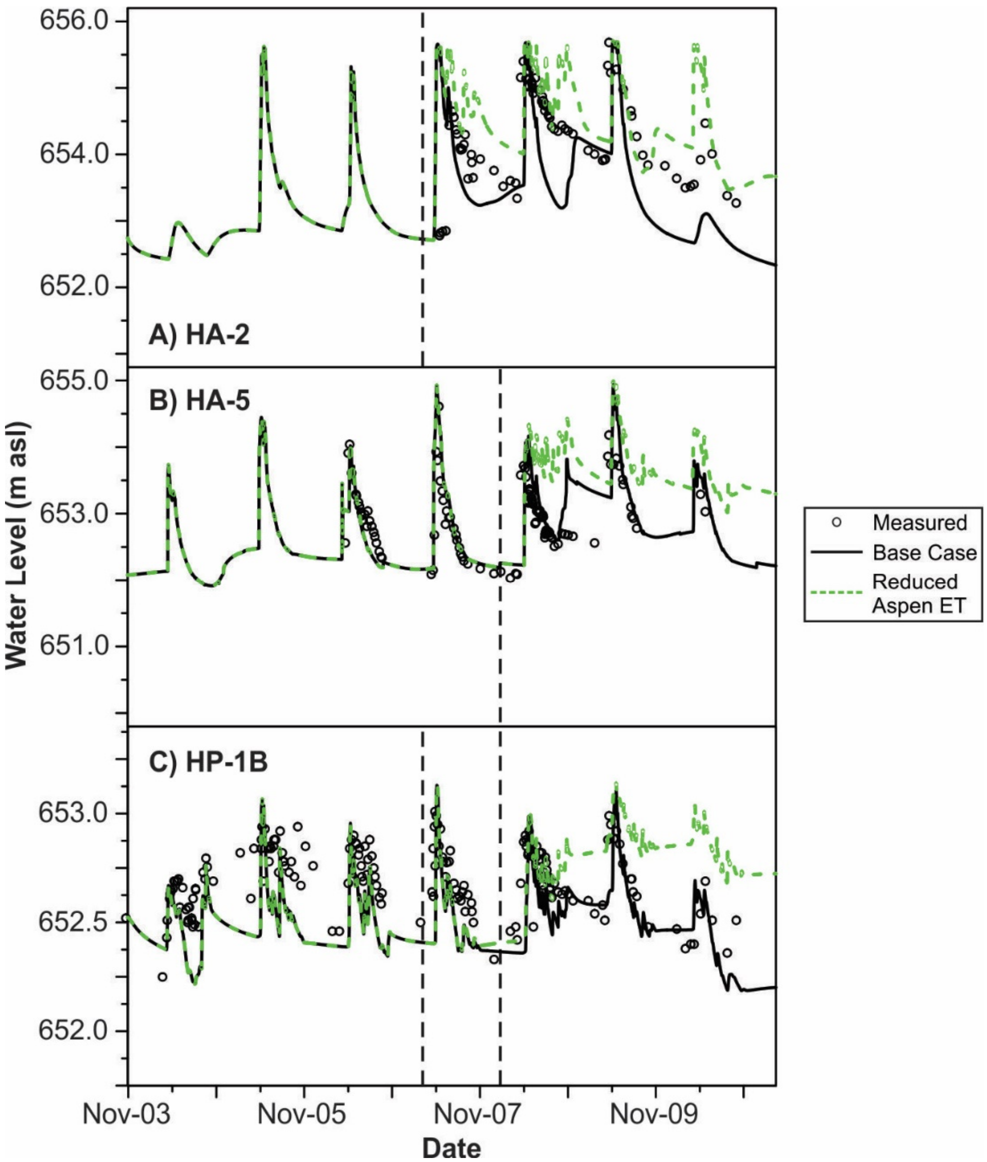

Figure 4-7. Predicted influence of reduced regenerating aspen evapotranspiration on groundwater levels at A) the harvested northern hillslope, B) the harvested southern hillslope, and C) the peatlands. Monitoring well locations are shown in Figure 4-1. Vertical dashed lines indicate the timing of aspen harvesting which was simulated in the base case.

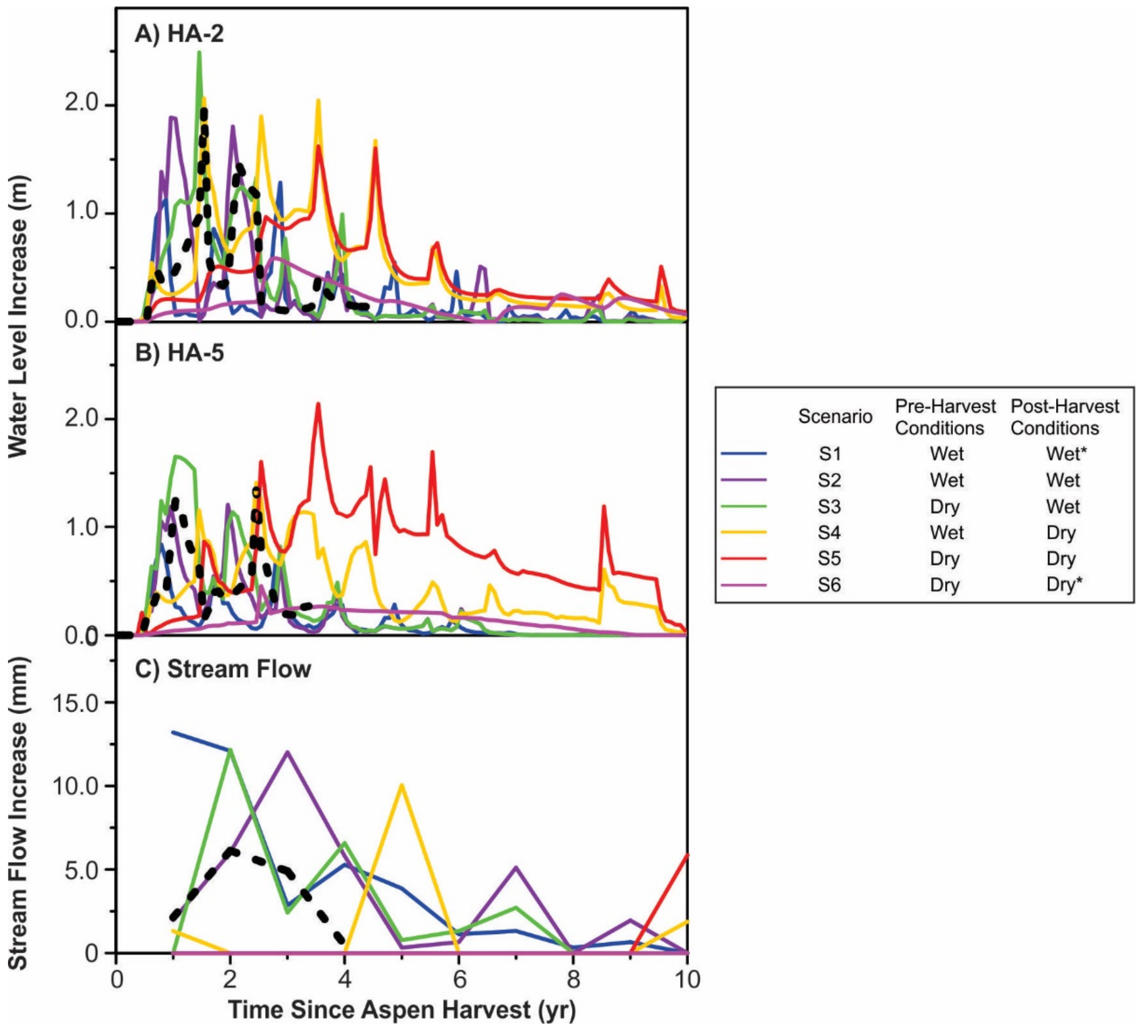

Figure 4-8. Predicted influence of different atmospheric conditions on A) monthly groundwater levels in the northern hillslope, B) monthly groundwater levels in the southern hillslope (B), and C) annual stream flows. Plotted groundwater levels and stream flows depict differences between synthetic harvested and unharvested cases. Post-harvest conditions marked with an * indicate cases with the wettest and driest years on record simulated sequentially. Predicted differences for the base case are also shown for reference (dashed lines).

CHAPTER 5

Potential Influence of Climate Change on Ecosystems within the Boreal Plains of Alberta[4]

5.1. Introduction

The landscape of the Boreal Plains of Alberta, Canada, is characterized by abundant shallow lakes and ponds, large wetland complexes comprised of thick peat deposits, and extensive upland forests situated within landscapes dominated by three glacially-derived landscapes (i.e., coarse-textured outwash, fine-textured hummocky moraine, and lacustrine clay-plains). These ecosystems exist under a delicate hydrologic balance due to the prevalence of water deficit conditions within the existing sub-humid climate (Marshall *et al.*, 1999). In a future warmer climate they may be particularly sensitive to changes in the timing and magnitude of precipitation, as well as increases in evapotranspiration from a longer, warmer growing season. Concurrent increases in wildfire frequency (Flannigan *et al.*, 2013; Thompson and Waddington, 2013) and northward migration of pests and diseases previously limited by cold winters (Lemmen *et al.*, 2008) may further stress these ecosystems with the potential for permanent peatland and forest cover loss or degradation and drying of shallow lakes and ponds (Hogg and Hurdle, 1995; Hogg and Schwarz, 1997; Cerezke, 2009). The negative impact of the decline of these ecosystems will have implications both regionally and globally due to their importance as a large carbon store (Kleinen *et al.*, 2012), spatially variable influence on climate (Krinner, 2003), source of seasonal habitat for migratory birds (Smith and Reid, 2013), and source of commercial lumber (e.g., aspen; David *et al.*, 2001). Therefore, understanding how these ecosystems may respond to future changes in climatic conditions is important for effective long-term management.

Global temperatures have risen over the last century, with each of the past three decades being successively warmer than any previously on record (IPCC, 2013). Since 1948, the annual temperature across the Boreal Plains has risen between 1 to 3°C, similar to the average for the rest of Canada and more than double that observed globally

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(Hengeveld *et al.*, 2005; Lemmen *et al.*, 2008). By the 2050s, annual temperature in the Boreal Plains may increase by a further 2 to 5°C relative to existing climate normals (Barrow *et al.*, 2004; Barrow and Yu, 2005; Christensen *et al.*, 2007; Lemmen *et al.*, 2008). Warmer spring and fall temperatures are likely to reduce the length of the snow-covered season, resulting in earlier spring blooms and longer growing seasons (Beaubien and Freeland, 2000; Brown *et al.*, 2010) combined with reduced ground frost or frozen soil (Smerdon and Mendoza, 2010). Thus, as the temperature rises the disparity between annual precipitation and evapotranspiration may be further widened due to greater evaporative water loss.

Future increases in evapotranspiration, however, may be offset by a concurrent rise in precipitation. Since 1948, rising temperatures within the region have not, yet, been accompanied by an appreciable increase in annual precipitation (Hengeveld *et al.*, 2005; Lemmen *et al.*, 2008); although increased precipitation of generally less than 5% is projected through the 2020s, with some climate change scenarios predicting an initial decrease (Barrow and Yu, 2005; Mbogga *et al.*, 2010). In the longer term, however, most projections do forecast a rise in precipitation by the 2050s, although there is considerable uncertainty in the predictions (Barrow and Yu, 2005; Mbogga *et al.*, 2010).

The future health and sustainability of Boreal Plains' ecosystems may also be sensitive to the evolution of precipitation patterns. Most groundwater recharge is currently derived from spring snowmelt, with comparatively less recharge occurring during the growing season except following larger rain events (Smerdon *et al.*, 2008; Redding and Devito, 2011; Chapter 2). For future scenarios, precipitation falling outside of the summer months is generally predicted to increase (Barrow and Yu, 2005; Mbogga *et al.*, 2010); however, it is currently unclear if this increase will be sufficient to offset the rise in evapotranspiration.

The net impact of climate change on Boreal Plains' ecosystems will not only be measured by changes in temperature and precipitation, but will also be a function of the differential responses of ecosystem components and the interaction amongst them. Peatlands, which retain large volumes of water within the landscape through internal feedback processes that act to mediate water table depth and reduce evapotranspiration, may be both resistant and resilient to the changing climate (Waddington *et al.*, 2015). However, ponds and upland forests are less resistant to increased evapotranspiration losses, and may draw increasing amounts of water from adjacent peatlands as temperatures rise (Schneider *et al.*, 2015), potentially overwhelming peatland feedback processes.

In this study, numerical simulations were used to examine the potential influence of climate change on water movement and availability within a pond-peatland-aspen forest ecosystem characteristic of low permeability glacial moraine settings within Alberta's Boreal Plains. Glacial moraine landforms within the region are generally typified by hydrologically isolated ponds and wetlands, and potentially represent landforms where ecosystems are the most susceptible to climate change across the Boreal Plain (Winter, 2000). A fully-integrated groundwater-surface water model previously calibrated (Chapter 2) to more than a decade of hydrologic observations was used for the analysis. The goals of the analysis were to address the following questions:

1. How will climate change influence the interaction between typical Boreal Plains ecosystems and which are predicted to be the most susceptible to a warmer climate?
2. Will the small lakes and ponds present within lowland depression areas remain permanent fixtures within the landscape, or will the extended ice-free season lead to them becoming ephemeral features due to enhanced evaporative losses?
3. Will peatland water levels decline significantly in a future warmer climate and what impact will this have on pond permanence?
4. Will enough water be retained in upland areas to sustain healthy forests and prevent northward migration of prairie grasslands?

To address these questions, 2D simulations were performed to the end of 2090 for thirteen climate change scenarios to encompass the large uncertainty in future climate projections and the potential range in hydrologic responses of the different ecosystems.

5.2. Study Area

The study was conducted at the catchment of Pond 43, located within the Utikuma Region Study Area (URSA) approximately 350 km northwest of Edmonton, Alberta, Canada (Lat: 56.07 N, Long: 115.5 W), and 150 km south of the discontinuous permafrost region (Woo and Winter, 1993; Figure 5-1). The catchment is underlain by deep, heterogeneous, glacial disintegration moraine deposits ranging in depth from 40 to 50 m

(Pawlowicz and Fenton, 2005). The glacial materials are underlain by marine shales of the Upper Cretaceous Smoky Group (Vogwill, 1978). Situated on a regional topographic high within the gently undulating glacial topography, ecosystems within the catchment may be highly vulnerable to climate change as the area functions as a regional recharge area with predominantly vertical groundwater flow (Ferone and Devito, 2004; Chapter 2).

Mature forests dominated by aspen, along with some balsam poplar and white spruce, cover the hillslopes. Near-surface soils within the hillslopes are classified as gray Luvisols and overlie oxidized clay till 5 to 8 m in depth with unoxidized clay till beneath (Ferone and Devito, 2004). In flatter areas and depressions, peatlands with low density stunted black spruce are the dominant surface cover and directly overlie unoxidized clay till. The peatlands range in depth from 2 to 5 m and transition to gyttja more than 3 m thick in the vicinity of the pond. Surface drainage within the peatlands is generally poorly developed (Ferone and Devito, 2004) and regularly modified by beaver activities.

The region is characterized by cold winters and warm summers, with average January and July temperatures of -14.5°C and 15.6°C, respectively. The climate is sub-humid, with average annual potential evapotranspiration (517 mm) slightly greater than average annual precipitation (485 mm; Marshall *et al.*, 1999). Wetter years where precipitation exceeds evapotranspiration are infrequent, occurring every 10 to 25 years (Mwale *et al.*, 2009). Most precipitation (50-60%) falls as rain during the summer months, coinciding with the growing season when the evapotranspiration demand is greatest. The autumn months are generally drier, with infrequent precipitation events. Average winter snowfall (i.e., 137 mm; Marshall *et al.*, 1999), generally represents less than 30% of annual precipitation.

5.3. Numerical Model

Numerical simulations were performed using HydroGeoSphere (HGS; Aquanty, 2013). HGS is a physically-based, fully-integrated, finite-element groundwater-surface water code that simultaneously solves the diffusion-wave approximation of the Saint Venant equations in the surface water domain and the variably-saturated Richards' equation in the subsurface domain. The numerical formulation allows the user to specify boundary conditions in the form of atmospheric fluxes, rather than specifying net groundwater recharge rates and surface water levels that necessitate additional *a priori* assumptions. HGS was selected as the numerical simulator due to its integrated simulation capabilities and previous successful application within the URSA (e.g., Smerdon *et al.*, 2008; Chapter 2).

The 2D model developed in Chapter 2 was used to assess hydrological impacts that may occur as a result of climate change within the region (Figures 5-1 and 5-2). The model was calibrated to over a decade of data including dry through relatively wetter climatic conditions (Table 5-1) and used to assess the interactions occurring between the pond, peatlands, and uplands, as well as to evaluate the sensitivity of the system to a range of soil characteristics and catchment configurations common to the region.

The model domain was discretized using uniform 1 m finite-elements horizontally and vertical node spacing that varied from 0.05 m to 0.25 m. Daily boundary conditions were implemented with simulations performed using an adaptive time-stepping scheme available in HGS that allowed for refined temporal discretization (i.e., sub-daily time-steps). Further details regarding the data used to parameterize the model, the numerical model settings, and previous simulation results can be found in Chapter 2.

5.3.1. Boundary Conditions

Daily fluxes of precipitation and evapotranspiration were applied to the surface of the model. Daily precipitation consisted of rain and spring snowmelt, and varied by surface cover to account for rainfall interception within forested areas and differences in sublimation and snowmelt timing. Within the forested areas, the specified interception was based on throughfall data collected within the study catchment (Chapter 2). During the summer months when evapotranspiration is highest, daily interception was set to 2 mm, with the remaining quantity of rainfall reaching the ground. During the cooler month of May and the winter months, daily interception was reduced to 0.5 mm and 0 mm, respectively. Within the pond, rainfall interception was assumed to be negligible throughout the year.

The duration of the frozen season, neglecting seasonally frozen peat at depth, was estimated for each scenario using daily temperature projections. The onset of the frozen period was assumed to be the same at all locations, occurring each year once the daily temperature remained below 0°C. Within the aspen areas, the frozen season was assumed to continue until the mean daily temperature remained above 0°C. In the respective footprints of the pond and peatlands, the duration of the frozen season was assumed to extend 7 and 14 days longer based on field observations of the snowpack and water levels from the study area between 2000 and 2011. Use of the 0°C threshold yielded frozen seasons that were similar in duration at the start of the simulations to what has been observed under the existing climate (Chapter 2), and allowed its length to decrease in response to warming climatic conditions.

Snow accumulating during the winter was calculated manually external to the model with the assumption that all precipitation falling during the frozen period was incorporated into the snowpack. Snowmelt was specified to commence at the end of the winter period which varied by surface cover, and ranged from 7 to 14 days depending on the depth of snow accumulation. Combined snowmelt evaporation and sublimation equal to 25% and 30% of annual snowfall were assumed within the respective forested (i.e., aspen hillslopes and black spruce peatlands) and pond areas (Chapter 2).

Daily evapotranspiration applied to the model's surface was calculated using the temperature-based Hamon (1963) method and was scaled by surface cover (Chapter 2) based on available measurements within the study catchments (Brown *et al.*, 2014; Brown *et al.*, 2010; Petrone *et al.*, 2007). Within the peatlands, evapotranspiration was applied throughout snow-free periods (Brown *et al.*, 2010). In aspen dominated areas evapotranspiration was restricted to the growing season, which increased in duration in response to rising temperatures as observed within the region (Beaubien and Freeland, 2000). The aspen growing season was assumed to extend from 1 month after the spring thaw to 1 month prior to the onset of the snow-covered period. Under this assumption, the length of the growing season at the start of the simulations was similar to what has been observed within the study area, extending from mid-May to mid-September (Brown *et al.*, 2014). The evapotranspiration depth was specified based on typical rooting depths within the aspen hillslopes (3 m; Debyle and Winokur, 1985), peatlands (0.5 m; Lieffers and Rothwell, 1987) and pond (0.1 m) and was focused at near-surface nodes. The actual evapotranspiration was limited by the available water, with the maximum rate occurring at saturations greater than the field capacity (i.e., -33 kPa) and decreasing to zero at the permanent wilting point (i.e., -1500 kPa).

Additional boundary conditions included discharge from the pond constrained by water level, specified heads at the base of the domain, and zero discharge conditions along the remaining model edges. The discharge boundary applied within the pond was utilized to simulate stream outflow. Use of this boundary type allowed water to discharge from the model whenever the pond stage rose above the elevation of the outflow channel.

At the model base, a uniform constant hydraulic head was applied to simulate connection with the regional groundwater flow system based on field measurements (Chapter 2). Along the lateral edges, zero discharge boundaries were specified as groundwater flow has been interpreted to be predominantly vertically downwards.

5.3.2. Material Properties

The distribution of material zones (Figure 5-3) and assigned parameters (Table 5-2) were unmodified from the calibrated base case simulation detailed in Chapter 2. Three primary hydrogeologic units were specified within the model including peat, gyttja, and glacial till. Highly decomposed organic materials at the edge of the peatland were also included within the till, as their hydraulic properties were found to be similar to the till (i.e., low permeability and porosity, high bulk density). Within aspen areas, a 0.1 m thick forest floor was also specified above the glacial till. Within each material, the saturated hydraulic conductivity was specified to decrease with depth based on observed depth trends. An anisotropy ratio of 10:1 and 100:1 was assumed for the organic materials and glacial till, respectively. Soil water characteristic and unsaturated hydraulic conductivity curves were derived from the literature for the peat, gyttja, and forest floor (Silins and Rothwell, 1998; Price *et al.*; 2010), and glacial till using the properties of a clay loam (Carsel and Parrish, 1988).

5.3.3. Climate Change Scenarios

Thirteen climate change scenarios were used for this study, with the selected scenarios chosen to bracket the range in variability of future projections for the region (Table 5-3). The software ClimateAB (Mbogga *et al.*, 2009, Mbogga *et al.*, 2010) and ClimateWNA (Wang *et al.*, 2012; Wang *et al.*, 2014) were used to extract predictions for the study area from each of the selected scenarios. Projections from these scenarios indicate a warming trend will occur over the next several decades. Relative to the existing climate normal, similar increases in annual temperature between scenarios are predicted for the 2020s (i.e., 1 to 2°C, Table 5-3) and 2050s (i.e., 2 to 5°C) for most scenarios. The range in projections for the 2080s is similar to the 2050s with the exception of the warmest scenario (i.e., CCSRNIES-A1F1), which has a predicted increase of 10°C relative to the existing climate. Seasonally, a larger range in future temperature projections is present, with some projections predicting larger rises in the winter and spring months (e.g., CCNRIES-A1F1, Figure 5-3A), and others predicting a larger increase during the summer (e.g., HADCM3-A2a).

Large variations exist between scenarios for projected changes to annual precipitation. In the 2020s, the CGCM2-B23, CCSRNIES-A1F1, and CCSM4-RCP2.6 scenarios predict a small decrease in precipitation, while the remaining scenarios predict an increase of up to 9% (Table 5-3). By the 2050s, annual precipitation is predicted to be elevated above the existing climate normal by 1 to 9% for all scenarios except for CGCM2-B23 and CCSM4-RCP2.6, where it is predicted to decline by a further 3 and 1%, respectively. However, by the 2080s, annual precipitation is predicted to rise in all scenarios, ranging from 2 to 20% greater than the current climate normal. On a seasonal basis, most scenarios predict that precipitation will either increase or remain similar to the existing climate normal in the months of October to May (Figure 5-3B). Despite the projected general rise in winter precipitation, the proportion falling as snow is generally predicted to decrease by 4% (i.e., CCSM4-RCP2.6) to 50% (i.e., CCNRIES-A1F1) by the 2080s as the climate warms. Exceptions include the HadCM3-B2b and HadCM3-A2a scenarios, where snowfall is predicted to increase by 7 and 23%, respectively. During the months of June through September, more variability is present in the projections, where a roughly equal number of scenarios predict either an increase or a decrease in precipitation.

5.3.4. Simulations Scenarios

Simulations were performed for the time period from the middle of March 2011 to the end of 2090, with results from the end of the calibrated simulation detailed in Chapter 2 used as an initial condition. Climatic conditions prior to 2011 were relatively dry, with nine of the twelve preceding years receiving less than average precipitation (Table 5-1). However, a series of near average to wet years occurred from 2005 to 2010. Thus, the initial conditions were interpreted to be equivalent to approximately average conditions.

Thirteen simulations were performed which incorporated the climate change projections summarized in Table 5-3. A further base case climate scenario that assumed no change to the existing climate was included to directly evaluate the impact of each scenario. The long-term climatic dataset available from Fort McMurray, Alberta (Environment Canada, 2016; Figure 5-4), was used to provide daily boundary conditions for the base case climate simulation and as a basis for developing synthetic future conditions for the different scenarios. Inputs for the climate change simulations were generated by perturbing the daily base case climate dataset using projected changes in monthly precipitation and temperature (i.e., relative to existing climate normals). Using this method, historical fluctuations in climatic conditions were preserved while allowing the precipitation and temperature to evolve as predicted by each climate change scenario (Figure 5-5).

Further simulations were also conducted to evaluate the sensitivity of model predictions to a number of assumptions incorporated into the analysis. For each sensitivity scenario, simulations were performed for the CCNRIES-A1F1, CGCM2-B23, and HADCM3-B2b scenarios to evaluate the overall sensitivity of the hydrologic system under a range of climatic conditions. These three scenarios were chosen because they approximately bracket the range in projected climate change for the study area (Figure 5-5).

5.4. Simulation Results

Simulation results indicate that in the near-term, a warming climate may exert limited influence on water levels within study area ecosystems (Figures 5-6 and 5-7). To the end of the 2020s, little change in water levels (i.e., centimeters) relative to the base case simulation is predicted within the pond and peatlands for all scenarios (Figure 5-7A and B). Base case climatic conditions during this period are similar to climatic normals for the region, with approximately an equal number of years receiving either below or above average precipitation (Figure 5-4). As a result, sufficient near-surface water is available within the peatlands to satisfy the evapotranspiration demand, preventing water levels from being significantly impacted by the small initial increases in evapotranspiration. Variability in evapotranspiration between scenarios does result in subtle differences in predicted stream flow from the pond (Figure 5-8A), although the range is relatively small. Somewhat greater variability in water levels between scenarios is predicted within the aspen area (Figure 5-7C) as insufficient soil moisture is available to satisfy the entire evapotranspiration demand, leading to water being drawn upwards from the water table.

Beginning in the 2030s, larger differences in water levels are predicted to develop for most scenarios relative to the base case that persist until the late 2040s (Figure 5-7). This period coincides with the start of a simulated prolonged dry period that continues to the mid-2040s before a series of wetter years occur (Figure 5-4). During this period, the pond is predicted to go dry for extended periods of time (i.e., weeks to months) in the warmer and/or drier scenarios. As a result, limited difference in stream flows are predicted as little to no outflow occurs (Figure 5-8A).

The greatest decline in water levels during the dry period is predicted for the warmer (e.g., CCNRIES-A1F1 and CSIRO-Mk3.6.0-RCP8.5) and drier (e.g., CGCM2-B2) scenarios, where changes in precipitation are insufficient to offset increasing evapotranspiration. In contrast, for the wetter and/or cooler scenarios (e.g., HADCM3-B2b) small differences in water levels are predicted relative to the base case, particularly in the pond and peatlands (Figure 5-7A and B). Thus, in these scenarios, the increased precipitation and resulting available net precipitation (i.e., precipitation minus evapotranspiration; Figure 5-9) is sufficient to reduce the impact of the dry period and prevent significant water level declines. For the remaining scenarios, subtle differences in the projected timing and magnitude of precipitation and temperature changes are predicted to lead to similar declines in water levels from 2030 to the late 2040s, although the decline is generally subdued.

From the late 2040s to the mid-2050s, water level differences (Figure 5-7) between the climate change scenarios and the base case simulation are predicted to be smaller as a series of wetter years occur (Figure 5-4). During this period, sufficient near-surface water is available in storage to satisfy the evapotranspiration demand, preventing large declines in water levels in a fashion similar to the initial years of the simulations. However, greater differences in pond and peatland net precipitation (Figure 5-9A and B) between scenarios result in more variability in predicted stream flows (Figure 5-8A).

For the remainder of the simulations, predicted water levels continue to be lowest for the warmest scenarios (i.e., CCNRIES-A1F1 and CSIRO-Mk3.6.0-RCP8.5) which have the lowest net precipitation near the end of the century (Figure 5-9). In these scenarios, the increasing precipitation is unable to keep pace with the more rapidly increasing evapotranspiration, leading to the pond drying out frequently towards the end of the simulations (Figure 5-8B). By comparison, water level differences are predicted to become more modest in the latter part of the simulation for the previously dry CGCM2-B23 scenario, as the projected annual precipitation begins to rise above the existing climate normal in the 2080s. At the same time, water levels for the wetter scenarios are predicted to continue to remain similar to the base case climate simulation, although stream flow from the pond does begin to drop from about 2070 to 2080 as the climate continues to warm.

5.4.1. Sensitivity Simulations

5.4.1.1. Peatland Evapotranspiration

Within the base climate change simulations, peatland evapotranspiration was assumed to decline to zero at the permanent wilting point of -1500 kPa. Although black spruce present within the peatlands may continue to draw water at this pressure (Lamhamedi and Bernier, 1994), mosses forming the surface cover of the peatlands may become water-stressed at lower soil tension and may reduce evapotranspiration losses through a number of internal processes (Waddington *et al.*, 2015). For Sphagnum mosses, which are the dominant surface cover within the peatlands (Petrone *et al.*, 2008), the biological limit of soil water pressure may range from -58 kPa to -9 kPa (McCarter and Price, 2014). Therefore, for these sensitivity simulations, evapotranspiration was specified to be limitied by the saturation corresponding to a capillary pressure of -58 kPa. The base case climate scenario was also run using this same limiting condition within the peatlands to allow direct comparison of the results.

Simulation results indicate that peatland water levels are sensitive to this parameter. Modification of the limiting saturation is predicted to result in peatland water levels that are similar to the base case climate simulation, as less water is drawn upwards through evapotranspiration (Figure 5-10A). The influence of this parameter is particularly evident during the dry period of the 2030s, where peatland water levels are predicted to be almost 0.3 m higher for CCNRIES-A1F1 and CGCM2-B23 relative to the base climate change simulations. As a result, the decline in pond water levels is also predicted to be reduced as more water is available within the peatlands to maintain the pond. However, for CCNRIES-A1F1, the increased peatland discharge is predicted to be insufficient to offset the elevated pond evapotranspiration by the 2070s, resulting in the pond drying out frequently (results not shown).

5.4.1.2. Sublimation and Snowmelt Evaporation

The second set of sensitivity simulations examined the influence of the assumed proportion of water lost to the atmosphere during the winter through sublimation and melt water evaporation. While snow sublimation may decrease in forested areas in response to warmer winter temperatures (Rasouli *et al.*, 2014), total water loss could rise due to increased winter evaporation (Arain *et al.*, 2003), particularly where surface ponding may be enhanced by shallow ground ice (Brown *et al.*, 2010). To simulate this situation, the atmospheric water loss was assumed to linearly increase by 15% to a total of 40 and 45% of the winter snowpack by 2090 for the respective forested areas and pond.

Simulation results indicate that on an annual basis, the hydrologic system is not very sensitive to this boundary assumption (Figure 5-10B), with the exception of CCNRIES-A1F1 after about 2075. Although the spring peak in water levels is lower, for the remainder of the year water levels are generally only marginally lower as less water is removed by evapotranspiration. However, declines in stream flow (results not shown) are apparent for each case, particularly for the wetter HADCM3-B2b scenario. For CCNRIES-A1F1, decreased spring snowmelt combined with increasing evapotranspiration is predicted to prevent peatland water levels from recovering after the series of simulated dry years in the early 2070s, leading to the pond and peatland water levels declining earlier relative to the base simulation.

5.4.1.3. Precipitation Intensity

Projections for the region indicate that precipitation may be increasingly focused within a series of wet days (i.e., days with precipitation >1 mm) as climate change progresses (Sillmann *et al.*, 2013). The greatest change is projected to occur in the spring (Mailhot *et al.*, 2010), with 5-day precipitation totals increased by up to 25% (Sillmann *et al.*, 2013). However, the projections also indicate that this likely will not be accompanied by a change in the return period of extreme events or the duration of consecutive dry days (Mailhot *et al.*, 2010; Sillmann *et al.*, 2013). Although these projections were performed at coarse scales (i.e., nationwide to global), they do provide an indication of potential future trends. Therefore, for these sensitivity simulations, 5-day precipitation totals during wet periods in the months of May and June were manually increased by 25% by the end of 2090 by modifying the distribution of precipitation. Using this method, total monthly precipitation and the duration of consecutive dry days for each scenario were unchanged from the base climate simulations.

Results indicate that the hydrologic system is not sensitive to spring precipitation intensity. In all scenarios, sufficient storage is available in the spring months of May and June to prevent large increases in both water levels across the catchment (Figure 5-10C) and stream flows from the pond. The simulations did not, however, incorporate the potential effect of frozen materials (i.e., surface ice or snow, subsurface ice lenses), which might restrict infiltration and lead to generation of overland flow and potential flooding. Furthermore, the daily boundary implementation used within this analysis limits the model's predictive ability with regards to short-term effects.

5.5. Discussion: Future Vulnerability of Boreal Plains' Ecosystems

5.5.1. Ponds: Diminishing Water Resources

Study results indicate that less water will be stored in the small, shallow ponds characteristic of Alberta's Boreal Plains as the climate warms and the duration of the ice-free season increases, with the potential for encroachment of neighboring terrestrial vegetative species. Particularly in the warmer scenarios with smaller net precipitation (e.g., CCNRIES-A1F1 and CSIRO-Mk3.6.0-RCP8.5), pond water levels are predicted to be low later in the century, with the pond becoming largely ephemeral after about 2075 (Figures 5-7A and 8B).

Cumulative net precipitation for the pond is projected to remain near or below zero from about the 2060s for about half of the climate change scenarios considered (Figure 5-9A). Thus, the pond may be susceptible to drying out during short-term drought conditions. Stream flows from the pond are also predicted to decline over this time (Figure 5-8A), which may further impact ponds located downstream (e.g., Pond 48, Figure 5-1). Discharge of water from the peatlands may aid in maintaining pond water levels. However, sensitivity simulation results indicated that the pond would still dry out frequently late in the century for the warmer CCNRIES-A1F1 scenario despite higher peatland water levels. Furthermore, during extended dry periods such as that simulated from the mid-2030s to the mid-2040s, large (i.e., 0.1 to 0.6 m) pond water level declines were predicted in most scenarios. As the climate continues to warm later in the century and beyond, the drop in pond water levels may be enhanced where similar dry conditions occur.

Dynamic biotic (e.g., beaver, ungulates such as moose, peat accumulation) and abiotic influences (e.g., sedimentation rates, shallow frozen ground) not considered within this analysis may act to mitigate (e.g., dam building by beaver; Hood and Bayley, 2008) or enhance (e.g., pond bank erosion by ungulates; Ireson *et al.*, 2015) climate change effects on shallow Boreal Plains ponds. However, any mitigating factors that lead to increased surface water cover during the warmer summer seasons will also lead to increases in evaporation (Ireson *et al.*, 2015), potentially enhancing overall ecosystem water loss. Thus, suitable habitat for aquatic organisms, mammals, and migratory birds may be diminished in some years, with potentially cascading impacts on neighboring forest ecosystems (e.g., through reduced bird insect predation; Ireson *et al.*, 2015).

5.5.2. Peatlands: A Story of Resistance and Resilience?

Study results suggest that, in contrast to the pond, the impact of a changing climate will be less on the neighboring peatlands. Although traditional thinking would suggest that they, too, are highly vulnerable to the effects of climate change due to their primary reliance on precipitation (Winter, 2000), the lower evapotranspiration and higher net precipitation (Figure 5-9B) of the peatlands results in predicted water levels generally dropping by less than 0.2 m relative to the existing climate, except during sustained dry periods (Figure 5-7B). In the warmer and drier scenarios, greater water level declines of up to 1.1 m are predicted. However, the ability of peat-forming mosses to reduce evapotranspiration losses as the water table drops (Waddington *et al.*, 2015) may reduce water level declines in even the warmest scenarios (Figure 5-10A), making them resistant to changing climatic conditions (Schneider *et al.*, 2015).

Despite the predicted generally low decrease in peatland water levels, the greater depth to water table will expose more peat to aerobic conditions, with the potential for increased peatland decomposition (Cerezke, 2009) and susceptibility to fire (Flannigan *et al.*, 2013). Furthermore, the encroachment of trees and/or shrubs (e.g., aspen) may enhance peatland water level declines through increased interception and evapotranspiration (Waddington *et al.*, 2015) as recently observed by Kettridge *et al.* (2015). Losses at the forest-peatland edge may be offset by encroachment of the peatlands into adjacent dry ponds with the potential for increased carbon storage. Several peatland processes may act to moderate changes in water table depth by reducing peat hydraulic conductivity (e.g., compression, general decrease in hydraulic conductivity with depth), lowering evapotranspiration losses by increasing peat surface albedo, and releasing soluble phenols and methane through decomposition that may slow or even halt further degradation (Waddington *et al.*, 2015). This could result in maintaining peat water content and increasing overall peatland resilience within a future warmer climate (Kettridge and Waddington, 2014). However, the relative strength of these mechanisms requires further evaluation and may be expected to vary both spatially and temporally (Kettridge *et al.*, 2015; Waddington *et al.*, 2015).

5.5.3. Aspen Forests: An Ominous Future?

Results of the numerical simulations indicate that water levels within aspen-forested hillslopes may frequently be reduced by 0.5 to 1 m relative to the existing climate (Figure 5-7C) as the available water declines due to the increased growing season length and warmer climate. As a result, near-surface soils are predicted to become progressively drier making transpiration increasingly difficult. Tabulated cumulative net precipitation further suggests that water deficit conditions may become the norm, particularly after 2070 when all scenarios project a negative trend (Figure 5-9C). Thus, existing aspen stands within the region are likely to become increasingly water-stressed, making them more susceptible to other maladies such as pests and disease (Lemmen *et al.*, 2008; Worrall *et al.*, 2013) as well as projected increases in fire frequency (Flannigan *et al.*, 2013), which may potentially lead to declines in productivity and increased mortality (Bernier *et al.*, 2006; Barr *et al.*, 2007; Hogg *et al.*, 2008; Schneider *et al.*, 2015).

Rapid decline and mortality of aspen stands have already been observed in recent years across widespread areas of Alberta and Saskatchewan, eastern Canada, and the southwestern United States (Rehfeldt *et al.*, 2009). In most of these regions, the decline has been attributed to prolonged and often severe periods of drought, although secondary contributing factors such as defoliating insects and disease have exacerbated the situation (Worrall *et al.*, 2013). Further climate modelling suggests that the climate may become unsuitable for the continued persistence of existing aspen stands throughout large portions of Western Canada's Boreal Plains (Bergengren *et al.*, 2011; Worrall *et al.*, 2013), thus presenting a potentially ominous future for the aspen resource with possible northward migration of prairie grasslands (Hogg and Hurdle, 1995; Bergengren *et al.*, 2011) and invasion of agronomic grass species planted along the thousands of kilometers of roadsides, pipeline right-of-ways, and seismic lines within the region (Schneider *et al.*, 2015).

Although dieback of existing aspen stands may become increasingly common, mortality may be initially mediated by the ability of aspen to draw water over large lateral distances (i.e., upwards of 30 m) through their clonal root system (Debyle and Winokur, 1985); but this may be limited by hydraulic impairment/failure of distal roots (Anderegg *et al.*, 2012) over prolonged droughts. There may be opportunity for migration northward and further up in elevation where shifting climatic conditions may become suitable for aspen development (Worrall *et al.*, 2013). However, northward migration rates (i.e.,

~10 km per century; Price *et al.*, 2013) may lag behind the northward shift of viable climatic conditions for aspen growth (i.e., >500 km by the 2050s; Gray and Hamann, 2013), potentially necessitating the need for assisted migration (Cerezke, 2009). The drying of neighboring depression areas currently occupied by ponds and peatlands may also provide additional suitable habitat (Cerezke, 2009). Furthermore, deterioration of existing stands may be mitigated through silvicultural (e.g., periodic thinning, removal of diseased trees) and pesticidal treatments (Cerezke, 2009); although these measures may only provide temporary relief as climate change progresses.

5.5.4. Study Limitations and Future Research

Results of this study provide an indication of the hydrologic response that may be typical within ecosystems characteristic of Alberta's Boreal Plains as climate change progresses; however, the analysis employed several assumptions that lead to a degree of uncertainty within the predictions. The numerical simulations were driven by climatic projections from thirteen different scenarios superimposed on an historic climate dataset. Although it appears clear that temperatures are rising (IPCC, 2013), prediction of future precipitation patterns remains inherently difficult (Barrow and Yu, 2005; Mbogga *et al.*, 2010), and future trends may differ from historical patterns (e.g., frequency of wet and dry years, drought duration). Further investigation into the evolution of regional precipitation trends is warranted, as future climatic trends may lie outside the range investigated here with the potential for greater long-term impact.

The future evolution of potential evapotranspiration was estimated using the Hamon equation along with projected changes in temperature. However, the Hamon method may be limited in its predictive ability, as changes in temperature are not always strongly correlated to changes in potential evapotranspiration (Shaw and Riha, 2011; McAfee, 2013). Future studies could benefit from a more rigorous estimation of future potential evapotranspiration.

Further assumptions incorporated into the analysis include the exclusion of secondary factors that may act to amplify and/or moderate climate change impacts. Examples include enhanced virulence of pests and disease (Lemmen *et al.*, 2008), increased fire frequency (Flannigan *et al.*, 2013; Thompson and Waddington, 2013), peatland evaporative, hydraulic conductivity, and decomposition feedbacks (Waddington *et al.*, 2015) vegetation succession (Hogg and Hurdle, 1995; Hogg and Schwarz, 1997;

Cerezke, 2009), seasonal subsurface freezing (Petrone *et al.*, 2008), biotic and abiotic influences on pond water levels (Hood and Bayley, 2008; Ireson *et al.*, 2015), and encroachment of anthropogenic developments (Devito *et al.*, 2012). Each of these factors may prove important over a range in spatial and temporal scales with regards to ecosystem health and stability, with a large number of potential scenario trajectories. Possible influences of these secondary factors on model predictions have been hypothesized in the preceding discussion sections; however, further research into the relative magnitude of their influence and to how they may be represented within existing and future simulation tools is warranted.

5.6. Conclusions

Numerical simulations were used to investigate the potential influence of climate change on water movement and availability within ecosystems at a catchment characteristic of Alberta's Boreal Plains. Climate projections from thirteen scenarios were simulated from 2011 to 2090 and compared to a base case scenario that assumed no change to the existing climate. Primary conclusions that can be drawn from the results of the study include:

1. Water levels within the small ponds and lakes that are ubiquitous to the region will be reduced as the climate warms, particularly if the future climate manifestation is similar to the warmer and drier scenarios considered (e.g., CCNRIES-A1F1). During extended dry periods and later in this century (i.e., after 2075), the simulated pond is predicted to dry out frequently, with negative implications for migratory bird species that frequent the region. Predicted concurrent decreases in stream flow may further negatively impact downstream ecosystems.

2. The large wetland complexes that occupy flat and depression areas within the region may be resistant to changing hydrological conditions associated with a warming climate. Internal peatland mechanisms that act to regulate the water table depth and maintain peat water content may prevent them from undergoing large decreases in water levels, at least in the first half of this century. However, this conclusion is predicated on the assumption that these internal mechanisms will not be outweighed by external stresses such as fire, encroaching trees or shrubs, and anthropogenic disturbance (e.g., resource development). Inclusion of these processes in future simulation tools would help reduce the uncertainty in model predictions.

3. Increasing evapotranspiration losses from the aspen-dominated hillslopes will lead to near-surface soils becoming progressively drier, rendering them more frequently water-stressed and susceptible to fire, pests, and disease. Simultaneous declines in productivity and increased mortality may reduce the economic viability of the aspen resource, with the potential for northward migration of prairie grasslands and by invasion of agronomic grass species.

Differences in the timing and magnitude of responses in ponds and aspen forests relative to adjacent peatlands may have further profound impacts on the evolution of inter-ecosystem interactions. The net impact of differential responses to climate change in this setting is difficult to determine, but could result in the development of novel landscapes with a configuration of ecosystems not currently observed (Schneider *et al.*, 2015).

Results of this study and the conclusions drawn should be considered in the context of the processes considered along with their inherent uncertainty. Although indicative of potential future trends in regional ecosystem health related to water availability, several simplifying assumptions and excluded processes have been identified. Future research into regional ecosystem health related to climate change could benefit from the development of improved numerical tools capable of extending the processes considered.

Table 5-1. Summary of study area annual climatic conditions during model calibration period.

Year	Precipitation (mm)	Temperature (ºC)	Evapotranspiration [a] (mm)
2000	442	0.9	410
2001	258	1.8	440
2002	289	-1.0	405
2003	442	-0.5	421
2004	377	0.8	391
2005	491	2.8	430
2006	432	2.9	411
2007	530	1.6	463
2008	505	1.3	450
2009	391	0.4	437
2010	282	2.4	481

Notes:

[a] Pond evapotranspiration.

Table 5-2. Subsurface parameters for hydrogeologic units used in the numerical models and shown in Figure 5-2.

Material	Depth Range (m)	Hydraulic Conductivity (m/s)		Porosity (-)	Specific Storage (m^{-1})
		Horizontal	Vertical		
Peat	0.0 - 0.1	3×10^{-3}	3×10^{-4}	0.90	1×10^{-4}
	0.1 - 0.3	3×10^{-4}	3×10^{-5}	0.82	5×10^{-5}
	0.3 - 0.5	8×10^{-5}	8×10^{-6}	0.72	8×10^{-6}
	0.5 - 1.0	4×10^{-5}	4×10^{-6}	0.60	2×10^{-6}
	1.0 - 1.5	2×10^{-6}	2×10^{-7}	0.50	5×10^{-7}
	1.5 - 2.0	3×10^{-8}	3×10^{-9}	0.45	2×10^{-7}
	2.0 - Base	1×10^{-8}	1×10^{-9}	0.40	1×10^{-7}
Gyttja	0.0 - 1.0	1×10^{-6}	1×10^{-7}	0.45	3×10^{-6}
	1.0 - 2.0	3×10^{-8}	3×10^{-9}	0.30	4×10^{-7}
	> 2.0	5×10^{-9}	5×10^{-10}	0.22	1×10^{-7}
Glacial Till	Upper	1×10^{-5}	1×10^{-7}	0.20	1×10^{-4}
	Mid	5×10^{-7}	5×10^{-9}	0.20	1×10^{-4}
	Lower	1×10^{-8}	1×10^{-10}	0.20	1×10^{-4}
Forest Floor	0.0 - 0.1	1×10^{-4}	1×10^{-4}	0.80	1×10^{-4}

Notes:

[a] Specified specific storage values for peat and gyttja are low; however, the influence on simulation results is negligible due to small fluctuations in water levels (i.e., less than 1 m).

Table 5-3. Summary of annual temperature and precipitation for climate change scenarios used for the numerical simulations relative to regional climate normals (1961 to 1990) for T (0.7 °C) and P (485 mm).

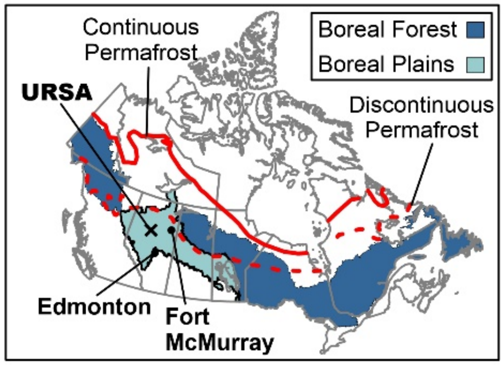
Climate Modeling Center	Scenario[c]	Acronym	Annual Temperature (°C)			Annual Precipitation (mm)		
			2020s	2050s	2080s	2020s	2050s	2080s
Center for Climate Research Studies/National Institute for Environmental Studies (Japan)	A1F1	CCSRNIES-A1F1*	1.5	5.3	10.5	475	516	575
Commonwealth Scientific and Industrial Research Organization (Australia)	8.5	CSIRO-Mk3.6.0-RCP8.5	1.4	3.6	6.1	487	491	494
Canadian Center for Climate Modelling and Analysis (Canada)	A2	CGCM2-A2	1.9	3.2	5.0	495	499	508
Hadley Center for Climate Prediction and Research (UK)	A2a	HadCM3-A2a	1.5	2.3	4.3	504	528	528
Canadian Center for Climate Modelling and Analysis (Canada)	B23	CGCM2-B23*	2.3	3.1	3.9	483	469	502
Canadian Center for Climate Modelling and Analysis (Canada)	B2	CGCM2-B2	2.0	2.9	3.8	491	496	498
Hadley Center for Climate Prediction and Research (UK)	B2b	HadCM3-B2b*	1.6	2.8	3.7	517	528	544
National Center for Atmospheric Research (USA)	A1B	NCARPCM-A1B	1.8	3.1	3.6	498	523	515
Meteorological Research Institute (Japan)	8.5	MRI-CGCM3-RCP8.5	1.1	2.0	3.5	499	515	528
Canadian Center for Climate Modelling and Analysis (Canada)	B1	CGCM2-B1	1.9	2.6	3.3	488	493	496
Max-Planck Institute for Meteorology (Germany)	B2	ECHAM4-B2	1.5	2.0	2.6	528	529	529
Meteorological Research Institute (Japan)	4.5	MRI-CGCM3-RCP4.5	0.5	1.4	2.1	489	508	502
National Center for Atmospheric Research (USA)	2.6	CCSM4-RCP2.6	1.6	1.8	2.0	480	475	499

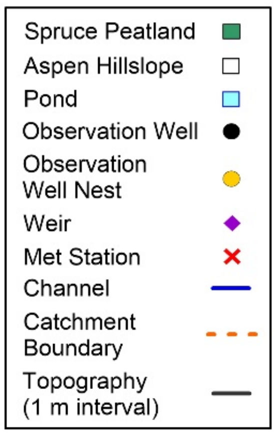
Notes:

[a] Scenarios ranked by 2080s temperature then by precipitation.

[b] Three scenarios indicated with an * were used for the sensitivity simulations.

[c] Special report on emissions scenarios (SRES) and representative concentration pathways (RCP; IPCC, 2013).





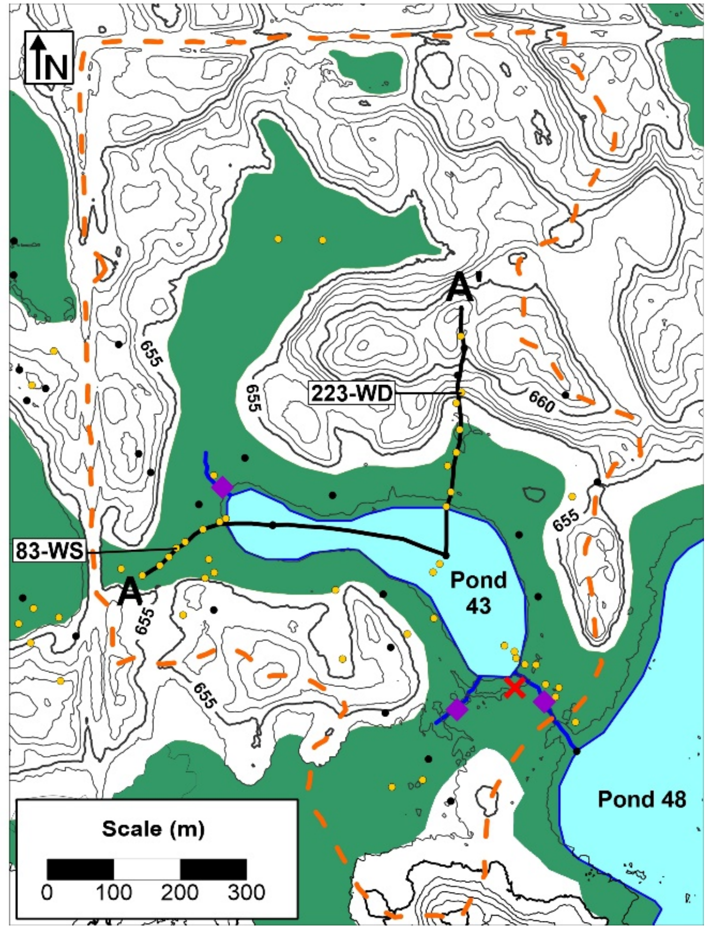


Figure 5-1. Left: Location of the Utikuma Region Study Area (URSA) within the Canadian Boreal Plains relative to the discontinuous permafrost zone. Right: Enlarged map showing the Pond 43 study area including instrumentation, vegetative cover, selected observation well locations, surface water catchment boundaries, and the location of the 2-D numerical model (Section A-A').

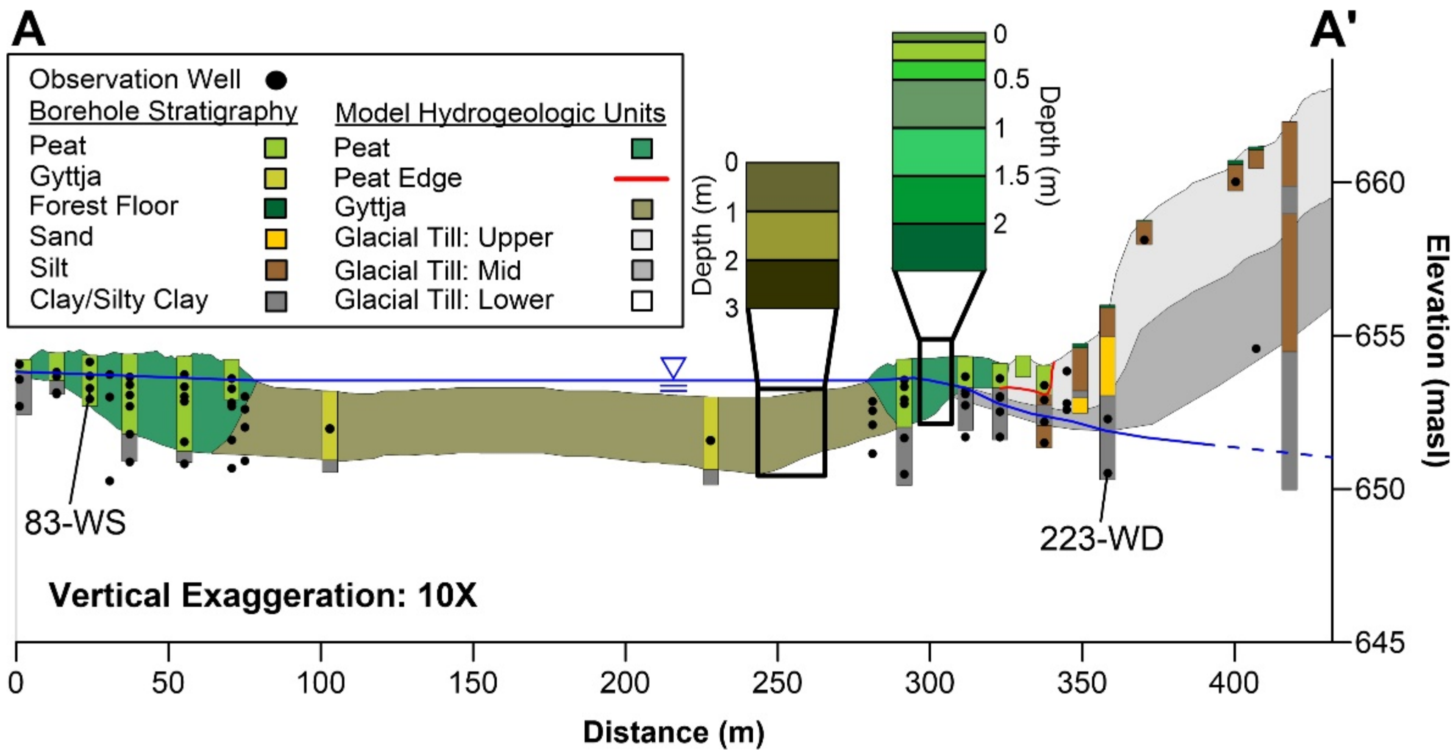


Figure 5-2. Model domain for cross-section A-A', including hydrogeologic units, observation points, and average water table. Inset columns show layering specified within the peat and gyttja. Note that the base of the model has been truncated at 645 m asl for illustration purposes.

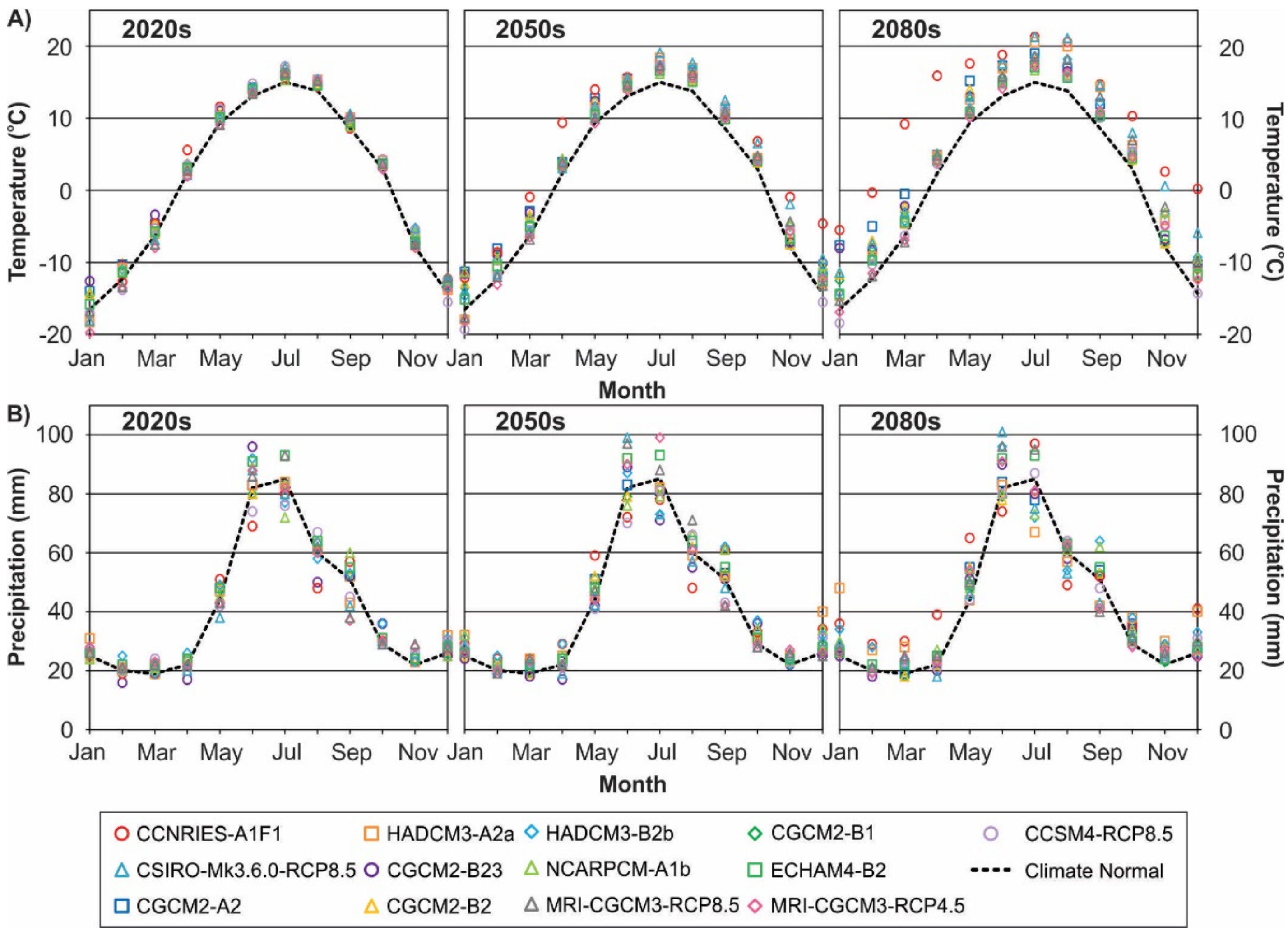


Figure 5-3. Monthly predicated (a) temperature and (b) precipitation for the 2020s, 2050s, and 2080s for 13 climate change scenarios considered along with climatic normals.

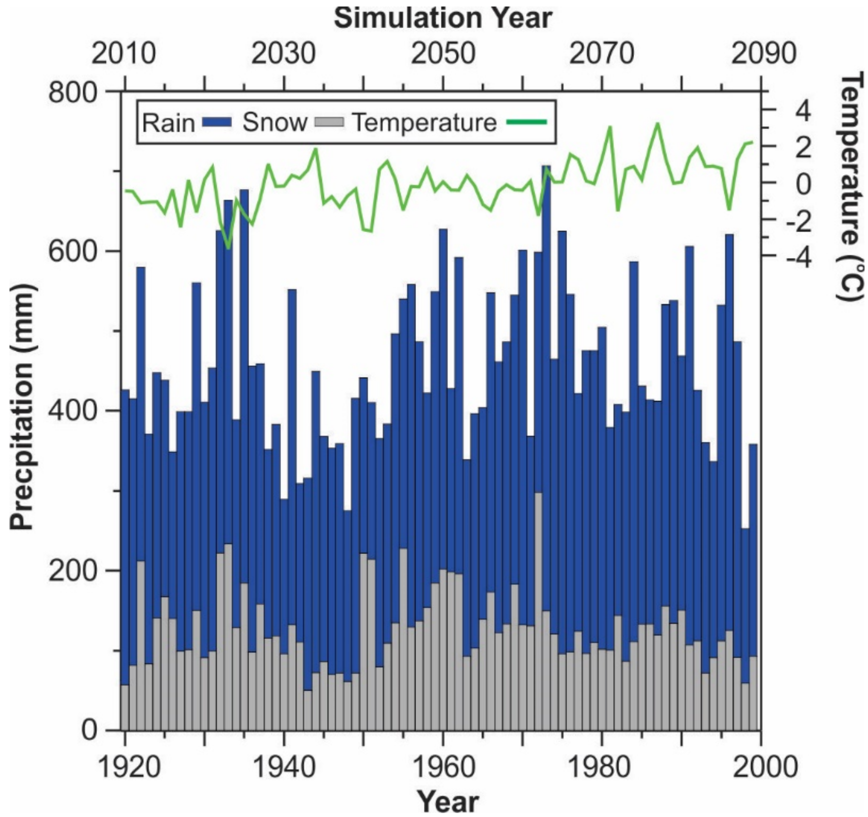


Figure 5-4. Annual precipitation and average annual temperature from Fort McMurray (Environment Canada, 2016) used to generate inputs for the climate change simulations. The year the data were obtained is indicated by the bottom horizontal axis, whereas the corresponding simulation year is indicated by the top horizontal axis.

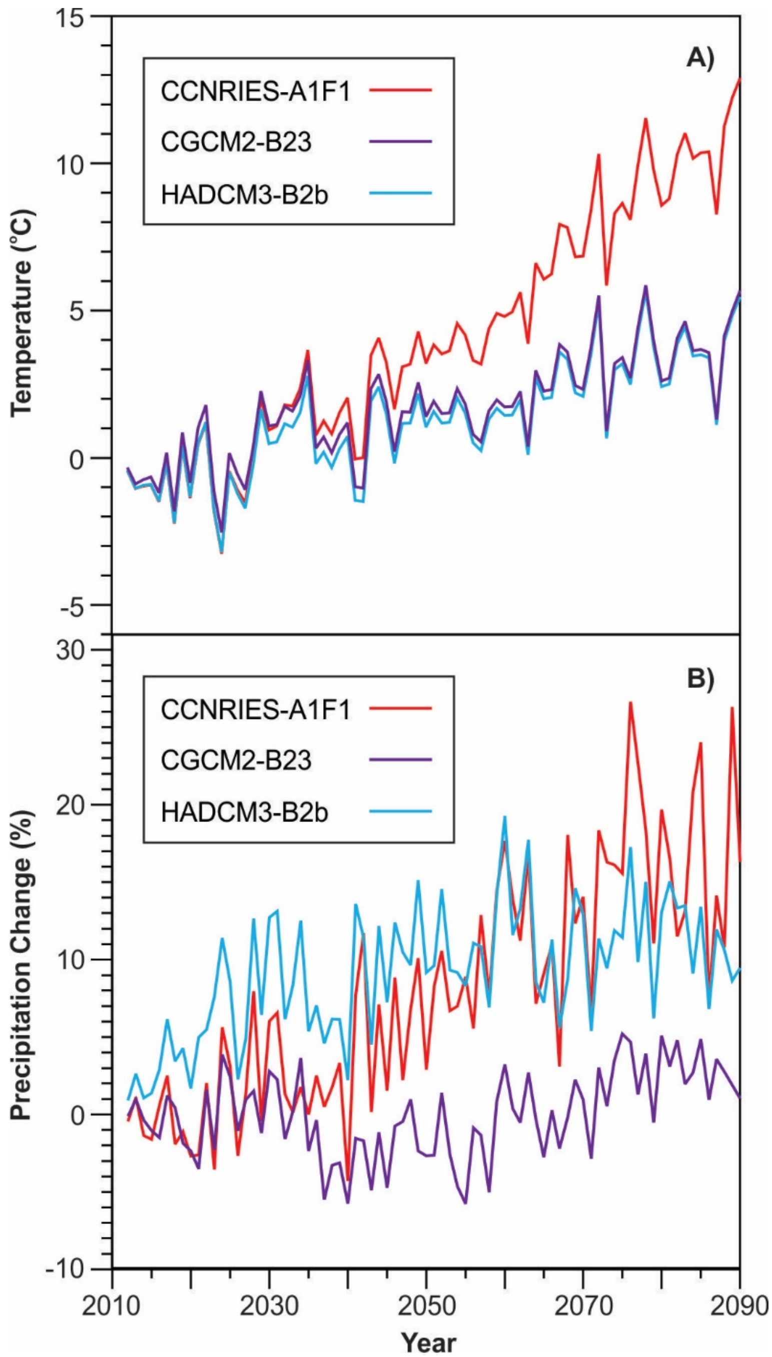


Figure 5-5. Calculated (a) average annual temperature and (b) change in annual precipitation relative to the base case climate for selected climate change scenarios. Scenarios shown approximately bracket the range in predicted climate changes for the study area.

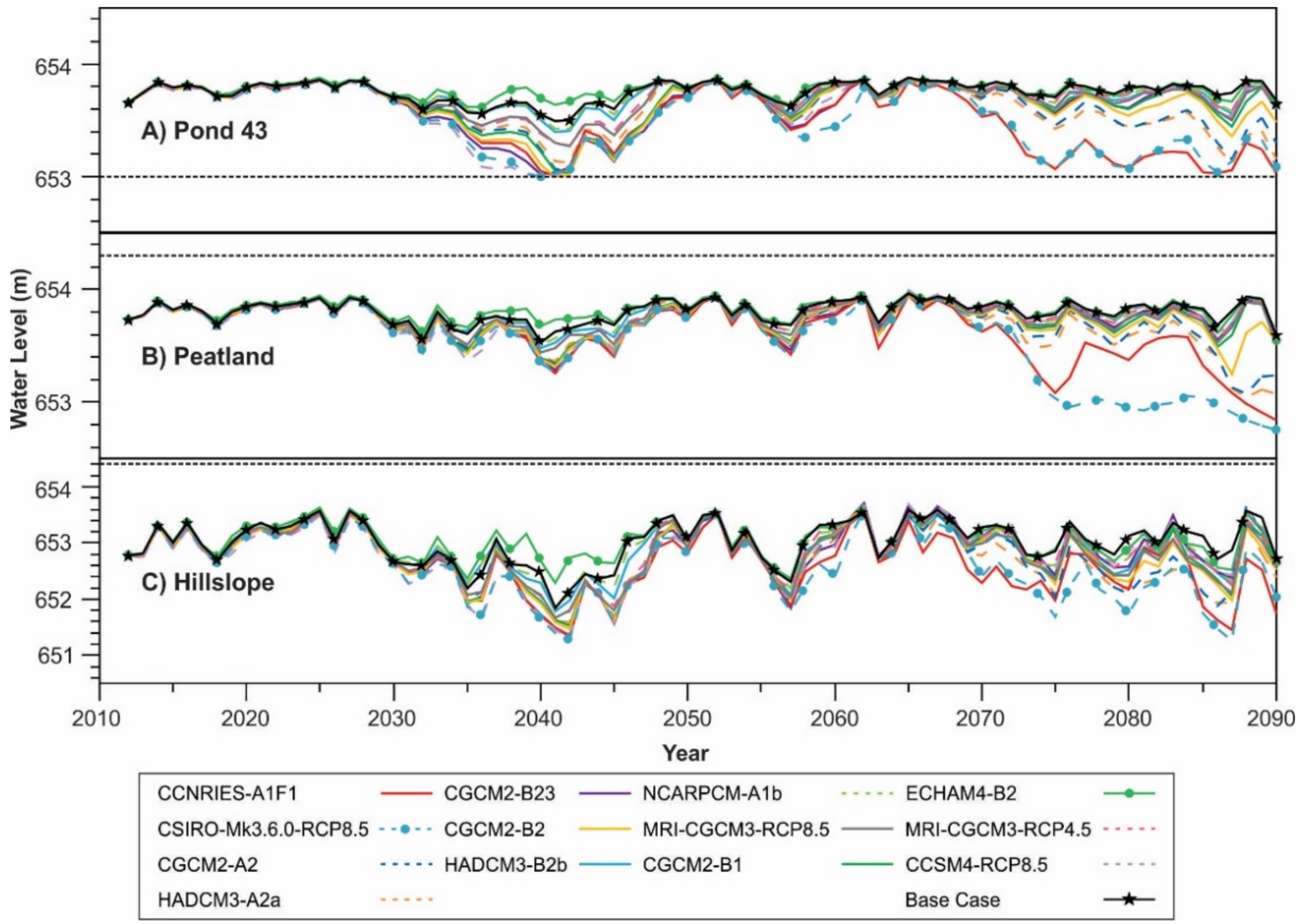


Figure 5-6. Predicted average annual water levels within (a) Pond 43, (b) the peatlands, and (c) the toe of the hillslope for the simulated climate change and base case scenarios. Dashed line indicates pond bottom, peat surface, or ground surface elevation.

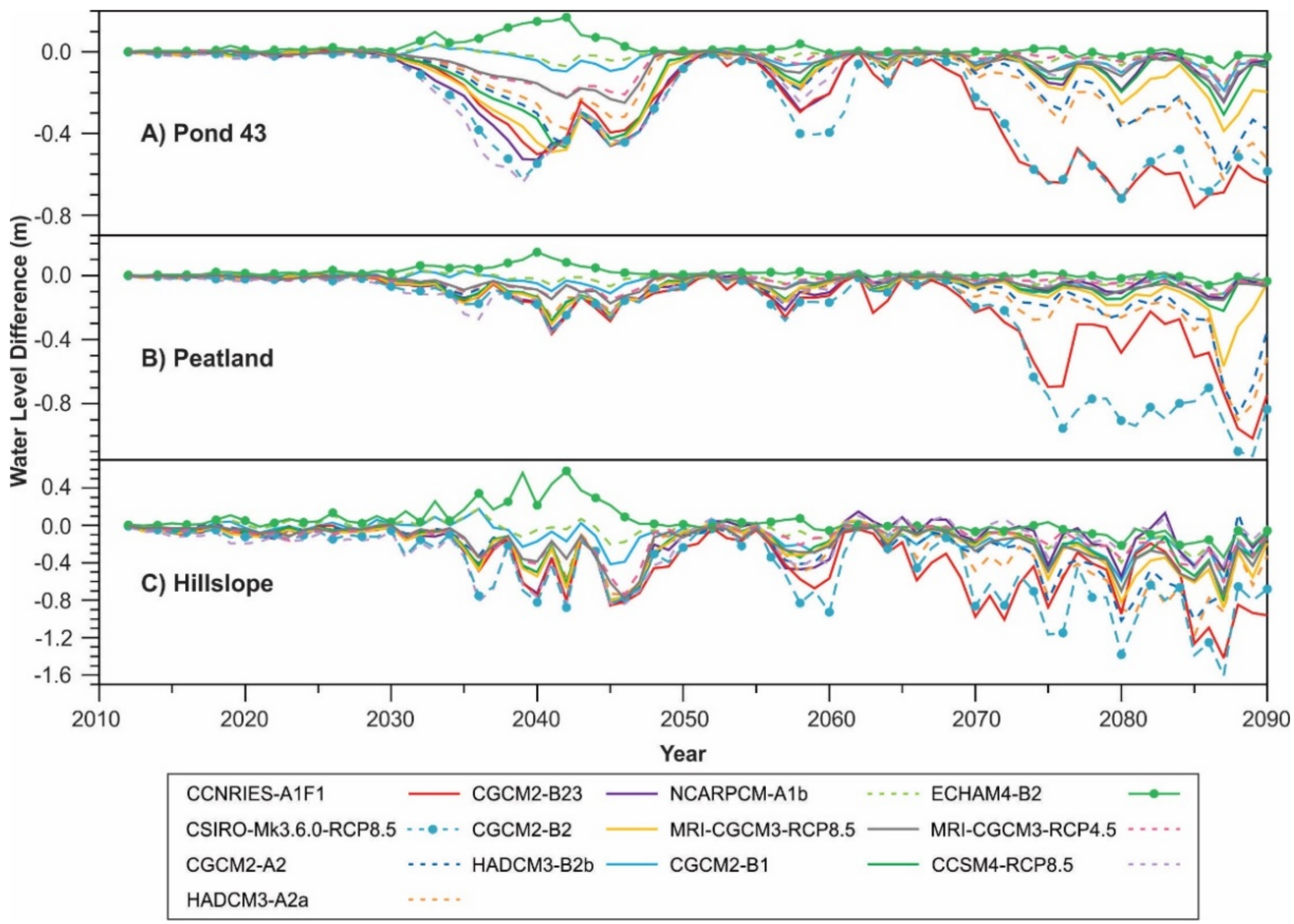


Figure 5-7. Predicted difference in average annual water levels relative to the base case simulation within (a) Pond 43, (b) the peatlands, and (c) the toe of the hillslope for the simulated climate change scenarios. Positive values indicate an increase in water level relative to the base case simulation, whereas negative values indicate a decrease.

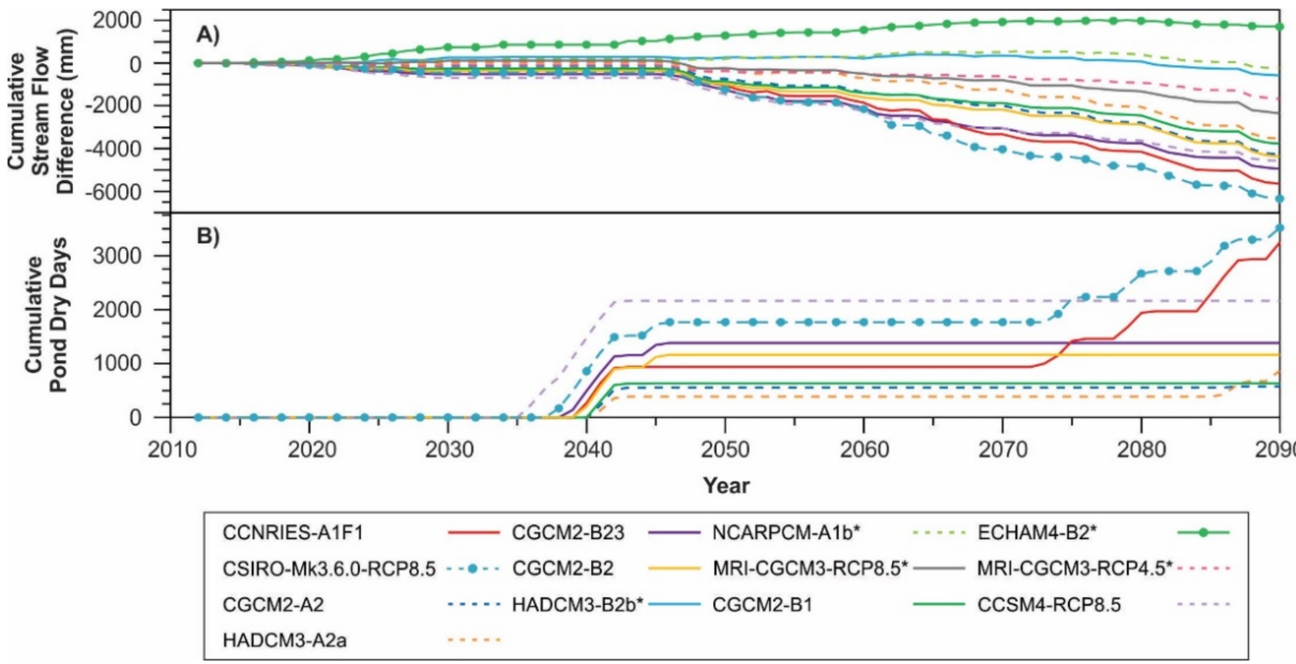


Figure 5-8. Predicted cumulative difference in (a) annual stream flows and (b) pond dry days for the simulated climate change scenarios. The base case and five scenarios indicated with an * had zero pond dry days and are not plotted. Positive values indicate an increase in stream flows relative to the base case simulation, while negative values indicate a decrease.

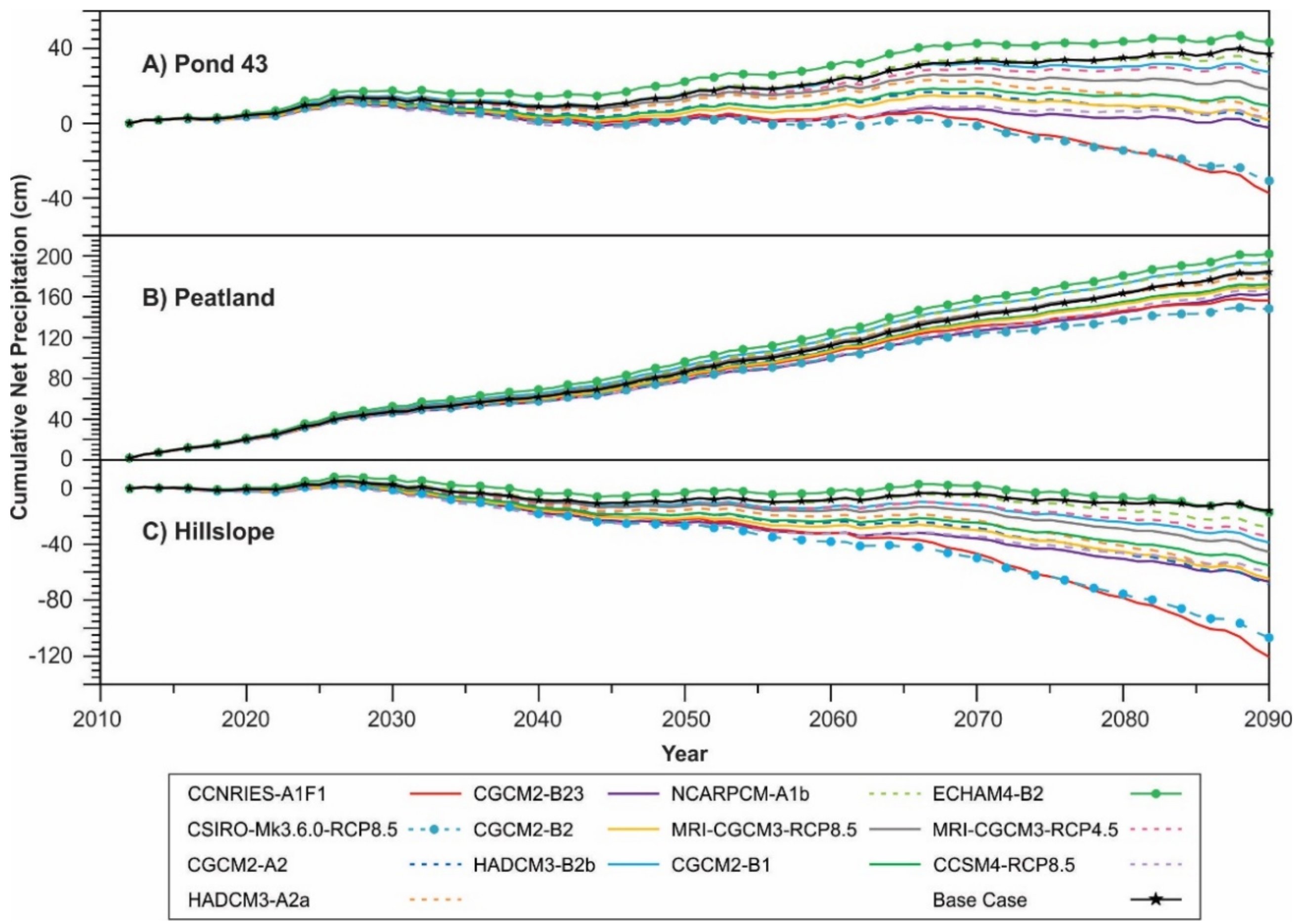


Figure 5-9. Cumulative net precipitation over the simulation period for (a Pond 43, (b) the peatlands, and (c) the aspen hillslope for each climate case. A positive slope indicates excess water, whereas a negative slope indicates water deficit conditions.

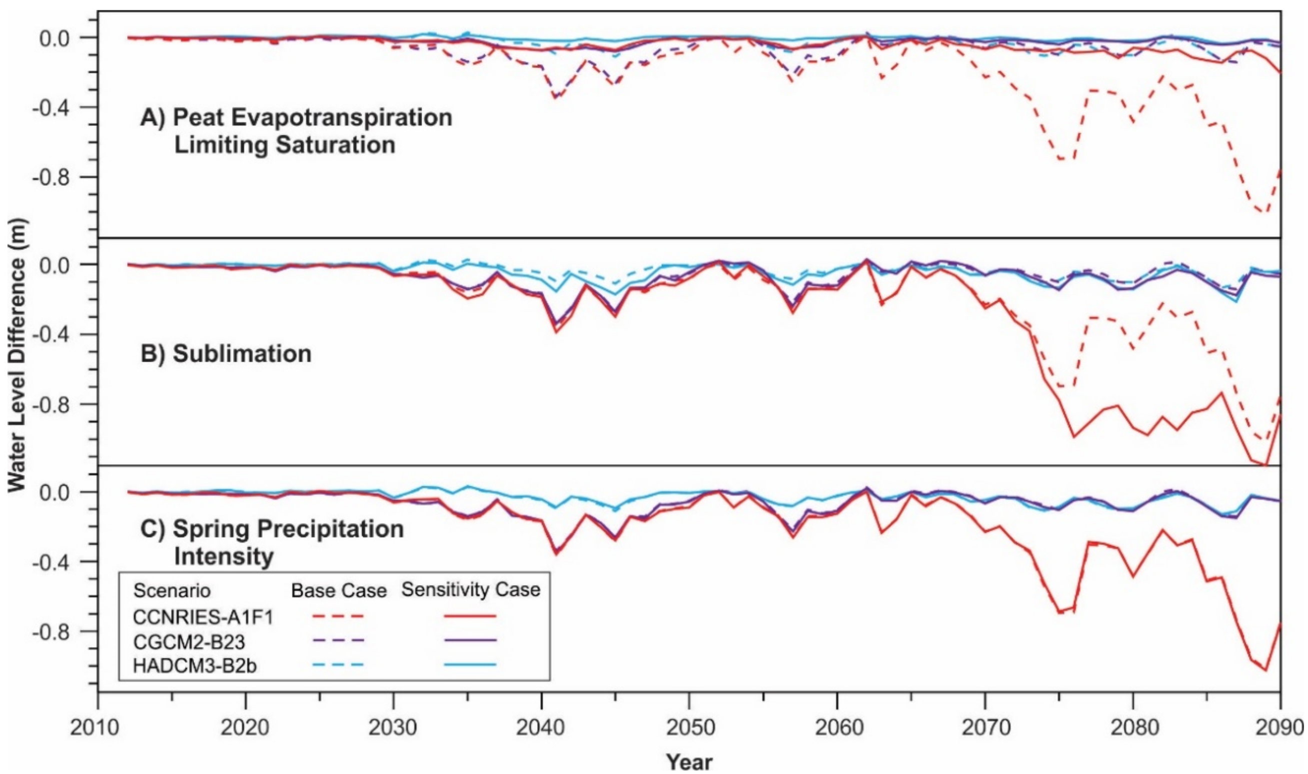


Figure 5-10. Predicted difference in average annual peatland water levels relative to the base case simulation (dashed lines) for sensitivity simulations (solid lines) with increase (a) limiting saturation for peat evapotranspiration, (b) snowpack sublimation, and (c) spring precipitation intensity. Positive values indicate an increase in water level relative to the base case simulation, whereas negative values indicate a decrease.

CHAPTER 6

Summary and Conclusions

The objective of this thesis was to advance the understanding of hydrologic interactions occurring within catchments situated in fine-grained glacial moraine settings within the Boreal Plains of north-Central Alberta. As introduced in Chapter 1, these catchments are composed of a mosaic of landscape units consisting of peatlands, ponds, and predominately aspen upland forests (NRC, 2006; Devito *et al.*, 2017) that persist within a sub-humid climate (Marshall *et al.*, 1999) with low frequency of large storms and synchronized peaks in precipitation and evapotranspiration during the growing season. Within this climatic framework, characterization of the processes governing the movement of water within and between these landscape units is paramount for proper management of existing ecosystems and restoration of those which have already been disturbed. Using a combination of field data and numerical models developed for two fine-grained glacial moraine catchments in the Utikuma Region Study Area (URSA), Chapters 2 to 5 explored the hydrological functioning and linkages between landscape units, evaluated the role of the seasonally frozen peatlands, and assessed potential ranges in hydrologic response caused by anthropogenic development (e.g., timber harvesting) and climate change. The following sections summarize key findings related to these topics. The final section provides several recommended avenues of future research that could be pursued to further the understanding of these hydrological systems.

6.1. Hydrologic Functioning and Linkages of Boreal Plains Glacial Moraine Landscapes

Within fine-grained glacial moraine settings, peatland, pond, and aspen upland forest landscape units play important and contrasting roles in maintaining a delicate hydrologic balance within the sub-humid climate of the Boreal Plains. Field data and numerical simulations described throughout this thesis support the concept that the peatlands play an vital role in the hydrologic functioning of these landscapes (Devito *et al.*, 2005a; Alberta Environment, 2008; Johnson and Miyanishi, 2008; Devito *et al.*, 2012; Ireson *et al.*, 2015), supplying water to ponds, streams, and adjacent hillslopes following snowmelt and rain events, and conserving water within the landscape during periods of drought (Gracz *et al.*, 2015; Devito *et al.*, 2017). Conversely, study results confirm that ponds and aspen upland forests are dominated by high rates of evapotranspiration

(Petrone *et al.*, 2007; Zha *et al.*, 2010; Brown *et al.*, 2014), and represent net water sinks within the landscape (Devito *et al.*, 2012).

Interactions between the ponds and peatlands are dynamic and driven by precipitation and evapotranspiration, thus reflecting recent weather trends with frequent reversals in flow direction (Ferone and Devito, 2004). The magnitude of the pond-peatland water exchange rate varies over temporal scales ranging from days to months, and is dependent on peatland water table depth and depth-dependent peatland hydraulic conductivity (Waddington *et al.*, 2015), on the degree of peatland hydraulic conductivity reduction due to ice (Woo and Marsh, 2005), and on whether the hydrologic connection is above or below the surface of frozen peat substrate (Hayashi, 2013; Ireson *et al.*, 2013).

In contrast to pond-peatland interactions, the hydrologic linkage between the peatlands and aspen forested uplands is limited. Within the uplands, the high evapotranspiration combined with the deep glacial soils within the region (Vogwill, 1978) result in deep upland water tables that do not follow topography and often decline away from adjacent ponds and peatlands (Ferone and Devito, 2004). Consequently, following snowmelt and rain events the upland hydrologic response is dominated by fluctuations in storage, with little potential for generation of overland flow (Devito *et al.*, 2005b; Redding and Devito, 2008). Furthermore, subsurface peatland-upland water exchange is limited by the low hydraulic conductivity of the glacial till and deeper peat, as well as seasonal freezing of the peat substrate.

6.2. Seasonal Peatland Freezing

Seasonal freezing of peatlands is an important process affecting water availability within Boreal Plains catchments, where the relatively dry sub-humid climate and synchronized peaks in precipitation and evapotranspiration occur during the growing season and magnify the importance of springtime replenishment of ponds and the subsurface. Numerical simulations conducted in Chapter 3 indicate that seasonal freezing maintains higher water table conditions within the peatlands by restricting infiltration of snowmelt and spring precipitation, thereby supporting higher rates of spring evapotranspiration and discharge at the peat surface as surface ponding and overland flow, with implications for flood generation (Woo and Marsh, 2005), nutrient loading rates (Kane *et al.*, 2010; Plach *et al.*, 2016), and peatland fire susceptibility (Waddington *et al.*, 2015). Simulation results also indicate subsurface water exchange between peatlands and ponds is restricted due to the lower hydraulic conductivity of the frozen peat.

These results highlight the dichotomous nature of peatland freezing with respect to pond-peatland hydrologic connectivity; freezing both promotes greater connection above the ice through generation of overland flow and restricts water exchanged within and below the ice. The relative importance of surface versus subsurface pond-peatland connectivity is difficult to discern and is likely to vary based on fluctuations in weather at sub-annual to decadal timescales. However, based on existing climate normals for the region, where winter precipitation typically comprises less than 25% of annual precipitation (Marshall *et al.*, 1999), overland flow generation is unlikely to be a significant contributor to pond water budgets in most years. Furthermore, ponded spring meltwaters and precipitation will be exposed to greater surface water evaporation rates (Petrone *et al.*, 2007), and may infiltrate the peat at breaks in the ice and as the ice recedes, thereby reducing the quantity available for generation of overland flow. In comparison, the frozen peat may restrict subsurface pond-peatland exchange throughout the majority of the year, potentially exerting a much greater influence on ponds levels. As a result, ponds that are dependent on contributions from neighboring peatlands may be prone to drying out during ice-rich years. Conversely, in different hydrologic settings, for example where pond flow-through conditions (Townley and Trefry, 2000) are prevalent, seasonally reduced subsurface pond-peatland hydrologic connectivity may be important for restricting pond seepage losses and maintaining pond water levels (Smerdon *et al.*, 2007).

6.3. Responses to Development and Climate Change

Landscape disturbance associated with anthropogenic development (e.g., oil and gas resources and timber harvesting) and climate change is increasingly impacting Boreal Plains ecosystems. Numerical simulations were used to explore the range of hydrological response to aspen harvesting (Chapter 4) and climate change (Chapter 5) across a range of respective historical climate conditions and climate change scenarios. Results of the timber harvesting simulations indicate that aspen harvesting has limited impact on groundwater levels and stream flows within these hydrologic systems because of the sub-humid climate with low-frequency of large storms, large soil-moisture storage capacity of heterogeneous glacial materials, and high evapotranspiration rates of regenerating aspen. However, the magnitude of increases in groundwater levels and stream flows were sensitive to regenerating aspen evapotranspiration rates, which can be enhanced by appropriate harvesting techniques. Results of sensitivity simulations conducted in Chapter 2 indicate that the primary impacts of forestry may be due to placement of roads within flat-lying peatlands, which may disrupt shallow groundwater flows and potentially lead to isolated pond and peatland areas that are highly susceptible to drought (Lieffers and MacDonald, 1990; Smerdon *et al.*, 2009).

Results of the climate change simulations indicate that the peatlands may be resistant to changing hydrological conditions associated with a warming climate. Internal peatland mechanisms that act to regulate the water table depth and maintain peat water content may prevent them from undergoing large decreases in water levels, at least in the first half of this century. However, this conclusion is predicated on the assumption that these internal mechanisms will not be outweighed by external stresses such as fire, encroaching trees or shrubs, and anthropogenic disturbance. Conversely, high evapotranspiration losses from the aspen upland forests result in near-surface soils becoming increasingly drier towards the end of the century. Thus, the aspen may frequently be water-stressed and increasingly susceptible to secondary maladies such as pests and disease. The results also indicate that water levels in the ponds that are ubiquitous to the region will be reduced as the climate warms, with the development of ephemeral conditions becoming increasing frequent in warmer and drier scenarios. Concurrent decreases in stream flow may have a compounding cumulative impact on downstream ecosystems. Differences in the timing and magnitude of responses in ponds and aspen forests relative to adjacent peatlands may have further profound impacts on the evolution of inter-ecosystem interactions but could result in the development of novel landscapes with a configuration of ecosystems not currently observed (Schneider *et al.*, 2015).

6.4. Landscape Reconstruction and Reclamation

Large areas of the Boreal Plains landscape have been disturbed by open-pit mining of oil sands in the Fort McMurray region which will require landscape reconstruction and reclamation on an unprecedented scale over the next 30 to 50 years (Kelln *et al.*, 2008). Regulatory requirements specify that the disturbed land be returned to an “equivalent capability”, with the reestablishment of a vegetative regime similar to what was present prior to disturbance (OSWWG, 2000; Alberta Environment, 2008). The strategy of engineering new ecosystems is not to re-create the landscape as it existed before but to construct a landscape in which the physical environment of geomorphic, hydrological, and biogeochemical processes will be able develop sustainably (Johnson and Miyanishi, 2008). Concepts that have been developed from natural analogs such as the URSA catchments studied as part of the thesis are crucial for design of reclamation plans that incorporate a higher probability of successful implementation.

A key conclusion from Chapter 2 is the concept that Boreal Plains peatlands can be self-sustaining within the sub-humid climate, and do not necessarily require regional groundwater discharge or significant flow contributions from adjacent uplands for long-term maintenance. However, in the context of a reconstructed landscape, this conclusion is predicated on the assumption that the peatlands be underlain by materials of sufficiently low hydraulic conductivity. Sensitivity simulations indicate that water levels within the peatland are in part maintained by the low hydraulic conductivity of the underlying glacial substrate. Furthermore, although natural established peatlands in the region have continued to persist despite experiencing extended drought conditions, prolonged periods of drought may produce conditions that are hostile to newly placed peat that is less well established (Price *et al.*, 2010; McCarter and Price, 2014). Therefore, an external source of water may be needed to maintain wetland vegetation if initial reclamation is followed by drier climate cycles. Traditional reclamation plans have specified the placement of small hills and hummocks within the reclaimed landscape to allow for runoff generation (OSWWG, 2000; Alberta Environment, 2008). However, the inclusion of forested uplands within the reclaimed landscape may prove detrimental to peat development, removing water from the system rather than supplying it to the peatlands as intended.

6.5. Recommendations for Future Research

The Boreal Plains region is expected to be an area of maximum ecological sensitivity in the 21st century and will thus require a thorough understanding of the interaction between hydrology, climate, and biology for successful for climate adaptation and sustainable forest management (Ireson *et al.*, 2013). Feedbacks between the terrestrial carbon cycle, including the large carbon pool stored within northern peatlands, and climate remains one of the largest uncertainties in future climate projections (Dorrepaal *et al.*, 2009; Waddington *et al.*, 2015). Numerical simulations were used to predict hydrologic responses to aspen harvesting (Chapter 4) and climate change (Chapter 5); however, secondary factors (e.g., enhanced virulence of pests and disease, increased fire frequency, vegetation succession) that may act to amplify and/or moderate these responses were excluded. Possible influences of these secondary factors on model predictions were hypothesized in Chapters 4 and 5; however, further research into the relative magnitude of their influence and to how they may be represented within existing and future simulation tools is warranted. Consideration of the interaction and feedback among climatic, hydrologic, geochemical, and biologic processes (Hayashi, 2013) with a focus on quantifying their relative magnitude (Chapter 5) is the next step to identifying probable scenario trajectories.

Seasonal peatland freezing is an important process affecting the hydrology of Boreal Plains catchments. Analyses conducted in Chapter 3 represented peatland freezing using a simplified approach that did not allow feedback between the thermal state of the peat and the simulated groundwater flow system. While suitable for assessing the influence of peatland ice on catchment hydrology during the study period where the depth of ice was constrained, this approach precluded the predictive use of the model to assess the future evolution of the hydrologic system to changing conditions (e.g., anthropogenic development and climate change). As noted in Chapter 3, future research could benefit from use of an integrated groundwater-surface water model that incorporates fully coupled flow and heat transport which could allow representation of processes such as rain-on-snow/ice events and mid-winter snowmelt that may play important roles on the thermal state of the subsurface, peatland ice persistence, overland flow generation, and water levels. Future studies could also benefit from inclusion of a robust representation of snowpack dynamics, which may have both a cooling (e.g., high albedo, high emissivity) or warming (e.g., high absorptivity of long waver radiation) influence on temperature at the land surface (Zhang, 2005; Ireson *et al.*, 2013).

Numerical simulations detailed in Chapters 2 to 5 were completed using several two-dimensional models. Expansion of the model domains into catchment-scale three-dimensional models could prove insightful to the overall hydrological functioning of these hydrologic systems. However, computational time remains a challenge, and may be expected to increase as further processes are incorporated into the simulation tools.

Lastly, studies detailed throughout this thesis benefitted from a large, multiyear, hydrologic and geologic dataset that allowed parameterization of material properties and boundary conditions within several numerical models. Nevertheless, data gaps were identified that required simplifying assumptions. Future studies conducted within the study catchments and other Boreal Plains' locations could benefit from direct measurement of snow interception and sublimation, increased measurement frequency of stream flow, and regular measurement of peatland ice thickness. However, it is recognized that the logistics of collecting such detailed datasets over the many years necessary is a daunting task.

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Appendix A

Estimation of Peatland Ice Depth

This appendix documents the methodology used to estimate the depth and duration of peatland freezing as simulated in Chapter 3. The methodology consists of three sequential components, which are used to derive loosely coupled daily estimates of (1) snowpack depth, (2) ground surface temperature, and (3) peatland ice distribution. Snowpack depth was computed using the degree-day method using daily measurements of precipitation and average air temperature from the study area. Ground surface temperature and peatland ice distribution were computed using numerical simulations conducted with the finite-element model SUTRA (Voss and Provost, 2010). SUTRA is a variably-saturated groundwater flow and heat transport code that has recently been enhanced to include porewater freezing and thawing (McKenzie *et al.*, 2007b) in both the saturated and unsaturated zones (Kurylyk *et al.*, 2014). This enhanced version of the code has been utilized to study subsurface freeze-thaw dynamics in both permafrost environments (Ge *et al.*, 2011; McKenzie and Voss, 2013; Wellman *et al.*, 2013; Briggs *et al.*, 2014, Kurylyk *et al.*, 2016) and regions with seasonal freezing (Kurylyk *et al.*, 2014).

Field data collected within the study watershed to support this analysis included measurements of precipitation, air-temperature, and peatland ice depth, along with groundwater and surface water levels (Chapters 2, 3, and 4). Precipitation data were collected using 1 to 2 automated tipping bucket rain gauges and were checked using manual measurements from bulk rain gauges distributed across the site. Snow surveys were used to measure the snowpack each winter and spring, with snow water equivalent (SWE) determined from composite samples obtained each survey. Air temperature was measured on a 30-minute interval and recorded using automated loggers for the duration of the study period. Depth to ice measurements were obtained from 2003 to 2009 by pushing a metal rod of known length through the peat to the ice surface. Using this method, the maximum depth of ice that could be detected was about 0.7 m. Weekly to monthly measurements were collected most years from early spring to late summer at peatland locations distributed across the study area.

A.1 Snowpack Depth

The depth of the snowpack was estimated from measured daily precipitation and snowmelt computed using the degree-day method. The degree-day method consists of a simple, temperature-based approach to estimate snowmelt that has been used in various forms for nearly a century (Rango and Martinec, 1995), with daily snowmelt (*m*) calculated as:

$$m = M(T_a - T_t)$$

where *M* is the melt factor that accounts for different factors affecting the melt rate, and *T* is temperature with subscripts *a* and *t* denoting measured air temperature and the threshold temperature above which snowmelt occurs, respectively. The melt factor *M* was calculated as outlined in Carrera-Hernandez *et al.* (2011), including constants that were derived through calibration to data from Fort McMurray, Alberta. The snowpack calculations assumed that all precipitation falling during the winter period is incorporated into the snowpack and that the influence of sublimation on snowpack depth is negligible. Further assumptions incorporated into the calculations included spatial and temporal homogeneity in snowpack properties and depth, and constant daily air temperature.

The estimated depth of the snowpack was compared to average snow depth measurements from snow surveys distributed across the Pond 43 catchment. In general, the observed snowpack depth was adequately represented (Figure A-1A), with overall trends in snow accumulation and melt adequately captured. However, peak snowpack depths were underestimated in some years.

A.2 Ground Surface Temperature

One-dimensional (1-D) thermal conduction simulations were performed using SUTRA to estimate the average daily ground surface temperature beneath the snowpack within the peatlands, pond, and hillslope, with a separate model developed for each location. Each model extended 20 m below ground surface, with finite-elements varying from 0.1 m thick in the upper 3 m to 1 m thick at the model base. Within each model, the mesh was extended 1 m above ground surface using 0.05 m thick finite-elements to allow representation of the snowpack. The distribution of materials assigned within the 1-D models is summarized in Table A-1. Thermal properties assigned to each material were assumed to be uniform and are summarized in Table A-2. Boundary conditions consisted of air temperature specified at ground/pond surface or the surface of the snowpack during snow-covered periods and constant temperature of 4 °C at the model base. During the snow-covered periods, nodes assigned the air temperature boundary varied daily and were raised and lowered with the height of the snowpack based on the estimated depth of the snowpack. To represent conduction alone, hydrostatic conditions were assumed with no flow throughout the model domain.

Simulated ground surface temperatures during snow-covered winter periods were highly dependent on snowpack depth (Figure A-1A). During low snowfall years (e.g., 2001 and 2010) the snowpack was predicted to provide little insulation, which resulted in predicted ground surface temperature reaching lows of about -20 °C. Conversely, in years with higher snowfall (e.g., 2007 and 2008), the insulating effect of the snowpack is apparent, as surface temperatures are predicted to remain near -5 °C despite air temperatures regularly dropping below -20 °C.

A.3 Peatland Ice Distribution

Peatland ice distribution was estimated using a two-dimensional (2-D) SUTRA model of groundwater flow with thermal transport including water-ice phase change in the peatlands. The model domain and material distribution were the same as utilized in simulations described in Chapter 2, 3, and 5. The finite-element mesh was discretized using uniform 1 m nodal spacing horizontally. Vertically, 0.05 m elements were specified to a depth of 4 m to encompass the peatlands where freezing was simulated to occur. Below this depth, finite-element layers were gradually increased to a maximum thickness of 2.5 m.

Specified permeability and porosity values (Table A-2) were consistent with simulations discussed in Chapter 3. Properties governing heat transport (i.e., thermal conductivity, heat capacity) were specified to be uniform within each hydrogeologic unit (Table A-2). Thermal dispersion was assumed to be negligible (i.e., thermal dispersivity was set to 0 m) consistent with Kurylyk *et al.* (2016). Simplified linear constitutive relationships for pressure-saturation, saturation-relative permeability, and temperature-ice saturation were defined based on the literature (Carsel and Parrish, 1988; Silins and Rothwell, 1998; Price *et al*; 2010; Smerdon and Mendoza, 2010).

SUTRA does not include a method of simulating saturation-constrained evapotranspiration; therefore, a flux-based set of surficial boundary conditions (i.e., groundwater recharge and evapotranspiration) could not be used to adequately represent the seasonality in groundwater levels. Consequently, groundwater flow boundary conditions at the surface of the model consisted of specified daily surface pressures derived from a combination of measured groundwater levels at wells and piezometers located along the 2-D model, measured stage of Pond 43, and calibrated simulation results from Chapter 2. Remaining flow boundary conditions consisted of constant pressure at the base and no flow conditions along the lateral edges consistent with Chapters 2 and 5.

Heat transport boundary conditions consisted of daily temperature applied to the model surface, constant temperature of 4 °C at the base based on measured groundwater temperatures, and no heat flow across the lateral edges. At the model surface, the daily surface temperature was assumed equal to air temperature when no snow cover was present. During snow-covered periods, the surface temperature was specified using the results of the 1-D SUTRA simulations (Section A.2). Initial conditions were generated by transiently spinning up the model and running it to steady state.

Simulation results were evaluated by comparing the predicted peatland ice depth to available measurements. The results indicate that observed peatland ice depths were adequately represented (Figure A-1B). The onset of peatland freezing was generally predicted to occur in late October, with the frozen period extending to late June to early September. The extent of the frozen period was predicted to be greatest during low snowpack years due to colder temperatures at the base of the snowpack, reaching almost three continuous years at some locations from the fall of 2002 to the spring of 2005. Similarly, the predicted depth of frozen peat was greatest during low snowpack years, ranging from 0.8 m in 2009 to more than 1.8 m in 2005.

Table A-1. Material distribution within 1-D SUTRA models.

<table>
<tr><th>Depth Range (m)</th><th>Peatland</th><th>Pond</th><th>Hillslope</th></tr>
<tr><td>-1 to 0</td><td>Snow</td><td>Snow</td><td>Snow</td></tr>
<tr><td>0 to 1</td><td rowspan="2">Peat</td><td>Water</td><td rowspan="4">Glacial Till</td></tr>
<tr><td>1 to 2</td><td rowspan="2">Gyttja</td></tr>
<tr><td>2 to 3</td><td rowspan="2">Glacial Till</td></tr>
<tr><td>3 to 20</td><td>Glacial Till</td></tr>
</table>

Table A-2. Summary of parameters within the SUTRA simulations.

Material	Depth Range (m)	Permeability (m^2)		Porosity (-)	Thermal Conductivity (W/m°C)	Specific Heat (J/kg°C)
		Horizontal	Vertical			
Water	-	-	-	-	0.6	4182
Ice	-	-	-	-	2.14	2108
Snow [a]		1×10^{-40}	1×10^{-40}	0.005	0.15	2090
Peat	0.0 - 0.1	3×10^{-10}	3×10^{-11}	0.90	0.3	1920
	0.1 - 0.3	3×10^{-11}	3×10^{-12}	0.82	0.3	1920
	0.3 - 0.5	8×10^{-12}	8×10^{-13}	0.72	0.3	1920
	0.5 - 1.0	4×10^{-12}	4×10^{-13}	0.60	0.3	1920
	1.0 - 1.5	2×10^{-13}	2×10^{-14}	0.50	0.3	1920
	1.5 - 2.0	3×10^{-15}	3×10^{-16}	0.45	0.3	1920
	2.0 - Base	1×10^{-15}	1×10^{-16}	0.40	0.3	1920
Gyttja	0.0 - 1.0	1×10^{-13}	1×10^{-14}	0.45	0.3	1920
	1.0 - 2.0	3×10^{-15}	3×10^{-16}	0.30	0.3	1920
	> 2.0	5×10^{-16}	5×10^{-17}	0.22	0.3	1920
Glacial Till	Upper	1×10^{-12}	1×10^{-14}	0.20	2.9	920
	Mid	5×10^{-14}	5×10^{-16}	0.20	2.9	920
	Lower	1×10^{-15}	1×10^{-17}	0.20	2.9	920
Forest Floor	0.0 - 0.1	1×10^{-11}	1×10^{-11}	0.80	0.3	1920

Notes:

[a] Snow represented in 1-D conduction simulations.

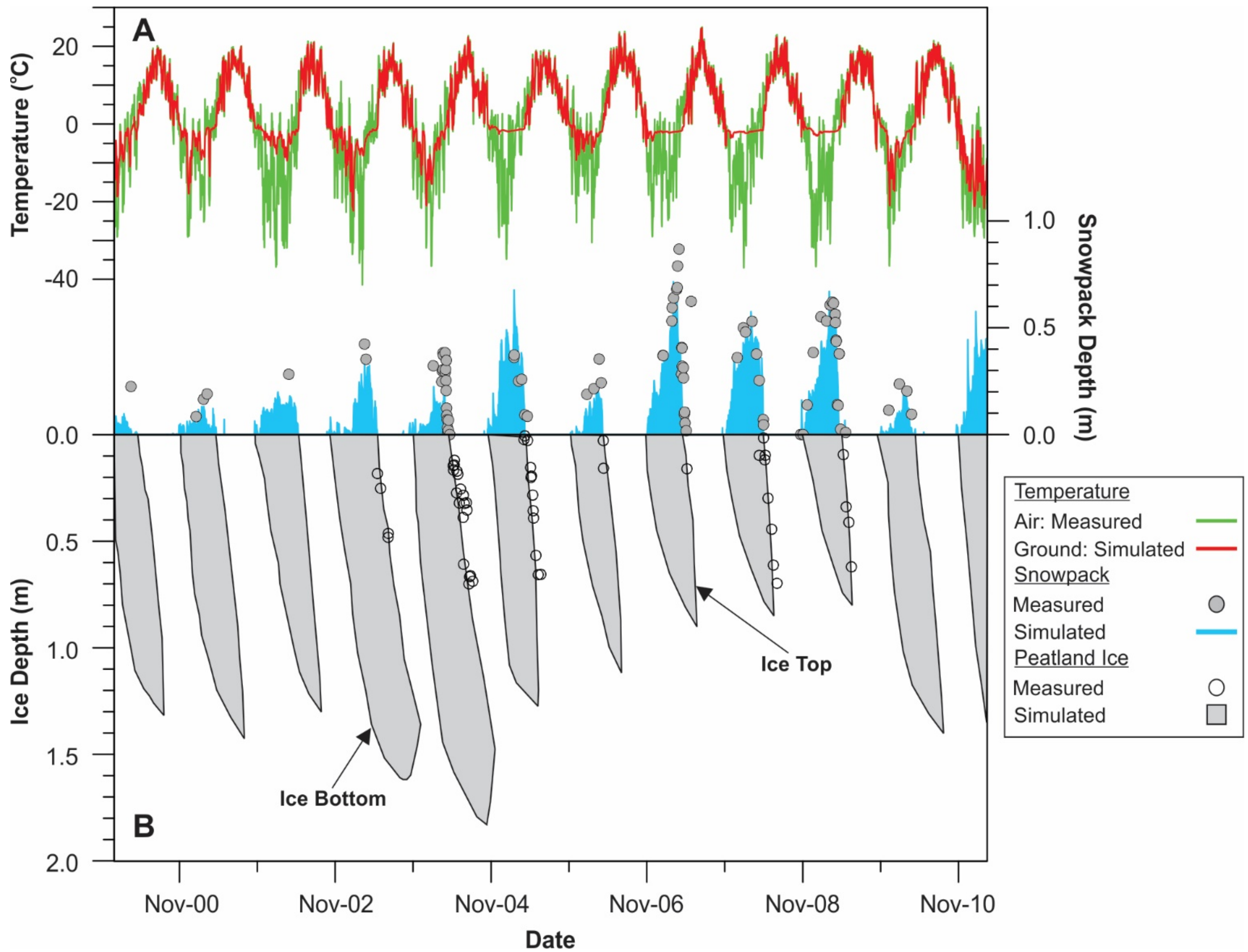

Figure A-1. Comparison of measured and simulated A) snowpack depth along with estimated peatland surface temperature relative to measured air temperature and B) average peatland ice depth.