University of Alberta

The influence of climate on water cycling and lake-groundwater interaction in an outwash landscape on the Boreal Plains of Canada

by

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Abstract

The influence of climate and land use on water cycling and lake-groundwater interaction were determined for a glacial outwash landscape on the Boreal Plains of Canada. Increased landscape disturbance (from logging activity, road building, and gravel pit construction) and a need to assess potential climate change effects, initiated study of hydrological processes at a lake in the Utikuma Research Study Area (URSA). The URSA has a sub-humid climate ($P \leq PET$), which controls many hydrological processes. Lake-groundwater interaction was quantified, climatic influence on groundwater recharge was investigated, and the influence of landscape disturbance on lake source water was determined.

On hummocky, poorly drained, outwash landscapes, surface water and wetlands are maintained by groundwater flow systems. Hydrometric measurements and stable isotopic analyses of waters collected during a 3-year lake budget study indicate that evaporation was the dominant hydrologic flux in summer months. The onset and duration of evaporation from the lake was a major controlling factor on groundwater exchange, to such a degree that the shallow lakes function as evaporation windows in this landscape, creating groundwater capture zones. Transient hydrologic responses of the flow-through lake and three-dimensional groundwater flow were simulated to study controls on lake-groundwater interaction. Replication of the flow regime required both an anisotropy ratio of 10:1 for the outwash deposits and the inclusion of storage effects within riparian peatlands. These peatlands govern lake-groundwater interaction and maintain surface water on permeable northern landscapes.

The control of spatially and temporally variable groundwater recharge and evapotranspiration processes on groundwater recharge was investigated through development of one- and two-dimensional models of variably-saturated flow. Groundwater recharge and upflux from the water table were found to depend on climate history and water table depth. In summer months, when transpiration and canopy interception were considered explicitly, groundwater recharge was negligible, and groundwater was predicted to be drawn vertically upward due to evapotranspiration. Simulations of isotopic transport identified that localized landscape disturbance has started to alter the water source to the study lake, and that climate and groundwater flow transience must be considered when applying isotope-mass balance approaches for quantifying hydrological processes on the Boreal Plains.

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Chapter 1

Introduction

1.1 Western Boreal Forest

The Boreal Forest is the largest land-based ecosystem in the world, and covers approximately 35% of Canada (Canadian Forest Service, 2006). This forested landscape has abundant wetlands, which are distributed across physiographic and climatic regions that vary greatly across the country (National Wetlands Working Group, 1988). The Boreal Plain ecozone of the Western Boreal Forest extends across three provinces in west-central Canada (Natural Resources Canada, 2006), and is experiencing increased anthropogenic impacts from development in the natural resource and energy industries (Alberta Environmental Protection, 1998; Canadian Forest Service, 2006). Natural variability in water cycling processes, and equilibrium in the aquatic and forested ecosystems on the Boreal Plains rely on a delicate balance between the atmosphere and the hydrosphere (Devito et al., 2005b; Smerdon et al., 2005). Effective land management and prediction of the impact of anthropogenic disturbances on ecosystems within the Boreal Plains require an understanding of present and historic hydrologic processes unique to this ecozone (Devito et al., 2005a).

Climate conditions and the geological environment of the Boreal Plains differ from other sections of Canada's Boreal Forest (i.e., shield, cordillera, tiaga, and tundra), which creates a unique hydrologic landscape (Winter, 2000) that must be considered in hydrologic analyses (Devito et al., 2005a). The Boreal Plains are characterized by a sub-humid climate (i.e., precipitation $\leq$ potential evapotranspiration; Ecoregions Working Group, 1989) with an average air temperature close to 0°C. The sub-humid climate controls the annual differences between precipitation (P) and evapotranspiration (ET), and the seasonal synchronicity of P and ET (Devito et al., 2005b), which are the dominant driving forces in the hydrologic cycle. The landscape is comprised of glacial sediments with low topographic relief that are 80 to 240 m thick (Klassen, 1989; Pawlowicz and Fenton, 2002), overlying shale bedrock of Upper Cretaceous age (Hamilton et al., 1999). The glaciated landforms (Figure 1.1) vary from predominant compositions of fine-textured sediments (e.g., moraines of clay and silt) to coarse-textured sediments (e.g., outwash sand and gravel), with heterogeneity due to the presence of fine-textured lenses within the outwash, and coarse-textured lenses within the moraine. The influence of climatic and geologic characteristics on hydrological processes in the Boreal Plains is the central theme of ongoing research projects at the Utikuma Research Study Area (URSA), in northern Alberta, Canada (Figure 1.1). Studies on lacustrine plain and moraine (Devito et al., 2005a; Ferone and Devito, 2004), and outwash landscapes (Gibbons, 2005; Smerdon et al., 2005) have revealed a complex relationship between climate and the terrestrial hydrologic environment.

Natural resources are ubiquitous in the Western Boreal Forest. Situated within a broad band around much of the Earth's northern hemisphere, a cool climate promotes growth of various tree species (spruce, pine, aspen and poplar), which supply Canada's forest industry with timber and fibre (Natural Resources Canada, 2006). Co-existing with the forested ecosystem, abundant pond-wetland complexes form one of the largest breeding grounds for waterbirds in north America, and sustain migratory birds annually (Ducks Unlimited, 2006). Underlying this diverse terrestrial and aquatic ecosystem, vast deposits of mineral aggregate and hydrocarbons exist, which are continually explored for and produced by the energy industry. These resources (forestry, waterfowl, and energy) are exploited to supply numerous products to global markets, and are linked by the hydrologic cycle, which may be disturbed by increased anthropogenic activity. Thus, accurately quantifying and simulating the dominant hydrological processes in this region is paramount to predicting future hydrologic and ecosystem responses.

1.2 Interaction of Climate, Surface water, and Groundwater

Effective management of Earth's water resources requires an understanding of how moisture and energy are transferred between the atmosphere and the hydrosphere. In all landscapes, these natural cycles are controlled by climate, vegetation, surficial geology, and the interaction of surface water and groundwater. Research has shown that surface-water/groundwater interaction is dynamic (Sophocleous, 2002; Winter, 1999), and that surface water bodies are both sources and sinks for groundwater. In the following studies, I sought to advance the understanding of water cycling processes on coarse-textured landscapes within the Boreal Plains and investigate representing key hydrological processes in simulation models of water flow.

The interaction of surface-water and groundwater on coarse-textured glacial deposits creates very dynamic hydrologic flow systems, which have been studied extensively in the USA (e.g., Cherkauer and Zager, 1989; Kenoyer and Anderson, 1989; Krabbenhoft et al., 1990; LaBaugh et al., 1997; Pint et al., 2003). Observed manifestations of surface-water/groundwater interaction include reversals of groundwater flow (Anderson and Munter, 1981; Devito et al., 1997), transient water geochemistry (Devito and Hill, 1997; Krabbenhoft and Webster, 1995); and sustainability of groundwater dependent ecosystems (Hayashi and Rosenberry, 2001; Rodriguez-Iturbe, 2000; Rosenberry et al., 2000), at various spatial and temporal scales.

On glaciated terrain, coarse-textured substrate is common and is considered to be a significantly transmissive hydrologic unit in a watershed (Winter, 2001) because of more active groundwater interactions than normally associated with fine-textured substrate (Winter et al., 2003). On the Boreal Plains, extensive coarse-textured landforms occur naturally as isolated segments of outwash from intra- and post-glacial melting (Figure 1.1), and as discrete lenses within moraines. In addition, large-scale, constructed landscapes of coarse-textured sediments are being created following hydrocarbon extraction from the oil sands deposits in northeast Alberta (e.g., Syncrude Canada's southwest sand storage facility; List et al., 1997).

A principal driving force for the hydrologic cycle is regional climate. Understanding the dominant hydrological processes within a given climatic region is paramount to predicting changes caused by anthropogenic influences (Winter et al., 2001). Holistic approaches for integrating climate, surface water and groundwater have recently emerged in hydrology (Devito et al., 2005a; Euliss et al., 2004), which have broadened the scope of hydrologic research programs (e.g., Allen et al., 2004; Cohen et al., 2006; York et al., 2002) that could ultimately guide water and land management practices. The underlying theme of the studies presented in this thesis was to quantify surface water/groundwater interaction and the influence of climate on water cycling for a coarse-textured landscape on the Boreal Plains of Canada.

1.3 Thesis Objectives and Format

For a glacial outwash environment, the fundamental research questions addressed in this thesis are:

1) What are the dominant hydrologic controls on water cycling?

2) What landscape and hydrologic data are required to replicate historic observations and support predictive simulations of future/alternative scenarios?

3) How can dominant processes be described and quantified?

These questions were addressed by studying the hydrology of a shallow lake and groundwater flow system located on a glacial outwash plain at the URSA. The Lake 16 study site (Figure 1.1) contains landscape hydrologic features that are characteristic of coarse-textured areas on the Boreal Plains: ponds, lakes, fens, and peatlands. The study site also contains a gravel pit, which is a typical form of landscape disturbance in glacial outwash areas (Figure 1.2).

This thesis follows the *paper format* style, and has been organized into six chapters, which address the hydrological aspects of climate and water cycling on the Boreal Plains. In these studies, I have elected to "follow the flow of water" at a spatial scale that includes a lake, rather than investigate fundamental physical properties of porous media at a smaller scale (e.g., individual transect or plot research combined with laboratory study). Following the introduction, Chapters 2, 3, 4 and 5 present separate studies that focus on specific objectives, utilizing field and modelling methods:

1) To determine a water budget for an outwash lake (Chapter 2). Fluid fluxes were measured for each component of the water budget independently, and a conceptual model for groundwater flow and surface-water/groundwater interaction was developed for the URSA outwash landscape.

2) To simulate observed lake-groundwater interaction and a portion of the outwash flow system with a numerical flow model (Chapter 3). The conceptual flow model developed in Chapter 2 was tested by simulating/replicating transient field observations. The degree of spatially distributed climatic data and landscape heterogeneity required for replication was determined, in addition to investigating

the applicability of an integrated hydrologic model for sub-humid regions with relatively flat topography.

3) To determine the effect of climate on groundwater recharge fluxes for coarse-textured deposits at the URSA (Chapter 4). The relationship between climate and groundwater was studied by elucidating potential canopy interception and actual evapotranspiration rates required to generate observed water table responses to short- and long-term precipitation events. Groundwater recharge rates were subsequently determined for the past 70 years and used to develop a probability distribution of recharge for coarse-textured deposits for the Boreal Plains.

4) To examine the effect of a transient flow system when using stable isotopes in water balance models (Chapter 5). The impact of gravel pit construction and subsequent alteration of the stable isotopic values of groundwater discharge to Lake 16 was used to test the assumption of steady state flow when modelling flow systems in lake-dominated areas.

Finally, a summary of these studies is presented in Chapter 6 with implications of the results for quantifying water cycling on coarse-textured deposits in sub-humid regions of the Boreal forest, and suggestions for future research.

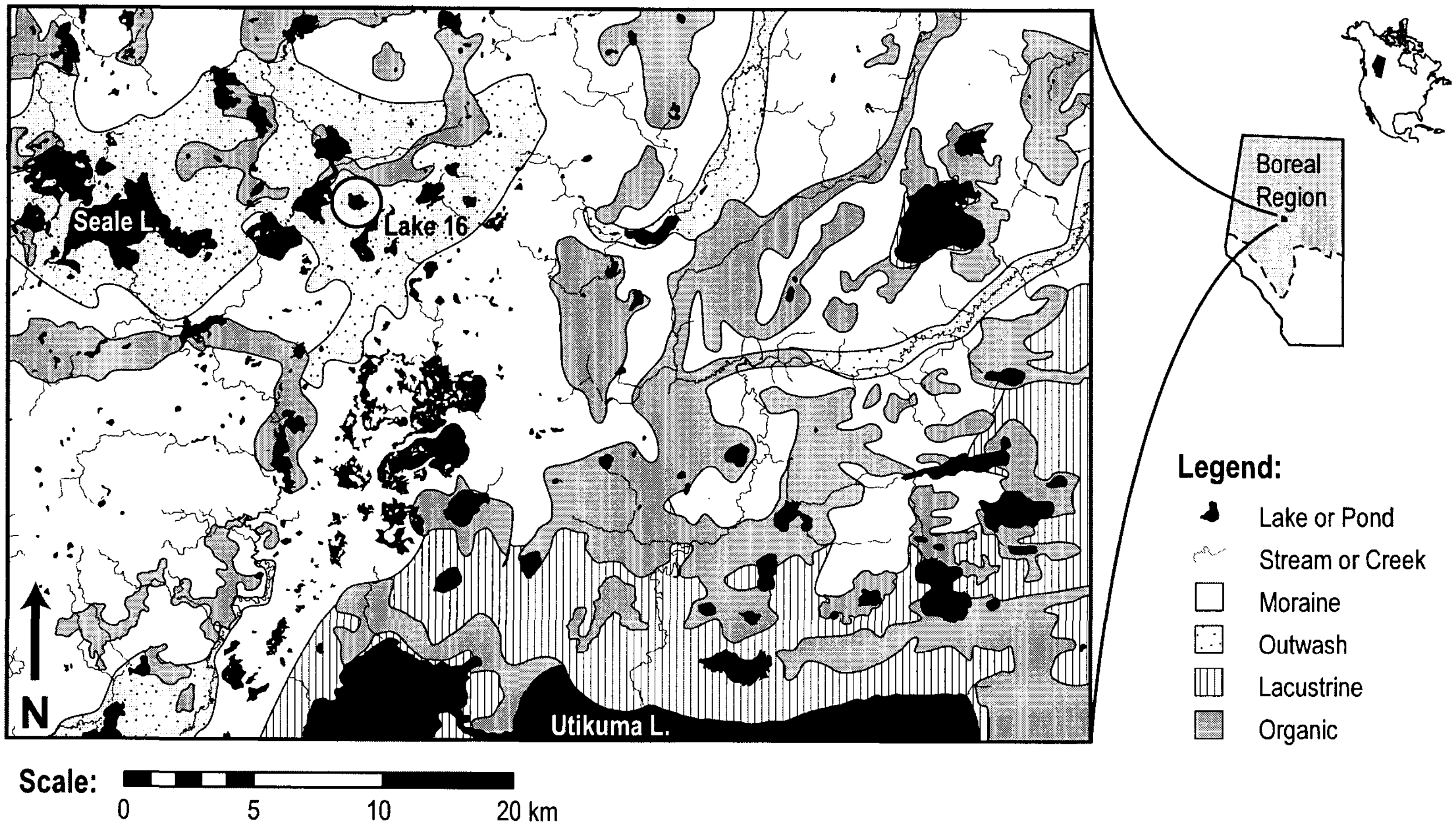
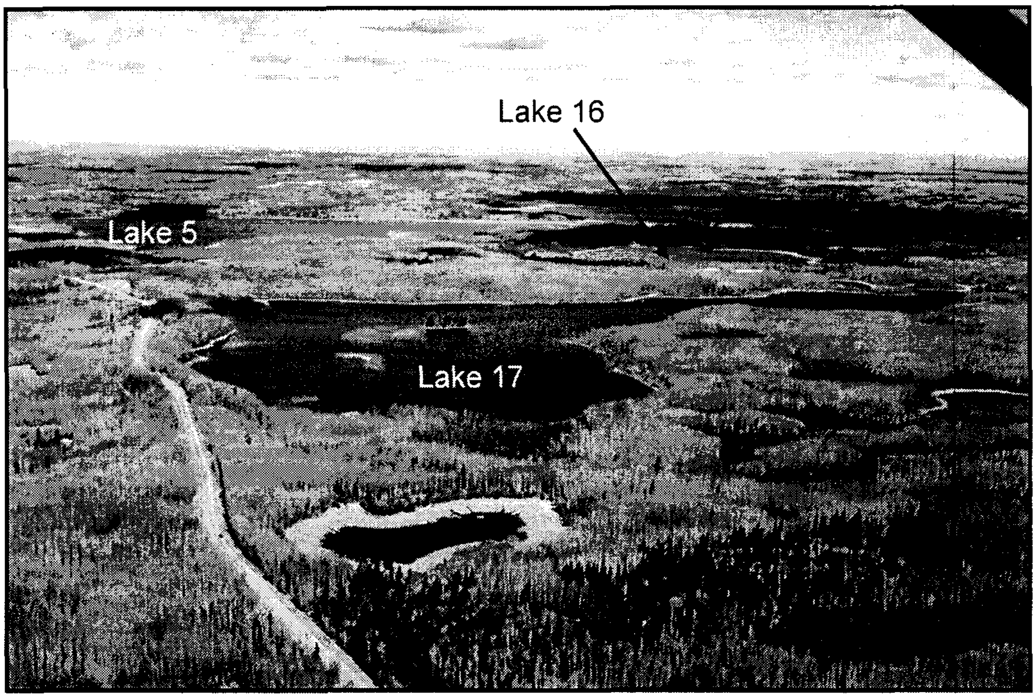


Figure 1.1 Location and surfical geology of the Utikuma Research Study Area (URSA). Lake 16 identified on the outwash plain (circled).

(a) Outwash plain at the URSA



(b) Lake 16 Study Site

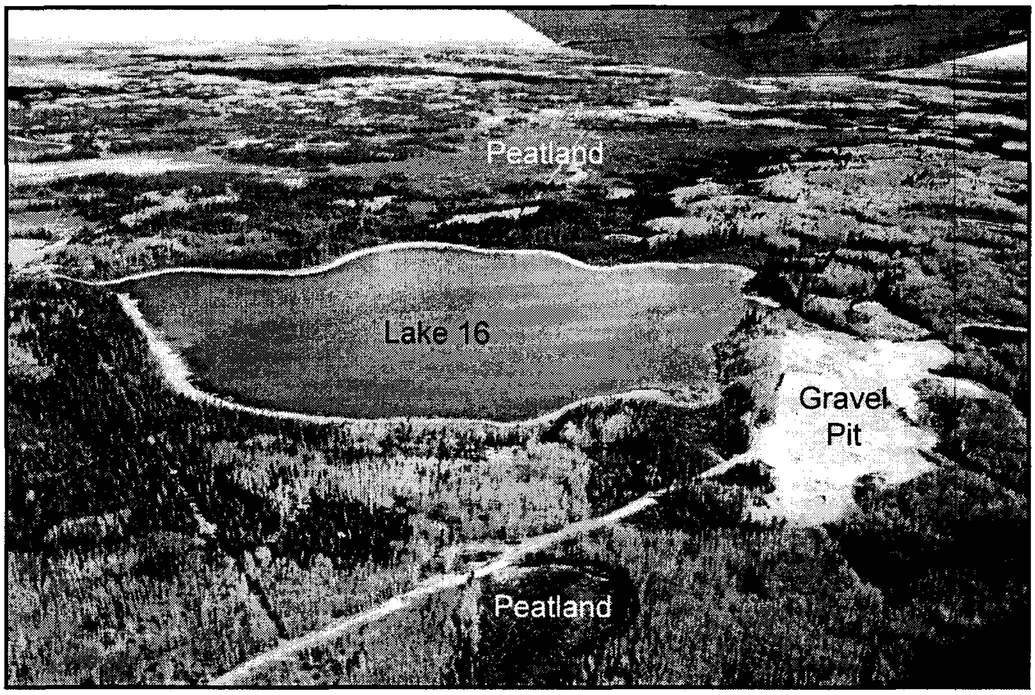


Figure 1.2 (a) Photograph of glacial outwash plain at the URSA (facing west). (b) Photograph of the Lake 16 study site (facing northwest).

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Chapter 2

Interaction of groundwater and shallow lakes on outwash sediments in the sub-humid Boreal Plains of Canada*

2.1 Introduction

The Boreal Forest covers approximately 35% of the land in Canada and contains physiographic and climatic regions that vary greatly across the country (National Wetlands Working Group, 1988). The Boreal Plain ecozone of the Western Boreal Forest (extending across three provinces in west-central Canada), is susceptible to changes from increased rates of natural resource development in the forestry and energy industries (Alberta Environmental Protection, 1998), and future scenarios of climate warming. Thus, to assess potential impact, an understanding of the natural variability in near-surface hydrologic systems is needed. The Boreal Plain ecozone differs from other Boreal areas with very low topographic relief, because it is underlain by thick glacial sediments (Klassen, 1989), and the average long-term annual precipitation is equal to, or less than average annual open-water evaporation (Winter and Woo, 1990). Within this sub-humid climate (Ecoregions Working Group, 1989), soil water storage and evapotranspiration generally control regional water budgets, and surface runoff is low (Woo et al., 2000). Thus, the high density of wetlands, ponds, and shallow lakes in the Plains region (National Wetlands Working Group, 1988) reflect complex interactions with shallow surface and groundwater systems (Ferone and Devito, 2004) that are susceptible to landscape disturbance or climate change.

The combination of climatic and geologic characteristics of the Boreal Plain is unique in the Canadian Boreal Forest, compared to adjacent Boreal Shield and Cordillera regions, and was the impetus for the Hydrology, Ecology and Disturbance (HEAD) project in northern Alberta. Hydrologic study at the Utikuma Research Study Area (URSA) initially focussed on locations of fine-textured soil, where interaction with regional groundwater was minimal (Devito et al., 2005; Ferone and Devito, 2004). The

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present study of the hydrologic controls on shallow lake and pond permanence in coarse-textured sediments followed this earlier work, and explicitly considered an area of sand and gravel outwash deposits and sand dunes. Such coarse-textured substrate is common on glaciated plains and is considered to be a significantly transmissive hydrologic unit in a watershed (Winter, 2001; Winter et al., 2003) because of more active groundwater interactions than normally associated with fine-textured substrate.

The interaction between surface water and groundwater in sandy outwash or dune sediments has been studied extensively in Minnesota (LaBaugh et al., 1997; Siegel and Winter, 1980), Wisconsin (Cherkauer and Zager, 1989; Jaquet, 1976; Kenoyer and Anderson, 1989; Krabbenhoft et al., 1990; Winter et al., 2001), and Nebraska (LaBaugh, 1986; Winter, 1986). These studies have clearly demonstrated that connection with groundwater flow systems is very dynamic at various scales, and can result in flow reversals (Anderson and Munter, 1981) and variable water geochemistry (Krabbenhoft and Webster, 1995). Understanding the dominant processes within the hydrologic cycle for a climatic region is necessary to predict the potential impact of anthropogenic changes and to sustainably manage water resources (Winter et al., 2001; Winter et al., 2003).

The objective of this study was to investigate the hydrologic controls on the interaction of a shallow lake within a groundwater flow system situated on glacial outwash deposits in the sub-humid Boreal Plain. I studied the connection between the lake and groundwater system, quantified components of the lake water budget for two consecutive hydrologic years, and determined the major controls on lake-groundwater interaction. Knowledge gained from this study, along with that from previous HEAD projects on fine-textured sites, will contribute to understanding hydrological processes in forested, sub-humid regions, to enable assessment of the potential impacts from forestry and energy industry disturbance or anticipated changes in climate.

2.2 Study Area

2.2.1 Utikuma Research Study Area

The outwash lake is located at the Utikuma Research Study Area (URSA), 370 km north of Edmonton, Alberta, Canada (56°6' N, 116°32' W; Figure 2.1). This research area is within the Plains region of the Boreal Forest and approximately 150 km south of the discontinuous permafrost zone (Woo and Winter, 1993). Overall, cold winters and warm summers characterize the climate, which has an average monthly temperature that ranges from –14.6°C to 15.6°C (Environment Canada, 2003). Annual potential evapotranspiration (517 mm; Bothe and Abraham, 1993) is reported to be slightly greater than average annual precipitation (503 mm; Environment Canada, 2003). Once every 10 to 15 years, annual precipitation is greater than evapotranspiration (Devito et al., 2005); however, most years experience a water deficit.

The URSA is characterized by low relief and is composed of glacially derived sediments overlying marine shales of the Upper Cretaceous Smoky Group (Hamilton et al., 1999). Surficial glacial sediments include glaciofluvial, glaciolacustrine, and moraine deposits, from 40 to 240 m thick (Pawlowicz and Fenton, 2002). The study site is located on a coarse-textured outwash plain that covers approximately 200 km^2 and is surrounded by fine-textured sediments from a stagnant ice moraine (Fenton et al., 2003). These coarse-grained sediments were deposited by a glacial meltwater stream that was partially in contact with glacial ice, forming an esker along the southern boundary of the outwash plain.

The outwash sediments consist of stratified sand and gravel, with some silt, whereas the esker sediments are poorly sorted and range in size from sand to cobble. The uppermost sand materials have been reworked by eolian processes since glacial deposition and have formed well-sorted, fine-grained dunes. In the vicinity of the study site (Figure 2.1), the outwash landscape shows evidence of various ice-melt structures, including kettles formed by slumping, outwash hummocks and remnants of an esker. Topographic relief is about 20 m, from the deepest depression to the highest upland area, and the outwash has a total thickness that varies from 10 to 30 m. Based on a shallow, seismic-geophysical survey and borehole drilling to 20 m, the sediments underlying the outwash are fine-textured and generally flat-lying (Domes and Schmitt, 2004). The kettle depressions are either dry, or filled with gyttja-bottomed lakes or peatlands, depending on intersection with the water table.

2.2.2 Lake 16 Study Site

Lake 16 is a 39 ha lake with groundwater springs occurring along the southeast shore and the presence of a small gravel pit adjacent to the lake (Figure 2.1). Surface drainage on the outwash plain is poorly developed, with no evidence of surface inflow to Lake 16. On the west side of Lake 16, a 1 m wide channel may discharge to a fen when the lake is ice-free (i.e., April to October). The lake basin is an average of 4.4 m deep with a maximum, localized, depth of 8.3 m within the eastern half. Although the outwash sediments are sandy, the lake basin is lined with finer-textured silt and silty-sand, and overlain by loose, flocculent gyttja to within approximately 1.5 m of the lake level. Most of the lake surface is open water; however, some tall emergent and floating macrophytes have developed. The perimeter of the lake consists of a 10 to 30 m wide riparian peatland (0.6 to 2.0 m thick), with organic materials that transition from a floating-marsh with grasses directly adjacent to the lake, to mosses, sedges, and small shrubs on peat adjacent to the upland mineral soils. Riparian vegetation is narrower (< 10 m) along the east shore at the location of groundwater discharge.

Transition from riparian areas to upland mineral soils is abrupt along the south and east margins of the lake basin. The slopes are approximately 10% and rise to 5 to 10 m above the lake level. Along these segments the upland sediments are well sorted, fine to medium sand, with occasional gravel and cobble zones and very few clay or silt lenses. The upland materials form sand ridges, associated with the remnants of an east to west trending esker, and are mostly forested with pine and aspen on a thin (< 0.1 m) organic layer (i.e., forest floor). Isolated depressions in these sandy uplands have developed into bog-peatlands, from 1.6 to 2.2 m thick, containing mostly black spruce, Labrador tea, and mosses. Linear depressions between the sandy ridges are generally narrow and have a closed canopy, while elliptical depressions are generally open and contain more shrubs than spruce trees. Along the north and west margins of the lake basin, gradual slopes grade into more extensive peatlands and the topography is generally flat. Sandy upland ridges are less common, and generally only 2 to 3 m higher than the lake level. The vegetation pattern is similar to that south of the lake (i.e., pine and aspen on the uplands, and spruce, mosses and sedges on the peatlands).

2.3 Field Methods

2.3.1 Groundwater

At 41 locations in the study area, 70 piezometers were installed in 0.051 to 0.089 m diameter boreholes to depths ranging from 1.2 to 9.1 m (Figure 2.1). All piezometers were constructed of 0.025 m inside diameter PVC, each having machine-slotted, 0.3 m long screens wrapped in filter cloth. All piezometers were installed with the screened interval below the water table, and most were installed in loose, sandy substrate that would typically collapse into the borehole below the water table. In these instances, the piezometer was driven through the loose, collapsed sandy material to the bottom of the borehole using a hammer. This technique allowed piezometer tips to be surrounded by native sediments, requiring no additional sand pack. For the few locations where finer-textured geologic materials were encountered, a sand pack was installed to 0.05 m above the slotted interval. The remaining borehole annulus, above the collapse depth or sandpack interval, was filled to the ground surface with medium bentonite chips, to ensure that piezometer tips were isolated from water movement along the outside of the casing. For the three locations in the centre of the lake, 0.025 m diameter PVC piezometers were driven to the approximate base of the gyttja, the depth of which was determined from probing with ridged metal rods. Additionally, 0.019 m diameter PVC piezometers with pointed tips were driven 1.2 to 2.0 m deeper into the lakebed sediments and fastened to steel pipe driven into the lakebed for stability during lake ice formation and break-up.

Water levels in all piezometers were measured manually, once or twice per month during the ice-free months (April to October) from 2001 to 2003, and every two months during winter. Eight piezometers were outfitted with pressure transducers and recorders (GlobalWater model WL-14) to continuously record hydraulic head data.

Saturated hydraulic conductivity values, for 25 piezometers completed in the mineral soil and two piezometers completed in the gyttja, were calculated by the Hvorslev (1951) method. Particle-size distributions from four boreholes were determined by sieve analysis of samples collected every 0.9 m, to 9.1 m depth. Weight fractions for 11 particle diameters (i.e., sieves), ranging from 0.063 to 2 mm, were measured. Grain diameters greater than 2 mm, such as those encountered while drilling in the gravel pit, were measured with a ruler. The presence and abundance of finer-textured grain size fractions, such as lakebed sediments and clay lenses within the sandy mineral soil, were estimated in the field.

2.3.2 Lake-Groundwater Seepage

Groundwater discharge along the east shore was measured using four seepage meters (Lee, 1977), made from plastic 55-gal drums, cut to 0.25 m high, and inserted into the lakebed. Collection bags (3.8 L) were pre-loaded with 0.5 to 1.0 L of lake water and emptied of air, as suggested by Shaw and Prepas (1989). Discharge was measured three times per month during the ice-free season. Seepage meters were installed into sandy lakebed sediments at locations with minimal organic accumulation, to ensure a proper seal between the seepage meter and the sediments, and thus to measure groundwater discharge directly at the lakebed interface.

For eight segments along the perimeter of the lake basin and three areas of the lakebed, seepage to and from the lake was estimated using piezometers and Darcy's Law (Freeze and Cherry, 1979):

$$Q = -KA\frac{dh}{dL} \qquad [2.1]$$

where Q is the discharge (m^3/s), K is the (saturated) hydraulic conductivity (m/s), A is the cross-sectional seepage area (m^2), and dh/dL is the hydraulic head gradient (dimensionless). For shoreline segments, the horizontal gradient was determined from the lake to an adjacent upland water table position. At the location of groundwater springs on the east shore, and for the three lakebed areas, vertical gradients between piezometers completed in the mineral soil or deep gyttja, and the lake level were used.

The net groundwater inflow was also estimated for periods when the lake was covered with ice (Rosenberry et al., 2000). This technique minimized the uncertainty associated with using Darcy's Law, by considering seepage for a period when the majority of typical hydrologic budget components were inactive (i.e., no summer precipitation, lake evaporation, or surface outflow). The net inflow was determined by measuring the difference in the lake level immediately before and immediately after the onset and break-up of lake ice.

2.3.3 *Surface Water*

The stage of Lake 16 and the two adjacent lakes in the study area (Lakes 5 and 17) were referenced to angle-iron staff-gauges driven into the mineral soil beneath each lake. Lake levels were continuously recorded during ice-free months using a pressure transducer and recorder (GlobalWater model WL-14) housed in a 0.051 m diameter PVC stilling well at each staff gauge. At Lake 16, the pressure transducer was lowered during winter months, beginning in 2002, to a depth sufficient to prevent it becoming frozen into the ice, and allow collection of a continuous record of hydraulic head beneath the lake ice.

A 1 m wide plywood flume was constructed on the outflow channel that discharges west of Lake 16. The outflow area was covered in wetland-marsh vegetation and organic soil; however, the small channel appeared to have existed for many years. The flume penetrated 0.9 m below the wetland surface into compacted peat that directly overlies sand, at 1.2 m depth. Flume walls converged at 45°, to a 0.7 m wide, parallel throat, where velocity, depth, and temperature were recorded continuously using a Doppler-style flowmeter and datalogger (Unidata Starflow model 6526) during ice-free months. Stage and velocity measurements were periodically corroborated using a staff gauge installed downstream of the flume and floats placed in the flume, respectively.

The depth of the lake basin was measured at 22 locations in March 2002, when the lake was covered with ice. At each location, depth to loose gyttja, firm gyttja, and mineral soil were measured. For this study, loose gyttja was defined as the depth at which a 152 mm diameter secchi-disk rested under its own weight (500 g) and firm gyttja was defined as the depth at which ridged, metal probing rods rested under their own weight (approx. 1800 g). Depth to mineral soil was determined by inspecting the bottom of the ridged metal rods for inorganic sediment remains after they were driven into the lakebed material and removed. These estimates of overall lake basin depth were combined with the known elevation of the lake ice during the survey and used to generate lake bathymetry by contouring depth measurements, which allowed for determination of a lake depth-volume function.

2.3.4 *Precipitation and Evaporation*

From May 2001 to October 2003 a weight-recording gauge at a meterological station provided a continuous record of summer and winter precipitation, 10 km west of the site. Summer precipitation at Lake 16 was measured using a bulk rain gauge in 2002 and both a bulk rain gauge and a tipping-bucket rain gauge (Jarek model 4025) in 2003 (Figure 2.1). The difference between the cumulative rainfall between May and September from the weight-recording gauge and the measurements made at the study site was 10 mm. Snow depth was measured at a minimum of 5 locations at the centre of the lake in March of 2001, 2002 and 2003. In addition, snow depth was measured along 3 snow courses from the lake edge to the upland areas in March 2003. Snow water equivalent (SWE) values were calculated from composite samples each year from the lake's centre. These values were treated as estimates, because water at the base of the snowpack could have become part of lake ice during the winter months.

Evaporation from the lake was estimated with a Class A evaporation pan that was placed on the east edge of the lake, similar to Hayashi et al. (1998). During the summer of 2001 the pan was partially submersed; however, during 2002 the pan was largely exposed because of a lower lake stage. To account for possible errors of the exposed pan, two pans were used in 2003: one exposed on the lake edge and a second partially submersed in the lake. In addition, the temperatures of the water in each pan, the lake and the air were recorded (Onset H8 Logger and External Temperature Sensors). Water levels were recorded continuously with a pressure transducer-datalogger device (GlobalWater model WL-14), and confirmed with periodic manual measurements. The water level in each pan was kept to within 0.1 m of the top rim. The differences observed between these two pans were used to modify evaporation estimates determined from the single exposed pan in 2002.

2.3.5 *Stable Isotopes*

Water samples were collected in October 2001 from Lakes 16 and 17, selected piezometers, and seepage meters. Piezometers were located directly upgradient and downgradient of the lake, and included shallow piezometers at the lake margin, in the upland areas and two deep piezometers. Shallow piezometers were sampled with a portable vacuum pump, and deep piezometers were sampled with an inertial-lift tubing system (Waterra). Samples obtained from the seepage meters were collected in pristine sample bags, not subjected to the preloading technique described previously. Samples were stored in 20 mL scintillation vials with a thin film of mineral oil to prevent evaporation.

Stable isotope ratios for hydrogen and oxygen were determined by the Isotope Science Laboratory at the University of Calgary. Hydrogen isotopes were determined by the chromium reduction and mass spectroscopy technique (2 per mil error), and oxygen isotopes were determined using the CO_2-H_2O and mass spectroscopy technique (0.2 per mil error). Results were expressed as per mil difference (‰), relative to Vienna Standard Mean Oceanic Water (VSMOW).

2.4 Results

2.4.1 Outwash Sediment Properties

Grain-size distribution curves were plotted for the predominant mineral soils, which include sandy ridge, dune, and esker deposits (Figure 2.2). Of the 33 samples analyzed, 29 were fine sand, three were silty fine sand, and one was coarse sand with gravel. The approximate pore-throat diameter (d_{10}), average grain size diameter (d_{60}), and hydraulic conductivity (K) values are summarized on Figure 2.2 for each of the three material types. Sediments from sand ridges on the south side of the lake were very well sorted, fine sand, with a narrow range in the d_{10} and d_{60} values (0.06 to 0.13 mm, and 0.19 to 0.21 mm, respectively). Dune and esker sediments were poorly sorted and contained a more heterogeneous particle-size distribution. Dune sand along the west side of the lake was composed of sand similar to the ridges, but also contained thin silt and clay lenses (up to 0.05 m thick), which increased the cumulative distribution range (d_{10} from <0.01 to 0.13 mm and d_{60} from 0.14 to 0.30 mm). Sand samples collected from the esker had a similar range in particle diameter (d_{10} from 0.03 to 0.25 mm and d_{60} from 0.18 to 0.30 mm), but also contained coarse sand and gravel size fractions, up to 51 mm diameter. This heterogeneity was confirmed at other locations along the esker, such as the gravel pit, where grains from 1 to >100 mm in diameter were observed.

Hydraulic conductivity of the mineral soils in the study area varied by six orders of magnitude (10^{-3} to 10^{-8} m/s). Porous media have been sub-divided into five primary groups, based on soil texture (Figure 2.3a). Most of the landscape is composed of sandy materials, which have mean K of 4.0×10^{-5} m/s. Sandy areas with a higher proportion of silt, have mean K of 3.4×10^{-6} m/s, whereas the coarser sand and gravel sediments, associated with the gravel extraction area, are estimated to have mean K greater than 1.0×10^{-3} m/s. The hydraulic conductivity of lakebed materials decreases with depth, such that the compacted gyttja (4.3×10^{-6} m/s) is able to more easily transmit water than lakebed mineral sediments (1.3×10^{-8} m/s).

2.4.2 Precipitation and Evaporation

Annual precipitation for the 2002 and 2003 hydrologic years was 283 mm and 370 mm, respectively (Figure 2.4). Hydrologic years were defined from November to October to capture the typically active events (spring melt, summer rainfall and evaporation) within each reported year (similar to Hayashi et al., 1998). The values recorded in this study are lower than the long-term average of 503 mm (Environment Canada, 2003), but are similar to the findings of Ferone and Devito (2004) for 2000, in the same region (344 mm). More than 70% of the precipitation occurs between May and October each year (239 mm and 287 mm), from short-duration, convective-cell, summer storms. During this study (May 2001 to Oct 2003), 99% of the daily precipitation events were less than 10 mm. There were only six days where precipitation exceeded 20 mm (Figure 2.4). SWE was 44 mm and 83 mm in March 2002 and 2003, respectively, which compared well to winter precipitation (42 and 88 mm) recorded at the weight-recording gauge, 10 km west of the study area.

For summer 2003, cumulative evaporation measured from the submersed and exposed evaporation pans were 465 mm and 403 mm, respectively, and the average difference between water temperature in the evaporation pan and the lake was only 0.35°C. These data suggest that actual lake evaporation was 1.2 times higher than that recorded using an exposed evaporation pan. The difference in evaporation estimates is primarily attributed to the presence of grasses around the elevated pan, which would have affected the microclimatological conditions (Petrone, personal communication, 2004).

Measured pan evaporation for 2002 was 336 mm, using the coefficient for an exposed pan (i.e., 1.2).

2.4.3 Lake Levels and Surface Water Discharge

The lakes within the study area exist as a series of 'steps', where Lake 16 was 1.8 m higher than Lake 5, and 2.4 m lower than Lake 17 (Figure 2.3a, Figure 2.4). Over the course of the study, the stage of Lake 16 decreased by 0.18 m. Most lake level decline occurred in summer months, with some increase through the winter months (0.09 m in 2002, and 0.08 m in 2003). A greater decrease was observed in Lake 17 (0.57 m over the study); however, there was not an associated increase throughout the winter months, suggesting diminished or absent groundwater input to the upgradient lake. The 0.14 m increase in Lake 17 recorded during spring 2003 could be a result of groundwater discharge during the winter months. The stage of the much larger, downgradient Lake 5 remained nearly constant throughout the study, suggesting lake level maintenance by connection with regional groundwater, and proportionally less groundwater exchange relative to lake surface area.

Daily surface discharge from Lake 16 varied from 0 to 29 mm (i.e., up to 0.05 m^3/s), and was mostly below 5 mm (Figure 2.5). Peak discharge rates were observed when the lake stage was above 643.80 m, as seen in the last week of June 2002. There is no empirical relationship between rainfall events and discharge from the lake. In 2002, when the lake stage was above 643.80 m, the only increased outflow event following rainfall occurred from August 3 to 4 (37 mm outflow) following 52 mm of rain in the last week of July. Once the lake stage declined below 643.80 m, similar amounts of rain at the end of summer 2002 and for the duration of 2003 did not produce the same outflow response from the lake basin as observed in 2002. These data suggest that surface outflow exists only when the lake level is above 643.80 m (i.e., the critical outlet elevation).

2.4.4 Water Table Configuration

Figure 6 illustrates the configuration of the water table around Lake 16 for spring (May), summer (July), and autumn (October) months of 2001 to 2003. Overall, the groundwater flow direction is from southeast to northwest; however, there were various degrees of interaction with the lake over the course of the study. The horizontal hydraulic gradient is steeper on the south side of the lake (0.007), than on the north side of the lake (0.001), as is evident in the water table hydrographs shown in Figure 2.7. A small rise in the water table north of the lake formed an elevated ridge of groundwater that focused groundwater flow toward the lake in May 2002 and July 2003 (Figure 2.6 and piezometer 28 on Figure 2.7).

Between Lakes 17 and 16, the water table configuration and flow direction remained consistent throughout the study. Because the stage of Lake 17 decreased more than Lake 16, the average horizontal gradient between the lakes decreased from 0.0052 to 0.0045; however, this decrease was not apparent in wells adjacent to Lake 16 (Figure 2.7). Most of this gradient decrease occurred between October 2001 and October 2002 (0.0051 to 0.0046). Between Lakes 16 and 5, the average horizontal hydraulic gradient only decreased from 0.0032 to 0.0028 between October 2002 and October 2003. In this downgradient area, the water table sloped from Lake 16 to Lake 5, except for a narrow water table ridge that formed periodically, southwest of Lake 16. The water table ridge formed downgradient of the lake in the spring of each year and created a reversal of the horizontal hydraulic gradient along most of the west side of Lake 16 (Figure 2.6 and piezometer 4 on Figure 2.7). By the summer months, this groundwater ridge decreased in size, except for summer 2002, when a large decline in lake stage and the adjacent water table completely reduced the groundwater ridge. In 2003, the groundwater ridge was maintained through the summer and into the autumn months, in response to 2 times SWE (83 mm) for winter 2003, compared to previous years for this area.

2.4.5 Lake-Groundwater Hydraulic Gradients

The larger upgradient and downgradient lakes establish the overall southeast to northwest groundwater flow direction. The resulting 'staircase' of lakes follows the groundwater flow direction and creates a flow-through style lake that has an average horizontal hydraulic gradient of 0.002 across the entire study area, over the course of the study (Figure 2.3b). Upgradient of Lake 16 (southeast), vertical hydraulic head gradients are negligible (piezometer nest 01-P8 on Figure 2.3b), indicating the dominance of lateral groundwater flow. Along the southeast margin of the lake basin, the vertical gradient shifts to -0.035 (piezometer 37, Figure 2.3b), indicating upward groundwater flow, confirmed by the presence of springs along the shore of the lake. Downgradient of Lake 16 (west), vertical gradients increase with distance from the lake, indicating downward groundwater flow and recharge conditions: +0.014 to +0.021 at PZ-55 and +0.225 at piezometer nest 01-P9 (Figure 2.3b).

Temporal vertical hydraulic head gradients for the gyttja and underlying lakebed sediments are shown on Figure 2.8 for three representative piezometer nests. Hydraulic gradients between the compacted gyttja and lake water frequently alternate between recharge and discharge conditions. However, the magnitude was generally less than 0.01, which is considered within measurement error, and likely represents minor changes to fluid flow within the gyttja layer. Between the underlying mineral lake sediments and the gyttja, the three representative piezometers nests report different vertical gradients. At PP4 the gradient from the gyttja to the lake indicated consistent groundwater discharge conditions (-0.019 to -0.031). Conversely, at PP3 groundwater recharge conditions were recorded (+0.038 to +0.025). For the area represented by PP2, lakebed to gyttja gradients varied from +0.004 to -0.009, with the recharge conditions occurring early in the summer and the discharge conditions occurring late in the summer. The data from this lakebed segment indicated that this area experiences small gradients with frequent reversals of direction.

2.4.6 Groundwater Seepage Measurements

Exchange of groundwater with Lake 16 was estimated for the eight shoreline segments and three lake areas (Figure 2.7) using the measured hydraulic gradients and equation 2.1. Division of lakebed segments was based on the position of adjacent shoreline segments that identified a division between groundwater recharge and discharge conditions, and the water table configuration. Because most groundwater inflow occurred in the vicinity of the springs on the east margin of the lake, calculated seepage volumes were compared to average discharge estimates obtained from the array of seepage meters for ice-free months in 2001 to 2003. Specific discharge estimates from the seepage meters and flux calculated from the Darcy's Law approach varied between

5×10^{-8} and 3×10^{-6} m/s for the duration of this study and were within an order of magnitude for any specific time. Calculated seepage values (using Darcy's Law) were also compared to the net groundwater discharge entering the lake during winter months, when Lake 16 was covered in ice (Figure 2.4). Using this method, calculated groundwater discharge rates were only 20% less than the estimated winter seepage for 2002 and 2003. The corroboration of seepage meter measurements and net winter groundwater discharge, supports using the fluxes calculated with equation 2.1 as for an estimate of the lake water budget.

2.4.7 Summary of Hydrologic Budget Components

Monthly estimates of each component of the hydrologic budget for Lake 16 (after LaBaugh et al., 1997) were made for two hydrologic years beginning November 1, 2001 (Figure 2.9):

$$\pm r = P + GW_{in} - E - GW_{out} - SW_{out} - \Delta V \qquad [2.2]$$

where P is precipitation, GW_{in} is groundwater discharge to the lake, E is lake evaporation, GW_{out} is groundwater recharge from the lake, SW_{out} is surface water discharge, ΔV is the change in lake volume, and r is the residual. The annual budgets shown on Figure 2.9 have residuals of 39 mm and 41 mm (8% and 7% of total inputs).

Annual E exceeded P for each year, as found by Ferone and Devito (2004) within the same region in 2000, suggesting a multi-year water deficit within a decadal climate record (Devito et al., 2005). During the first hydrologic year, SW_{out} (293 mm) also exceeded P, and therefore, storage within Lake 16 decreased by 109 mm by the end of the year. Because the lake level decreased below the outlet elevation (643.80 m) in the first year, SW_{out} decreased in the second year (117 mm), falling from the second largest water flux to the fourth. With decreased SW_{out}, the change in lake storage during the second year was minimized, and Lake 16 gained 33 mm by October 2003.

Figure 2.9 shows that P, E, and SW_{out} were dominant during the summer, or ice-free months. Groundwater discharge to the lake was generally consistent throughout each year and contributed 228 mm and 220 mm, for each of the respective hydrologic years. Likewise, groundwater recharge was consistent (30 mm and 26 mm) and was an order of magnitude lower than groundwater discharge.

2.4.8 Evaporated Water Sources

The isotopic ratios have been plotted as $\delta^{18}O$ against $\delta^{2}H$ (Figure 2.10), relative to the local meteoric water lines calculated for Edmonton, Alberta (Maulé et al., 1994), and the Utikuma Research Study Area (Devito, unpublished data: 1999 - 2002). Lake water and groundwater discharge had $\delta^{18}O$ values heavier than the local meteoric water line, indicating enrichment by evaporation. Within this set of near surface samples, there was also a difference of 17 ‰ $\delta^{18}O$ and 3 ‰ $\delta^{2}H$, between groundwater discharge and water from Lake 16.

Deeper groundwater samples had depleted $\delta^{18}O$ and $\delta^{2}H$ signatures compared to the lake water and near surface groundwater discharge. Samples from 15 and 18 m depth, at locations both upgradient and downgradient of the lake, were very similar to the local meteoric water line, suggesting that the isotopic enrichment at the surface had not affected deeper groundwater.

2.5 Discussion

2.5.1 Utikuma Outwash Groundwater Flow System

Outwash lake and groundwater interaction have been studied for decades in Northern Wisconsin (e.g., Cherkauer and Zager, 1989; Jaquet, 1976; Kenoyer and Anderson, 1989; Krabbenhoft et al., 1990), and the results have guided how similar studies are executed in other regions. The geologic setting, hydraulic properties, and topography of the study areas in Wisconsin outwash sediments are likely similar to those at the URSA, which allows the opportunity for a comparison of hydrologic controls on lake and groundwater interaction. The URSA receives less than half of the average annual precipitation, compared to the hydrologic study areas on outwash deposits in Wisconsin, and frequently has higher annual potential evapotranspiration than precipitation. Therefore, differences in regional climate are a major controlling factor on surface-water/groundwater interaction.

The series of lakes within the Utikuma outwash study area are well connected to local and larger-scale groundwater flow systems, as anticipated for highly permeable sediments (Smith and Townley, 2002; Tóth, 1963). Along the topographic gradient between the three lakes in the study area, Lake 17 receives little to no input from groundwater sources and is sensitive to annual precipitation as a hydrologic input, as shown by the decline in stage over the duration of this study. For the same climatic conditions, lakes progressively lower in the topographic gradient should capture more groundwater flow system (Townley and Davidson, 1988). Thus, lower lakes are more resistant to periodic water deficit conditions than shallow, upgradient ponds within the same climate zone. Cheng and Anderson (1994) demonstrated such landscape position behaviour with a numerical model. Isotopic evidence suggests that groundwater deeper in the outwash sediments may not mix with near-surface groundwater, indicating a local-scale groundwater-flow system within the outwash (Tóth, 1963), and different flowpath residence times (Pint et al., 2003; Walker et al., 2003).

The coarse-textured, hummocky, outwash sediments allow the formation of water table mounds downgradient from the lake during spring months, similar to those described by Anderson and Munter (1981) and Winter (1983). The narrow, ridge shape of the groundwater mound observed is a function of the relative dominance of recharge, compared to the moderating horizontal hydraulic gradient between Lake 16 and Lake 5. The seasonal formation of a water table ridge and its complete disappearance in the summer of 2002 indicates that hydrologic systems in the sub-humid Boreal Plains region, exhibit groundwater phenomena caused by processes associated with humid regions (i.e., spring recharge and mounding, (Anderson and Munter, 1981; Cherkauer and Zager, 1989)) and drought-prone regions (i.e., evaporative driven disappearance of the water table mound, (Meyboom, 1967; Rosenberry and Winter, 1997). The presence of these hydrologic phenomena, within a single year illustrates the sensitivity of shallow hydrologic systems to subtle annual variations in precipitation and evaporation, and the dependence on classifying subsurface hydrologic conditions in this region to evaluate disturbance.

2.5.2 Lake-Groundwater Interaction

Groundwater discharge supplies Lake 16 with a consistent source of water throughout the year. It is the dominant hydrologic flux from November to April. My approach to quantifying the amount of groundwater interaction relied on calculations using hydraulic head gradients measured between piezometers, which has been used in similar studies (e.g., LaBaugh et al., 1997). Although there is some uncertainty with this method (Hunt et al., 1996; Winter, 1981), seepage flux calculations are useful in understanding the broad interaction of lakes and groundwater (Kishel and Gerla, 2002).

Most groundwater discharge occurs along one edge of the lake margin and creates a flow-through system (Born et al., 1979), as expected for lakes in connection with intermediate or regional groundwater flow systems (Schuster et al., 2003; Smith and Townley, 2002). The amount of groundwater discharge for Lake 16 in the Boreal Plain is 77% lower than groundwater discharge at Williams Lake in Minnesota (Schuster et al., 2003; Siegel and Winter, 1980), where the average water table gradient is on the same order of magnitude (Siegel and Winter, 1980). However, the stage of Lake 16 periodically rises above the outlet elevation, thereby causing the amount of annual surface outflow to increase, and exceed annual groundwater input. This dynamic relationship modifies the traditional definition of a flow-through lake and creates a situation where groundwater may discharge to surface water, then exit the lake basin as surface water, and either recharge the groundwater regime, or be captured by high evapotranspiration through fen vegetation during the growing season. These interactions of surface water and shallow groundwater are complex, have implications for near-surface biogeochemistry, and need to be considered when representing fluid exchange using a numerical flow model.

Interaction between groundwater and the Lake 16 is controlled in part by lake stage, which depends on precipitation and groundwater inputs, and evaporation and surface flow outputs. Lake stage decreased by 0.2 m during the two hydrologic years in this study as a result of surface outflow and evaporation in summer 2002. During this same time, the average horizontal gradient between the lakes decreased, and annual groundwater discharge to the lake decreased by 8 mm (approximately 4% of groundwater input).

These transient fluxes support the concept of a dynamic, sensitive relationship between surface water and groundwater in the Boreal Plain, and show that hydrologic response to climate variability could be different than has been shown for the humid and semi-arid research stations in the mid-west USA (Winter et al., 2001), because of the fine balance between water sources and sinks. Furthermore, use of isotopic analyses to refine lake water budgets (e.g., Krabbenhoft et al., 1994) must consider such transience, and have not been explored extensively for groundwater-fed lakes in the Boreal region (Gibson et al., 2002).

2.5.3 Evaporation Windows

The largest hydrologic flux was evaporation from the lake surface, which exceeded precipitation for each of the two hydrologic years during this study. However, in the Boreal Plains region, the precipitation and evaporation components are only active from the months of April to October. For the remainder of the year, groundwater inflow was the dominant hydrologic component. The onset, duration, and timing of evaporation are important controlling factors on lake stage, which in turn, is a controlling factor on groundwater exchange. In the second hydrologic year, an earlier ice-free season allowed for higher evaporation rates during April and May, resulting in an additional 46 mm of evaporation compared to the same two months in the previous year. However, with a decreased lake stage, a slight reduction in both groundwater input and output was calculated between the first and second hydrologic year (8 and 6 mm), thereby illustrating the control that evaporation has on surface-water/groundwater interactions.

Manifestation of evaporative control on lake stage in this sub-humid region can also be seen from the downgradient position of hinge lines in the water table (Gosselin and Khisty, 2001; Townley and Davidson, 1988). Such features indicate that groundwater is preferentially drawn toward the lake basin, where evaporation is the dominant flux. In this manner, the shallow lakes in this region appear to exist as 'evaporation windows', a concept which is corroborated by isotopically enriched $\delta^{18}O$ and $\delta^{2}H$ values of shallow groundwater, compared to local meteoric water, or groundwater deeper in the flow system.

2.6 Conclusions

Shallow lakes and ponds on coarse-textured glacial deposits in the Boreal Plain exist in a fine balance between precipitation and evaporation. The presence and long-term sustainability of these surface water features in a sub-humid climate are a result of topographic depressions that are lined with low permeability mineral sediments and compacted gyttja, and supplied consistently by groundwater throughout the year. The hydraulic connection with larger-scale groundwater flow systems (i.e., beyond the local scale), and potentially large capture zones protect the shallow surface water bodies against precipitation-deficit conditions, which appear to exist for multiple years in the Boreal Plains region. However, the buffering capacity provided by the groundwater regime may conceal changes imposed by anthropogenic disturbance, within the natural variability. Therefore, understanding the transient hydrologic response of shallow lakes and ponds in these geologic settings is paramount for land-use management and assessing climate change in the Boreal Plains, which will be simulated with a detailed numerical model (see Chapter 3).

Lake and groundwater interaction within the outwash sections of the Boreal Plains has hydrological characteristics that have been observed in both humid and semi-arid environments. The dynamic relationship between precipitation, groundwater interaction, and surface flow is controlled by the evaporative flux from lake surfaces. In this regard, shallow lakes and ponds on the Boreal outwash are considered evaporation windows during ice-free months and are very sensitive to the onset of warm spring air temperatures, which will be enhanced considering future scenarios of climate change in northern environments. Additional water balance techniques using stable isotopes or simulation modelling of surface and groundwater flow may be valuable in determining potential impact to shallow hydrologic systems, but require further study before application (see Chapter 3 and Chapter 5).

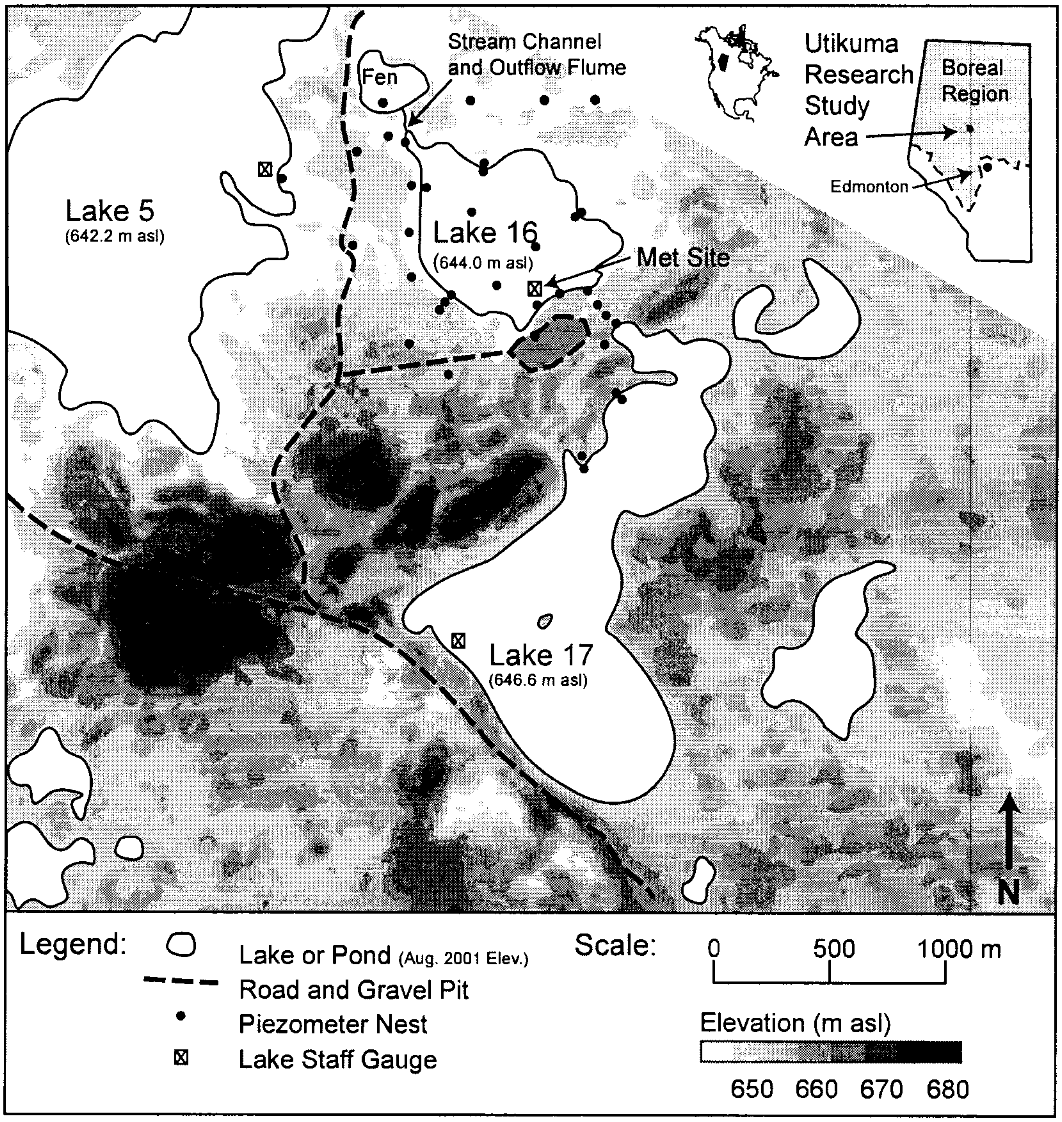

Figure 2.1. Lake 16 study site with topography and hydrologic instrumentation (URSA and Boreal Plains region in Alberta, Canada on inset).

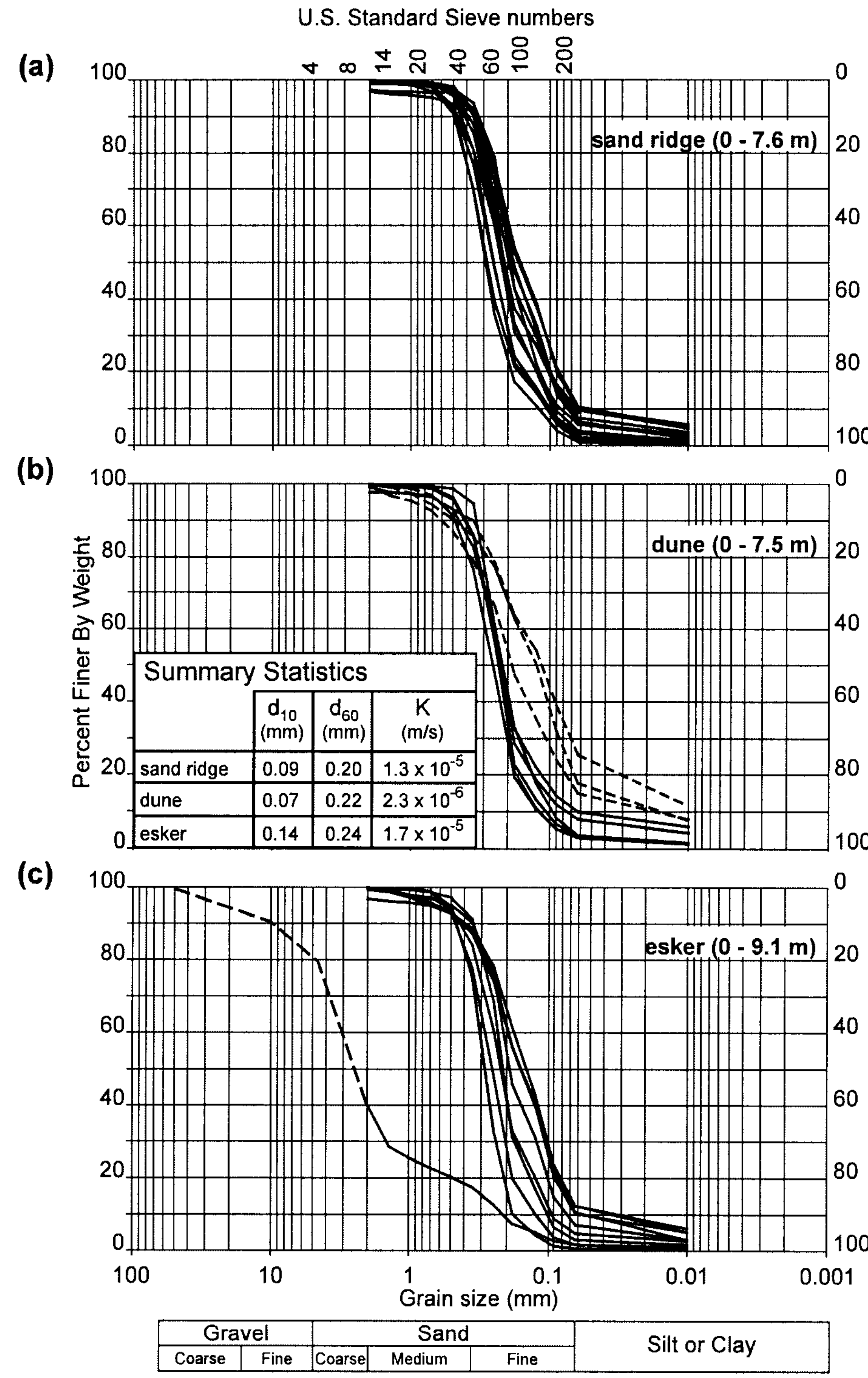

Figure 2.2. Grain-size distributions for (a) sand ridge, (b) sand dune (dashed lines for samples with silt lenses), and (c) esker deposits surrounding Lake 16 (dashed line for gravel and cobble). Arithmetic average grain-size distributions and measured hydraulic conductivity values on inset table.

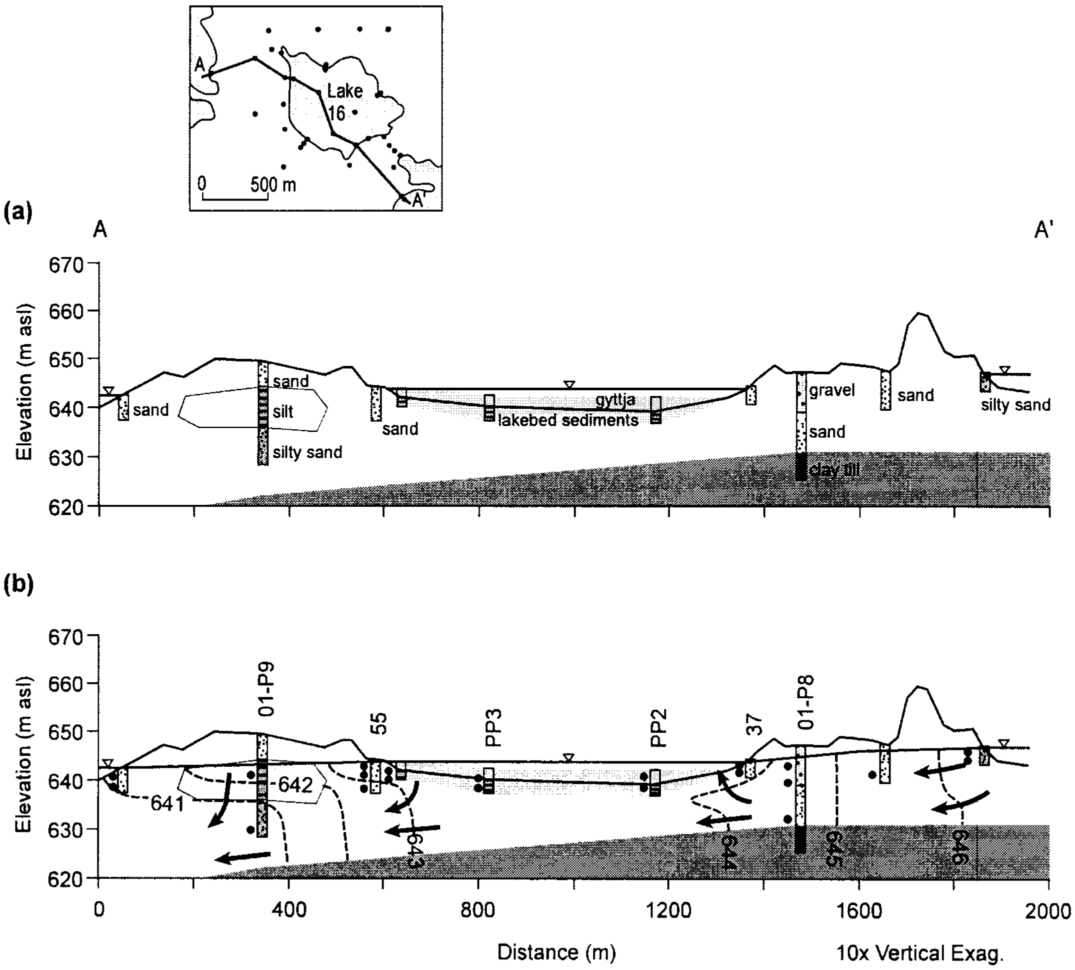

Figure 2.3. (a) Geologic cross section through Lake 16, with porous media zones. Cross section location shown on inset. (b) Hydraulic head distribution and interpreted groundwater flowpaths for July 2002. Piezometers (dots) referenced in the text have been labeled.

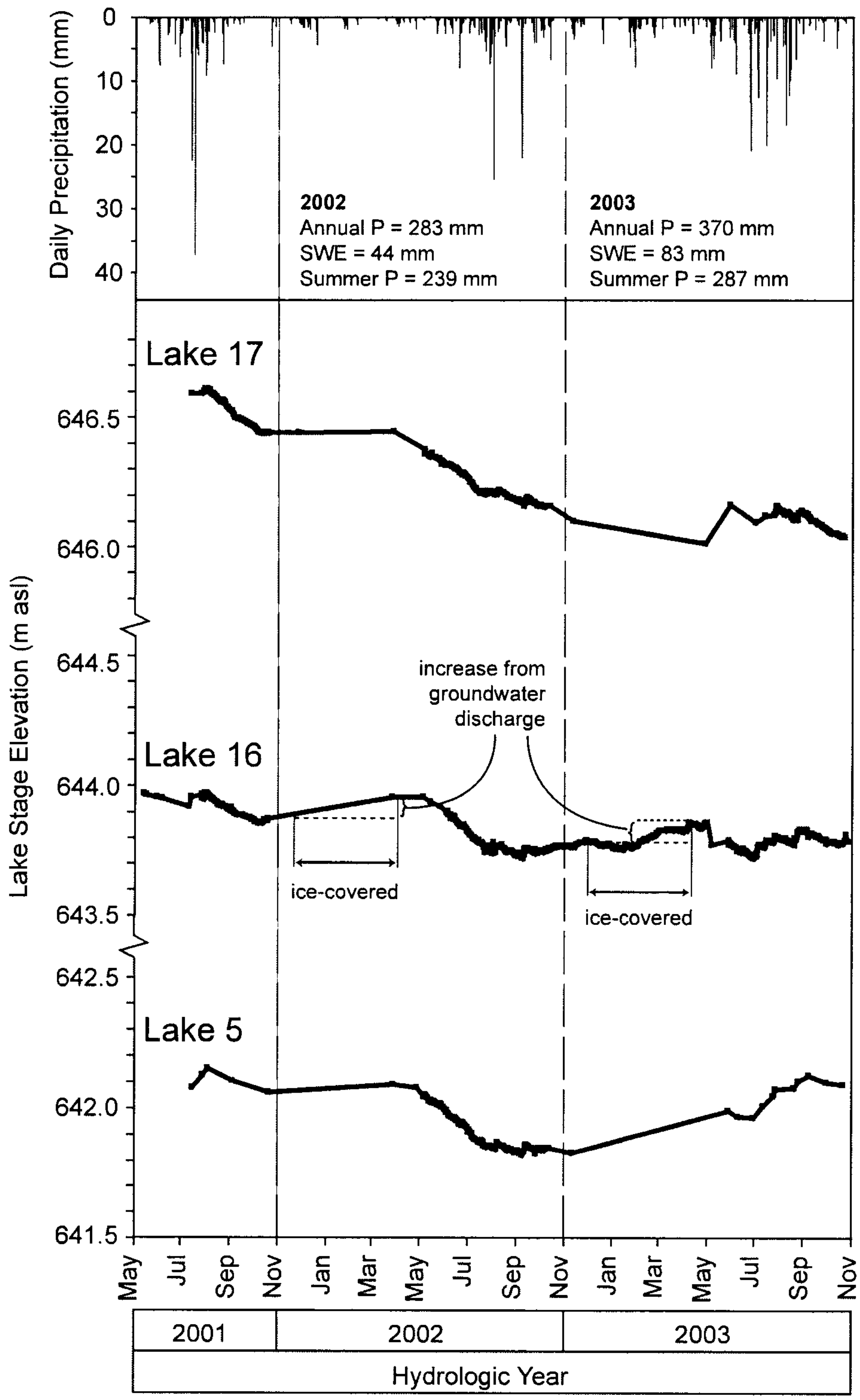

Figure 2.4. Daily precipitation and lake stages for duration of study. Time-series divided into hydrologic years with dashed lines. Increase due to net groundwater discharge to Lake 16 during winter months shown for 2002 and 2003.

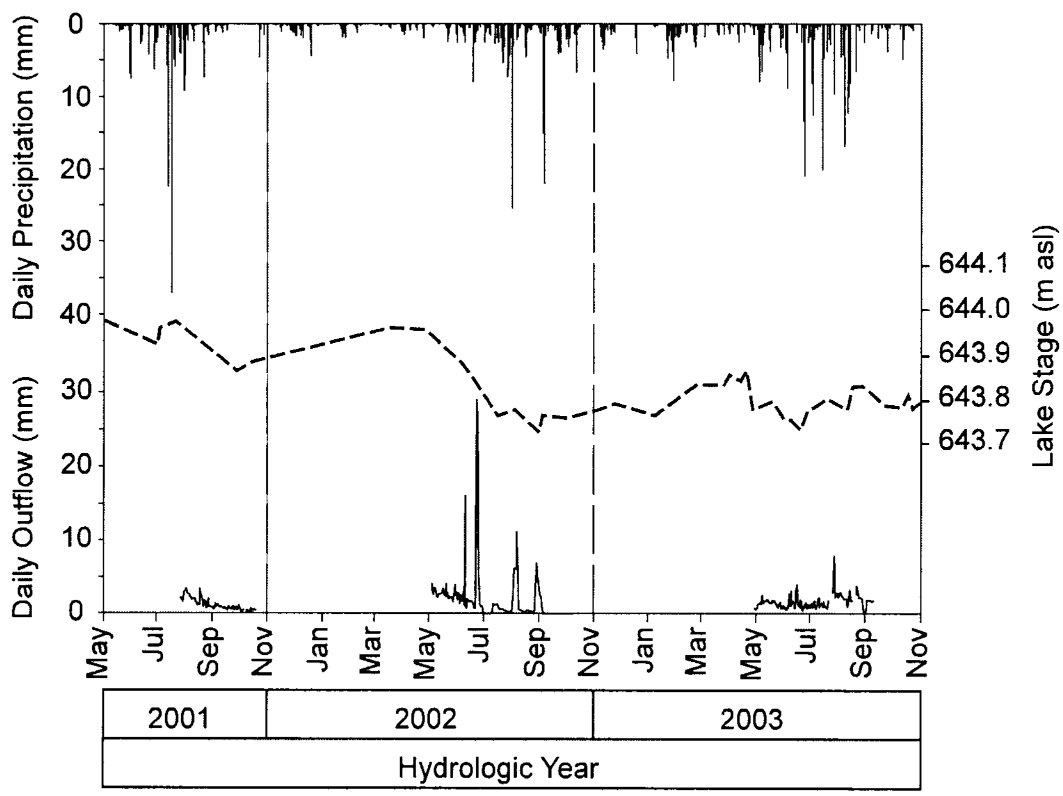

Figure 2.5. Daily surface water discharge (solid line), daily precipitation (shaded bars), and lake stage (dashed line).

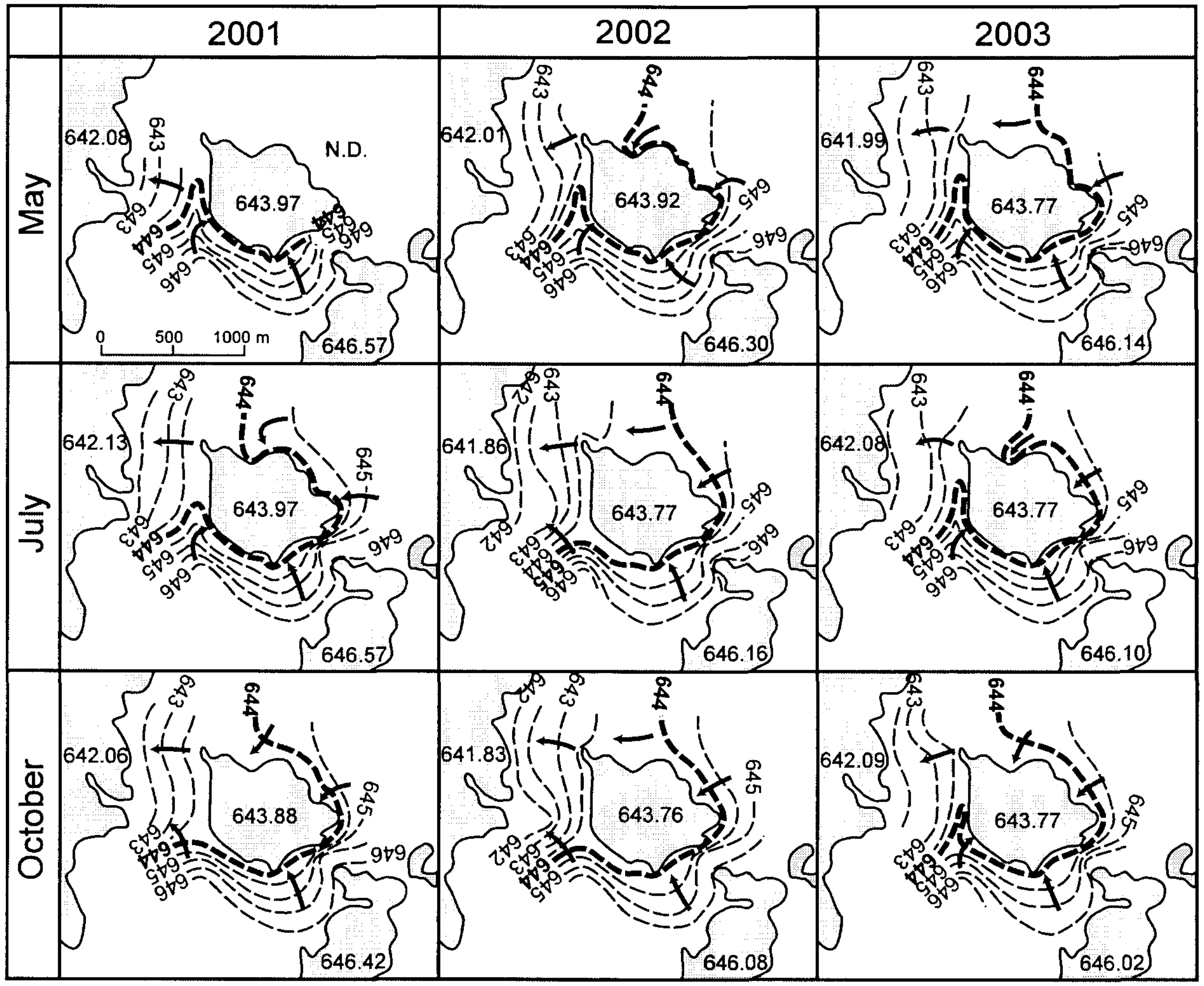

Figure 2.6. Water table contours (dashed lines), lake stages, and interpreted groundwater directions (arrows) for May, July, and October (0.5 m contour interval). Bold 664 m contour line illustrates transient variation in the capture zone of the lake.

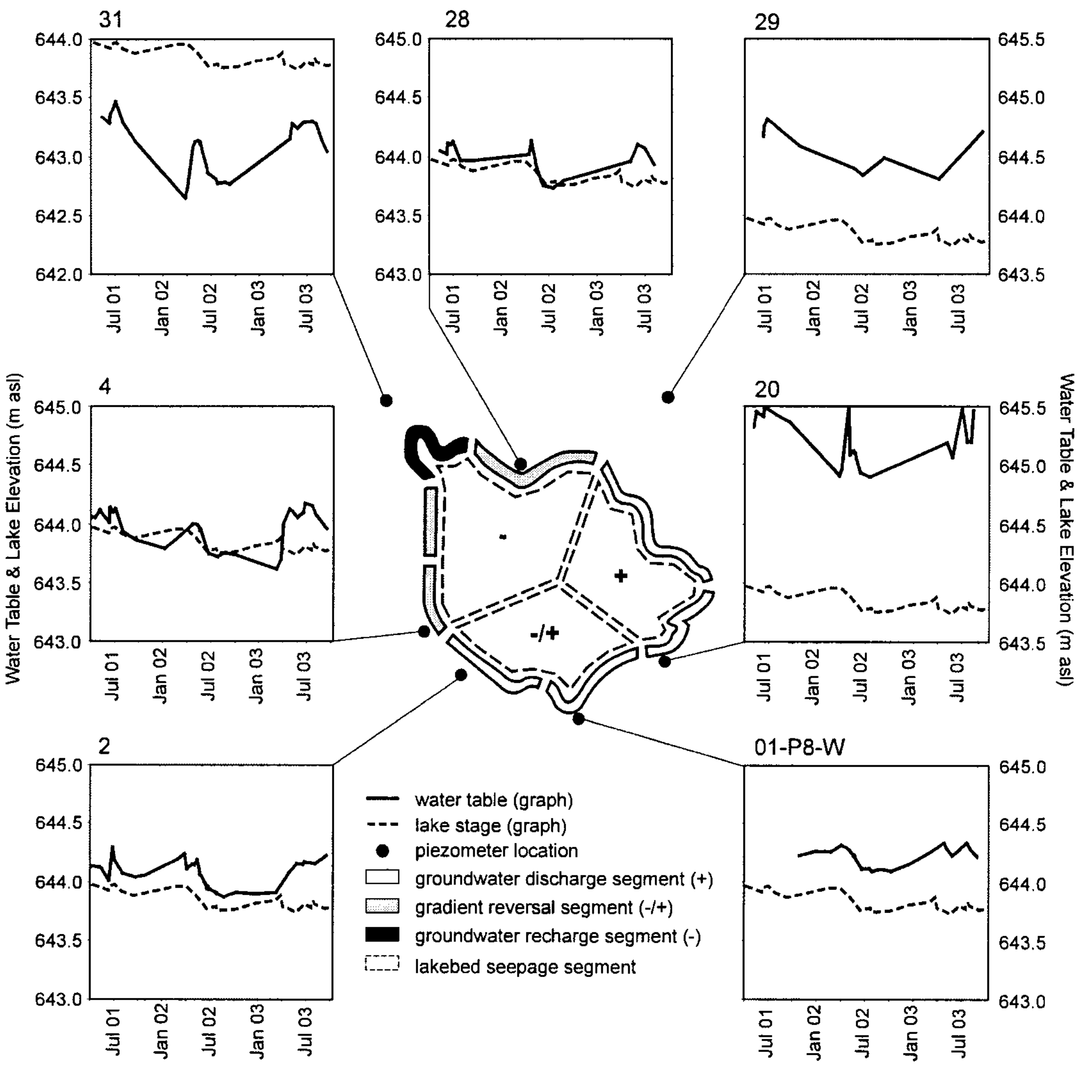

Figure 2.7. Groundwater interaction with Lake 16 for 8 shoreline segments, and 3 lakebed areas. Water table hydrographs shown for selected piezometers with Lake 16 stage, with consistent 2 m absolute water level range.

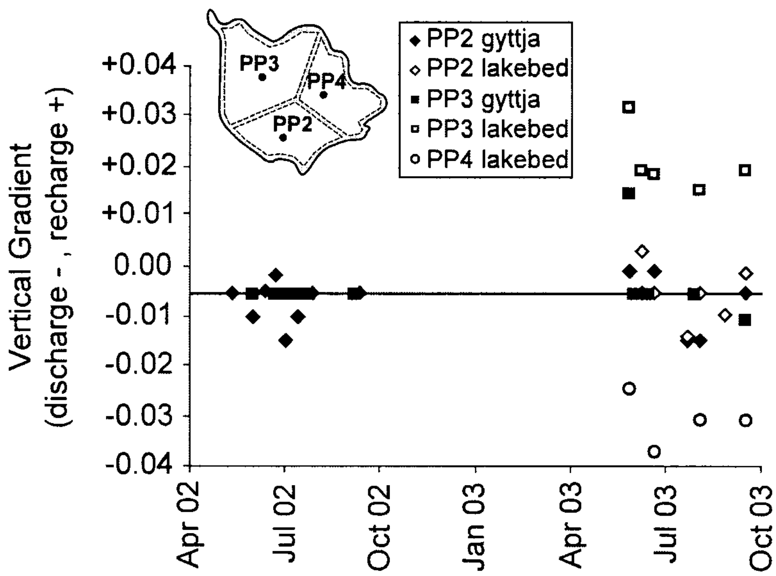

Figure 2.8. Vertical hydraulic head gradients (dimensionless) for piezometers completed below the base of the lake (locations shown on inset map). Groundwater recharge conditions plot above the zero gradient line; groundwater discharge conditions plot below.

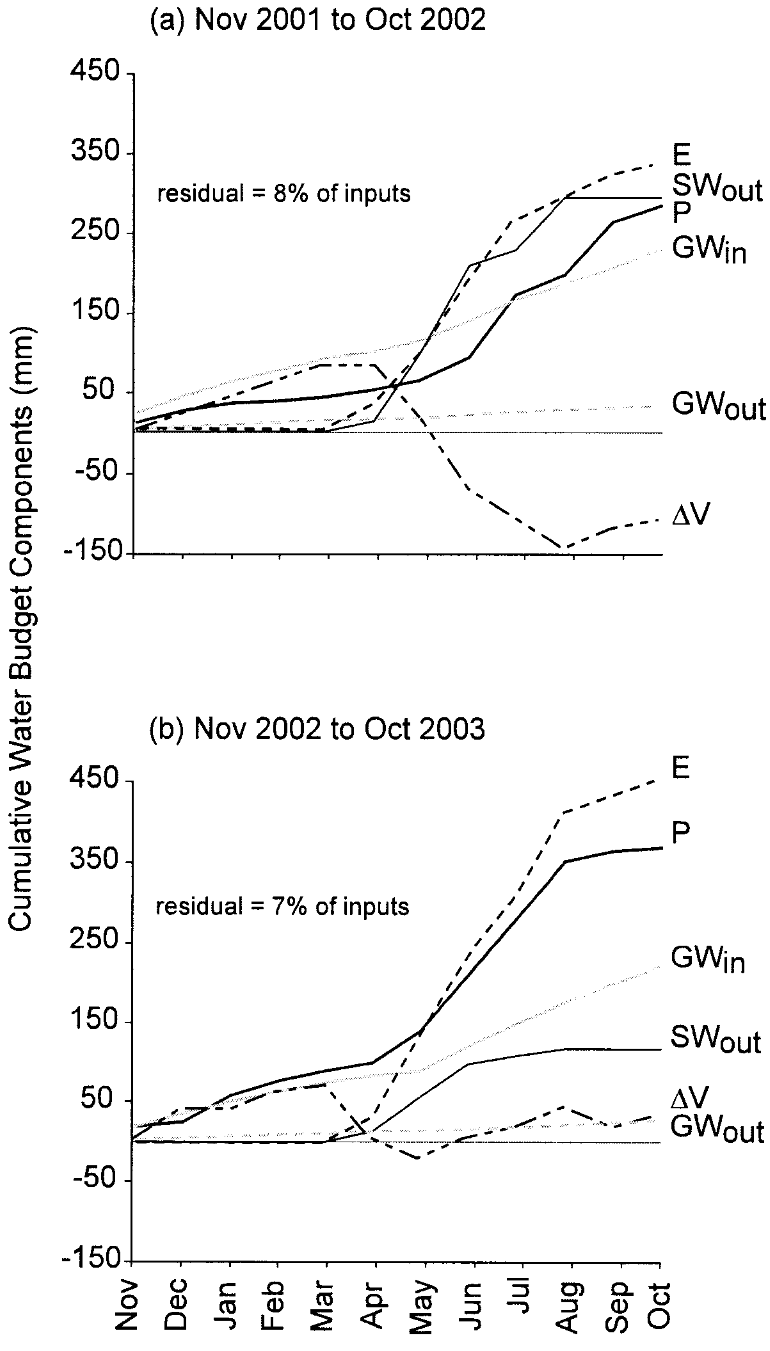

Figure 2.9. Monthly lake budget components for two hydrologic years. Fluxes have been plotted as cumulative positive values for comparison between each budget component. Cumulative change in lake storage (ΔV) has a negative slope and values because of loss of lake water throughout each year.

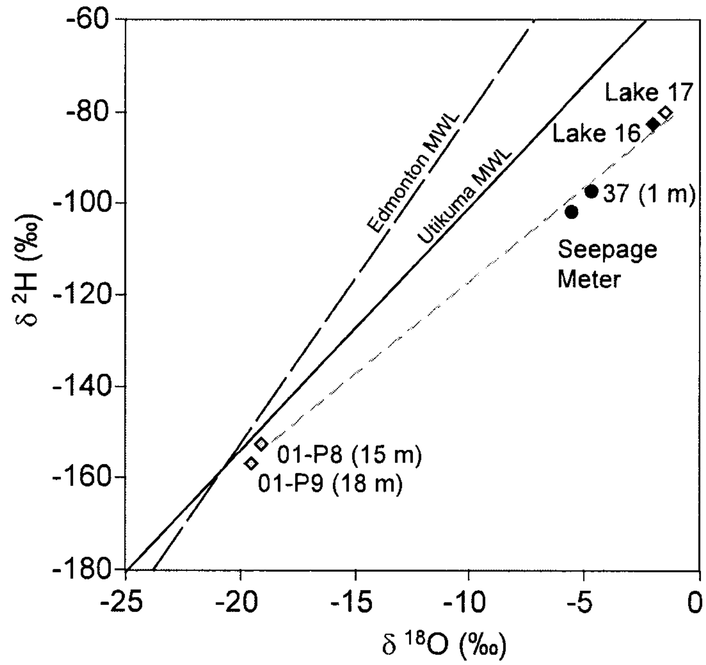

Figure 2.10. Stable isotopic ratios of $\delta^{18}O$ and $\delta^{2}H$ for autumn 2001: lake water, shallow groundwater discharge, and deep groundwater. Meteoric water lines for Edmonton, Alberta (Maulé et al., 1994) and the URSA (Devito, unpublished data) are shown.

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Chapter 3

Simulations of fully-coupled lake-groundwater exchange in a sub-humid climate with an integrated hydrologic model*

3.1 Introduction

Effective management of water resources on all landscapes requires an understanding of water cycling processes, which are controlled by climate, geology, and the interaction of surface water and groundwater (Devito et al., 2005a). Research at the Utikuma Research Study Area (URSA), on the Boreal Plains region of northern Alberta, Canada, reveals complex surface and groundwater interactions (Ferone and Devito, 2004; Smerdon et al., 2005), which are largely controlled by regional climate (Winter et al., 2005; York et al., 2002) and a heterogeneous landscape. Assessment of impacts to hydrologic systems on this landscape, from the rapid expansion of oil and gas exploration and production infrastructure, to exploitation from the forestry industry, requires models that can adequately represent integrated surface water/groundwater flow; a landscape dominated by widely distributed lakes, ponds and wetlands; and, a sub-humid climate. To simulate water cycling processes on the Boreal Plains, and to predict changes to surface water bodies for scenarios of landscape disturbance or variation in climate, spatially distributed models must also be inclusive of the ubiquitous peatlands, which greatly add to landscape heterogeneity and control surface water/groundwater interaction (Ferone and Devito, 2004).

The principles of simulating lake-groundwater interaction have been well documented by Winter (Winter, 1976; Winter, 1983), Townley (Smith and Townley, 2002; Townley and Trefry, 2000), and through development of the Lake Package for the MODFLOW groundwater model (described in Hunt et al., 2003). In this study, I seek to enhance modelling of lake-groundwater exchange further, by using a numerical model capable of determining the location and depth of surface water bodies (i.e., lakes and ponds) and surface-subsurface exchange fluxes, without prior definition (i.e., fully-

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coupled surface water/groundwater interaction). That is, I did not want to specifically define water bodies as hydraulic boundary conditions, but instead have lake location and stage determined as part of the simulation. This methodology is relatively new in hydrologic modelling, and has not been done for a lake- and wetland-dominated area with abundant peatlands and an irregular surface geometry. This fully-coupled approach allows lake-groundwater interaction to be investigated for a study lake at the URSA without defining the study lake as a boundary condition, thereby minimizing the amount of numerical intervention reflected in the solution of governing flow equations (Loague and VanderKwaak, 2004).

The objectives of this study were to simulate lake-groundwater interaction for a sub-humid hydrologic system on forested, glacial outwash terrain, and to determine the degree of spatially distributed atmospheric flux data (i.e., evapotranspiration and precipitation throughfall) and landscape heterogeneity needed to replicate field observations. These objectives were met through modelling groundwater conditions, lake levels, and lakebed seepage fluxes for an instrumented study lake at the URSA, for 2002 and 2003 (presented in Chapter 2). The findings represent an intermediate step toward modelling scenarios of future land use and climate change.

3.2 Study Area: URSA Lake 16

The Boreal region containing the URSA is characterized by a sub-humid climate (Ecoregions Working Group, 1989), marine shale bedrock (Upper Cretaceous Smoky Group: Hamilton et al., 1999) that is covered by 80 to 240 m of heterogeneous glacial sediments (Pawlowicz and Fenton, 2002), and low topographic relief. Average annual precipitation (P) and potential evapotranspiration (PET) are 481 mm (Environment Canada, 2003) and 518 mm (Bothe and Abraham, 1993), respectively; however, annual precipitation varies year to year (318 to 529 mm in 1997 to 2001; Devito et al., 2005b). The URSA is approximately 150 km south of the discontinuous permafrost zone (Woo and Winter, 1993), and has average monthly temperature that ranges from –14.6°C to 15.6°C (Environment Canada, 2003), with the winter season occurring from approximately November to April.

Lake 16 (39 ha) is located within the URSA, 370 km north of Edmonton, Alberta, Canada (56°6' N, 116°32' W; Figure 3.1), on a 200 km^2 coarse-textured glacial outwash plain, which is bounded by fine-textured sediments of a disintegration moraine (Fenton et al., 2003). In the vicinity of the study area, the outwash landscape is hummocky, with remnants of an east to west trending esker, and a maximum topographic relief of 20 m. Surface drainage is poorly developed, with no evidence of surface inflow to Lake 16, and ephemeral outflow through a narrow channel to a fen on the west side of Lake 16. The study area contains three lakes that exist in a series of 'steps', where Lake 16 was 1.8 m higher than Lake 5, and 2.4 m lower than Lake 17 in early 2002 (Figure 3.1). Groundwater springs occur along the southeast shore, adjacent to a small gravel pit, which was constructed by removing part of the esker deposit.

Measurements of hydraulic head at 70 piezometers, lake level at each of the 3 lakes, and groundwater seepage estimates from 8 lakeshore segments were made from April 2001 to October 2003 (Chapter 2). Hydraulic conductivity (K) was found to vary over six orders of magnitude (10^{-3} to 10^{-8} m/s), for 8 soil types (Figures 3.2 and 3.3). Annual P was 283 mm and 345 mm (including 44 mm and 83 mm of snow water equivalent), and annual lake evaporation was 336 mm and 455 mm for 2002 and 2003, respectively (Figure 3.4a). Over the summer of 2002, lake levels declined by 0.26 m, 0.18 m, and 0.34 m for Lakes 5, 16, and 17. Throughout 2003, Lake 5 increased by 0.14 m, and Lakes 16 and 17 remained at nearly a constant elevation.

Groundwater moved from southeast to northwest at an average horizontal hydraulic head gradient of 0.002, with the formation of small water table mounds (from snowmelt recharge) downgradient of the lake in spring months of each year (e.g., Anderson and Munter, 1981). Evaporation from the lake was the largest hydrologic flux each year, exposing captured groundwater to the atmosphere via evaporation windows (i.e., lakes), and causing water table hinge lines to be located downgradient from the lake (Chapter 2). This dynamic relationship between precipitation, groundwater interaction, and evaporation controls the degree of lake-groundwater exchange, and is expected to be very sensitive to regional climate conditions, which may be altered under future land use and climate change scenarios.

3.3 Numerical Simulations

Lake-groundwater exchange was simulated for a three-dimensional, 1800 m by 1800 m by 30 m (high) section of the outwash landscape, including Lake 16 (Figures 3.1 and 3.2). Two separate timeframes were considered: March 22 to November 2, 2002, when Lake 16 declined by 0.18 m, and April 25 to November 1, 2003, when the level of Lake 16 remained nearly stable. These periods coincide with abundant time-series field data and the water budget analysis in Chapter 2, but exclude the winter season, which would have additional cryospheric processes to consider. Results of simulated hydraulic heads, water table elevation, Lake 16 level, and seepage fluxes were compared to the field data. For the 2002 timeframe, field measured and/or assumed hydraulic input parameters were initially used, and were then adjusted slightly to better match simulation results with field observations (i.e., calibration); however, for the 2003 timeframe only new boundary conditions (based on field measurements) were specified to assess applicability of the model to represent field conditions without additional calibration (i.e., validation; Anderson and Woessner, 1992).

3.3.1 Lake 16 Flow Model

The Integrated Hydrology Model (InHM; VanderKwaak, 1999), was used to solve fully-coupled equations for variably-saturated porous media flow and overland flow, using a first-order flux relationship, as given by

$$-\nabla q \pm q^{b} \pm q^{e}_{ps} = \frac{\partial \phi S_w}{\partial t} \qquad [3.1]$$

$$-\nabla q_s \psi_s \pm a_s q^{b} \pm a_s q^{e}_{sp} = \frac{\partial (S_{ws} h_s + \psi_s^{store})}{\partial t} \qquad [3.2]$$

where q is Darcy flux, q^b is a specified boundary condition (i.e., source or sink within the porous medium or on the land surface), ϕ is porosity, S_w and S_{ws} are water saturation for the porous media and land surface, t is time, q_s is overland flow (implemented as a two-dimensional diffusive wave equation approximated with Manning's equation; see VanderKwaak 1999), ψ_s is depth of mobile surface water, ψ_s^{store} is depth of immobile surface water stored in centimetre-scale microtopography h_s. The exchange flux between the porous media and land surface is given by q^e, where the subscripts *ps* and *sp* denote flux from the porous media (p) to surface (s), or vice-versa, and a water exchange coefficient (a_s). Thus, the model for the URSA Lake 16 area simulated fully-coupled hydraulic head, water saturation, lake depth, and lakebed seepage for a three-dimensional, finite element mesh.

The finite element mesh used in this study was comprised of 243,000 subsurface prism elements and 16,200 triangular surface elements (i.e., the top of the prisms) that were uniformly spaced at 20 m horizontally, and variably spaced in the vertical direction. Elements had finer vertical spacing (0.25 m) to a depth of 1 m beneath the lake basin and within the expected range of water table depths. The topography of the surface of the mesh was constructed from a 1 m horizontal resolution digital elevation model of the land surface combined with lake bathymetry measurements (Chapter 2), which were smoothed using a Spline algorithm, to 20 m horizontal resolution to reduce the complexity of the surface features. The bottom surface of the mesh sloped from east to west (Figure 3.2b), corresponding to the base of the glacial outwash as determined from geophysical field investigations (Domes, 2004).

Subsurface heterogeneity of the glacial outwash was represented by 8 zones in the model, the parameters of which were based on field investigation. Each zone was specified with a hydraulic conductivity (K) value measured in the field and estimates of specific storage (S_s), ϕ, and ratio of horizontal to vertical hydraulic conductivity (i.e., K_{xy}:K_z; Figure 3.3a). Within the lake basin, the bathymetry (i.e., top of model domain) corresponds to the top of the gyttja, thereby allowing fluid flow, and more importantly, lakebed seepage flux, to be simulated through organic and mineral soil layers present in the lake basin (Squires et al., 2006). Gyttja and lakebed sediment K decreased with depth, such that the gyttja more easily transmits water than lakebed mineral sediments. Tabulated relationships for capillary pressure, water saturation, and relative permeability were required for 3 zones expected to be variably saturated (Figures 3.3b and 3.3c), and were based on published data of similar soil texture: outwash sand (Borden sand: Abdul, 1985); silty sand (Carsel and Parrish, 1988); and peat (Silins and Rothwell, 1998). Manning's n values used in equation 3.2 were assumed to be 0.04 for the narrow stream outlet on the west side of the lake (Dingman, 2002), and 0.4 for the remainder of the ground surface (Woolhiser, 1975). Microtopography (h_s) and mobile water depth were each assumed to be 10^{-2} m, and the surface-coupling coefficient (a_s) was set to 10^{-4}.

3.3.2 Boundary Conditions

Fluid flow boundary conditions include time-varying specified hydraulic heads of Lakes 5 and 17 (Figures 3.1 and 3.4c) along the east and west edges of the mesh (reported in Chapter 2), and a specified hydraulic head of 640 m within the lower sand zone (Figure 3.2b) to allow subsurface connection with the larger-scale, regional groundwater flow system existing beneath the west of the study area. The mesh has been oriented in the direction of groundwater flow; therefore, the remaining edges parallel to groundwater flow were specified as no-flow boundaries. The bottom of the mesh has also been set as a no-flow boundary, because it corresponds with the base of the outwash deposits, which were underlain by clay and have significantly lower K than the outwash sand.

Daily precipitation and evaporation was applied to the top of the model domain as time-series, net fluid fluxes (i.e., the sum of positive precipitation fluxes and negative evaporative fluxes). For areas of the ground surface that were saturated or that had surface water, and were not covered by a forest (e.g., fen and lake), specified net fluxes corresponded to the precipitation and open water evaporative fluxes (Figure 3.4a) reported in Chapter 2. Net fluxes specified for the gravel pit area were assumed to consist of precipitation only (i.e., no evaporative flux), because the area is un-vegetated. For the remaining forested land surfaces, daily net fluxes (Figure 3.4b) accounted for interception and evapotranspiration, which were dependent on the time of year. Forest interception and actual evapotranspiration were based on hydrologic research at another site in the Western Boreal Forest (Price et al. 1997). From March to June 2002, and October 2002 in the first simulated year, and April to June, and October 2003 in the second simulated year, 100% of daily P measured at the URSA was assumed for the forested areas, to account for limited interception and evapotranspiration. For the summer months (i.e., June to August) the net flux was assumed to be 80% of daily P greater than 15 mm, and for September net flux was assumed to be 95% of daily P greater than 5 mm. These net fluxes are lower than observed precipitation throughfall rates at the

URSA (85% to 95% of P: Devito, unpublished data), in order to account for canopy interception and actual evapotranspiration, which could not be simulated explicitly for unsaturated subsurface (i.e., porous media) nodes (Loague et al. 2005).

3.3.3 Initial Conditions

Initial conditions were required for hydraulic head, water saturation, and lake depth for each node in the model domain, and were determined by assigning the initial hydraulic heads of Lake 5 and Lake 17 to those measured in the spring of either 2002 or 2003, and running the model to steady state. For these steady state simulations, the average sum of atmospheric fluxes (i.e., precipitation and evapotranspiration) was assumed to be zero. The resulting hydraulic heads and lake depths were compared to March 22, 2002 and April 25, 2003 field observations, and subsequently input as initial conditions for each of the transient simulations.

3.4 Results

3.4.1 Steady State Initial Conditions

As the simulation converged from an initially saturated condition, to the steady state solution, the lake level came into equilibrium conditions with the simulated groundwater flow field (Figure 3.5). At steady state, the water table configuration, and level of Lake 16, were similar to field observations for March 2002. For 19 wells/piezometers distributed around the study site and Lake 16 (Figure 3.2a.), the root mean squared error (RMSE) was 0.29 m, which is on the order of observed water table and lake level fluctuations. Similar results were obtained for the initial conditions of April 2003 (RMSE = 0.30 m).

The simulated flow system adequately represented the flow-through interaction between the lake and groundwater, as well as the large groundwater capture zone (Figure 3.5). These initial conditions were sensitive to K and anisotropy of the outwash sand, the dominant zone of porous media. For the field-measured outwash K, an assumed anisotropy of 10:1 provided the most representative results. Variation of the anisotropy ratio to 1:1 or 100:1 (Figure 3.6a), or variation of K by half an order of magnitude, resulted in an incorrect regional water table gradient and lake level.

3.4.2 Comparison of Transient Hydraulic Heads

Simulated hydraulic head values for 19 observation locations around the lake, and the computed elevation of Lake 16, matched most of the time-series data from corresponding field observation points (Figure 3.7). For the riparian peatland zone directly adjacent to Lake 16 (2B, 14, 20, 23, 28, 29, 43 on Figure 3.7), the general trend in hydraulic head response appears to mimic field observations, except during summer months when simulated and observed hydraulic head data were out of phase. Considering that hydraulic heads are represented well for the spring and autumn months, and that the general trend of hydraulic head decrease is similar to field observations, I suspected that an additional transient storage mechanism is operating in the field that is not represented in the simulation. At a location downgradient of Lake 16, water table response was not as well represented in the simulation as is observed in the field (1; Figure 3.7), and improved characterization of the unsaturated hydraulic parameters in this porous media zone is likely warranted.

The RMSE between simulated and observed hydraulic heads were calculated for 3 times in each simulation (May, July, and October). RMSE varied from 0.24 to 0.31 m for the 2002 data, and from 0.21 to 0.42 m for the 2003 data (Figure 3.8). In each simulation, the RMSE increased by approximately 0.1 m for the July period, suggesting the occurrence of a mid-summer phenomenon occurring in the field that might not be replicated suitably in the numerical model.

3.4.3 Water Table Configuration

Simulated water table contours in the vicinity of Lake 16 were plotted for spring (May), summer (July), and autumn (October) months of 2002 and 2003 (Figure 3.9). As anticipated from the formulation of this boundary-value-problem, groundwater flow is from southeast to northwest (Lake 17 to Lake 5) across the model domain, with various degrees of interaction between the surface water and groundwater. Because of the dominance of evaporation from the lake, the water table contours wrap more than half way around Lake 16, resulting in hinge lines that point in the downgradient direction, indicative of the lake's subsurface capture zone (Gosselin and Khisty, 2001; Townley and

Davidson, 1988). The general shape of the water table compares favourably with the water table maps presented in Chapter 2 for the same time periods.

In the 2002 simulation, the lakes on the model domain boundary (Lakes 5 and 17) decline due to regional drought conditions (Figure 3.4c). As expected, the simulated water table configuration adjusts as the level of Lake 16 decreased, shown by the position of the 643.5 m asl contour on Figures 3.9a and 3.9b. In response to 2 days of heavy precipitation, late in September 2002, the elevation of Lake 16 increased by 0.05 m, and the position of the 643.5 m asl contour was relocated downgradient of the lake basin. In the 2003 simulation, the balance between fluxes of groundwater discharge and evaporation maintain the elevation of Lake 16 to a more consistent level than observed in 2002 (Figure 3.7), with nearly a steady water table configuration (Figure 3.9).

3.4.4 Lake-Groundwater Seepage

Simulated exchange fluxes, which were calculated between the subsurface and surface elements in the numerical model, represent lakebed seepage for the Lake 16 area within the model domain (Figure 3.9). Groundwater discharge measurements from 3 seepage meters installed along the southeast lakeshore varied from 2.4×10^{-6} to 5×10^{-8} m/s, with a geometric mean of 6.3×10^{-7} m/s for the summer months of 2002 and 2003 (Chapter 2). Simulated seepage fluxes at corresponding locations within the flow model were 2.5×10^{-7} m/s (Figure 3.10). Although the simulated seepage varied over a narrower range than either of the seepage estimates reported in Chapter 2, the magnitude of groundwater discharge and spatial pattern of groundwater recharge and discharge across the lakebed, honour field observations.

Shaded contours of lakebed seepage (Figure 3.9) indicate that groundwater discharge occurs over at least 75% of the lakebed area. Throughout the summer of 2002, and between 2002 and 2003, the fraction of the lakebed area that discharged groundwater to the lake basin increased from 0.75 to 0.83, and is associated with a noticeable movement in the hinge line between discharge and recharge fluxes across the lakebed (Figure 3.9).

3.5 Discussion

3.5.1 Landscape Heterogeneity

Eight zones of heterogeneity were required to represent the hydrologic flow regime of Lake 16. The broad-scale flow system was adequately defined by the hydraulic parameters of the most predominant zone of porous media (Winter and Pfannkuch, 1984), which in these simulations was the 'bulk' outwash K. The outwash sediments at the URSA were predominately well sorted fine sand, with a narrow range in approximate pore throat diameter and average grain size (d_{10} and d_{60} of 0.09 and 0.20 mm). Replicating the field observations of initial conditions and the transient response of Lake 16 and hydraulic heads within the surrounding outwash, required an anisotropy ratio of 10:1, which was an order of magnitude lower than values reported by (Winter and Pfannkuch, 1984) for an outwash field site. For the coupled lake-groundwater model developed in this study, a higher anisotropy ratio for the outwash sediments (100:1) resulted in an underestimate of hydraulic heads in 2002, and isotropic conditions resulted in overprediction of hydraulic heads (Figure 3.6b). The narrow range in grain size distributions reported for the URSA outwash sediments (Chapter 2), with low fractions of silt and clay, corroborate the low anisotropy (10:1) used in this study is representative of the predominant porous medium. Furthermore, Weeks (1969) measured aquifer anisotropy at an outwash deposit in Wisconsin, U.S.A., and found that it ranged from 5:1 to 40:1, providing additional corroboration of the 10:1 anisotropy ratio determined from this study.

3.5.2 Riparian Peatlands

Lake-groundwater interaction is further controlled by the presence of riparian peatlands at the margin of the lake and the stratigraphic sequence of lakebed deposits (upper and lower gyttja overlying fine grained mineral sediments). These heterogeneous features on the outwash landscape were required for the accurate simulation of Lake 16's elevation and the surrounding hydraulic heads, especially when transient responses were considered. The most sensitive of these zones of porous medium were the riparian peatlands. Variation of the peatland K by a factor of 5 had a large influence on the transient response of the lake elevation, the water table configuration adjacent to the lake, and nearshore lakebed seepage fluxes. In effect, riparian peatland K governs the hydraulic connection between the lake and adjacent groundwater regime. If the peatland K were increased slightly (i.e., having a lower contrast with the adjacent outwash sediments), the flow system was more efficiently connected laterally. The net effect was enhanced lateral fluid exchange with groundwater at the lake margin, which created more groundwater discharge along the lakeshore than observed in the field. With a slightly lower peatland K, the lake became more isolated from the transience of the outwash groundwater system, and began to appear more hydraulically disconnected. Thus, higher K riparian peatlands provide lower buffering ability or hydraulic storage capacity between the groundwater regime and the lake. Limited hydraulic storage capacity will have significant influence on biogeochemical reactions (Devito et al., 2000, Gibbons, 2005). High K riparian peatlands will promote nutrient movement between aquatic (i.e., lake) and groundwater zones, which are different biogeochemical environments.

When field observations of hydraulic head response were compared with transient modelled response (observation points with shaded boxes on Figure 3.7), it appears that the value of peatland K and specific storage (S_s) could be transient within a seasonal timeframe, as the riparian peat material transitions from a frozen to thawed state in the mid-summer. Considering that the riparian peatlands remain nearly water saturated continuously, below freezing air temperatures in the winter months (November to April) will cause these porous materials to freeze in solid form, with reduced permeability. Field observations and preliminary lab study of the peatland state (Smerdon, unpublished data) corroborate the timing of a transition from frozen to thawed states, and further study of the effects of seasonal thermal effects on water transmission properties (e.g., McCauley et al., 2002) is warranted.

The peatland zones were parameterized with relatively high water retention characteristics determined from a different peatland in the same climatic region (Silins and Rothwell, 1998). These water retention characteristics, high water saturation, shallow water table depth (i.e., nearly at surface), and close proximity to the lake, allow the peatland materials to retain water during regional water deficit conditions. Although the peatland material has been simulated as a traditional porous medium in this study, (Price, 2003) showed that peatlands will maintain saturation through a complex selfpreservation feedback mechanism that utilizes vertical compression during drought, and flotation when conditions are more favourable for sphagnum growth. The presence of peatland material around most of Lake 16 created a potential for water preservation on the outwash landscape. When the sub-humid climate is considered, these peatland features are responsible for buffering lake-groundwater exchange, and are a critical feature for maintenance of wetlands, ponds and lakes on a permeable outwash landscape.

3.5.3 Lake-Groundwater Seepage

Findings from other studies of lake-groundwater interaction, including Winter (1976, 1983), Cheng and Anderson (1994) and Townley and Trefry (2000), have shown that lakebed seepage patterns are largely controlled by position of a lake within a groundwater flow system, lakebed slope, and lakebed sediment K. However, the magnitude of seepage is more often controlled by regional hydraulic head gradient (Smith and Townley, 2002) and aquifer anisotropy (Winter and Pfannkuch, 1984). In this study, many of these controlling factors (such as the groundwater flow system and lakebed sediment K) were constrained by field data. Spatial distribution of seepage at Lake 16 is similar to the seepage patterns predicted by Townley and Trefry (2000) for lakes in a flow-through groundwater regime, and the magnitude of seepage flux was limited by the stratigraphy of lakebed deposits (gyttja overlying mineral sediments). Discrepancy between groundwater discharge measured in the field, and seepage flux predicted in the simulations (Figure 3.10), is within half an order of magnitude, and any subtle difference is attributed to the amount of spatial lakebed heterogeneity not represented in the model. Discrete field measurements capture the range of seepage through the lakebed at specific locations, representing lateral heterogeneity (Kishel and Gerla, 2002; Rosenberry, 2005) not specified in the model, whereas the numerical model produces an average result.

Previous numerical studies of lake-groundwater seepage have often described hypothetical, circular lakes, in homogeneous aquifers (e.g., Genereux and Bandopadhyay, 2001; Townley and Trefry, 2000), under steady state flow conditions. Field observations at Lake 16 (Chapter 2; Smerdon et al., 2005) and the present simulations illustrate that there is a transient interaction between lakes and the outwash groundwater flow system at the URSA, which is a common hydrologic phenomenon for lake-groundwater settings (e.g., Anderson and Cheng, 1993; Gosselin and Khisty, 2001). The fully-coupled transient solution of hydraulic heads and lakebed seepage fluxes were determined in these simulations simultaneously, which allowed for a more realistic representation of the hydrodynamics of lake-groundwater exchange than previously reported (e.g., Genereux and Bandopadhyay, 2001). The finite element formulation of the InHM is suitable for areas of irregular topography that must be considered in simulation, such as lakebed bathymetry, and for areas of varying seepage face locations (Romano et al., 1999), such as the dynamic lakeshore boundary. At the interface of the lake and the land surface, groundwater, surface water, and lakebed slope control the physics of water flow, and thus, lake-groundwater interaction (Winter, 1981). These have all been explicitly considered in this model.

3.5.4 Climatic Controls on Lake-Groundwater Exchange

Replication of the transient water table configuration, lake level and lakebed seepage, supports the hypothesis that shallow lakes act as evaporation windows on the Boreal Plains landscape. The observed lake level decline of 0.18 m in 2002 was caused by water deficit conditions imposed by the sub-humid atmosphere. Although the 2002 drought occurred only over one season, the rate of lake level decline (0.2 m/yr) was the same as observed during a severe (4 yr) drought at the glacial outwash Williams Lake in Minnesota, USA (Winter et al., 2005). The similarities in lake level decline between Lake 16 and Williams Lake offers the opportunity to elucidate hydrodynamic response of the lake-groundwater system at Lake 16 to drought conditions that could develop in many glaciated regions in North America. With a reduction in the volume of Lake 16, there was a corresponding increase in lakebed area contributing groundwater discharge (10% increase in seepage area) in the following year, which contributed groundwater inflow to the lake basin. Although simulated groundwater discharge fluxes were nearly constant throughout the simulation timeframes, the reconfiguration of the water table contours and decrease in groundwater outflow from the lake illustrate the role of regional climate on lake-groundwater interaction in this sub-humid climatic zone.

The sensitivity analysis by Genereux and Bandopadhyay (2001) of lakebed seepage suggests that the seepage patterns observed for Lake 16 (i.e., strong exponential decrease with increased distance from shore) are predominantly found in lakes with steeper lakebeds. The shallow slope of Lake 16's bed (0.012) appears to have similar patterns as those lakes with moderate (0.013) to steep (0.02) slopes in Genereux and Bandopadhyay (2001), suggesting that a large outflow of water from Lake 16 exists, and that Lake 16 is being replenished by the groundwater system. Simulation results and the corroborating field data indicate that there is minimal to no surface outflow and diminished groundwater outflow from Lake 16 (Figure 3.9) for 2002 and 2003. Therefore, given the shallow lakebed slope, seepage pattern indicative of a lake that would typically have a steeper slope, and field observations of minimal fluid outflow, the seepage pattern also confirms a dominance of summer evaporation (i.e., outflow via evaporation) and that climate has a significant influence on lake-groundwater interaction.

Lenters et al. (2005) suggest that it is critical to understand "why" lake levels change in response to climate variation for effective water resource management. For landscapes composed of high permeability lithology, lake level changes will be an integrated response of atmospheric conditions (precipitation and evaporation), groundwater interaction, and overland flow. In this study, water cycling for a coupled lake-groundwater system was simulated, and results compared favourably with transient observations from an instrumented field site. For coarse-textured landscapes in sub-humid climates, lake level changes are controlled by regional climate, and are further maintained by an active groundwater flow regime. The efficiency at which groundwater can maintain lakes, ponds and wetlands on this landscape, is controlled by the presence of landscape features that have high water retention, such as riparian peatlands, and by the duration of active lake evaporation (i.e., length of season). For shallow lakes in Northern Canada (which cover 37% of the landscape), Rouse et al. (2005) found that the lake evaporation season was 22 weeks per year, and that shallow lakes have a higher heat capacity and can partition water to the atmosphere more readily than wetlands. In these simulations, evaporation was applied to areas where the water table is at the land surface (similar to York et al., 2002), for 22.8 and 23.3 weeks in 2002 and 2003, respectively. Although the lake level decline occurred in 2002, due largely to a regional drought that also lowered adjacent Lakes 5 and 17, it appears that a lowered lake level and increased water deficit condition (110 mm for 2003, compared to 53 mm for 2002) caused increased groundwater inflow.

3.5.5 Representing Atmospheric Boundary Conditions

Lake level and hydraulic head response of the outwash groundwater flow systems were sensitive to the timing of the net atmospheric fluxes (described in section 3.3.2). Although actual evapotranspiration fluxes from the forested uplands were not explicitly simulated in this work, distinct periods of assumed boundary condition data (i.e., net fluxes that accounted for interception and evapotranspiration) had to be specified to accurately match field observations. The timing of these periods correspond to: spring (prior to June), when accumulated snowfall would melt and recharge the subsurface; summer (June to September), when only relatively large precipitation events (80% of daily P greater than 15 mm) were assumed to enter the subsurface; and, fall (from September to the end of each simulation), when reduced forest evapotranspiration and the loss of leaves would allow for more precipitation to reach the subsurface than summer months (95% of daily P greater than 5 mm). Although a good transient water table and lake level response was simulated, based on reasonable assumptions of upland forest water partitioning, future work should explicitly consider throughfall (as net precipitation reaching the ground surface) and actual evapotranspiration fluxes in a sub-humid climate, and is presently being studied by the Western Boreal Hydrology Group at the University of Alberta. Improved representation of the atmospheric boundary conditions could be applied to a larger-scale model that extends to natural watershed boundaries, and be used to investigate scenario simulations of landscape disturbances and changes in climate.

3.6 Conclusions

For lake-dominated hydrologic systems, many different styles of models have been developed to investigate lake-groundwater interaction. This study investigated simulation of lake-groundwater interaction through use of a fully-coupled, finite element, surface water/groundwater model to represent the hydrologic regime of a glacial outwash setting in a sub-humid environment. Formulation of the numerical model allowed simultaneous simulation of subsurface hydraulic heads and surface water depths, and the corresponding exchange fluxes between the surface and the subsurface, without assuming any explicit hydraulic exchange fluxes a-priori. Replication of three-dimensional, transient fluid flow, for a study area in northern Alberta, Canada, relied on accurately representing the "bulk" porous media with anisotropy that suitably represented structure of the porous media, and two key landscape heterogeneities that control lake-groundwater interaction: riparian peatlands and stratified lakebed deposits. To mimic time series field measurements, a lower anisotropy ratio (10:1) for glacial outwash was required than previously suggested for groundwater flow-through lake environments. The inclusion of riparian peatlands was found to govern lake-groundwater fluid exchange, and the peatlands appear to have seasonally transient hydraulic parameters. I hypothesize seasonal variation of peatland K values (i.e., higher K when thawed and lower K when frozen) as a mechanism for maintaining water in the sub-humid, Boreal environment. Field observations and preliminary lab study of the peatland state corroborate the timing of such a transition (i.e., from frozen to thawed states), and further study of the effects of seasonal thermal effects on lake-groundwater exchange is warranted.

Simulated patterns of lakebed seepage for Lake 16 confirm predictions made by others, for hypothetical lakes of more simplified geometry. The magnitude and spatial distribution of seepage fluxes computed in these simulations compare favourably with field observations and, when combined with calibrated hydraulic head data, provide further corroboration that the model for the lake in this study represented the hydrologic regime adequately. The transient flow system is largely driven by high summer lake evaporation, which caused reconfiguration of water table contours around the lake basin and increased lakebed seepage following lake level decline. Simplified atmospheric boundary conditions were assumed for the forested upland areas, indicating that three distinct timeframes (spring, summer and fall) are required in simulations, which correspond to seasonal variation of throughfall and evapotranspiration. Considering the dominance of evaporation from lakes and shallow water table areas in this landscape, further investigation into the relationship between temporal throughfall and evapotranspiration fluid fluxes is needed, to establish the rate and timing of groundwater recharge. In a lake-dominated landscape that appears to be controlled by evaporation in summer months, the timing of groundwater recharge is likely also an important controlling factor for maintenance of lakes and wetlands, which requires further investigation to more fully understand water cycling processes in this region.

This study illustrated that InHM can be applied to lake-dominated hydrologic systems, and used to investigate landscape and atmospheric controls on hydrologic processes. For larger areas, specific hydrologic landscapes (Winter, 2001), and longer-term applications (i.e., landscape management and reclamation), the modelling framework presented here would be appropriate, because it is not hampered by excessive numerical intervention (i.e., minimizing a-priori assumptions). The physical boundaries of the model domain need not be surface catchments or watersheds (Devito et al., 2005a; Winter et al., 2003); however, successful application will depend greatly on the ability to define spatially variable, subsurface hydraulic properties. Thus, robust models such as this will provide insight for investigating response of hydrologic systems in areas where anthropogenic changes (imposed by landscape disturbance or variation in climate) might be masked or subdued by natural variation in water cycling.

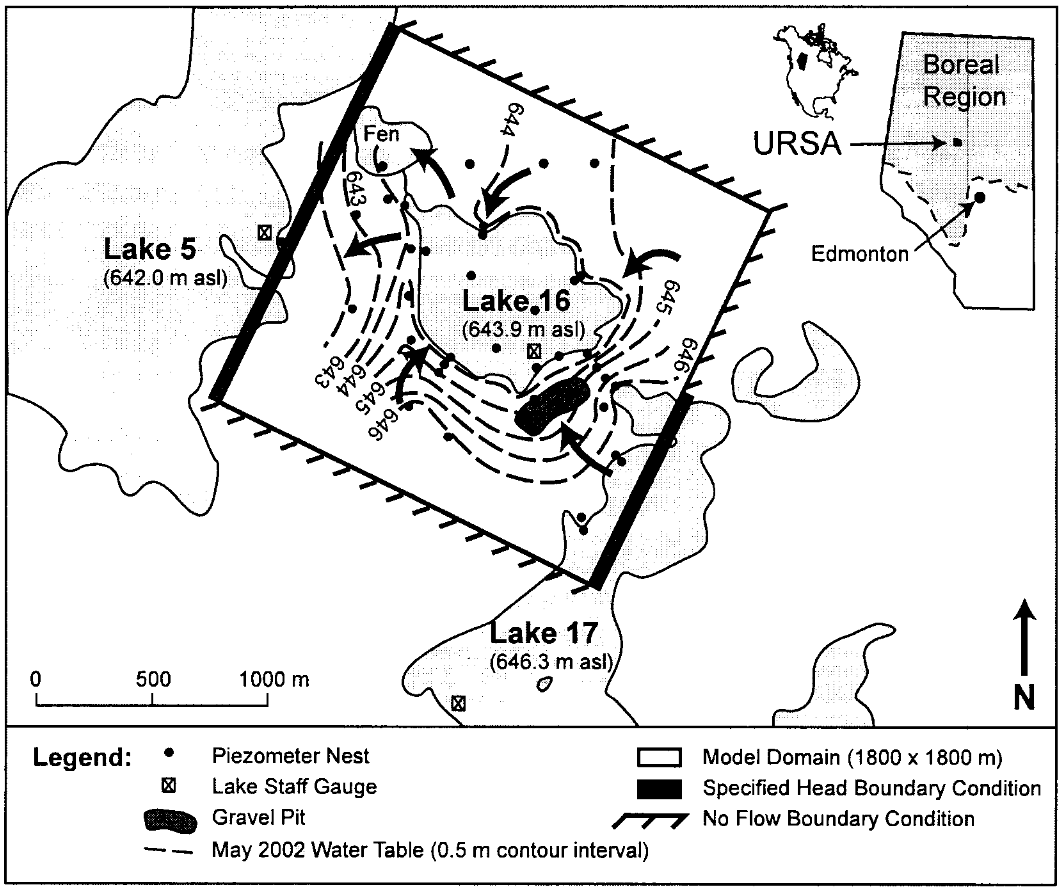


Figure 3.1. Lake 16 study site with selected field instrumentation, numerical model domain, and boundary conditions. Arrows indicate groundwater flow direction. URSA and Boreal Plains region in Alberta, Canada on inset.

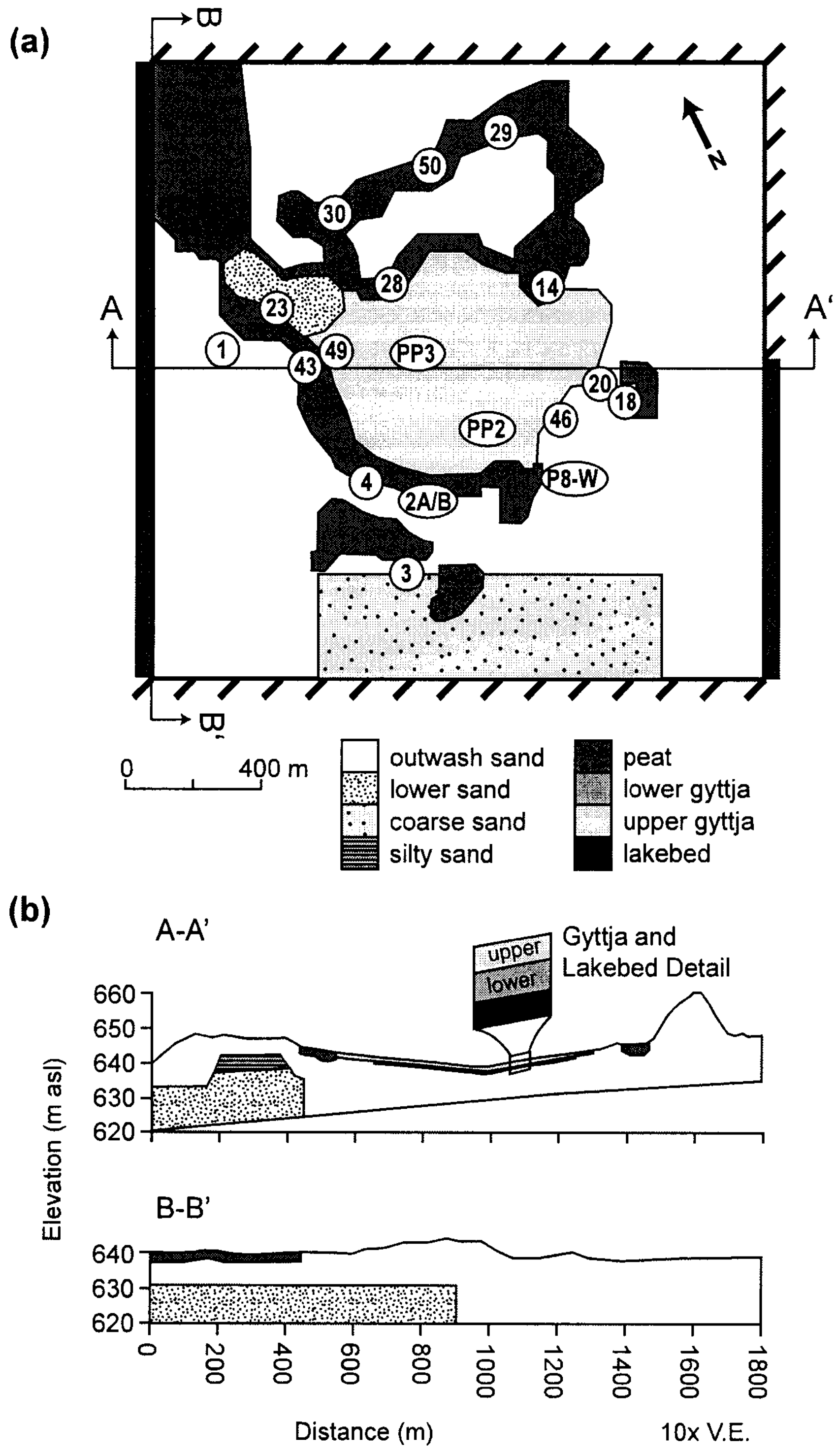


Figure 3.2. (a) Plan view and (b) cross sections of model domain, showing porous media zones (shaded), observation points (circled), and boundary conditions from Figure 3.1. Details of upper and lower gyttja and mineral lakebed zones shown on inset on (b).

(a)

Zone	K (m/s)	$K_{XY}:K_Z$	S_S (m^{-1})	ϕ
outwash sand	$1x10^{-5}$	10:1	10^{-5}	0.35
lower sand	$5x10^{-5}$	1:1	10^{-5}	0.35
coarse sand	$1x10^{-4}$	1:1	10^{-6}	0.35
silty sand	$2x10^{-6}$	50:1	10^{-5}	0.35
lakebed	$1x10^{-8}$	1:1	10^{-4}	0.45
upper gyttja	$3x10^{-6}$	1:1	10^{-4}	0.45
lower gyttja	$1x10^{-7}$	1:1	10^{-4}	0.45
peat	$3x10^{-6}$	1:1	10^{-4}	0.90

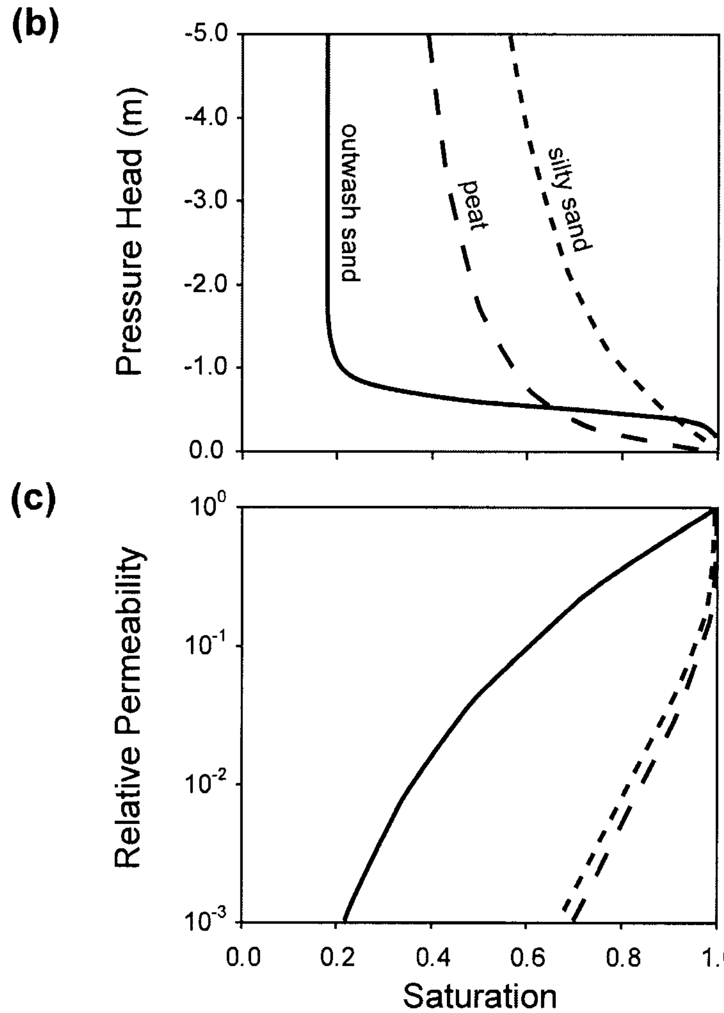


Figure 3.3. (a) Hydraulic parameters for each porous media zone, and (b) pressure head and (c) relative permeability relationships for variably-saturated zones.

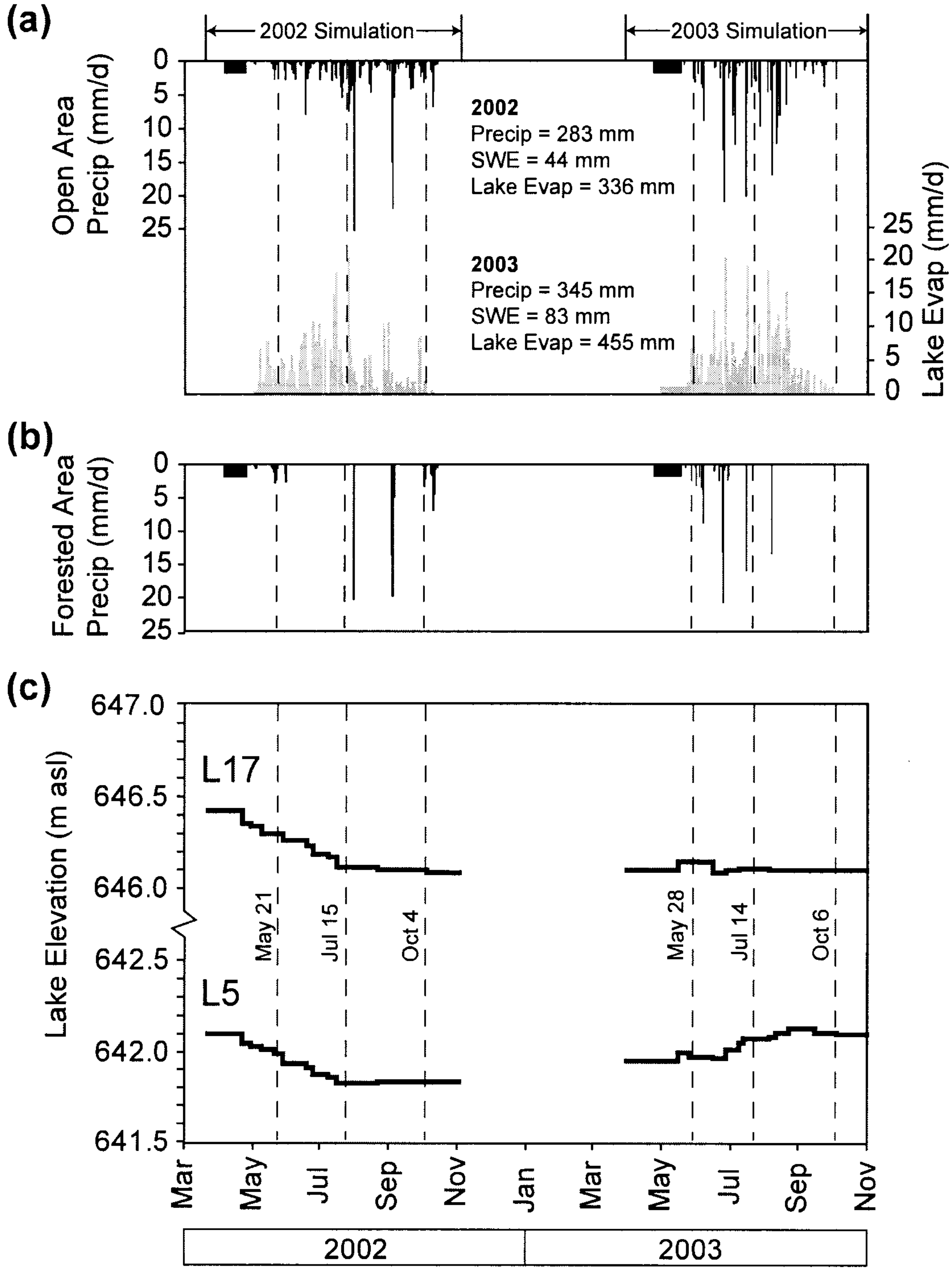


Figure 3.4. (a) Specified daily precipitation (including snow water equivalent – SWE) for open areas and specified lake evaporation. (b) Net precipitation for forested areas. (c) Specified hydraulic heads for Lakes 5 and 17. Specific output times for each year (May, July, and October) shown with dashed lines.

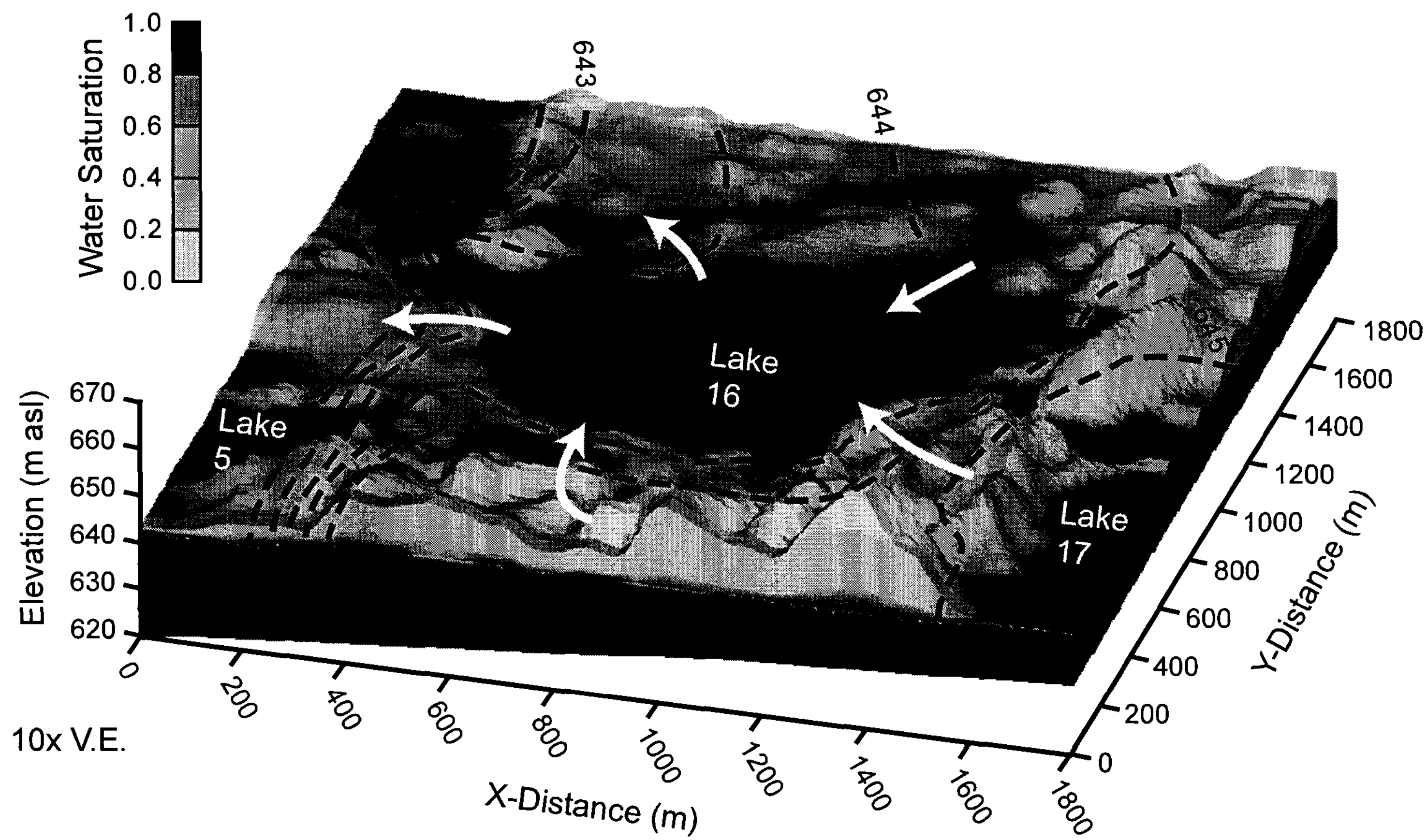


Figure 3.5. Three-dimensional shaded perspective of simulated water saturation overlain on model topography, simulated water table contours (0.5 m asl dashed lines), and approximate groundwater flowpaths interacting with Lake 16 (arrows) at model equilibrium.

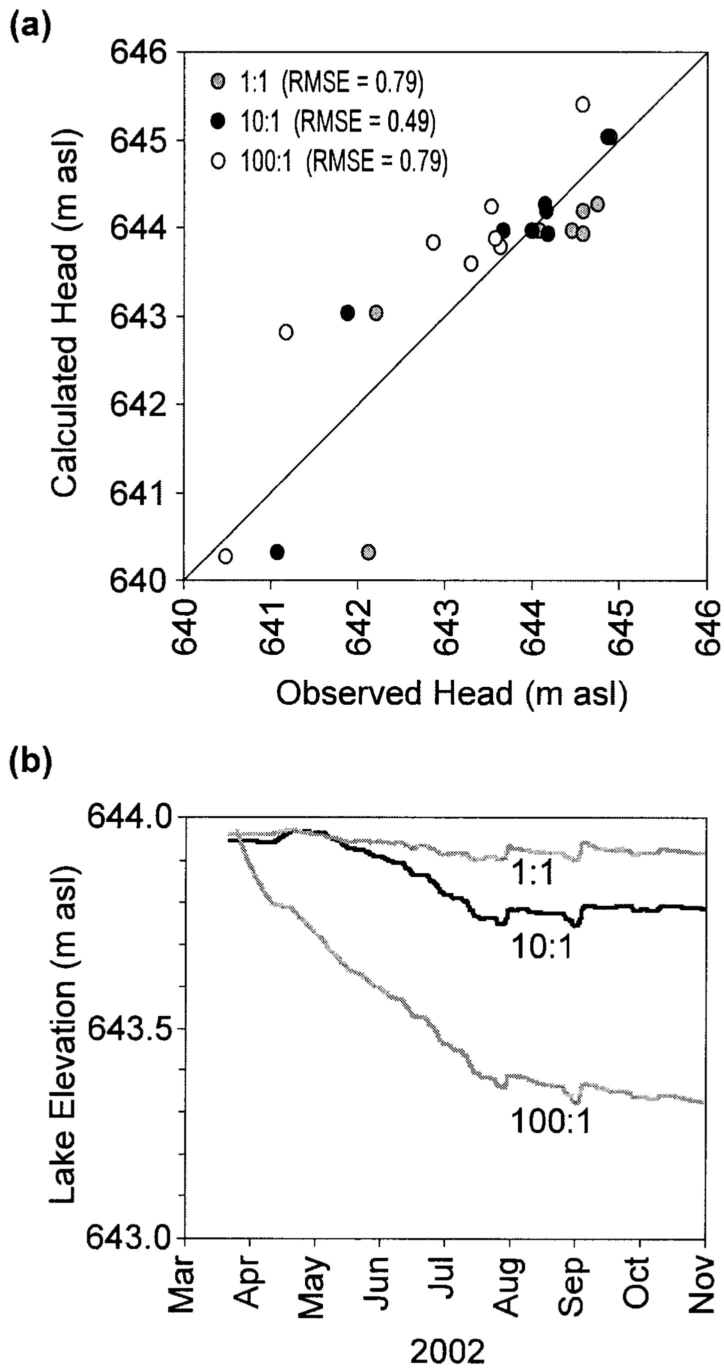


Figure 3.6. Effect of anisotropy (K_{xy}:K_Z) on (a) steady state solution of hydraulic heads, and (b) transient response of Lake 16 elevation.

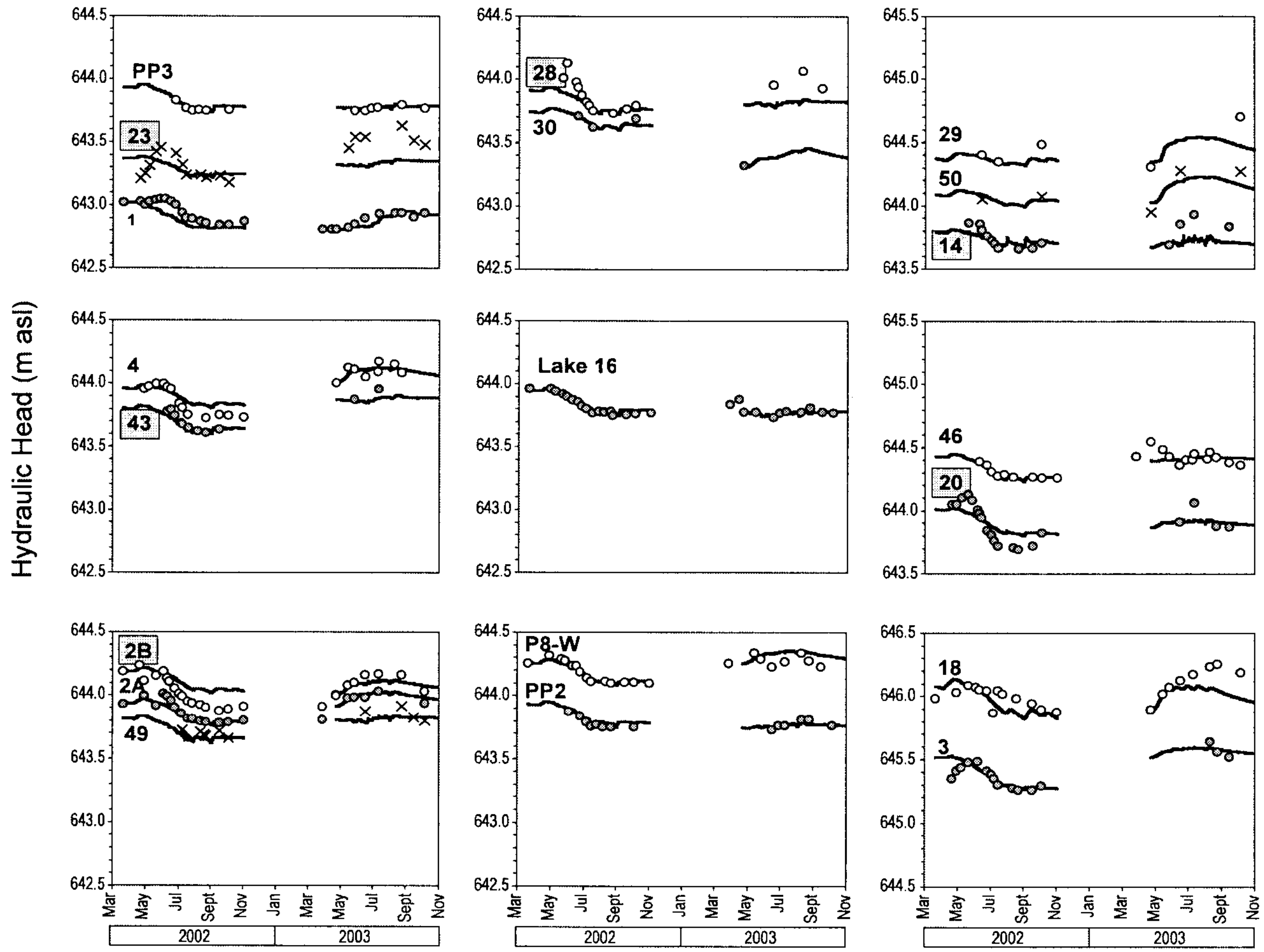
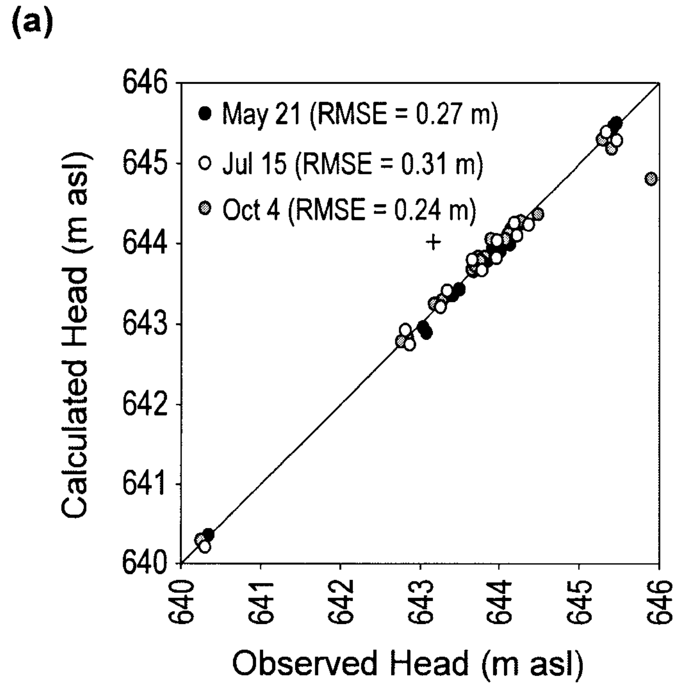


Figure 3.7. Time series model output (solid lines) and field measurements (symbols) of hydraulic head.



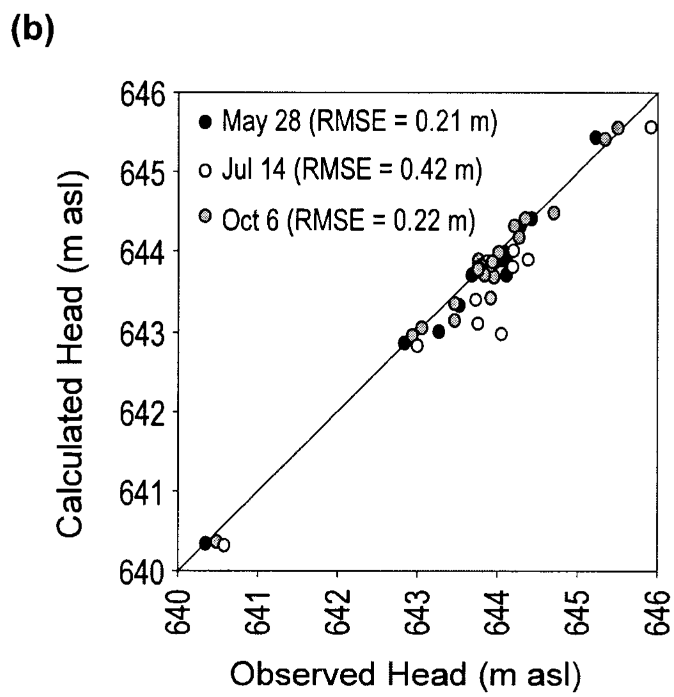


Figure 3.8. Root-mean-square-error (RMSE) for specific output times for (a) 2002 and (b) 2003.

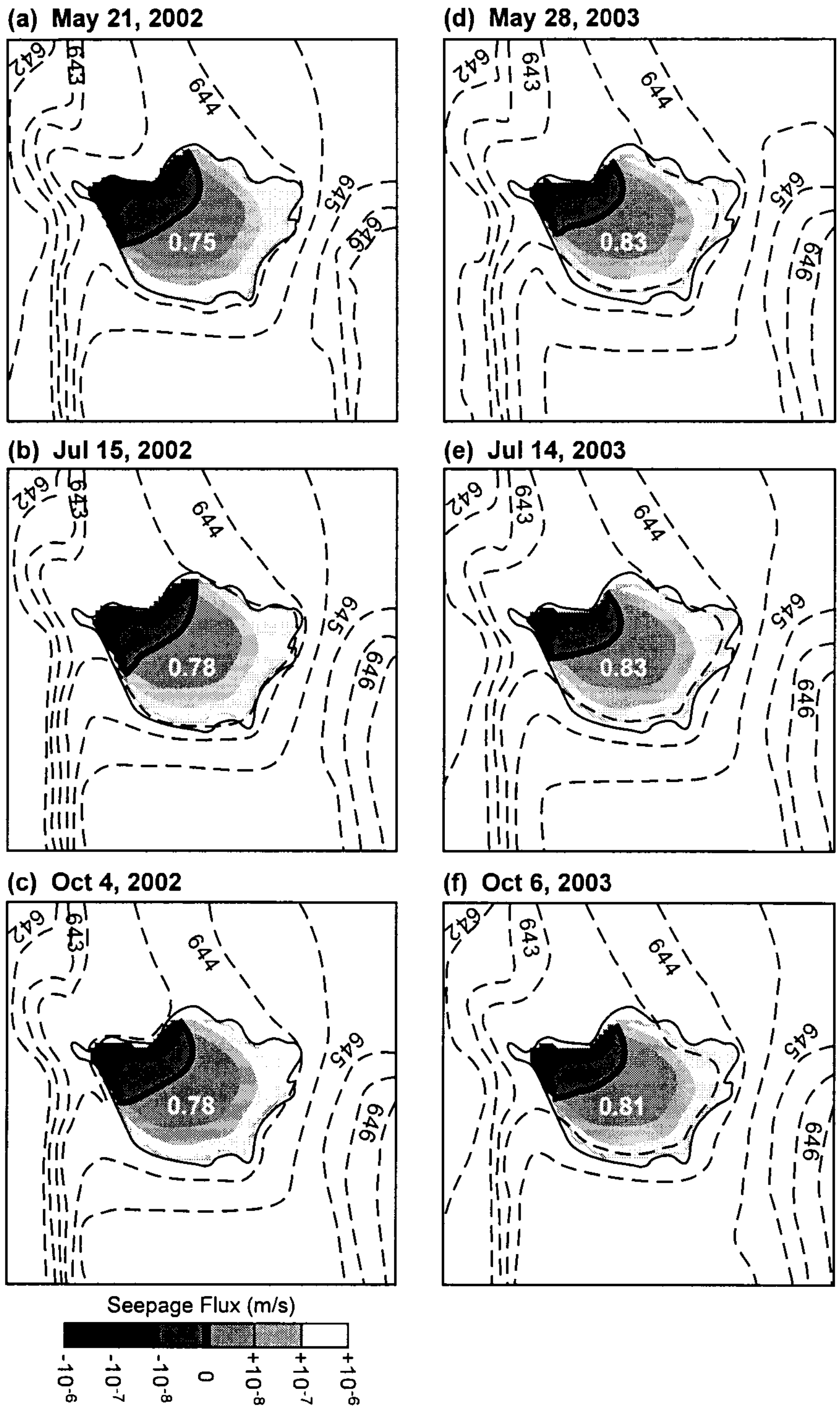


Figure 3.9. Simulated water table elevation (0.5 m asl contour interval) and shaded lakebed seepage flux. Fraction of lakebed area contributing groundwater to Lake 16 shown in white text. Transition between groundwater discharge areas and groundwater recharge areas shown as a thick black line.

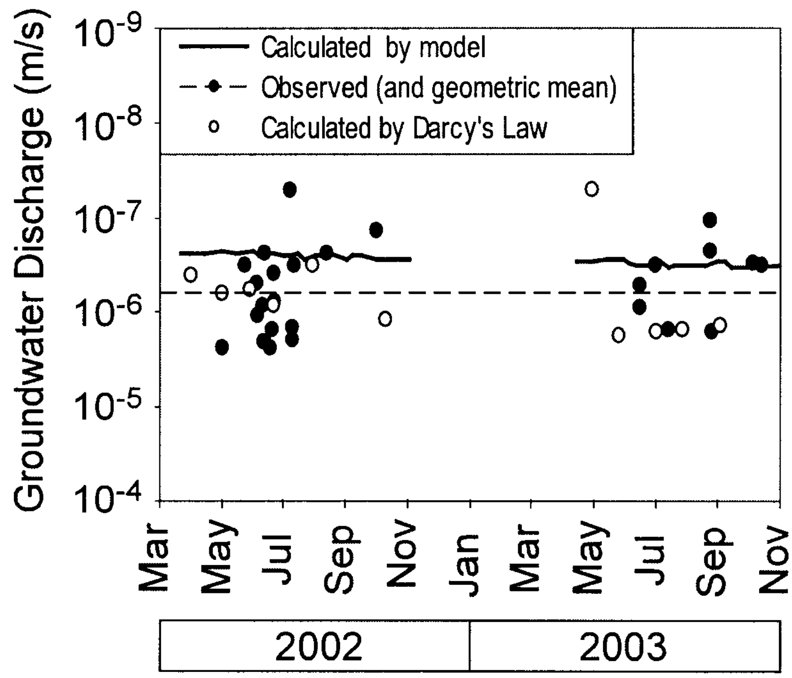


Figure 3.10. Simulated and observed groundwater discharge fluxes along southeast shore of Lake 16. Values observed from individual seepage meter measurements or calculated using Darcy's Law are from Smerdon et al. (2005).

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Chapter 4

Influence of climate and water table depth on groundwater recharge to an outwash plain in the Boreal Forest of Canada

4.1 Introduction

Groundwater recharge is an essential hydrologic process, which describes the portion of water fluxes from the atmosphere that migrate to the groundwater regime. Estimation of recharge at various spatiotemporal scales is vital for understanding deep percolation, developing conceptual models of water cycling, and for balancing large-scale watershed budgets by relating inputs (i.e., recharge) to outputs (e.g., regional evaporation, discharge through major river systems). Groundwater recharge is often a prevailing boundary condition for simulating the movement of fluids and solutes in the subsurface and requires knowledge of spatial and temporal distribution for application to hydrologic problems at the watershed scale (de Vries and Simmers, 2002). On the Boreal Forest landscapes of Canada, hydrological processes are sensitive to conditions of a sub-humid climate (Devito et al., 2005b), geology, and vegetation. Following a hierarchy proposed by Devito et al. (2005a), the influence of climatic conditions on groundwater recharge was investigated for a glacial outwash plain at the Utikuma Research Study Area (URSA), in northern Alberta, Canada.

On landscapes that have a relatively dry climate (e.g., climate is classified as semi-arid or sub-humid), an intricate relationship between the atmosphere and hydrosphere is created, which controls recharge and the configuration of the water table (Winter, 1986). As on most landscapes, precipitation may be intercepted by vegetative cover or may infiltrate into the subsurface. However, the exchange of energy and moisture at the interface of the atmosphere and the subsurface is complex and controls partitioning of moisture between subsurface, vegetative and atmospheric regimes. In dry regions, infiltrating water may be drawn up by vegetation (through evapotranspiration) or be stored in the unsaturated zone under diminished migration potential (due to low relative permeability associated with dry soils). In outwash landscapes, the timing of groundwater recharge has been found to be a controlling factor in the long-term maintenance of lakes and wetlands (Chapter 3; Anderson and Munter, 1981; Winter et al., 2001). Therefore, understanding knowledge of the hydrological processes affecting groundwater recharge will advance conceptual models of water flow (e.g., Chapter 2) and improve simulations of water cycling on outwash deposits at the URSA (e.g., Chapter 3).

The objective of this study was to determine the influence of climatic conditions on the subsurface moisture regime by quantifying groundwater recharge. Forest canopy interception and actual evapotranspiration rates were assumed, and estimates of groundwater recharge were determined by simulating transient water table observations from an instrumented field site at the URSA. The role of climatic conditions was investigated by simulating infiltration (i.e., unsaturated flow) driven by 71 years of historic climate data, and predicting water fluxes at various water table depths. The scope of the investigation was limited to a field site at the URSA (Lake 16), which had been studied previously (Chapter 2; Chapter 3; Smerdon et al., 2005). Field observations of water table response to specific precipitation events measured at the URSA constrained the development of generalized models of infiltration. Groundwater recharge rates determined in this study are summarized as probability distributions, which can be incorporated into larger-scale hydrologic and/or ecosystem simulation models, as a method of upscaling site-specific knowledge. Furthermore, the methodology described in this study is suitable for studying the response of landscapes of different soil texture and vegetative land cover (through modelling).

4.2 URSA Outwash Field Site

The URSA is located in the Boreal Forest of Canada, 370 km north of Edmonton, Alberta, Canada (56°6' N, 116°32' W; Figure 4.1), and is characterized by a sub-humid climate, low topographic relief, and thick heterogeneous glacial sediments (80 to 240 m: Pawlowicz and Fenton, 2002) overlying bedrock. Within the URSA, landscape heterogeneity encompasses lacustrine plains, moraines, and an outwash plain. The outwash field site is situated on the eastern margin of a 200 km^2 coarse-textured glacial outwash plain, which is hummocky, has remnants of an east to west trending esker, and a maximum topographic relief of 30 m. Within the specific study site, a gravel pit was constructed by mining sand and gravel from the east to west trending esker immediately southeast of Lake 16 (Figure 4.1). As determined from periodic aerial photographs of the study area, available from the 1940s to 2002, most of gravel pit development occurred in 1996, and the excavated area appears to have remained approximately constant.

The outwash study site contains three lakes that exist in a series of 'steps', where Lake 17 was 2.4 m higher than Lake 16, which was 1.8 m higher than Lake 5 in early 2002 (Figure 4.1). Measurements of hydraulic head at 70 piezometers, lake levels, and groundwater seepage estimates were made from April 2001 to October 2003, as part of the water budget study presented in Chapter 2. Groundwater moved from southeast to northwest (Figure 4.1) at an average horizontal hydraulic head gradient of 0.002, with groundwater springs occurring along the southeast shore of Lake 16. The groundwater flow system between Lakes 17 and 16 is defined by hydraulic head data from a subset of 9 of the 70 piezometers around the Lake 16 study site (Figure 4.2). The stage of Lakes 16 and 17, and the water table elevation in the vicinity of the gravel pit (P8-W; Figure 4.2), were recorded with vented pressure transducers and dataloggers (GlobalWater model WL-14) every 60 minutes. In this study, hydraulic head responses to rainfall events in summer 2002 were used as calibration data for transient simulations of groundwater recharge.

In north-central Alberta, annual precipitation varied from 318 to 529 mm from 1997 to 2001 (Devito et al., 2005b), and measured open water evaporation at the URSA varied from 336 to 438 mm from 1999 to 2003 (Ferone and Devito, 2004; Smerdon et al., 2005), below the estimated annual average potential evapotranspiration (PET) of 518 mm reported by Bothe and Abraham (1993). From 2001 to 2003, precipitation was recorded with a weight-recording gauge at a meteorological station 10 km west of the site. Historic climate data (i.e., prior to 2001) were recorded at an Environment Canada meteorological station in Slave Lake, Alberta, located approximately 100 km south of the Lake 16 study site (Environment Canada, 2002). In this study, precipitation that occurred for the months of November through April was assumed to accumulate as snow.

4.3 Outwash Hydrologic Modelling

4.3.1 Conceptual Models

Three-dimensional simulations of the outwash flow system (Chapter 3) identified values for the hydraulic parameterization of the outwash landscape, and illustrated that lake-groundwater interaction was controlled by highly permeable sandy sediments, riparian peatlands and stratified lakebed deposits. Simulation models of the hydrologic flow system at Lake 16 required that climatic boundary conditions (i.e., precipitation and evaporation) be divided into forested and non-forested areas, and that the forested areas had three distinct timeframes (spring, summer and fall), corresponding to seasonal variation of throughfall and evapotranspiration (Chapter 3). Water cycling was sensitive to the timing and magnitude of these seasonal periods of precipitation and evaporative fluxes and was further investigated in the present study, by quantifying groundwater recharge with three different flow models, including:

i) ***Present Day* model:** a two-dimensional cross-sectional representation of the present day outwash landscape in between Lakes 16 and 17, including the gravel pit.

ii) ***Historic* model:** a two-dimensional cross-sectional representation of the outwash landscape in between Lakes 16 and 17, prior to construction of the gravel pit.

iii) ***Recharge* models:** a series of one-dimensional models that represented vertical columns of unsaturated outwash.

Each conceptual model is presented in the following sections, with descriptions of the model configuration, boundary and initial conditions, and sources of field data. Each model was simulated for different finite-element domains using the HydroGeoSphere numerical code (Therrien et al., 2005), which solved Richards' equation for transient, variably-saturated groundwater flow in a porous medium. The two-dimensional, cross-sectional model domains, representing the outwash in between Lakes 16 and 17 (location shown on Figure 4.1), were oriented parallel to the direction of groundwater flow, as determined in Chapter 2. Common to each model, the subsurface hydrostratigraphy was represented by the same hydrostratigraphic zones defined in Chapter 3, which were based on interpretation of site geology (Chapter 2). Each zone was assigned hydraulic parameters utilized in Chapter 3, including a hydraulic conductivity (K) value, which was measured in the field, and estimates of specific storage (S_s), porosity (ϕ), and the ratio of horizontal to vertical hydraulic conductivity (i.e., K_X:K_Z; Figure 4.3a). Tabulated relationships for capillary pressure, water saturation, and relative permeability were required for the sandy outwash sediments, and were assumed to be similar to those reported for the Borden sand (Figure 4.3b; Abdul, 1985).

4.3.2 Canopy Interception and Actual Evapotranspiration

A key finding from Chapter 3 was that further investigation into the relationship between temporal throughfall and evapotranspiration fluid fluxes was necessary to define the rate and timing of groundwater recharge. Recharge to the groundwater regime depends on the partitioning of infiltrating water to evapotranspiration, soil moisture storage (in the unsaturated zone), and deeper percolation. Partitioning is governed by the complex exchange of energy and moisture at the interface of the atmosphere and unsaturated zone. At the time of this study, the temporal and spatial variation of moisture and energy transfer, forest canopy interception, and actual evapotranspiration (AET) had not been characterized at the URSA. Thus, assumed rates of AET and canopy interception (implemented as net precipitation: P_{net}) were required to further investigate groundwater recharge processes for the outwash landscape.

Actual evapotranspiration rates (Figure 4.4) were assumed to follow seasonal AET rates determined by Amiro et al. (2006) for an aspen canopy (BOREAS) site in Prince Albert, Saskatchewan, Canada. The study by Amiro et al. (2006) was selected from a review of hydrologic and meteorological publications (Redding et al., 2006) because of similarity in forest species (aspen upland), soil texture (sandy), and climate as the Lake 16 study site. In the simulation models, AET was either specified as a negative fluid flux at the ground surface (i.e., sink for water). Net precipitation (i.e., P_{net} = precipitation minus interception) was assumed to be 90% of total precipitation, based on a range of observed precipitation throughfall rates at the URSA (85% to 95% of P: Devito, unpublished data).

In the simulations, the negative time-series fluxes were assumed to represent flux of water removed from the subsurface by actual evapotranspiration (i.e., AET). The sensitivity of results was assessed by bracketing assumed values in each simulation with slightly higher and lower values. Net precipitation was varied from 85 to 95% of P. The weekly AET values were varied differentially such that the total annual AET was 50 mm higher and lower than the base case (Figure 4.4), resulting in a total of 350 to 450 mm/yr compared to 400 mm/yr estimated by Amiro et al. (2006). The combinations of a higher AET and lower P_{net} (to force *drier* soil moisture conditions) and a lower AET and higher

P_{net} (to force *wetter* soil moisture conditions), created the *Dry Case* and *Wet Case*, compared to the *Base Case*.

4.3.3 Present Day 2D Flow Model

Simulation of the *Present Day* flow system was necessary to confirm the hydraulic parameters assigned to the zones of porous media (from Chapter 3), and assumed AET and P_{net} values for the forested portions of the outwash landscape. Two separate timeframes were considered: July 30 to August 1, 2002, when a 35 mm rainfall event on July 31 caused an observed water table rise at P8-W *(Present Day A)*; and, August 14 to September 18, 2002, when there were two rainfall events (25 mm and 45 mm) separated by approximately two weeks *(Present Day B)*. For each scenario, the simulated transient water table response at P8-W was compared to time series of field data.

The two-dimensional, cross-sectional model extended from the middle of Lake 16 to the edge of Lake 17 (Figures 4.1 and 4.2), in the principal direction of groundwater flow (Chapter 2; Chapter 3). Subsurface elements were uniformly spaced at 1 m horizontally, and variably spaced vertically (0.1 to 0.9 m), with finer vertical spacing from the ground surface to a subsurface elevation below the position of the water table. The topography of the model domain was constructed from a 1 m horizontal resolution digital elevation model of the land surface combined with lake bathymetry measurements (Chapter 3), and interpretation of the bottom of the outwash sediments from geophysical data (Domes, 2004).

Fluid flow boundary conditions included time-varying specified hydraulic heads of Lakes 16 and 17, which were defined at the left- and right-hand sides of the model domain, and lake area (Figures 4.2b, 4.5a, and 4.6a). The lakes were simulated as specified hydraulic heads to constrain the groundwater flow setting by time series hydraulic head values recorded by the pressure transducers and dataloggers (Figures 4.5a and 4.6a). The bottom of the mesh was set as a no-flow boundary because it corresponds with the base of the outwash deposits, which are underlain by clay and have significantly lower hydraulic conductivity than the outwash sediments (Figure 4.3a). Initial hydraulic head and water saturation conditions were determined by running the model to equilibrium conditions with the specified head boundary conditions of Lakes 16 and 17 for July 30, 2002 and assuming that the average sum of atmospheric fluxes (i.e., P and AET) was zero (Chapter 3). Simulated initial conditions were compared to field observations for each of the 9 observation points.

Net precipitation was applied as a time-series fluid flux for each simulation (Figures 4.5a and 4.6a). For the single 35 mm rainfall event *(Present Day A)*, hourly precipitation data were specified uniformly across the entire ground surface, and AET and canopy interception were assumed to be negligible during the event. Simulated water table response to this single event was intended to confirm hydraulic parameters defined for the subsurface over a short duration, in the absence of complicating AET and interception factors. For the *Present Day B* simulation, daily precipitation data and various combinations of assumed AET and interception values were specified, as described in section 4.3.2. Positive (P_{net}) and negative (AET) time-series flux values were applied to the top of the ground surface, depending on whether the ground surface was a forested upland area or the gravel pit (Figure 4.2b). The simulated water table position at P8-W was compared to time-series field data obtained from the pressure transducer at this location, thereby using water table response to confirm assumptions of AET and canopy interception.

4.3.4 Historic 2D Flow Model

Using a similar model domain to that described above, subsurface flow conditions were simulated for 1983 to 1996, prior to construction of the gravel pit, to investigate *Historic* water table response to net climatic fluxes (i.e., P_{net} minus AET) for steady state and time-varying lake boundary conditions. The top surface of the model domain was assumed to include a forested esker, up to 10 m thick, which was subsequently removed to form the gravel pit (Figure 4.2b). The model used subsurface elements of the same spatial discretization as the *Present Day* model.

To simulate the natural flow system prior to gravel pit construction and to determine the sensitivity of the flow solution to lake boundary conditions, the flow system was simulated with: i) average lake levels (i.e., steady state boundary conditions); and, ii) monthly lake levels that fluctuated (i.e., transient boundary conditions). For each simulation, monthly climatic boundary conditions for the region were input as net fluid fluxes at the ground surface, and the water table fluctuation was simulated at P8-W. Because this period of time occurred prior to field research at the URSA, lake levels were estimated from field records, assuming a simplified relationship to average monthly P and potential evapotranspiration (PET) calculated by the Thornthwaite (1948) method from historic climate data (Figure 4.7a). These hypothetical changes in lake level were calculated for each month (Figure 4.7b), assuming that lake level was entirely controlled by P and PET (i.e., assuming a monthly balance of other inputs and outputs, such as groundwater seepage). Temporal AET rates (Figure 4.4) and P_{net} (90% of P), were assumed to be identical each year, and were implemented as net monthly fluid fluxes (i.e., positive or negative values) assigned to the top of the model domain.

Research at the Lake 16 study site (Chapter 2), and three-dimensional simulations of this lake-groundwater system (Chapter 3) have shown that the elevation of Lake 16 is dependent on groundwater; however, a dominant hydrologic control was open water evaporation. Thus, the objective of the *Historic* model was not to reconstruct the subsurface flow regime as accurately as possible, but rather to investigate the transient behaviour caused by fluctuating lake levels when combined with assumed AET and interception values. Considering the sub-humid climate, I hypothesized that when the landscape was completely forested, groundwater recharge would be minimal and that the boundary Lakes 16 and 17 would control the water table position (Haitjema and Mitchell-Bruker, 2005).

4.3.5 1D Recharge Models

Long-term groundwater recharge, for varying water table depths, was quantified with a series of one-dimensional models. These models allowed development of a more generic representation of groundwater recharge for coarse-textured deposits in this region than determined in the two-dimensional *Present Day* and *Historic* flow models. The series of *Recharge* models represent virtual columns of unsaturated outwash sediment, extending upward from the water table to the ground surface. Using the historic record for precipitation in this region, climatic boundary conditions were applied to the top of 11 columns, varying in height from 2 to 12 m (each model domain incrementally increased in height by 1 m), and the fluid flux at the base of the column (i.e., the water table) was simulated for 71 years (1933 to 2004).

The models had 0.2 m vertically spaced elements, and were defined with the hydraulic parameters of outwash sand and gravel (Figure 4.3). The base of the model was specified with a hydraulic head equal to the base elevation of the column to represent the (fixed) water table. Separate positive and negative time-series fluid fluxes were applied to the top of the model domain, corresponding to monthly P_{net} and the assumed seasonal AET, respectively (Figure 4.8b).

4.4 Results

4.4.1 Outwash Flow System (2D Models)

Simulated initial conditions for the *Present Day* model were similar to field observations for July 2002 (Figure 4.9). For 9 observation points located along the cross-section modelled, the root mean squared error (RMSE) for hydraulic heads was 0.16 m. The response of the hydraulic head at P8-W to the single rainfall event *(Present Day A)* also compared well with field data (Figure 4.5b). These results substantiate the distribution of subsurface hydrostratigraphy shown on Figure 4.2b, and the corresponding hydraulic parameters (Figure 4.3). Compared to previous lake-groundwater modelling (Chapter 3), the zones of porous media and hydraulic parameters were nearly unchanged. The addition of a more permeable upper section of outwash sand (labelled sand and gravel in the present study) provided a better representation of the outwash deposit in the vicinity of borehole P8-W (Chapter 2). The addition of this porous media zone improved the transient fit of simulated water table response to field observations at a finer temporal resolution than considered in Chapter 3 (Figure 4.5b), and was the only calibration necessary.

For the *Present Day B* simulation, hydraulic head response at P8-W compared favourably with field observations when assumed P_{net} and AET values were considered (Figure 4.6b). The *Base Case* P_{net} and AET values provided a close match to field data and were bracketed by *Wet* and *Dry Cases*. The observed water table response was only replicated when AET was considered explicitly. When AET was not considered, simulated hydraulic heads remained higher than observed in the field (Figure 4.6b:

August 15 to 30, 2002), and only appeared to mimic the transience of the lake boundary conditions (Figure 4.6b: September 5 to 18, 2002). These results justify my assumed P_{net} and AET values.

Results from the *Historic* model reveal that when the landscape was completely forested, water table fluctuations at the present day monitoring point (P8-W) should have been less than 0.10 m, and controlled by the boundary lake levels (Figure 4.7c). When Lakes 16 and 17 were simulated as constant level boundary conditions, the water table at P8-W remained at a constant level and did not respond to recharge originating at the ground surface (Figure 4.7c). Thus, for unsaturated zone thicknesses of 2 m or greater, implementation of a fixed water table boundary condition was a suitable simplification (for less than 0.10 m fluctuation) and will not have a large influence on the *Recharge* models.

4.4.2 Groundwater Recharge (1D Models)

Positive fluid fluxes at the water table indicate conditions when the saturated groundwater zone was a sink for infiltrating water (i.e., groundwater recharge), and negative values indicate conditions when the saturated groundwater zone was a water source (herein referred to as upflux; Jaber et al., 2006). For the *Base Case*, a wide variation in monthly fluid flux was found for water table depths of 4 m or less (Figure 4.8c), including positive and negative values. For water table depths of 4 m or less, the highest (positive) flux at the water table occurred in March each year and the minimum (negative) flux between the months of June and August. Absolute values varied considerably when a shallower water table was considered (2 m: Figure 4.8c), with peak groundwater recharge rates of 10 to 100 mm occurring in the spring, followed by upflux rates of 10 to 40 mm occurring anytime from June to August in the summer months, depending on the climate conditions of the preceding years.

For a water table depth of 4 m, recharge was up to 40 mm, occurring slightly later within the year (May and June) compared to the 2 m water table depth. Similarly, maximum upflux occurred later in the year (October) for a water table depth of 4 m. For a water table depth of 6 m, seasonal variation between recharge and upflux conditions existed; however, the peak flux rates were shifted to the late summer months (August and

September) for recharge, and spring months in the following year for upflux. For water table depths greater than 6 m, the seasonality of positive groundwater recharge and negative groundwater upflux was damped considerably (12 m: Figure 4.8c), and monthly groundwater recharge rates decreased to less than 10 mm, with upflux conditions occurring infrequently.

For water table depths less than 6 m, the amount of recharge and upflux depended on climatic boundary conditions of the present and immediately preceding year. As depth to the water table increased, the influence of climatic conditions became increasingly damped. This relationship between climatic conditions and flux at the water table was clearly illustrated for the late 1960's period, when moisture deficit conditions from 1963 to 1966 were followed by moisture surplus conditions from 1967 to 1972 (Figure 4.10a). Prior to the onset of moisture surplus conditions in 1967, groundwater recharge was negligible (<10 mm/month), with 10 to 25 mm/month of upflux occurring in the summer months. At the onset of moisture surplus conditions (marked by a grey bar on Figure 4.10a), groundwater recharge occurred within the same year for water table depths of 2 and 4 m, but there was a delayed increase in recharge at greater depths (see line markers on Figure 4.10b). Evaluation of annual precipitation (Figure 4.8a) revealed that for the years 1963 through 1966, annual snow accumulation was less than 50 mm/year, which is less than normal for the 71 year timeframe. Moisture surplus conditions were created by a return to normal thickness of snow accumulation (approximately 100 mm/year), and an increase in precipitation in the spring months of 1973 (251 mm) and July 1974 (150 mm). The increase in groundwater recharge, associated with moisture surplus conditions, was delayed by 2 and 3.5 years for water table depths of 6 and 12 m (see line markers on Figure 4.10b), respectively, indicating a strong time-lagged response.

The damped response of changes to fluxes at the water table, with increasing depth, was also observed when moisture surplus conditions reversed from moisture surplus to deficit conditions in 1998 (marked by a grey bar on Figure 4.10c). The onset of moisture deficit conditions was created by a decline in annual snow accumulation (approximately 65 mm/year from 1998 to 2000), and caused recharge to diminish in 1998 for water table depths of 2 and 4 m (Figure 4.10d). However, the moisture deficit (drought) conditions were found to decrease recharge for water table depths of 6 and 12 m within 1.5 and 2 years, respectively. Thus, compared to development of moisture surplus conditions (in the late 1960's), the onset of drought influenced fluxes at deeper water tables 0.5 to 1.5 years earlier. The effect of this drought in the late 1990's and early 2000's also caused upflux conditions to be favoured over groundwater recharge, with 25 mm/month of upflux simulated for a water table depth of 4 m in nearly every year from 1998 to 2003 (Figure 4.10d).

4.4.3 Net Effect on Annual Fluid Fluxes

The overall net effect on water cycling in coarse-textured landscapes, and the sensitivity to slightly wetter and drier conditions (represented by the *Wet* and *Dry Cases*), was apparent at annual (or longer) timescales. Although there are frequent years of moisture deficit conditions, annual net fluxes reaching the ground surface favour moisture surplus conditions (Figure 4.11a), as expected in a sub-*humid* climate. Fluid fluxes at a water table depth of 2 m vary from –80 to +150 mm (negative values are upflux and positive values are recharge). When the *Wet* and *Dry* climatic conditions were considered, annual flux values were 30 to 50 mm higher or lower than the *Base Case* (Figure 4.11b). For a water table depth of 4 m, annual values vary from –120 to +265 mm, with wider variability in *Wet* and *Dry Cases* (30 to 160 mm higher or lower than the Base Case; Figure 4.11c). When the water table depth is greater than 4 m, the variation between *Wet* and *Dry Cases* was only 40 to 80 mm higher or lower than the *Base Case* (Figure 4.11d and 4.11e).

At shallow water table depths (i.e., 2 m), the large monthly variability of fluid fluxes (Figure 4.8c) was reduced at the annual timescale (Figure 4.11b). The net effect was less variation in annual recharge for a 2 m deep water table, compared to 4 m. Annual variation in recharge and upflux was large for a 4 m deep water table in each of the scenarios (*Base, Wet* and *Dry Cases*), indicating a higher sensitivity to atmospheric moisture conditions than observed for shallower water table depths. These findings illustrate the control of climate conditions on moisture movement in the upper 4 m of the unsaturated zone. Below water table depths of about 4 m, the annual variability of recharge and upflux is similar to 2 m. The net effect was that subtle variations in climatic conditions would have a larger effect in areas where the water table is approximately 4 m deep. When the water table was shallower, the seasonal cycle of recharge and upflux are nearly equal, and when the water table was deeper, the time lag reduces annual variability.

For a water table depth of 12 m, simulated fluid fluxes were predominantly positive (i.e., recharge), with potential upflux only occurring in drought conditions for the *Base* and *Dry Cases*. Variability was less frequent at the annual timescale, and appeared to follow a 30 to 40 year period (Figure 4.11e). For the *Base Case*, the net recharge was between 60 to 100 mm/year from 1938 to 1958, and from 1978 to 1998.

The net effect was that the magnitude of groundwater recharge depended on the depth to the water table, which can be seen in Figures 4.8 and 4.11. The correlation between water table depth and recharge can be succinctly shown on probability distributions (Figure 4.12), utilizing the complete simulation timeframe of the 1D Models (71 years). Summary statistics of the probability distributions (mean and standard deviation) identify 4 well-defined groups that correspond to increasing water table depth. In areas with a shallow water table (2 to 3 m; Figure 4.12a), annual net fluid flux appeared to be normally distributed, with a mean of 25 mm/year and a standard deviation of 50 mm. As the depth to water table was increased to 4 to 5 m, the mean fluid flux and standard deviation increased to 60 and 80 mm/year, respectively. For water table depths of 6 m and below, the mean flux is higher (75 to 80 mm/year), the standard deviation is lower (30 to 40 mm), and the distributions appear to be skewed towards to negative flux values indicating a propensity for occasional upflux events.

4.5 Discussion

4.5.1 Effect of Sub-humid Climate and Forest Cover on Recharge

Groundwater recharge at the URSA is controlled by recent climate conditions, the presence of a forested landscape, and depth to the water table. The sub-humid climate is a dominant factor in determining the magnitude of water available for recharge, and can vary from water deficit to water surplus conditions at monthly and decadal timescales. Seasonal variation of P_{net} and AET controls the occurrence of recharge or upflux conditions, and ultimately constrains the flux of water at the water table. The recharge models in the present study of glacial outwash sediments utilize a seasonal distribution of AET that was determined for a similar site on the Boreal Plains (Amiro et al., 2006). Although these AET values were not site specific and were assumed to be identical each year, the simulations illustrated the importance of the timing of forest canopy activity for the potential for groundwater recharge. The maximum potential for groundwater recharge occurred when the influence of AET was limited, such as during spring melt (when the forest canopy has not developed much leaf area coverage), and in the fall months. Gosselin et al. (1999) had similar findings for a site in Nebraska, where groundwater recharge was greatest between growing seasons. Within the growing season, groundwater recharge may be possible on coarse-textured landscapes at the URSA. However, the occurrence of recharge would be limited to periods in the summer months when extreme precipitation events significantly exceed canopy interception and AET (e.g., 1996 on Figures 4.8a, 4.11c, and 4.11d).

The recharge simulations contained two intervals of water deficit conditions (drought), which occurred in the late 1960's and in the late 1990's (Figure 4.11a) and appeared to be caused by decreased snow accumulation (Figure 4.8a). Each of these drought conditions resulted in diminished groundwater recharge and increased upflux from the water table (Figure 4.10), a hydrological process which has been measured at up to 40 mm/yr in humid regions with agricultural land use (Jaber et al., 2006). The reliance of groundwater recharge on snowmelt and/or large summer precipitation events is illustrated further in the moisture surplus conditions of the early 1970's (Figure 4.10b), and relatively high precipitation of 1996 (Figure 4.10d).

Finally, for a water table depth of 12 m, there is a strong relationship between the pattern of annual snow accumulation and long-term groundwater recharge (Figure 4.11e). Thus, for coarse-textured landscapes in sub-humid parts of the Boreal Forest, the occurrence of recharge depends on climate history and the amount of annual snow accumulation. After the snowmelt recharge event, the forested portions of the landscape can remove water though AET and a decrease in recharge is followed by upflux conditions in the summer months. Large summer precipitation events may have the potential to generate recharge to the groundwater regime; however, snow accumulation (and subsequent rapid melting in the spring) appears to occur prior to the moisture demands and interception of a forested landscape.

4.5.2 Correlation of Recharge With Water Table Depth

The hummocky topography and a nearly planar water table between lakes on the URSA outwash add complexity to the spatial distribution of groundwater recharge. As illustrated along the cross sectional model in the 2D flow simulations (Figure 4.2), the depth to water table may vary from 0 to 12 m. Areas of shallow water table depth (i.e., 2 m or less) have a potential for variable recharge and upflux monthly, and the magnitude will depend on the amount of snowmelt and large precipitation events in the current year. In a semi-arid region having annual P of 340 mm (similar to deficit years at the URSA), Cook et al. (1989) found that most rainfall is captured by AET in the uppermost 2 m of the soil column. For a 2 m deep water table, the long-term simulations indicate that upflux were as high as 92 mm/month in drought years on the URSA outwash. As the depth to water table increased, the probability of groundwater recharge also increases because groundwater recharge in the spring months will have an opportunity to infiltrate to a depth at which AET can no longer remove water. When the water table is below 6 m depth, groundwater recharge rates were less sensitive to the *Wet* and *Dry Cases* (compared to shallower depths), and the strong seasonal response was subdued.

The magnitude of groundwater recharge is dependent on the depth to the water table, and can be grouped into 4 distinct zones of different recharge/upflux (Figure 4.12a). In areas of shallow water table depth (2 to 3 m), the probability of recharge or upflux is nearly equal, having a mean of only 25 mm/year (recharge). The largest standard deviation in recharge/upflux was predicted for a water table 4 to 5 m deep. Within this depth range, higher magnitudes of recharge or upflux are possible, and the occurrence depends on climatic conditions and integrating the moisture movement in the upper 2 m of unsaturated soil (as suggested by Cook et al., 1989). The highest probability for long-term recharge conditions occurred when the water table is greater than 6 m below the ground surface. A 6 m thick unsaturated zone allows for integration of moisture surplus and deficit conditions, and for snowmelt to migrate to a sufficient depth to not be drawn upward by evaporative processes near the ground surface.

Spatially variable water table depth, which is caused by the hummocky topography of outwash deposits, and spatially and temporally variable land cover, further complicate the effect of climate on groundwater recharge (Winter, 1986). For any landscape, including such spatial and temporal variations in a distributed flow model is desirable (de Vries and Simmers, 2002), and is often required to better represent local hydrologic conditions. Models such as HELP (Schroeder et al., 1994) have aided the quantification of groundwater recharge by integrating spatially variable land cover and common climatic variables (e.g., Jyrkama et al., 2002). However, upflux conditions are not predicted by the HELP model, and would require a separate model for evapotranspiration to be estimated.

Simulated recharge fluxes were not steady, and variation appeared to follow average climate conditions from the most recent decade, and whether there were extended periods of water deficit or surplus, compared to periods of near balance. These unsteady state results corroborate the findings of a only a weak relationship between mean annual temperature (indicative of recent climate conditions) and recharge by McMahon et al. (2006). Therefore, the present work illustrates that recharge and/or upflux could be simulated in one-dimension using atmospheric flux data and knowledge of soil profiles for a region, and subsequently applied to a larger regional model, based on approximate/average water table depth and knowledge of a heterogeneous land cover.

4.5.3 Variation in Soil Texture and Vegetation Type

Rates of recharge and upflux (Figures 4.10 and 4.11) and the associated probability distributions (Figure 4.12) were developed from an extension of field observations of water table response to specific precipitation events to generalized 1D models. These models link forest canopy interception, evapotranspiration, and groundwater recharge for *one* soil texture, which has a relatively high K (1×10^{-3} m/s), and a given relationship for capillary pressure, water saturation, and relative permeability. Variation in the parameterization of this soil texture will result in different unsaturated flow characteristics, which will yield different predictions of groundwater recharge. Considering the relatively high K of the outwash sand and gravel deposit, I would anticipate that sediments with higher K to exhibit infiltration characteristics similar to those predicted in this study (i.e., seasonal recharge driven by snowmelt, and development of upflux conditions in shallow water table areas following drought). For sediments with a lower K and/or more gradual capillary pressure-saturation curve, infiltration may be slowed, and the probability of recharge to the groundwater regime may be lower than predicted in this study. A lower infiltration rate will reduce the seasonal pulse of recharge from snowmelt that is predicted for the coarse-textured outwash, and possibly allow snowmelt water to participate in the dynamic moisture transfers occurring between the shallow soil and vegetation. The net effect might be that nearly equivalent recharge and upflux rates would exist for unsaturated zone thicknesses greater than 6 m (as was predicted in this study), and requires further investigation.

Although actual evapotranspiration was not modelled for the climate data in this study, reasonable AET values for a similar Boreal Forest site were specified. The relationships between atmospheric conditions, soil moisture and vegetation are complex, and actual evapotranspiration rates will vary annually depending on climate and soil moisture conditions, and the growth stage of the vegetative cover. For the 71 year timeframe considered in the one-dimensional infiltration models, AET rates would have varied depending on age of the forested ecosystem, density of trees and other vegetation, and climate conditions. As additional knowledge of site specific AET processes is acquired, future models of the coupled relationship between energy and moisture transfer could be developed, and explicitly consider succession of vegetative species for simulation of multiple decades and climate cycles.

4.5.4 Implications for Water Cycling in the Boreal Forest

Understanding the spatial and temporal distribution of groundwater recharge (and upflux) is necessary for applying hydrologic models at the watershed scale (de Vries and Simmers, 2002). The recharge and upflux rates predicted in this study, and probability distributions add to a growing understanding of water cycling in sub-humid regions of the Boreal Forest. Using the probability distributions, results of this study can be scaled up to regional models of hydrologic and/or ecosystem simulation. The methodology described in this study is suitable for studying the response of different landscapes and vegetative land cover (through modelling) to climatic cycles. Recharge processes could be quantified as shown here for any soil texture, and compared using probability distributions.

For example, at Syncrude Canada's Southwest Sand Storage Facility (SWSS), near Fort McMurray, Alberta, recharge water was found to interact with salt water present in the tailings dam (Price, 2005). Flushing of salts from this coarse-textured, constructed landform is expected to occur over 300 years (Price, 2005), and will control the effectiveness of landscape re-vegetation. Predicted flushing times and understanding the process of solute mixing with infiltrating fresh water and migration through the SWSS facility, relies on understanding groundwater recharge processes on coarse-textured deposits (e.g., Delin and Landon, 2002). The present study indicates that a water table depth of 6 m or more would be required to promote consistent recharge conditions. Although the soil texture is finer at the SWSS, compared to the URSA outwash, this study indicated that the probability of recharge conditions occurring is about 0.8 (Figure 4.12b) for areas where the water table is less than 6 m below ground surface, and generally lower for shallower water table depths (i.e., less than 6 m). While infiltrating fresh water may be favourable for re-vegetation, the high annual variation in fluid fluxes at water table depths of 2 to 4 m (Figure 4.12a) might favour the flushing of salts in some areas of Syncrude Canada's SWSS facility.

4.6 Conclusions

Rates of groundwater recharge and/or upflux depend on the climate history, the depth to the water table, and geology, and were largely driven by annual snowmelt. The influence of the present year, the preceding year, or the past decade will control the amount of fluid flux passing across the water table (positive or negative). On coarse-textured deposits in the Boreal Forest, groundwater recharge will also depend on the timing of precipitation events compared to seasonal AET. The sub-humid climate, with inter-annual and longer-term variation between water surplus and water deficit conditions, causes both groundwater recharge and upflux from the water table to occur within a single year for areas with a shallow water table. Recharge and/or upflux rates will depend on the current and previous year for water table depths of about 6 m or less. At depths greater than 6 m, recharge and/or upflux will depend on climate conditions from the most recent decade; however, it appears that 75 to 80 mm of recharge has been frequent in recent decades.

Fluid fluxes at the water table (recharge or upflux) were sensitive to small changes in annual climatic boundary conditions, especially in areas where the water table was 2 to 4 m below ground surface. The relationship between recharge (and upflux) to water table depth has implications for larger-scale hydrologic modelling on the Boreal Plains, and the probability distributions presented for distinct depth ranges will aid capturing site-specific knowledge and spatiotemporal variability in larger scale regional models. Detailed knowledge of ground surface topography may be useful for defining areas where the water table is expected to be shallow (i.e., <6 m depth), which will have more dynamic recharge/upflux functions than areas with deep water tables. However, considering that surface water features, such as wetlands, ponds and lake, control the configuration of the water table, lower resolution DEMs could be used, thus increasing the availability of resources for field-level reconnaissance. Furthermore, recharge and/or upflux could be modelled in one-dimension using atmospheric flux data and knowledge of soil profiles for a large region, and subsequently applied to a larger regional model, as a probabilistic function based on water table depths.

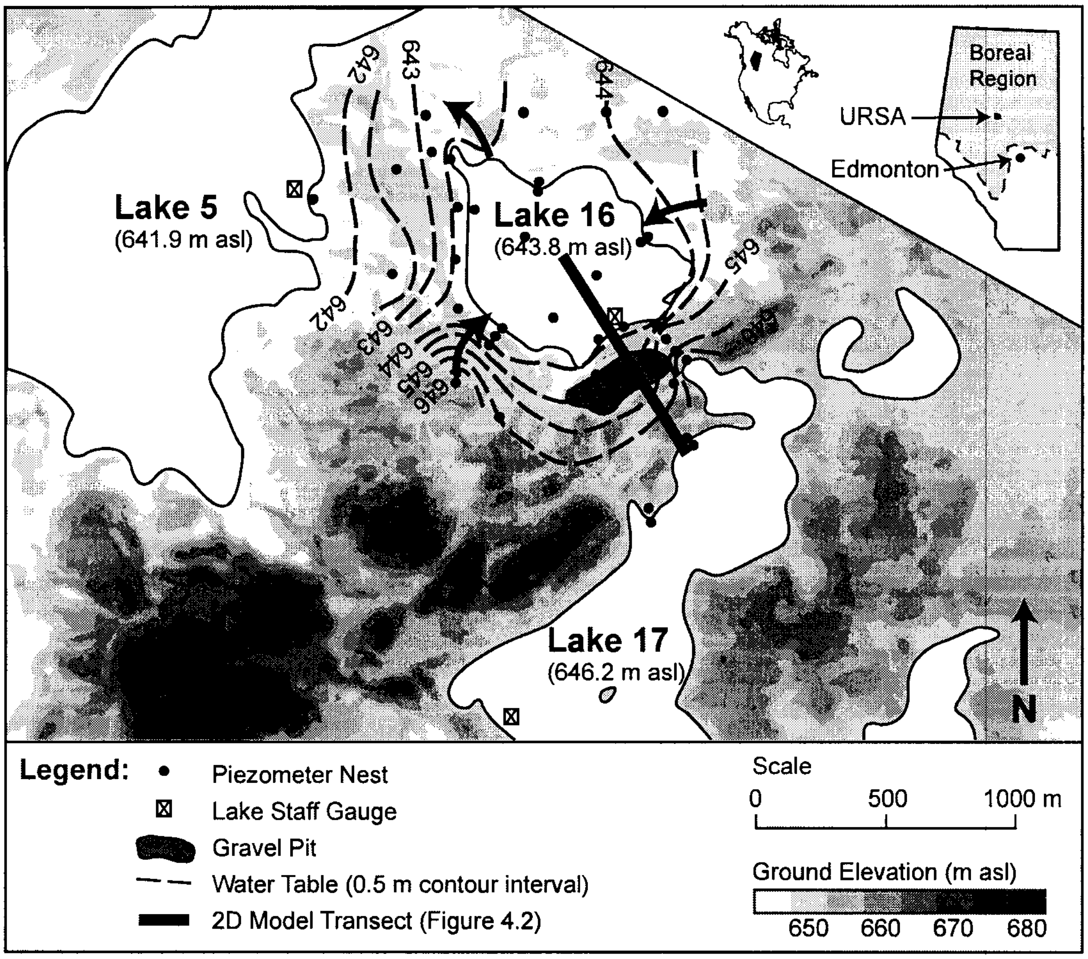

Figure 4.1. Lake 16 study site with topography, selected field instrumentation, and cross sectional numerical model domain. Water table contours and arrows indicating groundwater flow direction for July 2002. URSA and Boreal Plains region in Alberta, Canada on inset.

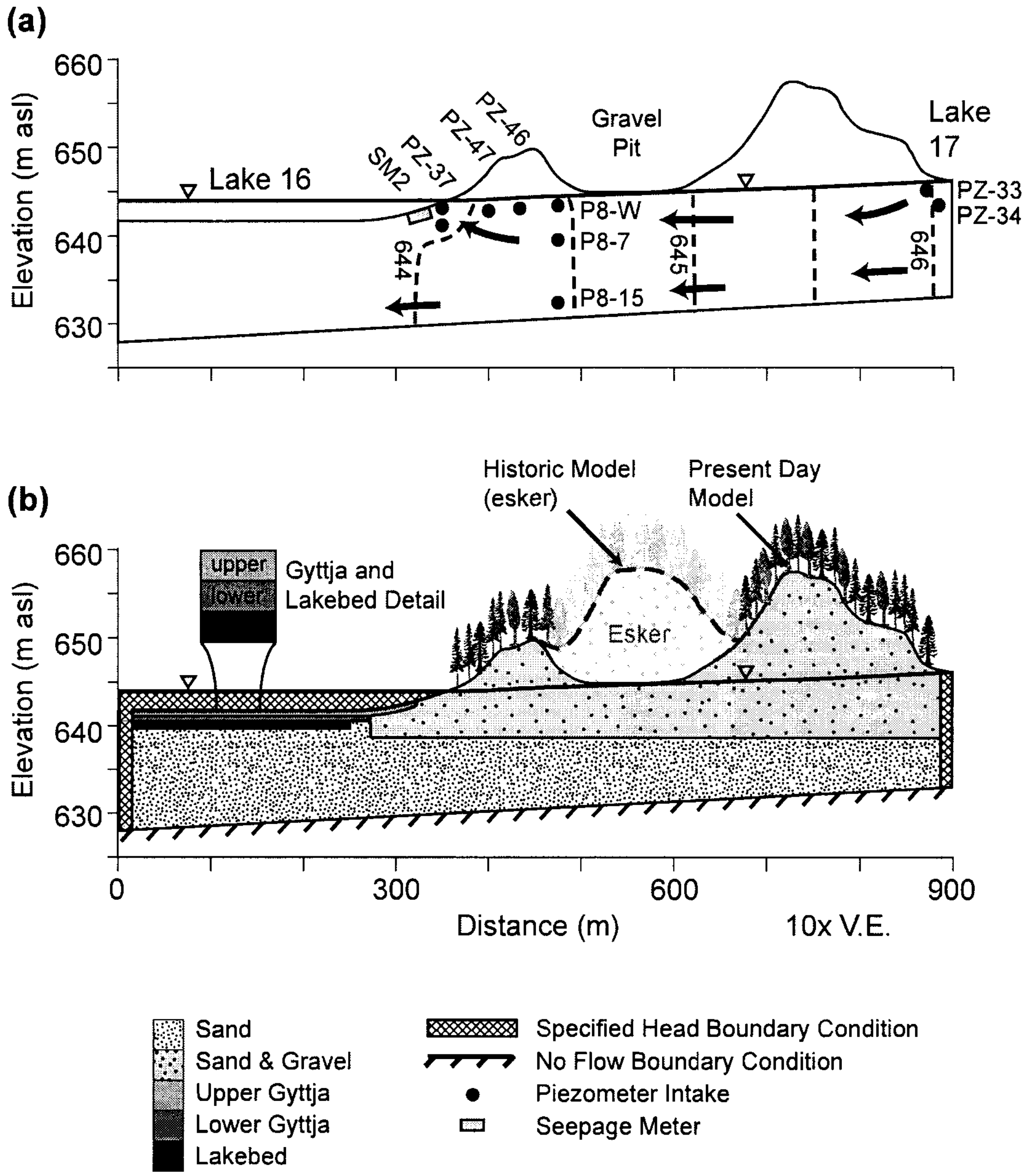

Figure 4.2. Cross section of *Present Day* and *Historic* models (location shown on Figure 4.1). (a) Observation points and contours of simulated hydraulic head for July 2002. (b) Shaded porous media zones, boundary conditions, and upland forested areas. Forested esker on (b) is shaded lightly to illustrate the difference between each model.

(a)

Zone	K (m/s)	K_X:K_Z	S_S (m^{-1})	ϕ
Sand	$1x10^{-5}$	10:1	10^{-5}	0.35
Sand & Gravel	$1x10^{-3}$	10:1	10^{-5}	0.35
Upper Gyttja	$3x10^{-6}$	1:1	10^{-4}	0.45
Lower Gyttja	$1x10^{-7}$	1:1	10^{-4}	0.45
Lakebed	$1x10^{-8}$	1:1	10^{-4}	0.45

(b)

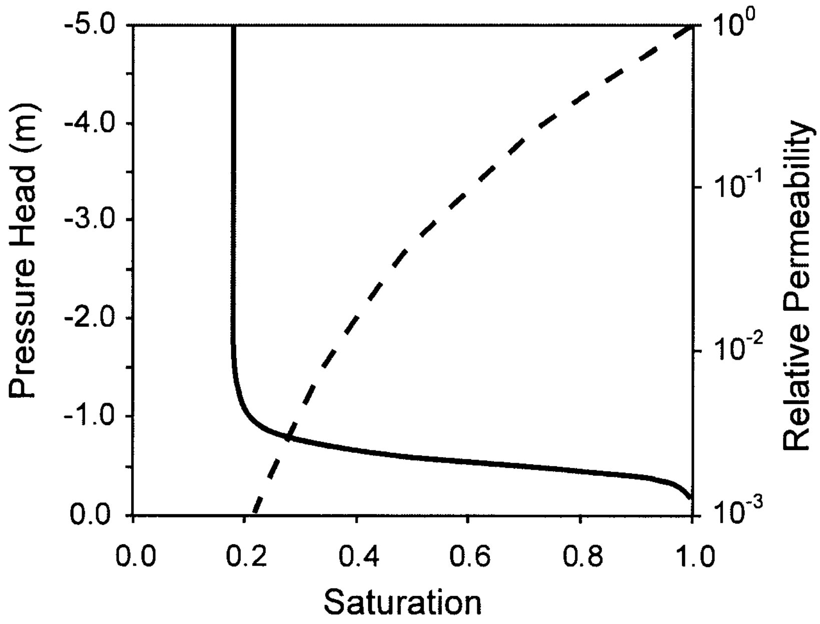

Figure 4.3. (a) Hydraulic parameters for each porous media zone, and (b) pressure head and relative permeability curves for sand and gravel zone.

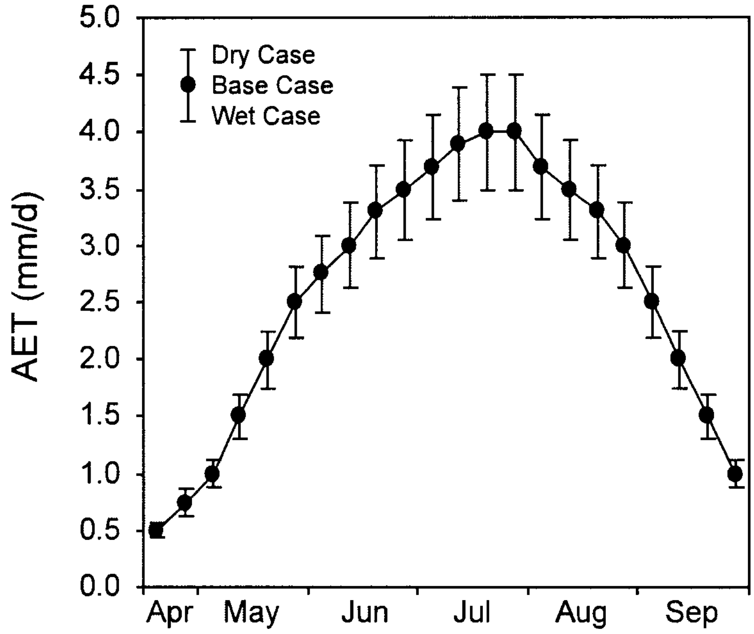

Figure 4.4. Specified weekly AET, based on Amiro et al. (2006). Solid dots represent values used in *Base Case* simulations, and the error bars denote *Dry* and *Wet Cases*. The *Base Case* represents 400 mm AET annually (+/- 50 mm for *Dry* and *Wet Cases*).

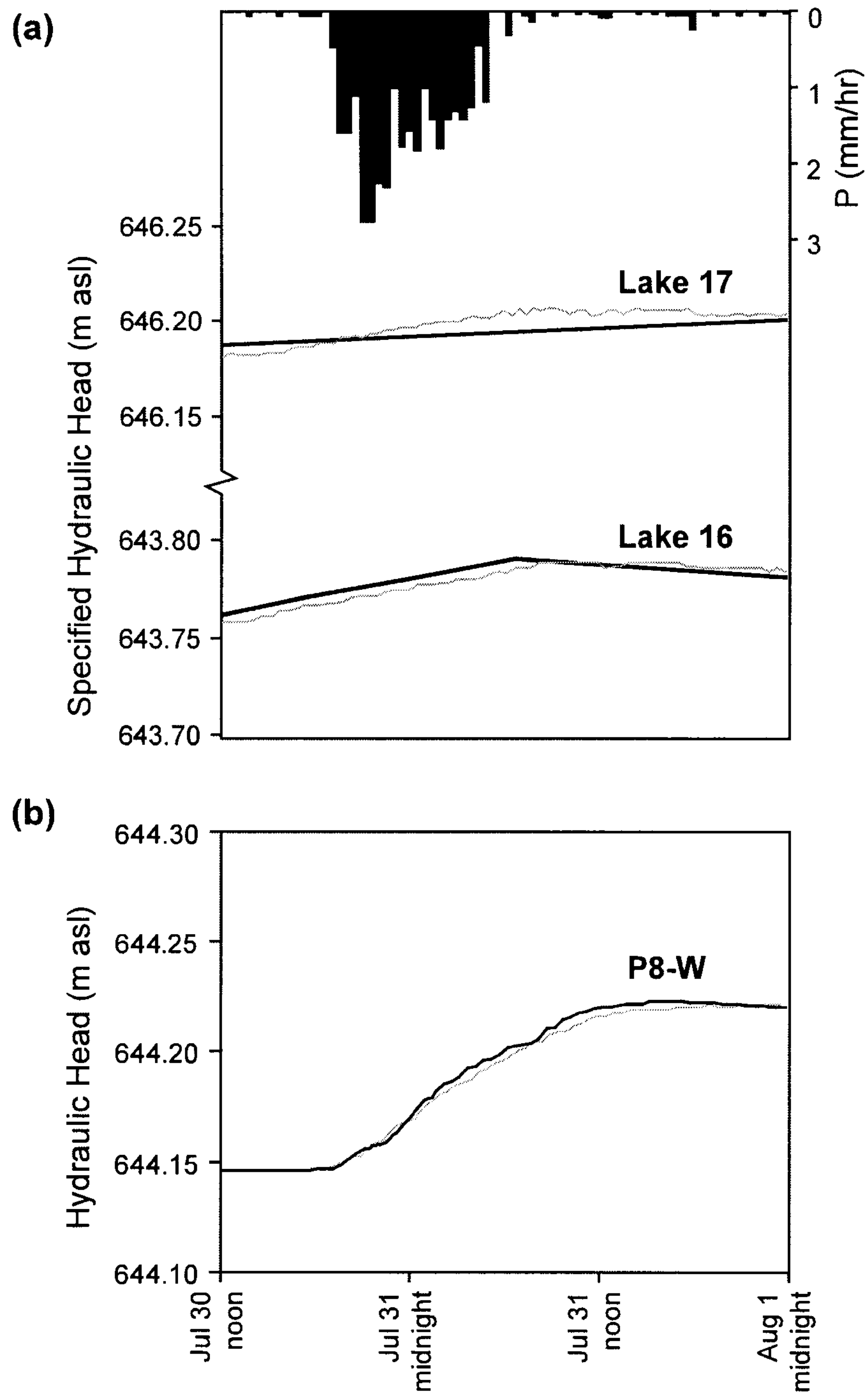

Figure 4.5. (a) Boundary conditions for *Present Day A* model, including precipitation and specified hydraulic heads for Lakes 16 and 17. Thick black lines represent time series of average specified hydraulic heads to represent the continuous field record (thin grey lines). (b) Time-series of simulated (thick black line) and observed (thin grey line) hydraulic head response for P8-W.

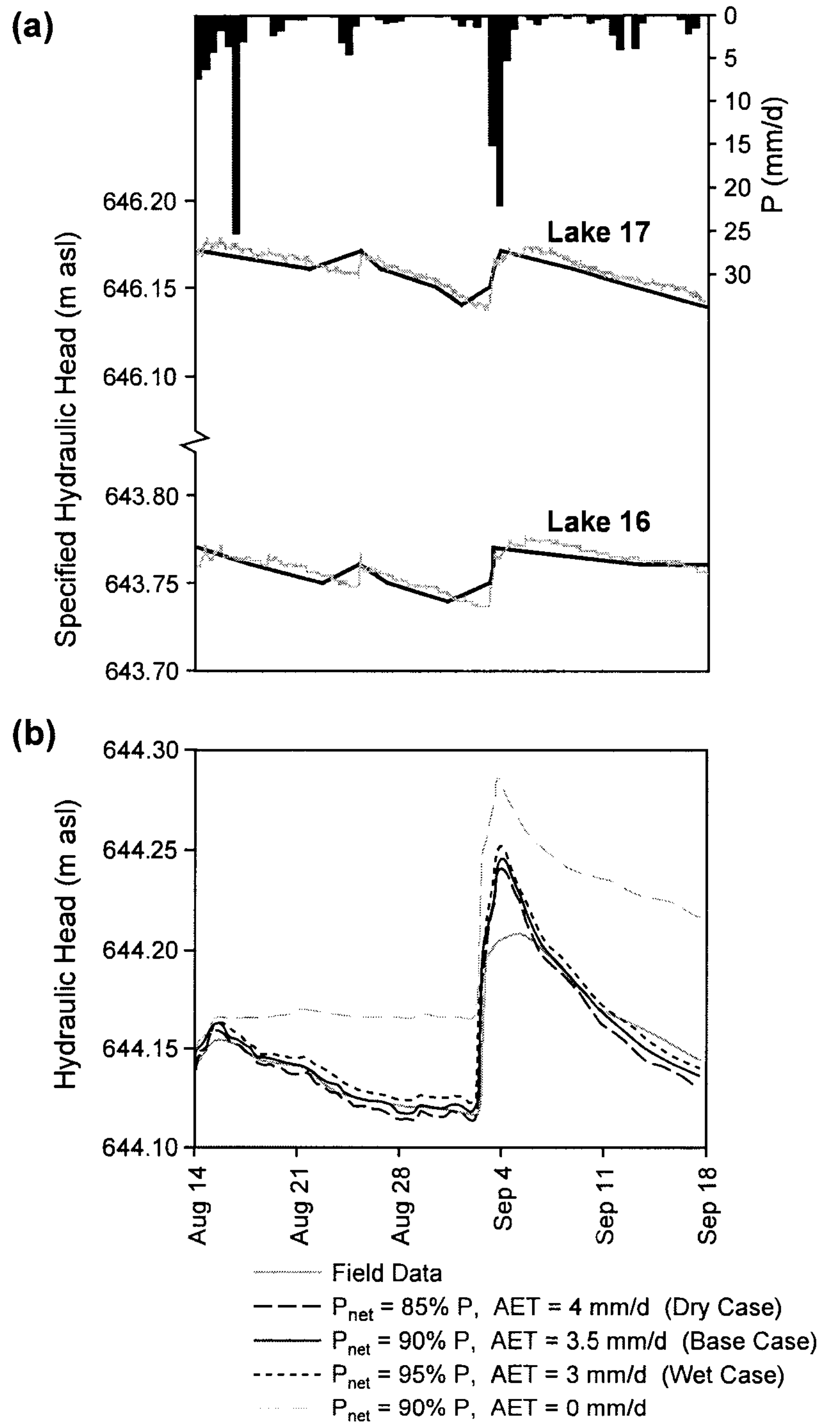

Figure 4.6. (a) Boundary conditions for *Present Day B* model, including daily precipitation and specified hydraulic heads for Lakes 16 and 17. Thick black lines represent time series specified hydraulic heads from the continuous field record (thin grey lines). (b) Sensitivity of simulated hydraulic head response at P8-W to different P_{net} and AET values.

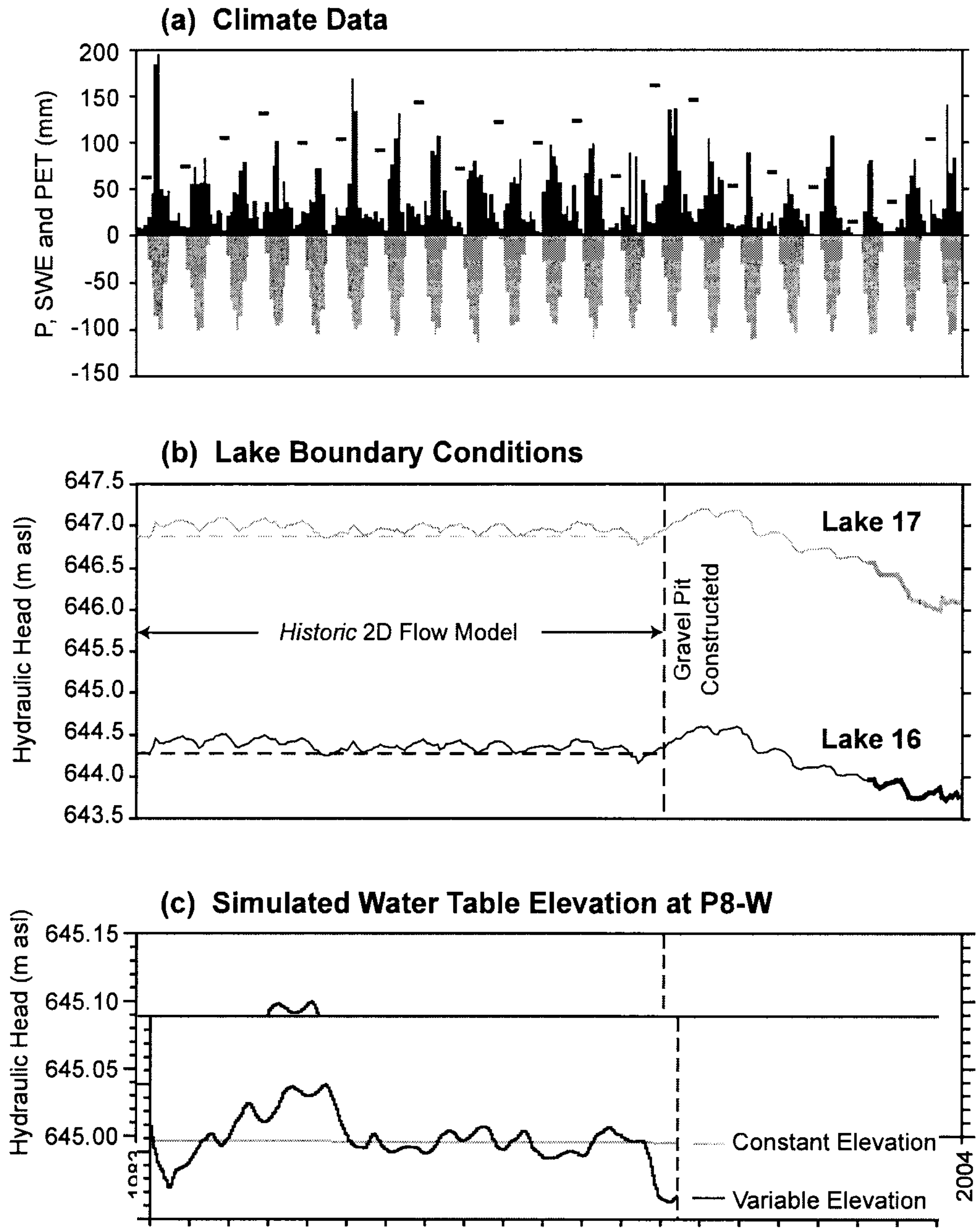

Figure 4.7. (a) Monthly precipitation, potential evapotranspiration (grey bars, calculated by the Thornthwaite method), and annual snow water equivalent, SWE (i.e., November to April precipitation expressed as mm/yr). (b) Constant (dashed line) and time-varying (solid line) specified hydraulic head boundary conditions for *Historic* model. Variable lake elevations were estimated before 2001 (thin lines), and measured from 2001 to 2004 (thick lines). (c) Simulated water table elevation at P8-W for constant (grey line), and variable lake elevations (black line). Time of gravel pit construction denoted by vertical dashed line.

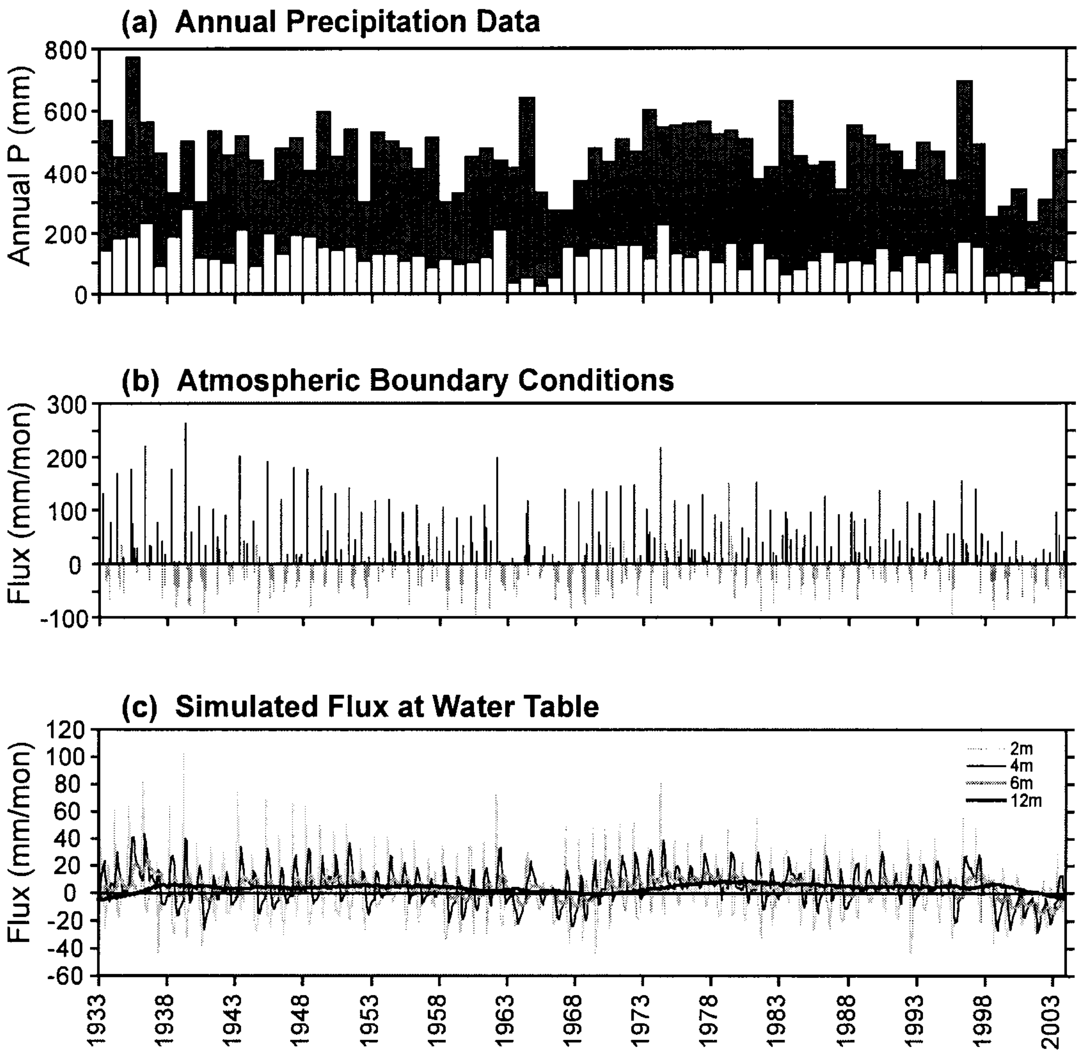

Figure 4.8. (a) Annual precipitation, separated into snow accumulation from November to April (white bars) and rainfall (black bars). (b) Net monthly atmospheric boundary conditions for the *Base Case* (positive fluxes shown as black bars, negative fluxes shown as grey bars). (c) Monthly fluid fluxes simulated for water table depths of 2, 4, 6 and 12 m below the ground surface.

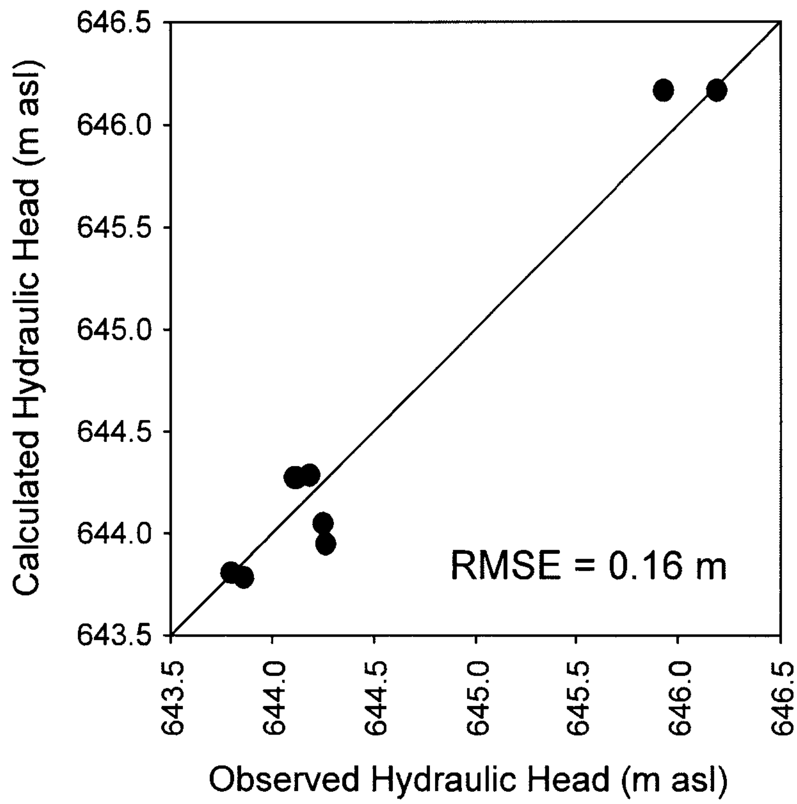

Figure 4.9. Simulated and observed hydraulic heads for July 2002 (initial conditions for *Present Day* models). Root mean squared error (RMSE) shown for 9 observations points.

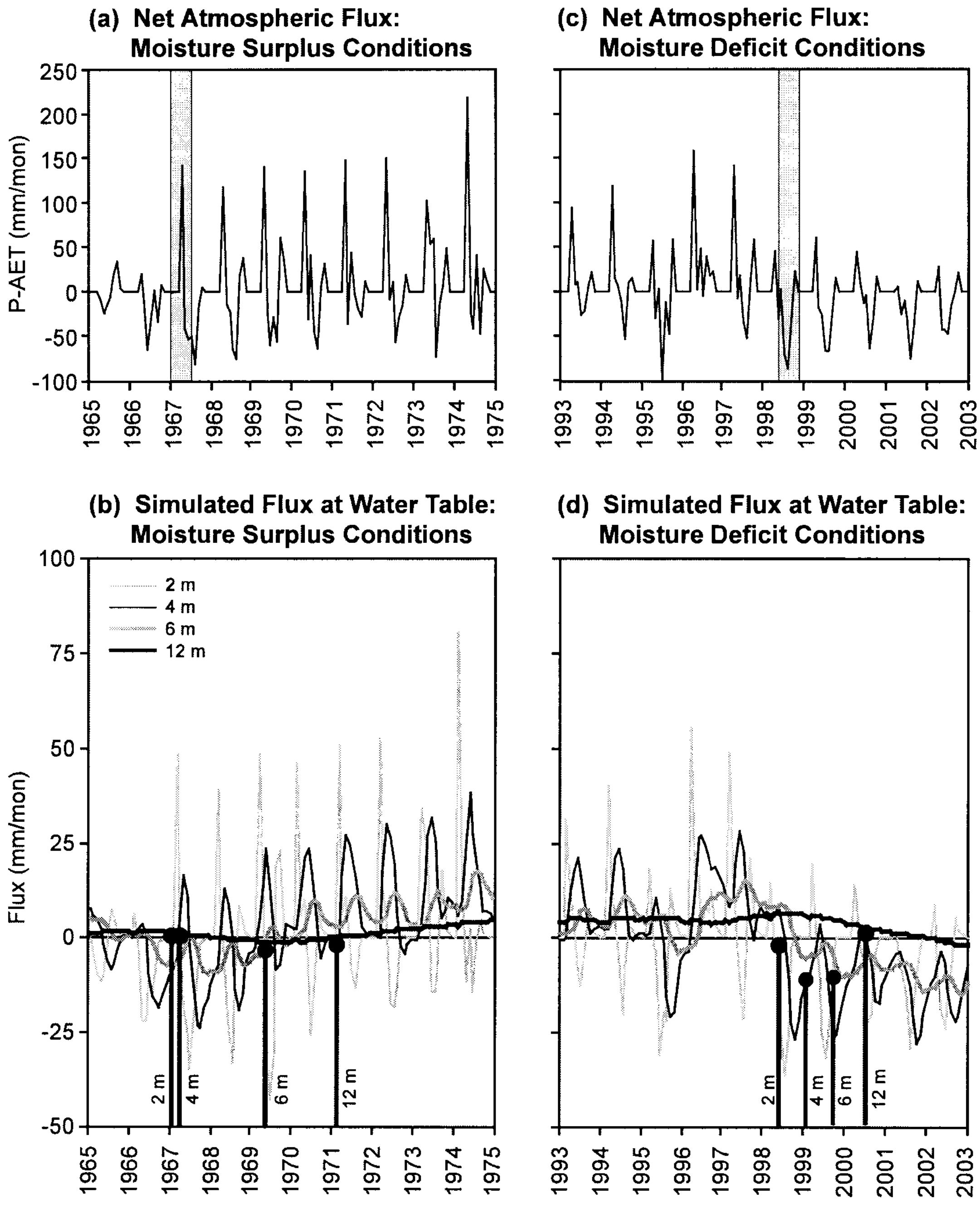

Figure 4.10. Net climatic flux (a) and simulated flux at water table depths of 2, 4, 6 and 12 m (b) for development of moisture surplus conditions in 1967. Climatic (c) and simulated water table fluxes (d) for development of moisture deficit conditions in 1998. Time lag flux at the water table associated with changes in climatic conditions identified by markers on (b) and (d).

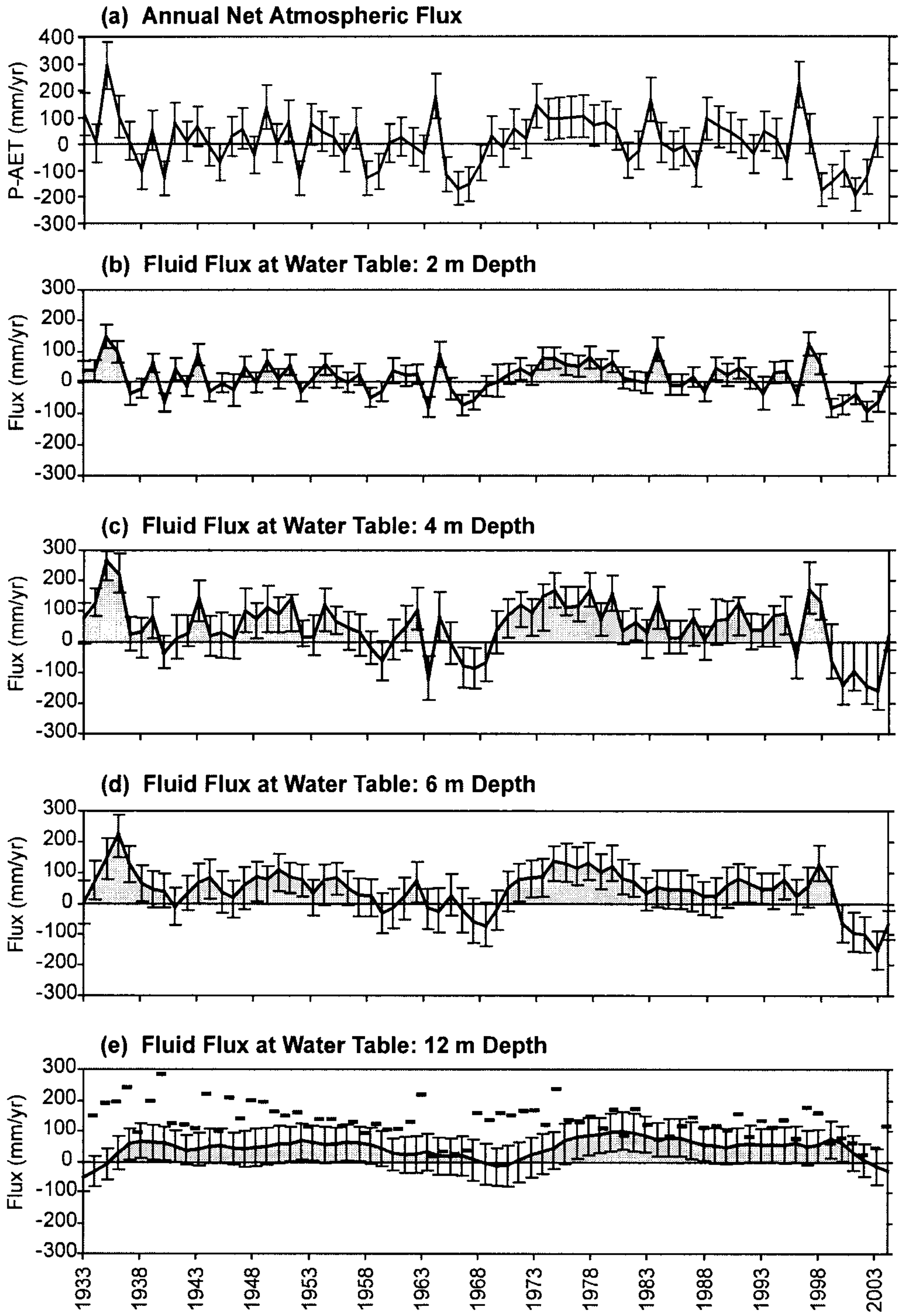

Figure 4.11. (a) Net annual atmospheric boundary conditions (i.e., precipitation minus actual evaptranspiration) for 1D models. (b) to (e) Simulated fluid fluxes at the water table for unsaturated thicknesses of 2, 4, 6 and 12 m, with duration of groundwater recharge shaded (grey). Solid lines represent the *Base Case* and error bar limits denote results of *Dry* and *Wet Cases*. Annual snow water equivalent shown as bars on (e).

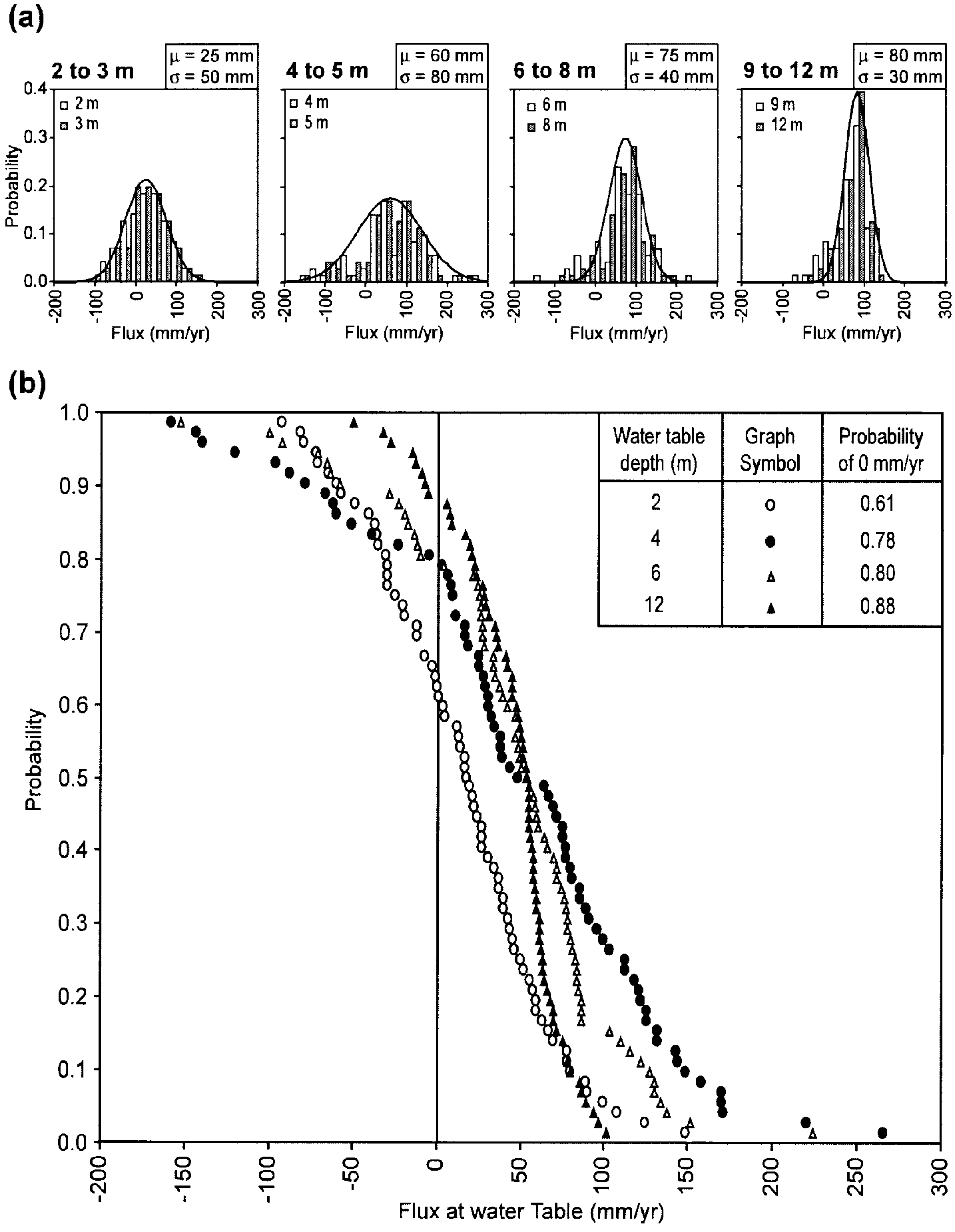

Figure 4.12. (a) Probability distributions of annual fluxes for various water table depths (shaded bars). Groupings based on water table depth ranges that have similar mean flux values (μ) and standard deviation (σ), assuming a normal distribution (solid lines). (b) Probability of exceeding minimum fluid flux for water table depths of 2, 4, 6 and 12 m. Positive values indicate conditions of groundwater recharge and negative values indicate conditions of upflux from the water table.

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Chapter 5

Predicting the effects of landscape disturbance on the Boreal Plains: Steady state versus transient analyses

5.1 Introduction

The Plains region of the Western Boreal Forest extends across three provinces in Canada, and is undergoing rapid rates of natural resource development in the forestry and energy industries (Alberta Environmental Protection, 1998). Such anthropogenic disturbances will have an impact on natural water cycling and aquatic ecosystems on the Boreal Plains, which are sustained in a delicate balance between climate and the interaction with groundwater and soil moisture storage (Devito et al., 2005; Smerdon et al., 2005; Chapter 2). Effective management of hydrologic systems on the Boreal Plains, and prediction of impacts from landscape disturbance (e.g., road construction, forestry), landform reclamation (e.g., following open pit mining and forest harvest), and climate change, will require a thorough understanding of historic and present water cycling processes. Through understanding and modelling historic and present conditions accurately, suitable modelling strategies can be developed to assess future hydrologic response (Chapter 3).

Water cycling on the Boreal Plains is sensitive to conditions of a sub-humid climate (Devito et al., 2005), and has been studied at the Utikuma Research Study Area (URSA) in northern Alberta (Chapter 2). Previous regional surveys and hydrologic modelling of Boreal lake catchments (Gibson et al., 2002) have relied on isotopic mass balance techniques that assumed steady state conditions. However, for lake-dominated environments, hydrologic response has been shown to be transient (e.g., Anderson and Cheng, 1993; Krabbenhoft and Webster, 1995; Gibson et al., 1998; Smerdon et al., 2005; Chapter 3). Although simplified, steady state models are more easily applied in landscape management practices, there is a need to ensure that key hydrological processes are represented, which can be established by testing models at well characterised field sites (e.g., Loague and VanderKwaak, 2002; Reddy et al., 2006).

The objective of this study was to examine the impact of different conceptual flow models when using stable isotopes in water balance models. To meet this objective, the influence of landscape disturbance on the water source to a shallow flow-through style lake (Born et al., 1979) on glacial outwash terrain was investigated at an URSA field site. In addition to a previous lake water budget (Smerdon et al., 2005; Chapter 2), numerical models have been developed to investigate the three-dimensional flow system (Chapter 3) and the influence of sub-humid climate on groundwater recharge (Chapter 4). The present study builds on these earlier studies and was constrained to a groundwater flow system adjacent to a small lake at the URSA. Hydrometric and isotopic field data supported development of models to test the assumptions of steady state when predicting movement of an isotopic tracer ($\delta^{18}O$), and potential changes to a flow system following landscape disturbance.

5.2 URSA Field Data

5.2.1 Lake 16 Study Site

Lake 16 is situated on a 200 km^2 coarse-textured glacial outwash plain within the URSA (56°6' N, 116°32' W; Figure 5.1), on the Boreal Plains of Canada. The regional climate is sub-humid (Ecoregions Working Group, 1989) with average annual precipitation (P) and potential evapotranspiration (PET) of 481 mm (Environment Canada, 2003) and 518 mm (Bothe and Abraham, 1993), respectively; however, annual values are highly variable (Devito et al., 2005; Chapter 4). The outwash landscape is hummocky, with remnants of an east to west trending esker, and a maximum topographic relief of 30 m.

The study site contains three lakes that exist in a series of 'steps', where Lake 17 was 2.4 m higher than Lake 16, which was 1.8 m higher than Lake 5 in early 2002 (Figure 5.1). Immediately southeast of Lake 16 (Figure 5.1), the natural landscape was disturbed when mineral aggregate (sand and gravel) was mined from the east to west trending esker (Figure 5.2). From an analysis of historic periodic aerial photographs of the study area (available from the 1940s to 2002) most of gravel pit development occurred in 1996, and the excavated area appears to have remained approximately constant. There was some further extraction during the spring of 2001, to support highway construction and access road maintenance in the region, at which point it was observed that excavation depths extended to within a couple of metres of the water table.

Hydrometric data were collected from April 2001 to April 2004 during the field campaign described in Chapter 2, and included measurements of hydraulic head at 70 piezometers, lake levels, and estimates of groundwater seepage. Groundwater was found to move from southeast to northwest (Figure 5.1 and 5.2) at an average horizontal hydraulic head gradient of 0.002. Groundwater springs were observed along the southeast shore of Lake 16 and were found to supply the lake with a consistent source of water throughout the year.

5.2.2 Hydrologic Data

The stages of Lakes 16 and 17, and the water table elevation in the vicinity of the gravel pit (P8-W; Figure 5.2), were recorded with vented pressure transducers and dataloggers (GlobalWater model WL-14) every 60 minutes from 2001 to 2004. Hydraulic head and the vertical hydraulic head gradient relative to Lake 16 were measured manually, once or twice per month during the ice-free months (April to October) from 2001 to 2003, at a piezometer at the interface of the lake-groundwater system (PZ-37; Figure 5.2). Groundwater discharge along the east shore of the lake was measured with seepage meters, made from plastic 55-gal drums, cut to 0.25 m high, and inserted into the lakebed (Smerdon et al., 2005). Discharge was measured three times per month during the ice-free season by pre-loading collection bags (3.8 L) with 0.5 to 1.0 L of lake water (Shaw and Prepas, 1989) and following the seepage measurement method described by Lee (1977).

From 2001 to 2004, precipitation was recorded with a weight-recording gauge at a meteorological station 10 km west of the site. Other climate data (including air temperature and precipitation prior to 2001) were recorded at an Environment Canada meteorological station in Slave Lake, Alberta, located approximately 100 km south of the Lake 16 study site (Environment Canada, 2002).

5.2.3 Stable Isotopic Measurements

Samples of surface water (from Lakes 16 and 17) and groundwater were collected in the fall, spring, and summer months from October 2001 to April 2004. Groundwater discharge samples were obtained from the shallowest intake of PZ-37 (installed 0.5 m below the lakebed) using a portable vacuum pump, and from seepage meter 2 (SM2) using pristine sample bags that were not subjected to the preloading technique described previously. Deeper piezometers within the flow system (P8-7 and P8-15) were sampled twice in 2002 using an inertial-lift tubing system (Waterra). All samples were stored in 20 mL scintillation vials with a thin film of mineral oil to prevent evaporation.

The Isotope Science Laboratory at the University of Calgary determined stable isotope ratios for hydrogen and oxygen. Hydrogen isotopes were determined by the chromium reduction and mass spectroscopy technique (2 per mil error), and oxygen isotopes were determined using the CO_2-H_2O and mass spectroscopy technique (0.2 per mil error). Results were expressed as per mil difference (‰), relative to Vienna Standard Mean Oceanic Water (VSMOW).

5.3 Predicting Landscape Disturbance

5.3.1 Groundwater Flow and Isotopic Tracer Model

The assumption of modelling a flow system as steady state was tested by simulating fluid flow and transport of an isotopic tracer through the outwash sediments, between Lakes 17 and 16 at the URSA. Results of different simulation models with steady and unsteady boundary conditions (hydraulic and isotopic tracer) were compared with data from the field site. The objective of the simulations was to determine whether field observations of the stable isotopic composition of groundwater discharge to Lake 16 were adequately simulated using steady state approximations, or whether a full transient analysis, that included migration of an isotopic tracer described by advective-dispersive processes, was required. Four separate simulations were considered:

- *i)* ***Constant*** *lake levels and* ***constant*** *isotopic tracer sources*
- *ii)* ***Constant*** *lake levels and* ***variable*** *isotopic tracer sources*
- *iii)* ***Variable*** *lake levels and* ***constant*** *isotopic tracer sources*
- *iv)* ***Variable*** *lake levels and* ***variable*** *isotopic tracer sources*

Field monitoring at the Lake 16 study site began in 2001; however, landscape disturbance (i.e., construction of the gravel pit) occurred in 1996. Thus, the influence of landscape disturbance was investigated by simulating the groundwater flow system, and migration of a passive isotopic tracer ($\delta^{18}O$) from January 1983 to October 2003.

Simulated $\delta^{18}O$ discharging to Lake 16 was compared with results of $\delta^{18}O$ analyses for groundwater at PZ-37 (i.e., at the location of groundwater discharge at the field site).

The saturated portion of the groundwater flow system on Figure 5.2 was modelled for a cross-sectional area of 12 m by 900 m (Figure 5.3a). Simulations were completed using the HydroGeoSphere numerical code (Therrien et al., 2005), which solved the groundwater flow equation and transport by advection and dispersion by the finite element method. Finite elements were uniformly spaced every 1 m in the horizontal direction and every 0.5 m in the vertical direction. Subsurface hydrostratigraphy was represented between Lakes 16 and 17 by sand and gravel overlying sand (Chapter 4; Figures 5.2 and 5.3a). Each zone was assigned a hydraulic conductivity (K_x) value, which was measured in the field, and estimates of specific storage (S_s), porosity (ϕ), the ratio of horizontal to vertical hydraulic conductivity (i.e., K_x:K_z), and dispersivity (α) in longitudinal and transverse flow directions (Figure 5.3b). Numerical solutions to the advection-dispersion equation may be sensitive to dispersivity values (especially the ratio of longitudinal to transverse dispersivity); however, variation of these parameters was not investigated in detail in this study.

5.3.2 Fluid Flow Boundary Conditions

The effect of a sub-humid climate (Chapter 4) and effect of landscape position (Chapter 2; Cheng and Anderson, 1994) caused lake levels to fluctuate through time and caused water table configuration to depend more on adjacent lake levels and limited recharge, than on ground surface topography (Haitjema and Mitchell-Bruker, 2005; Winter, 1986). These factors establish a transient groundwater flow system between Lakes 17 and 16 at the URSA. Thus, Lakes 17 and 16 were simulated as specified hydraulic heads, having either constant, or time-varying levels (Figure 5.4b), to investigate the effect of steady state and transient flow boundary conditions on simulating movement of an isotopic tracer. Specified hydraulic heads were applied to the lake area, and left- and right-hand sides of the model domain (Figures 5.3a). Definition of Lakes 17 and 16 as specified hydraulic heads implicitly considered precipitation and evaporation on these open water areas, and constrained hydraulic head values in solution of groundwater flow equation. These constraints imposed a groundwater flow field based on the conceptual models presented in Chapter 2, and allow comparison between steady state and transient analyses (and isotopic tracer migration).

Time-series hydraulic head values recorded by the pressure transducers and dataloggers were input as linear time-varying values for Lakes 16 and 17. Lake levels prior to 2001 (i.e., before field study) were estimated from field records, assuming a relationship between average monthly precipitation (P) and potential evapotranspiration (PET) using historic climate data (Figure 5.4a) described in Chapter 4. For scenarios of constant lake level, Lakes 16 and 17 were arbitrarily assumed to be 643.8 and 646.2 metres above sea level (m asl), respectively (equivalent to July 2002 levels). Initial hydraulic head conditions were determined by running the model to equilibrium conditions with the fixed head boundary conditions of 643.8 and 646.2 m asl (i.e., the beginning of the transient simulated timeframe), and assuming that the average sum of climatic fluxes (i.e., P and AET) was zero (Chapter 4).

Net monthly fluid fluxes were applied to the water table, as positive and negative values depending on the thickness of the esker sediments overlying the water table (Figure 5.4). These fluid fluxes represent an integration of moisture transfer at the ground surface and within the unsaturated zone (i.e., groundwater recharge or upflux conditions; Jaber et al., 2006), which were determined for January 1983 to October 2003 in Chapter 4. For areas that were undisturbed (Figure 5.2), fluid fluxes determined in Chapter 4 were specified for January 1983 to October 2003 (i.e., the complete simulated timeframe). For the gravel pit area (Figure 5.2), fluid fluxes determined in Chapter 4 were specified for January 1983 to April 1996, representing the unsaturated, forested esker. From May 1996 to October 2003, only precipitation fluxes were specified for the gravel pit area, which was bare mineral soil and assumed to have no evapotranspiration. This approach allowed the simulation of a landscape that had a change to the surface topography, and thus fluid fluxes at the water table, to be represented within one transient simulation model.

5.3.3 $\delta^{18}O$ Boundary Conditions

An isotopic tracer in the groundwater flow system (assumed to be $\delta^{18}O$) was simulated by advective and dispersive transport processes, which originated as either:

i) precipitation that infiltrated to the water table (groundwater recharge); or, ii) water from Lake 17 (Figure 5.3a). Each of these water sources was specified with assumed $\delta^{18}O$ values that were unique to the source. Fluid flux to the water table (groundwater recharge) was assigned $\delta^{18}O$ values based on seasonal variation, and water from Lake 17 was assumed to have either constant or time-varying $\delta^{18}O$ values. The objective of simulating constant and variable $\delta^{18}O$ values was to investigate the sensitivity that steady state versus transient boundary conditions had on predicted discharge to Lake 16.

Water sources originating from groundwater recharge were specified with $\delta^{18}O$ values determined from a simplified relationship between mean air temperature and observed $\delta^{18}O$ for precipitation at the URSA (Figure 5.6b). Samples of precipitation collected at the URSA (Figure 5.6a; Devito, unpublished data) had a seasonal distribution that was related to the mean daily air temperature for the time of sample collection (Figure 5.6b). Using this relationship, values of $\delta^{18}O$ were calculated for months when the mean air temperature was above freezing, from 1983 to 2003. The relation between air temperature and $\delta^{18}O$ was assumed to be consistent between daily and monthly timeframes, to calculate $\delta^{18}O$ values for each month. At the monthly scale, daily fluctuations were therefore averaged, creating distributions of $\delta^{18}O$ values (Figure 5.6c) that are within the range of values observed at the URSA (Figure 5.6a). The resulting $\delta^{18}O$ values were input as a Type III transport boundary condition when fluid flux values were positive (i.e., groundwater recharge conditions). If fluid flux values were negative (i.e., upflux), $\delta^{18}O$ was not removed from the groundwater zone or enriched (by evaporation).

Lake 17 was specified to have either constant or time-varying $\delta^{18}O$ (Type III transport boundary condition), to investigate the influence of steady state and transient transport boundary conditions. For the scenarios of constant $\delta^{18}O$ sources, transport boundary conditions were specified as -9‰ $\delta^{18}O$ in the upper half of the lake boundary and -19 ‰ in the lower half, corresponding to measured stable isotopic ratios for outwash groundwater (Smerdon et al., 2005). For scenarios of time-varying $\delta^{18}O$, monthly values for Lakes 16 and 17 followed an assumed seasonal cycle of isotopic enrichment and depletion (Figure 5.6c), which was assumed to be consistent each year.

In all simulations, the isotopic tracer ($\delta^{18}O$) was specified as a fixed value (i.e., Type I transport boundary condition) for Lake 16 and the left-hand side of the model domain. The sensitivity of imposing this type of transport boundary condition, to the simulated value of isotopic tracer in the groundwater discharge to Lake 16, was considered during model development. The proximity of observation point PZ-37 within the model, which corresponded with groundwater discharge to Lake 16, to the Lake 16 transport boundary, must be large to avoid influencing the result. A sensitivity analysis during model development showed that the distance from the boundary condition to the observation point must be greater than 5 m. At distances smaller than 5 m (determined by trial and error), observed tracer values appeared to mimic the boundary condition value, rather than vary depending on the upstream boundary value and Lake 16. To ensure that the Lake 16 transport boundary condition would not influence simulated isotopic values of groundwater discharging to Lake, the observation point (i.e., the point within the model that corresponded to PZ-37 in the field) was located 10 m from the Lake 16 boundary.

Initial $\delta^{18}O$ values within the model were assumed to be -9 ‰ in the upper zone and -19 ‰ in the lower zone of the model, based on field measurements at piezometer nest P8 (Smerdon et al. 2005; Chapter 2) and likely sources.

5.3.4 Predicting the Effect of Disturbance

In lake-dominated hydrologic systems, previous research has identified the complications of quantifying isotopic values for lakes (Krabbenhoft et al., 1990) that rely on multiple water budget components (i.e., differing magnitudes of surface water and groundwater inflow and outflow). Thus, by specifying either constant or an assumed seasonal distribution of $\delta^{18}O$ for transport boundary conditions in this study (Figure 5.6c), I sought to investigate the influence of how these boundary conditions might influence the predicted $\delta^{18}O$ discharge for a landscape that was disturbed.

For each different combination of hydraulic and transport boundary conditions (i.e., steady state and/or transient), simulated $\delta^{18}O$ in the area of groundwater discharge was compared with results of isotopic analyses from water samples collected from PZ-37. For the same timeframe (1983 to 2003), the landscape was assumed remain as it was prior to gravel pit construction (i.e., the sand and gravel esker was assumed to remain as part of the landscape). Each simulation was completed using the net fluid fluxes determined in Chapter 4 (i.e., no increased recharge associated with tree removal and aggregate mining), to illustrate the effect that creating a gravel pit had on groundwater discharge to Lake 16.

5.4 Results and Discussion

5.4.1 Stable Isotopic Composition

A cross-plot of the stable isotopic ratios for Lake 16 ($\delta^{18}O$ and $\delta^{2}H$), relative to the local meteoric water line (LWML) for the URSA, illustrated a cyclical shift that occurred during the study period (Figure 5.7a). From fall 2001 to spring 2004, the isotopic signature of water in Lake 16 progressively shifted from values heavier than the LMWL, to values nearly equivalent to the LWML. In each year, the shift toward the LMWL in the spring was followed by an isotopic shift away from the LWML in the summer months. A similar trend was observed in the isotopic composition of Lake 17. However, the difference in $\delta^{18}O$ between Lake 16 and 17 for May 2003 was 2.1 ‰, compared to only 0.5 ‰ in November 2001 (Figure 5.7b).

Shallow groundwater, collected from the discharge zone along the southeast shore of Lake 16 (PZ-37 and SM2; Figure 5.7b), exhibited a similar seasonal shift towards more depleted isotopic ratios (i.e., values becoming more negative over time). However, it does not appear that shallow groundwater became as enriched as Lake 16 during summer months (e.g., summer 2002: Figure 5.7b), with $\delta^{18}O$ values that were 1 to 2 ‰ lower than Lake 16. Seepage collected from SM2 had a similar $\delta^{18}O$ as groundwater collected from PZ-37 (0.5 m below the lakebed), indicating that groundwater at PZ-37 was representative of groundwater discharge to Lake 16. The similarity allowed a more continuous record of isotopic data for groundwater discharge, because water samples could not be collected from SM2 in late 2002 and 2003 because a 0.2 m decline in lake level that exposed the seepage meters above water.

These data suggest that the source of water to Lake 16 had shifted to a higher proportion of shallow groundwater that was derived from snowmelt (i.e., a progressively higher proportion of meteoric water) than prior to gravel pit construction, potentially caused by enhanced recharge in the gravel pit area. Such an isotopic shift could also be caused by an increasing proportion of lake ice (causing more depleted $\delta^{18}O$ values), a higher fraction of groundwater in Lake 16, or even long-term time-lag effects from the predominant upgradient water source (Lake 17).

Groundwater collected from greater depths in the outwash sediments (P8-7 and P8-15) had depleted $\delta^{18}O$ values (Figure 5.7b: -18 to -20 ‰) compared to the shallow groundwater discharge, and Lake 16, indicating that groundwater in the outwash was stratified into an upper zone that experienced periodic mixing, and a lower zone that appeared more consistent (Figure 5.7b). In 2002, samples of deeper groundwater varied by only 0.6 ‰ $\delta^{18}O$ between spring and fall, suggesting that seasonal isotopic processes at the land surface had not influenced deeper groundwater.

5.4.2 Steady State vs. Transient Analysis

Stable isotopic tracers have frequently been used to determine groundwater residence times (Maloszewski and Zuber, 1982; McGuire et al., 2002; Reddy et al., 2006). Output of an isotopic tracer from a watershed or groundwater system is related to input through a system response function (Maloszewski and Zuber, 1982) that describes piston flow, exponential flow, or dispersive flow (described in McGuire et al., 2002) at steady state. Input is often described by a sinusoidal function that represents the seasonal variation of an isotopic tracer (e.g., seasonal variation in $\delta^{18}O$) following the amplitude-attenuation approach, with a scaling factor that accounts for variable amounts of groundwater recharge. Reddy et al. (2006) investigated the use of this method to estimate the residence time of groundwater at a well-characterised lake-dominated glacial outwash landscape in Minnesota, and found it was primarily useful when $\delta^{18}O$ variation was not erratic or complex.

In each simulation, the $\delta^{18}O$ values of groundwater discharging to Lake 16 (i.e., at PZ-37) indicated an increase in the proportion of depleted water (Figure 5.8). These results, and corroborating field results (Figure 5.7a) suggest that the water in Lake 16 is a mixture of two distinct compositions: enriched water from an upgradient lake that experiences evaporation, and meteoric recharge water. For the most simplified representation of $\delta^{18}O$ sources in the outwash flow system (i.e., steady state), the simulated $\delta^{18}O$ of groundwater discharge to Lake 16 was nearly constant (Figures 5.8a and 5.8b: minimal effect from seasonal recharge). Following an increase in groundwater recharge in 1996, groundwater discharge at Lake 16 began to show an annual "pulse" of depleted $\delta^{18}O$ water, which was 1 to 2 ‰ below the nearly constant levels prior to disturbance. The initial shift in isotopic composition, and subsequent annual pulses of depleted water, arrived at Lake 16 in December of each year. For the scenario of constant lake levels and constant $\delta^{18}O$ water sources (Figure 5.8a), the seasonal pulse appeared to be slightly more damped than when variable lake levels and constant $\delta^{18}O$ sources were considered, which is illustrated from 2001 to 2003, during a period of lake level decline (Figure 5.8b).

For scenarios when $\delta^{18}O$ was simulated as a time-varying source, the composition of groundwater discharge was forced to have an annual cycle of enrichment and depletion (Figures 5.8c and 5.8d). With lake levels maintained at a constant level, groundwater discharge had maximum values (enriched) in January and minimum values (depleted) in July. When fluctuating lake levels were considered, the maximum and minimum $\delta^{18}O$ values arrived at the lake in December and May, respectively. In the year following construction of the gravel pit, there was a decrease in $\delta^{18}O$ for both methods of modelling lake level boundary conditions (steady state or transient). For the case of constant lake levels, there was a return to the cyclical pattern of enrichment and depletion, and $\delta^{18}O$ of groundwater discharging to Lake 16 varied less on a seasonal basis than groundwater discharging prior to construction of the gravel pit (-8.5 to -11 ‰). For the case of time-varying lake levels, seasonal pulses became less frequent, and the cyclical pattern appeared to diminish (1999 and 2000; Figure 5.8d).

5.4.3 Simulating Transport of an Isotopic Tracer ($\delta^{18}O$)

Compared to field observations of $\delta^{18}O$ for groundwater discharge to Lake 16, the simulations with steady state $\delta^{18}O$ sources had less seasonal variation than observed. Simulations that had fluctuating lake levels appeared to better replicate the decreasing $\delta^{18}O$ values that were observed for 2001 to 2003. The scenario of time-varying lake levels and time-varying $\delta^{18}O$ sources replicated field observations well. Compared to

$\delta^{18}O$ of groundwater discharging to Lake 16, $\delta^{18}O$ peak values arrived at location P8-W approximately 5 months earlier, indicating a consistent travel time of about half of a year from the edge of the gravel pit to Lake 16. At P8-W, the time series of $\delta^{18}O$ generally appeared less subdued than at PZ-37, indicating less mixing of $\delta^{18}O$ from two different sources (i.e., recharge from the land surface and water from upgradient Lake 17) than predicted for groundwater discharging to Lake 16. Deeper groundwater within the outwash sediments remained constant in each simulation, and did not appear to be influenced by seasonal variation or the presence of the gravel pit in the upper section of the saturated zone.

To adequately replicate the time-varying measurements of $\delta^{18}O$, a transport model that supported transient analysis of advective-dispersive process in the subsurface was required. Although these simulations have not addressed the sensitivity of the results to variation in the transport parameters (i.e., dispersivity values and porosity), which would be expected to alter the timing of $\delta^{18}O$ to Lake 16, the general systematics of tracer migration for time-varying input (compared to steady state) have been addressed. The next stage in modelling isotopic tracers could include a better representation of transport boundary conditions, such as: free outflow instead of specified concentration; a convertible transport boundary condition at the water table (to allow tracers to be removed by upflux conditions); and, a more rigorous analysis of sediment properties.

Transient analyses are typically not represented by isotopic mass balance models (e.g., Gibson et al., 2002; Krabbenhoft et al., 1990) or mean residence time models (e.g., McGuire et al., 2002; Reddy et al., 2006). The impact of inadequately representing a transient flow system with steady state assumptions illustrates the importance of a well-defined conceptual hydrologic model, which is often based on intensive field study and prior flow modelling. Thus, to represent mixing and migration of isotopic tracers on coarse-textured landscapes of the Boreal Forest, some complexity of a flow system must be incorporated into simulation models. The methodology employed in this study may not be readily applicable for larger-scale surveys; however, the importance of simulating isotopic time-series data with steady state assumptions, and the more realistic transient approach has been clearly demonstrated.

5.4.4 Predicting the Effects of Landscape Disturbance

An increase in groundwater recharge was assumed to be the cause of the observed isotopic shift in the groundwater discharge to Lake 16 (Figure 5.7). Although the isotopic variability observed in the most realistic simulation (Figure 5.8d) required assumptions of boundary condition data (Figure 5.6c), the effect of localized landscape disturbance could still be adequately predicted. Following creation of the gravel pit, the combined effect of having no forest evapotranspiration and a very shallow water table resulted in enhanced recharge to the groundwater regime, thus accentuating the seasonal pulse of snowmelt recharge. A higher proportion of snowmelt in the shallow groundwater zone has started to alter the isotopic composition of groundwater discharge to Lake 16. These findings illustrate the buffering capacity that high permeability deposits have on surface water bodies, by potentially having more than one water source, and also confirms the presence of an "interactive" shallow mixing zone approximately 5 m below the water table (Harvey et al., 2006).

For the scenarios without landscape disturbance (i.e., if the esker was not removed), the predicted $\delta^{18}O$ discharging to Lake 16 exhibits no fluctuation for constant $\delta^{18}O$ sources (Figures 5.9a and 5.9b). The simulated discharge follows the trend of the highest $\delta^{18}O$ values, indicating that the enhanced conditions of groundwater recharge that were caused by the presence of the gravel pit, had not occurred. When time-varying $\delta^{18}O$ sources were considered, the simulated discharge to Lake 16 retained a seasonal cycle that only appeared to vary with lake level (Figures 5.9c and 5.9d).

Increased cycling of meteoric water may have geochemical implications. Infiltrating water at the gravel pit has the potential to dissolve oxidized minerals that were present within the unsaturated esker material. Depending on the mineralogy, leached solutes (e.g., iron) could have been transported to Lake 16 within a few years of gravel pit construction, and possibly have started to alter lake water chemistry (Gibbons, 2005).

5.5 Conclusions

The interaction of lakes and groundwater on coarse-textured landscape at the URSA is transient. Temporally and spatially variable recharge, and fluctuating lake levels are controlled by the interaction of climate, a forested ecosystem, and landscape hydrostratigraphy. On the URSA outwash deposit, these factors have led to the development of a stratified flow system. Shallow groundwater, to an approximate depth of 5 m below the water table, interacts more with recent meteoric water than deeper groundwater, creating an upper zone with isotopic and geochemical mixing, and a lower zone with less mixing. To model watershed processes and simulate water cycling on the Boreal Plains, this isotopically stratified, hydrologically transient flow system must be accurately represented. Natural cycling of water, on these coarse-textured sediments includes fluctuating lake levels and variable amounts of groundwater recharge. Predicting the influence of disturbance will require models that can accommodate not only transience in the flow regime, but also a significant change to the land surface.

For the case of gravel pit development upgradient of Lake 16, groundwater recharge was enhanced following construction and may be responsible for an observed isotopic shift in the water source to Lake 16. Without forest canopy interception and seasonal evapotranspiration, enhanced recharge has the potential to dissolve mineral precipitates contained in the sand and gravel esker, and transport dissolved solutes to Lake 16 within a few years of disturbance.

The sub-humid climate, which has variable length periods of water deficit and water surplus, controls water cycling on coarse-textured deposits on the Boreal Plains. Wetlands, ponds, and lakes are maintained in a complex interaction between climate and groundwater. Relatively high landscape permeability allows groundwater to buffer hydrologic changes by readily cycling water to surface water for evaporation, and to be periodically recharged in water surplus climate conditions. Many isotopic models (mass balance and/or mean residence time) do not consider the transient characteristics of this dynamic hydrologic system. Thus, there is opportunity to develop models of flow, advective-dispersive, and reactive transport for well characterized sites on the Boreal Plains, in order to elucidate key hydrological processes that can be incorporated into more simplified models suitable for landscape management.

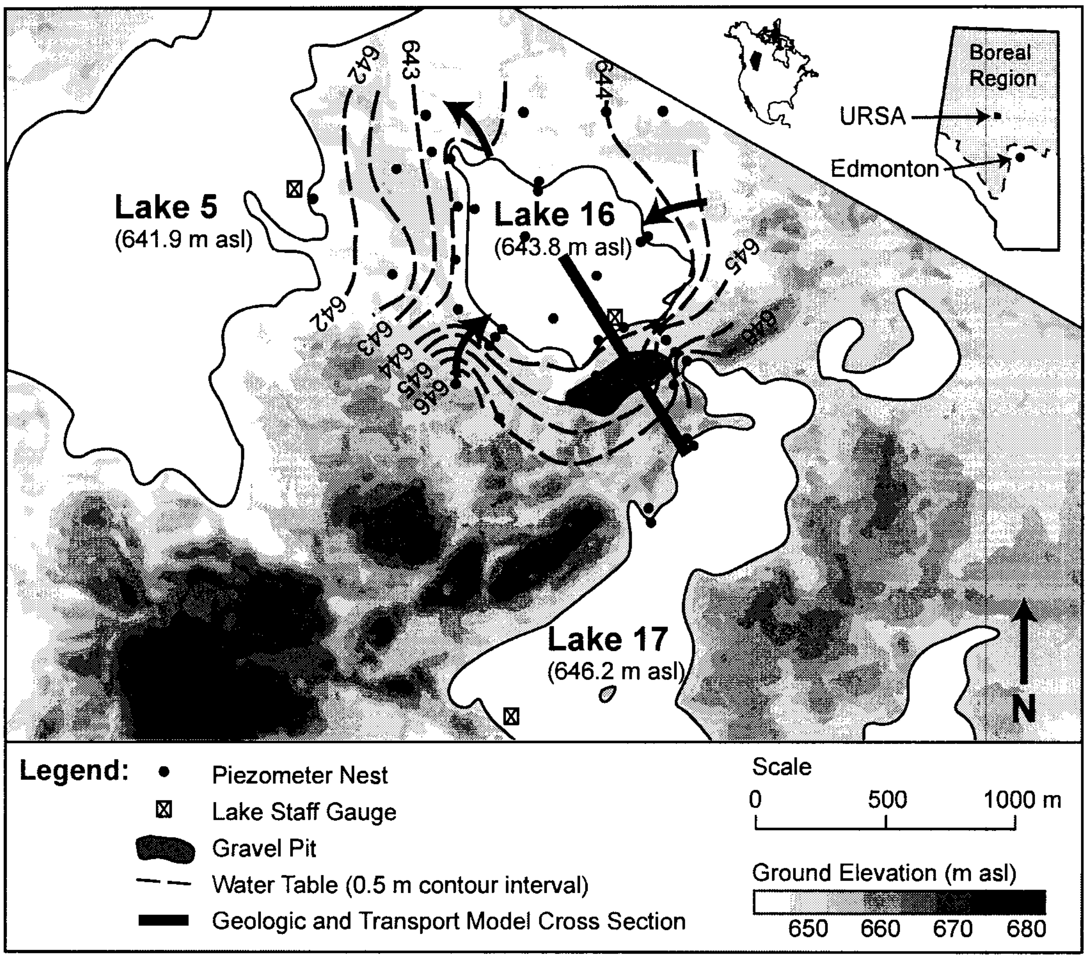


Figure 5.1. Lake 16 study site with topography, selected field instrumentation, and location of cross sectional numerical model domain. Water table contours and arrows indicating groundwater flow direction for July 2002. URSA and Boreal Plains region in Alberta, Canada on inset.

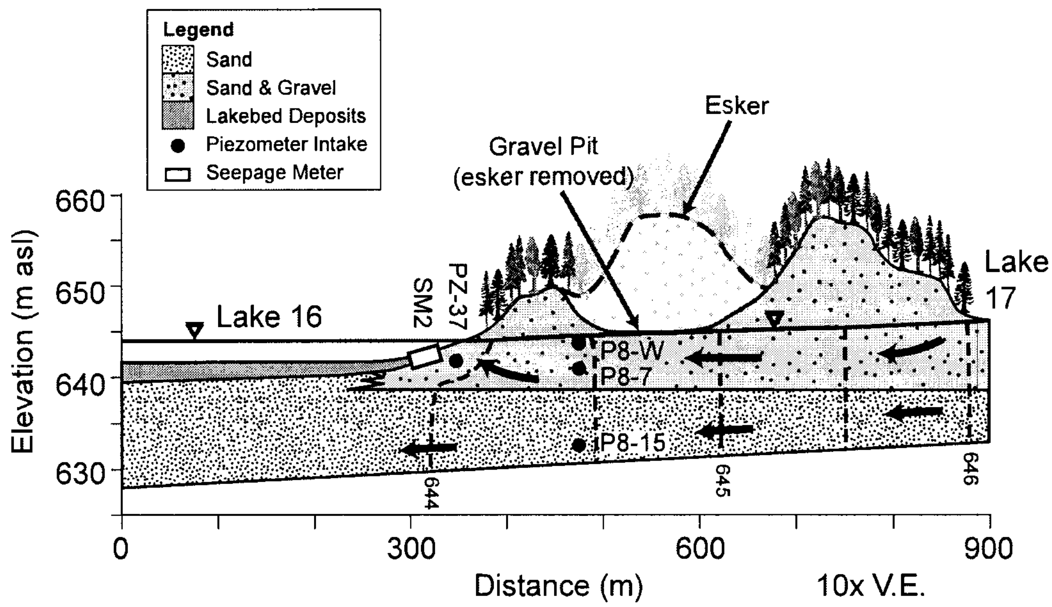
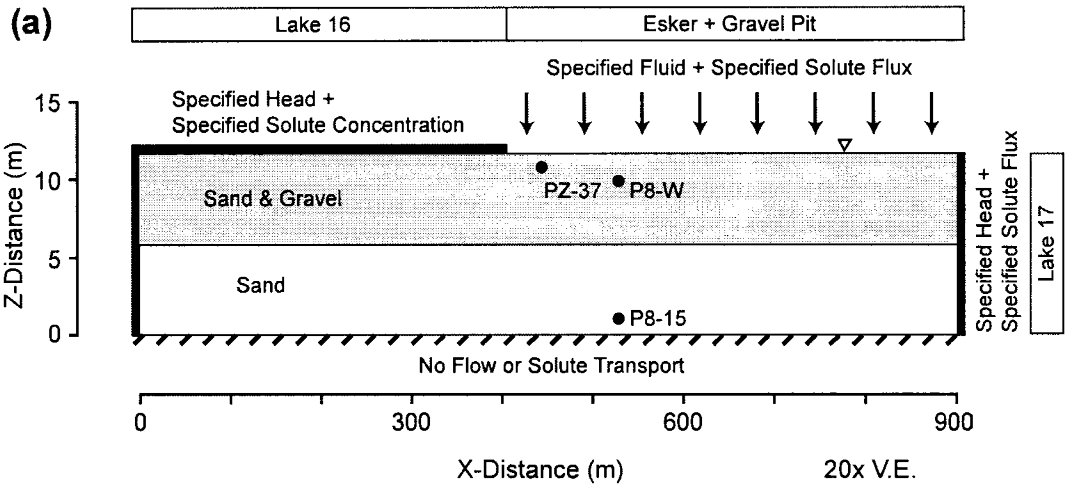


Figure 5.2. Geologic cross-section through the gravel pit area with hydraulic head distribution (0.5 m contour intervals) and interpreted groundwater flowpath (arrows) for July 2002. Ground surface topography shown with present day gravel pit and the assumed historic topography of the esker (lightly shaded).



(b)

	Sand & Gravel	Sand
K_x (m/s)	$1\text{x}10^{-3}$	$1\text{x}10^{-5}$
K_x:K_z	10:1	10:1
S_s (m^{-1})	10^{-5}	10^{-5}
ϕ	0.35	0.35
α_L (m)	3	1
α_T (m)	0.03	0.001

Figure 5.3. (a) Isotope Transport Model domain (simplified from the saturated zone shown on Figure 5.2a) with porous media zones, boundary conditions, and observations points (dots). (b) Hydraulic parameters for each porous media zone.

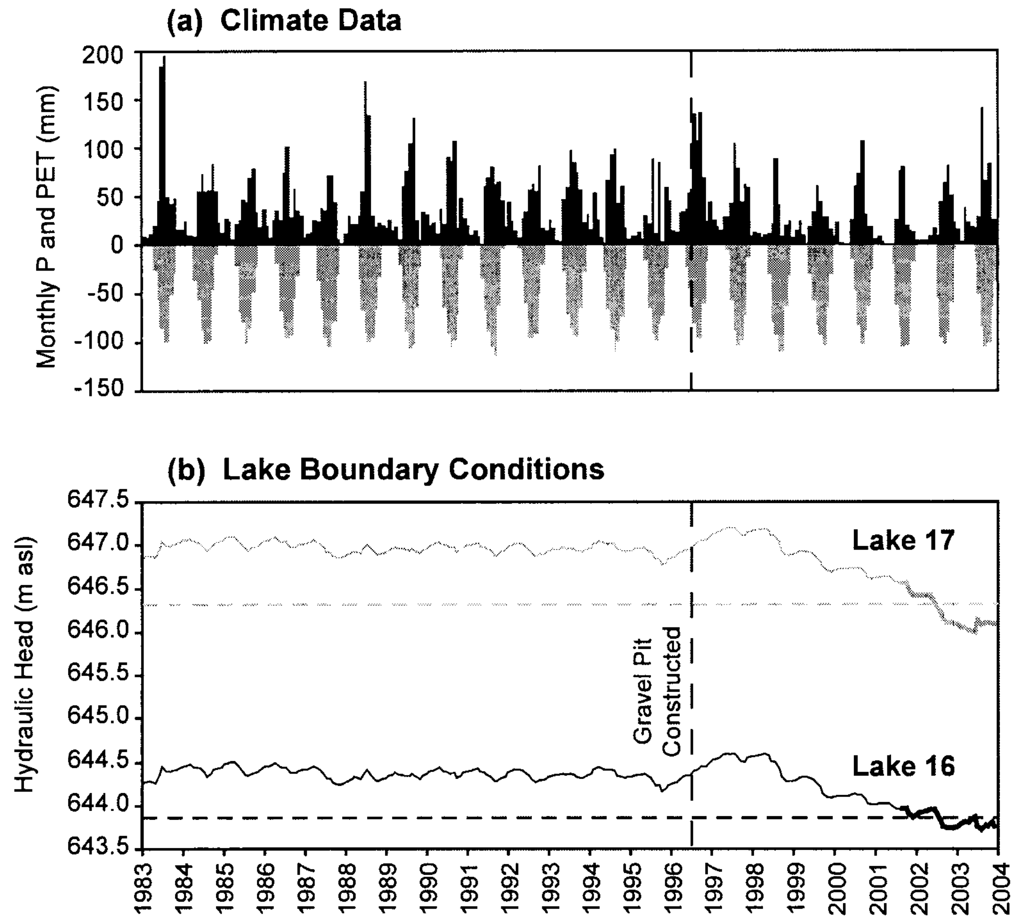


Figure 5.4. (a) Monthly precipitation and potential evapotranspiration. (b) Steady state (dashed line) and transient (solid line) specified hydraulic head boundary conditions for Lakes 16 and 17. Variable lake elevations were estimated before 2001 (thin lines), and measured from 2001 to 2004 (thick lines).

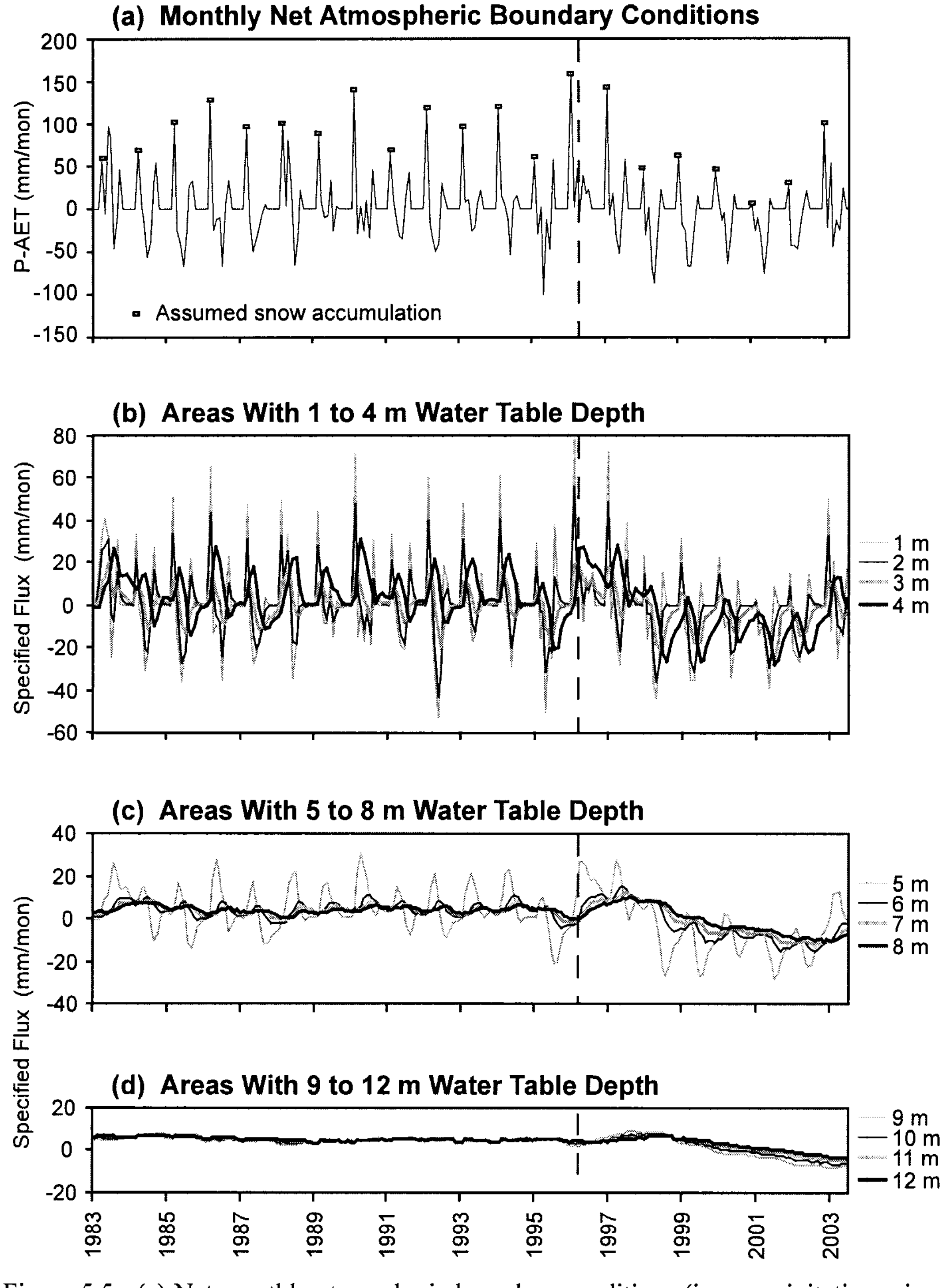


Figure 5.5. (a) Net monthly atmospheric boundary conditions (i.e., precipitation minus estimated actual evaptranspiration) and assumed snow accumulation from Figure 5.4. (b) to (d) Specified fluid flux for the water table boundary shown on Figure 5.3 for varying water table depths (1 to 12 m). Time of gravel pit construction denoted by vertical dashed line.

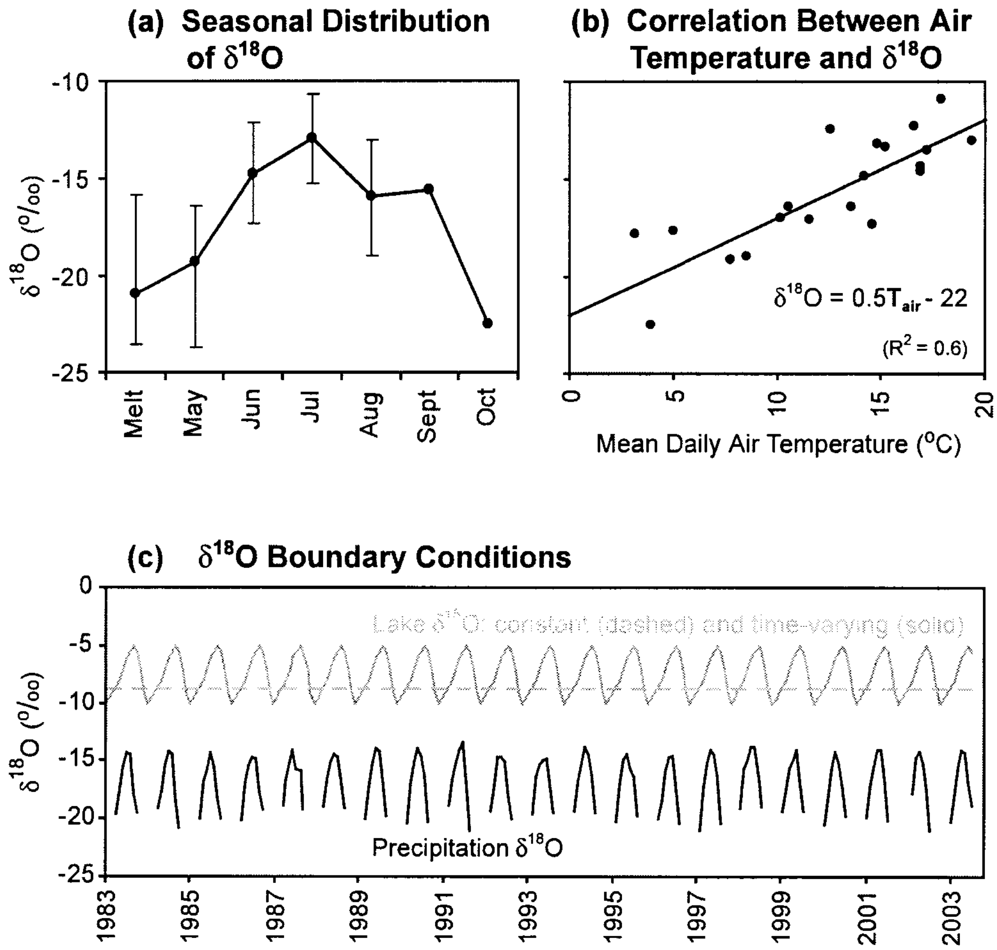


Figure 5.6. (a) Average (dots) and range (error bars) in $\delta^{18}O$ of snowmelt and monthly precipitation at the URSA from 1999 to 2004 (Redding and Devito, unpublished data). (b) Relationship between mean daily air temperature and $\delta^{18}O$ for precipitation samples above freezing (Devito, unpublished data). (c) Specified $\delta^{18}O$ boundary conditions for precipitation (black line), calculated from the mean monthly air temperature using relationship shown on (b), and lakes (grey line) for transport model. Constant $\delta^{18}O$ shown as a dashed line and time-varying shown as a solid line.

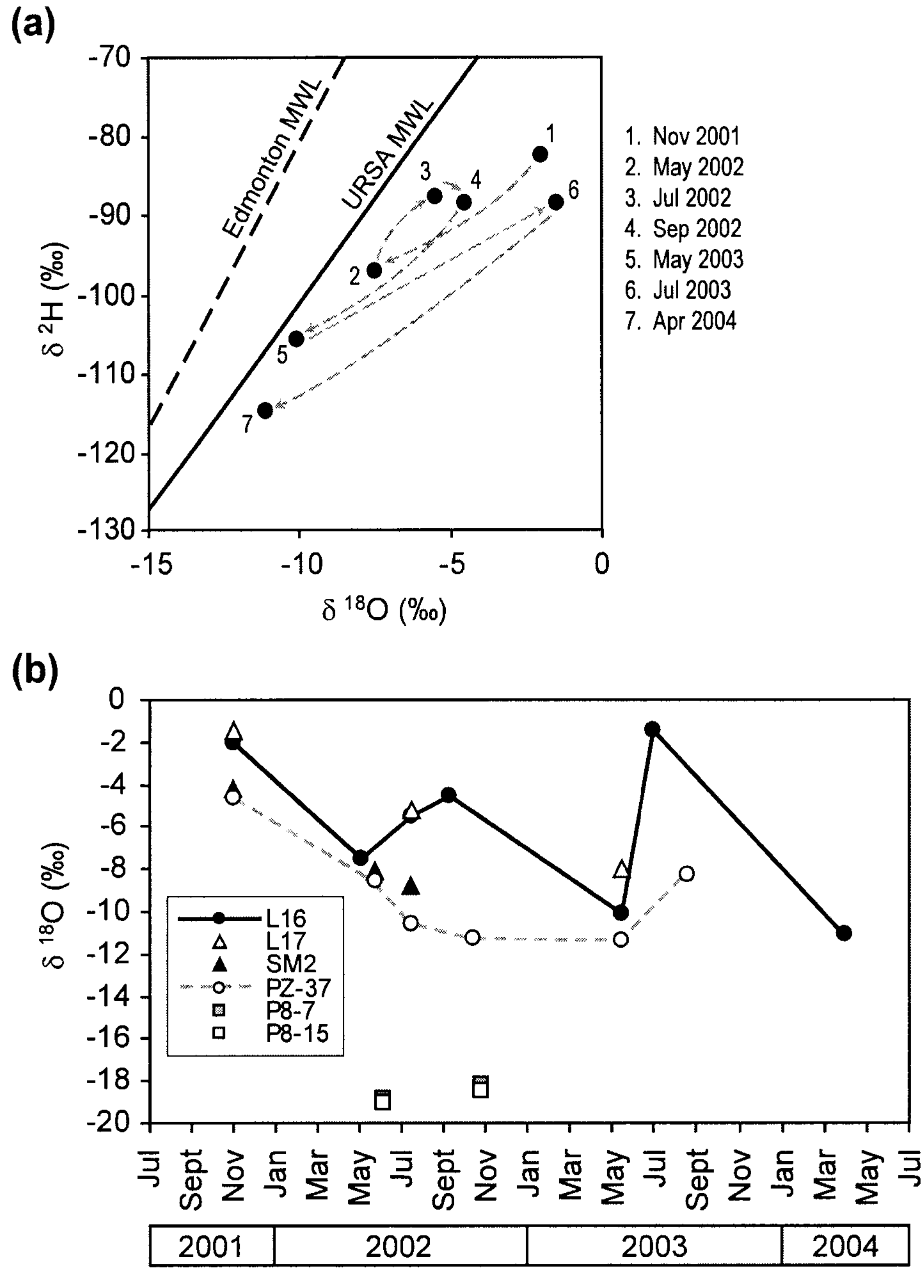


Figure 5.7. (a) Stable isotopic ratios ($\delta^{18}O$ and $\delta^{2}H$) of Lake 16 with meteoric water lines for the URSA and Edmonton, Alberta. (b) Times-series of $\delta^{18}O$ for Lakes 16 and 17, and various piezometers.

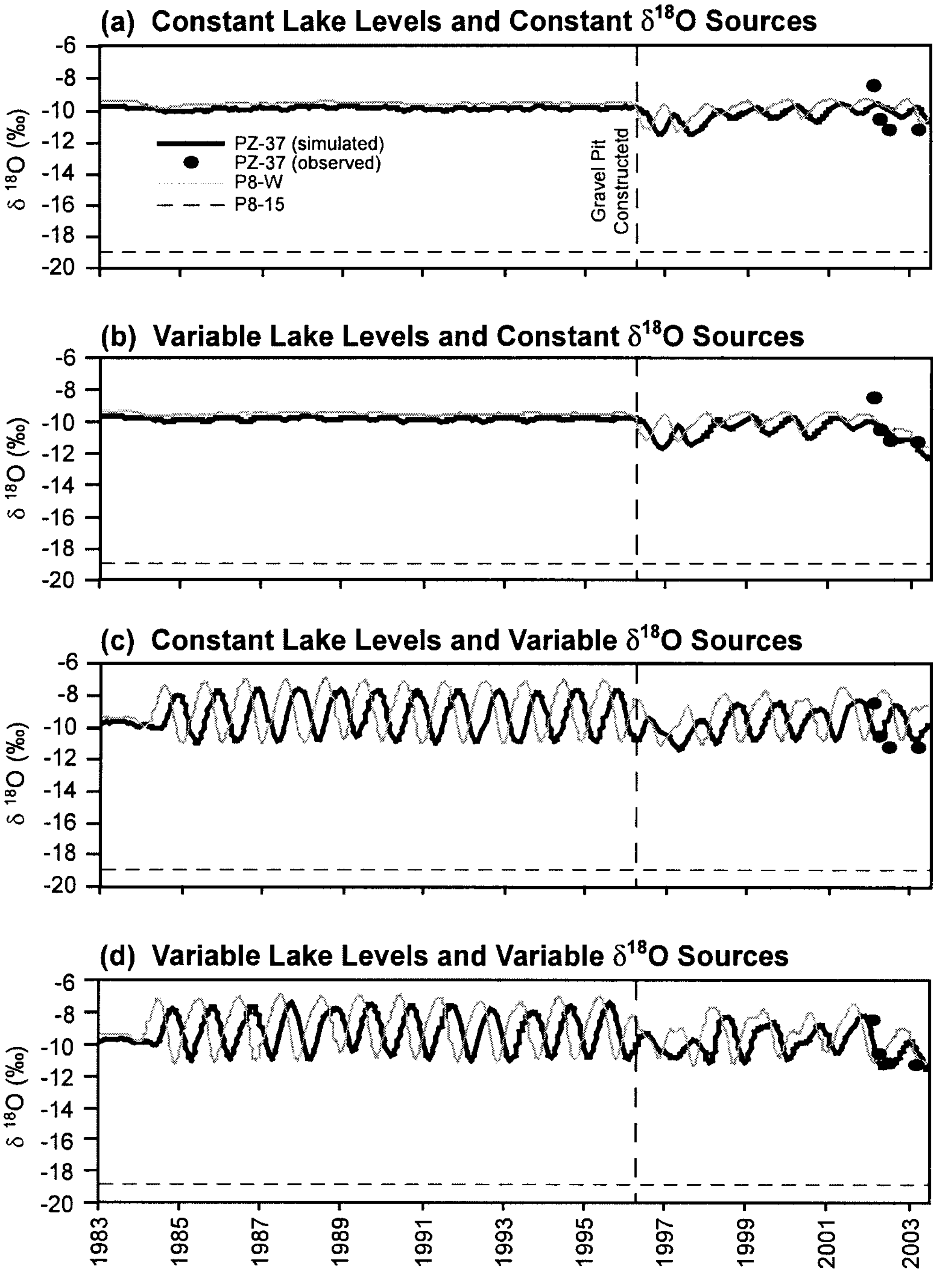


Figure 5.8. Sensitivity of simulated $\delta^{18}O$ to varying boundary conditions. Simulated $\delta^{18}O$ for shallow piezometers PZ-37 and P8-W, and deep piezometer P8-15 shown with $\delta^{18}O$ measured in the area of groundwater discharge. Effect of modelling boundary lakes as constant $\delta^{18}O$ sources (a and b) compared with variable sources (c and d), for constant lake levels (a and c) and variable lake levels (b and d). Time of gravel pit construction denoted by vertical dashed line.

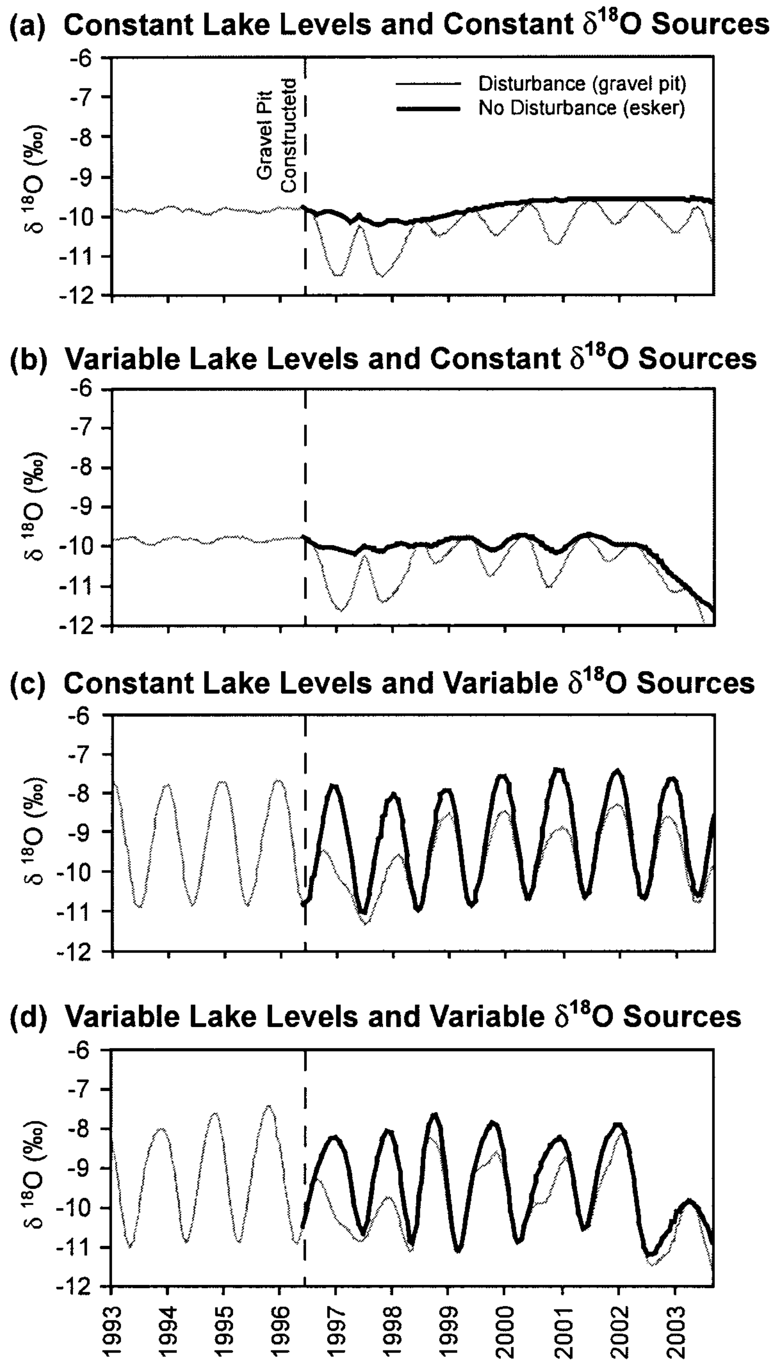


Figure 5.9. Effect of landscape disturbance on water source to Lake 16. Simulated $\delta^{18}O$ for PZ-37 using constant $\delta^{18}O$ sources (a and b) and variable sources (c and d), for constant lake levels (a and c) and variable lake levels (b and d). Time of gravel pit construction denoted by vertical dashed line.

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Chapter 6

Summary and Conclusions

6.1 Influence of Climate on Water Cycling

Hydrological processes on the Boreal Plains are controlled by climate and geology (Devito et al., 2005a). As predicted from modelling analyses (Winter, 1983) and similar to observations at a research station located on glacial outwash in a humid region (Anderson and Munter, 1981), seasonal periods of groundwater recharge from snowmelt formed small water table mounds at the URSA. Groundwater recharge and water table mounding will periodically alter hydraulic head gradients and groundwater flow directions in coarse-textured deposits, which needs to be reflected in water and nutrient budgets (Anderson and Munter, 1981). For the outwash deposits on the Boreal Plains, groundwater recharge depends on the timing of precipitation events, compared to the seasonal activity of the forested landscape. Temporal and spatial variation of forest canopy interception and evapotranspiration will govern the occurrence and magnitude of recharge at the water table. The most favourable conditions for long-term groundwater recharge occur following winter seasons that have relatively high accumulations of snow (approximately > 80 mm snow water equivalent), and a water table depth greater than 6 m. Long-term maintenance of lakes and wetlands will depend on the accumulation of snow each year in areas that have a shallow water table (< 6 m), and approximately a decade of non-drought climate conditions for areas with a deeper water table (> 6 m).

As shown in semi-arid regions, large evaporative fluxes (e.g., open water evaporation and evapotranspiration) force relatively quick dissipation of seasonal water table mounds (Rosenberry and Winter, 1997), and force the lakes and ponds in the URSA outwash to function as evaporation windows. Compared to hydrologic research stations on outwash in humid regions that have twice the annual precipitation (e.g., Cherkauer and Zager, 1989; Jaquet, 1976; Kenoyer and Anderson, 1989; Krabbenhoft et al., 1990), the sub-humid climate at the URSA causes open water evaporation to be dominant in the water budget of Lake 16. In the past 70 years, two multi-year intervals of water deficit (drought) conditions were caused by low snow accumulation. Drought conditions in 2002 lowered lake levels, which caused an increase in the contributing area of groundwater discharge to Lake 16 by 10%, and reconfiguration of the water table around the lake in the following year. These observations clearly illustrate the transient effect on outwash hydrology caused by climatic conditions. The climate on the Boreal Plains has experienced inter-annual and longer-term variation between water surplus and water deficit conditions (Devito et al., 2005b). Therefore, lake level fluctuations and water table reconfiguration are within the range of natural variability. In addition to the transient relationship with lakes and ponds, seasonal and longer-term water deficit conditions are able to cause upward migration of moisture from the water table through evapotranspiration (i.e., upflux). Although this hydrologic process has not yet been measured at the URSA, simulations suggest a high probability for upflux conditions to occur. The net result is that evaporative-dominant processes are ubiquitous on the Boreal Plains, and govern water cycling rates within the landscape.

Climatic influence on water cycling also extends into winter months. Average monthly temperatures that range from -14.6°C to 15.6°C (Environment Canada, 2003), and below freezing air temperatures for the winter months (November to March) cause open water and wetland areas to freeze each year. Although this effect on water cycling has not been addressed in this thesis, I hypothesize that seasonally frozen riparian peatlands may be a mechanism for maintaining water on coarse-textured substrates. Gibson et al. (2002) have suggested that lake ice cover, which eliminates evaporation in winter months, aids the permanence of northern lakes. Thus, a relatively cold climate influences water cycling by promoting seasonally variable lake-groundwater interaction and seasonal water storage as ice.

Water cycling through the outwash landscape is dependent on an ever-changing dominance between water sources (i.e., groundwater recharge, seasonal ice) and water sinks (i.e., open water evaporation, evapotranspiration) at monthly, yearly, and decadal timescales. Lake, pond and water table elevations; rates of surface-water/groundwater exchange; the magnitude of hydraulic head gradients; and, the migration time of water, nutrients and other solutes through coarse-textured watersheds, are entirely governed by the sub-humid climate. Each of these hydrologic components exists within the interplay between water entering, and water exiting the Boreal landscape. The results of my research contribute toward a *holistic* view of the hydrologic processes that function on coarse-textured deposits in the Boreal Forest. Specifically, my research has shown that:

- Water cycling on the coarse-textured landscapes in the Boreal Plains is governed by the difference between precipitation and evaporation. The alternating dominance of either precipitation or evaporative processes is responsible for the water table configuration, surface-water/groundwater interaction, and timing and amount of groundwater recharge.

- Integrated field and modelling studies are necessary to determine the controls on water cycling, and are needed to adequately predict response of hydrologic systems in the Boreal Plains.

- Site-specific analyses of lake-groundwater interaction, utilizing field and modelling methods, confirm conceptual models developed previously in hydrologic literature, which were often based largely on hypothetical settings.

- Detailed studies at the site-scale are essential for quantifying hydrological processes, which provide the basis for upscaling to examine the impacts of anthropogenic influences (e.g., climate change and landscape disturbance).

6.2 Modelling Water Cycling on the Boreal Plains

Representing water cycling processes on the Boreal Plains, and assessing impacts to hydrologic systems (from anthropogenic disturbances), requires models that can simulate a range of characteristics, which include: integrated surface-water/groundwater flow; a landscape dominated by widely distributed lakes, ponds and wetlands; and, a sub-humid climate. In this thesis, I tested my understanding of water cycling processes by simulating 2 years of field observations, and examined the effect of groundwater flow transience when using stable isotopes in traditional water balance models.

The hydraulic head distribution and lakebed seepage rates for the Lake 16 area were simulated with a recently developed surface-water/groundwater model (VanderKwaak, 1999). Formulation of the numerical model allowed simultaneous simulation of subsurface hydraulic heads and surface water depths, and the corresponding exchange fluxes between the surface and the subsurface, which realistically represented the hydrology of the study area, and is innovative in hydrological modelling. Lake-groundwater flow modelling identified:

- The sensitivity of parameterization of the bulk porous medium (outwash) and riparian peatlands. The glacial outwash required a lower anisotropy ratio (10:1) than previously suggested (Winter and Pfannkuch, 1984), and the riparian peatlands appeared to have seasonally variable storage and permeability values that likely correspond with the freezing and thawing process.

- The importance of spatially and temporally variable evapotranspiration from upland forests, which governs the occurrence, timing, and magnitude of recharge to the landscape.

Although simultaneous simulation of surface and subsurface flow was conceived over 30 years ago (Freeze and Harlan, 1969), only recently has this concept been realized in numerical flow models (e.g., Therrien et al., 2005; VanderKwaak, 1999). To my knowledge, the study presented in Chapter 3 represents the first application of a fully-coupled surface-water/groundwater model in a lake-dominated region, and is an intermediate step toward simulating the response of Boreal ponds and lakes to variation in climate at a larger scale.

The assumption of steady state that many isotopic mass balance models (e.g., (Gibson et al., 2002; Krabbenhoft et al., 1990) and mean residence time models use (e.g., McGuire et al., 2002; Reddy et al., 2006) was tested by simulating an isotopic shift observed in groundwater discharge to Lake 16. Simulated $\delta^{18}O$ of the groundwater discharge to Lake 16 was best replicated when transient factors, including fluctuating lake levels and seasonally variable $\delta^{18}O$ sources were explicitly considered. The results of Chapter 5 illustrated the importance of accurately representing a subsurface flow system (i.e., *honouring* the hydrology) when utilizing stable isotopic values for water budget or transit time calculations. Through use of a traditional advective-dispersive transport model, erratic and complex migration of a tracer could be predicted, which was not possible in simplified versions of isotopic mass balance and mean residence time models.

6.3 Predicting the Effects of Anthropogenic Disturbance

Process-based, site-scale field studies, such as those presented in this thesis, Devito et al. (2005b), and Ferone and Devito (2004) illustrate the natural variability in water cycling processes in the Boreal Plains. Completing field-based research is an essential step toward predicting the impact of anthropogenic disturbances to hydrologic systems and ecosystems. The effect of a typical landscape disturbance for glacial outwash (mining of mineral aggregate) was investigated by predicting changes to the water source of Lake 16, potentially caused by the presence of an upgradient gravel pit. In the few years following the gravel pit development, the effect of enhanced recharge to the groundwater regime (from logging and excavation), resulted in a shift in the water source to Lake 16. Enhanced recharge water in the gravel pit area has the potential to dissolve oxidized minerals (e.g., iron) that were present in the unsaturated esker deposit (that was exposed during mining), and potentially alter lake water chemistry. Landscape disturbance may have caused the increased levels of dissolved iron to discharge to Lake 16, which may have led to the low concentrations of soluble-reactive-phosphorous recorded by Gibbons (2005). Such a shift in the chemistry of source water will influence the aquatic ecosystem, including changes in the growth of aquatic vegetation and invertebrates, which provide shelter and sustenance to migratory birds.

Hydrologic and ecosystem response to changes in climate may be predicted using the modelling methodology described in Chapters 3 and 4. Under anticipated scenarios of climate change, a 2 to 3°C increase in the annual air temperature is predicted for much of North America, which will shorten the winter season in the northern hemisphere, and alter the near-surface hydrologic regime (Trenberth, 1999). Increased ground surface and subsurface temperatures could result in a higher frequency of mid-winter melting; a shift in the timing and magnitude of spring runoff; longer duration of evaporation from wetlands, ponds and lakes; and, alternative probabilistic functions for groundwater recharge. With an increased knowledge of present and historic water cycling on coarse-textured landscapes, the net effects on the hydrologic cycle can now be investigated.

6.4 Implications for Landscape Management and Reclamation

Effective land management practices and successful reclamation of disturbed landscapes requires understanding the hydrologic processes *unique* to the Boreal Plains ecozone. The glaciated terrain of the Boreal Plains is comprised of an assemblage of fine-textured sediments (e.g., moraines of clay and silt) and coarse-textured sediments (e.g., outwash sand and gravel), each also having extensive peatlands. Water cycling processes quantified in this thesis can be upscaled to characterize larger areas of the Boreal Plains, based on the predominant sediment soil texture (i.e., sand and gravel). At larger scales, management of hydrologic and ecologic systems can utilize the water budget components for surface-water/groundwater exchange, groundwater recharge and evapotranspiration.

Surface water on coarse-textured landscapes can be expected to function as evaporation windows (Chapter 2), and for shallow ponds and lakes, open water evaporation may be 1.2 times higher than observed in an evaporation pan (i.e., higher than potential evapotranspiration). Upland forests have an important role in partitioning precipitation into shallow soil moisture (available for transpiration) and deeper percolation (for groundwater recharge). The recharge estimates determined in Chapter 4 relied on assumed evapotranspiration fluxes from a study by Amiro et al. (2006), and further quantification of the influence of evapotranspiration on subsurface hydrology will benefit from additional study of site-specific evaporative process (e.g., Petrone et al., 2006). At larger scales, on any landscape, groundwater recharge could be modelled with the methodology developed in Chapter 4, and distributed spatially based on knowledge of coarse-scale digital elevation models that could be used to map water table depths. This approach would aid identification of areas of potentially high (and low) recharge that could be incorporated into regional models of water balance and landscape management.

For the *constructed* coarse-textured landforms that have been created following hydrocarbon extraction from the oil sands deposits in northeast Alberta (e.g., Syncrude Canada's southwest sand storage facility; List et al., 1997), an understanding of water cycling on a *natural* coarse-textured landscape will aid development of long-term reclamation strategies. Two engineering challenges at Syncrude Canada's southwest sand storage facility are centred on drainage of water-saturated tailings sand and recharge of fresh water to sustain re-vegetation and flush salts (Price, 2005). Naturally occurring outwash landscapes appear to evaporate large volumes of water to the atmosphere via open water bodies (ponds and lakes). For an engineered tailings dam that contains no internal drainage (List et al., 1997), a hummocky terrain with shallow ponds and lakes will enhance long-term water loss in warm summer months, especially if climate-warming trends continue. Consistent, long-term groundwater recharge will likely be limited to areas with a water table depth greater than 6 m; however, the sandy sediments at the Syncrude facility are finer-textured than the URSA outwash. Thus, the dynamics of recharge/upflux would require some investigation (by the modelling methodology used in Chapter 4) to adequately predict future response. These findings provide some context for flow rates and fluxes in the hydrologic cycle that can be expected for natural coarse-textured deposits in a sub-humid climate.

6.5 Recommendations for Future Research

Evaporative processes, either from ponds and lakes, or transpiration from wetland and forest vegetation are a dominant component of the hydrologic cycle on the Boreal Plains. In these research projects, estimates of open water evaporation were made for the study lake, and evapotranspiration fluxes were assumed for the simulation models. Because evaporative processes represent a link between the atmosphere and the terrestrial hydrosphere, additional research into quantifying *actual* evapotranspiration from different parts of the outwash landscape would be of interest. Refined estimates, based on field measurements and additional models of evapotranspiration, when coupled with moisture and energy transfer in the unsaturated and saturated zones in the subsurface, would allow for a more complete description of water movement and storage in this sub-humid environment. Further investigation at the hillslope and entire pond/lake spatial scale would begin to encompass some of the lateral heterogeneity present on the Boreal Plains.

Additional simulation models of the impact of landscape disturbance also warrant further investigation. Many isotopic models (mass balance and/or mean residence time) do not consider transient flow in multiple dimensions. Thus, there is opportunity to develop models of fluid flow, advective-dispersive, and reactive transport for well-characterized sites on the Boreal Plains (such as URSA Lake 16), to explore hydrogeochemical processes. The modelling presented in Chapter 5 illustrated the effect of steady state assumptions for a transient flow system. The analysis could be improved with a better representation of isotopic tracer boundaries (e.g., free outflow instead of specified concentration and a convertible transport boundary condition at the water table to allow tracers to be removed by upflux conditions), and sensitivity analysis of the parameters used to describe the porous media. A three-dimensional analysis of flow coupled with reactive transport, would aid investigation of the dissolution and migration of iron precipitates from the esker materials, which could have caused the reduced soluble-reactive-phosphorous concentrations measured in the study by Gibbons (2005). Multi-dimensional analysis would also support further investigation of the transient flow and $\delta^{18}O$ sources assumed in Chapter 5, especially if isotopic fractionation could be calculated during transient simulations.

Additional characterization of the physical properties of the outwash landscape is also justified. Knowledge of the spatial distribution of soil moisture characteristics would eliminate assumed relationships for capillary pressure, water saturation, and relative permeability. Site-specific knowledge of unsaturated flow could improve simulation models. Additional field investigation to evaluate anisotropy (using physical and tracer methods) would support assumptions in multi-dimensional models of flow and transport, especially when combined with a sensitivity analysis of dispersivity.

Lastly, field observations of riparian peatland freezing and thawing cycles corroborate the hypothesis of seasonally transient hydraulic parameters. Therefore, further study of the seasonal thermal effects (e.g., McCauley et al., 2002) on lake-groundwater exchange is warranted, especially considering that climate warming will alter the thermal regime of the subsurface in northern climates. Coupled cryospheric-hydrologic watershed processes could be investigated by simulating transient thermal and hydrologic field observations from various URSA sites. This could be achieved by adapting an existing, integrated watershed model (such as HydroGeoSphere; Therrien et al., 2005) to represent subsurface heat transport and seasonal freezing with the energy- and mass-transfer approach described by Harlan (1973), and freezing-characteristic-curve relationships (Spaans and Baker, 1996). This kind of watershed model would have unique capabilities that do not exist at the moment, and allow exploration of coupled cryospheric and hydrologic processes, and potential changes occurring from climate warming.

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