University of Alberta

Assessment of surface water-groundwater interaction at
perched boreal wetlands, north-central Alberta

by
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Abstract

Wetlands with perched water tables in the Western Boreal Forest, Alberta, were assessed using field and numerical modeling methods. These wetlands are important water resources and support diverse plant and animal communities. The study site sustains numerous wetland/aspen-upland complexes that are isolated from regional groundwater flow systems. Hydrometric measurements were made to quantify groundwater, surface water, and soil moisture regimes adjacent to and beneath the wetlands, and the hydrologic processes sustaining perched wetlands. These data and numerical simulations show that a laterally extensive, low-permeability confining layer is necessary for long-term permanence of these wetlands and that the magnitude and spatial distribution of water fluxes is controlled by the temporal and spatial distribution of soil water storage potential. This study developed a conceptual model of perched wetland hydrologic response characteristics, facilitating more accurate regional scale modeling efforts expected to help refine water resource impact assessment and conservation strategies.

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Chapter 1

Introduction

1.1 Western Boreal Forest

The Western Boreal Forest is a large ecosystem covering 35% of Canada (Canadian Forest Service, 2006) that contains nearly one quarter of the world's undeveloped forested area (Bryant *et al.*, 1997). This forest region represents an extremely important ecosystem for fresh water resources and carbon storage crucial for global climate regulation (Metcalfe and Buttle, 1999). The Boreal Plain ecozone of the Western Boreal Forest supports a variety of vegetation types (upland forest, shallow lakes, peatlands, and ponds) on three common landforms including coarse-grained glaciofluvial outwash deposits, fine-grained disintegration moraines and low-lying glaciolacustrine plains (Devito *et al.*, 2005a). This ecozone extends across the provinces of Alberta, Saskatchewan, and Manitoba in west-central Canada (National Wetlands Working Group, 1988), and is currently subject to rapid development and potential impacts from the proliferation of natural resource based industries (Alberta Environmental Protection, 1998). Recent hydrologic research completed in the Boreal Plain has shown that temporal climate patterns and geology control hydrologic processes driving wetland sustainability, vegetation patterns, and regional groundwater recharge (Ferone and Devito, 2004; Devito *et al.*, 2005b; Smerdon *et al.*, 2005; Petrone *et al.*, 2007; Smerdon *et al.*, 2007). Quantifying impacts caused by anthropogenic development in the area requires a detailed understanding of the natural variability of hydrologic processes and water cycling unique to this region, under past, present, and future climatic conditions.

The Western Boreal Plain (WBP) has many physical characteristics that necessitate different hydrologic analyses than other Boreal ecozones (Winter, 2001; Devito *et al.*, 2005a; Smerdon *et al.*, 2007). The combination of the sub-humid climate (i.e., precipitation ≤ potential evapotranspiration; Winter and Woo, 1990), low relief, deep glacial sediments (20 to 240 m thick; Klassen, 1989; Pawlowicz and Fenton, 2002), and its location south of the discontinuous permafrost zone (Woo and Winter, 1993) creates a unique hydrologic environment that has achieved a temporal and volumetric balance of atmospheric water fluxes, vegetative water demand, and soil water storage that differs from other ecozones (Smerdon *et al.*, 2005). To further investigate this hydrologic environment, a research area within the Boreal Plain ecozone has been established called the Utikuma Region Study Area (URSA), which encompasses the spectrum of physiographic landforms commonly found in the region and the WBP. Current multi-disciplinary research at the URSA includes concurrent characterization of atmospheric flux patterns, surface water-groundwater interaction, recharge characteristics, vegetative controls, and soil water storage. The determination of the influence of climatic and geologic characteristics on the ecohydrology of several wetland/upland complexes located across the geologic and physiographic gradients observed in the Boreal Plain will improve our ability to scale-up hydrologic responses to natural variability and disturbances.

Several industries are currently developing vast areas of the Boreal Plain, with major disturbances from the forestry and conventional oil/gas industries, as well development of oil sands resources (Dúcks Unlimited Canada, 2006). This rapid development of the Boreal Plain presents challenges in preserving its ecological characteristics for storing fresh water (in wetlands) and carbon, and for providing habitat for millions of waterfowl and songbirds (Blancher and Wells, 2005). A disconnect exists between the intrinsic economic and ecologic values of the Boreal Plain, requiring development of a standard conservation methodology that considers the diversity and complex hydrology of the landscape.

Current research ongoing in the Western Boreal Plain is divergent from the paradigm of topographically driven flow systems and the fundamental hydrologic landscape unit (FHLU) concept as described by Winter (2001). Completing process-based research at plot and catchment scales allows the development of hydrologic response unit (HRU; Devito *et al.*, 2005a) characteristics for a variety of land features in this environment. These characteristics will guide generalizations implemented within larger-scale hydrologic modeling, while improving our ability to predict hydrologic responses and natural variability at the landscape scale. Ultimately, this approach is intended to improve sustainable development of the Boreal Plains.

1.2 Perched wetland systems

Perched aquifers and flow systems have been identified in early hydrogeologic studies, but have infrequently been the focus of hydrologic research (Fetter, 1994). Recently, many studies are beginning to acknowledge the hydroecologic importance of perched flow mechanisms in supporting biodiversity and regional groundwater resources (Buttle, 1989; Brooks and Hayashi, 2002; Rains *et al.*, 2006; Wolfe *et al.*, 2006). Despite becoming the focus of new hydrologic research, a distinct knowledge gap in the biogeochemical and ecohydrological function of perched wetlands has been identified (Rains *et al.*, 2006). Studies have characterized ephemeral perched wetlands (vernal pools) in a wide range of climates and geological contexts, from humid shield regions in the eastern U.S. and Canada (Brooks and Hayashi, 2002), to arid loess landscapes of Idaho (O'Geen *et al.*, 2003; Rockafeller *et al.*, 2004), and Mediterranean valley climates of Oregon and California (Rains *et al.*, 2006). However, few, if any, studies have identified or characterized non-ephemeral, permanent perched wetlands in any of the aforementioned climatic and/or geologic contexts. Recently, permanent perched wetlands have been identified at the URSA within a sub-humid climate and glacial derived landforms of the Boreal Plain. There is a strong indication that these wetlands are currently important, widely distributed elements of the hydrologic landscape of the Western Boreal Forest of Canada (Wolfe *et al.*, 2006). The intent of this thesis is to study such permanent perched wetlands to advance our understanding of their ecohydrologic role in the Boreal Plain.

This study offers a unique opportunity to study the hydrologic processes that sustain permanent perched wetlands. Unlike many wetlands that may be ephemerally perched with periodic connection to a fluctuating water table, it appears that these wetlands are solely sustained by atmospheric fluxes and geological characteristics with no connection to regional flow systems. The climate, geology, and resulting hydrologic processes sustaining permanent perched wetlands at the study site are undocumented within the hydrologic literature, presenting many challenges in the interpretation of water cycling and surface water-groundwater interaction. Adaptation of conceptual models associated with previously studied perched wetlands and flow systems was required (Brooks and Hayashi, 2002; Rains *et al.*, 2006). Literature from the well-researched prairie pothole region from the Great Plains Region of North America (e.g., Meyboom, 1966; Millar, 1971; Mills and Zwarich, 1986; Hayashi *et al.*, 1998), was employed to help interpret the hydrology of the study wetlands as there appear to be similarities in their hydrologic setting and function. This study highlights the parallels between existing perched and applicable non-perched hydrologic literature, but also elucidate the differences and the fine balance of highly transient hydrologic processes that sustain permanent perched wetlands in the Boreal Plains.

This research follows a holistic, ecohydrologic conceptual/numerical modeling approach that attempts to compartmentalize perched wetland areas based on hydrologic response, which is a function of many variables as described by Devito *et al.* (2005a) and Euliss *et al.* (2004). It is hoped that this research will add to an emerging framework that will improve evolving water resource management and land reclamation practices.

This work will define the HRU characteristics for perched wetland areas. These characteristics can be utilized to improve our ability to assess long-term hydrologic response in the Boreal Plains region and predict impacts from direct anthropogenic land surface disturbances, and the potential for indirect impact from climate change. With an increased understanding of hydrologic linkages of perched wetland to adjacent landforms and underlying groundwater flow systems, we can interpret how anthropogenic impacts at a variety of scales will impact water resources across the region.

The wetlands selected for study are located in a transition zone from disintegration moraine to a large (200 km^2), outwash sand and gravel deposit located at the URSA. The study site lies between a well-characterized chain of regional groundwater fed outwash lakes (Gibbons, 2005; Smerdon *et al.*, 2007), and moraine ponds with previous and ongoing studies (Ferone and Devito, 2004; Petrone *et al.*, 2007). The apparent isolation of the perched wetlands from the deep regional water table warranted further study to assess the near surface hydrologic processes that sustain the upland/wetland complexes as permanent wetland features.

1.3 *Thesis objective and format*

The objective of this thesis is to enhance our general understanding of perched wetlands, and to fill a knowledge gap regarding boreal wetland sustainability. Two permanent perched wetlands with contrasting accumulation of organic sediments (i.e., a fen and a shallow pond-marsh), were selected for detailed hydrological study. This thesis allowed the following research questions to be addressed:

1) What are the hydrologic processes and their controls, that sustain permanent perched wetlands in a sub-humid climate?
2) What are the water fluxes that must be considered within a conceptual model to assess hydrologic budgets for perched boreal wetlands?
3) Can the conceptual model developed from interpretation of field data be implemented in a numerical model, and honour the observed wetland-upland water flux dynamics?

This thesis follows the paper based format and has been organized into four chapters including this introduction which provides a brief overview of the Western Boreal forest and perched wetland hydrology, and how perched wetlands fit into the ongoing research.

The field research element of the thesis, including independent measurements and interpretation of the critical water flux components to determine water budgets for both a perched pond and a perched peatland are presented in Chapter 2. In addition to the determination of the water budgets, this chapter presents a "portable" conceptual model that can be broadly applied to other perched boreal wetlands, thought to be prevalent in the region.

Chapter 3 used a focused numerical modeling approach to test/validate the conceptual model, hydrologic response area characteristics, and interpretations of wetland-upland flux dynamics developed through the field research (Chapter 2). This approach allowed further refinement of our understanding of the hydrological processes sustaining the perched boreal wetlands and the drivers of observed natural variability.

Finally, Chapter 4 provides a summary of the two main chapters and elaborates on the applicability of this research to evolving water resource management practices, and landscape reclamation approaches currently ongoing in the region.

1.4 References

Alberta Environmental Protection, 1998. The boreal forest natural region of Alberta. Natural Resource Services, Recreation and Protection, Special Report.

Blancher, P. and Wells, J., 2005. North America's bird nursery: The Boreal Forest Region and its global responsibility toward sustaining bird populations. Boreal Songbird Initiative and the Canadian Boreal Initiative. Ottawa, Ontario.

Brooks, R.T. and Hayashi, M., 2002. Depth-area-volume and hydroperiod relationships of ephemeral (vernal) forest pools in southern New England. Wetlands, 22(2): 247-255.

Bryant, D., Nelson, D., Tangley, L., 1997. The last frontier forests: Ecosystems and economics on the edge. Washington D.C., World Resources Institute.

Buttle, J.M., 1989. Soil moisture and groundwater responses to snowmelt on a drumlin sideslope. Journal of Hydrology, 105(3-4): 335-355.

Canadian Forest Service, 2006. The state of Canada's forests 2004-2005: The boreal forest. Canadian Forest Service, Ottawa, Ontario.

Devito, K., Creed, I., Gan, T., Mendoza, C., Petrone, R., Silins, U. and Smerdon, B., 2005a. A framework for broad scale classification of hydrologic response units on the Boreal Plains: Is topography the last thing to think of? Hydrological Processes, 19(8): 1705-1714.

Devito, K.J., Creed, I.F. and Fraser, C.J.D., 2005b. Controls on runoff from a partially harvested aspen-forested headwater catchment, Boreal Plain, Canada. Hydrological Processes, 19(1): 3-25.

Ducks Unlimited Canada. 2006. Waterfowl of the Boreal Forest. Ducks Unlimited Canada, Stonewall, Manitoba. 107p.

Euliss, N.H., LaBaugh, J.W., Fredrickson, L.H., Mushet, D.M., Laubhan, M.K., Swanson, G.A., Winter, T.C., Rosenberry, D.O. and Nelson, R.D., 2004. The wetland continuum: A conceptual framework for interpreting biological studies. Wetlands, 24(2): 448-458.

Ferone, J.M. and Devito, K.J., 2004. Shallow groundwater-surface water interactions in pond-peatland complexes along a Boreal Plains topographic gradient. Journal of Hydrology, 292(1-4): 75-95.

Fetter, C.W., 1994. Applied Hydrogeology. Prentice Hall, New Jersey, 691 p.

Gibbons, Z., 2005. The influence of groundwater flow on phosphate dynamics in three riparian wetlands surrounding an outwash lake in Northern Alberta. M.Sc. Thesis, University of Alberta, Edmonton, Alberta, 95p.

Hayashi, M., van der Kamp, G. and Rudolph, D., 1998. Water and solute transfer between a prairie wetland and adjacent uplands, 1. Water balance. Journal of Hydrology, 207(1-2): 42-55.

Klassen, R.W., 1989. Quaternary geology of the southern Canadian interior plains. In: R.J. Fulton (Editor), Quaternary Geology of Canada and Greenland. Geology of North America. Geological Survey of Canada, Ottawa, Canada: 138-174.

Metcalfe, R.A. and Buttle, J.M., 1999. Semi-distributed water balance dynamics in a small boreal forest basin. Journal of Hydrology, 226(1-2): 66-87.

Meyboom, P., 1966. Unsteady groundwater flow near a willow ring in hummocky moraine. Journal of Hydrology, 4(1): 38-40.

Millar, J.B., 1971. Shoreline-area ratio as a factor in rate of water loss from small sloughs. Journal of Hydrology, 14(3-4): 259-284.

Mills, J.G. and Zwarich, M.A., 1986. Transient groundwater flow surrounding a recharge slough in a till plain. Canadian Journal of Soil Science, 66: 121-134.

National Wetlands Working Group, 1988. Wetlands of Canada, Ecological Land Classification Series No. 24, Ottawa.

O'Geen, A.T., McDaniel, P.A., Boll, J. and Brooks, E., 2003. Hydrologic processes in valley soilscapes of the Palouse Basin in northern Idaho. Soil Science, 168(12): 846-855.

Pawlowicz, J.G., Fenton, M.M., 2002. Drift thickness of the Peerless Lake map area (NTS 84B), Map 253, Alberta Geologic Survey, Edmonton, Alberta, Canada.

Petrone, R.M., Silins, U., Devito, K.J., 2007. Dynamics of evapotranspiration from a riparian pond complex in the Western Boreal Forest, Alberta, Canada. Hydrological Processes, 21(11): 1391-1401.

Rains, M.C., Fogg, G.E., Harter, T., Dahlgren, R.A. and Williamson, R.J., 2006. The role of perched aquifers in hydrological connectivity and biogeochemical processes in vernal pool landscapes, Central Valley, California. Hydrological Processes, 20(5): 1157-1175.

Rockafeller, P.A., McDaniel, P.A. and Falen, A.L., 2004. Perched water table responses to forest clearing in Northern Idaho. Soil Science Society of America Journal, 68: 168-174.

Smerdon, B.D., Devito, K.J. and Mendoza, C.A., 2005. Interaction of groundwater and shallow lakes on outwash sediments in the sub-humid Boreal Plains of Canada. Journal of Hydrology, 314(1-4): 246-262.

Smerdon, B.D., Mendoza, C.A. and Devito, K.J., 2007. Simulations of fully coupled lake-groundwater exchange in a subhumid climate with an integrated hydrologic model. Water Resources Research, 43(1): W01416, doi:10.1029/2006WR005137.

Winter, T.C., 2001. The concept of hydrologic landscapes. Journal of the American Water Resources Association, 37(2): 335-349.

Winter, T.C. and Woo, M.-K., 1990. Hydrology of lakes and wetlands. In: M.G. Wolman and H.C. Riggs (Editors), The Geology of North America Vol O-1: Surface Water Hydrology. Geological Society of America, Boulder, Colorado.

Wolfe, B.B., Hall, R.I., Last, W.M., Edwards, T.W.D., English, M.C., Karst-Riddoch, T.L., Paterson, A. and Palmini, R., 2006. Reconstruction of multi-century flood histories from oxbow lake sediments, Peace-Athabasca Delta, Canada. Hydrological Processes, 20(19): 4131-4153.

Woo, M.-K. and Winter, T.C., 1993. The role of permafrost and seasonal frost in the hydrology of northern wetlands in North America. Journal of Hydrology, 141(1-4): 5-31.

Chapter 2

An evaluation of the hydrologic processes sustaining permanent perched boreal wetlands, north-central Alberta

2.1 Introduction

The Western Boreal Forest ecoregion of Northern Alberta, Canada, contains a variety of landforms, vegetation, and wildlife that support an important natural resource based industry (Alberta Environmental Protection, 1998). The area also provides one of the most significant waterfowl habitats in North America (Ducks Unlimited Canada, 2006), and the largest carbon storage pool in the world (Gorham, 1991). This area is currently subject to rapid development of natural resource based industries including conventional oil and gas exploration and production, oil sands mining, forestry, recreation, and agriculture (Alberta Environmental Protection, 1998). These ongoing anthropogenic developments have the potential to impact wildlife habitat, and influence global climate change through alteration of energy, water, and greenhouse gas (CO_2 and CH_4) exchanges (Metcalfe and Buttle, 1999).

The complex hydrology of boreal wetlands situated in heterogeneous land cover and glacial deposits characteristic of the Western Boreal Plain (WBP) makes the impact from anthropogenic activities difficult to predict (Smerdon *et al.*, 2005). Up to 50% of the landscape is covered by ponds, shallow lakes, or peatlands (Kuhry *et al.*, 1993), situated on three common landforms, including coarse-grained glaciofluvial (outwash) deposits, fine-grained glaciolacustrine (clay plain), and disintegration moraine deposits (Fenton *et al.*, 2003; Devito *et al.*, 2005b; Petrone *et al.*, 2007). Improving our understanding of the hydrological processes sustaining boreal wetlands is essential to assess potential impacts to water resources and boreal ecosystems. The utilization of a Hydrologic Response Unit (HRU) approach as described by Arnold *et al.* (2001) and Devito *et al.* (2005a), allows a framework to assess hydrologic impacts to different ecohydrologic landscape features in the WBP essential to sustainable development of the natural resources in the area.

Recent studies completed in the WBP have shown that the sub-humid climate (average annual precipitation is less than potential evapotranspiration), geology, water storage, and surface water-groundwater interaction control the sustainability of shallow lakes, ponds, and peatlands (Ferone and Devito, 2004; Smerdon *et al.*, 2005; Petrone *et al.*, 2007; Smerdon *et al.*, 2007). However, these components have varying degrees of control on the hydrological processes depending on the environmental position within a soil texture gradient (coarse to fine) and degree of hydrologic connection to larger scale flow systems.

Little research has been completed on perched wetlands and flow systems in the WBP. Perched wetlands in this region appear to be essentially uncoupled from regional groundwater flow systems representing an end member wetland type within textural and landscape position gradients. Despite their isolation from large-scale flow systems, they are thought to be widely distributed and important to water resources in the sub-humid, heterogeneous landscape of the Western Boreal Plains (Wolfe *et al.*, 2006).

The function of perched wetlands in the WBP has not been adequately assessed, but may be more widely distributed than suggested in the literature (Wolfe *et al.*, 2006). Detailed study of the localized hydrologic processes that sustain permanent and ephemeral perched wetlands is required to quantify hydrologic characteristics such as recharge, water storage potential, and biogeochemical function. Subsequent integration of knowledge gained on these landscape features into a HRU framework will facilitate more accurate landscape scale modeling, and assist in sustainable resource development. The results of this study will apply to ongoing land reclamation associated with the widespread natural resource based industry operating in the region such as de-watering of coarse-grained tailings sand from oil sands extraction (Price, 2005). In addition, knowledge obtained from this study will guide decisions regarding future development of roads, geophysical cut lines, and logging features to minimize ecological impacts.

The objective of this study was to determine the water budget and hydrological processes sustaining permanent, perched wetlands in the WBP. In order to accomplish this objective, existing conceptual models required adaptation to account for the unique wetland-upland water fluxes and interaction of perched groundwater and evapotranspiration to firstly, complete water budgets, and secondly, assess the role of perched wetlands within the boreal landscape.

Perched wetlands including an open pond-marsh, and a fen peatland, were selected on a landscape with poorly defined drainage networks and closed basins. In some ways, these perched wetlands are similar to wetlands commonly referred to as prairie potholes, which have been well studied since the 1960's (e.g., Meyboom, 1966; Millar, 1971; Mills and Zwarich, 1986; Winter and Rosenberry, 1995; Rosenberry and Winter, 1997; van der Kamp and Hayashi, 1997; Hayashi *et al.*, 1998). However, these perched wetlands can accumulate peat and are situated in a highly heterogeneous landscape with different upland vegetation characteristics and climate. These differences require that a unique conceptual model be tailored to the ecohydrologic characteristics observed to be present at the study site, including climate, hydrogeologic setting, and vegetation patterns to properly identify the sources and sinks of water at these wetlands. Partitioning of atmospheric fluxes into actual evapotranspiration (*AET*), throughfall or net precipitation (P_{net}), and interception loss, is a crucial component of the conceptual models previously developed for boreal environments (Elliott *et al.*, 1997; Smerdon *et al.*, 2007). The highly heterogeneous geologic environment of the study wetlands creates variable soil storage and vegetal patterns that govern the redistribution of atmospheric fluxes via surface water and perched groundwater flow, and thus, the sustainability of perched wetlands. The conceptual model necessitated that independent water budgets for catchment sub-components be considered to account for spatial variation of atmospheric fluxes and accurately assess overall catchment water budgets.

2.2 Study area

The perched wetlands studied are located 370 km north of Edmonton AB, Canada (56°6′ N, 116°32′ W) at the Utikuma Region Study Area (URSA). The URSA is located in the Mixed-wood Boreal Plains ecoregion and lies approximately 150 km south of the discontinuous permafrost zone (National Wetlands Working Group, 1988; Woo and Winter, 1993). The climate is characterized by seasonally variable average monthly temperatures ranging from -14.6 °C in January to 15.6 °C in July (Environment Canada, 2003) and is considered sub-humid with normal annual precipitation (481 mm; Environment Canada, 2003) which is exceeded by annual potential evapotranspiration (517 mm; Bothe and Abraham, 1993). This net deficit in atmospheric fluxes is typical of most years, with atmospheric surpluses occurring approximately every 10 to 15 years, resulting from a combination of cool summer temperatures and greater than normal annual precipitation (Devito *et al.*, 2005b). Most precipitation (i.e., >60%) generally falls in the summer months (June, July, and August), and snow typically accounts for <25% (<100 mm) of the average total annual precipitation (Devito *et al.*, 2005a). The surficial geology of the URSA is composed of glacial derived sediments including glaciofluvial, moraine, and glaciolacustrine deposits that vary in thickness from 20 to 240 m (Pawlowicz and Fenton, 2002). The land surface is hummocky with low relief and poorly defined surface drainage patterns.

2.2.1 Perched wetland study site

The study site is 2.3 km^2 with numerous perched wetland features, including several peatlands, sedge peatlands, one pond (P19), and a portion of a large shallow lake (L17; Fig. 2.1). Geologically, the study site is situated in the transition between extensive (200 km^2) coarse textured outwash deposits to the northwest and fine-textured sediments deposited by a stagnant ice moraine to the east/southeast (Fenton *et al.*, 2003; Smerdon *et al.*, 2005). This transition zone is situated on a topographic high within the URSA (Devito *et al.*, 2005a), with fine-textured moraine deposits (silts and clays) sourced from the moraine to the east, overlying the coarse-grained sand and gravel deposits, the surficial landform to the west (Fig. 2.2).

The study site has two gravel roads and numerous linear geophysical cut-lines, typical of much of Northern Alberta (Alberta Environmental Protection, 1998; Fig. 2.1). Maximum relief at the study site is approximately 25 m, from the lowest depression to the highest upland areas. Deep borehole data indicate a laterally extensive till surface that ranges in depth from approximately 15 to >40 m from land surface. Deep piezometer data indicate a deep laterally continuous water table 15 to 30 m from surface. The topographic position of the study site suggests that the deep water table at the study site represents a regional groundwater flow divide, coincident with a topographic high in the till surface and aquifer thickening to the southwest. These data also indicate an extensive (10 – 20 m thick) unsaturated zone between the perched and regional water tables (Fig. 2.2). The upland vegetation at the study site is dominated by trembling aspen (*Populus tremuloides*) with some black poplar (*Populus nigra*) and high bush cranberry (*Viburnum trilobum)* as the dominant understory species. Most wetlands at the study site are thicket swamps or peatlands with variable vegetation cover depending on peat depth. The drier riparian areas, near the upland base support grasses, willow (*Salix* sp.), and raspberry (*Rubus* sp.).

Saturated riparian areas near pond edges have *Typha* sp. and *Carex* sp. vegetation. Peatlands accumulate 0.5 – 3.0 m of organic sediments underlying sphagnum moss, sedges, Labrador tea (*Rhododendron groenlandicum*), stunted black spruce (*Picea mariana*), and alder (*Alnus rugosa*) thickets. There is also a swampy area of mixed-wood forest with both white spruce and aspen located in the centre of the study area. Small, ephemeral wetlands present in depressions and valley bottoms (draws) have large white spruce (*Picea glauca*), a variety of grasses, and swamp horsetail (*Equisetum fluviatile*). Paper birch (*Betula papyrifera*), willow (*Salix* sp.), and alder (*Alnus* sp.) are also common at wetland/upland transitions.

2.2.2 Pond 19 (P19)

P19 is a small pond that has a mineral soil riparian zone with no peat development and a confined basin area of 5.6 ha. A portion of the riparian zone is connected to a small (0.7 ha) ephemeral draw, which periodically conveys perched groundwater and surface water into P19. The ephemeral draw lies in a valley bottom between aspen uplands and flows into the pond only during times of water surplus such as spring melt and high-intensity summer storms. The basin and draw areas are surrounded by aspen upland creating a closed basin with no surface outflow.

Analysis of historic air photographs from the last 50 years indicates that the open water pond area has varied in area from 1.4 ha to 5.6 ha. The average pond elevation is 650 m a.s.l., and the average pond area over the period of this study was approximately 1.7 ha, with the remainder of the basin area occupied by riparian zone vegetation (grasses, nettles, willow, alder, minor paper birch). The catchment slope varies from 2.5 degrees in the wetland basin, to 8-10 degrees in upland areas. The wetland basin is up to 3.2 m deep and both the pond and riparian zone have thin (0.1-0.2 m) layers of loose gyttja and organic soil respectively, which is underlain by mineral soils. Seale Lake Road is located on the shoulder of the upland surrounding P19 approximately 100 m south of the edge of the P19 basin (Fig. 2.1).

2.2.3 Wetland 1 (W1)

W1 (1.75 ha) has a peat accumulation to a maximum depth of 2.5 m in the centre of the oblong basin, and is approximately 656 m a.s.l. (6 m higher than P19). W1 has vegetation representative of the majority of the other peatlands found at the study site including sphagnum moss, Labrador tea, stunted black spruce, and alder. Aspen uplands surround W1. The lack of either inflows or outflows creates a closed basin that has no apparent slope across the peat surface, but has significant micro-topography, with 0.3 m high lawns and depressions. Standing water is generally present in the depressions for most of the year, except for extended periods of hot, dry weather. The surrounding upland has a much shallower slope (2-3 degrees) than that of P19. The transition from peatland vegetation to the upland is more abrupt than P19, with a narrow (5 m) riparian zone. Despite the narrow width of the riparian zone, it has similar transitional vegetation to P19, including birch, willow, and alder trees between the peatland and the upland vegetation. W1 is a similar distance from Seale Lake Road as P19 (80 m; Fig. 2.1).

2.3 Methodology

2.3.1 Lithology and groundwater

The lithology, and both local (perched) and regional groundwater flow regimes (Tóth, 1963) of the study site were determined from drilling boreholes at 67 locations (Fig. 2.1) and installing 130 piezometers. Distinct well and piezometer locations were given a location number, followed by a dash, and a well designation (W) for long screen intervals that straddled the water table or the total depth (in cm) below ground surface for piezometers (e.g., 11-W and 11-590 for the well and piezometer respectively, at location 11). Five locations had deep boreholes (15 – 38 m) intersecting the regional water table. The remaining 62 locations were shallow to medium depths (i.e., <1 – 15 m) allowing characterization of near surface sediments and hydrology. Three methods were used to advance the boreholes, depending on the target depth. A track mounted drill rig with 0.15 m solid stem augers was used for deep boreholes. Either a trailer-mounted auger rig, or a bucket-style hand auger was used for 0.06 m diameter, intermediate to shallow depth boreholes. An attempt was made to continuously log soil characteristics when advancing boreholes, with samples retained for grain size analysis. Samples were also retained from shallow depth soil pits dug to install soil moisture equipment. Particle size and distributions of 50 samples were analyzed with a Micrometrics Corporation Sedigraph 5100. This method is appropriate for fine-grained materials such as clay, silt, and fine sand as it measures a grain size range from 1 – 300 μm. Samples containing coarse sand or organic debris that would clog the Sedigraph were analyzed using sieve analyses.

The regional groundwater flow regime was characterized with six deep piezometers at five separate locations, constructed of 0.0375 m inside diameter PVC pipe with 1.5 m screens installed below the regional water table. The annulus of all regional piezometers was filled with silica sand to 0.3 m above the screened interval, and the remaining annulus to surface was filled with bentonite chips. Similarly constructed shallow and intermediate depth wells/piezometers were installed coincident with these deep piezometers to quantify the extent of the unsaturated zone separating the regional and perched water tables. Shallow to intermediate depth wells and piezometers were constructed of 0.025 m inside diameter PVC pipe. Wells had variable length screened intervals to straddle the water table position and piezometer screens were 0.3 m or 0.6 m in length, installed below the perched water table surface. Both well and piezometer screens were wrapped in filter cloth, and the annulus was filled with silica sand to 0.1 m above the screened interval. The remaining annulus was filled with hydrated bentonite chips to surface.

Manual water level measurements were made in all piezometers and wells throughout the duration of the field program. The frequency of measurement varied from bi-weekly in 2004, to approximately 50 times during the ice-free period from April to November in 2005 and 2006. In addition, unvented, automated pressure transducers (Hobo Model U20-001-01) were installed in 8 wells to collect continuous hydraulic head data over the same period. An additional pressure transducer of the same type was installed in open air for barometric pressure compensation of unvented transducers. Frozen wells and shallow soil frost were noted in the wetland and riparian areas of both wetlands from mid-winter through approximately June 1, of both 2005 and 2006. Seasonal soil frost has been shown to have a significant effect on runoff regimes of other boreal and sub-arctic catchments (Woo and Winter, 1993; Carey and Woo, 2001). As such, depth to ice, and ice thickness measurements were made during the spring, by hammering a 0.5 cm steel rod into the soils along the main transects of both wetlands, and ephemeral draw (Fig. 2.1).

Saturated hydraulic conductivity values were determined for 20 piezometers, installed in areas saturated by perched groundwater using the Hvorslev (1951) method. Two tests were completed for humic/mesic peat, with the remainder being completed in silt/clay mineral soils. Lateral hydraulic head gradients, and saturated hydraulic conductivities were then used to estimate the saturated flow of perched groundwater at the study site using Darcy's Law (Freeze and Cherry, 1979);

$$(Eq.1) \quad q = -K\frac{dh}{dL}$$

where q is the Darcy flux (m/s), K is the saturated hydraulic conductivity (m/s), and dh/dL is the hydraulic head gradient (dimensionless).

2.3.2 Soil moisture

Radial soil moisture instrumentation transects, extending from the wetland margin into the upland areas similar to Hayashi *et al.* (1998), were established at P19 and W1. The instruments were configured to follow the saturated/unsaturated interface in the transition area from wetland to upland vegetation. These transects were used to determine the unsaturated flux of perched groundwater into the upland and changes in unsaturated zone water storage at the wetland peripheries and within the upland rooting zone. Each transect had 9 time-domain reflectometer (TDR) probes (Campbell Scientific CS-616), and 9 tensiometers (Soil Measurement Systems model SW-03) fitted with electronic pressure transducers (Honeywell 26PC). The bi-hourly volumetric water content data and soil tension data were logged using Campbell Scientific CR-10X data loggers.

The TDR probes were installed in four soil pits located at each wetland periphery extending into the surrounding aspen upland. TDR probes were pushed directly into the undisturbed pit wall at depths ranging from 0.2 – 1.5 m below the organic forest floor in the rooting zone (Fig. 2.3). The pit nearest to each wetland had a single TDR probe, while the other pits had nests of 2 – 4 probes allowing spatial and temporal changes in soil moisture storage to be determined (Fig. 2.3).

Nested tensiometers were installed at three locations coincident with the nested TDR probes (Fig. 2.3). Tensiometers were installed using a 0.025 m hand-auger to drill to the target depth, where the 0.025 m diameter porous cup of each tensiometer was pushed directly into the soil to ensure effective coupling with the substrate. The pressure transducers fitted to the tensiometers were calibrated before installation each year using a vacuum pump and pressure gauge assembly to develop suction pressure versus mV output functions for each transducer.

Unsaturated hydraulic head or soil tension was measured bi-hourly from August 2005 through the remainder of the study period, which enabled the determination of lateral and vertical gradients in soil tension at the wetland peripheries. Darcy's Law was employed to determine the specific unsaturated discharge at the wetland peripheries. Unsaturated hydraulic conductivity values were estimated for the soils in the rooting zone, which was based on the observed soil texture and saturation ratio. Soil pit observations were used to determine the depth of the rooting zone. The contributing area for which unsaturated flux into the upland areas occurred was determined by multiplying the rooting zone depth by the wetland basin circumference in order to assess a volumetric flux.

2.3.3 Precipitation and interception loss

Total precipitation (*P*) was measured at both P19 and W1 from May 2004 to October 2006 using a bulk rain gauge. A tipping bucket (Jarek model 4025) was also used to measure precipitation at P19 from June 2005 through the duration of the study. Data from a nearby (4 km) tipping bucket within URSA were used for 2004 to estimate the timing of spring precipitation in early 2005, prior to the installation of the on-site tipping bucket. Late winter snow surveys were used to estimate the maximum snow water equivalent (SWE) determined gravimetrically (snow depth and density measurements) just prior to spring melt. Real-time climate data from Environment Canada Red Earth Creek weather station (www.weatheroffice.ec.gc.ca/city/pages/ab-40_metric_e.html) was used to determine the timing of late February 2005, and early March 2006 snow surveys. SWE was determined along three transects at P19, extending from the upland, or ephemeral draw area, across the pond/riparian zone and into the opposing upland area. Two perpendicular transects were established from upland to upland, across the W1 peatland. These transects were used to assess mean and median SWE for each catchment sub-component (i.e., upland, riparian, pond, and peatland areas) within each wetland catchment to account for potential redistribution and/or interception of snow. Melting rates were estimated based on air and ground temperature profile data from 2 m thermistor strings with 6 sensors each (RST Instruments model TH0006-8-HP) installed at each study wetland.

Spatial and temporal distributions of interception (*INT*) losses vary significantly depending on site-specific variables including vegetation canopy (leaf area index), surface soil (litter) properties, and climate (rainfall duration, rainfall intensity, and potential evapotranspiration; Crockford and Richardson, 2000; Dingman, 2002). Boreal interception and throughfall literature document highly variable interception loss for individual sub-components of a forest (canopy, understory, and litter layer). Interception loss can also be determined through the use of intensively parameterized models (Price *et al.*, 1997; Hashino *et al.*, 2002; Carlyle-Moses *et al.*, 2004). However, this study required a summation of the interception loss, or the proportion of event precipitation that did not infiltrate the mineral soil underlying the forest floor. A gross estimate of the interception loss for upland and wetland (riparian and peatland) areas was developed by measuring the difference between total event precipitation and the corresponding change in soil moisture content measured by the time domain reflectometer (TDR) probe installed below the forest floor or litter layer in upland and riparian areas. These data were used to develop a daily interception threshold (in mm) for upland and riparian areas, and determine infiltration.

2.3.4 Evaporation

Potential evapotranspiration from P19 was measured from May 2005 to October 2006 using a Class A evaporation pan placed at the south edge of the pond with approximately half of the pan height submerged in standing water similar to Smerdon *et al.* (2005). An automated pressure transducer (Hobo model U20-001-01) was also placed at the bottom of the evaporation pan to record continuous water level data in the evaporation pan between manual measurements and temperature data to ensure that thermal equilibrium with the pond water was maintained. Potential evapotranspiration at W1 was measured with a 0.3 m by 0.45 m lysimeter that was filled with 0.2 m of the active peat layer and a sealed bottom preventing percolation. In the summer months, changes in storage were measured by weighing the lysimeter weekly to determine potential evapotranspiration.

In addition, shielded air temperature/relative humidity sensors (Campbell Scientific model CS-500L) measured hourly air temperature over both wetlands such that potential evapotranspiration (*PET*) could be determined using Thornthwaite's formula and average monthly temperatures measured on-site. Thornthwaites's formula is an empirical monthly water-balance approach to determine potential evapotranspiration (De Marsily, 1986);

$$(Eq.2)\quad PET = 1.6 \times \left[\frac{10T_a}{I}\right]^a \qquad I = \sum_{12}^{1}\left(\frac{T_a}{5}\right)^{1.5}$$

Where PET in mm, T_a is the average monthly temperature in °C, $\boldsymbol{I}$ is the annual heat index, and a is equal to $(0.49 + 0.0179\,\boldsymbol{I} + 7.71{*}10^{-5}\,\boldsymbol{I}^2 + 6.75{*}10^{-7}\,\boldsymbol{I}^3)$.

The dependence of the perched hydrological processes on climate required careful consideration of the spatial distribution of actual evapotranspiration, and the partitioning of evapotranspiration flux into evaporation of intercepted precipitation and evapotranspiration sourcing water from the mineral soils in the rooting zone within each catchment sub-component. The scope of the project required that literature values be used for *AET* in the riparian zones, upland areas, and the ephemeral draw. These values were scaled based on measured *PET* values measured on-site. The riparian zone *AET* was assigned a monthly variable daily *AET* rate for May through September measured at mid-latitude and Boreal fens (Lafleur *et al.*, 1997; Rouse, 2000). Aspen upland *AET* rates were based on measured aspen *AET* of approximately 440 mm/yr from the BOREAS SSA site near Prince Alberta, Saskatchewan (Amiro *et al.*, 2006). This *AET* seemed reasonable when compared to measured *PET* over the course of the study. This *AET* estimate also fell within the range of aspen *AET* rates of 226 to 487 mm/yr reported in the Western Boreal Plains ecoregion (Devito *et al.*, 2005b). For each catchment sub-component, *AET* was then broken into rooting zone evapotranspiration (AET_{RZ}), which is equal to the total evapotranspiration (*AET*), minus interception loss (*INT*) to account for observed storage changes in mineral soils.

2.3.5 Surface water

P19 stage was measured manually throughout the study with a staff gauge mounted to a steel stake driven deep into the mineral soils below the pond. Continuous stage data were also recorded with a pressure transducer (Hobo model U20-001-01) in a stilling well fixed to the staff gauge during the ice-free months of 2005 and 2006. The depth and extent of ephemerally saturated areas in the study area were monitored using permanent stakes or wells in areas of standing water. Surface water flows were only observed at the base of the ephemeral draw draining into the P19 basin. Surface flows were measured by capturing surface flow with a 40 L pail where it was constrained to a narrow natural weir. Sporadic flow measurements were taken by recording the volume captured by the pail in a given time interval, and the stage in a stilling well immediately upstream of the weir were used to develop an approximate stage-discharge relationship. A pressure transducer recorded continuous water level data from the stilling well to quantify surface flows into P19 if flow was not measured directly. Electrical conductivity (EC) and pH measurements of surface water were made at P19, W1, upper/lower portions of the ephemeral draw, and any small ephemerally saturated areas to assist in determining the flowpath of observed surface waters. Ten spot measurements were made from May through August 2006 at all monitoring points that had surface water present at the time of measurement.

2.3.6 Water balance conceptual model

The water budget was calculated for the ephemeral draw and wetland basin areas of each catchment (Fig. 2.4), according to the following formula:

$$(Eq.3) \quad r = P_{net} - AET_{RZ} + SGW_{in} - SGW_{out} - UGW_{out} + SW_{in} - SW_{out} - \Delta V_{wetland}$$

where r is the residual, P_{net} is net precipitation (P - INT), AET_{RZ} is actual evapotranspiration from the wetland rooting zone (and open water in the P19 basin), SGW_{in} is saturated groundwater in, SGW_{out} is saturated groundwater flow out, UGW_{out} is unsaturated groundwater flow out, SW_{in} is surface water in, and ΔV is change in volume of stored water within the wetland areas. Water budgets for upland areas of both catchments (Fig. 2.4), were calculated using the following formula:

$$(Eq.4) \quad r = P_{net} - AET_{RZ} + UGW_{in} - SW_{out} - \Delta V_{upland}$$

where UGW_{in} represents a positive flux equivalent to the UGW_{out} term in the wetland basin water budget formula, AET_{RZ} is the actual evapotranspiration from the aspen rooting zone, and ΔV is the change in soil moisture in the uplands.

Water budgets were completed for P19 for both 2005 and 2006 hydrologic years. The P19 catchment was divided into three sub-components including the wetland basin (BAS), the upland area surrounding the pond (UPLD), and the ephemeral draw (DRW) imbedded within the upland area (Fig. 2.4). The 5.6 ha wetland basin area was defined by the lateral extent of the perched groundwater lens, which corresponded to the edge of the riparian zone vegetation, and the break in slope at the base of the upland. The upland areas of each catchment were topographically defined, honouring symmetry boundaries of unsaturated flow into upland areas from adjacent ephemeral or permanent perched wetlands, and the topographic flow divide for potential runoff between adjacent perched wetlands. The 7.2 ha upland area at P19 was further sub-divided into the upland area and the 0.7 ha ephemeral draw area to directly account for the fluxes of perched groundwater and surface water from the ephemeral draw into the P19 basin. W1 had an upland area of 2.1 ha, and a wetland basin area of 1.7 ha defined in the same way as the P19 basin and upland.

2.4 *Results*

2.4.1 *Sediment properties*

The sediments at the study site are highly heterogeneous and poorly sorted. They originated from ice blocks left at the margin of a disintegration moraine (Fenton *et al.*, 2003). The proximity of the study site to the outwash plain to the northwest indicates reworking of the sediments in a glaciofluvial environment which concentrated fine-grained materials in depressions and deeper in the soil column (Pawlowicz and Fenton, 2002). Sediment collected from near the surface (0-0.5 m) had high organic content/debris and contained minor coarse sand. Deeper sediments were classified as sandy silt, silt, and clay materials (Fig. 2.5). All samples retained from shallow to intermediate depths contained a high fraction of fine-grained material (i.e., silt or clay). In contrast, sieve analyses of deep borehole samples showed that underlying sand and gravel deposits are well-sorted fine to medium sand with gravel, similar to the nearby outwash materials described by Smerdon *et al.* (2005).

An unfractured, unoxidized, highly plastic, blue-grey clay lithology was identified in all boreholes completed in wetland areas (i.e., P19, W1 – W8). A clay lithology with similar grain size distribution was also encountered in all upland boreholes; however, it was light-grey in colour, and had oxidized fractures. The soil texture data from all boreholes were used to krige the surface elevation of this low-permeability clay lithology as a laterally extensive confining layer (Fig. 2.6a). The difference between the upper surface of the confining layer, and a LIDAR generated digital elevation model of the land surface was used to produce a depth to confining layer map (Fig. 2.6b).

The soils overlying the confining layer are primarily silt mineral soils with a gradational change from sandy silt near surface, to clayey silt above the confining layer. P19 has a thin (<0.2 m) veneer of loose, organic soil in the riparian zone, and 0.1 – 0.3 m of gyttja on the pond bottom overlying the mineral soils (Fig. 2.7a). W1 has a peat accumulation of 2.5 m in the basin centre with an acrotelm ranging from 0.2 to 0.5 m in depressions and lawns respectively. Underlying the peat there are thin (0.1 - 0.2 m), layers of fine sand and silt immediately overlying the confining layer (Fig. 2.7b). Other peatlands (W2 – W8) have peat accumulations ranging from 0.4 – 3.0 m. Upland soil profiles are similar throughout the study site, with 0.1 – 0.3 m of forest floor at surface, and a mixture of fine sand, silt, and clayey silt overlying the confining layer, which is encountered at depths from 2 – 12 m from the surface.

The hydraulic conductivity (*K*) of the porous media, as determined from slug testing, ranged from 10^{-6} m/s for sandy silt, 10^{-7}-10^{-8} m/s for silts, and 10^{-9}-10^{-10} m/s for clays (Fig. 2.5). Thin (<0.4 m) silty sand layers were also observed near surface, below the forest floor layer and in the ephemeral draw. However, *K* values were not determined for these layers through in-situ testing. Hydraulic conductivity values for these materials are estimated to be 1×10^{-5} m/s based on grain size data. The aspen forest floor in upland areas contains wood and leaf debris and has extremely high porosity of approximately 0.9, and a hydraulic conductivity of 1×10^{-4} m/s (Redding *et al.*, 2005). The hydraulic conductivity of the active peat layer was estimated to be 1×10^{-4} to 1×10^{-5} m/s after Ferone and Devito (2004). The catotelm (0.6 – 2.5 m) has a hydraulic conductivity of 10^{-8} to 10^{-9} m/s as determined from four separate slug tests.

2.4.2 *Atmospheric fluxes*

2.4.2.1 *Precipitation and infiltration*

Hydrologic years were defined from November 1 to October 31 for both 2005 and 2006 to capture hydrologic events characteristic of the climate, including spring melt, spring/summer precipitation, and evaporative fluxes (Hayashi *et al.*, 1998; Smerdon *et al.*, 2005). The total annual precipitation was 411 mm and 394 mm for 2005 and 2006, respectively (Fig. 2.8a). Summer precipitation (April to October) measured by the on-site tipping bucket agreed with the bulk precipitation gauges to within 8 mm in 2005 and 10 mm in 2006. Both years had total annual precipitation that was lower than the long-term average. Pre-melt snow surveys showed that SWE was similar throughout the study site on a variety of slope aspects in both upland and wetland areas. In 2005, there was 135 mm SWE accounting for 33% of total annual precipitation, with the remainder of total annual precipitation falling as rain between May and October, with 5 daily precipitation values exceeding 10 mm. In contrast, in 2006, there was 66 mm SWE accounting for only 17 % of the total annual precipitation, with 12 daily precipitation values exceeding 10 mm (Fig. 2.8a).

Interception losses in wetland areas are not well documented, necessitating a reasonable approximation based on soil moisture data from shallow riparian soils, and qualitative observations made while at the study site during and after rain events. The wetland areas within each catchment, including the riparian zone of the P19, the ephemeral draw, and the W1 basin had 1 mm removed from daily precipitation totals for June through September, as interception loss. It was assumed that P_{net} over the pond area of P19 was equal to total measured precipitation.

In upland areas, daily precipitation totals less than 15 mm did not induce changes in soil moisture measured by upland TDR stations (N3 and N4) installed in the mineral soil rooting zone below the forest floor (Fig. 2.3). These data revealed that the aspen canopy, the understory, and forest floor intercept 100% of daily precipitation totals below approximately 15 mm. Infiltration into mineral soils below the forest floor occurred only for precipitation events exceeding 15 mm, with infiltration equal to the event total minus the infiltration threshold, which corresponded to measured changes in soil moisture. This approximation of upland interception loss and threshold behaviour resulted in 43% and

57% of the total precipitation, less the SWE, entering the mineral soils of the uplands in 2005 and 2006 respectively. An annual summation of positive changes in soil moisture from upland TDR probes showed that a similar proportion (approx. 50 %) of total precipitation infiltrated into mineral soils below the forest floor. This interception loss threshold is realistic as aspen canopy and understory integrated throughfall values of 5 mm have been determined by Devito (*unpublished data*), and the storage properties of 1.5 mm/cm of aspen forest floor materials as described by Redding *et al.* (2005). Using the developed interception loss relationship, the magnitude of calculated changes in upland storage related to summer precipitation correlate well to that of the measured storage changes (TDR data; Fig. 2.9).

2.4.2.2 Evapotranspiration

Cumulative potential evapotranspiration (*PET*) estimated from the evaporation pan data using a pan coefficient of 0.9, similar to Smerdon *et al.* (2005), was 454 and 500 mm in 2005 and 2006, respectively. Using Thornthwaite's method to determine *PET*, values of 440 mm in 2005 and 498 mm in 2006 were obtained. These values are within 10% of the measured pan evaporation values for both 2005 and 2006.

PET was estimated for W1 in 2006 by both a lysimeter during the growing season, and well as Thornthwaite's method for the entire year, which indicated a total evapotranspiration loss estimate of 328 mm for the W1 basin during 2006. Thermistor string data indicated that the top 0.2 – 0.5 m of W1 was frozen, with minimal standing water, until the end of May, thus limiting significant *AET* prior to June. Thornthwaite's method estimated 375 mm of *PET* for 2006, omitting pre-May data. The lysimeter estimated 250 mm, or 6 % less evapotranspiration than Thornthwaite's method *PET* estimation of 264 mm over the summer months of 2006. The lysimeter was not frequently monitored in September or October of 2006, so the *PET* estimate for these months was scaled based on the summer lysimeter observations. Aspen upland *AET* was estimated to be equal to approximately 440 mm from May through September for both 2005 and 2006 after Amiro *et al.* (2006). The riparian area of P19 and the ephemeral draw area were assigned monthly *AET* rates measured at a mid-latitude fen (Rouse, 2000) for May through September, totaling 370 mm annually.

Data from 2004 indicated a total annual precipitation of 364 mm, lower than annual totals from 2005 and 2006. However, an unusually high percentage of annual precipitation fell in late summer/early fall. From August 19 to October 31, 133 mm or 36.5% of the annual precipitation fell. Only 48.5 mm or 11.8% fell during the same time period in 2005, and similarly in 2006, late year precipitation accounted for only 37.5 mm or 9.5% of the annual precipitation. Continuous evaporation data was not collected on site during 2004. However, evaporation pan measurements for July and August 2004 indicated similar summer evaporative fluxes to those observed in 2005 and 2006.

2.4.3 Saturated zone

2.4.3.1 Water table configuration

Eleven laterally discontinuous, isolated perched water tables were identified in the P19 area where the low-permeability confining layer is close (<2 m depth) to the ground surface (Fig. 2.2; Fig. 2.5b). Synoptic transects across the margins of seven wetlands, including P19, peatlands (W1-W5, and W8), and two ephemerally saturated wetlands (Fig. 2.1) all show that saturated conditions terminate at the transition from wetland peripheries to upland areas (Fig. 2.2). Unsaturated upland areas, where the confining layer is at a greater depth from surface, are present between perched water tables. Intermediate depth and regional piezometers indicate that all the wetlands at the study site have similar perched/inverted water table configurations (Fig. 2.2; Fig. 2.7). The regional water table only varied <0.15 m over the duration of the study, indicating little to no hydraulic connection to the overlying perched water tables.

Detailed transects at P19 and W1 had enough nested wells and piezometers to determine the location of the inverted water table (Fig. 2.7). Deep piezometers (6 – 8 m) were installed at each detailed transect to define the location of the inverted water table surface. At P19, piezometer 11-590 remained dry throughout the study period, indicating unsaturated conditions beneath the inverted water table surface (Fig. 2.6a). This suggests that the inverted water table surface occurs near the bottom of the confining layer. Similarly at W1, piezometer 72-860 remained dry throughout the study indicating inverted water table surface exists between the screened intervals of piezometers 54-330 and 72-860 also suggesting the inverted water table follows the bottom of the confining layer at W1 as well as the other perched wetlands in the study area (Fig. 2.6b).

Four less detailed transects, extending from the pond to the upland, were monitored at P19 (Fig. 2.1). These transects showed water table configurations similar to those observed at the detailed transect, except for the one transect extending southeast into the ephemeral draw. Thus, the detailed transect was representative of water table configuration and flow regime for the pond periphery, with the exception of the 30 m wide area connecting the pond basin to the ephemeral draw. The water table configuration at the ephemeral draw outflow at the edge of the P19 riparian zone was highly transient. During intermittent periods of water surplus, a saturated tongue of perched groundwater developed, extending into the ephemeral draw. During frequent periods of water deficit, the saturated conditions in the draw dissipated as it drained into the P19 basin via saturated groundwater/surface water flows, and into the surrounding upland by unsaturated groundwater flow (Fig. 2.4c; Fig. 2.8b). As such, the area where the draw connects to the P19 basin had lateral fluxes that varied in both magnitude and direction, depending on atmospheric flux patterns. For water balance calculations, this transect was considered independently to account for the variable fluxes entering the P19 basin.

At W1, two additional transects were monitored to ensure that the detailed transect was representative of the water table configuration at its periphery (Fig.2.1). The perched water table configuration measured at the W1 transect was similar to that observed at P19 and is likely representative of the entire periphery of W1. The perched water table of W1 showed minimal hydrologic response to seasonal, and daily climate patterns. In 2005, saturated conditions were present in the riparian/upland transition area or well (55-Wb), for four months (June through September) due to large SWE (Fig. 2.8b). In 2006, only two months of saturated conditions were observed at 55-Wb, as it remained dry until mid June due to a small SWE, and water stored as frozen peat present until the end of June (Fig. 2.8b).

2.4.3.2 Saturated groundwater flow regime

The transient saturated hydraulic head responses of P19, riparian zone wells, the ephemeral draw, W1 peatland, and periphery (peatland/upland transition), and the regional water table are shown in Figure 2.8b. Saturated flow of perched groundwater only occurred in three areas of the study site: the P19 ephemeral draw, and in the riparian zones of both P19 and W1. In the riparian zones, groundwater flowed either towards or away from each wetland centre, depending on the atmospheric flux patterns, and pond or peatland water levels. Radial flow away from both wetlands was observed over most of the study period, which was rate limited by the unsaturated flow component given the unsaturated nature of the uplands. Groundwater ridging, ranging from 0.1 to 0.3 m above the stage of P19 and W1, was observed several times near the base of the upland after rain events >10 mm/d, causing saturated flow towards the wetlands. Such flow reversals towards the wetlands, denoted by (FR) on Figure 2.8b, lasted for 1 – 5 days until equilibrium was reached between the pond/peatland water levels, and the receding groundwater ridge. Subsequently, the flow pattern reverted to radial flow from the pond to the upland. This intra-basin saturated groundwater flow was not explicitly considered in water budget calculations as it was accounted for in the annual storage change of both P19 and W1.

Continuous water level data from mid-riparian well 16-SW at P19 (Fig. 2.7a) showed quick-flow occurring over shallow concrete ground frost in response to high melting rates or early spring rain events. Groundwater ridging and quick-flow in unfrozen, high porosity soils appear to be important local processes to maintain pond water levels (Fig. 2.10). Flow reversals occurred less frequently at W1 relative to P19 due to the narrow (5 m) riparian zone, resulting in only minor saturated flow contributions to the wetland basin.

A method similar to Hunt *et al.* (1996), using Darcy's Law, was used to provide a gross estimate of the volume of saturated vertical flow of perched groundwater lost to the underlying unsaturated zone. At P19, large vertical gradients of 1.12 were observed in the confining layer at 50-W (Fig. 2.7a). Similarly, at W1, gradients ranging from 0.75 to 1.14 were observed at 52-W, 53-W, and 54-W (Fig. 2.7b). A vertical hydraulic conductivity (K_v) of 1×10^{-11} m/s was estimated due to anisotropy as slug testing is more representative of horizontal hydraulic conductivity (K_h). However, vertical conductivity (K_v), may be 2 to 20 times less than K_h in some porous media such as clay (Freeze and Cherry, 1979). The existence of a capillary break below the inverted water table surface suggests that the effective hydraulic conductivity should be at least an order of magnitude lower than the saturated K_v. Resultant vertical leakage through the confining layer using the estimated K_v is very small, accounting for <1 mm lost annually from each wetland basin, despite large vertical gradients. Piezometers installed below the confining layer in the unsaturated zone in the P19 riparian zone, at the edge of W1, and in the ephemeral draw show that the perched flow regimes are not coupled with regional groundwater flow systems.

Outside the wetland basins, lateral saturated groundwater flow was limited to the ephemeral draw area near P19. The confining layer morphology in the ephemeral draw generally follows the land surface contours limiting soil water storage in the overlying mineral soils, resulting in frequent periods of saturation. Continuous head data from both hydrologic years were used to complete a transient analysis of the hydraulic head gradient between the draw outflow and the pond to calculate the volume of perched groundwater entering the pond basin via saturated flow. In 2005, 430 m^3 of perched groundwater flowed into the P19 basin. In 2006 significantly less saturated flow of perched groundwater was observed, with only 110 m^3 of groundwater entering the basin.

2.4.4 Surface water flow regime

Surface runoff from upland areas was not observed during the study period, similar to results from upland runoff and infiltration experiments completed at a site approximately 4 km away (Redding and Devito, *in press*). Their experiments show that runoff generation as lateral flow towards the wetland basins only occur in response to storms occurring every 50 to 100 years. The experimentally determined rain intensity of >50 mm/h required to generate down-slope lateral flow is far greater than the maximum precipitation intensity observed during the study period of <10 mm/h. Ice probing measurements in the upland areas indicate only localized, laterally discontinuous soil frost development in the well drained, near surface soils.

Active flow of surface water was spatially limited to the pond edge and base of the ephemeral draw draining into the P19 basin (Fig. 2.10). Surface flows resulting from quick flow over frozen soil or saturated overland flow mechanisms were present for 96 days in 2005 with a maximum flow rate of 0.45 L/s, and for 12 days in 2006 with a maximum flow rate of 0.20 L/s. The difference in duration and magnitude of observed surface flows accounted for a larger positive lateral flux of surface water into the pond basin in 2005 relative to 2006. In 2005, the ephemeral draw area contributed 890 m^3 to the pond basin. A portion of the overall surface water volume (290 m^3) measured entering the pond basin is flow-through component occurring as over-ice flow in the top 0.05 m of soil or as surface flow, generated by rapid melting of snow banks along the road and spring rain events. By contrast, in 2006, only 60 m^3 of surface water entered the P19 basin, due to minimal soil frost development, small SWE, and less saturated overland flow generation in the ephemeral draw during the summer months.

The pH and EC of surface water in the ephemeral draw indicate that extensive surface water/groundwater interaction occurs as water flows towards the pond. Surface water was measured to increase from a pH of 6.3 and an EC value of 100 μS/cm at its upper reaches, to a pH of 7.0 and EC of 1200 μS/cm at its outflow area, The surface water present in the ephemeral wetland to the northwest of the pond (Fig. 2.1) had EC values ranging from 60 – 100 μS/cm and pH values of 6.2 to 6.3, indicating that ephemerally saturated areas result from rain water exceeding their minimal storage capacity above the shallow confining layer.

The pond pH was 5.8 with an EC value of 105 μS/cm in late May 2006. Spring melt and heavy precipitation events caused an influx of local (perched) groundwater, increasing the pH to 7.9 and diluting the EC by 30 μS/cm after the bulk of the 2006 summer precipitation. Periods of net atmospheric flux deficit showed a trend of decreasing pH and demonstrated evapo-concentration effects coincident with the trends in decreasing pH during strongly evaporative periods in the summer of 2006. The pond water EC increased from 74 μS/cm on July 29 to 140 μS/cm on August 31, a period with a net atmospheric flux deficit of 35 mm (Fig. 2.11).

2.4.5 Soil moisture regime

The soil moisture regime was characterized within the transition area from wetland to upland vegetation along the detailed groundwater transects (Fig. 2.7). These transition areas were approximately 25 m wide and occurred near the saturated/unsaturated interface at the peripheries of both P19 and W1 (Fig. 2.3; Fig. 2.7). Data from the soil moisture transects (Fig. 2.12), show that both P19 and W1 had similar spatial and temporal soil moisture regimes at their peripheries during 2006.

2.4.5.1 *Soil tension and unsaturated groundwater flow*

All the 2005 and 2006 data showed identifiable spatial and temporal patterns of soil tension at the periphery of P19 and W1. Hydraulic head in the unsaturated zone was calculated from measured soil tension data. At P19 nests PN3, PN4 (Fig. 2.3), showed that the hydraulic head in the centre of the rooting zone, or the intermediate depth tensiometer cup, was >6 m lower than the head at the cups installed approximately 0.4 m directly above and below the intermediate depth. At W1, nests WN3 and WN4 consistently showed a similar pattern with the tension data indicating approximately 5 m lower hydraulic head in the centre of the rooting zone. Resultant vertical hydraulic head gradients of 10 – 15, indicated highly focused unsaturated flow in opposite directions towards the centre (i.e., 0.5 – 0.6 m from surface) of the rooting zone at both PN3 and WN3 (Fig. 2.12). The influence of infiltration events was observed at both tensiometer nests. Vertical gradients dissipated to almost zero in response to infiltration following precipitation. These events were only observed following rainfall events >15 mm/day. These large precipitation events exceeded the interception capacity of the upland vegetation and forest floor allowing infiltration of precipitation into the underlying mineral soils. These infiltration events would also occur after prolonged periods (2 – 5 days) of precipitation. When the precipitation ceased, depending on the magnitude of the event, the large gradients (10 – 15) were re-established in the aspen rooting zone within 3-5 days at P19 and slightly longer at W1. The time until the focused gradients were reestablished was dependant on the daily aspen evapotranspiration rate which consumed infiltrated water from soil water storage (Fig. 2.9; Fig. 2.12).

At P19, lateral hydraulic head gradients in the rooting zone varied from 0.83 between PN2 and PN3, and 1.7 between PN3 and PN4. This difference is likely related to the finer soil texture in the upland soils between PN3 and PN4 supporting a larger gradient than that measured between PN2 and PN3. Similar lateral gradients were observed at W1 with a gradient of 1.2 between WN3 and WN4. Lateral gradients were not determined between tensiometer nests at WN2 and WN3 at W1, as WN2 was very close (0.1 to 0.3 m) to the perched water table and likely outside of the active aspen rooting zone due to high water content.

Using measured lateral gradients, and an assumed unsaturated hydraulic conductivity value of approximately 5×10^{-7} m/s (Section *2.3.2*) for the rooting zone soils, the specific discharge or unsaturated Darcy flux of perched groundwater was determined for the periphery of each wetland. At P19, the unsaturated Darcy flux was estimated to be 3.4×10^{-4} m^2/s, which was similar to the estimate at W1 (1.1×10^{-4} m^2/s). The same methodology was used to determine the unsaturated flow from the ephemeral draw to the surrounding upland area. No tensiometers were installed at the draw to upland transition, so the same unsaturated Darcy flux as the P19 transect was assumed to be representative of the unsaturated flow regime at the draw to upland transition.

Unsaturated hydraulic head data calculated from soil tension data collected in August and September 2005 appeared to be similar to that from the same time period in 2006. However, an estimation of the volume perched groundwater lost to the upland from the P19 basin was required due to incomplete data from earlier in 2005. This was done by scaling the volume of perched groundwater lost in 2006 based on the ratio of respective *PET* values measured each year. This analysis indicated that the P19 basin could have lost 1700 m^3 to the surrounding upland in 2005. It is estimated that the ephemeral draw also lost an additional 290 m^3 of water from its periphery into the surrounding upland during the 55 days in which saturated conditions were observed (Fig. 2.13). Analysis of data collected throughout the entire growing season of 2006 indicated that the P19 basin lost 2300 m^3 from May to October 2006. The ephemeral draw was saturated for 45 days in 2006, and an unsaturated flux volume of 240 m^3 was estimated to have entered the surrounding aspen upland (Fig. 2.13). During 2006, the W1 basin lost 790 m^3 to the surrounding upland over the growing season (Fig. 2.13).

2.4.5.2 Vadose zone soil moisture

All near surface (0.2 – 0.3 m) TDR probes at both wetlands had more pronounced response to snowmelt and precipitation event response and minimal annual change in water content relative to deeper installations (Fig. 2.9; Fig. 2.12). The melt response measured by TDR probes is larger than the storage change calculated for snow melt (Fig. 2.9). This is likely due to a concurrent influence from snow melt entering soil storage from infiltration, and the lateral movement of the perched water table and capillary fringe towards the upland. At P19, TDR probes at PN1 and PN2 located close to the perched water table in the capillary fringe, recorded significant soil water content change over 2006 reflective of the contraction of the perched water table measured in the pond basin. At W1, TDR probes at WN1 and WN2 showed negligible changes in soil moisture storage corresponding to the water table position not changing over 2006 (Fig. 2.12). At both wetlands, TDR installations at PN3/WN3 and PN4/WN4 measured soil moisture changes representative of the upland rooting zone, rather than the capillary fringe. These data reveal that changes in soil moisture were relatively small over the 2006 hydrologic year (Fig. 2.12).

2.4.6 Summary of hydrologic budget components

Transient (daily) analyses of the flux terms in each catchment sub-component are summarized on Figure 2.13 as cumulative volumes (m^3) within each catchment sub-component P19 (2005 and 2006) and W1 (2006). The overall interaction between catchment sub-components and the water balance for the entire catchment, expressed in mm of water, is summarized in Table 1.

2.4.6.1 P19 hydrologic budget summary

Despite loosing 31 mm as UGW_{out} to the surrounding upland in 2005, the P19 basin had a measured positive change in storage of 26 mm. This positive storage within the pond basin was due to 24 mm of lateral inflow (SW_{in} and SGW_{in}) from the ephemeral draw. Without the addition of positive lateral fluxes to the basin, only one quarter of the observed storage change would have occurred, as all but 6 mm of the net atmospheric surplus of 37 mm ($P_{net} - AET_{RZ}$) over the pond basin area was lost to surrounding uplands.

The ephemeral draw had a large negative change in water storage over the 2005 hydrologic year. Heavy precipitation in late summer and fall 2004 created high antecedent moisture conditions and limited storage, thus "priming" the ephemeral draw for spring melt in 2005. The large ΔV of between -130 to -190 mm for the draw indicates a delayed release of water stored as soil frost from late fall precipitation in 2004 via lateral fluxes to the pond basin, and as unsaturated flux into the surrounding uplands. The ephemeral draw lost 190 mm as lateral fluxes (SW_{out} and SGW_{out}) to the pond basin, and 41 mm as UGW_{out} to the upland area. This results in a large (57 mm) residual within the ephemeral draw, but due to the small contributing area of the ephemeral draw, has minimal effect on the overall catchment residual. By the end of the 2005 hydrologic year, dry antecedent moisture conditions were present in the draw due to low fall precipitation values, typical of the climate patterns of the region.

In 2005, the upland area had a net atmospheric deficit of 31 mm. However, the calculated and observed ΔV terms for the upland area were minimal due to the unsaturated flow in (UGW_{in}) from the pond basin contributing 27 mm, plus an additional 4 mm of UGW_{in} from the ephemeral draw, relative to the upland area. The sum of the UGW_{out} components from the P19 basin and draw area is 31 mm, which correlated well with the negligible changes in storage observed in the upland during the growing season (May – September) despite the estimated annual net atmospheric flux deficit.

The hotter, more evaporative, conditions in 2006 resulted in a small net atmospheric flux surplus of 7 mm over the basin area contrasting the large surplus observed in 2005. There was a corresponding decrease in basin water storage of 29 mm. The loss of water from the basin was related to the net atmospheric deficit created by the additional demand of the UGW_{out} term, and insignificant contribution from the ephemeral draw (i.e., SGW_{in} of 3 mm and SW_{in} of 1 mm), and a larger UGW_{out} term of 41 mm from the P19 basin.

The small flux contributions to the P19 basin from the ephemeral draw in 2006 contrasted the 2005 observations because of very dry antecedent moisture conditions in the spring, and less than average SWE and spring precipitation. These factors resulted in shorter periods of saturation in the ephemeral draw and a smaller UGW_{out} term of 3 mm relative to the upland area over the summer months. This resulted in a negligible change in ephemeral draw storage, with only 1 mm residual between calculated and observed storage change for 2006.

Summing the UGW_{out} terms from the P19 basin and the draw in 2006 results in an UGW_{in} value of 39 mm for the upland, which nearly offsets the 47 mm net atmospheric deficit in the upland for the year. The calculated ΔV term indicates a loss of 8 mm of stored water within the upland rooting zone. TDR data from the shallow rooting zone indicate a loss of 4 mm, with greater loss from deeper upland TDR probes at the base of the rooting zone, suggesting good agreement between calculated storage losses and the soil moisture data from 2006 (Fig. 2.9; Fig. 2.12).

2.4.6.2 *W1 hydrologic budget summary*

The water budget for W1 was only completed for the 2006 hydrologic year, as it was instrumented in 2005. The water budget for the W1 catchment required consideration of only the wetland basin and upland area subcomponents because no inflow features such as the ephemeral draw at P19, were identified. In 2006, the W1 basin had a net atmospheric surplus of 47 mm (Table 1). However, there was no observed change in storage in the wetland basin over 2006. The UGW_{out} term at W1 was slightly larger proportional to basin area relative to the P19 UGW_{out} term. It accounted for a loss of 45 mm from the wetland basin, resulting in a 2 mm residual for the W1 basin. There were no SGW_{in} or SW_{in} terms associated with the W1 basin, as these were positive flux terms associated with the ephemeral draw area at P19, rather than upland areas.

The 2.1 ha upland area gained 38 mm from the unsaturated flux from the W1 periphery, which is proportional (in mm) to the volume of water lost at P19 relative to the contributing area at each wetland periphery (Table 2.1). This resulted in a 6 mm residual between calculated and observed upland storage changes. However, calculated upland storage changes are based on TDR data near the wetland edges (Fig. 2.9; Fig. 2.12), versus the upland crests, which may have greater negative changes in storage.

2.5 *Discussion*

2.5.1 *Requisite conditions for perched wetlands in the WBP*

Relatively little is known about the role of perched wetlands and groundwater flow systems in sustaining regional groundwater resources (O'Geen *et al.*, 2003). Perched flow systems are typically documented as highly transient elements of larger scale flow systems and are seldom the focus of hydrologic research (Rockafeller *et al.*, 2004). Furthermore, little, if any literature exists on the hydrology of permanent perched boreal wetlands and their role in the WBP. This study has shown that the hydrologic environment of perched boreal wetlands show greater similarities to non-perched moraine wetlands in the WBF (Ferone and Devito, 2004) and prairie pothole region (Mills and Zwarich, 1986; Hayashi et al., 1998), than other documented perched wetlands (Brooks and Hayashi, 2002; O'Geen *et al.*, 2003; Rockafeller *et al.*, 2004: Rains *et al.*, 2006). However, results also show permanent perched wetlands have water table configurations and unsaturated flow regimes that are unique manifestations of the climate, spatial distribution of storage potential, and vegetation patterns in the WBP.

The study site has hydrostratigraphy that is similar to other documented perched wetland flow systems, such as vernal wetlands in humid forest regions of the eastern U.S. and Canada (Brooks and Hayashi, 2002), and in arid regions of the western U.S. (O'Geen *et al.*, 2003; Rockafeller *et al.*, 2004; Rains *et al.*, 2006). These perched systems exist in coarse-grained sediments overlying near surface confining layers that impede vertical flow or recharge to underlying regional water tables. During seasonal surplus conditions, laterally extensive perched groundwater flow systems develop within thin soil veneers (0.1 -1.5 m) overlying low-permeability bedrock (Brooks and Hayashi, 2002), or hardpan horizons (Rockafeller *et al.*, 2004; Rains *et al.*, 2006). Perched wetlands are usually surface expressions of perched flow systems and are found where subtle thinning of the permeable soils over the confining layer occurs (Rains *et al.*, 2006). This study also shows that a laterally extensive clay confining layer exists (Fig. 2.2, Fig. 2.6a); however, there is <1 to >10 m of relatively coarse porous media overlying the confining layer. This variable depth creates large spatial variations in effective soil storage (Fig. 2.6b). This hydrostratigraphic and climatic framework leads to permanent, laterally isolated, perched water tables, and governs the hydrology of the site. Within this climate, perched wetlands represent area for saturated conditions to develop in an unsaturated environment. The correlation of wetland occurrence with areas with <2 m of porous media overly the confining layer indicate that storage potential dictates wetland position. Alternately, unsaturated upland areas are found in areas with higher soil storage potential where the confining layer is encountered at depths >2 m from ground surface. Unlike vernal pool landscapes, where the confining layer morphology and perched flow-through systems follow topographic gradients, perched boreal wetlands and groundwater flow systems only exist where net atmospheric inputs and local hydrologic processes can exceed soil storage.

The low relief, hummocky topography of the boreal landscape is similar to the prairie pothole region, with poorly defined surface drainage networks (Winter and Rosenberry, 1995; Hayashi *et al.*, 1998). However, perched boreal wetlands fundamentally differ from prairie potholes as shallow (<5 m) regional water tables within upland areas of the prairies. The prairie wetlands, therefore, locally modify the regional groundwater flow systems (Meyboom, 1966; Mills and Zwarich, 1986; Rosenberry and Winter, 1997; Hayashi *et al.*, 1998). In prairie pothole catchments, the presence of shallow, low permeability soils, and the depth to the regional water table dictate the spatial distribution of soil water storage. In the boreal forest, the regional water table is often located at large depths from surface, so the depth to a confining layer and temporal climate patterns control soil water storage potential, independent of regional water table position. The hydrogeologic and climate characteristics at URSA create saturated wetland areas, adjacent to upland areas with large storage potential with tens of metres of unsaturated porous media. Thus, the soil storage potential, frequent net atmospheric flux deficit, interception loss, and high evaporative demand of upland vegetation prohibit laterally extensive perched flow systems and saturated lateral flow in uplands, as observed in prairie pothole landscapes (Mills and Zwarich, 1986).

2.5.2 Hydrologic response unit characteristics

The perched wetland flow systems and wetland-to-upland flux dynamics observed in this study are similar to wetlands in the sub-humid to semi-arid Great Plains region of North America. Studies by Meyboom (1966), Mills and Zwarich (1986), Winter and Rosenberry (1995), Rosenberry and Winter (1997), and van der Kamp and Hayashi (1997), and Hayashi *et al.* (1998) document saturated flow of groundwater from the wetland centre to its periphery because of upland water table depression from willow ring evapotranspiration. Similar water table depression at wetland margins is observed at the study wetlands and other non-perched wetlands at URSA (Ferone and Devito, 2004; Petrone *et al.*, 2007). However, the flux dynamics documented during this study rely exclusively on unsaturated flow to recharge upland areas, as there is no upland water table, and no significant saturated groundwater flow in upland areas. The high storage potential of the uplands also limits upland contributions to the wetland basins from runoff, even during snow melt. As such, upland vegetation largely draws water from soil storage, resulting in upward vertical gradients in the vadose zone above the perched water table, similar to those noted by Schuh *et al.* (1993) and Hayashi *et al.* (1998). However, given the absence of an upland water table, lateral unsaturated hydraulic head gradients were also observed in shallow soil horizons within the upland rooting zone. This indicated lateral unsaturated flow from the saturated/unsaturated interface at wetland peripheries as opposed to prairie systems where saturated flow is the dominant upland recharge mechanism (Meyboom, 1966; Hayashi *et al.*, 1998).

Spatial and temporal distributions of soil storage and local perched flow systems exert the dominant control on the water flux dynamics within the catchments. This has not been documented in prairie pothole regions. Previous studies have concluded that although similar processes are observed, they only modify the overall flux dynamics controlled by a regional scale flow system and (Mills and Zwarich 1986; van der Kamp and Hayashi, 1997). With the exception of the inverted water table surface, the perched water table configuration at the study wetlands closely resembles those documented on other fine-grained landforms (moraine and clay plain) in the Western Boreal Forest (Ferone and Devito, 2004, Devito *et al.*, 2005a; Petrone *et al.*, 2007). This suggests that the documented wetland-to-upland flux dynamics, including lateral unsaturated flow into uplands, and limited regional groundwater recharge are likely representative of the many of wetlands in the region.

The intra-catchment flux dynamics and isolation from regional flow systems differ from depression focused recharge wetlands in prairie landscapes, but the hydrologic function of perched boreal wetlands is similar. Prairie potholes are critical for recharging uplands (2 – 40 mm annually; van der Kamp and Hayashi, 1997), which are connected to regional groundwater flow systems. Perched boreal wetlands play a minimal role in recharging regional groundwater flow systems through vertical leakage, but are crucial for storing water, buffering water deficit conditions, and sustaining upland vegetation. Ephemeral "recharge rings" at the edge of perched groundwater lenses may occur during infrequent periods of extreme water surplus. Recharge may occur via deep macro pores that develop in upland porous media from deep root penetration and fracturing due to cyclical desiccation/rewetting at wetland peripheries. This is evidenced by oxidized soil fractures and higher moisture contents relative to upland sediments in soil obtained from boreholes completed near perched wetland peripheries. This type of recharge is thought not to have occurred during the study period, as the perched groundwater lens did no extend in to upland areas. In addition, the water balance indicates that vertical recharge to the underlying flow system is a negligible relative to lateral "upland recharge".

The upland-to-wetland flux dynamics also differ from prairie potholes, where snowmelt runoff from uplands dominates water budgets (Woo and Rowsell, 1993; Hayashi *et al.*, 1998). TDR data indicated storage increases proportional to SWE from melt (Fig. 2.12), suggesting the uplands surrounding the perched wetlands did not contribute significant positive fluxes to the wetland basins during the study period. Water budget analyses and snowmelt runoff generation experiments completed at URSA (Redding and Devito, 2005), indicate that snowmelt contributions to ponds and wetlands are much smaller in magnitude relative to prairie potholes. However, small positive contributions from the base of the uplands may occur if concrete frost develops adjacent to the wetland areas. Runoff generation is limited to areas with small storage potential that can be overcome by precipitation and snowmelt, or that maintain high antecedent moisture conditions late in the year allowing concrete frost to develop. Saturated overland flow and flow over surficial ice lenses was observed in the ephemeral draw and riparian areas, similar to Devito *et al.* (2005b). Concrete frost observed at the ephemeral draw/upland transition zone in 2005 due to high antecedent moisture conditions from heavy fall precipitation in late 2004. This allowed some upland snowmelt run-off as a flow-through (*FT*) contribution in the 2005 water budget, measured as surface flow from the ephemeral draw (Fig. 2.11a). Summer run-off was restricted to the ephemeral draw area and small contributions from the riparian zone where high water table position limited soil storage, as observed by Hayashi *et al.* (1998) (Fig. 2.10; Fig. 2.13). These results indicate perched wetlands on the boreal plain have dynamic flux regimes that vary with temporal distribution of climatic inputs and spatial distribution of available soil storage within each catchment.

2.5.3 Conceptual model, ecohydrology, and disturbance

Although vertical water fluxes dominate the water movement in this environment, the lateral flux of surface water and perched groundwater also play important roles in wetland development, connectivity, and permanence (Fig. 2.4). Assessing the water budgets of the study wetlands confirmed that the conceptual model correctly identified atmospheric fluxes and soil storage as the hydrologic drivers within perched wetland catchments. Wetland susceptibility can also be assessed based on potential disturbance to secondary processes such as unsaturated fluxes, focused run-off generation, and wetland connectivity. The application of the same conceptual model and low water budget residuals to both study wetlands indicate that the interpretation of field data and assumptions employed, provide a reasonable, broadly applicable conceptual model for perched boreal wetlands. The surface water and perched groundwater chemistry observations also corroborate the interpretation that the spatial distribution of storage potential controls groundwater/surface water interaction and flow path.

Because of the importance associated with atmospheric fluxes within the conceptual model, the model is sensitive to spatial and temporal designation of AET and P_{net} to the various sub-components of each catchment as they are applied over large areas and dominate the water budgets. Although the scope of this study did not allow for precise measurement of AET in all of the sub-components of each catchment, gross estimates of AET_{RZ} based on TDR data interpretation provided reasonable values when compared to literature values for AET measured at sites with similar climate, ecoregion, and soil characteristics. Similarly, P_{net} has a large influence on the water budgets given that the climatic forcing of the perched systems and the interception losses are highly site specific and dependant on the local climate (Crockford and Richardson, 2000). The observed storage changes measured by upland TDR probes correlate well with the calculated change in upland storage for summer precipitation events. Inconsistencies during the summer precipitation period (Fig. 2.9) are likely related to the generalized AET estimate utilized (Amiro *et al.*, 2006). Average daily evapotranspiration rates for each month do not account for the magnitude or intensity of precipitation events, air temperature variation, or the transient nature of canopy development, all of which could introduce error into the developed interception loss relationship. Despite the error inherent to the atmospheric flux estimates, the cited AET estimates, and the method used to determine interception loss allowed adequate characterization of storage changes for upland and riparian areas (Fig 2.9).

The redistribution of water in the unsaturated zone past the wetland/upland transition is not addressed here because it can not be determined whether lateral gradients exist farther up the upland slope. Future work requiring an increased instrument array and study of the aspen root physiology would be required to determine the spatial and temporal redistribution of the UGW_{in} flux term within the upland areas.

The hydrology of the study site is concurrently determined by the intensity and distribution of precipitation and evapotranspiration relative to the spatial/temporal distribution of water storage at the study site. Both permanent and ephemeral perched wetlands/flow systems occur in areas where soil moisture storage can be overcome by atmospheric inputs and local hydrologic processes. Unsaturated uplands are located in areas where 2 – 15 m of well-drained coarse textured mineral soils overly the confining layer (Fig. 2.6b). The permanence and type of perched wetland (pond or peatland) depends on the proximity to a perched groundwater flow divide (Fig. 2.4) created by a topographic high in the confining layer, basin morphology, and/or basin volume. For instance, a cluster of peatlands (W1, W2, and W3) exists near a ridge-like feature in the confining layer topography. This feature limits the area from which potential saturated groundwater flow and runoff can be generated, and conveyed into wetland areas (Fig. 2.6a), limiting the potential for periodic inundation of the peatland basins. Consequently, annual storage changes are minimized as excess precipitation either enters the uplands via unsaturated flow, or is stored in the peat as its volume increases in response to the pore pressure (Ingram *et al.*, 1974). Fractured and oxidized soils were focused within <2 m of the current water table position at several peatland transects in the study area, indicating that water table positions do not vary significantly, and annual storage oscillations are minimal.

P19 is located further down gradient from the perched groundwater flow divide relative to W1. This creates a larger area from which runoff can be potentially generated. The ephemeral draw, located between P19 and the perched groundwater flow divide, has a sloped, trough-like confining layer. This morphology promotes saturated flow and connectivity to the P19 basin during surplus conditions. Lateral saturated flow from the ephemeral draw and the riparian zone (areas of low soil storage potential and high runoff potential) cause inundation and the major oscillations in storage within the P19 basin that are observed from air photo analysis. Soil properties indicate that there is an extensive (15 m) zone of fractured and highly oxidized soils at the wetland periphery extending into the upland also confirm significant temporal variation of the perched water table location.

The water table oscillations result in a permanent pond that temporally varies in area and volume, with an adaptable riparian area that thrives in variably saturated conditions.

These factors indicate that the basin morphology for a permanent perched boreal pond requires focused lateral flux inputs and a small degree of wetland connectivity to ephemeral wetland features to prevent long-term terrestrialization. Under typical climatic conditions, soil storage and transpiration demands exceed P_{net} in unsaturated upland areas (Price *et al.*, 2005). This water deficit in upland areas effectively prevents development of laterally extensive perched aquifers, through-flow of perched groundwater, and extensive wetland connectivity.

The hydroperiod, of ephemeral depression wetlands at the study site, or portion of the year in which surface saturation or hydric soil is present, is correlated to basin area and volume, mimicking observations of vernal pool wetlands by Brooks and Hayashi (2002). The ephemeral wetland basins at the study site had insufficient basin storage to maintain standing water throughout the year, particularly given the loss of water to the uplands through unsaturated groundwater flux out of ephemeral wetlands.

The presence of these ephemeral features beside permanent wetlands, suggests that the perched boreal wetland hydroperiod and riparian evaporative regimes follow the "shoreline loss" concept developed by Millar (1971) for prairie sloughs. This concept states that the smaller the wetland, the larger the role of "willow ring" evapotranspiration (i.e., UGW_{out}) in its water budget. Perched wetlands must be of a critical size to be permanent features. That is, saturated conditions in larger wetlands are maintained by storage volume and evaporative loss regulation mechanisms that limit UGW_{out} as the water table contracts away from the aspen roots ability to draw water from storage or the saturated/unsaturated interface. Riparian zone vegetation AET is also limited as the water table recedes from the surface during highly evaporative periods. Peatlands also regulate evaporative loss by increasing the albedo of the peat, which becomes blanched and lighter in colour, during periods of water deficit (Rouse, 2000).

Disturbance scenarios such as deforestation for agriculture in northern Idaho have increased seasonal perched water table levels above shallow fragipan horizons by 6 – 107 % in upland areas, potentially altering the timing and magnitude of hydrological processes (Rockafeller *et al.*, 2004). However, development of laterally extensive perched water tables in upland areas due to aspen canopy removal is unlikely at the study site. The magnitude of available storage in mineral soils, interception loss from residual understory species, and forest floor interception (which accounted for >60% of total upland interception loss) would likely buffer canopy removal effects sufficiently until aspen regeneration commences.

2.6 Conclusions

Perched wetlands common in the Western Boreal Plain are ecohydrologic features that persist in the delicate balance between atmospheric fluxes, storage, and localized hydrologic processes. The environment sustaining permanent, perched wetlands features includes a fine textured, low permeability soil horizon (confining layer) within a coarse textured host material and a deep regional water table. Although this framework is quite specific, it is not likely to be unique as this environment is found between moraine and outwash plain deposits, two common physiographic landforms in the Western Boreal Plain. The hydrostratigraphic and climatic framework of the study site create complex, transient local hydrologic processes requiring the development of a highly specialized conceptual model to complete a water budget, and assess the sustainability of the wetlands.

This study only encompassed two hydrologic years. However, this work has contributed to our understanding of the roles that several hydrologic processes play in sustaining perched boreal wetlands. Subtle processes, such as interception loss, unsaturated flow of perched groundwater, and localized lateral flow generation mechanisms at the study site result in permanent, isolated hydrologic systems. These systems represent HRU's with no broad-scale hydraulic or biogeochemical connection to other catchments or regional groundwater resources. The large amount of storage associated with these perched wetland catchments result in a localized landscape function that sustain wetland/upland complexes, independent from large-scale flow systems. This indicates perched boreal wetlands are even more sensitive to variations in precipitation and evaporative demand than the wetlands and shallow lakes on the outwash plain area of the URSA (Smerdon *et al.*, 2005), and that they act as windows for increased upland recharge within the sub-humid climate. The scope of this study did not allow for assessment of the long-term sustainability or connectivity of perched boreal wetlands;

however, ongoing intensive field studies and numerical modeling approaches will facilitate further study.

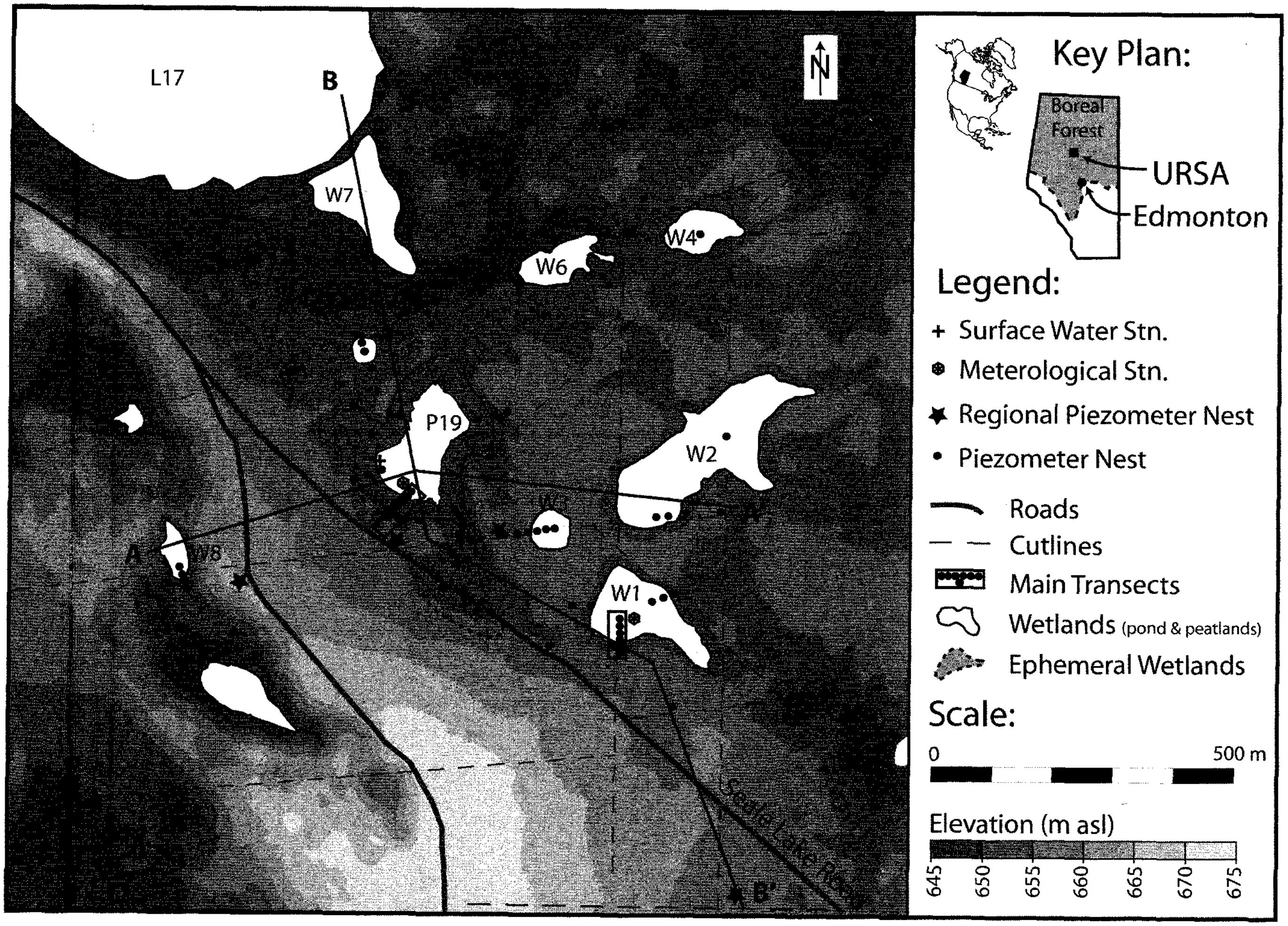

Figure 2.1. P19 study site with topography and hydrologic instrumentation. A - A' and B - B' represent plan view section lines of cross-sections shown on Figure 2.2. Utikuma Region Study Area (URSA) and Boreal Forest of Alberta, Canada in legend.

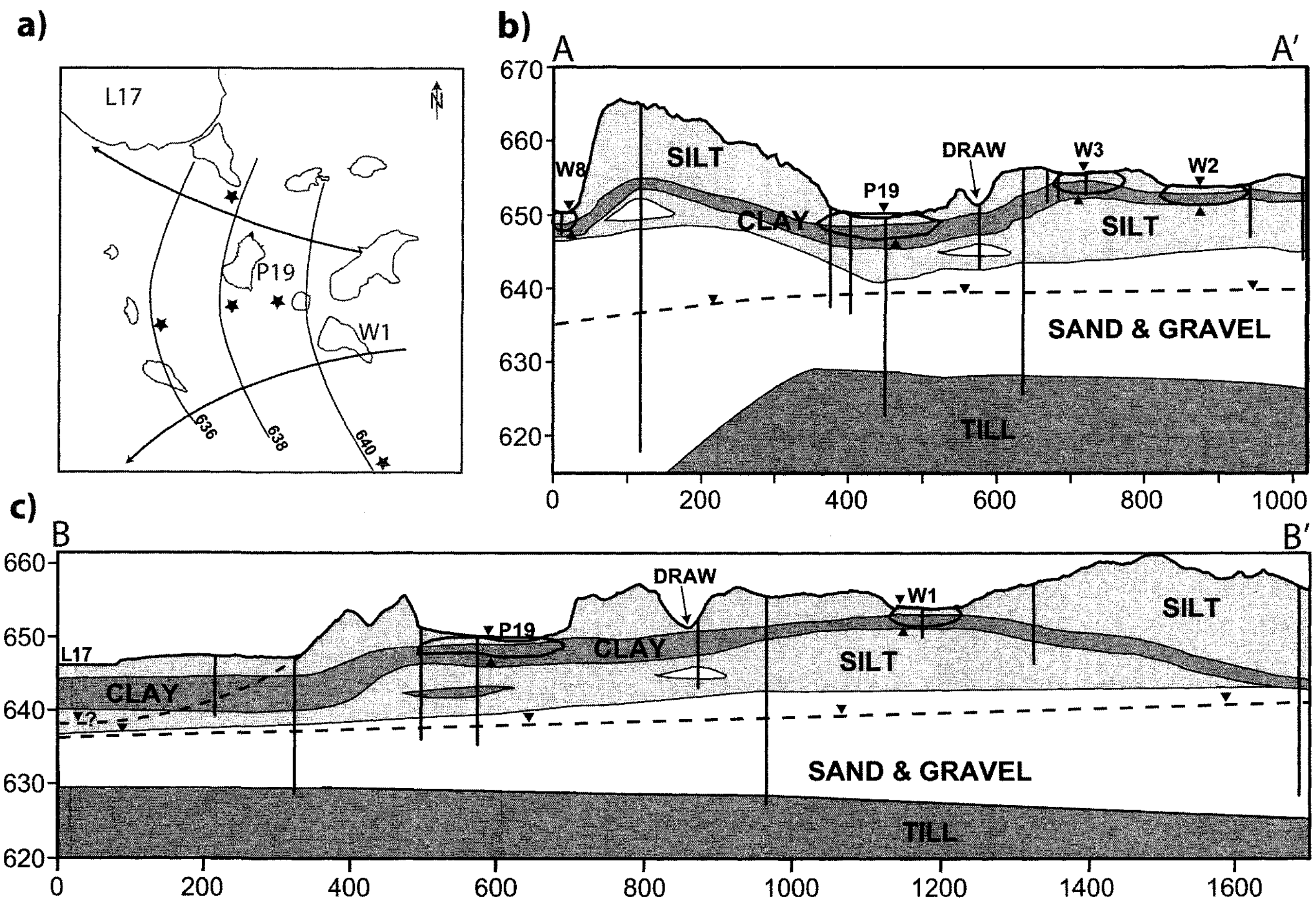

Figure 2.2. a) Regional water table elevation contours and flow direction showing regional groundwater flow divide based on deep piezometer data. Cross-sections A-A' shown in b), B-B' shown in c), with perched water tables, wetlands, the regional water table, unsaturated zone, transition zone geology, and hydrostratigraphic units based on data projected borehole data (vertical black lines). Section locations shown on Figure 2.1.

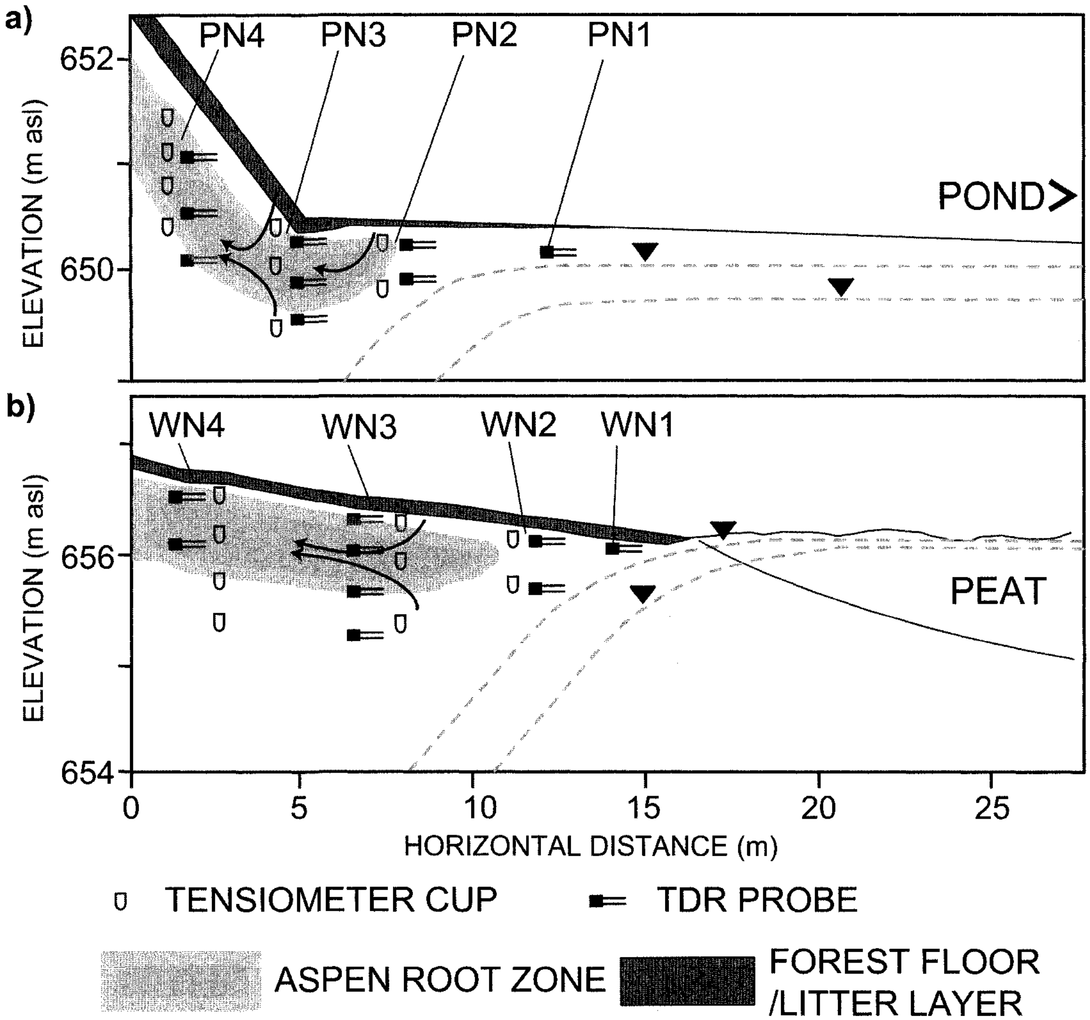

Figure 2.3. Detailed soil moisture transect instrumentation locations at a) P19, and b) W1. Maximum and minimum observed water table positions over the study period are marked by the dashed lines. Arrows represent unsaturated flow patterns in the aspen upland rooting zone at the wetland to upland transition (based on soil tension gradients).

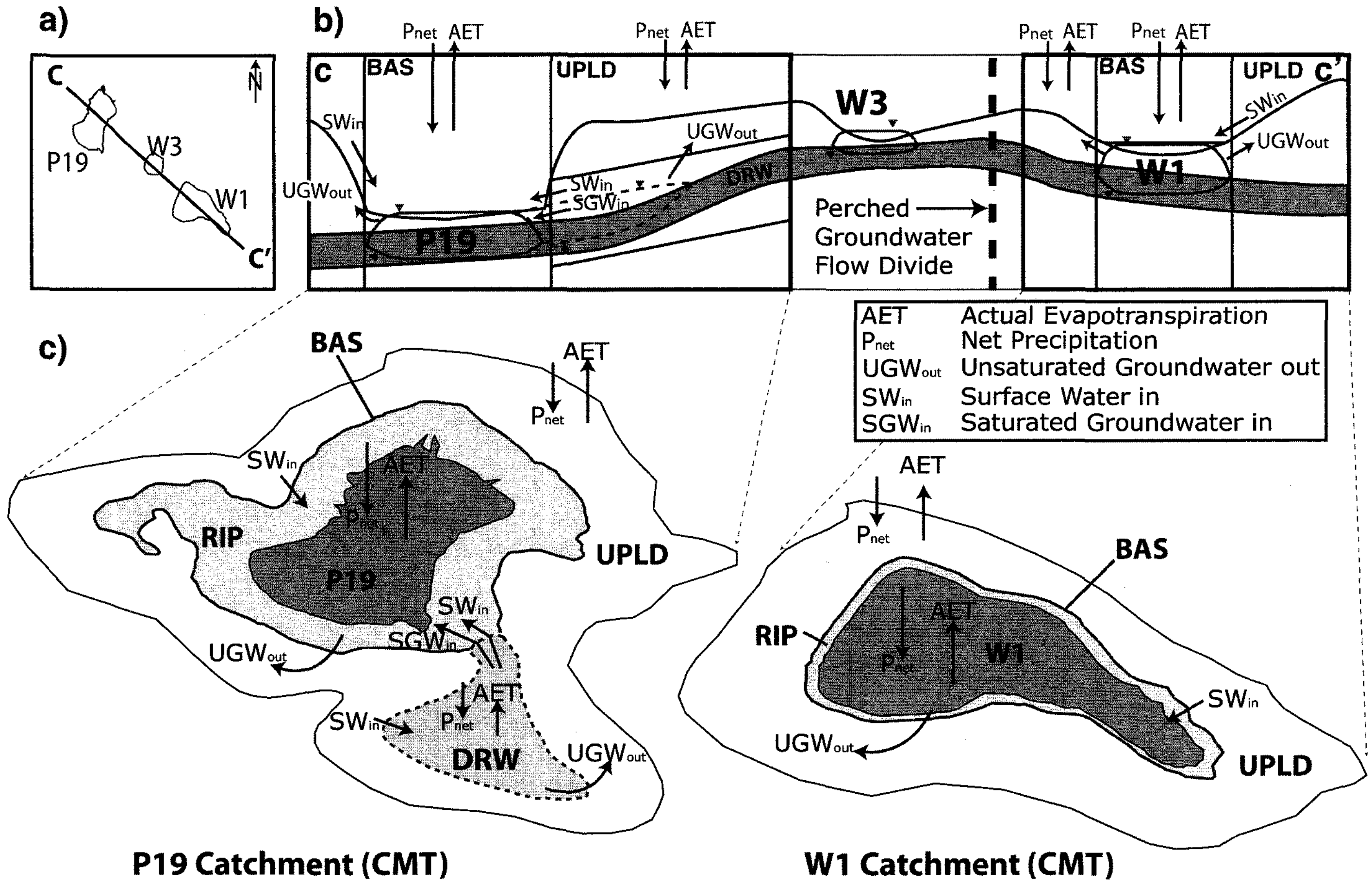

Figure 2.4. Schematic representation of perched wetland catchment boundaries and conceptual model along C -C' transect (a), in cross section (b), and in oblique plan view (c) showing catchment sub-components (bolded), and water flux terms and intra-catchment flux dynamics within each catchment.

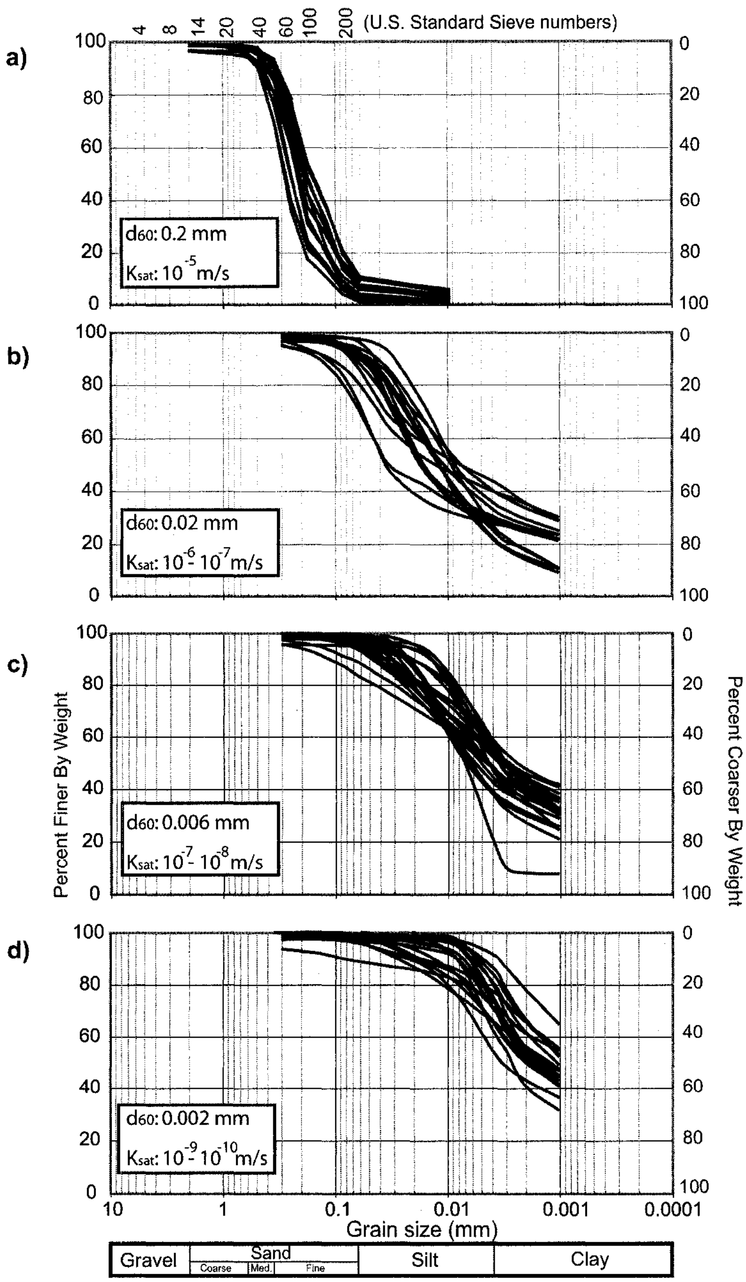

Figure 2.5. Grain size distributions for a) outwash deposits (Smerdon *et al.*, 2005), b) sandy silt, c) silt, and d) clay materials. Arithmetic average grain-size distributions and measured hydraulic conductivity ranges on inset tables for each material type.

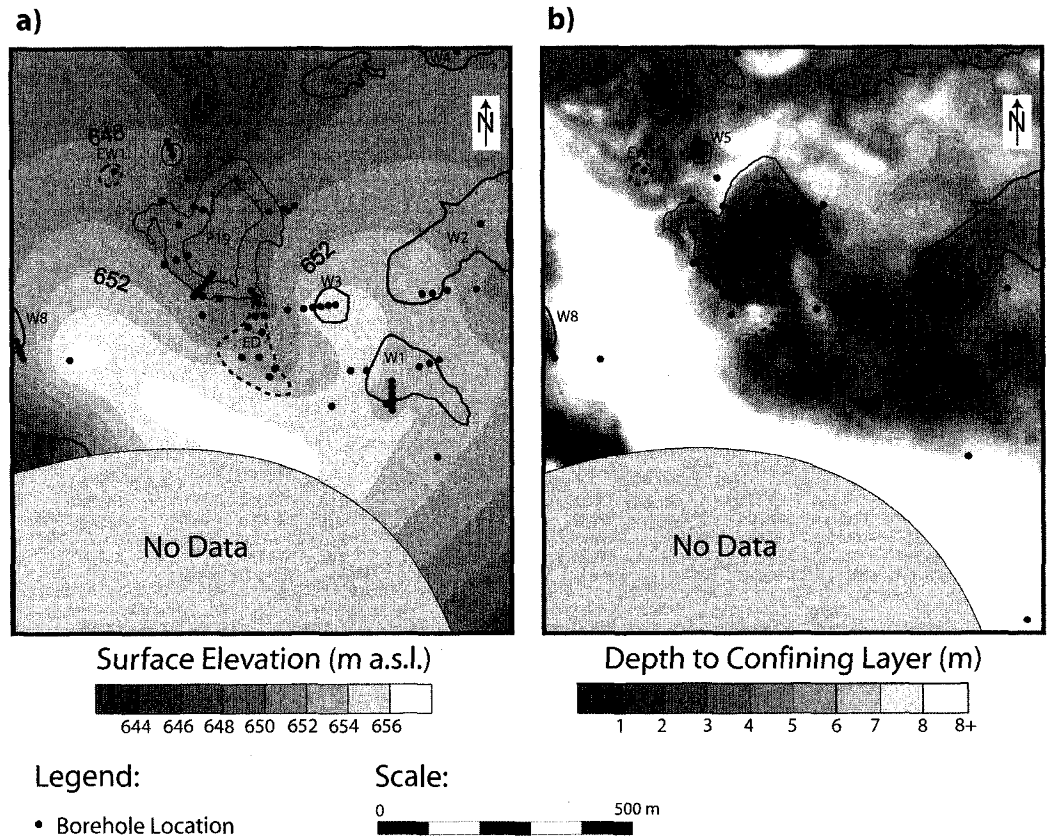

Figure 2.6. a) Kriged surface elevation map of top of clay confining layer as determined from borehole data marked by black dots. b) Depth to clay confining layer from land surface across study site generated from Map a), and LIDAR digital elevation model. (LIDAR data provided by I.F. Creed)

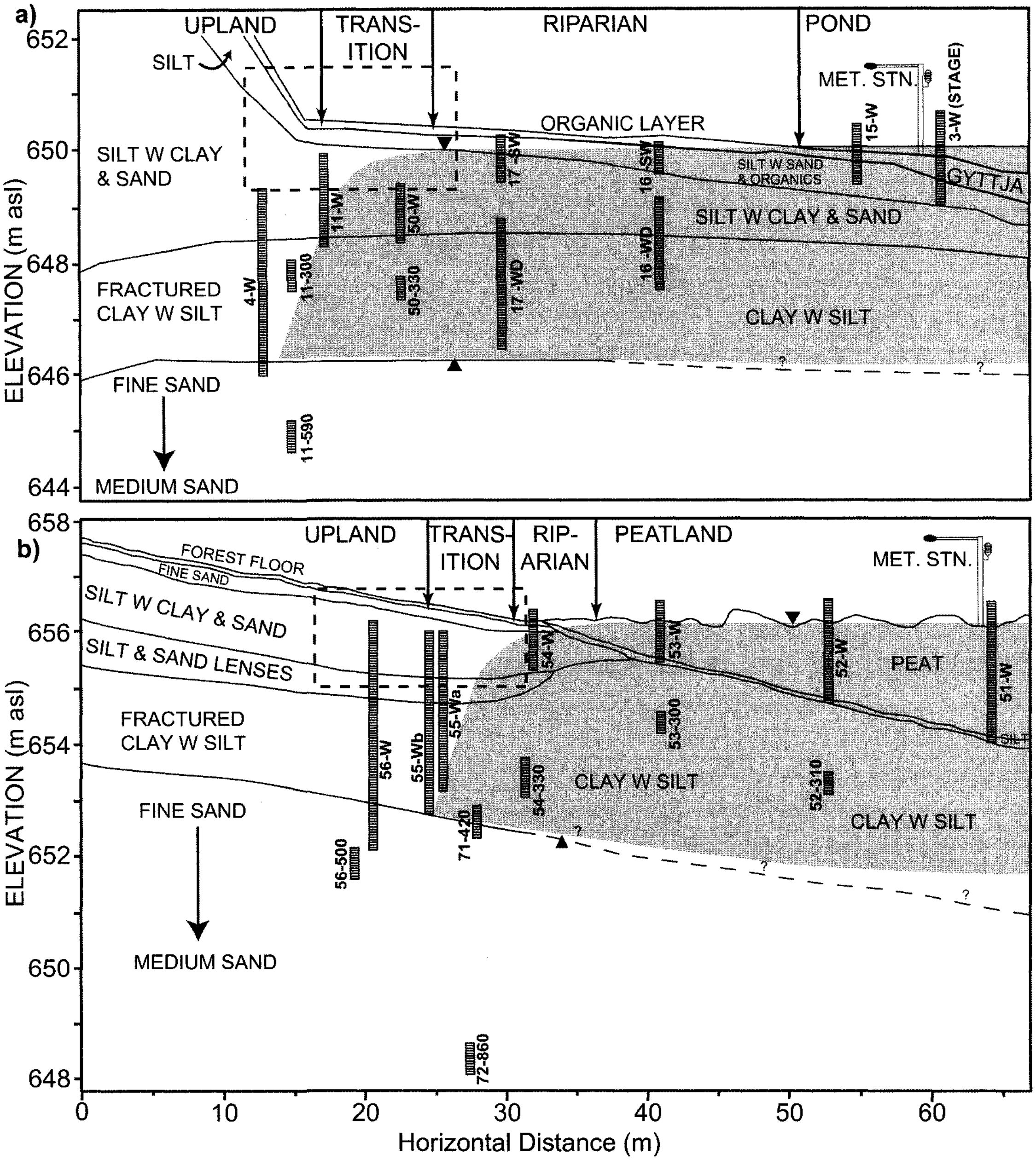

Figure 2.7. Instrumentation transects at a) P19, and b) W1 showing lithology, average perched water table configurations (transparent grey), and piezometer/well location and screen depths. Vegetation zones annotated as riparian, transition, and upland areas. Dashed boxes indicated location of soil moisture transect instrumentation shown in detail on Fig. 2.3.

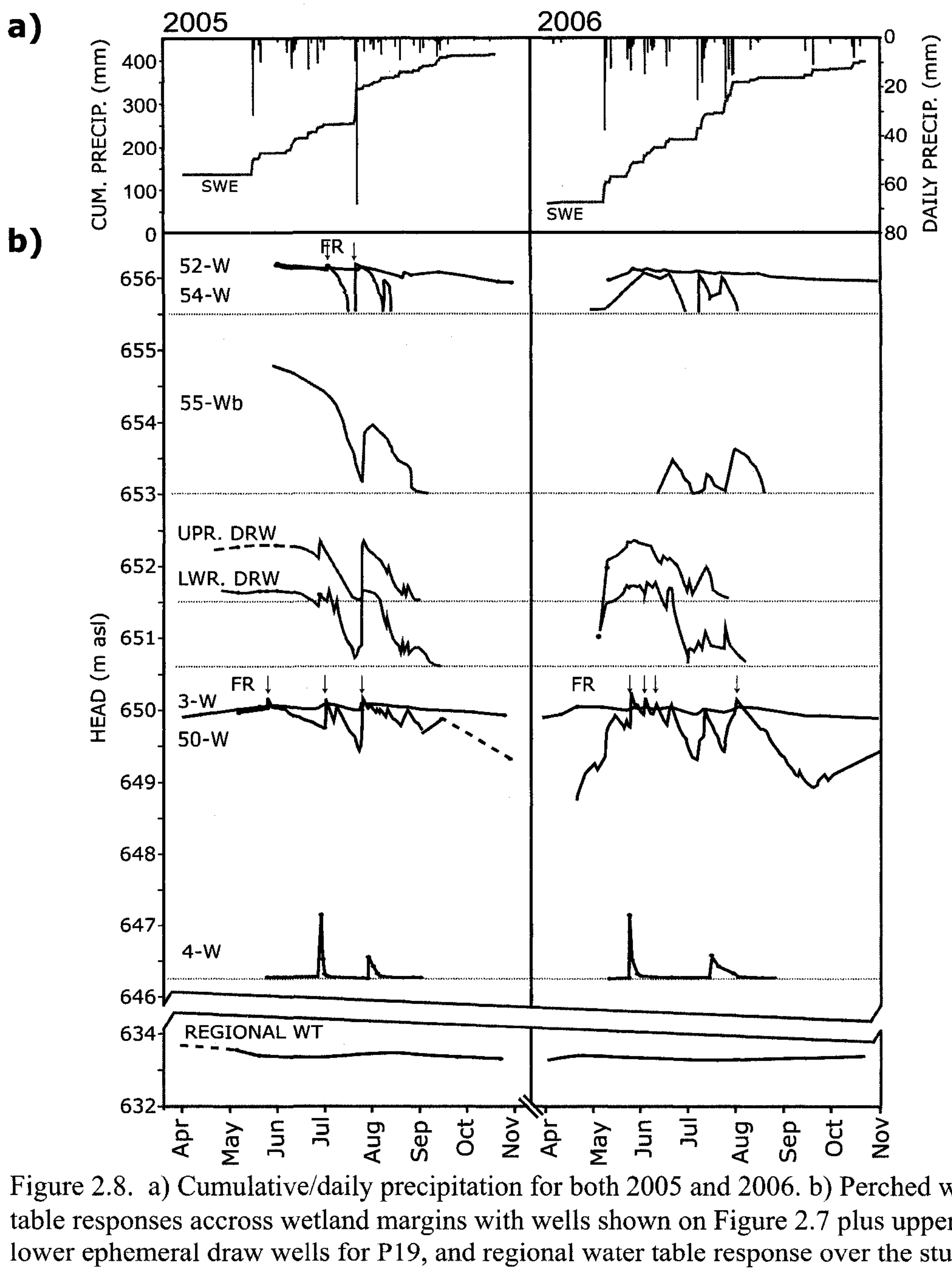

Figure 2.8. a) Cumulative/daily precipitation for both 2005 and 2006. b) Perched water table responses accross wetland margins with wells shown on Figure 2.7 plus upper and lower ephemeral draw wells for P19, and regional water table response over the study period. The faint lines indicate dry wells and flow reversals (FR) or riparian groundwater ridging indicated by arrows.

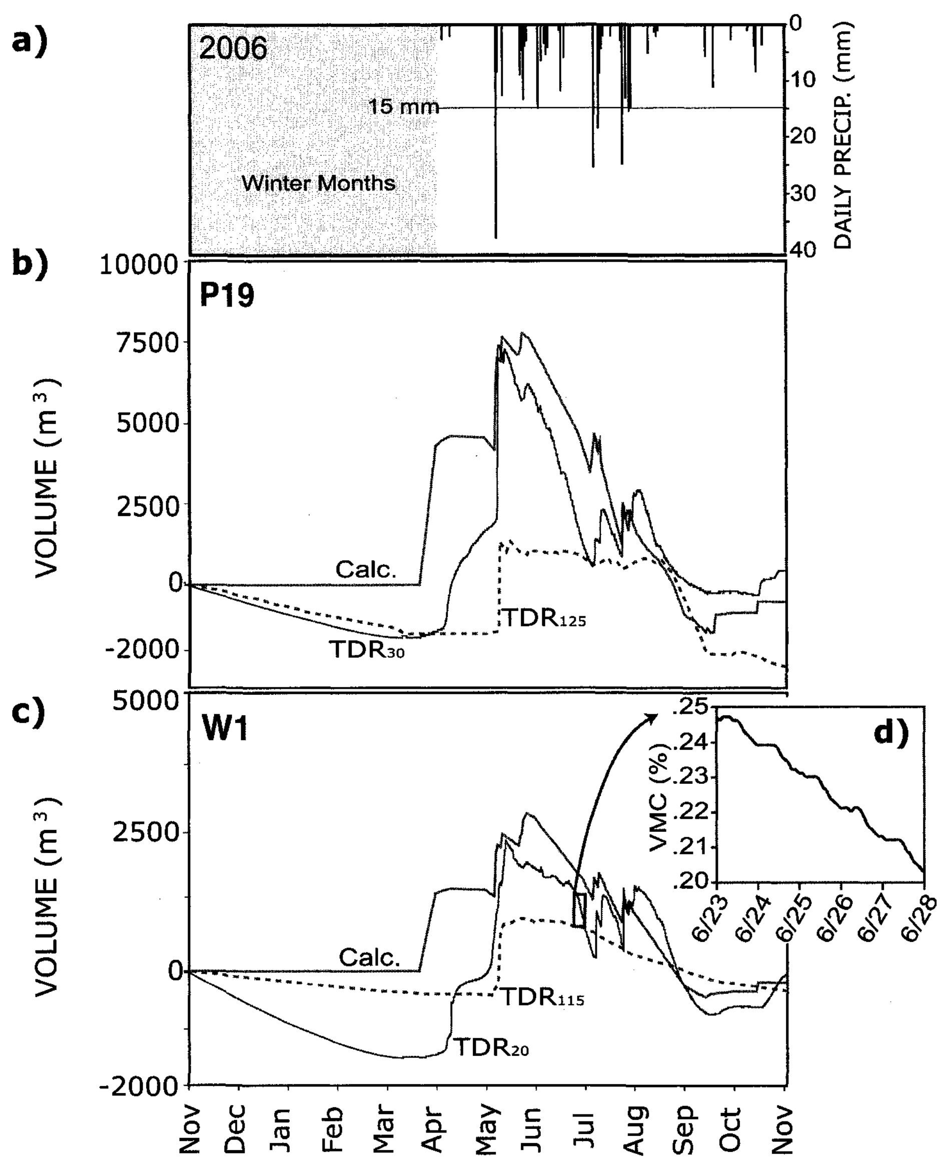

Figure 2.9. a) 2006 precipitation events and the 15 mm threshold. b) Calculated storage change for the rooting zone (Calc.), measured changes within (TDR_{30}), and at the base of the rooting zone (TDR_{125}) for P19 upland. c) Calculated, and measured changes from (TDR_{20}) and (TDR_{115}) in the rooting zone of the W1 upland. Inset d) shows diurnal soil moisture response observed at both wetlands related to aspen root uptake starting midnight of June 23 through June 28, 2006.

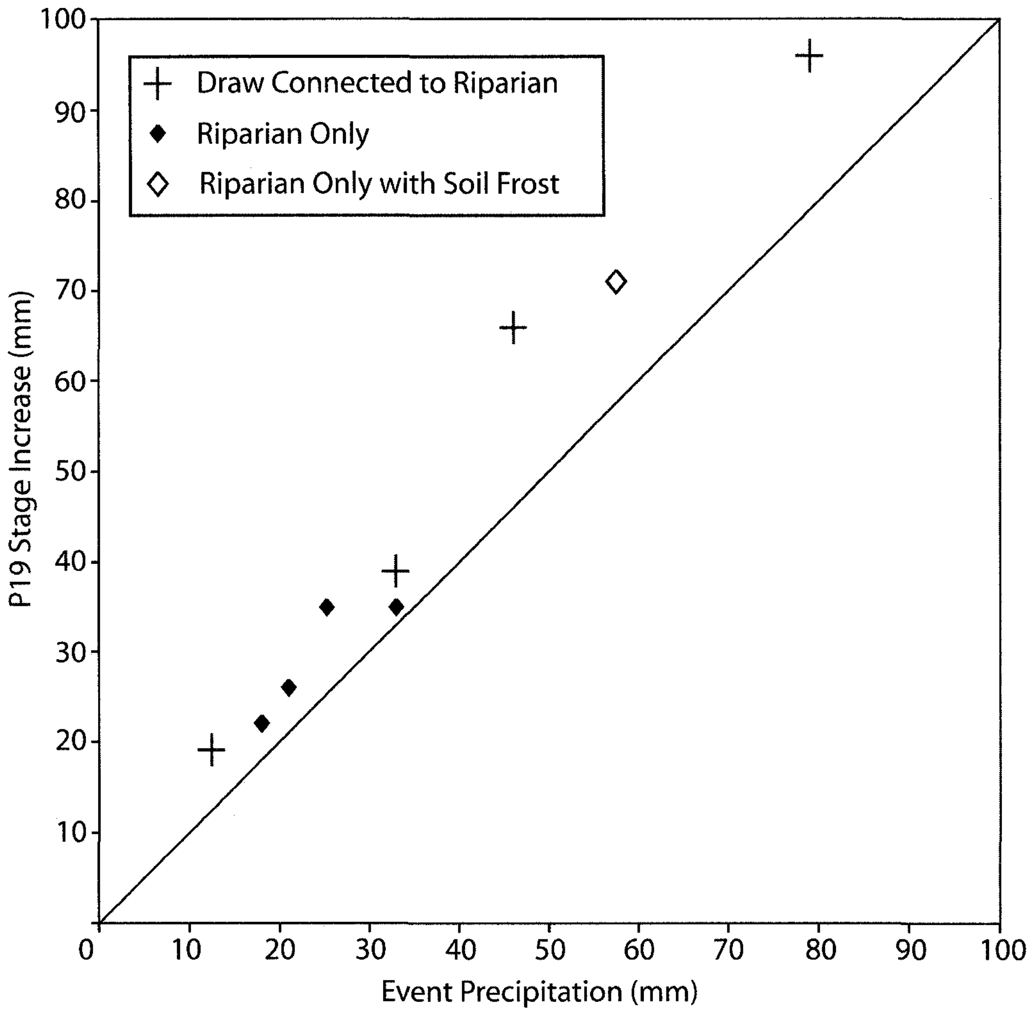

Figure 2.10. Total event precipitation versus P19 stage increases indicating minor runoff contributions from low soil water storage areas after large rain events (>10 mm). Runoff contributions from the riparian when connected to the ephemeral draw (with or without soil frost) are shown by crosses. Runoff contribution from the riparian area only without soil frost are shown by closed symbols. The open symbol shows the runoff contribution from the riparian zone only, during the soil frost period in early spring.

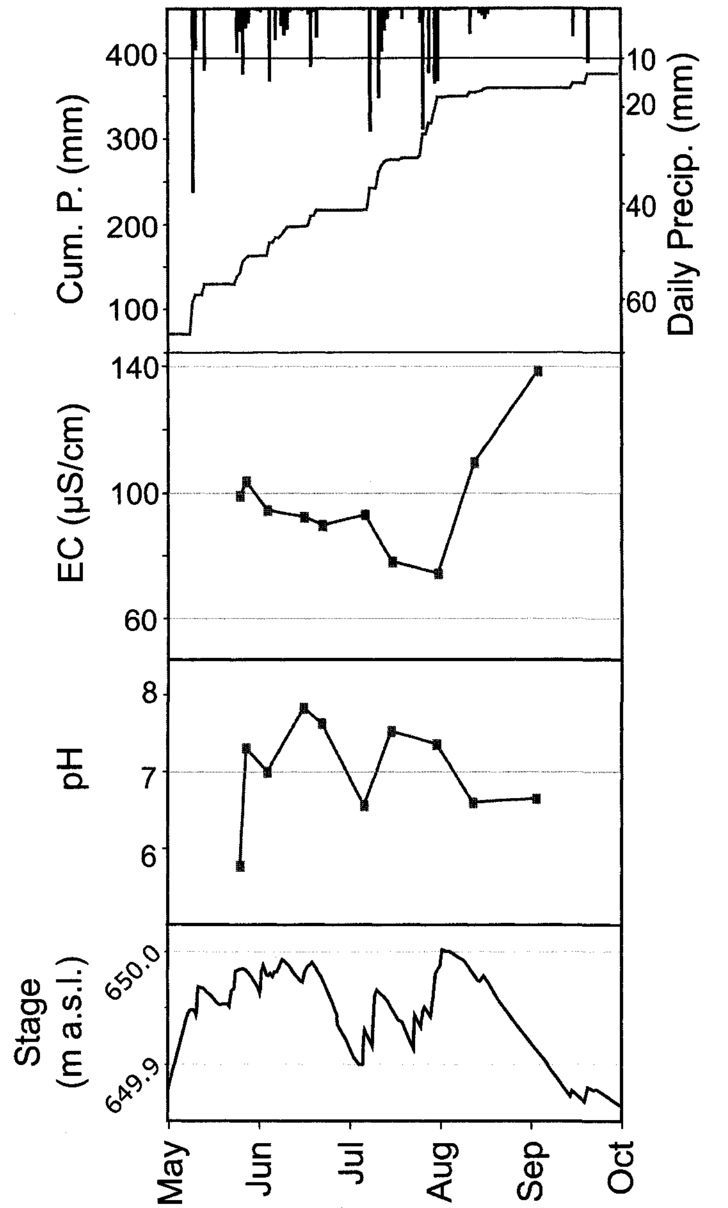

Figure 2.11. Cumulative precipitation, daily precipitation totals, electrical conductivity (EC), pH, and stage of P19 for ice free portion of 2006 hydrologic year.

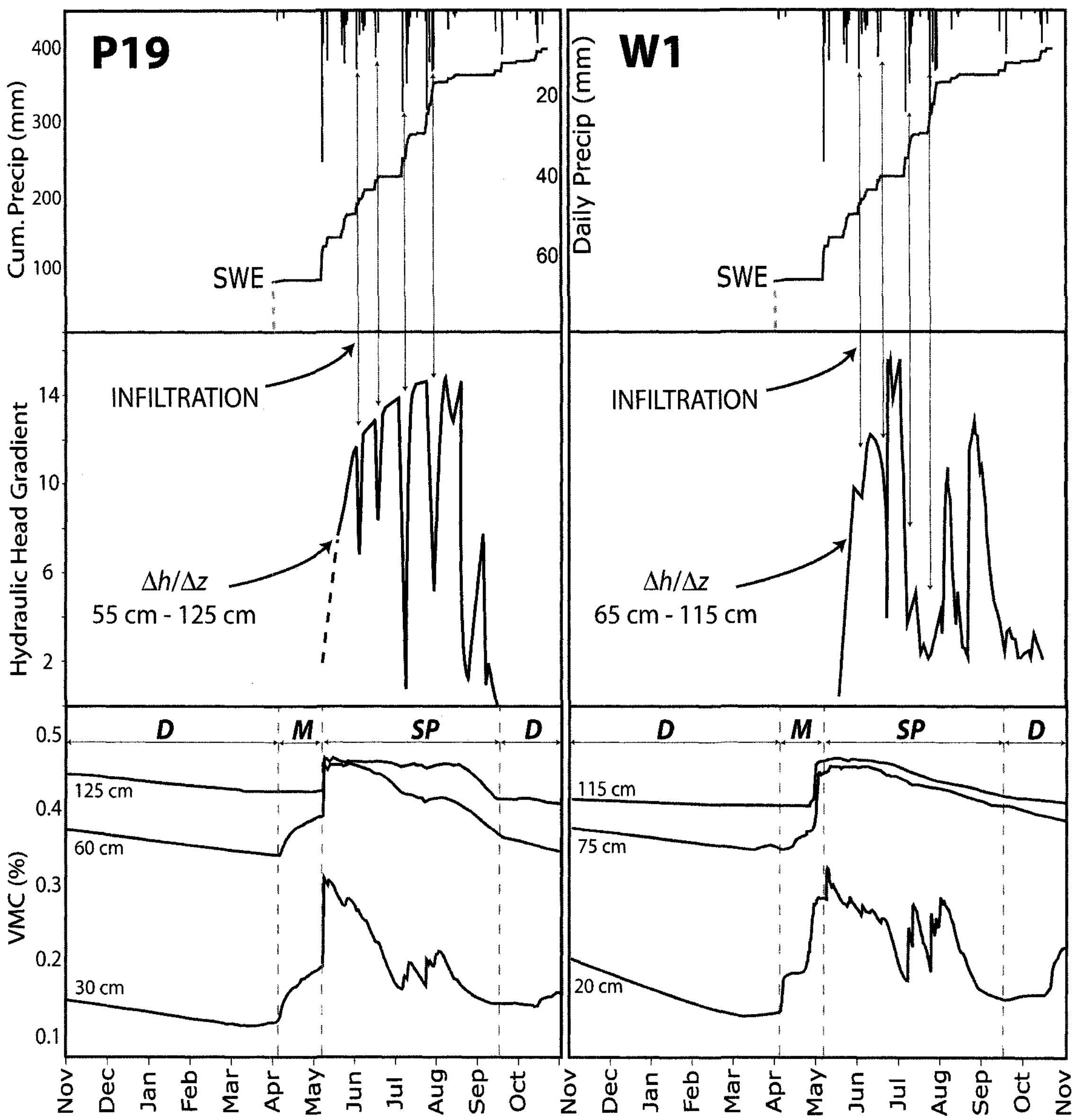

Figure 2.12. 2006 cumulative/daily precipitation, vertical soil tension gradients, and TDR soil moisture responses from N3 at both P19 (left) and W1 (right). Gradients indicate upward flow from the base of the rooting zone to the upper portion where root density is the highest. Reduced vertical tension gradients mark infiltration events from precipitation events. Storage changes due to drainage (***D***), melt and/or storm infiltration (***M***), and summer precipitation/evapotranspiration (***SP***) marked on TDR curves.

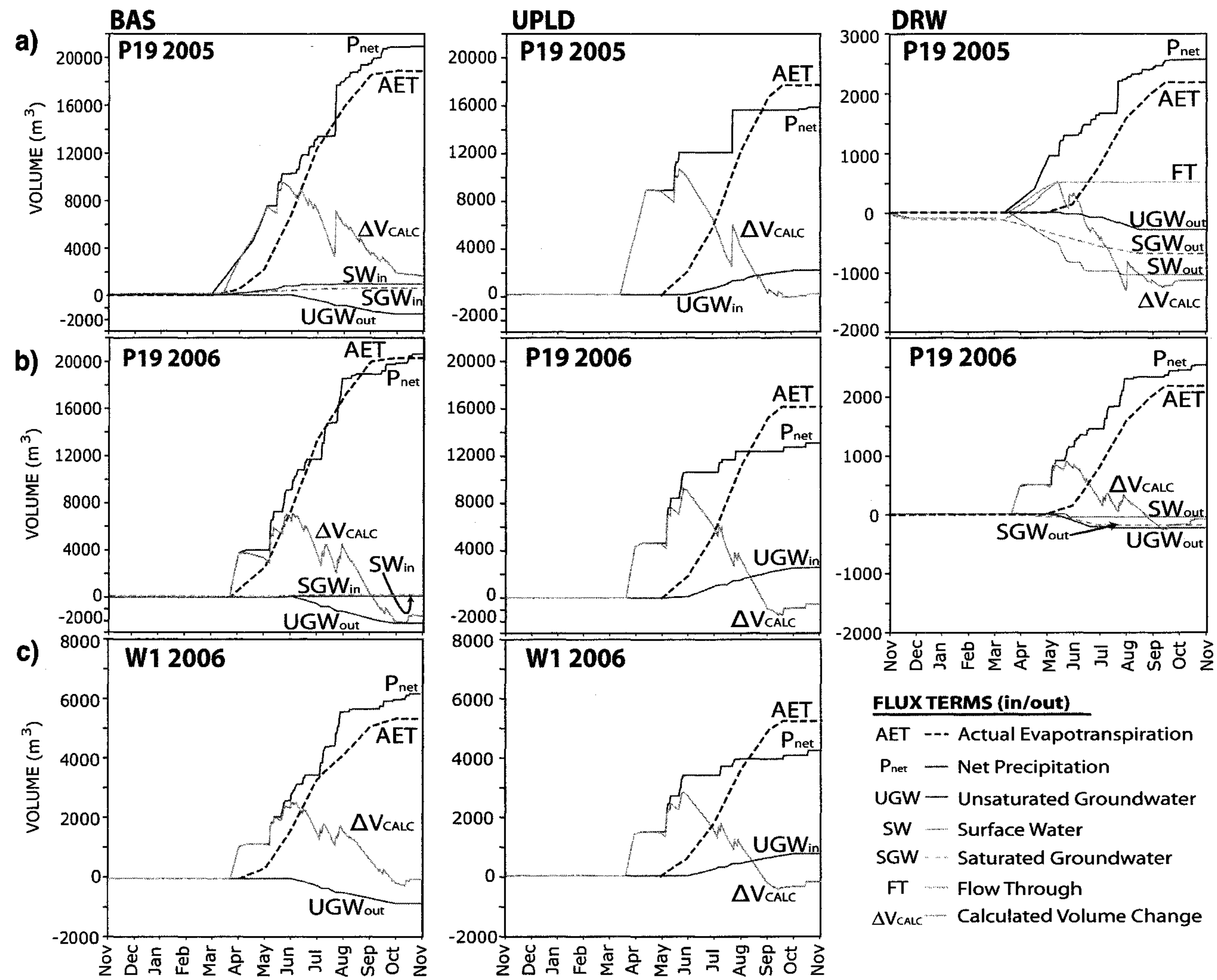

Figure 2.13. Graphs showing volumes lost/gained from each catchment sub-component from flux terms in water budget equation for a) P19 catchment 2005, b) P19 catchemnt 2006, and c) W1 catchment 2006

	P19 2005				P19 2006				W1 2006		
	CMT	BAS	UPLD	DRW	CMT	BAS	UPLD	DRW	CMT	BAS	UPLD
Area (ha)	12.8	5.6	6.5	0.7	12.8	5.6	6.5	0.7	3.8	1.8	2.1
AET	303	340	271	310	301	362	247	310	273	305	247
Pnet	307	377	240	365	283	369	200	360	268	352	200
INT	104	34	171	46	111	25	194	34	126	42	194
UGWout	-	31	0	41	-	41	0	34	-	45	0
UGWin	-	0	31	0	-	0	39	0	-	0	38
SWin	-	16	-	41	-	1	-	0	-	0	-
SWout	-	0	4	127	-	0	0	8	-	0	0
SGWout	-	0	-	63	-	0	-	21	-	-	-
SGWin	-	8	-	-	-	3	0	0	-	-	-
ΔV (obs.)	1	26	1	-192	-15	-29	-4	-12	-1	0	-3
ΔV (calc.)	-4	30	-4	-135	-18	-30	-8	-13	-5	-2	-9
residual (mm)	5	4	3	57	3	1	4	1	4	2	6

Table 2.1. Distribution of intra-catchment flux terms as shown on Figure 2.4 relative to catchment areas (CMT), wetland basin areas (BAS), upland areas (UPLD), and ephemeral draw area of P19 (DRW) for each P19 and W1 catchments.

2.7 *References*

Alberta Environmental Protection, 1998. The boreal forest natural region of Alberta. Natural Resource Services, Recreation and Protection, Special Report.

Amiro, B.D. *et al.*, 2006. Carbon, energy and water fluxes at mature and disturbed forest sites, Saskatchewan, Canada. Agricultural and Forest Meteorology, 136: 237-251.

Arnold, J.G., Muttiah, R.S., Srinivasan, R. and Allen, P.M., 2001. Regional estimation of base flow and groundwater recharge in the Upper Mississippi river basin. 227(1-4): 21-40.

Bothe, R.A. and Abraham, C., 1993. Evaporation and evapotranspiration in Alberta, 1986 - 1992 Addendum, Alberta Environmental Protection, Edmonton, Alberta.

Brooks, R.T. and Hayashi, M., 2002. Depth-area-volume and hydroperiod relationships of ephemeral (vernal) forest pools in southern New England. Wetlands, 22(2): 247-255.

Carey, S.K. and Woo, M.-k., 2001. Slope runoff processes and flow generation in a subarctic, subalpine catchment. Journal of Hydrology, 253(1-4): 110-129.

Carlyle-Moses, D.E., Flores Laureano, J.S. and Price, A.G., 2004. Throughfall and throughfall spatial variability in Madrean oak forest communities of northeastern Mexico. Journal of Hydrology, 297(1-4): 124-135.

Crockford, R.H. and Richardson, D.P., 2000. Partitioning of rainfall into throughfall, stemflow and interception: effect of forest type, ground cover and climate. Hydrological Processes, 14(16-17): 2903-2920.

De Marsily, G., 1986. Quantitative Hydrogeology. 403p.

Devito, K., Creed, I., Gan, T., Mendoza, C., Petrone, R., Silins, U. and Smerdon, B., 2005a. A framework for broad scale classification of hydrologic response units on the Boreal Plains: Is topography the last thing to think of? Hydrological Processes, 19(8): 1705-1714.

Devito, K.J., Creed, I.F. and Fraser, C.J.D., 2005b. Controls on runoff from a partially harvested aspen-forested headwater catchment, Boreal Plain, Canada. Hydrological Processes, 19(1): 3-25.

Dingman, S.L., 2002. Physical Hydrology. Prentice Hall, Upper Saddle River, New Jersey, 646 p.

Ducks Unlimited Canada. 2006. Waterfowl of the Boreal Forest. Ducks Unlimited Canada, Stonewall, Manitoba. 107p.

Elliott, J.A., Toth, B.M., Granger, R.J. and Pomeroy, J.W., 1997. Soil moisture in mature and replanted sub-humid boreal forest stands. Canadian Journal of Soil Science, 78: 17-27.

Environment Canada, 2003. Canadian Climate Normals 1971 - 2000: Slave Lake, Alberta. www.climate.weatheroffice.ec.gc.ca/climate_normals/results_e.html.

Fenton, M.M., Paulen, R.C. and Pawlowicz, J.G., 2003. Surficial geology of the Lubicon Lake area, Alberta (NTS 84B/SW). Alberta Geological Survey.

Ferone, J.M. and Devito, K.J., 2004. Shallow groundwater-surface water interactions in pond-peatland complexes along a Boreal Plains topographic gradient. Journal of Hydrology, 292(1-4): 75-95.

Freeze, R.A. and Cherry, J.A., 1979. Groundwater. Prentice-Hall Inc., Englewood Cliffs, New Jersey, 604 p.

Gorham, E., 1991. Northern peatlands: role in the carbon cycle and probable responses to climatic warming. Ecological Applications(1): 182–195.

Hashino, M., Huaxia, Y. and Yoshida, H., 2002. Studies and evaluations on interception processes during rainfall based on a tank model. Journal of Hydrology, 255(1-4): 1-11.

Hayashi, M., van der Kamp, G. and Rudolph, D., 1998. Water and solute transfer between a prairie wetland and adjacent uplands, 1. Water balance. Journal of Hydrology, 207(1-2): 42-55.

Hunt, R.J., Krabbenhoft, D.P. and Anderson, M.P., 1996. Groundwater inflow in wetland systems. Water Resources Research, 32(3): 495-507.

Hvorslev, J.M., 1951. Time Lag and Soil Permeability in Ground-water Observations. Bulletin No. 36, US Army Corps of Engineers, Vicksburg, Mississippi.

Ingram, H.A., Rycroft, D.W. and Williams, D.J., 1974. Anomalous transmission of water through certain peats. Journal of Hydrology, 22(3-4): 213-218.

Kuhry, P., Nicholson, B.J., Gignac, L.D., Vitt, D.H. and Bayley, S.E., 1993. Development of Sphagnum-dominated peatlands in boreal continental Canada. Canadian Journal of Botany, 71: 10-22.

Lafleur, P.M., McCaughey, J.H., Joiner, D.W., Bartlett, P.A. and Jelinski, D.E., 1997. Seasonal trends in energy, water and carbon dioxide at a northern boreal wetland. Journal of Geophysical Research, 102(24): 29009-29020.

Metcalfe, R.A. and Buttle, J.M., 1999. Semi-distributed water balance dynamics in a small boreal forest basin. Journal of Hydrology, 226(1-2): 66-87.

Meyboom, P., 1966. Unsteady groundwater flow near a willow ring in hummocky moraine. Journal of Hydrology, 4(1): 38-40.

Millar, J.B., 1971. Shoreline-area ratio as a factor in rate of water loss from small sloughs. Journal of Hydrology, 14(3-4): 259-284.

Mills, J.G. and Zwarich, M.A., 1986. Transient groundwater flow surrounding a recharge slough in a till plain. Canadian Journal of Soil Science, 66: 121-134.

National Wetlands Working Group, 1988. Wetlands of Canada, Ecological Land Classification Series No. 24, Ottawa.

O'Geen, A.T., McDaniel, P.A., Boll, J. and Brooks, E., 2003. Hydrologic processes in valley soilscapes of the Palouse Basin in northern Idaho. Soil Science, 168(12): 846-855.

Pawlowicz, J.G. and Fenton, M.M., 2002. Drift thickness of the Peerless Lake map area (NTS 84B). Alberta Geological Survey.

Petrone, R.M., Silins, U., Devito, K.J., 2007. Dynamics of evapotranspiration from a riparian pond complex in the Western Boreal Forest, Alberta, Canada. Hydrological Processes, 21(11): 1391-1401.

Price, A.C., 2005. Evaluation of groundwater flow and salt transport within an undrained tailings sand dam. M.Sc. Thesis, University of Alberta, Edmonton, Alberta, 129p.

Price, A.G., Dunham, K., Carleton, T. and Band, L., 1997. Variability of water fluxes through the black spruce (Picea mariana) canopy and feather moss (Pleurozium schreberi) carpet in the boreal forest of Northern Manitoba. Journal of Hydrology, 196(1-4): 310-323.

Price, J.S., Branfireun, B.A., Waddington, J.M. and Devito, K.J., 2005. Advances in Canadian wetland hydrology, 1999-2003. Hydrological Processes, 19(1): 201-214.

Rains, M.C., Fogg, G.E., Harter, T., Dahlgren, R.A. and Williamson, R.J., 2006. The role of perched aquifers in hydrological connectivity and biogeochemical processes in vernal pool landscapes, Central Valley, California. Hydrological Processes, 20(5): 1157-1175.

Redding, T.E. and Devito, K.J., 2005. Snowmelt infiltration and runoff from forested hillslopes, boreal plain, Alberta. Proceedings of the Annual Eastern Snow Conference, 62: 105-108.

Redding, T.E. and Devito, K.J., 2007. Lateral and vertical flow thresholds for aspen forested hillslopes in the Western Boreal Plain, Alberta, Canada. Hydrological Processes, *in press*.

Redding, T.E., Hannam, K.D., Quideau, S.A. and Devito, K.J., 2005. Particle density of aspen, pine, and spruce forest floors in Alberta, Canada. Soil Science Society of America Journal, 69: 1503-1506.

Rockafeller, P.A., McDaniel, P.A. and Falen, A.L., 2004. Perched water table responses to forest clearing in Northern Idaho. Soil Science Society of America Journal, 68: 168-174.

Rosenberry, D.O. and Winter, T.C., 1997. Dynamics of water-table fluctuations in an upland between two prairie-pothole wetlands in North Dakota. Journal of Hydrology, 191(1-4): 266-289.

Rouse, W.R., 2000. The energy and water balance of high-latitude wetlands: control and extrapolation. Global Change Biology, 6: 59-68.

Schuh, W.M., Meyer, R.F., Sweeney, M.D. and Gardner, J.C., 1993. Spatial variation of root-zone and shallow vadose-zone drainage on a loamy glacial till in a sub-humid climate. Journal of Hydrology, 148(1-4): 1-26.

Smerdon, B.D., Devito, K.J. and Mendoza, C.A., 2005. Interaction of groundwater and shallow lakes on outwash sediments in the sub-humid Boreal Plains of Canada. Journal of Hydrology, 314(1-4): 246-262.

Smerdon, B.D., Mendoza, C.A. and Devito, K.J., 2007. Simulations of fully coupled lake-groundwater exchange in a subhumid climate with an integrated hydrologic model. Water Resources Research, 43(1): W01416, doi:10.1029/2006WR005137.

Tóth, J., 1963. A theoretical analysis of groundwater flow in small drainage basins. Journal of Geophysical Research, 68(16): 4795-4812.

van der Kamp, G. and Hayashi, M., 1997. The groundwater recharge function of small wetlands in the semi-arid northern prairies. Great Plains Research, 8: 39-56.

Winter, T.C. and Rosenberry, D.O., 1995. The interaction of groundwater with prairie pothole wetlands in the Cottonwood Lake Area, east-central North Dakota, 1979-1990. Wetlands, 15: 193-211.

Wolfe, B.B., Hall, R.I., Last, W.M., Edwards, T.W.D., English, M.C., Karst-Riddoch, T.L., Paterson, A. and Palmini, R., 2006. Reconstruction of multi-century flood histories from oxbow lake sediments, Peace-Athabasca Delta, Canada. Hydrological Processes, 20(19): 4131-4153.

Woo, M.-K. and Rowsell, R.D., 1993. Hydrology of a prairie slough. Journal of Hydrology, 146(1-4): 175-207.

Woo, M.-K. and Winter, T.C., 1993. The role of permafrost and seasonal frost in the hydrology of northern wetlands in North America. Journal of Hydrology, 141(1-4): 5-31.

Chapter 3

Testing a conceptual model for perched boreal wetland hydrology using fully coupled numerical simulations

3.1 *Introduction*

Canada's Boreal ecosystems have been traditionally perceived as isolated from anthropogenic impacts and development (Bryant *et al.*, 1997). However, recent funding, conservation, and research efforts have begun to reflect the rate of economic development within this sensitive ecosystem (Ducks Unlimited Canada, 2006). The Boreal Plains ecozone has been subject to rapid expansion of infrastructure associated with conventional oil and gas, forestry, and oil sands industries (Alberta Environmental Protection, 1998).

Ongoing research in the Western Boreal Plain (WBP) at the Utikuma Region Study Area (URSA) has shown that the landscape of wetlands and forested uplands has complex hydrology controlled by climate, geology, and surface water-groundwater interaction (Ferone and Devito, 2004; Devito *et al.*, 2005a) as observed on other landscapes (York *et al.*, 2002). The highly heterogeneous, low relief landscape of the Boreal Plains varies from fine, to coarse textured porous media associated with widespread glaciolacustrine, moraine, and glaciofluvial outwash deposits (Pawlowicz and Fenton, 2002). The region's complex geology, ubiquitous lakes and ponds, diverse vegetal patterns, and temporal climate variations result in large spatial variation that presents challenges for adequate assessment of hydrological impacts resulting from industrial land surface disturbances.

Models that are used to evaluate impacts must capture the natural variability of boreal wetlands driven by climate, soil storage, and surface water-groundwater interaction. As such, it is important that they honour physical processes that are spatially and temporally variable in the WBP (Devito *et al.*, 2005b; Smerdon *et al.*, 2005; Chapter 2). Furthermore, it is desirable that fully coupled surface water-groundwater models quantify interaction between near surface flow systems and wetlands, and regional groundwater flow systems based on estimated atmospheric fluxes and geology (Smerdon *et al.*, 2007). The use of such models is required to quantify water fluxes associated with permanent and ephemeral perched wetlands, which are important water resources in the region (Wolfe *et al.*, 2006; Chapter 2).

Previous efforts to model perched systems have been application specific, largely pertaining to the impact of perched groundwater on long-term nuclear waste repository evaluations (Hinds *et al.*, 1999; Orr, 1999). These simulations could ignore surficial hydrologic processes, and used generalized, steady-state climatic input functions. In these studies, prescribed, steady state flux boundary conditions were used to evaluate unsaturated flow regimes of deep perched aquifers separated from surface, under long-term climate change scenarios (Hinds *et al.*, 1999).

The evaluation of perched boreal wetland sustainability presents similar challenges to accurately characterize and model, as nearby study sites (Ferone and Devito, 2004; Smerdon *et al.*, 2005; Smerdon *et al.*, 2007). The complex hydrology requires the use of a fully coupled hydrologic model that will enable the determination of the location and extent of surface water features and to quantify atmospheric flux/surface water-groundwater relationships, and changes in soil water storage. Unlike the glacial outwash deposits to the west of the study site studied by Smerdon *et al.* (2007), there is no overall regional groundwater flow field sustaining wetland water levels. Thus, relative to the nearby outwash wetlands, perched wetlands may be even more sensitive to temporal climate and soil storage variation, and localized hydrologic processes.

The objective of this Chapter was to test the conceptual model and water budget analysis at URSA Pond 19 (P19) and to further elucidate the role of different water cycling processes or flux components through the use of a fully coupled hydrologic model. In Chapter 2, the results of rigorous hydrologic data collection and water budgets for two hydrologic years at P19 were presented. Potential evapotranspiration, surface water fluxes, saturated/unsaturated groundwater flow regimes, and soil temperature data were shown. Analysis of these data allowed for comparison of numerical simulation results to the temporal and spatial distribution, and the magnitude of hydrologic processes observed in the field. Key interpretations of the hydrologic processes sustaining perched boreal wetlands, such as localized runoff generation, vegetative water table depression, and riparian zone flow reversals were also further investigated through the use of numerical simulations. Calibration of the hydrologic model was also intended to further elucidate the control of climate and soil storage on hydrologic processes, and provide a basis to evaluate overall pond sustainability, and impact susceptibility. This modeling also assisted in refining the Hydrologic Response Unit (HRU; Devito *et al.*, 2005a) characteristics developed from field studies, and the direction/magnitude of hydrologic connections of perched boreal wetlands to adjacent landscape features. Atmospheric interactions, hydraulic connections to large-scale groundwater flow systems, and the biogeochemical function of perched boreal wetlands can then be further evaluated through future scenario modeling.

3.2 Study area

The Utikuma Region Study Area (URSA) is located in the Boreal Plain region of the Western Boreal Forest in north-central Alberta, Canada. The URSA encompasses a variety of landforms common within the Boreal plains ecoregion (Devito *et al.*, 2005a). Salient characteristics include a sub-humid climate (Ecoregions Working Group, 1989), deep (20 – 240 m) heterogeneous glacial sediments (Pawlowicz and Fenton, 2002), and low topographic relief. The climate is continental, with average monthly temperatures ranging between -14.6° C and 15.6° with winter periods lasting approximately from November to April (Environment Canada, 2003). Annual precipitation is highly variable, ranging from 318 to 529 mm from 1997 to 2001 (Devito *et al.*, 2005b), with long-term average annual precipitation (*P*) of 481 mm (Enviroment Canada, 2003). Average annual potential evapotranspiration (*PET*) of 518 mm creates frequent deficit conditions. However, wet conditions and significant runoff generation occur every 10 to 15 years due to greater than average precipitation and cool summer temperatures (Devito *et al.*, 2005b).

The P19 study site is situated in a geologic transition zone between glaciofluvial outwash deposits and stagnation moraine deposits. Fine-grained moraine sediments with a laterally extensive clay-rich confining layer overlie 10 to 20 m of sand and gravel. The regional water table is >20 m from surface in the sand and gravel deposits (Chapter 2). The permeability contrast between the outwash and stagnation moraine deposits, and the deep regional water table create a hydrostratigraphic framework where laterally discontinuous perched wetlands and flow systems develop. As such, P19, and numerous other perched wetlands at the study site are separated from the regional water table by a 15 m thick unsaturated zone (Chapter 2).

Radial instrumentation transects, extending from P19, through the riparian zone, and into the surrounding upland of the closed basin, were monitored. Through field observations, a conceptual model of surface water-groundwater interaction, saturated flow, and unsaturated flow regimes between the pond/riparian zones and unsaturated upland areas was established (Fig. 3.2; Chapter 2). The water budget was dominated by precipitation (rain/snow) and evaporative fluxes. However, to balance the pond catchment water budget, spatially limited, but critical lateral water fluxes required consideration. Unsaturated groundwater flow from wetland margins to the upland was an important water cycling mechanism within the perched pond catchment, similar to prairie potholes (Meyboom, 1966; Mills and Zwarich, 1986; Hayashi *et al.*, 1998; Chapter 2). Groundwater ridging and corresponding saturated groundwater flow reversals, and saturated overland flow generation occurred after precipitation events of 10 to 15 mm in the riparian zone, which appear to be critical processes for pond permanence and sustainability. Lateral flow inputs from outside the pond basin were spatially limited to an ephemeral draw area that only drained into the pond/riparian areas during times of water surplus, such as spring melt and large summer storms.

Small annual water storage changes were observed in the pond basin over the study period as the wetland basin (i.e., the pond and riparian zone) gained 26 mm in 2005, and lost 29 mm in 2006. Aerial photograph analysis shows the current measured pond area of approximately 1.7 ha is close to the average pond surface area over the last 50 years. However, the pond has been as large as 5.6 ha after several years of water surplus, and as small as 1.4 ha after extended drought conditions, indicating P19's sensitivity to climate when compared to historic climate data from the nearby town of Slave Lake, Alberta.

Borehole data show that highly heterogeneous surface sediments overlying the clay confining layer range in thickness from 1 to 12 m, creating spatially variable water storage potential (Chapter 2). Mapping of the confining layer indicates that wetlands that are permanently or ephemerally saturated develop where low permeability clay is near surface (<2 m), whereas unsaturated upland areas have significant storage potential because the clay is deeper (>2 m; Chapter 2). Highly permeable and porous aspen forest floor materials present in upland areas are 0.1 to 0.2 m thick. A similar loose organic soil is found in the riparian zone. The pond bottom is composed of a silt-rich gyttja having an average thickness of 0.3 m. Underlying the organic soils, a mix of poorly sorted till material is present, becoming finer in texture and less permeable with depth including silty sand, silt loam, and silty clay. The confining layer is approximately 2 m thick, and is underlain by a thin gradational layer separating it from the underlying sand deposits. Beneath the perched groundwater lens, and extending to the edge of the riparian zone, the confining layer is composed of highly plastic, unfractured blue-grey clay. Beneath the upland areas, the confining layer is composed of unsaturated, light grey clay with oxidized desiccation fractures, and macropores from deep aspen roots.

3.3 Numerical simulations

Transient surface water-groundwater and atmospheric exchanges were simulated for a 100 m long, 20 m high, unit width domain representing a two-dimensional cross section of the main instrumentation transect at P19 (Fig. 3.3). Each hydrologic year (2005 and 2006) was simulated as a separate timeframe with unique initial conditions. This was done because the permeability field of near surface soils was expected to be transient due to concrete soil frost development, and temporally sparse field data in the winter months (November through March) were expected to make calibration difficult. The modeled timeframes were concurrent with high temporal resolution field data and water budget analyses completed in Chapter 2. Simulations began on March 1 and ran for 245 days until October 31 of each year to capture spring melt and the unfrozen portion of the hydrologic year in northern plains regions similar to Hayashi *et al.* (1998) and Smerdon *et al.* (2005).

Boundary conditions representing atmospheric fluxes were based on measured daily precipitation values and estimated evaporative fluxes. Spatial distributions of porous media and their estimated hydraulic conductivities were based on grain size analysis of borehole samples, and in-situ measurements (Chapter 2). The highly heterogeneous sediments within the riparian zone were generalized into four homogeneous porous media zones for the model, requiring calibration of their hydraulic parameters. Hydraulic conductivity, anisotropy ratios, and porosity were adjusted in the calibration process relative to those estimated from in-situ field measurements and grain size analyses of the porous media samples from this portion of the domain. The hydraulic parameters for the model were calibrated to hydrologic data from 2005. This was achieved by comparing simulation results to field data for the water table configuration, and both the saturated and unsaturated flow regimes. No further calibration was completed on the 2006 transient simulation. The 2006 data were used for validation.

3.3.1 P19 flow model

HydroGeoSphere (HGS; Therrien *et al.*, 2007) was used for the transient flow simulations of P19 for two hydrologic years solving fully coupled equations for flow in variably saturated porous media, and overland flow. The model is based on FRAC3DVS code (Sudicky *et al.*, 2005; Therrien *et al.*, 2007), which solves for flow, and solute transport in dual continuum, variably saturated porous media using a modified form of the Richard's equation. Fractures, secondary porous continuum, and solute transport were not considered in this study. Thus, the flow equation is simplified to conserve groundwater mass, which is given by;

$$(Eq.1) \quad -\nabla q + \Gamma_o \pm Q = \frac{\partial \phi S_w}{\partial t}$$

where q is the Darcy flux, Γ_o is the flux from or into the subsurface from the surface domain, Q is a specified boundary condition (domain sources and sinks), t is time, ϕ is porosity, and S_w is water saturation of the porous media. A dual node approach is used for the nodes on the surface of the domain, whereby the model couples surface flow equations with the subsurface flow equations via a leakage or emission term across a surficial skin layer (Γ_o). Overland flow is approximated by the Saint Venant equation for two-dimensional diffusive waves (see Therrien *et al.* (2007)), which is coupled to the subsurface domain via Γ_o when the dual node approach is used.

The model domain was composed of a finite-element mesh with 11 200 triangular prism elements with uniform spatial increments of 1 m in the horizontal direction, and variable spatial increments in the vertical direction. The mesh had 56 element layers ranging in thickness from 0.05 to 0.50 m. Finer vertical discretization was required at the top of the domain to honour important fine-scale heterogeneities, and to resolve complex unsaturated flow parameters and behaviour resulting from transient atmospheric fluxes and water table responses. The surface topography of the domain was based on bathymetric and level surveys with 1 m lateral resolution from the pond centre to the top of the surrounding upland (Fig. 3.3; Chapter 2). Two small micro-topographic (<0.5 m deep) features were not represented in the riparian area of the model domain, as field data show they had little effect on the water table configuration. The riparian slope was smoothed using linear interpolation from elevation data adjacent to each depression, reducing numerical stress from localized ponding in these small depressions, which aided in model convergence.

The heterogeneous sediments were simplified to a cross sectional geologic model containing 9 porous media zones (Fig. 3.3). Eight of the porous media zones were assigned hydraulic parameters modified from Carsel and Parish (1988) to approximate observed hydraulic conductivities and texture, with the exception of the forest floor material properties, which are modified from Samran *et al.* (1995) and Redding *et al.* (2005). Porous media properties including porosity, residual saturation, hydraulic conductivity, and anisotropy ratios are tabulated with capillary pressure and relative permeability versus saturation curves (Fig. 3.4).

Elements in the top layer in the upland and riparian zone were specified as a 0.05 to 0.20 m thick forest floor porous media zone to represent loose highly porous organic soil present at the study site. These high porosity soils required accurate spatial representation due to their large water storage potential, high hydraulic conductivity, and high infiltration rates (Redding *et al.*, 2005). Three zones underlie the forest floor, with porous media ranging from high permeability loamy sand, to less permeable silt loam and clayey silt with increasing depth from surface. The loamy sand and silt loam lithologies (zones 2 and 4; Fig. 3.3b) were assigned anisotropy ratios with larger horizontal hydraulic conductivities due to field observations indicating fewer roots and macropores.

The confining layer beneath the riparian and pond area (zone 6; Fig. 3.3b) was specified as low-permeability clay to represent the plastic, blue-grey clay lithology observed in the field. Beneath the upland area, the confining layer was assigned a slightly higher permeability (Fig. 3.4), based on field observations of fractures caused by desiccation/re-wetting cycles from pond storage oscillations and macropores from deep aspen root development (zone 8; Fig. 3.3b; Chapter 2).

The bottom portion of the domain represents the sandy deposit (zone 1; Fig. 3.3b) that hosts the deep regional water table. Thin layers of progressively coarser grained materials separate the bottom of the confining layer (zones 6 and 8) from the sand (zone 1), as observed from borehole data (inset; Fig. 3.3b). A laterally extensive till surface underlying the sand deposit is not represented in the model domain. The sand/till interface is approximately 10 m below the bottom of the model domain (Chapter 2).

3.3.2 Boundary conditions

No flow boundaries were used at both ends of the domain: within the upland area ($x = 0$), and in the "centre" of the pond ($x = 100$), to represent symmetry boundaries. In the case of the upland, the symmetry boundary condition represents a topographic divide for potential runoff processes, and a stagnation point for unsaturated flow from the margins towards upland crests between adjacent permanent and ephemeral perched wetlands (Chapter 2). The regional water table was represented by a static specified head boundary condition. It was assumed that the regional water table was hydrologically uncoupled from the perched water table for the duration as it only varied 0.1 m over the duration of the study, with no precipitation event responses or seasonal patterns (Chapter 2).

Analyses of field data show that atmospheric fluxes dominate the fluid movement at perched wetlands, which vary between the upland, riparian, and pond areas because of differences in *AET* and interception (Fig. 3.2; Chapter 2). Field results also show that the perched flow system responded rapidly (i.e., < 1 day) to measured precipitation events. As such, a net atmospheric flux was specified for each day of the simulations to each of the pond, riparian zone, and upland areas. These daily fluxes represented the net atmospheric flux estimates for each day of the study period made from field measurements of precipitation, potential evapotranspiration (*PET*), and lateral flux from the ephemeral draw.

Field measurements of total precipitation were made from a tipping bucket (Jarek model 4025). *PET* was measured using a Class A evaporation pan, and on-site meteorological data. These measurements and published *AET* estimates for similar landforms in boreal regions were used to bound AET estimates (Lafleur *et al.*, 1997; Rouse, 2000; Eaton and Rouse, 2001; Amiro *et al.*, 2006; Chapter 2). The lateral flow from the ephemeral draw (Chapter 2) needed to be considered as an additional source of water external to the model domain. Lateral flux inputs from the ephemeral draw into the basin determined from field measurements were added to the net specified flux applied to the pond and riparian areas.

Separate, daily time series representing the sum of positive fluxes and negative evaporative fluxes acting on each of the pond, riparian zone, and upland areas were generated. Positive fluxes included snow melt, P_{net} (or total precipitation minus interception), and lateral fluxes from the ephemeral draw. Evaporation directly from canopy foliage was removed by subtracting intercepted precipitation from the AET, such that AET_{RZ} represented evaporative flux from the rooting zone in the porous media and/or open water. The atmospheric flux regime applied to each area within the domain (i.e., pond, riparian, and upland) for 2005 and 2006 are shown in Figure 3.5.

The flux regime for the pond area was based on daily estimates of evaporative fluxes using a partially submerged evaporation pan, which allowed reasonable approximation of AET over the pond using a pan coefficient of 0.9 (Chapter 2). Snow melt timing was based on the air temperature from on-site meteorological sensors, snow survey data, and when the ice on the pond had completely melted. Lateral flux in the form of surface water and saturated groundwater inputs from the ephemeral draw were incorporated into the flux regime (Fig. 3.5b), based on field hydrometric measurements (Chapter 2). The net atmospheric flux over the pond area was applied to the top of the domain as a single time series as a positive or negative flux, depending on observed atmospheric patterns.

Synthesis of the atmospheric fluxes acting on the riparian zone was more complex because interception and reduced evapotranspiration from the rooting zone or porous media required consideration. Total AET_{RZ} from the riparian zone was estimated to be 2 to 4 mm/d from mid May to mid July, based on compiled boreal fen evaporation data (Lafleur *et al.*, 1997; Rouse, 2000, Eaton and Rouse, 2001), and small diurnal riparian water table responses. An interception threshold of 1 mm for each rain event, as developed from field data (Chapter 2), was removed from total daily precipitation and

AET. Resultant P_{net} and AET_{RZ}, and lateral flux inputs from the ephemeral draw were applied to the riparian zone (Fig. 3.5c). Unsaturated hydraulic head and snow survey data indicated that the riparian zone snow melt began approximately 15 days before the upland area (Fig. 3.5). The net atmospheric flux series for the riparian zone was applied to the top nodes of the model domain in the riparian zone because of the shallow rooting zone and water table position determined from soil pitting and borehole drilling.

The atmospheric fluxes applied to the upland area consisted of snow melt, P_{net}, and AET_{RZ} (Fig. 3.5). An interception threshold of 5 mm was used in aspen upland areas from leaf-out in mid May, to the end of September to account for interception loss (Devito, *unpublished data*). AET_{RZ} for the upland area is based on published boreal aspen *AET* rates (Amiro *et al.*, 2006), minus the precipitation intercepted by the canopy (Chapter 2). The net atmospheric flux time series for the upland area was broken into two time series representing positive and negative net atmospheric fluxes. The positive flux series was applied to the surficial nodes on the upland area to honour important vadose zone storage within the highly porous forest floor material. Negative fluxes from the aspen rooting zone (AET_{RZ}; Chapter 2), were assigned to nodes approximately 0.3 m from surface in the upland area, based on field observations of rooting zone characteristics. Soil moisture data also supported the interpretation that upland evaporative fluxes were drawn from the mineral soil interface between sandy loam (zone 3; Fig. 3.3) and silt loam (zone 4; Fig. 3.3; Chapter 2).

3.3.3 Initial conditions

Each node in the model domain requires a starting value for hydraulic head and saturation ratio, with surface nodes in the domain also requiring an initial value for surface water depth. Unlike other modeling studies of deep, perched systems (Hinds *et al.*, 1999; Orr, 1999), the initial conditions for this model could not be achieved in a steady state simulation. Inherent to the conceptual model, the initial conditions in the domain are a snapshot of a highly transient system, requiring that the simulation replicate the balance of atmospheric fluxes that can generate and maintain the perched water table configuration.

To achieve initial conditions, simulations were designed to replicate the hydrologic environment in which P19 developed. The P19 study site has surficial sediments that developed in a peri-glacial environment (Fenton *et al.*, 2003; Petrone *et al.*, 2005), assumed to have shallow regional groundwater levels. The pond's current perched or "recharge" position >15 m above the regional water table, indicates that regional groundwater levels have receded since surface sediment deposition. As such, the simulations were designed to start with saturated conditions in the majority of the domain, and subsequently undergo drainage with concurrent positive atmospheric fluxes to generate a perched groundwater lens.

Long-term simulations (30 000 d) were utilized to lower regional groundwater levels from surface to their current levels at the base of the domain (Fig. 3.3), through the use of a specified head function representing a receding water table. Concurrent 2nd type boundary conditions representing positive atmospheric fluxes were applied to the top of the domain as the regional groundwater level receded to create a perched groundwater lens above the low-permeability confining layer.

Interpreting field measurements to quantify pre-melt conditions was difficult due to the presence of soil frost, frozen wells/piezometers, and redistribution of water during winter months due to unsaturated flow. Pond elevation was based on the water level taken through a hole cut in the ice adjacent to the staff gage, and depths to ice in frozen riparian zone wells were used as a gross estimate of riparian zone hydraulic head distributions for both years.

3.3.4 Model calibration and evaluation

The saturated flow regime predicted by the model was evaluated at three key monitoring locations. These monitoring locations correspond to shallow groundwater wells in the field. Wells with high temporal resolution hydraulic head data for the pond (3W), mid-riparian zone (16-SW), and the upland toe area (50-W; Fig. 3.6) were chosen. This monitoring point distribution allowed the model to be evaluated on how well it predicted pond stage, the perched water table configuration, and groundwater ridging in the riparian zone.

The unsaturated flow regime predicted by the model was also evaluated at three monitoring points. These monitoring points correspond to nested tensiometers and volumetric water content probes installed at the riparian to upland transition (PN2 – PN4; Fig. 3.6), near the saturated/unsaturated interface as described in Chapter 2. These monitoring points allowed comparison of predicted and observed vertical and lateral unsaturated hydraulic head gradients and volumetric soil water storage. Saturation profiles of the domain were also used to determine if the predicted perched/inverted water table position corresponded to field observations.

3.4 Results

3.4.1 Initial conditions

The hydraulic head distribution was difficult to measure in the winter months preceding the spring melt. This required estimation of "actual" or unfrozen initial conditions at the start of the transient simulations on April 1 of 2005 and 2006. Initial conditions were estimated for each year using fall (i.e., November 2004 and 2005) pond and riparian head data, as well as pre-melt pond stage, and ice elevation data in riparian zone wells. Saturated hydraulic head output from the initial condition simulations were compared to hydraulic head data at the key observation points, including those outlined in the previous section.

The modeled initial conditions had very similar water table configurations and saturation profiles relative to field observations in both 2005 and 2006. The initial conditions for 2006 had a slightly higher pond level and higher hydraulic heads in the riparian zone. Because each year was modeled as a separate timeframe, the 2005 initial condition simulation was used as a template to generate the 2006 initial conditions. To replicate the slightly wetter initial conditions for 2006, the additional volume of water stored in the basin relative to 2005, was added over the last 5 days of the long-term simulation.

Comparison of initial conditions generated by the model showed that the initial saturated hydraulic heads at all three observation points were within 0.05 m of measured head and/or ice elevations at the three key monitoring locations for both 2005 and 2006. Given the estimation of "actual" initial conditions, the simulated initial conditions provided a reasonable starting point for the transient simulations. The simulations also reasonably represented the position of the inverted water table, with saturated conditions ending near the base of the confining layer. The inflection point at the perched/inverted water table transition was located close to the base of the upland slope, within the confining layer, as was observed in the field (Chapter 2).

3.4.2 *Saturated flow regime*

Daily simulation output from the transient simulations for hydraulic head was compared to continuous field data where available, as well as interpolated time series made from manual water level measurements (Fig. 3.6).

The modeled response to snow melt produced lower than observed water levels for pond stage from March 1 to just prior to June in both 2005 and 2006 when soil frost was present in the riparian zone. After June 1 of both 2005 and 2006, pond stage was reasonably simulated by the model with a maximum error of <0.03 m between modeled and observed water levels (Fig. 3.6).

Conversely, the modeled response slightly overestimated the hydraulic head and riparian storage in the mid-riparian well from March 1 until the beginning of June relative to discrete manual field measurements in 2005, and continuous field data in 2006. Interpolation of sporadic manual hydraulic head data in 2005 at the mid-riparian well made comparison of the rapid water table response difficult. Nonetheless, simulated responses to late summer storms appeared to be of the appropriate magnitude and timing, relative to observed water table responses (Fig. 3.6). In 2006, the simulated hydraulic head output in the mid-riparian zone also mimicked the timing and magnitude of continuous field data after the disappearance of soil frost (Fig. 3.6).

The upland toe well was installed in early summer of 2005, requiring estimation of water levels using nearby wells and the observed melt response from 2006 data, prior to June 1, 2005. In both years, lower antecedent moisture conditions prohibited extensive soil frost development near this well. The modeled responses reasonably replicate the timing and magnitude of observed responses throughout the simulation periods, including early spring, unlike the mid-riparian and pond observation points (Fig. 3.6). Maximum error between the magnitude of observed and modeled water table response was approximately 0.4 m. However, the modeled response was generally within 0.1 to 0.2 m of the measured hydraulic head, and showed a similar range in water table elevation (approximately 1.4 m) over the duration of the study, (Fig. 3.6).

In 2005, modeled responses at the mid-riparian and upland toe wells diverged from the observed responses after mid August, with heads being underestimated (Fig. 3.6). However, in 2006 the modeled response at these monitoring points closely matched the observed head data in late summer. Air temperature and soil tension data collected during mid August and early September 2005 show that there was likely a cessation of aspen evapotranspiration because of cool air temperatures (i.e., daily maximum of 5 – 10 °C). The generalized aspen *AET* estimate used for the net atmospheric flux boundary conditions did not account for daily temperature variations, likely overestimating riparian and upland evapotranspiration during this late summer period. Consequently, anomalously low head values in the riparian zone were simulated by the model.

Despite small errors regarding the timing of hydrologic responses in early spring and in the magnitude of perched water table response after the soil frost periods in each year, the model reasonably predicted hydrologic processes within the saturated (perched) flow regime that are critical to pond sustainability (Fig. 3.6). Spatial and temporal distribution of saturated groundwater flow reversals, and pond stage/storage changes were comparable to field observations made over the study period.

3.4.3 Unsaturated flow regime

The unsaturated flow regime is an equally important element of the conceptual model to honour the documented interaction between upland vegetation and perched groundwater (Chapter 2). Modeled results for volumetric water content and both vertical and lateral tension gradients were compared to field observations made at nested TDR probes and tensiometers (PN3: Fig. 3.6) located at the transition from riparian to upland vegetation. Model output for soil tension gradients and volumetric water content were plotted against field data (Fig. 3.7). The simulation results indicated upward flow at PN3 (Fig. 3.7b), into the rooting zone where evaporative flux was removed from the domain. Although the simulated magnitude of vertical gradients is smaller than observed in the field, large downward gradients of 2 – 3, indicating upward unsaturated flow similar to Schuh *et al.* (1993), were generated by the model. Given the fact that the porous media zones in the riparian zone and upland toe areas were homogenized, some error in the gradients was expected. Timing of infiltration events resulting from precipitation events were accurately predicted, although relative to the field data, they over-predict downward flow, or infiltration into deep storage for large precipitation events >10 mm.

Lateral unsaturated groundwater flow gradients predicted between PN2 and PN3, were also similar to those observed in the field. Unsaturated hydraulic head gradients of

0.6 were simulated by the model between PN2 and PN3, which is comparable to the measured gradient of 0.9 measured in the field. Between PN3 and PN4, a gradient of 1.1 was predicted by the model, compared to the gradient of 1.7 measured between these tensiometer nests in the field. Despite minor error in the magnitude of these gradients, the model simulated the lateral unsaturated flow regime with reasonable results.

Model output for volumetric water content also resembled soil moisture content data collected from TDR probes monitored at PN3. The model output matched field data to within <10 % of the volumetric water content measured by TDR probes (Fig. 3.7c). More importantly, despite deficit conditions within upland areas (Chapter 2), the soil moisture storage did not change significantly over the 2006 hydrologic year, and the trends in soil moisture conditions predicted by the model closely matched field observations.

Comparison of initial soil moisture conditions at PN3 revealed that the initial conditions for the transient simulation were wetter than observed in the upper soil TDR installation depth, resulting in overestimation of melt response in the mid and lower TDR installation depths (Fig. 3.7c). Furthermore, the model overestimated the increases in soil moisture in response to precipitation events (Fig. 3.7c), indicating systematic error with increased infiltration and soil storage responses. However, despite smaller simulated tension gradients relative to those observed in the field (Fig. 3.7b), the model reasonably represents the overall unsaturated flux at the wetland margins perhaps due to the wetter soils having a higher relative permeability.

Finally, the unsaturated flow regime was evaluated using daily saturation profiles for the entire domain. The saturation profiles showed that the inverted water table position did not change during the simulations, corresponding to field data showing that it remained static throughout the study period. Although the location of the inflection point at the perched to inverted water table transition is difficult to determine from field data (Chapter 2), interpretation of field data indicates that its position was transiently variable. During wet periods such as spring melt and large summer storms (>10 mm), the saturated/unsaturated interface could move as much as 0.75 m towards the upland. Conversely, during highly evaporative periods it would recede towards the pond. Model results showed that the lateral extent of the perched groundwater lens extended into the upland approximately 1 m over the duration of each of the 2005 and 2006 transient simulations, but did not retract as indicated by field observations.

3.4.4 Surface water flow regime

The numerical model was also used to test the conceptual model for predicting the spatial distribution of runoff. The conceptual model states that runoff should be limited to areas with low soil water storage potential, which is controlled by the water table depth and the porous media properties (i.e., the depth to the confining layer, forest floor materials; Fig. 3.2; Chapter 2). No Hortonian overland flow was generated during any simulations, even in response to summer storms as large as 66 mm/day measured in July 2005 (Fig. 3.5). Infiltration occurred in both the riparian zone and upland areas in response to precipitation and snow melt, as shown by saturated hydraulic head, soil tension, and volumetric moisture content data (Fig. 3.6; Fig. 3.7). The only surface runoff generation predicted by the model occurred during the summer months, resulting from saturated overland flow processes. This occurred where available soil water storage was minimal in a small, dynamic zone within 1 – 5 m of the pond edge where the water table is near surface (<0.1 m) replicating field observations (Chapter 2), and similar to Hayashi *et al.* (1998). However, the model did not predict the near surface lateral runoff that occurred in early spring soils when concrete soil frost was present, which limited storage in much of the riparian zone.

3.4.5 Sensitivity analysis

A sensitivity analysis is necessary to test the influence of assumptions inherent to the conceptual model and numerical model. Insight was gained from both the calibration of the model and a formal sensitivity analysis. Analyses were completed to test the sensitivity of the numerical model results to the hydraulic characteristics assigned to each of the 9 porous media zones and the boundary conditions applied to the domain.

Field observations were used to assign spatial distribution of porous media types and to estimate initial hydraulic characteristics, including hydraulic conductivity, anisotropy ratio, and porosity. Simulations were run to systematically test each porous media zone by varying the hydraulic conductivity values one order of magnitude higher and lower relative to initial estimates, and anisotropy ratios were varied from 1:1 to 1:20 (K_x:K_z). Simulations were also run to determine the effect of porosity, which was varied based on published ranges (Freeze and Cherry, 1979) for each porous media type. The van Genuchten parameters (α and β) were not changed during the sensitivity analysis, as there was no physical basis or measurements in which to guide changes relative to those presented by Carsel and Parish (1988).

During the initial calibration process, it was evident that the model was not sensitive to the hydraulic characteristics of the porous media zones representing the high permeability sediments near the domain surface (i.e., zones 2, 4, and 5; Fig. 3.3b). The model also demonstrated little sensitivity to the hydraulic characteristics of the sand underlying the confining layer (zone 1; Fig. 3.3b), as there was little to no active flow in this portion of the domain.

The model was not sensitive to the specified head condition representing the static regional water table. Transient simulations were run both with and without the specified head boundary. These results showed that over the duration of the study, leakage through the confining layer was negligible. As such, the perched flow regime was unaffected, reinforcing field observations that the perched and regional flow systems were hydraulically uncoupled (Chapter 2).

The model displayed sensitivity to the hydraulic characteristics of the silty clay lithology (zone 7; Fig. 3.2b), located directly above the confining layer where much of the perched groundwater flow occurred. Heterogeneities including thin (< 0.1 m), high permeability lenses (i.e., ~10^{-3} m/d), and subtle, gradational changes in soil texture, were represented by a homogeneous porous media zone. The vertical hydraulic conductivity, or anisotropy ratio of the silty clay zone controlled the magnitude and timing of water table responses. When a K_x:K_z ratio of 1:1 was used, the timing and magnitude of the modeled perched groundwater table responses did not resemble those observed in the field. Generally, interflow was predicted to occur in the upland toe area at the interface with the overlying silt loam layer (zone 4; Fig. 3.3b). Interflow was not observed to occur in this area based on available field data. Using isotropic media in this zone, interflow could only be prevented using hydraulic conductivity values of 10^{-3} to 10^{-2} m/d. Such hydraulic conductivities are unrealistic given the dominant soil texture and reduced the ability of this porous media zone to support the perched water table elevation observed in the field. Therefore, the initial hydraulic conductivity with a 1:8 anisotropy ratio value was used (Fig. 3.4a). This resulted in vertical flow into zone 7 such that infiltration occurred, and simulated soil moisture data match field observations. The resulting perched groundwater table configuration was also reasonable relative to field observations.

The model was also sensitive to the hydraulic properties of the upland confining layer (zone 8; Fig. 3.3b). Without the calibration of the vertical hydraulic conductivity of this zone, the shape of the saturated groundwater lens observed in the field could not be achieved. During simulations used to develop initial conditions, perching occurred laterally across the entire domain without a specifying higher permeability relative to the plastic, unfractured, blue-grey clay confining layer beneath the wetland (zone 6; Fig. 3.3b). The calibrated hydraulic conductivity of this zone was an order of magnitude higher than the unfractured confining layer, and had an anisotropy ratio of 1:2 (Fig. 3.4a).

The transient simulations reinforced field observations that atmospheric flux patterns, and storage potential are the primary controls on the hydrologic response of P19 (Chapter 2). Several simulations were run to test the sensitivity of the model to the estimated evaporative fluxes (annual *AET* values) applied to the upland, riparian, and pond areas. The estimate of pond evaporation derived from the evaporation pan (Chapter 2), had the least uncertainty of the three zones within the model domain. Therefore, the pond *AET* was increased and decreased only 10% from the measured value. Simulations showed that the pond level was slightly sensitive to the applied evaporative flux, but the overall perched flow regime was not.

Literature values of Boreal aspen *AET* can vary significantly (e.g., 318 to 529 mm/yr; Devito *et al.*, 2005b). However, the *AET* of the aspen upland areas at the study site were not water limited during the study due to unsaturated flux (recharge) from the wetland margin (Chapter 2). Therefore, similar to the pond, the estimated AET_{RZ} (based on 440 mm/yr; Amiro *et al.*, 2006) was varied slightly from the initial estimate in different simulations. These simulations showed that the overall perched flow regime was not sensitive to the applied upland evaporative flux.

The highly transient conditions in the riparian zone, including the variable water table position, shading, seasonal soil frost, and multiple plant species made it difficult to use *AET* estimates from published values. Therefore, different simulations were run with a range of different AET_{RZ} based on published AET estimates for mid-latitude and boreal fens, ranging from as low as 140 mm/yr (Eaton and Rouse, 2001) to as high as 380 mm/yr (Rouse, 2000). The transient hydrologic response of the numerical simulations was extremely sensitive to the net atmospheric flux regime applied to riparian evapotranspiration estimate. This is likely due to the riparian zone encompassing 40% of the domain surface. Application of high AET_{RZ} values caused severe water table depression in the riparian zone, which is not observed in the field (Chapter 2). Conversely, applying low AET_{RZ} values caused excessive groundwater ridging and flow reversals, and unrealistic pond levels after minor precipitation events. A small range of riparian AET_{RZ} (225 to 250 mm/yr), combined with the measured pond evaporation and initial upland AET_{RZ} estimate, resulted in reasonable predictions of the observed hydrologic response. Due to the sensitivity to the riparian evaporation, the model was also sensitive to the sum of positive fluxes, which included the lateral flux from the ephemeral draw. In 2005, without the additional flux (24 mm) from the draw, the model predicted pond and riparian water levels lower than those observed in the field.

3.5 *Discussion*

3.5.1 Study limitations

HydroGeoSphere (HGS) was chosen for its ability to accurately predict surface water-groundwater interaction and the unsaturated flow regime. The intent of this study was to test the interpretations made from rigorous field studies facilitating generalizations that honour plot scale processes such that up-scaling is computationally possible. However, the complexity and scale of the localized hydrologic processes being investigated required intensive data collection to calibrate the model.

The version of HGS used for the modeling is not capable of simulating soil frost development, which appears to be an important hydrologic process in the WBP. Although the spatial distribution of lateral flow was limited, field studies (Chapter 2) showed that it is an important process sustaining P19. Given the dominance of vertical water flux in this environment (Ferone and Devito, 2004; Petrone *et al.*, 2007; Chapter 2), predicting runoff related to soil frost is likely an important process in interpretation of wetland connectivity and larger scale problems.

Depth to ice and ice thickness measurements indicate seasonal, laterally continuous concrete soil frost in much of the riparian zone. Concrete soil frost was present in the near-surface (<0.2 m) mineral soils from the pond edge to between the mid-riparian well (16-SW; Fig. 3.6), and upland toe (50-W; Fig. 3.6) monitoring wells until June 10, and May 21 in 2005 and 2006 respectively (marked as soil frost in Fig. 3.6).

Field data indicate minimal storage changes occurred in the mid-riparian area as a result of over-ice quick flow into the pond from the mid-riparian area during the spring melt (Fig 3.6). The model's inability to represent the transient permeability field due to concrete frost resulted in over prediction of riparian storage changes in the mid riparian well and underestimation of the volume of water stored in the pond until the end of May both years (Fig. 3.6). Despite this limitation, the annual storage change predicted by the model in the pond basin and upland transect was relatively close to that observed in the field.

The complex flux dynamics at the interface between the wetland and upland were difficult to represent in the model. Actual evaporative loss from soil storage in the upland toe area occurs as a diffuse flux within the top metre of upland porous media, which is concentrated in a thin (0.2 – 0.4 m) dense rooting zone, rather than from a sheet of nodes as implemented in the numerical model. Furthermore, the complex aspen root physiology, and potential redistribution of water in the upland areas is not represented in the model. Deviations between observed and simulated vertical unsaturated head gradients, and lateral movement of the saturated/unsaturated interface into the upland, might be eliminated if a diffuse evaporative flux including deep root water uptake were implemented. This warrants further study.

3.5.2 Conceptual model as an HRU

A primary objective of this study was to validate the important elements of the conceptual model pertaining to wetland/upland flux dynamics developed through interpretation of field data (Chapter 2). Validation of the conceptual model required interpretation of intensive temporal and spatial data, even though the model domain consisted of only a small, but representative portion of the field study area. Modeling of the flux dynamics between the catchment sub-components and the underlying regional groundwater flow system improves our understanding of the hydrologic linkages and function of perched systems within the boreal landscape.

The conceptual model developed for perched boreal wetland systems identifies hydrologic drivers that are similar to those documented in other boreal forest basins (Metcale and Buttle, 1999; Ferone and Devito 2004, Devito *et al.*, 2005b). These dominant mechanisms lead to a model that differs from the fundamental hydrologic landscape unit (FHLU) described by Winter (2001). The definition of a FHLU revolves around topographically driven groundwater flow and surface water flow regimes commonly found in other climates and ecozones. Boreal landscapes exhibit large variability in evaporative fluxes, precipitation/melt timing and magnitude, depression storage, antecedent moisture conditions in unsaturated areas, and soil frost development as the dominant controls hydrologic process controls (Metcalfe and Buttle, 1999; Carey and Woo, 2001; Devito *et al.*, 2005b). Winter and LaBaugh (2003) acknowledge that there are wetlands within a FHLU that can be isolated from the overall hydrologic system due to poor connection to surface water and groundwater flow systems. The low topographic relief, poor drainage network, and landforms with coarse-grained over fin-grained material common in the WBP suggest that topographically driven flow systems and wetlands may be subsidiary to these so-called "isolated" wetlands (Haitjema and Mitchell-Bruker, 2005).

The climate, hydrostratigraphy, and hydrologic processes described in Chapter 2 create a unique HRU with characteristics that differ from other documented Boreal systems (Metcalfe and Buttle, 1999; Carey and Woo 2001; Buttle *et al.*, 2004). Modeling of perched boreal wetlands has shown that despite similarities to moraine and clay-plain pond-peatland complexes described by Ferone and Devito (2004), they represent an end member HRU. Perched boreal wetland systems have less wetland connectivity, landscape scale runoff potential (surface connectivity), and negligible recharge to regional groundwater flow systems relative to moraine wetland/upland complexes. Thus, wetlands found on regional topographic highs with deep regional water tables in the WBP should make use of these empirically derived HRU characteristics to improve broad scale modeling efforts and water resource management practices.

This study has shown that an accurate assessment of the hydrologic response and budgets of perched wetland systems requires detailed understanding of the initial conditions and hydrostratigraphy associated with perched boreal wetlands. Temporal and spatial variations in storage potential are the primary control on hydrologic processes sustaining perched wetlands and surrounding aspen uplands (Chapter 2). Potential linkages to the rest of the landscape that might arise, if the climate and/or physical system were perturbed, are difficult to assess without the use of a coupled surface water-groundwater model. Conventional hydrologic models would have difficulty resolving the lack of wetland connectivity and the complex redistribution of water within individual catchments.

3.5.3 Applicability

The tremendous rate of development of the natural resource industry in the Boreal Forest (Alberta Environmental Protection, 1998), makes understanding the hydrologic function of all wetlands important. Modeling of perched systems in this environment has confirmed field observations that perched boreal wetlands are essentially an isolated HRU with minimal linkage to adjacent wetlands and landforms. Unidentified perched wetland areas in regional scale modeling efforts would negatively impact water resource predictions as they are a crucial HRU in maintaining the ecohydrologic balance in the Boreal Plains (Wolfe *et al.*, 2006; Chapter 2). Independence from large-scale flow systems thus limits direct impact to perched wetlands from ongoing regional scale anthropogenic activity, with the exception of climate change.

The independence from broad scale impacts and the ability to buffer annual climate variation inherent to perched boreal wetland conceptual model presents a new design concept for use in landscape reclamation strategies. Oil sands mining operations in the northeast portion of Alberta have disturbed the hydrogeology, topography, and land cover of huge tracts of boreal forest. Regulatory bodies mandate that these lands be reclaimed to self-sustaining landforms with the land resource development capacity of the undisturbed ecosystem. Such, the perched boreal wetland HRU is a robust alternative to use in the design of portions of the reclaimed landscape, in particular on man-made tailings sand landforms. Building portions of the landscape into a mosaic of perched wetlands and forested uplands would create a reclaimed landscape similar natural ecosystem found in the boreal plains. The isolation of perched wetland areas from long-term anthropogenic impacts related to alteration of regional flow systems associated with oil sand mining, could allow development of sustainable ecosystems on topographically elevated, coarse-grained, man-made structures.

Expanding the numerical model used in this study to three dimensions would assist in reclamation design by accurately assessing the hydrostatigraphic framework, storage requirements, and spatial distribution of wetland/upland complexes. It would also allow prediction of hydrologic response of both natural and reclaimed perched wetlands to be ascertained under various scenarios of climate change and induced hydrogeologic regimes. Scenario modeling is also useful tool for the forestry, and conventional oil and gas industries to accurately assess impacts from harvesting and road construction common in the area.

3.6 Conclusions

The complex interplay of spatially variable vegetative water demand, precipitation distribution, and heterogeneous sediments is difficult to represent and accurately implement in HGS. However, it was the goal to ascertain if the conceptual model developed in Chapter 2 is valid based on fundamentals of the groundwater flow equations, without forcing the solution with overly prescriptive boundary conditions, or poor assumptions. Despite summing all external fluxes (i.e., atmospheric and lateral fluxes) within net atmospheric flux boundary conditions and generalizing the complex geology, the surface water flow, groundwater flow, and runoff regimes were reasonably represented over the study period.

Modeling has reinforced field interpretations that the perched boreal ponds represent an HRU with important and unique features. Perched boreal wetland/upland complexes have permanent, laterally discontinuous water tables, and highly localized hydrologic processes. They sustain both the wetland and upland sub-components of the catchment in hydraulic isolation from regional groundwater. The overall hydrologic response of the systems is highly dependent on climate, which is strongly buffered by basin and upland water storage. This isolation lowers their susceptibility to impacts from regional disturbance, with the exception of climate change. These findings can be incorporated into a framework that will assist in refining sustainable resource development practices within the Western Boreal Forest.

This study has highlighted the dominance of climatic forcing in hydrologic processes in the Boreal ecosystems. Without accurate representation of the temporal and spatial distribution of storage as a function of atmospheric fluxes, hydrologic modeling efforts will only provide, at best, approximate predictions of hydrologic response and natural variability. The results indicate that the numerical model could not capture the hydrologic responses associated with soil frost development and over-ice quick flow, thought to be important mechanisms for long-term pond sustainability. This also indicated the limitations of using generalized *AET* estimates to simulate the hydrologic response and natural variability of perched boreal wetlands in long-term scenarios such as climate change. To best predict long-term hydrologic response of perched boreal wetlands, these processes need to be reasonably simulated. This suggests that coupling a hydrologic model such as HGS with a near surface energy balance or Soil Vegetation Atmosphere Transfer (SVAT) model, would improve predictions of the Western Boreal Plain's long-term hydrologic response.

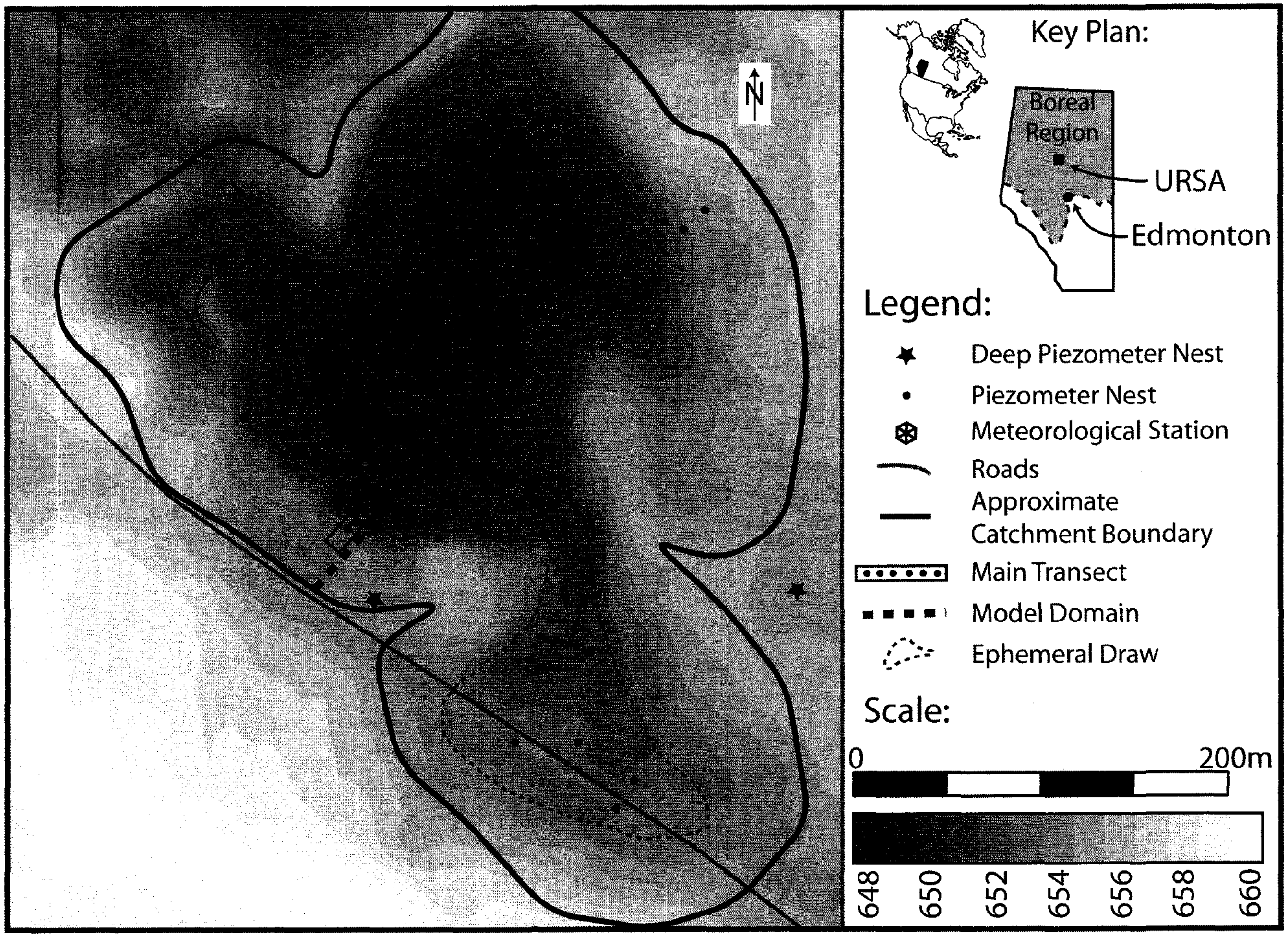

Figure 3.1. P19 study site with topography and main hydrologic instrumentation transect. Key plan shows the URSA and Boreal Region of Alberta, Canada.

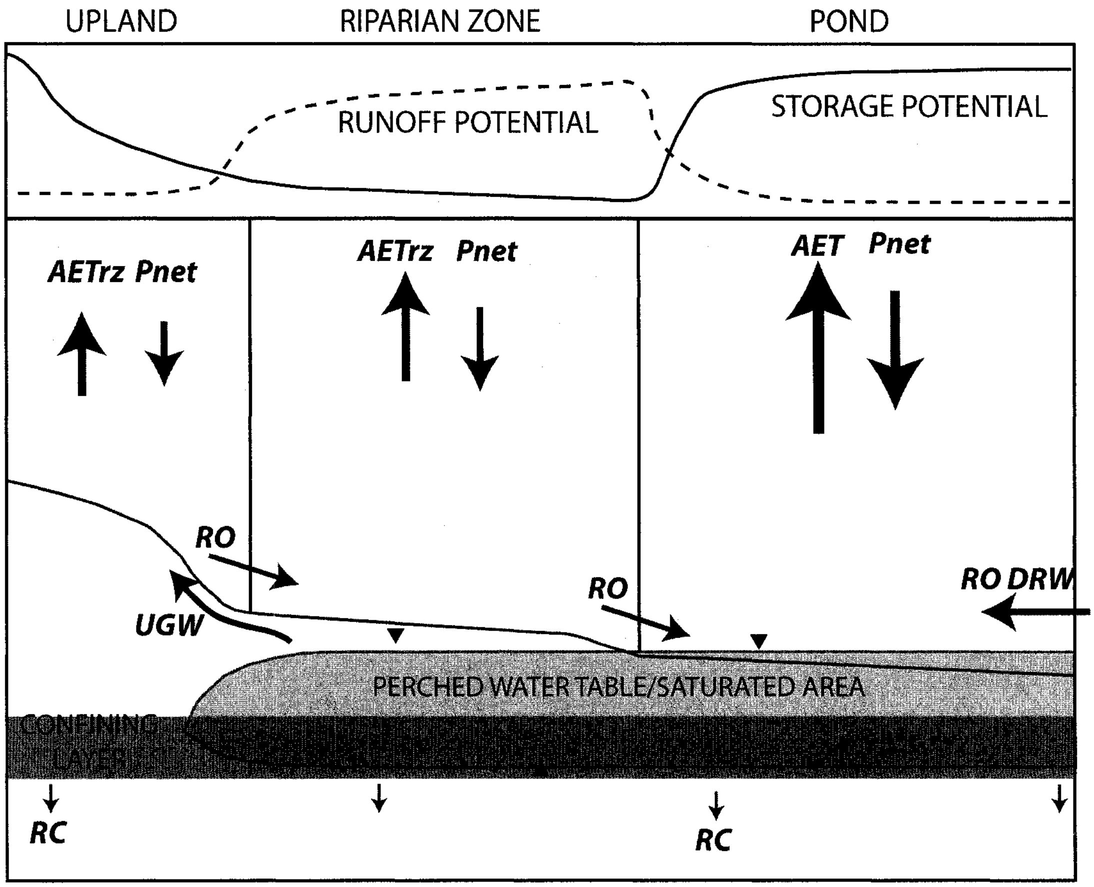

Figure 3.2. Cross-sectional representation of the conceptual model developed for perched wetland-upland flux dynamics (Chapter 2), showing upland, riparian zone, and pond areas with chart showing runoff, and storage potential. Atmospheric fluxes schematically represented by bold arrows where ***AET*** is actual actual evapotranspiration, ***AETrz*** is actual actual evapotranspiration from the rooting zone, and ***Pnet*** is net precipitation. ***RO DRW*** is specified runoff from the ephemeral draw. Intra-catchment fluxes resolved by the model are shown by grey arrows where ***RO*** is runoff to the pond area from the riparian zone, ***RC*** is drainage to the regional groundwater table, and ***UGW*** is unsaturated groundwater flux.

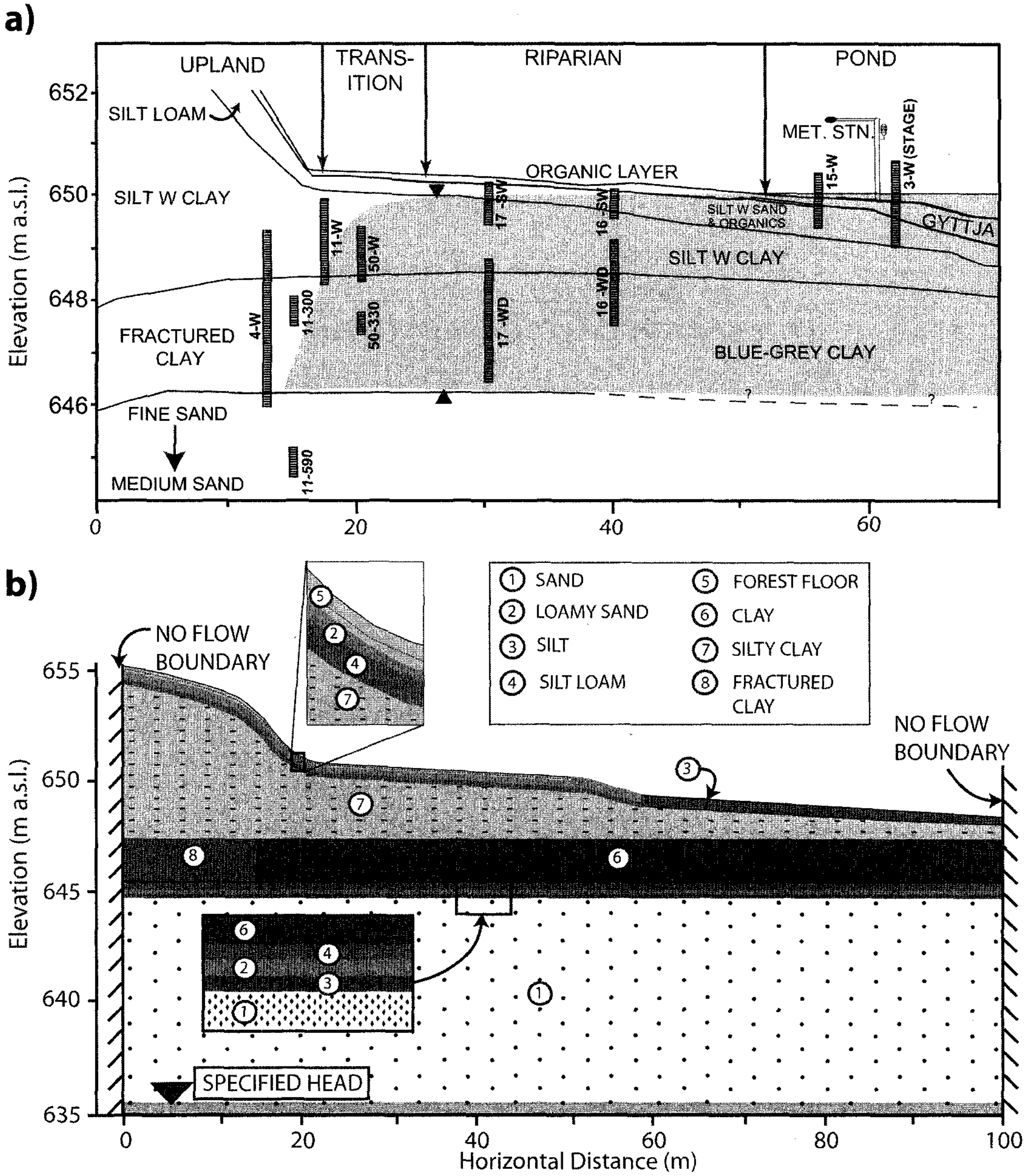

Figure 3.3. a) Detail of main instrumentation transect on Pond 19 (Fig. 3.1) showing geology, instrumentation locations, and average water table configuration. b) 2-D model domain representing main instrumentation transect and upland (Fig. 3.1) with porous media zones and boundary conditions.

a)

Line #	1	2	3	4	5	6	7	8
Porous Media	**Sand**	**Loamy Sand**	**Silt**	**Silt Loam**	**Forest Floor**	**Clay**	**Silty Clay**	**Frac. Clay**
α	14.5	12.4	1.6	2	N/A	0.8	0.5	0.5
β	2.68	2.28	1.3	1.41	N/A	1.09	1.09	1.09
ϕ_r	0.17	0.20	0.21	0.24	0.34	0.34	0.40	0.40
η	0.35	0.42	0.46	0.43	0.90	0.38	0.50	0.50
K (m/d)	4.3×10^{1}	1.0×10^{1}	8.9×10^{-3}	8.4×10^{0}	1.2×10^{1}	7.1×10^{-5}	7.1×10^{-4}	7.1×10^{-4}
Kx:Kz	1:1	2:1	1:1	10:1	1:1	1:1	1:8	1:2
Line								

b)

c)

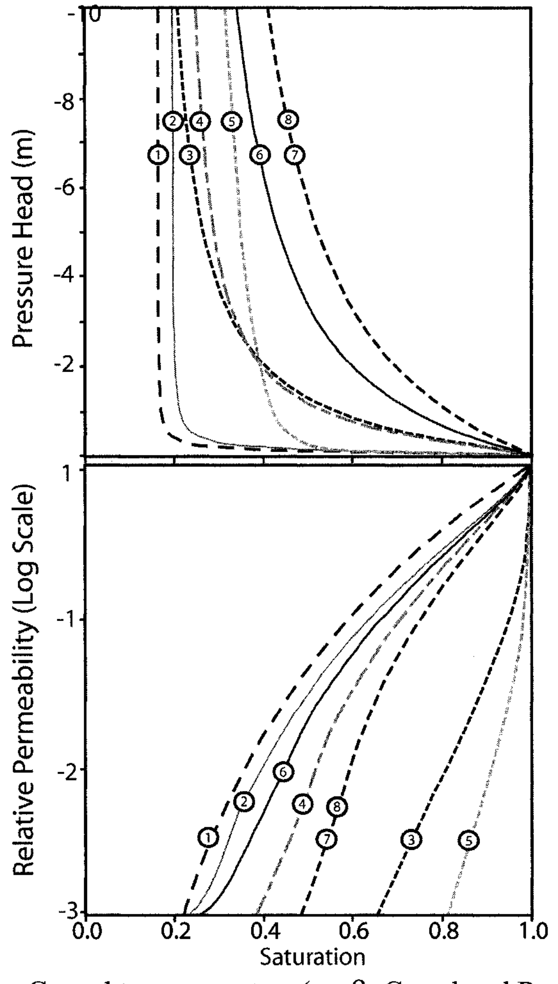

Figure 3.4. a) Tablulated van Genuchten parameters (α , β; Carsel and Parish, 1988), and hydraulic parameters for lithologies zones shown on Figure 3.3. b) Capillary pressure versus saturation curves and graph c) Relative permeability versus saturation curves for the lithologies in Table a).

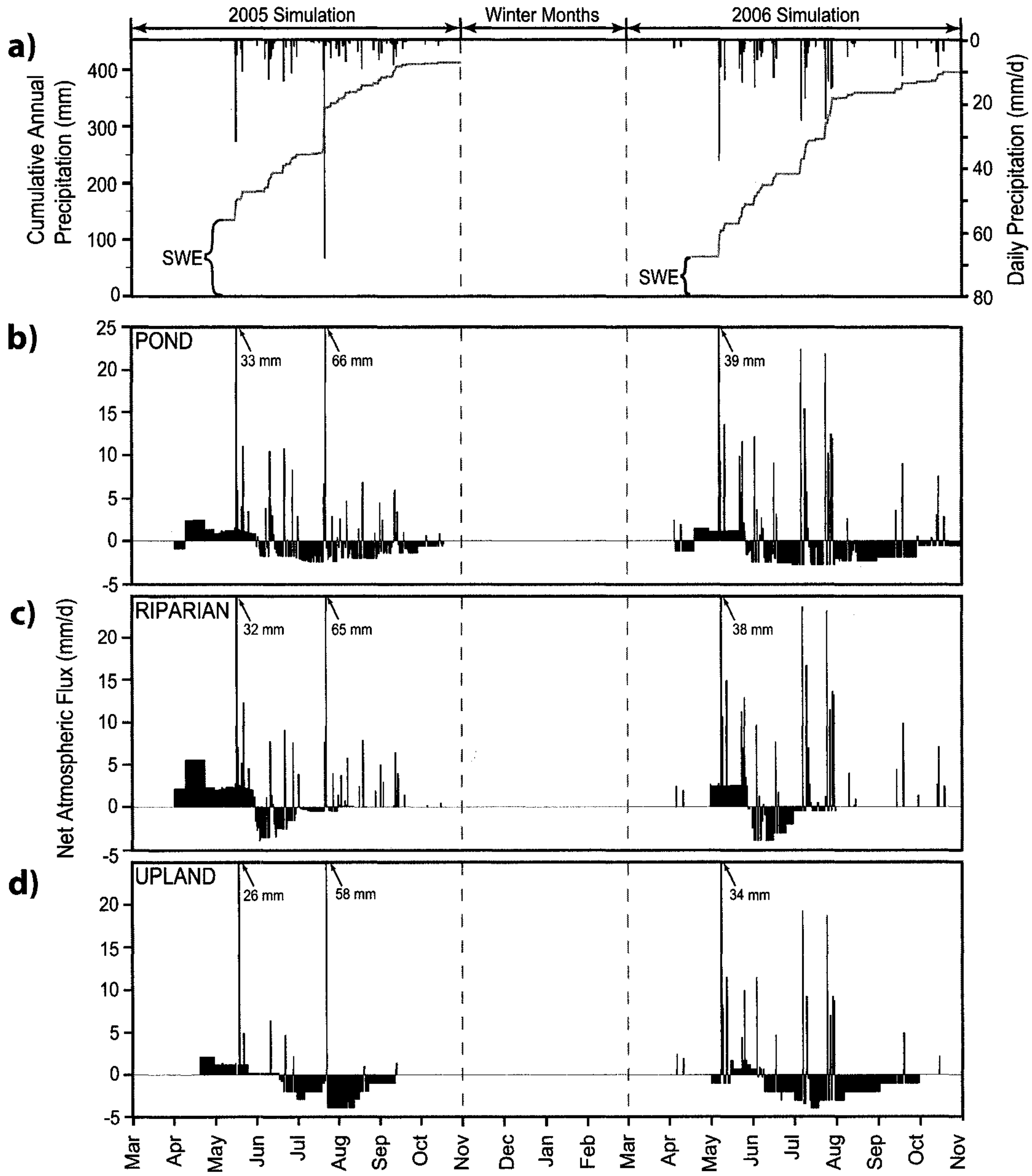

Figure 3.5. Four time series graphs over the study period showing a) event and cumulative precipitation and b), c), and d) the net atmospheric flux regimes applied to the pond area, the riparian zone and upland area respectively within the model domain. Note that lateral fluxes from the ephemeral draw are incorporated into the net fluxes applied to the pond and riparian zone.

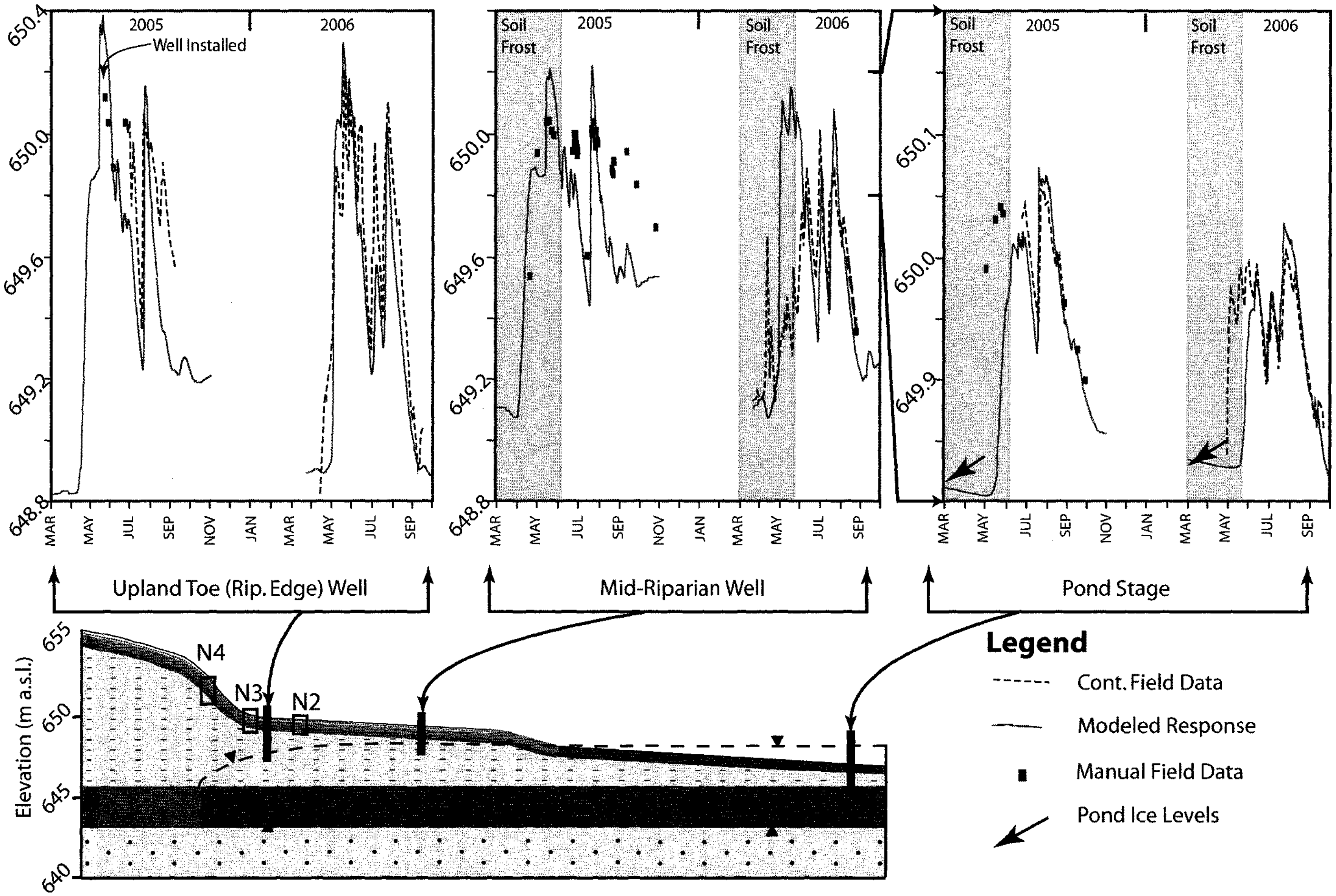

Figure 3.6. Field observations and model output for three monitoring points including hydraulic head for the upland toe, mid-riparian wells, and the pond stage for both hydrologic year simulations. Pond ice levels marked by arrows and the location and duration of observed soil frost as per field observations (Chapter 2). Soil moisture (PN2, PN3, and PN4) instrumentation nests and model observation locations (see Chapter 2). Note the exaggerated (4x) vertical scale for pond stage.

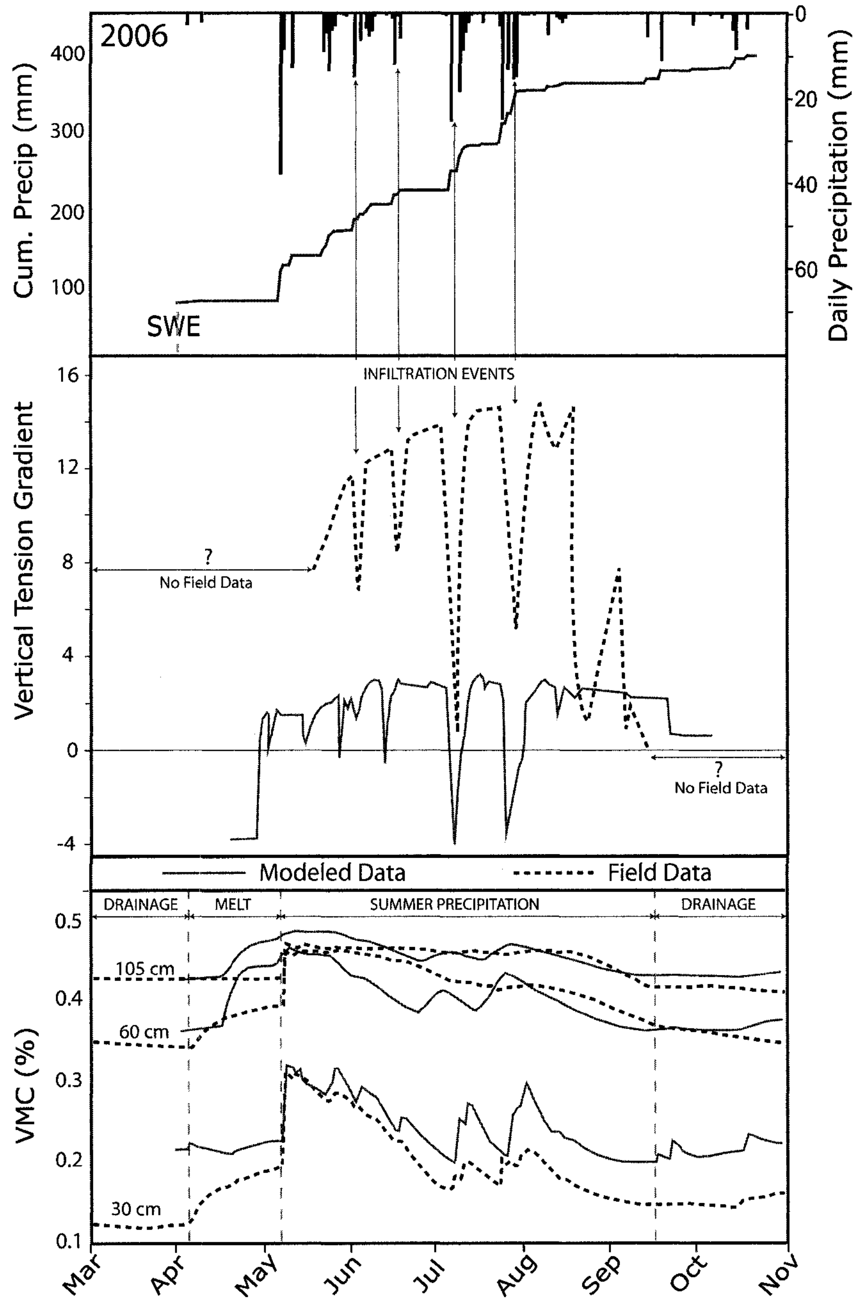

Figure 3.7. a) 2006 precipitation b) 2006 observed (dashed) versus modeled (solid) vertical tension gradients c) Observed (dashed) versus modeled (solid) soil storage response at 3 depths at PN3 (See Fig. 3.6)

3.7 *References*

Alberta Environmental Protection, 1998. The boreal forest natural region of Alberta. Natural Resource Services, Recreation and Protection, Special Report.

Amiro, B.D. *et al.*, 2006. Carbon, energy and water fluxes at mature and disturbed forest sites, Saskatchewan, Canada. Agricultural and Forest Meteorology, 136: 237-251.

Bryant, D., Nelson, D., Tangley, L., 1997. The last frontier forests: Ecosystems and economics on the edge. Washington D.C., World Resources Institute.

Buttle, J.M., Dillon, P.J. and Eerkes, G.R., 2004. Hydrologic coupling of slopes, riparian zones, and streams: an example from the Canadian Shield. Journal of Hydrology, 287(1-4): 161-177.

Carey, S.K. and Woo, M.-k., 2001. Slope runoff processes and flow generation in a subarctic, subalpine catchment. Journal of Hydrology, 253(1-4): 110-129.

Carsel, R.F. and Parrish, R.S., 1988. Developing joint probability distributions of soil water retention characteristics. Water Resources Research, 24(5): 755-769.

Devito, K., Creed, I., Gan, T., Mendoza, C., Petrone, R., Silins, U. and Smerdon, B., 2005a. A framework for broad scale classification of hydrologic response units on the Boreal Plains: Is topography the last thing to think of? Hydrological Processes, 19(8): 1705-1714.

Devito, K.J., Creed, I.F. and Fraser, C.J.D., 2005b. Controls on runoff from a partially harvested aspen-forested headwater catchment, Boreal Plain, Canada. Hydrological Processes, 19(1): 3-25.

Ducks Unlimited Canada. 2006. Waterfowl of the Boreal Forest. Ducks Unlimited Canada, Stonewall, Manitoba. 107p.

Eaton, A.K., and Rouse, W.R., 2001. Controls on evapotranspiration at a sub-arctic sedge fen. Hydrological Processes, 15: 3423-3421.

Ecoregions Working Group, 1989. Ecoclimatic Regions of Canada, Ecological Land Classification Series No. 23.

Environment Canada, 2003. Canadian Climate Normals 1971 - 2000: Slave Lake, Alberta. www.climate.weatheroffice.ec.gc.ca/climate_normals/results_e.html.

Fenton, M.M., Paulen, R.C. and Pawlowicz, J.G., 2003. Surficial geology of the Lubicon Lake area, Alberta (NTS 84B/SW). Alberta Geological Survey.

Ferone, J.M. and Devito, K.J., 2004. Shallow groundwater-surface water interactions in pond-peatland complexes along a Boreal Plains topographic gradient. Journal of Hydrology, 292(1-4): 75-95.

Freeze, R.A. and Cherry, J.A., 1979. Groundwater. Prentice-Hall Inc., Englewood Cliffs, New Jersey, 604p.

Haitjema, H.M., Mitchell-Bruker, S., 2005. Are water tables a subdued replica of the topography. Ground Water 43(6): 781-786.

Hayashi, M., van der Kamp, G. and Rudolph, D., 1998. Water and solute transfer between a prairie wetland and adjacent uplands, 1. Water balance. Journal of Hydrology, 207(1-2): 42-55.

Hinds, J.J., Ge, S. and Fridrich, C.J., 1999. Numerical modeling of perched water under Yucca Mountain, Nevada. Ground Water 37(4): 498-502.

Lafleur, P.M., McCaughey, J.H., Joiner, D.W., Bartlett, P.A. and Jelinski, D.E., 1997. Seasonal trends in energy, water and carbon dioxide at a northern boreal wetland. Journal of Geophysical Research, 102(24): 29009-29020.

Metcalfe, R.A. and Buttle, J.M., 1999. Semi-distributed water balance dynamics in a small boreal forest basin. Journal of Hydrology, 226(1-2): 66-87.

Meyboom, P., 1966. Unsteady groundwater flow near a willow ring in hummocky moraine. Journal of Hydrology, 4(1): 38-40.

Mills, J.G. and Zwarich, M.A., 1986. Transient groundwater flow surrounding a recharge slough in a till plain. Canadian Journal of Soil Science, 66: 121-134.

Orr, B.R., 1999. A transient numerical simulation of perched ground-water flow at the test reactor area, Idaho national engineering and environmental laboratory, Idaho 1952-94. U.S. Geological Survey Water-Resources Investigations Report 99-4277: 51p.

Pawlowicz, J.G. and Fenton, M.M., 2002. Drift thickness of the Peerless Lake map area (NTS 84B). Alberta Geological Survey

Petrone, R.M., Devito, K.J., Kaufman, S., Macrae, M.L. and Waddington, J.M., 2005. Potential carbon losses from boreal pond and riparian areas: influence of temperature and drought. Dynamics and Biogeochemistry of River Corridors and Wetlands, IAHS Redbook Series: 10-18.

Petrone, R.M., Silins, U., Devito, K.J., 2007. Dynamics of evapotranspiration from a riparian pond complex in the Western Boreal Forest, Alberta, Canada. Hydrological Processes, 21(11): 1391-1401.

Redding, T.E., Hannam, K.D., Quideau, S.A. and Devito, K.J., 2005. Partical density of aspen, pine, and spruce forest floors in Alberta, Canada. Soil Science Society of America Journal, 69: 1503-1506.

Rouse, W.R., 2000. The energy and water balance of high-latitude wetlands: control and extrapolation. Global Change Biology, 6: 59-68.

Samran, S., Woodward, P.M. and Rothwell, R.L., 1995. The effect of soil water on ground fuel availability. Forest Science, 41: 255-267.

Schuh, W.M., Meyer, R.F., Sweeney, M.D. and Gardner, J.C., 1993. Spatial variation of root-zone and shallow vadose-zone drainage on a loamy glacial till in a sub-humid climate. Journal of Hydrology, 148(1-4): 1-26.

Smerdon, B.D., Devito, K.J. and Mendoza, C.A., 2005. Interaction of groundwater and shallow lakes on outwash sediments in the sub-humid Boreal Plains of Canada. Journal of Hydrology, 314(1-4): 246-262.

Smerdon, B.D., Mendoza, C.A. and Devito, K.J., 2007. Simulations of fully coupled lake-groundwater exchange in a subhumid climate with an integrated hydrologic model. Water Resources Research, 43(1): W01416, doi:10.1029/2006WR005137.

Sudicky, E.A. *et al.*, 2005. On the challenge of integrated surface-subsurface flow and transport modelling at multiple catchment scales. Abstracts with programs - Geological Society of America, 37(7): 28.

Therrien, R., McLaren, R.G., Sudicky, E.A. and Panday, S.M., 2007. HydroGeoSphere: A three-dimensional numerical model describing fully-integrated subsurface and surface flow and solute transport. Groundwater Simulations Group, University of Waterloo. 350p.

Winter, T.C., 2001. The concept of hydrologic landscapes. Journal of the American Water Resources Association, 37(2): 335-349.

Winter, T.C. and LaBaugh, J.W., 2003. Hydrologic considerations in defining isolated wetlands. Wetlands, 23(3): 532-540.

Wolfe, B.B., Hall, R.I., Last, W.M., Edwards, T.W.D., English, M.C., Karst-Riddoch, T.L., Paterson, A. and Palmini, R., 2006. Reconstruction of multi-century flood histories from oxbow lake sediments, Peace-Athabasca Delta, Canada. Hydrological Processes, 20(19): 4131-4153.

York, J.P., Person, M., Gutowski, W.J. and Winter, T.C., 2002. Putting aquifers into atmospheric simulation models: an example from the Mill Creek Watershed, northeastern Kansas. Advances in Water Resources, 25: 221-238.

Chapter 4

Summary and Conclusions

4.1 Hydrologic characteristics of perched boreal wetlands

Perched boreal wetlands are not well documented; however, they are likely ubiquitous throughout the Western Boreal Plain (WBP), making them an important component of local and regional water resources (Wolfe *et al.*, 2006; Chapter 2). This work shows that perched wetlands can persist in a sub-humid climate, without regular contributions from regional groundwater flow systems, or overbank contributions as documented in the Peace-Athabasca River delta in northern Alberta (Wolfe *et al.*, 2006). The heterogeneous Quaternary geology and sub-humid climate of northern Alberta creates a hydrologic framework capable of sustaining permanent perched wetlands that are hydraulically isolated from large-scale groundwater flow systems, despite frequent water deficit conditions inherent to the region. The hydrological processes that sustain perched wetlands in this environment are primarily controlled by the temporal and spatial distribution of water storage as a function of atmospheric fluxes and geology.

The hydrostratigraphy of the study area consists of relatively coarse-grained porous media overlying a low permeability, clay, confining layer. This low permeability layer impedes the vertical flow of water. Furthermore, the variable thickness of the coarse-grained deposits creates a spatially variable distribution of potential soil storage. Temporal atmospheric flux patterns of precipitation and evapotranspiration superimposed on the spatial distribution of water storage potential, thus controls the spatial and temporal distribution of saturation, vegetation, and lateral water fluxes. These observations emphasize the importance of understanding the function of wetland riparian areas with low storage potential in terms of predicting perched boreal wetland sustainability and susceptibility to impacts. Furthermore, field and numerical modeling studies have shown that seasonal soil frost develops in low storage potential areas such as riparian areas and ephemeral draws. This reinforces the need to understand and resolve the water distribution of vadose zone storage or antecedent moisture conditions of perched wetland areas to interpret this important lateral flow generation mechanism.

Despite having hydrostratigraphy similar to other perched wetlands and flow systems (e.g., O'Geen *et al.*, 2003; Rockafeller *et al.*, 2004; Rains *et al.*, 2006), the hydrologic behaviour is quite different. The perched flow systems at the Utikuma Region Study Area (URSA) are largely permanent and laterally discontinuous, which contrasts the laterally extensive, topographically driven perched flow systems that develop above shallow fragipan layers documented in loess landscapes of Idaho (O'Geen *et al.*, 2003) and in California (Rains *et al.*, 2006). Unlike these flashy, flow-through systems documented in the U.S., perched flow systems in the WBP capture and store water from limited lateral water fluxes that occur in this environment due to their lack of connectivity to other wetlands and/or drainage networks. The study wetlands also differ from perched boreal wetlands associated with the Peace-Athabasca river delta. Although they are both found in similar climatic and geologic context, the study wetlands are unique in that they do not have periodic hydrologic linkage to a high order river system (Wolfe *et al.*, 2006). This, again, reinforces the importance of local hydrologic processes in sustaining the wetland/upland complexes accounting for large areas of the WBP.

Detailed field study of perched boreal wetlands enabled development of a conceptual model to assess water cycling within closed wetland catchments. The study wetlands exist in closed (i.e., undrained) basins with hummocky topography and high evaporative demands and show analogous hydrologic processes to prairie pothole wetland models described in the literature (Meyboom, 1966; Millar, 1971; Mills and Zwarich, 1986; Hayashi *et al.*, 1998). The perched wetlands also display similarities to moraine wetlands at the URSA. These similarities include vegetative water table depressions at wetland margins, flow reversals in riparian treed bog/fen areas surrounding ponds (Ferone and Devito, 2004), and frost development in areas with low soil-water storage potential such as ephemeral draws (Devito *et al.*, 2005b). This conceptualization of perched boreal wetlands highlights the importance of the spatial and temporal distribution of subtle hydrologic processes and domination of vertical water fluxes (Redding and Devito, *in press*). Further, this conceptual model also indicates that concurrent interpretation of near surface geology and topography is necessary to evaluate local water cycling processes, similar to Haitjema and Mitchell-Bruker (2005) and Devito *et al.*, (2005a).

Division of each study wetland catchment into sub-components was required to assess complex spatial patterns and intra-catchment exchanges via unsaturated groundwater flow and surface water-groundwater interaction. Furthermore, evapotranspiration, interception loss, and storage/runoff potential varied between catchment sub-components, requiring careful consideration to complete accurate water budgets. Thus, there is a need to understand the spatial and temporal distribution of stored water. This is crucial to anticipating hydrologic linkages via saturated/unsaturated groundwater flow, saturated overland flow, and quick-flow over shallow soil frost. Assessment of these linkages is crucial to understand the possible impacts from land surface disturbances.

Chapter 2 illustrated that within the sub-humid climate of the WBP vertical water fluxes dominate at perched wetlands. However, the basic answer to the first of the three research questions posed in Chapter 1, is that localized lateral flow generation processes sustain perched boreal wetlands through long periods of drought common in the region (Devito *et al.*, 2005a). These processes occur in areas with low soil-water storage potential, such as riparian zones, ephemeral draws or valley bottoms. Flow over concrete frost that develops in areas of high antecedent moisture similar to Devito *et al.* (2005b) and Hayashi *et al.* (1998) is an important lateral flow generation mechanism, especially in the spring when evaporative fluxes are low.

Chapter 2 also identified the need to consider the spatial variability of net atmospheric fluxes, such as net precipitation, rooting zone evapotranspiration, open water evaporation, and interception, between catchment sub-components. This is required in order to accurately assess water budgets for perched wetlands. Further, quantifying the plant-water interaction, or the unsaturated groundwater flow at wetland margins driven by frequent net water deficits in upland areas (Price *et al.*, 2005), is crucial to properly understand the intra-catchment water cycling and long-term sustainability of perched wetland/upland complexes.

Chapter 3 confirmed that a fully coupled numeric hydrologic model can simulate the conceptual model of water cycling processes sustaining perched boreal wetlands based primarily on atmospheric fluxes. Similar to numerical simulations of nearby boreal wetlands on outwash deposits (Smerdon *et al.*, 2007), perched wetland simulations confirmed the sensitivity of the study wetlands to climate. The simulation results also confirm that local lateral flow generation processes such saturated overland flow and quick-flow over shallow frozen soil are important to pond sustainability. However, similar to Smerdon *et al.* (2007), it confirmed limitations inherent to the numerical model, such as its inability to predict the development and effects of frozen soils, and intensive data requirements required for calibration.

In summary, the objectives outlined in Chapter 1, including the determination of the hydrologic processes and their controls, important water cycling proceses, and confirmation of a new conceptual model for perched boreal wetlands, were satisfactorily addressed. This study has added to the growing knowledge base of boreal hydrology, which is critical given the ongoing disturbance and need to refine land conservation and management strategies.

4.2 Threshold behaviour, future work, and implications

The study of P19 encompassed two hydrologic years. One year experienced an annual water surplus; while the other experienced an annual water deficit. Field measurements made throughout the observed climatic variation between years, allowed characterization of the hydrologic processes and their controls that sustained the pond over the duration of the study. However, the short duration of the study did not allow for full assessment of the natural variability and threshold behaviour that maintain this pond as a permanent landform under extreme climatic conditions such as prolonged periods of drought, or water surplus.

Aerial photographs of the study area from the last 50 years suggest that hydrologic thresholds exist that help to explain the permanence of perched wetlands. Historic climate data was compiled from the nearby (100 km away) town of Slave Lake Alberta. These data were generalized to capture long-term trends in water deficit (drought) and water surplus (deluge). Over the catchment area, P_{net} was represented as 90% of the total annual precipitation, and *AET* was estimated to be 400 mm/year. From these generalizations, yearly surplus or deficit values were derived, and plotted as a yearly time series (Fig. 4.1). The aerial photographs were used in conjunction with a depth-area relationship developed from bathymetric survey data to determine the stage and volume of water stored in the pond basin. These photographs indicate that the volume of water stored in the pond during the study period is representative of "average" climatic conditions when compared to the generalized historic climate data (Fig. 4.1). These photographs also indicate that P19 has the potential to store nearly an order of magnitude more water, filling the pond basin during prolonged surplus conditions as observed in 1977 (Fig. 4.1).

The aerial photographs also indicate that even during severe drought conditions, the pond appears maintain an area of approximately 1.4 ha, only slightly smaller than the area of 1.7 ha observed during periods of "average" climatic conditions (Fig 4.1). This suggests that as the pond area gets smaller, the area in which the confining layer is near surface, relative to the pond area, gets larger. In effect, the *AET* from the pond basin gets smaller, while the catchment area that is likely to generate lateral flow towards the pond gets larger due to the basin morphology. These data also suggest a negative feedback cycle limiting actual evapotranspiration from the pond surface with decreasing open water area, in addition to reduced *AET* from riparian and upland vegetation due to water stress as the saturated/unsaturated interface contracts with the pond. The rate of contraction of the perched water table lens towards the pond centre exceeds the rate of growth of the upland plant community, and limits its ability to draw water from the saturated/unsaturated interface.

Moreover, the pond likely has feedback mechanisms that limit the maximum volume of water stored in the pond basin. During times of extreme deluge, the effective catchment area likely to generate runoff gets smaller relative to *AET* from the large open water area. This is demonstrated in 1977 when the pond area swelled to 5.6 ha, which filled the pond basin completely and resulted in higher catchment-wide *PET*, and leaving only the high storage potential upland area to generate runoff (Fig. 4.1). Similarly, during periods when the pond basin is completely inundated, there is greater potential for a sub-surface hydraulic linkage or "outflow". Thus, minor saturated groundwater flow may be initiated above the confining layer in upland areas where the confining layer morphology promotes flow away from the pond. However, the "recharge ring" concept presented in Chapter 2 may increase linkage to underlying groundwater flow systems during wet conditions. If so, the additional recharge via leakage through the confining layer at the wetland periphery may limit both pond size, and upland area water table development, and thus, inter-wetland connectivity.

These cyclical storage oscillations suggest that the water table configuration, runoff regime, and plant-water demand in the catchment are highly transient which is important given the dominance of evaporative fluxes in perched wetland water budgets. For example, the upland species surrounding the pond would terrestrialize the pond area over the long-term without sporadic inundation caused by climatic variations. As such, inundations cause mortality of encroaching upland species in the pond basin, thus resetting the vegetation to a more phreatophytic plant community much like what is currently observed in the P19 riparian zone. The frequency or likelihood of these inundation events appears to be the main ecohydrologic distinction between the permanent perched fen peatlands and the permanent or ephemeral perched open marsh wetlands with no peat development.

Given the duration and scope of this study, concrete statements regarding the aforementioned threshold behaviour and feedback mechanisms can not be made. However, the findings and conceptual model of this study of perched boreal wetlands add to the foundation of ecohydrology, reinforcing the emerging need for interdisciplinary science to address water resource management issues in many environments (Hannah *et al.*, 2007). These findings also allow further investigation of the long-term sustainability of the perched boreal wetlands under natural and anthropogenic disturbance scenarios through the use of numerical models. Future work can employ the processes elucidated in this study to refine, and expand numerical modeling efforts to address the long-term hydrologic function and sustainability of perched boreal wetlands. This will likely include long-term simulations using HydroGeoSphere (HGS) to determine if the model can reasonably simulate the natural variability observed over the last 50 years (Fig. 4.1) by specifying spatially fixed evaporative regimes and ignoring the effects of soil frost. However, it is suspected that a more complex numerical model with an energy balance component or coupling with an existing Soil Vegetation Atmosphere Transfer (SVAT) model (Arora, 2002) would be required to accurately simulate the natural variability of P19, under long-term simulations or climate change scenarios.

The rate of development in the WBP emphasizes the need to utilize processed-based hydrologic research such as this, to improve the balance of ecosystem conservation with industrial and economic pressures. The wetlands and the hydrologic processes sustaining them described in this study are an ideal example of how conventional paradigms or single disciplinary science may not satisfactorily explain the integration of hydrology with other earth systems. Accurate hydrologic assessment of prairie potholes and boreal wetlands require intensive spatial and temporal data collection to characterize complex patterns in saturated and unsaturated groundwater flow, and plant-water-atmosphere interaction (Rosenberry and Winter, 1997; Price *et al.*, 2005; Chapter 2). However, the empirically derived HRU characteristics for perched boreal wetlands and numerical models capable of explicitly simulating atmosphere-surface water-groundwater interactions will provide a solid platform to continue improving our understanding of hydrology of the region.

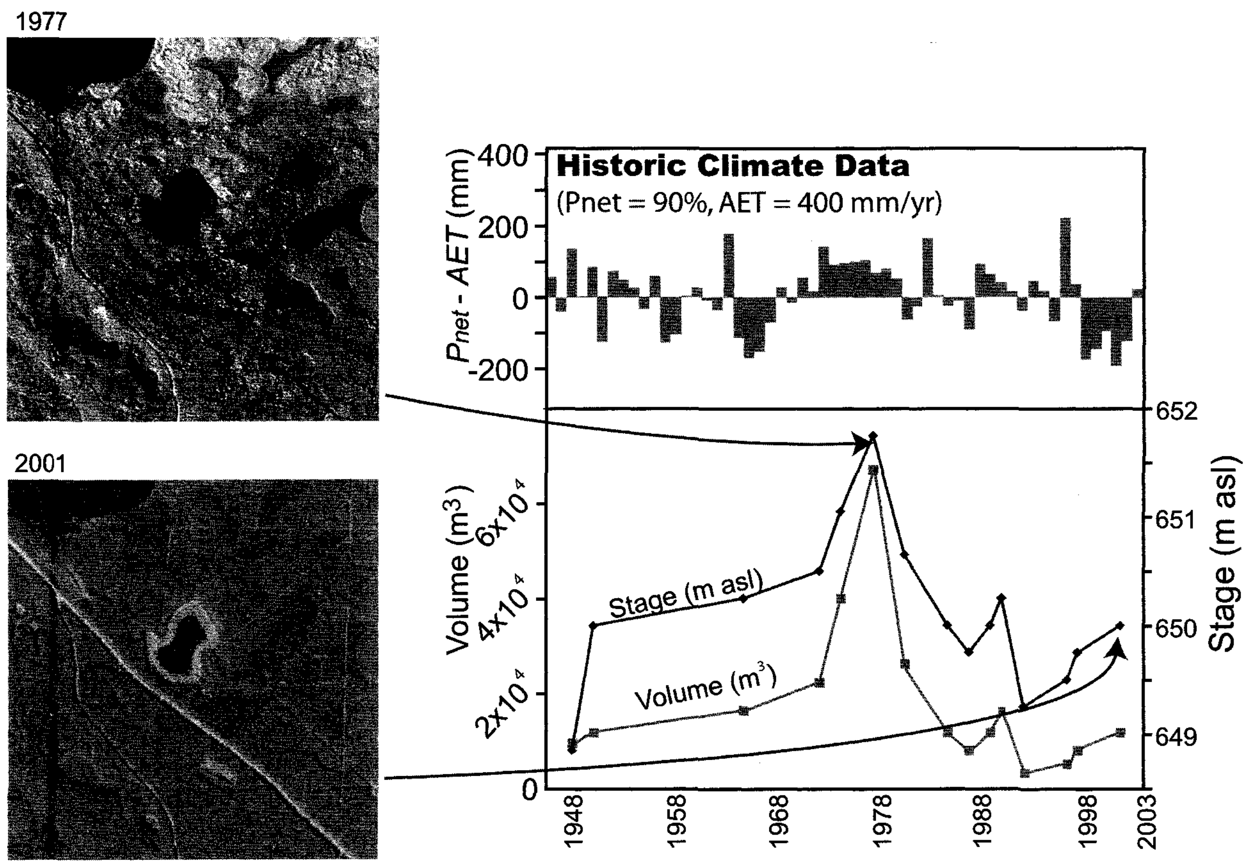

Figure 4.1. Historic climate data from Slave Lake, Alberta compared to air photograph analysis showing pond stage, and volume of water sored in the basin showing threshold behaviour in response to drought and water surplus conditions.

4.3 References

Arora, V., 2002. Modeling vegetation as a dynamic component in soil-vegetation-atmosphere transfer schemes and hydrological models. Reviews of Geophysics, 40(2): 1006, doi:10.1029/2001RG000103

Devito, K., Creed, I., Gan, T., Mendoza, C., Petrone, R., Silins, U. and Smerdon, B., 2005a. A framework for broad scale classification of hydrologic response units on the Boreal Plains: Is topography the last thing to think of? Hydrological Processes, 8(19): 1705-1714.

Devito, K.J., Creed, I.F. and Fraser, C.J.D., 2005b. Controls on runoff from a partially harvested aspen-forested headwater catchment, Boreal Plain, Canada. Hydrological Processes, 19(1): 3-25.

Ferone, J.M. and Devito, K.J., 2004. Shallow groundwater-surface water interactions in pond-peatland complexes along a Boreal Plains topographic gradient. Journal of Hydrology, 292(1-4): 75-95.

Hayashi, M., van der Kamp, G. and Rudolph, D., 1998. Water and solute transfer between a prairie wetland and adjacent uplands, 1. Water balance. Journal of Hydrology, 207(1-2): 42-55.

Hannah, D.M., Sadler, J.P. and Wood, P.J., 2007. Hydroecology and ecohydrology: apotential route forward. Hydrological Processes, 21: 3385-3390.

Haitjema, H.M., Mitchell-Bruker, S., 2005. Are water tables a subdued replica of the topography. Ground Water 43(6): 781-786.

Meyboom, P., 1966. Unsteady groundwater flow near a willow ring in hummocky moraine. Journal of Hydrology, 4(1): 38-40.

Millar, J.B., 1971. Shoreline-area ratio as a factor in rate of water loss from small sloughs. Journal of Hydrology, 14(3-4): 259-284.

Mills, J.G. and Zwarich, M.A., 1986. Transient groundwater flow surrounding a recharge slough in a till plain. Canadian Journal of Soil Science, 66: 121-134.

O'Geen, A.T., McDaniel, P.A., Boll, J. and Brooks, E., 2003. Hydrologic processes in valley soilscapes of the Palouse Basin in northern Idaho. Soil Science, 168(12): 846-855.

Price, J.S., Branfireun, B.A., Waddington, J.M. and Devito, K.J., 2005. Advances in Canadian wetland hydrology, 1999-2003. Hydrological Processes, 19(1): 201-214.

Rains, M.C., Fogg, G.E., Harter, T., Dahlgren, R.A. and Williamson, R.J., 2006. The role of perched aquifers in hydrological connectivity and biogeochemical processes in vernal pool landscapes, Central Valley, California. Hydrological Processes, 20(5): 1157-1175.

Redding, T.E. and Devito, K.J., 2005. Lateral and vertical flow thresholds for aspen forested hillslopes in the Western Boreal Plain, Alberta, Canada. Hydrological Processes, *in press*.

Rockafeller, P.A., McDaniel, P.A. and Falen, A.L., 2004. Perched water table responses to forest clearing in Northern Idaho. Soil Science Society of America Journal, 68: 168-174.

Smerdon, B.D., Mendoza, C.A. and Devito, K.J., 2007. Simulations of fully coupled lake-groundwater exchange in a subhumid climate with an integrated hydrologic model. Water Resources Research, 43(1): W01416, doi:10.1029/2006WR005137.

Wolfe, B.B., Hall, R.I., Last, W.M., Edwards, T.W.D., English, M.C., Karst-Riddoch, T.L., Paterson, A. and Palmini, R., 2006. Reconstruction of multi-century flood histories from oxbow lake sediments, Peace-Athabasca Delta, Canada. Hydrological Processes, 20(19): 4131-4153.