Assessing hydrological processes controlling the water balance of lakes in the Peace-Athabasca Delta, Alberta, Canada using water isotope tracers

by

Matthew D. Falcone

A thesis

presented to the University of Waterloo

in fulfillment of the

thesis requirement for the degree of

Master of Science

in

Earth Sciences

Waterloo, Ontario, Canada, 2007

I hereby declare that I am the sole author of this thesis. This is a true copy of the thesis, including any required final revisions, as accepted by my examiners.

I understand that my thesis may be made electronically available to the public.

Assessing hydrological processes controlling the water balance of lakes in the Peace-Athabasca Delta, Alberta, Canada using water isotope tracers.

ABSTRACT

One of the world's largest freshwater deltas (~4000 km^2), the Peace-Athabasca Delta (PAD), is located at the convergence of the Peace and Athabasca rivers and Lake Athabasca in northern Alberta, Canada. Since the early 1970s, there has been increasing concern regarding the ecological impacts on the PAD after flow regulation of the Peace River began in 1968, decreased discharge in the Peace and Athabasca rivers as a result of hydroclimatic changes in Western Canada, and increased Athabasca River water usage by oil sands development to the south. This thesis is part of an ongoing, multi-disciplinary project assessing current and past hydrological and ecological conditions in the PAD. Research conducted in this thesis aims to better understand the processes controlling water balance of lakes in the PAD using mainly stable water isotope data collected from lakes and their input sources. Isotope data are used to describe and quantify hydrological processes for individual lakes (seasonal and annual) and across the delta and are supported by other chemical and hydrometric data.

An isotopic framework in $\delta^{18}O$-$\delta^{2}H$-space is developed for the PAD using evaporation-flux-weighted local climate data, and isotopic data collected from a reference basin, lakes throughout the PAD, and lake input sources (i.e., snowmelt, rainfall, and river water). The framework is comprised of two reference lines, the Local Meteoric Water Line, which is based on measured isotopic composition of precipitation, and the Local Evaporation Line, which is based on modelled isotopic composition of reference

points. Evaporation pan data is used to assess short-term variations in key isotopic reference values, which are important for addressing short-term changes in the isotopic signature of shallow basins. This framework is used in subsequent chapters including assessment of seasonal and annual water balance of two hydrologically-contrasting shallow lakes, and to quantify the impacts of flood water and snowmelt on a set of 45 lakes in spring 2003.

Five years of isotope data using time-series analysis and the isotopic framework suggested that a perched (isolated) lake and its catchment (forest and bedrock) in the northern, relict Peace sector captured sufficient rain, snow, and runoff to maintain a relatively stable water balance, and also that a low-lying lake in the southern, active Athabasca sector was regularly replenished with river water in both spring and summer. Snowmelt and rainfall were found to have diluted the perched basin by an average of 16% and 28 % respectively, while spring and summer floods were found to almost completely flush the low-lying lake.

Using the spring 2003 regional dataset, flooded lakes were separated from snowmelt-dominated lakes through use of suspended sediment concentrations, isotope data, and field observations. Application of an isotope mixing model translated $\delta^{18}O$ values into a range of replenishment amount by either river water or snowmelt, which compared well with hydrological conditions at the time of sampling and previously classified drainage types of the lakes. Spatial mapping of replenishment amounts illustrated flooding of much of the Athabasca sector due to ice-jams, except for two sub-regions isolated from flooding by artificial and natural northern diversion of flow from the Athabasca River. It is also shown that most of the relict landscape of the Peace sector was replenished by snowmelt except for a few low-lying lakes close to the Peace River and its tributaries. Overall, improved understanding of lake and regional hydrology in the PAD, especially the ability to quantify the affects of various lake inputs, will improve the ability to develop effective guidelines and management practices in the PAD as lakes respond to future changes in climate and river discharge.

ACKNOWLEDGEMENTS

I would like to thank my supervisors, Dr. Brent Wolfe and Dr. Thomas Edwards, for their guidance and long-term companionship, and for providing me the opportunity to partake in such interesting research. I am also grateful to the many field assistants who have helped collect data throughout the duration of the study and the University of Waterloo's Environmental Isotope Laboratory for its rapid and precise analyses. Great appreciation goes out to Roland Hall and Rich Petrone for their helpful comments.

I express my thanks to the financial support provided by the Ontario Graduate Scholarship (research), Northern Scientific Training Program under Indian and Northern Affairs Canada (field logistics), Natural Sciences and Engineering Research Council of Canada (undergraduate student research award program and Northern Research Chair Program; field logistics and analysis), Polar Continental Shelf Project (helicopter time), British Columbia Hydro and Power Authority (field logistics and analysis), and Wood Buffalo National Park (field logistics).

Last but not least, I would like to extend a special thank-you to my family for their continuous guidance and support through the years. A very special thank-you goes out to my fiancée, Tricia Stadnyk, for her encouragement and emotional support over the past few years, and for critiquing versions of this dissertation.

TABLE OF CONTENTS

LIST OF TABLES

LIST OF ILLUSTRATIONS

1.0 INTRODUCTION

Introduction

The Peace-Athabasca Delta (PAD) is one of the largest freshwater deltas in the world, and is located mainly within the Wood Buffalo National Park at the convergence of the Peace and Athabasca rivers and Lake Athabasca in northern Alberta (Figure 1). It is recognized by both the International Ramsar Convention on Wetlands and United Nations Educational, Scientific and Cultural Organization (UNESCO) for its ecological, historical and cultural significance (Wetlands International, 2005). The delta is home to large populations of muskrat, beaver, and free-ranging wood bison, and lies along one of the main North American migratory bird flyways, as well as being the breeding ground for a variety of unique waterfowl. The local settlement, Fort Chipewyan, has been home to First Nations and Metis communities for ~200 years and played a crucial role in the success of the fur trade industry in Canada during the 1700s to the early 1900s. Important to the community's way of life is the delta's biological productivity and ecological diversity, which are thought to be reliant on water and nutrients from periodic river flood events (PAD-TS, 1996).

Since the early 1970s, there has been increasing concern regarding the ecological impacts of flow regulation of the Peace River that began in 1968, which has implications for the traditional way of life of local residents (Townsend, 1975) and management of the delta. Concerns include changing of the seasonal pattern of Peace River discharge (e.g., PAD-PG, 1973), desiccation of isolated and perched basins (Townsend, 1975), changes in vegetation and wildlife (PAD-PG, 1973), and absence of a major flood event between 1975 and 1995 (PAD-TS, 1996).

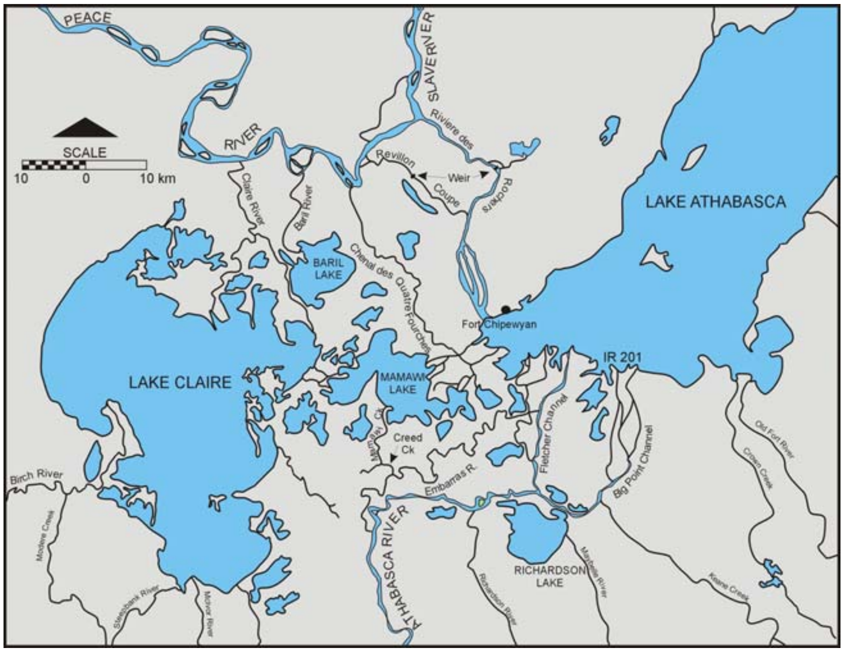


Figure 1. Location and major hydrological features of the Peace-Athabasca Delta, Alberta, Canada.

Several environmental studies have been completed in the PAD over the past few decades to address these concerns, most of them aimed at restoring water levels to perched basins in the PAD. Studies have focussed on evaluating the effects of river impoundment on the Peace River flow regime (PAD-PG, 1973; PAD-IC, 1987; PAD-TS, 1996; Prowse and LaLonde, 1996; Leconte *et al.*, 2001; Peters and Prowse, 2001; Prowse *et al*., 2006), and its effect on ice-jam flood development, frequency and severity during spring break-up of river ice (PAD-TS, 1996; Prowse and Conly, 1998; Prowse *et al.*, 2002; Beltaos, 2003). An important outcome of these studies was the identification of ice-jam flooding as a primary mechanism for water and nutrient replenishment in perched basins, and that damping of the hydrograph of the Peace River, and the absence of large flood events between 1975 and 1995, is likely due to a combination of river regulation and climatic variation. Climatic variation is thought to have increased temperatures during the ice-cover season, reduced snow-pack depths, and caused a change in the intensity and duration of the pre-melt period (Prowse and Conly, 1998; Prowse *et al.*, 2002; 2006). Accordingly, regulation of the Peace River and climate variability signify competing hypotheses to explain what are perceived to be unusually dry conditions in the PAD over much of the last 35 years. In fact, Timoney (2002) introduced the idea that drying of the PAD is only a 'paradigm' that needs more scientifically sound context. In an effort to understand the roles of river flooding and climate on water balances in the PAD, paleoenvironmental research has been conducted (Hall *et al*., 2004; Wolfe *et al*., 2005; Wolfe *et al*., 2006). Overall, these studies indicate that large, multi-decadal fluctuations in flood frequency and intensity have occurred over the past few centuries, and that the recent multi-decadal periods of low flood frequency lie within the range of natural variability.

Unlike the multitude of studies conducted on modelling flow regimes and understanding ice-jam flood development, there have been only a limited number of contemporary hydrologic studies conducted on discrete water bodies (lakes, ponds, open wetlands) of the PAD (Bennett *et al.*, 1973; Pietroniro *et al.*, 1999; Peters, 2003; Wolfe *et al*., 2007b). Although studies that have been conducted on these water bodies have provided important information regarding relationships between river flow and flooding and on the spatial distribution of water balance types, these studies were restricted in their ability to determine the relative importance of the various hydrological processes controlling the water balances of lakes in the PAD. More specifically, these studies were unable to effectively assess the role that recharge processes (including snowmelt, rainfall, catchment runoff, and river flooding) and evaporation have on seasonal, annual, and spatial variability of water balances of lakes. More studies are required to objectively evaluate the importance and effects of recharge processes at the lake-catchment and delta-wide scales. In fact, early on it was recognized that there was a need for more rigorous hydrological assessment of shallow basins to better understand the internal hydrology of the PAD (Townsend, 1975; 1984; Musik, 1991). More recently, Pietroniro *et al.* (1996) and Prowse *et al*., (2006) emphasized the need to understand and properly characterize the unique hydrological processes like snowmelt and ice-jam activity in cold regions like the PAD.

Catchment-sourced water has been modelled as a main input to some lakes in the PAD (Wolfe *et al.*, 2007b) and has been used to explain why a perched basin in the Peace sector of the delta has remained in existence for much of the past ~200 years (Wolfe *et al*., 2005). However, catchment-derived snowmelt and rainfall runoff have been disregarded or considered "minor components of the water balance of perched basins" for some mass balance modelling simulations because of the narrow levees and low-relief surrounding many of the basins (Peters, 2003; Peters *et al*., 2006, pg. 230). The inconsistent view regarding the importance of catchment-derived water on perched basin water balance requires a more rigorous assessment. It is especially important to understand the role of runoff processes on lake water balances with future predictions of higher temperatures, increased evaporation and transpiration, and more variable precipitation patterns for northern Canada (CCCMA, 2006). Indeed, the current deficiency of seasonal, annual, and regional understanding of hydrological processes in the PAD hinders the ability to predict the response of lakes and wetlands to changes in river discharge and climate. Of more recent concern are the potential hydroecological impacts on the southern Athabasca sector of the PAD due to expected expansion of the tar sands projects a few hundred kilometres upstream of the PAD (Parks Canada, 2006). Overall, the diverse hydrological and ecological conditions in PAD make it a model system for evaluating anticipated climate change and river regulation projects in northern Canada, and elsewhere in the world.

The use of classic, topographically-driven models to assess and monitor water balance (PAD-PG, 1973; Gibson *et al.*, 1996; Peters, 2003), or approaches based on remote sensing techniques (Pietroniro *et al.*, 1999), are difficult to apply in low-relief (generally <2m), expansive areas with complex drainage networks, such as the PAD. For example, acquiring high-resolution topographic data for ground 'truthing' of these models is field-intensive and costly. There has been some success with combining Radar and satellite imagery to map water and flood extents in the PAD (Töyrä *et al.*, 2001, 2003; Töyrä and Pietroniro, 2005), although these methods have been complicated by extensive emergent aquatic plant communities, which tend to obscure identification of standing waters from satellite imagery; and by backscatter caused by cloud-cover, ice and waves on water surfaces in the spring. In addition, major solute and nutrient tracers, which have been used previously in northern environments to assess the effects of hydrological change, are complicated by contact with the substrate and are sensitive to the time of sampling (Marsh and Hay, 1989; Lesack *et al.*, 1998). Alternatively, analysis of water isotope tracers ($\delta^{18}O$, $\delta^{2}H$) is a practical alternative to assess the hydrology of lakes in a remote environment like the PAD because a rapid survey of many water bodies can be obtained without the need for field-intensive studies (Gibson, 2002). In addition, stable isotopes preserve hydrological information well after other traditional tracers have lost their sensitivity. Stable isotopes of water have been used to describe temporal and spatial variability in the water balance of basins in other regions of northern Canada (Gibson *et al.*, 1993, 1996; Gibson, 2001; Gibson and Edwards, 2002; Edwards *et al.*, 2004).

Wolfe *et al.* (2007b) used water isotope data in combination with water-chemistry data taken at the end of a thaw-season from 57 shallow basins to estimate the water balance of lakes in the PAD during a year not marked by spring ice-jam flooding. The water balances were described as evaporation-to-inflow ratios using an isotope-mass balance model tailored to accommodate basin-specific input water compositions. Evaporation-to-inflow ratios in conjunction with limnologic data were then used to classify the basins into four drainage types (largely with respect to their degree of connectivity with rivers as described in the 'basin hydrology' section of the Site Description chapter below). A new, fourth drainage type, which was dominant in the central portion of the delta, included very shallow, thaw-season precipitation-dominated basins. Spatial representation of evaporation-to-inflow ratios clearly depicts the "diluted" basins in the Athabasca sector from the "evaporative" basins in the Peace sector. The strong relationships identified between water balances and limnological conditions suggests that past and future changes in hydrology are likely to be coupled with changes in water chemistry and, hence, the ecology of the aquatic environments in the PAD. Although this study demonstrated that water isotope tracers are a sensitive tool for assessing hydrology in this complex environment, especially quantification of water balances, more research is needed to quantify and assess the relative importance of the various hydrological processes in the PAD over seasonal, annual, and spatial scales. It is especially important to understand the role of spring melt processes, like snowmelt and river flooding, because this period sets the stage for the remainder of the thaw-season. More detailed understanding of hydrological processes affecting lakes in the PAD will facilitate interpretations of ongoing monitoring and process-based hydroecological investigations, as well as the reconstruction of site-specific hydroecological histories from multi-proxy lake sediment records (Wolfe *et al*., 2005; 2007a).

This thesis is part of an ongoing multi-disciplinary project including seasonal and annual hydrological and ecological process studies and paleolimnological investigations on the various lake types in the PAD. The aim of this larger project is to gain a better understanding of the past and present hydrology, ecology, and climate of the PAD. This thesis addresses several objectives of the larger project by helping to 1) interpret historical hydrological and ecological information from sediment cores extracted from lakes, 2) understand the complex temporal and spatial interactions between hydrology and limnology of lakes, 3) develop strategies for future management of the PAD under anticipated climatic and river discharge shifts, and perhaps most importantly, 4) improve the understanding of hydrological processes controlling lake water balances in the PAD. In fact, the methods of assessing and quantifying hydrological processes used in this thesis offer an alternative or supplemental means to investigate complex hydrological situations in almost any environment.

Research objectives and outline

The major objective of this thesis is to use information gleaned from water-isotope data to differentiate the relative importance of key hydrological processes that control temporal and spatial variability in lake water balances in the PAD. To meet the larger objective, the main body of this thesis was segregated into three main chapters.

The purpose of the first main chapter (Chapter 4) is to develop an appropriate isotope framework in δ^2H -$\delta^{18}O$ space that will be used in subsequent chapters of this thesis, including for 1) estimation of the isotopic input composition of input water following changes in lake water balance, 2) evaluation of the importance of snowmelt, rainfall, flooding, and evaporation on the water balance of lakes, and 3) differentiation of snowmelt-dominated lakes from river-flooded lakes.

To understand the seasonal and annual behaviour of lakes in the PAD, the purpose of the second main chapter (Chapter 5) is to qualitatively and quantitatively assess the influence of various hydrological processes (including evaporation and input sources: snowmelt, river-water, and thaw-season precipitation) on two hydrologically-contrasting lakes. The first study lake is used to assess snowmelt and rainfall since it is a closed-drainage lake perched in the relict Peace sector, surrounded by bedrock and wetlands. The second study lake is used to assess spring and summer river flooding in the low-lying, fluviodeltaic landscape of the Athabasca sector because it has frequent connection to a distributary channel of the Athabasca River during periods of elevated discharge.

Finally, to assess spatial variability of water balances across space, the purpose of the third main chapter (Chapter 6) is to develop an effective method to quantitatively and visually assess the spatial impact of river flooding and snowmelt on lakes during spring snowmelt. A secondary objective of this chapter is to determine if there is a correspondence between the main input source to a lake and previously-identified drainage classification of the lakes under normal flow conditions (Wolfe *et al*., 2007b).

Although each main chapter individually addresses important hydrological elements of the PAD, collectively they explore how typical hydrological processes in northern Canadian environments affect water balances of shallow lakes over time and space. Improved understanding of hydrology in the PAD will help address complex environmental concerns that are common in the north, especially regarding the impacts of climate change and river water use. It is anticipated that research findings will provide decision-makers with valuable information concerning future environmental stewardship of the PAD and other northern ecosystems.

2.0 SITE DESCRIPTION

The Peace-Athabasca Delta (PAD), northern Alberta, Canada (59°N and 112°W), occupies a flat-lying area of about 3900 km^2 (Figure 1). The PAD lies mostly within the confines of Canada's largest national park, Wood Buffalo National Park (WBNP), at the confluence of the Peace, Athabasca, and Birch rivers. The following provides information regarding the hydroclimate, geological setting, hydrology, and ecology of the PAD.

Hydroclimate

Located within the sub-humid, mid-boreal ecoclimatic region (EWG, 1989), the climate of the PAD is characterized by long, cold winters and short, warm summers. Based on climate normals from 1971 to 2000 from the Environment Canada weather station located at Fort Chipewyan, Alberta, mean annual air temperature is -1.9°C, with mean monthly minimum and maximum of -23.2°C and 16.7°C in January and July, respectively (Environment Canada, 2005). Total precipitation averages ~390 mm annually, with about 60% falling as rain during the ice-free season, which typically extends from early May to late October.

Geological setting

Development of the PAD began ~10,000 years ago following the recession of the Laurentide Ice-Sheet and creation of Glacial Lake McConnell. Through continued isostatic rebound, runoff from the eastern Cordillera and northern Great Plains, and sedimentation by major contributing rivers, the PAD was formed (Smith, 1995). The main tributaries include the Athabasca River to the south, the Peace River to the north, and the Birch River to the west. These rivers converge at the western arm of Lake Athabasca and are drained by a myriad of active and transitory channels to form an extensive fluvio-deltaic plain with a mosaic of terrestrial, wetland, and lake habitats. The Peace sector of the delta lies to the north and is a relict, bedrock-laden fluviodeltaic landscape that only receives widespread flooding from river water during large ice-jam flood events on the Peace River. The Athabasca sector to the south is low-lying and is actively prograding into central lakes and Lake Athabasca, and receives frequent flood inputs from the Athabasca River and its distributaries. The PAD straddles the contact between eroded and exposed Precambrian igneous and metasedimentary gneisses and granites of the Canadian Shield in the east, and mostly buried flat-lying Paleozoic limestone to the north, gypsum in the west, and sandstone to the southeast (PADPG, 1973; Alberta Geological Survey, 2006). The topography of areas underlain by Canadian Shield rocks is irregular and knobbly with local relief of around 50 m. Most of the surface of the delta (except for exposed bedrock) is mantled by Gleyed Cumulic Regosol, which is a sticky clay soil that lacks well-developed horizons and is characteristic of wet environments from fluvial activity (Natural Regions Committee, 2006). However, deposits of very-fine to medium-grained sands are found deposited in the active and inactive areas of the Athabasca sector (Peters *et al*., 2006)

Hydrology

Annual open-water evaporation is between 400 and 500 mm (den Hartog and Ferguson, 1978). The presence of an active layer from permafrost is limited because of the moderating effects imposed by large, nearby water bodies on air and soil temperatures

(Peters, 2003). The quantity of regional groundwater from the nearby Birch and Caribou Mountains is minimal because soft marine shale, having very low permeability, extends from the bed of the Peace River to the mountaintops (Nielson, 1972). Groundwater flow is also restricted in most catchments owing to low gradients and fluvial sediments of low hydraulic conductivity ($K=1.0\times10^{-8}$ to 1.0×10^{-10} ms^{-1}; Peters, 2003; Nielson, 1972) due to occasional sediment-laden deposits of silt and clay from floodwaters. Subsurface transfer between water bodies has been found to be negligible (Nielson, 1972; Peters, 2003). However, there may be cases where lakes are surrounded by contributing areas that support lateral movement of water (i.e., forests), and these lakes may receive some local soil or groundwater.

The fluvial hydrology of the PAD is highly complex because of its very low topographic relief, and highly variable water levels and discharges of the Peace and Athabasca rivers. The landscape is marked by intricate patterns of superimposed active and inactive channels, meander scrolls commonly bordered by densely forested levees, and numerous meadows interspersed with thousands of shallow lakes and wetlands. Described below are general features of the river and basin hydrology of the PAD.

River hydrology

Both the Peace and Athabasca Rivers originate from the eastern Rocky Mountains (Figure 1), possessing flow regimes that are strongly driven by spring snowmelt. Peak discharges usually occur between late April and early May during ice break-up, and again between late May and early June in response to upstream snowmelt runoff in elevated regions (Prowse and Lalonde, 1996). For most of the year, the Peace River bypasses the north end of the PAD, continuing north into the Slave River. As referenced at Peace

Point, the Peace River's estimated drainage area and mean annual discharge are ~293,000 km^2 and 2109 m^3/s, respectively (Peters and Prowse, 2001). Since 1968, Peace River flows have been partially controlled by the W.A.C. Bennett Dam in northern British Columbia.

The main tributary of the PAD is the Athabasca River. Unlike the Peace River, the Athabasca is unregulated and follows a more typical discharge pattern for high latitude rivers, whereby discharge is highest in the spring and lowest in the winter. As referenced from Fort McMurray, the Athabasca River's estimated drainage area and mean annual discharge are ~133,000 km^2 and 656 m^3/s, respectively (PAD-TS, 1996). In recent decades, an increasing proportion of Athabasca River water has flowed northward directly into the delta due to an engineered cut-off of a meander called the Athabasca Cut-off (1972), and a later natural bifurcation of the Embarras River called the Embarras Breakthrough (1982).

Discharge in the channels flowing through the Peace sector of the delta (namely Rivière des Rochers, Revillon Coupé and Chenal des Quatre Fourches) is proportional to the difference in water levels between the open lakes (i.e., Mamawi Lake, Lake Claire and Lake Athabasca) and the Peace River. Under normal flow conditions these lakes drain northward into the Peace River. However, temporary flow reversals can occur if the Peace River rises sufficiently above the level of the western arm of Lake Athabasca. This is commonly associated with excessive discharge, or backwater flooding behind spring ice-jams on the Peace River during the spring break-up period (Andres, 1995). If flooding is extensive enough, much of the relict landscape of the Peace sector of the PAD can become inundated with river water. Flooding of areas in the active Athabasca sector is more common because much of the Athabasca sector is low-lying as compared to the relict, high-relief Peace sector. The flooding that results, which has been known to last up to 47 days (Leconte, 2001), is thought to be important for replenishing water and nutrients in perched basins within the PAD (Prowse and Lalonde, 1996; PAD-TS, 1996). However, it has recently been shown that a perched lake in the PAD which has not received frequent flood waters for over 200 years has managed to maintain both water and productivity (Wolfe *et al*., 2005), apparently because catchment runoff offsets evaporation at this lake. More research is required to fully understanding the amount and type of runoff that may play a role in regulating water loss in perched basins in the delta. The last delta-wide ice-jam-induced floods occurred in the springs of 1996 and 1997.

Basin hydrology

Lakes within the delta have an average depth of 1.5 m and surface areas ranging from 13 to 1980 ha (Carter, 1996). Lake bottoms are generally lined with a partially decomposed organic matter and are underlain by fine-grained clays and silts. Conceptually, delta lakes can gain water from direct precipitation, catchment runoff, and/or river flooding; and lose water through evaporation and, in cases where the lake bottom is sandy, seepage. However, each of these fluxes has differing relative importance depending on the nature and frequency of river connection, variations in catchment and basin characteristics (i.e., volume, substrate), and local hydroclimate. In perched basins, where evaporative flux plays an important role in the water budget, it is estimated that thaw-season water levels can decrease due to evaporation by 1 to 6 mm/day, or ~390 mm over the thaw-season (Peters, 2003).

Lakes are situated at varying elevations and distances from river channels, and thus span a broad hydrologic continuum ranging from permanently 'open-drainage' basins that are hydrologically connected to river channels under all flow conditions, to 'closed-drainage' basins that are influenced by river waters only at times of extensive overland flooding (Figure 2). A broad intermediate class called 'restricted-drainage' basins have intermittent channelized or overbank river connection during elevated flow conditions, including those caused by ice-jams and snowmelt, excessive precipitation or water released from the Bennett Dam.

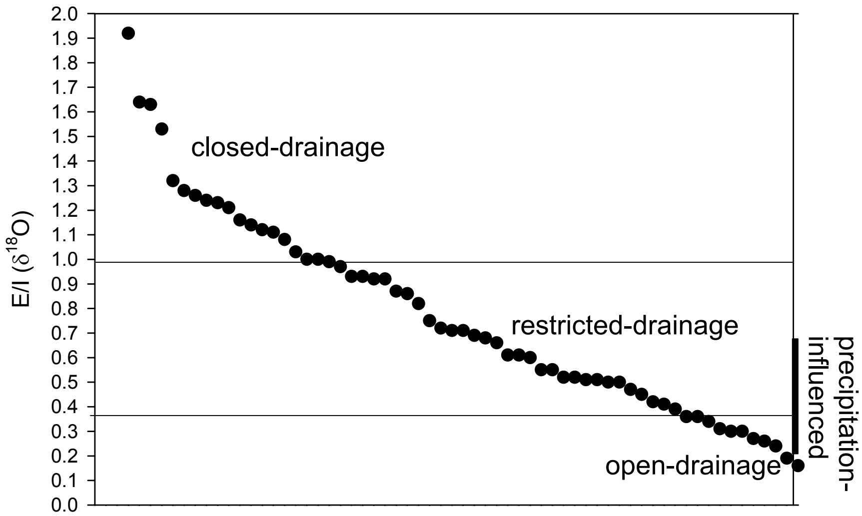


Figure 2: Categorization of drainage regimes of PAD lakes in October 2000

Adapted from Wolfe *et al.* 2007b. Drainage regimes are defined by connectivity to the main river system and were based on evaporation-to-inflow (E/I) ratios calculated using an isotope-mass balance model from water samples collected in October 2000. The various drainage types identified for basins in the PAD include: 'closed-drainage', which are evaporation-dominated and only influenced by river waters at times of extreme overland flooding; 'open-drainage', which are perennially connected to rivers under all flow conditions (e.g., Mamawi Lake); 'restricted-drainage', which are connected to rivers by ephemeral channels or over-bank flooding during high-water conditions; and 'precipitation-sensitive', which are shallow lakes highly sensitive to inputs by thaw-season precipitation.

This operational classification of basins in the PAD based on river-connection status was first used by early technical studies conducted in the PAD (PADPG, 1973), and has since been used by many researchers (e.g., Pietroniro *et al.*, 1999; Wolfe *et al.*, 2007b). Wolfe *et al.* (2007b) found that open-drainage basins are located in the interior of the delta, where several large rivers and creeks discharge; most closed-drainage basins populate the irregular, bedrock-laden topography of the Peace sector of the delta; and most restricted-drainage basins are located in the active Athabasca sector where the landscape periodically floods from Athabasca River and its tributaries. Wolfe *et al.* (2007b) also identified a fourth classification consisting of very shallow ponds that are highly sensitive to precipitation, which were located near the centre of the delta close to the open-drainage lakes (Figure 2). These four classifications of lakes will be used when discussing the study lakes in this thesis.

Ecology

The PAD has a rich biodiversity due to various factors including periodic nutrient recharge from flood waters, resuspension and recycling of nutrients, strong seasonal changes in climate, and highly fertile soils. At least 250 species of vascular plants, 215 species of birds, 44 species of mammals, 18 species of fish, and thousands of species of insects and invertebrates are found in the delta (Wetlands International, 2005). Grasses, sedges (*Carex*), and willows (*Salix*) are the dominant vegetation types in the active and semi-active portions of the PAD, similar to that of the widespread fens and bogs located within the Continental High Boreal wetland region of Canada (Timoney, 2002). Coniferous (mainly *Picea* and *Pinus*), deciduous (mainly *Alnus* and *Populus*), and mixed-forests grow on bedrock outcrops and along high levees in the less active areas.

Beds of *Equisetum fluviatile* grow along the river edges, and stands of *Phragmites australis* grow in areas that are frequently flooded. The shallow open-water ponds are dominated by extensive macrophyte communities including *Potamogeton* spp., ("pondweeds"), *Lemna* spp., *Utricularia vulgaris* L., and *Ceratophyllum demersum* L. (Timoney, 2002).

The following section provides necessary background information on isotope hydrology, including a description of an isotopic framework and how it is generally developed, and isotope and mass balance equations that are used throughout the manuscript. The section ends with an explanation of how water samples were collected and analyzed for isotopic analysis.

3.0 ISOTOPE HYDROLOGY - BACKGROUND AND SAMPLE COLLECTION

Isotope hydrology background

The use of stable oxygen and hydrogen isotopes (namely ^{18}O and ^{2}H) as tracers in hydrological studies has expanded over the past 50 years, much of this due to the development of a robust, physically-based understanding of isotopic partitioning (fractionation) in the water cycle (e.g., Edwards *et al.,* 2004), and advancements in instrumentation. To a great extent this understanding stems from the initial explanation of systematic variations in isotope compositions of global precipitation (Craig 1961; Dansgaard, 1964), development of the theory describing isotopic fractionation during evaporation (Craig and Gordon, 1965), and testing and validation of isotope modelling under an array of field conditions, which are commonly compared with conventional methods (e.g., Gat, 1981, Clark and Fritz, 1997; Kendall and McDonnell, 1998; Gibson and Prowse, 1996; Gibson *et al.* 1993, 1996, 1998, 2002).

Water isotopes have been especially useful to investigate runoff generation processes and lake water balance in strongly seasonal environments (Gat, 1970; Gibson *et al*., 1993; Gibson and Edwards, 2002; Wolfe *et al*., 2007b; Brock *et al*., 2007). Although measurement of only one water isotope provides important information, more detailed information of various hydrological processes can be gathered by superimposing both water isotopes in $\delta^{2}H$–$\delta^{18}O$ space against the backdrop of an isotopic framework. The key components of this framework include two reference lines representing the isotope labelling of both precipitation and evaporating surface water bodies. These lines exist because of systematic mass-dependent partitioning of water isotopes in the water cycle. The formation of the framework is described in further detail below.

The isotopic framework: isotopic labelling of precipitation and evaporating surface waters

The Local Meteoric Water Line (LWML)

Regression of $\delta^{18}O$ and $\delta^{2}H$ values obtained from weighted-monthly samples at several hundred stations worldwide (Rozanski *et al*., 1993) cluster in a linear array, termed the Global Meteoric Water Line (GMWL), which is closely approximated by $\delta^{2}H = 8\delta^{18}O + 10$ (Craig, 1961; Dansgaard, 1964). The GMWL exists because atmospheric moisture arising from the sub-tropic ocean surface undergoes progressive rainout of mass and heavy isotopes during subsequent inland and pole-ward transport. Both the linearity and slope of the GMWL reflect the predominant influence of temperature-dependent equilibrium partitioning of the two heavy isotope species between atmospheric vapour and condensing precipitation. On a more local scale, a regression line through amount-weighted $\delta^{18}O$ and $\delta^{2}H$ values of precipitation from a particular locale forms a Local Meteoric Water Line (LMWL) (Figure 3). Slight variations of a LMWL from the GMWL arise from differing histories of distinct regional air masses including moisture sources and recycling. At high latitudes, precipitation generally has a seasonal isotopic cycle, where winter precipitation is generally more depleted in heavy-isotope content compared to rainfall during the thaw-season.

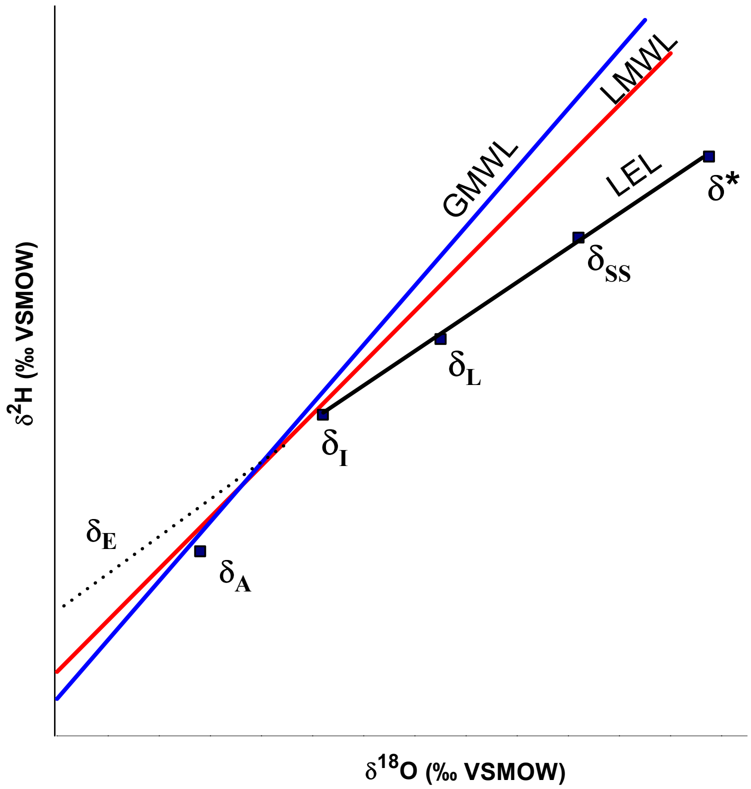


Figure 3: Isotope labeling in δ^2H-δ^{18}O space of water balance components for a hypothetical high-latitude environment.
The isotopic composition of the evaporating lake water (δ_L) at the time of sampling is offset from the mean annual input composition (δ_I) along a Local Evaporation Line (LEL) having a slope around five. The evaporative flux from the lake (δ_E) is modelled using the Craig and Gordon model (1965) because it cannot be sampled. Mass conservation dictates that δ_E will lie on the extension of the LEL to the left of the Local Meteoric Water Line (LMWL). Evaporation and exchange occur in association with ambient atmospheric moisture (δ_A). Also shown is the isotopic composition of a terminal lake in isotopic and hydrologic steady-state (δ_{SS}), and the limiting isotopic composition (δ^*), which is the maximum achievable heavy-isotope enrichment that a water body can reach under prevalent local atmospheric conditions.

The Local Evaporation Line (LEL)

A strong correspondence in both δ^{18}O and δ^2H variations is also evident from the development of characteristic heavy-isotope build-up in neighbouring surface waters that have undergone evaporation (Figure 3). Although each evaporating water body in a particular catchment possesses a unique evaporation line that deviates to the right of the LMWL (along a slope in the range of four to six in northern environments), it is useful to create a general, catchment-wide Local Evaporation Line (LEL) as a reference line for study purposes. This catchment-wide LEL is commonly defined by a linear regression through the isotopic composition of evaporating lake waters taken over some time of interest (Leng and Anderson 2003; Maric, 2003; Diefendorf and Patterson, 2005). Unfortunately retrieving sufficient timely and spatially representative data from lakes in northern environments can be difficult because of its seasonality and vastness. Alternatively, the LEL can be calculated based on a best-fit line through the isotopic composition of average input water to a catchment (δ_I), and the limit of isotopic enrichment (δ^*) a lake can undergo under existing atmospheric conditions (Gibson and Edwards, 2002; Wolfe *et al.*, 2007b; Brock *et al.,* 2007). The formation of this 'theoretical' or 'predicted' LEL is based on the idea that a lake would move along the LEL and eventually reach δ^* in the absence of δ_I. δ^* is predictably controlled by local atmospheric conditions during the ice-free season, principally relative humidity and the isotopic composition of ambient atmospheric moisture (δ_A).

Superimposing lake water isotope compositions on a δ^2H-$\delta^{18}O$ diagram containing the LMWL and LEL allows for the assessment of water balance controls on lakes in a particular region. For example, departure of lake water isotope composition *above* or *below* the LEL can provide information regarding recent contributions by various source waters. This is because upon receiving new water from a different source, a lake will deviate towards the isotopic composition of that source water. Alternatively, displacement of a given lake's isotopic composition *along* the LEL (away from δ_I) provides an index of water balance change due to evaporation. Highly useful information regarding a lake's hydrologic status at a particular point in time is also gained by assessing the position of lake water isotope composition versus key isotopic datums: namely δ_I, δ_{SS}, and δ^*. The following reviews the isotope-mass balance equations, and associated considerations, that will later be used to evaluate lake water balance variations based on analyzed lake water samples.

Isotope and mass balance equations

If it is assumed that the density of water remains constant, the water- and isotope-mass balance for a well-mixed reservoir may be written, respectively, as:

$$\frac{\mathrm{d}V}{\mathrm{d}t} = \mathrm{I} - \mathrm{Q} - \mathrm{E} \tag{1}$$

$$\frac{\mathrm{d}(V\delta_\mathrm{L})}{\mathrm{d}t} = \mathrm{I}\delta_\mathrm{I} - \mathrm{Q}\delta_\mathrm{Q} - \mathrm{E}\delta_\mathrm{E} \tag{2}$$

Where, V is the volume of the reservoir at time t; dV is the change in volume over the time interval dt; I, Q, and E are rates of inflow, outflow, and evaporation, respectively; δ_L is the isotopic composition of the evaporating liquid; δ_I is the weighted-mean isotopic composition of input waters to the system; δ_Q the isotopic composition of the liquid outflow; and δ_E is the isotopic composition of water vapour evaporating from the surface of the system. Most of these parameters can be measured directly, except for δ_E which, therefore, must be estimated using indirect measures. The most common way of estimating δ_E is through application of the layered resistance model of Craig and Gordon (1965). This model has been well-validated under a range of environments and conditions (Gat and Levy, 1978; Gat, 1981; Gibson and Edwards, 2002), and has been especially useful for hydrological studies in northern Canada (Gibson *et al.,* 1993, 1996; Gibson, 2001; Gibson and Edwards, 2002; see also Edwards *et al.* 2004). The model accounts for the differing volatilities of the water isotopologues ($^{1}H_2{}^{16}O$, $^{1}H^{2}H^{16}O$ and $^{1}H_2{}^{18}O$) as they pass through the air-water interface as a combination of mass-dependent variations in equilibrium vapour pressures ("equilibrium effects"), and variations in molecular diffusivities arising from the combination of differing absolute mass and the distribution within water molecules ("kinetic effects"). Assuming negligible resistance to liquid-phase mixing, δ_E can be defined as (Gonfiantini, 1986; Yi *et al.*, 2006):

$$\delta_E = \frac{((\delta_L - \varepsilon^*)/\alpha^*) - h\delta_A - \varepsilon_k}{1 - h + \varepsilon_k} = \frac{\delta_L/\alpha^* - h\delta_A - \varepsilon_T}{1 - h + \varepsilon_k} \quad (3)$$

Where, $\varepsilon_T = \varepsilon^*/\alpha^* + \varepsilon_k$, ε^* and ε_k represent the respective equilibrium and kinetic effects, expressed as decimal fraction (i.e., -0.01, rather than -10‰) differences between the liquid and vapour phase; α^* is the equilibrium liquid-vapour isotope fractionation ($\alpha^* = 1 + \varepsilon^*$); h is the atmospheric relative humidity (ranging from 0 to 1) normalized to the saturation vapour pressure at the temperature of the air-water interface; and δ_A is the isotopic composition of atmospheric moisture. ε^* values are temperature-dependent and can be calculated from known empirical relationships (Horita and Wesolowski, 1994). For $\delta^{18}O$, $1000 ln\alpha^* = -7.685 + 6.7123\ (10^3/T) - 1.6664\ (10^6/T) + 0.35041\ (10^9/T)$ for $\delta^{18}O$, and for δ^2H, $1000 ln\alpha^* = 1158.8\ (T^3/10^9) - 1620.1\ (T^2/10^6) + 794.84\ (T/10^3) - 161.04 + 2.9992\ (10^9/T^3)$, where T represents the interface temperature in degrees Kelvin. The kinetic separation factor, ε_k, is controlled by diffusive transfer mechanisms and can be approximated as a function of the relative humidity deficit based on $\varepsilon_K = C_k\ (1-h)$, where C_k is a parameter describing the relative transport resistance of the heavy and common isotopic species into a fully-turbulent atmosphere. C_k values of 14.2 for $\delta^{18}O$ and 12.5 for δ^2H are typically used (Gonfiantini, 1986; Araguás-Araguás *et al.*, 2000), and are expected to be representative for the open-water evaporation conditions in the PAD.

Substituting δ_E from equation (3) into equation (2), and noting that the isotopic composition of outflow is similar to that of the reservoir (i.e., $\delta_Q=\delta_L$), provides an index of water balance represented by:

$$E/I = (\delta_L - \delta_I)/m(\delta^* - \delta_L) \tag{4}$$

Where m is an enrichment slope adapted from Welhan and Fritz (1977), and Allison and Leaney (1982), and is defined by $m = (h - \varepsilon_k - (\varepsilon^*/\alpha^*)) / (1- h + \varepsilon_k) = (h - \varepsilon_T)/(1- \mathrm{h} + \varepsilon_k)$.

δ^* is the limiting isotopic enrichment attainable by a desiccating water body under prevailing atmospheric conditions, and is defined by (Gat, 1995):

$$\delta^* = (h\delta_A + \varepsilon_k + \varepsilon^*/\alpha^*) / (h - (\varepsilon_k + \varepsilon^*/\alpha^*)) = (h\delta_A + \varepsilon_T) / (h - \varepsilon_T) \tag{5}$$

As with δ_E, the isotopic composition of atmospheric moisture (δ_A) has proven difficult to measure directly due to logistical complications (Gibson *et al*, 1999). However, δ_A can be modelled through the rearrangement of Equation (4), knowledge of h and T, and measurement of the isotopic composition of a local terminal water body in hydrologic and isotopic steady state (δ_{SS}). According to Gonfiantini (1986), Gibson (1996), Gibson *et al.*, (1998; 1999; 2002), the equation to model the isotopic composition of a closed-drainage lake at any given time ($\delta_L(t)$) is:

$$\delta_L(t) = \delta_{SS} - (\delta_{SS}-\delta_o)\exp[-(1+m))(Et/V)] \tag{6}$$

Where δ_o is the initial isotopic composition of the reservoir; Et is the volume of water lost during evaporation in a time-step; V is the volume of the reservoir; and δ_{SS} is the measured or modelled isotopic composition of a water body in hydrologic and isotopic steady state. According to Gibson *et al.*, (1998; 1999; 2002) δ_{SS} can be measured from a local water body (or an evaporation pan) in steady-state, or modelled using:

$$\delta_{SS} = (\delta_I + m\delta^*) / (1+m) \qquad (7)$$

To model δ_A, parameters on the right side of equation (6) are adjusted so that the left side ($\delta_L(t)$) match the known (measured) isotopic composition of a lake in hydrologic and isotopic steady-state. In the case of the PAD, the latter was measured from a reference lake in hydrologic and isotopic steady-state which was assumed to be in equilibrium with atmospheric moisture, as described below. Since $\delta_{o,}$ *m*, Et, and V are constant in equation (6) the only parameter that can be changed is δ_{SS} using equation (7). Since δ_I and *m* are constant in equation (7), this leaves only δ^* to adjust. To adjust δ^*, equation (5) is used. Since ε_T and *h* are constant in equation (5) this leaves only δ_A to adjust. Essentially, each equation has one unknown to model until the equation that includes δ_A is reached. The modelled composition of δ_A is taken once it is adjusted so that $\delta_L(t)$ from equation (6) is equal to the measured value of the reference basin in steady-state. On the isotopic framework this would shift the position of δ_A to correspond with the observed Local Evaporation Line.

As shown above, many of the parameters of the isotope-mass balance equations require real data collected from input sources and evaporating water bodies in the environment of interest. Even modelled parameters like δ_A stem from measured data. Since the difference in isotopic composition between two measured samples is often small, analytical uncertainties must be minimal. Accordingly, the following section describes how water samples were collected from the field and analyzed in a state-of-the-art isotope laboratory.

Isotope sample collection and analysis

Water samples were collected manually from 10-20 cm below the water surface at the centre of 60 lakes, and from the middle of 12 river channels on several occasions between October 2000 and September 2005 for analysis of oxygen and hydrogen isotope composition. Lakes spanned the full range of apparent river connectivity. Water samples were also collected from Lake Athabasca, a local reference lake, an evaporation pan, and input sources (snow, rain, rivers) in order to help characterize key isotope and hydrological components for the PAD. All water samples were collected in air-tight 30 mL high-density polyethylene Nalgene® bottles, and were transported to the University of Waterloo Environmental Isotope Laboratory for stable isotope analysis using conventional techniques (Epstein and Mayeda, 1953; Coleman *et al.*, 1982). Oxygen values were determined using a conventional CO_2-equilibration technique on a VG MM 903 mass spectrometer (Drimmie and Heemskerk, 1993), while deuterium was analyzed using a EuroVector EA3028 elemental analyser interfaced in continuous-flow mode to a GV Instruments (previously Micromass) IsoPrime® Stable Isotope Ratio Mass Spectrometer. Data reduction was conducted using a three-point calibration with three internal standards corrected to international standards, and oxygen and hydrogen isotope compositions were normalized to -55.5 ‰ and -428 ‰, respectively, for Standard Light Antarctic Precipitation (Coplen, 1996). Analytical uncertainties based on sample repeats are ±0.1 ‰ for $\delta^{18}O$ and ±2.0 ‰ for $\delta^{2}H$. Results are expressed using δ–notation, representing a difference in per mil (‰) (i.e., per thousand) from Vienna Standard Mean Ocean Water (VSMOW), such that $\delta^{18}O$ or $\delta^{2}H = [(R_{sample}/R_{VSMOW})-1] \times 1000$; where R refers to the $^{18}O/^{16}O$ or $^{2}H/^{1}H$ ratio in the sample and VSMOW, respectively. VSMOW approximates the isotopic composition of the modern global ocean reservoir (invariant at human time scales), and has $\delta^{18}O$ and $\delta^{2}H$ values both defined by exactly 0‰.

4.0 DEVELOPMENT OF AN ISOTOPIC FRAMEWORK FOR THE PEACE-ATHABASCA DELTA: THAW-SEASON 2003

Synopsis

The objective of this chapter is to assemble data to develop an appropriate isotopic framework for the Peace-Athabasca Delta that is required in subsequent analyses, including qualitatively and quantitatively assessing the influence of hydrological processes (i.e., rainfall, snowmelt, river flooding, and evaporation) on water balance of selected PAD lakes over seasonal and inter-annual time-scales (Chapter 5) and over the delta landscape (Chapter 6). A representative Local Meteoric Water Line (LMWL) and Local Evaporation Line (LEL), which make up the isotopic framework, are formed using measured δ^2H and $\delta^{18}O$ data collected from precipitation and a reference basin, and evaporation-flux-weighted hydrometeorological data. The chapter is organized with respect to how the isotopic framework is developed, with a discussion of temporal and spatial variability of hydroclimate and the isotopic composition of snow (δ_{Snow}), rain (δ_{Rain}), river water (δ_{River}), atmospheric moisture (δ_A), average thaw-season isotopic composition of maximum evaporative enrichment (δ^*), and a terminal lake in isotopic and hydrologic steady state (δ_{SS}).

It was determined that an isotopic framework developed for the 2003 thaw-season is the most appropriate for assessing hydrological processes between 2000 and 2005 because 2003 had 1) the most comprehensive isotopic data of input sources (including snowmelt, summer precipitation, and river water); 2) radiation data to accurately calculate daily evaporation, which are used to evaporation-flux-weight monthly relative humidity and temperature; and 3) evaporation pan data, which are used to assess short-term variations in key isotopic datum values, and the appropriateness of using a local reference basin for modelling purposes. Moreover, it was also found that hydroclimate conditions of 2003 were similar to the average between 2000 and 2005, and that the slopes of the LWML and LEL are very similar to the ones developed for the 2000-2005 time period (using a more limited dataset). The common framework was then developed using a regression line through $\delta^{18}O$ and $\delta^{2}H$ data collected from the various input sources in 2003 (to form the LMWL), and a line from the average input composition of the PAD (δ_I) and δ^*, modelled using a reference basin (to form the LEL). Evaporation-flux-weighted relative humidity and temperature were used in the Craig and Gordon (1965) isotope-flux and fractionation model to calculate δ_A and δ^*.

Superimposing measured isotope data from 11 lakes in 2003 onto the modelled framework determined that lakes received input waters of varying isotopic compositions. Since snowmelt plotted down the LMWL from δ_I, lakes that received recent contributions by snowmelt were found to have plotted below the LEL. On the other hand, since rainfall typically plotted up the LMWL from δ_I, lakes that received recent rainfall were found to have plotted above the LEL. The degree to which a lake deviated away from the LEL was thought to depend on its size and catchment characteristics. An application of the isotopic framework includes modelling the isotopic composition of input water to each lake following a change in lake water balance. This was conducted by taking the intersection point at the LMWL from the isotopic trajectory of lake water before and after a hydrological event (i.e., snowmelt).

Introduction

Coupling of both water isotopes (^{2}H and ^{18}O) has been especially useful to investigate runoff generation processes, evaporation, and lake water balance in strongly seasonal environments (Gat, 1970; Gibson, 2001a; Gibson *et al*., 1993a, 1993b; Gibson and Edwards, 2002; Maric, 2003; Wolfe *et al.*, 2007b; Brock *et al*., 2007). By using both isotopes, more detailed information about the various hydrological processes in a catchment (especially northern environments) can be obtained. Isotope hydrology using both tracers is generally conducted by superimposing both water isotopes from water bodies of interest (i.e., lakes, groundwater, rivers, etc.) in $\delta^{2}H$–$\delta^{18}O$ space against the backdrop of an isotopic framework. However, many isotopic frameworks have been developed using climate normal data or data from outside the study of interest, which may not be appropriate for assessing short-term lake water balance changes in an environment sensitive to hydroclimate such as the PAD. Many of the difficulties associated with developing a representative framework for a northern region have been a direct cause of not having adequate hydroclimate and isotope data and/or reference basins to develop key components of the isotopic framework (i.e. evaporation pan or a local basin in steady-state). The objective of this chapter is to show the importance of acquiring local isotope and hydroclimate data and to assemble this data to develop an appropriate isotopic framework for the Peace-Athabasca Delta. This framework will be used in subsequent analyses, including qualitatively and quantitatively assessing the influence of hydrological processes (i.e., rainfall, snowmelt, flooding, and evaporation) on water balance of selected PAD lakes over inter- and intra-annual time scales (Chapter 5), and across the delta landscape (Chapter 6).

A theoretical isotopic framework for the Peace-Athabasca Delta (PAD) region has been developed by Wolfe *et al.* (2007b) to assess cumulative thaw-season water balance conditions of lakes in 2000. However, the researchers were confined to limited hydroclimate and isotope data, and were unable to assess variations in key isotope datums on the framework. For example, they used long-term hydroclimatic data (i.e., 1971-2000 climate normals), an interpolated value of rainfall based on a network of precipitation sampling stations over hundreds of kilometres away, an isotope composition of atmospheric moisture based on the assumption that moisture is in equilibrium with average summer rainfall, and a LMWL from Fort Smith, Northwest Territories. Although these approximations were justified given their set of objectives and results, the methods may not be suitable for assessing seasonal and annual changes in the isotopic composition of the PAD lakes because atmospheric conditions and associated isotopic parameters (especially δ_A and δ^*) are known to vary in such northern environments (Gibson *et al*., 1999).

The most significant difference from Wolfe *et al*, (2007) regarding the development of an isotopic framework for the PAD is the construction of the LEL. Specifically, the key isotope parameters used to form the LEL are determined using hydroclimate data from 2003 (a year within the range of years being investigated), and modelled isotope datums using a local terminal basin thought to be in isotopic and hydrologic steady state. The use of a reference lake (also known as an index lake) is recommended for establishing and assessing local isotopic parameters needed for isotope-mass-balance modelling (Dincer, 1968; Welhan and Fritz, 1976). The techniques and values used to develop the isotopic framework are compared with other conventional methods and values from other isotope-based studies conducted on sites with similar climatic conditions as the PAD (e.g., southern Northwest Territories, Gibson *et al*., 1998).

In addition to developing an isotopic framework for use in later chapters, this chapter also seeks to address variability and sensitivity of key isotope parameters that lie on the isotope framework, and to discuss some of the uses of the isotopic framework. Thus, the chapter is organized with respect to how the isotopic framework is developed, with consideration of the variability in the data and the components that make up the framework, including hydroclimate and the isotopic composition of snow (δ_{Snow}), rain (δ_{Rain}), river water (δ_{River}), atmospheric moisture (δ_A), the limit of evaporative enrichment (δ^*), and a terminal lake in isotopic and hydrologic steady state (δ_{SS}). The chapter ends with a discussion of how the isotopic framework is used to constrain the isotopic composition of input water following a hydrological event that affects the lake (i.e., a rainstorm or snowmelt), a previously difficult task for isotope-based hydrological studies. This is done by taking the intersection point at the LMWL from the isotopic trajectory of a lake following a hydrological event (leading to isotopic dilution). In other words, a line is taken from the isotope composition of a lake before and after a hydrological event, and the point at which this line intersects the LMWL is the isotopic composition of the input water. The ability to determine the isotopic composition of a *specific* hydrological event is an alternative to Wolfe *et al*. (2007b) who utilized a conservative means of assessing lake-specific *average* thaw-season input water by taking the intersection point on the LWML from a parallel line to the predicted LEL from the isotopic composition of a lake at the end of the 2000 thaw-season. Taking the intersection point using a direct isotopic trajectory may be an alternative means for assessing recent contributions. The isotopic composition of input water may be more precise than taking a parallel line along the LEL since isotope-mass balance considerations stipulate that a lake undergoing evaporation should follow an isotopic trajectory towards δ* from the isotope composition of recent input water (Figure 3).

Methods

Hydroclimatic data collection and processing

In 2003, daily relative humidity (*h*), air temperature (T), and precipitation were measured in Fort Chipewyan by standard instrumentation maintained by Environment Canada (Environment Canada, 2005). Monthly *h* and T were flux-weighted to evaporation as recommended by Gibson *et al.* (2005) for seasonal environments. Use of the evaporation-flux-weighted approach takes into account the differences in monthly evaporation rates. Monthly evaporation rates were tabulated based on daily evaporation rates (Appendix A) estimated using the Priestly and Taylor (PT) method (1972). The PT method has been tested and validated as an appropriate means of estimating the evaporation rates of small lakes and wetlands in northern Canada over periods of weeks and months (Roulet and Woo, 1986; Gibson et al., 1996; Peters, 2003). The PT method is largely driven by the available energy (i.e., incoming radiation) and is expressed as:

$$E = \alpha \frac{s(T_a)}{s(T) + y}(K_n + L_n)\frac{1}{\rho_w \lambda_v} \tag{8}$$

where E is evaporation (cm), α is a coefficient (an evaporability factor) relating actual to equilibrium evaporation, s(Ta) is the slope of the saturation-vapour pressure curve at air temperature (mB/°C), γ=0.66 is the psychrometric constant, K_n= short-wave radiation (cal/cm^2), L_n is long-wave radiation (cal/cm^2), ρ_w=1 is the density of water (g/cm^3), and λ_v=597.3-0.564T_a is the latent heat of vaporization (cal/g). The evaporability factor, α, is taken as 1.26 as suggested originally by Priestly and Taylor (1972) for mean evaporation from large, saturated land surfaces. This value has been verified for small lakes and wetlands in northern Canada (Roulet and Woo, 1986; Gibson et al., 1996), and more specifically, for the PAD for the thaw-seasons of 1995 and 1996 based on measurements taken beside an open-water pond (Peters, 2003). Radiation data for the PT calculations in this thesis were obtained from a weather station maintained in Fort Chipewyan by the Wood Buffalo Environmental Association (WBEA, 2005). Daily temperature, radiation, and evaporation are found in Appendix A.

Local Evaporation Line parameter assessment

To develop the upper-end of the LEL, δ^* was modelled using the measured isotopic composition of a reference lake in hydrologic and isotopic steady state (δ_{SS}), which are further explained below. A reference lake must be large enough in volume such that there is sufficient isotopic inertia to buffer short-term variations in atmospheric conditions. Use of the 'index' lake method has been shown to be highly valuable for assessing key isotopic parameters over annual periods (Dincer, 1968; Welhan and Fritz, 1976). This technique assumes that the nearby lakes are affected by the same atmospheric and exchange processes.

Data from an evaporation pan were also assessed to understand short-term changes in δ_{SS}, δ_A, and δ^*, and how variations of these parameters may affect evaluation of lake water isotope data. The evaporation pan was maintained in the backyard of the research facility in Fort Chipewyan between May 21st and August 18th, 2003. Evaporation pans have been shown to be reliable for estimating unknown water balance parameters (Gat, 1970; Welhan and Fritz, 1977; Gibson, 1999), specifically flux-weighted estimates of δ_A for periods of several days to weeks (Allison and Leaney, 1982). The fact that pan evaporation rates may not be identical to natural lake evaporation rates is not a problem, providing that mass transfer mechanisms are similar and differences in water temperature are accounted for (Gibson *et al.*, 1999). Details regarding the characteristics and maintenance of the index lake and evaporation pan are provided below.

Greenstar Lake

Greenstar Lake (GSL; informal name; 58°53.72' N, 111°21.45' W) is a comparatively deep ($Z_{maximum}$ = 9.2 m), closed-drainage basin with a surface area of ~30 ha (Figure 4). GSL is underlain by, and perched on, Canadian Shield bedrock in the Peace sector of the PAD. The lake basin has steep-sided morphometry such that littoral areas are generally restricted to a very narrow margin around the lake circumference. Water marks <20 cm above the current water level on shoreline rocks, and the presence of a continuous mature conifer forest surrounding the lake, suggest that the lake has not received flood water in recent years. SURFER 8.0® (Golden Software Inc.) was used to construct a bathymetric contour map and to estimate lake volume and lake area (Figure 4). These characteristics were based on GPS mapping around the lake's circumference, and water levels taken from several locations. Because GSL has at least a decadal residence time (due to its depth and closed-drainage), this basin will likely approximate the long term average isotopic composition of a terminal basin in hydrologic steady state (i.e., δ_{ss}), an isotopic datum that is useful for subsequent evaluation of lake water isotopic values elsewhere in the PAD.

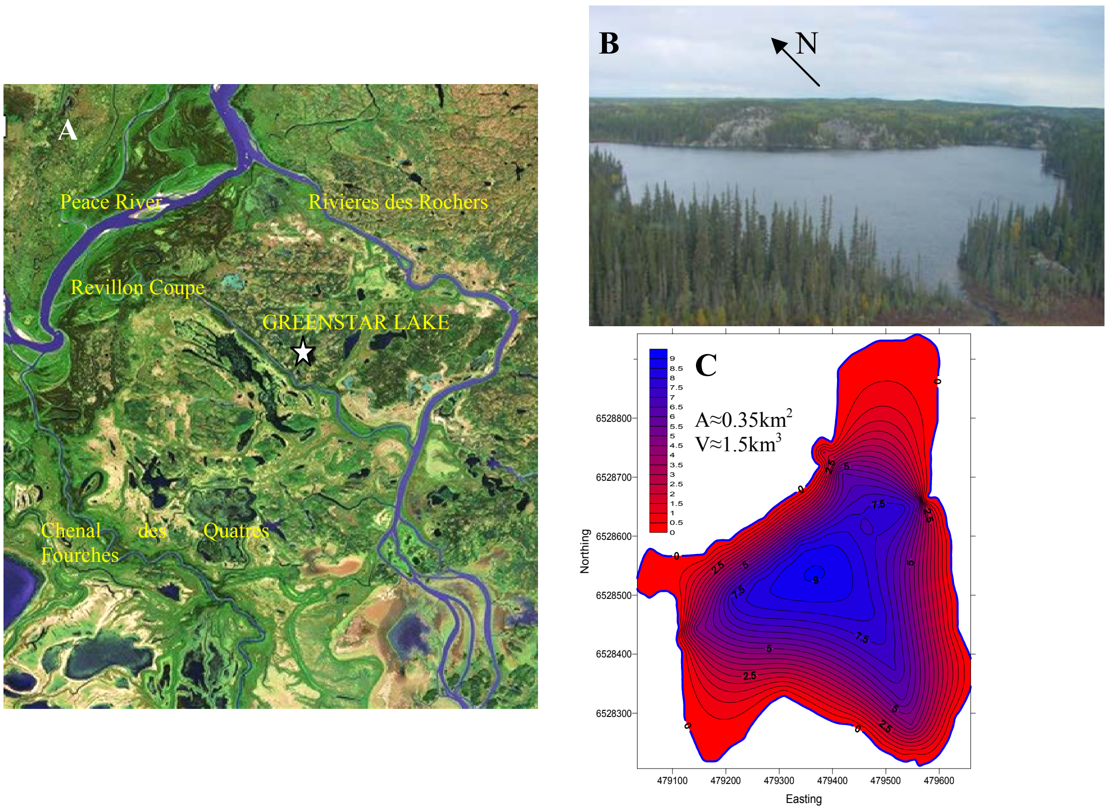


Figure 4: Characteristics of Greenstar Lake, a local basin in hydrologic and isotopic steady-state.

a) location relative to major river channels (obtained from GoogleEarth®), b) photograph of GSL taken in July of 2003, and c) bathymetry based on depth measurements between June and August of 2003. Bathymetry is based on average water depths of 2003.

Evaporation pan

The evaporation pan is a steel cylinder approximately 25 cm deep and 123 cm in diameter (220 L capacity; Appendix B). The water level was maintained at 19 cm on a fixed-point gauge by adding or removing a measured volume of water after evaporation or precipitation occurred. Pan and tap water samples were collected every 1 to 2 weeks for isotopic analysis. The isotopic composition of input water to the pan was determined through weighting the amount of rainwater or tap water the pan needed to offset evaporation in a certain time period, and by taking into account the isotopic composition of rainwater and tap water, the latter of which remained constant (i.e., only changed by ±0.4‰ for $\delta^{18}O$ and ±3‰ for $\delta^{2}H$). Evaporation only reduced pan depth up to a daily maximum of 1.58 cm (6% of original depth), which would have negligible quantitative effects in isotope-mass balance calculations. The pan was used for addressing short-term variations of δ_A, δ^*, and δ_{SS}.

Results and Discussion

After reviewing all available hydroclimate and isotope data between 2000 and 2005, it was decided that an isotopic framework based on data from the thaw-season of 2003 would be the most appropriate for assessment of lake-hydrology between 2000 and 2005. Thaw-season 2003 had 1) radiation data to accurately calculate evaporation, which will be used to evaporation-flux-weight relative humidity and temperature as recommended by Gibson *et al.* (2005) for seasonal environments; 2) the most temporally and spatially representative isotopic data of input sources (including snowmelt, summer precipitation, and river water); and 3) evaporation pan data, which will be used to assess short-term variations in key isotopic datum values and the appropriateness of using a local reference basin for modelling purposes. In addition, average climate conditions of 2003 were very similar to the average between 2000 and 2005, as discussed below.

Hydroclimate: 2003

Precipitation

Total thaw-season precipitation during 2003 (246 mm) was similar to the average total between 2000 and 2005 (232 mm) and climate normals (247 mm). However, monthly precipitation totals for 2003 are generally different than the average between 2000 and 2005 and climate normals (Figure 5). For example, July and August 2003 received only 66% and 64%, respectively, of the average between 2000 and 2005, whereas October received 260%. The lower than average precipitation in mid-summer, and higher than average rainfall in autumn 2003, are discussed when assessing isotope data from two hydrologically-contrasting lakes in Chapter 5. However, since the total precipitation in the thaw-season of 2003 was similar to the average between 2000 and 2005, it is practical to use average precipitation in 2003 as a comparison for assessing lake-hydrology between 2000 and 2005.

Temperature

Average evaporation-flux-weighted temperature during the thaw-season of 2003 (12.7°C) is similar to the average between 2000 and 2005 (12.0°C) and climate normals (12.5°C). Daily temperature rose consistently from early May until July, and then remained fairly stable at around 15°C until late August (Figure 6). y the beginning of September, the temperature dropped rapidly and remained low (i.e., <8°C) for the remainder of the thaw-season. Although the difference in weighted (12.7°C) and non-weighted (11.1°C) average monthly temperatures was not substantial in 2003, flux-weighting is still recommended to accurately represent air temperature. In cases where non-flux-weighted temperature is significantly different from the flux-weighted value, calculation of key isotopic datums on the isotopic framework would not be as precise, ultimately reducing the accuracy of interpretation of lake water isotope data.

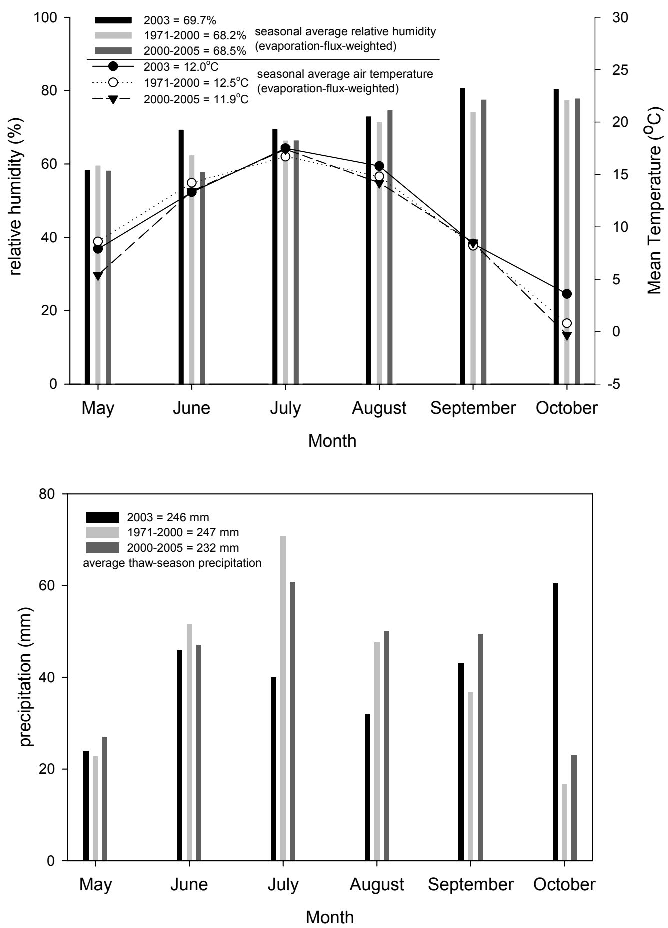


Figure 5: Climate conditions of the thaw-season in 2003 compared with climate normals
Thaw-season relative humidity (*h*) and temperature (T) of 2003 were evaporation-flux-weighted to monthly estimates of evaporation using the Priestly-Taylor (1972) method, whereas monthly climate normals are evaporation-flux-weighted using the Thornthwaite (1948) method. *H, T* and precipitation were measured in Fort Chipewyan by standard instrumentation maintained by Environment Canada (Environment Canada, 2005).

Relative humidity

Unlike precipitation and temperature, flux-weighted relative humidity was slightly higher in 2003 (69.7%) versus the average between 2000 and 2005 (68.1%) and 1971 to 2000 climate normals (68.2%). Over the duration of the thaw-season, non-flux-weighted relative humidity rose from an initial value of around 50% in early May to around 80% by October, with an average of 72.5%. After evaporation flux-weighting, the average thaw-season relative humidity was 69.7%. As will be shown subsequently, relative humidity is the most sensitive of the hydroclimatic parameters in calculations of isotopic datum values. It is especially sensitive in the calculation of δ^*, which affects the interpretation of displacement of lake water data along the LEL (Gibson, 2002b; Brock *et al*., 2007). Therefore, it is essential to evaporation-flux-weight relative humidity if evaporation data are available. Relative humidity was found to rise and fall by up to 20% daily, with peaks associated with precipitation events and lows associated with extended periods without precipitation (Figure 6).

Evaporation

By ice-off (early May) in 2003, daily evaporation rates based on the PT method were already similar to mid-July values (~2.5 mm/day) (Figure 6). High evaporation rates in May resulted from high incoming radiation (Appendix A). The potential for strong lake evaporation early in a thaw-season is important to recognize when assessing the degree to which snowmelt refills a lake in the spring. At 2.5 mm/day, evaporation could cause ~75 mm of water loss in the month of May, which is very similar to the average amount of water that snowmelt was estimated to have contributed to lakes between 2000 and 2005 (68 mm based on a 7.5 to 1 snow-to-water equivalent ratio). Consequently, in a year with high radiation in May, rain could be required to accommodate for most of the evaporative loss of a lake in the remainder of the thaw-season. By mid-August, calculated evaporation began to decline and continued to do so up until the end of October when evaporation became essentially negligible. Total thaw-season evaporation for 2003 was estimated as 401 mm, which corresponds very well with estimates of open-water evaporation for the PAD in 1996 (391 mm) and 1997 (405 mm) as reported by Peters (2003). Based solely on evaporation and rainfall the ratio would be 1.63 parts evaporation to 1 part rain (i.e., 401 mm/246 mm); calculations which support previous findings that open-water evaporation exceeds precipitation in the PAD (PAD-TS, 1996). If this water balance was sustained, the PAD would have likely undergone more significant water loss than was witnessed in 2003 (and between 2000 and 2005). Since relative humidity is one of the main controls on the ability of water to evaporate, and precipitation was similar to climate normals and the 2000 and 2005 average, it is expected that evaporation during 2003 was slightly lower than climate normals and the average between 2000 and 2005.

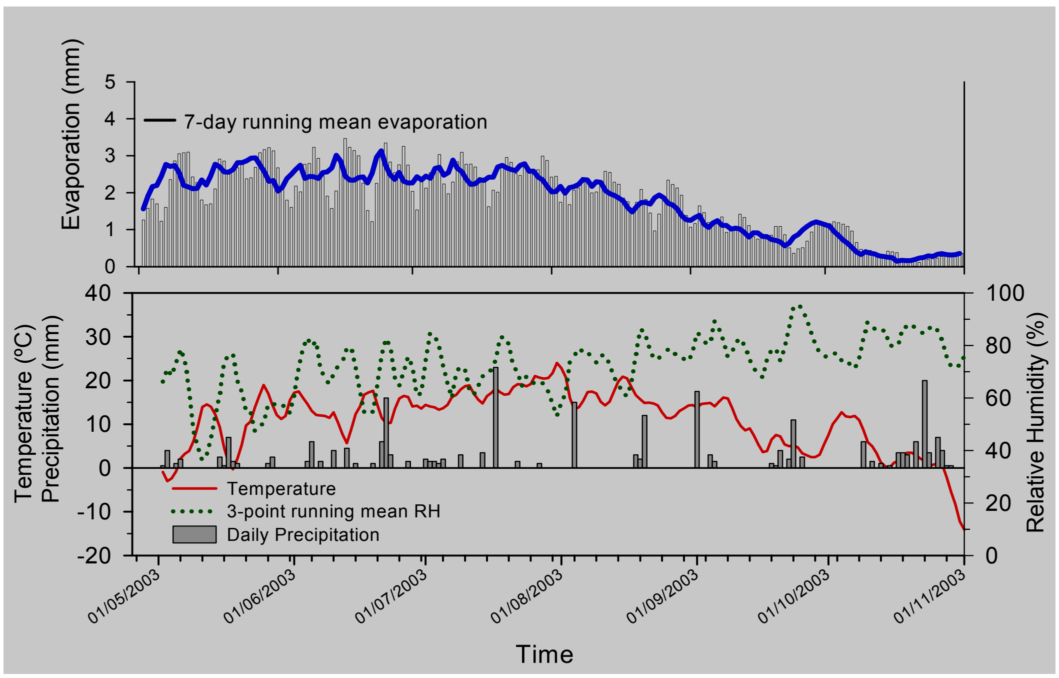


Figure 6: Hydroclimatic conditions in Fort Chipewyan between May 1st and October 31st, 2003.

Local Meteoric Water Line: 2003

Development of the LMWL

As shown on Figure 7, the LMWL is defined by $\delta^2H = 7.0\delta^{18}O - 16.2$ based on a regression through the isotopic compositions of rainfall and snowmelt collected in 2003, which is in good agreement with the regression through rain (n=38) and snow (n=22) samples collected between 2000 and 2005 ($\delta^2H = 6.9\ \delta^{18}O - 21.2$; Appendix C) and the reported LWML for the closest Canadian Network for Isotopes in Precipitation station in Fort Smith, Northwest Territories station (CNIP, 2005): $\delta^2H = 6.7\ \delta^{18}O - 19.2$. Scatter of rainfall isotopic data around the LMWL is caused by differences in the isotopic composition of the clouds that formed the rain, and difference and subsequent partial evaporation and re-equilibration of rain drops.

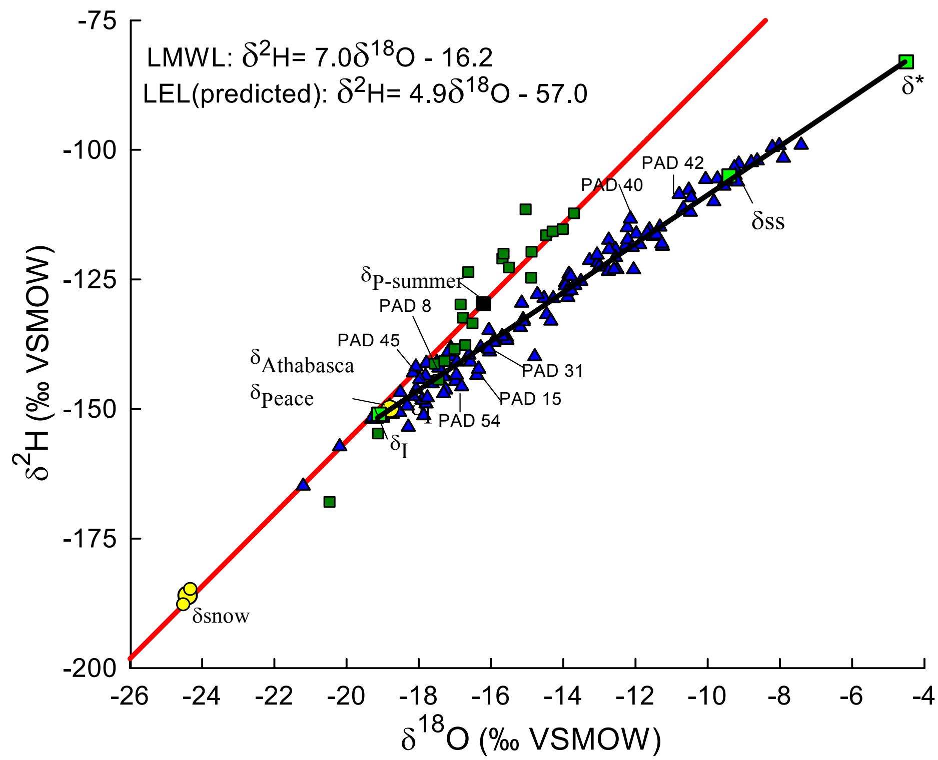


Figure 7: Isotope labeling of precipitation (squares and circles) and 10 lakes (triangles) in δ^2H-$\delta^{18}O$ space collected between June and September of 2003.

The LMWL is based on a regression through isotopic compositions of rain and snow (Appendix C). Rain water was amount-weighted to account for the volume of water that fell at the time of each rainstorm, using volumes collected in rain buckets containing oil to prevent evaporation. $\delta_{P\text{-summer}}$ represents average thaw-season rainfall. The LEL is based on a projected line from δ_I to δ^* (see text). δ_I is taken as the intersection point at the LMWL from the regression line through all measured isotopic compositions of lake water (-19.1‰ for $\delta^{18}O$ and -151‰ for δ^2H). See Appendix F and G for isotopic compositions of the 10 lakes. The isotopic composition of thaw-season δ^* is -5.5‰ for $\delta^{18}O$ and -83‰ for δ^2H, based on equation (5) in Chapter 3 where δ_A (-25.4‰ for $\delta^{18}O$ and -202‰ for δ^2H) was approximated by fitting δ_L in equation (6) from Chapter 3 to match δ_{SS}. Thaw-season (beginning of May to end of October) evaporation-flux-weighted *h* and T were calculated as 69.7% and 12.7°C, respectively. Thus, based on a line between δ_I and δ^*, the predicted LEL is defined by $\delta^2H = 4.9\ \delta^{18}O - 57.0$. A few examples of lakes that were affected by rain, snow, or river water are included. See text for details.

Snow and δ_{Snow}

The majority of snowmelt occurs in the PAD once air temperatures are maintained above zero for a significant period of time (April to early May). It is expected that snowmelt will affect all delta lakes to some degree, depending on their volume and catchment size. Although processes like melting and refreezing of the snowpack can result in isotopic fractionation that can cause isotopic variability within the snowpack (Cooper, 1998), these processes can be disregarded in this study because assessment of snow in lakes depends on the bulk isotope signature of the snowpack. The average isotopic composition of snow (δ_{Snow}) in 2003 (n=2) was -24.4‰ and -186‰ for $\delta^{18}O$ and $\delta^{2}H$, respectively. (-24.4‰ will be used in the isotope mixing model in Chapter 6 as the value for snowmelt). These values fall very close to the average composition of snow collected between 2000 and 2005 (-24.5‰, -189‰), and thus should represent a good approximation for analysis during other thaw-seasons. However, it should be noted that between 2000 and 2005, snow ranged from -19.4‰ to -30.6‰ for $\delta^{18}O$ and from -141‰ to -233‰ for $\delta^{2}H$ (Appendix C). The large range is likely because the snow samples were taken at different times of the winter and from different areas within the delta. Similar to rain, the moisture source for the air mass that deposited the snow in this northern environment may be different for different snow storms. Though, most snow samples were ±2‰ in $\delta^{18}O$ and ±20‰ in $\delta^{2}H$ from the average, so using an average value is appropriate for subsequent modelling purposes. Overall, multiple snow samples from throughout a study area of interest should be taken for properly determining the isotope composition of snowmelt in each area of interest. Note that, as expected, the average isotopic composition of snowmelt lies on the LMWL and is more depleted than rainfall or river water (Figure 7).

On the isotopic framework, data from a lake that is dominantly influenced by snowmelt should plot on a trajectory towards the isotopic composition of snowmelt (see Wolfe *et al*., 2007b). An example of this can be seen on Figure 7 where a few samples, taken in mid- to late-summer from two oxbow lakes (PAD 15 and PAD 54), plot below the LEL. Both catchments have large contributing areas in relation to their surface areas, and are surrounded by mature forest. Because of their physical and catchment characteristics, these lakes are more likely to be fed by sustained snowmelt-derived local groundwater input throughout the thaw-season. One of these lakes (PAD 15) was found to be dominantly snowmelt-influenced in the thaw-season of 2000 by Wolfe *et al*. (2007b).

Rain and $\delta_{p\text{-}summer}$

Similar to snow, rain should influence all lakes to some degree. The average isotopic composition of rainfall for 2003 (n=20) that will be used later during quantitative assessment of input sources is -16.2‰ and -130‰ for $\delta^{18}O$ and $\delta^{2}H$, respectively. This is similar to the average composition of precipitation collected between 2000 and 2005 (-15.7‰, -129‰; Appendix C), and thus should represent a good approximation for analysis during the other thaw-seasons. In addition, this corresponds well to the interpolated isotope composition of thaw-season precipitation in the PAD region using Canadian Network for Isotopes in Precipitation stations (-15‰ and -120‰ for $\delta^{18}O$ and $\delta^{2}H$, respectively; Gibson and Edwards, 2002).

The isotopic compositions of rainfall ($\delta_{p\text{-summer}}$) measured from around the PAD in 2003 ranged from -13.9‰ to -20.5‰ for $\delta^{18}O$ and -112‰ to -168‰ for $\delta^{2}H$. The largest rain event on one day (22 mm) occurred on July 17th. This rain event is discussed during isotopic analysis of lakes in Chapter 5, and was also used as an example of how to characterize rainfall as an input source to a lake in the absence of rainfall data. Slight variability of $\delta_{p\text{-summer}}$ around and along the LMWL (Figure 7) reflects variations in the condensation temperature and air mass vapour composition of individual precipitation events. In fact, steeper temperature gradients between the PAD and source of vapour in the winter explain why snow is generally more depleted in heavy-isotope content compared with thaw-season rain. Cool, continental Arctic air masses likely dominate the vapour mass compositions in the winter, whereas, warm, relatively moist Pacific air masses (westerly winds) likely dominate in summer months. Rainfall samples that plot to the right of the LMWL and closer to the LEL (Figure 7) represent light rain storms (<2mm) where raindrops have undergone partial evaporation as they passed through under-saturated atmosphere (Araguas-Araguas *et al.,* 2000).

Since the isotopic composition of rain generally plots near the upper end of the LMWL, it is expected that a lake that has received recent contributions from rain will plot above the LEL. In 2003, most samples that plot above the LEL were taken following moderate to large precipitation events from lakes that exhibit sensitivity to direct rainfall input. Examples include two shallow lakes surrounded by catchments capable of holding rain water following a rainfall in late September 2003 (PAD 40 and PAD 42) (Figure 7).

Rivers and δ_{River}

Influence by river water is common for lakes with low levees near river reaches. The rivers in the Athabasca sector that were sampled were the Athabasca River, Fletcher Channel, Embarras River and Mamawi Creek. The average isotopic composition of rivers in the Athabasca sector ($\delta_{Athabasca}$) in the thaw-season of 2003 was -18.2‰ for $\delta^{18}O$ and -145‰ for $\delta^{2}H$ (Appendix D). The rivers in the Peace sector that were sampled were Rivière des Rochers, Revillon Coupé, Chenal des Quatre Fourches, Baril River, and Claire River. The average isotopic composition of rivers in the Peace sector (δ_{Peace}) was -18.8‰ for $\delta^{18}O$ and -150‰ for $\delta^{2}H$, similar to the isotopic composition of rivers in the Athabasca sector. However, the average isotopic compositions of rivers in the Athabasca sector in the spring of 2003 (-19.2‰ for $\delta^{18}O$ and -152‰ for $\delta^{2}H$) differed from the compositions of rivers in the Peace sector (-21.2‰ for $\delta^{18}O$ and -165‰ for $\delta^{2}H$). The difference in average composition between the rivers in each sector in the spring could have been due to evaporatively-enriched discharge that enters the Athabasca River from Lesser Slave Lake (Wolfe *et al*., 2007b), and/or greater proportions of isotopically-depleted glacier and/or snowmelt water flowing into the Peace River from the Rocky Mountains of British Columbia. The range of isotopic compositions for rivers within each sector was minimal (i.e., Peace sector rivers ranged from -19.1‰ to -19.3‰ for $\delta^{18}O$ and -151‰ to -153‰ for $\delta^{2}H$, and Athabasca sector rivers ranged from -21.0‰ to -21.3‰ for $\delta^{18}O$ and -164‰ to -166‰ for $\delta^{2}H$). Conservation of isotopic compositions of the rivers in each sector is consistent with field observations made on April 30th and May 1st 2003, whereby ice-jam formation on the Athabasca and Embarras Rivers caused the northward flow of water into rivers in the Athabasca sector, and low water levels of

Lake Athabasca and Mamawi Creek caused the Peace River to flow south into rivers in the Peace sector. Accounting for the differences in the isotopic compositions of rivers in each sector in the spring of 2003 was found to be important for acquiring more accurate results in Chapter 5 and 6 using an isotope mixing model.

It is expected that a lake that has received recent contributions from rivers will plot close to the isotopic composition of the river water it has received. This is because river water is expected to either partially or entirely flush a lake during high water levels. A lake that has received river water should also plot near the LEL since the isotopic compositions of rivers generally plot close to the LEL. Indeed, in 2003 through-flow lakes (e.g., PAD 45-Mamawi Lake) or frequently flooded lakes (e.g., PAD 8 and PAD 31-Johnny Cabin Pond) plotted close to the isotopic composition of river water, and thus close to the LEL for much of the 2003 thaw-season (the cluster of values near δ_{Peace} and $\delta_{Athabasca}$ on Figure 7).

Local Evaporation Line: 2003

Recall that in the absence of inputs from any input source, a lake will tend towards δ^*. Therefore, displacement of lake water data along the LEL in relation to δ_I and δ^* reflects differences in heavy-isotope accumulation due to evaporation, and therefore can be used to semi-quantitatively evaluate lake water balance status. Recall also that the isotopic composition of a terminal lake in hydrologic and isotopic steady-state (i.e., δ_{ss}) under prevailing hydroclimate should lie somewhere on the LEL. The following describes the formation of the predicted LEL for the PAD in 2003, which is based on a trajectory from average input water (δ_I) to the limiting enrichment value (δ^*) (Figure 7).

Average input composition (δ_I)

The amount-weighted isotopic composition of input to lakes (δ_I) in the PAD was estimated using a conventional method based on the intersection point of the regression line through the isotopic compositions of lakes in 2003 (PAD 1, 5, 8, 9, 15, 18, 23, 31, 45, and 54) (Appendix F) and the LMWL: -19.1‰ for $\delta^{18}O$, -151‰ for δ^2H. This is in good agreement with the intersection point of the regression line through the isotopic compositions of these lakes between 2000 and 2005 (Appendix F) and the LMWL (-18.9‰ for $\delta^{18}O$, -150‰ for δ^2H); and the average composition of the three main inputs (rain, snow, river water) (-19.7‰ for $\delta^{18}O$ and -155‰ for δ^2H). The value also approximates average amount-weighted input water to Fort Smith, Northwest Territories: -18.8‰ for $\delta^{18}O$, -147‰ for δ^2H (CNIP, 2005).

Steady-state composition (δ_{SS})

The average isotopic composition of thaw-season δ_{SS} in 2003 was assumed to be equal to the average isotopic composition of the reference basin (Greenstar Lake, GSL) in 2003 (-9.4‰ for $\delta^{18}O$, -103‰ for δ^2H). GSL maintained a relatively steady water level (9.2±0.04 m) and isotope composition (-9.3±1.1‰ for $\delta^{18}O$ and -103±22‰ for δ^2H) during the thaw-season of 2003 based on measurements during the monitoring period (October 2000 to September 2005). GSL's average measured δ_{SS} isotopic value (-9.4‰ for $\delta^{18}O$, -103‰ for δ^2H) in 2003 corresponds well with a modelled δ_{SS} value, based on equation (7) in Chapter 3 (-9.9‰ for $\delta^{18}O$, -112‰ for δ^2H), which relies on the assumption that atmospheric moisture is in isotopic equilibrium with summer precipitation (i.e., $\delta_A = (\delta_{Rain} - \varepsilon^*)/\alpha^* \approx \delta_{Rain} - \varepsilon^*$; see Gibson and Edwards, 2002). The fact that the average isotopic composition of GSL (i.e. δ_{SS}) plots almost exactly on the predicted LEL (Figure 7), provides support that this lake can be used as a monitor of long-term average isotope and hydroclimatic conditions within the PAD.

Limiting enrichment composition (δ^*)

The isotopic composition of δ^* in the thaw-season of 2003 is -5.5‰ for $\delta^{18}O$ and -83‰ for δ^2H, based on equation (5) in Chapter 3 where δ_A (-25.4‰ for $\delta^{18}O$ and -202‰ for δ^2H) was approximated by fitting δ_L in equation (6) (Chapter 3) to match δ_L of GSL in 2003 (i.e., δ_{SS}). Recall, thaw-season evaporation-flux-weighted *h* and T (Figure 5) for 2003 were calculated as 69.7% and 12.7°C, respectively.

Thus, based on a line between δ_I and δ^*, the predicted LEL based on hydroclimatological analysis for the thaw-season of 2003 is defined by $\delta^2H = 4.9\ \delta^{18}O - 57.0$. This line will be used on the isotopic framework for all subsequent analyses. The predicted LEL corresponds well with the empirical way of generating an LEL, by taking a regression line through evaporating study lakes in the region of interest, which is defined by $\delta^2H = 4.7\ \delta^{18}O - 61.7$. Correspondence of these lines suggests that the reference basin experienced hydroclimatic conditions as the rest of the study lakes and, consequently, its use to model δ_A to calculate δ^* was appropriate.

Comparing isotopic frameworks: 2003 versus 2000-2005

Although the LMWL and LEL were formed using 2003 information because data were abundant and an evaporation pan was maintained, it should be noted that they are almost identical using the less comprehensive, scattered dataset available from 2000 to 2005. The LWML based on regression through rain (n=38) and snow (n=22) samples collected between 2000 and 2005 is $\delta^2H = 6.9\ \delta^{18}O - 21.2$, which is very close to the LEL derived for 2003: $\delta^2H = 7.0\delta^{18}O - 16.2$. The predicted LEL for 2000 to 2005 is $\delta^2H = 4.8\delta^{18}O - 60.6$, which also compares very well with the predicted LEL of 2003: $\delta^2H = 4.9\ \delta^{18}O - 57.0$. The LEL for 2000 to 2005 is based on δ^* as -4.7‰ for $\delta^{18}O$ and -83.4‰ for δ^2H using δ_A as -25.5‰ for $\delta^{18}O$ and -207‰ for δ^2H, which is based on isotopic data from the reference basin and equation (5) and (6) in Chapter 3; and δ_I as -18.9‰ for $\delta^{18}O$ and -150‰ for δ^2H, which is based on the intersection point of the LMWL and a regression line of lake water data taken between 2000 and 2005. Regardless of whether an isotopic framework was developed using 2003 data or 2000 to 2005 data, analysis of lake water isotopic data would be identical. However, when analyzing lake water isotope data on a shorter time scale use of specific thaw-season hydroclimate and isotopic data is important (Brock *et al*., 2007). Over weekly to monthly scales it is expected that there will be short-term changes in isotope parameters used on the isotopic framework, especially δ^* due to it's sensitivity to variations in relative humidity and δ_A.

Short-term variability of δ_{SS}, δ_A, and δ^*: Evaporation Pan

The effects of short-term changes in hydroclimate and isotope parameters on lake water balance conditions are important to acknowledge when conducting water balance investigations on shallow lakes in northern environments (Gibson *et al.*, 1999). These changes are especially important for assessing seasonal time-series data. Since evaporation is strongly controlled by relative humidity, it is one of the most sensitive hydroclimate parameters that effect shallow lakes in seasonal environments. A change in relative humidity can significantly alter the isotopic composition of δ^* (equation 7, Chapter 3). For example, a change of 5% in relative humidity from the average in 2003 would result in a net change in the isotopic composition of δ^* by ~2.5‰ in $\delta^{18}O$ and ~12‰ in δ^2H. A change of this magnitude could significantly alter the interpretation of lake water data along the LEL. δ^* plays a key role in interpreting how far a lake is from steady-state and, more importantly, how close it is to desiccation. For example, if the isotopic composition of lake water was taken after a particularly dry period (low relative humidity) and was assessed using a more depleted average thaw-season value of δ^*, the lake may not be identified as undergoing strong, non-steady-state evaporation. However, a considerable change in relative humidity does not significantly alter the slope of a predicted LEL, and thus would not have a significant effect on how the position of lake water isotope data around the LEL is interpreted. For example, if relative humidity was 5% higher in the thaw-season of 2003, the LEL would have been $\delta^2H = 5.0\delta^{18}O - 55.0$ instead of $\delta^2H = 4.9\delta^{18}O - 57.0$. This minor change in slope would be effectively negligible at the usual scale at which isotopic frameworks are prepared.

One of the most useful ways of quantifying short-term variability of δ^*, as well as δ_A and δ_{SS}, is through hydrologic and isotopic maintenance of an evaporation pan (Gibson *et al.*, 1999). The following includes a discussion on results from an evaporation pan maintained in Fort Chipewyan between May 21st and August 10th, 2003.

Table 1 and Figure 8a provide hydroclimatic measured (δ_{SS}) and modelled (δ_A and δ^*) data during the period of maintenance of the evaporation pan. The modelling of δ_A and δ^* was conducted for each sampling interval, which was, on average, 12 days in length. Isotopic measurement of pan water (δ_{PAN}) after each sampling interval provided a means to assess short-term changes in the isotopic composition of a water body at steady-state (i.e., δ_{SS}). The pan was assumed to be in steady-state with ambient atmospheric conditions by June 10th (-10.9‰ for $\delta^{18}O$ and -117‰ for δ^2H) because the June 19th sample (-10.8‰ for $\delta^{18}O$ and -116‰ for δ^2H) had an almost identical isotopic composition. After June 19th, δ_{PAN} became gradually more enriched, resulting in an August 10th composition of -8.6‰ for $\delta^{18}O$ and -103‰ for δ^2H. Slight isotopic enrichment is reasonable considering relative humidity was ~2% lower in late July and early August as compared with June and early July. In fact, relative humidity often rises throughout the thaw-season (e.g., 2003 on Figure 5), and thus δ_{SS} should become more depleted as the thaw-season progresses.

It can be assumed that there were similar hydroclimatic conditions controlling both the evaporation pan and reference lake since the average measured pan composition (-9.7‰ for $\delta^{18}O$ and -109‰ for δ^2H) was almost identical to the average thaw-season composition of the reference lake (-9.6‰ for $\delta^{18}O$ and -106‰ for δ^2H; Table 1). This also verifies that the reference basin was in isotopic steady-state during this period, and was in equilibrium with relative humidity and temperature thereby validating the use of the reference basin for modelling long-term changes δ_A and δ^*. It also shows that the maintenance of a class-A evaporation pan is ideal for assessing key isotopic data when there is no reference lake available in a study area.

Table 1: Modelling results of key isotopic parameters in 2003 using the evaporation pan.

Interval	T_{fw} (°C)	h_{fw} (%)	δ_{I}-weighted (‰)	δ_{PAN} (δ_{SS}) (‰)	δ_{A}-Pan (‰)	δ_{A}-precip-equil (‰)	δ^{*} (‰)
June 10th to June 18th	13.3	67.5	-17.1 -142	-10.9 -117	-30.0 -225	-28.5 -233	-7.6 -102
June 19th to June 23rd	11.4	75.3	-16.7 -129	-10.8 -116	-27.0 -222	-27.1 -215	-9.0 -111
June 24th to July 12th	15.8	70.3	-16.6 -135	-9.1 -105	-26.2 -207	-26.3 -212	-5.7 -89
July 13th to July 18th	16.1	69.3	-14.3 -117	-9.3 -106	-27.1 -213	-24.4 -202	-7.4 -100
July 19th to August 10	19.1	69.4	-15.9 -131	-8.6 -103	-25.5 -209	-24.5 -204	-5.3 -90
Average	***15.5***	***70.4***	***-16.1*** ***-131***	***-9.7*** ***-109***	***-27.2*** ***-215***	***-26.1*** ***-213***	***-7.0*** ***-98***
Thaw-season Average	***12.7***	***69.7***	***-19.1*** ***-151***	***-9.6***[A] ***-106***[A]	***-25.4***[A] ***-202***[A]	***-26.1*** ***-219***	***-5.5***[A] ***-83***[A]

$\delta_{A\text{-pan}}$ represents the δ_A value derived from the pan, and was calculated by adjusting δ_A to force the right side and left side of equation (6) to equal. δ^* and δ_{SS} were evaluated using equations (5) and (7), respectively, where δ_{SS} was treated as δ_{pan}. $\delta_{I\text{-weighted}}$ takes into account the amount of rain water that was captured by the pan between sampling intervals.
[A] based on modelling results of the local reference basin (Greenstar Lake).

The isotopic compositions of modelled atmospheric moisture (δ_A) in each sampling interval using δ_{PAN} as δ_L in equation (7) are reasonable and corresponded well (r^2=0.76) with traditional assessment of δ_A by assuming δ_A is in approximate isotopic equilibrium with local rain (i.e., $\delta_A = (\delta_{Rain}-\varepsilon^*)/\alpha^* \approx \delta_{Rain}-\varepsilon^*$) (Gibson *et al.*, 1993). This provides compelling evidence that the pan water isotopic composition can be used to calculate variations in ambient atmospheric moisture isotope composition. On the whole, $\delta^{18}O$ of δ_A enriched from an initial value of -30‰ to -25.5‰ by the end of the summer (both $\delta^{18}O$ and δ^2H behaved very similarly). Enrichment of atmospheric moisture throughout the summer is likely caused by the advancement of isotopically-enriched, warm continental air masses dominating as summer progresses, compared to the more isotopically-depleted, cold arctic air masses that influence conditions in the spring.

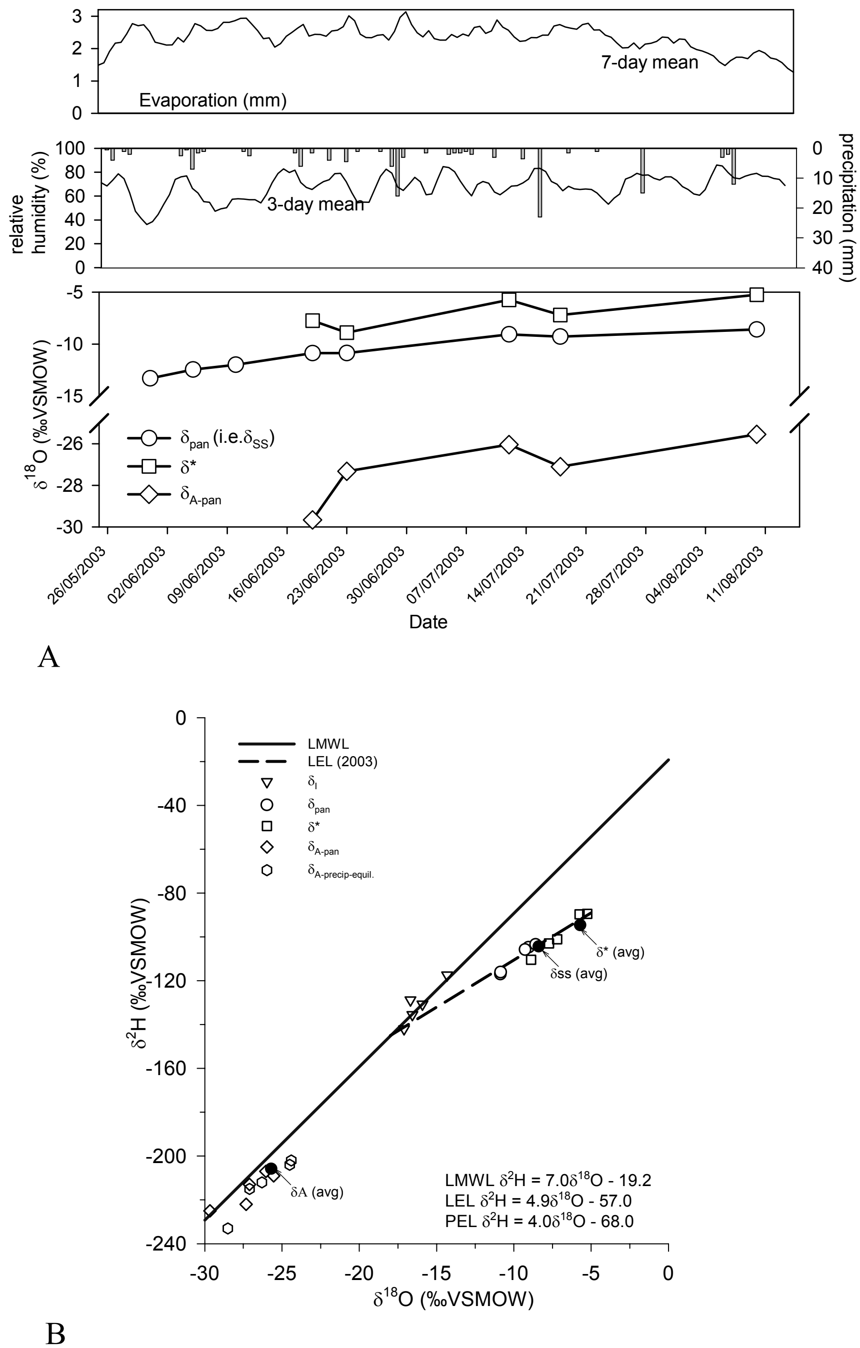


Figure 8: Evaporation pan modelling results. a) Hydroclimatic data and time series of $\delta^{18}O$, b) Pan results in $\delta^{2}H$-$\delta^{18}O$ space.

For example, an isotopically-enriched rainstorm (17 mm over 12-hours) on July 17th (-14.9‰ in $\delta^{18}O$) would have temporarily changed the isotopic composition of δ_A to a more enriched value. It is this short-lived, isolated rainstorm that explains the discrepancy between the values of δ_A using the precipitation-equilibrium assumption (-24.4‰ in $\delta^{18}O$), and the modelled δ_A composition using the evaporation pan (-27.1‰ in $\delta^{18}O$) from the July 13th to July 18th interval. The discrepancy can be explained by the fact that the precipitation-equilibrium method only accounts for δ_A during rainfall, whereas the evaporation pan records δ_A continuously. In this case, it is likely that after the saturated air mass passed, the more common isotopically-depleted local air mass representing most of the 5 day time interval prevailed. Therefore, since the pan-derived δ_A records are continuous and reflect interim periods when relative humidity is less than 100%, use of an evaporation pan is more accurate for estimating short-term variations in δ_A than the precipitation-equilibrium approximation.

Over the duration of the evaporation pan experiment, δ^* oscillated between -5.3‰ and -9.0‰. As expected from equation 5 (Chapter 3), the value of δ^* is mainly dependent on relative humidity and δ_A. To verify this relationship, it can be seen that the most depleted value of δ^* occurred during the June 19th to 23rd interval, when the relative humidity was highest (75.3%); and the most enriched value of δ^* occurred during the July 19th to August 10th interval when δ_A was enriched (estimated at -25.5‰ for $\delta^{18}O$ and -209‰ for δ^2H). Differences in relative humidity during the time the pan was maintained (68.0%) and the relative humidity signal that would have been captured by GSL during sampling in 2003 (last three thaw-season average=70.4%) may help to explain why the average modelled isotopic composition of δ^* (-7.0‰) was slightly more depleted than the δ^* based on GSL (-5.5‰). This clearly shows the effect of residence time of each reference basin and that each captures information important at different hydrological scales.

Although short-term changes in hydroclimate have been shown to cause temporary changes in key isotope data, the use of average thaw-season hydroclimate values and isotope data to develop an isotopic framework is still very practical for understanding the importance of the various hydrological processes in a complex environment like the PAD. However, awareness of short-term changes in isotope values allows for more accurate interpretation of lake water data. This is especially important during periods of short-lived, irregular hydroclimate, such as extensive arid periods when lakes are more strongly affected by evaporation. It has been shown that monitoring of the isotope and mass budgets of a class-A evaporation pan provides the means to accurately assess short-term changes in isotope data. An application of the evaporation pan results and short-term changes in isotopic parameters will be revisited in Chapter 5 when assessing why the isotopic composition of an isolated lake exceeded average thaw-season δ^*, a theoretically impossible occurrence.

Use of the isotopic framework

Recall, the position of the isotopic composition of surface waters around, and along the LEL in reference to δ_I, δ_{SS}, and δ^* provides a means for assessing the importance of key hydrological processes controlling lake water balances. Lake water oxygen and hydrogen isotope data collected between June 5th and September 22nd in 2003 (Figure 7) are generally localized between δ_I and δ^*, which suggests that the predicted-LEL spans the continuum of water balances in the PAD. Lake water data also plot around the predicted LEL, suggesting the LEL provides a reasonable assessment of local hydrology and isotope hydroclimate. The later verifies that the predicted LEL is practical for assessing how local hydrological processes and hydroclimate control lake water balances over the thaw-season.

The existence of slight scatter about the LEL likely reflects basin-specific variations in the isotopic compositions of the various input sources at the time of sampling (Wolfe *et al*., 2007b). The distance away from the LEL has been shown to reflect the physical characteristics of the lake (i.e., surface area to volume ratio) and its catchment characteristics (i.e., surrounded by forests or wetlands). For example, a lake with a small volume would be more sensitive to snowmelt or rain, and would therefore deviate from the LEL more than a large lake. Regardless of lake size, hydrological studies that rely on isotope data always require knowledge of input composition since isotope-mass balances require a lake to tend towards δ^* from δ_I as evaporation occurs. Knowledge of the distance from a lake's input source along its own LEL allows for quantitative assessment of instantaneous lake water balance (e.g., Wolfe *et al*., 2007b).

The following will describe one method to constrain the average input composition to a lake over a specified time period, and another method to constrain event-specific input composition to a lake. The first method, which is the conventional means of determining δ_I, involves determining the point at which a regression line of lake water data over some period of time (i.e., a thaw-season) intersects the LMWL.

However, this method only yields the average input composition, which can be derived from various amounts and types of input sources (i.e., rainfall, snowmelt, etc.) that often have very different isotopic compositions through time. The second method allows for differentiation of input sources and can be used if measured isotope data of a lake are sparse, or if the isotopic composition of the event water is needed (i.e., for an isotopic mixing model). In this method, the point at which the trajectory from the *pre-event* isotope composition and *post-event* isotope composition intersects the LWML (Figure 9) can be used. This method relies on the concept that the isotopic composition of a lake will be drawn towards the isotopic composition of event water on the isotopic framework. Although this method does provide a means to assess the isotopic composition of input water to a lake following a significant hydrological event, it requires timely sampling of the pre- and post-event isotopic compositions of the lakes in an environment such as the PAD because of sensitivity to evaporative enrichment. This method may help to address temporal and spatial variability of the isotopic composition of input sources to lakes in high-latitude environments, an uncertainty identified with isotope mixing models by Hayashi (2004). The LWML-intersection method will be used to identify input source composition required for an isotopic mixing model following rain and snow events on an isolated lake in the Peace sector (Chapter 6).

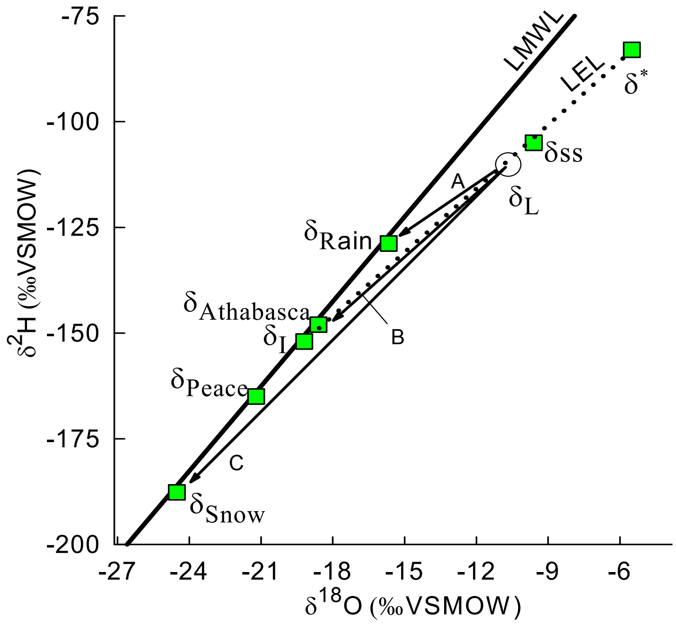


Figure 9. Expected isotopic depletion trajectories of PAD lakes in response to a) thaw-season rain, b) river flooding in the Athabasca sector, or c) snowmelt input, superimposed on the isotopic framework for the thaw-season of 2003.

The isotopic composition of parameters on the framework include, the limit of isotopic enrichment (δ^*; -5.5‰ for $\delta^{18}O$ and -83‰ for δ^2H), a basin in hydrologic and isotopic steady-state (δ_{ss}; -9.4‰ for $\delta^{18}O$ and -103‰ for δ^2H), average rainfall (δ_{Rain}; -16.2‰ for $\delta^{18}O$ and -130‰ for δ^2H), average local input (δ_I; -19.1‰ for $\delta^{18}O$ and -151‰ for δ^2H), and average local snowmelt (δ_{Snow}; -24.4‰ for $\delta^{18}O$ and -186‰ for δ^2H). For illustrative purposes, the average isotopic composition of rivers in the Athabasca sector ($\delta_{Athabasca}$, -19.2.4‰ for $\delta^{18}O$ and -152‰ for δ^2H) and Peace sector (δ_{Peace}, -21.2‰ for $\delta^{18}O$ and -165‰ for δ^2H) in spring 2003 were used.

Conclusions and implications

Overall, average temperature and precipitation during 2003 were similar to the average between 2000 and 2005, as well as to climate normals. However, average relative humidity in 2003 was slightly higher than 2000 to 2005 and the climate normals. Since the limit of evaporative-enrichment (δ^*) of lake water is strongly controlled by relative humidity, δ^* in 2003 would be slightly more depleted than the average between 2000 and 2005 and the climate normals. It was shown that a considerable change in relative humidity would not significantly alter the slope of the LEL, and thus would not have a significant affect on how lake water isotopic data would be interpreted around the LEL. On the other hand, it was shown that evaluation of lake water data along the LEL could be affected by changing δ^*. The implication of using the more depleted value of δ^* from 2003 to compare isotope data from other years is that the degree of evaporation from a lake may be misinterpreted. However, maintenance of an evaporation pan, and availability of a local reference lake in isotopic and hydrologic steady-state, allows for evaluation of δ^* over various time scales.

Intuitively, it was expected that evaporation would be the lowest in spring and the highest in mid-summer, when temperatures were higher and day-light hours were longer. However, low relative humidity and high radiation in May enabled evaporation rates during the spring to be similar to mid-summer values. In the spring, a lake may simultaneously undergo evaporative-enrichment while still receiving isotopically-depleted snowmelt and/or river water. By mid-September, evaporation was negligible. Therefore, PAD lakes underwent strong evaporation for four months from the onset of the open-water season to the beginning of the wet fall season. Since evaporation is strong during the thaw-season in the PAD, it is suggested that evaporation-flux-weighting of relative humidity and temperature be conducted. Indeed, it was shown that flux-weighting of these parameters provides highly comparable observed and predicted LEL, and similar modelling results of δ_A and δ^* using the evaporation pan and the reference basin.

The low precipitation in mid-summer of 2003 was compensated by the wet and cold autumn. The high autumn rainfall would have reset the lakes with isotopically-depleted water after undergoing various degrees of evaporative enrichment throughout the summer. Wet ambient moisture conditions in the fall would also have established saturated conditions for the melt period in the following spring, increasing snowmelt runoff. Notably, water levels in some of the lakes that were not flooded in spring 2004 were found to be moderately high (e.g., PAD 3, 30, 37, 41) as was observed during water sampling.

Use of evaporation-flux-weighted climate data and isotope data collected from precipitation and an index lake in the thaw-season of 2003 provided the means to form a representative isotopic framework for assessing water balance conditions from 2000 to 2005. A LMWL was formed by taking the regression line of the isotopic composition of rain and snow in 2003, which corresponded well to a LMWL that was developed for the 2000 to 2005 thaw-seasons. The isotopic compositions of rivers in both sectors of the delta were similar during summer months, but differed in the spring. Differences in spring isotopic composition of the rivers in the two sectors are important for more accurate results using an isotopic mixing model in subsequent chapters. A LEL was extended between the isotopic composition of input water and the limiting isotopic composition of a water body under prevailing hydroclimatic conditions. The average input composition was determined through the intersection point at the LMWL from a regression of all lake water data in 2003, which corresponded well with the average isotopic composition of snowmelt, rainfall and river water measured in 2003. The limit of evaporative enrichment was calculated using evaporation-flux-weighted relative humidity, and by modelling the isotopic composition of atmospheric moisture based on the measured isotopic composition of a reference basin in isotopic and mass balance.

Localization of 2003 lake water data to the predicted LEL revealed that lake water balances were being controlled over the season by ambient hydroclimate. The existence of slight scatter around the LEL allows for assessment of the input source specific to each basin at the time of sampling. It is the ability to accurately determine the input source to a lake that will allow for differentiating the relative importance of key hydrological processes controlling lake water balances in the PAD over seasonal and inter-annual timescales. Lake water oxygen and hydrogen data collected over the thaw-season of 2003 were mainly localized between δ_I and δ^*, which showed that the predicted LEL spanned the continuum of water balances in the PAD.

The methods and data used for formation of the isotopic framework are a significant improvement to the more rudimentary framework developed by Wolfe *et al.* (2007b), providing a means to properly assess water balance controls on lakes in the PAD over various temporal and spatial scales. Multiple years of data and the use of a local reference lake in long-term isotopic and hydrologic steady-state provided the main information to make these improvements to the framework. An evaporation pan also afforded a means to assess the sensitivity of lakes to short-term fluctuations in hydroclimate. Awareness of the sensitivity of isotopic data on the LEL (especially δ^*) will facilitate interpretation of lake water data during irregular periods of hydroclimate (i.e., extensive evaporation periods). In addition to being able to assess short-term variations of key isotopic values, the isotopic framework will be used in subsequent modelling exercises to constrain the input composition of source water following individual hydrological events. This allows for assessment of the main input source to a lake following a change in water balance, which is especially important for water balance studies in northern environments.

Superimposing lake water data against the isotopic backdrop presents an opportunity to semi-quantitatively assess lake water balance changes in the PAD over various time- and spatial-scales in subsequent chapters of this thesis, a task that is otherwise difficult, or impossible, to conduct using conventional methods. Offset from the LMWL along the LEL will provide a means to determine the relative proportion of lake water lost to evaporation over some time period (Chapter 5), and the isotopic trajectory of a lake on the isotopic framework will be used to determine the main hydrological input that is dictating isotopic change over seasonal and annual time-scales (Chapter 5), and over the landscape (Chapter 6). The ability to identify the main input source to a basin has broad implications in a wide range of environments, including the tundra and croplands, which are susceptible to changes in climate.

5.0 CHARACTERIZING THE SEASONAL WATER BALANCE OF TWO BASINS DIFFERING IN FLOOD SUSCEPTIBILITY USING WATER ISOTOPE TRACERS

Synopsis

Seasonal and annual variation in lake water balance in the Peace-Athabasca Delta (PAD) remains largely unknown. This chapter explores the use of stable water isotopes to assess variations in the water balances of two hydrologically-contrasting lakes in the PAD: a perched, closed-drainage lake in the Peace sector of the PAD, and a low-lying, restricted-drainage lake in the Athabasca sector. An evaluation of the relative contributions of snowmelt, river-water, and thaw-season rain, and the role of evaporation, is conducted through analysis of the isotopic compositions of input sources and lake waters collected between October 2000 and September 2005. Event-specific isotopic compositions are determined by assessing the trajectory of post-event and pre-event isotope data on the isotopic framework (as described using Figure 9 in Chapter 4). This is followed by quantification of lake replenishment by the main hydrological input using an isotope mixing model. Interpretation of results is assisted by water level data, field observations, and meteorological data.

Results suggest that in the absence of input from river flooding, the isolated lake in the Peace sector and its catchment capture sufficient precipitation to maintain hydrologic and isotopic quasi-steady-state conditions over the five year monitoring period; a lake type that was expected to lose water over the study timeframe since local evaporation exceeds precipitation. Analysis using the isotopic mixing model illustrates that rain dilutes this lake by an average of 16%, while snowmelt dilutes this lake by an average of 28% and thus plays a dominant role in offsetting subsequent thaw-season evaporative losses. In contrast, frequent inputs of river water in the spring and summer reset the water balance of the low-lying Athabasca sector basin throughout the duration of the study period. The ability of summer floods to almost entirely replace lake water in this basin is significant because these events are not well acknowledged and are generally not considered as significant as spring flood events in the PAD.

Discussions in this chapter address 1) the concerns of using isotope data based on average conditions to assess lake water balance of hydrologically-sensitive lakes, and 2) reasons that heavy-isotope enrichment trends differ between the two basin types under similar hydroclimatic conditions. Practical applications include constraining assumptions and interpretation of current and future paleohydrological studies on PAD lakes, as well as assisting in the design of management practices for the PAD under anticipated changes to climate and river use.

Introduction

Flooding has been identified as a key hydrological process controlling the water balance of perched lakes in the PAD (PAD-PG, 1973; Prowse *et al*., 1996; Pietroniro *et al*., 1999; Peters, 2003; Peters *et al*., 2006). However, these studies were based on observations and hydroclimatic data from an unusually dry period between the 1960s and 1990s may have underestimated the importance of catchment-derived runoff, especially from snowmelt. Peters (2003) and Wolfe *et al.* (2007b) suggested that more continuous water balance studies should be conducted in hydrologically isolated basins to determine the relative importance of varying catchment runoff processes in controlling lake water balance. As described in Chapter 1, Wolfe *et al*. (2007b) investigated the hydrological regimes of a suite of PAD lakes by assessing their instantaneous water balances at the end of the thaw-season of 2000. They found that evaporation-sensitive lakes were dominant in the largely relict landscape of the Peace sector, and flood-susceptible lakes were located in the more hydrologically-active Athabasca sector. Although their study provides highly useful information regarding time-integrated water balance conditions of lakes at the end of a thaw-season, it does not provide a sense of event-specific, seasonal or annual water balance variability.

The goal of the present chapter is to evaluate the effect of evaporation and various source waters (including snowmelt, rainfall, and river flooding) on lake water balances at the lake-catchment scale, and to assess how these processes effect seasonal and annual hydrological variability. Hydrological investigations were carried out between October 2000 and September 2005 on two basins, which are end-members of the broad spectrum of water balances that can be found for basins within the PAD. These lakes were targeted because they are representative of other lake basins in their respective sectors and they are expected to provide appropriate information regarding the range of hydrological variability in the PAD. The first lake (Spruce Island Lake), isolated and perched in the Peace sector with no apparent channelized input or output, was studied to objectively evaluate the importance of internal catchment hydrology; including snowmelt, rainfall, and evaporative influences. The importance of evaluating the effects of internal catchment hydrology of this lake stems from a paleolimnological study that concluded that there was no evidence that this lake had dried up in the last 200 years (Falcone, 2002; Wolfe *et al*, 2005). The contrast between the high evaporation-to-rainfall ratio of 1.63

(determined in Chapter 4), but little evidence of sustained water drawdown, also prompted the interest in investigating the role of catchment runoff in resetting and/or sustaining water in perched lakes. In the PAD, perched basins have received a great deal of attention since their water balances are thought to be highly reliant on replenishment by flood waters (Prowse and Conly, 1998; Prowse *et al*., 2002, Beltaos, 2003; Peters, 2003; Peters *et al*., 2006), and because they provide important wildlife habitats. In hydrological contrast, a second low-lying lake (Johnny Cabin Pond) adjacent to a river channel was studied to assess the importance of flood water influence in controlling the water balance of flood-prone lakes in the Athabasca sector of the delta. This lake is also in a prime location to evaluate possible changes in flow patterns of the Athabasca River, as indicated by paleolimnological data (Hall *et al*., 2004; Wolfe *et al*, in review)

The main body of this chapter is separated by analysis of the two study lakes. Within each section, hydrologic variability of the lake is discussed over the five years of the study, and is followed by a more detailed assessment of hydrologic variability within the 2003 and 2004 thaw-seasons. Snowmelt, rainfall and flood events are semi-quantitatively evaluated using the isotopic framework derived in Chapter 4, and then examples of these events are quantified using an isotopic mixing model.

Methods

Data collection

Water samples were collected from lakes and rivers on several occasions for isotopic analysis between October 2000 and September 2005 from ~10 cm below the water surface. Samples of snowpack were taken during various times of the winter, and samples of rain were captured in 1L buckets from around much of the delta. The buckets were filled with ~50 mL of vegetable oil to prevent evaporation of the rainwater. Continuous records of lake-level fluctuations from the lakes were acquired in the thaw-season of 2003 and 2004 using standard pressure-transducer data loggers. Lake area and volume for both lakes was calculated using SURFER™ 8.0 (Golden Software Inc.) based on lakeshore surveys using GPS and lake depth measurements taken along multiple transects.

Two-component mixing model

Volumetric dilution during major hydrological events (e.g., snowmelt, rain events or floods) was determined using a standard two-component isotopic mixing model pioneered by Dincer *et al.* (1970). The technique has been used extensively by the hydrological community to track the components of runoff (e.g., Sklash, 1990; Rodhe, 1998), to trace water in unsaturated soils (e.g., Buttle and Sami, 1990), and to evaluate possible consequences of the new-water/old-water ratio for stream chemistry and surface water quality (e.g., Kendall *et al.,* 1995). More recently, a two-component mixing model was used to separate ground water, surface water plus rainfall, and direct snowmelt in streamflow in a wetland-dominated catchment in the lower Liard River Basin, Northwest Territories (St. Amour *et al*., 2005; Stadnyk *et al.,* 2005).

A two-component mixing model was appropriate to quantify the contributions of source water since the three hydrological inputs (i.e., snowmelt, rainfall, and river water) have appreciably different isotope compositions versus that of the isotopic compositions of the lakes before a hydrological event (generally >8‰). The model assumes only one source of water influences the lake during a hydrological event (i.e., 100% contribution of this component), the isotopic composition of the event component maintains a constant isotopic content, and that unmeasured components (e.g., seeping soil water) do not significantly contribute to alteration of the lake's isotope composition. The model also assumes that the volume of the lake is constant. The mixing model is defined by:

$$\% \text{ dilution} = [(\delta_{Lnew} - \delta_{Lbefore})/ (\delta_{source} - \delta_{Lbefore})]*100 \quad (9)$$

Where, δ_{Lnew} and $\delta_{Lbefore}$ are the $\delta^{18}O$ ($\delta^{2}H$) values of lake water at the time of sampling and before a hydrological event, respectively; and δ_{source} is the $\delta^{18}O$ ($\delta^{2}H$) value of the hydrological event. δ_{source} was either determined, or verified, based on the point at which the isotopic trajectory from $\delta_{Lbefore}$ and δ_{Lnew} intersected with the LMWL in $\delta^{2}H$-$\delta^{18}O$ space (Chapter 4). In cases where there was a prolonged period between sampling of the lakes prior to a hydrological event, $\delta_{Lbefore}$ was inferred. To infer $\delta_{Lbefore}$, a sensitivity analysis was conducted using the known isotopic composition of the lake before and after a significant hydroclimate event. The response of the lake to unknown events was established using weighting factors based on the size of the hydroclimatic event (i.e., rainfall amount and/or period of evaporation loss), and taking into account the timing of sampling. If available, water level data were used to improve judgement of the expected change in the isotopic composition of lake water. Examples of inferring lake water isotopic composition are provided in subsequent text. Uncertainties regarding assumptions of this model are presented in the 'Discussion' section of this chapter.

Spruce Island Lake

Site description

PAD 5, or Spruce Island Lake (SIL; informal name) (58° 50.82' N, 111° 28.84' W) is a small (~23 ha), shallow ($Z_{max\text{-}2003}$ = 0.9 m) lake isolated from all but extreme ice-jam flood events. It is located in a Precambrian Canadian Shield bedrock basin between the Chenal des Quatre Fourches and the Revillon Coupé within the Peace sector of the PAD (Figure 10). SIL is surrounded by mature white spruce and trembling aspen forest on three sides, with only one possible inflow or outflow point via a low-lying wetland with willow thickets to the northeast. A 30 to 40 m wide sedge meadow borders the lake. SIL was nutrient-rich (~350 µg/L total kjeldahl N, ~55 µg/L total P, and ~4500 µg/L dissolved organic carbon), highly oxygenated (generally >85% saturated), alkaline (pH ~9.0), and maintained a warm temperature (~20°C) through most of the 2003 ice-free season. SIL supports extensive growth of filamentous algae and submerged aquatic macrophytes including *Potamogeton* spp., *Ceratophyllum* spp., and *Chara* spp. (Wiklund, in preparation). Cattail and bulrush are limited to shallow nooks between the bedrock outcrops that border most of the shoreline. Multiple lines of paleolimnological evidence such as diatom profiles and isotope modelling indicate standing water has persisted in SIL for at least the past ~200 years, attributable to sufficient direct precipitation and runoff amounts offsetting evaporative losses (Wolfe *et al.*, 2005). Between October 2000 and September 2005 the water level of SIL fluctuated ~0.20 m above and below its maximum thaw-season depth (~1.0 m) based on water mark fluctuations on surrounding bedrock. The lake did not receive river flood waters during the monitoring period.

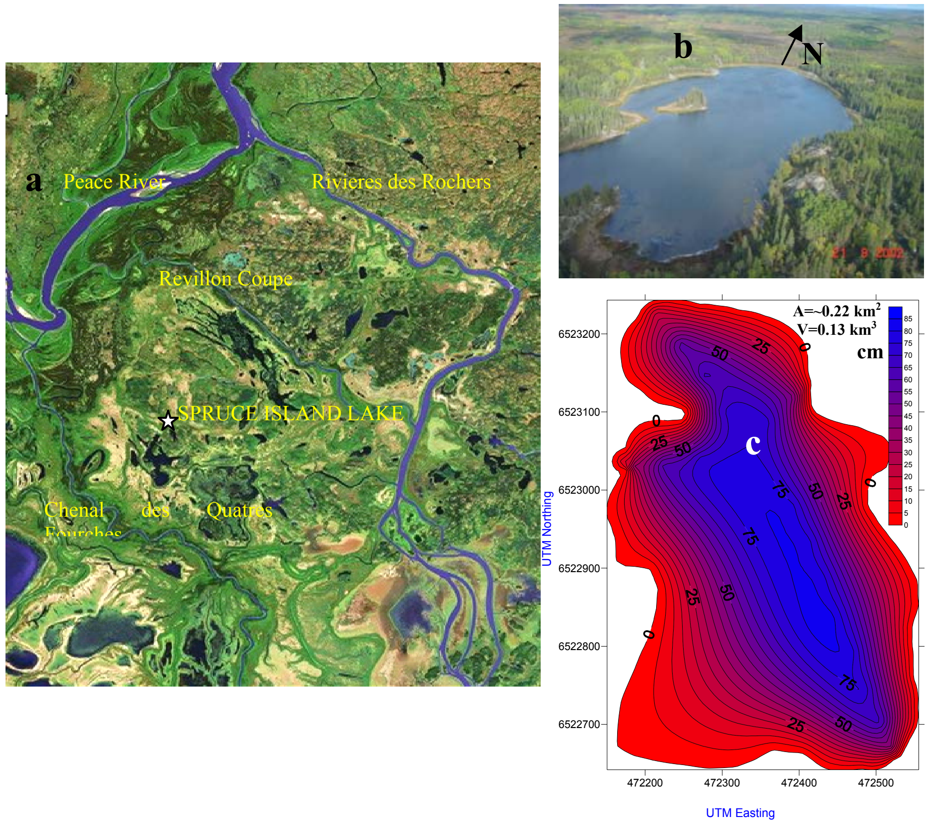


Figure 10: Characteristics of Spruce Island Lake, an isolated lake in the Peace sector. a) Location (GoogleEarth®.ca); b) aerial photograph; and c) basin bathymetry, based on average water depths of 2003.

Results and interpretation

Inter-annual variability: using the isotopic framework

As identified in Chapter 4, superimposing lake water $\delta^{18}O$ and $\delta^{2}H$ data in $\delta^{2}H$-$\delta^{18}O$-space provides a semi-quantitative means for assessing changes in thaw-season lake water balance. The most salient information can be attained by assessing the location of lake water data along, and about the reference Local Evaporation Line (LEL) and in relation to δ_I, δ_{SS} and δ^*. In the case of SIL, data collected between October 2000 and September 2005 (n=20; Appendix G) span much of the LEL (range ~9.6‰ in $\delta^{18}O$; ~50‰ in $\delta^{2}H$), reflecting SIL's strong sensitivity to seasonal hydrological variability

(Figure 11). Many values also cluster close to δ_{SS} suggesting that, during many times of sampling, the lake was in quasi-steady state. Isotopic data are also localized about the predicted LEL (R^2=0.97). This suggests that, on average, SIL captures water ($\delta_{I(SIL)}$ on Figure 11) with a similar isotopic composition to that of the average water captured by the ten other lakes ($\delta_{I(PAD)}$ on Figure 11) sampled regularly in thaw-season 2003. This implies that SIL is an ideal lake to assess some of the input sources affecting lake water balances of lakes in the PAD. Indeed, the estimated isotopic composition of input water based on the intersection point of the regression line through measured lake water and the LMWL (-20‰ in $\delta^{18}O$ and -159.2‰ in δ^2H) compares well with the average of all inputs to the PAD (-19.1‰ in $\delta^{18}O$ and -151‰ in δ^2H). Strong localization of data collected between 2000 and 2005 to the predicted LEL suggests that use of the 2003-based LEL is suitable for assessing the five years of data. This is not surprising considering that the flux-weighted relative humidity and temperature in 2003 (69.7%, 12.7°C) were similar to the average between 2000 and 2005 (68.1%, 12.0°C; Environment Canada, 2006).

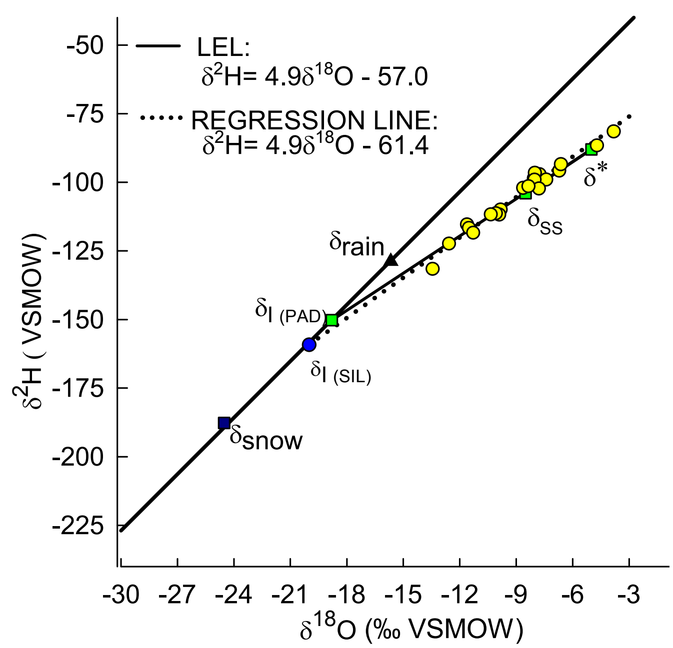


Figure 11: Stable isotope results on a δ^2H-$\delta^{18}O$ plot for SIL collected between October 2000 and September 2005.
Isotopic compositions of SIL are found in Appendix G.

Further inspection of Figure 11 reveals that many data points are slightly offset below and above the LEL. These offsets are mostly attributed to influences by inputs differing from the average delta-wide input to the PAD ($\delta_{I\text{-}PAD}$=-19.1‰ for $\delta^{18}O$, -151‰ for δ^2H). If SIL was only replenished with $\delta_{I\text{-}PAD}$, then during evaporation the isotopic composition of SIL would have been further constrained to the LEL (as it moves towards δ^*; see Chapter 4 for further discussion). As mentioned in the site description above, SIL has two main hydrological inputs: snowmelt and thaw-season precipitation. During the study period, snowmelt influence was observed to dilute the isotope composition of SIL below the LEL because snow ($\delta_{Snow(2003)}$=-24.4‰ for $\delta^{18}O$, -189‰ for δ^2H) plots below $\delta_{I\text{-}PAD}$, further down the LMWL. For example, between the September 23rd, 2004 sample and May 15th, 2005 sample, the isotopic composition of SIL was drawn below the LEL towards δ_{Snow} by 5.1‰ and 30‰ in $\delta^{18}O$ and δ^2H, respectively (Figure 12). On the other hand, displacement of data points above the LEL can be explained by recent influences of summer precipitation ($\delta_{P\text{-}summer(2003)}$=-15.7‰ for $\delta^{18}O$, -129‰ for δ^2H). Following 96 mm of rain between sampling on August 22nd, 2002 and September 21st, 2002 (~double the normal amount for this period based on 1971-2000 climate normals), the isotopic composition of SIL was depleted by 1.4‰ and 6‰ in $\delta^{18}O$ and δ^2H, respectively, following a trajectory towards the isotopic composition of summer rainfall (Figure 12).

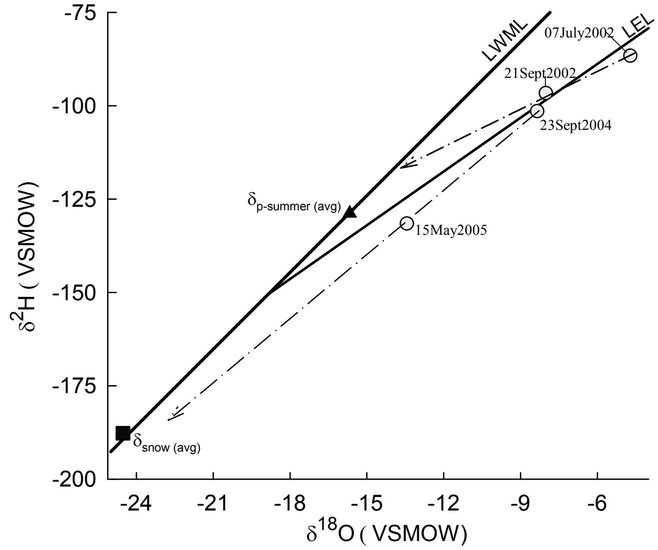


Figure 12. Examples of isotope dilution by snowmelt and rainfall offsetting the isotopic composition of SIL from the LEL.
A line drawn from the isotopic composition of SIL before an event through the value measured following the event should intersect the LMWL near the isotopic composition of the event water. The average isotopic composition of rain and snow are shown. Note: For illustrative purposes, the scales of δ^2H and $\delta^{18}O$ are larger than previous figures.

Inter-annual variability: using timeseries data

Additional insight into controls of water balance of SIL over seasonal time-scales can be gleaned from examining the time-series of $\delta^{18}O$ between 2000 and 2005 (Figure 13; Appendix G). (The time-series of δ^2H is not included because it follows a pattern nearly identical to $\delta^{18}O$). The dotted line on Figure 13 shows the inferred $\delta^{18}O$ composition of lake water during times when samples were not taken and a significant change in water balance was expected to have occurred (e.g., due to a rainstorm, an extended period without any rain, or flooding in the case of Johnny Cabin Pond). Although the main reason for estimating $\delta^{18}O$ during times when measured data did not exist is to help with understanding the hydrological behaviour of a lake, it also demonstrates another utility of water isotope data. To show how $\delta^{18}O$ was inferred, two examples are provided below. The first example is provided to demonstrate how the isotopic composition of SIL was inferred before lake freeze-up in 2003, a value that is especially important for assessing the impact of snowmelt once ice thaws in the following spring. It was determined that $\delta^{18}O$ of SIL prior to freeze-up (estimated to be around November 30th) would have dropped by ~0.7‰, having a final value of -8.7‰ compared to the last sample taken on September 23rd, because the PAD received ~80 mm of rain and underwent ~35 mm of evaporation. The second example describes the inferred $\delta^{18}O$ change of SIL between the samples taken on July 16th and September 23rd 2004. It was determined that SIL would have changed from -7.8‰ to -5.7‰ by August 15th due to evaporative enrichment. However, due to isotopic depletion on July 30th by a 32.5 mm rainstorm with an isotopic composition of ~-15‰ the value was adjusted to -6.9‰. Then between August 15th and September 23rd, lake water $\delta^{18}O$ was expected to enrich to ~-5‰, but a 40 mm rain event between September 18th and 19th with a measured composition of ~-17‰ ultimately left the $\delta^{18}O$ composition of SIL at ~-8.4‰. The time-series of GSL is also included on Figure 13 as a comparison of a nearby terminal lake in long-term (i.e., ~decadal) hydrologic and isotopic steady-state (i.e., δ_{SS} as described in Chapter 3). Note that $\delta^{18}O$ of GSL remains relatively stable over the five years of data, but tends towards a slightly more depleted $\delta^{18}O$ composition. The slight depletion may reflect an increase in contributions by an isotopically depleted source, which will be addressed in the 'Discussion' section of this chapter.

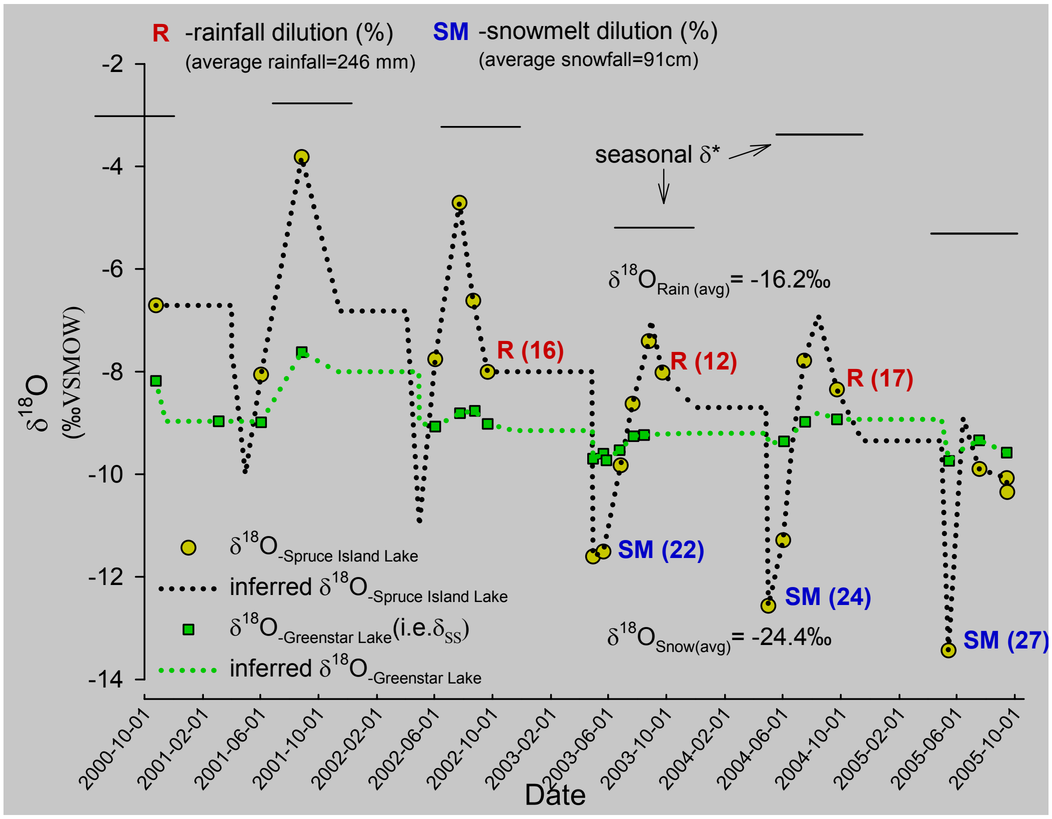


Figure 13. Five year time-series of the δ^{18}O composition of SIL.
Included is the inferred isotopic composition of lake water based on knowledge of climatic conditions between sampling and the response of SIL to inputs (see text for explanation). Year-specific δ^* values are provided to help assess the degree of evaporation SIL underwent and was calculated using equations 5, 6, and 7 (Chapter 3) based on year-specific h and T and δ_A modelled using GSL's thaw-season average δ_L value. Isotopic compositions of SIL are located in Appendix G.

The isotopic evolution of SIL in any one thaw-season exhibited a fairly systematic cycle beginning with rapid isotope-dilution by snowmelt following ice-off in late-April or early-May, followed by heavy-isotope enrichment caused by evaporation which was often punctuated by isotopic dilution from rain events. This pattern of alternating isotopic dilution and enrichment has been found in other small, isolated lakes in northern Canada (Gibson *et al.*, 1996b; Gibson, 2002a). Rapid isotopic enrichment was common during dry periods such as between the June and August 2001 samples when precipitation was less than 65% of climate normals (1971 to 2000; Environment Canada, 2006).

Nevertheless, strong enrichment due to evaporation was often offset by ^{18}O-depleted rain events. For example, in 2002 multiple influential rain events (i.e., 10 were >5 mm; Environment Canada, 2006) through July and August caused SIL's late September $\delta^{18}O$ composition to return to the early thaw-season composition observed in early-June. Actually, elevated rainfall (and consequently high relative humidity) in the thaw-season of 2005 (280 mm total; 122 mm of which occurred in August; Environment Canada, 2006) was likely the reason that the $\delta^{18}O$ signature of SIL remained slightly more depleted than GSL, the reference lake in hydrologic and isotopic steady-state (Figure 13). The latter likely indicates that rainfall and snowmelt exceeded evaporation for much of the 2005 thaw-season.

Prolonged catchment release of snow and rain and direct rainfall on SIL also appear to play a role in controlling the long-term isotopic evolution. Upon more detailed inspection of Figure 13, it can be observed that the $\delta^{18}O$ values of SIL (and GSL) became slightly more depleted over the duration of the study. This observed trend is mostly attributable to higher contribution by isotopically-depleted input sources (i.e., wetter conditions) since temperature and relative humidity remained relatively consistent between 2000 and 2005, except for 2000 and 2004 when relative humidity was lower by ~3% (Table 2). Lower relative humidity in these years would actually serve to increase evaporative enrichment and counter the isotopically depleted inputs, making it more plausible that wetter conditions were the cause of the long-term trend towards more isotopically-depleted values. Since rainfall did not have a significant linear increase, it is expected that snowmelt is the reason for a net decrease in $\delta^{18}O$ by 2005. And, in fact, between 2000 and 2005 there was a general increase in snowfall between October 1st and

May 1st (Table 2). Snowfall amounts are as follows: 2000-63 cm, 2001-76 cm, 2002-87 cm, 2003-111 cm, 2004-94 cm, and 2005-110 cm. Using an accepted 7.5 to 1 snow to

Table 2: Hydroclimatic conditions of 2000 to 2005: Temperature, relative humidity, rainfall during thaw-seasons (i.e., May 1st to October 31st), and yearly snowfall.

	T_{fw} (°C)	h_{fw} (%)	Snowfall* (cm)	Rainfall (mm)
2000	12.0	65.7	63	254
2001	13.6	69.0	76	189
2002	11.8	68.3	87	253
2003	12.7	69.7	111	209
2004	10.8	66.0	94	194
2005	11.4	69.7	110	293
Average	**12.0**	**68.1**	**90**	**232**
Normal	**12.5**	**68.2**	**69**	**246**

Evaporation flux-weighting (fw) was conducted on monthly values of temperature and relative humidity (Environment Canada, 2005) using the Thornthwaite method (1948). Climate normals are based on 1971-2000 data (Environment Canada, 2005). Environment Canada uses a snow water equivalent between 7 and 10 based on snow density from snow surveys in the area. *Snowfall accounts for snow that fell in the previous autumn of a year (i.e., 2000 snowfall is made up of 40 cm in autumn 1999 and 23 cm in spring 2000)

water equivalent ratio in northern environments (Young *et al*., 2005), then melting snow would have contributed approximately 47, 57, 65, 83, 71, and 83 mm of water each year, respectively. The fact that SIL retained water for at least five years and maintained a relatively stable water balance (i.e., tracked a nearby lake in hydrologic steady-state) without flood waters is contrary to results by Prowse *et al.* (1996), Pietroniro *et al.* (1999), Peters (2003), and Peters *et al.* (2006) who predicted that without replenishment by flood water, perched basins should dry out in between four to nine years depending on hydroclimatic conditions. Conversely, maintenance of water in SIL is consistent with results by Wolfe *et al*. (2005), which indicated that standing water has persisted in SIL for at least the past ~200 years. Upon further assessment of Prowse *et al.* (1996), Pietroniro *et al.* (1999), Peters (2003), and Peters *et al.* (2006) it appears that the role of snowmelt and rainfall runoff were neglected as they were believed to be minor components of the water balance because of the flat bathymetry and narrow levees surrounding most basins. However, many of the perched basins in the PAD are surrounded by bedrock, forests, and in some cases wetlands, all of which should contribute runoff. Recognition of the importance of snowmelt and rainfall as direct inputs, and as runoff, stimulated an interest in quantifying the effects rain and snow on controlling SIL's seasonal water balance. The following section includes an application of the isotope mixing model to quantify the amount of dilution following selected snowmelt and rainfall events, and is followed by a discussion of the results.

Quantifying the role of snowmelt and rainfall on Spruce Island Lake's water balance

Table 3 shows the percent dilution of SIL following examples of snowfall and rainfall events. Ascertaining δ_I specific to each event period was determined by the intersection point on the LMWL of a line from δ_{before} through δ_{new} in δ^2H-$\delta^{18}O$ space, as illustrated in Figure 12. Table 3a shows the percent of replenishment that SIL underwent during the 2003, 2004, and 2005 snowmelt periods, and compares these values with cumulative snowfall during the winter prior to sampling. The average replenishment by snowmelt during these three freshet periods is 28%, which is in close agreement with estimates of the contributions to streamflow by snowmelt using water isotope tracers in two small, wetland-dominated basins in Fort Simpson, Northwest Territories, Canada (25-30%, Gibson *et al.,* 1993b; 25-44%, Hayashi, 2004). This translates to an average increase of 0.13 m in water level of SIL utilizing the volume function in Surfer 8.0, which corresponds well with the common height of recent high water marks (0.15 m) observed on the bedrock surrounding two sides of the lake during sampling in May and June 2003 and 2004. The higher water level height increase (0.13 m) versus the available water by average snowpack (79 mm or 0.079 m) is attributed to additional snowmelt contributed from the bedrock, forest, and from the wetland at the northeast side of SIL. Contributions of snowmelt from the forest and wetland may also explain why there is not a direct correspondence between dilution amount and snowfall accumulation (Table 3a). For example, a slightly lower snowfall amount in winter 2004 as compared with 2003 and 2005 led to more dilution of SIL. This may be explained by antecedent conditions of the forest and wetland from the previous autumn. In the case of the higher replenishment in 2004, it was likely that saturation of the forest and wetland caused by 96 mm of rain that fell between August 22nd and September 21st in 2002 caused more snowmelt to drain directly into the lake.

Table 3. Estimates of dilution of SIL following (a) snowmelt, and (b) rainfall events.

a)

Date	**Snowfall (cm)** **(Water equivalent in mm)**	δ_{Lnew} **(‰)**	$\delta_{L\text{-}before}$ **(‰)**	δ_{source} **(‰)**	$\delta_{snowmelt\text{-}average}$ (‰)	**Dilution (%)**
May 01 2003	111 (83)	-11.6	-8.0	-21.3	-24.4	27
May 02 2004	94 (71)	-12.6	-8.7	-21.4	-25.2	30
May 16 2005	110 (83)	-13.4	-8.4	-28.1	-23.7	26

b)

Interval	**Rain (mm)**	δ_{Lnew} **(‰)**	$\delta_{L\text{-}before}$ **(‰)**	δ_{source} **(‰)**	$\delta_{rain\text{-}average}$ (‰)	**Dilution (%)**
Aug 22 to Sept 21 2002	96	-8.0	-6.6	-13.1	---	21
Aug 25 to Sept 23 2003	43	-8.0	-7.0	-13.0	-15.4	11
Aug 15 to Sept 23 2004	71	-8.4	-6.9*	-15.9	-15.3	16

* inferred to account for changes in isotopic composition due to evaporation and input from the last sampling; see text.

Table 3a also compares the isotopic composition of measured snowmelt water and modelled isotopic composition of input water to SIL during springmelt. Slightly more enriched modelled δ_{Source} values in 2003 and 2004 compared to measured snowmelt compositions, may be caused by isotopically-enriched rain storms in late 2002 and 2003 being "frozen-in" to the lake and the wetland to the northeast. In fact, Gibson *et al.* (1993b) noted 4 to 5‰ more enriched $\delta^{18}O$ values in snowmelt runoff in pipes and rills through wetlands and hillslopes as compared with snowpacks, and attributed this to mixing of the pre-event subsurface water in micro-depressions on the hillslopes. On the other hand, modelled δ_{Source} in 2005 was more depleted (-28.1‰ for $\delta^{18}O$) than the average measured snowfall for that year. This is likely because sampling was conducted in mid-May (two weeks after the 2003 and 2004 samples were taken) when most of the snow was already thawed and drained from the surrounding forest. During this period, 23 of the 30 days prior to sampling were greater than 0°C (Environment Canada, 2006), conditions conducive to strong snowmelt. It is likely that modelled δ_{Source} is more representative of the snow that melted into SIL than the average measured value (-23.7‰ for $\delta^{18}O$) because some of the snow samples that were used to derive the average value were taken from on top of lake ice in March, where snow is susceptible to sublimation and subsequent isotopic enrichment. Indeed, a protected sample taken from 15 to 25 cm snowpack depth in March 2005 from the forest surrounding a lake close to SIL (PAD 54) was found to have an isotopic composition of -27.5‰ for $\delta^{18}O$, very close to the modelled δ_{Source}. Premature sampling of SIL before most of the surrounding snowpack (especially in the forest) had melted in spring of 2003 and 2004 suggests that estimates for snowmelt dilution in these two years are slightly underestimated. As such, it is suggested that future sampling of lakes for assessment of snowmelt contribution be conducted after much of the ice has melted from the lake, and much of the snow has melted from the contributing catchment.

Table 3b shows the percent dilution of SIL following three intervals with substantial rain. The average replenishment amount during these periods is 16%. Utilizing the volume function in Surfer 8.0, this translates to a rise in SIL water level of ~0.09 m (or 90 mm), on average, during these periods of rain. There is evidence that SIL is sensitive to antecedent moisture conditions of the wetlands to the northeast, and the size of the rain event. The highest dilution in a sampling period was 21%, which was found after SIL received 71 mm of rain between August 15th and September 23rd, 2004 (Environment Canada, 2006), where much of this rain (46 mm) occurred during a four-day period between September 18th and 22nd. This temporarily raised the water level of SIL by 43 mm based on water level recorder data. The corresponding amount of rain and water level height increase is consistent with observations of already saturated wetlands surrounding SIL due a high amount of rainfall in July (85.5 mm; ~double climate normals; Environment Canada, 2006), allowing for maximum influence of direct rain on SIL. Other studies in northern Canada have shown that the magnitude of runoff from wetland areas increases as the storage capacity in the substrate decreases (Woo and Young, 1998). Further discussion of the sensitivity of SIL to rain events is presented in the subsequent section during short-term assessment of the thaw-season of 2003.

Results up to this point show that SIL is responsive to snowmelt and thaw-season rainfall, and that it likely relies on both these sources to maintain its water balance *between* years. Based on quantitative analysis, it is shown that snowmelt is a more influential input to SIL than rainfall. This is consistent with previous studies in sub-arctic regions which found that snowmelt played a dominant role in maintaining water in shallow aquatic basins (Spence and Woo, 2003; Hayashi *et al.*, 2004). However, there is a difference between the effectiveness of 1 mm of water from snowmelt than that of rainfall. In Table 3, 10 mm of snowmelt water equivalent would dilute SIL by 3.5%, whereas 10 mm of rain water would only dilute SIL by 2.3%. This could be explained by misrepresentation of $\delta_{L\text{-before}}$ and $\delta_{L\text{-new}}$. Samples were not always collected immediately before or after a rainfall event and in some cases it was likely that the isotopic composition of the lake enriched, for example, between the time of the rainstorm and when the sample was collected. Nevertheless, it is expected that rain would be less effective because some of the rain would be absorbed and intercepted by the forests and wetlands surrounding SIL, whereas snow would effectively drain from soils and wetlands still wet from the previous autumn rainfall. To understand the effectiveness of snowmelt and rainfall the following section provides a more detailed analysis of the short-term isotopic variability of SIL. The analysis also helps to demonstrate how the water budget of SIL is maintained *within* a thaw-season, which may provide insight into why SIL is able to maintain water over longer time periods.

Intra-annual variability: isotope-mass balance modelling for 2003

Between the sampling period following the initial snowmelt pulse (~May 1st) and August 25th of 2003 (~70% of the ice-free season), evaporation of 275 mm (Figure 6; Appendix A) in the PAD exceeded total rainfall of 142 mm (Environment Canada, 2006), yielding a net water loss of 133 mm. However, the $\delta^{18}O$ value of SIL measured on August 25th of 2003 reverted back to the $\delta^{18}O$ value prior to permanent ice-cover in 2002, before it was affected by snowmelt (Figure 14). For this to be accomplished, the initial input pulse by snowmelt would need to have contributed the 133 mm needed to balance the water budget. Based on results presented in the previous section, dilution by snowmelt during 2003 was estimated to be ~27% (Table 3), which translates to ~110 mm of water using the volume function in Surfer 8.0. As previously mentioned, it is thought that the estimation of water level increase exceeds snow water equivalent in 2003 (i.e., excess of ~63 mm) due to additional direct runoff contribution from the surrounding bedrock and wetlands. Since the water balance of SIL over an entire water year was closed by adding the contribution of snowmelt, this suggests that snowmelt is critical for maintaining water in perched basins.

Strong sensitivity of SIL's water level to a few significant rain events in 2003 was observed (Figure 14). For example, the water level of SIL rose rapidly and dropped extremely fast following two heavy rainfall episodes on August 4th and August 17th (i.e., water level was back to its pre-event level 1 to 2 days after the storms ended; Appendix H). The large increase in water level is most attributable to direct precipitation on the lake plus rapid runoff generated bedrock-covered areas of the catchment. However, evaporation does not account for the rapid decrease in water level back to its original pre-rainfall height. This could possibly be attributed to absorption by bordering wetlands (bogs) after the rainfall has stopped (i.e., a "sponge" effect). This process would be accelerated by the temporary increase in water level difference between the lake and adjacent wetland to the northwest following a large rain event. Wetlands to the northeast of SIL were commonly found to be saturated, even following extended periods without rainfall. Although further testing is required, hydrological buffering by wetlands may provide an explanation as to why many of the basins perched in the Peace sector having similar characteristics to SIL (i.e., perched basins surrounded by or connected to wetlands) have retained water over the past five years, as observed since isotopic monitoring was initiated. This phenomenon of regulation by wetlands may also explain why water has remained in this basin for (at least) the past 200 years (Wolfe *et al*., 2005). Similar behaviour and patterns of wetland connectivity have been observed by Quinton *et al.* (2003), Hayashi *et al.* (2004), and Stadnyk *et al*., (2005) in similar boreal environments.

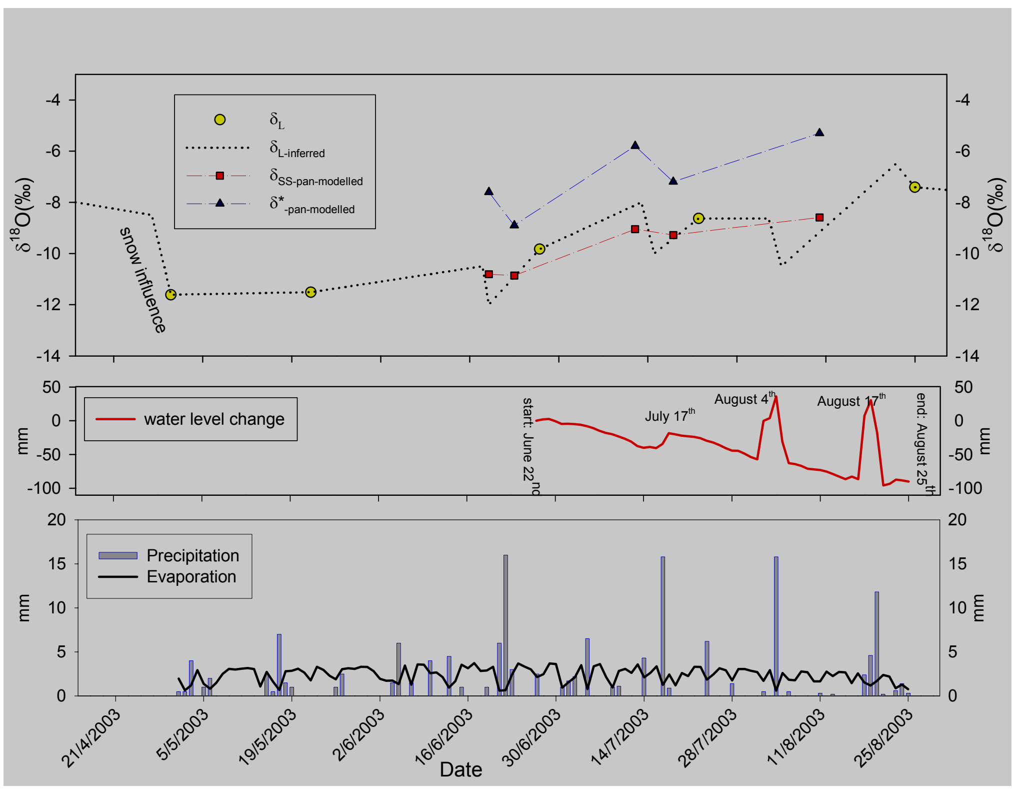


Figure 14: Stable isotope results for Spruce Island Lake collected in 2003 compared with hydroclimatic conditions and water balance of the lake.
Isotope data can be found in Appendix G, and water level data can be found in Appendix H.

The previous two paragraphs have shown that the water balance of SIL is sensitive to periodic meteorological and hydrological events. Much of the sensitivity of SIL to these events may be due to its shallow depth and small volume relative to surface area. It is these attributes that make SIL sensitive to short-term variability in atmospheric conditions, as compared with GSL, for example, while their isotopic responses over the longer term are remarkably similar. To demonstrate how varying atmospheric conditions can affect the interpretation of SIL's isotopic composition, isotope parameter modelling results (δ_{SS} and δ^*) from the evaporation pan maintained in 2003 (Table 1 and Figure 8 from Chapter 4) are also shown on Figure 14. Between the period that the evaporation pan was in apparent steady-state (June 19th to August 10th), the measured isotopic composition of SIL remained close to the composition of the pan. This means that SIL was responding to transient changes in atmospheric conditions and maintaining steady-state over short periods of time (weeks to months). Knowledge of short-term variability in δ_{SS} and δ^* is important when evaluating water balance conditions of shallow basins that are known to be sensitive to atmospheric changes because use of long-term average values may cause generalizations of a lake's water balance record.

SIL's response to temporary changes in atmospheric conditions prompted an interest in addressing the samples that plotted in a cluster between δ_{SS} and δ^* (or beyond δ^* in a few cases) on Figure 11 in δ^2H-δ^{18}O-space because this indicates strong evaporation (i.e., net evaporative loss) prior to sampling. However, the water level of SIL during sampling was rarely found to be significantly different than it was on any previous visit. During one of the extreme cases, the isotopic composition of SIL (-3.8‰ and -81‰ in δ^{18}O and δ^2H, respectively) in August 2001 exceeded δ^* based on 2003 hydroclimate data (-5.5‰ for δ^{18}O, -83‰ for δ^2H), which is theoretically impossible since δ^* represents the isotopic composition of the final residual of water in a lake prior to desiccation. This would imply that SIL was about to completely desiccate, yet in reality, the water level was only slightly lower than it had been on the previous visit in mid-June. This deviation from the 2003-derived isotopic framework suggests that SIL, and all other PAD study lakes, in 2001 evolved isotopically under hydroclimatological conditions which differed from 2003 conditions. During the two weeks prior to sampling, the relative humidity was low (67.7%), and rainfall was very low (18 mm; Environment Canada, 2006). These conditions would have temporarily caused the isotopic compositions of δ_{SS} and δ^* to be higher. Indeed, based on these hydroclimate conditions, the modelled compositions for δ_{SS} and δ^* leading up to the August sample were calculated to be -4.1‰ (-83‰) and -0.2‰ (-64‰), respectively, based on equations (7) and (5), respectively. Because the measured isotopic composition of SIL in August 2001 was similar to the revised δ_{SS}, it can be assumed that SIL would have lost only a small fraction (i.e., <5%) of its water to evaporation prior to sampling in 2002. As such, it is expected that in all cases where the isotopic composition of a sample implied that SIL was undergoing strong evaporation, that in fact the weeks leading up to sampling were warmer and/or drier than average conditions used to develop the LEL on the isotopic framework. These results emphasize the need to account for short-term variability in atmospheric conditions when assessing water balance conditions of all shallow basins in the PAD. This is noted to be especially prevalent following arid periods (i.e., low relative humidity, and/or low rainfall). Similar observations and conclusions were made by Brock *et al.* (2007) for a number of shallow floodplain lakes in the Slave River Delta.

Johnny Cabin Pond

Site description

Johnny Cabin Pond (JCP; informal name) (58° 29.84' N, 111° 31.15' W) is a small (~25 ha), shallow ($Z_{max-2003}$=1.4 m), low-lying lake separated to the east from Mamawi Creek by a 0.5 to 1.5 m levee and ~200 m of young forest within the Athabasca sector of the PAD (Figure 15). Shallow wetlands and floating vegetation border the remainder of the lake. JCP has become more susceptible to flooding by Mamawi Creek because of the Embarras Breakthrough in 1982, which is routing increasing volumes of water into Creed and Mamawi creeks as a consequence of northeastward over-extension of the Athabasca sector of the PAD (PAD-TS, 1996; Wolfe *et al*., in review). Low hydraulic conductivity clays and silts on Mamawi Creek's levee prevent the exchange of creek and JCP lake water under normal flow conditions. In 2003, JCP was nutrient-rich, relatively alkaline ($pH_{average-2003}$= 9.22), maintained a warm temperature ($T_{average-2003}$=20.2°C) and supported extensive submerged weeds including *Potomogeton* spp., *Ceratophyllum* spp., and *Myriophyllum* spp. (Wiklund, in preparation). JCP's water level was observed to have risen following high river stages during the monitoring period, including spring 2003 and 2005, and summer 2001 and 2004. Evidence that JCP floods periodically includes 1) the absence of mature trees in the extensive lowlands surrounding the lake, 2) bordering grass, cattail, willow, and a mixture of dead and living reeds, and 3) adventitious rooting on submerged shoots of willows. Multi-proxy analyses of a sediment record (e.g. grain size analysis, isotope data, diatoms) from JCP have also shown that water levels have varied over the past ~300 years, attributed to periodic river influence (Hall *et al*., 2004). Observations by Prowse et *al.* (1996) have also indicated that JCP's water level is more stable in dry years as compared to some other lakes in the Athabasca sector of the PAD.

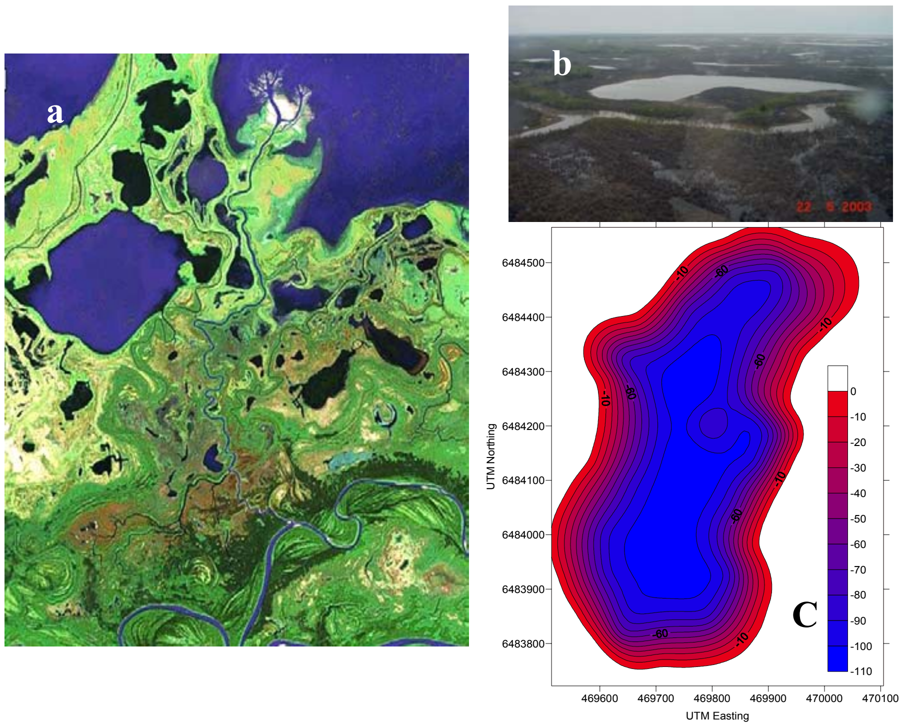


Figure 15: Characteristics of JCP, a low-lying lake in the Athabasca sector.
a) Location (GoogleEarth®.ca); b) aerial photograph; c) and basin bathymetry (cm), based on average water depths of 2003.

Results and interpretation

Inter-annual variability: using the isotopic framework

Results from stable isotope analyses on water samples from JCP taken between October 2000 and September 2005 (n=27; Appendix G) span a broad range of values (~8‰ in $\delta^{18}O$; ~42‰ in $\delta^{2}H$) in $\delta^{2}H$-$\delta^{18}O$ space, reflecting its strong sensitivity to hydrologic variability (Figure 16). Most of the data plot between δ_I and δ_{SS} along the LEL, with the majority clustering around the average isotopic composition of Mamawi Creek of -17.8‰ and -143‰ for $\delta^{18}O$ and $\delta^{2}H$, respectively. To confirm that the main input source to JCP was from periodic river water influence from Mamawi Creek, the intersection point of the regression line drawn through lake water data and the LMWL was determined. This was found to occur at -17.9‰ and -144‰, which is essentially identical to Mamawi Creek's isotopic composition.

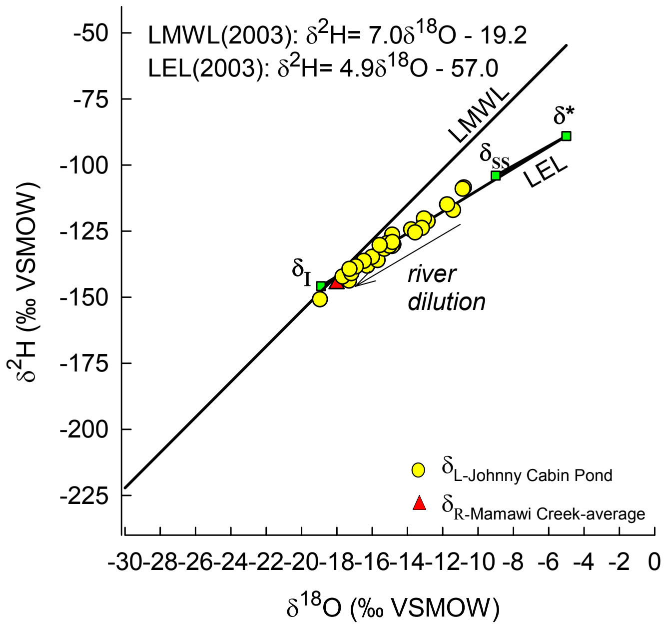


Figure 16: Stable isotope results on a δ^2H-$\delta^{18}O$ plot for JCP collected between October 2000 and September 2005. Isotopic compositions of JCP are located in Appendix G.

JCP was thought to have flooded during the summer of 2001 based on wet conditions in the area, and was witnessed flooding again during the spring of 2003. Figure 17 shows the isotopic compositions of JCP before ($\delta_{L\text{-before}}$) and after ($\delta_{L\text{-new}}$) these events (similar isotopic changes occurred during the floods that occurred in summer 2004 and spring 2005). The close correspondence between δ_{new} and the isotopic composition of Mamawi Creek water ($\delta_{L\text{-source}}$), as well as the large difference from $\delta_{L\text{-before}}$, indicate that water in JCP was almost entirely replaced by river water during these floods. During the 2003 spring sampling period, JCP was observed to be directly connected to Mamawi Creek via low-lying ground to the west (Figure 18). The spring 2003 flood event was

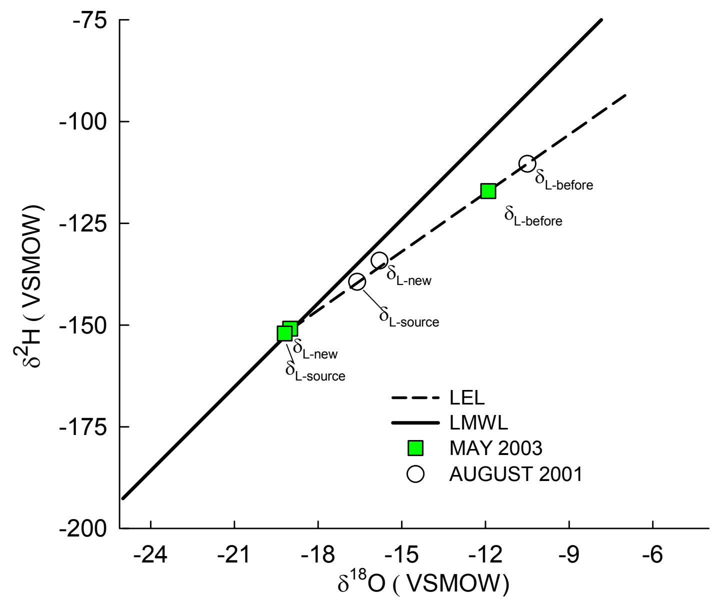


Figure 17: Examples of identifying spring and summer flood events in JCP using the isotopic framework.
A line drawn from the isotopic composition of JCP before an event through the value measured following the event should intersect the LMWL near the isotopic composition of the event water. For illustrative purposes the scales for $\delta^{18}O$ and $\delta^{2}H$ are smaller than other figures.

attributed to backwater effects caused by ice-jams along tight meanders on the Athabasca and Embarras Rivers. In spring 2005, the area surrounding JCP was also highly saturated as it had been in 2003, and thus it was assumed that localized flooding reoccurred in the vicinity of JCP. On the other hand, during the periods surrounding sampling of JCP in the summers of 2001 and 2004, there was no local knowledge of flooding in the Athabasca sector, but isotope data of JCP suggest otherwise. This illustrates the strength of water isotope tracers for assessing flood timing and severity. Such summer flood events are likely due to high river stages on Mamawi Creek which can result from delayed snowmelt in upstream catchments of the Athabasca River, shown to have occurred in the past (Peters, 2003); or sudden, high-intensity rainfall events in the upstream regions of the Athabasca River (i.e., ~200% of the 1961-1990 average between

July 1st and July 31st 2001 and 2004 in the Lesser Slave Lake region; AE, 2006). It should also be noted that the heavy rainfall events surrounding summer flood events likely played a role in diluting the isotopic composition of JCP since the PAD itself received 15 mm of rain five days prior to sampling in August 2001, and 31 mm of rain over the ten days prior to sampling in July 2004 (Environment Canada, 2006). Actually, on Figure 17, the August 2001 sample plots slightly above the LEL, illustrating that rainfall likely did play a small role in the JCP water balance. However, river water would have influenced JCP much more than rainfall based on the fact that the isotopic composition of JCP was almost identical to that of Mamawi Creek, and that the capability of rain to replenish lake water has been shown to be only moderate (i.e., <25%) based on results from analyses of SIL, which has a similar surface area and volume as JCP. Indeed, similar to SIL, it was observed that the isotopic composition of JCP during non-flood periods was occasionally offset above and below the LEL on Figure 16. This demonstrates that JCP is not exempt from temporary influences by rainfall and snowmelt during non-flood conditions. Nevertheless, because the isotopic composition of JCP is commonly drawn back to the isotopic composition of Mamawi Creek (2001, 2003, 2004, and 2005), it appears that river water is in fact the dominant input to JCP.

Figure 18: Photograph showing turbid river water cascading over the west bank of Mamawi creek into JCP.
Photograph was taken on May 1st, 2003, courtesy of Dr. Tom Edwards, University of Waterloo.

Inter-annual variability: using time-series data

Temporal changes in the water balance of JCP can be gleaned from analyses of $\delta^{18}O$ between 2000 and 2005 (Figure 19; Appendix G). The four flood events of spring 2003 and 2005, and summer 2001 and 2004 are observed as depressions of $\delta^{18}O$ close to the $\delta^{18}O$ value of Mamawi Creek. In fact, it is probably because of these floods that the $\delta^{18}O$ value of JCP remained more depleted than the isotopic composition of a closed-basin lake in long-term hydrologic steady-state (i.e., GSL; Figure 19) between 2000 and 2005. The fact that JCP maintained a neutral water balance does not discount the role of evaporation in its water balance. Pronounced isotopic enrichment did occur during periods without significant water input, indicating a strong evaporative flux out of JCP. For example, between June 3rd and September 21st of 2002 (a non-flooding year), JCP underwent more net $\delta^{18}O$ increase (+4‰) than SIL (-0.3‰). Differences in enrichment behaviour of JCP and SIL are discussed at the end of this chapter.

Much of the above results and discussions have been qualitative in nature. Although qualitatively assessing lake water isotope data provides valuable hydrological information, studies on JCP and other flood-susceptible lakes in the PAD would be much more informative if it was known *how much* of the lake volume was replenished by river water. As such, the following section involves quantitatively estimating the replenishment of JCP by spring and summer flood events using the isotope mixing model (equation 9).

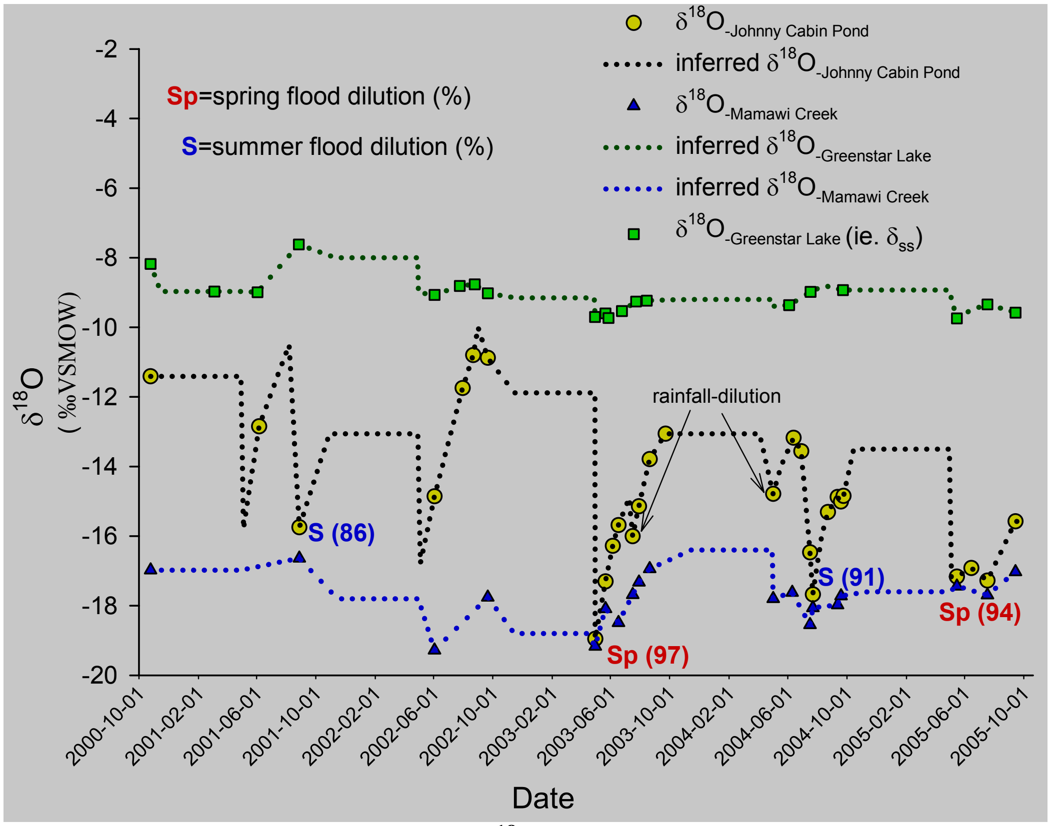


Figure 19: Five year time-series of the $\delta^{18}O$ composition of JCP.
Included is the inferred isotopic composition of lake water based on knowledge of climatic conditions, the $\delta^{18}O$ composition of Mamawi Creek, and a closed-basin lake in hydrologic and isotopic steady-state (GSL, Chapter 4). Isotopic compositions of JCP are located in Appendix G.

Quantifying the role of river flooding on Johnny Cabin Pond's water balance

The estimated degree of replenishment caused by the spring and summer floods can be found in Table 4. Since it is assumed that river water dominates as an input source during flooding, the isotopic composition of Mamawi Creek at the time of flooding was used as δ_{source}. Mamawi Creek's composition was also used instead of the 'intersection-approximation' (employed for determining event-specific δ_I for SIL) because the isotopic composition of Mamawi Creek sometimes plotted slightly up the LEL and away from the LMWL (δ_{source} for August 2001 on Figure 17). As can be seen in

Table 4, calculations show that water in JCP was almost entirely replaced by river water during all flood events. This was not surprising for the spring floods since the area surrounding JCP had been observed to be entirely inundated with river water due to backwater effects from ice-jams on the Embarras River. However, the fact that summer flood events were also capable of flushing almost all of JCP's existing volume is an important finding. First of all, this means that JCP and other low-lying lakes would be rejuvenated with water up to their controlling sills, which would reduce encroachment of willows and other terrestrial plants. Secondly, summer flooding would likely supply fresh nutrients and minerals utilized by aquatic biota. This latter point is being investigated (Wiklund, in preparation). Unlike spring events, in which murky waters inundate large expanses of land for days at a time, and ice-jams which are highly visible and easy to locate, summer floods may be isolated or short-lived. The only evidence of a summer flood event may simply be a significant increase in water level of a lake; an observation that on its own, however, is difficult to attribute to a specific cause. Hence, most small- to moderate-sized summer floods have likely gone undetected in the past because they have not been visually observed or documented. Identifying and quantifying summer flood events are yet two more strengths of using water isotopes in the PAD.

Table 4. Estimates of dilution of JCP following high water events on Creed Creek.

Month/Year	δ_{Lnew} (‰)	*$\delta_{L\text{-}before}$ (‰)	**δ_{source} (‰)	Dilution (%)	Notes
August 2001	-15.8	-10.5	-16.6	85.5	Discharge of Embarras River rose 3.5 times its July 23rd value between July 23rd and August 5th. Mamawi Lake rose ~0.6 m.
May 2003	-19.0	-11.9	-19.2	97.1	Ice-jams on Embarras and Athabasca River flooded regions along Mamawi Creek. Note: Mamawi lake rose ~1.9 m between April 15th and May 4th.
July 2004	-17.7	-13.6	-18.1	91.4	High precipitation in sub-catchments along the Athabasca River. Creed Creek rose ~1.4 m between July 8th and July 19th. Mamawi Lake rose ~1.3 m.
May 2005	-17.2	-13.5	-17.4	93.9	Ice-jams on the Athabasca River backed-up water and consequently caused major flooding along Mamawi Creek and tributaries.

*value may be inferred to account for evaporative-enrichment or precipitation-dilution since the last available sample.

**δ_{source} is measured, not modelled as is the case for SIL.

It should be noted that flood intensity estimates are more accurate when the isotopic compositions of JCP and Mamawi Creek are measured at the same time (as they were in this study) since the isotopic composition of Mamawi Creek has been found to differ slightly from spring to summer (Table 4). The average summer composition of Mamawi Creek between 2000 and 2005 was -17.5‰ for $\delta^{18}O$ and -141‰ for $\delta^{2}H$, whereas, the average spring composition was -18.4‰ and -147‰. The more depleted isotopic composition in the spring is caused by snowmelt contributions originating from the upstream reaches of the Athabasca River, whereas in the summer, slight evaporative-enrichment and comparatively isotopically-enriched rainfall and runoff would cause Mamawi Creek to have a more enriched isotopic signature. This complex hydrological regime and isotopic differences in summer and spring is indicative of many of the rivers in the Athabasca sector. So care should be taken when assuming the isotopic composition of the tributaries is the same as the main river in a system.

Intra-annual variability: example of a spring flood

Figure 20 shows $\delta^{18}O$ values of JCP in 2003 as compared to the water level of Mamawi Lake, and daily evaporation and precipitation in the PAD. Because Mamawi Creek drains directly into Mamawi Lake, the latter is used as a surrogate water body for analysis of water levels in Mamawi Creek for 2003. Mamawi Lake's high water level on May 1st, 2003 (2 m rise in the previous 11 days) corresponded with field observations which documented major volumes of the Athabasca River flowing north into Mamawi Lake, consequently inundating much of the Athabasca sector north of the Embarras River (including JCP). However, Mamawi Lake's water level during this time (210.0 m.A.S.L.) was likely lower than Mamawi Creek's since the lake has a large holding capacity and drains quickly via a multitude of channels. Flooding by Mamawi Creek caused the $\delta^{18}O$ value of JCP to become $\delta^{18}O$-depleted by 7‰ from its composition prior to freeze-up in 2002, resulting in a dilution of ~97%. Perhaps more interesting though is that during the remainder of the 2003 thaw-season, JCP underwent an equal degree of $\delta^{18}O$-enrichment, as observed by similar isotope compositions of the pre-May 1st sample and the September 23rd sample. If the evaporative-enrichment trend had not been offset by rainstorms, including a 17 mm rainstorm on July 17th, the lake would have likely undergone around 8‰ enrichment, exceeding the initial $\delta^{18}O$-dilution by Mamawi Creek. This result illustrates that frequent river flooding is important to offset the effects of evaporation in this particular basin.

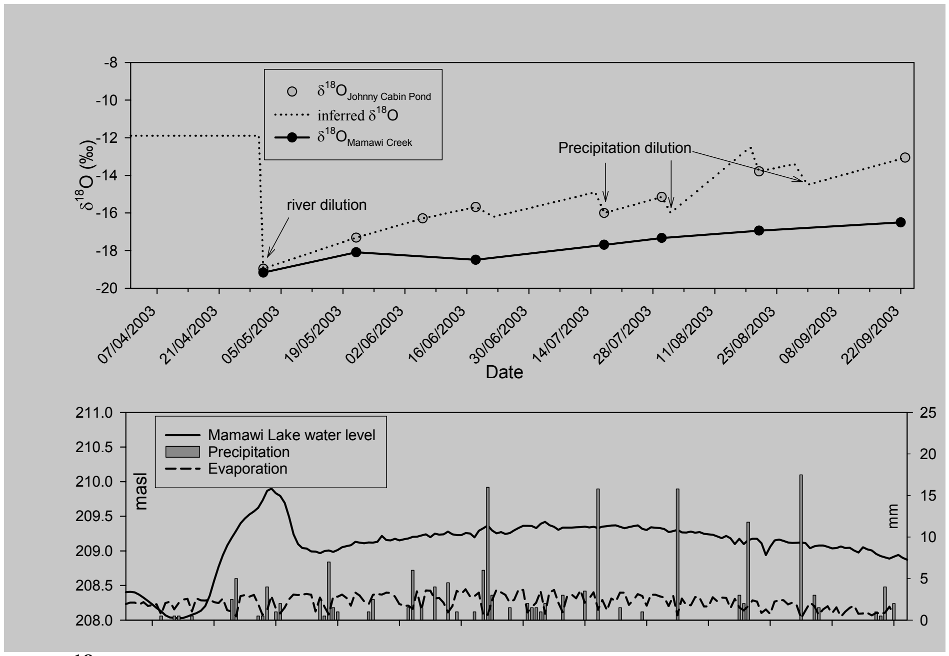


Figure 20: $\delta^{18}O$ results of JCP during 2003, compared with daily rainfall amount and water level of Mamawi Lake.

Water level data of Mamawi Lake can be found in Appendix I, and is courtesy of Jay Joyner from the British Columbia Hydro and Power Authority.

Intra-annual variability: example of a summer flood

Figure 21 shows the $\delta^{18}O$ composition for 2004 as compared to water level fluctuations of both Creed Creek (which flows directly into Mamawi Creek from the Embarras River) and JCP. It is apparent that the isotopic-enrichment trend due to mid-summer evaporation has been interrupted by an isotopically-depleted source. The magnitude of change in the isotopic composition of JCP suggests it was river water rather than rainfall that was the main input (Figure 17). Flooding likely began on July 14th, based on the simultaneous peaking of water levels in Creed Creek and JCP, and the concurrent depletion of JCP's $\delta^{18}O$ signature.

By coupling water level and isotope data, the threshold water level of Mamawi Creek required to flood JCP can be determined. Quantification of the Mamawi Creek water level required to flood JCP has implications for paleohydrological studies being conducted on JCP to determine paleo-water levels and discharge patterns of the Athabasca River. Based on the height of Creed Creek at the instant JCP's water level began to rapidly increase, Mamawi Creek's bank elevation can be equated to ~210.8 m.A.S.L.. However, it is probable that complete flushing of JCP's volume by Mamawi Creek took ~9 days to complete because the $\delta^{18}O$ value of JCP did not equal the $\delta^{18}O$ value of Mamawi Creek until July 23rd, 2004. This finding contrasts the rapid flushing of JCP during the spring floods as witnessed during fieldwork, and evidenced by isotope data. Slow, delayed flushing makes visual identification of summer flood events very difficult, a discovery that may be particularly important for assessing connections between light availability and the productivity of various biota.

There is evidence that JCP had already been partially flooded prior to the July 14th event; namely a rapid rise in both Creed Creek and JCP water levels beginning on June 17th (Appendix J and K). This summer flood event was not indicated in the isotope dataset because a sample had not been acquired from JCP until June 29th, by which time the $\delta^{18}O$ value of JCP likely would have been evaporatively-enriched back (or close) to its pre-flood composition. Nonetheless, flooding was expected during this period because the water level of Creed Creek recorded when JCP began to rise (211.0 m.A.S.L.) was almost identical to the recorded water level during the July 14th flood event (210.8 m. A.S.L.). Peters (2003) determined the threshold height of Mamawi Creek required to flood a lake neighbouring JCP (i.e., PAD 30; ~1.5 km upstream and ~50 m from Mamawi Creek) during a 1996 summer flood event was 210.5 m. A.S.L., nearly identical to results derived from analysis of the JCP flood events.

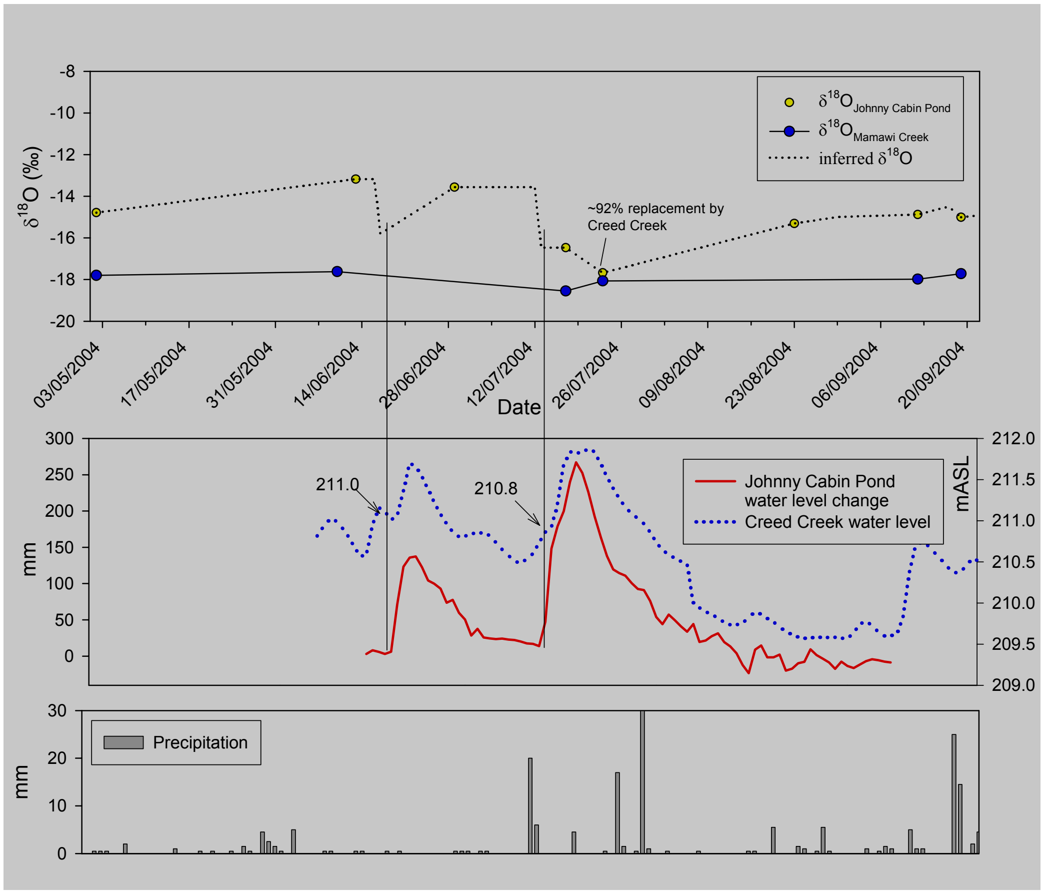


Figure 21: Stable isotope results of JCP during 2004 compared with precipitation and water level of JCP and Creed Creek.
Water level data of Creed Creek is courtesy of Jay Joyner from the British Columbia Hydro and Power Authority and can be found in Appendix J. Water level of JCP can be found in Appendix K. Note: the threshold height of Mamawi Creek required to flood JCP is estimated at 210.9 m.A.S.L.

Discussion

Results have shown that the isotopic evolution, and thus water balances of lakes in the PAD can vary significantly depending on the magnitude and isotopic composition of the primary input fluxes that the lake receives: snowmelt, rainfall, and flood water, all of which act to offset isotopic enrichment by evaporation. This chapter provides examples of the use of isotope tracer data alone, and in coupling physical water level data with isotope data, which together can be a very powerful tool. The examples used show an alternative means for assessing hydrological features of floodplain lakes with multiple water balance controls, like the effects of snowmelt, rainfall, wetlands, and flooding (including timing). These controls are often challenging to assess in such environments as the PAD using traditional methods (such as dissolved organic carbon, nutrient concentrations, and chlorophyll: Marsh and Hey, 1989; major solute concentrations: Lesack *et al.*, 1998) because such tracers are non-conservative and are highly sensitive to the timing of sampling. Water isotope data, on the other hand, preserves hydrological changes in lakes for days to weeks.

SIL, an elevated lake in the Peace sector, exhibited a relatively systematic seasonal cycle of isotopic-depletion from snowmelt and isotopic-enrichment caused by evaporation. Summer rain events also periodically interrupted the course of isotope-enrichment by evaporation. Findings based on analysis of isotopic data from SIL contrast traditional water-balance assessments from other perched basins in the Peace sector that were previously published by the Peace-Athabasca Delta Project Group (PAD-PG, 1973), Peters (2003), and Peters *et al.* (2006), who concluded that perched basins ~0.8 m deep should dry up within five to nine years without replenishment by river water inputs. Results in this chapter show that even in the absence of input from river flooding, which has likely not occurred since at least 1997, SIL's catchment captured sufficient precipitation and runoff (snow and rain) to maintain hydrologic quasi-steady-state over at least five years. It is likely that the assumptions of PAD-PG (1973), Peters (2003), and Peters *et al.*, (2006) that snowmelt and rainfall runoff are minor contributors of water to perched lakes are invalid, particularly for lakes in the northern Peace sector where bedrock relief, forests, and vast wetlands serve to accumulate snowfall and permit direct drainage to lakes. In the case of Peters (2003) and Peters *et al*. (2006), this assumption was based on the idea that isolated lakes in the PAD were similar to lakes in the Mackenzie Delta (Marsh and Lesack, 1996). However, lakes in the Mackenzie Delta are surrounded by deep soils with high infiltration capacity in small catchments, characteristics different than many perched basins found in the PAD. Most perched lakes in the PAD are generally surrounded by moderate-sized catchments with substrate that can provide, or exchange, stored water (i.e., soils of forests and shrubs and/or wetlands). In the case of SIL, the lake is bordered on three sides by bedrock and forest which provides direct and sustained runoff water, and a shrub wetland which may have provided a source of water during dry periods that can serve to buffer the magnitude of water level declines. Although more studies need to be completed on wetland hydraulics in the PAD, it is likely that the wetlands (and forests) that surround the hundreds of basins in the isolated regions of PAD play a role in maintaining water during arid times.

A much less predictable seasonal isotopic behavior occurred for JCP, a low-lying lake in the Athabasca sector. Results showed that this lake underwent considerable changes in water balance due to evaporation and frequent flooding from a bordering river over the duration of the study period. Although snowmelt and rainfall were shown to have contributed some water, qualitative and quantitative assessment illustrated that river water played the largest role in replenishing water in JCP. Summer flood events that were not known to have occurred were also captured by the isotopic data. These summer flood events contributed to JCP's ability to maintain a neutral water balance status between 2000 and 2005. Flooding of, and a net neutral water balance in JCP is consistent with expected hydrologic conditions in the active fluviodeltaic Athabasca region, where many basins receive river water input on a recurrent basis. Mamawi Creek's ability to flood JCP and other nearby lakes in the summer is a result of the diversion of considerable volumes of water into Creed and Mamawi Creeks following the Embarras Breakthrough in 1982 (PAD-TS, 1996; Wolfe *et al*., in review). These geomorphic changes have been identified as the principal driver behind why a few lakes in the southeast corridor of the Athabasca sector have become less susceptible to flooding over the past 20 to 30 years and appear to be drying up (Wolfe *et al*., in review). Findings highlight the hydroecological sensitivity of the Athabasca sector to changes in the flow regime of the Athabasca River superimposed by additional effects caused by climate warming.

Using the 'intersection-approximation' method described in Chapter 4 to estimate δ_{Source} for the isotopic mixing model has provided an opportunity to approximate lake water replenishment by various hydrological processes in the PAD. This method of estimating water replenishment of lakes is useful for other studies examining water balance changes in hydrologically-complex environments, especially ones having sparse hydroclimatic and hydrometric data. However, under the assumption that a lake receives only one input source at a time, the isotopic mixing model may slightly underestimate the amount of new water input to lakes. In addition, the model does not account for volumetric changes. Conceptually, as the water level rises, the ability of the isotopically-depleted water source to dilute the isotopic composition of the lake decreases. By not taking volumetric change into account the mixing model will underestimate the amount of dilution that the lake underwent. Lastly, the $\delta^{18}O$ of input water for a specific input source was assumed to be temporally and spatially constant, yet Chapter 4 showed that the isotopic composition of snow and rain in the PAD can vary across time and space. As such, it is likely that the calculated dilutions of flooded lakes which were not completely replenished by river water were overestimated since isotopically-depleted snow would have, in reality, mixed with river water as it flowed into a lake.

Much of the interpretation in this chapter has relied on the position of lake water isotope data as compared with reference lines (LMWL and LEL) and values of δ_I, δ_{SS}, and δ^* within a seasonal isotopic framework developed using average hydroclimatic and isotopic data from 2003. However, results from an evaporation pan experiment presented in Chapter 4 showed that the isotopic compositions of δ_{SS} and δ^* can be highly sensitive during arid intervals because these variables are primarily controlled by relative humidity and the isotopic composition of atmospheric moisture. Therefore, care must be taken when using a general isotopic framework for analyzing measured high-resolution isotope data from shallow lakes in seasonal environments. Fortunately, on average, hydroclimate and isotope parameters in the PAD during 2003 were found to be generally representative of the period between 2000 and 2005. For example, between 2000 and 2005, δ^* and δ_I were only 0.8‰ and 0.7‰ more enriched, respectively, for $\delta^{18}O$ than in 2003. Thus, the resultant slope of the LEL for 2000 to 2005 would have been 4.8 instead of 4.9 for the 2003 thaw-season (see Chapter 4 for more details). This difference in slope is not significant and would not change the interpretation of lake water isotope data.

As mentioned in the 'Results' section, isotopic-enrichment trends were often different between SIL and JCP over the same period of time. For example, between May 1st and September 23rd, 2003 (a non-flooding period), JCP underwent 7.0‰ enrichment in ^{18}O, whereas, SIL underwent only 3.5‰ enrichment. Since both lakes have similar volumes, differences in evaporative-enrichment could be explained by differences in catchment morphology that would effect snow accumulation and melt runoff, fetch, rainfall runoff potential, and wetland exchange. In the case of SIL, the bedrock and forest surrounding the three sides of the lake, as well as an extensive wetland on the other side, likely played a role in reducing net evaporative-enrichment. For example, cold bedrock and the sheltering nature of the forests surrounding SIL may have protected snowdrifts from melting as quickly, thereby supplying isotopically-depleted melt water for a sustained period after the seasonal snow cover had already melted from around the highly-exposed JCP catchment. During sampling in early May 2004, snow was more common in the Peace sector, where many lakes are bordered by bedrock and forests. The bedrock and forests surrounding much of SIL may have also protected SIL from wind and, thus, reduced moisture removal from the boundary layer above the lake; ultimately reducing diffusive transfer of the heavy isotopes. In fact, winds at SIL were often weaker than JCP during monitoring visits in thaw-season of 2003. The bedrock surrounding SIL would have also allowed for additional water from direct runoff during rainstorms and snowmelt. Water stored in the bog that is situated to the northeast of SIL may have released stored rain or melt water during times when SIL's water level was temporarily decreased during prolonged dry periods. Bogs play an important role in maintaining water in water bodies in other high-latitude environments like the PAD (Quinton *et al.*, 2003). Although JCP is also surrounded by wetlands, these wetlands likely have lower water storage capacities since they are shallower than SIL's and often inundated with silty flood waters which would clog pore spaces in the organic peat. In addition, the more concentrated, and wider (30 to 40 m wide) sedge meadow partially bounding SIL may reduce evaporation and, hence, ^{18}O-enrichment. This idea stems from a study by Rouse (2000) who attributed a 15% reduction in evaporation from a shallow wetland in James Bay to albedo effects caused by the recent growth of sedge. It is thought that emergent light-coloured sedge (and bedrock) surrounding SIL may increase the reflection of incoming solar radiation as compared to the muddy, partially rotted and often silted vegetation surrounding JCP. Thus, the darker surroundings of JCP may, in fact, induce evaporation loss. Although wetland buffering, and the type of wetland (Quinton *et al.*, 2003), is likely a control on some PAD lake water balances, at least to some extent, determining this role would have required rigorous fieldwork and expensive instrumentation that was beyond the scope of this project.

Conclusions and implications

Studies in this chapter have 1) improved the understanding of how key hydrological inputs (snow, rain, river water) in the PAD control the water balance of two basins differing in flood susceptibility, and 2) provided a means to quantify these inputs using an isotope mixing model. Combined, the hydrological behaviour of these two lakes represents the typical hydrological processes that affect lakes in northern environments. The seasonal isotopic behaviour of the basins were shown, in general, to vary as a result of three main hydrologic inputs, each of which acts to offset isotopic-enrichment due to evaporation. These include, 1) early introduction of isotopically-depleted snowmelt runoff, 2) abrupt introduction of moderately-depleted river water during spring and summer high river water events, and 3) episodic introduction of slightly-depleted precipitation during the thaw-season. The degree to which these hydrologic inputs played a role in other lakes in the delta is likely dependent mostly on catchment characteristics, and the proximity to the channel network.

Results from SIL reveal that in the absence of input from river flooding, this isolated, elevated basin in the relict Peace sector captures sufficient snowmelt and rainfall to maintain hydrologic quasi-steady-state over many years. It was suggested that snowmelt may have been slightly underestimated due to premature sampling of SIL, suggesting that in the future sampling of lakes should occur after much of the surrounding snowpack has melted. The equivalent amount of water from snowmelt versus rainfall was found to be more effective in replenishing water to SIL, a result most indicative of the catchment characteristics. Analyses of SIL over short periods of time showed that 1) the forest and bedrock making up the catchment provided additional runoff for this perched basin during the springmelt to offset evaporation during the thaw-season, 2) water stored in a wetland to the northeast of the SIL acted like a sponge to provide water to SIL during periods of low rainfall, and 3) care must be taken when using a general isotopic framework to assess shallow lakes because components of such a framework are sensitive to short-term variations in hydroclimatic conditions. These findings have clear implications for considering the effects of climate change on shallow perched basins of the Peace sector, which are driven predominantly by local and regional climatological variability at inter-annual to decadal timescales (Wolfe *et al.*, 2005; 2007b; in review). Overall, there is a tendency for the undulating landscape of the northern Peace sector, marked by bedrock outliers and wetlands, to capture snowdrifts which provide an important source of runoff to perched basins. This may be true for many lakes separated from the main drainage network in aquatic environments of the north. On one hand, reduced precipitation (especially snow) would reduce runoff from bedrock or forests surrounding many northern basins and on the other hand, wetlands may provide water for some time during periods of low precipitation.

The results from JCP have shown that flood waters play the dominant role in maintenance of the water balance in this low-lying lake in the hydrologically-active Athabasca sector. Initially, JCP was thought to have flooded only once each in the springs of 2003 and 2005, and the summer of 2001 and 2004. However, results have shown that JCP also flooded early in the summer of 2004 (i.e., June 17th), despite no local knowledge of the flood event. Analyses also showed that a minimum water level of ~210.9 m.A.S.L. of Mamawi Creek is required before JCP will flood. Knowledge of the threshold water levels needed to overtop river levees, and of the effects of spring and summer flood events, provides useful information for interpretation of paleo-discharge variations based on stratigraphic records from JCP. The strong influence of river water to JCP, and possibly other areas in the PAD, is due to a flow diversion occurring following the Embarras Breakthrough in 1982 that has led to increasing northern diversion of water into Mamawi Creek, and as such may indicate that many of the low-lying lakes in the Athabasca Sector are highly susceptible to flooding during high river stages on the Athabasca River (Falcone, 2004). Unlike the Peace sector, it is suspected that the very flat terrain of the Athabasca sector is less effective in generating snowmelt runoff to counteract evaporation. Subsequent analysis involves utilization of the mixing model to quantify the effects of river flooding and snowmelt over sub-regions of the PAD (Chapter 6).

Differences in the nature and magnitude of isotopic data between SIL and JCP are explained by variations in the isotopic compositions of input waters, and by differences in catchment morphology and characteristics. The existence of bedrock, a large elevated forest, and a bordering wetland are thought to play a role in controlling evaporative losses, and thus isotopic enrichment, of isolated lakes in the Peace sector. Bedrock and forests surrounding many of the perched basins in the Peace sector likely play key roles in providing additional and sustained water to these basins, versus the highly exposed lakes with poorly drained catchments in the Athabasca sector. Future research may include assessment of the hydrological function (i.e., exchange capability and storage capacity) of the different types of organic terrain present in the PAD. Understanding differences in local organic substrate may explain why there are differences in the responses of lakes to snowmelt, large rainfall events, and evaporation, and may also serve to facilitate a better understanding of the role water holding capacity and exchange capability of the various wetland types has to play in lake water balance. Identification of wetland types (and bedrock) could feasibly be conducted through radar analysis which can be verified through field surveys, and wetland characteristics could be determined through installation of a few instrumented monitoring wells around various lakes.

Given the likelihood of continued changes of climate and Peace and Athabasca River discharges, the latter potentially subject to accelerated change from increased industrial pressures along the river's downstream reaches, it is expected that hydroecological conditions in the PAD will continue to evolve. Because each sector has distinctive landscape characteristics, each will react uniquely to atmospheric drying and warming and river flow pattern changes (Wolfe *et al.*, in review). Hydroecological changes of basins in the PAD will be directed by local landscape features. To understand the larger landscape scale hydrological processes, including the replenishment amounts by river water and snowmelt, the following chapter utilizes a regional isotopic dataset.

6.0 DIFFERENTIATING THE IMPACT OF RIVER FLOODING AND SNOWMELT ON LAKE WATER BALANCES IN THE PEACE-ATHABASCA DELTA : SPRING 2003

Synopsis

This chapter uses water isotope tracers and inorganic suspended sediment analysis to assess the impact of river flooding and snowmelt input to 45 lakes across a 3400 km^2 region of the Peace-Athabasca Delta (PAD) in spring 2003. The spring 2003 dataset was ideal for estimating contributions by flood water because a major spring ice-jam flood occurred in the Athabasca sector and lake ice had melted by the time the flood occurred and for snowmelt because air temperatures remained above zero for much of the two weeks prior to sampling, allowing for rapid and strong snowmelt.

The differentiation of flooded lakes from snowmelt-influenced lakes was determined through the coupling of lake water $\delta^{18}O$ and suspended sediment content, and was verified through field observations. Replenishment percent of lake water from either river water or snowmelt was then estimated using an isotopic mixing model. The ability to quantify the degree to which snowmelt and river water replenish lakes is important for identifying sub-regional landscape units and how these areas may be uniquely sensitive to changes in climate and river discharge.

Lakes that flooded were identified by having suspended sediment concentrations between that of non-flooded lakes and river water, and low stable-isotope compositions close to the isotopic composition of the nearby river. Most lakes that did not flood possessed low suspended sediment contents and variable, but relatively enriched, heavy-isotope concentrations. Variability in isotope compositions from these snowmelt-dominated lakes is attributed to differences in lake volume and catchment characteristics. Most importantly, it was found that non-flooded lakes were replenished by ~30% by snowmelt. This significant contribution by snowmelt must be accounted for when conducting water balance modelling of non-flooded lakes in the PAD. The main input source and replenishment amounts compared well with previously classified lake drainage types under non-flood conditions, and spatial mapping of replenishment amount depicted regional differences in flood susceptibility. It was determined that it may be more appropriate to separate the restricted-drainage classification into 'highly-restricted' for the Peace sector and 'moderately-restricted' for the Athabasca sector.

Introduction

High-latitude, freshwater deltaic and floodplain environments are often strongly influenced by spring-thaw hydrological processes, including river flooding and snowmelt. The Peace-Athabasca Delta (PAD) is no exception. After a period of warming in the spring, much of the water stored in snowpacks in the catchments of the Peace and Athabasca Rivers melts, causing elevated streamflow pulses. These pulses of water are periodically met with jammed ice at river meanders, leading to flow reversals into distributary channels throughout the PAD. If the water levels in the channels surpass the height of their levees, then flooding will occur.

The average date of Peace River ice break-up (measured between 1973 and 1992) has been estimated as April 28th (Prowse and Conly, 1998), which is consistent with the timing of the peak Peace River discharge (Prowse and Lalonde, 1996). Although ice break-up is a highly complicated process (Beltaos, 2003), the timing of ice break-up is likely very similar for the Athabasca River. Indeed, on April 30th 2003, the Athabasca River was backed-up due to jammed ice at a sharp meander, which enhanced the flow of water into the Embarras River. A second ice-jam on the Embarras River then caused large volumes of water to flow into distributary channels which normally drain into Athabasca and Mamawi lakes. The distributary channels were unable to drain the water fast enough, thus a vast area of the Athabasca sector became inundated with river water.

Melting of the local snowpack is another main hydrological process that plays an important hydrological role during the spring in the PAD. Much of the snow that accumulates over the winter melts during a relatively short period between April and early-May when the air temperature remains above zero. At the same time, ice-cover has begun melting on most lakes, allowing for direct runoff of snowmelt into the lakes.

It was shown in the previous chapter that periodic river flooding plays an important role in maintaining the water balance of a low-lying lake in the Athabasca sector, and that snowmelt runoff plays an important role in offsetting long-term evaporation of a perched lake in the Peace sector. However, it is inappropriate to extrapolate these findings over a large spatial extent in the PAD since hydrology in the PAD is extremely complex, particularly during the spring freshet period. Thus, this chapter explores the application of isotope signatures of lake water as an indicator of water balance variability over much of the delta landscape. The ability to assess and quantify spring hydrological processes over the delta landscape is particularly useful for ongoing biological and water balance studies being conducted in the PAD. The ability to quantify the impacts of spring processes on lake water balances is also important in other northern environments because spring processes commonly play a significant role in lake and catchment water balance. Moreover, quantitative hydrological information is a prerequisite for hydrologic modelling, which has become an effective means of assessing hydrology in northern environments (e.g. Stadnyk *et al*., 2005).

The main objectives of this study are to 1) use isotope and chemical data to differentiate flooded lakes from snowmelt-dominated lakes, 2) quantify the amount of water that was replenished in each lake, and 3) develop a means to assess spatial patterns of flooding and snowmelt. A supplementary objective of this chapter is to evaluate whether the drainage classification of the lakes proposed by Wolfe *et al*. (2007b), based on data from a year without flooding, is also meaningful in the context of a year with a major flood event. This is of interest because the classifications are based on connectivity to the main river system, and thus a flood year should be ideal for testing such classifications. It is also important to identify if drainage classifications under non-flood conditions are appropriate under flood conditions because lake have been used as dependent variables in statistical assessments of the hydroecology of lakes in the PAD (e.g. Wolfe *et al*., 2007b; Wiklund, in preparation). Moreover, comparison of the drainage classifications and replenishment amounts may help ascertain if drainage classifications can be further broken down in both the relict deltaic landscape of the Peace sector, as well as the active Athabasca sector, as suggested by previous researchers (Peters *et al*, 2006).

Isotope data have been used effectively for regional lake water balance assessment (Gibson and Edwards, 2002; Leng and Anderson, 2003; Diefendorf and Patterson, 2005), particularly in assessing river water and snowmelt contributions (Wolfe *et al.*, 2007b). The methods in this chapter offer an innovative, field-efficient means for assessing the spatial extent, and quantitative impact, of flood water and snowmelt using isotope data. Notably, the application of water isotope tracers presented in this chapter is highly recommended for long-term ecosystem management for the PAD and other high-latitude environments experiencing changes in river discharge and climatic conditions. The ability to quantify the degree to which snowmelt and river water replenish lakes in these environments is important for identifying sub-regional landscape units and how these areas may be uniquely sensitive to changes in climate and river discharge.

Analyses in this chapter rely mostly on the isotopic composition of lake samples collected during a water sampling survey in the spring of 2003 (April 30th and May 1st) from a representative suite of lakes throughout the PAD. This data set was ideal for assessing spring hydrology since flooding occurred in the Athabasca sector, and air temperatures remained above freezing for much of the two weeks prior to sampling, which allowed for lake ice to thaw and most of the catchment snow to melt. Isotope data were used in the mixing model described in Chapter 5, which afforded the ability to translate the new isotopic signature of the basins into a quantitative proportion of 'new' water. Isotope mixing models have been especially useful for the hydrological community in assessing the contributions of 'new' and 'old' water in stream hydrograph separation studies (Sklash, 1990; Buttle and Sami, 1990; Kendall *at al*., 1998; St. Amour *et al*., 2005; Stadnyk *et al.*, 2005). This thesis is the first application of using water isotopes to quantify contributions of river flood water and snowmelt to floodplain lakes in northern environments.

Methods

Sample collection and analysis

Surface-water samples were collected from 45 lakes and 10 rivers in the PAD on April 30th and May 1st 2003 for analysis of oxygen and hydrogen isotope composition and total suspended solids (TSS) (Figure 22). Sampled basins span the full range of apparent river influence in the PAD, covering an area of around 3400 km^2. Detailed field notes were made during sampling, including observations of turbidity of the lake water and surrounding areas, and the height of water relative to previous visits (Appendix L).

Preparation of samples for analysis of inorganic TSS involved filtering a known volume of lake water on a 0.45μm glass membrane filter, drying overnight at 100°C, followed by heating at 500°C for 4 hours (ASTM, 1999). A measure of TSS was obtained by weight differences before and after a sample was subjected to high-temperature ignition.

Figure 22: Location of the 45 lakes sampled on April 30th and May 1st 2003.

Isotope mixing model

The replenishment of lake water by river water and snowmelt was determined using an isotopic mixing model. This model was appropriate for assessing river flooding and snowmelt because these two input sources had appreciably different isotopic compositions than the isotopic compositions of the lakes prior to freeze-up in 2002 (generally >10‰). Recall from Chapter 5 that the model is defined by

$$\% \text{ dilution} = [(\delta_{Lnew} - \delta_{Lbefore}) / (\delta_{source} - \delta_{Lbefore})]*100$$

where in this case, δ_{Lnew} and $\delta_{Lbefore}$ are the $\delta^{18}O$ values of lake water at the time of sampling and before freeze-up the previous year, respectively; and δ_{source} is the average $\delta^{18}O$ signature of input water specific to each basin. The 'trajectory' approximation in Chapters 4 and 5 could not be used because the full suite of lakes (n=63) was not sampled prior to freeze-up in the previous thaw-season. Instead, the average $\delta^{18}O$ value derived from all rivers originating from the Athabasca River (-19.2‰; n=4; range=-19.2‰ to -19.3‰) in spring 2003 was used for Athabasca lakes that flooded, and the average $\delta^{18}O$ value derived from all rivers originating from the Peace River (-21.2‰; n=6; range= -21.5‰ to -21.0‰) was used for Peace lakes that flooded. The isotopic composition of snowmelt from 2003 (-24.4‰; n=2; range=-24.4‰ to -24.5‰) was used for lakes that did not flood (i.e., dominated by snowmelt input). The isotopic composition of a lake prior to sampling in May 2003 ($\delta_{Lbefore}$) was represented by samples from late September 2002 provided it had been measured. For lakes that did not have isotopic compositions measured in September 2002, the isotopic composition of these lakes in October 2000 was used. The October 2000 dataset was chosen as a surrogate given that lakes that had $\delta^{18}O$ values measured in both October 2000 and September 2002 had similar isotopic compositions (i.e., average $\delta^{18}O$ difference of -0.2‰). The percent dilution results using the mixing model would be very similar if $\delta^{2}H$ was used because of the strong linear relationship of $\delta^{18}O$ and $\delta^{2}H$ (see Chapter 3).

Lake water replenishment mapping

Mapping the spatial distribution of water replenishment that each lake underwent was performed using Surfer 8.0® (Golden Software Inc). Radial Basis Function gridding with a Multi-quadratic basis kernel type was used to compute and assign weights to grid nodes within the PAD basin, as defined by the geographical limits of the study lakes. This technique is commonly used for small datasets that have non-uniform spacing between points since the weighting factor assigned to every point is based on the number of nearest neighbours, which allows for interpolation that avoids generating a 'bulls-eye' effect where data are sparse in a particular region of the grid (Franke, 1982; Carlson and Foley, 1991). This combination was also used by Wolfe *et al.* (2007b) for spatial assessment of evaporation-to-inflow ratios of lakes in the PAD.

Results

Separating flooded lakes from non-flooded lakes

A combination of $\delta^{18}O$ values and total suspended inorganic solids data from the spatial survey in May 2003 was compared with field notes (Appendix L) in order to distinguish between flood water-influenced basins, and those dominantly affected by snowmelt (Figure 23, Table 5). Overall, river waters exhibited the lowest isotopic compositions ($\delta^{18}O_{avg}$=-20.0‰) and very high inorganic mineral concentrations (average=11.35 mgL^{-1}). The isotopic compositions and suspended sediment concentrations varied between distributaries of the Athabasca River and rivers deriving their water from the Peace River. The rivers in the Athabasca sector ($\delta^{18}O_{avg}$= -19.2‰, $\delta^{2}H_{avg}$= -152‰; Fletcher Channel, Embarras River and Mamawi Creek) had average isotopic compositions that were slightly higher as compared to rivers in the Peace sector (-21.2‰, -165‰; i.e., Rivière des Rochers, Revillon Coupé, and Chenal des Quatre Fourches, Baril River, Claire River). This could have been due to evaporatively-enriched discharge entering the Athabasca River from Lesser Slave Lake, and/or greater proportions of isotopically-depleted glacial and/or snowmelt water flowing into the Peace River from the Rocky Mountains of British Columbia. Nonetheless, the isotopic compositions and suspended sediment concentrations of the rivers and creeks within each respective sector were found to have been conserved (i.e., had very similar isotopic compositions). Dominance of water from the main river in each sector is consistent with field observations, whereby ice-jam formation on the Athabasca and Embarras rivers caused increased northward flow of water into rivers in the Athabasca sector (Figure 24), and lower water levels of Lake Athabasca and Mamawi Lake caused some of the Peace River to flow south into rivers in the Peace sector. Overall, most flooded basins (n=18) showed $\delta^{18}O$ values (average=-19.2‰) that were similar to the $\delta^{18}O$ values of adjacent rivers (~-20.0‰; Figure 23). The average percent TSS of flooded lakes (2.76 mgL^{-1}) was between the average inorganic concentrations of snow-melt influenced lakes (0.38 mgL^{-1}) and river water (11.35 mgL^{-1}). The lower inorganic content in flooded lakes versus rivers could be due to various factors including incomplete flushing of a lake by river water,

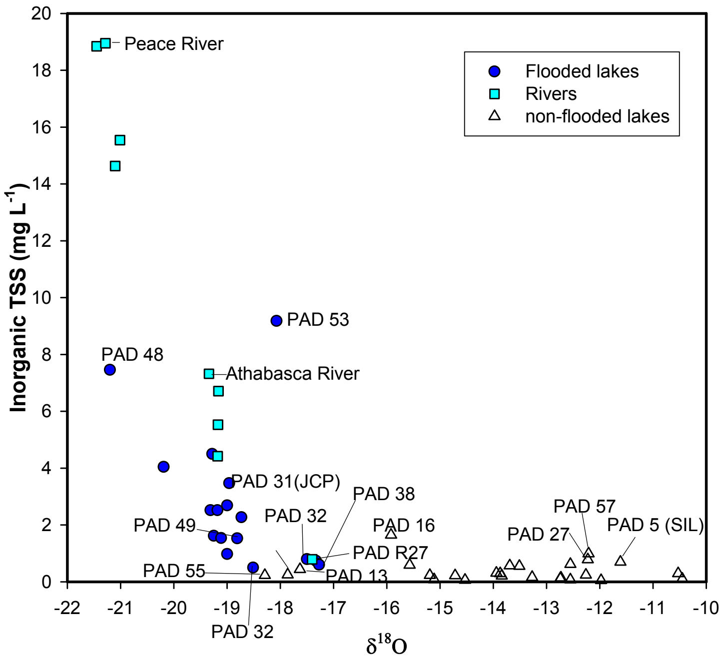


Figure 23. Total inorganic suspended sediment data versus $\delta^{18}O$ data based on sampling on April 30th and May 1st 2003.
These data, in conjunction with field observations at the time of the sampling, provide information to help differentiate river-flooded lakes from snowmelt-influenced lakes. Labelled sites are discussed in the text.

snowmelt influence, or settling of coarser inorganic grains as flood water lost energy upon travelling through interim storages such as wetlands or dense vegetation (brush). Examples of lakes that had a low percent inorganic content but had been flooded include PAD 25, PAD 32, and PAD 38 (Figure 23). These lakes had evidently flooded because the adjacent areas around them were inundated with turbid river water and the $\delta^{18}O$ values were within 0.1‰ of nearby rivers. Wetlands and willow thickets surrounding these basins may have acted as settling zones, where energy is lost and much of the suspended particles in river water would accumulate prior to reaching the lake.

Basins that did not flood (n=27) exhibited more-enriched and variable $\delta^{18}O$ compositions ($\delta^{18}O_{avg}$=-13.8‰), and diluted suspended sediment concentrations (average=0.38 mgL^{-1}) compared with flooded lakes (Figure 23). The more-enriched $\delta^{18}O$ values may be attributed to the premise that snowmelt would not have the same dilution capacity as river

Figure 24: Photographs of ice-jam in the Athabasca sector of the PAD. a) ice floe on Embarras River, b) water diversion at Creed Creek diversion from Embarras River.
Area in the background of photograph (b) shows flooded regions south of Mamawi Lake.

water would during flooding. Snowmelt would only dilute existing lake water, whereas flooding often entirely replaced lake water. The variable $\delta^{18}O$ values are likely due to differences in the physical characteristics of the lakes and their catchments. For example, it is expected that a large-volume lake that is surrounded by absorbent flora would become less diluted by initial inputs of isotopically-depleted snowmelt relative to a small-volume lake surrounded by bedrock. Although a snow sample was not tested for TSS content, it is expected that snow would have a low inorganic content, causing lakes that did not flood to have low suspended sediment contents. However, a few non-flooded lakes had elevated inorganic sediment contents compared with other non-flooded lakes. Higher suspended sediment content in these lakes is likely due to environmental features and sampling logistics during the sampling where this was observed. These lakes include PAD 27 (0.78 mgL^{-1}) and PAD 57 (0.99 mgL^{-1}), which are very shallow and were likely stirred up by the helicopter during sampling; and PAD 16 (1.6 mgL^{-1}), which was sampled through an opening of surface ice, below which there was a cluster of sediment-bearing vegetation.

δ^2H-$\delta^{18}O$-space

As was discussed in Chapter 4, superimposing lake water isotope data in δ^2H-δ^{18}O-space provides a semi-quantitative means for assessing lake water balance. The most salient information can be obtained by considering the location of lake water data in relation to δ_{SS} and δ^* along the reference Local Evaporation Line (LEL), and the location of lake water relative to the LEL. Figure 25 shows the isotopic composition of lakes sampled on April 30th and May 1st 2003 in δ^2H-δ^{18}O-space. It is apparent that all lakes were influenced to some degree by isotopically-depleted water since they plot down the LEL, below $\delta_{SS.}$ More importantly though, is that many of the flooded lakes plot separate from the lakes that did not flood. Basins that flooded plot in a tight cluster down and around the LEL, close to the isotopic composition of rivers in the PAD.

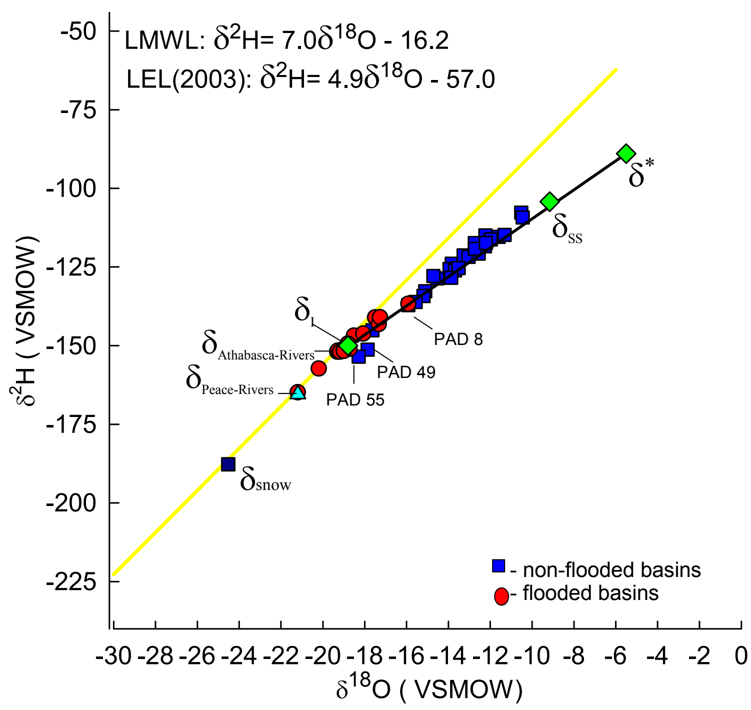


Figure 25. Isotopic composition of water samples from 45 water bodies in the PAD taken in May 2003 in relation to the isotopic framework.
The average isotopic composition of snowmelt and rivers in the Athabasca sector and Peace sector are included. The isotopic framework is based on precipitation and hydroclimatic conditions in the thaw-season of 2003.

On the other hand, most basins that did not receive flood water (snowmelt-dominated) plotted in a linear spread along the LEL between δ_{SS} and the most enriched flooded lake (i.e., PAD 8, $\delta^{18}O$=-15.9‰, $\delta^{2}H$=-137‰). Some of the visible spread may be due to spatial variability in the isotopic composition of snowmelt, which would not have been captured in the two samples taken in 2003 (average 2003 δ_{Snow}=-24.5‰, -189‰). Indeed, it was captured during the May 2004 sampling campaign where snow ranged from -23.6‰ to -28.7‰ in $\delta^{18}O$ and from -184‰ to -225‰ in $\delta^{2}H$, depending on the location and time of winter the sample was taken. However, variability in the isotopic composition of snow cannot entirely explain the broad extent of isotope compositions of the lakes in spring 2003 along the LEL. The spread is presumably also reflecting the degree to which lakes were diluted by snowmelt, which will depend on various factors including lake volume and catchment characteristics as described in the subsequent section. For example, the least snowmelt diluted lake was PAD 18, which has the largest volume of all perched lakes in the dataset.

There are a few lakes that did not flood but yet plotted down the LEL close to lakes that were known to flood. These lakes can be separated from flooded lakes by their clear water during sampling supported by low concentration of suspended solids (~0.25 mgL^{-1} compared to flooded lakes at 2.76 mgL^{-1}), but also based on their location around the LEL. As was shown in Chapter 4, lakes that receive snowmelt water influence should plot below the LEL. Indeed, a few highly-depleted, non-flooded lakes plotted below the LEL (PAD 49 and 55 on Figure 25). However, on Figure 25, most data representing lakes that had not flooded plotted on, or slightly above, the LEL, which required some further investigation. Recall that displacement of data above the LEL suggests recent input by thaw-season rain water ($\delta_{P\text{-summer}}$=-15.7±4.1‰, -129±28‰), which is usually more isotopically-enriched than the average input water to the PAD ($\delta_{I\text{-PAD}}$= -19.0‰, -152‰). Two weeks prior to the sampling campaign in September 2002 (prior to freeze-up) the PAD received 90 mm of rainfall, which is approximately double the average rainfall for September (Environment Canada, 2006). Figure 26 illustrates the effect that late, thaw-season rainstorms likely have on non-flooded lakes by comparing isotopic compositions from September 2002 and May 2003 for non-flooded water bodies in δ^2H–$\delta^{18}O$ space where data existed for both sampling campaigns. Note that most of these lakes were drawn above the LEL in September 2002. Overall, lakes that had been influenced by snowmelt were pulled along a trajectory towards the isotopic composition of snowmelt, away from their initial isotopic composition above the LEL. For illustrative purposes two examples of snowmelt-influenced lakes have been shown on Figure 26: one from the Peace sector (PAD 9), and the other from the Athabasca sector (PAD 39). Results in this section serve to illustrate the importance of knowing the isotopic composition of a lake prior to a hydrological event like spring melt.

Although a great deal of information can be gained from the analysis of lake water isotope data in δ^2H–$\delta^{18}O$ space, more information can be gleaned from quantifying the amount of replenishment that each lake underwent following both river water inundation and spring snowmelt. The subsequent section discusses the results of lake water replenishment estimates during the spring of 2003 using the isotope mixing model (equation 9 in Chapter 5).

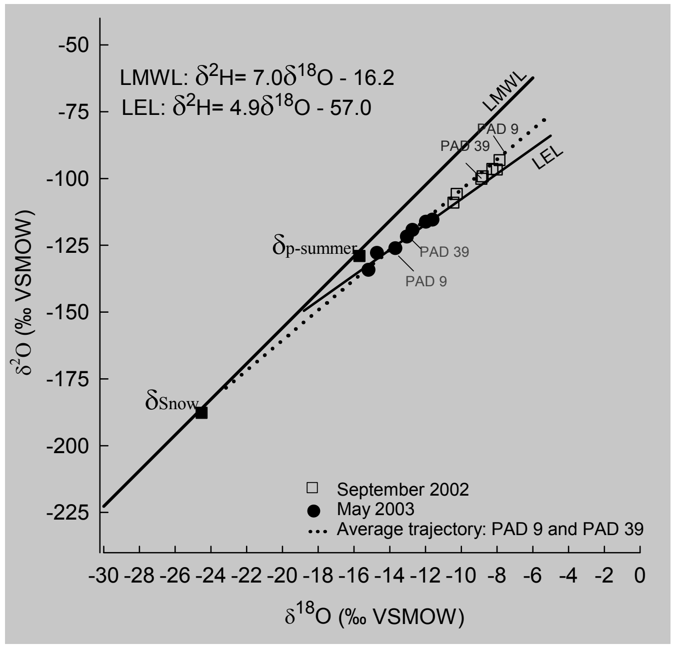


Figure 26: Isotopic composition of lakes that did not flood in May 2003 and that were also sampled in September 2002.
A basin in the Peace (PAD 9) and Athabasca sector (PAD 39) are shown. Notice that the isotopic compositions of lakes in May 2003 are on a trajectory across the LEL towards the composition of snow.

Quantifying replenishment of lake water

$\delta^{18}O$ results from the water sampling campaign in spring 2003 were used in the isotope-mixing model to quantify the amount of dilution that each lake underwent (Table 5; modelling using δ^2H produced similar results). The primary hydrological input to each lake was based on classifications made in the previous section. On average, flooded basins were strongly replenished by ~88%, which seems reasonable given many of the basins in the Athabasca sector were observed to be highly inundated with river water (e.g., PAD 31, Johnny Cabin Pond). The two least diluted lakes which flooded were PAD 38 (70%) and PAD 53 (68%). PAD 38 (Richardson Lake), a relatively large lake, only had river water flowing into it from a distributary channel from the opposite side of the lake from where the sample was taken and so some of its isotopically-enriched signature from the previous thaw-season was likely preserved. PAD 53 (Baril Lake), a relatively shallow basin, was definitely flooded to some extent as it had a high TSS concentration (9.18mgL^{-1}) indicative of river water. Indeed, Baril River, which generally drains Baril Lake was ~1 m higher than normal and was witnessed draining into Baril Lake. However, the sample was taken on the other side of the lake near the shore where some of the isotopically-enriched water from the previous thaw-season was protected in clumps of emergent plants.

In contrast, basins thought to be only snowmelt-influenced were found to be variably diluted, ranging from 9% to 57% dilution, and averaging about 31% dilution. This corresponds well with results reported by Hayashi *et al.* (2004), who estimated snowmelt runoff contributions to be 25% to 44% of the total runoff from a wetland-dominated catchment in the Mackenzie River basin. It is thought that snowmelt runoff estimates may be slightly underestimated for some lakes since the snow is known to linger in protected areas and on top of ice cover. Although it is very difficult to determine the size of every lake sampled, there seems to be a moderate correlation between the volume of the lake and how much it was affected by snowmelt. For example, the two lowest dilution amounts were from PAD 18 (9%) (a deep lake) and PAD 6 (16%) (a broad lake); and the highest dilutions were from PAD 49 (57%) and PAD 55 (55%) which were both very shallow ponds. Of course, catchment area and type would also play a role in how much snowmelt would runoff into a lake, but catchment area is very difficult to estimate in a flat and hydraulically-complex environment like the PAD. Based on results from Chapter 5, lakes with catchments formed in bedrock and/or forests are likely to have supplemental and sustained contributions from snowmelt.

Table 5. Hydrological variables and replenishment amounts of 45 PAD study basins sampled April 30th and May 1st, 2003.

Lake	Sector P=Peace A=Athabasca	Drainage type*	Suspended inorganic concentration (mg/L)	Dilution process F=Flooded SM=Snowmelt	$\delta_{L\text{-new}}$ $\delta^{18}O$	$\delta_{L\text{-before}}$ $\delta^{18}O$	$\delta_{I\text{-lake-specific}}$ $\delta^{18}O$	Replenishment (%)
1	P	C	0.21	SM	-13.8	-7.9	-24.4	36
2	P	C	0.16	SM	-13.3	-7.9	-24.4	32
3	P	C	0.08	SM	-15.1	-8.0	-24.4	43
4	P	C	0.07	SM	-12.6	-7.2	-24.4	31
5	P	C	0.7	SM	-11.6	-8.0**	-24.4	22
6	P	C	0.29	SM	-10.5	-7.8	-24.4	16
7	P	C	0.31	SM	-13.9	-8.2	-24.4	35
8	P	R	na	F	-15.9	-11.5**	-21.2	45
9	P	C	0.23	SM	-15.2	-8.8**	-24.4	40
11	P	R	0.59	SM	-15.6	-8.7	-24.4	43
12	P	C	0.57	SM	-13.7	-10.4**	-24.4	23
13	P	R	0.44	SM	-17.6	-8.8	-24.4	56
14	P	R	0.06	SM	-14.6	-9.1	-24.4	35
15	P	C	2.27	F	-18.7	-10.4**	-21.2	77
16	P	R	1.65	SM	-15.9	-8.8	-24.4	48
17	P	R	0.24	SM	-12.3	-9.4	-24.4	22
18	P	na	0.10	SM	-10.5	-8.9**	-24.4	9
20	A	R	2.52	F	-19.3	-11.5	-19.2	101***
22	A	R	1.62	F	-19.3	-9.8	-19.2	100
23	A	R	na	SM	-13.0	-10.1	-24.4	21
24	A	R	0.22	SM	-14.7	-10.0	-24.4	33
25	A	O	0.50	F	-18.5	-13.6	-19.2	88
26	A	O	1.53	F	-18.8	-12.3	-19.2	94
27	A	R	0.78	SM	-12.2	-9.4	-24.4	18
28	A	R	0.11	SM	-12.2	-9.4	-24.4	22
29	A	na	0.80	F	-17.5	-12.1	-19.2	76
30	A	R	1.54	F	-19.1	-11.2	-19.2	98
31	A	R	3.47	F	-18.9	-10.9**	-19.2	97
32	A	R	0.74	F	-17.3	-9.0	-19.2	82
33	A	R	4.5	F	-19.3	-12.1	-19.2	101***
34	A	R	2.52	F	-19.2	-11.8	-19.2	100
35	A	R	0.05	SM	-12.0	-8.3**	-24.4	23
36	A	R	0.62	SM	-12.6	-9.5	-24.4	21
38	A	O	0.6	F	-17.3	-12.7	-19.2	70
39	A	R	0.14	SM	-12.7	-7.8**	-24.4	29
40	A	R	2.69	F	-19.0	-10.9	-19.2	98
42	A	Pr	0.98	F	-19.0	-11.8	-19.2	97
48	P	R	7.46	F	-21.2	-11.0	-21.2	100
49	P	C	0.25	SM	-17.9	-8.9	-24.4	57
50	P	C	0.55	SM	--13.5	-7.5	-24.4	35
52	P	R	0.29	SM	-13.9	-9.3	-24.4	30
53	P	Pr	9.18	F	-18.1	-11.4	-21.2	68
54	P	R	4.04	F	-20.2	-11.7**	-21.2	89
55	P	R	0.23	SM	-18.3	-10.6**	-24.4	55
56	P	C	na	SM	-11.3	-8.8	-24.4	16
57	P	C	0.99	SM	-12.2	-7.5	-24.4	28

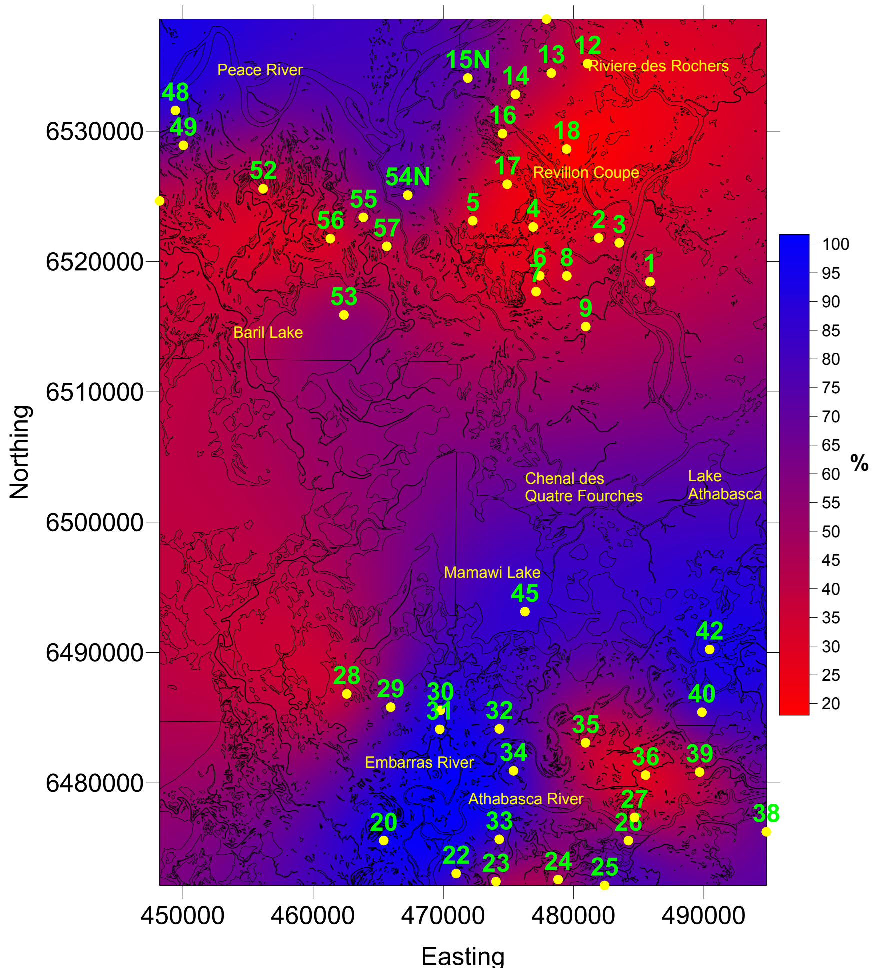
* Hydrological drainage categories based on isotopic and limnological data from samples collected near the end of the ice-free season of 2000 (C=closed; R=restricted; O=open; Pr=precipitation-sensitive; see Wolfe *et al.*, 2007b). PAD 18 and PAD 29 were excluded from the classification designation because PAD 18 is not a deltaic lake and PAD 29 had incomplete limnological information.

** $\delta_{L\text{-before}}$ is the isotopic composition of the basin in late September of 2002, the rest are from October 2000.

*** >100% dilution is likely due to the combined influence of snowmelt and river flooding (see text).

Mapping replenishment of lake water

The blue regions on Figure 27a contain lakes that received substantial input of isotopically-depleted water in the spring of 2003. These areas include a few flooded lakes in the Peace sector, numerous flooded lakes in the Athabasca sector, and a few snowmelt-sensitive lakes in the Peace sector. The less-replenished (red) regions in the Peace sector represent the snowmelt-influenced lakes that are generally surrounded by wetlands and perched in bedrock-lined basins. The less-replenished areas in the Athabasca sector are likely the result of the catchments having less snow-accumulating terrain (i.e., forests and shrubs), and a higher degree of isolation of the lakes following engineered and natural changes in river morphology. The latter is supported by paleolimnological analyses that indicate a gradual drying of PAD 23 and PAD 39 (basins within one of these isolated areas in red) since the early 1980s (Hall *et al.*, 2004; Wolfe *et al.*, in review). Figure 27b provides a schematic which outlines the approximated areas covered by flood waters in 2003. The approximated coverage of flooding is based on all available data, including suspended sediment concentrations, field observations, isotope data, and professional judgement based on the location and elevation of water bodies.



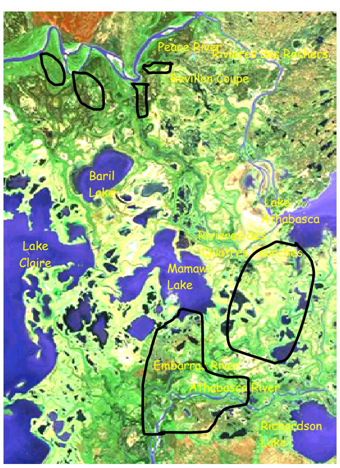


Figure 27: a) Spatial representation of dilution percents of 45 basins in the PAD during spring of 2003. b) A rough outline of areas that were witnessed with turbid river flood water on April 30th and May 1st 2003.

Dilution percents are based on results of the isotopic mixing model that uses lake-specific input sources. Flooding around the Peace and Athabasca Rivers and their distributaries is denoted by shades of blue, while the weaker effects of snowmelt are depicted by shades of red. PAD 18 is included as a reference point for comparison because it is large enough to buffer changes by input waters during spring.

Comparing lake water replenishment to lake drainage classification

It is expected that there will be relationships between the input source, and degree of replenishment, of lakes in spring 2003 with the drainage classification of the lakes based on data at the end-of-thaw-season 2000 (Wolfe *et al.,* 2007b; Table 5). For example, a closed-drainage basin as described by Wolfe *et al* (2007b) is a basin only influenced by river waters at times of extreme overland flooding and, therefore, would be fed dominantly by snowmelt during the spring. On the other hand, open-drainage basins are classified as being perennially connected to rivers under all flow conditions (e.g., Mamawi Lake) and, therefore, are expected to be flooded during ice-jam flood conditions. In between closed- and open-drainage lakes are restricted-drainage basins, which are connected to rivers by ephemeral channels or over-bank flooding during high-water conditions. Depending on the type and height of sill protecting these lakes and the distance of the lake from a flooded river, these lakes would be variably susceptible to flooding. Comparison of the October 2000 data set and the May 2003 dataset are ideal for assessing if drainage types are appropriate under any hydrological conditions because hydroclimate (i.e., temperature, relative humidity and rainfall) of the thaw-season of 2000 was similar to the thaw-season of 2002 (Table 2 from Chapter 5) and $\delta^{18}O$ compositions of lakes measured in both October 2000 and September 2002, where data existed, had very similar isotopic compositions (average difference of -0.2‰). This means that at the end of the thaw-season of 2002 the water balances of lakes would have been similar to that of October 2000. Thus, in spring 2003 lakes should have reacted to spring processes as defined by their drainage category.

As expected, most closed-drainage basins (13 of 14 basins) were not flooded in 2003, and were found to be diluted by about 35%. At around one-third of the annual input to lakes snowmelt is indeed an important water balance contributor. Many of these lakes occupied perched depressions with catchments made-up of bedrock and forest in the northeast Peace sector of the PAD (e.g., PAD 4, PAD 5 - Spruce Island Lake) and hence, would have been replenished mostly by snowmelt runoff. Lakes with the highest snowmelt dilution amounts (i.e., PAD 49 = 57%, PAD 55 = 55%) were very shallow ponds surrounded by terrain that was subject to snow accumulation (i.e., willow thickets), and were therefore capable of producing large volumes of direct snowmelt input as compared with the initial lake volumes. As is expected (based on Chapter 4), these lakes plotted well below the LEL on Figure 25. Unlike the high snowmelt dilution of PAD 49 and PAD 55, two other shallow, closed-drainage basins were observed to have undergone less dilution than expected (PAD 6 = 16% and PAD 56 = 16%). These lakes were noted to have lower than normal water levels in September 2002, which was probably attributable to evaporation. This would make the isotopic signatures of these lakes more enriched in late 2002, as compared with late 2000 (the surrogate value) and, thus, snowmelt replenishment is expected to be higher. In cases where spring hydrology is being assessed using water isotopes, it is recommended to have lake water isotope data prior to freeze-up in the previous year. The only closed-drainage lake that flooded was PAD 15 (77% replenished), a closed-drainage oxbow lake with a low sill elevation in the Peace sector that is normally separated from the main river network during the thaw-season. During sampling, this lake was witnessed to have received turbid water from the Revillon Coupé.

Also as expected, large, open-drainage basins (n=3, e.g., Mamawi Lake and Richardson Lake) found in the central part of the delta were flooded and were diluted by ~84%, on-average. However, dilution values lower than 100% indicate that a proportion of the enriched-isotope signature established prior to freeze-up in 2002 was preserved. This finding affirms that even though a lake may be directly connected to a river, it can not be assumed to be fully flushed. Incomplete flushing of lakes certainly has implications for assessing the water chemistry and ecology of such lakes. For example, certain protected areas of open-drainage lakes (oftentimes in shallower shoreline areas where aquatic plants thrive) may have a similar chemical and biological make-up as restricted-drainage lakes, introducing errors during statistical analysis.

Restricted-drainage basins (n=24) had an average dilution value intermediate to the open- and closed-drainage basins (58%). This is because many of the restricted basins in the Athabasca sector flooded (n=9), and many of the restricted basins in the Peace sector were snowmelt-influenced (n=9). This clearly illustrates that the reasons for, and degree of restriction from a river are highly dependent on the location of a lake within the delta. It is suggested that it may be more appropriate to separate the restricted-drainage classification into 'highly-restricted' for the elevated, relict Peace sector; and 'moderately-restricted' for the low-lying, active Athabasca sector.

Interestingly, two of the lakes that were found to be precipitation-sensitive because they were influenced by late thaw-season precipitation in October 2000 (i.e., PAD 42 and PAD 53, Wolfe *et al*., 2007b) were flooded in the spring of 2003. This illustrates that the modelled drainage regime of some lakes in the PAD can change depending on the time of year and hydrological conditions at the time of sampling. As was shown in Chapter 5, the isotopic composition and water balance of two shallow lakes in the PAD underwent both sudden and seasonal changes. Therefore care should be taken when utilizing water balance classifications based on only one sampling dataset. Overall, drainage classifications at the end of a thaw-season correspond well with active hydrological conditions in the spring. It is suggested that restricted-drainage lakes in the Peace sector should be separated from the Athabasca sector. To gauge the appropriateness of further classification, multi-season and multi-year datasets are required and would greatly improve the ability to understand the controls on water balances in the PAD. These datasets would also allow for a more accurate characterization of how lakes will respond to long-term changes in hydrology and climate.

Discussion

The primary water sources affecting a set of shallow lakes in the PAD in spring 2003 were identified using their suspended sediment content and stable-isotope concentrations, as well as field observations. Determining flooded lakes from lakes that did not flood is an important initial step for isotope-based water balance studies in the PAD and other northern fluviodeltaic environments. However, caution should be exercised if the turbidity of lakes is the only way to differentiate flooded from non-flooded environments. Suspended material can settle quickly, generally a few days after a flood event, and therefore may not be an appropriate proxy. In addition, predicting the timing of flooding in order to get an accurate turbidity sample is nearly impossible. Using such non-conservative tracers is particularly problematic working in an environment like the PAD because logistical challenges do not always allow for optimal timing of fieldwork. The method worked in this particular investigation because field personnel were able to sample for suspended material concentration during the time, or very soon after, the lakes became flooded. Alternatively, assessing both water isotope tracers measured from lakes against the backdrop of an isotopic framework provides a supplementary, less time-sensitive means to identify flooded lakes from non-flooded lakes.

Following identification of the primary water sources, the isotopic signature of lake water was used in an isotopic mixing model to provide first-order estimates of the contributions of river water and snowmelt. Lake water replenishment by river water was generally close to 100%, which was not surprising considering many of the lakes that flooded were highly inundated with river water at the time of sampling. However, a few lakes that flooded were found to have been replenished by greater than 100% (PAD 20, PAD 22, and PAD 33). This finding reveals the uncertainties associated with the isotope-mixing model: it only accounts for one source of input water and neglects changes in lake volume following an event. For example, replenishment values greater than 100% were likely caused by moderate inputs of melting snow (snowmelt is ~5‰ more depleted in $\delta^{18}O$ versus rivers) in addition to river floodwaters, which may originate from melting of snow by the warmer river water or by drainage from the surrounding terrain.

Lakes that did not flood were ~30% replenished by snowmelt. Although it is not expected that snowmelt would have had the same capacity to fully and initially restock the lake water as floods did, it was expected that snowmelt from the catchment would play a key role in offsetting subsequent thaw-season evaporation of lakes in 2003. This was expected because many of the snowmelt-influenced lakes were observed to be surrounded by 1) terrain of high relief (i.e., bedrock in, for example PAD 18 and PAD 5), and 2) moderate-size catchments with high water holding capacities (i.e., wetlands; e.g., PAD 1, PAD 35), or forests that delay snowmelt (e.g., PAD 15, PAD 54). Such catchment characteristics were used in Chapter 5 to explain why Spruce Island Lake underwent only half of the net isotopic enrichment of Johnny Cabin Pond, a relatively similar size lake. To confirm that snowmelt was playing a key role in the water balance of these same lakes in the spring, the spring 2005 regional dataset that was taken two weeks after the spring 2003 dataset was assessed. In spring 2005 the same set of snowmelt-influenced lakes generally underwent greater snowmelt dilution (average=~39% versus 30%) (Appendix M). This also verifies the previous suggestion that the spring 2003 dataset may have underestimated the amount of snowmelt that replenishes isolated PAD lakes.

Sustained contribution of snowmelt to basins in high-latitude, wetland-rich environments is not a newly identified phenomenon. Previous studies on the runoff response of northern peatland-dominated basins have shown that the areal extent of organic terrain greatly influences the basin runoff response (Roulet, 1986; Chapman, 1987). Nevertheless, results in this chapter do provide new insights into the hydrological processes important for lake water balance in the PAD, which has a more complex hydrological regime than many watersheds. Overall, snowmelt runoff must be accounted for as an important input to water balance studies conducted in the PAD, especially for basins which do not receive regular contributions of flood water. It is likely because of snowmelt that many such lakes in the PAD have persisted for many more years than expected in a highly evaporative environment.

Like any hydrological study, assumptions must be made and uncertainties do exist. Dilution by snowmelt may actually be slightly offset as an artefact of poor representation of the $\delta^{18}O$ value of lakes prior to freeze-up in 2002 since some of the lakes were represented by the October 2000 dataset. However, because the difference in $\delta^{18}O$ value between snowmelt and $\delta_{L\text{-before}}$ is large (15.7‰, on average), a small difference in $\delta_{L\text{-before}}$ by using the October 2000 data would cause only a minimal change in the calculated amount of lake water replenishment. Changing $\delta_{L\text{-before}}$ by ±2‰ only changes the percent replenishment by ±4%. Another uncertainty associated with using the isotopic mixing model to estimate the replenishment of non-flooded lakes is that the isotopic signature of snowmelt may vary across the delta, and through the snowpack (Buttle and Sami, 1990; Chapter 4). While rigorous assessment of uncertainties of the model associated with input parameters would require detailed site-specific investigations that are beyond the scope and objectives of this study, results provide a first-order approximation of the spatial distribution of flooding that would be difficult to obtain using other techniques.

On the whole, there exists a good correspondence between the previously-identified drainage regime of the study lakes (Wolfe *et al*., 2007b), and the type and amount of water a lake received from the major spring thaw processes of 2003. Lakes normally with open connections to the main river system, and many lakes normally restricted from the main river system in the Athabasca sector, were almost entirely replaced by river flood water. On the other hand, restricted-drainage lakes within the Peace sector were mostly snowmelt-dominated. Thus, it is appropriate to distinguish highly-restricted (Peace sector) lakes from moderately-restricted (Athabasca sector) lakes. Treating the Peace sector and Athabasca sector as separate regions within the PAD has been suggested by other researchers (Peters, 2003; Peters *et al*., 2006; Wolfe *et al*., 2007a,b).

Lakes that did not flood water were variably influenced by snowmelt depending on their storage capacity and the characteristics of their catchments. It should be noted that the degree of influence by river water or snowmelt in the spring is not necessarily a reflection of how a lake will respond to subsequent evaporation. As Chapter 4 and 5 revealed, water balances of basins vary significantly along the hydrological continuum in response to other prevailing atmospheric and hydrological processes on seasonal to annual timescales. Nevertheless, the initial pulse of snowmelt input, and subsequent release of snowmelt from the catchment, most likely plays a key role in offsetting subsequent thaw-season evaporation to lakes that do not often receive floodwater (e.g., SIL from Chapter 5).

One of the most useful features of using stable water isotope data to map flood water extent is that stable isotopes conserve a signature that reveals flooding well after a lake has hydraulically separated from the drainage network. This is a distinct advantage over other techniques, like remote sensing (e.g., LIDAR), which are often confronted with problems with surface water reflectivity associated with changing turbidity concentrations. Spatial mapping of lake water replenishment clearly identified the areas that were flooded and the areas where snowmelt was the dominant source of input water. The ability to quantify the degree to which snowmelt and river water replenish lakes is in northern environments is important for identifying sub-regional landscape units and how these areas may be uniquely sensitive to changes in climate and river discharge.

Conclusions and implications

The use of water isotope data collected during the sampling campaign of spring 2003, in conjunction with suspended sediment concentrations and field observations, has provided a means to separate the two main mechanisms that lead to lake water replenishment in PAD lakes in the spring: river flooding and snowmelt. Interpretation of lake water isotope data in relation to key linear relationships in δ^2H-$\delta^{18}O$-space revealed that the coupling of both water isotopes can be used, with some certainty, to separate flooded basins from non-flooded basins. Furthermore, an isotopic mixing model allowed for first-order estimates of the quantitative influence that snowmelt or river flood waters have on the water balance of lakes in the PAD during spring of 2003. The type and amounts of input to a lake were shown to be related to catchment characteristics and formerly defined basin drainage types (Wolfe *et al*., 2007b).

Of the 46 lakes sampled, 18 lakes flooded and 27 lakes received snowmelt as their primary input during the spring freshet period of 2003 in the PAD. Most of the flooded lakes normally maintained open-connections to the main river network, or were separated from the river by low-lying ground, and hence were almost entirely replaced with river water. Of the flooded lakes, 13 were found located in the Athabasca sector, northwest of ice-jams on the Athabasca and Embarras rivers. The other five were two oxbow lakes and three low-lying basins close to the Peace River or its distributaries. The flooding of these lakes was caused by Peace River ice-jamming at major meanders and flow reversals south into Lake Athabasca and connected delta lakes (Lake Claire and Mamawi Lake), which were temporarily below the level of the Peace River. Of the 28 snowmelt-influenced lakes, 21 were found in the Peace sector and were either closed-drainage lakes or were highly-restricted from rivers under normal flow conditions. Although non-flooded lakes were variably affected by snowmelt in spring 2003, their volumes were approximately one-third replenished as compared to ~80 to 100% for flooded lakes. Many of these lakes were perched in bedrock knolls and surrounded by wetlands. The seven snowmelt-influenced basins in the Athabasca sector were mainly situated in areas isolated from the main river network. Explanation of anomalies in the data (e.g., lakes that were not behaving like other similar lakes) illustrated the importance of prudent and detailed field notes during sampling in an environment as hydraulically complex as the PAD.

Although the application of the water isotope tracers used in this chapter (isotope mixing model and spatial mapping) were targeted to a particular situation, the methods have broader applicability for other studies in northern environments. Isotope applications are especially useful in watersheds that tend to be poorly monitored with respect to conventional hydrometric data. From a management perspective, routine regional water sampling of the PAD would allow for monitoring of evaporation and the frequency and spatial impact of key hydrologic events, including large precipitation events and floods. This has particular local relevance as results could provide an effective means to examine the effects of flooding on socioeconomic activities of community members, such as patterns of trapping and hunting. In addition, isotopic monitoring of a large suite of lakes would help assessment of changing climate and the effects on spatial and temporal patterns of vegetation.

7.0 SUMMARY, IMPLICATIONS, AND RECOMMENDATIONS

Summary

Chapter 4: Development of an isotopic framework for the Peace-Athabasca Delta: thaw-season 2003

The purpose of this chapter was to assemble data to develop an appropriate isotopic framework for the Peace-Athabasca Delta (PAD) that could be used in subsequent analysis of the thesis, including qualitatively and quantitatively assessing the influence of hydrological processes (i.e., rainfall, snowmelt, flooding, and evaporation) on PAD lakes over inter- and intra-annual time scales (Chapter 5) and across space (Chapter 6). Together information gained from this thesis will facilitate an ongoing, multi-disciplinary project assessing current and past hydrological and ecological conditions in the PAD.

It was determined that an isotopic framework developed for the 2003 thaw-season was appropriate for comparing hydrological processes between 2000 and 2005 because 2003 had 1) the most comprehensive isotopic data of input sources (including snowmelt, summer precipitation, and river water), 2) radiation data to accurately calculate evaporation, which was used to evaporation-flux-weight relative humidity and temperature, 3) evaporation pan data, which was used to assess short-term variations in key isotopic datum values and the appropriateness of using a local reference basin for modelling purposes, and 4) hydroclimatic conditions that were very similar to the average between 2000 and 2005.

- The general framework was developed using a regression line through $\delta^{18}O$ and δ^2H data collected from rain and snow in 2003 (to form the Local Meteoric Water Line or LMWL), and a line from the average input composition to the delta (δ_I) and the maximum attainable isotopic enrichment of a lake under 2003 thaw-season conditions (δ*) modelled using a reference basin (to form the predicted Local Evaporation Line or LEL). The latter was calculated using equations from the well-developed, flux-based Craig and Gordon Model (1965).
    - The LMWL for the PAD compared well with a LMWL developed for Fort Smith, NorthWest Territories. The predicted LEL was almost identical to the observed LEL based on simple regression through the isotopic compositions of numerous lakes in the thaw-season of 2003. This indicates that use of a predicted LEL is suitable if collecting isotope data from a large number of water bodies over the extent of a thaw-season is logistically difficult.
    - The methods of development of the LMWL and LEL were an improvement to Wolfe *et al*. (2007b) because of use of 1) local hydrometeorological data, 2) measured isotopic composition of local precipitation, and 3) a local reference basin in hydrologic and isotopic steady-state to model δ_A and δ^*.
- Temperature and precipitation of 2003 were similar to the average between 2000 and 2005, and relative humidity was slightly higher. Modelling showed that reasonable changes in relative humidity do not significantly alter the slope of the LEL, and thus would not change the evaluation of lake water data around the LEL.

However, it was shown that using the raw relative humidity in 2003, rather than flux-weighting for monthly evaporation, would have changed modeled δ^*, which could significantly alter the assessment of lake water data at the upper end of the LEL. In northern environments relative humidity is often highest in autumn (September and October), and evaporation has often slowed since days are shorter and radiation is lower. Therefore, not flux-weighting relative humidity would result in a higher apparent relative humidity. This would cause the calculated value of δ^* to be more depleted. Lakes that have undergone isotopic-enrichment due to evaporation and that plot up the LEL may be artificially assessed as being close to desiccation.

- Results from an evaporation pan determined isotope compositions of atmospheric moisture (δ_A), the limit of evaporative enrichment (δ^*), and a terminal lake in isotopic and hydrologic steady state (δ_{SS}) can vary by a few per mil in $\delta^{18}O$ over short-time periods.
    - It was determined that since the pan-derived δ_A records are continuous and reflect interim periods when relative humidity is less than 100%, use of an evaporation pan is more accurate to model δ_A than the more common precipitation-equilibrium approximation.
- Isotope data from lakes in 2003 were locally scattered around the LEL, revealing that lake water balances were affected by different input sources. These input sources include snowmelt, which draws the isotopic composition of a lake below the LEL; river water, which shifts a lake almost directly down the LEL; and rainfall which usually offsets a lake above the LEL. The degree of isotopic change of a

lake from the various input sources is likely a strong function of catchment characteristics (e.g., bedrock versus forest runoff) and physical properties of the lake (e.g., surface area to volume ratio).

- To characterize the composition of input water to a lake, the intersection point was taken at the LMWL from the isotopic trajectory of lake water from before and after a hydrological event. The ability to determine lake-specific and event-specific input composition is important for supplemental calculations using an isotopic mixing model used to determine the degree to which a lake was affected by a hydrological event.

Although forming an isotopic framework is necessary for analyses of lake water isotope data, its precision relies on a multitude of data which are often difficult to acquire in northern environments. To achieve the most accurate analyses of high-resolution lake water data seasonal isotopic frameworks should be developed. In addition to collecting samples from lakes, new projects incorporating water isotopes should plan to acquire data for development of an isotopic framework. As was shown in this thesis, the most important information on the isotopic framework can be obtained from monitoring a local basin in hydrologic and isotopic steady-state.

Chapter 5: Characterizing the seasonal water balance of two basins differing in flood susceptibility using water isotope tracers.

The purpose of this chapter was to assess the influence of the various hydrological processes (including snowmelt, river-water, thaw-season precipitation, and evaporation) on two hydrologically-contrasting lakes in the PAD over seasonal and annual timescales.

Comparing two distinct lakes in each sector within the delta will improve the understanding of seasonal hydrology within in each sector, and will also inform the role in which local catchment characteristics play on lake water balance.

- In the absence of input from river flooding, an isolated lake in the Peace sector of the PAD (Spruce Island Lake, SIL) captured sufficient precipitation and runoff (rain and snow) to maintain a water balance close to steady-state over the five year study period.
    - Results of an isotopic mixing model showed that 70 mm of rainfall replenished SIL by 16%, and 74 mm of snow-water equivalent replenished the lake by 28% during 2003. More effective replenishment by snowmelt is indicative of the antecedent moisture conditions of the soils in the wetlands and forest from the preceding autumns. Autumn in the PAD is often wet and is often accompanied by at least one large rain storm. Strong snowmelt replenishment reveals that snowmelt can play a critical role in replenishing water to SIL, and other similar lakes, which serves to offset subsequent evaporative enrichment. Water balance modelling indicates that snowmelt does indeed account for the water needed to close SIL's water balance in a given year.
    - Between 2000 and 2005, SIL (and Greenstar Lake) responded positively to consecutively deeper snowpack accumulation in the PAD. This emphasizes that local climatology plays a key role in controlling the water balance of isolated lakes in the PAD.

- Snowmelt and rainfall must be accounted for in all water balance studies in northern environments, but is especially important for the numerous shallow lakes that are surrounded by precipitation-storing forests and wetlands.

- Short-term variations (i.e., weekly) in δ^* modelled using evaporation pan data explain why SIL periodically exceeded the 2003 seasonal δ^*:
  - Transient changes in key isotopic reference values should be accounted for when analyzing high-resolution isotope results from shallow basins in the PAD and other environments that are susceptible to hydroclimate variations.

- Periodic river water inundation provided enough water to allow a low-lying basin in the Athabasca sector (Johnny Cabin Pond, JCP) to maintain a neutral water balance status between 2000 and 2005.
  - Results of an isotopic mixing model reveal that the spring 2003 and spring 2005 floods replenished JCP by 97% and 94%, respectively, and flooding in the summers of 2001 and 2004 replenished JCP by 86% and 91%, respectively. The two summer floods, and their ability to almost entirely replace JCP lake water, are significant because summer flood events are not well acknowledged and are generally not considered as significant as spring flood events in the PAD.

- Isotope data corroborated by water level data of JCP and a local river distributary revealed that the average river water level required to flood JCP was about 210.9 m.A.S.L.. Knowledge of river water levels needed to flood low-lying lakes in the Athabasca sector is important for future management of river water usage upstream of the PAD.

    - The ability to determine when a lake was replenished by river water and the water level required to top the banks of a river has implications for water managers of wetland-dominated and deltaic environments. Engineers often require flood levels for determining when to release water from a upstream reservoir or for deciding on the strength of man-made levees.

- Differences in the nature and magnitude of isotopic enrichment between SIL and JCP can be explained by the existence of bedrock, a large elevated forest, and a bordering wetland around SIL. It is thought that bedrock and forests surrounding many of the perched basins in the Peace sector likely play a role in providing additional, and sustained water to these basins, versus the highly exposed lakes with poorly drained catchments in the Athabasca sector.

Given changes of climate and Peace and Athabasca River discharges, the latter accelerated from increased industrial pressures along the river's downstream reaches, it is expected that the temporal evolution of lakes in the PAD will change. Because each sector has distinctive landscape characteristics, lakes within each sector will react uniquely to atmospheric drying and warming and river flow pattern changes. Continued monitoring of lakes within each sector is crucial for determining the fate of lakes under human and natural changes.

Chapter 6: Differentiating the impact of river flooding and snowmelt on lake water balances in the Peace-Athabasca Delta: Spring 2003

The purpose of this chapter was to provide an effective method to quantitatively and visually assess the spatial impact of river flooding and snowmelt on lakes during spring of 2003. Spring processes in northern environments are often complex and require volumetric analysis to determine their importance in lake water balance. In fact, quantitative information is a requirement for water balance studies using hydrologic modelling, which has become an effective means of assessing hydrology in northern environments. Recent development of user-friendly spatial software permits generation of flood maps, which are highly useful for water management of the PAD and other flood susceptible areas of the world.

- The two mechanisms of replenishment of lake water were separated using water isotope data in conjunction with suspended sediment concentrations and field observations:
  - Lakes that flooded (n=18) possessed elevated suspended sediment concentrations (average=2.76 mgL^{-1}), and isotopically-depleted compositions ($\delta^{18}O_{avg}$ = -19.2‰) close to the isotopic composition of nearby rivers ($\delta^{18}O_{avg}$ = -20.0‰). During sampling these lakes were inundated with turbid water evidently derived from high loads of suspended material carried by floodwaters of rivers.
  - Non-flooded lakes (n=28) contained low suspended sediment concentrations (average = 0.38 mgL^{-}), and variable but generally more enriched isotope signatures ($\delta^{18}O_{avg}$ = -13.8‰) than flooded lakes. Most

of these lakes were clear during sampling. Low suspended sediment concentrations were expected because snowmelt was expected to have low sediment concentrations.

  - The less isotopically-depleted signatures of non-flooded lakes are a reflection of the weaker dilution capability of snowmelt in comparison with river flooding. Variable isotopic signatures were attributed to differences in lake storage capacity (i.e., volume) and catchment characteristics, which would influence the degree of snowmelt runoff, and its effect on lake water isotope composition. Lakes with catchments formed in bedrock and/or forests, which are often found in the Peace sector, are likely to have supplemental and sustained contributions from snowmelt. On the other hand, snowmelt may by-pass still frozen, low-lying lakes with catchments susceptible to early snowmelt, which are often found in the Athabasca sector.

- Superimposing lake water $\delta^{18}O$ and $\delta^{2}H$ data from spring 2003 in $\delta^{2}H$-$\delta^{18}O$-space demonstrated a key application of an isotopic framework in assessing spring hydrology in complex, low-relief, fluvio-deltaic environments: the ability to separate flooded from non-flooded lakes.
  - Flooded lakes plotted in a cluster close to the river water isotope composition and were tightly constrained on the LEL. They were constrained to the LEL because river water plotted near the intersection of the LEL and LMWL.

- Non-flooded lakes generally plotted higher up the LEL because they maintained a degree of their evaporative signature from freeze-up 2002 and due to the fact that replenishment by snowmelt is typically less than river floodwaters. A heavy rainstorm prior to thaw-season freeze-up of 2002 caused many lakes to remain above the LEL in spring 2003. However, these lakes plotted in an isotopic trajectory towards the isotopic composition of snowmelt from their value prior to freeze-up in 2002. A few shallow lakes surrounded by bedrock plotted below the LEL, which is more characteristic of a lake influenced by snowmelt.

- Varying $\delta^{18}O$ values of lake water were translated into a range of estimated percent of new water using an isotopic mixing model. Values compared well with field observations at the time of sampling, and with previously classified drainage types if the lakes based on Wolfe *et al.* (2007b). Replenishment values were also plotted in space to provide a visual aid in examining the effects of flooding and snowmelt across the delta landscape.
  - Many of the flooded lakes (average ~88% replenished) were open-drainage lakes found in the centre of the delta, or were restricted-drainage lakes that flooded in the active, fluviodeltaic Athabasca sector of the PAD.
  - Many of the snowmelt-influenced lakes (average ~31% replenished) were closed- or restricted-drainage lakes that occupied depressions behind elevated river levees in the largely relict landscape of the Peace sector.

Documentation of flooded versus non-flooded lakes and the amounts that river and snowmelt play on the various lakes sets the stage for identifying the effect of these processes on lake water balances in the PAD as they evolve over the course of the open-water season, and for distinguishing subregions of the delta that are sensitive to the effects of hydrologic versus climatic change. Mapping the distribution of floodwaters provides a useful illustrative tool for examining connections between this important process and the local community's way of life, including trapping and hunting patterns.

Implications of research and recommendations for future research

This thesis demonstrates the effectiveness of stable water isotopes for assessing and tracing temporal and spatial variations of lake water balance and hydrology in a large, remote environment with a complex drainage network. In fact, much of the work on understanding the intricacies of hydrology in the PAD over temporal and spatial scales would have been very difficult, if not impossible, using conventional methods. Other approaches would have required high-resolution topographic data because of the PAD's subtle topography, which is very difficult to acquire; multiple satellite images, which are often obscured by clouds and water reflectivity; or an abundance of chemistry data, which can be complicated by micro-scale biological and physical processes. Straightforward interpretation of $\delta^{18}O$ alone and $\delta^{18}O$ coupled with $\delta^{2}H$ permitted qualitative assessment of water balance changes over seasonal scales, use of an isotope mixing model allowed for quantitative assessment of input sources to lakes, and mapping allowed spatial assessment of key spring processes. Understanding the complex hydrological processes in the PAD over both scales has clear implications for future lake and water management practices in and around the delta, especially under expected changes in climate. As such, isotopic monitoring of basins is as an effective means to assess hydrological changes in the PAD's and other northern ecosystems. Isotope monitoring will foster more informed analysis, decision making on resource management, and environmental stewardship.

Future management of the PAD and Mackenzie Basin Drainage system

Since the PAD is a key node in the Mackenzie Basin drainage system, continued isotopic monitoring of the water bodies in the PAD and area should be conducted to further unravel some of the complex hydrological characteristics of the system, and to distinguish the long-term hydrological effects of climate change and future upstream development. One of the most prevalent issues in northern Alberta that will likely influence hydrology of the PAD and Mackenzie Basin is the future expansion of the tar sands projects in Fort McMurray. Projects plan significant water withdrawals from the Athabasca River for oil extraction purposes. The Regional Aquatics Monitoring Program (RAMP) was initiated by multiple stakeholders in 1998 to "determine, evaluate, and communicate the state of the aquatic environment and any changes that may result from cumulative resource development within the Regional Municipality of Wood Buffalo" (RAMP, 2006). Although RAMP covers a broad spectrum of research needs, including for example, water and sediment quality, fish and wildlife monitoring, and river water drawdown, there seems to be a deficiency in knowledge regarding how changes in river water usage may effect downstream ecosystems. In fact, Griffiths and Woynillowicz (2003) argue that there is a gap in the understanding of how water withdrawal will affect hydrology in, and downstream of, the oil sands development region. As such, they recommended that Alberta Environment should ensure that research needs are identified and carried out. They advocate comprehensive water management plans to be initiated in areas of concern (which includes the PAD). Based on the limited research that was conducted by RAMP with respect to climate and hydrology, they stated that "surface water hydrology in the RAMP study area [which includes the Athabasca sector of the PAD] has exhibited minor changes, some of which may be a result of oil sands development" (RAMP, 2006). To develop a water management plan and to further understand surface water hydrology in the RAMP study area, it is suggested that a comprehensive water isotope monitoring program be initiated. Although, the initial reconnaissance dataset will provide highly valuable information for characterizing modern controls on water balances of study basins (e.g. Wolfe *et al.*, 2007b), it is the temporal insight gained from collecting isotope samples over many years that is necessary to gauge the response and resiliency of the study area to various environmental stressors. Therefore, to ensure development in the area is sustainable, RAMP should request federal funding agencies and oil sands development decision makers to come forth and provide additional funding.

Management of water in the delta will also be necessary as longer thaw-seasons and increased summer precipitation develop in the North (CCCMA, 2006). Research predicts that northern regions will shift to a more pluvial (precipitation-driven) system, reducing the ability of flow systems to replenish riparian ecosystems in the spring, especially river deltas (CCCMA, 2006). For the PAD, these changes will likely result in a less intense freshet period which could reduce spring flood capabilities and snowmelt runoff in the PAD, which have been shown to be essential for replenishing water to lakes in the Athabasca sector. However, with increased summer precipitation, summer floods may become more common. While research on the effects of summer flooding in the PAD and other northern ecosystems is its early stages, a shift to more pluvial systems may mean deltas become less and less reliant on spring flooding processes, which may have important implications for water management in these areas.

To monitor the effects of upstream river water use and regulation and climate change on the PAD, it is highly recommended to continue the regional isotope sampling program in the PAD that was initiated in 2000 (see Wolfe *et al*., 2007b). Currently, the program includes sampling of lakes that span the spectrum of local water balances, as well as sampling of main rivers in the PAD. With continuation of the program, regional patterns of evaporation, flooding, and snowmelt could be assessed in different years, and long-term lake water balances could be observed. To translate isotopic compositions of lakes into a meaningful quantitative expression of water balance, it is highly recommended that evaporation-to-inflow ratios of the lakes be calculated using the Craig and Gordon model (1965) and well-known isotope-mass balance models (Wolfe *et al*., 2007). Spatial assessment of evaporation-to-inflow ratios at different times of the thaw-season over many years may reveal systematic spatial trends reflecting upstream river water use and climate change. Additional analysis should include more rigorous quantitative assessment of water isotope data to more precisely model the isotopic composition of input water to each lake. Current research involves coupling of both water isotope tracers and further considerations of isotope-mass-balance models, and the geometry of isotopic data on the isotopic framework to characterize lake-specific input compositions (Yi *et al*., in preparation). Knowledge of input sources will guide water managers of the delta with respect to how sensitive a basin may be due to changes in river water usage and climate, having potential implications when conducting a cost-benefit analysis for consumptive water use. To provide a long-term baseline of hydrology in the PAD and to monitor how lakes have responded to changes in hydroclimatic conditions the reference basin in isotopic and hydrologic steady-state that was used in this thesis should continue to be monitored.

Integrating the methods in which hydrology has been assessed in the PAD could easily be integrated into RAMP's program. Including water isotope analysis would be a great asset to the program, and it could help meet some of their objectives, including "monitoring of the aquatic environment in the oil sands area to detect and assess cumulative effects and regional trends, and collecting baseline data to characterize variability in the oil sands area" (RAMP, 2006). Importantly, isotopic composition of water can also provide a cost-effective, scientifically rigorous approach to assess water resources, which is absolutely necessary for protecting and expanding oil sands development. As such, RAMP needs to request funding from oil sands managers and stakeholders to initiate isotope sampling. Simple questions need to be answered by the stakeholders like, what hydrological effects are incurred to the water bodies within and around the catchment isolated by diversion for oil sands operating practices? Will areas of development be usable by future generations? Monitoring efforts could be used to develop a regional database to be used by oil sands operators for their environmental management programs. Isotopic analysis may provide information to developers regarding water use for oil sands expansion and how to sustain this resource. The isotopic sampling program could be integrated into their existing sampling set-up which includes the Athabasca Delta, major and minor tributaries of the Athabasca River, and a regional set of shallow lakes in the vicinity of current and planned oil sands development. In new development areas where little research has been conducted a new sampling program may be required. It should be initiated and set up in a similar manner as that of the PAD. As such, it is recommended to take water samples from a suite of lakes and ephemeral tributaries that span the hydrological spectrum in the area of interest. This should include non-influenced basins to provide an adequate hydrological baseline, and potentially-influenced basins to assess the effects of water usage from the main contributing river. Isotope data could be converted to evaporation-to-inflow ratios to translate water balances into a meaningful expression and to assess landscape patterns (Wolfe *et al*., 2007b), which over time could be used to determine which lakes deserve more detailed attention and provide information for continuation of a monitoring program. To obtain the maximum amount of hydrological information in the area, the sampling should be conducted at the minimum, after ice-off, during the summer, and at the end of the thaw-season.

Monitoring of seasonal and spatial hydrology of the PAD is also important for Parks Canada's commitment to protect the heritage of national parks to ensure that they remain healthy and undamaged under new pressures. As part of this commitment, the WBNP is in the early stages of a renewed park management plan process that is expected to be fully updated by 2008 (Parks Canada, 2006). One of the major directions is in the form of maintenance and restoration of ecological integrity through protection of its natural processes and natural resources. Since there is a strong connection between hydrology and limnology in the PAD (i.e., ecological health, Timoney, 2002; vegetation type, Töyrä and Pietroniro, 2005; water chemistry, Wolfe *et al*., 2007b), ongoing research and monitoring of hydrology is considered to be fundamental. Continued isotopic sampling of lakes would be valuable to document and understand long-term hydrological and ecological changes in the PAD. More specifically, continuation of the water isotope sampling program could be used to 1) observe and evaluate the importance of snowmelt and rain (direct and runoff), and flooding on the water balance of lake under different hydroclimatic situations, 2) further understand the relationship water balances have on ecological changes, 3) provide more data for use with spatial software to visually assess hydrological processes across space, and 4) facilitate development and validation of water resource and ecosystem models used for predicting hydrological change under various climate scenarios (e.g. Prowse *et al*., 2006).

However, the application of water isotopes has rarely been incorporated into programs from which decisions are made by water and environment policy makers. Many policy makers have treated isotope hydrology as a largely academic pursuit. As this thesis has shown, stable water isotopes are highly useful to derive time-series and spatial snap-shots of hydrological variability in the PAD, a requisite during the planning and execution stages of park management. Indeed, stable isotope tracer approaches offer a reliable method to assess complicated hydrological processes in remote environments so that ecosystem managers can generate an integrated view of the surface water system in time and space, not easily afforded by conventional methods.

Research surrounding flood importance and extent has been perhaps one of the most challenging undertakings of research in the PAD due to its flat terrain and large size, but is very important for assessing the ecological response of the various species in the PAD. Landscape-scale assessment of water balances and flooding in this thesis could contribute to further understanding of the relationships between hydrology and limnology. For example, mapping of isotope data could be used to corroborate interpretation of other approaches like remote sensing data (i.e., geomatics) recently used to assess the spatial extent and duration of flooding (Töyrä *et al.*, 2005). As a component of needed ground 'truthing' data, isotope data would provide substantial amounts of valuable information far beyond simply verifying the existence of water, including where and how water is moving, the source of water, and if the water has undergone evaporative loss. The latter is important to differentiate water loss due to evaporation from that simply caused from natural drainage or seepage, which other methods cannot easily determine. If a number of maps are generated on a regular basis, relationships between vegetation and hydrological parameters (i.e., source water versus growth patterns) could be explored and changes monitored. In addition, recent advancements in the resolution of elevation data (i.e., LIDAR) may allow for combining isotope data with elevation data to answer questions regarding the water levels required for overbank-flooding at various points in the delta and the spatial extent of flooding caused by a particular river stage level. This has direct implications for water management in the delta

Paleohydroecology

Insights gained from detailed modern hydrological analysis of lakes have been, and can continue to be, used to inform and constrain interpretations of past hydrologic and ecological changes of the PAD (Falcone *et a*l., 2004a; Hall *et al*., 2004; Wolfe *et al*., 2005; Wolfe *et al*., in review). Ongoing paleolimnological studies rely on stratigraphic analysis of multiple proxies including geochemical and biological indicators in lake sediments, which often have direct relationships with the source of input water and water balance status of a lake at the time of the proxy's' formation. To realize the relationship between lake water and the proxy (e.g., diatoms, plants, aquatic cellulose) a paleolimnologist will collect samples of the proxy in the water, surface sediment, and biota from an array of lakes that cover the range of water balances in the region of interest (Hall *et al*., 2004). Information gained from this 'calibration' data set is then transferred to assess the sediment profile data. It is very important that the paleolimnologist knows the hydrological status of the lake at the time of the proxies formation, as well as the sensitivity of the lake to changes in hydrology and climatology. This is where isotope tracer data and isotope mass balance modelling are highly useful. For example, time series isotope data were used to explain discrepancies between the oxygen isotope data from lake-bottom aquatic cellulose and lake water from various shallow lakes in the PAD (Falcone *et al.*, 2004). It was suspected that the timing of main cellulose production may be different in lakes depending on the time of ice-off, source of input water, and turbidity of the lake. Many multidisciplinary paleolimnological studies are now incorporating modern isotopic hydrology to assess water balance controls and variability (Falcone, 2002; Wolfe *et al*., 2005; Wolfe *et al*., in review). For example, isotopic analysis of lake water from SIL was used to verify the assumptions and interpretations of a sediment core representing ~300 years of paleohydrologic history (Wolfe *et al.,* 2005). Most importantly, the reconstructed isotopic compositions of lake water over the past ~300 years fit within the measured range of modern isotope values. Also, both measured and reconstructed lake water isotope data suggested that this lake was sustained at quasi-steady-state by catchment-sourced precipitation, even in the absence of river flooding.

There are current and future paleolimnological studies being conducted on lakes in the PAD that should continue to assess contemporary hydrology of the lakes to aid in interpretation and modelling of sediment records. For example, various sediment proxies (mostly diatoms) are being analyzed from a core from JCP to reconstruct flood events in the Athabasca sector (Wiklund, in preparation; Wolfe *et al*., in review). It is generally assumed that when a river-dominant species is found in the sediment, that it was migrated during spring flood events. However, results in this thesis showed that flooding in the summer is also possible (and likely common) in many parts of the Athabasca sector. Overall, it is recommended that analysis of lacustrine proxy records is paired with modern water balance information in order to determine whether variability in paleorecords are reflective of changing lake hydrological conditions or local/regional climate variations (Hall *et al*., 2004; Wolfe *et al*., 2005; Wolfe *et al*., in review).

Also, with the advent of more routine, and higher resolution analyses of lake sediment aquatic cellulose oxygen isotope composition promoted by recent advancements in sample preparation techniques and mass spectrometry (Edwards *et al*., 2004; Wolfe *et al*., 2007a), more detailed understanding of modern hydrology of a lake will be critical. Perhaps more importantly, a detailed understanding of modern hydrology will be realized with new developments of new archive materials (including lipids from organic matter) used to reconstruct δ^2H lake water isotopic history from lake sediments (Yi, Ph.D. in progress). With quantitative knowledge of both tracers the primary isotopic effects caused by shifts in the isotopic of meteoric water can be separated from those associated with secondary hydrological processes such as evaporative isotopic-enrichment. In δ^2H–$\delta^{18}O$ space this translates to MWL-parallel shifts, which are commonly temperature-dependent but can also reflect air mass circulation characteristics, from LEL-parallel shifts, which are generally associated with changes in relative humidity and water balance (Wolfe *et al*., 2007a). A more rigorous assessment of the reasons for and sensitivity of changes along and about the LEL are needed. This could be completed through more detailed isotope-mass balance modelling and sophisticated characterization of input sources (Yi, Ph.D. in progress).

Although this thesis has provided an excellent template for understanding the various hydrological processes controlling temporal and regional water balance of lakes in the PAD, there is much more to learn about the hydrology of the PAD and how it will respond to various environmental stressors. Continued hydrological research on the PAD is warranted because of its distinctive hydrology, international importance, and its recognition as a model system for assessing the various hydrological processes that occur in complex northern environments. Specifically, the two unique sectors in the delta provide an excellent means to assess the effects of natural and anthropogenic changes. The Peace sector is more driven by local and regional climatological variability, while the Athabasca sector should be far more sensitive to continued Athabasca River discharge changes, accelerated by industrial pressures. For residents, community liaisons, and researchers confronting the intricacies and uncertainties of climate change and river usage and its impacts in the North, continued isotope monitoring will foster more informed analysis and decision making on resource and environmental management. In conjunction with what was learned in this thesis, continued isotopic monitoring will provide the opportunity to understand more about the linkages between hydroclimatic drivers and hydroecological responses. Overall, improved understanding of seasonal and regional hydrology in the PAD, especially the ability to quantify lake water balance controls, improves the ability to develop effective guidelines and management practices in other northern environments as they respond to future changes in climate and river discharge.

REFERENCES

Alberta Environment, 2006. Mountain snow conditions and water supply: August 2001 and July 2004 data from http://www3.gov.ab.ca/env/water/ws/watersupply/historical/Aug01.pdf. Accessed January 2006.

Alberta Geological Survey, 2006. Geology of the Peace-Athabasca River Delta Region, Alberta. Alberta Energy and Utilities Board. Authors: Bayrock, L.A and Root, J.D.. http://www.ags.gov.ab.ca/publications/ABSTRACTS/OFR_1972_01.html. Last modified February 10, 2006. Retrieved March 10th 2007.

Allison, G.B., and Leaney, F.W. 1982. Estimation of isotopic exchange parameters, using constant-feed pans, *Journal of Hydrology* 55: 151–161.

Araguás-Araguás, L., Froehlich, K. and Rozanski, K. 2000. Deuterium and oxygen-18 isotope composition of precipitation and atmospheric moisture. *Hydrological Processes* 16: 1341-1355.

ASTM. 1999. Standard Test Method for Determining Sediment Concentration in Water Samples. Annual Book of Standards, Water and Environmental Technology, 1999, 11.02: 389-394.

Andres, D.A. 1995. A case study: freeze-up and jamming on the Peace River at Peace River. In *River Ice Jams*, Beltaos, S., editor, Water Resources Publications, 5-23.

Beltaos, S. 2003. Numerical modelling of ice-jam flooding on the Peace-Athabasca delta. *Hydrological Processes* 17: 3685-3702.

Bennett R.M., Card J.R., Hornby D.M. 1973. Hydrology of Lake Athabasca: past, present and future. *Hydrological Sciences Bulletin XVIII.* 39: 337-345.

Brock, B.E., Wolfe, B.B., and Edwards, T.W.D. 2007. Characterizing the hydrology of shallow floodplain lakes in the Slave River Delta, NWT, using water isotope tracers. *Arctic, Antarctic and Alpine Research.* (in press)

Buttle, J.M., and Sami, K. 1990. Recharge processes during snowmelt: an isotopic and hydrometric investigation. *Hydrological Processes* 4: 343–360.

CNIP (Canadian Network for Isotopes in Precipitation). 2005. Accessed October 2005 from http://www.science.uwaterloo.ca/~twdedwar/cnip/cniphome.html

Carlson, R.E., and Foley, T.A., 1991. Radial Basis Interpolation Methods on Track Data, Lawrence Livermore National Laboratory. UCRL-JC-1074238.

Carter, T. 1996. Peace Athabasca Delta Perched Basin Surveys. Appendix in report entitled "Modelling the water balance of Peace-Athabasca Delta perched basins" prepared for the Peace-Athabasca Delta Technical Studies, Fort Chipewyan, Alberta, Canada, 45 pp.

CCCMA. 2006. Canadian Centre for Climate Modelling and Analysis. Canadian General Circulation Model 2: simulations of anticipated future climate change in 2070-2099. Data downloaded January 2005 from http://www.cccma.bc.ec.gc.caldatalcgcm2.

Chapman, A.R. 1987. Some hydrological and hydrochemical characteristics of boreal forest watersheds in the sub-arctic of west-central Canada. M.Sc. thesis, Trent University, Peterborough, Canada, 139 pp.

Clark, I.D., and Fritz, P. 1997. Environmental Isotopes in Hydrogeology, CRC Press, 328 pp.

Coleman, M.L., Shepherd, T.J., Durham, J.J., Rouse, J.E., and Moore, G.R., 1982. Reduction of water with zinc for hydrogen isotope analysis. *Analytical Chemistry* 54: 993-995.

Cooper L.W. 1998. Isotopic fractionation in snow cover. In Kendall C., and McDonnell, J., editors, *Isotope Tracers in Catchment Hydrology*. Elsevier, Amsterdam, 119–136.

Coplen T.B. 1996. New guidelines for reporting stable hydrogen, carbon, and oxygen isotope-ratio data. *Geochimica et Cosmochimica Acta* 60: 3359-3360.

Craig, H. 1961. Isotopic variations in meteoric waters. *Science*, 133: 1702-1703.

Craig, H. and Gordon L.I. 1965. Deuterium and oxygen-18 variations in the ocean and marine atmosphere. In Tongiorgi, E., editor, *Stable Isotopes in Oceanographic Studies and Paleotemperatures,* Spoleto. 9-130.

Dansgaard, W. 1964. Stable isotopes in precipitation, *Tellus* 16: 436–468.

den Hartog G. and Ferguson H.L. 1978. Evaporation Plate 17 – Hydrological Atlas of Canada. Department of Fisheries and Environment, Ottawa, Ontario, Canada.

Diefendorf, A.F., and Patterson, W.P. 2005. Survey of stable isotope values in Irish surface waters. *Journal of Paleolimnology* 34: 257-269.

Dinçer, T. 1968. The use of oxygen-18 and deuterium concentrations in the water balance of lakes. *Water Resources Research* 4: 1289-1305.

Drimmie, R.J. and Heemskerk, R.A. 1993. Water ^{18}O by CO_2 Equilibration, Technical Procedure 13.0, Rev. 02. Environmental Isotope Laboratory: Department of Earth Sciences, University of Waterloo, 11 pp.

Environment Canada, 2005. Climate data in Fort Chipewyan, Alberta for 2003. Accessed October 2005 from http://www.climate.weatheroffice.ec.gc.ca/climateData.

Edwards, T.W.D., Wolfe, B.B., Gibson J.J., and Hammarlund D. 2004. Use of water isotope tracers in high-latitude hydrology and paleohydrology. In Pienitz, R., Douglas, M., and Smol, J.P., editors, *Long-term Environmental Change in Arctic and Antarctic Lakes*, Springer, 187-207.

Epstein, S., and Mayeda, T.K., 1953. Variations in the $^{18}O/^{16}O$ ratio in natural waters. *Geochimica et Cosmochimica Acta* 4: 213.

Falcone, M. 2002. *A hydrological reconstruction of a perched basin in the Peace Athabasca Delta, Alberta, Canada using stable isotopes*. B.Sc. Thesis, McMaster University, 38 pp.

Falcone, M., St. Amour, N., Edwards, T.W.D., and Wolfe, B.B. 2004a. Comparing estimates of the isotopic composition of atmospheric moisture in northern Canada. *Canadian Geophysical Union* - Student Hydrology Section, Kingston, Ontario. Abstract.

Falcone, M., Edwards, T.W.D., and Wolfe B.B. 2004b. Quantifying transient water balance in floodplain lakes of the Peace-Athabasca Delta using isotope tracers. International Association for Great Lakes Research Annual Conference, Waterloo, Ontario. Abstract.

Franke, R. 1982. Scattered Data Interpolation: Test of Some Methods. *Mathematics of Computations*, 33: 181-200.

Gat, J. R., and Levy, Y. 1978. Isotope hydrology of inland sabkhas in the Bardawil area, Sinai. *Limnology and Oceanography*. 23: 841– 850.

Gat, J.R. 1981. Lakes: Stable Isotope Hydrology- Deuterium and Oxygen-18 in the Water Cycle. Gat, J.R., and Gonfiantini, R., editors, IAEA Technical Report Series No. 210, Vienna, pp. 203-221.

Gat, J.R. 1995. Stable isotopes of fresh and saline lakes. In Lerman, A., Imboden, D., Gat, J.R., editors, *Physics and Chemistry of Lakes*, Springer-Verlag, New York, 139-166.

Gibson, J.J. 2001a. Forest-tundra water balance signals traced by isotopic enrichment in lakes. *Journal of Hydrology* 251: 1-13.

Gibson, J.J. 2002a. Short-term evaporation and water budget comparisons in shallow Arctic lakes using non-steady isotope mass balance. *Journal of Hydrology* 264: 242-261.

Gibson, J.J. 2002b. A new conceptual model for predicting isotope enrichment of lakes in seasonal climates. International Geosphere Biosphere Programme IGBP PAGES News, 10, 10–11.

Gibson, J.J., Edwards, T.W.D., Bursey, G.G., and Prowse, T.D. 1993a. Estimating evaporation using stable isotopes: quantitative results and sensitivity analysis for two catchments in northern Canada. *Nordic Hydrology* 24: 79-94.

Gibson, J.J., Edwards, T.W.D., Prowse, T.D. 1993b. Runoff generation in a high boreal wetland in northern Canada. *Nordic Hydrology* 24: 213-224.

Gibson, J.J., Edwards, T.W.D., and Prowse, T.D. 1996a. Development and validation of an isotopic method for estimating lake evaporation. *Hydrological Processes* 10: 1369-1382.

Gibson, J.J., Prowse, T.D., and Edwards, T.W.D. 1996b. Evaporation from a small lake in the continental arctic using multiple methods. *Nordic Hydrology* 27: 1-24.

Gibson, J.J., Reid, R.R.. and Spence, C. 1998. A six-year isotopic record of lake evaporation in the Canadian subarctic. *Hydrological Processes* 12, 1779 – 1992.

Gibson, J.J., Edwards, T.W.D., and Prowse T.D. 1999. Pan-derived isotopic composition of water vapour and its variability in northern Canada. *Journal of Hydrology* 217: 55-74.

Gibson, J.J., and Edwards, T.W.D. 2002. Regional surface water balance and evaporation-transpiration partitioning from a stable isotope survey of lakes in northern Canada. *Global Biogeochemical Cycles* 16: 10.1029/2001GB001839.

Gibson, J.J., Prepas, E.E., and McEachern, P. 2002a. Quantitative comparison of lake throughflow, residency, and catchment runoff using stable isotopes: modeling and results from a regional survey of boreal lakes. *Journal of Hydrology* 262: 128-144.

Gibson, J.J., Edwards, T.W.D., Birks, S.J., St. Amour, N.A., Buhay, B., McEachern, P., Wolfe, B.B., and Peters D.L. 2003. Progress in isotope tracer hydrology in Canada. *Hydrological Processes* 19: 303-327.

Gonfiantini, R. 1986. Environmental isotopes in lake studies. In Fritz, P., and Fontes, J.-Ch., editors, *Handbook of Environmental Isotope Geochemistry, Volume 3*, Elsevier, New York, 113-168.

Griffiths, M., and Woynillowicz, D. 2003. Oil and Troubled Waters: Reducing the impact of the oil and gas industry on Alberta's water resources. Pembina Institute for Appropriate Development. Accessed December, 2006 from http:www.pembina.org/pdf/publications/oilandtroubledwaters.pdf.

Hall, R.I., Wolfe, B.B., Edwards, T.W.D., Karst-Riddoch, T, Vardy S., Falcone, M., and coauthors. 2004. A Multi-Century Flood, Climatic, and Ecological History of the Peace-Athabasca Delta, Northern Alberta, Canada. Final report to BC Hydro, 163 pp.

Hayashi, M., Quinton, W.B. Pietroniro, A., and Gibson, J.J. 2004. Hydrologic functions of wetlands in a discontinuous permafrost basin indicated by isotopic and chemical signatures. *Journal of Hydrology* 296: 81-97.

Jaques, D.R. 1989. Topographic mapping and drying trends in the Peace-Athabasca Delta, Alberta using LANDSAT MSS imagery. Report prepared Ecostat Geobotanical Surveys Inc. for Wood Buffalo National Park, Fort Smith, Northwest Territories, Canada, 36 pp.

Kendall, C. and McDonnell, J.J. (Editors). 1998. Isotope Tracers in Catchment Hydrology.
Elsevier Science B.V., Amsterdam, 839 pp.

Leconte, R., Pietroniro, A., Peters, D.L., and Prowse, T.D. 2001. Effect of flow regulation and hydroclimatic conditions on the Peace-Athabasca Delta 1996 summer flood. *Regulated Rivers: Research and Management* 1: 1751-65.

Leng, M.J., and Anderson, N.J. 2003. Isotopic variation in modern lake waters from western Greenland. *The Holocene* 13: 605-611.

Lesack, L.F.W., Marsh, P., and Hecky, R.E. 1998. Spatial and temporal dynamics of major solute chemistry among Mackenzie Delta lakes. *Limnolology and Oceanography* 43: 1530-1543.

Maric, R. 2003. *Coupled isotope-mass balance studies in tundra lakes, Lac de Gras area, NWT, Canada.* M.Sc. Thesis, University of Waterloo, 145 pp.

Marsh, P, and Hey, M. 1989. The flooding hydrology of Mackenzie Delta lakes near Inuvik, N.W.T., Canada. *Arctic.* 42: 41-49.

Marsh, P., and Lesack, L.F.W. 1996. The hydrologic regime of perched lakes in the Mackenzie Delta: potential responses to climate change. *Limnology and Oceanography* 41: 849–856.

Natural Regions Committee, 2006. Natural Regions and Subregions of Alberta. Compiled by D.J. Downing and W.W. Pettapiece. Government of Alberta. Pub. No. T/852. Retrieved March 2007. http://www.cd.gov.ab.ca/preserving/parks/anhic/docs/NRSRcomplete%20May_06.pdf

Nielson G., 1972. Groundwater investigation, Peace-Athabasca Delta. In: Peace-Athabasca Delta Project Group, Technical Report and Appendices: Volume 1, Hydrological investigations, 1973, Governments of Canada, Alberta and Saskatchewan, Section J.

PAD-PG (Peace-Athabasca Delta Project Group), 1973. Peace-Athabasca Delta Project, Technical Report and Appendices. Volume 1: *Hydrological Investigations*, Volume 2: Ecological Investigations.

PAD-IC. 1987. Peace-Athabasca Delta Water Management Works Evaluation. Final Report, Appendix A: Hydrological Assessment, Appendix B: Biological Assessment, Appendix C: Ancillary Studies. Peace-Athabasca Delta Implementation Committee, Governments of Alberta, Saskatchewan and Canada. 63 pp.

PAD-TS (Peace-Athabasca Delta Technical Studies). 1996a. MacMillan S. (Editor). Final Report. PADTS Steering Committee, Fort Chipewyan, Alberta. Word Picture Communications. 106 pp.

Parks Canada, 2006. “Parks Canada: Report on the state of conservation of Wood Buffalo National Park. Periodic report on the application of the world heritage convention”. Section 2. accessed January 2006. Website: <www.pc.gc.ca/docs/pm-wh/rspm-whsr/rapports-reports/r11_e.asp>

Pietroniro, A., Prowse, T., and Peters, D.L., 1999. Hydrologic assessment of an inland freshwater delta using multi-temporal satellite remote sensing. *Hydrological Processes* 13: 2483-2498.

Peters, D.L., and Prowse, T.D. 2001. Regulation effects on the lower Peace River, Canada. *Hydrological Processes* 15: 3181-3194.

Peters, D. 2003. *Controls on the persistence of water in perched basins of the Peace-Athabasca Delta, northern Canada*. Ph.D Thesis. Trent University, Ontario, Canada. 196 pp.

Peters, D.L., Prowse, T., Marsh, P., Lafleur P.M., and Buttle, J. 2006. Persistence of water within perched basins of the Peace-Athabasca Delta, northern Canada. *Wetlands Ecology and Management* 144: 221-243.

Priestley, C.H.B., and Taylor, R.J. 1972. On the assessment of surface hear flux and evaporation using large-scale parameters. *Monthly Weather Review* 100:81-92.

Prowse, T.D., and Lalonde, V. 1996. Open-water and ice-jam flooding of a northern Delta. *Nordic Hydrology* 27: 85-100.

Prowse T.D., Peters D.L., Marsh P. 1996. Modelling the water balance of Peace-Athabasca Delta perched basins. Peace-Athabasca Delta Technical Studies, National Hydrology Research Institute.

Prowse, T.D., and Conly, F.M. 1998. Impacts of climatic variability and flow regulation on ice-jam flooding of a northern delta. *Hydrological Processes* 12: 1589-1610.

Prowse, T.D., Peters, D., Beltaos, S., Pietroniro, A., Romolo, L., Töyrä, J., and Leconte, R. 2002. Restoring ice-jam floodwater to a drying delta ecosystem. *Water International* 27: 58-69.

Prowse, T.D., Beltaos, S., Gartner, J.T., Gibson, J.J., Granger, R.J., Leconte, R., Peters, D.L., Pietroniro, A., Romolo, L.A., and Toth, B. 2006. Climate change, flow regulation, and land-use effects of the hydrology of the Peace-Athabasca-Slave system; Findings from the Northern Rivers Ecosystem Initiative. *Environmental Monitoring and Assessmen*t 113: 167-197.

Quinton, W.L., Hayashi, M., and Pietroniro, A. 2003. Connectivity and storage functions of channel fens and flat bogs in northern basins. *Hydrological Processes* 17: 3665–3684.

RAMP, 2006. Regional Aquatics Monitoring Program. Hatfield Group. Accessed November 2006 from http://www.ramp-alberta.org.

Rodhe, A. 1998. Snowmelt dominated systems. In Kendall C., and McDonnell, J.J., editors, *Isotope Tracers in Catchment Hydrology*, Elsevier.

Roulet, N.T. 1986. *Hydrology of a low arctic drainage basin*. Ph.D. thesis, McMaster University, Hamilton.

Roulet, N.T., and Woo, M.K., 1986. Hydrology of a wetland in the continuous permafrost region. *Journal of Hydrology* 89: 73–91.

Rouse, W.R., 2000. Progress in hydrological research in the Mackenzie GEWEX. study. *Hydrological Processes* 14: 1667–1685.

Rozanski, K., Araguás-Araguás, L., and Gonfiantini, R. 1993. Isotopic patterns in modern global precipitation. In Swart P.K., McKenzie J., Lohmann K.C. and Savin S., editors, *Climate Change in Continental Isotopic Records*, Geophysical Monograph 78, American Geophysical Union, Washington, 1-36.

Sklash, M.G. 1990. Environmental isotope studies of storm and snowmelt runoff generation. In Anderson, M.G., and Burt, T.P., editors, *Process studies in hillslope hydrology*. John Wiley and Sons, 410-435.

Smith, D.G. 1995. Glacial Lake McConnell: paleogeography, age, duration, and associated deltas, Mackenzie River basin, western Canada. *Quaternary Science Reviews* 12: 829-843.

Spence, C., and Woo, M.K. 2003. Hydrology of subarctic Canadian Shield-soil-filled valleys. *Journal of Hydrology* 279: 151-166.

Stadnyk, T., St. Amour, N., Kouwen, N., Edwards, T.W.E., Pietroniro, A., and Gibson, J.J. 2005. A groundwater separation study in boreal wetland terrain: The WATFLOOD hydrological model compared with stable isotope tracers. *Isotopes in Environmental and Health Studies* 41:1, 49-68.

St. Amour, N.N., Gibson, J.J., Edwards, T.W.D.. Prowse, T.D., and Pietroniro, A. 2005. Isotopic time-series partitioning of streamflow components in wetland-dominated catchments, lower Liard River Basin, Northwest Territories, Canada. *Hydrological Processes* 19: 3375-3381.

Taylor, S., Xiahong, F., Williams, M., and McNamara, J. 2002. How isotopic fractionation of snowmelt affects hydrograph separation. *Hydrological Processes* 16: 3683-3690.

Timoney, K. 2002. A dying delta? A case study of a wetland paradigm. *Wetlands* 22: 282-300.

Thornthwaite, C.W. 1948. An approach toward a rational classification of climate. *Geographical Review* 38: 55-94.

Townsend, G.H. 1975. Impact of the Bennett dam on the Peace-Athabasca Delta. *Journal of the Fisheries Board of Canada* 32 : 171-176.

Townsend, G.H. 1984. Wildlife resources of the Slave and Peace-Athabasca Delta. *Journal of National and Provincial Parks Association in Canada* 20: 5-7.

Töyrä, J., Pietroniro, A., and Martz, L.W. 2001. Multisensor hydrologic assessment of a freshwater wetland. *Remote Sensing of Environment* 75: 162-173.

Töyrä, J., Pietroniro, A., Hopkinson, C., and Kalbfleisch, W. 2003. Assessment of airborne scanning laser altimetry (LiDAR) in a deltaic wetland environment. *Canadian Journal of Remote Sensing* 29: 718-728.

Töyrä, J., and Pietroniro, A. 2005. Towards operational monitoring of a northern wetland using geomatics-based techniques. *Remote Sensing of Environment* 97: 174-191.

Welhan, J.A., and Fritz, P. 1977. Evaporation pan isotopic behavior as an index of isotopic evaporation conditions. *Geochimica et Cosmochimica Acta* 41: 682-686.

WBEA, 2005. Climate data. Retrieved 3 October from Wood Buffalo Environmental Association website. Website: http://www.wbea.org/html/air/station8.html.

Wetlands International, 2005. Wetlands International: Ramsar Sites Information Service. Website:http://www.wetlands.org.

Wiklund, J. In preparation. Hydroecological studies in shallow floodplain lakes of the Peace-Athabasca Delta. Ph.D. thesis, University of Waterloo.

Wolfe, B.B., Karst-Riddoch, T.L., Vardy, S., Falcone, M., Hall, R.I., and Edwards T.W.D. 2005. Impacts of climate and river flooding on the hydro-ecology of a floodplain basin, Peace-Athabasca Delta, Canada since A.D. 1700. *Quaternary Research* 64:147-162.

Wolfe, B.B., Falcone, M.D., Clogg-Wright, K.P., Mongeon, C.L., Yi, Y., Mark, W.A., St. Amour, Brock, B., N., Wiklund, J. Yi, Y., and Edwards, T.W.D. 2007a. Progress in isotope paleohydrology using lake sediment cellulose. *Journal of Paleolimnology* 37: 221-231.

Wolfe, B.B., Karst-Riddoch, T.L., Hall, R.I., Edwards, T.W.D., English, M.C., Palmini, R., McGowan, S., Leavitt, P.R., and Vardy, S.R. 2007b. Classification of hydrologic regimes of northern floodplain basins (Peace-Athabasca Delta, Canada) from analysis of stable isotopes (δ^{18}O, δ^{2}H) and water chemistry. *Hydrological Processes* 21: 151-168.

Wolfe, B.B., Hall R.I., Edwards T.W.D., Vardy, S.R., Falcone M.D., Sjunneskog, C, Sylvestre, F., McGowan, S., Leavitt, P.R., and van Driel, P. In review. Hydroecological responses of the Athabasca Delta, Canada, to changes in river flow and climate during the 20th century. *Ecosystems*.

Woo, M.K., and Young, K.L. 1998. Modeling arctic snow distribution and melt at the 1 km grid scale. *Nordic Hydrology* 35: 295-307.

Yi ,Y., Brock, B.E., Falcone, M.D., Wolfe, B.B., and Edwards, T.W.D. In preparation. A δ^{18}O-δ^{2}H coupled method to characterize source water to shallow lakes in northern delta environments.

Young, K.L., Yang, D., Killingtveit, A., Bolton, W.R., Gieck, R.E., and Shutov, V.A. 2005. A review: precipitation and snowcover in northern circumpolar basins as related to water balance studies. Northern Research Basins Symposium and Workshop, Sweden.

Zimmerman, U., 1979. Determination by stable isotopes of underground inflow and outflow and evaporation of young artificial groundwater lakes. *Isotopes in Lake Studies*, IAEA, Vienna, 87–94.

APPENDICES

APPENDIX A: Daily evaporation rates and associated variables for calculations using the Priestly-Taylor method: May 1st to October 31st 2003.

s is the slope of the saturation-vapor pressure curve at temperature *t* and L is the latent heat of vaporization at temperature *t*. Radiation is the combination of incoming long-wave and short-wave radiation.

Date	Air temp	radiation	s	L	Evaporation
	C	cal/cm^2			mm
01/05/2003	2.40	9.36	124.43	595.95	**1.97**
02/05/2003	-1.20	3.01	97.30	597.98	**0.63**
03/05/2003	-3.90	5.71	80.49	599.50	**1.19**
04/05/2003	-3.90	14.01	80.49	599.50	**2.92**
05/05/2003	0.60	6.56	110.14	596.96	**1.38**
06/05/2003	1.00	3.86	113.18	596.74	**0.81**
07/05/2003	4.90	7.07	146.96	594.54	**1.49**
08/05/2003	3.80	11.90	136.65	595.16	**2.51**
09/05/2003	3.10	14.62	130.42	595.55	**3.08**
10/05/2003	11.70	14.10	227.10	590.70	**3.00**
11/05/2003	14.00	14.48	261.67	589.40	**3.09**
12/05/2003	16.10	14.81	297.12	588.22	**3.16**
13/05/2003	13.50	14.31	253.79	589.69	**3.05**
14/05/2003	12.30	4.98	235.71	590.36	**1.06**
15/05/2003	10.50	12.79	210.69	591.38	**2.72**
16/05/2003	5.10	7.75	148.91	594.42	**1.63**
17/05/2003	-1.70	3.23	93.97	598.26	**0.68**
18/05/2003	0.10	13.33	106.43	597.24	**2.79**
19/05/2003	0.80	13.51	111.65	596.85	**2.84**
20/05/2003	5.50	14.63	152.87	594.20	**3.09**
21/05/2003	9.50	12.47	197.81	591.94	**2.65**
22/05/2003	14.10	8.27	263.27	589.35	**1.76**
23/05/2003	8.20	15.65	182.09	592.68	**3.32**
24/05/2003	19.10	13.75	354.94	586.53	**2.95**
25/05/2003	21.90	10.65	417.40	584.95	**2.29**
26/05/2003	15.90	8.83	293.58	588.33	**1.89**
27/05/2003	13.40	14.28	252.24	589.74	**3.04**
28/05/2003	14.20	14.67	264.88	589.29	**3.13**
29/05/2003	8.70	14.50	188.00	592.39	**3.07**
30/05/2003	10.70	15.58	213.35	591.27	**3.31**
31/05/2003	16.30	15.36	300.71	588.11	**3.28**
01/06/2003	19.50	13.11	363.33	586.30	**2.81**
02/06/2003	15.80	9.06	291.82	588.39	**1.94**

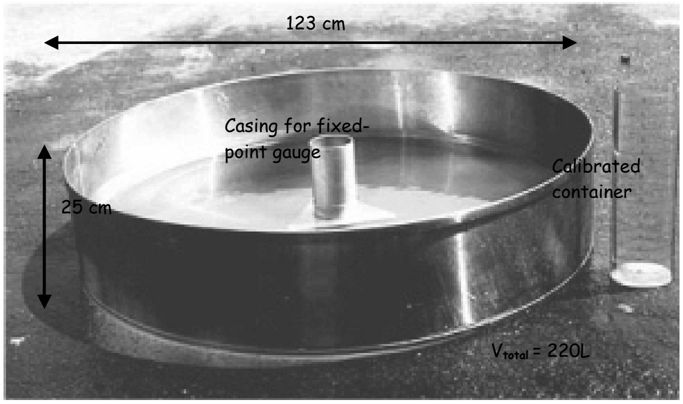
Date	Air temp	radiation	s	L	Evaporation
	C	cal/cm^2			mm
03/06/2003	17.10	8.02	315.42	587.66	**1.72**
04/06/2003	15.20	8.18	281.45	588.73	**1.75**
05/06/2003	11.70	6.33	227.10	590.70	**1.35**
06/06/2003	12.20	16.26	234.26	590.42	**3.46**
07/06/2003	12.40	5.96	237.17	590.31	**1.27**
08/06/2003	11.40	16.84	222.90	590.87	**3.58**
09/06/2003	11.80	16.60	228.51	590.64	**3.53**
10/06/2003	11.20	12.11	220.13	590.98	**2.57**
11/06/2003	15.00	12.57	278.07	588.84	**2.68**
12/06/2003	4.90	10.03	146.96	594.54	**2.12**
13/06/2003	3.20	4.43	131.29	595.50	**0.93**
14/06/2003	8.90	7.91	190.41	592.28	**1.68**
15/06/2003	14.20	16.57	264.88	589.29	**3.53**
16/06/2003	13.90	14.77	260.08	589.46	**3.15**
17/06/2003	13.90	17.43	260.08	589.46	**3.72**
18/06/2003	22.40	13.14	429.49	584.67	**2.83**
19/06/2003	15.50	13.50	286.59	588.56	**2.88**
20/06/2003	15.00	15.44	278.07	588.84	**3.30**
21/06/2003	12.40	2.80	237.17	590.31	**0.60**
22/06/2003	7.00	3.11	168.55	593.35	**0.66**
23/06/2003	11.20	11.34	220.13	590.98	**2.41**
24/06/2003	12.70	17.33	241.61	590.14	**3.69**
25/06/2003	16.50	15.58	304.33	587.99	**3.33**
26/06/2003	18.40	14.10	340.64	586.92	**3.02**
27/06/2003	14.50	9.98	269.76	589.12	**2.13**
28/06/2003	15.70	11.52	290.07	588.45	**2.46**
29/06/2003	11.90	17.33	229.94	590.59	**3.69**
30/06/2003	15.10	16.96	279.76	588.78	**3.62**
01/07/2003	13.70	4.39	256.92	589.57	**0.94**
02/07/2003	14.50	7.30	269.76	589.12	**1.56**
03/07/2003	13.90	9.89	260.08	589.46	**2.11**
04/07/2003	13.20	16.41	249.16	589.86	**3.50**
05/07/2003	12.80	3.62	243.10	590.08	**0.77**
06/07/2003	14.90	15.69	276.39	588.90	**3.35**
07/07/2003	15.50	16.97	286.59	588.56	**3.63**
08/07/2003	18.20	9.91	336.65	587.04	**2.12**
09/07/2003	16.40	4.51	302.51	588.05	**0.96**
10/07/2003	19.20	13.18	357.02	586.47	**2.83**
11/07/2003	20.50	14.32	385.08	585.74	**3.08**
12/07/2003	16.90	12.24	311.68	587.77	**2.62**
13/07/2003	13.60	16.82	255.35	589.63	**3.59**
14/07/2003	16.40	9.86	302.51	588.05	**2.11**
15/07/2003	14.20	12.31	264.88	589.29	**2.63**
16/07/2003	18.60	15.63	344.68	586.81	**3.35**
17/07/2003	18.70	5.86	346.71	586.75	**1.26**

Date	Air temp	radiation	s	L	Evaporation
	C	cal/cm^2			mm
18/07/2003	17.10	11.30	315.42	587.66	**2.42**
19/07/2003	16.10	5.56	297.12	588.22	**1.19**
20/07/2003	17.00	12.15	313.54	587.71	**2.60**
21/07/2003	17.60	10.56	324.92	587.37	**2.26**
22/07/2003	20.80	15.46	391.81	585.57	**3.32**
23/07/2003	19.00	15.33	352.86	586.58	**3.29**
24/07/2003	17.50	8.65	323.00	587.43	**1.85**
25/07/2003	19.50	11.28	363.33	586.30	**2.42**
26/07/2003	20.50	14.61	385.08	585.74	**3.14**
27/07/2003	23.00	13.48	444.40	584.33	**2.90**
28/07/2003	18.40	8.18	340.64	586.92	**1.75**
29/07/2003	19.80	14.24	369.74	586.13	**3.06**
30/07/2003	23.40	14.18	454.59	584.10	**3.06**
31/07/2003	21.90	13.30	417.40	584.95	**2.86**
01/08/2003	6.80	12.52	549.59	582.18	**2.71**
02/08/2003	19.70	7.94	367.59	586.19	**1.70**
03/08/2003	15.80	13.70	291.82	588.39	**2.93**
04/08/2003	12.20	2.84	234.26	590.42	**0.60**
05/08/2003	12.10	12.17	232.81	590.48	**2.59**
06/08/2003	17.40	8.58	321.09	587.49	**1.84**
07/08/2003	16.50	8.31	304.33	587.99	**1.78**
08/08/2003	18.10	12.94	334.67	587.09	**2.77**
09/08/2003	17.70	12.58	326.85	587.32	**2.69**
10/08/2003	15.30	7.71	283.16	588.67	**1.65**
11/08/2003	12.30	7.73	235.71	590.36	**1.65**
12/08/2003	15.30	12.93	283.16	588.67	**2.76**
13/08/2003	17.90	10.53	330.74	587.20	**2.25**
14/08/2003	19.10	12.67	354.94	586.53	**2.72**
15/08/2003	22.20	12.43	424.62	584.78	**2.67**
16/08/2003	21.20	6.79	400.96	585.34	**1.46**
17/08/2003	18.00	11.96	332.70	587.15	**2.56**
18/08/2003	16.50	7.20	304.33	587.99	**1.54**
19/08/2003	15.40	5.48	284.87	588.61	**1.17**
20/08/2003	15.10	7.91	279.76	588.78	**1.69**
21/08/2003	14.10	11.15	263.27	589.35	**2.38**
22/08/2003	15.60	10.32	288.33	588.50	**2.21**
23/08/2003	14.60	4.06	271.41	589.07	**0.87**
24/08/2003	12.70	6.07	241.61	590.14	**1.29**
25/08/2003	10.70	3.54	213.35	591.27	**0.75**
26/08/2003	11.00	10.47	217.40	591.10	**2.23**
27/08/2003	12.30	11.28	235.71	590.36	**2.40**
28/08/2003	12.40	11.26	237.17	590.31	**2.40**
29/08/2003	14.40	8.82	268.13	589.18	**1.88**
30/08/2003	14.00	9.87	261.67	589.40	**2.10**
31/08/2003	14.40	8.50	268.13	589.18	**1.81**

Date	Air temp	radiation	s	L	Evaporation
	C	cal/cm^2			mm
01/09/2003	14.70	1.09	273.06	589.01	**0.23**
02/09/2003	12.40	5.43	237.17	590.31	**1.16**
03/09/2003	16.80	9.82	309.83	587.82	**2.10**
04/09/2003	15.00	7.81	278.07	588.84	**1.67**
05/09/2003	12.80	2.89	243.10	590.08	**0.62**
06/09/2003	14.40	6.09	268.13	589.18	**1.30**
07/09/2003	18.30	7.81	338.64	586.98	**1.67**
08/09/2003	15.50	3.89	286.59	588.56	**0.83**
09/09/2003	13.90	7.19	260.08	589.46	**1.53**
10/09/2003	11.80	1.98	228.51	590.64	**0.42**
11/09/2003	7.90	4.86	178.62	592.84	**1.03**
12/09/2003	9.00	7.00	191.63	592.22	**1.48**
13/09/2003	8.80	8.23	189.20	592.34	**1.74**
14/09/2003	9.30	3.62	195.31	592.05	**0.77**
15/09/2003	4.20	3.87	140.32	594.93	**0.82**
16/09/2003	2.50	4.05	125.27	595.89	**0.85**
17/09/2003	4.20	2.73	140.32	594.93	**0.57**
18/09/2003	5.00	4.22	147.93	594.48	**0.89**
19/09/2003	10.20	3.65	206.75	591.55	**0.77**
20/09/2003	7.40	4.21	172.96	593.13	**0.89**
21/09/2003	4.10	7.63	139.40	594.99	**1.61**
22/09/2003	4.60	3.69	144.08	594.71	**0.78**
23/09/2003	6.20	0.97	160.02	593.80	**0.21**
24/09/2003	4.50	2.53	143.14	594.76	**0.53**
25/09/2003	3.40	1.65	133.06	595.38	**0.35**
26/09/2003	2.20	2.72	122.77	596.06	**0.57**
27/09/2003	3.20	2.93	131.29	595.50	**0.62**
28/09/2003	2.10	4.23	121.94	596.12	**0.89**
29/09/2003	2.20	6.20	122.77	596.06	**1.30**
30/09/2003	4.80	6.20	146.00	594.59	**1.31**
01/10/2003	9.20	4.84	194.08	592.11	**1.03**
02/10/2003	8.60	5.83	186.80	592.45	**1.24**
03/10/2003	12.30	5.74	235.71	590.36	**1.22**
04/10/2003	13.10	5.59	247.64	589.91	**1.19**
05/10/2003	12.80	5.42	243.10	590.08	**1.15**
06/10/2003	9.50	5.20	197.81	591.94	**1.10**
07/10/2003	12.60	4.74	240.12	590.19	**1.01**
08/10/2003	13.60	3.61	255.35	589.63	**0.77**
09/10/2003	6.60	0.91	164.24	593.58	**0.19**
10/10/2003	5.70	2.09	154.89	594.09	**0.44**
11/10/2003	5.40	1.87	151.87	594.25	**0.39**
12/10/2003	3.70	2.30	135.74	595.21	**0.48**
13/10/2003	1.00	1.32	113.18	596.74	**0.28**
14/10/2003	-1.00	0.92	98.66	597.86	**0.19**
15/10/2003	0.90	1.45	112.42	596.79	**0.31**

Date	Air temp	radiation	s	L	Evaporation
	C	cal/cm^2			mm
16/10/2003	0.50	3.60	109.39	597.02	**0.75**
17/10/2003	2.20	0.67	122.77	596.06	**0.14**
18/10/2003	2.10	1.13	121.94	596.12	**0.24**
19/10/2003	4.40	0.67	142.19	594.82	**0.14**
20/10/2003	4.00	0.65	138.47	595.04	**0.14**
21/10/2003	2.10	0.27	121.94	596.12	**0.06**
22/10/2003	2.70	1.09	126.97	595.78	**0.23**
23/10/2003	1.60	0.27	117.89	596.40	**0.06**
24/10/2003	0.60	1.63	110.14	596.96	**0.34**
25/10/2003	-0.80	0.80	100.03	597.75	**0.17**
26/10/2003	2.50	0.68	125.27	595.89	**0.14**
27/10/2003	1.50	1.79	117.10	596.45	**0.38**
28/10/2003	-1.70	1.44	93.97	598.26	**0.30**
29/10/2003	-6.20	1.66	68.24	600.80	**0.34**
30/10/2003	-8.70	1.78	56.80	602.21	**0.37**
31/10/2003	-9.90	1.15	51.94	602.88	**0.24**

APPENDIX B: Schematics of the class-A evaporation pan.



APPENDIX C: Isotope results of snowmelt and rain between October 2000 and September 2005.

SNOW

Date	**$\delta^{18}O$**	**$\delta^{2}H$**
25-Oct-00	-19.7	-158.0
25-Oct-00	-20.2	-154.2
25-Oct-00	-19.7	-158.0
25-Oct-00	-20.2	-154.2
4-Mar-01	-21.5	-173.8
4-Mar-01	-25.4	-190.7
4-Mar-01	-28.8	-223.1
5-Mar-01	-25.7	-191.6
5-Mar-01	-29.0	-225.5
5-Mar-01	-30.6	-233.4
6-Mar-01	-23.1	-184.3
6-Mar-01	-27.0	-207.2
6-Mar-01	-28.3	-215.0
7-Mar-01	-21.9	-177.4
7-Mar-01	-25.4	-190.2
7-Mar-01	-27.8	-215.7
30-Apr-03	-19.4	-140.7
20-May-03	-24.5	-187.7
20-May-03	-24.3	-184.8
2-May-04	-24.7	-188.8
2-May-04	-28.7	-224.7
2-May-04	-23.7	-184.7
2-May-04	-23.6	-183.6
16-Mar-05	-21.3	-165.8
16-Mar-05	-16.3	-176.1
16-Mar-05	-25.1	-196.3
16-Mar-05	-22.9	-180.3
16-Mar-05	-24.6	-189.2
12-Mar-05	-15.9	-145.7
12-Mar-05	-22.8	-177.7
12-Mar-05	-22.5	-176.8
12-Mar-05	-24.4	-195.5
12-Mar-05	-21.9	-172.4
12-Mar-05	-27.5	-212.7

RAIN

Date	**$\delta^{18}O$**	**$\delta^{2}H$**
25-Oct-00	-12.8	-116.3
25-Oct-00	-12.6	-116.1
25-Oct-00	-14.9	-124.9
5-Jun-03	-17.6	-141.3
10-Jun-03	-17.3	-140.7
16-Jun-03	-19.1	-154.8
18-Jun-03	-14.9	-124.7
23-Jun-03	-16.6	-123.6
25-Jun-03	-16.8	-132.4
27-Jun-03	-17.0	-138.4
28-Jun-03	-16.8	-129.8
3-Jul-03	-16.5	-133.5
5-Jul-03	-15.7	-121.0
6-Jul-03	-15.7	-120.0
15-Jul-03	-16.7	-137.7
18-Jul-03	-13.7	-112.3
22-Jul-03	-14.9	-119.6
23-Jul-03	-15.5	-122.7
24-Jul-03	-15.0	-111.5
30-Jul-03	-14.5	-116.5
15-Aug-03	-17.5	-144.4
18-Aug-03	-14.3	-115.7
18-Aug-03	-14.0	-115.3
20-Aug-03	-20.5	-167.9
14-Jul-04	-14.0	-121.7
16-Jul-04	-13.3	-119.7
18-Jul-04	-14.2	-126.5
19-Jul-04	-14.1	-122.6
22-Jul-04	-12.3	-119.3
23-Jul-04	-13.8	-125.3
1-Aug-04	-18.1	-142.6
12-Sep-04	-16.9	-136.3
13-Sep-04	-16.4	-132.3
15-Sep-04	-16.7	-135.7

RAIN

Date	**$\delta^{18}O$**	**$\delta^{2}H$**
16-Sep-04	-16.1	-134.9
17-Sep-04	-17.2	-138.5
22-Sep-04	-16.4	-130.8
02-Jun-05	-11.1	-79.2
20-Jul-05	-16.5	-131.4
16-Sept-05	-10.8	-113.6

APPENDIX D: Isotope results of rivers between October 2000 and September 2005.

River	Date	$\delta^{18}O$	$\delta^{2}H$
R1	25-Oct-00	-15.3	-131.2
R2	25-Oct-00	-17.9	-149.6
R3	25-Oct-00	-15.7	-136.7
R4	25-Oct-00	-16.9	-140.1
R5	25-Oct-00	-16.9	-141.7
R6	25-Oct-00	-16.9	-140.1
R7	25-Oct-00	-17.0	-142.4
R8	25-Oct-00	-18.5	-146.7
R6	05-Jun-01	-17.4	-139.1
R5	09-Jun-01	-18.1	-149.7
R4	07-Jun-01	-18.1	-149.1
R2	04-Jun-01	-15.5	-131.3
R1	27-Aug-01	-14.4	-128.5
R2	27-Aug-01	-17.7	-146.1
R3	27-Aug-01	-15.3	-132.2
R4	28-Aug-01	-17.1	-139.4
R5	28-Aug-01	-16.8	-138.3
R6	28-Aug-01	-17.3	-140.0
R7	28-Aug-01	-16.6	-139.4
R11	27-Aug-01	-14.5	-128.2
R1	03-Jun-02	-17.1	-140.7
R2	03-Jun-02	-20.1	-158.8
R3	03-Jun-02	-17.7	-145.9
R4	03-Jun-02	-19.2	-153.6
R5	03-Jun-02	-19.2	-153.5
R6	03-Jun-02	-19.3	-150.3
R7	03-Jun-02	-19.3	-151.8
R8	03-Jun-02	-20.4	-158.0
R9	03-Jun-02	-9.8	-110.0
R10	03-Jun-02	-19.6	-154.9
R11	03-Jun-02	-17.6	-142.2
R27	03-Jun-02	-16.0	-133.7
R1	21-Sep-02	-16.6	-135.0
R2	21-Sep-02	-18.2	-143.7
R3	21-Sep-02	-16.6	-135.0
R4	21-Sep-02	-17.8	-140.5
R5	21-Sep-02	-17.7	-142.0
R6	21-Sep-02	-17.7	-142.4
R7	21-Sep-02	-17.8	-141.8
R11	21-Sep-02	-15.7	-131.4
R27	21-Sep-02	-15.2	-131.8

River	Date	$\delta^{18}O$	$\delta^{2}H$
R1	30-Apr-03	-21.5	-165.9
R4	30-Apr-03	-19.3	-152.8
R5	01-May-03	-19.2	-151.2
R6	01-May-03	-19.2	-151.9
R7	01-May-03	-19.2	-152.1
R8	01-May-03	-21.1	-164.7
R9	01-May-03	-21.1	-165.1
R10	01-May-03	-21.3	-166.5
R11	30-Apr-03	-21.0	-165.0
R27	30-Apr-03	-17.4	-142.9
R2	02-May-04	-19.1	-150.3
R3	02-May-04	-19.0	-151.1
R4	02-May-04	-17.6	-144.8
R5	02-May-04	-17.9	-143.5
R6	02-May-04	-18.1	-144.1
R7	02-May-04	-17.8	-143.7
R8	02-May-04	-19.0	-151.8
R9	02-May-04	-15.0	-138.5
R10	02-May-04	-18.5	-154.8
R11	02-May-04	-17.4	-142.0
R27	02-May-04	-16.8	-140.9
R27	31-May-04	-16.3	-139.4
R11	31-May-04	-16.9	-142.6
R4	5-Jun-04	-17.3	-139.0
R2	5-Jun-04	-18.7	-150.7
R3	5-Jun-04	-16.0	-135.4
R27	7-Jun-04	-16.0	-137.8
R7	10-Jun-04	-17.6	-139.8
R1	11-Jun-04	-16.8	-139.7
R3	11-Jun-04	-16.5	-138.8
R 1	4-May-05	-17.6	-143.1
R 2	4-May-05	-19.0	-151.4
R 3	4-May-05	-16.6	-138.1
R 4	4-May-05	-17.5	-142.0
R 7	4-May-05	-17.4	-142.8
R 11	4-May-05	-17.9	-144.7
R27	4-May-05	-17.8	-146.4

APPENDIX D (continued):

River	Date	$\delta^{18}O$	$\delta^{2}H$
R 1	18-Jul-05	-16.7	-136.3
R 2	4-Aug-05	-18.9	-148.8
R 3	18-Jul-05	-17.3	-140.0
R 4	18-Jul-05	-17.6	-140.9
R 7	18-Jul-05	-17.7	-141.8
R 11	4-Aug-05	-15.8	-133.3
R 27	20-Jul-05	-16.8	-138.7
R 1	Sep-14-05	-15.4	-130.9
R 2	Sep-14-05	-18.4	-147.1
R 3	Sep-14-05	-16.0	-134.1
R 4	Sep-14-05	-17.0	-137.5
R 7	Sep-14-05	-17.0	-138.3
R 11	Sep-14-05	-15.3	-130.6
R 27	Sep-14-05	-15.5	-132.2

APPENDIX E: Isotope results of PAD 18 (Greenstar Lake) between October 2000 and September 2005.

Date	$\delta^{18}O$	$\delta^{2}H$
10/25/2000	-8.2	-104.2
3/6/2001	-9.0	-105.8
6/3/2001	-9.0	-103.5
8/27/2001	-7.6	-101.7
6/3/2002	-9.1	-101.8
7/25/2002	-8.8	-100.7
8/25/2002	-8.8	-101.9
9/21/2002	-9.0	-100.6
4/30/2003	-10.5	-109.2
5/22/2003	-9.6	-105.5
5/28/2003	-9.7	-105.6
6/25/2003	-9.5	-106.1
7/24/2003	-9.3	-103.4
8/15/2003	-9.2	-104.9
6/4/2004	-9.4	-108.0
7/18/2004	-9.0	-104.0
9/23/2004	-8.9	-103.9
5/16/2005	-9.7	-103.6
7/18/2005	-9.3	-104.7
9/14/2005	-9.6	-105.7

APPENDIX F: Isotope results between October 2000 and September 2005 from lakes used to develop the isotopic framework.

Spruce Island Lake (PAD 5) and Johnny Cabin Pond (PAD 31) were also used and can be found in Appendix H.

PAD 1
("Devils Gate Pond")

Date	$\delta^{18}O$	$\delta^{2}H$
10/25/2000	-7.9	-98
6/3/2001	-8.6	-101
8/27/2001	-4.3	-83
6/3/2002	-8.2	-97
4/30/2003	-13.8	-124
5/22/2003	-13.7	-126
6/8/2003	-12.5	-123
6/30/2003	-11.4	-116
7/18/2003	-10.5	-112
8/20/2003	-8.8	-102
5/2/2004	-13.8	-126
6/7/2004	-12.1	-118
7/17/2004	-9.0	-104
7/28/2004	-10.7	-110
8/1/2004	-9.3	-107
8/21/2004	-8.9	-105
9/15/2004	-9.0	-106
9/23/2004	-9.6	-110
5/16/2005	-15.2	-139
6/5/2005	-14.1	-130
7/16/2005	-12.2	-122
7/20/2005	-12.2	-121
9/14/2005	-12.8	-122

PAD 8
("Chilloney's Creek Pond")

Date	$\delta^{18}O$	$\delta^{2}H$
10/25/2000	-11.4	-114
3/7/2001	-13.7	-128
6/3/2001	-12.3	-117
6/5/2001	-9.6	-105
8/27/2001	-9.8	-108
6/3/2002	-14.2	-126
7/22/2002	-11.1	-115
8/22/2002	-11.3	-114
9/21/2002	-11.5	-113
4/30/2003	-15.9	-137
5/22/2003	-17.0	-145
6/1/2003	-16.6	-141
6/13/2003	-16.1	-138
7/12/2003	-15.1	-133
8/9/2003	-13.9	-127
5/2/2004	-18.0	-149
5/31/2004	-15.7	-137
5/31/2004	-15.7	-137
6/7/2004	-15.5	-135
6/28/2004	-14.3	-130
7/17/2004	-13.4	-126
7/22/2004	-13.9	-128
8/21/2004	-12.7	-122
9/16/2004	-13.2	-124
9/23/2004	-14.2	-127
5/16/2005	-18.1	-150
6/5/2005	-16.4	-140
7/16/2005	-14.7	-134
7/20/2005	-14.7	-132
9/14/2005	-13.8	-127
9/14/2005	-14.7	-125

PAD 9

Date	$\delta^{18}O$	$\delta^{2}H$
10/25/2000	-7.5	-101
6/3/2001	-9.1	-104
6/3/2001	-9.1	-103
8/27/2001	-4.9	-88
6/3/2002	-8.8	-101
7/25/2002	-5.7	-86
8/25/2002	-7.0	-93
9/21/2002	-8.9	-100
5/1/2003	-15.2	-134
5/22/2003	-14.3	-133
5/27/2003	-13.9	-126
6/26/2003	-11.3	-119
7/21/2003	-9.6	-107
8/26/2003	-7.9	-102
9/23/2003	-8.2	-100
5/2/2004	-14.7	-133
6/1/2004	-13.4	-131
7/17/2004	-7.8	-104
9/22/2004	-9.5	-107
5/16/2005	-17.1	-143
7/20/2005	-12.1	-122
9/14/2005	-10.9	-113

PAD 15

Date	$\delta^{18}O$	$\delta^{2}H$
10/25/2000	-8.7	-111
3/5/2001	-9.2	-116
6/3/2001	-9.7	-109
8/27/2001	-8.4	-105
6/3/2002	-12.3	-119
7/24/2002	-10.3	-110
8/24/2002	-10.0	-111
9/21/2002	-10.4	-110
4/30/2003	-18.7	-151
5/22/2003	-16.4	-142
6/2/2003	-16.0	-139
8/7/2003	-12.6	-123
9/23/2003	-13.0	-123
6/14/2004	-13.2	-127
5/2/2004	-13.3	-128
7/14/2004	-13.2	-127
7/16/2004	-10.9	-119
7/29/2004	-11.3	-118
9/23/2004	-10.2	-111
5/16/2005	-16.1	-140
7/18/2005	-13.9	-130
9/14/2005	-12.9	-125

PAD 23

Date	$\delta^{18}O$	$\delta^{2}H$
10/25/2000	-10.1	-108
6/7/2001	-10.2	-108
8/28/2001	-7.0	-97
6/3/2002	-9.3	-102
7/23/2002	-8.2	-97
8/23/2002	-8.5	-98
9/21/2002	-8.8	-99
4/30/2003	-13.0	-122
5/22/2003	-11.9	-118
5/29/2003	-11.7	-116
6/28/2003	-10.7	-111
7/23/2003	-10.1	-106
8/26/2003	-9.2	-106
9/23/2003	-9.1	-103
5/2/2004	-11.6	-117
6/5/2004	-11.0	-112
7/19/2004	-9.0	-105
9/23/2004	-9.1	-105
5/16/2005	-13.3	-126
7/18/2005	-11.4	-115
9/14/2005	-11.0	-113

PAD 45 (Mamawi Lake)

Date	$\delta^{18}O$	$\delta^{2}H$
10/25/2000	-17.3	-139
6/7/2001	-13.4	-120
6/3/2002	-19.1	-153
5/22/2003	-17.8	-144
5/26/2003	-18.1	-143
6/10/2003	-17.4	-142
6/24/2003	-18.0	-144
7/11/2003	-17.8	-141
8/12/2003	-17.2	-139
5/2/2004	-17.5	-145
6/9/2004	-17.4	-141
7/14/2004	-18.7	-147
7/17/2004	-18.4	-146
9/13/2004	-17.9	-142
9/19/2004	-17.6	-141
5/16/2005	-17.4	-141
7/20/2005	-15.7	-132

PAD 54 ("Horseshoe Slough")

Date	$\delta^{18}O$	$\delta^{2}H$
10/25/2000	-11.2	-115
3/4/2001	-11.3	-121
6/4/2001	-11.0	-117
8/27/2001	-12.0	-121
6/3/2002	-12.9	-123
7/24/2002	-11.8	-118
8/24/2002	-11.6	-116
9/21/2002	-11.7	-116
5/1/2003	-20.2	-157
5/22/2003	-18.5	-151
6/4/2003	-17.9	-148
6/16/2003	-17.8	-148
7/3/2003	-17.3	-146
7/15/2003	-16.8	-146
9/23/2003	-15.6	-137
5/2/2004	-15.4	-138
6/11/2004	-15.4	-139
7/16/2004	-14.5	-128
7/28/2004	-14.3	-133
9/17/2004	-13.5	-130
9/19/2004	-13.5	-129
5/16/2005	-17.0	-145
7/18/2005	-15.7	-136
7/28/2005	-15.6	-137
9/14/2005	-15.2	-134

APPENDIX G: Isotope results of Spruce Island Lake, Johnny Cabin Pond, and Mamawi Creek between October 2000 and September 2005.

Spruce Island Lake (PAD 5)

Date	$\delta^{18}O$	$\delta^{2}H$
10/25/2000	-6.7	-95.8
6/3/2001	-8.1	-98.6
8/27/2001	-3.8	-81.5
6/3/2002	-7.8	-97.1
7/24/2002	-4.7	-86.6
8/22/2002	-6.6	-93.5
9/21/2002	-8.0	-96.6
4/30/2003	-11.6	-115.4
5/22/2003	-11.5	-116.7
6/27/2003	-9.8	-110.0
7/22/2003	-8.6	-102.1
8/25/2003	-7.4	-99.0
9/23/2003	-8.0	-99.1
5/2/2004	-12.6	-122.3
6/2/2004	-11.3	-118.4
7/16/2004	-7.8	-102.3
9/23/2004	-8.4	-101.5
5/15/2005	-13.4	-131.6
7/19/2005	-9.9	-111.8
9/14/2005	-10.1	-111.4
9/15/2005	-10.4	-111.8

Johnny Cabin Pond (PAD 31)

Date	$\delta^{18}O$	$\delta^{2}H$
10/25/2000	-11.4	-117.1
6/6/2001	-12.9	-121.3
8/28/2001	-15.8	-134.2
6/3/2002	-14.9	-126.3
7/31/2002	-11.8	-114.9
8/22/2002	-10.8	-108.5
9/21/2002	-10.9	-109.0
5/1/2003	-19.0	-150.9
5/22/2003	-17.3	-143.8
6/6/2003	-16.3	-138.1
6/18/2003	-15.7	-136.0
7/17/2003	-16.0	-134.8
7/30/2003	-15.2	-129.6
8/21/2003	-13.8	-124.4
9/23/2003	-13.1	-120.3
5/2/2004	-14.8	-130.1
6/13/2004	-13.2	-123.8
6/29/2004	-13.6	-125.5
7/17/2004	-16.5	-136.3
7/23/2004	-17.7	-142.2
8/23/2004	-15.3	-131.7
9/12/2004	-14.9	-130.7
9/19/2004	-15.0	-130.5
9/24/2004	-14.9	-129.1
5/16/2005	-17.2	-140.9
6/15/2005	-16.9	-138.4
7/18/2005	-17.3	-139.3
9/14/2005	-15.6	-130.3

Mamawi Creek (PAD R7)

Date	$\delta^{18}O$	$\delta^{2}H$
10/25/2000	-17.0	-142.4
8/28/2001	-16.6	-139.4
6/3/2002	-19.3	-151.8
9/21/2002	-17.8	-141.8
5/1/2003	-19.2	-152.1
5/22/2003	-18.1	-143.3
6/18/2003	-18.5	-147.9
7/17/2003	-17.7	-142.1
7/30/2003	-17.3	-137.8
8/21/2003	-16.9	-135.8
5/2/2004	-17.8	-143.7
6/10/2004	-17.6	-139.8
7/17/2004	-18.6	-146.4
7/23/2004	-18.1	-145.4
9/12/2004	-18.0	-143.1
9/19/2004	-17.7	-142.2
5/16/2005	-17.4	-142.8
7/18/2005	-17.7	-141.3
9/14/2005	-17.0	-138.3

APPENDIX H: Water level change (mm) of Spruce Island Lake between June 27th and August 25th 2003.

Date	Water Level Change
6/27/2003	0
6/28/2003	2
6/29/2003	3
6/30/2003	-1
7/1/2003	-5
7/2/2003	-4
7/3/2003	-5
7/4/2003	-6
7/5/2003	-8
7/6/2003	-11
7/7/2003	-15
7/8/2003	-18
7/9/2003	-20
7/10/2003	-23
7/11/2003	-27
7/12/2003	-31
7/13/2003	-37
7/14/2003	-40
7/15/2003	-39
7/16/2003	-40
7/17/2003	-34
7/18/2003	-18
7/19/2003	-20
7/20/2003	-22
7/21/2003	-23
7/22/2003	-24
7/23/2003	-26
7/24/2003	-30
7/25/2003	-32
7/26/2003	-36
7/27/2003	-40
7/28/2003	-44
7/29/2003	-44
7/30/2003	-49
7/31/2003	-54
8/1/2003	-57
8/2/2003	-0.1
8/3/2003	4
8/4/2003	36
8/5/2003	-31
8/6/2003	-63
8/7/2003	-64
8/8/2003	-67
8/9/2003	-71
8/10/2003	-72
8/11/2003	-73
8/12/2003	-75
8/13/2003	-79
8/14/2003	-83
8/15/2003	-87
8/16/2003	-83
8/17/2003	-87
8/18/2003	7
8/19/2003	31
8/20/2003	-18
8/21/2003	-96
8/22/2003	-93
8/23/2003	-87
8/24/2003	-88
8/25/2003	-90

APPENDIX I: Water level (m.a.s.l.) of Mamawi Lake between April 10th and October 27th 2003.

Water level data is courtesy of Jay Joyner of British Columbia Hydro and Power and Authority

Date	Water Level	Date	Water Level	Date	Water Level	Date	Water Level
4/10/2003	208.07	6/1/2003	209.18	7/23/2003	209.33	9/14/2003	208.98
4/11/2003	208.04	6/2/2003	209.15	7/24/2003	209.34	9/15/2003	209.05
4/12/2003	208.03	6/3/2003	209.17	7/25/2003	209.36	9/16/2003	209.02
4/13/2003	208.03	6/4/2003	209.18	7/26/2003	209.37	9/17/2003	209.01
4/14/2003	208.03	6/5/2003	209.20	7/27/2003	209.32	9/18/2003	208.96
4/15/2003	208.05	6/6/2003	209.21	7/28/2003	209.30	9/19/2003	208.93
4/16/2003	208.08	6/7/2003	209.23	7/29/2003	209.34	9/20/2003	208.91
4/17/2003	208.10	6/8/2003	209.24	7/30/2003	209.33	9/21/2003	208.89
4/18/2003	208.14	6/9/2003	209.20	7/31/2003	209.33	9/22/2003	208.92
4/19/2003	208.20	6/10/2003	209.25	8/1/2003	209.32	9/23/2003	208.94
4/20/2003	208.35	6/11/2003	209.24	8/2/2003	209.27	9/24/2003	208.90
4/21/2003	208.57	6/12/2003	209.24	8/3/2003	209.29	9/25/2003	208.87
4/22/2003	208.77	6/13/2003	209.28	8/4/2003	209.31	9/26/2003	208.89
4/23/2003	208.94	6/14/2003	209.24	8/5/2003	209.27	9/27/2003	208.88
4/24/2003	209.08	6/15/2003	209.23	8/6/2003	209.27	9/28/2003	208.87
4/25/2003	209.19	6/16/2003	209.23	8/7/2003	209.28	9/29/2003	208.86
4/26/2003	209.29	6/17/2003	209.26	8/8/2003	209.26	9/30/2003	208.85
4/27/2003	209.40	6/18/2003	209.25	8/9/2003	209.27	10/1/2003	208.78
4/28/2003	209.47	6/19/2003	209.20	8/10/2003	209.25	10/2/2003	208.77
4/29/2003	209.53	6/20/2003	209.29	8/11/2003	209.24	10/3/2003	208.79
4/30/2003	209.57	6/21/2003	209.33	8/12/2003	209.23	10/4/2003	208.79
5/1/2003	209.62	6/22/2003	209.37	8/13/2003	209.21	10/5/2003	208.78
5/2/2003	209.73	6/23/2003	209.30	8/14/2003	209.18	10/6/2003	208.82
5/3/2003	209.87	6/24/2003	209.25	8/15/2003	209.22	10/7/2003	208.85
5/4/2003	209.90	6/25/2003	209.27	8/16/2003	209.18	10/8/2003	208.76
5/5/2003	209.83	6/26/2003	209.24	8/17/2003	209.10	10/9/2003	208.71
5/6/2003	209.79	6/27/2003	209.26	8/18/2003	209.18	10/10/2003	208.68
5/7/2003	209.69	6/28/2003	209.30	8/19/2003	209.10	10/11/2003	208.68
5/8/2003	209.49	6/29/2003	209.33	8/20/2003	209.16	10/12/2003	208.67
5/9/2003	209.24	6/30/2003	209.36	8/21/2003	209.18	10/13/2003	208.68
5/10/2003	209.10	7/1/2003	209.36	8/22/2003	209.18	10/14/2003	208.68
5/11/2003	209.04	7/2/2003	209.36	8/23/2003	209.11	10/15/2003	208.68
5/12/2003	209.04	7/3/2003	209.33	8/24/2003	208.94	10/16/2003	208.67
5/13/2003	208.99	7/4/2003	209.39	8/25/2003	209.06	10/17/2003	208.67
5/14/2003	208.99	7/5/2003	209.42	8/26/2003	209.15	10/18/2003	208.67
5/15/2003	208.97	7/6/2003	209.37	8/27/2003	209.16	10/19/2003	208.69
5/16/2003	209.00	7/7/2003	209.35	8/28/2003	209.15	10/20/2003	208.66
5/17/2003	209.01	7/8/2003	209.31	8/30/2003	209.12	10/21/2003	208.70
5/18/2003	208.99	7/9/2003	209.34	8/31/2003	209.12	10/22/2003	208.68
5/19/2003	209.01	7/10/2003	209.34	9/1/2003	209.12	10/23/2003	208.75
5/20/2003	209.05	7/11/2003	209.34	9/2/2003	209.11	10/24/2003	208.73
5/21/2003	209.07	7/12/2003	209.34	9/3/2003	209.04	10/25/2003	208.69
5/22/2003	209.08	7/13/2003	209.34	9/4/2003	209.07	10/26/2003	208.70
5/23/2003	209.13	7/14/2003	209.35	9/5/2003	209.08	10/27/2003	208.66
5/24/2003	209.12	7/15/2003	209.34	9/6/2003	209.08		
5/25/2003	209.11	7/16/2003	209.35	9/7/2003	209.06		
5/26/2003	209.12	7/17/2003	209.33	9/8/2003	209.04		
5/27/2003	209.12	7/18/2003	209.35	9/9/2003	209.04		
5/28/2003	209.13	7/19/2003	209.36	9/10/2003	209.06		
5/29/2003	209.22	7/20/2003	209.37	9/11/2003	209.04		
5/30/2003	209.16	7/21/2003	209.37	9/12/2003	209.05		
5/31/2003	209.15	7/22/2003	209.35	9/13/2003	209.01		

APPENDIX J: Water level (m.a.s.l.) of Creed Creek between June 7th and September 22nd 2004.

Water level data is courtesy of Jay Joyner of British Columbia Hydro and Power and Authority.

Date	Water Level	Date	Water Level	Date	Water Level
6/7/2004	210.82	7/23/2004	211.69	9/7/2004	209.60
6/8/2004	210.92	7/24/2004	211.53	9/8/2004	209.60
6/9/2004	211.01	7/25/2004	211.40	9/9/2004	209.63
6/10/2004	211.00	7/26/2004	211.27	9/10/2004	209.82
6/11/2004	210.92	7/27/2004	211.17	9/11/2004	210.36
6/12/2004	210.82	7/28/2004	211.08	9/12/2004	210.69
6/13/2004	210.69	7/29/2004	211.03	9/13/2004	210.74
6/14/2004	210.57	7/30/2004	210.96	9/14/2004	210.70
6/15/2004	210.58	7/31/2004	210.86	9/15/2004	210.62
6/16/2004	210.95	8/1/2004	210.74	9/16/2004	210.54
6/17/2004	211.16	8/2/2004	210.64	9/17/2004	210.45
6/18/2004	211.11	8/3/2004	210.59	9/18/2004	210.38
6/19/2004	211.01	8/4/2004	210.54	9/19/2004	210.36
6/20/2004	211.06	8/5/2004	210.51	9/20/2004	210.43
6/21/2004	211.36	8/6/2004	210.47	9/21/2004	210.52
6/22/2004	211.70	8/7/2004	209.99	9/22/2004	210.52
6/23/2004	211.66	8/8/2004	209.94		
6/24/2004	211.54	8/9/2004	209.89		
6/25/2004	211.38	8/10/2004	209.86		
6/26/2004	211.22	8/11/2004	209.81		
6/27/2004	211.08	8/12/2004	209.77		
6/28/2004	210.97	8/13/2004	209.73		
6/29/2004	210.86	8/14/2004	209.74		
6/30/2004	210.81	8/15/2004	209.75		
7/1/2004	210.81	8/16/2004	209.83		
7/2/2004	210.84	8/17/2004	209.89		
7/3/2004	210.85	8/18/2004	209.86		
7/4/2004	210.85	8/19/2004	209.81		
7/5/2004	210.82	8/20/2004	209.77		
7/6/2004	210.74	8/21/2004	209.71		
7/7/2004	210.65	8/22/2004	209.65		
7/8/2004	210.58	8/23/2004	209.62		
7/9/2004	210.51	8/24/2004	209.58		
7/10/2004	210.49	8/25/2004	209.57		
7/11/2004	210.53	8/26/2004	209.57		
7/12/2004	210.60	8/27/2004	209.58		
7/13/2004	210.73	8/28/2004	209.58		
7/14/2004	210.86	8/29/2004	209.58		
7/15/2004	210.92	8/30/2004	209.58		
7/16/2004	211.21	8/31/2004	209.57		
7/17/2004	211.68	9/1/2004	209.57		
7/18/2004	211.84	9/2/2004	209.63		
7/19/2004	211.82	9/3/2004	209.75		
7/20/2004	211.82	9/4/2004	209.78		
7/21/2004	211.88	9/5/2004	209.73		
7/22/2004	211.83	9/6/2004	209.65		

APPENDIX K: Water level change (mm) of Johnny Cabin Pond between June 16th and September 8th 2004.

Date	Water Level Change	Date	Water Level Change	Date	Water Level Change
6/15/2004	3.3	8/3/2004	57.4	1/9/2004	-13.5
6/16/2004	8.2	8/4/2004	49.8	9/2/2004	-16.5
6/17/2004	6.0	7/16/2004	178.9	9/3/2004	-11.5
6/18/2004	3.0	7/17/2004	199.8	9/4/2004	-6.8
6/19/2004	6.3	7/18/2004	240.1	9/5/2004	-4.3
6/20/2004	71.0	7/19/2004	266.9	9/6/2004	-5.5
6/21/2004	123.4	7/20/2004	252.6	9/7/2004	-7.5
6/22/2004	135.7	7/21/2004	225.8	9/8/2004	-8.6
6/23/2004	137.4	7/22/2004	192.5		
6/24/2004	122.5	7/23/2004	163.5		
6/25/2004	104.4	7/24/2004	138.0		
6/26/2004	99.9	7/25/2004	119.6		
6/27/2004	93.1	7/26/2004	114.5		
6/28/2004	73.7	7/27/2004	110.9		
6/29/2004	77.7	7/28/2004	100.6		
6/30/2004	59.8	7/29/2004	92.7		
7/1/2004	50.4	7/30/2004	91.1		
7/2/2004	28.4	7/31/2004	76.2		
7/3/2004	37.7	8/1/2004	53.9		
7/4/2004	25.8	8/2/2004	44.1		
7/5/2004	24.5	8/3/2004	57.4		
7/6/2004	23.7	8/4/2004	49.8		
7/7/2004	24.2	8/5/2004	41.2		
7/8/2004	22.8	8/6/2004	33.8		
7/9/2004	22.2	8/7/2004	44.4		
7/10/2004	20.4	8/8/2004	19.8		
7/11/2004	17.6	8/9/2004	21.5		
7/12/2004	17.2	8/10/2004	27.5		
7/13/2004	13.8	8/11/2004	31.6		
7/14/2004	47.4	8/12/2004	19.5		
7/15/2004	148.1	8/13/2004	13.5		
7/16/2004	178.9	8/14/2004	4.0		
7/17/2004	199.8	8/15/2004	-12.0		
7/18/2004	240.1	8/16/2004	-23.3		
7/19/2004	266.9	8/17/2004	9.0		
7/20/2004	252.6	8/18/2004	14.9		
7/21/2004	225.8	8/19/2004	-1.7		
7/22/2004	192.5	8/20/2004	-1.7		
7/23/2004	163.5	8/21/2004	2.1		
7/24/2004	138.0	8/22/2004	-19.9		
7/25/2004	119.6	8/23/2004	-17.2		
7/26/2004	114.5	8/24/2004	-9.7		
7/27/2004	110.9	8/25/2004	-7.6		
7/28/2004	100.6	8/26/2004	9.5		
7/29/2004	92.7	8/27/2004	1.5		
7/30/2004	91.1	8/28/2004	-3.5		
7/31/2004	76.2	8/29/2004	-8.5		
8/1/2004	53.9	8/30/2004	-17.5		
8/2/2004	44.1	8/31/2004	-7.5		

APPENDIX L: Results and description of hydrology during spring 2003 sampling.

Samples were taken on April 30th and May 1st, 2003.
SM = Snowmelt-influenced, F=Flooded, TSS=Total Suspended Solids

Basin	Lake/River name	^{18}O	^{2}H	TSS Conc.en (mg/L)	Status	Observations
1	Devils Gate Pond	-13.8	-124	0.21	SM	completely ice-free, water fairly clear, brown stained, level seems normal, fringe fairly wet
2		-13.3	-121	0.16	SM	some residual ice, water clear, level not drawn down, flooding shore fringe
3		-15.1	-133	0.08	SM	ice-free, water fairly clear, not drawn down, seems high level with lots of water in willows to the west
4		-12.6	-119	0.07	SM	completely ice-free, water fairly clear, level high
5	Spruce Island Lake	-11.6	-115	0.7	SM	still 80% ice-covered, water very clear, water level low, organic smell
6	Pushup Lake	-10.5	-108	0.29	SM	~50% ice cover, water looks clear, level seems down, veg'n frozen into ice, lots of water around
7		-13.9	-126	0.31	SM	some residual ice, water clear, water level high, flooding shore fringe
8	Chilloneys Creek Pond	-15.9	-137	na	F	water level very high, some residual ice, water clear, meadow flooded and snow remaining
9		-15.2	-134	0.23	SM	completely ice-free, water clear, level seems high , fringe very wet, mud lake to N very turbid
11		-15.6	-136	0.59	SM	clear, open water, much higher level than previous visits- good snowmelt signature, no turbid water in the area
12		-13.7	-126	0.57	SM	water very clear, ice-covered, level looks a bit high on shore, very wet around lake,rocher seems to be flowing S
13		-17.6	-145	0.44	SM	heavily ice-covered-level seems very low, large exposed hummucky muddy area, water very clear, some brown stains
14		-14.5	-129	0.06	SM	ice-covered, water clear, slightly brown stained, level down, exposure of muddy hummucks
15		-18.7	-151	2.27	F	channel turbid all the way in, N.arm very murky- versus ice cover on moat, S arm much clearer
16	Egg Lake	-15.9	-137	1.65	SM	clear, lots of water in basin, fairly high water level, almost confluent with PAD 17
17		-12.3	-118	0.24	SM	very clear, scattered ice, water level seems high, sedges submerged,
18	Greenstar Lake	-10.5	-109	0.10	SM	heavily ice-covered, water level appears same as past visits, sample taken in moat
20		-19.3	-152	2.52	F	fully flooded area of meander scrolls from jamming on Emb R. and Athab, willows ~3m high
22		-19.3	-152	1.62	F	lake is murky-surrounding area very wet, turbid water throughout trees, thin ice cover
23		-13.0	-122	na	SM	partial ice-cover, moat developing, water level very high, water clear, surrounding channel murky on N side
24		-14.7	-128	0.22	SM	residual pond, surrounded by meadows near blanche L, thin ice-covered (new) and residual thick ice (old)
25	Blanche	-18.5	-147	0.50	F	partial ice-cover, water is clear, turbid water in area to north, may have flooded on may 1st via channel near PAD 26
26		-18.8	-149	1.53	F	very turbid, direct river water into the area
27		-12.2	-115	0.78	SM	somewhat murky, no apparent river influence level is drawn down
28		-12.7	-117	0.11	SM	mixed clear and murky, bottom of lake visible, some thin (new) and thick residual ice
29		-17.5	-141	0.80	F	turbid water, bottom not visible, level high, sedge still exposed, very broad flooded area
30	Mamawi Creek Pond	-19.1	-151	1.54	F	fully flooded, turbid, same level as mamawi creek, no obvious inflow channel

Observations were taken by Dr. T. Edwards, University of Waterloo, Ontario, Canada.

Continued on next page

Basin	Lake/River name	^{18}O	^{2}H	Inorg. Conc. (mg/L)	Status	Observations
31	Johnny Cabin Pond	-19.0	-151	3.47	F	lake is very high, at same level as mamawi ck, sampled distant from river inflow
32		-17.3	-143	0.74	F	lake level high, water is variably mixed with turbid overbank flow from the embarras
33	Dagmar Lake	-19.3	-152	4.5	F	very murky, surrounding area completely flooded, level is high on shore fringe, ice remaining in NW end
34		-19.2	-152	2.52	F	fully flooded area, broad willow meadows all with standing turbid water from river
35		-12.0	-116	0.05	SM	lake is clear, no river flooding, level is moderately high
36		-12.6	-121	0.62	SM	lake is clear, no river flooding, level is moderately high, frozen in macrophytes
38	Richardson Lake	-17.3	-141	0.6	F	partial ice-covered, water level fairly high, plumes of river water backflowing into the lake from distributary channels
39		-12.7	-119	0.14	SM	water is very clear, can see bottom, doesn't seem to be direct river influence, 50% residual ice-cover
40	Grey Wavy Lake	-19.0	-151	2.69	F	really murky, lots of inflow from small channel off fletcher, lake level mederately high, must be rising quickly
42		-19.0	-152	0.98	F	lake is still partly clear (remnant snowmelt or old lakewater), starting to become murky (mixing evident),
48		-21.2	-165	7.46	F	flooded from channel from peace, very turbid, forest flooded
49		-17.9	-151	0.25	SM	clear, water level fairly high,, visible bottom
50		-13.5	-125	0.55	SM	clear, no ice, water level fairly high
52		-13.9	-128	0.29	SM	clear, no ice, small lake in large area of sedge and willow meadow
53	Baril Lake	-18.1	-146	9.18	F	no ice cover, somewhat turbid, water seems high, likely runoff
54	Horseshoe Slough	-20.2	-157	4.04	F	lake connected to river on N side, water very turbid, 80% ice cover, water level very high,
55		-18.3	-154	0.23	SM	clear water, brown-stained, runoff apparent
56		-11.3	-115	na	SM	water level seems low, lots of open water, some residual ice, water clear, bottom visible
57		-12.2	-117	0.99	SM	large lake in cluster, clear water, level fairly high
rivers						
R1	Revillon Coupe Weir	-21.5	-166	18.84		S flow, weir flat, extremely turbid
R4	Athabasca River	-19.3	-153	7.32		local ice jam flooding, banks full
R5	Embarras River	-19.2	-151	6.71		extremely turbid, no visible flow, debris and ice-jammed up
R6	Fletcher Channel	-19.2	-152	4.42		river entirely clear of ice, was over flowing but is stopping
R7	Mamawi Creek	-19.2	-152	5.53		very active, log jams, flowing over banks
R8	Peace River	-21.1	-165	20.77		extremely turbid, bank full, ice push at shore
R9	Claire River	-21.1	-165	14.63		very turbid in moat, heavily ice-covered, levees 2-3m above water level, channel 10-15m wide
R10	Baril River	-21.3	-167	18.95		very murky, river full, 1-1.5m free of levees, broken up ice cover, lots of open water
R11	Riviére des Rochers	-21.0	-165	15.54		river fully covered with narrow moat, very turbid, level high, well below levees, still flooding of shore
R27	Chilloneys Creek	-17.4	-143	0.79		creek is ice-covered with moats along shore, water not as turbid as main rivers, open fully to PAD 8

APPENDIX M: Replenishment of lakes in spring 2005.

Lake	δ_{Lnew} $\delta^{18}O$	$\delta_{Lbefore}$ $\delta^{18}O$	$\delta_{source\ water}$ $\delta^{18}O$	Replenishment (%)	Dilution Process
PAD 1	-15.2	-9.6	-22.4	43.6	SM
PAD 2	-14.5	-9.1	-22.4	40.7	SM
PAD 3	-13.6	-9.7	-22.4	31.0	SM
PAD 4	-14.9	-8.6	-22.4	46.1	SM
PAD 5	-13.4	-8.4	-22.4	36.2	SM
PAD 6	-13.6	-8.6	-22.4	36.1	SM
PAD 7	-14.6	-9.6	-22.4	38.9	SM
PAD 8	-18.1	-14.2	-18.0	102.7	F
PAD 9	-17.1	-9.5	-22.4	59.0	SM
PAD 11	-15.4	-10.5	-22.4	41.0	SM
PAD 12	-13.3	-9.0	-22.4	32.3	SM
PAD 13	-15.9	-10.9	-22.4	43.5	SM
PAD 14	-12.5	-10.0	-22.4	19.7	SM
PAD 15	-16.1	-10.2	-18.0	76.5	F
PAD 16	-17.1	-13.4	-22.4	41.2	SM
PAD 17	-16.0	-9.3	-22.4	50.9	SM
PAD 20	-17.3	-9.3	-17.5	98.2	F
PAD 22	-17.4	-9.0	-17.5	99.3	F
PAD 23	-13.3	-9.1	-22.4	31.0	SM
PAD 24	-17.1	-9.7	-22.4	58.6	SM
PAD 25	-17.6	-12.8	-17.5	102.4	F
PAD 26	-17.3	-13.8	-17.5	96.0	F
PAD 27	-13.7	-9.5	-22.4	32.7	SM
PAD 28	-16.0	-9.3	-17.5	82.1	F
PAD 29	-15.9	-9.7	-22.4	49.1	SM
PAD 30	-16.9	-10.1	-17.5	92.0	F
PAD 31	-17.2	-15.0	-17.5	88.3	F
PAD 32	-16.9	-9.7	-22.4	56.5	SM
PAD 33	-17.1	-14.6	-17.5	86.0	F
PAD 34	-16.2	-12.5	-22.4	37.1	SM
PAD 35	-13.3	-8.9	-22.4	32.5	SM
PAD 39	-14.1	-8.9	-22.4	38.6	SM
PAD 40	-17.2	-11.8	-17.5	96.0	F
PAD 42	-15.8	-9.7	-17.5	78.1	F
PAD 48	-16.4	-13.7	-22.4	30.7	SM
PAD 49	-17.2	-11.5	-22.4	51.8	SM
PAD 50	-15.0	-9.8	-22.4	40.9	SM
PAD 52	-14.1	-9.4	-18.0	54.5	F
PAD 53	-15.6	-10.9	-18.0	67.0	F
PAD 54	-17.0	-13.5	-18.0	79.1	F
PAD 56	-13.1	-9.2	-22.4	29.6	SM
PAD 57	-14.3	-9.5	-22.4	37.4	SM