The hydrology and geochemistry of a saline spring fen peatland in the Athabasca Oil Sands Region of Alberta

by

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Author's Declaration

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.

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Abstract

Due to the nature of the regional geology and the bitumen extraction process, the post-mined landscape of Canada's oil sands region will have a much higher concentration of dissolved salts than it did prior to mining. As a result, naturally saline wetlands may constitute appropriate reclamation targets and knowledge of saline wetland hydrology can provide important clues to their form and function. Furthermore, the presence of saline discharge features in the Athabasca oil sands region (AOSR) provides an opportunity to study more closely the nature of groundwater flow in a region of considerable hydrogeologic complexity, including the origin and flow history of brines and the link between springs, subsurface wastewater containment and surface water quality.

A low-flow saline-spring fen peatland located adjacent to a proposed in-situ oil extraction facility was examined south of the oil sands hub of Fort McMurray, Alberta. Hydrologically disconnected from underlying Devonian deposits that are a typical source of salinity, a saline groundwater plume originating from a Lower Cretaceous aquifer (the Grand Rapids Formation) was identified as a likely source for the accumulation of Na+ (mean of 6,949 mg L^{-1}) and Cl^- (mean of 13,766 mg L^{-1}) in fen groundwater. Considerable spatial variability in ground and surface water salinity was observed, with the concentration of dissolved salts decreasing by an order of magnitude in the direction of flow. A sharp decrease in near-surface salinity was found along the entire perimeter between the fen and adjacent freshwater wetlands. Patterns in deep groundwater flux were difficult to interpret due to possible inaccuracies associated with the piezometer network (e.g., time-lag errors in low hydraulic conductivity substrates), and rates of groundwater input were estimated to be small (< 1 mm over a season) due to the low conductivity of the underlying mineral till (5.5×10^{-7} cm s^{-1}). Water table dynamics were exaggerated in response to wetting and drying for both study seasons and the fen's small subsurface storage capacity was readily exceeded under periods of sustained rainfall. The large pond network functioned as an effective transmitter of surface water during periods of high water table but was a sink of groundwater during dry periods due to high rates of evaporation. Despite flooding conditions observed in 2012, groundwater exchange between the fen and adjacent wetlands was low and the rough microtopography worked to detain surface waters and restrict runoff in the fen's lower reaches. Together these mechanisms worked to isolate the saline fen

and restrict the flux of saline waters into the surrounding landscape. Elevated concentrations of dissolved salts in nearby wetland and river systems indicates that influence of saline discharge is not solely restricted to the region's major river systems. The flux of salt from saline wetlands may play an important role in the overall water quality of groundwater and receiving water bodies (e.g., nearby river systems).

The geochemical signature of fen groundwater points to halite as a source of salinity, as indicated by Cl^{-}/Br^{-} ratios in excess of 7,000. This is in contrast to what has been observed for regional formation brines that are typically related to evaporated seawater. Isotopic evidence and relatively low salinities compared to springs in the Wood Buffalo region suggests that fen discharge water may be significantly diluted as a result of mixing with freshwater sources. The contribution of evaporite to discharge water may be coming from somewhere deeper and further south in the basin. This has important implications for the disposal of wastewater by deep well injection, as disposal zones may be hydrologically linked to near-surface aquifers and discharge features well beyond the immediate production and storage area.

Acknowledgements and Dedication

It all started with a simple email in my 3rd year of undergraduate studies here at the University of Waterloo. It was from Dr. Price, my GEO 303 professor at the time. After the initial alarm of receiving personal correspondence from my professor had passed, I read the email to see that it was for a summer co-op opportunity helping graduate students with their fieldwork out in eastern Quebec. Fast forward four years later, and I'm completing my Masters research and working full-time for Jonathan. I often reflect upon that email, and wonder how different my life would be if I had never received it. Jon, although it has likely never crossed your mind, that email had a profound impact on my life. That opportunity, along with the countless others you've given me since that time have helped to completely reshape my career path and my future. Thank you.

A lot of fieldwork was required to get to where I am today and I couldn't have done it without the support and guidance of countless individuals. I'd like to extend a special thanks to Jonathan Goetz, Roxanne Andersen and Scott Ketcheson for their invaluable assistance and critical eye during that first summer in Fort McMurray when we were literally thrown into the unknown. Thanks to Pete Macleod for all the laughs and post-fieldwork rap shows. The help of Sarah Scarlett and Emma Bocking was immeasurable. Thanks to Dr. Duke at the University of Alberta's SLOWPOKE Nuclear Reactor Facility for his support with bromide analyses.

It almost goes without saying that I wouldn't be where I am today if it wasn't for the unwavering support of the Wells Clan. To Jared, Shannon, Elizabeth and Michael, thank you for pushing me and supporting me through the easy and not so easy moments of my seemingly endless academic career.

Lastly, thanks to Stephanie Cullen, my Fiancée and partner in crime, who has been here throughout the entire ordeal since the Quebec days. The fact that you stuck around even when I left you every summer for fieldwork is unbelievable and I'm forever thankful.

This thesis is dedicated to you, Stephanie.

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1.0 Background

1.1 The State of Development and Reclamation in The Athabasca Oil Sands and Reclamation

Alberta's Athabasca oil sands region (AOSR) is home to the largest bitumen deposit in Canada, covering roughly 23% of the entire provincial landmass (Johnson and Miyanishi, 2008). The deposits, which are considered a safe and secure source of oil for North America, are the second largest in the world and contain an estimated 173.2 billion barrels of recoverable oil (Government of Alberta, 2008). Today, approximately 1.6 million barrels of oil per day are produced through a combination of surface mining and in-situ extraction techniques, with production expected to double within the present decade (ERCB, 2011). Open pit mining operations, which involve the stripping and stockpiling of overburden in order to access the near-surface bitumen, are expected to affect roughly 2,000 km^2 of the Boreal landscape once fully operational (Woynillowicz et al., 2005). The sheer magnitude of development within Canada's oil sands will lead to significant changes in its landscape structure and will require reclamation at the ecosystem scale (Halsey, 2007; Johnson and Miyanishi, 2008). Due to the nature of the regional geology and the bitumen extraction process, the post-mined landscape will have a much higher concentration of dissolved salts than it did prior to mining. Marine deposits and deep saline aquifers become exposed and can leach dissolved salts to the surface while processed ore, recycled tailings water and stockpiles of marine shale overburden containing elevated ionic content have accumulated within the reconstructed landscape (Purdy et al., 2005; Trites and Bayley, 2009). The presence of these ions in elevated concentrations may pose a serious challenge for wetland reclamation due to the adverse effects of salts on endemic wetland species (Renault et al., 1998; Apostol et al., 2004; Purdy et al., 2005; Pouliot et al., 2012). Saline wetlands that grow spontaneously or are built on the post-mined landscape will not support most plant species typical of fens and bogs in northern Alberta. Thus, naturally saline wetlands represent potential reclamation targets for vegetation establishment (Trites and Bayley, 2009).

In-situ Extraction and Saline Springs in the AOSR

With the majority of the recoverable bitumen too deep for traditional surface mining techniques, it is expected that up to 80% of the proven bitumen reserve will be extracted using in-situ technologies (Jordaan, 2012). In-situ oil recovery was originally conceived for high-viscosity bitumen deposits that made conventional production methods impractical (Butler, 1994). While several methods of in-situ extraction exist, the general procedure involves the introduction of heat to a reservoir to reduce the viscosity of the semi-solid bitumen. Upon heating, the introduction of high pressures or the use of gravity aids in the evacuation of the bitumen through horizontal well pairs that bring it to the surface (Bachu et al., 1989; Butler, 2001). The application of in-situ technologies has increased rapidly in the AOSR, with at least four large-scale projects currently in operation and over 19 small scale or pilot projects in development since 2000 (Butler, 2001; Gordon et al., 2002). Following recovery of the bitumen, production water is treated at the surface and the remaining wastewaters are once again injected into the subsurface where they are stored in deep aquifers (Bachu et al., 1989). In the Fort McMurray region, subsurface wastewater injection has many challenges and is potentially hazardous to both industrial and domestic groundwater supplies.

Owing to the rapid expansion of in-situ projects, considerable industry attention has been given to the subject of wastewater disposal by deep-well injection (Gordon et al., 2002). The geology and hydrogeologic setting of the AOSR has created unique challenges for the successful containment of wastewater. Due to the relatively shallow depth of the regional stratigraphic package and the limited number of overlying and laterally confining layers isolating potential disposal zones from the surface, the potential for leakage and horizontal and upward migration of wastewater is large (Hackbarth and Nastasa, 1979; Bachu et al., 1989; Gordon et al., 2002). Carrigy and McLaw (1973) noted that, as a result of high injection pressures and the resulting increase in porosity and permeability in reservoirs post-extraction, groundwater flow rates and patterns can be impacted well beyond the immediate production area. In addition, the presence of structural complexities related to Paleozoic salt bed erosion, the outcropping of potential disposal zones at or near the surface and the presence of saline springs connected to deep aquifers intensify the risk for wastewater discharge (Gordon et al., 2002). Extra pressures may augment groundwater flow and establish new connections with deeper flow systems causing higher flow rates from springs as well as increasing the proportion of salt in discharge waters (Carrigy and

McLaw, 1973; Hackbarth and Nastasa, 1979).

As development expands, the likelihood of accidental process-affected wastewater discharge to the environment increases along with the possibility of freshwater contamination and subsequent animal toxicity and mortality. As a result, operators are held to a strict zero discharge policy that prohibits the release into the environment of a substance in an amount, concentration or level or at a rate of release that causes or may cause a significant adverse effect (Government of Alberta, 2000). In the AOSR, subsurface wastewater containment is considered a safe and viable disposal option only when the formation can be deemed suitable for maintaining the ongoing confinement of the disposal fluid (ERCB, 1994), a prerequisite that poses a challenge for developers in many parts of the region.

1.2 Geology and Fluid Flow in the AOSR

Regional Geologic Setting

The AOSR is located on the northeastern margin of the Alberta Basin, a sub-basin of the Western Canadian Sedimentary Basin that is bordered to the west by the Rocky Mountain Thrust Belt, to the northeast by the Precambrian Shield and to the southeast by the Bow Island Arch (Grasby and Chen, 2005; Connolly et al., 1990a) (Figure 1-1). The basin comprises a simple sedimentary wedge that rests unconformably on buried Precambrian rocks of the Canadian Shield. To the west, the sedimentary package exceeds 5,700 m but thins to a northeastern zero-edge where the Precambrian Shield becomes exposed as a result of depositional thinning and erosion (Connolly et al., 1990a).

The simplified modern-day hydrodynamic regime of the Alberta Basin is predominantly south to north through Paleozoic era Cambrian sandstones and Devonian through Mississippian carbonates. Thick sequences of shale and silt units form regional aquitards through most of the overlying Mesozoic strata; however, interbedded sandstones form local and in some cases regional scale aquifers (Grasby et al., 2006). Throughout most of the Alberta basin, local and regional scale flow processes are modified by highly permeable Upper Devonian and Carboniferous carbonate rocks that act as a low-fluid potential drain, channeling flow from the entire basin northward where it discharges in the AOSR (Hitchon, 1969).

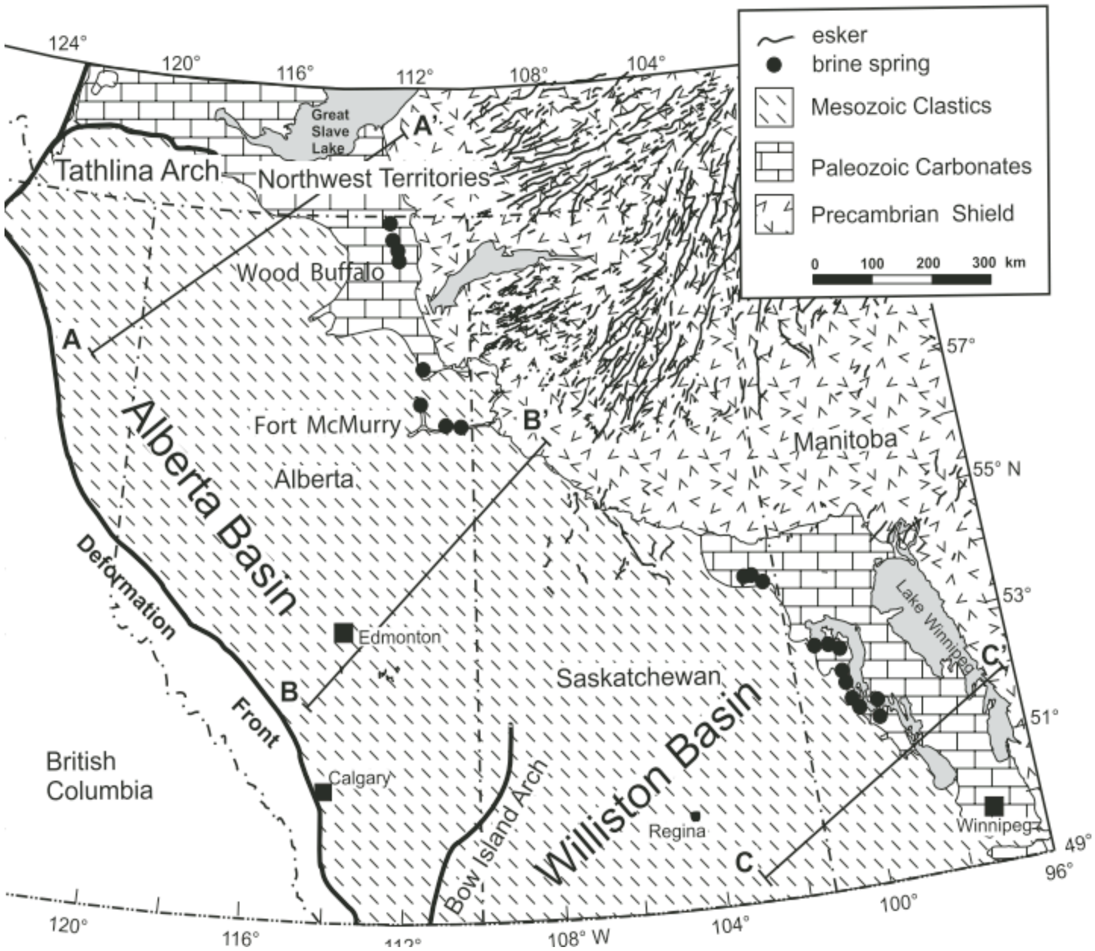

Figure 1-1 Western Canadian Sedimentary Basin showing location of Alberta sub-Basin. Brine spring locations are marked on the map as dark circles (Grasby and Chen, 2005).

Regional Scale Hydrogeology in Northeastern Alberta

A synthesis of aquifer salinity and regional, intermediate and local scale flow regimes for northeastern Alberta was conducted by Bachu et al. (1993). The following review provides an abbreviated version of their work on the hydrogeologic regime of the AOSR. The regional geology and flow regime is exceedingly complex and so for simplicity only the most relevant stratigraphic units and flow systems are highlighted. The reader is directed to the generalized stratigraphic chart in Figure 3-3 for reference on the various formations, their positions and nomenclature. A hydrostratigraphic dip cross-section of the various flow systems discussed below can be found in Figure 1-2.

Located at the northeastern edge of the Alberta basin, the AOSR has undergone considerable postglacial erosion, exposing deep Paleozoic strata to atmospheric conditions. In general, formation waters that follow the basin-wide trend begin to show a modification in their flow paths in the AOSR as a result of atmospheric exposure and the influence of topographic and physiographic features. Regional scale fluid flow is in a northeast direction, however, examination of regional hydrogeologic cross-sections, hydraulic head data and aquifer salinity distributions indicate that there are few Paleozoic aquifers that follow the true regional flow regime. The Contact Rapids-Winnipegosis system is a regionally trending high-salinity aquifer that exhibits minimal topographic influence in and around the oil sands hub of Fort McMurray. Salinities are exceptionally high, exceeding 350,000 mg L^{-1} at some locations due to the adjacent Elk Point group evaporitic unit that provides a local source of soluble ions. Above the Prairie Evaporite unit, which acts as an extensive regional aquitard, Paleozoic aquifers demonstrate intermediate flow characteristics and are under stronger topographic influence. Most relevant to the study area is the intermediate to local scale flow regime of the Beaverhill Lake-Cooking Lake aquifer system, a lower salinity (~160,000 mg L^{-1}) formation that is influenced by present-day topography. Flow is generally to the northeast, however, aquifer recharge and dilution occurs along topographic highs (i.e., Birch Mountains) and discharge has been observed where the formation subcrops along the Athabasca and Clearwater rivers.

Unlike Paleozoic flow systems in the AOSR, Cretaceous aquifers show strong local flow characteristics and correspondingly low salinities caused by the influx of meteoric waters. The McMurray-Wabiskaw system (bitumen bearing unit) is an important, regionally continuous aquifer system that is entirely under local control. High-potential zones can be found across the

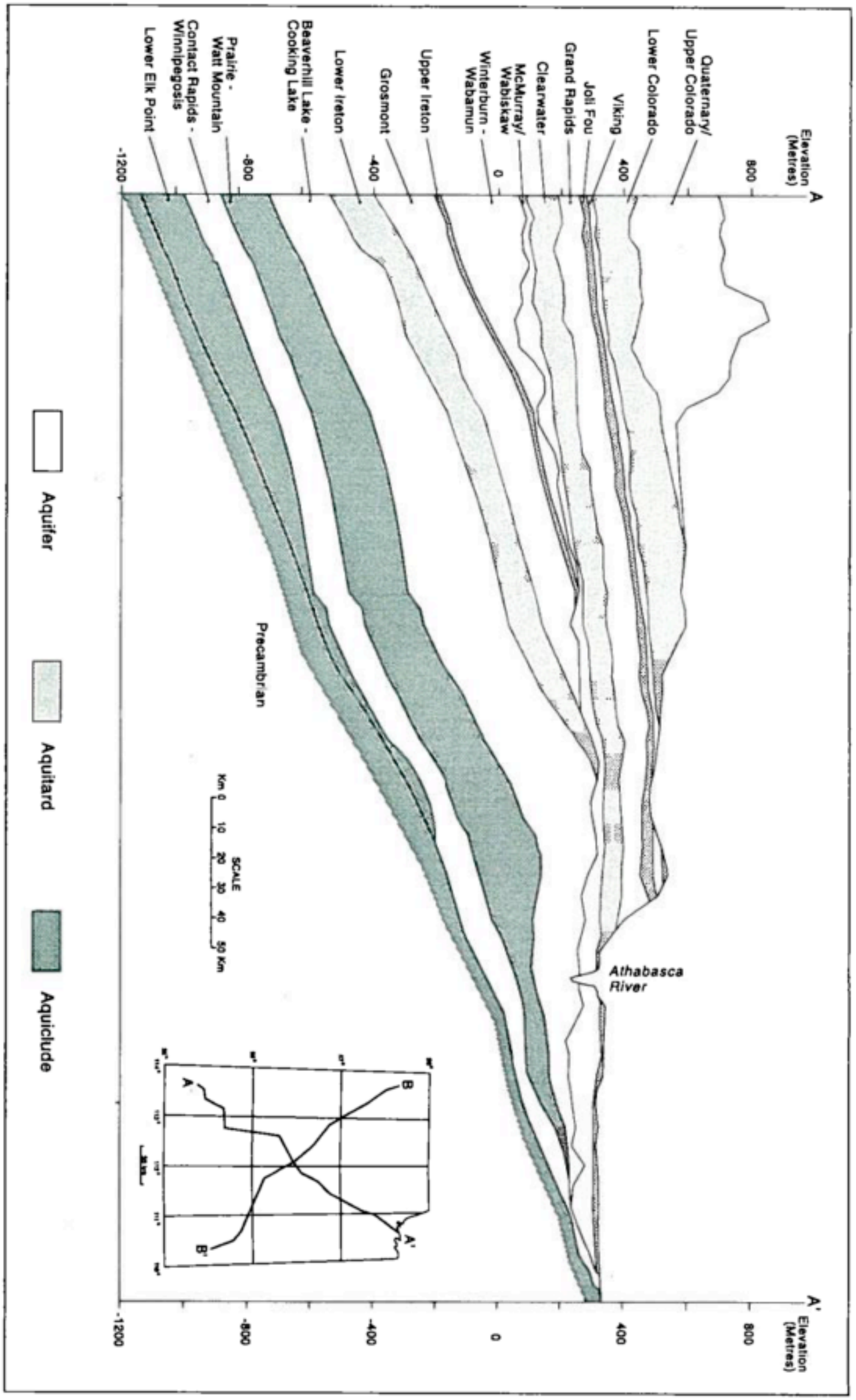

Figure 1-2 Hydrostratigraphic dip cross-section of the study region (Bachu et al., 1993)

Birch and Stony Mountain ranges, indicating groundwater recharge that is confirmed by very low salinities in the range of 5 to 10,000 mg L^{-1} near the city of Fort McMurray. Similar hydraulic head distributions and lower salinities are observed in the near-surface Grand Rapids aquifer, with discharge observed along low lying river systems and plains regions. The McMurray-Wabiskaw and the Grand Rapids aquifer are separated by the shale-dominated Clearwater aquitard that prevents cross-formational flow throughout most of the AOSR. However, silt and fine sand lenses within the Clearwater aquitard allow for hydraulic communication across large areas and may provide a channel for the discharge of high salinity groundwater to the surface. The discharge of saline groundwater is observed across the Fort McMurray and Wood Buffalo regions in connection to evaporite (salt) bearing Devonian outcrop belts (Hitchon et al., 1969; Grasby et al., 2006).

Evaporite Deposits

During the Early and Middle Devonian the Western Canadian Sedimentary Basin was the site of extensive evaporite deposition. Evaporite deposits are water-soluble sediments that contain an array of minerals including anhydrite, dolomite, limestone, shale, chloride, sulfate, potash and halite (sodium chloride) and are named in reference to their mode of deposition through the concentration of saline water by evaporation (Hamilton, 1971).

Depositional History

According to Hamilton (1971), the deposition of all evaporite deposits originate through the concentration and precipitation of seawater in embayments that become isolated from the ocean. Often, a continuous or intermittent influx of seawater occurs across an elevated sea-floor barrier and, as evaporation continues, the concentration of dissolved salts increases to form dense brines that sink to the bottom of the isolated basin. Over time, salts begin to precipitate out of solution as the basin becomes gradually more saline.

The thick succession of evaporitic deposits between Pre and Upper Devonian rocks underlying the AOSR is collectively known as the Elk Point Formation (Hamilton, 1971). The deposition of the Elk Point group began in early Middle Devonian time as the ancient Boreal Ocean encroached southeastward onto the emergent Interior Plains and developed as successive layers of salt over several wetting and drying cycles of evaporative enrichment (Hamilton, 1971). The last of these evaporitic cycles developed during a time of expanded seas and intense biologic activity, producing large biohermal reef systems that eventually cut off flow into an adjacent embayment. As brines underwent progressive concentration, they eventually produced the largest deposit of evaporite in the Elk Point group, the Prairie Evaporite halite/anhydrite deposit (Hamilton 1971; Meijer Drees, 1994).

Regional Stratigraphy of the Elk Point Group

The Elk Point group is a cyclic succession of evaporite, carbonate and clastic rock that contains an estimated total rock salt volume of $6x10^4$ km^3, the largest concentration of salt in the entire Western Canadian Sedimentary Basin (Grobe, 2000). The Elk Point group is divisible into distinct upper and lower subgroups. Restricted largely to east-central and northern Alberta, the Lower Elk Point depocenter is in east-central Alberta where its thickness reaches ~350 m (Hamilton, 1971). Five distinct salt beds are found in the Lower Elk point division and consist primarily of coarse crystalline halite that grade into red shale inclusions near their depositional edges (Grobe, 2000). The majority of the Lower Elk Point group has undergone subsurface dissolution resulting in thinning of its margins, most predominantly along the deposit's eastern extent where the Devonian formation crops out at the surface (Grobe, 2000).

The Upper Elk Point subgroup is more widespread and extends fromimmediately south of the Wood Buffalo region southeast through Saskatchewan and Manitoba. The Prairie Evaporite formation is the only salt unit within the subgroup and is composed primarily of halite but contains a large proportion of impurities, including shales, anhydrites and dolomite (Hamilton, 1971). Regionally, the deposit is thickest in northwestern Alberta, with the bulk of the salt found in east-central Alberta. Along its eastern edge the deposit terminates sharply as a result of the salt dissolution processes (Hamilton, 1971; Grobe, 2000). Evidence of groundwater discharge influenced by evaporite includes numerous saline springs and anomalously high sodium chloride levels in surface waters throughout the AOSR (Hitchon et al., 1969; Grasby et al., 2006; Jasechko et al.,, 2012).

Geochemical Signature of Formation Brines and Saline Springs

In the AOSR, saline springs are primarily found where groundwater discharges through subcropping or exposed Paleozoic carbonate rocks in the low-lying plains and the valleys of the Athabasca and Clearwater rivers (Grasby et al., 2006). These types of springs are karstic in origin due to the soluble nature of the carbonate rocks and the extensive dissolution that has occurred throughout evaporite deposits which lie close to the surface (Borneuf, 1983; Ford, 1998). Sodium chloride springs make up 4% of the springs in Alberta and are found exclusively in the northeastern region of the province (Borneuf, 1983), where the compositions of brines represent a mixture of geochemical end members related to seawater, evaporite dissolution and freshwater mixing (Rittenhouse, 1967; Connolly et al., 1990a; Michael et al., 2003; Gupta et al., 2012). Connolly et al., (1990a), in their study of the origin and evolution of formation waters in the Alberta basin, found that Devonian through Cretaceous reservoirs were all derived from subaerially evaporated brines and not influenced by evaporite dissolution. Similar interpretations were made by Lemay (2002) around the Fort McMurray area that showed Lower Cretaceous aquifers having a diluted seawater composition with no indication of evaporite. In a study by Michael et al., (2003), geochemical evidence pointed to the dissolution of halite as a source for salinity in deep basinal brines. Within the Fort McMurray and Wood Buffalo region, isotopic evidence and major ion composition of both high-and low-flow springs indicated that discharge water was not related to evaporated brine but was of meteoric origin, with elevated salinity a result of contact with buried evaporite beds, namely halite (Grasby, 2000; Last and Ginn, 2005; Grasby et al., 2006; Grasby and Londry, 2007; Berard et al., 2013).

The physical characteristics of springs in the AOSR are variable and largely a function of salinity and rate of discharge (Grasby and Londry, 2007). High-salinity springs (>100,000 mg L^{-1} total dissolved solids) are often characterized as mud flats with considerable halite development that surrounds a central domal outlet. Often devoid of typical Boreal vegetation, salt tolerant plants grow sparsely in mudflats that then grade sharply into grasslands and Boreal forest (Grasby et al., 2006; Grasby and Londry, 2007). In contrast, low-salinity springs are often found within large marshy regions where discharge flows through a series of poorly defined pools. In many cases, cyanobacterial mats grow around spring outlets and outflow channels, desiccating towards the edge of pools to form characteristic “brain” textured surfaces (Grasby and Londry, 2007). The rate of groundwater discharge varies both between springs and over time. Historical records have shown considerable fluctuation in discharge rates over the last 50 years, with many springs becoming inactive for extended periods (Grasby and Londry, 2007).

1.3 Peatlands

Abundance and Peatland Type

The boreal plains landscape of Alberta is a rich tapestry of landforms driven by gentle, rolling topographies overlying deep and heterogeneous geologic deposits under a prevailing sub-humid climate. A combination of these factors works to create an assemblage of wetland and upland sequences within which wetlands comprise over 40% of the landscape (Kuhry et al., 1993). Within the AOSR, wetlands comprise up to 60% of the landscape, the majority of which are peatlands (95%) dominated by fens (Suncor Energy, 2005; Johnson and Miyanishi, 2008). Compared to bog peatlands, where precipitation makes up the primary hydrologic input (National Wetlands Working Group, 1997), fens are defined as minerotrophic and receive inputs of water from both precipitation and surrounding and/or underlying mineral soils (Vitt and Chee, 1990). A diverse assemblage of fen peatland types, defined by both hydrologic and ecological characteristics, is found in the AOSR.

Fen peatlands exist in a variety of forms based on landscape position that influences hydrology as well as peatland biogeochemistry. The fen gradient is composed of three classes of fen peatlands, each defined by hydrologic, chemical and vegetative characteristics. Poor fens more closely resemble bog environments and are often acidic ($pH < 6.0$), low in base cations and distinguished by *Sphagnum* moss carpets and ericaceous shrub cover (Vitt, 2006). Opposite poor fens are extreme-rich fens fed by substrates rich in base cations. Low hydrogen ion concentrations create alkaline waters ($pH > 7.0$) that, in combination with high mineral content, allow for sedges and brown moss species to proliferate (Vitt, 2006). Moderate to moderate-rich fens with intermediate physiochemical attributes form the mid-point of the Alberta fen gradient. While the AOSR is dominated by freshwater peatlands, saline fens have been documented throughout the Fort McMurray and Wood Buffalo areas in connection to direct or indirect discharging of saline groundwater (Trites and Bayley, 2009). Groundwater composition in these systems is often slightly acidic to basic (pH of 6.6 to 10.3) and dominated by $NaCl$ and $NaSO_4$ and to a lesser extent Mg, Ca, and HCO_3. Due to elevated salinities and pH, species richness is often low and dominated by salt tolerant vegetation, with typical peat-forming mosses such as *Sphagnum,* absent (Trites and Bayley, 2009). In many cases, distinct discharge mounds surrounded by bacterial mats and evaporitic minerals can be found that flow outward through a series of poorly defined pools and channels (Grasby 2006; Grasby and Londry, 2007).

Controls on Water Flow in the WBP

Peatlands rely on a constant and long-term supply of water (Vitt, 2006) that is constrained by a set of dominant hydrologic controls in the AOSR. Excess water is generally minimal due to a sub-humid climate where annual potential evapotranspiration (PET) is greater than precipitation. In addition, excess water that is available interacts with deep glacial deposits with large storage potential. Thus, the hydrologic function of peatlands, i.e., the ability of peatlands to adapt and sustain themselves through extended periods of drought and periodic wet cycles, is controlled by the interaction of climate and geology (Devito et al., 2012).

Climate

Precipitation and evapotranspiration (ET) are major components in a wetland water balance and play a critical role in influencing the form and function of wetland environments (Devito and Mendoza, 2007). Due to its subhumid climate, the AOSR is in a long term water deficit (annual average PET:P = ~520 mm:480 mm) and peatland maintenance relies heavily on seasonal timing of precipitation as well as decadal and multi-decadal precipitation cycles (Devito et al., 2012). Annual precipitation is low compared to other parts of the province and averages ~450 mm a year near the oil sands hub of Fort McMurray (Environment Canada, 2012). Fall and winter seasons are typically dry, with fall and spring rains and snow accumulation accounting for less than 35% of the annual precipitation (Johnson and Miyanishi, 2008; Environment Canada, 2012). The majority of the annual precipitation (65-75%) falls during the summer growing season (May-September), when evapotranspiration rates are at their peak, in the form of variable convective-cell storms of short duration where daily rainfall rates are typically less than 10 mm (Smerdon et al., 2005; Devito et al., 2012). The synchronization of peak rainfall with maximum PET exacerbates the seasonal water deficit and limits the opportunity for overland flow (Devito et al., 2005a; Devito and Mendoza, 2007; Redding and Devito, 2005).

While extended periods of drought are typical in the AOSR, the timing and variability of annual rainfall and the cycling of infrequent wet years help sustain peatlands and regulate soil moisture deficits (Petrone et al., 2007). Despite the low annual precipitation input of less than 35%, the dry fall and winter non-growing season (October through April) represents an important hydrologic input for the year and helps regulate the annual water deficit due to low ET. Accumulated non-growing season precipitation (fall rains, snow, spring rain) can result in an average net surplus of approximately 100 mm of water for the landscape (Devito et al., 2012). Over the long term, large variations in annual precipitation (up to 400 mm between years) and minimal change in annual PET (~30 mm between years) can create exceptionally wet seasons that typically occur every 10 to 15 years (Petrone et al., 2007; Devito et al., 2012).

Geology

The hydrologic function of wetlands is controlled by their ability to store water effectively and/or receive ground and surface water from their surroundings. Collectively, these mechanisms are driven by the wetland's position in the landscape, or the wetland's geologic setting (Devito and Mendoza, 2007). In the AOSR, extensive erosion caused by Late Tertiary and recent glacial and fluvial processes have shaped a low-relief topography underlain by generally deep, unconsolidated sediments with variable storage and transmission properties (Andriashek, 2003; Devito et al., 2005b). Surficial deposits overlie sedimentary bedrock formations that are relatively permeable and composed primarily of carbonate minerals (Johnson and Miyanishi, 2008). This carbonate-rich sedimentary bedrock enriches groundwater with calcium, magnesium and carbonate cations, creating alkaline environments that foster the development of dense fen peatland complexes characterized by brown moss and sedge species (Vitt, 2006).

Unlike the geologic architecture of the Precambrian shield, where thin surficial deposits overlie impermeable crystalline bedrock, the deep, heterogeneous and low relief topography of the AOSR generates a wide range of complex flow systems that influence chemistry, recharge/discharge capacities and wetland permanence (Ferone and Devito, 2004; Devito et al., 2005a; Devito et al., 2005b). In areas of high relief, wetlands exhibit seasonally dynamic hydrologic conditions and function primarily as groundwater recharge systems that are driven by local flow systems influenced by adjacent topographic highs (Toth, 1963; Ferone and Devito, 2004). Conversely, wetlands in low-relief environments are often connected to large scale intermediate or regional groundwater flow systems that damp seasonal variability and drive the upward flow of water from underlying soils towards the land surface (Siegel, 1988; Ferone and Devito, 2004). The scale of these flow systems, with elevated concentrations of solutes such as salts found in groundwater connected to more regional hydrogeologic conditions, often dictates the chemistry of peatlands.

1.4 Thesis Framework and Project Role

The primary objective of this research is to investigate the hydrology of a saline spring wetland within the subhumid Boreal climate. As such, the two manuscripts that comprise this thesis examine a saline wetland from two different but connected scales. The first manuscript investigates the hydrologic function (i.e., the movement and storage of water and solutes) of a saline fen peatland at the local (site) scale. This is achieved through a detailed assessment of its individual water balance components as well as geochemistry. Patterns in fen ground and surface water geochemistry presented in the first manuscript provide the foundation for manuscript 2, which examines the connection of the saline spring fen and surrounding study area to regional groundwater flow systems. A detailed analysis of major ion composition, isotopic signatures and deep groundwater function provide clues to the origin and flow history of saline waters that dominate the fen.

Conception, planning, and implementation of the study were done with guidance from Dr. Price. Colleagues assisted the author with field data collection, while data analyses and writing of the manuscripts was done by the author. Dr. Price provided valuable feedback throughout the revision process.

2.0 Manuscript 1: Hydrologic function of a saline fen in the Athabasca oil sands region

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2.1 Overview

Due to the nature of the regional geology and the bitumen extraction process, the post-mined landscape of Canada's oil sands region will have a much higher concentration of dissolved salts than it did prior to mining. As a result, naturally saline wetlands may constitute appropriate reclamation targets and knowledge of saline wetland hydrology can provide important clues to their form and function. The overall objective of this study was to investigate the site-scale hydrologic and geochemical function of a saline fen in the sub-humid Boreal region of northeastern Alberta.

Ground and surface water salinities were dominated by Na^+ (mean of 6,949 mg L^{-1}) and Cl^- (mean of 13,766 mg L^{-1}) and to a lesser extent $SO^{2}{}_{4}^{-}$ (mean of 728 mg L^{-1}) and the concentration of dissolved salts decreased by an order of magnitude in the direction of flow, between the south and north sections of the fen. While the vertical groundwater component was difficult to interpret due to possible inaccuracies associated with the piezometer network (e.g., lag-time errors in low hydraulic conductivity substrates), the overall model of focused saline groundwater discharge in the south fen and the redistribution of saline groundwater northward is supported by the spatial patterns in ground and surface water geochemistry. For both seasons, water table fluctuations were exaggerated in response to wetting and drying and the fen's small subsurface storage capacity was readily exceeded under periods of sustained rainfall. The large pond network functioned as an effective transmitter of surface water during periods of high water table but was a sink of groundwater during dry periods. Despite flooding conditions, the rough microtopography and low rates of subsurface groundwater exchange between wetlands worked to detain surface waters and restrict runoff in the fen's lower reaches. The flux of groundwater between the fen and adjacent wetlands was spatially variable and flow reversals occurred often in response to seasonal and short-term weather changes. Groundwater mounds were important mechanisms that restricted the exchange of groundwater between wetlands and the actual flux was a minor component of the fen's water balance.

This study highlights the unique hydrologic and geochemical function of saline wetlands in the Boreal Plains region of Alberta. Currently, the form and function of saline wetlands is poorly understood in the oil sands region and developing an improved understanding of these systems will aid in the management of an important ecological resource while supporting compliance with environmental policies related to oil sands extraction and reclamation.

2.2 Introduction

Open-pit mining, which involves the stripping and stockpiling of overburden to access the near-surface bitumen, is expected to affect roughly 2000 km^2 of the Boreal landscape within the Athabasca oil sands region once fully operational (Woynillowicz et al., 2005). Wetlands are the dominant landscape unit within the region and the sheer magnitude of development will lead to significant changes in its landscape structure and will require reclamation at the ecosystem scale (Johnson and Miyanishi, 2008). Due to the nature of the regional geology and the bitumen extraction process, the post-mined landscape will have a much higher concentration of dissolved salts than it did prior to mining. Marine deposits and deep saline aquifers become exposed and can leach dissolved salts to the surface. In addition, processed ore, recycled tailings water and stockpiles of marine shale overburden containing elevated ionic content comprise a major component of the reconstructed landscape (Purdy et al., 2005; Trites and Bayley, 2009). The presence of these ions in elevated concentrations may pose a serious challenge for wetland reclamation due to the adverse effects of salts on endemic wetland species (Renault et al., 1998; Apostol et al., 2004; Purdy et al., 2005; Pouliot et al., 2012). Consequently, saline wetlands that grow spontaneously or are built on the post-mined landscape will not support vegetation typical of “freshwater” fens and bogs in northern Alberta. With oil sands production expected to double in the present decade (ERCB, 2011), wetland reclamation strategies will require a sound understanding of how salinity influences the form and function of wetlands in the post-mined landscape.

Within the Athabasca oil sands region (AOSR) of Alberta, wetlands comprise up to 65% of the landscape, the majority of which are fen peatlands that rely on hydrologic inputs from precipitation as well as surrounding and underlying mineral soils (Devito et al., 2012; Andriashek, 2003). Extensive peatland complexes function as important water conservation and redistribution mechanisms in the subhumid Boreal climate and comprise a massive pool of terrestrial carbon for Canada (Gorham, 1991; Devito et al., 2012). While treed and open freshwater fens make up the primary peatland classes within the AOSR, rare saline wetlands can be found throughout the region’s low-lying plains and river systems. Naturally occurring saline wetlands within the AOSR may serve as appropriate reference analogues for oil sands reclamation (Grasby and Londry, 2007; Trites and Bayley, 2009).

Naturally saline wetlands exist worldwide under a variety of hydrologic settings, the most common of which are coastal wetlands that receive inputs of dissolved salts through tidal action (Price and Woo, 1988; Mitsch and Gosselink, 2000). Saline marshes and peatlands are also common throughout North America's interior, such as within the Great Plains region, where salts accumulate over many years within regional and local discharge zones (Harvey et al., 2007; van der Kamp and Hayashi, 2008; Heagle et al., 2013). The clay-rich glacial till of the prairie region provides an abundant source of sulphate salts derived mainly from the oxidation of pyrite (Heagle et al., 2013). In northern Alberta, saline spring wetlands are primarily found where regional groundwater discharges to the surface through subcropping or exposed Paleozoic rocks (Hitchon et al., 1969; Timoney and Lee and Lee, 2001; Grasby et al., 2006). Near-surface evaporite deposits provide a source of dissolved salts conveyed to the surface by springs situated close to or along the banks of the Athabasca and Clearwater rivers, where erosion has deeply incised into the exposed carbonate sediments (Ozoray et al., 1980; Grasby et al., 2006; Jasechko et al., et al., 2013). The groundwater composition is often slightly acidic to basic (pH of 6.6 to 10.3) and dominated by Na^+ and Cl^- and $NaSO_4^-$ and to a lesser extent Mg^{2+}, Ca^{2+}, and HCO_3^- (Trites and Bayley, 2009). In many cases, distinct discharge mounds surrounded by bacterial mats and evaporitic minerals can be found that flow outward through poorly defined pools and channels (Grasby 2006; Grasby and Londry, 2007). Due to elevated salinities and pH, species richness is often low and dominated by salt-tolerant vegetation, with typical peat-forming mosses such as *Sphagnum* absent (Timoney and Lee and Lee, 2001; Trites and Bayley, 2009). In many cases these saline wetlands provide habitat for a variety of vegetation communities rare to northern Alberta, including samphire (*Salicornia europaea*), narrow reed grass (*Calamagrostis stricta*), seaside arrow grass (*Triglochin maritime*), Nuttal's salt meadow grass (*Puccinellia nutalliana*), and alkali marsh aster (*Aster pauciflorus*) (Timoney and Lee, 2001; Allen, 2012). In addition to rare plant communities, northern Alberta saline wetlands serve as important habitats and breeding sites for moose, the endangered Peregrine Falcon and other waterfowl (Sweetgrass Consultants Ltd., 1997; Timoney and Lee, 2001).

Saline spring wetlands have been observed throughout northeastern Alberta, with the study of the hydrogeologic connection of springs (Grasby and Chen, 2005; Grasby et al., 2006; Grasby and Londry, 2007) and their influence on nearby rivers well documented (Hitchon, 1969a; Jasechko et al.,, 2012). From an oil sands reclamation perspective, naturally saline wetlands have been the focus of a number of studies looking into the relationship between vegetation communities and their abiotic conditions both within natural and reclaimed settings (Timoney and Lee, 2001; Trites and Bayley, 2009; Purdy et al., 2005). Yet, despite the fact that hydrology is a fundamental control on wetland biogeochemistry, little is known about the hydrologic function of these saline systems at the site scale (Scarlett and Price, 2013). Knowledge of saline wetland hydrology can provide important understanding of wetland form and function, including wetland ecology. Furthermore, while saline wetlands are a relatively rare occurrence in the AOSR, generating hydrological data on a variety of peatland types provides a better understanding of how natural systems operate in a region under developmental stress, and contributes to the improved management of these critical ecosystems. With this in mind, the overall objective of this study was to investigate the hydrologic and geochemical function of a saline spring fen situated within a low lying till plain in the sub-humid Boreal region of northeastern Alberta.

2.3 Study Site

The saline fen (56°34'28.84" N, 111°16'38.39" W) is located approximately 10 km south-southeast of the AOSR hub of Fort McMurray, Alberta, Canada (Figure 2-1), within the Central Mixedwood Subregion of the Boreal Plains Ecozone (Natural Regions Committee, 2006). Cold winters and warm summers characterize the Fort McMurray climate, with thirty-year daily average temperatures of -18.8°C in January to 16.8°C in July (Environment Canada, 2012). Annual precipitation (P) is low compared to other parts of the province (455 mm, Environment Canada 2013) and the majority of annual P (65-75%) falls during the summer growing season (May-September) in convective cell storms where daily rainfall rates are typically less than 10 mm (Smerdon et al., 2005; Devito et al., 2012). Potential evapotranspiration (PET) is high (annual average of 520 mm) and exceeds P for most years, creating a long-term water deficit that is satisfied by infrequent wet years that occur on a 10-15 year cycle (Ferone and Devito, 2004). The synchronization of peak rainfall with maximum PET during the summer months exacerbates the seasonal water deficit and limits the opportunity for overland flow (Devito et al., 2012).

The fen lies at approximately 400 masl within the McMurray lowlands subdivision of the Dover Plains, a relatively flat region characterized by widespread, continuous organic deposits (Andriashek, 2003). Several prominent upland features roughly encircle the broad lowlands surrounding the Fort McMurray area, the closest of which is the Stony Mountain complex that lies approximately 15 km south of the saline fen (Hackbarth and Nastasa, 1979). Flowing northwestward from the Stony Mountain uplands through the study area are the primary tributaries of the Hangingstone River; the Saline River, Prairie Creek and Salt Creek basins (Value Creations Inc., 2012). Glacial till is relatively thin around the study area (15-20 m) and overlies Cretaceous shale and sandstones of the Clearwater and bitumen bearing McMurray formations. Devonian carbonates underlie the McMurray formation at a depth of approximately 180 m (Value Creations Inc., 2012) and are a possible source of saline groundwater at the fen due to the presence of extensive halite deposits found within adjacent strata. The Lower Cretaceous Grand Rapids formation typically forms the bedrock surface for most of the Dover Plains but extensive erosion has truncated the formation south of the study site.

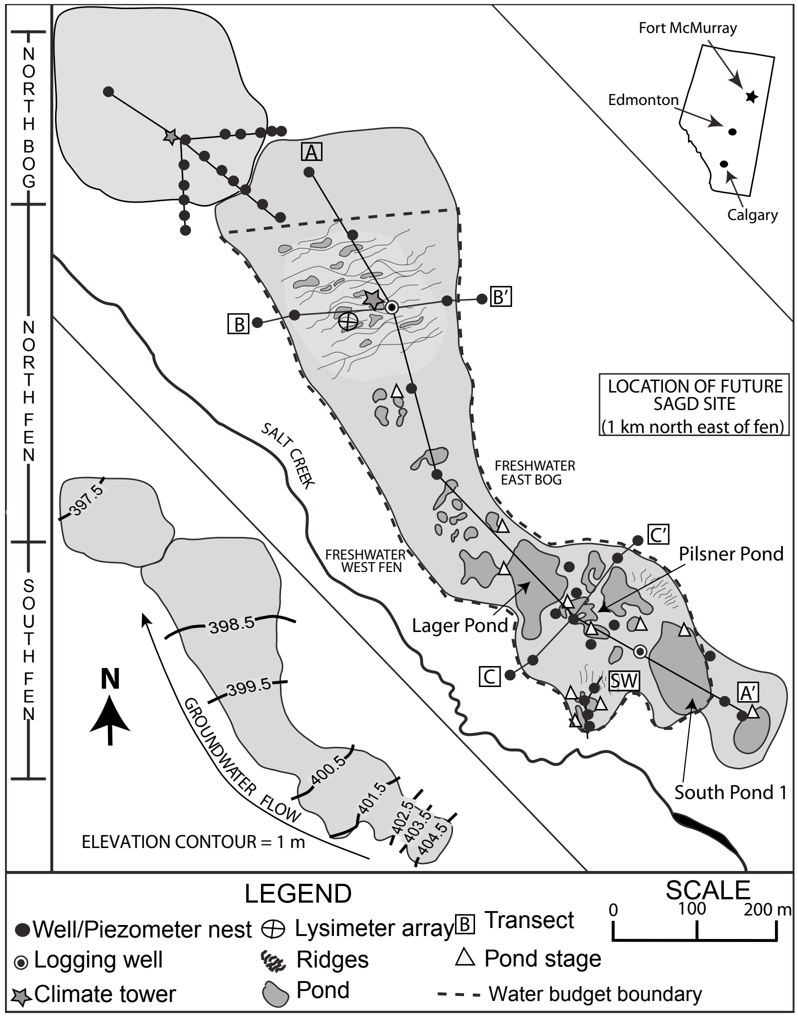

Figure 2-1 Map of the Saline Fen with site topography and location within Alberta on insets. Note the location of the future SAGD facility just east of the fen. Vertical legend to the left of the map divides the site into three distinct sections that are referred to as such throughout the text.

Comprising an area of approximately 27 ha, the saline fen lies adjacent to several other large peatland complexes of similar salinity and configuration. The surface elevation of the fen declines northward and is characterized by a steep gradient in the south (~8 m km^{-1}) that transitions into a gently slopping plain in the north half of the fen (~2.7 m km^{-1}) (Figure 2-1). A series of ridge and inter-ridge depressions run perpendicular to the predominant shallow groundwater flow direction and are most prominent in the northern section of the fen. A large pond network comprises 19% of the fen surface. Southward, the fen exhibits saline spring features, including halophytic vegetation, salt crusting and desiccated microbial mats that surround a series of what appear to be relict discharge features. The vegetation composition of the saline fen contains characteristic salt-tolerant vascular and non-vascular plant species. The dominant vegetation includes Sweetgrass (*Hierochloe hirta* ssp. *arctica*) around the outer fringe, Narrow reed grass (*Calamagrostis stricta*) and Foxtail barley (*Hordeum jubatum*) in the ridges, Seaside arrow grass (*Triglochin maritima*) and Redwool plantain (*Plantago eriopoda)* in the inter-ridge depressions, and Samphire (*Salicornia rubra*) in the salt flats and pools. Patches of birch (*Betula* spp.) and willow (*Salix* spp.) are found along the fen's outer margins and in low-salinity zones in the north half of the fen. Stands of black spruce (*Picea mariana*) and tamarack (*Larix laricina*) also occur along the margins which transition sharply into forested bog and/or fen peatlands. No significant moss cover is present in the main portions of the fen where there is elevated salinity.

2.4 Methods

Groundwater

Nests of wells and piezometers were installed in four transects across the fen and into adjacent wetlands (30 total nests; Figure 2-1) by pre-augering a hole and manually inserting the pipes. For installation into the underlying till or into peat that was particularly dense, a mallet was used to drive the pipes the remaining distance. All nests were constructed from 2.5 cm inner diameter polyvinyl chloride pipes with each nest consisting of a depth-integrated well and two to five piezometers with 17 cm slotted screens covered in well sock. Piezometer depths varied according to peat thickness but were usually centered at 0.50 and 0.75 m within the peat and between 1.0 and 3.0 m within the till. Nests located within zones of very high salinity had additional piezometers set at depths ranging from ~3.0 to 4.0 m. Continuous water level measurements were made at two wells located in the north and south ends of the primary transect (A-A’, Figure 2-1) using pressure transducers; manual measurement were taken at least once per week between June and September in 2011 and April and September in 2012. Fen topography and pipe top elevations were measured and referenced to sea level using a dual-frequency survey-grade GPS. Field estimates of horizontal saturated hydraulic conductivities were measured using bail tests for 30 piezometers in peat and 11 piezometers in the till using the method of Hvorslev (1951). For depths <50 cm, the modified cube method was used on samples processed in the laboratory for the determination of horizontal and vertical saturated hydraulic conductivities (Beckwith et al., 2003).

The vertical and horizontal fluxes of groundwater to and from the peatland were calculated using Darcy’s Law (Freeze and Cherry, 1979) as

$$Q = KA\ dh/dl \quad , \tag{1}$$

where Q is the discharge ($m^3\ s^{-1}$), K is the saturated hydraulic conductivity ($m\ s^{-1}$), A is the cross-sectional area (m^2) and dh/dl is the hydraulic gradient (dimensionless). For vertical fluxes, a geometrically averaged K obtained from the till was used. Vertical flux rates were determined by using the vertical hydraulic gradient (dh/dz) between the piezometers in the till and the water table measured within an adjacent well.

Due to elevated groundwater salinities within piezometers, hydraulic heads had to be corrected for differences in density and converted to freshwater equivalents before vertical gradients and fluxes could be calculated. The density of saline water is a function of temperature, salinity and pressure. In the case of the saline fen, groundwater pressures have a negligible effect on density due to shallow depth of piezometers and thus only the influence of temperature and salinity were considered. Salinity measurements were recorded for all piezometers in the field three times over the two-year study under dry (August, 2011 and June, 2012) and wet (August, 2012) conditions. Piezometers located within the till exhibited little variability in salinity within and between years so an average value was obtained and applied for the calculation of density for each piezometer. The average annual air temperature for Fort McMurray (0.7ºC) was applied to all density calculations, as it was determined that piezometers in the till were located close to or well past the zero annual amplitude depth of 1.5 m, calculated using

$$D = sqrt(K_{HS}*P_T) , \quad (2)$$

where D is the depth of zero annual amplitude (m), K_{HS} is the thermal diffusivity of the soil (m^2 s^{-1}) and P_T is the period of the temperature waves (s), in this case a year (Oke, 1987). Two soil probes located near the peat surface (~2.5 cm) and at 40 cm were used to record soil temperatures at half hourly intervals and K_{HS} of the peat was calculated as 9.0×10^{-8} m^2 s^{-1} using the time-lag method

$$K_{HS} = (P/4\pi)(\Delta z/\Delta t)^2 , \quad (3)$$

where Δz is the change in depth between the two probes (m) and Δt is the time lag between the crest of the temperature wave between the two depths (Oke, 1987). A sensitivity analyses was conducted to determine the influence of temperature on density calculations. Under both low and high salinity conditions, increasing the temperature from 1 to 15ºC resulted in a change in hydraulic head of equal to or less than 3 mm and is thus within the error of the survey GPS accuracy and manual water level measurements.

Once groundwater density was corrected based on the above approximations for each piezometer, hydraulic heads were corrected to freshwater equivalents using the equation

$$hf = (\rho_p/\rho_f)hp , \quad (4)$$

where hf is the freshwater pressure head (m), ρ_p is the density of the salt water hydraulic head (kg m^{-3}), ρ_f is the density of freshwater (kg m^{-3}) and hp is the field measurement of hydraulic head uncorrected for salinity (m) (Fetter, 2001). The computed freshwater hydraulic head was then added to the elevation head to obtain a freshwater corrected hydraulic head measurement. The conversion of saltwater head to freshwater head resulted in an average increase in hydraulic head of 3 cm. Groundwater with less than 10,000 mg L^{-1} total dissolved solids (TDS) at temperatures below 100°C are considered to have densities comparable to freshwater (1000 g L^{-1}), therefore, density corrections were made only on piezometers that exceeded this TDS threshold (Freeze and Cherry, 1979).

Micrometeorological Conditions

Meteorological parameters were monitored continuously between June and September, 2011 and May and September, 2012 using a data logger at a meteorological station centered within the north half of the fen (Figure 2-1). All instrumentation was measured at 60-second intervals and averaged every half hour. Summer precipitation (*P*) was measured continuously with a tipping bucket rain gauge (Hobo Onset) and supplemented with two manual bulk rain gauges. A tipping bucket malfunction in early July of 2011 required the use of rain gauge data (July to September) from the Fort McMurray airport weather station (AWOS-A configuration) 8 km northeast of the study site (Environment Canada, 2012). A comparison between the airport weather station and the fen tipping bucket when it was operational (June to early July) showed a strong correlation in the magnitude of each rain event (r^2 = 0.95), with the saline fen tipping bucket recording an average of 0.6 mm more rain per day then at the airport. Air temperature and relative humidity were measured at 1.0 and 3.0 m using a self-logging Hobo Onset logger. A net radiometer (NR-Lite2) and a wind monitor (RM Young 05103) were installed at the top of the weather station at 3.0 m to measure net radiation flux (Q*) and wind speed/direction. Soil heat flux plates were installed in an elevated ridge and in an inter-ridge depression to measure ground heat flux (Q_G).

Daily evapotranspiration (ET_a) for the peat surface was calculated using the Priestley and Taylor (1972) combination method, where

$$ET_a = \alpha(s/s+\gamma)(Q^*-Q_g/L_v\rho_w)\ , \qquad (5)$$

and where s is the slope of the saturation pressure-temperature curve (Pa °C^{-1}), γ is the psychrometric constant (kPa °C^{-1}), Q^* is the net radiation flux (J day^{-1}), Q_g is the ground heat flux (J day^{-1}), L_V is the latent heat of vaporization (J Kg^{-1}), and ρ is the density of water (kg m^{-3}). The coefficient of evaporability, α, represents the slope of the regression line relating actual evapotranspiration (ET_a) to equilibrium evaporation (ET_{eq}), which is the value of equation 5 when α=1. The coefficient α is determined empirically from independent measures of ET_a (Price and Maloney, 1994), using soil lysimeters. Once determined, α is used in equation 5 to determine daily ET_a. Six lysimeters filled with peat monoliths representative of the two major peat surface classes (ridge/lawns and inter-ridge depressions, 3 repetitions each) were installed in the north fen (see Figure 2-1) and used for determining ET_a and derivation of an α coefficient. The fen was divided into two peat surface classes (ridge/lawn and inter-ridge depression) based on observations of vegetation cover and water table position. Despite differences in surface elevation and thus water table position between microforms, during wet periods peat was typically moist throughout microforms. Vegetation was sparse within inter-ridge depressions and dominated by scattered patches of *Triglochin maritime*, *Puccinellia nuttaliana* and *Salicornia rubra*. Narrow reed grass (*Calamagrostis stricta*) and Foxtail barley (*Hordeum jubatum*) grew densely along ridges and dominated the lawn regions that comprised much of the middle and southern portions of the fen. Because vegetation cover and water table position below the ridges in the north fen were comparable to the ridge-lawn landscape of the south fen under baseflow conditions, ridge lysimeters in the north fen were considered representative of the south fen peat surface. Individual α values were derived for each surface type and an aerially weighted α value was derived for the entire peat surface, where:

$$\text{Site } \alpha = \Sigma(\alpha_i A_i) \tag{6}$$

and where A_i is the fractional aerial coverage of the ith surface class with an α_i coefficient of evaporability. A site scale α was then applied equation 6 to provide an estimation of total evapotranspiration from the peat surface at the fen.

The Penman equation (1948) was used to estimate daily potential evaporation (mm day^{-1}) for open water surfaces with no vegetation (E_o), where:

$$E_o = \frac{\Delta(Q^*) + \gamma\, \lambda_v\, \rho_w\, K_E\, v_a \{e^*_a - e_a\}}{\lambda_v\, \rho_w \{\Delta + \gamma\}} \tag{7}$$

and where Δ is the slope of the saturation vapor pressure versus temperature relationship at the ambient air temperature (kPa °C^{-1}), Q^* is net radiation flux (J m^2 day), γ is the psychrometric constant (kPa °C^{-1}), L_V is the latent heat of vaporization (J Kg^{-1}), ρ_w is the density of water (kg m^{-3}) K_E is the mass transfer coefficient (kPa^{-1}), v_a is the velocity of air (m day^{-1}), e^*_a is the saturation vapor pressure at ambient air temperature (kPa^{-1}) and e_a is the water vapor pressure (kPa^{-1}). K_E represents the aerodynamic component of the Penman equation and was calculated using the Thornthwaite and Holtzman equation (1939), where:

$$K_E = \frac{0.622\, \rho_a k^2}{p\, \rho_w\, [\ln\{(Z_a - Z_d)/Z_o\}]^2} \tag{8}$$

and where k is von Karmen's constant (0.4), ρ_a is the density of air (kg m^{-3}), p is atmospheric pressure (kPa), ρ_w is the density of water (kg m^{-3}), Z_a is the height of the velocity and temperature measurement (m), Z_d is the zero plane displacement (approximately 0 for water), and Z_o is the surface roughness height (~0.0003 m for open water) (Oke, 1987).

Daily estimates of ET for peat and E for open water were then fractionally weighted based on their aerial coverage and summed to obtain a site-scale estimation of daily ET for the entire fen. Due to extremely wet conditions in 2012, the relative aerial coverage of the different surface types (ridges, inter-ridge depressions and open water) changed as a consequence of rising water tables and the complete filling of the pond network. The majority of inter-ridge depressions in the north half of fen were completely inundated for the remainder of the study season and site-scale α values for the peat surface were adjusted to account for the reclassification of inter-ridge depressions as open water surfaces. The sparse vegetation of the inter-ridge depressions with relatively deep ponding (between 10 and 30 cm) allowed for the assumption that they were likely evaporating close to their potential rate without signification plant transpiration. The ridge/lawn surface class was then assumed to represent the peat surface, as water tables remained mostly below ridges in the north fen and rarely breached the peat surface in the south. Accordingly, the fractional weighting of the peat and pond surface classes were then adjusted to account for the increase in open water surface at the fen and a site-scale ET value was estimated.

Geochemistry

Groundwater was sampled from each piezometer in July of 2011 and in June and August of 2012 while pond surface water samples were obtained only in 2012 (both June and August) from selected ponds and pools that were also measured in-situ at least once per week during the season. Pond stage was also recorded for each in-situ measurement. Groundwater samples were extracted using foot-valves attached to plastic tubing that was rinsed thoroughly with distilled water prior to each use and piezometers evacuated several days before each sample period to ensure representative pore water was obtained. Unstable parameters (pH, temperature, electrical conductivity and salinity) were obtained in-situ by a handheld device (YSI 63 meter) that was calibrated before each field day. Electrical conductivity measurements were corrected in the field to 25°C. For the remaining parameters, samples were collected and preserved on ice in the field and frozen before laboratory analyses.

Major ion concentrations and alkalinity were determined at the University of Waterloo Ecohydrology Laboratory. Samples were passed through a 0.45 μm filter and saline samples were diluted before analyses. Alkalinity was determined using automated spectrophotometric flow injection analyses (Lachat Quickchem 8500). Cations were measured by inductively coupled plasma-optical emission spectrometry (ICP-OES, Thermo Scientific iCAP 6300) while anions were determined using a capillary ion chromatograph (Dionex ICS-5000). Analytical error in concentration measurements was determined to be less than 5%. The concentration of total dissolved solids (TDS) was estimated by summing the concentrations of the individual major ions (Fetter, 2001).

2.5 Results

Site stratigraphy and substrate characteristics

Peat thickness averaged 1.2 m and varied considerably across the fen, ranging from 1.5 m at its depocenter in the north half of the fen to almost zero along a thin band towards the fen's southwest margin. North of Lager Pond, the peat deposit thickened and the underlying mineral layer formed a basin-like morphology that quickly transitioned into an elevated mound along the bog-fen margin (Figure 2-2). Here, the ground surface followed the change in mineral elevation, creating a local topographic high where peat thickness decreased to around 0.5 m. Similar mineral mounds were observed along the B-B' transect in the north fen, where the mineral layer formed a distinct sinusoidal pattern that increased in elevation along the fen margins (Figure 2-3a). In this area, however, the surface topography did not match the substrate. The ground surface formed a gentle topographic high along the eastern edge of the fen and sloped gradually westward into the adjacent wetland. Southward (i.e., up-gradient), the overall peat thickness decreased and became more variable. Along the C-C' transect in the south fen, the ground elevation sloped westward, forming a wedge that tapered to around 0.5 m where the mineral substrate formed a mound similar to that observed in the north fen (Figure 2-3b).

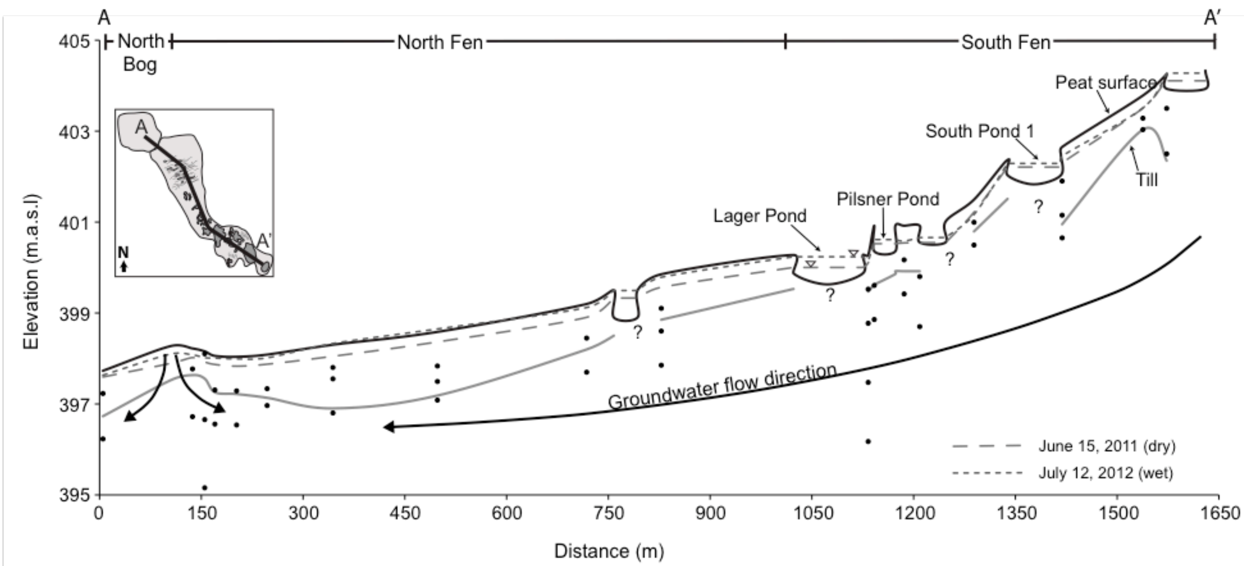

Figure 2-2 A cross-sectional profile of the north bog and saline fen along the primary A-A' transect. General fen boundaries are indicated by the legend along the top of the figure. (•) indicate the location of piezometer intakes while (?) refer to locations where depth to mineral till is unknown. The two dashed horizontal lines represent typical water table elevations under dry (June 15th, 2011) and wet (July 12th, 2012) conditions.

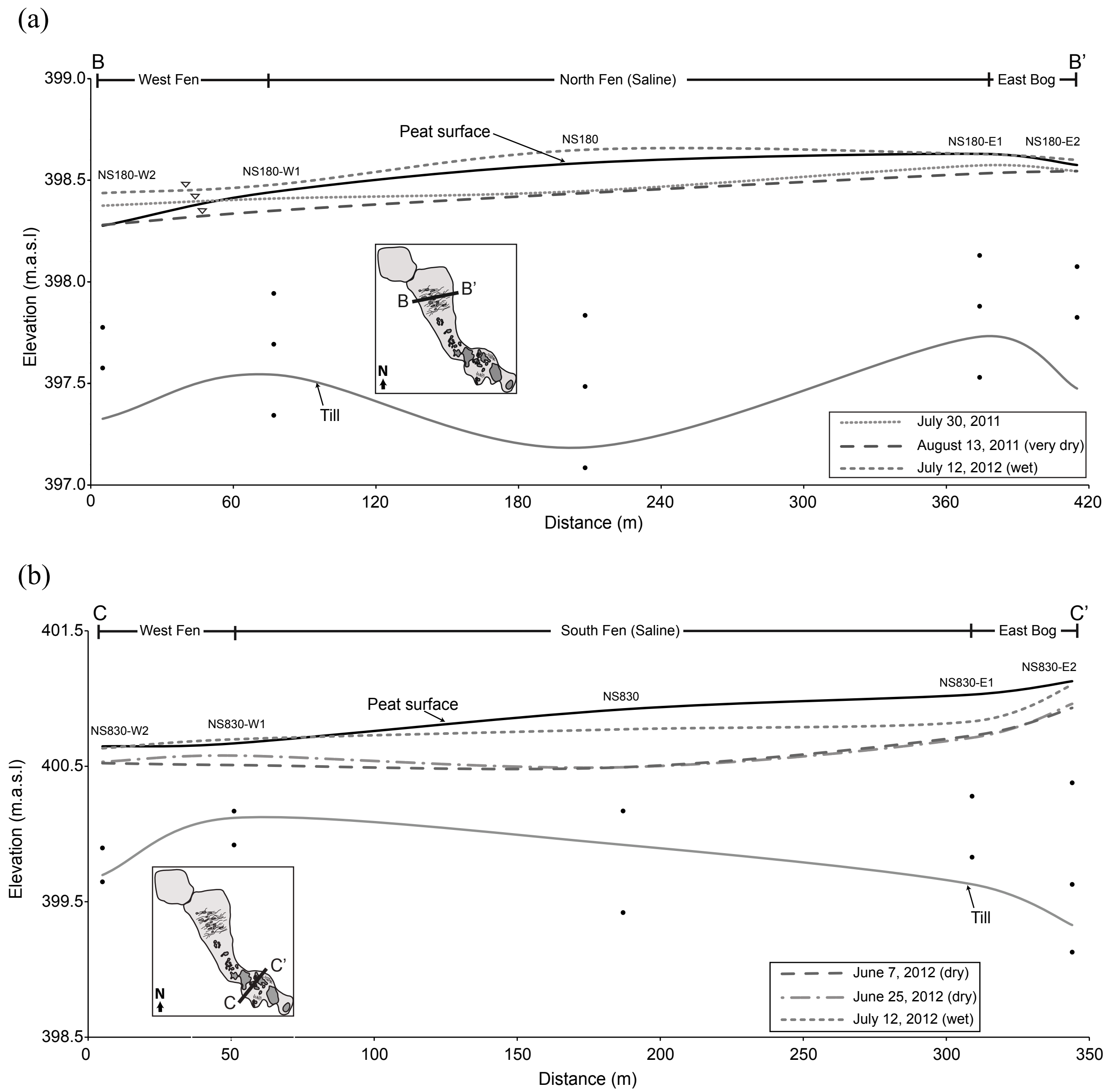

Figure 2-3 Cross-sectional profile of the B-B' in the north fen (a) and C-C' transect in the south fen (b) with typical daily water table trends over both study seasons. General fen boundaries are indicated by the legend along the top of the figure. (•) indicate the location of piezometer intakes while nest identifiers are located above each piezometer nest (e.g., NS180). Note the difference in scales on the x-axis between cross-sections.

The morphology of the pond network was highly irregular and the ponds ranged in size from as large as 1 ha to as small as 1 m^2 or less (Figure 2-1). Deep depressions with steep, almost vertical margins characterized many of the ponds in the south and central fen. These ponds were often small (<400 m^2) and formed roughly circular to oval depressions, with their long axis oriented transversely to the local gradient. In other cases, ponds formed thin bands parallel to water flow that tapered towards their edges and were generally shallow (<30 cm). In general, pond floors comprised a dense, decomposed peat underlain by mineral till.

Surface peat (0 to 10 cm) was weakly decomposed (bulk density ~ 0.12 g cm^3) and highly fibric and became more humic and amorphous at depth while maintaining its structure and firmness (Table 2-1). Unlike decomposed *Sphagnum* peat, the moderately decomposed sedge peat could rarely be extruded with pressure and many of the decomposing plant structures were still recognizable even at the peat-mineral interface. At all depths, decomposing roots and stalks within the sedge peat were predominantly vertically oriented. Average bulk densities of the lower peat layers (<50 cm from the surface) ranged from 0.15 g cm^3 to 0.22 g cm^3 and increased with depth. Peat specific yield was generally low and followed no discernable pattern with depth, averaging 0.05 (maximum of 0.12 and minimum of 0.02) for the entire fen (Table 2-1).

Table 2-1 Physical properties of peat for the saline fen. ρb is bulk density, *S*y is specific yield, K_{LH} is saturated horizontal hydraulic conductivity and K_{LV} is saturated vertical hydraulic conductivity. Values of saturated hydraulic conductivity were obtained in the lab. All data are means (geometric for K) of three or more samples.

Depth	Humification (Von Post)	ρb (g cm^{-3})	S_y	K_{LH} (cm s^{-1})	K_{LV} (cm s^{-1})
0-10	H2	0.12	0.05	$8.5x10^{-3}$	$1.5x10^{-2}$
10-20	H2	0.14	0.05	$5.6x10^{-3}$	$8.1x10^{-2}$
20-30	H3	0.15	0.04	$3.6x10^{-3}$	$5.5x10^{-3}$
30-40	H3	0.15	0.06	$2.5x10^{-3}$	$3.7x10^{-3}$
40-50	H4	0.16	0.05	$2.3x10^{-3}$	$3.5x10^{-3}$
50-60	H3	0.15	0.06	$2.4x10^{-4}$	$2.0x10^{-3}$
60-80	H4-5	0.22	0.07	$1.6x10^{-3}$	$2.2x10^{-3}$
80-100	H4-5	0.17	0.09		
100-130	H4-5	0.22	0.06		

Underlying the basal peat was a homogenous and laterally continuous mineral layer. The blue colour and high plasticity and cohesion of the mineral till indicated an abundant clay fraction. Variability in the underlying mineral layer was minimal but coarse-textured clay-silt till containing sand and pebble fractions was identified underlying two highly saline sections of the fen.

Field (K_H) and laboratory (K_{LH}) estimates of peat horizontal saturated hydraulic conductivities are shown in Figure 2-4. The geometric mean K_H of the fen peat was $2.6x10^{-4}$ cm s^{-1}, or 22 cm d^{-1} (n = 36) with considerable range both spatially across the fen and with depth. Higher K_{LH} in the poorly decomposed and highly fibric upper peat layers towards the surface was observed, with a geometric mean K_{LH} of $3.2x10^{-3}$ cm s^{-1} (280 cm d^{-1}, n = 18). Laboratory analyses of both horizontal and vertical saturated hydraulic conductivity (K_{LV}) provide a good indication of the level of anisotropy at the saline fen. Over 70% of the samples (n = 22) had greater K_{LV} (geometric mean of $7.1x10^{-3}$ cm s^{-1}) than K_{LH} (Table 2-1). The K_H of the underlying mineral layer varied spatially and by several orders of magnitude. A geometrically averaged field estimate of mineral K_H was found to be $5.5x10^{-7}$ cm s^{-1}, or 0.05 cm d^{-1} (n = 11). Spatially, mineral K_{LH} varied between 10^{-5} and 10^{-8} cm s^{-1}, with the highest K_{LH} observed in the high salinity zones in the fen's southern section.

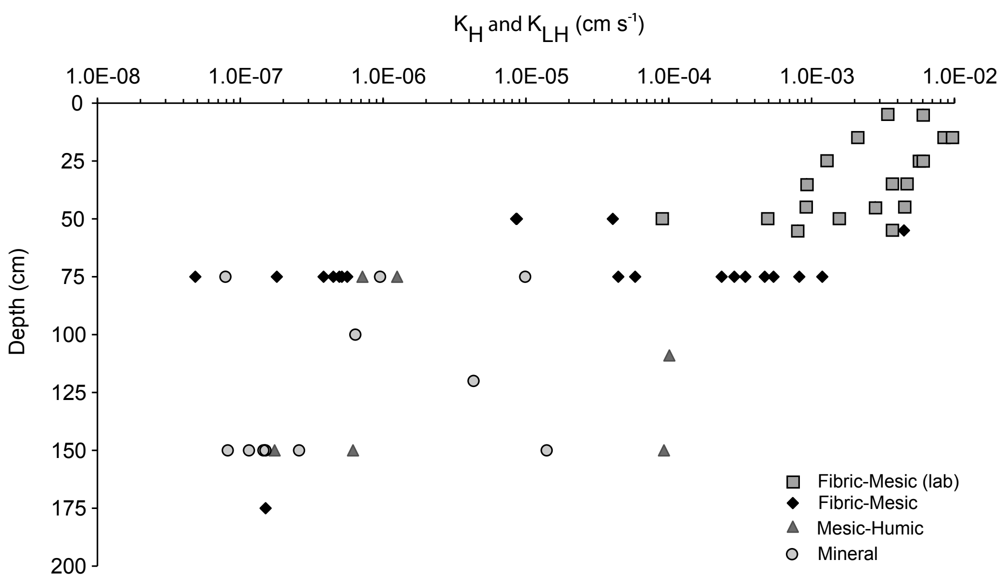

Figure 2-4 Field (>50 cm depth) and laboratory (<50 cm depth) estimates of saturated horizontal hydraulic conductivities for peat and mineral till. Fen peat was divided into two main classifications based on the von Post system of humification (von Post, 1924) and the Canadian three-part fiber content system (Lemasters et al., 1983).

Precipitation and Evapotranspiration

Over the two study seasons rainfall varied considerably, providing an opportunity to contrast dry and very wet conditions. During the 2011 study, between June and mid-September, 192 mm of rain fell compared to 438 mm in 2012 for the same period (Figure 2-5). In 2011, monthly rainfall totals were all well below the long-term average for the Fort McMurray region, with over 80% of the daily rain events falling as short-duration storms yielding less than 5 mm. The extended 2012 study season saw a total of 488 mm of rain between April 1st and September 17th and captured the spring and early summer rainfall input missed in 2011. An unusually wet April (36 mm) was followed by a dry period in May where only 13.5 mm of rain was recorded, compared to the long-term average of 75 mm. The most significant contribution to total rainfall at the fen occurred in July, with over 200 mm that was dominated by a series of high intensity storms that repeatedly exceeded 20 mm d^{-1}. Several storms in July, August, and September had massive inputs of rain over relatively short time periods. For example, over 130 mm of rain (24% of seasonal total) fell over a three-day period in July with an average rainfall intensity of 3.3 mm h^{-1}. September was another significant month for water input at the fen in 2012, recording almost 80 mm more rain than the monthly average.

The areal average daily ET_a rate for the fen peat surface (inter-ridge depressions and ridge/lawns) in 2011 was 3 mm (maximum 6 mm d^{-1}) based on a site-scale α coefficient of 1.19 for peat (Table 2-2). Between surface types, α coefficients were on average greater in the ridge/lawns (1.21) than in the inter-ridge depressions (1.15). As a result, calculated ET_a losses from peat were greatest in the ridge/lawns, totaling 332 mm compared to 314 mm for the inter-ridge depressions despite an equivalent average ET_a (Table 2-2). While water tables and pond storage did decline steadily towards the end of the 2011 study (Figure 2-5), with the exception of a few ponds in the fen's south end, the pond network remained saturated for the entire summer. The average rate of E_o from the ponds was 3 mm d^{-1} (maximum 7 mm d^{-1}) for a total of 368 mm for the season (Table 2-2). For the entire fen, including all peat surface classes and the pond network, the areal average ET_{site} rate in 2011 was 3 mm d^{-1}, reaching its maximum daily rate of 6 mm in late June. Between June 1st and September 17th, total ET from the fen was 333 mm, with maximum rates of ET_{site} occurring in late June through to the end of July, in synch with maximum rainfall input (Figure 2-5). Cumulative ET_{site} exceeded cumulative P for the entire 2011 season.

Taking into account an unusually dry spring in 2012, the areal average daily ET_a rate for the peat was 2 mm, with a maximum of 6 mm in early July. Unlike 2011, ET_a was greatest in the inter-ridge depressions (439 mm, maximum of 6 mm d^{-1}) compared to the ridge/lawns (423 mm, maximum of 6 mm d^{-1}) (Table 2-2). A much wetter season in 2012 led to flooding conditions in the north fen from early July until the end of the study season in mid September, filling the entire pond network and changing the surface conditions across the site. The average rate of E_o from the ponds was 3 mm d^{-1} (maximum 7 mm d^{-1}) for a total of 526 mm between April 1st and September 17th (Table 2-2). Prior to heavy rainfall in early July, a site-scale α of 1.13 was applied for estimation of peat ET_a. Extensive flooding filled the majority of the inter-ridge depressions completely in the north fen, resulting in a reevaluation of the surface class scheme and fractional weighting of ET_{site}. Since the ridges and lawns in the fen remained largely above the water table, an α of 1.12 was applied to estimates of peat ET_a (calculated from ridge/lawn lysimeters) while flooded inter-ridge depressions were reclassified as ponds and pools (see methodology). Based on this, the aerially weighted average ET_{site} rate for the entire fen in 2012, including all peat surface classes and the pond network, was calculated as 3 mm d^{-1} for a total of 449 mm between April 1st and September 17th. Between the end of April and late June, cumulative ET_{site} far exceeded P. High intensity rain events beginning in July and persisting for the remainder of the season resulted in cumulative P roughly matching ET_{site} until early September, following which P exceed ET_{site}.

Table 2-2 Percent cover, α values (from Priestley-Taylor method for peat), open water evaporation (E_o) for ponds and evapotranspiration rates (ET_a) for each surface type along with a site-scale seasonal ET_{site} rate for the entire saline fen. 2011 values are from June 1st to September 17th (108 days) and 2012 is from April 1st to September 17th (170 days).

		Inter-ridge Depressions	Ridge/Lawn	Ponds
	% Total Cover	33	48	19
2011	α	1.15	1.21	
	ET_a | E_o Rate (mm d^{-1})	3	3	3
	Total ET_a | E_o (mm)	314	332	368
	Seasonal ET_{site} Total (mm)†		333	
2012	α	1.15	1.12	
	ET_a | E_o Rate (mm d^{-1})	3	2	3
	Total ET_a | E_o (mm)	439	423	526
	Seasonal ET_{site} Total (mm)†		404	

† aerially averaged seasonal ET rate

Water Table and Storage Changes

Water table dynamics were similar between logged and manually observed wells for both the north and south sections of the fen and thus Figure 2-5 can be seen as a representative approximation of general water table trends for those areas (see Figure 3-1 for well locations). In 2011, a rapid increase in water table elevation was observed throughout the fen during the wet early summer period (June). While the largest daily rain event in early July (20 mm) raised the water table temporarily above the surface in the south fen, the same response was not observed in the north fen and surface storage features (e.g., inter-ridge depressions) remained dry throughout the season. For the remainder of July, low rainfall coupled with peak ET_{site} rates resulted in a steady mid-summer water table decline that was periodically interrupted by small rainfall events.

Continuous measurements over the winter in the north fen showed a small but gradual rise in the water table from late December to early February. This was followed by a stable water table regime up until the initiation of the 2012 spring freshet in late March, where snowmelt replenished the winter storage deficit and brought the water table to within several cm of the ground surface. Total rainfall in May 2012 was well below the monthly average for the region and water tables declined steadily until mid-June in response to continued ET_{site} (Figure 2-2 and 2-5). Over 40% of the total rainfall for the 2012 study season fell during the month of July, with 130 mm falling over a three-day period that satisfied the spring storage deficit developed in May. As a result of consistent, high-intensity rainfall, surface storage features were quickly filled and the majority of the pond network exceeded bankfull stage, flooding the adjacent peat surface and becoming connected to other ponds. A shift in the hydrologic regime between the north and south sections of the fen can be seen after this point, with the water table in the north fen persisting mostly above the peat surface for the remainder of the season (Figure 2-2 and 2-5). In the south, these flooding events were episodic and followed by a rapid decline in water table.

Pond Stage and Pond-Peatland Interaction

Ponds and pools comprised 19% of the fen surface and were a dominant landform feature within the south and central portions of the fen (Figure 2-1). Over the two-year study pond stage varied considerably, both in response to short-term weather variability as well as spatially between ponds. While pond stage was not measured in 2011, response to the sustained drying trend beginning in mid July was visually recorded. During this time, a dry late summer and early fall resulted in surface storage decline that led to some ponds drying up completely by the end of the study. In many cases adjacent or nearby ponds remained at least partially filled.

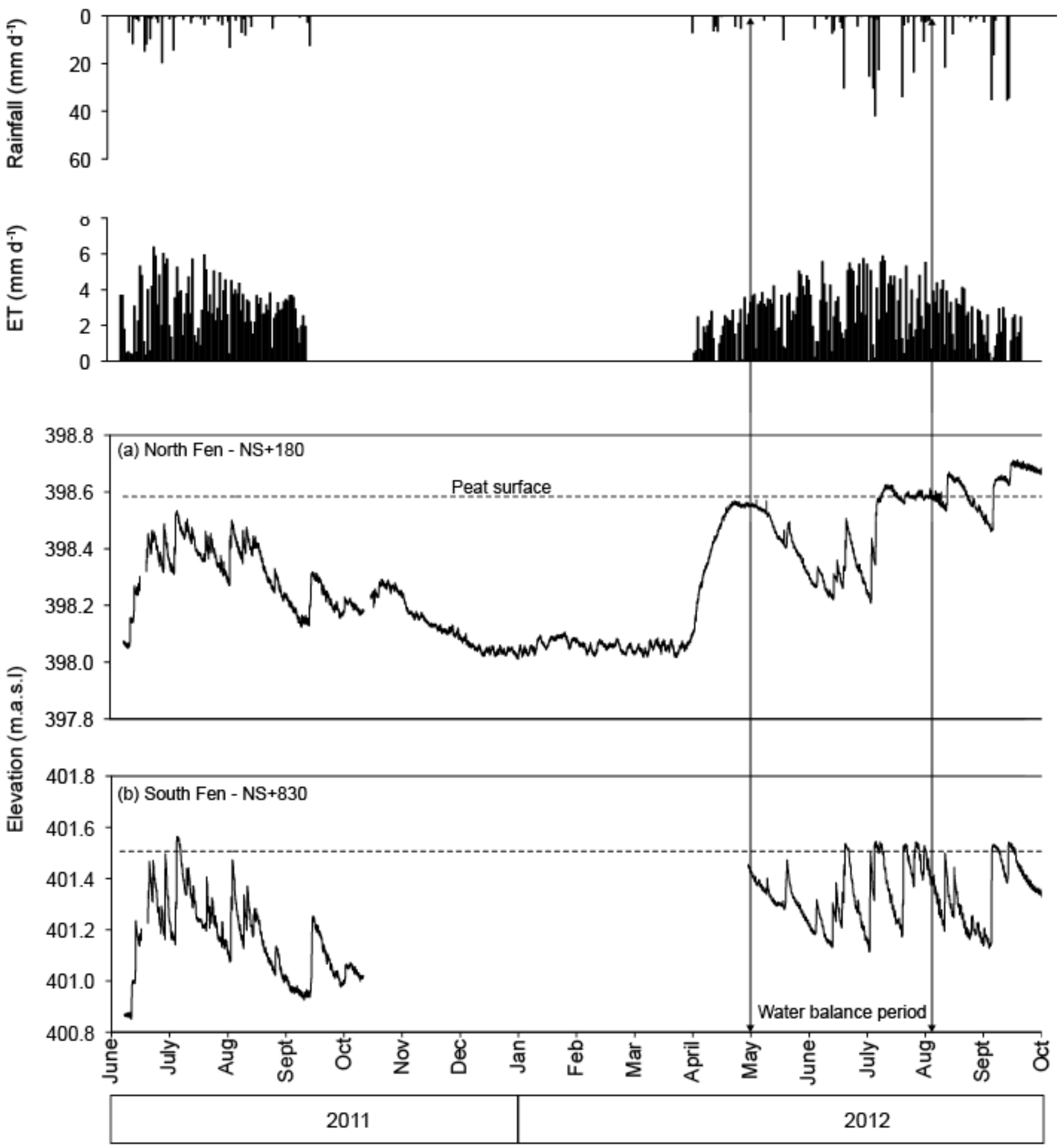

Figure 2-5 Micrometeorological conditions and water table elevations for the north (a) and south (b) portions of the fen during the 2011 and 2012 study seasons. Well locations can be found in Figure 3-1. The dashed horizontal lines on the hydrographs represent the peat surface.

The pond network exhibited a typical seasonal trend for the 2012 study season; the water level gradually declined after the spring freshet (Figure 2-5) until a series of high intensity rain events in early July (DoY 183-188, Figure 2-6). The majority of the ponds filled completely and in some cases spilled over their banks, creating a large network of interconnected ponds and pools. During rain events, some ponds exhibited an increase in stage that roughly matched the amount of rainfall for that period (e.g., 1:1 slope) while for other ponds, stage level increase exceeded precipitation depths. In some cases, pond stage continued to increase post rain events (DoY 187-194, Figure 2-6).

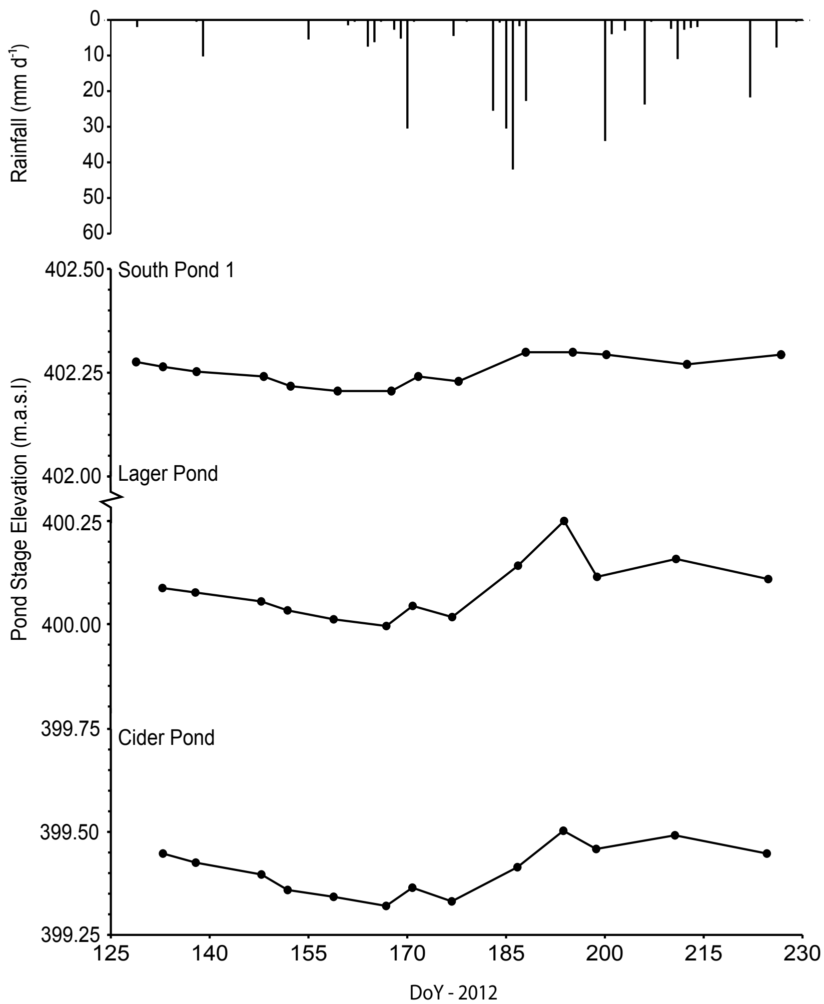

Figure 2-6 Pond stage elevations during the 2012 study period. Ponds were selected based on their position within the fen and can be found in Figure 2-1. Pond stage was measured manually.

However, pond stage was not measured continuously and only several rain events could be examined due to the number of days between typical measurement periods (between 5 to 10 days). In general, ponds of larger size and smaller perimeter-to-area ratios had smaller rates of stage recession following drying events compared to smaller ponds.

The complexity of pond-peatland interaction was studied in greater detail in 2012 by examining Pilsner Pond, a 0.25 ha, irregularly shaped saline pond in the south fen (Figure 2-7). The overall groundwater flow direction was from southwest to northeast following the local topographic relief. Horizontal hydraulic gradients (*dh/dl*) between the pond and adjacent well were on average positive (flowing from pond to peat) and steeper in the north side of the pond (0.04), compared to the south (-0.02) where groundwater discharged into the pond. The peatland water table was more sensitive to rain and following wetting events, horizontal *dh/dl* between the pond and the southern peatland margins became steeper (mean *dh/dl* of -0.05 on DoY 187). An opposite trend was observed along the northern margins, where a rapid rise of the peatland water table led to a decrease in horizontal *dh/dl* relative to pond stage (mean *dh/dl* of 0.02 on DoY 187). The pond was both a sink and store of groundwater over the season, feeding the peatland during periods of water deficit and functioning as a surface storage feature during stretches of elevated water table. Groundwater interaction between the pond and peatland was dynamic and complete flow reversals occurred in as little as six days during periods of low or intermittent rainfall.

Groundwater Dynamics

Lateral subsurface flow patterns were consistent between years throughout the fen despite variability in precipitation and water table position and the overall topographic gradient drove a south to north groundwater flow regime (Figure 2-2). Horizontal *dh/dl* along the primary transect (A-A') for both years were steepest (≥ 0.007) and more variable in the south fen (south of Lager Pond) where the topographic gradient is steeper and where the specific yield of peat is lower. Between years and throughout each season, average *dh/dl* remained stable under both wet and dry conditions and based on a geometrically averaged peat K_H of 2.6×10^{-4} cm s^{-1}, the average rate of specific discharge through the peat in the south section of the fen was calculated to be 2 mm d^{-1}. North of Lager Pond, the average horizontal *dh/dl* was also stable between years (<0.004) and reflected the mild relief, decreasing gradually toward the bog-fen margin where horizontal *dh/dl* reversed and groundwater between the fen and north bog converged (Figure 2-2). An average rate of specific discharge for the north fen was calculated as 0.6 mm d^{-1} and 0.04 mm d^{-1} under dry and wet conditions, respectively.

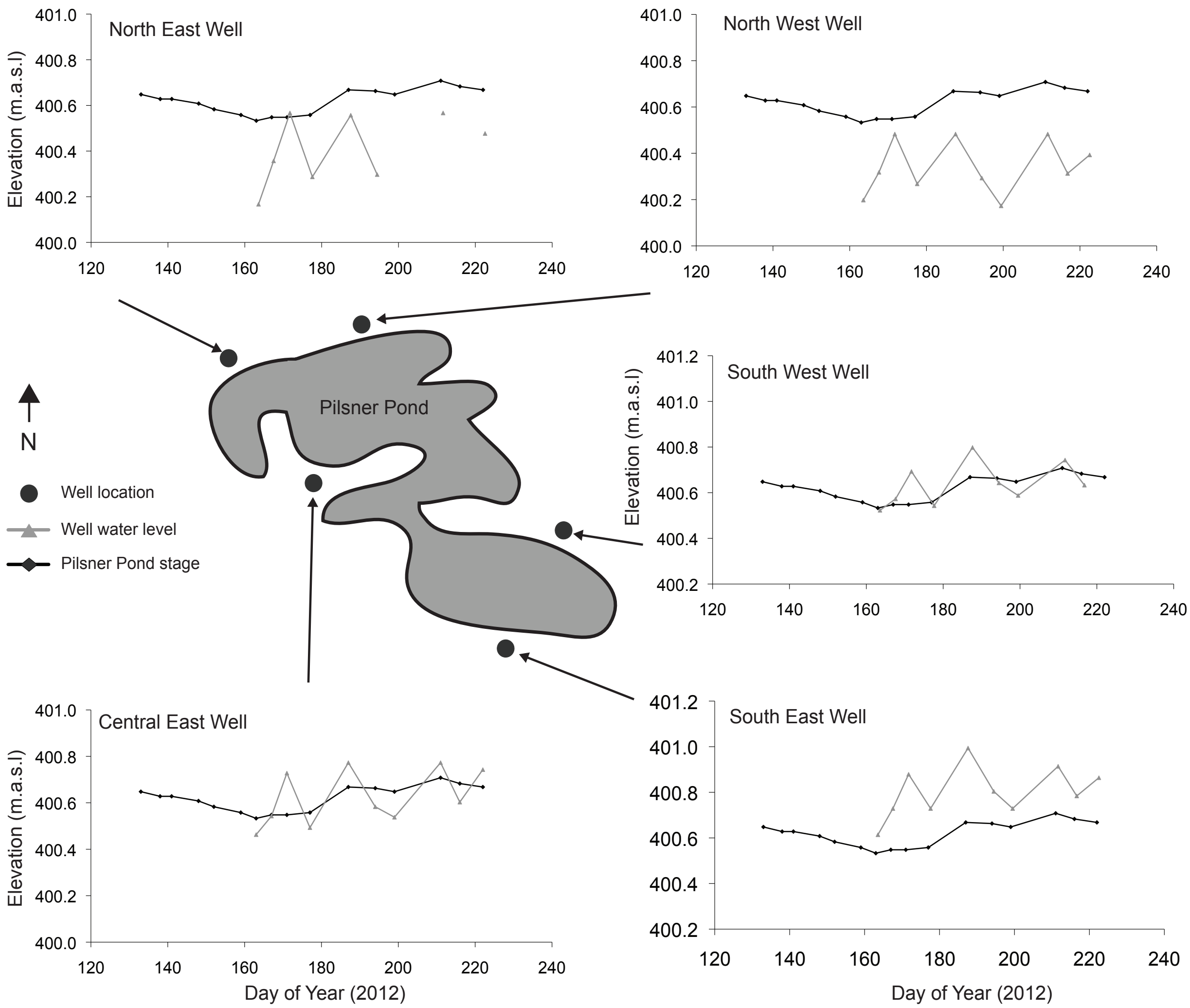

Figure 2-7 Pilsner Pond-groundwater interaction over the 2012 study season. Pond stage and water table elevations were measured manually.

Subsurface flow patterns between the saline fen and adjacent wetlands were variable and sensitive to short-term weather changes. Despite this variability, groundwater exchange was generally minor due to weak horizontal gradients. West of nest NS180 on the B-B' transect in the north fen, horizontal *dh/dl* (and hence flow direction) indicated the consistent discharge of groundwater into the adjacent treed west fen (nest NS180-W2) for both study seasons (Figure 2-

3a, all-dates). Horizontal *dh/dl* between the saline fen margin (nest NS180-W1) and NS180-W2 were weak (<0.002), discharging at an average rate of <1 mm d^{-1}, and marginally higher during periods of high water table. A gentle but persistent groundwater mound along the eastern margin (NS180-E1) of B-B' impeded subsurface flow towards the north fen from the treed east bog for the majority of 2011 and 2012 (Figure 2-3a, July 30th 2011). The ephemeral nature of this groundwater mound was pronounced in 2012 and horizontal *dh/dl* between the saline fen and east bog reversed several times throughout the study season (Figure 2-3a, July 12th 2012). Horizontal *dh/dl* (mean of 0.002 in 2011 and 0.006 in 2012) indicated flow from the saline fen margin to the east bog at a rate of <1 mm d^{-1}.

In the south fen (C-C' transect), groundwater consistently discharged into the saline fen margin (nest NS830-E1) from the east bog (nest NS830-E2) for both years at an average rate of 1.3 mm d^{-1} (Figure 2-3b, all-dates). West of C-C', reversals in horizontal *dh/dl* between the saline fen margin (nest NS830-W1) and the adjacent treed west fen (nest NS830-W2) were again driven by the cycling of wet and dry periods. Under baseflow conditions or drought periods, a gentle water table mound impeded flow from the west fen into the saline fen for the majority of 2011 and much of 2012 (Figure 2-3b, June 25th 2012). Following rain events, a rapid rise in water table within the south fen eliminated the mound, initiating complete east to west flow through conditions that were typically short-lived (i.e., ~5 days) (Figure 2-3b, July 12th 2012). During periods of extended drought (e.g., May 31st to June 11th, 2012), the water table west of nest NS830 in the south fen sloped against the peat surface and flowed eastward, into the saline fen at an average rate of less than 1 mm d^{-1} (Figure 2-3b, June 7th 2012). Flux between the wetlands along the western margins of C-C' was very low due to weak gradients (mean of 0.001 for 2011 and 2012, maximum of 0.002), and rates of specific discharge were typically less than 1 mm d^{-1}.

On average, vertical gradients (*dh/dz*) were generally positive throughout the fen in 2011, indicating downward (recharging) groundwater flow into the underlying till. The same general trend was observed for the north fen in 2012 with the exception of sporadic groundwater discharge conditions, which occurred more frequently during this season and were largely restricted to the region south of Lager Pond. For both years, vertical *dh/dz* were generally weak and sensitive to short-term seasonal weather changes; extended dry periods induced apparent reversals in vertical *dh/dz* at many nests, initiating discharge conditions that were weak and short-lived. However, because of the very low hydraulic conductivity of the underlying till (mean of 5.5×10^{-7} cm s^{-1}), the authors were cautious in interpreting vertical flux patterns in greater detail due to possible inaccuracies in *dh/dz* calculations (e.g., piezometer time-lag errors in low hydraulic conductivity substrates). Nevertheless, weak vertical *dh/dz* coupled with low hydraulic conductivity sediments resulted in rates of (specific) discharge and recharge that were negligible over the entire study (< 1 mm $season^{-1}$). The pond network in the south fen did not reveal any active spring outlets or discharge mounds.

Water Balance

The magnitude and relative importance of the fen's hydrologic components are compared through an assessment of its water balance. The water balance equation for the fen can be written as:

$$P + G_{in} + D_{in} - E - G_{out} - D_{out} = \Delta S + \xi$$

where P is precipitation, G_{in} and D_{in} are shallow horizontal groundwater and deep vertical inflow respectively, E is evapotranspiration, G_{out} and D_{out} are shallow horizontal groundwater and deep vertical outflow respectively, ΔS is change in storage, and ξ is the residual error term (Table 2-3). While the majority of inter-ridge depressions were inundated in the north fen from July until September of 2012, no surface runoff was observed between the fen and adjacent wetlands and thus is not included in the water balance equation. The water balance was calculated for an area of the fen bounded by South Pond 1 on the south, the winter road to the north and the margins of the adjacent bog and fen wetlands to the east and west (Figure 2-1). Computations were estimated from field data for 2012 only, between May 8th and August 9th, as it was the only study season with a complete data set.

The largest single component over the water balance period was precipitation (311 mm) followed closely by water loss through evapotranspiration (300 mm), and combined these variables dominated the water balance for the fen. Shallow groundwater inflow from fen margins comprised less than 20% (60 mm) of the total input over the study season and was the dominant source of non-meteoric water for the fen. The discharge of shallow groundwater to adjacent wetlands was also low, comprising less than 3% (7 mm) of total seasonal output. The low hydraulic conductivity of the underlying till restricted the vertical component of groundwater at the fen and both deep groundwater recharge and discharge were negligible over the water balance period. Change in storage (ΔS) over the study season was estimated using the specific yield (Sy) of the peat and change in water table elevation (Δh), such that $\Delta S = \Delta hSy$. For the calculation of ΔS for ponds, a Sy of 1 was used. ΔS was estimated to be 7 mm in the peat and 18 mm in the pools. Based on pool and peat surface area, the net storage change in the fen was approximately 8 mm, representing only 3% of precipitation. This was much less than change in storage calculated as a residual (65 mm) which gives rise to the error term (ξ) in the water balance equation; in this case 57 mm.

Table 2-3 Summary of hydrologic budget components for the 2012 study season. All values are in mm.

P	ET	GW_{in}	GW_{out}	D_{in}	D_{out}	ΔS	ξ	% Error
311	300	60	7	< +/- 1	< +/- 1	65	57	18

Salinity Distribution

The spatial distribution of groundwater salinity in the near surface peat (≤ 50 cm depth) is represented in Figure 2-8 as contours of electrical conductivity (EC). Near-surface groundwater EC measured in 2011 decreased from an average of 39 mS cm^{-1} in the south fen to 19 mS cm^{-1} in the north fen, with the area around Lager Pond marking the general boundary between brackish (<10,000 mg L^{-1} TDS) and saline groundwater (>10,000 mg L^{-1} TDS) (Table 2-4 and 2-5). The same spatial trend of northwardly decreasing groundwater EC was observed within the deeper peat layers (>50 cm) across the two-year study and salinity remained relatively stable over the three measurement periods (Table 2-4). Average groundwater EC in the south fen (34 mS cm^{-1}) was approximately double that of the north fen (16 mS cm^{-1}). The same trend was observed for the underlying mineral till.

Groundwater EC exhibited considerable variation over short distances throughout the fen. In the south fen, several 'hotspots' of elevated EC were observed (>30 mS cm^{-1}). These high salinity zones were distinct from the rest of the fen, exhibiting sparse vegetation cover, salt surface crusts and what appeared to be desiccated microbial mats surrounding irregularly shaped pools and ponds. North of Lager Pond, near-surface groundwater EC decreased rapidly in the direction of site-scale groundwater flow, from 30 to 5 mS cm^{-1} over a distance of less than 300 m (Figure 2-8). This dip in EC was followed by a relatively large region of elevated and stable salinity that comprised much of the fen's northern extent (between 15 and 20 mS cm^{-1}). A sharp decrease in near surface groundwater EC was observed approaching the fen's eastern and western margins and most notably towards the treed north bog where EC decreased from 5 to 0.5 mS cm^{-1} over a distance of several meters. Within the adjacent wetlands, groundwater EC was significantly lower than that of the saline fen and, on average, was slightly higher in the west fen than that of the east bog (Table 2-4).

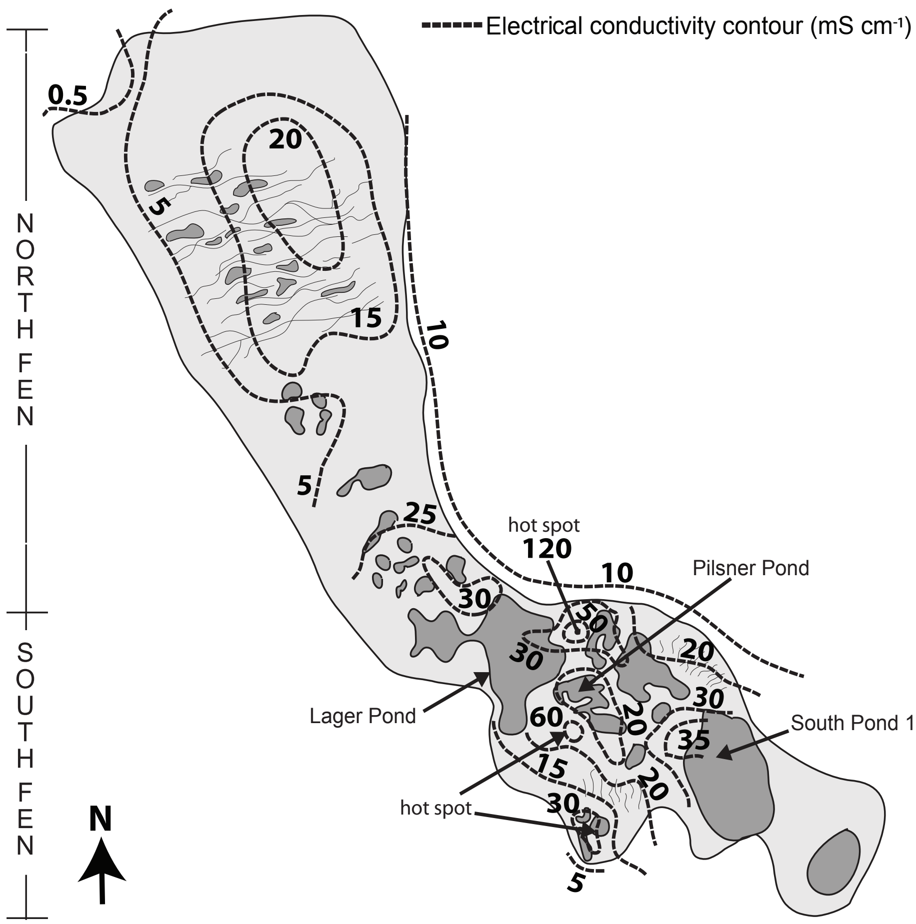

Figure 2-8 Near-surface groundwater salinity distribution shown as contours of electrical conductivity. Lager Pond marks the general boundary between brackish (north fen) and saline (south fen) groundwater. Note the location of the saline 'hot spots' in the south fen. The North Bog is not shown as it was not mapped for near surface salinity in 2011.

Table 2-4 Electrical conductivity and pH values averaged over three measurement periods (August 15, 2011 (dry) and June 19th (wet) and August 12th (wet), 2012) for the saline fen and adjacent wetlands. Each wetland parameter is divided into peat (P) and mineral (M) substrates. Electrical conductivity is in mS cm^{-1}. Measurements were obtained from piezometers at depths >50 cm from the peat surface.

				Saline Fen				Adjacent Wetlands				
	Date	Bog		North Fen		South Fen		West Fen		East Bog		Ponds
	(mm-yy)	P	M	P	M	P	M	P	M	P	M	
pH	08-11			6.4	6.2	6.6	6.6	6.6	6.2	6.2	6.3	
	06-12	4.1	6.9	6.4	6.6	6.4	6.7	6.6	6.5	6.5	6.4	7.7
	08-12		6.7	6.4	6.3	6.4	6.6	6.4	6.5	6.5	6.4	7.9
EC	08-11			16.3	21.0	29.8	49.6	7.1	25.9	6.0	17.9	
	06-12	.09	0.74	15.9	22.8	35.6	45.8	6.9	26.7	5.1	19.2	20.8
	08-12		0.80	14.4	20.4	33.5	51.6	9.8	27.8	6.4	18.4	25.4

In general, fen peat exhibited a weak relationship of increasing EC with depth (Figure 2-9 and Table 2-4). However, for some high salinity zones, such as the within one of the saline 'hotspots' where near-surface groundwater EC exceeded 100 mS cm^{-1}, this trend reversed and EC decreased linearly with depth (Figure 2-9). A marked increase in EC was observed within the underlying mineral till throughout the fen (Table 2-4). This increase was most pronounced within the adjacent east and west wetlands where mineral EC was twice that of the overlying peat. At one location in the adjacent east bog on the C-C' transect, EC increased from an average of 3 mS cm^{-1} at 75 cm (peat) to almost 20 mS cm^{-1} at 200 cm within the till.

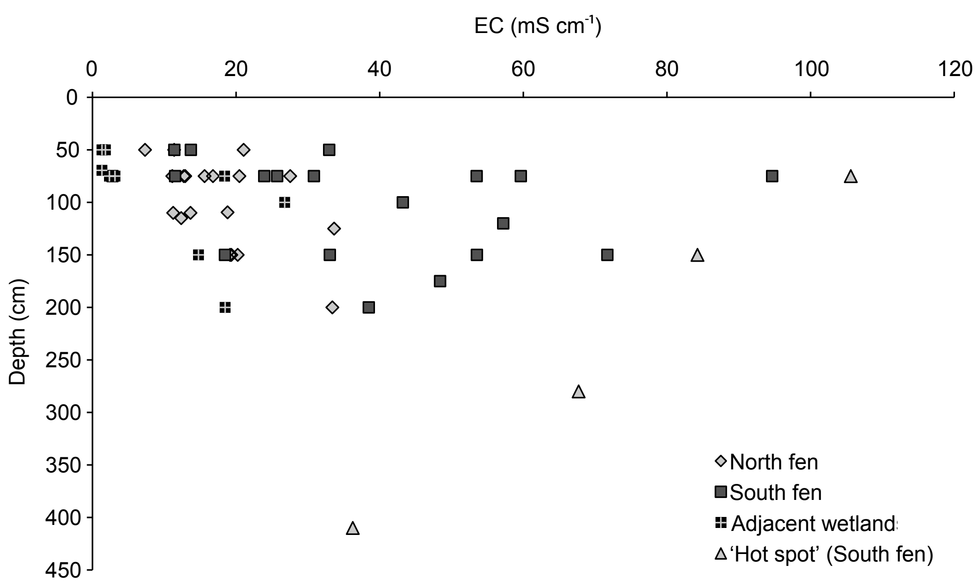

Figure 2-9 Electrical conductivity versus depth for the saline fen and adjacent wetlands.

Site-Scale Geochemistry

The saline fen had a site-scale average pH of 6.4. Over the two-year study, pH remained relatively stable and showed little variability across the fen or with depth (Table 2-4). Measurements of pond chemistry in 2012 did show that surface water features had a greater pH (mean pH of 7.8) than that of the surrounding groundwater. Considerable spatial variability in pond pH was also observed, with one pond in the south fen (Pilsner Pond) exhibiting an average pH of 4.5 (maximum of 6.3, minimum of 2.7).

Groundwater within the peat (≥ 50 cm depth), the underlying till and the entire pond network were dominated by Na^+ (195 to 25,680 mg L^{-1}) and Cl^- (1785 to 56,249 mg L^{-1}) and to a lesser extent SO_4^{2-} (28 to 3080 mg L^{-1}) (Table 2-5). Ca^{2+} and Mg^{2+} levels were also generally high throughout the fen (48 to 1696 and 25 to 2875 mg L^{-1}). In the north fen, Na^+ and Cl^- levels within the till were higher on average than in the overlying peat, while this trend was reversed in the south fen (Table 2-5). A distinct south-to-north decrease in dissolved salts was observed following the local groundwater flow path (Table 2-5). In the south fen, the region of elevated ion concentration was restricted to the regions bordered by Lager Pond to the north and South Pond 1 to the south. The ionic composition of the saline fen groundwater contrasted sharply with the freshwater north bog, where HCO_3^- formed the dominant anion, followed by Ca^{2+} and Mg^{2+} as the dominant cations (Table 2-5). In the wetlands bordering the fen’s eastern and western margins, major ion composition indicated that Na^+ and Cl^- also dominated the deeper peat layers despite a lower near surface groundwater EC and apparently freshwater vegetation cover (Table 2-4 and 2-5). Similar to what was observed in EC measurements at depth, major ion concentration spiked within the mineral till underlying the adjacent wetlands.

Table 2-5 Major ion and TDS concentrations for the saline fen (n=48) and adjacent wetlands (n=8) for the 2012 study season. Each wetland parameter is divided into peat (P) and mineral (M) substrates. Concentrations for each substrate type in the adjacent wetlands comprise a maximum of 3 samples and so only averages are provided. *Alk* is alkalinity. All concentrations are in mg L^{-1}.

		Saline Fen					Adjacent Wetlands					
		North Fen		South Fen		Pond	West Fen		East Bog		North Bog	
		P	M	P	M		P	M	P	M	P	M
Cl^-	Mean	5393	6943	26338	17799		306	10543	335	3895	4	3
	Max	10273	13193	56249	33884							
	Min	2647	2362	9491	4624							
HCO_3^-	Mean	179	135	203	132	25	71	65	80	99	397	347
	Max	391	241	627	318	84						
	Min	20	34	40	61	129						
SO_4^{2-}	Mean	80	96	1561	959	892	10	695	8		<1	3
	Max	157	157	3080	2741	418						
	Min	28	43	512	277	1590						
Ca^{2+}	Mean	183	243	721	534	48	24	650	30	157	36	22
	Max	262	435	1696	1235	150						
	Min	127	132	231	88	515						
Mg^{2+}	Mean	493	147	407	301	48	9	330	9	94	17	14
	Max	2875	299	903	758	73						
	Min	62	51	112	71	208						
Na+	Mean	2307	3491	13015	9090	4847	151	5148	167	1937	3	5
	Max	5285	6596	25680	15920	1838						
	Min	108	1153	5534	2644	6474						
Alk	Mean	147	111	166	108	52	59	54	65	81	325	285
	Max	321	197	514	261	69						
	Min	16	28	33	50	105						
TDS	Mean	8694	12155	42394	28827	42	629	17485	693	9961		
	Max	16844	21118	87971	54637	9364						
	Min	3916	4377	17547	8168	21117						

2.6 Discussion

General Hydrology of Saline Fen

The highly dynamic hydrologic and geochemical behaviour of the fen was largely a function of its structure and overall configuration within the landscape. The relatively thin peat profile (mean depth of 1.2 m) composed of dense sedge peat with a low Sy (mean of 0.05) meant that the fen's small subsurface storage capacity was readily exceeded under periods of sustained rainfall (Table 2-1, Figure 2-5). Water table fluctuations were exaggerated in response to wetting and drying across the fen but most notably south of Lager Pond where the peat profile thinned considerably and where the pond-pool network was most extensive (Figure 2-2, Figure 2-5). Rapid water table rises leading to episodic flooding occurred throughout 2012 and were largely restricted to the south fen until a series of intense storms in early July lead to a change in the fen's overall hydrologic regime. South of Lager Pond, water table response continued to follow its typical pattern of periodic inundation followed by steep recessions (Figure 2-5). The developed pond-pool network likely enhanced water table decline post rain events due to the function of open water features as evaporation windows in the subhumid Boreal Landscape (Smerdon et al., 2005; Devito et al., 2012). Sustained flooding within the fen's lower reaches was a result of the fen's position in the landscape and ridge-dominated microtopography that collects and detains surface water. Steeper gradients within the south fen supplied surface and subsurface flow northward following the local topography (Figure 2-1). Under baseflow conditions, the marked increase in slope south of Lager pond produced larger hydraulic gradients and greater subsurface flow rates. During periods of elevated water table, horizontal subsurface flow was enhanced due to the higher hydraulic conductivity of the upper peat layers (Figure 2-4). Low rates of groundwater exchange between adjacent wetlands limited water loss from the fen while the damming effect of the water table mound between the north fen and north bog effectively restricted shallow ground water outflow. Surface flow was restricted by the large storage capacity and damming effect of patterned microtopography. The ridge-depression pattern may be linked to groundwater and the associated accumulation of salts leading to changes in peat formation over time (Timoney and Lee, 2001). Similar to other studies, the ponds and pools became interconnected during very wet periods and functioned as a drainage network for surface flow northward, although the magnitude of surface flow was not quantified in this study (Price and Maloney, 1994). These infrequent and prolonged wetting events are likely important in the flushing and redistribution of solutes throughout the fen and the actual magnitude of solute flux warrants further exploration.

Over the two-year study, runoff and vertical groundwater recharge were negligible and *ET* represented by far the largest water loss at the fen. The synchronization of peak rainfall with maximum *ET* rates was observed for both study seasons, but its effect on the seasonal water deficit was most pronounced in 2011 (Devito et al., 2005a; Petrone et al., 2007; Brown et al., 2010; Redding and Devito, 2005). *ET* dynamics between peat microforms were variable between seasons in response to varying moisture regimes. Cumulative *ET* from ridge/lawns (332 mm) was slightly greater than from inter-ridge depressions (314 mm) in 2011 (Table 2-2). This could be explained by moisture availability, where the sparsely vegetated inter-ridge depressions were limited in their moisture supply due to lower-than-average precipitation. *ET* rates could be sustained along ridges and lawns due to their dense vascular growth that has a deeper moisture supply not significantly reduced until the water table drops below its rooting zone (Lafleur and Roulet, 1992). Higher *ET* rates because of greater vascular plant cover may also partly explain the steeper water table recessions found in the south fen compared to the north (Figure 2-5). *ET* dynamics between microforms in 2011 contrasts to what was observed in the wet 2012 study season. In a concurrent study looking into the spatial and temporal variability of *ET* among microforms at the saline fen, *ET* was slightly higher amongst depressions in May and June but greater in ridges during July and August (Phillips, 2013). This was attributed to changes in the dominant controls on *ET* over the growing season, where moisture availability in depressions allowed for greater *ET* rates despite inadequate vegetation growth during the spring. Following peak growing season in July and August, the more densely vegetated ridge sites dominated when the large water supply provided in July was coupled with increased transpiration. While a similar spring pattern was observed in this study, depression *ET* continued to exceed ridge *ET* for the entire study period. Higher *ET* rates in the inter-ridge depressions may be partly explained by their water storage characteristics and interaction with surrounding microtopography. The open water surface of the inundated inter-ridge depressions provides an ample moisture supply for sustained *ET* despite a lack of significant plant transpiration. Additionally, as a result of heat advection caused by adjacent peat ridges, evaporation can in some cases exceed the potential rate, allowing *ET* in water-filled depressions to exceed ridge/lawns (Price, 1991; Price and Maloney, 1994).

Hydraulic gradients between the saline fen and surrounding wetlands were spatially variable and sensitive to seasonal and short-term weather changes. In general, groundwater flow between the fen and adjacent wetlands was low and often restricted by the presence of transient groundwater mounds. The presence of groundwater mounds along fen margins mirrors the underlying substrate, where mineral ridges with low hydraulic conductivities create zones of higher water table (Figure 2-2, Figure 2-3a,b) (Scarlett and Price, 2013). Following a series of high intensity storms in July of 2012, flooding caused a rapid rise in water table in the north fen and saline water recharged the adjacent bog (Figure 2-3a). A similarly complex flow regime was observed at the western margin between the adjacent wetland and the south fen (transect C-C'). During wetting events, groundwater flow followed the local topographic gradient westward and recharged the adjacent wetland, but this pattern reversed under sustained drought conditions and the water table sloped against the surface topography (Figure 2-3b). Water tables that do not necessarily reflect topography are common to the western Boreal Plains region and support the notion that topographically derived catchment delineation may be invalid in such settings (Devito et al., 2005b). The sensitivity of groundwater flow to drought is similar to what has been observed in humid climates, where a sustained water deficit can initiate regular flow reversals in peatlands with shallow till and a reliance on local groundwater and precipitation (Devito et al., 1997). Local and seasonally transient flow reversals linked to large-scale wetting events, such as what was observed in the north fen, have also been recorded for other peatlands in the Boreal Plains region (Ferone and Devito, 2004). Despite these reversals, small *dh/dl* and the low hydraulic conductivity of the saline fen peat restricted the flux of shallow subsurface flow towards adjacent wetlands, which likely only occurs under atypically wet scenarios (Table 2-3). Low K_{LH} compared to K_{LV} is similar to some other studies of fen peat and is likely influenced by the vertical orientation of decomposing stalks and roots (Surridge et al., 2005; Kruse et al., 2008).

Ponds

Over the two-year study, the function and connectivity of the pond network was highly irregular. Following rain events some ponds would fill rapidly then dry gradually until the next wetting event, suggesting many are semi-permanent features that are sensitive to short-term weather changes. This was shown to be true for several ponds instrumented in 2012 and is similar to what has been observed for other shallow pond-peatland complexes in the WBP and northern Prairie regions (Ferone and Devito, 2004; van der Kamp and Hayashi, 2009). The influence of these open water ponds as net-loss features on fen hydrology is strong, as indicated by the high rates of evaporation for both years, and likely contributed to the flashy nature of water table response in the south fen.

Ponds that remained wet functioned as long-term water sinks, and the combination of groundwater and rainfall helped them to resist complete drying as a result of seasonal drought conditions and long-term water deficits typical for the region (Smerdon et al., 2005; Devito et al., 2012). A connection to shallow groundwater was reflected in some ponds by a rise in stage that exceeded precipitation depth for a given rain event or through a continual rise in stage for several days following a storm surge (Figure 2-9). The connection to groundwater is supported geochemically by salinity and major ion concentrations in their surface waters. Many ponds that remained saturated over the two-year study had far greater concentrations of dissolved salts than adjacent fen peat groundwater, although this was not always the case (Table 2-3, Table 2-4).

Typically, ponds and pools within groundwater discharge wetlands are areas of focused groundwater discharge. Variability in underlying substrate, such as the thinning of sand and gravel layers, often coincides with areas of groundwater discharge resulting in the pooling of surface waters (Lowry et al., 2009). Drilling and piezometer bail tests showed that a dense mineral till lined the majority of the smaller study ponds, with very low hydraulic conductivities that would restrict the vertical flux of groundwater (Figure 2-4). This type of surficial geologic setting is similar to ponds found in the northern Prairie regions, where clay-rich tills underline most wetlands (Hayashi et al., 1998; van der Kamp and Hayashi, 2009). Nevertheless, while it may be that the contribution of deep groundwater may have little effect on the ponds' water balance in the long term, even small amounts of groundwater input can have a strong influence on the ponds' mass balance of dissolved species (Hayashi and van der Kamp, 2007). This is reflected in the very high salinities of many of the ponds in the south fen. The horizontal exchange of groundwater between ponds and the surrounding peat was also shown to be highly variable within a given pond (Figure 2-7). Short-term pond flow reversals are common in Boreal Plain and northern Prairie regions and have important implications on pond water quality and ecology (Devito, 2004; Smerdon et al., 2005; van der Kamp and Hayashi, 2009). In the case of the saline fen, groundwater reversals and wetting/drying sequences may play an important role in the flux of salinity in pond surface waters. For a pond that functions largely as a store of groundwater, salinity increases over time as water is retained by the low hydraulic conductivity of the underlying till. As water levels decrease in response to seasonal drying trends, horizontal hydraulic gradients reverse and the pond functions as a temporary recharge feature. These small windows may provide an opportunity for ponds to regulate their salinity and may explain why some ponds maintain surface water salinities considerably lower than that of the surrounding peat.

The morphology and distribution of ponds throughout the saline fen is complex and their development can be attributed to both physical processes (e.g., fen structure, groundwater flow) and the effect of salt accumulation in the landscape. The saline fen fits well into the regional description of interior patterned saline marshes, where ridge-depression sequences develop over time on gentle gradients that initially promote the lateral flow of saline water (Timoney and Lee, 2001). The ridge-depression pattern found throughout the fen, with the orientation of ridges transverse to the local gradient, work to detain surface water where it pools. This is clearly demonstrated in the fen's northern reaches where the rough microtopography restricts surface flow despite flooding conditions. Irregularities in the sub-peat surface have also been suggested as a possible cause for the initiation of pool development and this may apply to some of the larger ponds at the fen (Moore, 1982; Belyea and Lancaster, 2002; Lowry et al., 2009). Over time, plant growth can be restricted in pooled depressions and peat accumulation ceases, while the oxygenated surface waters favor erosion and degradation processes (Pearsall, 1956; Foster et al., 1983). The formation of minor depressions results in the re-direction of shallow groundwater flow towards them during periods of high water table, leading to increased salt accumulation and depression development. The presence of heavily decomposed peat underlying many of the ponds suggests that they were at one time vegetated, while many of the inter-ridge depressions in the north fen have a very sparse vegetation cover compared to adjacent ridges.

Many of the ponds and pools at the fen are linked and the coalescing of surface water features can be reasonably explained by decomposition that deepens them and increases their extent over time. Eventually, ridges that separate individual ponds or pools are degraded, connecting multiple water features that form preferentially along the contour (Foster et al., 1983). The network appears to be concentrated within the central and southern portions of the fen and coincides with the areas of highest groundwater salinity (Figure 2-8). The development of a more distinct pond-pool network in this region may be the result of focused shallow groundwater discharge within the south fen (indicated by highest concentration of dissolved salts, Table 2-5) and the subsequent effect of salt toxicity on vegetation composition.

Salinity Distribution

The spatial distribution of solutes at the fen is counter to what would be interpreted by the local topographic gradient; elevated ground and surface water salinities occur at the topographic high point of the fen. While vertical groundwater flux patterns through the underlying till are difficult to interpret due to the possible inaccuracies associated with the piezometer network (e.g., lag-time errors), the overall model of focused saline groundwater discharge south of Lager Pond and the re-distribution of saline groundwater northward is supported by the spatial patterns in ground and surface water geochemistry (Figure 2-9, Table 2-3 and 2-4). The magnitude of both salinity and almost all major ions increased substantially over a short distance southward and the location of Lager Pond roughly marked the center point that bisected the system into two distinct geochemical sections (Figure 2-8, Figure 2-10). Elevated ground and surface water salinities and vertical profiles of decreasing concentrations of dissolved salts with depth are also consistent with the gradual upward migration and concentration of solutes near the surface (Figure 2-9, Table 2-4). This was supported by the presence of salt surface crusting that was restricted to the fen's southern half. The fen's overall structure promoted the lateral movement of saline groundwater northward and the gradual reduction in slope north of Lager Pond worked to retard subsurface flow where it was diluted by precipitation and recharged the surficial aquifer (Figure 2-2, Figure 2-9). Vertical hydraulic gradients were variable in the north fen but generally positive and the concentration of Na^+ and Cl^- in groundwater were on average lower in the peat than the till (Table 2-4). This increase in dissolved salts with depth is indicative of the downward flux of groundwater and influence of dilution by precipitation and high water tables. The steady northward migration of saline groundwater over time is supported anecdotally by a successive southward decrease in black spruce health along the local groundwater flow path; complete mortality has occurred for all trees within the fen while trees towards the outermost margins show symptoms of salt toxicity (cf. Redfield, 2001). The source of saline groundwater and the fen's function as a conduit for deep groundwater discharge forms the focus of a future paper.

Water Budget

While there is no absolute measure of error on the individual components of a water balance, the largest probable source of error at the fen over the 2012 thaw season was the estimation of evapotranspiration. The Priestly and Taylor (1972) method is accurate to within ±15% under ideal conditions. This error is compounded by impacts related to the accuracy of lysimeter measurements (e.g., weighing errors, periods of rainfall, duration between measurement periods) (Allan et al., 1991) and the definition of microform surface area for the calculation of site-scale α values. Due to the remoteness nature of the saline fen, extended periods of time took place between lysimeters measurements and periods of rainfall were not excluded. A relatively small sample size for lysimeter measurements could also have led to inaccurate estimations of *ET* at the fen for both years, especially since seasonal variation could not be accounted for. The Penman (1948) formula for open water evaporation has been estimated as having a probable error of ~8% over monthly time-steps (Linacre, 1993). Because the meteorological station was not located over or adjacent to a pond, wind speed, irradiance and other factors not representative of the true pond surface had to be used for the estimation open water evaporation. However, this error was likely reduced for much of the 2012 season when the north fen was flooded, mimicking more closely the conditions of the ponds. Nevertheless, the Penman formula does not take into account pond heat storage and assumes that net radiation is the driving force behind evaporation (Finch and Hall, 2001). This can lead to a de-coupling of *E* with net radiation both seasonally and diurnally, where rates of *E* can be higher at night and during autumn months due to the gradual release of energy from storage (McCuen and Asmussen, 1973; Hayashi and van der Kamp, 2007; Finch and Calver, 2008). It has been observed that this effect can be sometimes ignored in ponds with depths of less than 0.5 m, which is deeper than many of the ponds measured in this study (Finch and Calver, 2008).

Errors in estimating shallow groundwater inputs and outputs are large (as high as 70%, Ferone and Devito, 2004) and the calculation of margin ground and surface water flux at the fen may have been oversimplified. Preferred flow zones with hydraulic conductivities greater than what were estimated may have been missed, most notably during periods of elevated water table. Moreover, while surface runoff between the fen and adjacent wetlands was not observed, it is possible that some surface water exchange occurred during periods of inundation. Thus, underestimations in GW_{out} and surface runoff, particularly along the fen's western sloping margins (Figure 2-3a,b), may also lead to an underestimation of total outputs and an exaggerated storage term (Table 2-3). Nevertheless, despite flooding and complex shallow groundwater interactions between the saline fen and adjacent wetlands, the true flux of ground and surface water to and from the fen is likely still low compared to *P* and *ET*. This is supported by the sharp decrease in groundwater salinity and increase in freshwater vegetation composition along the fen margins (Figure 2-8, Table 2-4 and 2-5). A comparison between the fen tipping bucket used in this study and the north bog tipping bucket located ~500 meters away showed a measurement discrepancy of 26%, with the fen measuring 74 mm more rain then the bog. Based on this, precipitation could be off by as much as 80 mm, indicating that ET (300 mm) may have exceeded rainfall input for 2012. However, the true magnitude of error is likely smaller than what is suggested by the tipping bucket discrepancy, as it was clear that the fen received sufficient rainfall over the study (Figure 2-5). In addition, differences in canopy cover and tipping bucket placement between the bog and the fen may account for some variation between measurement totals. An experiment investigating variability in tipping bucket measurements was conducted within a small pond-peatland complex at the Utikima Research Study Area (URSA) in the WBP of northern Alberta. Between four tipping buckets located around the perimeter of two ponds measuring precipitation simultaneously between June 1st and August 31st 2005, a difference of 92 mm of rain was recorded between the maximum and minimum totals (mean of 324 mm, standard deviation of 44 mm) (Scott Brown, personal communication). Despite the above potential errors, the site's overall water budget clearly indicates that precipitation and evapotranspiration are the dominant inputs and outputs for the fen, followed by shallow subsurface horizontal groundwater exchange with adjacent wetlands. This type of hydrologic setting is common for wetlands in the subhumid climates of the WBP and northern Prairie regions.

2.7 Conclusion

Similar to other saline wetlands throughout northeastern Alberta, the saline fen exhibited a distinct net-patterned microtopography built on a gradual sloping plain, both of which influenced the distribution of salts. While a general trend of decreasing salinity northward was observed, the link between hydrologic function and geochemistry was not always clear. Estimated vertical groundwater discharge was negligible due to the presence of a dense mineral till, however, vertical hydraulic gradients corresponded weakly to observed ground and surface water salinity patterns. The absence of significant deep groundwater input at the fen suggests that shallow groundwater processes dominate its water balance. The accumulation of salts may be the result of slow but consistent discharge gradients over the long-term, that work to prevent the loss of salts from the fen. It is possible that the measured flux of deep groundwater input at the fen could be influenced by the effects of piezometer construction/response in dense tills and/or because of locally conductive zones that were not captured. The function of some of the larger ponds as possible discharge windows warrants a more detailed investigation. Nevertheless, this study shows that estimating the true magnitude of deep groundwater input in systems underlain by dense tills is complex, and makes establishing a baseline of the true magnitude of natural saline discharge a challenge in the AOSR.

Precipitation was by far the greatest input to the fen over the entire study period and, when combined with a small subsurface storage capacity, led to a rapid change in the fen's hydrologic regime under periods of sustained rainfall. The extensive pond network plays a dual role in the hydrologic function of the saline fen. The fill-and-spill mechanism of the ponds during flooding events or periods of elevated water table (e.g., spring) created an interconnected surface drainage network that transmitted surface flow and thus dissolved salts from the south fen northward. During dry periods, however, the pond network functioned as a strong sink of groundwater (i.e., evaporation windows) and did not contribute to the fen's long-term average moisture surplus. Despite flooding within the fen's lower reaches in 2012, runoff was not observed due to the damming effect of ridges and the large surface storage capacity of the inter-ridge depressions. The rough surface features may serve as a self-limiting mechanism, developing over time as they detain surface waters while slowing the conveyance of water northward and thus the overall extent of the saline fen. In this way, total salinity is re-distributed within the wetland but retained within its boundaries.

The pronounced boundary between adjacent freshwater wetlands and the saline fen was a result of a low net flux of ground and surface water. The net-pattern of ridges and inter-ridge depressions effectively prevented the lateral movement of saline surface waters into adjacent freshwater peatlands during periods of elevated water table, which occurred often due to a low subsurface storage capacity. Below the surface, groundwater exchange between wetlands was variable but generally limited and often restricted by the presence of transient groundwater mounds. The dynamic interaction between wetlands in response to seasonal and short-term weather changes, notably when water tables slope against the ground surface during extended dry periods, supports the observation that topography may be an inappropriate metric for catchment delineation in the Boreal Plains (Devito et al., 2005).

This study highlights for the first time the characteristic hydrologic and geochemical functionality of a saline wetland in the oil sands region of northeastern Alberta. Due to the likelihood of increased salinity within the post-mined landscape, natural saline fen ecosystems constitute potential reference analogues for oil sands reclamation. Furthermore, due to the impact of saline groundwater discharge on water quality, features like the saline fen should be incorporated into monitoring networks to better quantify the amount of naturally occurring saline discharge in the region. Developing an improved understanding of saline systems will aid in the management of an important ecological resource while supporting compliance with environmental policies related to oil sands extraction and reclamation.

3.0 Manuscript 2: Origin and flow history of a saline spring fen peatland in the Athabasca oil sands region of Alberta

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3.1 Overview

Saline springs can be found along the northeastern margin of the Western Canadian Sedimentary Basin, where regional and local scale flow systems that come into contact with near-surface evaporite beds discharge to the surface. Understanding the function and connectivity of these discharge features can provide important clues into the nature of groundwater flow, including the origin and flow history of brines and the link between the occurrence of saline spring wetlands and subsurface wastewater containment. A low-flow saline spring fen peatland located adjacent to a proposed in-situ oil extraction facility was examined south of the oil sands hub of Fort McMurray, Alberta. Hydrologically disconnected from underlying Devonian deposits, a saline groundwater plume originating from a Lower Cretaceous aquifer (the Grand Rapids Formation) was identified as a likely source for the accumulation of Na^+ (mean of 6,949 mg L^{-1}) and Cl^- (mean of 13,766 mg L^{-1}) at the fen surface. Deep groundwater flux was estimated to be very low due to the presence of low-conductivity clay till, and based on ground and surface water sampling, appeared to be largely restricted to a relatively small region in the fen's southern sector. The anomalously high concentration of dissolved salts within the surrounding wetlands and river systems indicate that the actual extent of saline groundwater discharge may be much greater than what occurs in the saline fen. The geochemical signature of fen groundwater points to halite as a source of salinity, as indicated by Cl/Br ratios in excess of 7,000. This is in contrast to what has been observed for regional formation brines that are typically related to evaporated seawater, with Cl/Br ratios closer to 250. Isotopic evidence in ground and surface waters similar to the regional average for precipitation and near-surface aquifers combined with relatively low salinities compared to springs in the Wood Buffalo region suggests that fen source water may be significantly diluted as a result of mixing with freshwater flow systems. Fen water geochemistry points to a complex fluid history where elevated salinities can be explained, in part, by evaporite dissolution. The contribution of evaporite to discharge water may be coming from somewhere deeper and further south in the basin. This has important implications for the disposal of wastewater by deep well injection, as disposal zones may be hydrologically linked to near-surface aquifers and discharge features well beyond the immediate production and storage area. The localized accumulation of salts in this region is indicative of the long-term discharge of groundwater and suggests that larger-scale flow systems may play an important role in ecosystem function. Further investigation into the saline plume south of the saline fen study area may provide more evidence of the origin and flow history of saline brines in the Athabasca oil sands region.

3.2 Introduction

Saline springs have been observed historically throughout the western Boreal Plains region, with the more recent study of the origin and connectivity of spring discharge (Grasby and Chen, 2005; Grasby et al., 2006; Grasby and Londry, 2007) and their influence on nearby rivers well documented (Hitchon, 1969a; Jasechko et al., 2012). While discrete, high-salinity springs have been studied in detail in Alberta's Boreal Plains, less is known about the hydrologic function of saline spring wetlands, whose discharge tends to be more diffuse (Scarlett and Price, 2013; Manuscript 1). Understanding the function and connectivity of these unique discharge features can provide important clues into the nature of groundwater flow in the Athabasca oil sands region (AOSR), including the origin and flow history of brines and the link between springs and subsurface wastewater containment (Carrigy and McLaw, 1973; Hackbarth and Nastasa, 1979; Gordon et al., 2002; Grasby et al., 2006; Gupta et al., 2012). Furthermore, as a result of unprecedented rates of industrial development within the oil sands landscape, the role wetlands play is becoming increasingly compromised (Johnson and Miyanishi, 2008; Ferone and Devito, 2004) and likely to be exacerbated by the predicted impacts of climate change (Stocks et al., 1998; Li et al., 2000). Generating hydrological data on a variety of wetland types contributes to the management of critical ecosystems while supporting compliance with environmental policies related to oil sands development and reclamation.

In the AOSR, saline springs are primarily found where groundwater discharges through subcropping or exposed Paleozoic carbonate rocks in the low-lying plains and Athabasca and Clearwater rivers (Grasby et al., 2006). These types of springs are karstic in origin due to the soluble nature of the carbonate rocks and the extensive dissolution that has occurred throughout evaporite deposits that lie close to the surface (Borneuf, 1983; Ford, 1998). Sodium chloride springs make up 4% of the springs in Alberta and are found exclusively in the northeastern region of the province (Borneuf, 1983), where the composition of brines represent a mixture of geochemical end members related to seawater, evaporite dissolution and freshwater mixing (Rittenhouse, 1967; Connolly et al., 1990a; Michael et al., 2003; Gupta et al., 2012). Connolly et al., (1990a), in their study of the origin and evolution of formation waters in the Alberta basin, found that Devonian through Cretaceous reservoirs were all derived from subaerially evaporated brines and not influenced by evaporite dissolution. Similar interpretations were made by Lemay (2002) around the Fort McMurray area based on showed Lower Cretaceous aquifers having a diluted seawater composition with no indication of evaporite. In a study by Michael et al (2003), geochemical evidence pointed to the dissolution of halite as a source for salinity in deep basinal brines. Within the Fort McMurray and Wood Buffalo region, isotopic evidence and major ion composition of both high-and low-flow springs indicated that discharge water was not related to evaporated brine but was of meteoric origin, with elevated salinity a result of contact with buried evaporite beds, namely halite (Grasby, 2000; Last and Ginn, 2005; Grasby et al., 2006; Grasby and Londry, 2007; Berard et al., 2013).

Saline springs typically consist of an observable discharge outlet connected directly to Paleozoic era Devonian carbonates that either subcrop or are exposed completely to the surface (Hitchon et al., 1969; Grasby et al., 2006). In many cases these springs are found in close proximity to the banks of the Athabasca and Clearwater Rivers, where erosion has deeply incised into the exposed carbonate rocks (Ozoray et al., 1980; Grasby et al., 2006; Jasechko et al., 2012). In contrast to the typical hydrogeologic setting of saline springs in the AOSR, Wells (Manuscript 1) studied the site-scale hydrology of a saline spring wetland complex immediately south of Fort McMurray, whose surficial aquifer was isolated from underlying evaporites. Nearby river systems were over 12 km north of the saline spring wetland and over 200 m of Quaternary and Lower Cretaceous sediments covered Devonian evaporite beds that were hydrologically disconnected from the surface (Hackbarth and Nastasa, 1979, Value Creations, Inc., 2012). Instead of receiving saline groundwater directly from underlying evaporite beds, a typical feature of saline springs in the region, a plume from a Lower Cretaceous aquifer (the Grand Rapids Formation) discharging at its erosional limit south of the fen was identified as a likely source for the accumulation of dissolved salts at the surface (Figure 3-1) (Value Creations Inc., 2012). Examining the geochemistry of a saline spring wetland hydrologically connected to a near-surface aquifer provides an opportunity to learn more about the origin and flow history of brines in the Alberta Basin. Furthermore, owing to the expansion of in-situ projects within the AOSR, considerable industry attention has been given to the subject of wastewater disposal and the potential for leakage and the upward migration of disposal fluids via spring discharge (Carrigy and McLaw, 1973; Hackbarth and Nastasa, 1979; Gordon et al., 2002). The gravity of this issue is particularly relevant to the saline fen that forms the basis of this study, where a steam assisted gravity drainage (SAGD) pilot plant aims to inject wastewater within the subsurface. In order to assess the potential impact of saline spring wetlands as conduits for subsurface wastewater discharge, a better understanding of their hydrology and linkage to deep groundwater is required. With this in mind, the specific objectives of this study are to (i) identify and describe the function of a saline spring wetland located near an in-situ extraction site and investigate its potential as a conduit for deep groundwater discharge; and (ii) link site-scale geochemistry to regional-scale fluid flow to identify the connection of discharge water to underlying formation waters.

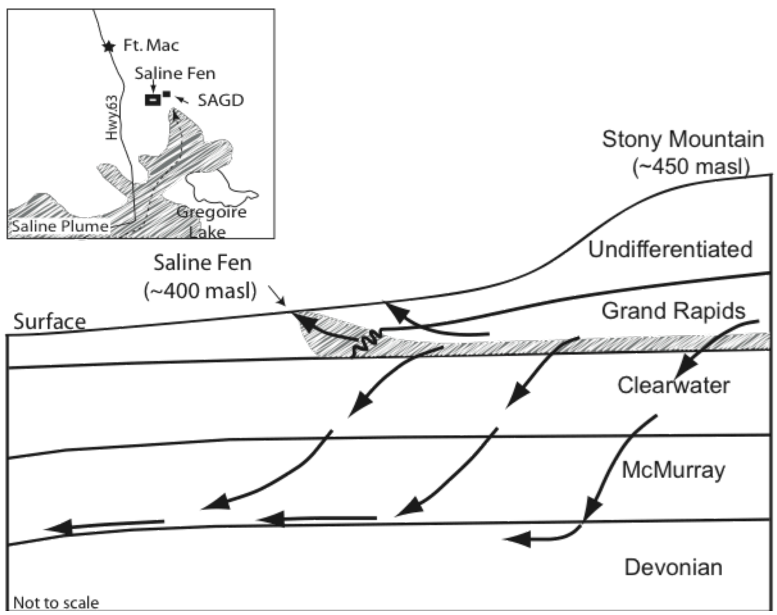

Figure 3-1 Simplified hydrogeologic cross section of the local groundwater regime and saline plume (grey hash marked region) for the study area. The inset map shows the approximate location of the saline plume and its overall flow direction, which is northward. Groundwater flow through the Grand Rapids formation is upward to the surface through the glacial drift where it crops out south of the fen and downward into the Clearwater caprock (adapted from Value Creations Inc., 2012).

3.3 Site Description and Hydrogeologic Setting

The saline fen (56°34'28.84" N, 111°16'38.39" W) is located approximately 10 km south-southeast of the AOSR hub of Fort McMurray, Alberta, Canada (Figure 3-2), and is described fully in Manuscript 1 of this thesis. Adjacent to the saline fen (~1 km east) is the Tristar steam-assisted gravity drainage (SAGD) pilot project, a three-year operation aiming to test the reliability of the SAGD in-situ process and demonstrate caprock integrity of the area (Figure 3-2). At peak production it will produce 1,000 barrels of crude oil per day from the McMurray formation, which will also serve as the disposal reservoir for process-affected wastewater (Value Creations Inc., 2012).

The fen lies at approximately 400 masl within the McMurray lowlands subdivision of the Dover Plains, a relatively flat region characterized by widespread, continuous and thick organic deposits (Andriashek, 2003). Several prominent upland features roughly encircle the broad lowlands surrounding the Fort McMurray area, the closest of which is the Stony Mountain complex approximately 15 km south of the saline fen (Hackbarth and Nastasa, 1979). Flowing northwestward from the Stony Mountain uplands through the study area are the primary tributaries of the Hangingstone River; the Saline River, Prairie Creek and Salt Creek basins (Value Creations Inc., 2012). Comprising an area of approximately 27 ha, the saline fen lies adjacent to several other large peatland complexes of similar salinity and configuration. The surface elevation of the fen declines northward and is characterized by a steep gradient in the south (~8 m km^{-1}) that transitions into a gently sloping plain in the north half of the fen (~2.7 m km^{-1}) (Figure 3-2). A large pond network comprises 19% of the fen surface. Salinity is high in the fen's southern limits and within this area a number of spring features are found, including salt surface crusting, halophytes and extremely saline 'hot spots' devoid of vegetation (Manuscript 1). Around the saline fen study area, clay-rich glacial till is thin (15-20 m) and overlies Cretaceous shale and sandstones of the Clearwater and bitumen-bearing McMurray formations (Figure 3-3) (Andriashek, 2003; Value Creations Inc., 2013). Devonian carbonates underlie the McMurray formation at a depth of approximately 180 m and are a possible source of saline groundwater at the fen due to the presence of extensive

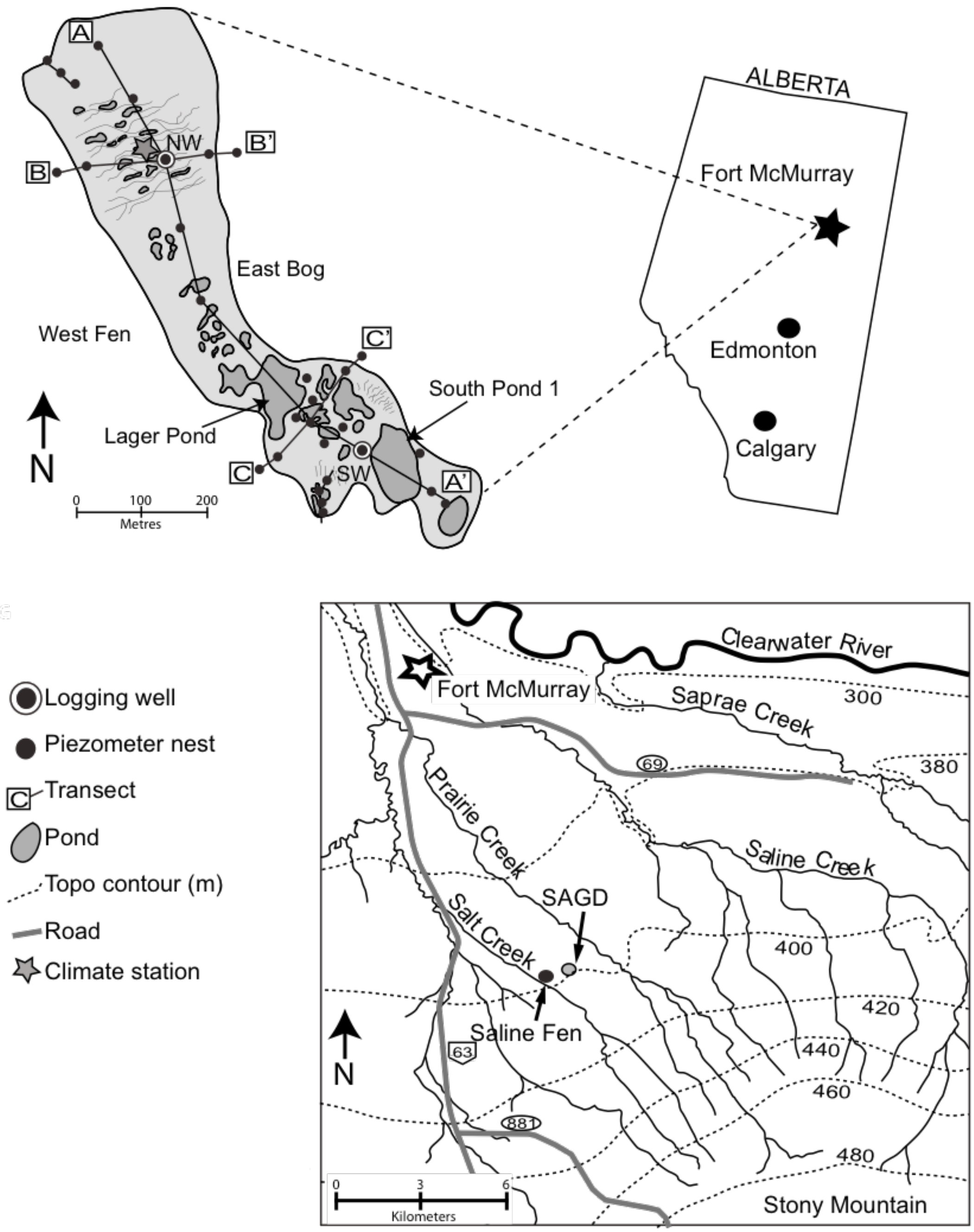

Figure 3-2 Local and regional map of the saline fen study area. Note the location of the adjacent SAGD facility.

halite deposits found within adjacent strata. The Lower Cretaceous Grand Rapids formation typically forms the bedrock surface for most of the Dover Plains but extensive erosion has truncated the formation south of the study site.

Regionally, the AOSR is located on the northeastern margin of the Alberta Basin, a sub-basin of the Western Canadian Sedimentary Basin that forms a simple sedimentary wedge resting unconformably on buried Precambrian rocks of the Canadian Shield. To the west, the sedimentary package exceeds 5,700 m but thins to a northeastern zero-edge where the Precambrian Shield becomes exposed as a result of depositional thinning and erosion (Grasby and Chen, 2005; Connolly et al., 1990a). The simplified modern-day hydrodynamic regime of the Alberta Basin is predominantly south to north through Cambrian sandstones and Devonian through Mississippian carbonates. Thick sequences of shale and silt units form regional aquitards through most of the overlying Mesozoic strata; however, interbedded sandstones form local and in some cases regional scale aquifers (Grasby et al., 2006). Throughout most of the Alberta basin, local and regional scale flow process are modified by highly permeable Upper Devonian and Carboniferous carbonate rocks that act as a low-fluid potential

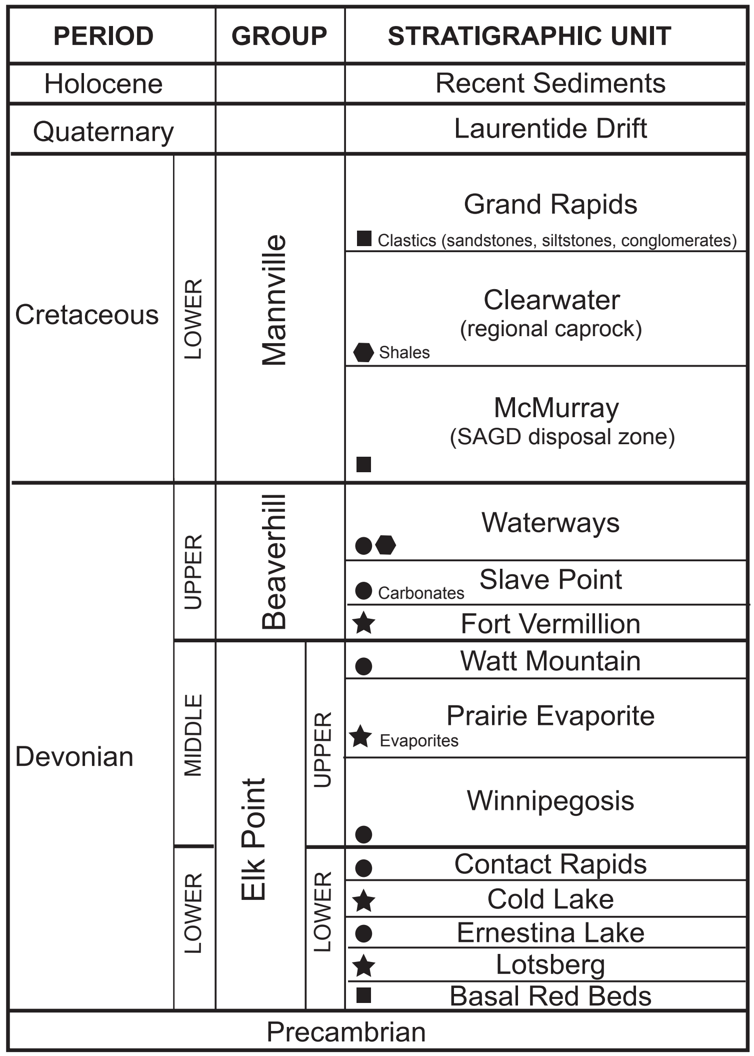

Figure 3-3 Simplified stratigraphic chart for the study area. Symbols underneath each unit refer to its primary lithology. The Lower to Middle Elk Point group is typically the source of dissolved salts at the surface in northeastern Alberta. Note the location of the McMurray Formation, which is the proposed disposal zone for the adjacent SAGD facility.

drain, channeling flow from the entire basin northward where it discharges in the AOSR (Hitchon, 1969). Considerable postglacial erosion within the AOSR has exposed Paleozoic strata to atmospheric conditions and, in general, formation waters that follow the basin-wide trend begin to show a modification in their flow paths due to atmospheric exposure and the influence of topographic and physiographic features (Bachu et al., 1993). Regional-scale fluid flow follows a northeast direction while near-surface Cretaceous aquifers show strong local flow characteristics and correspondingly low salinities caused by the influx of meteoric waters.

3.4 Methods

Groundwater

Nests of wells and piezometers were installed in four transects across the fen and into adjacent wetlands by the use of hand augers (30 total nests; Figure 3-2). For installation into the underlying till or into peat that was particularly dense, a mallet was used to drive the pipes the remaining distance. Polyvinyl chloride pipes (2.5 cm I.D) were used for the piezometer nests with each nest consisting of a depth-integrated well and two to five piezometers with 17 cm slotted screens covered in well sock. Piezometer depths varied according to peat thickness but were usually centered at 0.50 and 0.75 m within the peat and between 1.0 and 3.0 m within the till. Nests located within zones of very high salinity had additional piezometers set a depths ranging from ~3.0 to 4 m. Loggers were installed in two wells at opposite ends of the fen to measure water table continuously (A-A' transect, Figure 3-2) and manual measurements were taken at least once per week between June and September in 2011 and April and September in 2012. Fen topography and pipe top elevations were measured and referenced to sea level using a dual frequency survey-grade GPS. Field estimates of horizontal saturated hydraulic conductivities were measured using bail tests for 30 piezometers in peat and 11 piezometers in the till using the method of Hvorslev (1951). For depths <50 cm, the modified cube method was used on samples processed in the laboratory for the determination of horizontal and vertical saturated hydraulic conductivities (Beckwith et al., 2003).

Vertical fluxes rates were determined by using the vertical hydraulic gradient (*dh/dz*) between the piezometers in the till and the water table measured within an adjacent well. Due to elevated groundwater salinities within piezometers, hydraulic heads had to be corrected for differences in density and converted to freshwater equivalents before vertical gradients and fluxes could be calculated. A full description of the methodology can be found in Manuscript 1 of this thesis. The conversion of saltwater head to freshwater head resulted in an average increase in hydraulic head of 3 cm and conversions were largely restricted to only a few nests in the high-salinity zones within the south half of the fen. Groundwater with less than 10,000 mg L^{-1} total dissolved solids (TDS) at temperatures below 100°C are considered to have densities comparable to freshwater (1000 g L^{-1}), therefore, density corrections were made only on piezometers that exceeded this TDS threshold (Freeze and Cherry, 1979).

Geochemistry

Field sampling methods and laboratory protocols for the determination of major ion concentrations, alkalinity and total dissolved solids are described in full in Manuscript 1 of this thesis. Bromide was determined using neutron activation analyses at the University of Alberta's SLOWPOKE Nuclear Reactor Facility. Saturation indices were calculated in PHREEQC using the Pitzer aqueous model for high salinity waters (Parkhurst and Appelo, 2013). Stable isotope data were determined at the University of Waterloo Environmental Isotope Laboratory. $\delta^{18}O$ was determined by CO_2 equilibration measured on an IsoPrime continuous flow isotope ratio mass spectrometer system (CF-IRMS). Deuterium (δ^2H) was measured using hydrogen gas produced by chromium reduction. Highly saline samples were pre-processed using azeotropic distillation for the determination of $\delta^{18}O$ and δ^2H following the methods of Dewar and McDonald (1961). Results for $\delta^{18}O$ and δ^2H are referenced to the Vienna Standard Mean Ocean Water (SMOW) standard. Analytical errors for isotope analyses were estimated to be ± 0.2 ‰ for $\delta^{18}O$, ± 0.8 ‰ for δ^2H.

Local and Regional Hydrogeology

Hydrogeology and surface and formation water chemistry data for the study area was compiled from field assessments conducted for the TriStar Pilot Project by Value Creations Inc., (2012) and the Alberta Geologic Survey (Stewart and Lemay, 2011). Local geology was determined by drilling supplemented with borehole and petrophysical analyses. Groundwater direction was estimated by pressure tests and transducers installed in wells while formation water chemistry was determined by calibrating open-borehole resistivity logs to measurements of total dissolved solids.

3.5 Results

Hydraulic Conductivity

Field estimates of horizontal saturated hydraulic conductivities (K_H) for peat and mineral are shown in Figure 3-4. The geometric mean K_H of the fen peat was 2.6×10^{-4} cm s^{-1}, or 22 cm d^{-1} (n = 36) with considerable range both spatially across the fen and with depth. A geometrically averaged field estimate of mineral K_H was found to be 5.5×10^{-7} cm s^{-1}, or 0.05 cm d^{-1} (n = 11). Spatially, mineral K_H varied between 10^{-5} and 10^{-8} cm s^{-1}, with the highest K_H observed in the high salinity zones in the fen's southern section.

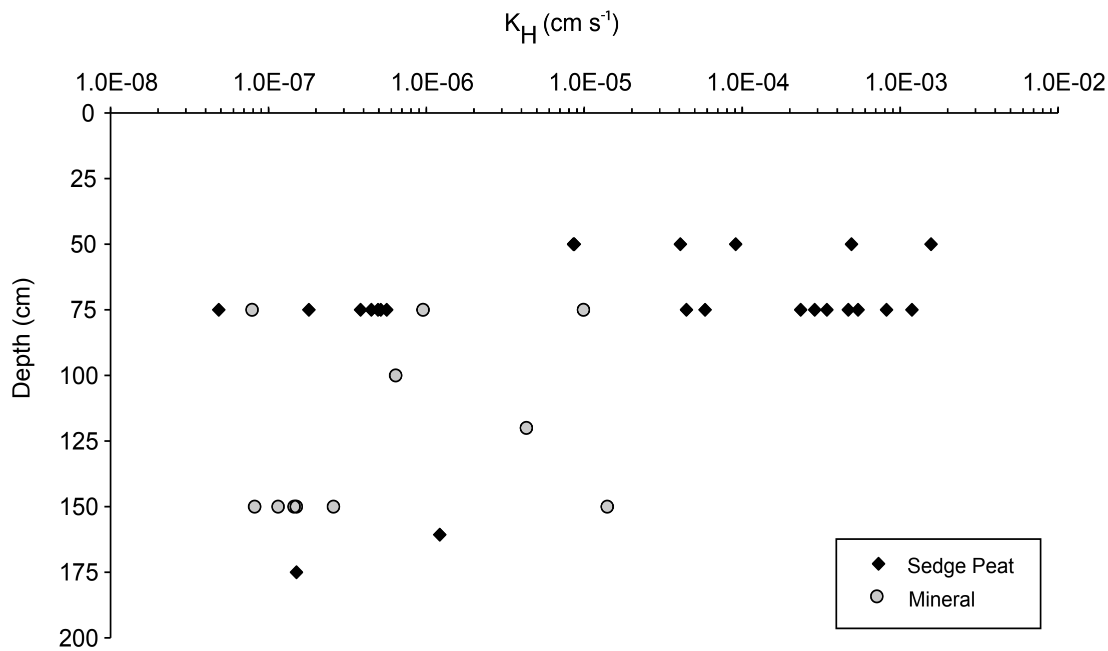

Figure 3-4 Field (>50 cm depth) estimates of saturated horizontal hydraulic conductivities for peat and mineral till.

Deep Groundwater

In general, vertical hydraulic gradients (*dh/dz*) between the underlying till and peat varied spatially and temporally in response to seasonal and short-term weather changes. Due to the low hydraulic conductivity of the underlying mineral till, *dh/dz* were interpreted with caution due to the possibility of piezometer lag-time errors that may have artificially indicated discharge conditions in response to changing boundary conditions. This was found to be true for many nests throughout the fen. Consequently, these were discarded for the analyses. Nevertheless, based on hydraulic conductivity estimates of the mineral till (mean $5x10^{-7}$ cm s^{-1}), the average flux of deep groundwater to or from the fen was estimated to be very low. For example, based on an average recharge (0.02) and discharge (-0.05) *dh/dz* for the fen over a 110-day study period in 2012, vertical flux through the mineral till was 2 and 3 mm, respectively. While the function of some of the larger ponds as discharge windows could not be confirmed, piezometers located within several smaller high salinity ponds did not show consistent discharge conditions during the 2012 season when the ponds were instrumented. To determine if time-lag errors between deep piezometers and the water table were ubiquitous across the site, individual nests were investigated more closely. In most cases, absolute change in hydraulic head decreased with depth across the fen. However, for some nests changes in hydraulic head within deep piezometers (i.e., within the till or basal peat) closely matched changes recorded at the water table, but in many cases showed groundwater recharge conditions (Figure 3-5c). For other nests with similar response in deep piezometers (e.g., NS+1000), discharge conditions were often consistent despite fluctuations in water table (Figure 3-5b). This pattern contrasted with other piezometers at different nests that showed a limited response to water table change (e.g., NS+920). In such cases, hydraulic heads continued to increase despite site-scale drying trends (e.g., between DoY 141 and 163), although seasonal trends could not be clearly determined since many piezometers did not recover fully following groundwater sampling (Figure 3-5a). No clear pattern was found between nest location and hydraulic head response to changing boundary conditions.

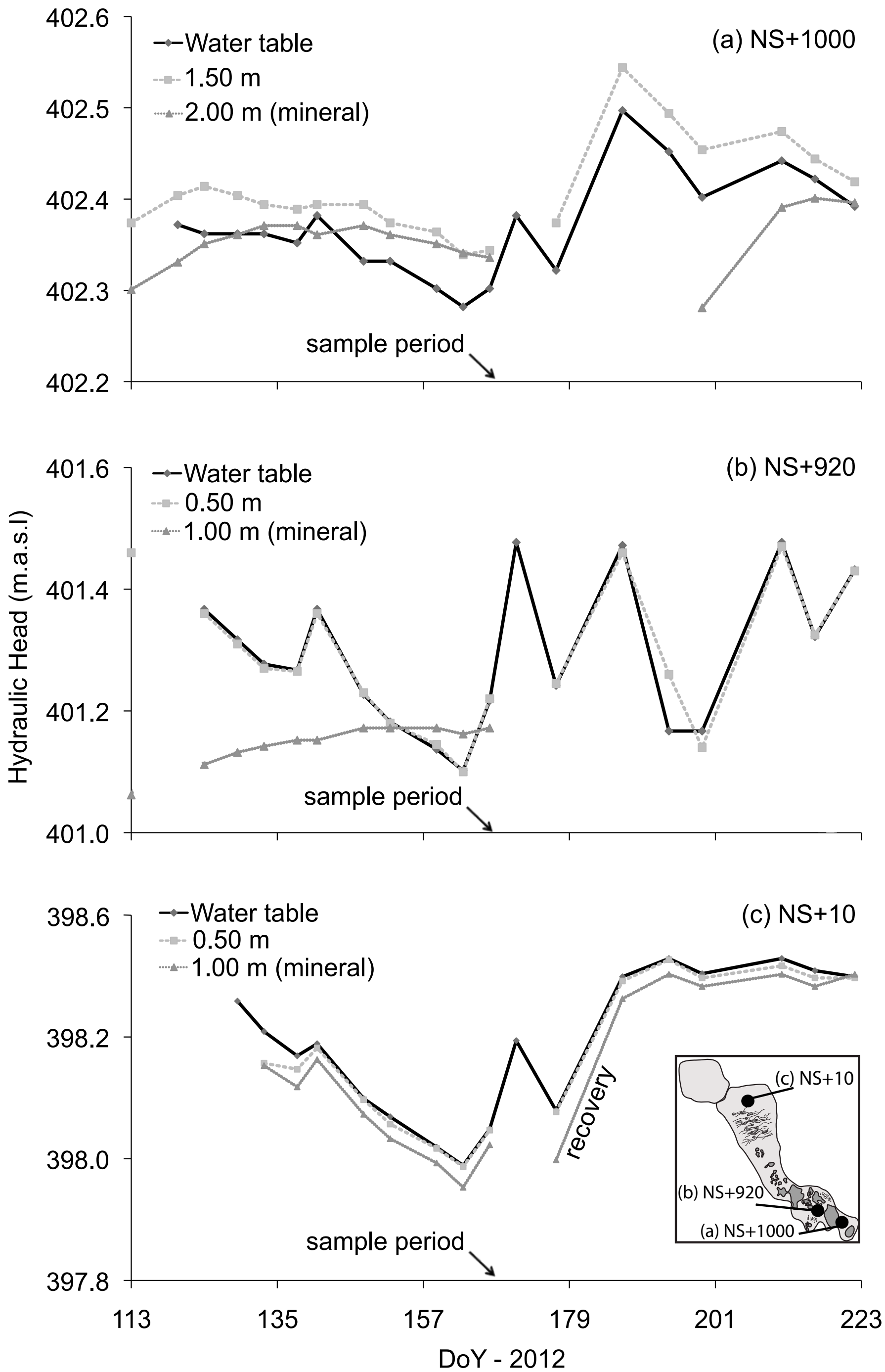

Figure 3-5 Comparison of absolute change in hydraulic head between deep piezometers and the water table for select nests for the 2012 study season (see inset map). (a) deep piezometer closely matches water table change and discharge is consistent throughout the season; (b) change in hydraulic head with depth is reduced and is less sensitive to drying trends: (c) deep piezometers closely match water table change but show consistent recharge over the season

Salinity Distribution

The total dissolved solids (TDS) content of ground and surface waters varied widely across the fen, from slightly brackish (2,893 mg L^{-1}) to hyper-saline (87,971 mg L^{-1}) (Table 3-1). For the entire fen, average groundwater TDS was 22,230 mg L^{-1} and TDS was only slightly greater within the fen peat on average (mean 26,668 mg L^{-1}) compared to the underlying mineral till (mean 24,464 mg L^{-1}). Surface water features (i.e., ponds and pools) were typically less saline than the adjacent peat, with some exceptions in the fen's southern region (Table 3-1). Along the primary transect (A-A'), TDS increased steadily southward. A sharp rise in TDS was observed in ground and surface waters in the relatively small area bordered by South Pond 1 and Lager Pond (Figure 4-6). Upslope of South Pond 1, TDS decreased sharply but remained generally high, increasing again towards the fen's southern-most margins. While the adjacent freshwater wetlands were generally less saline then the saline fen, TDS concentrations were still high within the underlying mineral till, most notably within the western fen (17,485 mg L^{-1}) (Table 3-1).

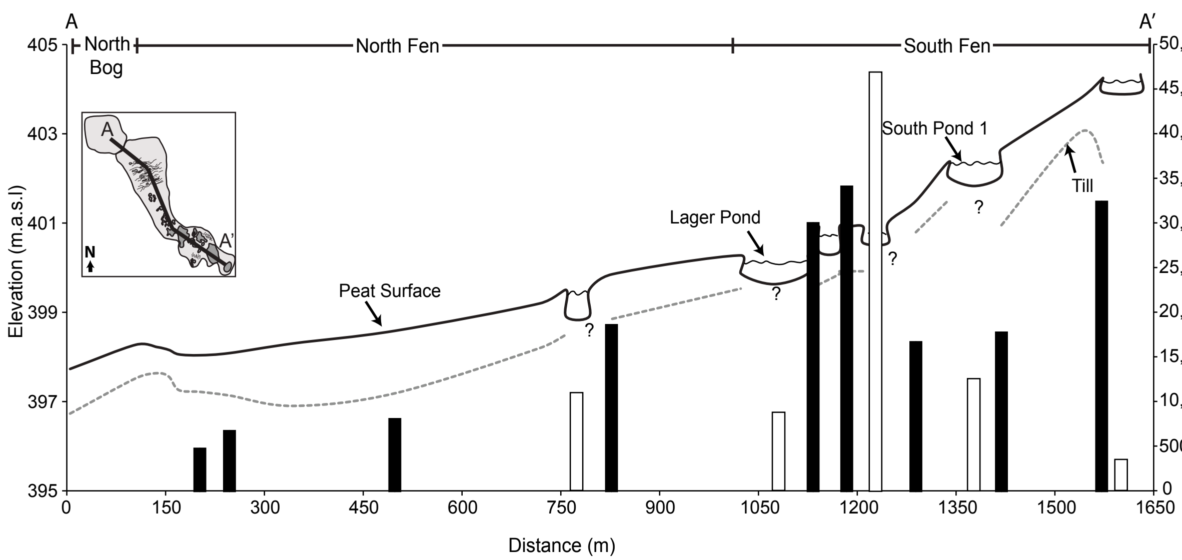

Figure 3-6 Concentration of total dissolved solids in groundwater within the fen's underlying mineral till (solid black vertical bars) and pond surface water (open vertical bars). Samples were extracted from piezometer nests along the primary A-A' transect (see inset map). (?) denotes locations under some larger ponds where the till could not be mapped.

Surface water samples collected during hydrogeologic investigations for the nearby SAGD project and for the Alberta Geological Survey were compared against samples collected at the saline fen to determine the spatial extent of saline discharge for the study area (Figure 3-7). In general, the presence of elevated salinity within surface features was largely restricted to the area around the saline fen and the Salt Creek Basin that runs through it. TDS increased from 500 mg L^{-1} upstream to 1,470 mg L^{-1} downstream along the reach adjacent to the saline fen. Here, a saline discharge feature similar to the saline fen with a TDS of 6,430 mg L^{-1} was found bordering the western edge of Salt Creek. Downstream, TDS remained high but decreased steadily up to the confluence between the eastern and western branches of Salt Creek. For the Prairie Creek and Saline Creek tributaries, TDS concentrations were typically no greater then 300 mg L^{-1}, with the exception of a distinct high salinity zone along the eastern branch of Saline Creek (Figure 3-7). A large saline wetland 500 m south east of the saline fen in between the Prairie Creek and Salt Creek basins also showed very high salinities similar to that of the saline fen (TDS 23,200 mg L^{-1}).

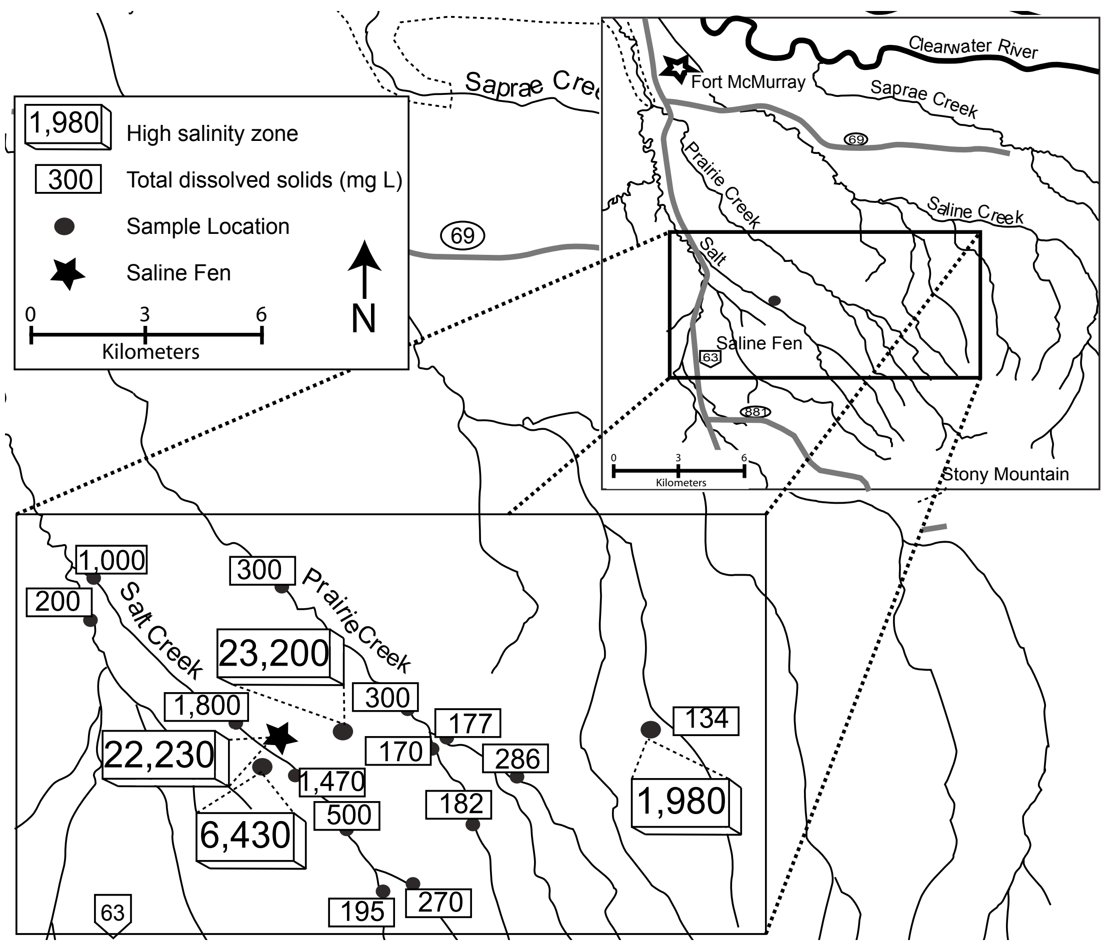

Figure 3-7 Concentration of total dissolved solids in surface waters around the saline fen study area. Samples were taken from adjacent river systems and other distinct saline surface features. Compiled from 2011-2012 field data from this work along with data from field investigations by Stewart and Lemay (2011) and Value Creations Inc. (2012).

Ground and Surface Water Chemistry

Ground and surface waters were all dominated by Na^+ and Cl^-, which on average accounted for over 90% of the TDS (Figure 3-8). The average concentration of Na^+ at the saline fen was 6,949 mg L^{-1} (minimum 195 mg L^{-1}, maximum 25,680 mg L^{-1}) while Cl^- was 13,776 mg L^{-1} (minimum 1,785 mg L^{-1}, maximum 56,249 mg L^{-1}) (Table 3-1). SO_4^{2-} was also quite high, averaging 728 mg L^{-1} with a maximum of 3,080 mg L^{-1}. The concentration of all dissolved ions, with the exception of Mg^{2+} and HCO_3^-, increased by at least an order of magnitude southward (Figure 3-9). Within the mineral till of the adjacent freshwater wetlands, one round of groundwater sampling revealed high concentrations of dissolved salts (Table 3-1). Groundwater related to the dissolution halite has a Na^+/Cl^- ratio close to 1:1. Ratios of Na^+/Cl^- (in meg/L) within fen groundwater averaged 0.80, increasing to 0.81 south of Lager Pond (minimum of 0.70, maximum of 0.98) and decreasing to 0.75 for the north fen (minimum of 0.60, maximum of 0.83). Despite the dominance of Na^+ and Cl^-, fen groundwater was several orders of magnitude undersaturated with respect to halite (saturation index > 0 when saturated). A modest but distinct increase towards saturation is observed moving southward, from -3.4 in the north fen to -2.8 in the south. The same trend is observed for gypsum and anhydrite, with groundwater in the south fen approaching saturation for both minerals (-0.97 and -1.38, respectively) (Table 3-2)

The conservative ions of Br^- and Cl^- were compared to deep formation waters (Connolly et al., 1990a) and saline springs (Grasby et al., 2006) throughout central and northeastern Alberta (Figure 3-10). The experimentally derived seawater evaporation/dilution trajectory (SET) is also shown (Carpenter, 1978). The distribution of formation waters on Figure 3-10 relative to the SET can indicate different origins and processes that have affected water chemistry during its fluid history. Similar to other springs, groundwater from the fen clusters to the left of the SET due to an excess of Cl^- relative to Br^-, with an average Cl^-/Br^- ratio of 7,500 due to Br depletion. This contrasts with formation brines that plot on or to the right of the trajectory, indicating an evaporated seawater component that is not derived from halite dissolution. The saline fen and Fort McMurray springs also show a more diluted signature compared to the halite saturated springs in the Wood Buffalo region.

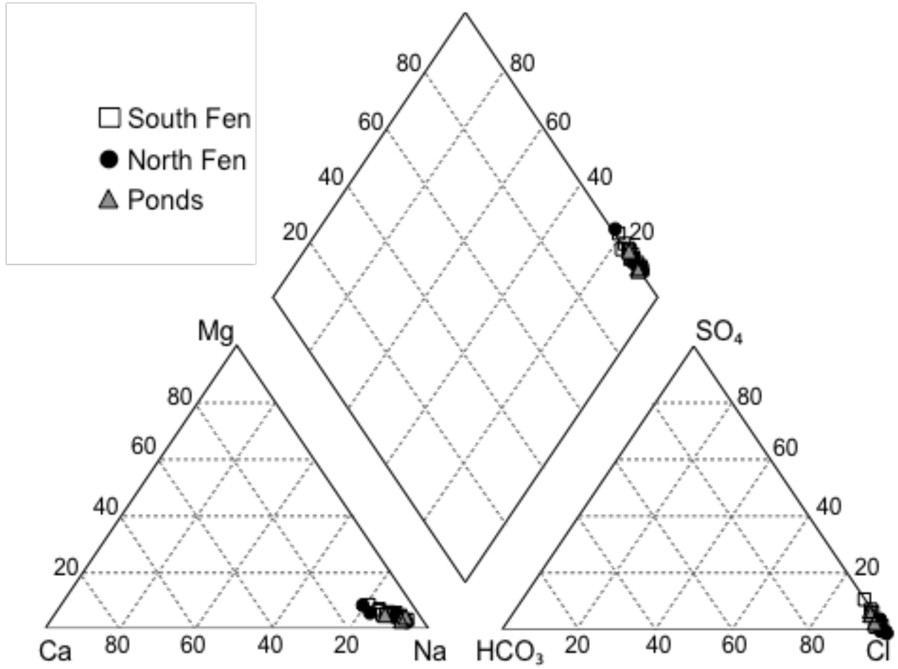

Figure 3-8 Piper plot showing the relative equivalent fractions of major ions for pond and peat water samples collected during the 2011 and 2012 study periods.

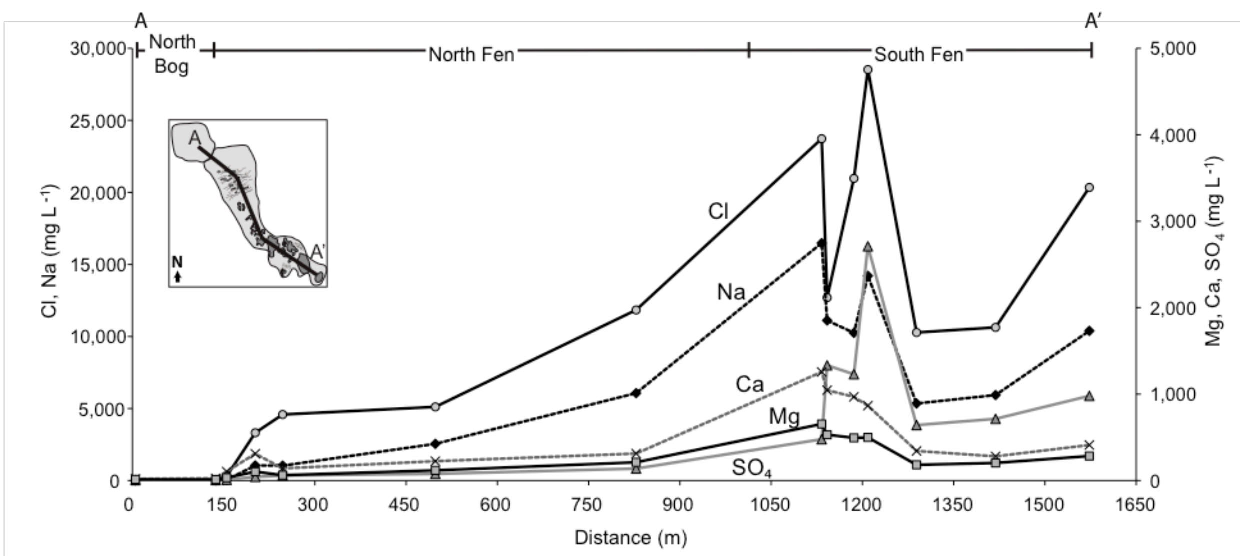

Figure 3-9 Spatial distribution of major ion concentrations within the mineral till along the primary A-A' transect (inset map). Each dot represents an average concentration for a piezometer obtained over two measurement periods in 2012.

Table 3-1 Major ion composition of saline fen groundwater and ponds along with wetlands to the west and east of the saline fen. Cl^-/Br^- and Na^+/Cl^- ratios are also shown for the saline fen.

		Saline Fen		Adjacent Wetlands	
				West Fen	East Bog
Parameter (mg L^{-1})		Groundwater§	Ponds¡	Groundwater†	
Cl^-	*Mean*	13776	5688	10543	3895
	Min	1785	1785		
	Max	56249	12095		
HCO_3^-	*Mean*	134	25	65	99
	Min	20	129		
	Max	627	84		
SO_4^{2-}	*Mean*	728	892	695	
	Min	28	1590		
	Max	3080	418		
Ca^+	*Mean*	391	48	650	157
	Min	48	515		
	Max	1696	150		
Mg^+	*Mean*	269	48	330	94
	Min	25	208		
	Max	2875	73		
Na^+	*Mean*	6949	4847	5148	1937
	Min	108	6474		
	Max	25680	1838		
Alk	*Mean*	110	52	54	81
	Min	16	105		
	Max	514	69		
TDS	*Mean*	22230	42	17488	9961
	Min	2893	21117		
	Max	87971	9364		
Cl^-/Br^-	*Mean*	6331	7654*		
	Min	4722	4411*		
	Max	9807	10896*		
Na^+/Cl^- (mEq/L)	*Mean*	0.79	0.78		
	Min	0.60	0.72		
	Max	0.98	0.84		

§ = Includes both mineral till and peat samples (55-60 samples)
¡ = Sample size of 8
† = Sample size of <3
* = Sample size of 2

Table 3-2 Saturation indices for halite, gypsum and anhydrite minerals in saline fen groundwater and ponds

		Halite	Gypsum	Anhydrite
North Fen	*Mean*	-3.42	-1.88	-2.31
	Max	-2.88	-1.59	-2.06
	Min	-4.21	-2.24	-2.49
South Fen	*Mean*	-2.86	-0.97	-1.38
	Max	-1.70	-0.19	-0.57
	Min	-3.94	-1.75	-2.14
Ponds	*Mean*	-3.46	-1.32	-1.72
	Max	-2.88	-0.55	-0.88
	Min	-3.94	-1.75	-2.44

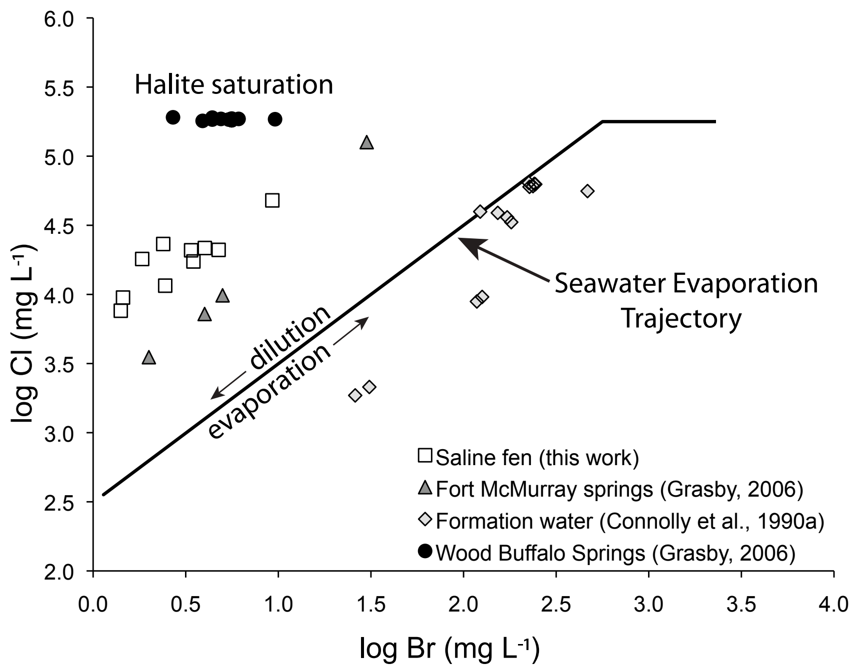

Figure 3-10 Log of chloride and bromide concentrations for fen groundwater plotted against the experimentally derived evaporation trajectory for seawater (Carpenter, 1978). Compositions for other springs in the Fort McMurray and Wood Buffalo regions are also shown (Grasby et al., 2006), along with values for formation waters found throughout the Alberta Basin (Connolly et al., 1990a).

O and H Stable Isotope Data

At the saline fen, $\delta^{18}O$ and $\delta^{2}H$ showed a broad range in values, generally falling far right of the local meteoric water line (AEMWL) for Edmonton (IAEA/WMO, 2001) with a low slope indicative of an evaporation trend ($\delta^{2}H = 4.6\delta^{18}O - 60.2$) (Figure 3-11). While the majority of samples exhibited patterns similar to those recorded by Grasby and Londry (2007) for low-flow springs with a strong evaporative component, several samples of pond surface water and groundwater in the till (depth > 1 m) within the south fen were plotted to the AEMWL. In a number of cases, samples within the basal peat and mineral till from the south fen were close to or well below the average value for local precipitation and near-surface aquifers.

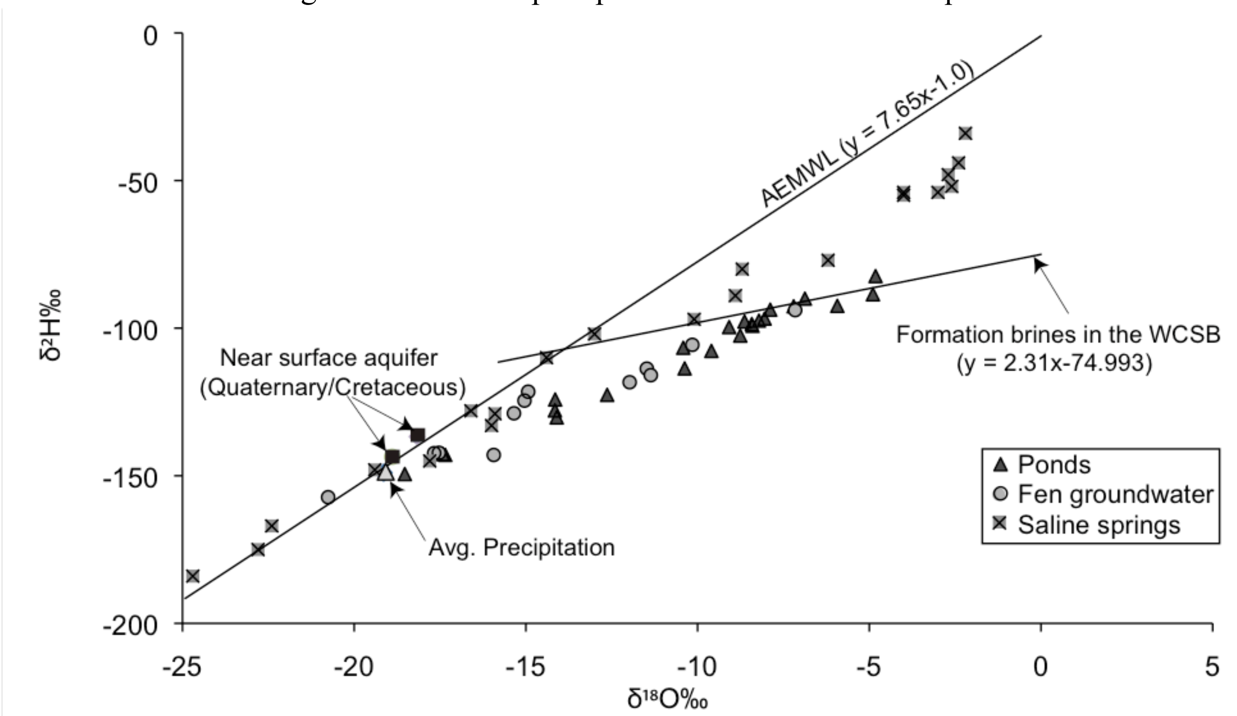

Figure 3-11 Relationship between oxygen and hydrogen isotope ratios for ponds (dark triangles) and groundwater taken from piezometers in saline fen (grey circles). Average values taken from the Fort McMurray region for Quaternary and the Lower Cretaceous Grand Rapids formation are also shown (Lemay, 2002), along with values from other springs found in the region (Grasby and Chen, 2005; Grasby and Londry, 2007). An average value for regional precipitation was calculated from data obtained from the Approximate Edmonton Meteoric Water Line (AEMWL). The formation water line obtained by Connolly et al., (1990b) is shown for comparison.

3.6 Discussion

Spring function

While the fen exhibited a similar configuration to other low-flow springs in the region – a sloped topography with a dense pond-pool configuration found along its topographic high – no active discharge outlet was observed in the pond-pool network over the two-year study (Timoney and Lee, 2001; Grasby and Londry, 2007). Due to the very low hydraulic conductivity observed in the till underlying the peat, deeper piezometers encounter a time-lag in trying to reach equilibrium, compared to the rapidly responding well; thus head differences, and consequently estimates of *dh/dz*, are artifacts of the design. A selection of piezometers that displayed relatively small time-lag did show consistent groundwater upwelling irrespective of water table changes (Figure 3-5) and in many cases were located in high-salinity zones south of Lager Pond. However, data from many of the piezometers had to be discarded from the analyses (because of excessive time-lag or never recharging post sampling), which made determining site-scale trends and estimates of saline groundwater discharge difficult. Nevertheless, on the basis of the few piezometers considered reliable, the overall flux of deep groundwater discharge was estimated to be very low and a minor component of the fen's water balance (Manuscript 1).

Mapping of the underlying substrate in 2011 revealed that a dense till was present throughout the fen, with very low hydraulic conductivities similar to the tills of northern prairie wetland regions (van der Kamp and Hayashi, 2008). Nevertheless, it is possible that fissures, weathered sections of till or locally thin or more permeable sediments may underlie some sections of the fen. The influence of these locally conductive zones can have a disproportionately large effect on the bulk hydraulic conductivity of a system, much more so than the estimated average hydraulic conductivity measured by piezometers (van der Kamp, 2001). This can partly be explained by the fact that piezometers have the tendency to underestimate hydraulic conductivity (Hanschke and Baird, 2001; Seo and Choe, 2001; Surridge et al., 2005), especially where there are localized higher-transmissivity zones. Consequently, the flux of saline groundwater from underlying units to the fen may be greater than what was estimated due to the presence of unidentified discharge windows and the incorrect estimates of hydraulic conductivity in dense tills. Transient conditions in spring discharge may also contribute to the very low measurements observed during this study. Historical records show that saline spring discharge can vary considerably over time, with rates fluctuating or stopping entirely for some springs over the last 100 years (McKillop et al., 1992; Grasby and Londry, 2007). Often, spring seeps are coated with extensive microbial biofabrics that tend to dry out and develop into characteristic ‘brain’ textured surfaces as spring discharge decreases (Grasby and Londry, 2007; Berard et al., 2013). At the saline fen, similarly textured microbial mats coat many of the dry adjacent areas surrounding what appear to be relict discharge features, which may be an indicator of the gradual decrease in groundwater discharge. In addition, the connection of discharge wetlands to regional groundwater flow systems can vary as a result of long-term drought and precipitation cycles (Winter and Rosenberry, 1998). It is possible that the minor groundwater input estimated at the saline fen may be due to the ephemeral nature of its discharge, which could be partly explained by the fact that the Boreal Plains region is currently in a long-term regional water deficit (Devito et al., 2012).

Despite uncertainties in the true magnitude of groundwater flux at the fen, strong geochemical evidence in ground and surface waters supports the hypothesis that the fen functions as a low-flow saline spring. This conceptual model is supported by the anomalously high concentrations of Na^+ and Cl^- within a regional landscape that is otherwise generally low in dissolved salts despite a similar hydrogeologic setting (Figure 3-7). An order of magnitude rise in Na^+, Cl^- and other ions within a relatively small region in the south fen (Figure 3-6; Figure 3-9), along with the presence of distinct spring-like features restricted to that area (e.g., halophytic vegetation, salt crusting, developed pond-pool network) suggests that discharge is or at one time was strongly focused in the fen’s southern extent. While patterns in deep groundwater flow observed at the fen must be interpreted with some caution, a significant amount of salt can still be transported via regional groundwater flow over thousands of years, despite the fact that it has little significance on the systems overall water balance (van der Kamp and Hayashi, 2008). Low but consistent dh/dz over the long-term in a system that detains solutes can accumulate salts despite a relatively low overall flux of groundwater. The fen’s current geochemical setting may be the result of a similar mechanism, where small dh/dz through a low-conductivity substrate can result in a net gain of salt over very long time-scales.

Regional Groundwater Connection

The geochemistry of the saline fen and other discharge features in northeastern Alberta is at odds with the majority of formation brines in the Alberta basin, whose salinities are often related to seawater (Hitchon and Friedman, 1969; Hitchon et al., 1971; Connolly et al., 1990a). Within the regional framework of the saline fen study area, a mix of NaCl-HCO_3 and NaCl water types dominated formation water brines in Lower Cretaceous aquifers (Lemay, 2002). Cl^-/Br^- ratios were consistent throughout the stratigraphic package and were similar to that of seawater (between 250 and 450), with no evidence of contact with halite (as indicated by low Cl^-/Br^- ratios). TDS concentrations within these formations were typically well below what would be expected for connate seawater (~35,000 mg L^{-1}), suggesting the brines have undergone some level of mixing with freshwater resulting in dilution (Lemay, 2002). Despite trends in regional groundwater, geochemical evidence still points to halite dissolution as a potential source of salinity at the saline fen (Table 3-1; Figure 3-10). This is similar to other saline springs in northeastern Alberta, east-central Saskatchewan and west-central Manitoba that receive dissolved halite in connection to Devonian carbonate outcrop belts (Grasby and Betcher, 2002; Grasby and Chen, 2005; Grasby et al., 2006; Grasby and Londry, 2007). Due to the relatively conservative nature of Na^+, Cl^- and Br^- in most groundwater settings, the relationship between these ions has proved useful in the determination of the origin and evolution of formation waters (Carpenter, 1978; Walter et al., 1990; Kesler et al., 1995; Davis et al., 1998; Gupta et al., 2012). At the saline fen, the average Cl^-/Br^- ratio of groundwater in the high salinity zone (south fen) was 7,500 and was well within the range of other subsurface brines influenced by halite dissolution (Carpenter, 1978; Davis et al., 1998). This is much higher then the average ratio of ~290 for seawater but is lower than Cl^-/Br^- ratios found in high salinity springs in the Wood Buffalo region that are typically supersatured with respect to halite (Grasby et al., 2006). Ratios as high as 70,000 in subsurface brines are due not only to the significant difference in the natural abundance between Cl^- and Br^- but also as a result of differences in solubility (Davis et al., 1998; Freeman, 2007). Br^- tends to be excluded from the crystal structure of halite as it precipitates due to the continual evaporation of seawater. As a result, meteoric waters that come into contact with and cause dissolution of halite will typically have very high Cl^-/Br^- ratios due to excess chloride relative to bromide. Evidence of halite dissolution is supported further through the relationship between Na^+ and Cl^-. While Na^+/Cl^- ratios at the fen are not as high as 1.0 and are lower than ratios found in other highly saline springs (>0.9), they are on average greater than seawater brines with ratios typically between 0.5 and 0.8 (Table 3-1) (Connolly et al., 1990a; Grasby and Betcher, 2002).

The relationships among Na^+, Cl^- and Br^- points strongly to halite dissolution as a source of salt at the fen but biogeochemical processes, such as sorption and plant uptake, can influence the concentration of these ions in the wetland setting. Since natural concentrations of Br^- are typically low and because Cl^- can be as much as 8,000 times as abundant, slight changes in the concentration of Br relative to Cl^- can cause significant changes in Cl^-/Br^- ratios and thus the interpretation of the origin of salt (Davis et al., 1998). Br^- has traditionally been considered non-reactive in soils; however, the sorption of Br^- by up to 10% has been reported in various materials, including clays and organic soils (Behl et al., 1990; Wilson and Gabet, 1991 in Davis et al., 1998). While it is possible that Br- concentrations obtained from fen groundwater could be underestimated (thus leading to an overestimation of Cl^-/Br^- ratios), an increase in Br^- by up to 25% results in a Cl^-/Br^- ratio of just under 6,000, still well above the minimum reported value for brines influenced by halite dissolution (Davis et al., 1998, Freeman, 2007). In their investigation of Br as an appropriate tracer in wetlands, Xu et al., (2004) found that typical wetland plants (*T. latifolia* and *P. australis*) can take up significant quantities of Br^- in root and leaf tissues. This is unlikely an issue at the fen since all samples used for Cl^-/Br^- analyses were taken well below the root layer within the basal peat (>1.0 m depth) or underlying mineral till. Moreover, Cl^- has been shown to strongly inhibit the uptake of Br^- in plants, which in highly saline environments like the saline fen may reduce or eliminate Br^- removal from groundwater (Xu et al., 2004).

The saline fen is a characteristic groundwater feature within the Fort McMurray area, whose geochemistry suggests halite dissolution as a salinity source with mixing of meteoric waters. Studies of the isotopic composition of deep basinal brines related to seawater within the Alberta basin indicate enriched $\delta^{18}O$ and $\delta^{2}H$ values that plot far to the right of the global meteoric water line (Connolly et al., 1990b) (Figure 3-10). This is in contrast to the discharge waters of saline springs associated with halite dissolution that are similar to or significantly lower than the mean regional average for precipitation (Figure 3-11) (Grasby and Chen, 2005). The cause for $\delta^{18}O$ values as low as -25‰ in some saline springs has been attributed to the process of infiltration and subsequent recirculation of Pleistocene meltwater that has dissolved deeply buried evaporite deposits (Grasby and Chen, 2005). Similarly low $\delta^{18}O$ values were not observed at the saline fen and the broad range in values with a low slope is indicative of kinetic fractionation caused by evaporation due to the weak or nonexistent nature of saline discharge (Figure 3-11). The influence of evaporation on discharge water has been observed for other low flow springs in central Manitoba (Grasby and Londry, 2007). However, several high salinity ponds and samples within the till (depth > 1 m) plotted close to the AEMWL and are similar to the regional average for precipitation and local aquifers. This is similar to other springs in the Wood Buffalo region, where more present-day flow systems are responsible for the dissolution of halite (Grasby et al., 2006). $\delta^{18}O$ and $\delta^{2}H$ values within the McMurray, Grand Rapids and Clearwater formations all fell close to the AEMWL, which was interpreted as result of mixing with meteoric waters despite a strong NaCl signature (Lemay, 2002). Saline discharge water at the fen may be significantly diluted as a result of mixing with freshwater sources similar to formation waters described by Lemay (2002). Salinities, Cl^-/Br^- and Na^+/Cl^- ratios lower than the highly saline springs north of Fort McMurray supports this hypothesis (Grasby et al., 2006). Hydrologic connections between deep, highly saline Paleozoic aquifers and low-salinity to freshwater near surface aquifers have been observed and considered feasible along the regional flow path southwest of the AOSR (Bachu et al., 1993). This connection could explain the presence of the saline plume within the near surface Grand Rapids formation and may explain the increased levels of salinity found not only within the saline fen but the surrounding study region. A detailed geochemical investigation of the Grand Rapids saline plume may reveal more information on its fluid history and confirm the origin of groundwater at the saline fen

3.7 Conclusions

The saline fen functions as a low flow discharge feature that provides an opportunity to examine the nature of groundwater flow within the AOSR. Its location south of the oil sands hub of Fort McMurray, away from major river system makes its hydrogeologic position different than other saline springs studied in the region. Due to a low-conductivity till underlying the fen and weak vertical hydraulic gradients, the present magnitude of saline groundwater discharge was estimated to be minimal. Patterns in ground and surface water geochemistry indicate strong spatial variability at the fen, where the greatest concentration of Na^+ and Cl^- was found within a relatively small zone in its southern extent. Elevated salinities combined with other spring-like features (e.g., halophytes, pools and ponds and salt crusting) indicate that discharge is strongly focused within this region. Based on the current hydrologic regime as characterized in Manuscript 1 it is unlikely that the fen would have any significant impact on the surface within the foreseeable future in the event that it functioned as a conduit for wastewater discharge. This is supported not only by minimal deep groundwater discharge but also through low rates of shallow groundwater flow between the fen and adjacent wetlands. However, it is possible that the true rate of deep groundwater discharge was underestimated due to locally conductive zones not captured or because of the ephemeral nature of spring discharge. Installation of a piezometer network constructed specifically for dense, low-conductivity materials (e.g., sand packs around piezometer openings, longer open screen length) with greater spatial coverage may allow for a better characterization of heterogeneity and thus a more accurate estimation of hydraulic conductivity. The anomalously high concentration of dissolved salts within the surrounding wetlands and river systems indicate that the actual extent of saline groundwater discharge may be much greater than what is occurring in the saline fen. The localized accumulation of salts in this region is indicative of the long-term discharge of groundwater and suggests that larger-scale flow systems play an important role in ecosystem function. Furthermore, the presence of saline springs and the elevated concentration of salinity in nearby river systems indicate that the influence of saline discharge is not restricted to major river systems. The flux of salt from these systems may play an important role in the over water quality of groundwater and receiving water bodies (e.g., nearby river systems). In a region under considerable developmental stress, effective management of these groundwater-dependent ecosystems will require a sound understanding groundwater flow systems at multiple scales.

Fen geochemistry differed from what has been observed for formation brines in the region, where elevated salinities and isotopic signatures are consistent with the interpretation that they represent residual evaporated seawater. This contrasts with the origin of salinity in saline springs, where the dissolution of near-surface evaporites by meteoric water results in high concentrations of Na^+ and Cl^-. Unlike other springs, however, the saline fen is hydrologically disconnected from underlying Devonian evaporite beds, and receives its groundwater from a near-surface plume in a Lower Cretaceous aquifer that subcrops just south of the fen. This makes the fen an unlikely source for the discharge of subsurface wastewater stored by adjacent SAGD facilities. Fen ground and surface water geochemistry point to a complex fluid history where elevated salinities can be explained, in part, by evaporite dissolution. The contribution of evaporite to discharge water may be coming from somewhere deeper and further south in the basin. This has important implications for the disposal of wastewater by deep well injection, as disposal zones may be hydrologically linked to near-surface aquifers and discharge features well beyond the immediate production and storage area.

4.0 Conclusions and Significance of Research

Disconnected from underlying Devonian evaporites and positioned much further south than other springs in the region, the saline fen's hydrogeologic position made it an intriguing discharge feature that had not previously been explored in the AOSR. The fen's overall configuration in the landscape, along with a net-patterned microtopography, influenced the spatial distribution of salts within ground and surface waters. Under baseflow conditions, steep gradients in the south fen supplied subsurface flow following the local topographic gradient. The fen's subsurface storage capacity was quickly exceed during extended rain events, which contributed to flooding of the fen's lower reaches during wet periods. The large depression storage capacity of the pond-pool network was rarely exceeded, but during periods of intense rainfall or snowmelt the fill-and-spill mechanism created an extended pond network that transmitted water northward. The re-distribution of salts and other solutes in shallow ground and surface water during wet periods likely plays a critical role in vegetation composition and biodiversity at the fen. Shallow groundwater flow was important for pond-pool permanence, as a dense low-conductivity till was found to underline the majority of the pond network that restricted the vertical flux of deep groundwater. The semi-permanence of many ponds suggests that shallow groundwater inflow is minor and meteoric input is the primary supply of water. Flow reversals were also common, and these temporary recharge features may provide an opportunity for ponds to regulate their salinity.

Precipitation and evapotranspiration dominated the spring fen's water balance and the flux of shallow groundwater between adjacent wetlands was small. Runoff was also negligible as a result of the rough surface topography that detained surface water despite flooding conditions. Together, these mechanisms provide a clearer picture as to how freshwater wetlands can occur in such close proximity to a saline system. Despite the high concentrations of dissolved salts and a clear spatial trend in salinity, the link between hydrologic function and geochemistry was not always clear. The presence of a low conductivity till made estimating the true magnitude of deep groundwater flux a challenge at the saline fen, while the existence of fractures or other discharge windows may have been missed. Installation of piezometer networks constructed specifically for dense tills with greater spatial coverage may allow for a more accurate characterization of the true magnitude of discharge at the fen and other systems underlain by similar substrates.

Within the low-lying plains regions of study area, thick and continuous organic deposits made up of treed and open freshwater fens overlie glaciolacustrine tills with high clay content. These low-conductivity clay-rich tills work to isolate surficial aquifers from larger-scale flow systems, developing extensive wetland complexes that are controlled by more localized groundwater regimes. Despite the apparent lack of regional groundwater-wetland interaction, saline systems like the saline fen and river systems with above-average concentrations of dissolved salts can be been found within the study region south of Fort McMurray. The localized accumulation of salts is indicative of deep groundwater discharge and suggests that larger-scale flow systems may play an important role in wetland permanence in the AOSR. Furthermore, the flux of salt from saline wetland features may play an important role in the overall water quality of groundwater and receiving water bodies (e.g., nearby river systems). For the saline fen, a lower Cretaceous formation (the Grand Rapids) that subcrops just south of the study area was identified as a likely source of saline groundwater discharge. The geochemical signature of fen groundwater is distinct from regional formation brines that are typically related to seawater but is similar to other saline springs that receive the bulk of their salinity from dissolved evaporite. The presence of dissolved halite in a system hydrologically disconnected from deep underlying Devonian evaporites points to a complex fluid history where discharge water may be coming from somewhere deeper and further south in the basin. This has implications for deep well injection, as disposal zones may be hydrologically connected to near surface aquifers and discharge features well beyond their immediate production area. A more detailed investigation of the Grand Rapids aquifer geochemistry may provide a better understanding of the extent and origin of discharge water at the fen.

Currently, the hydrologic function of saline wetlands is not entirely understood within the oil sands region. This study serves as a first approximation looking into the hydrology of a saline wetland, a governing variable that has strong control over biogeochemistry and ecology. With development expected to double within the coming decade, successful reclamation strategies will require a sound understanding of wetland form and function. The presence of salinity in the post-mined environment will create unique challenges for developers and the study of saline wetlands in particular can provide information critical to the successful establishment of reclaimed landscapes. From an in-situ extraction perspective, knowledge of how saline spring wetlands function as possible discharge features is crucial for developers aiming to successfully contain wastewater within the subsurface. Furthermore, due to the known impact of saline groundwater discharge on regional water quality, features like the saline fen should be incorporated into monitoring networks to better quantify the amount of naturally occurring saline discharge. As a rare ecosystem in an area under stress, greater knowledge of how saline wetlands operate within a regional context will allow for a more holistic approach to ecosystem management within the AOSR.

References

Allen, R.G., Pruitt, W.O., Jensen, M.E., 1991. Environmental requirements of lysimeters. In Allen, R.G., Howell, T.A., Pruitt, W.O., Walter, I.A., Jensen, M.E., (Eds). Lysimeters for Evapotranspiration and Environmental Measurements. ASCE, New York, NY. 170-181

Andriashek, L., 2003. Quaternary geological setting of the athabasca oil sands (in situ) area, Northeast Alberta. Alberta Energy and Utilities Board, EUB/AGS Geo-Note 2002-01.

AOC, 2011. Application for approval of the Athabasca Oil Sands Corp. Hangingstone project. Athabasca Oil Sands Corporation.

Apostol, K.G., Zqiazek, J.J., MacKinnon, M.D., 2004. Naphthenic acids affect plant water conductance but do not alter shoot Na+ and Cl- concentrations in jack pine (*Pinus banksiana*) seedlings. Plant and Soil, 263: 183-190.

Bachu, S., Underschultz, B.H., Cotterill, D., 1993. Regional-scale subsurface hydrogeology in northeastern Alberta. Alberta Geological Survey, Alberta Research Council, Edmonton.

Beckwith, C.W., Baird, A.J., Heathwaite, A.L., 2003. Anisotropy and depth-related heterogeneity of hydraulic conductivity in a bog peat. I: laboratory measurements. Hydrological Processes, 17(1): 89–101.

Behl, E., 1990. Chloride/Bromide ratios as a natural tracer of urban runoff in the Tucson basin, Arizona. M.S. thesis, University of Arizona.

Belyea, L.R., Lancaster, J., 2002. Inferring landscape dynamics of bog pools from scaling relationships and spatial patterns. Journal of Ecology, 90(2): 223-234.

Berard, G., Applin, D., Cloutis, E., Stromberg, J., Sharma, R., Mann, P., Grasby, S., Bezys, R., Horgan, B., Londry, K., Rice, M., Last, Bill., Last, Fawn., Badiou, P., Goldsborough, G., Bell, J., 2013. A hypersaline spring analogue in Manitoba, Canada for potential ancient spring deposits on Mars. Icarus, 224: 399-412.

Borneuf, D., 1983. Springs of Alberta. Alberta Research Council, Earth Sciences Report 82-3. Edmonton, Alberta, Canada.

Butler, R.M., 1994. Steam-assisted gravity drainage: concept, development, performance and future. Journal of Canadian Petroleum Technology, 33(2): 44-50

Butler, R., 2001. Some recent developments in SAGD. Journal of Canadian Petroleum Technology, 40(1): 18–22.

Bradbury, K.R., Muldoon, M.A., 1990. Hydraulic conductivity determinations in unlithified glacial and fluvial materials. In: Nielsen, D.M., Johnson, A.I., (Eds.). Ground Water and Vadose Zone Monitoring, ASTM STP 1053. American Society for Testing and Materials, Philadelphia: 138-151.

Brown, S.M., Petrone, R.M., Mendoza, C., Devito, K.J., 2010. Surface vegetation controls on evapotranspiration from a sub-humid Western Boreal Plain wetland. Hydrological Processes, 24: 1072-1085.

Carpenter, A.B., 1978. Origin and chemical evolution of brines II Sedimentary Basin. Society of Petroleum Engineers of AIME. SPE 7504.

Carrigy, M.A., McLaws, I.J., 1973. Athabasca tar sands study. The environmental impact of in-situ technology. Research Council of Alberta. Edmonton.

Connolly, C.A., Walter, L.M., Baadsgaard, H., Longstaffe, F.J., 1990a. Origin and evolution of formation waters, Alberta Basin, Western Canada Sedimentary Basin. I. Chemistry. Applied Geochemistry, 5(4): 375–395.

Connolly, C.A., Walter, L.M., Baadsgaard, H., Longstaffe, F.J., 1990b. Origin and evolution of formation waters, Alberta Basin, Western Canada Sedimentary Basin. II. Isotope systematics and water mixing. Applied Geochemistry, 5(4): 397–413.

Davis, S.N., Whittemore, D.O., Fabryka-Martin, J., 1998. Uses of chloride/bromide ratios in studies of potable water. Ground Water, 36(2): 338–350.

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

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

Devito, K.J., Waddington, J.M., & Branfireun, B.A., 1997. Flow reversals in peatlands influenced by local groundwater systems. Hydrological Processes, 11: 103–110.

Devito, Kevin, Mendoza, C., 2007. Appendix C-1 - Maintenance and dynamics of natural wetlands in western Boreal forests: synthesis of current understanding from the Utikuma Research Study Area. In: Harris, M.L., Guideline for wetland establishment on reclaimed oil sands leases (2nd Edition). Lorax Environmental for the Wetlands and Aquatics Subgroup of the Reclamation Working Group of the Cumulative Environmental Management Association, Fort McMurray, Alberta, Canada.

Devito, K., Mendoza, C., Qualizza, C., 2012. Conceptualizing water movement in the boreal plains - implications for watershed reconstruction. Synthesis report prepared for the Canadian Oil Sands Network for Research and Development, Environmental and Reclamation Research Group. 164 pp.

Dewar, W.A., McDonald, P., 1961. Determination of dry matter in silage by distillation with toluene. Journal of the Science of Food and Agriculture, 12: 790-795.

Drees, M., 1994. Devonian elk point group of the Western Canada Sedimentary Basin. In G., Mossop, Shetsen, I., (Eds.), Geological Atlas of the Western Canada Sedimentary Basin (Special Re.). Canadian Society of Petroleum Geologists and Alberta Research Council.

Environment Canada, 2012. Canadian climate normal 1971-2000 station data. Government of Canada

ERCB, 1994. Injection and disposal wells - well classifications, completions, logging, and testing requirements, Directive 051. Energy Resources and Conservation Board, Calgary, Alberta, Canada.

ERCB, 2011. Alberta's reserves and supply/demand outlook 2010–2020, ST98. Energy Resources and Conservation Board. Calgary, Alberta, Canada.

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

Fetter, C.W., 2001. Applied Hydrogeology 4th Edition. Prentice-Hall, Englewood Cliffs, N.J.

Finch, J., Calver, A., 2008. Methods for the quantification of evaporation from lakes. World Meteorological Organization Commission for Hydrology. United Kingdom.

Finch, J.W., Hall, R.L., 2001. Estimation of open water evaporation. Guidance for Environment Agency Practitioners. Great Britain. 45 pp.

Ford, D.C., 1998. Principal features of evaporite karst in Canada. Suppl. Geogr. Fis. Dinam. Quat., 4: 11–19.

Foster, D.R., King, G.A., Glaser, P.H., Wright, H.E Jr., 1983. Origin of string patterns in boreal peatlands. Nature, 306(17): 256-258.

Freeze, R.A., Cherry, J.A., 1979. Groundwater. Prentice-Hall, Englewood Cliffs, N.J.

Gordon, T.M., Kokot, S.L., Parks, K.P., 2002. Guide to recent publications on inorganic water-rock interactions relevant to deep-well wastewater disposal in carbonate-evaporite formations in the Athabasca Oil Sands Area. Alberta Energy and Utilities Board, EUB/AGS Special Report 52.

Gorham, E., 1991. Northern peatlands: role in the carbon cycle and probable responses to climate warming. Ecological Applications, 1(2): 182–195.

Government of Alberta, 2000. Environmental protection and enhancement act, revised statutes of Alberta 2000. Province of Alberta, Edmonton, Alberta, Canada.

Government of Alberta, 2008. Alberta's oil sands - resourceful, responsible.

Grasby, S.E., 2006. Brine springs of northern Alberta. In: Hannigan, P.K., (ed.), Potential for carbonate-hosted lead-zinc Mississippi Valley-type mineralization in northern Alberta and southern Northwest Territories. Geoscience Contributions, Targeted Geoscience Initiative, Geological Survey of Canada, Bulletin 591, p. 241–254.

Grasby, S.E., Betcher, R.N., 2002. Regional hydrogeochemistry of the carbonate rock aquifer, southern Manitoba. Candian Journal of Earth Science, 39: 1053-1063

Grasby, S.E., Londry, K.L., 2007. Biogeochemistry of hypersaline springs supporting a mid-continent marine ecosystem: an Analogue for martian springs? Astrobiology, 7(4): 662–683.

Grasby, S.E., Chen, Z., 2005. Subglacial recharge into the Western Canada Sedimentary Basin—impact of Pleistocene glaciation on basin hydrodynamics. Geological Society of America Bulletin, 117(3): 500-514.

Grobe, M., 2000. Distribution and thickness of salt within the Devonian Elk Point Group, Western Canada Sedimentary Basin. Alberta Energy and Utilities Board, Earth Sciences Report 2000-02.

Gupta, I., Wilson, A.M., Rostron, B.J., 2012. Cl/Br compositions as indicators of the origin of brines: Hydrogeology of the Alberta Basin, Canada. 65A Bulletin, 124: 200-212.

Hackbarth, D., Nastasa, N., 1979. The hydrogeology of the Athabasca Oil Sands Area, Bulletin 98. Alberta Research Council. Alberta, Canada.

Hamilton, W.N., 1971. Salt in east-central Alberta, Bulletin 29. Alberta Research Council. Edmonton, Alberta, Canada.

Han, H.H, Choe, J., 2001. Numerical simulations for effects of anisotropy and heterogeneity on well responses to slug tests. Environmental Geology, 40: 1066–1074.

Hanor, J.S., 1994. Origin of saline fluids in sedimentary basins. Geological Society, London, Special Publications, 78(1): 151–174.

Hanschke, T., Baird, A.J., 2001. Time-lag errors associated with the use of simple standpipe piezometers in wetland soils. Wetlands, 21(3): 412-421.

Harvey, F.E., Ayers, J.F., Gosselin, D.C., 2007. Ground water dependence of endangered ecosystems: Nebraska's eastern saline wetlands. Groundwater, 45(6):736-752.

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

Hayashi, M., van der Kamp, G., 2007. Water level changes in ponds and lakes: the hydrological processes. In Johnson, E.A., Miyanishi, K., (Eds.), Plant Disturbance Ecology: The Process and the Response. Academic Press, San Diego, 311-3

Heagle, D., Hayashi, M., van der Kamp, G., 2013. Surface-subsurface salinity distribution and exchange in a closed-basin prairie wetland. Journal of Hydrology, 478: 1-14.

Hitchon, B., 1969. Fluid flow in the Western Canada Sedimentary Basin 1. Effect of topography. Water Resources Research, 5(1): 186-195.

Hitchon, B., Billings, G.K., Kloven, J.E., 1971. Geochemistry and origin of formation waters in the western Canada sedimentary basin - III. Factors controlling chemical composition. Geochimica et Cosmochimica Acta, 35(520): 567–598.

Hitchon, B., Friedman, I., 1969. Geochemistry and origin of formation waters in the western Canada sedimentary basin - I. Stable isotopes of hydrogen and oxygen. Geochimica et Cosmochimica Acta, 33: 132-1349.

Hitchon, B., Levinson, A., Reeder, S., 1969. Regional variations of river water composition resulting from halite solution, Mackenzie River Drainage Basin, Canada. Water Resources Research, 5(6): 1395–1403.

Hvorslev, M. J., 1951. Time Lag and Soil Permeability in Ground-Water Observations, Waterways Experimental Station Bulletin 36. US Army Corps of Engineers, Vicksburg, Mississippi.

IAEA/WMO, 2001. Global network for isotopes in precipitation, the GNIP database. <http://www-naweb.iaea.org/napc/ih/IHS_resources_isohis.html>

Ingram, H.A.P, 1983. Hydrology. In: Gore, A.J.P. (Ed.), Ecosystems of the World 4A, Mires: swamp, bog, fen and moor. Elsevier, Amsterdam, 67-158.

Jasechko et al.,, S., Gibson, J.J., Birks, S.J., Yi, Y., (2012). Quantifying saline groundwater seepage to surface waters in the Athabasca oil sands region. Applied Geochemistry. 27: 2068-2076.

Johnson, E.A., Miyanishi, K., 2008. Creating new landscapes and ecosystems: the Alberta Oil Sands. Annals of the New York Academy of Sciences, 1143: 120–45.

Jordaan, S.M. 2012. Land and water impacts of oil sands production in Alberta. Environmental Science & Technology, 46(7): 3611–7.

Kasenow, M., 2001. Applied Ground-water hydrology and well hydraulics, 2nd edition. Water Resources Publications. Denver, Colorado, United States of America.

Kesler, S.E., Appold, M.S., Martini, A.M., Walter, L.M., Huston, T.J., Kyle, J.R. 1995. Na-Cl-Br systematics of mineralizing brines in Mississippi Valley – type deposits. Geology, 23(7): 641-644.

Kruse, J., Lennartz, B., Leinweber, P., 2008. A modified method for measuring saturated hydraulic conductivity and anisotropy of fen peat samples. Wetlands, 28(2): 527–531.

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

Last, W.M., Ginn, F.M. 2005. Saline systems of the Great Plains of western Canada: an overview of the limnogeology and paleolimnology. Saline systems, 1(10).

Lefleur, P.M., Roulet, N.T., 1992. A comparison of evaporation rates from two fens of the Hudson Bay Lowland. Aquatic Botany, 44: 59-69.

Lemasters, G.S., Bartelli, L.J., Smith, M.R., 1983. Characterization of organic soils as energy sources. In: Jarrett, P.M. (Ed.). Testing of Peats and Organic Soils, ASTM STP 820, American Society for Testing and Materials: 122-137.

Lemay, T., 2002. Geochemical and isotope data for formation water from selected wells, Cretaceous to Quaternary succession, Athabasca Oil Sands (in situ) area, Alberta. Alberta Energy and Utilities Board, EUB/AGS Geo-note 2002-02.

Li, C., Flannigan, M.D., Corns, I.G.W., 2000. Influence of potential climate change on forest landscape dynamics of west-central Alberta. Canadian Journal of Forest Research, 30: 105-1912.

Linacre, E.T., 1993. Data-sparse estimation of lake evaporation, using a simplified Penman equation. Agricultural and Forest Meteorology, 64: 237-256.

Lowry, C.S., Fratta, D., Anderson, M.P., 2009. Ground penetrating radar and spring formation in a groundwater dominated peat wetland. Journal of Hydrology, 373: 68-79.

McCuen, R.H., Asmussen, L.E., 1973. Estimating the effect of heat storage on evaporating rates. Hydrological Sciences Bulletin, 18(2): 191-196,

McKillop, W.B., Patterson, R.T., Delorme, L.D., Nogrady, T., 1992. The origin, physico-chemistry and biotics of sodium chloride dominated saline waters of the western shore of Lake Winnipegosis, Manitoba. Canadian Field Naturalist, 106: 454-473

Michael, K., Machel, H.G., Bachu, S., 2003. New insights into the origin and migration of brines in deep Devonian aquifers, Alberta Canada. Journal of Geochemical Exploration, 80: 193-219

Moore, P. D., 1982. Pool and ridge patterns in peat mires. Nature, 300: 110

National Wetlands Working Group, 1997. The Canadian wetland classification system, second edition. University of Waterloo, Waterloo, Ontario, Canada.

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/582.

Oke, T.R., 1987. Boundary layer climates, 2nd edition. Routledge, London.

Ozoray, G., Hackbarth, D., Lytviak, A., 1980. Hydrogeology of the Bitumount-Namur Lake Area, Alberta, Earth Science Report 78-6. Alberta Research Council.

Parkhurst, D.L., Appelo, C.A.J., 2013. Description of input and examples for PHREEQC version 3 - A computer program for speciation, batch-reaction, one-dimensional transport, and inverse geochemical calculations. US Geological Survey Techniques and Methods, Book 6, Chapter A: 497 pp.

Pearsall, W.H., 1956. Two-blanket bogs in Sutherland. Journal of Ecology, 44(2): 493-516.

Penman, H.L., 1948. Natural evaporation from open water, bare soil and grass. Proceedings of the Royal Society of London. Series A, Mathematical and Physical Sciences, 193(1032): 120-145.

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

Phillips, T., 2013. Quantifying and characterizing evapotranspiration in a saline fen within the western boreal plain, Fort McMurray, AB. M.Sc thesis, Wilfred Laurier University.

Pouliot, R., Rochefort, L., Graf, M.D., 2012. Impacts of oil sands process water on fen plants: Implications for plant selection in required reclamation projects. Environmental Pollution, 167: 132-137.

Price, J.S., 1991. Evaporation from a blanket bog in a foggy coastal environment. Boundary-Layer Meteorology, 57: 391-406.

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

Price, J.S., Ketcheson, S.J., 2009. Water Relations in Cutover Peatlands. Geophysical Monograph Series, 184: 277–287.

Price, J.S., Maloney, D.A., 1994. Hydrology of a patterned bog-fen complex in southeastern Labrador, Canada. Nordic Hydrology, 25: 313–330.

Price, J.S., Woo, M.K., 1988. Studies of a subarctic coastal marsh, I. Hydrology. Journal of Hydrology, 103: 275-292

Priestly, C.H.B., Taylor, R.J., 1972. On the assessment of surface heat flux and evaporation using large-scale parameters. Monthly Weather Review, 100(2): 81-92

Purdy, B., Macdonald, S.E., Lieffers, V.J., 2005. Naturally saline boreal communities as models for reclamation of saline oil sands tailings. Restoration Ecology, 13: 666-677

Redding, T., Devito, K., 2005. Snowmelt infiltration and runoff from forested hillslopes, Boreal Plain, Alberta. 62nd Eastern Snow Conference: 105–108.

Redfield, E.B., 2001. Tolerance mechanisms of black spruce (*Picea mariana*) seedlings exposed to saline oil sands tailings; M.Sc thesis, University of Calgary.

Reeve, A., Siegel, D., Glaser, P., 2000. Simulating vertical flow in large peatlands. Journal of Hydrology, 227(1-4): 207–217.

Renault., Lait, C., Zwiazek, J.J., MacKinnon, M., 1998. Effect of high salinity tailings waters produced from gypsum treatment of oil sands tailings on plants of the boreal forest, Environmental Pollution, 102: 177-184.

Rittenhouse, G., 1967. Bromine in oilfield waters and its use in determining possibilities of origin of these waters. The American Association of Petroleum Geologists Bulletin, 51: 2430-2440.

Rydin, H., Jeglum, J., 2006. The Biology of Peatlands*.* U.S: Oxford University Press.

Scarlett, SJ., Price, J.S., 2013. The hydrological and geochemical isolation of a freshwater bog within a saline fen in north-eastern Alberta. Mires and Peat, 12(4): 1-12.

Seo, H.H., Choe, J., 2001. Numerical simulations for effects of anisotropy and heterogeneity on well responses to slug tests. Environmental Geology, 40: 1066-1074.

Siegel, D. I., 1988. The recharge-discharge function of wetlands near Juneau, Alaska: Part II. Geochemical investigations. Ground Water, 26(5): 580–586.

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

Suncor Energy, 2005. Environmental Impact Assessment, Voyager Project*.* Suncor Energy Inc.

Surridge, B.W.J., Baird, A.J., Heathwaite, A.L., 2005. Evaluating the quality of hydraulic conductivity estimates from piezometer slug tests in peat. Hydrological Processes, 19(6): 1227–1244.

Stewart, S.A., Lemay, T.G., 2011. Inorganic water chemistry of saline fens in northeastern Alberta (NTS 74D). Energy Resources Conservation Board, ERCB/AGS Open File Report 2011-09. Edmonton, Alberta.

Stocks, B.J., Fosberg, M.A., Lynham, T.J., Mearns, L., Wotton, B.M., Yang, Q., Jin, J-Z., Lawrence, K., Hartley, G.R., Mason, J.A., McKenney, D.W., 1998. Climate change and forest fire potential in Russian and Canadian Boreal Forests. Climate Change, 38: 1-13.

Sweetgrass Consultants Ltd., 1997. Environmentally Significant Areas of Alberta Volume 2. Calgary, Alberta.

Thornthwaite, C.W., Holzman, B., 1939. The determination of evaporation from land and water surfaces. Monthly Weather Review, 67: 4-11

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

Timoney, K., Lee, P., 2001. String and net-patterned salt marshes: rare landscape elements of boreal Canada, 115(3): 406-412.

Trites, M., Bayley, S.E., 2009. Vegetation communities in continental boreal wetlands along a salinity gradient: Implications for oil sands mining reclamation. Aquatic Botany, 91(1): 27–39.

Value Creations Inc., 2012. TriStar Pilot Project - Application and Project Development Plan. Calgary, Alberta, Canada.

van der Kamp, G., 2001. Methods for determining the in situ hydraulic conductivity of shallow aquitards - an overview. Hydrogeology Journal, 9(1): 5–16.

van der Kamp, G., Hayashi, M., 2009. Groundwater-wetland ecosystem interaction in the semiarid glaciated plains of North America. Hydrogeology Journal, 17(1): 203–214.

Vitt, D.H., 2006. Functional characteristics and indicators of Boreal peatlands. Ecological Studies, 188: 9–24.

Vitt, D.H., Chee, W.L., 1990. The relationships of vegetation to surface water chemistry and peat chemistry in fens of Alberta, Canada. Vegetatio, 89(2): 87–106.

von Post, Lennart, 1924. Das genetische System der organogenen Bildungen Schwedens. Memoires sur la nomenclature et la classification des sols: 287-304.

Waddington, J.M., Roulet, N.T., 1997. Groundwater flow and dissolved carbon movement in a boreal peatland. Journal of Hydrology, 19: 122–138.

Walter, L.M., Stueber, A.M., Huston, T.J., 1990. Br-Cl-Na systematics in Illinois basin fluids: Constraints on fluid origin and evolution. Geology, 18: 315-318.

Wilson, T.P., Gabet, R.M., 1991. Bromide uptake by soils and shales. Eos, Transactions American Geophysical Union, 72(44): 177.

Winter, T.C., Rosenberry, D.O., 1998. Hydrology of prairie pothole wetlands during drought and deluge: a 17-year study of the Cottonwood Lake wetland complex in North Dakota in perspective to long term measured and proxy hydrological records. Climate Change, 40: 189-209.

Woynillowicz, D., Severson-Baker, C., & Raynolds, M., 2005. Oil sands fever: The environmental implications of Canada's oil sands rush. Alberta: The Pembina Institute.

Xu, Shangping, Leri, A.C., Satish C.B. Myneni, Jaffe, P.R., 2004. Uptake of bromide by two wetland plants (*Typha latifolia* L. and *Phragmites australis* (Cav.) Trin. Ex Steud). Environmental Science and Technology, 38(21): 5642-5648.