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A hydrologic assessment of a saline-spring fen in the Athabasca oil sands region, Alberta, Canada – a potential analogue for oil sands reclamation

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Abstract:

Canada's post-mined oil sands will have a higher concentration of salts compared with freshwater peatlands that dominate the landscape. While rare, naturally occurring saline wetlands do exist in Alberta's Boreal Plains and may function as analogues for reclamation, however, little is known about their hydrology. This paper investigates the geochemical and hydrologic characteristics of a natural saline-spring peatland in Alberta's oil sands region. The fen is located within a saline groundwater discharge area connected to the erosional edge of the Grand Rapids Formation. Na^+ (195–25,680 mg l^{-1}) and Cl^- (1785–56,249 mg l^{-1}) were the dominant salts, and the fen transitioned sharply to freshwater along its margins because in part of subsurface mineral ridges that restricted shallow groundwater exchange. Salinity decreased from hypersaline to brackish along the local groundwater flow path but no active spring outlets were observed over the two-year study. Vertical groundwater discharge was minimal because of the very low permeability of the underlying sediments. Subsurface storage was exceeded during periods of high flow, resulting in flooding and surface runoff that was enhanced by the ephemerally connected pond network. These findings have implications for reclamation, as mechanisms such as subsurface mineral ridges may function as effective saline groundwater-control structures in the post-mined environment. Incorporating saline wetlands into regional monitoring networks will help to better quantify natural discharge, which has implications for belowground wastewater storage related to *in situ* bitumen extraction. Copyright © 2015 John Wiley & Sons, Ltd.

KEY WORDS wetland; salt; springs; water balance; oil sands; reclamation

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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 (AOSR) once fully operational (Woynillowicz *et al.*, 2005). Because of 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 (Purdy *et al.*, 2005; Trites and Bayley, 2009a). High concentrations of these ions may pose a serious challenge for wetland reclamation because of 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' peatlands within the AOSR. With oil sands production expected to double in the present decade (ERCB, 2011), reclamation strategies will require a sound understanding of the form and function of saline wetlands in the post-mined landscape.

Within the AOSR, wetlands comprise up to 65% of the landscape, the majority of which are fens (Andriashek, 2003; Devito *et al.*, 2012). 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 low-lying plains and river systems and may serve as appropriate reference analogues for reclamation (Grasby and Londry, 2007; Trites and Bayley, 2009a; Wells and Price, *in review*).

Naturally saline wetlands exist worldwide under a variety of settings, the most common of which are coastal wetlands that receive inputs of salts through tidal action

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(Price and Woo, 1988; Mitsch and Gosselink, 2007). 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, 2009; Heagle *et al.*, 2013). In northern Alberta, saline-spring wetlands are often found close to or along major river systems, where groundwater discharges to the surface through subcropping or exposed Paleozoic rocks (Hitchon *et al.*, 1969; Timoney and Lee, 2001; Grasby, 2006; Jasechko *et al.*, 2012). In some cases, these wetlands can be found far removed from riverbanks, as was observed by Wells and Price (in review), where a low flow discharge wetland received its salts from a near-surface aquifer. Species richness is often low and dominated by salt-tolerant vegetation, with typical peat-forming mosses such as *Sphagnum* absent (Timoney and Lee, 2001; Trites and Bayley, 2009a). 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 maritima*), Nuttall's salt meadow grass (*Puccinellia nuttalliana*), and alkali marsh aster (*Aster pauciflorus*) (Timoney and Lee, 2001).

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; Purdy *et al.*, 2005; Trites and Bayley, 2009a, 2009b). 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; Phillips, 2013). With this in mind, the overall objective of this study was to investigate the hydrologic and geochemical functions of a natural-saline ecosystem that could be used as a reference analogue for the saline post-mined landscape. Specific objectives are to (1) examine the source, storage, and transmission characteristics of a saline wetland through an assessment of its water balance components; (2) characterize spatial and temporal patterns of salinity; and (3) identify key processes in natural-saline wetlands that can be used to improve land-use decisions in the AOSR, particularly in the post-mined reclamation setting.

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 1). Cold winters and warm summers characterize the climate, with 30-year daily average temperatures of -18.8°C in

January to 16.8°C in July (Environment Canada, 2012). Annual precipitation (P) is low (455 mm), and the majority of annual P (65–75%) falls during the summer growing season (May–September) as convective cell storms with daily rainfall rates of >10 mm (Smerdon *et al.*, 2005; Devito *et al.*, 2012; Environment Canada, 2012). Potential evapotranspiration 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- to 15-year cycle (Ferone and Devito, 2004; Petrone *et al.*, 2007).

The fen lies at approximately 400 m.a.s.l within the McMurray lowlands subdivision of the Dover Plains, a relatively flat region characterized by widespread and continuous organic deposits (Andriashek, 2003). Glacial sediment is relatively thin around the study area (15–20 m) and overlies Cretaceous shale and sandstones of the Clearwater and bitumen bearing McMurray formations (Value Creations Inc., 2012).

The saline fen covers an area of approximately 27 ha, with a large pond network comprising ~19% of the fen surface. The fen's surface elevation declines northward and is characterized by a steeper gradient in the south (0.6% slope) that transitions into a gently sloping plain in the north half of the fen (0.2% slope) (Figure 2). The vegetation composition of the saline fen contains characteristic salt-tolerant vascular and non-vascular plant species. The dominant vegetation includes sweetgrass (*Hierochloa hirta* ssp. *arctica*) around the outer fringe, narrow reed grass (*C. stricta*) and foxtail barley (*Hordeum jubatum*) in the ridges, seaside arrow grass (*T. maritima*) and redwood 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 non-saline forested bog and/or fen peatlands. No significant moss cover is present in the main portions of the fen where there is elevated salinity (Borkenhagen, pers. comm.).

METHODS

Groundwater

Nests of wells and piezometers (2.5-cm inner diameter polyvinyl chloride) were installed in four transects across the fen and into adjacent wetlands (30 total nests; Figure 1) using hand augers. Each nest consisted of a slotted well and between 2 and 5 piezometers (17-cm slotted screens covered in well sock). Pipes were

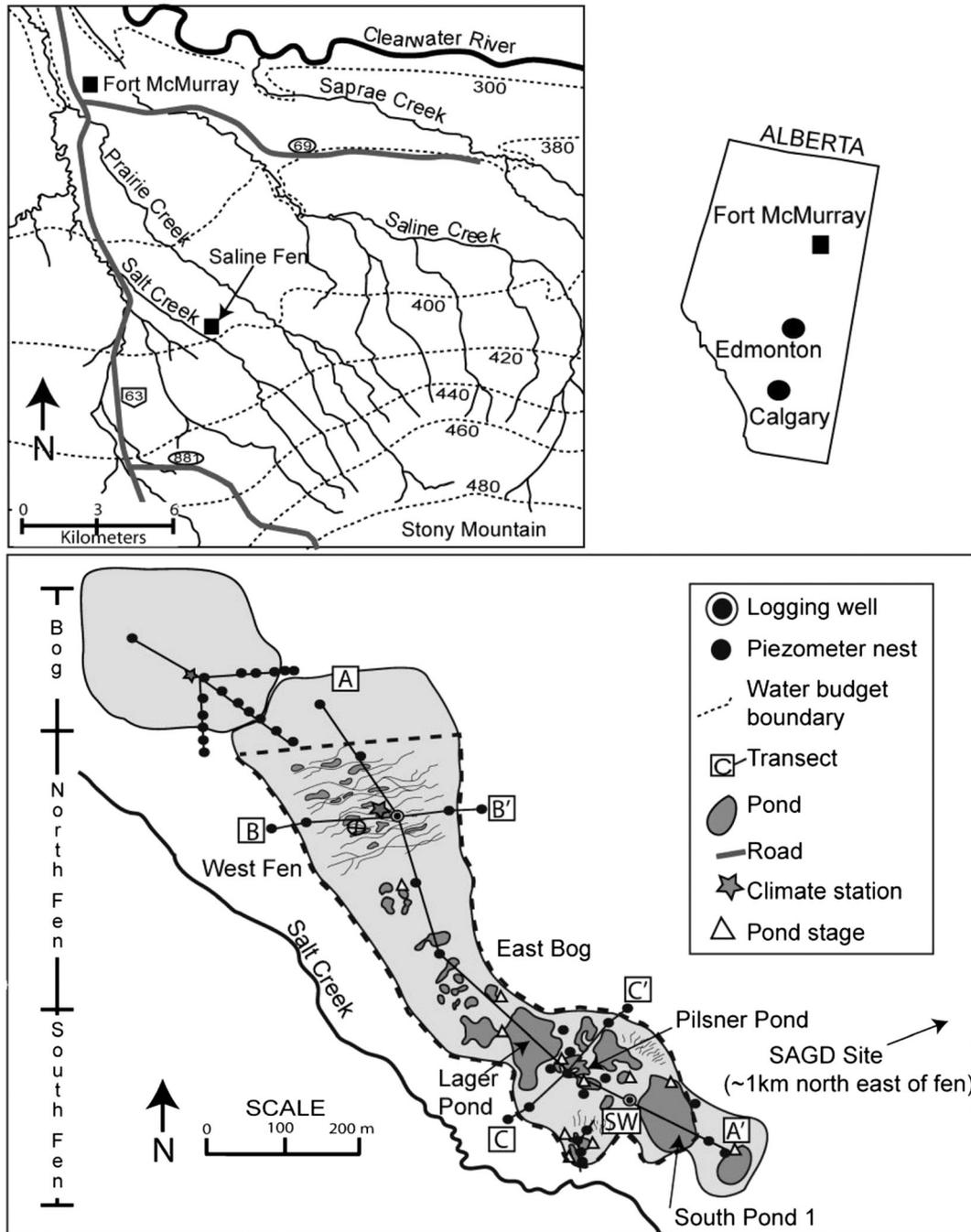


Figure 1. Map of the Saline Fen with site topography and location within Alberta on insets. Note the location of the future steam assisted gravity drainage (SAGD) facility just east of the fen. Vertical legend to the left of the map divides the site into three distinct sections that are referred to as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

measured manually at least once per week between June and September in 2011, and April and September in 2012. Continuous water level measurements were made at two wells located in the north and south ends of the primary transect (A–A', Figure 1) using pressure transducers. At the end of each study season, fen topography and pipe top elevations were measured and referenced to sea level using a dual-frequency survey-grade global positioning system.

Horizontal saturated hydraulic conductivities were measured in the field using bail tests for 30 piezometers in peat and 11 piezometers in the underlying till using the method of Hvorslev (1951). For shallow peat (<50 cm), the modified cube method was used (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

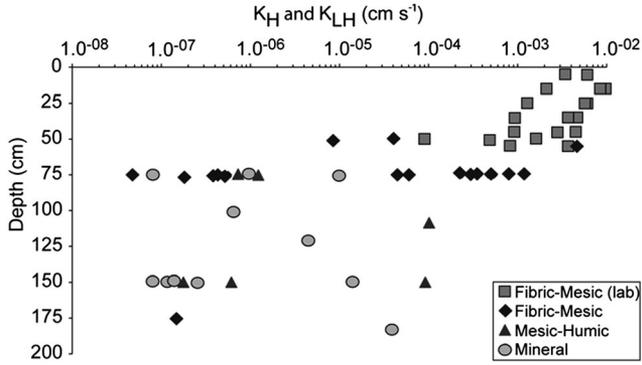


Figure 2. Field K_H (>50-cm depth) and laboratory K_{LH} (<50-cm depth) estimates for peat and mineral layer. Fen peat was divided into two main classifications based on the von Post system of humification (von Post, 1924) and the Canadian three-part fibre content system (Lemasters *et al.*, 1983)

$$Q = -KA \frac{dh}{dl} \quad (1)$$

where Q is the discharge ($\text{m}^3 \text{s}^{-1}$), 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.

Because of elevated groundwater salinities within deeper piezometers, changes in density had to be considered when calculating hydraulic heads. Groundwater with less than 10000 mg l^{-1} total dissolved solids (TDS) at temperatures below 100°C are considered to have densities sufficiently comparable with freshwater (1000 g l^{-1}); therefore, density corrections were only made for groundwater zones that exceeded this TDS threshold (typically within the south fen) (Freeze and Cherry, 1979). An average salinity value for each piezometer was based on one sampling period in 2011 (dry August conditions) and two sampling periods in 2012 (dry June and very wet August conditions). Groundwater temperatures for deep piezometers were assumed to be close to the average annual air temperature for Fort McMurray (0.7°C). Hydraulic heads were corrected to freshwater equivalents using the equation

$$hf = \left(\frac{\rho_p}{\rho_f} \right) hp \quad (2)$$

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 conversion of saltwater head to freshwater head resulted in an average increase in hydraulic head of $\sim 3 \text{ cm}$ and had a

negligible influence on estimation of vertical hydraulic gradients.

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 centred within the north half of the fen (Figure 1). Summer P was measured with a tipping bucket rain gauge supplemented with two manual bulk rain gauges. A tipping bucket malfunction in early July of 2011 required the use of rain gauge data (July–September) from the Fort McMurray airport weather station (AWOS-A configuration) 8 km northeast of the study site (Environment Canada, 2012). Air temperature and relative humidity were measured at 1.0 and 3.0 m, while net radiation (Q^*) and wind speed/direction were monitored at 3.0 m. 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

$$ET_a = \alpha \left(\frac{\Delta}{\Delta + \gamma} \right) \left(\frac{Q^* - Q_g}{L_v \rho_w} \right) \quad (3)$$

and where Δ is the slope of the saturation vapour pressure–temperature curve ($\text{Pa } ^\circ \text{C}^{-1}$), γ is the psychrometric constant ($\text{kPa } ^\circ \text{C}^{-1}$), 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_w is the density of water (kg m^{-3}). The coefficient of evaporability, α , is determined by regressing independent measures of ET measured empirically using soil lysimeters (Price and Maloney, 1994) to equilibrium evaporation (ET_{eq}), which is the value of Equation (5) under reasonably moist conditions with no advection ($\alpha=1$). Six lysimeters filled with peat monoliths representative of the two major peat surface classes (ridge/lawns and inter-ridge depressions, three repetitions each) were installed in the north fen (Figure 1) and used for determining ET_a and derivation of a site-scale α coefficient. The use of two major peat-surface types was based on field observations of surface elevation, vegetation cover and water table position. Air photo interpretation was then used to delineate ridge and inter-ridge depressions, lawns, and pond surface areas.

Because of installation and measurement errors associated with evaporation pans at the fen (for calculation of α for open water) the Penman equation (1948) was used to estimate daily potential evaporation (mm day^{-1}) for open-water surfaces with no vegetation (E_o), where

$$E_0 = \frac{\Delta Q^* + \gamma L_v \rho_w K_E v_a \{e_a^* - e_a\}}{L_v \rho_w \{\Delta + \gamma\}} \quad (4)$$

and where Δ is the slope of the saturation vapour pressure versus temperature relationship at the ambient air temperature ($\text{kPa}^\circ\text{C}^{-1}$), K_E is the mass transfer coefficient (kPa^{-1}) (Dingman, 2002), v_a is the velocity of air (m day^{-1}), e_a^* is the saturation vapour pressure at ambient air temperature (kPa^{-1}), and e_a is the water vapor pressure (kPa^{-1}). Estimates of ET_a 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. Because of the flooding in 2012, the fractional weighting of the peat and pond surface classes were adjusted to account for the increase in open-water surface at the fen.

Geochemistry & soil physical parameters

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. Piezometers were purged before sampling and groundwater samples were extracted using clear flexible PVC tubing connected to a foot valve rinsed with distilled water prior to each use. Unstable parameters (pH, temperature, electrical conductivity (EC)) were obtained *in situ* by a handheld device (YSI 63 m) calibrated before each day. Electrical conductivity measurements were corrected in the field to 25°C . Major ions were collected and frozen before laboratory analyses at the University of Waterloo. Samples were passed through a $0.45\text{-}\mu\text{m}$ filter, and saline samples were diluted before analyses. Alkalinity was determined using automated spectrophotometric analyses. Cations were measured by inductive-coupled plasma-optical emission spectrometry, while

anions were determined using a capillary ion chromatograph. Analytical error in concentration measurements was determined to be less than 5%. The concentration of TDS was estimated by summing the concentrations of the individual major ions (Fetter, 2001). For the determination of TDS, bicarbonate was converted by a gravimetric factor of 0.4917 (Hem, 1985). For determination of soil physical parameters, peat samples ($\sim 1\text{-m}$ length) were extracted along the A–A' transect using a Wardenaar coring device. Samples were frozen and shipped for processing at the lab. Cores were subdivided into 10-cm section, and physical parameters were determined using standard methods (e.g. Freeze and Cherry, 1979).

RESULTS

Substrate characteristics

Surface peat (0–10 cm) was weakly decomposed (bulk density $\sim 0.12\text{ g cm}^{-3}$) and highly fibric, and became more humic and amorphous at depth (Table I). At all depths, decomposing roots and stalks within the sedge peat were predominantly vertically oriented. Average bulk densities of the lower peat layers ($< 50\text{ cm}$ from the surface) ranged from 0.15 to 0.22 g cm^{-3} . Peat specific yield (S_y) was low and followed no depth-related pattern, averaging 0.05 (maximum of 0.12 and minimum of 0.02) for the entire fen (Table I).

The field-based geometric-mean horizontal hydraulic conductivity (K_{FH}) of the fen peat was $2.6 \times 10^{-4}\text{ cm s}^{-1}$, or 22 cm d^{-1} ($n=36$). Laboratory analyses of both horizontal (K_{LH}) and vertical (K_{LV}) saturated hydraulic conductivity for peat $< 50\text{ cm}$ provide a good indication of the level of anisotropy at the saline fen. Over 70% of the samples ($n=22$) had greater K_{LV} than K_{LH} (Table I). A geometrically averaged field estimate of mineral K_{FH} was

Table I. Peat physical properties and hydraulic conductivity measurements (K) for peat and mineral. ρ_b is bulk density and S_y is specific yield. The subscripts F and L for measurement of K refer to Field (*in situ*) and Lab, respectively. K_{LH} is saturated horizontal hydraulic conductivity, and K_{LV} is saturated vertical hydraulic conductivity. All data are means (geometric for K) of three or more samples

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

found to be $5.5 \times 10^{-7} \text{ cm s}^{-1}$, or 0.05 cm d^{-1} ($n=11$). Spatially, mineral K_{FH} varied between 10^{-5} and $10^{-8} \text{ cm s}^{-1}$, with the highest K_{FH} observed in the high-salinity zones in the fen's southern section (Figure 2).

Rainfall and evapotranspiration

During the 2011 study, between June and mid-September, 192 mm of rain fell compared with 333 mm in 2012 for the same period (Figure 3). In 2011, monthly rainfall totals were all below the long-term average for the Fort McMurray region, while the extended 2012 study season saw a total of 366 mm of rain between 1 April and 17 September. The most significant contribution occurred in July, when 40% (152 mm) of the study's total rainfall occurred.

The areal average daily ET_a rate for the fen peat surface in 2011 was 2 mm (maximum 5 mm d^{-1}) based

on a site-scale α coefficient of 0.75 for peat (Table II, Figure 3). Calculated ET_a losses from peat were greatest in the ridge/lawns, totaling 223 mm compared with 185 mm for the inter-ridge depressions despite an equivalent average ET_a (Table II). E_o from the pond network totaled 368 mm for the season (Table II). For the entire fen, including all peat surface classes and the pond network, ET_{site} totaled 236 mm between 1 June and 17 September. 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 5 mm in early July. Unlike 2011, ET_a was greatest in the inter-ridge depressions (406 mm) compared with the ridge/lawns (380 mm) (Table II). Total E_o from the pond network was 526 mm between 1 April and 17 September (Table II). The aerially weighted total ET_{site} for the entire fen in 2012 was 421 mm between 1 April and 17 September.

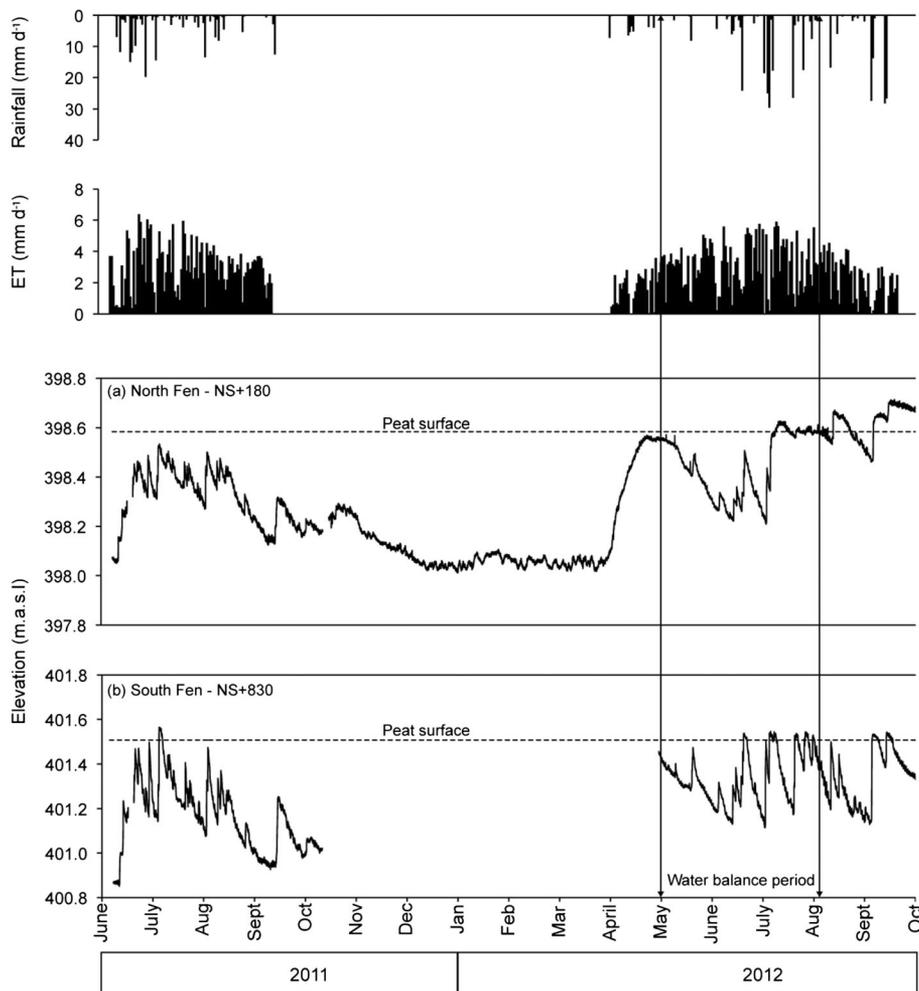


Figure 3. Micrometeorological conditions (P and ET_{site}) and water table elevations for the north (a) and south (b) portions of the fen during the 2011 and 2012 study seasons. Well locations can be found in Figure 1. The dashed horizontal lines on the hydrographs represent the peat surface. Dashed vertical lines bracket the water balance period calculated in 2012

Table II. Percent cover, α values for ET_a , E_o for ponds and ET_a for each peat-surface type along with a site-scale seasonal ET_{site} rate for the entire saline fen. 2011 values are from 1 June to 17 September (108 days), and 2012 is from 1 April to 17 September (170 days). Values of ET have been rounded to the nearest mm to account for uncertainty in methods

		Inter-ridge depressions	Ridge/lawn	Ponds
2011	% Total Cover	33	48	19
	α^\dagger	0.63	0.81	
	ET_a and E_o (mm d^{-1})	2	2	3
	Total ET_a and E_o (mm)	185	223	368
			236	
2012	Seasonal ET_{site} (mm^a)			
	α^\dagger	1.06	1.00	
	ET_a and E_o (mm d^{-1})	2	2	3
	Total ET_a and E_o (mm)	406	380	526
			421	

[†] Calculated only for peat

Groundwater and wetland connectivity

In 2011, the water table rose rapidly throughout the fen during the wet early summer period in June then declined gradually for the remainder of the summer season (Figure 3). A small but gradual rise in water table occurred between late December and early February, and this was followed by a stable water table regime up until the 2012 spring freshet in late March, where snowmelt replenished the winter storage deficit. Over 40% of the total rainfall for the 2012 study season fell during the month of July, satisfying the spring storage deficit developed in May. 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 4). In the south, these flooding events were episodic and followed by a rapid decline in water table.

Lateral subsurface flow patterns were consistent between years, and the overall topographic gradient drove a south-to-north groundwater-flow regime (Figure 4). Along the primary transect (A–A'), dh/dl were steepest (≥ 0.007) and more variable in the south fen (south of Lager Pond), with an average specific discharge rate of 2 mm d^{-1} . North of Lager Pond, dh/dl were also stable between years (< 0.004) and reflected the mild relief, decreasing gradually towards the bog-fen margin where dh/dl reversed and groundwater between the fen and north bog converged (Figure 4). An average rate of specific discharge for the north fen peat was calculated as 0.6 mm d^{-1} .

Groundwater exchange between the saline fen and surrounding wetlands was variable and sensitive to short-term weather changes; however, because of weak dh/dl , rates of specific discharge were typically no greater than 2 mm d^{-1} . In the north fen (B–B' transect), groundwater

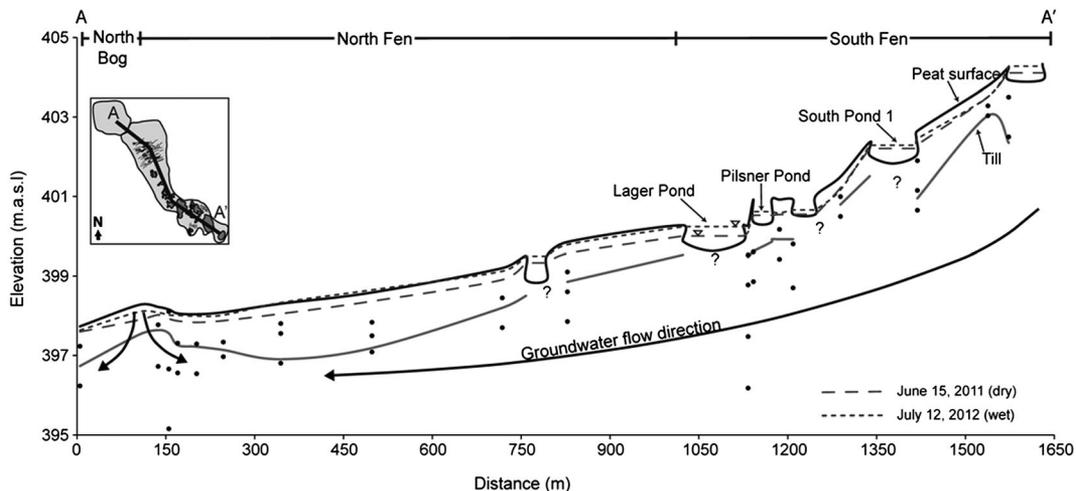


Figure 4. A cross-sectional profile of the fen along the primary A–A' transect. General fen boundaries are indicated by the legend along the top of the figure. (•) indicate the location of piezometer intakes while (?) refer to locations where depth to mineral till is unknown. The two dashed horizontal lines represent typical water table elevations under dry (15 June 2011) and wet (12 July 2012) conditions

discharged into the adjacent-treed west fen for both study seasons (Figure 5A, all dates), while along the east margin, ephemeral groundwater mounds typically impeded flow between wetlands (Figure 5A). In the south fen, water tables sloped towards the fen for both years (Figure 5B, all-dates). Groundwater mounding and flow reversals similar to the north fen were observed along the western margin of the C–C' transect (Figure 5B, 25 June 2012); during wet periods, these mounds were eliminated, initiating short lived (i.e. ~5 days) east-to-west flow through conditions (Figure 5B, 12 July 2012).

For both years, vertical dh/dz were generally weak, and strong spatial patterns were difficult to identify. However, because of the very low hydraulic conductivity of the underlying sediment (geometric mean of $5.5 \times 10^{-7} \text{ cm s}^{-1}$, minimum of $7.9 \times 10^{-8} \text{ cm s}^{-1}$), the authors were cautious in interpreting vertical flux patterns in greater detail because of possible inaccuracies in dh/dz calculations (e.g. piezometer time-lag errors in low hydraulic-conductivity substrates

and issues with piezometer recharge post-sampling). Nevertheless, the weak dh/dz observed coupled with low hydraulic-conductivity sediments resulted in estimates of deep-groundwater discharge and recharge that were negligible over the entire study. The pond network did not reveal any active spring outlets or discharge mounds.

Pond–peatland dynamics

Pond stage was spatially and temporally variable over the two-year study. While not directly measured in 2011, a steady summer decline in pond stage throughout the fen was visually recorded. In 2012, pond stage gradually declined after the spring freshet (Figure 3) until a series of high-intensity rain events in early July (day of year (DOY) 183–188, Figure 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

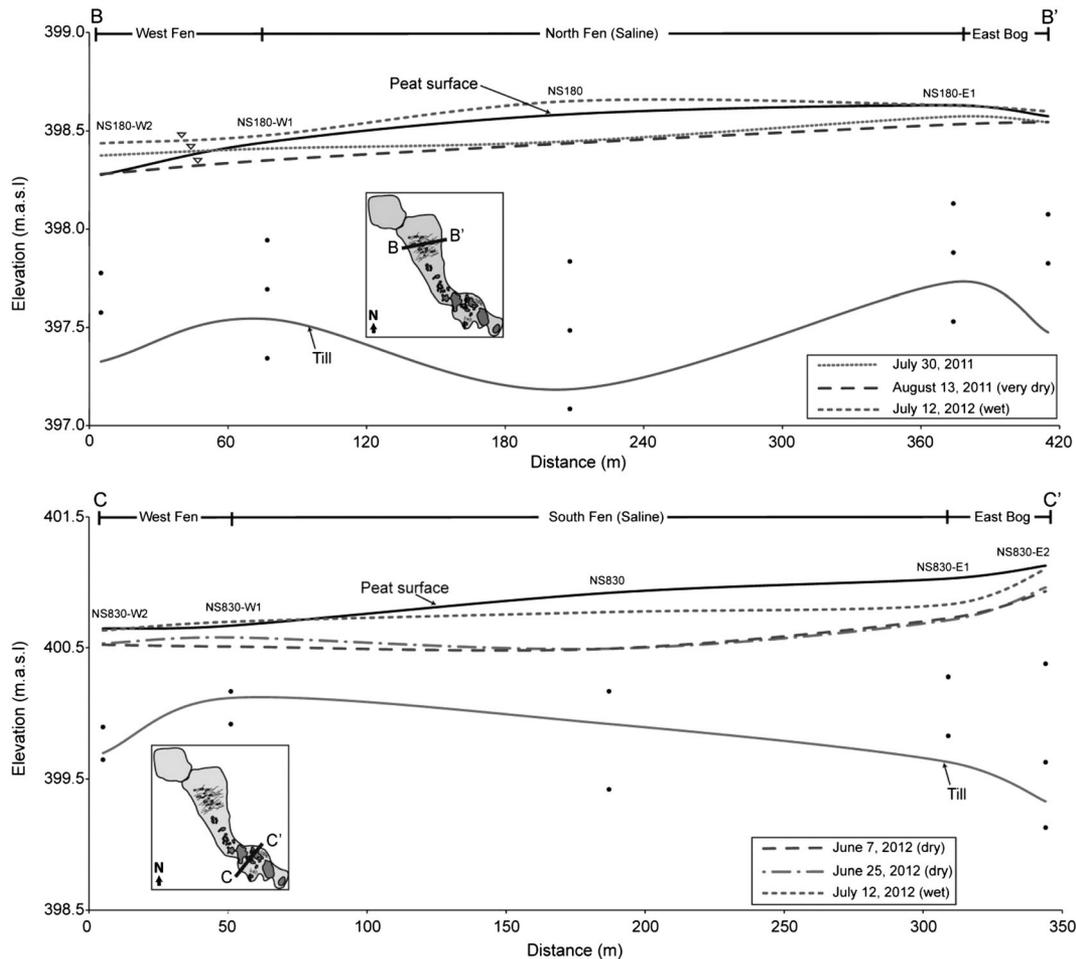


Figure 5. Cross-sectional profile of the B–B' in the north fen (a) and C–C' transect in the south fen (b) with typical daily water table trends over both study seasons. General fen boundaries are indicated by the legend along the top of the figure. (•) indicate the location of piezometer intakes, while nest identifiers are located above each piezometer nest (e.g. NS180). Note the difference in scales on the x-axis between cross-sections

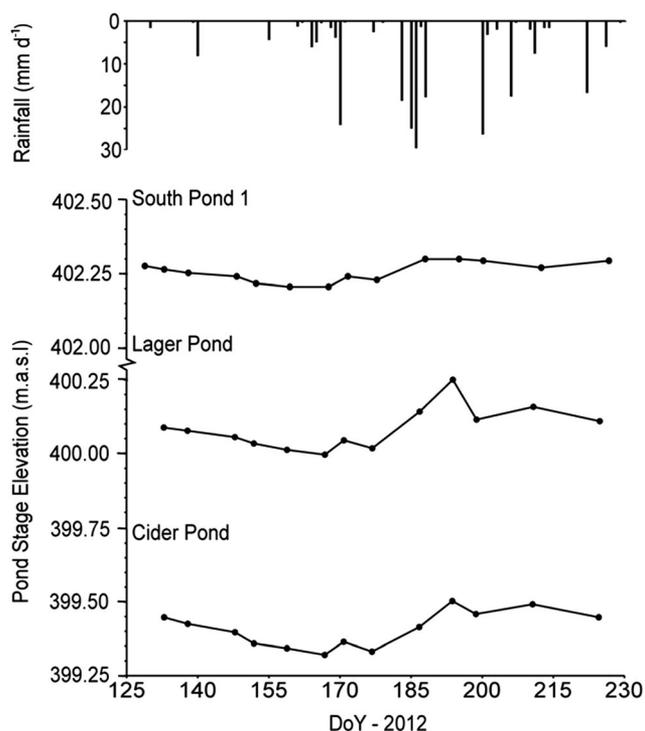


Figure 6. Pond stage elevations during the 2012 study period. Ponds were selected based on their position within the fen and can be found in Figure 1. Pond stage was measured manually

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 6).

Pond–peatland interaction was studied in detail in 2012 by examining a 0.25 ha, irregularly shaped saline pond in the south fen (Pilsner Pond, Figure 7). The pond's subsurface connection followed topography, receiving peat groundwater along its southern (up-gradient) margins (mean dh/dl of -0.02 , with $-dh/dl$ indicating groundwater inflow) while supplying fen groundwater along its northern (down-gradient) perimeter (mean dh/dl of 0.04). During wetting events, dh/dl along the pond's southern margins became steeper (mean dh/dl of -0.05 on DoY 187) while an opposite trend was observed along its northern margins (mean dh/dl of 0.02 on DoY 187). Groundwater interaction between the pond and peatland was dynamic, and complete flow reversals occurred in as little as 6 days during periods of low or intermittent rainfall.

Summary of water balance components

The water balance equation for the saline fen can be written as

$$P + G_{in} + D_{in} - E - G_{out} - D_{out} = \Delta S + \zeta \quad (5)$$

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 ζ is the residual error term. No surface runoff was observed between the fen and adjacent wetlands and thus is not included in the water balance equation.

Precipitation (217 mm) and water loss through evapotranspiration (280 mm) were the largest components over the water balance period (8 May 2012 to 9 August 2012). Subsurface connectivity between the saline fen and adjacent wetlands was minimal, with shallow groundwater inflow and outflow comprising less than 1 mm each over the 94-day period. Estimates of both deep groundwater recharge (2 mm) and discharge (5 mm) were low over the water-balance period because of the low hydraulic conductivity of the underlying mineral layer. Change in storage (ΔS) was estimated using the S_y of the peat and change in water table elevation (Δh), such that $\Delta S = \Delta h S_y$. ΔS was estimated to be 3 mm in the peat and 18 mm in the pools ($S_y = 1$). Based on pool and peat surface area, the net ΔS in the fen was approximately 6 mm and contrasts sharply with a negative ΔS calculated as the residual of Equation (5), in this case, -59 mm. A negative ΔS indicates that outputs exceeded inputs despite flooding in the north fen and an estimated ΔS of 6 mm, pointing to the possibility of an underestimation of inputs and overestimation of ET for this period.

Salinity patterns

Groundwater within peat (≥ 50 cm depth), the underlying mineral sediment, and the entire pond network was dominated by Na^+ (195 – $25\,680$ mg l^{-1}) and Cl^- (1785 – $56\,249$ mg l^{-1}) and to a lesser extent SO_4^{2-} (28 – 3080 mg l^{-1}) (Table III). In the north fen, Na^+ and Cl^- levels within the mineral sediments were higher on average than in the overlying peat, while this trend was reversed in the south fen (Table III). A clear spatial trend of decreasing EC and major ion concentration was observed northward (Figure 8, and Table III). Near-surface EC (< 50 -cm depth) 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). In the south fen, 'hotspots' of elevated EC were observed (> 30 mS cm^{-1}). EC and major ion concentrations were markedly lower within the adjacent wetlands; however, the concentration of Na^+ and Cl^- increased dramatically within the basal peat and underlying mineral sediment, most notably within the west fen (Table III). This trend was not observed for the north bog, where HCO_3^-

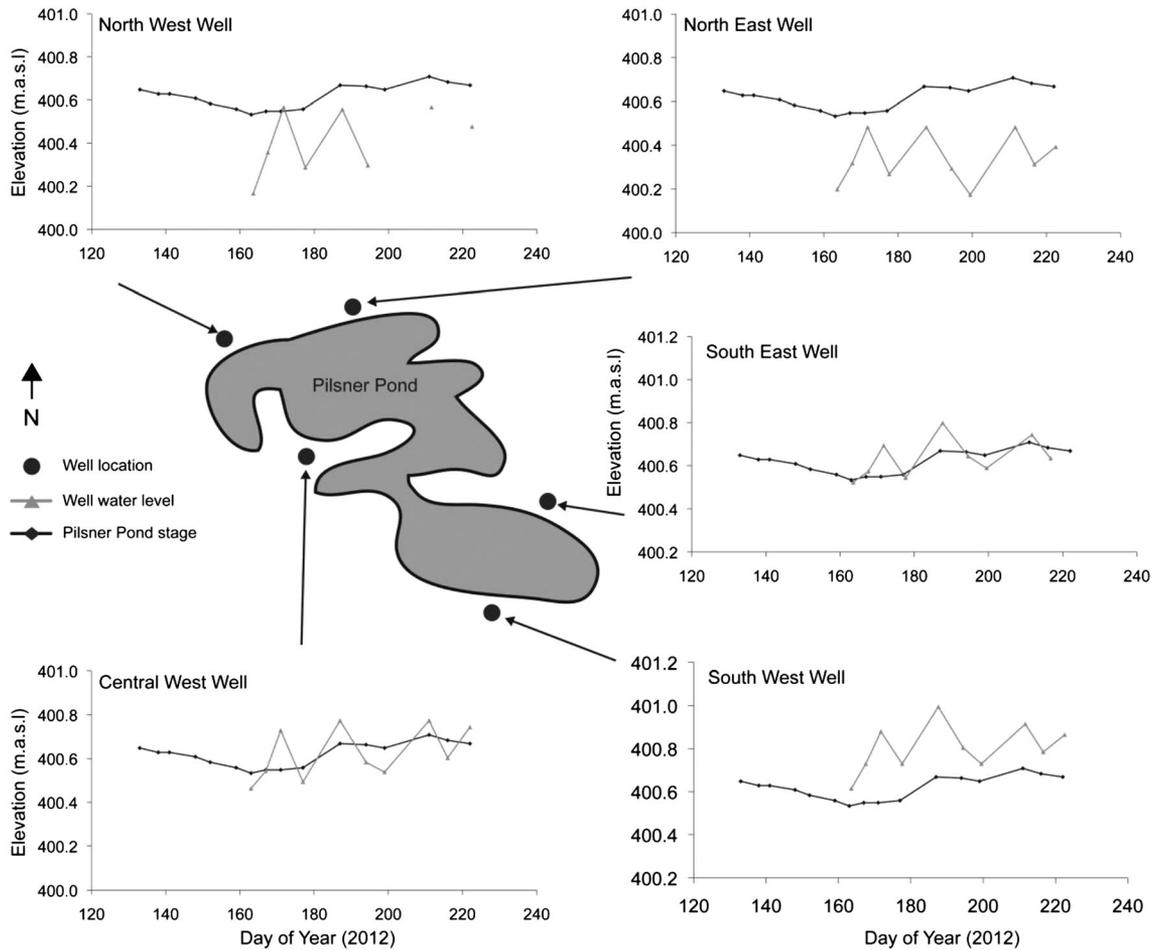


Figure 7. Pilsner pond-groundwater interactions over the 2012 study season. Pond stage and water table elevations were measured manually

formed the dominant anion, followed by Ca^{2+} and Mg^{2+} as the dominant cations (Table III).

DISCUSSION

Hydrologic setting

The dynamic hydrologic and geochemical characteristics of the fen were largely a function of its configuration within the landscape. A saline plume originating from the erosional edge of the lower Cretaceous Grand Rapids Formation was identified as a probable source for the accumulation of Na^+ and Cl^- salts at the surface (Wells and Price, in review). Despite the fen's proximity to a subsurface plume, deep-groundwater input was estimated to be low because of the influence of an underlying low-conductivity mineral layer (Figure 2; Figure 5), and no active discharge outlets were observed. However, it is unclear at the resolution of this study whether or not discrete spring outlets, diffuse seepage, or a combination of the two are driving the discharge of saline water at the fen. Potential errors with the piezometer network resulted

in many of the vertical dh/dz being discarded from the analysis. Thus, the true magnitude of discharge may be greater than what was estimated during this study and may represent a significant portion of inputs unaccounted for in the water balance. Sandy lenses with hydraulic conductivities several orders of magnitude greater than the site average were identified underlying the peat in the south fen, which may dramatically increase the bulk hydraulic conductivity of the system (Sharp, 1984; Keller *et al.*, 1988; Stephenson *et al.*, 1988; Haldorsen and Kruger, 1990; Hinton *et al.*, 1993).

Historically, variability in spring discharge has been observed throughout the region, with rates fluctuating or stopping entirely for some springs over the last 100 years (McKillop *et al.*, 1992; Grasby and Londry, 2007). Some ponds and high-salinity zones exhibited surface features associated with once active spring seeps (e.g. microbial biofabrics, Grasby and Londry, 2007), while preliminary analysis of microfossil assemblages in peat cores points to dramatic shifts in salinity over time (Wells and Price, in review; Volik, unpublished data). A period of low groundwater input at the fen may be indicative of the

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Table III. Major ion and total dissolved solids 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 three 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	26 338	17799	5668	306	10 543	335	3895	4	3
	Max	10 273	13 193	56 249	33 884	12 095						
	Min	2647	2362	9491	4624	1785						
HCO_3^-	Mean	179	135	203	132	84	71	65	80	99	397	347
	Max	391	241	627	318	129						
	Min	20	34	40	61	52						
SO_4^{2-}	Mean	80	96	1561	959	418	10	695	8		<1	3
	Max	157	157	3080	2741	1590						
	Min	28	43	512	277	48						
Ca^{2+}	Mean	183	243	721	534	150	24	650	30	157	36	22
	Max	262	435	1696	1235	515						
	Min	127	132	231	88	48						
Mg^{2+}	Mean	493	147	407	301	73	9	330	9	94	17	14
	Max	2875	299	903	758	208						
	Min	62	51	112	71	25						
Na+	Mean	2307	3491	13 015	9090	2903	151	5148	167	1937	3	5
	Max	5285	6596	25 680	15 920	6474						
	Min	108	1153	5534	2644	892						
Alk	Mean	147	111	166	108	69	59	54	65	81	325	285
	Max	321	197	514	261	105						
	Min	16	28	33	50	42						
TDS	Mean	10 284	12 004	42 114	28 857	9254	535	17 399	620			
	Max	16 327	20 801	87 708	54 361	20 947						
	Min	4175	4220	17 463	8060	2825						

sensitivity of the regional aquifer responsible for fen discharge to long-term drought and precipitation cycles (Winter and Rosenberry, 1998; Hayashi and Farrow, 2014), which is supported by the fact that the Boreal Plains region is currently in a long-term regional water deficit (Devito *et al.*, 2012).

Water storage and transmission

A thin peat profile (Figure 4) combined with a low S_y (Table I) reduced the fen's subsurface storage capacity, which was exceeded under periods of sustained rainfall (Figure 3). Water table fluctuations were exaggerated across the fen, most notably south of Lager Pond where the peat profile thinned and where the pond–pool network was most extensive (Figures 1 and 4). Flooding within the fen's northern reaches was enhanced by its position in the landscape and the ridge-dominated microtopography that collected and detained surface water. Steeper hydraulic gradients within the south fen supplied surface and subsurface flow northward, and this flow was enhanced when the water table was elevated because of higher hydraulic conductivities within upper peat layers

(Figures 2 and 4). In general, groundwater exchange was minimal and often restricted by transient groundwater mounds that mirrored the underlying mineral substrate, creating bands of elevated water table along the fen margins (Scarlett and Price, 2013) (Figures 4 and 5A and B). Groundwater exchange was sensitive to seasonal and short-term weather changes, and in some cases, water tables sloped against surface topography during drought conditions (Figure 5B, 7 June 2012), supporting the notion that topographically derived catchment delineation may be invalid in such settings (Devito *et al.*, 2005).

Similar to other studies, the ponds and pools became interconnected during very wet periods and functioned as a drainage network for surface flow northward (Price and Maloney, 1994). Although not quantified in this study, these infrequent wetting events are likely important in the flushing and redistribution of solutes throughout the fen, and the actual magnitude of solute fluxes during these events warrants further exploration. The connectivity of the pond network was spatially variable, and following rain events, some ponds filled rapidly then dried gradually until the next wetting event, suggesting many are semi-

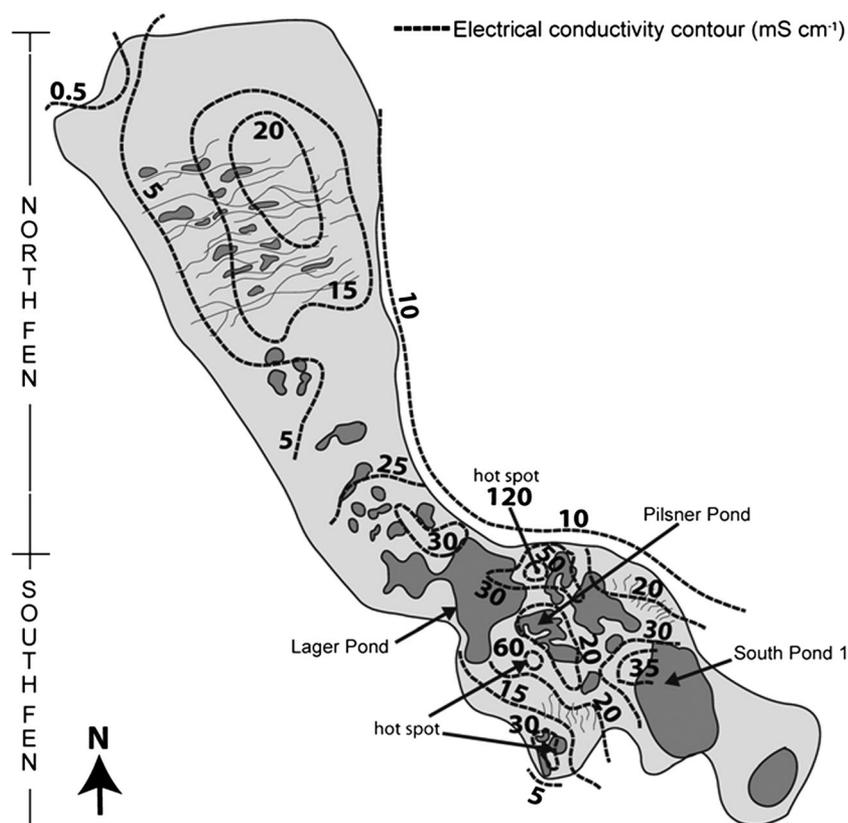


Figure 8. Near-surface groundwater salinity distributions for the fen in 2011 shown as contours of electrical conductivity. Note the location of the saline 'hot spots' in the south fen

permanent features that are sensitive to short-term weather changes. The influence of these open water ponds as water-loss features on fen hydrology is strong, as indicated by the high rates of evaporation for both years and the complete dry out of almost all ponds in 2011 (Table II). While groundwater helped other ponds resist drying out completely as a result of seasonal drought conditions and long-term water deficits typical for the region, these inputs were short-lived (Figure 7) (Smerdon *et al.*, 2005; Devito *et al.*, 2012). Thus, under periods of high water stress, even large ponds fed by both rainfall and groundwater can be susceptible to dry out within relatively short time-scales.

Water budget of a saline fen

The largest probable sources of error in the closure of the saline fen water balance were the calculation of site-scale ET and groundwater exchange. The Priestley and Taylor (1972) method for calculating ET is accurate to within $\pm 15\%$ under ideal conditions, and error is compounded by impacts related to the accuracy of lysimeter measurements (e.g. weighing errors) (Allen *et al.*, 1991). Because of the remote nature of the site, extended periods of time elapsed between lysimeter

measurements. A relatively small sample size could also have led to inaccurate estimations of ET at the fen for both years, especially because seasonal variation could not be accounted for. Because the meteorological station was not located over or adjacent to a pond, factors that are not representative of the true pond surface had to be used for the estimation open water evaporation; the influence of pond storage both diurnally and seasonally was not accounted for.

The definition of pond surface area and an underestimation of groundwater input likely contributed to the exaggerated negative ΔS term (-59 mm) in the fen water balance. The large, irregularly shaped pond network could not be monitored entirely, and a fixed pond surface area of 19% was used for open water E , assuming ponds were at bank-full width. Consequently, pond E was overestimated during baseflow or drought conditions when many ponds began to dry up. Errors in estimating shallow groundwater fluxes 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. 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 along the fen margins was

not observed, it is possible that some surface water exchange occurred during periods of inundation. The estimation of deep groundwater input through the underlying mineral sediments may also have been underestimated because of the potential influence of high-conductivity zones not captured by the piezometer network.

Salinity dynamics

Na and Cl dominated the ground and surface waters within the fen, consistent with other springs in the AOSR connected to regional aquifers found within subcropping Devonian-aged carbonate and Cretaceous formations (Hitchon *et al.*, 1969; Grasby, 2006). Salinity (EC, Cl^- , and Na^+ , Table III) was markedly higher than other natural-saline wetlands observed in the region (Purdy *et al.*, 2005; Trites and Bayley, 2009a, 2009b). While north fen salinities compared more closely to what has been observed in the natural-saline wetland setting, south of Lager Pond the concentration of dissolved salts were proportional to the $\text{Na}^+ - \text{Cl}^-$ dominated high flow-brine springs of the Fort McMurray and Wood Buffalo regions (Grasby and Londry, 2007). Despite such elevated salinities, the fen contained peat-forming plant assemblages (e.g. *T. maritime*, *P. eriopoda*, and *H. jubatum*) that have been shown to survive in oil sands process waters with typically much lower salinities (Purdy *et al.*, 2005; Trites and Bayley, 2009a). However, in the south fen where mean Cl^- was over $20\,000\text{ mg l}^{-1}$, a clear shift in vegetation was observed and halophytes such as *Salicornia* and other non-peat forming vegetation comprised a large portion of the landscape (Borkenhagen, unpublished data). In the context of vegetation establishment, the saline fen functions as a suitable reference analogue, with a natural spatial-salinity gradient that provides some indication to the upper limits of salinity that can be expected before dramatic shifts to non-peat forming vegetation may occur in the post-mined setting.

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, 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 and the fen's hydrogeologic position (Figure 8, Table III). The fen's topographic gradient promoted the lateral movement of groundwater northward, and the gradual reduction in slope north of Lager Pond worked to retard subsurface flow where it is diluted by precipitation (Figure 4). 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 (Redfield, 2001). Along fen margins near surface salinity was low, but within the basal peat and underlying mineral layers of the adjacent wetlands salts increased dramatically. Additionally, high TDS concentrations were found along nearby tributaries (e.g. Salt Creek), and discharge features similar to the fen were found throughout the study area (Wells and Price, in review), suggesting that saline discharge plays a much larger role in groundwater quality in the region.

CONCLUSIONS AND IMPLICATIONS FOR OIL SANDS RECLAMATION

The reclamation of peatlands within the post-mined setting will be complicated by elevated ground and surface-water salinities (Renault *et al.*, 1998; CONRAD, 1999; Purdy *et al.*, 2005). Consequently, reclamation strategies should focus on wetland designs that can adapt to high-salinity conditions and support salt-tolerant vegetation communities. The management of salt within the reclaimed landscape as a whole will also need to be considered. Although rare, saline wetlands do exist in the AOSR and serve as natural benchmarks by which wetland reclamation plans can be compared.

At the saline fen, the lack of peat-forming freshwater vegetation (e.g. *Sphagnum*) combined with a dense and generally shallow peat profile reflects what can be expected in the reclaimed wetland setting (Johnson and Miyaniishi, 2008; Trites and Bayley, 2009a). The low storage potential of the peat enhanced the fen's response to rainfall causing flooding, which more closely mimicked the shallow open-water marshes typical of current reclamation prescriptions (Rooney and Bayley, 2011). Managers should consider the effect of peat properties and depth on reclamation design, as low storage can potentially enhance the rapid transmission of water and salts across landscapes. However, the flushing of salt is also an important mechanism by which wetlands can effectively manage salinity. The capacity of a constructed wetland to flush salts should influence their placement within the landscape, where ecosystems sensitive to salt should be located away from the local flow path. Freshwater wetlands placed at the head of the watershed can provide enhanced flushing potential by providing a freshwater source to the saline wetland below. As was seen at the saline fen, a well-developed pond network can enhance this flushing potential during wet periods. However, reclamation managers should consider the

susceptibility of open water features to drying. A connection to a groundwater source may prevent complete drying but wetland–pond connectivity may be difficult to predict in the subhumid Boreal climate.

Belowground, the exchange of water between the saline fen and adjacent wetlands was controlled in some cases by small mineral ridges running along the wetland perimeter. Water table mounds formed as a result of these ridges effectively restricted groundwater exchange between the saline fen and adjacent freshwater wetlands, creating sharp boundaries between saline and non-saline wetlands. In situations where managers may need to limit the hydrologic connectivity between landscape units, such as when the placement of a reclaimed saline wetland beside a freshwater ecosystem cannot be avoided, subsurface mineral ridges along wetland boundaries may function as effective groundwater control structures. While the exchange of saline groundwater was minimal across the saline fen boundary, high salinities within nearby streams and basal peat and mineral layers of adjacent wetlands throughout the study area point to regional-scale saline discharge that would otherwise not be apparent based on landscape type and surface observations alone.

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. Despite flooding within the fen's lower reaches in 2012, runoff across the fen was minimal, because in part of 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. While a general trend of decreasing salinity northward (down-gradient) was observed, the link between hydrologic function and geochemistry was not always clear. Obtaining a reliable estimate of deep groundwater input was challenging because of the dense underlying mineral sediments and difficulty in capturing locally conductive zones accurately. This has implications for in-situ operators storing wastewater within the subsurface, as the potential for leakage and the upward migration of disposal fluids may be enhanced by the presence of surface features such as springs (Carrigy and McLaws, 1973; Hackbarth and Nastasa, 1979; Bachu *et al.*, 1989; Gordon *et al.*, 2002). The apparent difficulty in estimating groundwater input in systems underlain by low permeability sediments may complicate the development of an accurate baseline for natural-saline discharge, which would be of considerable interest to *in situ* operators in the AOSR.

The sharp salinity gradient observed at the fen, from slightly brackish in its northern reach to hypersaline along its topographic high, appears to parallel a shift to non-peat forming vegetation. Nevertheless, the fen supported herbaceous species determined to be best suited to the saline post-mined environment despite ground and surface-water salinities much higher than other natural-saline wetlands targeted as analogues for oil sands reclamation (Purdy *et al.*, 2005; Trites and Bayley, 2009a). Field-based evidence of the upper limits of salt tolerance in the natural setting allows reclamation managers to better gauge the trajectory of wetland development in the post-mined environment.

This study highlights the baseline hydrologic setting of a saline wetland in the AOSR. Because of the likelihood of increased salinity within the post-mined landscape, naturally saline wetlands constitute potential reference analogues for oil sands reclamation. Furthermore, because of the influence of salinity on ecosystem-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 reclamation strategies and design within the post-mined oil sands environment.

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