Journal of Hydrology 246 (2001) 142–154

Journal of Hydrology

www.elsevier.com/locate/jhydrol

Groundwater flow patterns in a large peatland

C.J.D. Fraser[a], N.T. Roulet[a,*], M. Lafleur[b]

[a]Department of Geography and the Centre for Global and Climate Change Research, McGill University, Burnside Hall, 805 Sherbrooke St. W., Montréal, Québec, Canada, H3A 2K6

[b]Department of Geography, Trent University, Peterborough, Ontario, Canada K9J 7B8

Received 27 June 2000; revised 28 December 2000; accepted 26 February 2001

Abstract

Groundwater flow patterns and geochemistry were studied in the Mer Bleue bog, near Ottawa, Ontario, Canada. Groundwater flow patterns alternated between recharge, i.e. head gradients producing flow from the surface of the peatland to the deeper peat, and discharge, i.e. head gradients indicating flow from the deeper peat towards the surface of the peatland, during the summer of 1998. The patterns were controlled by changes in precipitation, evapotranspiration and the differential head response of catotelm peat (lower layer) to changes in water table elevation. Above-average rainfall in the spring created recharge patterns of groundwater flow in the peatland. Evapotranspiration exceeded precipitation for a three-week period in mid-summer decreasing head at the water table and reversing flow from recharge to discharge. A sustained moisture deficit maintained a flow reversal for 32 days, until a 46 mm rainfall raised the water-table, reversing the vertical hydraulic gradients and restoring recharge flow.

The mixing of meteoric water and deeper groundwater controls geochemical profiles. Diffusion modeling shows that the peatland is a long-term recharge system. However, electrical conductivity and cation concentrations in peat pore-waters increase (up to 60%) during a flow reversal, and then decrease when recharge conditions are re-established. The redistribution of substrates from reversed flow is likely important to peatland biogeochemical function. © 2001 Published by Elsevier Science B.V.

Keywords: Peatlands; Groundwater; Flow reversals; Peat hydraulics; Biogeochemistry

1. Introduction

Northern peatlands store one-third, or approximately 450 Gt (C) of the global soil carbon pool, and cover an estimated $3.5 \times 10^6\ km^2$ (Gorham, 1991). The hydrology and biogeochemistry of peatlands can, in part, be attributed to their hydrogeologic setting and connection to local, intermediate and regional scale flow systems (Winter and Carr, 1980; Siegel, 1981; Winter, 1999). Hydro-biogeochemical studies have illustrated the importance of groundwater flow patterns in the distribution of vegetation (Glaser et al., 1981), the pattern and dynamics of water chemistry (Siegel, 1988a,b; Branfireun et al., 1996), trace gas exchanges (Romanowicz et al., 1993; Waddington and Roulet, 1997), and metal transport (Hill and Siegel, 1991). Peatlands can be considered important sources, sinks and transformers of key environmental metabolites. However, the limited number of studies of the hydro-biogeochemistry of peatlands makes extrapolation of results between different hydrogeologic settings and scales difficult.

Advances in peatland hydrology may have been

* Corresponding author. Fax: +514-398-7437.
E-mail address: roulet@felix.geog.mcgill.ca (N.T. Roulet).

0022-1694/01/$ - see front matter
PII: S0022-1694(01)00362-6

hindered by the misconception that groundwater advection in organic soils (catotelm) is negligible (Ingram, 1981), despite evidence showing the contrary (Charman et al., 1992; Chanton et al., 1995). Striking examples illustrating the importance of regional groundwater flow and advection in catotelm peat are groundwater flow reversals (Doss, 1993; Romanowicz et al., 1993; Siegel et al., 1995). These authors speculate that draw-down attributed to evapotranspiration is sufficient to decrease hydraulic potential in the local flow system (wetland) to induce discharge of regional groundwater through the underlying mineral substrate to the wetland. Such flow reversals can be sustained on seasonal to yearly time-scales, and dramatically alter peatland biogeochemical function (e.g. episodic CH_4 emissions (Romanowicz et al., 1993); changes in pore-water chemistry (Siegel et al., 1995)).

Flow reversals have also been documented in small peatlands (<0.12 km^2) isolated from regional groundwater systems (Devito et al., 1997). Devito et al. suggested that water deficits and subsequent water-table draw-down reverse groundwater flow patterns from recharge, i.e. head gradients producing flow from the surface of the peatland to the deeper peat, to discharge, i.e. head gradients indicating flow from the deeper peat towards the surface of the peatland. The discharge flow patterns that occurred in such isolated hydrogeologic settings were also important to peatland biogeochemical function. For example, Waddington and Roulet (1997) showed that a reversal of pore pressure gradients coincided with episodic methane emissions ~25 times greater than that of baseline fluxes.

To understand flow reversals, we carried out a hydro-chemical study in a large peatland in south-eastern, Ontario. Our objectives were to: (1) study the groundwater flow patterns in a large peatland that was isolated from regional groundwater flow; (2) determine the controls on groundwater flow patterns; and (3) assess whether changes in pore-water chemistry are traceable to recharge and discharge patterns of groundwater flow. We hypothesize that flow reversals could occur in large peatlands isolated from regional groundwater flow, and that the mechanisms controlling the flow patterns are differential variations in precipitation and evapotranspiration.

2. Study site and methods

2.1. Study site

The Mer Bleue bog (45°30′ N latitude and 75°25′ W longitude) is a 28 km^2 temperate peatland characterized by several raised peat domes and three drainage fingers that extend westward to the margins of the peatland (Fig. 1a). The fingers are remnant glacial meltwater streams that incised marine clay sediments of the glacial Lake Champlain bottom, after isostatic rebound forced sea waters eastward to the Atlantic Ocean ~9500 years ago (Mott and Camfield, 1969). Carbon dating of basal peat suggests that terrestrialization occurred on the saline sediments of the basin ~8500 years ago (Pierre Richard, pers. comm.). Seismic surveys and drilling in the Mer Bleue area indicate underlying marine clay thickness ranges from 12–45 m (Hobson, 1969; Belanger and Harrison, 1977).

The climate of the Mer Bleue is classified as cold humid continental, and mean annual temperature, mean annual precipitation and growing season length are 5.8 °C, 910 mm, and 193 days, respectively (Environment Canada, 1998). Dominant moss species on the bog are *Sphagnum* spp. (*fuscum, magellanicum* and *rubellum*), and heath and grass species such as Labrador tea (*Ledum groenlandicum*), leatherleaf (*Chamaedaphne calyculatae*), blueberry (*Vaccinium myrtilloides*), cotton grass (*Eriophorum* sp.) and sedge (*Carex oligosperma*) dominate the micro-canopy. There is a sparse cover of black spruce (*Picea mariana*), tamarack (*Larix laricina*) and white paper birch (*Betula populifolia*).

2.2. Hydrology

The north finger of the Mer Bleue bog was instrumented for measurements of runoff (R), evapotranspiration (ET) and precipitation (P) (Fig. 1a and b). The surface of this basin slopes from east to west, at a gradient of 0.0008. Surrounding uplands are small in area and well drained by networks of ditches, thus the contributing area was considered to be exclusively peatland (4.8 km^2). Stream velocities were gaged at the basin outflow and discharge estimates were determined from rating curves and continuous

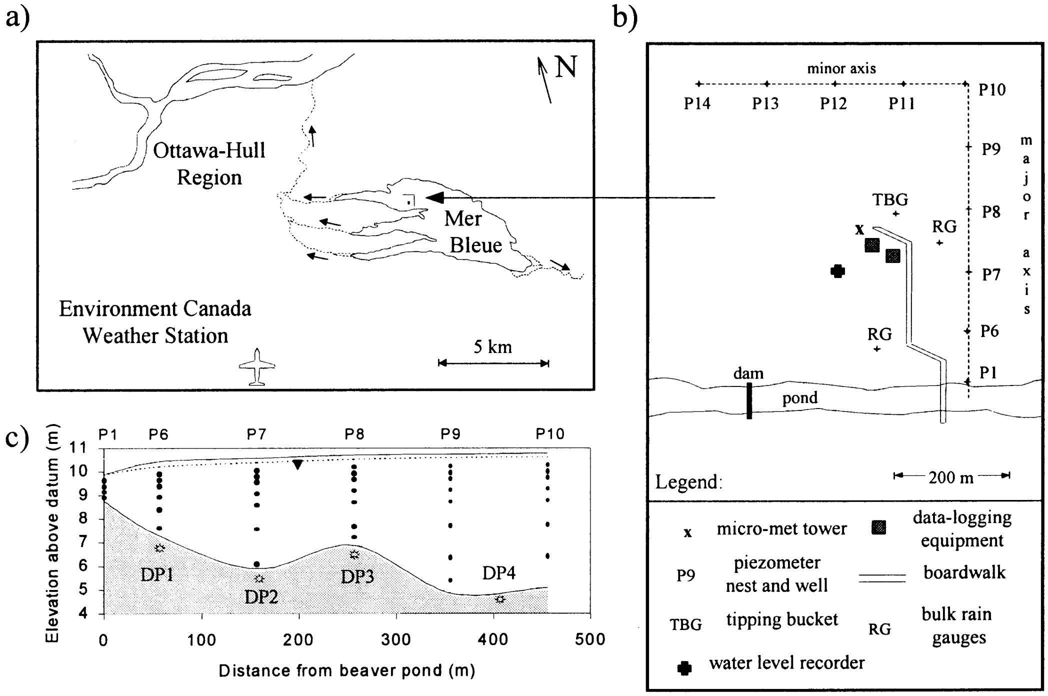

Fig. 1. (a) Map of the Mer Bleue bog showing the three western-draining basins and the eastern-draining basin with respect to Ottawa, Ontario, Canada. The peat dome on the northern drainage finger (see transects and arrow) was instrumented for this study. (b) Schematic of field instrumentation at the Mer Bleue bog. (c) Cross-section of the major axis illustrating topography, peat depth and piezometer coverage. Black dots denote piezometer locations, stars denote the locations of drive-point piezometers (DP).

measurements of stage. Runoff was calculated from discharge and basin area.

ET was estimated from continuous eddy covariance measurements of the latent heat flux (Q_E) (Baldocchi et al., 1988; Verma, 1990). Rainfall was measured with two bulk rain gages and a tipping bucket rain gage. Bulk gages agreed with rainfall estimates from the Ottawa International Airport, Environment Canada weather station. The patterns of rainfall measured by the tipping bucket rain gage correlated well with daily rainfall data from the weather station ($r = 0.74$), but underestimated total rainfall measured by bulk gages. To standardize the tipping bucket record for total rainfall, a multiplier was applied to daily rainfall measurements based on bulk and inter-station measurements. Snowfall was measured with a snow gage equipped with a Nipher shield.

Peatland topography was surveyed with a Total Station and the major and minor topographic gradients were determined. Piezometers and wells were inserted across two perpendicular transects to capture these gradients (see Fig. 1b). Piezometers were constructed from 1.25 cm ID SCH 80 PVC pipe coupled to 20-cm-long slotted heads and sheathed with 40 μm Nitex mesh. Manual observation wells were constructed from 5 cm ID PVC pipe (nests P6–P14, Fig. 1b). A well situated near the eddy-covariance tower was constructed from 15 cm ID PVC pipe and instrumented with a continuous water level recorder.

Piezometers inserted into peat were 0.5, 0.75, 1.0, 1.5, 2.0, 3.0 and 4.5 m long, where depth permitted. A cross-section of the major axis (P1, P6–P10) illustrates piezometer coverage, and the topography of the peat surface and peat-clay interface (Fig. 1c). Four drive-point piezometers were inserted into the underlying marine clay to sample interstitial pore-waters and assess groundwater exchange across the base of the peatland (Fig. 1c). For this, standard 30 cm Waterloo drive-point heads with 180 μm shields, sampling tube and 2 cm diameter galvanized pipe were used. A theodolite was used to determine the elevation of all piezometers, drive-points and wells on two occasions over the summer 1998.

Horizontal hydraulic conductivities (K_h) and mean hydraulic conductivities (K_m) were estimated from bail tests after Hvorslev (1951). K_h were determined from piezometers with slotted heads (P1, P6–P14), whereas K_m were determined from additional piezometers constructed without slotted heads (open at two ends, Nitex covered at lower extent). Vertical hydraulic conductivities (K_v) were calculated from estimates of K_h and K_m:

$$K_m = \sqrt{(K_h \cdot K_v)} \quad (1)$$

after Freeze and Cherry (1979), and anisotropy was estimated by comparing ratios of K_v:K_h.

3. Geochemistry

Measurements of electrical conductivity (EC) were performed on surface water, groundwater and precipitation samples using a Consort K220 Microcomputer Conductometer. Prior to measurement, the meter was calibrated for temperature and EC using a standard solution of 0.01 M KCl. Concentrations of Na^+, Mg^{2+} and Ca^{2+} in surface water, groundwater and precipitation samples were analysed at the Earth and Planetary Sciences Geochemistry Laboratory, McGill University, using flame atomic absorption analysis. Detection limits for Na^+, Ca^{2+} and Mg^{2+} analyses were 0.02, 0.04 and 0.01 mg L^{-1}, respectively.

Observed concentrations of solutes in peat profiles were compared to predictions from a one-dimensional diffusive transport equation to evaluate peatland recharge-discharge function (see Siegel, 1988b). The diffusion model as summarized by Crank (1975) is:

$$C = C_0/2[\mathrm{erfc} \cdot z/\sqrt{(D^* \cdot t)}] \quad (2)$$

where C is the calculated concentration, C_0 is the initial concentration at the base of the peat profile, z is a depth (L), D^* is the effective coefficient of diffusion (L^2/T), t is time and erfc is the complementary error function. D^* was determined by:

$$D^* = D_0 \cdot \theta^2 \quad (3)$$

where D_0 is the coefficient of molecular diffusion for cations in solution (L^2/T) and θ compensates for effective peat porosity and tortuosity (dimensionless). θ and D_0 were 0.22 and 2.0×10^{-9} $m^2\,s^{-1}$ respectively, after Freeze and Cherry (1979).

4. Results

4.1. Hydraulic conductivities and flow boundaries

Estimates of horizontal hydraulic conductivity (K_h) are shown in Fig. 2. Acrotelm (~0–0.45 m) K_h ranged from 10^{-7} to 10^{-3} m s^{-1} ($n = 36$), increasing in magnitude toward the peat surface. K_h estimates of catotelm peat ranged from 10^{-8} to 10^{-6} m s^{-1} ($n = 86$), but showed no pattern with depth or across the groundwater network. K_m estimates ($n = 9$) were typically an order of magnitude less than K_h measurements for corresponding depths. K_v estimates were 1–3 orders of magnitude less than K_h estimates (Eq. (1)), and average anisotropy was ~450.

K_h estimates of the underlying marine clay were $\leqslant 10^{-10}$ m s^{-1} ($n = 4$), and head measurements were higher in the peatland than in the clay. Since the marine deposits are continuous and thick (12–45 m), hydraulic gradients are from the peat to clay, and the K_h of the marine clay is very small, the flux of water

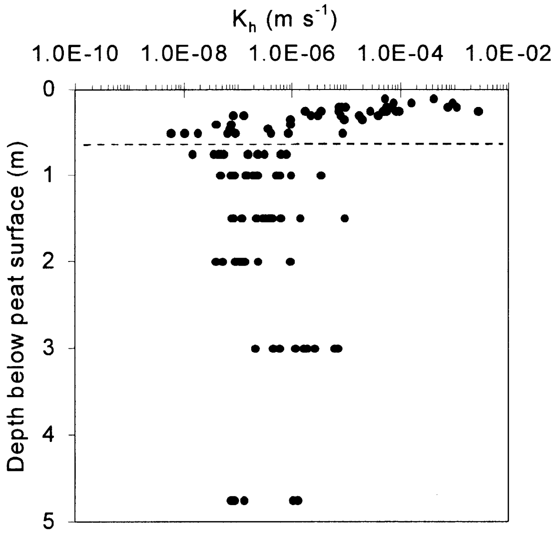

Fig. 2. Horizontal hydraulic conductivity (K_h). Acrotelm and catotelm K_h estimates are separated by the dashed line at ~0.45 m below the peat surface.

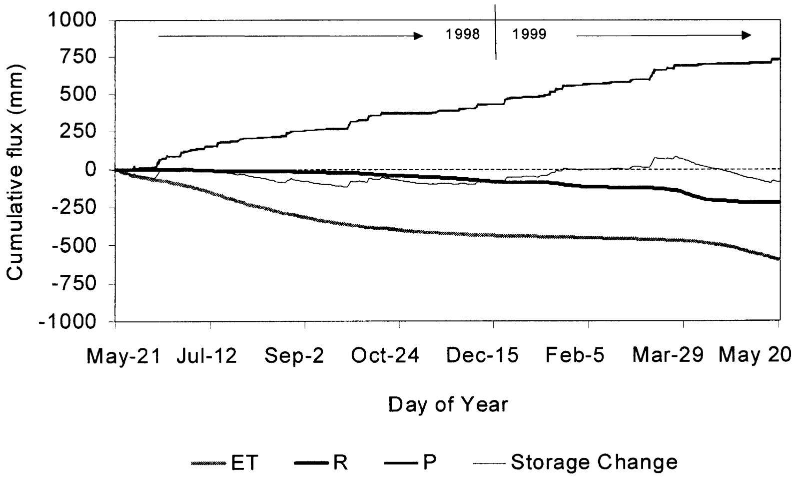

Fig. 3. Components of the water balance measured at the Mer Bleue bog (May 22, 1998 to May 21, 1999 inclusive). S is the residual of the measured components where $\Delta S = P - ET - R$.

into the peatland from the underlying aquiclude was assumed to be zero.

4.2. Hydrology

The water balance for the study period is shown in Fig. 3. Annual P, ET and R were 757, 598 and 222 mm, respectively, yielding a calculated change in storage (ΔS) of −63 mm. Close agreement (<20%) between peatland water-table position and ΔS from start to completion of the study gave confidence in P, ET and R measurements. It also supported the assumption that the peat-clay interface was a zero-flux plane for regional groundwater exchange.

Daily P, ET, R and water-table (WT) positions from June 1 to September 30, 1998 are shown in Fig. 4 to

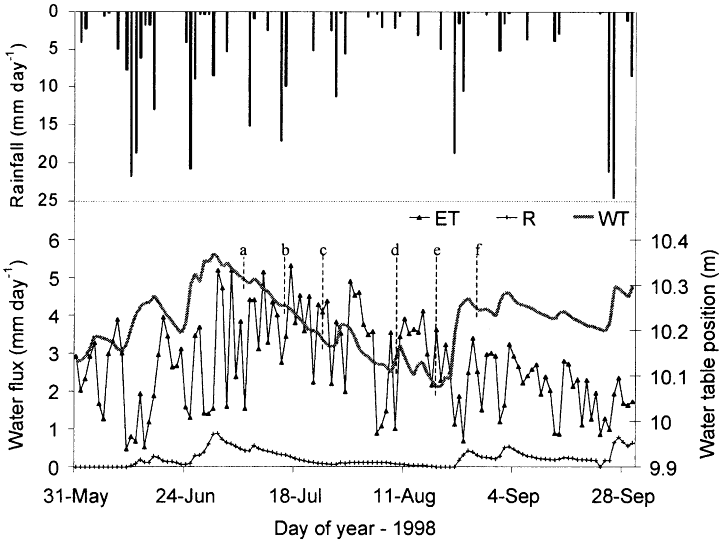

Fig. 4. Daily rainfall, evapotranspiration, runoff and water-table position measured at the Mer Bleue bog over the summer 1998. Water-table positions labeled (a–f) correspond to dates of the recharge-discharge plots shown in Fig. 5 (a–f).

characterize the hydroperiod and temporal patterns of water fluxes that influenced groundwater flow. *WT* draw-down occurred when *ET* exceeded *P* for the last two weeks of July and the majority of August. During this draw-down, measured *P* for July was similar to normal for that month (85 versus 88 mm), but measured *P* for August was half of normal (46 versus 92 mm). Water-table elevation decreased ~0.30 m during the evaporative draw-down to ~0.50 m below the peat surface, and then increased ~0.20 m on August 23, 1998 (Fig. 4).

P, *ET*, *R* and ΔS were 91, 160, 8 and −77 mm, respectively, for the fifty-day draw-down from July 4 to August 23. Six recharge–discharge plots from nest P10 are shown in Fig. 5 to illustrate the changes in vertical gradients. P10 was chosen to represent the similar pattern observed at nests P6–P14 during this time period. The peatland showed recharge with respect to hydraulic head of the water-table, prior to, and during the early part of the draw-down (Fig. 5a and b). Vertical hydraulic gradients were ~0.02–0.03 during maximum recharge. Discharge conditions developed over the period of water-table draw-down (Fig. 5c–e), achieving vertical hydraulic gradients of ~ − 0.03 to −0.02 during maximum discharge. Recharge conditions were present after August 23 (Fig. 5f).

The absolute changes in hydraulic head measured at given depths in the peatland decreased with increased depth into the peatland, but were related to changes recorded at the water-table (Fig. 6). Hydraulic head at the water-table, 0.5 and 0.75 m responded rapidly to changing boundary conditions, and the differences between maximum and minimum head in the upper peat profile were ~0.20–0.30 m (Fig. 6a — only water-table is shown). Intermediate depths (Fig. 6b — 1.5 m shown) also responded to changes at surface boundaries, but head differences were less (~0.12–0.19 m). The gross features of the hydroperiod were observed in hydraulic heads at

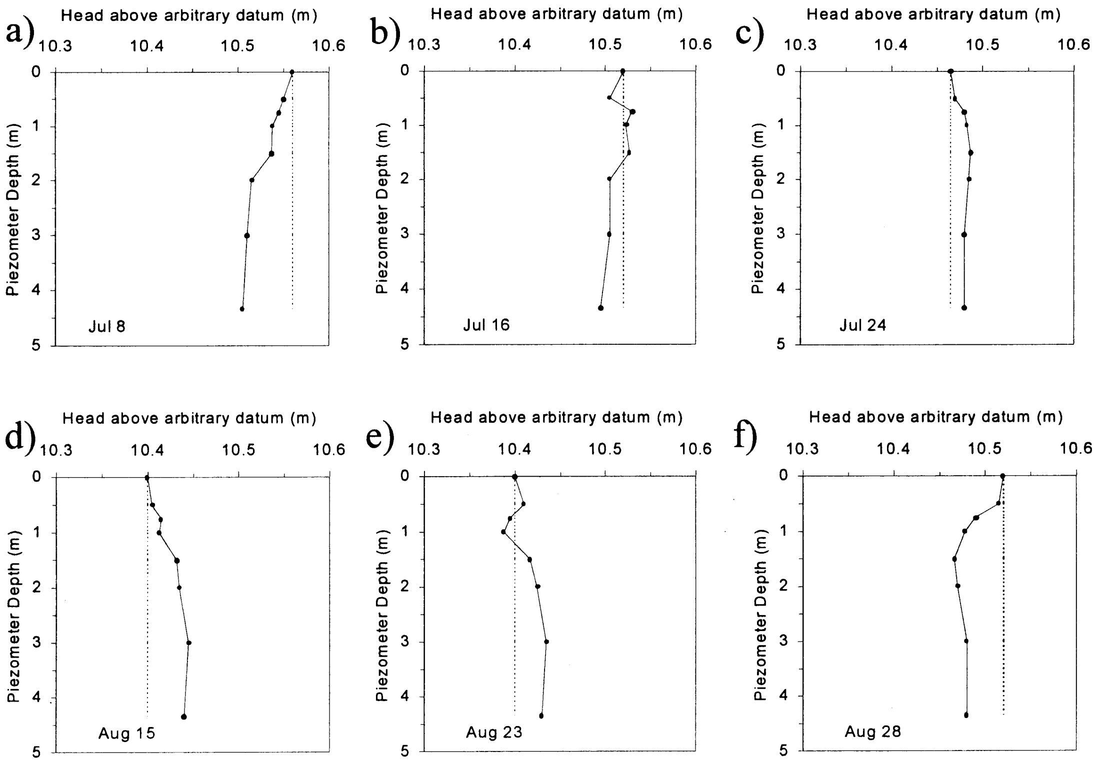

Fig. 5. Recarge–discharge plots illustrating the changes in vertical hydraulic gradients at nest P10 over the summer hydroperiod. Hydraulic potential of the water-table for each sample date is shown with a dashed line.

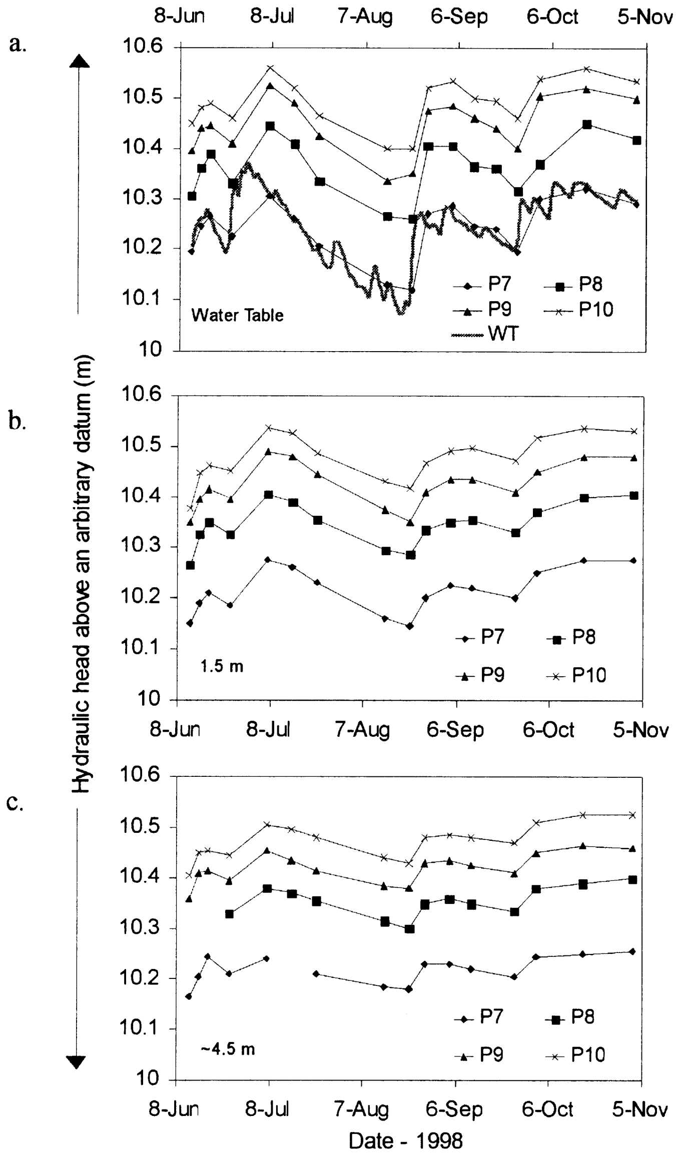

Fig. 6. (a) Manual measurements of water-table position at nests P7 through P10 for the growing season. The continuous record of water-table position measured at the tower is shown for reference. Regressions between manual observations at nests on the peatland and the continuous *WT* record ranged from 0.93 to 0.99. (b) Plots of hydraulic head from 1.5 m piezometers for the same hydroperiod and nests shown in (a). (c) Plots of hydraulic head from deepest piezometers inserted at piezometer nests for the same hydroperiod and nests shown in (a) and (b). In (c) piezometer length ranges from 3.8 to 4.5 m.

4.5 m (Fig. 6c), but the differences were even smaller (~0.09–0.12 m). Time lags between hydraulic responses at the surface boundary and deep peat were short.

Differential head response with depth (Fig. 6) caused over-pressured deeper zones relative to the surface boundary (Fig. 7), after *ET* exceeded *P* for 3 weeks. From the onset of maximum hydraulic head in surface peat (~July 1) until the observation of reversed gradients (July 21), measured *ET* and *P*

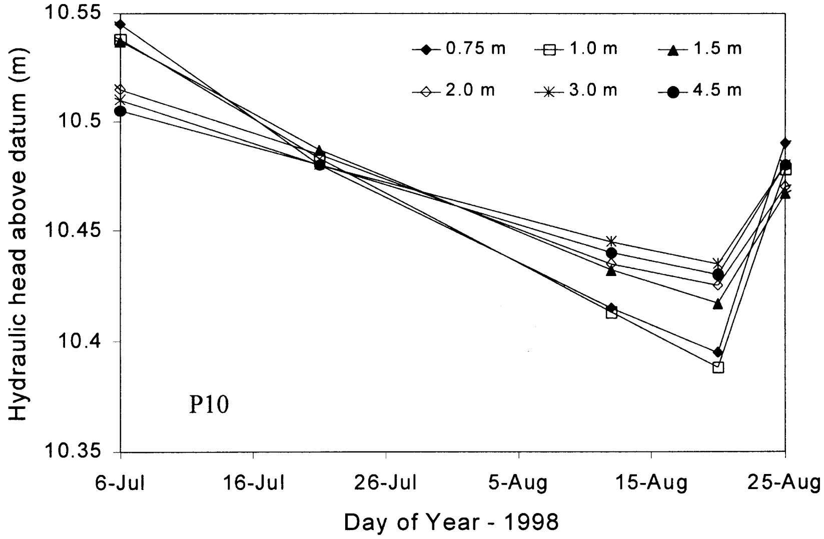

Fig. 7. Hydraulic head measurements from 0.75, 1.0, 1.5, 2.0, 3.0 and 4.5 m piezometers at nest P10. Evaporative draw-down of hydraulic head at the water table caused over-pressured deeper peats on July 19, 1998.

were 81 and 51 mm, respectively, and water-table draw-down was ~0.15 m. Continued $ET > P$, beyond the initial draw-down necessary to reverse vertical gradients, maintained groundwater discharge for 32 consecutive days.

4.3. Geochemistry

$[Na^+]$, $[Mg^{2+}]$, $[Ca^{2+}]$ and EC in the marine clay were as high as 2,200, 80, 75 mg L^{-1} and 4000 $\mu S\ cm^{-1}$, respectively (EC data shown in Table 1). In contrast, measurements of $[Na^+]$, $[Mg^{2+}]$, $[Ca^{2+}]$ and EC in precipitation were ~1–3, <1, ~2 mg L^{-1} and 10 $\mu S\ cm^{-1}$, respectively. These are typical of meteoric water in this region of North America (Dingman, 1993). In the peat, highest cation concentrations and values of EC correspond to the deepest piezometers and vice-versa (Fig. 8), indicating that profiles are the result of upward diffusion and mixing with meteoric water of low concentration. However, the Na^+, Mg^{2+} and EC profiles differed from the Ca^{2+} profile, suggesting that Ca^{2+} patterns are controlled only partially by end-member mixing.

The patterns of concentration (Na^+ and Mg^{2+}) and EC are consistent throughout the peatland as indicated by the regression of Mg^{2+} and EC versus Na^+ (Fig. 9). The strong relationship ($r^2 = 0.97$; $p = 0.001$) implies that mixing was constant over space. However, the relationship does not discriminate between the relative roles of advection and diffusion in maintaining the chemical profiles. This was tested using a chemical diffusion model.

Diffusion of Na^+ from the underlying marine clay provided an in situ tracer to assess the recharge–discharge function (see Siegel, 1988b) of the peatland with a one-dimensional diffusive transport model (Eq. (2)). We believed Na^+ was a conservative tracer because of the shape of the (Na^+) profile that results from end-member mixing (Fig. 8), and the importance of Na^+ to the charge balance of the bog. Calculations using concentrations of Na^+, average estimates of

Table 1
Summary of electrical conductivity measurements from drive-point piezometers in the underlying marine clay at the Mer Bleue bog

	Conductivity ($\mu S\ cm^{-1}$)			
Date	DP1	DP2	DP3	DP4
24-Jul	4335	3595	–	–
14-Aug	4804	3814	3644	3284
3-Sep	4470	4090	3760	3650
3-Nov	4501	4171	4171	4501
Mean	4527.5	3917.5	3858.3	3811.7
Standard Deviation	197.9	263.8	276.9	624.4
CV(%)	4.4	6.7	7.2	16.4

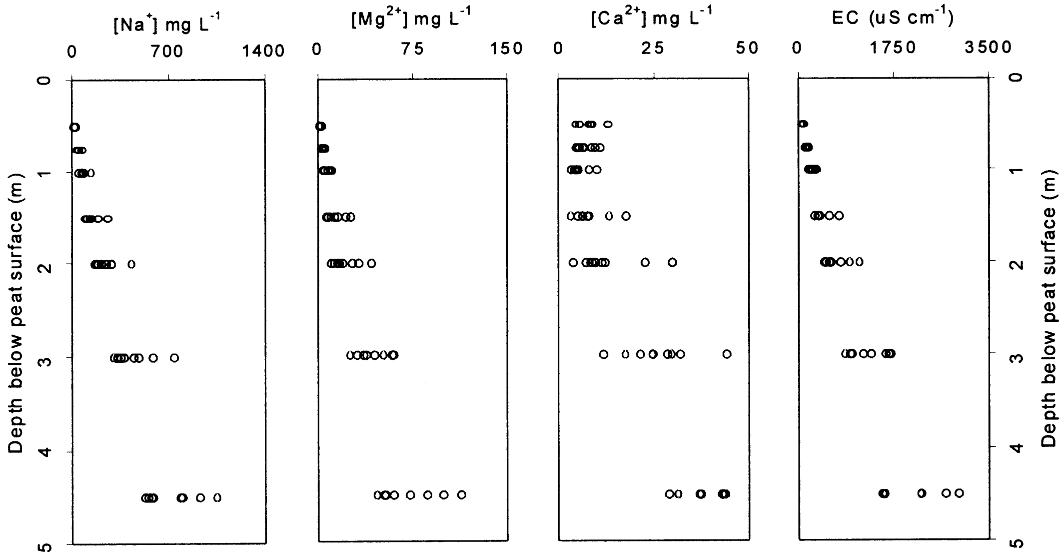

Fig. 8. Changes in Na^+, Mg^{2+}, Ca^{2+} and EC with depth based on measurements from all piezometers inserted in the peatland.

cation exchange capacity (CEC) in peat (100–200 meq/100 g; Lévesque et al., 1980), and typical peat properties (Letts et al., 2000) show that under complete exchange of Na^+ peat pore-water concentrations can change by only 2–7%. This is due to the high salinity. Assuming the tracer is conservative, if modeled concentrations are greater than observed concentrations, groundwater advection in the catotelm is essential to yield the observed profile. In contrast, if

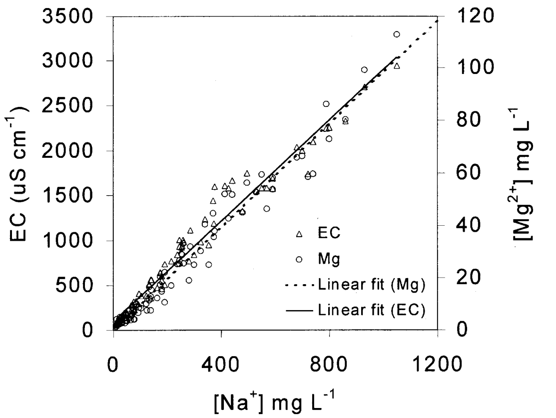

Fig. 9. Regression models of Mg^{2+} and EC versus Na^+ concentrations. The statistics from the EC-Na^+ model showed an $r^2 = 0.97$, F-stat = 2294.1 and significance (0.95) < 0.001. The statistics from the Mg^{2+}–Na^+ model showed an adjusted $r^2 = 0.97$, F-stat = 2378.3 and 0.95 significance (0.95) < 0.001.

observed concentrations are greater than modeled concentrations, upward advection > horizontal advection in the catotelm is required to yield the observed profile.

Modeling results showed that for plausible diffusion scenarios (peatland age ~8500 years; diffusion in the direction of peat growth), recharge and groundwater advection through the peat profile are required to yield the observed geochemical profile (Fig. 10). The model and assumptions inherent in the analysis are too uncertain to determine precisely the time required to develop the recharge–diffusion interaction, but clearly show the requirement of catotelm advection. The sensitivity of model output to ±50% changes in D^* are also shown in Fig. 10.

Groundwater advection attributable to the flow reversal changed the chemical profiles. EC increased during the groundwater flow reversal at all piezometer nests (Fig. 11), and decreased once recharge gradients re-established and were maintained into the autumn of 1998. The absolute magnitude of the EC change was larger at depth than near the surface of the peatland (e.g. 1250–1400 versus 50–90 $\mu S\ cm^{-1}$, respectively), but the relative change in EC was larger near the surface of the peatland than at depth (e.g. ~10–60 versus ~2–20%, respectively). The changes in EC from piezometers inserted between the surface and 2.0 m depths were also translated to estimates of

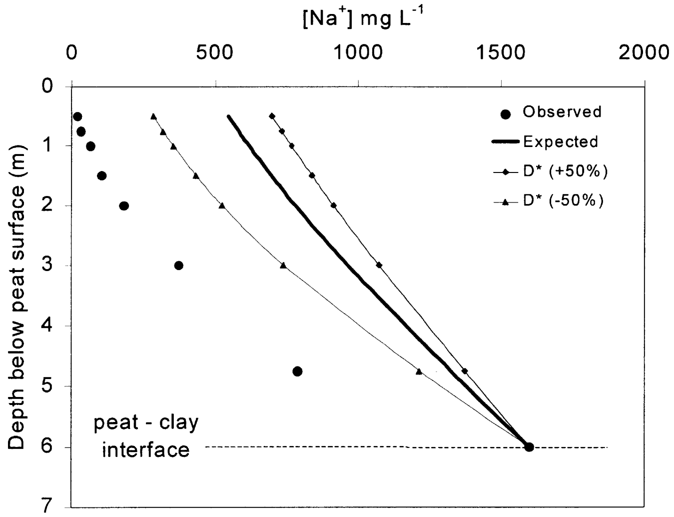

Fig. 10. Chemical assessment of groundwater flow patterns at the Mer Bleue bog using predicted chemical profiles from a one-dimensional diffusive transport model. The predicted Na^+ profile is compared with observed Na^+ concentrations at nest P12 ($D_O = 2.0 \times 10^{-9}\ m^2\ s^{-1}$; $\theta = 0.22; D^* = 1 \times 10^{-10}$). The sensitivities of the expected profile to ±50% changes to D_O (or D^*) are also shown. (Note: the ±50 sensitivity analysis also describes the scenario of a ±27% change in θ with no change in D_O or D^* (Eq. (3))).

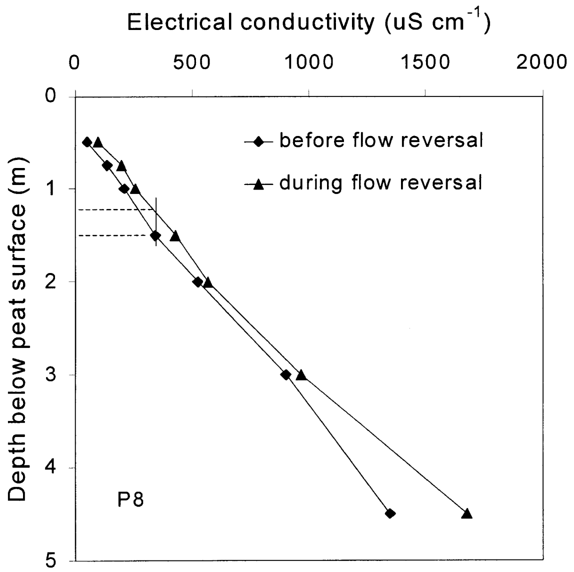

Fig. 11. Changes in EC measured at P6 before and during the groundwater flow reversal. EC measurements from June 23, 1998 (before reversal) and August 23, 1998 (during the reversal) are shown. Dashed lines between 1–2 m depth below the peat surface correspond to an estimate of vertical advection that occurred during the flow reversal.

vertical advection (see Fig. 11). Average displacement of solutes attributed to the flow reversal was 0.153 ± 0.044 m ($n = 29$).

5. Discussion

5.1. Hydraulic conductivity at the Mer Bleue bog

There was a discrepancy between depth-integrated K_v estimates back-calculated from diffusion modelling, and K_v determined from bail tests and Eq. (1) ($K_v \sim 10^{-9}\ m\ s^{-1}$). To fit a curve to observed [Na^+] in Fig. 10, the effective coefficient of diffusion (D^*) had to be reduced by a factor of five. Therefore, we estimated that downward advection must exceed upward diffusion by a similar factor to achieve the observed chemical profile. Assuming downward advection is four times the rate of D^* ($1 \times 10^{-10}\ m^2\ s^{-1}$) and average vertical hydraulic gradients are 0.02, Darcys equation yields an average, depth integrated K_v for the peat profile of $10^{-8}\ m\ s^{-1}$. We acknowledge our solution for K_v is non-unique, and the discrepancy between field and calculated K_v estimates could be explained by parameter uncertainty alone (i.e. D_o, C_o, θ, gradients). However, the estimate of K_v based on the positional change of cations during the reversal also differed from our field estimates of K_v. Given that average vertical displacement during the reversal was 0.153 ± 0.044 m per 40 days, and vertical gradients were ~ -0.02, K_v was estimated to be 10^{-7}–10^{-6} m s^{-1}.

It is possible that: (1) the vertical macroscale hydraulic conductivity (see Siegel et al., 1995; Neuzil, 1986) of the Mer Bleue, exceeds the hydraulic conductivity measured by a single piezometer, or (2) the peat was compacted during piezometer insertion and the K_m estimates were erroneous. To our knowledge, K_m is seldom measured in peatlands, and measuring K_m in the field can be problematic (Chason and Siegel, 1986). If the discrepancy can be attributed to macroscale variability or compaction, it may be beneficial to insert extra nests in groundwater networks for the purpose of 'building up' sample sizes prior to estimating K. Further, if these integrated estimates of K_v are representative of the macroscale K_v, it may be necessary to re-evaluate the rates of groundwater flow in catotelm peat, and the importance of groundwater advection to peatland biogeochemical function.

5.2. Controls on groundwater flow patterns

Groundwater flow patterns at the Mer Bleue bog were transient, and reversed during the summer. Patterns were controlled by hydraulic conductivities and gradients, but the gradients were transient because of precipitation, evapotranspiration, and a differential response with depth to head changes at the water-table (Figs. 6 and 7). The decreased draw-down of head at 1.5 m compared with 0.75 m during this time period could be explained by differences in hydraulic conductivity between the acrotelm and catotelm. For example, we found that in the upper 1.5 m of the peat profile, head responses to boundary changes decreased approximately two-fold (Fig. 6), and hydraulic conductivities decreased up to three orders of magnitude (Fig. 2). However, (1) the ranges of K_h were large (~3 orders of magnitude between acrotelm and catotelm), (2) there was little variation in hydraulic conductivity below the acrotelm, and (3) the position of the water-table was often below this boundary. Therefore, the differential head response may not be solely attributed to the physical properties of peat.

In situ CH_4 bubbles could explain the differential head response. Reynolds et al. (1992) showed that a four-fold increase in [CH_4] in peat pore-water decreased volumetric water content >30%, and decreased K from 10^{-6} to 10^{-8} m s^{-1} because peat pores were occluded by gas bubbles. Further, Brown et al. (1989) found that in situ [CH_4] increased 2–3-fold between 0.6 and 1.2 m peat depth at the Mer Bleue bog. Thus, spatial and temporal variations in pore occlusion could yield decreases in K through the peat profile and explain the differential head response.

If pore occlusion controls the differential head response, it is also probable that pore water pressure affects K values as well. Baird and Gaffney (1995) reported that a 5 kPa increase in pressure (~0.5 m water-table increase) increased hydraulic conductivity by 33%, because increased pressure decreased the size of CH_4 bubbles and the degree of pore occlusion. The draw-down that occurred during the flow reversal at the Mer Bleue bog (0.3 m) could have decreased K when pressure was reduced. These mechanisms can explain the differential head response that occurs through the profile, but more field studies are necessary to determine the spatial and temporal importance of CH_4 occlusion and pore-water pressure to peatland hydrological function.

5.3. Characteristics of flow patterns and geochemistry

The magnitude of vertical gradients at the Mer Bleue (0.03 to −0.03) is similar to gradients reported in other studies. For example, the gradients in the Glacial Lake Agassiz peatlands of northern Minnesota ranged from 0.03 to −0.04 (Romanowicz et al., 1993; Siegel et al., 1995). These vertical gradients are also similar to gradients reported for three small peat-complexes by Devito et al. (1997). However, Devito et al. reported vertical gradients as large −0.2 in one of the peat complexes, which was explained by hillslope connectivity.

Connectivity to regional groundwater allowed for flow reversals to be prolonged for several years in Minnesota (Siegel et al., 1995). In contrast, we found that reversals were a self-limiting process. At the Mer Bleue, drought is required to draw-down the water-table to invoke reversed gradients. However, prolonged drought eventually decreases the atmosphere–peatland coupling that is required to draw-down surface heads and maintain discharge gradients. We found that hydraulic potentials in deepest piezometers decreased linearly at $\sim$0.0023 m day^{-1} during water-table declines (Fig. 7), and hydraulic potentials in deepest piezometers were within 0.03 m of surface hydraulic potentials at the end of the reversal (Fig. 5). If drought were to persist (e.g. no storm August 23), reversed flow could be sustained no more than 15 additional days because hydrostatic conditions would have developed. We expect that flow reversals in similar hydrogeologic settings (Devito et al., 1997) are self-limiting, but may have longer duration if the initial vertical gradients are larger.

We found that a short flow reversal increased the concentrations of cations near the water-table by 10–60% (e.g. EC increased from 180 to 240 μS cm^{-1} at 1.0 m depth below the peat surface). In contrast, Siegel et al. (1995) found that prolonged drought for several years increased pH from <4.5 to 6.0, and specific conductance increased from <60 to 200 μS cm^{-1} ($\sim$400%). These data suggest that flow reversals affect pore-water chemistry in peatlands, but the significance of the pore water changes to biogeochemical processes is largely unknown. There is some convincing evidence (Charman et al., 1992; Chanton et al., 1995) that younger dissolved organic carbon (DOC) is transported to depth in peatlands that may experience flow reversals, and that this DOC plays a role in respiration deeper in the profile (Chanton et al., 1995).

6. Conclusions

Groundwater flow patterns were transient during the growing season, and $ET > P$ for 21 days was sufficient to reverse vertical gradients in July 1998 ($ET = 80$ mm, $P = 51$ mm). Flow patterns were controlled by P, ET, and a differential response with depth to head changes at the water-table. The time required to reverse vertical gradients was short ($\sim$21 days), and the magnitude of the water-table draw-down was small ($\sim$0.15 m). This would suggest that reversals at the Mer Bleue bog and peatlands with similar hydrogeologic settings are common.

End-member mixing controlled geochemical profiles, and the results from a diffusion model showed that the peatland is a long-term recharge system. However, changes in the direction of groundwater advection increased concentrations of pore-water chemistry as much as 60% during the flow reversal. This would suggest that flow reversals could be important for the redistribution of nutrients and limiting substrates in peatlands isolated from regional groundwater.

Acknowledgements

The authors are grateful for field assistance from Sarah Giffen and Kira Rich. We thank the National Capital Commission for access to the Mer Bleue, and K.J. Devito, J.-M. Ferone and T.R. Moore for comments on earlier drafts of this paper. We also thank Don Siegel and a second anonymous reviewer for their constructive criticisms. This research was supported by NSERC Strategic Grant (#0193069), a NSERC Research Grant to NTR and FCAR Equipe and Centres grants.

References

Baird, A.J., Gaffney, S.W., 1995. A partial explanation of the dependency of hydraulic conductivity positive pore water pressure in peat soils. Earth Surf. Proc. Land. 20, 561–566.

Baldocchi, D.D., Hicks, B.B., Meyers, T.P., 1988. Measuring biosphereatmosphere exchanges of biologically related gases with micrometeorological methods. Ecology 69, 13311340.

Belanger, J.R., Harrison, J.E., 1977. Bedrock geology, drift thickness trend and bedrock topography, Ottawa-Hull, Ontario and Quebec. GSC Paper, 77–11, 18pp.

Branfireun, B.A., Heyes, A., Roulet, N.T., 1996. The hydrology and methylmercury dynamics of a Precambrian Shield headwater peatland. Water Resour. Res. 32, 1785–1794.

Brown, A., Mathur, S.P., Kushner, D.J., 1989. An ombrotrophic mire as a methane reservoir. Global Biogeochem. Cyc. 3, 205–213.

Chanton, J.P., Bauer, J.E., Glaser, P.A., Siegel, D.I., Kelley, C.A., Tyler, S.C., Romanowicz, E.H., Lazrus, A., 1995. Radiocarbon evidence for the substrate supporting methane formation within northern Minnesota peatlands. Geochim. Cosmochim. Acta 59, 3663 –3668.

Charman, D.I., Aravena, R., Warner, B.G., 1992. Isotope geochemistry of gas and water samples from deep peats in Boreal Canada. Suo 43, 199–201.

Chason, D.B., Siegel, D.I., 1986. Hydraulic conductivity and related physical properties of peat, Lost River peatland, northern Minnesota. Soil Sci. 142, 91 –99.

Crank, J., 1975. The mathematics of diffusion. 2nd Ed. Clarendon Press, Oxford, p. 414.

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

Dingman, S.L., 1993. Physical Hydrology. Prentice Hall, NJ, p. 575.

Doss, P.K., 1993. The nature of a dynamic water table in a system of non-tidal, freshwater coastal wetlands. J. Hydrol. 141, 107–126.

Environment Canada, 1998. Canadian Climate Normals: Ottawa International Airport, 1938–1990.

Freeze, R.A., Cherry, J.A., 1979. Groundwater. Prentice-Hall, NJ, p. 604.

Glaser, P.H., Wheeler, G.A., Gorham, E., Wright Jr, H.E., 1981. The patterned mires of the Red Lake peatland, northern Minnesota: vegetation, water chemistry and landforms. J. Ecol. 69, 575–599.

Gorham, E., 1991. Northern Peatlands: role in the carbon cycle and probable responses to climatic warming. Ecol. Appl. 1, 182–195.

Hill, B.M., Siegel, D.I., 1991. Groundwater flow and the metal content of peat. J. Hydrol. 123, 211 –224.

Hobson, G.D., 1969. Bedrock features of the Mer Bleue area by seismic methods. Can. Field Nat. 84, 35 –38.

Hvorslev, M.J., 1951. Time lag and soil permeability in groundwater observations. U.S. Army, Corps Engng Waterways Exp. Stn, Bull. 36, Vicksburg, Mississippi, 50pp.

Ingram, H.A.P., 1981. Hydrology. In: Gore, A.J.P. (Ed.). Ecosystems of the World. Mires: Swamp, Bog, Fen and More. Elsevier, Amsterdam, pp. 67–158 General Studies, Vol. 4A.

Letts, M.G., Roulet, N.T., Comer, N.T., Skarupa, M.R., Verseghy, D.L., 2000. Parameterization of peatland hydraulic properties for the Canadian Land Surface Scheme. Atmos. Ocean 38, 141 –160.

Lévesque, M., Morita, H., Schnitzer, M., Mathur, S.P., 1980. The physical, chemical, and morphological features of some Quebec and Ontario peats. Research Branch, Agriculture Canada, 70pp.

Mott, R.J., Camfield, M., 1969. Palynological studies in the Ottawa area. GSC Paper, 69–38, 16pp.

Neuzil, C., 1986. Groundwater flow in low-permeability environments. Water Resour. Res. 22, 1163–1197.

Reynolds, W.D., Brown, D.A., Mathur, S.P., Overend, R.P., 1992. Effect of in situ gas accumulation on the hydraulic conductivity of peat. Soil Sci. 153, 397 –408.

Romanowicz, E.A., Siegel, D.I., Glaser, P.H., 1993. Hydraulic reversals and episodic emissions during drought cycles in mires. Geology 21, 231–234.

Siegel, D.I., 1981. Hydrogeologic setting of the Glacial Lake Agassiz Wetlands, northern Minnesota. USGS Water Resour. Investigation Rep. 81–24, 30pp.

Siegel, D.I., 1988a. The recharge-discharge function of wetlands near Juneau, Alaska: Part I. Hydrogeological investigations. Ground Water 26, 427 –434.

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

Siegel, D.I., Reeve, A.S, Glaser, P.H., Romanowicz, E.A., 1995. Climate-driven flushing of pore water in peatlands. Nature 374, 531 –533.

Verma, S.B., 1990. Micrometeorological methods for measuring surface fluxes of mass and energy. Rem. Sens. Rev. 5, 99115.

Waddington, J.M., Roulet, N.T., 1997. Groundwater flow and dissolved organic carbon movement in a boreal peatland. J. Hydrol. 191, 122 –138.

Winter, T.C., 1999. Relation of streams, lakes, and wetlands to groundwater flow systems. Hydrogeol. J. 7, 28–45.

Winter, T.C., Carr, M.R., 1980. Hydrologic setting of wetlands in the Cottonwood Lake area, Stutsman County, North Dakota. USG Water Resour. Investigation Rep. 80–99, 42pp.