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Modelling two-dimensional steady-state groundwater flow and flow sensitivity to boundary conditions in blanket peat complexes

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Abstract:

This study used a two-dimensional steady-state finite-element groundwater flow model to simulate groundwater flow in two Newfoundland blanket peat complexes and to examine flow system sensitivity to changes in water table recharge and aquifer properties. The modelling results were examined within the context of peat-forming processes in the two complexes. Modelled flow compared favourably with observed flow. The sensitivity analyses suggested that more highly decomposed bog peat along bog margins probably has/had a positive impact on net peat accumulation within bog interiors. Peat with lower hydraulic conductivity along bog margins effectively impedes lateral drainage, localizes water table drawdown to extreme bog margins, and elevates water tables along bog interiors. Peat formation and elevated water tables in adjacent poor fens/laggs currently rely on placic and ortstein horizons impeding vertical drainage and water flow inputs from adjacent bogs. Modest reductions in atmospheric recharge were found to govern bog-flow-system geometries in a way that would adversely affect paludification processes in adjacent fens/laggs. Copyright © 2004 Crown in the right of Canada. Published by John Wiley & Sons, Ltd.

KEY WORDS peatlands; groundwater flow; steady-state finite-element model; paludification; blanket bog; Newfoundland

INTRODUCTION

Groundwater hydrology is important to the maintenance and development of wetlands (Ivanov, 1978; Ingram, 1983; Mitsch and Gosselink, 1986). Numerical groundwater-flow modelling is often required to study flow dynamics in systems with complex boundary conditions and/or spatially variable aquifer properties (Freeze and Cherry, 1979). Aquifer properties such as hydraulic conductivity can be highly spatially variable and boundary conditions in wetlands can be difficult to specify (Romanov, 1968; Bay, 1969; Ingram, 1983; Price, 1992; Hoag and Price, 1995; Hunt *et al*., 1996). Numerical modelling of wetland flow systems, coupled with field observations, can help elucidate groundwater-flow sensitivity to changes in wetland recharge, changes in hydraulic properties and changes in morphological boundary conditions (Siegel, 1983; Beckwith *et al*., 2003b). Relatively recent interests in wetland conservation and the potential effects of climate change on wetland function require a better understanding of the processes governing groundwater flow and of the hydrological conditions required for wetland development and maintenance (Burt *et al*., 1990).

Peatlands are wetlands with soils composed predominantly of organic remains of decomposing plant material (peat). Peat accumulation, both laterally and vertically requires saturated or nearly-saturated surface conditions for prolonged periods of time (Ivanov, 1978). Blanket peatlands are extensive peat deposits that cover flat and gently undulating terrain (Crum, 1988) in cool oceanic regions where precipitation greatly exceeds

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evapotranspiration (Tansley, 1949; Crum, 1988). In Canada, true blanket bogs are located almost exclusively on the extreme coasts of southern Newfoundland (Davis, 1984). Blanket bogs have been considered climatic formations effectively independent of topography (Wells, 1981). The implication of such a consideration is that high net precipitation compensates for greater bog drainage resulting from greater slope gradients. However, recent investigations on the hydrology of Newfoundland blanket bogs suggest that site factors such as soil processes, substrate permeability and topography may govern the spatial development and maintenance of blanket peats to a much greater extent than previously considered (Lapen, 1998; Graniero and Price, 1999a, b; Lapen and Wang, 1999). The potential role that pedogenesis and topography play in bog development are not new concepts (e.g. Ugolini and Mann, 1979; Smith and Taylor, 1989); however, the sensitivity of the hydrological conditions required for bog maintenance and paludification (lateral expansion of wet, peat-forming environments over adjacent areas as a result of waterlogging) to spatially and temporally variable groundwater processes, has received somewhat less attention for mature blanket bogs and adjacent water shedding areas; especially for Newfoundland systems. In Newfoundland, contributions of acidic groundwater from currently established blanket bog can be considered critically important to the establishment of hydrophilic bog-forming vegetation (i.e. *Sphagnum* spp. and Cyperaceae), as well as the propagation and development of drainage impeding sesquioxide pans within adjacent fens/laggs and downslope receiving sites (Lapen, 1998; Lapen and Wang, 1999; Lapen *et al*., 2000).

This study focuses on steady-state groundwater flow in two blanket peat complexes on the southeastern tip of the Avalon Peninsula, Newfoundland. The complexes include an upland blanket bog–poor fen and a lowland blanket bog flow system. The general purpose of this investigation was to identify some of the critically limiting environmental mechanisms governing requisite site hydrological conditions for peat accrual in mature blanket bogs and adjacent fens/laggs where wet surface conditions are inducing establishment of bog-forming vegetation and peat formation. The specific objectives of this study are to model groundwater flow in the complexes in order to: (i) examine, within a parsimonious context, how changes in recharge and the spatial distribution of aquifer properties affect groundwater flow within the systems, (ii) elucidate the potential role of pedogenic processes on the distribution of aquifer properties, and (iii) assess relationships between groundwater dynamics and blanket peat formation.

STUDY SITE

The study site is located at Cape Race, Newfoundland (Figure 1). The average annual total precipitation at Cape Race is approximately 1379 mm (Environment Canada, 1982). Occult precipitation (OP) is common in the region and visibility <1 km owing to fog has been observed to occur, on average, 155 days a year at Cape Race (Banfield, 1983).

Blanket bogs and heathlands are the major community types in the region (Meades, 1983; Wells and Pollett, 1983). Blanket bogs are generally <2 m deep. Poor fens and/or laggs with relatively shallow peat deposits ($<0{\cdot}3$ m deep), often receive acidic discharge from upslope bogs (Pollett, 1968; Lapen *et al*., 1996; Lapen and Wang, 1999).

The bogs are carpeted primarily by *Sphagnum fuscum* and the peat is composed of poorly to highly-decomposed *Sphagnum* spp., sedges (Cyperaceae), sweet gale (*Myrica gale* L.) and heath-type (Ericales) vegetation. The upland and lowland bogs are underlain by Placic Humic and Placic Humo-Ferric Podzols. Placic horizons, commonly called iron pans (Soil Survey Staff, 1975), occur approximately 0·3 m beneath the peat–mineral soil contact and effectively perch the groundwater (Lapen *et al*., 1996; Lapen and Wang, 1999). Water tables in the bogs generally occur within 0·05 to 0·20 m depth during the summer growing season (Northlands Associates Ltd, 1989; Price, 1992; Lapen *et al*., 2000).

The vegetation in the upland poor fen (part of upland bog complex) consists largely of *Sphagnum* spp., with higher proportions of sedges, heath, sweet gale and more minerotrophic vascular vegetation than the bogs. Peat and muck, ranging roughly between 0·10 to 0·30 m thick, overly sandy Placic Humo-Ferric Podzols.

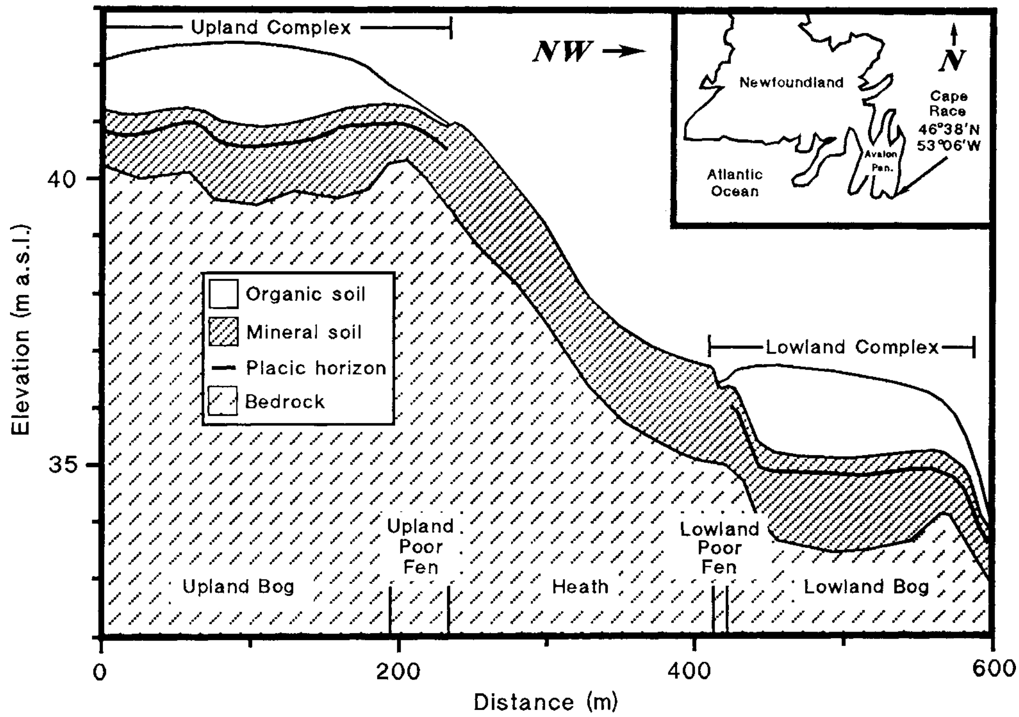

Figure 1. Study site and location of blanket peat complexes

The lowland poor fen consists of approximately 0·05 to 0·20 m of peat and muck overlying poorly drained Ortstein Ferro-Humic and Ortstein Humic Podzols. Vegetation in the lowland poor fen is primarily sedges, sweet gale and *Sphagnum* spp.

METHODS AND MATERIALS

Field and laboratory measurements

The hydrological data reported in this study were acquired in the field from 24 June to 12 August 1993. Water table recharge (D) via the atmosphere to the upland and lowland bog complexes during this period was calculated according to

$$D = R + OP - ET - \Delta SWS \quad (1)$$

where R is rain precipitation, OP is occult precipitation, ET is evapotranspiration and ΔSWS is change in water storage in the unsaturated zone ((+) values = increase in storage, (−) values = decrease in storage).

Rain precipitation was measured by a calibrated tipping-bucket rain gauge set at ground level and one manual rainwater collector. Wind corrections were not applied to the rain gauge measurements. Wet deposition by occult precipitation (OP) was obtained, using wind speed and surface roughness data measured at the site (Lapen, 1998), in a sedimentation–turbulent transfer model (Unsworth and Crossley, 1987). Bog ET was determined from (net radiation (Q^*) − soil heat flux (Q_g)) versus latent heat flux (Q_e) regression relationships developed from Bowen ratio/energy balance (BREB) measurements over bog during the summer of the previous year. Water table depth versus BREB Q_e associations confirmed suitability for predicting Q_e from $Q^* - Q_g$ during water table conditions examined over the period of this study. Evapotranspiration from the upland poor fen was determined using peat/*Sphagnum* spp. filled lysimeters (0·09 m^2 by 0·2 m deep) with impermeable bottoms and outflow reservoirs (in case of overflow), according to

$$ET = R + OP - \Delta LWS \tag{2}$$

where ΔLWS is change in lysimeter water storage ((+) values = increase in storage, (−) values = decrease in storage). See Lapen (1998) and Lapen *et al.* (2000) for more detailed discussions of lysimeter and micrometeorological measurements.

Changes in water storage in the peats/*Sphagnum* spp. above the water table were determined by time domain reflectometry (TDR) using a 1502B Tektronix cable tester and twin-balanced wave guides. Wave guides, of various lengths (0·30, 0·15 and 0·10 m lengths), were inserted vertically and diagonally into the peat/*Sphagnum* spp. above the water table to determine ΔSWS. See Lapen *et al.* (2000) for measurement details.

Water table and hydraulic heads were determined from screened wells (0·015, 0·025 and 0·13 m i.d. PVC) and piezometer nests. Piezometers installed in the peat were constructed from 0·015 m i.d. PVC pipe with 0·06 m screened intakes. The piezometers were manually inserted into the peat and it was assumed the peat provided a natural seal (Price, 1992). Piezometers installed in the mineral soils beneath the peat layers were made of 0·015 m i.d. PVC pipe and had 0·05 or 0·10 m screened intakes and bentonite seals. Owing to potential for excessive site damage by heavy machinery, only two piezometer nests were installed in the deeper mineral substrate (Lapen *et al.*, 1996). All piezometers and wells were open at the bottom. Saturated hydraulic conductivity (K_{sat}) was obtained by pumping tests on piezometers and wells at various water table elevations (Freeze and Cherry, 1979). Measurements of K_{sat} were also performed at 0·06 m depth increments over the saturated profile at 11 additional locales in the bogs to help account for spatial variability. Precise differences between vertical (K_y) and horizontal (K_x) K_{sat} could not be distinguished precisely from the pumping tests used. In some cases it was difficult to measure the K_{sat} of the highly permeable living *Sphagnum* spp. carpet and surface peats using the pumping tests; consequently, K_{sat} (K_x via preferential orientation of the sample in the permeameter) values of these materials were determined by constant-head permeameters (0·076 m diameter by 0·05 m length cores) (Freeze and Cheery, 1979). Soil cores (0·076 m diameter by 0·076 m length) were taken laterally in mineral soils from soil pits immediately adjacent to the bogs in the poor fens to acquire estimates of K_x of soils between the basal peat deposits and the placic horizons ($N = 6$).

Daily groundwater storage changes (ΔGWS) in fen and bog were determined according to methods described in Price (1992). We used 0·3 m thick by 0·09 m^2 soil lysimeters filled with 'undisturbed' bog or fen materials to approximate specific yields of soils in the dynamic water table fluctuation zone (<0·3 m depth) ($N = 2$ lysimeters per peatland unit). Water table depth changes were determined from wells. Additional information on water release characteristics of the wetland soils was determined from drainable porosity of soil cores. Drainable porosity was calculated as the volume of gravity drained water over a 24 h period divided by the volume of the saturated sample ($DP_{(24\ h)}$)(Lapen *et al.*, 2000). Measurements of $DP_{(24\ h)}$ were made on 0·05 m thick by 0·004 m^2 and 0·05 m thick by 0·09 m^2 samples from the surface to maximum depths of 0·25 m for bog and 0·15 m for fen. These samples were taken from 20 spatially distributed locations in each peatland unit.

The degree of peat decomposition was assessed using the von Post scale of decomposition (Parent and Caron, 1993). Redox potentials were measured using a Jensen Model P5E oxygen/ORP meter with a Ag/AgCl reference electrode and Pt microelectrodes. The electrodes were inserted into the peat/soil and allowed to equilibrate up to a maximum of 1 h. The geometry of subsurface features within the flow systems was delineated by ground-penetrating radar (GPR) and traditional point observation techniques (Lapen *et al.*, 1996).

Steady-state flow model

A steady-state two-dimensional finite element flow model (FLOTRANS) (Frind and Matanga, 1985; Guiguer *et al.*, 1994) was used in this study. The finite element technique was selected because it reproduces irregular geometry and small hydraulic gradients more accurately than the finite difference approach. The FLOTRANS model uses the Galerkin finite-element method to solve the governing equations for steady-state two-dimensional flow in heterogeneous, saturated and anisotropic porous medium (Guiguer *et al*., 1994). To summarize Guiguer *et al*. (1994), the model-governing equations, in dual formulation, are

$$\frac{\partial}{\partial x}\left[K_{xx}\frac{\partial \phi}{\partial x}\right]+\frac{\partial}{\partial y}\left[K_{yy}\frac{\partial \phi}{\partial y}\right]=0 \tag{3}$$

and

$$\frac{\partial}{\partial x}\left[\frac{1}{K_{yy}}\frac{\partial \psi}{\partial x}\right]+\frac{\partial}{\partial y}\left[\frac{1}{K_{xx}}\frac{\partial \psi}{\partial y}\right]=0 \tag{4}$$

where x and y are the horizontal and vertical coordinate directions, respectively (L), ϕ is the hydraulic head (L), ψ the stream function (L^2 T^{-1}) and K_{xx} and K_{yy} are principal components of the hydraulic conductivity tensor (L T^{-1}).

The groundwater flux, q (L T^{-1}) in Darcy's form can be expressed as

$$q_x=-K_{xx}\frac{\partial \phi}{\partial x} \quad \text{and} \quad q_y=-K_{yy}\frac{\partial \phi}{\partial y} \tag{5}$$

and in terms of streamfunctions, q can be given by

$$q_x=\frac{\partial \psi}{\partial y} \quad \text{and} \quad q_y=-\frac{\partial \psi}{\partial x} \tag{6}$$

Given that discharge in a streamtube (defined by two boundary flownet streamlines) ΔQ (L^2 T^{-1}), equals the streamfunction increment $\Delta\psi$ (L^2 T^{-1}), q can be calculated according to

$$q=\frac{\Delta Q}{\Delta p}=\frac{\Delta \psi}{\Delta p} \tag{7}$$

where Δp is streamtube width (L).

Boundary conditions for Equation (3) are required over the four boundaries of the flow system. The model derives equivalent streamfunction boundary conditions internally. The boundary conditions for Equation (3) can be fixed-head or constant-flux. The boundary conditions can vary between these two types over all boundary nodes. However, the system can be uniquely solved only if at least one boundary node is fixed-head. The base of the flow system is considered impermeable.

A constant-flux boundary is specified as a flow per unit length of boundary per unit width of flow system. This flow rate is multiplied by the contributing length of the element. The constant flux boundary is commonly applied at the surface of the flow system where it represents water table recharge (Guiguer *et al*., 1994).

In the solution to Equations (3) and (4), an iterative approach is used where the flow domain is allowed to deform vertically to conform to the equilibrium water table position. The hydraulic head solution algorithm will iterate until a user-defined convergence criterion is reached. The condition to be satisfied at the water table is

$$\phi_{\mathrm{w}}=y_{\mathrm{w}} \tag{8}$$

where ϕ_{w} is the head and y_{w} the water table

At each iteration, the computed heads at the water table are compared with the water table elevation, and the grid vertically deforms to account for any difference. After the hydraulic head solution has converged, the linear streamfunction solution will proceed. Mass balance errors are provided for the hydraulic potential and streamfunction solutions to verify accuracy.

STEADY STATE MODELLING APPROACH AND IMPLEMENTATION

General approach

Modelling steady-state flow involved, (i) developing a conceptual model for each flow system, (ii) defining the finite-element grid domains, (iii) applying constant-flux boundaries, or average study period recharge (D) to flow system surface nodes (Table I), (iv) specifying constant-head values (average observed) at boundary nodes, (v) assigning aquifer properties based on field measurements (assuming heterogeneous–isotropic K_{sat} flow systems) (e.g. Figure 2) and (vi) calibrating K_{sat} layers until modelled and observed heads/water table reasonably matched.

Although anisotropy is a real component in bog flow systems (Beckwith *et al*., 2003a,b), it was not incorporated in modelling endeavours in this study. Logistical efforts were focused more strongly on defining hydraulic property heterogeneity, as heterogeneity was deemed, and has been demonstrated (Beckwith *et al*., 2003b), to be more influential than anisotropy for capturing the complex steady-state flow conditions in bogs.

Implementation (lowland bog complex)

Detailed representations of the observed 'steady-state' lowland bog two-dimensional flow system is given in Figure 3. No surface flow outlets or drainage channels occurred along either of the flow system transects. From approximately the 583 m mark (Figure 3) to the pond (*c.* 595 m mark), the lowland bog was characterized by hummocky, highly decomposed and strongly eroded peat

The model grid of the lowland bog flow system was approximately 3 m in depth, 159 m in length, and unit width was assumed. The lowland poor fen (Figure 1) was not included in the modelling exercise because of its limited spatial extent and that the ortstein horizons in the fen could not be considered effectively impermeable (Lapen and Wang, 1999), thus complicating the boundary conditions. The grid had 53 grid columns and 10 grid rows. Average recharge ($D = 3{\cdot}4$ mm day^{-1}) was applied to surface nodes as a constant-flux boundary.

Table I. Summary statistics of atmospheric fluxes, storage changes, and water table recharge during the 50 day study period

Variable	System	Mean (mm day^{-1})	Standard deviation (mm day^{-1})	Total (mm)	Minimum (mm day^{-1})	Maximum (mm day^{-1})
R	Upland bog	4·7	11·1	234·6	0·0	64·3
	Upland Poor Fen	4·7	11·1	234·6	0·0	64·3
	Lowland Bog	4·7	11·1	234·6	0·0	64·3
OP	Upland Bog	0·6	0·5	29·2	0·00	1·8
	Upland Poor Fen	0·7	0·6	35·7	0·00	2·2
	Lowland Bog	0·6	0·5	29·2	0·00	1·8
ET	Upland Bog	2·0	0·9	101·0	0·6	3·8
	Upland Poor Fen	2·6	1·2	130·2	0·6	4·8
	Lowland Bog	2·0	0·9	101·0	0·6	3·8
Δ GWS[a]	Upland Bog	−0·2	10·0	−10·1	−28·9	52·6
	Upland Poor Fen	−0·1	6·8	−4·0	−20·7	34·7
	Lowland Bog	−0·2	9·4	−9·6	−28·2	48·2
Δ SWS[b]	Upland Bog	−0·3	2·1	−13·5	−2·8	7·0
	Upland Poor Fen	−0·4	2·6	−21·6	−3·2	7·2
	Lowland Bog	−0·3	2·1	−8·5	−3·5	6·2
D	Upland Bog	3·5	9·9	176·2	−2·7	57·9
	Upland Poor Fen	3·2	9·7	161·7	−2·4	56·7
	Lowland Bog	3·4	9·8	171·5	−2·5	57·4

[a] derived from flow length weighted averages.
[b] values representative of conditions dominating centres of respective flow systems.

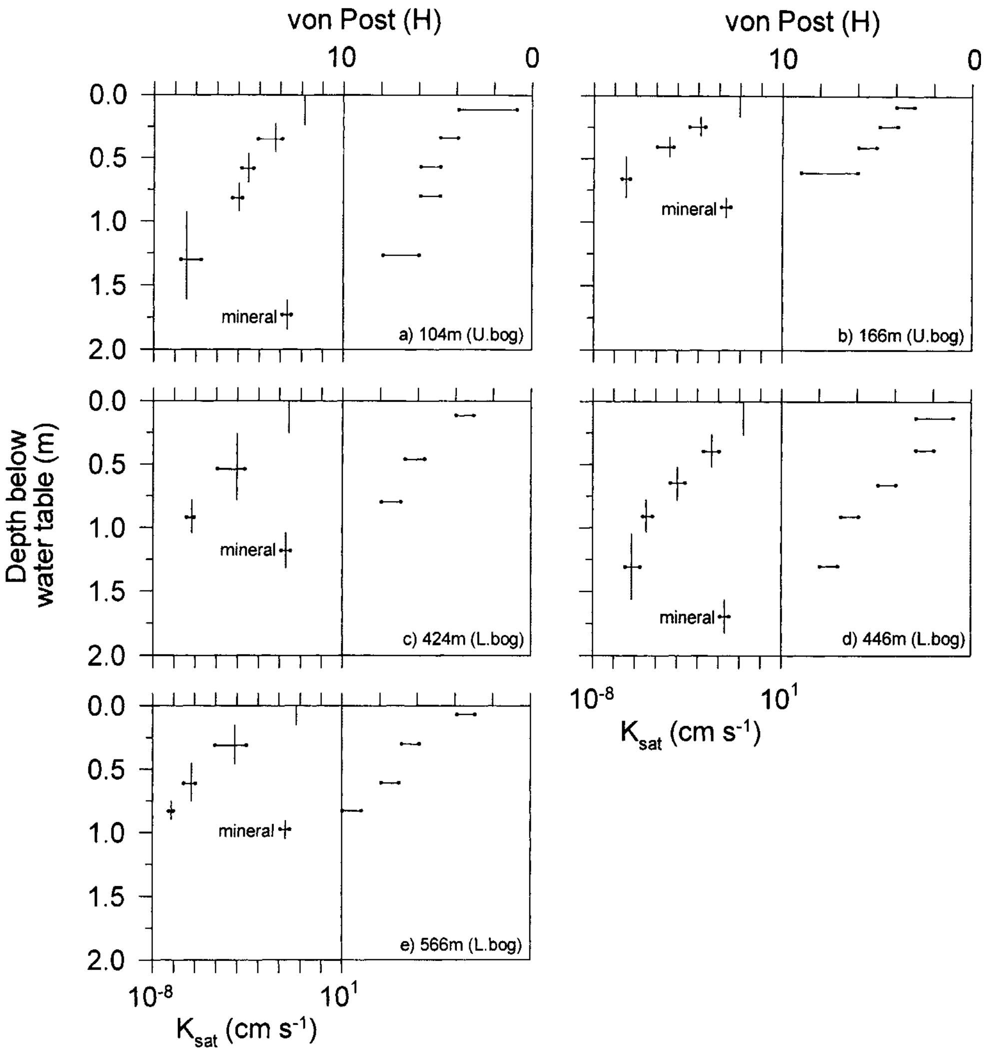

Figure 2. Observed and calibrated K_{sat} and von Post (H) layer profiles representative of deformed lateral bog flow system segments. Layer depths are based on grid deformed depths from water table to surface of impermeable layer. Each K_{sat} layer (layer depth represented by length of vertical bar) below the top Ksat layer is represented by the geometric mean of the equivalent Kx (Freeze and Cherry, 1979) of the 0·06 m depth increment values for the given layer at several locales along the complex: i.e., there were at least four equivalent Kx sites for each layer for each respective lateral flow system segment (segment definitions below). The range of K_{sat} and von Post values for each layer are also given. The K_{sat} of the top K_{sat} layer is given as the model *calibrated* K_{sat} value. The depth profiles are representative for: (a) 104 m mark (representative of 104 to 165 m mark in upland bog); (b) 166 m mark (representative of 166–193 m mark in upland bog); (c) 424 m (representative of 424 to 445 m mark in lowland bog; (d) 446 m mark (representative of 446 to 565 m mark in lowland bog); (e) 566 m mark (representative of 566 to 583 m mark in lowland bog)

A groundwater divide associated with the domed water table of the bog was observed at the 456 m mark; as a result, constant-heads (average water table elevations during the study period) were designated at nodes representing the right-hand and left-hand boundaries of the flow domain. As the right-hand boundary of the model did not represent the 'actual' right-hand outflow boundary of the lowland bog flow system (the stream is the actual end of the system), constant-head values were specified at all right-hand outflow nodes. Modelling was not performed in the zone of dynamic peat erosion between the 583 m mark and the stream; as collection of field data and model parameterization in this area would have been problematic.

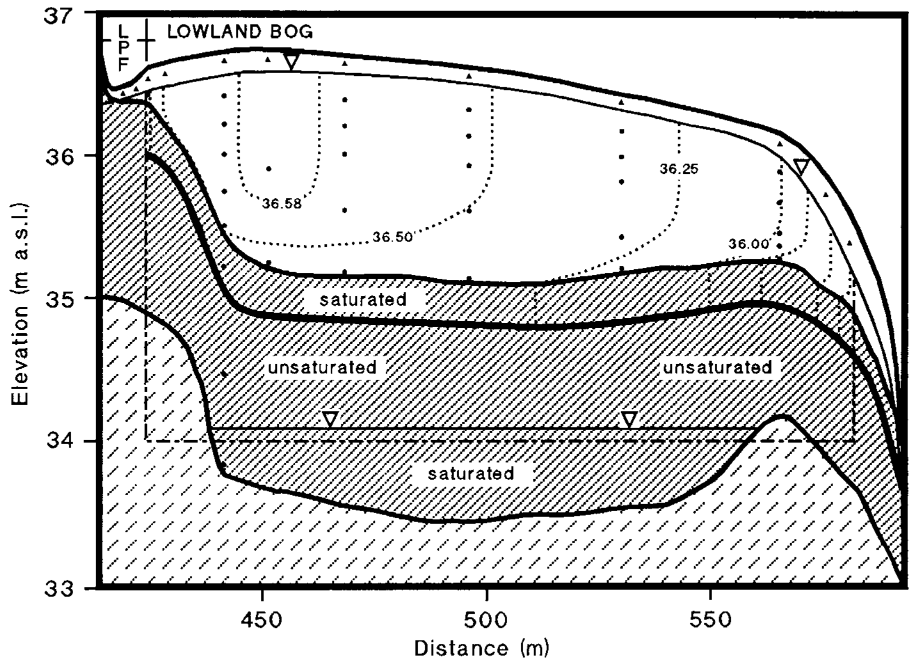

Figure 3. Observed 'steady-state' flow regime for lowland blanket peat complex. See Figure 1 for material legend. Equipotential lines are given as dotted lines: ● = permanently installed piezometers; ▲ = permanently installed wells

The flat impermeable lower boundary condition in the model was used to represent the geometry of the thin impermeable placic horizon, the actual lower no-flow boundary in both peatland complexes. In the model these were represented as a series of non-deformable layers (between gridlines) with permeabilities of 1×10^{-10} cm s^{-1}. The K_{sat} of the remaining layers were initially set according to the measured geometric average values shown in Figure 2. Individual K_{sat}-layers were considered to be homogeneous and isotropic. Once the flow domain was parameterized, the model was run and the simulated water table allowed to deform. The model was calibrated by modifying the K_{sat} of the top K_{sat}-layer (approximately top 0·3 m of the flow domain), where K_{sat} was observed to be largest (Figure 2). As water tables during large recharge events often occurred very near the surface of the peatlands, our initial K_{sat} input values for the top K_{sat} layer reflected water transmission properties of materials above the average water table elevation.

Implementation (upland bog/upland poor fen complex)

The upland flow system (Figure 5) grid encompassed a region 3 m thick, 128 m in length and 1 m in width. The flow system grid had 84 grid columns and 10 grid rows. Average D for the study period on the upland bog and upland poor fen were 3·5 and 3·2 mm day^{-1}, respectively. These constant-fluxes were applied to the surface nodes of each respective peatland system. A groundwater divide at the 104 m mark in the upland bog was a symmetry boundary. Thus, only the right-hand side of the dome was modelled. Constant-head nodes were specified at all nodes along the right-hand (outflow) boundary of the flow system at the 232 m mark because it was assumed, like the lowland bog, that outflow into the well-drained heathlands (Lapen and Wang, 1999) was essentially parallel to the surface. The K_{sat}-layer configuration for layers beneath the surface zone initially was defined in a manner depicted in Figure 2 and placic horizon specification and model calibration were performed as discussed in the previous section.

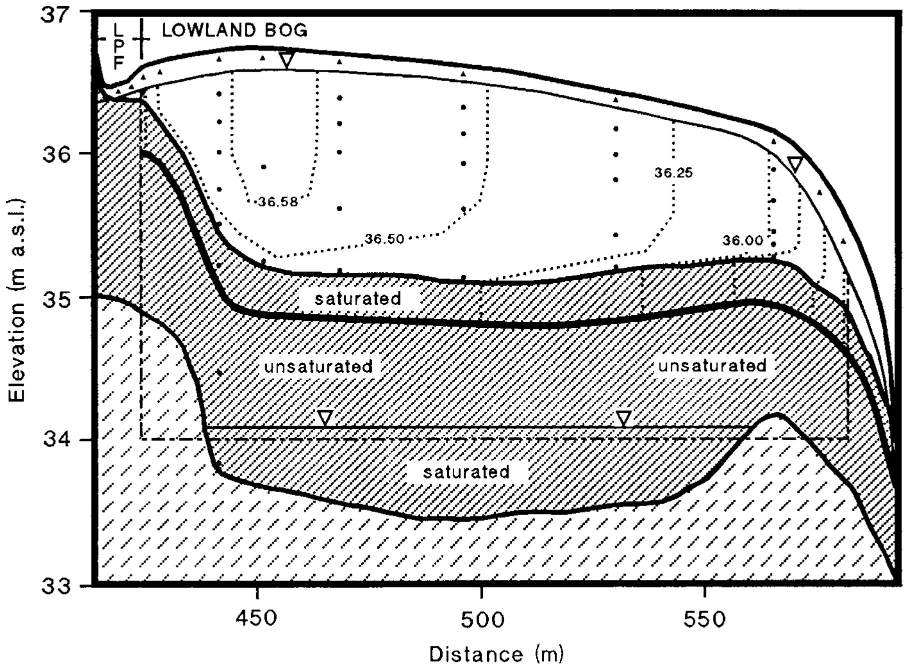

Figure 4. Modelled 'steady-state' flow regime for lowland blanket peat complex. Modelled flow regime was superimposed on Figure 3 site map. See Figure 1 for material legend. Equipotential lines are given as dotted lines ● = permanently installed piezometers, ▲ = permanently installed wells

STEADY-STATE MODELLING RESULTS

Recharge and storage

Field data indicated that total precipitation ($R + OP$) averaged 5·3 and 5·4 mm day^{-1} for the study period for the bogs and upland poor fen, respectively (Table I). The 30-year (1951–1980) normal total precipitation (rainfall only) for July and August at the Cape Race meteorological station, was 2·8 and 4·5 mm day^{-1}, respectively (Environment Canada, 1982). Observed rainwater inputs were somewhat higher, but within the normal range of variability. Rain and/or OP occurred on all but 7 days of the 50 day study period. Evapotranspiration (ET) was 38% of total precipitation ($R + OP$). Soil water storage losses and ET were slightly higher in the upland poor fen, relative to bog (Table I).

Average water table depths in the centres of the lowland bog, upland bog and upland poor fen were approximately 0·11(SD = 0·02 m), 0·12 (0·02), and 0·07 (0·01) m, respectively (Lapen *et al.*, 2000). These average depths were similar to those reported for other blanket bogs in the region during the summer growing season (Northlands Associates Ltd, 1989; Price, 1992).

Owing to the relatively small net groundwater storage changes (Table I), average hydraulic heads and recharge were considered to represent, for modeling purposes, 'steady-state' conditions during the study period. Net study period changes in water table depth near the centres of the upland bog, upland poor fen and lowland bog were approximately −0·03, −0·02 and −0·03 m, respectively. The average specific yield of the soil within zones of water table fluctuation for all three peatlands was around 0·2. On a weekly time-scale, net ΔGWS in Newfoundland blanket bogs generally oscillates around zero (Northlands Associates Ltd, 1989; Price, 1992).

Modelling lowland bog complex

Based on the initial boundary conditions previously discussed, the modelled head distribution closely matched the observed heads (Figure 3 and 4). Moreover, model mass-balance errors for hydraulic head and streamfunction solutions were $<10^{-6}$ %. The final calibrated deformed surface K_{sat} layer values are given in Figure 2. The rounded geometric mean (cm s^{-1}) and range of observed K_{sat} values corresponding to a nominal 0 to 0·3 m depth below surface in bog were approximately 0·5(1·2–1·4 × 10^{-1} cm s^{-1}) for the 424–445 m extent, 1·0(1·3–6·2 × 10^{-1} cm s^{-1}) for the 446–565 m extent, and 0·5(1·0–7·4 × 10^{-2} cm s^{-1}) for the 566–583 m extent, as determined from well tests when the water table was within approximately 0·05 m of the bog surface. The range in observed K_{sat} resulted largely from spatial variability in hummock and hollow peat/*Sphagnum* spp. properties and the variable nature of living and dead surface vegetation representative of the acrotelm (Ingram, 1982,1983); a zone where the peat is less strongly decomposed and more porous. The acrotelm of the bog, in particular, has pronounced vertical variability in K_{sat} as it is composed of both live (highly permeable live *Sphagnum* spp.) and decomposed (lower relative permeability) vegetation. Some estimates of acrotelm K_{sat} at 10 to 20 mm depth from surface are of the order of 100 cm s^{-1} (Ivanonv, 1978). However, calibrated K_{sat} values for the top layer determined from steady-state modelling in this study (Figure 2), reflected more strongly K_{sat} values of more decomposed peat associated with the peat at depths between average water table elevation and 0·3 m depth, which ranged between 10^{-1} and 10^{-3} cm s^{-1}. Moreover, this finding is consistent with observations that the live *Sphagnum* spp. layer (typically dominating the 0·05 m of the surface) only transmitted water during very short time periods during the largest observed flow events (hence, 'steady state' water table depth was approximately 0·11 m below surface). Anisotropy may have contributed to differences in calibrated versus observed K_{sat}. Our well data were primarily K_x measures and Beckwith *et al.* (2003a) and Schlotzhauer and Price (1999), respectively, found average anisotropy ($\log_{10}(K_x/K_y)$) values of 0·55 to 0·57 for bog peat.

Figures 3 and 4 demonstrate that groundwater flow is effectively parallel to the water table in the permeable upper K_{sat} layers of the flow system. According to model streamfunction distributions, approximately 90% of steady state flow occurred within the acrotelm of the lowland bog (i.e. within the top 0·3 m of the flow system).

Flow within the deeper, more decomposed, lower K_{sat} catotelm peat was directed predominately downward. The catotelm is a zone where water storage changes are usually negligible (Ingram, 1982,1983). In the saturated mineral soils beneath the peat and above the placic horizon, flow was horizontal because the K_{sat} of these soils is relatively high (0·02 cm s^{-1}).

Modelled horizontal specific discharges were relatively high along bog margins. For example, in the upper ≈0·4 m of flow system at the 566 m mark, specific discharge was approximately 2·4 × 10^{-3} cm s^{-1}. At similar depths in the centre of the lowland bog these respective specific discharges were around an order of magnitude lower (≈ 4 × 10^{-4} cm s^{-1}). These differences largely reflect various degrees of water table drawdown and that steady state flow along bog margins occurred in the highly permeable surface peat. Vertical specific discharges in the catotelm were very low, being approximately 10^{-7} to 10^{-8} cm s^{-1}. Where flow moved vertically downward, the very low K_{sat} catotelm peats, which make up a majority of the saturated profile, significantly reduced velocities.

Modelling upland bog/upland poor fen complex

Based on the previously noted conditions, the modelled head distribution closely matched the observed head distribution (Figures 5 and 6), and once again, the model mass balance errors for hydraulic head and streamfunction solutions were small ($<10^{-11}$ %). Overall, the general shape of the flow regime in the upland complex is similar to that in the lowland complex. Calibration of K_{sat} for modelling was, again, only performed on the upper K_{sat} layer of the bog flow system (initially 0 to 0·3 m) and for the bulk saturated layers for the fen. Final calibrated surface K_{sat} layer values in bog are given in Figure 2. For fen, the calibrated values for the deformed saturated layers above the discretized placic horizon layers were 4·9 × 10^{-2} and 6·6 × 10^{-2} cm s^{-1} for the 194–211 and 212–232 m fen extents, respectively. The rounded geometric mean (cm s^{-1}) and range of observed K_{sat} values corresponding to a nominal 0 to 0·3 m depth below surface in bog were approximately 0·7 (1·7–5·0 × 10^{-1} cm s^{-1}) for the 104–165 m extent and 0·4 (1·2–4·3 × 10^{-2} cm s^{-1}) for the 166–193 m

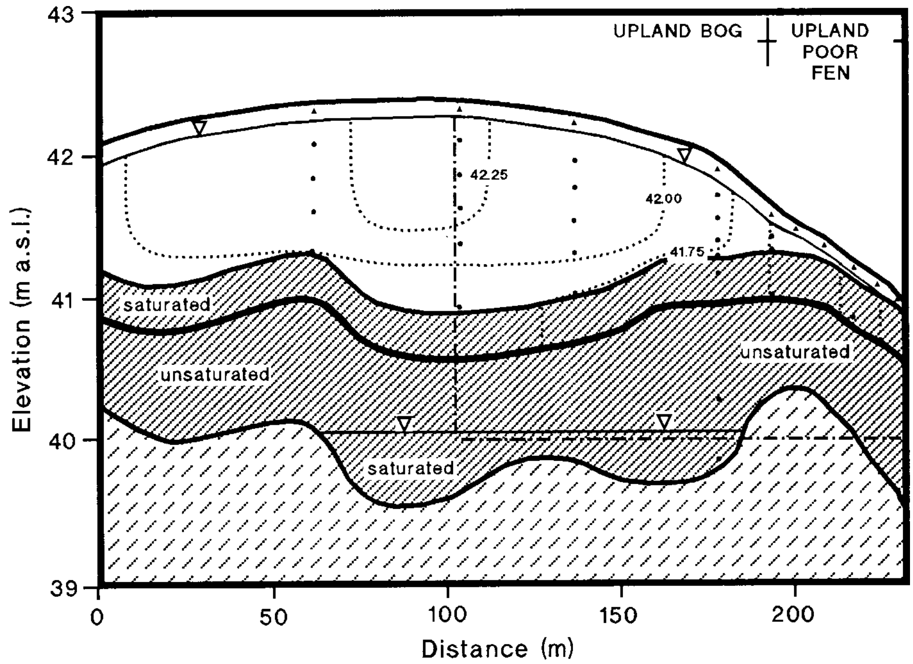

Figure 5. Observed 'steady-state' flow regime for upland blanket peat complex. See Figure 1 for material legend. Observed equipotential lines are given as dotted lines: ● = permanently installed piezometers; ▲ = permanently installed wells

extent. For fen, observed K_{sat} (range) values from well tests for peat/mineral soils above the placic horizon when the water table was observed at the surface was 1·2 (4·2–6·5 × 10^{-3} cm s^{-1}) for the 194–211 m and 1·1 (5·1–3·1 × 10^{-3} cm s^{-1}) for the 212–232 m extent in fen. Maximum observed values were associated with hummocks. The high values for the fen were linked to porous and hummocky *Sphagnum fuscum, S. fimbriatum* and *S. nemoreum* and undecomposed surface litter materials that directly overlayed a highly decomposed organic/mucky layer and mineral soil mix. During high flow events, water transmission through this surface material was probably considerable. Permeameter measures of fen *Sphagnum* spp. K_x had orders of magnitude of 10^0 and 10^1 cm s^{-1}. Some surface flow was observed in some of the hollows in the fen during the largest rain events, as well. Lowest observed K_{sat} for the upland bog–upland poor fen complex flow zones were predominantly associated with hollows: hollows in fen dominated by more highly decomposed peat–muck–mineral soils.

Groundwater flow in the acrotelm was parallel with the slope of the water table, and like the lowland bog, approximately 90% of steady-state discharge occurred there (i.e. within the top ≈0·3 m of flow system). In the catotelm, flow was directed downwards. Shallow flow in the relatively permeable upland poor fen was parallel with the water table, except at the upland bog–upland poor fen margin where there was some vertical flow towards the water table.

Horizontal specific discharges for modelled conditions in the upper ≈0·4 m layer at the 125 and 176 m marks were around 1·6 × 10^{-4} and 2 × 10^{-3} cm s^{-1}, respectively. Discharges at the 220 m mark in the upland poor fen were ≈8 × 10^{-4} cm s^{-1}. The steeper hydraulic gradients in the fen and along bog margins accounted for, in large part, the higher flow velocities.

Like the lowland bog complex, the upland bog–poor fen complex was predominantly a recharge zone; however, at the upland bog–upland poor fen boundary (193 m mark), flow was directed slightly upward toward the water table. Modelled net-discharge was (−0·096 m day^{-1}) at this location.

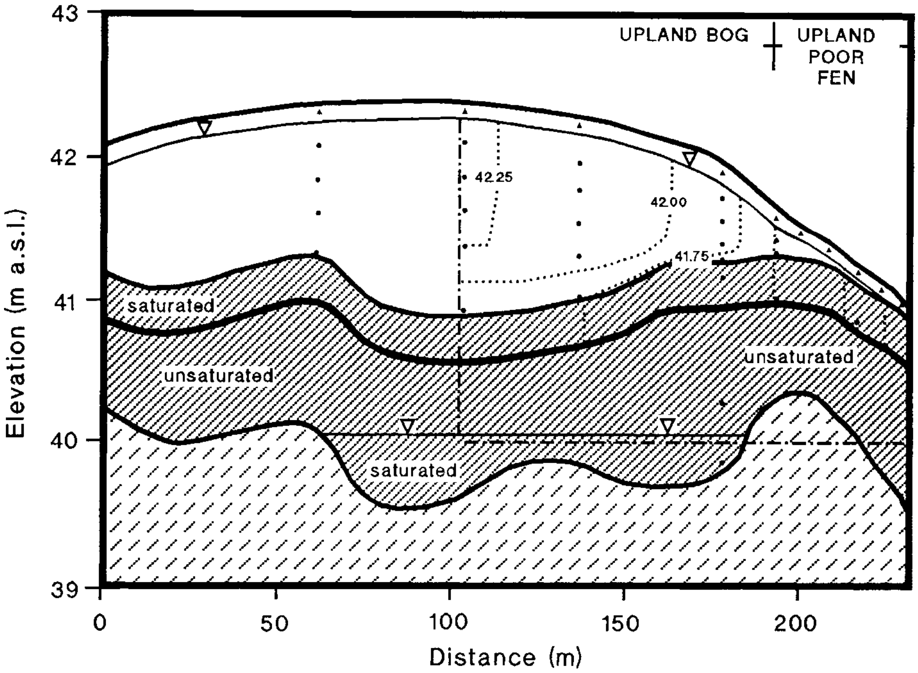

Figure 6. Modelled 'steady-state' flow regime for upland blanket peat complex. Modelled flow regime was superimposed on Figure 5 site map. See Figure 1 for material legend. Equipotential lines are given as dotted lines ● = permanently installed piezometers, ▲ = permanently installed wells

SENSITIVITY ANALYSIS RESULTS

Role of peat hydraulic properties

Simple sensitivity tests were conducted using the flow model to further assess the sensitivity of flow and water table geometry to changes in flow regime K_{sat}. The tests involved systematically increasing K_{sat} over certain flow domains while keeping steady-state model recharge rates constant. To illustrate, the K_{sat} of the upper three K_{sat} layers of the margin (566 to 583 m mark) in the lowland bog (Figure 2) were increased, from steady-state model values to steady-state model values similar to those respective layers within the central portion of the bog (446–565 m mark) (Figure 2). This resulted in a 0·15 and 0·33 m lower water table at the centre (493 m mark) and margin (566 m mark), respectively. Equivalent increases in the K_{sat} of the upper three K_{sat} layers in the middle of the lowland bog (446 to 565 m mark), while maintaining steady-state model K_{sat} along the margins constant, resulted in only a 0·08 m decline in the water table at the centre (493 m mark) and a 0·05 m increase in the water table at the margin (566 m mark). Sensitivity tests conducted on the upland bog and south-east portion of the lowland bog gave similar type results. The tests suggest that shallower, low K_{sat} peat along sloping bog margins, coupled with water table drawdown into less permeable peats, helped to maintain a more elevated, lower gradient water table within bog interiors.

Role of recharge to the water table

Tests were also conducted to evaluate potential sensitivity of groundwater flow to changes in recharge (D) while keeping the modelled K_{sat} regimes constant. Under the steady-state conditions observed in the models, the upland and lowland poor fens received flow inputs from adjacent bogs amounting to approximately 0·31 and 0·11 m^3 day^{-1} per m width, respectively. By applying a water table recharge flux of 2·0 mm day^{-1} (i.e. approximately 60% of observed D) to the entire upland complex, the flow patterns did not change substantially; however, drainage to the upland poor fen from the bog decreased by about 0·13 m^3 day^{-1} per m width. In the lowland bog complex, however, flow patterns experienced important changes under a more systematic recharge modification (steady-state recharge was systematically decreased by 0·1 mm day^{-1} from a maximum of 3·4 mm day^{-1}). Groundwater flow from the lowland bog to the lowland poor fen was found to cease altogether when surface recharge approximated 2·2 mm day^{-1} (64% of steady-state D). The steady-state constant head boundary conditions were not adjusted for this particular sensitivity analysis. In a more conservative manner, and one probably approaching more realistic outflow boundary conditions during lower recharge rates (Lapen and Wang, 1999), the constant heads specified along lowland bog outflow boundaries were lowered by 0·3 m and systematic decreases in recharge were applied to the flow system in the manner previously described. It was found that contributing areas to the lowland poor fen decreased systematically with reductions in recharge as the water table mound in the bog subsided. Flow contributions from the lowland bog to the lowland poor fen ceased completely when recharge rates were approximately 1·7 mm day^{-1} (50% of steady-state D).

Sensitivity tests conducted by systematically increasing recharge rates (by 0·1 mm day^{-1} from steady-state values) did not result in flow reversals in the lowland bog and increased water table elevations in both complexes. To illustrate, under a 5·0 mm day^{-1} water table recharge scenario (keeping steady-state constant-head boundary conditions and K_{sat} regime constant), the contributing area for the lowland poor fen increased by approximately 12 m^2. In the following section, the implications of the sensitivity tests will be discussed in the context of peatland development.

GENERAL IMPLICATIONS: PALUDIFICATION AND BLANKET BOG MAINTENANCE/DEVELOPMENT

Bog peat K_{sat} was found to be inversely related to von Post H values in this study (Figure 2), as noted elsewhere (eg. Baden and Eggelsmann, 1963; Boelter, 1965). The spatial pattern of K_{sat}, therefore, appears to be related to the spatial pattern of decomposition. Decomposition was observed to be greatest at bog margins. It is speculated that these areas experienced deeper and more sustained water table drawdown and greater gradient induced flow velocities during high water table conditions, thus increasing the oxygen content of acrotelm peats (Table II). In support of this contention, redox potentials below the water table near the centre of the bogs were generally smaller than those along the steeper margins. In the fen, minimum soil redox potentials in the unsaturated zone were lower than for bog; albeit the range in redox values were more variable (more variable vegetative and peat properties in the fen, relative to bog).

Table II. Ranges of redox potentials in the unsaturated (0·02 to 0·12 m depth) and saturated zones (0·02 to 0·05 m below water table) of blanket peat complexes. Measurements were made during non-rainfall conditions when the bog and fen water tables were within approximately 0·08 to 0·15 m and 0·03 to 0·10 m below the surface, respectively

Site	Redox potential (mV)	
	Unsaturated zone	Saturated zone
Lowland bog		
438 m	754–820	−9 to 701
490 m	788–813	−101 to 620
575 m	554–809	3 to 777
Upland bog		
106 m	742–808	−97 to 711
174 m	697–814	−13 to 808
Upland poor fen		
219 m	30–708	89 to 746

The model results and field observations demonstrate how changes in K_{sat} distribution and water table recharge within established blanket bogs potentially could affect peatland developmental processes within and immediately adjacent to mature blanket bog. The sensitivity of system hydrology to K_{sat} along bog margins suggests that the decomposition of bog-margin peats, and resulting reduction of peat K_{sat}, could have a positive impact on peat accumulation processes within bog interiors by reducing drainage and augmenting water table doming. As it was shown previously, water table geometry largely governs flow direction, and by extension, water inputs essential for paludification and pedogenesis (sesquioxide pan formation) in adjacent laggs and fens (Lapen, 1998; Lapen and Wang, 1999).

The suite of recharge sensitivity tests indicate that decreases in water table recharge probably would have an adverse impact on drainage from bog to adjacent fens. This is especially true for the lowland poor fen because of the nature of lowland bog water table geometry and the fen's relative topographic position with respect to a limited groundwater contributing area. There, in addition to lower direct precipitation, it was shown that moderate decreases ($D = 50$ and 65% of steady-state model recharge) in water table recharge could substantially minimize contributing areas and even eliminate important flow contributions from the lowland bog to adjacent fen entirely. This is because of the change in contributing area brought about by the shift in the peak of the groundwater mound directing water to the lowland poor fen. To underscore flow regime sensitivity to recharge rates, if the average of the 30-year July and August daily rainfall normals are used in Equation (1) (keeping other recharge parameters measured during the field season constant) instead of the daily rainfall inputs observed during the study period, water table recharge to the lowland bog would be approximately 2·4 mm day^{-1}. Using this water table recharge rate as a baseline, recharge rates at which flow from the lowland bog to the lowland poor fen would cease entirely, represent conservatively $>71\%$ of the baseline rate: notwithstanding systematic decreases in contributing area with a systematic decrease in water table recharge.

Fens in the region have a very limited spatial extent (Graniero and Price, 1999a). However, these narrow areas are not distributed randomly over the landscape but typically are spatially associated with blanket bog margins where hydraulic gradients within the bog evidently support groundwater inputs to fen. As topography was found to explain statistically about 22% of the present distribution of bog in this landscape (Graniero and Price, 1999b), both pedogenic factors and upslope water inputs from bog must be significant paludification mechanisms (Lapen and Wang, 1999).

Sesquioxide pans in the mineral substrate can impede vertical drainage, which over some sites (heathlands lacking placic horizons), is critically limiting with respect to surface saturation because K_{sat} of the sandy Podzols is typically $>10^{-3}$ cm s^{-1} (Lapen and Wang, 1999). The presence of sesquioxide pans would favour peat accumulation, as would a change to a wetter climatic condition. Nevertheless, there also appears to be a link, at this particular site, between sesquioxide pan formation and inputs of acidic, low redox and organic drainage waters from adjacent wetland systems. In this context, the sensitivity tests in this study indicate that moderate decreases in water table recharge could potentially have a more detrimental impact on pedogenic processes in the lowland bog, compared with the upland bog. Along with the greater length of time it took for the lowland poor fen, relative to the upland poor fen, to establish a domed morphology suitable to establish flow inputs to contemporary fen locales, this might partially explain why peat accrual, growth of *Sphagnum fuscum* (main bog carpet species), Fe and Al depletion in the surface mineral soil horizons, and sesquioxide pan development in the lowland poor fen are currently less advanced or established relative to the upland poor fen (Lapen and Wang, 1999; Lapen *et al.*, 2000).

For future studies, a running hypothesis could be stated: contemporary paludification-induced blanket bog development, and perhaps in some areas, sesquioxide pan formation adjacent to mature blanket bogs, will be greater where bog contributing areas are larger and/or have greater temporal establishment. As it was, paludification was more advanced in the observed upland system, relative to the lowland system, probably for these very reasons.

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