Nordic Hydrology, 25, 1994, 313-330

Hydrology of a Patterned Bog-Fen Complex in Southeastern Labrador, Canada

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A 0.05 km^2 patterned bog and a 0.29 km^2 ribbed fen were compared to determine the influence of their morphometry on runoff and evapotranspiration. The bog hydrology was dominated by irregularly located and poorly linked pools $\leq$1.0 m deep, separated by prominent ridges containing local water-table maxima. This small basin had an appreciable lag-to-peak time (*e.g.* 6 h), and a low runoff ratio (0.1 to 0.2), because of the large depression storage of the pools, and a long detention storage caused by the weak pool-to-pool linkage. The ribbed fen was dominated by a 1.3 km water track, which consisted of shallow pools ($\leq$0.25 m deep), separated by low ridges that did not produce local water table maxima. The pools were well linked, but their sequential arrangement resulted in a very large detention storage which trimmed the runoff ratio to only 0.003 to 0.07. The lag-to-peak time was relatively short (approx. 3 h), because of sheet flow in the lower water track, just above the weir. The large depression and detention storage of both systems enhanced evapotranspiration losses. Between July 5 and August 14, 1990, evapotranspiration from the bog was 97 mm, compared to 126 mm from the fen, while runoff was only 12 and 28 mm for the bog and fen, respectively. Evapotranspiration from ridges was 67 and 84% of pool evaporation, in the bog and fen respectively.

Introduction

Patterned peatlands are wetlands which are characterized by alternating peat ridges and pools oriented parallel to the contours of surface elevations. They occur

frequently in boreal and subarctic regions of the northern hemisphere. In south-eastern Labrador, for example, they cover up to 25% of the land surface (Foster 1985; Foster and Glaser 1986). Patterned peatlands, therefore, play an important ecological and hydrological role, although they remain poorly understood.

Patterned peatlands comprise both bogs and fens. Bog peatlands are normally disconnected from regional surface and groundwater inputs by virtue of their topographic position within the landscape (Ivanov 1975). Being fed only by rain, they are generally oligotrophic, and are characterized by acid waters. The surface is dominated by non-vascular plants such as *Spagnum* (moss), and *Cladonia* (lichen), and some vascular woody plants (Jansens and Glaser 1986). Fens typically receive groundwater and surface water inflow (Siegel and Glaser 1987), which enhances their mineral nutrient status, and support a more diverse assemblage of vascular plants (Malmer 1986). Hydrological processes operating within a bog basin are entirely responsible for the nature of its discharge regime. However, the proportion of upland area in a fen basin, and the hydrological processes operating therein, may profoundly affect the fen system (Carter 1986). In turn, the fen system modulates basin runoff (Price 1987).

In summer, flow from bogs may cease, and produce no runoff except following large rain events (Bay 1969). Surface and groundwater inflows to fens sustain the water table at a higher elevation, producing a more regulated runoff regime (Verry and Boelter 1975). The position of the water table within the peat profile controls the rate of lateral drainage, thus runoff characteristics (Ingram 1983). This results from the high hydraulic conductivity of the acrotelm (Boelter 1969), which is the upper layer of dead and poorly decomposed vegetation (Ingram 1978). In contrast, the deeper, more decomposed plant material in the catotelm (Ingram 1978), is much less permeable. The hydrological processes which characterize patterned peatland systems are dominated by low slope, rough microtopography, and the numerous pools which impart a large depression storage capacity (Price *et al.* 1991). During periods of high water table, as in the spring, depression storage may be exceeded so that pools become interconnected, effectively delivering water from the basin (Quinton 1991). During periods of lower water table, the pools become hydrologically disconnected, and runoff is generated by small source areas immediately upstream of the outflow (Quinton 1991).

The position of the water table also affects the loss of water to evapotranspiration. The higher water table generally observed in fens results in larger evaporative losses than in bogs (Ingram 1983). The high water-storage capacity of pools in patterned peatlands presents a large open water surface where evaporation may exceed the potential rate due to heat advection from peat ridges (Price *et al.* 1991).

The morphometry of patterned bogs is distinct from patterned fens, partly due to the size and organization of pool and ridge microtopography (Maloney and Price, under review). The comparative effect of these differences on the hydrological regime have not been well documented. The objective of this study is to compare

and contrast the hydrological processes operating in patterned bog and fen peatlands. Specificially, the effect of the ridge and pool microtopography and basin morphometry will be examined to determine their effect on runoff and evapotranspiration from two small, adjacent bog and fen systems in S.E. Labrador.

Study Site

The study focuses on a 0.29 km^2 patterned fen basin and a 0.05 km^2 patterned bog basin located on the south side of Lake Melville, 65 km northeast of Goose Bay, Labrador, Canada (59°29'N, 53°33'W) (Fig. 1). The peatlands are situated on a raised marine terrace that has rebounded isostatically since deglaciation between 8640 and 5330 BP (Fulton and Hodgson 1979). The terrace is approximately 15 to 30 m a.s.l. with a slope of approximately 0.005. The peat covering the terrace is approximately 1.5 m thick, and underlain by a thin layer of sand and lag deposits ($\leq$ 0.1 m) at the base of the mountains, which in turn cover deep deposits of marine sediments that can be in excess of 65 m (Grant 1975; Vilks *et al.* 1987). The peatlands are backed by the Mealy Mountains which rise abruptly to an elevation of 600 m, and are composed mainly of anorthosite criss-crossed by gabbro dykes and sills (Lopoukhine *et al.* 1977).

The bog can be classified as *domed bog* (NWWG 1987), which has become isolated from regional surface and groundwater inflow by deeply incised channels which cut through the marine terrace. The string bog basin boundary was estimated from the elevation of pools in the area, by selecting boundary pools with local elevation maxima (see Price *et al.* 1991). The ridges are covered primarily with *Sphagnum spp.* and lesser amounts of *Ericaceae spp.* and *Cladonia spp.*

The fen basin is dominated by a water track that extends from the base of the Mealy Mountains to a point 1,250 m downslope. The water track is a locally depressed zone of preferred flow, which has developed over former channel beds in the mineral substrate (Maloney and Price, under review). Within the water track, vegetation is primarily sedges (*Carex spp.*), mosses (*Sphagnum spp.*), and cottongrass (*Eriophorum spissum*). Outside the water track, vegetation consists of mosses (*Sphagnum spp.*), and cottongrass (*Eriophorum spissum*). The forested section of the fen basin extends 400 m upslope onto the colluvial deposits of the Mealy Mountains, reaching an elevation approximately 50 m above the fen. The vegetation is predominantly black spruce (*Picea mariana*) with some balsam fir (*Abies balsamea*). Identification of plants follow Crum and Anderson (1981) for mosses, and Ryan (1978) for vascular plants.

The climate of the study area is controlled by a mixture of Maritime tropical, Bermudian subtropical and Continental sub-arctic anticyclones during the summer, and the Continental sub-arctic anticyclone and Icelandic low air masses during the winter (Banfield 1981). The 30-year (1951-1980) mean annual temperature at

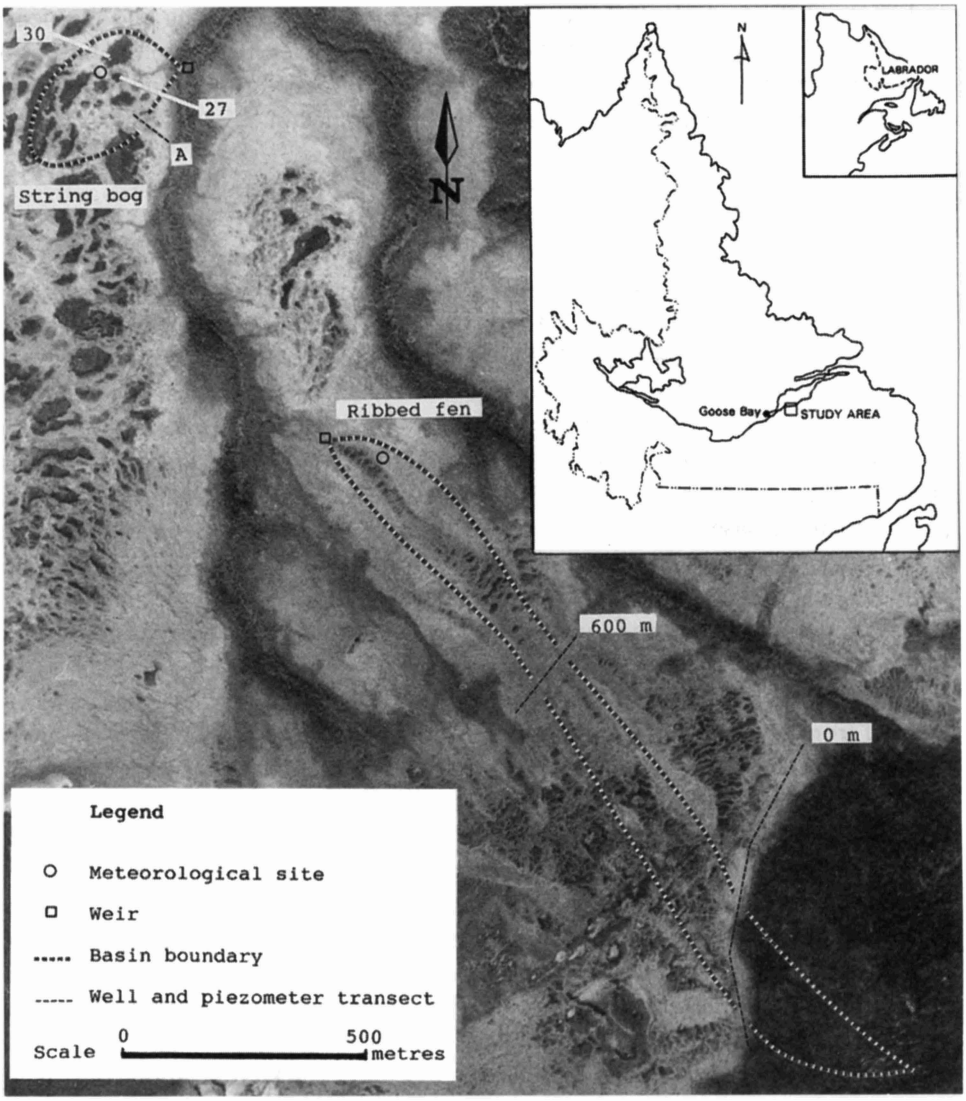

Fig. 1. The study area. NW portion of photo shows raised ombrotrophic peatland and patterned bog basin. NE to SW trending fen basin shows water-track, and its connection to the upland forested portion of the fen basin (SE).

Goose Bay is 0.0° C with mean January and July temperatures of −16.4° C and 15.8° C respectively (Atmospheric Environment Service 1984). The study area receives 946 mm of precipitation annually, of which 45% occurs as snow (Atmospheric Environment Service 1984).

Methods

Precipitation, evapotranspiration and runoff data were collected from July 5 to August 14, 1990. Precipitation was recorded by a tipping bucket rain gauge located approximately 0.5 m above the surface of the bog. Two manual rain gauges were deployed in the fen, one at either end of the water track.

Four soil and four pool lysimeters (each 0.6 m^2 and 0.15 m deep) were used at both the bog and fen to determine actual evapotranspiration. Soil lysimeters were filled with undisturbed monoliths of peat with representative surface covers. These were weighed daily to determine evapotranspiration for the previous 24 hours. Water was added to the lysimeters as necessary to keep them at a level of saturation similar to their surroundings. Pool lysimeters consisted of clear colourless bowls supported by a 1.5 × 1.5 × 0.06 m styrofoam floatation/splashguards. Evaporation for the previous 24 hours was calculated by measuring the amount of water necessary to return the water level in the lysimeter to a fixed mark. Data for which there was a > 20% discrepancy from the mean of the four values were rejected (Roulet and Woo 1986). Estimates of daily evapotranspiration from pool and peat surfaces of each basin were made using the combination model of Priestley and Taylor (1972), such that

$$E = \alpha \frac{s}{s+q} \frac{Q^* - Q_G - Q_W}{L\rho} \tag{1}$$

where E is the evapotranspiration rate (mm d^{-1}), L is the latent heat of vaporization (J kg^{-1}), ρ is the density of water (kg m^{-3}), s is the slope of the saturation vapour pressure-temperature curve (Pa $°C^{-1}$), q is the psychrometric constant (0.0662 KPa $°C^{-1}$ at 20° C), Q^* is the net radiation flux (J d^{-1}), Q_G is the ground heat flux (J d^{-1}), Q_W the pool heat storage (J d^{-1}). When $\alpha = 1$, Eq. (1) represents equilibrium evaporation, which is the condition when there is no vapour pressure deficit in the near surface atmosphere. The ratio of actual (lysimeter) and equilibrium evapotranspiration provides an empirical coefficient (α) which can be used in Eq. (1) to estimate evapotranspiration when direct measurements are unavailable, but when net radiation, substrate heat flux, and air temperature are available.

Net radiation was recorded at both sites with radiometers 1.0 m and 0.5 m over peat and pool surfaces, respectively. The ground-heat flux was measured by soil-heat flux plates 2 cm below peat surfaces typical of each site. The air temperature was measured at both sites by a shielded thermistor 1 m above the peat surface. Pool-heat storage was determined as

$$Q_W = C_W \sum \frac{\Delta T}{\Delta t} \Delta z \tag{2}$$

where C_W is the heat capacity of water (J kg^{-1} $°K^{-1}$), ΔT is the change in water temperature (°K), Δt is the time interval (s) over which the change in water temperature was recorded, and Δz the depth of the water layer (m) over which the

temperature changed (Oke 1978). A string bog pool approximately 0.5 m deep, had thermistors located at its surface and base; the soil heat flux at the base of the string bog was assumed to be zero. A ribbed fen pool 0.2 m deep had a thermistor located 0.10 m depth below the surface and the heat flux into the bottom peat was measured by a soil heat flux plate. Evapotranspiration from the forested section of the fen basin was determined by the Priestley and Taylor (1972) method (Eq. (1)) based upon parameters derived from the literature. The forest was assumed to receive an average of 15% more net radiation than open areas (McCaughey 1981), and have an $\alpha = 1.0$ as it was not water stressed during the field season (*e.g.* Munro 1986).

Streamflow from both basins was measured with V-notched weirs calibrated against the water elevation behind them, by collecting flow over a timed interval. In the bog, a 0.3 m high weir board was installed across the rivulet fed by a soil pipe draining pool 30 and the bog basin. In the fen a weir box was constructed to direct surface and subsurface flow from the water track to the weir. The box was installed 0.4 m below the surface, and was open at the upslope end.

Subsurface flow was determined by Darcy's Law

$$Q_{ss} = -K \frac{\Delta h}{\Delta l} w d \qquad (3)$$

where K is hydraulic conductivity (m d^{-1}), $\Delta h/\Delta l$ is the hydraulic gradient (dimensionless), w is the width of the flow face over which drainage is occurring (m), and d is the depth of the saturated flow face (m). One transect containing 8 wells (21 mm i.d.) and 9 piezometers (16 mm i.d.) was located across the water track at the head of the ribbed fen (0 m), and another transect containing 4 groundwater wells and 1 piezometer was located across the water track 600 m downslope (600 m) (Fig. 1). Within the string bog, the transect at the basin boundary contained 7 groundwater wells ("A" in Fig. 1). Wells were slotted over their entire length and penetrated the peat layer to the mineral substrate. Piezometers had a slotted intake length of 100 mm and were set to the bottom of the peat layer. One piezometer was installed 0.5 m into the sediments of the marine terrace. The water level in the wells and piezometer was measured daily. Hydraulic conductivity, K, was determined in the field with bail tests using the method of Hvorslev (1951). The influence of water-table elevation on hydraulic conductivity was determined on the bog, in the fen-water track, and in the fen outside the water track. This was done by evaluating the hydraulic conductivity at a given location at periods of different water-table elevation.

Fig. 2. a) Oblique photo of bog basin looking south. More elevated and dryer portion of ridges support more lichen, and appear lighter than the *Spagnum* carpet ringing some pools. Pool 30 (foreground) is the lowest pool in the basin, and drains through a pipe (dark line going to left of photo). b) Fen water-track near the weir. Note shallow pools, low ridges, and general flooded appearance, and presence of vascular plants.

Results

The hydrological data reported below describe 1) the water flow and storage processes *within* the fen and the bog basin; and 2) the water exchanges *to* and *from* the individual basins, *i.e.* comprising the water balance components.

Internal Water Dynamics

The patterned bog contains approximately 67% peat cover and ridges, and 33% open water in pools which are typically 80 m long and 40 m wide. The pools are generally $\leq$ 1 m deep, and separated by peat ridges 0.3 to 0.5 m high and 10 to 20 m wide (Fig. 2a). No direct channels or soil pipes connecting bog pools were observed. The hydraulic connection between pools was by seepage under ridges. This was measured between pools 27 and 30 of the bog (Figs. 1 and 2a). The water table in the ridge was always above the water level of the upper pool (No. 27) during the study period. This precluded pool-to-pool water flow through the acrotelm, even though there was a 0.26 m elevation difference between pools 15 m apart. Water transfer was limited to deeper flow through the catotelm, where the mean hydraulic conductivity is only 8.0×10^{-6} m s^{-1}. Therefore, flow across the 30 m long, 0.5 m high submerged ridge face was estimated to be approximately 0.2 m^3 d^{-1}.

In contrast to the weak pool-to-pool water transfer, a vertical pipe located on the saturated carpet fringing pool 30, drains pool water into the incised channel to the east. The horizontal reach of the pipe is indicated by woody vegetation (Fig. 2a), which exploits the locally recessed water table. From the surface, it appears as a dry linear depression. The pipe exits close to the incised channel (not visible in Fig. 2a) at the level of the mineral substrate. (Runoff from the bog was measured at this point with a weir).

The fen basin consists of 49% peat ridges, 13% pools and 38% upland forest. It can be classified as *northern ribbed fen* (NWWG 1987). The forested section occurs at the head of the basin, on the colluvial slopes of the Mealy Mountains. The fen basin is dominated by a water track, a zone of preferred flow characterized by a locally high water table, and segmented by low ridges oriented perpendicular to the flow direction. The pools at the head and base of the fen differ both in dimension and shape. At the head of the fen, pools are typically rectangular in shape measuring 25 to 50 m long, 5 to 7 m wide, and < 0.25 m deep, separated by ridges 0.1 to 0.4 m high. Near the fen outlet pools are more circular, 5 to 10 m across, and 0.1 to 0.2 m deep, separated by ridges 0.05 to 0.2 m high (Fig. 2b).

The water track begins where water from the colluvial slopes emerges onto the coastal plain (Fig. 1). Vertically and horizontally oriented pipes 0.05 to 0.1 m in diameter were observed to discharge gently into pools and short open channels leading to pools. Along the fen-water track, pool elevation decreased sequentially toward the outlet. No groundwater mounds were observed in the ridges, therefore water flowed readily across ridges, which in the water track were of relatively high

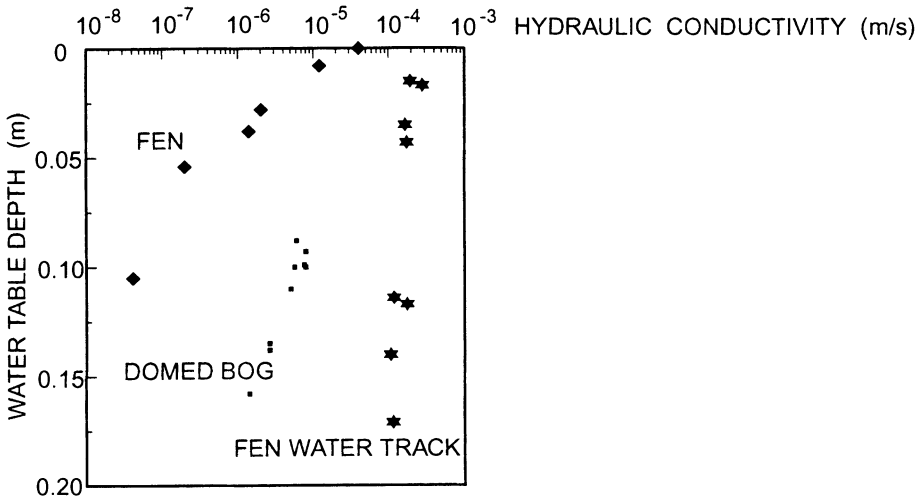

Fig. 3. Hydraulic conductivity plotted as a function of water-table depth below the surface. The water-table depths encompass the ranges encountered during the full measurement period. The three locations include the fen water-track, the fen outside the water-track, and the bog (transect "A").

hydraulic conductivity (Fig. 3). However, the poor definition of the water track at the 600 m range of the main transect (Fig. 1), affects the connection between the upper and lower section of the fen basin. At the 600 m transverse transect, the water table varied between 0.05 and 0.23 m below the surface of the 0.8 m thick peat layer, and had a hydraulic gradient of 0.006. The basin is 180 m wide at this location, the water track portion is 60 m wide. Hydraulic conductivity of the water track was only marginally affected by water table position (Fig. 3), but more so outside the water track. Based on the above information, subsurface flow through this section was estimated to be approximately 2.5 m^3 d^{-1}, or about 3% of the daily runoff.

Vertical hydraulic gradients were measured at transverse transects at 0, 600, 800, and 1,250 m. There are discontinuous interbedded sand layers underlying pools at the footslope (0 m transverse transect), which had an average (± standard deviation) upward hydraulic gradient of 0.002±.006 (two of eight piezometers). Elsewhere at this transverse transect, the average hydraulic gradient was −0.076±.032 (downward). Piezometers elsewhere on the fen had an average hydraulic gradient of −0.024±.031. This average includes other sections of the water track, where occasionally there was an upward gradient.

Basin Water Inputs and Losses

Precipitation recorded between July 5 and August 14 totalled 120 mm. The maximum variation in total gauge catch for any of the three gauges employed was ±7%.

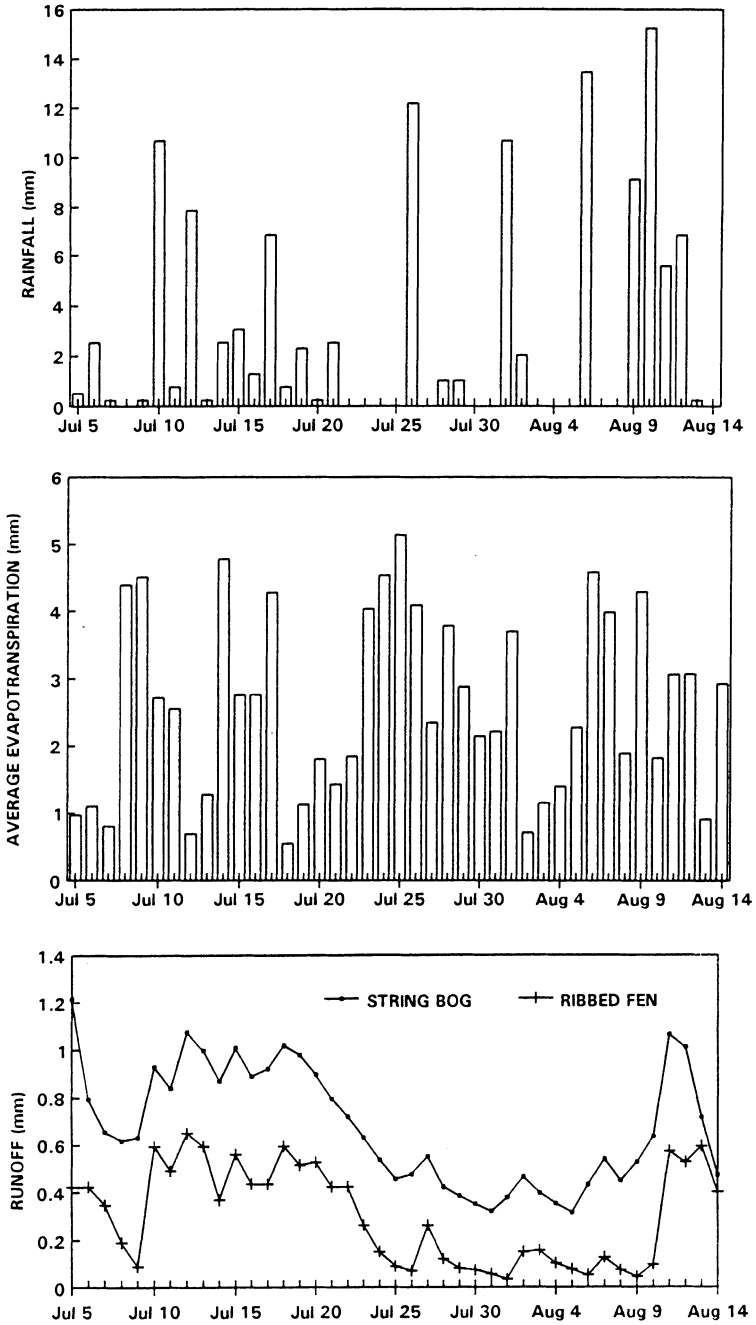

Fig. 4. Rain (upper), evaporation (middle), and discharge (lower) for the bog and fen study basins between July 5 and August 14, 1990.

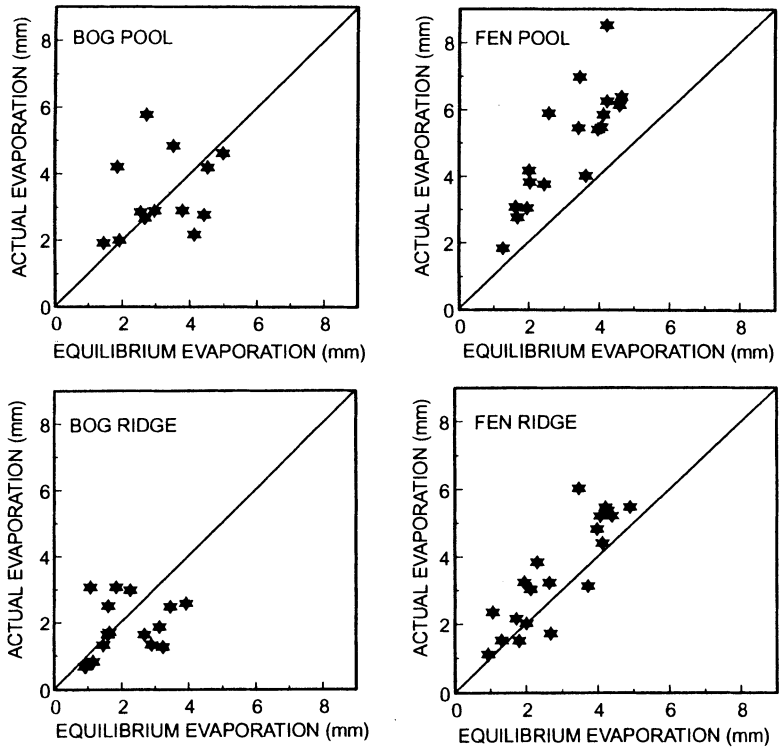

Fig. 5. Actual (lysimeter) *vs.* equilibrium evapotranspiration for pool and ridge locations in the bog and fen.

July and August rainfall at Goose Bay were 91 mm and 111 mm, respectively, compared to the 30-year mean of 105 mm and 103 mm (Atmospheric Environment Service 1984, 1992). Daily precipitation events were generally small, none exceeding 16 mm (Fig. 4).

Direct evapotranspiration from the lysimeters is plotted against equilibrium evaporation (Fig. 5). Compared to the fen, there was more scatter in the data from the bog, and the values were smaller. The slope of the best fit line in each plot in Fig. 5 represents the α co-efficient of evaporability (Eq. (1)). This is the quantity by which equilibrium evapotranspiration must be multiplied to estimate actual evapotranspiration (Table 1). The evaporative efficiency from ridges was only 67% and 84% of pools, in the bog and fen, respectively. The areal average evapotranspiration for bog and fen peatland was 2.3 and 3.4 mm d^{-1}, respectively, with values ranging up to 5 mm d^{-1} (Fig. 4). Based on the assumed parameters for the forested section of the fen basin (net radiation 15% higher, α = 1.0; see "Methods" section), average forest evapotranspiration was 2.5±1.3 mm. The areal average over the fen basin, including peat, pond and forest surfaces, was 3.0 mm. Total evapo-

Table 1 – Mean ± standard deviation of the Priestley-Taylor (1972) α value, and evapotranspiration, between July 5 and August 14, 1990

	Bog Pool	Bog Ridge	Fen Pool	Fen Ridge
n†	13	15	26	28
α	1.20±0.50	1.00±0.60	1.55±0.48	1.27±0.35
E mm d^{-1}	3.0±1.7	2.0±1.0	3.7±2.0	3.2±1.6
E_{AVG} mm d^{-1}‡	2.3		3.0*	

† n = number of days of lysimeter data.
‡ Areal average evapotranspiration rate.
* includes forested upland portion of basin (3.4 mm without)

transpiration for the bog and fen basins between July 5 and August 14 was 97 and 126 mm, respectively.

Runoff from the bog was consistently higher than from the fen (Fig. 4). Peak runoff of 1.2 mm d^{-1} occurred on July 5 from the bog, and about 0.7 mm d^{-1} on July 13 from the fen (Fig. 2). The cumulative depth of runoff from the bog and fen during the study period was 28 and 12 mm, which corresponds to values of 0.98 and 0.37 L d^{-1}, respectively. There was no period during which flow ceased entirely. Runoff ratios for precipitation events that occurred during low, medium and high water table positions within each basin are presented in Table 2. The fen had a consistently lower runoff ratio than the bog, although this increased an order of magnitude during periods of higher water table. The runoff ratio in the bog doubled at high water table periods. The time lag (lag-to-peak) between a one-hour storm (July 26) delivering 7 mm of rain, and the peak runoff, was 3 hours for the fen, and 6 hours for the bog.

Because of the integrity of the basin divide, groundwater outflow from the fen basin occurred as a component of the surface outflow, which was measured at the weir (Fig. 1). Seepage below and around the weir were thought to be minimal. In

Table 2 – Runoff ratio for rain events at relatively dry, intermediate, and wet conditions

	Bog§			Fen‡		
Relative water table position†	Low 68 mm	Medium 52 mm	High 43 mm	Low 12 mm	Medium 8 mm	High 1 mm
Date	01 Aug	26 Jul	17 Jul	01 Aug	26 Jul	17 Jul
Rain (mm)	10.8	12.2	6.5	10.8	12.2	6.5
Runoff (mm)	1.4	1.5	0.95	0.03	0.07	0.47
Runoff Ratio	0.1	0.1	0.2	0.003	0.006	0.07

† Elevation below a local datum.
§ At pool 27.
‡ At the fen weir.

the bog basin, juxtaposition of pools alongside an incised channel resulted in some lateral seepage losses from the basin. This was assessed at transect "A" (Fig. 1) based on Darcy's law, where daily flows were calculated based on water-table position, which was used to determine hydraulic conductivity (Fig. 3), and saturated thickness. Given the average hydraulic gradient of 0.004, the cumulative depth of subsurface outflow from the bog over the study period was approximately 1 mm. Deep seepage into the mineral substrate was also negligible, since the mineral sediments have a very low hydraulic conductivity (1×10^{-9} m s^{-1}).

Water Balance

The water balance was calculated as

$$P - E - Q - Q_{SS} = \Delta S + \xi$$

where P is rainfall (mm), E is evapotranspiration (mm), Q is surface discharge (mm), Q_{SS} is subsurface discharge (mm), ΔS is the change in storage (mm) and ξ is the residual term (mm) (Table 3). Precipitation for the water-balance period of July 5 – August 14 totalled 120 mm. Evapotranspiration accounted for 126 and 97 mm of water from the fen and bog, respectively, whereas surface discharge was 12 mm and 28 mm respectively. Subsurface discharge from the bog basin was about 1 mm and there was no seepage loss from the fen. Change in storage was rather complex for the fen. Above the 600 m point along the transect, the water table in the fen dropped 13 mm, whereas below that point, the water table rose 3 mm. Since the 600 m point approximately divides the upper and low fen into halves, the respective changes were weighted equally. Therefore, storage change in the peat ($S_Y = 0.26$) was estimated to be −1 mm, and in the pools −5 mm. On the basis of pool and peat surface area, the net storage change in the fen portion of the basin was approximately −2 mm. In the bog, the water table in the peat rose 18 mm, which for $S_Y = 0.16$, represents a storage change of +3 mm. Change in storage in bog pools averaged +19 mm. The areal weighted average net storage change in the bog was therefore approximately +7 mm.

Discussion

The physical structure of these patterned peatlands stongly influenced their hydrological behaviour, mainly through their effect on the depth and timing of water storage. The bog basin is small (4.7 ha) and circular in shape. These properties are typically associated with responsive and efficient drainage (Ingram 1983). However, the bog was neither, as evidenced by the low runoff ratio (0.1 to 0.2), which is comparable with those of Bay (1969) for low and medium water table elevations in a 9.7 ha unpatterned bog. Bay (1969) reported a runoff ratio of approximately 0.4

when the water table was above the surface of hollows, at which time overland flow pathways are linked (Kadlec *et al.* 1981). It is unlikely that pools in this system become similarly linked because of the high ridges, so the runoff ratio in unlikely to match the value reported by Bay (1969).

The lag-to-peak time of 6 hours following the July 26 storm is a relatively short duration for this basin, compared to 1989. Then, the pool No. 30 water level was 58 mm lower, on average, and the lag-to-peak was 24 to 42 hours (Price *et al.* 1991). In either year of study, the lag-to-peak time of runoff in this basin is much longer than reported by Bay (1969), who found values typically between 1 and 3.5 hours. The limited runoff ratio, and the long lag-to-peak, are due to the presence of large and disconnected pools which comprise 33% of this basin. These impart a large depression and detention storage capacity. The pools do not have a direct or well organized drainage network, but are linked through water flow in the deeper peat (lower hydraulic conductivity) beneath the ridges. Because bog pools are not sequentially linked, and none are distant from the outlet, they may all make a small contribution to runoff in a given event. This partly explains the higher runoff ratio than noted for the fen. The time required for the water level to reach the elevation of the opening of the soil pipe that drains pool 30, and thus the basin, also affected the timing and magnitude of runoff events (*cf.* Woo and DiCenzo 1988), but there are insufficient data to investigate this further.

The physical structure of the fen has some similarities to the bog, but some important differences. The pool-ridge sequence of the fen water-track has considerable depression storage in the upper reaches, where the ridges are higher, but only limited depression storage in the lower reach, near the outlet. The low ridges in the lower fen were frequently flooded (Fig. 2a), thus during storms provided minimal resistance to sheet flow. This explains the relatively short lag-to-peak time (3 hours). Vegetation growth during the summer in the pools and on the ridges may reduce the runoff response from the fen, but there are insufficient events to quantify the effect. Such changes are unlikely in the bog, where surface flow is not important. Prolonged detention storage in the extremely elongate basin (*cf.* Ingram 1983), the large depression storage in the upper reaches of the fen, and the poor hydraulic connection between upper and lower fen (at the 600 m reach of the water-track), increase the time available for evapotranspiration. This is the explanation for the very low runoff ratio (0.003 to 0.07).

Evaporation from fen pools ($\alpha = 1.55$) was greater than from bog pools ($\alpha = 1.2$), because of their smaller size and shallow water depth. Smaller pools experience an "oasis effect" caused by lateral advection of heated air (Bello and Smith 1990). Shallow water is more rapidly heated, thus encouraging convective energy losses. There was considerable scatter in bog pool α values, probably due to splash effects in the lysimeters. However, they were considerably smaller than reported by Price *et al.* (1991) ($\alpha = 2.1$), for the warmer and dryer 1989 season. Warmer and dryer conditions increase the advective heat transfer to pools. In this study, evapo-

Table 3 – Water budget between July 5 and August 15 1990. Terms are as previously defined. All values in mm

Basin	P	E	Q	Q_{ss}	ΔS	ξ	% Error
Fen	120	126	12	0	−2	−20	13
Bog	120	97	28	1	7	1	1

transpiration from bog ridges (α = 1.0) was also less than from fen ridges (α = 1.27). Bog ridges are more elevated and drier than fen ridges (Fig. 2a and b), and have a significant component of non-vascular plants (*Sphagnum, Cladonia*). The overall evaporation rate was therefore significantly greater from the fen (3.4 mm d^{-1}), than from the bog (2.3 mm d^{-1}).

The water balance provides a useful tool to assess the relative importance of the hydrological processes. It is recognized that in spite of the small calculated residual error in this study (Table 3), there is no effective check on the errors of the individual components. Precipitation was measured at three locations, and the difference between the highest and lowest total precipitation was less than 10%. Evaporation estimates incorporate errors especially in the lysimetric analysis, but under ideal conditions the Priestley and Taylor (1972) method is accurate to within ±15% (Stewart and Rouse 1976). Discharge can be measured accurately with a weir (Ackers *et al.* 1978), but determination of basin area (*i.e.* to calculate runoff depth) is subject to errors in areas of low relief. The fen basin is reasonably well defined, so errors are probably within ±15%. There is more uncertainty in defining the drainage area for the bog, but it is likely within 25% of the true value. The effect of this relatively large error is not great, however, since the streamflow values are so small. The same argument can be used for subsurface losses. In spite of the potential for error, the overall impression provided by the water budget is still valid. Evapotranspiration was clearly the dominant water loss, and runoff was a distant second. This is reflected as well, in the previous discussion of large depression and detention storage. The large amount of free surface water (depression storage) enchances the evaporation, and the extensive delay in draining it (detention storage) allows evaporative processes to continue longer.

Conclusion

The movement of water within, and out of patterned peatlands was strongly controlled by the nature and position of pools and ridges within the basin. The patterned bog had relatively high ridges separating large pools. Groundwater mounds within the ridges partly isolated pools, and restricted their inter-linkage to slower, deep groundwater flow. Depression and detention storage were thus increased. However, since no pools were far from the outflow, the detention storage

was less than in the fen. Smaller, sequentially linked pools occurred in a water-track within the fen. This zone conveyed almost all of the flow toward the basin terminus. Higher ridges in the upper part of the fen provided substantial depression storage, whereas low, wet ridges in the lower part of the fen provided little depression storage. Storm events flooded the lower fen, and water was quickly discharged. The long flow path of the fen water-track sufficiently detained water so that most was lost by evapotranspiration.

Soil pipes played an important role in both fen and bog systems. In the bog, a soil pipe delayed water loss from the basin, until the water level in the pool rose sufficiently to drain directly into it. In the fen, soil pipes fed by water from the colluvial footslopes of the Mealy Mountains, discharged onto the terrace. As in the bog, these occur only in the zone of higher hydraulic gradients. The opening of these pipes forms the locus for a series of water tracks (Fig. 1), one of which dominates the fen study basin. The role of pipes requires further study, so that their effect on runoff can be quantified.

Acknowledgements

The authors thank Fred Downey and Brent Keeping for the assistance they provided in gathering the data. The research was funded in part by the Natural Science and Engineering Research Council and by a Northern Training Grant provided by the Department of Northern and Indian Affairs. Valuable criticism on an earlier draft of this manuscript by anonymous reviewers is kindly acknowledged.

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First received: 14 September, 1993
Revised version received: 17 May, 1994
Accepted: 7 June, 1994

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