

Spring and Summer Runoff Hydrology of a Subarctic Patterned Wetland

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Abstract

Patterned wetlands occur extensively throughout the Canadian subarctic and boreal regions. These areas are important because of their hydrological and chemical interaction with other ecosystems, aquatic and terrestrial. However, there is a lack of knowledge on the hydrology of many wetland types in the boreal and subarctic regions, particularly patterned fens. This paper examines the relationship between the flux and storage of water both within a patterned wetland and between it and the adjacent ecosystems. It is hypothesized that patterned wetlands alternate between two hydrological phases. The first occurs when the water supply exceeds the depression storage capacity of the patterned wetland pools, and the wetland drains as a single source area, while receiving inputs from adjacent nonwetland portions of the basin. As a result, basin discharge during such periods is relatively high. The second phase occurs when the loss of water through seepage and evaporation leads to disconnection of the wetland pools into separate microcatchments within the pool-ridge complex. Under this condition, basin discharge is relatively low, maintained only by a limited source area immediately upstream of the outlet. Through an explanation of the hydrological processes and linkages among sub-systems, an understanding of the processes governing runoff from a basin with a patterned wetland at its outlet is obtained. It is demonstrated that the discharge response of patterned wetlands is distinctive from that of other wetland types because the large storage capacity of the wetland pools delays runoff contributions from large portions of the wetland and surrounding basin.

Introduction

Twelve percent of Canada's land surface (111.3 million ha) is wet organic terrain. In the north, wetland coverage often exceeds 50% (National Wetlands Working Group, 1988). In the northern boreal and subarctic regions, many wetlands form patterns, which are characterized by low, narrow ridges, separated by pools that traverse the wetland at right angles to the direction of water flow (Tarnocai, 1980). The peat ridges rise between 10 and 30 cm above adjacent pool surfaces and are spaced between five and tens of meters apart. The pool-ridge sequence imparts a large depression storage on patterned fens (Price and Maloney, 1994). Studies of patterned wetlands have examined the micro- and macrotopography (Ivanov, 1975) and the relationship between hydrology and the evolution of patterns (Foster et al., 1983; Swanson and Griegel, 1988), but there have been no published studies detailing the surface and subsurface flow systems of patterned peatlands.

With the relatively impermeable frost table at or near the surface for an extended period, wetlands offer little attenuation to flow during snowmelt (Woo, 1986). Consequently, the snowmelt hydrographs of subarctic basins containing a large proportion of wet organic terrain have larger peaks than nonwetland basins (Roulet and Woo, 1988). Loss of water to drainage and evapotranspiration following the snowmelt period, combined with the increase in soil and depression storage capacity, results in low summer discharge. Subsurface runoff from many northern wetland types is very low due to the low wetland gradient and low peat hydraulic conductivity (Roulet and Woo, 1986).

Patterned wetlands typically occupy the outlet of the flow pathway of peatland dominated drainage basins. This makes the

wetland an integral component of the surface hydrology of such basins, since most water discharging from the basin must pass through or over the wetland. Woo and Heron (1987) demonstrated that when the storage capacity of wetland pools is exceeded, water passes over and around the ridges, leading to a diffuse pattern of runoff from the wetland. Studies of the ecology of patterned peatlands have inferred that the ridges form an impermeable barrier to subsurface flow (Foster et al., 1983; Swanson and Griegel, 1988), but these studies ignored the possibility of water movement through the basal sediments of the pools. The objective of this study was to develop an understanding of the factors that control the surface and subsurface hydrology of a patterned wetland from the onset of snowmelt to late summer.

On the basis of hydrological studies of other wetland types and given the surface microrelief of patterned wetlands, we hypothesize that the surface flow pattern will have two distinct hydrological phases. When the water supply exceeds the depression storage capacity of the wetland pools, the pools coalesce, effectively making the flow system fully connected. This phase occurs during the spring melt period and following large summer rain events. As the pool water level drops, individual pools disconnect into independent microcatchments separated by ridges. In this phase, little discharge is produced. Between the two phases, there is a transition phase where the flow system is partially connected. We tested this hypothesis in a valley bottom, patterned fen in subarctic Quebec, Canada.

Study Area

The study was conducted in a 24.2 ha drainage basin (Fig. 1), 5 km north of Schefferville, Quebec (54°48'N, 66°49'W).

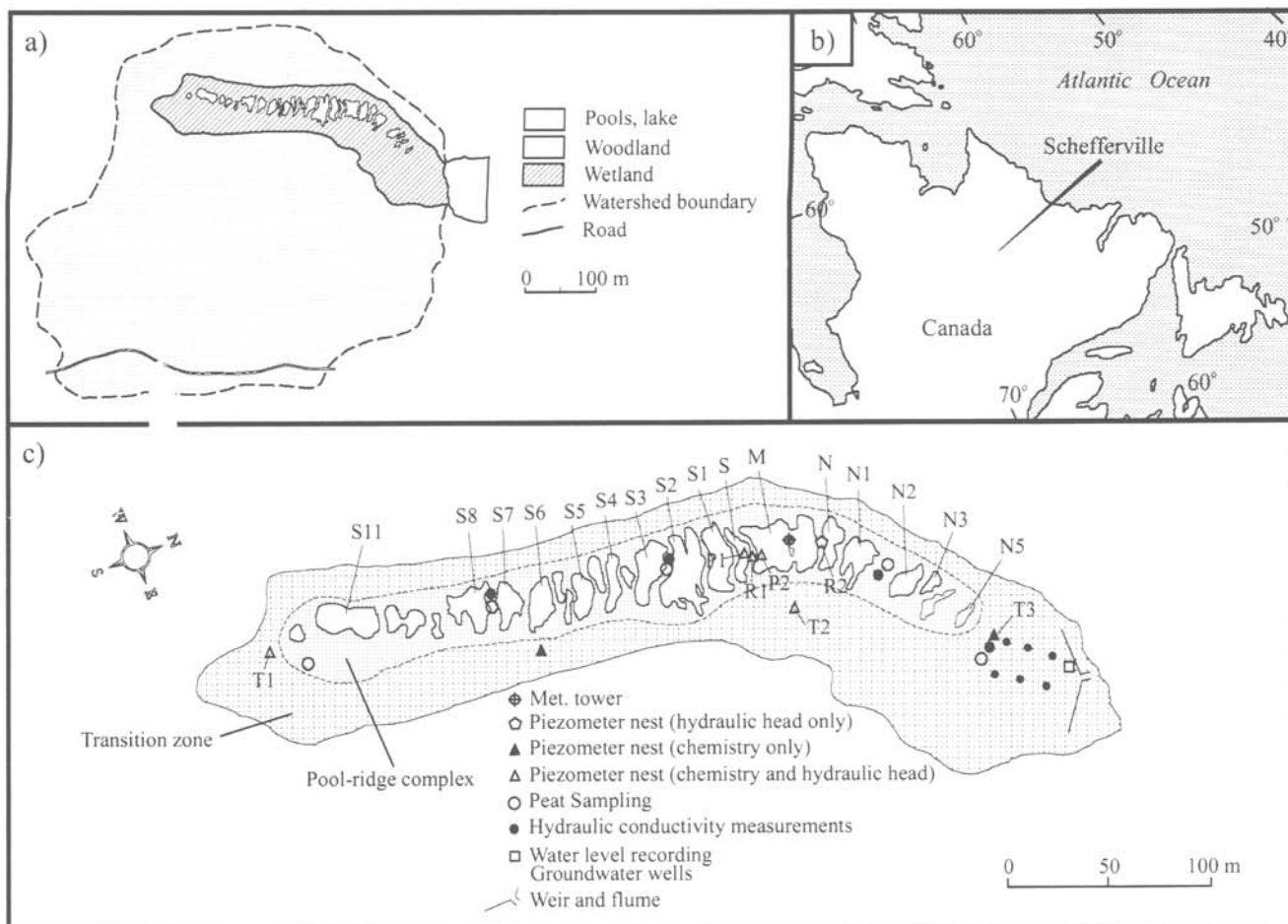


FIGURE 1. The study basin (a), its regional setting (b), and the location of instrumentation and sampling locations within the wetland (c). The major pools with staff gauges are identified. The wetland is unofficially named Pelletier Wetland.

The Schefferville area comprises 25 to 30% wetlands, and is located on the boundary of the Atlantic boreal and high subarctic wetland regions (National Wetland Working Group, 1988). Schefferville is in the zone of widespread discontinuous permafrost, but permafrost is absent in the valley bottoms where the wetlands are found.

Open and closed lichen woodland underlain by shallow Dystric Brunisol (<30 cm depth) occupy 88% (21.3 ha) of the study basin. The remaining 12.5% (3 ha) is occupied by an elongated patterned wetland, consisting of a central pool-ridge complex of 16 major pools, and a transition zone between the hill-slope base and the pool-ridge complex (Fig. 1c). The surface areas of the land and pool portions of the wetland relative to the area of the basin are 10% (2.4 ha) and 2.5% (0.6 ha), respectively. By late summer, the mean pool depth was ~0.3 m, and the average elevation of the ridges above their adjacent pool surfaces was ~0.1 m.

Twelve principal species were noted to occur within the wetland: black spruce (*Picea mariana*), blueberry (*Vaccinium uliginosum*), bog bean (*Menyanthes trifolium*), bog laurel (*Kalmia polifolia*), bog lily (*Smilacina trifolia*), labrador tea (*Ledum groenlandicum*), leatherleaf (*Chamaedaphne calyculata*), dwarf birch (*Betula michaux*), mosses (*Sphagnum* spp.), salmonberry (*Rubus chamaemorus*), sedges (*Cyperaceae* spp.), and tamarack (*Larix laricina*) (Pelletier, 1977). The sedges, leatherleaf, tamarack, and mosses occupied the ridge here as bog bean, sedges, and mosses dominated near the pools and in wet areas.

In the longitudinal direction, the wetland slope is 0.004. Elevation surveys were conducted across the wetland in several locations. The wetland surface is slightly concave in cross-section, with the central pool-ridge complex at an elevation ~0.15 m below the surface of the surrounding transition zone. Peat depths were measured at 19 ridge crests over a transect that traversed the long axis of the wetland. Average peat depth is 1.30 m; the minimum and maximum depths are 0.64 to 2.00 m. The basal material is iron-rich clay. The peat at the bottom of the pools is unconsolidated and loose for the first 0.2 m, but becomes more consolidated with depth.

Methods

Figure 1c identifies the major physical components of the wetland under study and the locations of instrumentation. Total water equivalent of basin snow cover was estimated from late winter snow surveys. To isolate areas with common snowpack characteristics, the basin was divided into its three major biophysical components: the pool-ridge complex, the transition zone, and the surrounding wooded upland (Fig. 1a, c). Daily snowmelt was measured in each terrain component using the procedure outlined in Heron and Woo (1978). Rainfall was measured with 12 bulk rain gauges distributed throughout the wetland. The gauges were checked following each rain event, and the average rainfall depth was computed.

The 16 major pools (denoted with letters, Fig. 1c) each

contained a staff gauge of known elevation above the basin outlet. From daily staff-gauge readings, a water-level record was compiled for each pool. During periods of high pool stage, daily observations were made to determine if there was surface flow between adjacent pools. During this procedure, the principal surface flowpaths between connected pools (e.g. over ridge crests, around ridge flanks) were noted.

A weir conveyed surface flow through a flume box at the basin outlet (Fig. 1c). The basin discharge hydrograph was computed from daily measurements of discharge through the flume and continuous stage readings. Flow velocity was measured with an Ott current meter, and discharge was computed from the velocity-area method. Stage was monitored with a chart recorder mounted on a stilling well. A total of 98 stage-discharge measurements were made, from which several rating curves were defined ($r^2 = 0.78$ to 0.99). Each rating curve related to a particular period of the overall flow record.

Subsurface discharge from the wetland was computed using the Dupuit-Forcheimer method (Freeze and Cherry, 1979). A network of six groundwater wells were installed in the transition zone between the flume box and the pool closest to the outlet (Fig. 1c). Water levels and frost depths were measured manually each day. The hydraulic conductivity was estimated from the Hvorslev (1951) method (Freeze and Cherry, 1979). This was applied at four locations in the wetland, including a location close to the network of groundwater wells (T3) near the basin outlet (Fig. 1c).

The daily evaporative loss from pools N and M was obtained by measuring the volume of water required to restore the water level to a fixed mark on floating transparent polyethylene lysimeters. The accuracy of the technique was hampered when windy conditions caused splash and spillage. Data was rejected for days when there was more than a 20% discrepancy among the 12 lysimeter readings (Roulet and Woo, 1986). For days yielding acceptable data, an average measured evaporation value was computed. The near-surface water temperature of the lysimeters and of the adjacent pool water was recorded throughout the study period in order to check that the lysimeter water was thermally representative of the pool water.

A 5-m meteorological tower was erected in pool M on the surface of what appeared to be a degraded ridge, that had become submerged over much of its length. Net radiation (Q^*), relative humidity, air temperature, and wind speed were measured at 1.5 m above the surface. All variables were sensed every minute, and averaged each hour using a Campbell Scientific 21X datalogger. Measurements were assumed to be representative of both pool and ridge surfaces.

The average daily evaporation measurements were regressed with estimates of equilibrium evaporation calculated for the same diurnal periods using the Priestley-Taylor (1972) approach

$$E = \alpha [\Delta(Q^* - Q_G) / \rho_w L_v (\Delta + \gamma)]$$

where E is the evaporation rate (mm d^{-1}), α is an empirical coefficient equal to the ratio between the measured (Q_e) and equilibrium evaporation (Q_{EQ}) rates (i.e. $\alpha = Q_e/Q_{EQ}$), Δ is the slope of the saturation vapor pressure curve, Q_G is the ground heat flux, ρ_w is the density of water, L_v is the latent heat of vaporization, and γ is the psychrometric constant. Δ was calculated using air temperature and relative humidity data from the meteorological tower using coefficients supplied by Buck (1981). The coefficient α was then used in the above equation for days when direct (i.e. lysimeter) measurements were not made.

The ground heat flux (Q_G) was measured directly with a heat flux plate inserted 5 cm below the surface of the ridge where the meteorological tower was located. The pool heat storage was computed for each day from

$$Q_w = \rho_w C_w Z_M \Delta T_L / \Delta t$$

where $T_L/\Delta t$ is the average daily temperature change of pools N and M. T_L was measured using a thermistor probe. Z_M is the mean depth of the two pools.

Three indirect methods were used to evaluate the hydrological flux between the pools and underlying groundwater: water balance calculations for all pools, chemical analysis of pool and peat-pore water, and calculations of groundwater flow from measured hydraulic gradients and conductivities.

Pool water balances were computed from the difference between the observed pool water level and the value calculated from the difference between precipitation and evaporation for the period while the pools were hydrologically disconnected (24 May to 18 July). Electrical conductance, pH, and concentrations of calcium (Ca) and magnesium (Mg) were measured on water sampled weekly at seven piezometer nests within the wetland (Fig. 1c). Three piezometer nests were located in the transition zone (T1, T2, and T3), two on ridge crests (R1 and R2), one in pool S (P1), and one in pool M (P2). Variations in Ca and Mg concentrations were used to infer the direction of groundwater flow. If groundwater was moving toward the surface, we assumed that the chemical composition of the migrating water would reflect that of the underlying sediment, which is known to be rich in Ca and Mg.

At the six piezometer nests where hydraulic head was measured (Fig. 1c), weekly measurements were used to compute hydraulic gradients, and the specific groundwater discharge (Q_z) is estimated from Darcy's Law with the assumption that the vertical hydraulic conductivity is 0.1 of the measured (horizontal) conductivity (Boelter, 1965). Hydraulic heads measured at P1, P2, and the intervening ridge R1 were used to construct weekly flownets. These piezometer nests were ideal for flownet construction since they provided a cross section of the subsurface flow field beneath a ridge. Hydraulic head measurements were used to plot equipotential lines from which hydraulic gradients were estimated. Ten separate flownets were constructed—one for each week between 24 May and 18 August. For each flownet, an average hydraulic gradient (dh/dz) was calculated, and the daily specific groundwater flux (Q_z) was computed from the hydraulic gradients and the hydraulic conductivity.

Results

ATMOSPHERIC INPUTS

The cumulative snowmelt input to each land component and the rainfall input are shown in Figure 2. The areally-weighted basin snow storage was 268 mm. For wetland and woodland portions of the basin, the snow water equivalents were 243 and 272 mm, respectively. The rate of snowmelt accelerated following 28 April. By 13 May, less than 10% of the pool-ridge complex was snow covered, while snow persisted on approximately 95% of the woodland surface. On the final day of ablation measurements, the snowcover occupied less than 5% of the pool-ridge complex (16 May), ~30% at the transition zone (19 May), and ~25% in the woodland (21 May). The 34 rainfall events recorded over the study period contributed a total of 211 mm. The rainfall event of 19 July (59 mm) was nearly three times larger than the next largest event.

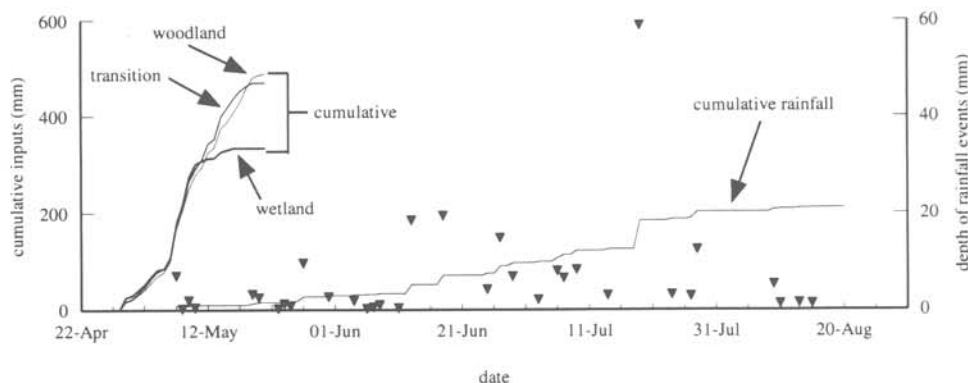


FIGURE 2. Cumulative inputs from snowmelt to the major biophysical basin components. The rainfall events (black triangles) and the cumulative rainfall input are also shown.

EVAPORATION

Forty days of acceptable evaporation measurements remained after rejecting the days with a >20% discrepancy among lysimeter readings. For these days, the cumulative evaporative flux was 181 mm (~ 4.5 mm d⁻¹), and the equilibrium flux was 108 mm. The Priestley-Taylor alpha value computed for the same period was 1.67 ± 0.60 (Quinton, 1991). The near-surface water temperature of the lysimeters closely matched that of the pools ($r^2 = 0.966$, slope = 1.037, y-intercept = 0.382, $N = 73$) suggesting that errors in pool evaporation measurements were not likely to have arisen from differences between the lysimeter and pool thermal regimes. The value of α for the pools is large compared with the value for the adjacent non-pool surfaces (i.e. ridge crests and transition zone) of 1.34 (Quinton, 1991), suggesting that there might be considerable advective enhancement.

CONNECTIVITY SURVEY

The pool stage level record began on 10 May. At that time the snowmelt water supply was sufficient to maintain hydrological connection among all 16 pools. This lasted until 21 May when 5 pools became disconnected. By 23 May, 12 pools were disconnected and by 25 May, there was no surface flow between any pools.

Flow between pools occurred in two ways: over ridge crests (diffuse flow through the vegetation layer), and around ridge flanks. In many cases, flow around flanks produced flooded areas which extended outward, beyond the pool-ridge complex into the transition zone. Many pools merged on both eastern and western flanks, although flooding was more common on the east side. The flooded pool extensions were connected by a system of poorly defined channels flowing toward the basin outlet along the margin between the pool-ridge complex and the transition zone. The large rain event of 19 July enabled the re-connection of seven pools with one or both of their adjacent pools. The water level record of each pool is shown in Figure 3.

POOL STAGE RECESSIONS

An analysis was conducted to determine if the rate of recession from a rise in pool stage (i.e. stage-recession) is affected by the pool being in a connected or disconnected condition. The stage-recession limbs of eight major peaks in pool water level were examined (Fig. 3). Stage-recessions 1, 2, and 7 were at least initially associated with the connected-transitional hydrologic phase. Stage-recessions 3, 4, 5, 6, and 8 occurred while the pools were disconnected. The stage-recession limbs of an arbitrarily chosen series of pools (S2 to N2) are plotted on semi-log paper, with the stage-recessions of the two phases shown

separately (Fig. 4). The stage-recessions of the disconnected phase are shown to have the greater average slope value.

The stage-recession coefficient (b) (Quinton, 1991) of the 8 major storm events was calculated for all 16 pools. An ANOVA showed that at each pool, b was significantly different over time ($P = 0.0001$). Storm 8 was omitted from the ANOVA since it did not produce a hydrograph rise at all pools. A multiple comparison analysis was necessary to categorize the storm events which produce a similar range of stage-recession coefficients. This was achieved using Duncan's Multiple Range Test which individually compares the means of each event with the means of the other six events (Table 1). Table 1 shows that the seven events are classified into 2 groups: Type 1 (events 3, 4, 6), and Type 2 (events 1, 2, 5, 7). With the exception of event 5, the Type 2 category produced by the multiple range test contains the storms of the connected phase. The Type 1 category contains the events that occurred while the pools were disconnected. The fifth water level rise was generated by two events in close succession. As a result, although event 5 occurred during the disconnected phase, the stage-recessions from this event were protracted, and therefore resembled the Type 2 (i.e. connected phase) stage-recessions. The ranking of the events in order of decreasing average values of b (Table 1), shows that among the events of the type 2 category, event 5 is closest to the Type 1 (i.e. disconnected phase) values.

GROUNDWATER INTERACTION

Budgets

In computing pool water balances, it was assumed that the pools were hydrologically isolated from one another once the surface flow between them ended, because of the very low horizontal hydraulic conductivity (10^{-2} to 10^{-1} m d⁻¹) of the intervening ridges and the shallow slope of the wetland (0.004). This assumption is supported by the pool water level record (Fig. 3), which shows that the rate of water loss was roughly equal for all pools. If the pools were not isolated, those at the southern end of the wetland would have drained at a higher rate than those downstream.

A water balance for the wetland pools was computed using precipitation, evaporation, and the average change in storage of the 16 pools. The residual was assumed to be an estimate of the vertical exchange between the wetland and groundwater. Since the 59 mm storm of 19 July caused several pools to merge, the water balance was calculated for the period between 24 May and 18 July. The difference between pool evaporation (-210 mm) and precipitation (109 mm) was -101 mm, while the mean observed pool water loss was only 69 mm. The difference of 32 mm is attributed to the net groundwater input. For the 54-d pe-

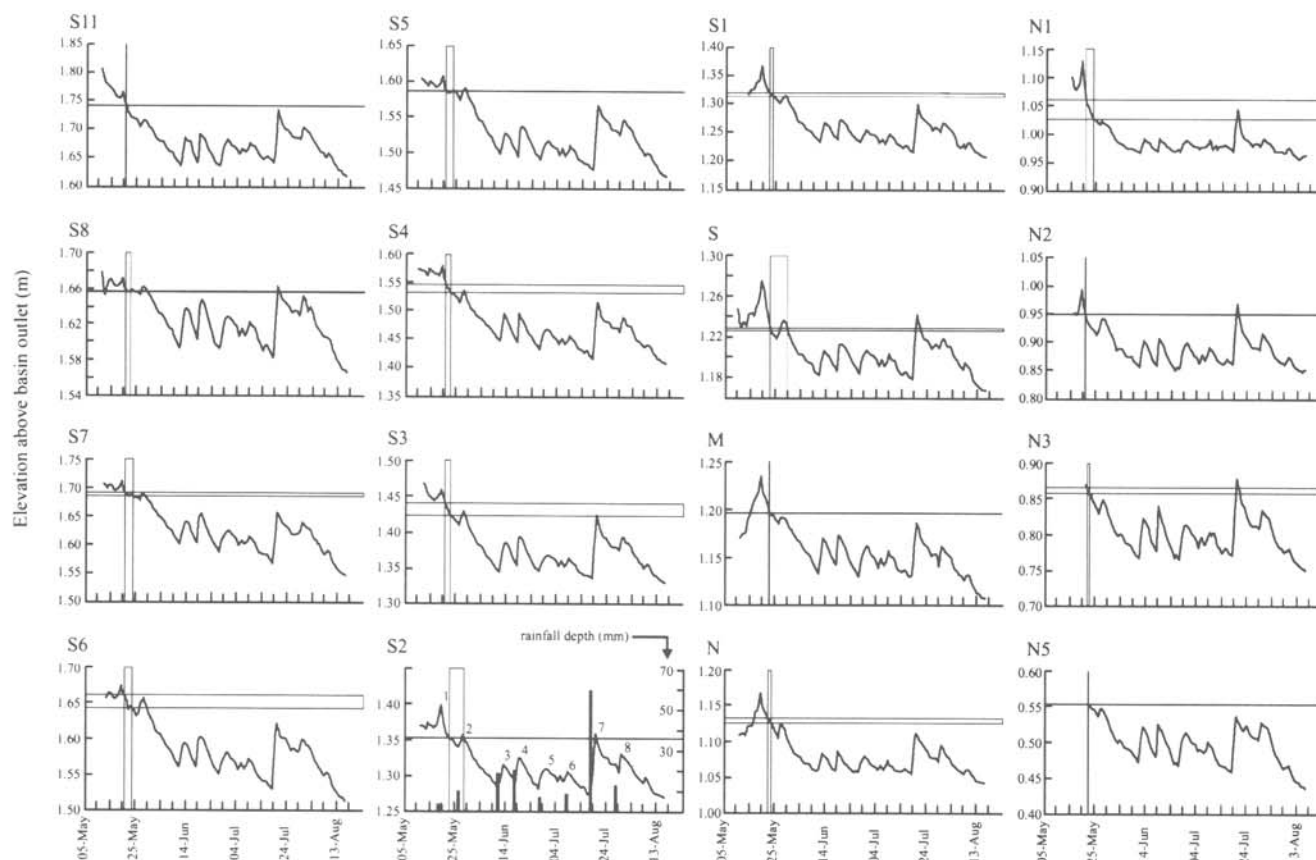


FIGURE 3. The water level record at each pool. The horizontal bars separate the stage levels of the connected and disconnected phases and indicate the range of water levels between them. The vertical bars separate the connected and disconnected phases, and indicate the duration of the transition period. Single horizontal and vertical lines are shown for pools where disconnection with the upstream and downstream pool occurred on the same date. The eight major rainfall events and the resulting water level rises are identified in Pool S2.

riod of pool disconnection, the estimated average groundwater input was therefore 0.6 mm d^{-1} .

Hydraulic Analysis

In calculating Q_z , it was assumed that the vertical hydraulic conductivity of $6.69 \times 10^{-3} \text{ m d}^{-1}$ at R1 (estimated as 0.1 of the horizontal hydraulic conductivity of the peat at 0.75 m depth) was representative of the other five sites as well. The spatial variation of Q_z was greatest at the beginning of the season and decreased steadily as the season progressed (Fig. 5). Three flow-nets, constructed from measurements taken at P1, P2, and R1, representing the early, middle and late study period are shown in Figure 6. Over the study period, total Q_z was 131.4 mm, or 1.88 mm d^{-1} . This value is likely a high estimate for the wetland, since Figure 5 shows that the wetland includes both groundwater recharge and discharge zones, and that R1 has a relatively large upward Q_z flux.

Groundwater Chemistry

Ion concentrations of Ca and Mg, electrical conductance, and pH values increased with time over the sampling period. An analysis of variance (ANOVA) showed that the values of the chemical variables showed no statistical difference among the sampling sites. Therefore the data of just one site (T3) is presented.

Figure 7 illustrates the change in groundwater chemistry

with depth over the 10-wk period of sampling from T3. Figure 7a shows that in late June, electrical conductance was highest at 1.75 m ($97.7 \mu\text{S}$). By mid-July, electrical conductance was highest at 0.25 m ($49.0 \mu\text{S}$). The peaks in electrical conductance move steadily to shallower depths, suggesting an upward migration of water bearing a higher electrical charge, hence groundwater discharge.

The concentrations of Ca and Mg are shown in Figure 7b. The contribution of Ca and Mg to the electrical conductance was found to be greatest near the bottom of the profile, where the values of electrical conductance were higher (Quinton, 1991). Similarities exist between Figure 7a and 7b in terms of the timing of fluctuations. However, the differences are sufficient to suggest that other ions contribute to electrical conductance. This is especially evident closest to the surface (0.25 m depth), where in spite of only minimal fluctuations in Ca and Mg, electrical conductance rose steadily throughout the season.

Figure 7c shows that the variation in pH with time decreases with depth. Near the surface, pH values increased over the sampling period; while at depth, the reverse occurred. The hydrogen ion activities calculated from the pH values showed that [H] makes a relatively small contribution to the overall electrical charge (Quinton, 1991).

RUNOFF

Basin discharge (Fig. 8) began on 6 May, 8 d following the start of snowmelt (28 April). Discharge peaked at 55 L s^{-1} , 3 d

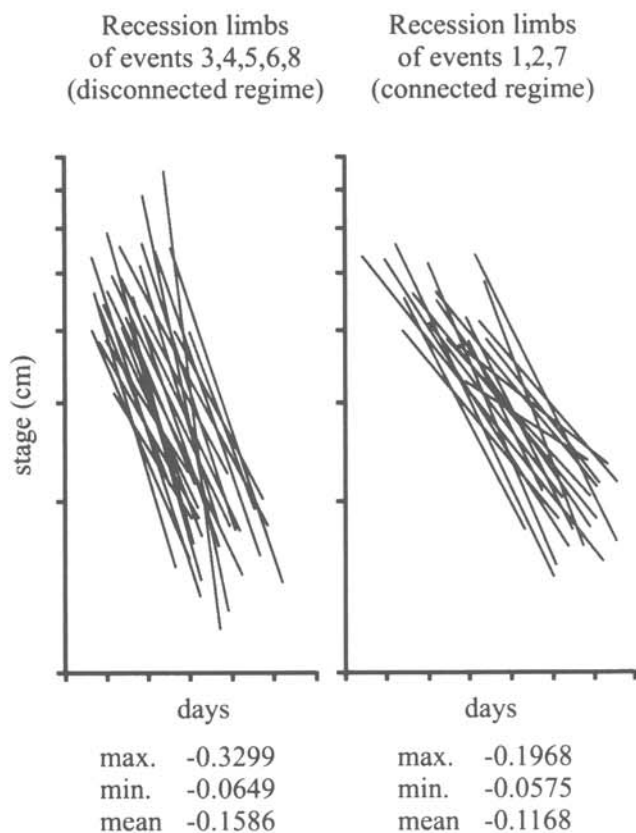


FIGURE 4. A comparison of the stage-recession limb slopes of the connected and disconnected phases. The descriptive statistics of their slopes are provided for each phase.

later (9 May). The snowmelt portion of the hydrograph was characterized by four peaks (periods of increased meltwater input). Although a high level of discharge continued until 24 May, the hydrograph was in recession by 20 May, when the wetland pools began to hydrologically disconnect. Summer discharge occurred at a lower level, rising periodically in response to precipitation. On the final day of study (17 August), discharge was 0.16 L s^{-1} . The cumulative basin surface discharge for the study period was $47,795 \text{ m}^3$ (196 mm of runoff). Total subsurface runoff over the same period was approximately 4.1 m^3 ($\sim 0.02 \text{ mm}$ of runoff).

Discussion and Conclusions

Several factors contribute to the temporal trend of runoff from a basin dominated by a patterned wetland. Location of the

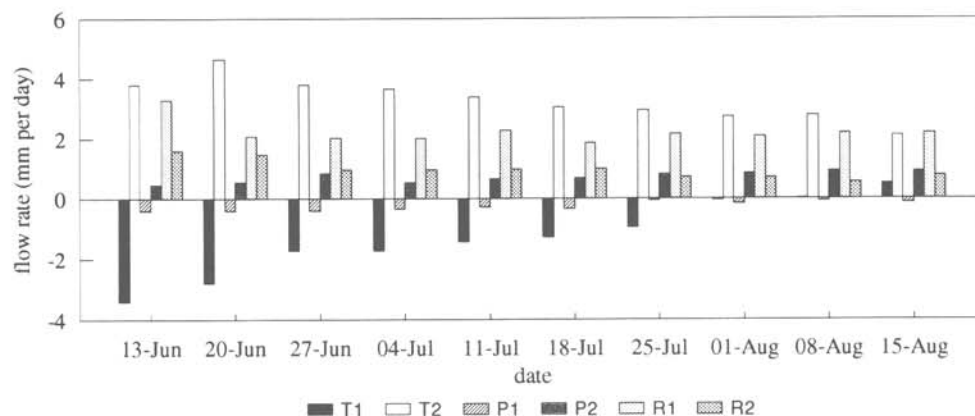


FIGURE 5. Spatial and temporal variations in the vertical groundwater flux (Q_z) at the six piezometer nests.

TABLE 1

Duncan's Multiple Range Test for stage-recession coefficients^{a,b}

Event	N	b (mean)	Critical value	Type
4	16	0.007191	0.001454	1
3	16	0.006983	0.001409	1
6	16	0.006249	0.001515	1
5	16	0.004726	0.001487	2
2	16	0.004245	0.00134	2
7	16	0.004229	0.001538	2
1	14	0.003478		2

^a The events are ranked in order of decreasing mean value of the stage-recession coefficient (b). N is the number of stage-recessions included in the analysis for each event (equal to the number of pools where stage-recessions were recorded).

^b Alpha = 0.05, degrees of freedom = 103.

wetland is important: much of the hydrological impact of the wetland is due its location at the outlet of the basin. All water exiting the basin is therefore routed through the wetland, and as a consequence the storage-runoff relationship for the wetland controls the basin discharge. The wetland being of the patterned form exerts a further influence above and beyond what would be expected from some other wetland forms. The pool-ridge sequence creates, at times of lower water levels in the pools, a large storage capacity. There has been considerable study of these landforms because of their wide spread occurrence in northern Canada (National Wetlands Working Group, 1988), and northern Eurasia, but why they form over other peatlands is still not clear (Foster et al., 1983). It is clear that hydrology plays an important role (Swanson and Grigal, 1988) and that the orientation of ridges and pools is always perpendicular to the direction of water flow, i.e. the hydraulic gradient.

That wetlands reduce runoff, has been a popularly held misconception. Bay (1969) and others (e.g. Roulet and Woo, 1986; Verry et al., 1988) have demonstrated that like any other hydrological system, runoff from wetlands is dependent on moisture conditions, and that wetlands have very little available storage capacity because of the large volume of stored water. The main hydrological function of wetlands comes from the lack of gradient, resistance to surface flow and the amount of depression or detention storage that results from the surface topography of a wetland (Verry et al., 1988; Winter and Woo, 1990). The ability of a patterned wetland to reduce and delay runoff, and thus to attenuate peak flows, depends on the connectivity of the pools, which in turn is a function of the amount of water stored in the pools relative to their storage capacity.

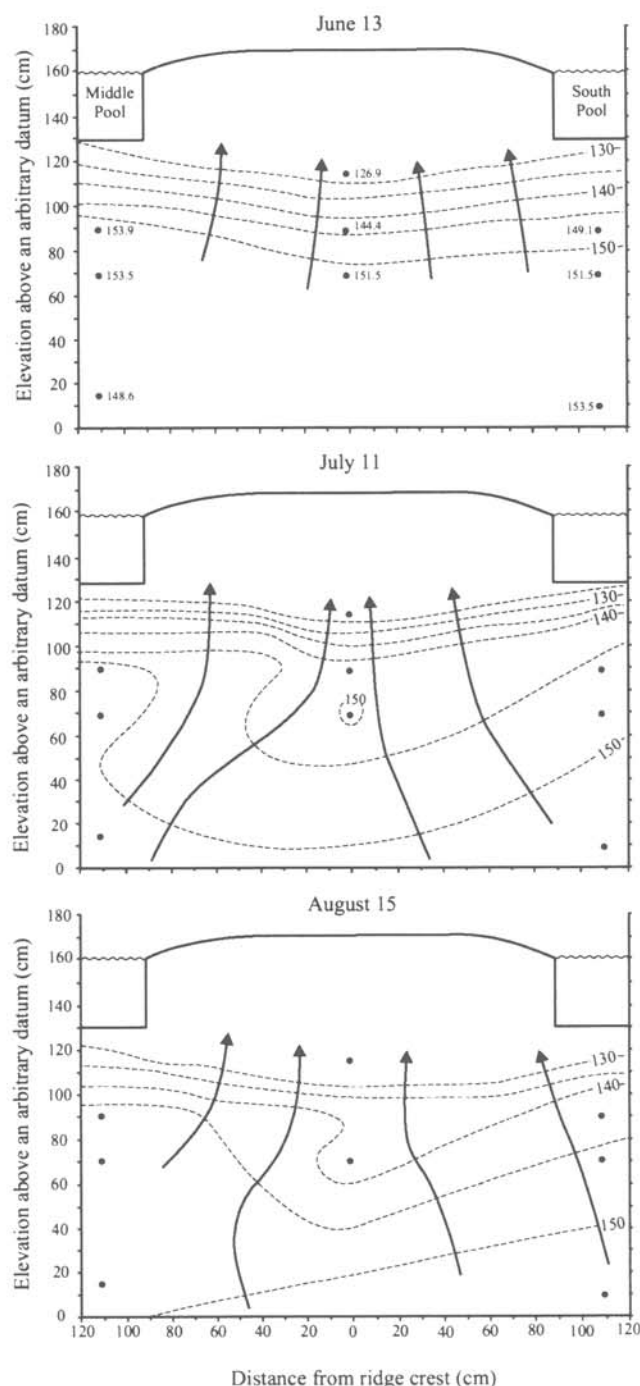


FIGURE 6. Hydraulic head distribution and flownets determined for a piezometer nest (R1) for the early, middle, and late summer periods.

When the wetland is in the hydrologically connected phase (Fig. 9a) each pool receives surface input from upstream and laterally from the transition. Direct precipitation and groundwater can also contribute to the pools. Output from a pool during the connected phase includes surface runoff to the next downstream pool and evaporation. During the disconnected phase (Fig. 9b), a pool receives input from precipitation and groundwater and loses water to evaporation. Under this condition, the wetland pools have only a storage function: all water entering a pool contributes to storage and evaporation, and no surface runoff is generated. When the pools were disconnected, loss of water to evaporation far exceeded the small inputs from ground-

water flow. Therefore, the amount of available storage in the pools increased with time following the hydrological disconnection of pools. The basin runoff response to rain events was therefore smaller during the disconnected phase, and as time progressed, increasingly larger rain events were required to generate flow among the pools.

There is a marked difference in the rate of recession of pool water levels of the two phases, owing to a change in the balance between hydrological inputs and outputs. During the disconnected phase, recessions are relatively steep since groundwater input, the only inflow, is relatively small ($\sim 0.5 \text{ mm d}^{-1}$) compared to evaporation ($\sim 5 \text{ mm d}^{-1}$), the only outflow. If during the connected phase, runoff input = runoff output, as is the case during the disconnected phase (i.e. when both are zero), then the recessions of the two phases would have the same slope, since both would be controlled by the difference between groundwater input and evaporation loss. However, since the recessions of the connected phase are shallower, it appears that the net effect of surface flows is that the rate of water loss from the pools is offset. This would only occur if runoff inflows exceed the runoff outflows. The difference between runoff inputs and outputs would have to be small, since the overall water balance of the pools is negative, but large enough to reduce the rate of recession.

If it had not been for the groundwater input, the recession in pool water levels during the disconnected phase would have been even steeper. The patterns of hydraulic head and the Ca and Mg profiles indicate a small but significant input of groundwater to the pools. The input is larger early in the year and then decreases, indicating the groundwater input is of local origin, i.e. contained within the topographically defined basin (see Toth, 1963; Freeze and Witherspoon, 1967). Surface water-groundwater connections in peatlands was not commonly reported in the literature until recently. This is likely the result of some researchers relying on the water balance to compute groundwater as a residual, while the magnitude of the groundwater flux, being secondary to precipitation, surface flow, evaporation, and changes in storage, would have been well within the errors of most water balances. However, direct measurement of the pattern of hydraulic heads in and around many peatlands clearly shows groundwater interaction is common (e.g. Siegel and Glaser, 1987; Roulet, 1990).

The hydrological phase of a patterned wetland determines the amount of runoff generated seasonally and for single storm events. The flow system was disconnected 81% of the 84-d study period (6 May–17 August). The wetland was hydrologically connected for only 14% of the time: 14 d during the snowmelt runoff period, plus a ~ 1 -d period in response to the 59 mm rain event of 19 July that was sufficiently large to move many of the wetland pools back into the connected phase. For approximately 5% of the time (5 d) the wetland was in transition between the two phases. However, the amount of runoff attributable to each phase is almost the reverse of the duration of each phase. Approximately 77% of the runoff observed during the study period occurred during the connected phase. The remaining 23% was contributed by drainage from the pool-ridge complex during the transition phase, and by drainage from the transition zone while the pool-ridge complex was in the disconnected phase. Rain events of a lesser magnitude than the event of 19 July, did not generate any measurable discharge when the hydrological system was disconnected, but much smaller rain events did produce a response during the connected phase (e.g. the rain events of 19 and 20 May that produced the first of the eight large pool water level rises shown in Fig. 3).

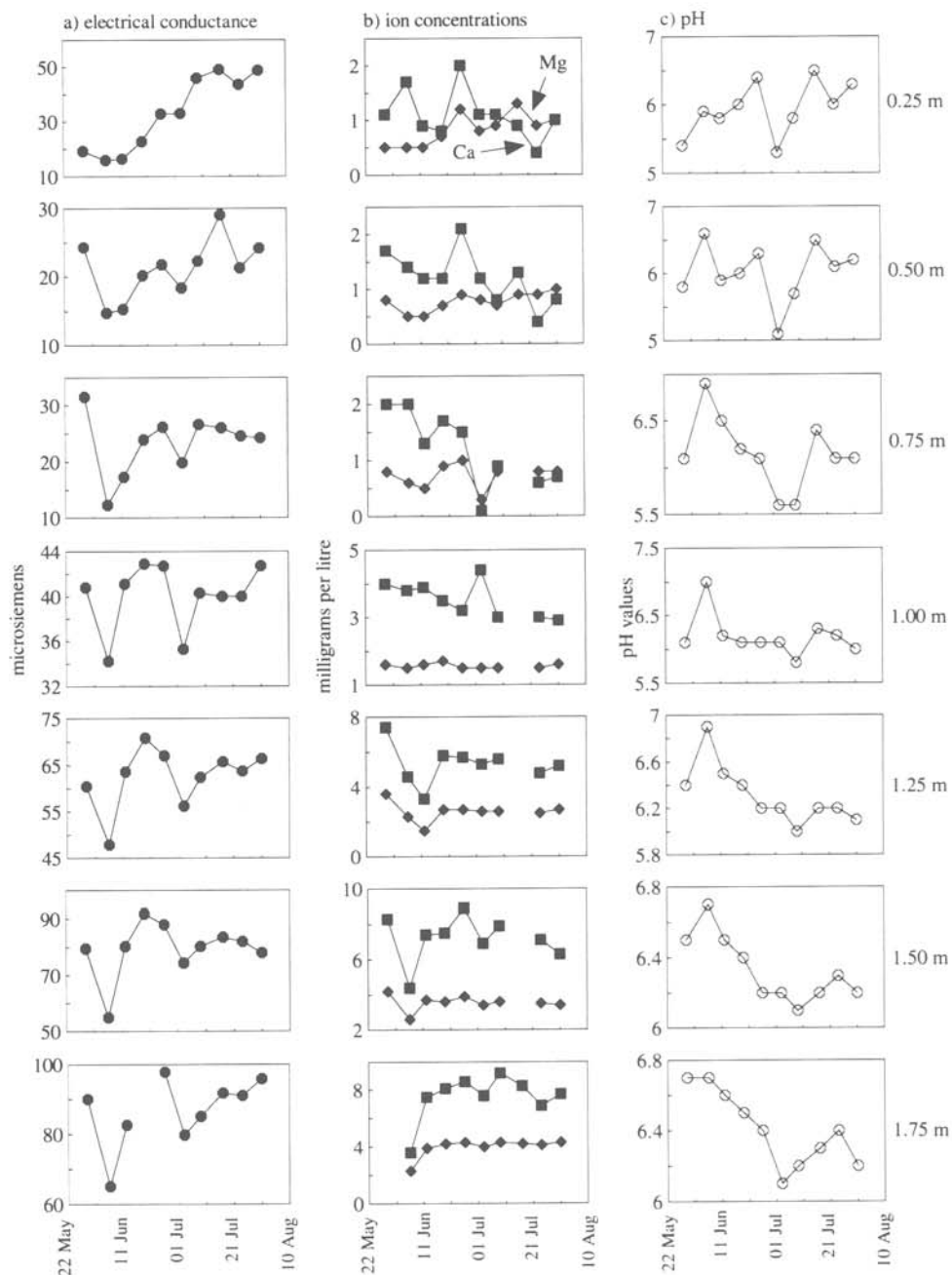


FIGURE 7. Variations in groundwater chemistry with depth over the 10-wk period of sampling at the location closest to the basin outlet (T3).

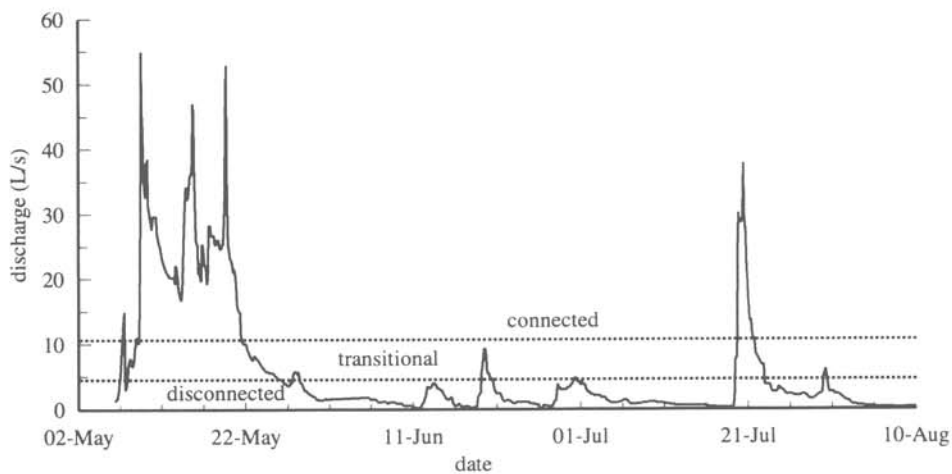


FIGURE 8. The basin discharge hydrograph. The horizontal lines relate the discharge hydrograph to the hydrological phase of the wetland pools.

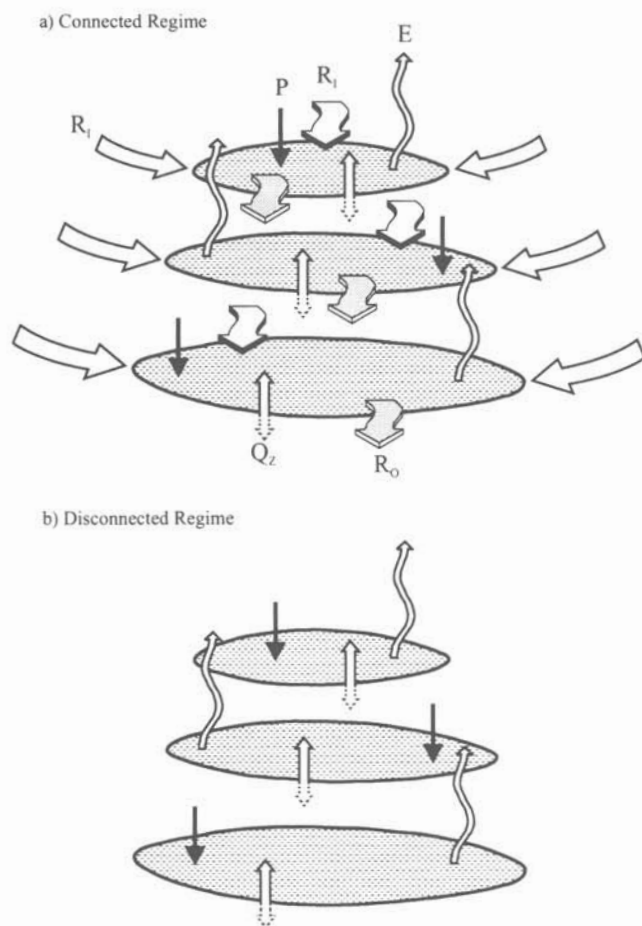


FIGURE 9. An illustration of the difference between the connected and disconnected hydrological phases in terms of the relationship between the flux and storage of water. The fluxes include surface input from the upstream pool and the transition zone (R_1), direct precipitation (P), groundwater input (Q_z), surface runoff to the downstream pool (R_0), and evaporation (E).

Branfireun and Roulet (1998) reported two different hydrological responses for boreal bogs. They identified a low flow regime when storm runoff response was isolated to the wetland itself, and a high flow regime when the wetland not only generated flow *in situ*, but also received a large amount of water from the adjacent upland hillslopes. In their study, they refer to the connection or disconnection occurring on the scale of wetland-hillslope interaction. This scale of connectedness is also observable in the present study. The cumulative discharge for the connected phase of spring runoff ($32,307 \text{ m}^3$) is approximately five times larger than the runoff that could have been produced (6678 m^3) by the difference between the late winter snow storage on the wetland (7290 m^3 calculated as the depth of water equivalent times the area of the wetland) and the capacity for storage of water in the pools just before spring melt (612 m^3 calculated from the product of the average late-summer depression storage capacity of the pools, and the total area of the pools within the wetland). This indicates that approximately 80% of the water discharged during snowmelt came from the upland portion of the basin. The same analysis can be done for the basin response to the rain event of 19 July. Cumulative basin discharge in response to this event was 4533 m^3 , while the maximum possible response from direct precipitation on the wetland, assuming that the wetland pools were all completely filled with

water, was 1770 m^3 (wetland area times the depth of rainfall). Hence, at least 61% of the storm runoff originated from beyond the perimeter of the wetland. The value of 61% is a minimum estimate of the upland contribution to basin discharge, because the pools were not full prior to the rain event of 19 July (Fig. 3), and a certain amount of input water was required to bring the pools to this condition before runoff from the pool-ridge complex could proceed. The pool storage records also suggest the occurrence of substantial lateral input from outside the pool-ridge complex, since the average water level rise of the 16 pools in response to the 59 mm event of 19 July was 83 mm.

In conclusion, the hydrology of this patterned wetland is similar in many respects to that observed in nonpatterned, peat accumulating wetlands. The major difference is that the storage available within a patterned wetland is larger than the depression storage of subarctic fens (Winter and Woo, 1990), and the storage attributed to the depressions created in the hummock—hol-low topography of a bog (Verry et al., 1988). Much of the runoff from this patterned wetland was in response not to runoff generated *in situ* but to the input of water from adjacent uplands. This is similar to the observation of Branfireun and Roulet (1998): when the wetland is hydrologically coupled to the upland, basin discharge is large. Because the pool-ridge topography of a patterned wetland is a result of how the wetland develops through differential peat accumulation, it follows that the storage-runoff relationships found in the present study are a function of the stage of development of this particular wetland. However, the results should be relevant to most patterned wetlands once the developmental history of the wetland and the hydrological setting of the wetland in its basin are accounted for.

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