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Blanket bog in Newfoundland. Part 2. Hydrological processes

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ABSTRACT

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The water balance was calculated for a 9.4 ha blanket bog basin between 5 and 28 July 1989, and between 16 May and 24 June 1990. Rainfall for these periods was 174 mm and 231 mm, respectively, and fog deposition added 31 mm and 23 mm of water to the inputs, respectively. Runoff was the largest loss, accounting for 70% and 63% of inputs in 1989 and 1990, respectively, compared with 24% and 28% for evaporation, and 6% for groundwater seepage. Fog simultaneously enhances the inputs and reduces the evaporative loss. Evaporation occurs at the potential rate for a very short period following fog, but decreases quickly as the surface dries, making it difficult to predict. Pipe-flow and high near-stream gradients, coupled with the high transmissivity of the elevated water table, produce a flashy hydrograph. The maritime climate maintains a high water table, despite relatively steep gradients, which is essential to the development of blanket bog systems in Newfoundland.

INTRODUCTION

Blanket bogs are extensive peat deposits that occur more or less uniformly over gently sloping hills and valleys (Environment Canada, 1987). Their occurrence is restricted to cool oceanic locales, where climate is apparently more important than topography in limiting their spatial extent. Although commonly occuring in Ireland, Scotland and Wales (Moore, 1982), they occur sporadically in North America along the Pacific coasts of British Columbia and Alaska. In eastern North America they are restricted to the Burin and Avalon peninsulas of Newfoundland (Davis, 1984).

Since blanket bog formation is at least partly independent of topography, their range is sensitive to the climatic conditions which ensure saturation at the surface. This is necessary to limit the rate of decay of organic material requisite for peat development. Total annual precipitation alone is not an

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adequate indicator, since the south coast of insular Newfoundland, for example, receives 200–400 mm more than the Burin and Avalon peninsulas (Unpublished maps, Newfoundland and Labrador Department of Environment, 1987), yet is without blanket bog. However, the blanket bog zone of Newfoundland corresponds to the foggiest area in Canada (Hare, 1952; Joe, 1985, cited in Barrie and Schemenauer, 1986), indicating a more complex relationship with the climate.

The objective of this paper is to characterize the hydrological processes operating on a Newfoundland blanket bog, and to better understand its relationship with the climate that sustains it. The results have possible commercial implications for a nearby fuel peat mining operation at St. Shotts.

HYDROLOGICAL PROCESSES

Previous hydrological research on blanket bog has most often been associated with drainage operations (Burke, 1972; Mulqueen, 1986), including bogs at St. Shotts, Newfoundland (Northland Associates Ltd., 1989). Few comprehensive hydrological studies of undrained blanket bog exist. However, some data are available on streamflow and water-table fluctuations.

An undrained Irish blanket bog (Burke, 1972) experienced water-table fluctuations ranging from 5 to 40 cm below the surface. The water table at a virgin bog at St. Shotts, Newfoundland, varied between 2 and 11 cm between June and September 1988, compared with values of 5 and 19 cm at a slope bog/domed bog complex at Markland, 20 km north of the Atlantic Oceanic Wetland Zone (Environment Canada, 1986), which corresponds to the zone of blanket bog occurrence. The high water table on the blanket bog is indicative of the different climatic and hydrological regime. In a study of the runoff from a Newfoundland basin with blanket bog, Barnes (1984) noted that runoff from the peatlands was sustained through the summer, which is consistent with the relatively high water-table position noted at St. Shotts. Burke (1972) found that Irish blanket bogs have a flashy hydrograph, and that flow in summer is discontinuous, requiring heavy rainfall to produce runoff. This is not unlike the continental bogs of Minnesota reported by Bay (1969).

Fog has a double role in the hydrological regime. Price (1992, Part 1), reported that fog is advected over the bogs during warm onshore winds in summer. The predominant southwesterly air flows during this season, passing over frigid offshore water, produced fog on 40% of all half-hourly measurement periods between mid-May and mid-July 1989. Fog-water deposition between late June to early August averaged 1.8 mm day^{-1}. Furthermore, Price (Part 1) showed that the average evaporation rate during daytime foggy periods was 1.1 mm day^{-1} compared with 2.5 mm day^{-1} during equivalent periods with no fog. Evaporation from wetlands in general is poorly understood (Lafleur, 1990), and this is especially true for bogs, which are dominated by non-vascular *Sphagnum* spp. Ingram (1983), in summarizing chiefly European studies, suggests that because of the loose pore structure of *Sphagnum* moss, there is little opportunity for sustained water supply to the surface under strongly evaporative conditions. This is supported by Price (Part 1), who noted that under clear (no fog) conditions on blanket peat, the Bowen ratio exceeded unity, and that the surface layer had a relatively high resistance to vapour transport. Combined with the severe reduction in evaporotion during fog, the seasonal evaporation from oceanic bogs is likely to be small in comparison with precipitation and runoff.

The relatively steep slopes associated with some blanket bogs commonly result in soil piping, erosion, and even massive slope failure (Moore, 1982). Seigel (1988) suggested that ground water can discharge from underlying sediments into local pipes, enhancing slope drainage. However, the topographic position of blanket bogs suggests they are groundwater recharge zones. Seigel (1988) found that downward hydraulic gradients in a vertical profile of an Alaskan blanket bog produced between 0.08 and 0.5 mm of daily recharge to underlying alluvial sediments. Wells (1976) noted that the mesotrophic state of Newfoundland blanket bogs suggests groundwater discharge. However, he failed to account for the high dissolved ion concentrations in fog (J.S. Price, unpublished data, 1990), which elevate the eutrophic state in maritime peatlands (Vitt et al., 1990).

STUDY AREA

The general characteristics of climate and geology are described in the companion paper (Price, 1992). The 9.4 ha basin chosen for the study has a total relief of 11 m, much of this attributable to the heath-covered hillocks which represent 10% of the basin area (Fig. 1). The peat-covered area (79%) has a relief of about 7 m, with the maximum gradient in the lower portion of the basin of 0.04. This is an order of magnitude greater than most continental bogs. Ponds and bog pools represent 11% of the area. The bog stream drains into Cripple Cove Creek, which is a 4.75 km^2 basin characterized by peatland (48%), heathland (42%), stunted balsam fir and spruce krumholtz (6%), and lakes (4%).

METHODS

The measurement period was from 26 May to 16 August 1989, and 16 May to 25 June 1990. However, site visits during mid-winter and early spring

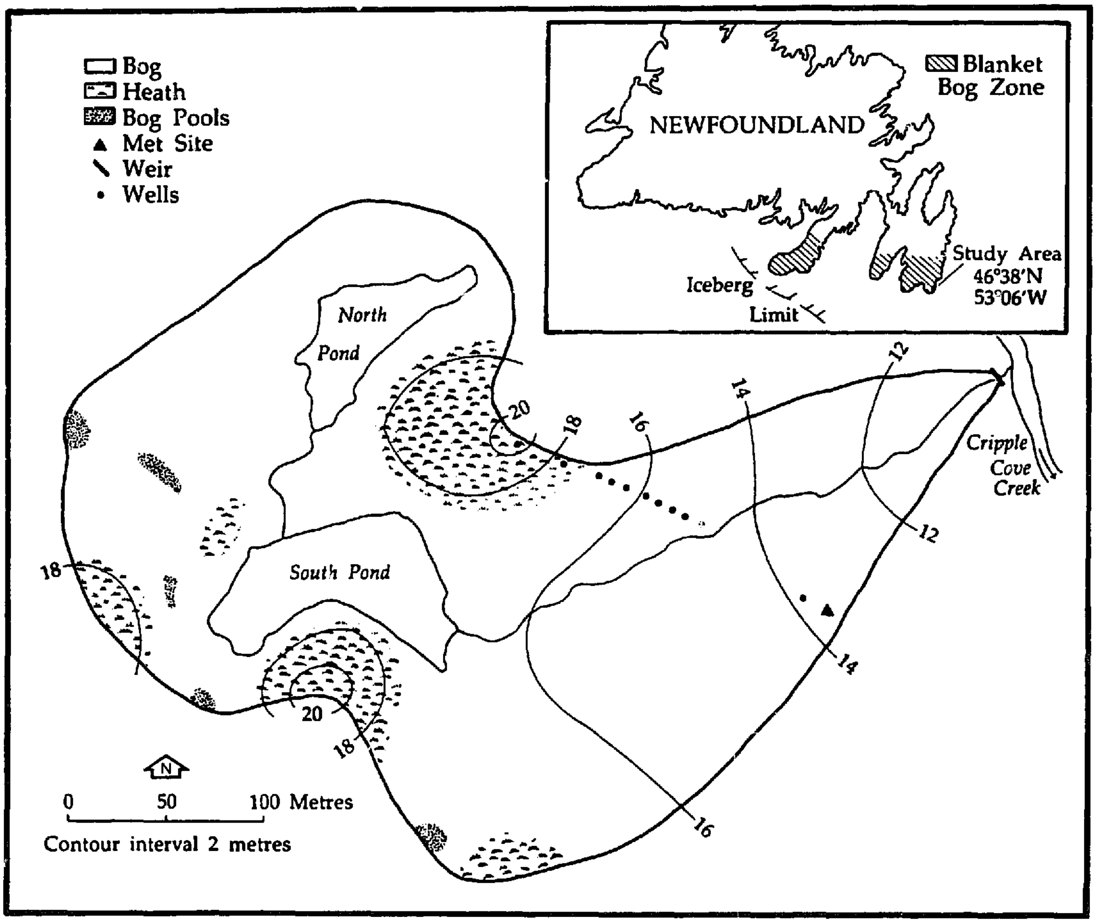

Fig. 1. The study area at Cape Race, Newfoundland.

provided additional data of a qualitative nature. Rain and fog measurements for the 1989 field season are described in the first paper (Part 1) of this series (Price, 1992). Similar procedures were followed in 1990, except five smaller lysimeters (dimensions 0.3 × 0.2 × 0.15 m deep) were used to estimate fog deposition to the surface. Data from these smaller lysimeters were invalidated when rain during the measurement period exceeded 4 mm, causing the vessel to overfill. A fog collector described by Price (Part 1) recorded timing and relative depth of fog, but was not representative of the actual surface deposition.

Discharge from the study basin was measured at a v-notched weir just above the confluence with Cripple Cove Creek (Fig. 1), by collecting overflow in a vessel of known volume over a timed interval. Higher flows were measured with a propeller-type current meter below the weir. Both methods were used to develop a stage-discharge rating curve, and stage was measured behind the weir with an electronic water level recorder. Discharge was also measured in Cripple Cove Creek, 500 m downstream of the confluence with the bog stream, using a current meter and stage recorder.

Groundwater wells and piezometers were set in a transect at 0.5, 1.0, 5, 15, 25, 45, 65, and 75 m from the stream bank (see Fig. 1). Wells (19 mm diameter) penetrated the entire peat deposit, and were slotted throughout their length. Piezometers (13 mm diameter) were set at the bottom of the peat layer, each having a slotted intake length of 100 mm. Rising and falling head tests were performed on wells and piezometers, respectively, and hydraulic conductivity was determined using Hvorslev's (1951) method. Specific yield was determined from a block of undisturbed peat in a lysimeter (dimensions $0.3 \times 0.3 \times 0.3$ m). The monolith was initially saturated to the surface, then covered with plastic wrap to prevent evaporation. A fixed volume of water was removed daily from a small (13 mm diameter) well located in one corner, and the water table allowed to equilibrate for 24 h before its elevation was measured. Specific yield (S_y) was calculated as

$$S_y = \frac{\Delta \mathrm{Vol}/A}{\Delta h} \tag{1}$$

where ΔVol is the daily volume of water removed, Δh is the change in water-table elevation, and A is the area of the lysimeter.

Evaporation (E) in both seasons was measured with a Bowen ratio/energy balance approach (see Part 1). The Priestley and Taylor (1972) evaporation model is a commonly used operational procedure to estimate evaporation from wetlands (e.g. Roulet and Woo, 1986; Price and Woo, 1988). It assumes that there is no vapour pressure deficit at the surface, the slope of the vapour pressure–temperature curve (S) can define the gradient of vapour pressure, and

$$E = \alpha \frac{S(Q^* - Q_G)}{L_V(S + \gamma)} \tag{2}$$

where Q^* and Q_G are net radiation and ground heat flux, respectively, γ is the psychrometric constant, and L_V is the latent heat of vaporization. When $\alpha = 1$, eqn. (1) becomes the equilibrium evaporation model, which describes evaporation when there is no vapour pressure deficit in the atmosphere. $\alpha = 1.26$ represents potential evaporation. To test the validity of this model in the maritime blanket bog environment, α was evaluated by substituting actual evaporation on the left-hand side of eqn. (2), then solving for α.

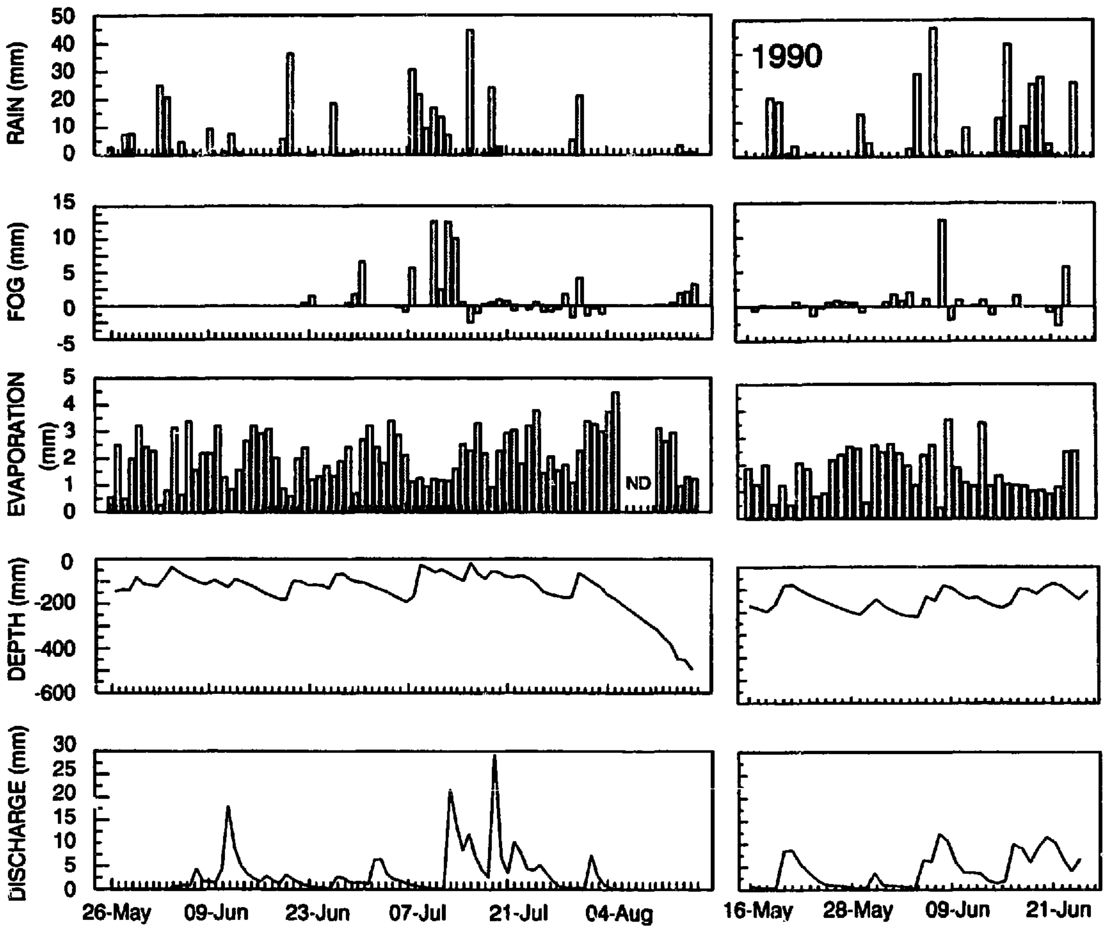

Fig. 2. From top to bottom: rain, fog, evaporation, water-table elevation, and streamflow for 1989 (left) and 1990 (right). All values in millimetres. ND indicates no data.

RESULTS

Precipitation

Snow accumulation is highly spatially variable, since snow redistribution by wind on this barren terrain redeposits most of the snow into the stream channels, similar to arctic and prairie snowpacks. Snow accumulation in general is limited, however, because of the frequent rain and melt events during winter. Stream flow does occur in winter, although it is unknown if it is continuous.

Rain (P) occurred very frequently during both years (Fig. 2). Rain at this location tends to be of frontal origin, and is generally of lower intensity, but of longer duration. Total rainfall for June and July 1989 was 136 mm and 125 mm, respectively, and for June 1–25 1990 was 199 mm. These totals are all high compared with the monthly average values at the government weather

TABLE 1

For Cape Race, Newfoundland, average monthly temperature (T), total monthly rainfall (P) (Environment Canada, 1982), and average number of days with observations of fog (visibility reduced below 1 km), between 1951 and 1980 (Environment Canada, 1984)

	Month											
	J	F	M	A	M	J	J	A	S	O	N	D
T (°C)	−2.9	−3.6	−1.7	0.9	3.9	8.1	12.0	13.4	11.0	7.1	3.8	−0.3
P (mm)	139	107	96	106	106	81	85	140	104	138	146	129
Fog (no. of days)	7	8	10	14	19	21	24	22	12	11	9	7

station 2 km away (Table 1), but the published values probably seriously underestimate the true value, since the gauge is at an exposed location on the cliff overlooking the ocean. Only the rainfall from 1 to 15 August 1989 (5.1 mm) is significantly below average.

Fog deposition (F_{fog}) determined by the lysimeter method is subject to errors, sometimes resulting in negative values in the range of −2 mm (Fig. 2), suggesting a daily range of error of ±2 mm. Since the error appears to be random, cumulative totals over a month or longer period approach the true values. Notwithstanding the error, daily accumulations of fog reached up to 10 mm. During foggy periods (e.g. 21 June–14 July 1989), fog deposition accounted for about 50% of rainfall, but over longer periods it was less important.

Evaporation

The evaporation ranged from 0.3 to 4.5 mm day^{-1}, and the overall daily average was 2.1 mm day^{-1} and 1.8 mm day^{-1} during the 1989 and 1990 field season, respectively (Fig. 2). However, during foggy periods, daily evaporation averaged only 1.1 mm, compared with 2.5 mm during clear periods (i.e. no fog or rain). Days with prolonged heavy fog lost $\ll 1$ mm to evaporation, although clear days following evening and early-morning fog typically had the greatest evaporation rates (e.g. 3.8 mm on 10 June 1990). High evaporation rates in periods following fog reflect the availability of water on the surface of the non-transpiring mosses which dominate the surface. The availability of water during these periods allows the surface to temporarily evaporate near the potential rate. This is demonstrated in Fig. 3, which indicates that half-hourly α values for the Priestley and Taylor (1972) evaporation model decline from the potential rate (approximately $\alpha = 1.26$) following the predawn fog, to below 0.9 by 09:00 h. Typically, the period of potential evaporation is very

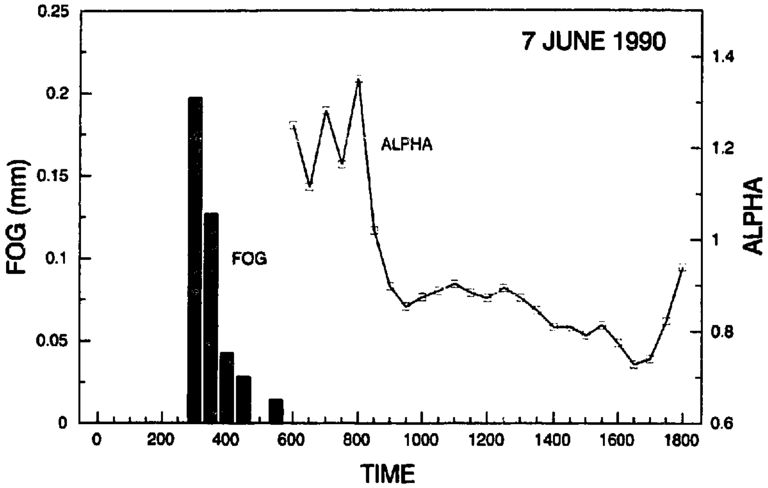

Fig. 3. Morning fog registered in the fog collector (Price, 1992), and α, the Priestley and Taylor (1972) evaporability parameter. Potential evaporation (α = 1.26) occurs from the wetted surface, but declines rapidly as the surface dries.

short; α falls below unity after very small water losses. Figure 4 shows the seasonal trend of α values in 1989 during fog ($\bar{\alpha}$ = 0.99) and clear ($\bar{\alpha}$ = 0.87) periods.

Ground water: storage and seepage

Water storage in the peat is reflected by the height of the water table (h) and specific yield (S_y), such that

$$\Delta S_{meas} = \Delta h S_y \tag{3}$$

where ΔS_{meas} is the measured storage change. The water-table position (h) ranges from 17 to 480 mm during the study, both those extremes occurring during 1989 (Fig. 2), which given the average specific yield of 0.25 (Fig. 5), represents a dynamic storage capacity of 115 mm of water. More typically the water table fluctuated between 80 and 200 mm, and was generally lower in 1990.

Peat in the vertical column was chiefly of *Sphagnum* spp. remains, and displayed a general, although not marked increase in the degree of humification with depth. This is reflected by the lack of a significant difference (Students $t_{0.90}$) between the average hydraulic conductivity in wells of 0.67×10^{-5} cm s^{-1} (on 17 June 1990), and in piezometers, which was 1.3×10^{-5} cm s^{-1} (excluding the piezometer near the interface of peat and heathland). There is a distinct trend of decreasing hydraulic conductivity from the heath hillocks toward the stream (Fig. 6). This may be related to the older (hence more decomposed) peat deposits generally found in lower lying areas

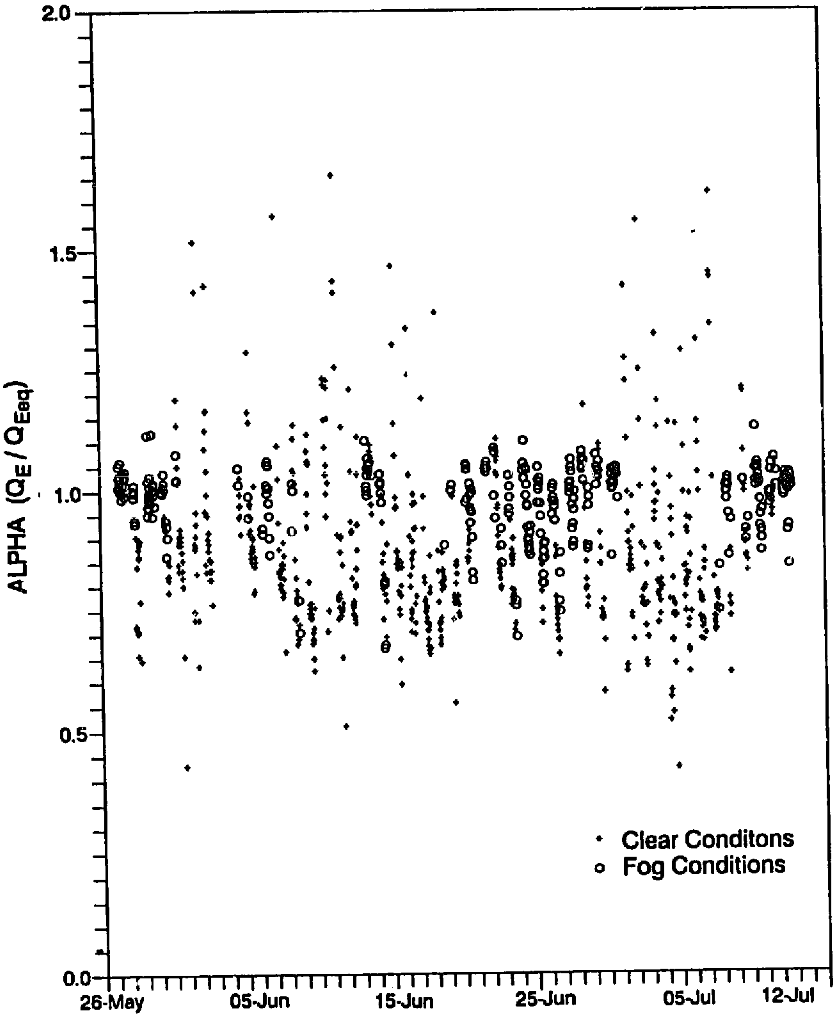

Fig. 4. Time series of Priestley and Taylor's (1972) evaporability parameter, α, during clear and fog conditions. Note limited variability during fog ($\bar{\alpha}$ = 0.99) compared with clear periods ($\bar{\alpha}$ = 0.87).

(T.E. Irwin, personal communication, 1989). The transmissivity (T) of a peat deposit to lateral water movement depends on the height of the saturated column (b), and the hydraulic conductivity (K), where $T = Kb$. Since hydraulic conductivity is higher in the upper layers, T varies with water-table elevation (Fig. 7). Unfortunately, no data are available during very high water-table periods, when transmissivity is most important in terms of water delivery to the channel.

Figure 8 depicts the water table and the potentiometric surface of water measured at the peat/sediment interface. The upper profile (16 June 1990) was measured during a rain event, and there is a strong downward hydraulic gradient averaging 0.16. On 18 June and 3 August 1990, which represent typical conditions, being neither very wet nor dry, the average gradients were

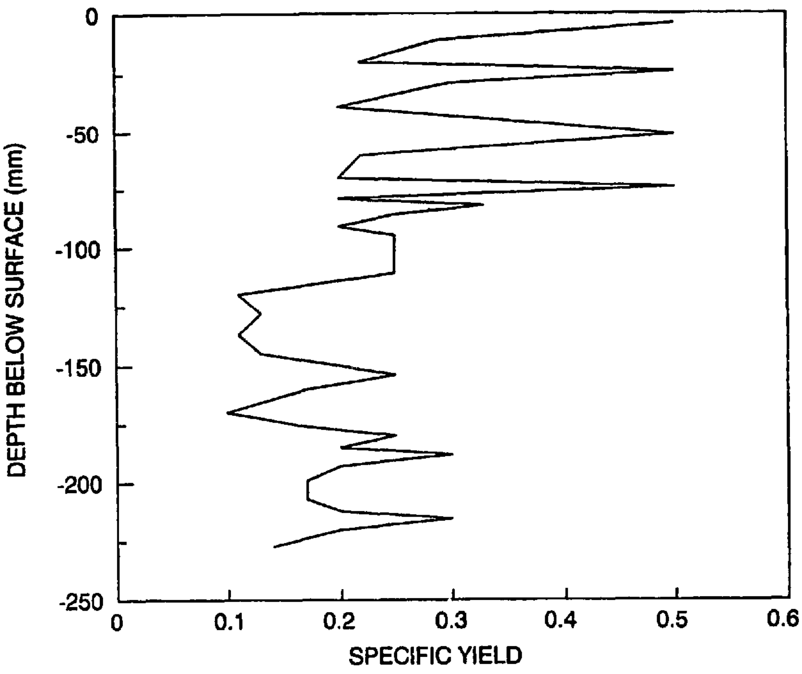

Fig. 5. Specific yield of peat.

0.07 and 0.08, respectively. Based on the latter gradients, and the average hydraulic conductivity of the peat profile, the rate of deep groundwater recharge (GW) is approximately 0.4 mm day^{-1}.

Streamflow

Stream discharge (Q) is flashy as in other bogs (Fig. 2), and is closely tied

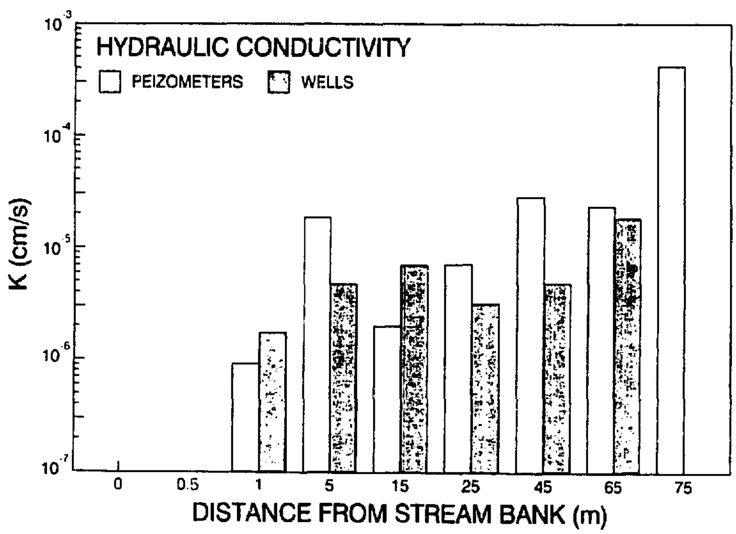

Fig. 6. Hydraulic conductivity in wells and piezometers along the transect.

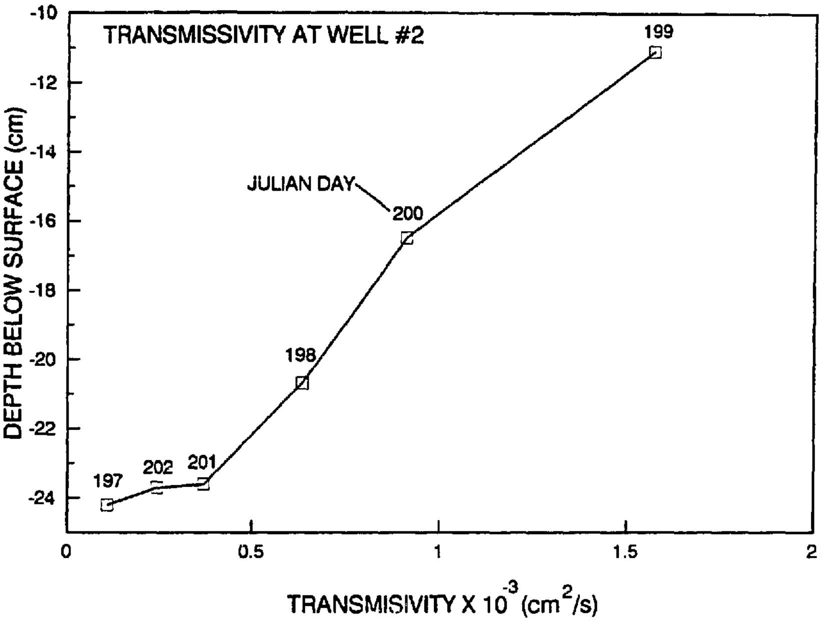

Fig. 7. Transmissivity at the well 5 m from the stream bank, under different water-table elevations.

to the water-table position. The average daily discharge was 3.7 mm and 4.2 mm for the measurement periods in 1989 and 1990, respectively. The lower figure in 1989 is due to the extended dry periods in June and August of that year. As with the water table, the highest flow (29.3 mm day^{-1}) and lowest (zero) occurred in 1989. Extreme values in 1990 were 12.7 and 0.3 mm day^{-1}. Low flows were typically sustained between 1 and 2 mm day^{-1}, even during extended dry periods. The exception was in August 1989, when extremely dry conditions were coupled with unusually high evaporation rates. Flows above 20 mm day^{-1} only occurred in 1989, and did so only following a series of wet days which raised the water table within 5 cm of the surface. Heavy rainfall in 1990 occurred during periods of lower water table and thus had a muted streamflow response. Streamflow from the bog had a more variable regime than Cripple Cove Creek (Fig. 9), into which the bog stream discharges. During peak flows, the bog streamflow was generally much higher, but between events Cripple Cove Creek released more water. Between 25 May and 10 August 1989, the cumulative discharge was 3.7 mm day^{-1} and 4.2 mm day^{-1} for the bog stream and Cripple Cove Creek, respectively. Discharge from Cripple Cove Creek was not measured in 1990.

Water balance

Water balance calculations were made between 5 and 28 July 1989, and between 16 May and 24 June 1990, corresponding to periods when the data

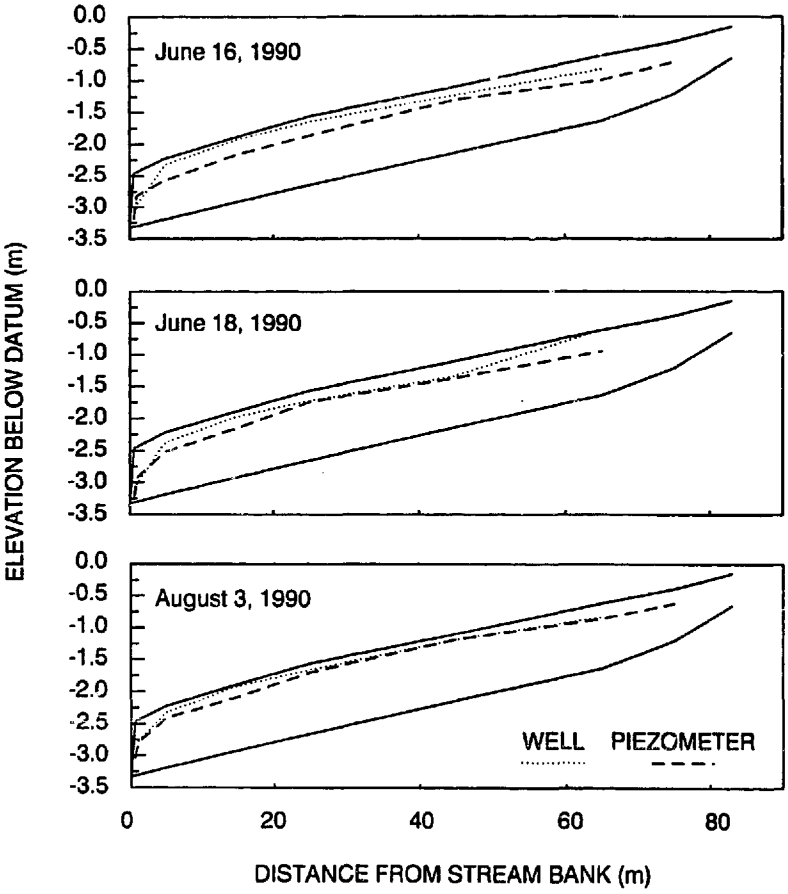

Fig. 8. Water table, and hydraulic potential at base of peat following rain (top), and under more 'typical' conditions (centre and bottom). Note downward gradient persists except very near the stream bank.

records were complete (Fig. 10). The water balance is

$$\Delta S_{cal} = P + F_{fog} - Q - E - GW \tag{4}$$

where ΔS_{cal} is the calculated storage change, or residual. The measured storage change (ΔS_{meas}) compared favourably with the calculated value in both years (Table 2). Since the final level of storage returned to near the starting value in both years, this component has little influence on the outcome of the balance. The two water balance periods share common traits which characterize this type of system. Rain dominates the inputs, and fog is significant, amounting to 18% and 10% of the rainfall in 1989 and 1990, respectively. Runoff dominated the losses, followed by evaporation (34% and 43% of runoff in 1989 and 1990, respectively), and ground water (10% of runoff in both periods).

DISCUSSION AND CONCLUSIONS

The hydrological regime of this blanket bog is strongly governed by its wet

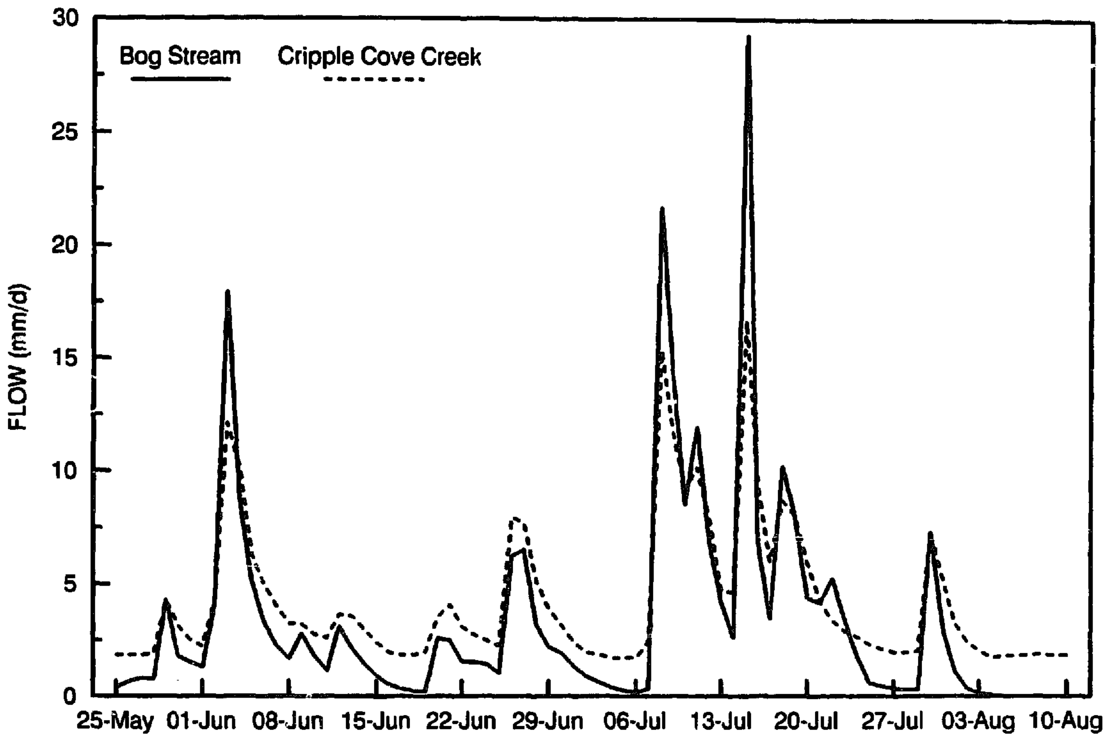

Fig. 9. Average daily streamflow from the bog, and in Cripple Cove Creek in 1989.

maritime location; fog has a multiple role. Fog provides a direct input of water to the system, and it also reduces the evaporative losses. While evaporation is low during fog, periods following fog experience evaporation at the potential rate until the surface moisture is consumed (Fig. 3). Thereafter, the rate diminishes quickly because of the high resistance to vapour diffusion through the peat (Price, 1992). This makes evaporation rates highly variable, and thus difficult to predict. This is evident from Fig. 4, which illustrates that during fog evaporation occurs at the equilibrium rate ($\bar{\alpha}$ = 0.99), and is controlled almost exclusively by the available energy (which is low), since the vapour pressure deficit is essentially nil. When it is clear ($\bar{\alpha}$ = 0.87), net radiation and the vapour pressure deficit are much higher, but water supply to the surface cannot keep up with the atmospheric demand, so α drops below unity. Furthermore, α is highly variable during clear periods, due to the changing moisture content of the peat, which affects the rate of water movement to the surface. The low α value during clear periods demonstrates that even in a relatively moist maritime environment, peatlands do not normally evaporate at or near the potential rate. Peatlands in non-maritime locations are probably subject to even more stringent limitations. Therefore, use of the Priestley and Taylor (1972) method in peatlands dominated by

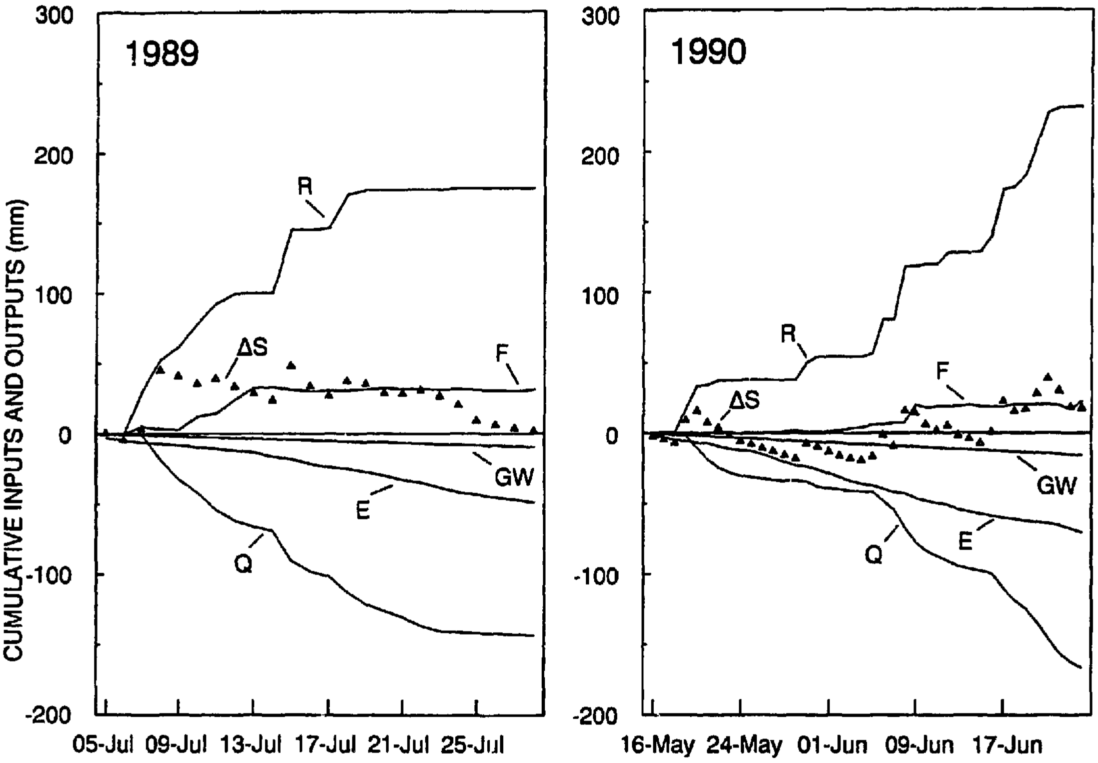

Fig. 10. Cumulative inputs and outputs of the water balance for 1989 (left), and 1990 (right).

non-transpiring plants, is subject to large errors unless α is determined experimentally.

The combination of enhance inputs, and limited evaporation losses result in higher water-table elevation in blanket bog, compared with bogs outside this zone (e.g. Northland Associates Ltd., 1989), even though the blanket bog surface gradients are steeper. Excluding the unusually dry period in August 1989, the water-table fluctuations are not extreme. These do not differ appreciably from continental bogs (e.g. Dai et al., 1974), except for a generally higher position. However, in continental bogs, where the water-table gradients are lower, recession is less dominated by runoff than in the blanket

TABLE 2

Water balance summary for the periods 5–28 July 1989 and 16 May–24 June 1990 (values are in millimetres)

Year	P	F_{fog}	Q	E	GW	ΔS_{calc}	ΔS_{meas}
1989	174	31	142	49	12	14	2
1990	231	23	160	70	16	17	7

bog, and more responsive to evaporation, especially in bogs with trees (Dai et al., 1974).

Water-table elevation is closely allied to streamflow fluctuations. The major streamflow peaks (exceeding 15 mm day^{-1}) all occurred in 1989, and did so when the water table came within 5 cm of the surface. This corresponds to the zone of greatest water mobility, or acrotelm (Ingram, 1978), as implied by the high specific yield in that layer (Fig. 5). In 1990, the water table did not come to within 5 cm of the surface, and consequently, the runoff peaks were lower. Delivery of rain water to the stream is a function of the transmissivity of the peat, which is clearly a function of water-table position (Fig. 7). However, although the slopes are relatively steep, they are also long (Fig. 1), so it is unlikely that all but a small segment near the stream makes a contribution within 24 h of a storm event. Even if the hydraulic conductivity of the acrotelm is as high as 10^{-2} cm s^{-1}, the gradients along the transect (≈ 0.04) drive the water less than 1 m day^{-1}. Since runoff is the most significant process of water loss, amounting to 63% in both summers, near-stream processes and pipe-flow must be considered. While no evidence of piping was evident along the bog stream downstream from South Pond (see Fig. 1), a large pipe (0.3 m diameter) exists between South Pond and North Pond, which is not active except during high-flow periods. No estimate of its contribution could be made. Other undetected pipes may also exist. At the transect, the stream is incised 1.5 m through the peat, down to the bouldery till substrate. The water table rises about 1.3 m within 5 m of the stream bank, and this gradient increases during periods of rainfall. The high gradient, coupled with the higher transmissivity (Fig. 7) during these periods is responsible for a significant portion of the runoff production.

Streamflow was at base-flow levels (< 1 mm day^{-1}) when the water table dropped below 15 cm from the surface, where specific yield and hydraulic conductivity are appreciably lower (Figs. 5 and 6), corresponding to the catotelm (Ingram, 1978). The general trend of downward groundwater seepage is also reversed at the stream bank (Fig. 8), and groundwater discharges from the peat into the stream, helping to sustain flow during dry periods. Water is also released from the ponds during these times.

The streamflow regime of the bog is quite different from Cripple Cove Creek (Fig. 9), exhibiting much less storage effect. While some of the difference is due to the larger basin size of Cripple Cove Creek, hydrological processes near the surface also differ. A water table was never observed in the heath during the two study periods. In two soil pits dug in heathland terrain, which represents 42% of the surface of the larger basin, the water table was not encountered at 1.5 m. The storage capacity of the relatively large unsaturated layer provides a mechanism to dampen the hydrograph response, which is evident in Fig. 9. The heath is typically very dry, which when coupled with a low water table, results in evaporation rates which are probably lower than in the bog. This is reflected by the higher streamflow loss from Cripple Cove Creek.

The water balance (Table 2) provides a tool for assessing the relative importance of the hydrological processes operating on the bog. The water balance indicates that fog increases the total precipitation by 10–20% during the summer months, and that increases to 50% during foggy periods. If fog input is sustained at these rates, the published historical value of May–August rainfall (412 mm; see Table 1) could be enhanced by 40–200 mm. However, taking into account the reduction in evaporation (approximately 1.4 mm day^{-1}) caused by fog, this could represent a suppression of up to 170 mm of evaporation. The enhanced input and suppressed losses thus incurred could tip the water balance by up to 300 mm compared with a site outside the fog-bound zone. It is this combination of enhanced inputs, and suppressed losses which provides the conditions necessary for blanket bog development in Newfoundland.

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REFERENCES

Barnes, J.L., 1984. The hydroclimatological characteristics of a peatland dominated catchment in southeastern Newfoundland. M.Sc. Thesis, Memorial University of Newfoundland, St. John's, 300 pp.

Barrie, L.A. and Schemenauer, R.S., 1986. Pollutant wet deposition mechanisms in precipitation and fog water. Water, Air, Soil Pollut., 30: 91–104.

Bay, R.R., 1969. Runoff from small peatland watersheds. J. Hydrol., 9: 90–102.

Burke, W., 1972. Aspects of the hydrology of blanket bog. In: Hydrology of Marsh Ridden Areas, Proc. Minsk, Symp., Paris, 1975. UNESCO–IAHS, pp. 171–182.

Dai, T.S., Haavisto, V.F. and Sparling, J.H., 1974. Water level fluctuation in a Northeastern Ontario Peatland. Can. J. For. Res., 4: 76–81.

Davis, A.M., 1984. Ombrotrophic peatlands in Newfoundland and Canada: Their origins, development and trans-Atlantic affinities. Chem. Geol., 44: 287–309.

Environment Canada, 1982. Canadian Climatic Normals 1951–1980: Temperature and Precipitation, Atlantic Provinces. Atmospheric Environment Service, Downsview, 136 pp.

Environment Canada, 1984. Canadian Climatic Normals 1951–1980: Soil Temperature, Lake Evaporation, and Days with . . . Fog. Atmospheric Environment Service, Downsview, 109 pp.

Environment Canada, 1986. Canada's Wetland Regions (map). Canada Committee on Land Classification, Lands Directorate, Environment Canada, Ottawa.
Hare, F.K., 1952. The climate of the island of Newfoundland. A geographical analysis. Can. Dept. Mines Tech. Surv., Geogr. Bull., 2: 36–38.
Hvorslev, M.J., 1951. Time lag and soil permeability in groundwater observations. US Army Corps of Engineers Waterways Exp. Stn. Bull., 36, Vicksburg, MS.
Ingram, H.A.P., 1978. Soil layers in mires: Function and terminology. J. Soil Sci., 29: 224–227.
Ingram, H.A.P., 1983. Hydrology. In: A.J.P. Gore (Ed.), Mires: Swamp, Bog Fen and Moor. Elsevier, New York, pp. 67–158.
Lafleur, P.M., 1990. Evaporation from wetlands. In: N.T. Roulet (Ed.), Focus: Aspects of the Physical Geography of Wetlands. Can. Geogr., 34: 79–88.
Moore, P.D., 1982. Beneath the blanket bogs of Britain. Nat. Hist., 91(11): 48–55.
Mulqueen, J., 1986. Hydrology and drainage of peatland. Environ. Geol. Water Sci., 9: 15–22.
Northland Associates, Ltd., St Johns, 1989. Research on Bog Preparation and Peat Sod Drying Techniques in Newfoundland (1986–1988), Energy, Mines and Resources (Canada). Northland Associates, Ltd., St. Johns, 161 pp.
Price, J.S., 1992. Blanket bog in Newfoundland. Part 1. The occurrence and accumulation of fog-water deposits. J. Hydrol., 135: 87–101.
Price, J.S. and Woo, M.K., 1988. Studies of a subarctic coastal marsh. I. Hydrology. J. Hydrol., 103: 275–292.
Priestley, C.H.B. and Taylor, R.J., 1972. On the assessment of surface heat flux and evaporation using large scale parameters. Mon. Weather Rev., 100: 81–92.
Roulet, N.T. and Woo, M.K., 1986. Wetland evaporation in the low arctic. Arct. Alp. Res., 18: 195–200.
Seigen, D.I., 1988. The recharge-discharge function of wetlands near Juneau, Alaska: Part I. Hydrogeological Investigations. Ground Water, 26: 427–434.
Vitt, D.H., Horton, D.G., Slack, N.G. and Malmer, N., 1990. *Sphagnum*-dominated peatlands of the hyperoceanic British Columbia coast: patterns in surface water chemistry and vegetation. Can. J. For. Res., 20: 696–711.
Wells, E.D., 1976. A classification of peatlands in eastern Newfoundland. M.Sc. Thesis, Memorial University of Newfoundland, St. John's, 201 pp.