EVAPORATION FROM A BLANKET BOG IN A FOGGY COASTAL ENVIRONMENT

(*Research Note*)

JONATHAN S. PRICE

Geography Department, Queen's University, Kingston, Ontario, Canada, K7L 3N6

(Received in final form 18 March, 1991)

Abstract. Frequent fog severely restricts evaporation from blanket bogs in Newfoundland because it more than halves the radiant energy input, and it eliminates the vapor pressure deficit, resulting in evaporation at the equilibrium rate (average $\alpha = 0.99$ during fog). During these periods, there is no surface resistance to evaporation because the bog has been wetted by fog drip, and although the latent heat flux dominates over sensible heat (average $\beta = 0.8$), both are small. In contrast, the surface dries during clear periods, increasing the surface resistance to evaporation so that sensible heat becomes more important ($\beta = 1.05$). When the mosses are dry, evaporation is below the equilibrium rate ($\alpha = 0.87$), although the higher available energy ensures that actual evaporation is higher. During clear periods, daily evaporation averaged 2.5 mm, compared to 1.1 and 0.7 mm for fog and rain, respectively. The suppressed evaporation at this site is important in maintaining appropriate hydrological conditions for blanket bog development.

1. Introduction

Bogs are a class of peatland. They normally develop without lateral surface or subsurface water inflows, and thus require a limited set of atmospheric water flux conditions, where evaporation during the warm season does not exceed rainfall (Romanov, 1968). Topographic conditions are normally such that water is detained at or near the surface for long enough to promote peat formation. Only under exceptional circumstances are conditions sufficiently wet and cool that peat can blanket the entire landscape, including higher relief areas. Such peatlands are called blanket bogs, and these have a limited occurrence in eastern North America, being restricted to the southern Avalon and Burin Peninsulas in Newfoundland (Davis, 1984). While these are notably cool and wet locations, they are less so than other Newfoundland locales not supporting blanket bog. However, the blanket bog region has an extraordinarily high frequency of fog which adds a significant quantity of water during the summer (Price, 1991). The addition of fog drip, coupled with suppressed evaporation may sufficiently modify the water balance to produce the threshold moisture conditions necessary for blanket bog formation. Indeed, the recent difficulties experienced with inadequate bog drainage and sod drying in fuel peat operations at St. Shotts (Northland Associates, 1989) are directly related to these poorly understood microclimatic processes. Thus the objectives of this study are to quantify the rates and variability of daily and seasonal evaporation, and to examine the energy balance of the blanket bog surface in order to

Boundary-Layer Meteorology **57**: 391–406, 1991.

understand better the role of advection in a maritime environment, and especially the role of advective fog.

Unfortunately, the state of knowledge of evaporation from bogs in general is poor, and data are conflicting and often misleading, perhaps due to the use of oversimplified methods. For example, Nichols and Brown (1980) found that evaporation from moss peat monoliths in a growth chamber was less (*sic*) when the water table was at the surface than when it was 5 to 15 cm below. However, the bulk of the field evidence indicates that evaporation is reduced as the water table is lowered (Virta, 1960, 1966, quoted from Ingram, 1983; and Williams, 1970). Romanov (1968) found that the evaporation rate decreased markedly when the water table dropped 15–20 cm below the surface, which is the limit of the root structure of vascular plants on bogs. He further noted that the moss structure is such that the pore size is very large in the layer 3 to 9 cm below the surface. Because of the low capillarity of such large pores, "there is no sustained supply of water to the surface . . . under strongly evaporative conditions" (Ingram, 1983, p. 81). Hence, it is doubtful that bogs can maintain an evaporation rate at or near the potential rate. Nevertheless, Ingram (1983), in his extensive survey of mostly European and Soviet literature, concluded that actual evapotranspiration from bogs is approximately equal to potential evapotranspiration. This points to a poor understanding of the water transfer processes occurring within the moss, and its variable resistance to evaporation. Lafleur (1990) indicates that a non-transpiring surface such as sphagnum experiences surface resistance (in the upper layer), which is a function of the gas and liquid diffusivity of the soil and the temperature gradient above the water table. Resistance to evaporation occurs even in transpiring wetland plants where there is no soil water deficit, due to stomatal control (Rouse *et al.*, 1987). Furthermore, Lafleur and Rouse (1988) noted that in a wetland site, the increase in canopy resistance corresponding to seasonal vegetation growth dampened the evapotranspiration flux.

The role of advection in a marine environment was considered by Rouse and Bello (1985), and Rouse *et al.* (1987). They found that cold onshore winds enhance the sensible heat flux over the latent flux. While vegetative control was acknowledged, the role of small saturation deficits was important, so can be expected to be very significant under an advective fog regime. Lafleur and Rouse (1988) found that surface resistance was lower in a subarctic coastal marsh during onshore advective regimes, but that it dominated over aerodynamic resistance. The relative effect of these resistances over a moss surface in a coastal environment is unknown.

2. Theoretical and Empirical Models of Evaporation

Evaporation represents the consumption of latent energy and thus is part of the surface energy balance, which is given as

$$Q^* = Q_H + Q_E + Q_G \tag{1}$$

where the terms are net radiation, sensible, latent and ground heat flux, respectively. Direct measurement of Q_H and Q_E is difficult, but their ratio Q_H/Q_E is proportional to the gradient (σ) of dry-bulb air temperature (T_a) and vapour pressure (e), such that

$$\beta = \frac{Q_H}{Q_E} = \gamma \frac{\sigma T_a}{\sigma e} \tag{2}$$

where β is the Bowen ratio and γ is the psychrometric constant. Vapour pressure was determined as

$$e = e^* - \gamma(T_a - T_w)$$

where e^* is the saturation vapour pressure at T_w, the wet-bulb temperature. Rearranging Equation (1), then substituting Equation (2), we get

$$Q_E = \frac{Q^* - Q_G}{1 + \beta}, \tag{3}$$

then

$$E = \frac{Q_E}{L_V} \tag{4}$$

where E is the evaporation, and L_V is the latent heat of vaporization.

The Penman–Monteith combination model provides a one-dimensional description of the latent heat flux from a surface by considering the resistance to vapour flow from aerodynamic sources (r_a), and resistance from the canopy (r_c) such that

$$Q_E = \frac{S(Q^* - Q_G) + \rho C_p VPD/r_a}{S + \gamma(1 + r_c/r_a)} \tag{5}$$

in which S is the slope of the vapour pressure-temperature curve at the air temperature, ρ is the air density, C_p is the specific heat of air at constant pressure, and VPD is the vapour pressure deficit. Aerodynamic resistance is a function of the surface roughness, which controls the amount of turbulence for a given wind velocity, such that under neutral stability conditions, when bluff body effects are ignored (Thom, 1975)

$$r_a = \frac{[\ln(z - d)/z_0]^2}{k^2 U_z}, \tag{6}$$

where z_0 is the roughness length, d is the zero plane displacement, k is von Karman's constant, and U_z is the wind velocity at elevation z.

Canopy resistance (r_c) is closely related to the resistance produced by individual leaves within the canopy (Monteith, 1965). Since this is difficult to determine, the bulk surface resistance (r_s) can be computed from meteorological measurements; re-arranging Equations (3) and (5)

$$r_s \approx r_c = (\beta + 1)r_i + \left[\beta\left(\frac{S}{\gamma}\right) - 1\right] r_a \tag{7}$$

where r_i is the climatological resistance, given as

$$r_i = \frac{\rho C_p}{\gamma} \cdot \frac{VPD}{Q^* - Q_G} \tag{8}$$

and the other terms are as previously defined. r_i is not a true resistance, but is a measure of the dominating overhead climatological conditions, and quantifies the relative importance of VPD and the available energy $(Q^* - Q_G)$.

The application of Equation (7) assumes that the individual resistances of surfaces within the canopy can be treated as a unit, and the resistance is determined by the properties of the heat and vapour fluxes measured in the constant flux layer above that surface. This concept can be extended to a non-transpiring moss surface, where the bulk surface resistance, determined in a similar manner, represents the resistance that the moss provides to vapour diffusion. The values so determined are useful as a comparative tool for studying the role of variable surface wetness. Although the internal processes of vapour diffusion through a moss surface are not the same as stomatal control in a vascular canopy, lumping the processes by evaluating Equation (7) permits intercomparison with other surfaces such as wetland sedges and trees (Lafleur and Rouse, 1988), or heath vegetation (Miranda *et al.*, 1984) that have also been evaluated within this framework.

The Penman–Monteith combination model (5) can be simplified under certain conditions. If the vapour pressure deficit and surface resistance approach zero, the right hand-term in (5) drops out, and

$$Q_{Eeq} = \frac{S(Q^* - Q_G)}{S + \gamma} \tag{9}$$

where Q_{Eeq} refers to equilibrium evaporation. Q_{Eeq} approximates Q_E when conditions are not too wet or too dry (Wilson and Rouse, 1972). Actual evaporation is related to equilibrium with a coefficient of evaporability (α) (Priestley and Taylor, 1972), so that

$$Q_E = \alpha \frac{S(Q^* - Q_G)}{S + \gamma}. \tag{10}$$

Combining (9) and (10) then solving for α gives

$$\alpha = \frac{Q_{Eeq}}{Q_E}. \tag{11}$$

When $\alpha \approx 1$, (10) represents the equilibrium evaporation condition where the surface is neither excessively wet or dry. Open water and many wet surfaces which

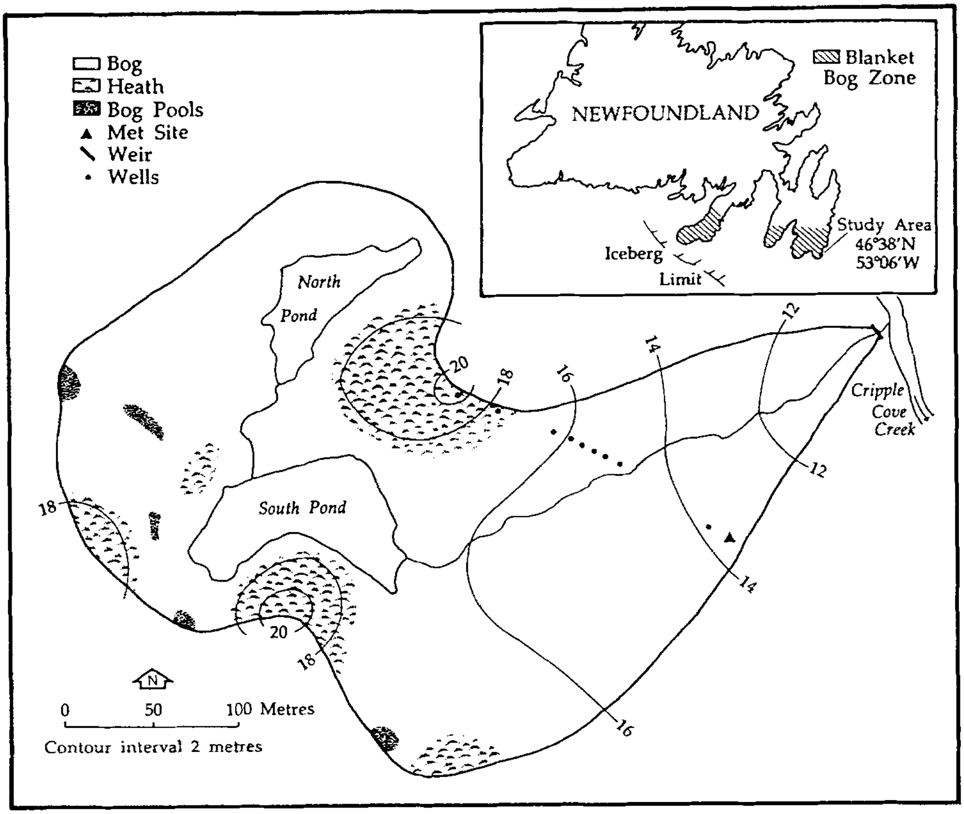


Fig. 1. The study area. Inset shows the annual iceberg limit, which indicates the extent of cold water of the Labrador Current which encircles the blanket bog zone of Newfoundland, and which is responsible for heavy, frequent fog (Farmer, 1981).

evaporate at the potential rate have α equal to 1.26 (Priestley and Taylor, 1972; Brutsaert, 1982). These findings were confirmed for a wet sedge surface by Stewart and Rouse (1977), but later work by Rouse *et al.* (1987) indicates that α depends on temperature and *VPD*, and that in maritime environments it typically falls below unity during cooler onshore winds. There is a paucity of information on α values for moss surfaces. Price *et al.* (1991) found peat ridges in a Labrador string bog to have $\alpha = 1.1$.

3. Study Area

The study location (Figure 1) was at Cape Race (46° 38′ N, 53° 06′ W), which lies within the southeast climatic zone of Newfoundland, and is characterized by cool summers with persistent fog (Banfield, 1981). The climate is strongly affected by the Labrador current, which encircles the Avalon and Burin Peninsulas, bringing cold water and ice in the spring and summer (Farmer, 1981) (see Figure 1). The predominantly southwesterly airflow over the cold ocean produces advection fog.

Coastal exposures in the study area indicate that approximately 6 m of stony glacial till overlies sandstone bedrock. The topography is characterized by gently rolling hills, with a relief ranging from 30–50 m at the cliff crest, to about 60 m at the study site, which lies 750 m from the coast at its nearest point.

The blanket bog is extensive (Wells, 1976), and its surface is predominantly *Spagnum fuscum*, with a variety of small ericacea such as *Empetrum spp.*, and patchy cover of *Rubus spp.* and *Cladonia spp.* There is a thin but fairly even cover of *Scirpus spp.* The bog is ombrogenous, but its trophic state has been elevated by generous amounts of solute-enriched fog (Price, unpublished data).

4. Methods

The study was performed between 26 May and 11 July 1989. Evaporation was determined using the Bowen ratio/energy balance method. Q^* was measured directly with a REBS net radiometer which was factory-calibrated prior to installation, and cross-calibrated with a new Middleton net radiometer before and after the field season. There was no appreciable drift. Q_G was determined with a REBS soil heat flux plate which was factory-calibrated prior to installation. It was embedded 1 cm beneath the surface of the *Sphagnum* carpet, which was remarkably uniform at the site. During fog and rain, measurement of Q^* was affected by water on the domes. These were dried periodically during and immediately after fog, but during rain, they were not attended. Some measurement inaccuracy is inevitable during these conditions, which, when coupled with the small humidity gradients reduces the reliability of the evaporation estimates. However, in terms of total vapour flux, the error is small because of the low absolute values at these times. An error analysis is presented later.

Temperature and vapour pressure were measured with a thermocouple and psychrometer system installed on a tower at 0.5, 1.0, 1.5, and 2.0 m. Dry-bulb (T_a) and wet-bulb temperatures (T_w) were measured at each level with a potted thermocouple, and the wet bulb was covered by a saturated wick. Both the wet- and dry-bulb thermocouples were housed in a shielded chamber aspirated with an electric fan. The multi-level system was within the surface boundary layer, as the fetch in all directions exceed 200 m.

Bowen ratios were determined from the profiles of half-hourly temperature and vapour pressure, using a computer routine to calculate the mean slope of the temperature vs. vapour pressure curve at all levels. During rain, the gradients were small (see Table I), so β at these times is least reliable. The fog is of advective origin, and passes over a weakly to moderately heated surface. Thus, during daylight hours, the gradients of temperature and vapour pressure were sufficient (e.g., Table I) to determine the Bowen ratio. A wind profile system with cup anemometers at 0.2, 0.4, and 0.8 m was employed to find the surface roughness parameter z_0 when the vegetation cover was fully developed (early August).

Rain was recorded with a tipping bucket rain gauge. Fog was collected with a

TABLE I

Average air temperature, *VPD* and windspeed between 0600–1800 h, and gradients of temperature and vapour pressure from 0.5 to 1.5 m at noon

Date Condition	16 June clear	28 June fog/clear	29 June fog
Temperature (°C)	9.5	14.0	12.8
VPD (kPa)	0.35	0.02*–0.18§	0
dT/dz (°C m^{-1})	0.98	1.06	0.30**
de/dz (kPa m^{-1})	0.053	0.080	0.032
Windspeed (m s^{-1})	9.9	4.1*–5.6§	5.6

* 0600–1100 (fog).
§ 1130–1800 (clear).
** Also the wet bulb gradient at this time.

fog collector consisting of 2000 m of nylon monofilament line, strung onto 3 vertically oriented concentric cylindrical aluminum frames (Goodman, 1985) positioned over a funnel, which was directed into a tipping-bucket rain gauge. The depth of fog deposition was determined as the volume of water divided by the area of the collecting funnel, which was 0.65 m dia. If rain was registered simultaneously, fog deposition was ignored. The fog collector was used to record relative volumes and time of fog, but was not calibrated to the peat surface.

Three conditions were tabulated. 'Clear' indicates that fog or rain was not occurring, 'rain' was identified by tips in the rain gauge, and 'fog' was indicated by tips in the fog collector's gauge when none was occurring in the rain gauge. Measurements were recorded with a Campbell Scientific 21X electronic data logger, which gave sensor output signals every 20 s, and averaged them (or totalled them where appropriate) over each 1/2 h period.

5. Results

Between 26 May and 11 July 1989, fog was experienced 41.4% of the time, and rain 11.3%, based on measurements at half-hour intervals. Averages of the half-hourly data between 0600–1800 h indicate that the radiation balance is strongly affected by the atmospheric condition (Figure 2a). Average net radiation during clear periods (i.e., when fog or rain was not occurring) was 321 W m^{-2} compared to 133 and 57 W m^{-2} for fog and rain periods, respectively. Average vapour pressure was high and varied within a limited range, being highest for fog (1.30 kPa), intermediate for rain (1.22 kPa), and lowest for clear conditions (1.18 kPa). Saturation occurred under fog and rain, so *VPD* was essentially zero. The average *VPD* was small even during clear conditions (0.22 kPa) compared to values at other sites (e.g., Lafleur and Rouse, 1988).

The Bowen ratio (Figure 2c) indicates how available energy ($Q^* - Q_G$) was partitioned into sensible heat (Q_H) and latent heat (Q_E). On average, clear conditions ($\beta = 1.05$) marginally favoured sensible heat. However, under fog ($\beta =$

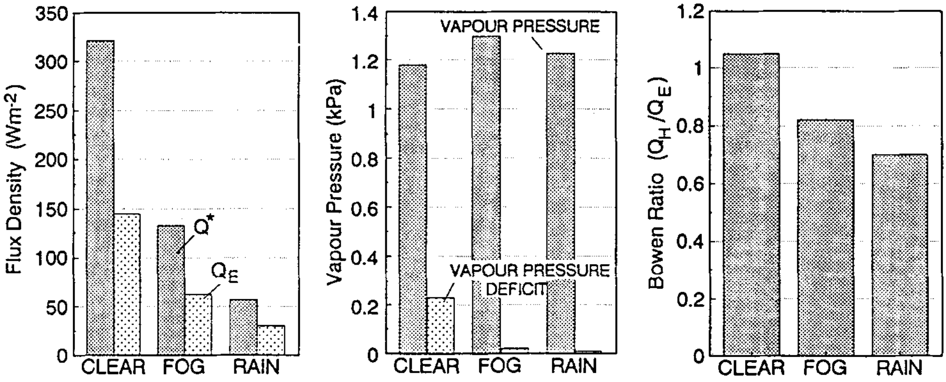


Fig. 2. Half-hourly averages (26 May–11 July, 1990) between 0600–1800 h of (a) net radiation and latent heat, (b) vapour pressure and vapour pressure deficit, and (c) the Bowen ratio.

0.82) and rain ($\beta = 0.70$), latent heat flux dominated. Nevertheless, the weak radiant energy during fog and rain produced relatively low values of Q_E (Figure 2a) so that the mean evaporative flux was small. In contrast, more evaporation was possible during clear periods not only because of the greater available energy, but also because of the higher vapour pressure deficit. Clear conditions produced an average evaporation of 2.5 mm d^{-1}, compared to 1.1 and 0.7 mm d^{-1} for fog and rain, respectively.

Daily values of available energy, vapour pressure deficit and evaporation are shown in Figure 3. The available energy was generally low compared to seasonal values at St. John's (Banfield, 1981). Daily values below 100 W m^{-2} correspond to days with rain, and values above 250 W m^{-2} are associated with clear days, and the intermediate values with fog. The vapour pressure deficit was generally under 0.2 kPa, which is much lower than a subarctic wetland on James Bay (Lafleur and Rouse, 1988), which had values commonly ranging between 0.2–0.5 kPa for influxes of maritime air. Here, values above 0.2 kPa were associated with unusually warm days. Evaporation over the study period ranged from 0.2 to 3.4 mm d^{-1}. On a seasonal basis, most of the variance in evaporation is explained by the available energy ($r^2 = 0.97$), slightly modified by *VPD* ($r^2 = 0.51$).

The partitioning of the energy balance can be addressed by examining the half-hourly data for selected days. The data show the expected diurnal trends, but modified by advection, which affects humidity, temperature and turbulence. Consider three typical days with different atmospheric conditions (Figure 4). June 16 was clear and mostly cloudless. The average temperature was 9.5 °C, peaking at 14.2 °C at 1330 h. The surface was dry because of the absence of fog or rain during the preceding 2 days. The sensible heat flux was relatively strong ($\beta = 1.2$–1.6) due to heating of the dry surface. Even though Q_H dominated, Q_E was also relatively large because of the high available energy on this day. A relatively strong evaporative flux of 3 mm was recorded.

On 28 June (Figure 4b), the average air temperature was 14.0 °C, varying

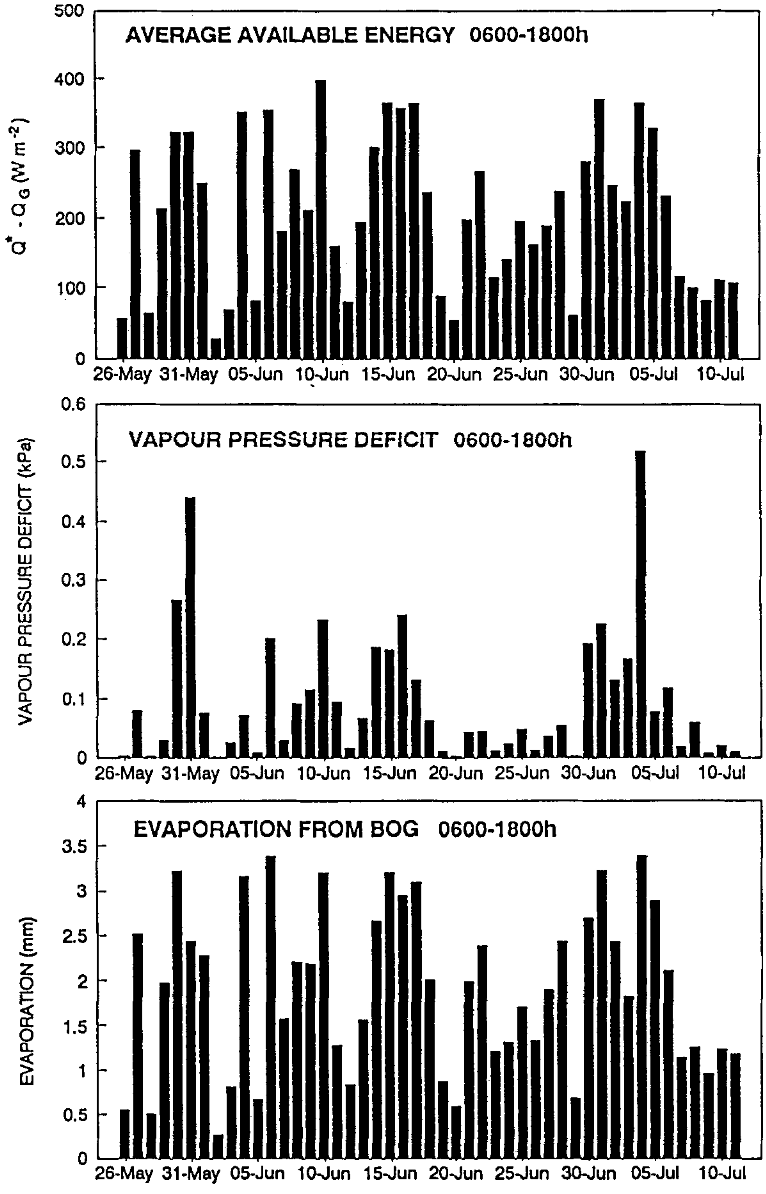


Fig. 3. Daily (a) available energy, (b) vapour pressure deficit, and (c) evaporation, between 26 May and 11 July 1990.

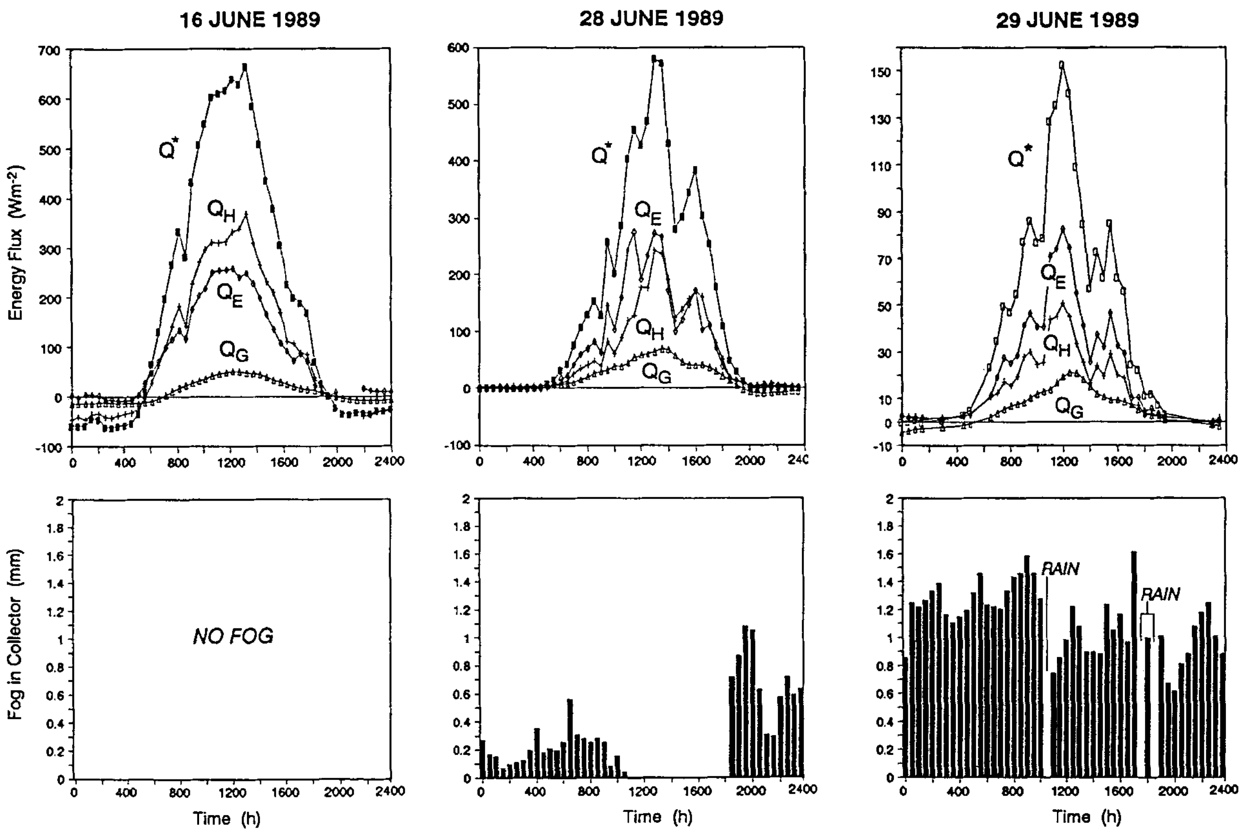


Fig. 4. Energy partition (top) under different advective regimes as indicated by fog (bottom). Note the different scales for the energy flux density.

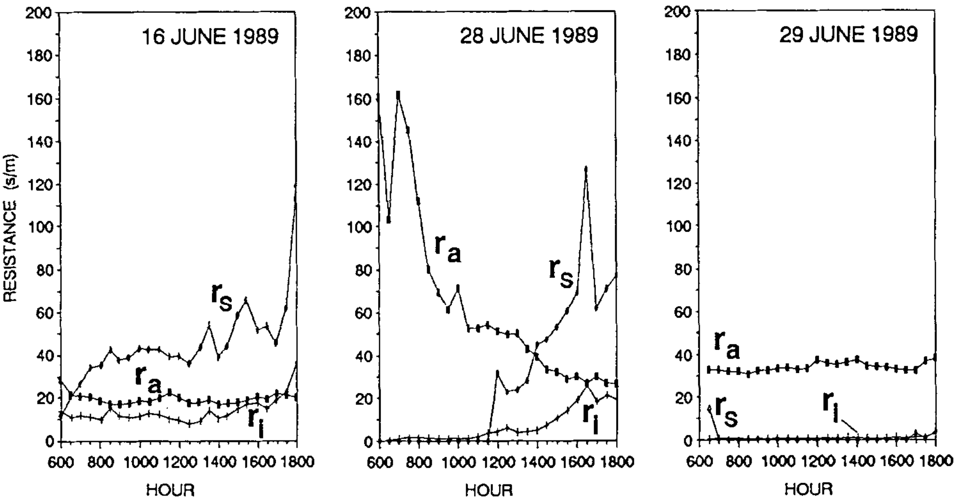


Fig. 5. Resistances corresponding to conditions presented in Figure 4a, b, and c. r_s is surface resistance, r_a is aerodynamic resistance, and r_i is climatological resistance.

between 12.9 and 15.4 °C except during the period 1530–1730 h when it rose to 19.2 °C after the sky cleared. There had been light fog all night which moistened the surface, so that during the morning, Q_E dominated ($\beta = 0.6$). By mid-afternoon when surface moisture had been removed by evaporation, the Bowen ratio approached unity. On this day, 2.4 mm of evaporation occurred.

On 29 June (Figure 4c), heavy fog was recorded all day, severely limiting the radiative input (153 W m^{-2} at 1200 h). The average temperature was 12.8 °C, varying by less that ±1 °C during daylight hours. On this day, the surface was perpetually moist, and the latent heat flux dominated ($\beta = 0.6$–0.7), but daily evaporation was only 0.7 mm.

These typical days demonstrate that the condition of the atmosphere has three notable effects; (1) during clear periods, cool maritime air enhances sensible heat flux, although latent heat is still relatively high because more energy is available, (2) advection of fog reduces the energy at the surface and favours latent heat, but at the same time severely limits the rate of latent heat transfer, and (3) fog drip moistens the moss surface, which enhances evaporation in subsequent clear periods.

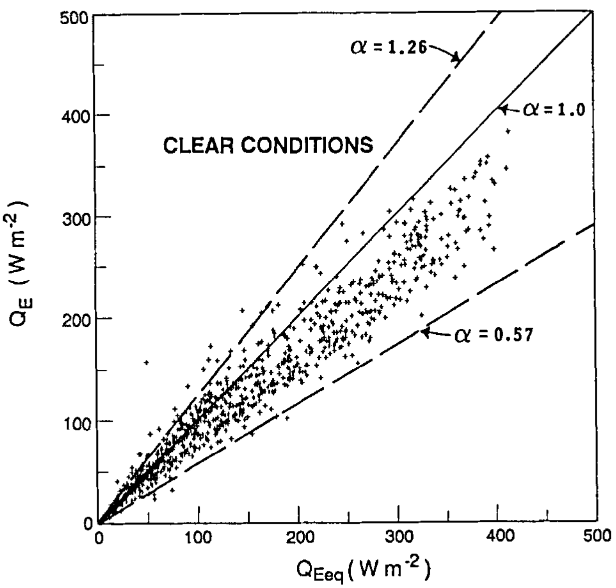
The role of resistance on evaporation from the surface is shown in Figure 5. On 16 June, the relatively high *VPD* is reflected in the climatological resistance (r_i) term, which was larger than on foggy days. In contrast, on the morning of 28 June, and on all of 29 June, there was little or no *VPD* because of fog, which pushed r_i toward zero (Figures 5b and c). Thus, although r_i was generally small, periods of higher r_i were associated with higher Q_E.

Evaluation of aerodynamic resistance (r_a) and hence bulk surface resistance (r_s) normally require adjustment for stability conditions as indicated by the Richardson number (Oke, 1978). Richardson numbers could not be determined with the instrumental setup available. However, stability could be ignored on 16 June because of high average windspeed (9.9 m s^{-1}), and on 28 and 29 June, because of overcast and foggy skies and a moderate airflow (4.1 and 5.6 m s^{-1}, respectively). The dominance of latent heat on the latter two days (Figure 4) also minimized surface heating, hence thermal buoyancy effects. Aerodynamic resistance (r_a) was generally larger than r_i, but was moderated by the high wind conditions typical of this site. r_a was most important during periods of fog, when r_i and r_s were negligible. Comparison of evaporation during clear and fog periods demonstrates the importance of surface moisture, which becomes limiting on evaporating moss surfaces. On clear days (e.g., 16 June) when the surface dried, the surface resistance (r_s) dominated (Figure 5a). Warm temperatures (e.g., 28 June) produced r_s values exceeding 100 s m^{-1}. Under these conditions, there can be little capillary rise of water in the dry surface mosses, so moisture transport therein occurs by vapour diffusion. In contrast, there was effectively zero surface resistance during fog periods (Figure 5b and c), but this increased rapidly as the surface dried (i.e., 29 June in the afternoon).

The high surface resistance (e.g., Figure 5a) depresses Q_E below the equilibrium rate (Figure 6a). The assumption that $\alpha = 1.26$ on wetland surfaces is clearly not applicable when the surface is moss. Figure 6a indicates that most of the values are between $\alpha = 1.26$ and 0.57 during clear periods, tending toward lower α values during higher Q_E (i.e., warmer) periods when capillary rise cannot match evaporative loss. The overestimate by the equilibrium model for dry moss surfaces is reflected in the low average α (0.87). In contrast, fog drip moistens the surface, which decreases r_s to zero (Figure 5c), and with essentially no *VPD*, the Penman–Monteith model, Equation (5), collapses to equilibrium evaporation (Figure 6b). The good fit of Q_E to Q_{Eeq} during fog is demonstrated by the average evaporability parameter α (0.99) and the low scatter (Figure 6b). The close fit between Q_{Eeq} and Q_E during fog is potentially useful for estimating evaporation during periods when evaporation is most difficult to measure, and confirms the assumption that equilibrium conditions hold during rain and fog.

6. Conclusions

Between 26 May and 11 July 1989, the average daily evaporation from the blanket bog near Cape Race, Newfoundland, was 1.7 mm. During clear periods, it averaged 2.5 mm, compared to 1.1 and 0.7 mm during fog and rain, which occurred 41.4 and 11.3% of the time, respectively. The error associated with these values depends on the accuracy of available energy and Bowen ratio measurements. Assuming Q^* is accurate to within ±5% (Latimer, 1972), Q_G is within ±20% (including spatial variability), and given that Q_G is 10 and 14% of Q^* during daytime clear and fog periods, respectively, the error in available energy is 8 and 9%, respectively (Angus and Watts, 1984). Under moist conditions, the Bowen



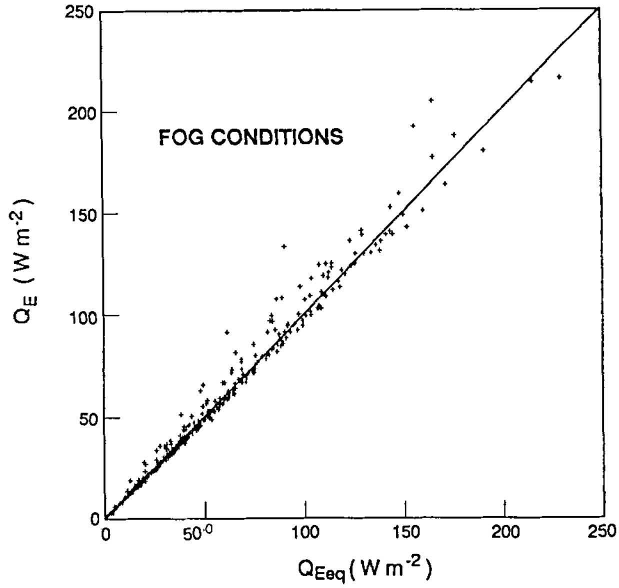


Fig. 6. Actual evaporation versus equilibrium evaporation during clear (top) and fog conditions (bottom).

ratio is small, so the significance of errors in the gradient measurements is minimized (Fuchs and Tanner, 1970). Assuming that the Bowen ratio was measured to within ±30%, and considering the average energy and β conditions presented previously, the total error in measuring Q_E is approximately 16 and 20% for clear and foggy conditions, respectively.

The magnitude of evaporation reported here is low compared to other wetland sites. A concurrent peatland study 900 km north of this site (Price *et al.*, 1991) found a range of evaporation 0.6 to 4.5 mm d^{-1} for an average of 2.5 mm d^{-1}. An earlier study of an arctic fen by Roulet and Woo (1986) yielded values as high as 7.3 mm, and a seasonal average of 4.5 mm d^{-1}. There are two important reasons for the low vapour fluxes reported here, relating to (1) the role of advection fog, and (2) the inability of the moss to supply the surface with moisture.

Depression of evaporation occurs when the characteristics of the airflow cause the evaporation rate to fall. This is not unexpected in a maritime environment during onshore winds because the airstream is adjusted to the offshore environment, which during summer is cooler than the land. Such conditions enhance the temperature gradient and limit the vapour pressure gradient, favouring sensible heat transport over evaporation. However, even though sensible heat flux dominated over latent heat under these conditions, the latent flux was relatively large because of the high available energy. In contrast, warm onshore airflows which produced more fog (Price, 1991) caused evaporation to decrease.

During fog conditions, evaporation occurred at the equilibrium rate ($\alpha = 0.99$) because the vapour pressure deficit was nil. The good relationship between actual and equilibrium evaporation during fog (Figure 6b) indicates that the latter can provide a much simpler approach to estimating evaporation. Similar results were reported by Miranda *et al.* (1984) for a heath surface on wet days, which evaporated at the equilibrium rate. Evaporation at the potential rate ($\alpha = 1.26$) was infrequent here, and was restricted mainly to periods of low radiant energy in periods following fog, when the surface was wet. During clear periods when evaporation dries the surface, α drops to well below the equilibrium rate, because of the high surface resistance of drying mosses. Daily evaporation on clear days is therefore more variable, and requires a better understanding of processes operating below the surface.

Because of the high frequency of fog, and the moderately high aerodynamic resistance here, the Penman–Monteith model indicates that the evaporation from this site is dominated by the available energy, and this is confirmed by the low climatological resistance (r_i). This is also reflected in the excellent linear relationship ($r^2 = 0.97$) between available energy and evaporation, which could be used as an operational model to estimate evaporation at this site. Nevertheless, the relationship between evaporation and available energy is more complex because the latter is strongly affected by atmospheric conditions (i.e., fog).

The advective conditions have important implications for the climate, hydrology, ecology, and morphology of this environment. The frequent fog actually provides a significant source of water to the system (Price, 1991), and simultaneously curtails the evaporative loss; thus the water table is maintained at a higher level than it otherwise would have. This provides the requisite condition for peat development. Here, it is manifest in the extreme, and maintains an adequate water supply at or near sloping and upland surfaces for long enough for blanket bog to develop.

Acknowledgements

This project was funded by Operating and Equipment grants from the Natural Science and Engineering Research Council (NSERC) of Canada. The field assistance of John Crant is appreciated, and the helpful discussions with Dr. Peter Lafleur are gratefully acknowledged. Dr. Lafleur also generously provided the computer routine which determined the Bowen ratios.

References

Angus, D. E. and Watts, P. J.: 1984, 'Evapotranspiration – How Good is the Bowen Ratio System', *Agric. Water Mgmt.* **8**, 133–150.

Banfield, C. E.: 1981, 'The Climatic Environment of Newfoundland', in A. G. MacPherson, and J. B. MacPherson (eds.), *The Natural Environment of Newfoundland, Past and Present*, Memorial University of Newfoundland, St John's, pp. 83–153.

Brutsaert, W.: 1982, *Evaporation Into the Atmosphere*, D. Reidel, London, 299 pp.

Davis, A. M.: 1984, 'Ombrotrophic Peatlands in Newfoundland and Canada: Their Origins, Development and Trans-Atlantic Affinities', *Chem. Geol.* **44**, 287–309.

Farmer, G. H.: 1981, 'The Cold Ocean Environment of Newfoundland', in A. G. MacPherson and J. B. MacPherson (eds.), *The Natural Environment of Newfoundland, Past and Present*, Memorial University of Newfoundland, St John's, pp. 56–75.

Fuchs, M. and Tanner, C. B.: 1970, 'Error Analysis of Bowen Ratios Measured by Differential Psychrometry', *Agr. Meteorol.* **7**, 329–334.

Goodman, J.: 1985, 'The Collection of Fog Drip', *Water Resour. Res.* **21**, 392–394.

Ingram, H. A. P.: 1983, 'Hydrology', in A. J. P. Gore (ed.), *Mires: Swamp, Bog Fen and Moor*, Elsevier Scientific, 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. Geographer* **34**, 79–88.

Lafleur, P. M. and Rouse, W. R.: 1988, 'The Influence of Surface Cover and Climate on Energy Partitioning and Evaporation in a Subarctic Wetland', *Boundary-Layer Meteorol.* **44**, 327–347.

Latimer, J. R.: 1972, 'Radiation Measurement. International Field Year for the Great Lakes', Technical Manual Series No. 2, CNC-IHD, National Research Council of Canada, 53 p.

Miranda, A. C., Jarvis, P. G., and Grace, J.: 1984, 'Transpiration and Evaporation from Heather Moorland', *Boundary-Layer Meteorol.* **28**, 227–243.

Monteith, J. L.: 1965, 'Evaporation and Environment', *Symp. Soc. Exp. Biol., XIX*, Cambridge University Press, London, pp. 205–304.

Nichols, D. S. and Brown, J. M.: 1980, 'Evaporation from a Sphagnum Moss Surface', *J. Hydrol.* **48**, 289–302.

Northland Associates Ltd.: 1988, 'Research on Bog Preparation and Peat Sod Drying Techniques in Newfoundland (1986–1988)', Energy, Mines and Resources (Canada), 161 pp.

Price, J. S.: 1991, 'The Occurrence and Accumulation of Fog Deposits on Blanket Bog Near Cape Race, Newfoundland', accepted with revision, *J. Hydrol.*

Price, J. S., Maloney, D. A., and Downey, F. G.: 1991, 'Peatlands of the Lake Mellville Coastal Plain, Labrador', *Proc. Northern Hydrol. Symp.*, Saskatoon, July 1990 (in press).

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.

Romanov, V. V.: 1968, *Hydrophysics of Bogs*, Israel Program for Scientific Translations, Jerusalem, 299 p.

Roulet, N. T. and Woo, M. K.: 1986, 'Wetland Evaporation in the Low Arctic', *Arctic and Alpine Res.* **18**, 195–200.

Rouse, W. R. and Bello, R. L.: 1985, 'Impact of Hudson Bay on the Energy Balance of the Hudson Bay Lowlands and the Potential for Climatic Modification', *Atmosphere-Ocean* **23**, 375–392.

Rouse, W. R., Hardill, S. G., and Lafleur, P.: 1987, 'The Energy Balance in the Coastal Environment of James Bay and Hudson Bay During the Growing Season', *J. Climatol.* **7**, 165–179.

Rouse, W. R., Mills, P. F., and Stewart, R. B.: 1977, 'Evaporation in High Latitudes', *Water Resour. Res.* **13**, 909–914.

Stewart, R. B. and Rouse, W. R.: 1976, 'Simple Models for Calculating Evaporation From Dry and Wet Surfaces', *Arctic and Alpine Res.* **8**, 263–274.

Thom, A. S.: 1975, 'Momentum, Mass, and Heat Exchange of Plant Communities', in J. L. Monteith (ed.), *Vegetation and the Atmosphere*, Vol. 1, Academic Press, New York, pp. 57–110.

Wells, E. D.: 1976, 'A Classification of Peatlands in Eastern Newfoundland', M.Sc. Thesis, Memorial University of Newfoundland, 201 pp.

Williams, G. P.: 1970, 'The Thermal Regime of a Sphagnum Peat Bog', *Proc. 3rd Int. Peat Congr.*, Quebec, 1968, pp. 195–200.

Wilson, R. G. and Rouse, W. R.: 1972, 'Moisture and Temperature Limits of the Equilibrium Evapotranspiration Model', *J. Appl. Meteorol.* **11**, 436–442.