

Evapotranspiration from a lakeshore *Typha* marsh on Lake Ontario

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Abstract

Evapotranspiration (E_t) was measured over a *Typha* marsh bordering Lake Ontario, and open water evaporation was estimated from a pond ($E_{o(\text{POND})}$) within the marsh system, and from Lake Ontario ($E_{o(\text{LAKE})}$). Evapotranspiration from the marsh averaged 4.8 mm day⁻¹, whereas open water evaporation from the pond was estimated to be 4.9 mm day⁻¹, and from Lake Ontario between 0.4 and 1.3 mm day⁻¹. The ratio of $E_t/E_{o(\text{POND})} = 0.97$ ($r^2 = 0.96$) indicates that E_t was very close to the potential rate, as defined by open water evaporation. The ratio $E_t/E_{o(\text{LAKE})}$ ranged between 3.7 and 12.5. This wide range reflects uncertainty in the net radiation over the lake. Nevertheless, evaporation from Lake Ontario was low, owing to the very large heat storage, which was 76–92% of net radiation. The average Bowen ratio measured at the marsh was 0.17, and canopy and aerodynamic resistance were 74 s m⁻¹ and 31 s m⁻¹, respectively. Aerodynamic resistance over open water was estimated to have an average value of 201 s m⁻¹. Although there were different resistance networks over the *Typha* and the pond surfaces, the end result is that similar values of latent heat flux were produced.

1. Introduction

Evapotranspiration from lakeshore marshes is an important component of the hydrological and ecological regime of both lake and wetland. Atmospheric water exchanges affect marsh water levels, which determine the strength and direction of lateral water fluxes between lakes and marshes, and hence the exchange of nutrients and contaminants.

Research on the efficiency of evapotranspiration (E_t) relative to evaporation (E_o) from adjacent open water has provided a wide range of E_t/E_o ratios. Differences in E_t/E_o may be caused by local influences such as vegetation types (Brezny

et al., 1973), density (Idso and Anderson, 1988) and stage of growth (Eisenlohr, 1966), and advective effects (Linacre et al., 1970). Experimental results based on lysimeters typically exhibit $E_t/E_o > 1$ (e.g. Brezny et al., 1973; Van der Weert and Kamerling, 1974; Bernatowicz et al., 1976; Anderson and Idso, 1987), but this method overestimates E_t because of advective effects and inadequately equilibrated water levels (Priban and Ondok, 1985). Estimates of E_t based on energy balance and aerodynamic methods typically indicate $E_t/E_o \leq 1$ (e.g. Eisenlohr, 1966; Rijks, 1969; Linacre et al., 1970; Lafleur, 1990). The latter methods may be less influenced by local advection, as care is normally taken to ensure adequate fetch. However, all methods may be subject to errors in the estimation of E_o (Granger, 1989).

The processes governing evapotranspiration from wetland and open water environments have some notable differences. A large open water surface (i.e. without emergent vegetation) may experience larger daytime net radiation than vegetated systems, owing to its lower albedo and lower surface temperature (Oke, 1978). However, the energy available for evaporation may be offset by convective mixing in the upper layer of water, which rapidly carries away heat from the surface. In wetland systems dominated by tall emergent vegetation such as *Typha*, *Scirpus*, etc., direct evaporation from the surface is limited by shading and sheltering within the lower part of the canopy, and transpiration is limited by stomatal effects forming canopy resistance. However, compared with open water, the aerodynamically rougher vegetation favours larger convective fluxes, more efficiently carrying away latent heat. Thus, there are differences in energy status and the resistance network for open water and vegetated wetland systems.

The net effects of these complex and interrelated processes are poorly understood, and have led to confusion and contradictory interpretations regarding the efficiency of evapotranspiration from wetland systems. Therefore, the objectives of this paper are (1) to determine the relative evapotranspiration from a *Typha*-dominated marsh, and adjacent open water systems, and (2) to improve our understanding of evapotranspiration from this type of wetland system through an examination of the resistance networks.

2. Study area

The study was located on the southeast shore of Wolfe Island (44°12'N, 76°22'W) (Fig. 1). The area is underlain by Ordovician limestone, with approximately 2 m of laminated clay originating from glacial lake Iroquois (Dalrymple and Carey, 1990). Poorly decomposed peat overlies the mineral sediments, ranging from 0 m at the interface with the mineral soil upland to 5 m at the Lake Ontario margin. Along the landward periphery of the marsh, a shallow (less than 1 m depth) pond with surface area of approximately 0.3 ha is connected to Lake Ontario by a shore lagoon and channel, which is typical of this 160 ha wetland complex located at the head of semi-enclosed Bayfield Bay (Fig. 1). There is at least 750 m fetch in the direction of the prevailing south and southwesterly winds

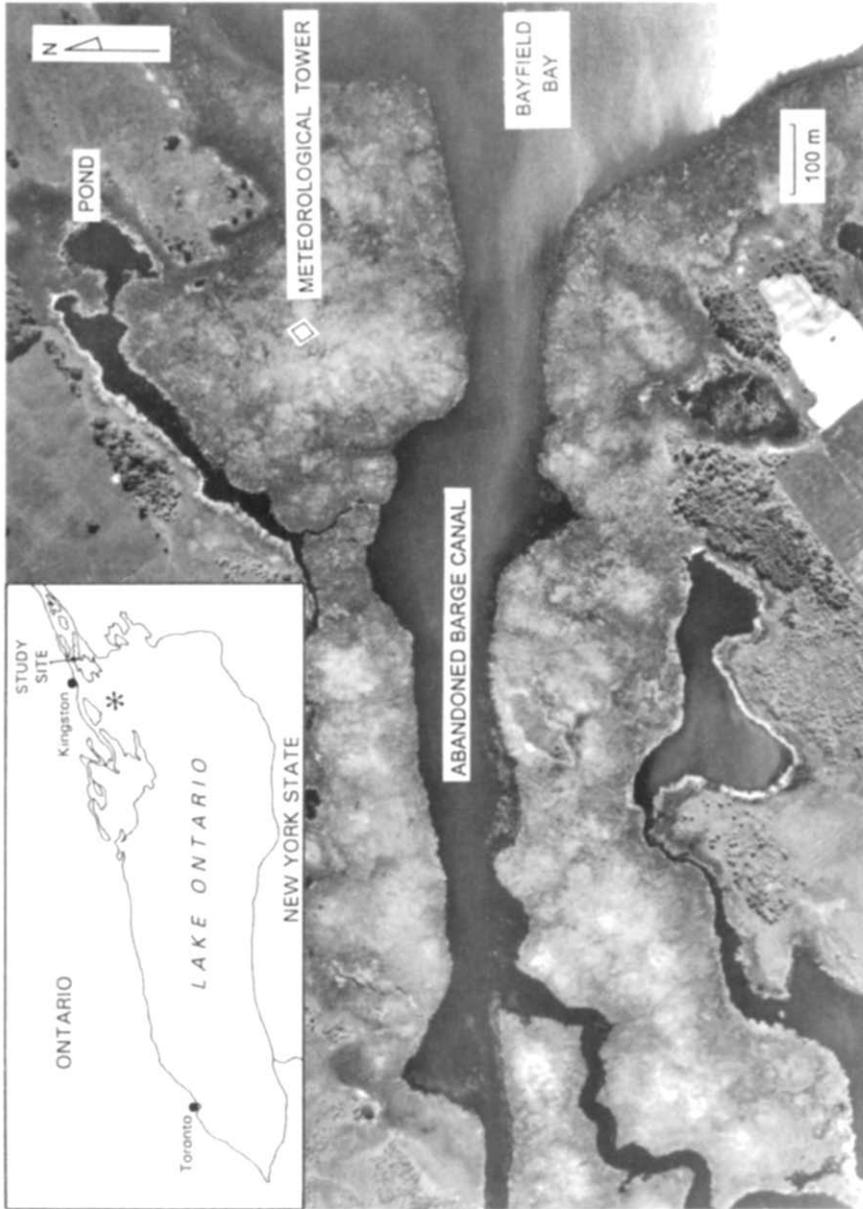


Fig. 1. Study location map. The Lake Ontario water temperature profiles were measured at the point indicated with an asterisk.

(Environment Canada, 1991), although this is interrupted by an abandoned barge canal (approximately 200 m distant from the tower at its nearest point). Minimum fetch is 200 m to the north and east. Prevailing winds in summer are from the southwest over Lake Ontario, but pass over rolling agricultural lands, mostly pasture, before reaching the marsh complex. The marsh is dominated by *Typha latifolia* L., *T. angustifolia* L., and hybrid *T. glauca* Godr.

3. Methods

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

$$Q^* = Q_H + Q_E + Q_G \quad (1)$$

where the terms are net radiation, and 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

$$\beta = \frac{Q_H}{Q_E} = \gamma \frac{\sigma T_a}{\sigma e} \quad (2)$$

where β is Bowen's ratio and γ is the psychrometric constant. Rearranging Eq. (1), then substituting Eq. (2), we obtain

$$E = \frac{Q_E}{L_V} = \frac{Q^* - Q_G}{1 + \beta} \quad (3)$$

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

Penman (1948) described evaporation from open water with an equation of the form

$$Q_{E_o} = \frac{S(Q^* - Q_G) + \frac{C_p \text{VPD}}{r_a}}{S + \gamma} \quad (4)$$

in which Q_{E_o} is the latent heat flux from open water, S is the slope of the vapour pressure–temperature curve at the ambient air temperature, C_p is the heat capacity of air, VPD is the vapour pressure deficit, and r_a is the aerodynamic resistance. 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),

$$r_a = \frac{[\ln(z-d)/z_o]^2}{k^2 U_z} \quad (5)$$

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

Eq. (4) was modified for vegetated systems by Monteith (1965), who incorporated a term representing the canopy resistance (r_c), such that

$$Q_{E_o} = \frac{S(Q^* - Q_G) + C_p \text{VPD}/r_a}{S + \gamma(1 + r_c/r_a)} \quad (6)$$

Canopy resistance (r_c) is closely related to the resistance produced by individual leaves within the canopy (Monteith, 1965), and can be computed from Eq. (6) when the other variables are all known.

The study was done between 14 June and 6 August 1991. Evaporation was determined using the Bowen ratio–energy balance method. Q^* was measured directly with a Swissteco net radiometer calibrated before installation. When the water table was above the surface (15 May–5 July) Q_G was determined from thermocouples placed at 10, 1, 0, –1, –10 and –50 cm from the surface, based on the relationship

$$Q_G = C_w \frac{\Delta T_w}{\Delta t} h + C_p \sum_{i=1}^n \frac{\Delta T_s}{\Delta t} z_i \quad (7)$$

where T_w and T_s are the temperature of the water and soil, respectively, h is the depth of water above the surface, z_i is the thickness of the layer of soil represented by each sensor, and

$$C_p = C_o X_o + C_w X_w \quad (8)$$

where C_w and C_o , and X_w and X_o are the heat capacity and volumetric fraction of water and organic material, respectively. Water table elevation was measured with a continuously recording well near the tower. When the water table was below the surface, two Radiation Energy Balance Systems heat flux plates embedded 1 cm below the surface were used to measure ground heat flux.

Temperature and vapour pressure were measured with a thermocouple and psychrometer system installed on a tower at 2.7, 3.7, and 4.7 m above the peat surface. Dry-bulb (T_a) and wet-bulb temperature (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 that was aspirated with an electric fan.

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. The Bowen ratio was determined only for the half-hour periods between 08:00 and 18:00 h because Bowen ratios calculated from the small temperature and vapour pressure gradients during the night-time were frequently rejected. A wind profile system with anemometers at 3.07, 5.92, and 6.20 m above the peat surface was employed to find the surface roughness parameter z_o . Rain was recorded with a tipping bucket rain gauge, and water level was recorded continuously in a well 15 m from the tower.

Water temperatures on the marsh (64 samples) and in the pond–lagoon system (16 samples) were taken on 14 June and 1 August. Temperature profiles were

measured in Lake Ontario 35 km southwest of the study site. Measurements were made at nine depths in 38 m of water on 18 June and again on 23 July 1991 (unpublished data from the Inland Waters Directorate of Environment Canada).

4. Results

For the period June–August 1991, average temperature and total rainfall measured at the Kingston, Ontario, weather station 10 km to the northwest were 20.1°C and 193 mm, compared with 30 year mean values of 18.7°C and 168 mm, respectively (Environment Canada, 1982, 1991).

The average energy fluxes over the marsh at half-hour intervals are shown in Fig. 2, and average daytime values of energy fluxes, air temperature, VPD, and β are given in Table 1. The energy fluxes peaked in early afternoon, following the general trend of Q^* . The Bowen ratio (Q_H/Q_E) was very low, and rose from 0.03 at 07:00 h, peaked at 0.35 in mid-morning (10:30 h), then declined steadily to 0.12 at 18:00 h (Fig. 3a). The VPD rose steadily through the morning, and peak values of approximately 1.3 kPa were sustained into late afternoon (Fig. 3b). Aerodynamic resistance (r_a) over the *Typha* canopy was high in the early morning, but dropped rapidly as the wind velocity increased (at about 08:00 h), then decreased slowly from 27 s m⁻¹ to 20 s m⁻¹ over the day (Fig. 3 and Table 1). Canopy resistance (r_c) was low in the morning, and increased steadily over the day. Thus, at about 08:00 h r_a and r_c were similar, but diverged so that canopy resistance was about five times greater than aerodynamic resistance over the canopy at 16:00 h.

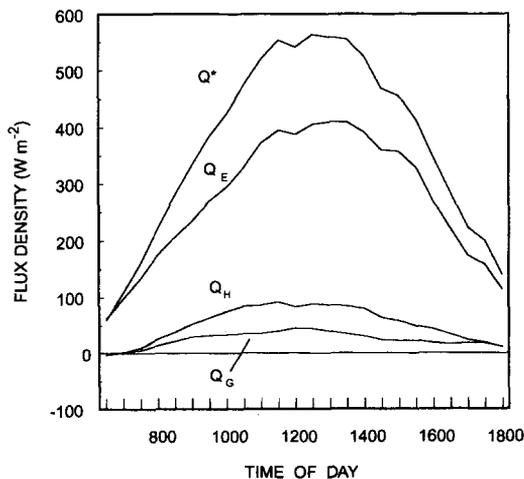


Fig. 2. Averages of the half-hourly measurements of net radiation (Q^*), and latent (Q_E), sensible (Q_H) and ground (Q_G) heat flux. The energy balance components were measured over the *Typha* canopy between 14 June and 6 August 1991.

Table 1

Mean and standard deviation of half-hourly values of net radiation (Q^*), latent heat flux (Q_E), sensible heat flux (Q_H), ground heat flux (Q_G), dry-bulb temperature (T_D), Bowen ratio (β), alpha (α), canopy resistance (r_c), atmospheric resistance over the canopy (r_a), and atmospheric resistance over water (r_{aw})

	Time of measurement			
	06:00–18:00 h	09:00 h	12:00 h	15:00 h
Q^* ($W m^{-2}$)	355.1 ± 100.4	331.2 ± 116.9	538.8 ± 169.6	420.8 ± 162.6
Q_E ($W m^{-2}$)	276.0 ± 80.4	254.7 ± 95.8	404.5 ± 118.1	337.3 ± 138.6
Q_H ($W m^{-2}$)	57.7 ± 26.4	49.2 ± 40.8	88.4 ± 59.7	20.7 ± 42.7
Q_G ($W m^{-2}$)	21.0 ± 9.0	18.4 ± 19.7	38.0 ± 18.2	22.9 ± 17.7
T_D ($^{\circ}C$)	22.8 ± 2.3	21.3 ± 2.3	24.0 ± 2.4	24.2 ± 2.5
VPD (kPa)	1.01 ± 0.34	0.71 ± 0.34	1.22 ± 0.42	1.29 ± 0.51
β	0.17 ± 0.09	0.22 ± 0.17	0.28 ± 0.14	0.18 ± 0.11
r_c ($s m^{-1}$)	74.4 ± 30.0	41.4 ± 25.2	51.4 ± 21.9	80.7 ± 63.6
r_a ($s m^{-1}$)	30.5 ± 10.7	26.0 ± 14.4	22.7 ± 15.2	20.6 ± 10.3
r_{aw} ($s m^{-1}$)	201.5 ± 122.4	216.8 ± 170.7	157.2 ± 93.5	156.3 ± 121.6
E/E_o	1.0 ± 0.1	1.08 ± 0.21	0.96 ± 0.09	0.95 ± 0.10
n	2015	44	43	43

n , Number of samples. Measurement period was 14 June–6 August 1991.

Open water, or potential evaporation, was estimated with Eq. (4) for a shallow pond within the marsh ($E_{o(POND)}$), and for Lake Ontario ($E_{o(LAKE)}$), assuming $Z_o = 0.003$ m for open water (Lafleur and Roulet, 1992). Energy balance components were not measured over the pond, so energy and humidity gradients measured over the *Typha* canopy were used. These assumptions are considered in the Discussion. The latent energy flux over the *Typha* (Q_{E_t}) was closely related to that from the pond ($Q_{E_o(POND)}$), such that $Q_{E_t}/Q_{E_o(POND)} = E_t/E_{o(POND)} = 0.97$ ($r^2 = 0.96$) (Fig. 4). The average evapotranspiration was 4.8 mm day^{-1} , compared with average evaporation of 4.9 mm day^{-1} .

An estimate of open water evaporation was also made for Lake Ontario at the location indicated in Fig. 1 (inset), for the period 18 June–23 July 1991. During this period the average water temperature increased $3.05^{\circ}C$, over the 38 m depth, equivalent to a heat sink averaging 156 W m^{-2} . Based upon 24 h averages of Q^* (170 W m^{-2}), T ($21.3^{\circ}C$), VPD (0.61 kPa) and windspeed measured 6.2 m above the marsh (2.28 m s^{-1}) for the same period, $Q_{E_o(LAKE)} = 12 \text{ W m}^{-2}$. This represents an evaporation rate of only 0.4 mm day^{-1} , and so $E_t/E_{o(LAKE)} = 12.5$.

5. Discussion

The small magnitude of the Bowen ratio (Fig. 3) indicates the dominance of the latent heat flux in the partition of energy from the *Typha*-dominated wetland. The low Bowen ratio persisted under all weather conditions, and during periods of low water table, such as later in the summer. The daily trend of latent and

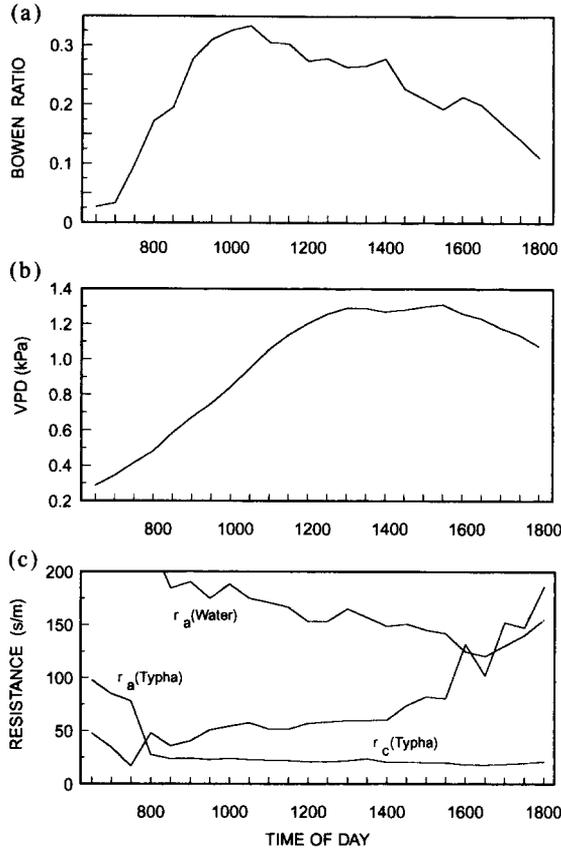


Fig. 3. Averages of the half-hourly measurements of (a) the Bowen ratio; (b) vapour pressure deficit; (c) canopy resistance (r_c) and aerodynamic resistance (r_a). Values based on measurements between 14 June and 6 August 1991.

sensible heat flux was asymmetrical compared with Q^* , such that the Bowen ratio peaked in mid-morning. The canopy was frequently wet with dew in the early morning, and most available energy was used to evaporate this water. As the supply of intercepted water (dew) became exhausted, the canopy heated, and the Bowen ratio peaked (Fig. 3a). As the windspeed and VPD increased over the day, and as transpiration cooled the canopy, the Bowen ratio declined. During the early morning, when the canopy was typically wet, the canopy resistance (r_c) was small, but atmospheric resistance (r_a) was large owing to low windspeeds. The canopy resistance (r_c) continued to increase over the day, but was partly offset by a reduction in atmospheric resistance (r_a) as windspeed picked up. The increase in r_c over the day made it the dominant resistance — a resistance never experienced by open water.

Evaporation from the pond ($E_{O(\text{POND})}$) was estimated on the basis of energy and VPD measurements over the canopy, and with roughness length $z_o = 0.003$

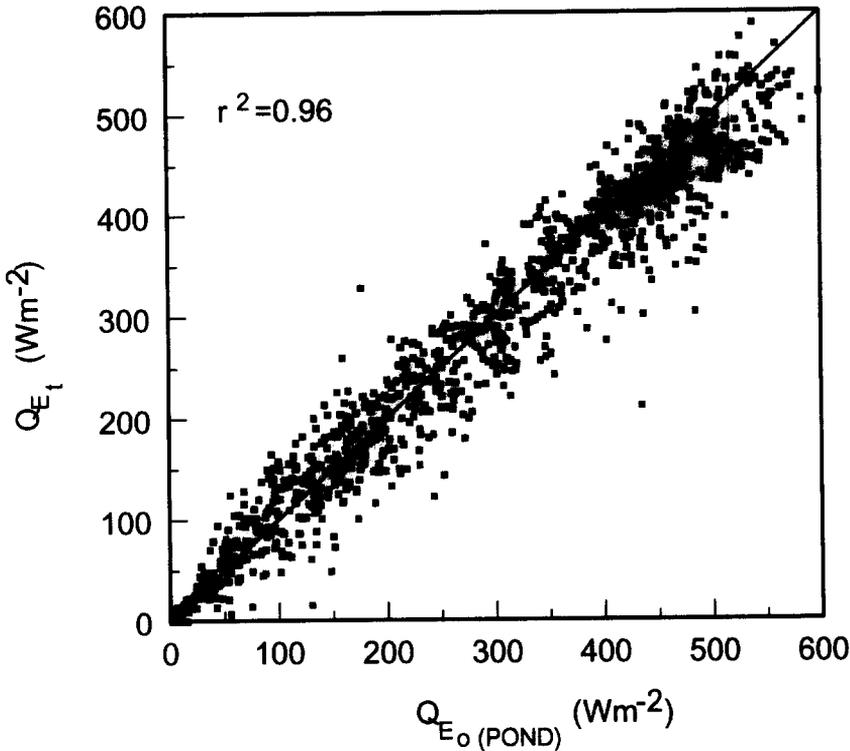


Fig. 4. Latent heat of evapotranspiration (Q_{E_t}) vs the latent heat flux over open water ($Q_{E_o(POND)}$).

m. The similarity of air temperature, VPD, and windspeed is justified on the basis of the small area of the pond, over which there is too little distance for boundary layer adjustment to occur. The available energy was probably not very different from that measured beneath the canopy, as water temperatures measured in the pond and on the marsh surface (14 June and 1 August) were not significantly different ($\alpha=0.05$). Evapotranspiration from the marsh paralleled open water evaporation from the pond ($E_t/E_{o(POND)}=0.97$). The equivalent latent heat flux occurred in spite of high aerodynamic resistance over the water (Table 1), because water experiences no canopy resistance. Although there is a significant canopy resistance associated with *Typha* canopies, a feature confirmed directly by Anderson and Idso (1987), the lower atmospheric resistance was a compensating factor. As E_t was at the potential rate (E_o) as defined by Penman's (1948) formula, this demonstrates that significant canopy resistance does not preclude evapotranspiration at the potential rate.

The initial estimate of evaporation from Lake Ontario ($E_{o(LAKE)}$) was made using net radiation, temperature, and VPD measured over the marsh, but substituting heat storage (Q_s) in the water column for Q_G in Eq. (4). The average lake surface temperature for the period 18 June–23 July was similar to that measured

over the marsh during this period. VPD over the lake was probably smaller than the measured value, but windspeed higher, thus partially compensating for each other in terms of their effect on $E_{o(LAKE)}$. Based on these assumptions, $Q_{E_{o(LAKE)}} = 12 \text{ W m}^{-2}$, ($E_{o(LAKE)} = 0.4 \text{ mm day}^{-1}$). However, Q^* was probably underestimated. If Q^* over water is assumed to be 20% higher than that measured over the reeds, the corresponding latent energy flux (from Eq. (4)) was 36 W m^{-2} (1.3 mm day^{-1}). Therefore, on this basis, the range of estimates of the ratio $E_t/E_{o(LAKE)}$ is 3.7–12.5.

The energy balance derived over Lake Ontario was very different from that of the pond described above, owing to the immense heat storage that occurred in Lake Ontario during this phase of the annual regime. The average heat storage (Q_s) for the period 18 June–23 July (156 W m^{-2}) compares favourably with the June and July averages measured by Bruce and Rodgers (1962, in Miller, 1977), which were 190 W m^{-2} and 130 W m^{-2} , respectively. They found the available energy ($Q^* - Q_s$) at the surface to average 10 and 70 W m^{-2} for these months; most of this presumably went to the latent heat flux. The estimate in this study ($Q^* - Q_s = 14 \text{ W m}^{-2}$; $Q_E = 12 \text{ W m}^{-2}$) is comparatively low, probably because Q^* was underestimated. Adjusting Q^* upward by 20% to account for the lower reflectivity and smaller longwave loss from the cooler water surface results in an estimate of latent energy flux of 36 W m^{-2} , which is comparable with the values published by Bruce and Rodgers (1962, in Miller, 1977). The likely range of $E_t/E_{o(LAKE)}$ (i.e. marsh evapotranspiration/Lake Ontario evaporation) is therefore between 3.7 and 12.5. Clearly, the vapour flux from Lake Ontario is far less than that from the marsh vegetation, or from open water ponds within the marsh. It should be noted that if Lake Ontario evaporation were used to describe potential conditions in the *Typha* canopy, the results would add to the confusion about the role of aquatic macrophytes such as these.

6. Conclusion

The following conclusions have been reached: (1) the relative evapotranspiration from a *Typha*-dominated marsh is approximately equivalent to that of ponds within the marsh system ($E_t/E_{o(POND)} = 0.97$), but is much smaller than from adjacent Lake Ontario ($E_t/E_{o(LAKE)} = 3.7\text{--}12.5$); (2) evapotranspiration from the marsh is at the potential rate, based on Penman's (1948) model; (3) the resistance network maintains a balance between open water evaporation and evapotranspiration from the marsh when the energy regimes are similar. The former system has no canopy resistance, but high aerodynamic resistance as it is smooth. The vegetated system experiences a moderate canopy resistance later in the day, but as it is aerodynamically rougher, aerodynamic resistance is small.

The relatively high evapotranspiration rate from the wetland will lower its water level faster than evaporation lowers the level of Lake Ontario. Depending on the other mechanisms controlling the lake water level, this situation could cause lake-

water to flow into the marsh to replenish the loss, thus influencing marsh chemistry.

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