# Evapotranspiration from a riparian fen wetland

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**Abstract** Evapotranspiration rates were measured in a riparian fen wetland dominated by vascular vegetation and surrounded by open agricultural areas and forests. The wetland is situated on a floodplain in central Denmark. Measurements were taken throughout the growing season (April–September) of 1999.

Evapotranspiration rates were higher than those published for most other wetland types, with an average of 3.6 mm d$^{-1}$ during the growing season and a peak rate of 5.6 mm d$^{-1}$. Daily average evapotranspiration was 110% of Penman's potential open water evaporation.

Evapotranspiration was the dominant sink in the energy balance of the wetland studied. During the day, evapotranspiration accounted for 82% of the available radiant energy, $R_n$. Due to the presence of deposited fine-grained sediments, soil-water availability was kept high at all times which resulted in moderate canopy resistances, $r_c$ (overall mean $= 32$ s m$^{-1}$). Evapotranspiration was controlled by a combination of driving forces: $R_n$, saturation vapour pressure deficit, $D$, and $r_c$.

It is hypothesized that the results presented in this study are conditioned by the proximity of the wetland to drier upland areas. During periods with high evaporative demand and low precipitation, warm, dry air is formed over the upland areas and wetland evapotranspiration rates are enhanced by local advection. Indicative evidence for the hypothesis is presented. Although the absolute magnitude of the results reported is only directly relevant to similar sites in Denmark, the processes and controls described are considered to be representative of riparian wetlands subjected to frequent flooding and/or with a high groundwater table, with vascular vegetation, and which are narrow corridors in open agricultural landscapes.

**Keywords** Bowen ratio; energy balance; evapotranspiration; wetland

## Introduction

The use of constructed and natural wetlands for water pollution control is becoming an accepted technology worldwide (IWA 2000). The Danish Parliament decided in 1998 to restore 16 000 ha of wetlands with the chief purpose of reducing agricultural nitrate loading to the aquatic environment. It is expected that many of these restored wetlands will be located on or along watercourses in riparian areas, which are often narrow corridors in landscapes otherwise dominated by agriculture. The processes of nutrient turnover and retention in a wetland all depend on hydrology, first and foremost due to water being the main dissolving and transporting agent (LaBaugh 1986; Hoffmann 1998). In order to quantify the nutrient balance of a wetland it is therefore necessary to quantify the water balance. Evapotranspiration is a major component of the water balance of wetlands (e.g. Lafleur 1990; Campbell and Williamson 1997) and it is thus important to have insight into the wetland energy balance and the factors governing the process of evapotranspiration.

This paper describes the results of an energy balance study in a Danish vegetated riparian fen. Evapotranspiration was estimated during the growing season (April to September) of 1999 using the Bowen ratio energy balance method. The purpose of the study was to (i) quantify the range in evapotranspiration rates for this wetland type and (ii) identify the major controls on evapotranspiration rates. The processes and controls described may be representative of other riparian wetlands subjected to frequent flooding and/or with a high groundwater table, with vascular vegetation, and located in close proximity to extensive agricultural areas.



## Study site

A riparian fen wetland in the lower River Gjern catchment (114 km$^2$) in Jutland, Denmark (56°22′N, 9°41′E) was selected as study site (Figure 1). The site is located 500 m from the outlet of the catchment in Lake Sminge (0.18 km$^2$). At this place, the floodplain is 250 m wide and, to the north and south, delineated by 7 m high terraced hillslopes. The river having a width of 4 m is bordered by levees. The floodplain is generally flat except for a very gentle slope (0.4%) from the levees toward the hillslopes. The upland area consists of sandy soils dominated by agriculture and smaller plantations of mainly coniferous trees. Upland groundwater depth is generally several metres during summer (Grant *et al.* 1996). The hillslopes are vegetated with *Picea abies* and *Pinus silvestris*. The wetland covers the entire floodplain and consists of poorly drained peat of a depth of approximately 2 m overlying quarternary fluvial deposits of sand. Frequent flooding by the river (e.g. 21 events of an average duration of 4.6 d during the period 1995–1999) has created a compact layer of fine inorganic material mixed with organic matter on top of the peat. This layer is wedge-shaped with a maximum depth of 0.4 m close to the river and is absent as a distinct layer 80 m from the river. Soil types range from sandy loam in the levees (*Gleyic Fluvisol* (FAO 1998)) to clay loam overlying hemic peat and with a fibric A-horizon in the central part (*Histic Gleysol*) to

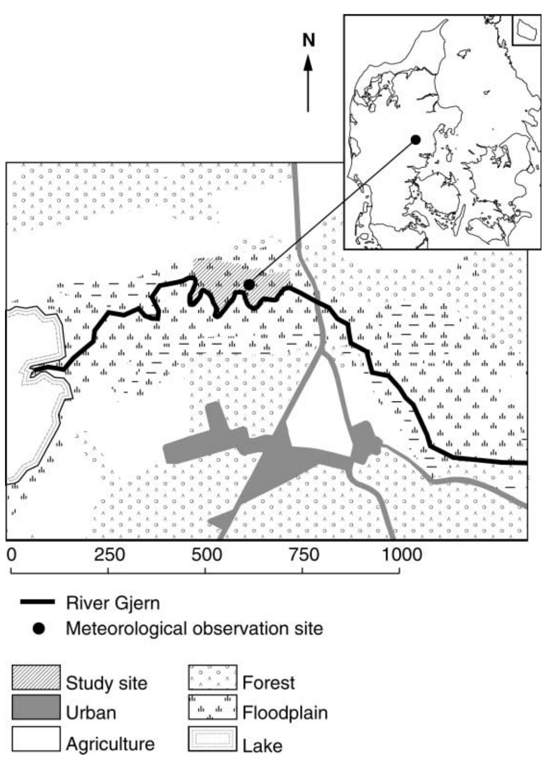


**Figure 1** Location of the study area. The floodplain is bordered by forests and open, cultivated areas. Length scale unit is m

hemic peat overlying fibric peat (*Histic/Fibric Histosol*) in a very wet band of 25–30 m bordering the hillslope (back swamp). The direction of groundwater flow is generally from the hillslope towards the river. During high river water levels the flow was reversed. Groundwater is close to or above the surface most of the year with a maximum depth of 0.5 m in the central part of the wetland in midsummer. The wetland vegetation is governed by the hydric gradient (Schröder *et al.* 2005). The driest parts are dominated by *Alopecurus pratensis* and the central main part by *Deschampsia caespitosa.* The very wet area is vegetated with *Glyceria maxima* followed by *Alnus glutinosa* in the back swamp. The dryer parts of the wetland are usually mowed once a year although this was not done in 1998 and 1999. Species richness ranges from 5–10 species $m^{-2}$ in the driest parts and up to 40 species $m^{-2}$ close to the wet *Glyceria maxima*-dominated zone (Schröder *et al.* 2005). The selected site is fairly typical of riparian areas in the lower parts of Danish catchments, provided the intensity of drainage and river channelization has been low.



## Methods

### Actual evapotranspiration

The Bowen ratio energy balance (BREB) approach was used to estimate evapotranspiration from the wetland. A Bowen tower was erected on the central part of the wetland 50 m from the river. To the north (N) and south (S) fetch, i.e. the upwind distance to a change in vegetation, was 40 m and 125 m, respectively, whereas the fetch was several hundred metres to the east (E) and west (W). As the prevailing wind direction was E–W along the riparian corridor, the requirements for micrometeorological measurements were typically met (fetch:depth of constant flux layer ratio = 70 for the forested hillslope–wetland transition (Monteith and Unsworth 1990)).

According to the BREB method, evapotranspiration is estimated via its energy equivalent, the latent heat flux $\lambda E$ ($\mathrm{W\,m^{-2}}$), within the framework of the surface energy balance:

$$R_n = H + \lambda E + G \quad (\mathrm{W\,m^{-2}}) \tag{1}$$

where $R_n$ is net radiation, $H$ is the sensible heat flux and $G$ is the surface soil heat flux. Minor components of the energy balance, such as photosynthesis and biomass storage, were ignored. The Bowen ratio ($\beta$) describes the ratio of sensible to latent heat fluxes and

$$\beta = H/\lambda E = \gamma \Delta T/\Delta e \tag{2}$$

where $\gamma$ is the psychrometric constant and $\Delta T$ and $\Delta e$ are measured differences in air temperature and vapour pressure, respectively, between two heights above the canopy. $\Delta T$ and $\Delta e$ were measured with the Campbell Scientific Inc. (CSI) Bowen ratio system. The energy balance can thus be rewritten to give the latent heat flux as

$$\lambda E = (R_n - G)/(1 + \beta). \tag{3}$$

Net radiation was measured at a height of 1.5 m (Radiation and Energy Balance Systems (REBS) Q-7 net radiometer). Soil heat flux ($G_{0.08}$) was measured by two soil heat flux plates buried at 0.08 m (REBS). Changes in soil temperature ($\Delta T_s$) above the plates were measured by four parallel thermocouples (CSI). The soil heat flux at the surface was then calculated by adding the measured flux at 0.08 m to the energy stored in the layer above the heat flux plates:

$$G = G_{0.08} + d(\Delta T_s/t)\rho_b(c_s + wc_w) \tag{4}$$

where $d = 0.08$ m is the thickness of the soil layer, $t$ is the time (s) between consecutive measurements of soil temperature, $\rho_b$ is the soil bulk density, $c_s$ is the specific heat of the dry soil material, $c_w$ is the specific heat of water and $w$ is the gravimetric soil moisture content estimated from measurements of soil water pressure potentials (see below).

All the above sensors were operated by a CR10X data logger (CSI) and data were recorded as 20 min averages.

At a climate station close to the river bank precipitation was measured at 1.5 m (Rimco, Rauchfuss Instruments) and wind speed (Vector Instruments), wind direction (Malling Kontrol), air temperature and relative humidity (Vaisala) were measured at 2 m. Precipitation was summed every 10 min. The other variables were 10 min averages. Incident short-wave radiation (Kipp and Zonen, CM5) was recorded as instantaneous values every 10 min. All data were stored on a Mylog datalogger (Dansk Elektronik Design).



#### Groundwater level and soil water pressure potential

Groundwater level was measured weekly by hand along a transect of five screened PVC pipes of 32 mm running from the river bank and 100 m N to the area dominated by *Alnus glutinosa*. A similar transect consisted of six nests, each of 2–4 water-filled tensiometres inserted to depths of 10, 25, 40 and 55 cm. From 7 May to 30 September, the soil water pressure potential head was measured automatically at intervals of 30 min by pressure transducers (RS Components) connected to a Mylog datalogger.

#### Vegetation structure

Prior to the study period, species composition of the vegetation was mapped and a transect traversing all identified vegetation zones was laid from the river bank, extending 100 m to the N. Eight times during the growing season at approximately fortnightly intervals, the canopy height was measured along the transect and biomass was harvested in 7 plots, each measuring $0.25\,\mathrm{m}^2$. The harvested samples were subsequently sorted into dead and live biomass, and were dried for 24 h at 105°C and weighed. Leaf Area Index (LAI) was measured *in situ* using a LAI-2000 Plant Canopy Analyzer (LI-COR Instruments) at approximately the same times as the harvest.

#### Theoretical considerations

In previous studies of wetland evapotranspiration, Penman's potential open water evaporation, $E_0$, has been calculated in order to develop simple predictive models (Koerselman and Beltman 1988; Lafleur 1990; Campbell and Williamson 1997; Souch *et al.* 1998). This use of a simple index across wetland studies facilitates comparisons to be made, thus addressing the inconsistencies in findings noted in reviews by Ingram (1983) and Linacre (1976). $E_0$ was computed from the formula given in Koerselman and Beltman (1988) as follows:

$$E_0 = t\frac{\Delta(R_n - G) + \gamma(3.7 + 4u)D}{\lambda(\Delta + \gamma)} \quad (\mathrm{mm\,s^{-1}}) \tag{5}$$

where $t$ is the time period (s), $\Delta$ is the slope of the saturation vapour pressure versus temperature, $\gamma$ is the psychrometric constant, $u$ is the wind speed at reference height (2 m), $D$ is the saturation deficit of the atmosphere at the reference height and $\lambda$ is the latent heat of vaporization.

The original Penman model (Penman 1948, 1956) was modified by Monteith (1965) to incorporate aerodynamic and surface resistance controls, thus describing evapotranspiration from a vegetated surface in any state of water stress (the Penman–Monteith equation):

$$\lambda E_{\mathrm{PM}} = \frac{\Delta(R_n - G) + \rho_a c_p(D/r_a)}{\Delta + \lambda(1 + r_c/r_a)} \quad (\mathrm{W\,m^{-2}}) \tag{6}$$

where $\rho_a$ is the air density, $c_p$ is the specific heat of air, $r_c$ is the canopy resistance derived from Eq. (6) when the relevant climatological parameters are known and $\lambda E$ is measured, and

$r_a$ is the aerodynamic resistance estimated from measured wind speed at 2 m, roughness length $z_0 = 0.123h$ and zero displacement height $d = 0.667h$, given the measured canopy height, $h$. Following Monteith and Unsworth (1990),

$$r_a = r_b + r_{aM} \tag{7}$$

where $r_b$ is the excess resistance to heat and vapour transfer compared to momentum transfer. $r_b$ is calculated by Thom's (1972) empirical equation, $r_b = 6.2u^{*-0.67}$. $r_{aM} = u/u^{*2}$ is the aerodynamic resistance to momentum transfer from the measurement height to the canopy and $u^*$ $\kappa u/\ln((z-d)/z_0)$ is the friction velocity, where $\kappa = 0.41$ is the von Karman constant.

Penman's potential open water evaporation, $E_0$ (Eq. (5)), as calculated in various studies, cannot, however, be considered a true reference since net radiation has been actually measured at the experimental sites over vegetated surfaces with albedos differing from a free water surface. FAO (Allen *et al.* 1998) has recommended the Penman–Monteith method as a new international standard for reference evapotranspiration ($ET_0$). The FAO Penman–Monteith method defines the reference crop as a hypothetical crop with an assumed height of 0.12 m, with surface resistance of 70 s m$^{-1}$ and an albedo of 0.23. In the present study $ET_0$ is based on calculated values of net radiation in order to get a common reference. The calculation of net radiation used measured values of incident short-wave radiation, air temperature and relative humidity at a reference height (2 m) as prescribed by Allen *et al.* (1998).

To identify the main forces driving $\lambda E$, McNaughton and Jarvis (1983) rewrote Eq. (6) incorporating a 'decoupling coefficient', $\Omega$, $0 < \Omega < 1$:

$$\lambda E_{\mathrm{PM}} = \Omega \frac{\Delta + (R_n - G)}{\Delta + \gamma} + (1 - \Omega) \frac{\rho_a c_\rho D}{\gamma r_c} \quad (\mathrm{W\,m^{-2}}) \tag{8}$$

where $\Omega$ is defined by

$$\Omega = \frac{\Delta + \gamma}{\Delta + \gamma(1 + r_c/r_a)} \tag{9}$$

When $\Omega = 1$, $\lambda E$ is referred to as the equilibrium rate, i.e. the rate of evapotranspiration that would be obtained if the heat budget of a surface was completely dominated by the radiative term. This may happen when short well-watered vegetation is exposed to bright sunshine, humid air and a light wind. The surface is then said to be 'decoupled' from the prevailing weather. When $\Omega = 0$, $\lambda E$ is the 'imposed rate' driven solely by $D$ and $r_c$.

## Results and discussion

### Weather conditions during the study period

The temperature of the growing season of 1999 (April–September) averaging at 13.3°C, was 0.9°C above normal (Figure 2). Especially hot were July and September, at 1.2°C and 3.0°C, respectively, above normal. Precipitation during the period was normal (432 mm) with a dry July (27 mm, 53 mm less than normal), whereas September received 104 mm of rain (normal 83 mm), which almost all fell during the last two weeks of the month. Due to the throughflow of cold groundwater and the higher water content and the consequently higher soil heat capacity, the soil temperature of the wetlands is expected to be lower than that of the mineral upland soils. This leads to reduced outgoing long-wave radiation, resulting in higher net radiation over the wetland. In this study, net radiation made up 59% of total short-wave irradiance (global radiation) during the period April–September 1999 (24-hour values) compared to a normal value (1955–1974) of 49% for the same months measured at Taastrup (The Climate and Water Balance Station, Taastrup, 55°40′N and 12°18′E) over a grass crop (albedo = 0.24) (Jensen 1996). Albedo was not measured in this study but it is assumed to

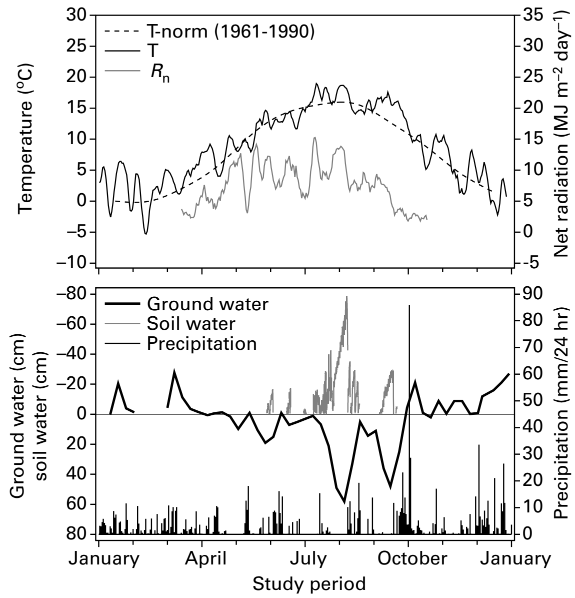


**Figure 2** Upper part: net radiation and air temperature shown as five days' moving averages. Normal temperature (1961 – 1990) for Denmark, monthly values. The study period, April – September 1999, is indicated. Lower part: daily precipitation, groundwater depth (missing values due to ice) and soil water pressure potential head at a depth of 10 cm (only shown when negative)

be below 0.2 (based on a review in Burba *et al.* (1999)). Lower soil temperature and a lower albedo resulted in a 20% higher net radiation on the wetland site than on the grass field.

#### Vegetation structure

Vegetation growth begins in late April (Figure 3). At this time, vegetation cover is very patchy, with LAI ranging from 0 (herb communities) to above 2 (grasses) with a mulch of dead vegetation covering the ground surface in the open spots. With maximum LAI = 5 attained at the beginning of July the wetland vegetation forms a dense cover. Biomass production is high and with a maximum area weighted average of 856 g dry matter m$^{-2}$ comparable to a cultivated grass crop (Andersen 1987). Maximum canopy height is 0.85 m (average for the study site).

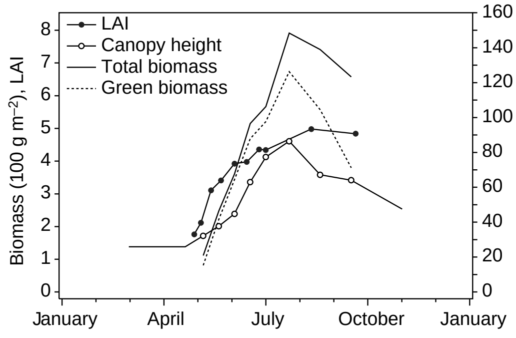


**Figure 3** LAI, canopy height (broken line: extrapolated values) and total and green biomass. All values are area weighted averages of measurements in seven distinct vegetation zones

#### Evapotranspiration

Evapotranspiration was measured during the period April–September 1999. In order to avoid periods when $\beta$ approaches $-1$ and $\lambda E$ estimates approach infinity, only daylight values (global radiation $> 0\,\mathrm{W\,m^{-2}}$) were used for the present analysis. However, from dusk to dawn little radiative energy is available to drive the exchange of latent heat, and the percentage of the daily flux missed is small. Bidlake *et al.* (1996) found for a marsh site the daylight evapotranspiration rate to be 1.7% higher than the daily rate. Negligible vapour pressure gradients were observed during rainy periods, leading to excessively large $\beta$. In early mornings on fine summer days, condensation occurred within intake hoses, leading to errors in the vapour pressure measurement. Data affected by these two types of error were omitted. When ponding water is present, the measurement setup for $G$ underestimates the true value since it does not take into account heat storage in water on top of the soil surface. Flooding periods are consequently omitted. A total of 110 days were left for further analysis. The days rejected accounted for 20.1% of the available energy ($R_n$) during April–September 1999. It was not possible to quantify errors associated with the monthly evapotranspiration rates that were due to incomplete sampling.

The seasonal course of evapotranspiration is presented in Table 1 as monthly mean values. Average for the growing season of 1999 was $3.6\,\mathrm{mm\,d^{-1}}$ with a maximum rate of $5.6\,\mathrm{mm\,d^{-1}}$. These evapotranspiration rates are higher than published values for most other wetlands (Table 2). On average, evapotranspiration is higher than Penman's potential open water evaporation, $E_0$ (Table 1 and Figure 4). On a daily basis, $E$ is well correlated to $E_0$ with $r^2 = 0.94$ (intercept $= 0.31$, slope $= 0.99$). The average ratio $E/E_0 = 1.10$ is considerably higher than for other wetland types (Table 2). Evapotranspiration is also higher than the FAO reference ($ET_0$) - average ratio $= 1.28$. However, $E$ is slightly positively biased relative to $ET_0$, since $ET_0$ is calculated on a daily basis, thus including nighttime where net radiation and evapotranspiration usually are negative. Furthermore, $ET_0$ has been found in a study by Detlefsen and Plauborg (2001) to be 7–13% below the evapotranspiration calculated by the Penman formula, when the Penman formula was fed with data measured above a well-watered crop of short green grass. Under Danish conditions the Penman formula has been found to yield almost identical results as lysimeter measurements of evapotranspiration from short, well-watered grass (Mikkelsen and Olesen 1991). Latent heat flux, $\lambda E$, consumes on average 82% of net radiation, $R_n$, for the growing season of 1999. These ratios are also considerably higher than those measured in most other wetland studies (Table 2), but they are, however, within the published limits and agree well with Oke (1978), who states that for a grass crop not short of water grown at 50°N, latent heat flux consumes 82% of net radiation during summer.

#### Soil water availability

Figure 2 shows that the groundwater table in midsummer may be more than 0.5 m below the surface. However, the soil water pressure potential head never exceeds $-80$ cm, even in periods with a high evaporative demand (in July 1999, for example, the difference between precipitation and evapotranspiration was $-104$ mm). This is due to the capillary characteristics of the fine-grained deposition layer covering the peat in the central wetland. This layer has a clay content of 36% and thus a high air-entry value which creates a wide capillary fringe. Andersen (2004) estimated the thickness of the capillary fringe to be greater than 15 m. This means that the capillary fringe intersects the ground surface with very rapid response to infiltrated precipitation as a result (Figure 2), since only a small fraction of air-filled porosity exists (Gerla 1992). Moreover, the roots of the vascular vegetation (depth of rooting zone $= 0.5$ m) are capable of extracting soil water so as to maintain a very high transpiration rate even though the water table drops. This is in contrast to the situation in

**Table 1** Monthly average daylight sums of energy balance components ($R_n$, $G$, $H$ and $\lambda E$) and evapotranspiration rates (E), including range, and ratios $E/E_0$, $E/ET_0$ and $\lambda E/R_n$. In parentheses is given the number of days selected for analysis

| | April (19) | May (12) | June (17) | July (22) | August (22) | September (18) |
|---|---|---|---|---|---|---|
| $R_n$ (MJ m$^{-2}$ d$^{-1}$) | 8.8 | 12.5 | 12.2 | 12.6 | 10.7 | 7.9 |
| $G$ (MJ m$^{-2}$ d$^{-1}$) | 0.8 | 0.9 | 0.6 | 0.6 | 0.4 | 0.4 |
| $H$ (MJ m$^{-2}$ d$^{-1}$) | 0.8 | 1.5 | 1.7 | 1.7 | 1.5 | 0.9 |
| $\lambda E$ (MJ m$^{-2}$ d$^{-1}$) | 7.3 | 10.0 | 9.9 | 10.3 | 8.8 | 6.6 |
| $E$ (mm d$^{-1}$) | 2.9 (0.8–4.9) | 4.1 (2.1–5.2) | 4.0 (3.1–5.4) | 4.2 (2.5–5.6) | 3.6 (2.3–5.2) | 2.7 (1.2–3.7) |
| $E/E_0$ | 1.21 | 1.11 | 1.15 | 1.06 | 1.02 | 1.07 |
| $E/ET_0$ | 1.41 | 1.29 | 1.27 | 1.20 | 1.18 | 1.30 |
| $\lambda E/R_n$ | 0.82 | 0.80 | 0.81 | 0.82 | 0.82 | 0.84 |

**Table 2** Mean daily evapotranspiration rates including range, $E/E_0$ and $\lambda E/R_n$ reported in wetland studies

| Wetland type | $E$ (mm d$^{-1}$) | $E/E_0$ | $\lambda E/R_n$ | Author |
|---|---|---|---|---|
| Riparian fen wetland, Denmark | 3.6[a] (0.8–5.6) | 1.10 | 0.82 | This study |
| Sub-Arctic coastal wetland | | | | Lafleur (1990) |
| Dry site | 2.6[b] (1.0–4.5) | 0.74 | 0.50 | |
| Wet site | 3.1[b] (1.4–6.0) | 0.90 | 0.57 | |
| Quaking fens, Netherlands | 2.5[c] (1.0–4.1) | 0.77 | | Koerselman and Beltman (1988) |
| Transition peatland, Japan | 2.5[d] (max = 4.6) | | 0.37–0.68[e] | Takagi *et al.* (1999) |
| Open water marsh/sedge meadow, Indiana Dunes National Lakeshore | 3.58[f] | 1.03 | 0.44 | Souch *et al.* (1998) |
| Prairie wetland, Nebraska | 3.75[g] (0.5–6.5) | | 0.8–1.0[h] 0.6–0.8[i] 0.3[j] | Burba *et al.* (1999) |
| Lakeshore *Typha* marsh, Ontario | 4.9[k] (3.5–6.3) | 1.0 | 0.78 | Price (1994) |
| Czechoslovakia | | | | Priban and Ondok (1986) |
| Willow carr | 3.5[l] (2.4–4.9) | | 0.86 | |
| sedge-grass marsh | 3.0[l] (2.3–3.7) | | 0.72 | |
| Raised peat bog, New Zealand | 1.54[m] (max = 2.13) | 0.34 | 0.23 | Campbell and Williamson (1997) |
| Dry canopy | | | | |

[a] Daylight values, April–September
[b] Daylight values, May–August
[c] Daily values, April–October. Average for three vegetation types
[d] Daily values, June–October
[e] Read from Figure 6 in Tagaki *et al.* (1999)
[f] Daylight values, June
[g] Daily values, June–October
[h] Early and peak growth
[i] During senescence
[j] After senescence
[k] 0600–1800 h, June–August
[l] Daily values, July–September. Range of monthly means
[m] Daylight values, November–March

*Sphagnum*-dominated wetlands. Several studies report that the level of the water table affects the evaporative capacity of the *Sphagnum* surface (Ingram 1983; Phersson and Petterson 1997). This is due to the lack of vascular tissue and low matric potentials, which means that much higher surface resistances are imposed as the water table declines below the ground surface.

**Canopy resistance**

The seasonal course of $r_c$ is shown in Figure 5. The daily minimum value usually occurred between 1100–1300 h. $r_c$ varies from 1–85 s m$^{-1}$ with an overall mean of 32 s m$^{-1}$. The lowest values are found in April when the water table is at or very near the ground surface (see Figure 2). The highest values are found at the beginning of August 1999 following a long, dry period coinciding with maximum depth to the water table and maximum saturation vapour pressure deficit. These values are well within the limits reported in the literature for other wetland types (see Table 3 in Campbell and Williamson (1997)), which range from 4 s m$^{-1}$ to 608 s m$^{-1}$, with most values below 100 s m$^{-1}$. The overall median (1000–1800 h) aerodynamic resistance, $r_a$, is 39 s m$^{-1}$.

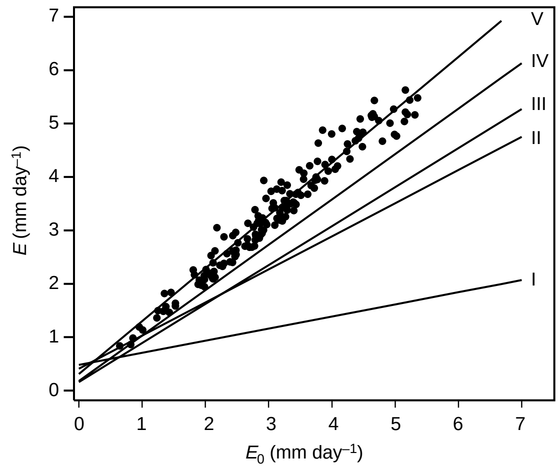


**Figure 4** Measured evapotranspiration, $E$, vs Penman's potential open water evaporation, $E_0$. Lines are: I. Campbell and Williamson (1997), $E = 0.48 + 0.227E_0$, II. Lafleur (1990), dry site, $E = 0.41 + 0.62E_0$, III. Koerselman and Beltman (1988), $E = 0.16 + 0.73E_0$, IV. Lafleur (1990), wet site, $E = 0.18 + 0.85E_0$, V. this study, $E = 0.31 + 0.99E_0$, $r^2 = 0.94$

The implication of increasing $r_c$ is shown in Figure 6. When $r_c$ is less than about 10 s m$^{-1}$, evapotranspiration rates are well above Penman's potential open water evaporation, thus illustrating the aerodynamically rough nature of the wetland vegetation relative to open water. This extra roughness results in more efficient transport of the latent heat flux away from the surface and explains the higher $E/E_0$ in April and June. In April there is still open water in depressions, while in June precipitation at the beginning of the month raises the water table to the ground surface. In April, the average $r_a$ and $r_c$ are 47 and 12 s m$^{-1}$, respectively. The figures for June are 52 and 20 s m$^{-1}$, respectively. The combined resistances are thus considerably lower than the aerodynamic resistance over an extensive open water surface (which does not experience surface resistance). Open water aerodynamic resistance is around 200 s m$^{-1}$ (Priban and Ondok 1986; Price 1994). As $r_c$ is increased to more than 25 s m$^{-1}$ $E/E_0$ is unity or slightly below. Attempts at modelling evapotranspiration rates for this wetland type should therefore not be based on Eq. (5), but on the Penman–Monteith model (Eq. 6) with sub-models for the calculation of $r_a$ and $r_c$.

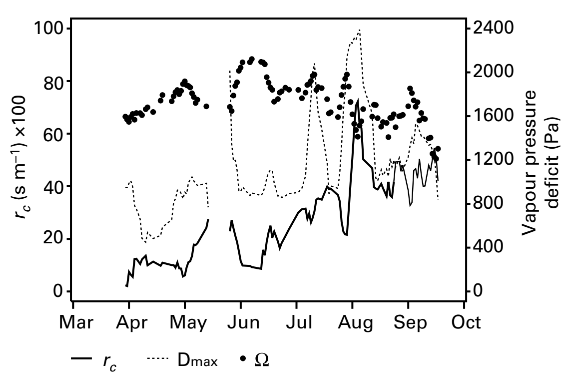


**Figure 5** Canopy resistance, $r_c$ (daily minimum, represented by mean from 1100–1300 h), decoupling coefficient, $\Omega$ (mean of 1000–1800 h) and saturation vapour pressure deficit, $D_{max}$ (daily maximum). All values are five days' moving averages

### Advection

The relationship between $\lambda E/R_n$ (daily daylight values) and $r_c$ (daily minimum represented as the mean of 1100–1300 h) is shown in Figure 7. Contrary to intuitive expectations based on Figure 6, increasing canopy resistance does not affect evapotranspiration efficiency, i.e. the proportion of net radiation used for evapotranspiration. This result is not due to a coincidence between high $r_c$ and low $R_n$, since a regression between $R_n$ and $r_c$ does not yield a slope significantly different from zero. The finding is rather taken as an indication of the effect of advection, $A$. If the advective energy input is included as $\lambda E/(R_n + A)$ and there is a coincidence of increasing $r_c$ and presence of advection, then $\lambda E/(R_n + A)$ will decrease with increasing $r_c$. During periods with high evaporative demand and low precipitation, upland areas having sandy soils and a deep groundwater table become considerably drier than the wetland. With water less available for evapotranspiration, a larger proportion of $R_n$ is transferred as sensible heat flux, resulting in the formation of warm, dry air over the upland areas. The wetland, on the other hand, with its plentiful water supply is cooled by evaporation (Oke 1978). Thus the wetland is cooler than the surrounding air in which it is embedded. Consequently, there is a downward directed flux supplying sensible heat to the wetland surface, thus enhancing evapotranspiration rates (i.e. dry-air advection (McNaughton and Jarvis (1983)). The observed $D$ values are high and quite similar to those measured over extensive areas of agricultural crops grown on sandy soils with a deep groundwater table (comparison with data from Agricultural Research Centre Foulum (56°29′N, 9°35′E) June–September 1999 (Jensen personal communication)). Bowen ratios are low (e.g. daylight average for the hot and dry July of 0.09), indicating a suppression of $\Delta T$ by atmospheric supply of sensible heat. Advection is thought to be local on the scale of the floodplain as opposed to regional advection, since the majority of the wind (>75% in

**Table 3** Energy balance components (MJ $m^{-2} d^{-1}$) and $\lambda E/R_n$ for 6 August 1999 presented in Figure 8

| | $R_n$ | $G$ | $H$ | $\lambda E$ | $\lambda E/R_n$ |
|---|---|---|---|---|---|
| Daylight values (global radiation: $>0$ W $m^{-2} s^{-1}$) | 13.48 | 0.68 | 1.16 | 11.64 | 0.86 |
| Daily (24 h) values | 13.10 | 0.20 | 1.28 | 11.62 | 0.89 |

Missing data were estimated by interpolation

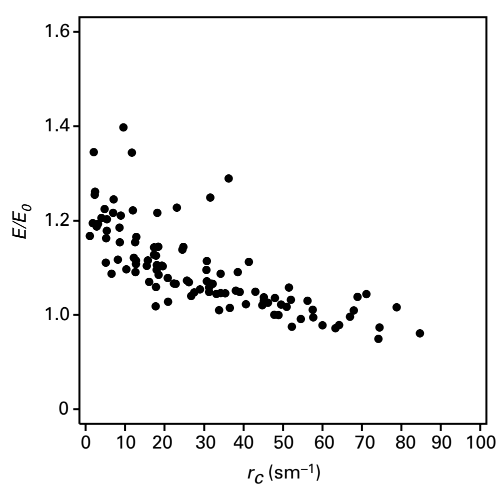


**Figure 6** $E/E_0$ (daily daylight values) vs canopy resistance, $r_c$ (daily minimum, represented by mean of 1100–1300 h)

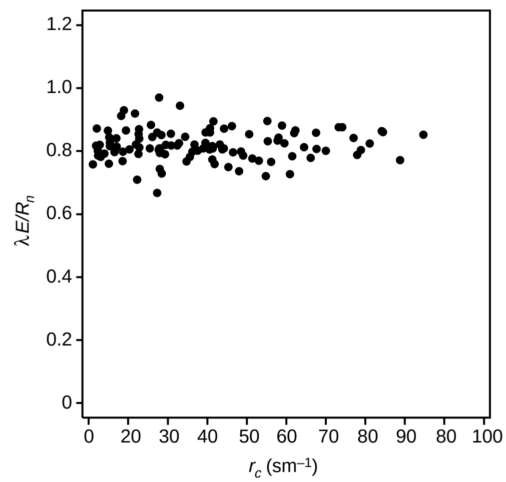


**Figure 7** $\lambda E/R_n$ (daily daylight values) vs canopy resistance, $r_c$ (daily minimum, represented by mean of 1100–1300 h)

July 1999) was in an E–W direction along the riparian corridor. Danish riparian wetlands are often narrow bands not meeting fetch requirements and surrounded by open agricultural land. Advection is therefore believed to be a common phenomenon in this type of wetland under certain weather conditions. This is a condition opposite to that reported in a number of other wetland studies (e.g. Lafleur 1990; Souch *et al.* 1998) in which proximity to large water bodies results in humid sea/lake breezes over the wetland during the daytime that prevent advective enhancement of evapotranspiration rates. The observed rates are thus very close to equilibrium rates (Souch *et al.* 1998).

It should be stressed, however, that the presence of advective enhancement of evapotranspiration rates in the wetland studied is a hypothesis for which there is indicative evidence, but which could not be proved by the experimental setup. Enhanced turbulent mixing due to the proximity of the wetland to the forested hillslopes and edge effects could also be part of the explanation of the high evapotranspiration rates. Additionally, evapotranspiration estimates could be positively biased by errors in measuring the variables that are necessary for the application of the BREB method. Bidlake *et al.* (1996), using similar BREB sensors as in this study, performed a sensitivity analysis using the three most important input variables for the BREB method: $R_n$, $\Delta e$ and $\Delta T$. They found, based on estimated potential measurement errors, a percentage change in average computed evapotranspiration of $-10$ to $+13$, $-6$ to $+7$ and $-9$ to $+13$ for $R_n$, $\Delta e$ and $\Delta T$, respectively.

#### Energy balance

The diurnal course of the energy balance fluxes on a clear-sky summer day at the end of a long dry period is shown in Figure 8. $\lambda E$ peaks around midday coincide with maximum $R_n$. According to Kelliher *et al.* (1993) this is typical for grasslands while, for forests, maximum $\lambda E$ is reached by mid-afternoon, coinciding with maximum $D$. $\lambda E$ is negative between 1900–2200 h when dew is formed, while it is negligible during the night.

Bowen ratios, $\beta$, are small; the peak value of 0.21 is reached at 1130 h and $\beta$ declines steadily to 0.01 at 1800 h. As indicated by these low $\beta$ values, $\lambda E$ is the dominant sink of available energy, consuming 89% of $R_n$ (Table 3). 10% of the available energy is transferred as sensible heat, $H$. A mat of dead vegetation and the dense canopy effectively isolates the soil surface from incident radiation. $G$ is thus a very small component in the energy budget of

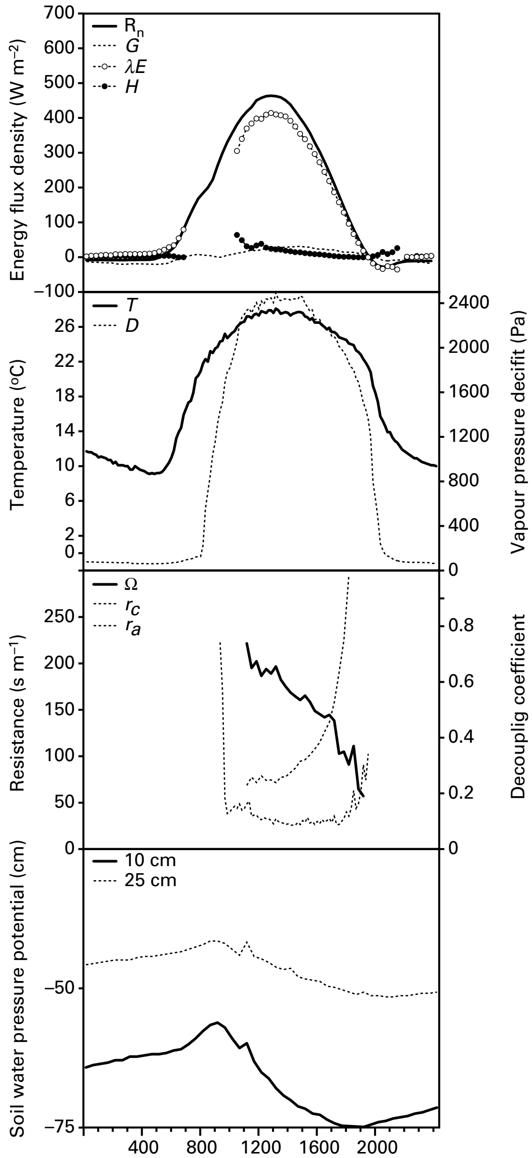


**Figure 8** Diurnal energy balance, saturation vapour pressure deficit (*D*), air temperature (*T*), aerodynamic resistance ($r_a$), canopy resistance ($r_c$), decoupling coefficient (Ω) and soil water pressure potential head in 10 and 25 cm (broken line)

this wetland. In wetlands with open water, *G* may act as a considerable sink of energy during daytime ($G/R_n = 0.2$–$0.4$) (Souch *et al.* 1998; Burba *et al.* 1999). On a daily (24 h) basis Burba *et al.* (1999) found *G* to be small, whereas Souch *et al.* (1998) found *G* to consume 30% of $R_n$, even on a daily basis.

Canopy resistance, $r_c$, is at its minimum before midday with a value of $75\,\mathrm{s\,m^{-1}}$ and increases steeply during the afternoon, adjusting to increasingly negative soil water pressure potential and to *D*, which is maintained at a high, almost constant, level until the evening. As there is no wind at night, $r_a$ takes a very high value until 0800 h when the wind starts blowing. The minimum value of $25\,\mathrm{s\,m^{-1}}$ is reached at midday and is maintained into the afternoon. $\Omega$ decreases during the daytime from 0.7 at 1100 h to below 0.2 at 1900 h, illustrating a shift in evapotranspiration control from available energy to $r_c$ and $D$.

#### Controls on evapotranspiration

The diurnal course of $\Omega$ shown in Figure 8 with a shift in controls on evapotranspiration during the afternoon is typical for all days. The overall mean daily $\Omega$ (mean of 1000–1800 h) is 0.72. This value is an intermediate between values given in the literature (see the review by Jones (1992)) for various vegetated surfaces with the limits set by grass (height 0.1 m): $\Omega = 0.9$ and forest (height 30 m): $\Omega = 0.1$. This implies that the surface of this wetland to some degree is 'coupled' to the prevailing weather conditions and that $\lambda E$ is controlled by a combination of driving forces $R_n$, $D$ and $r_c$ which therefore all have to be considered when modelling evapotranspiration from this wetland.



### Conclusions

Evapotranspiration is the dominant sink in the energy balance of the studied wetland. During the day, evapotranspiration accounts for 82% of the available radiant energy with a peak summertime rate of 5.6 mm d$^{-1}$ and an average rate for the growing season (June–September) of 3.6 mm d$^{-1}$. Due to the presence of deposited fine-grained sediments, soil water availability is kept high at all times, resulting in moderate canopy resistances (overall mean $=$ 32 s m$^{-1}$). Evapotranspiration is controlled by a combination of driving forces $R_n$, $D$ and $r_c$. Consequently, modelling of evapotranspiration rates requires an approach such as the Penman–Monteith equation.

It is hypothesized that the high evapotranspiration rates presented in this study are conditioned by the proximity of the wetland to drier upland areas. In periods with high evaporative demand and low precipitation, warm dry air is formed over the upland areas and wetland evapotranspiration rates are enhanced by local advection. Indicative evidence is presented to support the hypothesis. The results are considered to be representative of other Danish riparian wetlands subjected to frequent flooding and/or with a high groundwater table, with vascular vegetation, and which are narrow corridors in open agricultural landscapes.

### References


Allen, R.G., Pereira, L.S., Raes, D. and Smith, M. (1998). Crop evapotranspiration. *Guidelines for Computing Crop Water Requirements. FAO Irrigation and Drainage Paper 56*. Food & Agriculture Organisation, United Nations, Rome.

Andersen, H.E. (2004). Hydrology and nitrogen balance of a seasonally inundated Danish floodplain wetland. *Hydrol. Process.*, **18**, 415–434.

Andersen, S. (1987). *Landbrugsplanterne*. DSR Forlag, Copenhagen (in Danish).

Bidlake, W.R., Woodham, W.M. and Lopez, M.A. (1996). Evapotranspiration from areas of native vegetation in west-central Florida. *US Geological Survey Water-supply Paper* 2430.

Burba, G.G., Verma, S.B. and Kim, J. (1999). Surface energy fluxes of *Phragmites australis* in a prairie wetland. *Agric. Forest Met.*, **94**, 31–51.

Campbell, D.I. and Williamson, J.L. (1997). Evaporation from a raised peat bog. *J. Hydrol.*, **193**, 142–160.

Detlefsen, N.K. and Plauborg, F. (2001). Reconsideration of methods for calculation of reference evapotranspiration (potential evapotranspiration). *Danish Institute of Agricultural Sciences. DJF rapport Markbrug* 61 (in Danish).

FAO (1998). *World Reference Base for Soils. World Soil Resource Report* 84. Food & Agriculture Organisation, United Nations, Rome and International Society of Soil Science & International Soils Reference & Information Centre, Wageningen.

Gerla, P.J. (1992). The relationship of water-table changes to the capillary fringe, evapotranspiration, and precipitation in intermittent wetlands. *Wetlands*, **12**(2), 91–98.

Grant, R., Laubel, A., Kronvang, B., Andersen, H.E., Svendsen, L.M. and Fuglsang, A. (1996). Loss of dissolved and particulate phosphorus from arable catchments by subsurface drainage. *Wat. Res.*, **30**(11), 2633–2642.

Hoffmann, C.C. (1998). Nutrient retention in wet meadows and fens. PhD thesis, National Environmental Research Institute, Silkeborg, Denmark.

Ingram, H.A.P. (1983). Hydrology. In *Ecosystems of the World vol 4A Mires, Swamp, Bog, Fen, and Moor*, A.P.J. Gore (Ed.), Elsevier Scientific Publishing Company, Amsterdam, The Netherlands, pp. 67–158.

IWA Macrophyte Specialist Group (2000). Constructed wetlands for water pollution control. *Scientific and Technical Report*. International Water Association.

Jensen, S.E. (1996). *Agroclimate at Taastrup 1961–1990*. Jordbrugsforlaget, Copenhagen.

Jensen, T. (personal communication). Agricultural Research Center Foulum, Blicher Allé 50, 8830 Tjele, Denmark.

Jones, H.G. (1992). *Plants and Microclimate*. Cambridge University Press, Cambridge.

Kelliher, F.M., Leunig, R. and Schulze, E.-D. (1993). Evaporation and canopy characteristics of coniferous forests and grasslands. *Oecologia*, **95**, 153–163.

Koerselman, W. and Beltman, B. (1988). Evapotranspiration from fens in relation to Penman's potential free water evaporation (*E*.) and pan evaporation. *Aquatic Botany*, **31**, 307–320.

LaBaugh, J.W. (1986). Wetland ecosystem studies from a hydrologic perspective. *Wat. Res. Bull.*, **22**(1), 1–10.

Lafleur, P.M. (1990). Evapotranspiration from sedge-dominated wetland surfaces. *Aquatic Botany*, **37**, 341–353.

Linacre, E.T. (1976). Swamps. In *Vegetation and the Atmosphere Case Studies*, J.L. Monteith (Ed.), vol 2, Academic Press, New York, pp. 329–347.

McNaughton, K.G. and Jarvis, P.G. (1983). Predicting effects of vegetation changes on transpiration and evaporation. In *Water Deficits and Plant Growth*, T.T. Kozlowski (Ed.), vol 7, Academic Press, New York, USA, pp. 1–42.

Mikkelsen, H.E. and Olesen, J.E. (1991). Sammenligning af metoder til bestemmelse af potentiel vandfordampning. *Tidsskrift for Planteavls Specialserie. Beretning nr*. S 2157 (in Danish).

Monteith, J.L. (1965). *Evapotranspiration and Environment. Symp. Soc. Exp. Biol. XIX*. Cambridge University Press, Cambridge, 205–304.

Monteith, J.L. and Unsworth, M. (1990). *Principles of Environmental Physics*. Edward Arnold, London.

Oke, T.R. (1978). *Boundary Layer Climates*. Methuen, London.

Penman, H.L. (1948). Natural evaporation from open water, bare soil and grass. *Proc. R. Soc.*, **A 193**, 120–145.

Penman, H.L. (1956). Evaporation: an introductory survey. *Neth. J. Agric. Sci.*, **4**, 8–29.

Phersson, M. and Pettersson, O. (1997). Energy and water balances of a bog in central Sweden. *Nord. Hydrol.*, **28**(4/5), 263–272.

Priban, K. and Ondok, J.P. (1986). Evapotranspiration of a willow carr in summer. *Aquatic Botany*, **25**, 203–216.

Price, J.S. (1994). Evapotranspiration from a lakeshore *Typha* marsh on Lake Ontario. *Aquatic Botany*, **48**, 261–272.

Schröder, H., Andersen, H.E. and Kiehl, K. (2005). Rejecting the mean – estimating the response of wetland plant species to environmental factors by non-linear quantile regression. *J. Veget. Sci.* in press.

Souch, C., Grimmond, C.S.B. and Wolfe, C. (1998). Evapotranspiration rates from wetlands with different disturbance histories: Indiana Dunes National Lakeshore. *Wetlands*, **18**(2), 216–229.

Takagi, K., Tsuboya, T., Takahashi, H. and Inoue, T. (1999). Effect of the invasion of vascular plants on heat and water balance in the Sarobetsu Mire, northern Japan. *Wetlands*, **19**(1), 246–254.

Thom, A.S. (1972). Momentum, mass and heat exchange of plant communities. In *Vegetation and the Atmosphere Principles*, J.L. Monteith (Ed.), vol 1, Academic Press, London, pp. 57–109.