DOI: 10.1002/jpln.200621992
J. Plant Nutr. Soil Sci. 2006, 169, 762–774

Measurement and modeling of soil-water dynamics and evapotranspiration of drained peatland soils

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Dedicated in memory of Prof. Dr. Rienk R. van der Ploeg

Accepted July 23, 2006
PNSS P199/2B

Summary

Natural peat soils serve as important sinks for nutrients, organic components, and water. Peat soils can pose major environmental problems when they are drained for agricultural production, which may change their role in the landscape from a sink to a source. To successfully restore and conserve peat soils, it is important to understand the soil-moisture dynamics and water demand of drained peat soils for different climate and groundwater conditions. For this purpose, we conducted a series of lysimeter experiments with peat soils subject to different groundwater levels. Evapotranspiration (ET) rates and upward capillary fluxes in peat soils under grass were measured, while TDR probes and tensiometers were used to monitor the soil-water dynamics in the lysimeter during the growing season. The lysimeter data were simulated using an extended version of HYDRUS-1D to enable ET calculations using the Penman-Monteith equation. A physically based approach was tested to predict the canopy resistance as a function of the average pressure head of the soil root zone. The numerical simulations closely followed the observed soil-moisture dynamics in the lysimeter and were consistent with measured differences in ET rates for different groundwater levels. Besides average climate conditions, the effects of extreme dry and wet weather conditions on ET and groundwater recharge during the growing season were evaluated using the calibrated numerical model for different groundwater levels. Evapotranspiration rates during dry years depended very much on upward capillary flow from the water table and hence on the soil hydraulic properties. During wet years, however, ET was controlled mostly by the evaporative demand of the atmosphere, and much less by the soil hydraulic properties.

Key words: modeling / soil-water dynamics / evapotranspiration / peatland soils

1 Introduction

About 10% of NE Germany is covered by fens. Many fens in Germany in the past suffered from inadequate reclamation, overexploitation, and overall degradation. As a result of intensive agricultural use, the ecological function of fens has slowly changed over the years from being a sink of various nutrients and organic components to becoming a source. This transformation of the function of peat soils has become a major environmental concern. Many studies suggest that groundwater-level regulation can significantly influence the release of C (*Walter* et al., 2001; *Renger* et al., 2002; *Lafleur* et al., 2005). Restoration and well-balanced management of peat soils requires a thorough understanding of the hydrological processes involved (*Joris* and *Feyen*, 2003), particularly the evaporative fluxes from the wetlands, in order to achieve the best use of available water resources for ecological and environmental purposes (*Gavin* and *Agnew*, 2004).

Evapotranspiration is a major component of the water budget of both natural and managed bogs and fens (*Drexler* et al., 2005). Although several studies have focused on the influence of groundwater levels and plant communities on evapotranspiration (ET) from wetlands (*e.g.*, *Spieksma* et al., 1997; *Lott* and *Hunt*, 2001; *Acreman* et al., 2003; *Lafleur* et al., 2005), there is still a general lack of knowledge about the hydrology of fens. Especially little is known about dynamic interactions between the depth of the water table, soil-moisture conditions in the root zone, ET, and climate conditions (*Kellner*, 2001). Furthermore, the importance of transient soil-moisture conditions in the unsaturated zone of peat soils and their effect on the water budget is often underestimated and/or neglected in hydrological studies (*Bradley* and *Gilvar*, 2000; *Baird* and *Gaffney*, 2000). The lack of comprehensive physically based hydrologic models may well have hampered the development and implementation of more effective conservation and restoration management practices of wetlands (*Kellner* and *Halldin*, 2002).

The most common methods for measuring ET in wetlands are the Bowen ratio energy balance and eddy covariance methods (*cf.* review by *Drexler* et al., 2005). Evapotranspiration may also be measured from the diurnal fluctuations of the shallow water table in wetlands (*e.g.*, *Lott* and *Hunt*, 2001) or by the use of the lysimeter technique but their installation is often very expensive. In contrast to conventional lysimeters, we developed a lysimeter that can be installed more easily and with minimal technical and financial effort (*Schwär-*

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zel and *Bohl*, 2003). This type of field lysimeter was used to measure the soil-water dynamics of peat soils at different groundwater levels.

The aims of this study were (1) to apply a physically based approach to predict the canopy resistance as a function of the pressure head of the soil root zone and (2) to incorporate and use the approach in the HYDRUS-1D numerical software (*Šimunek* et al., 1998a) to predict root water uptake and ET for various groundwater and climate conditions. Experimental data from field lysimeters were used to study interactions between ET, groundwater recharge, soil hydraulic properties, and the groundwater level. The calibrated numerical model was subsequently used to examine the influence of groundwater level and climate on the water consumption of peat soils. The numerical results should provide more reliable estimates of the water demand of peat soils for more effective and environmentally friendly land use.

2 Material and methods

2.1 Study area

The lysimeter experiments were conducted in Rhinluch, a fen area of about 87,000 ha ca. 60 km NW of Berlin, Germany, in the Havelland basin (52°85' N, 12°8' E). Peat soils in the area were first drained more than 250 y ago. Drainage was expanded considerably in the 1970s by means of a dense network of open ditches and subsurface perforated pipes. From the 1970s until the collapse of the communist regime in E Germany, agricultural use of fen areas became much more intensive. To enable better access to agricultural machinery, the water tables were lowered during the growing season to a depth of 100 cm below the soil surface. Land use in the early 1990s changed again when hay crops became very popular in the area.

Peat in the Rhinluch area formed under the influence of paludification (swamp formation). The average thickness of the peat is ca. 120 cm. Glacifluvial sands (mostly fine sand) and limnic sediments such as detritus or calcareous mud underlay the peat. The upper peat layers are strongly decomposed and were altered pedogenically by the intensive land use and

Table 1: Surrounding water table and target groundwater depths in three lysimeters during three growing seasons.

Growing season	Surrounding	Lysimeter 1	Lysimeter 2	Lysimeter 3
		— Target groundwater depth [cm] —		
1995	40–60	60–80	–	–
1996	30–40	<30	<30	-
1997	30–40	<30	30–50	60–80

drainage. Deeper layers are dominated by sedge (*Carex* sp.), reed peats (*Phragmites* sp.), or often a mixture of both types of peat. The field site was vegetated predominantly with reed canary grass (*Phalaris arundinacea*).

The climate of the region is characterized by an average annual temperature of 8.1°C, an average annual precipitation rate of 526 mm, and an average annual crop reference evaporation of ca. 600 mm. The Rhinluch area has one of the lowest precipitation rates in Germany. The climatic water balance (precipitation *minus* crop reference ET) is highly negative during the summer months from April to October. We calculated an average value of –170 mm between 1993 and 1998.

2.2 Field and laboratory methods

Three groundwater-table lysimeters (ca. 1 m × 1 m × 1 m) were equipped with fully automated tensiometers (176PC28Honeywell) and manually operated TDR probes (*Easy Test*, 1993), both installed horizontally in two replicates at depths of 10, 20, 30, 40, 50, 60, 70, and 80 cm (Fig. 1). TDR readings were translated into water-content values using the third-order polynomial

$$\theta = 3.93 \cdot 10^{-2} + 2.84 \cdot 10^{-2} \times \varepsilon - 4.19 \cdot 10^{-4} \cdot \varepsilon^2 + 2.77 \cdot 10^{-6} \cdot \varepsilon^3, \quad (1)$$

which was fitted to observed data of the dielectric constant ε [-] and the volumetric water content θ [cm^3 cm^{-3}] for peat

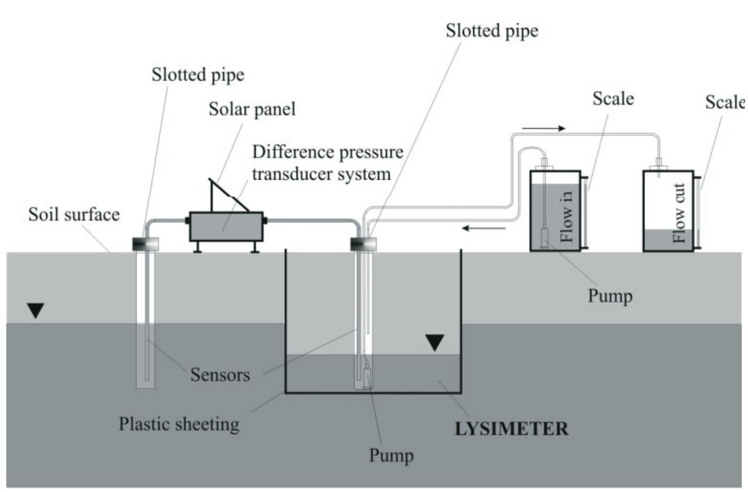

Figure 1: Experimental setup of the groundwater lysimeters.

soils at our site. Equation (1) had a standard error of calibration of 0.022 [cm^3 cm^{-3}].

Groundwater levels in the lysimeters and the surrounding soil were monitored using slotted PVC pipes connected *via* flexible tubes to a differential pressure transducer having a precision of 1 cm. Following procedures outlined by *Feddes* (1971), the water levels in the lysimeters were either automatically adjusted to the surrounding groundwater level (a reference level) or fixed at a certain depth. Surrounding water table and target groundwater depths in three lysimeters during three monitored seasons are summarized in Tab. 2. Because of the shallow surrounding water table we assumed that so-called oasis effects were minimal during our lysimeter experiments. In some periods, we observed higher deviations from the target water table. Such measurement periods were not used for quantification of ET.

Pressure differences recorded by a control unit either triggered operation of a bilge pump in the lysimeter pipe or released a valve in a supply-water container to increase the water level in the lysimeter. The volume of water added or removed was measured by recording water levels in the containers. Hysteresis in the imposed groundwater level was at most 2 cm.

The average actual ET rate (ET_{act}) between times t_1 and t_2 was calculated as follows

$$\int_{t_1}^{t_2} ET_{act}\,dt = \int_{t_1}^{t_2} [P+Q]\,dt - \int_0^L \int_{t_1}^{t_2} \frac{\partial\theta}{\partial t} dt dz \tag{2}$$

where P is the precipitation rate [mm d^{-1}], Q represents the net inflow of water [mm d^{-1}] into the lysimeter needed to maintain the water table inside the lysimeter equal to that of the surrounding groundwater, L is the depth of the lysimeter [cm], and z is the vertical coordinate [cm]. More information about the groundwater lysimeter and its operation is given in *Schwärzel* and *Bohl* (2003).

At the site, continuous measurements of air temperature, humidity, wind speed, and global radiation were made at the height of 2 m above the ground. The precipitation rates were measured using a Hellmann rain gauge.

Evaporation experiments on undisturbed soil samples were conducted in the laboratory to determine the unsaturated hydraulic conductivity (*Schwärzel* et al., 2006). The unsaturated soil hydraulic properties were described with the van Genuchten-Mualem (VGM) model (*van Genuchten*, 1980; *Mualem*, 1976):

$$S_e = \frac{\theta-\theta_R}{\theta_S-\theta_R} = \frac{1}{(1+\alpha h^n)^m} \quad h\leq 0, \tag{3}$$

$$S_e = 1 \quad h > 0 \tag{4}$$

$$K(S_e) = K_S S_e^{0.5}\left[1-\left(1-S_e^{1/m}\right)^m\right]^2 \tag{5}$$

$$m = 1-1/n \quad n>1 \tag{6}$$

where S_e is effective saturation [-], θ is the volumetric water content [cm^3 cm^{-3}], θ_R is the residual water content [cm^3 cm^{-3}], θ_S is the saturated water content [cm^3 cm^{-3}], h is the pressure head [cm], K is the hydraulic conductivity [cm d^{-1}], K_S is the saturated hydraulic conductivity [cm d^{-1}], and α [cm^{-1}], n [-], and m [-] are empirical parameters. Parameter optimization (inverse) methods (*Šimunek* et al., 1998b) were used to fit θ_R, θ_S, α, n, and K_s to the laboratory evaporation data. The final set of soil-hydraulic parameters used in this study is given in Tab. 2.

2.3 Numerical model

2.3.1 Governing equations

Variably saturated flow in the lysimeters was simulated using HYDRUS-1D (*Šimunek* et al., 1998a) which numerically solves the Richards equation:

$$\frac{\partial\theta}{\partial t} = \frac{\partial}{\partial z}\left(K\frac{\partial h}{\partial z}+K\right) - S, \tag{7}$$

where t is time [d], z is the vertical coordinate (positive upward) [cm], S represents the volume of water taken up by roots per unit bulk volume of the soil in unit time [cm^3 cm^{-3} d^{-1}], and where the other variables are as defined before. The solution of Eq. 7 requires a set of initial and boundary conditions. The initial condition in our study was given in terms of the measured pressure head, h_i:

$$h(z,t) = h_i(z) \qquad t=0 \tag{8}$$

Table 2: Soil-physical and soil-hydraulic parameters for various horizons of the studied peat soil (*Schwärzel* et al., 2006). θ_s is the saturated water content, K_S is the saturated hydraulic conductivity, and [cm^{-1}] and n are empirical parameters in the *van Genuchten* (1980) function.

Depth [cm]	Peat type	*BD* [g cm^{-3}]	*X* [%]	θ_s [cm^3 cm^{-3}]	α [cm^{-1}]	*n*	K_s [cm d^{-1}]
0–20	very humified peat	0.312	77	0.797	0.020	1.23	14.0
20–25	very humified peat, compacted	0.371	76	0.730	0.012	1.12	6.69
25–30	humified peat, compacted	0.386	77	0.741	0.005	1.15	0.92
>30	weakly humified, reed-sedge peat	0.194	84	0.891	0.003	1.16	104

For all cases, θ_R = 0.0 and λ = 0.5
BD = bulk density
X = ignition loss

The following atmospheric boundary condition was specified at the soil surface:

$$\left| -K\frac{\partial h}{\partial z} - K \right| \leq E \quad \text{at } z = 0 \tag{9}$$

and

$$h_A \leq h \leq h_S \qquad \text{at } z = 0, \tag{10}$$

where E is the maximum potential rate of infiltration or evaporation under the current atmospheric conditions [cm d^{-1}], and h_A and h_S are, respectively, the minimum and maximum permitted pressure heads at the soil surface [cm]. Following *Šimunek* et al. (1998a), the values of h_A and h_S were assumed to be -10^6 and 0 cm, respectively. A time-variable pressure-head boundary condition consistent with our measurements was specified at the lower boundary.

2.3.2 Root water uptake

The sink term, S, in Eq. 3 represents the volume of water removed from a unit volume of soil per unit time due to the root water uptake. *Feddes* et al. (1978) defined S as

$$S\,(h) = \alpha(h)\;S_p, \tag{11}$$

where $\alpha\,(h)$ is a prescribed dimensionless function ($0 \leq \alpha \leq 1$) accounting for the effects of water stress on root water uptake, and S_p is the potential root water-uptake distribution [d^{-1}]. When the potential root water-uptake rate is uniformly distributed over the root zone, S_p is given by

$$S_p = \frac{1}{L_r} T_p, \tag{12}$$

where L_r is the depth [cm] of the root zone and T_p is the potential transpiration rate [cm d^{-1}]. Equation 12 may be generalized by introducing a nonuniform distribution of the potential root water-uptake rate over a root zone of arbitrary shape as follows:

$$S_p = b(z)T_p, \tag{13}$$

where $b(z)$ is the normalized root water-uptake distribution [cm^{-1}]. To ensure that $b(z)$ integrates to unity over the root zone, any measured or prescribed root distribution function, $b'(z)$ over the root zone may be normalized into $b(z)$ using the equation

$$b(z) = \frac{b'(z)}{\int_{L_r} b'(z)dz}. \tag{14}$$

A number of soil pits were dug to estimate the mean root depth depending on groundwater conditions. In the most cases, a well-defined transition between the very densely rooted topsoil and the weakly rooted subsoil was found. Therefore, we assumed a uniform root distribution in our modeling studies. Based on our field observations, we chose for groundwater levels ≤30 cm a root depth of 10 cm, for groundwater levels between 40 and 60 cm a root depth of 30 cm, for groundwater levels between 60 and 80 cm a root depth of 35 cm, and for groundwater levels >90 cm a root zone depth of 40 cm.

The potential root water-uptake function S_p in Eq. 11 corresponds to the potential transpiration rate T_p [cm d^{-1}] of a growing canopy as follows

$$\int_{L_r} S_p dx = T_p, \tag{15}$$

in which T_p depends on weather conditions, plant height, and canopy resistance against water loss. The actual root water uptake rate S as a function of depth is obtained by substituting Eq. 13 into Eq. 11, while the actual transpiration rate T_a [cm d^{-1}] is calculated by integrating S over the root zone:

$$T_a = T_p \int_{L_r} \alpha(h)b(z)dz, \tag{16}$$

which shows that T_a equals T_p only when $\alpha = 1$ over the entire root zone (*i.e.*, no water stress in any part of the root zone).

2.3.3 Daily evapotranspiration

Evapotranspiration ETI [cm d^{-1}] is equal to the sum of transpiration by plants T [cm d^{-1}], evaporation from the soil surface E [cm d^{-1}], and evaporation of water intercepted by the canopy surface I [cm d^{-1}], further referred to as interception. The potential transpiration (maximum possible water uptake) rate can be calculated as follows

$$T_p = ETI - I - E. \tag{17}$$

In this study, we assume that evaporation from the soil surface is negligible because of the presence of an essentially complete plant cover.

The daily ET rate [cm d^{-1}] from a dry canopy was estimated using the Penman-Monteith equation (*Monteith*, 1965) as follows:

$$ET_{PM} = \frac{1}{\lambda}\frac{\left[s(R_n - G) + \frac{\rho c_p}{r_a}(e_a - e_d)\right]}{\left[s + \gamma\left(1 + \frac{r_c}{r_a}\right)\right]}, \tag{18}$$

where λ is the latent heat of vaporization [J g^{-1}], R_n is the net radiation flux at the canopy surface [J m^{-2} d^{-1}], G is the soil heat flux [J m^{-2} d^{-1}], ρ is the atmospheric density [g cm^{-3}], c_p is the specific heat of humid air [J g^{-1} °C^{-1}], $(e_a - e_d)$ is the vapor pressure deficit [kPa], r_c is the crop canopy resistance [s m^{-1}], r_a is the aerodynamic resistance [s m^{-1}], s is the slope of the vapor-pressure curve [kPa °C^{-1}], and γ is the psychrometric constant [kPa °C^{-1}]. The FAO (*Allen* et al., 2000) suggested a procedure for evaluating the Penman-Monteith equation using daily values of air temperature [°C^{-1}], air humidity [kPa], wind speed [m s^{-1}], and solar radiation [J m^{-2} d^{-1}]. This procedure was modified as described below and implemented into HYDRUS-1D.

The daily evapotranspiration of a wet canopy (with $r_c = 0$) was calculated as follows (*Monteith*, 1965; *Rijtema*, 1965):

$$ET_{wet} = \frac{1}{\lambda}\frac{\left[s(R_n - G) + \frac{\rho c_p}{r_a}(e_a - e_d)\right]}{s + \gamma}. \tag{19}$$

The transpiration rate from a wet canopy is generally larger than the corresponding value from a dry canopy. This reduction in the potential transpiration rate from a dry as compared with a wet canopy is given by the ratio of Eqs. 18 and 19 (see also *Feddes*, 1971):

$$\frac{ET_{PM}}{ET_{Wet}} = \frac{s+\gamma}{s+\gamma\left(1+\frac{r_c}{r_a}\right)}. \tag{20}$$

During periods with precipitation, part of the incoming energy is used for the evaporation of intercepted water, which leaves less energy for potential transpiration relative to dry periods under the same conditions. The potential transpiration rate may then be calculated as (*cf.*, *Rijtema*, 1965; *Feddes*, 1971):

$$T_p = \frac{(s+\gamma)}{s+\gamma\left(1+\frac{r_c}{r_a}\right)}(ET_{wet} - I). \tag{21}$$

The interception (I) of precipitation (P) was calculated according to *Hoyningen-Hüne* (1983), *Braden* (1985), and *van Dam* et al. (1997) as follows:

$$I = aLAI\left(1 - \frac{1}{1+\frac{bP}{aLAI}}\right), \tag{22}$$

where a is an empirical coefficient [cm], LAI is the leaf area index [-], and b is the soil cover fraction [-]. We assumed a value of 0.5 for a and 1.0 for b. The leaf-area index (LAI) for grass was calculated from the vegetation height h_{Veg} [cm] using a modification of an equation proposed by *Menzel* (1997):

$$LAI = 0.7862 + 0.4123h_{Veg} - 5.49\times10^{-3}h_{Veg}^2. \tag{23}$$

The mean vegetation height of the site was measured weekly from 1993 until 1998 during the growing season.

Knowledge of the canopy resistance r_c is needed in order to calculate the actual ET rate of a growing crop using the Penman-Monteith equation. When ET is evaluated using Eq. (2), the canopy resistance r_c can be derived from weather data and the known plant height. This approach assumes that for given conditions evaporation from a fully covered soil can be neglected. The canopy resistance r_c for precipitation-free periods can then be obtained assuming that the measured ET rate is equal to the actual transpiration rate as follows (*cf.*, also *Henning*, 1992):

$$r_c = r_a\left(1+\frac{s}{\gamma}\right)\left(\frac{ET_{wet}}{[ET_{measured}]} - 1\right). \tag{24}$$

3 Results and discussion

3.1 Soil-water dynamics

In this section, we discuss our lysimeter data describing the soil water regime of a drained fen during the summer of 1995. Interactions between weather conditions, the groundwater hydrograph, and soil-moisture dynamics are presented and analyzed.

3.2 Soil water regime during the summer

Figure 2 shows daily values of precipitation, depth of the groundwater table, and measured values of the soil water content and pressure head at various depths in one of the lysimeters covered by plants during the growing season. In addition, calculated daily values of the crop reference ET (*Allen* et al., 2000) are also presented.

In early July, the groundwater level was located between 70 and 75 cm depth. An extremely intense rain storm (25 mm in 4 h) in the middle of July caused a sudden rise of the groundwater table to a depth of ca. 20 cm (Fig. 2b). The water table during the next several days dropped to the target groundwater level of 70 cm depth. Except for a period with technical problems in the beginning and the middle of August, this target groundwater level was maintained until the end of September.

Figures 2c and 2d show that drying of the soil during the months of June and July was mostly limited to the upper 10 cm in spite of the high evaporative demand of the atmosphere (Fig. 2a). The relatively high level of the water table and the comparatively frequent precipitation events prevented more substantial changes in the water content of the subsoil. Until early July, observed pressure heads were very similar at all depths (Fig. 2d). Afterwards, due to the high evaporative demand, pressure heads decreased faster at a depth of 10 cm than at the other measurement depths. Within a few days, the measurement limit (ca. –700 cm) of the tensiometer located at the shallowest depth was reached. The rapid decrease in the pressure head at a depth of 10 cm was caused by drastic reductions in the unsaturated hydraulic conductivity as the soil dried out, with a simultaneous decrease in upward flow from the water table through the capillary fringe. At the end of the dry period in the middle of summer, the soil water content at a depth of 10 cm reached the wilting point (pF 4.2). Figure 2c shows that the water content was reduced by nearly 50% within 2 months.

The soil profile rewetted slowly following several precipitation events at the end of the dry period, with a delay of several days. The rain water was found to infiltrate very fast without fully wetting the peat, thus causing a rapid rise of the water table. The pressure head during this time period increased quickly at all depths. However, the water contents increased only very slowly because of peat water repellency (Figs. 2c, d). Our findings agree well with those of *Kellner* and *Halldin* (2002) who for a Swedish *Sphagnum* mire found that rain water moved rapidly to layers just above the water table without any significant storage in the upper horizons. The slow increase in water content is related to enhanced hysteresis in the soil water-retention curve (*cf.,* also *Kellner*, 2001). As further described in detail by *Schwärzel* et al. (2002), differences in water content of up to 30% were observed between the draining and wetting branches of the retention curve at the same pressure head.

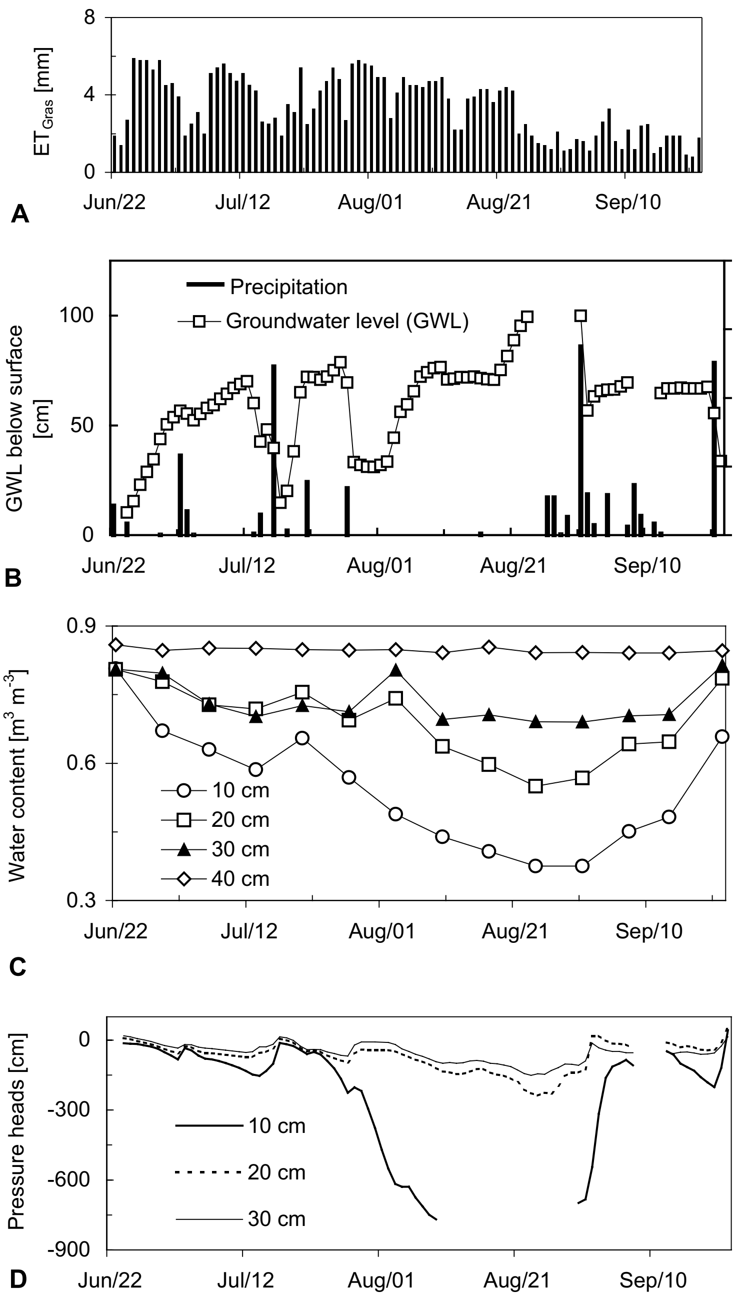

Figure 2: Plots showing (a) calculated daily values of the potential evapotranspiration rate (ET_{grass}), measured daily values of (b) the precipitation rate and the groundwater table, (c) the soil water content, and (d) observed pressure heads in one of the lysimeters (from June 22 through September 21, 1995).

3.3 High-temporal resolution soil-plant-atmosphere studies

Our understanding of the transient nature of soil-water dynamics in peat soils can be improved with high-resolution values of pressure head, precipitation, and evapotranspiration *vs.* time. Figure 3 shows such data, where precipitation was measured hourly, while ET was calculated hourly according to *Penman* (1948) as modified by *Wendling* (1991). Daily fluctuations of ca. 100 cm in the pressure head were recorded at a depth of 20 cm in mid August. These daily fluctuations were a result of root water uptake during the day and more continuous, stable capillary rise of water from the water table. The decrease in the pressure head during daytime followed the ET peaks with a time delay of a few hours (Fig. 3). At noon, when the evaporative demand of the atmosphere was highest, the pressure head started to decrease at a depth of 20 cm. At this time, the upward flux from deeper soil zones was clearly smaller than the water uptake rate by plant roots. Water continued to be taken up by the roots until the evenings (although at a diminishing rate), at which time the lowest pressure head values were reached (Fig. 3b).

The increase in pressure head overnight is an indication that water absorbed by plants is being replenished by capillary rise and other redistribution processes. Yet, it is also striking that daily amplitudes of the pressure head decreased notwithstanding the relatively high evaporative demand of the atmosphere (Fig. 3). This may have been caused by the shape of the water-retention function, with the soil water

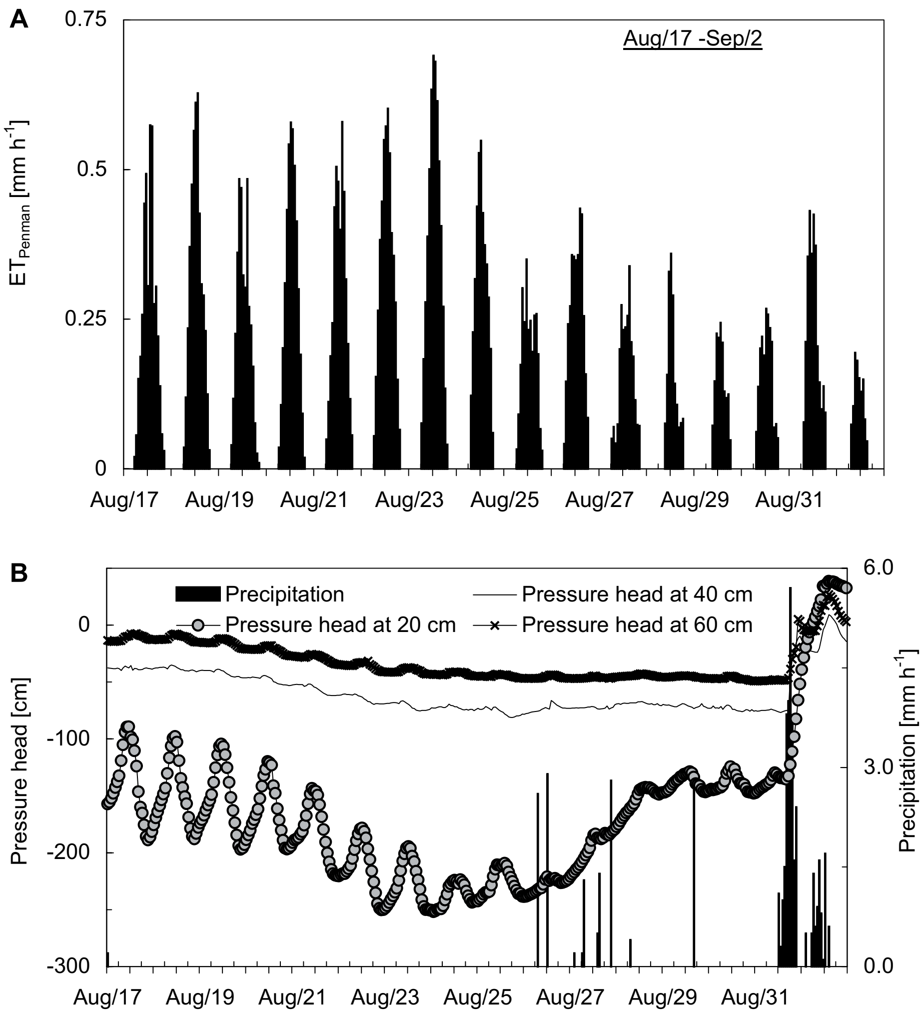

Figure 3: Plots showing (a) calculated hourly values of the potential evapotranspiration rate and (b) measured hourly values of the precipitation rate and observed pressure heads at three depths in one of the lysimeters.

capacity ($d\theta/dh$) increasing with decreasing pressure heads. At the same time, the maximum pressure heads gradually decreased (Fig. 3b). During this time period, the daily water-uptake rate by plant roots exceeded the daily capillary supply of water from the groundwater table. This is due to a reduction in the hydraulic conductivity caused by the lower water contents, thus limiting upward capillary flow from the water table. Only when the evaporative demand of the atmosphere declined markedly at the end of August, upward capillary flow provided more water than was taken up the roots. As a result, the daily maximum values of the pressure head at that time gradually increased again (Fig. 3b). The rapid increase in the pressure head at a depth of 20 cm corresponded with the first precipitation events after a long dry period (August 26 and 27). Within a few hours, the large amounts of precipitation on August 31/September 1 caused a substantial increase in the pressure heads at all depths (Figs. 2d, 3b), occasionally causing positive pressures. The groundwater level then also increased substantially (Fig. 2b).

Our interpretation of the high temporal pressure-head measurements during the precipitation free period, and of the observed minimum values in the evening, confirm results of

Table 3: Values of the canopy resistances of grass as compiled from the literature.

Type of Grass	r_c [s m^{-1}]	Reference
Reed canary grass	22	This study
Mesotrophic grassland	50–100	*Acreman* et al. (2003)
Grass	75–90	*Zenker* (2003)
Japanese moor grass	111	*Takagi* et al. (1998)
Reed canary grass-nettle reeds	15	*Böhm* and *Quast* (1998)
Purple moor grass	22	*Spieksma* et al. (1997)
Grass	40–85	*Menzel* (1996)
Grass, 12 cm	70	*Allen* et al. (1994)
Field weeds	74	*Henning* (1992)
Tall feccue	1–11	*Feldhake* and *Boyer* (1986)
Meadow knawel	11–29	*Feldhake* and *Boyer* (1986)
Creeping finger grass	20–48	*Feldhake* and *Boyer* (1986)
Grass, cut	26	*Szeizc* and *Long* (1969)

Kellner (2001) who derived hourly values of the canopy resistance in Eq. 18 from energy-budget measurements. In that study, the canopy resistance often had its diurnal maximum in late afternoon or the evening when the vapor deficit had passed its daily maximum. *Kellner* (2001) suggested that with increasing evaporative forcing, the capillary-flow capacity was too low to deliver water to the canopy.

3.4 Canopy resistance and water budget for the measurement periods

Evapotranspiration rates of lysimeter 2, measured in the summer of 1996 when the groundwater level was <30 cm below the soil surface, were used to calculate the minimal canopy resistance because the water supply can be assumed to be unlimited under these conditions. From these calculations, using Eq. 24 we obtained an average value of 22 s m^{-1} for the minimal canopy resistance r_c. The minimal canopy resistance of a crop can be interpreted as the maximum conductivity of the canopy to deliver water at unlimited soil-water availability. A compilation of canopy resistances published for various grasses is presented in Tab. 3, with our value being in the middle of the range of reported values. The value of 22 s m^{-1} is also close to the value of 15 s m^{-1} found by *Böhm* and *Quast* (1998) for very similar reed canary grass–nettle reeds vegetation.

Table 4: Simulated and measured water-balance components for different groundwater levels (July 29 through September 16, 1997). Average daily values are given in parentheses (in mm d^{-1}).

Period	Groundwater depth [cm]	measured [mm]		simulated [mm]	
		ETI	$Q_{capillar}$	ETI	$Q_{capillar}$
7/29–9/16/97	<30	225 (4.6)	199 (4.1)	201 (4.1)	163 (3.3)
	30–50	226 (4.6)	159 (3.2)	197 (4.0)	138 (2.8)
	60–80	174 (3.6)	116 (2.4)	166 (3.4)	106 (2.2)

$Q_{capillar}$ = water flow into the lysimeter to keep constant groundwater level

Numerical simulations of the water balance were carried out using HYDRUS-1D and the experimentally determined canopy resistance of 22 s m^{-1}. Figure 4 compares modeling results with values collected in lysimeter 1 during the summer of 1996. Measured and calculated cumulative ET rates agreed well, with the cumulative ET rate being underestimated by <5% over a period of 70 d. The dynamics of soil water contents was also predicted satisfactorily by the numerical model.

Evapotranspiration rates were next calculated for various time periods with groundwater levels set to maintain a canopy resistance, r_c, of 22 s m^{-1}. In Fig. 5, the measured and calculated ET rates are plotted against the averaged pressure heads in the root zone. In Fig. 6, $ETI_{optimal}$ is the ET rate calculated for conditions of unlimited supply of water using the estimated canopy resistance of 22 s m^{-1}. The hand-fitted curve through the data closely resembles the piece-wise linear water-stress response function $a(h)$ initially proposed by *Feddes* et al. (1978). Near saturation, the measured ET gradually declined compared to the calculated ET rate as indicated by the $a(h)$ function. Despite the scatter, the data clearly show that water uptake by the reed canary grass crop is reduced when pressure heads in the root zone increase to ca. –15 cm or higher. Notice that the measured evapotranspiration rate was equal to the calculated ET only in a comparatively small range of pressure heads. The ratio of measured and calculated ET rates started to decline already below pressure heads of ca. –70 cm.

Our finding that the ET rate depends strongly on the soil-water pressure head has, to our knowledge, not been found in other wetland studies. Several previous studies (*e.g.*, *Mundel*, 1982; *Ingram*, 1983) suggest that the depth of the water table to a large extent controls peatland ET, while more recently, *Gavin* and *Agnew* (2004) observed a considerable effect of peat surface moisture on ET of a temperate wet grassland. In contrast, *Lafleur* et al. (2005) showed for a shrub-covered bog in Canada that the ET rate appeared unaffected by the water table until it dropped below 65 cm. For groundwater tables <65 cm, the moisture supply to the vascular plants became limiting and transpiration decreased. Key to understanding these different results appears to be the amount of water available for root water uptake. When evaluating the role of groundwater in the supply of water to vegetation in peat soils, one must consider the upward capil-

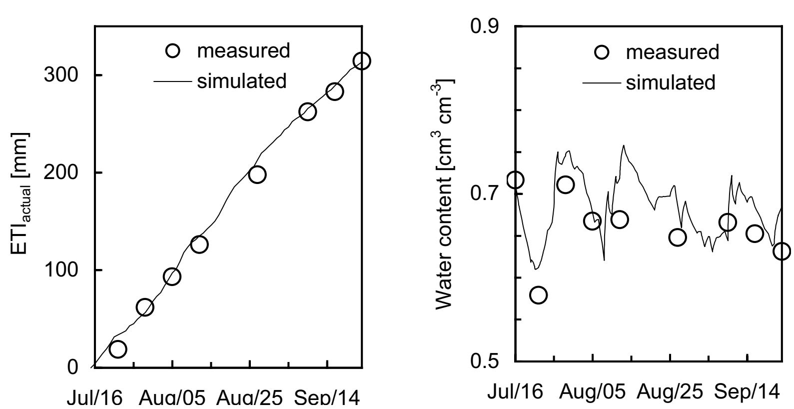

Figure 4: Simulated and measured cumulative evapotranspiration rates (left) and water contents in the top horizon (right) during the 1996 water-balance period. The average depth of the water table during this time interval was 25 cm.

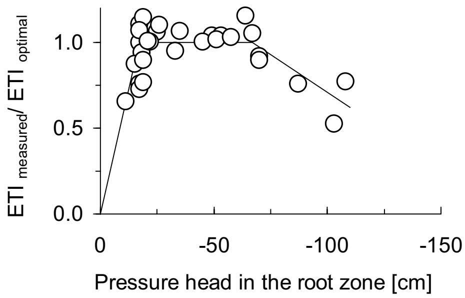

Figure 5: Plot of the ratio of measured and calculated evapotranspiration rates as a function of the average root-zone pressure head. The curve was hand-fitted through the data.

lary flow of water from the water table into the root zone. This upward flow depends on the unsaturated hydraulic conductivity of the peat soil and the hydraulic gradient, which are controlled by both the weather conditions and depth to groundwater. In general, the larger the vertical distance between water table and the bottom of the root zone, the less capillary rise should be expected.

In previous papers (*Wessolek* et al., 2002; *Schwärzel* and *Bohl*, 2003), we showed that for an average groundwater level of 40 cm, the highest capillary delivery rates occurred during periods with high evaporative demand. In some instances, the amount of capillary rise was independent of the evaporative demand of the atmosphere. The capillary delivery from groundwater was delayed in time when a period of high evaporative demand was followed by a period of low evaporation and *vice versa.* Capillary rise is negligible during and after rainfall events. The plant water demand is then directly satisfied by precipitation water infiltrating the root zone. These transient flow conditions explain why in some studies, no significant relation was found between the depth of the water table and the daily ET rate (*e.g.*, *Lafleur* et al., 2005), while in other studies (*e.g.*, *Lapen* et al., 2000), the ET rate depended on soil water, as we found here. Still, *Kellner* (2001) observed no correlation between soil moisture, as measured by TDR, and the rate of ET. However, our results show that the average pressure head of the root zone is a more sensitive indicator for the relation between evaporative demand of the atmosphere and the water consumption of fens and bogs; small changes in the water table and soil moisture may lead to a relatively large and easily detectable change in the pressure head.

To consider the observed reduction in ET (Fig. 5) in numerical models, we suggest the introduction of a critical pressure head, h_{crit}, of the root zone. This parameter should indicate when soil water conditions in the root zone are no longer optimal for the standing vegetation. Plants respond to suboptimal conditions by increasing their minimal canopy resistance to water loss. When the actual pressure head in the root zone, h_{act}, reaches values $< h_{crit}$ (*i.e.*, more negative), then the actual canopy resistance $r_{c,act}$ of reed canary grass is calculated using

$$r_{c,act} = r_{\mathrm{c,minimal}}[1 + \delta(|h_{act}| - |h_{crit}|)], \qquad (25)$$

where the conversion factor d is equal to 10^{-1} m^{-1}, $r_{c,minimal}$ is 22 s m^{-1}, and h_{crit} is –70 cm. This equation is a modification of a similar expression proposed by *Sambale* (1998). Our field measurements (Fig. 5) indicate that Eq. 25 is valid for pressure heads >–120 cm. Figure 6 further shows that when the pressure head in the root zone becomes >–15 cm, root water uptake is reduced linearly because of poor aeration, similarly as in the approach of *Feddes* et al. (1978).

To test the reduction approach, simulations were carried out for a period of 7 weeks in 1997 for lysimeters in which the groundwater levels were kept at average depths of 20, 45, and 65 cm. Table 4 compares calculated and observed values of the ET rate and the water flux into the lysimeters to maintain constant groundwater levels. Differences in ET between the three lysimeters were predicted well with the model. Deviations between measured and calculated cumulative ET rates were <10% during the 49 d period. The largest deviations between measured and calculated values occurred for the scenario with the shallowest depth of the groundwater level after the high-precipitation events (not shown). Furthermore, the model underestimated capillary rise for the same scenario by ca. 15%, probably due to underestimation of the hydraulic conductivity near saturation as obtained with the laboratory tests. Measured and calculated values for upward flow rates (capillary rise) agreed well when the water tables were deeper. Results listed in Tab. 4 show that for shallow groundwater conditions ca. 90% of the evaporated water are supplied by capillary rise from groundwater. When the groundwater level is deeper, 65% of root water uptake still comes from groundwater.

Simulated and measured water contents and pressure heads at depths of 10 and 20 cm for conditions where the average water table was kept at about 45 cm below the soil surface are presented in Fig. 6. Pressure heads at a depth of 10 and 20 cm were underestimated with the model, relative to the measurements at early times. Rewetting of the peat after precipitation late in the summer caused a rapid increase in pressure-head values, a tendency which was captured well with the model. In contrast, the soil only gradually wetted as a result of air entrapment and the water-repellency problems.

3.5 Water-budget simulations dependent on groundwater and climate conditions

3.5.1 Water balance

The tested model was next used to investigate the influence of various groundwater and weather conditions on ET. For this purpose, we used meteorological data from years 1993/94 and 1997/98 to represent relatively wet and dry years, respectively, and assumed that the water table remained constant at specific depths. Figure 7 shows the dependence of the annual ET and groundwater recharge rates on the groundwater level for the dry and wet years. The ET rate was generally higher in a dry year than in a wet year due to higher solar radiation and higher temperatures. While during the dry year, 80 mm less water evaporated when groundwater was at a depth of 70 cm compared to a depth of 30 cm, the difference was only 50 mm for the wet year.

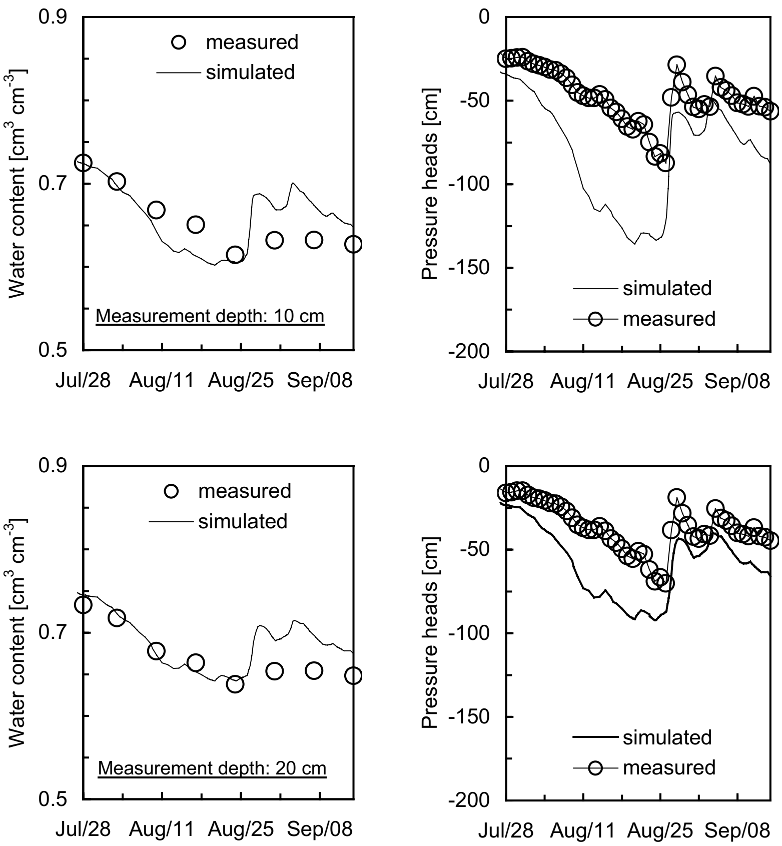

Figure 6: Simulated and measured soil water contents and pressure heads at two depths in one of the lysimeters with the average depth of the water table maintained at 45 cm during the 1997 water-balance period.

The water balance of drained fens during a dry year can be extremely negative (Fig. 7). Considerable capillary rise can then occur, even with groundwater table at 120 cm. In contrast, we calculated groundwater recharge of up to 240 mm y^{-1} in the wet years with much more precipitation and low actual ET rates. From Fig. 8, we conclude that in dry years the water supply of plants (and hence ET) depends much more on capillary rise from groundwater than in wet years.

The legend of Fig. 7b shows also calculated values of the crop reference evapotranspiration ET_0 (*Allen* et al., 2000). These annual values underestimate the simulated data at water tables ≤60 cm below the soil surface. This is not surprising because the approach of *Allen* et al. (2000) is a climatic parameter and does not consider the plant characteristic and the soil factors.

3.5.2 Water demand of drained peat soils

Since peat mineralization accelerates when the water table is deep and the soil surface dry, raising the groundwater table and rewetting the soil profile are measures that protect the peat (*Renger* et al., 2002). Simulations over many years may be used to obtain estimates of the average water demand needed to successfully implement such measures. Figure 8 presents the results of such simulations, showing the water

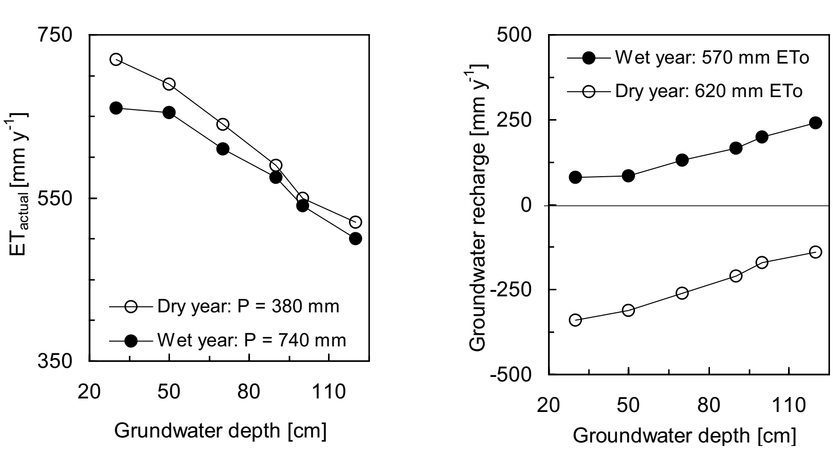

Figure 7: Simulated components of the soil water balance of a peat soil for relatively dry and wet years. ET_0 is the crop reference evapotranspiration (*Allen* et al., 2000).

demand needed to establish a given groundwater level for various climate conditions. As an indicator of the evaporative demand of the atmosphere (or "climate stress") during the growing season, we used the climate water balance (CWB), which is defined here as the difference between precipitation and crop reference ET (ET_0). Assuming a mean groundwater depth of 1 m, Fig. 8 shows that 210 mm water must be supplied for a CWB of –330 mm during the summer months to reach a groundwater level of 30 cm. This is two times higher compared to a slightly negative CWB value of –50 mm. To maintain the water table at only 70 cm, the water demand decreases to 130 mm and 70 mm for CWB values of –330 and –50 cm, respectively.

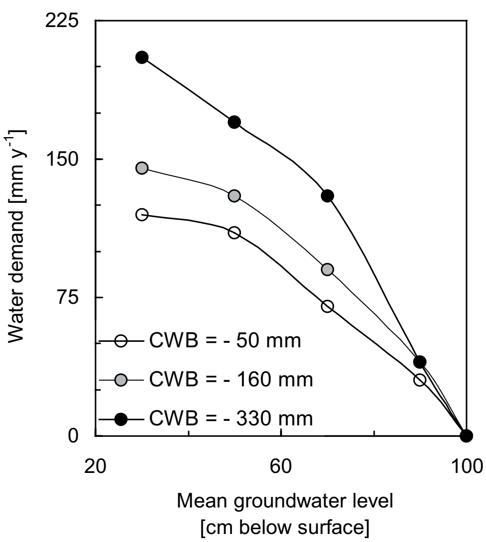

Figure 8: Water demand required to maintain the water table at a desired level for various climate conditions. CWB is the climate water balance during the growing season (precipitation *minus* crop reference evapotranspiration). The average depth of the natural groundwater level was assumed to be 1 m.

4 Conclusions

We extended the HYDRUS-1D model to allow calculations of soil water-balance components (evaporation, groundwater recharge, capillary rise) of drained fens from meteorological data. Potential ET rates were calculated using the Penman-Monteith equation. Estimates of the canopy resistance required in this equation were derived from our lysimeter experiments. Field measurements showed that the soil-water dynamics depended on the depth of the groundwater level and the evaporative demand of the atmosphere. We were able to demonstrate that for *Phalaris arundinacea*, the ET rate is controlled by the pressure head of the root zone and the evaporative demand of the atmosphere. A physically based approach was tested to relate the canopy resistance of plants to the average pressure head in the root zone, the latter being continuously adjusted in the model.

Model simulations described the observed lysimeter data relatively well. Some deviations between measured and calculated soil water contents occurred during wetting after relatively long drying periods. These deviations were presumably due to wettability problems of the water-repellent peat. These problems of water repellency were not included in the present version of HYDRUS-1D.

During dry years, a nearly linear relationship was found between the depth of the water table and the amount of ET. For groundwater levels between 30 and 90 cm, the ET rate was controlled mostly by capillary rise of water from groundwater, and thus on the soil-hydraulic properties. During wet years, relatively small differences in ET rates existed for groundwater levels between 30 and 70 cm. Additional water supply from groundwater was found to be far less important in wet years than in dry years. In wet years, the amount of ET was mainly controlled by the energy demand of the atmosphere and less by the soil hydraulic properties.

Lower soil water contents in the root zone of peat soils are known to promote peat mineralization and the release of CO_2 into the atmosphere (*Renger* et al., 2002). Shallow groundwater levels are essential for environmentally friendly land use with minimal C release. This strategy, unfortunately, requires large amounts of water during the summer months. This water may not be available for large peat areas such as the Rhinluch fens in Germany. Hence, a common practice of wetland management is to protect only parts of the peat soils from mineralization. Deeper groundwater levels are then allowed in the remaining areas. Our findings, however, show that at least temporarily during drought periods, peat areas in Central Europe will have a great demand for water. This is an important finding that needs to be considered in the design and implementation of sustainable restoration/conservation practices for wetlands, particularly with regard to the predictions of climate change.

Acknowledgments

This study is dedicated to *Rienk van der Ploeg* who in the late 1970s introduced in Germany the first numerical models applicable to soil-hydrologic problems. Gerd Wessolek is thankful for the time at Georg August University in Göttingen when *Rienk* taught a relatively small group how to solve problems in water and solute transport. *Rienk's* vigorous support of the use of advanced conceptual models and computations tools has lasted through, among other things, his organization of yearly HYDRUS courses at the University of Hannover. Rien van Genuchten and Jirka Šimůnek always very much enjoyed visiting Hannover and lecturing at those courses. *Rienk van der Ploeg* was a very giving, collaborative, and humorous colleague who always encouraged us to use innovative methods to address soil and environmental problems such as those described in this publication. Thanks, *Rienk*, for the time that we could spend with you.

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