# Analysis of Soil Water Response to Grass Transpiration

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**Abstract**: This paper focuses on numerical modelling of soil water movement in response to the root water uptake that is driven by transpiration. The flow of water in a lysimeter, installed at a grass covered hillslope site in a small headwater catchment, is analysed by means of numerical simulation. The lysimeter system provides a well defined control volume with boundary fluxes measured and soil water pressure continuously monitored. The evapotranspiration intensity is estimated by the Penman-Monteith method and compared with the measured lysimeter soil water loss and the simulated root water uptake. Variably saturated flow of water in the lysimeter is simulated using one-dimensional dual-permeability model based on the numerical solution of the Richards' equation. The availability of water for the root water uptake is determined by the evaluation of the plant water stress function, integrated in the soil water flow model. Different lower boundary conditions are tested to compare the soil water dynamics inside and outside the lysimeter. Special attention is paid to the possible influence of the preferential flow effects on the lysimeter soil water balance. The adopted modelling approach provides a useful and flexible framework for numerical analysis of soil water dynamics in response to the plant transpiration.

**Keywords**: evapotranspiration; root water uptake; water stress function; Richards' equation; preferential flow; dual-permeability model; lysimeter

The improvement of our current understanding of the soil-plant-atmosphere interactions represents one of the most important research objectives of the relevant disciplines of natural, agricultural, and technical sciences (i.e. hydrology, plant physiology, meteorology, climatology, etc.). An adequate description of the mass, momentum, and energy fluxes on the land-atmosphere interface, including the transfer of water vapour through the atmospheric boundary layer and soil water movement in the rhizosphere, is of major importance in the studies of the soil retention capacity, vegetation productivity, land use change impacts, climate change feedbacks, and many others.

Quantitative analysis of the flux of water through soil-plant-atmosphere system is often based on the concept of potential evapotranspiration, defined as the evaporative demand of the atmosphere and derived from the actual atmospheric conditions. While the potential evapotranspiration is supposed

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to be controlled exclusively by meteorological factors, the actual evapotranspiration is limited mainly by the supply of water to the plant roots.

Actual evapotranspiration can be measured either as the rate of the loss of water from the land surface or as the rate of the gain of water vapour by the atmosphere. The former approach is often carried out by means of soil lysimeters. Although lysimeters are considered to be relatively accurate devices, they provide the measure of the water loss from a very limited sample area. Therefore, more representative values of evapotranspiration can be obtained only when the plant density, height, vitality, and leaf area of vegetation inside the lysimeters are similar to those in the surrounding area. The differences between the conditions inside and outside the lysimeter, such as the soil characteristics, moisture availability, soil temperature, and rooting characteristics, are the most frequent causes of dissimilarity.

One of the first quantitative studies of the water extraction by plant roots was published by Gardner (1960), who presented a microscopic model of individual rootlet extraction. Since the detailed geometry of the root system and its development in time are usually unknown, using a macroscopic approach is more appropriate. Several more or less sophisticated models of the root water uptake focus on the macroscopic formulation of the extraction term in Richards' equation. This approach has led to the formulation of the extraction function by Molz and Remson (1970), a semi empirical function including soil and plant resistance to water uptake by Rowse *et al.* (1978), modification of Nimah's and Hanks' (1973) water transport model by Feddes *et al.* (1974), and a simplified extraction function by Feddes *et al.* (1976, 1978). The degree of domination of the root resistance in relation to the extraction function and the role of the root morphological parameters is widely discussed by Molz (1981), Novák (1994), Jackson *et al.* (1996), and others.

More complex models accounting for the process of the growth and development of the root system have been developed e.g. by Clausnitzer and Hopmans (1994) and Wilderotter (2003). Vertical variability in the root-density distribution, root-uptake efficiency, and atmospheric demand are widely investigated in the recent literature (e.g. Lai & Katul 2000).

The main objective of our study is to refine, test, and assess one of the existing approaches to modelling the transpiration-driven soil water movement. Our choice of the modelling approach is based on reasoning that the resulting model should be relatively complex as to reflect the complicated soil-plant-atmosphere interactions, yet reasonably realistic in terms of the input information required. The selected approach is applied to the controlled volume exposed to natural conditions, represented by a field lysimeter.

In the approach presented, the atmospheric demand for evaporation is evaluated using the conventional concept of potential evapotranspiration. This is done independently of the soil water flow model. The availability of water for the root water uptake is than determined by the evaluation of the plant water stress function, incorporated directly in the soil water flow model.

Three different lower boundary conditions are tested to asses the functionality of the lysimeter and to compare the soil water response to transpiration inside and outside the lysimeter. The impact of the preferential flow on the soil water dynamics is taken into account through dual-permeability formulation of the governing equation of variably saturated soil water flow.

## MATERIAL AND METHODS

### Experimental site

The experimental site is located in a small headwater catchment Uhlířská. The catchment is situated in the Jizera Mountains, North Bohemia. The drainage area of the catchment is 1.87 km$^2$, whose significant part was deforested in early 80's. The average altitude is 822 m above sea level. The average annual temperature is 6.5°C. With the mean annual precipitation of 1379 mm and more than 200 rain days in a year, the catchment belongs to the most humid regions in the Czech Republic (Czech Hydrometeorological Institute 1999).

The deforested part of the catchment is covered with grass (*Calamagrostis villosa*) and young spruce trees. The shallow hillslope soil, classified as Dystric Gleyic Cambisol, turns into Histosol on the valley floor. A typical hillslope soil profile consists of about 10–20 cm of black peaty layer, 0–10 cm of greyish gleyic layer, 30 cm of ocher loamy-sand layer, and 30 cm of yellowish-brown layer gradually changing into weathered granite bedrock.

The time period of interest (vegetation seasons 2002 and 2003) was relatively warm and dry. For all months in the seasons studied, the precipitation total was below the long-term average, except August and October 2002, and October 2003. The hydrologic year 2003 was the driest year since 1961. The precipitation total reached only 65% of the long-term average.

### Lysimeters

Two subsurface lysimeters have been installed to monitor the soil water fluxes and plant transpiration at the experimental site in Uhlířská (Tachecí 2002). Both lysimeters were put into operation in spring 1998. They are of cylindrical shape with the surface area of 1 $m^2$. One of them is 35 cm deep and the other 70 cm deep. The lysimeter setup (Figure 1) follows the Thornthwaite concept (Gangopadhyaya *et al.* 1966). Both lysimeters were filled with the original soil (which had been put aside during the excavation of the lysimeter pit). The vertical wall is made of impermeable plastic foil. Grass, growing in lysimeters, is identical to the typical species of the site.

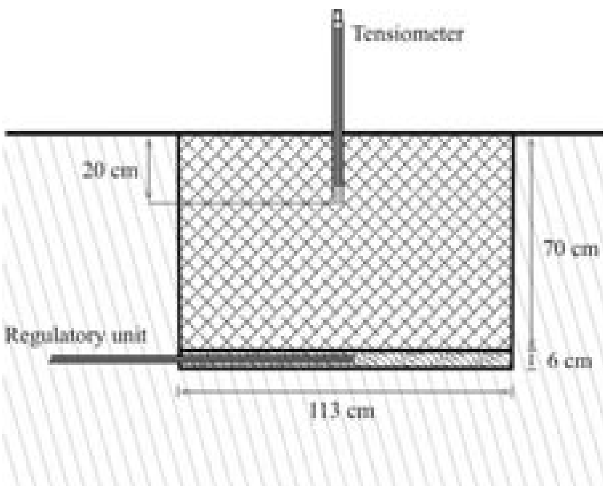


Figure 1. Schematic of the lysimeter

The bottom part of the lysimeters is filled with a drainage/irrigation layer made of sandy gravel. The layer was designed to keep the bottom of the lysimeter permanently saturated to compensate for a capillary rise during rainless periods. This is achieved through a regulatory unit, which is connected to the supply tank. The water recharge is monitored by a pressure transducer (measuring the water level fluctuations in the tank). The outflow from lysimeters is measured by a tipping bucket flowmeter.

Precipitation data are measured continuously by tipping bucket ombrometer near the lysimeters. The soil water pressure inside the lysimeters is measured by tensiometers inserted in the depth of 20 cm. Additional tensiometers are installed in the undisturbed soil near the lysimeters, for comparison. The measurements are automatically recorded by a datalogger at 10-minute intervals (Šanda *et al.* 2005).

All the input and output fluxes are measured (both at the top and the bottom of the lysimeter), except the evapotranspiration from the lysimeter surface, which can be determined indirectly from the water balance of the lysimeters.

The lysimeter water balance is given by:

$$LS_i - LS_0 = LR_i + RT_i - LD_i - LE_i \qquad (1)$$

with lysimeter moisture storage difference on the left hand side (m), and cumulative input/output fluxes on the right hand side (m). In this equation, $i$ stands for the time level, $LS_0$ is the initial lysimeter moisture storage, LR is the lysimeter recharge via capillary rise, RT is the cumulative rainfall, LD is the lysimeter drainage and LE is the cumulative lysimeter evapotranspiration. In our study, the left hand side of the lysimeter balance equation was neglected (set equal to zero), since the single tensiometer reading did not provide sufficient information for the accurate determination of the instantaneous moisture storage. It is, however, evident that the relative importance of the storage difference in the lysimeter balance equation considerably decreases with time, and so does the error caused by this simplifying assumption.

### Estimation of potential evapotranspiration

The term potential evapotranspiration is used here in its most general sense, i.e. as evaporation that would occur from a well vegetated surface, when moisture supply to the plant roots was not limited (e.g. Chow *et al.* 1988).

The exact description of the energy, momentum, and mass transfer processes, associated with the transfer of water vapour from the evaporative surface to the atmosphere, is fairly complicated due to the complex vegetation, soil, and atmospheric factors. We use the Penman-Monteith method (Monteith 1965) to estimate the daily potential evapotranspiration from the meteorological observations available. The method is based on Penman solution, which was extended by including an additional vapour transfer coefficient, the surface resistance. The Penman-Monteith formula is usually written as:

$$Q_\lambda = \lambda \rho_w E = \frac{\Delta(R_n - Q_G) + \rho_a c_p \dfrac{p_{vs} - p_v}{r_a}}{\Delta + \gamma\left(1 + \dfrac{r_s}{r_a}\right)} \qquad (2)$$

where:
$Q_\lambda$ – latent heat flux (W/m$^2$)
$E$ – evapotranspiration intensity (m/s)
$R_n$ – net radiation supply (W/m$^2$)
$Q_G$ – ground heat flux (W/m$^2$), in our case taken equal to zero
$p_v$ – vapour pressure (Pa)
$p_{vs}$ – saturated vapour pressure (Pa)
$\rho_w$ – water density (kg/m$^3$)
$\rho_a$ – air density (kg/m$^3$),
$\lambda$ – specific latent heat of vaporisation (J/kg)
$c_p$ – specific isobaric heat capacity of air (J/kg/K)
$r_a$ – aerodynamic resistance (s/m)
$r_s$ – surface resistance (s/m),
$\Delta$ – slope of the saturated vapour pressure curve (Pa/K)
$\gamma$ – psychrometric constant (Pa/K)

The coefficient $r_s$ represents the combined resistance to the flow of water vapour through the plant stomata and the soil surface (Allen *et al.* 1998). The aerodynamic resistance $r_a$ is associated with the turbulent diffusion through the atmospheric boundary layer.

To estimate the effective values of the resistance parameters, highly approximative approaches are often used. In the present study, we estimate the resistance coefficients as follows: The surface resistance coefficient is (Monteith 1981):

$$r_s = \frac{r_1}{\mathrm{LAI}_{\mathrm{active}}} \qquad (3)$$

where:
$r_1$ – bulk stomatal resistance of a well-illuminated leaf (s/m), here assumed to be equal to 100 s/m
$\mathrm{LAI}_{\mathrm{active}}$ – index of the sunlit leaf area (m$^2$/m$^2$) that actively contributes to the surface heat and vapour transfer

Based on the measurements of the grass height during the vegetation season, the following empirical formula was developed by Herza (2005), based on the research by Holub (2000):

$$\mathrm{LAI}_{\mathrm{active}} = 5\left[(h_{\max} - h_{\min})\sin\left(\ln\left(1 + 8\frac{i}{N}\right)\right) + h_{\min}\right] \qquad (4)$$

where:
$h_{\max}$ – maximum grass height in the respective vegetation season (m)
$h_{\min}$ – height measured at the beginning of the season (m)
$i$ – stands for the sequential number of the day in the vegetation season
$N$ – duration of the vegetation season in days

The $h_{\max}$ and $h_{\min}$ values were set equal to 0.35 m and 0.05 m, respectively. The time development of the values of the leaf area index and the grass height is shown in Figure 2.

Assuming neutral stratification of the atmospheric boundary layer, the aerodynamic resistance is estimated using the formula (Shuttleworth 1993):

$$r_a = \frac{\ln\dfrac{(z_u - d)}{z_0}\ln\dfrac{(z_h - d)}{z_{0h}}}{k^2 u} \qquad (5)$$

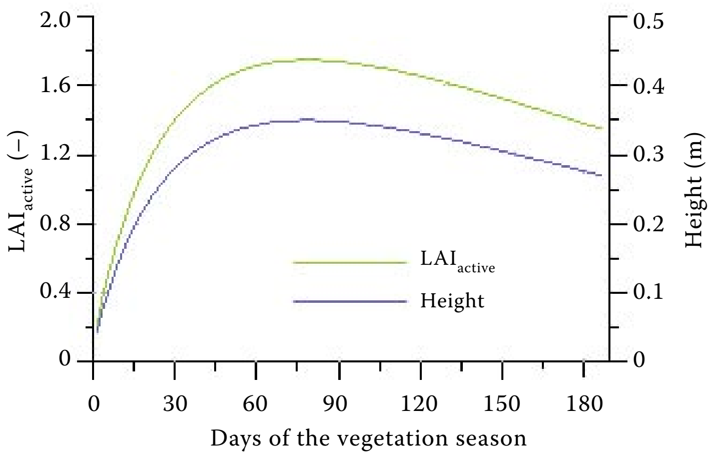


Figure 2. Development of active leaf area index and average plant height for *Calamagrostis villosa* during vegetation season

where:

$k$ – von Karman constant, usually taken as 0.41

$u$ – wind velocity at the height $z_u$

$z_h$ – height of the humidity measurements

$z_0$, $z_{0h}$ – roughness lengths of the momentum transfer and the transfer of vapour, respectively (m)

$d$ – zero plane displacement height (m)

The following empirical relationships were used to evaluate $r_a$:($z_0$, $z_{0h}$, $d$) = ($0.123h_0$, $0.0123h_0$, $0.67h_0$), where $h_0$ is the height of the vegetation. The application of the equation (5) for short time periods (not our case) may require the inclusion of stability corrections for stable or unstable stratification of the atmospheric boundary layer.

Equation (2) was evaluated for each day of the vegetation season, based on the mean daily values of the meteorological variables. Whenever possible, the mean daily vapour pressure deficit was calculated from hourly temperature and humidity measurements, and the mean daily net radiation from the measurements taken at ten-minute intervals.

The missing net radiation data in the season 2002 were replaced by estimates obtained by the Angstrom method (Martinez-Lozano *et al.* 1984). According to this method, the total incoming shortwave solar radiation $R_s$ (W/m$^2$) is calculated from the extraterrestrial radiation $R_a$ (W/m$^2$) by means of a simple empirical formula:

$$R_s = (a_s + b_s \frac{n}{N}) R_a \tag{6}$$

where:

$n$ – actual duration of the sunshine (h)

$N$ – maximum possible duration of the sunshine (h)

$a_s$. $a_s + b_s$ – fractions of the extraterrestrial radiation reaching the earth surface on overcast and clear days, respectively

The extraterrestrial radiation is calculated from the geographic latitude and the solar declination. In our analysis, we used the values of $a_s$ and $b_s$ recommended for the average climate ($a_s$ = 0.25, $b_s$ = 0.50).

The subsequent calculation of the net radiation $R_n$ followed the procedure recommended e.g. by Shuttleworth (1993) or Allen *et al.* (1998) (with the value of surface albedo equal to 0.24).

The missing humidity data of the same period were estimated from daily minimum air temperatures. Detailed procedures for estimating the missing data at the site of interest are discussed by Herza (2005).

## Root water uptake

The root water uptake due to transpiration is modelled through the sink term, $S$, in the governing equation of the soil water flow (equation (15)). Feddes *et al.* (1978) defined $S$ as:

$$S(h) = a(h)S_p \tag{7}$$

where:

$a(h)$ – prescribed dimensionless function of pressure head $h$ ($0 \le a(h) \le 1$), called plant water stress function

$S_p$ – potential water uptake rate in the root zone (s$^{-1}$)

$S_p$ is equal to the actual soil water uptake rate during the periods of no water stress when $a(h) = 1$, which also means (if satisfied everywhere in the root zone) that the actual plant transpiration is equal to the potential transpiration. When the potential water uptake rate is equally distributed over the root zone, $S_p$ becomes, according to Feddes *et al.* (1978):

$$S_p = \frac{1}{z_R} E_p \tag{8}$$

where:

$E_p$ – potential transpiration rate (m/s)

$z_R$ – depth of the root zone (m)

Equation (8) can be generalised to account for the non-uniform distribution of the potential water uptake rate over the root zone (Vogel 1987):

$$S_p = b(z)E_p \tag{9}$$

where:

$b(z)$ – normalised potential root water uptake distribution function (m$^{-1}$)

This function describes the spatial variation of the potential extraction rate, $S_p$, over the root zone, and is given by:

$$b(z) = \frac{b'(z)}{\int_{\Omega_R} b'(z)\,dz} \tag{10}$$

where:

$\Omega_R$ – area occupied by the root zone

$b'(z)$ – prescribed distribution function (possibly taken as proportional to the average root density distribution function)

Normalising the water uptake distribution according to (10) ensures that $b(z)$ integrates to unity over the flow domain. From the equations (9) and (10) it follows that $S_p$ is related to $E_p$ by:

$$\int_{\Omega_R} S_p\, dz = E_p \tag{11}$$

The actual water uptake distribution is obtained by substituting (9) into (7):

$$S(h,z) = a(h)b(z)E_p \tag{12}$$

whereas the actual transpiration rate, $E_a$, is obtained by integrating (12) as follows:

$$E_a = \int_{\Omega_R} S\, dz = E_p \int_{\Omega_R} a(h,z)\, b(z)\, dz \tag{13}$$

The argument $z$ in $a(h, z)$ indicates the possible variation of the shape of the function $a$ with depth e.g. due to the soil stratification.

Figure 3 shows two alternative water stress functions recommended by Feddes *et al.* (1978) and van Genuchten (1987), respectively. In the former stress function, the pressure head interval ($h_3$, $h_2$) defines the range of optimal soil water conditions for the plant transpiration (no water stress), $h_4$ is the plant wilting point, and the interval ($h_2$, $h_1$) defines the range of the reduced transpiration due to the lack of aeration in the root zone. The position of $h_3$ depends on the intensity of the potential transpiration. The latter stress function is given by:

$$a(h) = \frac{1}{1 + \left(\dfrac{h}{h_{50}}\right)^P} \tag{14}$$

where:

$h_{50}$ – pressure head at which the extraction rate is reduced by 50%

$P$ – empirical parameter determining the slope of the curve

A similar equation can be used also for the salinity stress function (van Genuchten & Hoffman 1984).

## Soil water flow model

The movement of water in a variably saturated soil profile with preferential pathways is modelled by dual set of Richards' equations (Gerke & van Genuchten 1993a; Vogel *et al.* 2000):

$$\frac{\partial \theta_i}{\partial t} = \frac{\partial}{\partial z}\left( K_i \left( \frac{\partial h_i}{\partial z} + 1 \right) \right) - S_i - \Gamma_{wi} \tag{15}$$

in which:

$$\Gamma_w = w_f \Gamma_{wf} = -w_m \Gamma_{wm} \tag{16}$$

where:

subscript $i$ – equal to $m$ for the soil matrix domain and $f$ for the preferential flow domain (often called macropore domain)

$h_i$ – soil water pressure head (m)

$K_i$ – unsaturated hydraulic conductivity (m/s)

$\theta_i$ – volumetric water content (m$^3$/m$^3$)

$S_i$ – sink term, used to account for the root water uptake (s$^{-1}$)

Both last variables stated are defined per unit volume of a particular domain. $\Gamma_w$ is the soil water

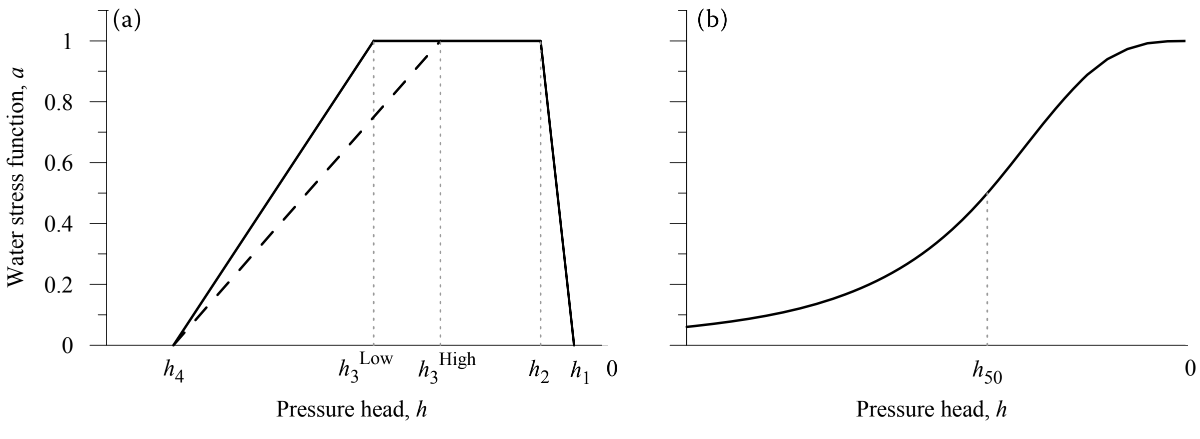


Figure 3. Plant water stress function $a$ ($h$) as suggested by: (a) Feddes *et al.* (1978), and (b) van Genuchten (1987)

transfer term ($s^{-1}$), which is defined as the volume of water moving from the preferential flow domain to the matrix domain per unit volume of the bulk soil per unit time. The terms $w_m$ and $w_f$ are the volume fractions of the matrix domain and the preferential flow domain per unit volume of the bulk soil, respectively, with $w_m + w_f = 1$.

The hydraulic functions $\theta(h)$, $K(h)$ and $S(h)$ are defined separately for each of the two flow domains. The composite water content $\theta$ is equal to $\theta_m w_m + \theta_f w_f$.

The soil water transfer term is dynamically varying proportionally to the pressure difference between the soil matrix and the macropore domain. The following first-order approximation is used to estimate the rate of the water exchange:

$$\Gamma_w = \alpha_w (h_f - h_m) \tag{17}$$

in which

$$\alpha_w = \alpha_{ws} K_{ar}(\theta_a) \tag{18}$$

where:

$K_{ar}$ – effective relative conductivity of the interface separating the soil matrix domain from the preferential flow domain (Gerke & van Genuchten 1993b)

$\alpha_{ws}$ – transfer coefficient at full saturation

$\theta_a$ – effective saturation of the interface

The dual set of Richards' equations is solved numerically by the computer code S1D DUAL (Vogel 1999, Documentation of the S1D DUAL code – version 1.1, CTU Prague, internal report), which is the extended version of the HYDRUS 5 code (Vogel *et al.* 1996).

**Simulation of soil water movement**

The S1D DUAL code was used to simulate the soil water movement in the deeper of the two lysimeters. The simulations were carried out for the vegetation seasons 2002 and 2003.

The initial pressure head distribution, required for the numerical simulations, was derived from the tensiometric observations. The upper boundary condition involved the instantaneous rainfall intensities and the potential evapotranspiration rates calculated by Penman-Monteith method. The precipitation intensities observed were resampled to form a 1-hour series. The potential evapotranspiration data were organised in a daily series. The root water uptake distribution function, $b(z)$, was approximated by a simple piecewise linear relationship, with maximum value between the depth of 1 cm and 20 cm below the soil surface and zero at the depth of 40 cm and deeper. This corresponds quite well to the measured root density in the soil profile, which is often used as an indicator of the distribution of the root water uptake.

The piecewise linear distribution model of $b(z)$ used in this study is but one of several frequently used models (such as constant, linear decrease and exponential decrease models). As proved by the results of numerical experiments (not presented in this paper), the choice of the root water uptake distribution model is not crucial, provided that the basic geometric properties of these distributions are similar (especially the vertical range and the position of the centre of mass).

A more accurate formulation of the atmospheric evaporative demand would require disaggregating the daily potential evapotranspiration values into hourly values (for instance, to allow for the differences between day and night). The effect of such disaggregation was tested for the year 2003 (in which highly resolved net radiation data were available) and found to be insignificant. This is probably due to the fact that the effect of the time distribution of the evaporative demand on the root water uptake is moderated by the soil water response in the root zone. Greater effects can be expected for dryer conditions where the vegetation is exposed to more severe moisture stresses.

Two different lower boundary conditions were tested for the lysimeter system (constituting two basic simulation scenarios): (i) the zero pressure boundary condition and (ii) the seepage face boundary condition. The first simulation scenario, in which the soil water pressure was set equal to zero at the bottom of the lysimeter, represents the situation of the perfect function of the lysimeter regulatory unit (i.e. in full agreement with the lysimeter design). The second scenario represents the situation in which the unit fails to keep the bottom of the lysimeter saturated. The seepage face condition allows water to drain from the lysimeter in the wet parts of the season (under the atmospheric pressure), and assumes zero flux when the bottom of the lysimeter becomes unsaturated during dry parts of the season.

Note, that neither of the two simulation scenarios made use of the measured bottom flux information. The reason for omitting this information from simulations was twofold: (i) relatively low accuracy of the measured fluxes, due to the problems connected with construction/maintenance of the regulatory unit (closely related to rather extreme weather conditions at the site), and (ii) the expected presence of preferential flow effects, reported in the previous studies (Císlerová *et al.* 1997). The possible impact of the preferential flow on the water transfer through the lysimeter was assessed separately by comparing the drainage fluxes obtained from the dual-permeability approach with those simulated using the single-permeability approach.

An additional scenario was set up to simulate the soil water flow in the natural soil profile outside the lysimeter. In this case, the vertical flow domain was slightly extended below the bottom of the lysimeter to reach next (less permeable) soil horizon. The lower boundary condition was formulated as zero pressure gradient condition (commonly called free-drainage condition). This simulation scenario is henceforward referred to as the reference scenario.

Since all the simulations were done for the fully developed vegetation cover, the contribution of the soil surface evaporation to the combined soil and plant evaporation was minor. In principle, this contribution could be incorporated in the model via the surface flux boundary condition; however, in our study we assumed that the actual evapotranspiration was equal to the actual transpiration.

In general, there is a great uncertainty associated with the determination of the soil hydraulic properties of the partially consolidated soil materials in lysimeters (unless undisturbed soil column is used instead of the repacked material). The values of the soil hydraulic parameters, used in our numerical analysis (Table 1), were adopted from the previous studies on the soil water flow in several locations of the same experimental site (Šanda 1999; Vogel *et al.* 2003). For the modelling purposes, we assumed that these parameters were also valid for the lysimeter soil profile. The parameters were not optimised for the lysimeter system (i.e. the model was not calibrated). The additional parameters needed for the dual-permeability formulation were estimated to be ($w_f$, $\theta_{ws}$) = (0.05, 0.01).

The fllowing values of the plant water stress function parameters were used: ($h_4$, $h_3^{\mathrm{Low}}$, $h_3^{\mathrm{High}}$, $h_2$, $h_1$) = (–12 000, –700, –600, 0, 0) cm, respectively. The most important stress function parameters $h_3^{\mathrm{Low}}$ and $h_3^{\mathrm{High}}$ were set equal to the values, which correspond well with the previously reported values obtained under similar conditions (Tesař *et al.* 2003; Kroes & van Dam 2003).

## RESULTS AND DISCUSSION

The potential evapotranspiration (ETp), estimated by Penman-Monteith method, was compared with the actual evapotranspiration from the lysimeters (ETa), calculated from the measured components of the soil water balance. The comparison was made for two successive vegetation seasons 2002 and 2003, of which the season 2003 was considerably dryer then the preceding one. The results of the comparison, as obtained for the deeper lysimeter, are shown in Figures 4 and 5. Due to the lysimeter design, relatively small differences between ETp and ETa were expected (especially in the wet season), however, this expectation is not fully consistent with the reality. While the estimated ETp gradually increases during the season, the calculated lysimeter ETa shows periods with negligible or even negative rates as well as instances of fast increases of ETa following major rainfall events.

Table 1. Soil hydraulic parameters for the soil matrix domain – M, and the preferential flow domain – PFD

| Domain | Layer | Depth (cm) | $\theta_r$ (–) | $\theta_s$ (–) | α (cm$^{-1}$) | $n$ (–) | $K_s$ (cm/day) | $h_s$ (cm) |
|---|---|---|---|---|---|---|---|---|
| M | 1 | 0–8 | 0.20 | 0.69 | 0.050 | 2.00 | 1700 | 0.00 |
| | 2 | 8–20 | 0.20 | 0.67 | 0.050 | 1.50 | 200 | –0.69 |
| | 3 | 20–70 | 0.20 | 0.61 | 0.020 | 1.20 | 50 | –1.48 |
| | 4 | 70–75 | 0.20 | 0.51 | 0.020 | 1.20 | 4 | –1.88 |
| PFD | – | 0–75 | 0.045 | 0.43 | 0.145 | 2.68 | 1000 | 0.00 |

$\theta_r$, $\theta_s$ – residual and saturated soil water contents; α, $n$ – van Genuchten's empirical parameters; $K_s$ – saturated hydraulic conductivity; $h_s$ – air entry value – for details on this parameterisation see Vogel *et al.* (2001)

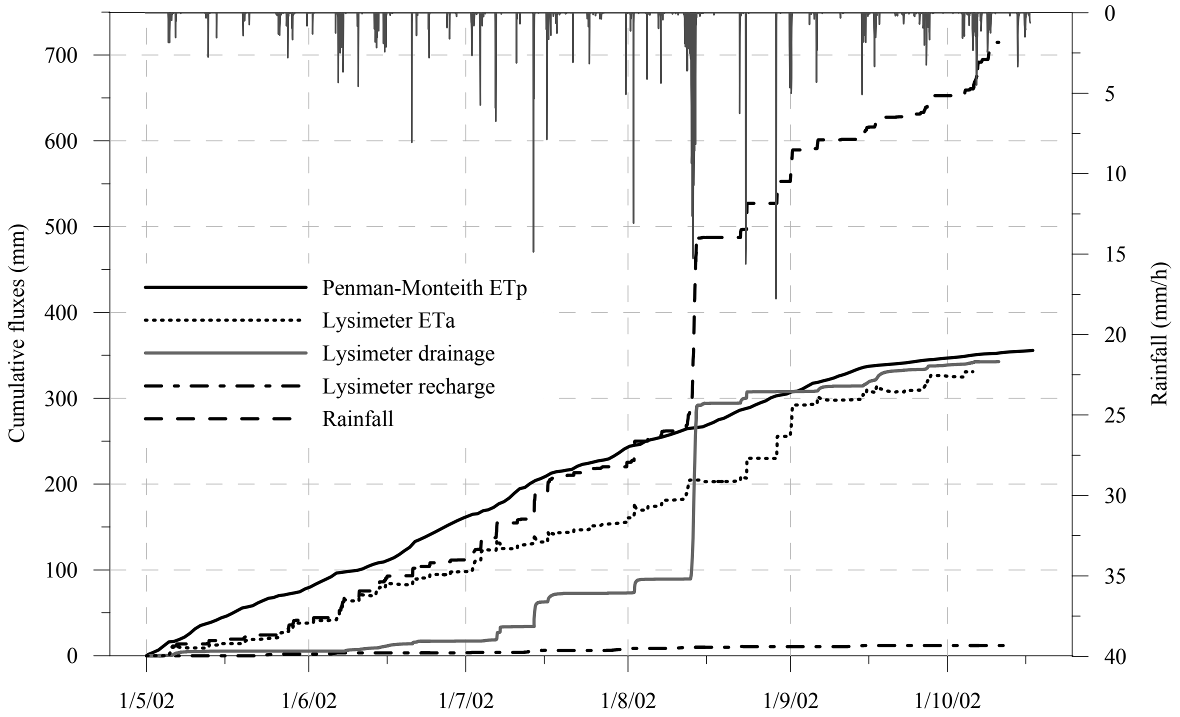


Figure 4. Components of the lysimeter water balance in 2002; the actual evapotranspiration (ETa) is determined as evaporation loss in the balance equation; the lysimeter recharge takes place through the capillary rise from the bottom of the lysimeter

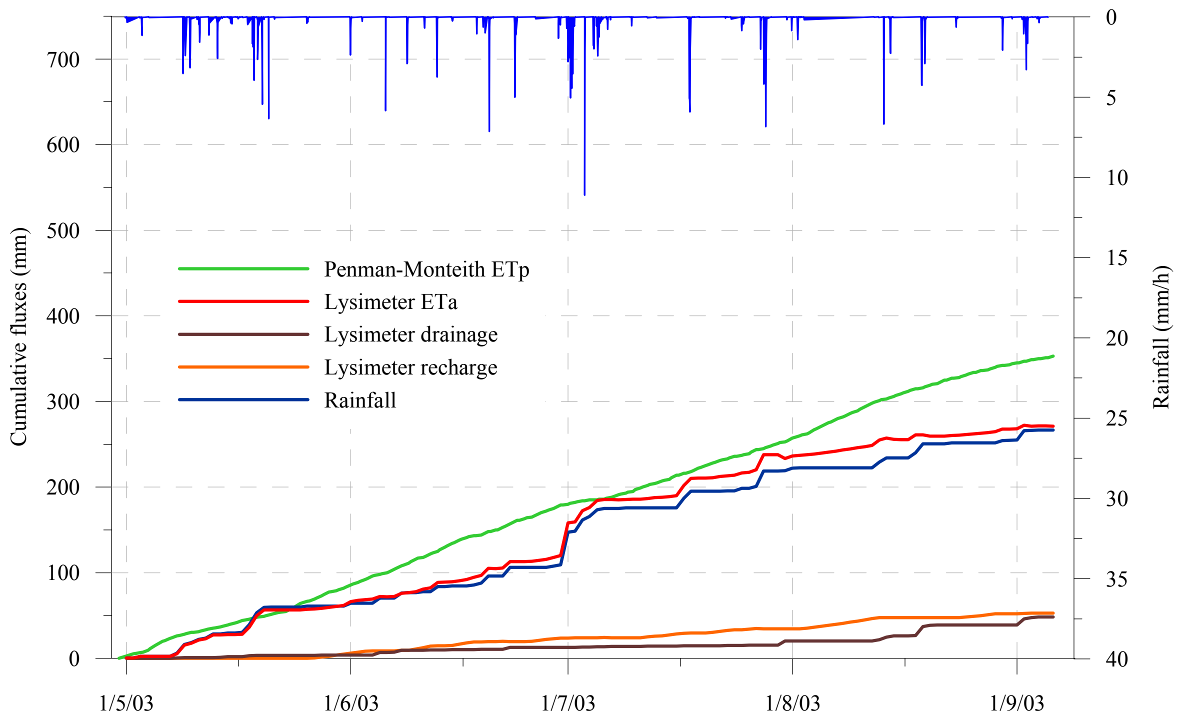


Figure 5. Components of the lysimeter water balance in 2003

During the dry periods, the artificial bottom water supply is apparently insufficient, and the plants are exposed to a moderate water stress. Consequently, the ETa is smaller than ETp.

Additional ETa estimates were generated by the S1D DUAL code for both lysimeter scenarios as well as the reference scenario, representing the natural conditions outside the lysimeter. The resulting cumulative ETa rates differ only slightly from the estimated ETp. The model predicts a moderate reduction of transpiration (13 mm in total) during the season 2003 (i.e., Eta < ETp), due to the water stress, while practically no reduction is predicted for the season 2002 (the grass is transpiring optimally).

The measured and the simulated soil water pressure variations in the deeper lysimeter for the vegetation seasons 2002 and 2003 are shown in Figures 6 and 7, respectively. Had the lysimeter been functioning according to the design, the simulation scenario involving zero pressure bottom boundary condition should have provided a closer agreement between the measured and the simulated pressures than the scenario based on the seepage face boundary condition. Since this was obviously not the case, the simulation results support our suspicion that the water supply from the regulatory unit is, during relatively dry periods, insufficient. None of the two scenarios performed well for the extremely wet period of September 2002. This could, perhaps, be explained by the clogging of the drainage layer at the bottom of the lysimeter, leading to almost complete saturation of the lysimeter.

The failure of the water supply is probably related to the hydraulic properties of the drainage layer at the bottom of the lysimeter, which failed to provide a sufficient capillary contact with the soil in the lysimeter, so that the lower boundary of the lysimeter may have become unsaturated during the dryer periods of the season.

In addition to the pressure development in the lysimeter, Figures 6 and 7 also show the simulated variation of the soil water pressure outside the lysimeter using the free-drainage boundary condition (i.e. for the reference simulation scenario). The simulated pressures are superimposed over the pressure variations measured which reflects

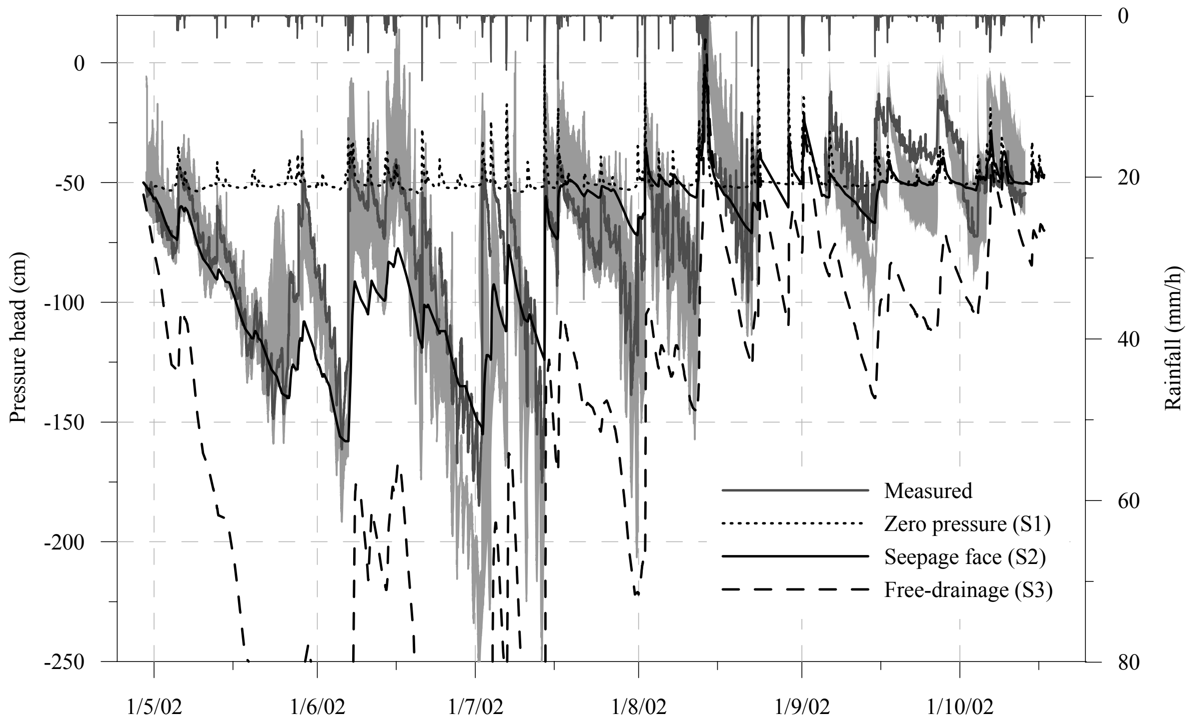


Figure 6. Comparison of observed and simulated variations of soil water pressure in the lysimeter and in the reference soil profile; three simulation scenarios (denoted as S1, S2 and S3) are associated with different boundary conditions applied at the bottom of the lysimeter (S1 and S2) and at the lower boundary of the reference soil profile outside the lysimeter (S3); the shaded area represents the envelope of the pressure heads observed at several locations nearby the lysimeter at approximately the same depth (about 20 cm)

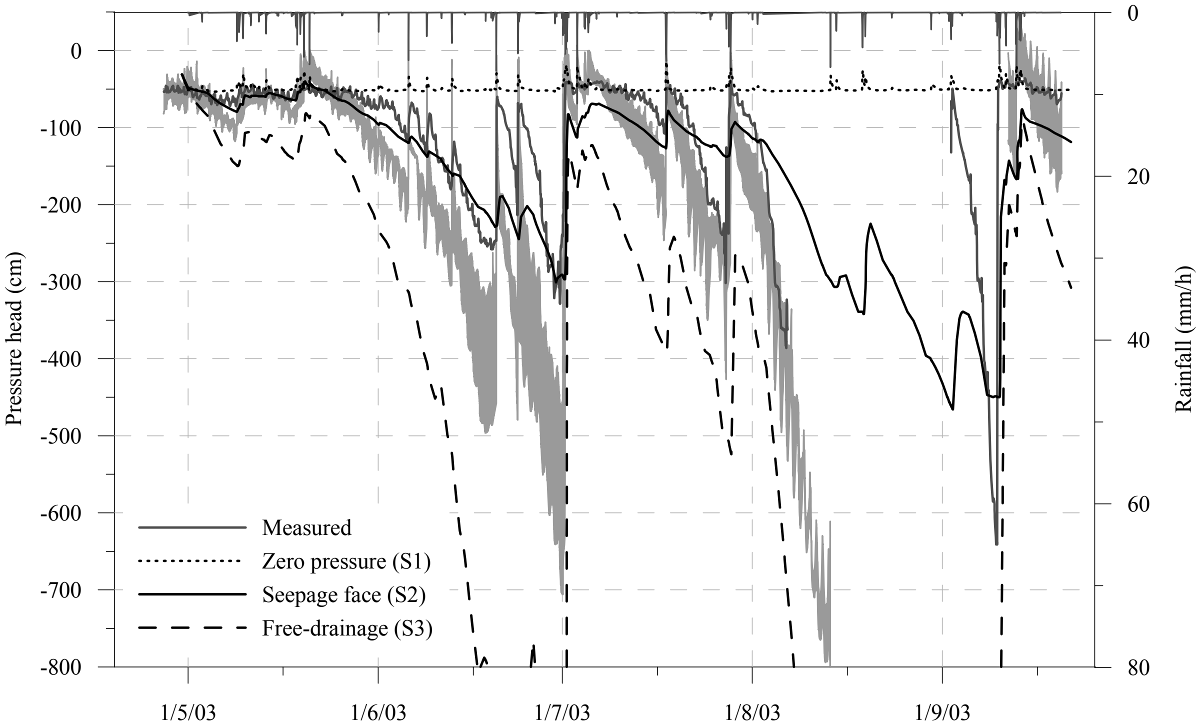


Figure 7. Comparison of observed and simulated variations of soil water pressure in 2003

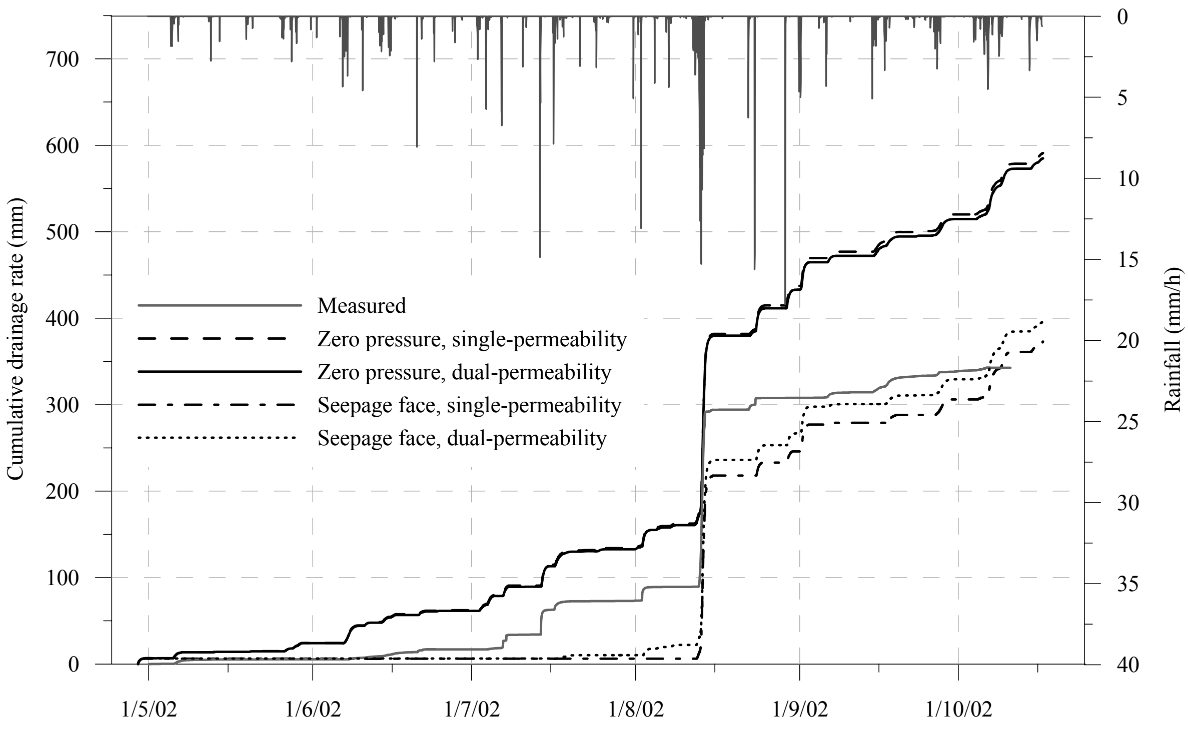


Figure 8. Measured and simulated cumulative drainage rates from the lysimeter; two different boundary conditions (seepage face and zero pressure) as well as two alternative flow approximations (single-permeability and dual-permeability) are compared

the spatial variability of the soil water pressure outside the lysimeter.

When comparing the simulation results obtained by means of the two lysimeter scenarios (denoted in the figures as S1 and S2, respectively) with the results of the reference scenario (S3), we can see that the lysimeter scenario involving the seepage face boundary condition (S2) produces a variation pattern of the soil water pressure quite similar to that of the reference scenario based on the free-drainage boundary condition (S3). This is true especially for the wet season 2002 and less so for the dryer season 2003. Generally, the application of the seepage face and free-drainage conditions leads to quite distinct soil water dynamics. In our particular case, however, the similarity is caused by the fact that the lower boundary of the reference system is located in the soil horizon with considerably lower hydraulic conductivity than is the conductivity at the bottom of the lysimeter (cf. Table 1).

The observed and the simulated drainage discharges from the lysimeter are compared in Figure 8. The figure shows that the drainage rates are consistently underestimated when the seepage face boundary condition is applied, and overestimated when the zero pressure head condition is used. This also corresponds well with the above discussed comparison between the observed and the simulated soil water pressure responses. The figure also suggests that the dual-permeability model combined with the seepage face lower boundary condition provides the best approximation of the lysimeter discharge measured. The difference between the single-permeability and dual-permeability flow systems is negligible in the case that the zero pressure boundary condition is used.

## CONCLUSIONS

Our study confirms that the adopted modelling approach provides a useful and flexible framework for numerical analysis of the soil water dynamics in response to the atmospheric forcing (precipitation input and transpiration demand).

The approach was tested by simulating the soil water flow in the lysimeter and comparing the computed and the measured soil water pressure heads and boundary fluxes. Two different lower boundary conditions were used for the bottom of the lysimeter: the seepage face and the zero pressure boundary condition. Although the latter of the two conditions seems to be more appropriate for the hydraulic design of the lysimeter, the application of the seepage face condition resulted in better model predictions. This was attributed to the influence of the hydraulic properties of the drainage layer at the bottom of the lysimeter, which failed to provide a sufficient capillary contact with the soil in the lysimeter, the lower boundary of the lysimeter becoming unsaturated in the dryer periods of the season.

The soil water status in the natural soil profile outside the lysimeters was reasonably well simulated using the free-drainage boundary condition.

It was shown that the discharge rates from the lysimeter are simulated slightly better by the dual-permeability model, i.e. when the preferential flow effects are taken into account.

The macroscopic concept of the root water uptake applied, relying on the moisture stress function used in conjunction with the normalised spatial distribution function, seems to represent an adequate technique for the hydrological estimation of the actual plant transpiration. A wider use of this technique, however, requires further progress in the development of experimental methodology for the determination of the soil and plant specific root water uptake parameters (especially the moisture stress parameters).

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