Int. Agrophysics, 1998, 12, 33-36

EXCHANGE OF WATER IN THE SOIL-PLANT-ATMOSPHERE SYSTEM

M. Kutílek[1], *V. Novák*[2]

[1]Institute of Hydrology, Nad Patankou 34, 16400 Prague, Czech Republic
[2]Institute of Hydrology, Slovak Academy of Sciences, Racianska 75, P.O.Box 94, 83008 Bratislava, Slovak Republic,
E-mail: novakv@uh.savba.sk

A b s t r a c t. The developed methodology (the scaling procedure) was shown how to solve the problems of fluxes and of soil water storage in systems with heterogeneous soil-plant-atmosphere boudary.

K e y w o r d s: soil-plant-atmosphere system, exchange of water

INTRODUCTION

Solution of fluxes in the soil-plant-atmosphere system (SPAS) are dominantly influenced by two characteristic features:

a) physical heterogeneity in each part of the system,

b) the strongly non-linear type of equations describing fluxes in each part of the system.

All deterministic solutions dealing with the transports in the SPAS are then approximative. This contribution is not discussing detailed procedures but it tries to offer some new views upon the opportunities to solve fluxes and storage of water in the heterogeneous SPAS.

SCALING OF THE RICHARDS EQUATION (RE) IN EVAPORATION

When we consider evaporation from the bare soil as the simplest process on the boundary with the atmosphere, we are tackling two types of heterogeneities: Heterogeneity of the soil and heterogeneity of the boundary condition. The second one we are discussing here and the method of scaling of the Richards equation RE by the flux boundary condition is applicable. In addition to the dependence upon soil hydraulic characteristics, evaporation intensity depends on evaporative demand of the atmosphere, mostly on net radiation and air temperature [7]. Both are heterogeneous in space. The flux on the boundary is then non uniform in space, too, and we are usually measuring both the flux and the soil water content, and potential on one vertical. The 'extension' of our data to realistic areal scale is therefore plausible. We can do it by transforming the RE to the invariant form with regard to the variable flux on the boundary [4]. Hydraulic characteristics of the soil are expressed as power functions of the soil water content and of the residual soil water content θ_r. The functions are simply transformable to the equations of Brooks and Corey [1], BCE. Equations of van Genuchten [8], VGE, are only approximatively identical with our power functions and they are applicable for $\theta \leq 0.9\ \theta_s$, where θ_s denotes the saturated soil water content. Soil hydraulic characteristics are in our scaling procedure:

soil water diffusivity $D[L^2T^{-1}]$

$$D(\theta) = D_o(\theta - \theta_r)^g \tag{1}$$

retention curve with pressure head h $[L]$

$$h(\theta) = -p(\theta - \theta_r)^{-u} \tag{2}$$

and unsaturated hydraulic conductivity K $[LT^{-1}]$

$$K(\theta) = K_o(\theta - \theta_r)^v \quad (3)$$

D_o, K_o, g, p, u, v are fitting parameters which have functional relationships to parameters in BCE and VGE. Variables of the *RE* are scaled with regard to the flux q_o on the top boudary:

$$t = q_o^{\alpha} T^*, z = q_o^{\beta} Z^*, (\theta - \theta_r) = q_o^{\gamma} \theta^* \quad (4)$$

with α, β, and γ as scaling factors, t is time, z is the vertical coordinate, oriented from $z = 0$ on the surface downward. With (4) in Eqs (1) to (3) we obtain *RE* in the scaled form:

$$q_o^{\gamma-\alpha} \frac{\partial \theta^*}{\partial T^*} = q_o^{\gamma(1+g)-2\beta} \frac{\partial}{\partial Z^*}\left[D^*(\theta^*)\frac{\partial \theta^*}{\partial Z^*}\right] - q_o^{\gamma v-\beta} \frac{\partial K^*(\theta^*)}{\partial Z^*}. \quad (5)$$

In order to get (5) invariant in q_o and using scaled boundary conditions and scaled $D(\theta) = K(\theta)\, dh/d\theta$, we obtain:

$$\alpha = (-2u - g)/v, \beta = -u/v, \gamma = 1/v, v = g + u + 1. \quad (6)$$

The scaled RE invariant to the top boundary condition is then:

$$\frac{\partial \theta^*}{\partial T^*} = \frac{\partial}{\partial Z^*}\left[D^*(\theta^*)\frac{\partial \theta^*}{\partial Z^*}\right] - \frac{\partial K^*(\theta^*)}{\partial Z^*} \quad (7)$$

with boundary conditions:

$$T^* \le 0, \quad Z^* \ge 0, \quad \theta^* = \theta_s^*$$

$$T^* \ge 0, \quad Z^* \to \infty, \quad \theta^* = \theta_s^*$$

$$T^* \ge 0, \quad Z^* = 0,$$

$$1 = D^*(\theta^*)\frac{\partial \theta^*}{\partial Z^*} - K^*(\theta^*) \quad (8)$$

In addition to the solution of evaporation, the solution is applicable as approximative for evapotranspiration, too, when the leaf area index LAI $\le$ 0.8. The space heterogeneity is then increasing due to the non uniform distribution of plant canopy, and the importance of the proposed procedure is increasing, too.

SCALING OF THE SIMPLIFIED RE IN EVAPOTRANSPIRATION

When the soil is covered by vegetation and LAI $\ge$ 0.8, the RE is not fully adequate to the deterministic description since $\theta(z,t)$ as well as the fluxes in the soil are influenced by the uptake of water by the plant roots. The net of roots in the soil creates a system of 'channels' in which water is flowing at velocities by orders higher than in the neighbouring soil [7]. Roots are a new kind of heterogeneity. The continuity equation is therefore completed by the extraction term S and:

$$\frac{\partial \theta}{\partial t} = -\frac{\partial q}{\partial z} - S(z,t). \quad (9)$$

In order to get rid of complicated term of extraction, we substitute the soil water content θ by the soil water storage W $[L]$. Thus we average the heterogeneous q_o and θ fields in the soil-root system:

$$W = \int_o^L \theta \, dz \quad (10)$$

where L is the depth of the main rooting zone. The procedure is applicable mainly for cultural plants. Then [5]:

$$\frac{\partial W}{\partial t} = -K(W)\frac{\partial H}{\partial z} \quad (11)$$

where the total potential $H = h(W) + z$. [2] have demonstrated that the known various forms $K(\theta)$ are applicable for $K(W)$, too. We express the term $K(W)$ as the sum of resistances in SPAS, $K(W) = L/R(W)$ and:

$$R(W) = \frac{L}{\overline{K}_{soil}(W)} + R_p + R_o \quad (12)$$

where $\overline{K}_{soil}(W)$ is the mean weighted hydraulic conductivity of the soil over the depth (O,L), R_p

is the resistance in cells of plants, usually negligible and therefore the length of the path of the flux of water from the root up to the canopy can be neglected. R_O is the regulated resistance of stomatal openings, in the first approximation R_O (W). In (4) we replace the expression of $(\theta\text{-}\theta_r)$ by water storage $W = q_o^{\gamma} W^*$. Further on:

$$K(W) = K_o W^{\nu}$$

and in scaled form

$$K(W) = q_o^{\nu\gamma} K^*(W^*) \qquad (13)$$

$$h(W) = -pW^{-u}$$

and

$$h(W) = q_o^{-u\gamma} h^*(W^*). \qquad (14)$$

Equation (11) is then transcribed into the scaled form:

$$q_o^{\gamma-\alpha} \frac{\partial W^*}{\partial T^*} = q_o^{\nu\gamma} K^*(W^*) q_o^{\beta-u\gamma}$$

$$\frac{\partial h^*(W^*)}{\partial Z^*} - q_o^{\nu\gamma} K^*(W^*) \qquad (15)$$

With the top boundary condition:

$$T^* \geq 0,$$

$$q_o = -\frac{\partial W^*}{\partial T^*} q_o^{\gamma-\alpha} \qquad (16)$$

we obtain for the condition of invariancy of Eqs (15) and (16) to the flux on the top boundary q_O the scaling parameters:

$$\beta = u/\nu, \ \gamma = 1/\nu, \ \alpha = (1-\nu)/\nu. \qquad (17)$$

The scaled equation is then invariant to the flux on the boundary. If we are measuring at intervals in depth and time, we are frequently replacing differentials by finite differences in (15), and we continue further on in following approximations: The soil water pressure head h (matrix potential) is measured at $z = L/2$, ∂z is replaced by L and $\partial Z^* = q_o^{\beta} L^*$.

The term $\partial h^*(W^*)/\partial Z^*$ is then replaced by $\Delta h^*/L^*$ with Δh^* being the scaled form of the difference of the pressure head at the centre of the root zone and pressure head (potential) at the transpiring surface, i.e., $(h_{atL/2} - h_{atm})$. We are assuming that h_{atm} is given by the atmospheric conditions at the top level of the canopy.

CONCLUSIONS

The scaling procedure enables to solve following problems:

1. Space structure of time T_R when the potential evapotranspiration ET_p is reduced to actual evapotranspiration ET_A, where ET_p is defined as the flux q_o on the boundary. Since ET_{pi} as well as q_{oi} at location i are space-variable due to local climate and microclimate, due to the non-monocultural character of the vegetation and due to the non-uniform plant growth at i, we expect space variability of T_{Ri}. This variability can be estimated from scaled T^*, once T_R is known together with estimates of soil hydraulic functions at one pedon-scale location.

2. In situation of Chapter 1 (evaporation) we can estimate soil water distribution with the depth $\theta(z)$, which is either measured, or numerically estimated on one vertical (pedon scale). From this known $\theta(z)$ for the given q_o we estimate $\theta(z)$ by the scaled RE for all locations for which actual fluxes q_o are either measured or estimated from meteorologic data.

3. For each measured $\theta(z)$ and known physical characteristics of the soil, the flux on the boundary $q_O(=ET_A)$ is estimated using the methods of inverse problem solution, see Chapter 1 (evaporation).

4. Hydraulic parameters of the soil-root zone $K(W)$ and $h(W)$ are not identical with independently measured parameter, of soil only (i.e., when the sink term due to absorption of water by roots is not considered). If $W(t)$ in soil with vegetation is measured together with $h_{atL/2}$ and with h_{atm}, we can estimate $K(W)$ and $h(W)$ functions of the soil-root zone applying the inverse procedure to the scaled approximative equation of Chapter 2 (evapotranspiration). These characteristics are usable for estimation of T_R, etc.

The developed methodology offers more applications in solving the problems of fluxes and of soil water storage in systems with heterogeneous soil-plant-atmosphere boundary.

REFERENCES

1. **Brooks R.H., Corey A.T.:** Hydraulic properties of porous media. Hydrology paper 3, Colorado St. Un., Fort Collins, 1964.
2. **Katul G.G., Wendroth O., Parlange M.B., Puente C.E., Folegatti M.V., Nielsen D.R.:** Estimation of in situ hydraulic conductivity function from nonlinear filtering theory. Water Resour. Res., 29, 1063-1070, 1993.
3. **Kutílek M., Nielsen D.R.:** Soil Hydrology. Geoecology Textbook. Catena Verlag, Cremlingen-Destedt, Germany, 1994.
4. **Kutílek M., Zayani K., Haverkamp R., Parlange J.Y, Vachaud G.:** Scaling of the Richards equation under invariant flux boundary conditions. Water Resour. Res., 27, 2181-2185, 1991.
5. **Nielsen D.R., Kutílek M., Parlange M.B.:** Surface soil water content regimes: Opportunities in soil science. J. Hydrol., 184, 35-55, 1996.
6. **Novák V.:** Evapotranspiration and Methods of Its Estimation (in Slovak). Veda, Bratislava, Slovak Republic, 1995.
7. **Novák, V., Hurtalová T., Matějka F.:** Sensitivity analysis of the Penman type equation for calculation of potential evapotranspiration. J. Hydrol. Hydromech., 1997 (in press).
8. **Van Genuchten M.T.:** A closed form equation for predicting the hydraulic conductivity of unsaturated soils. Soil Sci. Soc. Am. J., 44, 892-898, 1980.