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SOLVING THE WATER TRANSPORT PROCESSES OCCURRING IN SOIL-ATMOSPHERE AND SOIL-PLANT SYSTEMS

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Abstract. This work deals with solutions to the problems of water flow in the soil profile. It takes into account processes occurring at the soil surface as well as water uptake by plants. Soil surface processes are delt with using a modification of a model proposed by Feddes et al. [4]. Practical application of the proposed model is illustrated by an example of a layered soil with variable pipe-drain spacing. Calculations were performed for the growing season of grasses, assuming two different root systems.

INTRODUCTION

The processes connected with water flow on the soil-atmosphere and soil-plant boundaries are among the most difficult to solved. This is because the external limitations (i.e. boundary conditions) cannot be precisely determined, in most cases, except in potential form which means the maximum possible flow. The actual response of the system to a given set of conditions depends on its ability to transport water upwards (in case of evaporation) or downwards (in case of infiltration). Also, in case of water flow from the soil into the plant, the filtrability of a soil plays an important role in the actual flow, besides the availability of soil water to plants. With this in mind, we can use the following relations:

$$q^{pot} \geq q^{act} \tag{1.a}$$

$$ET^{pot} \geq ET^{act} \tag{1.b}$$

where q^{pot} - potential infiltration or evaporation flow, q^{act} - actual infiltration or evaporation flow, ET^{pot} - potential transpiration, ET^{act} - actual transpiration. These conditions (1.a) must be supplemented with an additional limitation, which is

$$h \geq h_{lim} \tag{2}$$

where h is the actual soil moisture pressure head at the soil boundary with the atmosphere, and h_{lim} is the boundary value of the pressure calculated from Kelvin' s formulae (see also Zaradny, [18]):

$$h_{lim} = \frac{R \cdot T_a}{M} \cdot \ln \frac{e_o}{e} \tag{3}$$

where R - universal gas constant (=82 10^3 hPa cm^3 mol^{-1} $^oK^{-1}$), T_a - air temperature ($^oK^{-1}$) M -molar water volume (=18 cm^3 mol^{-1}), e_o- actual vapour pressure of the air (hPa), e - saturated vapour pressure of the air at temperature T_a.

If we would consider a problem where there is no possibility for water to pond on the soil surface, which means the unlimited surface runoff; then condition (2) will be

$$0 \geq h \geq h_{lim.} \tag{4}$$

When considering water flow from the soil to the plant, the sink term S is often introduced into the Richard's equation. There are a variety of proposals to define this term. Most of them are based on analogies to Ohm's law. The rate of water uptake by plant roots is then assumed to be directly proportional to the difference between the soil water pressure and the root suction pressure, to the actual hydraulic conductivity of the soil and to some empirical function such as 'root effectiveness', 'root density', etc. However, these proposals (described in detail by Molz [7], and Zaradny *et al.* [20]) have significant shortcomings; they need a lot of input data which are difficult to measure [4]. Furthemore, data such as 'effectiveness and density of roots' for the same plant species often depend on the soil profile, the plants' growth stage, external conditions, etc.

The results of testing Gardner's model [5] and Feddes' model [4] have been presented in a paper by Zaradny *et al.* [20]. Especially interesting results were obtained for Feddes' model with respect to Bielnik within the Żuławy Depression in the Delta area of the Vistula River [15]. Since this model has practical importance, it is discussed in detail later in this paper. This paper also presents basic equations for the description of water flow in soil, taking into account uptake by the plant and boundary conditions. The equations were used to compute moisture dynamics in a layered soil profile containing a variable drain-pipe spacing.

THEORY

Equations of water flow in the soil with plant water uptake

For soils with growing plants, water flow is described by Richard's equation, supplemented by the sink term, which reflects water uptake by plant roots* (see e.g. Zaradny, [18]):

$$\frac{\partial}{\partial z}\left\{k(h)\frac{\partial h}{\partial z}\right\}+\frac{\partial k(h)}{\partial z}+S=\left\{C(h)+S_s\beta\right\}\frac{\partial h}{\partial t} \qquad (5)$$

where $k(h)$ - hydraulic conductivity, h - soil water pressure head, $C(h)=\frac{\partial\Theta}{\partial h}$ - soil water capacity (for soils without hysteresis $C(h)=\frac{d\Theta}{dh}$), Θ - volumetric water content in the soil, S_s - elastic capacity of the system soil-water, β-coefficient $\beta=\Theta/\Theta_s$, where Θ_s- water content at saturation, t - time, z - vertical coordinate.

The general solution of Eq. (5) is parabolic. For the particular case, $\Theta=\Theta_s$ and S_s=O, the solution becomes elliptic. Eq. 5 is nonlinear because the soil parameters k, C, and S depend on the function $h(z,t)$.

The sink term S represents the volume of water taken up by the roots per unit bulk volume of the soil per unit of time; ($cm^3cm^{-3}s^{-1}=s^{-1}$). Models from the literature, e.g., Zaradny *et al.* [20], can be used to calculate this term. The model proposed by Feddes *et al.* [4] was used here.

The model of water uptake by plants from the soil according to Feddes *et al.* [4]

This model was developed in 1977 and published in 1978. The main aim, as emphasized by the authors, was to overcome the short comings and difficulties connected with the application of other more 'exact' models.

According to Feddes, the sink term S in Eq. (5) depends on boundary conditions (potential transpiration ET^{pot}), rooting depth of the plant (L_k) and the distribution of soil water pressure head (h) in the root zone. This is the simplest version of the model. A modified version is also available. Here root distribution density as a function of depth,

* For simplification the formula has been written for only one-dimensional systems.

RDF (z), and the effect of daily transpiration rate on the availability of soil water are also considered. The function is shown in Fig. 1.

To find a particular solution, the values h_1, h_2, h_3, and h_4 depicited in Fig. 1 must be estimated. Values for h_1 and h_4 are relatively easy to find. Assume $h_1 = h(\Theta_s - \Delta\Theta_1)$, where Θ_s is the water content at saturation and $\Delta\Theta_1$ is a small water content of the order 0.01-0.02 $cm^3\ cm^{-3}$. Higher values of $\Delta\Theta_1$ *are* recommended only for soils showing poor structure, for example sandy soils. The value of h_4 is the wilting point, i.e. the state of soil moisture where pF=4.2 (pF = log h → h = $10^{4.2}$ = 15, 850 cm, hence h_4= -15, 850 cm).

The value of h_2 can be estimated by the soil's relationship between gas diffusion and water content. From this one can determine the ability of the soil to diffuse oxygen to the roots. If the oxygen diffusion coefficient is smaller than 1.5 x 10-4 $cm^2\ s^{-1}$ [1], then plant growth will be hampered. This value of the diffusion coefficient corresponds to the following air-filled pore space in the soil:

$\Delta\Theta_2$ = 0.05 cm^3cm^{-3} for structured soils, and

$\Delta\Theta_2$ = 0.10 cm^3cm^{-3} for structureless (single-grained) soils.

Therefore we propose $h_2 = h(\Theta_2 - \Delta\Theta_2)$, where $\Delta\Theta_2$ will be in the range <0.05; 0.10>, depending on soil structure. The remaining value h_3 and h_3' is assumed to be equal to pF-values of 2.6 and 3.0 (-400≥h ≥-1,000 cm). Studies by Yand and de Jong [14] have pointedoutthath3 depends on the evaporative demand of the atmosphere, that is on values of potential transpiration (ET^{pot}). Generally lower values of h_3 are assumed for lower ET pot values. For interpretation of this phenomenon see Zaradny's paper [18].

The studies carried out during the growing season for grasses at Bielnik in Żuławy [15] had a good fit of calculated results to data at h_3=-400 cm (for $ET^{pot} \geq 5.0$ mm/day) and at h_3'=-1 000 cm (for $ET^{pot} \geq 1.0$ mm/day) and for the linear relationships of h for intermediate values of ET^{pot}:

$$h^* = h_3 + \frac{5.0 - ET^{pot}}{5.0 - 1.0}(h_3 - h_3'). \quad (6)$$

Using this formula as a sink term, water uptake by plants is highest only in the soil profile, within the rooting zone, where soil water pressure head (h) reaches the range $h^* \leq h \leq h_2$. Thus we can write the following equation:

$$ET^{pot} = ET^{act} = \int_0^{L_k} S(h)\, dz. \quad (7)$$

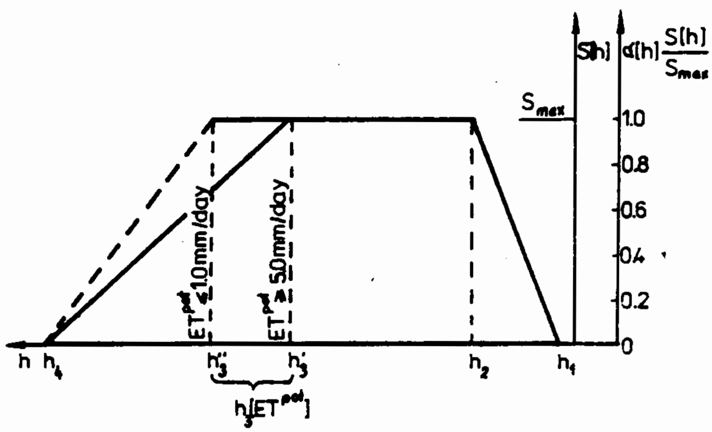

Fig. 1. The sink term S(h) according to Feddes et al. [4].

Introducing a dimensionless variable $\alpha(h)$ such that

$$S(h) = S_{max} \cdot \alpha(h); \; 0 \le \alpha(h) \le 1 \tag{8}$$

Equation (7) becomes

$$ET^{pot} = ET^{act} = S_{max} \int_0^{L_k} \alpha(h)\, dz \tag{9}$$

This equation is true for the time interval for which ET^{pot} values are determined in the calculations, most often 1 day. Accounting for limitations of the dimensionless variable, the maximum value of the integral expression Eq. (9) will be $S_{max} \cdot L_k$ and hence

$$S_{max} = \frac{ET^{pot}}{L_k} \tag{10}$$

where L_k is the rooting depth.

The values of $\alpha(h)$ are then as follows (see Fig. 1):

$$\alpha(h) = \begin{cases} 0 & \text{for } h_1 \ge h \text{ and } h_4 \le h \\ \dfrac{h_1 - h}{h_1 - h_2} & \text{for } h_2 \le h \le h_1 \\ \dfrac{h_4^1 - h^2}{h_4 - h^*} & \text{for } h_4 \le h \le h^* \\ 1 & \text{for } h^* \le h \le h_2 \end{cases}$$

where

$$h^* = \begin{cases} h_3' \text{value for } ET^{pot} \le 1.0 \text{ mm/day} \\ h_3 \text{value for } ET^{pot} \ge 5.0 \text{ mm/day} \\ h_3 + \dfrac{5.0 - ET^{pot}}{5.0 - 1.0}(h_3' - h_3) \text{ for } 1.0 \end{cases}$$

$$< ET^{pot} < 5.0 \text{ mm/day}.$$

The function $S(h)$ may be modified by introduction of a root distribution function RDF(z)

$$\int_0^{L_k} RDF(z)\, dz = L_k. \tag{11}$$

By using the $RDF(z)$ in the earlier Eq. (9) we obtain

$$ET^{pot} \ge ET^{act} = S_{max} \int_0^{L_k} \alpha(h)\, RDF(z)\, dz. \tag{12}$$

The value of the sink term for any coordinate z' in the root zone ($0 \le z' \le L_k$) can be calculated from the relationship:

$$S(h)\Big]_{z'} = S_{max}\alpha(h)\Big]_{z'} RDF(z'). \tag{13}$$

Boundary conditions of soil-atmosphere and soil-plant interfaces

Boundary conditions at the soil-atmosphere interface depend on actual meteorological data and the kind, stage, and condition of the crop. Potential transpiration ET^{pot} can be calculated from the relation

$$ET^{pot} = E^{pot} - ES^{pot} \tag{14}$$

where E^{pot} is potential evapotranspiration from both the crop and the soil. ES^{pot} is potential evaporation from only the soil surface. The E^{pot} values are most often derived from Penman's equation [10]. This uses a combination of energy balance and transport of water vapour. Basically, Penman applied this combination method to a water and a soil surface. In 1965 Monteith [8] Rijtema [12], independent of each other, extended this method to soil with a plant cover. The final formula for E^{pot} is (for details see Zaradny's papers: [16,18]):

$$E^{pot} = \frac{\delta\, Rn + c_p \rho_a (e_o - e_d)/r_a}{(\delta + \gamma) l} \tag{15}$$

where E^{pot}- potential evapotranspiration flux (kg m^{-2} s^{-1}) δ - coefficient numerically corresponding to the derivative of the saturated vapour pressure es with air temperature T_a (hPa $^oK^{-1}$), Rn - net radiation (W m^{-2}), c_p - specific heat of the air at constant pressure (J kg^{-1} $^oK^{-1}$), ρ_a - density of the air (kg m^{-3}), ε_o - water vapour pressure of saturated air at air temperature T_a (hPa), e_d - actual vapour pressure (hPa), γ - psychrometer constant (hPa $^oK^{-1}$), L - latent heat of evaporation (J kg^{-1}), and r_a - aerodynamic resistance (s m^{-1}).

The value r_a can be determined under conditions of natural stability if the plant height and the wind velocity are known (for instance from the formulae and tables given by Feddes *et al.*, [4]). Potential evaporation of a soil under a crop cover can be computed from the formula proposed by Ritchie [13]:

$$ES^{pot} = \frac{\delta}{(\delta + \gamma) L} Rn \exp(-0.39\, LAI) \qquad (16)$$

where LAI is the leaf area index, which depends on soil surface shaded by leaves (Sc ≤ 1).

In most studies performed until now, direct measurements of Rn have not been made. Therefore, this value must often be derived with the empirical formulae

$$Rn = (1 - \beta)\, Rs - Rt \qquad (17)$$

where Rs - short-wave radiation (W m^{-2}), Rt - thermal radiation (W m^{-2}), β-reflection coefficient, 'albedo', which depends on the kind of surface (see also Eagleson [3]; Tables 3 and 4).

The value Rt can be derived from the following formula [2, 10]:

$$Rt = 5.67\text{x}10^{-8}\, T_a^4\, (0.56 - 0.08\sqrt{e_d})$$

$$(0.1 + 0.9\, n/N) \qquad (18)$$

where T_a - air temperature (^{o}K), e_d - actual water vapour pressure in the air (hPa), n - actual duration of sunshine (h), and N - maximum (optimal) possible duration of sunshine.

In practice daily Rs values are not often measured (not only in Poland). Thus, in many cases the Rs radiation is calculated from empirical formulae such as the Kimball expression:

$$Rs = (A + B\, n/N)\, Rst \qquad (19)$$

where A and B are coefficients, determined for a chosen site from a regression analysis, Rst - radiation (insolation) at the top boundary of the terrestrial atmosphere, dependent on the latitude and the time of the year.

This way of determining the Rs values has been described in detail by Zaradny and van der Ploeg [19]. In water melioration practice, this was used for the Northern region of Poland (the Delta area of the Vistula River, called Żuławy). Aside from the models mentioned above [8,12], one can find simpler expressions on the subject in the literature such as that of Priestley and Taylor [11]. The application of this formula as well as those described earlier allowed us to elaborate new calculating programs, such as OBEV and EVAPOT. They were presented by Zaradny [16].

Practical application of the models presented

These models with their assumptions were used for simulation of the effectiveness of a drainage system based on the level of the ground water table, the water supply according to plant needs (grasses), and on the field workability, i.e. the accessibility of the soil to field work and traffic operation. For some select simulations, assume that the drainage system can only drain water from the field. This means that outflows of the drains are not flooded with water. In the simulation consider pipe-drain spacings of L = 5, 10, 15, 20, 25 and 30 m, with the following assumptions: the pipe diameter is equal to 7.5 cm and the depth of the pipe is 100 cm (top of the pipe). Furthermore, assume that the root system is either uniform(RDF (z) = 1 = const., L_k = 60 cm) or nonuniform, as from studies carried out by Olszta and Zawadzki [9], (Fig. 2).

Simulations were carried out for one growing season (April 1 - September 30) on a shallow heavy alluvial soil, laying on silt. The relationships of hydraulic conductivity and pressure head on water content are presented in Fig. 3.

The 'wet year' (app. 666 mm of rainfall during the vegetative period, for instance in July - precipitation (P) =164 mm, August - P=153 mm, and in September - P=118 mm) appeared to be critical for the locality under consideration. This confirms results of earlier studies [15]. Moreover, three mowings of grass were simulated for the growing season (in two-month intervals). That directly affected the degree of soil surface cover (Sc), and indirectly affected ES^{pot} and ET^{pot} in the total evapotranspiration flux E^{pot} [15].

Simulation studies were performed with the HZARG program [17], which includes the method of infinite elements for two-dimensional systems x-z. The results obtained for the nonuniform rooting case are presented in Figs 4 and 5. Fig. 4 illustrates the position of the ground water table, h_{zw}, and the values of ratio ET act/ET^{pot}. Fig. 5 presents the ranges of soil water pressure heads (h) at the surface.

The results for uniform root distribution are depicited in Figs 6 and 7.

The uniform, deeper rooting system had a slightly higher drop in ground water table during the initial part of the growing season (April-June); 234 mm in comparison to 226 mm for nonuniform, shallower rooting. However, despite a considerable drop of the water table during this period, the availability of water to plants was still good. In the later period, as the ground water level rose, water availability was similar for both root systems. Close to the end of the season (August-September), a slightly lower availability of water to plants was noticed for the deeper, uniform rooting system, expressed by lower values of ET^{act}/ET^{pot} ratio.

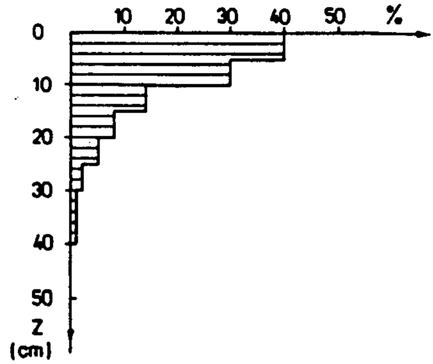

Fig. 2. Root distribution of grasses acc. to Olszta and Zawadzki [9].

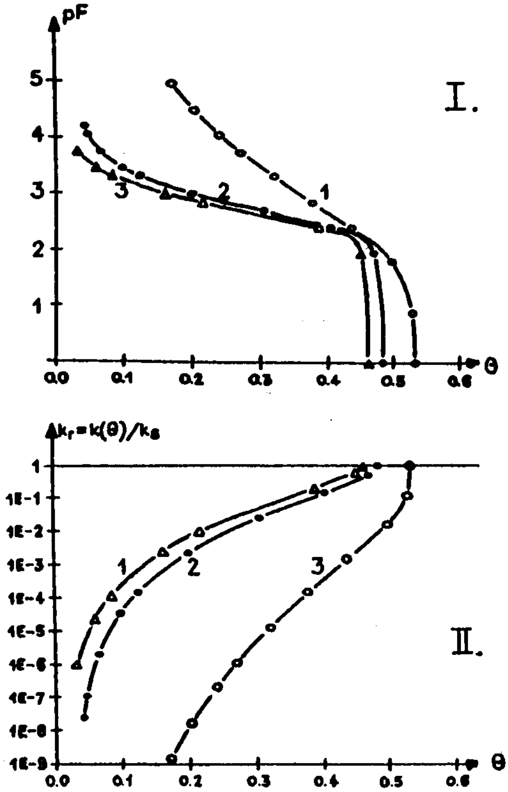

Fig. 3. The pF curves (I) and relative hydraulic conductivity (II) of the soil profile used for the simulation studies.

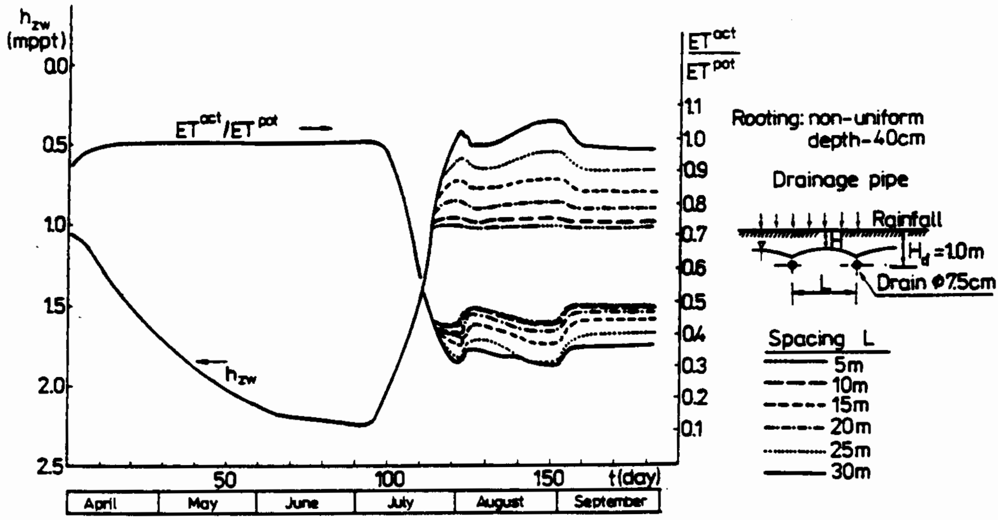

Fig. 4. The calculated position of the ground water level h_{zw} and the values of ET^{act} / ET^{pot} for nonuniform rooting.

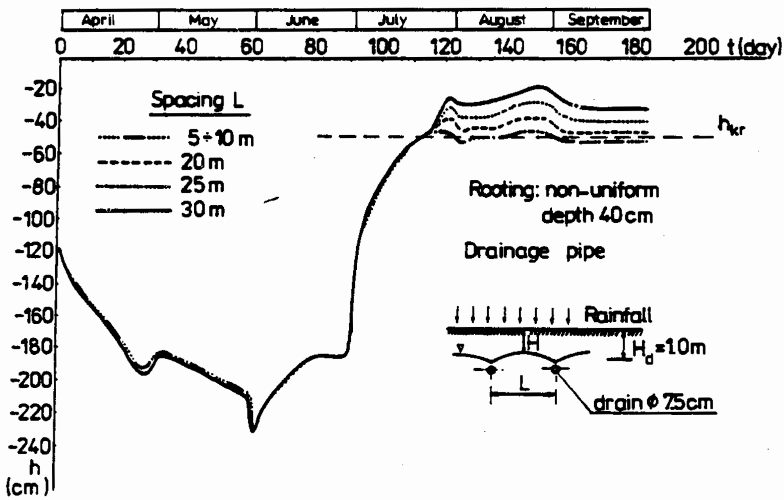

Fig. 5. The range of soil water pressure head at the soil surface (nonuniform rooting).

The ratio of ET act/ET pot reached the minimum value, i.e., 0.21, for spacing L=30 m. The availability of water to plants at the nonuniform and shallower rooting depth at a narrower pipe-drain spacing (L<30m) was slightly lower. The conditions of accessibility for field operations would be worse for the uniform' deeper rooting system. This was proven by values of soil water pressure head depicited in Figs 5 and 7.

From this data it can be seen that soil moisture conditions in the soil profile will

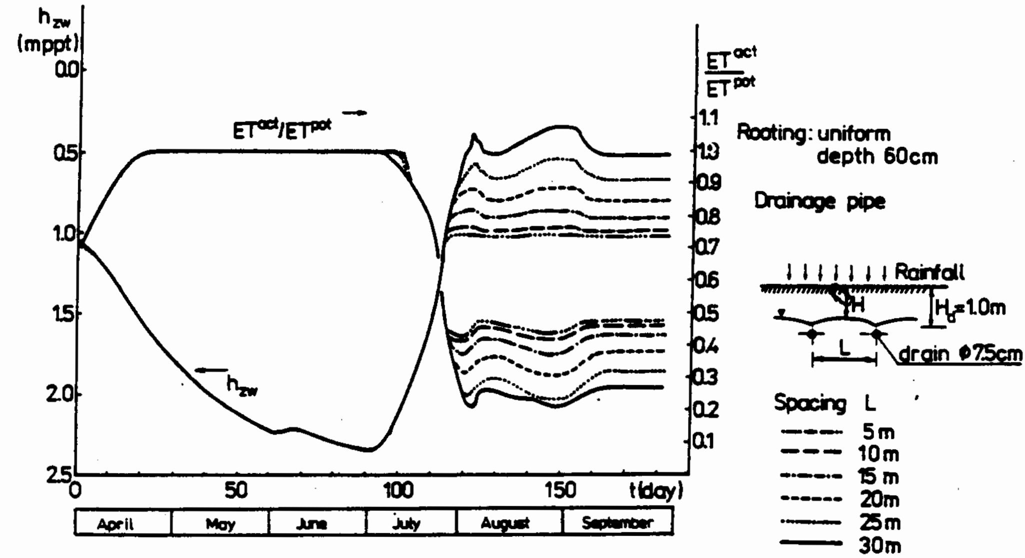

Fig. 6. The calculated position of water level (h_{zw}), and the values of ET^{act}/ET^{pot} for uniform root.

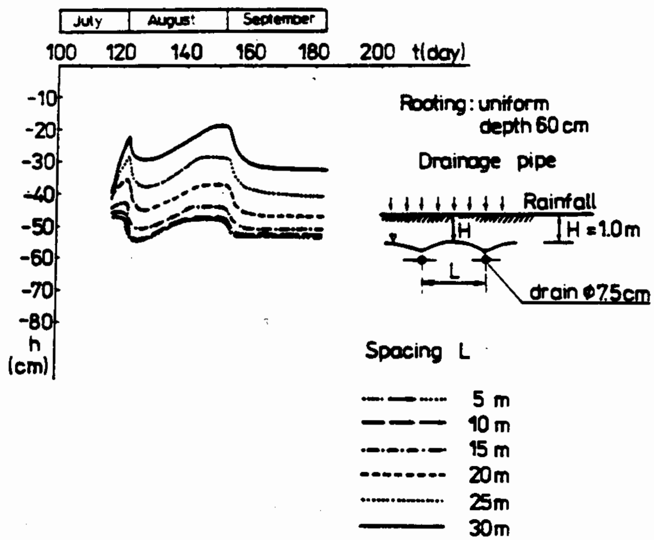

Fig. 7. The range of soil water pressure heads at the soil surface of the area studied for uniform rooting.

not be favourable for both the rooting systems considered during the second half of the vegetation period (starting from the end of July). Improvement can be obtained with deeper drainage and by growing plants characterized by shallower rooting systems.

CONCLUSIONS

The results obtained under project CPBP 05.03.01.2 allow us to solve complex problems for water flow in the soil-atmosphere and soil-plant systems. Theoretical descriptions containing computer programs are aimed at solving this problem. The applicability of these methods have been proven by the results presented in annual reports. Some of them are presented in this paper. Based on these results, we may conclude that these methods can be used to help explain the agrophysical basis for soil and plant productivity.

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ROZWIĄZYWANIE ZAGADNIEŃ TRANSPORTU WODY ZACHODZĄCEGO NA KONTAKCIE GLEBA-ATMOSFERA ORAZ GLEBA-ROŚLINA

W pracy przedstawiono rozważania związane z rozwiązywaniem zagadnień przepływu wody w profilu glebowym z uwzględnieniem procesów zachodzących na kontakcie z atmosferą oraz poboru wody przez rośliny. Te ostatnie proponuje się rozwiązywać wychodząc z modelu Feddesa i inn., 1978, z jego późniejszymi modyfikacjami. Praktyczne wykorzystanie proponowanych modeli zilustrowano na przykładzie uwarstwionego profilu glebowego, będącego pod działaniem sieci drenarskiej o zmiennej rozstawie. Obliczenia przeprowadzono dla sezonu wegetacyjnego, przy założeniu dwóch różnych układów korzeniowych.