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THEORETICAL ANALYSIS OF THE IRRIGATION OF SOILS WITH VARIOUS STRUCTURES

Summary

This paper reports the results of research into water flow in three various soil types. Soil data were obtained on the basis of a classification by Wösten and van Genuchten [38]. The mathematical model of water flow in soil was formulated using Richards' equation. The data for hydraulic conductivity and suction pressure of unsaturated soil were obtained using the van Genuchten model. The method of finite difference and an original calculation program were applied to solve equations. The time of water transfer through soil and depth of groundwater enabling water transport to root-zone were studied. These phenomena were examined for different boundary conditions. The unsteady character of irrigation process was analyzed. Water flow in soil following short-term and intense precipitation was one of analyzed problems.
Key words: *irrigation, numerical calculations, Richards' equation, water in soil*

1. Introduction

The presence of water in soil and its flow is essential for life plants and their growth.

Water infiltration in soil is directly associated with the transport of ingredients indispensable for the plant growth, e.g. mineral and organic fertilizers. This water flow in soil also leads to the spread of pollutants and their infiltration into the subterranean aquifers.

The mathematical model of water transfer through unsaturated soil is based on Richards' equation [15, 29]. This equation expresses the relationship between soil water pressure head and volumetric water content and requires an adequate selection of the boundary and initial conditions.

Richards' equation is a strongly non-linear, comprising source terms, boundary conditions for modeling of the irrigation systems [1, 4, 25], or the water intake by plants [3, 5, 7, 16, 17]. Strong non-linearity of Richards' equation results from the changes of orders in the soil water pressure head and the soil hydraulic conductivity along with the changes in water content in soil. In turn, these parameters vary considerably according to the type of soil as well as its components and structure [12, 26, 32, 38].

The relation between soil water pressure and the volumetric water content – i.e. so called, retention curve - and also the relation between its hydraulic conductivity and the volumetric water content, is most often expressed by means of the van Genuchten's equations [34]. These equations account for a few physical qualities and coefficients depending on the soil type and its composition. They find application in many studies with regard to the modeling of water flow in soil [14, 18, 20, 28 31, 35]. However, other models are also applied to relate water pressure with soil water content, and also other relations are used to express hydraulic conductivity as a function of water content in soil. In this context we can mention relations involving power such as ones developed by Brooks and Corey [2] or Gardner [10]. Another solution could be involution model of retention curve, using fractal dimension of soil structure as the exponent [8].

Any statement regarding boundary conditions in the soil layer is associated with problems as it needs to account for various manners of water exchange between soil and atmosphere over time. While modeling water flow in soil one must note that plants draw different amount of water in different vegetation stages. The intensity of water intake by plants is also dependent on the weather conditions.

One characteristic of soil is associated with its diversification along with the profile depth. This fact is most often taken into consideration when soil is divided into layers, depending on its grain size components and its physical properties. Examples of numerical solutions of Richards' equation for these soil profiles can be found, among others in [10, 14, 24, 29].

A vast majority of studies tackle the one-dimensional problem considering the variation of the physical parameters of soil in the function of soil depth and time. Studies describing the complex irrigation systems and plant root zones mostly consider the two-dimensional problem [1, 5, 25]. The issues of soil irrigation on a sloping area [6, 23] or in levee (river embankment) [11] have also been tackled to represent two-dimensional ones.

A number of studies attempted to systematize equations describing the retention curve and the soil hydraulic conductivity as well as the coefficients and exponents in these equations for varied soil types. However, the relations themselves and coefficients developed for these purposes do not give the immediate practically useful information. Numerical solutions of Richards' equation most often define the front position of soil saturation in the function of time [3, 11, 14, 23] or changes of soil water pressure head [25, 35]. Taking a practical view, valuable studies could potentially involve search for vital information concerning the time after which dry soil adsorbs water up to a certain depth as a result of irrigation and the indispensable minimal flux of water to maintain it.

This paper contains the numerically derived indicators specific for water flow in soil such as: time of water relocation from soil surface into the depth of 100 cm, depth of the underground water generating a given flux of water into plant root zone, and flux of water inside the soil corresponding to the minimal, indispensable soil aeration. Soil parameters as well as their values are measured and calculated on the basis of determined curves of the hydraulic conductivity obtained from studies by Wösten and van Genuchten [38].

For the soil types examined here, water flow and moisture changes in the function of time are determined here. The time of moisture reaching different field depth levels and the amount of water supply is described as well.

2. Material and methods

2.1. Mathematical model of water flow in soil

The mathematical model of water flow in soil was based on the Richards' equation [15, 29]. This equation, after transformations to account for the negative value of water pressure head h_s and axis z directed towards the soil surface, takes the form (equation 1):

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z}\left(K(\theta) \cdot \frac{\partial h_s(\theta)}{\partial z} \right) - \frac{\partial K(\theta)}{\partial z} - S(\theta) \tag{1}$$

In the above equation, θ stands for the soil moisture defined as a volumetric part of soil volume filled with water, $K(\theta)$ stands for hydraulic conductivity of unsaturated soil. The h_s value is the soil water pressure head, $S(\theta)$ is the source term. The relation between the soil water content and its soil matric potential is experimentally specified and presented in a form of the water retention curve (pF).

The $h_s(\theta)$ and $K(\theta)$ functions were adapted in this work, according to equations derived by van Genuchten [34]. The following dependences result from the mathematical description of interaction between water and solid in porous material, and, also from the adoption of functional dependence between the soil water pressure head and volumetric water content recognized in the form of dimensionless parameters (equations 2 and 3):

$$h_s(\theta) = -\frac{1}{\gamma} \cdot \left(\left(\frac{1}{s} \right)^{\frac{1}{m}} - 1 \right)^{\frac{1}{n_g}} \tag{2}$$

$$K(\theta) = K_s \cdot s^{\eta} \cdot \left(1 - \left(1 - \frac{1}{s^m} \right)^m \right)^2 \tag{3}$$

in which γ, n_g , $m=1-n_g^{-1}$, η are empirical parameters. Value S is defined as follows (equation 4):

$$S = \frac{\theta - \theta_r}{\theta_s - \theta_r} \cdot \tag{4}$$

In the above formula θ_s and θ_γ denote the maximum soil moisture (soil saturated with water) and its minimum moisture, respectively.

In an interpretation of the parameters found in (2) and (3), we can assume that K_s may be considered as the soil percolating flux, with its maximum moisture saturation. Magnitudes γ, n_g and η form the empirical parameters. It is essential to point out that the values of these parameters, depending on the type of soil, vary considerably. Information concerning the value of these parameters for various soil types can be found in various sources [12, 14, 38, 36].

2.2. Indicators of soil retention

Soil conductivity in saturation state, and also other mentioned above parameters are only qualitative parameters relating to water flow in soil. Therefore, the value criteria which would practically facilitate evaluation of water retention in different soil formations were introduced.

To evaluate the degree of water retention Wösten et. al. [37] introduced the following functional criteria:

a) The time of water relocation (transfer time T_γ) from its surface to the depth L = 100 cm (equation 5):

$$T_r = \frac{\bar{\theta} \cdot L}{\bar{q}_d} \tag{5}$$

where $\bar{q}_d$ = 0,14 cm/d is an average flux for Danish winter conditions. An average soil moisture $\bar{\theta}$ corresponds to flux $\bar{q}_d$ for the experimentally determined soil characteristics $K = f(h_s)$, $h_s = f(\theta)$.

a) Depth of water table L_c, enabling the transport of water to root zones. This depth is specified by the formula (equation 6):

$$L_c = \int_0^{h_c} \frac{1}{1 + \dfrac{\bar{q}_u}{K(h_s)}} \cdot dh_s \tag{6}$$

b) On the basis of Darcy's Law, soil matric pressure h_c was assumed to be equal to -500 cm, and average steady-state upward water flux $\bar{q}_u$ = 0,2 cm·d^{-1}.

c) Downward water flux q_d, corresponding to the water contents in the soil 0,05 cm^{3}cm^{-3} less than value of soil fully saturated with water. This is minimum allowed value corresponding to the conditions of soil aeration. Such defined flux q_d is not much smaller than K_s – fully saturated soil conductivity.

The above indicators characterize the hydrological potential of soil conditions. During irrigation process it is relevant to estimate the variation in initially dry soil in the function of time and its depth. The mathematical model developed in this paper was applied for the analysis of irrigation of soils with various structures. An attempt was made to numerically determine the indicators defined before.

2.3. Numerical solutions of mathematical model

Numerical methods are the successful tool in solving several issues associated with water transfer through soil. Both the finite difference method [1, 23, 31, 33] and the finite elements method have been applied [11, 20, 25]. Some of shared programs e.g. HUDRUS-1D [31], HYDRUS-2D [13] and MODHMS [22] offer research tool in moisture soil examinations.

Equations of the mathematical model were solved using finite difference method [9, 21] with the help based on the original program developed in the algorithmic lg-FORTRAN. The finite difference method is most effectively applied to solve numerous mathematical-physics problems, including fluid movements in porous media.

First-order explicit method was applied to obtain the solution of the equation (1) of the mathematical model in time. Non-linear diffusion member in this equation was approximated applying the central differences. Finally, the differential equation (1) corresponds to the equation (7) below:

$$\theta_z^{n+1} = \theta_z^n + \Delta t\left(\left(\frac{(K_{z+1}^n + K_z^n)\cdot(h_{s,z+1}^n - h_{s,z}^n}{2\Delta z} - \frac{K_z^n + K_{z-1}^n)\cdot(h_{s,z}^n - h_{s,z-1}^n}{2\Delta z}\right)/\Delta z - \frac{(K_{z+1}^n - K_{z-1}^n)}{2\Delta z} + S_z^n\right) \tag{7}$$

In the above mentioned equation *Δt* and *Δz* represent the time and spatial steps in a differential matrix, and the indices n and z respectively denote: time level *n+1*"new", *n* "old" and node on the axis z, respectively.
The water flux was calculated from the equation (8):

$$q_z^n = \frac{(K_{z-1}^n + K_z^n)\cdot(h_{s,z}^n - h_{s,z-1}^n)}{2\Delta z} - \frac{(K_{z-1}^n + K_z^n)}{2}. \tag{8}$$

Both initial and boundary conditions depend on the problem being considered. During surface irrigation of the soil, moisture is set to *θ(1,t)* **-** Dirichlet condition, and at the depth of 125 cm: $\frac{\partial\theta}{\partial z} = 0$**-** Neumann condition.

Simulating the instances of precipitation, the soil moisture level at the surface is designated as the boundary condition on the basis of the following formula (9):

$$\theta_1^n = \theta_1^{n-1} + \frac{(q_r(t) + q_2^n}{\Delta z}\cdot\Delta t \tag{9}$$

Where $q_r(t)$ is the level of precipitation in time and q_2^n is the water flux (negative) leaving the surface level. When $\theta_1^n > \theta_s$ we can assume that $\theta_1^n = \theta_s$, and excess of water in calculations is introduced into soil, after irrigation we can assume that $\theta_1^n = \theta_s$. This condition persists until the water is depleted from the soil surface.

3. Results and Discussion

On the basis of a mathematical model and calculation program, an attempt was made to determine coefficients of water retention. Calculations were conducted for the same soil formations, discussed earlier, the parameters of which are given in Table 1. These are the values adopted from Wösten, van Genuchten paper [38]. Columns were supplemented in this table with the numerical results. The differences, in comparison to the characteristics values indicated by Wösten and van Genuchten, result, among others, from the manner of water flux calculation in soil. In the numerical calculation program, the flux is evaluated from Darcy's law, written in the final differences method and involving gravity force interaction. Although, the calculated flux value is close to the observed condition, hydraulic conductivity *K(θ)* (according to (3)) varies to some extent on the value *K(θ)* reported by Wösten and van Genuchten. Nevertheless, this is the way to obtain the values approximated to the ones given by the authors.

The data found in Table 1 below indicate that in case of the third parameter, its values for individual soil formations differ by three rows. In the case of the first indicator T_r, there is the qualitative consistency of calculation results with the measured values especially when comparing the results of numerical calculations with measurements.

As far as layer thickness L_c is concerned, the differences of measured and calculated values are significant, especially for the case of the results of numerical calculations for the coarse-textured soil. The agreement of the numerical calculations with measurement results is better for the calculation of q_d water flux, which refers to coarse-textured and dusty soils. For the medium-textured soil the calculations from paper [38] as well as from the numerical calculations differ by more than twice compared to the measured data. The differences found in the study may occur due to other reasons, such as from uncertainty of curve tracing *h = f(θ)* and *K(θ)*, especially at the place of soil saturation.

Table 1. Properties of tested soils, measured and calculated values of parameters T_r, L_c, q_d [38]
Tabela 1. Cechy charakterystyczne badanych gleb, zmierzone i obliczone wartości parametrów T_r, L_c, q_d [38]

Soil	Coarse-textured soil			Medium–textured soil			Fine-textured soil		
Parameters	n = 2,43 K_s = 90 cm·d^{-1} α = 0,0551cm^{-1} η = 0,31 γ = 0,0551cm^{-1} θ_r = 0,05 θ_s = 0,40			n = 1,25 K_s = 17 cm·d^{-1} α = 0,0624 cm^{-1} η = -3,30 γ = 0,0624 cm^{-1} θ_r = 0,07 θ_s = 0,45			n = 1,08 K_s = 8 cm·d^{-1} α = 0,0740 cm^{-1} η = -9,38 γ = 0,074 cm^{-1} θ_r = 0,30 θ_s = 0,60		
measurements /calculations	measured	calculated	numerical calculations	measured	calculated	numerical calculations	measured	calculated	numerical calculations
T_r [d]	143	121	94.8	300	257	253	407	421	417,9
L_c [cm]	145	120	54	138	65	60	22	12	9,75
q_d [cm·d^{-1}]	30	23	28,7	0,2	0,5	0,439	0,002	0,010	0,0019

The calculated values T_r, L_c and q_d are directly calculated on the basis of the retention curve as well as from the dependence, $K = f(\theta)$. Uncertainty in the estimation of these functions affects the uncertainty in the T_r, L_c and q_d. This uncertainty is considerable, especially for q_d (downward flux). This results from high variability of h and K in the θ function, and from the use of the logarithmic scale. Close to saturation point θ_s**,** small changes are associated with changes in soil water pressure by an order or two. This impinges on the uncertainty of K designation as volumetric water content function.

The designation uncertainty of the average value of T_r, L_c and q_d for the collection of soil samples of different types depends on the above standard uncertainty, and on the number of tested samples. For example, as reported in [37], in case of sandy soil the flux (q_d) changes in the range 12,2-32,3 $cm \cdot d^{-1}$ and the half width of the 90% prediction interval is 118,5 $cm \cdot d^{-1}$. In case of clay loam soil and silty clay soil the range of changes q_d results from 0,04 to 0,09 $cm \cdot d^{-1}$ and the half width of the 90% prediction interval is 0,24 $cm \cdot d^{-1}$. The uncertainty of measurement exceeds the value of the measured parameter.

The mathematical model and the calculation program were used to investigate water flow in soil. The simulation of surface irrigation was performed for each soil with parameters defined in Table 1. An assumption was made that initially the surface of soil is fully saturated, and in addition, the whole volume of soil is dry, that is, its relative moisture level corresponds to pF = 4,2.

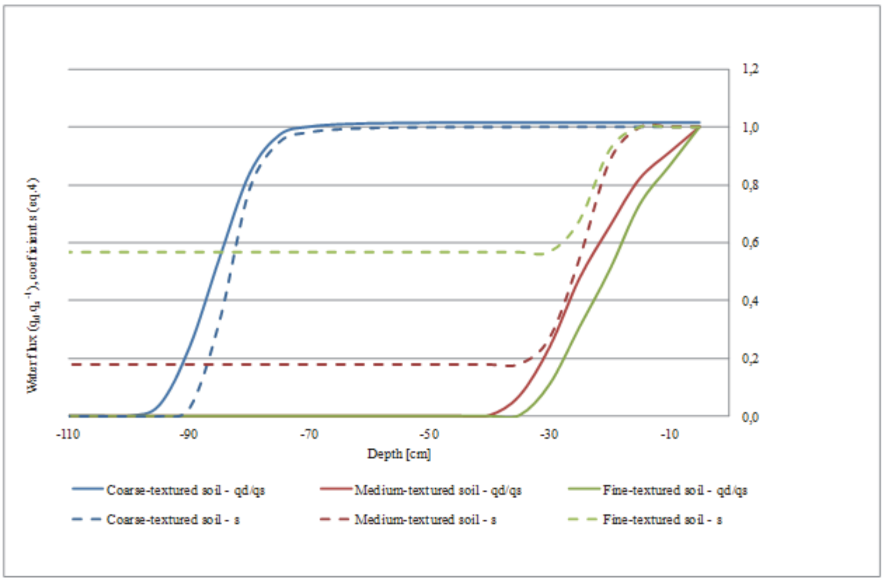
Figure 1 shows water fluxes and volumetric water content after 0,25 d for the analyzed soil textures. The front line of soil saturation by water is relatively flat and at relatively short section θ (coefficients) changes the value characteristic for dry soil into the value which equals to its moisture saturation. The soil water pressure head also changes in the same way. This characteristic change of volumetric water content with its intensive water saturation are also shown by the calculations, presented in other papers [11, 18, 19, 23, 36].

Soil saturation changes from the value practically θ_r at the lower parts of soil, to the value equal θ_s. The value change takes place at a layer depth of 30 cm. In case of coarse-textured soil moisture saturation of the crop root-zone, which is up to the depth of 40-50 cm, practically takes place after two hours. The calculations also show that after not even 24 hours the deep soil moisture saturation occurs. In case of medium-textured soil and coarse-textured soil, the course of water saturation of the soil is similar, the differences occur when the water infiltrates till the respective depths. The course of the curves shows that small changes of water content result big changes of flux.

The calculations show that the water infiltration until the depth of 40-50 cm is reached by water after about 18 hours, thus making totally it saturated with moisture. Soil is saturated to depth of 1m after about two days. Moisture saturation for the fine-textured soils occurs in a similar manner. The q and θ courses run analogously as for the case of the medium-textured soil, hence, its saturation takes more time. The depth of 40-50 cm is reached after about one day and soil saturation at 1m depth after about three days.

Fig. 2 shows the time after which the field moisture at the depth of 50 cm is reached in the tested soils (pF ≈ 2,5). Initially we assumed that the soil is dry (pF = 4,2) and the steady amount of moisture on the soil surface was established. The level of this moisture was expressed by the value of parameter s (4). At the depth of 125 cm the Neumann boundary condition was set: $\frac{\partial \theta}{\partial z} = 0$.

There are irrigation systems in which the surface moisture is maintained at a lower level than its saturation. This moisture may be due to types of irrigation system, for example: trickle irrigation [1, 25]. These limit water loss, both as a result of evaporation, and soaking into deeper soil layers. Figure 2 presents the calculated results indicating that the time may vary fundamentally, expressed in days. It depends essentially on soil type being examined. The value of surface moisture also significantly influences the time of soil wetting. A decrease in moisture from s = 1 to s = 0,9 change this time to hours for the coarse-textured soil, days for medium-textured soil, and practically prevents irrigation in case of fine-textured soil.



Source: own study / *Źródło*: opracowanie własne

Fig. 1. Water flux $q_d \cdot q_s^{-1}$ and coefficient s (4) after 0,25 d irrigation

Rys. 1. Strumień przepływu $q_d \cdot q_s^{-1}$ i współczynnik s po czasie irygacji równym 0,25 doby

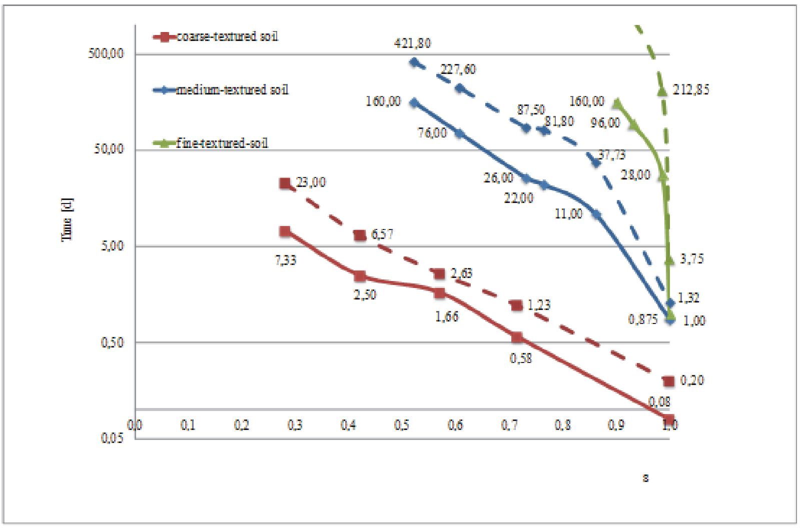


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Fig. 2. Irrigation time needed to reach field moisture, at the depth of 50 cm

Rys. 2. Czas nawadniania potrzebny do osiągnięcia wilgotności polowej na głębokości 50 cm

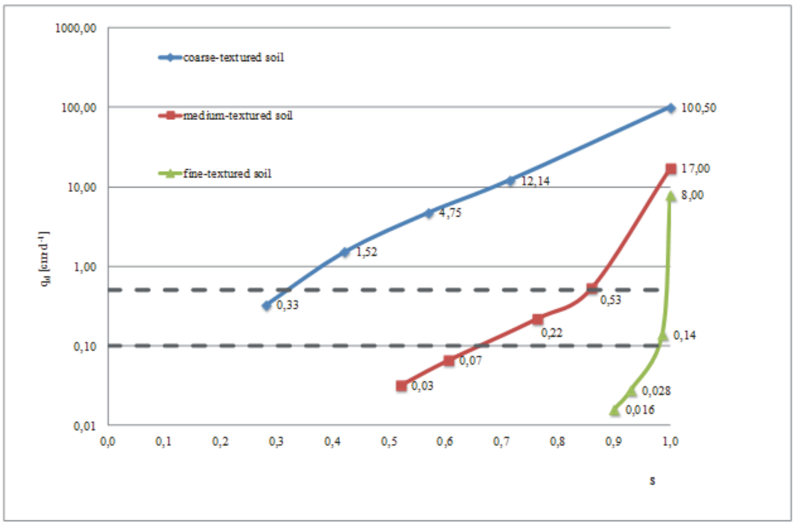

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Fig. 3. Water flux q_d through the surface, at the moment of field moisture at the depth of 50 cm

Rys. 3. Strumień przepływu q_d przez powierzchnię w momencie osiągnięcia wilgotności polowej na głębokości 50 cm

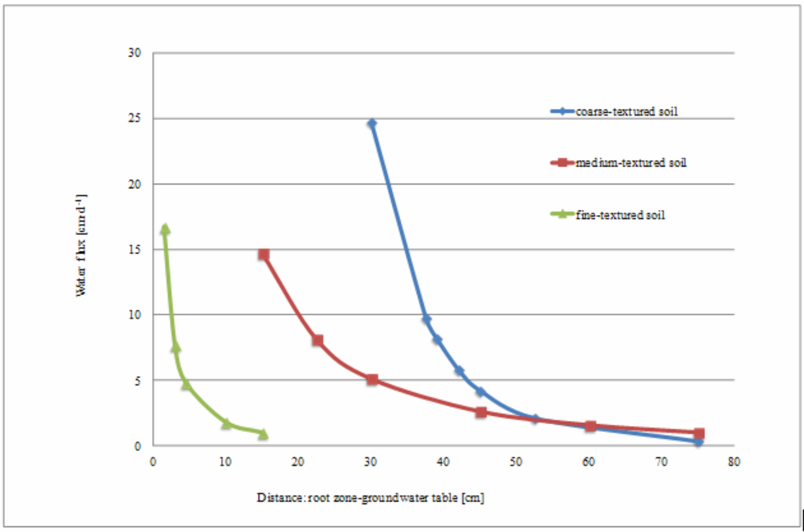

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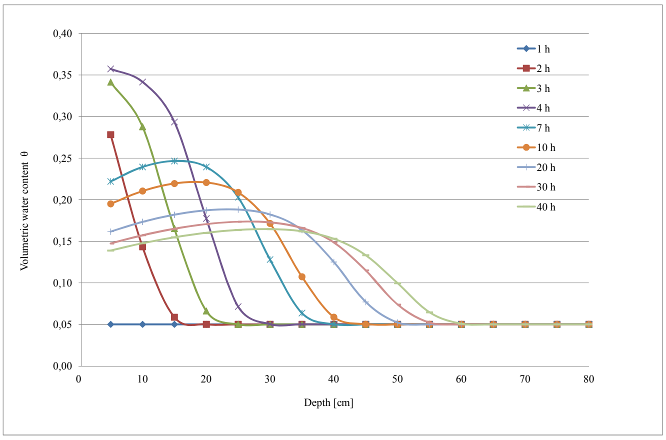
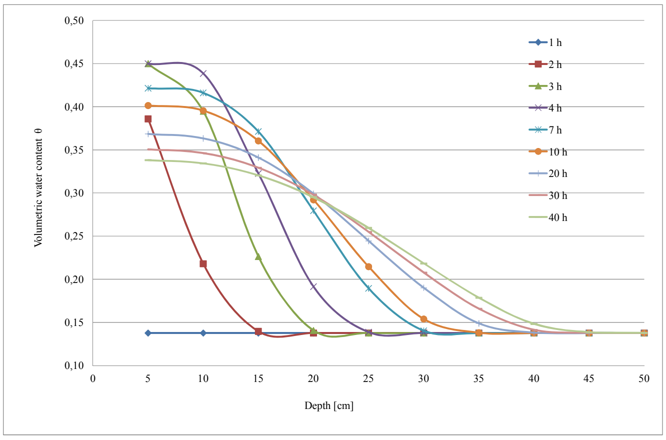
Fig. 4. Soak flux of field moisture in the root-zone, in function of distance from root-zone to groundwater table level

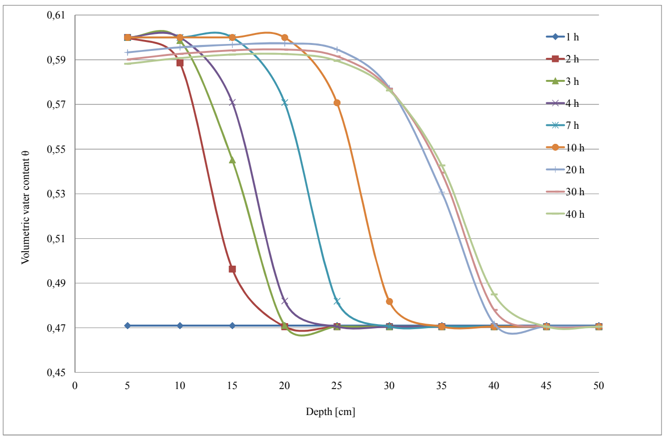
Rys. 4. Podsiąk w strefie korzeni, w funkcji odległości od poziomu strefy korzeni do lustra wody gruntowej a)



b)



c)



Source: own study / *Źródło*: opracowanie własne

Fig. 5. Changes in volumetric water content for: coarse- (a), medium- (b) and fine-textured soil (c) in time

Rys. 5. Zmiany w objętościowej zawartości wilgoci w czasie dla gleby: grubo- (a), średnio- (b) i drobnoziarnistej (c)

Fig. 2 also presents the time of water movement in the steady state (marked with dotted line), as well. Then, in the whole area, the soil moisture remains the same as set on the surface. The time of moisture relocation T_r (for $L = 50$ cm) was derived from formula (5). We can see that the time is apparently longer. This results from a considerable suction pressure gradient present in the soil-root zone. It occurs while introducing water into this zone, at the initial point when the soil is dry.

Fig. 3 shows values of water flux flowing through the surface of tested soils in the moment of reaching field moisture at the depth of 50 cm. For moisture saturated soil, this flux equals to the soil hydraulic conductivity in the state of saturation. It diminishes substantially along with the decrease in moisture in the root-zone. Fig. 3 presents also the range of flux values ($0{,}1$ - $0{,}5$ $cm \cdot d^{-1}$), in which, on average, there is a variation in the demand in plants for water at the different stages of their vegetation. For the case of fine-textured soil being tested, the flux practically corresponds to the full soil moisture saturation. For the medium-textured soil, the section occurs with $s = 0{,}8$. The need to reach this level for coarse-textured soil requires permanent irrigation with the soil moisture $s < 0{,}3$, or periodic irrigation considering the soil retention capacity.

For the case of ground water presence and drainage system, it is possible to change the irrigated plants through the regulation of capillary soaking magnitude. It requires a change in the groundwater table level [27]. Figure 4 presents the numerical calculation of capillary soaking flux considering different distance from groundwater table level to the root-zone. The soil moisture corresponding to field moisture was achieved into this zone. We can see that in the case of fine-coarse soil the irrigation requires the ground water table to be at a distance of less than 20 cm from the root zone.

Fig. 5 (a, b, c) depicts moisture changes, for coarse-, medium-, and fine-textured soil respectively, as a function of time resulting from irrigation with intensive rain. Soil was assumed to be dry (pF = 4,2) in these calculations. Irrigation with 5 cm of water in time of three hours was simulated. From one to fourth hour a layer of water is formed when the physical conditions do not allow temporary water penetration of the total rain flux through the soil surface. Water is absorbed gradually by the surface layer while water flows deeper into the soil. This situation is characteristic for fine-textured soil.

For the case of coarse-textured soil, moisture saturation of the soil surface did not reach this level resulting from this kind of irrigation. The maximum water flux in soil corresponds approximately to the irrigation flux and occurs in the third hour of irrigation. The maximum achievable soil moisture equals to about 0,36. Field moisture is reached at the depth of 50 cm after 23 hours. After 12½ days soil water flow is lower than $0{,}1$ $cm \cdot d^{-1}$, and field moisture reaches a level of 90 cm.

For medium-textured soil, the state of maximum moisture saturation at the surface layer was gained as a result of its irrigation. This condition is reached at the end of irrigation and is short-lived directly after it. After 8¾ days soil water flux is lower than 0.1 cm/d, and field moisture reaches the depth of 28 cm.

The above mentioned irrigation of fine-textured soil causes its instant moisture saturation at the surface layer, which still lasts for the next nine hours. After two days the water flow is lower than 0.1 $cm \cdot d^{-1}$, and field moisture reaches a depth of 35 cm.

4. Summary and conclusions

The description of water flow in soil was based on the Richards' equation, to which empirical dependencies were integrated. These include: soil water pressure head and hydraulic conductivity of soil as the function of volumetric water content. A program was developed to perform numerical solutions for the model equations, based on the finite difference method. Numerical studies for water flow in three different soil formations were conducted. Their properties were taken from studies by Wösten's and van Genuchten's [38]. Such parameters as: time of water infiltration through soil T_r depth of subterranean water L_c, enabling water transport to root zone, as well as flux q_d; these were then compared with the measured values, thus, quality and quantity compliance was obtained in majority cases. Water fluxes in soils were examined numerically, their saturation changes were analyzed for this purpose. The level of soil field moisture over time using different ways of irrigation, as well as the amount of accumulated water for the analyzed soils were presented. The applied research methodology is useful in analyzing complicated processes taking place in different soil formation structures.

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TEORETYCZNA ANALIZA NAWADNIANIA GLEB O RÓŻNYCH STRUKTURACH

Streszczenie

W artykule zaprezentowano wyniki badań dotyczących przepływu wody w trzech różnych typach gleb. Dane dotyczące badanych gleb uzyskano na podstawie publikacji Wöstena i van Genuchtena [39]. Matematyczny model przepływu wody został sformułowany przy wykorzystaniu równania Richardsa. Dane dla przewodności hydraulicznej oraz ciśnienia ssącego gleby nienasyconej otrzymano wykorzystując model van Genuchtena. W celu rozwiązania równania Richardsa wykorzystano metodę różnic skończonych przy zastosowaniu oryginalnego programu obliczeniowego. Zbadano czas przepływu wody przez glebę oraz głębokość wody gruntowej pozwalających na transport wody do strefy korzeni. Zjawiska te zostały zbadane dla różnych warunków brzegowych. Poddano analizie nieustalony charakter procesu nawadniania. Ponadto jednym z problemów poddanych analizie był przepływ wody w glebie po krótkotrwałych i intensywnych opadach.
Słowa kluczowe: *nawadnianie, obliczenia numeryczne, równanie Richardsa, woda w glebie*