

София, 2015. Sofia

Estimating groundwater recharge using HYDRUS-1D

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Ю. Шимунк. 2015. Оценка на подхранването на подземните води посредством компютърен код HYDRUS-1D. *Инженерна геология и хидрогеология*, 29, стр. 25 - 36.

Резюме. Моделирането на водния баланс е един от методите за количествена оценка на подхранването на подземните води. Това е от изключителна важност за определяне на водните ресурси и уязвимостта от замърсяване на водоносните хоризонти. Освен това, определянето на водния баланс близо до повърхността е от ключово значение при оценката на въздействията на изменението на климата и промяната на земеползването върху потока на подземните води.

Целта на статията е да представи математическото описание на процесите участващи в подхранването и включени в числовия код HYDRUS-1D като се представя и пример на базата на експериментални данни от района на гр. Ломел, Белгия. В този пример се разглежда водния поток през еднометров почвен профил при два варианта – без растителност и с тревно покритие. Представените резултати демонстрират, че подхранването на подземните води е по-голямо при гола почва. Тревната покривка “връща” повече вода обратно в атмосферата посредством транспирация по сравнение с физическото изпарение от почва без растителност.

Ключови думи: поток в ненаситена среда, подхранване на подземните води, моделиране, HYDRUS-1D

Abstract. Water balance modeling is one of the few available tools that can be used to predict groundwater recharge, which is critical for assessing water resources and aquifer vulnerability to contamination. Quantification of the near surface water balance is extremely important for evaluating land-atmosphere interactions, and the impact of climate change and land-use change on the subsurface flow.

The objective of this paper is to provide a mathematical description of processes implemented into the numerical model HYDRUS-1D and to provide a simple example of its application to experimental data from Lommel, Belgium. In this example, we consider transient water flow through a one meter deep soil profile, which is assumed either bare or covered with grass. Using this example we demonstrate that the soil profile with plants returns more water back to the atmosphere via transpiration than the bare soil profile via evaporation, resulting in more groundwater recharge in a bare soil profile.

Keywords: water flow, unsaturated zone, groundwater recharge, modeling, HYDRUS-1D

Introduction

Water balance modeling is one of the few available tools that can be used to predict groundwater recharge, which is critical for assessing water resources and aquifer vulnerability to contamination (Scanlon et al., 2002). Quantification of the near surface water balance is extremely important for evaluating land-atmosphere interactions, and the impact of climate change and land-use change on subsurface flow. There are different types of models that can be used to evaluate the near-surface water balance. These models vary in complexity from simple, analytical codes and storage-routing or bucket-type codes to more complex codes that use the Richards' equation.

To evaluate the near-surface water balance and estimate groundwater recharge for specific soil, vegetation, and climate conditions is one of the very common applications of the HYDRUS-1D model (Šimůnek et al., 2008) (see for example, Adomako et al., 2010; Assefa et al., 2013; Awan et al., 2013; Dickinson et al., 2013; Ficklin et al., 2010; Gates et al., 2011; Huo et al., 2014; Ibrahim et al., 2014; Jimenez-Martinez et al., 2009; Kurtzman et al., 2011; Le Coz et al., 2013; Leterme et al., 2012; Lu et al., 2011; Mastrocicco et al., 2010; Scanlon et al., 2002; Scott et al., 2000; Turkeltaub et al., 2014; Wang et al., 2009).

The objective of this brief note is to provide a mathematical description of processes implemented into the numerical model HYDRUS-1D and to provide a simple example of its application to experimental data from Belgium.

Model Description

Governing Flow Equation

One-dimensional uniform water movement in a partially saturated rigid porous medium is described in HYDRUS-1D (Šimůnek et al., 2008) by a modified form of the Richards' equation using the assumptions that the air phase plays an insignificant role in the liquid flow process and that water flow due to thermal gradients can be neglected:

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z} \left[K \left(\frac{\partial h}{\partial z} + 1 \right) \right] - S \quad (1)$$

where h is the water pressure head [L], θ is the volumetric water content [$L^3 L^{-3}$], t is time [T], z is the spatial coordinate [L] (positive upward), S is the sink term [$L^3 L^{-3} T^{-1}$], and K is the unsaturated hydraulic conductivity function [$L T^{-1}$] given by

$$K(h, x) = K_s(z) K_r(h, z) \quad (2)$$

where K_r is the relative hydraulic conductivity [-] and K_s the saturated hydraulic conductivity [$L T^{-1}$].

Root Water Uptake

The sink term, S , is defined as the volume of water removed from a unit volume of soil per unit time due to plant water uptake. Feddes et al. (1978) defined S as

$$S(h) = \alpha(h) S_p \quad (3)$$

where the root-water uptake water stress response function $\alpha(h)$ is a prescribed dimensionless function of the soil water pressure head ($0 \leq \alpha \leq 1$), and S_p the potential water uptake rate [T^{-1}]. Water uptake is assumed to be zero close to saturation (i.e., wetter than some arbitrary "anaerobiosis point", h_1). For $h < h_4$ (the wilting point

pressure head), water uptake is also assumed to be zero. Water uptake is considered optimal between pressure heads h_2 and h_3 , whereas for pressure head between h_3 and h_4 (or h_1 and h_2), water uptake decreases (or increases) linearly with h . The variable S_p in (3) is equal to the water uptake rate during periods of no water stress when $\alpha(h)=1$.

When the potential water uptake rate is non-uniformly distributed over the root zone, S_p becomes

$$S_p = b(z)T_p \quad (4)$$

where T_p is the potential transpiration rate [$L T^{-1}$] and $b(z)$ is a normalized water uptake distribution [L^{-1}]. This function describes the spatial variation of the potential extraction term, S_p , over the root zone, and is obtained by normalizing any arbitrarily measured or prescribed root distribution function.

The actual water uptake distribution is obtained by substituting (4) into (3):

$$S(h, z) = \alpha(h, z) b(z)T_p \quad (5)$$

whereas the actual transpiration rate, T_a , is obtained by integrating (5) over the rooting depth, L_R , as follows:

$$T_a = \int_{L_R} S(h, z) dz = T_p \int_{L_R} \alpha(h, z) b(z) dz \quad (6)$$

Unsaturated Soil Hydraulic Properties

The unsaturated soil hydraulic properties, $\theta(h)$ and $K(h)$, in (1) are in general highly nonlinear functions of the pressure head. HYDRUS permits the use of five different analytical models for the hydraulic properties. In this work we use the soil-hydraulic functions of van Genuchten (1980) who used the statistical pore-size distribution model of Mualem (1976) to obtain a predictive equation for the unsaturated hydraulic conductivity function in terms of soil water retention parameters. The expressions of van Genuchten (1980) are given by

$$\theta(h) = \begin{cases} \theta_r + \frac{\theta_s - \theta_r}{[1 + |\alpha h|^n]^m} & h < 0 \\ \theta_s & h \geq 0 \end{cases} \quad (7)$$

$$K(h) = K_s S_e^l [1 - (1 - S_e^{1/m})^m]^2 \quad (8)$$

where

$$m = 1 - 1/n, \quad n > 1 \quad (9)$$

and the effective saturation S_e is defined as follows:

$$S_e = \frac{\theta - \theta_r}{\theta_s - \theta_r} \quad (10)$$

The above equations contain five independent parameters: θ_r , θ_s , α , n , and K_s . The pore-connectivity parameter l in the hydraulic conductivity function was estimated (Mualem, 1976) to be about 0.5 as an average for many soils.

Initial and Boundary Conditions

The solution of Eq. (1) requires knowledge of the initial distribution of the pressure head within the flow domain:

$$h(z, t) = h_i(z) \quad t = t_0 \quad (11)$$

where h_i [L] is a prescribed function of z , and t_0 is the time when the simulation begins.

The surface boundary is exposed to atmospheric conditions. The potential fluid flux across this interface is controlled exclusively by external conditions. However, the actual flux depends also on the prevailing (transient) soil moisture conditions near the surface. The soil surface boundary condition may change from a prescribed flux to a prescribed head type condition (and vice-versa). The numerical solution of (1) is obtained by limiting the absolute value of the surface flux by the following two conditions:

$$\left| -K \left(\frac{\partial h}{\partial z} + 1 \right) \right| \leq E \quad \text{at } z = L \quad (12)$$

and

$$h_A \leq h \leq h_S \quad \text{at } z = L \quad (13)$$

where E is the maximum potential rate of infiltration or evaporation under the current atmospheric conditions [$L T^{-1}$], and h_A and h_S are, respectively, minimum and maximum pressure heads at the soil surface allowed under the prevailing soil conditions [L]. The value for h_A is determined from the equilibrium conditions between soil water and atmospheric water vapor, whereas h_S is usually set equal to zero; if positive, h_S represents a small layer of water ponded which can form on top of the soil surface during heavy rains before initiation of runoff. One option in HYDRUS is to assume that any excess water on the soil surface above zero will be immediately removed. When one of the end points of (13) is reached, a prescribed head boundary condition will be used to calculate the actual surface flux.

Another option in HYDRUS is to permit water to build up on the surface. If surface ponding is expected, a “surface reservoir” boundary condition of the type

$$-K \left(\frac{\partial h}{\partial z} + 1 \right) = q_0(t) - \frac{dh}{dt} \quad \text{at } z = L \quad (14)$$

may be applied. The flux q_0 in this equation is the net infiltration rate, i.e., the difference between precipitation and evaporation. Equation (14) shows that the height $h(L, t)$ of the surface water layer increases due to precipitation, and reduces because of infiltration and evaporation.

A free drainage (zero pressure head gradient) boundary condition is usually specified at the bottom of the soil profile. This boundary condition assumes that the groundwater table is deep below the considered soil profile and that the bottom flux is driven only by gravity.

Evaluation of Potential Evapotranspiration

Hargreaves Formula

The potential evapotranspiration can be evaluated using the Hargreaves formula (e.g., Jensen et al., 1997):

$$ET_p = 0.0023R_a (T_m + 17.8) \sqrt{\Delta T_m} \quad (15)$$

where R_a is the extraterrestrial radiation in the same units as ET_p [e.g., mm d⁻¹ or J m⁻² s⁻¹], T_m is the daily mean air temperature, computed as an average of maximum and minimum air temperatures [°C], ΔT_m is the temperature range between mean daily maximum and minimum air temperatures [°C]. The extraterrestrial radiation, R_a [J m⁻² s⁻¹], can be calculated as follows:

$$R_a = \frac{G_{sc}}{\pi} d_r (\omega_s \sin \varphi \sin \delta + \cos \varphi \cos \delta \sin \omega_s) \quad (16)$$

where G_{sc} is the solar constant [J m⁻² s⁻¹] (1360 W m⁻²), φ is the site latitude [rad], ω_s is the sunset hour angle [rad], d_r is the relative distance between Earth and Sun [-], and δ is the solar declination [rad]. The last three variables are calculated as follows:

$$\omega_s = \arccos(-\tan \varphi \tan \delta) \quad (17)$$

$$d_r = 1 + 0.033 \cos\left(\frac{2\pi}{365} J\right) \quad (18)$$

$$\delta = 0.409 \sin\left(\frac{2\pi}{365} J - 1.39\right) \quad (19)$$

where J is the number of the day in the year [-].

The minimum pressure head at the soil surface allowed under the prevailing soil conditions, h_A [L], is calculated from the air humidity, H_r [-], as follows:

$$H_r = \exp\left[-\frac{h_A M g}{RT}\right]$$

$$h_A = -\frac{RT}{Mg} \ln(H_r) \quad (20)$$

where M is the molecular weight of water [g mol⁻¹] (=0.018015 kg mol⁻¹), g is the gravitational acceleration [L T⁻²], (=9.81 m s⁻²), and R is the universal gas constant [J mol⁻¹ K⁻¹] (=8.314 J mol⁻¹ K⁻¹) [M L² T⁻² mol⁻¹ K⁻¹].

Penman-Monteith Combination Formula for Reference Evapotranspiration

Defining reference evapotranspiration (ET_0) as the rate of evapotranspiration from a hypothetical crop with an assumed crop height of 12 cm, a fixed canopy resistance of 70 s m^{-1} and an albedo of 0.23, closely resembling the evapotranspiration from an extensive surface of green grass of uniform height, actively growing, completely shading the ground and not short of water, the estimation of the ET_0 can be determined with the combination formula based on the Penman-Monteith approach. When combining the aerodynamic and radiation terms, the combination formula can be noted as (FAO, 1990):

$$ET_0 = \frac{0.408 \Delta (R_n - G) + \gamma \frac{900}{T + 273} U_2 (e_a - e_d)}{\Delta + \gamma (1 + 0.34 U_2)} \quad (21)$$

where ET_0 is the reference crop evapotranspiration [mm d^{-1}], R_n is the net radiation at crop surface [$\text{MJ m}^{-2} \text{d}^{-1}$], G is the soil heat flux [$\text{MJ m}^{-2} \text{d}^{-1}$], T is the average temperature [$^{\circ}\text{C}$], U_2 is the wind speed measured at 2 m height [m s^{-1}], $(e_a - e_d)$ is the vapour pressure deficit [kPa], Δ is the slope vapour pressure curve [$\text{kPa } ^{\circ}\text{C}^{-1}$], γ is the psychrometric constant [$\text{kPa } ^{\circ}\text{C}^{-1}$], and 900 is the conversion factor.

When no measured radiation data are available, the net radiation can be estimated as follows:

$$R_n = R_{ns} - R_{nl} \quad (22)$$

$$R_{ns} = 0.77 \left(0.25 + 0.50 \frac{n}{N} \right) R_a \quad (23)$$

$$R_{nl} = 2.45 \cdot 10^{-9} \left(0.9 \frac{n}{N} + 0.1 \right) \left(0.34 - 0.14 \sqrt{e_d} \right) (T_{kx}^4 + T_{kn}^4) \quad (24)$$

$$G = 0.14 (T_{month, n} - T_{month, n-1}) \approx 0 \quad (25)$$

where R_n is net radiation [$\text{MJ m}^{-2} \text{d}^{-1}$], R_{ns} is the net shortwave radiation [$\text{MJ m}^{-2} \text{d}^{-1}$], R_{nl} is the net longwave radiation [$\text{MJ m}^{-2} \text{d}^{-1}$], R_a is the extraterrestrial radiation [$\text{MJ m}^{-2} \text{d}^{-1}$], n/N is the relative sunshine fraction [-], T_{kx} and T_{kn} are the maximum and minimum temperatures [K], respectively; e_d is the actual vapour pressure [kPa], and G is the soil heat flux [$\text{MJ m}^{-2} \text{d}^{-1}$].

Partitioning of ET_p into E_p and T_p

Potential evaporation and transpiration fluxes can also be calculated from potential evapotranspiration using Beer's law that partitions the solar radiation component of the energy budget via interception by the canopy (Ritchie, 1972) as follows:

$$T_p = ET_p (1 - e^{-k*LAI}) = ET_p SCF$$

$$E_p = ET_p e^{-k*LAI} = ET_p (1 - SCF) \quad (26)$$

where ET_p , T_p , and E_p are potential evapotranspiration, transpiration and evaporation fluxes [$L T^{-1}$], respectively, LAI is the leaf area index [-], SCF is the soil cover fraction [-], and k is a constant governing the radiation extinction by the canopy [-] as a function of sun angle, the distribution of plants, and the arrangement of leaves (between 0.5-0.75).

Demonstrative Example

To demonstrate the use of the HYDRUS-1D software package to evaluate the near-surface mass balance and to estimate groundwater recharge, we have adopted an example from the HYDRUS short course, which uses experimental data from Seuntjens (2000). In this example we consider transient water flow through a 1-m deep multi-layered Podzol soil profile. Transient flow is induced by atmospheric boundary conditions. Either a fully plant-covered (grass) or bare soil profile is considered, thus restricting the atmospheric boundary conditions to daily values of precipitation, and transpiration or evaporation, respectively.

Input Data

Soil hydraulic and physical parameters (Table 1; Fig. 1) of the dry Spodosol located at the “Kattenbos” site near Lommel, Belgium were taken from Seuntjens (2000, Tables 3.1 and 7.1).

The initial water content conditions (Fig. 2) are defined assuming a constant flux of 0.12 cm day⁻¹ and a free-drainage lower boundary condition. The flux corresponds to the long-term (1972-1981) actual infiltration rate (precipitation - actual evapotranspiration). The upper boundary condition involves daily values of precipitation and potential evaporation (for a bare profile) or potential transpiration (for a planted profile) fluxes defined using meteorological data from the Brogel station weather (Belgium) for 1972 (Fig. 3). In a simulation with plants, it is assumed that roots are uniformly distributed in the top 30 cm.

Table 1. Soil hydraulic and physical properties of seven soil horizons

Таблица 1. Хидравлични свойства и физични параметри за седем почвени хоризонта

Horizon	Depth	Bulk density, ρ	Organic Carbon	Residual water content, θ_r	Saturated water content, θ_s	Empirical parameter, α	Empirical parameter, n	Saturated hydraulic conductivity, K_s
	cm	$g\ cm^{-3}$	%	-	-	cm^{-1}	-	$cm\ d^{-1}$
A	0 – 7	1.31	2.75	0.065	0.48	0.016	1.94	95
E	7 – 19	1.59	0.75	0.035	0.42	0.015	3.21	311
Bh1	19 – 24	1.30	4.92	0.042	0.47	0.016	1.52	39
Bh2	24 – 28	1.38	3.77	0.044	0.46	0.028	2.01	864
BC	28 – 50	1.41	0.89	0.039	0.46	0.023	2.99	1210
C1	50 – 75	1.52	0.12	0.030	0.42	0.021	2.99	1210
C2	75 – 100	1.56	0.08	0.021	0.39	0.021	2.99	1210

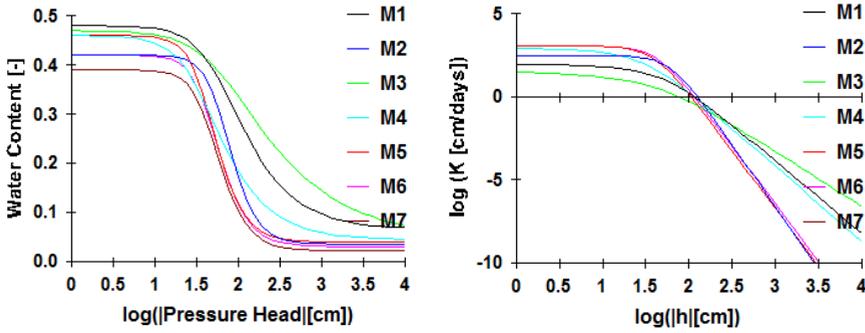


Fig. 1. Soil hydraulic properties (retention curves – left; hydraulic conductivity functions – right) of the seven soil horizons of the Podzol soil profile
 Фиг. 1. Хидравлични свойства (криви на влагозадържащата способност на почвата $\theta = f(h)$ - вляво; зависимост на коефициента на влагопренасяне от всмукващия потенциал $K = f(h)$ - вдясно) на седемте почвени хоризонти на подзолистта почва

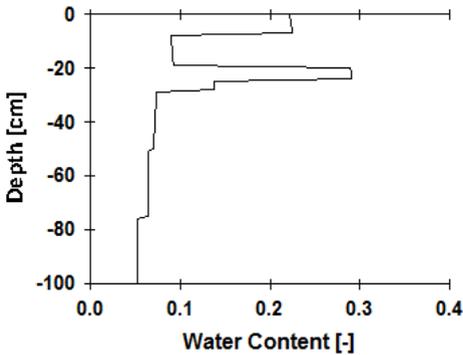


Fig. 2. Initial water contents
 Фиг. 2. Начална обемна влажност

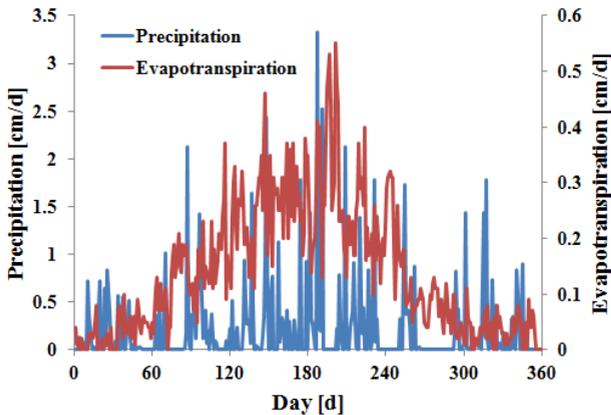


Fig. 3. Daily values of precipitation and potential evapotranspiration fluxes defined using meteorological data from the Brogel weather station (Belgium) for 1972
 Фиг. 3. Ежедневни стойности на валежите и потенциалната евапотранспирация определени по метеорологични данни за станция Brogel (Белгия) за 1972 г.

Output Data

Figure 4 shows water contents and water fluxes at depths of 50 and 100 cm during the one-year long simulation for a bare soil profile. Both water contents and water fluxes reflect the surface drivers of water flow in the soil profile, i.e., variable precipitation.

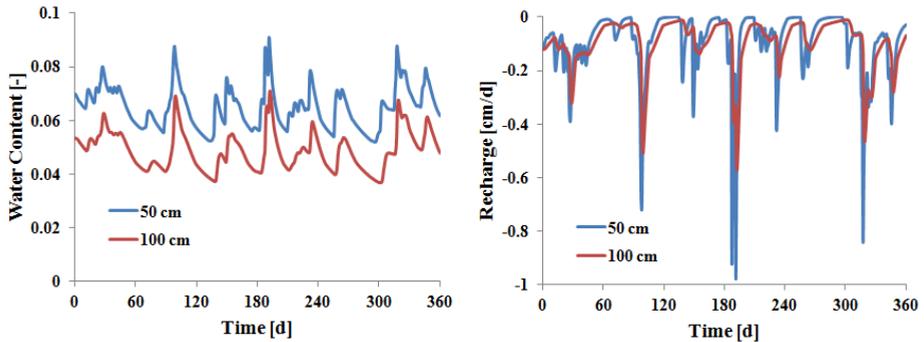


Fig. 4. Water contents (left) and water fluxes (right) at depths of 50 and 100 cm
Фиг. 4. Обемна влажност (вляво) и водни потоци (вдясно) на дълбочини 50 и 100 cm (вдясно)

Figure 5 shows main cumulative water fluxes for both bare and planted soil profile. For the bare soil profile, Figure 5 shows potential and actual cumulative surface fluxes, cumulative infiltration and evaporation, and cumulative recharge. The potential surface flux (positive downwards in Fig. 5) is the sum of precipitation and potential evaporation. Since the saturated hydraulic conductivity of the surface layer is significantly higher than the highest values of precipitation and thus all precipitation water infiltrates into the soil profile, any difference between the potential and actual surface fluxes is the result of the reduction of evaporation from its potential to actual values. Note that this reduction is about 10 cm (less water leaving the soil profile, and thus overall more water moving downwards through the soil surface). Also note that since the soil profile has little retention capacity, the cumulative recharge follows closely the actual surface flux. Any difference between these two fluxes is the change

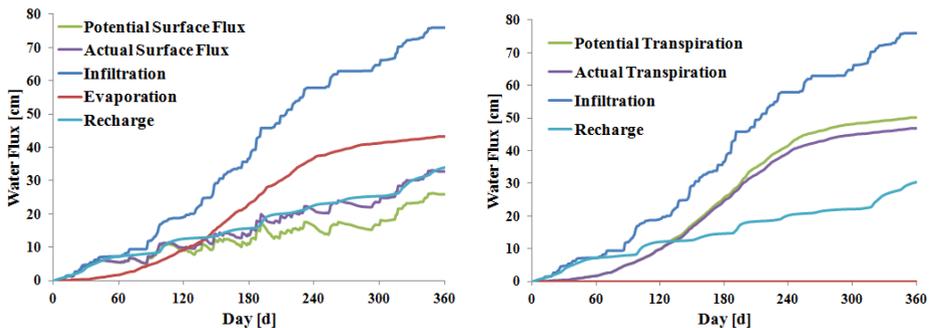


Fig. 5. Main cumulative water fluxes for the bare soil profile (left) and the soil profile with grass (right)
Фиг. 5. Основни кумулативни потоци за гол почвен профил (вляво) и затревен почвен профил (вдясно)

in water storage in the soil profile.

For the planted soil profile, Figure 5 shows potential and actual cumulative transpiration, cumulative infiltration, and cumulative recharge. Infiltration is equal to both potential and actual surface fluxes, since evaporation is considered to be zero for cropped soil profile, and again, since the saturated hydraulic conductivity of the surface layer is significantly higher than the highest values of precipitation, all precipitation water infiltrates into the soil profile. Note that the difference between potential and actual transpiration (about 3 cm) (for a cropped profile) is significantly smaller than the difference between potential and actual evaporation (about 10 cm) (for a bare soil profile). This indicates that the soil profile with plants returns more water back to the atmosphere than the bare soil profile.

Figure 6 shows cumulative recharge for the bare and planted soil profiles, reflecting the processes discussed above, i.e., that the soil profile with plants returns more water back to the atmosphere via transpiration than the bare soil profile via evaporation, resulting in more recharge in a bare soil profile.

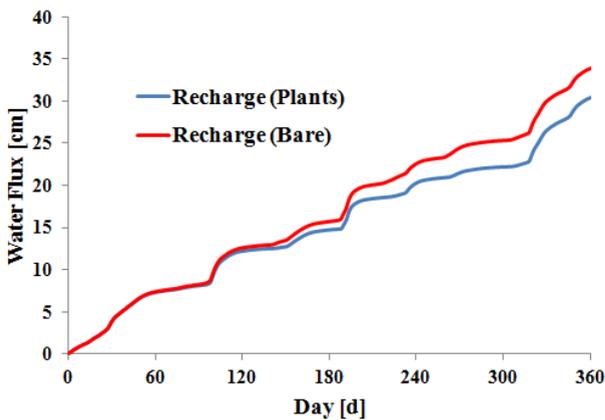


Fig. 6. Cumulative recharge for the bare and planted soil profiles

Фиг. 6. Кумулативно подхранване на подземните води за гол почвен профил и за затревен почвен профил (вдясно)

Summary

In this brief note we have provided mathematical description of processes implemented into the numerical model HYDRUS-1D and a simple example of its application to experimental data from Belgium. In this simple example, we considered transient water flow through a 1-m deep multi-layered Podzol soil profile, which was assumed either bare or covered with grass. Using this example we demonstrated that the soil profile with plants returns more water back to the atmosphere via transpiration than the bare soil profile via evaporation, resulting in more groundwater recharge in a bare soil profile.

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Постъпила: 02.11.2015

Received: 02 November 2015

Приета: 07.12.2015

Accepted: 07 December 2015