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A model for Evaporation and Drainage investigations at Ground of Ordinary Rainfed-areas

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ABSTRACT

Quantification of water losses through evaporation and drainage from bare soils in arid and semi-arid regions is very important for an effective management strategy to conserve soil water. In this study, a model for Evaporation and Drainage investigations at Ground of Ordinary Rainfed-areas (hereafter E-DiGOR) is presented. Daily potential evaporation (E_p) from bare soils was calculated using the Penman–Monteith equation with a surface resistance of zero. Actual soil evaporation (E_a) was computed according to Aydin et al. [Aydin, M., Yang, S.L., Kurt, N., Yano, T., 2005. Test of a simple model for estimating evaporation from bare soils in different environments. Ecol. Model. 182 (1), 91–105; Aydin, M., Yano, T., Evrendilek, F., Uygur, V., 2008. Implications of climate change for evaporation from bare soils in a Mediterranean environment. Environ. Monit. Assess. 140, 123–130]. Deep drainage (D) was simply calculated by the mass balance, taking field capacity into account. In order to test the performance of the model mainly for drainage estimations, a micro-lysimeter-experiment was carried out under field conditions. The experimental terrain was nearly flat, with no appreciable slope. Estimated and measured water balance components such as actual evaporation ($R^2 = 91.4\%$; $P < 0.01$), drainage ($R^2 = 88.5\%$; $P < 0.01$) and soil water storage ($R^2 = 89.7\%$; $P < 0.01$) were in agreement. This agreement supported the model hypothesis, thus rendering the model useful in estimating soil evaporation, drainage and water storage in an interactive way with a few parameters.

Once the estimated and measured data from the experiment had been compared for validation, simulations were carried out continuously for the period of 1994–2006 in a semi-arid environment of Turkey. E_p rates were lower during the winter season because of the lesser evaporative demand of the atmosphere. However, E_a rates were mainly found to be functions of the rainfall pattern, and presumably soil wetness in addition to atmospheric evaporative demand. D volumes below 120 cm soil depth were high during rainy months, with a peak in January. Annual E_p varied between 850.6 and 909.8 mm during the period of 13 years. E_a ranged from 248.0 to 392.9 mm with a mean annual value of 302.5 mm. D substantially varied inter-annually (150.5–757.4 mm) depending on the intensity and frequency of rainfall events and especially antecedent soil wetness. The next logical step in model development would be the inclusion of runoff for sloping lands.

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1. Introduction

The most important transport processes are characterized by a simultaneous change in the amount of energy and/or material with time and place (Aydin and Huwe, 1993). Soil water and energy balances interact since they are integrative aspects of the same processes within the same environment. For example, soil water potential plays a role in water flow, and water potential mainly depends on water content (Campbell, 1985). Soil evaporation is an important driving force for changes in water content. There exist many methods for direct determinations of evaporation (Boast, 1986). However, instead of adopting a method to directly measure evaporation, many researchers prefer a practical model to estimate actual evaporation. In general, models of soil water evaporation have expressed the rate of loss from cropped areas rather than from bare soils. However, in semi-arid regions, the soil evaporation represents a large fraction of the total water loss from bare fields. In regions where summer fallow is practiced, the soil water evaporation accounts for most of the incoming precipitation affecting directly soil water storage (Hanks, 1992; Hillel, 1998). Thus, many simple models are available for the reasonable predictions of soil evaporation (Gardner, 1959; Gardner and Hillel, 1962; Hanks and Gardner, 1965; Black et al., 1969; Ritchie, 1972; Van Keulen and Hillel, 1974; Hillel, 1975; Jackson et al., 1976; Campbell, 1985; Brisson and Perrier, 1991; Katul and Parlange, 1992; Malik et al., 1992; Alvenas and Jansson, 1997). The evaporation from bare soils depends not only on atmospheric conditions but also on soil properties. Parameterization of evaporation from a non-plant-covered surface is very important in the hierarchy strategy of modelling land surface processes (Mihailovic et al., 1995).

The evaporation and drainage from bare soils are strongly interdependent, as they occur sequentially and simultaneously. Unfortunately, it is almost impossible to measure seepage rate directly. Indirect experimental techniques may be used, but are cumbersome and require sophisticated equipment (Aydin, 1994). One can study water budget of a soil under natural conditions by theoretical means (Mwendera and Feyen, 1997; Eilers et al., 2007; Moret et al., 2007). A major requirement for studying vertical soil water flow is the solution of the Richards' equation (Yang and Yanful, 2002; Varado et al., 2006). This equation has a clear physical basis and is subject to specified initial and boundary conditions, with known relations among the volumetric water content, soil water potential and hydraulic conductivity. In many practical situations, however, the detailed information concerning the hydraulic conductivity and water retention relations necessary for the solutions is not readily available. In these situations, much simpler but not necessarily less precise models are required (Aydin et al., 2005). If models have a sound basis in physical science, the processes involved are accurately depicted within the models (Kite et al., 2001). Depending on the amount and quality of information available for input and required for output, either a very simple model or a very complex model may be appropriate (Aydin et al., 2005). There is no single way that is likely to be applicable to all situations. However, when the models are to be included in soil-water management systems, they need to be relatively simple and require readily available parameters (Armstrong et al., 1993).

In this study, a model for Evaporation and Drainage investigations at Ground of Ordinary Rainfed-areas (called E-DiGOR hereafter) is therefore presented and validated for the descriptions and predictions of the past, present and future dynamics.

2. Description of the model

Daily potential evaporation from bare soils can be calculated using the Penman–Monteith equation (Allen et al., 1994) with a surface resistance of zero (Wallace et al., 1999; Aydin et al., 2005):

$$E_p = \frac{\Delta(R_n - G_s) + 86.4\rho c_p \delta / r_a}{\lambda(\Delta + \gamma)} \tag{1}$$

where E_p is potential soil evaporation (kg m^{-2} d^{-1} ≅ mm d^{-1}), Δ is the slope of saturated vapour pressure–temperature curve (kPa °C^{-1}), R_n is the net radiation (MJ m^{-2} d^{-1}), G_s is the soil heat flux (MJ m^{-2} d^{-1}), ρ is the air density (kg m^{-3}), c_p is the specific heat of air (kJ kg^{-1} °C^{-1} = 1.013), δ is the vapour pressure deficit of the air (kPa), r_a is the aerodynamic resistance (s m^{-1}), λ is the latent heat of vaporization (MJ kg^{-1}), γ is the psychrometric constant (kPa °C^{-1}), and 86.4 is the factor for conversion from kJ s^{-1} to MJ d^{-1}.

Initially evaporation from a wet soil proceeds at the potential rate. With time, the soil surface becomes progressively drier, and drying front moves into the soil with a time-lag. Thus, soil water potential at the top surface layer declines. In order to estimate soil water potential at the top surface layer, Aydin et al. (2008) tested a model originally described by Aydin and Uygur (2006):

$$\psi = -\left[\frac{(1/\alpha)\left(10\sum E_p\right)^3}{2(\theta_{fc} - \theta_{ad})\left(D_{av}t/\pi\right)^{1/2}}\right] \tag{2}$$

where Ψ is soil water potential (cm of water) at the top surface layer, α is a soil-specific parameter (cm) related to flow path tortuosity in the soil, $\sum E_p$ is cumulative potential soil evaporation (cm), and θ_{fc} and θ_{ad} are average-volumetric water content (cm^3 cm^{-3}) at field capacity and air-dryness, respectively. Field capacity is defined as the amount of water, which the soil can hold against gravitational forces. D_{av} is average hydraulic diffusivity (cm^2 d^{-1}) determined experimentally, t is the time since the start of evaporation (days), and π is 3.1416.

The term, $2(\theta_{fc} - \theta_{ad})(D_{av}t/\pi)^{1/2}$, in Eq. (2), represents water supplied from deeper layers to the soil surface. This term is similar to the equation given by Gardner (1959) for the calculation of cumulative actual soil evaporation. There are two options for starting the calculations: (1) calculations start on a day that a considerable soil depth (e.g. top 5 cm) is at nearly saturation as is the case after heavy rainfall and (2) if soil water retention curve is available, an initial quantity for $\sum E_p$, which results in a Ψ value corresponding to initial water content of the top surface layer, can be used for the first day of the study period. As can be seen from Eq. (2), it was assumed that (i) water potential near the soil surface decreases as a cubic function of potential soil evaporation and (ii) upward transport of