Hydrol. Earth Syst. Sci. Discuss., 7, 6887–6923, 2010
www.hydrol-earth-syst-sci-discuss.net/7/6887/2010/
doi:10.5194/hessd-7-6887-2010

Hydrology and Earth System Sciences Discussions

Quantifying the impact of groundwater depth on evapotranspiration in a semi-arid grassland region

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Received: 12 August 2010 – Accepted: 30 August 2010 – Published: 14 September 2010

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Published by Copernicus Publications on behalf of the European Geosciences Union.

Abstract

The interactions between shallow groundwater and land surface processes, mediated by capillary rise processes from groundwater, may play an important role in the eco-hydrology of riparian zones in both humid and semi-arid ecosystems. Some recent land surface models (LSM) incorporate the contribution of groundwater to land surface processes with varying levels of complexity. In this paper, we examine the sensitivity of evapotranspiration at the land surface to the depth of groundwater using three models with different levels of complexity, two widely used representative soil hydraulic parameter sets, and four soil textures. The selected models are Hydrus-1D, which solves the Richards equation, the Integrated Biosphere Simulator (IBIS), which uses a multi-bucket approach with interactions between buckets, and a single-bucket model coupled with a classic simple capillary rise flux approximation. These models are first corroborated with field observations of soil moisture and groundwater elevation data from a site located in south-central Nebraska, USA. We then examine the sensitivity of the Richards equation to node spacing, as well as the relationship between groundwater depth and the ratio of actual to potential evapotranspiration (ET) for various soil textures and water table depths. The results show that selecting one representative soil parameter set over another may result in up to a 70% difference in actual ET (relative to the potential ET) when the depth to water table is in 0–5 m depending on the soil type. Moreover, solution type of the Richards equation and node spacing have also effect on surface ET up to 50% and 30% respectively depending on the depth-to-groundwater and node spacing. Therefore, further studies are needed to understand the sensitivities of land surface and atmospheric models to the existence of saturated layers, including studies with more field validation in regions with different climates and land cover types.

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1 Introduction

Shallow groundwater may interact with soil, vegetation, and climate through either capillary rise or direct root water uptake from the water table, impacting land surface processes and states (e.g., riparian zones and wetlands). Unlike deep water table conditions, a shallow groundwater table maintains elevated soil moistures in the root zone (Chen and Hu, 2004). Since land surface processes (e.g., evapotranspiration, runoff, and infiltration) are mainly dependent on soil moisture, incorporating groundwater in land surface models (LSMs) can be crucial (Niu et al., 2007; Yeh and Eltahir, 2005; York, 2002). Yet, little is known about the impacts of groundwater on land surface fluxes over different time and space scales. In the absence of comprehensive data, numerical models are currently used to explore the role of groundwater on land surface fluxes (Fan et al., 2007; Liang et al., 2003).

For a shallow unconfined aquifer, water can move upward from the water table to relatively drier soil surface layers through capillary rise driven by soil matric potential gradients. Quantifying capillary flux to the root zone depends on soil hydraulic properties, groundwater table depth, and soil matric potential at the lower boundary of the root zone. A number of approaches, which vary significantly in their model complexity and parameterization, have been proposed to simulate this process in various LSMs by linking groundwater and soil moisture in the root zone. The majority of recent LSMs employ the Richards equation to simulate water movement in the unsaturated zone by representing the groundwater as a simple unconfined, lumped aquifer and treating the water table as a constant-head lower boundary condition by keeping lower layers saturated in the soil column (Yeh and Eltahir, 2005; Niu et al., 2007; Fan et al., 2007). Moreover, Maxwell and Miller (2005) presented a fairly complicated approach integrating groundwater, subsurface, and surface flow components. An LSM that solves the Richards equation was coupled with a 3-dimensional aquifer model by replacing the soil column/root zone soil moisture formulation of the LSM with the groundwater model's formulation. They concluded that the coupled LSM and the groundwater flow model yields more realistic simulations of soil moisture and runoff.

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The Richards equation is the most widely accepted physically-based model used to simulate variably saturated flow in porous media:

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

where θ is volumetric water content [$L^3 L^{-3}$]; $K(h)$ is unsaturated hydraulic conductivity [LT^{-1}]; h is the water pressure head [L]; z is the vertical coordinate [L] and S is the rate of root water uptake [$L^3 L^{-3} T^{-1}$]. The Richards equation can be defined in three basic forms: (1) a pressure-based form (i.e., h-based), (2) a volumetric water content-based form (i.e., θ-based), and (3) a mixed form as shown by Eq. (1) (Celia et al., 1990).

Solving the Richards equation requires the knowledge of the relations among θ, h, and K (Brooks and Corey, 1966; Clapp and Hornberger, 1978; van Genuchten, 1980; Rawls et al., 1982). However, due to the highly nonlinear nature of the relations among θ, h, and K, analytical solutions of the Richards equation only exist for few simplified boundary conditions and specific soil hydraulic descriptors (Zlotnik et al., 2007). Therefore, numerical techniques are needed to solve the Richards equation for more general applications (Warrick, 2003).

Many numerical studies have used either h-based or θ-based forms of the Richards equation to describe the flow in unsaturated zone using variety of solution techniques (e.g. Hills et al., 1989; Kirkland et al., 1992). Overall, the numerical solution of the θ-based Richards equation yields more accurate mass balance and computational efficiency in very dry soils and is, therefore, the preferred form in most LSMs (Dickinson et al., 1993; Sellers et al., 1996). However, the application of the θ-based form is problematic when dealing with saturated soil layers, as unlike hydraulic pressure heads, soil moistures do not change within homogeneous and inelastic saturated porous media (Celia et al., 1990; Pan and Wierenga, 1995; Zeng and Decker, 2009; de Rooij, 2010). Nevertheless, the θ-based Richard equation has been used in some LSMs with groundwater aquifers represented by saturated soil layers below the unsaturated zone (Kim and Eltahir, 2004; Yeh and Eltahir, 2005). In spite of the drawbacks of h- and

θ-based forms of the Richards equation, many studies have focused on combining the advantages of the two forms (e.g. Allen and Murphy, 1986; Celia et al., 1990). The mixed form of the Richards equation provides solution in terms of hydraulic pressure heads, while it is considered as mass conserving.

On the other end of the spectrum, simple analytical solutions have also been employed to couple groundwater and land surface processes in some LSMs, such as the Gardner-Eagleson model (G-E model) that can be used to calculate steady soil moisture fluxes from the water table (Gardner, 1958; Eagleson, 1978; Famiglietti and Wood, 1994). The G-E model is derived from the Darcy-Buckingham equation based on the assumptions of steady-state conditions and a completely dry surface at the upper boundary, which, however, greatly limits its applications. As such, the G-E model is often used to calculate fluxes under mean annual conditions. Recently, similar models with varying degrees of complexity have been proposed to relax the requirement of the upper boundary in the G-E model. Also based on the Darcy-Buckingham equation, Bogaart et al. (2008) offered a set of closed-form equations, which account for both root zone soil moisture and depth to the water table. Vervoort and van der Zee (2008) provided a piecewise linear equation for calculating soil moisture fluxes from the water table, which depends on the potential capillary flux and the actual evaporative demand, and coupled the proposed model to a stochastic soil moisture accounting model (Laio et al., 2001). Ridolfi et al. (2008) suggested an analytical framework to couple soil moisture dynamics and groundwater fluctuations under bare soil conditions, which was later extended to vegetated conditions (Laio et al., 2009).

Despite the previous efforts, there is still a paucity of research on assessing different numerical schemes for calculating soil moisture fluxes from the water table within a consistent framework, which is critical for the application of those numerical schemes. Therefore, this study aim to investigate the impacts of different model structures and parameterizations on quantifying groundwater-land surface interactions, and to examine sensitivity of those models to various soil parameters and evapotranspiration models. Four models are selected in this study, including the Hydrus-1D model

(Simunek et al., 2005), the Integrated Biosphere Simulator (IBIS) model (Foley et al., 1996), and two modified G-E models, along with two widely used representative soil hydraulic parameters set (Clapp and Hornberger, 1978; Rawls et al., 1982). Here, the Hydrus-1D model is used as a "benchmark" model with the most complex parameterization for subsurface flow. The IBIS model serve as an intermediate-complexity model with multiple buckets that exchange soil moisture based on the Richards equation. Finally, the G-E model is coupled with a single-bucket model (Laio et al., 2001), which assumes successive steady-state conditions at each time step. Different from previous studies that used synthesized data (Shah et al., 2007; Bogaart et al., 2008), a detailed hydrometeorological data set is used in this study to examine the impact of water-table depths on annual and mean annual evapotranspiration in a semi-arid region.

2 Model descriptions

2.1 Benchmark model: Hydrus-1D

In this study, the Hydrus-1D model (Simunek et al., 2005) is used as a benchmark model because it has been previously verified by analytical techniques (Zlotnik et al., 2007) and used in numerous studies (Desilets et al., 2008; Twarakavi et al., 2009). On the basis of Eq. (1), the Hydrus-1D model solves variably saturated flow in homogenous and rigid porous media. The term *S(h)* in Eq. (1) is simulated according to the method of Feddes et al. (1978):

$$S(h) = \alpha(h) S_p, \tag{2}$$

where S_p is the potential root water uptake rate [$L^3 L^{-3} T^{-1}$], which is assumed to be equal to the potential transpiration. The term $\alpha(h)$ is a dimensionless, prescribed function of soil water pressure head ($0 \leq \alpha(h) \leq 1$), which relates the root water uptake under different soil moisture levels to the maximum potential rate using a piecewise empirical relationship:

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$$\alpha(h) = \begin{cases} 0 & h \le h_{\text{wilt}} \\ \left(\frac{h-h_{\text{wilt}}}{h^*-h_{\text{wilt}}}\right) & h_{\text{wilt}} < h \le h^* \\ 1 & h^* < h, \end{cases} \quad (3)$$

where h_{wilt} and h^* are water pressure heads at the wilting point and incipient stomata closure point, respectively. Below h_{wilt} plants cannot extract water, and $\alpha(h)$ equals zero. Between h_{wilt} and h^*, root water uptake is limited by soil moisture and increases linearly with hydraulic head as the soil gets wetter. Above h^*, the stomata – and likewise the root water uptake – are not constrained by soil moisture.

In our model simulations, the Clapp and Hornberger (1978) model is used to relate soil matric potential to soil moisture and unsaturated hydraulic conductivity:

$$\theta(h) = \theta_s \left(\frac{h}{h_{ae}}\right)^b, \quad (4a)$$

$$K(\theta) = K_s \left(\frac{\theta}{\theta_s}\right)^{\frac{2}{b}+3}, \quad (4b)$$

where θ_s is saturated soil moisture content [$L^3\,L^{-3}$], h_{ae} is an air entry (bubbling) pressure [L], b is a pore size distribution index [–], and K_s is saturated hydraulic conductivity [LT^{-1}]. This form of the Richards equation is used in various models in the recent literature (Tripp and Niemann, 2008; Wang et al., 2009), sometimes with other water retention curves such as the van Genuchten model (van Genuchten, 1980).

2.2 Integrated biosphere simulator (IBIS)

IBIS is a dynamic global vegetation model that integrates various terrestrial ecosystem processes within a single, physically consistent framework (Foley et al., 1996). IBIS simulates land surface physics, terrestrial carbon balance, vegetation dynamics and phenology, and canopy physiology. Detailed descriptions of the IBIS model can be

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found in Foley et al. (1996), Kucharik et al. (2000) and Lenters et al. (2000). Here the components of IBIS that are most relevant to the focus of this paper are discussed.

In its standard version, IBIS simulates upper and lower vegetation canopies, 3 snow layers, and 11 soil layers. The total thickness of the soil layers is 2.5 m, and the dynamics of soil water content and ice content were simulated in each layer. The model solves the θ-based form of the Richards equation for soil moisture flow between layers:

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z}\left(D(\theta)\frac{\partial \theta}{\partial z}\right) + \frac{\partial K(\theta)}{\partial z} - S(h), \tag{5}$$

where $D(\theta) = K(h)\left(\partial h / \partial \theta\right)$ is the diffusion coefficient [$L^2\,T^{-1}$].

The sink term in IBIS can be described as a function of plant transpiration (T) through

$$S(h) = T \cdot F \tag{6}$$

where F is the water uptake fraction, which is a function of root distribution and soil water content. The transpiration function in IBIS is based on the work of Pollard and Thompson (1995):

$$T_u = \frac{\rho s_u}{(1 + r_u s_u)}\left(1 - f_u{}^{wet}\right)\left(q_{sat}(L_u) - q_{12}\right)\mathrm{LAI}_u \tag{7a}$$

$$T_l = \frac{\rho s_l}{(1 + r_l s_l)}\left(1 - f_l{}^{wet}\right)\left(q_{sat}(L_l) - q_{34}\right)\mathrm{LAI}_l \tag{7b}$$

where the subscripts u and l represent the upper and lower canopy, respectively, ρ is the density of near-surface air [ML^{-3}], f_{wet} is the fraction of leaf area wetted by intercepted water or snow [–], L is the leaf temperature [T], s is the heat/vapor transfer coefficient between canopy and air [LT^{-1}], r is stomatal resistance per unit leaf area [TL^{-1}], $q_{sat}(K)$ is the saturation specific humidity at ambient pressure [MM^{-1}], q_{12} and q_{34} are upper and lower canopy specific humidity [MM^{-1}], respectively, and LAI is canopy leaf area index (single sided) [$L^2\,L^{-2}$]. Total transpiration is calculated by summing T_u and

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T_I. Total "actual" evapotranspiration ($\mathrm{ET_a}$) is calculated as the sum of total transpiration, evaporation of water intercepted by vegetation, and evaporation from the soil surface (Pollard and Thompson, 1995).

IBIS originally has 11 soil layers with varying thicknesses from 5 cm to 50 cm. To capture the soil moisture gradient and the groundwater level change, and to make the IBIS simulations compatible with Hydrus-1D simulations that use smaller node spacings, we change the thicknesses of each layer to 2.5 cm and the number of the soil layers to 100. In our model simulations, the water table is positioned by saturating the soil layers below the top of the capillary fringe, in which soil is saturated, rather than the water table. The average thicknesses of the capillary fringes for sand, silt loam, silt clay loam and clay are taken as 5 cm, 32.5 cm, 45 cm and 32.5 cm respectively (Mausbach, 1992 as cited in Tiner, 1999).

2.3 Coupled root-zone and steady-state groundwater upward flux models

2.3.1 Gardner-Eagleson (G-E) model

The G-E model offers an analytical solution to calculate the upward flux of soil moisture from the water table. The derivation of this model was first given by Gardner (1958) and later modified by Eagleson (1978). Under steady-state conditions, the change in water content becomes zero ($\partial\theta/\partial t = 0$). Ignoring root water uptake, and assuming that the capillary flux (G) [$\mathrm{LT^{-1}}$] is constant and only in the vertical upward direction, the Darcy- Buckingham equation as follows:

$$G = K(h)\left(\frac{\partial h}{\partial z} - 1\right), \tag{8}$$

Eq. (8) can be rearranged and integrated from the water table to the upper boundary at the soil surface as:

$$z = \int_0^{h_\mathrm{r}} \frac{dh}{1 + (G/K(h))}, \tag{9}$$

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where h_r is the root zone water potential [L] and G is the steady capillary flux [LT^{-1}]. In order to solve Eq. (9), Gardner (1958) used an empirical expression relating $K(h)$ with h through $K(h) = a/(h^n + c)$, where a and c are constants. In this equation, c is a relatively small number compared to h_n (and is often neglected), and a is defined as $K_s h_{ae}{}^n$, where K_s is saturated hydraulic conductivity, h_{ae} is air entry pressure, and n is a soil index which can be calculated from the pore size distribution index (b) according to $n = 2+3/b$. With these modifications, the $K(h)$ equation of Gardner (1958) reduces to the form of Brooks and Corey (1966) and Campbell (1974):

$$K(h) = K_s \left(h_{ae}/h\right)^n. \tag{10}$$

Substituting Eq. (10) into Eq. (9), gives:

$$z = \int_0^{h_r} \left[1/\left(1+\alpha h^n\right)\right] dh, \tag{11}$$

where $\alpha = G/(K_s h_{ae}{}^n)$. Gardner (1958) showed that Eq. (11) can be solved analytically for certain n values. He suggested a simple analytical solution for calculating capillary rise, in which the soil surface is assumed to be completely dry (i.e. h_r is minus infinity), thereby maintaining a strong upward gradient and resulting in the following equation for capillary flux:

$$G = BK_s \left(h_{ae}/z_{gw}\right)^n, \tag{12}$$

where z_{gw} represents the depth to groundwater and B is a parameter – often taken from a look-up table (Table 1) – that depends solely on the value of n.

In addition to the solution of Gardner (1958), Ripple (1972) suggested various graphical solutions to Eq. (11), and Anat et al. (1965) developed some approximate solutions in the case of $n > 1$ for Eq. (9). Warrick (1988) extended the analytical solutions of Eq. (9) for various n values, assuming the Brooks-Corey retention curve model. However, these solutions cannot be explicitly written in terms of $q(h,z)$ and $h(q,z)$ except for the case of $q \ll K_s$. Approximate analytical models were presented by Eagleson (1978)

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and later modified by Salvucci (1993). Eagleson (1978) suggested a continuous relationship to extend B over the full range of soil index values using the following empirical function:

$$B = 1 + \left[(3/2)/(n-1)\right]. \tag{13}$$

Substituting Eq. (13) into (12) gives the G-E model that we use in this paper:

$$G = \left(1 + \left[(3/2)/(n-1)\right]\right) K_s \left(h_{ae}/z_{gw}\right)^n. \tag{14}$$

It is important to note here that in Eq. (14), the soil surface is assumed to be dry with an infinitely low soil matric potential. In this study, the original G-E model and a modified version of it are coupled to a bucket-type vadose zone hydrology model, in which the depth of the bucket is defined by rooting depth. This coupling is necessary, as our intention is to examine the influence of groundwater on evapotranspiration. In the modified model, the pressure head, which is calculated from the depth-averaged soil moisture by using a soil water potential relationship in the root-zone, is used as the upper limit of the integral in Eq. (11). This effectively replaces the dry soil assumption in the G-E model with more realistic surface soil moisture values. It is assumed that the depth-averaged soil moisture in the root zone characterizes the soil suction pressure for the boundary condition to the G-E equation. Numerical integration of Eq. (11) is used to apply this modification to the model, as the upper boundary is no longer taken as infinity. For numerical integration of Eq. (11), we use the trapezoidal rule to calculate the capillary flux to the root zone.

2.3.2 Bucket soil moisture model

The G-E capillary flux model was coupled to a leaky bucket-type soil moisture model. With the steady-state groundwater capillary flux added to the root zone, the rate of change in the depth-averaged soil moisture in the root zone can be calculated (Brolsma and Bierkens, 2007) according to:

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$$nZ_r \frac{ds}{dt} = I(s) - \mathrm{ET}(s) - \mathrm{L}(s) + \mathrm{G}(s, z_{gw}), \tag{15}$$

where n is porosity [$\mathrm{L}^3\,\mathrm{L}^{-3}$], Z_r is rooting depth [L], s is the degree of saturation (volumetric soil moisture normalized by porosity) [–], I is infiltration rate, ET is evapotranspiration, L is drainage rate and G is capillary flux rate (all in [LT^{-1}]). Infiltration rate is defined as rainfall limited by the infiltration capacity and in excess of canopy interception:

$$I(s) = \begin{cases} \min[p, I_c] & 0 \le s < 1 \quad P > C_I \\ K_s & s = 1 \quad P > C_I \\ 0 & 0 \le s \le 1 \quad P \le C_I \end{cases}, \tag{16}$$

where p is rainfall rate [LT^{-1}], I_c is infiltration capacity [LT^{-1}], P is total rainfall [L], and C_I is canopy interception [L]. Runoff is generated when the canopy can no longer intercept additional precipitation and the rainfall rate exceeds the infiltration capacity, or the soil is fully saturated.

Drainage from the bottom of the root zone is assumed to occur at the same rate as the saturated hydraulic conductivity when the soil is saturated. For unsaturated soils ($s < 1$), the drainage rate is represented by the unsaturated hydraulic conductivity according to Campbell (1974):

$$L(s) = \begin{cases} K_s & s = 1 \\ K_s s^{(2b+3)} & s_{fc} < s \le 1 \end{cases}, \tag{17}$$

where s_{fc} is the degree of soil saturation at field capacity. Actual evapotranspiration follows widely used soil moisture limitation function (e.g., Laio et al., 2001):

$$\mathrm{ET}(s) = \begin{cases} 0 & s_h < s \le s_w \\ \mathrm{ET}_p \left(\frac{s - s_w}{s^* - s_w} \right) & s_w < s \le s^* \\ \mathrm{ET}_p & s^* < s < 1 \end{cases}, \tag{18}$$

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where s_h is soil hygroscopic capacity, s_w is wilting point and s^* is the soil moisture threshold for incipient stomata closure under plant water stress.

In the coupling the bucket model and the G-E model, we assume that capillary flux occurs between the water table and the bottom of the root zone and we run simulations at daily time steps using grass as the prescribed plant parameters.

Two forms of the G-E model are used. In the first application, we use the original G-E model by setting the upper boundary to have infinite soil matric potential. This represents a one-way coupling in which the G-E model provides water flux into the root-zone only. In the second application, we update the upper boundary for the modified G-E model by setting it to the depth-averaged soil matric potential in the root zone. In this way, the G-E model provides upward flux into the root-zone at a rate conditioned on the root zone soil moisture. In both cases evapotranspiration is calculated as a function of root zone soil moisture following Eq. (18).

The soil and plant parameters used in the bucket model and the Hydrus-1D model are shown in Table 2. (Hereafter, the "leaky bucket type soil moisture model coupled with the G-E model" and the "modified G-E model" are referred to simply as the "G-E model-1" and the "G-E model-2," respectively.)

3 Field site and data

3.1 Study area

Our field data were collected at a wetland in the Republican River basin in south-central Nebraska, USA. The study area is located at 40°17.91′ N and 99°57.90′ W and has an average elevation of 640 m above sea level in a region of relatively low relief (Fig. 1). The climate is semi-arid, with a long-term mean annual precipitation of 430 mm. The majority of the precipitation falls in the summer season. Croplands are dominant in the region, with limited trees except in the riparian zone near the Republican River and other areas where the water table is near the surface. Regions in which the water table is above the ground surface are generally occupied by tall grasses and open water (maximum depth ~1 m). A variety of data have been collected at our field site since 2008, including groundwater levels, soil water content and meteorological data. Data collected during the 2009 growing season were used for validation purposes (Fig. 2), and a High Plains Regional Climate Center (HPRCC) meteorological station near Champion, Nebraska was used to drive model sensitivities and comparisons over a longer, 10-year period (1999–2008).

3.2 Model verification against field observations

At the study site, the soil moisture probes were horizontally located at depths of 10 cm, 20 cm, and 50 cm below the surface. The soil moisture values from the three probes are averaged together to compare the simulation results. Similarly, we create three observation points in the IBIS and Hydrus-1D models – located at the same depths as the field soil probes – and average them together for comparison with the observations. The rooting depth in the G-E models is set to 50 cm, and no averaging is required since the G-E models already provide depth-averaged soil moisture values in the root zone.

The Hydrus-1D model, G-E models, and IBIS are evaluated with field observations collected from the wetland site (Fig. 1). Daily groundwater levels and observed precipitation data used to drive the models are shown in Fig. 2. To examine the role of groundwater, we first run the models assuming free drainage conditions (i.e., no groundwater influence). Subsequently, we replace the lower boundary with the time series of observed water table depth. Under free drainage conditions (no groundwater), the G-E models become reduce to a lumped root zone soil moisture model. Soil type in the models is set to sand using the representative soil parameter values from Rawls et al. (1982), and the soil parameters are not adjusted to do any calibration. We only calibrate parameters that are specific to each model, such as the fraction of bare soil (set to 0.3) in the bucket model and the extinction coefficient in Beer's Law equation (set to 0.5) to calculate potential evaporation and transpiration from potential evapotranspiration (Ritchie, 1972) in the Hydrus-1D model. Hourly input data are used

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in both the Hydrus-1D and IBIS models, while daily input data are used in the G-E models.

Separate, detailed parameter optimizations – which could have improved the simulation results for the IBIS and Hydrus-1D models – are not applied to the various models. However, the same soil parameters are used in all models. Despite the range of complexities among the models, each one showed improvements in the soil moisture simulation when groundwater is represented as the lower boundary condition (Fig. 3). The G-E models, in particular, show very good agreement with the observed soil moisture values.

Although the simulated soil moisture values are close to the observed data for all models, the responses of each model to groundwater level rise and precipitation pulses are slightly different. While the soil moisture response to precipitation is relatively strong in Hydrus-1D and IBIS, the G-E models response to precipitation events is more muted. This may be related to the fact that when the G-E models receive a precipitation pulse, it instantaneously averages the pulse into its root zone. On the other hand, groundwater level changes have a stronger and more immediate effect on soil moisture in the G-E models than in the other models. The steady state assumption in the bucket model capillary rise equation is the most likely reason for this more immediate response to groundwater level changes. In other words, when the water table elevation changes, the calculated amount of capillary rise flux based on the new water table depth is immediately adjusted in the root zone soil moisture model. IBIS, however, does not seem to be as sensitive as the other models to groundwater level changes.

3.3 Model sensitivity experiments

We conduct two sets of sensitivity experiments. In the first set of simulations, we examine the models' responses to widely used soil representative soil hydraulic parameter sets with variety of soil textures, and water table depths on evapotranspiration, as simulated by Hydrus-1D. Moreover, using 2 different node spacing, a sensitivity analysis is done to examine the effect of node spacing on evapotranspiration. In the second set of experiments, we compare the G-E models and IBIS with Hydrus-1D separately to investigate the responses of the models with various complexities in areas where water table is shallow.

The atmospheric forcing from the High Plains Regional Climate Center (HPRCC) station near Champion, Nebraska is used in each simulation. The model results are presented as the ratio of the long-term mean annual actual to potential evapotranspiration (i.e., ETa/ETp). The primary reason for using this ratio is to examine whether the land surface is water limited or energy limited. In other words, the ETa/ETp ratio indicates what percentage of the available energy (ETp) is partitioned into latent heat flux (ETa) for a given simulation.

The Hydrus-1D and coupled bucket/G-E models require ETp and precipitation as inputs for the upper boundary condition. ETp is calculated using the Priestley-Taylor method, which requires net radiation and air temperature. IBIS requires incoming solar radiation, air temperature, relative humidity, precipitation, and wind speed as inputs. All of these variables are taken directly from the HPRCC station observations. Constant head lower boundary condition is used to represent the groundwater table for the models. To minimize the effects of initial soil moisture in Hydrus-1D, the simulation of the first year of the time period is repeated several times until the soil moisture profile reached an equilibrium. Similarly, for the IBIS runs, several years of "spin-up" time are simulated prior to the time period of interest.

In the Hydrus-1D and G-E models simulations, we vary the depth-to-groundwater from 1 to 14 m (in 1-m increments) and we use Rawls et al. (1982) representative soil hydraulic parameters. ETp values are calculated on a daily time step for the 10-year period (1999–2008).

IBIS uses a multilayer soil characteristics datasets (CONUS-SOIL) for regional simulations like many other climate and land surface models. The maximum depth for this dataset is 2.5 m from the surface. Hence, we compare Hydrus-1D's results with IBIS' results over a limited range of depths. Thickness and number of soil layers are modified to constant node spacing of 2.5 cm and number of soil layers are adjusted to 100 layers. Nine simulations are run by increasing the water table from 2 m to the surface.

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Prescribed vegetation (C3 grass) is used in IBIS, and the soil thickness in Hydrus-1D is also set to 2.5 m. Both models are run at hourly time steps for the 10-year period 1999–2008 using data from the HPRCC station. Since IBIS uses Rawls et al. (1982) soil hydraulic parameters as a default, we use these parameters in Hydrus-1D as well, and we also run the models with four different soil textures. Calculation of evapotranspiration in IBIS is done quite differently than in Hydrus-1D. While Hydrus-1D calculates ETa based on available water content and prescribed ETp values (using Eqs. 2 and 3), IBIS calculates ETa based on Eq. (7) and the primary equations of the Pollard and Thompson (1995) land surface transfer scheme. To ensure use of the same atmospheric forcing in both models, we performe a set of IBIS simulations with all soil layers saturated (and for various soil textures) and then use the IBIS-simulated ETa as ETp inputs for Hydrus-1D.

Finally, we note that in the application of the Richards equation, there is usually some level of sensitivity to node spacing. Experimenting with the Richards equation, van Dam and Feddes (2000) reported that a node spacing of ~5 cm or larger may not correctly estimate evaporation and infiltration, especially in layers close to the surface with a shallow groundwater table. We performe our own tests with Hydrus-1D by examining the difference in ETa rates for small (1.5-cm) and large (30-cm) finite element node spacings. To reduce the uncertainty of model predictions due to the selection of the root profile and its empirical parameterization, we use a uniform root distribution, with a root depth of 50 cm. The Hydrus-1D model is run at daily time steps for a 10-year period (1999–2008). As in the first set of experiments, 56 simulations are run for each of the two node spacings (i.e., 14 water table depths and four soil textures). In order to also examine the effect of soil parameters, we repeate this same set of experiments using an alternative to the Rawls et al. (1982) representative soil parameter set – namely that of Clapp and Hornberger (1978).

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4 Results and discussion

4.1 Evapotranspiration as function of groundwater depth: influence of soil texture and node spacing

First we explore the role of soil texture and node spacing on ETa using the Hydrus-1D model. Daily Hydrus-1D model output for 10 years are used to calculate the mean annual evaporation fluxes from the root zone, which are then normalized by ETp and plotted as a function of depth-to-groundwater. There are two widely used representative soil parameter sets that utilize soil texture information – namely, Clapp and Hornberger (1978) and Rawls et al. (1982). Both soil parameter sets are separately incorporated into our Hydrus-1D simulation, with various water table depths. The results are shown in Fig. 4. Although it has been suggested that the use of soil texture may be inappropriate for estimating soil hydraulic parameters (Gutmann and Small, 2005), the availability of global soil texture maps makes soil texture a commonly used predictor of soil hydraulic parameters.

Kollet and Maxwell (2008) described the "critical zone" as the region in which a strong correlation between ETa and water table depth exists, and they found this zone to occur at depths of 1–5 m in their study area with generally loamy sand and loam soil textures. Figure 4 suggests that the thickness of the so-called critical zone depends on the soil type and selected representative soil parameter set. The critical zones in Fig. 4 generally concur with the results of Kollet and Maxwell (2008), with a few exceptions. For example, when the Clapp and Hornberger (1978) soil parameter set is used, the critical zones are deeper than when the Rawls et al. (1982) soil parameter set is used (for all four soil types). Thus, models that use the Clapp and Hornberger (1978) parameter set would produce more surface evapotranspiration because of the higher contribution of groundwater to soil moisture. The differences in ETa, as a result of using a different soil parameter set for the same soil texture, are particularly large in the critical zone (Fig. 4), and would cause significant differences in the partitioning of available energy into latent and sensible heat flux in LSMs. The magnitude of these

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differences, at least in our model simulations, is up to 70% for the ETa/ETp ratios, especially in the case of sand and silt loam. This implies that the model is highly sensitive to the choice of soil hydraulic parameters (especially when the water table is in the critical zone) and that the soil moisture in the root zone varies significantly between the two sets of model simulations. On the other hand, it is also interesting to note that the two sets of model simulations converge to identical ETa/ETp as the water table deepens. (The case for silt loam appears to converged as well, though not by a depth of 15 m.) This implies that the choice of soil hydraulic parameter set becomes less important as the depth-to-groundwater increases beyond the critical zone, and the land surface and groundwater becomes decoupled.

The sensitivity analysis of the effect of node spacing indicates that the use of thicker soil layers leads to higher surface ETa, particularly in the near-surface layers (0–3 m; Fig. 4). Solving the Richards equation with the coarser node spacing of 30-cm, the Hydrus-1D results show both higher root uptake and lower infiltration rates, which causes up to a 30% increase in ETa/ETp.

It has been shown that root-zone soil moisture affects atmospheric processes such as surface temperature (Chow et al., 2006), winds, cloud cover, and precipitation dynamics (Chen and Avissar, 1994: Maxwell et al., 2007). Therefore, recent studies have focused on coupling LSM's with groundwater models that have varying degree of complexities. Importantly, our results also show that the selection of soil parameters and the solution scheme of the Richards equation may result in significant differences in soil moisture and ETa/ETp, especially in areas where the water table is close to the surface (such as wetlands and riparian corridors). These differences may, in turn, cause significantly different feedbacks to atmospheric models that could modify the response to external forcings, such as those associated with climate change. However, which of these soil databases would lead to better predictions in regions underlain by shallow groundwater table is yet to be explored in detailed field studies.

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4.2 Inter-model comparisons

To investigate potential errors in the numerical solution of commonly used mass-conservative, θ-based forms of the Richards equation when the soil domain contains saturated layers, we compare the results of the IBIS and Hydrus-1D model simulations (both run at hourly time steps for a 10-year period). The results are shown in Fig. 5. As the water table gets closer to the surface, the ETa/ETp ratios from the two model simulations converge toward each other. In the case of sand, the solutions also converge as the depth-to-groundwater gets large. However, the difference between the ETa/ETp ratios from the two models can be as high as 50% if the water table is in the so-called critical zone. The main reason for this significant difference in ETa/ETp is the difference in solution schemes of the Richard equations between the two models. This implies that the use of θ-based forms of the Richards equation may create significant errors in simulating soil moisture when the water table is in the critical zone. And since surface evapotranspiration is maintained through capillary rise from the water table in dry seasons, this is a very critical issue in terms of vegetation dynamics, as well as surface energy, water, and carbon fluxes (Nepstat et al.,1994).

As the water table approaches the surface, the ETa/ETp ratios converge to a similar solution, since the models do not need to simulate capillary flux to increase the root zone soil moisture (i.e., the layers are already saturated). Similarly, as the water table deepens, the modeled ETa/ETp ratios become more similar because of the inverse relationship between capillary rise and depth-to-groundwater. To examine the ETa/ETp ratios without the effect of groundwater, simulations were also run in which free-gravity drainage is allowed to occur in the lower boundaries of the models without the existence of any saturated layer (Table 3). In this case, the average difference in ETa/ETp ratios between the two models becomes as low as 7%. This close match between the two models and the advantages of the θ-based form of the Richards equation (such as robustness and computational efficiency) reflect some of the reasons why most LSMs use the θ-based form over the h-based form when the effects of groundwater is not considered.

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In Fig. 6, we compare the ETa/ETp ratio simulations of G-E model-1 and G-E model-2 with that of Hydrus-1D. The soil parameter values used in the models are identical. Each point represents a ratio of the 10-year mean ETa to ETp for various water table depths, ranging from 0 to 14 m in 1-m increments. In general, the agreement among the three models is very good, especially in clay and silty clay loam soil types, implying that the G-E capillary rise flux model gives satisfactory results for this particular study area despite the model's simplicity and computational efficiency. The G-E models predict slightly lower values than Hydrus-1D, even when the water table is below the critical zone. This disparity can be attributed to the bucket hydrology model used to represent the surface soil moisture dynamics. With further exploration of the underlying causes of this could even improve the consistence between the two models.

5 Summary and conclusion

In this study, we investigate the effects of groundwater depth on surface evapotranspiration for areas where the water table is shallow. These effects are studied for various soil types, representative soil hydraulic parameter sets and models. Two models are tested against a benchmark model (Hydrus-1D) to compare their accuracies and sensitivities – one being IBIS, with the lower boundary condition modified to introduce a constant groundwater table, and the other being a bucket-type model coupled with a groundwater component. In addition, the solution of the benchmark model is tested by changing its soil parameters and node spacing.

Representative soil hydraulic parameter sets from Clapp and Hornberger (1978) and Rawls et al. (1982) are used separately with various water table depths in the Hydrus-1D simulations. We find that selecting one soil hydraulic parameter set over another may result in up to a 70% difference in actual evapotranspiration (relative to potential ET), when the water table is located in the critical zone (i.e., roughly 0–5 m to 0–15 m, depending on the soil type). This result indicates that the Hydrus-1D model is highly sensitive to soil hydraulic parameters, especially when the water table is in the critical

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zone. Moreover, this significant difference in ETa/ETp stems from the difference in simulated soil moisture, which is known to affect atmospheric processes such as surface temperature, wind, cloud cover, and precipitation. Therefore, this difference is likely to be important for simulating feedbacks to atmospheric models and studying responses to external forcings such as climate change.

Generally, most LSMs are not designed to simulate the effects of groundwater on soil moisture. Thus, their numerical solution of the Richards equation utilizes the θ-based form, which yields a more accurate mass balance and higher efficiency in dry soils as compared to the h-based form. However, it has been shown to not be an appropriate solution form when saturated layers exist (Celia et al., 1990; Pan and Wierenga, 1995). Despite this deficiency of the θ-based form of the Richards equation in simulating the vertical movement of water in areas where the water table is shallow, the θ-based form is widely used in current LSMs, sometimes to simulate the effects of groundwater on the surface energy balance. Our own comparison of IBIS (an LSM which uses the θ-based form of the Richards equation) and Hydrus-1D (a vadose zone model which solves the mixed form of the Richards equation) indicates that the difference between ETa/ETp ratios may be as high as 50% if the water table is located at a depth of 75–200 cm. The thickness of the soil layers also has an effect on partitioning the surface energy balance when the water table is in the critical zone. Solving the Richards equation with a coarser node spacing causes both higher root uptake and lower infiltration rates, and our sensitivity analysis shows an increase of up to 30% in ETa/ETp ratios. These results suggest that other LSM models that have soil moisture solution schemes similar to IBIS may need further modifications to simulate groundwater effects appropriately, such as improvements in the solution type of the Richards equation and an increase in the number of soil layers.

Solving the Richards equation can be numerically demanding, especially when used in a high spatial resolution setting. Therefore, capillary flux type models – such as the G-E model used in the current study – can be used as an alternative way to link groundwater and soil moisture. In this study, comparisons of the G-E model (coupled

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with a bucket-type model) with Hydrus-1D showed satisfactory results, even with the assumption that the soil is completely dry in the root zone. In the comparisions with observed soil moisuture and groundwater depths in our study site, the G-E based models give more accurate results than both IBIS and Hydrus-1D.

Finally, we note that it is well known that neglecting the groundwater component in LSMs may result in significant errors in the surface energy and water balance, especially in areas where the water table is shallow. In addition, we find that coupled models may also lead to inaccurate results depending on the choice of soil parameters and solution methods for simulating the interaction between saturated and unsaturated zones. Hence, further studies are needed to understand the sensitivities of LSMs to the existence of saturated layers, including studies with more field validation in regions with different climates and land cover types.

Acknowledgements. The authors would like to thank Gregory Cutrell, Kyle Herrman, Steven Walters, and Durelle Scott for their help with the field studies, as well as the High Plains Regional Climate Center (HPRCC) for supplying some of the meteorological data. This research was supported through funding from the Nebraska Environmental Trust (NET), the University of Nebraska Water Resources Advisory Panel (WRAP), and the University of Nebraska Rural Initiative.

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Table 1. *B* values used to determine the capillary flux for certain soil indices (n) in the case of a completely dry surface, based on the analytical solution of Gardner (1957) (as given in Eq. 10).

n	B
3/2	3.77
2	2.46
3	1.76
4	1.52

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Table 2. Soil parameters utilized in the Hydrus-1D and bucket models.

	K_s (m/day)	h_{ae} (m)	n	b	s_h [a]	s_w [a]	s^* [b]
			Clapp and Hornberger (1978)				
Sand	15.21	0.121	0.395	4.05	0.107	0.144	0.335
Silt Loam	0.62	0.786	0.485	5.30	0.259	0.374	0.617
Silty Clay Loam	0.15	0.356	0.477	7.75	0.358	0.403	0.649
Clay	0.11	0.405	0.482	11.40	0.503	0.559	0.754
			Rawls et al. (1982)				
Sand	5.20	0.072	0.437	1.69	0.004	0.007	0.053
Silt Loam	0.16	0.207	0.501	4.74	0.166	0.214	0.440
Silty Clay Loam	0.04	0.326	0.471	6.62	0.297	0.356	0.595
Clay	0.01	0.373	0.475	7.63	0.355	0.415	0.648

[a] Caylor et al. (2005).
[b] Rodriguez-Iturbe and Porporato (2004).

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Table 3. The ratios of ETa/ETp as simulated by IBIS and Hydrus-1D using free drainage lower boundary conditions (total soil thicknesses of the models are 2.5 m).

	IBIS	Hydrus-1D
Silt Loam	0.370	0.409
Clay	0.328	0.391
Silty Clay Loam	0.372	0.421
Sand	0.250	0.382

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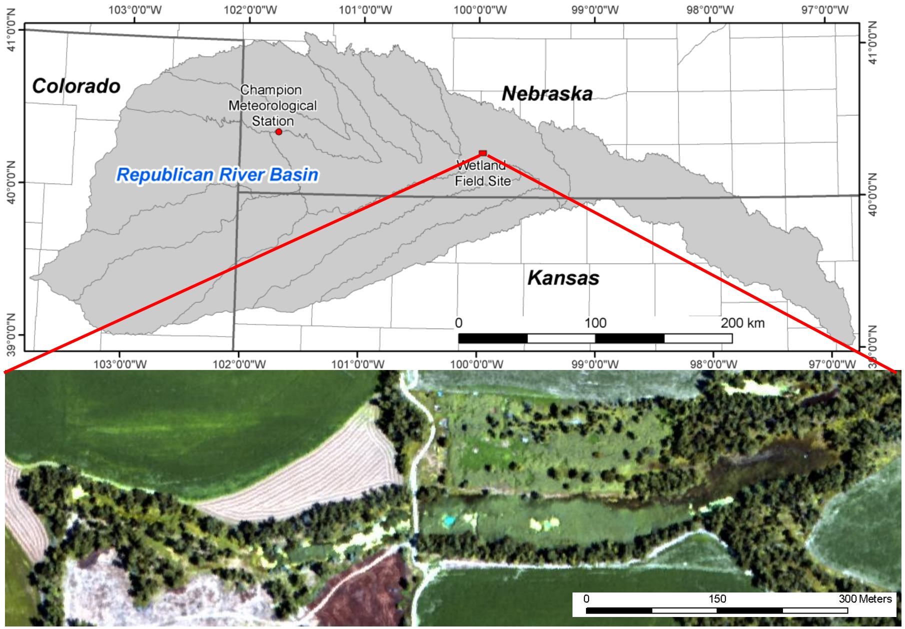

Fig. 1. Map showing the wetland field site in the Republican River basin and the location of the HPRCC meteorological station in Champion, Nebraska.

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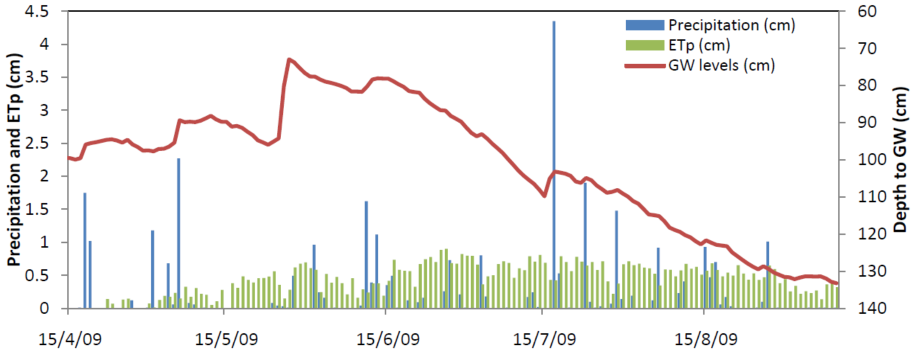

Fig. 2. Observed precipitation and calculated ETp and depth-to-groundwater during the 2009 growing season for the study site in south-central Nebraska.

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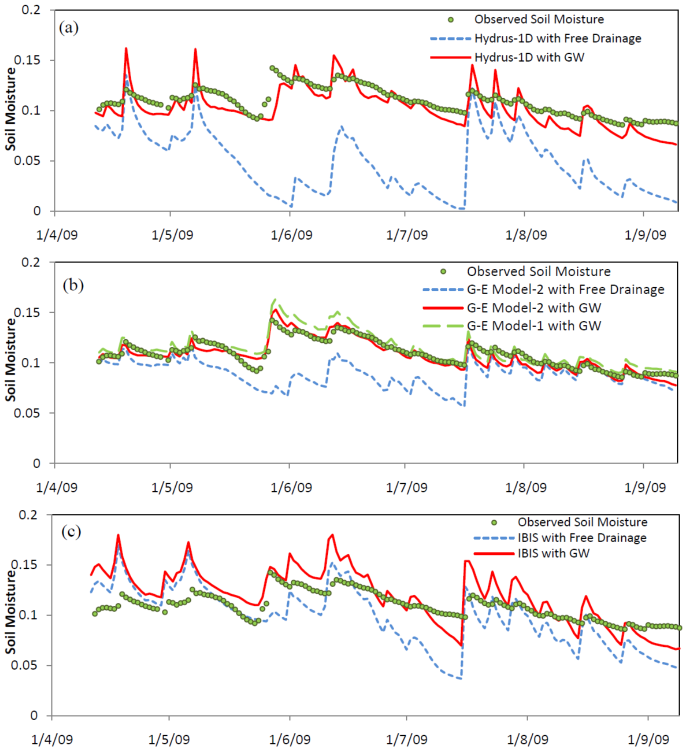

Fig. 3. Observed and simulated soil moisture for the wetland site in south-central Nebraska. Simulation results using free drainage and observed groundwater table depths as the lower boundary condition are represented by dashed and solid lines, respectively. Green dots show the observed soil moisture. **(a)** Hydrus-1D simulations, **(b)** G-E models simulations, and **(c)** IBIS simulations.

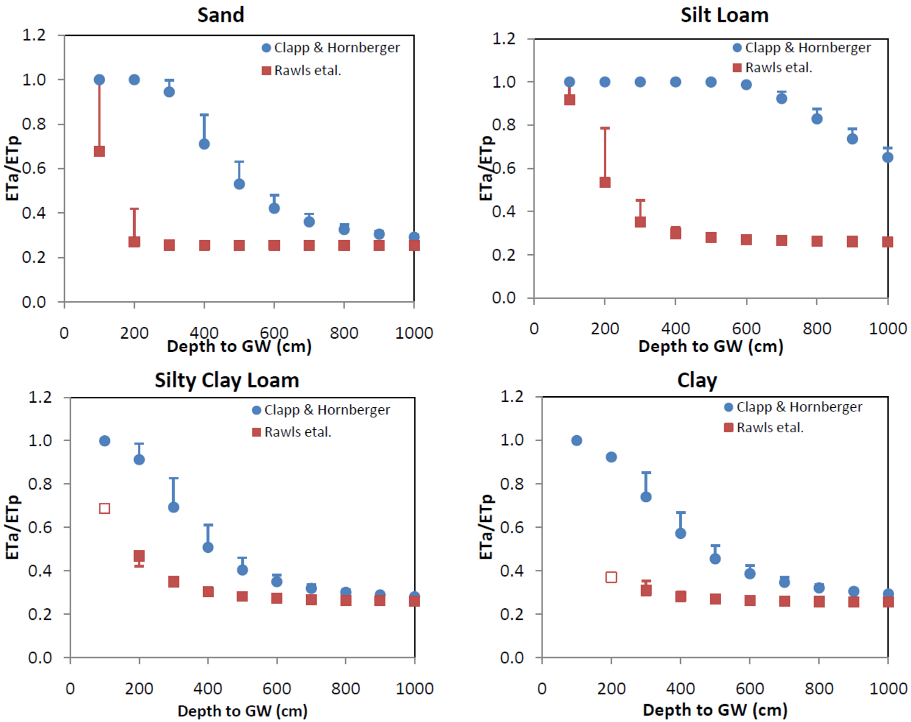

Fig. 4. Hydrus-1D mean annual ETa/ETp versus depth-to-groundwater. Circles and squares indicate model solution using Clapp and Hornberger (1978) and Rawls et al. (1982) soil parameters, respectively (both with 1.5-cm node spacing). Upper error bars represent the solution using 30-cm node spacing, while empty squares indicate that these solutions did not converge using a 30-cm nodal distance.

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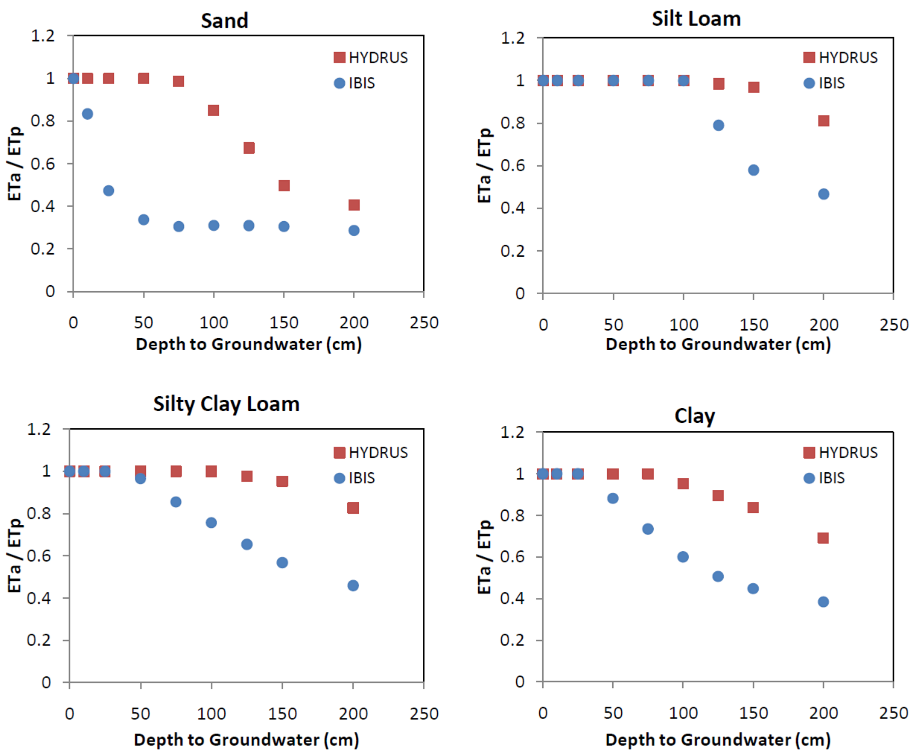

Fig. 5. Hydrus-1D and IBIS results showing mean annual ETa/ETp versus depth-to-groundwater (using hourly time steps and 10-year simulations). The Rawls et al. (1982) soil parameter set is used for all simulations.

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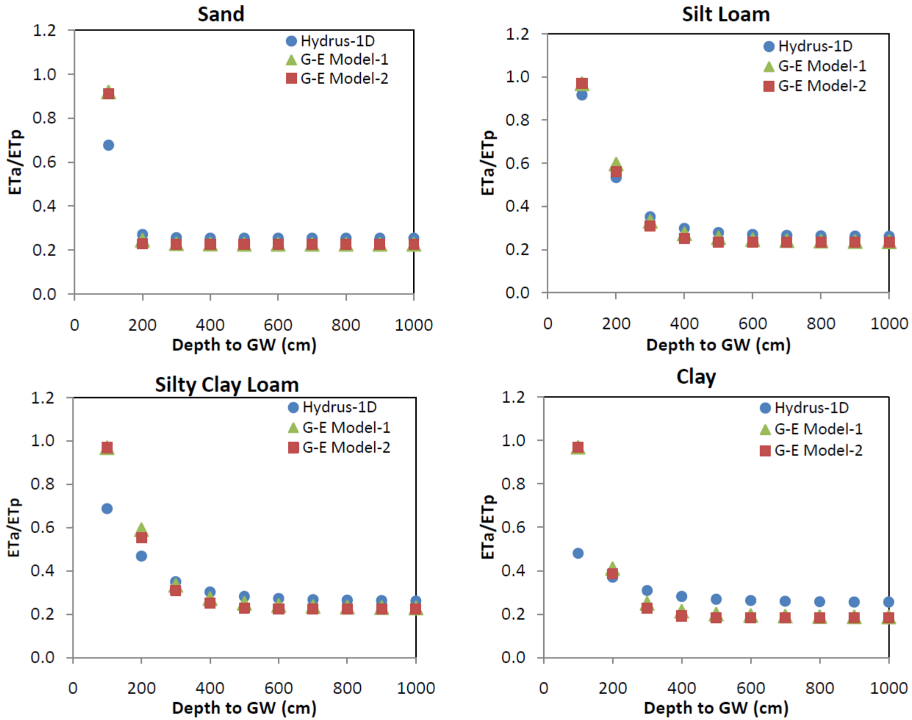

Fig. 6. Results from the Hydrus-1D and G-E models showing mean annual ETa/ETp versus depth-to-groundwater (using daily time steps for 10 years). The Rawls et al. (1982) soil parameter set is used for all simulations.

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