



Stochastic soil water dynamics of phreatophyte vegetation with dimorphic root systems

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Received 25 June 2008; revised 26 June 2009; accepted 6 July 2009; published 29 October 2009.

[1] As the direct uptake of deep groundwater by vegetation may be essential in semiarid regions, we incorporated this process in stochastic root zone water balance models. The direct water uptake by vegetation via deep tap roots is simulated using one additional empirical parameter. This is considered for the case of feedback with root zone saturation and without such feedback. The model that accounts for feedback between shallow root zone saturation and groundwater uptake by deep roots takes up less water if the shallow root zone is wet. The behavior of the models demonstrates that for certain combinations of climate and groundwater depths this feedback becomes important in determining differences in total evapotranspiration (ET). This feedback mechanism also captures hydraulic redistribution processes. The range of relative contributions of groundwater to ET predicted by the models was similar to values derived in isotope studies.

Citation: Vervoort, R. W., and S. E. A. T. M. van der Zee (2009), Stochastic soil water dynamics of phreatophyte vegetation with dimorphic root systems, *Water Resour. Res.*, 45, W10439, doi:10.1029/2008WR007245.

1. Introduction

[2] In semiarid regions, small differences in water availability may significantly affect the coverage of the land surface by vegetation, the feedback on partitioning of latent and sensible heat, and the albedo of the land surface [McAlpine *et al.*, 2007; Maxwell and Kollet, 2008]. Consequently, feedback relationships between land surface and atmosphere processes involved in energy and water balances are receiving considerable attention [Pitman *et al.*, 2004; Lawrence *et al.*, 2007; McAlpine *et al.*, 2007; Maxwell and Kollet, 2008].

[3] Most ecohydrological models have ignored interaction of vegetation with groundwater [e.g., Rodriguez-Iturbe and Porporato, 2004]. Water uptake by trees from groundwater or from soil moisture in deeper soil layers can be important for survival. Trees in semiarid areas often use groundwater for 50% of their actual transpiration [Mensforth *et al.*, 1994; Thorburn *et al.*, 1995; Zencich *et al.*, 2002; Lamontagne *et al.*, 2005]. Since root water uptake from shallow soil layers may be physiologically preferable, as it is energetically more efficient and shallow root zone water commonly contains more nutrients [Mensforth *et al.*, 1994; Zencich *et al.*, 2002; O'Grady *et al.*, 2006a], riparian trees are often facultative users of groundwater [Busch *et al.*, 1992; Dawson and Pate, 1996; Snyder and Williams, 2000; O'Grady *et al.*, 2006b]. They rely on groundwater during rather dry periods when shallow soil water is insufficiently available.

[4] Two mechanisms can be postulated with regard to the availability of groundwater to vegetation: (1) capillary fluxes

of water lead to enhanced water supply into the root zone after which it can be taken up by the vegetation [Ridolfi *et al.*, 2008; Vervoort and Van der Zee, 2008] and (2) part of the root mass is in direct contact with the groundwater or capillary fringe and is able to take up water without requiring capillary fluxes in soil. If only capillary fluxes are providing the water needed for evapotranspiration (case 1), the main variables of interest are the depth of groundwater and the force with which water is pulled upward (evapotranspiration demand or root zone water potential) [Eagleson, 1978; Salvucci, 1993; Thorburn *et al.*, 1995; Vervoort and Van der Zee, 2008]. The capillary flux is only comparable to common values of the evapotranspiration demand, if the lower boundary of the root zone is at a relatively short distance from the groundwater table. For most soils, meaningful capillary fluxes will not occur for distances exceeding 2 m [Vervoort and Van der Zee, 2008].

[5] Case 2 relates to the direct uptake of groundwater, which may involve tap roots that extend significantly below the upper active root zone. Uptake is mainly governed by the ability of roots near the groundwater table to take up water. Therefore, knowledge of the relative root density distribution may become important, but also water movement to roots, specific behavior of tap roots, and energy expenditure for plants to invest in developing tap roots. Detailed numerical simulations provide insight into such specific root water uptake processes [de Jong van Lier *et al.*, 2006]. However, analytical approaches are less data intensive and can provide a quick insight into ecohydrological trade-offs between soil moisture and groundwater.

[6] In many ecohydrological applications, the active root depth (Z_r) of vegetation is often taken as 50 or 100 cm [Eagleson, 1978; Entekhabi *et al.*, 1992; Milly, 2001; Rodriguez-Iturbe and Porporato, 2004], and roots are assumed to be distributed homogeneously over this depth. The "active soil volume" where roots extract water [Guswa *et al.*, 2002] is subsequently defined by ϕZ_r , where ϕ is the

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porosity. The reported root depth for many trees can reach up to 68 m with a mean value of 7 m [Canadell *et al.*, 1996; Zeng, 2001]. While overall root depths can be quite large, root densities generally decline very rapidly with depth [Schenk and Jackson, 2002].

[7] Several relative root density models have been suggested in the literature [Jackson *et al.*, 1996; Feddes *et al.*, 2001; Zeng, 2001; Schenk and Jackson, 2002]. However, on the basis of the literature the following observations can be made: (1) all models basically follow a power model or exponential decline with depth, probably because they are all based on the same empirical data and (2) all models indicate few roots below 1 m, possibly because the empirical data included few semiarid deep groundwater situations, or because tap roots may have quite different functionalities.

[8] In one of the few observational studies in south Australia, the root density declined exponentially with depth for a mature Jarrah tree over deep groundwater (15 m) [Carbon *et al.*, 1980]. There is also substantial evidence of a “dimorphic root system” of *Eucalyptus* and *Banksia* spp. [Dawson and Pate, 1996] which consists of many shallow lateral roots and a few deep sinker or tap roots [Knight, 1999; Schenk and Jackson, 2002]. Such dimorphic root system would have a substantially lower root density in the subsoil compared to the root distributions in the upper zone [Lee *et al.*, 2005] and thus root density would decline rapidly with depth.

[9] Given the above and as solid experimental evidence that suggests otherwise is lacking, we assume that the relative root density ($R(z)$) for the vegetation can be described simply as [Feddes *et al.*, 2001]

$$R(z) = \exp(-c \cdot z/100) \quad (1)$$

where z is the depth (cm) and c is the e -folding parameter (m^{-1}) which describes the decline of the relative root density with depth z .

[10] While root density is important, root functionalities can override the importance of the root density in water uptake. For example, root systems may have a compensating ability, which means they can take up more water in areas where water availability is larger [Guswa, 2005; Teuling *et al.*, 2006]. Physiologically, this is possible as the hydraulic conductivity of tap roots can be much greater than for lateral roots [Pate *et al.*, 1995; Dawson and Pate, 1996; McElrone *et al.*, 2004], and low root densities can still take up significant amounts of water as is demonstrated by isotope studies [Dawson and Pate, 1996; Mensforth and Walker, 1996; Zencich *et al.*, 2002; O’Grady *et al.*, 2006b]. Such complexities can be simulated in a zero-dimensional model using a compensation factor [Guswa, 2005].

[11] Our work is motivated by our interest in the impact of a few deep tap roots beyond the upper active root zone, i.e., in the behavior of a dimorphic root system under different climate and water table conditions. In particular, we are interested in the feedback between root zone moisture and direct uptake of groundwater on the transpiration of vegetation. Rather than using a multi layer approach [Baudena and Provenzale, 2008] we choose to follow the ecohydrological framework of Rodriguez-Iturbe and Porporato [2004] by considering a perfectly mixed root zone layer, similar as we did earlier [Vervoort and Van der Zee, 2008].

If this appears useful, such an approach could be extended to a multilayer model later. The aim of this study is therefore to develop a simple but relatively realistic water balance model for water-limited vegetation with groundwater (or deep subsoil water) uptake using tap roots and to determine the relative contribution of the uptake from groundwater to the transpiration. Therefore, we consider three conceptual models which differ with regard to their treatment of direct, groundwater uptake by the root system and the level of feedback between upper root zone soil saturation and groundwater uptake.

2. Methods

2.1. Theory

[12] As a start, we consider a homogeneous soil with an active shallow root zone that extends from soil surface to a depth Z_r (cm). A groundwater table occurs at a depth Z (cm) below the soil surface (Figure 1a) and the distance from the soil surface (z , cm) is positive downward. The shallow root zone ($0 < z < Z_r$) can be seen as the depth in which the majority of all root activity takes place and is most sensitive to climate forcing [Teuling *et al.*, 2009]. From equation (1) we derive the fraction of roots present in the root zone layer $0 < z < Z_r$ which we then distribute homogeneously over this layer. As pointed out by Guswa [2005], this assumption could overestimate the transpiration, and we discuss later how this assumption affects our results. We will assume that the soil water profile below the upper root zone (Z_r) is continuously at steady state, which implies that the perturbations in the groundwater table are at a much larger temporal scale than the perturbations in the climate (i.e., days and weeks for climate versus years for groundwater) and drained water instantaneously reaches the groundwater. Hence, in contrast to [Ridolfi *et al.*, 2008], the vegetation and the climatic forcing have little impact on the groundwater level because the groundwater level is relatively deep or the aquifer volume large relative to the vegetation uptake. The groundwater may be brackish or saline, but we do not yet account for salinity induced by capillary fluxes [Shah *et al.*, 2009]. The soil is characterized by the hydraulic properties h_b (the capillary rise or bubbling pressure), the saturated hydraulic conductivity K_s and by exponential soil water retention and unsaturated hydraulic conductivity functions [Salvucci, 1993].

[13] Throughout this study, we use the equation given by Teuling and Troch [2005] to calculate evapotranspiration from the shallow root zone, ET_{ZR} , from the potential evaporation rate E_p :

$$ET_{ZR} = f_r \beta (1 - e^{-0.45\xi}) E_p \quad (2)$$

In equation (2) we used the value of 0.45 for the solar radiation extinction coefficient [Teuling and Troch, 2005], ξ is the leaf area index (LAI), f_r is the fraction of the soil containing roots, and β in equation (2) is the soil water stress function defined as

$$\beta = \begin{cases} 0 & 0 \leq s < s_w \\ \frac{s - s_w}{s^* - s_w} & s_w \leq s < s^* \\ 1 & s^* \leq s < 1 \end{cases} \quad (3)$$

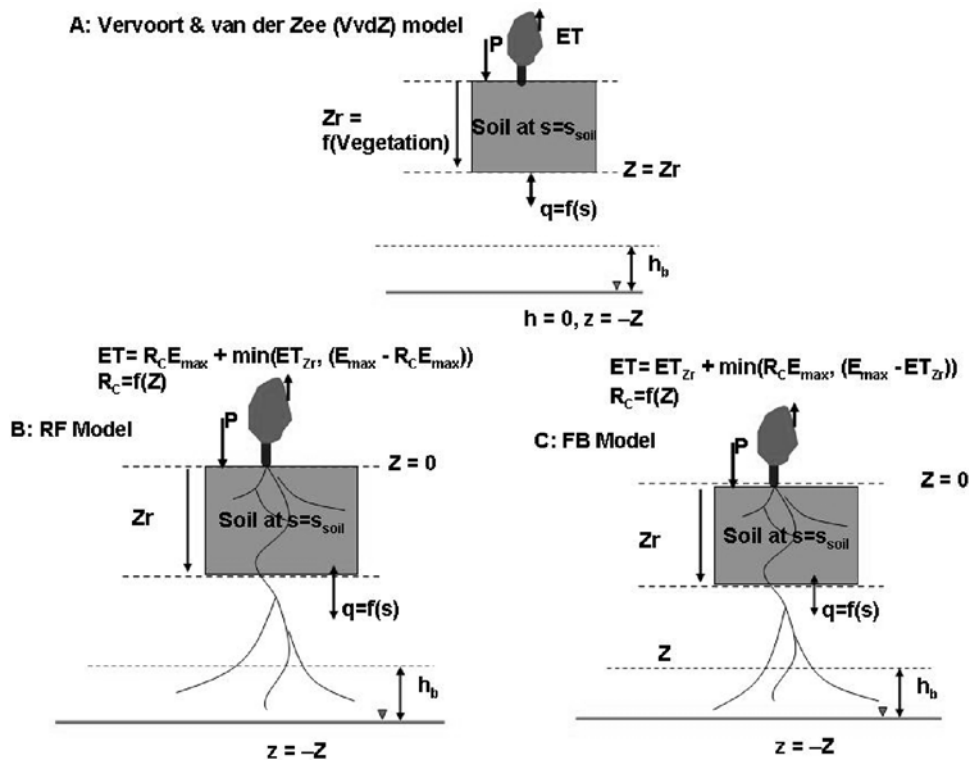


Figure 1. Conceptual models of the different groundwater, deep root zone, and upper root zone combinations in this paper.

with s^* the soil saturation level at which the transpiration is reduced and s_w is the wilting point [Porporato *et al.*, 2001]. Equation (2) ignores soil evaporation (i.e., $ET = 0$ if $\xi = 0$). In semiarid regions, evaporation from the soil surface is concentrated in a shallow top layer and the rapid decrease of s and hydraulic conductivity reduce evaporation strongly for s dropping below the field capacity (s_{fc}) [Wythers *et al.*, 1999; Lauenroth and Bradford, 2006; Yopez *et al.*, 2007]. It seems clear to us, that bare soil evaporation may be significantly reduced by such an evaporation barrier, but we do not suggest it to be insignificant. For the entire water balance and particularly for the amount of water that may leach, soil evaporation may be important. For results such as the probability density of s , we believe that the errors (on the dry part of the saturation range) are acceptable.

[14] Here, we have initially assumed that the relative root density model in equation (1) can be interpreted as a model which combines the relative root density and the local relative root water uptake capacity. This will then involve an adjustment of the parameter c in equation (1) to account for both the relative root density and the root water uptake capacity.

2.1.1. Eagleson Model: Capillary Fluxes From Groundwater to the Root Zone

[15] The conceptual Eagleson model (Figure 1a) is the classical model of groundwater interaction with the root zone through capillary rise as defined by Eagleson [1978]. Capillary flux is driven by the depth of the water table, but soil water losses through evapotranspiration (ET) only occur in the root zone in the upper part of the soil column as no roots extend beyond Z_r (Figure 1a). This implies that $ET = ET_{Z_r}$. Then, the capillary flux is constant over time and depends only on the soil hydraulic properties. The capillary flux (q_{cap})

at Z_r is calculated from the steady state approximation on the basis of Darcy's law:

$$Z - Z_r = \int_0^{h(z=Z_r)} \frac{dh}{1 + q_{cap}/K(h)} \quad (4)$$

which describes the maximum height $Z - Z_r$ for which a designated capillary flux (q_{cap}) can be supplied for the particular soil hydraulic conductivity function $K(h)$ and potential h at Z_r . This results in the following approximation [Eagleson, 1978]:

$$q_{cap} = K_s \alpha_e \left(\frac{h_b}{Z - Z_r} \right)^{2b+3}$$

$$n = 2 + 3/b$$

$$\alpha_e = 1 + \frac{3/2}{n - 1} \quad (5)$$

where α_e and n are parameters in the groundwater uptake function related to the exponent b of the exponential soil water function. The total flux at the bottom of the root zone (q) can be both positive (drainage from the root zone to the water table) and negative (capillary uptake) [cf. Eagleson, 1978, p. 729, equation (62)]:

$$q = -K_s \left(s^{2b+3} - \alpha_e \left(\frac{h_b}{Z - Z_r} \right)^n \right) \quad (6)$$

where s is the soil saturation. This approach assumes that the potential at the upper boundary (in our case Z_r) is constant at $-\infty$. While this would not be the case in our model

(Figure 1a), the potential at the bottom of the root zone would still be several orders of magnitude lower than the potential at the groundwater table.

2.1.2. RF Model: Direct Groundwater Uptake Based on Root Fraction

[16] Besides the Eagleson (E) model, we also consider a model where, in addition to capillary fluxes, deep roots also take up water to meet evaporative demand. This uptake is determined by the ability of the roots in or near the capillary fringe to supply water to the vegetation (Figure 1b). We again assume a homogeneous root distribution in the shallow root zone ($0 < z < Z_r$), similar to the E model. Drainage and capillary fluxes at Z_r therefore follow equation (6). However, below Z_r a few roots extend to the groundwater and in the root fraction (RF) model, water uptake from the groundwater is a constant fraction of the maximum possible ET . We continue to call $0 < z < Z_r$ the shallow root zone, despite a few roots crossing the layer below Z_r toward the groundwater. The capacity of the deep roots (below Z_r) to take up water at depth Z will be fully accounted for through a function $R_C(Z)$.

[17] We visualize that this capacity is limited by two components: (1) the fraction of roots below Z_r and (2) the gravity driven hydraulic gradient that the vegetation needs to overcome to pump the water into shallow the root zone [Lee *et al.*, 2005] or to the leaves. The second component implies that groundwater uptake; hence, $R_C(Z)$ decreases with increasing Z .

[18] An appropriate approximation of $R_C(Z)$ would be the probability density of the root distribution between Z and Z_r based on equation (1) weighted by the potential difference due to gravity:

$$R_C(Z) = \frac{Z}{Z - Z_r} \frac{\int_{Z_r}^Z R(z) dz}{\int_0^\infty R(z) dz} \quad (7)$$

$$R(z) = \exp(-c \cdot z/100)$$

In equation (7) any roots predicted by $R(z)$ below Z are considered inactive. Given the exponential form of $R(z)$ this possibly introduces errors for small values of c and Z . Whereas for an impression of model assumptions, parameters can be freely varied, it is unlikely that all parameter combinations are equally probable, or even realistic. For instance, most plants (except species such as rice and reed) do not grow roots below the water table, in view of anaerobicity. Hence, small c values and relatively shallow groundwater tables (small Z values) are not probable, for typical semiarid phreatophyte vegetation.

[19] With $(Z/(Z - Z_r))$ we have only considered the gravitational component on the basis of a linear relationship and have assumed that all other components of the potential difference between root and leaf can be seen as vegetation specific [Porporato *et al.*, 2001]. Other models for $R_C(Z)$ are possible, for example, if the vegetation is able to compensate for the gravitational difference through increased hydraulic conductivity in the root system [Dawson and Pate, 1996] than the gravitational gradient could be ignored and groundwater uptake would be independent of Z . Alterna-

tively a nonlinear relationship with Z could be considered, where groundwater uptake decreases sharply with depth, if field data would support this.

[20] Equation (7) assumes that the actual conductivity of the root system is independent of the potential gradient or the saturation of the root itself. In view of the continuous supply of groundwater, loss of conductivity due to cavitation is unlikely in the root system [Hultine *et al.*, 2003]. Although we assume an exponential model for the root density (i.e., sharply decreasing with depth), another shape for $R(Z)$, e.g., where the highest density of the deep roots is concentrated at the capillary fringe or groundwater fringe [Carbon *et al.*, 1980; Dawson and Pate, 1996; Knight, 1999; Schymanski *et al.*, 2008], would also fit to our RF model. In fact, we do not make any necessary limiting assumptions regarding the exact distribution of the roots below Z_r and assume complete compensation below Z_r [Guswa, 2005]. This means we assume that the deep roots will access the location with the maximum available water, whether this is groundwater or another deep store.

[21] In view of equation (7), it is logical to consider a relationship between the parameter f_r in equation (3) and the function $R(z)$. If the cumulative root density in the root zone is higher, the fraction of roots in the deeper layers will be smaller according to equation (1), and also the capacity to extract water from the groundwater decreases according to (7). Hence, f_r in equation (2) is defined as

$$f_r = \frac{\int_0^{Z_r} R(z) dz}{\int_0^\infty R(z) dz} \quad (8)$$

While f_r is a root fraction or root density distribution, the function $R_C(Z)$ describes the capacity of the deep roots to take up groundwater and satisfy part of the total evaporative demand. The function $R_C(Z)$ is related to the root density distribution below Z_r , but cannot be interpreted to be identical to a root density distribution.

[22] For vegetation that maximizes groundwater uptake (i.e., a phreatophyte), the partitioning of the total evapotranspiration is

$$ET_Z(Z) = R_C(Z) \cdot (1 - e^{-0.45\xi})E_p$$

$$ET_{Z_r}(s) = \min[(1 - R_C(Z)), f_r\beta(s)] \cdot (1 - e^{-0.45\xi})E_p \quad (9)$$

Here $ET_Z(Z)$ is the maximum transpiration that can be provided by the deep roots, which is only limited by the water uptake capacity of the roots below Z_r via $R_C(Z)$. This is further supplemented by $ET_{Z_r}(s)$, the evapotranspiration between $0 - Z_r$, to achieve maximum ET ($ET_p = (1 - e^{-0.45\xi})E_p$, which is governed by the vegetation species and climate). Vegetation maximizing groundwater use might have competitive advantages over shallow rooted trees as groundwater is more reliable [Thorburn and Walker, 1994].

2.1.3. FB Model: Uptake With Root Zone Saturation or Direct Groundwater Uptake Feedback

[23] A limitation of the suggested RF model is that the deep roots in this model act quite independently; and do

Table 1. Overview of Literature Values of Relative Uptake of Groundwater by Different Vegetation Species as Measured Using Isotope Analysis^a

	Relative GW Uptake (%)	GW Depth (m)	Annual Rainfall (mm)	Vegetation Species	Location
Snyder and Williams [2000]	45–100	1–4	330	<i>Salix gooddingi</i>	Arizona, United States
Snyder and Williams [2000]	70–100	1–4	330	<i>Populus fremonti</i>	Arizona, United States
Snyder and Williams [2000]	70–100	1–4	330	<i>Prosopis velutina</i>	Arizona, United States
Cramer et al. [1999]	40	1.6–3	647–811	<i>Casurina Glauca</i>	Queensland, Australia
Cramer et al. [1999]	25	1.6–3	647–811	<i>Eucalyptus Camaldulensis</i>	Queensland, Australia
Zencich et al. [2002]	20–80	2.5	870	<i>Banksia Illicifolia</i> and <i>attenuata</i>	Western Australia
Zencich et al. [2002]	20–40	4	870	<i>Banksia Illicifolia</i> and <i>attenuata</i>	Western Australia
Mensforth et al. [1994]	27	2.8	260	<i>Eucalyptus Camaldulensis</i>	New South Wales, Australia

^aGW, groundwater.

not take into account the water availability in the root zone above Z_r . While this model is interesting as an extreme (limiting) case of groundwater uptake, this approximation is unlikely to reflect the majority of phreatophyte vegetations. Decoupling of the deep root uptake from the root zone saturation leads to an upper root zone that on average remains relatively wet and thus the system would experience increased leaching. Plants would have to invest energy in taking up deep water at the expense of energetically more favorable shallow (root zone) water both in developing roots and overcoming the gravitational gradient (e.g., equation (7)). Additionally, the associated loss of nutrients through increased leaching seems far from optimal in nutrient deficient ecosystems. It is therefore plausible that plants in such ecosystems have adapted to maximize uptake of root zone moisture and of the nutrients dissolved in it [Mensforth et al., 1994; Zencich et al., 2002; O'Grady et al., 2006a, 2006b].

[24] Our third model, feedback (FB), therefore extends the RF model and introduces a feedback with upper root zone saturation. During wetter periods, the vegetation would take up water from the upper root zone and reduce groundwater uptake. Conversely, during dry periods, the uptake from groundwater increases relative to the uptake from the upper root zone [Mensforth et al., 1994; Zencich et al., 2002; Lamontagne et al., 2005] (Figure 1).

[25] An appropriate alternative model to model RF (equation (9)) then becomes

$$\begin{aligned} ET_{ZR}(s) &= f_r \beta(s) (1 - e^{-0.45\epsilon}) E_p \\ ET_Z(Z, s) &= \min[(1 - f_r \beta(s)), R_C(Z)] \cdot (1 - e^{-0.45\epsilon}) E_p \end{aligned} \quad (10)$$

In this case, ET_Z is now also regulated by the saturation in the shallow root zone (above Z_r). If this zone is drier, then $(1 - f_r \beta(s))$ will be greater and thus more water will be taken up by the tap roots, $R_C(Z)$ permitting.

2.2. Model Parameterization

[26] We defined generic trees using $Z_r = 100$ cm, $\Psi_{s,s^*} = -0.12$ MPa and $\Psi_{s,sw} = -5$ MPa to represent *Eucalyptus* spp. as a common phreatophyte vegetation species in Australia. A representative leaf area index (LAI) of 2.5 was used, and rainfall interception Δ is 0.25 cm on the basis of LAI [Slavich et al., 1999; Whitehead and Beadle, 2004]. Rainfall input was generated assuming a Poisson distribution of rainfall events and exponentially distributed rainfall depths [Rodriguez-Iturbe et al., 1984] based on representative

semiarid climate types from Australia. Using data from the Bureau of Meteorology for climates ranging from the semiarid Mediterranean areas in northwestern NSW to more tropical in the Northern Territory, the average occurrence of rain storms, λ , ranged between 0.14–0.22 d^{-1} and the average storm depth α ranged between 0.74–1.47 cm. Related potential evaporation (E_p) ranged from 0.5 to 0.8 $cm d^{-1}$. We only report the results for one soil type in this paper, a sandy clay loam. We tested for more soil types, but most of the results changed in a predictable way. The hydraulic properties of the soil were derived using standard Australian soils in “Neurotheta” [Minasny and McBratney, 2002], resulting in $\phi = 0.37$, $K_s = 52.8$ $cm d^{-1}$, $b = 6.41$, the bubbling or air entry pressure, $\Psi_s = -1.2 \times 10^{-3}$ MPa, and the field capacity, $s_{fc} = 0.72$. Simulations generally covered 10,000 days and the first 365 days of the simulations were discarded as a warm-up period.

[27] Probability density functions (pdf's), means and variances of s , ET from groundwater (ET_Z) and root zone (ET_{ZR}) were calculated numerically from the water balance time series. Relative ET_Z was calculated by dividing ET_Z by the total actual evapotranspiration ($ET_Z + ET_{ZR}$).

[28] One way to estimate the parameter c in the function $R_C(Z)$ is to do (expensive) field studies, as there are very few studies for groundwater-dependent vegetation which partition vegetation water use and provide sufficient hydrological data. On the basis of our review of literature we identified a few studies which indicate relative uptake of groundwater by groundwater-dependent or riparian tree systems, mostly on the basis of isotopic data (Table 1). We further used a suggested range of literature values for the parameter c in equation (1) as a starting point (0.26 to 1.09 m^{-1} for *Eucalyptus* spp. [Knight, 1999] to 2.17 m^{-1} for Mediterranean vegetation [Feddes et al., 2001]). As a limited sensitivity analysis, models RF and FB were used to simulate ET_Z for a range of c values and the results will be compared to the range of relative groundwater uptake values on the basis of isotope data reported in the literature (Table 1).

3. Results and Discussion

3.1. Different Loss Functions and the Role of the c Parameter

[29] We first investigated the nonlinear effect of parameter c , and the groundwater depth (Z) on relative ET_Z (i.e., $ET_Z/(ET_Z + ET_{ZR})$, where $(ET_Z + ET_{ZR})$ is the total actual ET). The effect of c is complex as it determines not only R_C ,

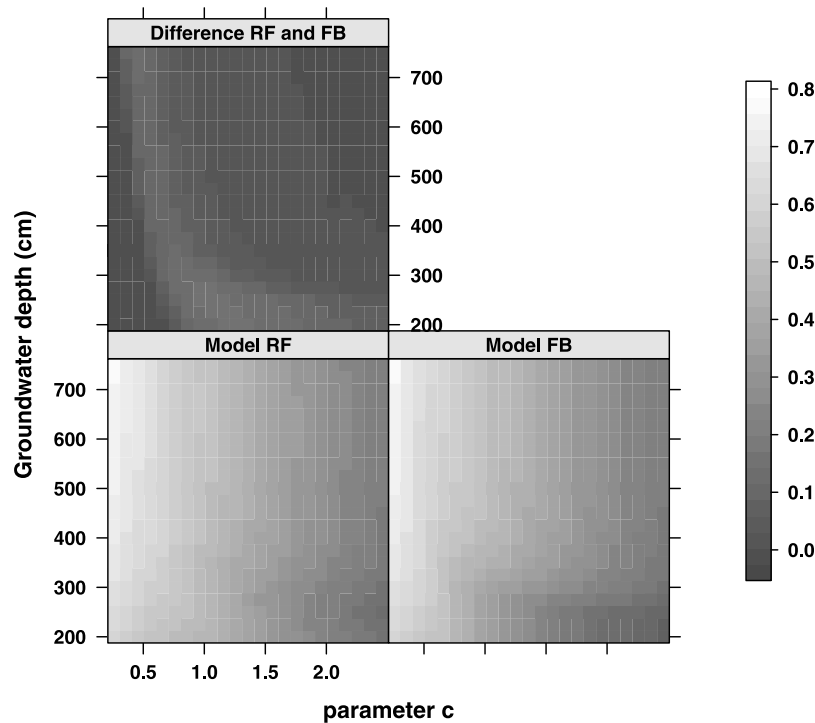


Figure 2. Change in relative ET_Z ($ET_Z / (ET_Z + ET_{ZR})$) for model RF and model FB as a function of the c parameter and the groundwater depth Z (i.e., equations (9) and (10)).

but also the factor f_r in equation (3). Smaller values of c imply larger root densities below the root zone. Therefore relative ET_Z increases for both models (RF and FB) with a decrease in c (Figure 2). As Z increases relative ET_Z changes only little. This is because, while equation (7) includes a linear decrease in ET_Z with increasing Z , ET_{ZR} also decreases with increasing Z because of the decrease in the capillary flux, q_{cap} . At shallow groundwater tables (smaller Z) the feedback mechanism in model FB will reduce ET_Z as the root zone is relatively wet (because of increased q_{cap}). As a result, relative ET_Z actually increases with increasing Z for low values of c , because in this case ET_Z will dominate total ET if the Z increases.

[30] We checked this process further by switching off the capillary flux. Shallow groundwater tables in the absence of capillary fluxes leads to a much drier shallow root zone [see also Vervoort and Van der Zee, 2008], and thus there is little difference between model FB and model RF. Furthermore, at $Z = 1000$ cm, where capillary fluxes do not play a role, ET is only governed by the groundwater uptake capacity of the deep roots ($R_C(Z)$). In this case there is little difference between the models (both RF and FB) with and without capillary fluxes. The implication is that the feedback in model FB is particularly important in the presence of capillary fluxes.

[31] In model RF, ET_Z is constant for each combination of c and Z , but increases with increasing Z and decreasing c . Changes in relative ET_Z are more complex. If c is small, ET_Z dominates total ET and ET_{ZR} decreases with Z because of the reduction in q_{cap} . As a result relative ET_Z increases with Z , because roots in $R(z)$ below Z are inactive. As a result, with greater Z , $R(z)$ and $R_C(Z)$ increase as more and more roots become active. In contrast, at large values of c ,

ET_{ZR} dominates and decreases with Z because of the decrease in q_{cap} . As a result relative ET_Z actually increases for $200 \text{ cm} < Z < 400 \text{ cm}$ until greater Z beyond 400 cm reduce ET_Z through equation (7) (Figure 2).

[32] If c gets very small (i.e., close to the lowest literature value of 0.26 m^{-1}) we would expect almost no effect of the groundwater depth in model RF as the roots are now relatively evenly distributed between the upper and lower root zone, however because of the fact that roots below Z are inactive this is not the case (Figure 2). The greatest feedback effects occur with relatively small Z (high q_{cap}) and with relatively small c (as in this case in model RF the roots would “oversupply”). Hence the differences between models RF and FB in terms of relative ET_Z are greatest at smaller Z for all $c > 0.5$ and for all Z at $c \sim 0.5 - 1$ (Figure 2).

[33] The literature values for relative ET_Z in Table 1 range from 20% to more than 100%. On the basis of Figure 2, the upper value can only be reached with very small values of c indicating very high root densities or water uptake capacities by the deep roots [Dawson and Pate, 1996] or very shallow water tables as indicated in Table 1. The general range of values in Table 1 is, however, well covered by the range of values for c suggested in the literature [Knight, 1999; Feddes et al., 2001].

[34] The loss function for the models, which captures the moisture loss from the soil above Z_r (i.e., $ET_{ZR} + q$, where q is the difference between the drainage losses and the capillary fluxes, q_{cap}) indicates some distinct differences (Figure 3). Model E predicts a decrease in the overall losses at shallow water tables (i.e., in general less than 3 m from the bottom of the root zone) due to capillary fluxes [Vervoort and Van der Zee, 2008]. This effect becomes rapidly smaller as Z becomes greater than 2 m (Figure 3). The difference

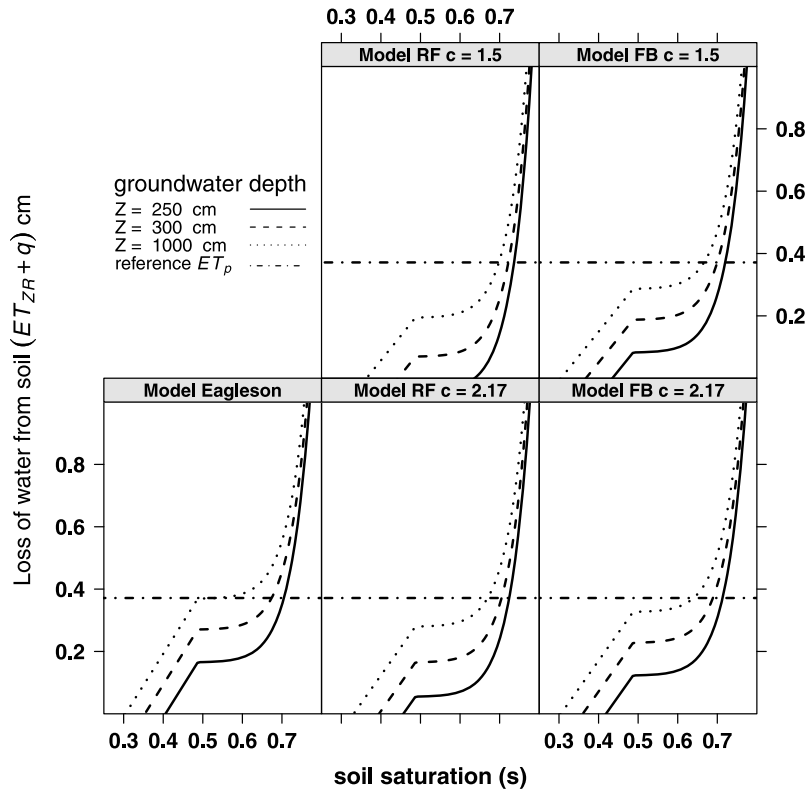


Figure 3. Functions describing the loss of water from the root zone above Z_r for the different models for a sandy clay loam soil. The loss functions are calculated for three different groundwater depths (curves) and, in the case of model RF and model FB, for two different c values: (bottom) $c = 2.17 \text{ m}^{-1}$ and (top) $c = 1.5 \text{ m}^{-1}$. A reference line at ET_p has been inserted to improve comparison of the different graphs.

between model E and the other two models is the greater loss of water from the soil above Z_r in model E, at any groundwater depth. In model RF and FB part of the overall ET is supplied by the deep roots through ET_Z , and thus ET_{ZR} is always smaller than ET_p .

[35] The difference between model RF and FB is more subtle. It is mainly visible at the shallower groundwater depth and smaller c parameter (1.5 m^{-1}), i.e., when the root water uptake capacity ($R_C(Z)$) below Z_r is greater (Figure 3, top). Model RF loses less water from the root zone because more of the evaporative demand is met by the constant groundwater uptake than in model FB, as there is no feedback between the deep root uptake and the soil saturation above Z_r . In contrast, model FB includes a feedback with the saturation above Z_r which reduces the groundwater uptake on the wet end of the loss curve (equation (10)). This means the loss function of Model FB generally lies in between model E and model RF as the water is preferentially taken up from the root zone above Z_r , if sufficient water is available.

3.2. Differences in the Evapotranspiration and Soil Saturation Distributions

[36] We first compare model Eagleson (E) and model RF and comment on the differences in behavior before discussing the need to include the additional complexity of the root zone moisture feedback in model FB. We focus on the numerically derived distributions of the relative distribution of ET from groundwater (ET_Z). We will later discuss the capillary fluxes (q_{cap}) for the three different models.

3.2.1. Differences Between Model E and Model RF

[37] The average root zone saturation for model E and RF for the intermediate climate ($\lambda = 0.21 \text{ d}^{-1}$ and $\alpha = 0.89 \text{ cm}$) indicates that groundwater uptake by deep roots increases the average saturation of the root zone above Z_r (Table 2). In model RF, the direct water uptake from groundwater only depends on c , Z and Z_r (equations (7) and (9)). In this case, we vary Z and the resulting change in ET_Z implies that the overall loss of soil water in the root zone (ET_{ZR}) changes (Figure 3). The probability density function (pdf) indicates that the overall ET (defined as $ET_Z + ET_{ZR}$) for model RF is always greater than for model E (Figure 4), as model E lacks the supply from deep groundwater during dry periods, which means excursions of s below s^* occur more often. Even for quite deep groundwater tables ($Z = 1000 \text{ cm}$) there is still a significant difference in the pdf of total ET between the two models (Figure 4), based on a two-sided Kolmogorov-Smirnov (KS) test [Massey, 1951].

[38] The difference in average root zone (above Z_r) saturation between model E and model RF increases with increasing rainfall occurrence (λ) and depth (α) in the climate (Table 2). As the direct uptake of groundwater by deep roots in model RF is not regulated by the root zone saturation, the increase in soil moisture input from the climate only affects ET_{ZR} and results in an increase in average s (compared with no direct groundwater uptake). In dry climates, ET_{ZR} will be reduced as the deep roots will be unable to supply sufficient water to the vegetation to reach ET_p . Wetter conditions also reduce the number of dry periods involving

Table 2. Means and Standard Deviations of the Relative ET_Z ($ET_Z/(ET_Z + ET_{ZR})^a$)

	Mean s			Mean Relative ET_Z		
	Model E	Model RF	Model FB	Model E	Model RF	Model FB
<i>Depth of Groundwater, Z</i>						
250 cm	0.48 (4.9)	0.65 (3.3)	0.59 (5.5)	0	0.34 (0.0)	0.21 (1.24)
300 cm	0.41 (3.6)	0.47 (5.8)	0.45 (4.5)	0	0.36 (4.6)	0.34 (8.1)
500 cm	0.38 (3.5)	0.41 (4.2)	0.40 (4.0)	0	0.43 (8.9)	0.42 (9.5)
1000 cm	0.36 (3.5)	0.38 (3.9)	0.38 (3.9)	0	0.46 (12.3)	0.46 (12.4)
<i>Climate, Z = 350 cm</i>						
$\alpha\lambda = 0.11 \text{ cm d}^{-1}$	0.37 (2.4)	0.39 (2.7)	0.39 (2.7)	0	0.47 (7.4)	0.47 (7.5)
$\alpha\lambda = 0.2 \text{ cm d}^{-1}$	0.40 (4.4)	0.44 (6.1)	0.43 (5.5)	0	0.39 (7.5)	0.38 (9.3)
$\alpha\lambda = 0.375 \text{ cm d}^{-1}$	0.49 (8.3)	0.59 (8.5)	0.56 (8.9)	0	0.31 (2.1)	0.25 (5.7)

^aValues were derived from a 10,000 day time series simulation for the different models using $c = 1.5 \text{ m}^{-1}$ and $\alpha\lambda = 0.2 \text{ cm d}^{-1}$. E, Eagleson; RF, root fraction; FB, feedback. $Z = 350 \text{ cm}$ was used unless otherwise specified. Statistical differences (parentheses, $\times 10^{-2}$) were analyzed using a log transformation of the data. Mean s and ET_Z are significantly different between models across all Z and by Z . Across all climates, mean s is significantly different between all models and by climate type. Mean ET_Z is different between models by climate.

reduced ET_{ZR} and evapotranspiration smaller than the ET_p (i.e., periods with $s < s^*$). As a result relative ET_Z decreases (Table 2) with wetter conditions toward a minimum ET_Z/ET_p . This is also indicated by the pdf's of ET (Figure 5), which indicates that for model RF ET approaches ET_p for $\alpha\lambda = 0.375 \text{ cm d}^{-1}$, while for model E there is always more spreading of the pdf (Figure 5). While the difference in mean root zone saturations between the models seems small (Table 2), mean s are significantly different ($p < 0.001$) across all climates and groundwater depths between the models. Because of the influence of the deep roots there are also significant differences ($p < 0.001$ based on a two-sided KS test) in the total ET pdf's between the models (Figure 5).

3.2.2. Differences Between Model RF and FB

[39] The main difference in the structure of model RF and model FB is that in model RF ET_{ZR} is dependent on ET_Z , while in model FB this is reversed. As a result, the difference in the average root zone s between model RF and FB is only small at deeper groundwater tables because in this case s would be predominantly smaller than s^* (Table 2). As a result, there is little feedback and the overall ET between the models is quite similar at those depths. However, at shallow groundwater tables average root zone s for model RF is much higher than for model FB because of the lack of feedback in model RF. Under drier climate conditions (small $\alpha\lambda$) the mean s above Z_r for the two models converge (Table 2),

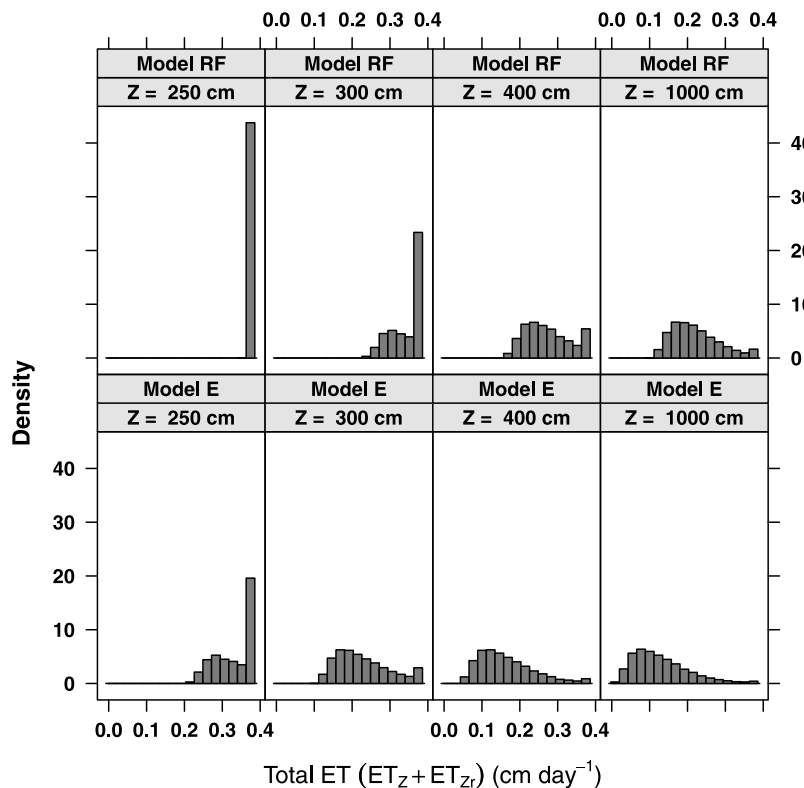


Figure 4. Numerically derived probability density functions (pdf's) of ET_Z for model E and model RF with different groundwater depths (Z). Pdf's are based on 10,000 day water balance simulations using $\alpha\lambda = 0.2 \text{ cm d}^{-1}$ and $c = 1.5 \text{ m}^{-1}$.

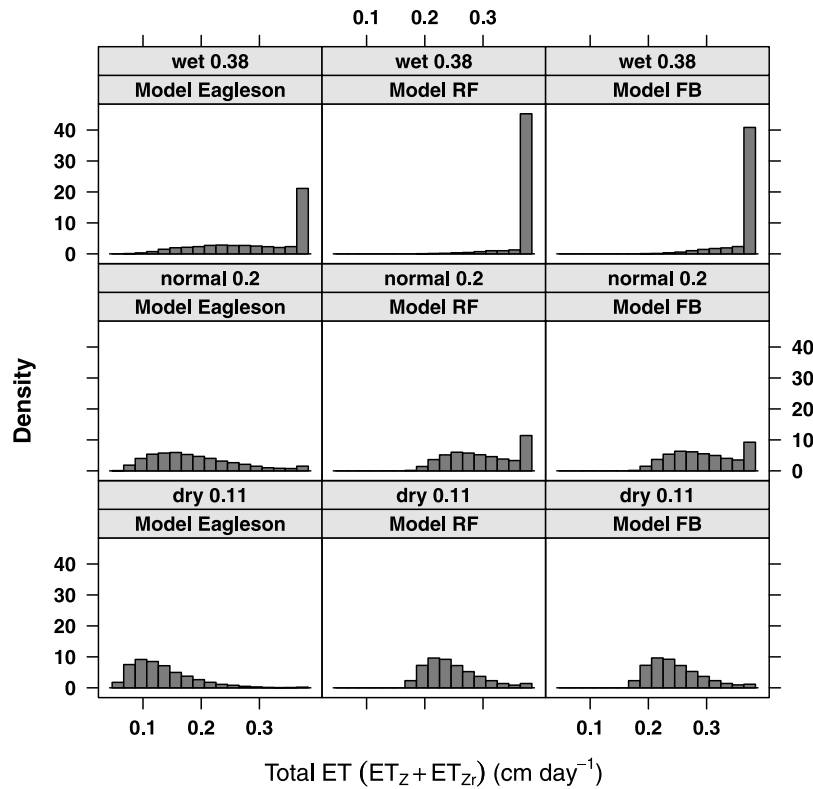


Figure 5. Numerically derived pdf's of the total ET under different climates (in cm d^{-1}) based on 10,000 day water balance simulations for the three models using $c = 1.5 \text{ m}^{-1}$ and $Z = 350 \text{ cm}$.

as s would be predominantly $< s^*$. In this case the feedback between ET_Z and ET_{ZR} in equation (10) would be unlikely to occur. In contrast, relative ET_Z is smaller for model FB than for model RF under wetter climates and thus the pdf for total ET is mid way between model E and model RF (Table 2 and Figure 5).

[40] In general, relative ET_Z will range from ET_Z/ET_p in a wet climate end to 1 in a very dry climate for model RF, while the wet end boundary for model FB becomes $(ET_p - ET_{ZR})/ET_p$; that is, the difference between RF and FB at the wet end is mainly dependent on c . However, in most cases relative ET_Z will be in between these two values (Figure 6). The difference between the two models becomes smaller with greater Z and in drier climates (Figure 6). At very shallow groundwater tables the difference between the models is further complicated by capillary fluxes. This results in a reduction in relative ET_Z in model FB because of increased ET_{ZR} and feedback (Figure 6 and equation (10)). In model RF, relative ET_Z also decreases because ET_{ZR} increases but there is no feedback so the effect is not as strong.

[41] The differences between the models imply that for particular combinations of climate and groundwater ET_Z is significant and is about equal for model RF and FB (Figure 6). In drier climates ($\lambda < 0.2 \text{ d}^{-1}$) and at $Z > 300 \text{ cm}$, model RF and FB are similar and ET_Z/ET_p is the highest. For very deep groundwater levels the relative ET_Z is reduced because of the lower root density at depth and thus, relatively speaking, ET_{ZR} increases for both model RF and model FB (this is not shown in Figure 6). If the climate gets wetter ($\lambda > 0.2 \text{ d}^{-1}$), model FB has lower relative ET_Z compared to model RF as the root zone can supply more

water for ET_{ZR} . In addition, for model RF, relative ET_Z decreases with Z at the wet climate end as ET_{ZR} is relatively constant. For shallow groundwater depths ($< 300 \text{ cm}$) the difference between model RF and FB is clearly evident. In model RF, relative ET_Z decreases with groundwater depth and this is independent of the changes in λ . In contrast, in model FB, relative ET_Z decreases with increasing λ and increases with increasing groundwater depth, because of the built-in feedback between s and R_C (equation (10)).

[42] The isotopic data from field studies (Table 1) also indicate that a range of relative ET_Z values exists bounded by ET_Z/ET_p and 1. In fact, the data suggests that the vegetation studied in Arizona [Snyder and Williams, 2000] under shallower groundwater tables reach the upper limit, which would also agree with the arid climate in that region (330 mm of annual rainfall). The results from two of the vegetation studies in Australia [Cramer et al., 1999; Zencich et al., 2002] indicate a lower range of relative ET_Z probably due to the deeper groundwater tables and higher rainfall (Table 1), confirming the theoretical results in this study. The only exception is the study on the Chowilla floodplain [Mensforth et al., 1994], which indicates low rainfall but also low relative ET_Z . However, this discrepancy might be explained by the reported high salinity of the groundwater at this location [Mensforth et al., 1994]. This could have caused a reduced groundwater uptake compared to our current model, which does not consider such salinity effects, but this is part of our ongoing research [Shah et al., 2009]. Another explanation is that the parameter c is much greater than 1.5 m^{-1} used in the simulations (i.e., a much higher root density above Z_r and only very few deep roots

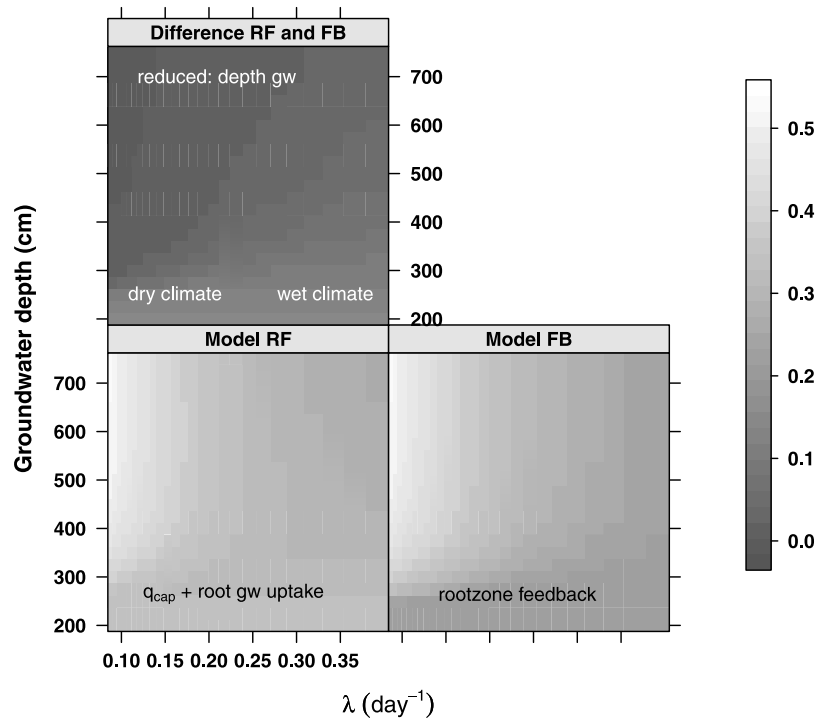


Figure 6. Variation in the average relative ET_Z for models RF and FB over a 5000 day water balance simulation for a range of climates (characterized by λ in d^{-1}) and groundwater depth using $c = 1.5 m^{-1}$.

or a limited water uptake capacity of the deep roots). Direct assessment of sap flow and transpiration, supported with a close monitoring of root zone water saturation is therefore an important future research area.

[43] As a result, end-member model E (including only drainage and capillary fluxes) will suffice to model vegetation regardless of the groundwater depth if deep roots are not considered. On the other end of the spectrum, in very dry climates ($\lambda < 0.2 d^{-1}$ and $\alpha < 1 cm$), ET_Z dominates ET as the average $s < s^*$. In this case there is very little difference between end-member model RF and model FB as s is never large enough to limit the groundwater uptake through feedback. However, if the climate is not very arid, model RF tends to be somewhat simplistic. Because a feedback with s is included model FB gives important differences over model RF in climates which have $\lambda > 0.2 d^{-1}$ and $\alpha > 1 cm$. In this case s influences the relative ET_{ZR} , which means model FB takes an intermediate position between the end-member models E and RF and implicitly takes into account hydraulic redistribution processes [Burgess *et al.*, 2001; Ryel *et al.*, 2002].

3.3. General Discussion

3.3.1. Effect of ET_Z on Capillary Fluxes

[44] The differences in s (Table 2) also result in differences in q_{cap} between the models (Figure 7). In this case the pdf's of the relative capillary flux ($q_{cap}/(ET_{ZR} + ET_Z)$) indicate that for model E this is a larger contribution than for model RF and FB. Model RF and model FB have higher s under the same climate (Table 2) and as a result lower q_{cap}/ET . In addition, in model RF and FB, the overall ET is also determined by ET_Z . In both models, the relative contribution of ET_Z and $q_{cap}/(ET_{ZR} + ET_Z)$ would increase with increasing

dryness, and more so in model FB than in model RF, with the overall contribution of the q_{cap} therefore always remaining lower than for model E.

3.3.2. Role of Compensation in the Root Zone Above Z_r and the Location of s_{cr}

[45] In our model, we have considered full compensation in the upper root zone. Complete compensation has two major effects on the daily ET from the root zone [Guswa, 2005]: The ET values reach E_p too often and their decay is too rapid. Basically, the lack of vertical discretization in the zero-dimensional models makes the fluctuations in ET too coarse and results in a relatively wetter root zone [Teuling *et al.*, 2006]. In our case, these limitations would be particularly true for the deep groundwater and arid climate scenarios. Thus, the pdf's presented in Figure 5 and for $Z = 1000 cm$ (in Figure 4) would shift to lower total ET (left) as the number of days at high ET_{ZR} is overestimated [Guswa, 2005]. In the case of model FB, this means that the impact of the feedback is possibly overestimated, but this impact is only small in dry climates and for deep water tables.

[46] For shallower groundwater tables, the contribution of ET_Z in model RF and q_{cap} in all three models reduces the impact of ignoring the root density distribution above Z_r . The impact of the climate inputs on s becomes smaller as the influence of q_{cap} and ET_Z increases. Guswa [2005, p. 548, equation (13)] indicates how the spatial infiltration index I_Z , which is the ratio between the average depth of rainfall to the available soil storage, can be used to estimate whether the root distribution can be ignored. In the presence of capillary fluxes, the spatial infiltration index changes to

$$I_Z = \frac{\alpha + q_{cap}}{(s - s_w)\phi Z_r} \quad (11)$$

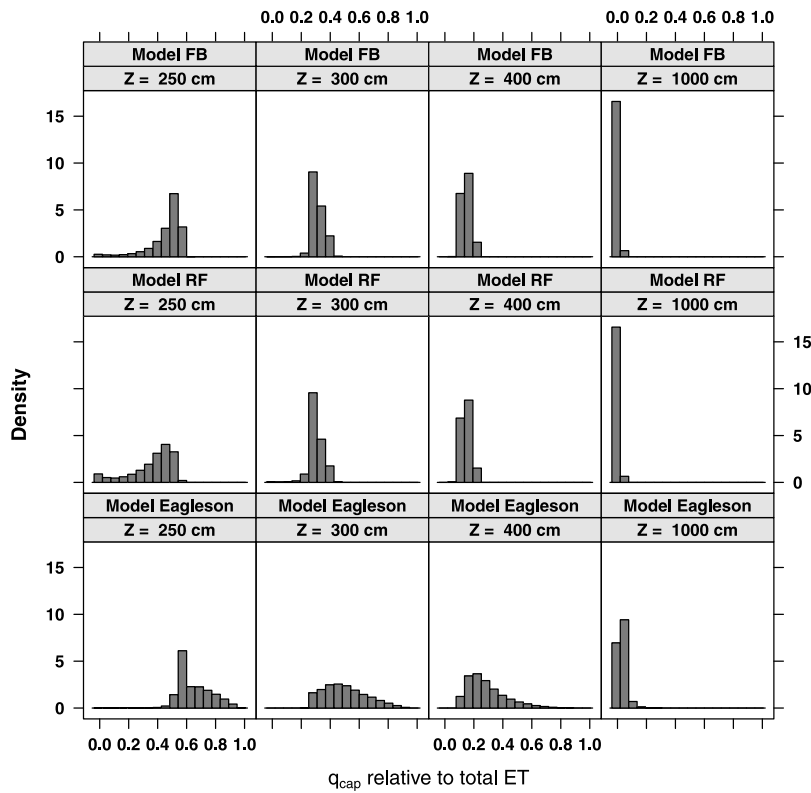


Figure 7. Numerically derived pdf's of the relative capillary flux ($q_{cap}/(ET_{ZR} + ET_Z)$) for the different models. Pdf's were based on 10,000 days of water balance simulations using $\alpha\lambda = 0.2 \text{ cm d}^{-1}$ and $c = 1.5 \text{ m}^{-1}$.

indicating that increasing q_{cap} increases I_Z and thus reduces the need for a root distribution in the model. Another way of looking at this is that q_{cap}/α increases with increasing q_{cap} and thus reduces the relative contribution of the rainfall to s . In model RF this is further strengthened by the fact that at shallower water tables the contribution of ET_{ZR} to total ET decreases. As a result the effect of ignoring a root distribution in the upper root zone on total ET also decreases.

[47] In our earlier paper [Vervoort and Van der Zee, 2008], we indicated a specific soil saturation s_{cr} at which ET equaled q_{cap} , or where the root zone losses equal zero. In the presence of groundwater uptake by deep roots this point shifts as ET_{ZR} is reduced because of ET_Z . For model FB the recalculation is straight forward as this only involves inclusion of f_r in the earlier equation [Vervoort and Van der Zee, 2008] as ET_{ZR} is independent of ET_Z :

$$s_{cr} = \frac{q_{cap}}{f_r \cdot ET_p} (s^* - s_w) + s_w \quad (12)$$

In contrast, for model RF, the situation is slightly more complex, as we need to consider the dependency of ET_{ZR} on ET_Z . This means there is a minimum groundwater level for each value of c above which s_{cr} is above s^* and therefore $\beta(s)$ and $R_C(Z)$ and thus ET_Z/ET_p equal 1. The minimum value of Z at which this occurs can be found by equating equation (7) to ET_p . For any groundwater levels deeper than this minimum level, s_{cr} can be found through equation (12) and there is no difference between model RF and FB in this respect (Figure 3)

[48] The models we have presented are simplified realizations of complex interactions. However, in view of the feedbacks and the generality of the description, the models presented here could be incorporated into larger climate models after further testing. The models might well improve the climate models as the functional dependencies between deep roots and surface roots are well represented, without the discretization errors of current climate models (for example using four soil layers and consider each continent to be covered by a few soil types). This is a potential area for further research.

4. Conclusions

[49] We have demonstrated the behavior of low-parameter analytical ecohydrological models to study the effects of groundwater uptake by deep roots and capillary fluxes on vegetation ET . In particular, our results indicate that feedback between the upper root zone saturation and the uptake of groundwater by deep roots becomes important in climates which are slightly wetter. We hypothesize that this is probably important ecologically because of the impact on nutrient uptake and leaching of salts. In more arid climates feedback is less important as the upper root zone is never wet enough to experience drainage.

[50] As a result of groundwater uptake by deep roots, the overall water use of the vegetation remains higher than for vegetation that lacks such deep roots once the groundwater levels move beyond the range of significant capillary fluxes ($Z > 300 \text{ cm}$). The relative importance of capillary fluxes

also decreases in the presence of deep roots and more so without feedback as the groundwater uptake of deep roots suppresses ET_{ZR} .

[51] The exact range of possible values for the root density parameter c is an area for further experimental research even though some literature values are available. Using the middle of the range of literature values ($c = 1.5 \text{ m}^{-1}$) the models predicted that between 20 and 50% of the total ET would be contributed by groundwater for a range of climates. Literature data for relative groundwater uptake based on isotopic studies ranged from 20–100%, but the very high values were based on vegetation over very shallow water tables in arid climates and probably lump capillary fluxes and deep root groundwater uptake. Our models also predicted the highest groundwater contributions for such situations.

[52] **Acknowledgments.** This work was initiated while the first author was on sabbatical leave from the University of Sydney and working at the Soil Physics, Ecohydrology and Groundwater Management Group at Wageningen University (Netherlands) funded by the Wageningen Institute for Environment and Climate Research (WIMEK). The second author acknowledges support of the International Research Training Network NUPUS, funded by the German Research Foundation (DFG) (GRK 1398) and Netherlands Organisation for Scientific Research (NOW) (DN 81–754).

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