DYNAMIC MODELING OF ROOT WATER UPTAKE USING SOIL MOISTURE DATA

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ABSTRACT

In this work a dynamic model of water uptake from plants growing in naturally vegetated areas subjected to a rainfall and evaporation time series is described. The model results are compared and contrasted with popular pre-existing models. Also, the effects of the uptake pattern on the movement of water across multiple soil layers are also analyzed. The results showed that contrary to common modeling approaches, root water uptake is both a function of root distribution and variability in water content.

Keywords - root water uptake, vadose zone, soil moisture data

1. INTRODUCTION

Over the past two centuries rapid increase in human population coupled with associated water resources development activities has resulted in severe degradation of ecosystems on a global scale (Zalewiski, 2000). Several studies have shown that the mechanisms of interaction of the biota with their surroundings contribute to their spatiotemporal patterns (Rodriguez-Iturbe and Porporato, 2004). Hence, knowledge about specie specific interaction with its environment is of utmost importance for successful restoration efforts.

Historically, hydrology and ecology have evolved as two distinct sciences with little or no connection with each other (Baird and Wilby, 1999). As an example, for a hydrologist, plants on the river bed may have meant nothing but Mannings' roughness coefficient, while many ecologists have considered soil is as simply a reservoir of water. It is this difference in perspective that has limited our ability to forecast changes, assess impacts and develop mitigation strategies. Traditional relationships used for quantifying hydrological processes, though very useful, are based more on empiricism rather than actual experimental approaches. Estimating evapotranspiration from pan measurements (Doorenbos and Pruitt, 1977), specifying extinction depths based on qualitative rules (Anderson and Woessner, 1991), and estimation of recharge to groundwater as a calibration parameter (e.g. MODFLOW (Harbaugh et al., 2000)) are some of the relationships that have been in use in hydrology primarily because plants physiology is ignored or over simplified. Recent studies such as that of Shah et al. (2007) and Nachabe et al. (2005) have shown that hydrologic processes of evapotranspiration and recharge, for example are strongly a function of the type and condition of vegetative cover and climate. Ignoring the land cover effects can hence lead to erroneous estimate of these fluxes.

To cater to this need, interdisciplinary work in ecology and hydrology has been initiated. Zalewiski et al. (1997), Rodriguez-Iturbe and Porporato (2004) have shown promising results from seminal research in this new area called 'Eco-hydrology' thereby increasing confidence in the use of ecohydrological framework for understanding species dynamics. Despite the recent progress, our knowledge about species interaction, especially that of plants in ecotones and response of an ecosystem to the change in ambient conditions remains limited.

An important gap that remains in the eco-hydrological framework is the ability to successfully simulate the spatial and temporal patterns of root zone soil moisture. Fundamental to the modeling of the soil moisture dynamics in the root zone is the knowledge of the water uptake patterns by roots. Two major class of root water uptake models that are in use are the microscopic scale models (Steudle 2000) where water movement along single root hair is modeled and the other is the macroscopic model where instead of a root hair, a section of roots is considered (e.g. Feddes et al. 1978). The former class of models even though more accurate require much information and hence become infeasible while modeling at watershed scale (~10 km^2). The latter class (watershed scale) of models are mostly empirical, suffer from data uncertainty, and when applied on large scale rarely consider plant physiology and hence often exhibit poor performance dictating low user confidence for assessment, planning or predictive capability. More physically based watershed scale models have a better the capability to simulate moisture conditions in the unsaturated vadose zone incorporating variability in soil, plant and atmospheric conditions. However, empirical conceptualization of root water uptake in these models is still preveland thereby casting doubt on the validity of the model results.

The objective of this paper is thus to (1) discuss the empirical root water uptake models used, (2) to describe a methodology involving field data to calculate root water uptake, (3) use field data to compute root water uptake values, and (4) compare and contrast the model derived estimate from those derived from field data.

2. FORMULATION OF ROOT UPTAKE MODEL

The governing equation for soil moisture dynamics in the unsaturated soil zone is the Richards's equation (Richards 1931). Richards's equation is derived from Darcy's law for flow in porous media and the continuity equation. What follows is a brief description on Richards's equation and how can it incorporates root water uptake. For more detailed information about formulation of Richards's equation, including its derivation in three dimensions, the readers are directed to any text book on soil physics e.g. Hillel (1998).

Due to ease of measurement and conceptualization, energy of water (E) is represented in terms of height of liquid column and is called the hydraulic head (h). It is defined as the total energy (potential energy, kinetic energy, osmotic potential etc.) of water per unit weight. Mathematically hydraulic head, h, can be represented as

$$h = \frac{E}{\rho_W g} \tag{1}$$

where ρ_W is the density of water and g is the acceleration due to gravity. The flow of water always occurs along decreasing head. In soil physics the fundamental equation used to model the flow of water along a head gradient is known as Darcy Law (Hillel 1998). Mathematically the equation can be written as

$$q = K\frac{\Delta h}{l} \tag{2}$$

where q [$L^3L^{-2}T^{-1}$] is known as the specific discharge and is defined as the flow per unit cross-sectional area, K[LT^{-1}] is termed as the hydraulic conductivity, which indicates ease of flow, Δh [L] is the head difference between the points of interest and l[L] is the distance between them. The quantities with square brackets represent the dimensional units wherein L is length, and T is time. Darcy's Law is analogous to Ohm's law with head gradient being analogous to the potential difference and current being analogous to specific discharge and hydraulic conductivity being similar to the conductance of a wire.

The continuity aspect of Richards's equation is based on the law of mass conservation, and for any given volume it states that net increase in storage in the given volume is the difference between inflow and outflow together with any sink present in the volume of soil. Mathematically it is this sink term that allows the modeling of water extracted from the given volume of soil.

In one dimension for flow occurring in the vertical direction (z axis is positive downwards) Richards's equation can be written as

$$\left(\frac{\partial \theta}{\partial t}\right)=\left(\frac{\partial}{\partial z}K\left(\frac{\partial h}{\partial z}+1\right)\right)-S \tag{3}$$

where θ is the water content and it is defined as the ratio of volume of water present with total volume of the soil element , t is time, S represents the sink term while other terms are as defined before.

If the flow is also considered in X and Y directions Richards's equation in three dimensions can be derived, analogously. Solution of the partial differential equation derived above can theoretically provide the spatial and temporal variability of moisture in the soil. Due to high degree of non linearity of the Richards's equation an analytical solution is not feasible; however, numerical techniques are used to find an approximate solution. For a numerical solution of Richards's equation two essential properties that need to be defined a-priori are (a) the relationship between soil water content and hydraulic head, also known as, soil moisture retention curves, and (b) a model that relates hydraulic head to root water uptake. While much of the literature and field data exist describing soil moisture retention curves, relatively sparse information exists about the root water uptake models describing the source or sink function. Root water uptake models, especially, on watershed scale, are mostly empirical and lack field verification. The main reason for this can be attributed to the fact that only recently instrumentation and understanding of basic plant physiology have evolved to a point to improve the hydrological modeling. Details about the soil moisture retention curves and numerical techniques used to solve Richards's equation can be found in Simunek et al. (2005). The focus of this paper will be on the root water uptake models and field data that contradict the existing models.

One common approach used to model root water uptake is to define sink term S (see (3)) as a function of hydraulic head using the following equation

$$S(h) = \alpha(h) S_p \tag{4}$$

where S(h)[$L^3L^{-3}T^{-1}$] is the root water uptake (RWU) from roots subjected to hydraulic or capillary pressure head *h*. On the right hand side of the equation S_p [$L^3L^{-3}T^{-1}$] is the maximum (also known as potential) uptake of water by the roots. The $\alpha(h)$ is a root water uptake stress response function, which varies between 0 and 1.

The conceptualization of (4) is based on three basic assumptions. The first assumption is that as the soil becomes dryer the amount of water that can be extracted decreases proportionally, due to the increase sorption potential of the soil. Secondly, the amount of water extracted by the roots is affected by the ambient climatic conditions. Drier and hotter conditions result in more water loss from the surface of leaves, hence, initiating more water extraction from the soil. The third and the final assumption is that the uptake of water from a particular section of a root is directly proportional to the amount (mass and surface area) of roots present.

A root water stress response function α is the result of the first assumption. Two models commonly used to define α are the Feddes model (Feddes et al. 1978) and the van Genuchten model (van Genuchten 1987). Figure 1 (a and b, respectively) shows that the variation of α with decreasing hydraulic head which also implies decreasing water content or increasing soil dryness. Both models for α are empirical and do not involve any plant physiology to define the thresholds or variability exhibited by the water stress response functions. An interesting contrast, due to empiricism that is clearly evident, can be seen in the value of α during saturated conditions. While the Feddes model predicts the value of α to decrease to zero the van Genuchten model predicts maximum uptake with α increasing to unity under saturated conditions.

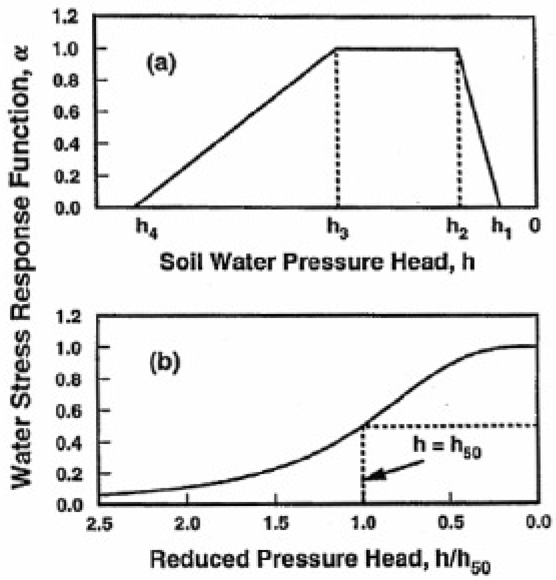


Figure 1. Water stress response functions as conceptualized by (a) Feddes et al. 1978 and (b) van Genuchten (1980) [Adapted from Simunek et al. 2005].

Recently several models (e.g., Li et al., 2001 and Li et al., 2006) have been reported to overcome the empiricism in α of these popular models. However, these models are more a result of observation fitting and yet fail to bring in the plant physiology, which is what causing the changes in the water uptake rate due to the variation in soil moisture conditions.

Combining the second and the third assumptions in (4) results in the definition of layer specific potential (S_p). For any section of roots, S_p, is defined as the product of root fraction in that section and the maximum possible water loss by the plant which is also known as the potential evapotranspiration. Potential evapotranspiration is a function of ambient atmospheric conditions and standard models such as Penman-Monteith (Allen et al. 1998) are used to calculate the potential evapotranspiration rate. For any given value of potential evapotranspiration, rate limiting the value of *Sp* by the fraction of roots restricts the amount of water that can be extracted from a particular section. This as will be shown later using field data is a big limitation especially during dry periods when the top soil with maximum root mass becomes dry while deeper soil layers with much less root mass still have soil moisture available for extraction.

3. MATERIAL AND METHODS

3.1 Study Site

For the current study, field data were obtained from a site located in Hillsborough County, Florida, near the Tampa Bay Regional Reservoir in Lithia. As a part of small scale intensive data collection effort to understand surface and ground water interactions, hydrologic data were collected at different locations surrounding the reservoir. Detailed information about the study site and data collection efforts can be found in Trout and Ross (2005) and Said et al. (2005).

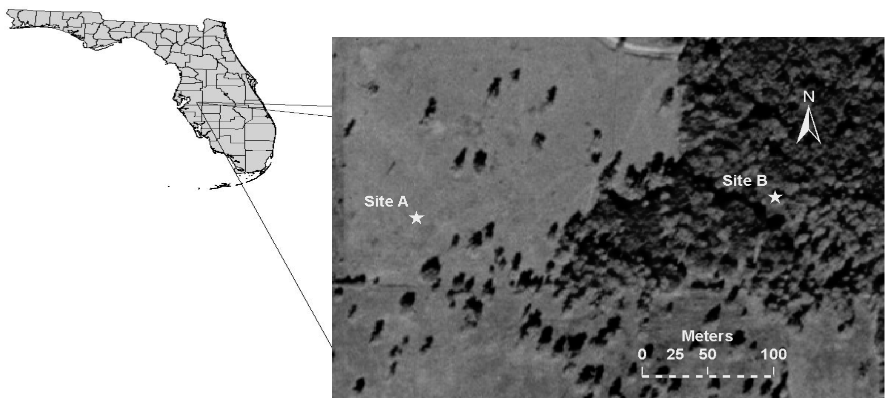


Figure 2.Regional location and aerial image showing the location of the instrumentation and land cover.

For this particular analysis data from two locations, one in a grassland site (site A) and other in a forested land cover site (B), were collected. The study locations were instrumented with a water table observation well and a soil water content probes in proximity to each other. The water table observation well housed a Northwest Inc (Kirkland, WA) submersible pressure transducer. The transducer was calibrated to measure the pressure from 0-34 kPa (0-5 psi) with an accuracy of 0.034 kPa (0.005 psi). To prevent air compression inside the well, the well was vented so that the water surface remained in direct connection with the atmosphere. Soil water content was measured using an EnviroSMART (Sentek Pty. Ltd., Adelaide, Australia) soil moisture probe. The probe was fitted with eight sensors set at 10, 20, 30, 50, 70, 90, 110, 150 cm, respectively, below the land surface to measure soil water content over 10 cm at corresponding levels. The sensors work on the principle of Frequency-Domain Reflectometry and were found to measure volumetric water content ranging from oven dryness to saturation with a resolution of 0.1% [Buss, 1993].The operation and accuracy of these sensors have been tested extensively in the laboratory as well as under field conditions (Starr and Paltineanu, 1998). Fares and Alva (2000) found no significant difference (<0.5%) in volumetric water content as measured by capacitance sensors and gravimetric methods for the fine sands typical in central Florida. The factory supplied equations were used for the calibration of these sensors. Hourly averaged data at four hour time step were used for the analysis in this study.

Extensive soil investigations including in-situ and laboratory analysis were performed for the study site. The soil in the study area is primarily sandy marine sediments (Myakka fine sand) with high permeability in the surface and subsurface layers. Detailed information about soil and site characteristics can be found in Said et al. (2005), and Trout and Ross (2005). Data for period of record, January 2003 to December 2003, were used in this analysis.

3.2 Methodology

Soil matrix has voids which can be filled with water or air. In soil physics the ratio of the volume of voids and total volume of soil matrix is defined as porosity. If all pores (or voids) are filled with water, the soil matrix is termed saturated and the water content in the soil matrix is called saturated water content, represented by θ_s. As the soil dries the water content (θ) reduces below θ_s. As the small pores in the soil matrix do not necessarily make a continuous network not all of the water can be removed from the soil under natural conditions (Hillel 1998). Hence, even under extreme dry conditions, soils do not become completely dry. The minimum water content that remains, for a given soil in particular atmospheric conditions is called the residual water content, and is represented by θ_r.

A common technique used to represent the observed water content is to normalize it using the following equation, hence confining the values between 0 and 1.

$$S_e = \frac{\theta - \theta_r}{\theta_s - \theta_r} \tag{5}$$

Here, S_e is called the normalized water content, varying between 0 and 1. θ is the observed water content, while θ_r and θ_s are the residual and saturated water content values respectively.

An important implication of varying water content which greatly affects the soil moisture dynamics is the fluctuations in the value of hydraulic conductivity of soil. When the soil is saturated all of the pores are well connected and hence the water can flow thorough the soil matrix more easily. However, as the soil dries the paths get blocked due to intermittent air pockets that develop due to evaporation of water from the pores. The net result is that the water carrying capacity of soil is reduced, which is manifested as reduced hydraulic conductivity (Hillel, 1998).

Hence, with increasing soil dryness, which increases soil suction head (negative pressure head), both water content and hydraulic conductivity is reduced. van Genuchten (1980) proposed a model relating the water content and hydraulic conductivity with suction head and is represented by the following equations

$$h(\theta) = \frac{[(S_e^{1/m}) - 1]^{\frac{1}{n}}}{\phi} \tag{6}$$

$$K(h) = \begin{cases} K_S S_e^l [1 - (1 - S_e^{1/m})^m & h < 0 \\ K_S & h \geq 0 \end{cases} \tag{7}$$

where m = 1 – 1/n for n > 1, S_e [-] is the normalized water content, K_S [LT^{-1}] is the hydraulic conductivity when the soil matrix is saturated, l[-] is the pore connectivity parameter assumed to be 0.5 as an average for most soils (Mualem, 1976), and ϕ[L^{-1}], n[-] and m[-] are the van Genuchten empirical parameters. Negative values of hydraulic head means water is able to be held against the action of gravity. For saturated water content in the soil matrix the hydraulic heads shows positive values. From (6) and (7), it is clear that for each type of soil, five parameters, namely, K_S, n,ϕ, θ_r and θ_s have to be determined to uniquely define the relationships of hydraulic conductivity and water content with soil suction head.

Before further discussing how the parameter values were determined, a brief overview of the system being studied is in order. Figure 3 shows the schematic of the vertical soil column which is monitored using eight soil moisture sensors plus one pressure transducer to measure water table elevation at each of the two locations (Site A and B). Shown also in Figure 3 is the zone of influence of each sensor along with the elevation of water table and arrows showing possible flow directions.

For the purpose of defining moisture retention and hydraulic conductivity curves, each section is treated as a differently soil layer and independently parameterized. Hence, for each of the two locations for this particular study eight soil cores from depths corresponding to the zone of influence of each sensor were taken and analyzed using the methods described below.

3.2.1 Saturated and Residual Water Content

Actual water content measurement for all the eight locations was available for each of the two sites, for the two and half years of record (Jan 2002-June 2004), with pronounced wet and dry seasons. Hence, from the observed data the maximum and minimum water content were set up as saturated and residual water content respectively.

3.2.2 Saturated Hydraulic Conductivity

Saturated hydraulic conductivity K_S for different soil layers at the study locations was calculated using falling head permeameter analysis as described in Das (2002). The falling head permeameter test is a standard technique to determine the saturated hydraulic conductivity. Multiple tests were done (replicates) and the results were averaged to determine the most appropriate value of saturated hydraulic conductivity for each of the soil layers at both the study locations.

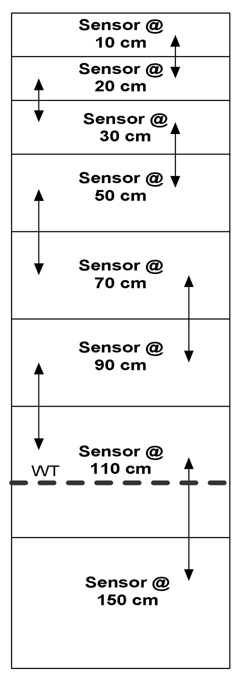


Figure 3. Schematic of the vertical soil column with location of the soil moisture sensors and water table.

3.2.3 van Genuchten Parameters

To determine the values of parameters n, ϕ the soil cores taken were saturated and rotated in a centrifuge. The centrifugal force and corresponding water content values of the soil sample were used to generate the moisture retention curves (as in Carlisle et al., 1989). Moisture retention curve from them measure data was then plotted and fitted with (6) and the best fit values of n, ϕ, were taken as the parameter value for the respective soil layer.

Table 1(a) and (b) shows parameters values that were obtained following the analysis of all soil tests.

Sensor Location Below Land Surface (cm)	θs (%)	θr (%)	Φ (cm^{-1})	n (-)	K_s (cm/hr)
10	35	3	0.03	1.85	4.212
20	35	3	0.07	1.7	2.520
30	32	3	0.07	1.7	2.520
50	34	3	0.03	1.6	0.803
70	31	3	0.03	1.6	0.005
90	32	3	0.05	1.9	0.005
110	32	3	0.05	1.8	0.005
150	30	3	0.05	1.8	0.001

(a)

Sensor Location Below Land Surface (cm)	θs (%)	θr (%)	Φ (cm^{-1})	n (-)	K_s (cm/hr)
10	38	3	0.02	1.35	0.0100
20	34	3	0.03	1.35	0.0100
30	31	3	0.03	1.35	0.0100
50	31	3	0.07	1.9	0.0100
70	31	3	0.20	2.2	0.0100
90	31	3	0.20	2.2	0.0004
110	33	3	0.20	2.2	0.0004
150	35	3	0.20	2.1	0.0012

(b)

Table 1. Soil parameters for study locations in (a) Grassland and (b) Forested area.

3.2.4 Calculation of Root Water Uptake

Once the soil parameterization was completed, the root water uptake from each section can be calculated. For any given soil layer in the vertical soil column (Figure 3), above the observed water table, observed water content and (6) can be used to calculate the hydraulic head. For soil layers below the water table, hydraulic head is the same as the depth of soil below the water table due to assumption of hydrostatic pressure. Similarly using (7) hydraulic conductivity can be calculated. Hence, at any instant in time hydraulic head in each of the eight soil layers can be calculated. To determine total head, gravity head which is the height of the soil layer above a common datum, has to be added to the hydraulic head. For this particular study the datum was arbitrarily selected as 2000 cm below the land surface. As water flows in a direction of decreasing head, observing total head values of the adjacent layers yields the direction of water flow for a given soil layer.

To quantify flow across each soil layer, Darcy's Law for porous flow (2) is used. Average head values between two consecutive time steps are used to determine the head difference. Also, head is assumed to be at the midpoints of each layer. Hence, to determine the head gradient ($\Delta h/l$), the distance between the midpoints of each soil layer is used. The last component needed to solve Darcy's Law is the value of hydraulic conductivity. For flow occurring between layers of different hydraulic conductivities equivalent hydraulic conductivity is calculated by taking harmonic means of the hydraulic conductivities of both layers (Freeze and Cherry 1979). Hence, for each time step harmonically (8) averaged hydraulic conductivity values were used to calculate the flow across soil layers.

$$K_{eq} = \frac{2K_1K_2}{K_1 + K_2} \tag{8}$$

Where K_1 [LT^{-1}]and K_2 [LT^{-1}]are the two hydraulic conductivity values for any two adjacent soil layers and K_{eq} [LT^{-1}]is the equivalent hydraulic conductivity for flow occurring between those two layers.

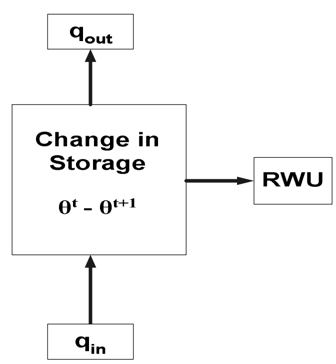


Figure 4. Schematic of a section of vertical soil column showing fluxes and change in storage.

Figure 4 shows a typical flow layer with inflow and outflow marked. Now by measuring mass balance changes via changes in water content between two consecutive time steps and considering vertical flow, net change yields the root water uptake (assuming no other sink is present). Equation 9 can hence be used to determine root water uptake from any given soil layer as

$$RWU = (\theta^t - \theta^{t+1}) - (q_{out} - q_{in}) \quad (9)$$

Using the described methodology the root water uptake from each soil layer throughout the study period was determined at both study locations (sites A and B).

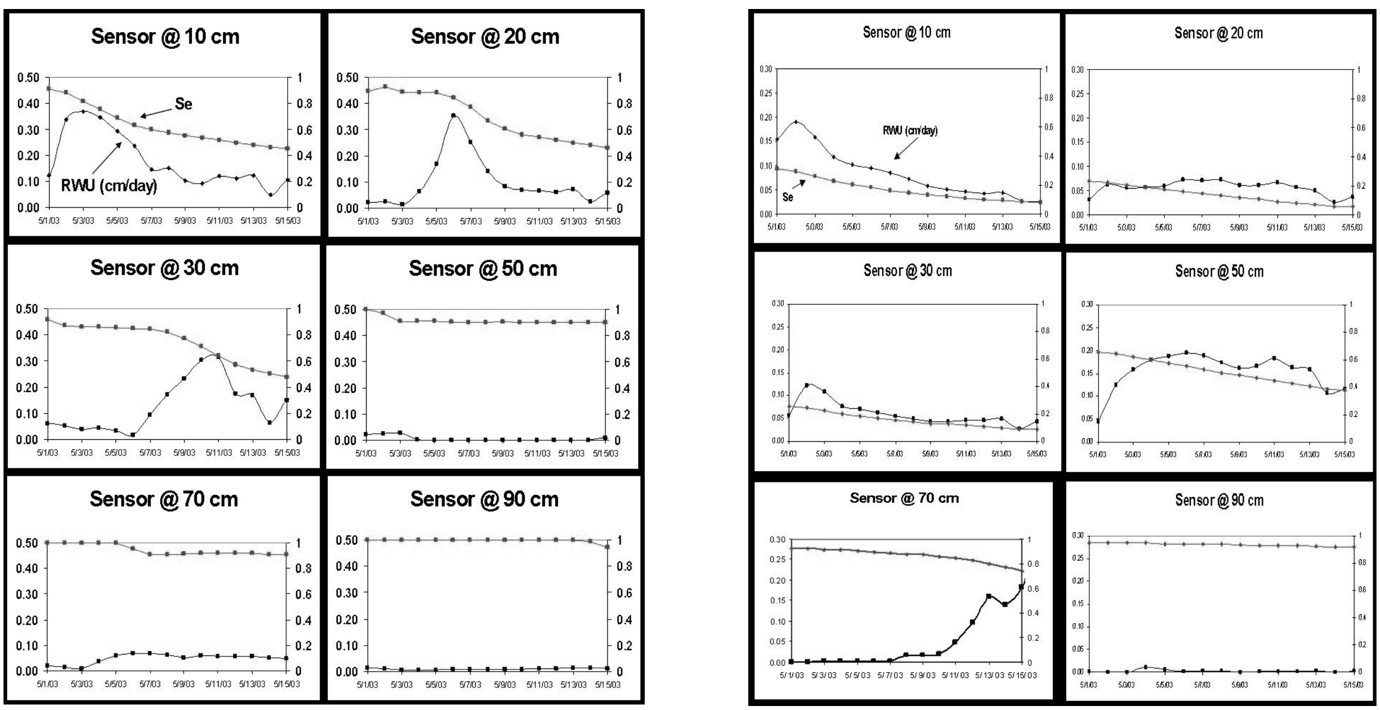
The time step for calculation of the root water uptake was set as four hours (i.e. every forth hourly measurement of total moisture was used) and the root water uptake values obtained were summed to get a daily value for each soil layer. The results section, which follows, describes the findings of the study.

4. RESULTS and DISCUSSION

Using the above methodology, root water uptake was calculated from each section of roots for tree and grass land cover from January to December 2003 at a daily time step. Figure 5 (a and b) shows the variation of root water uptake for a representative period from May 1st to May 15th 2003, This particular period was selected as conditions were dry with no rainfall observed. Figure 5(a) and (b) show root water uptake variation from corresponding sections. Also plotted on the graphs is the normalized water content, which also gives an indication, of water lost from the section.

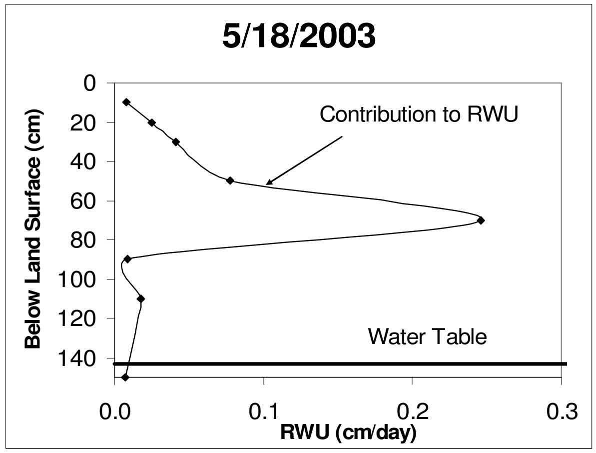
Figure 5(a) shows root water uptake from the grassed site while Figure 5(b) plots RWU from the forested area. From figure 5(a and b) it can be seen that in both cases grass and forest, the root water uptake varies with water content starting from the top layers, then, progressively shifting towards the lower layer so as to maintain constant root water uptake. This suggests that plant compensation takes place (plants are able to derive full potential from a fraction of the roots) contrary to the widely used empirical models. Another important point to note is that in Figure 5(a) root water uptake from top three sensors accounts for almost the entire water uptake while in Figure 5(b) the contribution from fourth and fifth sensor is also significant. Also, as will be shown later, in case of forested land cover, root water uptake is observed from the sections that are even deeper than 70 cm below land surface. This is expected owing to the differences in the root system of both land cover types. While grasses have shallow roots, trees tend to have deeper root zones to meet higher water consumptive use and overcome dryer and deeper water table conditions.

Figure 6, shows an interesting scenario when a rainfall event occurs right after a long dry stretch that exhibits dry upper soil layers. Figure 6(a) shows the root water uptake profile on 5/18/2003 for forested land cover with maximum water being taken from a section of soil profile corresponding to 70 cm below the land surface. A rainfall event of 2.5 cm (1inch) took place on 5/19/2003. As can be clearly seen in Figure 6(b) the maximum water uptake rate shifts right back up to the top 10 cm of soil, clearly showing that the ambient water content directly and instantaneously affects the root water uptake distribution. Figure 6(c) shows a snapshot on 5/20/2003, a day after the rainfall where the root water uptake starts redistributing and shifting toward deeper wetter layers. In fact this kind of behavior was observed for all the data analyzed for the period of record for both the grassed and forested land cover. In this manner roots take water from deeper wetter layers until the shallower layer is re-wetted shifting uptake back to the top layers.

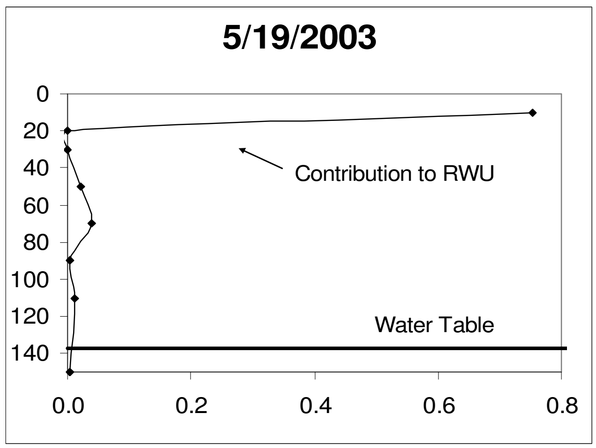


(a) (b)

Figure 5. Root water uptake from sections of soil corresponding to each soil moisture sensor for (a) Grassed and (b) Forested land cover



(a)



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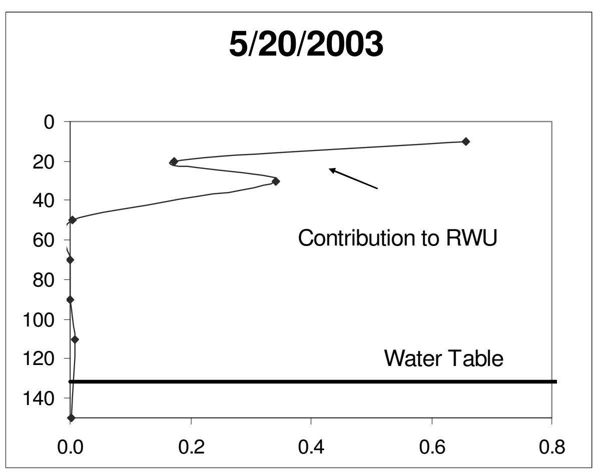
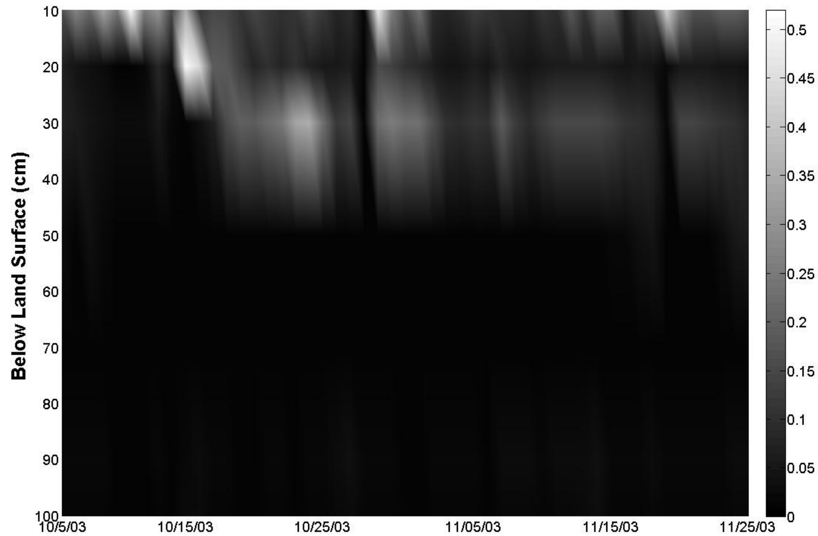
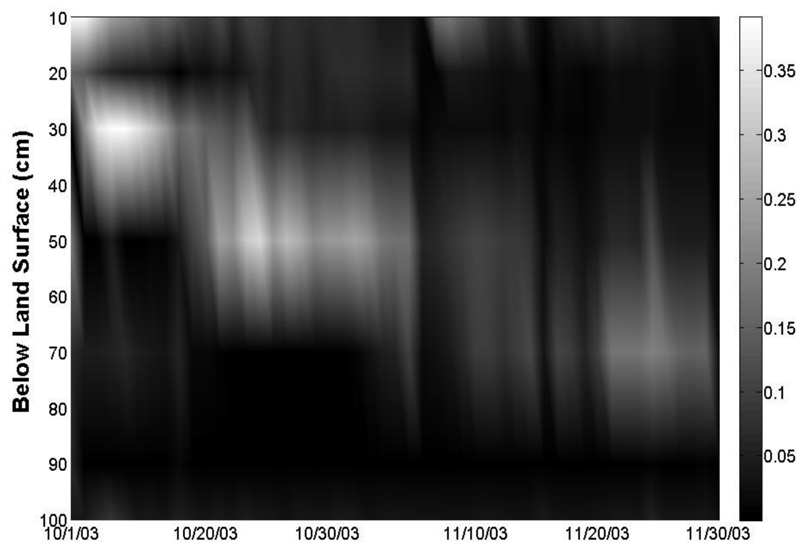


(c)

Figure 6. Root water uptake variation due to a 2.5 cm (1inch) rainfall even on 5/19/2003.



(a)



(b)

Figure 7. Daily root water uptake variation for October and November 2003 for (a) Grassed and (b) Forested land cover.

Figure 7(a) and (b) show a long duration record spanning 2 months, (starting October to end November), with the lighted shaded areas indicating higher root water uptake. From both figures it is reiterated that water uptake significantly shifts in lieu of drier soil layers especially in the case of forest land cover (Figure 7(b)). However in the case of grass land, uptake is primarily concentrated in the top layers, because the root zone is shallower.

5. CONCLUSIONS

This particular analysis provides great insight and poses important questions about the behavior of root water uptake and clearly shows the inadequacy of the presently used models. Because plant compensate for available moisture in different zones, it is necessary to provide model which describes the shifting of the potential evapotranspiration to different layers subject to moisture availability and shifting back of the demand to wetter upper layers in case of rainfall

The methodology presented here elucidates the non linear variation of root water uptake. It also reveals that water uptake is just not directly proportional to amount of the roots but also depends on the ambient water content. Under dry conditions roots can easily take water from deeper wetter soil layers with no reduction in total rate. Hence, traditionally used models are not adequate as such, to model this behavior. Changes in modeling techniques as well as new conceptualizations are required. Plant physiology is one area that needs to be further understood and incorporated to more precisely model water uptake.

The analysis was done using limited set of data from two sites. Therefore there is a strong need to do more studies, especially for different types of land cover for longer duration to improve the robustness of the proposed methodology and increase the confidence in the findings. Also plant root distribution needs to be determined to come up with better idea of root fractions and the mutual relationships between the ambient water content, root fraction and the root water uptake.

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