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Official Sources

1. Spelling: Webster's *New Collegiate Dictionary*
2. Amendments to the U.S. system of soil taxonomy (Soil Survey Staff, 1975) have been issued in the *National Soil Survey Handbook* (NRCS, 1982–1996) and in *Keys to Soil Taxonomy* (Soil Survey Staff, 1996). Updated versions of these and other resources are available at http://www.statlab.iastate.edu/soils/index.html
3. Scientific names of plants: *A Checklist of Names for 3000 Vascular Plants of Economic Importance* (USDA Agric. Handb. 505, see also the USDA Germplasm Resources Information Network database, http://www.ars-grin.gov/npgs/searchgrin.html)
4. Chemical names of pesticides: *Farm Chemicals Handbook* (Meister Publishing, revised yearly)
5. Soil series names: *Soil Series of the United States, Including Puerto Rico and the U.S. Virgin Islands* (USDA-SCS Misc. Publ. 1483, http://www.statlab.iastate.edu:80/soils/osd)
6. Fungal nomenclature: *Fungi on Plants and Plant Products in the United States* (APS Press)
7. Journal abbreviations: *Chemical Abstracts Service Source Index* (American Chemical Society, revised yearly)
8. *The Glossary of Soil Science Terms* is available both in hard copy (SSSA, 1997) and on the SSSA Web page (www.soils.org/sssagloss/). It contains definitions of more than 1800 terms, a procedural guide for tillage terminology, an outline of the U.S. soil classification system, and the designations for soil horizons and layers.

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This issue's cover: In semi-arid regions where irrigation water amount is a limiting factor, efficient water use is possible with close monitoring to minimize water stress. Currently, such monitoring is conducted in the soil, however the use of trees stems to observe the water stress is being researched. Three-rod TDR probes (70-mm) were installed in the trunks of 5-yr old lemon [*Citrus limon* (L.) Burm. F] trees to examine water stress reflected in the stem water content. Please see "Evaluation of TDR Use to Monitor Water Content in Stems of Lemon Trees and Soil and Their Response to Water Stress" by A. Nadler, E. Raveh, U. Yermiyahu, and S.R. Green, p. 437–448.

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Recently, the *SSSAJ* production editing staff has changed systems for preparing accepted manuscripts for typesetting.

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DIVISION S-1—SOIL PHYSICS

Modeling Soil Water Redistribution during Second-Stage Evaporation

A. A. Suleiman* and J. T. Ritchie

ABSTRACT

Calculating the dynamics of soil water content (θ) near the surface and modeling soil water evaporation (E_s) are critical for many agricultural management strategies. This study was performed to develop a model to simulate soil water redistribution during second-stage evaporation (SSE). In this model, the daily change of θ was estimated from the difference between the initial θ (θ_i) and air-dry θ (θ_{ad}), multiplied by a conductance coefficient (C). The C represents the fraction of the remaining soil water ($\theta_i - \theta_{ad}$) that can be removed in 1 d during SSE and is a power function of soil depth. Testing the dependency of C and α (the slope of cumulative evaporation [E_c] vs. square root of time [$t^{1/2}$]) on soil characteristics was done using theoretical and laboratory data. Then the whole model was evaluated in laboratory and field conditions by measuring θ for different soils at different depths during SSE. Linear relationships with zero intercept were found between α and drained upper limit θ (θ_{dul}) with slope and r^2 = 1.19 and 0.69 and 1.39 and 0.95 for laboratory and theoretical data, respectively. Conductance coefficient and θ_{dul} were correlated with $r^2 > 0.9$. Root mean square error (RMSE) between measured and estimated θ in the field was highest (0.014 cm^3 cm^{-3}) at depths of 3 and 6 cm and lowest (0.005 cm^3 cm^{-3}) at the 9-cm depth. The model gave reasonable estimates of both water redistribution and E_s during SSE and is expected to work well for soils for which the diffusivity theory holds.

SOIL WATER EVAPORATION, on one hand, can be a major component of the water balance because most crops have incomplete cover throughout a significant part of a growing season (Ritchie and Johnson, 1990; Qiu et al., 1999). Accurate modeling of E_s is needed to find management strategies that minimize water losses. On the other hand, E_s impacts θ near the surface. Therefore, estimates of E_s and the dynamics of θ during E_s are required for the assessment of soil water management practices such as irrigation scheduling (Lascano and Hatfield 1992; Chanzy and Bruckler, 1993; Bonsu, 1997).

Water evaporation from a soil surface can be divided into two stages: (i) the constant-rate stage in which E_s is limited only by the supply of energy to the surface and (ii) the falling-rate stage in which water movement to the evaporation sites near the surface is controlled by the soil moisture conditions and soil hydraulic properties (Ritchie, 1972; Brutsaert, 1982; Jury et al., 1991; Lockington, 1994; Porte-Agel et al., 2000). Experimental results agreed well with the two-stage model of evaporation (Brutsaert and Chen, 1995; Salvucci, 1997; Menziani et al., 1999; Wythers et al., 1999; Snyder et al., 2000; Ward and Dunin, 2001). The second-stage evaporation can be attributed to the increase in resistance to evaporation (van de Griend and Owe, 1994) and to the decreasing rate of water movement to the surface (Rose, 1996). The constant-rate stage of evaporation varies not only with the prevailing atmospheric environment, but also with soil surface features such as soil surface color and aerodynamic roughness (Mcllroy, 1984). The falling-rate stage of evaporation requires an internal movement of water to the regions where vaporization is actually occurring (near-soil surface) (Mcllroy, 1984).

For many agricultural systems especially those where rainfall events are sparse, most of soil evaporation occurs during second-stage evaporation because first-stage evaporation does not usually last long (Brutsaert and Chen, 1995) after rainfall or irrigation events. The evaporation rate during second-stage evaporation is lower than during first-stage evaporation, but the cumulative evaporation during second-stage evaporation can be very significant within a growing season. Also, the change of θ near the surface (the top 10 cm) can be profound during second-stage evaporation.

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Abbreviations: C, conductance coefficient; DOY, day of year, DSSAT, Decision Support System for Agrotechnology Transfer; E_c, cumulative evaporation; E_s, soil water evaporation; PVC, polyvinyl chloride; SSE, second-stage evaporation; θ, soil water content; θ_{ad}, air-dry soil water content; θ_{dul}, drained upper limit soil water content; θ_i, initial soil water content.

Several mechanistic models have been reported in the literature that estimates E_s using the general equation of water flow (Rose, 1968a; Gardner and Gardner, 1969; van Bavel and Hillel, 1976; Hillel and Talpaz, 1977; Feddes et al., 1978; Norman and Campbell, 1983; Hanks, 1991; Evett and Lascano, 1993; Farahani and Ahuja, 1996). On the contrary, functional models for calculation of E_s using a capacitance approach are rare in the literature (Ritchie and Johnson, 1990) and only a few evaluations have been conducted on such functional models (Gabrielle et al., 1995). Ritchie (1972) developed a simple functional model to estimate daily E_s under second-stage evaporation based on the diffusivity theory. This model has been widely used (e.g., Shouse et al., 1982 [referred to by Jury et al., 1991]; Yunusa et al., 1994) to estimate E_s because of its validity and simplicity.

The Ritchie (1972) model assumes a linear relationship with zero intercept between cumulative evaporation (E_c) and the square root of time ($t^{1/2}$). The value of the slope of this relationship (α) is needed to use Ritchie (1972) model. Yunusa et al. (1994) and Brutsaert and Chen (1995) among others have reported experimental values of α for different soils. Estimating α from other soil properties can be useful. Bonsu (1997) observed that (i) α is correlated to soil texture but the correlation was statistically insignificant and (ii) α increased exponentially with water content of the soil. No attempt has been made to investigate the relationship between θ_{dul} (soil water content at the end of a drainage cycle, closely related to field capacity), see Ritchie et al. (1999) for a description of θ_{dul}, and α. In this study we will examine this relationship because it is expected to be significant.

Rose (1968b) showed that the diffusivity theory relates α to soil water dynamics during the second-stage evaporation (more details are provided in the theory section). He also presented curves of soil water contents at different depth vs. Boltzmann transform. These curves confirmed clearly that soil water contents at different depth had a single relationship with the Boltzmann transform. This implies that the change of θ at any depth and any time during second-stage evaporation is related to α.

The objective of this research was to develop a model based on the diffusivity theory to calculate soil water dynamics during second-stage evaporation. This model does not require water retention curves or soil hydraulic conductivity or soil water diffusivity functions of soil water or matric potential. What it requires is α and drained upper limit, air dry, and initial soil water contents. In case α is unknown, it may be estimated from θ_{dul} as will be shown in the Results section below. This second-stage evaporation model can be incorporated into a water balance of functional crop models such as those of the Decision Support System for Agrotechnology Transfer (DSSAT) family (Boote et al., 1998; Ritchie et al., 1998; Tsuji et al., 1994). In many agricultural fields, especially those with a restricted soil layer in the root zone, saturated layers may impact soil water redistribution and evaporation rate during second stage, and hence the impact of saturated layers on the applicability of diffusivity theory was investigated as well.

THEORY

The generalized isothermal vertical flow equation can be written as follows (Philip, 1957):

$$\frac{\partial\theta}{\partial t} = \frac{\partial}{\partial z}\left(D(\theta)\,\frac{\partial\theta}{\partial z}\right) - \frac{\partial K(\theta)}{\partial z} \qquad [1]$$

where $D(\theta)$ ($m^2\ d^{-1}$) and $K(\theta)$ ($m\ d^{-1}$) are soil water diffusivity and hydraulic conductivity, respectively, θ ($m^3\ ^{-3}$) is soil water content, and t (d) and z (cm) are time and distance, respectively.

Second-Stage Evaporation

When a semi-infinite soil column $z > 0$, initially at a uniform water content θ_{dul}, subsequently has its surface maintained at water content θ_{ad} in equilibrium with the vapor pressure of the atmosphere, the initial and boundary conditions governing flow rate are:

$$\begin{array}{lll} \theta = \theta_{dul} & z \geq 0 & t = 0 \\ \theta = \theta_{ad} & z = 0 & t > 0 \\ E_{sa} < E_p & & t > 0 \end{array} \qquad [2]$$

where θ_{ad} ($m^3\ m^{-3}$) is air-dry volumetric soil water content, E_{sa} is actual soil water evaporation ($m\ d^{-1}$), and E_p is potential soil water evaporation ($m\ d^{-1}$).

The solution of Eq. [1] subject to these conditions assuming that the second term of Eq. [1] is negligible is,

$$z(\theta,t) = \sum_{n=1}^{\infty} \lambda_n t^{n/2} \qquad [3]$$

where $\lambda = zt^{-1/2}$ is the Boltzmann transform. The λ_n are all single-valued functions of θ, and the series converges so rapidly that, except when $t \to \infty$, only the three or four leading terms of the series are needed to describe flow problems, for example, infiltration, or capillary rise above a water table. When gravity can be ignored (e.g., horizontal flow) or neglected without serious errors (e.g., drying of a vertical column of well-structured soil with $\theta_{ad} < \theta_i \leq \theta_{dul}$) only the first term of the series is needed as follows (Rose, 1968b),

$$z = \lambda(\theta)t^{1/2} \qquad [4]$$

Thus, cumulative evaporation (E_c, cm) is given by

$$E_c = \int_{\theta_i}^{\theta_{ad}} z\,d\theta = \alpha t^{1/2} \qquad [5]$$

and the evaporation rate (E, cm d^{-1})

$$E = \frac{1}{2}\,t^{-1/2}\,\alpha \qquad [6]$$

where α, soil water desorptivity (Lisle et al., 1987), is a constant for a given soil (Brutsaert, 1982) for a particular θ_i, and can be described as follows:

$$\alpha = \int_{\theta_{ad}}^{\theta_{dul}} \lambda(\theta)\,d\theta \qquad [7]$$

It is worth noting that the value of α for any soil cannot be obtained from theory (Brisson and Perrier, 1991). However, it can be assessed from second-stage evaporation experiments as will be shown in the results section.

This theory is valid when, for a given soil, evaporation yields water content profiles invariant with $zt^{-1/2}$, that is, when $\lambda(\theta)$ is uniquely dependent on θ (Philip, 1957; Rose, 1968b).

Soil Water Redistribution

The θ at any z, subject to Eq. [2], has an exponential relationship with t as follows:

$$\theta = \theta_{ad} + (\theta_{dul} - \theta_{ad}) \exp(-Ct) \quad [8]$$

where C (d^{-1}) is a conductance parameter that can vary among soils. Figure 1 shows θ as function of t for a soil with three contrasting conductance parameters using Eq. [8]. The value of C is expected to decrease with z because the change of θ decreases with z during second-stage evaporation. The value of C should approach 1 when z approaches 0 and the value of C approaches 0 when z approaches infinity. A power function would describe the relationship between C and depth (z) when $z \geq 1$ cm, because the $\Delta\theta$ is expected to change exponentially with depth:

$$C = az^{b} \quad [9]$$

where a and b are constants.

A problem using Eq. [8] is that it becomes difficult to determine an initial value for t if initial soil water content was $<\theta_{dul}$. However the daily change in θ can be expressed in a form independent of t by taking the first derivative of Eq. [8]:

$$\Delta\theta = C\,(\theta_i - \theta_{ad}) \quad [10]$$

where θ_i is initial θ. The daily evaporation rate, E, is

$$E = \sum_{i=1}^{i=n} \Delta\theta_i \Delta z_i \quad [11]$$

where $\Delta\theta_i$ is the daily change of θ at a particular layer, Δz_i is the thickness of the soil layer being considered, n is number of layers. The cumulative evaporation, E_c, is

$$E_c = \sum_{t=1}^{t=m} \sum_{i=1}^{i=n} \Delta\theta_i \Delta z_i \quad [12]$$

where m is number of days.

At the end of the first day ($t = 1$), according to Eq. [5], E_c equals α. As a result, α can be described as follows, assuming that the thickness of each soil layer is 1 cm

$$\alpha = (\theta_{dul} - \theta_{ad}) \sum_{i=1}^{i=n} C \quad [13]$$

Equation [13] demonstrates the relationship between α and C and θ_{dul}. This equation does not imply that an experimental evaporation cycle of 1 d is enough to come up with values for C or α because the possible errors that may result from such a short experiment.

According to Black et al. (1969):

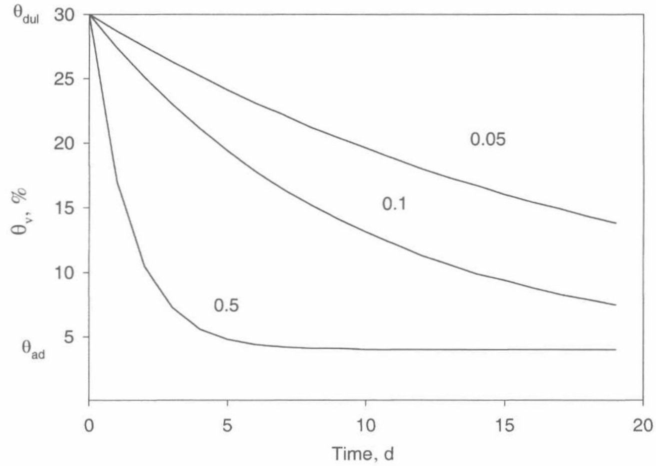

Fig. 1. Soil water profiles with different C values.

$$\alpha = 2(\theta_{dul} - \theta_{ad})\left(\frac{D}{\pi}\right)^{1/2} \quad [14]$$

where D is weighted-mean diffusivity, which for a desorption process, is related to the true soil water diffusivity ($D[\theta]$) by the integral:

$$D = \frac{1.85}{(\theta_{dul} - \theta_{ad})^{1.85}} \int_{\theta_{ad}}^{\theta_{dul}} D(\theta)(\theta_{dul} - \theta)^{0.85}\, d\theta \quad [15]$$

MATERIALS AND METHODS

Laboratory and field experiments were conducted to study the soil water dynamics during drying cycles. Besides the laboratory and field experiments, data from six different soils (Rose, 1968b) and twelve theoretical soils (thereafter mean soils) were used to develop a relationship between α and θ_{dul}. Some properties of Rose (1968b) soils are shown in Table 1. For Rose (1968b) soils, the curves of soil water contents at different depth vs. Boltzmann transform (presented in Rose [1968b]) were used to find values of α by solving Eq. [7] because α values were not readily available. Some soil properties of mean soils are provided in Table 2. The hydraulic parameters were obtained by Rosetta software found at http://www.ussl.ars.usda.gov/models/rosetta/rosetta.htm (Schaap et al., 2001) using texture and bulk density values. These hydraulic parameters were needed to find θ_{dul} and α for each soil of the mean soils. For mean soils, θ_{dul} was assumed equal to soil water content after 10 d of simulated drainage. Drainage was simulated numerically assuming a unit hydraulic gradient and

Table 1. Description and some properties of Rose (1968b) soils.

Soil	Soil description	Aggregate size	Particle density	Apparent density	Soil water content at saturation	θ_{dul}	α
		mm	g cm^{-3}		cm^3 cm^{-3}		cm d$^{-1/2}$
Highfield	A clay loam pasture soil with an excellent and stable structure, under grass for several centuries.	1–2	2.47	0.86	0.651	0.323	0.30
Lansome	A sandy market garden soil dressed annually with farmyard manure, but with a structure easily broken down by mechanical abrasion.	0.5–1	2.54	1.09	0.570	0.185	0.21
Greatfield	A sandy clay loam arable soil on old grassland.	1–2	2.58	1.04	0.595	0.263	0.20
Ignited soil	Greatfield soil crumbs ignited at 850°C for 1 h.	0.5–1	2.56	0.94	0.630	0.236	0.22
Subsoil clay	Saturated subsoil from 1 m below the Greatfield plowed layer weathered into crumbs by alternate freezing and thawing.	1–2	2.7	1.01	0.623	0.322	0.52
Sepiolite	Nonswelling magnesium silicate mineral.	1–2	2.47	0.58	0.766	0.428	0.56

Table 2. Mean soils properties (D_b is bulk density) and hydraulic parameters†.

Soil	Clay	Sand	D_b	θ_{dul}	θ_r	θ_s	α_1	n	K_o	L	C_1	C_2
	%		g cm^{-3}	cm^3 cm^{-3}					cm d^{-1}			
Clay	80	10	1.15	0.389	0.116	0.565	0.0249	1.213	7.314	−2.71139	0.0684	15.062
Clay loam	35	35	1.32	0.263	0.087	0.465	0.0135	1.417	3.785	−0.63673	0.0631	22.140
Loam	20	40	1.41	0.202	0.062	0.401	0.0095	1.526	3.074	−0.22231	0.0951	27.205
Loam sand	7	83	1.64	0.069	0.045	0.350	0.0387	1.842	26.607	−0.95713	0.0684	86.577
Sandy	3	95	1.79	0.050	0.049	0.303	0.0322	3.013	22.870	−0.83726	0.1526	93.136
Sandy clay	40	55	1.33	0.284	0.091	0.477	0.0251	1.292	8.037	−1.6431	0.0780	20.089
Sandy clay loam	25	65	1.42	0.215	0.071	0.433	0.0229	1.391	8.923	−1.04902	0.1143	24.847
Sandy loam	10	65	1.56	0.139	0.040	0.364	0.0331	1.423	17.614	−1.15236	0.2886	32.742
Silty	5	5	1.54	0.191	0.050	0.407	0.0077	1.603	3.068	0.11813	0.0864	29.301
Silty clay	45	10	1.23	0.313	0.101	0.521	0.0137	1.380	3.787	−0.72923	0.0517	19.152
Silty clay loam	35	10	1.27	0.282	0.092	0.493	0.0091	1.488	2.401	−0.22006	0.0537	21.096
Silty loam	15	20	1.41	0.201	0.060	0.400	0.0051	1.676	1.665	0.22157	0.1014	27.505

† θ_{dul}, drained upper limit soil water content; θ_r, residual soil water content; θ_s, saturated soil water content; α_1, curve shape parameter; n, shape parameter; K_o, hydraulic conductivity; L, curve parameter; C_1 and C_2, constants.

a time step of 1 min using van Genuchten (1980) hydraulic conductivity equation. Equation [16] (van Genuchten et al., 1991) can be used in Eq. [15] to find a weighted-mean diffusivity, which is required to find α from Eq. [14]. However, soil diffusivity obtained by Eq. [16] goes to 0 at θ_r not at θ_{ad}, which may result in an inaccurate weighted-mean diffusivity and α. To overcome such a problem and to make it mathematically feasible to integrate Eq. [15], soil water diffusivities obtained from Eq. [16] for soil water contents between θ_{dul} and $0.5 \times (\theta_{dul} + \theta_r)$ were fitted into an exponential equation (Eq. [17]) assuming that soil water diffusivity goes to 0 at θ_{ad}. This exponential equation of soil water diffusivity has been used for desorption experiments (Jalota and Prihar, 1991).

The van Genuchten et al. (1991) soil water diffusivity function is as follows:

$$D(S_e) = \frac{(1-m)K_o S_e^{1-1/m}}{\alpha_1 m(\theta_s - \theta_{rs})}\left[\left(1 - S_e^{1/m}\right)^{-m} + \left(1 - S_e^{1/m}\right)^{m} - 2\right] \quad [16]$$

where $D(S_e)$ is soil water diffusivity (m^2 d^{-1}), S_e is relative saturation [$= (\theta - \theta_r)/(\theta_s - \theta_r)$], θ_s and θ_r are saturation and residual soil water content (m$^{3\ -3}$), respectively, K_o is hydraulic conductivity at saturation point (m d^{-1}), and m ($= 1 - 1/n$), α_1, and l are curve shape parameters. Values of the different hydraulic parameters for mean soils are shown in Table 2.

The exponential soil water diffusivity ($D[\theta]$) can be written as:

$$D(\theta) = C_1 \exp(C_2\theta) \quad [17]$$

where C_1 and C_2 are constants and values of these for mean soils are shown in Table 2.

Laboratory Experiments

Two different soils from Michigan were used to investigate soil water redistribution during second-stage evaporation in 1997. These soils were a Misteguay (Fine, mixed, calcareous, mesic Aeric Endoaquepts) soil, obtained from the Saginaw area, and a Capac (Fine-loamy, mixed, active, mesic Aquic Glossudalfs), obtained from Lansing area. Saginaw soil was a loamy soil (25.4% clay, 43% sand, and 1.31 g cm^{-3} bulk density) and Lansing soil was a sandy loam soil (9.4% clay, 65.4% sand, and 1.44 g cm^{-3} bulk density). The two soils were air-dried, sieved through a 2-mm screen, and then assembled uniformly into insulated polyvinyl chloride (PVC) columns of 60 cm in height and 30 cm in diameter by adding soil in small increments while continuously shaking the columns until the soil stopped settling. Twenty-centimeter time domain reflectometry (TDR) probes were installed horizontally at depths of 3, 6, 9, 12, and 15 cm from the surface. The top 25 cm of the soil columns were saturated by adding water on soil surface, which then was covered with a black plastic sheet to avoid evaporation. The soils were allowed to drain for 10 d, to obtain initial conditions similar to Eq. [2], and then the soil surface was uncovered. A flourescent light source and a table fan were directed toward the soil surface of each column to ensure high potential evaporative losses (≈15 mm d^{-1}). A constant air temperature (25°C) and atmospheric relative humidity (20) were maintained constant throughout the experiment. Soil water content was monitored at the five depths every 15 min for about 2 mo.

The two soils were also used to evaluate the effect of saturated layers on soil water distribution and evaporation rate during second-stage evaporation. The air-dried soils were assembled uniformly into insulated PVC columns of 150 cm in height and 30 cm in diameter using the same procedure mentioned above. Twenty-centimeters TDR were installed horizontally at depths of 3, 6, 9, 12, 15, 25, 35, 45, 55, 65, and 75 cm from the surface. The soils were saturated from the bottom using a constant head of 150 cm. The soils were allowed to drain for 10 d while the soil surface was covered and then the soil surface was uncovered. A fluorescent light source and a table fan were directed toward the soil surface of each column to ensure high potential evaporative losses (≈15 mm d^{-1}). A constant air temperature (25°C) and atmospheric relative humidity (20) were maintained constant throughout the experiment. Soil water content was monitored at the 11 depths every 20 min for 18 d. A data logger (CR21X, Campbell Scientific Inc., Logan, UT) controlled the digital TDR (Tektronix Model 1502B, Tektronix Inc., Beaverton, OR) and the multiplexing system. For, more details about automating and multiplexing soil moisture measurement by TDR, you may refer to Baker and Allmaras (1990).

Field Experiment

Twenty-centimeter TDR probes were installed horizontally at depths of 3, 6, 9, 12, and 15 cm from the surface of a Capac soil (9.4% clay, 65.4% sand, and 1.44 g cm^{-3} bulk density) in the Lansing area on 10 July 1997. The θ was measured at these depths every 20 min for 2 mo using automated system similar to that in the laboratory experiment. Daily solar irradiance, maximum and minimum air temperatures, and rainfall from day of year (DOY) 200 (19 July) to 280 (7 October) of 1997 are shown in Fig. 2. Rainy days and days after those rainy days, on which soil water drainage was occurring in the top 50 cm of the soil profile were not considered good days for second-stage evaporation. Also, days on which solar irradiance was low resulting in a low potential evaporation which the soil could meet were not considered good days for second-stage evaporation. Overall, the change of θ at the five depths mentioned above was assumed to be due to second-stage evaporation for only 17 d because the boundary conditions for second-stage evaporation did not prevail during other days.

RESULTS AND DISCUSSION

Laboratory Results

A relationship between the slope of cumulative evaporation vs. $t^{1/2}$ (α) and θ_{dul} was investigated and shown in Fig. 3. The value of α for laboratory data (Rose [1968b] and loamy and sandy loam soils) ranged from 0.21 to 0.55 cm $d^{-1/2}$ for soils with θ_{dul} of 0.18 to 0.42 cm^3 cm^{-3}. For mean soils, α ranged from 0.07 to 0.53 cm $d^{-1/2}$ for soils with θ_{dul} of 0.05 to 0.39 cm^3 cm^{-3}. Yunusa et al. (1994) found that α was 0.4 cm $d^{-1/2}$ for a fine-textured Xeralfic Alfisol. Ritchie (1972) presented α values for four soils namely Adelanto clay loam (coarse-loamy, mixed, superactive, thermic Xeric Haplargid) (α = 0.508 cm $d^{-1/2}$), Yolo (fine-silty, mixed, superactive, nonacid, thermic Mollic Xerofluvent) loam (α = 0.404 cm $d^{-1/2}$), Houston black clay (very-fine, smectitic, thermic Oxyaquic Hapludert) (α = 0.350 cm $d^{-1/2}$), and Plainfield sand (Mixed, mesic Typic Udipsamment) (α = 0.334 cm $d^{-1/2}$). Although Yunusa et al. (1994) and Ritchie (1972) did not mention θ_{dul} values for the above

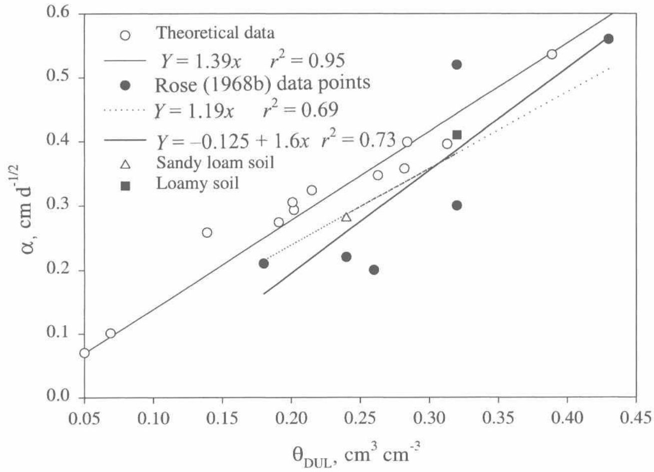

Fig. 3. Relationships between α and θ_{dul} of laboratory soils (eight soils: Loamy and Sandy loam soils from this study and six soils from Rose [1968b]) and mean soils.

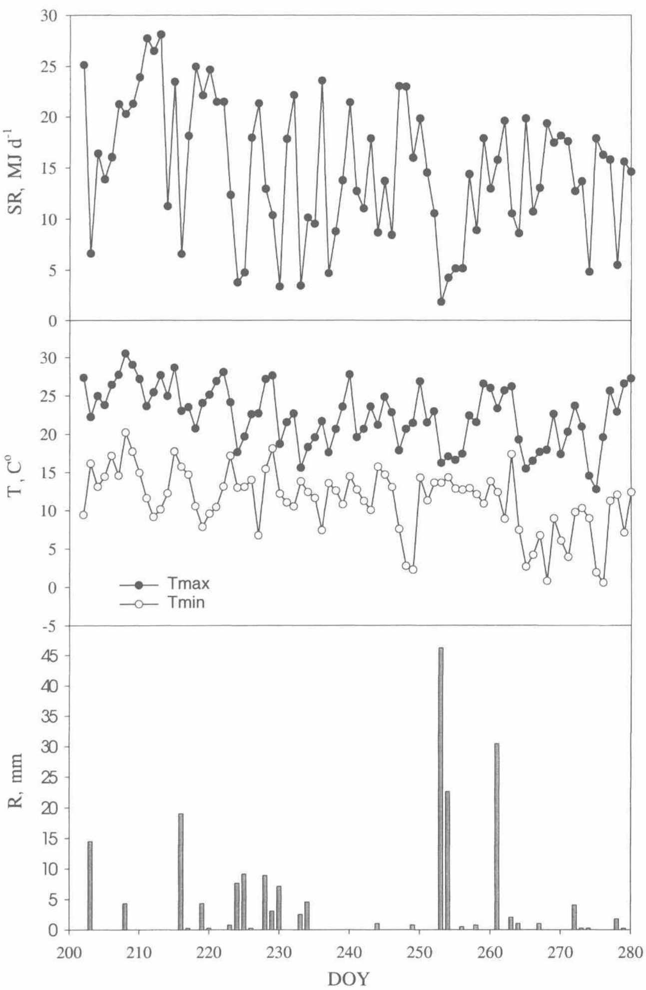

Fig. 2. Daily solar radiation (SR), maximum (Tmax) and minimum (Tmin) air temperatures, and rainfall (R) from DOY 200 to 280 in 1997.

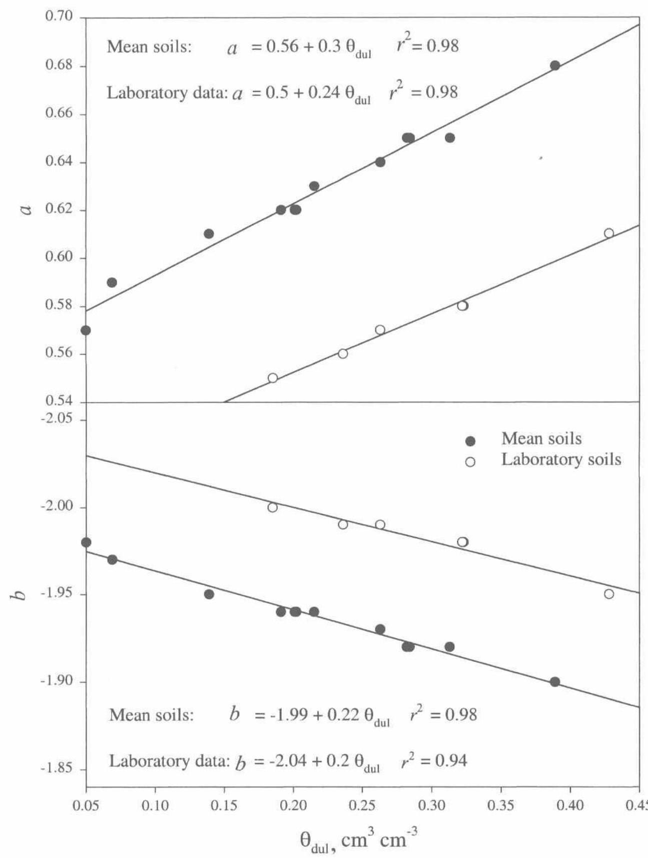

Fig. 4. Relationships between *a* and *b* with θ_{dul} for laboratory soils (eight soils: Loamy and Sandy loam soils from this study and six soils from Rose [1968b]) and mean soils.

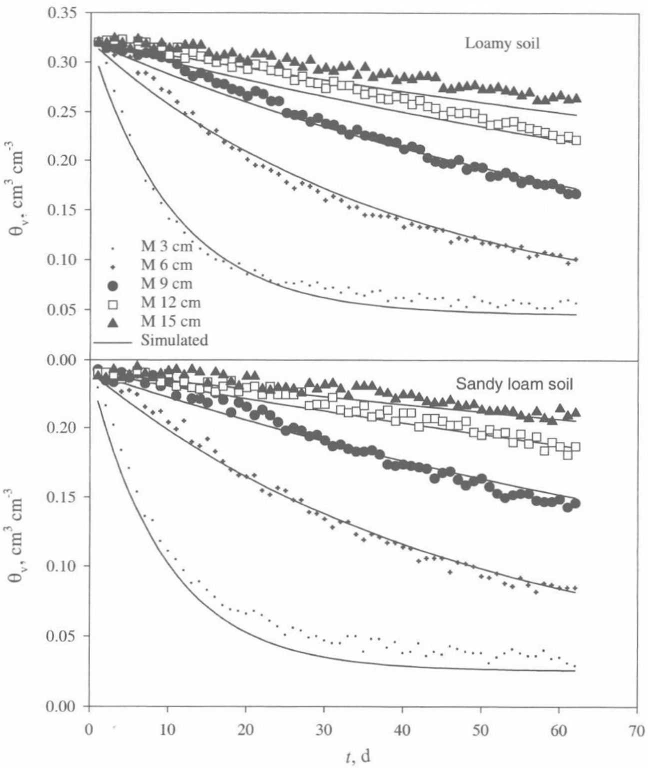

Fig. 5. Measured and simulated soil water content during second-stage evaporation.

soils, the α values they found were within the range of our α values.

For laboratory data, a linear relationship was found between α and θ_{dul} with $r^2 = 0.73$. Another linear relationship with zero intercept was developed between α and θ_{dul} with slope = 1.19 and $r^2 = 0.69$ (Fig. 3). For the mean soils, the slope of the linear relationship with zero intercept between α and θ_{dul} was 1.39 with $r^2 = 0.95$. The difference between the slope of α and θ_{dul} (with zero intercept) between laboratory and mean soils can be contributed to a faster decrease of soil water diffusivity of Rose (1968b) soils than mean soils because

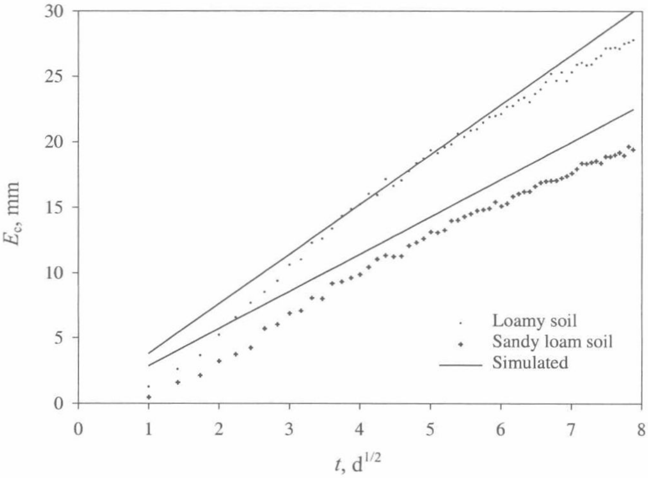

Fig. 6. Measured and simulated cumulative evaporation of loam and sandy loam soils during second-stage evaporation.

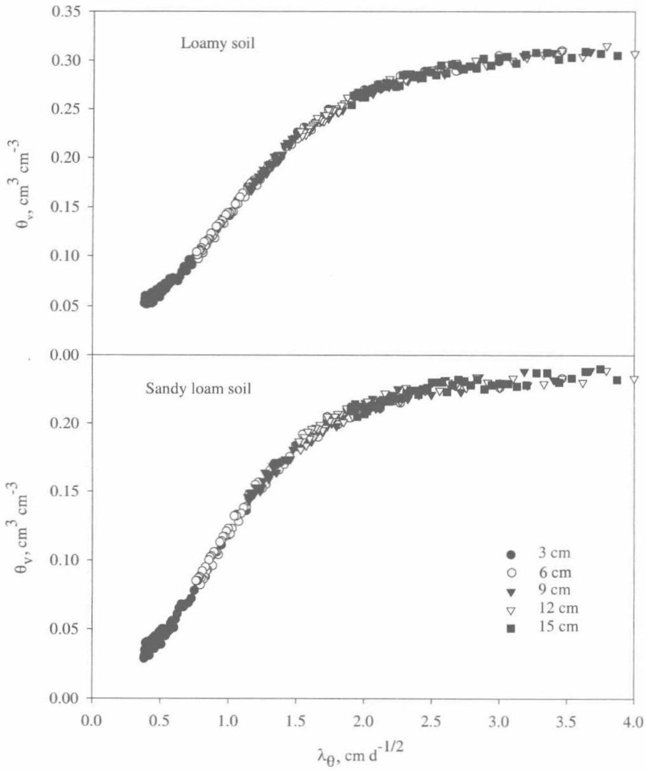

Fig. 7. Soil water content profiles of 60-cm columns of loamy and sandy loam soils vs. Boltzmann transform during second-stage evaporation.

Rose (1968b) soils were formed of aggregates. Two reasons make the fit with zero intercept more appealing: (i) Eq. [13] shows that α goes to 0 when θ_{dul} goes to 0, and (ii) simply no soil would have negative soil water evaporation. We recommend a slope of 1.39 to be used to estimate α from θ_{dul} if α was not measured.

The developed relationships with zero intercept between α and θ_{dul} was used to obtain simulated values of a and b for the eight different laboratory soils and 12 mean soils. Trial and error was used to obtain a and b values at the different θ_{dul}. A set of a and b values was accepted if the simulated E_c = E_c calculated from the developed relationship between α and θ_{dul}, the diffusivity theory was preserved, and a linear relationship between E_c and $t^{1/2}$ was kept. Linear relationships were found between a and θ_{dul} and between b and θ_{dul} with $r^2 > 0.94$ for laboratory and mean soils as shown in Fig. 4. The higher the θ_{dul} are, the greater the a and b values are (b closer to zero). Soils with same θ_{dul} and higher α have higher a and b values. For both laboratory and mean soils, b could be estimated from a using $b = 0.8 \times a - 2.44$. The ratio between a of mean soils and a of laboratory soils equals 1 + 0.86 ln(the ratio of α of mean soils to α of laboratory soils). The relationship between a and b and a of laboratory data and a for mean soils (or any other soil for which α is known) can be very useful in finding the appropriate a and b for a particular soil. The relationships of a and b with θ_{dul} were evaluated and validated for soils with θ_{dul} ranging from 0.05 to 0.42 cm^3 cm^{-3}.

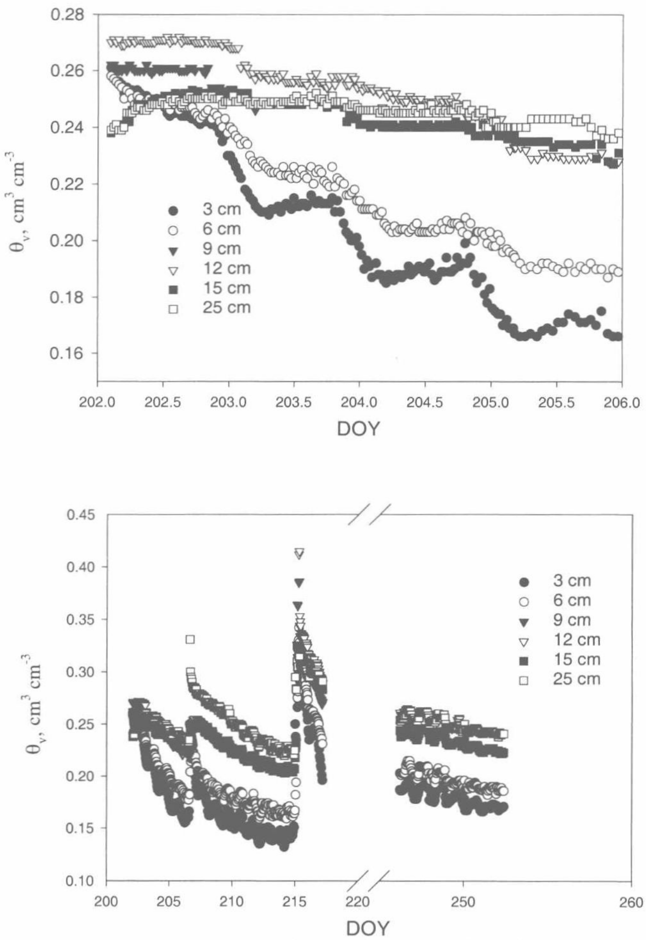

Fig. 8. Soil water content profiles at six depths of sandy loam soil.

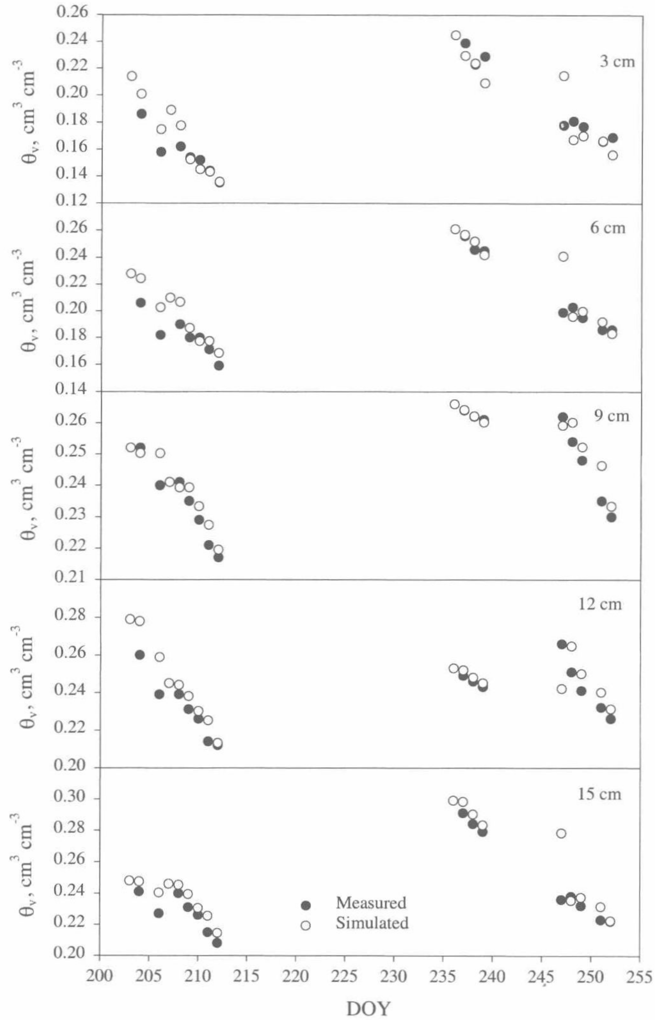

Fig. 9. Measured and simulated soil water content of a bare soil in Lansing field during second-stage evaporation.

The daily θ at 3-, 6-, 9-, 12-, and 15-cm depths for 2 mo during second-stage evaporation for loamy and sandy loam soils in the laboratory is shown (Fig. 5). Soil water content went from θ_{dul} toward θ_{ad} at all soil depths. The θ_{dul} was about 0.32 cm^3 cm^{-3} for loamy soil and about 0.24 cm^3 cm^{-3} for sandy loam soil. The θ_{ad} was about 0.05 cm^3 cm^{-3} for loamy soil and about 0.03 cm^3 cm^{-3} for sandy loam soil. The change of θ decreased with increasing depth. In other words, the daily change of soil water content was greater at 3 cm than at 6 cm, and at 6 cm was greater than at 9 cm, and so on. For instance, the change of θ at 3 cm in 2 mo was 0.27 cm^3 cm^{-3} for loamy soil and 0.21 cm^3 cm^{-3} for sandy loam soil. While, the change of θ at 15 cm in 2 mo was 0.04 cm^3 cm^{-3} for loamy soil and 0.03 cm^3 cm^{-3} for sandy loam soil.

Using the developed relationships between a and b with θ_{dul}, a and b were 0.58 and −1.98 for loamy soil and 0.56 and −1.99 for sandy loam soil. Using these values of a and b, the new model produced simulated water contents close to the measured ones at 3-, 6-, 9-, 12-, and 15-cm depths for loamy and sandy loam soils (Fig. 5). The model is expected to do as well at other soil depths as long as the diffusivity theory holds because θ at any depth and time is function of Boltzmann transform. For instance, θ at the 3-cm depth after 1 d of evaporation subject to Eq. [2] is equal to θ at the 18-cm depth after 36 d.

Loamy soil evaporated more water than sandy loam soil (Fig. 6). At Day 1, measured E_c was too low because the change of θ near the surface (0–2 cm) was not included in computing E_c since the closest TDR probe to surface was at 3 cm. The relationship between E_c and $t^{1/2}$ was linear as expected from the theory. The developed model gave good estimates of E_c for both loam and sandy loam soils for a 60-d period.

Volumetric water contents at 3-, 6-, 9-, 12-, and 15-cm depths had the same relationship with λ_θ for loam and sandy loam soils for about 62 d (Fig. 7). This supports that diffusivity theory for uniform and isotropic soil drying under second-stage evaporation and is in agreement with Rose (1968b) and Black et al. (1969). Significant changes of θ started when λ_θ was about 2.5 cm $d^{-1/2}$ for loamy soil and at about 2 cm $d^{-1/2}$ for sandy loam soil. Soil hydraulic characteristics determine the rate at which soil water moves upwards.

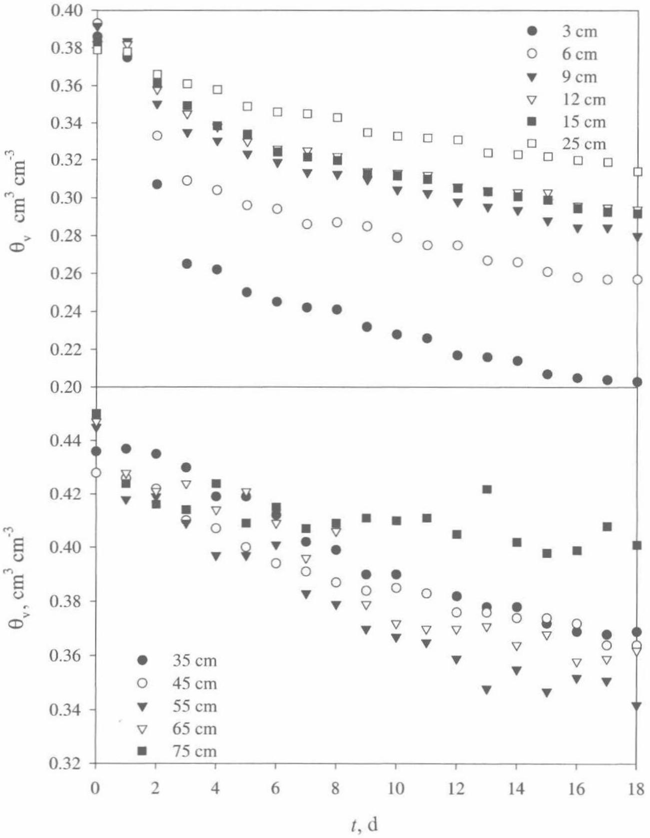

Fig. 10. Soil water content profiles of 150-cm columns of loamy soil during second-stage evaporation.

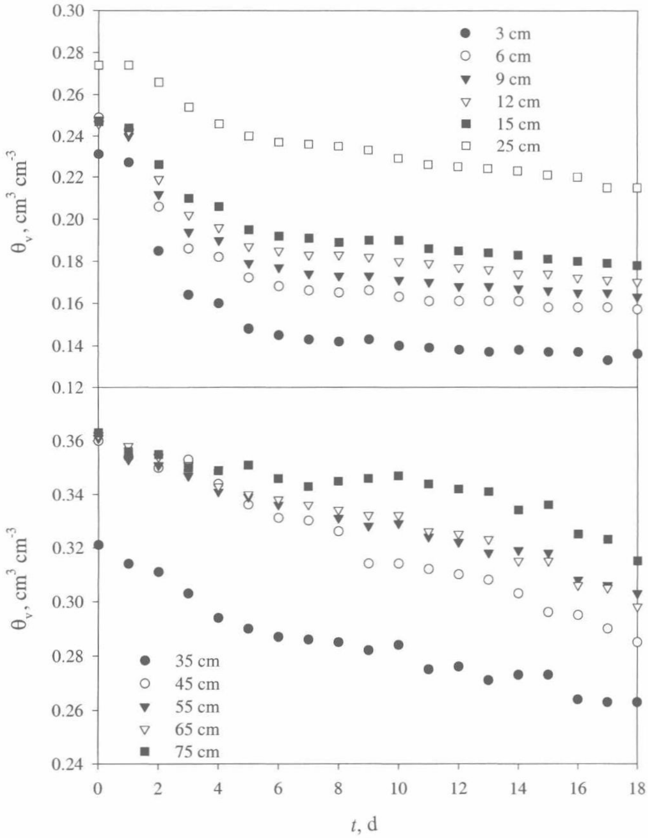

Fig. 11. Soil water content profiles of 150-cm columns of sandy loam soil during second-stage evaporation.

Field Results

Field data were used to evaluate the developed model under field conditions. Soil water content profiles were shown at six depths (Fig. 8). A 4-d period, on which daily solar irradiance was >15 MJ d^{-1}, average air temperature was about 20°C, and rainfall was 0, was selected and plotted to show the fluctuation of θ between day and night. It was clear that θ at the 3-cm depth and, to lesser extent, at 6 cm increased at night as a result of upward soil water flow. The driving force of such movement is the soil hydraulic gradient.

Simulated soil water contents in the field showed good agreement with the measurements (Fig. 9). The RMSE (cm^3 cm^{-3}) was 0.014, 0.014, 0.005, 0.01, and 0.012 at 3-, 6-, 9-, 12-, and 15-cm depths, respectively, using the new model with $a = 0.56$ and $b = -1.99$. The RMSE values indicate that the new model calculates θ distribution reasonably well under second-stage evaporation.

Effect of Water Table

If one or more of the boundary conditions of second-stage evaporation was violated, the above relationships may not be applicable. For instance, having a shallow water table may violate the boundary condition of semi-infinite soils. Because a shallow water table is evident in many agricultural fields, the impact of a shallow water table (or saturated layers within the root zone) on the diffusivity theory under second-stage evaporation was investigated.

Figures 10 and 11 show volumetric θ at 11 depths for loam and sandy loam soils. The θ_i was not uniform but rather increased from θ_{dul} at 3 cm to about saturation at 75 cm for loamy soil (Fig. 10) and increased from θ_{dul} at 3 cm to about saturation at 45 cm for sandy loam soil (Fig. 11). Soil water content was function of depth and time and the change of θ decreased with depth and time under second-stage evaporation.

To test the validity of diffusivity theory under such conditions, volumetric θ was plotted against λ_θ as shown in (Fig. 12). Soil water content at any depth for loam and sandy loam soils was going from its initial value toward a certain soil water content higher than θ_{ad}.

That θ was about 0.19 cm^3 cm^{-3} for loamy soil and about 0.12 cm^3 cm^{-3} for sandy loam soil. Soil water content had a different relationship with λ_θ at each depth when $\lambda_\theta \geq 2$ cm $d^{-1/2}$ since initial soil water was different at different depths (Fig. 10 and 11). It was concluded that Boltzmann transform couldn't be formulated as in Eq. [4] since there was no single-valued function between soil water content and λ_θ.

A linear relationship with zero intercept was found between E_c and $t^{1/2}$ for loam and sandy loam soils (Fig. 13). This suggested that soil water evaporation was limited by θ and soil characteristics. Evaporation from a loam soil was higher than that from a sandy loam soil.

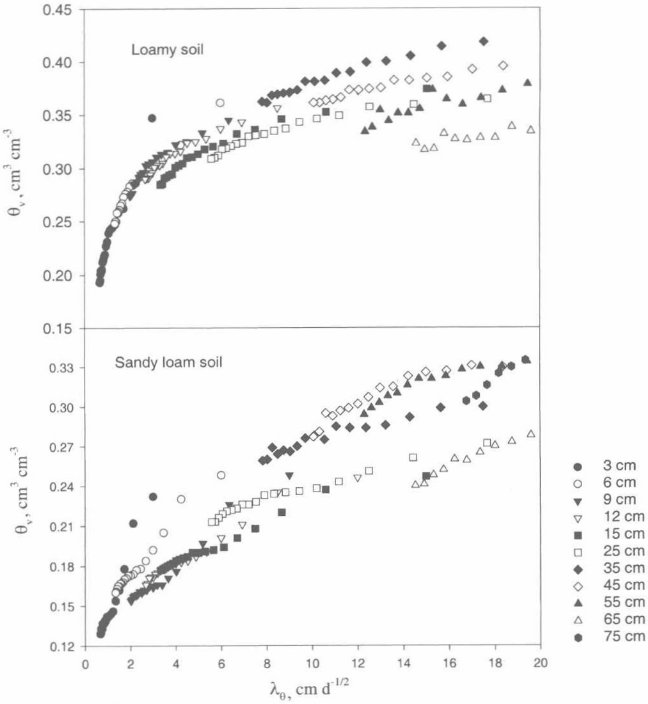

Fig. 12. Soil water content profiles of 150-cm columns of loamy and sandy loam soils vs. Boltzmann transform during second-stage evaporation.

The slope of the best-fit line was 15.4 mm d$^{-1/2}$ for a loamy soil and 12.1 mm d$^{-1/2}$ for sandy loam soil. Hence, α for soils affected by shallow soil water table was about four times greater than α for semi-infinite soils under second-stage evaporation. This led us to conclude, that the relationships that developed for semi-infinite soils were not applicable for soils affected by a shallow water table.

CONCLUSIONS

A model was developed on the basis of the diffusivity theory to simulate the soil water redistribution dynamics during second-stage evaporation. The developed model was tested in field and laboratory conditions. Three parameters (α, a, and b) characterized the soil water dynamics during second-stage evaporation. The value of α and the two constants (a and b) were different for soils with different θ_{dul}. Linear relationships between α, a, and b with θ_{dul} were developed. These relationships enabled the developed model to simulate soil water redistribution and soil water evaporation for diverse soils accurately during second-stage evaporation. It was found that using only the first term of the series in defining Boltzmann transform was an inappropriate approximation when some soil layers within the profile were saturated because soil water contents at different depths had different relationships with Boltzmann transform. Further studies should be conducted on modeling evaporation from soils that have shallow water table.

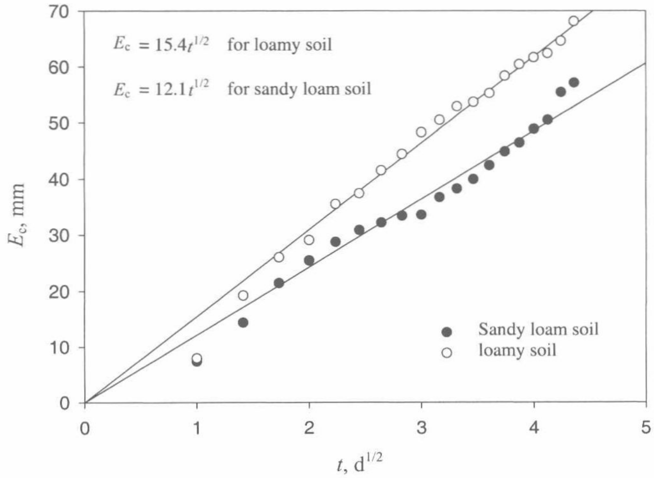

Fig. 13. Measured and simulated cumulative evaporation of loam and sandy loam soils during second-stage evaporation.

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Root-System Development and Water-Extraction Model Considering Hydrotropism

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ABSTRACT

A two-dimensional model that combines root-system development and water extraction by roots is proposed to simulate the dynamic interaction between root growth and soil-water flow. Both of hydrotropism and gravitropism were considered as the controlling factors of root growth in the proposed root-system development model. The finite-element method was employed to compute the soil-water flow caused by water extraction, evaporation, and irrigation. We succeeded in simulating the plagiogravitropic elongations of lateral roots under a plane condition, and the asymmetric architecture of root system under a slope condition by the proposed model in which the root hydrotropism is considered. On the other hand, we cannot simulate such morphological characteristics of a root system by the use of the conventional model in which a random elongation factor is employed, and root hydrotropism is not considered. The results support the importance of hydrotropism in root-system development and the availability of the proposed model in which the hydrotropism is considered.

THE INTERACTION between plant-root systems and soil, especially soil with moisture, is very important in many respects. For example, root-water uptake from the soil plays an important role in the hydrological process of water flow through soil, plants, and air. In the field of crop science, many experimental studies indicate that variations in soil-water conditions affect the root-system architecture of various kinds of crops, and that architecture affects absorption efficiency. However, the interaction between the roots and soil, which is hidden under the ground, is difficult to investigate and studies of them are lagging when compared with the studies of shoots or leaves. The modeling of plant-root system development and of soil-water extraction by plant roots is therefore very useful for understanding the interaction between plant-root systems and soil.

The water extraction by plant-root system can be calculated by adding a term of water-extraction intensity, S (s^{-1}), to the Richards' equation, which is the fundamental equation of water flow in unsaturated soil. Various models that give the extraction intensity S have been proposed (Gardner, 1964; Herkelrath et al., 1977a; Herkelrath et al., 1977b; Feddes et al., 1978). Herkelrath et al. (1977a, 1977b) proposed that the extraction intensity S is proportional to the potential difference between roots and soil, volume saturation of the soil space, and root length per unit soil volume. Feddes et al. (1978) introduced water-extraction efficiency as a function of soil-water potential and succeeded in showing the behavior of soil-water extraction by a plant-root system under certain conditions. However, because root-system development is not considered in these water-extraction models, the root length per unit soil volume should be given a priori as a function of soil depth. That is, the water-extraction model alone cannot simulate the behavior of soil-water extraction by roots with active growth, which changes the distribution of the root system over time.

As a pioneer study of modeling plant-root system development, Lungley proposed a two-dimensional model (Lungley, 1973). Diggle and Pages developed three-dimensional root-system architecture models, and they successfully simulated crop plant-root system development and morphological architecture three-dimensionally (Diggle, 1988; Pages et al., 1989). It seems that the fundamental part of root-system development modeling has been completed by these models. Recently, root-system development models focus on the interactions between plant-root systems and soil-water flow (Clausnitzer and Hopmans, 1994; Doussan et al., 1998), or nutrient supply (Somma et al., 1998), which combines the water uptake models or nutrient uptake models. The models shown above are not the only ones that exist. Many other models have been proposed and some of them have succeeded in simulating root-system development in various plant species, under various conditions (e.g., Lynch et al., 1997; Jourdan and Rey, 1997).

However, all applications of these models were made to the root systems developing under plane conditions, and no application under slope conditions can be found. It has been generally known that a plant that grows on a slope has an asymmetric root system, and this asymmetric architecture of the root system has been confirmed by some experimental studies (Yamadera, 1990; Scippa et al., 2001). Recently, the contribution of plant-root system architecture to slope stability has become one of the main interests in the fields of erosion control and revegetation technology. Some studies have shown that plant-root systems growing under hillside slope contributes to the slope stability, increasing the soil strength by the their architecture, and decreasing the soil-water content by water uptake (Greenway, 1987). This contribution has been investigated, considering root strength, growth, and rate of decay (Watson et al., 1999).

To simulate the root-system development and soil-water flow under a slope condition, it is necessary to take into account the effect of the slope condition on root growth. Some experiment results have shown that root growth is influenced by hydrotropism, which is the root elongation toward water (Takahashi, 1994; Takano et al., 1995). Because soil moisture exhibits asymmetric distribution under slope conditions, root hydrotropism can be one of the main factors causing the asymmetric root-system development.

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