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CALIBRATION OF A ROOT WATER UPTAKE MODEL IN SPATIALLY VARIABLE SOILS[1]

J.W. HOPMANS and E. GUTTIÈREZ-RAVÉ

LAWR Department, Veihmeyer Hall, University of California, Davis, CA 95616 (U.S.A.)
Department de Suelos y Riegos, Apartado 240, 14071 Córdoba (Spain)

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ABSTRACT

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A procedure is discussed to calibrate a root water uptake model in unsaturated water flow modeling, thereby taking into consideration the variability of soil hydraulic properties. The Monte-Carlo analysis employed, in combination with a trial-and-error procedure to optimize the parameters of the Feddes reduction function, resulted in improved values for the actual transpiration in the Hupsel watershed as compared with a single site calibration. It is shown that the site originally chosen for the calibration of the reduction function behaved quite differently than most other naturally occurring soils in the watershed. The hydraulic functions of this calibration site very much limited transpiration and, therefore, cannot be regarded as representative.

INTRODUCTION

The modeling of water flow in unsaturated cropped soils requires the description of water uptake by plant roots. Because of the complexity of root growth and water transport through the plant, often a simplified model is chosen to simulate root water uptake.

One of these models simply assumes that the actual transpiration rate is a function of soil water pressure head and potential transpiration only, thereby assuming that the root density is constant with depth and time. In such a model, root water uptake is described from a macroscopic viewpoint. It considers a bulk volume of soil and roots, rather than the microscopic approach in which the transport mechanism through single roots is analyzed.

The advantage of the macroscopic approach is clear. It does not require complete insight in the physical process of root water uptake and, therefore, eliminates the need for difficult to measure soil and plant parameters. It is, however, still needed to calibrate such macroscopic model for every different soil–plant system.

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[1] Contribution from the Department of Hydraulics and Catchment Hydrology, Wageningen Agricultural University, Nieuwe Kanaal 11, Wageningen (The Netherlands)

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The macroscopic root water uptake model to be discussed in this paper was introduced by Feddes et al. (1978). Although the documentation of this model in Belmans et al. (1981) gives values for the calibration parameters, neither this report nor other related publications (Belmans et al., 1983) give indications how to match simulated and actual root water uptake for large areas.

In a previous study (Hopmans and Stricker, 1988) the influence of spatially variable soil hydraulic properties on the soil-water regime in the 650 ha Hupsel watershed was described. The unsaturated water flow model SWATRE (Belmans et al., 1981) was used which incorporates the macroscopic root water uptake model. One of the conclusions of the work by Hopmans and Stricker was that differences between simulated and measured actual transpiration were possibly caused by the calibration procedure of the root water uptake model. The calibration was based on soil physical measurements at one site only, thereby assuming that it was representative for the whole watershed. However, it was previously found that the soil hydraulic properties in the Hupsel watershed exhibited a large variability (Hopmans and Stricker, 1987). Since it is well known that the hydraulic functions have a major influence on soil water flow, it is therefore proposed to calibrate the root water uptake model while taking into account the variability of the soil hydraulic properties.

The macroscopic volume to which the water uptake model then applies is the root zone of the total watershed. Although not completely so, it is further assumed that the only crop grown in the watershed is grass.

MATERIALS AND METHODS

One-dimensional unsaturated water flow is simulated with the numerical model SWATRE (Belmans et al., 1983). This model was chosen for various reasons: (1) the soil water domain can be differentiated into layers with different soil hydraulic properties; (2) various types of boundary and initial conditions can be used; (3) it incorporates a sink term to simulate root water uptake; and (4) soil water transport can be simulated during a whole growing season with a reasonable amount of computer time.

SWATRE numerically solves the following unsaturated flow equation:

$$\frac{\partial \psi}{\partial t} = \frac{1}{C(\psi)} \left\{ K(\psi) \left[\frac{\partial \psi}{\partial z} + 1 \right] \right\} - \frac{S(\psi)}{C(\psi)} \quad (1)$$

where: ψ = soil water pressure head (cm, < 0 in unsaturated soil); z = vertical coordinate (cm) with origin at soil surface and directed positively upwards; t = time (d); K = unsaturated hydraulic conductivity (cm d^{-1}); C = differential moisture capacity (cm^{-1}); and S = sink term (cm^3 cm^{-3} d^{-1}).

To solve eqn. (1) for specific initial and boundary conditions a functional form for $S(\psi)$ must be defined. Feddes et al. (1978) described this function as:

$$S(\psi) = \alpha(\psi)\, S_{max} \quad (2)$$

where $\alpha(\psi)$ is a dimensionless prescribed reduction function and S_{max} is the maximum possible root water extraction for the governing meteorological conditions. Feddes et al. (1978) assumed a homogeneous distribution of S_{max} with depth, according to:

$$S_{max} = \frac{E_{pot}}{D_r} \tag{3}$$

where E_{pot} is the potential transpiration rate (cm d^{-1}), and D_r is the root zone depth (cm).

Under nonoptimal conditions, i.e., when the soil is either too dry or too wet, S_{max} is reduced by the ψ-dependent α-function. The shape of the Feddes reduction function is shown in Fig. 1. Root water uptake is limited between ψ_1 and ψ_2 by oxygen deficiency, and between ψ_3 and ψ_4 by decreased water availability. For ψ-values between ψ_2 and ψ_3 root water uptake is optimal. The ψ_3-value is dependent on the evaporative demand of the atmosphere and varies linearly with E_{pot}.

Proper calibration of the sink term requires parameter values for ψ_1, ψ_2, ψ_3-0.5 (ψ_3 for $E_{pot} = 0.5$ cm d^{-1}), ψ_3-0.1 (ψ_3 for $E_{pot} = 0.1$ cm d^{-1}), and ψ_4. Usually these parameter values are determined using soil physical data of only one site. In case eqn. (2) is applied to a scale as large as a watershed, we believe that the spatial variation of the soil properties should be taken into account.

As an alternative, the reduction function from the model MUST (De Laat, 1985) was also implemented in the SWATRE-model. Differences between this reduction function and the one described by Feddes et al. (1978) are twofold. Firstly, α does not depend only on ψ, but also on the hydraulic conductivity of the root zone. Secondly, even for very dry soil moisture conditions, the actual transpiration will never be less than one third of E_{pot} ($\alpha \geq 1/3$). An example of the MUST reduction function is depicted in Fig. 2.

In this study, measured and computer simulated cumulative transpiration values over a growing season were compared for the watershed Hupselse Beek. The Hupselse Beek catchment is situated in the eastern part of the Netherlands

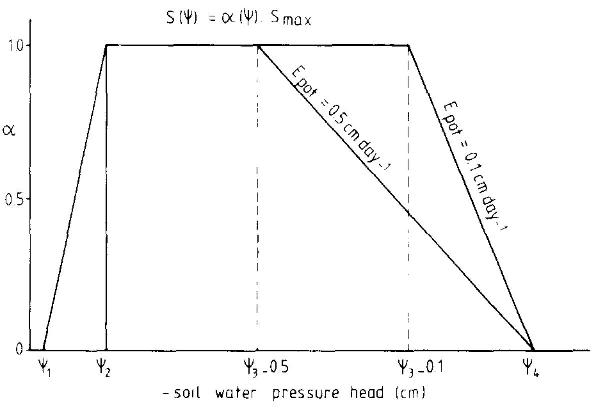

Fig. 1. General shape of the Feddes reduction function $\alpha(\psi)$.

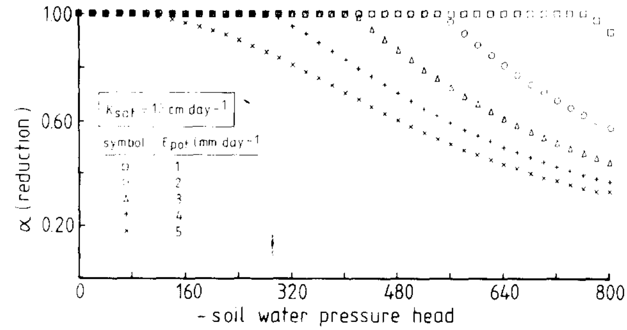

Fig. 2. General shape of MUST reduction function $\alpha(\psi)$ for a surface soil with a value of 10 cm d^{-1} for the saturated hydraulic conductivity.

and covers approximately 650 ha. Underlain by water-tight Miocene clay, the valley is ideal for hydrological studies. During the Pleistocene period part of the area was covered by boulder clay, while gullies eroded during that same period were filled with coarse fluvioglacial sands. At the end of the Pleistocene the whole area was covered with aeolian sands. All these soil deposits occur close enough to the present soil surface to affect the movement of water in the unsaturated zone. Groundwater is within 1.5 m from the soil surface most of the year and influences crop growth by capillary rise. Approximately 70% of the watershed is permanently grown with grass.

Actual transpiration (*EA*) was determined from an energy balance and mass balance of the watershed. In a previous study (Van der Graaf and Feddes, 1984) actual transpiration was also calculated using the SWATRE model from measured daily E_{pot}-values, assuming a rootzone thickness of 30 cm, and applying the sink term described in eqns. (2) and (3) with the Feddes reduction function. The parameter values $\psi_1 = 10$, $\psi_2 = 25$, ψ_3-0.5 = 200, ψ_3-0.1 = 800, and $\psi_4 = 8000$ cm were determined from comparison of measured and simulated *EA* at the meteorological station (meteo site).

However, instead of using this deterministic approach, the parameters of the sink term in the present study were determined in a stochastic way. In this approach, the one-dimensional SWATRE-model is used to simulate unsaturated soil-water transport a number of times, thereby generating spatially variable soil hydraulic properties in each simulation. The use of repeated simulations with a deterministic model in combination with randomly generated input data is called Monte Carlo (MC)-analysis. A MC-analysis results in a probability density function (PDF) of each of the output variables, such as for cumulative actual transpiration (*EA*) and the cumulative reduction in plant transpiration (*RED*).

The determination and description of the stochastic input for the MC-analysis will be discussed only in general terms, since the emphasis of this study is to describe a procedure for calibrating the sink term in spatially variable soils. A detailed analysis and discussion with respect to the stochastic input can be found in Hopmans and Stricker (1988).

The variability of the soil hydraulic properties in the Hupselse Beek was determined by scaling the measured hydraulic data (Warrick et al., 1977). Using the scaling technique, the pattern of spatial variability can be described by a set of scale factors, relating the soil hydraulic properties at each measured location to a representative mean or reference curve. Application of the similar media concept (Miller and Miller, 1956) to soils with different microscopic characteristic lengths connects variation in the soil water retention curve to the hydraulic conductivity function by the scale factor value. Methods to determine scale factors were presented by Warrick et al. (1977), Russo and Bresler (1980), and Hopmans (1987).

Soil hydraulic properties were measured at 20 locations within the watershed (Hopmans and Stricker, 1987). One of the conclusions of this study was that scale factors as determined from water retention data compared fairly well with scale factors calculated from conductivity data alone. Therefore, the scale factor derived from the water retention data were assumed to be sufficient to describe the variability of the soil hydraulic properties. A soil survey of the watershed (Wösten et al., 1985) in combination with the hydraulic data indicated that dominant soil types could be physically characterized by a A and BC horizon, both with a sandy or loamy sand texture, and a D-horizon consisting of a sandy or silty clay. Therefore, scaling was applied to each of these horizons independently. The mean and standard deviation of each such set of scale factors, together with the type of PDF (normal or lognormal) and the reference water retention and conductivity curves parameterized by the Van Genuchten (1978) model, then serve as stochastic input for the MC-simulations. The reference water retention and hydraulic conductivity curves for the three described horizons are shown in Figs. 3a and b, respectively. Scale factor values for each of these three horizons were lognormally distributed with a standard deviation for the A, BC, and D-horizon of 0.3376, 0.2541, and 0.2142, respectively.

Apart from daily values of E_{pot} and precipitation, the soil hydraulic properties and the sink term, SWATRE also requires a lower boundary condition. A lower boundary condition which will partly be influenced by the soil hydraulic properties is a relationship between groundwater level (h) at a specific location and the watershed discharge (q), or:

$$q = ae^{b|h|} \tag{4}$$

where a and b are parameters. Equation (4) was used as a flux boundary condition, where the change in groundwater level is calculated from the mass balance of the soil profile (Belmans et al., 1983).

Since groundwater levels exhibited a large variation, also the $q(h)$-relation in eqn. (4) was scaled for each of the 83 groundwater level observation points in the watershed. The scaling resulted in a reference $q(h)$-relation and a set of scale factors relating the reference curve to each individual curve. From the PDF with its mean and standard deviation, a spatially variable lower boundary condition was defined and was used in combination with the randomly variable

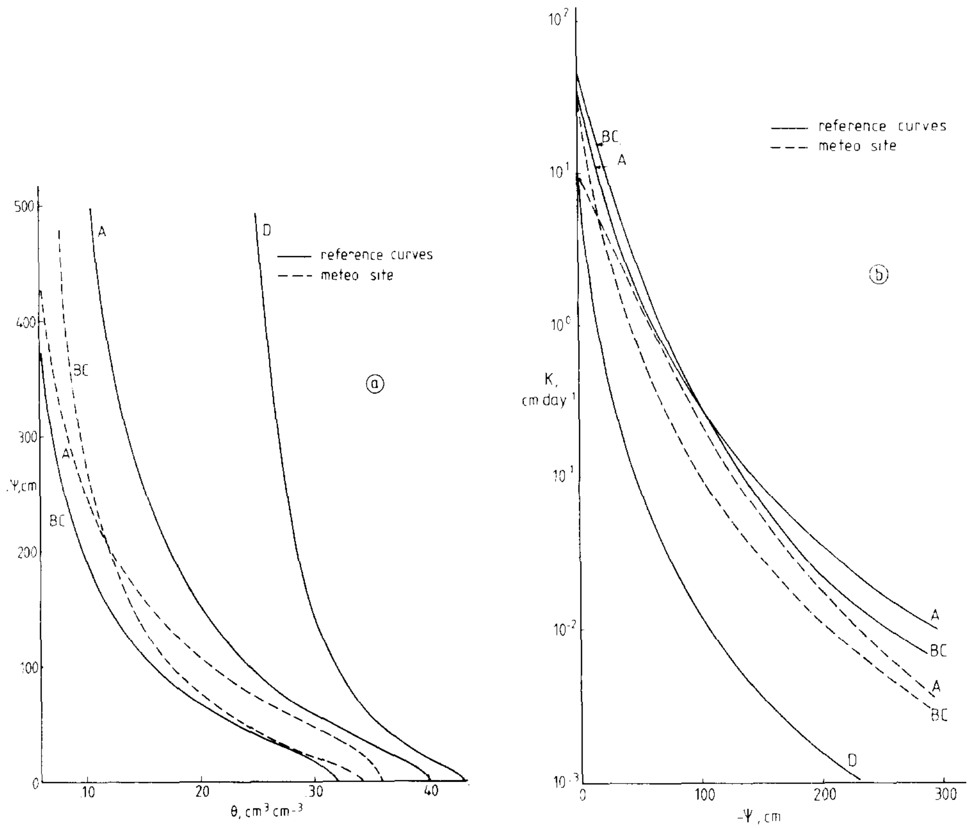

Fig. 3. Water retention curves (a) for A, BC, and D horizon of reference soil (solid lines) and meteo site (dashed lines). Hydraulic conductivity curves (b) for A, BC, and D horizon of reference soil (solid lines) and meteo site (dashed lines).

soil hydraulic properties. Again, for more specific information with regard to the implementation of the variable lower boundary condition one is referred to Hopmans (1987).

All simulations started April 1 and ended October 1. It was assumed that the soil profile was in hydraulic equilibrium at the start of the simulations, i.e., the initial soil water pressure head profile was calculated from the groundwater level. Variations in soil hydraulic properties and lower boundary condition were simulated through generation of scale factors from the statistical properties of both random processes, while an ensemble of realizations was obtained by MC-analysis.

Since thickness of the BC-horizon varies with the starting depth to clay (D-horizon), the Hupsel watershed was mapped into five regions with different clay depths. Thus each region was characterized by a different soil profile type. A MC-analysis was carried out for each soil profile type, after which a PDF is obtained for a number of output variables of each profile type. Output variables of special interest with respect to this study were cumulative *EA* and *RED*, and the groundwater level at the end of the growing season (*GWL1*). A MC-analysis was performed for two years, 1976 and 1982. In Dutch weather conditions, 1982 is a fairly dry year while 1976 was extremely dry.

The parameters of the Feddes reduction function (ψ_3-0.5, ψ_3-0.1, and ψ_4) were optimized by trial and error. For various combinations of ψ_3 and ψ_4, ensemble averages of *EA* were compared with measured *EA* for the two simulated years, thereby obtaining a root water uptake model that is valid for an area with spatial variable soils and which does not depend on the choice of one particular calibration site.

RESULTS AND DISCUSSION

In an earlier study (Hopmans and Stricker, 1988) the soil water regime in the Hupsel watershed was investigated using the MC-analysis described. However, instead of using the Feddes reduction function shown in Fig. 1, the ψ-dependent root water uptake was described by the reduction function used in the unsaturated–saturated water flow model MUST (De Laat, 1985). It should be pointed out, however, that both reduction functions were calibrated such that the measured and simulated *EA* were approximately equal for the soil profile of the meteo site.

The MC-analysis results using the MUST reduction function are shown in Table 1 (from Hopmans and Stricker, 1988). Table 1 shows that *EA*-values are

TABLE 1

Results MC-analysis 1982 and 1976 using MUST reduction function

Soil profile type	Starting depth to clay (cm)	*EA* (mm)		*RED* (mm)		*GWL1* (cm)	
		μ	σ	μ	σ	μ	σ
1982 (E_{pot} = 443.8 mm, precipitation = 254.3 mm)							
1 (34)*	No clay	428.8	15.8	15.0	15.8	202.5	34.7
2 (12)	140	428.8	16.7	15.0	16.7	198.6	17.9
3 (27)	100	430.6	15.3	13.2	15.3	185.3	19.5
4 (23)	60	419.9	16.9	23.9	16.9	169.6	19.3
5 (4)	30	381.4	13.5	62.4	13.5	150.0	27.0
Weighted average		425.3		18.5		188.0	
Measured		419.8		24.0		174.0	
1976 (E_{pot} = 504.4 mm, precipitation = 213.9 mm)							
1 (34)	No clay	429.8	36.2	74.6	36.2	206.6	28.6
2 (12)	140	420.3	42.4	84.1	42.4	207.2	24.4
3 (27)	100	440.6	24.1	63.8	24.1	194.7	19.7
4 (23)	60	391.9	27.3	112.5	27.3	179.1	24.2
5 (4)	30	346.3	21.0	158.1	21.0	163.6	28.3
Weighted average		419.5		84.9		195.4	
Measured		369.5		134.9		191.2	

* Areal percentage.

not affected by the presence of clay if the latter occurs at a depth greater than 1.0 m below the soil surface. Decreasing *EA*-values and thus increasing *RED*-values were calculated for the simulations of profile types 4 and 5. As is expected, the mean groundwater level at the end of the growing season (*GWL1*) decreases with a decreasing depth to clay. With clay higher in the soil profile, less water will be transported upwards and, therefore, the fall in groundwater table is also less.

In contrast to the simulation results of 1982, the weighted average *EA* in 1976 (419.5 mm) was considerably higher than its measured value of 369.5 mm. The discrepancy may be caused by the calibration procedure of the $\alpha(\psi)$-function. The reduction function was determined such that for the groundwater levels and soil hydraulic properties measured at the meteo site, simulated and measured *EA* were approximately equal. However, a soil profile with different soil physical characteristics will generally lead to different *EA*-values. Since it was concluded from the water balance study and energy balance measurements that the measured *EA* was valid for the whole watershed, the calibration of $\alpha(\psi)$ would only be correct if the soil physical properties of the calibration site were also representative. This, however, is doubtful, and it may therefore be more appropriate to calibrate $\alpha(\psi)$ while taking into consideration the variation in soil hydraulic properties and groundwater levels.

Since the Feddes reduction function has far less parameters than the MUST reduction function, we verified the above by optimizing the parameters of the Feddes reduction function so that the weighted averages for *EA* would become approximately equal to the measured values for both years. For that, we assumed that ψ_1 and ψ_2 have values equal as were used before (i.e., $\psi_1 = -10$ and $\psi_2 = -25$ cm). Since both years are dry to very dry, the influences of wetness and oxygen deficiency in the root zone were of minor importance.

The mean and standard deviation of *EA* and *RED* of both years and for soil profile type 2 with different choices of ψ_3-0.5, ψ_3-0.1 and ψ_4 are shown in Table 2. To reduce the number of required simulations, only those for this profile type were carried out. Thus, when combining the results for all five soil profile types after optimizing the reduction function parameters the weighted average *EA* will be less than shown in Table 2. In Table 2 the parameter values of 200, 800, and 8000 were originally obtained by calibration at the meteo site. Clearly, those parameter values are much larger than used for our optimization. Also listed in Table 2 are the *EA*-values when using the MUST reduction function. These values are fairly close to those obtained with the Feddes reduction function, using the originally determined parameter values. The initial estimates of ψ_3-0.5, ψ_3-0.1, and ψ_4 (Fig. 1 and Table 2) were derived from the reference water retention curves of the A-horizon in Fig. 2. At a soil water pressure head of -125 cm the root zone has lost already almost 50% of the total available water. Thus for $\psi_3 = 125$ cm root water uptake will be less than optimal for $E_{pot} \geq 0.5\,\mathrm{cm\,d^{-1}}$, when the volumetric water content (θ_v) in the root zone is less than half of the saturated water content ($0.40\,\mathrm{cm^3\,cm^{-3}}$). For

TABLE 2

Comparison of *EA* and *RED* values as obtained from MC-analysis using different trial values for the parameters of Feddes reduction function; statistics were calculated from 50 simulations for soil profile type 2 (starting depth to clay: 140 cm)

Coefficients			1976				1982			
			EA		*RED*		*EA*		*RED*	
ψ_3-0.5	ψ_3-0.1	ψ_4	μ	σ	μ	σ	μ	σ	μ	σ
200	800	8000	447.9	36.5	56.5	36.5	427.5	22.6	16.3	22.6
125	250	600	399.2	33.4	105.2	33.4	406.9	23.5	36.8	23.5
150	300	750	409.6	42.2	94.8	42.2	411.1	22.6	32.7	22.6
100	400	800	403.0	37.3	101.4	37.3	414.9	24.9	28.9	24.9
75	425	800	415.5	32.2	88.9	32.2	411.1	25.6	32.7	25.6
measured			369.5		134.9		419.8		24.0	
$\alpha(\psi)$-MUST			439.8	31.5	64.6	31.5	428.8	15.8	15.0	15.8

$E_{pot} = 0.1\,\mathrm{cm\,d}^{-1}$ reduction will occur if $\theta_v \leq 0.15$ at ψ_3-0.1 = 250 cm and finally if $\theta_v \leq 0.09$ at ψ_4 = 600 cm root water uptake is zero for any E_{pot}value.

Other trial values for the reduction function parameters were chosen to obtain better agreement between simulated and measured *EA* for both years. To reach the objective, we needed to find parameters such that the simulated *EA* was severely reduced in 1976, but only slightly so in 1982. Note, however, that the choice of parameter values was limited. For a trial set of parameter values, reduction in 1976 will also lead to reduction for 1982. Based on the results of profile type 2, values for ψ_3-0.5, ψ_3-0.1, and ψ_4 were set to 100, 400, and 800, respectively, and using these values also the MC-simulations for the four other profile types were carried out. The results are shown in Table 3. The weighted average *EA*-values while applying the two reduction functions are compared with the measured *EA* in Table 4. Clearly, we have only partly succeeded in matching simulated with measured *EA*. After optimizing the reduction function parameters, *EA* in 1976 improved from 419.5 to 394.4 mm, but is still too large when compared with the 369.5 mm measured. For 1982 the optimized reduction function led to an *EA*-value of 404.9 mm, as compared to the weighted average using the MUST reduction function of 425.3 mm and the 419.8 mm measured (Table 4).

To demonstrate differences in soil hydraulic properties between the calibration site (meteo site) and a representative profile, we have shown in Table 5 simulation results for the soil of the meteo site and for a soil characterized by the scaled mean or reference curves of a A-BC soil profile. For both years values for *EA*, *RED*, and *GWL1* are listed using the MUST reduction function and the Feddes reduction function with the old and new parameters. The water retention and conductivity functions of the horizons for the meteo site are combined with those of the reference curves in Figs. 3a and b.

Clearly, the soil hydraulic properties of the meteo site create less favourable

TABLE 3

Results MC-analysis 1982 and 1976 using Feddes reduction function with optimized parameter values

Soil profile type	Starting depth to clay (cm)	*EA* (mm)		*RED* (mm)		*GWL1* (cm)	
		μ	σ	μ	σ	μ	σ
1982							
1	No clay	406.9	26.9	36.9	26.9	196.3	30.9
2	140	413.5	26.4	30.3	26.4	190.8	20.2
3	100	414.4	18.8	29.4	18.8	181.3	16.9
4	60	397.3	24.5	46.5	24.5	167.3	16.8
5	30	342.8	24.1	101.0	24.1	154.7	20.9
Weighted average		404.9		38.9		183.3	
Measured		419.8		24.0		174.0	
1976							
1	No clay	411.3	34.9	93.1	34.9	205.3	34.2
2	140	406.8	34.2	97.6	34.2	204.0	23.5
3	100	396.6	39.4	107.8	39.4	200.0	23.7
4	60	373.8	23.8	130.6	23.8	175.7	19.5
5	30	317.1	18.0	187.3	18.0	159.2	19.5
Weighted average		394.4		110.0		195.1	
Measured		369.5		134.9		191.2	

TABLE 4

Comparison of weighted average *EA* using MUST and optimized Feddes reduction function with measured *EA*

Reduction function	Weighted average *EA* (mm)	
	1976	1982
MUST	419.5	425.3
Optimized Feddes	394.4	404.9
Measured	369.5	419.8

growing conditions than the soil with the scaled mean hydraulic functions. Comparison of the A-horizon water retention curve of the reference soil with the meteo site (Fig. 3a) shows that in dry conditions (ψ small) less water is available for root water uptake at the meteo site (compare dashed with solid curve). Also, the conductivity curves for the meteo site lie below those of the reference curves (Fig. 3b), indicating that the soil of the meteo site restricts upward water movement towards the root zone more than the reference soil.

Thus, when applying a single site calibration method the choice of the meteo station as a calibration site for the root water uptake model will yield a simulated *EA* at most other locations, that is higher than the measured *EA*. If

TABLE 5

Comparison of the various applied reduction functions for meteo site and soil profile type 1 with reference soil hydraulic functions

	1976			1982		
	EA	*RED*	*GWL1*	*EA*	*RED*	*GWL1*
Meteo site						
MUST	392.1	112.3	178.2	412.9	30.9	160.8
Feddes (200–800–8000)	398.2	106.2	171.0	433.6	10.2	168.0
Feddes (100–400–800)	395.5	108.9	171.0	410.7	33.1	162.0
Reference curves (soil profile type 1)						
MUST	441.2	63.2	199.0	435.8	8.0	184.0
Feddes (200–800–8000)	453.1	51.3	202.7	442.5	1.3	186.4
Feddes (100–400–800)	427.6	76.8	197.0	429.3	14.5	181.0

alternatively a representative profile is chosen to match simulated with measured *EA*, roughly half of the simulations at locations other than the representative calibration site, as also the meteo site, will result in a calculated *EA* lower than the measured *EA*.

From Table 5 it is further more clear that the optimized Feddes reduction function results only in little improvement when the simulated *EA* with the other reduction functions was already low. Table 5 again shows that the simulation results using the MUST reduction function are in fairly good agreement with those when applying the Feddes reduction function with the old parameters. In fact, the difference in *EA* between simulations with the old and optimized Feddes reduction function is even larger than between simulations with the MUST and optimized Feddes reduction function.

CONCLUSIONS

In this paper we discussed the application of a procedure to calibrate a root water uptake model in unsaturated water flow modeling, thereby taking into consideration the variability of soil hydraulic properties. In previous studies the reduction function was calibrated at only one site and it was assumed that the resulting sink term is representative for the whole study area. However, in general large areas show an increasing soil variability and may require a different approach.

A Monte-Carlo analysis in combination with a trial-and-error procedure to optimize the parameters of the reduction function resulted in improved values for the actual transpiration when compared with the single site calibration. However, there was still a considerable deviation from measured transpiration if the proposed analysis was applied to the extremely dry growing season of 1976.

Various reasons can be conceived to explain such deviation. Firstly, it must be emphasized that although the MC-analysis incorporates soil variability, it does not treat other sources of variation as in precipitation and cropping pattern, and is as such not a complete realistic analysis. The simulation model assumes that roots remain active in water uptake, while it has been observed that plants died in 1976 because of drought stress. Secondly, although the Feddes reduction function is simple in its use and implementation, other or more sophisticated models may be required to simulate root water uptake under soil moisture conditions as limited as during part of the 1976 growing season.

It was demonstrated that the site chosen for the calibration of the reduction function (meteo site) behaved quite differently than most other sites within the study area. The hydraulic functions of the soil of the meteo site are highly unfavourable for optimal growing conditions in dry periods. The soil of the meteo site is, therefore, not representative for the study area Hupselse Beek, and should not be used to calibrate a root water uptake model.

Moreover, it has been shown that the Feddes reduction function with the old parameters gave approximately the same results as when using the MUST reduction function.

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