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Parameterizing the soil – water – plant root system

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Abstract

Root water uptake is described from the local scale, to the field scale and to the regional and global scales. Locally, water uptake can be considered at two different Darcian scales, referred to as the mesoscopic and the macroscopic scales.

At the *local mesoscopic scale*, root water uptake is represented by a flux across the soil–root interface, resulting from the mesoscopic Richards equation describing the flow of water from soil to plant root, supplemented by appropriate initial and boundary conditions. The mesoscopic model involves two characteristic lengths describing the root-soil geometry, and two characteristic times describing, respectively, the capillary flow of water from soil to plant roots and the ratio of supply of water in the soil and demand by plant roots. Generally, at a certain critical time, uptake will switch from plant-atmosphere demand-driven to soil supply-dependent. The resulting expressions for the evolution of the average water content can be used as a basis for upscaling from the mesoscopic to the macroscopic scale.

At the *local macroscopic scale*, the root water uptake is represented by a sink term in the macroscopic Richards equation. Reduction of water uptake due to water and salinity stresses is incorporated by either linear or non-linear response functions. The root water uptake is strongly related to root length density, but it is easier to obtain root mass density, and therefore a conversion relationship has been established between the bulk root mass and root length densities, both in space and in time. The local macroscopic model can be incorporated in Soil–Plant–Atmosphere Continuum (SPAC) numerical models, like the SWAP, HYSWASOR, HYDRUS, ENVIRO-GRO and FUSSIM models. These SPAC models in turn can be used for upscaling, first to the *field scale* and from there to *the regional and global scales*. As Global Climate Models (GCMs) show a strong sensitivity to continental evaporation, closer root water-uptake modeling might improve soil vegetation control instead of uncontrolled continental evaporation.

It is concluded that the relationships between mesoscopic and macroscopic descriptions of water uptake provide a useful framework for interpreting laboratory and field data and formulating macroscopic sink terms. One-dimensional analysis of the root-zone water balance is now well developed and has strong computational capabilities. Joint stresses can be computed easily by multiplication of water stress with salinity stress. The implementation at the macroscopic scale of the mesoscopic

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description of simultaneous uptake of water and solutes is promising. Present Soil–Vegetation–Atmosphere Transfer (SVAT) schemes should be improved with respect to root water-uptake descriptions, using existing global data sets of root and soil properties. A first priority is to establish firmly the relationships of root biomass, rooting depth, root distribution and root functions with land-use type, soil type, soil texture, topography and climate.

Introduction

Water for evapotranspiration from land surfaces is supplied mainly by the soil. The soil water reservoir balances the episodic excesses of water supply from rainfall against the more smoothly varying atmospheric demand for evapotranspiration. The role of soil water within the soil–plant–atmosphere system depends on the size of the soil water reservoir and the availability of water in that reservoir which, in turn, depends on the texture, structure and organic matter of the soil and the characteristics of the root system.

One key function of plant roots is their ability to link the soil, where water and nutrients reside, to the organs and tissues of the plant, where these resources are used. Hence roots serve to connect the soil environment to the atmosphere by providing a link in the pathway for water fluxes from the soil through the plant to the atmosphere. Fluxes along the soil–plant–atmosphere continuum are regulated by above-ground plant properties like the leaf stomata, which can regulate plant transpiration when interacting with the atmosphere, and plant root-system properties like depth, distribution and activity of roots, as well as soil properties like the soil water retention and hydraulic conductivity characteristics.

Plant root systems show a remarkable ability to adapt to soil depth and to changes in the availability of water and nutrients and the chemical properties (e.g. salinity) in soils. Root response to soil properties, in turn, affects the uptake of soil water and nutrients and the storage of carbon below ground. Root distribution may change when ecosystems respond to greenhouse warming and carbon-dioxide fertilization. For example, at higher atmospheric CO_2 concentrations, stomata of plants can contract somewhat for a given influx of CO_2. Transpiration thus decreases and, coupled with generally higher photosynthesis at higher atmospheric CO_2 concentration, water-use efficiency can increase dramatically (Field, Jackson and Mooney 1995). Increased water-use efficiency will potentially feed back to changes in root characteristics (Chaudhuri, Kirkham and Kanemasu 1990), with the possibility of further substantial changes in the water and energy balances. Exploration of such feedbacks has only begun.

Generally one is interested in water movement and solute transport at local, field and regional to global scales, with appropriate models for each scale (see Raats, Smiles and Warrick 2002):

- The local scale (approximately 1 m^2) refers to several scales: the *microscale* of solid particles and pores and of plant roots at the tissue and cellular level, the *mesoscale* of soil structural elements and of individual roots and associated volumes of soil, and the *macroscale* of the soil profile, including the entire root system;
- The field scale (approximately hectares) is the scale of farm management, i.e. of maintaining an optimal soil structure, adjusting the availability of water and nutrients, applying pesticides and herbicides, etc.;

- The regional to global scale (larger than 1 km^2) is the scale at which e.g. pollution of ground and surface water becomes evident, but which also is of interest in land surface modeling from the perspective of the hydrological cycle and climate.

Temporal scales range from below hourly in fundamental studies, to diurnal and seasonal in agricultural applications, to annual and decadal in climate studies.

Within this context, one objective of this position paper is to explore the level of detail that needs to be included to parameterize properly models of water and energy flux at the local, field, regional and global scales. In particular, we will investigate whether the concept of a single one-dimensional (1D) root water-uptake model with effective parameters is feasible for application at regional scales. Toward this goal we also examine the existing databases on plant rooting depth and distribution, and some of the key models that can use these data.

Liquid water flow inside the plant: soil and plant resistances

A simple *electric analogue* can be used to describe water flow through the entire soil–plant–atmosphere system (Van den Honert 1948; Kirkham 2002). This model assumes that the water-flux density v (cm^3 cm^{-2} s^{-1}) through the rooted soil zone and the root–stem–leaf–stomata path is proportional to the total water potential, i.e. head difference Δh_{total} (cm) and inversely proportional to the total resistance R_{total} (expressed in seconds) of the system. Thus considering a *one-dimensional steady-state flow in a series network*, the liquid flow equation is:

$$T = v = -\frac{\Delta h_{\text{total}}}{R_{\text{total}}} = -\frac{\left(h_{\text{root}} - h_{\text{soil}}\right)}{R_{\text{soil}}} = -\frac{\left(h_{\text{leaf}} - h_{\text{root}}\right)}{R_{\text{plant}}} = \frac{h_{\text{leaf}} - h_{\text{soil}}}{R_{\text{soil}} + R_{\text{plant}}} \tag{1}$$

where T (cm s^{-1}) is the transpiration rate, h_{soil}, h_{root} and h_{leaf} (cm) are pressure heads in the soil, at the root surface and in the leaves, respectively, R_{soil} and R_{plant} (s) are liquid-flow resistances of the soil and the plant. Hence R_{plant} does not include stomatal resistance. When the transpiration demand of the atmosphere on the plant system is high or when the soil is rather dry, R_{soil} and R_{plant} influence h_{leaf} in such a way that transpiration is reduced by closure of the stomata. Eq. (1) can be applied to the root system as a whole by measuring T, h_{soil}, and h_{leaf} during two periods, thus obtaining two equations with two unknowns, from which R_{soil} and R_{plant} can be computed. The relative magnitude of R_{soil} and R_{plant} is an important object of study. Under wet conditions R_{soil} is close to zero. Generally one can state, except for very dry soil, that $R_{\text{plant}} > R_{\text{soil}}$. Most of the *plant resistance is concentrated in the roots*, to a lesser extent in the leaves, and a minor part in the xylem vessels, although in some monocotyledonous crops longitudinal resistance may hamper the uptake of water from deeper layers (Wind 1955; Richards and Passioura 1989). Feddes and Rijtema (1972, Table 5) have concluded that R_{plant} increases with progressive drying of the soil and decreases when the transpiration rate is higher. This finding (see Figure 1) has been confirmed by e.g. Hansen (1974). It must be emphasized that this is a controversial issue as other researchers think that R_{plant} depends neither on soil water content nor on transpiration rate.

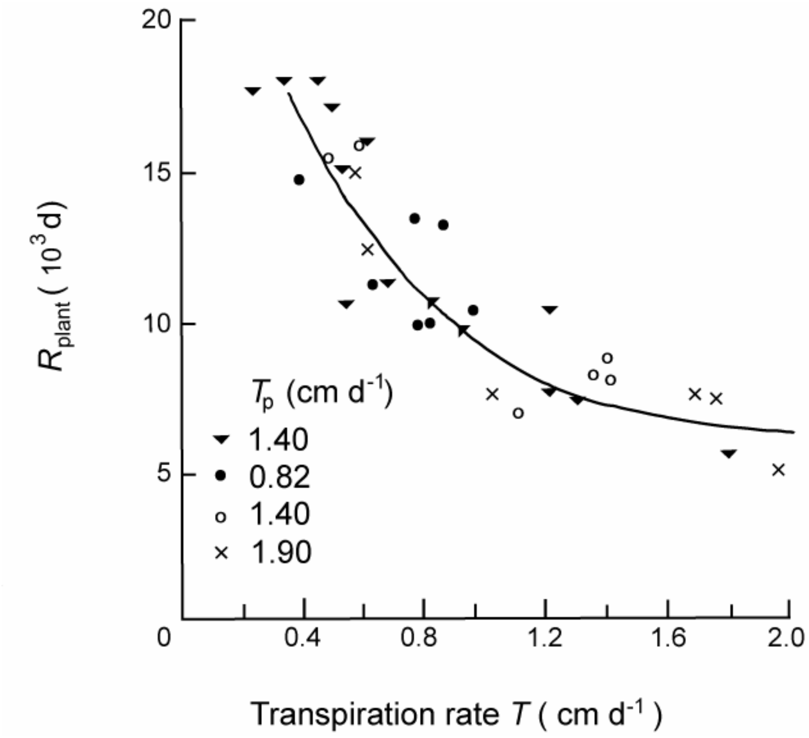

Figure 1. Plant resistance R_{plant} versus transpiration rate T as computed from an Italian-ryegrass column experiment, showing the inverse proportionality of R_{plant} with T (after Hansen 1974)

About 30 years ago it was already realized that the uptake of water is complicated by the presence of solutes in the water. On the one hand solutes tend to be sieved out by root membranes; on the other hand they may be actively taken up by the plant. Inspired by the partly analogous problem of simultaneous flow of water and solutes in clays (Bolt and Groenevelt 1969), Dalton, Maggio and Piccinni (2000) and Fiscus (Fiscus 1975) formulated a model for simultaneous uptake of water and solutes by plant roots, accounting both for salt sieving and active uptake of salt. Dalton, Maggio and Piccinni (Dalton, Maggio and Piccinni 2000) found qualitative agreement between calculations based on these equations and a variety of observations on simultaneous uptake of water and solutes. In particular *the theory accounts for the often observed nonlinear relationship between the volume flux of the aqueous solution and the salt concentration difference between the soil–root interface and the xylem.* In recent years the model has been used in mechanistic studies of salt tolerance (e.g. Dalton, Maggio and Piccinni 2000) and water and nutrient management of greenhouse crops (Van Ieperen 1996). Extensions of the theory are possible, including a description of the accumulation of solutes at the soil–root interface and the polarity of transport processes associated with series arrays of membranes (Raats in prep.).

Macroscopic description of root water uptake

The more hydrologically oriented approach at the macroscale regards the root system as a diffuse sink that penetrates each layer of soil uniformly, though not necessarily with a constant strength throughout the root zone. Root water uptake can then be represented as a sink term that is added to the vertical water-flow equation. One has to realize, however, that 1-D root-system models may fail when lateral transport of water by subsurface or overland flow occurs. In case of catchments with complex sloping terrain and groundwater tables, a vertical flow model has to be coupled with either a process or a statistically based scheme that incorporates lateral water transfer.

The macroscopic way of solving the root water-uptake problem is to introduce in the differential form of the water mass balance a sink term representing water extraction by plant roots:

$$\frac{\partial \theta}{\partial t} = -\frac{\partial v_z}{\partial z} - S_z \tag{2}$$

where z (cm) is the vertical co-ordinate taken positively upward, v_z (cm d^{-1}) is the Darcian soil water-flux density taken positively upward, and S_z (cm^3 cm^{-3} d^{-1}) is the root water-uptake rate. S_z thus depends on depth, time, soil water pressure head, root density or a combination of these variables.

The volumetric flux v_z is given by Darcy's law:

$$v_z = -k(h)\left(\frac{\partial h}{\partial z} + 1\right) \tag{3}$$

where k is hydraulic conductivity (cm d^{-1}) and h is soil water pressure head (cm). As written, (2) and (3) refer to a one-dimensional, homogeneous system, but these limitations are easily removed, particularly in the transition to numerical formulations.

Root water uptake S_z generally can be derived by applying the *water-conservation equation* to a given volume of soil. Commonly it is assumed that, in an unsaturated soil, water flows only in the vertical direction z. Let us consider a volume of soil of unit cross-sectional area in the horizontal plane, bearing vegetation, with a lower boundary at $z = 0$ (for example at the groundwater table or a level with constant pressure head) and an upper boundary at the soil surface at $z = z$ (where z is both the vertical co-ordinate and the position of the soil surface). Water flow through the roots ($\bar{v}_{\text{roots}}$) can then be calculated as the **measured total flow through soil and roots (**$\bar{v}_{\text{total}}$) diminished with the **calculated flow through the soil (**$\bar{v}_{\text{soil}}$). Based on (1) the water-conservation equation can then be written for every height z and for a short time interval t_2 - t_1 as

$$\begin{aligned}\int_0^z \bar{S}_z\,\mathrm{d}z &= \bar{v}_0 - \frac{\int_0^z \left[\theta(t_2) - \theta(t_1)\right]\mathrm{d}z}{t_2 - t_1} - \bar{v}_z \\ \bar{v}_{\text{roots}} &= \bar{v}_{\text{total}} - \bar{v}_{\text{soil}}\end{aligned} \tag{4}$$

where $\bar{S}_z$ (cm^3 cm^{-3} d^{-1}) is the time-averaged volume of water taken up by the roots per unit bulk volume of soil in unit time and considered positive from the soil into the roots, $\bar{v}_0$ (cm d^{-1}) is the time-averaged volumetric flux density through the lower boundary, θ (cm^3 cm^{-3}) is the bulk volumetric soil water content, $\bar{v}_z$ (cm d^{-1}) the time-averaged volumetric flux density through the upper boundary.

For the calculation of root water-uptake patterns, one needs to know the $k(h)$ curve and the profiles of θ and h. The latter two can be measured by well-known methods such as time domain reflectometry and soil water tensiometry. When the moisture retention curve is used, it suffices to measure either θ or h and infer the unknown one via the retention curve. The flux $\overline{v_0}$ at the bottom should be derived from either measurements (e.g. a lysimeter) or by means of (3). In some experiments use was made of a zero-flux plane (e.g. Richards, Gardner and Ogata 1956; Cooper 1979; 1980). Application of (4) gives the integrated root water uptake over a given depth interval in the root zone. Differentiation of (4) gives the water uptake rate by the roots at height z. A sample of calculation of water extraction by roots at depth z is given for red cabbage on clay for the period of 18 through 25 July 1967 (Feddes 1971).

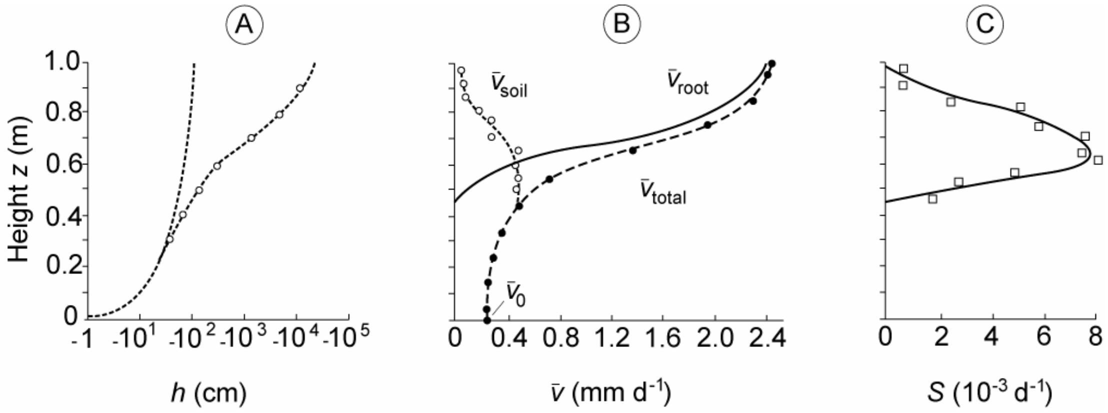

Figure 2. Profiles of time-averaged pressure head h (A); time-averaged cumulative water withdrawal of both a cabbage crop and clay soil $\overline{v}_{\mathrm{total}}$, of soil only $\overline{v}_{\mathrm{soil}}$ and of roots only $\overline{v}_{\mathrm{root}}$ (B); time-averaged extraction rate S (C); for red cabbage on clay for the period of 18 through 25 July 1967 (after Feddes 1971)

The mean pressure head during this period is given in Figure 2A, the time-averaged cumulative withdrawal patterns of both crop and soil $\overline{v}_{\mathrm{total}}$, soil only $\overline{v}_{\mathrm{soil}}$, and crop only $\overline{v}_{\mathrm{roots}}$ are presented in Figure 2B, while the time-averaged extraction rate at heights z, $S(z)$, is given in Figure 2C. From Figures 2B and 2C it follows that the effective rooting depth can be found at the height where the cumulative withdrawal function of total flow and soil flow intercept. From Figure 2C it can be seen that the maximum extraction rate occurs at a height of 0.30 to 0.35 m below the soil surface.

The root extraction rate is generally small near the top of the profile, unless the soil is wet just after rainfall or irrigation. Downwards, the extraction rate increases to a certain maximum and then decreases to zero at the bottom of the root zone.

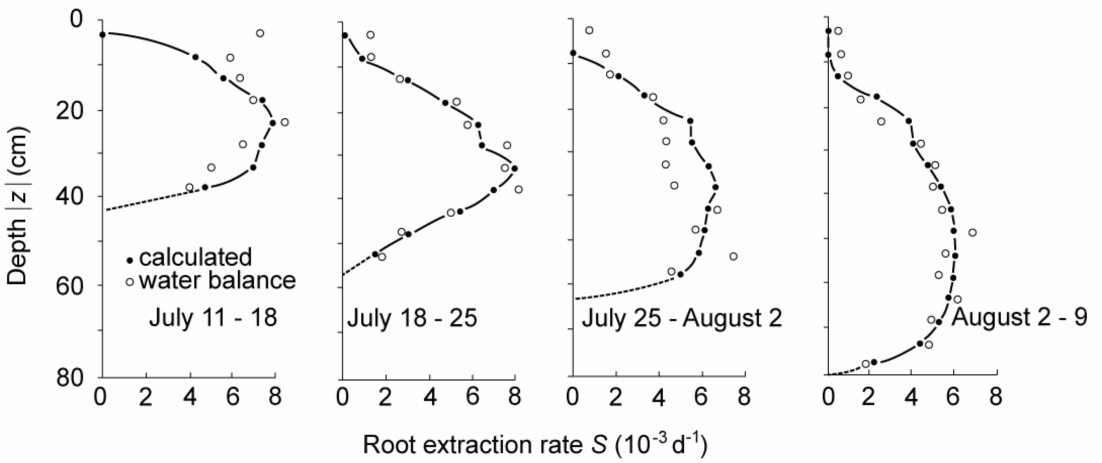

Figure 3. Example of measured variations of root water uptake with depth and time of red cabbage grown on a clay soil with the groundwater table at 90-110 cm depth and obtained from water-balance studies over 4 consecutive weeks (after Feddes 1971)

As the soil dries (Figure 3) the zone of maximum root water uptake moves from smaller to larger depths in dynamic correspondence with the downward progression of roots into deeper moist soil. The maximum extraction rate appears to depend on the demand the atmosphere exerts on the plant system, on the depth to which the roots penetrate, and on the soil water pressure-head distribution. Later water uptake from the upper layers becomes relatively less important. Most of the water is absorbed from the zone of low tensions near the groundwater table. Thus, a relatively small part of the root system can be responsible for most of the plant water uptake.

Modeling root water uptake at mesoscopic and macroscopic scales

Introduction

For overviews of root water-uptake models the reader may refer to Feddes (1981), Molz (1981), Hopmans and Bristow (2002). Uptake of water by plant roots can be considered at two different Darcian scales, referred to as the *mesoscopic* and *macroscopic* scales (Raats 1990; 2002; Raats, Smiles and Warrick 2002). At the *macroscopic* scale, the uptake of water by plant roots is represented by a sink term in the volumetric mass balance Eq. (2).

At the *mesoscopic* scale, this same uptake is represented by a flux across the soil–root interface. That flux is a consequence of the interaction of processes in the soil and in the plant. In the soil, flow of water towards or away from individual plant roots may be described by a nonlinear diffusion equation, subject to appropriate initial and boundary conditions. The mesoscopic model involves at least *two characteristic lengths* describing the root–soil geometry and *two characteristic times* describing, respectively, the capillary flow of water from soil to plant roots and the ratio of supply of water in the soil and uptake by plant roots. Generally, at a certain critical time, uptake will switch from demand-driven to supply-dependent. The resulting mesoscopic expressions for the evolution of the average water content can be used as a basis for upscaling to the macroscopic scale.

The mesoscopic analysis considers the convergent radial flow of soil water toward and into a representative individual root, taken to be a line or narrow-tube sink uniform along its length, i.e. of constant and definable thickness and absorptive properties. The macroscopic root system as a whole can then be described as a set of such individual roots, assumed to be regularly spaced in the soil at specified distances, that may vary within the soil profile. The Richards equation for flow of water at the mesoscale is solved for the distribution of soil water pressure heads, water contents, and fluxes from the root outward. In cylindrical co-ordinates, the pressure head (h), water content (θ), and matric flux potential $\left(\varphi = \int k \mathrm{d}h = \int D \mathrm{d}\theta\right)$ forms of the Richards equation are:

$$\frac{\partial\theta(h)}{\partial t} = \frac{1}{r}\frac{\partial}{\partial r}\left\{rk(h)\frac{\partial h}{\partial r}\right\}, \quad \frac{\partial\theta}{\partial t} = \frac{1}{r}\frac{\partial}{\partial r}\left\{rD(\theta)\frac{\partial\theta}{\partial r}\right\}, \quad \frac{\partial\theta(\varphi)}{\partial t} = \frac{1}{r}\frac{\partial}{\partial r}\left\{r\frac{\partial\varphi}{\partial r}\right\} \quad (5)$$

where r is the radial coordinate from the center of the root.

Geometrical and mass-based parameters describing root systems

Depending on the scale of interest and the purpose, a variety of parameters have been introduced to describe aspects of the spatial–temporal distribution of plant roots. In the following we give definitions of some commonly used parameters and relationships among these. In particular, we describe relationships among the *geometrical* parameters used in soil–water–plant–atmosphere models and relate these to the *root mass*-based parameters often measured by ecologists.

For an individual plant root in soil, two characteristic lengths can be identified (Figure 4):

- r_0 (cm), radius of the plant root and also the internal radius of the equivalent cylindrical shell of soil associated with the plant root;
- r_1 (cm), external radius of the equivalent cylindrical shell of soil associated with the plant root.

The following parameters can be used to characterize the root system in the soil:

- L_r (cm cm^{-3} = cm^{-2}), the root length per unit volume of soil;
- σ_r (cm^2 cm^{-3} = cm^{-1}), the root surface area per unit volume of soil;
- ϕ_r (cm^3 cm^{-3} = -) and ϕ_s (cm^3 cm^{-3} = -), the root and soil volume fractions, being subject to the constraint:

$$\phi_\mathrm{r} + \phi_\mathrm{s} = 1 \quad (6)$$

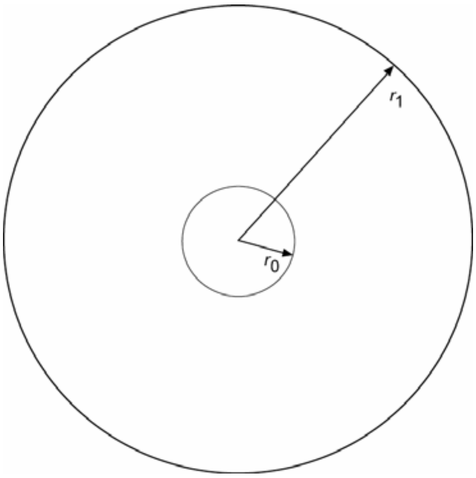

Figure 4. Two characteristic lengths for an individual plant root: the internal radius of the plant root r_0 and the external radius of the soil associated with the plant root r_1

For uniformly distributed roots, the root length density is the inverse of the volume of root plus soil, associated with unit length of root:

$$L_\mathrm{r} = \left(\pi r_1^2\right)^{-1} \tag{7}$$

The root volume fraction ϕ_r is equal to the ratio of the volume of a unit length of root and the volume of this unit length of root plus the cylindrical shell of soil associated with it:

$$\phi_\mathrm{r} = \left(r_0/r_1\right)^2 = \pi r_0^2 L_\mathrm{r} \tag{8}$$

The constraint (6) on the volume fractions implies

$$\phi_\mathrm{s} = 1 - \phi_\mathrm{r} = 1 - \left(r_0/r_1\right)^2 = 1 - \pi r_0^2 L_\mathrm{r} \tag{9}$$

The specific root surface area σ_r is the ratio of the surface area of a unit length of root divided by the volume of root plus the cylindrical shell of soil associated with unit length of root:

$$\sigma_\mathrm{r} = 2r_0/r_1^2 = 2\pi r_0 L_\mathrm{r} = 2\left(\pi\phi_\mathrm{r} L_\mathrm{r}\right)^{1/2} \tag{10}$$

From (7) and (10) it follows that:

$$r_0 = \sigma_\mathrm{r}\left(2\pi L_\mathrm{r}\right)^{-1} \qquad r_1 = \left(\pi L_\mathrm{r}\right)^{-1/2} = \left(\frac{2r_0}{\sigma_\mathrm{r}}\right)^{1/2} \tag{11}$$

Clearly, *the characterization of a uniform root system by the radii* r_0 *and* r_1 *is equivalent to the characterization by the root length density* L_r *and the specific root surface area* σ_r. Solving (8) for r_0/r_1:

$$r_0/r_1 = \phi_\mathrm{r}^{1/2} = \left(1 - \phi_\mathrm{s}\right)^{1/2} \tag{12}$$

Thus, to characterize the *geometry of a uniform root system*, either the radii r_0 and r_1 or two of the parameters L_r, σ_r, and ϕ_r or ϕ_s can be used. *In experimental practice, the root length density, and the root radius* r_0 *are usually measured.* Then (7) can be used to calculate r_1 and (9), (8), and (9) can in turn be used to calculate σ, and ϕ_r or ϕ_s.

The nature of all the parameters just introduced is geometrical. The *bulk root mass density* ρ_r i.e. the mass of roots per unit volume bulk soil, can be obtained by multiplying the root volume fraction ϕ_r by the intrinsic root density γ_r, i.e. the mass of roots per unit volume of root:

$$\rho_\mathrm{r} = \gamma_\mathrm{r}\phi_\mathrm{r} = \gamma_\mathrm{r}\left(r_0/r_1\right)^2 = \gamma_\mathrm{r}\pi r_0^2 L_\mathrm{r} \tag{13}$$

A complicating factor related to the intrinsic density γ_r is the often considerable gaseous volume fraction of root tissue, in particular in plants adapted to growing in wet soils. Eq. (13) relates the geometrical parameters root radius r_0 and root length per unit volume L_r, used in water-uptake models, to the bulk root mass volumetric density ρ_r, often measured by ecologists. From this it follows that root-system databases for use in specific types of soil–water–plant–atmosphere models should contain, in addition to the spatial–temporal distribution of $\rho_\mathrm{r}(z,t)$, some information on the intrinsic density γ_r and the root radii r_0. Mathematical expressions and databases for $\rho_\mathrm{r}(z,t)$ are discussed later.

Mesoscopic analysis of uptake by a single root and its macroscopic implications

To get some insight into the acquisition of water from soil, both Philip (1957b) and Gardner (1960) considered uptake of water by a plant root with radius r_0 surrounded by a cylindrical shell of soil with outer radius r_1. Linearizing (5) and assuming $r_0 << r_1$, they considered an approximate solution of (5) subject to a constant flux into a line sink (see Raats, Smiles and Warrick 2002).

Philip (1957b) used this solution to analyse the time course of the water content θ_0 at the soil–root interface, in particular the instant t_w at which this water content reaches the critical value θ_{0w} at which the plant wilts. He demonstrated that the average water content $\bar{\theta}_w$ at wilting does not only depend on θ_{0w}, but also on the soil water diffusivity D, the geometrical parameters r_0 and r_1, and the water demand u. Thus, from such an analysis at the mesoscale, he concluded that “uncritical use of the ‘wilting point’ as an invariant index of the lower limit of the availability of soil moisture to plants can be very misleading”.

Denmead and Shaw (1962) verified the predictions that water was not equally available to plants in the range between field capacity and the permanent wilting point, and that transpiration could be restricted and plants could wilt over a wide range of soil water contents, depending on root density, the soil hydraulic properties, and last but not least, on the transpiration demand of the atmosphere.

Tanner (1967, his Figure 5) nicely summarized other data available at that time relating the ratio of actual and potential evapotranspiration to the amount of water in the soil.

Transient soil water depletion regarded as a succession of *steady-state* profiles

Gardner (1960) used the line sink solution to calculate water-depletion patterns around individual roots. But more importantly, he also used it as a point of departure for the formulation of a simpler model, in which the depletion resulting from uptake by a single root is treated as a series of steady flows in the cylindrical shell of soil surrounding the root, with the soil–root interface at the inner edge and the water coming from the outer edge. This simple model has been used ever since as a point of departure for more sophisticated mesoscopic as well as macroscopic models of water uptake.

In the single-root model of Gardner (1960), the root is viewed as a cylinder of uniform radius r_0 and infinite length having uniform water-absorbing properties. For *steady-state conditions* $(\partial\theta/\partial t = 0)$ in the soil shell surrounding the root with water flowing from the outer cylindrical surface at $r = r_1$ to the inner cylindrical soil–root interface at $r = r_0$, the solution of $(5)_1$ under the assumption of constant k gives the following expression for the flux q_r at the soil–root interface *at the mesoscale*:

$$q_\mathrm{r} = Bk\left(h_1 - h_0\right) \tag{14}$$

where q_r (cm^3 cm^{-1} d^{-1}) is the rate of water uptake per unit length of root, h_1 (cm) is pressure head of the soil at $r = r_1$, h_0 (cm) is the pressure head at the soil–root interface, and the dimensionless geometric and root distribution factor B is given by:

$$B = \frac{2\pi}{\ln\left(r_1/r_0\right)} \tag{15}$$

In hydrology, an equation analogous to (14) is used for the steady-state flux towards a well per unit length of well.

Cowan (1965) realized that the assumption of constant k used in deriving (14) can be avoided. In essence, he replaced (14) by:

$$q_\mathrm{r} = B\overline{k}\left(h_1 - h_0\right) = B\overline{D}\left(\theta_1 - \theta_0\right) = B\left(\varphi_1 - \varphi_0\right) \tag{16}$$

where the average hydraulic conductivity $\overline{k}$ and average diffusivity $\overline{D}$ are defined by:

$$\overline{k} = \frac{\left(\int_{h_0}^{h_1} k\,\mathrm{d}h\right)}{\left(h_1 - h_0\right)}, \qquad \overline{D} = \frac{\left(\int_{\theta_0}^{\theta_1} D\,\mathrm{d}\theta\right)}{\left(\theta_1 - \theta_0\right)} \tag{17}$$

The integrals in (17) can be easily evaluated for any of the commonly used expressions to represent the $k(h)$ and $D\left(\theta\right)$ relationships for specific classes of soils.

At the *macroscale*, for a discrete soil layer of thickness Δz and with root length density L_r (cm cm^{-3}), the water-uptake rate Δv_r (cm d^{-1}) can be written as:

$$\Delta v_\mathrm{r} = L_\mathrm{r} \Delta z q_\mathrm{r} \tag{18}$$

Introducing Eq. $(16)_1$ in (18) gives:

$$\Delta v_\mathrm{r} = BL_\mathrm{r} \Delta z \overline{k}\left(h_1 - h_0\right) \tag{19}$$

In an attempt to account for the location of the soil layer, one might be tempted to replace h_0 in (19) by $h_\mathrm{rootsystem} - z$, where $h_\mathrm{rootsystem}$ is the pressure head in the root system evaluated at the soil surface. However, as will become apparent from an analysis of the hydraulic connection between the soil–root interface and the xylem discussed later on, this cannot be justified.

Transient soil water depletion as a succession of *steady-rate* profiles

An attractive alternative, to the depletion approach as a succession of steady state profiles, is to regard the withdrawal of water to be uniformly distributed over the cylindrical shell of soil. In fact, if the Richards equation were linear and the water demand were constant, there would be a solid basis for this at intermediate times when the memory of the details of the initial spatial distribution of the water content has already disappeared and the constant demand can still be met.

Assuming *steady-rate conditions* $\left(\partial\theta/\partial t = \text{constant}\right)$ and with zero flux at the outer cylindrical surface at $r = r_1$ of the soil shell surrounding the root, the solution of (5) for the flux q_r at the soil–root interface is (cf. Jacobsen 1974):

$$q_\text{r} = B^*\overline{k}\left(h_1 - h_0\right) = B^*\overline{D}\left(\theta_1 - \theta_0\right) = B^*\left(\varphi_1 - \varphi_0\right) \tag{20}$$

where the dimensionless geometric and root distribution factor B^* is given by:

$$B^* = \frac{2\pi}{\left[1-(r_0/r_1)\right]^{-1}\ln\left(r_1/r_0\right) - 1/2} \tag{21}$$

The corresponding steady-rate mesoscopic expression for Δv_r is:

$$\Delta v_\text{r} = B^* L_\text{r}\Delta z\overline{k}\left(h_1 - h_0\right) \tag{22}$$

For $r_0 << r_1$, the factor B^* can be approximated by:

$$B^* \approx \frac{2\pi}{\ln\left(r_1/r_0\right) - 1/2} \approx \frac{2\pi}{\ln\left(0.6064\,r_1/r_0\right)} \tag{23}$$

Comparison of the expressions for B in (15) and for B^* in (23) shows that whereas in the *steady-state* approximation all the water is assumed to originate from $r = r_1$, in the *steady-rate* approximation it seemingly originates from $r = 0.6064\,r_1$.

The steady-rate feature has long been known as an important term in the full solutions of linear equations subject to flux boundary conditions for heat conduction and flow to wells. Cowan (1965) demonstrated its relevance for flow to plant roots, even for a diurnally varying, but on average constant demand. Passioura and Cowan (1968) compared the predictions of the water content at the soil–root interface based on the approximate steady-state and steady-rate solutions and the exact numerical solution, finding reasonable agreement among them, with the steady-rate method being somewhat more accurate than the steady-state one.

The steady-rate approximation has been extensively used by De Willigen and Van Noordwijk (1987; 1995), De Willigen et al. (2000), Heinen (1997), and Heinen and De Willigen (1998; 2001) not only for uptake of water, but also for the uptake of nutrients. They have also considered extensions to cases where the flow to the individual roots is not radially symmetric, either due to partial soil–root contact or due to spatially non-uniform root distributions.

Constant and falling rate phases of uptake

If the Richards equation were linear, the distribution of the water content around an individual root resulting from a constant rate of uptake would be the sum of two parts:

- an exponentially fading infinite series with characteristic diffusion time $t_d = r_1^2 / D$, which guarantees that the initial condition is satisfied;
- a water-content profile of time-invariant form, with the water content decreasing linearly with time everywhere at the same rate, corresponding to the steady-rate profile discussed above.

The deviation from this simple structure will depend on the degree of nonlinearity. In any case, linear or nonlinear, the *constant-rate phase* can only last up to the instant that the water content at the soil–root interface reaches a critical value θ_{0w} and the *falling-rate phase* starts. If the Richards equation were linear, the solution in the falling-rate phase would again be the sum of two parts:

- an exponentially fading infinite series with characteristic diffusion time $t_d = r_1^2 / D$, which guarantees that the intermediate condition at the transition from the constant-rate phase to the falling-rate phase is satisfied;
- a term that has the form of the product of two terms, one depending on the radial co-ordinate and one depending exponentially on time, again involving the characteristic diffusion time $t_d = r_1^2 / D$.

The change from the *constant-rate phase* to the *falling-rate phase* is basically a change from the *plant-dominated* boundary condition to the *soil-dominated* boundary condition. Similar switches of boundary conditions are well known in other environmental physics problems. For example, evaporation from a saturated bare soil, starting at a constant rate being determined only by atmospheric conditions, may at a certain instant switch to a falling-rate phase in which the evaporation rate depends only on the hydraulic properties of the soil (Philip 1957a). For infiltration of water into soil, an analogous switch occurs at the so-called ponding time (see e.g. Broadbridge and White 1988).

Hydraulic connection between the soil–root interface and the xylem

Taylor and Klepper (1975) used a steady-state single-root model to interpret experimental data for uptake of water by cotton root systems. Specifically, they used the following equation of the same form as (14) and (16):

$$q_\mathrm{r} = B_\text{soil-rootsystem}\, k_\text{soil-rootsystem} \left(h_1 - h_\text{xylem} \right) \tag{24}$$

where h_xylem is a value obtained from shoot water-potential measurements, $k_\text{soil-rootsystem}$ is the hydraulic conductivity of the combined soil–root pathway, and the dimensionless geometric and root distribution factor $B_\text{soil-rootsystem}$ is given by:

$$B_\text{soil-rootsystem} = \frac{2\pi}{\ln\left(r_1 / r_\text{stele} \right)} \tag{25}$$

where r_stele is the radius of the root stele, which includes all of the tissues inside the cortex. The analysis of the data gave $k_\text{soil-rootsystem}$ as a function the water content θ.

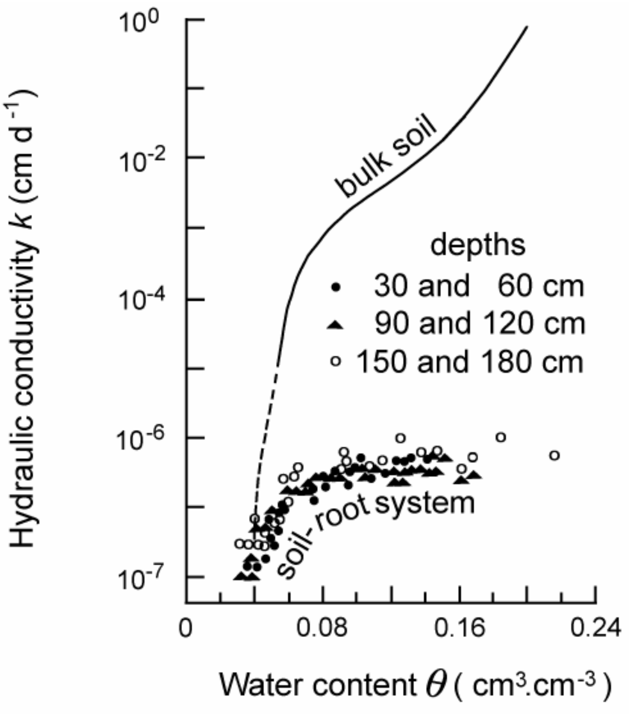

Figure 5. A comparison of the hydraulic conductivity of a soil–cotton root system obtained at various depths with the soil hydraulic conductivity (*k*) for various soil moisture contents (after Taylor and Klepper 1975)

Figure 5 shows that in the wet range $k_{\text{soil-rootsystem}}(\theta)$ is up to six orders of magnitude smaller than $k(\theta)$ for just the soil. This strongly suggests that most of the resistance resides in the plant root, except when the soil is very dry and $k_{\text{soil-rootsystem}}$ becomes comparable to $\bar{k}$ given by $(17)_1$.

Herkelrath, Miller and Gardner (1977a) investigated the influence of θ and h upon root water uptake in a laboratory experiment with winter wheat. They used wax layers, penetrable by roots but not by water, to divide their soil column in different sections. Their observations imply a large resistance between the soil–root interface and the xylem and a small resistance between the xylem and the leaves. Water uptake began to decrease rapidly when θ fell below about 0.10, corresponding to about $h = -100$ cm, in agreement with the data of Taylor and Klepper (1975) for cotton.

To account for their observations, Herkelrath, Miller and Gardner (1977b) proposed to complement the steady-state models (16-19) or the steady-rate models (20-23) with the following expression to describe the flow from the soil–root interface to the xylem:

$$q_{\mathrm{r}} = C\frac{\theta_0}{\theta_s}\left(h_0 - h_{\text{xylem}}\right) \tag{26}$$

where $C(\mathrm{s}^{-1})$ is the conductance of the region *between the soil and the xylem*. The degree of saturation θ_0/θ_s of the soil at the soil–root interface is a factor accounting for the contact between the soil and the root. In view of the θ_0 proportionality in (26), it is interesting to note that for $\theta > 0.03$, Taylor and Klepper (1975) found a more or less linear increase of the rate of uptake with θ. Eq. (26) is similar to an equation commonly used to describe the steady flow of water across a surface crust into the underlying soil where the water is at a uniform water content and pressure head and steadily flows downward under the influence of gravity. Raats (1974) described a simple graphical procedure to analyse such steady infiltration into a crusted soil. Van Noordwijk (1983; see also De Willigen and Van Noordwijk 1987; and Raats 1990) used a similar procedure to analyse the implications of (26). For a critical value of $h_{\text{xylem}} = h_{\text{xylemw}}$ at which wilting starts, it follows from (17) that:

$$h_{0\text{w}} = h_{\text{xylemw}} + \frac{\theta_{\text{s}}}{\theta_{0\text{w}}} \frac{q_{\text{r}}}{C} \tag{27}$$

The graphical or equivalent iterative procedure consists of reducing the seemingly infinity of possible pairs $(h_{0\text{w}}, \theta_{0\text{w}})$ to a single pair by determining the intersection of (27) with the soil water-retention curve:

$$h_{0\text{w}} = h(\theta_{0\text{w}}) \tag{28}$$

The root contact model (26) has been used extensively by Jensen et al. (Jensen, Henson and Hansen 1990; 1993). The concept of contact resistance is also relevant for heavier textured soils in which the roots have a strong tendency to grow at the surface of aggregates.

In view of (18), (22) and (26) the flow from the soil to the xylem can be described by

$$\Delta v_{\text{r}} = B^* \bar{k} L_r \Delta z \left(h_1 - h_0\right) = C \frac{\theta_0}{\theta_s} L_r \Delta z \left(h_0 - h_{\text{xylem}}\right) \tag{29a}$$

Assuming that the soil conductance is large in comparison with the root conductance, i.e. $B^* \bar{k} L_r \Delta z >> C\left(\theta_0/\theta_s\right) L_r \Delta z$ or $B^* \bar{k} >> C(\theta_0/\theta_s)$, then the water status around the individual roots is nearly uniform, i.e. $h_0 \approx \bar{h} \approx h_1$ and $\theta_0 \approx \bar{\theta} \approx \theta_1$, where $\bar{h}$ is the pressure head and $\bar{\theta}$ is the water content at the macroscopic scale. Eq. (26) may then be replaced by

$$\Delta v_{\text{r}} = C \frac{\bar{\theta}}{\theta_s} \left(\bar{h} - h_{\text{xylem}}\right) \tag{29b}$$

If the xylem conductance is large in comparison with the root conductance $C\left(\theta_0/\theta_s\right) L_r \Delta z$, so that the xylem pressure head h_{xylem} at depth z can be written as the sum of the root pressure head h_{r} evaluated at the soil surface and the gravitational head z, i.e. $h_{\text{xylem}} = h_{\text{r}} + z$, then (26) reduces to:

$$\Delta v_{\text{root}} = C \frac{\bar{\theta}}{\theta_s} L_r \Delta z \left(\bar{h} - h_{\text{root}} + z\right) \tag{30}$$

Clearly, (30) should only be used if the soil resistance and the xylem resistance are negligible.

An alternative method for dealing with limited soil–root contact was explored by De Willigen and Van Noordwijk (1987, Chapter 10). The fraction of the potentially available water that can be acquired by the root system at a certain rate decreases with decreasing soil–root contact and increasing spatial clustering of roots. Veen et al. (1992) showed that limited soil–root contact leads to a decrease of uptake of water and nitrate, explaining the existence of an optimum in the response curve of crops to soil compaction.

Rappoldt (1992) developed a practical method for the description of transport processes at the mesoscale in *aggregated soils*. The method is based on a simplification of the complex geometry of natural structural soils, in such a manner that the essential geometric and structural characteristics governing the diffusion process are preserved. This is done as follows. The geometry of the natural system is replaced by a model system of cylinders. The distribution of the radii of the cylinders is chosen such that the structured soil and the model system have the same distance distribution. Finally calculations of diffusion processes are carried out in the model system. On this basis, Rappoldt (1992) analysed in detail various aspects of aeration of heterogeneous soils (see also Rappoldt 1990; Rappoldt and Verhagen 1999). To illustrate the generality of the method, he also used the method to analyse the uptake of water and nutrients by an irregularly distributed root system (Rappoldt 1992, Chapter 7).

Hydrologically oriented modeling approach for root water uptake at macroscale

Introduction

We now return to the *macroscopic* description of root water uptake based on the mass-balance equation (2) and the Darcy equation (3). Combination of (2) and (3) results in the Richards equation:

$$\frac{\partial \theta}{\partial t} = C(h)\frac{\partial h}{\partial t} = \frac{\partial\left[k(h)\left(\frac{\partial h}{\partial z}+1\right)\right]}{\partial z} - S(z,t) \qquad (31)$$

where $C = \mathrm{d}\theta/\mathrm{d}h$ (cm^{-1}) is the differential water capacity, i.e. the slope of the soil water characteristic. Van Genuchten (1980) has provided analytical expressions for the non-linear soil hydraulic characteristics $\theta(h)$ and $K(h)$. To obtain a solution of (31) one has to supplement it with conditions for the initial situation and for the upper and lower boundary of the flow system. A major difficulty in solving (31) for $\theta(z,t)$ stems from the unknown form of the function $S(z,t)$. Feddes, Bresler and Neuman (1974) assumed $S(z)$ to be proportional to k, the difference between $h_1 - h_0$ as in (14), and to an empirical root effectiveness function b'. This function is proportional to root mass and varies nearly exponentially with depth. They found from 1-D finite difference simulations that calculated cumulative transpiration was in good agreement with measured field data. Neuman, Feddes and Bresler (1975) extended the 1-D flow expression (31) to 2-D flow for rigid unsaturated or partly saturated soils by means of the finite-element method. Feddes, Neuman and Bresler (1975) showed the capabilities of this method in applications to complex but realistic 2-D flow situations.

At the upper boundary the vegetation plays a dominant role in the partitioning of the various fluxes. Hence, in solving (31) one needs in principle a coupling of the soil water-balance model with a daily vegetation-growth model. Only in this way can a proper prediction of vegetation development and growth dependent on the actual prevailing soil water conditions be obtained, thus assuring the proper feedback.

Reduction of water uptake due to water stress

Under *optimal* water conditions the maximum possible root water extraction rate $S_p(z)$, integrated over the rooting depth, is equal to the *potential transpiration rate*, T_p (cm d^{-1}), which is governed by atmospheric conditions. Feddes et al. (1976) first described S (for each time t) as a function of soil water content θ, and improved this later (Feddes, Kowalik and Zaradny 1978) as a function of soil water pressure head h according to:

$$S(h) = \alpha(h)\, S_p \tag{32}$$

where $\alpha(h)$ is a dimensionless prescribed function of soil water pressure head and S_p is the potential, i.e. maximal possible, water extraction by roots (cm^3 cm^{-3} d^{-1}).

Under *non-optimal conditions*, i.e. either too dry or too wet, S_p is reduced by means of the factor $\alpha(h)$. The shape of this function has been proposed by Feddes, Kowalik and Zaradny (1978) as shown in Figure 6.

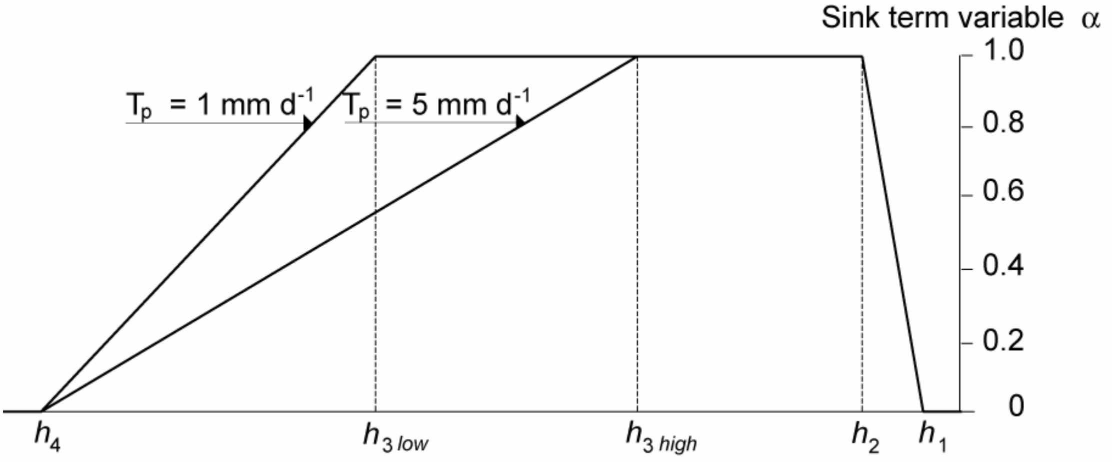

Figure 6. Dimensionless sink term variable α as a function of the soil water pressure head h (after Feddes, Kowalik and Zaradny 1978)

Water uptake below $|h_1|$ (oxygen deficiency, as shown in wheat experiments by Yang and De Jong (1971) and Ehlers (1976) with h_1 approaching zero in sandy soils) and above $|h_4|$ (wilting point) is set equal to zero. Between $|h_2|$ and $|h_3|$ (reduction point) water uptake is maximal. Between $|h_1|$ and $|h_2|$ and between $|h_3|$ and $|h_4|$ a linear variation is assumed. The value of h_3 depends on the water demand of the atmosphere (compare with Denmead and Shaw 1962) and thus varies with T_p. For an indication of these h_1 - h_4 values for 5 crops see Table 1. Kroes and Van Dam (2003, Appendix 3) give high and low estimates of h_3 for a large number of crops, based on a compilation by Taylor and Ashcroft (1972, Table 14.3) of pressure heads at which “water should be applied for maximum yields of various crops grown in deep, well-drained soil that is fertilized and otherwise managed for maximum production”.

Table 1. Critical pressure-head values h (cm) of the sink-term function $\alpha\,(h)$ depicted in Figure 6 for some main agricultural crops (after Wesseling 1991)

Crop	h_1	h_2	$h_{3,\,\mathrm{high}}$	$h_{3,\,\mathrm{low}}$	h_4
Potatoes	-10	-25	-320	-600	-16000
Sugar beet	-10	-25	-320	-600	-16000
Wheat	0	-1	-500	-900	-16000
Pasture	-10	-25	-200	-800	-8000
Corn	-15	-30	-325	-600	-15000*)

*) after Veenhof and McBride (1994)

A more detailed discussion on the modeling of reduction of root water uptake not only by water stress $\alpha\,(h)$, but also by salinity stress $\alpha\,(h_\mathrm{o})$ and by combined water and salinity stress $\alpha\,(h_\mathrm{o})$ will be given later on (see 38 etc.).

As in the past detailed data on root-system properties like root mass, root length density and their variations with depth and time were hardly available, Feddes, Kowalik and Zaradny (1978) assumed a *homogeneous* root distribution over the rooting depth and defined S_p according to Figure 7):

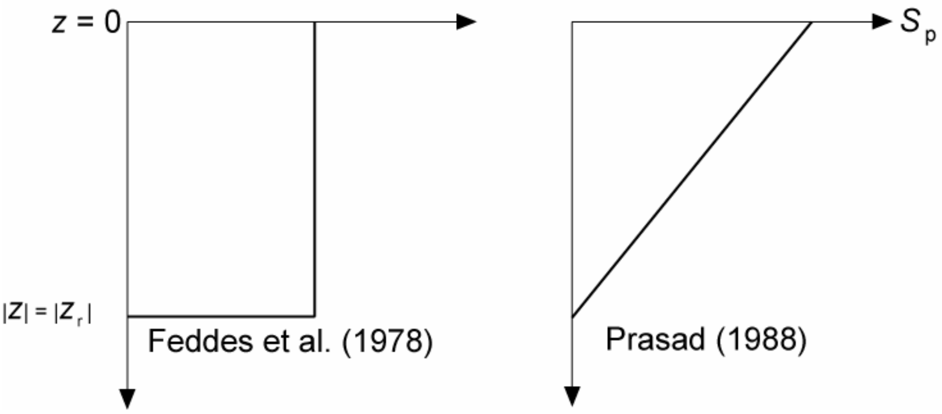

Figure 7. Schematic view of root water uptake under a homogeneous (left) and a heterogeneous (right) root distribution, both under **optimal** soil water conditions

$$S_\mathrm{p} = \frac{T_\mathrm{p}}{|z_\mathrm{r}|} \tag{33}$$

where T_p is the potential transpiration rate (cm d^{-1}) and $|z_\mathrm{r}|$ is the root-zone depth (cm). Hoogland, Feddes and Belmans (1981), Prasad (1988) and Hayhoe and De Jong (1988) took care of the fact that in a moist soil the roots can mainly extract water from the upper soil layers, leaving the deeper layers relatively untouched. Taking root water uptake at the bottom of the root zone (z_r) equal to zero, Prasad’s description reads as (Figure 7):

$$S_{\mathrm{p}}(z)=\frac{2T_{\mathrm{p}}}{|z_{\mathrm{r}}|}\left\{1-\frac{|z|}{|z_{\mathrm{r}}|}\right\} \tag{34}$$

The potential root water extraction rate at a certain depth, $S_{\mathrm{p}}(z)$ (d^{-1}), may *for non-homogeneous root distributions* be determined by either the root mass or the root length density, $L_{\mathrm{r}}(z)$ ($\mathrm{cm\ cm}^{-3}$), as fraction of the total root length density over the rooting depth $|z_{\mathrm{r}}|$ (e.g. Nimah and Hanks 1973)

$$S_{\mathrm{p}}(z)=\frac{L_{\mathrm{r}}(z)}{\int_{z_{\mathrm{r}}}^{0}L_{\mathrm{r}}(z)\,\mathrm{d}z}T_{\mathrm{p}} \tag{35}$$

Assuming that changes in storage of water in the vegetation are negligible, the actual rate of transpiration T at time t is given by the spatial integral of the volumetric rate of water uptake $S(z)$ over the entire profile:

$$T(t)=\int_{z_{\mathrm{r}}}^{0}S(z)\,\mathrm{d}z \tag{36}$$

Root water uptake under minimum energy conditions

Under conditions of high evaporative demand of the atmosphere, i.e high T_{p}, and/or conditions of water and/or salinity stress, limited availability of water in part of the root zone may be compensated by uptake in the remainder at a rate that is higher than would be expected on the basis of the relative root length density distribution only. In that case $L_{\mathrm{r}}(z)/L_{\mathrm{profile}}$ still is a good initial estimate in an iterative procedure to determine all factors in (32) and (35).

Experimental evidence by Dirksen (1985) on alfalfa plants grown in intermediate ranges of salinity and water content, showed that these plants took up water according to distributions that kept the uptake-weighted mean total (= pressure + osmotic) head as high as possible. In other words: plant water uptake occurs with the *minimum amount of work possible*. Hence, if the osmotic head is uniform, plants will extract relatively more water from the wetter than from the drier soil water zones. To account for this plant behavior, a mathematical (numerical) procedure needs to be developed that minimizes the total rate of energy expenditure during root water uptake.

Ghali (1986; 1989) assumes that the entire root zone is divided into n concentric cylinders of very small radial thickness. If the total root extraction from cylinder k is denoted D_k and the corresponding work to be done by plants is W_k, then total volumetric root extraction D_{r} and the total work to be done by plants W_{r} to extract this amount can be expressed by $D_{\mathrm{r}}=\sum_{k=1}^{n}D_k$ and $W_{\mathrm{r}}=\sum_{k=1}^{n}W_k$. According to this concept the *objective is to maximize the state variable* D_{r} *subject to a set of constraints* i.e. Max $D_{\mathrm{r}}=\sum_{k=1}^{n}D_k$ subject to $D_k<D_k^{\max}$ and $W_k=W_k^{\max}$, where $D_k^{\max}=T_{\mathrm{p}}$ from cylinder k and $W_k^{\min}$ is the minimum possible work to be done to extract $D_k^{\max}$. For the constraints above, each of the concentric cylinders is subdivided into n_l finite cylinders stacked in vertical direction. If the rate of extraction from each

subcylinder i is represented by a sink term S_i , then $D_k = \sum_{i=1}^{n_l} S_i \cdot V_i$ where V is the volume of subcylinder i. To satisfy the first constraint, S_i should not exceed an upper value, i.e. $S_i < S_\mathrm{p}$ for $i = 1,2 ......n_l$. The minimization process can finally be expressed by: Min $W_\mathrm{r} = \sum_{i=1}^{n_l} W_i$ subject to $h_i \le W_i \le h_0$, where h_0 is the initial value and $W_i \ge h_\mathrm{cr}$, being the critical pressure head. To complete the minimization process, weights (pressure-head-dependent) must be selected and applied to S_i to ensure that in each zone i the root extraction is proportional to the work to be done.

Adiku et al. (2000) applied a Gardner (1960) type of root uptake model (i.e keeping the pressure head at the root surface constant throughout the root zone), and solved the minimization problem by a so-called dynamic-programming framework. They showed that in uniform wet soils root water extraction patterns follow the root distribution, but when the top soil dries out increased root activity occurs at greater soil depths.

Li, De Jong and Boisvert (2001) pursued a different approach. They modified an exponential root water-uptake model by incorporating an extra parameter, i.e. a weighted-stress index, which accounts for *both* root distribution and soil water stress. This model was represented as a function of T_p, soil water availability and root length density. Simulated soil water contents improved considerably with the new model.

Van Wijk and Bouten (2001) considered actual root distributions as a reflection of the optimization strategy of the plant. They defined ‘optimal root distribution’ of trees as the one that *maximizes* root water uptake over a period of 10 years. They tested this assumption by means of 10-year simulations with a root water-uptake model based on (35) and the function depicted in Figure 6. Preferential water uptake from wetter soil layers was a main factor in the interpretation of the simulated root distribution.

Reduction of water uptake due to combined water and salinity stress

Introduction

For crops subject to water stress, Hanks (1983) found the relative actual/potential yield Y/Y_p to be equal to the relative transpiration T/T_p :

$$\frac{Y}{Y_\mathrm{p}} = \frac{T}{T_\mathrm{p}} \tag{37}$$

Please note that in (37) the magnitudes of T_p and Y_p differ from year to year according to the prevailing meteorological conditions, and that relative yield is related to water availability as well as to salinity stress. The yield reduction due to salinity stress also leads to a reduction of the water uptake. Therefore, water uptake depends on both water availability and salinity stress. Hence, our task is to relate the local water uptake rate on the one hand to the demand at the vegetation scale and on the other hand to the limitations imposed by local water availability and salinity stress.

In analogy with (32 and 33) the local sink strength $S(z)$ at any time t may thus be assumed to be given by

$$S(z) = \alpha\left[h(z), h_{o}(z)\right] S_{p}(z) = \alpha\left[h(z), h_{o}(z)\right] \frac{L_{r}(z)}{\int_{-z_{r}}^{0} L_{r}(z)\partial z} T_{p} \tag{38}$$

where $\alpha\left[h(z), h_{o}(z)\right]$ is the water and salinity stress reduction function which will depend on both the distributions of the water pressure head $h(z,t)$ and the osmotic pressure head $h_0(z,t)$.

In the absence of water and salinity stress anywhere in the root zone, i.e. if $\alpha\left[h(z), h_{o}(z)\right] = 1$, an obvious choice for the potential rate of water-uptake function $S_{p}(z,t)$ is the relative root length density distribution function L_{r}/L_{profile}. It will be shown later on that for L_{r}/L_{profile} various expressions can be obtained from (72-76).

Root water uptake due to water and salinity stress: *linear* reduction functions

To account for the influence of the composition of the soil solution in the macroscopic expressions, Nimah and Hanks (1973) just replaced the pressure head in the soil *h* by the *sum of h and* h_o, being sometimes referred to as an 'additive model'. The basic problem with this model is that it lacks a proper theoretical justification. Feng, Meiri and Letey (2003) write about the Nimah-Hanks model: "…this type of water-uptake function was shown to be insensitive to salinity and generally inadequate to properly evaluate plant water uptake under saline conditions". It is unfortunate that the extrapolation of the Gardner (14) / Cowan (16) model beyond its intended range of application in the Nimah-Hanks model seems to discourage the further development and use of the mesoscopic/macroscopic approach to simultaneous flow of water and transport of solutes in soil–plant root system.

The piecewise linear reduction function of Feddes, Kowalik and Zaradny (1978), Figure 6, is given by:

$$\alpha\left(h, T_{p}\right) = \begin{array}{cc} 0 & h \geq h_{1} \\ \dfrac{\left(h_{1}-h\right)}{\left(h_{1}-h_{2}\right)} & h_{1} > h \geq h_{2} \\ 1 & h_{2} > h \geq h_{3}\left(T_{p}\right) \\ \dfrac{\left(h-h_{4}\right)}{\left(h_{3}\left(T_{p}\right)-h_{4}\right)} & h_{3}\left(T_{p}\right) > h \geq h_{4} \\ 0 & h_{4} > h \end{array} \tag{39}$$

Inserting (39) in (38) yields the root water uptake at the various pressure-head ranges.

The representation of crop salt-tolerance data in terms of relative yields by Maas and Hoffman (1977) has served as a point of departure for describing reduction of water uptake due to salinity stress.

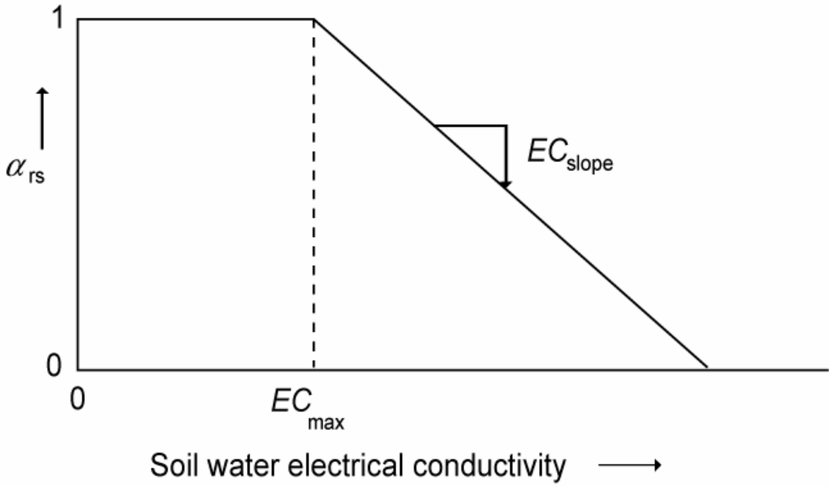

Figure 8. Reduction coefficient for root water uptake, α_{rs}, as function of the electrical conductivity *EC of the soil saturation extract* (after Maas and Hoffman 1977). To convert the *EC*-based slope into an osmotic-head-based slope one may apply a factor 360 (Richards 1954)

The response function can be written in terms of concentration, or electrical conductivity (see Figure 8) of either the soil water or the soil saturation extract, or osmotic pressure head (Maas 1986; 1990; Maas and Grattan 1999; Maas and Hoffman 1977; Van Genuchten and Hoffman 1984). In terms of *osmotic pressure heads*, the piecewise linear yield response function of Maas and Hoffman (1977) is given by:

$$\frac{Y(h_\mathrm{o})}{Y_\mathrm{p}}=\frac{T(h_\mathrm{o})}{T_\mathrm{p}}=\begin{cases}1 & 0\geq h_\mathrm{o}\geq h_\mathrm{ot}\\ 1-\dfrac{(h_\mathrm{ot}-h_\mathrm{o})}{(h_\mathrm{ot}-h_\mathrm{o0})} & h_\mathrm{ot}>h_\mathrm{o}\geq h_\mathrm{o0}\\ 0 & h_\mathrm{o0}>h_\mathrm{o}\end{cases} \tag{40}$$

where $Y(h_\mathrm{o})$ is the yield at the osmotic pressure head h_o, Y_P is the potential yield, h_ot is the threshold value of h_o above which $Y(h_\mathrm{o})=Y_\mathrm{p}$ and h_o0 is the threshold value of h_o below which $Y(h_\mathrm{o})=0$ is zero.

Just as (37) served to justify the water-stress reduction function $\alpha(h,T_p)$ given by (39), the yield response function (40) may serve as the basis for formulating the salinity-stress reduction function $\alpha(h_\mathrm{o})$ for the macroscopic rate of water uptake:

$$\alpha(h_\mathrm{o})=\begin{cases}1 & 0\geq h_\mathrm{o}\geq h_\mathrm{ot}\\ 1-\dfrac{(h_\mathrm{ot}-h_\mathrm{o})}{(h_\mathrm{ot}-h_\mathrm{o0})} & h_\mathrm{ot}>h_\mathrm{o}\geq h_\mathrm{o0}\\ 0 & h_\mathrm{o0}>h_\mathrm{o}\end{cases} \tag{41}$$

Such a correspondence between the yield response and water-uptake reduction functions due to salinity stress were first proposed by Van Genuchten (1987), who, from the start however, emphasized the nonlinear reduction functions that will be discussed in the following subsection.

Root water uptake due to water and salinity stress: *non-linear* reduction functions

Alternative, nonlinear S-shaped and exponential yield response functions due to salinity stress were proposed and tested by Van Genuchten and Hoffman (1984). The S-shaped salinity-stress yield reduction function is given by:

$$\frac{Y(h_\mathrm{o})}{Y_\mathrm{p}} = \frac{1}{1+\left(h/h_{\mathrm{o}0.5}\right)^{p}} \tag{42}$$

where at $h_{\mathrm{o}0.5}$ is the value of h_o at which the yield has declined by 50% and p is an empirical parameter. Re-analysis of the Maas-Hoffman data set showed that for most crops $p = 3$, which gives very satisfactory results (see Van Genuchten and Hoffman 1984; Van Genuchten 1987, viz. in particular the 1984 USSL Annual Report in Appendix A). Van Genuchten and Gupta (1993) found that the S-shaped response function (42) is equally good or better than the piecewise linear equation (41) of Maas and Hoffman.

This nonlinear response function can also be used as the basis for formulating the reduction function $\alpha(h_\mathrm{o})$ due to salinity stress for the macroscopic rate of water uptake. In fact, Van Genuchten (1987) proposed the following pair of S-shaped reduction functions $\alpha(h)$ for water stress and $\alpha(h_\mathrm{o})$ for salinity stress:

$$\alpha(h) = \frac{1}{1+\left(h/h_{0.5}\right)^{p}} \qquad \alpha(h_\mathrm{o}) = \frac{1}{1+\left(h_\mathrm{o}/h_{\mathrm{o}0.5}\right)^{p_\mathrm{o}}} \tag{43}$$

where $h_{0.5}$ and $h_{\mathrm{o}0.5}$ are the values of h and h_o at which $\alpha(h)$ and $\alpha(h_\mathrm{o})$ are reduced to 0.5.

Dirksen and Augustijn (1988) and Dirksen et al. (1993) modified the nonlinear S-shaped functions of Van Genuchten (1987) by introducing values h^* and h_o^* of h and h_o above which there is no reduction of uptake:

$$\alpha(h) = \frac{1}{1+\left(\dfrac{h^*-h}{h^*-h_{0.5}}\right)^{p}} \qquad \alpha(h_\mathrm{o}) = \frac{1}{1+\left(\dfrac{h_\mathrm{o}^*-h_\mathrm{o}}{h_\mathrm{o}^*-h_{\mathrm{o}0.5}}\right)^{p_\mathrm{o}}} \tag{44}$$

Homaee (1999; see also Homaee, Dirksen and Feddes 2002; Homaee, Feddes and Dirksen 2002a) further modified the equation of Dirksen and Augustijn (1988) and Dirksen et al. (1993) by introducing maximum values h_max and $h_{\mathrm{o}\,\mathrm{max}}$ at which the reduction factors reach minimum values:

$$\alpha(h)=\frac{1}{1+\dfrac{1-\alpha(h_{\max})}{\alpha(h_{\max})}\left(\dfrac{h^*-h}{h^*-h_{\max}}\right)^p}\qquad \alpha(h_{\mathrm{o}})=\frac{1}{1+\dfrac{1-\alpha(h_{\mathrm{o}\max})}{\alpha(h_{\mathrm{o}\max})}\left(\dfrac{h_{\mathrm{o}}^*-h_{\mathrm{o}}}{h_{\mathrm{o}}^*-h_{\mathrm{o}\max}}\right)^{p_{\mathrm{o}}}}. \quad (45)$$

The curves $\alpha(h)$ and $\alpha(h_0)$ pass through the cut-off points $[h_{\max}, \alpha(h_{\max})]$ and $[h_{\mathrm{o}\max}, \alpha(h_{\mathrm{o}\max})]$. However, the experimental proof of the existence of this cut-off point is rather weak. Also equations (45) can be rewritten in the forms of (44) with $h_{0.5}$ and $h_{\mathrm{o}0.5}$ defined as:

$$h_{0.5}=\left(h^*-\left(\frac{1-\alpha(h_{\max})}{\alpha(h_{\max})}\right)^{-1/p}\left(h^*-h_{\max}\right)\right)$$

$$h_{\mathrm{o}0.5}=\left(h_{\mathrm{o}}^*-\left(\frac{1-\alpha(h_{\mathrm{o}\max})}{\alpha(h_{\mathrm{o}\max})}\right)^{-1/p}\left(h_{\mathrm{o}}^*-h_{\mathrm{o}\max}\right)\right) \quad (46)$$

In other words, the Eqs. (45) are equivalent to the Eqs. (44) with cut-off points $[h_{\max}, \alpha(h_{\max})]$ and $[h_{\mathrm{o}\max}, \alpha(h_{\mathrm{o}\max})]$.

We now consider *reduction factors that depend on* ***both*** *h and h_{o}*:

$$\alpha=\alpha(h,h_{\mathrm{o}}) \quad (47)$$

In line with early concepts of simultaneous water and salinity stress, Van Genuchten (1987) considered a root water-uptake reduction function with a dependence on a *weighted sum* $(a_1h+a_2h_{\mathrm{o}})$ of h and h_{o}:

$$\alpha(h,h_{\mathrm{o}})=\frac{1}{1+\left\{\dfrac{(a_1h+a_2h_{\mathrm{o}})}{h_{0.5}}\right\}^p} \quad (48)$$

where a_1 and a_2, respectively, are the weighting factors of h and h_{o}, p is an empirical parameter, and $h_{0.5}$ is the value of the hydraulic pressure head h at which $\alpha(h,h_{\mathrm{o}}=0)=0.5$. This reduction function is not separable in the multiplicative form $\alpha\left[h(z,t),h_{\mathrm{o}}(z,t)\right]=\alpha\left[h(z,t)\,\alpha h_{\mathrm{o}}(z,t)\right]$. But it is not simply additive either, as is the Nimah-Hanks model. In Eq. (48), setting $a_1=1$, $1/a_2=a$ and $h_{0.5}/a_2=ah_{0.5}=h_{\mathrm{o}0.5}$ gives:

$$\alpha(h,h_{\mathrm{o}})=\frac{1}{1+\left\{\dfrac{(ah+h_{\mathrm{o}})}{h_{\mathrm{o}0.5}}\right\}^p}=\frac{1}{1+\left\{\dfrac{h}{h_{0.5}}+\dfrac{h_{\mathrm{o}}}{h_{\mathrm{o}0.5}}\right\}^p} \quad (49)$$

where $h_{o0.5}$ is the value of the osmotic pressure head h_o at which $\alpha\left(h=0,h_o\right)=0.5$.

Cardon and Letey (1992a) showed that the combined water and salinity stress reduction function (49) gives more satisfactory results than the Nimah-Hanks model. Based on the salt-tolerance data for corn given by Maas (1986), they assumed $h_{o0.5}=-4300$ cm. Based on data of Ehler (1983), they expected $h_{0.5}$ to be in the range of -2500 to -6500 cm. Since a sensitivity test showed that the simulation results were rather insensitive to the choice of $h_{0.5}$ in the range of -2500 to -6500 cm, Cardon and Letey (1992a) adopted $h_{o0.5}=h_{0.5}=-4300\,\text{cm}$, thus setting $a=1$ in (49). The same values were used by Cardon and Letey (1992b) in a test of the model using field data for sweet corn reported by Shalhevet, Vinten and Meiri (1986) for alfalfa. Cardon and Letey (1992c) evaluated a greenhouse experiment. Again based on Maas (1986), they adopted $h_{o0.5}=-6400$ cm.

According to (49) the water and salinity stress reduction function α is less than unity, except for $h=h_o=0$. Seemingly inspired by Feddes et al. (1976) and the Maas-Hoffman yield response data, Pang and Letey (1998) introduced threshold values h_t and h_{ot} of h and h_o above which there is no reduction of uptake. They also gave a recipe for estimating h_t, $h_{0.5}$, h_{ot}, and $h_{o0.5}$, and hence the value of a in (49). For corn they give the following estimates: $h_t=-500\,\text{cm}$, $h_{0.5}=-1376\,\text{cm}$, $h_{ot}=-1224\,\text{cm}$ and $h_{o0.5}=-4250\,\text{cm}$. This implies $a=h_{o0.5}/\text{h}_{0.5}=3.09$. Pang and Letey (1998) also used a procedure for compensating for reduced uptake due to water and/or salinity stress in part of the root zone by removing extra water from the remainder. The value $a=h_{o0.5}/\text{h}_{0.5}=3.09$ in (49) contrasts with the assumed value $a=1$ in Cardon and Letey (1992a; 1992b; 1992c). In line with this, Feng, Meiri and Letey (2003) used the methods of Pang and Letey (1998) to re-evaluate the experiment reported by Shalhevet, Vinten and Meiri (1986).

Van Genuchten (1987) also proposed, “as an equally plausible approach”, a root water-uptake reduction function with a *multiplicative* dependence on the pressure h and the osmotic head h_oi.e. $\alpha\left(h,h_o\right)=\alpha\left(h\right)\alpha\left(h_o\right)$. Hence

$$\alpha\left(h\right)\alpha\left(h_o\right)=\frac{1}{1+\left(\dfrac{h}{h_{0.5}}\right)^{p}}\times\frac{1}{1+\left(\dfrac{h_o}{h_{o0.5}}\right)^{p_o}} \tag{50}$$

where p and p_o are empirical parameters, and $h_{0.5}$ and $h_{o0.5}$ are empirical parameters at which, respectively, $\alpha\left(h\right)=0.5$ and $\alpha\left(h_o\right)=0.5$. This combined water and salinity stress reduction factor was used by Šimůnek, Suarez and Sejna (1996).

Dirksen and Augustijn (1988) and Dirksen et al. (1993) modified (50), by introducing values h^* and h_o^* of, respectively, h and h_o above which there is no reduction:

$$\alpha\left(h\right)\alpha\left(h_o\right)=\frac{1}{1+\left(\dfrac{h^*-h}{h^*-h_{0.5}}\right)^{p}}\times\frac{1}{1+\left(\dfrac{h_o^*-h_o}{h_o^*-h_{o0.5}}\right)^{p_o}} \tag{51}$$

This in effect anticipated the introduction of the similar threshold values h_t and h_ot by Pang and Letey (1998) discussed above.

Based on (45) Homaee (1999), Homaee and Feddes (2001) and Homaee, Dirksen and Feddes (2002; 2002b; 2002c; 2002a) proposed:

$$\alpha(h)\alpha(h_\mathrm{o}) = \frac{1}{1+\dfrac{1-\alpha(h_{\max})}{\alpha(h_{\max})}\left(\dfrac{h^*-h}{h^*-h_{\max}}\right)^{p}} \times \frac{1}{1+\dfrac{1-\alpha(h_{\mathrm{o}\max})}{\alpha(h_{\mathrm{o}\max})}\left(\dfrac{h_\mathrm{o}^*-h_\mathrm{o}}{h_\mathrm{o}^*-h_{\mathrm{o}\max}}\right)^{p_\mathrm{o}}} \qquad (52)$$

However, (52) reduces to (51) with $h_{0.5}$ and $h_{\mathrm{o}0.5}$ given by (46). In other words, these equations are equivalent with cut-off points $[h_{\max}, \alpha(h_{\max})]$ and $[h_{\mathrm{o}\max}, \alpha(h_{\mathrm{o}\max})]$.

Shani and Dudley (1996) and Dudley and Shani (2003) compared (49) with the Nimah-Hanks model and also with a multiplicative model like (50) with the water stress factor replaced by the Gardner / Cowan model.

Van Dam et al. (1997; see also Van Dam 2000) proposed a root water-uptake reduction function with a multiplicative dependence on the pressure h and the osmotic head h_o by multiplying the Feddes water stress reduction function and the Maas-Hoffman-Van Genuchten salinity stress reduction function. In its most general form this reduction function is obtained by multiplying (39) and (41):

$$\alpha(h)\alpha(h_\mathrm{o}) = \left(\begin{array}{cc} 0 & h \ge h_1 \\ \dfrac{(h_1-h)}{(h_1-h_2)} & h_1 > h \ge h_2 \\ 1 & h_2 > h \ge h_3(T_p) \\ \dfrac{(h-h_4)}{(h_3(T_p)-h_4)} & h_3(T_p) > h \ge h_4 \\ 0 & h_4 > h \end{array}\right) * \left(\begin{array}{cc} 1 & 0 \ge h_\mathrm{o} \ge h_\mathrm{ot} \\ 1-\dfrac{(h_\mathrm{ot}-h_\mathrm{o})}{(h_\mathrm{ot}-h_\mathrm{o0})} & h_\mathrm{ot} > h_\mathrm{o} \ge h_\mathrm{o0} \\ 0 & h_\mathrm{o0} > h_\mathrm{o} \end{array}\right) \qquad (53)$$

Van Dam et al. (1997; see also Van Dam 2000) wrote this multiplicative model in the simple form:

$$\alpha(h)\alpha(h_\mathrm{o}) = \left(\frac{h-h_4}{h_3-h_4}\right) * \left[1-b^{-1}(h_\mathrm{ot}-h_\mathrm{o})\right] \qquad (54)$$

where $b^{-1} = (h_\mathrm{ot}-h_\mathrm{o0})$ is a crop-dependent parameter, and the expression is valid for $h_3 \ge h \ge h_4$ and $h_\mathrm{ot} \ge h_\mathrm{o}$. In this form it is also used in the current version of SWAP (Kroes and Van Dam 2003).

Homaee (1999) and Homaee, Dirksen and Feddes (2002; 2002b; 2002c; 2002a) modified (53) to account for the fact that plants wilt at higher water pressure head in the presence of salinity than without salinity. To this end they shifted the wilting point by an amount equal to h_o and thus proposed (see Figure 9):

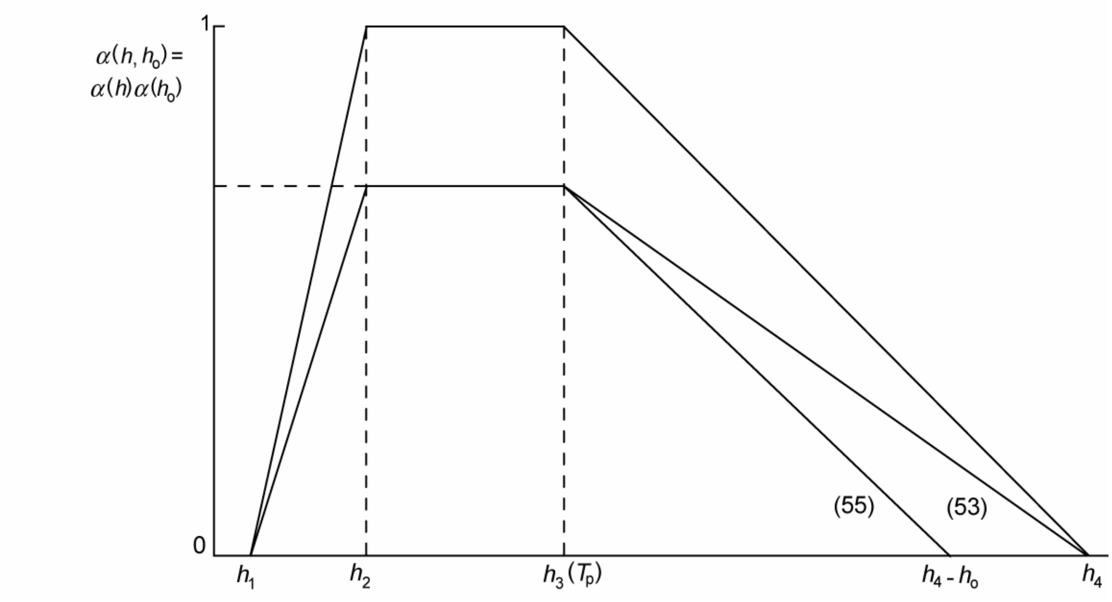

Figure 9. Sketch of the multiplicative reduction factor $\alpha(h,h_0) = \alpha(h)\alpha(h_0)$ according to the equations (53) and (55)

$$\alpha(h,h_o) = \alpha(h)\,\alpha(h_o) = \left(\frac{h-(h_4-h_o)}{h_3-(h_4-h_o)}\right)*\left(1-b\left(h_{ot}-h_o\right)\right) \tag{55}$$

They also pointed out three alternatives resulting from replacement of the factor $\left(1-b^{-1}\left(h_{ot}-h_o\right)\right)$ by the osmotic reduction factors $\alpha(h_o)$ in, respectively, (43), (44) and (45).

Spatial–temporal description and interconversion of bulk root mass density and root length density data

Spatial–temporal description of bulk root biomass density

Earlier we discussed that root water uptake is strongly related to root length densities. Such data are often not available. Root mass, however, is easier to obtain and, therefore, we investigate the *relationship between root mass and root length density*. Moreover, it may elucidate the different ways researchers characterize root systems.

Gale and Grigal (1987) introduced the following expression for a *time-invariant* cumulative root biomass fraction $Y[-]$ as a function of depth $z[\text{cm}]$:

$$Y(z) = \frac{B(z,t)}{B_\infty(t)} = 1 - \beta^{\gamma z} \tag{56}$$

where $B(z,t)$ is the root biomass per unit area of land *above* z at time t, $B_\infty(t) = B(\infty,t)$ is the *total* root biomass per unit area of land at time t, and β [–] and γ [cm^{-1}] are vegetation-dependent parameters. Higher values of β^{γ} correspond to a deeper root system.

To render those data useful in root water-uptake models, they should be converted first to spatial–temporal distributions of the bulk root *mass* density $\rho\left(z,t\right)$ and next to spatial–temporal distributions of the root *length* density $L_{\mathrm{r}}\left(z,t\right)$. In the following first the bulk root mass density $\rho\left(z,t\right)$ corresponding to (56) is derived and interpreted. Next the reformulation and generalization of the root distribution model in exponential form by Arora and Boer (2003) is presented and interpreted.

Partial differentiation of $B\left(z,t\right)$ with respect to z gives an expression for $\rho\left(z,t\right)$:

$$\begin{aligned}&\rho\left(z,t\right)=\frac{\partial B\left(z,t\right)}{\partial z}=\frac{\mathrm{d}Y\left(z\right)}{\mathrm{d}z}B_{\infty}\left(t\right)=\\&=\left(\ln\beta^{-\gamma}\right)\beta^{\gamma z}B_{\infty}\left(t\right)=\left(\ln\beta^{-\gamma}\right)\left(1-Y\left(z\right)\right)B_{\infty}\left(t\right)\end{aligned}\tag{57}$$

Eq. (57) implies that the surface root biomass density $\rho\left(0,t\right)$ = $\left(\ln\beta^{-\gamma}\right)B_{\infty}\left(t\right)$. Note that an equivalent uniform root system with root biomass density $\rho\left(z,t\right)=\left(\ln\beta^{-\gamma}\right)B_{\infty}(t)$ in the depth range of $0\le z\le\left(\ln\beta^{-\gamma}\right)^{-1}$ will have the same total root biomass $B_{\infty}(t)$. *Thus the inverse of* $\ln\beta^{-\gamma}$ *in (57) can be interpreted as the characteristic rooting depth based on the surface root biomass density and the total root biomass.* Furthermore (57) implies that, at any time t and at any position z, the relative time rate of change of the root biomass density $\rho\left(z,t\right)$ is equal to the relative time rate of change of the total root biomass $B_{\infty}(t)$:

$$\frac{1}{\rho\left(z,t\right)}\frac{\partial\rho\left(z,t\right)}{\partial t}=\frac{1}{B_{\infty}\left(t\right)}\frac{\mathrm{d}B_{\infty}\left(t\right)}{\mathrm{d}t}\tag{58}$$

Eq. (58) shows that the model of Gale and Grigal (1987) implies a *spatially uniform* allometric relationship between $\rho\left(z,t\right)$ and $B_{\infty}(t)$, *independent of time and depth.* Note that the model of Gale and Grigal (1987) also *implies that roots are present at all depths at all times* $t>0$. For crops and vegetations in the early stages this is not realistic.

Arora and Boer (2003) developed a more versatile root distribution model. First they rewrote the model of Gale and Grigal (1987) in exponential form (cf. Gerwitz and Page 1974). To this end they *replaced* in (57) and (58) *the two-parameter factor* β^{γ} *by* e^{-a}, *so that the inverse length* $a=\ln\beta^{-\gamma}=-\gamma\ln\beta$. The bulk root mass density $\rho\left(z,t\right)$ then reads:

$$\rho\left(z,t\right)=aB_{\infty}\left(t\right)\exp\left(-az\right)\tag{59}$$

Note that (59) implies that $\rho\left(0,t\right)$ = $aB_{\infty}(t)$. The cumulative root biomass fraction $Y(z)$ can then be formulated as:

$$Y(z) = \frac{B(z,t)}{B_\infty(t)} = 1 - \exp(-az) \tag{60}$$

An equivalent uniform root system with $\rho(z,t) = aB_\infty(t)$ in the depth range of $0 \le z \le a^{-1}$ will have the same total root biomass $B_\infty(t)$. Therefore similar to the interpretation of (57) the *inverse of* the *parameter* a in (59) *can be interpreted as the characteristic rooting depth* based on the surface root biomass density and the total root biomass. The exponential model (59) still satisfies the spatially uniform allometric relationship between $\rho(z,t)$ and $B_\infty(t)$ expressed by (58). Also, (59) still implies that at all times $t > 0$ roots are present at all depths. *The cumulative root biomass fraction* $Y(z,t)$ *in (60) is time-invariant.*

Next, Arora and Boer (2003) *generalized the exponential root growth model* (59) to allow for the common observation that rooting depth increases in the course of time. To this end they introduced a power-law dependence of the parameter a upon the total root biomass $B_\infty(t)$, namely:

$$a(t) = bB_\infty^{-\alpha}(t) \tag{61}$$

where b and α are constants, thus generalizing (59) to:

$$\rho(z,t) = a(t)B_\infty(t)\exp\left[-a(t)z\right] = bB_\infty^{(1-\alpha)}(t)\exp\left[-bB_\infty^{-\alpha}(t)\,z\right] \tag{62}$$

Eq. (62) implies that the surface root biomass density $\rho(0,t) = a(t)B_\infty(t) = bB_\infty^{(1-\alpha)}$. An equivalent uniform root system with root biomass density $\rho(z,t) = a(t)B_\infty(t) = bB_\infty^{(1-\alpha)}$ in the depth range of $0 \le z \le a^{-1}(t) = b^{-1}B_\infty^{\alpha}(t)$ will have the same total root biomass $B_\infty(t)$. Again the inverse of the parameter $a(t) = bB_\infty^{-\alpha}(t)$ in (62) can be interpreted as the *time-dependent characteristic rooting depth* based on the surface root biomass density and the total root biomass.

The cumulative root biomass fraction $Y(z,t)$ corresponding to (62) is:

$$Y(z,t) = \frac{B(z,t)}{B_\infty(t)} = 1 - \exp\left(-a(t)\,z\right) = 1 - \exp\left(-\frac{b}{B_\infty^{\alpha}(t)}z\right) \tag{63}$$

Note that, unlike in (56) and (60), the *cumulative root biomass fraction now depends on time*, whenever the total root biomass B_∞ depends on time. The relative time rate of change of the root biomass density corresponding to (62) in place- and time-dependent form reads as:

$$\begin{aligned}\frac{1}{\rho(z,t)}\frac{\partial\rho(z,t)}{\partial t} &= \left[(1-\alpha)+\alpha a(t)\,z\right] \qquad \frac{1}{B_\infty(t)}\frac{\mathrm{d}B_\infty(t)}{\mathrm{d}t}\\ &= \left[(1-\alpha)+\alpha b B_\infty^{-\alpha}(t)\,z\right] \quad \frac{1}{B_\infty(t)}\frac{\mathrm{d}B_\infty(t)}{\mathrm{d}t}\\ &= g(z,t) \qquad\qquad \frac{1}{B_\infty(t)}\frac{\mathrm{d}B_\infty(t)}{\mathrm{d}t}\end{aligned} \tag{64}$$

Clearly, unlike in (58), in (64) the relative time rates of change of $\rho(z,t)$ and $B_\infty(t)$ are no longer equal to each other. Eq. (64) shows that the model of Arora and Boer implies a spatially non-uniform temporally variable allometric relationship between $\rho(z,t)$ and $B_\infty(t)$, depending on α, b, $B_\infty(t)$ and z. Following Huxley (1972), we may refer to $g(z,t)$ in (64) as the growth coefficient. For a given time and depth, $g(z,t)$ describes the relative time rate of change of the mass density $\rho(z,t)$ in comparison with the relative time rate of change of the total root biomass $B_\infty(t)$. However, note that the generalized model of Arora and Boer (2003) *still implies that roots are present at all depths and at all times* $t>0$. Again, for crops and vegetations in the early stages this is not realistic.

Two important special cases of (61-64) are:

- If $\alpha=0$, (61) implies that $a=b$ and (62) implies that the generalized exponential model of Arora and Boer (2003) reduces to the simple exponential model, i.e. (62), (63) and (64) reduce to (59), (60) and (58).
- If $\alpha=1$, (61) implies that the characteristic rooting depth $a^{-1}(t)$ is proportional to the total root mass density $B_\infty(t)$ and (62), (63) and (64) reduce to:

$$\rho(z,t)=b\exp\left(-bB_\infty^{-1}(t)\,z\right) \tag{65}$$

$$Y(z,t)=\frac{B(z,t)}{B_\infty(t)}=1-\exp\left(-a(t)\,z\right)=1-\exp\left(-\frac{b}{B_\infty(t)}z\right) \tag{66}$$

$$\begin{aligned}\frac{1}{\rho(z,t)}\frac{\partial\rho(z,t)}{\partial t} &= a(t)\,z \qquad \frac{1}{B_\infty(t)}\frac{\mathrm{d}B_\infty(t)}{\mathrm{d}t}\\ &= bB_\infty^{-1}(t)\,z\,\frac{1}{B_\infty(t)}\frac{\mathrm{d}B_\infty(t)}{\mathrm{d}t}\\ &= g(z,t) \qquad \frac{1}{B_\infty(t)}\frac{\mathrm{d}B_\infty(t)}{\mathrm{d}t}\end{aligned} \tag{67}$$

Eq. (65) implies $\rho(0,t)=b$, i.e. the surface root biomass density is constant. An equivalent uniform root system with root biomass density $\rho(z,t)=b$ in the depth range $0\le z\le b^{-1}B_\infty(t)$ will have the same total root biomass $B_\infty(t)$. Thus the inverse of $bB_\infty^{-1}(t)$ in (65) can be interpreted as the time-dependent characteristic rooting depth based on the surface root biomass density and the total root biomass. In other words: *for* $\alpha = 1$ *the surface root density is constant and the characteristic rooting depth increases linearly with the total root biomass* $B_\infty(t)$. Note that in (67) the growth coefficient $g\left(z,t\right)$ is proportional to depth z and inversely proportional to the total root biomass $B_\infty(t)$.

Interconversion of bulk root *mass* density and root *length* density data

The bulk root mass density ρ can be obtained by multiplying the root volume fraction ϕ_r by the intrinsic root density γ_r:

$$\rho = \gamma_\mathrm{r}\phi_\mathrm{r} \tag{68}$$

where the root volume fraction ϕ_r is the product of the root length density L_r and the root cross section πr^2:

$$\phi_\mathrm{r} = \pi r^2 L_\mathrm{r} \tag{69}$$

Combining (68) and (69) gives:

$$L_r = \frac{\rho}{\pi r^2 \gamma_r} \tag{70}$$

The denominator $\pi r_0{}^2\gamma_\mathrm{r}$ (m g^{-1}) in (70) represents the *root mass per unit length of root*. Van Noordwijk (1987) refers to it as the specific root length. Van Noordwijk and Brouwer (1991) account for two potential sources of error:

- presence of a potentially significant air-filled volume fraction in root tissue;
- the variance of the root radius, requiring in (70) the use of the square of the quadratic average root radius which is equal to the square of the linear average root radius plus the variance of the root radius.

Thus dividing both sides of (57), (59), (62) and (65) gives the following four expressions for the root length density:

$$L_\mathrm{r} = \frac{\rho(z,t)}{\pi r^2 \gamma_\mathrm{r}} = \frac{\left(\ln \beta^{-\gamma}\right)\beta^{\gamma z} B_\infty\left(t\right)}{\pi r^2 \gamma_\mathrm{r}} \tag{71}$$

$$L_\mathrm{r} = \frac{\rho(z,t)}{\pi r^2 \gamma_\mathrm{r}} = \frac{a B_\infty\left(t\right)\exp\left(-az\right)}{\pi r^2 \gamma_\mathrm{r}} \tag{72}$$

$$L_\mathrm{r} = \frac{\rho(z,t)}{\pi r^2 \gamma_\mathrm{r}} = \frac{a\left(t\right) B_\infty\left(t\right)\exp\left(-a\left(t\right)z\right)}{\pi r^2 \gamma_\mathrm{r}} = \frac{b B_\infty{}^{(1-\alpha)}\left(t\right)\exp\left(-b B_\infty{}^{-\alpha}\left(t\right)z\right)}{\pi r^2 \gamma_\mathrm{r}} \tag{73}$$

$$L_\mathrm{r} = \frac{\rho\left(z,t\right)}{\pi r^2 \gamma_\mathrm{r}} = \frac{b\exp\left(-b B_\infty{}^{-1}\left(t\right)z\right)}{\pi r^2 \gamma_\mathrm{r}} \tag{74}$$

Note that (71) and (73) correspond, respectively, to the models of Gale and Gringal (1987) and Arora and Boer (2003), and that (73) reduces to (72) and (74), respectively, if $\alpha = 0$ and $\alpha = 1$. *Eqs. (71)-(74) can be used to convert bulk root mass density data to root length density data and vice versa.*

Integrating (70) over the entire rooting depth gives an expression for the root length L_{profile} (cm cm^{-2}) per unit area of land (cf. Newman 1969):

$$L_{\text{profile}} = \int_0^\infty L_r \mathrm{d}z = \int_0^\infty \frac{\rho}{\pi r^2 \gamma_{\mathrm{r}}} \mathrm{d}z = \frac{B_\infty}{\pi r^2 \gamma_{\mathrm{r}}} \tag{75}$$

Root water uptake at regional and global scales

The importance of roots in atmospheric models

Research by Milly and Dunne (1994) suggests a strong sensitivity of continental evaporation to water capacity. Hence, closer root modeling in Global Climate Models (GCMs) might improve soil–vegetation control instead of uncontrolled continental evaporation that extracts water from a deep soil water reservoir. GCMs typically use shallow rooting depths around 2 m. However, deep-rooted vegetation (of up to 68 m) has been found in the Tropics. Therefore, Kleidon and Heimann (2000) investigated the effects of larger rooting depths associated with the incorporation of deep roots, on the surface energy balance and the atmosphere using a GCM. The incorporation of deep rooted vegetation in the GCM leads to large-scale differences in the simulated surface climate and the atmospheric circulation, especially in the seasonal humid tropics.

Various Soil–Vegetation–Atmosphere Transfer (SVAT) schemes have been developed for use in GCMs and numerical weather prediction models. Their weakest component remains their link with the lower boundary. These schemes face various difficulties, including: comparable complexity between system components; scaling incongruities between atmospheric, hydrological and terrestrial components; and validation of SVATs at appropriate time and space scales (Kalma et al. 1999). SVATs, which sometimes are over-parameterized, use a variety of different methods to represent the relationship between roots, soil water and transpiration. Moreover, SVAT parameters are generally highly variable in space and difficult to measure.

It is therefore not surprising that the Project for Intercomparison of Land-surface Parameterization Schemes (PILPS) showed that SVATs/Land Surface Schemes (LSS) driven by the same meteorological forcing of air temperature, humidity, wind speed, incoming solar radiation, longwave radiation and rainfall, produced remarkably different surface energy and water balances (Chen et al. 1997; Koster and Milly 1997; Pitman et al. 1999).Therefore the question raised in this context is: *how important is the role of roots?*

Vegetation root distribution is one of the factors that determine the overall water-holding capacity of the land surface and the relative rates of water extraction from different soil layers by transpiring vegetation. Despite its importance, significantly different root distributions are used by different land-surface models.

Sensitivity of land-surface models and climate models to the representation of roots

Using the SWAP-model approach of Van Dam and Feddes (2000), one can evaluate the effect of root distribution on the course of actual transpiration in time. We take as an example a grass vegetation (covering the soil completely) with a rooting depth of 80 cm growing on a loamy sand of 2 m depth containing 10% clay (< 2 μm). Free drainage prevails at the bottom. As initial condition throughout the profile the soil water pressure head *h* = -200 cm, implying a rather wet soil. Then at the soil surface a potential transpiration rate $T_p = 4$ mm d^{-1} is applied for two different relative root density distributions: Root 1, where most of the roots are located in the topsoil, the root density decreasing fast with depth, and Root 2 where there is a constant, homogeneous root distribution with depth. The results of the simulations are shown in Figure 10.

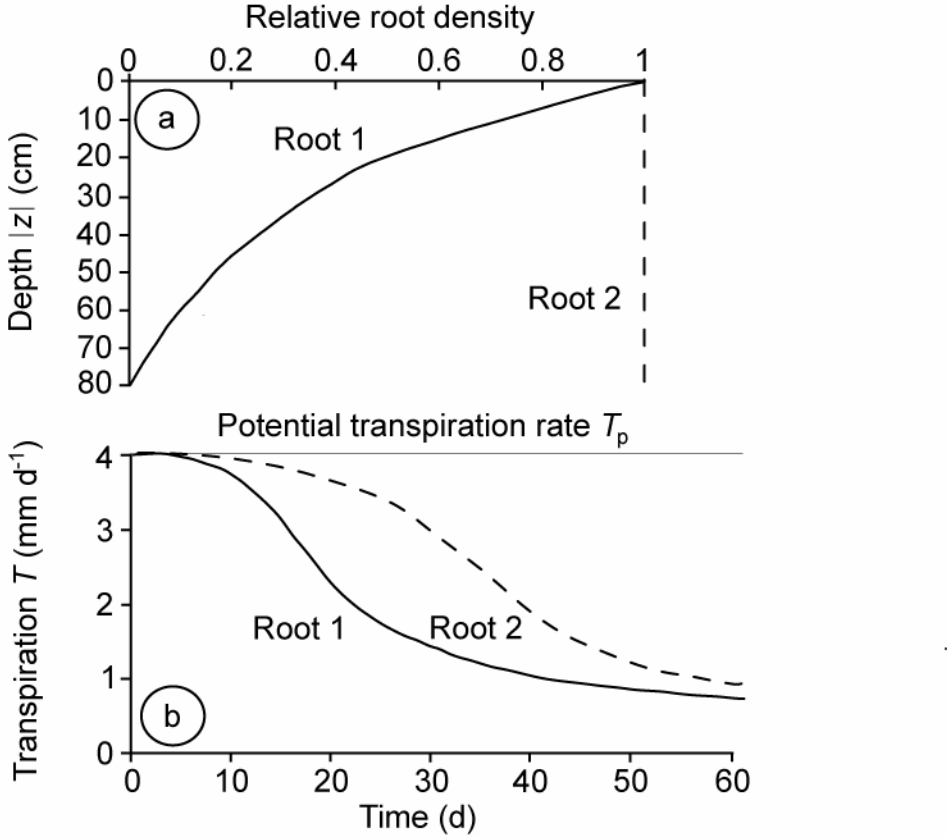

Figure 10. (a) Two different relative root density distribution functions (Root 1 and Root 2) adopted for grass, with a rooting depth of 80 cm, growing on loamy sand; (b) Simulated actual transpiration rate *T* in time for these two root distribution functions, taking potential transpiration rate $T_p = 4$ mm d^{-1} as upper boundary condition

Transpiration is generally more sensitive to the moisture content of the densely rooted soil layer than to that in the remainder of the root zone, hence Root 1 produces an earlier onset of moisture stress than Root 2. In fact, after 30 days the actual transpiration rate of Root 1 is about half that of Root 2. Similar results were reported by Desborough (1997). This is a clear demonstration that *roots can influence the behavior of a land-surface model, the role of roots being particularly important when soil water limits evapotranspiration.*

De Rosnay and Polcher (1998) have presented root water uptake in the LSS Sechiba that receives climatic forcing from the GCM and explicitly takes into account sub-grid-scale variability of vegetation and root profiles. Normalized root length density $L_\mathrm{r}\left(z\right)/L_\mathrm{profile}$ (varying between 0 at the bottom of the rooting zone and 1 at the soil surface where $z = 0$) is here assumed to depend exponentially on positive soil depth z as

$$\frac{L_\mathrm{r}(z)}{L_\mathrm{profile}} = e^{-cz} \tag{76}$$

where c is a fitting constant depending on the biome considered. The use of a relatively easily observable parameter such as c makes the development of a global data set for root water-uptake parameterization feasible.

De Rosnay et al. (2002) presented in their new Sechiba scheme soil–plant interactions that result from root–soil water profile interactions being represented on a rather fine vertical resolution. The grid boxes of the model included different vegetations and soil textures. This allowed for a better control of evapotranspiration (it reduced the overestimation of continental evapotranspiration, which typically occurs in GCMs when soil water capacity is increased).

Overall, *there is significant evidence that roots play an important role in the simulations by land-surface models*. Understanding the interactions of root distributions with water use, as well as potential feedbacks with climate, is important for climate simulations as we determine which systems in nature actually rely on relatively deep soil water (Greenwood 1992; Nepstad et al. 1994; Jackson et al. 2000).

Deep roots play a major role in ecosystems such as forests and need to be studied globally in more detail. They are likely to have somewhat more effect in the tropics than in mid-latitudes. On the other hand, the role of shallow roots seems to be overestimated in many land-surface schemes.

A number of regional and global *soil* databases that are currently available are listed in Table 1 of Feddes et al. (2001). In addition, see also the website http://sis.agr.gc.ca/cansis/.

What root information exists?

Because of the sensitivity of simulated transpiration in climate models, global datasets of root and soil properties are increasingly needed. Cumulative root distributions Y from the soil surface down to rooting depth here defined as the depth where Y reaches an arbitrary value, for example, 99%, have been fitted for various biomes with a vegetation-dependent coefficient β to the asymptotic equation of Gale and Grigal (1987) as:

$$Y = 1 - \beta^{|z|} \tag{56 b}$$

Eq. (56b) has been widely used to represent *cumulative* root biomass distribution data (Jackson 1999; Jackson et al. 1996; 1997; 2000; Feddes et al. 2001; Zeng et al. 1998; Zeng 2001). Values for β (see Figure 11) and properties like the ratio fine/total root biomass, root length, maximum rooting depth, root/shoot ratio, and nutrient content of different terrestrial biomes, can be found in the above-cited references.

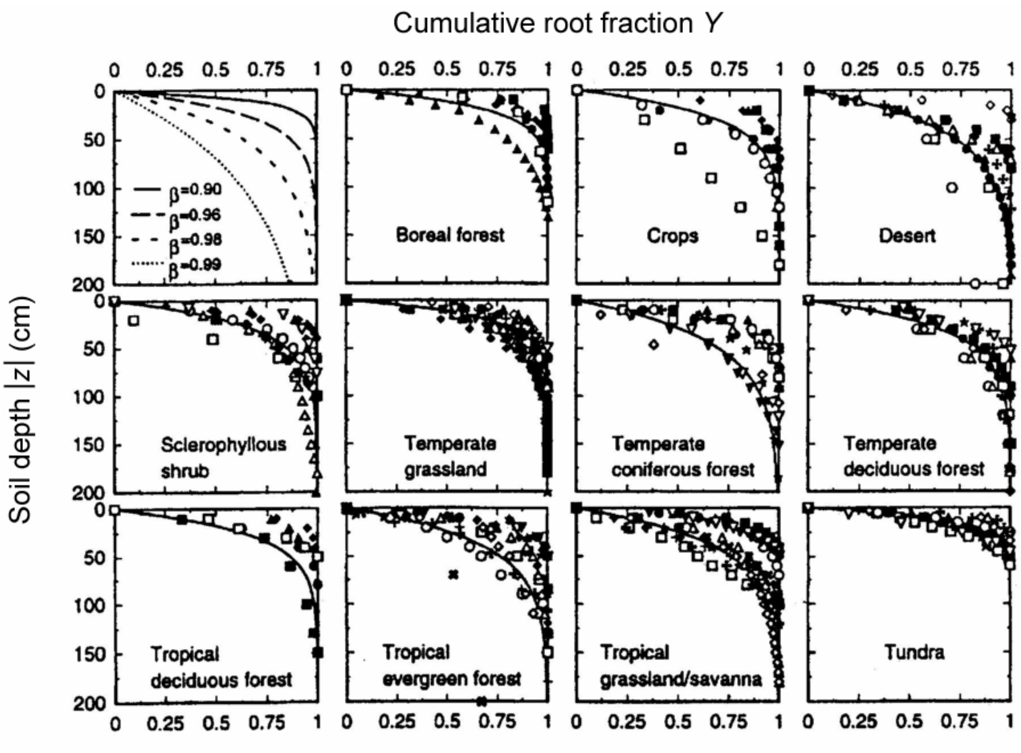

Figure 11**.** Cumulative root distributions as a function of soil depth for 11 terrestrial biomes and for the theoretical model of Gale and Grigal (Eq. 56). Crop data are simply examples from a few studies (after Jackson 1999)

To date a root database (Jackson et al. 2000; 2000) of more than 1000 profiles exists that covers various combinations of maximum rooting depth of fine and coarse roots, root length densities, root biomass (and surface area in a small subset of the data), as well as root nutrient concentrations by biome and plant life form (e.g. Jackson et al. 1996; 1997; Canadell et al. 1996; Gill and Jackson 2000). There is, however, a lack of information on annual crops within the database. This root database currently has no spatial expression.

Newman's (1969) expression for L_{profile} per unit area of land (Eq. 75) is related to the total live fine root length $L_{\mathrm{lf}\infty}(t)$ reported in Jackson, Mooney and Schulze (1997). Newman's Table 2 shows L_{profile} data from 11 literature sources, giving measurements made on plants grown under field conditions in closed or nearly closed stands. The values of L_{profile} are largest for gramineous herbs, intermediate for non-gramineous herbs, and smallest for woody plants. They range from about 5 to 100 cm cm^{-2} for woody plants and from 50 to 4000 cm cm^{-2} for herbs.

Although not available in a formalized and centralized database, many root distribution data, rooting-depth data, etc for *annual (agricultural) crops* can, according to R. de Jong (personal communication) be found in numerous published manuscripts, reports, books etc. Perhaps there is an *opportunity* to pull this material together in a single data base, along the same lines that Van Genuchten, Leij and Wu (1999) did for soil hydraulic properties. Preferably, the climatic, soil and management conditions under which the root data were collected should also be recorded in such a data base.

Analysis of Strengths and Weaknesses, Opportunities and Threats (SWOT)

Liquid water flow inside the plant: soil and plant resistances

S: The electric-analogue model of Van den Honert (1948) stimulated the evaluation of the relative importance of hydraulic resistances to flow in soil and plant roots.

W: a) Soil and plant resistances are not necessarily constant but depend on the actual environmental conditions.
   b) The strong focus on the hydraulic aspects diverted attention away from the osmotic aspects.

O: The membrane theory introduced in the 1970s provides an opportunity to avoid simple addition of hydraulic and osmotic potentials in theories of flow and transport. The membrane theory can possibly also contribute to modeling the functioning of *aquaporins* (Maurel 1997).

T: Current organizational and funding structures in the USA, Europe and Australia seem not to stimulate close co-operation between plant scientists and hydrologists: they appear to be working in parallel rather co-operatively.

Macroscopic description of root water uptake

S: a) Root water uptake can relatively easily be measured in dependence of depth and time.
   b) Empirical descriptions of root water uptake at the macroscopic scale are convenient for use in numerical models.

W: Ideally one would derive such macroscopic descriptions from underlying process-based mesoscopic and microscopic models.

O: The ongoing miniaturization of instruments which can be used over a large moisture range, such as the osmotic tensiometer, allows one to measure root water uptake in great detail. Also the work with e.g mini-rhizotrons and heat-pulse probes is promising (e.g. Clothier and Green 1997).

Modeling root water uptake at mesoscopic and macroscopic scales

S: Relationships between mesoscopic and macroscopic descriptions of water uptake have been explored much over the last half century. This has provided a useful framework for interpreting laboratory and field data and formulating macroscopic sink terms.

W: a) The focus on the hydraulic aspects of root water uptake seems to have diverted attention away from the osmotic aspects.
   b) The evaluation of degree and implications of limited soil–root contact gives only qualitative results.
   c) The mesoscopic approach requires details about root geometry and soil heterogeneity that are usually not available.
   d) Soil hydraulic properties, like the water-retention and hydraulic-conductivity characteristics, at the mesoscale may be quite different from the commonly available macroscale soil hydraulic properties.
   e) Most solutions of the flow equations at the mesoscale involve linearization.

O: a) There are several possibilities for coping with limited soil/root contact and/or heterogeneity. One way to deal with limited soil–root contact is to define an ‘effective conductivity’, which represents the conductivity of the soil–root interface and the root tissue. The justification of this should be explored further.

b) Steady-rate solutions giving water-content distributions of time-invariant form, with the water content decreasing linearly with time everywhere at the same rate, can also be found for roots that are in contact with the soil over only a part of their circumference or roots that are not uniformly distributed in the soil. As for the cylindrically symmetric case, an instant at which the *constant-rate phase* terminates and the *falling-rate phase* starts can be evaluated.

c) Techniques for measurement of hydraulic properties at the mesoscale should be explored further (Hartge and Stewart 1995).

d) Linearization can be avoided by relying on numerical solutions. Also, general inferences from analytical solutions of the nonlinear diffusion equation remain a worthwhile objective.

T: Funding of the mechanistic studies needed, including expensive experiments, is very problematic. The question arises: why is not sufficient funding forthcoming?

Hydrologically oriented modeling approach for root water uptake at macroscale

S: a) One-dimensional analysis of the root-zone water balance is well developed and has strong computational capabilities.

b) Co-operation between hydrology and meteorology in defining upper boundary conditions for root water uptake (i.e. potential transpiration) is well developed.

c) Limited availability of water in part of the root zone may be compensated by uptake in another, wetter part. Plant water uptake generally occurs with the minimum amount of work/energy possible. Different numerical procedures to compute root water uptake under those minimum-energy conditions have now been developed.

d) Joint stresses can be computed easily by multiplication of the water stress with the salinity stress.

e) Based on salt-tolerance studies, we have a fairly good knowledge of the salinity stress reduction function for a wide variety of crop species.

W: a) Multi-dimensional analysis of the root-zone water balance is often needed, i.e. for isolated trees, row crops, patterns in vegetation.

b) A sound biophysical theory is lacking. It is not completely clear whether the water and salinity stresses are additive or multiplicative. There is a severe lack of data for implementing the mesoscopic description of simultaneous uptake of water and solutes by a variety of plant species.

c) For wild species the salt tolerance has not been established in the same systematic fashion as for crops, and hence the salinity stress reduction functions are not as easily available.

d) Co-operation between hydrology and plant sciences, in particular plant nutrition and plant genetics, needs to be strengthened.

O: a) The implementation at the macroscopic scale of the mesoscopic description of simultaneous uptake of water and solutes is promising.

b) The already developed schemes that assure that root water uptake will be realized under minimum-energy conditions should be implemented in all currently used numerical macroscale models.

c) There is a need for further evaluation of the various root water uptake reduction functions, but experimental data are lacking.

   d) The successful use of combined water and salinity stress reduction functions holds promise for the introduction of reduction functions related to nutritional status and crop pests.

T: a) Funding of mechanistic studies is problematic.

Spatial–temporal description and interconversion of bulk root mass density and root length density data

S: Many empirical data are available on root mass and length distribution.

W: a) Little is known about the effectiveness of roots as a function of age, size etc.
   b) Root distribution functions are handled differently by different researchers. This causes considerable confusion among the users. However, a general framework for a better understanding has been presented in this paper.
   c) There is a lack of systematic root information on annual vegetations.

O: a) The extent and shape of rooting systems and their changes play a major role in determining uptake patterns. Root growth evolves in time and space, which in models is often not accounted for. Hence, models may fail in predicting spatial variations and the dynamics of soil-water – plant-growth interactions.
   b) Root water-uptake dynamics is usually related to root length density, ignoring uptake control by root surface area and root age.
   c) 3D-modeling is limited by the need of many additional soil and plant parameters (Hopmans and Bristow 2002).

Root water uptake at regional and global scales

S: a) Root water uptake can be represented in Land Surface Schemes (LSS) that receive climatic forcing from a GCM and explicitly take into account sub-grid-scale variability of vegetation and root profiles.
   b) A number of regional and global soil and root data bases exist that can be implemented in present land-surface schemes (e.g. Feddes et al. 2001). These should be used in combination with databases on e.g. regional / global land use, topography.

W: Large-scale aggregation of soil and root data that takes into account sub grid variability of soil and roots, requires improved root and soil data in combination with better knowledge of the interaction of these two systems.

O: a) As there exists a strong sensitivity of continental evaporation to water capacity, closer root modeling in GCMs might improve soil–vegetation control instead of uncontrolled continental evaporation that extracts water from a deep soil moisture reservoir.
   b) A first priority is to establish firmly the relationships of root biomass, rooting depth, root distribution and root functions with land-use type, soil type, soil texture, topography and climate.

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