THE DISTRIBUTION OF THE UPTAKE OF WATER BY PLANTS: INFERENCE FROM HYDRAULIC AND SALINITY DATA *

P. A. C. Raats **

1. *Recent advances in the technology of irrigation*

Methods of irrigation can be devided roughly in two groups: 1) methods which depend on flow of water over the land or in furrows to get proper distribution of the water, and 2) methods which supply the desired amounts directly at each point. Flooding of entire fields, or parts of fields, or in furrows falls in the first category. As soon as water is supplied at the end of a field or at the end of a furrow, infiltration starts at that edge or end. At all other points infiltration starts only after arrival of the water. At every moment the cumulative amount of water supplied to the land is equal to the sum of the amount of water on the land and the amount of water that has already infiltrated:

$$qt = cL + \int_0^L I\,dx, \tag{1}$$

where t is a time scale with its origin at the instant the irrigation was started, x is a spatial coordinate with its origin at the point where the irrigation was started, q is the rate of supply, L is the distance the water has travelled on the land at time t, c is the depth of the water on the land, and I is the cumulative infiltration at any point up till time t. The cumulative infiltration is a function of the type of soil, the distribution of its initial water content, and of the time the water has been available for infiltration. Even if this opportunity time is everywhere the same, cumulative infiltration will vary widely as a result of the spatial variability of the relationships between the water content, the pressure head, and the hydraulic conductivity. In practice typical efficiencies of flood and furrow irrigation, defined as per-

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** Soil Physicist, Institute for Soil Fertility, Postbus 30003, 9750 RA Haren (Gr.). The Netherlands.

centage of water required for evaporatranspiration and leaching of excess salts relative to amount of water delivered to the field, range from 40 to 60 percent. The large amounts of drainage waters often lead to waterlogging, saline seeps, and deterioration of the quality of water in aquifers and rivers.

The only way one can circumvent the dependence on the infiltrability is to deliver known quantities of water to all points on the field. Overland flow should be prevented by applying water at a rate less than the soil infiltrability of the area for which the water is intended or by forming basins, say around trees. The delivery of the water requires pressurizing the water, conveying it with pipes, and distributing it with sprinklers, drippers, or bubblers.

Sprinklers can be used with any crop density. Their main disadvantages are distortion of the distribution patterns from wind and leaf damage from salts. Solid-set sprinklers require a relatively large number of sprinkler heads. Fewer emitters are needed for mobile center pivot and side roll sprinklers. A disadvantage of solid-set sprinklers is that their operation is usually intermittent and that as a result the pipes have to be relatively large. Mobile sprinklers can be used more or less continuously.

More continuous operation can also be achieved with surface drip or trickle systems. These systems can be based on relatively small pipes, but require a large number of emitters. They are most economical on orchards and widely spaced and/or valuable row crops. The small diameter orifices clog easily and generally the water needs to be filtered. Accumulation of salts in the «dead zones» between the emitters may also pose a problem, in particular if such salts are washed into the rootzone by an occasional rainstorm.

Localized application can also be achieved with subsurface application with porous tubes and emitters. The advantage is that there is no interference with field operations. Reduction of evaporation from the soil surface is also often cited as an advantage. However, this will in part be offset by increased transpiration, since the weather tends to dictate the sum of the evaporation from the soil surface and the transpiration. One problem is that the surface needs to be kept moist to get germination and the consequent leaching losses may be large. More serious are the uncertainty about subsurface clogging and the accumulation of salts at the soil surface. The tubes or emitters need to be closely spaced, the more so if the soil is coarse textured.

A quite general experience is that continuous or frequent application of water leads to a small root system or at least to a small volume in which the roots are active. This could make plants vulnerable to sudden periods of high demand for water, poorer quality water, or misplacement of fertilizer. For these reasons there is in recent years quite a lot of interest in supplying water at intermediate frequencies from once every few days to once every week. In this context bubblers for small basins around trees are often used.

2. *Some recent theoretical and experimental studies of localized irrigation*

The flow of water from surface and subsurface sources has been treated in numerous theoretical as well as experimental studies (cf. reviews by Bresler, 1977; Raats and Warrick, 1981; Warrick and Raats 1981). Raats and Warrick (1981) chronologically list 28 papers giving two- and three-dimensional solutions of the flow equation using linearizing assumptions. The influence of uptake of water by plant roots is considered in a multidimensional setting only in 3 of those 28 papers (Gilley and Allred, 1974; Thomas et al., 1977; Warrick et al., 1979). However, in a one-dimensional setting the uptake of water has received much more attention (Rawlins, 1973; Raats, 1974 a, b; Warrick, 1974; Rawlins and Raats, 1975; Raats, 1976; Jury et al., 1977; Loman and Warrick, 1978; Dirksen et al., 1981 a, b, c, d). Also the associated buildup of salt in the profile due to exclusion of salts by the plant roots has been treated in some detail (Raats, 1974 a, b; Oster et al., 1974; Raats, 1975; Rawlins and Raats, 1975; Jury et al., 1977, 1978 a, b, c).

In the remainder of this paper I will restrict myself to one aspect of the uptake of water, namely a discussion of some attempts to infer the distribution of the water uptake from hydraulic and salinity data.

3. *Inference of the distribution of the water uptake from hydraulic data*

At any point in the soil, the balance of mass for the water may be written as:

$$\delta\theta / \delta t = - \nabla \cdot (\theta \mathbf{v}) - \lambda T , \tag{2}$$

where t is the time, ∇ is the vector differential operator, θ is the volumetric water content, $\mathbf{v}$ is the velocity of the water, T is the rate of transpiration, and λ is the spatial distribution function for the uptake of the water. The flux, $\theta \mathbf{v}$, of the water is given by:

$$\theta \mathbf{v} = - k \nabla h + k \nabla z , \tag{3}$$

where h is the tensiometer pressure head, k is the hydraulic conductivity, and z is the position in the gravitational field.

Methods for determining the dependence of the hydraulic conductivity upon the water content, including those based on the Boltzmann transform, on measured instantaneous water content and pressure head profiles, and on calculation from water retention data were reviewed by Klute (1972). Arya et al. (1975) introduced a Boltzmann transform method involving rapid evaporation. This method has been used in many recent studies (e. g. Rossi-Pisa, 1978). Also very promising is a method involving a series of infiltration runs with different pressure heads at the input (Dirksen, 1979). If the distribution of the water uptake is to be inferred at a par-

The left hand side of (6) represents the time-averaged rate of uptake from a soil column of unit cross section and extending from z_o to z. The first term on the r.h.s. of (6) represents the time averaged change in water content. The second and third tems on the r.h.s. of (6) represent the time averaged fluxes into and out of the column.

A review of the uses of these methods in later work is in preparation (Raats, 1981). The spatial variability and the errors of the measurements imply that the best results are obtained if either the term $\delta\,\theta/\delta\,t$ is dominant or the flow is steady.

In a drip irrigation system the transport of water may be regarded as periodic, with important periods arizing from the scheduling of the irrigations, the diurnal variation of evapotranspiration, and the seasonal variations associated with crop development and weather. To a first approximation, the behavior of a periodic system may be described in terms of quantities averaged over the period P. For the mass balance this is no problem (Raats, 1974 a), but to evaluate the average flux would require very good data for the space/time distribution of the water content and the pressure head and for the dependence of the hydraulic conductivity upon the water content. It appears that hysteresis of the retention of water would have to be taken into account (C. Dirksen, pers. communication, 1979).

The space/time distribution of the tensiometer pressure sometimes gives a qualitative indication of the water uptake, as will now be illustrated with data from a single orange tree in a large field experiment (see Hoffman et al., 1978, and Van Schilfgaarde, 1977 for details concerning the experiment). To determine the main features of the instantaneous flow pattern, 63 tensiometers were placed under a particular tree in three planes through the trunk of the tree; 22 tensiometers were placed in the row, 23 perpendicular to the row, and 18 along a diagonal. Figures 1 and 2 show the observed total head distributions on April 3, 1975 and on July 31, 1975. The most notable difference between the two days is that on July 31 the soil is everywhere wetter. This is consistent with the sparse tensiometer readings for other trees and as expected from steady flow theory in view of the larger flux and correspondingly higher hydraulic conductivity and water content (Raats, 1974 a, b). Assuming that the soil is isotropic, the streamlines will be everywhere perpendicular to the equipotentials, in the direction of more negative heads. In the region below the drippers, the flow is downward, with the magnitude of the gradients everywhere larger than unity, as a simple, one-dimensional steady flow theory would suggest. On April 3, there is upward flow below about 1,5 m in both planes and also outward flow into a region near the soil surface outside the wetted area in the plane perpendicular to the row. On July 31 the flow is everywhere downward in the row, but again upward at larger depth and outward in the plane perpendicular to the row. In the plane along the diagonal (data not shown) the outward flow is even more pronounced. The data suggest that the leaching below the bottom of the root zone is very nonuniform and, whenever it occurs, largely concentrated in a cylinder with a cross section slightly larger than the wetted surface area. An estimate based on steady

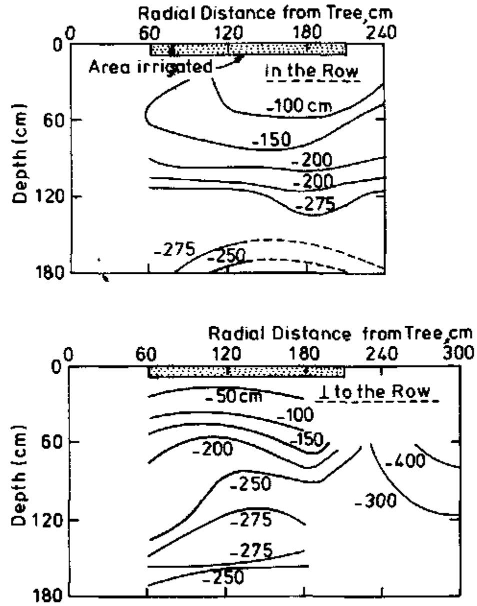

Fig. 1 - Hydraulic head distribution under the center tree of plot H4 in vertical planes in the row and perpendicular to the row on April 3, 1975.

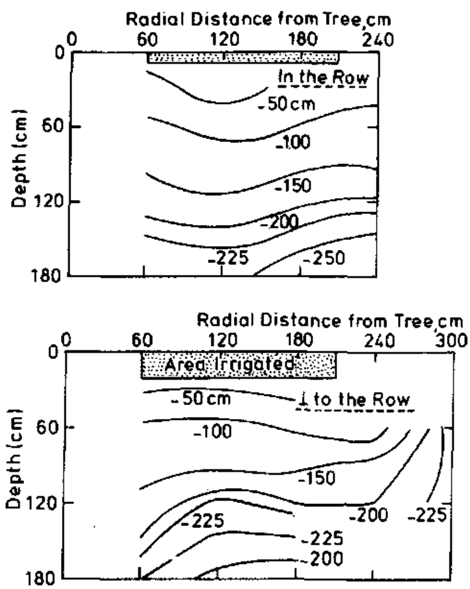

Fig. 2 - Hydraulic head distribution under the center tree of plot H4 in vertical planes in the row band perpendicular to the row on July 31, 1975.

flow in uniform, unsaturated soil suggests that, in the absence of a water table, the flow pattern would approach that of a strip source under the row of trees at a depth of about 5 m, and that uniformity over the whole area cannot be expected above depths of 10 to 20 m. However, stratification probably enhances the lateral flow. (In 1976 there were 21 additional tensiometers installed under the same orange tree; for data see Van Schilfgaarde, 1977 and Dirksen et al., 1979).

4. *Inference of the distribution of the water uptake from salinity data*

At any point in the soil, the balance of mass for a solute may be written as:

$$\frac{\delta}{\delta t}\theta c + \frac{\delta}{\delta z}F_s = -\frac{\delta}{\delta t}\mu_a - \frac{\delta}{\delta t}\mu_f, \qquad (7)$$

where c is the concentration of the solute in the aqueous phase, F_s is the flux of the solute, μ_a and μ_f are the densities of the solute per unit bulk volume in the adsorbed and immobile phases, respectively. The flux F_s is assumed to be the sum of a convective component $\theta v_w c$ and a diffusive component $-D\,\delta c/\delta z$:

$$F_s = \theta v_w c - D\,\delta c/\delta z. \qquad (8)$$

In writing equation (7), it is assumed that the solute is not taken up by the plants. The uptake of the water does affect the distribution of the solute. Roughly, as water is taken up and solute is excluded by the plant roots, the concentration of the solute increases. This implies that the space/time distribution of the solute will in part reflect the distribution of the uptake of the water. This suggests that the distribution of the uptake of the water can perhaps be inferred from the distribution of the solute.

Solving equation (8) for the water flux θv_w gives

$$\theta v_w = F_s/c + D\frac{\delta}{\delta z}\ln c. \qquad (9)$$

Introducing equation (9) into equation (2) and solving for λT gives

$$\lambda T = -\frac{\delta}{\delta t}\theta - \frac{\delta}{\delta t}F_s/c - \frac{\delta}{\delta z}D\frac{\delta}{\delta z}\ln c. \qquad (10)$$

Integration of the mass balance for the solute expressed in equation (6) gives an expression for F_s:

$$F_s = F_{so} - \int_{z_0}^{z} \frac{\delta}{\delta t} (\theta c + \mu_a + \mu_f)\, dz \tag{11}$$

$$= F_{so} - \frac{d}{dt} \int_{z_0}^{z} (\theta c + \mu_a + \mu_f)\, dz . \tag{12}$$

Equations (11) and (12) simply show that the flux of the solute at depth z is equal to the flux of solute at depth z_0 minus the time rate of change of storage between z_0 and z. Introducing equation (12) into equation (10) gives

$$\lambda T = - \frac{\delta}{\delta t} \theta - F_{so} \frac{\delta}{\delta z} c^{-1} - \frac{\delta}{\delta z} D \frac{\delta}{\delta} \ln c$$

$$- \frac{\delta}{\delta z} c^{-1} \frac{d}{dt} \int_{z_0}^{z} (\theta c + \mu_a + \mu_f)\, dz . \tag{13}$$

If the flow is steady and the effects of dispersion and precipitation are negligible then equation (13) reduces to:

$$\lambda T = - F_{so} \frac{d}{dz} c^{-1} . \tag{14}$$

According to equation (14), the rate of water uptake may be calculated as the product of the salt flux F_{so} and the negative of the slope of the dilution profile, $- dc^{-1}/dz$. Gardner (1967) appears to have been the first the realize this. He wrote: «Equation (5) (= (14) above) gives us a relation between the water uptake pattern with depth and the concentration distribution. Since it is easier to measure the concentration than to measure the flux directly, the concentration gradient may give a better measure of w (= λ T above) than the divergence of the flux density. Furthermore, the lower limit of the water uptake can be ascertained from the depth at which the concentration becomes constant».

It appears that equation (14) was not noticed for the following seven years (Raats, 1974 a, b). Oster et al. (1974) applied the method to bromegrass under high frequency irrigation in outdoor lysimeters. The cumulative water uptake distribution estimated from chloride data was 60, 80, and 90% for depths of 15, 30, and 45 cm, respectively. Evaporation losses in the 0-1 cm depth interval accounted for about half of the applied water. Plots of the log of the cumulative uptake as a function of depth were approximately linear. This meant that the distribution of the rate of uptake could be approximated by (cf. Raats 1947 a, b):

$$\lambda = \delta^{-1} \exp - (z/\delta) . \tag{15}$$

For two different lysimeters the rooting depth parameters δ turned out to be 8.5 cm and 9.6 cm, respectively. Assuming a dispersion coefficient of 0.05 cm^2/day, the third term on the right hand side of (13) had a negligible effect on the estimate of δ. A subsequent calculation showed that only if the leaching fraction is very small the influence of dispersion will be noticeable (Raats, 1977). If electrical conductivity data are used as a basis, then dissolution/precipitation described by the last term on the right hand side of equation (12) also needs to be considered. In the lysimeters the sum of the mineral equilibria and diffusion corrections to the rate of uptake was zero to the 15 cm depth. At greater depths the mineral equilibria correction was dominant and increased the calculated rate of uptake by as much as 30%.

The steady state distribution of chloride was also used to estimate the distribution of the water uptake under the orange tree for which some tensiometric data are shown in Figs. 1 and 2 (Van Schilfgaarde, 1977; Dirksen et al., 1979). The cumulative relative water uptakes were 64, 86, 93, 97, and 98%, respectively for depths of 0.3, 0.6, 0.9, 1.2, and 1.5 m, respectively, corresponding roughly to a rooting depth δ of 0.4 m. This information can in turn be used in a simple model for the time course of a parcel of water through the root zone (Raats, 1975). Assuming $\theta = 0.5$, $T = 7$ mm/day, and $\delta = 0.4$ m, figure 3 shows such time courses for leaching fractions 5, 10, and 20% (cf. Fig. 15 of Van Schilfgaarde, 1977).

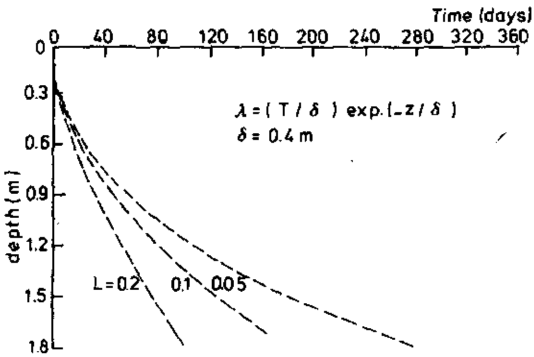

Fig. 3 - Travel times of water.

In a laboratory study of space/time distributions of matric and osmotic potentials of daily irrigated alfalfa, Dirksen et al. (1981 b) estimated the distribution of the water uptake on two different days from hydraulic data and from the salt flux and salinity sensor readings on another day. The agreement between the two estimates was good. Between 80 and 90% of the uptake occured above 0.50 m.

Most recently Jury et al. (1980 a, b, c) used soil salinity sensor and chloride data to estimate fractional water uptake above a depth of 5 cm and in the layer 0-20 cm, respectively, in a lysimeter experiment with wheat and sorghum. Corrections for precipitation were made by using the chemical equilibrium model of Oster and Rhoades (1975) and calculating the electrical conductivity of the resulting mixed salt solutions by the method of McNeal et al. (1970). It turned out that 50% or more of the water was evaporated or was taken up within 5 cm from the soil surface.

Thus far, only one-dimensional flows have been discussed in this section. To infer anything about the distribution of the water uptake from the distribution of the salinity, one must have separate information about the flow pattern. For example, it can be shown that the proper generalisation of equation (14) to multidimensional flows is given by

$$\lambda = -(A_o/A)\, F_{so} \frac{\delta c^{-1}}{\delta s}, \tag{15}$$

where s is the directional derivative along a streamline, A is the cross sectional area of a stream tube, and the subscript o indicates a reference point along the same stram tube. A logical next step would be to set up some experiments involving localized irrigation and sufficiently detailed measurements of the salinity distribution and the flow pattern so that equation (15) can be used.

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