In equation (28) θ is expressed as a function of place (x) and time (t). The solution of this diffusion equation, provided the diffusion coefficient is constant can be obtained for a variety of initial and boundary conditions. Such a solution is either comprised of a series of error functions or related integrals, in which case it is most suitable for numerical evaluation at small times, or it is in the form of a trigonometrical series which converges most satisfactorily for large values of time.

The transformation described above and used by various investigators provides an equation from which solutions in the form of profiles of θ at given terms t may be obtained. For these solutions one first has to know how K varies with the moisture content. Secondly, with this information the differential equation has to be solved subject to the initial and boundary conditions appropriate for the case under considerations.

In employing the diffusivity concept, and all relationships derived from it, we must remember that the diffusivity equations fail wherever the hysteresis effect is appreciable or where the soil is layered, or in the presence of thermal gradients, since under such conditions flow bears no simple of consistent relation to the decreasing water-content gradient and may actually be in the opposite direction to it. On the other hand, an advantage in using the diffusivity equations is in the fact that the range of variation of diffusivity is smaller than that of conductivity. The maximum value of $D(\theta)$ found in practice is of the order of 10^4 cm^2/day. $D(\theta)$ generally decreases to about 1-10 cm^2/day at the lower limit of wetness, normally encountered in the root zone. It thus varies about a thousandfold rather than about a millionfold as does the hydraulic conductivity in the same wetness range. Furthermore, the wetness and its gradient are often easier to measure in practice, and to relate to volume fluxes, than the matric potential and its gradient.

The solution of eqn. (24) and all derived relationships in because of its non linearity, much more difficult to solve than are the classical linear differential equations for the flow of heat or electricity. The solution of such equations depends as already indicated on the boundary and initial conditions of the problem. Analytical solutions are available for simple boundary conditions. They are based either on the Laplace transform or Boltzmann transformation.

The Laplace transform is a mathematical method by which the unsaturated diffusion equation, including its boundary and initial conditions are transformed into an ordinary differential equation with corresponding boundary and initial condition. The in this way derived ordinary differential equation will then be solved classically subject to its boundary and initial conditions. Of this solution the socalled inverse Laplace transform will be taken to yield the solution to the original partial differential equation.

In 1894 BOLTZMANN showed that for certain boundary conditions, provided the diffusivity is a function of moisture content (θ) only, θ may be expressed in terms of a single variable $x/t^{\frac{1}{2}}$ and that eqn. (27) may therefore be reduced to an ordinary differential equation by the introduction of a new variable, $\lambda(\theta)$ where:

$$\lambda(\theta) = x/t^{\frac{1}{2}} \tag{29}$$

The transformation, eqn. (29), can be used when diffusion takes place in infinite or semi-infinite media of uniform initial wetness; $\theta = \theta_i$ for $x > 0$, $t = 0$; and $\theta = \theta_o$ for $x = 0$, $t > 0$. The new variable, λ, is simply a mathematical consequence of the form of the differential equation. When an actual experiment fails to conform accurately to the $x/t^{\frac{1}{2}}$ relation, the discrepancy can only attributed to an imperfect description of the behavior of the soil system by the assumed differential equation and/or its assumed boundary conditions, or to errors of the experiment.

For more complicated conditions solutions are either unknown on they become very complex. Solutions may then be found through numerical methods. The numerical solutions of the diffusion equations are obtained by replacing some or all of the derivatives by finite-difference or finite-element approximations. Numerical methods of solution of the diffusion equation for various kinds of problems (i.e., initial and boundary conditions) may be found in specialized books. Numerical solutions can be highly accurate but as the calculations are lengthy, they often require the use of a computer. However, in all the solutions available, ranging from the simple to very complex situations, with moving boundary conditions, it is assumed that D, the diffusivity, is an unique function of soil moisture. The validity of this assumption is that hysteresis is not present.

As pointed out above, first of all the relation between the hydraulic conductivity and the moisture content must be known for solving flow problems in unsaturated soil. In order to calculate the diffusivity the moisture characteristic must be available. Otherwise the conductivity may be derived from D-values and the moisture characteristic of the soil.

There are various methods applied to determine the hydraulic conductivity and diffusivity of unsaturated soil. The principles of the methods described in literature are given in next section.

4. The determination of hydraulic conductivity and diffusivity in unsaturated soil.

The methods for measuring unsaturated hydraulic conductivities or soil water diffusivities can be placed into two groups-those based on steady state methods and those based on transient state procedures. Some steady state

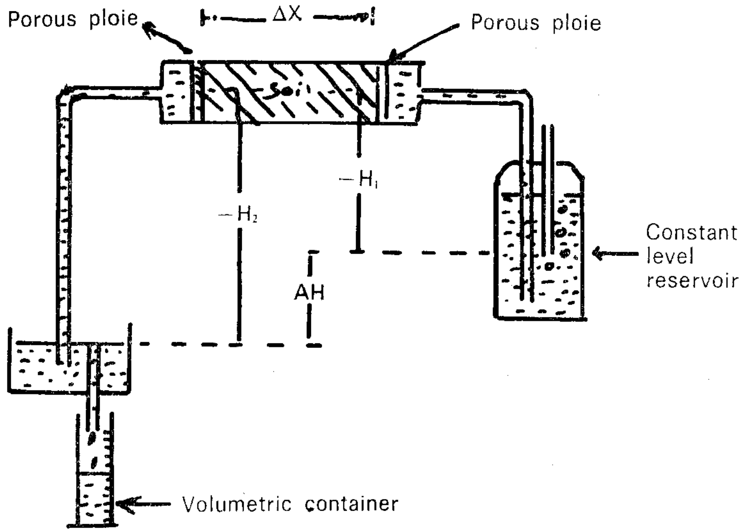

Figure 7.: A model illustrating unsaturated flow under a suction gradient in a horizontal column.

methods make use of a long soil column, generally saturated at one end with controlled water loss at the other. The amount of water flowing through the soil and the potentials within the soil are measured and the conductivity calculated. If only water contents are measured, then D can be calculated. In another steady state method a soil sample is clamped between two porous plates which are maintained at different soil-water potentials. The amount of flow and the potentials within the soil are again measured and the soil conductivity calculated. Steady-state methods are primarily laboratory methods. Both disturbed and undisturbed samples may be used. The use of small size core samples introduces a serious sampling problem when one wishes to characterize a field soil volume. Long-column methods are therefore more suitable.

In the unsteady-state methods the time dependence of some aspect of the flow system is used to obtain the conductivity or soil-water diffusivity. The methods may be subdivided into: (1) outflow, - inflow methods and (2) instantaneous profile methods. Results can be obtained in a shorter time with the outflow methods, and different water contents are easily obtained. However, there are more uncertainties in this method than in the steady state methods.

Steady state methods.

In this methods a potential gradient is maintained across a sample by applying a different suction to the end plates (see figure 7). When the flow rate is constant, the Darcy equation can be used, written in finite difference form, to calculate the hydraulic conductivity.

$$v = Q/At = -\overline{K}\,\frac{\Delta\psi_m}{\Delta x} = -\overline{K}\,\frac{\Delta h}{\Delta x} \tag{30}$$

The potentials can be taken from the tensions at the end plates or from tensiometers inserted in the soil. The value of the conductivity obtained is associated with the mean suction heads and mean water content at which the flux and gradient were measured provided that the column of figure 7 is sufficient short.

The conductivity function is mapped by proceeding through a series of steady-state flows with progressively more and more negative values of h, beginning with $h \geq 0$. The bubbling pressure of the porous barriers (the pressure required to force air through the wetted barrier) must be at least as large as the magnitude of the most negative pressure head to be used in the instrument. This apparatus can be used for cores of undisturbed soil. It is limited to suctions below 1 bar, and in practice to below 0.5 bar. Furthermore, it is best if the pressure head difference should be made as small as possible in order to minimize the effect of the gravitational force on the moisture distribution in the vertical direction. When this effect is real, the conductivity in the sample is then not only a function of position along the flow column, but also of position in the vertical direction across the sample.

A greater range in potentials can be maintained by supplying water at one end of a soil column and allowing evaporation at the other (see figure 8). At steady state, for a vertical soil column, the appropriate finite difference approximation of the flow equation is:

$$v = -\overline{K}\left(\frac{\Delta\psi_m}{\Delta z} + 1\right) = -\overline{K}\left(\frac{\Delta h}{\Delta z} + 1\right) \tag{31}$$

When the system reaches steady state (indicated by no further change in tensiometer readings or change in moisture profile), the tensiometer readings are converted to weight matric potentials and plotted as a function of height above the water table. Using the slope of this curve as the gradient in eqn. (31) allows one to calculate the unsaturated conductivity for a large number of matric potential values, whereas if the data were not plotted, only a limited number of gradients are available for unsaturated hydraulic conductivity calculations. Unfortunately, the $K - \psi_m$ relation that can be measured in this way is limited through the limite range of measurement of tensiometer (0,85 bars). This range can be extended if thermocouple psychro-

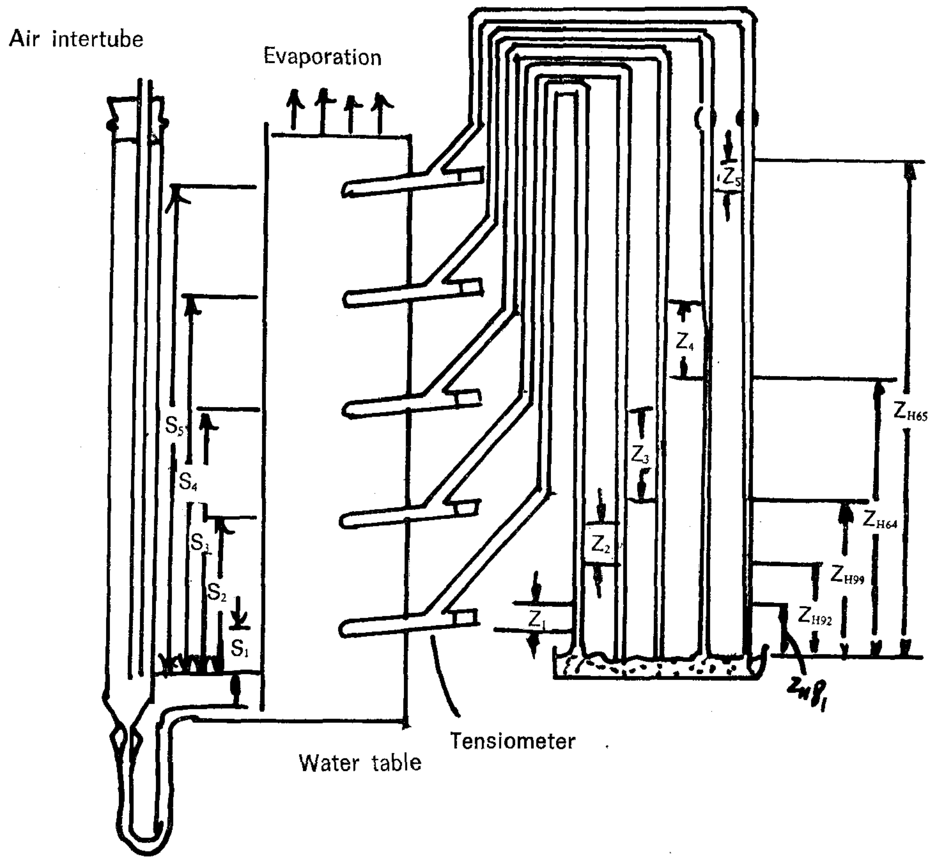

Figure 8.: Apparatus for measuring unsaturated hydraulic conductivity based upon steady state evaporation from a soil column.

meters which can measure potentials between — 1 and — 20 bar may be applied. Alternatively, the soil column can be sliced for θ determination or θ can be measured non-destrustively by gamma ray-attentuation. A range of water contents is obtained from wet at the inlet to nearly dry at the evaporation end. For any two adyacent slices of soil:

$$D = (Q/At) \cdot (L/\theta_2 - \theta_1) \qquad (32)$$

where: L = distance between slices:
θ_1 , θ_2 = volumetric water content of adjacent slices.

From these values a curve of D vs. θ can be drawn. Values of K can be obtained with the additional measurement of the water retention curve on other samples. The later introduces the problem of variability between samples.

YOUNGS (1964) have described another method to evaluate the hydraulic properties of the soil medium as a whole. The principle of the method is that a continued supply of water to the soil, as under sprinkling, at a constant rate lower than the saturated hydraulic conductivity of the soil eventually results in the establishment of a steady moisture distribution in the conducting profile. Once steady state conditions are established, a constant flux exists. In a uniform soil the suction gradients will tend to zero so that the

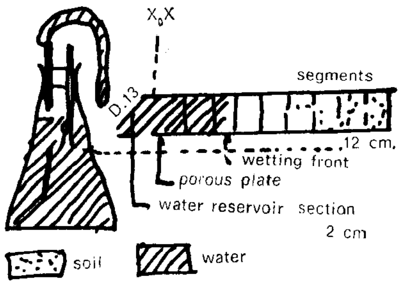

Figure 9.: Experimental set-up for horizontal infiltration. The method is based on the introduction of the Boltzmann transform ($\Lambda = x/t^{\frac{1}{2}}$) to transform the soil-water diffusivity equation to an ordinary differential equation. Under the conditions of
$t = o,\ x > o,\ \theta = \theta_i$
$t > o,\ x = o,\ \theta = \theta.$
$> \theta_i$
advances a plane of constant water content proportionally to the square root of the infiltration time. This can easily be verified by plotting the advance of the wetting front (a plane of constant water content) as a function of the square root of infiltration time. Usually this plot is a straight line unless the soil exhibits considerably swelling when wetted.

hydraulic conductivity becomes essential equal to the flux. If this test is carried out for a series of application rates (sprinkling intensities) it becomes possible to obtain different values of hydraulic conductivity corresponding to different values of soil moisture content. The difficulty of the steady sprinkling infiltration test in the field is that it requires rather elaborate equipment which must be maintained in continuous operation for considerable periods of time. The operation becomes increasingly difficult as one attempts to reduce the application rate to the order of 1mm/hour or even less.

The infiltration methode described by HILLEL and GARDNER (1970) makes use of the hydraulic resistance of a membrane or crust at the soil surface to decrease the soil water content, and correspondingly the pressure head, the conductivity values of the infiltrating profile. The lower the hydraulic conductivity of the crust, the more negative is the pressure head in the soil. Estimates of K and D can be obtained during the transient stage (= unsteady state method) of infiltration. However, the most reliable measurements are obtained by allowing the infiltration process to proceed to a steady state, when the flux becomes equal to the conductivity. The use of a series of crusts of progressively lower resistance can give progressively higher K-values corresponding to higher water contents up to saturation. Such a serie of tests, can be carried out if the soil is initially fairly dry, either successively in the same location or concurrently on adjacent locations.

The main disadvantage of steady-state methods is that they require relatively long times to establish steady flow. During this time, changes can occur in the hydraulic properties of the sample due to biological activity. The use of mercuric chloride, phenol or thymol helps to reduce this effect.

Unsteady state methods.

The outflow method was first described by GARDNER (1956) and subsequently refined by many other research-workers. The system used consist of a soil sample in a pressure membrane apparatus. The outflow of water as a function of time is measured and from this it is possible to evaluate the permeability. For small pressure steps, i.e. small $\Delta\psi_m$ values, the conductivity $K(\psi_m)$ can be assumed to be constant over the interval and the flow equation becomes:

$$\frac{\delta\theta}{\delta t} = \frac{\delta\left[K(\psi_m)\,\frac{\delta\psi_m}{\delta x}\right]}{\delta x} = K(\psi_m)\,\frac{\delta^2\psi_m}{\delta x^2} \tag{33}$$

This equation has the form of an ordinary diffusion equation and solutions are available for a wide variety of boundary and initial conditions (cfr. GARDNER, 1956). By repeating the measurements over a succession of small increments of potentials ($\triangle\psi_m$) a series of permeability-potential values is obtained.

BRUCE and KLUTE (1956) described a method in which the spatial distribution of water content at a fixed time in a horizontal infiltration flow system was used to calculate the diffusivity function (see figure 9).

The method is based on the introduction of the BOLTZMANN transform ($\lambda = x/t^{\frac{1}{2}}$) to transform the soil-water diffusivity equation to an ordinary differential equation. Under the conditions of

$$t = 0,\ x > 0,\ \theta = \theta_i$$

$$t > 0,\ x = 0,\ \theta = \theta_o > \theta_i\,,$$

advances a plane of constant water content proportionally to the square root of the infiltration time. This can easily be verified by plotting the advance of the wetting front (a plane of constant water content) as a function of the square root of infiltration time. Usually this plot is a straight line unless the soil exhibits considerably swelling when wetted.

The solution of eqn. (27) if the water content at the inflow boundary remained constant is:

$$D(\theta_x) = -\frac{1}{2} \cdot \frac{1}{(d\theta/d\lambda)_{\theta_x}} \cdot \int_{\theta_i}^{\theta_x} \lambda \, d\theta \tag{34}$$

where: $D(\theta_x)$ is the diffusivity corresponding to θ_x, θ_x being the moisture content at time t at a distance x from the inflow boundary;

$(d\theta/d\lambda)_{\theta x}$ represents the slope of the curve relating θ with λ at the point where $\theta = \theta_x$;

$\int_{\theta_i}^{\theta_x} \lambda d\theta$ can be determined, for each, value of θ, from the shaded area below the curve (λ, θ) between θ_i and θ_x (see figure 10).

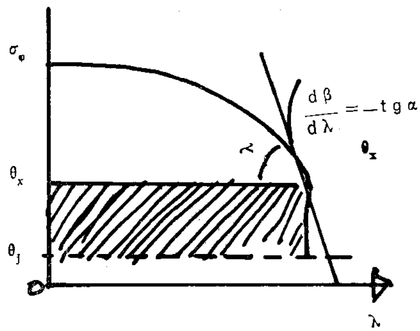

Figure 10.: Water content distribution at end of infiltration test with set up of figure 9.

While the method based on eqn. (34) is suitable for gravimetric sampling for θ, the water content can also be measured with gamma absorption measurements. The soil columns must be uniform with respect to water-holding and transmitting properties. The method has usually been applied to disturbed samples in relatively long columns (20 cm or more). The initial water content is usually that of the air-dry soil material and it is most convenient to maintain the boundary x = 0 at or near saturation. The latter can be done with a screen or porous plate with a very low bubbling pressure and high conductivity. With inflow or infiltration only the wetting $D(\theta)$ function is obtained. However, ROSE (1968) has applied the same principles to evaporation of water from columns of soil aggregates. Experimental tests

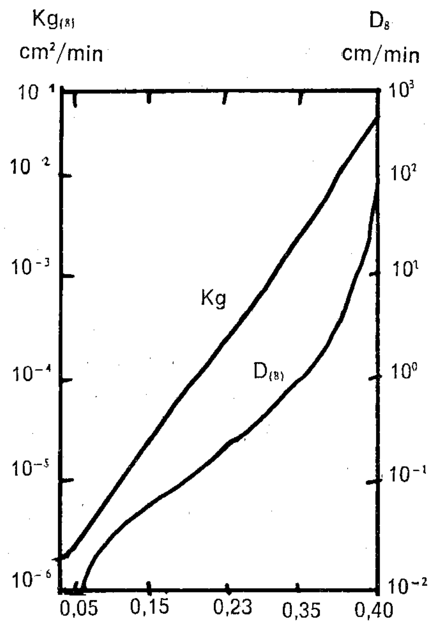

Figure 11.: Experimental values of K and D as functions of obtained from the θ (Λ) function of figure 10. (Columbia silt loam, c.f. Davidson et al., 1963).

showed that θ was a function of λ only as the columns remained effectively semiinfinite and thus equation (34) could be applied to calculate the drying function of $D(\theta)$.

The direct result of these methods is a diffusivity function $D(\theta)$. If a conductivity function is needed, the water capacity function must be obtained from a water retention curve measured on other samples. Resulting curves for K and D as function of θ are given in figure 11.

The hydraulic conductivity can also be determined in the field or the laboratory, over the entire range of water contents whatever the nature of the soil profile by means of the instantaneous profile method, described by WATSON (1956). In this method the hydraulic conductivity-water relationship is determined from measurements obtained during the drainage of an initially saturated column covered at the surface with a plastic sheet to prevent any water flux across the soil surface. The approach utilizes instantaneous profiles of the macroscopic flow velocity, the potential gradient and the water content at any instant of time after the commencement of drainage. Once these are known for a particular time, it is then possible to find the instantaneous hydraulic conductivity for each elevation by dividing the appropriate velocity value by the potential gradient value. Since the water content profile is known at the same time, a series of points on the instantaneous hydraulic-conductivity-water content relation is available.

The method is based on the applicability of Darcy's law in unsaturated soil:

$$q = -K \frac{d\psi_h}{dz} \tag{35}$$

where:

q = the water flux in $cm.sec^{-1}$

K = the hydraulic conductivity in $cm.sec^{-1}$

$\frac{d\psi_h}{dz}$ = the potential gradient, with ψ_h the hydraulic potential ($= \psi_z + \psi_m$).

Application of the water conservation equation, that in an one dimensional unsaturated flow system may stated as:

$$\frac{\delta\theta}{\delta t} = -\frac{\delta q}{\delta z} \tag{36}$$

leads to:

$$q_{z_i} = \int_0^{z_i} \frac{\delta\theta}{\delta t}\, dz \tag{37}$$

where: q_{z_i} = the flux at a depth z_i.

Combinaton of eqn. (35) and (37) gives:

$$q_{z_i} = \int_0^{z_i} \frac{\delta\theta}{\delta t}\, dz = -K \frac{\delta\psi_h}{\delta z} \tag{38}$$

or:

$$K = -\frac{v_{z_i}}{\delta\psi_h/\delta z} = -\frac{\int_0^{z_i} \frac{\delta\theta}{\delta t}\, dz}{\delta\psi_h/\delta z} \tag{39}$$

The ratio of the flux and hydraulic gradient at the given position and time is then the hydraulic conductivity at the water-content found at that position.

The calculations proceed as follow:

— Obtain the distribution of (θ, t) and (ψ_m, t) for several column elevations (z). These curves give the variation of water content and matric potential with time for each section and/or registrating depth of the pressure head.

— From the plot (θ) versus (t), evaluate for each section $\delta\theta/\delta t$ at a series of values of (t).

— From these data the soil moisture flux through each section can be calculated as $\Delta z\ (\delta\theta/\delta t)$. At the required time (t), the flux throughout the profile can be obtained by summing up foregoing values or $q = \Sigma\ \Delta z\ (\delta\theta/\delta t)$.

— Using this information it is a straightforward matter to find the hydraulic conductivity, after having plotted (ψ_h) versus depth (z) at the corresponding values of (t).

This method seams to offer the best possibility for hydraulic characterisation of relatively homogeneous field soils. The method as described is not applicable where lateral movement of soil moisture is appreciable and relatively slowly permeable soil horizons, such as commonly occuring plow layers or certain genetic B-horizons, result in unequal wetting of the soil during initial saturation. Under this circumstances it is impossible to saturate completely the profile, which implies that this method for horizons below a slow permeable horizon will never yield K-values for potentials between saturation and $-$ 35 cm to $-$ 50 cm. The crust test, which performs well at high soil water potentials, but which is time-consuming and unreliable at very low flow rates could previde these data.

Water content should be measured with the neutron method, while the pressure head with tensiometers for values below 0.8 bar, or taken from a water-retention curve at the measured θ for higher suctions.

Several variations in the flow system (drainage, evaporation, infiltration and redistribution) and data analysis have been employed to the instantaneous profile method in laboratory and field measurements. A complete review is given by KLUTE (1972).

Due to the difficulty of removing samples of field soil without disturbance, and the difficulties in measuring K have lead soil physicists to look for methods of calculating K from other soil properties. There are many publications that deal with the relationship of conductivity to various aspects of pore space geometry often obtained from water retention data. See BRUTSAERT (1967) for a review.

According to TAYLOR (1972) we may conclude that: ‘ All of the methods now in use for measuring the unsaturated hydraulic conductivity or the water diffusivity have some disadvantages. Since non of them is completely satisfactorily, it is propably best to use a combination of methods in order to get the desired relations. If the values agree reasonably well when determined by two or more methods, then one would have reasonable assurance that the results are valid.

5. A PRACTICAL APPROACH FOR TRICKLE IRRIGATION MANAGEMENT.

One of the basic aims of trickle irrigation is that water should discharged out an orifice at a controlled rate directly on the soil surface so that the area across which infiltration takes place is very small compared with the total soil surface. As a result, one has a case of three-dimensional transient infiltration of water into the soil. Under the assumptions that:

— the soil is a stable, isotropic, and homogeneous porous medium;

— the initial water content of water is negligible everywhere in the system;

— that the water suction and the hydraulic conductivity of the soil are single-valued, unique and continuous functions of soil water content; what is realistic if the soil water content at any point in the system cannot decrease with time, the flow of water in the system can be expressed through the differential eqn. in terms of the diffusivity form (see eqn. 24).

The mathematical tools to analyze such a multi-dimensional transient infiltration from a trickle source have been developed. BRANDT et al (1971) solved the conventional diffusion type water flow equation numerically for both plane and cylindrical flow. WARRICK (1974) linearized the unsteady-state unsaturated flow equation by introducing the concept of the matric flux potential as did GARDNER (1958), and assuming that the differential water capacity is a constant. A condition valid if the values of θ (or ψ_m) very over a small range or at least a small range about a steady state. Such conditions might well exist in the case of trickle or high-frequency irrigation. The advantage of the linearized form of the moisture flow equation is that analytical solutions exist for many problems of interest.

The main reason of these theoretical mathematical analysises is to get an idea of the moisture front advance in function of the local soil dynamic parameters, either under continuous or cyclic water application.

The analysises become very complicated if multiple sources are in use to replenish the root zone, or if one wants to take into account the plant-water withdrawl. Although the addition of simple water extraction patterns would appear to be a logical extension to the excisting comprehensive numerical models, I am afraid that the solutions based either on finite difference or finite element schemes, on linearizing assumptions or other excisting mathematical methods will still fail to simulate the evolution of the moisture profile, as long as one cannot incorporate hysteresis, the inhomogeneity of field soils and the real soil-moisture extraction pattern. Last parameter variable with time, depends upon the crop being grown, its stage of grouwth and other crop factors.

Anyhow, if one wants to approach the shape of the wetted soil volume, the moisture distribution within the wetted volume and the plant-water withdrawl by means of model studies a more than advanced mathematical and soil physical knowledge is required. Question is, if one is not able to look at the water movement from point sources or even several point sources in a more simplified fashion. Much can be learned from tests carried out on existing installations if there are such in the vicinity of a new project. If there are none, and if a typical area can be found, a simple set of field test will enable the engineer to arrive at a satisfactory design.

Before starting the experiment it is desirable to remove the native vegetation if one is only interested in the relation between the wetting and the water conducting characteristics of the soil. Provision must also be taken that the soil surface is a smooth horizontal plane, that water is applied at typical trickle irrigation flow rates for selected durations and that the number of emitters per plant is in relation to the spatial extent of the rooting pattern.

At the termination of each water application a trench can be dug approximately 20 cm from the center of the wetted soil volume. The trench will provided access to a vertical diameter plane through the wetted volume. After the trench is dugged, the shape of the wetted pattern can easily be identified by the color change between wet and dry soil. A 15 cm grid can be imposed on the wetted pattern and horizontal soil samples taken, at each node point. These samples are necessary, to determine the soil moisture content on a dry weight basis at these points. By taking the soil samples the first 5 cm should be discarded since it will been exposed to the air and dried due to evaporation. Horizontal soil samples should be also taken in 15 cm increments outside the wetted volume to determine antecedent soil moisture conditions.

Furthermore the moisture release characteristics for the soil should be determined using a hanging water column as described by BAKER et al (1974). Saturated and unsaturated hydraulic conductivity are also necessary parameters to explain the experimental results. This can be done by one of the methods described in previous section.

Suppose that in a first approach the unsaturated conductivity may be described by

$$K(\theta) = K(\theta_S) \left[\frac{\theta - \theta_{FC}}{\theta_S - \theta_{FC}}\right]^3 \tag{40}$$

where:

$K(\theta)$ = the unsaturated conductivity at the moisture content θ expressed in cm.day^{-1};

$K(\theta_S)$ = the saturated conductivity in cm.day^{-1};

θ = the moisture content at any location between saturation and field capacity in volume percent;

θ_{FC} = the moisture content at field capacity;

θ_S = the moisture content of the saturated soil.

Eqn. (40) allows for $\theta = \theta_{FC}$, $K(\theta) = 0$, which implies that at field capacity the conductivity decreases to a small value, negligible compared to $K(\theta_S)$. Transformation of eqn. (40) leads to:

$$\left[\frac{K(\theta)}{K(\theta)_S}\right]^{1/3} = \frac{\theta - \theta_{FC}}{\theta_S - \theta_{FC}}$$

or:

$$(\theta_S \cdot \theta_{FC}) \cdot \left[\frac{K(\theta)}{K(\theta_S)}\right]^{1/3} = \theta - \theta_{FC}$$

$$\theta_S - \theta_S - \theta + \theta_{Fc} = -(\theta_S - \theta_{FC}) \cdot \left[\frac{K(\theta)}{K(\theta_S)}\right]^{1/3}$$

$$\theta_S - \theta = (\theta_S - \theta_{FC}) - (\theta_S - \theta_{FC}) \cdot \left[\frac{K(\theta)}{K(\theta_S)}\right]^{1/3} \tag{41}$$

$$(\theta_S - \theta) = (\theta_S - \theta_{FC}) \cdot \left[1 - \left[\frac{K(\theta)}{K(\theta_S)}\right]^{1/3}\right]$$

where:

$\theta_S - \theta$ = represent the air filled pore space at the moisture content θ and

$\theta_S - \theta_{FC}$ = the air filled pore space at field capacity.

Assume that we may replace the unsaturated conductivity through the infiltration flux (q), eqn. (41) may be written as:

$$(\theta_S - \theta) = (\theta_S - \theta_{FC}) \cdot \left[1 - \left[\frac{q}{K(\theta_S)}\right]^{1/3}\right] \tag{42}$$

From eqn. (42) we learn that for very low values of q the ratio $q/K(\theta_S)$ to the power 1/3 is negligible so that under this circumstances

$$\theta_S - \theta = \theta_S - \theta_{FC} \tag{43}$$

Which physically means that during infiltration at a certain radial distance from the point source, where the moisture flux justified this assumption, the air filled porosity will never decrease below 10 or 12 percent of the pore volume, which in normal agricultural soils is filled with air at field capacity. In soils in which less than 10 or 12 percent of their volume is filled with air aeration is in adequate. It is necessary that during operation of a trickle system this requirement is met in at least 30 to 50% of the rooting volume. Furthermore if the water supply is stopped when the active root zone is replenished to field capacity deep drainage losses due to redistribution will be low there according to eqn. (40) $K(\theta) = 0$, for $\theta = \theta_{FC}$. However, if irrigation water is saline, this criteria does not held. Deep drainage is necessary to leach the excess of salt.

From this we learn that the experimental set up, combining emitter discharge, number of emitters, geometrical situation of the emitters and duration of a single application, that fulfill these two requirements can be selected as the most adequate for the given crop on the given soil. Once these parameters are fixed, it is a straightforward procedure to work out a complete trickle design. Attention should be paied that for each combination of emitter discharge several plots will be necessary in order to test the length of a single application.

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