SOME PHYSICAL ASPECTS OF TRICKLE IRRIGATION

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INTRODUCTION.

Water stress affects practically every aspect of plant growth. In summary decreasing water content is accompanied by loss of turgor and wilting, cessation of cell enlargement, closure of stomata, reduction in photosynthesis, and interference with many basic metabolic processes. Eventually, continued dehydration causes disorganization of the protoplasm and death of most organism.

Plant water stress or water deficits develop when water loss exceeds absorption. Temporary midday water deficits occur in rapidly transpiring plants because the resistance to water movement through roots causes absorption to lag behind transpiration even in moist soil. Longer-term and more severe water deficits develop when decreasing soil water potential and hydraulic conductivity cause decreased absorption of water. Daily cycles in water stress are controlled chiefly by transpiration; long-term cycles are due to severe water deficits, chiefly because of decreasing availibility of soil water.

The objective of a good trickle irrigation system is to provide that during the growing season there is always enough water available to the roots. The system tries to apply the water in such a way that the soil at a specific site near each plant is kept between field capacity and a value of the matrix potential depending on the crop being grown. That value of the matric potential is meant at which water should be applied for maximum yields. The system can even be adjusted to replace water continually as it is used up by the plant. Under this circumstances the rate of water supply to the roots will be sufficient to prevent severe water deficits in the plant.

Midday stress can only be avoided by reducing temporary the transpiration rate. It means adjusting the water loss rate to the water uptake.

Briefly this can be adchieved by:

— growing plant varieties that transpire less;

— reducing the drying capacity of the air over the crop;

— using chemical antitranspirants. The present available chemical antitranspirants work by clossing stomata, forming a transparent barrier over the stomata to the escape of water vapor or by cooling the leaf with a reflecting coat that reduces the amount of solar energy absorbed. Unfortunately, the use of antitranspirants is mainly experimental today. Their use would be a step forwards in reducing moisture stress during water-sensitive growth stages (e. g. transplanting of germination). However, it requires more study before it will become feasible on a large scale, if indeed this can ever be done.

The question arises of an uniform level of water potential near field capacity is always desirable between germination and harvesting. A good knowledge of the effect of soil moisture potential on the plant at different stages of its development is necessary. An answer to this and the water requirements of different crops are included in the section ' Plant needs for water '.

The design of a trickle system, as well as any other irrigation system, depends on the availability of soil moisture to the plant, which itself is a function of the soil water potential and of the hydraulic conductivity of the soil. An understanding of the retention and transports of water in the unsaturated zone of the soil enables one to define an application rate that assures a sufficient water supply to the root system while deep drainage is prevented. Furthermore, for each design, an optimal relation had to be sought between:

— the emitter discharge;

— the number of emitters;

— the duration of a single application and

— the location of the emitters with respect to the plant.

These parameters should be selected in function of the physical behavior of the soil, the rooting pattern of the crop and the abstraction rate of the roots in such a way that at the end of each irrigation cycle only the rooted soil volume is re-wetted to the desired potential. Only a detailed study part-empirical, part-theoretical can give an answer to these questions.

1. Plant needs for water.

Transpiration removes much water from plant tissue. 95% of the water absorbed by roots moves up through the plant and passes into the atmosphere as water vapor. The potential of the water in the plant tissue is reduced rapidly when the water loss is not immediately replaced. The decrease of water potential in the xylem sap causes loss of guard cell turgor and closure of stomata. This results in a compensating decrease in water loss by transpiration. This illustrates that the rates of absorption and transpiration are kept in balance through the continuity of water in the conducting system which provides a communication system between roots and shoots. Thus when transpiration increases, the demand for an increased water

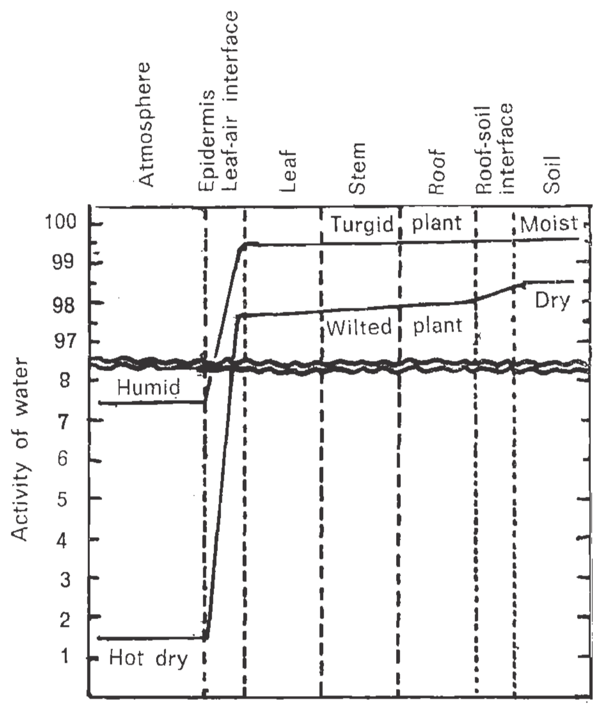

Figure 1.: **Diagram** of the activity of water (water potential) along the transpiration stream.

supply to the leaves is transmitted to the roots by a decrease in water potential in the xylem sap, which causes an increase in absorption. Actually, there is some lag in response which depends on the conductivities in the plant water system (the reciprocal of the root-, stem- and leaf resistance) and on the supply form the soil to the root. The replenishment rate also depends on the water potential gradients that develop along the flow path form the transpiring tissue to the source of water in the soil. It is assumed that the movement of water through the plant to the air is proportional to the difference in water potential, which is the driving force, and inversely proportional to the resistance in the pathway. As shown in figure 1, the water potential drop is proportional to the resistance encountered along the flow path. The resistance to flow is relatively low in the plant, being highest in the roots, intermediate in the leaves, and lowest in the stems, where movement is largerly in the vascular system. The main resistance to flow is encountered in the vapor pathway, at the leaf-air interface.

Transpiration is the dominant factor in plant water relations because it produces the energy gradient which causes the movement of water into and through plants. In addition transpiration does increase the rate of upward movement of salt, and it might be asked whether enough salt would reach the leaves of plants in the absence of transpiration. Furthermore, the transpiration tends to cool the leaves. Leaves which are transpiring very lowly because wilting, closure of stomata or other reasons are usually considerably warmer than the air. We can agree with CURTIS (1926) that transpiration is an unadvoidable evil, unadvoidable because of the structure of the leaves and evil because it often results in water deficits and injury by desiccation.

Although basically the transport of water through the entire soil-plant-atmosphere system is passive, it is affected by physical factors. The chief environmental factors are light intensity, vapor pressure and temperature of the air, wind and water supply to the root. Plant factors include the extent and efficiency of the root system as an absorbing surface; leaf area, leaf arrangement and structure, and stomatal behavior. If any one of these factors is altered enough to cause a change in rate of transpiration, other factors will be changed, causing further adjustments in the transpiration rate.

The rate of transpiration depends on the supply of energy to vaporize water, the water vapor pressure or concentration gradient which constitutes the driving force, and the resistances to diffusion in the vapor pathway. Leaf resistance is variable. Changes in light intensity cause variation in leaf resistance, through its effect on stomatal aperture. Normally, during daytime, light intensity is high enough to have no effect on the transpiration rate. Leaf resistance is more frequently affected by the supply of water to the roots. In generally it is assumed that in soils near field capacity, the movement of water toward roots is sufficiently to meet the desiccating rate of the atmosphere. As soil water is depleted, the supply of water to the roots may become a limiting factor and cause decrease in the rate of transpiration. As mentioned earlier, the reduction occurs because restricted absorption results in leaf water stress and closure of stomata, increasing leaf resistance. According to this view, the ability of the soil to supply water exert a controlling influence on transpiration, which may results in reduction in photosynthesis and disturbance of other physiological processes. Plant water stress caused by inadequate absorption of water leads, in function of the growth stage and the length of the period that stress occurs, to growth stagnation.

It may be that the plant is under stress for only short periods during daylight when the drying power of the atmosphere is greater than the absorption rate; the plant may recover quickly at night. If the soil is drier, its ability to supply water is reduced, and the rate of supply to the plant remains under stress for much of the night. It is now widely recognized that plant growth is directly related to the water balance in plant tissue. This balance is determined by the rate of water uptake and water loss.

Actually the soil should supply the plant at a rate that depends upon atmosphere conditions. Foregoing statement involves that water needs are high on hot, dry, windy days with bright sunshine and are low on cool, humid, calm days, with overcast skies. Thus, plant needs for water are greater in arid climates, than in humid.

It is questionable of the effect of water stress on the production is equal during the entire growing season. Numerous investigators have shown that the need for water is not uniform for each growing stage. For most crops a high positive correlation may be found between available soil water and total vegetative growth, providing that other factors (e.g. fertility, light and temperature) are not limiting. Vegetative growth and moisture supply are mostly directly related to the final yield. When the soil is at its maximum wetness immediately after planting is usually the most efficient. Much of the benefical effect of the early water supply may be attributed to the fact that it causes early germination and development. Nevertheless, excessive growth may affect the production negative. As a general rule, the flowering stage is less sensitive to water stress. Contrary during the ripening stage, water should be available at a sufficient rate to increase the production of the marketable yield as much as possible. Without excess of course, because this may cause rotting or undesirable vegetative growth.

It is practically impossible to give general guidelines what the mean moisture content should be during the succeeding growing stages which are valid for all crops at the same time. The main reason for this is the great variety between the different crops. Each crop has its own specific water needs. Therefore it is not surprising that research on this subject has been comprehensive only for the more important crops. A small review of field crops, vegetable crops, forage crops and fruit trees follows.

Wheat, cotton and peanuts, represent the three most important cereal, fiber and annual legume crops, respectively which are grown in arid and semi-arid regions.

The critical threshold of available moisture reserve below which wheat suffers from water stress has not been sharply defined, but it appears that this threshold occurs after the utilization of 2/3 to 3/4 of the available moisture. It was found that moisture stress during early stages (tillering and stem elongation) can cause damage which can be partially remedied by a subsequent, favorable moisture regime. Furthermore a drought at the soft-dough stage, after heading may cause spikelet abortion and grain shriveling, whereas a late water supply may increase kernel weight.

Experiments carried out in countries of a Mediterranean type of climate show that the first water supply by corn can be delayed until the beginning of flowering, provided that the depletion of the available water in the main root zone does not exceed 50%. Excessive vegetative growth due to excessive water supply may cause boll loss due to shedding. During the main flowering and fruiting period water should be supplied when about 65-70% of available moisture in the main root zone is extracted. This growth stage is the time of maximum water consumption. Roots have attained near maximum depth, leaf area is greatest and air temperatures are highest. The mid-season moisture regime will determine the size and weight of the bolls. Consumptive use declines in late season as a result of limited new growth, cooler weather and maturation of the crop. Moisture deficits at this time are not harmfull, and may even have a benefical effect by retarding vegetative growth without causing measurable decrease in seed or lint yields.

Various investigations have been conducted to determine the effect of soil water stress on growth and yield of peanut, when it is imposed at one of the three main stages of development recognized in the peanut plant. This stages are: from planting until the beginning of flowering, the flowering period and from beginning of pod growth until harvest. It has been found that for maximum yields, there must be an adequate supply of soil moisture during flowering and seed development. There are indications that yield may not be affected by wide variations in soil water between germination and flowering.

Vegetable crops are usually more sensitive to water shortage than other crops. This relative sensibility is probably due to two mains reasons:

— most vegetable crops are shallow-rooted, although there are several exceptions such as tomatoes , water melons, artichokes and asparagus;

— the marketable product is usually the fresh fruit or tubers, or the vegetative portions of the plant; contrary to dry-matter products (grain, fiber, sugar) the yield of these organs is more sensitive to water deficit.

Table 1 gives, for different vegetable crops, the lowest admissable value that the matric potential (joules/kg) in the root zone may obtain before water should be applied. When the matric potential does not decrease under this critical value during the different growth stages maximum benefit may be derived, if at the same time the farmer should use good seeds or plants of selected varieties, fertilize properly, control weeds and insects and employ other improved management practices.

The water use of forage corn varies according to the stage of plant development, with maximum consumption occuring between tasseling and the hard-dough stage of grain production. Then, as physiological activity declines,

TABLE 1.: *Maximum yield obtained at specific matric potentials for different crops.*

Crop	Matric potential (joules/kg)	Equivalent matric suction (centibars)	Reference
VEGETABLE CROPS			
Beans	-75 to -200	75 to 200	Vittam et al (1963)
Broccoli			
Early	-45 to -55	45 to 55	Pew (1958)
After budding	-60 to -70	60 to 70	Pew (1958)
Cabbage	-60 to -70	60 to 70	Vittam et al (1963) and Pew (1958)
Canning peas	-30 to -50	30 to 50	Taylor (1972)
Carrots	-55 to -65	55 to 65	Pew (1958)
Carrots			
During seed year at 60cm depth	-400 to -600	400 to 600	Hawthorn (1951)
Cauliflower	-60 to -70	60 to 70	Pew (1958)
Celery	-20 to -30	20 to 30	Marsh *, and Marsh (1961)
Lettuce	-40 to -60	40 to 60	Marsh *, Vissar (1959) and Pew (1958)
Onions			
Early growth	-45 to -55	45 to 55	Pew (1958)
Bulbing time	-55 to -65	55 to 65	Pew (1958)
During seed year			
at 7cm depth	-400 to -600	400 to 600	Hawthorn (1951)
at 15cm depth	-150	150	Hawthorn (1951)
Potatoes	-30 to -50	30 to 50	Taylor et al (1963)
Tomatoes	-80 to -150	80 to 150	Vittam et al (1958), and Vittam et al (1963)
FRUIT CROPS			
Avacadoes	-50	50	Richards et al (1962)
Bananas	-30 to -150	30 to 150	Schmeuli (1953)
Deciduous fruit	-50 to -80	50 to 80	Marsh *, and Vissar (1959)
Grapes			
Early season	-40 to -50	40 to 50	A.W. Marsh *
During maturity	-100	100	A.W. Marsh *
Lemons	-40	40	A.W. Marsh *
Oranges	-20 to -200	20 to 200	Stolzy et al (1963)
Strawberries	-20 to -30	20 to 30	A.W. Marsh *, and Marsh (1961)

* Unpublished research, see Taylor and Ashcroft (1972).

Note: Where two values are given, the higher value is used when evaporative demand is high and the lower value when it is low; intermediate values are used when the atmospheric desiccating rate is intermediate.

water use gradually decreases. It seems that from emergence to tasseling it is safe to extract up to 60% of the available water in the main root zone (00-90cm), and even a greater amount as the plant appreaches maturity. However, from tasseling to silking, the extraction of available water should not be allowed to exceed 40-50%.

Alfalfa will maintain itself over a wide range of moisture conditions since the foliage of the plant is the part harvested, rapid growth should be promoted. It is considered a good practice to replenish the root zone to field capacity immediately after each cutting to supply sufficient moisture for rapid growth of new stems and leaves. If the matric potential in the root zone between two cuttings approaches 400 to 800 centibars it is advisable to replenish the available water.

The water needed by fruit trees vary with climatic conditions under which they are grown, with the age and size of trees, the stage of growth (vegetative-, bloom- and fruit setting stage, developing stage of the fruit), the depth and type of the soil and rooting habits of the tree. Generally not more than 40% of the available water may be depleted during the vegetative period. Brought during the growth stage reduces sharply shoot and leaf growth. Furthermore the uptake of nutrients as nitrogen and potassium in insufficient which lead to light leaf color and thin leaf set. The plant may even never recover the uptake of nitrogen and potassium if a delay in the uptake took place during the stage of greatest vegetative growth. Moreover, drought after blossoming may induce the abcission of the weaker fruits. Re-activation of the vegetative growth through excessive water supply during this period can affect also fruit-drop. Severe fluctuations in moisture stress should be avoided during flowering to fruit setting. A high moisture level should be maintained in the soil profile during the period of early blossom and fruit set. Many authors report that a light stress before flowering induce blossoming. Fruit growth is less sensitive to poor water supply. Irrigation should be applied before wilting point is reached in the root zone. Not only fruit size will be adversely affected by deficient soil water but also the bud initiation and consequently the yield of the following year. Anyhow, attention should be paied on the cumulative response of fruit trees to moisture regime on a long-term basis, namely the effect of irrigation practices from one year to another.

The data, presented in this paper, for the different crops serve to illustrate their specific water needs in function of the growth stage. Most of the data are applicable to Mediterranean conditions. According to the experience gained in the determination of water requirements, it is clear that the actual timing of water supply to obtain maximum yield must be based on local conditions. Generally best growth can be expected when depletion of available moisture in the root zone does not exceed 30 to 70% depending from soil type. Moreover, if nutrients are not available at all times and in sufficient quantity, little will be gained be maintaining a high moisture level in the soil. The more fertile the soil, the greater the benefits to be expected from the application of water.

2. The availability of water to plants.

Two important concepts resulted from early work relating soil water to plant response. The first was the field capacity concept (Israelsen and West, 1922; Veihmeyer and Hendricson, 1927) which became identified with the upper limit of available water, assuming that the amount in excess of field capacity drains away too quickly to be of any use. This is somewhat misleading, however, because all water that is not held tigthly in the soil is available for plant growth as long as it is in contact with the roots. The second, the wilting coefficient (Briggs and Shantz, 1911), led to the permanent wilting percentage (Hendrickson and Veihmeyer, 1927) that gained wide acceptance as the lower limit of available water. It is the water content of the soil at which the leaves of plants growing in it show wilting and fail to recover when placed overnight in near saturated atmosphere. The utility of these two concepts is in determining practical irrigation schedules.

The availability of soil moisture to plants is of particular importance for agriculture, mainly in arid regions. The problem connected with soil moisture availability has been the subject of diverse opinion between opposing schools. Veihmeyer and his co-workers claimed that soil water is equally available over the available range between field capacity and permanent wilting point. On the other hand, Richards and Wadleigh (1952) made a strong case for the concept of decreasing availability as the soil water content decreased from the upper to the lower limit.

A third approach is represented by a group of irrigation scientists who suggested the existence of a ' critical soil moisture level ' (in the available soil moisture range). Below this critical level a significant decrease in yield is noted. In fact, each of these situations may prevail under certain conditions. The three approaches are presented schematically in figure 2.

Curve A represents the concept of equal soil moisture availability, in the range between field capacity (F.C.) and permanent wilting point (P.W.P.). Curve B describes the existence of a critical soil moisture within the available soil moisture range. Curve C represents the decrease in yield as soil moisture decrease gradually. Nowadays, most irrigation specialist adhere to the approach illustrated by curve B. The amount of water in a soil is in itself no effective indication of its availability; a better indication is the force with which the water is kept by the soil.

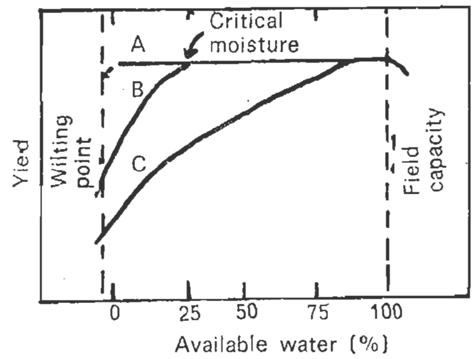

Figure 2.: The relation between available water and yield (adapted from « Water-Soil-Plant Relations » Calif. Agric. Vol. 11, 1957).

The amount of available water that the plant can extract easily (namely that amount stored in the root zone between field capacity and the critical matric potential) depends from the position of the critical moisture level and the shape of the water retention curve.

The relationship between water content and matric or soil water suction is typical for each soil, and is usually referred to as soil moisture characte-

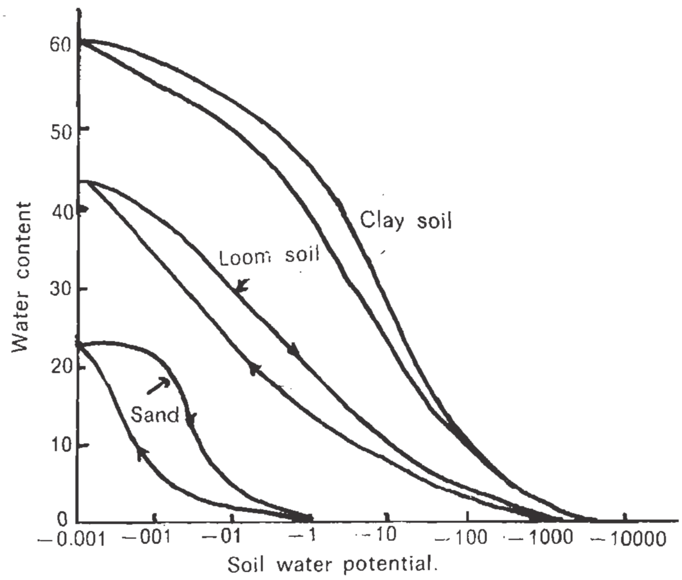

Figure 3.: Representative water-retention curves.

ristic or water retention curve. The matric forces, by which the soil can retain water against the gravitational force, can be subdivided into capillary forces, osmotic or adsportive forces and adhesion forces. The adhesion forces and the adsorptive action of adsorbed counterions in the double layer act over a short range. They cause a strong binding of a very thin film but of little importance at the higher water content which the plant can extract. The larger amounts of water in the soil are retained due to the presence of air-water interfaces similarly as exist in sponges. Surface tension acting at the air-water interface provide the mechanism of water retention. In other words, the retention of water between soil particles can be tought as a capillary phenomenon; whereas the capillary force increases with decreasing diameter of the pore. What means that water is more tightly held in narrow pores (fine textured soils) than in wide ones (coarse textured soils). Some typical soil moisture characteristics are shown in figure 3.

The sandy soil has the well-known chair like shape, which is indicating that there is a prevailing pore size as can be expected for well graded sands. It appears from the pF-curves that the sandy soil is already fairly dry at field capacity, that the quantity of available water is small and that the matric potential range (pF-range) between a wet and a fairly dry soil is relatively narrow. Therefore to use such soils for agricultural purposes a rather high groundwater-level had to be maintained or irrigation is needed. The buffer-capacity of the loamy soils is much greater, resulting in a greater availability of water. These soils are more suited for agricultural use. High yields can obtained even without or with small irrigation applications. As can be seen from these curves, high clay contents results in a much stronger binding of the soil moisture. The differential water capacity ($C_{(\theta)} = \delta\theta / \delta\psi_m$; the tangent of the water characteristic curve at any value of Ψ_m or the absolute value of the (negative) rate of change of water content with matric potential) of clay soils is in the low suction range of optimal plant growth less than of loamy soils. With other words in this range plants on clay soils must exert more power (being the rate of doing work) than on loam soils and even much more than on sandy soils) to extract equal quantities of water.

The most important applications of the moisture characteristic are briefly:

— the determination of field capacity, wilting point and available moisture storage;

— the determination of the pore size distribution to characterize soil structure. From the capillary-rise equation can be computed for each moisture tension an effective pore size. Then the pF-scale of the moisture characteristic is to transform into a pore size scale and for each pore size range can be read the volume occupied by the pores in that range.

The problems involved by using moisture characteristics is that the relation between moisture tension and moisture percentage determined on drying samples and moistening samples give not the same result. This effect, which is called hysteresis, is due to one or more of the following four factors:

— the tension at which a pore space cell is emptied is determined by the size of the greatest pore, which gives entry into the cell; the tension however at which the cell is filled completely is determined by the greatest diameter of the cell, which is greater than that of the greatest pore; therefore a lower tension is necessary in the moistening stage than in the drying one to get the same moisture percentage (reversible hysteresis);

— a variation of the soil particles in the drying sample (irriversible hysteresis);

— soils containing clay and humus are remoistening slowly and often partly after previous drying and

— enclosed air, which is dissolving gradually in the soil moisture.

By plotting both the retention curves for wetting and drying one gets the hysteresis loop which represents the extreme changes in the capillary potential for any given soil water content. The external branches of the hysteresis loop represents the envelope loops for all the intermediate curves, called ' scanning curves '.

Soil water hysteresis imposes severe difficulties in analyzing flow systems in the soil as well as the soil water energy status in the soil under field conditions. It is common to work either on the drying or the wetting branch

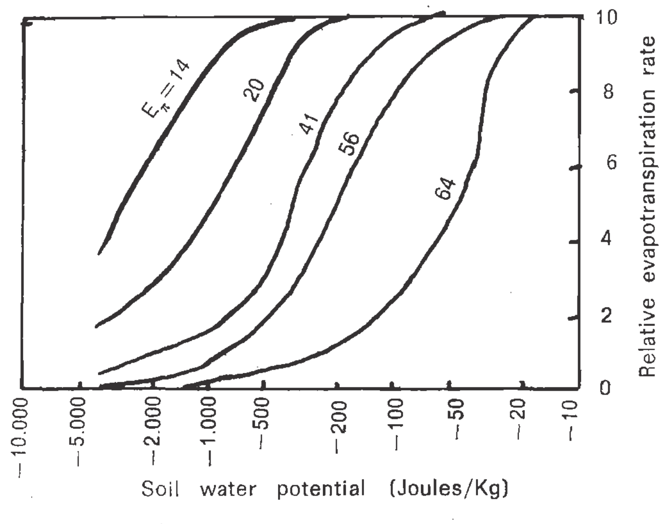

Figure 4.: The relative evapotranspiration rate as function of soil water potential for different potential evapotranspiration conditions. The E_{TP}-values for the curves are expressed in millimeters of water evapotranspired in a 24-hour period.

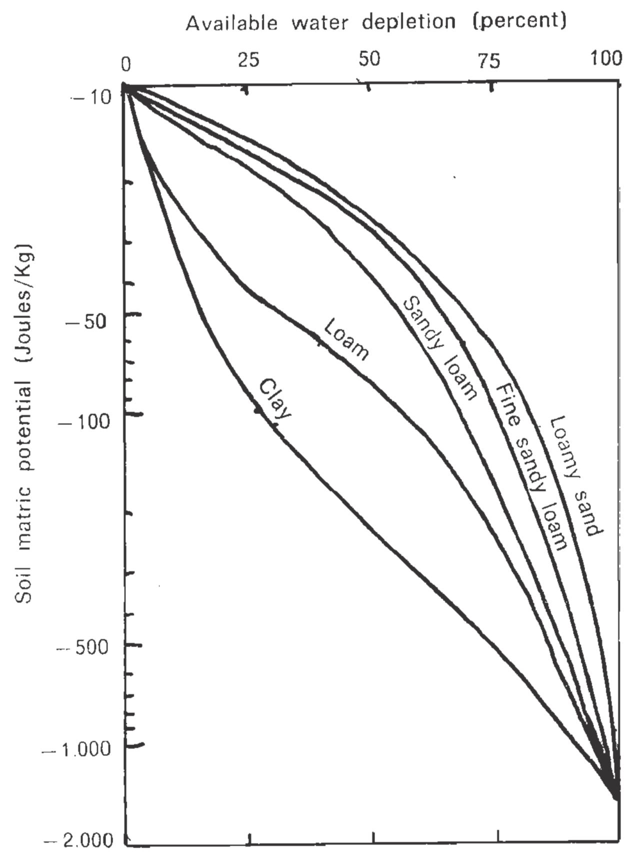

Figure 5.: **The relation between available water depletion percent and soil matric potential (adapted from Richards and Marsh, 1961).**

of the soil-water potential functions. There are some recent experimental indications showing that, due to fluctuations in the soil, water potential, caused by water transfer the hysteresis loops tends to shrink or degenerate into a ' relaxed valued ' curve.

As explained in foregoing section, the actuel water loss by the canopy (transpiration rate) decreases when the water potential in the soil diminish. If we consider as DENMEAD and SHAW (1962) the transpiration rate at field capacity as the potential evapotranspiration rate, it is possible to plot the relative evapotranspiration rate E_T/E_{TP} , for various atmospheric and soil water conditions. Figure 4, after DENMEAD and SHAW (1962), shows the relative evapotranspiration rate as a function of soil water potential for different desiccating atmospheric conditions.

These curves demonstrate clear that from a critical value of the matric potential further depletion causes a drastic reduction in evapotranspiration, consequently also in production. They are very practical guides in the deduction of the available water that may be used before irrigation water should be applied. In order to that, the water retention curves are represented in a different form. The available water depletion expressed as a percentage of the total available water storage (between F.C. and P.W.P.) is plotted against the matric potential. Figure 5 presents water retention curves for severals solis plotted in terms of percent available water removed.

It can be seen that the soil moisture potential at 50% available water depletion in a clay soil is considerably lower than in loam and sandy loam soils.

Suppose that, to obtain maximum yield, the relative evapotranspiration rate during the growing season may not be lower than 0.75, we can deduce from figure 4 in function of the potential evapotranspiration rate the matric potential to which the plant may deplete the soil. Entering this value in the ordinate of figure 5 gives immediately the percentage depletion of available water. The use of soil water suction or potential overcomes much of the difficulty in applying results from one area to another. Furthermore plant response is better correlated with water potential than with soil water content. The measurement of soil water potential or suction provides an universal tool for sheduling irrigation.

As we discussed it here, the impression may arise that the ability of a soil to supply water depens only from its potential. The decrease in water supply is primarily a result of increased resistance in the path of water flow rather than a result of reduction in the water potential. As the soil water content and matric potential decrease, the hydraulic conductivity decreases very rapidly, as shown in fig. 6, so that when $\psi_m = -1500$ joules/kg, K is only about 10^{-6} of the value at saturation. According to PHILIP (1957) the rapid decrease in conductivity occurs because the larger pores are emptied first, greatly decreasing the cross section available for liquid flow. When the continuity of the films is broken, liquid flow no longer occurs.

The rate at which the soil supply water depends on the interaction between hydraulic conductivity K and the gradient of the soil water or matric potential. If the soil in the root zone is uniform moist, K will be large and fairly constant with distance so that water will be supplied from a large soil volume, ensuring a constant transport to the root system. Rapid absorption, due to an increased water loss by the leaves, causes a steep decrease in water content close to the roots. The faster decrease of K compared with the increase of the potential gradient cutts off the supply, leaving untapped water in the soil not far distant from the root.

The lower limit to available water ' will soon be reached in the sur-

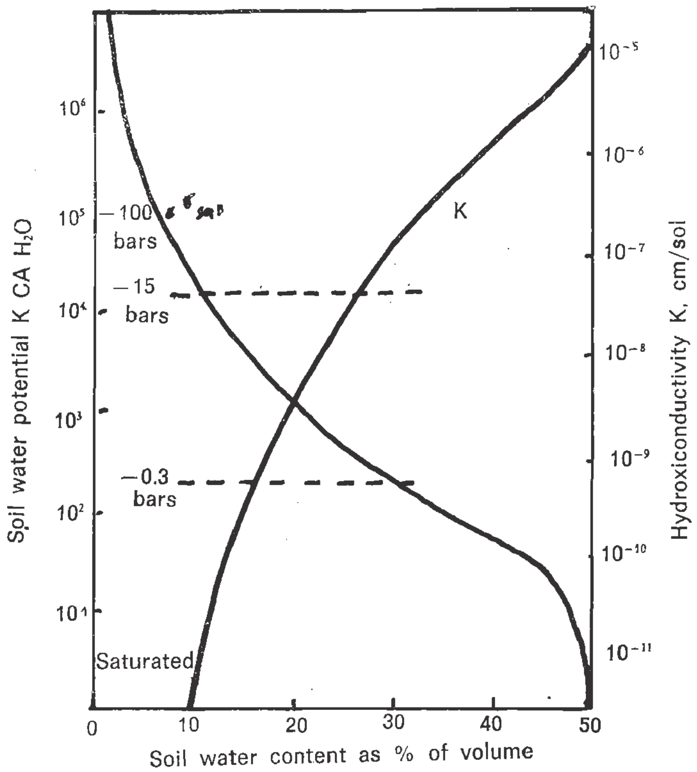

Figure 6.: Decrease in hydraulic conductivity K and matric potential Ψ_m with decrease in soil water content.

roundings of the root. Just behind this barrier to water transport a wetting front exist, a region of very steep moisture gradients. There will be a more continuous flow of water in the soil to the plant system if the root environment is kept nearly uniform moist without markedly moisture changes in distance. It means between field capacity and the critical matric potential, corresponding the crop grown.

Finally we can say that the availability of soil water depends chiefly on the hydraulic conductivity of the soil and on its potential, both of which are closely related to the water content of the soil. Beside, many other factors are involved in the total amount of water that moves to the roots that it has been practical impossible to describe this phenomenon mathematically. The root systems of actively growing plants continually occupy new regions of soil, and there is more rapid absorption in the more permeable regions of roots, hence the soil masse occupied by a root system often consists of several zones of different water contents. As a result most of the water absorption by a given root system may occur from a small fraction of the soil mass occupied by it. This situation is accentuated if irrigation (typical for trickle systems) recharges only a part of the root zone. Under such conditions the roots are in contact with a wide range of water potentials. It is most probably that the water potential measured at the base of the stem, gives a reasonable estimate of the water potential in the root tissue at the wettest area of the root zone where most of the water is being absorbed (Slatyer, 1960).

The problem of soil water flow in the root zone is complicated; simple models have been derived, which appear to give good agreement with experimental data. The simulation of the soil moisture extraction of an irrigated crop by a point source is even more complex and yet unsolved. Up to now, there have been several trials to approach mathematically the moisture movement within the soil profile from a trickle source. Before dealing with this complex analyses it is better to discuss briefly the physical laws governing the flow of water in the unsaturated soil.

3. The physical laws governing the flow of water in unsaturated soil.

The fundamental law describing the movement of groundwater through a soil was given by Darcy in 1856. Darcy observed that the amount of water flowing through a sand sample in unit time (in other words the rate of flow or the discharge) was proportional to the difference Δh between the fluid heads at the inlet and outlet faces of the sample and inversely proportional to the length of the sand sample (the flow path). This proportionality can be expressed mathematically as follows:

$$Q = K \frac{\Delta h}{L} A \qquad (1)$$

where: Q = the rate of flow through the sample (L^3t^{-1})
Δh = the head loss (L)
L = the length of the sample (L)
A = the cross-sectional area of the tube (L^2)
K = a proportionality constant, depending of the nature of the sand and the fluid (Lt^{-1}).

The quantity Q/A represents the discharge or flow rate per unit of cross-sectional area and is called apparent velocity, sometimes also called effective flow velocity, specific discharge or flux. It is denoted by the symbol v. Hence

$$v = Q/A \qquad (2)$$

The term $\Delta h/L$ represents the loss of head per unit length of flow path and is called gradient or hydraulic head. Denoting this hydraulic gradient by i and substituting it into equation (2) yields what is known as Darcy's law of lineair resistance:

$$v = -Ki \tag{3}$$

Darcy's law states that the apparent velocity is directly proportional to the derivative of the hydraulic head in the direction of flow. The negative sign indicates that the flow is in the direction of decreasing head. The dimension of v is (Lt^{-1}), while i is dimensionless. Hence the dimension of K is that of velocity (Lt^{-1}). The proportionality constant K is known as the coefficient of permeability or, preferably, hydraulic conductivity. This coefficient may be considered as a combination of two factors:

— the intrinsic permeability k, which is a function of the solidphase geometry, and

— the fluid factor ($\rho\ g/\eta$) which is a function of the dynamic viscosity.

The above combination is represented by the relation:

$$K = k\left(\rho \frac{g}{\eta}\right) \tag{4}$$

where: ρ = the fluid density (ML^{-3})
η = the dynamic viscosity ($ML^{-1}t^{-1}$).

It should be noted that the flow velocity in the individual pores of the soil greatly exceeds the apparent velocity, which, in fact, is a hypothetical velocity that the water would have if flowing through the given flow column quite unobstructed by solid particles. The actual velocity of the water particles, v_a, follows from:

$$v_a = Q/nA = v/n \tag{5}$$

where: n = the porosity of the soil (dimensionless).

Since n is always smaller than 1, it can readily be seen that the actual velocity of the water is always greater than the water flux.

Darcy's law is valid for laminar flow, and since groundwater generally is moving at a low speed, laminair flow conditions prevail; consequently, Darcy's equation may be applied without any appreciable error.

Equation (3) applies also to an unsaturated soil where K is dependent on the soil water content (θ) or matric potential (Ψ_m). The conductivity stated as a function of potential is a highly hysteretic relation as compared to K-θ

which is practically unique. Equation (3) for an unsaturated soil will then become introducing the concept of potentials:

$$v = -K(\theta)\frac{d\psi_t}{ds} \tag{6}$$

where: ψ_t = the total potential of water, defined as the work required to transfer a unit quantity of water from a standard reference, where the potential is taken zero, to the situation where the potential has the defined value.

The total potential of soil water in an equilibrium system is the sum of all potentials acting on water. It can be represented as:

$$\psi_t = \psi_z + \psi_p + \psi_m + \psi_g + \psi_o \tag{7}$$

in which ψ_z is gravitational potential, ψ_p is the submergence potential, ψ_m is the matric potential, ψ_g is thepneumatic potential and ψ_o is the solute potential. The pneumatic potential takes into account the effect of an external gas pressure, different from atmospheric pressure. The fundamental units of the potential of soil water depend on what unit quantity is taken in the potential's definition. If the potential is expressed per unit mass ψ_t has the units of erg. g^{-1}; expressed per unit volume ψ_t has the same nature and units as pressure, namely dyne. cm^{-2}; expressed per unit weight ψ_t has the dimension of length (cm). It is more convenient to define the potential on unit weight basis, rather than on unit mass or volume basis.

Is we consider only the liquid phase soil-water movement the solute or osmotic potential may be excluded, because due to diffusion equality of solute concentration and therefore of osmotic potential will tend to be adchieved. In the presence of differential permeable membranes, such as plant roots and the air-water interface osmotic potential effects the liquid and/or vapour movement. In practice it is only differences in potential that are significant. The pressure in the soil air adjacent to the soil water is in most cases equal or close to the atmospheric pressure. Therefore there may be some doubts as to the usefulness of introducing the concept of pneumatic potential in field studies.Ignoring both the osmotic and pneumatic potential the total potential is defined as:

$$\psi_t = \psi_z + \psi_p + \psi_m \tag{8}$$

Submergence and matric potential are mutually exclusive possibilities, this follows immediately from their character. The submergence potential is the potential that results from liquid pressure in saturated soil, while the matric potential expresses the pressure in the soil water above the water table in the unsaturated zone. If either of these potentials is non-zero, the other must be zero.

Correspondingly equation (8) should be adapted to the situation. For saturated conditions, the total potential is given by:

$$\psi_t = \psi_z + \psi_p = \psi_h \tag{9a}$$

and for unsaturated conditions:

$$\psi_t = \psi_z + \psi_m = \psi_h \tag{9b}$$

where ψ_h is called the hydraulic potential.
Substituting eqn. (9b) in eqn. (6) gives Darcy's law for an unsaturated soil:

$$v = -K(\theta)\frac{d\psi_h}{ds} \tag{10}$$

The Darcy equation, irrespective of its form, i.e. eqn. (1), (3) or (10), assumes a steady flow rate under steady total head and moisture content, meaning a steady state. Yet under most conditions the total potential (and so the water content) changes with time and position. For the solution of the Darcy equation under such conditions two additional equations incorporating the same variables are required. The first additional equation is obtained from the principle is conservation of matter. This equation, called the equation of continuity reads for a three dimensional flow:

$$\frac{\delta\theta}{\delta t} = -\frac{\delta v_x}{\delta x} - \frac{\delta v_y}{\delta y} - \frac{\delta v_z}{\delta z} \tag{11}$$

where:

$\frac{\delta\theta}{\delta t}$ = the net change in moisture content (cm^3) through the unit volume of soil (cm^3) per unit of time (sec);

v_x, v_y and v_z = the flux of water respectively in the x-, y-, and z- direction or the velocity components in the rectangular coordinate system ($cm \cdot sec^{-1}$).

According to Darcy we may assume that:

$$v_x = -\ K(\theta)_x \frac{\delta\psi_h}{\delta_x} \tag{12a}$$

$$v_y = -\ K(\theta)_y \frac{\delta\psi_h}{\delta_y} \tag{12b}$$

$$v_z = -\ K(\theta)_z \frac{\delta\psi_h}{\delta_z} \tag{12c}$$

From eqns. (11) and (12a, b, c):

$$\frac{\delta\theta}{\delta t} = \frac{\delta}{\delta x}\left[K(\theta)_x \frac{\delta\psi_h}{\delta x}\right] + \frac{\delta}{\delta y}\left[K(\theta)_y \frac{\delta\psi_h}{\delta y}\right] + \frac{\delta}{\delta z}\left[K(\theta)_z \frac{\delta\psi_h}{\delta z}\right] \tag{13}$$

If we regard the flow in a saturated soil, and we assume that the fluid is incompressible, than is $\theta = \theta$ saturated $=$ a constant or $\dfrac{\delta\theta}{\delta t} = 0$.

Taking $K(\theta)_x$, $K(\theta)_y$ and $K(\theta)_z$ as constant (which is true in a saturated homogeneous anisotropic medium) we may write eqn. (13) as:

$$K(\theta)_x \frac{\delta^2\psi_h}{\delta x^2} + K(\theta)_y \frac{\delta^2\psi_h}{\delta y^2} + K(\theta)_z \frac{\delta^2\psi_h}{\delta z^2} = 0 \tag{14}$$

Is the soil isotropic, we then obtain:

$$\frac{\delta^2\psi_h}{\delta x^2} + \frac{\delta\psi_h^2}{\delta y^2} + \frac{\delta\psi_h^2}{\delta z^2} = 0 \tag{15}$$

Which is the Laplace equation for three-dimensional saturated flow. For two-dimensional flow it reduces to:

$$\frac{\delta^2\psi_h}{\delta x^2} + \frac{\delta^2\psi_h}{\delta y^2} = 0 \tag{16}$$

Laplace's equation is also written as:

$$\nabla^2 h = 0 \tag{17}$$

where the symbol ∇, called ' del ', is used to denote the differential operator

$$\frac{\delta}{\delta x} + \frac{\delta}{\delta y} + \frac{\delta}{\delta z}$$

and ∇^2, called ' del squared ' is used for

$$\frac{\delta^2}{\delta x^2} + \frac{\delta^2}{\delta y^2} + \frac{\delta^2}{\delta z^2}$$

which is called the Laplacean operator.

For an unsaturated soil two flow situations may prevail. First the steady state, which occurs while water is flowing through the soil and no change in the soil water content and the potential distribution occur with time. For example, the steady state evaporation from a soil column with groundwater table. Since under these circumstances $\frac{\delta\theta}{\delta t} = 0$ and the components of the potential gradient in each point do not change in time eqn. (13) may be written as:

$$\frac{\delta}{\delta x}\left[K(\theta)_x \frac{\delta\psi_h}{\delta x}\right] + \frac{\delta}{\delta y}\left[K(\theta)_y \frac{\delta\psi_h}{\delta y}\right] + \frac{\delta}{\delta z}\left[K(\theta)_z \frac{\delta\psi_h}{\delta z}\right] = 0 \qquad (18)$$

and the flux must be a constant. However, the components of the conductivity can not be assumed as constants. They are a function of the moisture content or matric potential, which changes from location. For one-dimensional vertical flow eqn. (18) reduces to:

$$v = -K(\theta)_z \frac{d\psi_h}{dz} \qquad (19)$$

This equation is valid for each point of the flow path. Note that the value of the hydraulic conductivity and the potential gradient are only independent from time. According to eqn. (19) the product of both must be constant throughout the profile. Upward flow in a soil column with a water table, drying under the influence of evaporation is possible due to a gradient in the hydraulic potential built up between the water table and the evaporating surface. Hence the matric suction increases (i.e. potential decreases) from the water table to the surface. Since the conductivity decreases more rapidly than the matric suction increases, the gradient of the hydraulic potential should become very great close to the surface to maintain a constant flux.

The graphical representation of the soil moisture potential with depth will have an exponential form (being normal to the water table and approaching the surface assymptotic), because most empirical expressions derived for the $(K - \psi_h)$ or $(K - \theta)$ are exponential.

The second flow situation that normally occurs in the unsaturated zone si the unsteady state. Under this circumstances eqn. (13) has to be applied. This equation includes three interrelated parameters (i.e. θ, ψ_h and $K(\theta)$) and therefore it is difficult to solve. A second additional relationship is required for the solution of eqn. (13), which is found in the water retention curve and the conductivity.

Before introducing the concept of the diffusion coefficient of water in soil the hydraulic potential in eqn. (13) may be written in terms of the matric and gravity potential. Since the gravitational force only acts in the vertical direction $\left(\frac{\delta\psi_z}{\delta z} = 0, \frac{\delta\psi_z}{\delta y} = 0\right)$ eqn. (13) changes into:

$$\frac{\delta\theta}{\delta t} = \frac{\delta}{\delta x}\left[K(\theta)_x \frac{\delta\psi_m}{\delta x}\right] + \frac{\delta}{\delta y}\left[K(\theta)_y \frac{\delta\psi_m}{\delta y}\right] + \frac{\delta}{\delta z}\left[K(\theta)_z \frac{\delta\psi_m}{\delta z} + K(\theta)_z \frac{\delta\psi_z}{\delta z}\right] \tag{20}$$

Expressed on a weight basis, the gravitational potential is equal to the height z above the reference level. Accordingly, the gradient will be the height difference between every two locations under consideration.

$$\frac{\delta\theta}{\delta t} = \frac{\delta}{\delta x}\left[K(\theta)_x \frac{\delta\psi_m}{\delta x}\right] + \frac{\delta}{\delta y}\left[K(\theta)_y \frac{\delta\psi_m}{\delta y}\right] + \frac{\delta}{\delta z}\left[K(\theta)_z \left(\frac{\delta\psi_m}{\delta z} + 1\right)\right] \tag{21}$$

Splitting the gradient of the matric potential into two parts

$$\left(\frac{\delta\psi_m}{\delta z} = \frac{\delta\psi_m}{\delta\theta} \cdot \frac{\delta\theta}{\delta z}\right)$$

using the chain rule of differentiation one obtains:

$$\frac{\delta\theta}{\delta t} = \frac{\delta}{\delta x}\left[K(\theta)_x \frac{\delta\psi_m}{\delta\theta} \cdot \frac{\delta\theta}{\delta x}\right] + \frac{\delta}{\delta y}\left[K(\theta)_y \frac{\delta\psi_m}{\delta\theta} \cdot \frac{\delta\theta}{\delta y}\right] +$$
$$\frac{\delta}{\delta z}\left[K(\theta)_z \frac{\delta\psi_m}{\delta\theta} \cdot \frac{\delta\theta}{\delta z}\right] + \frac{\delta K(\theta)_z}{\delta z} \tag{22}$$

Doing this we assume implicitly that the matric potential is an unique function of the moisture content. Both $K(\theta)$ and $\frac{\delta\theta}{\delta\psi_m}$ (the tangent of the moisture characteristic of the soil at any particular value of wetness θ), also defined as the differential water capacity $C(\theta)$) are dependent on the moisture content and therefore the ratio must also be. Indicating the ratio by D (called the soil-water diffusivity) eqn. (22) may be written:

$$\frac{\delta\theta}{\delta t} = \frac{\delta}{\delta x}\left[\frac{K(\theta)_x}{C(\theta)} \cdot \frac{\delta\theta}{\delta x}\right] + \frac{\delta}{\delta y}\left[\frac{K(\theta)_y}{C(\theta)} \cdot \frac{\delta\theta}{\delta y}\right] + \frac{\delta}{\delta z}\left[\frac{K(\theta)_z}{C(\theta)} \cdot \frac{\delta\theta}{\delta z}\right] + \frac{\delta K(\theta)_z}{\delta z} \tag{23}$$

or:

$$\frac{\delta\theta}{\delta t} = \frac{\delta}{\delta x}\left[D(\theta)_x \frac{\delta\theta}{\delta x}\right] + \frac{\delta}{\delta y}\left[D(\theta)_y \frac{\delta\theta}{\delta y}\right] + \frac{\delta}{\delta z}\left[D(\theta)_z \frac{\delta\theta}{\delta z}\right] + \frac{\delta K(\theta)_z}{\delta z} \tag{24}$$

For water movement in a vertical column eqn. (24) reduces to:

$$\frac{\delta\theta}{\delta t} = \frac{\delta}{\delta z}\left[D(\theta)_z \frac{\delta\theta}{\delta z}\right] + \frac{\delta K(\theta)_z}{\delta z} \tag{25}$$

for z positive upward, and to:

$$\frac{\delta\theta}{\delta t} = \frac{\delta}{\delta z}\left[D(\theta)_z \frac{\delta\theta}{\delta z}\right] - \frac{\delta K(\theta)_z}{\delta z} \tag{26}$$

for z positive downward.
For horizontal flow, in the absence of gravity, eqn. (24) reads obviously as:

$$\frac{\delta\theta}{\delta t} = \frac{\delta}{\delta x}\left[D(\theta)_x \frac{\delta\theta}{\delta x}\right] \tag{27}$$

In the special case that the diffusivity remains constant (though it is not safe to assume this except for a very small range of wetness) eqn. (27) can be written in the form of Fick's second diffusion equation:

$$\frac{\delta\theta}{\delta t} = D \frac{\delta^2\theta}{\delta x^2} \tag{28}$$

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

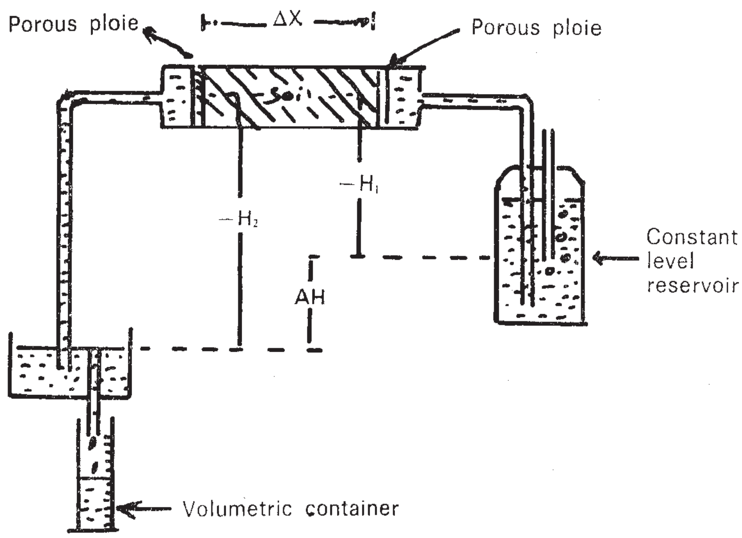

Figure 7.: A model illustrating unsaturated flow under a suction gradient in a horizontal column.

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.

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

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:

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

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-

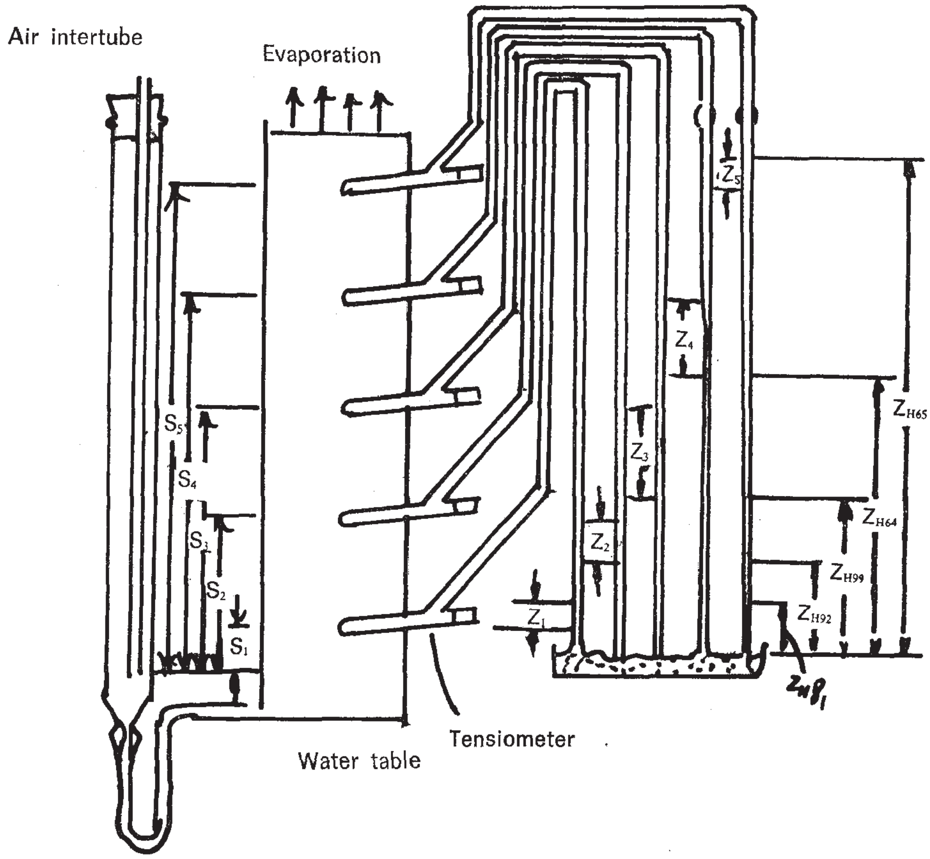

Figure 8.: Apparatus for measuring unsaturated hydraulic conductivity based upon steady state evaporation from a soil column.

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

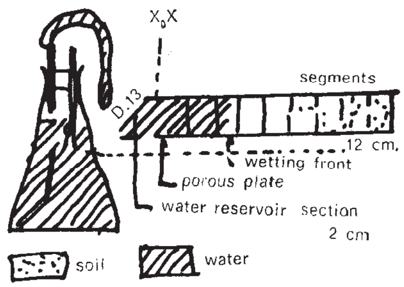

Figure 9.: Experimental set-up for horizontal infiltration. The method is based on the introduction of the Boltzmann transform ($\Lambda = x/t^{\frac{1}{2}}$) to transform the soil-water diffusivity equation to an ordinary differential equation. Under the conditions of
$t = o,\ x > o,\ \theta = \theta_i$
$t > o,\ x = o,\ \theta = \theta.$

$> \theta_i$

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:

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

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).

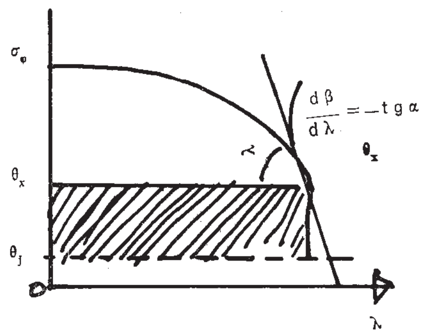

Figure 10.: Water content distribution at end of infiltration test with set up of figure 9.

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

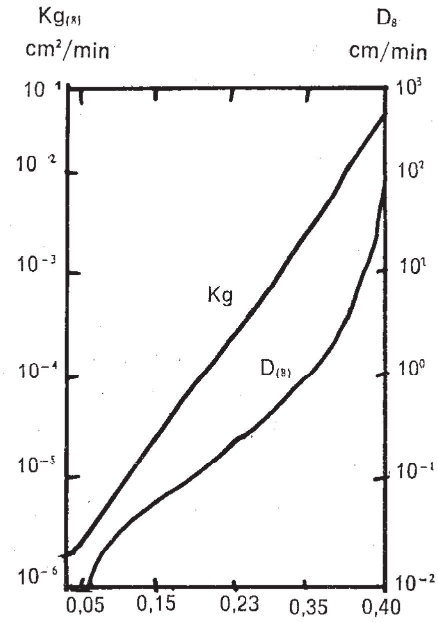

Figure 11.: Experimental values of K and D as functions of obtained from the θ (Λ) function of figure 10. (Columbia silt loam, c.f. Davidson et al., 1963).

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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