UNIT 4 SOIL WATER MOVEMENT

Structure

4.1 INTRODUCTION

Water movement in soil is a complex phenomenon. This movement plays a significant role in the availability of water to the plants, base flows and water table build up, etc.; and above all it is ultimately lost tothe atmosphere. Physico-chemical and hydraulic properties of soil significantly affect the movement of water through soils.

Objectives

After studying this unit, you should be able to :

- state the laws governing soil water movement,
- describe basic concepts of water movement,
- explain unsaturated flow of water,
- describe process of loss of water from a bare surface,
- describe water uptake by plants, and
- elaborate on drainage, infiltration and conductivity of water.

4.2 SOIL WATER MOVEMENT LAWS

Flow of water takes place due to a gradient in soil water potential from one soil zone to another. The direction of flow is from a zone of higher to one of lower moisture potential. Saturated flow takes place when the soil pores are completely filled (saturated) with water. Unsaturated flow takes place when the pores in even the wettest soil zones are only partially filled with water. In each case energy soil moisture relations are dominant.

4.2.1 Poiseuille's Law

Early theories of fluid dynamics were based on the hypothetical concepts of a **perfect** (i.e., ideal) fluid which is one that is frictionless and incompressible. In such case the contacting layers can exhibit no tangential force (shearing stresses) but only normal forces (pressures). Such fluids do not in fact exist in nature.

In the flow of **real** fluids adjacent layers do transmit tangential stresses (drag), and the existence of intermolecular attractions causes the fluid molecules in contact with a solid wall to adhere to it, rather than slip over it. The flow of real fluid is associated with the property of viscosity which characterises the fluids resistance to flow.

To maintain the relative motion at a constant velocity it is necessary to apply a tangential force, that force having to overcome the frictional resistance in the fluid. This resistance per unit area,τ, of the plate is proportional to the velocity of the upper plate u, and inversely proportional to the distance y. Thus shearing stress, τ, at any point is proportional to the velocity gradient du/dy. The viscosity η is the proportionality factor between τ, and du/dy :

$$\tau = \eta \frac{du}{dy}$$

This relationship can be used to describe the flow through a straight cylindrical tube with a constant diameter D = 2 R. The velocity is zero at the wall because of adhesion, maximum at its axis and constant on cylindrical surfaces which are concentric about the axis. The adjacent cylinderical layers, moving at different velocities, slide over each other. A parallel motion of this kind is called **laminar** flow.

In the poiseuille flow, the mean velocity is propotional to the pressure drop per unit distance.

Laminar flow prevails only at relatively low flow velocities and in narrow tubes. As the radius of tubes and flow velocities are increased, the point is reached at which the mean flow velocity is no longer proportional to the pressure drop and the parallel laminar flow change into a turbulent flow with fluctuating eddits. Consequently, the laminar flow is a rule rather than the exception in most water flow processes taking place in soils because of the narrowness of the soil pores.

4.2.2 Darcy's Law

Soil pores do not resemble uniform, smooth tubes but are highly irregular, tortuous and intricate. Flow through soil pores is limited by numerous constrictions or 'necks' and occasional 'dead ends' spaces. Hence, the actual geometry and flow pattern of a typical soil specimen is too complicated, as the fluid velocity varies drastically from point to point even along the same passage. For this reason flow through complex porous media is generally described in terms of a 'macroscopic flow velocity vector' which is the average over the total volume of the soil.

Let us assume a horizontal column of soil, through which a steady flow of water is occurring from an upper reservoir to a lower one in each of which the water level is maintained constant. Thus, the discharge rate Q, being the volume V flow through the soil column per unit time, is directly proportional to the cross-sectional area and to the hydraulic head drop ΔH and inversely proportional to the length of the soil column L :

$$Q = \frac{V}{t} \propto \frac{A \, \Delta H}{L}$$

Hydraulic head drop across the system is measured, determining the head of inflow boundary H_i and at the outflow boundary H_o relative to some reference level. H is the difference between these two heads

$$\Delta H = H_i - H_o$$

Obviously no flow, occurs in the absence of a hydraulic head difference i.e. when $\Delta H = 0$.

The head drop per unit distance in the direction of flow ($\Delta H / L$) is the 'hydraulic gradient'. The specific discharge rate Q/A (i.e. volume of water flowing through a unit cross-sectional area per unit time) is called the **flux density** or **flux**, and is indicated by q. Thus the flux is proportional to the hydraulic gradient :

$$q = \frac{Q}{A} = \frac{V}{A\,t} \propto \frac{\Delta H}{L}$$

The proportionality factor K is generally designated as the "hydraulic conductivity":

$$q = K \frac{\Delta H}{L}$$

This is known as **Darcy's Law**, after Henery Darcy, the French Engineer about 150 years ago who obtained this relationship.

SAQ 1

State the laws governing fluid flow in soil.

4.3 WATER MOVEMENT IN SOIL

Water movement in soil is a physical phenomenon. It is possible through the head gradient and conductance of the soil. In the profile soil water moves horizontally or vertically. These processes are discussed here.

4.3.1 Total Hydraulic Head

Flow in the horizontal column occurs in response to a pressure head gradient (i.e., energy gradient). Flow in a vertical column can be caused by gravitation as well as pressure. The gravitational head (H_g) at any point is determined by the height of the point relative to some reference plane, while the pressure head is determined by the height of the water column resting on that point. The total hydraulic head H is the sum of these two heads :

$$H = H_g + H_p$$

To apply Darcy's law to vertical flow we must consider the total hydraulic head at the inflow and the outflow boundaries :

$$H_i = H_{pi} + H_{gi}$$

$$H_o = H_{po} + H_{go}$$

Darcy's Law thus becomes :

$$q = K\,[\,(H_{pi} + H_{gi}) - (H_{po} + H_{go}\,)\,]\,/L$$

The reference plane (level) is generally set at the bottom of the colomn so that gravitational potential can always be positive. On the other hand the pressure head of water is positive under a free water surface (i.e. a water table) and negative above it. A negative hydraulic head signifies a pressure smaller than atmospheric and it occurs whenever the soil becomes unsaturated.

4.3.2 Flow in a Vertical Column

In order to calculate the flux according to Darcy's law we must know the hydraulic head gradient, which is the ratio of the hydraulic head drop (between outflow and inflow boundaries) to the column length. The Darcy's equation for this case (Figure 4.1) is

$$q = K\frac{\Delta H}{L} = K\frac{(H_1 + L)}{L}$$

$$q = K\,\frac{H_1}{L} + K$$

Comparison of this case with the horizontal one shows that the rate of downward flow of water in a vertical column is greater than in horizontal column by the magnitude of hydraulic conductivity. It is also apparent that if the ponding depth H_1 is negligible, the flux is equal to the hydraulic conductivity. This is due to the fact that in the absence of a pressure gradient the only driving force is the gravitational head gradient which in a vertical column has the value of unity.

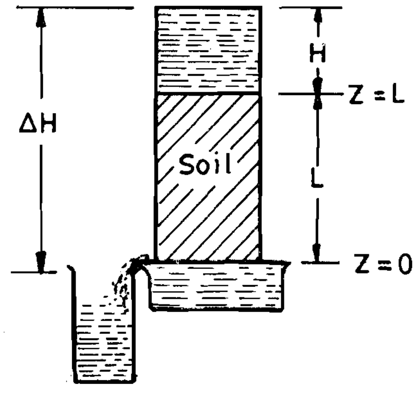

Figure 4.1 : Flow of Water in a Saturated Vertical Column

We shall now examine the case of upward flow in a vertical column as shown in Figure 4.2

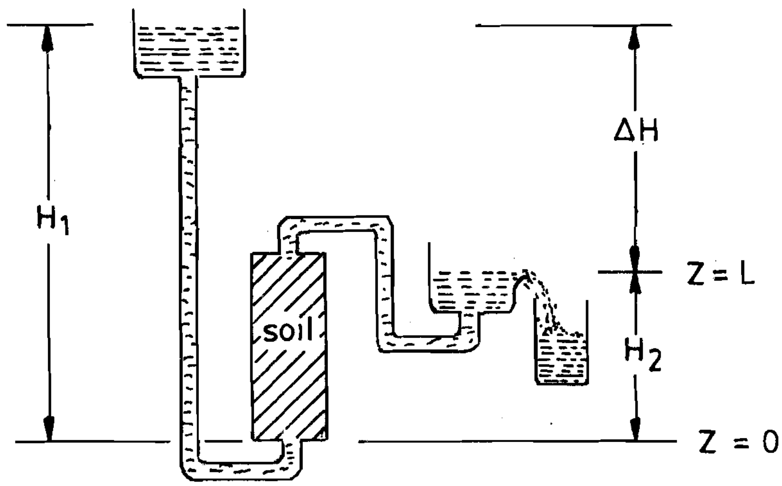

Figure 4.2 : Upward Flow in a Saturated Vertical Column

The Darcy's equation is thus :

$$q = K\frac{(H_1 - L)}{L} = K\frac{H_1}{L} - K$$

$$q = K\frac{\Delta H}{L}$$

4.3.3 Flux, Flow Velocity and Tortuosity

The **flux** is the volume of water passing through a unit cross-sectional area (perpendicular to the flow direction) per unit time. The dimensions of flux are :

$$q = \frac{V}{A\,t} = \frac{L^3}{L^2\,T} = [\,L\,T^{-1}\,]$$

These are the dimensions of velocity, yet we prefer the term flux to flow velocity, the later being an ambiguous term. Since soil pores vary in shape, width and direction the actual flow velocity in the soil is highly variable (e.g. wider pores conduct water rapidly and the liquid in the centre of each pore moves faster than the liquid in proximity to the particles).

Tortuosity can be defined as the average ratio of the actual roundabout path to the apparent or straight flow path i.e. it is the ratio of the average length of the pore passages to the length of the soil specimen. Tortuosity is thus the dimensionless geometric parameter of porous medium, which though difficult to measure precisely is always greater than one and may exceed two. The tortuosity factor is sometime defined as the inverse of what we have defined above.

4.3.4 Hydraulic Conductivity, Permeability and Fluidity

The hydraulic conductivity is the ratio of the flux to the hydraulic gradient. The dimension of flux is [LT^{-1}] where as the hydraulic conductivity depends upon the dimension assigned to the driving force. In a saturated soil the hydraulic conductivity is characteristically constant e.g. sandy soil 100 – 1000 cm/sec and clayey soil 10^{-4}— 10^{-7} cm/sec. In many soils, the hydraulic conductivity does not in fact remain constant for chemical, physical and biological change as water permeats and flows in the soil.

The hydraulic conductivity (K) is not an exclusive property of the soil alone, since it depends upon the attribute of the soil (total porosity, distribution of the pore size, tortuosity and pore geometry) and fluid (fluid density and viscosity) together. Theoretically, K could be separated into two factors :

intrinsic permeability of the soil k and **fluidity** of the liquid or gas f.

$$K = kf$$

where, K is expressed as cm/sec (LT^{-1}), k is expressed in cm^2 [L^2] and f is in 1/cm sec) [$L^{-1}T^{-1}$].

Fluidity is inversely proportional to viscosity :

$$f = \rho g / \eta$$

hence,

$$k = K\eta / \rho g$$

where, η is the viscosity (dyne sec/cm^2), ρ is the fluid density (gm/cm^3) and g is the gravitational acceleration (cm/sec^2).

In an ordinary liquid, the density is nearly constant (though it varies with temperature and solute concentration) and changes in fluidity are likely to result primarily from changes in viscosity. In the compressible fluids such as gases, on the other hand, changes in density due to pressure and temperature variations can also be considerable.

In the past the term permeability has often been applied synonymously with hydraulic conductivity. Permeability has been qualitatively defined as the readyness with which a porous medium transmits water or various other fluids.

While fluidity varies with composition of the fluid and with temperature, the permeability is ideally an exclusive property of the porous medium and its pore geometry alone provided the fluid and the solid matrix do not interact in such a way as to change the properties of either. In a complete stable porous body, the same permeability will be obtained with different fluids e.g., with water, air or oil.

SAQ 2

Differentiate among gravitational, pressure and total hydraulic heads

4.4 FLOW OF WATER IN UNSATURATED SOIL

Most of the processes involving soil water interactions in the field, and particularly the flow of water in the rooting zone of most crop plants, occur while the soil is in an unsaturated condition. Unsaturated flow processes are in general complicated and difficult to describe quantitatively, since they often entail changes in the state and content of soil water during flow. Such changes involve complex relations among the variable soil wetness, suction, and conductivity, whose interrelations may be further complicated by hysteresis. The formulation and solution of unsaturated flow problems very often require the use of indirect methods of analysis, based on approximations or numerical techniques.

4.4.1 Comparison of Flow in Saturated and Unsaturated Soils

The soil water flow is caused by a driving force resulting from a potential gradient, and the flow takes place in the direction of decreasing potential, and the rate of flow (flux) is proportional to the potential gradient and is affected by the geometric properties of the pore channels through which flow takes place. These principles apply to unsaturated, as well as saturated, soils.

The driving force in a saturated soil is the gradient of a positive pressure potential. On the other hand, water in an unsaturated soil is subject to a subatmospheric pressure, or suction, which is equivalent to a negative pressure potential. The gradient of this potential likewise constitutes a driving force. Matric suction is due to the physical affinity of water to the soil particle surfaces and capillary pores. Water tends to be drawn from a zone where the hydration envelopes surrounding the particles are thicker to where they are thinner, and from a zone where the capillary menisci are less curved to where they are more highly curved. In other words, water flows spontaneously from where matric suction is lower to where it is higher. When suction is uniform all along a horizontal column of soil the column is at equilibrium and there is no moving force. Not so when a suction gradient exists. In that case, water will flow in the pores which are water filled at the existing suction and will creep along the hydration films over the particle surfaces, with a tendency to equilibriate the potential. Vapour transfer is an additional mechanism of water movement in unsaturated soils. In the absence of temperature gradients, it is likely to be much slower than liquid flow as long as the soil is fairly moist. In the surface zone, however, where the soil becomes dry and strong temperature gradients occur, vapour transfer can become the dominant mechanism of water movement.

The moving force is greatest at the wetting front zone, where water invades and advances into a originally dry soil. Perhaps the most important difference between unsaturated and saturated flow is in the hydraulic conductivity. When the soil is saturated, all of the pores are water filled and conducting, so that continuity and hence conductivity are maximal. When the soil desaturates, some of the pores become air filled and the conductive portion of the soil's cross-sectional area decreases correspondingly. For these reasons, the transition from saturation to desaturation generally entails a steep drop in hydraulic conductivity.

At saturation, the most conductive soils are those in which large and continuous pores constitute most of the overall pore volume, while the least conductive are the soils in which the pore volume consists of numerous micropores. Thus, as is well known, a saturated sandy soil conducts water more rapidly than a clayey soil. However, the very opposite may be true when the soils are unsaturated. In a soil with large pores, these pores quickly empty and become non-conductive as suction develops, thus steeply decreasing the initially high conductivity. In a soil with small pores, on the other hand, many of the pores remain full and conductive water even at appreciable suction, so that the hydraulic conductivity does not decrease as steeply and may actually be greater than that of a soil with large pores subjected to the same suction.

4.4.2 Relation of Conductivity and Suction

Let us consider an unsaturated soil in which water is flowing under suction. In this model, the potential difference between the inflow and outflow ends is maintained not by different heads of positive hydrostatic pressure, but by different imposed suctions. In general, as the suction varies along the sample, so will the wetness and conductivity. If the suction head at each end of the sample is maintained constant, the flow process will be steady but the suction gradient will vary along the sample's axis. Since the product of gradient and conductivity is constant for steady flow, the gradient will increase as the conductivity decreases with the increase in suction along the length of the sample (Figure 4.3).

Since the gradient along the column is not constant, as it is in uniform saturated systems, it is not possible, strictly speaking, to divide the flux by the overall ratio of the head drop to the distance ($\Delta H / \Delta x$) to obtain the conductivity. Rather, it is necessary to divided the flux by the exact gradient at each point to evaluate the exact conductivity and its variation with suction. In the following treatment, however, we shall assume that the column of Figure 4.3 is sufficiently short to allow us to evaluate at least an average conductivity for the sample as a whole (i.e., $K = q \dfrac{\Delta x}{\Delta H}$).

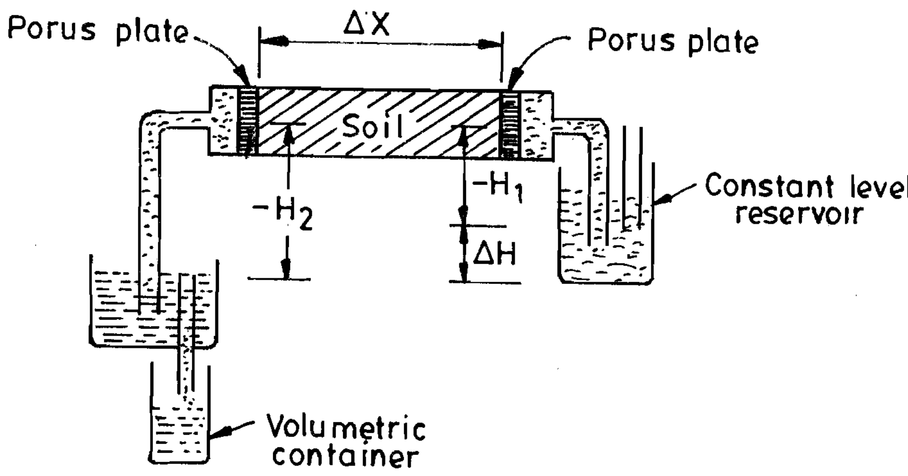

Figure 4.3 : Unsaturated Flow in a Horizontal Column - An Illustrative Mode'

The average negative head, or suction, acting in the column is

$$\bar{H} = \frac{1}{2}(H_1 + H_2)$$

We shall further assume that the suction everywhere exceeds the air entry value so that the soil is unsaturated throughout.

4.4.3 Hydraulic Diffusivity

To simplify the mathematical and experimental treatment of unsaturated flow processes, it is often advantageous to change the flow equations into a form analogous to the equations of diffusion and heat conduction, for which ready solutions are available in some cases involving boundary conditions applicable to soil water flow processes. To transform the flow equation, it is sometimes possible to relate the flux to the water content (wetness) gradient rather than to the suction gradient.

4.4.4 Vapour Movement

As we know that liquid water moves in the soil by mass flow, in a process by which the entire body of a fluid flows in response to differences in hydraulic potential. In certain special circumstances, water vapour movement can also occur as mass flow; for instance, when wind gusts induce bulk movement of air and vapour mixing in the surface zone of the soil. In general, however, vapour movement through most of the soil profile occurs by diffusion, a process in which different components of a mixed fluid move independently, and at times in opposite directions, in response to differences in concentration (or partial pressure) from one location to another. Water vapour is always present in the gaseous phase of an unsaturated soil, and vapour diffusion occurs whenever differences in vapour pressure develop within the soil.

The diffusion equation for water vapour is given below

$$q_d = -D_{vap} \frac{\Delta P_{vap}}{L}$$

where, ΔP_{vap} is the vapourpressure difference between two points at a distance L apart and D_{vap} is the diffusion coefficient for water vapour and q_d is the diffusion flux . D_{vap} in the soil is lower than in open air because of the restricted volume and the tortuosity of air filled pores.

SAQ 3

Differentiate between hydraulic conductivity and suction.

4.5 EVAPORATION FROM BARE SURFACE

Evaporation in the field can take place from plant canopies, from the soil surface, or, more rarely, from a free-water surface. Evaporation from plants, called transpiration, is the principal mechanism of soil-water transfer to the atmosphere when the soil surface is covered with vegetation. In the absence of vegetation, and when the soil surface is subject to radiation and wind effects, evaporation occurs directly and entirely from the soil.

4.5.1 Physical Conditions

Three conditions are necessary if the evaporation process from a given body is to persist. First, there must be a continual supply of heat to meet the latent heat requirement (which is about 590 cal/g of water evaporated at 15°C). This heat can come from the body itself, thus causing it to cool, or as is more commonly the case, it can come from outside the system in the form of radiated or advected energy. Second, the vapour pressure in the atmosphere over the evaporating body must remain lower than the vapour pressure at the surface of that body (i.e., there must be a vapour-pressure gradient between the body and the atmosphere), and the vapour must be transported away, by diffusion or convection, or both. These two conditions – namely, supply of energy and removal of vapour are generally external to the evaporating body and are influenced by meteorological factors such as air temperature, humidity, wind velocity, and radiation, which together determine the atmospheric evaporativity which being the maximal flux at which the atmosphere can vapourize water from a free-water surface.

The third condition is that there be a continual supply of water from or through the interior of the body to the site of evaporation. This condition depends upon the content and potential of water in the body as well as upon its conductive properties, which together determine the maximal rate at which the body can transmit water to the evaporation site. Accordingly, the actual evaporation rate is determined either by external evaporativity or by the soil's own ability to deliver water, whichever is the lesser and hence these are the limiting factors.

Among the various sets of conditions under which evaporation may occur are the following :

1) A shallow ground water table may be present at a constant or variable depth or it may be absent.
2) The soil profile may be uniform (homogeneous and isotropic).
3) The profile may be shallow, i.e., of finite depth, resting on bedrock or some other impervious floor, or it may be deep (semi-infinite).
4) The flow pattern may be one-dimensional (vertical), or it may be two or three – dimensional, as in the presence of vertical cracks which form secondary evaporation planes inside the profile.
5) External evnironmental conditions may remain constant or fluctuate.
6) Soil moisture flow may be governed by evaporation alone or by both evaporation (at the top of the profile) and internal drainage, or redistribution, down below.
7) The surface may or may not be covered by a layer of much (e.g., plant residues) differing from the soil in hydraulic, thermal and diffusive properties.
8) Finally, the evaporation process may be continuous over a prolonged period of time or it may be interrupted by regularly recurrent or sporadic episodes of rewetting (e.g., intermittent rainfall or scheduled irrigation).

4.5.2 Reduction in Evaporation from Bare Soils

In principle, the evaporation flux from the soil surface can be modified in three basic ways; by controlling energy supply to the site of evaporation (e.g., modifying the albedo through colour or structure of the soil surface; shading the surface) by reducing the potential gradient, or the force driving water upward through the profile (e.g., lowering the water table, if present, or warming the surface so as to set up a downward acting thermal gradient), or by decreasing the conductivity or diffusivity of the profile, particularly of the surface zone (e.g., tillage and mulching practices).

The actual choice of means for reduction of evaporation depends on the stage of the process one wishes to regulate; whether it be the first stage, in which the effect of meteorological conditions on the soil surface dominates the process, or the second stage, in which the rate of water supply to the surface, determined by the transmitting properties of the profile, becomes, the rate-limiting factor. Methods designed to affect the first stage do not necessarily serve during the second stage and vice-versa.

Covering or mulching the surface with vapour barriers or with reflective materials can reduce the intensity with which external factors, such as radiation and wind, act upon the surface. Thus, such surface treatments can retard evaporation during the initial stage of drying. A similar effect can result from application of materials which lower the vapour pressure of water. Retardation of evaporation during the first stage can provide a greater opportunity to the plants to utilise the moisture of the upper most soil layers, an effect which can be vital during the germination and establishment phases of plant growth. The retardation of initial evaporation can also enhance the process of internal drainage, and thus allow more water to migrate downward into the deeper parts of the profile, where it is conserved longer and is less likely to be lost by evaporation.

Shallow cultivation designed to pulverize the soil at the surface often has the immediate effect of causing the loosened layer to dry faster and more completely but many over a period of time help conserve the moisture of the soil underneath.

During the second stage of drying, the effect of surface treatments is likely to be only slight and reduction of the evaporation rate and of eventual water loss will depend on decreasing the diffusivity or conductivity of the soil profile in depth. Deep tillage, for instance, by possibly increasing the range of variation of diffusivity with changing water content, may reduce the rate at which the soil can transmit water toward the surface during the second stage of the drying process.

An irrigation regime having an excessively high irrigation frequency can cause the soil surface to remain wet and the first stage of evaporation to persist most of the time, resulting in a maximum rate of water loss. Water loss by evaporation from a single deep irrigation is generally smaller than from several shallow ones with the same total amount of water. However, water losses due to percolation are likely to be greater from deep irrigations than from shallow ones. New water application methods such as drip (or trickle) irrigation, which concentrate the water in a small fraction of the area while maintaining the greater part of the soil surface in a dry state, are likely to reduce the direct evaporation of soil moisture very significantly.

SAQ 4

i) Differentiate between the loss of water for cropped and uncroped land

ii) Discuss the factors affecting reduction of evaporation from bare soil surface.

4.6 UPTAKE OF WATER BY PLANTS

Nature, despite its celebrated laws of conservation, can in some ways be exceedingly wasteful, or so it appears at least from our own partisan viewpoint. One of the most glaring examples is the way it requires plants to draw quantities of water from the soil far in excess of their essential metabolic needs. In dry climates, plants growing in the field may consume hundreds of tons of water for each tone of vegetative growth. That is to say, the plants must inevitably transmit to an unquenchably thirsty atmosphere most (often well over 90%) of the water they extract from the soil. The loss of water vapour by plants, a process called transpiration, is not in itself an essential physiological function, nor a direct result of the living processes of the plants. In fact, plants can thrive in an atmosphere saturated or nearly saturated with vapour and hence requiring very little transpiration. Rather than by plant growth per se, transpiration is caused by the vapour pressure gradient between the normally water-saturated leaves and the often quite dry atmosphere. In other words, it is extract of the plants by the evaporative demand of the climate in which they live.

In a sense, the plant in the field can be compared to the wick in an old fashioned lamp. Such a wick, its bottom dipped into a reservoir of fuel while its top is subject to the burning fire which consumes the fuel, must constantly transmit the liquid from bottom to top under the influence of physical forces imposed upon the passive wick by the conditions prevailing at its two ends. Similarly, the plant has its roots in the soil-water reservoir while its leaves are subject to the radiation of the sum and the sweeping action of the wind which require it to transpire unceasingly.

4.6.1 Soil Plant Atmosphere Continuum (SPAC)

Current approaches to the problem of soil-water extraction and utilisation by plants are based on recognition that the field with all its parts soil, plant and atmosphere taken together forms a physically integrated, dynamic system in which various flow processes occur interdependently like links in a chain. This unified system has been called the SPAC (soil-plant-atmosphere continuum). The universal principle which operates consistently throughout the system is that water flow always takes place spontaneously from regions of higher to regions of lower potential energy.

The various terms used to characterize the state of water in different parts of the soil-plant-atmosphere system are merely alternative expressions of the energy level, or potential of water. Moreover, the very occurrence of differences, or gradients, of this potential between locations in the system constitutes the force inducing flow within and between the soil, the plant, and the atmosphere. This principles applies even though different components of the overall potential gradient are effective in varying degrees in different parts of the soil-plant-atmosphere system.

The flow path includes liquid water movement in the soil toward the roots, liquid and perhaps vapour movements across the root-to-soil contact zone, absorption into the roots and across their membranes to the vascular tubes of the xylem, transfer through the xylem up the stem to the leaves, evaporation in the intercellular spaces within the leaves, vapour diffusion through the substomatal cavities and out the stomatal perforations to the boundary air layer in contact with the leaf surface, and through it, finally, to the turbulent atmosphere which carries away the water thus extracted from the soil.

4.6.2 Plant Water Relations

Close observation of how higher plants are built reveals much about how they function in their terrestrial environment. To begin with, it is intriguing to compare the shape of such plants to the shape of familiar higher animals. The characteristic feature of animal bodies, in contrast with plants, is their minimal area of external surface exposure. Apart from a few protruding organs needed for mobility and sensory perception, animals are rather compact and bulky in appearance. Not so is the structure of plants, whose vital functions require them to maximize rather than minimize surface exposure both above and below ground. The aerial canopies of plants frequently exceed the area of covered ground by several fold. Such a large surface helps the plants to intercept and collect sunlight and carbon dioxide, two resources which are diffused rather than concentrated.

Even more striking is the shape of roots, which proliferate and ramify throughout a large volume of soil while exposing an enormous surface area: a single annual plant can develop a root system with a total length of several hundred kilometers and with a total surface area of several hundred square meters. The need for such exposure becomes apparent if we consider the primary function of roots, which is to continuously gather water and nutrients from a medium that often provides only a meager supply of water per unit volume and that generally contains soluble nutrients only in very dilute concentrations. And while the atmosphere is a well-stirred and thoroughly mixed fluid, the soil solution is a sluggish and unstirred fluid which moves toward the roots at a grudgingly slow pace, so that the roots have no choice but to move towards it. Indeed, roots forage constantly through as large a soil volume as they can, in a constant quest for more water and nutrients. Their movement and growth, involving proliferation in the soil region where they are present and extension into ever new regions, are affected by a host of factors additional to moisture and nutrients, e.g. temperature, aeration, mechanical resistance, the possible presence of toxic substances, and the primary roots' own geotropism (i.e. their preference for a vertically downward direction of growth).

Although all plants are absolutely dependent on water, different types of plants differ in adaptation to environments with varying degrees of water availability or abundance. **Hydrophytes**, or aquatic plants, inhabit water-saturated domains. Plants adapted to drawing water from shallow water tables are called **phreatophytes**. Plants which grow in aerated soils, generally in semi-humid to semi-arid climates, are called **mesophytes**. Most crop plants belong to this category. Mesophytes control their water economy by developing extensive root systems and optimizing the ratio of roots to shoots (the former supply water and nutrients, the latter photosynthesise and transpire) and by regulating the aperture of their stomates. On the dry end of the scale are **xerophytes**, which are adapted to growing in desert environments. Such plants generally exhibit special features (called xeromorphic) designed to minimise water loss (e.g., thickened epidermis and a waxy cuticle, recessed stomates and reduced leaf area, and specialised water storage or succulent tissues).

Only a very small fraction (generally less than 1%) of the water absorbed by plants is used in photosynthesis, while most (often over 98%) is lost as vapour, in a process known as transpiration. This process is made inevitable by the exposure to the atmosphere of a large area of moist cell surfaces, necessary to facilitate absorption of carbon dioxide and oxygen, and hence transpiration has been described as a necessary phenomena. Mesophytes are extremely sensitive, and vulnerable, to lack of sufficient water to replace the amount lost in transpiration. Water deficits impair plant growth and, if extended in duration, can be fatal. It was once believed that transpiration is beneficial in that it induces a greater uptake of nutrients from the soil. However, it now appears that the two processes of water uptake and nutrient uptake are largely independent. The absorption of nutrients into roots and their transmission up the shoots are not merely a passive consequence of the transpiration-induced water flow but are effected by metabolically active processes. A more likely beneficial effect of transpiration is prevention of excessive heating of leaves by solar radiation. Indeed, when water deficit occur and transpiration is reduced by stomatal closure, plant-canopy temperature rises measurably.

4.6.3 Root Uptake and Transpiration

The rate of water uptake from a given volume of soil depends on rooting density (the effective length of roots per unit volume of soil), soil conductivity, and the difference between average soil-water suction and root suction. If the initial soil-water suction is uniform throughout all depths of the rooting zone, but the active roots are not uniformly distributed, the rate of water uptake should be highest where the density of roots is greatest. However, more rapid uptake will result in more rapid depletion of soil moisture, and the rate will not remain constant very long.

The rooting system can be divided into two layers: an upper layer, in which root density is greatest and nearly uniform and in which water depletion is similarly uniform, and a lower layer, in which the roots are relatively sparse and in which the rate of water depletion is slow as long as the water content of the upper layer is fairly high. The water content of the lower layer is depleted by two, sometimes, simultaneous processes; uptake by the roots of that layer, and direct upward flow in the soil itself, caused by suction gradients.

A mature root system occupies a more or less constant soil volume of fixed depth so that uptake should depend mainly upon the size of this volume, its water content and hydraulic properties, and the density of the roots. On the other hand, in young plants, root extension and advance into deeper and moisture layers can play an important part in supplying plant water requirements.

One possible reason for the differences observed between the responses of pot-grown and of field-grown plants to the soil-water regime is the difference in root distribution with depth. In a pot, root density can be fairly uniform, while in the field, it generally varies with depth. Furthermore, the roots present in different layers may exhibit different water uptake and transmission properties. For instance, the roots of the deeper layers may offer greater resistance to water movement within the plant than the roots of the upper layers. The possible contribution of moist sublayers underlying the rooting zone can be especially significant where a high water table is present.

4.6.4 Soil Water Availability to Plants

Soil water is equally available throughout a definable range of soil wetness, from an upper limit (field capacity) to a lower limit (the permanent wilting point), both of which are characteristic and constant for any given soil. This indicates that the plant functions remain unaffected by any decrease in soil wetness until the permanent wilting point is reached, at which plant activity is curtailed abruptly. This schematised model, though based upon arbitrary limits, enjoys widespread acceptance for last many years, particularly among workers in the field of irrigation management.

However, now we know that soil-water availability to plants actually decreases with decreasing soil wetness, and that a plant may suffer water stress and reduction of growth considerably before the wilting point is reached. Attempts to divide the so-called 'available range' of soil wetness into 'readily available' and 'decreasingly available' ranges, and search for a 'critical point' somewhere between field capacity and wilting point as an additional criterion of soil-water availability is now also accepted.

4.6.5 Water Balance in Root Zone

Water balance in a root zone of defined given volume of soil can be expressed as

$$\Delta W = W_{in} - W_{out}$$

Where, ΔW is the change in water content, W_{in} is the water added and W_{out} is the water extracted from the soil mass in a particular period of time. The above equation can be

$$\Delta W = \underbrace{P + I}_{W_{in}} - \underbrace{[N + F + (E + T)]}_{W_{out}}$$

further rewritten as

where,

- P = Precipitation,
- I = Irrigation,
- N = Runoff,
- F = Deep Percolation,
- E = Evaporation from Surface, and
- T = Transpiration.

SAQ 5

i) Describe the SPAC in the context of water uptake by crop plants.

ii) What are the uptake processes of water in plant root?

iii) Write the water balance equation for soil root zone.

4.7 GROUND WATER DRAINAGE

4.7.1 Drainage

The term "Drainage" is used in a general sense to denote outflow of water from soil. More specifically, it can serve to describe the artificial removal of excess water, or the set of management practices designed to prevent the occurrence of excess water. Ground water drainage refers to the outflow or artificial removal of excess water from within the soil, generally by lowering the water table or by preventing its rise.

4.7.2 Benefits

Excess water in the soil tends to block soil pores and thus retard aeration and in effect strangulate the roots. In water logged soils, gaseous exchange with the atmosphere is restricted to the surface zone of the soil, while within the profile proper oxygen may be almost absent and carbon dioxide may accumulate. Under anaerobic conditions, various substances are reduced from their normally oxidised states. Toxic concentrations of ferrous sulphide, and manganeous ions can develop. These, in combination with products of the anaerobic decomposition of organic matter (e.g., methane) can greatly inhibit plant growth. At the same time, nitrification is prevented, and various plant and rot diseases (especially fungal) are more prevalent.

The occurrence of a high water table condition may not always be clearly evident at the very surface, which may be deceptively dry even while the soil is completely water logged just below the surface zone. Where the effective rooting depth is thus restricted, plants may below the surface zone. Where the effective rooting depth is thus restricted, plants may suffer not only from lack of oxygen in the soil, but also from lack of nutrients. If the water table drops periodically, plants growing in water logged soils may even paradoxically, suffer from occasional lack of water, especially when the transpirational demand is very high. High moisture conditions at or near the soil surface cause the soil to be susceptible to compaction by animal and machinery traffic.

SAQ 6

Define drainage describing adverse effect of excess water on plants growth.

4.8 INFILTRATION

Infiltration is the entry of the water into soil through the soil surface. Nearly all water which enters the soil water via infiltration, either rainfall or from the water, directed to the soil surface via irrigation. Knowledge of the infiltration characteristics of a soil is basic information required for designing efficient irrigation systems or predicting runoff. The soil infiltration rate will affect advance and recession time, deep percolation and tailwater runoff in furrow and border system, advance and ponded times in level borders (basins), and maximum allowable application rate for sprinkler or drip irrigation systems.

The rate at which water is absorbed by the soil is the infiltration rate, I, which has units of volume per unit area per unit time, $cm^3/cm^2/min$, or equivalent depth per unit time, cm/min. The cumulative infiltration, z, is the total accumulated depth infiltrated in the given time period. Cumulative infiltartion is thus the integration of the infiltration rate, and conversely, I is the derivative of z.

$$z(t) = \int I(t)\, dt$$

$$I(t) = \frac{dz}{dt}$$

The rate at which water infiltrates is usually decreases over time until a relatively steady basic infiltration rate is asymptotically approached. The decrease is caused by the decreasing capillary pressure of the soil as it becomes wet. The constant rate is due to the constant gravity force, the other component of driving force.

Due to the change in driving forces (hydraulic gradient) and the varying moisture content of the soil, infiltration rate of a soil cannot be directly related to its hydraulic conductivity. This is why a specialised measurement must be made. Layers with low conductivities, either at or below the surface, will limit the infiltration rate – especially the basic rate. Consequently, surface crusts, silt deposits, implements tracks, plow layers, or clay layers will have a major effect on infiltration.

Pre-existing soil moisture content will affect the initial infiltration rate. Drier soils will have higher initial rates, due to their high capillary pressure. Saturated soils initially infiltrate water at near the basic rate.

Infiltration rate within one area or field are highly variable – variation as high as an order of magnitute in one field are not uncommon due to non-homogeneities in the soil (both textural and structural) and variation in the above-mentioned factors. This variation is both areal and temporal. The infiltration rate of a wheel track furrow is usually much less than that of a less compacted furrow. The intake rate during the first irrigation after cultivation is usually higher than during later irrigation. Consequently, time and locations for infiltration measurement must be chosen carefully in order for the results to be representative and meaningful. Some factors to cosider include :

1) Soil moisture content at the time of irrigation
2) Soil texture
3) Subsurface strata (plow pan, clay layers)
4) Surface soil compaction (wheel tracks)

5) Surface sealing (sedimentation, erosion, dispersion)
6) Soil cracking
7) Soil structure (tillage operations)
8) Irrigation method (sprinkler vs. surface; border vs. furrow)
9) Crops and surface mulches
10) Soil and water salt ion concentrations
11) Soil and water temperature

Due to the variability, several measurerments are required to achieve a reliable estimate of the infiltration rate for a field.

Several mathematical relationship or models have been proposed for infiltration rate and cumulative infiltration, including the empirical Kostiakov-Lewis equation:

$$z = K t^{a}$$

The empirical relationship used by the U.S. Soil conservation services is

$$z = K t^{a} + C$$

where,

z = accumulated infiltration rate in time t,

t = elapsed time, and

a, K, C are characteristic constants.

The typical mass infiltration curve can also be drawn using the expression

$$f = f_c + (f_o - f_c)\, e^{-kt}$$

where,

f = infiltration capacity [LT^{-1}],

f_c = the constant infiltration capacity at t approaches infinity,

f_o = infiltration capacity at the onset of infiltration,

k = a constant for a given soil and initial condition, and

t = time.

4.8.1 Methods

Several methods can be used to determine the infiltration characteristics of a soil under field conditions. Methods frequently used for basin and border irrigation are:

1) Cylinder (ring) infiltrometer, and
2) Basin infiltrometer.

The two common methods used for furrow irrigation are:

3) Blocked furrow infiltrometer, and
4) Inflow-outflow measurments to both border and furrow.

An additional method applicable to both border and furrow irrigation is:

5) Volume balance techniques based upon the rate of advance of the water front and estimated or measured values of the volume of water in surface storage.

Cylinder and blocked furrow infiltrometers are used due to their ease of use, precision, and for their simplicity. Now we shall discss about ring and blocked furrow infiltrometers.

Ring Infiltrometer

The cylinder or ring infiltrometer is a metal cylinder which is driven into the soil. It measures primarily the vertical rate of water movement into the soil surface (one-dimensional) from the pond it encloses. In cylinder infiltrometer, infiltration data are obtained by measuring the depth of ponded water inside the cylinder at time intervals.

After water penetrates the soil to the depth of the bottom of the cylinder, it will begin to spread radially as well as vertically and the infiltration rate change accordingly. Another pond can be constructed by forming an earthen dike around the cylinder or by driving a larger diameter cylinder into the soil concentric with the cylinder infiltrometer. Water is maintained in the area between the two cylinder at about the same depth as that in the inner cylinder.

Two problems may make infiltration rates measured by cylinder infiltrometers unrepresentative of true infiltration rates. One is that the soil may be disturbed when the rings are driven into the ground. The second is that the ponded water may not duplicate real conditions. For example, although infiltration rates are not too sensitive to the water pressure at the surface, or equivalently the depth of water on the surface, water pressure will have some effect. The infiltration rate measured with an infiltrometer with 20 cm of ponded water will not be equivalent to that in a border with 5 cm water depth, or that under a sprinkler where no water is ponded. Also air which is trapped under a level basin may escape laterally from around the infiltrometer. Finally, under real conditions, water usually moves across the soil surface rearranging surface particles. This reason and aggregation which does not occur in ponded infiltration tests tends to seal soil surface and reduce infiltration rates.

Blocked Furrow Infiltrometer

The blocked furrow infiltrometer measures the infiltration of water into the soil profile from a short segment of a furrow (about 1 meter). The water is ponded in the section by barriers driven into the soil across the furrow. Furrow infiltration is two-dimensional because it includes both vertical and horizontal movement of water into the soil. After the water penetrate the soil surface it will begin to spread laterally as well as vertically. In order to prevent excess lateral movement of water and simulate the actual irrigation furrows adjacent to the tested furrow are filled with water. These are known as buffer furrows.

Since the wetted surface will vary as the water depth in the furrow varies, maintaining the water level in the furrow near normal depth is important. The normal flow depth should be determined, if possible, while the furrow is being used during a normal irrigation.

The main problem with blocked furrow infitrometer measurments is that they are made with ponded water. Water running down furrow will erode & deposit surface sediments which can cause furrow cross section and surface infitration charecterstics to change.

4.8.2 Equipment and Procedure

Equipment

1) A cylinder infiltrometer, 20 to 30 cm in diameter and 30 to 40 cm in length made of smooth steel, rigid enough to allow it to be driven into the ground, but still thin enough to enter the soil with minimum distance. Two millimetre thick (14 gauge) steel will usually work, but may need reinforcement around the upper edge. The larger the diameter of the cylinder and the deeper the cylinder penetrates the soil, the less will be the edge effects and the greater will be the accuracy of measurements.
2) A buffer cylinder having a diameter at least 30 cm greater than the infiltrometer and a length of about 20 cm. Construction should be similar to that of the infiltrometer. Sections of 220 litre drums can also be used as buffer cylinders. Alternately, the buffer 'ring' can also be a diked surrounding the cylinder infiltrometer.
3) A water level gauge for measuring the changes of water level in the cylinder infiltrometer. A simple hook gauge or other arrangements are also possible to meet this requirement.
4) Two metal sheets, about 90 cm wide, 30 cm high and 0.20 cm thick, with reinforcement at the upper end.
5) Marriott syphon or other regulated supply reservoir with attached water level gauge.
6) Staff gauge or other means to indicate the constant water level required in the furrow.
7) Equipment for installation of the cylinders and furrow dams such as a metal plate or heavy timber and a sledge hammer.
8) A plastic sheet or other waterproof membrane.
9) Source of water (about 200 litres for each ring infiltrometer and 800 litres for each blocked furrow test).
10) Bucket, shovel, watch.

Procedure

1) Select a repersentative location for each cylinder and examine carefully for signs of usual surface disturbance, such as stones that might damage cylinders, cracks which might give nonrepresentative readings, void area that may have been affected by usual animal or machinery traffic. Note the soil type and condition (tillage, vegetation, moisture).

2) Drive the infiltrometer cylinder to a depth of at least 15 cm. The cylinder should be installed as vertically as possible. This can be assured by checking the alignment frequently during the installation procedure. Do not drive the cylinder into the soil irregularly so that first one side then the other goes down. This procedure produces poor contact between the cylinder wall and the soil, and may disturb the cylinder gates out of alignment while driving.

3) Set the buffer cylinder around the infiltrometer and drive it into the soil. This outside cylinder need not be driven as deep as the cylinder infiltrometer – generally, 5 to 10 cm into the soil will be adequate.

4) Place a plastic sheet or other waterproof membrane on the soil within the cylinder infiltrometer so that it forms an inner cylinder to hold the water. The plastic should be in contact with the soil at the bottom of the infiltrometer and extend up the walls of the infiltrometer at least 15 cm.

5) Fill the buffer pond with water to a depth roughly equal to the depth desired in the inner ring. Maintain roughly equivalent depths throughout the period of observation.

6) Fill the cylinder infiltrometer with water to a depth of about 10 cm; in extermely porous (high intake) soils, a greater depth may be needed.

7) Place the hook gauge board on the cylinder and set the point of the gauge at the water level. Read the scale from top edge of the clip. This is the initial depth reading and will correspond to the initial time reading.

8) Quickly, but gently, remove the plastic membrane recording the time at which this is done. This is the initial time reading. Make a hook gauge to do so, and compare this reading to the initial one to insure that there were no air space below the plastic.

9) Make additional hook gauge reading at periodic intervals and record the hook gauge and time readings. Intervals between observations should be increased to 10 to 20 min. After about 2 hours, measurements made at 30 to 50 min intervals will usually be sufficient.

10) When the water level goes down by 4 to 5 cm in the cylinder infiltrometer, carefully add a sufficient volume of water to return the water surface to the approximate initial level. The known volume of the cylinder, gives the depth of water added to the cylinder. If a volumetric container is not available the added depth can be closely estimated by plotting the depth reading over time just before and after the water was added. The volume added and corresponding depth added, or the projected depth added should be recorded on the data collection sheet.

11) Where abnormally high or low infiltration values are indicated by the soil examined, cause for a high rate of infiltration can be seen, such as water rising outside the cylinder, the test should be terminated immediately and the cylinder moved to a near by location and reinstalled.

12) The measurements should be continued until the change in depth over time becomes fairly constant. The time equivalent will generally vary from 3 to 6 hours, depending on the pre–existing moisture condition and soil hydraulic conductivity.

4.9 SUMMARY

Soil water movement is a dynamic property and fluid dynamic laws, i.e., Darcy's and Poiscuille's laws govern the flow of water in soils. Flow of water in saturated or unsaturated condition is governed by the hydraulic conductivity of soil and operating head/suction head. Water is lost from the soil surface and is chiefly affected by evaporative demand of the atmosphere, soil water available for evaporation and movement of water at sub-atmospheric pressure. Uptake of water by the plant is affected by the biological activity of plant, evapotranspiration demand, rooting activity, moisture release pattern of soil, soil

water movement etc. Downward entry of water is important soil water behaviour for irrigation and drainage also the ability of a soil to transmit water is important behaviour in soil water movement.

4.10 KEY WORDS

Perfect Fluid	:	One that is frictionless and incompressible.
Poiseuille's Law	:	The total flow rate of water through a capillary tube is proportional to the fourth power of the radius while the flow rate per unit cross-sectional area of the tube is proportional to the square of the radius.
Gravitational Head	:	Pressure head caused due to force of gravity.
SPAC	:	Soil Plant Atmosphere Continuum.
Readily Available Moisture	:	Soil Moisture between field capacity and allowable depletion limit.
Anaerobic Condition	:	A condition without oxygen.
Hydraulic Conductivity	:	It is the measure of how easily water can move through soils. This is proportional to the driving force.

4.11 ANSWERS TO SAQs

Please refer the preceding text for answers of all the SAQs.

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