

**SAWAM-PAK**

**SUSTAINABLE AGRICULTURAL WATER MANAGEMENT IN PAKISTAN:  
ADDRESSING FOOD SECURITY AND SOCIAL INSTABILITY**

# **LECTURE ON BASIC PRINCIPLES OF SOIL HYDROLOGY**



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# 1 SECTION IV: The physical and mechanical characteristics of soil

The physical condition of the soil can in some circumstances have an overriding effect on plant growth, irrespective of the chemical fertility of the soil. If the physical condition is unsuitable, then plant growth will be reduced or even stopped completely. The 'physical fertility' of the soil must be maintained as well as the chemical fertility.

As illustrated in Figure 1, the soil system consists of three basic phases: solids, liquid and gas. Each of these phases will be examined in detail before the management of soil physical fertility is discussed.

## 1.1 Basic definitions

The following basic terms will be used throughout the succeeding chapters and are defined together in this section to allow easy reference. A list of the symbols used is given at the end of the book. The schematic diagram of soil in Figure 1 will be used as the basis for the definitions and should be examined closely. The three basic phases have been separated out and their mass and volume relations are shown at either edge of the diagram. The total mass of soil ( $M$ ) is the sum of the mass of solids ( $M_s$ ) plus the mass of water ( $M_w$ ) and the relatively small, but nevertheless defined, mass of air ( $M_a$ ). The total volume of the three phases ( $V$ ) is the sum of the volume of solids ( $V_s$ ) plus the volume of water ( $V_w$ ) and the volume of air ( $V_a$ ). The volume of pores ( $V_r$ ) is equal to the volume of water plus air.

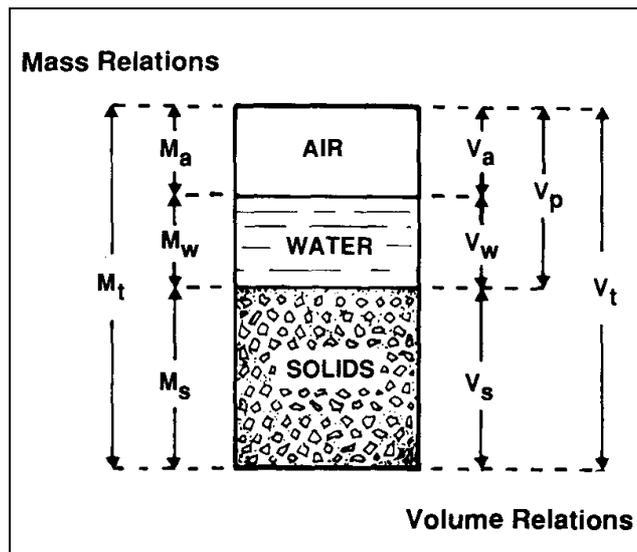


Figure 1. Schematic diagram showing the volumes and masses of the three phases in soil.

### 1.1.1 Particle Density ( $\rho_p$ )

$$\rho_p = \frac{\text{mass of solids}}{\text{volume of solids}} = \frac{M_s}{V_s} \quad (5.1)$$

The average particle density is the ratio of the total mass of solids to the volume of solid material. It is usually expressed in units of grams per cubic centimetre ( $\text{g cm}^{-3}$ ). For example, if 1 cubic centimetre of soil solids has a mass of 2.65 grams then the particle density is  $2.65 \text{ g cm}^{-3}$ .

Although particle density can be measured relatively easily, for most mineral soils it is sufficient to assume a standard value of  $2.65 \text{ g cm}^{-3}$ . This is because the quartz, feldspar and silicate minerals which make up the major proportion of the particles in mineral soils have approximately this density. The particle density of unweathered pumice varies with the amounts of heavy minerals present, but values from under  $2.0$  to over  $3.5 \text{ g cm}^{-3}$  are possible. The particle: density of weathered

pumice is similar to that of other mineral material (i.e. 2.65 g cm<sup>-3</sup>). Organic matter has a particle density of about 1.3 g cm<sup>-3</sup>.

### 1.1.2 Soil dry bulk density (P<sub>b</sub>)

$$\begin{aligned} \rho_b &= \frac{\text{mass of dry soil}}{\text{total volume of soil}} = \frac{M_s}{V_t} \\ &= \frac{M_s}{(V_s + V_w + V_a)} \end{aligned} \quad (5.2)$$

The soil dry bulk density is the ratio of the mass of dry soil to the total volume of soil (i.e. solids plus pores), not just the volume of solids as in the particle density. Bulk density is normally expressed in grams per cubic centimetre.

Because bulk density takes into account the pore space in the soil it can give an indication of the level of compaction or, conversely, porosity of a soil. A recently cultivated topsoil is likely to have a bulk density in the range from 0.9 to 1.2 g cm<sup>-3</sup>, whilst a compact subsoil may have a bulk density of 1.6 g cm<sup>-3</sup> or even more.

The dry bulk density of a soil can be measured by taking an undisturbed core of soil, using a tube or ring of known volume, and measuring the mass of soil removed after it has been dried at 105°C. For example, if a 2.5 cm radius (r) tube was used to sample a 5 cm depth of soil (h) the volume collected would be 98.2 cm<sup>3</sup>. If the mass of dry soil was 147.2 grams then the bulk density would be 1.5 g cm<sup>-3</sup>.

Table 1 gives some examples of the bulk density values commonly found in New Zealand soils. It can be seen from the table that the bulk density is lower in all topsoils compared with their subsoils and that soils derived from volcanic ash (e.g. Egmont) have lower bulk densities than soils derived from many other parent materials, such as greywacke alluvium (e.g. Temuka). Volcanic ash soils can typically contain 15-20% of organic matter by mass in the topsoil, and this can reduce the bulk density further.

### 1.1.3 Porosity (E)

$$\begin{aligned} \epsilon &= \frac{\text{volume of pores}}{\text{total volume of soil}} = \frac{V_p}{V_t} \\ &= \frac{(V_w + V_a)}{(V_s + V_w + V_a)} \end{aligned} \quad (5.3)$$

Porosity is the ratio of the volume of pores to the total volume of soil. Most soils have a porosity of between 0.3 and 0.6, which is often expressed as a percentage (i.e. 30% and 60%). Porosity is a useful descriptor of the soil and will be discussed further in Chapter 8. Porosity varies inversely with bulk density and some values for New Zealand soils are given in Table 1.

Porosity can be calculated from dry bulk density and the average particle density of the soil using the following equation:

$$\epsilon = 1 - \frac{\rho_b}{\rho_p} \quad (5.4)$$

For example, a soil with a bulk density (P<sub>b</sub>) of 1.40 g cm<sup>-3</sup> and a particle density (p<sub>r</sub>) of 2.65 g cm<sup>-3</sup> would have a porosity of 47.2%.

#### 1.1.4 Air-filled porosity ( $\epsilon_a$ )

$$\begin{aligned}\epsilon_a &= \frac{\text{volume of air}}{\text{total volume of soil}} = \frac{V_a}{V_t} \\ &= \frac{V_a}{(V_s + V_w + V_a)}\end{aligned}\quad (5.5)$$

The air-filled porosity of a soil is the ratio of the volume of air to the total volume of soil and gives an indication of the soil's aeration status, as will be discussed. In practical terms, if the air-filled porosity of a soil is below 0.1(10%) at field capacity then drainage is recommended.

#### 1.1.5 Void ratio ( $e$ )

$$\begin{aligned}e &= \frac{\text{volume of pores}}{\text{volume of solids}} = \frac{V_p}{V_s} \\ &= \frac{(V_w + V_a)}{V_s}\end{aligned}\quad (5.6)$$

The void ratio is the ratio of the volume of pores to the volume of solids. This is a useful index when the total volume of soil changes due to compaction. The void ratio is more commonly used in engineering and mechanics, whilst porosity is used in agriculture. The following relations are often useful:

$$e = \frac{\epsilon}{1 - \epsilon} \quad (5.7)$$

$$\epsilon = \frac{e}{1 + e} \quad (5.8)$$

**Table 1.** Bulk density and porosity of some New Zealand soils (adapted from Soil Bureau, 1968).

Soil series	Group	Horizon	Texture	Bulk density (g cm <sup>-3</sup> )	Porosity (%)
Conroy	Brown-grey earth	A	sandy loam	1.08	57.1
		B	silt loam	1.28	53.8
Timaru	Yellow-grey earth	A	silt loam	1.01	60.0
		B	silt loam	1.68	37.4
Puketeraki	Yellow-brown earth	A	silt loam	0.58	74.7
		B	silt loam	1.01	61.7
Temuka	Gley	A	silty clay loam	0.93	60.2
		B	silty clay loam	1.41	47.2
Taupo	Yellow-brown pumice soil	A	loamy sand	0.57	74.3
		B	sandy loam	0.66	71.0
Egmont	Yellow-brown loam	A	loam	0.66	70.0
		B	loamy sand	0.73	69.5
Hamilton	Brown-granular loam	A	clay loam	0.86	63.6
		B	clay	0.84	67.9

Arapohue	Rendzina	A	clay	0.73	67.5
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### 1.1.6 Soil water content: mass basis ( $\theta_m$ )

$$\theta_m = \frac{\text{mass of water}}{\text{mass of dry soil}} = \frac{M_w}{M_s} \quad (5.9)$$

When the soil water content is expressed on a mass basis it is calculated as the ratio of the mass of water to the mass of solids and is called the **gravimetric water content**. This is often expressed as a percentage and typical values at **saturation** range from 30 to 60% in mineral soils. Much higher values, over 100%, are possible in organic soils and in volcanic soils.

### 1.1.7 Soil water content: volume basis ( $\theta_v$ )

$$\begin{aligned} \theta_v &= \frac{\text{volume of water}}{\text{total volume of soil}} = \frac{V_w}{V_t} \\ &= \frac{V_w}{(V_s + V_w + V_a)} \end{aligned} \quad (5.10)$$

When the soil water content is expressed on a volume basis it is calculated as the ratio of the volume of water to the total volume of soil and is called the volumetric water content. The volumetric water content value may range from 40% in saturated sandy soil to 60% in a saturated clay soil. The measurement and use of both the gravimetric and volumetric water content will be discussed in more detail further.

To convert from a mass to a volume basis or *vice versa* the following relations may be used:

$$\theta_m = \theta_v \times \frac{\text{density of water } (\rho_w)}{\text{soil bulk density}} = \frac{\theta_v}{\rho_b} \quad (5.11)$$

$$\theta_v = \theta_m \times \frac{\text{soil bulk density}}{\text{density of water } (\rho_w)} = \theta_m \rho_b \quad (5.12)$$

where the density of water ( $\rho_w$ ) is assumed to be  $1.0 \text{ gcm}^{-3}$ .

### 1.1.8 Degree of saturation (s)

$$s = \frac{\text{volume of water}}{\text{volume of pores}} = \frac{V_w}{V_p} = \frac{V_w}{(V_w + V_a)} \quad (5.13)$$

The degree of saturation is the ratio of the volume of water to the volume of pores. The degree of saturation can range from about zero % in a dry soil to about 100% in a completely saturated soil. The degree of saturation and volumetric water content are related by the following equation:

$$s = \frac{\theta_v}{\epsilon} \quad (5.14)$$

### 1.1.9 Soil particle size, texture and specific surface area

Soils contain mineral particles of different sizes. The smallest particles are in the **clay** fraction and it is this material which gives the soil many of its key chemical and physical properties. The larger sized **silt** and **sand** particles, although relatively inactive in chemical terms, do provide the soil with certain essential physical characteristics.

*1.1.9.1 Particle size*

In soil, any mineral material greater than 2 mm is generally referred to as gravel, stones or boulders (see Chapter 1). Smaller sized material, which usually makes up the bulk of the soil, is classified according to its equivalent particle diameter. Several classification schemes exist, as shown in Figure 2. The International Soil Science Society scheme will be followed in this text as it is the scheme most commonly used in New Zealand.

Sand-size particles range from 2 mm to 0.02 mm (or 2000-20  $\mu\text{m}$ : where 1  $\mu\text{m}$  = 0.001 mm) and can be further classified as coarse, medium or fine sand. Sands are generally sharp angular fragments of weathered rock and mainly consist of the mineral quartz, with smaller amounts of other minerals such as feldspars and micas. Sand grains are generally resistant to weathering and are chemically inactive except when coated with clay-sized material. This means that they are not important in the supply or storage of plant nutrients. Nevertheless they can have a major influence on other soil properties such as soil drainage and aeration.

Silt particles range in size from 0.02 mm to 0.002 mm (or 20-2  $\mu\text{m}$ ) and are usually similar in mineralogy to sand grains. Silt normally provides little in the form of nutrient supply or storage, unless they are coated with clay material.

Clay material is less than 0.002 mm (or 2  $\mu\text{m}$ ) in size and is the most surface-active soil mineral component.

Clay particles mostly consist of weathered mineral material referred to as secondary minerals. Clay has the ability to adsorb and hold nutrients as well as to hold water. Although clay is obviously of prime importance to the productivity of a soil, too much clay can produce a soil which is poorly drained and is difficult to cultivate. There is a need for a balance of sand, silt and clay in the soil to ensure that it has optimum growing properties.

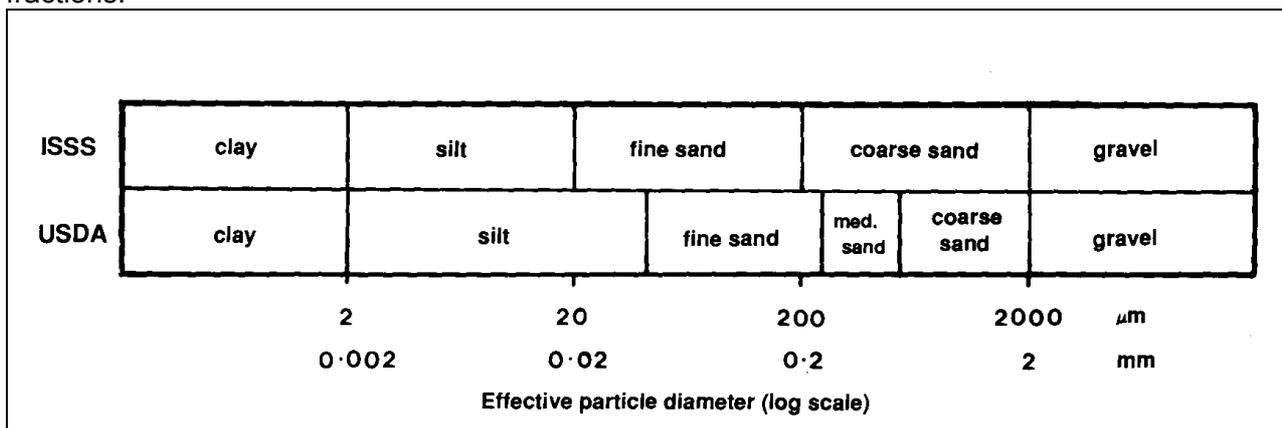
*1.1.9.2 Soil texture*

The relative proportions of sand, silt and clay give the soil its **texture**. Soil texture is an important characteristic because it gives a good indication of other soil properties such as water storage, drainage and nutrient supply. It is also a stable soil property and is not likely to change with normal soil management.

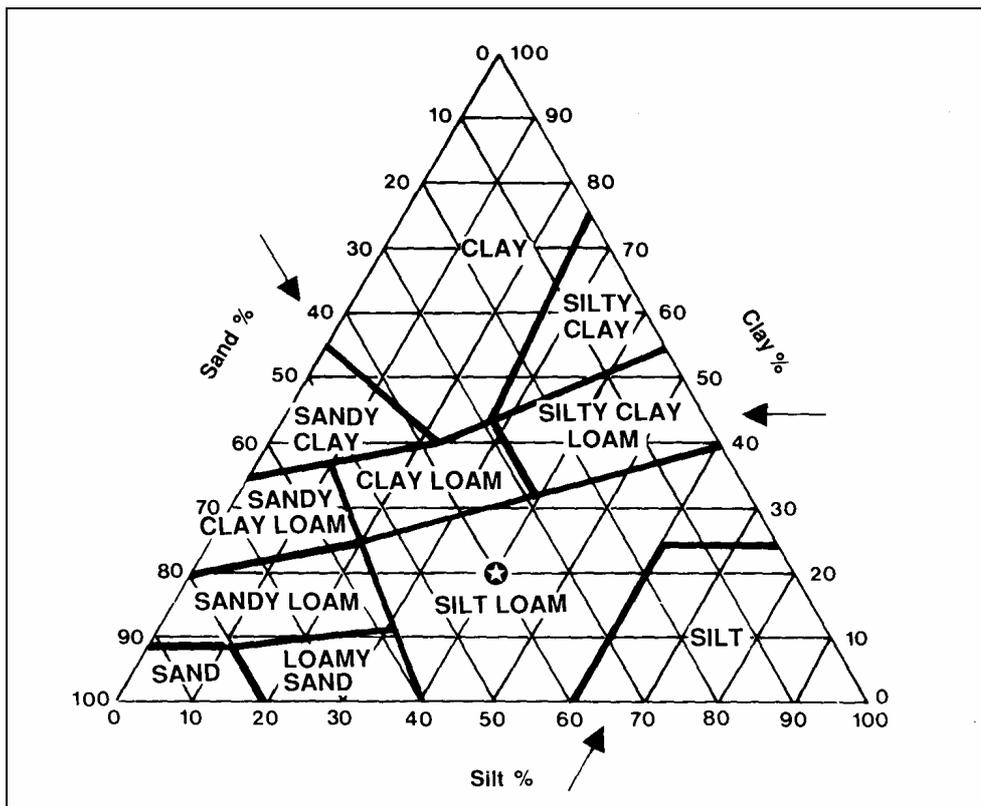
Soil texture can be estimated in the field by some practical tests involving the 'feel' of the soil, or it can be determined in the laboratory by measuring the amounts of each size fraction of soil material. The latter method is called **mechanical analysis** and will be described in the next section.

Soil texture classes are defined according to the **texture triangle** shown in Figure 3. To use this triangle the relative percentages of sand, silt and clay must be measured or known and these are read off the appropriate axis. In the example shown, a soil with 40% sand, 40% silt and 20% clay is classified as a silt loam. A further example is a soil with 60% sand, 20% silt and 20% clay which is classified as a sandy loam.

As can be seen from the texture triangle, each soil texture class has some mixture of all three size fractions.



**Figure 2.** Particle size classification. The International Society of Soil Science (ISSS) scheme is generally used in International environment. The United States Department of Agriculture (USDA) scheme is shown for comparison.



**Figure 3.** Texture triangle. Soils contain a range of particle sizes and the relative proportions of sand, silt, and clay define the soil texture. In the example shown, a soil with 40% sand, 40% silt, and 20% clay is classified as a silty loam.

Soils which have a 'uniform' mixture of sands, silts and clays are generally called loams and are probably the most important for agriculture. Soils which have an extreme of one size fraction are often more difficult to manage (e.g. clay or sand). Those which have a dominantly sandy texture (e.g. sandy loam) are often referred to as 'light' and those with a higher proportion of clay as 'heavy'. These terms, light and heavy, refer to the soil's ease of cultivation and do not indicate the relative density or mass of the soil particles.

*1.1.9.3 Particle-size distribution and mechanical analysis*

**Particle-size distribution (PSD) curves** - As can be seen in Figure 2, soil particles are grouped into different size classes. Soil particles, however, rarely exist in discrete size classes, and usually form a continuum of sizes. For this reason it is often more appropriate to construct a 'particle-size distribution curve' which shows the cumulative percentage of particles less than a specific size. Such a curve is shown in Figure 4 and gives the sizes on a logarithmic scale to allow for the wide range of sizes to be presented on one graph. The steeper the curve the greater the amount of particles of that particular size. Examples of particle-size distributions in a clay, silt loam and sandy loam are shown in Figure 4. The PSD curve can also show whether a soil is 'well-graded, i.e. composed of a continuous array of particle sizes, or whether the soil is 'poorly graded', i.e. composed of distinct groups of size fractions.

**Mechanical analysis** - Laboratory methods of measuring the particle-size distribution of soil samples are well established and are commonplace in many soils and engineering laboratories. Some of the texts listed at the end of this chapter give detailed methods but most follow the procedure outlined below. Firstly, the soil sample is air-dried and passed through a 2 mm sieve to remove any stones or gravel. The fine earth material which is left is still in the form of soil aggregates and the second stage is for these aggregates to be dispersed. Soil aggregates are held together by organic matter, clay and other soil components. Therefore, the organic matter is removed by oxidation, usually with hydrogen peroxide, and the clays are dispersed by the addition

of a dispersion agent, such as sodium hexametaphosphate, and by mixing or shaking. Ultrasonic dispersion is used in some laboratories to produce a similar suspension of dispersed soil in an aqueous solution.

Figure 4. Particle-size distribution curves for three different textured soils.

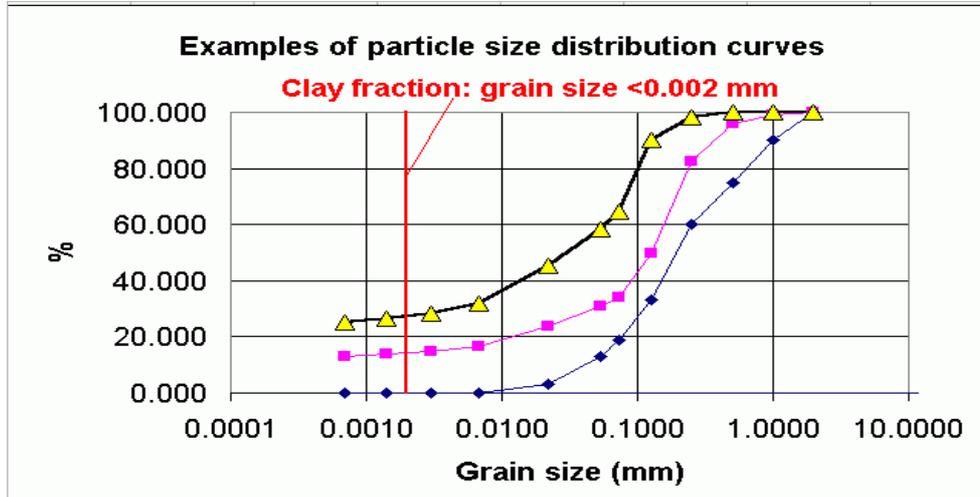
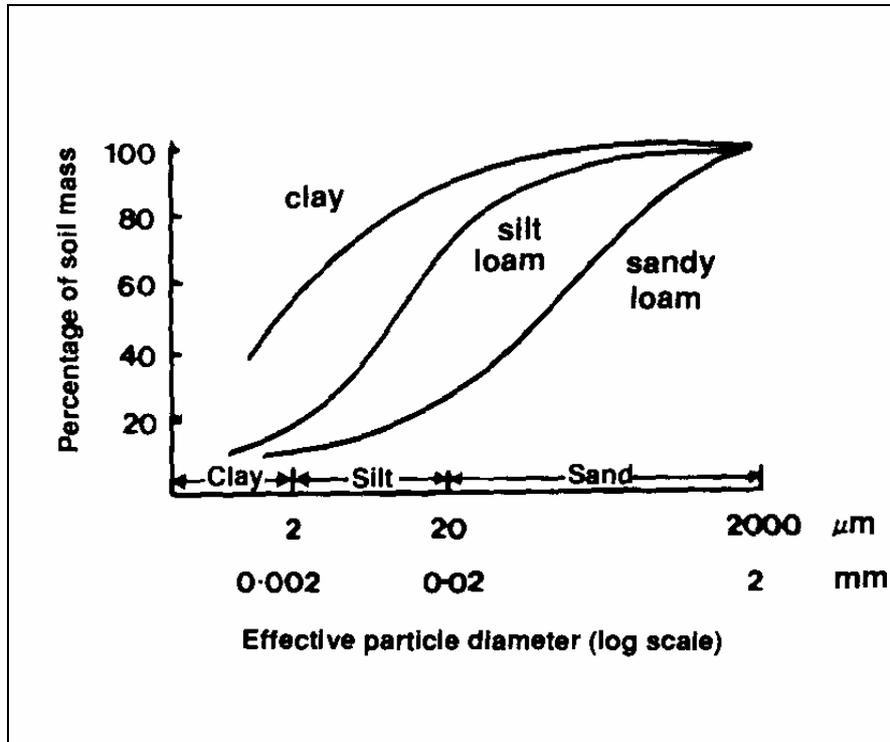


Fig. 4b. Examples of particle-size distribution curves.

The sand-sized material is usually removed by passing the solution through a bank of sieves. The finer material is then separated by sedimentation. This process relies on different-sized particles taking different times to settle, or fall out of, suspension in the liquid. The physical principles involved are well established and follow **Stokes Law**. This law states that the terminal velocity ( $v$ ) of a spherical particle falling through water under the influence of gravity ( $g$ ) is proportional to the square of its radius ( $r$ ):

$$v = \frac{2(\rho_p - \rho_w) gr^2}{9\eta} \tag{5.15}$$

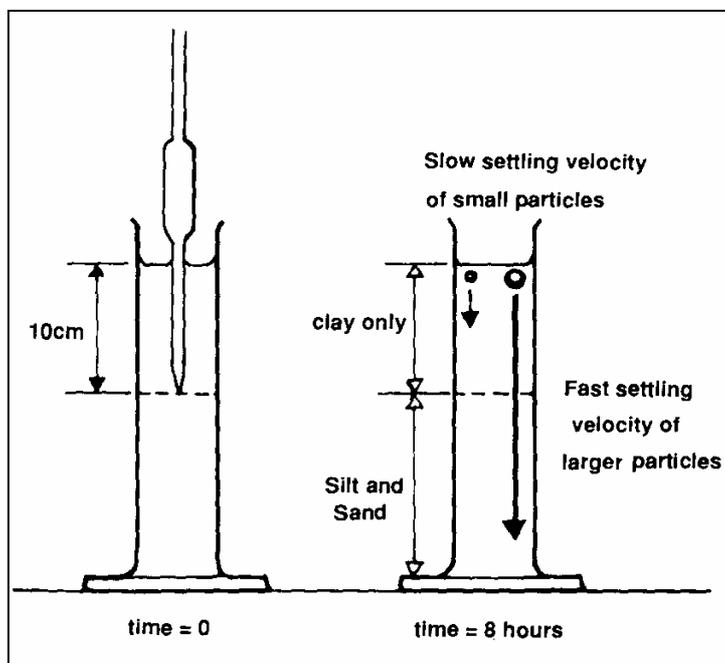
where  $p$ , is the density of the particle,  $p_w$  is the density of the water, and  $\eta$  is the viscosity of water.

This finer material can be measured in a number of ways. In the **pipette method** a tall cylinder is used to contain the suspension and a pipette inserted at a controlled depth (Figure 5.5). Samples of the suspension are taken at specific times following mixing. According to Stokes Law, after a certain time period has elapsed only particles which are smaller than a defined diameter may remain in suspension at the particular depth of sampling. All larger material must have settled past this depth due to their faster terminal velocities. For example, after a time period of eight hours only clay-sized particles can remain in suspension at the depth of the pipette shown in Figure 5.

Another method involves the use of a hydrometer to measure the density of the suspension at various times and thus give a measure of suspended sediment of each particular size. Automatic instruments are also available which measure the suspension density and give the particle-size distributions.

There are, however, some limitations to the application of Stokes Law to soil particles: (i) soil particles are not spherical and the diameter which is calculated is an **equivalent spherical diameter**, and (ii) all soil particles do not have the same density since that depends on their mineralogy. Despite these and some other limitations, allowing only approximate results, mechanical analysis is still regarded as an important means of characterizing soil.

**Figure 5.** Mechanical analysis. After mixing, the soil suspension is allowed to settle. Samples of the suspension are then taken after different time periods. In this example, after 8 hours only clay sized particles remain in suspension above the depth of sampling.



Mechanical analysis of volcanic ash soils is, however, fraught with difficulties. The proportion of sand- and siltsized particles can be measured but there are experimental problems in determining the clay content of soils with a large amount of amorphous or short-range order material.

#### 1.1.9.4 Specific surface area

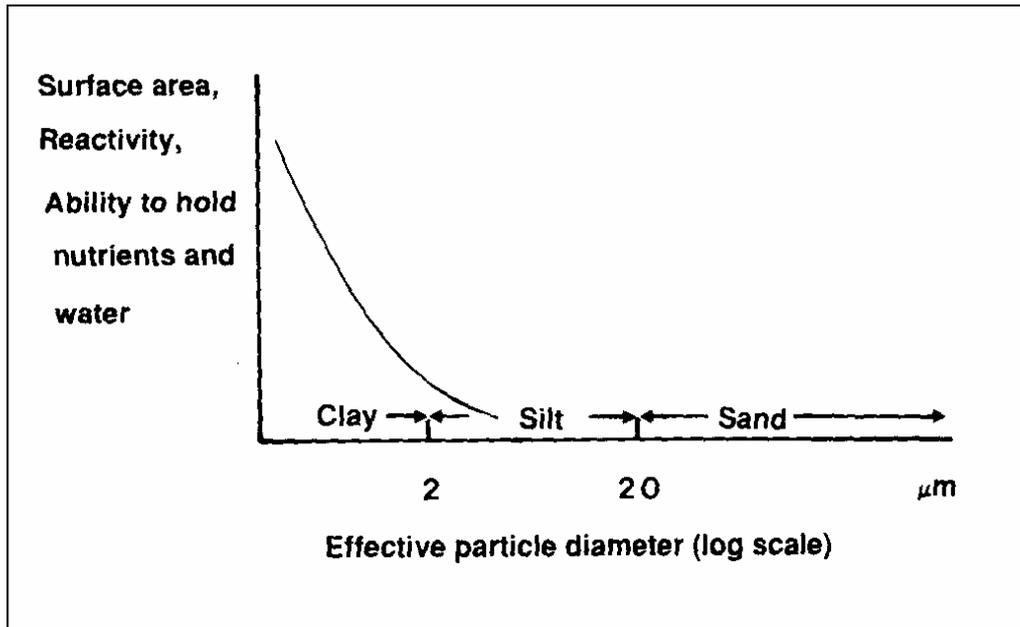
In general, the smaller the soil particle the more reactive it is and the greater its ability to retain nutrients and water (Figure 6). As will be described later, certain types of clay have the ability to hold on to large amounts of nutrients and this is partly due to the large surface area they have for adsorption. Probably a good measure of characterizing a soil would be to measure the amount of surface area which is present per unit mass or volume of soil. If, for example, a sand grain was broken down into clay-sized particles and the total area of all the surfaces produced were measured, it would exceed many times the original surface area of the sand grain. The surface area of particles per unit mass of soil ( $a_m$ ) is called the **specific surface area** and is defined by Equation 5.16. Surface area can also be calculated per unit volume ( $a_v$ ) by Equation 5.17.

$$a_m = \frac{A_s}{M_s} \quad (5.16)$$

$$a_v = \frac{A_s}{V_t} \quad (5.17)$$

where A, is the surface area, M, is the mass of solids, and V<sub>t</sub> is the total volume of soil.

**Figure 6.** General relationship between particle size and surface area or 'reactivity' of soil



Specific surface areas of soil material can be measured in the laboratory using various adsorption and desorption techniques and some examples of the specific surface area, of different soil components, are given in Table 2. Sand-sized particles have small specific surfaces of about 0.1 m<sup>2</sup> g<sup>-1</sup>, whilst certain types of clay can have specific surface areas as high as 800 M<sup>2</sup> g<sup>-1</sup>. It is thus the clay fraction which dominates the specific surface of most soils.

**Table 2.** Specific surface areas (adapted from Talibudeen, 1981; Warkentin and Maeda, 1980).

Particle size	Specific surface (m <sup>2</sup> g <sup>-1</sup> )
Coarse sand	0.01
Fine sand	0.1
Silt	1.0
Clay – kaolinite	10-20
illite	90-130
vermiculite	750-800
montmorillonite	750-800
allophane	500-700
Fe and Al hydrous oxides	25-42
Pumice (weathered)	600
Pumice (unweathered)	100

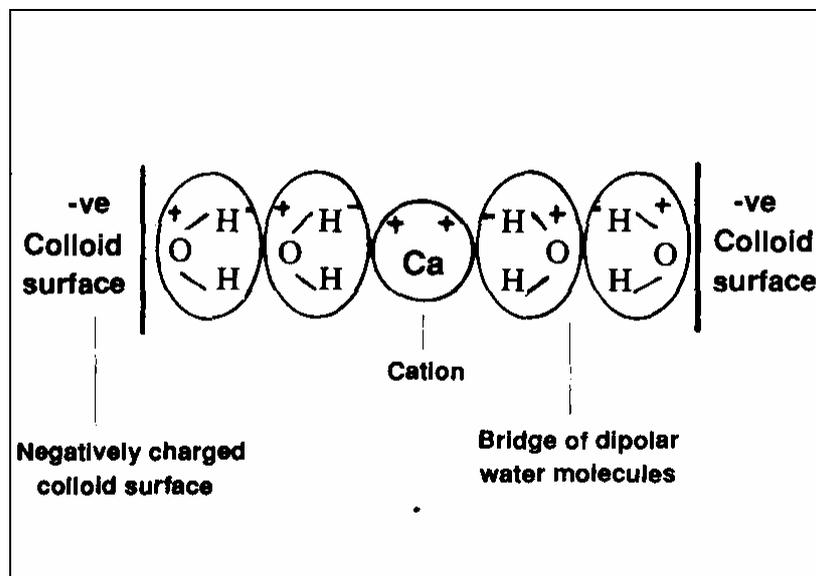
### 1.1.10 Soil structure and pore space

Soil structure has already been described in Chapter 1 and the types of structure forms found in soils shown. The shape and packing of these aggregates define the pore space within the soil. The pore system in turn is essential for drainage, aeration and root growth, as discussed later. In order that the pore system remains open, the structural aggregates must be stable and must not break down, blocking the soil pores.

#### 1.1.10.1 Structure formation and stability

In order to understand structure formation it is best to examine the soil at three different scales: (i) the atomic or molecular level, (ii) the microscopic level, and (iii) the macroscopic or visible scale.

At the molecular level, soil colloids (e.g. clay and organic matter) normally have a negative charge and thus attract cations and dipolar water molecules (see next Chapters). As shown in Figure 7, water molecules are able to link cations and clays together by orientating themselves in such a way that they act as bridges between the colloids. When the soil dries out the number of water molecules making up the bridge decreases and the clay surfaces are brought together. This enables **flocculation** of clay particles and is a first step in structure formation.



**Figure 7.** Water molecules and cations can both link soil colloidal material (clays) together to assist in flocculation. When the soil dries the bridge of water molecules becomes shorter and the colloid surfaces are brought closer together becoming strongly linked by the cation.

The type of cations which are present can have a marked effect on the extent of flocculation which can occur. Calcium, for instance, encourages rapid flocculation of colloidal particles whilst sodium does not. The reasons for this will be explained later, but it is partly due to the di-valent charge on the  $\text{Ca}^{2+}$  ion compared with the mono-valent charge on the  $\text{Na}^{+}$  ion. Indeed, a soil which has a high calcium content will generally have

a good structure, whilst a soil with a high sodium content (a saline soil) will have a poor structure.

The influence of calcium on soil structure is not, however, solely due to its effect on flocculation. Calcium also encourages the activity of soil organisms which, in turn, are responsible for assisting in structure formation and providing aggregate stability.

Organic molecules themselves are also important in bringing clay particles together. Long chain polymers, such as humic acid and polysaccharides, have flexible structures and are also able to come into close contact with the clay, linking particles together as shown in Figure 8. Aluminium oxides and hydroxides can perform a similar function in some soils, creating complex linkages between soil colloids and thus assisting in structure formation.

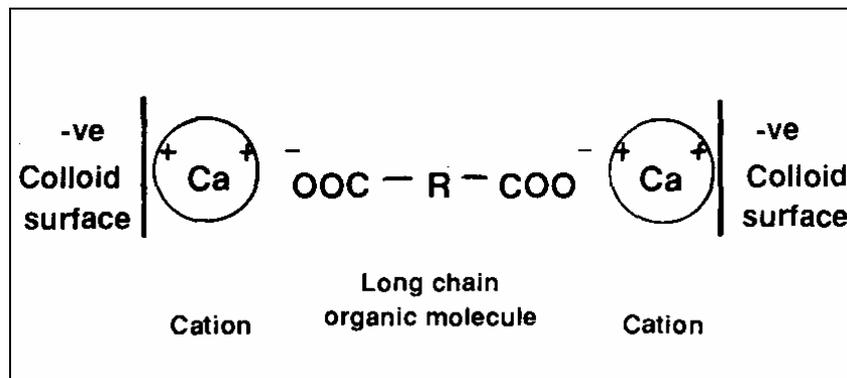


Figure 8. Organic molecules also assist in flocculation.

Sand and silt particles have no charge but at the microscopic level can be linked into **micro-aggregates** if they have a coating of clay or organic matter on their surface.

All of these processes should be thought of as acting together, as in the model presented in Figure 9. The clay and organic matter act as bridges between the sand and silt particles, producing micro-aggregates of soil. The clay is often present as **domains** in which the particles may be orientated to provide a series of plates which, although linking the sand together, also keeps them some distance apart. This is essential if the soil aggregates are to have micro-porosity.

At the macroscopic level the soil micro-aggregates are bound together into **macro-aggregates** by fungal hyphae, plant roots and other stabilizing agents (Figure 10). It is generally the form or shape of these macro-aggregates that is described by soil scientists since they are important for providing the soil with a system of pores for drainage, etc.

Soil aggregate formation thus requires the following:

(i) the presence of soil colloids, (ii) flocculation of these colloids, (iii) micro-aggregation, and (iv) macroaggregation. A number of factors are recognized as being important to structure formation:

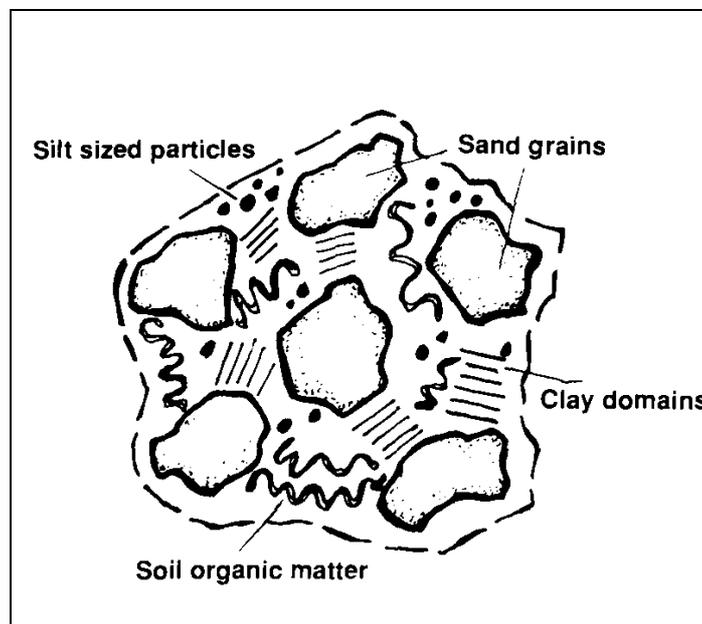
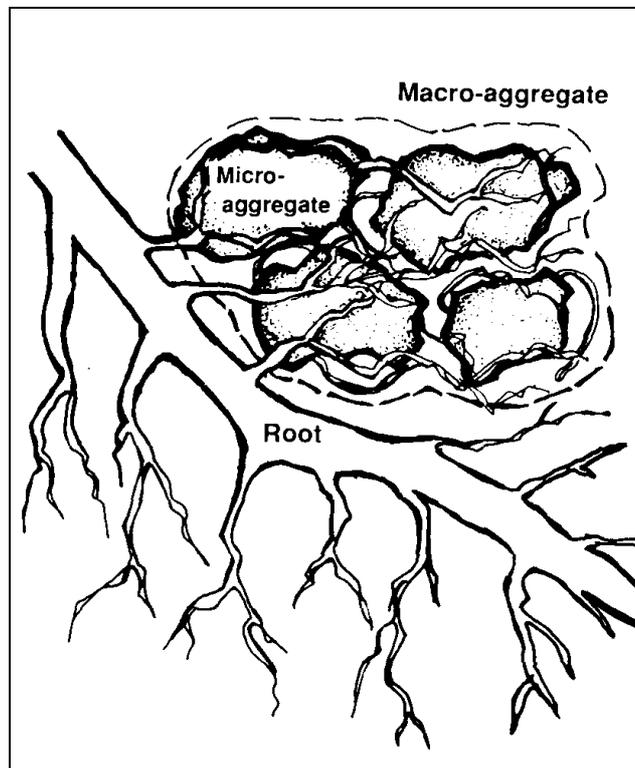


Figure 9. 'Inside' a soil aggregate (after Emerson, 1959).

**Wetting and drying cycles** - Wetting and drying cycles help to break down clods of soil and produce finer aggregates. Since the outside of a clod wets up first, the clay which is there swells quickly and this creates a differential pressure build up which cracks and disrupts the clod. In addition, entrapped air within the clod is compressed as water is absorbed inwards from the edges causing an explosive pressure build up and disruption of the clod (Figure 11). Drying of the soil as already described assists in the **aggregation** of the soil into microaggregates. It also causes closer approach and rearrangement of particles, promoting interparticle bonding.

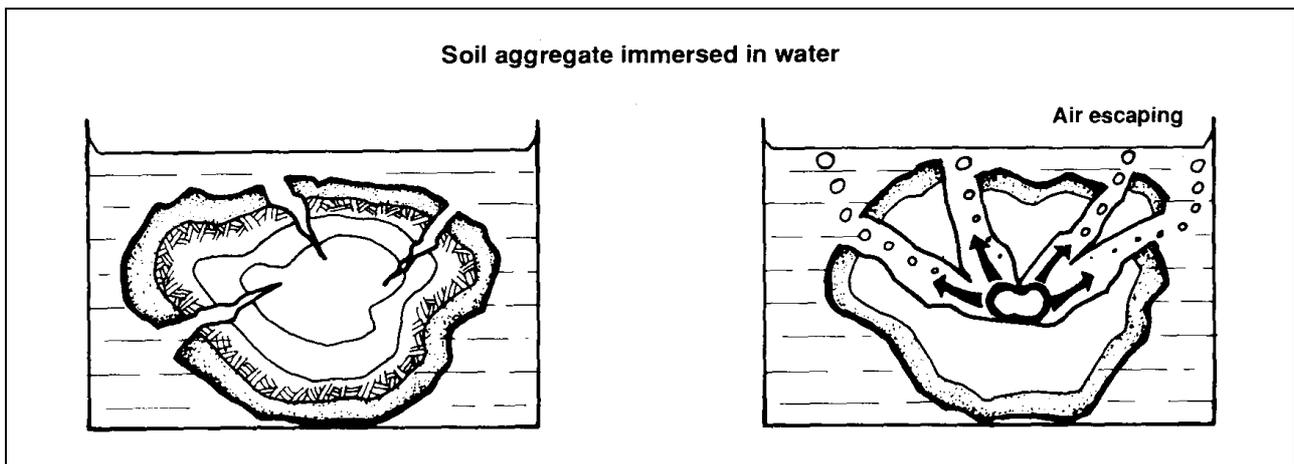
**Freezing and thawing cycles** - When water freezes to ice it expands and when this occurs within the restricted space of a soil pore it can cause a pressure build up which may disrupt a clod. Ice formation also attracts water from the soil immediately around the ice crystal. This effectively dries out the soil and helps stabilize the microaggregates. Farmers will often leave a cloddy surface over winter and allow the natural processes of freezing and thawing to produce a fine tilth by spring.

**Root growth** - As roots grow and expand in soil pores they create pressure on the surrounding soil. This together with plant water uptake helps to bind the soil into aggregates.



*Figure 10. Binding of micro-aggregates into macroaggregates. Roots, root hairs and fungal hyphae provide reinforcing to help hold micro-aggregates together to form macro-aggregates.*

In the plant root rhizosphere, microbial activity produces various organic glues or 'mucigels' and these also help stabilize the surrounding aggregates. **Organic matter and the activity of soil organisms** - Soil microbes which break down organic matter, such as dead plant roots and other residues, are essential to the production of the organic glues described previously. The addition of plant residues alone to a soil will do nothing to assist the stability of the aggregates if the soil microbes are unable to break them down. Fats and waxes produced in the break down give aggregates the ability to resist wetting and thread-like fungal mycelium help to bind the aggregates together. Different types of organic matter have been found to be responsible for providing different levels of stability. Humus provides the long term stability, polysaccharides and other glues provide stability for a few months, whilst fungus and other growths may provide stability for only a few weeks.



**Figure 11.** Aggregate breakdown following immersion in water. Rapid wetting causes differential swelling which produces cracks. Rapid wetting causes compression of air within the aggregate producing a small explosion.

Earthworms ingest soil and in the process bind particles together to produce very stable earthworm casts. Although these are most obvious on the soil surface the effects of earthworm activity within the whole of the soil is extremely important.

**Tillage** - Physical changes to the soil caused by tillage operations can be beneficial in producing a fine tilth. However, if cultivations are excessive or conducted when the soil is not at the correct moisture content then tillage can cause damage to the soil structure (see Chapter 9).

#### *1.1.10.2 Structure assessment*

In the field, the shape of soil aggregates is usually assessed by visual comparison with the aggregates main types (es. Prismatic, sub-angular blocky, granular, etc.). In the laboratory, aggregates are often examined under a microscope using a 'thin section' of soil. This technique (called **micromorphology**) allows examination of the sand grains and the clay domains, as well as the aggregates themselves (e.g. Plate 1).

The size of soil aggregates can either be measured directly using a scale, or by passing a sample of soil through a series of graduated sieves.

The stability of soil aggregates is more difficult to assess than aggregate size distribution and a number of tests have been developed. In general, these tests involve assessing the extent of aggregate breakdown following the application of an artificial force which is designed to simulate field conditions. A common test is that of wet **sieving**. This involves the gentle agitation of a known mass of soil on a sieve, or bank of sieves, immersed in a bath of water. After a short period of time the mass of soil

remaining on each sieve is measured and the proportion of stable aggregates calculated. Other tests which are used involve a visual assessment of the extent of aggregate breakdown which occurs: (i) over time following immersion of aggregates in a dish of water, or (ii) with bombardment of water drops from a standard height.

#### *1.1.10.3 Soil pore space*

The way in which the soil units (aggregates and primary particles) are arranged relative to each other defines the pore space in the soil. Pore space is essential as it provides the soil with the ability to store water and air, as well as allowing drainage and root growth to occur. The total pore space (porosity) was defined by Equation 5.4 and examples of porosity values for some New Zealand soils were given in Table 1.

The total porosity, however, is of less importance than the relative distribution of pore sizes within the soil. A sandy soil, for example, has a lower total porosity than a clay soil yet it usually drains faster than a clay soil. This is because soil consists of a range of pore sizes and it is only the larger pores which are effective in drainage. The sandy soil has a higher

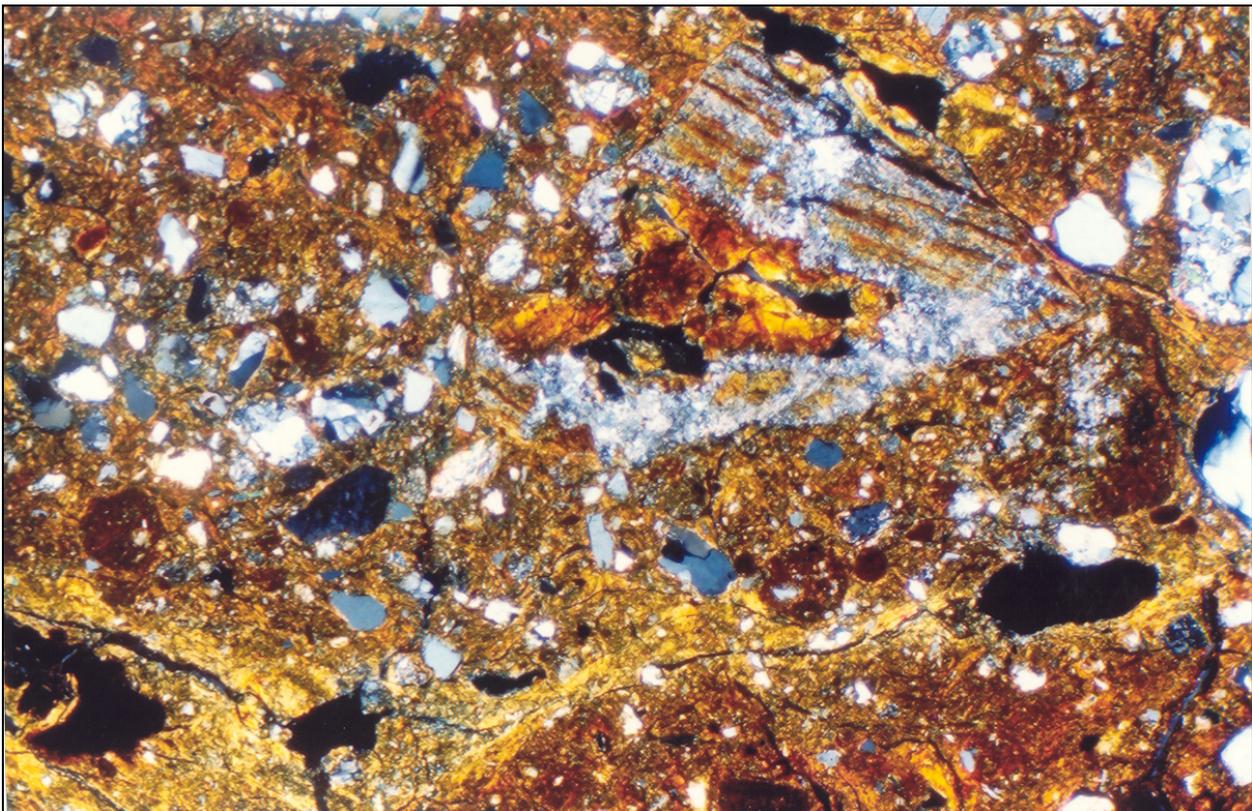
proportion of large soil pores than the clay and thus drains faster. Small pores are, however, essential to a soil too, as they provide the ability to store water. Indeed, the sandy soil is often prone to drought because of its lack of smaller pores. Water storage and drainage will be discussed in more detail in the next chapter, as will the method of measuring the pore-size distribution of a soil.

Pores can be classified according to size or function and these are presented in Table 3. The larger of the pores (**macropores**:  $>30\ \mu\text{m}$  diameter) are usually airfilled and do not contain water except when the soil is saturated or is draining. The smaller pores (**micropores**:  $<30\ \mu\text{m}$  diameter) include those which are responsible for storing water in the soil and are generally found within, rather than between, the soil aggregates.

Pores may be created by faunal activity, such as the burrowing of earthworms, and this is generally responsible for the creation of the macropores. Swelling clay soils often have large cracks in them and these are created when the clays shrink upon drying.

Cultivation usually results in an increase in the amount of macropores; however this often collapses over the season. Continuous cultivation can cause soil structure to break down, resulting in a loss of macroporosity, and this will be discussed further in Chapter 9.

*Plate 1. Micromorphology involves the examination of a 'thin-section' of soil under a microscope. This plate shows a layered coating of clay on the surfaces of a soil pore. The white irregular shapes are sand and silt particles within the soil matrix (Photo: R.Napoli)*



**Table 3.** Classification of pores according to size and function.

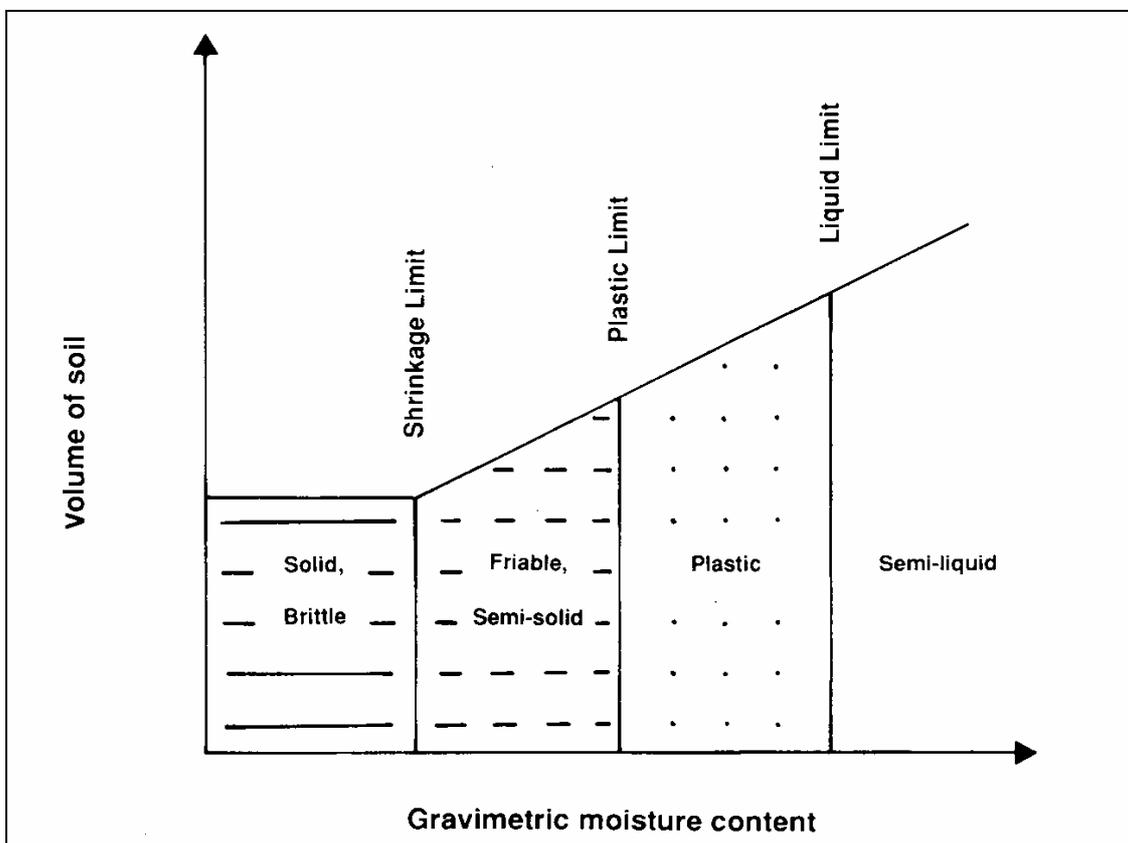
Pore diameter ( $\mu\text{m}$ ) <sup>*</sup>	Pore description	
>300	Air pores	
300-30	Transmission pores	<b>(macropores)</b>
30-0.2	Storage pores	
<0.2	Residual pores	<b>(micropores)</b>
* $1\mu\text{m} = 0.001 \text{ mm}$		

### 1.1.11 Soil consistency and soil strength

#### 1.1.11.1 Soil consistency

Soil consistency was described in pedological terms, and was stated to vary with the relative moisture content of the soil. Soil consistency is also used more precisely to describe the **mechanical** state of the soil, that is whether it behaves as a solid, a plastic, or a liquid. Consistency is a composite property of soil and describes how a soil behaves when acted upon by a force, or when it is put under stress. For example, when a soil is dry it is **hard** and resists the force of a tillage tool, such as a plough, which makes it difficult to cultivate. When moist the same soil becomes **friable** and is easy to cultivate. If the soil is cultivated when it is wet it may 'deform' rather than break up and in this state is called **plastic**. Very wet soil may start to flow when a force is applied to it, for example around the wheels of a vehicle which has become 'bogged', and in this state is termed **liquid**. These changes in state have been studied widely and are illustrated in Figure 12.

**Figure 12.** Atterberg limits in relation to moisture content and volume for a soil with shrink/swell properties (adapted from Hillel, 1980).



The Atterberg consistency limits referred to in Figure 12 are described as:

The **shrinkage limit** is the gravimetric moisture content (i.e. grams water per gram dry soil) at which a soil changes from being friable into being rigid or hard. This is measured by sequentially drying a pat of re-moulded finely ground (<0.4 mm) soil until no further shrinkage in the pat occurs. The **plastic limit** is the gravimetric moisture content at which a soil changes from being 'friable' into being 'plastic'. It therefore represents the minimum amount of water at which puddling will occur and the maximum amount of water that a soil can hold when it is friable. This plastic limit is also referred to as the 'lower plastic limit' and is assessed by measuring the water content of a sample of finely ground soil when it can just be rolled into a thread 3 mm in diameter without breaking. Although this test is usually conducted on a glass plate in the laboratory, some farmers test whether their soil is too wet to plough by rolling a thread of soil on their hand. If the thread does not break, then the soil is too wet to cultivate (i.e. it is in a plastic state and would be damaged).

The **liquid limit** is the gravimetric moisture content at which the soil changes from being plastic and starts to behave as a viscous liquid. The liquid limit is also referred to as the 'upper plastic limit' and is measured using a Casagrande apparatus or a drop cone penetrometer.

The **plasticity index** is defined as the difference between the liquid and plastic limits and is occasionally used to assist with the classification of soil properties, particularly in engineering.

In general, the liquid limit, plastic limit and plasticity index all increase as the amount of clay and organic matter in the soil increases. Examples of these consistency limits are given for a range of soils in Table 4. Volcanic ash soils which contain significant amounts of allophane have very high liquid and plastic limits, above the ranges given in Table 4.

**Table 4.** Soil consistency limits for a range of soils (adapted from Archer, 1975).

Soil texture	Clay content (%)	Shrinkage Limit (% w/w)	Plastic Limit (% w/w)	Liquid Limit (% w/w)	Plasticity index
Clay	51	13	36	83	47
Sandy clay loam	23	18	25	40	15
Sandy loam	12	14	16	21	5

#### 5.4.2 Soil strength

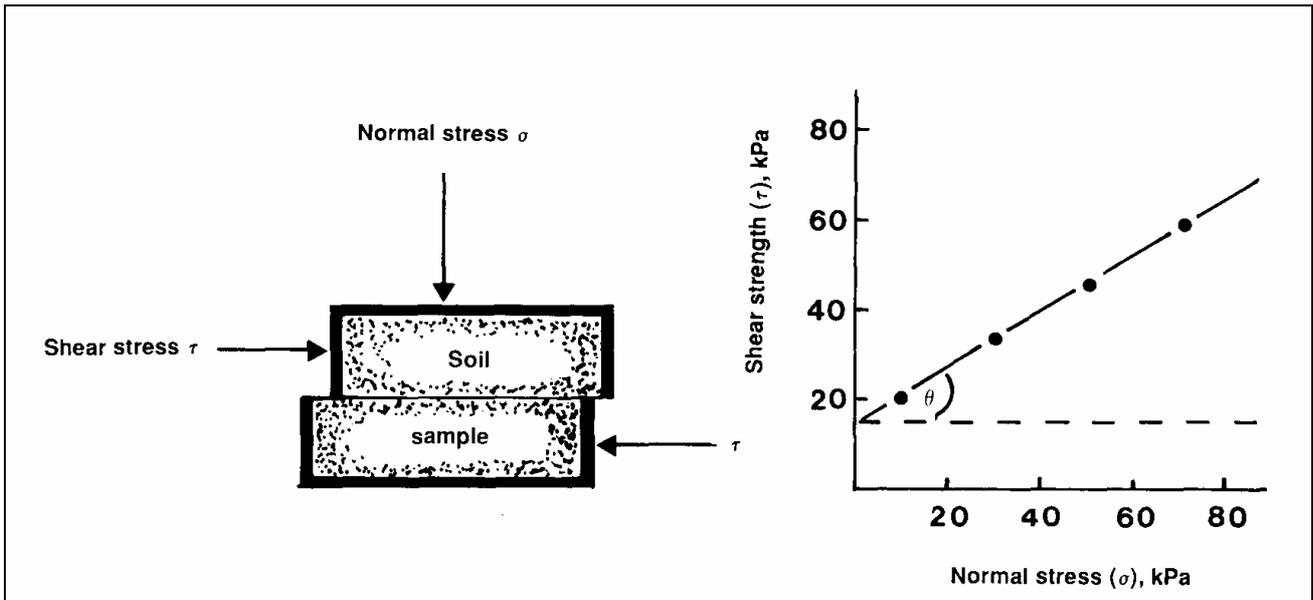
Soil strength is defined as the ability of soil to resist a force without **shearing** (or flowing if wet). The shear strength of a soil is derived from a combination of its **cohesive** strength (i.e. bonding between particles) and its **internal friction** (i.e. the friction which occurs when particle surfaces are forced to slide over each other). Wet clay soils, for example, have good cohesion but only a low internal friction. Pure dry sand, at the other extreme, has no cohesion but derives its strength from internal friction. Most agricultural soils lie somewhere between these two extremes and therefore have both cohesion plus internal friction. Volcanic ash soils have lower cohesion than soils with predominantly crystalline clay and following cultivation this can increase their susceptibility to erosion.

The shear strength of a soil can be described by the **Móhr-Coulomb equation**.

$$\tau = C + \sigma \tan \phi \quad (5.18)$$

where  $\tau$  is the shear strength required for failure to occur,  $\sigma$  is the stress normal to the shear plane,  $\tan\phi$  is the coefficient of internal friction, and  $C$  is the shear stress required to cause failure at zero normal load (often called **soil cohesion**).

Figure 13. Soil strength measurement and results.



Various methods have been developed to measure soil strength and although the **direct shear method** is illustrated in Figure 13 because of its simplicity, other laboratory methods such as the **triaxial shear test** are more reliable.

Field measurement of soil strength can be made using an instrument called a **vane-shear tester** which is sketched in Figure 14. The vanes of the instrument are pushed gently into the required depth in the soil and then the handle is turned. The torque at which the soil shears (i.e. when the blades turn) is then taken as a measure of soil strength.

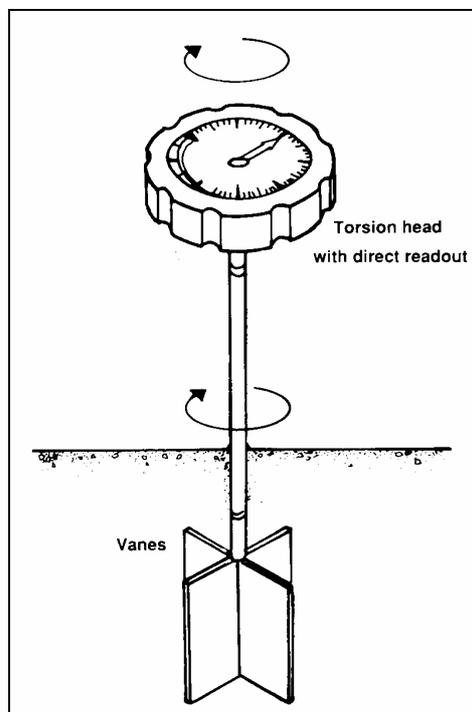
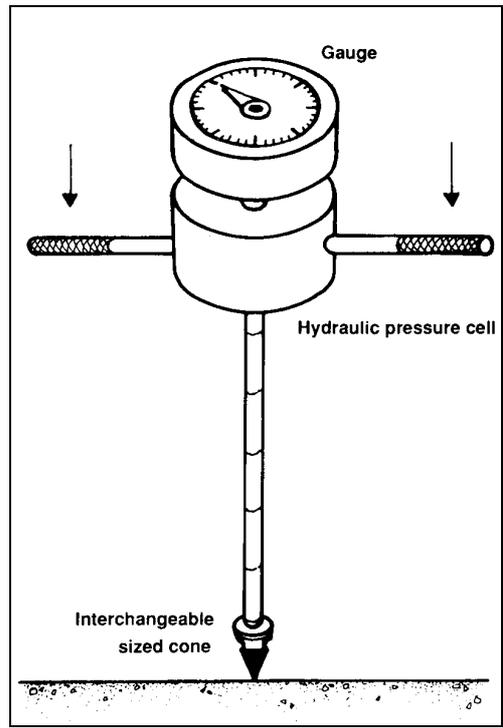


Figure 14. Vane shear tester for measuring soil strength.

Soil strength can also be assessed indirectly by an instrument called a **penetrometer** (Figure 15). The force required to push the tip of the penetrometer into the soil is recorded. Although this measure also includes forces of compression in front of the probe, and friction between the probe

and the soil, the penetrometer is a useful instrument for estimating resistance of soil to root penetration, traffic loading and to tillage. It should be remembered that shear strength and penetration resistance are also strongly dependent on the water content of the soil.



**Figure 15.** Penetrometer for measuring soil strength.

## 2 SECTION V: The storage of water in soil

### 2.1 Introduction to the hydrological cycle

Water is added to the soil as precipitation (i.e. rainfall, hail, snow, dew) or as irrigation. Rainfall in Italy varies tremendously depending on location and altitude. It can be as high as 2000 mm/ year in the Alps, or as low as 300 mm/ year in Southern mediterranean regions.

When water is applied to a dry soil, a certain amount will be absorbed or stored before drainage starts. If the storage capacity and drainage rate of the soil is exceeded, then water will either pond on the surface, or run off downslope.

Water vapour may be lost to the atmosphere either through direct evaporation from the soil surface or through **evaporation** of water from plant leaves, which is called **transpiration**. The combined loss to the atmosphere from plant and soil is termed **evapotranspiration**.

These major gains and losses of water to the **hydrological cycle** are shown in Figure 16. If the amount of rainfall exceeds both the soil storage capacity and evapotranspiration loss, then drainage will occur. Alternatively, if evapotranspiration loss exceeds rainfall, then either the soil must supply water from its store, or irrigation must be applied in order to maintain full plant production.

### 2.2 Physical properties of water

Water is essential to life, and an understanding of its basic properties is essential to the study of soils and plants. Despite being such a common substance, it is, in fact, unusual. According to its chemical formula ( $H_2O$ ) alone, it should not be a liquid at normal temperatures, but a gas. This and many of the other unusual properties of water can be explained by examining the structure and bonding of water molecules.

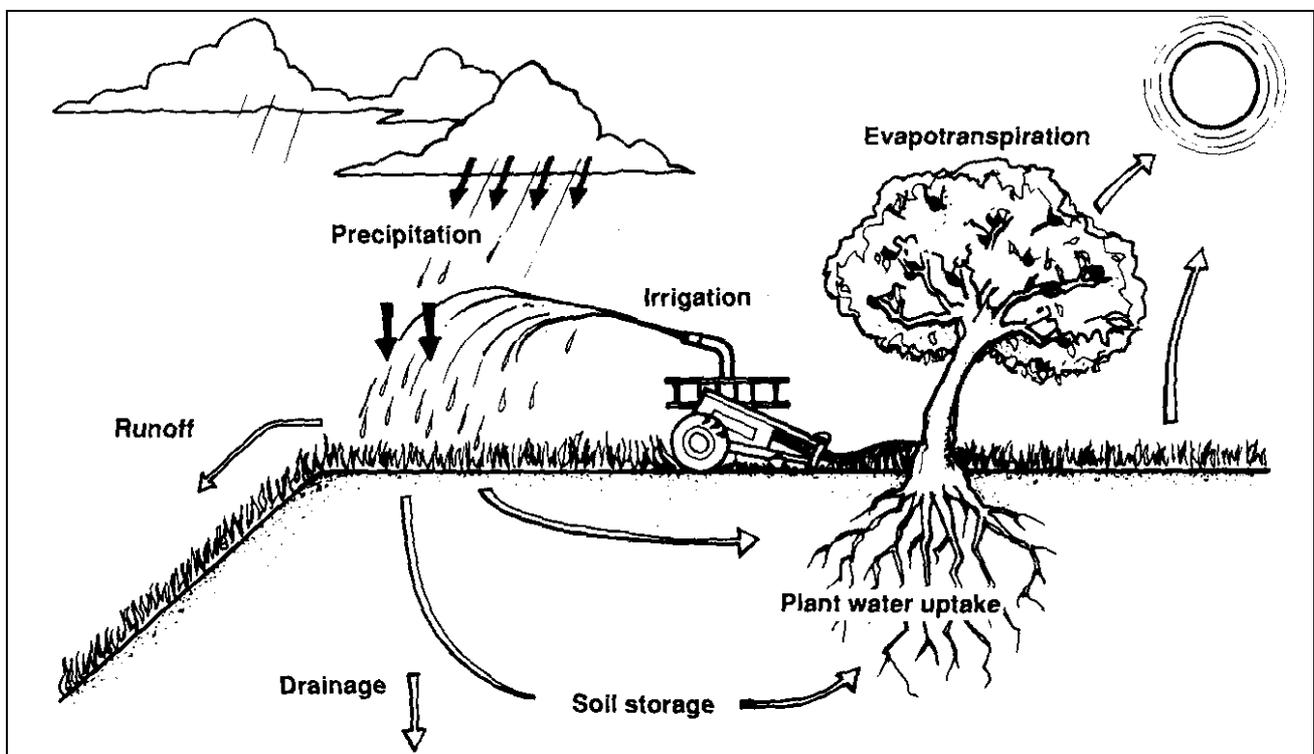
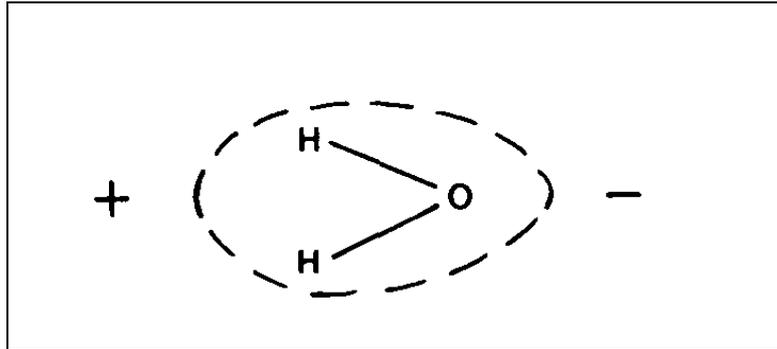


Figure 16. The Hydrologic Cycle in agricultural systems.

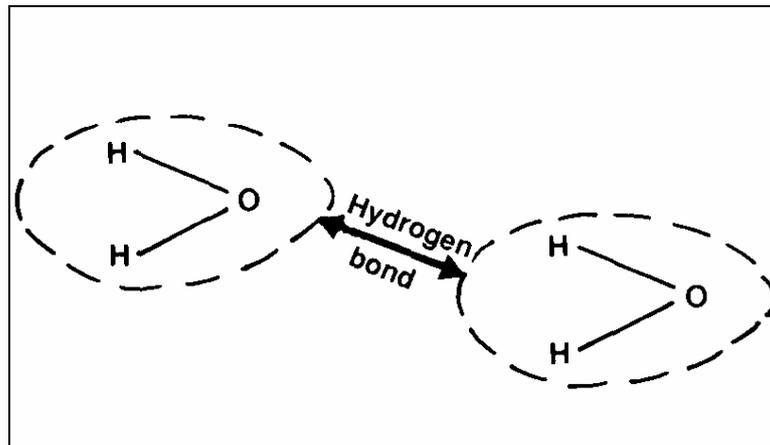
### 2.2.1 Molecular structure, polarity and hydrogen bonding

Water molecules have an asymmetrical arrangement of hydrogen and oxygen atoms, as shown in Figure 17. The H-O-H bond is set at an angle of  $105^\circ$  and this means that the two hydrogen atoms are located at one side of the molecule, with the oxygen atom at the other. Water molecules are electrically neutral but because of the relatively large attractive force (electronegativity) of the oxygen atom, the single electrons of both the hydrogen atoms are displaced towards it. This gives the side of the molecule with the hydrogen ions a slight positive charge and the side with the oxygen atom a slight negative charge (Figure 17). As a result water molecules are said to be **polar**.

The polar nature of water molecules causes them to orientate themselves in the liquid or solid in such a way that the oxygen atom of one molecule is closest to the hydrogen atom of another (Figure 18). This produces a linkage between the individual molecules called **hydrogen bonding**.



*Figure 17. Asymmetrical water molecule. Positive charge at the hydrogen end and negative charge at the oxygen end.*



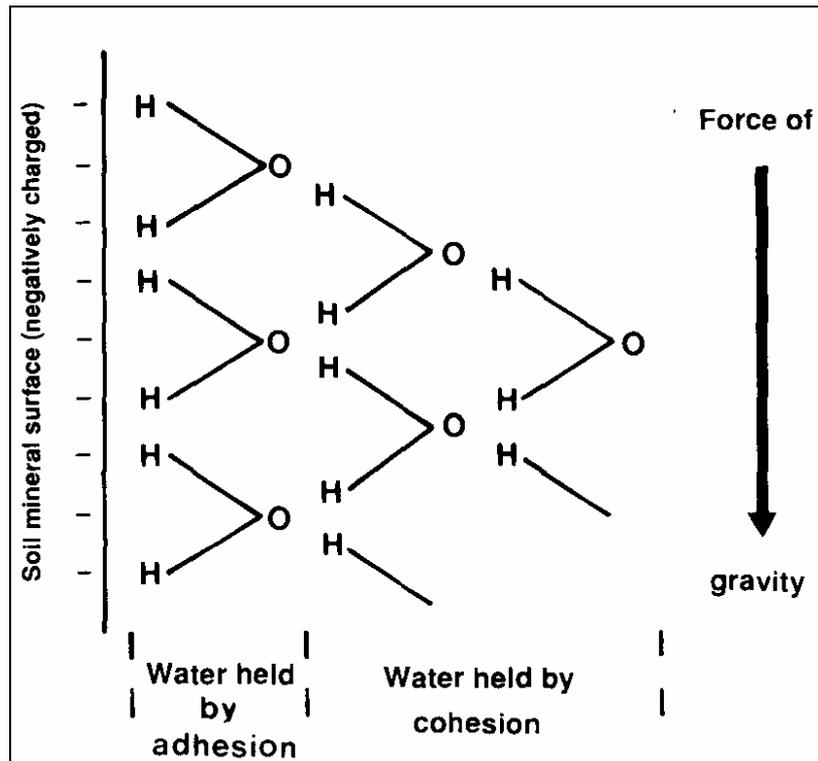
*Figure 18. Hydrogen bonding creates an intermolecular link between water molecules.*

Each hydrogen atom thus shares in two bonds, one covalently with the oxygen of its own molecule and one through dipole attraction to the oxygen of an adjacent molecule.

In ice, hydrogen bonding holds the water molecules together in a rigid tetrahedral structure. In liquid water, this structure is less apparent but hydrogen bonding is still responsible for holding the molecules together and keeping water as a liquid at normal temperatures. The heat required to boil water and turn it into a gas is used to break these hydrogen bonds, destroying the intermolecular structure and producing separate water molecules.

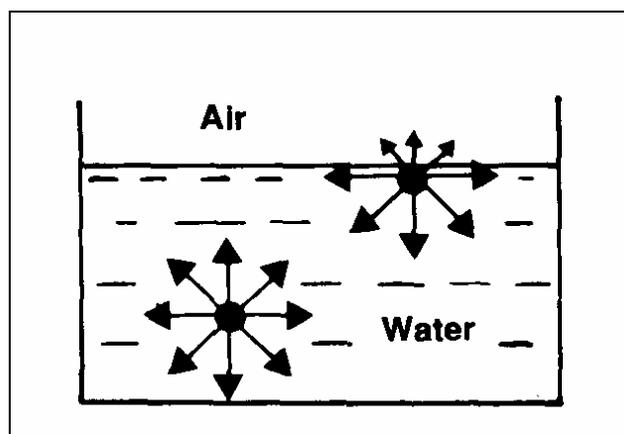
### 2.2.2 Adhesion, cohesion and surface tension

The polar nature of water results in the attraction of water molecules for each other (called **cohesion**) and the attraction of water for other surfaces, such as glass or clay (called **adhesion**). Adhesion and cohesion are important because, amongst other things, they provide the soil with the ability to store water. Figure 19 shows how some water molecules are held by adhesion on to the surface of a soil mineral whilst others are held by cohesion between the water molecules.



**Figure 19.** Adhesion and cohesion hold water molecules against mineral surfaces.

Within a body of water the cohesive forces acting on each molecule are equal in all directions, as indicated by the arrows, or vectors, in Figure 20. However, at the surface, the water molecules have a greater attraction from the other molecules within the liquid than from the air above. This results in the water molecules close to the surface having a higher energy than those in the bulk body of water. To minimize the energy, the surface area is minimized. It is as if the surface area were covered with a 'skin' with a tension force minimizing the area.



**Figure 20.** Cohesion causes an equal attraction of water molecules for each other within the bulk liquid. At the surface, the water molecule is attracted with greater force towards other water molecules than towards the air.

### 2.2.3 Capillarity

If a narrow (capillary) tube is dipped into the water bath represented in Figure 21, then water will move up into the tube, above the height of the free water surface. This phenomenon, called **capillarity**, is due to the adhesive attraction of water for the solid surface of the tube and the cohesive attraction of water molecules for each other.

The shape of the meniscus within the capillary tube in Figure 21 suggests that the pressure of the

water in the tube ( $P_w$ ) is less than the pressure of the atmosphere ( $P_a$ ) and is less than the pressure of the free water outside the tube. The relatively higher pressure of the water outside the tube (equivalent to  $P_a$ ), compared to that below the meniscus ( $P_w$ ), causes water to be forced up into the capillary tube. This upward movement continues until the weight of water inside the tube provides a sufficient force to balance the pressure difference between  $P_a$  and  $P_w$ . The smaller the internal radius of the tube the greater the height of rise, and this may be calculated by the following equation:

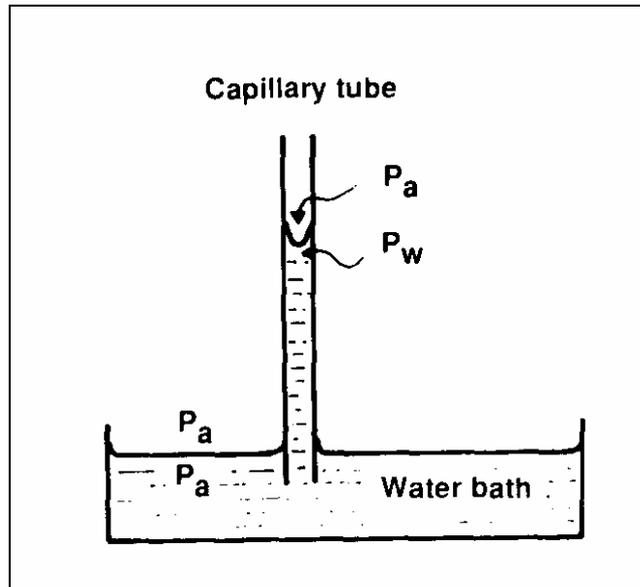
$$h = \frac{2\gamma}{r\rho_w g} \quad (6.1)$$

where  $h$  is the height of capillary rise in the tube,  $\gamma$  is the surface tension,  $r$  is the radius of the tube,  $\rho_w$  is the density of water, and  $g$  is the force of gravity. For water this equation approximates to:

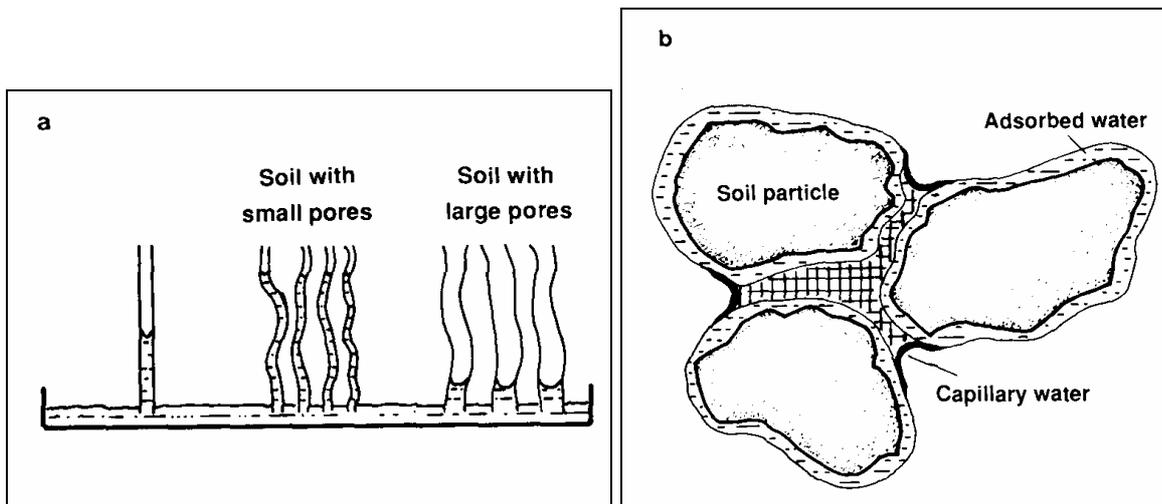
$$h = \frac{0.15}{r} \quad (6.2)$$

where  $h$  and  $r$  are both in cm.

Because there is a negative pressure under the meniscus (i.e.  $P_w < P_a$ ) the water there is said to be under **suction** or **tension**.



**Figure 21.** The attraction of the capillary tube walls for water causes the pressure in the water below the meniscus ( $P_w$ ) to be less than the pressure in the water bath ( $P_a$ ). Water is therefore drawn up into the capillary tube.



**Figure 22.** Water is held in soil pores and on the surfaces of soil material: (a) The smaller the pore diameter the greater the rise due to capillarity; (b) Adsorbed water on the surfaces of colloidal material with capillary water in the pore spaces between material.

The principle of capillarity explains why soil is able to hold water against the force of gravity. Although the pores which hold the water are not as symmetrical as in the capillary tube model, the height of rise and the suction with which the water is held is still greater in a soil with predominantly small pores compared with a soil consisting mainly of large pores (Figure 22). In a saturated soil free water can be present in the macropores but in an unsaturated soil these are empty and the water which remains in the smaller pores is held under suction.

## 2.3 Soil water content and potential

### 2.3.1 Soil water content

The water content of soil is important not only because it affects plant growth but also because it influences soil properties such as aeration, temperature, consistence, and others which will be described later. Soil water content can be measured in a number of different ways and some of these will be described here. **Gravimetric method** - This is the most common method of measuring soil water content. It involves collecting a sample of soil from the field and weighing it to obtain the mass of wet soil. The sample is then dried in an oven at 105°C and the mass of dry soil recorded. The basis of expression of water content ( $\theta_m$  has already been given in Equation 5.9 and the method of calculation is:

$$\theta_m = \frac{M_w}{M_s} = \frac{(\text{mass of wet soil} - \text{mass of dry soil})}{\text{mass of dry soil}} \quad (6.3)$$

where  $M_w$  is the mass of water and  $M_s$  is the mass of solids. The units used are  $\text{kg kg}^{-1}$  or  $\text{g g}^{-1}$ , but it is also often expressed as a percentage.

One problem with this simple method of measurement is that the soil profile has to be destructively sampled and this makes it difficult to monitor changes in water content at any one location.

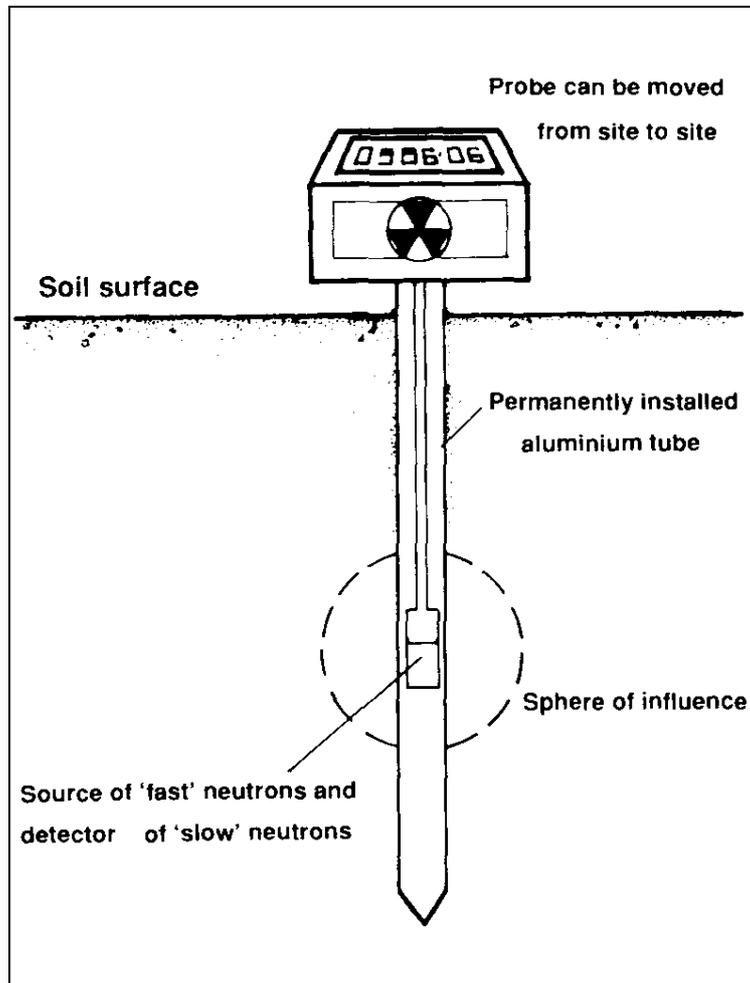
**Neutron scattering method** - This method allows easy monitoring of changes in profile water content and has been used both in research and in routine irrigation scheduling. Water content is measured on a volumetric basis, as expressed in Equation 5.10, and therefore can provide a direct indication of the amount of water which may be available to plants in a given volume of soil.

A neutron probe, consisting of a radioactive source of "fast neutrons" and a detector of "slow neutrons", is lowered down an access tube permanently installed in the soil (see Figure 23). When fast neutrons are emitted into soil they collide with small atoms, such as hydrogen, which cause them to be 'slowed' and scattered back towards the probe. The flux of slow neutrons which returns to the probe is measured by the detector. In soil, it is mainly the hydrogen atoms contained in water molecules which are responsible for slowing the neutrons and therefore the flux is a function

of the water content.

The volume of soil measured at each depth, called the 'sphere of influence', can be as large as 200 mm radius. This creates a problem in measuring water content in the surface soil, as some neutrons can escape through the surface, giving a low reading. However, the large volume of measurement does give the method an advantage in scheduling irrigation because it is generally more representative of the field soil than small cores taken for gravimetric analysis.

Other methods of measuring *in situ* water content, such as **time domain reflectometry** and **gamma ray attenuation**, are also available and the reader is referred to the texts listed at the end of this chapter for details of these methods.



**Figure 23.** Neutron moisture probe for measuring soil water content.

### 2.3.2 Soil water potential

In order to understand the complexities of water in the soil-plant-atmosphere system, scientists have adopted a concept of **potential energy** being associated with the water in the system. The basis of this concept may be appreciated by considering the following example.

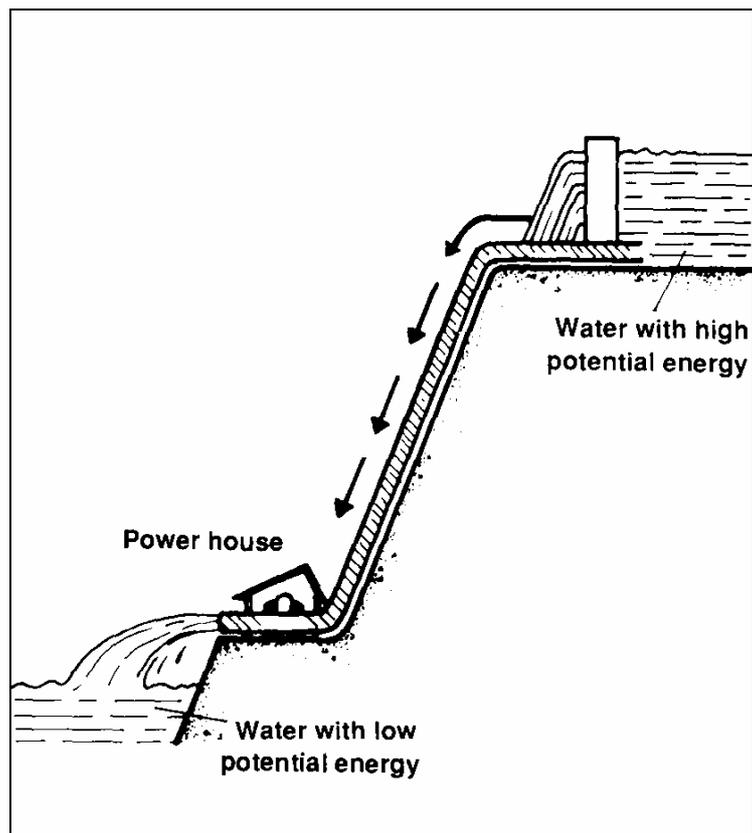
Water stored in the hydro-electric dam represented in Figure 24 has potential energy because of its relatively high position compared with the elevation of the powerhouse below. The water in the dam has the 'ability to do work' and this potential energy eventually results in the turbines in the power-house being driven to produce electricity.

Water in the soil also has potential energy, though obviously in smaller amounts, and like the water in the dam flows from a position of high potential to one of low potential. The word **potential** is used as an abbreviation for 'potential energy' in soil physics. When water moves from one equilibrium system to another there is a change in the energy level of each system.

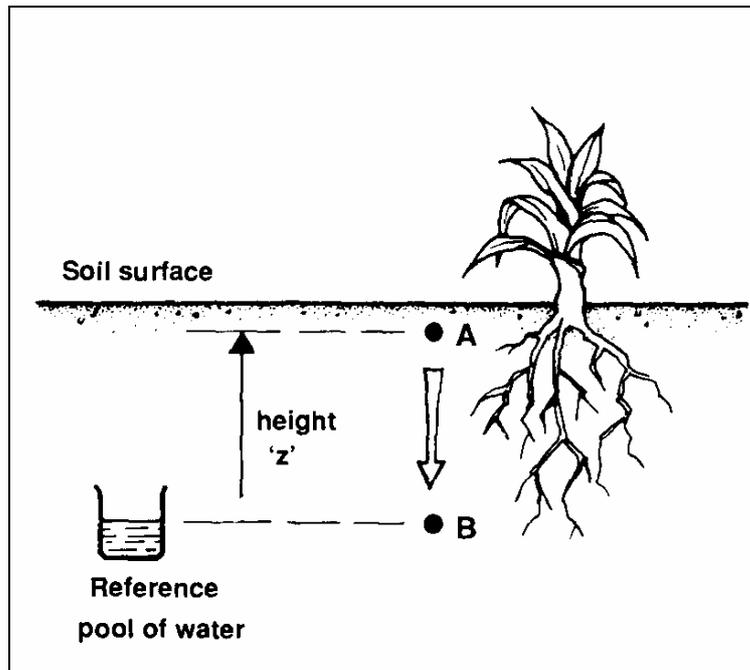
This potential energy approach to understanding soil water provides a good basis for examining

most aspects of water in soil. It provides a useful tool to study water storage, water movement, and water uptake by plants. The fundamental concept to remember is that water moves from a zone of high potential to a zone of lower potential.

In our example of water in a hydro-dam, it is not the absolute height of the water above sea level that is important, but the relative height of the water above the power-house which defines its potential energy. Similarly for soil water it is not an absolute amount of energy which is important but a relative amount. The potential energy stored by soil water is expressed relative to a hypothetical body of water at atmospheric pressure and at the same temperature, and at a given reference elevation. For instance, in Figure 25 the reference body of water is at the base of the soil profile. Water will move readily from level A, at the soil surface, to level B, at the same level as the reference, because it is moving from a high potential to a lower potential.



**Figure 24.** Water held in the dam has a high potential energy. Water flows from a position of high potential to one of low potential energy.



**Figure 25.** Water moves downward from point A to point B because water at A has a higher potential than the reference pool of water at the same elevation as point B. Water always moves from a position of high potential to one of lower potential.

In unsaturated soil (Figure 26) water can, however, move from level B to a higher level, C, if point C has a lower water potential than point B. Obviously this would not be the case if the potential was dependent only on the relative height (or gravity force). Since water does move upwards and laterally (e.g. towards plant roots) then the potential must be dependent on other forces as well as gravity. The essential components of the **total soil water potential** will be detailed first before describing the methods of measurement or units used. **Total soil water potential ( $\psi_t$ )** - The total soil potential is due to the sum of all the forces acting on the soil water :

$$\psi_t = \psi_g + \psi_m + \psi_p + \psi_s + \dots \quad (6.4)$$

where  $\psi_t$  is the total potential,  $\psi_g$  is the gravitational potential,  $\psi_m$  is the matric potential,  $\psi_p$  is the pressure potential,  $\psi_s$  is the solute potential, and the dots indicate other potentials which are mostly insignificant and can be ignored for our purposes.

**The gravitational potential ( $\psi_g$ )** - The gravitational potential is defined as:

$$\psi_g = \rho_w g z \quad (6.5)$$

where  $\rho_w$  is the density of water,  $g$  is the acceleration due to gravity, and  $z$  is the height above the reference body of water (e.g. see Figure 25).

Informally then, the gravitational potential is the amount of work that can be done by water moving from a location at depth  $z$  to a body of identical water at a reference level. If the reference level is taken to be below the soil surface, as in Figure 25, then any free water located above that level will have a positive potential. In other words, it can do work in moving to the reference position. If the water is below the reference level then work must be performed in moving it up to the reference level in order to overcome gravity, and the gravitational potential is negative.

**The matric potential ( $\psi_m$ )** - The matric potential results from the forces of adhesion and capillarity which occur in soil. These forces are due to the interaction of soil water with the solid soil particles

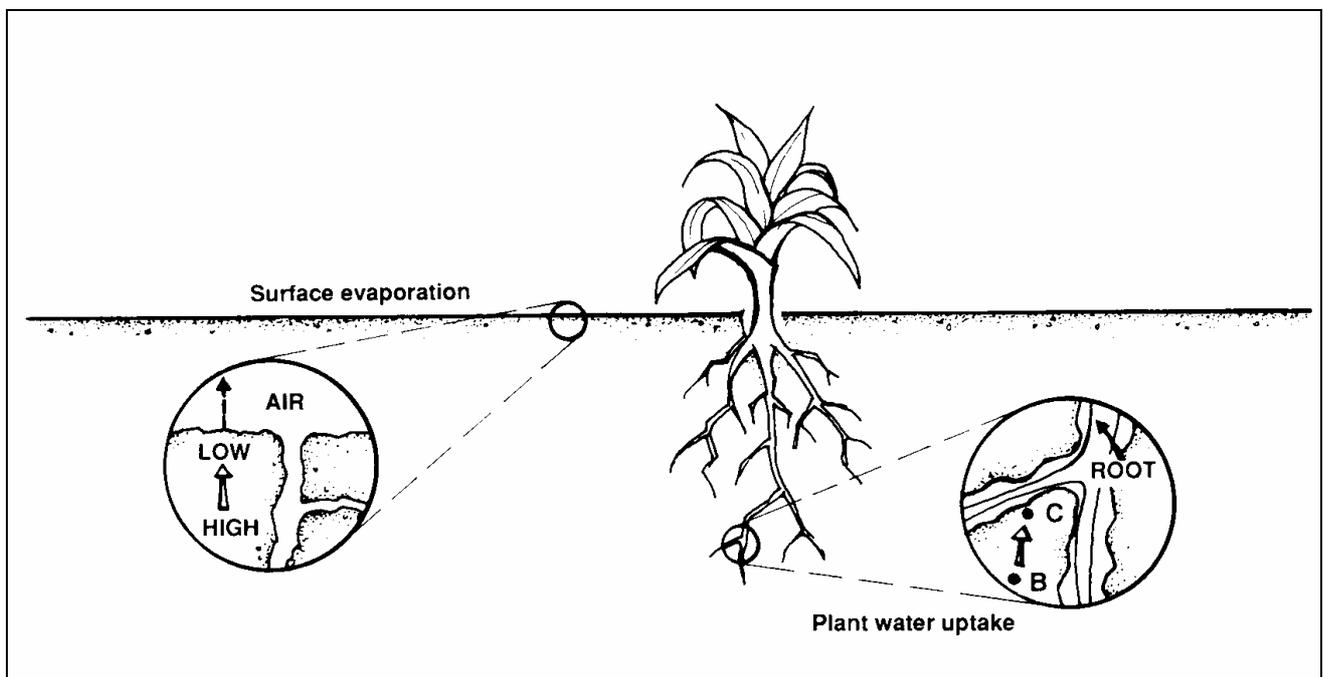
(the soil matrix), as described previously.

Because water is held strongly by adsorption on to soil particles and by capillarity in soil pores it is understandable that work would need to be done to remove this water. Water which is surface-adsorbed or held by capillarity thus has a lower potential than that in a reference body of free water. This means that the matric potential is always negative.

Free water will be quickly adsorbed by a dry soil due to the lower (negative) matric potential of the dry soil. Similarly water will move from a wet zone in the soil towards a drier zone because it is moving from a zone of high matric potential to a zone of lower (more negative) matric potential.

This is illustrated in Figure 26 which shows water movement from a zone of moist soil towards zones of lower matric potential associated with surface evaporation and plant water uptake.

The soil matric potential is important as it is this which frequently must be overcome by plants in order to be able to obtain water for growth. When water is removed from the soil by plant uptake, the water which remains is at an even lower potential and this makes it more difficult for further uptake by the plant.



**Figure 26.** Water moves from wet zones within soil aggregates (high potential energy) to drier zones (lower potential) produced by surface evaporation or plant water uptake.

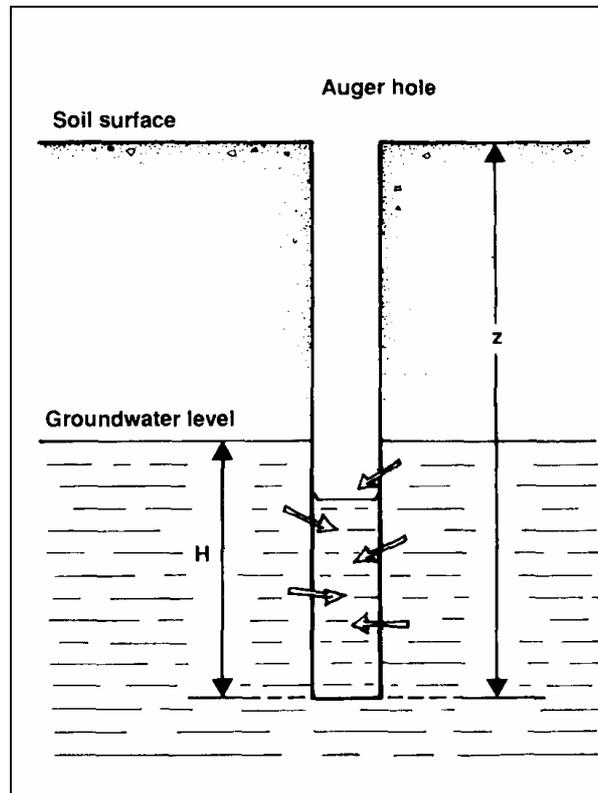
**The pressure potential ( $\Psi_p$ )** - A pressure potential in a field soil is always positive and develops as a result of a **hydraulic head**. For example, at any given point below the water table of a saturated soil, there is a pressure potential which is represented by the size of the hydraulic head (i.e. the depth below the ground water level in Figure 27).

The pressure potential does not include any effects of the soil matrix and in an unsaturated field soil the pressure potential is always zero. (Soil matric potential can however be expressed as a negative pressure potential, but this will be ignored for simplicity).

**The solute (osmotic) potential ( $\Psi_s$ )** - The solute, or osmotic, potential results from the effect of solutes in the soil solution. These solute ions or molecules attract water by hydration and this reduces the potential energy of the hydrated water molecules.

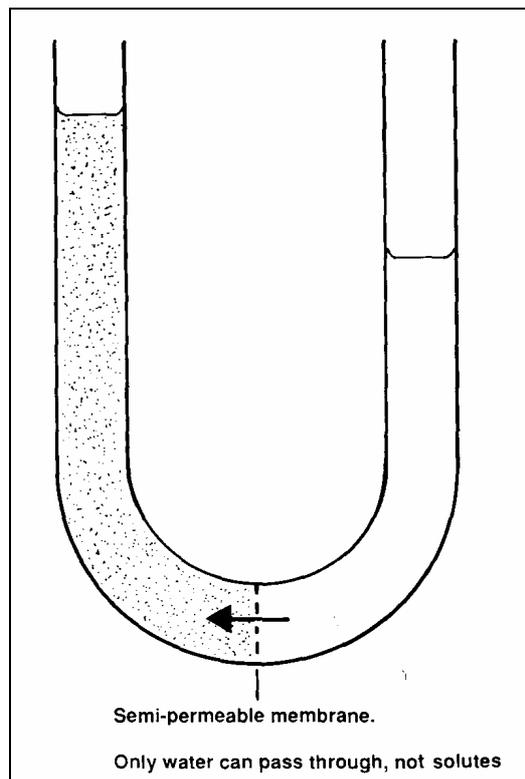
To understand this better it is useful to consider the process of 'osmosis' which is illustrated in Figure 28. The membrane between the pure water and the solution is 'semi-permeable', that is, it allows water molecules through but not solutes. Water molecules are able to move in either direction. However, over time more water molecules will move from the pure water to the solution,

than the reverse. This is due to the osmotic 'suction' of the solution and is a result of the reduction in energy of the water associated with the solute molecules. Since the solute potential of the pure water in Figure 28 is zero and water molecules must be moving from it to a zone of lower potential (i.e. across the membrane) then the osmotic potential must be negative. Similarly in a soil, water would move from a reference body of pure water into the soil solution. However, for the reverse to occur, work would have to be done to overcome the osmotic potential. The osmotic potential in soil is therefore always negative. The osmotic potential does not affect water movement in soil to any great extent but is extremely important to plant water uptake.



**Figure 27.** The pressure potential is represented by the hydraulic head ( $H$ ) (i.e. the depth below the water table).

An example of the extreme effects of soil osmotic potential on plant growth is found in saline soils. The high salt concentrations in the soil solution (low potential) mean that the plant cannot easily remove water from the soil. If the salt concentration in the soil is extremely high the soil water potential may be so low that the plant may have a relatively higher osmotic potential. This can result in the transfer of water from the plant into the soil (i.e. from high to low potential), resulting in the subsequent death of the plant.



**Figure 28.** Osmotic attraction of water through the semipermeable membrane causes there to be a larger amount of water in the left arm of the manometer. This attraction of solutes for water lowers the potential energy status of the water which causes it to flow from right to left.

**Units of soil water potential** - In the previous description of soil water potential we purposely avoided the use of units as it is important for the reader to understand the concepts first, before examining the units of measurement which are used.

Since soil matric and osmotic potentials are always negative they are often considered as 'suctions' or 'tensions'. Suctions and tensions are however always expressed as positive values (see Table 5).

As soil physics has evolved, different units of measurement have been used. It is suggested that Table 6.1 be examined closely and referred to whenever necessary, as it will help the reader to compare units.

Three ways of expressing soil water potential (energy) are used:

(i) **Mass basis.** Potential per unit mass is expressed as joules per kilogram ( $\text{J kg}^{-1}$ ).

(ii) **Volume basis.** The osmotic potential and matric potential are often expressed on a volume basis as joules per cubic metre ( $\text{J m}^{-3}$ ), or, since this has the dimensions of pressure, as newtons per square metre ( $\text{N m}^{-2}$ ) or pascals (Pa). The latter is the preferred unit but as a pascal is a small unit, kilopascals (kPa:  $1 \times 10^3$  Pa) are often used. Potential per unit volume is also given in bars, which is another pressure unit. One bar is equivalent to 100 kPa or approximately one atmosphere, which is the average air pressure at sea level. (iii) **Weight basis.** It has often been considered convenient to express water potential on a weight basis, in terms of the pressure created by a hydraulic head. A pressure of one atmosphere is equivalent to a column of water 10.2 m high.

Expressing potential in terms of hydraulic head is frequently used for the osmotic or matric potential since the method of measurement often involves the suction created by a column of water, as will be described in the next section.

**Table 5.** Potential energy levels of soil water in different units and corresponding maximum pore size full of water at each suction.

Soil water potential		Soil water suction		Approximate max. pore diameter full of water
Volume basis (kPa)	Mass basis (J kg <sup>-1</sup> )	Hydraulic head (cm H <sub>2</sub> O)	Pressure (bars)	(µm)
0	0	0	0	All pores
-1	-1	10.2	0.01	300
-5	-5	51.0	0.05	60
-10	-10	102.0	0.1	30 (a)
-100	-100	1020.0	1.0	3
-500	-500	5100.0	5.0	0.6
-1000	-1000	10200.0	10.0	0.3
1500	-1500	15300.0	15.0	0.2 (b)

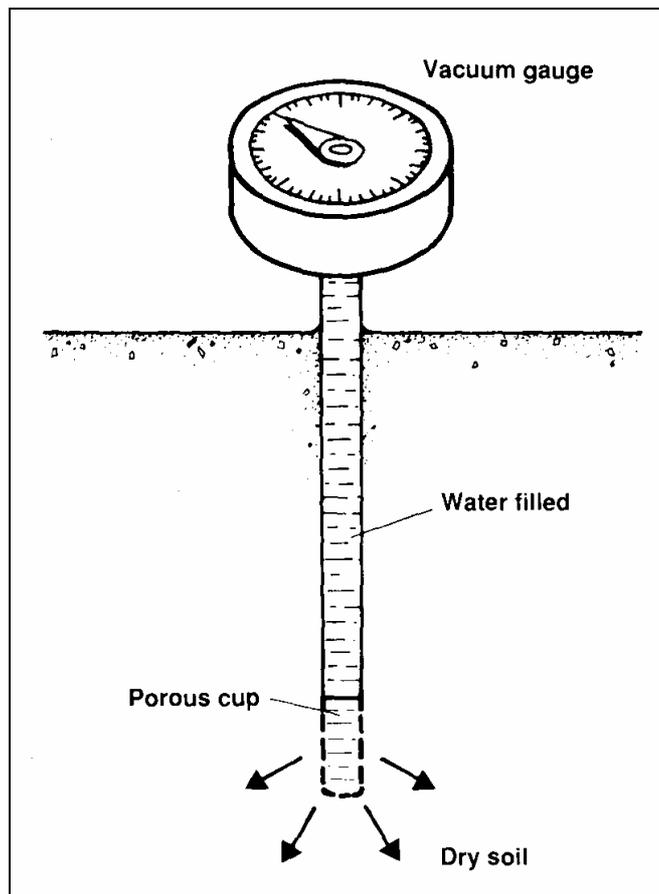
(a) = approximately equal to 'field capacity' for many soils.

(b) = approximately equal to 'permanent wilting point'.

**Soil water potential measurement** - Measurement of the soil water potential provides useful information about the availability of water to plants and about the direction of flow of water in the soil. Although a number of methods of measurement have been developed for research purposes, only a few have been applied in routine irrigation scheduling.

(i) **Tensiometer method.** Tensiometers have been used with considerable success for scheduling irrigation and represent an excellent example of scientific principles and technology being transferred directly to the farmer or grower.

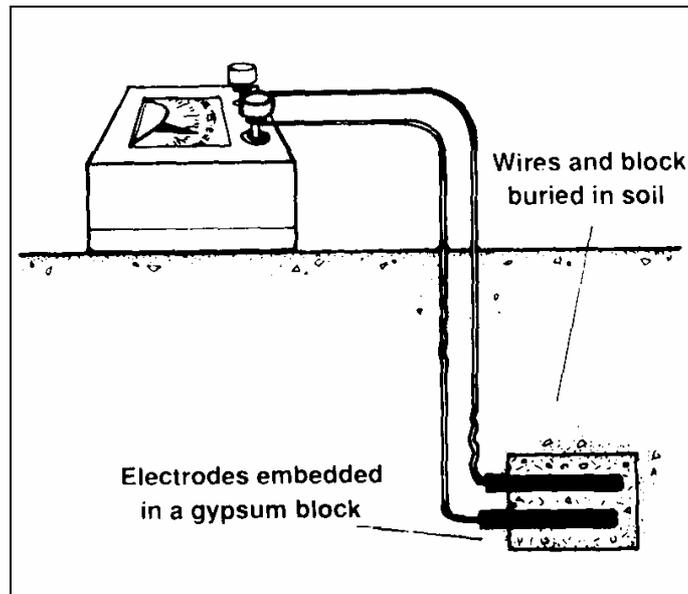
The tensiometer shown in Figure 29 consists of a porous cup connected by a tube to a vacuum gauge. The instrument is filled with water and installed carefully in the soil, ensuring that a good contact is established between the cup and the surrounding soil. In a dry soil a small amount of the water in the tensiometer is 'sucked' out through the porous cup which causes a suction to be registered by the vacuum gauge. In terms of water potential, the water in the tensiometer is at a higher potential than that in the soil so water flows from it out into the surrounding soil. The drier the soil is, the greater the potential gradient and the higher the suction reading on the gauge. Changes in the soil matric potential due to plant water uptake or irrigation can be monitored by the tensiometer.



**Figure 29.** Tensiometer for measuring soil water potential. In a dry soil water is 'drawn' through the porous cup creating a suction to be registered on the vacuum gauge.

Tensiometers are restricted to measurements of matric potential above about -80 kPa (i.e. suctions of <0.8 bar). However, this does not limit their usefulness for irrigation scheduling in many soils since the range from 0 to -80 kPa generally accounts for over 50% of the amount of water available to plants. Tensiometers do not measure water content directly, although they can be calibrated to give approximate values. (ii)

**Electrical resistance blocks.** Porous blocks with electrodes embedded in them (Figure 30) can be used to measure soil water potential ranges outside those covered by tensiometers (i.e. less than -80 kPa). The electrical conductivity of the porous block, made for example from gypsum, is dependent on the amount of water between the electrodes. When installed in soil the amount of water in the block is dependent on the matric potential of the surrounding soil. Therefore, providing the blocks have been suitably calibrated, a measurement of the electrical resistance of the block can indicate the matric potential of the soil.



**Figure 30.** Gypsum block for measuring low soil water potentials. Electrical resistance is measured between the two wires embedded in the gypsum block. Since the water in the block is in 'equilibrium' with the soil water, the drier the soil the greater the resistance between the wires.

### 2.3.3 Relation between soil water content and potential (the moisture characteristic curve)

The relationship between the soil water content and the soil water potential provides useful information about the **water holding capacity** and drainage characteristics of the soil. The amount of water held at any particular soil water potential is mainly dependent on the soil texture and structure since these properties define the sizes of pores in the soil and therefore the amount of water which can be retained in the soil.

The amount of water retained by the soil over a range of soil water potentials can be measured by applying a suction to the soil to overcome the forces of capillarity and adhesion holding the water. This can be achieved by the apparatus shown in Figure 31. A core taken from the soil, without disturbing the structure, is saturated with water and placed on the porous plate of the **tension** table. When the column of water is level with the soil all the soil pores remain full with water. When the column of water is lowered a suction is created which can remove water from the larger soil pores. As the suction (which is equivalent to a negative pressure potential) is increased, by lowering the tube further, water which is held in smaller and smaller pores (i.e. those with lower and lower water potentials) is gradually removed. The loss of water at each suction is easily measured by re-weighing the soil core once it has ceased to drain at each particular suction.

The diameter of the largest pores which can remain full of water ( $d$ ) at any particular suction ( $h$ ) can be calculated approximately as:

$$d(\text{cm}) = \frac{0.3}{h(\text{cm})} \quad (6.6)$$

The maximum pore diameters for a range of soil water suctions are given in Table 5.

The suction (or tension) technique is limited to suctions of less than 1 bar. Water held at lower potentials (i.e. less than -100 kPa) must be removed by a different technique. This involves transferring the cores to a **pressure-plate** apparatus, which is shown in Figure 32. The pressure vessel contains a porous plate which separates the upper pressurized section from the unpressurized section underneath. When the upper section is pressurized, water which is held in the soil is forced out into the lower part of the vessel which is at atmospheric pressure. The increased air pressure on the soil water effectively raises its potential above that of the free water below, causing it to flow out in that direction.

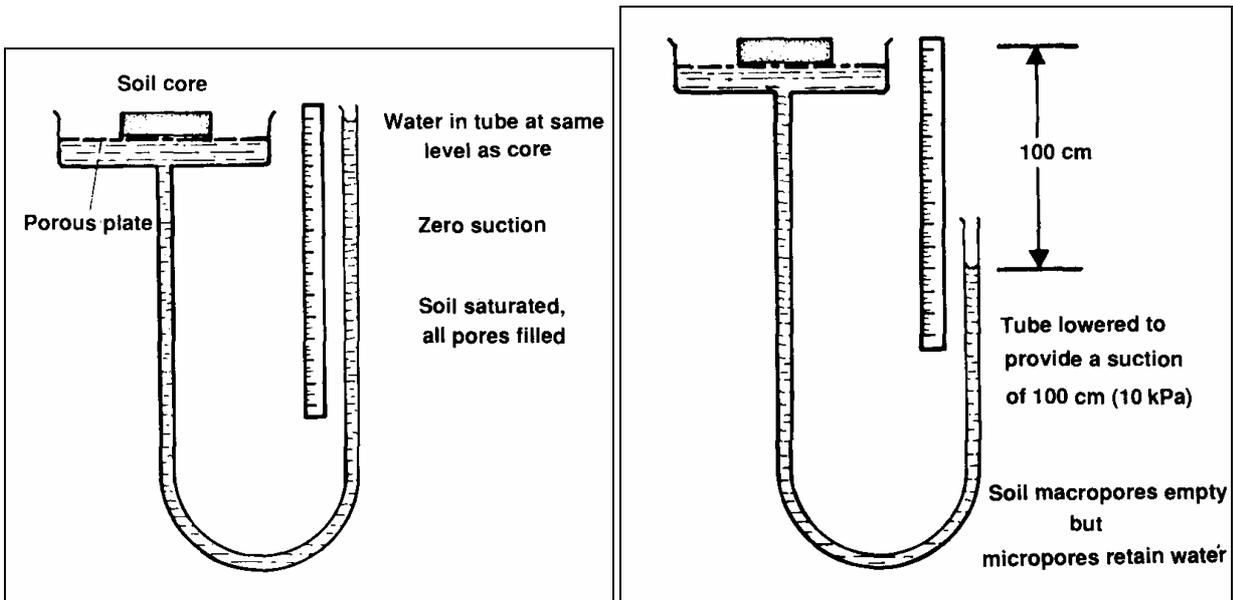


Figure 31. Measurement of the soil moisture release characteristic using tension table apparatus.

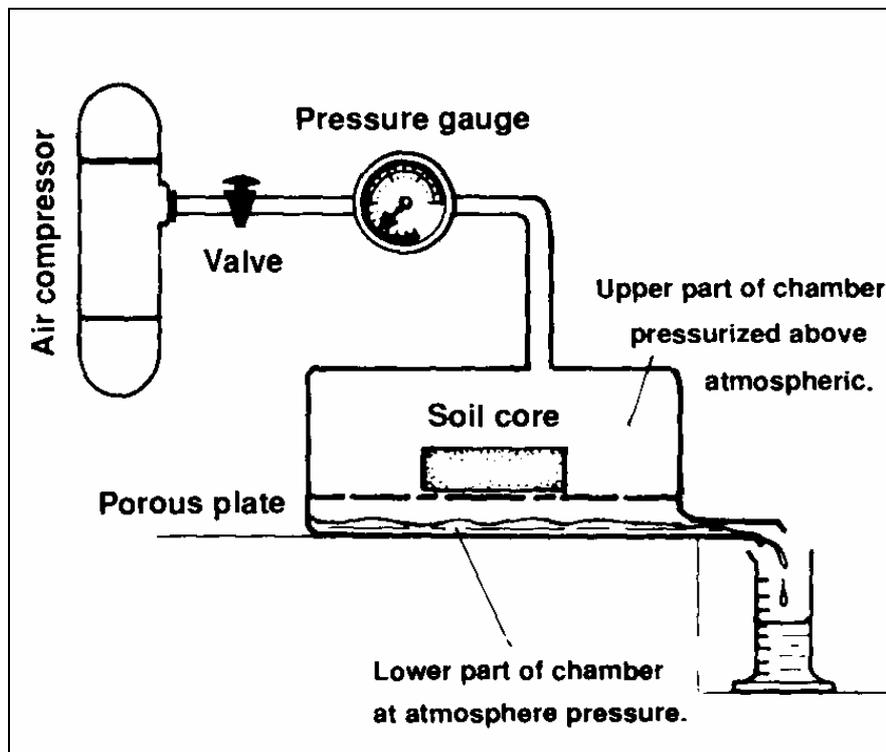


Figure 32. Pressure-plate apparatus for determination of soil water content at low potentials ( $< -100 \text{ kPa}$ ). Water is forced out of the soil by the greater pressure in the upper chamber.

The relationship between the amount of water held in the soil and the soil water potential is called the **moisture characteristic** and this is normally plotted as a curve, as for example in Figure 33. The clay soil, in Figure 6.18, has a higher water content at each potential than the sandy soil because the clay has a finer texture with smaller pores and therefore a greater amount of water held at the lower potentials. The sandy soil mainly consists of large pores and the water in these can be removed at relatively low suctions. The shape of the curve for the sandy soil reflects the restricted size range of pores, most of which empty at, or around, the same suction (e.g. 0.1 bars in Figure 33). The clay soil has a more gradual curve reflecting the more uniform distribution of pore sizes.

Soil structure affects the shape of the moisture release curve at low suctions (e.g. 0 to about 0.1 bar) because it is the shape and packing of the structural units which mainly defines the volume of soil macropores. Differences between soils at low suctions provide information about the relative drainage and aeration capacities of the soils, as will be described later.

The shape of the moisture characteristic curve also depends on whether the soil pores are being emptied or being filled with water. The most commonly reported curve is the 'desorption', or 'moisture release curve', as described earlier; however the relationship between soil matric potential and soil moisture content can also be measured by sequentially re-wetting the soil. The equilibrium soil moisture content at any particular suction is less under wetting (sorption) conditions than under drying (desorption) conditions and this phenomenon is called **hysteresis** (see Figure 34). Hysteresis will occur in field soils and is likely to influence the accuracy of moisture content calculations for soils which are undergoing sequential wetting and drying. It can also affect the conditions of water flow in the soil. Hysteresis arises due to: (i) the narrow openings of many soil pores (called the 'ink bottle' effect), (ii) entrapped air being present during wetting, and (iii) changes in the orientation of clay-sized particles during wetting and drying.

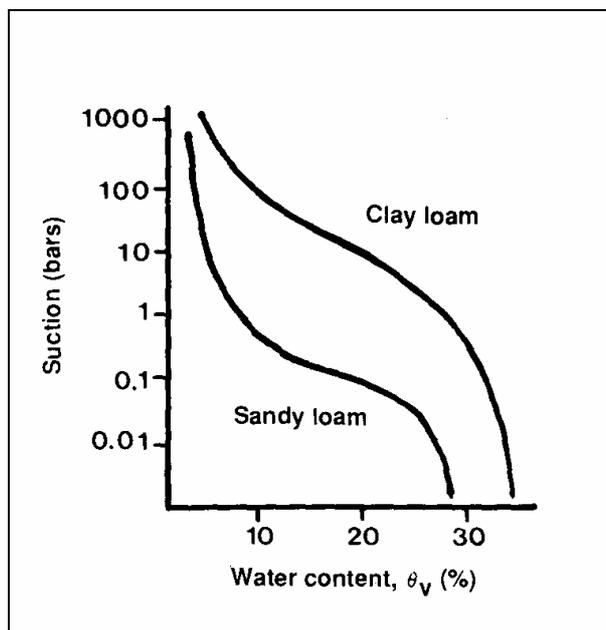


Figure 33. Soil moisture release characteristic curves showing the effect of soil texture.

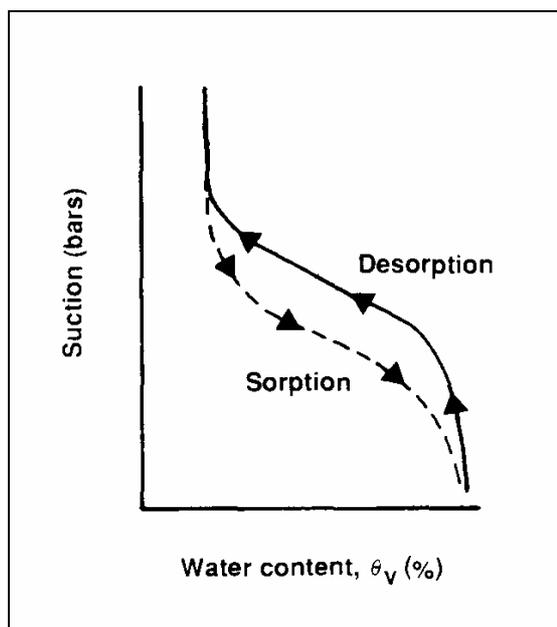


Figure 34. Desorption and sorption curves for the same soil illustrating the phenomenon of hysteresis.

## 2.4 Soil water classification and storage

### 2.4.1 Classification

The soil water potential concepts described in the previous section allow a precise definition of the soil water status. However, in practical situations some more general concepts are often used. In order to describe these it is useful to consider a soil which has recently received a heavy application of water, either as rainfall or irrigation, as shown in Figure 35. The following terms are often used to describe the moisture status of such a soil:

**Saturation (s)** - At the end of the irrigation, water would have replaced almost all the air in the topsoil and drainage would be occurring through the macropores. Field soils normally have some air trapped within them and they rarely reach a fully saturated state; nevertheless saturation can loosely be described as the amount of water in the soil *“when all the pores are full of water and no air remains in the soil”*. For most soils this is a temporary state since drainage would be occurring allowing air back in through the macropores. The matric potential of the soil at saturation is zero.

**Field capacity (FC)** - When the application of water is stopped, the water in the largest pores (i.e. water at the highest potential) drains rapidly. After a period of time, often about 1-2 days, the rapid drainage of macropores is complete and the soil is at a state called 'field capacity' (Figure 35). Although it is not possible to define precisely, field capacity can be described as the state of the soil *“after rapid drainage has effectively ceased and the soil water content has become relatively stable”*.

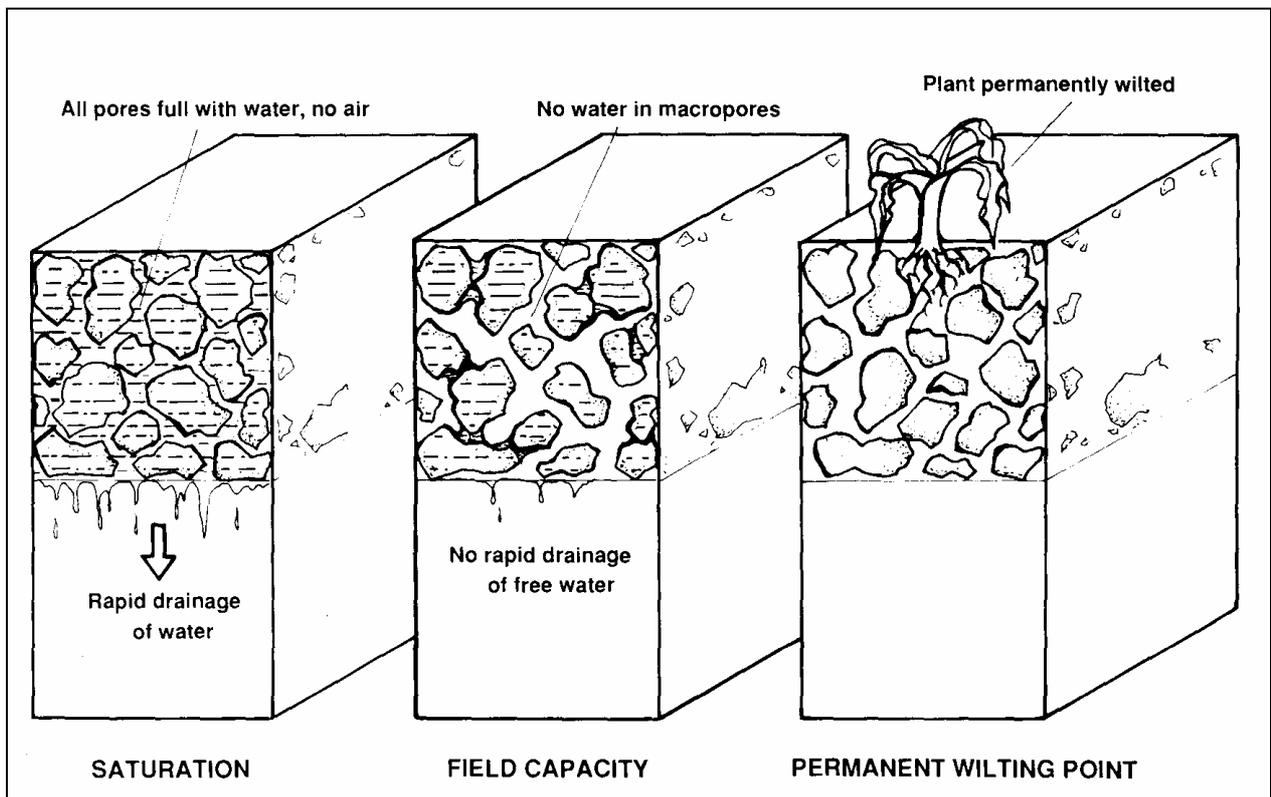
Free draining soils can reach field capacity relatively quickly but if there is a barrier to water movement, such as a clay pan, then field capacity takes longer to be attained, if it can be reached at all. Field capacity is not a true equilibrium condition but it can be determined by monitoring the change in water content of a field soil following saturation.

Laboratory methods of determining field capacity are not always reliable because they ignore profile changes of texture and structure, both of which influence drainage and water storage. Nevertheless, field capacity is often related to the soil moisture content at a potential of -10 kPa, although the most appropriate potential for a particular soil may be between -5 to -20 kPa (equivalent to 0.05-0.2 bars suction).

Examples of the approximate amounts of water held at field capacity in a range of New Zealand soils are given in Table 6.

**Permanent wilting point (PWP)** - Plant removal of water and evaporation from the soil cause it to dry below field capacity (Figure 35). As it does so, the water which remains in the soil becomes more and more difficult to abstract because it is held at greater suctions (i.e. the remaining water has a lower potential). Initially the plant may start to wilt during the day, when the demand for water is the highest, but eventually the plant may reach a point at which it is wilted both day and night. This is the 'permanent wilting point' and can be described as the *“amount of water in the soil at which plants are permanently wilted”*.

Like field capacity, the permanent wilting point is difficult to define because it can vary with soil characteristics, such as depth, and with plant environmental conditions. Nevertheless, the soil water potential at which permanent wilting point occurs is generally taken to be -1500 kPa (equivalent to 15 bars suction). The water remaining is thus water which is held tightly in small micropores and adsorbed on to soil particles (Figure 35). Some examples of the amounts of water at PWP in a range of New Zealand soils are given in Table 6.



**Figure 35.** Representation of soil water contents at Saturation, Field Capacity, and Permanent Wilting Point.

**Table 6.** Approximate moisture contents (% v/v) of selected N.Z. soils (A horizons only) (adapted from Soil Bureau, 1968 and Gradwell, 1976).

Soil type (A horizons)	Field capacity <sup>1</sup>	Permanent Wilting point <sup>2</sup>	Available Water capacity
Conroy Sandy Loam	29.4	7.0	22.4
Stratford Sandy Loam	51.5	28.4	23.1
Timaru Silt Loam	36.1	14.5	21.6
Marton Silt Loam	42.2	18.9	23.3
Templeton Silt Loam	33.8	15.5	18.3
Dannevirke Silt Loam	44.7	23.8	200.9
Temuka Silt Loam	47.0	23.9	23.1
Taita Clay Loam	42.3	20.8	21.5
Hamilton Clay Loam	36.3	18.5	17.7
Waiareka Clay	51.6	32.1	19.4
Ruatangata Clay	43.7	24.9	18.7
Egmont Black Loam	44.2	24.3	19.9

<sup>1</sup> = Moisture content 0.2 bar (-20 kPa)

<sup>2</sup> = Moisture content at 15 bar (-1500 kPa)

**Available water capacity (AWC)** - The available water capacity can be loosely defined as “the amount of water which a soil can store for plant growth”. It is numerically equal to the amount of water held at field capacity (FC) minus the amount held at permanent wilting point (PWP):

$$AWC\% = FC (\%, v/v) - PWP (\%, v/v) \quad (6.7)$$

Although the AWC can only be an approximate value, it nevertheless allows useful comparisons to be made between soils and also assists in scheduling irrigation. Examples of the amount of available water in some New Zealand topsoils are given in Table 6.

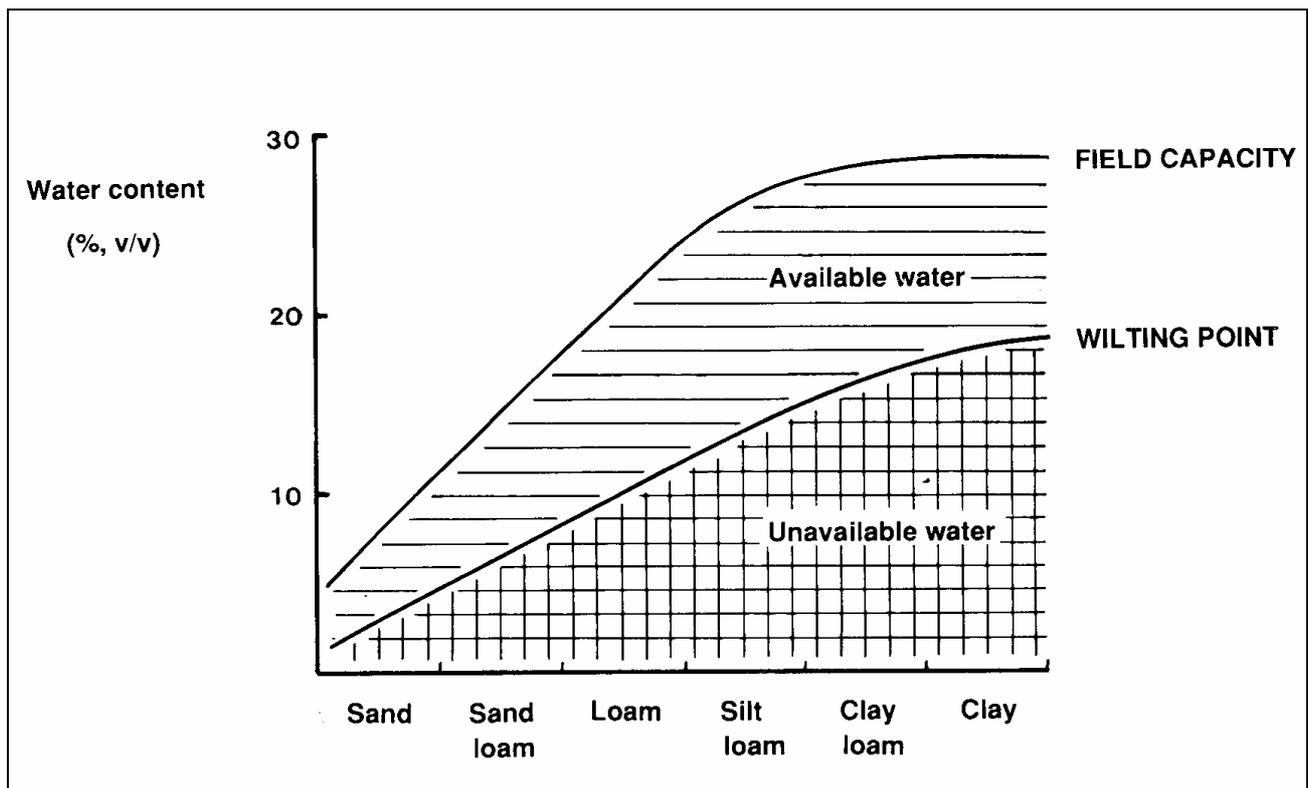
From an appreciation of soil water potential concepts it should be apparent that all the water within the AWC range is not equally available to plants. Water becomes more difficult for plants to absorb the closer the soil water potential moves toward wilting point (-1500 kPa). This will be discussed further.

The following factors have a major influence on the available water capacity of a soil:

(i) **Texture.** As shown in Figure 36, in general the finer the soil texture the greater the AWC. In a sand the difference between FC and PWP is small, whilst in a clay the difference is large, giving it a large AWC. Close examination of Figure 36 shows that the difference between FC and PWP is, in fact, greatest for a silt loam texture. The silt loam has a greater AWC than the clay because it has a smaller PWP value, whilst having a similar FC value.

(ii) **Organic matter.** In most mineral soils the main contribution organic matter makes to the AWC is in assisting in the development and maintenance of good soil structure. A mineral soil with a high organic matter content has a greater AWC than one with a low organic matter level. Organic particles attract water because of their surface charge but this water is generally unavailable to plants and it is the effects on soil structure which are important.

(iii) **Salt.** The effects of salts in soils will be discussed in detail, but at this stage it should be appreciated that a high concentration of salts in the soil solution lowers the water potential, making it less available to plants. High salt concentrations may occur naturally (e.g. in saline soils) or may be produced locally by the addition of fertilizers (e.g. potassium chloride).



**Figure 36.** General relationships between soil texture and soil water content at field capacity and permanent wilting point (adapted from Cassel, 1983).

#### 2.4.2 Amount of water storage in the soil profile

The amount of available water which can be stored in the soil is a function of six main factors: texture, structure, organic matter, soil depth, profile layering, and stone content\_ as described by the following equation:

$$\text{amount of available water in soil profile} = f \left[ \begin{array}{l} \text{texture, structure,} \\ \text{layering, depth,} \\ \text{organic matter, stones} \end{array} \right] \quad (6.8)$$

The first three factors have been discussed in the previous section and the remainder are described below.

**Depth** - The depth of the soil profile defines the overall volume of soil available for soil water storage. Assuming that all other factors in Equation 6.8 are constant, a deep soil holds more water than a shallow one. The following equation can be used to calculate the amount of available water stored in a uniform soil layer.

$$\begin{aligned} \frac{\text{amount of available water in soil layer (mm)}}{\text{(mm)}} &= \frac{\text{FC (\%)} - \text{PWP (\%)}}{100} \times \frac{\text{depth of layer (mm)}}{\text{(mm)}} \\ &= \frac{\text{AWC (\%)}}{100} \times \frac{\text{depth of layer (mm)}}{\text{(mm)}} \quad (6.9) \end{aligned}$$

where FC and PWP are expressed as volumetric water contents ( $\theta_v$ ).

The total amount of available water stored in a soil profile or within the rooting depth of a plant equals the sum of available water stored in each layer. The approximate amounts of available water held in some major New Zealand soil groups are shown in Table 7.

Examination of the data in Table 7 shows that there is a large difference in the amount of available water that can be stored in different soils. Organic soils can store very large amounts of available water whilst brown loams are unable to store much available water. Yellow-brown pumice soils are relatively drought resistant because both the A horizon and the B/ C horizons can store large amounts of available water. Yellow-grey earths conversely have a low storage capacity in the subsoil and are therefore more susceptible to drought. The high storage capacity of yellow-brown loams is due to the presence of allophane.

Calculation of the amount of available water stored in a soil profile is useful in scheduling irrigation. The production capacity of a soil can also be better appreciated once the water storage capacity is known and compared with the regional climatic conditions.

**Table 7.** Approximate amounts of available water held in some major New Zealand soil groups (adapted from New Zealand Standard 5130: 1973).

Soil group	Amount of available water (mm water per 100 mm depth of soil)	
	Topsoil	Subsoil
Northern yellow-brown earths	18	11
Northern podzols and podzolised soils	22	9
Brown loams on basalt	13	8
Brown-granular clays (N. Auckland)	18	15
Brown-granular loams (S. Auckland)	16	8

Yellow-brown loams	20	12
Yellow-brown pumice soils	26	22
Central and Southern yellow brown earths	20	11
Yellow-grey earths	22	11
Brown-grey earths	18	-
Organic soils (peat)	20-25	At least 20-25

The effective depth of profile may actually be dependent on the depth of rooting of the crop rather than on the soil itself. Some examples of the effective root depths of crops grown under irrigation are given in Table 8. It should be emphasized that these figures relate to situations where there is no soil restriction to root growth.

**Soil profile layering** - It is easy to appreciate how a clay layer, or a pan, which restricts drainage can give the soil an effectively greater water storage capacity. Somewhat less obvious is the effect of a coarse sand layer located below a finer textured soil such as a silt loam. Because of the influence of soil water potential on hydraulic conductivity the sand layer cannot transmit water as rapidly as the silt loam above. Therefore any water which is applied to such a soil will not drain away immediately but will accumulate above the boundary with the sand layer. The reasons for this will be discussed in more detail later, but the phenomenon means that many soils which overlie sands or gravels have a higher water storage capacity than would be expected from a calculation based on Equation 6.9 alone.

**Stones** - If the soil profile is stony then the volume of soil which can store water is reduced in proportion to the percentage stone content.

**Table 8.** Effective root depths of selected crops grown in an unrestrictive soil under irrigation (adapted from New Zealand Standard 5103: 1973).

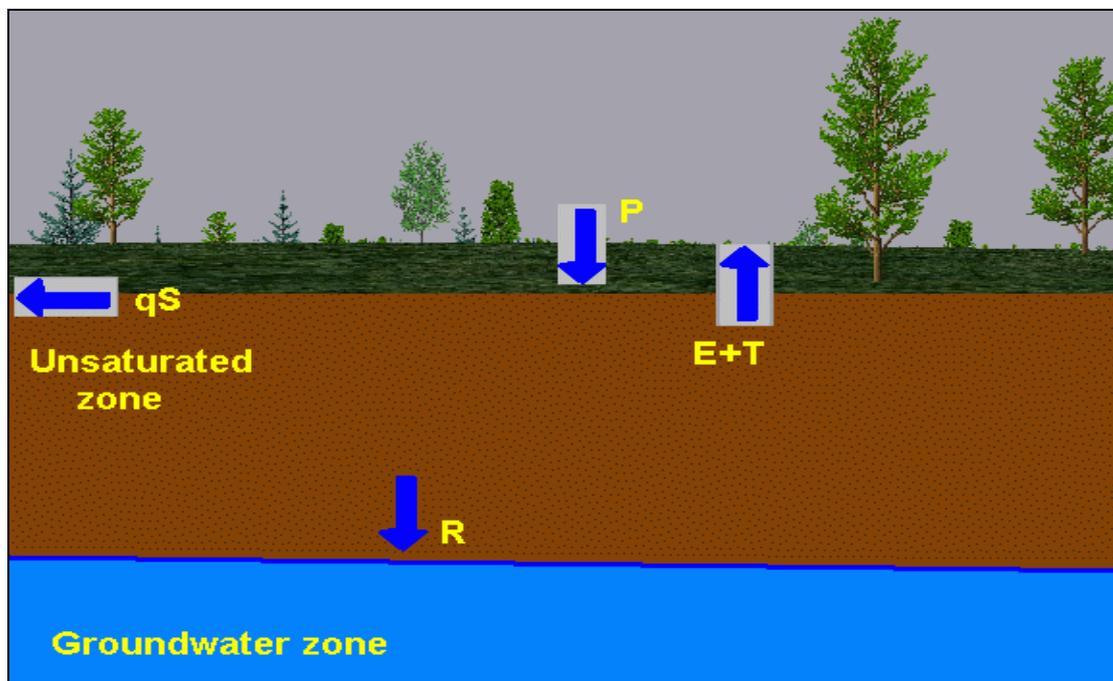
Crop	Root depth (cm)	Crop	Root depth (cm)
<b>FRUIT CROPS</b>		<b>VEGETABLES (continued)</b>	
Apples	75-120	Potatoes	60-90
Apricots	60-135	Squash	60-90
Citrus	60-120	Swedes	60-90
Grapes	45-90	Tomatoes	60-120
Peaches	60-120	Turnips	30-60
Pears	60-120		
Plums	75-120	<b>FIELD CROPS</b>	
Strawberries	30-45	Barley	90-110
		Lucerne	120-185
<b>VEGETABLES</b>		Maize	60-90
Beans	45-60	Oats	60-75
Brussel sprouts	45-60	Wheat	75-110
Cabbages	45-60	<b>PASTURE AND FODDER CROPS</b>	
Carrots	45-60		
Cauliflowers	45-60		
Cucumbers	45-60	Choumoellier	45-60
Lettuces	15-45	Pasture	30-75
Parsnips	60-90	Rape	45-60
Peas	45-60		

### 3 CALCULATION OF FLOW IN UNSATURATED ZONE AND ESTIMATION OF RECHARGE TERM FOR GROUNDWATER MODELS

#### 3.1 Introduction and terminology

##### 3.1.1 Recharge of infiltrated water through the unsaturated zone

In groundwater models discussed the term recharge  $R$  has been continuously used, but so far no exact way of estimating  $R$  has been given. Specifically the aim is to be able to calculate the variation of recharge term as a function of time. The purpose of this Chapter is to provide definition of the recharge term and provide methods for calculation of  $R$  in the case that other components of the hydrological cycle – precipitation  $P$ , evaporation from soil  $E$ , transpiration  $T$  and surface runoff  $q_s$  - are known. Recharge  $R$  is the amount of water from the total input (precipitation  $P$  + artificial groundwater recharge) that flows through the unsaturated zone to the groundwater table. The unsaturated zone (or vadose) is the hydrological connection between the surface water component of the hydrologic cycle and the groundwater component shown Figure 37.



*Fig. 37. Schematic diagram of the water balance in unsaturated soil profile.*

It has to be pointed out that recharge  $R$  is not same than the amount of water leaving the rooted zone (called deep percolation  $D$  here). For the first, there is a time lag between  $D$  and  $R$  because water has to move through the unsaturated profile. For the second, some part of  $D$  may flow upwards towards the rooted zone in periods when precipitation is smaller than sum of evaporation and transpiration, i.e. during periods when there is water deficit in the rooted zone. Therefore, accurate calculation of the magnitude and timing of  $R$  can be obtained only by calculating the water balance in the unsaturated soil profile by solving the so called Richard's equation.

The unsaturated zone plays a crucial role in the transfer of pollutants. Many pollutants present in the surface waters eventually find their way into the groundwater through the unsaturated zone. Accident spills of chemicals, application of fertilisers and pesticides on the land surface, leaks from gasoline storage tanks, septic tank drainage, leaching from landfills are examples of human activities that contribute to the leaching of contaminants through the unsaturated zone into the

groundwater (Haverkamp et al., 2002).

### 3.2 Soil water retention curve (pF-curve)

The so called pF-curve - soil moisture retention curve - gives the relationship between volumetric soil moisture content  $\theta$  ( $\text{m}^3\text{m}^{-3}$ ) and soil water pressure head  $h$  (cm). It describes the soil's ability to store or release water. The water retention characteristic is a highly non-linear 'S'-shaped curve. Soil is a porous medium that consists of solid particles and pores that are either water-filled or air-filled. The saturated water content,  $\theta_s$  ( $\text{m}^3\text{m}^{-3}$ ) depends on the packing density of the soil and the specific density of the individual soil particles. In a fully saturated soil, soil water pressure head is nil. As stated in the previous section, when  $h$  is lowered, the larger pores start to drain first.

NOTE! Usually saturated water content measured in laboratory is smaller than soil porosity  $\varepsilon$  due to the fact some air will always be entrapped in soil sample when  $\theta_s$  is measured. An approximation very often used is that  $\theta_s = 0.9\varepsilon$ .

The decrease in actual water content,  $\theta$ , when larger pores will be emptied, is different in different soils. Sandy soil has many wide pores and loses much of its moisture for soil water pressure head greater than e.g. -100 cm. Clay soil consists of fine particles and is finely porous. In clay soils wide pores are lacking and therefore clay soils release very small amount of water for soil water pressure head greater than -100 cm ( $pF=2$ ) (see Fig. 38). When soil water pressure head is still decreased, e.g. to -10 000 cm ( $pF=4$ ), the coarse sand has lost almost all of its moisture content, whereas in the heavy clay soil volumetric water content is still around 0.30.

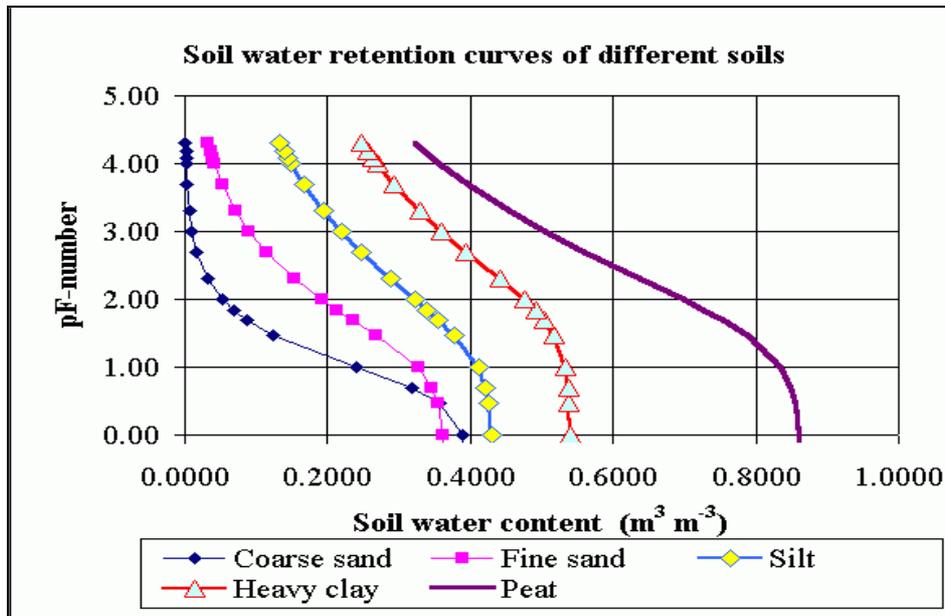
#### 3.2.1 Fraction of water that can be removed from profile by using subsurface drains

It is useful to estimate how much of the water available in the rooted zone can be removed by subsurface drains alone, i.e. assuming that evaporation is very small and water content in the soil surface is reduced by removing water via subsurface drains. The common depth of subsurface drains is 100 cm in Finnish conditions and therefore it can be assumed that at soil surface the soil water pressure will be -100 cm in the case that the drains have been operating long enough and a steady-state condition is prevailing between the drain level ( $h$  is 0 cm there) and soil surface ( $h = -100$  cm). If the shape or the parameters of the soil water retention curve are known it is possible to calculate what is the water content in the soil surface compared to saturated case. The relative water content remaining at soil surface (fractional value) is then

$f_R = \theta(-100) / \theta_s$  and fraction drainable is then  $f_D = 1 - f_R$ .

#### 3.2.2 Available water for plants

If the soil water retention curve is known it is possible to compute an estimate for the amount of water available for plant roots. The upper soil moisture limit may be considered to be at field capacity, i.e.  $pF=2$  or soil water pressure head  $h = -100$  cm. The lower soil moisture limit is the so called "wilting point", where  $pF=4.2$  or  $h = -16 000$  cm.



**Figure 38.** Some characteristic  $\theta(h) - h$  relationships for different soils (here y-axis is pF-number,  $h = -10^{pF}$ ). Parameters of the curves shown in Section 7.3 in Table 7-1.

### 3.3 Mathematical description of $\theta(h)$ -curve

#### 3.3.1 Van Genuchten's model

The soil water retention curve can be described mathematically by the so called van Genuchten's model shown below. The equations relate soil water pressure head  $h$  with volumetric water content  $\theta$  - i.e.  $\theta(h)$  - and unsaturated hydraulic conductivity  $K(h)$  (see Eq. (7-13)) using a concept called "relative saturation rate"  $S_R$ .  $S_R$  varies between 0..1. When  $S_R = 0$  then soil is very dry and  $\theta$  equals the residual water content  $\theta_R$  and when  $S_R = 1$ , soil is fully saturated and  $\theta = \theta_S$ .

$$S_R = \begin{cases} \left[ 1 + [\alpha|h - h_a|]^\beta \right]^{-\gamma} & ; \text{if } h < h_a \\ 1 & ; \text{if } h \geq h_a \end{cases}$$

$$\gamma = 1 - 1/\beta \tag{7-12}$$

$$\theta = \theta_R + (\theta_S - \theta_R)S_R \quad \text{or} \quad S_R = \frac{\theta - \theta_R}{\theta_S - \theta_R}$$

where  $S_R$  is the relative saturation rate (0..1),  $h$  is soil water pressure head (cm),  $h_a$  is a so called air entry point (cm); very often  $h_a = 0$  cm,  $\theta$  is the volumetric water content ( $\text{cm}^3 \text{cm}^{-3}$ ),  $\theta_S$  is the saturated water content ( $\text{cm}^3 \text{cm}^{-3}$ ),  $\theta_R$  is the residual water content ( $\text{cm}^3 \text{cm}^{-3}$ ),  $\alpha$  is a parameter ( $\text{cm}^{-1}$ ),  $\beta$  is a parameter (dimensionless) and  $\gamma = 1 - 1/\beta$ . Residual water content  $\theta_R$  can be defined as the asymptote of the pF-curve when  $h$  gets very high negative values. Usually  $\theta_R$  is very small - on the order of 0.001--0.02 for coarse soils but gets as high values as 0.10..0.20 for heavy clay soils.

### 3.4 Hydraulic conductivity curve

When  $h=0$  cm, i.e. the soil is fully saturated, soil hydraulic conductivity  $K(h)$  has its maximum value,

saturated hydraulic conductivity,  $K_S$ . As  $h$  decreases, the larger pores will be emptied first and they do not contribute to the water flow. This implies that the unsaturated  $K(h)$  is continuously decreased as  $h$  decreases. The measurement of the relationship between  $h$  and  $K(h)$  is difficult and time consuming. Therefore, mathematical description are used to relate soil water pressure head  $h$  and hydraulic conductivity  $K(h)$ .

$$K(h) = K_S K_R(h)$$

$$K_R(h) = S_R^\lambda \left[ 1 - \left( 1 - S_R^{1/\gamma} \right)^\gamma \right]^2 \quad (7-13)$$

where  $S_R$  is the relative saturation rate calculated from Eq. (7-12),  $K(h)$  is the unsaturated hydraulic conductivity ( $\text{cm d}^{-1}$ ),  $K_S$  is the saturated hydraulic conductivity ( $\text{cm d}^{-1}$ ) and  $K_R(h)$  is the relative hydraulic conductivity (0..1). Examples of relative values of  $K_R(h)$  (0..1) for different soils are shown in Figure 39. It is usually assumed that parameter  $\lambda$  is 0.5 in Eq. (7-13).

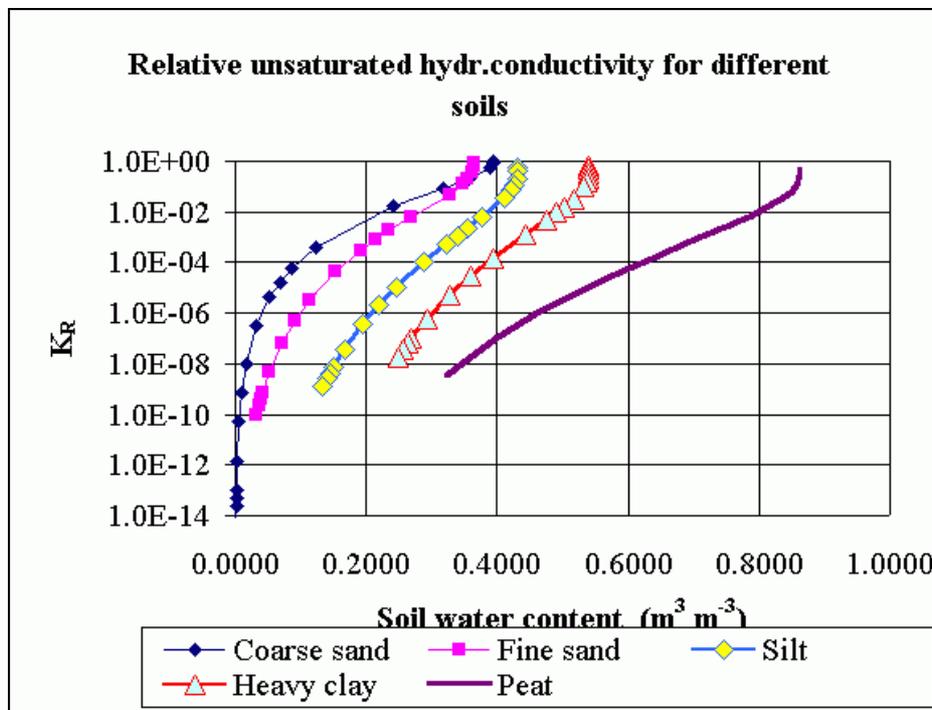


Figure 39. Some characteristic  $K_R(\theta)$  relationships for different soils (relative value between 0..1).

### 3.4.1 Saturated hydraulic conductivity $K_S$

The saturated hydraulic conductivity  $K_S$  is strongly related to soil structure. The saturated hydraulic conductivity is very much influenced by effects such as macropores, stones, fissures, cracks and other irregularities formed for various biological and mechanical reasons. Hence, it is the parameter which is the most difficult to predict. The models proposed in the literature either give estimations of the capillary conductivity value or are based on site and soil specific databases. The results should therefore be considered with caution when applied for field studies (Haverkamp et al. 2000). Mishra and Parker (1990) used the Mualem model (1976) with the Van Genuchten water retention function (Eq. (7-12)) to obtain a closed-form expression of the saturated hydraulic conductivity:

$$K_s = c_1 \frac{[\theta_S - \theta_R]^{2.5}}{h_{ag}^2} \quad (7-14)$$

where  $c_1$  is a constant including the effects of fluid characteristics and the porous media geometric factor; it has a value of 108 cm<sup>3</sup>/s when  $K_s$  is expressed in cm/s;  $h_{ag}=1/\alpha$ , where  $\alpha$  is the parameter of Van Genuchten-model (1980). Ahuja et al. (1985) used the general Kozeny-Carman approach to determine the saturated hydraulic conductivity from the effective porosity ( $\varepsilon - \theta_R$ ):

$$K_s = c_2 [\varepsilon - \theta_R]^{c_3} \quad (7-15)$$

where  $c_2$  is equal to 1058 cm/h when  $K_s$  is expressed in cm/h; and  $c_3$  takes a value of 4 or 5.

### 3.5 Solution of water balance in unsaturated-saturated soil profile

#### 3.5.1 Water movement in soil

For the flow of water in saturated or unsaturated soil, Darcy's law applies:

$$q_W = -K(h) \frac{dH}{dz} = -K(h) \left( \frac{d(h+z)}{dz} \right) = -K(h) \left( \frac{dh}{dz} + 1 \right) \quad (7-16)$$

where  $q_W$  is the flow velocity (cm d<sup>-1</sup>),  $K(h)$  is the hydraulic conductivity of the soil (function of soil water pressure head) (cm d<sup>-1</sup>),  $H$  is the hydraulic head ( $h+z$ ) (cm) and  $z$  is vertical coordinate, positive upwards (vertical flow in soil considered now) (cm).

#### Steady-state profile

**Example.** Consider a steady-state case when vertical flow in the unsaturated and in the saturated zone is zero. The thickness of the vertical profile is 250 cm and groundwater depth is 200 cm from the soil surface.

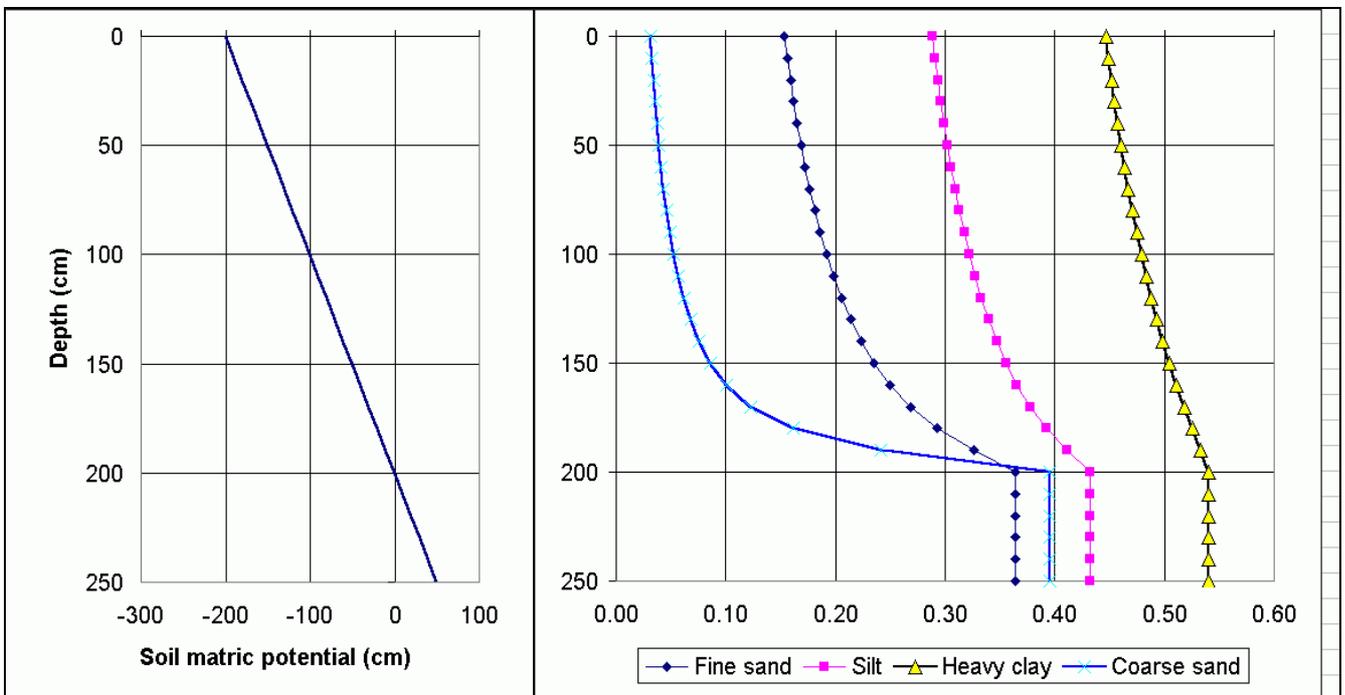
1. Calculate the soil water pressure head profile.
2. Calculate the corresponding soil water content profile for different soils types when the parameters of the soil water retention curve are given in Table 9.

Solution for a): Flow is zero when  $dH/dz$  in Darcy's equation (7-16) is zero, i.e.  $H$  is constant over depth  $z$ . Hydraulic head  $H$  is the sum of soil water pressure head  $h$  and elevation  $z$ , i.e.  $H=h+z$ , and soil water pressure head is zero at groundwater level, i.e. at depth 200 cm. Therefore at groundwater level  $H=50$  cm ( $z$  is zero at the bottom of the profile and  $z=250$  at soil surface). Since  $h+z$  is constant it is easy to see that  $h= -200$  cm at soil surface and  $h=50$  cm at the bottom of the profile. The soil water pressure is linear with respect to  $z$  and the graph is shown in Figure 40.

**Table 9.** Van Genuchten-model parameters for five different soil types.

Soil Description	$\theta_S$	$\theta_R$	$\alpha$	$\beta$
Coarse sand	0.395	0	0.1596	1.727
Fine sand	0.364	0	0.0632	1.337
Silt	0.432	0.02	0.045	1.190
Heavy clay	0.54	0.2141	0.0154	1.145
Peat	0.863	0	0.0327	1.152

The soil water pressure profile can be expressed as volumetric water content for different soil type. The parameters shown in Table 7-1 were used in the van Genuchten equation (7-12). The curves corresponding the steady-state pressure head profile are shown in Fig. 7-7 indicating that for coarse sand the profile is quite dry at soil surface (pressure head  $h$  is -200 cm). For heavy clay soil the soil water content is still quite high showing that with drainage alone it is not possible to drain a heavy clay soil profile.

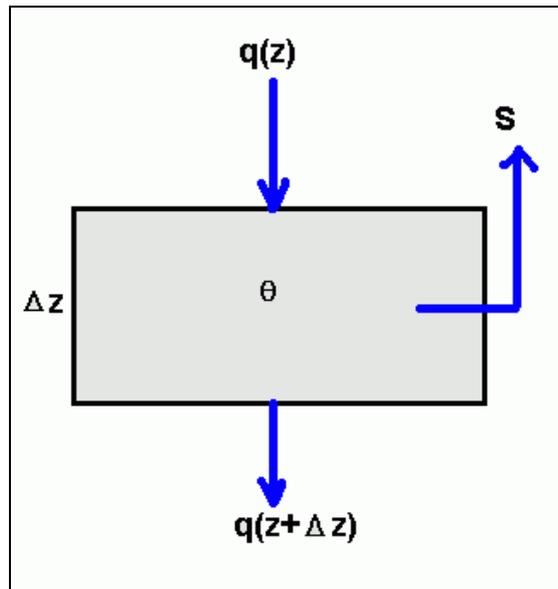


**Figure 40.** Steady-state soil water pressure head profile (left) and the corresponding soil water content profiles for different soils (right).

### 3.6 Derivation of the Richards' equation

Consider the water balance of a soil layer show in Figure 41. Inflow into the layer is denoted by  $q(z)$  ( $\text{cm d}^{-1}$ ) and outflow from the layer is  $q(z+\Delta z)$ .  $S$  is the amount of water taken up by soil roots ( $\text{cm}^3 \text{cm}^{-3} \text{d}^{-1}$ ). The change in water content  $\theta$  over time can then be written as

$$\frac{\partial \theta}{\partial t} = \frac{q(z) - q(z + \Delta z)}{\Delta z} - S \quad (7-17)$$



**Figure 41.** Water balance components of a single soil layer.

By taking into account the definition of partial derivative it is possible to write Eq. (7-17) in the form:

$$\frac{\partial \theta}{\partial t} = -\frac{\partial q(z)}{\partial z} - S \quad (7-18)$$

Eq. (7-16) is then substituted into Eq. (7-18):

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z} \left[ K(h) \left( \frac{\partial h}{\partial z} + 1 \right) \right] - S \quad (7-19)$$

The drawback of Eq. (7-19) is that it includes two dependent variables: soil water content  $\theta$  and soil water pressure head  $h$ . The solution to this problem is to eliminate  $\theta$  from Eq. (7-19) by adopting the concept of differential water capacity  $C(h)$ , which is defined as the derivative of the soil water retention curve:

$$C(h) = \frac{\partial \theta}{\partial h} \quad (7-20)$$

The Richards' equation is obtained by combining Eqs. (7-19) and (7-20):

$$C(h) \frac{\partial h}{\partial t} = \frac{\partial}{\partial z} \left[ K(h) \left( \frac{\partial h}{\partial z} + 1 \right) \right] - S \quad (7-21)$$

The numerical solution of Eq. (7-21) will be discussed next.

### 3.7 Numerical solution of the Richards' equation

There exist analytical solutions to Eq. (7-21) only in particular cases and in practical problems the Richards' equation has to be solved numerically. Finite difference method is the most often used method and in this case partly implicit and partly explicit solution method is adopted. The finite difference approximation of Eq. (7-21) is given here as follows (see also Figure 42):

$$C_{i,tot} \frac{h_i^{t+1} - h_i^t}{\Delta t} = \frac{1}{\Delta z_i} \left\{ K_{i,i+1}^t \left[ \frac{h_{i+1}^{t+1} - h_i^{t+1}}{z_{i+1} - z_i} + 1 \right] - K_{i-1,i}^t \left[ \frac{h_i^{t+1} - h_{i-1}^{t+1}}{z_i - z_{i-1}} + 1 \right] \right\} - S(h_i^t) \quad 7-22$$

where  $C_{i,tot}$  is a so called 'total derivative approximation' to the differential water capacity shown in Eq. (7-20) (determination of  $C_{i,tot}$  will be discussed in Section 7.3.3.1),  $\Delta t$  is the time step of calculation,  $\Delta z_i$  is the thickness of the  $i^{th}$  node,  $z_i$  refers to  $z$ -coordinate of node  $i$ ,  $K_{i-1,i}^t$  and  $K_{i,i+1}^t$  are the internodal hydraulic conductivities. Eq. (7-22) can be converted into a linear equation with three unknown values at nodes,  $i-1, i$  and  $i+1$  from time level  $t+1$ . The same type of equation can be written for all intermediate nodes  $i=2,..,N-1$ .

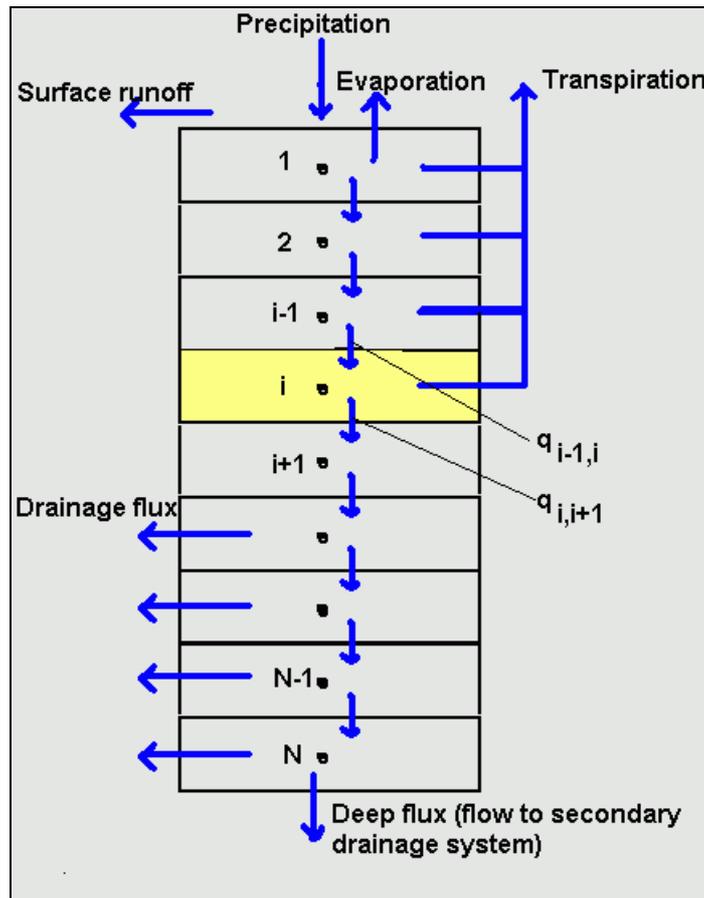
$$\alpha_i h_{i-1}^{t+1} + \beta_i h_i^{t+1} + \phi_i h_{i+1}^{t+1} = \varphi_i \quad , i = 2, \dots, N - 1$$

$$\alpha_i = - \frac{\Delta t K_{i-1,i}^t}{\Delta z_i (z_i - z_{i-1})}$$

$$\phi_i = - \frac{\Delta t K_{i,i+1}^t}{\Delta z_i (z_{i+1} - z_i)} \tag{7-23}$$

$$\beta_i = C_{i,tot} - \alpha_i - \phi_i$$

$$\varphi_i = C_{i,tot} h_i^t + \frac{\Delta t}{\Delta z_i} (K_{i,i+1}^t - K_{i-1,i}^t) - \Delta t [S(h_i^t) + S_D(h_i^t)]$$



**Figure 42.** Discretization of the soil profile for the numerical solution of the Richards' equation.

The hydraulic conductivity values of Eq. (7-22) are calculated using an explicit approximation, i.e.

using h-values from time level t. The approximation with respect to spatial derivative is taken as a fully implicit one leading to a convenient form of solution where the K(h)-values need to be calculated only once during each time step, which significantly reduces the computational burden. The same type of approximation has been used in the most recent version of the SWAP-model (van Dam et al. 1997; van Dam and Feddes 2000). The internodal hydraulic conductivities used in Eqs. (7-22) and (7-23) are defined as the arithmetic average between the nodes:

$$K_{i-1,i}^t = \frac{K(h_{i-1}^t) + K(h_i^t)}{2}$$

$$K_{i,i+1}^t = \frac{K(h_{i+1}^t) + K(h_i^t)}{2}$$
(7-24)

Eq. (7-23) is applied for all intermediate nodes  $i=2,\dots,N-1$  and appropriate boundary conditions need to be defined. In the upper boundary of the profile infiltration into the 1<sup>st</sup> node is calculated as an explicit approximation leading to a set of coefficients as given by (7-25):

$$\alpha_1 = 0$$

$$\phi_1 = -\frac{\Delta t K_{1,2}^t}{\Delta z_1 (z_2 - z_1)}$$

$$\beta_1 = C_{1,tot} - \phi_1$$

$$\varphi_1 = C_{1,tot} h_1^t - \frac{\Delta t}{\Delta z_1} (q_{inf} - K_{1,2}^t) - \Delta t [S(h_1^t) + S_D(h_1^t)]$$
(7-25)

Infiltration into the 1<sup>st</sup> node,  $q_{INF}$ , will be calculated by comparing several criteria and the following options will be calculated: 1° available pore space in the soil profile, 2° flux potential calculated using Darcy's law and assuming that at the soil surface soil water pressure head is zero and 3° available pore space in the 1<sup>st</sup> node plus the vertical flux downward from node 1 to node 2 (this option can restrict infiltration into very dry soil when the surface saturates first before the wetting front starts to advance). The value used in the model as  $q_{INF}$  is the smallest value from 1°-3°.

The purpose of the numerical solution adopted here is to calculate the recharge term for the groundwater flow models. Therefore, at the bottom boundary it is assumed that groundwater level is kept at a prescribed level,  $z_{WTD}$  (elevation from the reference level). The coefficients of the solution equation for node N are as follows:

$$\alpha_N = 0$$

$$\phi_N = 0$$

$$\beta_N = 1$$

$$\varphi_N = z_{WTD} - z_N$$
(7-26)

Eq. (7-26) implies in practice that soil water pressure head  $h_N$  at the bottom node equals the hydrostatic pressure. The soil profile must extend below the groundwater level to ensure that the lowest node is saturated throughout the computation.

Recharge term R(t) is calculated from Eq. (7-27):

$$R(t) = -\frac{K(h_{N-1}^t) + K(h_N^t)}{2} \left( \frac{h_N^t - h_{N-1}^t}{z_N - z_{N-1}} + 1 \right) \quad (7-27)$$

The boundary condition shown in (7-26) also means that with the solution it is possible to calculate the recharge term when  $z_{WTD}$  remains constant. In groundwater models the thickness of the unsaturated zone is not constant and therefore the solution shown here provides an approximate way of estimating the recharge term. It is necessary to calculate this model for several  $z_{WTD}$ -depths and provide different R-values for different zones of the model.

### 3.7.1 Approximation of the differential moisture capacity $C_{i,tot}$

In solving the Richards' equation, the differential water capacity  $C_{i,tot}$  in (7-23) has to be evaluated. Tests with numerical solution methods have shown that non-linear changes in C tend to produce mass conservation errors, whereas the non-linear changes in hydraulic conductivity mostly affect the internal water distribution (Jensen, 1983).

Karvonen (1988) has compared several methods to estimate C and the most accurate approximation for  $C_{i,tot}$  was:

$$C_{i,tot} = \frac{\theta_i^{t+1} - \theta_i^t}{h_i^{t+1} - h_i^t} \quad (7-28)$$

In fact  $C_{i,tot}$  calculated using equation (7-28) is the exact value which should be used when the change in soil moisture potential is  $h_i^{t+1} - h_i^t$  and the corresponding change  $\theta_i^{t+1} - \theta_i^t$  should be calculated. How to apply in the numerical solution an approximation of the form (7-28)? The solution method adopted here can be summarized as follows (Karvonen 1988):

1° At the beginning of each time step an explicit approximation of the change in water content is calculated for each nodal point. An estimate of the soil moisture content of each nodal point can be obtained, denoted by  $\theta_i^{\circ,t+1}$ .

2° Approximation of soil water potential for each nodal point can be calculated using new water content values estimated in stage 1° ( $h_i^{\circ,t+1}$  as a function of  $\theta_i^{\circ,t+1}$ )

3° Differential moisture capacities can be estimated now from (7-28) where  $\theta_i^{t+1}$  is replaced by  $\theta_i^{\circ,t+1}$  and  $h_i^{t+1}$  by  $h_i^{\circ,t+1}$ .

4° New pressure head values can be calculated by solving the Richards' equation. If convergence is not attained, a new iteration is necessary and within this new cycle the stages 1° through 3° are fulfilled again. At the end of the iteration procedure,  $h_i^{\circ,t+1}$  is equal to  $h_i^{t+1}$ .

The main merit of the procedure described above is the accurate calculation of mass balances. Hence, it is possible to use longer time steps and still maintain reliable water balance calculations. The validity of the proposed approximation method has been given by Karvonen (1988):

## 4 SECTION VI: Water and solute movement in soils and plants

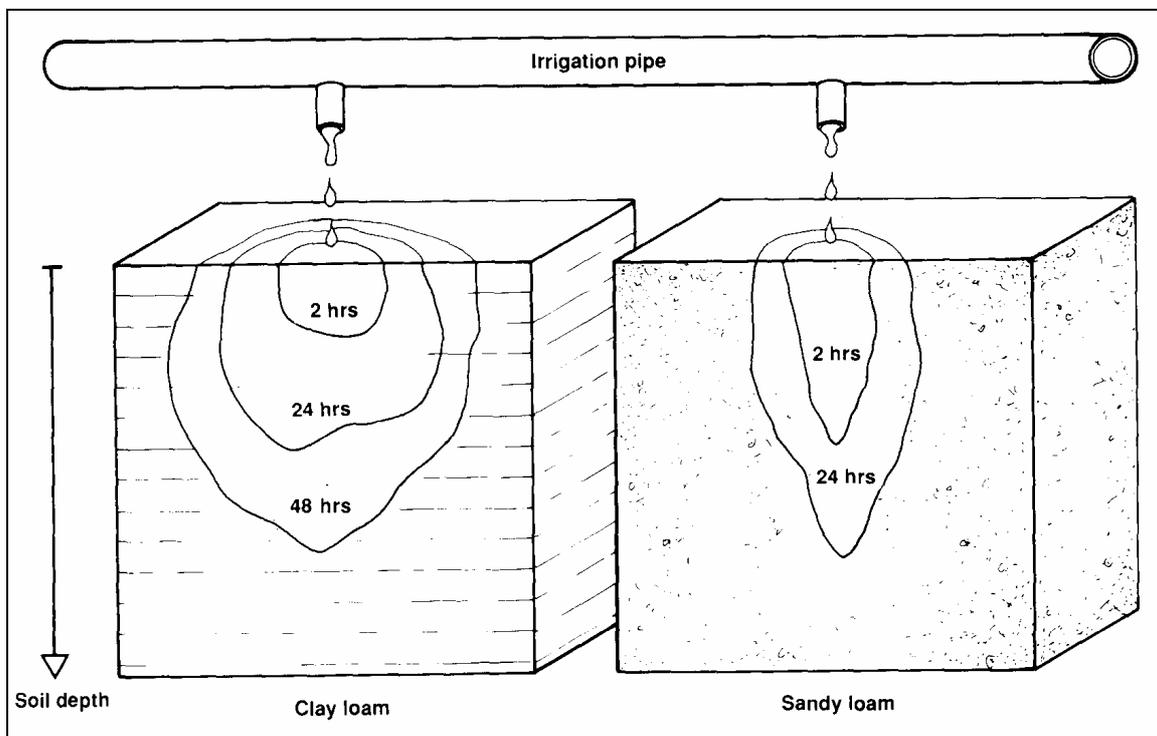
### 4.1 The principles of water movement in soil

In this section we will outline some of the basic principles of water flow in soil before examining the main processes involved in water infiltration, redistribution and drainage of field soils. Water movement from the soil into the plant and the atmosphere will also be described.

### 4.2 Saturated flow

A soil is said to be saturated when all the pores are full of water and there is no air present. Saturated conditions may occur throughout the soil profile or in isolated parts of the profile. In a poorly drained soil, for instance, the subsoil may be saturated and the topsoil unsaturated. Alternatively, after heavy rainfall the topsoil may be temporarily saturated whilst the subsoil is unsaturated.

Under normal drainage conditions the direction of water flow is predominantly downward, but horizontal flow can also occur in some circumstances. However, since it is not assisted by gravity, the rate of horizontal flow is generally less than the vertical flow. Figure 43 shows a typical pattern of flow through soil resulting from water supplied, for example, by an irrigation supply.



**Figure 43.** Comparison of water flow through a clay loam and a sandy loam soil. Vertical water movement from the point of infiltration is more rapid in the sandy loam. Extensive lateral flow of water only occurs in the clay loam soil.

The rate of water flow through soil can be described by Darcy's Law which states that the flux of water  $q$  is proportional to the hydraulic gradient multiplied by the conductivity or permeability of the soil:

$$q = \frac{Q}{A} = -K \frac{\Delta H}{L} \quad (7.1)$$

where  $Q$  is the rate of discharge through a cross-sectional area  $A$ ,  $\Delta H/L$  is the hydraulic gradient and  $K$  is the hydraulic conductivity of the soil. This is illustrated in Figure 44 which shows a column of soil with water ponded on the top and drainage occurring through the saturated soil. In this illustration the driving force is the hydraulic head gradient which is the ratio of the hydraulic head drop ( $\Delta H$ ) to the column length ( $L$ ).

A more generalized expression of Darcy's Law is in the differential form and for flow in  $one(x)$ -

dimension Equation 7.1 becomes:

$$q = -K \frac{dH}{dx} \quad (7.2)$$

The hydraulic conductivity of the soil (K) represents the ability of the soil to transmit water and is dependent on the nature of the pore system in the soil. It is usually expressed in units of  $m s^{-1}$ . It is equal to the rate of flow per unit area of soil ( $q$ ) when the hydraulic gradient is unity. The negative sign in Equation 7.2 is there by convention to indicate that the direction of water flow is from regions of higher potential to regions of lower potential.

The saturated hydraulic conductivity is an important measure of the drainage capacity of a soil. A sandy soil for example may have a saturated hydraulic conductivity ( $K_s$ ) value of  $10^{-5} m s^{-1}$  whilst a clay soil with poor structure could have a K, value as low as  $10^{-8} m s^{-1}$ . From Equation 7.1, when there is no pressure or suction gradient (i.e. a gravity gradient only), the maximum rate of flow ( $q$ ) through the sand would be  $36 mm hr^{-1}$  whilst the clay would have a maximum rate of only  $0.036 mm hr^{-1}$ . Therefore, although a rainfall of  $5 mm hr^{-1}$  would cause ponding on the clay it would quickly drain through the sand.

The rate of flow per unit area ( $q$ ) represents the volume of water discharged through a unit cross-sectional area per unit time. It can therefore have units of  $m^3 m^{-2} s^{-1}$ , or simply  $m s^{-1}$ . This is often referred to as the 'flux' because the actual flow velocity of water within the pores is extremely variable, depending on the sizes of the soil pores.

The rate of volume flow of water through a soil pore (Q) is proportional to the fourth power of the radius (r) of the pore:

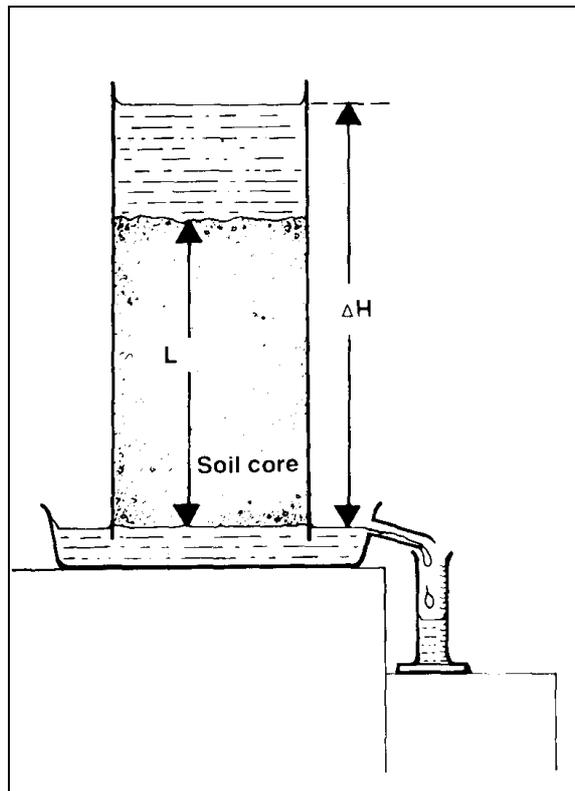
$$Q = \left( \frac{\pi r^4}{8\eta} \right) \left\{ \rho_w g \Delta H / L \right\} \quad (7.3)$$

where  $\eta$  is viscosity,  $\rho_w$  is the density of water,  $g$  is the acceleration due to gravity, and  $\Delta H / L$  is the hydraulic gradient along the length (L) of the pore.

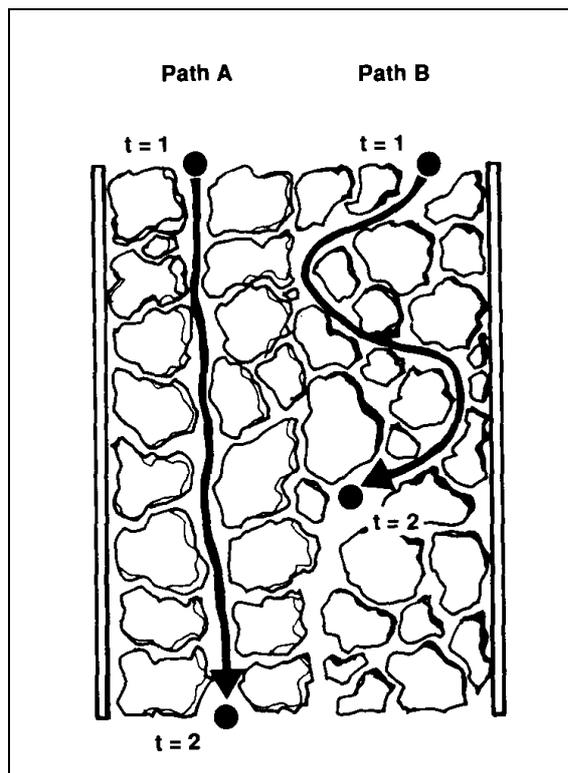
Thus the actual size of pores in the soil can have a large effect on its hydraulic conductivity. For example, a pore of 1 mm radius will have a flow rate 10000 times faster than a pore of 0.1 mm radius for the same head drop. This deserves some thought as it explains the relative importance of macropores compared to micropores in soil drainage. For example 10 000 pores of 0.1 mm would be required to provide the same contribution to hydraulic conductivity as one single pore of 1 mm radius.

Since soil structure has a large influence on the pore size range of a soil, it therefore has a large effect on the hydraulic conductivity of the soil. A soil with a tightly packed 'platy' structure, for example, will have a lower hydraulic conductivity than one with an open, porous 'granular' type structure. Earthworms and other soil organisms are also important since they can create relatively large channels in soil which can allow rapid water flow under saturated conditions. Similarly, plant roots may provide relatively large pathways through the soil which increase the hydraulic conductivity.

The geometry of the pore system is also important to the hydraulic conductivity. If the pores are orientated vertically and provide a relatively straight route from the top to the bottom, as in Figure 45a, then the rate of flow through the soil will be rapid. However, this rarely occurs and it is more likely that the pores take a more tortuous route, as shown in Figure 45b. The more tortuous the pathway through the pore system the slower the rate of flow.



**Figure 44.** Water flow through a column of saturated soil under a hydraulic head gradient ( $\Delta H/L$ ).



**Figure 45.** Pore tortuosity affects the rate of water flow. Over the same time interval ( $t_2-t_1$ ) the water molecule in the straight vertical pore (Path A) travels a greater vertical distance than the molecule moving through the more tortuous pore (Path B).

The hydraulic conductivity may not necessarily remain constant for any particular soil. For example, if the soil structure is unstable then the pores can become blocked when the aggregates break down and this will cause the hydraulic conductivity of the soil to decrease

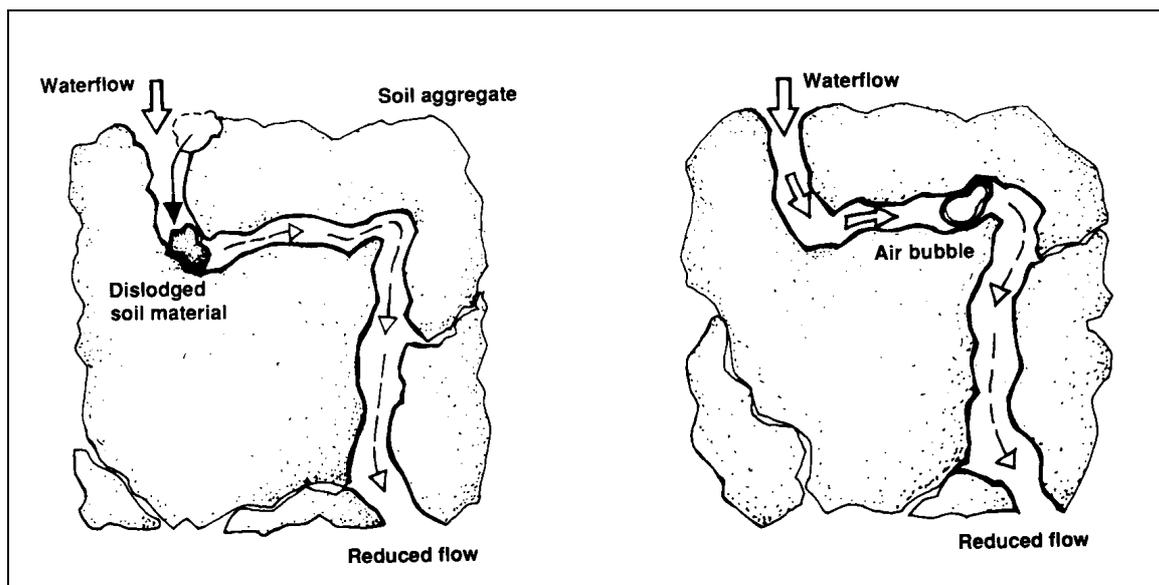
with time. Entrapped air in soil pores, as shown in Figure 46, can also constrict the pore and reduce the rate of flow. The hydraulic conductivity may also change with a change in the concentration of solutes in the soil solution.

Typical saturated hydraulic conductivity values can be grouped together to indicate the relative drainage properties of soil (Table 10). The use of these values in relation to the design of drainage schemes will be considered later but for the moment it is worthwhile noting the limitations on land use imposed by the  $K_s$  values, and the types of soils these relate to.

**Measurement of saturated hydraulic conductivity-** There are a number of methods which can be used to measure the saturated hydraulic conductivity of soil. Details of methods are described in the references listed at the end of this chapter and two common methods are outlined here:

(i) **Permeameters.** Soil cores are taken in the field and returned to the laboratory where they are installed in a permeameter, similar to that shown schematically in Figure 47. Re-packed cores of soil may also be used but have less meaning than undisturbed field cores. The volume of water flowing through the soil core is measured at short time intervals and the  $K_s$  value calculated from this and the size of the hydraulic gradient. The size of the hydraulic gradient can be adjusted by raising or lowering the height of the water reservoir.

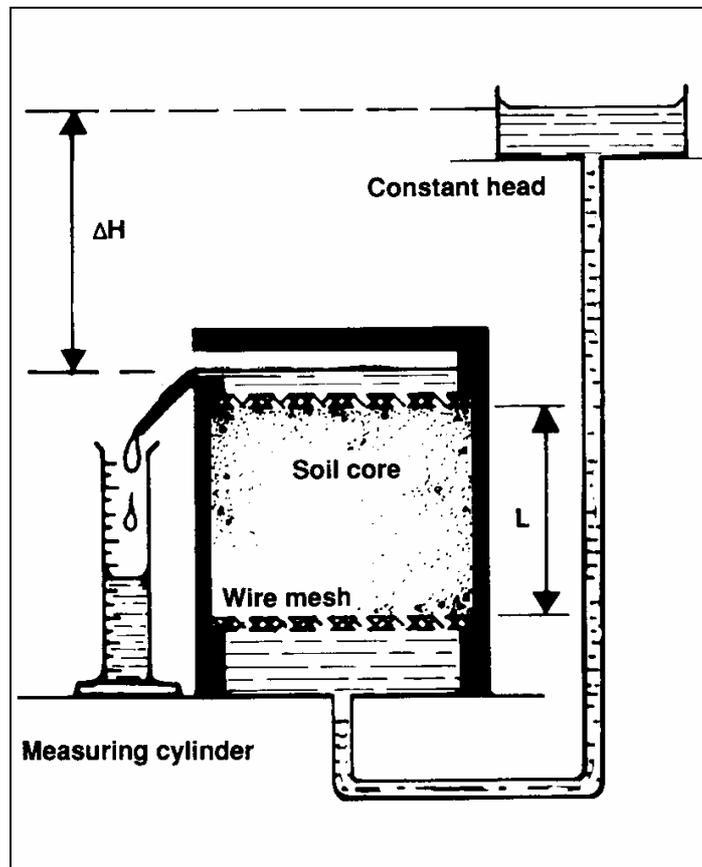
(ii) **Auger hole methods.** In a saturated soil the hydraulic conductivity can be measured by drilling a hole and measuring the rate at which the water rises in the hole (Figure 48a). This is often termed a 'piezometer' method. To measure the  $K_s$  value of soil above the water table, water can be supplied by a constant head device to fill the bottom of the hole and the rate at which this drains away can then be measured (Figure 48b)



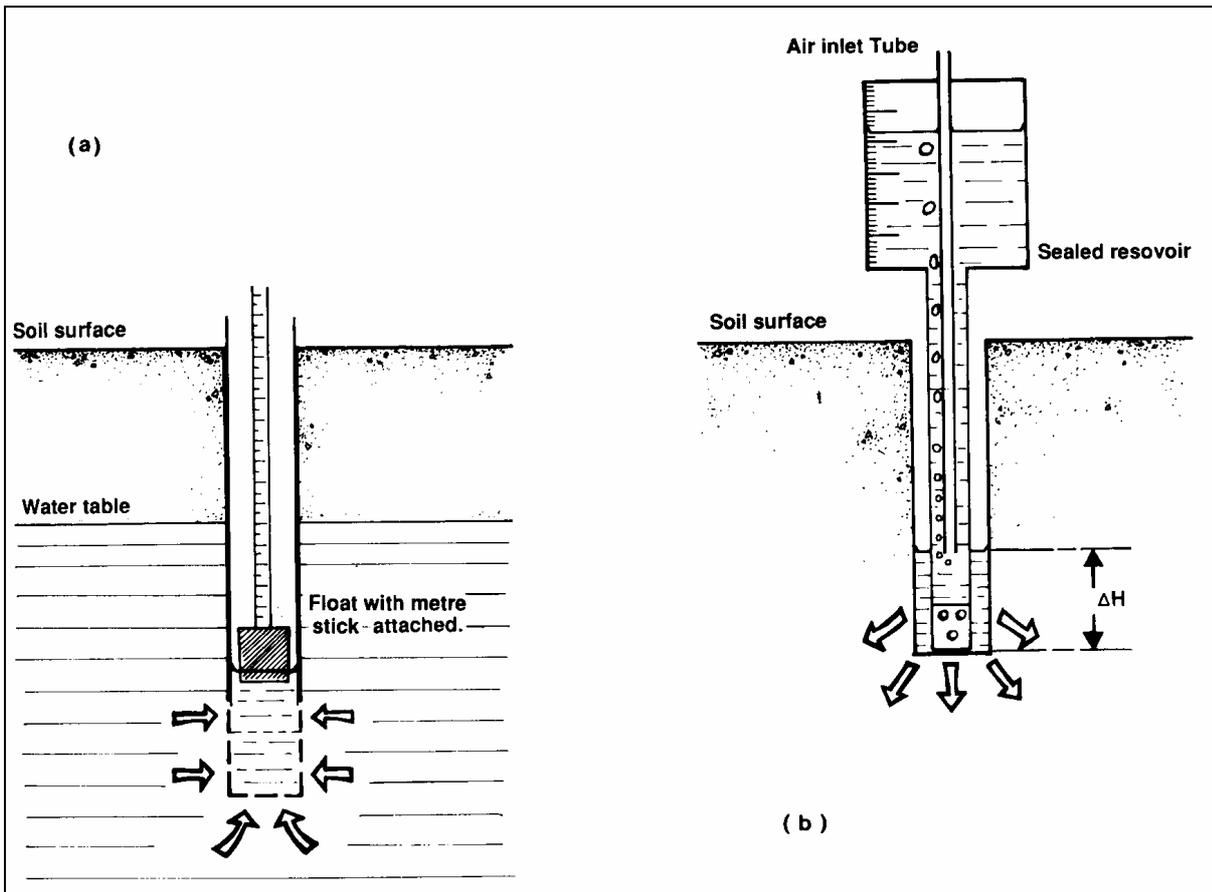
**Figure 46.** Soil pores may be restricted by dislodged soil material or even entrapped air bubbles.

**Table 10.** Hydraulic conductivity of saturated soil (adapted from Marshall and Holmes, 1979).

Range		Soil	Effects on land use in agriculture
$m\ S^{-1}$	$mm\ hr^{-1}$		
$<10^{-7}$	$<0.36$	Fine textured and poorly aggregated soil. Mottles common in subsoil.	Poor drainage restricts use to shallow rooted plants.
$10^{-7}$ to $10^{-4}$	0.36 to 360	Wide range of soil textures and structures.	Most land used for agriculture and horticulture in this range.
$>10^{-4}$	$>360$	Coarse textured soil.	Soil with poor water holding capacity restricts use to deep rooted plants.



**Figure 47.** Constant head permeameter for measuring the hydraulic conductivity of a soil core.



**Figure 48.** Field methods for measuring hydraulic conductivity: (a) Piezometer hole for measurements below the water table; (b) Constant-head well permeameter for measurements above the water table.

### 4.3 Unsaturated flow

Soils in the field are normally unsaturated, that is to say, not all the soil pores are filled with water. Water flow under unsaturated conditions is slow compared with that in saturated soil. However, it is nonetheless essential for the supply of water to plants and organisms in the soil.

#### 4.3.1 Liquid flow

In a saturated soil the driving force is the hydraulic gradient which acts to effectively push water through the soil. In an unsaturated soil there is no water in the large soil pores and the water which is present is at a negative potential (i.e. it is held within the soil matrix by adhesion and cohesion). In an unsaturated soil it is the potential gradient from one zone to another which acts as the driving force for water movement. Remember, water will always try to come to an equilibrium in soil by flowing from a zone of high potential to a zone of lower potential.

The driving force in an unsaturated soil is in fact many times greater than the gravitational potential. Water will not only flow downwards but can flow upwards or sideways. Even if the majority of the soil pores are empty, flow can still occur through water films attached to the edges of soil particles as well as in the vapour phase (see next section). During unsaturated flow both the volume and potential energy of the water at any one location within the soil may change. Since the rate of flow through a pore is proportional to the fourth power of the radius (Equation 7.3) any change in the sizes of pores filled with water will affect the hydraulic conductivity.

For any particular soil there is a relationship between the hydraulic conductivity and the matric potential. An example of this is shown in Figure 49. For example, when the water potential of the soil is high (e.g. near saturation) then the hydraulic conductivity is high. However, when the water potential decreases, the sizes of pores which are waterfilled also decreases resulting in a lower hydraulic conductivity. Hydraulic conductivity can therefore be thought of as a function of

water content or matric potential.

The relationship between conductivity and potential can vary considerably between soils as indicated by the shapes of the curves shown in Figure 49. The sandy soil has the greater hydraulic conductivity near saturation but only conducts water slowly when in an unsaturated state, owing to its relative lack of smaller pores. The curve for the clay soil shows a lower hydraulic conductivity near to saturation but a relatively higher conductivity when unsaturated.

Darcy's Law can be extended to describe unsaturated flow,

$$q = -K(\psi) \frac{dH}{dx} \quad (7.4)$$

where  $dH/dx$  is the hydraulic gradient which includes both a suction and gravitational component, and  $K(\psi)$  represents the relation of conductivity to matric potential.

Because of the effects of hysteresis on the soil moisture characteristic curve, and thus on the hydraulic conductivity, a more useful form of Equation 7.4 is:

$$q = -K(\theta) \frac{dH}{dx} \quad (7.5)$$

where  $K(\theta)$  represents the relation of conductivity to volumetric water content, which is less affected by hysteresis.

In unsaturated soil it is likely that the water content changes with both time and location. Indeed, this will be the most common condition found in field soils. In order to deal with these transient conditions an equation representing the law of conservation of matter, called the **continuity equation**, must be combined with the flow equation. In this case, the continuity equation simply states that the rate of change of water content of a small volume element, or cube of soil, is equal to the difference between the 'flux in' and the 'flux out', plus any sources of water and minus any sinks of water within the cube (Figure 50). Ignoring any sources or sinks of water within the cube the continuity equation can be expressed as

$$\frac{\partial \theta}{\partial t} = - \left( \frac{\partial q_x}{\partial x} + \frac{\partial q_y}{\partial y} + \frac{\partial q_z}{\partial z} \right) \quad (7.6)$$

where  $q_x$ ,  $q_y$  and  $q_z$  represent the fluxes in the x, y, and z directions shown in Figure 49.

When flow is restricted to *one-(x-)* dimensional horizontal movement, the gravitational component of the hydraulic gradient is zero and Equations (7.4) and (7.6) can be combined to give:

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial x} \left( K \frac{\partial \psi}{\partial x} \right) \quad (7.7)$$

where only the matric potential gradient appears.

In order to consider water content rather than potential, Equation 7.7 can be re-written as:

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial x} \left( D \frac{\partial \theta}{\partial x} \right) \quad (7.8)$$

where  $D$  (called **hydraulic diffusivity**) is the ratio of hydraulic conductivity to the so-called 'water capacity' of the soil (where the latter equals the slope of the moisture characteristic curve at any value of  $\theta$ ).

Although these equations can be used to solve a number of practical flow problems this is beyond the scope of this text and the reader is referred to the more advanced texts listed at the end of this chapter. We will now consider methods of measuring the unsaturated flow rate in soil.

#### 4.3.2 Measurement of unsaturated flow

Laboratory methods of measuring the unsaturated hydraulic conductivity of soil cores are available. However field methods are more appropriate because they allow for a larger and more realistic scale

of observation. Field methods generally produce unsaturated flow conditions by restricting the rate of water addition to the soil. This can be done by using a precision controlled sprinkler to apply water uniformly over the soil surface and at a rate which is lower than the saturated hydraulic conductivity. Alternatively, the rate of water addition to the profile can be restricted by producing a layer of puddled soil on the surface. The matric potential and hydraulic conductivity of the restrictive layer will not allow the larger soil pores of the soil below to fill and will thus exclude them from flow.

'Tension infiltrometers' have been developed which only supply water under suction and thus ensure that soil macropores do not fill with water and that unsaturated flow only can occur (for details see texts listed at the end of this chapter).

A further method involves frequent monitoring of the changes in soil water content (by neutron probe) and water potential (by tensiometers) which occur during the drainage of a uniformly saturated soil profile.

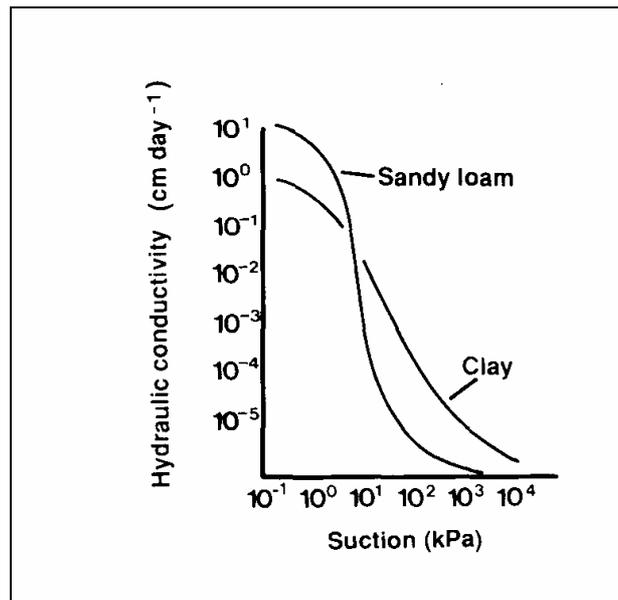


Figure 49. Relationship between hydraulic conductivity and suction for a clay soil and a sandy loam.

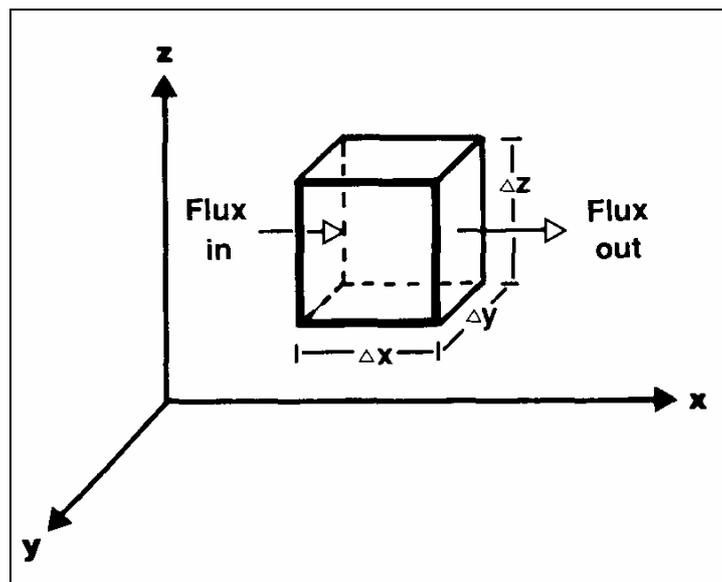


Figure 50. The continuity principle states that the change in water content of a hypothetical cube of soil must equal the difference between the 'flux of water in' and the 'flux of water out' of the cube in each of the x, y and z directions.

### 4.3.3 Vapour flow

Soil air generally has a **relative humidity** of nearly 100%. Despite this the amount of water normally held in vapour form is very small. Vapour flow may be important in the surface soil during evaporation, or as a 'bridging' mechanism for unsaturated liquid flow between isolated zones of waterfilled pores. Vapour flow does not usually make a significant contribution to the total amount of water which moves towards plant roots.

Vapour flow in soil occurs in response to vapour pressure differences. It can be considered as a diffusion mechanism in which the driving force is the vapour pressure gradient:

$$q_v = -D_v \frac{\partial \rho_v}{\partial x} \quad (7.9)$$

where  $q_v$  is the vapour flux,  $D_v$  is the water vapour diffusion coefficient,  $\rho_v$  is the **vapour density**, or vapour concentration.

Temperature has a large effect on vapour pressure. In a sealed container, for example, the pressure of the gas inside increases as the temperature rises. The same principle applies to soil. A change in temperature from 10 to 30°C can cause a three fold increase in saturated vapour pressure. This can produce a vapour pressure gradient in the soil resulting in vapour flow. During the day, water vapour will flow away from the warm soil surface towards the cooler subsurface layers (i.e. from a zone of higher pressure in the warm surface soil to a zone of lower vapour pressure in the cooler layers below). This effect can help to reduce the day time evaporation losses from soil by producing a dry surface layer which acts as a 'mulch' preventing further water losses. Vapour transport also causes the subsurface layers to develop a higher water content. At night when the surface is cooler there is a return flow of water vapour and possibly liquid water, due to the induced potential gradient, and this rewets the surface soil.

## 4.4 Infiltration

### 4.4.1 Infiltration processes

The rate at which water can enter the soil surface is called the **infiltration rate**. The infiltration rate is important, firstly, because it affects the rate at which a soil may 'recharge' with water and, secondly, because it affects the likelihood of surface runoff and hence erosion occurring during heavy rain or irrigation.

The infiltration rate of a soil may be described by the following equation:

$$i = St^{1/2} + K_i t \quad (7.10)$$

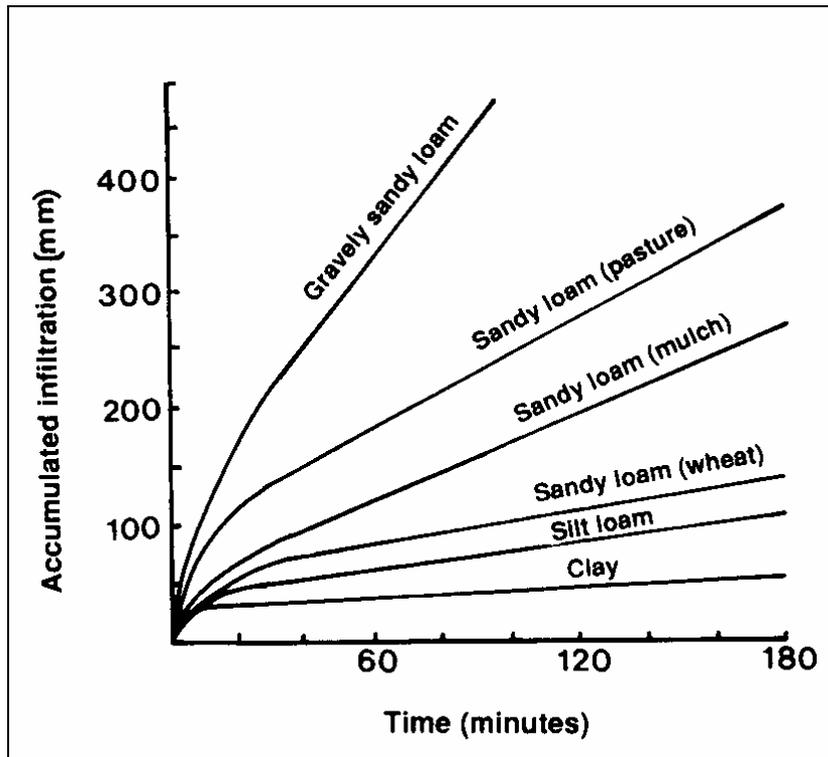
where  $i$  is the quantity of water infiltrating a unit crosssectional area of soil ( $i = Q/A$ ),  $S$  is the **sorptivity** of the soil (which indicates the capacity of the soil to absorb water), and  $K_i$  is the ability of the soil to transmit water (conductivity).

In the early stage of infiltration into an unsaturated soil, sorptivity ( $S$ ) is the dominant term in Equation 7.10 and can be approximated from the measured infiltration rate at early times in Figure 51. The early stage of infiltration thus represents a rapid flow of water into empty pores in the soil. This flow is in response to a matric potential gradient as well as the gravitational gradient and is therefore influenced by the initial soil water content (or potential). As the initial water content increases, the sorptivity and thus the early infiltration rate decreases.

At later stages of infiltration the parameter  $K_i$  (conductivity) becomes more important and eventually becomes the only significant term in Equation 7.10. At this stage the soil is effectively saturated and the value of  $K_i$  should be almost equivalent to the saturated hydraulic conductivity of the soil ( $K_s$ ). The value of  $K_i$  can be approximated from the measured rate of infiltration at later times in Figure 51, that is, once steady state has been established.

The final infiltration rate will be affected by many of the soil properties which have already been

outlined as affecting the hydraulic conductivity (e.g. texture, structure, pore characteristics). It is worth emphasizing that aggregate stability and the presence of swelling clays can have a large effect on infiltration.



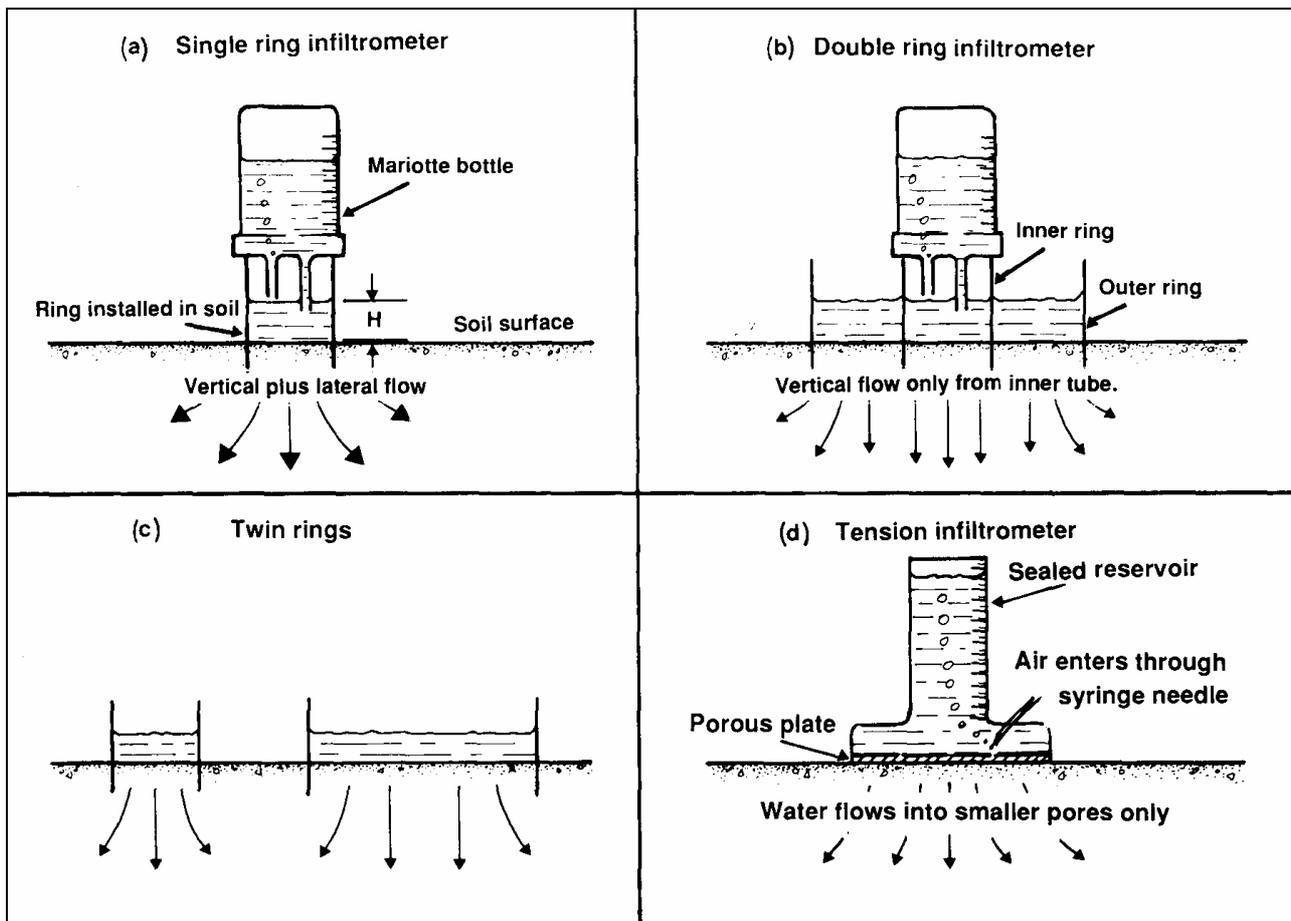
**Figure 51.** Typical infiltration curves for different soil textures and plant covers (adapted from Taylor and Ashcroft, 1972).

The surface soil is subject to raindrop impact as well as other factors, such as traffic and stock damage, which can reduce the infiltration capacity of the soil. Raindrops can impart considerable impact on the surface soil aggregates and these can easily break down if they are weak and unstable. Soils which have been heavily cultivated and are low in organic matter are particularly susceptible to aggregate breakdown and surface crust formation. A comparison of the infiltration rates of a cultivated soil with an uncultivated permanent pasture in Figure 51 illustrates the effect of over-cultivation. The presence of a layer of dead plant material, called a 'mulch', helps to reduce raindrop impact and maintains a high infiltration rate (Figure 51).

#### 4.4.2 Measurement of infiltration

The infiltration rate of a soil can be measured using a 'ring infiltrometer'. This consists of a metal ring (approximately 300 mm diameter), which is pushed about 20 mm into the soil, and a Mariotte bottle, which provides a constant head of water ponded on the soil within the ring (Figure 52a). Since a single ring allows water to move laterally, as well as vertically, a double ring infiltrometer is often used to reduce the amount of lateral flow (Figure 52b). The rate of water supply required to maintain a constant head (usually 10-20 mm) within the inner ring is measured and the infiltration rate calculated on the basis of the enclosed inner area.

Measurements of infiltration rate, field saturated hydraulic conductivity, and sorptivity can also be made using a system of 'twin-rings'. In this method the smaller ring is placed outside the larger one and the average field saturated hydraulic conductivity calculated from the results of a series of 'twin-rings' (Figure 52c). The unsaturated infiltration rate and the sorptivity of soil can be measured using a 'tension infiltrometer' device (Figure 52d).



**Figure 52.** Field methods for measuring infiltration rate: (a) Single-ring infiltrometer; (b) Double-ring infiltrometer; (c) Twin-ring infiltrometers; (d) Tension infiltrometer.

## 4.5 Water redistribution and drainage

### 4.5.1 Processes in a uniform soil

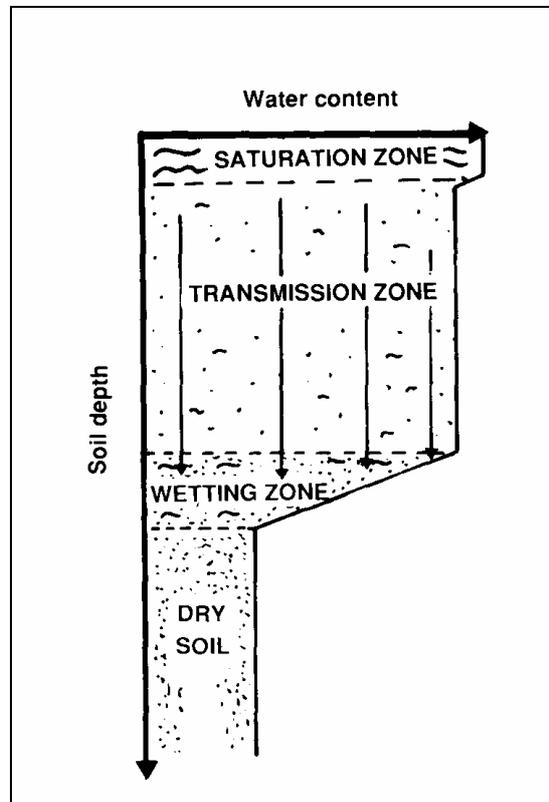
The processes involved in the redistribution of water following infiltration in the soil profile are complex. They can, however, be visualized in a simple manner by examining the changes in profile water = tent which would occur in a 'uniform' soil following an intense application of irrigation or rainfall. After a short period with water ponded on the soil surface, the profile has a range of water contents, such as those shown in Figure 53. The upper part of the profile is saturated with water whilst the lower part of the profile is unaltered from its initial water content. The 'transmission zone' in the middle of the profile has a uniform water content, almost at saturation, and a uniform matric potential of almost zero. Gravity is the main driving force for water flow in the transmission zone, and the rate of water flow, which is about equal to the saturated hydraulic conductivity, can be described by Darcy's Law. The 'wetting zone' shows a large change in water content due to the large matric potential gradient within this zone, allowing only the small pores, of low hydraulic conductivity, to fill with water. The rate of water flow is restricted until the micropores are filled and the larger pores are able to take part in flow.

In a free draining situation, once the profile is saturated and the application of water has ceased, then rapid drainage will continue for a period of perhaps one or two days. This represents the drainage of soil

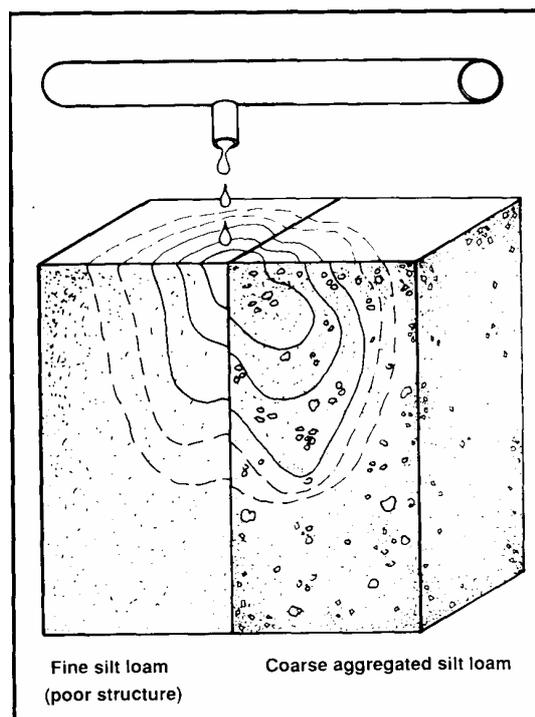
macropores which are unable to hold water against gravity. As discussed previously, once this rapid drainage has slowed to negligible levels, the soil is said to be at 'field capacity'.

#### 4.5.2 Processes in field soils

Soil profiles are, however, rarely uniform and the presence of cracks, earthworm burrows and root channels cause the redistribution process to be more erratic than that represented by Figure 53. Although difficult to describe quantitatively, the essential processes of water redistribution in non-uniform structured soils with cracks and channels can be appreciated by studying flow patterns in simplified soil models. In each of the models presented (Figures 54-55) the lines indicate the positions of the wetting front after increasing intervals of time.

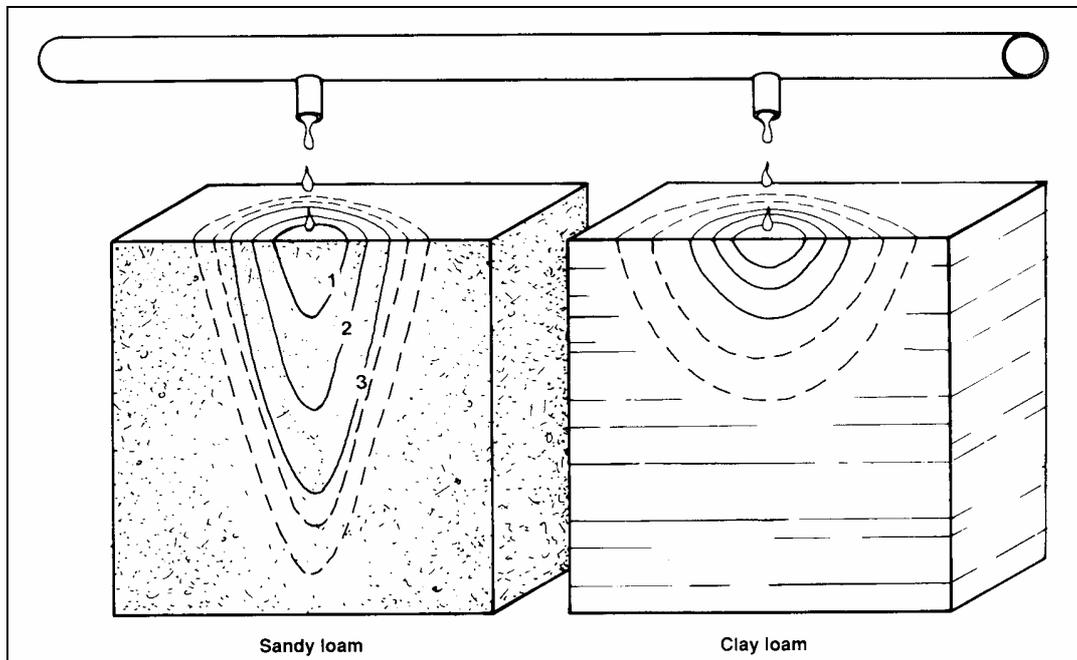


**Figure 53.** Redistribution of ponded water in a uniform dry soil profile (adapted from Bodman and Coleman, 1944).



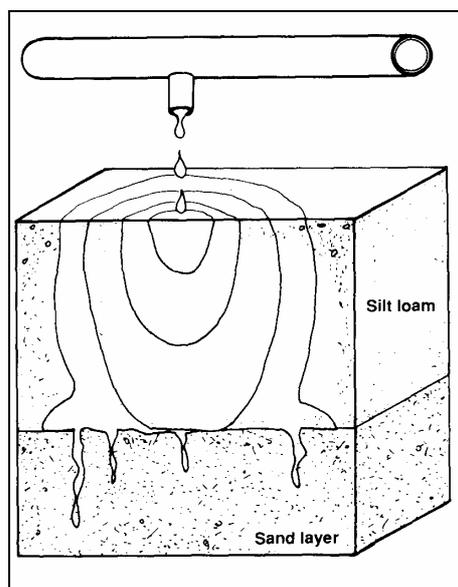
**Figure 54.** Water redistribution in soils with different structures. Whilst water is being applied, flow is faster through the

well structured soil than through the poorly structured soil. The packing arrangement of the structural aggregates leaves open spaces which provide numerous pathways for rapid water flow. The aggregated soil thus has a greater saturated hydraulic conductivity than the weakly aggregated silt loam with only small pores. Under unsaturated conditions however, when the interaggregate pores do not contain water, the rate of flow is faster in the weakly structured soil (as shown by the broken lines). The soil aggregates also absorb water and this retards the rate of progress of the wetting front in the strongly structured soil. Variations in forms of soil structure, as well as degree of structure development, will affect the rate of water flow through the soil as will be described further.



**Figure 55.** Water redistribution in soils of different texture. The positions of the wetting fronts after similar time intervals ( $t = 1, 2, 3$ ) show that whilst water is being applied, the rate of flow is faster through the sandy loam than through the clay loam soil. The sandy soil has a greater proportion of large pores which gives it a greater hydraulic conductivity whilst at or near to saturation.

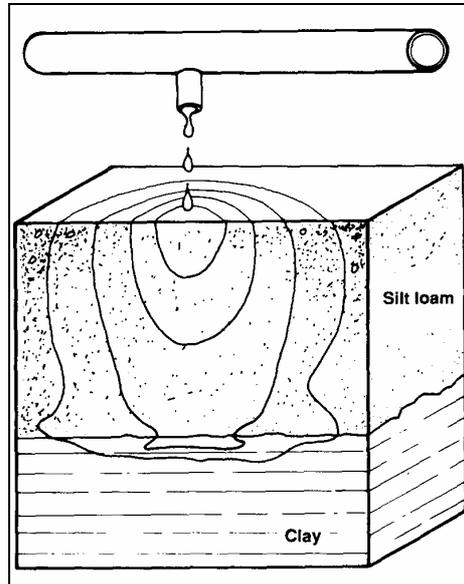
Once the water application ceases the flow in the sandy loam slows to less than that in the clay loam. This is indicated by the broken lines which represent the wetting front at longer time intervals after the application has ceased. Water continues to flow out of the wet zone, which has a high matric potential, into the dry soil, which has a lower (more negative) matric potential. The hydraulic conductivity of the finer textured soil is greater than the sandy soil at lower matric potentials and the wetting fronts are thus spaced further apart.



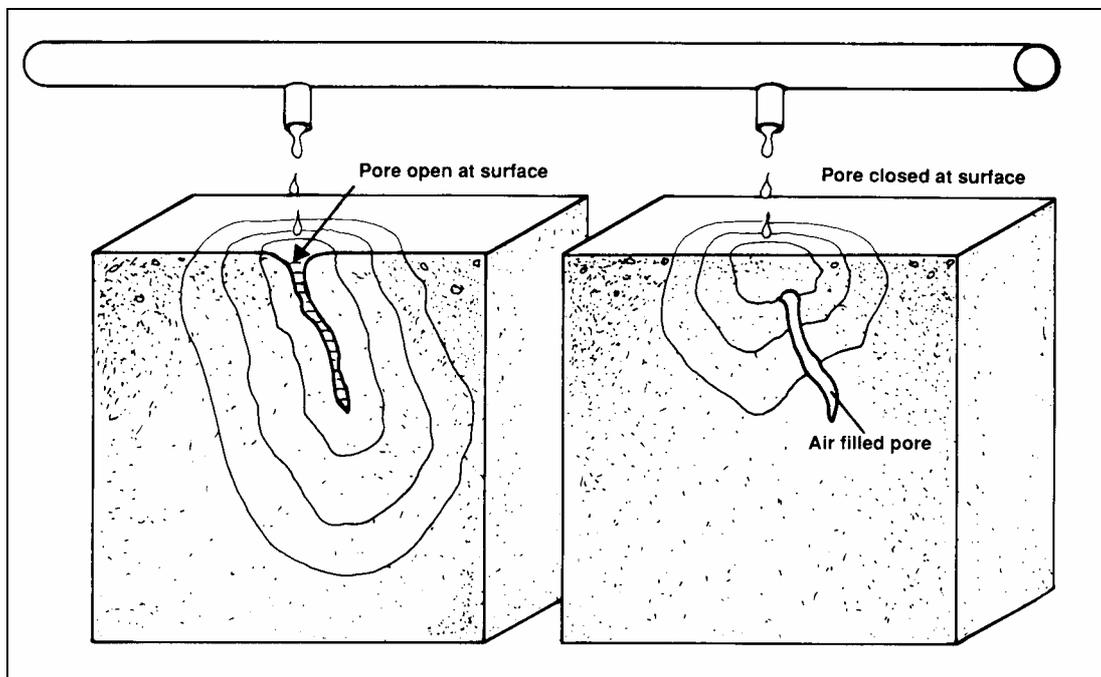
**Figure 56.** Water redistribution in soil with an underlying sand layer. The wetting front will move uniformly through the silt loam until it reaches the sand layer. On encountering the boundary, water in the silt loam side is initially at a fairly low

potential, so that the large rapidly-transmitting pores in the sand remain empty. As water builds up above the boundary, the water potential increases to a level which enables these larger pores to fill, allowing water to begin to flow in the sand layer.

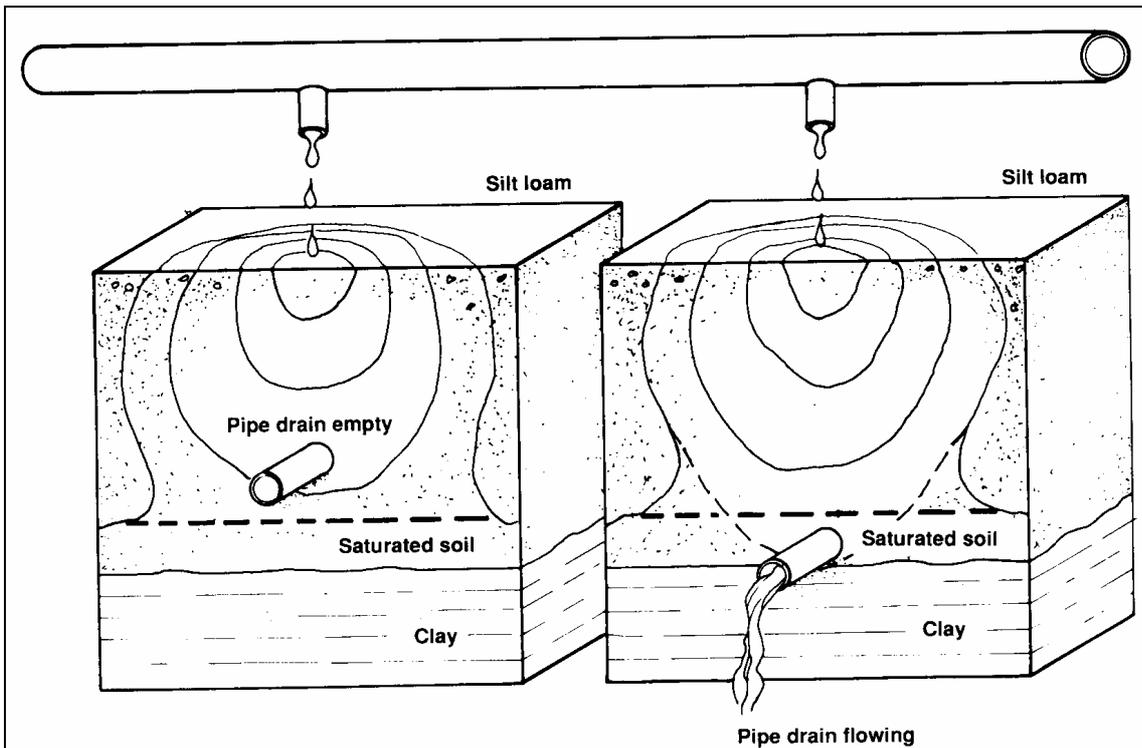
Similar effects may also occur at a boundary between soil and underlying layers of stones and gravels. These abrupt texture, or more precisely pore size, changes act in effect as 'check valve' mechanisms which allow water through when the soil above is saturated but assist in retaining water in the profile at other times.



**Figure 57.** Water redistribution in soil with an underlying clay layer. Soils which have a clay layer in the profile are commonly recognised as being poorly drained. Water movement into a dry clay layer from the soil above is initially rapid. However, once wet, the low hydraulic conductivity of the clay layer restricts the rate of water movement through the profile. Water therefore 'perches' above the clay layer.



**Figure 58.** Water redistribution in a soil with a vertical crack or channel. When water is ponded on the soil and the crack is open at the surface, then water will be conducted rapidly to depth in the soil. When the crack is not open at the surface it will only fill with water once the surrounding soil is almost saturated.



**Figure 59:** Water redistribution in a soil with a drainage pipe. Water cannot enter the pipe drain when it is placed above the level of the water table (i.e. water will not flow from a position of low potential in the soil to a position of higher potential). Water will only enter the drain when it is placed within the saturated zone (below the water table) and if there is sufficient hydraulic head.

## 4.6 Solute movement

When water moves through soil it generally transports solutes with it. This can lead to leaching losses of applied fertilizers and of native soil components. Although it is an important mechanism for soil development, leaching can also result in a reduction of soil fertility and the pollution of surface and groundwaters, as will be described in later chapters.

### 4.6.1 Principles

Solute movement occurs through a combination of three main mechanisms: convection, diffusion and dispersion. Convective transport - Convective transport results from the movement of solutes with the mass flow of water in soil. The convective flux of solutes can therefore be described by a modified form of Darcy's Law:

$$J_c = qc = -c \left( K \frac{dH}{dx} \right) \quad (7.11)$$

where  $J_c$  is the convective solute flux,  $c$  is the solute concentration,  $q$  is the water flux,  $K$  is the hydraulic conductivity and  $dH/dx$  is the hydraulic gradient.

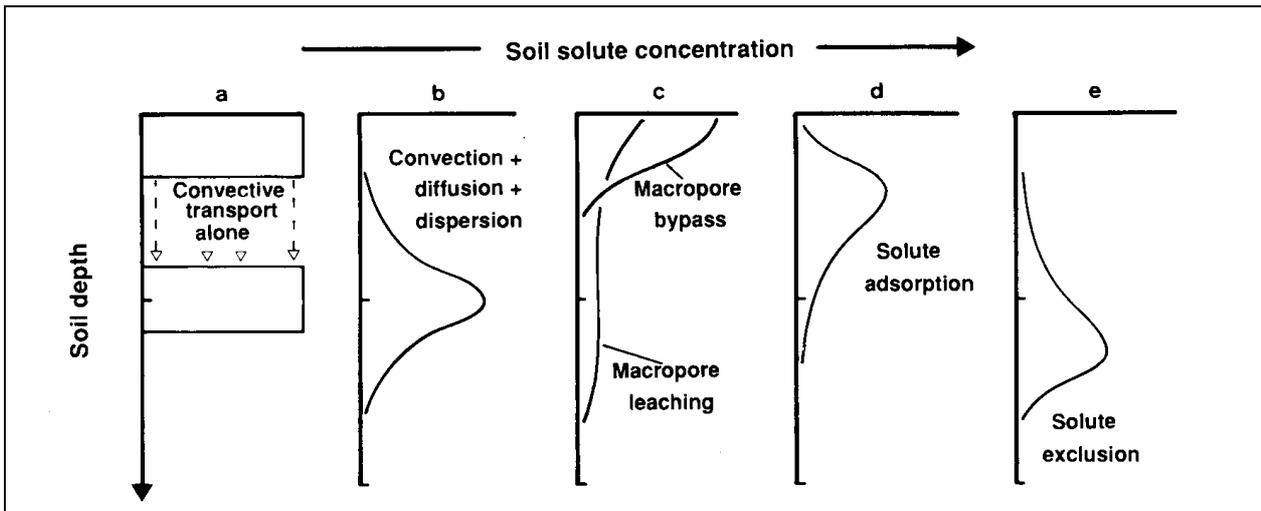
The distance of transport per unit time by convection depends on the average pore water velocity,  $U$ , where:

$$U = q/\theta_v \quad (7.12)$$

and  $\theta_v$ , is the volumetric water content.

Convective transport of a band of fertilizer is illustrated in Figure 60a and is often referred to as **piston displacement**. In reality, however, the solute does not remain as a sharp band but tends to spread throughout the profile (Figure 60b) due to the processes of **diffusion** and **hydrodynamic**

dispersion.



**Figure 60.** Schematic diagram of the various components of leaching: (a) convective transport alone; (b) convection - diffusion - dispersion; (c) macropore bypass and macropore leaching; (d) solute adsorption; (e) solute exclusion (adapted from Cameron and Haynes, 1986).

#### 4.6.2 Diffusive transport

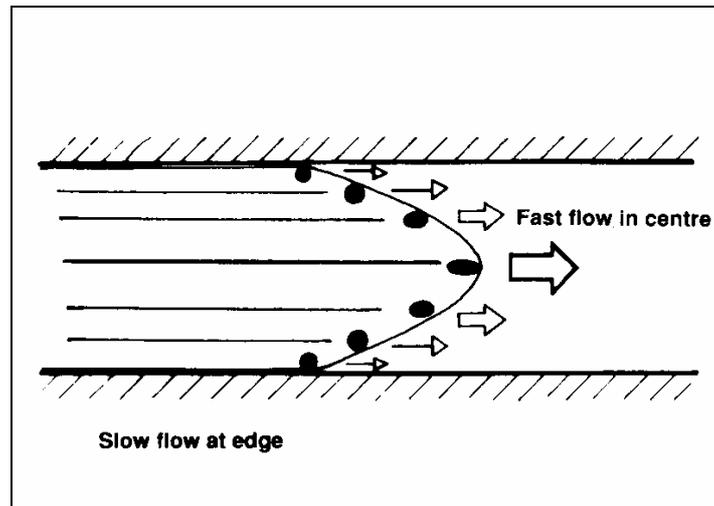
When there is an uneven distribution of solutes in a solution there is a diffusive flux of solute from areas of high concentration to areas of lower concentration. Movement by diffusion can be described by:

$$J_d = -D_s (\theta) \frac{dc}{dx} \quad (7.13)$$

where  $J_d$  is the rate of diffusion,  $D_s$  is the diffusion coefficient of the solute in soil and depends on the soil moisture content ( $\theta$ ), and  $dc/dx$  is the solute concentration gradient.

#### 4.6.3 Hydrodynamic dispersion

The mechanical action of a solution flowing through soil causes mixing and tends to equalize the solute distribution by a process called hydrodynamic dispersion. This process enhances the dispersive effect of diffusion and during flow it usually completely masks it. Hydrodynamic dispersion occurs because: (i) the flow velocity within a single pore is not uniform (Figure 61), (ii) the large variation in pore size in soil causes an extremely wide range of pore water velocities, and (iii) the tortuosity of pores results in a range of flow path lengths.



**Figure 61.** Flow velocity within a soil pore is uneven.

#### 4.6.4 Combined convective-diffusive-dispersive transport

In order to account for transient-state conditions, in which fluxes and concentrations can vary in time and space, the continuity equation must be combined with all the transport processes outlined above. The overall equation describing convective-diffusive-dispersive transport can then be written as:

$$\frac{\partial c}{\partial t} = D_a \frac{\partial^2 c}{\partial x^2} - U \frac{\partial c}{\partial x} \quad (7.14)$$

where  $D_a$  is the **apparent diffusion coefficient** and represents the sum of molecular diffusion plus hydrodynamic dispersion.

#### 4.6.5 Other factors affecting solute movement Macropore effects

Earthworm activity, root growth, freezing and thawing, and wetting and drying cycles can lead to the development of surface-connected macropores in the soil. Water flow through these macropores can have two distinct effects on leaching: (i) when solutes are present in the infiltrating water, or water is applied immediately after a solute, then macropore flow will lead to extensive leaching at a faster rate than normal (Figure 60c), (ii) when solutes are present within aggregate micropores then they may be bypassed by the bulk of the flowing water and thus protected from leaching (Figure 60c).

##### 4.6.5.1 Soil reaction

Reaction of the solute with soil components can reduce the rate of leaching, as illustrated in Figure 60d. Cations, such as  $\text{Ca}^{2+}$  and  $\text{K}^+$ , which are adsorbed by the cation exchange complex, are less prone to leaching than non-adsorbed anions such as  $\text{NO}_3^-$  and  $\text{SO}_4^{2-}$ . Since soil surfaces generally carry a net negative charge they therefore repel anions rather than attract them, a process which is termed **anion exclusion**. Anion exclusion leads to a more rapid rate of leaching because of the smaller pore volume effective in transport (Figure 60e).

##### 4.6.5.2 Transformations in soil

Some biological transformation processes which occur under normal soil conditions, such as immobilization, can remove solutes from the soil solution and thus reduce the rate of leaching through the profile. Other transformation processes, such as mineralization, may release nutrients into soil solution and thus enhance their rate of removal from the soil.

#### 4.6.5.3 Plant uptake of solutes

Plant uptake of solutes decreases their concentration in soil solution and therefore reduces their rate of leaching loss from the soil.

### 4.7 Plant water uptake from soil

Plants need to take up a tremendous amount of water to satisfy their transpiration demand. For every kilogram of dry matter (DM) produced a plant must transpire between 200 and 500 kilograms of water (200-500 litres). For a crop yielding 10000 kg DM ha<sup>-1</sup> this represents a minimum of 2 million l ha<sup>-1</sup>. Two processes are essential if plants are to obtain this amount of water from soil. The first is rapid root growth in order to reach zones of moist soil and the second is the occurrence of unsaturated flow of soil water in response to water potential gradients.

#### 4.7.1 Root growth and water flow

Plant roots grow outward in a random pattern. Where growing roots encounter a zone of moisture in a dry soil they tend to branch profusely and the dense root pattern which is produced is able to come into contact with more water for direct absorption into the plant.

Although plant roots may be able to elongate at a sufficiently rapid rate to come into contact with enough water for growth in a moist soil, the extent of direct contact is extremely small. Consequently much of the water required by the plant must move through the soil to reach the root rather than be directly absorbed by contact. The processes of unsaturated water flow in soil have been already described. It is worth remembering that as the plant root removes water from one zone of soil, the potential energy of the water remaining in that zone is decreased. The water in the soil surrounding that zone is then at a relatively higher potential and water will be induced to flow towards the drier zone. The distance of water movement is never very large and may in fact be only a few millimetres. However, considering the normal density of roots in soil, this is sufficient to supply enough water for plant growth.

#### 4.7.2 Water movement from soil into plant roots

Depending on the rate of transpiration of the plant, water movement into plant roots is caused by two different driving forces. At slow rates of transpiration (i.e. when there is a low demand for water) then **osmotic** or **active absorption** occurs. At high rates of transpiration **mass flow** or **passive absorption** occurs. Absorption (F) may be described by the following equation:

$$F = \frac{(\psi_m + \psi_s)_{\text{soil}} - (\psi_p + \psi_s)_{\text{plant}}}{r_s + r_p} \quad (7.15)$$

where  $\psi_m$  is the soil matric potential,  $\psi_s$  is the solute, or osmotic potential,  $\psi_p$  is the pressure potential within the plant, and  $r$  is the resistance to flow in either the plant or the soil. Depending on which terms are dominant in the equation, water absorption is either osmotic or passive. **Osmotic (active) absorption** - At slow transpiration rates, plant roots can have a significant concentration of solutes in the xylem. This lowers the water potential of the root below that of the surrounding soil and as a result water flows from the soil to the root by osmosis. **Passive absorption** - At fast transpiration rates the concentration of solutes in the root is reduced by the large amount of water which is taken up, causing the osmotic absorption mechanism to become ineffective.

In rapidly transpiring plants, the leaves are losing water extremely quickly by evaporation. This causes a decreased plant water potential to develop in the leaf cells and induces water to flow from the stem where the water potential is relatively higher. In turn, the removal of water from the stem causes water to flow from the root, where the water potential may be reduced to a value as low as -1500 kPa. The steep gradient between the plant and the soil water potential (the latter may only be -30 kPa for example) causes water to flow rapidly into the plant. In this case the plant root acts merely as an absorbing surface and water essentially moves as a continuous column through the plant, 'pulled' by the very low potential developed at the evaporating leaf surface.

With continued removal of water the soil dries out and its water potential decreases. When the soil water potential becomes too low then the potential gradient into the plant becomes so small that water flow is no longer induced. Increasing resistance to water flow in the soil and roots, as well as decreasing contact between the roots and soil, also decreases the rate of absorption. The plant then starts to wilt and will continue to do so until water is applied to the soil.

#### 4.8 Evaporation and evapotranspiration

Water may be evaporated into the atmosphere from the surfaces of plant leaves, soil, or open bodies of water, such as lakes or rivers. The rate at which evaporation occurs depends on climatic factors, such as radiation, temperature, humidity, and wind speed, as well as on various characteristics of the plant or soil surface.

##### 4.8.1 Mechanisms

Energy is required to change water from a liquid state into a gaseous state and in the evaporation process the sun supplies this energy in the form of radiant heat. For every gram of water vaporized at least 540 calories of energy are required in order to break the hydrogen bonds which link water molecules together in the liquid state. Energy from the sun must be absorbed by the evaporating surface for vaporization to occur and this depends on the reflectance or **albedo** of the surface. Fresh snow, for example, has a high albedo (80%) since it reflects most incoming solar radiation and absorbs only a small amount (20%) of it. In contrast, water in a lake can have an albedo of 5% and is thus able to absorb the vast majority (95%) of incoming solar radiation. The rate of evaporation from a lake is therefore higher than from a snow slope. The albedo values given in Table 11 show that when all other things are equal the rate of evaporation from pasture (albedo 25%) is likely to be less than from a forest (albedo 15%).

For evaporation to take place from any surface it is also necessary that the vapour pressure of the air above be less than at the evaporating surface itself. This vapour pressure difference is the driving force causing flow to occur and has already been discussed in relation to vapour flow within soil.

**Table 11.** Some approximate typical albedo values

Surface	Albedo(%)
Snow	80
Grass	25
Bare soil (dry)	20
Forest	15
Bare soil (moist)	10
Deep, clear water	5

The **evaporation rate** (E) from soil may be described in a simple manner by the following equation:

$$E = \frac{e_s - e_a}{r_s + r_a} \quad (7.16)$$

where  $e_s$  and  $e_a$  are the water vapour pressures in soil and in air respectively, and  $r_s$  and  $r_a$  represent any resistances to flow. The equation states that the rate of evaporation is proportional to the difference in vapour pressure between the soil and the air above, divided by the sum of the resistances to flow in the soil and air above.

**Transpiration** from plants ( $T_r$ ) can be described by a similar equation which includes the vapour pressure at the (inner) leaf surfaces ( $e$ ) and the effects of diffusive resistances within the leaf ( $r_l$ ) rather than the soil:

$$T_r = \frac{e_1 - e_a}{r_1 + r_a} \quad (7.17)$$

Vapour pressure, and therefore the evaporation rate (E), are influenced by the following climatic factors:

(i) **Humidity.** On humid days the  $e_a$  value is relatively high and therefore the rate of evaporation according to Equation 7.16 is slow compared to days when the  $e_a$  value is low (i.e. non-humid days).

(ii) **Temperature.** Temperature has a large effect on the vapour pressure of water in soil and plant leaves but has no direct effect on the vapour pressure of the air. On hot days the evaporation rate is greatly increased by the rise in vapour pressure at the soil or leaf surface. (iii) **Wind speed.** When a dry wind blows across an evaporating surface, the air closest to the surface, which is at the highest  $e_a$  value, is exchanged with air of lower vapour pressure above. The effect of the wind is therefore to maintain a steep vapour pressure gradient and encourage evaporation to occur. In areas such as Canterbury, which receive hot, dry winds, the evaporation losses can be extremely high.

When the soil is wet, the resistance term ( $r_s$ ) in Equation 7.16 is low and the rate of evaporation can be as great as from an open water surface. For evaporation to continue to occur, from either a soil or leaf surface, there must, however, be a continuous supply of water to that surface. In soil the development of a matric potential gradient initially ensures a supply of water for rapid evaporation to occur. However, as the surface soil matric potential decreases further due to drying the rate of hydraulic conductivity also decreases markedly (see Section 7.1.2). The slow rate of water flow through the soil micropore system eventually limits the rate of evaporation from the surface and increases the size of the resistance term ( $r_s$ ) in Equation 7.16.

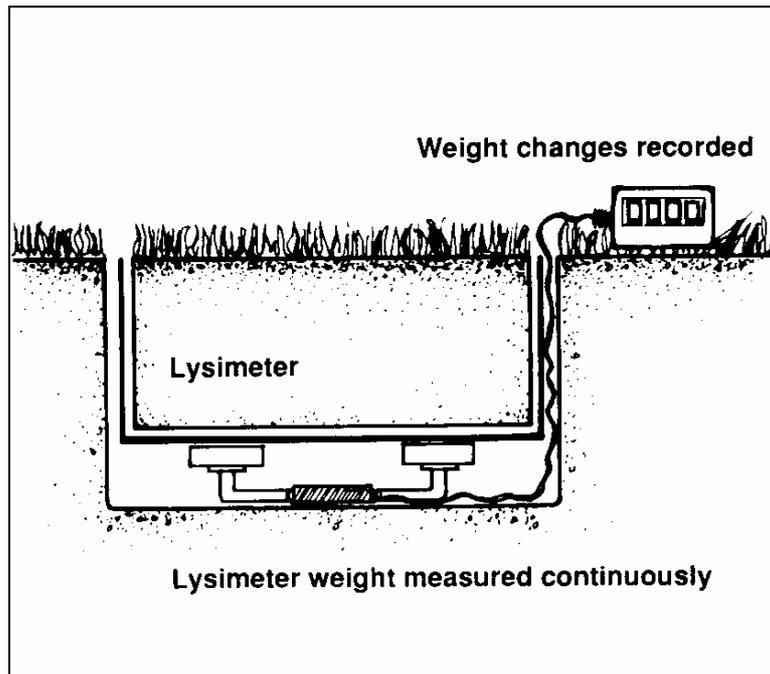
Once the surface 10-20 mm becomes air dry the rate of evaporation will be reduced considerably. The dry surface layers have a 'self-mulching effect' and reduce the loss of water from lower in the profile. This effect is often used by farmers to retain moisture in the soil for planting an autumn crop, such as a greenfeed, and simply involves leaving a bare cultivated soil surface over the summer.

In soils with a shallow water-table, at less than one or two metres depth, the evaporating surface may be continually supplied with water. This phenomenon, often called **capillary rise** because the water flows up through the soil pores by capillary attraction, can result in an excessive water loss and in some cases a build up of salts at the soil surface.

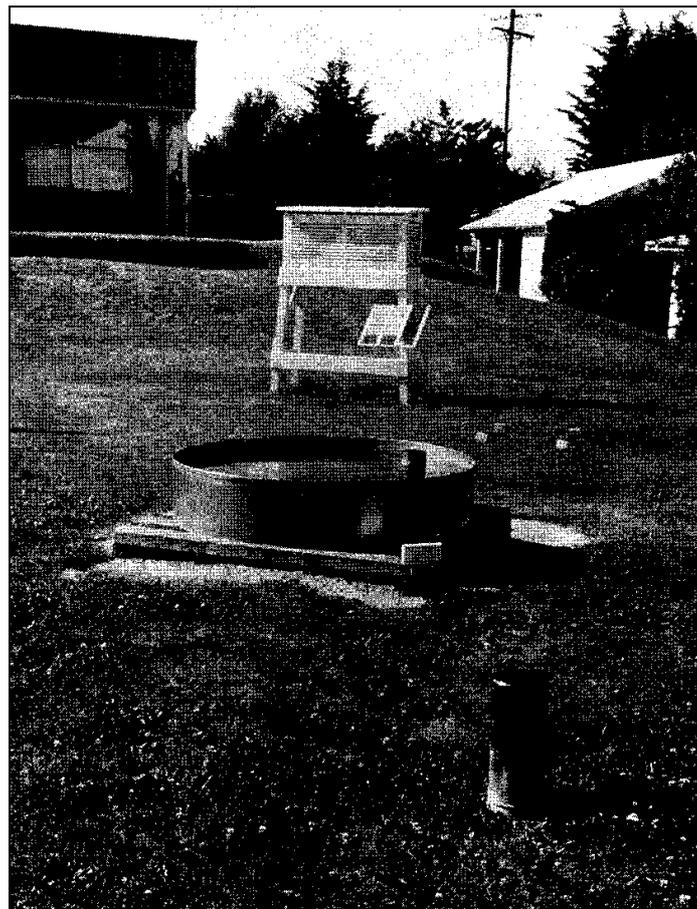
#### 4.8.2 Measurement

Evaporation is normally measured using a large tank of water (Plate 2). The drop in water level is recorded accurately each day and the amount of evaporation is recorded in units of cubic metres of water per square metre ( $m^3 m^{-2}$ ) of surface area, or simply millimetres.

Evapotranspiration rates can be measured directly by recording the weight change of a tank of soil which has a crop growing at the surface, such as that represented in Figure 62. These 'weighing lysimeters' are usually large, often  $4m^2$ , and are expensive to install and operate. It is therefore more common for evapotranspiration rates to be estimated using some form of theoretical, or empirical, model derived for the purpose. These methods are, however, beyond the scope of this book and the reader should refer to the texts listed at the end of this chapter for further information.



**Figure 62.** Weighing lysimeter for direct measurement of evapotranspiration.



**Plate 2.** Part of a meteorological station showing (i) rain gauge (ii) open pan evaporimeter, and (iii) Stevenson screen for housing thermometers. (Photo: K.C. Cameron)

#### 4.9 Water movement in the soil - plant atmosphere continuum (SPAC)

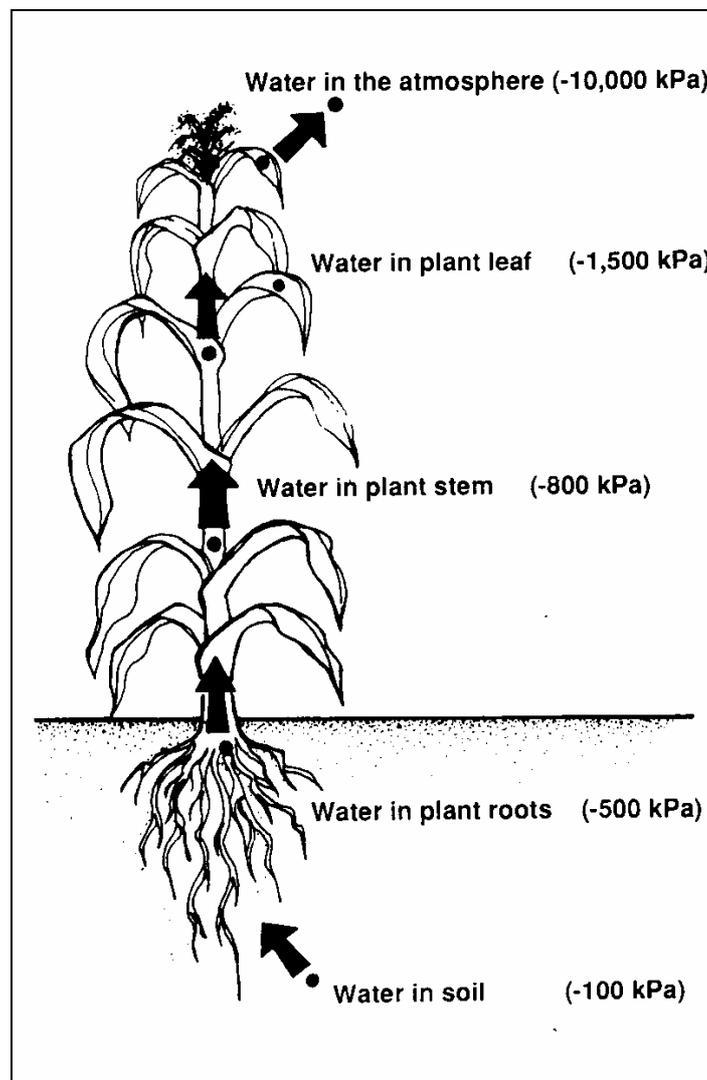
Discussion of water movement up to this stage has necessarily been restricted to specific parts of

the soil-plant-atmosphere continuum (abbreviated to 'SPAC'). However, it is now possible, and worthwhile, to consider the movement of water within the SPAC as a series of interrelated and interdependent processes.

As explained in the previous sections, water will flow from zones of high potential to zones of lower potential. In soil, for instance, water will flow from wet zones within aggregates towards the dry surfaces next to plant roots. Water can then be taken up into the plant because the root water potential is lower than that of the soil surrounding it. Within the plant, there is a continuation

of the potential gradient which causes water to flow from the roots up to the leaves. Once the water reaches the surface of the leaf it can be evaporated into the atmosphere because the difference in vapour pressure between the bulk air and the leaf surface represents a potential energy gradient.

Typical water potentials in each part of the soil-plant-atmosphere continuum are shown in Figure 63. For example, water in soil may be at -100 kPa, water in plant roots at a lower potential of -500 kPa, water in the stem at an even lower potential of -800 kPa, water in the leaf at -1500 kPa, and water in the atmosphere at a very low potential of -10000 kPa.



**Figure 63.** Water flow through the soil - plant - atmosphere continuum (SPAC). Water moves from a relatively high potential energy level in the soil (e.g. -100 kPa) and flows down a potential gradient into the plant roots (-500 kPa), plant stem (-800 kPa), and leaves (-1500 kPa), where it is eventually evaporated into the atmosphere (-10000 kPa).

The rate of water flow (the flux) in any part of the continuum is proportional to the driving force (i.e. the size of the potential energy gradient) and the conductivity of the medium. In soil, the hydraulic conductivity is affected by the pore sizes and their continuity, whilst in the plant

transpiration losses are affected by the size of the stomatal openings. Alternatively, the flux can be considered to be proportional to the driving force but inversely proportional to the resistances to flow within the medium. This may be summarized by the following equation:

$$\text{Flux} = \frac{\text{potential difference}}{\text{total resistance to flow}} \quad (7.18)$$

\* \* \*