united nations
educational, scientific
and cultural
organization

international atomic
energy agency

the
abdus salam
international centre for theoretical physics

H4.SMR/1304-7

"COLLEGE ON SOIL PHYSICS"

12 March - 6 April 2001

Lecture Notes: Soil Physics

R. Hartmann

University of Ghent
Laboratory of Soil Physics
Dept. Soil Management & Soil Care
Faculty of Agriculture & Applied Biological Sciences
Ghent, Belgium

UNIVERSITEIT
GENT

LECTURE NOTES : SOIL PHYSICS

by

R. HARTMANN[1]

---

[1] *Laboratory of Soil Physics*
Department of Soil Management and Soil Care
Faculty of Agricultural and Applied Biological Sciences
University Gent - Belgium

TEXTBOOKS

Rose, C.W., 1956.
Agricultural physics.

Slayter, R.O., 1967.
Plant-water relationships.
New York: Academic.

Childs, E.C., 1969.
An introduction to the physical basis of soil water phenomena.
New York: Intersciences.

Hillel, D., 1971.
Soil and water: Physical properties and processes.
New York: Academic.

Nielsen, D.R., Jackson, R.D., Corry, J.W. and Evans, D.D., 1972.
Soil water.
Madison, Wisconsin, Am. Soc. of Agronomy.

Baver, L.D., Gardner, W.H. and Gardner, W.R., 1972.
Soil Physics.
New York: Wiley.

Kirkham, D. and Powers, W.C., 1972.
Advanced soil physics.
New York: Interscience.

Taylor, S.A. and Ashcroft, G.L., 1972.
Physical edaphology.
San Francisco: W.H. Freeman and Company.

Drainage Principles and Applications, 1973.
I. Introductory subjects.
II. Theories of field drainage and watershed runoff.
III. Surveys and investigations.
IV. Design and management of drainage systems.
International Institute for Land Reclamation and Improvement, Wageningen, The Netherlands.

Hillel, D., 1974.
L'eau et sol.
Principes et processus physiques.
Louvain: Vander.

Yong, R.N. and Warketin, B.P., 1975.
Soil properties and behaviour.
Amsterdam: Elsevier.

Henin, S., 1977.
Cours de physique du sol, I et II.
Paris: Orstom.

Marshall, T.J. and Holmes, J.W., 1979.
Soil physics.
Cambridge: Cambridge University Press.

Hanks, R.J. and Ashcroft, G.L., 1980.
Applied soil physics.
Berlin: Springer-Verlag.

Hillel, D., 1980.
Fundamentals of soil physics.
New York: Academic.

Hillel, D., 1980.
Applications of soil physics.
New York: Academic.

Koorevaar, P., Menelik, G. and Dirksen, C., 1983.
Elements of Soil Physics.
Elsevier, Amsterdam.

Campbell, G., 1985.
Soil physics with basic.
Elsevier, Amsterdam.

Iwata, S., Tabuchi, T. and Warketin, 1988.
Soil-water interactions. Mechanisms and applications.
New-York and Basel: Marcel Dekker, Inc.

Ghildyal, B.P. and Tripathi, R.D., 1987.
Soil Physics.
John Wiley and Sons, New-York.

Jury, W.A., Gardner, W.R. and Gardner, W.H., 1991.
Soil Physics.
Fifth Edition: John Wiley & Sons, Inc.

Smith, K.A. and Mullins, C.E., 1991.
Soil Analysis (Physical Methods).
Marcel Dekker Inc., New York, Bazel, Hong Kong.

Topp, G.C., Reynolds, W.D. and Gran, R.E., 1992.
Advances in measurement of soil physical properties:
Bringing theory into practice.
SSSA Special Publication n° 30.

Hanks, R.J., 1992.
Applied Soil Physics.
Soil Water and Temperature Application, 2th edition, Springer-Verlag.

Kutilek, M. and Nielsen, D.R., 1994.
Soil Hydrology
Geo-ecology textbook, Catena Verlag.

Dirksen 1999
Soil physics measurements
Geoscience Publisher
Catena Verlag GMBH

TABLE OF CONTENTS

Introduction

In soil physics more than in most subjects there has been a strong symbiotic relationship between the development of theory and practical applications on the one hand and the development of experimental methods on the other. For example, acceptance and use of the concept of matric potential, originally outlined by Buckingham in 1907, were greatly accelerated following Richards' description in 1942 of simple methods for constructing reliable tensiometers. The development of the neutron probe, which allows rapid nondestructive field measurement of soil water content, has led to an enormous upsurge of interest in soil water measurements and their practical applications and to the accelerated development of the theory of soil water movement. In comparison with soil chemistry and mineralogy, the development of techniques for soil physical measurements has been held back by the necessity for many measurements to be carried out in the field.

Other techniques that require reliable 'undisturbed' cores or samples on which to make laboratory measurements have also required specification of a number of soil physical measurement techniques, which is currently being undertaken by the Soil Quality working group of the International Standards Organisation, is an indication of both the increasing importance of soil physical measurements and the coming of age of a number of physical measurement methods. These methods can be regarded as being relatively stable, in the sense that their limitations, reliability, and areas of application are now well established.

Despite these current attempts at standardisation, the choice of appropriate soil physical measurement techniques is often still uncharted ground full of pitfalls. Abandoned and unused soil physical equipment in research laboratories provides a sorry testimony to the difficulties faced by researchers who have been unable to obtain adequate advice on choice of appropriate methods. There are certainly many books, reviews, and technical notes that provide listings and descriptions of soil physical methods. However, most users require a brief guide to any necessary theory and to the techniques most suitable to their own particular application before they can make a choice and are ready to consult one of the many available complications of methods.

Recent books written to provide all the information required to allow the user to choose the technique most suited to his or her desired application are given in the reference list. These books contain mainly a critical review of relevant theory and measurement methods. Particular emphasis has been given to the merits, limitations, and range of application of each method. As well as the consideration of accuracy, this includes measurement time, case of use, and cost.

1. SOIL WATER CONTENT

Soil water is important in itself as a feature of the physical environment but especially prominent in its relationships with climatology and with the surface and the subsurface hydrologic regimes as a component in the terrestrial water balance. To understand the behaviour of soil water one must measure it. This have long proved a difficult task both from the instrumental aspect as because of the complexity of the soil body surpasses that of the vegetated layer and the atmosphere above.

There are numerous procedures and types for the determination of the soil water content. Since each method has its advantages and limitations it is well to consider both the purpose for which determinations are to be made and the features of each possible method including the cost of buying equipment and operating and maintenance costs.

Eight of these are listed in table 1. Only the most frequently applied methods will be discussed.

Table 1. Methods of measuring soil water content

1. Thermogravimetric method
2. Neutron scattering technique
3. Gamma-ray attenuation technique
4. Gamma-ray backscattering technique
5. Electrical resistance
6. Thermal conductivity
7. Time Domain Reflectometry
8. Capacitance method

The soil water content (wetness) can be expressed in terms of either mass or volume ratios or fractions.

- *Mass wetness or dry mass fraction of water (w).*

$$w = \frac{M_w}{M_s} \tag{1}$$

This is the mass of water relative to the mass of dry (105°C) soil particles where w is the mass wetness, M_w water mass and M_s dry (105°C) soil mass.

- *Volume wetness or volume fraction of water (θ).*

$$\theta = \frac{V_w}{V_s + V_w + V_a} \tag{2}$$

It is the dimensionless ratio of the water volume (V_w) relative to total bulk soil volume V_t. The latter is the sum of the volume of solids (V_s), water (V_w) and air (V_a) (figure 1).

The two expressions can be related to each other as follows:

$$\frac{\theta}{w} = \frac{V_w}{V_t}\frac{M_s}{M_w} \tag{3}$$

with : $M_s/V_t = \rho_b$ (mass of dry soil per unit bulk volume and usually lies in the range of 1.3 and 1.7 g cm^{-3})

$M_w/V_w = \rho_w$ (mass of water per unit volume of water and is approximately equal to 1 g cm^{-3})

Equation (3) becomes

$$\theta = \frac{w \ \rho_b}{\rho_w} \tag{4}$$

So in order to obtain the soil water content on a volume basis the mass wetness is multiplied with the dry bulk density. Both w and θ are usually multiplied by 100 and reported as percentages by mass or volume.
The usefulness of the soil water content by volume lies in the fact that it can be converted easily into head units of water therefore being compatible with quantities of rainfall or irrigation water applied. All calculations involving the water balance of the soil, including calculation of irrigation deficits, water application efficiency and recharge of the soil moisture reservoir by rainfall, involve the use of the volumetric soil water percentage (vol.% of water = mm of water per 10 cm soil depth).

If the soil water content is measured at different depths of the profile than the depth-interval dz_i for which θ_i is valid, is taken as the vertical distance between the measuring point and the points located half-way respectively the above and underlying measuring point (figure 2).

The soil water storage “S” is than equal to the equivalent waterhead (mm or cm) present in the profile until a depth z and is obtained through the summation of the moisture contents over each depth interval:

$$S = \sum_{i=1}^{i=n} \theta_i \ dz_i \tag{5}$$

with the same unit as that of z which is compatible with the amount of precipitation or evaporation.

Two moisture content profiles, measured at respectively time t_1 and t_2, allows us to calculate the change in water storage $\Delta S(z_1, z_2)$ during the period Δt.

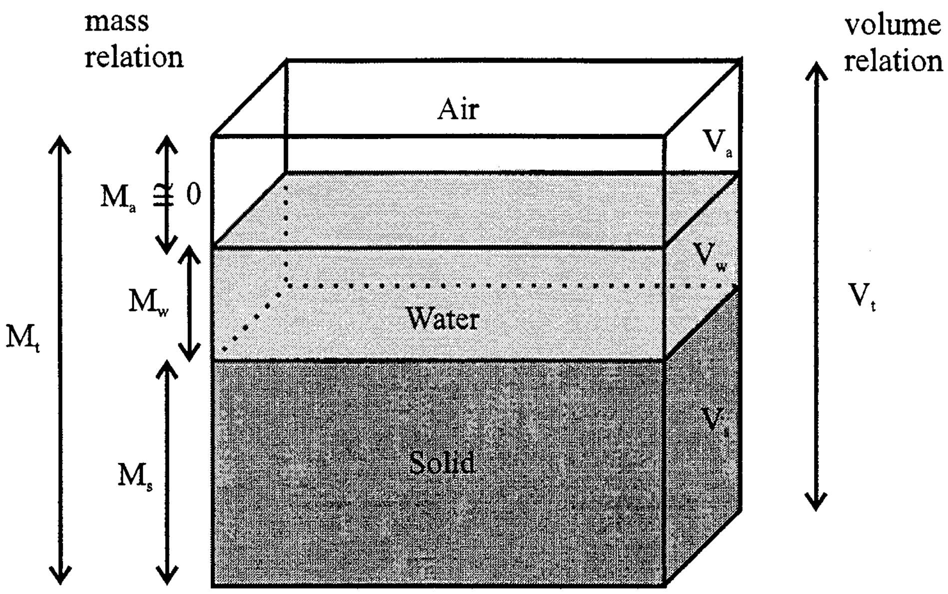


Figure 1. Schematic diagram of the soil as a three-phase system.

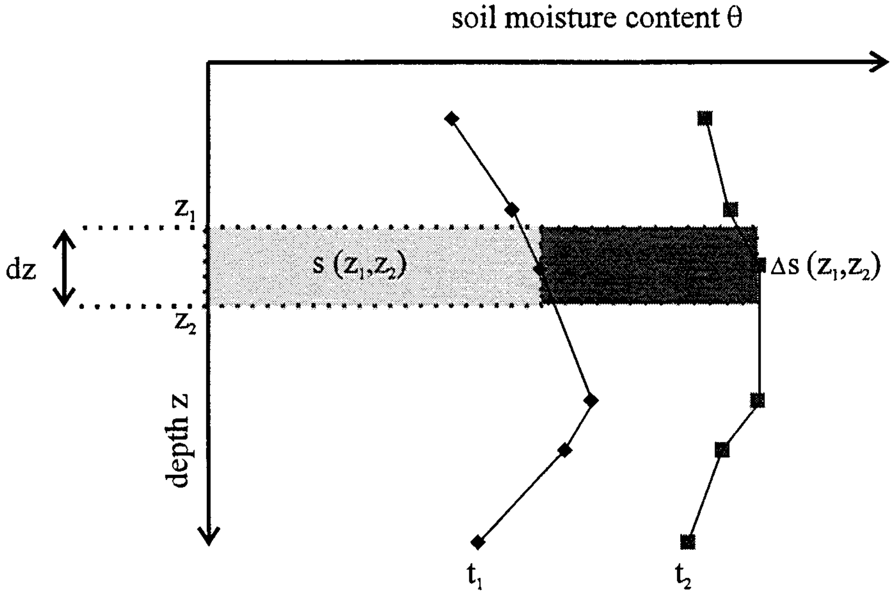


Figure 2. Soil moisture content profiles.

Thermogravimetric method

The thermogravimetric method of measuring soil water content consists of removing a sample by augering into the soil and then determining its moist and dry weight. The moist weight is determined by weighing the sample as it is at the time of sampling and the dry weight is obtained after drying the sample to a constant weight in an oven. The standard method of drying is to place the sample in an oven at 105°C for 24 hours.

The mass wetness is the ratio of the weight loss in drying to the dry weight of the sample:

$$w = \frac{wet\ weight\ -\ dry\ weight}{dry\ weight} = \frac{weight\ loss\ by\ drying}{weight\ of\ dried\ sample}$$

The method depending at it does on sampling, transporting and repeated weighings entails practically inevitable errors. It is also laborious and time consuming, since a period of at least 24 hours is usually considered necessary for complete oven drying. The standard method of oven drying is itself arbitrary. Some clays may still contain appreciable amounts of adsorbed water even at 105°C. On the other hand, some organic matter may oxidize and decompose at this temperature so that the weight loss may not be due entirely to the vaporization of water.

The errors of the gravimetric method can be reduced by increasing the sizes and the number of samples. However the sampling method is destructive and may disturb an observation or experimental plot sufficiently to distort the results. For these reasons many prefer indirect methods, which permit making frequent or continuous measurements at the same point and once the equipment is installed and calibrated with much less time, labor and soil disturbance.

Although the gravimetric determination of soil moisture content is rather laborious it is, because of its simplicity and reliability, the most extensively applied technique and is used as calibration standard for other methods.

2. SOIL WATER POTENTIAL

2.1 Introduction

Soil water content is not sufficient to specify the entire status of water in soil. For example, if soils with a same water content but with different particle size distribution are placed in contact with each other, water will flow from a coarse textured soil to a fine textured soil. One needs to define a property that will help to explain this observation.

Perhaps the following analogy will help. Heat content (analogous to soil water content) is a property of a material that is useful for many purposes. It will not, however, tell us directly whether heat will flow. Therefore a heat intensity term, temperature, has been defined which permits to determine the direction of heat flow. The soil water term that is analogous to temperature (i.e. the intensity with which the water is in the soil) is called the soil water potential. Water potential is a much more complicated property than temperature.

2.2 Energy state of soil water

Soil water, like other bodies in nature, can contain energy in different quantities and forms. Classical physics recognizes two principal forms of energy, kinetic and potential. Since the movement of water in the soil is quite slow, its kinetic energy, which is proportional to the velocity squared, is generally considered to be negligible. On the other hand, the potential energy, which is due to position or internal condition, is of primary importance in determining the state and movement of water in the soil.

The potential energy of soil water varies over a very wide range. Differences in potential energy of water between one point and another give rise to the tendency of water to flow within the soil. The spontaneous and universal tendency of all matter in nature is to move from where the potential energy is higher to where it is lower and to equilibrate with its surroundings. In the soil, water moves constantly in the direction of decreasing potential energy until equilibrium, definable as a condition of uniform potential energy throughout, is reached.

The rate of decrease of potential energy with distance is in fact the moving force causing flow. A knowledge of the relative potential energy state of soil water at each point within the soil can allow us to evaluate the forces acting on soil water in all directions, and to determine how far the water in a soil system is from equilibrium. This is analogous to the well-known fact that an object will tend to fall spontaneously from a higher to a lower elevation, but that lifting it requires work. Since potential energy is a measure to the amount of work a body can perform by virtue of the energy stored in it, knowing the potential energy state of water in the soil and in the plant growing in that soil can help us to estimate how much work the plant must expend to extract a unit amount of water.

Clearly, it is not the absolute amount of potential energy "contained" in the water which is important in itself, but rather the relative level of that energy in different regions within the soil. The concept of soil water potential is a criterion, for this energy. It expresses the specific potential energy (= per unit mass) of soil water relative to that of water in a standard reference state. The standard state generally used is that of a hypothetical reservoir of pure free water (i.e. water not influenced by the solid phase), at atmospheric pressure, at the same temperature as that of soil water (or at any other specified temperature) and at a given and constant elevation.

It is the convention to assign to free and pure liquid water a potential value of zero.

Since the elevation of this hypothetical reservoir can be set at will, it follows that the potential which is determined by comparison with this standard is not absolute, but by employing even so arbitrary a criterion we can determine the relative magnitude of the specific potential energy of water at different locations or times within the soil.

The concept of soil water potential is of great fundamental importance. This concept replaces the arbitrary categorizations which prevailed in the early stages of the development of soil physics and which purported to recognize and classify different forms of soil water : e.g. gravitational water, capillary water, hygroscope water.

New definition by the soil physics terminology committee of the International Soil Science Society provided more clarity in what used to be a rather complicated theoretical set of criteria. The total potential of soil water was defined as follows : " the amount of work that must be done per unit quantity (mass, volume or weight) of pure free water in order to transport reversibly and isothermally an infinitesimal quantity of water from a pool of pure water at a specified elevation at atmospheric pressure (standard reference state) to the soil water at the point under consideration in the soil-plant-atmosphere-system" (figure 3).

If work is required the potential is positive, but if water in the reference state can accomplish work in moving into the soil the potential is negative.

Soil water is subjected to a number of force field which cause its potential to differ from that of pure free water. Such forces result from the attraction of the solid matrix for water, as well as from the presence of dissolved salts and the action of the local pressure in the soil gas phase and the action of the gravitational field. Accordingly the total potential (ψ_t) of soil water relative to a chosen standard rate can be thought of as the sum of the separate contributions of the various components as follows :

$$\psi_t = \psi_g + \psi_o + \psi_m + \psi_{e.p} + \ldots$$

where : ψ_t = total soil water potential

ψ_g = gravitational potential

ψ_o = osmotic potential

ψ_m = matric potential

$\psi_{e.p}$ = external gas pressure potential

. . . = additional terms are theoretically possible

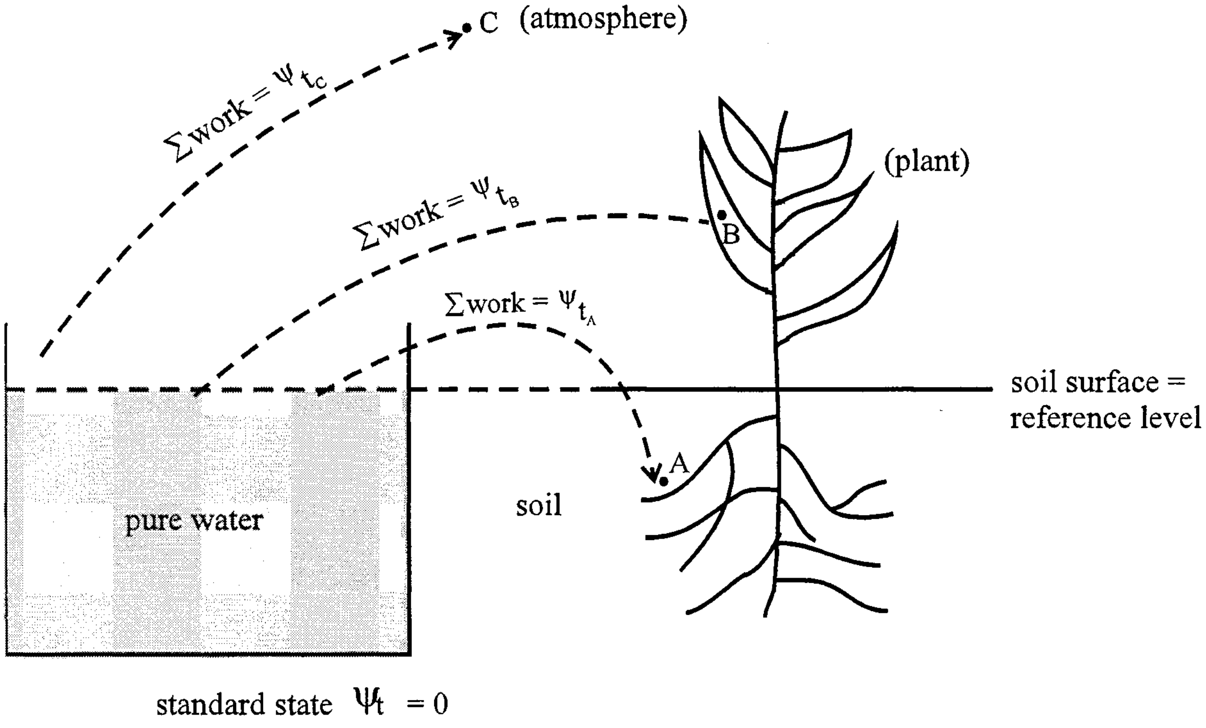


Figure 3. Potential of soil water, water in plant cell and water in the atmosphere.

The main advantage of the total potential concept is that it provides a unified measure by which the state of water can be evaluated at any time and every where within the soil-plant-atmosphere system.

2.3 Quantitative expression of soil water potential

The dimensions of the soil water potential are those of energy per unit quantity of water and the units depend on the way the quantity is specified. Common alternatives used are :

a. Energy per unit mass of water (J/kg)

This method of expression is not widely used.

b. Energy per unit volume of water (pressure) (J/m^3 or N/m^2)

This is the most common method of expressing potential and can be written with units of either Pascal or bar or atmosphere.

c. Energy per unit weight of water (head) (J/N = Nm/N = m)

This method of expressing potential is also common and has units of length.

For conversion from one unit to another knows that :

- 1 bar corresponds to 100 J/kg
- 1 bar = 10^5 Pa
- 1 bar corresponds to 10 m water head

2.4 Gravitational potential

Every body on the earth's surface is attracted towards the centre of the earth by a gravitational force equal to the weight of the body, that weight being the product of the body's mass by the gravitational acceleration. To rise a body against this attraction, work must be expended and this work is stored by the rised body in the form of gravitational potential energy. The amount of this energy depends on the body's position in the gravitational force field.

The gravitational potential of soil water at each point is determined by the elevation of the point relative to some arbitrary reference level. If the point in question is above the reference, ψ_g is positive. If the point in question is below the reference, ψ_g is negative. Thus the gravitational potential is independent of soil properties. It depends only on the vertical distance between the reference and the point in question.

Conversion table for units of soil water potential (*)

Specific potential units	Volumetric potential units				Weight potential units
Joule kg^{-1}	bar	millibar	Pa	atmosphere	m
0.0001	0.000001	0.001	0.1	0.000000987	0.001017 x 10^{-2}
1	0.01	10	10^{3}	0.00987	10.17 x 10^{-2}
100	1	1000	10^{5}	0.987	10.17
0.1	0.001	1	10^{2}	0.000987	1.017 x 10^{-2}
101.3	1.013	1013	1.013 x 10^{5}	1	10.30
0.09833	0.0009833	0.9833	98.33	0.0009703	10^{-2}

(*) The density of water was taken as 1.000 g cm^{-3}. This holds only at 4°C but is approximately correct at other temperatures.

At a height z below a reference level (e.g. the soil surface) the gravitational potential of a mass M of water, occupying a volume V is :

$$-Mgz=-\rho_w Vgz$$

where : ρ_w = density of water
g = acceleration of gravity

Gravitational potential can be expressed :

- per unit mass : $\psi_g=-gz$ (J/kg)
- per unit volume : $\psi_{gv}=\psi_g\,\rho_w=-\rho_w gz$ (Pa)
- per unit weight : $\psi_{gw}=\dfrac{\psi_g}{g}=-z$ (m)

2.5 Osmotic potential

The osmotic potential is attributable to the presence of solutes in the soil water. The solutes lower the potential energy of the soil water. Indeed, the fact that water molecules move through a semi-permeable membrane from the pure free water into a solution (osmosis) indicates that the presence of solutes reduces the potential energy of the water on the solution side (figure 4). At equilibrium sufficient water has passed through the membrane to bring about significant difference in the heights of liquid. The difference (z) in the levels represents the osmotic potential.

Since the osmotic potential of pure free water is zero the osmotic potential of a solution at the same temperature of free water is negative (water flow occurs from point of high potential to one with lower potential).

Differences in osmotic potential only play a role in causing movement of water when there is an effective barrier for salt movement between the two locations at which the difference in ψ_o was observed. Otherwise, the concentration of salts will become the same throughout the profiles by the process of diffusion and the difference in ψ_o will no longer exist. Therefore osmotic potential does not act as a driving force in water flux. This potential is of importance in water movement into and through plant roots, in which there are layers of cells which exhibit different permeabilities to solvent and solute.

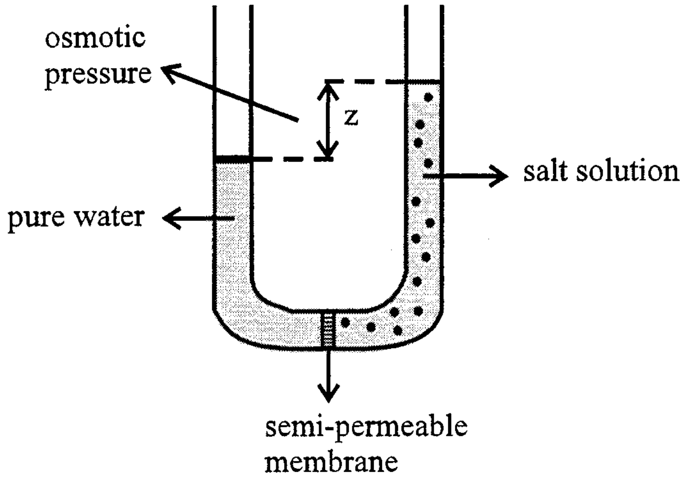


Figure 4. Schematic presentation of osmosis.

2.6 Matric potential

Matric potential results from forces associated with the colloidal matric and includes forces associated with adsorption and capillarity. These forces attract and bind water in the soil and lower its potential energy below that of bulk water. The capillarity results from the surface tension of water and its contact angle with the solid particles. In an unsaturated (three-phase) soil system, curved menisci form which obey the equation of capillarity :

$$P_i - P_a = \Delta P = \gamma(\frac{1}{R_1} + \frac{1}{R_2})$$

where : P_i = pressure of soil water, can be smaller than atmospheric
P_a = atmospheric pressure, conventionally taken as zero
ΔP = pressure deficit
γ = surface tension of water
R_1, R_2 = principal radii of curvature of a point on the meniscus, taken as negative when the meniscus is concave

As we assume the soil pores to have a cylindrical shape (figure 5) the meniscus has the same curvature in all directions and equation above becomes :

$$P_i = \Delta P = \frac{-2\gamma}{R}$$

since :

$$R = \frac{r}{\cos\alpha}$$

($\alpha = 0$: angle of contact between water and the soil particle surface)

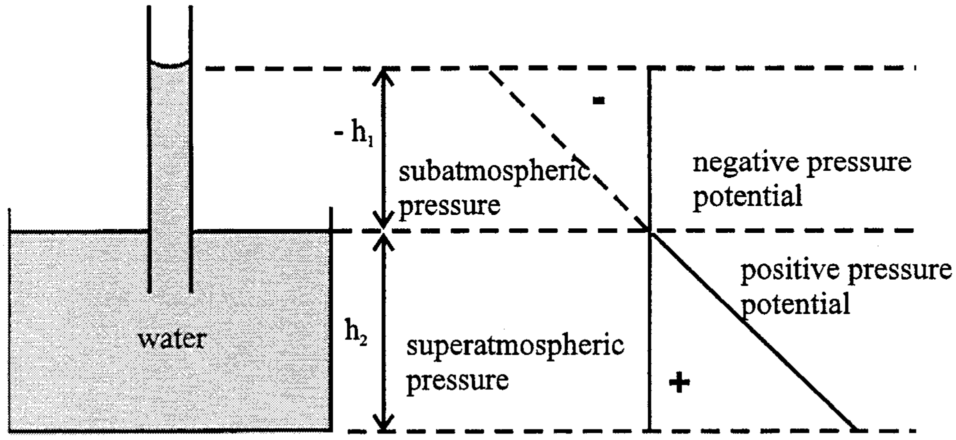
$$P_i = \Delta P = \frac{-2\gamma}{r} \text{ with } \Delta P \text{ equals } - h\, \rho_w\, g$$

where : r = radius of the capillary tube
h = height of capillary rise
ρ_w = density of water (10^3 kg/m^3)
g = acceleration of gravity (9.81 m/s$^2 \approx$ 10 m/s$^2 \approx$ 10 N/kg)

If the soil were like a simple bundle of capillary tubes, the equations of capillarity might be themselves suffice to describe the relation of the negative pressure potential or matric potential to the radii of the soil pores in which the menisci are contained. However, in addition to the capillarity phenomenon, the soil also exhibits adsorption, which forms hydration envelopes, over the particle surfaces. These two mechanisms of soil water interaction are illustrated in figure 6.

The presence of water in films as well as under concave menisci is most important in clayey soil and at high suctions or low potential, and it is influenced by the electric double layer and the exchangeable cations present. In sandy soils adsorption is relatively unimportant and the capillary effect predominates.

In general, however, the matric potential results from the combined effect of the two mechanisms, which cannot easily be separated since the capillary “wedges” are at a state of internal equilibrium with the adsorption “films” and the ones cannot be changed without affecting the others. Hence matric potential denotes the total effect resulting from the affinity of the water of the whole matric of the soil, including its pores and particle surfaces together.



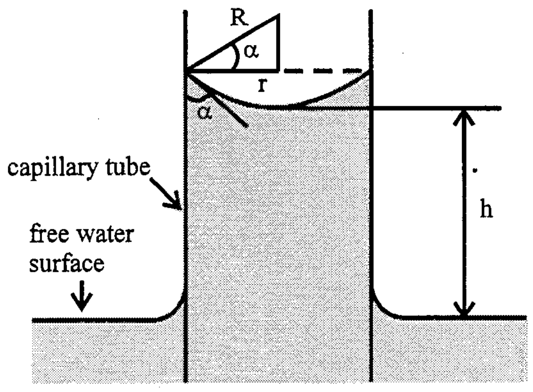


Figure 5. Capillary rise of water into a capillary tube.

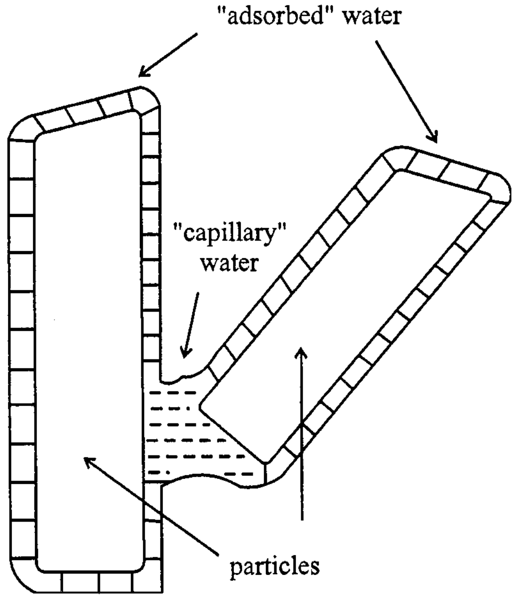


Figure 6. Water in an unsaturated soil is subject to capillarity and adsorption, which combine to produce a matric potential.

The matric potential can be expressed :

- per unit mass : $\psi_m = -gh = \frac{-2\gamma}{\rho_w r}$ (J/kg)

- per unit volume : $\psi_m \rho_w = -\rho_w g h = \frac{-2\gamma}{r}$ (Pa)

- per unit weight : $\psi_m \frac{1}{g} = -h = \frac{-2\gamma}{\rho_w g r}$ (m)

The matric potential is a dynamic property of the soil.

In saturated soil (below the ground water level) the liquid phase is at hydrostatic pressure greater than atmospheric and thus its pressure potential is considered positive (figure 5). Thus water under a free water surface is at a positive pressure potential (hydrostatic pressure potential ψ_h), while water at such a surface is at zero pressure potential (assuming atmospheric pressure in the soil) and water which has risen in a capillary tube above that surface is characterized by a negative pressure or matric potential.

Since soil water may exhibit either of the two potentials, but not both simultaneously, the matric and the hydrostatic pressure potential are referred to as the pressure potential (ψ_p).

Nevertheless it is an advantage in unifying the matric potential and hydrostatic pressure potential in that this unified concept allows one to consider the entire profile in the field in terms of a single unsaturated zone, below and above the water table.

2.7 External gas pressure potential

A factor which may affect the pressure of soil water is a possible change in the pressure of the ambient air. In general this effect is negligible in the field as the atmospheric pressure remains nearly constant small barometric pressure fluctuations notwithstanding. However, in the laboratory the application of excess air pressure to change the soil water pressure is a common practice resulting into the so called external gas pressure of pneumatic potential.

FINAL REMARKS

1. Matric potential and the former term matric suction are numerically equal - when expressed in the same units - but except for the sign.

2. The effect of an external gas pressure different from the atmospheric (reference) pressure is generally also included in the pressure potential so that :

$$\psi_p = \psi_m + \psi_h + \psi_{e.p}$$

Accordingly the total potential being :

$$\psi_t = \psi_g + \psi_o + \psi_p$$

characterizes fully the state of water in soil under the prevailing conditions ; the gradients of these three parameters are the basis for transport theory.

2.8 Hydraulic head

The total potential is obtained by combining the relevant component potentials :

$$\psi_t = \psi_g + \psi_o + \psi_p$$

Equilibrium, which is defined as the situation where mass transfer of water in the liquid phase is absent, is obtained when the value of the total potential at different points in the system in constant. Usually, sufficient condition is that the sum of the component potentials, ψ_o being ignored, is constant. The equilibrium condition states then that :

$$\psi_g + \psi_p = constant = \psi_H \qquad (6)$$

called hydraulic potential.

As already stated, the external gas pressure or pneumatic potential in the field may be assumed to be zero. Also the soil water within a profile may exhibit either matric or hydrostatic pressure potential (figure 5) but not simultaneously. Therefore it is an advantage in unifying both in a single continuous potential extending from the saturated region into the unsaturated region below and above the water table.

As it is often usual to designate the potential in terms of head, equation (6) becomes :

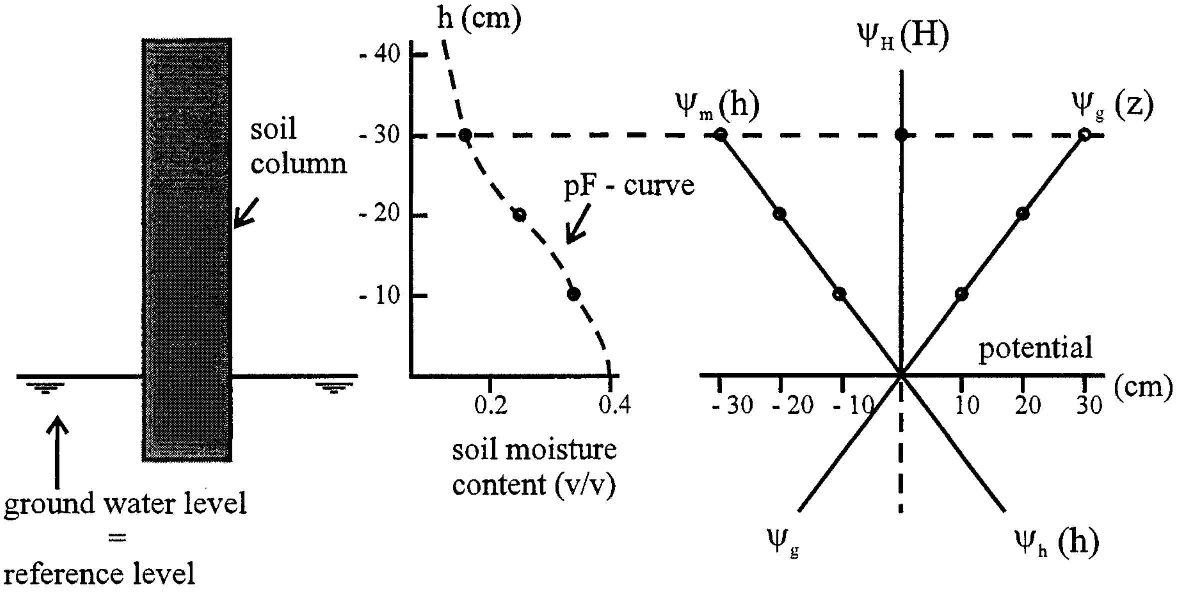
$$H = h + z$$

where : H = the hydraulic head (m)
h = the soil water pressure head (m)
> 0 under the water table (saturated zone)
< 0 above the water table (unsaturated zone)
z = the gravitational head (m)

The definition is very important because the hydraulic gradient between two points under consideration in a soil is the driving force for water movement.

In figure 7 the condition is applied to a vertical soil column in equilibrium with a water table. No water movement occurs in the column. The water table is taken as the reference level for the gravitational potential.

Under the water table matric potential equals zero, but a pressure potential called hydrostatic pressure potential occurs which can also be presented by a value of h but with always a positive sign.



height (cm)	ψ_g (z,cm)	ψ_m (h,cm)	ψ_h (h,cm)	ψ_H (H,cm)
30	30	-30	0	0
20	20	-20	0	0
10	10	-10	0	0
0 reference level	0	0	0	0
-10	-10	0	10	0
-20	-20	0	20	0

Figure 7. Equilibrium condition in a soil column.

3. TENSIOMETER

3.1 Hydrostatic pressure potential - Piezometer

As discussed earlier the hydrostatic (positive) pressure potential ψ_h under field conditions applies to saturated soils and is measured with a piezometer (figure 8).

A piezometer is a tube of a few cm inner diameter, open at both ends, which is installed in a soil profile. If the lower end is below the groundwater table, a piezometer is partially filled with water. By determining the height of the water level in a piezometer it is possible to calculate the (positive) hydrostatic pressure potential of the soil water at the lower end of the tube. The diameter of piezometer is chosen large enough that capillary rise and resistance to water flow are negligible. As a result, any variation in hydraulic potential that may arise inside the piezometer, is instantaneously equalized. Thus, even if the hydrostatic pressure potential at the lower end is changing rapidly, the water inside a piezometer goes through a series of static equilibria and at any moment it can be assumed that the hydraulic head is uniform and equal to the hydraulic head of the soil water at the open lower end. There exchange of water takes place such that the pressure is always locally uniform. The static hydraulic head in piezometer can be determined by measuring the depth of the water level, since at the flat air-water interface the pressure potential is zero.

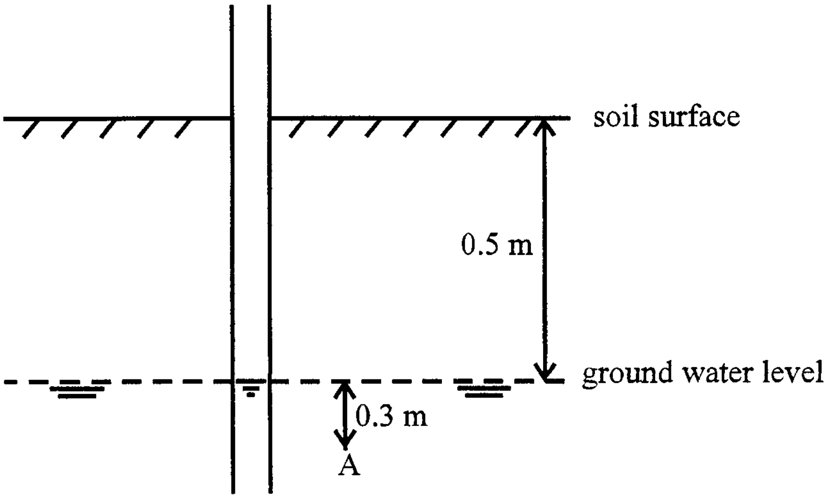


Figure 8. A piezometer in a soil profile.

Figure 8 shows a piezometer in a soil profile in which the water is at static equilibrium. The reference point z = 0 is taken at the soil surface. In the piezometer at the water level H = h + z = 0 - 0.5 m = - 0.5 m. Thus at point A, H must also be - 0.5 m (H = h + z = 0.30 m - 0.80 m = - 0.50 m).

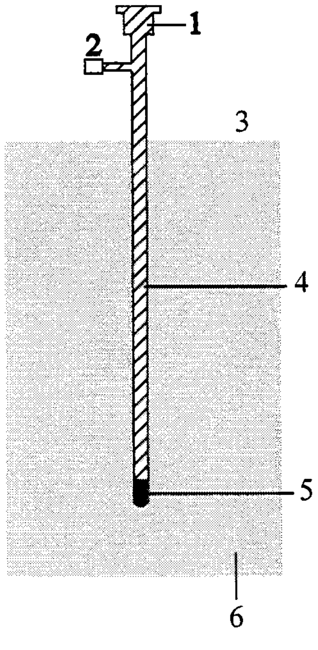
The hydrostatic pressure potential expressed per unit weight of water at any point in the soil under the water table is the distance between the point and the water level in the piezometer tube.

The water level in a piezometer tube is at the level of the groundwater table in a situation of static equilibrium, independent of the depth of the lower end.

3.2 Matric potential - Tensiometer

Piezometers cannot be used to measure negative pressure potentials because in unsaturated conditions, water flows out of the tube into the soil leaving the tube dry. The negative pressure or matric potential can be measured with the so-called tensiometer.

The tensiometer consists of a liquid filled porous cup, mostly of ceramic material and connected to a pressure measuring device such as a mercury manometer or vacuum gauge via a liquid-filled tube (figure 9).



1. Flexible water reservoir
2. Gauge to measure tension
3. Soil level
4. Filled with water
5. Porous end (cup) through which water can move
6. Soil

Figure 9. Tensiometer with a vacuum gauge (jet fill tensiometer).

If the ceramic cup is embedded in soil, the soil solution can flow into or out of the tensiometer through the very small pores in the ceramic cup. Analogously to the situation discussed for piezometer, this flow continues until the (negative) pressure potential of the liquid in the cup has become equal to the (negative) pressure potential of the soil water around the cup. Thus the (negative) pressure potential called matric potential ψ_m of soil water can be measured with a tensiometer, and is therefore also often called tensiometer pressure potential.

3.3 Principle of the tensiometer

When the cup is placed in a water reservoir (figure 10), the water inside the cup comes into hydraulic contact with the water in the reservoir through the water-filled small pores in the ceramic walls. The water level in the tube will indicate the level of the water in the reservoir.

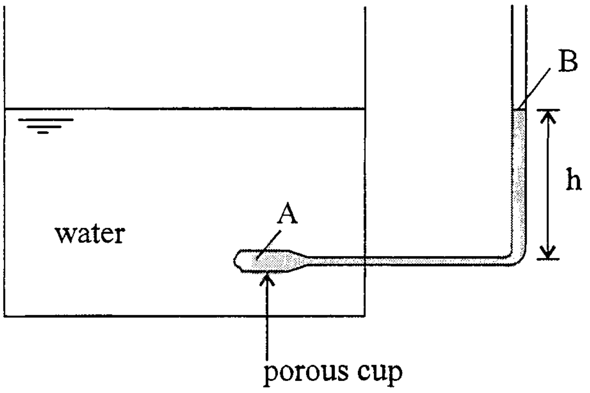


Figure 10.Porous cup connected with a piezometer tube for measuring pressure potentials under the water table.

The pressure is given by the height h of the water level above the middle of the porous cup and the pressure P_A equals :

$$P_A = \rho_w g h$$

where : ρ_w = density of water
g = acceleration due to gravity

If we place now the porous cup, connected with an U-shape water filled tube in a soil than the bulk water inside the cup will come in hydraulic contact with the liquid phase in the soil. When initially placed in the soil, the water in the tensiometer is at atmospheric pressure. Soil water in unsaturated soil has a negative pressure and therefore exercises a suction which drawn out a certain amount of water from the rigid and air-tight tensiometer, causing a drop in the water level at the open end of the U-tube (figure 11).

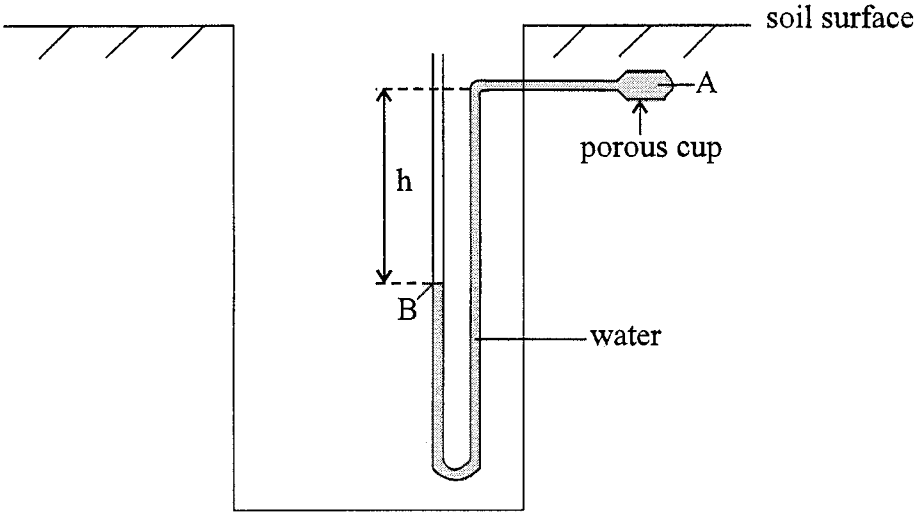


Figure 11.Tensiometer for measuring pressure potentials in soils.

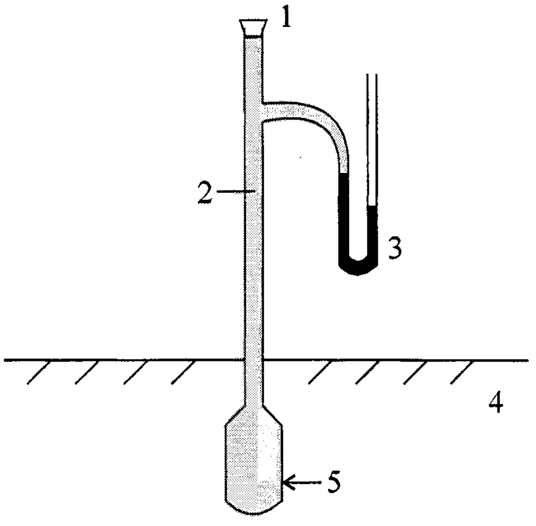
The drier the soil, the higher the suction and the lower the water level at equilibrium in the U-tube. The height h of the liquid column that has moved into ("sucked into") the soil in figure 11 is therefore an index of the magnitude of the potential, or :

$$P_A = - \rho_w \, g \, h$$

As h is measured downwards the minus sign is introduced so that P_A gives a negative pressure.

This type of tensiometer is very simple and useful to illustrate the basic principles involved. Practical applications often do not allow the use of the water manometer because the U-tube extends below the level of the tensiometer cup and measurements thus requires inconvenient, deep pits. Therefore open manometers, filled with immiscible liquids of different densities such as mercury are used so that these problems do not arise (figure 12).

Using mercury implies that a relatively short height indicated a relatively large pressure difference in the manometer (1 cm of mercury corresponds to 13.55 cm of water). Besides the simple water or mercury manometer a vacuum gauge or an electrical transducer is also used.



1. stop
2. water
3. mercury
4. soil
5. porous cup

Figure 12. Tensiometer with mercury manometer.

3.4 How to calculate the soil water pressure head h and the hydraulic head H

Let x be the height of the mercury in the manometer (cm) and z the vertical axis (figure 13). At the interface water-mercury in the manometer, the pressure is the same in water and in mercury (being P_B). The repartition of the pressure is hydrostatic in the water column between point B and the tensiometer cup (point A), but also between point B and the free surface of the mercury in the reservoir (point C).

Using the hydrostatic law for liquids in equilibrium one obtains per unit weight of liquid the following hydraulic head equation :

$$z+\frac{P}{\rho g}=constant$$

where : z = gravitational head
$P/\rho g$ = pressure head

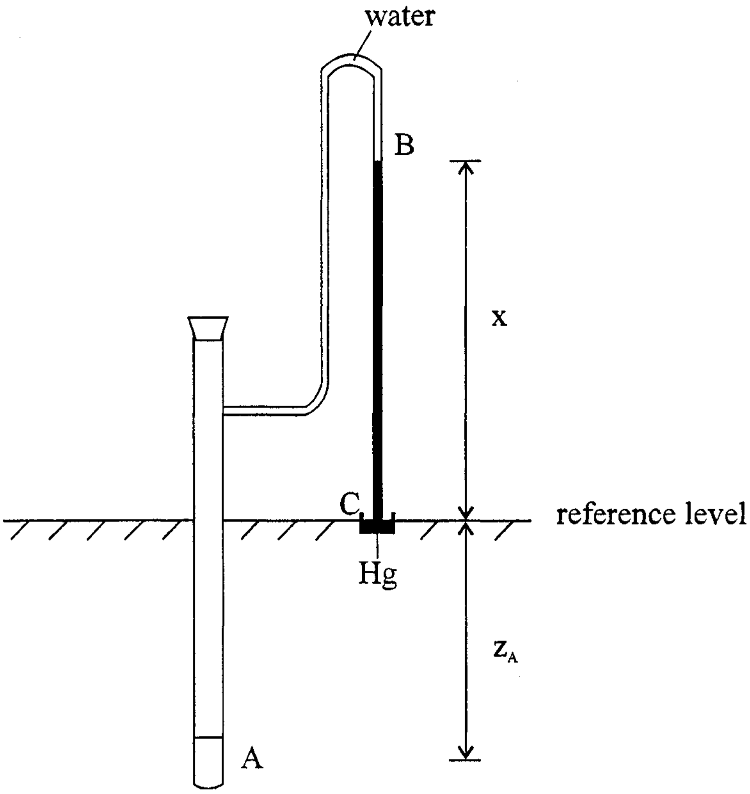


Figure 13. Tensiometer installation with the mercury level in the reservoir at the soil surface being the reference level.

From figure 13 one obtains :

- in water : $z_A + \frac{P_A}{\rho_w g} = z_B + \frac{P_B}{\rho_w g}$

Because soil surface is taken as reference level for the gravitational potential, and point A is located below that level, the gravitational head is negative (- z_A).

$$P_A - z_A \rho_w g = P_B + z_B \rho_w g$$

$$P_A = P_B + z_B \rho_w g + z_A \rho_w g = P_B + \rho_w g (z_A + z_B) \qquad (7)$$

- in mercury : $$z_B + \frac{P_B}{\rho_{Hg}\, g} = z_C + \frac{P_C}{\rho_{Hg}\, g}$$

because : z_C = reference level = 0
P_C = atmospheric pressure = 0

the equation becomes : $$z_B + \frac{P_B}{\rho_{Hg}\, g} = 0$$

or : $$P_B = -\rho_{Hg}\, g\, z_B \tag{8}$$

(8) in (7) gives :

$$\frac{P_A}{\rho_w\, g} = \frac{-\rho_{Hg}\, g\, z_B}{\rho_w\, g} + z_A + z_B \tag{9}$$

since : ρ_w = 1000 kg m^{-3}
ρ_{Hg} = 13600 kg m^{-3}

(9) becomes : $$h_A = -13.6 z_B + z_A + z_B$$

or : $$h_A = -12.6 z_B + z_A$$

$$h_A = -12.6 x + z_A \tag{10}$$

Normally the free surface of the mercury in the reservoir (point C) is located y cm above the soil surface (reference level) (figure 14).

Equation (10) becomes : $$h_A = -12.6 x + y + z_A \tag{11}$$

The hydraulic head H, being the sum of the pressure head h and the gravitational head z, (H = h + z), becomes :

$$H = -12.6 x + y + z_A + (-z_A)$$

$$H = -12.6 x + y \tag{12}$$

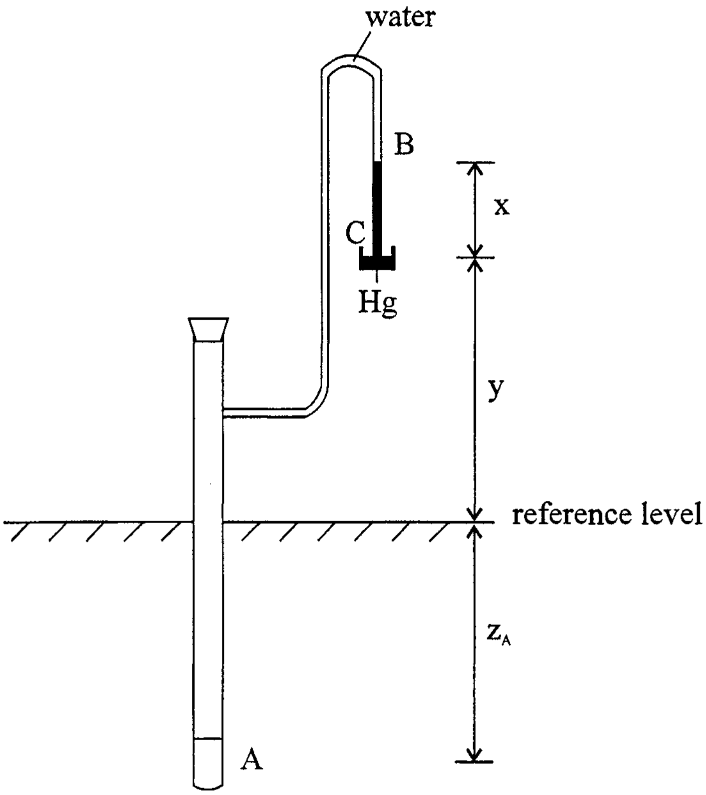


Figure 14. Tensiometer installation with the mercury reservoir y cm above soil surface.

If the soil water pressure head around the porous cup changes, the height of the mercury level will change consequently. For a situation where e.g. two tensiometers are connected to the same mercury reservoir, the value y is the same and from equation (12) it follows that :

$$H_1 - H_2 = -12.6(x_1 - x_2)$$

This means that tensiometers connected to the same mercury reservoir yield hydraulic head differences irrespective of there depths. So $H_1 - H_2 = 0$ where $x_1 = x_2$. Such installation provides at once the direction of the water flow: from the tensiometers with low mercury height to those with high mercury height.

The mercury manometer can be replaced by a Bourdon or dial vacuum gauge (figure 15) or by a pressure transducer (figure 16). For such manometers equation (11) and equation (12) became :

$$h = h_{man} + z_o = h_{man} + z_1 + z_2$$

$$H = h_{man} + z_1 + z_2 - z_2 = h_{man} + z_1$$

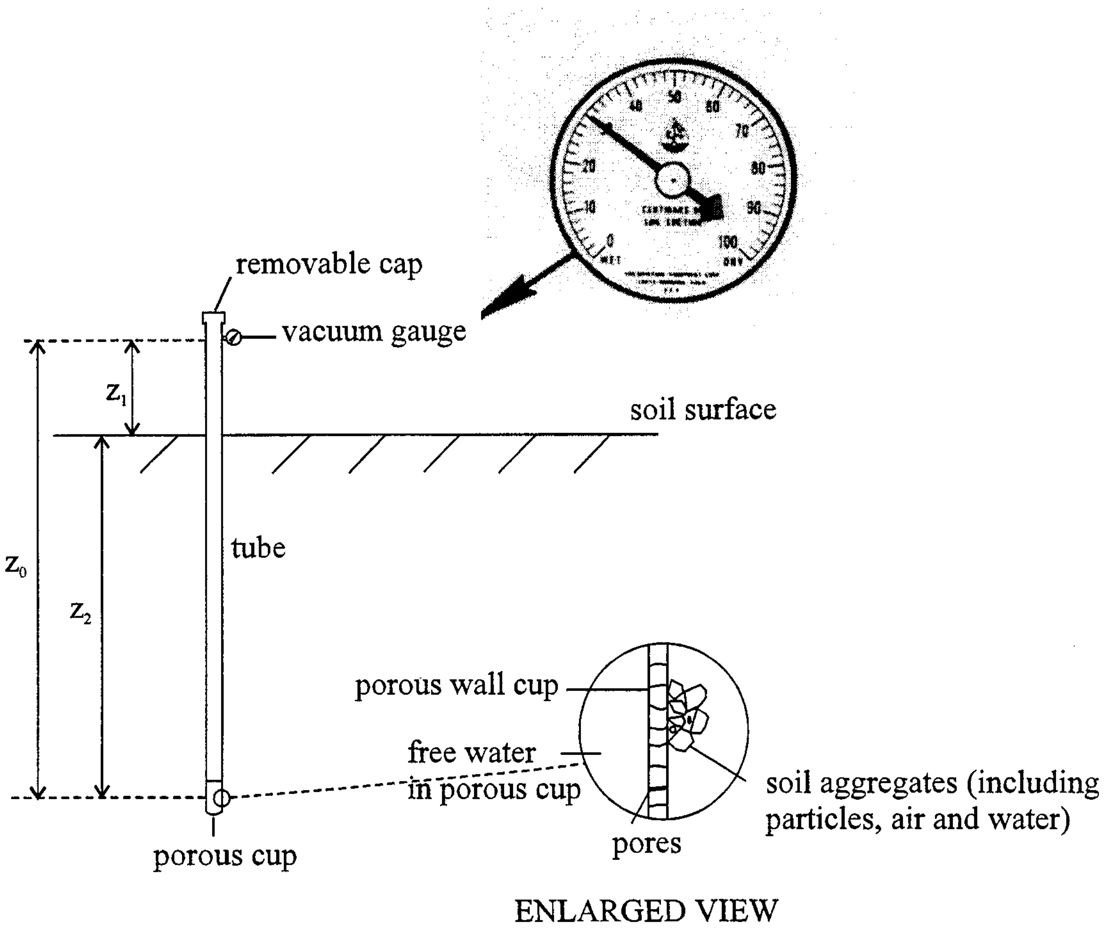
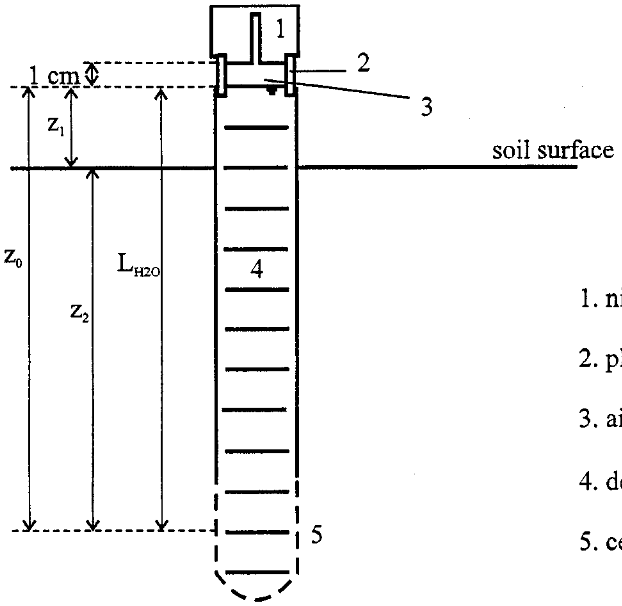


Figure 15. Principle of tensiometer with vacuum gauge.



1. nitril-caoutchouc septum stopper
2. plexi-glass tube (lucid)
3. air pocket (V_1) (end of syringe needle here)
4. deaerated water
5. ceramic cup

MEASURING DEVICE

1. air vent
2. containment with p = p (atm)
3. semiconductor element
4. steel membrane (D = 13 mm)
5. guiding tube
6. guiding tube for needle
7. syringe needle (D = 0.4 mm) dead volume (V_0)
8. guiding spring
9. membrane chamber (45 mm)
10. shielded four lead wire

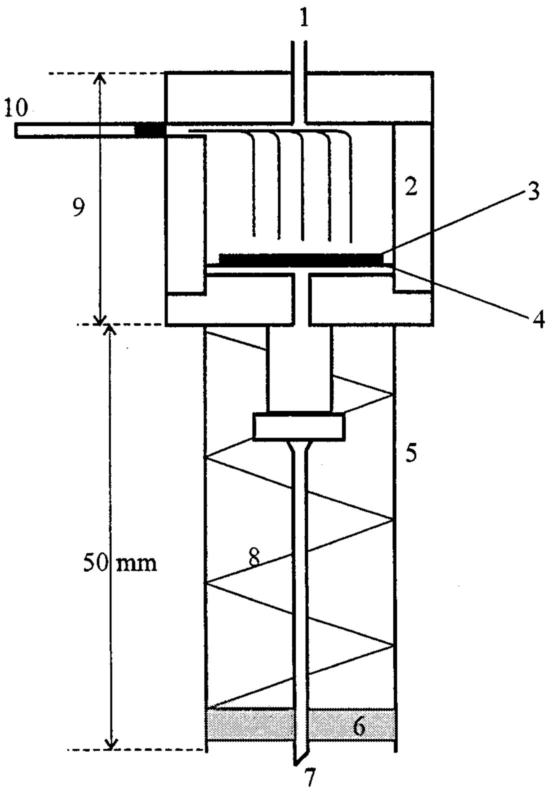


Figure 16.Diagram of tensiometer with septum stopper and pressure transducer with attached syringe needle.

3.5 Some characteristics of the tensiometer

3.5.1 Cup conductance "C"

Being the volume of water dV that has to be displaced per unit time between the tensiometer and the soil through the porous cup for the measuring device to register the change in soil water pressure head dh :

$$C = \frac{dV}{dt\,dh} \qquad (\mathrm{m}^2/\mathrm{s})$$

The surface area, conductivity and wall thickness of the porous cup can be varied to increase its conductance and consequently reduce the measuring time. However there are practical limits to this e.g. the air entry value of the porous cup.

3.5.2 Sensitivity of the manometer "S"

The sensitivity of the manometer determines the performance of the pressure measuring device :

$$S = \frac{dh}{dV} \qquad (\mathrm{m}^{-2})$$

where dV is the volume of water displaced needed by the measuring device to register a pressure head change dh.

- water/mercury manometer

The water manometer has the lowest sensitivity as it requires a large volume of water to be displaced per unit soil water pressure head change. If the diameter of the manometer tube is large enough to eliminate capillary effects, soil water pressure head can be measured with a precision of ± 1 mm. Since large change in soil water pressure head requires an exchange of a large volume of water with the soil it is convenient to replace water by mercury.

The mercury manometer reduces the height of the liquid column and thus the displaced volume of water by a factor 12.6 (equation 11) and so increases the sensitivity by the same factor. The overall precision of mercury manometers is often not better than 5 cm WH. Also the size of the manometer tube influences the sensitivity. Indeed if the cross-sectional area of a manometer tube is A cm^2 the water displacement head ΔV is A Δh so that :

$$S = \frac{\Delta h}{A \Delta h} = \frac{1}{A} \qquad (m^{-2})$$

- Bourdon/dial vacuum gauges

The sensitivity are generally comparable to that of the mercury manometer.

- Pressure transducers

They have very high sensitivity since they require very small displacements of their sensing element to register a full scale pressure range. However the sensitivity of the tensiometer system is different from that of the pressure sensor. The volume of water that needs to be exchanged with the soil is generally larger because the tubing and air present in the system expand and compress with changes in pressure.

- Response time "T_r"

It is the time needed to make an accurate measurement with a particular tensiometer system and is inversely proportional to the conductance of the porous cup of the tensiometer and the sensitivity of the pressure measuring device and is characterized as follows :

$$T_r = \frac{1}{CS} \qquad (s)$$

This is for a system were the tensiometer is the limiting factor (in water or wet soil). In a soil-limited system (dry soil) the response is much slower under decreasing soil water pressure heads due to small water flow rates to the tensiometer cup as a result of low soil hydraulic conductivities. The flexibility of the tensiometer tubing and the compressibility of air bubbles inside the tubing decrease the effective sensitivity and thus increase the effective response time of a tensiometer system.

3.6 Practices and limitations of tensiometers

The purpose of the measurements with tensiometers is to characterize the existing pressure potential of the soil water.

Water within the tensiometer should be continuous throughout the system to allow a correct transfer of pressure from the soil to the mercury. Occurrence of gas bubbles disrupts this continuity and makes the system inoperative. The fine porous cup has the function of not allowing penetration of air from the unsaturated soil into the water-filled tensiometer tube, even though water can and should move through it. The fine pores inside the wall of the ceramic cup have a high air-entry value which is the pressure needed to remove the water from the pores in the cup replacing it by air. Even with a high air entry value breakdown of the system occurs due to entrapped air within the tensiometer tube or to air coming out of solution at reduced pressure.
Due to the fact that the manometer measures a partial vacuum relative to the external atmospheric pressure, measurements by tensiometry are generally limited to about - 850 cm of waterhead. Use of tensiometers in the field is therefore only possible when pressures do not fall below this value. Moreover the pressure head in the tensiometer must stay above the air-entry value of the porous material which is gives as follows :

$$h = \frac{-4\sigma \cos\theta}{\rho_w g d_{max}}$$

where : h = pressure head at which air enters the cup through the largest pore
σ = surface tension of an air-water interface (0.0073 N m^{-1} at 25°C)
θ = contact angle
d_{max} = largest pore diameter

So to cover the practical tensiometer range to $h = -850$ cm for a completely wetting material ($\theta = 0$) d_{max} must be smaller than 3.6 µm. If the air-entry value is exceeded air can enter the tensiometer and the soil will drain the tensiometer.

However, the limited range of pressure measurable by the tensiometers is not as serious as it may seen at first sight. In many agricultural soils the tensiometer range accounts for more than 50% of the amount of soil water taken up by the plants. To what extend the available water range expressed e.g. as a percentage of the water between pF 2 and pF 4.2 is covered by the tensiometer depends on the shape of the moisture characteristic curve (pF-curve) as shown for three soil types in figure 17.

Thus where soil management (particularly in irrigation) is aimed at maintaining high pressure potential conditions which are mostly favourable for plant growth, tensiometers are definitely useful.

Air diffusion through the porous cup into the system requires frequent purging with deaired water. Tensiometers are also sensitive to temperature gradients between their various parts. Hence the above-ground parts should preferably be shielded from direct exposure to the sun. Therefore it is also suggested to make readings always at the same time of the day (e.g. at 08.00 a.m.).

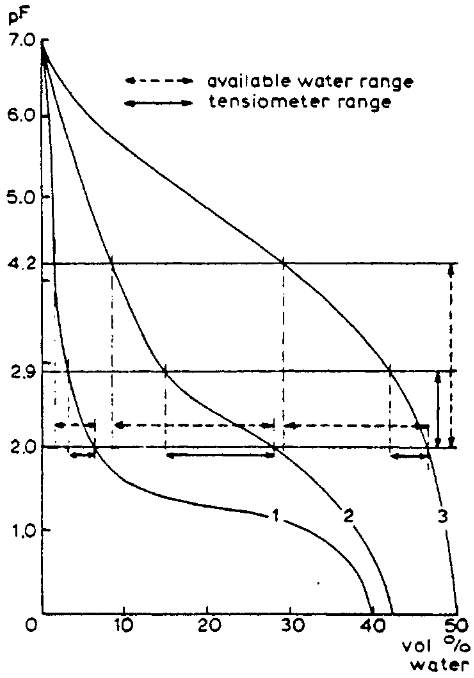


Figure 17. Part of the available moisture range covered by tensiometers, depending on soil type.

1. Sand 50% of available moisture
2. Loam 75% of available moisture
3. Clay 25% of available moisture

When installing a tensiometer it is important for proper functioning that good contact be made between the porous cup and the surrounding soil. Generally the porous cup is pushed into a hole with a slightly smaller diameter to ensure good contact. If the soil is initially rather dry and hard, prewetting of the hole may be necessary. In a stony soil a small excavation should be made and filled with very fine sand into which the tensiometer can be placed.

With mercury manometers, even when small diameter nylon tubing (+/- 2 mm) is used, often a considerable volume of water must be adsorbed by the soil (during water uptake or drying process) or by the porous cup (replenishing by rainfall or irrigation) before the potential that really exists can be read off correctly. A very convenient modern device, the electronic transducer can be used which reacts to very small changes in pressure and converts these changes in a small electrical current which can be registered and amplified by a voltmeter. This system is very accurate but also very sensitive to the occurrence of small air bubbles in the tensiometer system. Moreover is it rather expensive.

Since the porous cup walls of the tensiometer are permeable to both water and solutes, the water inside the tensiometer tends to assume the same solute composition and concentration as soil water, and the instrument does not indicate the osmotic potential of soil water.

3.7 Applications of measurements

3.7.1 Determination of the direction of water flow at different levels in the soil profile (figure 18)

The concept of the water potential is well suited for the analysis of water flow in soils, since all flow is a consequence of potential gradients. Darcy's law, though originally conceived for saturated flow only, was extended to unsaturated flow, with provision that conductivity is a function of soil water content θ.

For a vertical one dimensional water flow Darcy's equation can be written as follows :

$$q = -K(\theta)\frac{dH}{dz} \tag{13}$$

where : q = flux
$K(\theta)$ = hydraulic conductivity
H = hydraulic head
= $h + z$ with h = soil water pressure head
z = gravitational head

The minus sign in the equation indicates that the flow is in the direction of decreasing potential. This means also that if we have two tensiometers located at depths z_1 and z_2 ($z_1 < z_2$) :

- q will be negative (upward flow - evaporation) if $H_2 > H_1$; the rise of mercury in manometer nr 2 is lower than in manometer nr 1
- q will be positive (downward flow - percolation) if $H_2 < H_1$; the opposite situation is observed
- q will be zero (plane of zero flux) at a certain depth z when the curve H(z) will show a maximum or the rise of the mercury a minimum because $dH/dz = 0$. A graphical example is presented in figure 18

3.7.2 Flux control at a certain depth

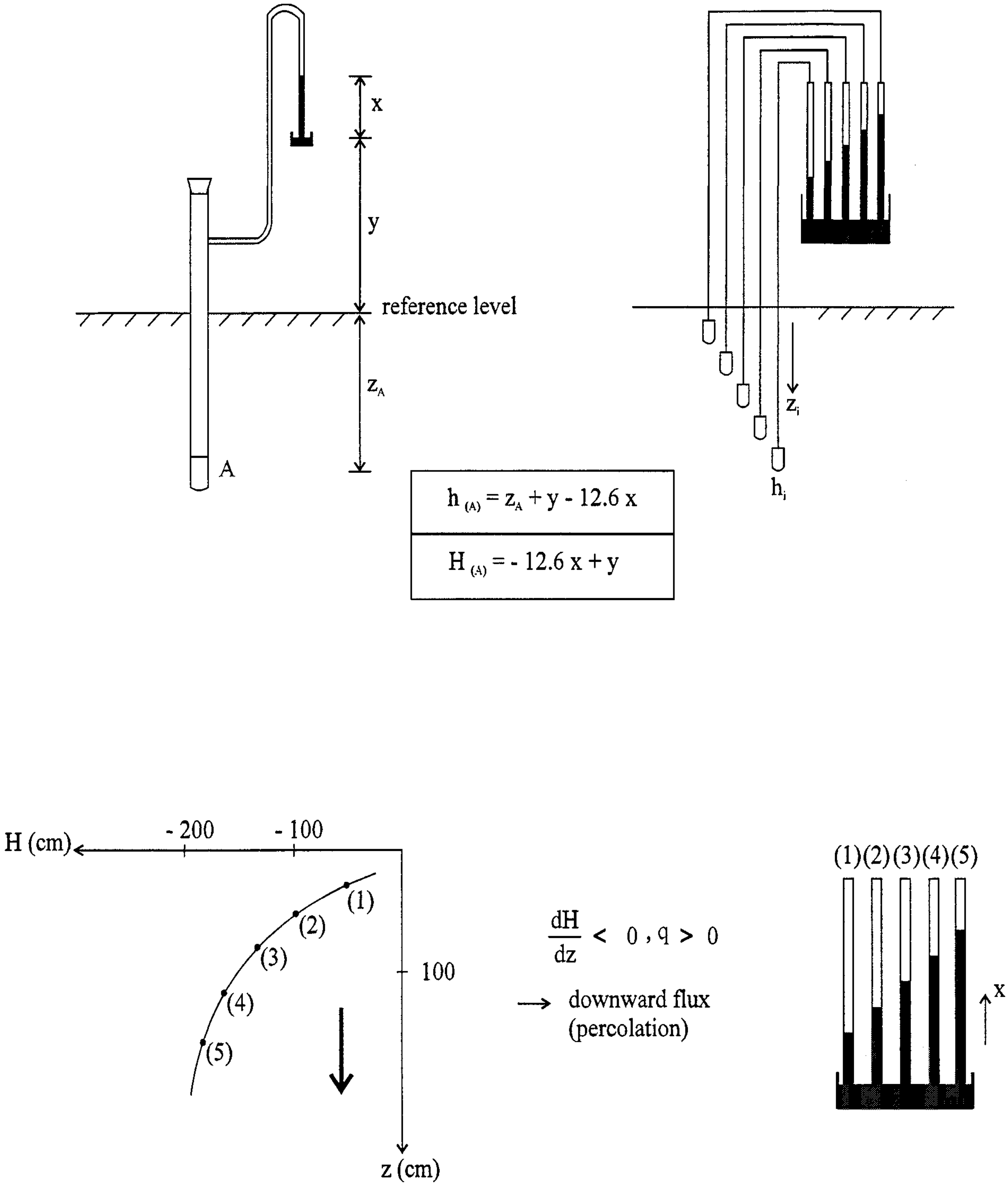
From agricultural point of view it could be of interest to know it there is a recharge of the water table or capillary rise. Therefore only 2 tensiometers are needed with a depth distance of say 25 cm in the control zone. A simple reading of the rise of mercury in the manometer will indicate the flow direction.

Knowing the moisture content θ at the depth between z_1 and z_2, the $K(\theta)$ relation of that soil and the hydraulic head gradient dH/dz, one can calculate the instantaneous water flow q (see equation 13).

3.7.3 Determination of the soil water characteristic curve (or retentivity curve)

The h-θ relation (retentivity curve) of a soil layer in situ can be established :

- knowing the soil water pressure head (h) using tensiometers (see equation 11)
- knowing the soil water content (θ) using the neutron moisture meter



$$h_{(A)} = z_A + y - 12.6\,x$$

$$H_{(A)} = -12.6\,x + y$$

Figure 18. Hydraulic head profiles. The manometers from left to right increase with depth.

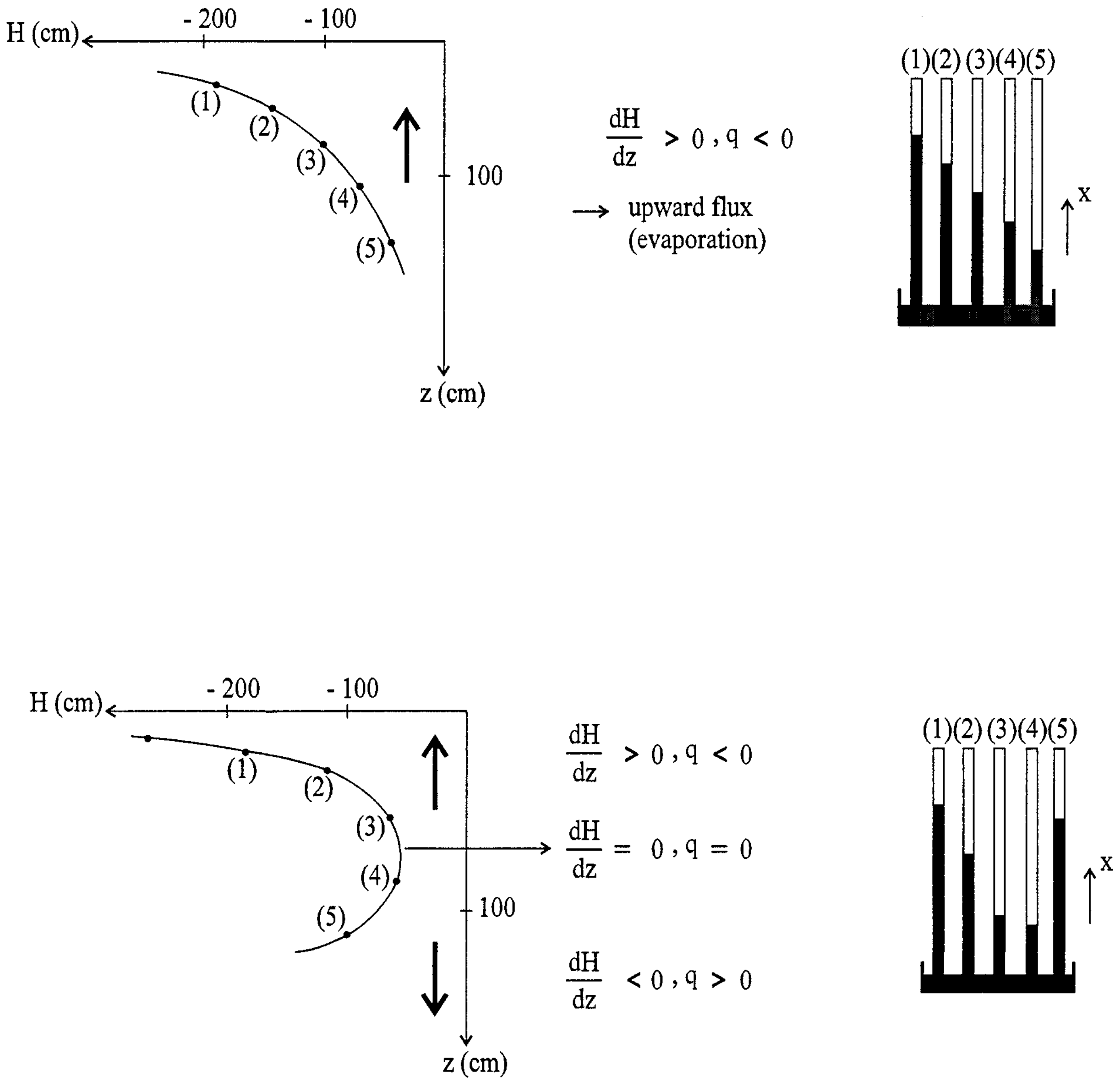


Figure 18bis. Hydraulic head profiles. The manometers from left to right increase with depth.

3.7.4 Scheduling irrigation

The root zone for most agricultural plants is limited to the unsaturated part of the profile because the plant roots do not proliferate in a saturated soil where aeration is limiting. Consequently in a non-saline soil the plant behaviour is largely determined by the matric potential of the soil water. Moreover the plant does not depend as much on the quantity of water present as it does on the water potential.

Water should be applied to the soil when the matric potential is still high enough that the soil can and does supply water fast enough to meet the atmospheric demands without placing the plant under a stress that will reduce yield or quality of the harvested crop.

Although the tensiometers function over only a limited part of the available water range (0 to - 800 cm water) it is usually in this range that plants should be irrigated.

From practical point of view tensiometers are installed at minimum 2 locations. One unit should be placed in the zone of maximum root activity and another near the bottom of the active rootzone.

The time to irrigate is determined by following the matric potential readings in the zone of the greatest root activity. The exact value of the matric potential at which water should be applied is not the same for every crop. A good approximation of that matric potential is available for many common crops. For most crops it is time to irrigate when the top tensiometer reads - 300 to - 500 cm water and the bottom tensiometer begins to indicate drying (Table 1).

Table 1. Matric potential at which water should be applied for maximum yields of various crops grown in deep, well-drained soil that is fertilized and otherwise managed for maximum production. Where two values are given, the higher value is used when evaporative demand is high and the lower value when it is low; intermediate values are used when the atmospheric demand for evapotranspiration is intermediate. (The values are subject to revision as additional experimental data become available). (TAYLOR and ASHCROFT, 1972)

Crop	Matric potential (joules/kg)	Equivalent matric suction (centibars)
Vegetative crops		
Alfalfa	-150	150
Beans (snap and lima)	-75 to -200	75 to 200
Cabbage	-60 to -70	60 to 70
Canning peas	-30 to -50	30 to 50
Celery	-20 to -30	20 to 30
Grass	-30 to -100	30 to -00
Lettuce	-40 to -60	40 to 60
Tobacco	-30 to -80	30 to 80
Sugar cane		
Tensiometer	-15 to -50	15 to 50
Blocks	-100 to -200	100 to 200
Sweet corn	-50 to -100	50 to 100
Turfgrass	-24 to -36	24 to 36
Root crops		
Onions		
Early growth	-45 to -55	45 to 55
Bulbing time	-55 to -65	55 to 65
Sugar beets	-40 to -60	40 to 60
Potatoes	-30 to -50	30 to 50
Carrots	-55 to -65	55 to 65
Broccoli		
Early	-45 to -55	45 to 55
After budding	-60 to -70	60 to 70
Cauliflower	-60 to -70	60 to 70

Crop	Matric potential (joules/kg)	Equivalent matric suction (centibars)
Fruit crops		
Lemons	-40	40
Oranges	-20 to -100	20 to 100
Deciduous fruit	-50 to -80	50 to 80
Avocadoes	-50	50
Grapes		
Early season	-40 to -50	40 to 50
During maturity	< -100	> 100
Strawberries	-20 to -30	20 to 30
Cantaloupe	-35 to -40	35 to 40
Tomatoes	-80 to -150	80 to 150
Bananas	-30 to -150	30 to 150
Grain crops		
Corn		
Vegetative period	-50	50
During ripening	-800 to -1200	800 to 1200
Small grains		
Vegetative period	-40 to -50	40 to 50
During ripening	-800 to -1200	800 to 1200
Seed crops		
Alfalfa		
Prior to bloom	-200	200
During bloom	-400 to -800	400 to 800
During ripening	-800 to -1500	800 to 1500
Carrots		
During seed year at 60 cm depth	-400 to -600	400 to 600
Onions		
During seed year at 7 cm depth	-400 to -600	400 to 600
at 15 cm depth	-150	150
Lettuce		
During productive phase	-300	300
Coffee	Requires short periods of low potential to break bud dormancy, followed by high water potential	

4. SOIL WATER CHARACTERISTIC CURVE

Text from "Soil Physics" by Jury W.R., Gardner W.R. and Gardner W.H. (John Wiley & Sons, Inc., New York)

4.1 Measurement

In rigid porous media, the matric potential as defined in the preceding represents the effect of adsorptive soil solid forces and interfacial curvature on water potential energy. The functional relationship between the matric potential and the gravimetric or volumetric water content is called the water characteristic function or matric potential-water content function $\psi_m(\theta)$. This function may be evaluated by measuring matric potential and water content simultaneously with the methods already discussed during a succession of water content changes. In the laboratory, $\psi_m(\theta)$ may be measured on replicated prepared samples over a large range of water contents. Virtually the entire range from water-saturated soil to very dry soil may be covered by using a hanging water column, a pressure membrane, and equilibration over salt solutions. These devices will be illustrated using the equilibrium principle.

4.1.1 Hanging Water Column (Range -100 cm $< h < 0$)

A hanging water column consists of a water-saturated, highly permeable porous ceramic plate connected on its underside to a water column terminating in a reservoir open to the atmosphere. Water-saturated samples of soil held in rings are placed in contact with the flat plate when the water reservoir height is even with the top of the plate. Then the reservoir is lowered to a new height a distance z below the top of the plate (figure 19).

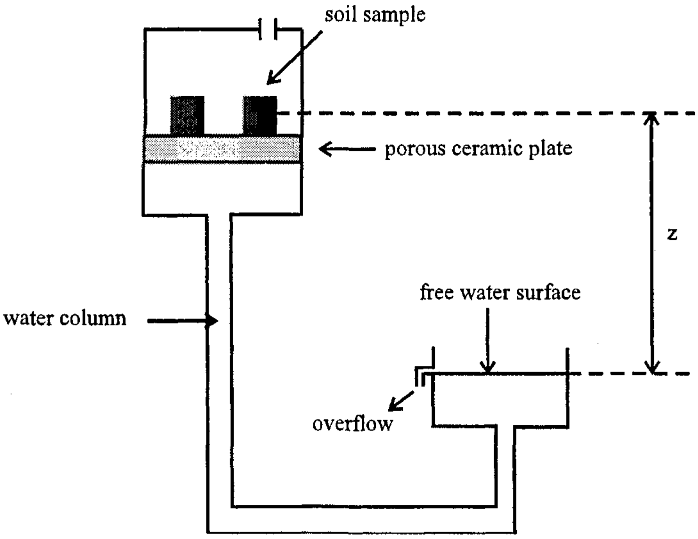


Figure 19.Desaturation of soil water samples to a desired energy state with a hanging water column.

By the equilibrium principle, water will flow from the soil samples through the ceramic to the reservoir until the total water potential of the system is constant. At this time the potential of the free reservoir may be set equal to zero, and at the soil sample height z we may write ($z = 0$; $P = P_{atm}$; neglect solutes) $\psi_m + \psi_z = 0 = \psi_m + \rho_w gz$, or $\psi_m = -\rho_w gz$. When equilibrium has been restored, some of the samples may be removed and their gravimetric or volumetric water content measured. The tube may then be lowered further and a new set of samples measured. If there is good contact between the soil and the ceramic, equilibrium will be reached rapidly (i.e. several hours) since the samples are quite moist. The range of the device is limited chiefly by the space available for lowering the water column.

4.1.2 Pressure Plate (Range - 15000 cm ≤ h ≤ - 300 cm)

The pressure plate consists of an air-tight chamber enclosing a water-saturated, porous ceramic plate connected on its underside to a tube that extends through the chamber to the open air. Saturated soil samples are enclosed in rings and placed in contact with the ceramic on the top side. The chamber is then pressurized, which squeezes water out of the soil pores, through the ceramic, and out the tube (figure 20). At equilibrium, flow through the tube will cease. We may set the total potential equal to zero at the point where the water exits the tube. Inside the plate we may write ($P = P_{atm}$; $z = 0$; neglect solutes) $\psi_{e.p.} + \psi_m = 0 = \psi_m + \Delta P$, or $\psi_m = -\Delta P$. When equilibrium is reached, the chamber may be depressurized and the water content of the samples measured. An assumption is made in this method that the matric potential of the sample does not change as the air pressure is lowered to atmospheric.

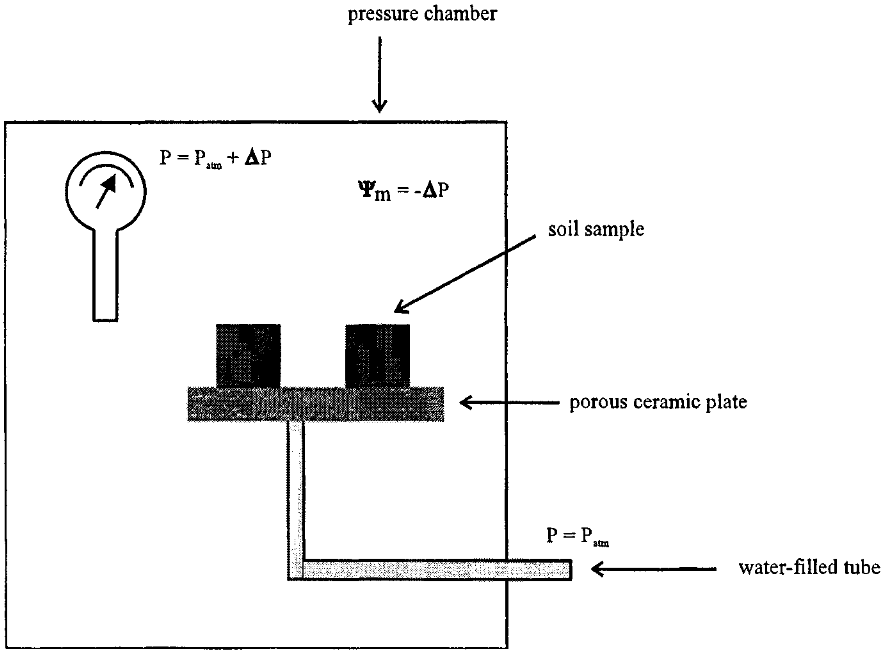


Figure 20.Desaturation of soil water samples to a desired energy state with a pressure plate.

This method may be used up to air gauge pressures of about 15 bars if special fine-pore ceramic plates are used. Since these devices have a very high flow resistance, it may require a substantial amount of time to remove the last small amount of water from the soil. Thus, the time of equilibrium is difficult to estimate.

4.1.3 Equilibration over Salt Solutions ($h < -15000$ cm)

By adding precalibrated amounts of certain salts, the energy level of a reservoir of pure water may be lowered to any specified level. If this reservoir is brought into contact with a moist soil sample, water will flow from the sample to the reservoir. If the sample and the reservoir are placed adjacent to each other in a closed chamber at constant temperature, water will be exchanged through the vapor phase by evaporation from the soil sample and condensation in the reservoir until equilibrium is reached. Since the reservoir is a pool of salt solution, at equilibrium the total potential will be $\psi_t = \psi_{so}$ of the solution. In the soil $\psi_t = \psi_m + \psi_o$ since the air-water interface acts as a solute membrane. Thus $\psi_m = \psi_{so} - \psi_o$, the difference between the solute potentials of the reservoir and the soil. In practice, the soil will usually not be saline enough for its solute potential to be significant compared to ψ_o in the range where these measurements are made. The equilibration time for this method can be shortened by creating a partial vacuum in the chamber. Care should be taken that the sample and the reservoir are at the same temperature, because even small temperature differences will cause the soil and salt solution to equilibrate at very different potentials. Figure 21 shows typical matric potential-volumetric water content curves for a sandy soil and a finer textured soil high in clay measured from soil initially at water saturation. The water characteristic function for a soil desorbed from saturation may be roughly divided into three regions, as shown in the figure. The air entry region corresponds to the region at saturation where the matric potential changes but the water content does not. The minimum suction that must be applied to a saturated soil to remove water from the largest pores is called the air entry suction, which varies from about 5 to 10 cm for sands to much higher values in unaggregated, fine-textured soils. After air begins to enter the system, incremental increases in suction on the soil sample will drain progressively smaller pores, and the water content will drop. This intermediate part of the curve is called the capillary region. When essentially all of the water held in pores has been drained, only the tightly bound water adsorbed to particle surfaces remains. Large changes in matric potentials in this region, called the adsorption region, are associated with small changes in water content. The differences in the shapes of the water characteristic function for the prototype sandy and clay soils in figure 21 may be explained by considering the properties of the bulk solid phases. The clay soil generally has a lower bulk density and hence a higher water content at saturation. The clay soil has very few large pores and a broad distribution of particle sizes. Hence, it decreases gradually in water content with decreases in matric potential. The sandy soil, on the other hand, has much of its water held in large pores that drain at modest suctions. Hence it will have a very rapid decrease in water content in the capillary region. Finally, the clay soil has a very large surface area compared to the sand and will have a large amount of water adsorbed to the surfaces.

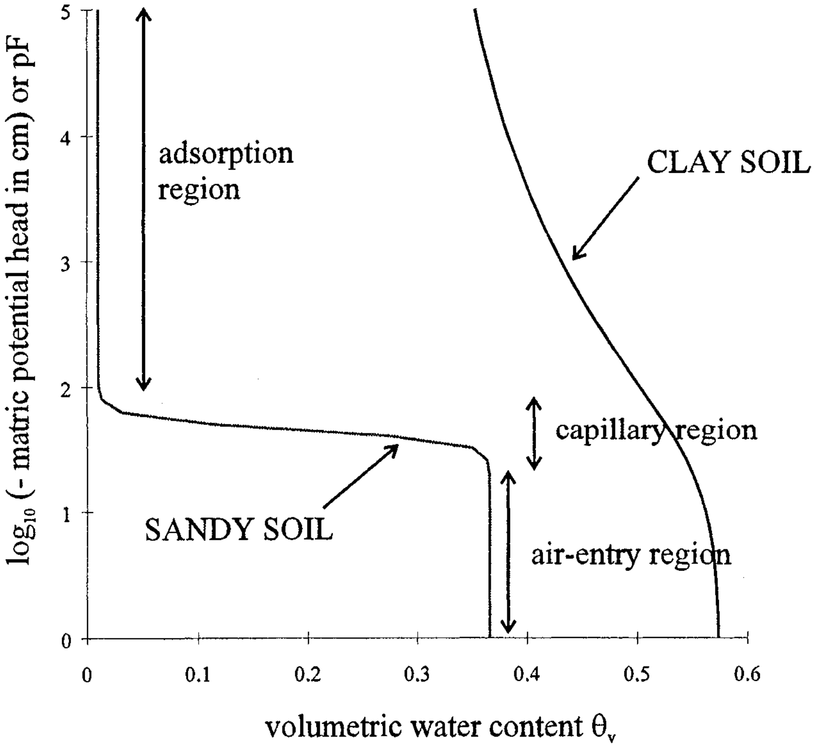


Figure 21.Matric potential-water content function (water characteristic function).

4.2 Hysteresis in Water Content-Energy Relationships

Water content and the potential energy of soil water are not uniquely related because the potential energy state is determined by conditions at the air-water interfaces and the nature of surface films rather than by the quantity of water present in pores. Soil pores are highly variable in size and shape and interconnect with each other in a variety of ways. Common to porous media are so-called bottleneck pores, which have large cavities but narrow points of connection to adjacent pores. Water is held most tenaciously in small pores, which fill first when water is admitted to a system. But they do not always empty again during drying in the same order as they were filled. The factors involved in hysteresis may be discussed most clearly by assuming that the soil is initially completely devoid of water and subsequently has no air phase (just liquid water and its vapor). If water is added at this point to the system, small pores fill first, followed by successively larger and larger pores until all pores are filled and the matric potential is zero. At intermediate values of saturation, with enough water in the system so that vapor-water interfaces can exist between particles and in small pores, the curvature of such interfaces is given by $\Delta P = - 2\gamma/R$ where the pressure difference ΔP refers to the difference in pressure between the vapor and the liquid water. Water content and water potential in such a system will follow the wetting curve in figure 22. Some small pores could be isolated during wetting, so that they might remain dry while larger pores are filled. However, this would not be the case at equilibrium in the absence of air, in as much as vapor transfer would assure the wetting of all pores small enough to retain water at a particular matric potential.

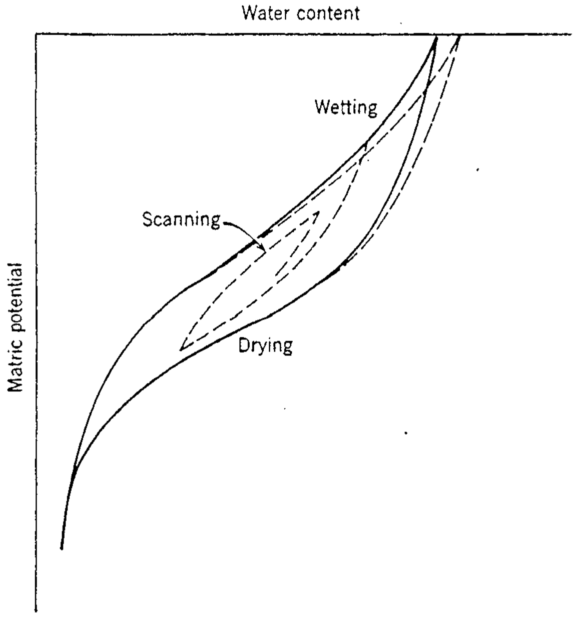


Figure 22. Wetting and drying curves and scanning curves.

If the system is dried either by evaporating water or by bringing the soil into contact with a dry, porous material that pulls water away from the system, pores will begin to empty, generally from large to small. However, liquid water may now be trapped in large pores in such a way that they will not empty in the order that they filled. Water will be held in large pores until conditions are reached where at least one interconnecting smaller pore can empty; at this time the larger pore quickly empties. The sudden release of a relatively large amount of water from a large pore floods surrounding pores and increases the matric potential in them temporarily. If matric potential were monitored in a small porous system having discrete differences in pore size, the matric potential-water content relationship for drying might be saw-toothed, as indicated by the drying curve in figure 22. In real soil systems, the pore size distribution contains many pores in all size ranges, and the water content and potential distributions tend to average out so that a smooth curve is obtained. However, the water content for a given matric potential is higher than for the wetting system, as is shown by the drying curve in figure 22. This principle is illustrated by the pore-water system in figure 23a. Here it may be observed that the curvature of the vapor-water interface in the small pores of two identical systems can be in equilibrium with each other even though their water contents are grossly different. The matric potential is determined by the curvature of the liquid interface which at equilibrium would be precisely the same in the small pores connecting with the large pore in each case. This ideal representation commonly is called the "ink bottle principle", which refers to the fact that an ink bottle has a small opening into a large cavity. Large pores that are interconnected by smaller pores are not required for hysteresis to occur. It is possible for a single pore to contain the same amount of water at two different water potentials, as is shown in figure 23b. Water that condenses initially into such a capillary from a humid environment is shown by the diagonal cross-hatched area. However, as condensation proceeds, water at the center finally coalesces and a concave meniscus is formed as a consequence of surface tension forces (vertical cross-hatching). Whereas positive pressure existed in the system before coalescence, the system suddenly goes under negative pressure as a consequence of its new configuration. In the example in figure 23b, water in a cylindrical pore about 1 cm in length and 0.1 cm in radius would have a slight positive potential of about 0.15 mbar immediately before coalescence and a potential of - 1.5 mbars immediately afterward.

Surface wetting can also induce hysteresis. Unless particle surfaces are meticulously clean, they will form a nonzero contact angle with water when wetted (figure 23c). This results in thicker films than would be present in the drying phase where water films are drawn tightly over the surface by adsorptive forces.

Thus far the discussion of hysteresis has not involved the presence of air in the system, which can introduce additional differences between water content at a given matric potential during wetting and drying. As small pores and interstices between particles fill with water, air may become entrapped in large pores. Continued water entry into such pores will cause a buildup of air pressure. Since air is slightly soluble in water, pressures in such pores may gradually be relieved, which will sometimes allow more complete pore filling. However, the order of filling and access to pores will be influenced by entrapped air, so that water content is still permanently affected despite the fact that some air may go into solution and disappear.

In the absence of air, the water potential-water content relationship for complete wetting and complete drying will follow approximately the dashed line-solid line loop shown in figure 22. However, this loop is not exactly reproducible because of the inherent difficulty associated with repetition of the exact order of pore filling over each cycle. When air is present in the system, the curves are offset somewhat toward the dry side (solid curve, figure 22). If a soil is completely wetted so that no air is present and then is dried, it will follow the dashed-solid curve down; upon rewetting in the presence of air, it will follow the solid wetting curve and will not return to the starting point because of the presence of entrapped air. If the process is reversed at any time during wetting or drying, curves like those in the interior (dotted curves) of the hysteretic envelope are produced. These interior curves have been called scanning curves; the curves that form the hysteretic envelope have been called characteristic curves or the soil moisture characteristic. The wetting curve of the hysteretic envelope is commonly known as a sorption curve and the drying curve a desorption curve.

Hysteretic phenomena also exist in soil materials as a consequence of shrinking and swelling, which can affect microscopic pore size geometry as well as overall bulk density. Both factors would lead to a volumetric water content for a given energy state that differs from that which would exist if the soil matrix remained fixed. Shrinking and swelling often take place slowly and usually irreversibly, particularly when organic matter is involved; this complicates the evaluation of their contribution to hysteresis. Experimental observations do not always reveal that measurements involve true hysteresis and permanent or semipermanent changes in the porous system.

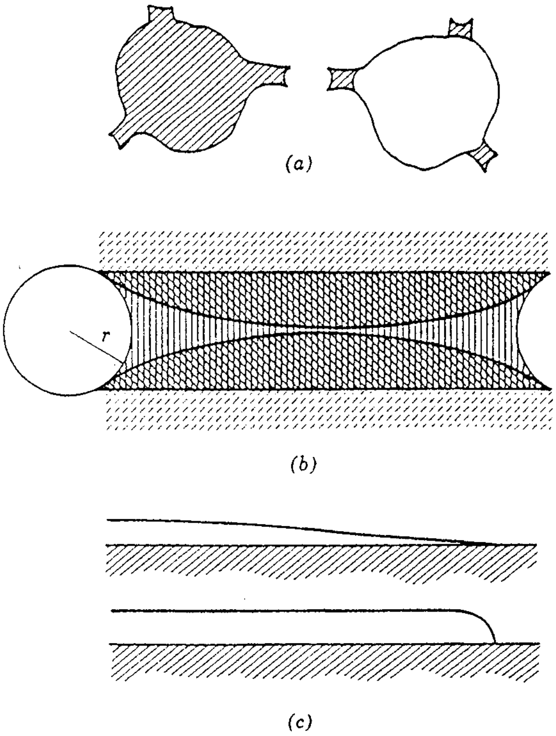


Figure 23.Diagrammatic representation of three forms of water content-matric potential hysteresis.

5. HYDRAULIC CONDUCTIVITY OF SATURATED AND UNSATURATED SOIL

5.1. Representative elementary volume

The movement of water through soils takes place in the tortuous channels between the soil particles with velocities varying from point to point. Darcy's law does not consider this microscopic flow pattern between the particles but instead assumes the water movement to take place in a continuum with a uniform flow averaged over space. It therefore describes the flow of water macroscopically in volumes of soil much larger than the size of the pores. In fact it can be used to describe only the macroscopic flow of water through soil regions of volume greater than some representative elementary volume that encompasses many soil particles.

The concept of representative elementary volume of a porous material is most easily illustrated by considering the measurement of the water content of a sample of unstructured 'uniform' saturated soil, starting with a very small volume and then increasing the sample size. For very small volumes comparable to the size of the soil particles, the sample volume would include only solid matter if located wholly within a soil particle, giving zero soil-water content. If located wholly in a pore, it would contain only water, giving a soil-water content of one. All values between zero and one are possible when the sample is located partly within a soil particle and partly within the pore. As the volume size is increased, the lower limit of measured water content increases while the upper limit decreases, as shown in figure 24. When the sample is large enough, repeated measurements on random samples of the soil give the same value of soil-water content. The smallest sample volume that produces a consistent value is the representative elementary volume. Measurements of hydraulic conductivity and other soil properties must be made on volumes larger than this volume. While additive soil properties, such as the water content, can be obtained by averaging a large number of measurements made on smaller volumes within the representative elementary volume, the hydraulic conductivity cannot be obtained in this way because of the interdependent complex pattern of flows that this property embraces.

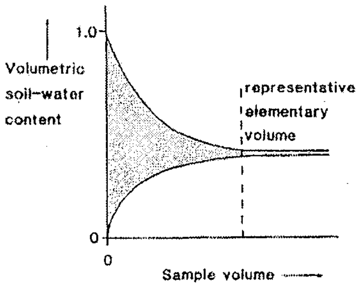


Figure 24.Measurement of soil-water content of a saturated 'uniform' soil (Smith and Mullins, 1991).

Figure 24 illustrates the variability of a soil physical property that exists in all porous materials at a small enough scale because of their particulate nature. Variability can also be present in soils at larger scales. For example, in aggregated and structured soils where a distribution of macropores between the aggregates or peds is superimposed on the interparticle micropore space, the soil-water content would vary with sample size as shown in figure 25: only when the sample size encompasses a representative sample of macropore space do we have a representative volume. This volume will be characteristic of the soil's structure that determines the hydraulic conductivity.

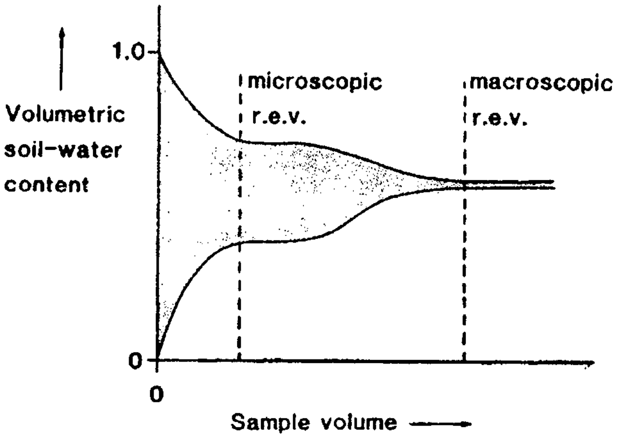


Figure 25. Measurement of soil-water content of a saturated soil with superimposed macrostructure (r.e.v. = representative elementary volume) (Smith and Mullins, 1991).

It is only in materials that show behaviour similar to that depicted in figure 24 that continuum physics, such as that implied by Darcy's law, can be applied macroscopically without difficulty to soil-water flow phenomena. In materials such as that illustrated in figure 6, boundary conditions at the surfaces of the aggregates and ever, for saturated conditions, as long as sufficiently large volumes are considered, continuum physics can still be applied to water flows at this larger scale using an appropriate value of hydraulic conductivity.

5.2. Saturated hydraulic conductivity 'Ks'

5.2.1 Laboratory methods

- constant water head
- variable water head

5.2.2 Field methods

- saturated zone
  - auger hole method
  - piezometer method

- unsaturated zone
  - double ring method
  - Guelph/disk permeameter

5.2.3 Models

- capillary tube model
- pedotransferfunctions

5.3. Unsaturated hydraulic conductivity K(θ)/K(h)

5.3.1 Steady state flow

- laboratory method
  - standard method
  - crust method
  - sprinkler method
- field method
  - crust method
  - sprinkler method

5.3.2 Non steady state/ transient flow

- laboratory method : evaporation method of wind
- field method
  - internal drainage method
  - water balance method

5.3.3 Models

- capillary tube model
- model of Mualem

5.4. Models for determination of the soil water characteristic curve and the unsaturated hydraulic conductivity K(θ)

5.4.1 Introduction

The soil moisture characteristic curve (or *pF* curve) can be determined based on a set of discrete measuring points determined in the laboratory or in the field. To represent the curve graphically, one can draw a best fitting curve by hand on mm paper or an analytical equation can be used to fit a curve through the set of measuring points. Because determination of the soil moisture characteristic curve in the lab or in the field is time consuming, models have been introduced to calculate the soil moisture characteristic curve based on readily available physical soil characteristics.

As regards the unsaturated hydraulic conductivity $K(\theta)$, can be determined in the laboratory (e.g. Wind method) or in the field (e.g. internal-drainage method). Since this is also quite laborious, models have been developed that calculate $K(\theta)$ based on the saturated hydraulic conductivity K_s and the soil moisture characteristic curve, or on readily available physical soil characteristics.

5.4.2 Fitting pF curve to a discrete set of measuring points by modelling

5.4.2.1 The model of van Genuchten

5.4.2.1.1 Principle

There exists a non-linear relationship between the soil moisture content and de matric potential. To fit a curve through a discrete set of measuring points closed-form analytical equations can be used. One of the most widely used equations is that of van Genuchten (1978, 1980) :

$$\theta = \theta_r + (\theta_s - \theta_r) \cdot \left(\frac{1}{1 + (\alpha \cdot |h|)^n} \right)^m \qquad (14)$$

where θ = volumetric moisture content ($m^3 \cdot m^{-3}$),

θ_r = residual volumetric moisture content ($m^3 \cdot m^{-3}$),

θ_s = volumetric moisture content at saturation ($m^3 \cdot m^{-3}$),

h = matric potential (dimensionless, but expressed in cm H_2O),

α, n en m = parameters (dimensionless).

If, as proposed by van Genuchten,

$$m = 1 - \frac{1}{n} \quad \text{for } n > 1 \tag{15}$$

equation (15) contains 4 independent parameters (θ_r, θ_s, α and n), which have to be estimated for the observed soil-moisture retention data. This can be done by applying a non-linear least-squares analysis according to the Marquardt (1963) algorithm. This is an iterative method implying an initial estimate of the parameters.

The Marquardt algorithm can be performed by using mathematical software programmes such as Mathcad®. On the other hand most statistical software packages, as e.g. SPSS® contain the Marquardt algorithm. Graphical software such as SigmaPlot® also have the possibility to perform a non-linear regression. Finally, the RETC code, written in the programme language Fortran, can be used to determine the parameters of the van Genuchten model (van Genuchten et al., 1991; Yates et al., 1992).

5.4.2.1.2 Graphical interpretation and physical meaning of the parameters

As stated above the model of van Genuchten contains, after imposing the restriction $m = 1\text{-}1/n$, four independent parameters, namely θ_r, θ_s, α and n. The value for the **volumetric moisture content at saturation** θ_s is, in theory, equal to the porosity ε of the soil. It is the latter value that will be taken as initial estimate for θ_s.

From a practical point of view the **residual volumetric moisture content** θ_r can be defined as the pressure head at some large negative value, e.g. at the permanent wilting point (h =-15,495 cm H_2O or $pF = 4.2$). In some cases, however, significant decreases on h are likely to result in further desorption of water, especially on fine-textured soils. It seems that such further changes in θ are fairly unimportant for most practical field problems.

To obtain estimates of the remaining parameters α and n a characteristic point P on the curve has to be defined (see figure 26). This point is located halfway between θ_r and θ_s. From equation (14) it follows that the **parameter** α is inversely proportional with the pressure head at P, h_P :

$$\alpha = \frac{1}{h_P} \cdot \left(2^{\frac{n}{n-1}} - 1 \right)^{\frac{1}{n}} \tag{16}$$

For n large enough, and hence m approaches 1, α is approximately equal to the inverse value of h_P. Most soils, however, have a m significant smaller than 1 ($n < 10$). The value for α then increases exponentially with decreasing m value.

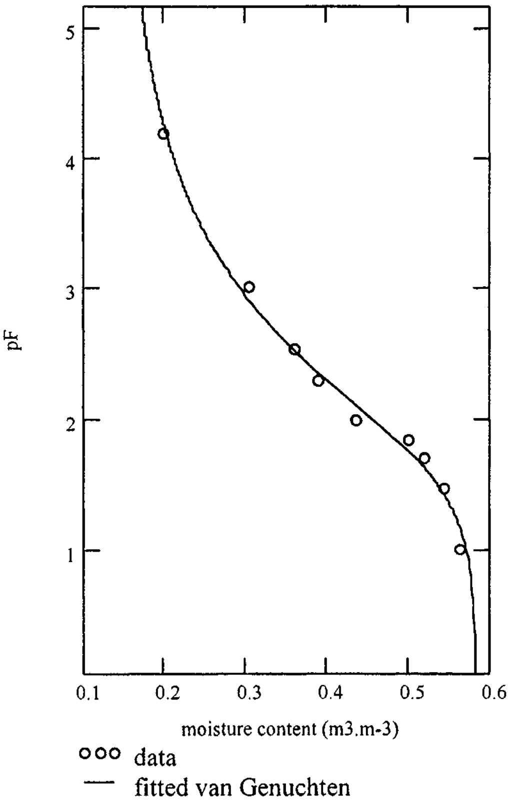


Figure.26 pF curve of a heavy sandy clay soil

Finally, the **parameter *n*** can be obtained be differentiation of equation (14) $(\mathrm{d}\theta \cdot \mathrm{d}h^{-1})$ and by substituting equation (16). This gives a relation between $h \cdot \mathrm{d}\theta \cdot \mathrm{d}h^{-1}$ and n. Further calculating results in following equation for n :

$$n = \begin{cases} e^{(0{,}8 \cdot S_P)} & (0 < S_P \leq 1) \\ \left(\dfrac{0{,}5755}{S_P} + \dfrac{0{,}1}{S_P^{\,2}} + \dfrac{0{,}025}{S_P^{\,3}} \right)^{-1} & (S_P > 1) \end{cases} \qquad (17)$$

where S_P = slope at point P.

The slope S_P is defined as :

$$S_P = \frac{1}{\theta_s - \theta_r} \cdot \left| \frac{\mathrm{d}\theta}{\mathrm{d}(\log h)} \right| \qquad (18)$$

From equation (17) and (18) it is clear that as the ratio $\mathrm{d}\theta \cdot \mathrm{d}(\log h)^{-1}$ increases, and hence the flatter the curve at P, the value for n will increase.

5.4.2.2 Other models

5.4.2.2.1 The model of Brooks and Corey

The model of Brooks and Corey (1964) can be expressed as :

$$\theta = \theta_r + (\varepsilon - \theta_r) \cdot \left(\frac{h_b}{h}\right)^{\lambda} \tag{19}$$

where ε = porosity ($m^3 \cdot m^{-3}$),
h_b = bubbling pressure or air entry value (cm H_2O),
λ = pore size index (-).

The model of Brooks and Corey, and the model of Campbell (see 5.4.2.2.2.) do not permit a representation of the total soil moisture retention curve - they only represent that part of the curve for pressure heads lower than the air entry value -, whereas the model of van Genuchten does so. The model of van Genuchten does not account for the air entry value but does have an inflection point, allowing the van Genuchten model to perform better than the Brooks and Corey, and the Campbell model for many soils, particularly for data near saturation.

5.4.2.2.2 The model of Campbell

The model of Campbell (1974) can be represented by :

$$\theta = \varepsilon \cdot \left(\frac{H_b}{h}\right)^{\frac{1}{b}} \tag{20}$$

where H_b = bubbling pressure or air entry value (cm H_2O),
b = constant.

5.4.3 Methods to obtain the pF curve without laboratory determinations

5.4.3.1 Soil texture reference curves

The simplest method for estimating the $h(\theta)$ relationship is to use soil texture reference curves. These curves give the soil water characteristic curve for several soil textural classes (see figure 27).

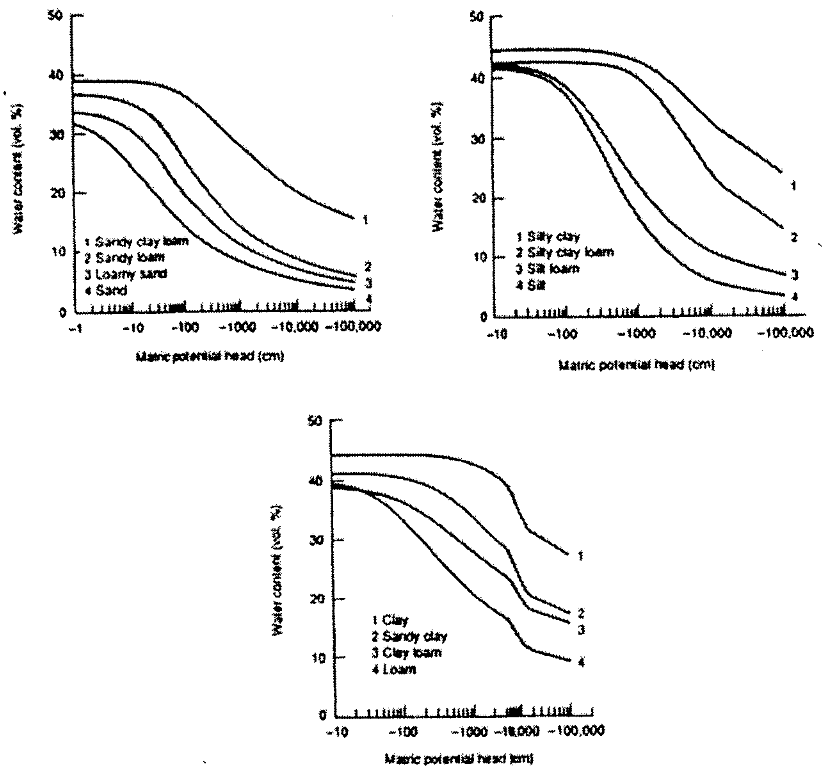


Figure 27 Soil water retention curves for USDA soil textures (Rawls et al., 1993)

5.4.3.2 Estimating soil water retention curves from soil physical characteristics

5.4.3.2.1 Introduction

The amount of water retained at high soil water potentials (> -10 kPa) depends primarily on the pore size distribution, and is thus strongly affected by soil structure, bulk density, and porosity. As the soil dries out (< -150 kPa), water adsorption becomes critical, and those soil properties that affect specific surface become important such as texture, organic matter content, and clay mineralogy. Most of the models that are developed to estimate the soil moisture characteristic curve from soil physical characteristics, the so called pedotransfer functions, make use of one or more of those characteristics.

Two approaches can be distinguished for estimating soil water retention characteristics from soil properties. The first approach estimates soil water retention *values* from soil physical properties using regression analysis. The second approach estimates *parameters* for water retention models (like van Genuchten, Brooks and Corey, and Campbell) from soil physical properties using regression analysis. An overview of both approaches is given in Rawls et al. (1991) and van Genuchten et al. (1992). More recently Kern (1995) evaluated six of the most cited models that use soil physical characteristics. These include the models of Gupta and Larson (1979), Rawls et al. (1982, 1983), De Jong et al. (1983), Cosby et al. (1984), Saxton et al. (1986) and Vereecken et al. (1989).

5.4.3.2.2 Estimation of specific points on the soil water retention curve

The most cited models to relate soil properties to the soil moisture held at specific matric potentials are those of Gupta and Larson (1979) and Rawls et al. (1982, 1983). To increase the accuracy of the regression equations using physical soil properties, Rawls et al. introduced the moisture content held at -1500 kPa (*pF* 4.2) or at both -33 kPa (*pF* 2.54) and -1500 kPa. Adding these variables, which require more costly and/or time-consuming laboratory procedures, increase the explained variation from 76% to 95%. The model of Rawls et al. can thus be written as :

$$\theta_x = a + b \cdot sand + c \cdot silt + d \cdot clay + e \cdot O.M. + f \cdot \rho_b + g \cdot \theta_{0{,}33\,\mathrm{bar}} + h \cdot \theta_{15\,\mathrm{bar}} \tag{21}$$

where θ_x = predicted soil moisture content for a given suction x ($\mathrm{cm^{-3} \cdot cm^{-3}}$),
$sand$ = percentage sand (%),
$silt$ = percentage silt (%),
$clay$ = percentage clay (%),
$O.M.$ = percentage organic matter (%),
ρ_b = bulk density ($\mathrm{g \cdot cm^{-3}}$),
$\theta_{0{,}33\,bar}$ = moisture content at a matric potential of - 33 kPa ($\mathrm{cm^{-3} \cdot cm^{-3}}$),
$\theta_{15\,bar}$ = moisture content at a matric potential of - 1500 kPa ($\mathrm{cm^{-3} \cdot cm^{-3}}$),
a tot h = regression coefficients from tables.

Thus, in its simplest form the model of Rawls et al. exists of the first 6 terms of equation (21). The values for regression coefficients differ depending on the chosen simplicity of the model (6, 7, or 8 terms) and the given matric potential. The tables of Rawls et al. include 12 different matric potentials, and hence a set of 12 discrete points can be determined. A best fitting curve can then be drawn through this set of points by modelling (e.g. van Genuchten).

The model of Gupta and Larson is similar with this difference that the intercept *a* is equal to zero (and hence the regression coefficients are different).

5.4.3.2.3 Estimation of soil water retention model parameters

With the increased interest on modelling, the need for a continuous function describing the soil water retention curve has become very important. Frequently water retention data is fitted to a water retention model (see 5.4.2) and the model parameters subsequently related to physical soil properties using regression analysis.

One approach to relate the model parameters to soil properties is to develop average parameter values as a function of soil texture classes. Tables are available with values for the parameters of the different models. Clapp en Hornberger (1978), De Jong (1982) and Raws et al. (1982) calculated, for the different USDA textural classes, values for the models of Brooks and Corey, and Campbell. Vereecken et al. (1989) calculated the values for the parameters of the van Genuchten models for the seven textural classes of the Belgian textural triangle (see table 2).

The second approach uses linear regression to relate the model parameters to soil physical properties. Examples of this approach are the pedotransfer functions of Bloemen (1977), Cosby et al. (1984) and Rawls and Brakensiek (1985) for the Brooks and Corey model. Using the correspondence between the model parameters given in 2, i.e. $H_b = h_b$, $b = 1/\lambda$, $\alpha = h_b^{-1}$ and $n = \lambda + 1$, the same pedotransfer functions can be used with the Campbell and the van Genuchten model. Bloemen developed empirical expression relating the model parameters to a particle size distribution index and the median of particle size, whereas in Cosby et al. the parameters are function of percentage clay and sand. Rawls and Brakensiek included also porosity. Saxton et al. (1986) used the percentage clay and sand to calculate the parameters of a model that was derived from the model of Campbell. Widely used pedotransfer functions for the model of van Genuchten are those of Vereecken et al. (1989) which were developed based on the physical characteristics of 182 horizons of 40 different Belgian soils :

$$\theta_s = 0.81 - 0.283 \cdot \rho_b + 0.001 \cdot clay$$

$$\theta_r = 0.015 + 0.005 \cdot clay + 0.014 \cdot C$$

$$\ln(\alpha) = -2.486 + 0.025 \cdot sand - 0.351 \cdot C - 2.617 \cdot \rho_b - 0.023 \cdot clay$$

$$\ln(n) = 0.053 - 0.009 \cdot sand - 0.013 \cdot clay + 0.00015 \cdot sand^2$$

where C = carbon content (%).

De Jong et al. (1983) followed a similar approach but calculated the parameters from percentage organic matter, silt and clay. These parameters are then substituted in a model that expresses gravimetric moisture content as a function of matric potential by linear regression.

A third more recent approach introduced by Tyler and Wheatcraft (1990) uses fractal mathematics and scaled similarities to show that the empirical constant in the Arya and Paris (1981) model is equivalent to the fractal dimension of the tortuous fractal pore. This information can be used to predict soil water retention from measured particle size distributions.

5.4.4 Determination of the unsaturated hydraulic conductivity K(θ) by modelling

5.4.4.1 Introduction

To determine the unsaturated hydraulic conductivity $K(\theta)$, several approaches can be distinguished. The most common technique is to determine $K(\theta)$ from the pore-size distribution and the saturated hydraulic conductivity K_s. The pore-size distribution can be derived from the soil moisture characteristic curve. The K_s has to be determined in the laboratory or in the field, or has to be calculated based on soil physical properties.

The $K(\theta)$ relationship can also be determined from readily available soil physical characteristics.

5.4.4.2 Estimating the unsaturated hydraulic conductivity K(θ) from water retention data

5.4.4.2.1 Introduction

A popular alternative to direct measurement of $K(\theta)$ is the use of theoretical methods which predict $K(\theta)$ from more easily determined field or laboratory water retention data. Theoretical methods are usually based on statistical pore-size distribution models which assume water flow through cylindrical pores, and incorporate the equations of Darcy and Poiseuille. Three broad groups of models predicting the unsaturated hydraulic conductivity from measured water retention data can be identified, i.e., the models proposed by Childs and Collis-George (1950), Burdine (1953) and Mualem (1976).

Table 2 Mean, minimum and maximum value for the parameters of the van Genuchten model for the seven textural classes according to the Belgian classification (N is the number of observations) with $m = 1$

Textural class	N	Mean	Minimum	Maximum
Heavy clay (U)	3			
θ_s		0.55	0.52	0.57
θ_r		0.27	0.25	0.29
$\alpha \cdot 10^{-3}$		1.60	0.60	2.40
n		0.66	0.57	0.71
Clay (E)	10			
θ_s		0.44	0.39	0.49
θ_r		0.16	0.00	0.30
$\alpha \cdot 10^{-3}$		2.00	2.00	7.77
n		0.63	0.30	0.96
Loam (A)	33			
θ_s		0.42	0.38	0.46
θ_r		0.11	0.04	0.14
$\alpha \cdot 10^{-3}$		1.56	0.40	4.20
n		0.80	0.42	1.44
Sandy loam (L)	55			
θ_s		0.41	0.36	0.48
θ_r		0.09	0.00	0.22
$\alpha \cdot 10^{-3}$		2.29	0.20	5.70
n		0.86	0.32	1.23
Light sandy loam (P)	10			
θ_s		0.37	0.35	0.42
θ_r		0.09	0.07	0.11
$\alpha \cdot 10^{-3}$		2.50	0.61	5.60
n		0.92	0.58	1.17
Loamy sand (S)	10			
θ_s		0.41	0.32	0.53
θ_r		0.09	0.07	0.14
$\alpha \cdot 10^{-3}$		7.42	3.41	20.61
n		1.21	0.76	1.27
Sand (Z)	62			
θ_s		0.39	0.30	0.54
θ_r		0.04	0.00	0.10
$\alpha \cdot 10^{-3}$		12.30	1.60	37.10
n		1.68	0.69	2.62

5.4.4.2.2 Estimating K(θ) based on the model of Childs and Collis-George

The original equation by Childs and Collis-George (1950) has been later modified by Marshall (1958), Millington en Quirk (1959) and Jackson (1972). Jackson formulated the unsaturated hydraulic conductivity at moisture content θ_i ,$K(\theta_i)$, as :

$$K(\theta_i) = K_s \cdot \left(\frac{\varepsilon_i}{\varepsilon_1}\right)^p \cdot \frac{\sum_{j=i}^{m}\left((2\cdot j + 1 - \cdot 2\cdot i) \cdot h_j^{-2}\right)}{\sum_{j=1}^{m}\left((2\cdot j - 1) \cdot h_j^{-2}\right)} \tag{22}$$

where K_s = saturated hydraulic conductivity ($\mathrm{m\cdot s^{-1}}$),
ε_l = amount of pores filled with water = θ_i ($\mathrm{cm^3\cdot cm^{-3}}$),
h_j = pressure head at midpoint of each θ increment $j = i$ (m),
i = number of moisture content class where $i = 1, 2, \ldots m$, and $i = 1$ at saturation and m = the total number of increments,
p = arbitrary constant assigned values of 0 to 4/3 by various workers (a value of unity was found to be satisfactory by Jackson) (-).

The saturated hydraulic conductivity K_s can be derived from field or laboratory measurements, or can be calculated based on statistical pore-size distribution models. Childs and Collis-George (1950) defined K_s as :

$$K_s = \frac{\gamma^2}{2\cdot\eta\cdot\rho_w\cdot g\cdot\tau} \cdot \Delta\theta \cdot \sum_{i=1}^{n} h_i^{-2} \tag{23}$$

where γ = surface tension of water ($\mathrm{N\cdot m^{-1}}$),
η = viscosity of water ($\mathrm{kg\cdot m^{-1}\cdot s^{-1}}$),
ρ_w = density of water ($\mathrm{kg\cdot m^{-3}}$),
g = gravitational acceleration ($\mathrm{m\cdot s^{-2}}$),
τ = tortuosity ($\mathrm{m\cdot m^{-1}}$),
$\Delta\theta$= θ increment ($\mathrm{m^3\cdot m^{-3}}$),
n = number of different capillary size classes in the bundle of tubes making up the soil column,
h_i = pressure head at midpoint of each θ increment i (m).

Marshall (1958) and Millington and Quirk (1959) defined K_s as :

$$K_s = \frac{\gamma^2}{2 \cdot \eta \cdot \rho_w \cdot g} \cdot \frac{\varepsilon_1^p}{n^2} \cdot \sum_{j=1}^{m} \left((2 \cdot j - 1) \cdot h_j^{-2} \right) \tag{24}$$

5.4.4.2.3 Estimating K(θ) based on the model of Burdine

Burdine's (1953) model was later applied by Brooks and Corey (1964) to derive their classical function for the unsaturated hydraulic conductivity $K(\theta)$:

$$K(\theta) = \left(\frac{\theta - \theta_r}{\varepsilon - \theta_r} \right)^n \tag{25}$$

where $n = 3 + 2/\lambda$.

5.4.4.2.4 Estimating K(θ) based on the model of Mualem

Van Genuchten (1980) used the model of Mualem (1976) to derive his well-known equation for unsaturated hydraulic conductivity $K(\theta)$ or $K(h)$:

$$K(\theta) = K_s \cdot \left(\frac{\theta - \theta_r}{\theta_s - \theta_r} \right)^{\frac{1}{2}} \cdot \left(1 - \left(1 - \left(\frac{\theta - \theta_r}{\theta_s - \theta_r} \right)^{\frac{1}{m}} \right)^m \right)^2 \tag{26}$$

$$K(h) = K_s \cdot \frac{\left(1 - \left(\alpha \cdot |h| \right)^{n-1} \cdot \left(1 + \left(\alpha \cdot |h| \right)^n \right)^{-m} \right)^2}{\left(1 + \left(\alpha \cdot |h| \right)^n \right)^{\frac{m}{2}}} \tag{27}$$

where α, n and m = parameters from the van Genuchten model.

5.4.4.3 Estimating the unsaturated hydraulic conductivity $K(\theta)$ from soil physical characteristics

5.4.4.3.1 Soil texture reference curves

As for the soil moisture retention curve, the K-θ relationship can be derived from soil texture reference curves (see figure 28) The saturated hydraulic conductivity K_s can also be derived from soil texture reference curves (see figure 29).

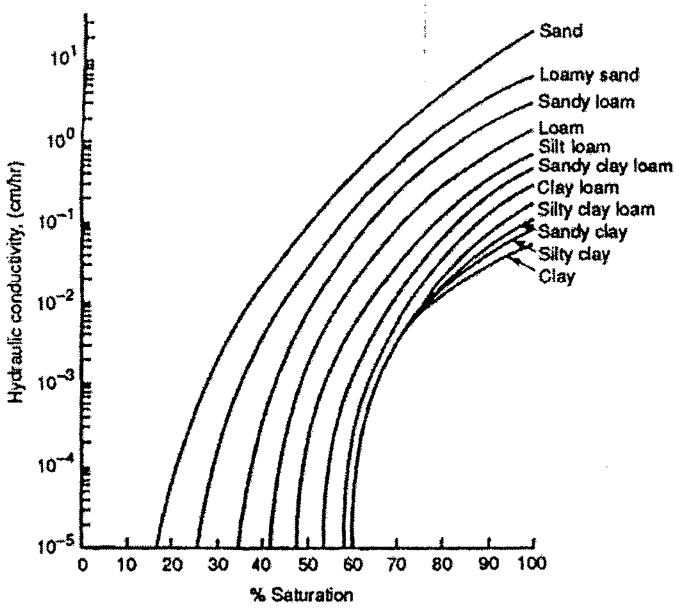


Figure 28 Hydraulic conductivity sorted by soil texture (Rawls et al., 1982)

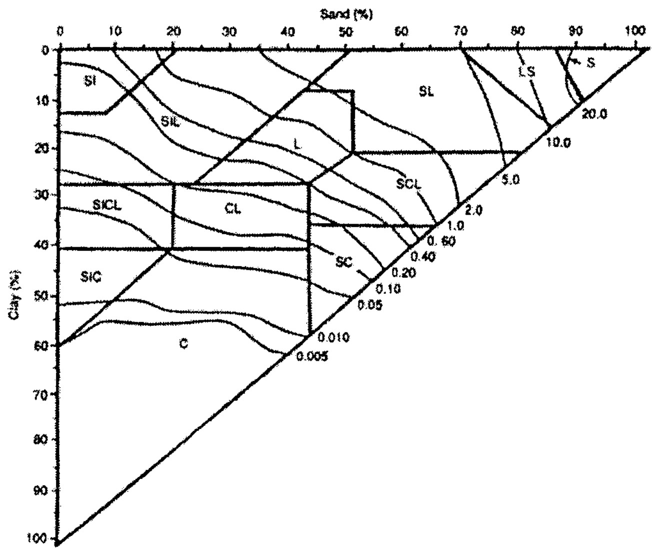


Figure 29 Saturated hydraulic conductivity K_s for USDA soil texture triangle (Rawls and Brakensiek, 1985)

5.4.4.3.2 Pedotransfer functions to estimate K(θ)

Very popular pedotransfer functions to determine K_s and $K(\theta)$ or $K(\psi)$ are those proposed by Campbell (1985) :

$$K_s = 4\cdot 10^{-3}\cdot\left(\frac{1.3}{\rho_b}\right)^{1.3\cdot B}\cdot e^{(-6.9\cdot m_c\,-\,3.7\cdot m_s)}\cdot 3.53\cdot 10^{-3} \tag{28}$$

$$K(\theta) = K_s\cdot\left(\frac{\theta}{\theta_s}\right)^{m} \tag{29}$$

$$K(\psi) = K_s\cdot\left(\frac{\psi_e}{\psi}\right)^{n} \tag{30}$$

where K_s = saturated hydraulic conductivity (cm·h^{-1}),
m_c = clay mass fraction,
m_s = silt mass fraction,
$B = -2\cdot\psi_{es} + 0.2\cdot\sigma_g$
ψ_{es} = air entry potential (J·kg^{-1}) for a standard bulk density of 1,300 kg·m^{-3},
$= -0.5\cdot d_g^{-0.5}$,
d_g = geometric mean particle diameter,
$= e^{\sum m_i\cdot\ln d_i}$,
σ_g = geometric standard deviation,
$= e^{\sqrt{\sum m_i\cdot(\ln d_i)^2 - \left(\sum m_i\cdot\ln d_i\right)^2}}$,
m_i = mass fraction of textural class i,
d_i = arithmetic mean diameter of class i,
$m = 2\cdot B + 3$,
$n = 2 + \dfrac{3}{B}$,
ψ_e = air entry potential (J·kg^{-1}),
$= \psi_{es}\cdot\left(\frac{\rho_b}{1.3}\right)^{0.67\cdot B}$.

5.4.5 References

Arya, L.M. and J.F. Paris (1981). A physicoemperical model to predict the soil moisture characteristic from particle-size distribution and bulk density data. Soil. Sci. Soc. Am. J., **45**, 1023-1030.

Bloemen, G.W. (1980). Calculation of capillary conductivity and capillary rise from grain size distribution. I. Real and theoretical values of the exponent in a formula of Brooks and Corey for the calculation of hydraulic conductivities. ICW Wageningen nota no. 952. II. Assessment of the values of the exponent in a formula of Brooks and Corey for the calculation of hydraulic conductivity from grain size distribution. ICW Wageningen nota no. 962. III. Air entry pressure and saturated conductivity calculated from grain size distribution and median grain size. ICW Wageningen nota no. 990. IV. Capillary rise in soil types and soil profiles. ICW Wageningen nota no. 1013.

Brooks, R.H. and A.T. Corey (1964). Hydraulic properties of porous media. Hydrology Paper 3, Colorado State University, Fort Collins, Colorado, USA.

Burdine, N.T. (1952). Relative permeability calculations for pore-size distribution data. Trans. AIME, **198**, 34-42.

Campbell, G.S. (1974). A simple method for determining unsaturated conductivity from moisture retention data. Soil Sci., **117**, 311-314.

Campbell, G.S. (1985). Soil physics with BASIC. Transport models for soil-plant systems. Developments in Soil Science 14. Dept. of Agronomy and Soils, Washington State University, Pullman, USA.

Childs, E.C. and N. Collis-George (1950). The permeability of porous materials. Proc. R. Soc. London Ser. A, **201**, 392-405.

Clapp, R.B. and G.M. Hornberger (1978). Empirical equations for some hydraulic properties. Water Resour. Res., **15**, 601-604.

Cosby, B.J., G.M. Hornberger, R.B. Clapp and T.R. Ginn (1984). A statistical exploration of soil moisture characteristics to the physical properties of soils. Water Resour. Res., **20**, 682-690.

De Jong, R. (1982). Assessment of empirical parameters that describe soil water characteristics. Can. Agric. Engin., **24**, 65-70.

De Jong, R., C.A. Campbell and W. Nicholaichuk (1983). Water retention equations and their relationship to soil organic matter and particle size distributions for disturbed samples. Can. J. Soil Sci., **63**, 291-302.

Gupta, S.C. and W.E. Larson (1979). Estimating soil water retention characteristics from particle size distribution, organic matter percent and bulk density. Water Resour. Res., **15**, 1633-1635.

Jackson, R.A. (1972). On the calculation of hydraulic conductivity. Soil Sci. Soc. Am. Proc., **36**, 380-383.

Kern, J.S. (1995). Evaluation of soil water retention models based on soil physical properties. Soil Sci. Soc. Am. J., **59**, 1134-1141.

Marquardt, D.W. (1963). An algorithm for least-squares estimation of non-linear parameters. J. Soc. Ind. Appl. Math., **11**, 431-441.

Marshall, T.J. (1958). A Relation between permeability and size distribution of pores. J. Soil Sci., **9**, 1-8.

Millington, R.J. and J.P. Quirk (1959). Permeability of porous media. Nature, **183**, 387-388.

Mualem, Y. (1976). A new model for predicting the hydraulic conductivity of unsaturated porous media. Water Resour. Res., **12**, 513-522.

Rawls, W.J., D.L. Brakensiek and K.E. Saxton (1982). Estimation of soil water properties. Trans. ASAE, **25**, 1316-1328.

Rawls, W.J., D.L. Brakensiek and B.Soni (1983). Agricultural management effects on soil water processes, Part I. Soil water retention and Green and Ampt infiltration parameters. Trans. ASAE, **26**, 1747-1752.

Rawls, W.J. and D.L. Brakensiek (1985). Prediction of soil water properties for hydrological modelling. In : E.B. Jones en T.J. Wards (eds.). Watershed management in the eighties. Proc. of Symp. Sponsored by Comm. on Watershed Management, I & D Division, ASCE. ASCE Convention, Denver, Colorado, USA, April 30-May 1, 293-299.

Rawls, W.J., T.J. Gish and D.L. Brakensiek (1991). Estimating soil water retention from soil physical properties and characteristics. Adv. Soil Sci., **16**, 213-234.

Rawls, W.J., L.R. Ahuja, D.L. Brakensiek and A. Shirmohammadi (1993). Infiltration and soil water movement. In : D.R. Maidment (ed.). Handbook of hydrology. McGraw-Hill, Inc., USA.

Saxton, K.E., W.J. Rawls, J.S. Romberger and R.I. Papendick (1986). Estimating generalised soil-water characteristics from texture. Soil Sci. Soc. Am. J., **50**, 1031-1036.

Tyler, S.W. and S.W. Wheatcraft (1990). Fractal processes in soil water retention. Water Resour. Res., **26**, 1047-1054.

van Genuchten, M.Th. (1978). Calculating the unsaturated hydraulic conductivity with a new closed-form analytical model. Research Report 78-WR-08, Water Research Program, Dept. of Civil Engineering, Princeton University, Princeton, New Jersey, USA.

van Genuchten, M. Th. (1980). A closed-form equation for predicting the hydraulic conductivity of unsaturated soils. Soil Sci. Soc. Am. J., **44**, 892-898.

van Genuchten, M.Th., F.J. Leij and S.R. Yates (1991). The RETC code for quantifying the hydraulic functions of unsaturated soils. EPA/600/2-91/065, R.S. Kerr Environmental Research Laboratory, Office of Research and Development, U.S. Environmental Protection Agency, ADA, Oklahoma, USA.

van Genuchten, M.Th., F.J. Leij and L.J. Lund (1992). Indirect methods for estimating the hydraulic properties of unsaturated soils. Univ. of California, Riverside, USA.

Vereecken, H., J. Maes, J. Feyen and P. Darius (1989). Estimating the soil moisture retention characteristic from texture, bulk density and carbon content. Soil Sci., **148**, 389-403.

Yates, S.R., M.Th. van Genuchten, A.W. Warrick and F.J. Leij (1992). Analysis of measured, predicted, and estimated hydraulic conductivity using the RETC computer program. Soil Sci. Soc. Am. J., **56**, 347-354.