INTERNATIONAL ATOMIC ENERGY AGENCY
UNITED NATIONS EDUCATIONAL, SCIENTIFIC AND CULTURAL ORGANIZATION
INTERNATIONAL CENTRE FOR THEORETICAL PHYSICS
I.C.T.P., P.O. BOX 586, 34100 TRIESTE, ITALY, CABLE: CENTRATOM TRIESTE

SMR.705 - 4

COLLEGE ON SOIL PHYSICS

(6 - 24 September 1993)

"I - Soil Water
II - Soil Water Potential
III - Tensiometer
IV - Soil Water Characteristic Curve"

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TRIESTE - ITALY

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LECTURE NOTES

by

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I. SOIL WATER

II. SOIL WATER POTENTIAL

III. TENSIOMETER

IV. SOIL WATER CHARACTERISTIC CURVE

I. SOIL WATER

Soil water is important in itself as a feature of the physical environment but is especially prominent in its relationships with climatology and with the surface and subsurface hydrologic regimes as a component in the terrestriel 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 and because of the complexity of the soil body surpasses that of the vegetated layer and atmosphere above.

Numerous techniques have been devised for measuring soil moisture content, and nine of these are listed in table 1. Of these, the first three have gained the greatest acceptance. Only the most frequently applied methods will be discussed.

Table 1. Methods of measuring soil water content.

1. Gravimetric
2. Neuton scattering
3. Gamma-ray attenuation
4. Gamma-ray backscattering
5. Electrical resistance
6. Thermal conductivity
7. Chemical (CaC_2)
8. Time-domain reflectometry

1. Soil water content.

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)

$$w = \frac{M_w}{M_s} \qquad (1)$$

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)

$$\theta = \frac{V_w}{V_t} = \frac{V_w}{V_s + V_w + V_a} \qquad (2)$$

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) (Fig 1).

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

$$\frac{\theta}{w} = \frac{V_w\, M_s}{V_t\, M_w} \qquad (3)$$

knowing: $\rho_b = \frac{M_s}{V_t}$ (mass of dry soil per unit bulk volume and usually lies in the range 1.1 and 1.7 g.cm^{-3})

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

Equation (3) becomes:

$$\theta = \frac{w\rho_b}{\rho w} \qquad (4)$$

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 depth

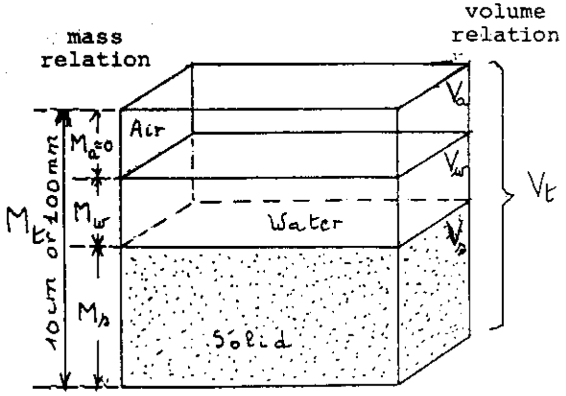

Figure 1. Schematic diagram of the soil as a three-phase system.

2. How to measure the soil water content?

There are numerous procedures and types of equipment 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.

2.1. Gravimetric Method

The gravimetric method of measuring soil water content consists of removing a sample by augering into the soil and then determining it moist and dry weights. The moist weight is determine 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:

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

The method depending as 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 vaporation of water.

The errors of the gravimetric method can be reduced by increasing the sizes and number of samples. However, the sampling method is destructive and may distrub 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.

2.2. Radiation Techniques

Will be discussed by Prof.Dr. K. Reichardt

2.3. Time-Domain Reflectometry(1)

The use of time-domain reflectometry (TDR) is coming into its own as an important new tool for soil water measurements. Early use of TDR stemmed from attempts to deduce molecular structures from knowledge of the frequency dependence of the dielectric constant of organic molecules. Discrete measurements made over a wide range of frequencies were time consuming and tedious. Fellner-Feldegg (1969) used TDR to obtain the complete dielectric frequency spectrum in a single experiment by recording the response to the launching of an electromagnetic pulse along a finite coaxial transmission line containing the organic solvent. Time-domain reflectometry is now being used as a method for measuring the dielectric constant and electrical conductivity of soil by determining the travel time and dissipation, respectively, of an electromagnetic pulse launched along parallel metallic probes

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(1) *Text from "Development of Time-Domain Reflectometry for Measuring Soil Water Content and Bulk Soil Electrical Conductivity" by F. N. Dalton in "Advances in Measurement of Soil Physical Properties: Bringing Theory into Practices" SSSA Special Publication*

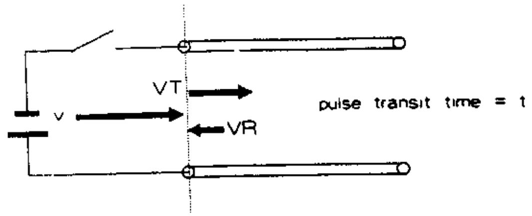

Fig. 2 Schematic illustration of time-domain reflectometry principles showing the launching of a voltage pulse V with the transmitted and reflected components V_T and V_R.

embedded in the soil. Theoretical analysis and experimental correlations show that the pulse travel time is proportional to the apparent dielectric constant of the soil and that signal dissipation is proportional to the bulk soil electrical conductivity. These two fundamental physical parameters are uniquely correlated with soil water content and pore water salinity.

A simplified symbolic representation of a TDR system used to measure these parameters is shown in Fig. 2 A voltage V is applied to the ends of a buried transmission line. A fraction of the applied voltage, V_T, propagates along the path defined by the parallel rods. Upon reflection of the pulse at the end of the rods in the soil, a fraction of the voltage, V_R, returns to its origins. Topp et al. (1980), Topp et al. (1982), and Topp and Davis (1985a) used TDR to measure this pulse transit time and showed that it is proportional to water content for a wide variety of soil types. Their work in this area was preeminent in promoting the use of TDR technology for water content measurements for agricultural purposes (Topp & Davis, 1985b). This technique is also finding unique applications for studying soil water in freezing environments, e.g. Patterson and Smith (1981), Baker et al. (1982), Hayhoe et al. (1983), Hayhoe and Topp (1983), Stein and Kane (1983), and Ryden (1986).

Early attempts by Dalton et al. (1984), Dasberg and Dalton (1985), and Dalton (1987) to use TDR techniques for water content measurements in highly conductive soils led to the development of methods for using other TDR waveform parameters to infer bulk soil electrical conductivity simultaneously with soil water content. They showed that the TDR waveform not only contained information about the pulse transit time, but also information concerning the dissipation of the electromagnetic energy that could be used to deduce the medium electrical conductivity and therefore soil water salinity. Topp et al. (1988) measured bulk soil electrical conductivity in coaxial sample chambers using an analysis that analytically accounted for a pulse reflection that was not explicitly considered in the previously described version. Recently Dalton et al. (1990) used an analysis derived from loaded transmission line theory to measure bulk soil electrical conductivity based on measurement of long time reflections. This will be discussed in some detail later in the paper.

Bulk soil electrical conductivity is a soil-type dependent function of both soil water content and pore water electrical conductivity (Rhoades et al., 1976). The unique ability of TDR methods to simultaneously measure soil water content and electrical conductivity over identical sampling volumes, using a single probe, makes TDR an important new research tool for agricultural and hydrologic studies.

WATER CONTENT MEASUREMENTS

Complex physical measurements are often reduced to correlations between an instrument transducer output and some independent measure of the physical property in question. With TDR, the soil-water content measurement is ultimately a correlation between an instrument response (i.e., pulse transit time over the path length of the TDR probe) and an independent measure of soil water content (i.e., gravimetric measurements). The fundamental physical property that affects the pulse transit time is the dielectric property of the medium. This is seen by equating pulse velocities derived from electrodynamics and mechanics. The electrodynamic expression for pulse velocity, v, is given in terms of the velocity of light in vacuum, c, and the relative dielectric constant of the medium, ϵ. The pulse velocity derived from mechanics is given in terms of a transit time, t, and path length, $2l$,

$$\text{electrodynamic pulse velocity: } v = c/\sqrt{\epsilon} \qquad [1]$$

$$\text{mechanical pulse velocity: } v = 2l/t. \qquad [2]$$

Equating Eq. [1] and Eq. [2] and solving for the dielectric constant gives,

$$\epsilon = [ct/2l]^2. \qquad [3]$$

Topp et al. (1980) measured this pulse transit time, t, in coaxial chambers of known length and correlated the calculated value of the dielectric constant, ϵ, of soil with their volumetric water content, θ. This work represented the first data showing the efficacy of this new technology for soil water measurements. A summary of this important work is given below in terms of their empirically derived regression equation, (Fig. 3)

$$\theta = -5.3 \times 10^{-2} + 2.92 \times 10^{-2}\,\epsilon - 5.5 \times 10^{-4}\,\epsilon^2 + 4.3 \times 10^{-6}\,\epsilon^3 \qquad [4]$$

This relation has been shown to be good, to within a few percent, for a wide class of soils. Using other calibration techniques, Roth et al. (1990) showed the uncertainty in the water content not to exceed 0.013 $cm^3\ cm^{-3}$. Drungil et al. (1989) even found this empirical relation to be valid for gravelly soils. Most likely this equation will require adjustment for peat and heavy clay

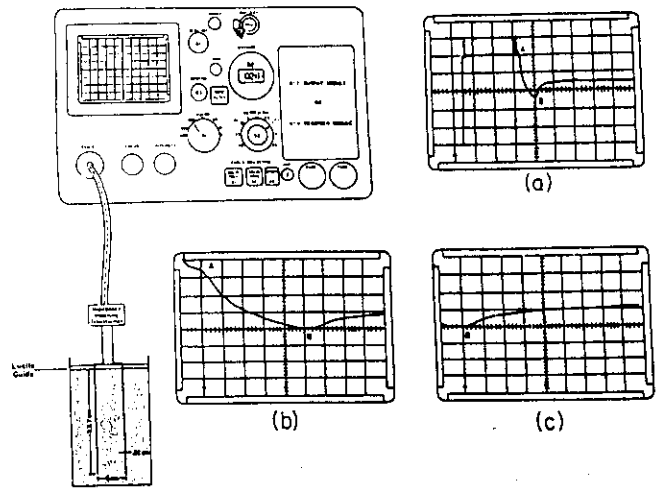

Fig. 4 Cable tester connected to parallel wire wave guide showing full and expanded wave-forms on cathode-ray tube used to measure moisture content. Point *A* results from pulse entering soil. Point *B* results from reflected voltage returning to source.

soils or in any case where greater accuracy is required. According to Eq. [4] and Eq. [3], the water content determination is reduced to the measurement of the transit time of a voltage pulse over a known path length. A typical pulse transit time for a path length of 30 cm in a wet soil is about 2 ns. This extremely small time interval can be accurately measured with commercially available equipment. Figure 4 shows a Tektronix 1502 cable tester[1] (Tektronix, Beaverton, OR) interfaced with parallel metallic probes embedded in a moist soil, and, the expanded wave forms and corresponding settings for the function switches to measure soil moisture. The TDR wave form contains characteristic pick off points that are used to measure the soil dielectric constant. The distance between points *A* and *B* (Fig. 4b) represent the pulse transit time over a distance twice the probe length and is used to calculate an effective dielectric constant for the soil using Eq. [3], and finally the water content by using Eq. [4]. Aggregate data obtained using these TDR techniques for calculating soil water contents at various levels of pore water salinity are shown in Fig. 5 The effects of moderate to low salinities can often be neglected for TDR determined water contents. However, under certain conditions, the effects of salinity cannot be neglected and departures from the relation described by Eq. [4] begin to occur. This condition is discussed in the next section.

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[1] Names of products are included for benefit of the reader and do not imply endorsement or preferential treatment by USDA.

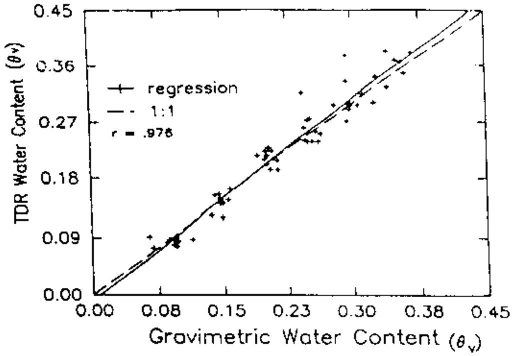


Fig. 5 Time-domain reflectometry water content versus gravimetric water content with soil water salinity ranging up to 10 dS m^{-1}.

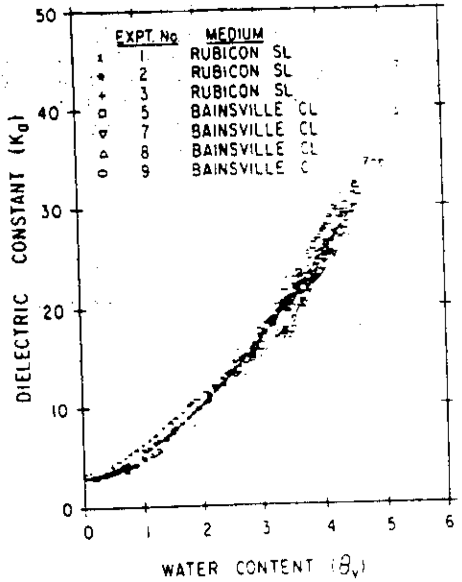


Figure 3 The relationship between dielectric constant and water content measured by Topp et al. (1980).

Figure 6 shows a block diagram of the TDR system and the idealized TDR output. The TDR source generates a fast rise time voltage step function as shown in Fig. 6 at A. The step propagates down the instrument transmission line through the receiver, B, to the transmission line filled with soil, C to D. All the signal is reflected at D, the end of the transmission line. The receiver used an electronic sampling technique to put out on the display an audio frequency facsimile of the radio frequency signal. The displayed signal was either photographed or recorded on an x-y plotter. The travel time of the step in the soil was obtained by measuring the distance C to D along the time axis in plots similar to that in Fig.6 . Eq. 3 was used to calculate the apparent dielectric constant.

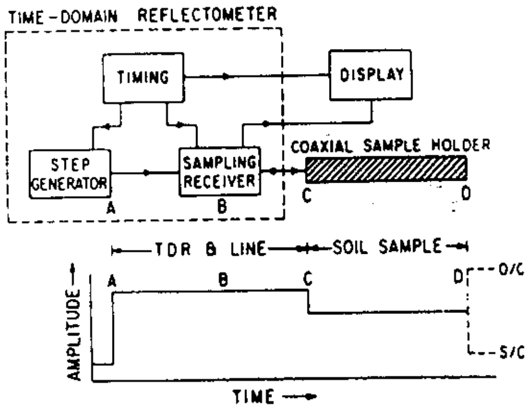


Figure: 6 Block diagram of the TDR system and its connection to a soil sample. The lower portion shows an idealized TDR output trace.

II. SOIL WATER POTENTIAL

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. 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, hygroscopic 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 tranport 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 1).

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:

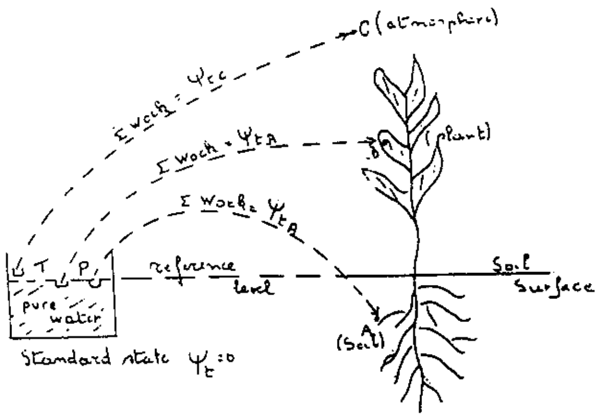


Figure 1. Potential of soil water, water in plant cell and water in the atmosphere.

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

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

The dots signify that additional terms are theoretically possible.

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.

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.

Joule kg^{-1}

This method of expression is not widely used.

b. Energy per unit volume of water (pressure).

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

Joule m^{-3} or N m^{-2} (Pa)

c. Energy per unit weight of water (head).

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

Joule N^{-1} = Nm N^{-1} = m

For conversion from one unit to another knows that:

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

4. Gravitational potential.

Every body on the earth's surface is attracted towards the center 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.

Conversion table for units of soil water potential (*)

Specific potential units	Volumetric potential units				Weight potential units
Joule/kg	bar	millibar	Pa	atmosphere	m
0.0001	0.000001	0.001	0.1	0.000000987	0.001017×10^{-2}
1	0.01	10	10^3	0.00987	10.17×10^{-2}
100	1	1000	10^5	0.987	10.17
0.1	0.001	1	10^2	0.000987	1.017×10^{-2}
101.3	1.013	1013	1.013×10^5	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.

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.

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:

$$- M g z = - \rho_w V gz$$

where ρ_w is the density of water and g the acceleration of gravity.

Gravitational potential can be expressed:

- per unit mass: $\psi_g = - gz \quad (J\ kg^{-1})$
- per unit volume: $\psi_{gv} = \psi_g \cdot \rho_w = - \rho_w g z$ (Pa)
- per unit weight: $\psi_{gw} = \psi_g \, 1/g = - z$ (m)

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 2). 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).

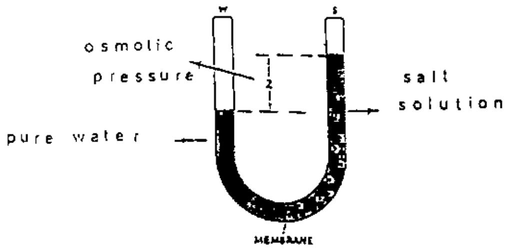


Figure 2. Schematic presentation of osmosis.

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_0 was observed. Otherwise, the concentration of salts will become the same throughout the profiles by the process of diffusion and the difference in ψ_0 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.

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

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

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

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

As we assume the soil pores to have a cylindrical shape (figure 3) 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)

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

where: h = height of capillary rise
ρ_w = density of water
g = acceleration of gravity
r = radius of the capillary tube

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 capillary 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 4.

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.

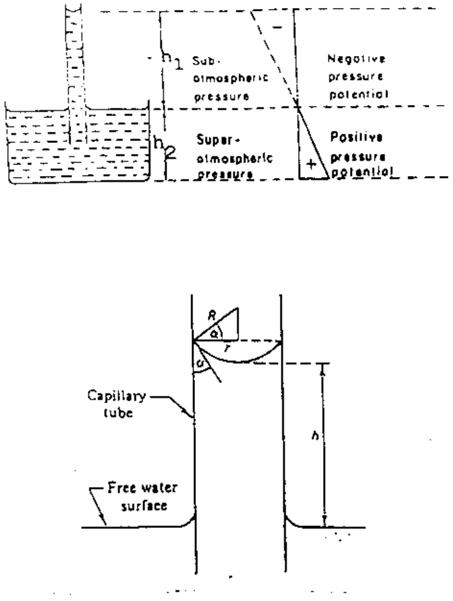


Figure 3 : Capillary rise of water into a capillary tube.

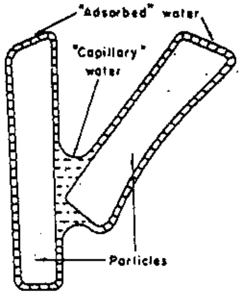


Figure 4 : Water in an unsaturated soil is subject to capillarity and adsorption, which combine to produce a matric potential

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.

The matric potential can be expressed:

- per unit mass:

$$\psi_m = g\ h = -\frac{2\gamma}{\rho_w r} \qquad (\mathrm{J\ kg^{-1}})$$

- per unit volume:

$$\psi_m\ \rho_w = \rho\ g\ h = \frac{-2\gamma}{r} \qquad (\mathrm{Pa})$$

- per unit weight:

$$\psi_m \frac{1}{g} = h = \frac{-2\gamma}{\rho_w\ g\ r} \qquad (\mathrm{m})$$

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 3). 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 continuous potential, extending from the saturated zone into the unsaturated zone, below and above the water table.

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 (See determination of the soil water characteristis curve by the pressure plate extraction apparatus).

FINAL REMARK.

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.

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 is constant. Usually, sufficient condition is that the sum of the component potentials, ψ_o being ignored, is constant. The equilibrium condition states then that:

$$\psi_g + \psi_p = \text{constant} = \psi_H \qquad (1)$$

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 3) 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 (1) becomes:

$$H = h + z$$

where: 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)
H = the hydraulic 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 5 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.

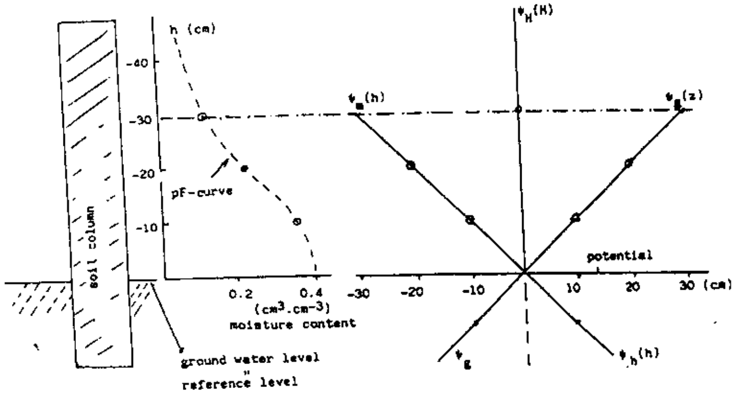


Figure 5. Equilibrium condition in a soil column.

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

III. TENSIOMETER

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 6).

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 piezometers 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 piezometers can be determined by measuring the depth of the water level, since at the flat air-water interface the pressure potential is zero.

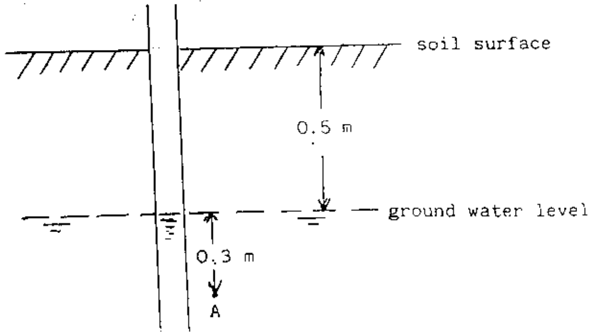


Figure 6. A piezometer in a soil profile.

Figure 6 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.

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 7).

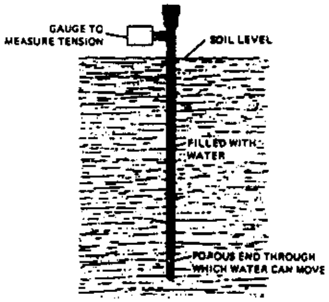


Figure 7. Tensiometers with a vacuum gauge.

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. Principle of the tensiometer.

When the cup is placed in a water reservoir (figure 8), 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.

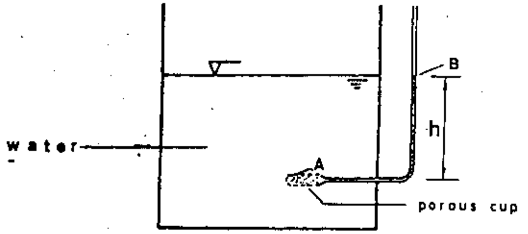


Figure 8. Porous cup connected with a piezometer tube for measuring pressure potentials under the water table.

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 a 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 exercices 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 9).

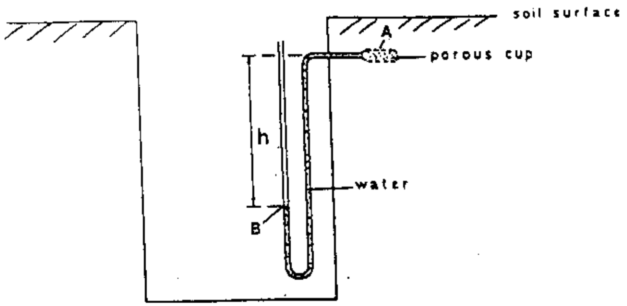


Figure 9. Tensiometer for measuring pressure potentials in soils.

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 9 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 10).

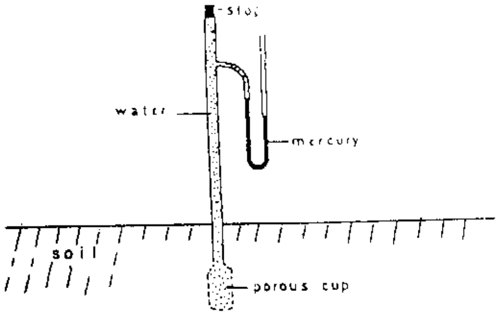


Figure 10. Tensiometer with mercury manometer.

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.

4. How to calculate the soil water pressure head h and the hydralic head H.

Let x be the height of the mercury in the manometer (in cm) and z the vertical axis. 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).

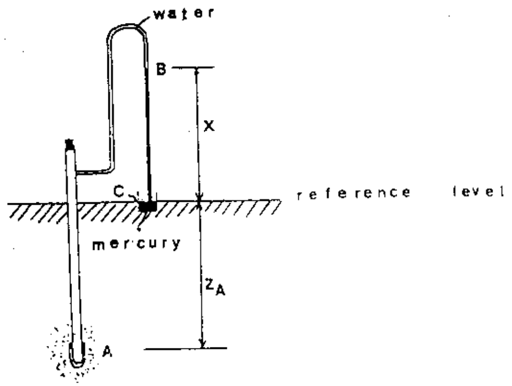


Figure 11. Tensiometer installation with the mercury level in the reservoir at the soil surface being the reference level.

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

$$z + P/\rho g = \text{constant}$$

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

From figure 11 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 ($- z_A$).

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

$$P_A = P_B + z_B \rho_w g + z_A \rho_w g = P_B + \rho_w g(z_A + z_B) \qquad (1)$$

- in mercury:

$$z_B + \frac{P_B}{\rho_{Hg} g} = z_C + \frac{P_C}{\rho_{Hg} g}$$

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

the equation becomes:

$$Z_B + \frac{P_B}{\rho_{Hg} g} = 0$$

or:

$$P_B = - \rho_{Hg} g\, z_B \qquad (2)$$

(2) in (1) gives:

$$\frac{P_A}{\rho_w g} = \frac{- \rho_{Hg}\, g\, z_B}{\rho_w g} + z_A + z_B \qquad (3)$$

Since: $\rho_{Hg} = 13600 \text{ kgm}^{-3}$
$\rho_w = 1000 \text{ kg m}^{-3}$

(3) becomes:

$$h_A = - 13.6\, z_B + z_A + z_B$$

or:

$$h_A = - 12.6\, z_B + z_A$$

$$h_A = - 12.6\, x + z_A \qquad (4)$$

Normally the free surface of the mercury in the reservoir (point C) is located y cm above the soil surface (reference level) (figure 12).

Equation (4) becomes:

$$h_A = - 12.6\, x + y + z_A \qquad (5)$$

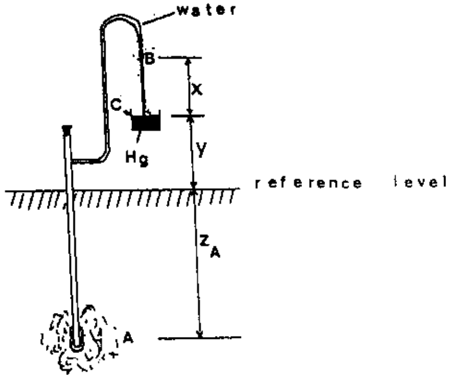


Figure 12. Tensiometer installation with the mercury reservoir y cm above soil surface.

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

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

$$H = -12.6\,x + y \qquad (6)$$

Example

Given: The porous cup of a tensiometer with mercury manometer is installed 0.3 m under the soil surface. The height of the mercury in the manometer is 0.31 m. The level of the mercury reservoir is located 0.4 m above the soil surface. Reference level is the soil surface.

Calculate:
- soil water pressure head h
- hydraulic head H

Solution:

$$h = (-12.6 \times 0.31 + 0.4 + 0.3)\ \text{m} = -3.20\ \text{m}$$

$$H = (-12.6 \times 0.31 + 0.4)\ \text{m} = -3.50\ \text{m}$$

5. Some characteristics of the tensiometer.

- cup conductance:

$$K' = \frac{\Delta V}{\Delta t\ \Delta h}$$

being the volume of water passing through the cupwall under a unit pressure difference per unit time ($m^3\ s^{-1}\ Pa^{-1}$)

- sensitivity of the manometer:

$$S = \frac{\Delta h}{\Delta V}$$

being the change in manometer reading per unit volume change of water ($Pa\ m^{-3}$)

- response time of the system:

$$T_R = \frac{1}{K'S}$$

is a measure of the over-all responsiveness of the instrument to a change of soil-water potential at the cup surface (s)

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. Occurence 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 water. Use of tensiometers in the field is therefore only possible when pressures do not fall below this value. 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 13.

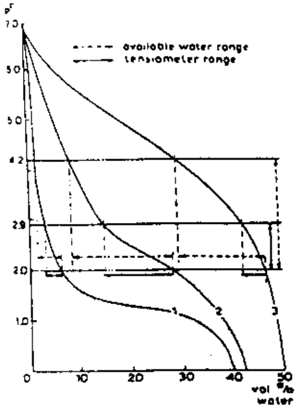


Figure 13. Part of the available moisture range covered by tensiometers, depending on soil type

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

Thus where soil management (particularly in irrigation) is aimed at maintaining high pressure potential conditions which are mostly favorable 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.).

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 occurence 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.

7. Applications of measurements.

By the use of several tensiometers at different depths the direction and magnitude of water movement can easily be as certained.

EXAMPLES

1. Determination of the direction of water flow at different levels in the soil profile (Figure 14).

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 the provision that the conductivity is a function of soil water content θ .

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

$$q = - K(\theta) \frac{dH}{dz} \qquad (7)$$

where: q = flux

$K(\theta)$ = hydraulic conductivity

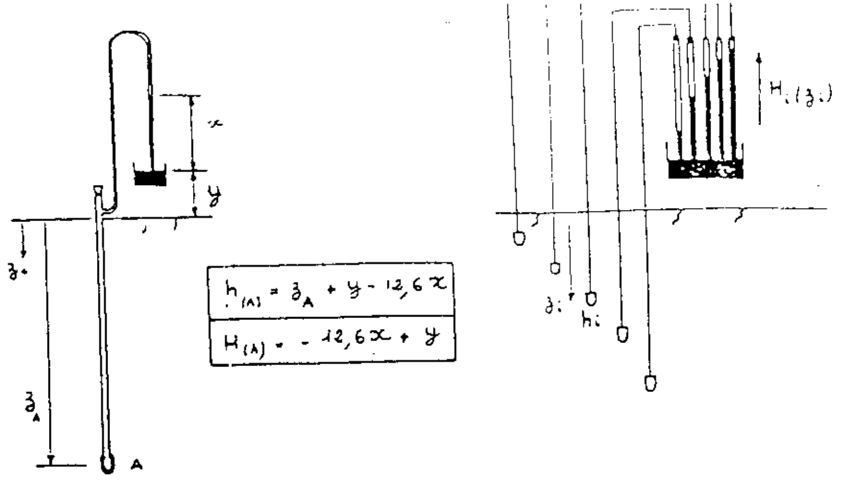
H = hydraulic head

= $h + z$ 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 n° 2 is lower than in manometer n° 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 14.



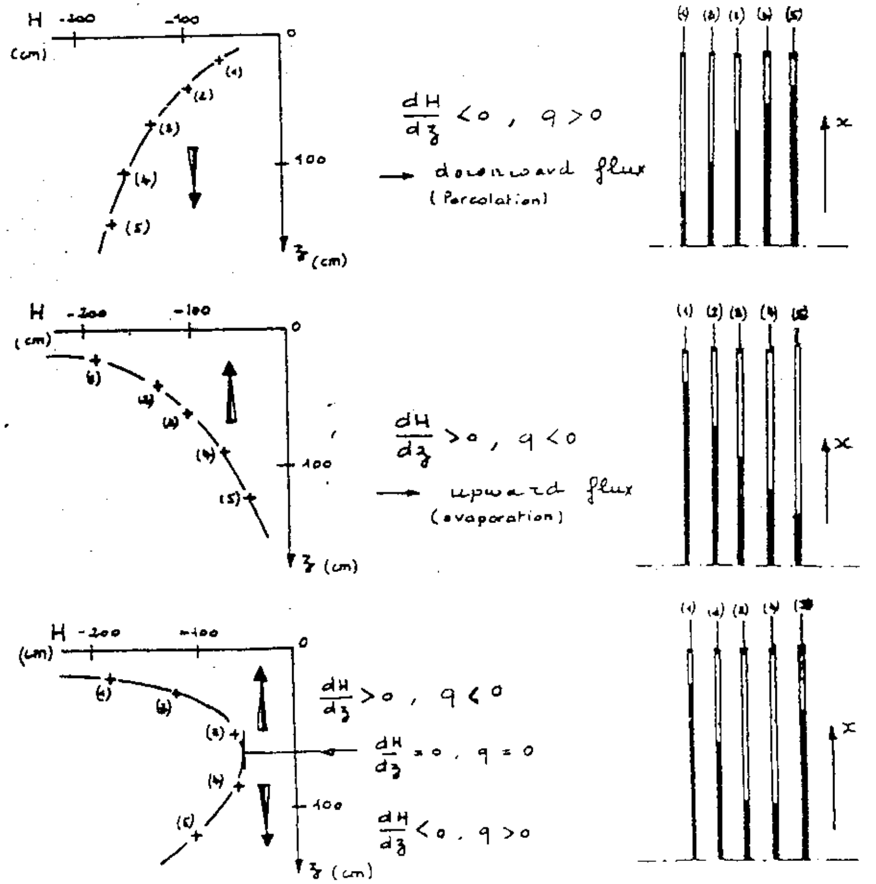


Figure 14. Hydraulic head profiles. The manometers from left to right increase with depth.

- Numerical example.

Reference level being the soil surface.

tensiometer z_1 = 50 cm x = 10.7 cm y = 20 cm
tensiometer z_2 = 75 cm x = 13.8 cm y = 20 cm

$H_1 = -\ 12.6\ x + y = -\ 1.05$ m
$H_2 = -\ 12.6\ x + y = -\ 1.54$ m

In this situation downward flow occurs since flow is towards the position with the lowest water potential.

2. Flux control at a certain depth.

From agricultural point of view it could be of interest to know if 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(θ) relation of that soil and the hydraulic head gradient dH/dz, one can calculate the instantaneous water flow q (see equation 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 5)
- knowing the soil water content (θ) using the neutron moisture meter.

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 atmopsheric 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 (table 1). 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. 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 ASHCRIFT, 1972).

Crop	*Matric potential (joules/kg)*	*Equivalent matric suction (centibars)*	*Reference*
Vegetative crops			
Alfalfa	−150	150	S. A. Taylor[a]
Beans (snap and lima)	−75 to −200	75 to 200	Vittam et al. (1963)
Cabbage	−60 to −70	60 to 70	Vittam et al. (1963), and Pew (1958)[b]
Canning peas	−30 to −50	30 to 50	S. A. Taylor[a]
Celery	−20 to −30	20 to 30	A. W. Marsh,[a] and Marsh (1961)
Grass	−30 to −100	30 to 100	Vissar (1959)
Lettuce	−40 to −60	40 to 60	A. W. Marsh,[a] Vissar (1959), and Pew (1958)[b]
Tobacco	−30 to −80	30 to 80	Jones et al. (1960)[c]
Sugar cane			
Tensiometer	−15 to −50	15 to 50	Waterhouse et al. (1954)[d]
Blocks	−100 to −200	100 to 200	Robinson (1963)
Sweet corn	−50 to −100	50 to 100	S. A. Taylor,[a] and Vittam et al. (1963)
Turfgrass	−24 to −36	24 to 36	Morgan (1964)
Root crops			
Onions			
Early growth	−45 to −55	45 to 55	Pew (1958)[b]
Bulbing time	−55 to −65	55 to 65	Pew (1958)[b]
Sugar beets	−40 to −60	40 to 60	S. A. Taylor[a]
Potatoes	−30 to −50	30 to 50	S. A. Taylor,[a] Vittam et al. (1963), and Pew (1958)[b]
Carrots	−55 to −65	55 to 65	Pew (1958)[b]
Broccoli			
Early	−45 to −55	45 to 55	Pew (1958)[b]
After budding	−60 to −70	60 to 70	Pew (1958)[b]
Cauliflower	−60 to −70	60 to 70	Pew (1958)[b]
Fruit crops			
Lemons	−40	40	A. W. Marsh[a]
Oranges	−20 to −100	20 to 100	Stolzy et al. (1963)
Deciduous fruit	−50 to −80	50 to 80	A. W. Marsh,[a] and Vissar (1959)
Avocadoes	−50	50	Richards et al. (1962)
Grapes			
Early season	−40 to −50	40 to 50	A. W Marsh[a]
During maturity	< −100	> 100	A. W. Marsh[a]
Strawberries	−20 to −30	20 to 30	A. W. Marsh,[a] and Marsh (1961)
Cantaloupe	−35 to −40	35 to 40	Marsh (1961), and Pew (1958)[b]
Tomatoes	−80 to −150	80 to 150	Vittam et al. (1958),[e] and Vittam et al. (1963)
Bananas	−30 to −150	30 to 150	Schmeuli (1953)[f]
Grain crops			
Corn			
Vegetative period	−50	50	S. A. Taylor[a]
During ripening	−800 to −1200	800 to 1200	S. A. Taylor[a]
Small grains			
Vegetative period	−40 to −50	40 to 50	S. A. Taylor[a]
During ripening	−800 to −1200	800 to 1200	S. A. Taylor[a]
Seed crops			
Alfalfa			
Prior to bloom	−200	200	Taylor et al. (1959)
During bloom	−400 to −800	400 to 800	Taylor et al. (1959)
During ripening	−800 to −1500	800 to 1500	Taylor et al. (1959)
Carrots			
During seed year at 60 cm depth	−400 to −600	400 to 600	Hawthorn (1951)[g]
Onions			
During seed year at 7 cm depth	−400 to −600	400 to 600	Hawthorn (1951)[g]
at 15 cm depth	−150	150	Hawthorn (1951)[g]
Lettuce			
During productive phase	−300	300	Hawthorn et al. (1956)[g]
Coffee	Requires short periods of low potential to break bud dormancy, followed by high water potential		Alvin (1960)

IV. 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)

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.

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 (Fig. 2.11).

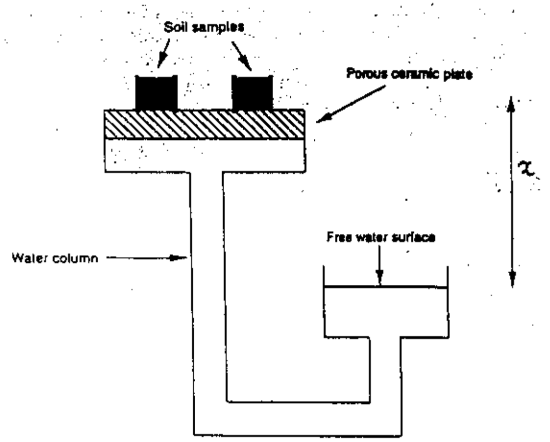


Figure 2.11 Desaturation of soil water samples to a desired energy state with a hanging water column.

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) $\psi_m + \psi_z = 0 = \psi_m + \rho_w g Z$, or $\psi_m = -\rho_w g Z$

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.

Pressure Plate (Range $-1500\overset{0}{}$ cm $\le h \le -300$ 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 (Fig. 2.12).

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) $\psi_{ep} + \psi_m = 0 = \psi_m + \Delta P$, 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.

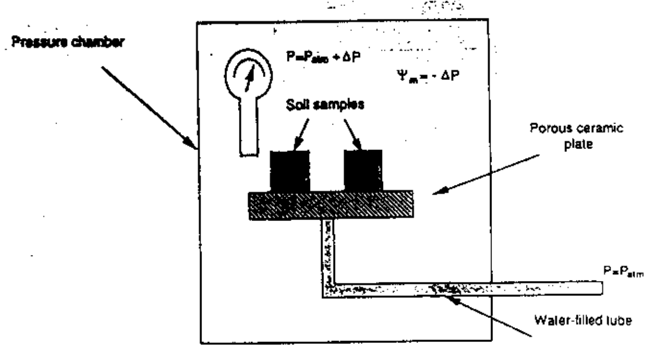


Figure 2.12 Desaturation of soil water samples to a desired energy state with a pressure plate.

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.

Equilibration over Salt Solutions ($h < -15{,}000$ 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 $\psi_T = \psi_o$ 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 2.13 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 Fig. 2.13 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.

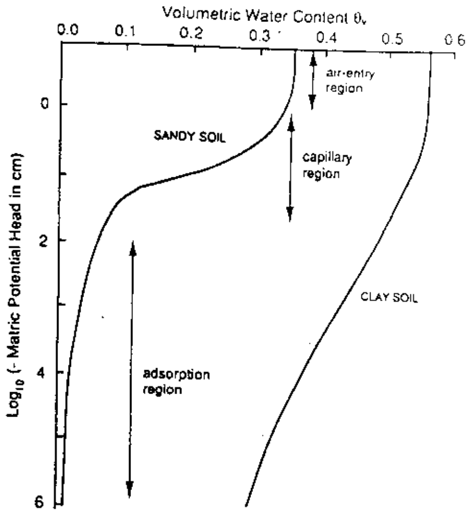


Figure 2.13 Matric potential–water content function (water characteristic function).

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 (2.5), 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 Fig. 2.14. 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, inasmuch as vapor transfer would assure the wetting of all pores small enough to retain water at a particular matric potential.

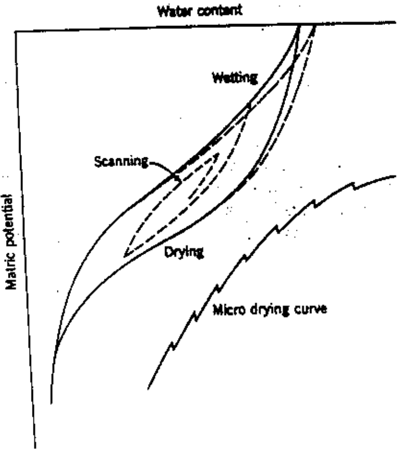


Figure 2.14 Wetting and drying curves and scanning curves.

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 Fig. 2.14.

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 Fig. 2.14. This principle is illustrated by the pore–water system in Fig. 2.15*a*. 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 (2.5), 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 Fig. 2.15*b*. 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 Fig. 2.15*b*, 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 (Fig. 2.15*c*). 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

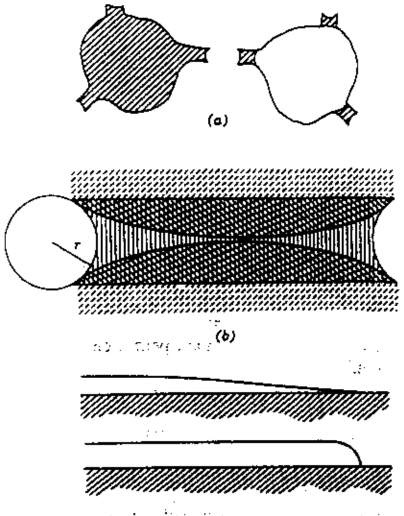


Figure 2.15 Diagrammatic representation of three forms of water content–matric potential hysteresis.

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 Fig. 2.14. 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, Fig. 2.14). 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 ([illegible]). The wetting curve of the hysteretic envelope is commonly known as a sorption curve and the drying curve a desorption curve. Since air is almost always present when such curves are produced experimentally, the solid curves shown in Fig. 2.12 are obtained. However, some ambiguity exists, because the starting point for many measurements is a wet and sometimes even puddled sample in which the degree of air removal is unknown.

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.