**FIGURE 10.8** A neutron moisture meter with sealer/rate meter device. Some models have a rate meter built within one assembly (Ibadan, Nigeria, 1972).

The common source of fast neutrons used in probe is either 2–5 millicurie mixture of radium-beryllium, which in addition to neutrons also emits γ-rays. These sources have an extremely long half-life of 1620 years. The slow neutrons are monitored by a detector filled with BF_3 gas, which cause the following reaction:

B+neutron=α (particle with helium nucleus)

$$(10.5)$$
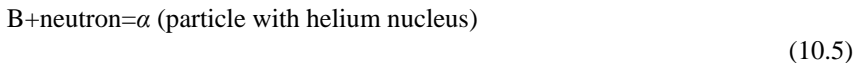
The emission of α particle creates an electrical pulse on a charged wire. The number of pulses generated over a measured time interval is counted by a scalar or indicated by a rate meter.

The technique has numerous merits. It is nondestructive, facilitates monitoring soil moisture content for the same site overtime, covers a large soil volume, and monitors volume of soil moisture (Fig. 10.9). However, there are numerous limitations of the technique. It is expensive, poses health hazards, requires specialized maintenance and repair, and there are specific problems with calibration (Lal, 1974; 1979b). The equipment calibration is influenced by texture, gravel content, stoniness, clay mineralogy, and soil's chemical constituents (Fig. 10.10). Some elements present in the soil can capture neutrons. These include gadolinium, cadmium, boron, chlorine,

FIGURE 10.9 A plastic covered plot is used to assess field water capacity using a neutron moisture meter. After saturing the plot with sufficient water, the plastic cover was used to prevent evaporation. (Ibadan, Nigeria, 1971)
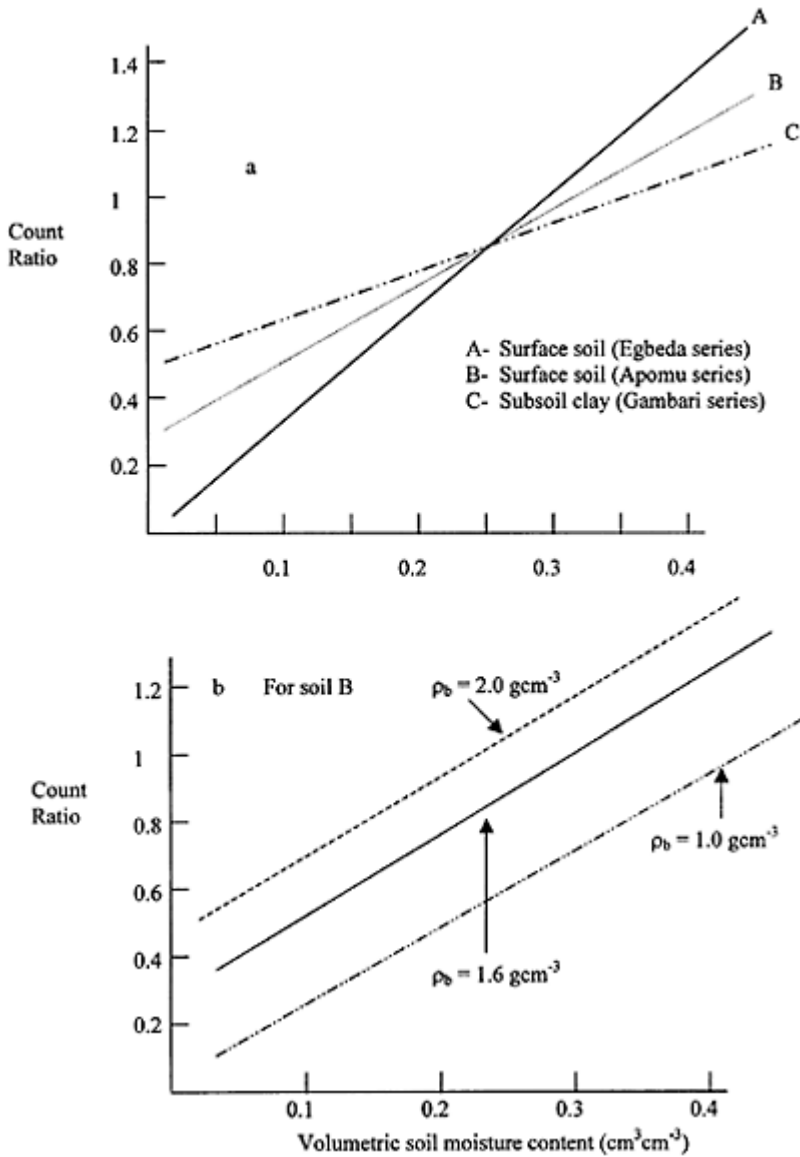
FIGURE 10.10 The effect of (a) soil texture and (b) bulk density on neutron probe calibration. (Redrawn from Lal, 1974.)

manganese, and iron. The measurements are also not very accurate for surface horizons, and in soils with high organic matter content (e.g., Mollisols, organic soils). There are,

however, surface neutron meters available to measure soil's moisture content for the plow layer. Lunar Prospector using the neutron spectroscope, reported the existence of water on the moon (Kerr, 1997). Feldman et al. (1998) used neutron spectroscopy to measure fluxes of fast and epithermal neutrons from Lunar Prospector and concluded that lunar poles contain water and ice. Nozette et al. (1996) used data from the Clementine bistatic radar experiment and arrived at the same conclusion. Nonetheless, existence of water on the moon remains to be a controversial issue (Eshelman and Parks, 1999).

Gamma Ray Attenuation. The degree to which the intensity of monoenergetic γ-ray is reduced when passed through soil is related to wet soil density. If the bulk density remains constant, then the intensity of γ-ray passing through the soil is related to its moisture content as per Eq. (10.6).

$$I = I_o e^{-\mu_w} \rho x$$

(10.6)

where I is the transmitted intensity, I_o is the incident intensity, μ_w is the mass absorption coefficient of water, ρ is density of the absorber, and x is thickness of the soil. Intensity of γ-radiation is usually measured in terms of the count rate registered by a scalar or a rate meter, and Eq. (10.6) can be rewritten as follows:

$$N = N_o e^{-\mu} \rho x$$

(10.7)

or

$$\rho_n N/N_o = -\mu \rho x$$

(10.8)

where N and N_o are counts corresponding to intensity I and I_o.

There are two types of γ-ray equipment. The single γ-ray attenuation method involves a single source (Gurr, 1962; Reginato and Van Bavel, 1964). The second type of equipment involves two sources so that simultaneous measurements can be made for bulk density and moisture content. There are two techniques available for dual γ-scanning. One involves independent measurements of γ-ray attenuation usually using [241]Am at 0.060 MeV and [137]C$_s$ at 0.662 MeV. It is important to know the mass absorption coefficients of soil *(μs)* and water *(μw)*. This technique is generally used under laboratory conditions. The second technique involves simultaneous measurement of two γ-rays at different energy levels using a multichannel analyzer. In this set up the [137]C$_s$ is placed behind the [241]Am source (Nofziger and Swartzendruber, 1974; Nofziger, 1978).

Equation (10.8) can be solved for both moisture content and soil bulk density. Let N_p, N_s, and N_{sw} be the count rates through an empty column, through a column packed with oven dry soil, and through a column containing soil and through the column containing soil and water or wet soil, respectively. Then Eq. (10.8) can be written for dry and wet soils as Eqs. (10.9) and (10.10), respectively.

$$\ln \frac{N_p}{N_s} = \mu_s \rho_s x$$

(10.9)

$$\ln \frac{N_p}{N_{sw}} = -(\mu_s \rho_s + \mu_w \theta) x$$

(10.10)

Dividing Eq. (10.10) by Eq. (10.9) yields Eq. (10.11) and Eq. (10.12).

$$\ln \frac{N_s}{N_{sw}} = \mu_w \Theta x \tag{10.11}$$

or

$$\theta = \frac{2.3}{\mu_w x} \log_{10} \frac{N_s}{N_{sw}} \tag{10.12}$$

The γ-scanning equipment has been designed for both laboratory and field use and details of such devices are available in Gardner (1986) and Catriona et al. (1991).

Merits and limitations of the γ-scanning technique are similar to those of the neutron scattering method. Perhaps the health hazards are more with γ-scanning than with neutron scattering method.

Dielectric Properties of Soil

The dielectric constant of a material is the ratio of the value of the capacitor with the material between the plates, compared with the value with air between the plates. In comparison with a metal, a dielectric material is an insulator. When subjected to an electric field, the positive and negative charges in a dielectric material are displaced with respect to each other and tiny electric dipoles are produced. The dipoles are aligned by the electric field and the dielectric medium as a whole becomes polarized. Therefore, the dielectric constant is a measure of the polarization of a substance. Some materials (e.g., water) whose molecules have a permanent dipole moment have a large dielectric constant. The dielectric constant of water is about 80 and that of the soil about 5 to 7 (Table 10.3).

Principal properties of a dielectric material are: (i) dielectric constant, (ii) dielectric loss, and (iii) dielectric strength. The dielectric constant is

TABLE 10.3 Dielectric Constant *(E)* of Some Materials at 20°C

Material	Dielectric constant K
Vacuum	1.0000
Air (1 atm)	1.0006
Paraffin	2.2
Rubber, hard	2.8
Vinyl (plastic)	2.8–4.5
Paper	3–7
Quartz	4.3
Glass	4–7

Porcelain	6–8
Mica	7
Ethyl alcohol	24
Water	80

Source: Adapted from Weast, 1987.

the factor by which the electric field strength in a vacuum exceeds that in the dielectric for the same distribution of charge. The dielectric loss is the amount of energy it dissipates as heat when placed in a varying electric field, and dielectric strength is the maximum potential gradient it can stand without breaking down.

Dielectric constant *(E)* is the ratio of the capacity of a condenser with that substance as dielectric to the capacity of the same condenser with a vacuum for dielectric. It is a measure, therefore, of the amount of electric charge a given substance can withstand at a given electric field strength. The dielectric constant is measured in units of hertz, which is a unit of frequency; 1 Hz equals 1 cycle/second. Two methods of soil moisture measurements are based on the dielectric properties of the soil. These methods are as follows.

The Capacitance Method. A capacitor is a device that can store electric charge. It consists of two conducting objects placed near each other but not touching. A typical capacitor consists of parallel plates of area *A* separated by small distance. When voltage is applied, the capacitor becomes charged. The amount of charge acquired by each plate is proportional to the potential difference *V (Q=CV)*. The constant of proportionality *C* is called capacitance. The capacitance method involves using the moist soil as a part of the dielectric of a capacitor. Measurement of the capacitance gives the dielectric constant, which changes with the soil's moisture content.

There is a wide range of capacitance electrodes (Schmugge et al., 1980). Rather than using probes or push-in electrodes inserted directly into the soil, electrodes or probe can be inserted into an access tube similar to that of the neutron moisture meter. However, there should be no or minimal air gaps between the access tube and the soil. Push-in electrodes are useful for measurement of soil moisture at shallow depths, where soil is highly heterogenous and measurements are extremely variable and unrepeatable. Using access tube is the best method of measurement (Thomas, 1966; Bell et al., 1987; Dean et al., 1987). The capacitance is usually measured by a bridge method at a frequency range of 30–3000 MHz.

The capacitance method has numerous advantages. It is economic, safe, without legal constraint, stable, and rapid by manual operations. Because it involves the use of an access tube, the operation is similar to that of the neutron probe but is much safer and free from legal/policy constraints. However, the techniques require calibration which may be influenced by the composition and density of soils. This method is also not sensitive to the water held by surface adsorption forces or in chemical association with humus, sesquioxides.

Time Domain Reflectometry (TR). This method is also based on the measurement of the dielectric constant of the soil (Topp et al., 1980; 1982; 1988; Topp, 1993; Dalton et al., 1984; 1986). High-energy electromagnetic pulse is fed into the soil between two

metal rods. A part of the pulse is reflected back up through the soil from the bottom of the rods and the time interval for the pulse to traverse back, or the time interval between the incident and reflected pulse, is measured. This time interval is related to the soil's moisture content. Major differences between the TDR and the capacitance methods are that the TDR method

> Measures an average dielectric constant over the length of the rod
> Uses a pair of parallel rods inserted in the ground
> Measures dielectric constant over a broad band of frequencies usually ranging from 100 to 1000 MHz
> Measures electrical conductivity and dielectric constant simultaneously.

The velocity (v) of an electromagnetic wave through a transmission line in a nonmagnetic medium is given by Eq. (10.13).

$v=C/K^{1/2}$

$$(10.13)$$

where C is the velocity of light $(3\times10^8$ m/s) and K is dielectric constant of the nonmagnetic medium, such as soil. For H_2O with a dielectric constant of 80, the v is 3.3×107 m/s. For applicaion to soil-moisture determinations, TDR is essentially a cable radar in which the velocity is computed to measure the time interval (t) for the wave to traverse back and forth in the rod of length L $(v=2L/t)$. Substituting $2L/t$ for v in Eq. (10.13), we can solve for dielectric K_a of the soil [Eq. (10.14)].

$$K_a = \left(\frac{Ct}{2L}\right)^2 \qquad (10.14)$$

where K_a is the apparent dielectric constant of the soil which varies with soil wetness. Topp et al. (1980) observed that the dielectric constant does not vary with texture, porosity, and proposed a polynomial equation relating K_a to Θ [Eq. (10.15a)].

$$\theta = -5.3 \times 10^{-2} + 2.92 \times 10^{-2}K_a - 5.5 \times 10^{-4}K_a^2 + 4.3 \times 10^{-6}K_a^3$$

$$(10.15a)$$

However, θ vs. K_a relationship is affected by soil's organic matter content especially for organic soils (Herkelrath et al., 1991), and the calibration may also be influenced by salinity (Baumhardt et al., 2000; Nadler et al., 1999). The technique can also be used for simultaneous measurement of soil's moisture content and soil-moisture potential (Noborio et al., 1999) (see Chapter 11). Details of the theoretical principles are outlined by Topp et al. (1980; 1982), Dalton et al. (1984), Catriona et al. (1991), Zegelin et al. (1992), Topp (1993); Topp et al. (2000), and Nadler et al. (2003). The technique is presently being used to assess water and solute transport, and penetrometer resistance in sols (Vaz and Hopmans, 2003; Vaz et al., 2002; Caron et al., 2002). This method has numerous advantages of the neutron scattering and γ-ray attenuation methods, yet is free from health hazard and nuclear regulation. However, calibration of the method and its reliability and reproducibility are still to be worked out.

The TDR technique is still in its evolutionary stage, and rapid progress is being made in alleviating methodological constraints (Malicki and Shierucha, 1989; Zegelin et al., 1989) and in automating the procedure (Baker and Allmaras, 1990).

Thermal Conductivity

Soil's thermal conductivity increases with an increase in soil's moisture content (see also Chapter 17), and this relationship can be used to measure soil wetness (Shaw and Baver, 1939). The temperature rise depends on the ability of the soil to conduct heat away from the source, which depends on soil's moisture content. A principal advantage of this method is that the measurement is not affected by soluble salts that are present in the soil, and the method also measures soil temperature, and the effect of soil temperature on moisture measurement can be accounted for. The technique involves placing a heating element and a temperature sensor in the soil, and the time required to increase soil temperature by a predetermined value is measured. There are two types of equipment based on: (i) encasement of the sensor and element in a porous medium (Sophocecus, 1979) and (ii) placement directly in the soil (Fritton, 1969). The first technique is more suited to measure soil-moisture's potential than moisture content because it reflects the equilibrium moisture content of the porous block. In contrast, the direct placement technique may have a limitation of the poor soil-probe contact, especially in soils with high swell-shrink capacity.

Remote Sensing

Methods of measuring soil moisture described in the previous sections are applicable at the pedon level for different depths or at plot level by simultaneous measurements at several locations. The in situ measurement of the distribution of soil moisture at a watershed scale is difficult because it requires the instruments that can remotely sense it with reasonable accuracy. Ulaby et al. (1996) described a technique of surface soil wetness. Reflectance properties (albedo) can be correlated to the degree of soil wetness. Remote sensing techniques involve use of airborne and satellite imagery procedures. Such can be used for estimating soil's moisture content of the surface layer to a maximum depth of only 0.3 m. These measurements are considerably influenced by ground cover, cloud cover, and other objects between soil and the sensing devices in the space (e.g., crop residue mulch). Remote sensing techniques estimate soil's moisture content over relatively large areas.

Potentials and limitations of remote sensing techniques have been discussed in detail by Myers (1983). These procedures are based on the following five techniques:

Digital Elevation Models (DEMs). Space borne differential interferometric synthetic aperture radar data (InSAR, C band) have the potential for measuring soil moisture at watershed scale (Nolan and Fatland, 2003). The differential InSAR is a powerful tool for making DEMs and is capable of separating surface deformations from static topography. The recent, more accurate DEMs can detect topographic noise to submillimeter range. The spatial variations of SAR are correlated in many locations where changes in soil moisture are expected such as in stream channels, farm boundary, and watershed divide. The underlying theory is that the changes in soil moisture affect soil permitivity

(dielectric constant) and the penetration depth. However, penetration depth varies inversely to the soil wetness and the relationship is nonlinear. The rapid advances in the global positioning system (GPS) and inertial motion compensation technology have the potential of increasing accuracy with the added benefit of acquiring the data at any temporal resolution (Nolan and Fatland, 2003).

γ-Radiation. Soils natural emission of γ-rays is related to soil moisture content changes overtime. This method may be accurate within 10% for the top 30 cm layer (Grasty, 1976; Zotimer, 1971; Carroll, 1981). The γ-ray flux can be measured by a sensor placed on a low-flying aircraft at 100–200 m altitude (Salomonsen, 1983). The spatial resolution for this technique is at least 200 m. Therefore, variations in moisture content due to differences in soil at small distances cannot be detected. This technique may be useful for large tracts of extremely homogenous soils (e.g., recent alluvial or loess deposits, Andisols, etc.).

Visible and Near Infrared Spectrum. Soil's color changes with its moisture content; moist soil is darker in color. This implies that the spectral reference of soil for the visible and near infrared wavelengths decreases with increase in soil's moisture content (Condit, 1970). However, soil color and its spectral characteristics also differ due to differences in soil's organic matter content, texture, cloud cover, ground cover, and lighting conditions (Evans, 1979; Moore et al., 1975). Soil's moisture content and soil type also affect polarization characteristics of visible light. The degree of polarization of light can also be related to soil's moisture content (Stockhoff and Frost, 1972).

Thermal Infrared Radiation. Changes in surface soil temperature due to differences in soil's moisture content can be monitored and related to soil wetness. Surface soil moisture content has been related to soil temperature using an airborne thermal scanner (Cihlar et al., 1979; Elkington and Hogg, 1981).

Microwave Techniques. Changes in dielectric properties of soil at different soil moisture contents are measured in terms of the microwave energy emitted (Schmugge et al., 1974; Njoku and Kong, 1977).

Acoustic Properties

The propagation of low-energy ultrasonic waves has been used as a non-destructive method for determining moisture content of soils. Such waves propagate at certain sinusoidal frequencies (megacycles), at which the propagated energy varies with soil moisture content. Energy propagated at frequencies of 16 to 20 megacycle/s is sensitive to changes in soil's moisture content in the low range of w from 0 to 10% by weight. Energy propagated at frequencies of 114 to 142 megacycle/s is sensitive to soil moisture content in the high range of w up to 50%. The energy propagated, however, is also influenced by the presence of soluble salts in the soil (Ghildyal, 1987).

Chemical Properties

Several direct and indirect methods of soil-moisture determinations are based on soil's chemical properties. Some of these methods include the following:

1. Changes in the concentration or specific gravity of alcohol (ethyl, methyl, or propyl) when placed in contact with wet soil are related to soil's moisture content.

2. The pressure of the acetylane gas generated in a closed system when calcium carbide is mixed with a moist soil depends on soil wetness [Eq. (10.15b)].

$$CaC_2 + 2H_2 \rightleftharpoons OC_2H_2 + Ca(OH_2)$$

$$(10.15b)$$
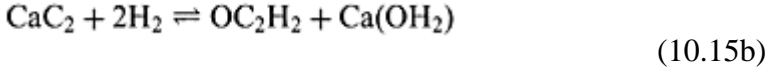
The equipment called Speedy Moisture Tester or Gas Moisture Tester is based on this principle. Known amount of soil, usually 10−25g, is mixed with about 25 g of CaC_2 and the pressure of the gas generated is measured and related to soil's moisture content.

3. The heat evolved when the wet soil is placed in a concentrated H_2SO_4 solution is also measured and related to soil's moisture content.

4. Changes produced in the electrical conductivity of the system when water in soil is displaced with alcohol, acetone, and other organic liquids can be related to soil wetness.

Volume Displacement Method

This method is based on assessing the increase in volume of water when a known amount of wet soil is immersed in a known volume of water, and all entrapped air is removed (Prihar and Sandhu, 1968).

$$\Delta V = V_s + V_w$$

$$(10.16)$$

$$\Delta V = \frac{M_s}{\rho_p} + V_w$$

$$\therefore V_w = M_w \quad \text{for } \rho_w = 1.0 \text{ g/cm}^3 \text{ and } M_{ws} \text{ is mass of wet soil}$$

$$\Delta V = \left(\frac{M_{ws} - M_w}{\rho_p}\right) + M_w$$

$$\Delta V \cdot \rho_p - M_w \cdot \rho_b = M_{ws} - M_w$$

$$\therefore M_w = \frac{(\Delta V \cdot \rho_p - M_{ws})}{(\rho_p - 1)}$$

$$w = \frac{M_w}{M_{ws} - M_w}$$

10.3 COMPARATIVE ADVANTAGES AND LIMITATIONS OF DIFFERENT METHODS

Among the wide range of methods available, the choice of an appropriate method of determination of soil's moisture content depends on numerous factors including the objectives, soil properties, site accessibility, resources available, and technical expertise. Further, different methods are suitable for specific soil characteristics. Merits and

limitations of different methods are outlined in Table 10.4. Special precautions should be taken for soils with gravel content. Most techniques are not suitable for soils with high gravel content. Furthermore, computations of volumetric moisture content (Θ) from gravimetric moisture content *(w)* require knowledge of ρ_b of the gravel-free fraction.

10.4 EXPRESSION OF RESULTS

There are numerous ways to express results of soil moisture content measurement. Among 14 methods listed in Table 10.5, the most useful and commonly used indices are those identified with an asterisk (*). Volumetric moisture content (expressed either as a fraction or a percentage) depth of soil moisture, and saturation percentage are the most useful and commonly used indices.

TABLE 10.4 Merits and Limitations of Different Methods of Determining Soil's Moisture Content

Method	Advantage	Disadvantages
Thermogravimetric	Simple, inexpensive, routine, and the most direct method	Time consuming, laborious, destructive sampling, high variability, measurement of ρ_b is necessary, same site cannot be measured.
Neutron moisture meter	Large soil volume, directly measures Θ, technically sound method, easily computerized	Expensive, health hazards, subject to nuclear regulations, not accurate for soil layers, neutron meter not suitable for organic soils.
Electrical conductance	Simple, low cost, easy to install, nondestructive	Not suitable for soils with high salt content, and soils of low pH, calibration changes with time.
TDR	Nondestructive, simple equipment (metal rods), no health hazards and nuclear regulation	Expensive, still evolving, limited depth range highly variable results,
Gamma scanner	Nondestructive, also measures soil bulk density	Very high health risks, cumbersome equipment especially with double source.
Thermal conductivity	Useful for saline soils, simultaneous measurement of soil temperature	Highly variable results due to poor contact, not applicable for soils with high swell shrink capacity due to contact problems on cracking.
Remote sensing	Large resolution, nondestructive rapid	The measurements cover a large area comprising several soils, results valid only for the surface layer, interference with cloud cover, vegetation and other land features.

TABLE 10.5 Methods to Express Soil's Moisture Content

1 and 2	Mass water fraction $(w)=M_w/(M_w+M_s)$ (fraction or %)
3 and 4*	Gravimetric moisture content $(w)=M_w/M_s$ (fraction or %)
5 and 6*	Volumetric moisture content $(\Theta)=V_w/V_t$ (fraction or %)
7*	Depth of water $(d)=(\Theta$ as fraction)×(depth of soil column/ profile/layer in units of length)
8	Soil moisture density $(\rho_m)=M_w/V_t$ (g/cm^3, Mg/m^3)
9 and 10	Saturated water holding capacity on gravimetric bases $(W_c)=M_w$ at $\Theta=s/M_s$ (fraction or %)
11 and 12	Saturated water holding capacity on volumetric bases $(\Theta)=V_w$ at $\Theta=s/V_t$ (fraction or %)
13*	Liquid ratio $(\Phi_\rho)=V_w/V_s$
14	Saturation percent$=(V_w/f_t)\times100$

* Important and very useful.

PROBLEMS

1. Compute soil moisture content of a 20 g of wet sample that registers an increase in volume by 5 cm^3. Assume ρ_s of 2.7 g/cm^3.

2. The following soil data were obtained for an irrigation experiment with corn. Irrigation of 10 cm was applied on 6/10/88 after monitoring the soil moisture.

Depth (cm)	Bulk density (g/cm^3)	Wilting point (w g/g)	Field capacity (w, g/g)	Soil moisture content (g/g)	
				6/10/88	6/20/88
0–30	1.2	0.10	0.30	0.10	0.20
31–50	1.3	0.12	0.32	0.15	0.25
51–80	1.4	0.14	0.28	0.25	0.20
81–150	1.6	0.15	0.25	0.20	0.15

(a) Calculate depth of penetration of irrigation water,

(b) Evaluate evapo transpiration of corn in mm/day,

(c) Determine drainable porosity at field capacity assuming ρ_s=2.65 g/cm^3.

(d) If irrigation is withheld as from 6/20/88, how long will it take for corn crop to exhaust the entire water reserves if the ET continues at the rate computed in 'b' above?

3. Plot a calibration curve for the neutron moisture meter from the following data:

	Volumetric moisture content (Θ_v)							
	0.5	0.4	0.3	0.2	0.1	0	Standard count	
Soil				CPM (10^3)				
A		34	28	24	15	8	1	20
B		60	50	48	37	30	20	

(a)   Develop an empirical relation for predictive purposes.

(b)   Estimate Θ_v for a count of 32×10^3 CPM.

(c)   Suggest possible reasons for differences in calibration curves among two soils.

4.   Describe theoretical principals and practical limitations of a neutron moisture meter.

5.   Prepare a matrix of the merits and demerits of different methods of moisture measurement for soil profiles with the following characteristics:

(a)   Gravelly soil

(b)   Soil with low pH

(c)   Saline/sodic soil

(d)   Peat soil

(e)   Soil with high contents of Fe and Mn

(f)   A layered profile

6.   Describe the TDR method giving its principles, equipment, and merits in relation to the neutron moisture meter.

7.   Why is expressing soil moisture content on volumetric basis more useful than mass or gravimetric basis?

8.   How do soil structure, aeration, and soil strength influence available water holding capacity?

9.   How do soil organic matter and clay contents influence plant-available water capacity?

10.   What technologies do you suggest to improve waterholding capacity of coarsetextured soils?

11.   Net weight of a wet soil core 7.5 cm in diameter and 7.5 cm deep is 600 g. Calculate wet and dry density and equivalent depth of water if the oven dry weight of the core is 500 g.

12.   A soil clod has a volume of 100 cm^3, gravimetric moisture content of 0.20, and bulk density of 1.5 mg/m^3. Calculate the degree of saturation (s) and air-filled porosity (f_a).

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11
Soils Moisture Potential

Soil's moisture content by itself, regardless of its method of expression in any of the 14 different ways, is not sufficient to describe the status of water in soil. There are several hydrological processes that cannot be fully explained on the basis of soil's moisture content alone. These processes include: (i) water absorption by plant roots, which differs among soils with different textures that have similar moisture content, (ii) water movement that may occur from one soil to another although their moisture contents are similar, and (iii) different soil moisture contents may occur in soils with similar management or environmental conditions. In addition to the moisture content, another property that is essential to a complete description of the soil water regime is the energy status of water in the soil. Soil's moisture content is similar to the heat content of a body. It is the index of a system's capacity in contrast to temperature, which is a measure of its intensity. Similarly, soil's moisture content is a measure of the capacity factor while the energy status of the water is an index of its intensity.

11.1 ENERGY STATUS OF SOIL MOISTURE

Soil water, similar to other natural bodies, possesses two forms of energy: (i) potential energy due to its position or configuration relative to a reference point and (ii) kinetic energy by virtue of its motion (equal to $1/2$ mV^2 where m is mass and V is velocity). In addition, change of state of water (e.g., solid, liquid, vapor) due to differences in temperature can also affect its kinetic energy. The gravitational potential energy of soil moisture is the product of its weight (mg) and height *(h)* above a reference point or mgh. The gravitational potential energy is the work done by gravity in moving the mass *m* of water from point A to point B, *h* distance apart. The potential energy depends on the vertical height of soil moisture above some reference level.

In practical terms, water in soil moves at a very slow velocity, and possesses an extremely low level of kinetic energy. Further, most processes involving soil-water and plant-water systems are primarily governed by changes in potential energy of soil water and can be addressed without considering the kinetic energy. This is especially true in systems, which are isothermal. In addition, the potential energy of soil water can be substantial and an important factor governing the status of soil water. Water movement under isothermal conditions in soil, both in terms of its direction and velocity, is to a

large extent governed by its potential energy. It is primarily because of the differences in this potential energy that water moves from one place to another in the direction of decreasing potential energy until it reaches an equilibrium state determined by equal potential energy at all points within a soil system connected via transmission pores. The driving force is the rate of change of potential energy with distance. It is not the absolute quantity of potential energy but the relative level of energy for one region vis-à-vis another that governs the rate, magnitude, and direction of water movement.

11.2 SOIL-MOISTURE POTENTIAL

Soil-moisture potential refers to this relative level of the potential energy contained in the soil water. It is a measure of the relative potential energy of water in the soil in comparison with pure water. In other words, soil-moisture potential is an expression or indicator of the potential energy contained in soil water relative to that of water in a standard reference state. The latter is a reservoir of pure water (no salts) at atmospheric pressure (not confined) and at the same temperature and level as the soil moisture.

Soil water is subject to the work-energy principle, which states that the work done by an object is equal to change in its energy status. If positive work is done on soil water, soil water's potential (energy) status increases equal to the work w done on it. If negative work w is done on soil water, the soil-moisture potential (energy) decreases by an amount w. In contrast with free water, soil water is held by the soil matrix because of the forces of adsorption involving cohesion, adhesion, and solution. Therefore, soil water is usually not capable of doing work W as can a reservoir of pure water. Consequently, soil water potential is usually negative.

Thus, soil water potential has the following characteristics:

> *Relative:* It is a relative quantity.
> *Negative:* It is usually negative.
> *Continuity.* It is a continuous entity without any abrupt discontinuities.
> *Driving force:* It is the driving force that moves soil water from one region within the soil to another.
> *Variability:* It is highly variable even over short distances within the soil.
> *Dynamic:* It is a highly dynamic entity.

In view of these characteristics, soil-moisture potential, hereafter designated by the symbol Φ, is defined as "the amount of work that a unit quantity of water in an equilibrium soil-moisture system is capable of doing when it moves to a pool of water in the reference state at the same temperature."

Total soil-moisture potential (Φ_t) is the amount of useful work per unit quantity of pure water that must be done by means of externally applied forces to transfer irreversibly and isothermally an infinitesimal amount of water from the standard state to the soil liquid phase at the point under consideration (Bolt, 1976). Total soil water potential is measured in units of energy, which can be expressed per unit mass, volume or weight basis as follows (see also Sec. 11.6):

1. Energy per unit volume is expressed as ergs/cm^2, dynes/cm^2, N/m^2,
2. Energy per unit mass is expressed as ergs/g or J/kg, and
3. Energy per unit weight is expressed in terms of height of water as cm or m.

11.3 COMPONENTS OF TOTAL SOIL-MOISTURE POTENTIAL

Total soil-moisture potential (Φ_t) consists of several components [Eq. (11.1)]:

$\Phi_t = \Phi_p + \Phi_m + \Phi_z + \Phi_\pi + \Phi_o$

$$(11.1)$$

where $t, p, m, z, n,$ and o refer to total, pressure, matric, height or position, osmotic, and overburden potential, respectively.

11.3.1 Pressure Potential (Φp)

Pressure potential (Φ_p) is defined as the water pressure exerted by the overlying saturated column of water on a specific position within a soil. It is equal to the water pressure exerted by the height of water above a specific point. If a volume Θ is transferred from a body of water where the gauge pressure is zero to one where it is $p,$ the work done against p is [Eq. (11.2)]

$\Phi_p = \text{work} = pv$

$$(11.2)$$

The work per unit volume is $pv/v = p$. The work done by water can also be computed by assuming this water to be displaced from a tube of length l and cross-sectional area A into water at pressure p. The work done in this hypothetical case against pressure p is $W = p \cdot A \cdot l = pv$.

Therefore,

Work (Φ_p) per unit volume $= pv/v = \rho gh$ dynes/cm^2

$$(11.2a)$$

Work (Φ_p) per unit mass $= \dfrac{pv}{\rho v} = (\rho gh)\, v/\rho v = gh$ ergs/g

$$(11.2b)$$

Work (Φ_p) per unit weight $= \dfrac{pv}{\rho g v} = (\rho gh) = v/\rho gv = h$ cm

$$(11.2c)$$

where p is density (g/cm^3) and g is acceleration due to gravity (cm/s^2). The pressure head is usually measured in units of length (cm, m), and exists and only under saturated soil conditions ($\Theta = s = 1$). The positive pressure potential usually occurs below the groundwater level and is called the piezometric head or the submergence potential. Under field conditions, the pressure potential is measured by a piezometric tube. A piezometer tube is a solid tube open at both ends, and a water table tube is a perforated tube open at both ends (Fig. 11.1). The pressure potential is the vertical distance from a specific point in the soil to the water surface of a piezometer
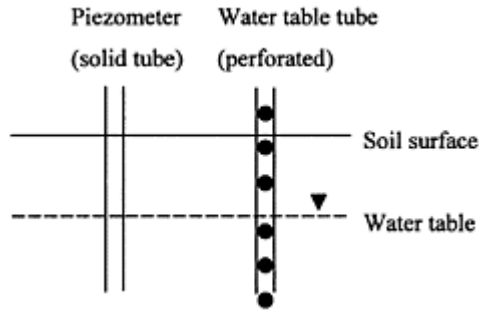
FIGURE 11.1 Piezometric pressure at different points.
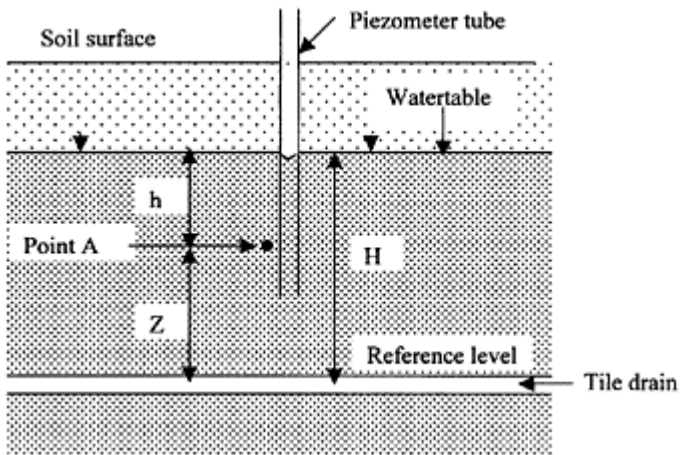

FIGURE 11.2 A piezometer tube showing the soil water pressure below the water table. At the reference point A, the pressure potential equals h ($\phi_p = h$ cm), gravitational potential z, and the total potential $H=(h+Z)$.

connected to the point in the question. The schematic in Fig. 11.2 shows the magnitude of Φ_p.

In the field situation, Φ_p is zero above and at the level of the water in the piezometer. It is positive and equal to the depth of the water column above, when the point is below the water table.

11.3.2 Matric Potential (Φ_m)

Matric potential exists only in unsaturated soils, and therefore, matric potential and pressure potential are mutually exclusive. Under specific soil-water conditions, a soil either has pressure potential (Φ_p) or matric potential (Φ_m), but not both. Soil matric potential is due to the effects of soil solids, interfacial curvature due to surface tension and forces of cohesion and adhesion of the soil matrix (Fig. 10.2). This negative pressure potential is also called *capillary potential*. Similar to the potential, the matric potential may be expressed in three units.

Φ_m per unit volume=ρgh dynes/cm^2

(11.3a)

Φ_m per unit mass=gh ergs/g

(11.3b)

Φ_m per unit weight=h cm

(11.3c)

Some soil physicists (Jury et al., 1991) argue that Φ_m comprises tensiometric potential (capillary potential) and air potential (pneumatic potential). The tensiometric potential is the work required to transfer reversibly and isothermally an infinitesimal amount of soil solution from a reservoir in soil to the point of interest in the soil. In comparison, the air pressure potential is the gauge pressure of the soil air relative to the standard state air pressure ($P_{\text{soil}}-P_{\text{atmosphere}}=\Delta P_a$). The gauge pressure of the soil air with reference to the ambient pressure, called *pneumatic potential* (Φ_a), is usually negligible. In unsaturated soils, therefore, the matric potential is the sum of capillary potential and the pneumatic potential. Under laboratory conditions, however, Φ_a is important. The Φ_a is used to measure soil moisture retention at different matric potentials (see Sec. 11.7). In that condition, $\Phi_a=\Phi_m$. In practical terms, however, the matric potential is the same as the tensiometric potential or the capillary potential because Φ_a is practically zero.

Matric potential is measured by tensiometers. Tensiometer is a device that measures potential energy of soil water relative to free water in a porous ceramic cup in equilibrium with soil water. A graphical representation of different types of tensiometers is shown in Fig. 11.3a–c. In Fig. 11.3a, Φ_m is the vertical distance between the point in the soil and the water surface of a manometer filled with water and connected to the soil point in question via a ceramic cup. This device is called a tensiometer or a ceramic cup tensiometer (Fig. 11.3).

A tensiometer consists of a porous cup and a monometer or a pressure gauge. The ceramic (or any other suitable porous material) cup and part of the manometer are filled with deaired water and buried in soil at the desired depth. Depending on the soil wetness, the water moves from the cup into the soil and develops a negative pressure or suction, which is measured by the depression in the height of the water in the manometer tube or in the gauge pressure attached to the cup.

Depending on the system used to measure the suction created by the movement of water from the ceramic cup to the soil, there are numerous types of tensiometers.

Mercury Manometer Tensiometer

These tensiometers use a combination of H_2O and Hg to measure the Φ_m as shown in Fig. 11.3b. The use of Hg is a health hazard. Therefore, this following description is merely to explain the underlying principles.
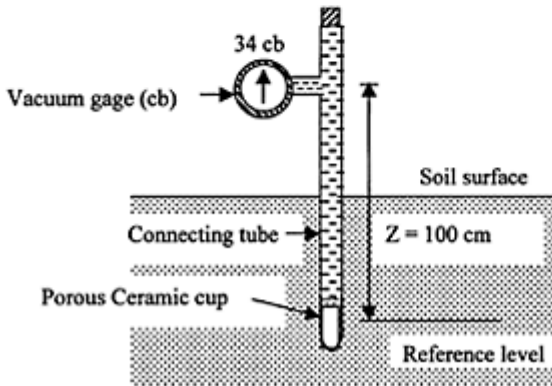
Z=distance from top of the mercury column to the center of the ceramic cup.

Z_{Hg}=distance from top of the mercury column to the surface of the mercury in the reservoir.
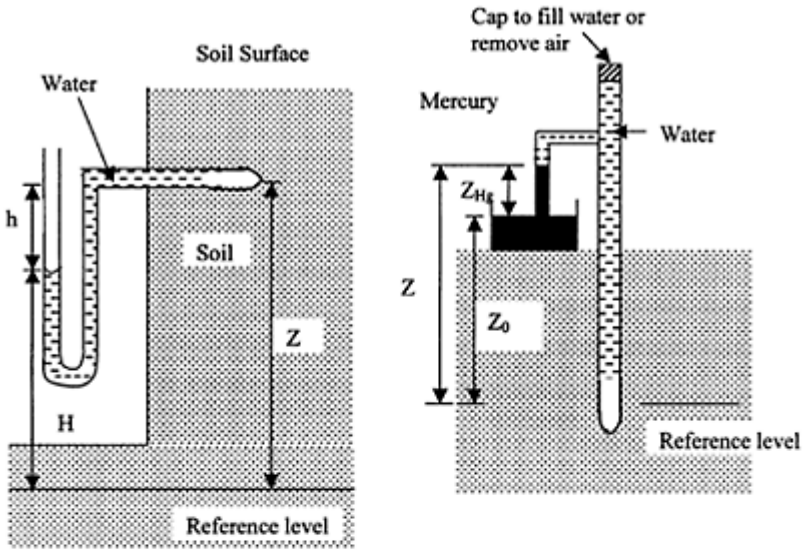
(a)

(b)

(c)

FIGURE 11.3 Different types of tensiometers: (a) a water manometer connected to a ceramic cup installed in soil at the designated depth, Φ_m equals $-h;$ (b) a mercury manometer connected to a ceramic cup installed in soil at the desired depth, $\Phi_m = -Z_{Hg} \times 13.6 + Z;$ and (c) a vaccum gauge tensiometer, $\Phi_m = -34 \times 10$ cm $+ 100$ cm $= -240$ cm.

$Z_o =$ distance from the top of the mercury level in the reservoir to the center of the ceramic cup.

$$\Phi_m = -Z_{Hg}\frac{\rho_{Hg}}{\rho_w} + Z \qquad (11.4)$$

$\rho_{Hg} = 13.6$ g/cm^3
$\rho_w = 1.0$ g/cm^3
$\Phi_m = -13.6\, Z_{Hg} + Z$

$$(11.5)$$

The distance Z varies as the height at mercury column changes. If the distance from the surface of the mercury reservoir to the center of the cup is kept constant (h_o) we have a constant for any tensiometer:

$Z = Z_o + Z_{Hg}$

$$(11.6)$$

Substituting Eq. (11.6) in (11.4)

$$\Phi_m = -Z_{Hg}\frac{\rho_{Hg}}{\rho_w} + Z_{Hg} + Z_o \qquad (11.7)$$

$$\Phi_m = -Z_{Hg}\left[\frac{\rho_{Hg}}{\rho_w} - 1\right] + Z_o \qquad (11.8)$$

$\Phi_m = -12.6\, Z_{Hg} + Z_o$

$$(11.9)$$

Example 11.1

If $Z_o = 20$ cm, $Z_{Hg} = 14.2$ cm, calculate Φm

Solution
$\Phi_m = -12.6 \times 14.2$ cm $+ 20$ cm
$= -17.9$ cm $+ 20$ cm $= -159$ cm

Vacuum Gauge Tensiometer

In this tensiometer, the Hg is replaced by a vacuum gauge, and the reading on the dial can be converted to Φ_m (Fig. 11.3c). The units of measurement must be carefully considered.

The dial is usually calibrated from 0 to 100, which on a weight basis corresponds to a range of 0 to −1000cm (0 to −100 centibars).

Example 11.2

Calculate Φ_m in Fig. 11.3c. 1 gauge reading=10 cm of Φ_m.

Solution

$\Phi_m=-10$ cm×(gauge reading)+Z
$\Phi_m=-10×34$ cm+100 cm
$\Phi_m=-240$ cm

Vacuum gauge may be also calibrated in inches of Hg rather than in cm or centibars.

Example 11.3

A tensiometer dial is calibrated from 0 to 30. If the gauge is 25 inches above the tensiometer cup and it reads 20 inches of Hg, calculate Φ_m.

Solution

$\Phi_m=-13.6×20+25$ inches
$\Phi_m=-247$ inches of water
$\Phi_m=-627$ cm of water

Most commercially available tensiometers may already be calibrated for the length of the tensiometer stem. There are two principal limitations of tensiometers. First concerns with the range of suction, or Φ_m, that can be measured with a tensiometer. The useful range is about 0 to 80 kPa, or 0 to 800 cm of water suction. As soil gets drier than this range, air enters the cup and water column in the tensiometer breaks. Soil moisture content corresponding to this suction varies widely among soils, depending on the texture and organic matter content. The second limitation is due to the response time of the tensiometer. In soils with rapidly changing Φ_m, tensiometers are usually slow to respond. The response time depends on hydraulic conductivity of the porous cup and sensitivity of the gauge or the suction-measuring devices.

11.3.3 Gravitational Potential (ΦZ)

The Φ_z is due to the position of soil water. It is the energy required to move an infinitesimal amount of pure, free water from the reference elevation to the soil water elevation. Therefore, the Φ_z of soil moisture is determined by the elevation of the point relative to the reference level. Three forms of expressing Φ_z are shown by Eq. (11.10).

Φ_z per unit volume=$\rho g Z$ dynes/cm^2

(11.10a)

Φ_z per unit mass=gZ ergs/g

(11.10b)

Φ_z per unit weight=Z cm

(11.10c)

The gravitational potential is usually measured by the height above or below an arbitrarily chosen reference point. The gravitational potential is positive if the specific point is above the reference level, and negative if the specific point is below the reference level. The Φ_z is strictly due to the position of a specific point in the soil, and is independent of the soil properties or atmospheric (ambient) conditions. Its magnitude depends on the vertical distance between the reference and the point in question.

11.3.4 Osmotic Potential (Φ_o)

Osmotic potential is due to the presence of solutes in soil moisture that affect its thermodynamic properties (e.g., entropy, enthalphy, free energy). Presence of solutes in soil lowers the vapor pressure of soil moisture and affects its Φ_o. The Φ_o refers to the change in energy per unit volume of water when solutes identical in composition to the soil solution at the point of interest in the soil are added to pure, free water at the elevation of the soil. Presence of solutes in soil moisture creates a suction that can suck water from a reservoir of pure water brought into contact with the solution through a semipermeable membrane. The ability of soil moisture to suck water from a reservoir of pure water depends on the concentration of solutes, which also determines decrease in its vapor pressure, increase in boiling point, and depression in its freezing point. The Φ_π can be expressed in three ways as per Eq. (11.11).

Φ_π per unit volume=$\rho g h \pi$ dynes/cm^2

(11.11a)

Φ_π per unit mass=$gh\pi$ ergs/g

(11.11b)

Φ_π per unit weight=$h\pi$ cm

(11.11c)

TABLE **11.1** Components of Total Soil-Water Potential

Soil	Components of Φ_t	Remarks
Saturated soil		
Nonswelling soil	$\Phi_t = \Phi_z + \Phi_p + \Phi_\pi$	Φ_π is zero for soils in the humid region.
Swelling	$\Phi_t = \Phi_z + \Phi_p + \Phi_o$	
Unsaturated soil		
Nonswelling soil	$\Phi_t = \Phi_z + \Phi_m + \Phi_\pi$	
Swelling soil	$\Phi_t = \Phi_z + \Phi_m + \Phi_o + \Phi_a + \Phi_\pi$	Φ_a is usually 0.

The osmotic potential is also discussed in Chapter 20 in section dealing with soil salinity.

11.3.5 The Overburden Potential (Φ_o)

The Φ_o is due to the mechanical pressure exerted by the unsupported solid material on the soil water. It is the change in energy per unit volume of soil water due to the weight of the unsupported soil above the soil water. The overburden pressure is usually significant only in swelling soils (see Chapter 20).

Components of Φ_t under different situations are shown in Table 11.1. For most saturated soil situations, Φ_t comprises only two components, the gravitational potential (Φ_z) and the pressure potential Φ_p [Eq. (11.12)]. Under this case Φ_t is called the *hydraulic head*.

Hydraulic head (Φ_t) = $\Phi_z + \Phi_p$

$$(11.12)$$

11.4 TOTAL SOIL-MOISTURE POTENTIAL UNDER FIELD CONDITIONS

Components of soil-moisture potential can be measured under field conditions for assessing the direction and magnitude of flow. A line joining all points with equal soil-moisture potential is called an *isobar*. Soil water flows perpendicular to the isobars. There is no water movement in the soil if Φ_t is equal at all points. Soil water moves in the direction of decreasing soil-moisture potential.

Example 11.4

With 10 cm of water ponding and maintained constant on the soil surface and a tile drain at 100 cm depth flowing full, plot the soil moisture potential profile.

Solution

Components of soil moisture potential in this case are pressure potential (Φ_p) and gravitational potential (Φ_z). Because it is saturated, flow Φ_m is zero. Taking soil surface as a reference point, components of Φ_t are as follows:

Φ_p (cm)	Φ_z (cm)	Φ_t (cm)
10	0	10
30	−20	10
50	−40	10
70	−60	10
90	−80	10
0	−100	−100

Example 11.5

Consider the situation in Example 11.4 when tile is plugged and not flowing. What is the Φ_t profile?

Solution
This will be a situation of steady state condition and Φ_p at 100 cm depth will be 110 giving a total water potential of 10 cm at all depths above the drain line.

Example 11.6

Consider Example 11.1 when there is no water ponded on the surface, and the drain is not flowing but a free water table exists at 100 cm depth. Calculate the Φ_t at all depths above the drain line.

Solution
Because drain is not flowing, therefore, Φ_t must be constant (same) at all points. This is based on the assumption that there is no soil evaporation. Under these conditions, components of Φ_t are as follows (all units are in cm).

Φ_z	Φ_m	Φ_p	Φ_t
0	−100	0	−100
−20	−80	0	−100
−40	−60	0	−100
−60	−40	0	−100
−80	−20	0	−100
−100	0	0	−100

−120	0	+20	−100
−140	0	+40	−100

Example 11.7

Assume a soil with water table at 100 cm depth and soil surface evaporating at a constant rate. Tensiometers are installed in the soil to measure Φ_m as shown in the Table below. Components of Φ_t are shown in the Table.

Φ_z	Φ_m	Φ_p	Φ_t
0	+800	0	+800
−20	−600	0	−620
−40	−400	0	−440
−60	0	0	−60
−80	0	20	−60
−100	0	40	−60

11.5 MEASUREMENT OF SOIL'S MATRIC POTENTIAL (Φ_m)

Techniques for measurement of Φ_m are outlined in Table 11.2, and described at length by Mullins (2001), Livingston (1993), Young and Sisson (2002), Andraski and Scanlon (2002), and Scanlon et al. (2002). *Tensiometers* are the most widely used for a low range of Φ_m from 0 to −80 KPa, and are relatively simple, inexpensive, easy to install, and have a sensitivity of about 0.1 KPa. Major limitations of tensiometers include the following: (i) insensitivity to soil solution osmotic potential rendering them unsuitable for measuring Φ_m in salt-affected soils, (ii) restricted measurement range of 0 to −80 KPa, (iii) long response time, (iv) poor soil contact in gravelly soils, (v) increase in Φ_m due to movement of water from cup into the adjacent soil as influenced by the soil's and cup's hydraulic conductivities, and (vi) the maximum limit of 4m depth to which a

TABLE 11.2 Techniques for Measurement of Soil Matric Potential

Technique	Principle	Range (kPa)	Limitations	Reference
Tensiometers	Measurement of vacuum created in the tensiometer tube due to absorption of water by the dry soil from porous cup.	0 to −85	Low Range Long response time Air entry due to poor contact	Klute and Gardner (1962)
Psychrometer	Monitoring relative	−80 to	Extremely	Rawlins and

	humidity of vapor in equilibrium with the liquid phase in soil.	−1500	sensitive to temperature	Campbell (1986)
Porous material sensors (filter paper, gypsum blocks)	Evaluating changes in matric potential with change in water content of a porous material.	−1 to −10^5	Hysteresis of the material Calibration of all material	Fawcette and Collis-George (1967); Hamblin (1993); Scholl (1978); Pereira (1951)
Heat dissipation in porous blocks	Assessing the rate of heat dissipation in a porous material 0 to −100 sensor.	0 to −100		Phene et al. (1971)

tensiometer can be inserted. The absolute pressure *(P)* inside the tensiometer is given by Eq. (11.13):

$$P = A - \Phi_m - h$$

(11.13)

where *A* is the atmospheric pressure and *h* is height above the tensiometer cup (Livingston, 1993). If a tensiometer cup is installed 3 m below the ground and the vacuum gauge is about 0.5 m above the soil surface (assuming that *A* is 10 m), the lowest pressure in the system will be about −0.0065 MPa. This limit is reduced for deeper installation.

Psychrometers compliment tensiometers with an upper limit of Φ_m of about −100 Kpa (Campbell and Gardner, 1971; Andraski and Scanlon, 2002). The total water potential is determined by measurement of relative vapor pressure of air in equilibrium with soil pores [Eq. (11.14)].

$$\Phi_m = \frac{\rho_w RT}{V_w \ \ln(p/p_o)}$$

(11.14)

where Φ_m is matric potential in MPa, *R* is the universal gas constant (8.314×10^{-6}MJ/mol/K), *T* is the absolute temperature (K), V_w is molar volume of water (1.8×10^{-5}m^3/mole), and p/p_o is relative humidity expressed as a fraction.

There are two types of psychrometers: (i) those that can be used for in situ measurements and are placed in the soil, and (ii) those in which soil samples are placed in the sample chamber and the Φ_m is determined after about 15 minutes of equilibrium time. The former, a soil psychrometer, consists of a small ceramic cup (1 cm in diameter and 1 cm long) that contains a single thermocouple (50–100 nm in diameter) constructed of chromal and constantan wires (Fig. 11.4). The reference junction usually comprises a Cu wire. The porous ceramic cup facilitates diffusion of water vapors from soil air to the thermocouple. Accurate measurements of air and soil temperatures are critical to psychrometric evaluations. A psychrometer measures the thermal electromotive force from the cooling of the junction in an enclosed space. The force is measured in microvolts (μv) and related to Φ_m. There are two principal limitations of the psychrometric technique. One, the relative humidity of the soil air changes only slightly from 94 to 100%. Two, differences in soil temperature can lead to large errors. A

difference in temperature of 1°C can lead to differences in Φ_m by 10 MPa (Campbell, 1979).

There are several miscellaneous methods of measuring Φ_m. Scanlon et al. (2002) describe seven different techniques based on heat dissipation sensors, electrical resistance sensors, frequency domain and time domain sensors, electrooptical methods, filter paper method, dew
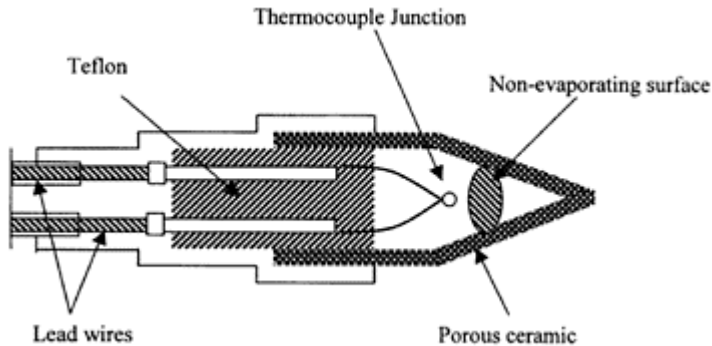

FIGURE 11.4 Soil psychrometer for measuring soil water potential in situ. (Redrawn from Campbell and Gardner, 1971.)

point potentiometer, and vapor equilibration method. A commonly used method is that of measuring electrical resistance. *Resistance blocks* for measuring Φ_m are similar to those described for soil-water measurements and are comprised of porous material such as gypsum, nylon, or fiberglass. Blocks can be used to measure Φ_m in soils drier than −50 KPa. These devices are simple, inexpensive, and provide nondestructive and continuous measurement of Φ_m. The electrical conductivity of porous blocks is zero for dry soil and increases with increase in Φ_m. Porous blocks have several limitations including: (i) unusable in salt-affected soils or those irrigated with saline water, (ii) change in calibration for each block over time, (iii) hysteresis of the porous material, (iv) long response time, (v) degradation of blocks over time, (vi) impact of variations in temperature, (vii) non-suitability of blocks in soils that develop large cracks, and (viii) the error may be large of the magnitude of ±100 to 500 KPa. Cracks are often formed in the vicinity of blocks rendering soils to dry out rapidly after the crack develops or wet quickly following rain or irrigation due to water flowing into the cracks.

The *filter paper method* uses a special type of porous material. This technique is described at length by Al-Khafaf and Hanks (1974), Hamblin (1981), and Greacen et al. (1987). The filter paper, of known porosity and soil moisture characteristic curve, is wrapped around a wedge and pushed into the soil at a desired depth. The filter paper takes 4 to 6 days to equilibrate with the soil following which it is removed and weighed to determine its moisture content. Soil matric potential is determined from the precalibrated soil moisture characteristic curve or the potential vs. θ relationship. It is a

simple and inexpensive method of measuring Φ_m within the range of -50 to -100 KPa. Because of the long equilibration time, however, it is useful only for soils with slow changes in Φ_m.

Similar to the filter paper method, the *heat dissipation technique* also involves a porous medium. The technique is based on measuring the heat dissipation within the porous material in which is located a heat sensor. The dissipation of short heat pulse applied to the sensor depends on thermal diffusivity or its moisture content. This technique is not sensitive to salt content, and therefore, can be used for salt affected soils. Theory, design, and construction of heat dissipation devices are given by Phene et al. (1971). These devices have a measuring range of 0 to -600 KPa with an accuracy of ±10 KPa in the low range (0 to -300 KPa) and of ±100 KPa in the high range (-300 to -600 KPa). However, the accuracy is influenced by hysteresis and contamination of the porous material. These devices are useful for scheduling irrigation (Phene and Beale 1976).

11.6 UNITS OF MEASUREMENT OF SOIL-MOISTURE POTENTIAL

All units of soil-moisture potential are defined with regards to the unit quantity of water. The specific unit depends on the way the unit quantity of water is defined; volume basis, mass basis, or weight basis.

Relationships among different ways to express soil-moisture potential are shown in Table 11.3. A common unit to express soil-moisture potential on volume basis is a "bar." Numerous ways to express one bar are listed in Table 11.4. Similarly, a common unit to express soil-moisture potential on weight basis is pF. The latter is computed as a logarithm to the base 10 of

TABLE 11.3 Units for Expressing Soil-Water Potential

Basis	Units
Volume $(P_v=P=\rho gh)^a$	dynes/cm^2, Pa, ergs/m^2, bar, J/m^3, N/m^2
Mass $(P_m=P/\rho=gh)$	ergs/g; J/Kg
Weight $(P_w=P/\rho g=h)$	cm, m

[a]Soil water potential on volume basis (P_v) is the work done against pressure P to transfer volume V is PV/V or $P=\rho gh$.
1 Dyne=1 g cm/s^2=10^{-5}N
1 N=1 Kgm/s^2
1 Pa=1N/m^2 (1kPa=1/Jkg^1)
1 J=1Nm=10^7 ergs=watts
1 Bar=10^5 Pa=0.987 atmosphere=29.53″ Hg=10^6 dynes/cm^2
1 Atmosphere=1,013,250 dynes cm^{-2} 101,325 N/m^2
1 Torr=1 mm Hg=1/760 atmosphere=1013,250/760 dynes/cm^2=133.22 microbars
1 Watt=J/s=107 erg/s
1 erg=1 dyne cm=g/cm s^2=10^{-4} J/kg