CHAPTER 11 — MEASUREMENT OF SOIL MOISTURE

CHAPTER 11

MEASUREMENT OF SOIL MOISTURE

11.1 General

One of the most significant factors influencing crop yield and watershed performance is the amount of water stored in the soil mantle. This soil moisture information is essential for determining irrigation schedules, for the evaluation of water and solute fluxes, and for partitioning of net solar radiation into latent and sensible heat components.

Soil moisture information is essential in hydrology for input to a range of hydrological models that determine catchment runoff. Also, in numerical models of the atmosphere, the modelling of interactions at the land/atmosphere interface requires, among other water-cycle variables, determination of soil moisture. Surface-based direct measurements of soil moisture are also required for verification of remotely-sensed estimates, e.g. from satellites.

Soil moisture determinations are typically characterized by measuring either the soil-water content or the soil-water potential. Soil-water content is an expression of the mass or volume of water in the soil while the soil-water potential is an expression of the soil-water energy status.

Determination of soil moisture is of great concern to a number of agricultural disciplines. To satisfy the widespread need of determining soil moisture status, a number of commercially-available instruments have been developed. The most commonly used instruments will be discussed, together with their advantages and disadvantages. In addition, a few state-of-the-art instruments and techniques will be briefly mentioned, as they may be widely used in the near future.

11.1.1 *Definitions*

Soil-water content

The simplest and most widely used method for measuring soil-water content is the gravimetric technique. Because this method is easy and based on direct measurements, it is the standard with which all other procedures are compared. Gravimetric soil moisture, θ_g, is typically determined on a dry mass basis and is expressed as:

$$\theta_g = \frac{M_{\text{water}}}{M_{\text{soil}}} \cdot 100 \qquad (11.1)$$

where M_{water} is the mass of the water in the soil sample and M_{soil} is the mass of oven-dry (100–110°C) soil that was contained in the sample. Gravimetric soil-water measurements of air-dry (25°C) mineral soil are typically less than 2 per cent, but as the soil approaches saturation, the water content may increase to values between 25 and 60 per cent. Unfortunately, gravimetric sampling is destructive, making it difficult to obtain an accurate measurement of soil-water content as soils approach saturation.

In many cases, soil moisture is expressed volumetrically. Since precipitation, evapotranspiration and solute transport variables are commonly expressed in terms of flux, volumetric expressions for water content may be more useful. The volumetric water content, θ_v, is expressed as:

$$\theta_v = \frac{V_{\text{water}}}{V_{\text{soil}}} \cdot 100 \qquad (11.2)$$

where V_{water} is the volume of water, and V_{soil} is the total volume of soil (soil + air + water). Volumetric soil-water content may range from less than 10 per cent for air-dry soil to between 40 and 50 per cent for mineral soils approaching saturation. Because of the difficulties of accurately measuring water and soil volumes, volumetric water contents are not usually determined directly.

Volumetric and gravimetric soil-water content are related. The relationship between gravimetric and volumetric water contents can be expressed as:

$$\theta_v = \theta_g \rho_b / \rho_w \qquad (11.3)$$

where ρ_b is the dry-soil bulk density and ρ_w is the soil-water density.

Soil-water potential

The soil-water potential is an important property which describes the energy status of the soil water and is critical to water transport analysis, water storage estimates, and soil-plant-water relationships. A difference in water potential between two soil locations indicates a tendency for water flow, from high to low potential. Since the water potential is reduced by drying (becoming more negative), the work that must be done to remove it increases, making water extraction by plants more difficult. As the plant water uptake becomes more difficult, the water potential in the plant drops, eventually resulting in plant stress and, in extreme cases, severe wilting.

Classically, the water potential is a measurement which describes the ability of soil water to perform work, or in the case of negative potential, the work required to remove the water from the soil. The total water potential, ψ_t (the combined effect of all force fields) can be expressed as:

$$\psi_t = \psi_z + \psi_m + \psi_o + \psi_p \qquad (11.4)$$

where z, m, o, and p denote the gravitational, matrix, osmotic, and pressure potentials, respectively (Nielsen, *et al.*, 1971). Not all of these separate potentials act the same way, and their separate gradients may not always be effective in inducing flow. For example, ψ_o requires a

semi-permeable membrane to induce flow, and ψ_p will exist in saturated or ponded conditions. As most practical applications are in unsaturated soil and do not involve a semi-permeable membrane, the total water potential is typically expressed as:

$$\psi_t = \psi_m + \psi_z \qquad (11.5)$$

11.1.2 *Units and scales*

Water contents are frequently assumed to be dimensionless, as they are typically used as percentages. However, in solving the mass balance or continuity equations for water it must be remembered that the water content is not dimensionless. Gravimetric water content is an expression of the grams of soil water contained in a gram of soil (g water/g dry soil). Likewise, volumetric water content is an expression of the volume of water contained in a specific volume of dry soil (cm^3 water/cm^3 dry soil).

The unit for expressing the water potential is typically the kilopascal (kPa), which is numerically equivalent to J kg^{-1}. Units commonly encountered in older literature are: bars, atmospheres, dynes per centimetre squared, ergs per gram, centimetres of water, centimetres of mercury, and pounds per square inch.

11.1.3 *Meteorological requirements*

The accuracy needed in water-content determinations is a function of the problem being resolved. At present, acceptable levels of accuracy range from 1 to 5 per cent soil moisture on a volume basis. The degree of acceptable accuracy for a problem depends on the scale of observation, soil texture, soil structure, frequency of observation, and the intended use.

The spatial and temporal resolutions required for soil moisture measurements also vary with application. Hydrologists calibrate catchment-scale runoff models by using data that often include soil moisture estimation. At the other extreme, atmospheric general circulation models require measurements on a continental scale to incorporate hydrological processes. Atmospheric general circulation models have surface property resolutions of the order of tens or hundreds of kilometres. Components of the hydrological cycle on a catchment or basin scale (of the order of tens of kilometres) are also required for atmospheric modelling. Available soil moisture, which effectively controls evapotranspiration, is measured only to a shallow depth (tens of centimetres) by remote sensing techniques, with a horizontal resolution of tens of kilometres (see Chapter 8, Part II).

In agriculture, both the quantity of water in the soil and its energy status are important. For hydrological and water-balance needs, as well as for effects upon soil properties (such as soil mechanical strength, thermal conductivity and diffusivity), it is the evaluation of the soil-water content that is most useful. For plant physiology and hydraulic problems connected with water movement, the matrix potential is the most critical measurement.

Most water-balance methods calculate the total plant-water content available in the soil profile, thereby neglecting the energy aspects of the availability and movement of water to plants. Soil water available to plants is commonly considered to be between field capacity and the permanent wilting point for the soil. Field capacity is usually defined as the amount of water remaining after free drainage has ceased (generally 24–48 hours after a precipitation event). Soil texture (clay type and content), soil structure, and organic matter content all affect the amount of water retained at field capacity. By definition, permanent wilting conditions occur when plant leaves are incapable of regaining turgor when placed in a water saturated atmosphere. In general, permanent wilting point is assumed to occur at a matrix potential of –1.5 megapascals. As a result, the permanent wilting point for sandy soil occurs at a θ_g of 3 per cent while for clay soil the same potential would result in θ_g closer to 30 per cent.

11.1.4 *Methods of measurement*

There are many instruments available to evaluate soil-water status. Soil θ_g is typically determined directly. Soil θ_v is usually determined indirectly by measurements of a soil property or by reaction from an object placed in the soil.

Indirect methods of determining soil moisture involve inference of θ_v from measurements of a soil property or reaction from an object placed in the soil which is affected by water content. Common indirect methods for determining water content include radiological techniques, time-domain reflectometry, and nuclear magnetic resonance. Indirect methods of measuring water potential include tensiometers, electrical resistance blocks, and soil psychrometers.

Regardless of the method used, it is impossible to represent a field-scale water content without acknowledging variability due to spatial heterogeneity. Although soil tends toward an equal soil-water potential, this does not relate well to equal water content. None the less, typical ranges of variability expressed as the coefficient of variation (standard deviation/mean) are 15 to 35 per cent. As water content appears to have a limited range of spatial dependence, water-content variability will decrease with a decreasing scale of observation. Fortunately, many field-scale processes can generate acceptable results with a 5 per cent level of accuracy in water contents. Field-scale estimates of evapotranspiration are one area where the latter point is valid (Doorenbos and Pruitt, 1976). However, the sensitivity of any evapotranspiration estimate is also a function of the model being utilized.

Estimates of chemical transit times are unfortunately strongly influenced by a number of localized soil properties and characteristics. A a result, a 5 per cent level of accuracy for water content estimates is not sufficient. In fact, even at the 3 per cent level of accuracy, unacceptable estimates of water content may be generated since only a

fraction of the soil pore space may be used in chemical transport (preferential or macropore flow). Instead, it may be necessary to formulate a distribution of water content; this distribution being characterized by its first moment (mean), the second moment (variance), and the type of distribution (normal, lognormal, etc.). The distribution of the water contents can, then, be used to calculate a distribution of pore water velocities, the pore water velocity being proportional to the input flux and inversely proportional to the water velocity. The distribution of water velocities can, then, be used to form a probability density function for a transfer-function model simulation (Jury and Roth, 1990).

Lysimeters (evapotranspirometers) are relevant to soil moisture determination. The lysimeter method is a non-destructive direct method in which a container filled with soil is weighed either occasionally or continuously to indicate changes in total mass in the container, which may be in part or totally due to changes in soil moisture. Lysimeters are discussed in more detail in Chapter 10, Part I.

11.2 Soil water content: direct measurement

In order to determine θ_g, soil samples are removed from the field with the most convenient tool. Typical tools include shovels, spiral hand augers, bucket augers, as well as power-driven coring tubes. The soil samples are then placed in a leak-proof, tare-weighed container suitable for transporting to a laboratory and drying in an electrically heated oven. The samples and container are weighed in the laboratory both before and after drying, the difference being the mass of water originally in the sample. The drying procedure consists in placing the open container in an electric oven at 105°C until the mass stabilizes at a constant value. The time required varies from 16 to 24 hours. However, if the soil samples contain considerable amounts of organic matter, excessive oxidation may occur and some of the organic matter will be lost from the sample. Although the specific temperature at which excessive oxidation occurs is difficult to specify, lowering the oven temperature from 105 to 70°C seems to be sufficiently low to avoid significant loss of organic matter.

Microwave oven drying for the determination of gravimetric water contents can also be used effectively (Gee and Dodson, 1981). In this method, soil water temperature is quickly raised to boiling point where it remains constant for a period of time due to the consumption of heat in vaporizing water. However, the temperature rapidly rises as soon as the energy absorbed by the soil water exceeds that consumed for vaporizing the water. Caution should be used with this method as temperatures can become so high that they can melt plastic containers if stones are present in the soil sample.

Although rarely used, there are other methods for the direct measurement of soil-water content. However, they are limited to special purposes and emergencies. One of these methods involves placing the soil in a tared container with a perforated bottom and weighing it to determine the wet mass. The soil samples are irrigated with methanol, which will eventually displace the water. The methanol is then ignited, and the procedure is repeated at least one more time. The sample is then again weighed to determine its dry mass. The amount of methanol needed to displace the water depends on a number of factors, such as the size of the sample, its water content, and its texture. The latter method is very susceptible to error as volatile soil components may be lost.

11.3 Soil water content: indirect methods

The capacity of soil to retain water is, among other variables, a function of soil texture and structure. In removing a soil sample, the soil being evaluated will be disturbed, and its water-holding capacity altered. Indirect methods of measuring soil water are beneficial as they allow information to be collected at the same location for each observation without disturbing the soil-water system.

11.3.1 *Radiological methods*

Two general radiological methods are widely used and available for measuring soil-water content. One is the neutron scatter method, which is based on the interaction of high-energy (fast) neutrons and the nuclei of hydrogen atoms in the soil. The other method utilizes the attenuation of gamma rays as they pass through soil. Both instruments use portable equipment for taking measurements at permanent observation sites and require careful calibration, preferably with the soil in which the equipment is to be used.

When using any radioactive emitting device, some precautions are necessary. All rules regarding radiation hazard laid down by the manufacturers and health authorities must be observed. When the guidelines and regulations are followed, there is no need to fear exposure to excessive radiation levels, regardless of the frequency of use. None the less, whatever the type of radioactive emitting device is used, the operator should wear some type of film badge that will enable the exposure levels to be evaluated and recorded on a monthly basis.

11.3.1.1 Neutron attenuation

There are two types of neutron soil moisture detecting device a soil surface meter and a depth probe. In both devices, high-energy (fast) neutrons are emitted and are eventually slowed down upon their interaction with matter (resulting in neutron thermalization) (Visvalingam and Tandy, 1972). The hydrogen nuclei, having about the same mass as neutrons, are by far the most effective soil components in slowing down neutrons upon collision. As a result, the density of slow neutrons in the vicinity of the neutron probe is nearly proportional to the volumetric soil-water content. The slow or thermalized neutrons form a cloud around the neutron-emitting device where its density and size

represent an equilibrium between the emission rate of fast neutrons and those thermalized. Within each neutron-emitting device is a thermalized neutron detector which determines the density of the thermalized neutron cloud. Unfortunately, the volume encompassed by the thermalized neutron cloud varies substantially with water content. For example, in wet soil, the radius of influence may be only 15 cm, while in dry soil, the radius may increase to 35 cm. Because the volume being measured varies with water content, this method lacks high resolution, making it impossible to localize water-content discontinuities. A particular problem occurs at the soil interface on account of the soil-air discontinuity. As a result, the neutron probe is not used in the top 18 cm of soil. However, the neutron surface meter is used exclusively for measuring water contents in the soil surface (0–30 cm). Unfortunately, where the soil surface is rough, precision falls off dramatically.

A neutron depth probe comprises a radioactive source of high-energy neutrons, and a detector of slow thermalized neutrons, typically in a cylindrical form. The probe is attached by cable to the main electronics so that the probe can be lowered into a previously installed access tube. Although several arrangements of source-detector are possible, it is best to have a probe with a double detector and a central source. This arrangement allows for a more spherical zone of influence and leads to a more linear response with soil-water content. The neutron surface meter usually has a thermalized neutron detector laid horizontally on the soil surface with a fast neutron source behind it.

The access tube should be seamless and thick enough (typically 1.25 mm) to be rigid, but not so rigid that the access tube itself is responsible for thermalizing neutrons. The access tube must be made of a non-corrosive material, such as stainless steel, aluminium, or some plastics, but polyvinylchloride should be avoided as it absorbs slow neutrons. The probe should be capable of being inserted into the tube without risk of jamming; usually a 4-cm diameter tube is sufficient. Care should be taken in installing the access tube to ensure that it is not bent.

Additionally, no air voids should exist between the access tube and the soil matrix. Approximately 15 cm of the tube should extend beyond the soil surface as the box containing the electronics fits on top of the access tube. All access tubes should be fitted with a removable cap to keep rainwater from entering the tubes.

In order to enhance experimental reproducibility, the soil-water content is not compared directly with the number of slow neutrons detected, but rather with a count ratio (*CR*). The count ratio is given by:

$$CR = \frac{C_{\text{soil media}}}{C_{\text{background}}} \qquad (11.6)$$

where $C_{\text{soil media}}$ is the count of thermalized neutrons detected in the soil, and $C_{\text{background}}$ is the detected thermalized neutrons from a reference platform. All neutron probe instruments now come with a reference platform on which these background counts can be made. Typically, the platform is part of the shipping box. The instrument is placed on the platform and a series of 10 short readings is taken. Although the duration of the readings is up to the investigator, typical observation times range from 30 seconds to one minute. The distribution of these readings should be normal, i.e., three out of 10 readings should be beyond the mean plus or minus one standard deviation. The mean of these 10 counts is then recorded and used as $C_{\text{background}}$ while $C_{\text{soil media}}$ is determined from averaging several soil readings at a particular depth/location. For calibration purposes, it is best to take three samples around the access tube and to average the water contents corresponding to the average *CR* calculated for that depth. A minimum of five different water contents should be evaluated for each depth. Although some calibration curves may be similar, a separate calibration for each depth increment should be conducted. Typical coefficients of determination (r^2) with a new probe/meter should be in the range of 0.90 to 0.99.

11.3.1.2 Gamma absorption

Whereas the neutron-attenuation method measures the volumetric water content in a large sphere, gamma absorption scans a 1-cm layer. Although it has a high degree of resolution, the small soil volume evaluated will exhibit more spatial variation due to soil heterogeneities (Gardner and Calissendorff, 1967). The single-probe gamma device measures attenuation by reflection and is no longer widely used. However, the dual-probe gamma device which measures both soil density and water content is still a widely accepted instrument. Gamma attenuation can be mathematically expressed as:

$$I = I_o e^{-\mu x \rho} \qquad (11.7)$$

where I is the measured intensity of the gamma beam, I_o is the intensity of the non-attenuated gamma beam, μ is the mass absorption coefficient for the absorbing material, x is the thickness of the absorbing material, and ρ is the density of the absorber.

Changes in gamma attenuation for a given mass absorption coefficient and absorber thickness can be related to changes in total density. As the attenuation of gamma rays is due to mass, it is not possible to determine water content unless the attenuation of gamma rays responding to dry soil density is known. Additionally, the dry density of the soil must remain unchanged with changing water content. If the dry soil density is known, then the soil-water content can be determined from the difference between the total and dry density values.

Unlike neutron attenuation, gamma-ray attenuation enables a high spatial resolution. Vertical measurements at 2.5 cm can be made with excellent precision. It also has the advantage of making accurate measurements 2.5 cm below the air-surface interface.

Additional caution should be taken with the use of gamma-emitting devices as they are potentially more dangerous than the neutron-emitting devices. The manufacturer will provide a shield which should be used at all times. The only time the probe leaves the shield is when it is lowered into the access tube.

11.3.2 *Soil-water dielectrics*

Because of the dramatic difference in the dielectric constants of water and dry soil (approximately 80 and 3.5, respectively), theoretical and empirical relationships relating soil volumetric water content to the dielectric constant of the soil-water system have been proposed. This approach allows reliable, fast, non-destructive measurements of the volumetric water content, without the potential hazard associated with radioactive emitting devices. In addition, these methods lend themselves to being fully automated for large-scale data-acquisition programs. At present, two newly developed instruments which evaluate soil-water dielectrics are commercially available and are being used on an international scale. The first instrument utilizes time-domain reflectometry (TDR) technology, while the other measures the dielectric constant at a specific microwave frequency.

11.3.2.1 TIME-DOMAIN REFLECTOMETRY

Time-domain reflectometry is a relatively new method which determines the dielectric constant of the soil by measuring the transmittal time of an electromagnetic pulse launched along a pair of parallel rods of known length embedded in the soil. As the sampling area is essentially a cylinder around the parallel probes, a large soil volume is examined. Theoretically, the dielectric constant is sensitive to soil surface area; however, time-domain reflectometry does not appear to be sensitive enough to require calibration for the range in surface areas typically found in soils. The most widely accepted dielectric response to soil-water content was proposed by Topp, Davis and Annan (1982) and is expressed as:

$$\theta_v = -0.053 + 0.029\varepsilon - 5.5 \cdot 10^{-4}\varepsilon^2 + 4.3 \cdot 10^{-6}\varepsilon^3 \quad (11.8)$$

where ε is the soil-water dielectric constant. This empirical relationship has been confirmed by other investigators and appears to be roughly independent of texture and gravel content (Drungil, Abt and Gish, 1989).

Generally, the parallel probes are separated by 5 cm and can vary in length from a few to over 30 cm. Additionally, the rods making the probe can be of any metallic substance; stainless steel is most frequently used. Although some care should be taken to ensure that the probes are parallel, slight deviations do not affect the resultant dielectric readings.

In theory, the attenuated time-domain reflectometry signal should be able to measure both soil-water content and the salinity independently from a single reading; however, this work is still in its infancy. Additional work is being evaluated which allows this technique to be automated by examining the water content from a buried set of probes, each placed horizontally at a different depth. The probes are then linked through a multiplexing device attached to a field data logger.

11.3.2.2 MICROWAVE PROBE

The microwave dielectric probe utilizes an open-ended coaxial cable and a single reflectometer at the probe tip to measure amplitude and phase at a particular frequency (typically in the microwave region). Soil measurements are referenced to air, and typically calibrated with dielectric blocks and/or liquids of known dielectric properties. One advantage of using the liquids for calibration is that a perfect electrical contact between the probe tip and the material can be maintained (Jackson, 1990).

As a single, small probe tip is used, only a small volume of soil is ever evaluated. As a result, this method is excellent for laboratory or point measurements but is likely to be subject to spatial variability problems if used on a field scale. Additionally, the probe evaluates a small soil volume; therefore, soil contact is critical.

11.4 Emerging technologies

Due to recent engineering advances, new methods are being developed which allow for the rapid measurement of soil moisture conditions. Two recent developments in soil moisture measurements are the use of pulsed nuclear magnetic resonance and microwave remote sensing.

11.4.1 *Pulsed nuclear magnetic resonance (PNMR)*

Still in the research and development stage, the use of PNMR may have practical application in the near future (Paetzold, Gish and Jackson, 1987). This measurement approach focuses on the interaction between hydrogen nuclear magnetic moments and a magnetic field. The sensor unit consists of an electromagnetic, radio-frequency coil, and a tuning capacitor. Essentially, this method allows for the instantaneous measurement of the volumetric water content in soil — independent of texture — organic matter content, and soil density.

The magnetic moment of a nucleus which contains an odd number of protons/neutrons behaves like a spinning bar magnet. When placed in a static magnetic field, the magnetic moment precesses about an axis parallel to the applied magnetic field. If an oscillating magnetic field equal to the precession frequency of a hydrogen atom is applied at right angles to the static magnetic field, it will force the magnetic moments of hydrogen to precess in phase. The oscillating magnetic field is produced by the radio-frequency generator. The amount of energy adsorbed by the sample can, then, be measured, as well as the decay signal of the oscillating field. The analysis of the resultant adsorption and decay signals yields information concerning the spin-spin and spin-lattice relaxation times which, in turn, are used to calculate the amount of hydrogen in the sample.

A tractor fitted with a prototype PNMR device has already been built and tested. This device could be used to determine soil-water content at the time of planting, or could be used to collect ground data for calibrating remote sensing instruments. Although the tractor PNMR system can accurately evaluate approximately 5 cm of surface soil moisture, precision drops off dramatically with increasing depth. The magnetic field must be homogeneous for PNMR techniques to work effectively, and obtaining a homogeneous magnetic field in undisturbed soil is the greatest limitation of this technique.

Laboratory PNMR instruments can be purchased but they are generally too expensive for practical applications.

11.4.2 *Remote sensing*

Measurements from space-borne instruments utilizing remote sensing techniques will be available in the near future for evaluating soil-water content, estimation of evapotranspiration rates, and evaluation of plant stress on a watershed scale (Jackson and Schmugge, 1989). Although infrared and microwave energy levels have been widely studied, only the microwave region has the potential for obtaining direct quantitative soil moisture measurements from a space platform.

Microwave techniques can be separated into passive (radiometric) and active (radon) radiation. Passive microwave techniques focus on analysing the natural microwave emissions from the Earth's surface, while active radiation refers to measuring the attenuation of a radar backscattering signal. Both approaches are based on the large differences that exist between the dielectric properties of liquid water and dry soil, and both are conducive to monitoring surface soil-water content over large areas of land.

The microwave radiometer response will range from an emissivity of 0.95 to 0.6 or lower for passive microwave measurements. For the active microwave measurements, an increase of about 10 decibels in return is observed as soil goes from dry to wet. The microwave emission is referred to as the brightness temperature, T_b, and is proportional to the emissivity, β, and the temperature of the soil surface, T_{soil}, or:

$$T_b = \beta T_{soil} \qquad (11.9)$$

where T_{soil} is in kelvin. Because β is dependent on soil texture, surface roughness, and vegetation, actual soil water contents will be empirically related to T_b.

For active microwave measurements of soil water content, the total backscatter signal must be separated into the portion due to vegetation, and that of the soil. Additionally, the vegetation canopy will influence the soil component. The volumetric water content is related to the total active backscatter, S_t, by:

$$\theta_v = L\,(S_t - S_v)\,(RA)^{-1} \qquad (11.10)$$

where L is the vegetation attenuation coefficient, S_v is the backscatter from vegetation, R is a soil surface roughness term, and A is a soil moisture sensitivity term. Unfortunately, there is no suitable independent means of measuring R and A. As a result, the active microwave response to soil-water content can be expressed as an empirical relationship.

11.5 Soil-water potential instrumentation

To date, only instruments capable of measuring the matrix potential are sufficiently inexpensive and reliable to use in a field-scale monitoring programme. In each case, there are severe limitations to the range in water potential in which the instrument functions properly. Care must, therefore, be exercised if osmotic potentials are significant.

11.5.1 *Tensiometers*

The most widely used and least expensive water-potential measuring device is the tensiometer. Tensiometers are simple, generally consisting of a porous ceramic cup and a plastic cylindrical tube connecting the porous cup to a recording device which seals the top of the cylinder. In view of their universal availability and low cost, a detailed description of their construction is unnecessary.

The tensiometer establishes a quasi-equilibrium condition with the soil-water system. The porous ceramic cup acts as a membrane through which water flows, and as such, must remain saturated if it is to function properly. Consequently, all the pores in the ceramic cup and the cylindrical tube are initially filled with de-aerated water. Once in place, the tensiometer will be subject to negative soil-water potentials, causing water to move from the tensiometer into the surrounding soil matrix. The water movement from the tensiometer will create a negative potential or suction in the tensiometer cylinder which will register on the recording device. The recording device can be a pressure transducer (Marthaler, *et al.*, 1983), a Bourdon-type vacuum gauge, or a simple U-tube filled with water and/or mercury. On the other hand, if the soil receives water, the soil-water potential may increase to where water moves from the soil back into the tensiometer, resulting in a less negative water potential reading. This exchange of water between the soil and the tensiometer, as well as the tensiometer's exposure to negative potentials, will cause dissolved gases to be released by the solution, forming air bubbles. The formation of air bubbles will alter the pressure readings in the tensiometer cylinder and will result in faulty readings. Consequently, the cylinders occasionally need to be refilled and de-aired with a hand-held vacuum pump.

Before installation, but after the tensiometer has been filled with water and degassed, the ceramic cup must remain wet. Wrapping the ceramic cup in wet rags or inserting it into a container of water will keep the cup wet during transport from the laboratory to the field. In the field, a hole of the appropriate size and depth is prepared. The hole should be large enough to be snug on

all sides of the cylinder and long enough for the tensiometer to extend several centimetres above the soil surface. Since the ceramic cup must remain in contact with the soil, it is generally beneficial to prepare a thin slurry of mud from the excavated site and to pour it into the hole before inserting the tensiometer. Care should also be taken to ensure that the hole is backfilled properly, thus eliminating any depressions that may lead to ponded conditions adjacent to the tensiometer. The latter precaution will minimize any water movement down the cylinder walls, which would produce unrepresentative soil-water conditions.

Tensiometers can measure only the matrix potential, because solutes can move freely through the porous cup. However, tensiometers can be purchased with additional features such as electrodes, placed either inside the ceramic cup or just above the ceramic chamber, thus allowing electrical conductivity within the tensiometer to be determined simultaneously. Obviously, it may take some time for these tensiometers to equilibrate with the soil environment. Another limitation is that the tensiometer has a practical lower limit of about –80 kPa. Beyond –100 kPa, water will boil at ambient temperature, forming water vapour bubbles which will destroy the vacuum inside the tensiometer cylinder.

The cylinder and recording portion of the tensiometer allow for appreciable changes in volume. Under drought conditions, appreciable amounts of water can move through the tensiometer to the soil. Thus, tensiometers can alter the very condition they were designed to measure. Typically, when the tensiometer acts as an irrigator, so much water is lost through the ceramic cups that a vacuum in the cylinder cannot be maintained, and the tensiometer gauge will be inoperative. Additional support of this process comes from excavated tensiometers which have accumulated large numbers of roots in the proximity of the ceramic cups.

The tensiometer is also sensitive to temperature. Although only a small portion of the tensiometer is exposed to ambient conditions, the interception of solar radiation may induce thermal expansion of the tensiometer cylinder. Additionally, temperature gradients from the soil surface to the ceramic cup may result in thermal expansion or contraction of the cylinder, thus inducing false water-potential readings. To minimize these effects, the tensiometer cylinder should be constructed of non-conducting materials and readings should be taken at the same time every day, preferably in the early morning.

11.5.2 *Resistance blocks*

Electrical resistance blocks, although insensitive to water potentials in the wet range, are excellent companions to the tensiometer. Electrical resistance blocks consist of electrodes encased in some type of porous material that will reach a quasi-equilibrium state with the soil. The most common block materials are gypsum, nylon fabric, and fibreglass (Perrier and Marsh, 1958).

Resistance blocks are relatively inexpensive and are good for field investigations. However, they do need to be calibrated before installation. This is generally accomplished by saturating the blocks in distilled water and then subjecting them to a predetermined pressure in a pressure-plate apparatus. After equilibration at a specific pressure, readings are taken, and the block is exposed to successively greater pressure potentials. This procedure should be repeated for at least five different pressures before installation. As the resistance-block calibration curves change with use, they need to be calibrated both before installation and after each investigation.

Unfortunately, resistance blocks are slow to equilibrate with soil, generating water-potential estimates that are more closely associated with the soil-drying curve. Consequently, this method is subject to errors where soil hysteresis may be an important factor. There is also a problem with shrinking and swelling soil which will break contact with the blocks. In addition, this approach determines water potential as a function of electrical resistance, and is sensitive to soil salinity. If saline conditions do exist, it must be remembered that added salts will decrease resistance, falsely indicating a wetter soil. The gypsum blocks are less sensitive to salts because the electrodes are consistently exposed to a saturated solution of calcium sulphate. However, gypsum blocks tend to deteriorate faster than fibreglass blocks.

When installing the resistance blocks it is best to dig a small trench for the lead wires before preparing the hole for the blocks. This will minimize water movement along the wires to the block, which could result in erroneous readings.

11.5.3 *Psychrometers*

Thermocouple psychrometers do not measure the soil-water potential directly, but measure the vapour phase with which it is in equilibrium (Rawlins, 1972). As a result, psychrometers are quick to equilibrate with the soil environment. As with electrical resistance blocks, this method is not sensitive to wet conditions but is well suited to a dry soil environment. It also lends itself to automated data acquisition.

Under non-saline conditions, the relationship between the matrix potential and the relative humidity can be determined from:

$$\psi_m = (RT/W) \ln (p/p_0) \qquad (11.11)$$

where W is the molecular weight of water (0.018 kg mole^{-1}), R is the ideal gas constant (8.31 J K^{-1}), T is the temperature in kelvin, p is the water-vapour pressure in equilibrium with the liquid phase, and p_0 is the saturated water-vapour pressure of the liquid phase.

Psychrometers consist of a miniature thermocouple placed within a small chamber. The thermocouple is cooled off by the Peltier effect, condensing water on a wire junction. As water evaporates from the junction, its temperature decreases and a current is produced which is measured by a voltmeter. Consequently, these

measurements are quick to respond to changes in soil-water potential, but are very sensitive to temperature and salinity.

At equilibrium, the relative humidity in the soil atmosphere does not change significantly over the growing season. The lowest water potential typically associated with active plant-water uptake is 1 500 kPa, which corresponds to a relative humidity of approximately 98.8 per cent. Thus, the range in relative humidity measured by the psychrometer is between 98 and 100 per cent. However, psychrometers are so temperature-sensitive that if the water potential is to be measured accurately to within 10 kPa, the temperature must be controlled to within 0.001 K. Fortunately, the development of new equipment in recent years has made this degree of precision possible (Brunini and Thurtell, 1982). None the less, diurnal temperature fluctuations can induce temperature gradients in the psychrometer as the components of the instrument differ in heat capacities. To minimize the latter effect, readings should be taken at the same time each day, preferably in the early morning.

11.6 Site selection and sample size

Soil moisture information will be most beneficial if values are collected frequently and if its spatial variability over the study area is assessed. Soil moisture information is reliable only at the point of measurement; therefore, a large number of samples may be required to describe adequately the soil moisture status of the site. However, the location of individual sampling points should be chosen so that the number of points required to reach the desired precision is minimized. Preliminary sampling may be necessary to estimate the number of soil samples needed to obtain a reliable field-scale soil-water content value, and/or to locate areas that need individual characterization. This preliminary sampling will generate a sample variance, s^2, which can be used to calculate the number of samples needed to estimate soil-water content at an observed level of accuracy (L). To calculate the number of samples required to achieve the desired level of accuracy, the sample size, n, can be estimated by:

$$n = 4\,(s^2/L^2) \qquad (11.12)$$

For example, suppose that a preliminary sampling yielded an s^2 of 25 and we wanted our level of accuracy to be within 2 per cent. Equation 11.12 indicates that we would need 25 samples from our site. By using this method, two common sampling schemes can be employed to describe soil-water content, each yielding a different level of confidence.

In the random sampling approach, all locations in the field theoretically have an equal chance of being selected as a sampling site. Unfortunately, this approach assumes that the water-content distribution is normally distributed. For large areas, this may not be a valid assumption, as soil morphological and pedological processes may differ significantly.

A second approach divides the area into strata based on the uniformity of relevant variables within the strata, e.g. similarity of hydrological response, soil texture, soil type, vegetative cover, slope, etc. Then each stratum can be sampled independently and the data recombined by weighing the results for each stratum by its relative area. Fortunately, topography is the most critical factor controlling the distribution of soil water in a small watershed. Therefore, the subdivision of low-sloping watersheds into spatial units of homogeneous response can be based on topography alone. Similarly, sloping rangeland will need to be more intensely sampled than a flat cropland. However, presence of vegetation tends to diminish the soil moisture variations caused by topography.

As direct methods generally require field collection and transport of soil samples to a laboratory for analysis, several precautions should be discussed. Immediately after taking the soil sample, it should be placed in a leak-proof, seamless container. As the samples are frequently placed in a convection oven, the container should be able to withstand high temperatures without melting or losing significant mass. The most common soil containers are aluminium cans, which can be numbered and tare-weighted. If the soil samples are to be transported for a considerable distance, then electrical tape should be used to seal the container to avoid moisture loss through evaporation. Microwave ovens can be used with non-metallic sample containers. In particular, they allow shorter drying times.

11.7 Special consideration

Water contents for stony or gravelly soils can be grossly misleading. When rocks occupy an appreciable volume of the soil, they modify the direct measurement of soil mass, without making a similar contribution to the soil porosity. For example, water content on a mass basis may be 10 per cent for a soil sample having a bulk density of 2.0 g cm^{-3}; however, the water content of the same sample based on the finer soil material (stones and gravel excluded) would be 20 per cent (assuming the density of the fine soil material was 1.62 g cm^{-3}).

Although the gravimetric water content for the finer soil fraction, $\theta_{g\ \text{fines}}$, is the value usually used for spatial and temporal comparison, there may also be a need to determine the volumetric water content for a gravelly soil. The latter value may be important in calculating the volume of water in a root zone. The relationship between the gravimetric water content of the fine soil material and the bulk volumetric water content is given by:

$$\theta_{v\ \text{stony}}\ \theta_{g\ \text{fines}}\ (\rho_b/\rho_w)(1 + M_{\text{stones}}/M_{\text{fines}}) \qquad (11.13)$$

where $\theta_{v\ \text{stony}}$ is the bulk volumetric water content of a soil containing stones, and/or gravel and M_{stones} and M_{fines} are the masses of the stone and fine soil fractions, respectively (American Society of Agronomy, 1976).

References

American Society of Agronomy, 1976: *Methods of Soil Analyis, Part 1, Physical and Mineralogical Methods* (A. Klute, ed.). Americal Society of Agronomy, Madison, Wisconsin, 188 pp.

Brunini, O. and Thurtell, G. W., 1982: An improved thermocouple hygrometer for *in situ* measurements of soil water potential. *Soil Science Society of America Journal*, 46, pp. 900–904.

Doorenbos, J. and Pruitt, W. O., 1976: Crop water requirements. *Irrigation and Drainage*, 24, Food and Agriculture Organization of the United Nations, Rome.

Drungil, C. E. C., Abt, K. and Gish, T. J., 1989: Soil moisture determination in gravelly soils with time domain reflectometry. *Transactions of the American Society of Agricultural Engineers*, 32, pp. 177–180.

Gardner, W. H. and Calissendorff, C., 1967: Gamma ray and neutron attenuation measurements of soil bulk density and soil water content. *Proceedings of the Symposium on Techniques in Soil Physics and Irrigation Studies*, Istanbul. International Atomic Energy Agency, Vienna, pp. 101–113.

Gee, G. W. and Dodson, M. E., 1981: Soil water content by microwave drying: a routine procedure. *Soil Science Society of America Journal*, 45, pp. 1234–1237.

Jackson, T. J., 1990: Laboratory evaluation of a field-portable dielectric/soil moisture probe. *IEEE Transactions on Geoscience and Remote Sensing*, 28, pp. 241–245.

Jackson, T. J. and Schmugge, T. J., 1989: Passive microwave remote sensing system for soil moisture: some supporting research. *IEEE Transactions on Geoscience and Remote Sensing*, 27, pp. 225–235.

Jury, W. A. and Roth, K., 1990: *Transfer Functions and Solute Movements Through Soil.* Birkhauser Verlag AG, Basle, Switzerland, 228 pp.

Marthaler, H. P., *et al.*, 1983: A pressure transducer for field tensiometers. *Soil Science Society of America Journal*, 47, pp. 624–627.

Nielsen, D. R., *et al.*, 1971: *Soil Water*. American Society of Agronomy, Madison, Winconsin, 175 pp.

Paetzold, R. F., Gish, T. J. and Jackson, T. J., 1987: NMR measurements of soil water content. *Proceedings of the International Conference on the Measurement of Soil and Plant Water Status*, Logan, Utah, 6–10 July 1978, 1, pp 255–260.

Perrier, E. R. and Marsh, A. W., 1958: Performance characteristics of various electrical resistance units and gypsum materials. *Soil Science*, 86, pp. 140–147.

Rawlins, S. L., 1972: Theory of thermocouple psychrometers for measuring plant and soil water potentials. In: Brown, R. W. and van Haveren, B. P., *Psychrometry in Water Relations Research.* Utah Agricultural Experiment Station, Utah State University, Logan.

Topp, G. C., Davis, J. L. and Annan, A. P., 1982: Electromagnetic determination of soil water contents using TDR: I. Applications to wetting fronts and steep gradients. *Soil Science Society of America Journal*, 46, pp. 672–678.

Visvalingam, M. and Tandy, J. D., 1972: The neutron method for measuring soil moisture content: a review. *Journal of Soil Science*, 23, pp. 499–511.