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Soil Water and Temperature Status

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Soil water status and temperature are critical factors that affect the activity of organisms in soils. Soil water content also has important effects on the quantity of water that eventually becomes available to surface water, both through its influence on infiltration/runoff relationships (wet soils have lower infiltration rates) and because of its potential contribution to the groundwater. Soil temperature influences soil water availability, the growth rates of plants and microorganisms, and rates of chemical/biochemical reactions.

A number of useful soil physics texts are available that cover the general theory and application of soil water and temperature (e.g., Baver et al. 1972; Hillel 1980a, 1980b; Marshall and Holmes 1988). In addition, several books contain detailed summaries of current measurement techniques (Klute 1986; Smith and Mullins 1991; Topp et al. 1992). These references should be used to supplement material in this chapter.

Soil water content has been measured for many years. Earliest measurements were semiquantitative. To the experienced fieldworker, the appearance and feel of the soil can provide valuable information on the amount of water it contains, especially the total water content. Wet soils are generally darker than dry soils and feel “smoother” and heavier. However, appearance and feel can be deceiving. Furthermore, they do not provide quantitative measures of soil water status. Earliest measures of soil water content involved collecting a moist field sample, weighing it, drying it in air or in an oven, and determining the weight loss. Weight loss on drying at 100–110 °C is usually attributable to water evaporation. The ratio of weight loss to dry weight has been termed *gravimetric soil water content*.

For soils with comparable textures, gravimetric water content corresponds well to the ability of plants to extract water from the soils. However, gravimetric water

Table 3.1. General Lower Limit of Soil Water Potential for Organism Activity

Organism	Soil Ψ matrix, -MPa
Sunflower	1.5
Creosote bush	5.0
Bacteria	2.0
Fungi	3.0
Nematodes	0.15
Protozoa	0.10

Note: For plants, this represents the lower limit for water uptake; for other organisms, processes other than strict survival are limited.

is not a reliable indicator of water availability across soils of differing textures, for example, when comparing clayey and sandy soils. Clays hold relatively high amounts of water even when plants are experiencing drought, while plants can take up water from sandy soils at a very low gravimetric water content.

To deal with water availability to plants and water movement in soils, the concept of *soil water energy potential*—often shortened to *soil water potential*—has been developed (see later discussion). The presence of solutes and the attraction of water molecules to the surfaces of soil particles lower the energy of soil water compared with that of pure water. The plant develops an energy gradient to extract water from the soil. Thus, in order to extract water from a dry, salty soil, the plant itself must become “drier” than the soil, lowering the energy of water in its leaves below that of the soil water. The activity of other soil organisms is also sensitive to soil water potential (Tab. 3.1).

In practical terms, these soil water measurements may be separated into two categories. For ecosystem studies, where knowledge of hydrologic balance is important, measurement of the *total soil water content* is appropriate. For ecophysiological studies, the appropriate measure is the *soil water potential* because it provides information on the instantaneous ability of the plant and microorganisms to extract water from the soil.

When considering the effect of soil water on organism activity at the landscape scale, the availability of water to organisms can be assessed by considering three parameters:

- the total amount of soil water available in the root zone during the growing season, which relates to the total quantity of transpiration that can occur, and therefore total potential net primary production;
- the duration of availability, when temperature and light are adequate for growth, which determines how the phenology of the plant interacts with soil moisture; and
- the depth and spatial location of available water, which determines how plant rooting patterns affect access to water.

Obviously these factors all interact strongly. However, if these attributes of the hydrologic cycle are known, the potential effect of soil water on plant growth can be predicted.

Available Protocols

Water in soils is generally described in terms of either (1) the total amount or (2) the energy status of the soil water, frequently called *water potential*. Decisions about which to measure, and how, are based on the research question of interest.

Water balance measurements depend on accurate assessment of the quantities of water in soil. These quantities are expressed in terms of

- mass water/mass soil (gravimetric water content, usually kg water/kg dry soil), and
- volume of water/volume of soil (volumetric water content, L water/L soil).

To convert from gravimetric to volumetric water content, multiply by the bulk density of the soil (Mg dry soil/m³ dry soil or g dry soil/cm³ dry soil). For example, a soil that contains 0.3 kg water/kg soil, with a dry soil bulk density of 1.4 Mg/m³, has a volumetric water content of 0.42 mL water/cm³ soil. The water content of soil is usually expressed using the symbol θ , with θ_g referring to gravimetric water content and θ_v referring to volumetric water content.

Soil Water Content

A variety of principles have been applied to estimate volumetric water content in field soils. Two of the most common rely on neutron analysis using a neutron probe and analysis of a soil's dielectric constant using capacitance or time domain reflectometry probes.

The neutron probe technique relies on the ability of hydrogen in water to slow fast neutrons to thermal neutrons. The fraction of neutrons slowed is proportional to water content. The probe is lowered into an access tube permanently installed in the soil, and readings are taken at specific depths, for example, 15, 30, 45, and 60 cm. The probe is held at a given depth for a period of time to count the number of returning thermal electrons. This number is proportional to the water content of the soil. Depending on the degree of textural differences in the field, different calibration curves may be required for different sites. Although it has proven to be rugged and safe in the field, use of the neutron probe requires a radiation license and operator certification with respect to the handling and transport of devices with radioisotopes.

The dielectric constant of soils changes as a function of soil water content (e.g., Whalley et al. 1992), and this constant can be measured by a variety of techniques to provide soil water content estimates. Soil capacitance probes, for example (Tab. 3.2), can be lowered into an access tube similar to that for the neutron probe to measure soil capacitance (a function of the dielectric constant of the soil) at different depths.

Other methods have also been developed for rapid, effective measurements of the soil dielectric constant. The most popular of these relatively new methods is termed *time domain reflectometry* (TDR; Dalton 1992). In TDR, two stainless steel rods are inserted parallel to one another to a given soil depth (10–100 cm); an electrical pulse sent through these wave guides generates an electrical response (read

Table 3.2. Suppliers of Time-Domain Reflectometers, Tensiometers, and Resistance Blocks

Method	Supplier	Model	Datalogger	Instrument Price Range	Sensors (each)
Time domain reflectometer	Soilmoisture	Trase		\$9,200	\$15-23
Tensiometer	Campbell Scientific	Water Content Reflectometer	Possible	220	NA
	Soilmoisture	Soilmoisture tensiometer (3')	No	50	NA
Resistance blocks	Irrrometer	Tensiometer Model RS (3')	No	65	NA
	Soilmoisture	Soilmoisture meter	No	285	11
	Irrrometer	Digital readout resistance meter	No	240	24
Thermistor	Onset Computer	Hobo, Stowaway	Built-in	63-329	NA
Laboratory psychrometer	Decagon Devices, Inc.	Decallogger	Built-in	2,500	NA

Notes: List is current as of 1998. Prices listed are for field-capable units without dataloggers, unless otherwise indicated.

NA = not applicable because sensors are integral to instrument.

Soilmoisture Equipment, Inc., Goleta, CA (www.soilmoisture.com)

Campbell Scientific, Inc., Logan, UT (www.campbellsci.com)

Irrrometer, Inc., Riverside, CA (www.irrometer.com)

Decagon Devices, Inc., Pullman, WA (www.decagon.com)

Onset Computer Corporation, Pocasset, MA (www.isa.org)

with an oscilloscope) that is characteristic of the dielectric constant for a soil at a particular water content. Probes can also be inserted laterally to provide water contents for a specific soil depth or horizon. The major drawback of TDR at this time is the high initial cost of equipment. However, its convenience makes the method extremely attractive, and further refinements are likely to lower the price.

The dielectric constant of the soil is generally proportional to the water content. Under some conditions, soil salinity will affect the signal, but instruments that are currently available minimize this effect. In fact, some instruments have turned this "interference" into a benefit and have been specifically designed to measure salinity (Dalton 1992).

Soil Water Potential

The energy status of water in soils is expressed as the energy embodied in water molecules per unit volume of water, compared with pure water at standard temperature and pressure. This is described in terms of the following equation (Nobel 1991):

$$\mu_w = \mu_w^* + V_w \times \Pi + V_w \times P + m_w \times g \times h$$

where

μ_w = energy potential of water (joules/mole)

μ_w^* = standard energy potential of water (reference is pure water)

V_w = partial molar volume of water (0.018 L/mole)

Π = total osmotic pressure (Pa)

P = hydrostatic pressure (Pa)

m_w = molar mass of water (1000 kg/m³)

g = acceleration of gravity (kg m/sec²)

h = height relative to reference (m)

This equation expresses the energy embodied in water, quantifying the effects of solutes, solid surfaces, and gravity.

The net energy potential of the water is usually expressed in terms of the difference between μ_w and μ_w^* , namely

$$\mu_w - \mu_w^* = V_w \times \Pi + V_w \times P + m_w \times g \times h$$

Further, when both sides of the equation are divided by the molar volume of water, we have

$$\mu_w - \mu_w^*/V_w = \Pi + P + (m_w \times g \times h)/V_w$$

The term P can represent the pressure relative to the location of the point in question with respect to the gravitational head. The combined term $m_w g h$ refers to the height of capillary rise in a tube of radius r , where r represents the radius of the small-

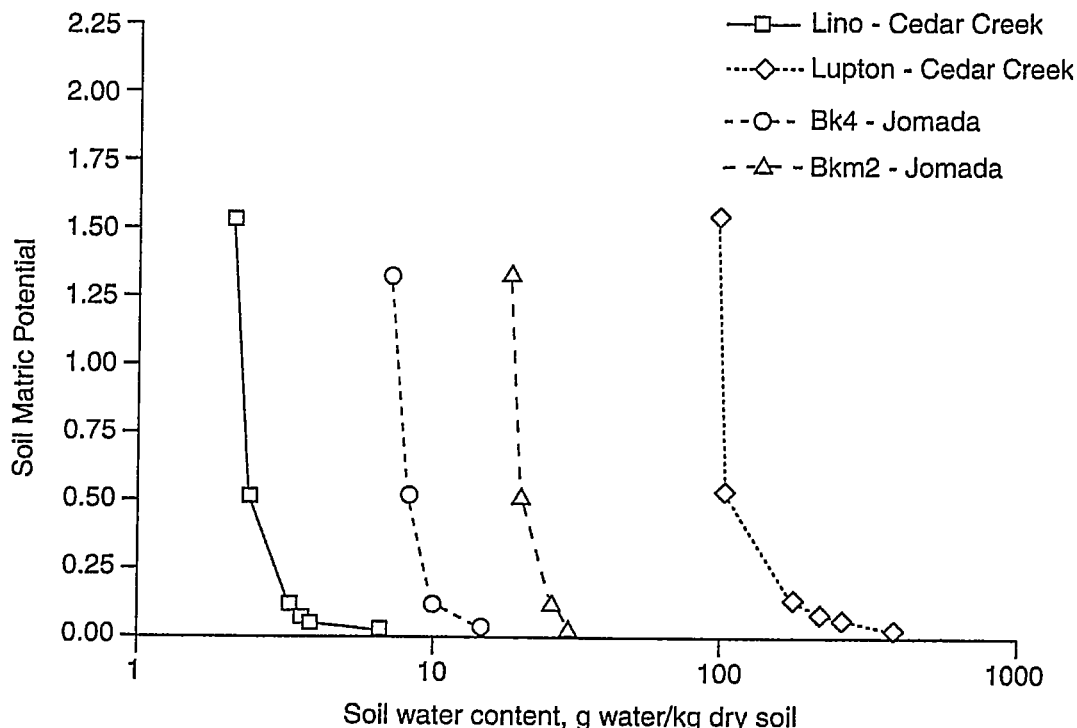


Figure 3.1. A soil water characteristic curve describing the relationship between soil water potential and soil water content.

est water-filled pore. This term expresses the contribution of the matric potential to lowering soil water potential.

The sign of the water potential under these conditions is always negative in unsaturated soils because μ_w is always less than μ_w^* .

The units of water potential are energy per unit of volume. In most cases, these units are converted to force/unit area, or pressure. The standard unit for soil water potential is now the pascal (Pa), or newtons/m². Most soils, particularly where biological processes are active, have soil water potentials between 0 and -2.0 MPa.

Soil water potential is typically measured using one of several techniques:

- *Tensiometer*. When soil water potential is between 0 and -0.08 MPa, a tensiometer can be used. Tensiometers consist of a ceramic cup attached to a sealed plastic tube, with a vacuum gauge attached to the tube. The tube is filled with water, avoiding entrainment of air bubbles, and sealed tightly. The tensiometer is placed in the soil in a hole slightly smaller in diameter than the tensiometer tube. As the soil dries, water moves out of the tube into the soil, through small pores in the ceramic cup. This creates a tension, or partial vacuum, within the tube, which is reflected by a change in the vacuum gauge reading.
- *Thermocouple psychrometer*. The thermocouple psychrometer is a thermocouple junction surrounded by a ceramic cup. The cup takes up water from the surrounding soil. The water evaporates inside the cup and creates a characteristic

humidity, which is related to total soil water potential by the psychrometric equation

$$\Psi = (R \times T/V_w) \times \ln (rh)$$

where

Ψ = total soil water potential

R = ideal gas constant, $8.31 \text{ J} \cdot \text{K}^{-1} \cdot \text{mol}^{-1}$

T = temperature ($^{\circ}\text{K}$)

rh = relative humidity of the air

Since rh is affected by both the matric potential and the solute potential, the psychrometer measurement senses both. Because the rh in most soils is 99% or higher, the measurement of humidity must be very precise, and temperature measurement or control is critical. Psychrometers work best in soils drier than -0.15 MPa .

- *Resistance block.* A solid block of material—frequently gypsum, or $\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$ —is cast around two electrodes. Water is absorbed into the pores in the block and dissolves some of the low-solubility gypsum, lowering the resistance to an electrical current passed between the electrodes. The more water that enters the block, the lower the resistance. A standard curve is created that relates the soil matric potential to the resistance in the block.

Resistance blocks are relatively inexpensive and work in many situations. Their limitations include the fact that the gypsum eventually dissolves and the standard curve shifts; in addition, in saline soils, readings will be artificially high because of high conductance of the soil water; the soil will appear to have a higher water content than it actually has.

The Soil Water Characteristic Curve

The relationship between the soil matric potential and total soil water content is described in a functional relationship called the *soil water characteristic curve* (SWCC). This curve varies as a function of soil texture (Fig. 3.1). This curve does not provide any information on the effects of salts on soil water availability but does provide information on water availability at different moisture contents.

The SWCC determination is especially sensitive to sample handling. Wherever possible, the SWCC is determined in the laboratory using intact soil cores collected in the field. Large pores are critically important in holding water near saturation, making maintenance of soil structure a fundamental concern (the aggregation of single grains of soil particles; see Chapter 4, this volume). If aggregates are broken down and macropores destroyed, the field water content at high water potentials will be underestimated.

The relationship between soil water potential and soil water content depends on whether the soil is in the process of wetting or drying. This asymmetry is termed *hysteresis*. During the wetting phase, there is resistance to water moving from

smaller to larger pores, so at a given water potential the soil water content is lower than at true equilibrium. During the drying phase, large pores retain water even when pores around them are drying.

The variability of the SWCC leads to problems in selecting a single method of soil water characterization, particularly if both quantity and energy status are important aspects of the study. If only water potential is measured, a poor SWCC may result in either overprediction or underprediction of total soil water content. If the soil water contains significant salt concentrations, converting total water potentials to total water contents will underestimate water contents, particularly under drier conditions.

Soil Temperature

Soil surface radiation balance can be represented as follows (Buchan 1991):

$$R_n = [(1 - \alpha) \times R_s] + (\epsilon \times L_d) - L_u$$

where

R_n = net radiation

α = short-wave reflection coefficient (albedo)

R_s = incident solar radiation

ϵ = long-wave emissivity of the soil surface

L_d = long-wave radiation input

L_u = long-wave radiation emitted

The net radiation may be partitioned into three terms as

$$R_n = H + (L_v \times E) + G_o$$

where

R_n = net radiation

H = sensible heat flux between soil and air

L_v = latent heat of vaporization

E = mass flux of water evaporated

G_o = heat flux into soil

We are typically concerned with temperatures throughout the soil profile, since temperature directly affects rates of biological and chemical reactions.

Historically, temperature measurements use the thermal expansion properties of liquids or metals to indicate temperature change. Alcohol, mercury, and bimetallic thermometers are still commonly used. Small, rugged bimetallic thermometers are favored by soil scientists because they can be inserted directly into soils without breaking.

A second approach to temperature measurement is the use of thermocouples, based on the Seebeck effect. If two welded junctions are created with specific metal alloys of wire, current will flow through the wires at a rate proportional to the difference in temperature of the two junctions. If the temperature of one junction is

controlled or known, the temperature of the other junction can be determined by measuring the current flowing in the wire. The two junctions together are termed the *thermocouple*.

A third method for measuring temperature involves thermistors, based on another electrical property, resistivity. Thermistors are materials in which resistance to current flow is a function of temperature. These materials are called *thermistors*. Once the relationship between temperature and resistance is known, the temperature of a given location in the soil can be assessed simply by determining the resistance to electron flow through the thermistor.

Sensor Installation—General Considerations

Installation of soil water sensors in the field requires attention to several factors that can influence moisture and temperature measurements. First is the changing fraction of large pores. This can occur either through compaction and loss of large pores in the vicinity of the sensor or by creating and not refilling large pores (air gaps) in the soil during installation. Ideally, the soil adjacent to the sensor should have the same pore size distribution as the bulk soil. This is particularly critical for conditions near saturation, since within this range the large pores fill and the fraction of total porosity present as large pores becomes even more critical. Minimal disturbance to the soil is always recommended. This includes keeping installation hole size to a minimum and not installing sensors when soils are wet and easily compacted or dry and susceptible to air gaps.

Second is the danger of modifying water movement through the soil. This may occur because of the creation of channels of large, continuous pores, e.g., adjacent to tubes or wires installed in the soil. These channels allow preferential flow, potentially accelerating wetting and accentuating drying compared with the bulk soil.

Third, be aware of ways that sensors can affect soil thermal conductivity. To avoid undesirable effects and to protect sensors from weather and animals in general, shield exposed portions of sensors appropriately. At the very least, sensors should be shielded from direct solar radiation.

Gravimetric Water Content

Gravimetric water content (Gardner 1986) is the most direct, laborious, and disruptive method for determining the instantaneous water content in soils. The soil sample is removed from the field, oven-dried, and reweighed. This method is most useful when one must determine water content of a sample collected for some other chemical, physical, or biological analysis.

Materials

1. Field coring device or shovel
2. Ziplock bags or other watertight containers

3. Weighing cans (preferably aluminum with tight-fitting lid, 250 mL capacity)
4. Balance with resolution to 0.01 g

Procedure

1. Collect samples from known depths. The depths selected may be based on soil horizons, in which case the horizon thicknesses and bulk densities must be known to allow conversion to volumetric and spatial units. Alternatively, samples may be collected at fixed depth increments (see Chapter 1, this volume).
2. Transport the samples to the laboratory in sealed containers that prevent evaporation. In the lab, weigh a portion of the field-moist soil (approximately 100 g) into a preweighed drying container with lid (frequently an aluminum can with a tight-fitting lid, although any drying container that maintains constant weight when dried at 105 °C can be used).
3. Place the open container in a drying oven set at 105 °C. When soil is dried to constant weight, remove the container from the oven, seal it with the lid (to avoid rehydration), allow to cool to room temperature, and then weigh the entire can plus lid.

Calculations

Gravimetric Water Content

After the weight of the container plus lid is subtracted from moist and dry samples, the gravimetric water content (always referenced to the oven-dry soil mass) is calculated as

$$\theta_g = [(g \text{ moist soil}) - (g \text{ dry soil})]/(g \text{ dry soil})$$

where θ_g = gravimetric water content as g H₂O/g dry soil.

Volumetric Water Content

Use soil bulk density information (see Chapter 4, this volume) to convert θ_g to volumetric water content θ_v :

$$\theta_v = \theta_g \times BD$$

where

θ_v = volumetric water content as mL H₂O/cm³ soil or m³H₂O/m³ soil

θ_g = gravimetric water content as g H₂O/g dry soil

BD = soil bulk density as g dry soil/cm³ soil or Mg dry soil/m³ soil.

If high precision is not required, one can assume clay soils have bulk densities around 1.1 g/cm³ and soils high in sand have bulk densities nearer 1.7 g/cm³. However, compaction can significantly increase bulk density, while well-aggregated soils high in organic matter can have much lower bulk densities (well below 1 g/

cm³). Under these conditions, bulk density should be determined directly to make the conversion accurate.

To convert from volumetric to areal values, θ_v is multiplied by the depth of each horizon for which θ_v has been calculated. For example, if θ_v is 0.42 m³ water / m³ soil and the depth increment is 0.2 m, this depth increment contains the equivalent of $(0.42 \times 0.2) = 0.084$ m of water, or 8.4 cm. By adding together the values for each depth increment, the total water storage in the profile can be calculated in terms of centimeters of water. To convert to a volume of water, simply multiply this value by the land area of concern, e.g., a square meter or hectare.

Special Considerations

While they are direct and require little equipment, gravimetric water content measurements are time-consuming and labor-intensive, create substantial disturbance, and can drastically alter the hydrologic cycle. In addition, gravimetric measures do not allow the user to obtain frequent readings from exactly the same site, cannot be automated, and require bulk density measurements for conversion to volumetric terms.

See Chapter 4 (this volume) to calculate water-filled pore space (WFP) from gravimetric water content values.

Time Domain Reflectometry (TDR)

The TDR approach (Dalton 1992) allows continuous or intermittent measurement of total soil water content with minimal disturbance. The technology is advancing quickly, and equipment prices are dropping. TDR represents the best available method for regular monitoring of soil water content. In addition, with appropriate adjustment (e.g., instrument frequency changes), one can also measure soil salinity.

Materials

1. TDR electronics (e.g., Tektronix line tester, Campbell Scientific)
2. Wave guides (probes). Guides can be cut from stainless steel rods or can be purchased commercially.
3. Installation guides

Procedure

1. Probes are inserted in parallel pairs a precise distance apart from one another; they typically are inserted vertically from the soil surface. Commercial probes vary in length from 15 to 70 cm. Differences in measured water contents among probes of different length can be used to estimate water contents over differences in depth increments. For example, a 15–30 cm depth increment can be estimated as the difference between readings from a 0–30 cm depth pair and a 0–15 cm depth pair.

For more precise measurements of soil water at particular depths, probes may be installed horizontally through the sidewalls of a soil pit or trench. Depth increments can be regular or based on major discontinuities in soil properties that affect hydrology, such as texture or structure. In most cases, probes 15–25 cm in length work well. Note that the presence of a pit or trench can severely disrupt local hydrology; the pit should be backfilled to conditions as close as possible to those of the undisturbed site.

If the objective is to understand quantities and patterns of water uptake in the plant root zone, sensors should be distributed through the entire depth of the effective zone of water use. In most cases, a minimum of three sensors equally distributed down a profile is necessary. In deeper or more heterogeneous profiles, other spacings may be necessary to capture the average pattern of water content as a function of depth. Run the sensor wire leads up to the soil surface; protect the lead ends with a waterproof cover.

2. Read water content at a frequency appropriate to the study and site. Sensor systems that are set up for continuous readings are usually interfaced through a multiplexer to a field datalogger unit. In most cases, hourly readings are adequate to describe rates of water loss from the soil through leaching or evapotranspiration. However, if detailed tracking of wetting patterns is a major concern, more frequent sampling intervals may be required, at least for short periods of time.

While there is a natural tendency to limit readings to the time of most active biological activity (frequently the moist part of the summer), there are good reasons for year-round monitoring. This is especially important where leaching patterns, including water loss, are evaluated.

3. The TDR unit should be periodically checked and calibrated against soil samples of known water content. This should be done initially using both *in situ* and laboratory techniques.

In situ check. When sensors are installed, a soil core can be taken near the site and sectioned into segments that represent equal distances between sensors or horizon depths. The gravimetric water content of these sections is then determined and compared with the initial TDR readings. Values should be within $\pm 2\%$ for both techniques.

Laboratory calibration. Insert probes into a variety of soil samples for which water content is known; preferably, soil samples should be intact cores or otherwise minimally disturbed samples from the site of interest. Compare readout of the device with the known water content and make calibration adjustments as needed. Individual cores can be moistened and then allowed to air-dry to provide a wide range of moisture conditions (moisture loss can be monitored gravimetrically).

Calculations

Modern TDRs typically provide direct readouts of volumetric soil water content. Extrapolating to depth of water in the effective root zone uses the same procedures as described earlier for gravimetric water content.

Special Considerations

Calibration can be a problem, especially for organic soils. While uncalibrated values can be obtained that may be useful, additional in-field or in-laboratory calibration is required for accurate measurements of volumetric water content at any particular site.

When choosing sensor sites, select representative areas of the landscape for moisture determinations; avoid anomalously low or high places unless this variation is an important element of the study.

Soil Water Potential—Tensiometers (0 to -0.08 MPa)

No devices are available that measure soil water potential over the entire range of values of interest in biology. As a result, we recommend tensiometers for measuring water potentials under wet conditions (0 to -0.08 MPa) and resistance blocks for measuring potentials in the dry range (-0.10 to -3.0 MPa).

Materials

1. Tensiometer with an appropriate tube length
2. Pump for priming the ceramic cup
3. Deaerated water (boiled)
4. Auger of diameter slightly smaller than tensiometer

Procedure

1. Prepare the tensiometer by placing its ceramic cup end in a large bucket of deaerated water. Unscrew the end cap from the tube, place the vacuum pump mouth over the tube, and pull a vacuum that will bring water through the ceramic cup, displacing air and filling all pores with water. Leave the vacuum on (at about -0.09 MPa) until no bubbles emerge from the ceramic cup (an advantage of using clear tubing material).

Finish filling the tube with deaerated water and screw the cap on tightly. Make sure no air bubbles are trapped in the water column. If air bubbles are present, the tensiometer will not respond to drying soil, and will always read wetter than actual soil conditions.

2. Bore a hole to the desired depth with the auger. Depths, as discussed previously for the TDR method, are selected as appropriate for characterizing the dynamics of water input and removal from the root zone of the plant. However, in practical terms, few tensiometers are longer than 1.5 m. Insert the tensiometer tube so that it fits snugly in the hole. Inserting the tube at an angle is recommended because soil directly above the ceramic cup is less disturbed. However, under no conditions should there be a gap between the tensiometer tube and the soil, because preferential flow will occur during precipitation events and the area around the ceramic cup will wet faster than the

bulk soil. Poor tube-soil contact can accelerate drying rates as well, since water vapor can rapidly move up and out of the soil.

3. Read the tensiometer at a frequency appropriate to the site and study. This will depend on the rate of soil drying and the specific research questions. If the soil is not near saturation initially, the manometer needle should depart from zero shortly after installation. If the soil is very moist, the needle may stay on zero until drainage or evapotranspiration decreases water potential below -0.002 MPa. If the soil becomes drier than approximately -0.08 MPa around the ceramic cup, air will enter the pores of the cup and the reading will rapidly drop back to zero. At this point the tensiometer has become useless, and it must be removed and refilled before again installing in the soil. The problem can be partly overcome by using ceramic cups with smaller maximum pore sizes, but their resistance to water movement results in slow attainment of equilibrium in soil.

Calculations

The tensiometer reads water potential directly (sometimes potential is described as soil suction), so no additional calculations are necessary.

Special Considerations

Do not install a tensiometer in the landscape or at depths where the soil will be drier than -0.08 MPa most of the time unless you are prepared to frequently remove, recondition, and reinstall the device after rewetting has occurred. In wetland soils, in soils above a shallow groundwater table, or in irrigated sites there may be extended periods when the soil will remain wet. In other cases, use information from the TDR measurements or, for generally dry sites, use gypsum blocks (see next section).

Soil Water Potential—Resistance Blocks (0 to -3.0 Mpa)

The use of resistance blocks (Campbell and Gee 1986) is straightforward. As indicated earlier, resistance blocks reach maximum conductance at approximately -0.1 MPa (-0.2 MPa in some cases) and therefore are unresponsive in wet soils. Use tensiometers (as described previously) for wet (0 to -0.48 MPa) soils.

Materials

1. Resistance meter
2. Gypsum or fiberglass resistance blocks with leads of required length
3. Soil auger (3 cm diameter)

Procedure

1. Bore auger holes to desired depths (see the TDR protocol, described earlier). Place the gypsum block in the bottom of the hole, ensuring that it fits snugly

with the soil. Backfill the hole with soil, making sure it is tamped firmly so it does not become a channel for preferential flow. Bring the leads to the surface and tie them together. Connect them directly to the resistance meter/datalogger device and place them in a protective container.

2. Make readings at a frequency appropriate to the study and site. To capture events that involve rapid infiltration, time steps may be on the order of 15 minutes, or at even shorter intervals in very sandy, high-porosity soils. For overall monitoring of soil water status, storing hourly data will provide adequate temporal resolution for most applications.

Calculations

The output may need to be converted from resistance readings to soil water potential using calibration curves provided with the blocks or developed as described below. The potential readings can then be converted to volumetric soil water content estimates using the soil water characteristic curve (see next section). However, depending on the degree of hysteresis exhibited by the soil and the spatial variability in the field, these conversions may result in serious errors in estimating soil water content. Therefore, if soil water content is the desired parameter, it is preferable to measure it directly.

Special Considerations

Over time, the calibration of the blocks may drift due to changes in the chemical composition of the block or to dissolution and loss of gypsum. In most cases the blocks should be removed and recalibrated on a yearly basis. The blocks should be buried in a sample of field soil at a known water potential (e.g., -0.3 MPa based on psychrometric measurements in the laboratory). After the reading has stabilized, if the resistance readings vary more than 5% from the standard curve accompanying the block, a new calibration curve should be developed (at least four points between -0.1 and -2.0 MPa), or the block should be discarded and replaced with a new unit.

The resistance block method is sensitive to soil salinity; any salts entering the block will give artificially wet (low resistance) readings. They are not recommended for soils with saturation extract specific conductance greater than 2 dS/m. Under these conditions, if soil water potential must be measured directly, soil psychrometers are recommended.

Soil Water Characteristic Curve—Psychrometry

In the laboratory, pressure and vacuum plate apparatuses have been used to measure water potential in soil samples (see Klute 1986). One important difference between classical pressure plate methods and the psychrometer is that the latter measures both the matric potential and the solute or osmotic potential—effectively the total potential of the system. The pressure plate measures only the matric potential and is thus less preferred.

Materials

1. Psychrometer
2. Sealable bags

Procedure

1. Add a known amount of water to a variety of dry soils to create a range of samples with known soil moisture contents. Alternatively, determine the gravimetric water content of a range of moist soil samples (see previous section "Time Domain Reflectometry"). Added water must be uniformly distributed throughout the soil. In some cases, freezing the dry soil and mixing with an appropriate mass of crushed ice can lead to a more uniform and rapid equilibration; this is especially appropriate under dry soil conditions.
2. Keep the soil in a sealed bag to allow equilibration for a period of 1–2 days.
3. Place duplicate samples in a psychrometer calibrated following the manufacturer's recommendations. The psychrometer measures the relative humidity of the atmosphere above the samples.

Calculations

The water potential is calculated from the following equation:

$$\mu_w = (R \times T/V_w) \times \ln (p/p_0)$$

where

μ_w = energy potential of water (joules/mole)

R = ideal gas constant, $8.31 \text{ J} \cdot \text{K}^{-1} \cdot \text{mol}^{-1}$

T = absolute temperature, °K

V_w = partial molar volume of water, 0.018 L/mole

p = water vapor pressure in equilibrium with the liquid phase

p_0 = saturated water vapor pressure of the liquid phase

This equation highlights the extreme temperature dependence of this relationship, both directly through T and indirectly through p_0 . Temperature must thus be known very precisely.

The relationship between soil water potential and soil water content is then graphed as the SWCC. Frequently the relationship is expressed as a semi-log graph, with the soil water potential graphed on a common log scale (see Fig. 3.1).

Soil Temperature

For long-term, continuous measurements in the soil, thermistor sensor arrays are generally preferred. The thermistor is part of an electrical circuit in which the resistance to current flow is proportional to its temperature (Buchan 1991; Taylor and Jackson 1986). Thermistors are rugged, reasonably precise, stable, and inexpensive.

With the advent of inexpensive self-contained datalogging thermistors, continuous monitoring of soil temperature has become easy and inexpensive.

Materials

1. Soil auger, smallest possible diameter
2. Resistance meter appropriately calibrated for specific thermistors to provide direct temperature readout
3. Thermistors with appropriate adequate cable lengths. Alternatively, integral temperature-dataloggers that can be installed in situ (e.g., the Hobo Unit from Onset Computer Corp.; see Table 3.2) can be substituted for items 1 and 2

Procedure

1. Auger a hole to the maximum depth to which sensors will be installed, retaining the soil. The depths selected will depend on the site and study. For near-surface processes, installation at 5, 15, 25, and 50 cm depths will usually suffice. At most sites, soil temperatures do not change diurnally below 1.5 m. At this depth, annual mean temperature is generally equivalent to 2° above the mean annual air temperature at the location.

Place a thermistor in the bottom of the hole and backfill with soil to the depth of the next thermistor. Place the second sensor in the hole and continue in the same manner until all sensors are in place. Replace soil in approximately the order in which it was removed from the hole.

Surface shading and soil water status will have dramatic effects on soil thermal properties. Make sure the site selected for installation is typical of the area to be characterized in terms of soil water relations and shading cover. In deserts or savannas, installations under and between perennial woody species may be required.

2. Bring all leads to the surface and connect to the temperature meter/datalogger. Protect the dataloggers from direct sunlight and precipitation.
3. For datalogging, storing 1 hour temperature averages is more than adequate to describe the essentials of diurnal and seasonal heating and cooling patterns. In some cases, less frequent measurement is satisfactory. Data collection should be made sufficiently frequent to provide means over time at various depths, temperature ranges, and initiation and duration of freezing if it occurs.

Calculations

Most thermistor units currently available are designed for meters that read out directly in degrees Celsius, so no calculations are necessary.

Special Considerations

Newer probes have extremely reproducible output, so calibration in the laboratory is less important than in the past. Nevertheless, calibration should be checked annually at two temperatures that bracket the temperature range of interest. For cali-

bration, place probes in a constant-temperature bath and compare the readings to those on a reference thermometer. If probes are not identical (within 0.2 °C of bath temperature), they should be tested at several temperatures to generate individual calibration curves relating bath temperature to output. In this case they must be individually marked in the field so the appropriate calibration or correction can be applied.

Be aware of the potential for installation effects on temperature profiles. Exposure of wire leads to direct sunlight, for example, can affect temperature readings, and precautions should be taken to avoid direct solar radiation. Also, take care to backfill augered soil carefully so as not to change the soil's thermal conductivity with depth.

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