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Soil Water and Monitoring Technology

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Soil water content and the energy potential of soil water are properties that are often determined in order to guide management of irrigation and drainage, and that are frequently necessary data in irrigation and drainage research. The soil water status affects the transpiration stream of crops and their ability to uptake nutrients from the soil and CO₂ from the atmosphere for yield formation. Because soil water status is reflected in crop water status, there are many links between soil water status and crop physiological response, including leaf turgor and orientation, growth through cell expansion and division, rooting, stomatal size, chemical (hormonal) processes, flowering, fruiting, senescence, canopy temperature, etc. Because soil water status is modified by crop water uptake and evaporation, the dynamics of soil water properties are affected by the dynamics of weather, both diurnally and over longer periods, and by crop growth and senescence as modified by soil fertility, pests, and diseases. Thus, observation of soil water dynamics may be useful for crop management. The objective of irrigation and drainage management is to control soil water status; but the goal for crop management may be to obtain maximum crop yield, maximum water use efficiency, soil dry enough for harvest operations, or a combination of these with other goals.

PHYSICAL PROPERTIES OF WATER

Water, known as the universal solvent, is the carrier for fluxes of nutrients, agrichemicals, and some disease vectors through the soil to plants. Composed of hydrogen and oxygen (H₂O), water reaches a maximum density of 1.00 g cm⁻³ at 4°C, is slightly less dense just before it freezes, and has a density of 0.958 g cm⁻³ at 100°C (Fig. 2-1). The density of water increases with its salt content, a factor that may cause saltier water to accumulate lower in an aquifer, resulting in increasing salinity problems as the aquifer is depleted. The density of ice is 0.917 g cm⁻³, which is why freezing and thawing of soil causes swelling and shrinking, respectively, resulting in soil property modifications; for example, a decrease in bulk density and “fluffing” of the near-surface soil in clayey soils after

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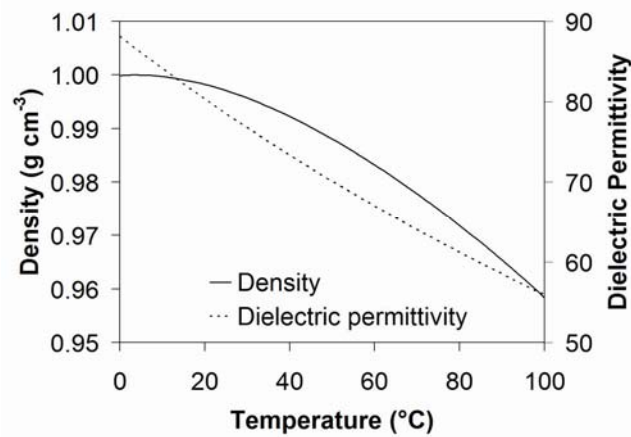


Fig. 2-1. Both density and dielectric permittivity of liquid water vary with temperature.

several freeze-thaw cycles. Natural ice often contains entrapped air that further reduces its density.

The measured electrical properties of water vary with temperature, phase state, amount and kind of dissolved substances, and the electrical frequency of the measurement system. This complicates the measurement of soil water content and soil electrical conductivity by electrical methods. For example, the dielectric permittivity (ϵ') of water is ~ 78.3 at 25°C , increasing to ~ 87.8 at 0°C for liquid water (Fig. 2-1) (Weast, 1971, p. E-61); yet the dielectric permittivity of ice is ~ 3 at the frequencies used in soil water sensing. Still, because the dielectric permittivity of water is so much larger than that for other soil constituents, the electromagnetic (EM) methods for soil water estimation have some usefulness (ϵ' is 2 to 5 for soil organic matter and 3 to 7 for soil mineral matter, Topp and Ferré, 2002). Pure water is essentially an insulator, but electrical conductivity of water increases greatly with minor additions of ionic compounds such as salts. This conductivity increases about 2% per $^\circ\text{C}$ in soil temperature for salts commonly dissolved in soil water. Unfortunately, such conductivity can interfere with the EM soil water sensing methods.

SOIL WATER POTENTIAL

The energy potential of water in soil varies with the soil water content. At smaller water contents capillary forces cause water to remain in smaller soil pores at relatively more negative values of potential relative to that of free water. Typically, the energy potential of water in the atmosphere is even more negative than that of water in the soil, so that water flux occurs from soil to plant roots and through the roots, stems, leaves and stomata to the atmosphere. This flux rate increases with the difference in energy potential between soil and atmosphere as moderated by resistances in the roots and leaves. Thus, when the soil water potential becomes more negative as the soil dries, the flux rate decreases and plants transpire less water. Because of the strong correlation between transpiration and yield formation, this translates into yield reduction. From this perspective,

irrigation can be seen as the process of increasing soil water potential so that the flux rate of water through the crop is increased.

The total energy potential of soil water, Ψ_T (kPa), is the sum of component potentials

$$\Psi_T = \Psi_M + \Psi_P + \Psi_O + \Psi_Z \quad [1]$$

where Ψ_M is the matric potential, related to the capillary force; Ψ_P is the pressure potential, related to variations in pressure; Ψ_O is the osmotic potential, related to variations in solute concentration; and Ψ_Z is the gravitational potential, related to position in the earth's gravitational field. If the total potential of water everywhere in the soil is equal, then the potential gradient between points is everywhere zero and water will not flow. When gradients in total potential exist, water flux exists in direct proportion to the size of the gradient and to the magnitude of the soil hydraulic conductivity.

In unsaturated soil, water and air both exist in the soil pores. The interface between water and air follows a compound curved surface, the degree of curvature being dictated by the surface tension of the water, inversely proportional to the size of the pore, and influenced by the surface material of the pore. If water tends to adhere to the surface of the pore, then that force is transmitted to the free water surface, exerting a pull, called the capillary force, on the water that tends to make the water move towards the air. Gravity exerts a counteracting force that pulls the water downward. The capillary force is inversely proportional to the size of the pore. The matric potential, Ψ_M , is the energy invested in this capillary force plus the energy of strictly absorptive effects. The latter become large at small water contents. In saturated soils there is no air-water interface and thus no capillary force, and absorptive forces are insignificant; thus the matric potential is zero. If water is repelled by the surfaces of the soil pores, as in hydrophobic soils, then the capillary force is negative and water does not tend to move into the soil unless the negative capillary force is overcome by pressure, such as that of sufficiently deep standing water.

The osmotic potential, Ψ_O , is the potential energy difference between pure water and water containing dissolved substances, both being open to the atmosphere. Pure water placed on one side of a semi-permeable membrane, will move across the membrane to saline water on the other side. This is because the osmotic potential of pure water is zero, and the osmotic potential of saline water is a negative value. Figuratively speaking, the water "falls down hill" along the energy gradient to the saline water side of the membrane. Because soil contains open pores, not semi-permeable membranes, differences in osmotic potential do not influence soil water flux. However, the pathway from the soil to the inside of plant roots does cross semi-permeable membranes; and the flux of water from the soil into the plant is thus decreased as the osmotic potential of soil water becomes increasingly negative, reducing the potential energy gradient from soil to atmosphere.

Pressure potential, Ψ_P , is the pressure on water with respect to a reference pressure, usually taken as atmospheric pressure. Water in the soil is usually under atmospheric pressure so that the value of Ψ_P is zero so long as the soil is unsaturated and continuity of air filled pore space to the atmosphere exists. However, basin or flood irrigation can trap a layer of air in the soil beneath the wetting front, effectively pressurizing the air as the water infiltrates. This pressure increases the pressure potential of soil water in the unsaturated soil beneath the wetting front. In saturated soil, the pressure potential at any point is the hydrostatic head of water above that point. Another example of

pressure potential is the use of pressure, applied with a pump, to saline water on one side of a reverse osmosis membrane. Normally, pure water on the other side of the membrane would move to the saline side because of its negative osmotic potential. However, when the pressure potential applied is larger than the osmotic potential of the saline water, the total potential of the water on the saline side becomes larger than the total potential of the water on the pure water side (which is open to the atmosphere) and water moves from the saline side to the pure water side. This process of desalinization of water is used to provide non-saline water to some irrigation projects.

In saturated soils, the value of Ψ_M is zero and $0 \leq \Psi_P < \infty$. In unsaturated soils, Ψ_P is zero and $0 \geq \Psi_M > -\infty$ (assuming the air in the soil pores is not being pressurized by an overlying saturated wetting front).

Several different units may be used for soil water potential; so it is important to be able to convert between them. If units of length are used, we write the potential as hydraulic head, H , usually with SI units of meters. If energy per unit volume is considered, we write the potential as Ψ , usually with units of Newtons per $\text{m}^{-2} = \text{N m}^{-2} = \text{Pascal} = \text{Pa}$. To convert between the two

$$\Psi = \rho_w g H \quad [2]$$

where ρ_w is the density of water (often taken as 1000 kg m^{-3}) and g is the acceleration due to gravity (9.81 m s^{-2}).

Useful conversions from Warrick (2003) are:

$$\begin{aligned} 10^5 \text{ Pa} &= 1 \text{ bar} \\ 10.22 \text{ m of water} &\approx 1 \text{ bar} \\ 10.35 \text{ m of water} &\approx 1 \text{ atm} \\ 1 \text{ J kg}^{-1} &\approx 1 \text{ kPa} \approx 0.1 \text{ m} \\ 1 \text{ bar} &\approx 100 \text{ J kg}^{-1} \end{aligned}$$

SOIL WATER CONTENT

For most uses and calculations in irrigation research and management, soil water content ($\theta_v, \text{m}^3 \text{ m}^{-3}$) is expressed as a volume fraction,

$$\theta_v = \frac{\text{volume of soil water}}{\text{total volume of soil}} \quad [3]$$

Volume per volume units are used by most models of soil water flux and crop water uptake, including irrigation scheduling computer programs. These units make it easy to convert moisture measured in a soil profile over a given depth, z , to an equivalent depth of water (θ_z) by multiplying the water content by the depth: $\theta_z = \theta_v z$. The units of θ_z are the units of z . For example, the depth of irrigation water, I_{zUL} , that a uniform soil can accept without large losses to deep percolation is limited on the upper bound by the depth of the root zone, z_r , and the difference between the mean

water content of the root zone, θ_r , and the water content at field capacity, θ_{FC} : $I_{zUL} = z_r(\theta_{FC} - \theta_r)$. For soils that are non-uniform, similar calculations can be done layer by layer using the various layer field capacity values and water contents.

Soil water content is also expressed as the mass of soil water per unit mass of oven dried soil,

$$\theta_g = \frac{\text{mass of water}}{\text{mass of oven-dry soil}} \quad [4]$$

This gravimetric soil water content, θ_g (Mg Mg^{-1}), is obtained if the soil sample volume is unknown. The mass of water is determined by weighing the sample before non-negligible water loss can occur, followed by oven drying it at 105°C for 24 h, and weighing the dried sample. The difference in masses is the mass of water; and the mass of the dried sample is the “mass of oven-dry soil” in Eq. [4]. Because it is difficult to obtain soil samples of known volume, gravimetric water contents are commonly obtained. It is also common to convert gravimetric water contents to volumetric using the density of water ($\rho_w \approx 1 \text{ Mg m}^{-3}$) and the bulk density of the soil, ρ_B (Mg m^{-3}),

$$\theta_v = \frac{(\text{mass of water})/(\rho_w)}{(\text{mass of oven-dry soil})/\rho_B} \approx \theta_g \rho_B \quad [5]$$

The first equality in Eq. [5] is exact if the actual density of water under the ambient condition is used and if the bulk density value is correct. The second equality is approximate, as it contains the assumption that the value of the density of water is 1 Mg m^{-3} . Values of θ_v obtained using Eq. [5] are often incorrect due to the use of bulk density values measured elsewhere or because bulk density has changed with time (tillage, compaction, freeze-thaw action, etc.). Soil bulk density is more thoroughly described in the next section.

PHYSICAL PROPERTIES OF SOILS

Soil physical properties determine the range of possible water contents, the range of plant-available water contents, and the freedom of water to move through the soil during infiltration from irrigation or precipitation, redistribution within the root zone after these events, movement between soil and roots, and movement downward away from the root zone or upward towards it. Both irrigation management and methods are tailored to best use the soil water reservoir and to accommodate the rate at which that reservoir can receive water from irrigation and release it to the crop.

Soil texture is quantified by the relative percentages by mass of sand, silt, and clay after removal of salts and organic matter. Both texture and structure determine the soil-water characteristic curve, which quantifies the relationship between soil water content and soil water potential. This relationship differs largely according to texture (Fig. 2-2), but can be strongly affected by organic

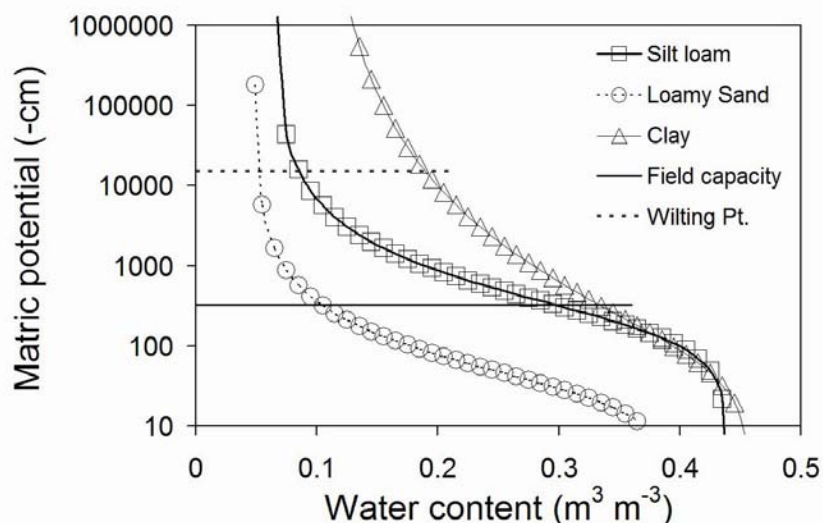


Fig. 2-2. The soil water content vs. soil water matric potential relationship for three soil textures as predicted by the Rosetta pedotransfer model (Schaap et al., 2001). Horizontal lines are plotted for the field capacity, taken as -333 cm (~ 33 kPa), and for the wilting point, taken as $-15\,000$ cm. <http://www.ars.usda.gov/Services/docs.htm?docid=8953>

matter and salt contents. The range of plant-available water (PAW) possible for a given soil is determined by two limits.

The upper limit, also known as the *field capacity*, is often defined as the soil water content of a previously saturated soil after 24 h of free drainage into the underlying soil. The field capacity can be viewed as the water content below which the soil does not drain more rapidly than the crop can take up water. In heavier textured, more clayey, soils, this limit is often characterized as the water content at -0.10 kPa soil water potential. In more sandy (“lighter”) soils, the upper limit may be more appropriately placed at -0.33 kPa potential. The difference in soil water potentials that are related to the upper limit of PAW is due to the relatively large conductivities for water flux in lighter soils near saturation, which means that lighter soils will drain more rapidly.

The lower limit of PAW, also known as the *permanent wilting point*, is often defined as the soil water content at which the crop wilts and cannot recover if irrigated. The soil water potential associated with the lower limit varies with both the crop and the soil; but is often taken to be -1500 kPa. The amount of PAW differs greatly by soil texture. For example, as illustrated in Figure 2-2, a clay soil may have a plant available water content range of 0.19 to 0.33 $\text{m}^3 \text{m}^{-3}$, or 0.14 $\text{m}^3 \text{m}^{-3}$ PAW; whereas a silt loam may have a larger PAW content range of 0.08 to 0.29 $\text{m}^3 \text{m}^{-3}$, or 0.21 $\text{m}^3 \text{m}^{-3}$ PAW. Sandy soils tend to have small amounts of PAW, such as the 0.04 $\text{m}^3 \text{m}^{-3}$ for the sandy loam illustrated in Fig. 2-2 or the 0.06 $\text{m}^3 \text{m}^{-3}$ reported by Morgan et al. (2001a) for an agriculturally important fine sand in Florida. Thus, irrigation management often focuses on applying smaller amounts of water more frequently on sandy soils.

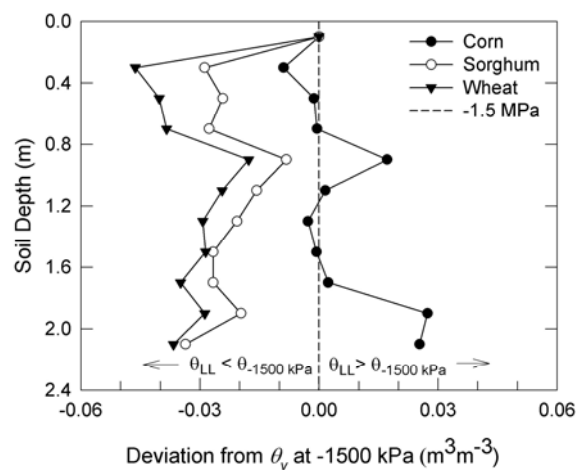


Fig. 2-3. Deviation of the lower limit of water extraction, θ_{LL} , measured in the field using a neutron probe, from that measured at -1500 kPa in the laboratory on soil cores taken at several depths in the soil. Data are for corn, sorghum and wheat crops grown in a Ulysses silt loam (Tolk, 2003).

Additionally, crops differ in their ability to extract water from the soil, with some crops not capable of extracting water to even -1500 kPa, and others able to extract water to more negative potentials (Ratliff et al., 1983, Tolk, 2003) (Fig. 2-3). Confounding this issue is the soil type effect on rooting density and on the soil hydraulic conductivity, both of which influence the lower limit of PAW for a particular crop. The fact that soil properties vary with depth and horization means that the lower limit of PAW may be best determined from field, rather than laboratory, measurements.

The *available water holding capacity* (AWHC) is a term used to describe the amount of water in the entire soil profile that is available to the crop. Because water in the soil below the depth of rooting is only slowly available, the AWHC is generally taken as the sum of water available in all horizons in the rooting zone, calculated for each horizon as the product of the horizon depth and the PAW for that horizon. For example, for a crop rooted in the A and B horizons of a soil the AWHC is the product of the PAW of the A horizon times its depth plus the PAW of the B horizon times the rooted depth in the B horizon (Table 2-1).

The relationship between soil water content and its potential is complicated by the fact that a drying soil will typically contain more water at a given potential than a wetting soil does (Fig. 2-4). This is called hysteresis. For example, the curves in Fig. 2-2 are for a drying soil; that is, they give

Table 2-1. Example calculation of available water holding capacity (AWHC) in the rooting zone of a crop rooted to 0.95-m depth in a soil's A and B horizons, each with a different value of plant available water (PAW).

Horizon	Depth range	Rooting depth	Rooted depth	PAW	AWHC
		cm		$\text{m}^3 \text{m}^{-3}$	cm
A, silt loam	0 to 20	0 to 20	20	× 0.21	4.2
B, clay	20 to 100	20 to 95	75	× 0.14	10.5
Sum					14.7

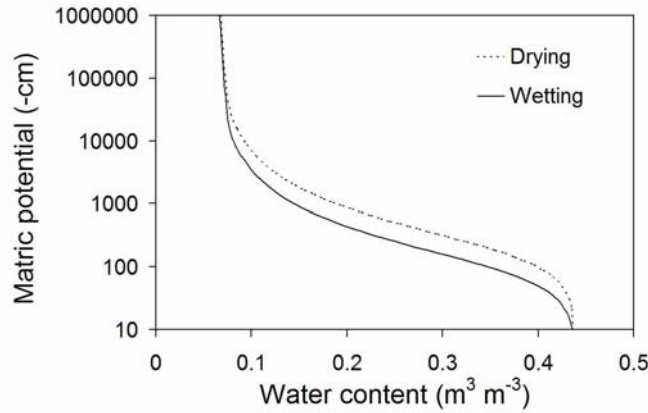


Fig. 2-4. Drying (d) and wetting (w) curves for a silt loam plotted using the van Genuchten equation with $\alpha_w = 2\alpha_d$, which is a first approximation to reproducing the boundary wetting and drying curves if only one is known (Warrick, 2003).

the content of water retained by the soil as it is dried to increasingly more negative values of soil water potential. Plotted relationships for both a wetting and a drying soil give separate curves (Fig. 2-4). Since a soil can be partially dried and then re-wetted, there exist many curves, called scanning curves, that fall between the wetting and drying curves to describe the relationship. Because of hysteresis, it is problematic to accurately convert soil water potential measurements to soil water content.

For soil at a particular water content, soil hydraulic conductivity ($K(h)$, L/T) is the rate constant determining the rapidity of soil water movement as calculated by the Buckingham-Darcy equation

$$J = K(h) \frac{\partial H}{\partial z} \quad [6]$$

where J is the soil water flux rate (L/T), and $\partial H/\partial z$ is the total hydraulic head gradient (dimensionless). The hydraulic conductivity can be seen as a function of soil water pressure head, $K(h)$, or of water content, $K(\theta)$. Hydraulic conductivity as a function of soil water status varies greatly with soil texture (Fig. 2-5), but is also very dependent on organic matter content and on soil structure, including macroporosity engendered by tillage, fauna and flora, cracking due to drying, etc. See Warrick (2003) for discussion on estimating soil water fluxes.

Hydraulic conductivity also typically decreases as soil bulk density increases. Bulk density (ρ_B , $M L^{-3}$) is the mass of dry soil per unit volume of soil.

$$\rho_B = \frac{\text{dry mass of soil solids}}{\text{total volume of soil}} \quad [7]$$

The total volume of soil includes the volume of air, water and other non-solid substances in the undisturbed soil volume. Bulk density is somewhat related to soil texture, with more sandy soils

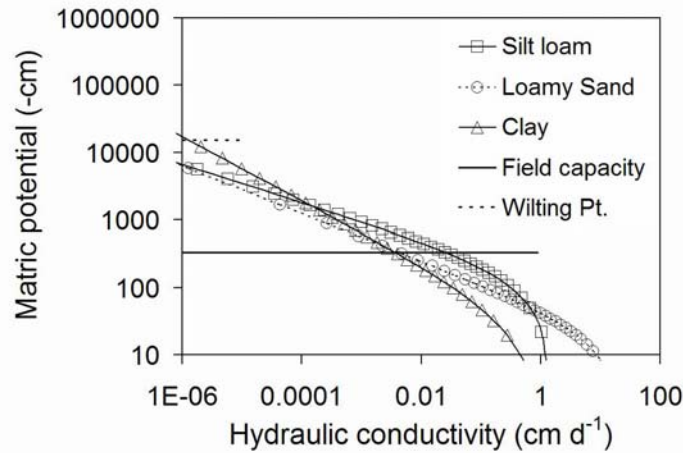


Fig. 2-5. Soil hydraulic conductivity (cm d^{-1}) vs. matric potential ($-\text{cm}$) for three soils.

typically exhibiting larger bulk densities, and more clayey soils smaller bulk densities. However, bulk density is also greatly influenced by tillage (either decreased by loosening or increased by compaction), slaking, rainfall impact on the surface, “fluffing” due to freezing, etc. Soil organic matter content often mediates bulk density. Generally, soils with lower organic matter content are more apt to consolidate, and attain larger bulk density values.

There is also an important relationship between bulk density and soil porosity (ϕ , dimensionless), which is the volume fraction of non-solid material in the soil.

$$\phi = 1 - \frac{\rho_B}{\rho_P} \quad [8]$$

where ρ_P is the particle density, commonly taken as 2.65 Mg m^{-3} , which is the density of quartz. For accurate conversion between bulk density and porosity, the particle density should be measured for a particular soil. Since in agricultural soils the non-solid materials are predominantly air and water, the porosity is nearly equivalent to the volumetric water content of a soil that is completely saturated, providing an important check on the upper limit of non-direct soil water content estimates. If a soil swells with increases in water content, its bulk density will decrease and porosity will increase.

The bulk electrical conductivity (σ_b , S m^{-1}) of soil can be an important indicator of its salt content, which influences crop growth, water uptake, and hydraulic conductivity; and so can be important for irrigation management. However, the value of σ_b depends also on soil water content, being essentially zero in very dry soil and increasing with wetness; and it varies with clay content and type. This is why electrical conductivity measurements, such as those made with mobile Wenner array resistivity measurement systems (eg. the Veris² system <http://www.veristech.com/>), can be

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used to map soil moisture and/or texture under conditions of fairly uniform salinity and texture, or can be used to map soil texture under conditions of uniform salinity and water content. If any two of these properties (texture, water content, and salinity) are non-uniform then the electrical conductivity measurement alone cannot be used to reliably map the third property. It is common that at least two, and often all three, are non-uniform.

SPATIAL AND TEMPORAL VARIATION IN SOIL PROPERTIES

As with all soil properties, those properties of interest for irrigation management and research are variable in three-dimensional space and in time. This variability complicates the tasks of measuring, modeling, estimating or forecasting of soil properties. This added complexity has been dealt with in numerous ways, including compositing of multiple samples into one, and through various statistical approaches. Sample compositing averages sample variability, but can have unintended consequences, as when sample mixing is incomplete or when sample value distribution is skewed. Statistical approaches range from simple descriptive statistics, such as the mean, range, and standard deviation, to more complex analyses involving estimation of the statistical distribution representing the samples (eg. Gaussian, Log-normal, Poisson, etc.), skewness and kurtosis of the distribution, or analyses in space or time such as spatial variogram analysis followed by kriging to derive maps of sample value estimates, or time-series analysis. A full discussion of statistical treatments is beyond the scope of this work, but useful discussions are given by several authors in Chapter 1 of *Methods of Soil Analysis* (Dane and Topp, 2002) on sampling theory, descriptive statistics, and geostatistics; and by Nielsen and Wendroth (2003) on time series and state space analysis and geostatistics. See also Nielsen and Bouma (1985) and Warrick (2002, 2003).

Classical research methods for dealing with soil spatial variability include selection of plot sizes large enough to average out small scale variability, blocking of plots (two or more areas, all of which include all of the experimental treatments), randomization of treatment plots within blocks, and inclusion of measurements of important properties that are correlated with the properties under study (covariate analysis). Statistical methods that include covariates include the general linear model as applied to analysis of covariance (ANCOVA), and covariogram analysis and cokriging.

Spatial variability studies usually find that variance between soil water content samples increases with the distance between samples, the separation distance. But, the same studies indicate that there is a nugget effect; that is, the variance between samples does not go to zero at small distances. For most soil water sensors, it is this small-scale (<1 m), non-zero variance that influences the variability of estimated water contents.

The sample support size or volume has a great effect on the ability to determine this small scale variability. Support volume is tied to the concept of the representative elemental volume (REV), illustrated in Fig. 2-6. For example, a sample size smaller than the size of soil pores could obtain a sample either in pore water, in soil solids, or in an air-filled pore. For the first, the water content would be $1 \text{ m}^3 \text{ m}^{-3}$, and for the latter two the water content would be zero. As sample volume increases, more and more of the small scale variability in the soil fabric is integrated into each sample, and the range of possible values decreases. The REV is the sample volume at which most of the small scale variability is integrated. The REV is different for different soil properties and changes over time for some properties such as soil water content.

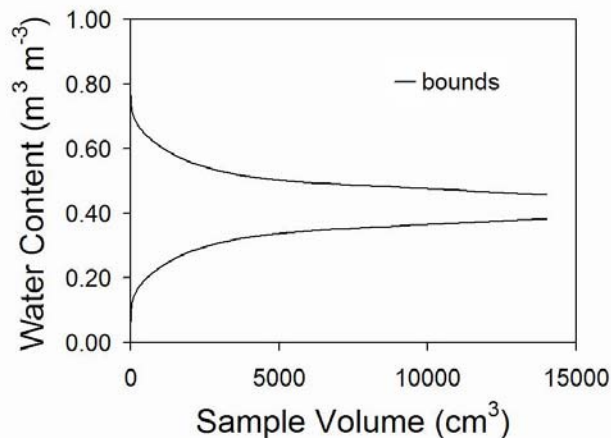


Fig. 2-6. Example of bounds on likely sample values as sample volume increases. The representative elemental volume (REV) can be chosen according to the acceptable variability in sample values.

The concept of an REV is supported by field measurements. Hawley et al. (1982) studied the relationship between sample volume and variance of water content samples, using eight different sample volumes ranging from 7 to 825 cm³, and concluded that variance increased for smaller volumes. The same was true when a 15-cm³ sampler was compared with a 60-cm³ sampler for neutron moisture meter (NMM) calibration (Allen et al., 1993; Dickey et al., 1993). Most other studies of soil water variability used only one sample size or did not report the sample size. The NMM measures, at minimum, a volume of ~14 000 cm³. Comparing this with the much smaller sampling volumes of most gravimetric methods, or time domain reflectometry (TDR) and capacitance probes, indicates that more measurements would be needed with these technologies to give a field or plot mean profile water content with precision comparable to that from neutron thermalization. This was recognized as early as the 1960s and was an important factor in the adoption of the NMM for crop water use measurements based on soil water balance (Calif. Dept. Water Res., 1963).

Thus, in comparing the variance in water content as measured by different methods, it is useful to keep in mind that measured variation of water content in a field is likely to increase as the volume of soil that is measured decreases. Approximately 24 soil samples (50-mm diameter and 0.3-m long = 589 cm³ or approximately the volume sampled by a 0.3-m TDR probe) would be needed to sample the same soil volume as one NMM measurement. Small scale variation of soil water is controlled by topography, vegetation, soil properties, sampling depth (Hawley et al., 1982); and for a particular location, variability increases with time since wetting (Schmitz and Sourell, 2000) and decreases as water content increases (Famiglietti et al., 1999; Hawley et al., 1982; Hupet and Vanclooster, 2002; Schmitz and Sourell, 2000). These studies indicate that more samples will be needed in drier soils to attain the same precision of measurement as in wetter soils. Thus, more sensors are required for a given precision of measurement at the same time that irrigation decisions are commonly made, when the soil has dried to the management allowable depletion level. No simple statement of the desired sample volume can be given, other than to state that fewer large

volume samples will be needed to determine the mean value within a given confidence interval than would be needed if smaller volume samples were obtained. For a parallel and useful discussion relevant to irrigation scheduling see Schmitz and Sourell (2000). Variance in observed soil properties can also affect the precision of calibration of water content sensors as is discussed in a later section.

Because of differing sensed volumes, different sensors will report water contents with different degrees of variance or standard deviation in the field. The variance due to sample volume size is in addition to other sources of error or variation. For n readings, x , of a soil water sensor, the sample standard deviation, s , is

$$s = \left[\frac{1}{n-1} \sum_{i=1}^n (x_i - \mu)^2 \right]^{0.5} \quad [9]$$

where μ is the sample mean and the subscript i indicates the i^{th} reading. For a given value of standard deviation, S , the number of readings, n , required to estimate a mean value with an error $< d$ can be estimated as

$$n = \left(\frac{u_{\alpha/2} S}{d} \right)^2 \quad [10]$$

where S is estimated by the sample standard deviation, $u_{\alpha/2}$ is the $(\alpha/2)$ value of the standard normal distribution, and $(1 - \alpha)$ is the probability level desired (eg. 0.95 or 0.90). Equation [10] is valid for normally distributed values that are independent of one another and for the population standard deviation, S , estimated from a large number of samples. In an experimentally based approach, Tollner et al. (1991) recast this in terms of the working range (WR) of interest for a particular use (irrigation scheduling) or study. They assumed that for useful results there was a certain percentage of the working range ($\%WR$) that should be discernible as statistically different at the $(1 - \alpha)$ probability level, so they computed

$$n = \left(\frac{2t_{\alpha(df)} \sqrt{EMS}}{WR \frac{\%WR}{100}} \right)^2 \quad [11]$$

where the square root of the error mean square (EMS) corresponds to s in Eq. [9], and $t_{\alpha(df)}$ is the value of the t -distribution for the α significance level with df degrees of freedom. They determined EMS with df degrees of freedom from a repeated measures analysis of measurements in order to remove the effect of time. This analysis uses the t -distribution, assuming that the number of samples used to estimate EMS is not large.

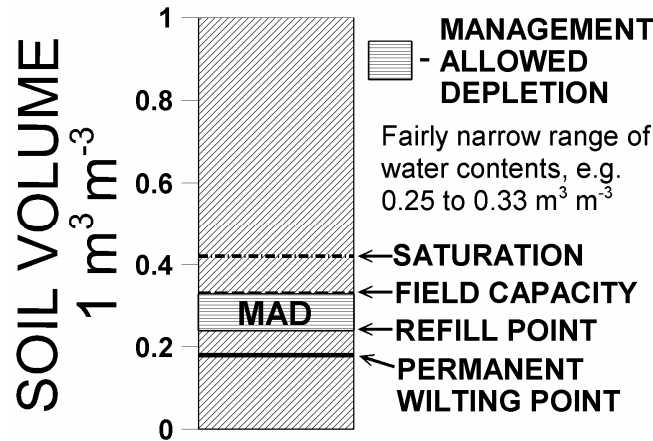


Fig. 2-7. Cartoon of the soil profile indicating fractions of the total soil volume (here represented by unity) that are occupied by water at four key levels of soil water content. For this silty clay loam, the soil is full of water at saturation ($0.42 \text{ m}^3 \text{ m}^{-3}$), drains easily to field capacity ($0.33 \text{ m}^3 \text{ m}^{-3}$), and reaches the permanent wilting point (15 bars) at $0.18 \text{ m}^3 \text{ m}^{-3}$ water content. To avoid stress in a crop such as corn, irrigations are scheduled when the soil water content reaches or is projected to reach $0.25 \text{ m}^3 \text{ m}^{-3}$, the value of θ_{MAD} for this soil and crop.

For irrigation scheduling using the management allowed depletion (MAD) concept (Fig. 2-7), irrigation is initiated when soil water has decreased to the θ_{MAD} level. The θ_{MAD} value may be chosen such that the soil never becomes dry enough to limit plant growth and yield, or it may be a smaller value that allows some plant stress to develop. It is common to irrigate at some value of water content, $\theta_{\text{MAD}+}$, that is larger than θ_{MAD} . This is done to ensure that the error in water content measurement, which may cause inadvertent over estimation of water content, is not likely to cause irrigation to be delayed until after water content is actually smaller than θ_{MAD} . Minimizing the difference, $d = \theta_{\text{MAD}+} - \theta_{\text{MAD}}$, allows irrigation interval to be increased. It is desirable to know the number of samples required to estimate the water content to within d of θ_{MAD} at the $(1 - \alpha)$ probability level. Knowing the standard deviation, s , of soil water content measurements, the required number of samples, n , is

$$n = \left(\frac{t_{\alpha(df)} s}{d} \right)^2 \quad [12]$$

The above three examples assume that samples are taken from an area small enough that large-scale spatial variability does not come into play. In the event that spatial variability is important, the number of samples must be increased such that an adequate number of samples is available for each spatially different area (Vauclin et al., 1984). Also, because these analyses depend on the sample standard deviation determined by repeated readings with a particular device, they encapsulate the variability of readings from that device; but they do not include bias (non-random error) that may be present in the device readings due to, for example, inaccurate calibration. Aside from large-scale spatial variability, the calibration is the largest source of error; and this error is not reduced by

Table 2-2. Example calculation† of management allowed depletion (MAD, $\text{m}^3 \text{m}^{-3}$) in three soils with widely different textures. The small range of MAD severely tests the abilities of most soil water sensors, particularly for the loamy sand soil.

Horizon	θ_{FC}	θ_{PWP}	θ_{PAW}		MAD		MAD
	-----	$\text{m}^3 \text{m}^{-3}$	-----		fraction		$\text{m}^3 \text{m}^{-3}$
silt loam	0.086	0.295	0.209	×	0.6	=	0.126
loamy sand	0.066	0.103	0.037	×	0.6	=	0.022
clay	0.190	0.332	0.142	×	0.6	=	0.085

† θ_{FC} , θ_{PWP} , and θ_{PAW} are the soil water content at field capacity and the permanent wilting point and the plant-available water

repeated sampling (Vauclin et al., 1984). Thus, careful field calibration is essential to minimize such bias (Hignett and Evett, 2002; Greacen, 1981). In most cases, these analyses may be applied to values of soil profile water storage that are calculated on the basis of samples at multiple depths.

For example using the data for the three soils in Fig. 2-2, the differences between the values of water content at field capacity, θ_{FC} , and at the permanent wilting point, θ_{PWP} , are the plant available water, θ_{PAW} (Table 2-2). Assuming that the management allowed depletion is 0.6 of θ_{PAW} , the allowable ranges of water content during irrigation scheduling are 0.126, 0.085, and 0.022 $\text{m}^3 \text{m}^{-3}$ for silt loam, clay, and loamy sand, respectively (Table 2-2). These narrow ranges place high demands on soil water sensing equipment. Assuming that soil-specific calibrations have been performed to minimize bias, and that the accuracy of calibration (as determined by the RMSE of regression \ll MAD range) is an acceptably small value, a specific sensor must still provide an acceptably precise mean value of field readings (that is, standard deviation of readings at multiple locations $<$ MAD range).

The ability to provide an acceptably precise mean value of field readings using a cost-effective number of access tubes or sensors in the soil is where some sensors are lacking (Table 2-3). In particular, the capacitance sensors appear to be very sensitive to variations in soil water content at scales smaller than the REV, and thus require many more access tubes to attain a precision equal to that attained with much fewer NMM or gravimetric samples. Another example is data from Australia showing that the standard deviation of profile water contents reported by the EnviroSCAN system was 12.36 cm compared with s of 0.93 cm for the NMM in the same flood irrigation basin (Evett et al., 2002b).

If no other information were available about soil water variability, sampling a field for profile water content would typically require many profiles to be sampled, either directly or using water content sensor(s) in access tubes. However, distribution of profile water content tends to be temporally stable in some fields (Vachaud et al., 1985; Villagra et al., 1995). This means that there are locations in the field where the profile water content is usually very representative of the mean for the field, or of the extremes (Fig. 2-8; Evett, 1989)). Irrigators recognize this when they observe the crop in a field for water stress or when they probe the soil for water content. For example, an irrigator may ignore drier crops at the edge of a field, or a low, wet corner of the field when

Table 2-3. Calculation of number of access tubes (N) needed to find the mean profile water storage in a field to a precision d (cm) at the $(1 - \alpha)$ probability level ($\mu_{\alpha/2}$ is the value of the standard normal distribution at $\alpha/2$) for a given field-measured standard deviation (s , cm) of profile storage calculated using Eq. [10]. Data are from ten access tubes for each device, spaced at 10-m intervals in transects that were 5-m apart.

Method	Soil condition	s cm	N	
			$\alpha = 0.05, \mu_{\alpha/2} = 1.96,$ $d = 1$ cm	$\alpha = 0.10, \mu_{\alpha/2} = 1.64, d$ $= 0.1$ cm
Diviner 2000†	Irrigated	1.31	6.6	464
	Dryland	2.42	22.5	1584
EnviroSCAN†	Irrigated	1.52	8.9	625
	Dryland	2.66	27.2	1914
Delta-T PR1/6†	Irrigated	2.72	28.4	2002
	Dryland	12.16	568.0	40006
Sentry 200AP†‡	Overall	3.78	54.9	3866
Trime T3	Irrigated	0.75	2.2	152
	Dryland	2.38	21.8	1533
Gravimetric by push tube	Irrigated	0.45	0.8	55
	Dryland	0.70	1.9	133
CPN 503DR NMM	Irrigated	0.15	0.1	6
	Dryland	0.27	0.3	20

† Capacitance type sensors

‡ Estimated from data of Evett and Steiner (1995)

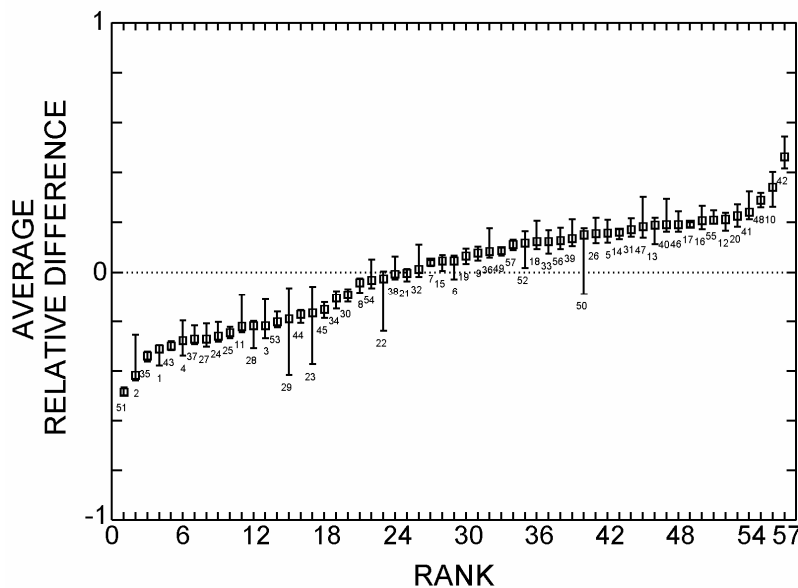


Fig. 2-8. Ranking of locations by their average relative difference from the field mean profile water content. Vertical bars indicate the range of values observed over the course of the experiment. Location 21 in particular was close to the mean profile water content at all times.

assessing the need to irrigate. The tendency is to make observations in places that show the mean behavior of the field. This is not an adequate way of choosing observation locations for a scientific experiment for which blocking, randomization, replication and other considerations are required for statistical validity. But, for irrigation management in production agriculture, the choosing of measurement locations on the basis of observed soil and plant properties that are representative of the field may be the most cost effective and efficient method. That said, the scheduling of irrigations on the basis of a single profile water content measurement in a field is prone to error. Also, there is strong evidence that actively growing vegetation can reduce or eliminate the temporal stability of water content, particularly in the root zone (Hupet and Vanclooster, 2002) and in fields with little topographic relief. A reasonable minimum for the NMM or gravimetric sampling is three to four profile water content measurements at locations chosen to be representative of the field (Tollner et al., 1991). For other methods, such as the capacitance sensors, that sense smaller volumes resulting in larger values of S , the number of profile measurements needed may be much greater.

SOIL WATER BALANCE

The soil water balance equation is usefully written for a control volume within which water storage, S , may change (Fig. 2-9). Changes in S (ΔS , often in mm) may be due to fluxes of water to the atmosphere via evaporation and plant transpiration (ET), lateral fluxes or vertical fluxes (F) into or out of the control volume, and infiltration (I). The fluxes and change in storage may be considered as instantaneous rates, or as integral values over some period of time. The sum of runoff and runoff is represented by R_o , which may be positive (more runoff than runoff) or negative. Infiltration is often

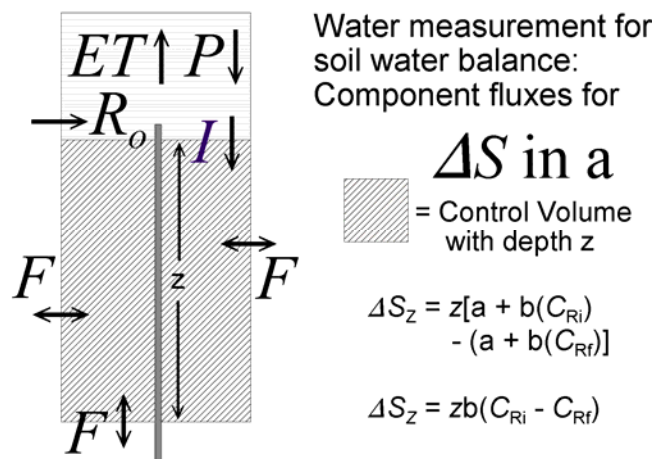


Fig. 2-9. Schematic depiction of the soil water balance defined by fluxes of water into and out of a control volume, and by changes of water storage (ΔS) between an initial time (subscript i) and a final time (subscript f). The water content in this simplified schematic is assumed uniform and is represented by the calibration equation: $\theta_v = a + b(C_R)$ where C_R is the measured response (frequency shift, count, etc.) and a and b are intercept and slope of a linear calibration equation. The change in storage in a control volume with defined profile depth, z , is ΔS_z . Fluxes of water are evapotranspiration (ET), precipitation (P), the sum of runoff and runoff (R_o), infiltration (I), and horizontal and vertical soil water fluxes (F).

considered as the sum of precipitation (including irrigation), P , and R_o , that is $I = P + R_o$, because precipitation and irrigation depths and runoff volumes are more easily measured than is infiltration. In Eq. [13], the sign convention for P , R_o , and F is that fluxes into or onto the surface of the control volume are positive, resulting in positive values of ΔS . In accordance with common practice, the sign convention for ET is that flux away from the control volume is positive.

$$\Delta S = -ET + P + R_o + F + \epsilon_{\Delta S} \quad [13]$$

When written to yield ΔS , the water balance equation is useful in forecasting of soil water content levels. Forecasts of soil water content are used in irrigation scheduling, agricultural management and policy making, and in weather forecasting. The error term, $\epsilon_{\Delta S}$, may be as large as the sum of the absolute values of errors in each of the component terms in the right-hand-side of Eq. [13]. Errors in ET estimation include errors in weather data (measurement errors or forecasting errors), and errors in calculation such as in reference ET or crop coefficient (K_C) values. Other errors are discussed below.

For determination of crop water use, the water balance equation may be written to solve for ET .

$$ET = -\Delta S + P - R_o + F + \epsilon_{ET} \quad [14]$$

The error term, ϵ_{ET} , is the sum of errors in each of the component terms in the right-hand-side of Eq. [14].

$$\epsilon_{ET} = \epsilon(\Delta S, P, R_o, F) \quad [15]$$

Sensing of soil water content throughout the depth, z , of the control volume, at the beginning and end of a time period, allows the computation of ΔS_z for that period. Shown in Fig. 2-9 is such a calculation in terms of a linear calibration equation for a soil water sensor. As with all soil water sensors, the sensor does not measure water content, but measures some property related to water content, shown here as a count, C_R , with the subscript i indicating the count at the beginning of the period and the subscript f denoting the end. Multiplying the depth of the control volume by the difference in water contents over the time period gives the change in storage in the control volume, ΔS_z for this simplified case, where the water content is assumed uniform throughout the control volume at any time. More realistically, water content and its change would vary with depth; and ΔS would be calculated for each layer ($j = 1$ to N), each of depth z_j , in which water content was sensed. The total value of ΔS_z would then be calculated as the sum of the values of $\Delta S_j \times z_j$. Also, the calibration equation may be nonlinear. In any case, the only coefficient of the calibration equation that does not affect the calculation of ΔS_z is the intercept term a . Errors in any other coefficient(s) will cause errors in ΔS_z . Non-uniform irrigation, as with microirrigation, furrow irrigation, or low energy precision applicator (LEPA) systems, greatly complicates the sensing and calculation of changes in soil water storage.

Typically, errors in estimated infiltration arise from errors in measurement of precipitation and runoff/runon. Precipitation rates may vary greatly over short distances (Plate. 2-1, top), leading to errors if rain gages are located even a few tens of meters from the site for which the water balance is calculated. Spatial variability of precipitation is a world-wide phenomenon (Plate 2-1, bottom).

While placement of rain gages close to the site is essential, the height of the rain gage above ground or vegetation surface is equally important because wind speed increases logarithmically with elevation above the surface (or above the zero plane displacement height in the case of vegetated or residue covered surfaces), leading to decreased catch for gages placed at greater elevations above the surface. Alternatively, a wind shield may be used to reduce wind velocity over the gage, in effect bringing the zero plane displacement height closer to the top of the gage (Environmental Protection Agency, 1990; Bigelow et al., 1990). In regions where snow makes important contributions to the soil water balance, gages may be heated to improve capture. Rain gage mountings on commercially available weather stations are frequently inappropriate and must be modified to improve accuracy of the catch (Fig. 2-11).

While standard U.S. Weather Bureau gages are reasonably accurate if well sited and shielded, they do not provide a record of rainfall over time. For rainfall over time, a tipping bucket rain gage is most widely used. Tipping bucket gages should be recalibrated periodically to reduce errors due to wear, dirt, etc. The concept of effective precipitation has relevance for irrigation scheduling and is discussed thoroughly by Dastane (1978).



Fig. 2-11. The shielded gage at lower left is heated so that snow and ice fall can be detected. As described at <http://205.156.54.206/asos/tipbuck.htm>, it is a tipping bucket gage. The gage at right is a standard USWB static gage. It and the wind shield at lower left are described at <http://www.nwstc.noaa.gov/METEOR/srg/rain8in.html>. The tipping bucket gage at center top is mounted on a weather station at 2-m height, inappropriate for accurate catch.

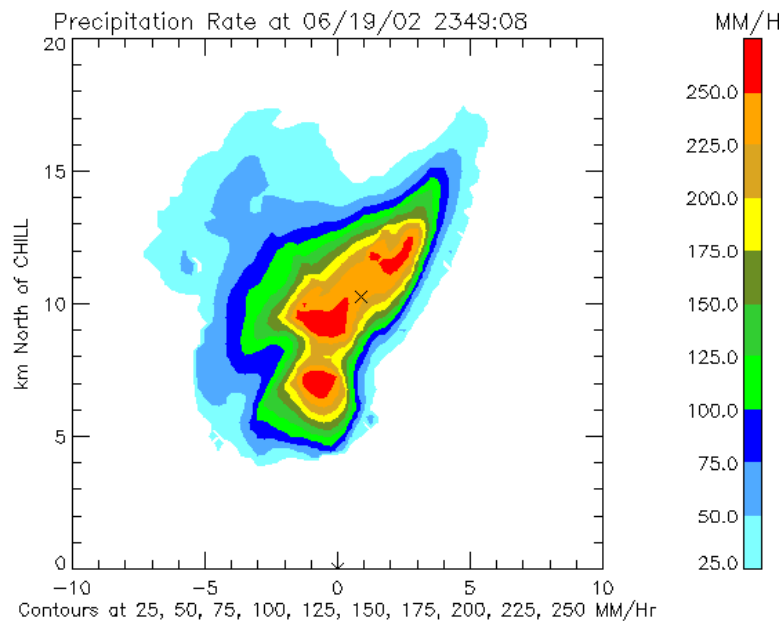


Plate 2-1. (Top) Contours of rainfall intensity in mm h^{-1} for a thunderstorm on 19 June 2002 near Greeley, Colorado as measured by the Colorado State University CHILL National Radar Facility. (http://chill.colostate.edu/pck_page/19jun2002/index.html). The horizontal axis is in km east (positive) and west (negative) from the radar site. (Bottom) A convective thunderstorm near Tashkent, Uzbekistan delivering rain to only a small portion of the landscape in 1999.

Runoff and runoff (R_o) are often not measured in water balance studies, which can lead to large errors. The H-flume illustrated in Fig. 2-12 measures runoff from a field of several hectares. This runoff measure is an integration of runoff from different parts of the field; but it is not directly applicable to any particular small area because it does not account for the spatial variability of infiltration. This is particularly true for fields irrigated by overland flow (furrow, graded border) for which the increased opportunity time for infiltration at the upper end of the field typically results in a reduction of infiltrated depth of water in the down slope direction such that $I \neq P + R_o$ for all places in the field for runoff measured at the outlet of the field. Approaches to control R_o include placing low earthen dikes around plots, placing metal borders around plots, and using furrow dikes to inhibit water movement along furrows. In some studies, furrow irrigation flows have been measured at the upper and lower ends of field plots using small flumes. This is painstaking and expensive work.

Lateral fluxes at the sides of the control volume or vertical flux at the bottom of the control volume can cause large errors in water balance calculations if not controlled or accurately estimated. Lateral fluxes can often be made negligible by careful choice of measurement location and size of plot. Weighing and non-weighing lysimeters control both lateral and vertical fluxes, and are discussed thoroughly in Allen et al. (1991). Drainage lysimeters are a type of non-weighing lysimeter that controls vertical and lateral fluxes. Such lysimeters typically have an upper edge that extends above the soil surface to control runoff and runoff. Measures of soil water content throughout the profile inside the lysimeter are then used to calculate the change in soil water storage. Studies in hill slope hydrology may encounter sizable lateral soil water fluxes that cannot be safely ignored. For field plot studies of agricultural crops, locations and plot sizes may be chosen to ensure insignificant



Fig. 2-12. An H flume, with stage recorder on left side, in operation.

rainfall exceeds crop water use or water tables are so near the surface as to contribute to crop water net lateral flux, even though soil water redistribution after irrigation can cause important amounts of lateral flux. This is one reason why the volume sensed is of importance in soil water measurements. Vertical flux errors can sometimes be controlled by measuring soil water content to depths well below that of rooting and of infiltration from irrigation. This becomes impractical in regions where rainfall exceeds crop water use or water tables are so near to the surface as to contribute to crop water use. For these cases, lysimeters may be used to provide measurements under conditions of constant vacuum drainage or to establish an internal water table that is regulated to match the water table in the field (e.g. Schneider et al., 1996).

There is no widely accepted method for measurement of soil water flux. Recently, heat pulse devices, analogous to the sap flow gages used to estimate plant transpiration, have been employed to estimate soil water flux (Hopmans, et al., 2002). A water flux meter utilizing a fiberglass wick and a tipping bucket to measure water flow has been proposed and demonstrated by Gee et al. (2000). However, the most common method of estimating vertical flux rates is to measure or estimate the soil matric potential at two depths, and then calculate the potential gradient between the depths for use in Eq. [6] to calculate the flux rate.

Vertical flux calculations involve several steps. One may use soil water potential measurements (discussed later) to establish the potential gradient at depth z_f from potentials above ($f+1$) and below ($f-1$).

$$\Delta\Psi_f = \Psi_{f+1} - \Psi_{f-1} \quad [16]$$

Or, one may infer the potential gradient from soil water content measurements and knowledge of the $\Psi(\theta)$ relationship (being always aware of errors due to hysteretic behavior of this relationship). Typically, the hydraulic conductivity K_f at z_f is inferred from $K(\theta)$ or $K(\Psi)$ functions; and then the flux rate, $q_f = -K_f(\Delta\Psi_f/\Delta z)$, is calculated, followed by the total flux, Q_f , over the period from time t_0 to time t_1 by integration.

$$Q_f = \int_{t_0}^{t_1} q_f \partial t \quad [17]$$

The reliability of estimates of Q_f is better if the $K(\theta)$ function has been measured for the soil layer in which Q_f is estimated. However, this is onerous to do; and these functions are typically estimated from pedotransfer functions using one of several levels of detail of the soil properties (e.g. the Rosetta computer program, <http://www.ars.usda.gov/Services/docs.htm?docid=8953>, Schaap et al., 2001). Use of a $K(\Psi)$ function, even if measured, will incur the error associated with ignoring hysteresis in the conversion of the measured water contents to values of Ψ . The errors associated with the use of pedotransfer functions to estimate Ψ may be very large. Perhaps the most useful application is in situations where it is possible to demonstrate that deep flux rates are negligible.

If Eq. [13] is to be solved for the change in soil water storage then ET or its components E and T must be measured or estimated. Evaporation from the soil surface may be measured directly

with microlysimeters, containers of soil small enough to fit between plants, and weighed at appropriate intervals to measure the mass loss due to evaporation during the intervening periods. Typically, microlysimeters are thin-walled cylinders of 7.5- to 10-cm diameter and 10- to 30-cm length, pressed vertically into the soil surface, removed with soil core intact, cleaned on the outside, capped on the bottom, weighed, and then returned to the hole made by extraction of the soil core or to another hole such that the soil surface inside the microlysimeter is at the same elevation as the field soil. Typical weighing times are sunrise and sunset or daily, although smaller time intervals have been used. Problems with microlysimeters include the difficulty of obtaining accurate weights in the field, divergence from the field soil condition due both to lack of active roots in the microlysimeter and to the lack of upward or downward water flux through the bottom cap, and divergence from the field energy balance due to poor design. Well designed microlysimeters will have cylinder walls of a relatively non-thermally conductive material (e.g. rigid polyvinyl chloride or other plastic) in order to inhibit vertical heat flux within the wall, and bottom caps of a thermally conductive substance such as metal in order to transmit heat between the microlysimeter and underlying soil (Evetts et al., 1995; Todd et al., 2000). Water content of short lysimeters (10 cm) quickly diverges from that of the surrounding field soil so that such lysimeters should be replaced daily. Longer lysimeters may be used for multiple days before replacement, up to nine days for 30-cm long lysimeters (Evetts et al. 1995). Evaporation from plant surfaces (other than transpiration) can be very difficult to measure. There is good evidence that evaporation from wetted plant surfaces during the daytime will cause a corresponding decrease in the transpiration rate (Tolk et al., 1995).

Measurement of transpiration has been approached in several different ways, including comparative lysimeters with the soil surface sealed in one lysimeter to suppress the E component of ET , clear glass or plastic chambers placed over the plants with sensors inside to measure the rate of increase of humidity, and sap flow gages (Baker and Van Bavel, 1987; Ham et al., 1990). Sap flow gages are based on either heat balance (Sakuratani, 1981) or heat pulse (Cohen et al., 1981) principles. Calibration of sap flow methods to give accurate estimates of transpiration is a difficult and not completely resolved issue.

Evapotranspiration (ET) is widely estimated using models that vary in complexity from detailed mechanistic computer models capable of modeling the surface energy balance and soil water and heat fluxes (eg. ENWATBAL; Evetts and Lascano, 1993) to more empirical approaches using a time-varying crop coefficient (K_c) multiplied by a daily reference evapotranspiration (ET_o). The reference ET is often estimated from the Penman-Monteith combination energy balance and aerodynamic equation, which uses solar radiation, wind speed, relative humidity and air temperature as input data (Allen et al., 1998, 2005).

MEASUREMENT OF SOIL WATER CONTENT

Direct Methods

Direct measurement of soil water content is the reference method against which all other methods are compared. Direct measurement involves taking a sample of soil, either of known or unknown volume, protecting the sample from evaporative loss of water, determining the mass of the sample, drying the sample at a specified temperature for a specified time (typically 105°C for 24 h or until mass stabilizes), and determining the mass of the dried sample. If the sample volume is unknown then Eq. [4] is used to calculate the gravimetric water content, θ_g ; and if the sample volume is known then Eq. [3] is used to calculate the volumetric water content, θ_v . Because it is time consuming, frequently difficult (particularly at depth), and destructive, direct sampling is often not the method of choice for soil water determination. However, all other methods are indirect; that is, they measure properties other than water content; and so must be calibrated against direct measurements in order to create a calibration equation relating the measured property to the water content.

There are two main methods of obtaining volumetric soil samples. One method is to use a metal cylinder, scoop or other device of known volume to obtain a sample – a so-called undisturbed core. This common method is subject to errors arising from sample compression or dilation. Some of the available sampling equipment is ill-designed to avoid compression. In particular, soil compression is likely using soil core samplers that employ metal cylinders inside a larger, cylindrical sampling body with a beveled cutting edge. Compression is due to the large cross-sectional area of the cutting edge normal to the axis of insertion. Compression can usually be avoided by using a thin-walled cylinder with an acutely beveled cutting edge. To reduce friction between the soil core and sampler wall, the cylinder should be machined behind the cutting edge to have a larger inside diameter than that of the cutting edge. Sample rings or cylinders are often cut to length to provide a known volume. The cross-sectional area of the cylinder wall should not exceed 5% of the cross-sectional area of the soil core obtained. Thus, we want a sampler for which the inside radius $r \geq 0.975 R$, the outside radius (Hignett and Evett, 2002) (Fig. 2-12). Another standard for coring rings suggests that

$$\frac{A_{wall}}{A_{core}} = (D_e^2 - D_i^2) / D_i^2 < 0.1 \quad [18]$$

where A_{wall} is the cross-sectional area of the cylinder wall, and A_{core} is the cross-sectional area of the soil core, and D_e and D_i are the external and internal diameters, respectively (Fig. 2-12; American Society for Testing and Materials, 1999). Equation [18] allows a slightly thicker tube wall ($r \geq 0.95 R$).

The Madera probe (Fig. 2-13) was developed by the USDA for neutron probe calibration, and has some of the qualities of a good volumetric sampler. It is constructed of thin-walled stainless

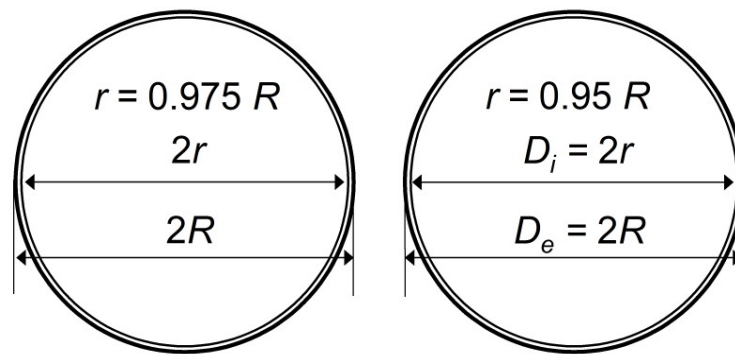


Fig. 2-12. Schematics depicting sampling cylinder cutting face relative inside and outside diameters that make for a small facial cutting area relative to the cross-sectional area of the sample, minimizing compaction. (Left: Hignett and Evett, 2002; Right: ASTM, 1999)



Figure 2-13. The Madera probe design was adapted to a larger diameter tubing to give an 80-cm³ sample volume at the IAEA Laboratory, Seibersdorf, Austria. Shown are the sampling tube with slots for cutting the sample to length, the two spatulas used to cut and retain the sample core, and a metal rod used to turn the sampler after it is inserted into the soil, thus breaking the soil core at the cutting face.

steel tubing, has a sharply beveled cutting edge, and has an inside diameter for most of its length that is larger than the diameter of the cutting edge. The latter characteristic reduces sample compression caused by friction between the soil core and the probe inside wall. The Madera probe has a bayonet connection on one end so that it can be attached to a shaft and used to obtain samples at the bottom of an augered hole. However, if the probe is used in this way, it is easy to compress the sample, shatter the sample, or sample loose material that has fallen to the bottom of the augered hole, all

without being aware of a problem. Better quality control results from inserting the Madera probe into the soil from the side of a pit, or vertically into the top of a soil layer, so that the soil inside the probe can be observed after insertion. If the soil inside the probe is at the same distance from the proximal end of the sampling tube as is the soil outside, then it is clear that sample compression did not occur. If the soil inside shatters during insertion, which would cause the bulk density to decrease (dilation), this too can be clearly observed. Compressed or shattered samples can then be discarded and replacements taken. The Madera probe differs from other designs mainly in that it has two slots that allow spatulas to be used to cut the soil core to a specific length, resulting in a 60 cm^3 sample. Soil in the probe outside of the section enclosed by the spatulas is removed, and the remaining 60 cm^3 volume is transferred to a soil can or bag for weighing. Two advantages ensue, i) hundreds of volumetric samples can be taken without having a sampling cylinder for each, ii) the method is much faster than using sampling cylinders and cutting the soil flush with each end of the cylinder to define the volume sampled.

Coring techniques may be difficult or impossible to use in dry, hard, or stony soils. Because of the difficulty of determining if any compression or shattering occurred, it is not recommended to sample in auger holes where the sampler may be out of sight. Hydraulically or manually pushed long, cylindrical probes may be used for deep sampling without trenching; but sample compaction is common. With care, long cores may be used to obtain volumetric samples by extracting the core intact from the probe tube into a tray and sectioning into sub samples of length appropriate to the study. However, values of water content thus obtained tend to be more variable than water contents obtained with shorter cores for which control over compaction is easier.

The other main method of volumetric sampling is to excavate a sample and measure the volume of the hole made by excavation. This is not commonly done, but is the only appropriate method for soils that are so stony or hard that undisturbed samples cannot otherwise be obtained. Several methods exist for measuring the volume of the excavation. The device shown in Fig. 2-14 (top) consists of a guide plate that is fixed in place on the soil surface, and a volume measurement device that is fitted with a graduated glass cylinder and an air pump. A rubber balloon is attached to the bottom of the cylinder. To use the device, the cylinder is partially filled with water; the guide plate is fixed in position over the soil surface; and an initial volume measurement is made with air pressure applied to the top of the cylinder so that the rubber balloon is forced to occupy all of the volume below. The cylinder and balloon are removed, leaving the plate in place, and a soil sample is excavated and saved, after which a second volume measurement is made. The difference between the two volumes is the volume of the excavated soil.

A similar method uses free water (Fig. 2-14, bottom; Grossman and Reinsch, 2002). The guide plate is placed over the foam ring and the threaded rods are forced into the soil through the three holes in the plate. The wing nuts on the rod are used to level the plate while forcing it firmly into contact with the foam ring. A thin plastic sheet is placed in the hole in the guide plate and filled with a measured volume of water. The hook gage is used to find the height of the water. After the soil has been excavated, the plastic sheet is again put in place and filled with water up to the point of the hook gage. The difference in the two volumes of water is the volume of the excavation.

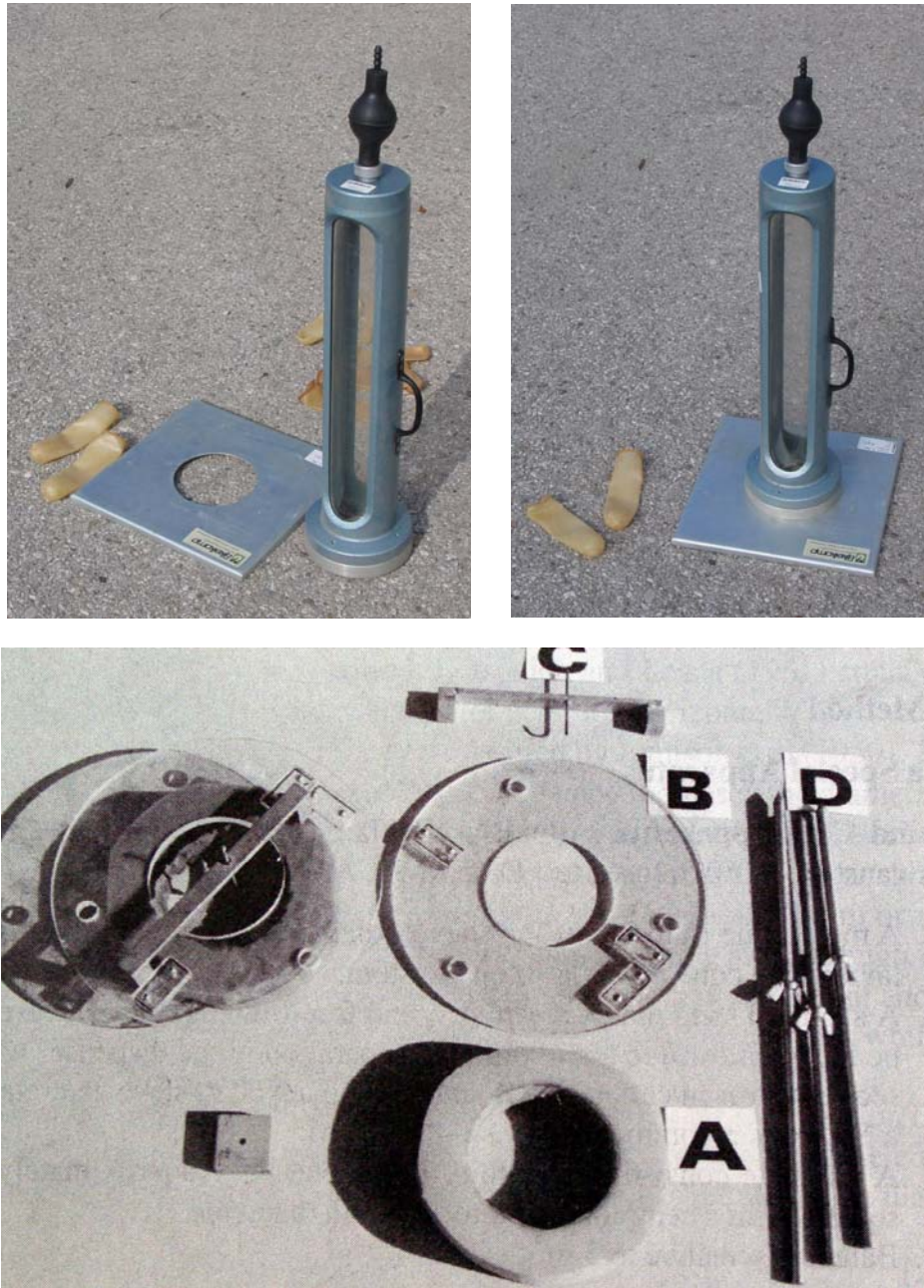


Fig. 2-14. Equipment for the balloon method for measuring excavation volume (top) includes a guide plate, balloons, and a volumetric cylinder that fits on the guide plate. Equipment for the compliant cavity method (bottom) includes a flexible foam ring (A), a guide plate (B), a hook gage (C), and threaded rods that are forced into the soil and which serve to level the guide plate while pushing it firmly in contact with the foam ring (Photo from Grossman and Reinsch, 2002).

For either excavation method, having determined the volume of the excavated soil, its volumetric water content is calculated by dividing the volume of the water lost on oven drying by the volume of the sample excavated. With care, the excavation methods can be accurate, the chief impediments being the difficulty in maintaining the soil left in the hole after the excavation in a state as similar to its original state as possible, and the difficulty of obtaining the sample rapidly enough to avoid evaporative loss of water. The characteristics of the soil being measured largely determine the success of the method.

For any of the direct methods, sample size relative to the REV is a concern. Soil structure, cracking, and other sources of macroporosity may influence the REV so that several samples may be needed to obtain a good mean value. This consideration also applies to the volume of soil sensed by indirect methods. The use of the data is also to be considered. For example, Evett and Steiner (1995) found that four Madera probe samples (volume of 60 cm³) adequately represented the volume sampled by the neutron moisture meter, but may have been taken outside the volume sampled by a capacitance type sensor from the same access tube. Although direct methods are the standard against which indirect methods are compared, there are many sources of error, including compression or dilation of the soil during sampling, possible loss of water before samples are weighed, loss of chemically bound water or volatilization of soil liquids or solids other than water during drying, etc.

Lastly, more subjective methods of estimating soil water content can be useful in experienced hands. The immediate feel and appearance of samples obtained in the field may be used to judge the remaining plant available water in a soil, at least for broad texture classes such as coarse, moderately coarse, medium, and fine textured soils (USDA-NRCS, 2001). Also, the soil probe may be used to determine the depth of the wetter layer, and so offers a subjective look at the amount of soil water storage (Robinson, 2003). The soil probe consists of a metal rod (e.g. 12-mm diameter) at the bottom end of which a metal ball has been attached. A T-handle is attached to the top end of the rod to facilitate pushing the rod into the soil. The metal ball has a slightly larger diameter than the rod, so that resistance to pushing is mostly due to the force required to move the ball through the soil, with little friction developed between the rod and soil. Since soil strength is inversely related to soil water content, depth of irrigation or rainfall infiltration can be assessed quickly in many points in the field.

Indirect Methods

Indirect methods provide estimates of soil water content based on measurements of soil properties that are assumed to be correlated with water content. However, there are many potential interferences, including temperature, bulk electrical conductivity (whether from salinity or clay type and content), bulk density, soil minerals and organic matter content, etc. For best accuracy, all indirect methods must be calibrated for the specific soil under study. Indirect methods can be divided into those employing neutron thermalization, those employing gamma ray attenuation, those related to soil thermal properties, and those employing electromagnetic (EM) methods to measure a signal frequency, an electronic pulse travel time, or power loss of a signal. The EM methods all depend on the effect that water content has on the dielectric properties of soil. A fifth class of sensors, which measure the electrical resistance of a porous body in hydraulic equilibrium with the soil, is more commonly calibrated vs. soil water matric potential. The most important methods will

be described here, but many more sensors are available than can be mentioned here. A wider compendium of moisture sensing technology with comments on applicability for irrigation scheduling was published by Charlesworth (2005).

Neutron Thermalization

The neutron moisture meter (NMM) consists of a source of fast neutrons (mean energy of 5 MeV) and a detector of slow neutrons (~ 0.025 eV or 300°K). High energy (fast) neutrons emitted from the source ($\sim 10^9$ /s) are either slowed through repeated collisions with the nuclei of atoms in the soil (scattering and thermalization), or are absorbed by those nuclei. A small fraction of scattered neutrons are deflected back to the detector. Of these, an even smaller fraction ($\sim 10^3$ /s) are slowed to thermal (room temperature) energy levels and can be detected. The most common atoms in soil (aluminum and silicon) scatter neutrons with little energy loss because they have much greater mass than a neutron. However, if a neutron strikes a hydrogen nucleus its energy is halved, on average, because the mass of the hydrogen nucleus is the same as that of the neutron. On average, 19 collisions with hydrogen are required to thermalize a neutron. Carbon, nitrogen, and oxygen are also relatively efficient as neutron thermalizers (about 120, 140 and 150 collisions, respectively, needed to bring a neutron to ambient energy level). On the time scales of common interest in irrigation research and management, changes in soil carbon and nitrogen content are minor and have little effect on the concentration of thermal neutrons. Also, on these time scales, changes in soil hydrogen and oxygen content occur mainly due to changes in soil water content. Thus, the concentration of thermal neutrons is most affected by changes in water content; and volumetric water content can be accurately and precisely related to the count of thermal neutrons through empirical calibration. Soil density has a small but measurable effect on the concentration of thermalized neutrons around the detector.

Because hydrogen and carbon effectively thermalize neutrons, the organic matter content of soil affects the calibration. Also, organic matter and most clays contain important amounts of hydrogen, some not in the form of water, that may not be driven off by heating to 105°C. So, separate calibrations are often required for soil layers that differ in organic matter or clay content from layers above or below. In arid or semi-arid zones, many soils have layers rich in CaCO_3 and CaSO_4 that require separate calibration. Atoms that absorb neutrons efficiently include boron, cadmium, chlorine, iron, fluorine, lithium and potassium. Although these usually comprise a small fraction of soil material, soils or soil horizons containing large or fluctuating amounts of such elements require separate calibrations or adjustments in data interpretation. For example, soils high in iron, such as Oxisols or soils rich in magnetite, typically require separate calibration, as may soils high in chloride salts. In a few US soils, boron is present in sufficient quantity to affect calibration; but boron toxicity prevents most irrigated agriculture in such soils.

Neutron moisture meter (NMM) equipment comes in two forms: i) a profiling meter with a source - detector pair assembled into a cylindrical probe that is lowered into a hole in the soil, and ii) a flat-based meter that is placed on the soil surface with the source and detector fixed at separate locations inside the base of the meter. The volume measured by the surface meter is roughly hemispherical and extends into the soil for a distance that decreases as soil water content and soil density increase, and which varies from ~ 0.15 m in wet soil to ~ 0.3 m in dry soil. The precision is

less than can be attained with a profiling meter; and it suffers even more when soil water content changes greatly with depth near the surface, a common occurrence. Good precision has been reported under fairly stringent conditions including: i) flattening the surface to fit the meter bottom with no air gaps, ii) marking the measurement site so that the meter can be repeatedly placed in identical position, and iii) using a neutron absorber shield made of cadmium around the meter (except for the bottom) to reduce effects of surrounding vegetation. However, the strong depth dependency of calibration coefficients and the inability to accurately estimate the depth of reading can lead to great uncertainty as to the accuracy of measurements (Nakayama and Allen, 1990).

More commonly used in irrigation work is the profiling NMM, which is operated at user-chosen depths in the soil (Fig. 2-15). A cylindrical access tube is used to line the hole, protecting the probe and ensuring a constant hole diameter. The probe is connected to a counter, data storage and display module by a cable that allows the probe to be lowered into the tube and stopped at intervals to measure the thermal neutron concentration. Attainable depths are limited only by cable length and have exceeded 100 m. Common probe diameters are 38 and 48 mm. When not in use, the probe is locked in the instrument shield, which comprises a block of high-density polyethylene or similar plastic, and which is commonly attached to the readout and control unit. In the probe, the source is either directly beneath the detector, or is centered around or on one side of it. The relative position of the source and detector affects the calibration; but for modern meters, source-detector geometry has little effect on the attainable precision. In modern meters the source is a mixture of americium-241 and beryllium with an activity ranging from 0.4 to 1.9 gigabecquerels. The nuclear reaction is $(^9\text{Be}(\alpha, n)^{12}\text{C})$ in which ^{241}Am emits an alpha particle that is absorbed by a Be atom, which then produces ^{12}C and a fast neutron. The measurement volume is approximately a sphere. For a soil of specified volumetric water content (θ_v , $\text{m}^3 \text{m}^{-3}$), about 95% of the measured slow neutrons are from a sphere of radius R (cm).

$$R = 15(\theta_v)^{-1/3} \quad [19]$$

Recently, Evett et al. (2003) showed that the axial distance of influence (A , cm) for a modern NMM may be smaller than that indicated by Eq. [19]

$$A = 9(\theta_v)^{-1/3} \quad [20]$$

Access tubing materials that have been used successfully include stainless steel, mild steel, polyvinyl chloride (PVC), polycarbonate, and polyethylene plastics, and aluminum. The hydrogen in plastics affects calibration as does the neutron absorber chlorine in PVC tubes. Aluminum is nearly transparent to neutrons, while the neutron absorber iron affects calibration in steel tubes. Thus, it is important that a NMM be calibrated in the same tubing as will be used in the field. Although calibration precision decreases slightly if plastic tubes are used, precision and accuracy are much more dependent on the tube installation and calibration methods employed than on tube material. Thus, choice of access tube should be based on considerations of cost, availability, corrosion resistance (if an issue) and uniformity, not on the material's minor effect on calibration. Recommendations for installation of access tubes are given in Hignett and Evett (2002).

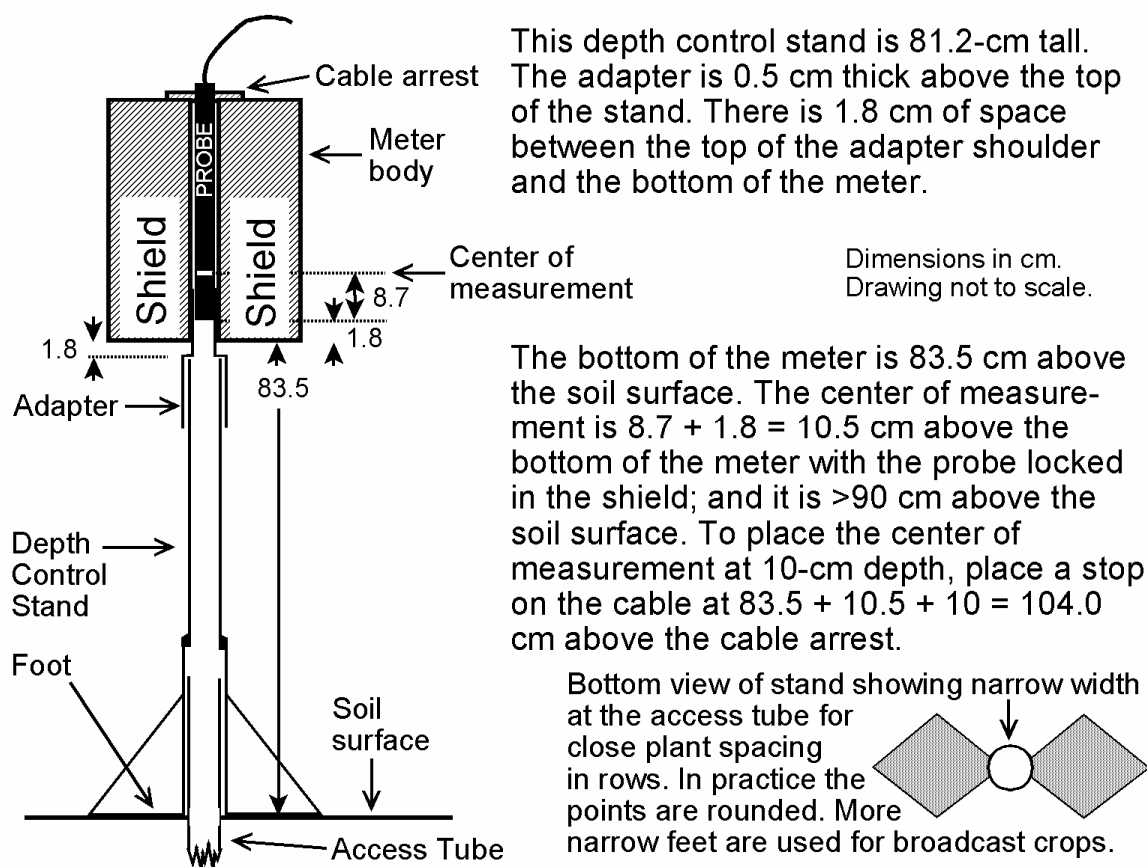


Fig. 2-15. Cartoon of a neutron moisture meter resting on a depth control stand, which provides a constant and known elevation of the meter above the soil surface. Guidance for adjusting cable stops to obtain measurements at desired depths below the surface is given in the figure.

It is common practice to place the NMM on top of the access tube near the soil surface before lowering the probe for readings. This practice is not recommended for two reasons. First, when the NMM is placed near the soil surface, the shield in the meter body may influence near-surface counts to a degree that depends strongly on the height of the meter above the soil. Second, in field use, the height of access tubes above the soil is likely to change with tillage, rainfall-induced settling, erosion or deposition, or other factors, resulting in an equivalent change in the depth of probe placement. For readings above 0.3-m depth, the depth of the probe will strongly influence the reading and the calibration equation due to loss of neutrons to the atmosphere.

These problems are addressed by using a depth control stand (Evetts et al., 2003). This device comprises a length of access tube fixed to a 0.2-m length of slightly larger tubing that is in turn supported by a foot resting directly on the soil (Fig. 2-15). The larger diameter of the lower length of tubing allows it to be slipped over the top of an access tube so that the foot rests on the soil surface. The NMM is then placed on top of the stand, and the probe lowered through the stand and access tube to a depth in the soil. Cable stops are arranged to achieve the desired depth placement of the

probe. This maintains the reading depth at an exact distance relative to the soil surface. Because standard counts taken with the meter too close to the soil surface may vary with the water content of the soil, the stand described is tall enough to be suitable for taking standard counts with the NMM mounted on the stand and the probe locked in the meter shield.

Manufacturers' calibration equations are seldom useful for soil water determination. Calibration of NMMs involves correlating measured count ratio values with independently determined volumetric water contents, θ_v ($\text{m}^3 \text{m}^{-3}$). For modern meters and the normal range of values of soil water content, the calibration is linear

$$\theta_v = a + bC_R \quad [21]$$

where a and b are the calibration coefficients as determined by linear regression, and C_R is the count ratio defined as

$$C_R = x/x_s \quad [22]$$

where x is the count in the measured material and x_s is a standard count taken with the probe within a standard and reproducible material. Count ratio values are used because the source activity and thus counts will decline over time, and because the detector efficiency is slightly temperature dependent. Recommendations for taking standard counts are given in Hignett and Evett (2002), as are recommendations for field calibration using the wet site - dry site method of Evett and Steiner (1995). Careful field calibrations done using the wet site - dry site method and the depth control stand should attain root mean squared errors $< 0.01 \text{ m}^3 \text{m}^{-3}$ and r^2 values greater than 0.9, even for depths near the surface (e.g. 10 cm in Evett et al., 2003). As with any indirect method, calibration involves obtaining independent volumetric water content values by direct sampling. For each depth of neutron probe reading, four or more samples should be taken such that the mean sample value provides a representative value integrating the volume of soil sampled by the neutron probe.

Safety concerns relate to radiation safety and to back and knee strains incurred during repeated bending and kneeling to operate meters placed on access tubes. The depth control stand described above allows users to work standing up, and has virtually eliminated physical injuries where it is used. Due to the low levels of radioactivity involved, the principle of reducing exposure to as low as reasonably achievable (ALARA) guides most radiation safety rules. Users may lower radiation received by increasing distance from the meter, decreasing time spent near the meter, and increasing shielding. The probe should always be locked into the shield except when it is lowered into an access tube. Users should be made aware that the source emits radiation at all times, even when the meter is turned off and batteries removed. Guidelines for ALARA use of the NMM are found in Evett (2000a). The USDA Radiation Safety Staff maintains an Internet site of useful information on radiation safety and hazardous materials transport (<http://www.rss.usda.gov/>) as does the International Atomic Energy Agency (<http://www.iaea.org/>).

Due to regulation, the method is not usable for automatic measurements. Due to its large measurement volume, the method is inappropriate where detailed vertical definition is required. This can be particularly important near the surface where water content often changes rapidly with depth. In such cases, the NMM can be used for deeper measurements in conjunction with time domain

reflectometry (TDR) measurement of the near-surface soil water content (Evetts et al., 1993; Musters and Bouten, 2000). The time and effort required to install access tubes and calibrate for each soil type is nontrivial. There is also a substantial cost for the equipment and for necessary training and licenses to handle and transport radioactive materials.

Despite its drawbacks, the NMM remains the best available tool for repeated non-destructive measurement of soil profile volumetric water content (IAEA, 2000) in a research setting because it can be field calibrated with high accuracy, works successfully to depths not easily attained with other methods, and works well in stony soils and cracking clays in which other methods work poorly. Also, the large measurement volume means that fewer replicates are required than for other methods to produce a given precision, that soil disturbance during tube installation has minimal effect on results (unlike electronic sensor methods), and that field calibration is successful because volumetric soil samples can be obtained from within the volume measured by the probe at each depth (unlike electronic methods used in access tubes that have much smaller measurement volumes). The technology is mature with a wide literature base describing applications and problems (e.g., Greacen et al., 1981; Hignett and Evett, 2002). The NMM is widely and successfully used for soil water balance and irrigation scheduling in research, and is used successfully by consultants to provide irrigation scheduling advice to farm managers. However, its on-farm use is limited due to the licensing, training, and administrative costs associated with the radioactive source.

Time Domain Reflectometry

In the time domain reflectometry (TDR) method, a very fast rise time (approx. 200 ps) step voltage increase is injected into a waveguide (usually coaxial cable) that carries the pulse to a probe placed in the soil or other porous medium (Fig. 2-16, left). The velocity of the pulse in the probe is measured and related to soil water content, with smaller velocities indicating wetter soils. In a typical field installation, probes are connected to the instrument through a network of coaxial cables and multiplexers. Part of the TDR instrument (e.g. Tektronix model 1502B/C) provides the voltage step and another part, essentially a fast oscilloscope, captures the reflected waveform. The oscilloscope can capture waveforms that represent all, or any part of, the waveguide (this includes cables, multiplexers and probes), beginning from a location that is actually inside the instrument. For example, Fig. 2-16 (left) shows a waveform that represents the waveguide from a point inside the cable tester, before the step pulse is injected, and extending beyond the pulse injection point to a point that is 4.2 m from the cable tester. The relative height of the waveform represents a voltage, which is proportional to the impedance of the waveguide. Although most TDR instruments display the horizontal axis in units of length (a holdover from the primary use of these instruments in detecting the location of cable faults), the horizontal axis is actually measured in units of time.

The TDR method relies on graphical interpretation of the waveform reflected from that part of the waveguide that is the probe (Fig. 2-16, right). An example of waveform interpretation for a 20 cm TDR probe in wet sand shows how tangent lines are fitted to several waveform features (Fig. 2-17). Intersections of the tangent lines define times related to: i) the separation of the outer braid from the coaxial cable so that it can be connected to one of the probe rods in the handle, $t_{l.bis}$; ii) the time when the pulse exits the handle and enters the soil, t_l ; and iii) the time when the pulse reaches the

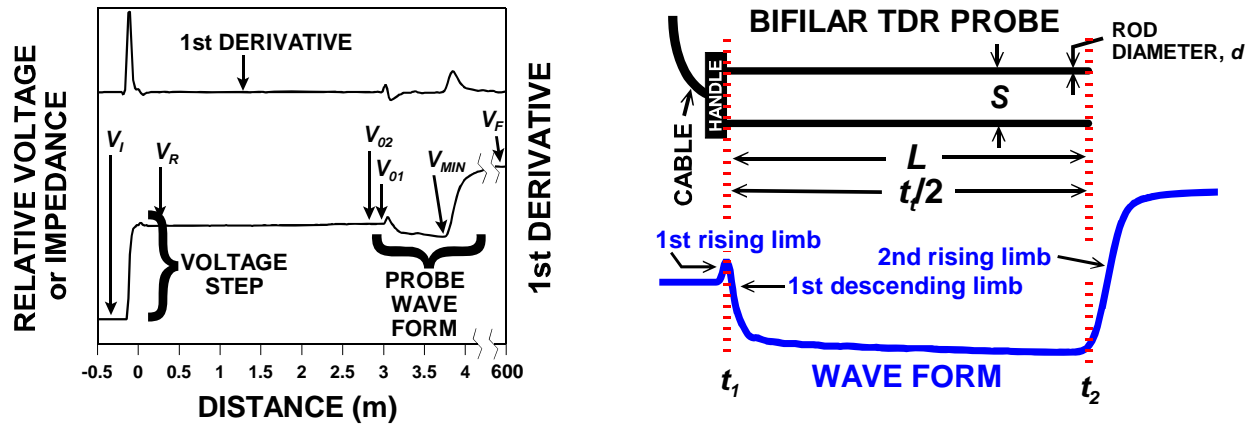


Figure 2-16. (Left) Plot of waveform and its first derivative from a Tektronix 1502C TDR cable tester set to begin at -0.5 m (inside the cable tester). The voltage step is shown to be injected just before the zero point (cable connector on instrument front panel). The propagation velocity factor, v_p , was set to 0.67 because electricity travels at 0.67 of the speed of light in the coaxial cable. At 3 m from the instrument, a TDR probe is connected to the cable. The relative voltage levels, V_I , V_R , etc. are used in calculations of the bulk electrical conductivity of the medium in which the probe is inserted. Inflections in the first derivative of the waveform are used in software or firmware to help determine pulse travel times, which, for the probe, are proportional to water content. (Right) Schematic of a typical bifilar TDR probe and the corresponding waveform, illustrating probe rod length (L), one-way travel time ($t_t/2$), rod spacing (s), and rod diameter (d).

ends of the probe rods, t_2 . The time taken for the step voltage pulse to travel along the probe rods, $t_t = t_2 - t_1$, is related to the pulse propagation velocity, v (m/s), as

$$t_t = 2L/v \quad [23]$$

where L is the length of the rods in the soil (Fig. 2-16, right), and the factor 2 signifies two-way travel. The pulse propagation velocity is influenced by the dielectric permittivity, ϵ , of the medium between the probe rods according to

$$v/c_0 = (\epsilon\mu)^{-0.5} \quad [24]$$

where c_0 is the speed of light in a vacuum, and μ is the magnetic permeability of the medium. For a TDR probe in a soil, the dielectric medium between the probe rods is a complex mixture of air, water and soil particles that exhibits a variable apparent permittivity, ϵ_a . Water is the largest determinant of permittivity in soils. It has a permittivity of approximately 80, whereas the permittivity of soil minerals is approximately 5 (Robinson and Friedman, 2003); the permittivity of organic matter is likewise low; and the permittivity of air is unity. Also, soil water is the only rapidly changing determinant of ϵ_a . Thus, soil water content can be related to estimates of ϵ_a using a calibration. The fact that frozen water has a low permittivity impedes accurate determination of frozen water content, but allows the use of TDR for investigations of freezing depth and extent (Spaans and Baker, 1995).

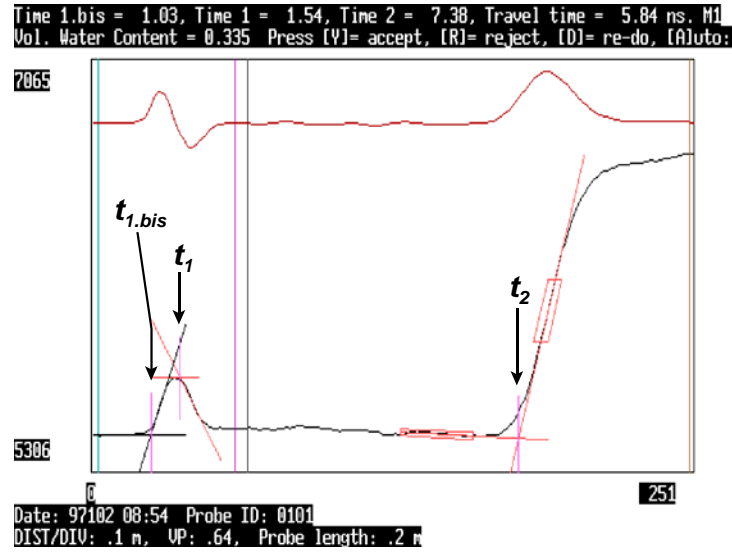


Figure 2-17. Example of graphical interpretation of a waveform from a probe in wet sand using the TACQ computer program (Evelt, 2000b, 2000c). Vertical lines denoting times $t_{1.bis}$, t_1 , and t_2 have been marked by arrows and labels. Time t_1 is the time at which the TDR pulse exits the probe handle and enters the stainless steel rods in the soil. Time t_2 is the time at which the pulse is reflected at the distal ends of the rods in the soil. Time $t_{1.bis}$ is the time at which the pulse is reflected from the point in the probe handle where the outer braid of the coaxial cable is separated and connected to the two outer rods of the trifilar probe. The first peak in the waveform occurs just before t_1 and just after $t_{1.bis}$. A horizontal line, drawn tangent to the waveform base line at the far left, intersects with a line drawn tangent to the first rising limb of the waveform to define $t_{1.bis}$. A horizontal line drawn tangent to the peak intersects with a line drawn tangent to the descending waveform after the peak to define t_1 . Alternatively, t_1 may be defined as the value of $t_{1.bis}$ plus the (independently determined) transit time of the TDR pulse in the handle. This alternative definition becomes essential in dry soil and is the preferred method of interpretation. Time t_2 is defined by the intersection of a line fitted to the waveform before t_2 , and a line fitted to the second rising limb of the waveform after t_2 . The difference $(t_2 - t_1)$ is the travel time, t_t , from which the water content is calculated using Eqs. [25-26]. The width of the waveform window in this example is 1 m, or 5.2 ns with the cable tester set to a propagation velocity factor $v_p = 0.64$.

Substituting ε_a and Eq. [23] into Eq. [24], and assuming $\mu = 1$, one sees that ε_a may be determined for a probe of known length, L , by measuring t_t

$$\varepsilon_a = [c_0 t_t / (2L)]^2 \quad [25]$$

Topp et al. (1980) found that a single polynomial function described the relationship between volumetric water content, θ_v , and values of ε_a determined from Eq. [25] for four mineral soils.

$$\theta_v = (-530 + 292\varepsilon_a - 5.5\varepsilon_a^2 + 0.043\varepsilon_a^3) / 10^4 \quad [26]$$

Since 1980, other researchers have noted that the quantity $[t_t / (2L)]$ in Eq. [25] is quadratic, and have shown that the relationship between θ_v and $t_t / (2L)$ is practically linear (e.g., Ledieu et al., 1986; Yu

et al., 1997). Several attempts have been made to predict ϵ_a of soils from theoretical considerations using dielectric mixing models that consider the volumetric proportions of soil mineral, organic, water, and air constituents, as well as soil mineralogy and particle shape and packing considerations (Dirksen and Dasberg, 1993; Friedman, 1998). Success could lead to a more universal calibration, but has been elusive (White et al., 1994). Equation [26] and like empirical calibrations for specific soils (particularly electrically lossy soils including clays with high charge, and organic soils) are still considered to be the accepted standards.

For most soils, excluding those very high in organic matter (OM>10%) or containing large amounts of high surface area clays, the TDR method provides water content in the range from 0 to $0.5 \text{ m}^3 \text{ m}^{-3}$ with accuracy better than $0.02 \text{ m}^3 \text{ m}^{-3}$ without calibration. With calibration, accuracy of better than $0.01 \text{ m}^3 \text{ m}^{-3}$ for a specific soil is attainable. Repeatability is excellent, with standard deviations of measurement ranging from $0.0006 \text{ m}^3 \text{ m}^{-3}$ (Evet, 1998) to $0.003 \text{ m}^3 \text{ m}^{-3}$ (Herkelrath et al., 1991).

The apparent permittivity in Eq. [25] is composed of a real part, ϵ' , and an imaginary part that includes an increase in permittivity due to relaxation losses, $\epsilon''_{\text{relax}}$, and the ratio of the electrical conductivity, σ_{dc} , of the medium to the measurement frequency, ω :

$$\epsilon_a = \frac{\mu\epsilon'}{2} \left(1 + \left\{ 1 + \left[\left(\epsilon''_{\text{relax}} + \frac{\sigma_{\text{dc}}}{\omega\epsilon_0} \right) / \epsilon' \right]^2 \right\}^{0.5} \right) \quad [27]$$

where ϵ_0 is the permittivity of free space. The real part of the permittivity is mostly related to the bulk water in the soil. However, the apparent permittivity, to which all EM sensors respond, increases with relaxation losses and with the bulk electrical conductivity ($\text{BEC} = \sigma_a$) of the soil. Also as measurement frequency decreases, the effect of σ_a increases. Disregarding relaxation effects and including effects of measurement frequency and σ_a , Evett et al. (2005) produced a calibration equation for conventional TDR in three soils:

$$\theta_v = -0.182 + 0.1271[c_{otl}/(2L)] - 0.004933[\sigma_a/(2\pi f_{vi}\epsilon_0)]^{0.5} \quad [28]$$

where the effective measurement frequency, f_{vi} , and the bulk electrical conductivity were both determined from the TDR waveform. This calibration was accurate to $0.01 \text{ m}^3 \text{ m}^{-3}$ for three soils varying in total clay content from 17 to 48% and made temperature dependency negligible.

Probe lengths reported in the literature range from 0.05 to 1.5 m. Accuracy and precision are reduced for very short probes due to problems with waveform interpretation and the limits of TDR instrument time base resolution. As probe length increases, signal loss increases, so that waveforms from long probes may not be interpretable. Common probe lengths of 20 and 30 cm reflect a compromise to ensure high-resolution measurements while not losing the waveform due to signal loss. Probe rod spacing, s , and rod diameter, d , may vary also, so long as $d/s \leq 0.1$ (Fig. 2-16, right) (Knight, 1992). As d/s becomes much smaller than 0.1, the volume of soil sensed becomes very small and TDR measurements may become overly sensitive to soil heterogeneity close to the rods.

Because of this flexibility in probe width and length, TDR probes may be designed to measure a wide range of soil volumes. Because the volume measured extends only 1 to 2 cm above and below the plane of the rods for most probe designs, TDR is ideal for measurements in thin layers near the soil surface. It is also very useful in root water uptake studies for which information from discrete parts of the root zone is desired. Because TDR accurately integrates soil water content changes occurring along the length of the probe rods, TDR probes may be inserted vertically into soils to accurately assess mean water content over the length of the rods, even in soils exhibiting sharp water content changes with depth. Probe designs are not confined to a bifilar or trifilar, parallel rod design, but may be customized in many ways. Ferré et al. (1998, 2000) gave sample areas of several probe designs and investigated their sensitivity and relative merit. Probe construction is relatively simple, and construction guides exist (e.g. Evett and Ruthardt, 2000), so that producing a customized probe design is a feasible task, at least for scientific work.

The dielectric permittivity of all materials is frequency dependent. Campbell (1990) showed that the effects of temperature, ionic conductivity, and soil type on the dielectric permittivity increased as signal frequency decreased. Thus, total coaxial cable length between the TDR instrument and probe is restricted due to attenuation of high frequency components of the TDR pulse by the cable, which acts as a low-pass electronic filter. This is particularly true for high surface area soils, which tend to exhibit larger BEC. Hook and Livingston (1995) determined that an RG6, 75- Ω coaxial cable had superior characteristics for TDR than the commonly used RG58, 50- Ω cable due to the five times faster signal rise time of the RG6. R.G. Schwartz (Conservation and Production Research Lab., Bushland, TX, personal communication, 2004) compared RG6, RG62, RG58 and RG8 cables and found that RG6 and RG8 offered the least signal attenuation. Total cable lengths <30 m are recommended for sandy soils, with the allowable total length decreasing to <10 m for soils with large BEC when wet. These lengths may be increased by ~50% when using RG6 or RG8 cable rather than RG58.

An important use of the TDR method is to calculate the soil bulk electrical conductivity (BEC) from values of the waveform relative voltage or impedance at various points along the waveguide (Fig. 2-16, left) (e.g. Dalton et al., 1984; Dalton and van Genuchten, 1986; Dasberg and Dalton, 1985; Wraith et al., 1993; Wraith, 2002). The bulk electrical conductivity is

$$\sigma_a = \frac{\epsilon_o c_o}{L} \frac{Z_o}{Z_u} \left(\frac{2(V_R - V_I)}{V_F - V_I} - 1 \right) \quad [29]$$

where ϵ_o is the permittivity of free space (8.854×10^{-12} F m⁻¹), c_o is the speed of light in a vacuum ($299\,792\,458$ m s⁻¹), L is the probe length (m), V_R , V_F and V_I are relative voltages measured from the wave form (Fig. 2-16), Z_o is characteristic impedance of the probe (Ω), and Z_u is the characteristic impedance of the cable tester (Ω). Calculation of σ_a from TDR data is still a subject of active research.

Although the TDR method has proven useful for irrigation research applications, its large installed cost, complexity of installation and requirement for careful and time consuming system management have combined to reduce its favor for on-farm irrigation management. About two thirds of the world's soils present little difficulty for TDR (Logsdon, 2000), and accuracies should be

within $\pm 2\%$ using Eq. [26] with minimal temperature effect if short cables are used. However, for soils with large amounts of organic matter, smectite clays, iron-rich soils, or clays derived from volcanic materials (Andisols, Miyamoto et al., 2003), Equation [26] is not accurate and soil-specific calibrations are necessary (Bridge et al., 1996). In the case of iron-rich soils, the apparent permittivity of the soil solids may be much different from that of the mineral soils studied by Topp et al. (1980). In the soils with large surface area, the large amounts of bound water have a different permittivity than does the mostly free water found in many mineral soils. Also, the BEC in such soils may vary with water content, even though the soils are non-saline, influencing the waveform analysis; and in these soils, the calibration tends to be temperature sensitive (Logsdon, 2000). However, the calibration of Evett et al. (2005) decreased temperature sensitivity to $<0.0006 \text{ m}^3 \text{ m}^{-3}$ (Eq. [28]). In any soil, for $\text{BEC} > 4 \text{ dS m}^{-1}$, the waveform will not be interpretable due to a loss of the second reflection. As long as BEC is small enough that the waveform can be interpreted, the effect of salinity on reported water content is small. Finally, because the waveform analysis for travel time is so critical to the method, it is recommended that software be used that allows for user control over the waveform analysis and the specific algorithms used. Two public domain software packages are TACQ (<http://www.cprl.ars.usda.gov/programs/>) and WinTDR (<http://129.123.13.101/soilphysics/wintdr/index.htm>).

Capacitance, Frequency Domain, and Power Loss Methods

These methods employ various forms of an electronic circuit called an oscillator, producing a repetitive waveform, usually sinusoidal, but sometimes with a short rise time pulse as in TDR. Even if the waveform has a short rise time similar to the step pulse used in TDR, the method is not TDR unless it captures all parts of the waveform associated with the probe and analyses the waveform for travel time. The methods differ in the particular circuit employed, the way in which the probe itself is employed in the circuit (as a capacitive element or as a wave guide), and the method of placement of the probe in the soil (by direct burial or insertion, or by moving the probe within an access tube). They also differ in the measured property, which may be a frequency, count of reflected pulses, phase angle, or power loss. As with TDR, none of these methods actually measures the soil electrical permittivity or water content. The base frequency is a concern. It should be $\gg 100 \text{ kHz}$ to minimize DC conductivity effects. Regardless of the frequency or particular method employed, all these methods are subject to the same interferences as is TDR, except more so because methods in this class employ frequencies below the broadband range inherent in TDR signals.

Most circuits employed use an oscillator coupled electrically to a capacitive element (C_2 in Fig. 2-18) that is either directly in contact with the soil or placed within an access tube. Typically, such circuits employ capacitive elements (C), inductive elements (L), and resistive elements (R), and so are called RLC circuits.

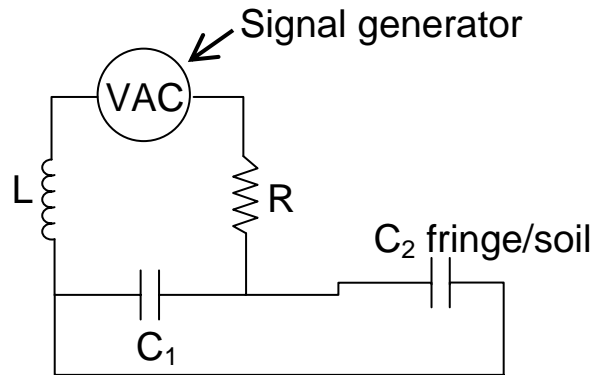


Fig. 2-18. Schematic of an RLC oscillator coupled to a capacitive element, C_2 , in contact with the soil, either directly or through the wall of an access tube.

A typical oscillator employs an RLC circuit with capacitive elements in parallel (Fig. 2-18). Capacitance C_1 is on the circuit board, and its value is well known. Capacitance C_2 is formed by the electrodes (rods, plates, cylinders, etc.) of the probe and in part by the soil itself, which acts as part of the dielectric medium for C_2 . The degree to which the soil acts as part of the dielectric medium for C_2 is determined by the probe design. If the probe capacitive element consists of two or three rods buried or inserted into the soil, then the soil makes up a large part of the dielectric medium for that element (Zegelin et al., 1989, Ferré et al., 2000). The probe handle makes up part as well. In the case of a capacitive element made up of two cylindrical plates, one above the other in an access tube, the soil may make up only a small part of the dielectric medium of the element. In the latter case, the soil is affected by only a part of the electromagnetic field between the plates; and this is called the fringing field.

A typical design is that of Dean et al. (1987), which used a capacitor made up of two cylindrical plates, one stacked above the other (Fig. 2-19). This was lowered into a plastic access tube, or could be buried directly in the soil. The capacitance of the soil-access tube system, $C(F)$, is given by:

$$C = g\epsilon_a \quad [30]$$

where ϵ_a is the system apparent permittivity, and g has units of farads and a value dependent on the geometry of the system. The resonant frequency, F (Hz), is (Dean et al., 1987):

$$F = [2\pi(L)^{0.5}]^{-1} (C^{-1} + C_b^{-1} + C_c^{-1})^{0.5} \quad [31]$$

where C_b and C_c are the electrode capacitances including the capacitances of internal circuit elements to which the electrodes are connected, C is the capacitance of the soil-access tube system defined in Eq. [31], and L is the inductance (henries) of the coil in the LC circuit.

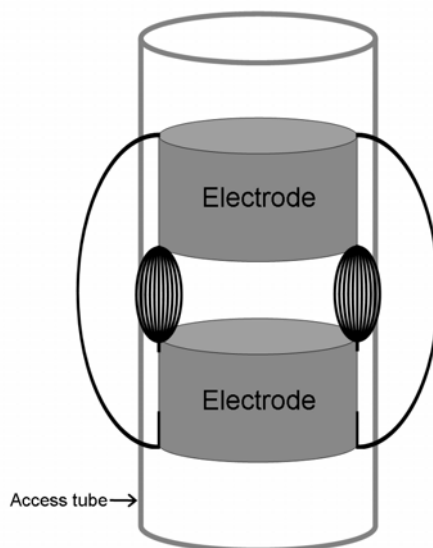


Fig. 2-19. Cartoon of a cylindrical capacitance probe in an access tube illustrating that most of the electromagnetic force lines go directly between the two cylindrical electrodes of the capacitor, with only a relatively small fringing field permeating the soil. Outmost dark curves indicate largest extent of fringing field.

A similar design is employed by the Sentek EnviroSCAN, Diviner 2000 and related systems; the Troxler Electronics Laboratories, Inc. Sentry 200AP; the Delta-T PR1/4, PR1/6 and PR2/6, and other proprietary systems. Typically, the volume sensed by capacitance systems used in access tubes is relatively small compared with the volume sensed by the NMM. For instance, Evett and Steiner (1995) in a field calibration of several NMMs of two manufactures and the Sentry 200AP, found that the NMMs could be calibrated with $RMSE < 0.01 \text{ m}^3 \text{ m}^{-3}$ and $r^2 > 0.9$, while the capacitance probe calibration r^2 values ranged from 0.041 to 0.712 with RMSE values ranging from 0.036 to 0.058 $\text{m}^3 \text{ m}^{-3}$. Soil samples were taken with the Madera probe, four samples at each measurement depth at each access tube. While these samples were taken as close to the access tube as feasible, they were not within the volume sensed by the capacitance probe. That this was true was shown by the high correlation between the four capacitance probes used. With r^2 values ranging from 0.96 to 0.99, the four probes all were sensitive to the same soil properties in the same way, but these properties were not representative of the REV for water content as sensed by the NMM and measured by volumetric sampling. Paltineanu and Starr (1997), working with the EnviroSCAN system, showed that over 80% of the sensitivity of an EnviroSCAN sensor was within 2.5 cm of the outside of the access tube, and over 90% of the sensitivity was within 3 cm of the access tube. The sensed volume decreased as water content increased. Kelleners et al. (2004a) found that most of the electromagnetic field from these sensors does not go into the soil outside of the access tube. Evett et al. (2002a) tested the axial sensitivity of the Diviner 2000 and PR1/6 capacitance probes along with the NMM and the Trime T3 tube probe. They found that the Diviner 2000 did not sense above and below the top and bottom of the sensor capacitor electrodes in dry soil, and that in saturated soil the axial response was actually less than the vertical height of the sensor body (Table 2-4). Range of axial sensitivity of the PR1/6 also decreased as water content increased, but was always larger than the distance between the top of

Table 2-4. Axial response to the soil-air interface† for two capacitance type soil water sensors (Delta-T PR1/6 and Sentek Diviner2000), the neutron probe, and a quasi-TDR type sensor (Trime T3) in air-dry and saturated (wet) soil. The measurement volume decreases as the soil becomes wetter for both the capacitance sensors and for the neutron probe.

Instrument	Sensor Height	Height (cm) of 90% <u>Response</u> <u>Window</u>		Ratio of Response to Sensor <u>Heights</u>	
		Dry	Wet	Dry	Wet
		----- cm -----			
Delta-T PR1/6	4.8	7.4	5.6	1.54	1.16
Sentek Diviner2000	6.2	6.2	3.1	1.00	0.50
Neutron probe	13.2	27.7	15.6	2.10	1.18
Trime T3	17.5	16.9	18.3	0.97	1.04

† Measured incrementally from >30-cm above to >30-cm below the surface.

the top electrode and the bottom of the bottom electrode. The volume sensed by other capacitance probes is small enough to make field calibration problematic as shown by the poor calibration results reported by Evett et al. (2002b) for the Diviner2000 and Whalley et al. (2004) for the PR1/6.

Thus, for many capacitance systems the volume sensed is small, may be smaller than the representative elemental volume for soil water content, and is largely within the zone that might be disturbed during access tube installation. The installation kits supplied with these systems are usually optimized to minimize such soil disturbance. However, the effect of air gaps between capacitance probes and soil is large, causing a decrease in sensed permittivity of as much as 28% for a gap of 0.2 mm (de Rosny et al., 2001) for one capacitance probe design. Calibration accuracies for soil-specific calibrations reported in the literature are on the order of 0.02 to 0.03 m³ m⁻³, somewhat larger than the values ≤0.01 m³ m⁻³ reported for the NMM and conventional TDR (Evett et al., 2006).

Methods Measuring Thermal Properties

Few commercially available sensors rely on thermal properties for soil water sensing. One is the model 229 thermal diffusivity sensor sold by Campbell Scientific, Inc. This sensor consists of a porous block within which are included a heating element and a thermocouple for measuring temperature. A measurement cycle consists of heating the element for a known time, and sensing the rise and fall of temperature within the block. The wetter the block the faster is the dissipation of heat away from the heater. The block is assumed to be in equilibrium with the soil water; that is, it is at the same energy potential, not necessarily the same water content as the soil. This sensor is thus calibrated vs. matric potential rather than water content. Its range is from 0.01 to 1 MPa. Sensors operating on similar principles, but having one or two probe rods that are inserted or buried directly in the soil, can be calibrated for soil water content. Research and development of such probes continues, with construction and calibration details still under investigation (Ham and Benson, 2004), and influences of soil properties being considered (Rena et al., 2003).

Measurement Interferences

Soil chemical composition, bulk density, bulk electrical conductivity, and temperature all interfere with indirect water content measurement methods; but the degree of interference varies widely among the various technologies and the specific implementations of these. For example, Evett et al. (2002a) found that the neutron thermalization and conventional TDR methods were not significantly sensitive to temperature changes, while the capacitance methods they studied were all affected to some degree, one of them to a great degree (sensitivity of $0.025 \text{ m}^3 \text{ m}^{-3}$ water content per $^\circ\text{C}$). Working with clay, peat, and sand soils, Persson and Berndtsson (1998) found that the largest temperature correction needed for TDR water contents was for sand, and was $-0.00269 \text{ m}^3 \text{ m}^{-3} \text{ }^\circ\text{C}^{-1}$; but they concluded that temperature-induced errors in TDR readings generally were small compared with other sources of error. Pepin et al. (1995), working with TDR in sand, loam, and peat soils, found that temperature sensitivity increased from near zero at small water contents to $0.008 \text{ m}^3 \text{ m}^{-3} \text{ }^\circ\text{C}^{-1}$ at a water content of $0.5 \text{ m}^3 \text{ m}^{-3}$. Differences in these results can be attributed to different experimental conditions and waveform analysis algorithms, and to the source of temperature dependency. The inverse relationship between the dielectric permittivity of water and temperature (Fig. 2-1) is in agreement with the negative coefficient found by Persson and Berndtsson (1998) in a sand. Because there is relatively little bound water in sand, the system is evidently responsive to free water. Positive relationships, such as those found by Pepin et al. (1995) may be due to the presence of important amounts of bound water in a soil system, and to the influence of clays with large surface charge in which BEC increases with temperature. Seyfried and Murdock (2001) also stated that the temperature effect on TDR was much smaller than that on a water content reflectometer that employed a bistable multivibrating oscillator. Mead et al. (1996), using the EnviroSCAN system, found a day-night fluctuation of up to $0.05 \text{ m}^3 \text{ m}^{-3}$ associated with temperature change in a loam soil. They also found that the effect of temperature was larger for a wet soil than for a dry soil, in agreement with both the observations of Evett et al. (2006) (Table 2-5) for four frequency domain sensors, and the results of Seyfried and Murdock (2001) for a model CS615 probe (Campbell Scientific, Inc., Logan, UT) for which the temperature effect was different for different soil types. The implications of these findings include: i) that soil type may affect the degree of temperature interference, ii) that temperature should be measured as a covariate during calibrations of frequency domain sensors, which should be done in such a way to avoid confounding the effects of water

Table 2-5. Temperature sensitivity[†] in air-dry and saturated soils.

Instrument	Dry soil ($0.05 \text{ m}^3 \text{ m}^{-3}$)			Saturated soil ($0.43 \text{ m}^3 \text{ m}^{-3}$)		
	Slope ($\text{m}^3 \text{ m}^{-3}$) $^\circ\text{C}^{-1}$	r^2	RMSE ($\text{m}^3 \text{ m}^{-3}$)	Slope ($\text{m}^3 \text{ m}^{-3}$) $^\circ\text{C}^{-1}$	r^2	RMSE ($\text{m}^3 \text{ m}^{-3}$)
Trime T3	0.009	0.52	0.005	0.0204	0.75	0.0012
Delta-T PR1/6	0.001	0.73	0.0013	0.0250	0.94	0.0002
EnviroSCAN	0.0009	0.76	0.0004	0.0010	0.88	0.00001
Diviner2000	0.0005	0.65	0.0003	0.0019	0.77	0.0001

[†] Measured at 25-cm depth.

content change and temperature change, and iii) that if temperature has an affect, it should be measured as a covariate during field measurements so that these measurements can be corrected. Still, measurement of temperature as a covariate may be insufficient to correct calibrations of EM sensors (Evelt et al., 2006) and measurement of BEC may also be required.

Soil bulk electrical conductivity (BEC) is the most important interference for all measurement methods that are based on electrical properties of soils. Capacitance and other frequency domain and power loss methods are all sensitive to various degrees to soil BEC. For example, capacitance systems typically exhibit increasing degrees of error as the salinity of soil water increases (Fig. 2-20) (Evelt et al., 2002b; Kelleners et al., 2004a). So far, attempts to incorporate salinity in capacitance sensor calibration have only been partially successful (Kelleners et al., 2004b). Another example of the large effect of salinity on sensor output is given by Miyamoto and Maruyama (2004) in which they showed that the output signal increased by 1.6 times as EC increased to 6 dS m^{-1} for the Campbell Scientific, Inc. model CS615. The same authors showed that coating the rods with plastic could allow for a useful calibration that was insensitive to salinity, though such coating did reduce the sensor's sensitivity to water content changes. The BEC of soils is strongly temperature dependent (Persson and Berndtsson, 1998) and also increases strongly with soil water content for a given value of soil solution EC (Rhoades et al., 1976, 1989) as shown in Figure 2-21 (Mmolawa and Or, 2000). Also, BEC increases with clay content for at least 14 San Joaquin Valley, CA soils (Rhoades et al., 1989). In contrast with most sensors that respond to changes in soil permittivity, conventional TDR with waveform analysis has been found to be relatively insensitive to $\text{BEC} \leq 2 \text{ dS m}^{-1}$ (soil solution $\text{EC} \leq 10 \text{ dS m}^{-1}$) (Nadler et al., 1999). At values of $\text{BEC} \geq 4 \text{ dS m}^{-1}$, the TDR waveform is so attenuated that analysis for travel time becomes impossible. Increasing cable length will increase the sensitivity of TDR to BEC and thus to temperature fluctuations. This occurs because cables act as low-pass filters, filtering out the high frequency components of the TDR signal and lowering the effective signal frequency.

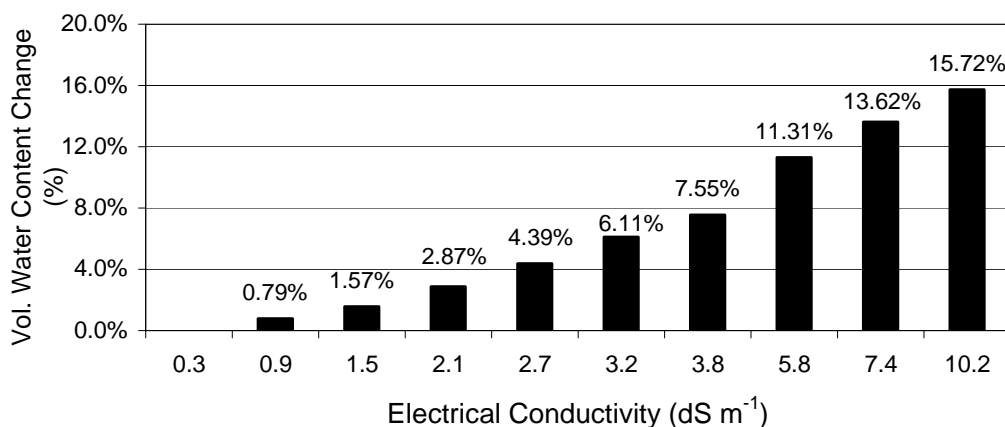


Figure 2-20. Percent change from the true water content as electrical conductivity (EC, dS m^{-1}) of the soil water increases for a Sentek Diviner 2000 capacitance system.

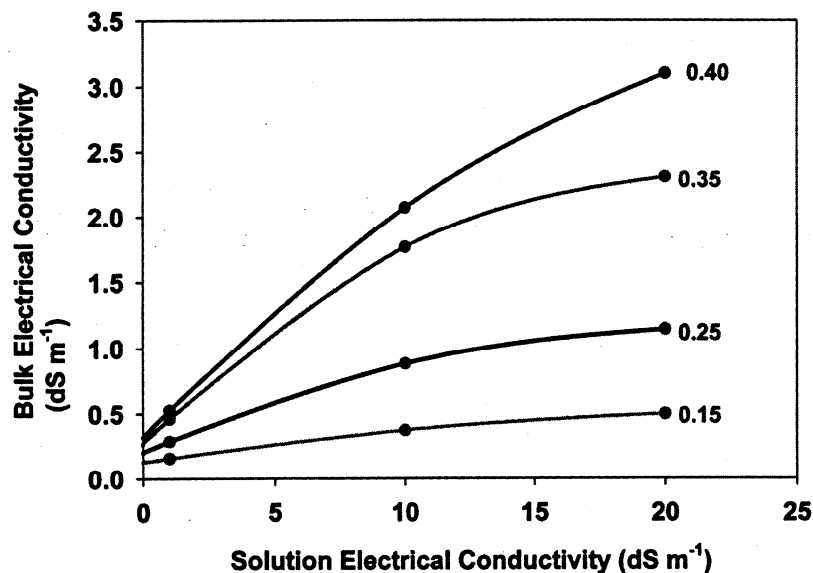


Figure 2-21. Soil bulk electrical conductivity (BEC) as a function of soil solution EC for water contents ranging from 0.15 to 0.40 $\text{m}^3 \text{m}^{-3}$ in a Millville silt loam soil (Figure 15 from Mmolawa and Or, 2000, with kind permission from Springer Science and Business Media).

World wide, 20% of irrigated soils are salt affected (Hachicha and Abd El-Gawed, 2003). Sensitivity to soil BEC limits the applicability of frequency domain or power loss sensors in many irrigated soils in which BEC varies across the field (Plate 2-2) and with time (Plate 2-3). Variations of BEC of as much as 12 dS m^{-1} can occur over distances of less than one meter (Burt et al., 2003), and differences equally as large can occur from year to year or even within an irrigation season in one location (Hanson et al., 2003). Abdel gawad et al. (2003) measured periodic soil solution EC variations of 5 to 6 dS m^{-1} under drip irrigation in Syria. Mmolawa and Or (2000) measured a BEC change from 0.3 to 2.3 dS m^{-1} in a few hours under drip irrigation of corn. While it is possible to calibrate most sensors for a particular BEC, in these situations of temporally and spatially variable BEC, such a calibration is not applicable. From the available data, it is clear that errors larger than 50% in water content at a single location, and errors similarly large in profile water content are possible given the range of BEC values measured. Spatial and temporal variations of BEC are not confined to drip irrigation, but are present under furrow, flood, and sprinkler irrigation as well.

Sensors based on frequency domain or power loss (of an oscillator) principles are often also sensitive to clay content and type even in non-saline soils. This is because clays exhibit varying degrees of charge and are associated with cations or anions in the soil solution to varying degrees. Commonly, clays exhibit negative charge and are associated with cations to a degree that is evaluated as the cation exchange capacity (CEC). As the soil content of high CEC clay increases, the soil becomes more electrically lossy, that is, more capable of affecting the movement of electrical fields. This affects the frequency of oscillation of capacitance systems or the power loss of power loss systems in a way that is separate from, but not completely independent of, the soil water content. Examples include the much different calibration equations developed for the several soils existing

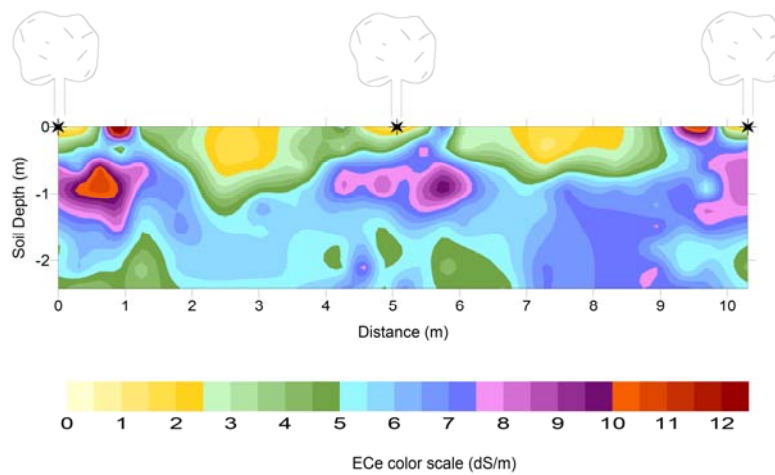


Plate 2-2. Variations in EC of saturation paste soil extracts (EC_e) in two dimensions of a pistachio plantation that was drip irrigated in California.

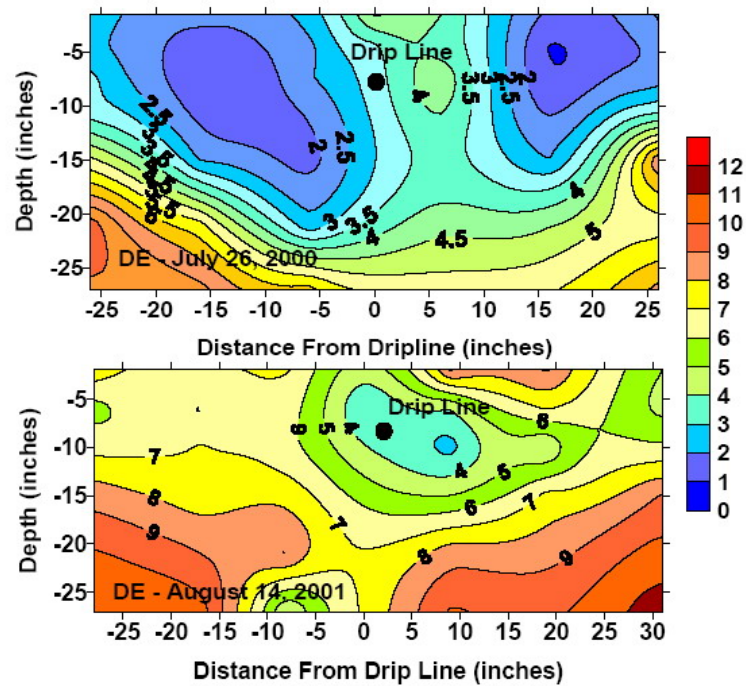


Plate 2-3. Variations in EC from saturation paste soil extracts from a single location in a drip-irrigated tomato field in California in two different years. No yield variation was found.

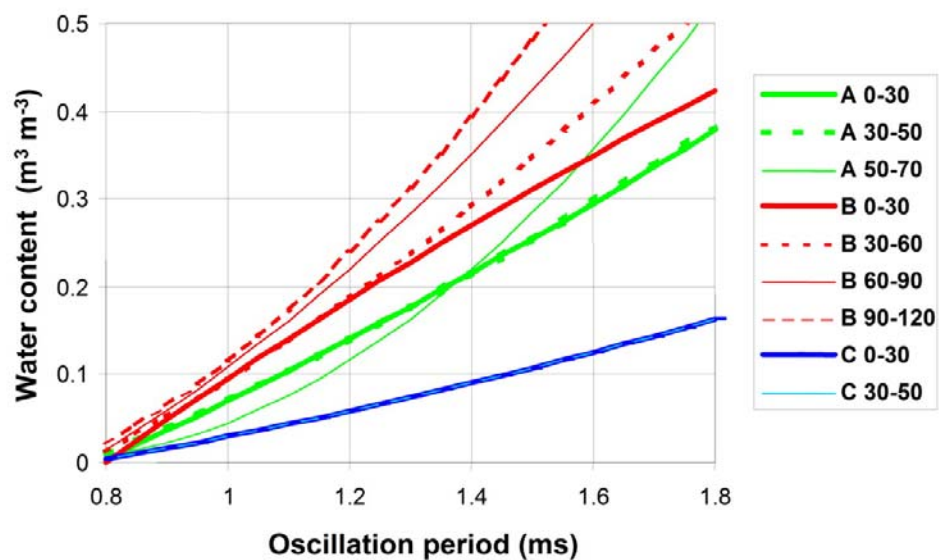


Plate 2-4. Calibrations of the model CS615 soil water probe from Campbell Scientific, Inc. in nine different soil layers of three different soils (A, B, and C), illustrating the wide variance in calibration equations for different layers in a particular soil and among soils (Ruelle et al., 2003, personal communication).

under one center pivot irrigation system in France (Plate 2-4) (Ruelle et al., 2003, personal communication), and the different calibration equations reported by Baumhardt et al. (2000) and Morgan et al. (1999) for the Sentek EnviroSCAN system.

Several studies (e.g. Farahani and Buchleiter, 2002) illustrate that BEC can be highly correlated with both clay content and water content in non-saline soils (rain fed or irrigated with high quality water). In non-saline soils, the spatial distribution of BEC can be temporally stable because it is dependent mostly on the distribution of clay content in the field. In this situation, there can be a unique relationship between BEC and water content at a particular location such that BEC measurements made with a TDR system can be correlated with water content, resulting in a system that can give good values of water content even when BEC is too large for the TDR waveform to be correctly interpreted for travel time.

MEASUREMENT OF SOIL WATER POTENTIAL

The volumetric soil water content is essential information for irrigation management because, when combined with knowledge of the soil water content vs. potential and water content vs. hydraulic conductivity relationships, it allows calculation of how much water can be added to the soil before losses to deep percolation become important. However, plants respond directly to soil water potential, not water content. Thus, various devices have been invented to sense the soil water potential: the tensiometer, the soil psychrometer, gypsum blocks, and granular matrix sensors being the four most common.

Tensiometers

Among the several methods for measuring soil water potential, the tensiometer is the oldest, dating to 1908 (Or, 2001). The tensiometer consists of a cup of porous material, usually ceramic, and often connected to a thick-walled plastic tube, filled with water, and attached for reading to a device for measuring the differential pressure between the water inside the tensiometer and the atmosphere. The pore size in the cup determines both its bubbling pressure and its hydraulic conductivity. The bubbling pressure is that at which air would force water out of the pores, allowing air to flow across the cup wall. Common tensiometer designs include a thick-walled, rigid plastic tube long enough to allow placement of the ceramic body at the desired depth while allowing a pressure gage to be attached at the top of the tube, which extends above the soil surface. Older designs used a mercury manometer to measure the pressure. Newer designs use digital pressure sensors; and at least one design replaces the pressure gage with a rubber septum through which a needle can be inserted for connection of a portable pressure gage (Fig. 2-22).

Tensiometers tend to fail at more than 70 kPa tension due to air passage through the cup. The upper limit of use is 100 kPa, at which tension the water inside would boil at room temperature (at sea level). Boiling the water before filling the tensiometer removes most of the dissolved air and decreases the tendency for bubble formation at tensions below 100 kPa. Because soil air will diffuse through the cup and dissolve in the water inside, this is not a permanent fix. A new design by Hubbell and Sisson (1998) extends the depth placement range by placing a pressure transducer at the

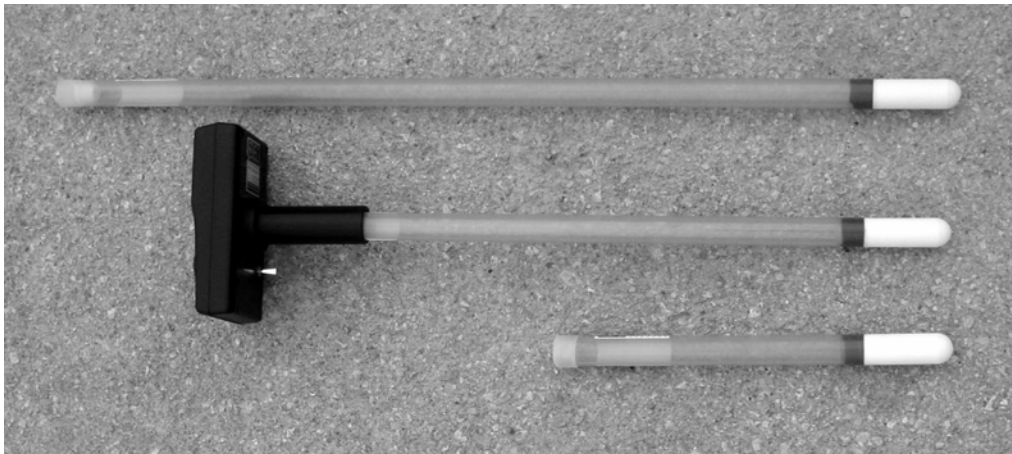


Fig. 2-22. Tensiometers for use at three depths. For each, the hollow ceramic body is the white portion on the right. These tensiometers have a rubber septum at the upper (left) end. A digital pressure gage has been connected to the middle tensiometer by pressing the gage needle through the septum.

level of the porous cup, eliminating the column of water from the cup to the soil surface. The ceramic body must make good contact with the soil to maximize tensiometer response to changes in soil water potential. In shrink/swell clay soils, contact is easily lost. Contact can also be a problem in very coarse-textured soils due to the capillary barrier created by the different sized pores of the cup and the sand. Tensiometers are temperature sensitive, either due to thermal expansion/contraction of the tube, expansion of air in the tensiometer, or due to temperature effects on the pressure sensor. Thermal expansion or contraction of the tube or air inside the tensiometer will cause water to move in from or out to the soil through the porous cup. Thus, the cup impedance and soil hydraulic conductivity will affect the amount of pressure fluctuation, with larger impedance and lower hydraulic conductivities leading to larger fluctuations (Warrick et al., 1998). Tensiometers require periodic refilling. Tensiometers have been successfully used for irrigation scheduling, particularly for high value, shallow rooted crops for which the labor intensive nature of these devices can be offset by yield and quality gains from careful water management.

Installation of tensiometers involves augering a hole to the depth of measurement, placing loose soil in the hole to ensure good contact between the cup and soil, and pressing the cup into the soil, then backfilling the hole. In coarse soils, a finer textured material may be placed in the bottom of the hole to establish good contact. Typically, the plastic tube extends to above the soil surface for measurements. The vertical distance between the height of measurement must be subtracted from the reading to account for gravitational potential and arrive at the matric potential. Calibration is of the pressure sensing system.

The use of automatic datalogging allows tensiometers to be monitored continuously so that they can be used for irrigation automation. However, this automation has pointed up the fact that tension readings fluctuate with temperature, particularly in drier soils (up to 35 cm fluctuation of H) due to the smaller hydraulic conductivities that impede water flow into and out of the porous cup (Warrick et al., 1998). This complicates the comparison of readings from different depths, as would

be necessary if the hydraulic gradient were to be assessed. Richards (1949) offered a guideline for the timing of readings: “The effect of diurnal fluctuations on the reading of field instruments can be minimized by making the reading at the same time of day, preferably in the morning, so as to follow a period of slow temperature change”. The first part of this guideline holds for most instruments, except for those subject to the lag of soil temperature fluctuations with depth. The second part, preferring morning, is contradicted by the results of Warrick et al. (1998), which showed that tensiometer readings fluctuated the most in early morning, and that readings at night would be the most stable. The numerical results of Warrick et al. (1998) also provide insight into the loss of hydraulic contact between tensiometer cup and soil in dry soils, which is relevant to gypsum blocks and granular matrix sensors as well: “...for a dry soil we might expect water to be driven away from the tensiometer (and not returning) owing to K changing as a function of hysteresis”.

Gypsum blocks

Gypsum blocks are just that, a block of calcium sulfate (gypsum), usually formed by mixing plaster of Paris with water and pouring into a mold. Embedded in the block are two wires, often connected to metal mesh electrodes. The porosity of the solidified plaster of Paris is such that the block will take up water from wet soil and release it as the soil dries. The electrical resistance of the block is related to the soil water potential through a calibration curve. Because the gypsum salt buffers the water in the block, the effects of soil water salinity on the electrical resistance measured are minimized. Gypsum blocks are highly variable in output from one block to the other, and must be calibrated. However, the calibration drifts over time as the block dissolves and its porosity changes. The soil water potential is a curvilinear function of the electrical resistance of the block. The range of useful readings is approximately -150 to -600 kPa matric potential. Gypsum blocks are temperature sensitive, which is less problematic with deeper installation. It is relatively easy to install gypsum blocks to various depths in auger holes. The blocks are read with a hand-held meter or connected to a data logging system for unattended data acquisition. Good contact between the block and soil is essential; and in some soils this contact may be problematic (sandy soils or cracking clays). While they have their place in irrigation scheduling, gypsum blocks are not accurate enough to determine the soil water potential gradient for soil water flux calculations.

Granular matrix sensors

Several types of granular matrix sensors (GMS) are on the market. The sensor consists of a porous medium in which are embedded two wires, often connected to wire mesh electrodes inside the sensor. The reading is of the electrical resistance in the medium between the wires or mesh electrodes. Often, a quantity of gypsum (calcium sulfate) is included to buffer the soil water solution and decrease effects of salinity on the resistance. The greater the soil water tension, the less water is in the porous medium, and the greater the electrical resistance. Calibration may be done in a porous medium covering a pressure plate, which is subjected to several values of pressure in a pressure chamber. Calibrations are soil specific, so it is wise to use the soil to be measured as the porous medium. Installation and contact problems are similar to those for a tensiometer or gypsum block,

including contact problems in coarse sands and shrink/swell clays. At tensions less than 30 kPa, Taber et al. (2002) found that tensiometers responded more rapidly than GMS sensors in silt loam, loam, and coarse sand. As with gypsum blocks, reading requires an alternating current to minimize effects of capacitive charge build up and ionization. Lack of precision and calibration drift over time may limit use of GMS for determining soil water potential gradients.

The useful range of readings is approximately -10 to -200 kPa matric potential, though Morgan et al. (2001b) were able to use GMS sensors to -5 kPa in a fine sand. Sensors may be manually read or data logged (resistance reading). Some hysteresis is noted with these sensors; and they are temperature sensitive (as much as 20 kPa per 10°C , Shock, 2003). Like gypsum blocks, GMS may be installed to practically any useful depth, limited only by wire length. Fewer problems with soil contact are noted with GMS. The usefulness of GMS systems for irrigation scheduling has been illustrated by work done with onions, potato (Fig. 2-23), alfalfa, and sugar beet in the Malheur Valley of Oregon (Shock, 2003; Shock et al., 2003). Because of soil and irrigation variability, at least six sensors should be used to provide data for irrigation scheduling (Shock, 2003). For irrigation science, the GMS can be useful if calibrated for the soil over a range of temperatures and soil water potentials, and if soil temperature is measured at the location of each sensor so that calibration corrections for temperature can be applied. Automatic irrigation scheduling has been successfully implemented using GMS for high-value row crops (Shock et al., 2002) and for landscapes (Qualls et al., 2001).

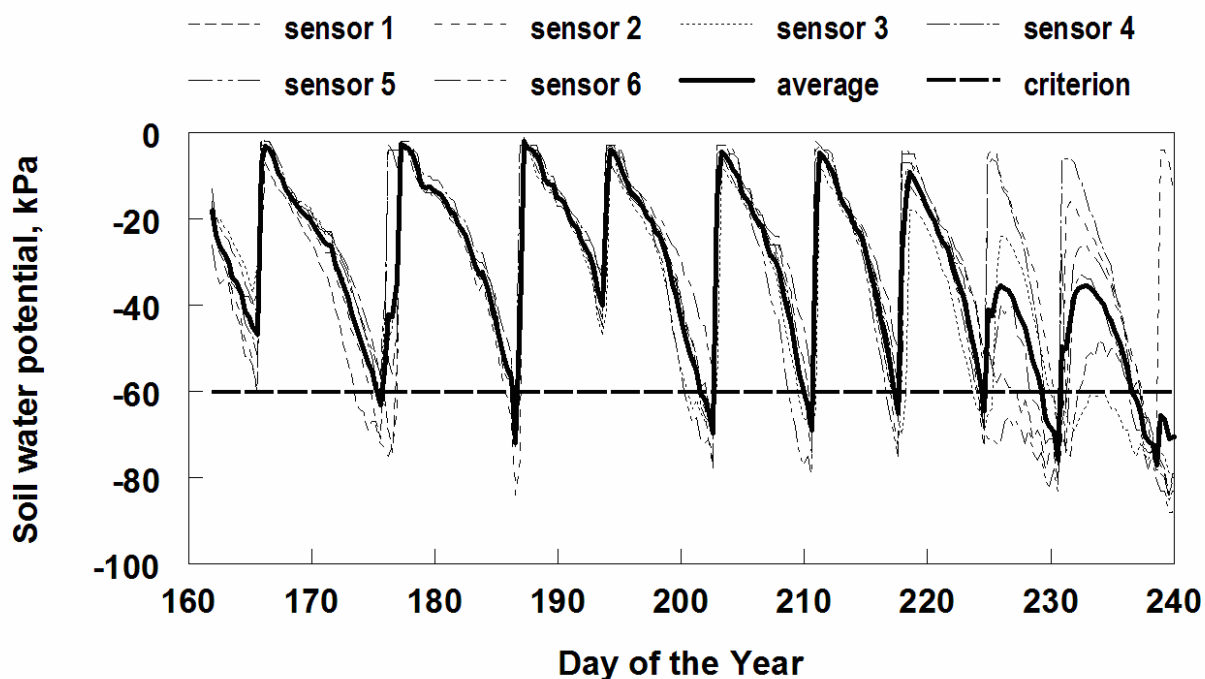


Fig. 2-23. Soil water potential in a sprinkler-irrigated potato field as sensed with six granular matrix sensors datalogged using a Hansen model AM400 data logger, showing very good control of soil water potential. Note the dry-down period at the end of the irrigation season (Shock et al., 2003)

Soil psychrometers

The soil psychrometer senses the relative humidity of the soil pore space; that is, it senses the vapor pressure of water in the soil. The vapor pressure, p (Pa), is related to both the matric potential, Ψ_m , and the osmotic potential, Ψ_o ,

$$\Psi_m + \Psi_o = \frac{RT\rho_w}{M_w} \ln(p/p_o) \quad [32]$$

where R is the universal gas constant ($8.314 \text{ J mol}^{-1} \text{ K}^{-1}$), T is the temperature (K), M_w is the molecular weight of water ($0.018 \text{ kg mol}^{-1}$), ρ_w is the density of water (1.000 kg m^{-3} at 20°C), and p_o (Pa) is the vapor pressure when the air is saturated with water at temperature T , i.e. 100% relative humidity. The ratio p/p_o multiplied by 100 is the percent relative humidity.

The thermocouple psychrometer is the most common type, and was described as early as 1936 (Powell). Psychrometers employ both a wet bulb and a dry bulb measurement. The dry bulb measurement gives the temperature of the soil air; and the web bulb measurement gives the (smaller) temperature of water in equilibrium with the water vapor in the soil air. The drier the air, the smaller is the wet bulb temperature, or the greater is the temperature depression – the difference between dry bulb and wet bulb readings. In the case of 100% relative humidity, saturation, these temperatures are equal. In traditional micrometeorological instruments, the wet bulb was literally covered with a cloth sock and wetted with pure water. In order to find the true temperature depression, the web bulb was ventilated to maximize the evaporative cooling. Powell (1936) and others found that the use of fine-wire (e.g. 0.025 mm diameter) thermocouples for the temperature sensor would allow ventilation to be eliminated with little ensuing error. Spanner (1951) demonstrated that the Peltier effect could be used to cool a thermocouple junction to below the dew point temperature, thus inducing water to condense out of the air onto the thermocouple. With this advance, the dual needs for a wetting system and a source of pure water were eliminated, and in-situ measurements were facilitated. Modern soil psychrometers are smaller than 1 cm in diameter, and have the thermocouples protected by a wire mesh or porous ceramic bulb, which in turn are in contact with the soil.

The origin of modern soil psychrometers can be traced to the work of Rawlins and Dalton (1967) and Dalton and Rawlins (1968). Various design improvements were described by Chow and de Vries (1973), McAneney et al. (1979), and Brown and Collins (1980). Still, accurate measurements remain difficult and subject to error (Brown and Chambers, 1987; Brown and Oosterhuis, 1992). The measurement range is from zero to approximately -8 MPa (Brown and Oosterhuis, 1992). Thermocouple psychrometer output voltages are in the microvolt range, requiring expensive analog to digital conversion circuitry. They must be carefully cleaned and maintained, and calibrated against standard salt solutions having a range of known osmotic potentials (Brown and Oosterhuis, 1992). They are sensitive to temperature gradients, and so are not recommended for use in the top 0.3 m of soil (Brown and Chambers, 1987). Because they respond to osmotic as well as matric potential, psychrometers are useful in the study of plant responses to soil salinity and saline waters, and in that context they are an essential research tool. But, their fragility, sensitivity to

temperature gradients and contamination, and requirement for expensive measurement circuitry combine to preclude their use in irrigation management.

CHOOSING A WATER CONTENT SENSOR

Several studies have compared soil water content sensors with an eye towards evaluating their use for irrigation research and on-farm water management. A common conclusion is that all sensors should be calibrated for the specific soil (including calibrations for different depths if soil properties change with depth) if water contents accurate to $\pm 0.02 \text{ m}^3 \text{ m}^{-3}$ or better are needed (Evett et al., 2002b, 2006; Hanson and Peters, 2000; Leib et al., 2003; Quinones et al., 2003). This conclusion comes despite manufacturer claims of ± 0.02 to $\pm 0.03 \text{ m}^3 \text{ m}^{-3}$ accuracy. A possible exception is conventional TDR (with waveform analysis) for many soils (Evett et al., 2002a), particularly if using the calibration in terms of travel time, effective frequency and BEC (Eq. [28]). Still, conventional TDR may need calibration in some silty clay (Hanson and Peters, 2000) or clay soils, depending on clay type, or in soils with large amounts of organic matter.

It has been argued that the change in water content (change in soil water storage in a profile) may be accurately estimated even if the water content values themselves are inaccurate. However, this is not true for linear calibrations if the slope is incorrect, or for nonlinear calibrations if the exponents and/or coefficients are incorrect (Hignett and Evett, 2002). For example, Hanson and Peters (2000) showed that the EnviroSCAN system did not accurately measure water content change in a silt loam and a silty clay using the factory calibration.

Many time domain, frequency domain, and resistance (potential) sensors can be used in automatic systems that provide multiple measurements per day, leading to graphical displays of the soil water dynamics (Fig. 2-24). If the sensor output is well correlated to actual water content, then set points for triggering irrigation can be established and can be reliable, even if the water contents and water storage values are inaccurate. However, this is far from foolproof and requires understanding and experience on the part of the irrigator. Also, the amount of irrigation required is not clear from a dynamic display, if the water contents, and thus water storage in the profile, are inaccurate. And, the set points so established are not transferable to other soils or other sensors, and cannot be related to information from extension and research publications (Lieb et al., 2003). In addition, intuitive analysis of such graphics relies on several days of data without irrigation so that patterns in crop water uptake can be discerned, such as relative daily rates of uptake that might be smaller in response to either soil saturation (lack of oxygen) or to lack of water, and that are larger during periods of maximal crop water use. Thus, irrigation management systems that require frequent irrigations are difficult to manage using the intuitive analysis, particularly for soils with low available water holding capacity. Finally, changes in weather from day to day cause differences in evapotranspiration rates and soil water uptake patterns that are difficult to separate visually from changes in the pattern of uptake that are due solely to soil water and plant status.

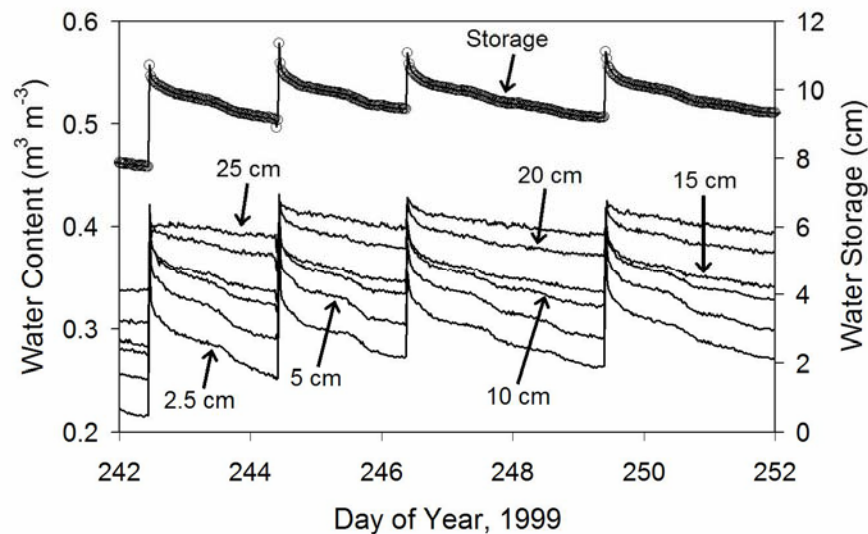


Figure 2-24. Dynamics of soil water content determined by TDR (Vadose System, Dynamax, Houston, TX) at six depths and of soil water storage in the surface to 27.5-cm depth range at Bushland, Texas. Four irrigations are visible, as are daily decreases in soil water content due to evapotranspiration from the alfalfa crop.

Given that most sensors must be calibrated before use, many factors besides the off-the-shelf accuracy should determine the selection of a water content sensor:

- Can the sensor be calibrated, and to what accuracy?
- Will calibration be needed for different soil types and depths in the field?
- How easy is the calibration?
- Are laboratory calibrations transferable to the field?
- Can calibration be done in the field?
- What are the accuracy of calibration, and accuracy and precision in the field?
- How many measurements or profile water content measurement locations will be needed to characterize water content to the needed level of accuracy?
- What are the interferences? Do they vary over time? Can they be overcome?
- Can the appropriate depths in the soil be measured?
- Is the volume measured appropriate for the desired use?
- What is the initial cost?
- What is the cost per measurement location?
- Can measurements be recording automatically?
- How easy is it to use?
- How much training is required?
- How durable is it?
- Will it interfere with farming or other operations?

Calibration issues

Most available sensors can be calibrated. Important statistics of calibration are the root mean squared error (RMSE) and coefficient of determination (r^2) of the regression of directly measured water contents vs. the sensor readings (eg. count, frequency, frequency shift, voltage, travel time, etc.). Of these, the RMSE is more important since it is a measure of the calibration accuracy. The r^2 value reflects both the degree to which the calibration equation explains the variability in the data and the degree to which there is a wide range of data values. Thus, a large r^2 value may not reflect a high degree of accuracy. Gypsum blocks and granular matrix sensors are calibrated for soil water potential, not water content. Lieb et al. (2003) showed how inaccurate the conversion from a potential reading to a water content can be if a generic relationship between soil water potential and soil water content is used for the conversion (root mean squared errors $\geq 0.17 \text{ m}^3 \text{ m}^{-3}$). Even if the soil water potential vs. water content curve is measured for a specific soil (an expensive and time consuming process), hysteresis in this relationship can cause large inaccuracies in the estimated water contents (e.g., Fig. 2-4).

It is important to consider the appropriateness of the calibration method. For example, Yoder et al. (1998) calibrated a Sentry 200-AP capacitance probe in two re-packed soils in weighed plastic columns and obtained coefficients of determination greater than 0.96. For three subsequent wetting and drying cycles, they found that this probe outperformed all others in their test, resulting in a mean absolute difference from mass balance water content of $0.017 \text{ m}^3 \text{ m}^{-3}$. They did not field test this calibration. In contrast, Heathman (1993) reported an r^2 of 0.62 for a field calibration of this probe. And, Evett and Steiner (1995) conducted a rigorous field calibration of four Sentry 200-AP probes in comparison with six neutron scattering (NS) meters, using wet and dry sites. Calibrations for the 200-AP probes exhibited low r^2 values, ranging from 0.04 to 0.71, and root mean squared error of calibration values of 0.036 to $0.058 \text{ m}^3 \text{ m}^{-3}$. The latter authors excavated the access tubes and verified the absence of air gaps between the tube walls and soil; so they attributed the poor calibration results to a combination of small sampling volume of the capacitance probe and small-scale heterogeneity of the soil. The electromagnetic field induced in the soil by a capacitance probe is influenced by boundaries between soil volumes having different permittivities (Dean et al., 1987). Thus, the size and shape of the sampled volume could be influenced by bulk density or water content variations on a small scale. Anomalous readings from capacitance probes applied to building materials were attributed to heterogeneities by Boot and Watson (1964). Wobschall (1978) pointed out that heterogeneous soils can cause poor results with capacitance sensing. In sum, it is important to realize that excellent laboratory calibrations done with re-packed soils, where tight soil-access tube contact is ensured and soil heterogeneities are minimized by the re-packing, may not be transferable to the field, particularly for capacitance or other frequency domain sensors, and especially in terms of the root mean squared error of regression, which may be much smaller for laboratory calibration than for field calibration.

Thus, field calibration is the preferred method because it places the sensor in the actual soil to be studied rather than in a more ideal laboratory setting. Any calibration will involve independent, direct measures of soil water content. The volumetric water contents are then related to corresponding values of the property measured by the sensor (frequency, travel time, etc.) through a statistical analysis (linear or non-linear regression). It is important in any calibration to ensure that

the soil volume sampled volumetrically is the same volume that is sampled by the sensor. For sensors that consist of rods that are buried or pushed into the soil, this can be accomplished by removing the sensor and sampling the soil into which the sensor had been inserted. However, for sensors that work from within an access tube, it may be that the volume sensed is smaller than the volume that can readily be sampled using volumetric samplers outside the tube. For example, the volume sensed by the EnviroSCAN capacitance sensor does not extend above or below the top of the upper electrode or the bottom of the lower electrode, respectively (Evetts et al., 2002a). The radial extent of the sensitivity is ~ 3 cm from the outer access tube wall for 92% response and is ~ 2.5 cm for 82% response (Paltineanu and Starr, 1997). Thus, the volume sensed is relatively small and concentrated near the access tube, where soil disturbance during tube installation is most likely and where it is not readily sampled using a volumetric soil sampler.

Sensors for which the calibration is linear are inherently easier to calibrate than those that have curvilinear calibrations. Linearity in the calibration allows the use of a wet-site/dry-site or two-point calibration. Examples in most soils are the NMM, for which water content is linear with count ratio, and conventional TDR with waveform analysis, for which water content is linear with the measured travel time. It has been claimed that since $(\epsilon_a)^{0.5}$ is essentially linear with water content for TDR, then values of permittivity estimated by other EM sensors will be linear with water content. However, there are two problems with this conjecture. First, the function $\theta_v(\epsilon^{0.5})$ is linear for TDR only because measurement frequency is large and BEC effects are relatively small. Second, many EM sensors do not provide an estimate of ϵ_a . Examination of Eqs. [27-28] shows that water content is non-linearly related to σ_a and the frequency of measurement. For EM methods, such as the capacitance methods, that use much smaller measurement frequencies than does TDR, it is expected that response would be non-linear, particularly in saline soils or those with appreciable amounts of high surface area clays.

If a probe is designed such that it can be inserted progressively into the soil, and if the probe response is linear along the length of the rods (eg. bifilar and trifilar probes TDR probes), then calibration data may be obtained by progressively inserting the probe into a soil of known water content, θ_i , typically saturated (Quinones et al., 2003). The progressive insertion method relies on calculation of an equivalent water content, θ_e , as $\theta_e = (x/L)\theta_i$, where L is the total length of the sensor rods and x is the length of rods inserted in the soil of water content θ_i . The sensor response (period, frequency, count, etc.) is measured for each insertion length and a corresponding equivalent water content calculated. The resulting set of values of sensor response and water content can be used in linear or nonlinear regression to determine coefficients for a calibration curve. Not all sensors exhibit a linear response along the length of the rods. Loiskandl et al. (2003) proposed a method for determining linearity in which they inserted the probe incrementally downward into free water. Insertion of 50% of the rod length should provide a 50% response (Fig. 2-25). Linearity of this sort is also essential when using the calibration method of Young et al. (1997), which involved upward wetting of a soil column in which a probe had been vertically inserted, and concurrent measurement of column mass and probe response (the height of the soil column and the length of the probe were identical).

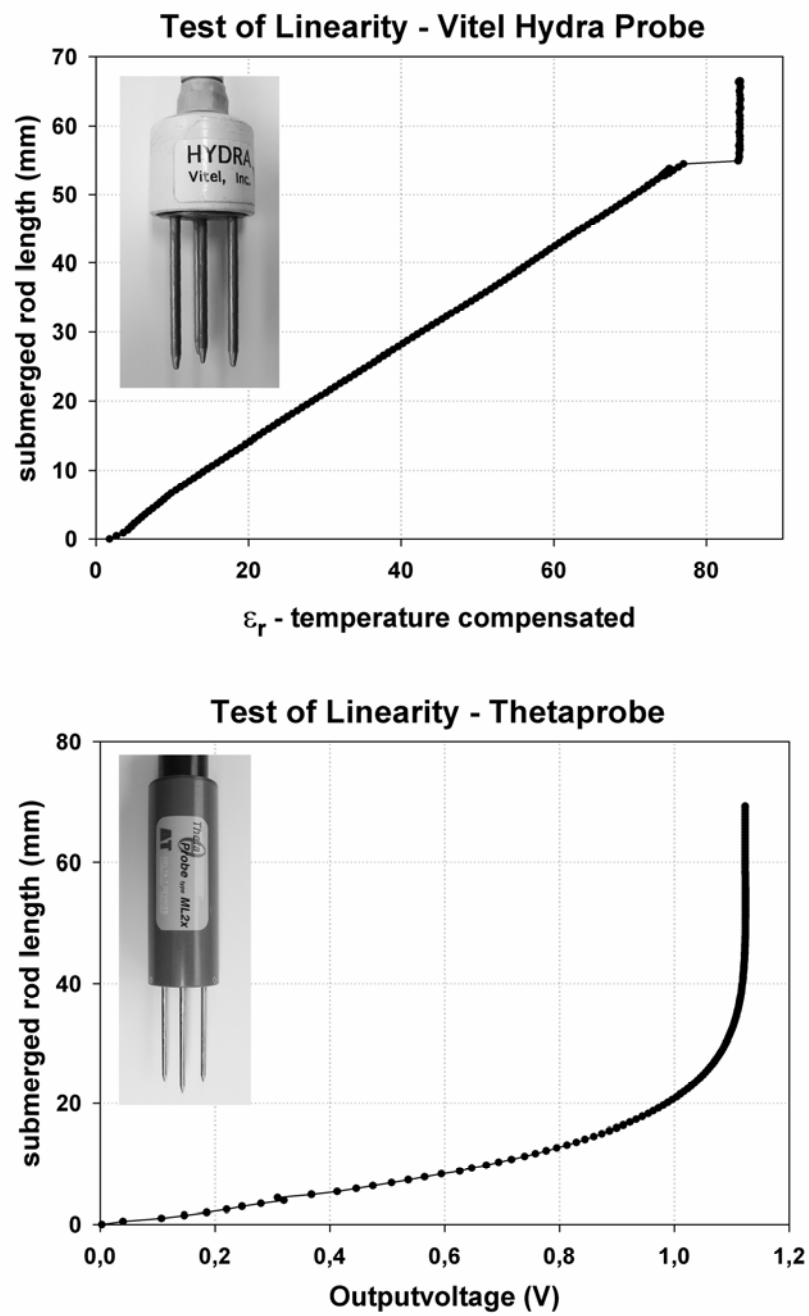


Fig. 2-25. Comparison of (top) linear response of a Vitel Hydra Probe and (bottom) non-linear response of a Delta-T Thetaprobe when inserted incrementally into a free water surface (with kind permission from Willi Loiskandl, BOKU-University of Natural Resources and Applied Life Sciences, Vienna, Austria).

Interference issues

All water content sensors are subject to interferences, but there are great differences among sensors as to the severity of the effect, the ease or possibility of correction, and the temporal nature of the interference. The NMM is relatively immune to interference from changes in temperature and salinity. It is subject to interference from changes in soil density, particularly in highly swelling clays such as Vertisols; and special attention is needed in those soils. There are interferences from soil salinity and temperature for all methods that measure properties (travel time, frequency, phase shift, etc.) of electrical circuits that are influenced by the electrical permittivity of the soil. These are aggravated in soils with large surface area (some clays, volcanic soils, soils high in organic matter) in which the bulk electrical conductivity (BEC) of the soil may change with both temperature and water content. In saline soils, the increase in BEC with water content is larger as the salinity level increases. It is important to understand that none of the electrical methods measures the soil bulk electrical permittivity or the soil water content. The severity of the interference differs greatly among specific sensor designs. But in general, the capacitance type sensors are more subject to interferences from inherent soil BEC, salinity-induced soil BEC, and temperature and water content effects on these properties. Conventional TDR with waveform analysis is probably the least subject to these interferences if cable lengths are kept short, but waveform analysis becomes impossible due to signal loss at soil BEC $>4 \text{ dS m}^{-1}$. In salt-affected soils, most of the electrical methods are simply impossible to use with accuracy due to the spatially and temporally variable patterns of BEC.

Depths of measurement

The depth to which water content sensing may be done and the depth-resolution of a particular method are often factors that limit the usefulness of that method. The depth of root water extraction varies with crop and soil, but can be $>2.4 \text{ m}$ for wheat, $>4 \text{ m}$ for alfalfa, $>2 \text{ m}$ for maize, etc. Equipment that limits measurements to a particular depth will not be suitable for water balance work with crops that have important root water uptake below that depth. For water balance work it is desirable to measure well below the maximum depth of root water extraction in order to close the water balance. Only a few methods allow much flexibility in depths of sensing or the maximum sensing depth. Two methods that do allow flexibility are neutron thermalization and the Trime T3 tube probe. Both of these use probes that are lowered in an access tube to any desired depth up to the length of the cable. Whereas cables for the NMM may be many tens of meters long, the length of the Trime cable is limited (approx. 3 m are provided) due to progressive loss of high frequency components of the TDR pulse along the length of the cable. Conventional TDR is also fairly flexible, allowing probes to be installed at any depth up to the length of the cable. However, soil disturbance is a limiting factor for installation of TDR or other probes that are inserted into or buried in the soil. Depth resolution is necessarily poor for those methods having a large measurement volume in all directions, notably the NMM. Conventional TDR probes and probes of similar methods (based on soil electrical properties) respond to only a small volume above and below the plane of the probe rods, and so are well-suited to measure thin layers of soil when installed

horizontally. The small volume of measurement of the EM methods does mean that sensors must be placed relatively close together in order to provide full coverage of the soil profile.

Other factors

Cost, ease of use, training requirements, durability, etc. are all factors that will influence the choice of a water content sensing method. Some systems, such as conventional TDR, have a large initial cost on the order of \$5,000 to \$10,000, but adding individual sensors is relatively inexpensive. With some frequency domain systems the cost of using one sensor is relatively low, but using many sensors becomes very costly. Automatic, stand alone data acquisition is usually possible with frequency domain and TDR sensors, but infeasible with the NMM due to regulations. Conventional TDR systems are typically more complicated to install, use and maintain than are systems built around frequency domain sensors.

Conjunctive Use

There is an unfortunate tendency to rely on a single soil water measurement system, but this need not be so. Several researchers have demonstrated the conjunctive use of two or more measurement systems. Usually, conjunctive use is based on combining the particular strengths of two or more systems. For instance, conventional TDR is quite capable of automatically recording the dynamics of water content in thin layers near the soil surface where water content changes most rapidly in most irrigation situations; whereas the NMM is capable of measuring deeply in the profile where changes are slower, and it provides very good coverage (no gaps in measurement) but cannot be used for unattended recording of dynamics. Combining two methods can provide complete coverage of soil water changes at appropriate time scales (Evett et al., 1993; Hunsaker et al., 2000; Kang et al., 2000; Murphy and Lodge, 2004; Suarez-Rey et al., 2000). Also, the NMM may be usefully combined with periodic surface soil sampling, either by gravimetric methods or by manually inserted and removed TDR probes. Another opportunity for conjunctive use occurs when a sensor designed for access tube use is found to provide a fairly faithful record of soil water content dynamics, but is inaccurate due to interferences. Calibration and use of the NMM in the same access tubes may provide a periodic record of more accurate water contents, which may be used to correct the record of soil water content dynamics.

CONCLUSIONS

Understanding and sensing of soil water content are essential for irrigation research and management. The science of soil water content is well developed, but the tools for sensing are undergoing rapid change, and the number of sensor choices is increasing. The reliable and well-understood NMM is being replaced in many situations by technologies based on electrical properties of soils. However, such systems are subject to many interferences related to the soil bulk electrical conductivity, which is highly variable in time and space for many soils. In particular, saline soils or soils irrigated with brackish waters are subject to large variations in BEC that defeat most sensors

based on soil electrical properties. All sensors must be tested for accuracy, and calibrated if not within acceptable limits for the intended use.

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