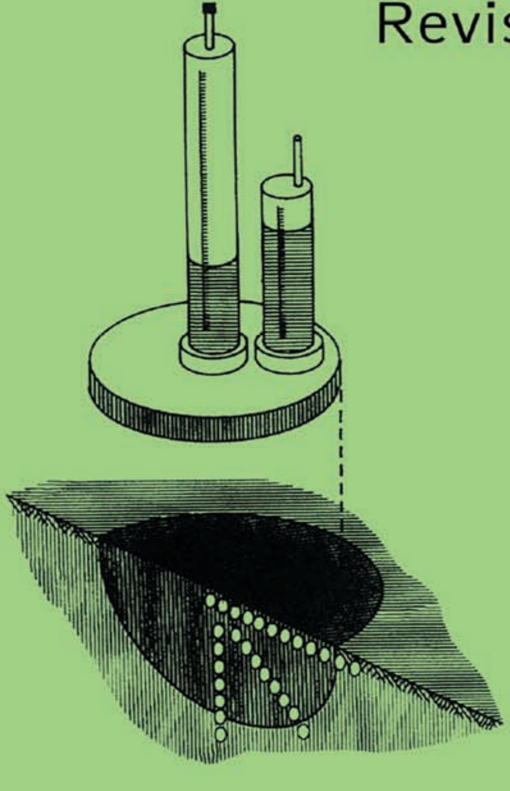
Soil and Environmental Analysis

Physical Methods

Second Edition
Revised and Expanded



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Soil and Environmental Analysis

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Preface

This second edition retains all of the topics covered in the first edition. Each chapter has been revised, to take account of new developments. The two separate contributions relating to penetrometer measurements have been combined into one chapter, and others have been somewhat shortened, in order to include new material on the measurement of infiltration, the measurement of soil strength and friability, and field methods of assessment of soil physical conditions. The chapter on gas movement and air-filled porosity now covers soil–atmosphere exchange of environmentally important gases, including radon and greenhouse gases.

While some topics have undergone relatively little change in terms of available methods or instrumentation in the period since the first edition appeared, some have changed considerably. The measurement of soil water, which has such an important role in soil physics and which underwent such a change when the neutron probe was developed, can now be undertaken with other sophisticated instruments. For example, time domain reflectrometry (TDR) and frequency domain systems, which share with the neutron method the desirable feature of allowing nondestructive measurements at the same site to study temporal variations, now provide a reliable alternative to the neutron probe, while avoiding the problems of radiation protection. The widespread availability and use of data loggers has also transformed our approach to many measurements, particularly water content, matric potential, penetrometry, and soil thermal properties, and placed a greater emphasis on those instruments that can be logged.

Like the previous edition, this book is aimed at the researcher or agricultural or environmental adviser working in environmental science, soil science, or a related field. It should also be useful to teachers and students in postgraduate courses in soil science, soil analysis, and environmental science. One of the significant trends of the past few years has been the development of interdisciplinary activities, in the attempt to improve understanding of complex phenomena in the life and environmental sciences. This places new emphasis on the concurrent measurement of physical, chemical and biological parameters. One typical example of this is the study of losses of nitrogen from soils into waters and the atmosphere, where information may be needed on soil water infiltration, saturated and unsaturated flow, and water-filled pore space—all of which require physical measurements—as well as on soil mineral nitrogen analysis and plant growth. Researchers who may have trained in chemistry or biological sciences now need to become informed about physical techniques as well. In this book we attempt to provide an introduction to each type of measurement, with enough theory to teach the principles behind the methods, and to help in the selection of methods appropriate to the task at hand.

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Soil and Environmental Analysis

1

Soil Water Content

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I. INTRODUCTION

Measurement of the water content of soil and the unsaturated zone is fundamental to many investigations in agriculture, horticulture, forestry, ecology, hydrology, civil engineering, waste management, and other environmental fields. While other factors related to soil water are important, probably the single most useful piece of information about soil water is knowing how much is present, either in a complete profile or within a well-defined volume.

The diverse range of applications means that there is a wide range of demands on the measurements. Some objectives require a single measurement of total soil water content in a field profile, whereas others demand repeated measurements of the spatial distribution of water content to track changes over time. The time scales may vary from minutes to months. Measurements may be undertaken in the laboratory, on loose or repacked samples, on undisturbed cores, in plant containers or lysimeters, or as part of field experiments, trials or larger, catchment scale, studies. The measurement precision and accuracy demanded varies widely and hence so does the sophistication of the methodology which must be employed. As a result of this wide range of demands, no one method can satisfy all requirements. However, three methods are used for the vast majority of determinations today: the thermogravimetric method, neutron thermalization, and a group of techniques based on measurement of soil dielectric properties.

The oldest established and the only truly direct method is the thermogravimetric method, which requires samples for oven-drying. The other two techniques rely on measurement of physical properties of the soil that depend on its water content. The neutron method was adopted for routine use in the 1960s and has been popular ever since, although the radiation hazard and cost preclude semipermanent installation and hence automation. The development of dielectric methods since 1980 has introduced opportunities for rapid collection of soil water content data at short time intervals, five minutes or less if required, and permitted automation and logging of measurements. The ability to log soil water content automatically is opening up ways of soil water monitoring and soil hydrological research that have hitherto been impossible.

In this chapter, the concept of soil water content, definitions of the water content of a block of soil, and the terminology and units used are described briefly. The relative merits of direct and indirect measurements and the spatial and temporal resolution that can be achieved by various methods are considered. The principles and practice of the three methods are then discussed in detail and applications of the neutron and dielectric methods are described. A summary of the more common alternatives to the three major ground-based methods for soil water content measurement, referred to above, is provided in Table 1. A review of techniques for remote sensing of soil water, which complement ground-based techniques, is also provided.

II. SOIL WATER CONTENT

A. Definition

The term "soil water content" is widely accepted as referring to the water that may be evaporated from a soil by heating to between 100 and 110°C, but usually at 105°C, until there is no further weight loss. This is the basis of the thermogravimetric method. It is important to be aware of the arbitrary nature of this definition, which is the standard reference against which other techniques are normally calibrated. As Gardner (1986) stated, "the choice of this particular temperature range appears not to have been based upon scientific consideration of the drying characteristics of soil." Its origin probably has more to do with the notion of ensuring evaporation of liquid or "free" water and the relative ease with which determinations can be made by oven-drying samples.

Water is present in soil as water vapor and liquid. In addition, water molecules are adsorbed in layers on the surfaces of colloidal materials, particularly clays, and molecules are incorporated with hydroxyl groups within clay lattice structures. The distinctions between thin films of water retained by surface tension and water that is adsorbed (bound water), and between bound and structural water, are less precise than this categorization suggests. Water vapor and structural water are disregarded in the conventional definition of soil water content. Structural water is immobile and is generally released only upon heating to temperatures

Table 1 Alternative Methods of Soil Water Content Measurement

Method	Use	Principle	Application	References/ Comments
Calcium carbide method	Lab/field, on samples	Calcium carbide mixed with soil in pressure chamber produces acetylene gas; gas pressure depends on soil water content	Civil engineering purposes as well as agricultural	Morrison (1983)
Sulphuric acid method	Lab, on samples	Concentrated sulphuric acid mixed with soil raises temperature; maximum temperature depends on soil water content	Mainly agricultural	Gupta and Gupta (1981)
Soil matric potential	Lab/field in situ	Soil matric potential measurements are translated into water content using the water release characteristic. As the matric potential–water content relationship is hysteretic, precise determination of water content is not possible. Assumes the soil water release characteristic is known	Field or lab where soil matric potential measurement required but water content measurement precluded, e.g., irrigation	See Chapter 2 for details of measuring soil matric potential
Gamma ray attenuation	Lab/field in situ	When soil is irradiated with gamma rays, the scattering and absorption which occur are primarily a function of soil density. In nonshrink–swell soils, temporal variation in total bulk density is due to water content change and therefore gamma ray attenuation or backscatter can be used to monitor water content	Experimental conditions only due to cost and radioactive hazard. Used in lab and field	See Chapter 7 for details of gamma ray methodology. Wood and Collis George (1980); Morrison (1983)
Nuclear magnetic resonance spectroscopy (NMR)	Lab samples, field in situ	Atomic nuclei change their energy levels when subjected to oscillating electromagnetic fields; different frequencies affect different nuclei, but hydrogen nuclei give the strongest response. Electronic detection of either the energy absorption or nuclear dipole excitation gives the NMR signal. NMR measurement of hydrogen concentration is related to water content by calibration	Geophysical use in boreholes. Experimental NMR equipment for field measurements on samples and of surface water content has been described	Paetzold and Matzkanin (1984); Paetzold et al. (1985)
Thermal conductivity	Lab/field in situ	An electrical heating element and a temperature sensor are placed in soil either directly, or encased in a porous block. The time for a given temperature to be achieved after heat is applied is measured. The rate of heat dissipation is a function of soil thermal diffusivity, which depends on soil water content	Mainly agricultural. Direct contact probes require good contact with soil; blocks respond to soil water potential—see above. Usable in very saline soils	Fritton et al. (1974); Sophocleus (1979)

between 400 and 800°C; an exception is gypsum, from which structural water is lost at only 80°C. Bound water does have a degree of mobility which becomes important at very low water contents and may be exploited by drought-resistant plants. Heating to 105°C is not normally sufficient to remove bound water; most is eliminated from clay surfaces at temperatures between 110 and 160°C.

The conventional definition of soil water content is not a limitation in most work because the quantities of bound and structural water are small relative to the "free" water content and can be assumed to be constant for most purposes. In practice it is usually changes of soil water content with time that are of interest (e.g., seasonal changes in field soils or change in response to irrigation). Alternatively, the quantity of water retained between specific thresholds may be required (e.g., between the liquid and plastic limits or between "field capacity" and "wilting point"). Several methods of water content determination, including the neutron probe and dielectric methods, are sensitive to all the water molecules present in a soil, although this information is effectively lost as they are calibrated against thermogravimetric determinations. Dielectric methods have the potential to discriminate between liquid, bound, and structural water, but this has yet to be exploited.

B. Units

Soil water content may be expressed on either a mass or a volumetric basis, that is, as a mass ratio, kg kg^{-1} (kg water per kg dry soil), or a volume fraction, $\text{m}^3\ \text{m}^{-3}$ (m^3 water per m^3 of bulk soil volume), respectively. In either case the value is a dimensionless fraction and can be multiplied by 100 to express it as a percentage. One can be obtained from the other if the dry bulk density of the soil, and the density of water, are known:

$$\theta = \frac{w\rho_\text{b}}{\rho_\text{w}} \tag{1}$$

where θ is volumetric soil water content (volume fraction), w is water content as a mass fraction, ρ_b is the dry bulk density of the soil (kg m^{-3}), and ρ_w is the density of free water (usually approximated as 1000 kg m^{-3}). For most purposes, expression as a volume fraction is more useful, since multiplying θ by the soil depth gives the "depth" of water in that depth of soil, a figure with the same (length) dimensions used to express rainfall, evaporation, transpiration, drainage, and irrigation.

Because the thermogravimetric method is used as a standard for calibration, soil bulk density as well as water content measurements are required to calibrate techniques that measure volumetric water content, unless undisturbed samples of known volume are obtained for oven-drying. This introduces an additional source of error into the calibration. Since a technique can be no more accurate than the procedure used to calibrate it, particular care is required in both the water content and the density determinations when undertaking calibrations.

If soil water content is monitored at several depths in a core or a soil profile, the depth interval z_i to which a measurement θ_i refers is normally taken as the vertical distance separating the two midpoints between the measurement depth and the depths of the measurements immediately above and below it. The water content of the soil profile, P, to a depth z, is obtained by summation of the water contents of each depth interval:

$$P = \sum_{0}^{z} \theta_i z_i \tag{2}$$

The effect of this integration of a step function of the water content is equivalent to trapezoidal integration; although little used, Simpson's rule can reduce the errors involved (Haverkamp et al., 1984).

C. Direct *Versus* Indirect Measurements

Direct measurements involve removal of soil water by evaporation, leaching, or a chemical process, and subsequent determination of the amount of water removed; the thermogravimetric method is the principal example. Direct measurements are beset with problems primarily due to the need for destructive sampling. Thus replicate samples must be taken to determine the variance of measurements made on a given occasion and whether they differ significantly from measurements made on other occasions. This replication can result in the handling of very large numbers of samples. Practical difficulties are compounded if determinations deep in the profile are required. Furthermore, repeated sampling within the same area may cause unacceptable damage to the soil or vegetation present. Provision must also be made for bulk density determinations if volumetric water content data are required. However, taking undisturbed cores of known volume to determine both water content and bulk density avoids this.

Indirect methods depend on monitoring a soil property that is a function of water content (e.g., the basis of the neutron method is detection of hydrogen nuclei in soil, most of which are present in water molecules). Indirect methods usually involve instrumentation placed in or on the soil, or remote techniques involving sensors mounted on a platform over the soil or on aircraft or satellites. Although indirect measurements require calibration, most have the considerable advantage that measurements on the soil in situ are possible and these can be repeated at the same place through time.

Another significant advantage is that change in soil water content is determined directly. The standard error of estimation of change of water content obtained from repeated measurements on the same n samples is simply

$$s.e.(\Delta\theta) = \sqrt{\frac{\mathrm{var}(\Delta\theta)}{n(n-1)}} \tag{3}$$

whereas the standard error associated with a change in water content based on direct measurements made on two sets of n_1 and n_2 independent samples, depends on the variances attached to both sets of samples:

$$s.e.(\overline{\theta_2} - \overline{\theta_1}) = \sqrt{\frac{\mathrm{var}(\theta_1)}{n_1(n_1-1)} + \frac{\mathrm{var}(\theta_2)}{n_2(n_2-1)}} \tag{4}$$

In the latter case, the variation in the water content on each measurement occasion is superimposed on the spatial variation of the change in water content. Therefore, if changes of water content are the focus of interest, rather than absolute water contents, indirect in situ measurements are preferable to direct measurement that involves removing samples.

D. Spatial Resolution of Measurements

The thermogravimetric, neutron, dielectric, and remote sensing methods between them cover various measurement scales in three dimensions (Fig. 1). Most measurements integrate over a volume around a position in the soil, the size of which depends on the technique used, or may be defined by the size of a sample or core taken to the laboratory. Oven-drying of a soil sample produces an integrated water content measurement for that sample. Most instruments integrate the water content unevenly over a volume of soil, with the largest contribution coming from the region close to the sensor. The size of the volume measured is frequently dependent on the water content of the soil. The neutron depth probe measures a sphere of soil, 0.2 to 0.5 m in diameter, centered approximately on the source. Many dielectric instruments have parallel rod type sensors that are usually most influenced by the soil between and immediately around the rods and so measure a roughly cylindrical volume, the length of which is determined by the length of the rods; the measurement integrates the water content along the sensor. Rod spacing in most equipment implies a cylinder of 50 to 100 mm diameter, and rod lengths range from 50 mm to 1 m. In deciding which measurement method to employ, it is important to consider the volume of soil that the measurements will represent and how water content or other gradients within that volume (e.g., wetting fronts, density, or mineralogical differences) may influence the measurements made.

Many techniques make what are referred to as "point measurements." In practice this is actually a measurement of soil water content within a finite volume which is small compared with the overall scale of the area and/or depth range

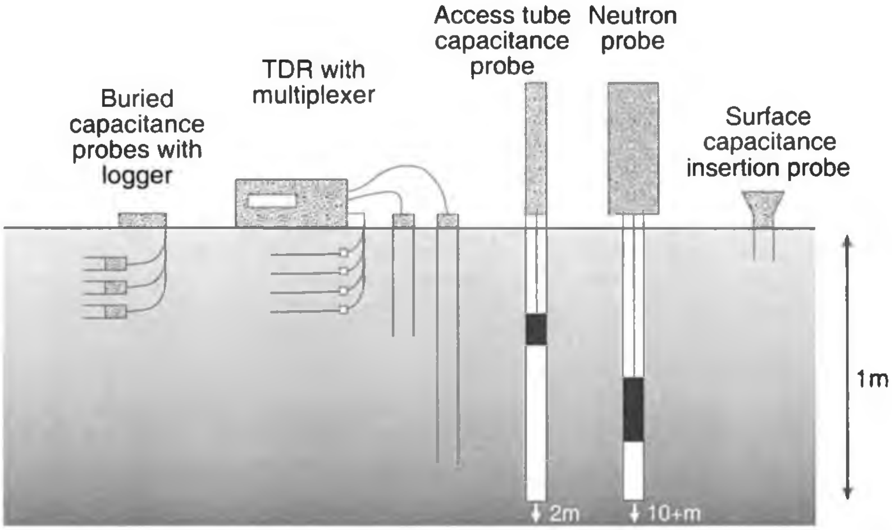


Fig. 1 Spatial arrangement of soil water sensors for in situ measurement. Sensors for dielectric methods (capacitance and time domain reflectometry, TDR) can be installed semipermanently and operated automatically. Installation of access tubes permits manual use of neutron or capacitance depth probes. Capacitance and TDR instruments can also be used for one-off readings at the soil surface.

under study. Water content information is often required over large areas, but research is only now addressing how to make the leap from "point" to areal measurements. Remote sensing techniques are potentially very useful in this respect; although they only allow measurement of water content in the surface soil, the combination of this with point measurements at greater depth, and/or modeling of changes with depth, has considerable potential that has yet to be fully realized.

III. THE THERMOGRAVIMETRIC METHOD

The thermogravimetric method is straightforward. A soil sample is placed in a heat-proof container of known weight, weighed, dried in an oven set at a constant temperature of 105°C, removed and allowed to cool in a desiccator, then reweighed. This procedure is repeated until the sample attains a constant mass (ISO, 1993). The water content, w, of the sample is the mass of water per unit mass of dry soil:

$$w = \frac{\text{Mass of wet soil} - \text{Mass of dry soil}}{\text{Mass of dry soil}} \tag{5}$$

If a sample of known volume obtained by coring is used, the volumetric water content can be obtained directly:

$$\theta = \frac{\text{Mass of wet soil} - \text{Mass of dry soil}}{\text{Soil volume}} \tag{6}$$

(ISO, 1997). An oven temperature of 105 ± 5°C and a 24 hour drying period are widely adopted. Drying time is influenced by the oven's efficiency and the condition, size, and number of samples in it. 24 hours may be insufficient for some soils and especially large wet samples (Reynolds, 1970), but unnecessarily long when making determinations on small or air-dried samples. Constant mass is defined as that reached when the change in sample mass, after drying for a further 4 hours, does not exceed 0.1% of the mass at the start of the 4 hours (ISO, 1993, 1997).

An oven ventilated by a fan that distributes the heat evenly is required. The drying temperature should be checked periodically using a thermocouple in a dry soil sample. Oven efficiency can be checked by loading it with subsamples of a well mixed moist soil and checking the variation in water content measured. A balance capable of weighing to better than 0.1% of the mass of the dried samples is required. Analyses of the random errors accompanying gravimetric water content determination due to varying degrees of weighing precision and accuracy were provided by Gardner (1986).

Recommended sample sizes range from 10 to 100 g (Australian Water Resources Council, 1974), but 50 to 100 g is preferable for moist samples. If volumetric water content is to be obtained, undisturbed cores of at least 20 cm^3 should be collected and dried. For stony soils, larger samples are necessary; recommendations according to the dimensions of the aggregates and stones in the moist soil are available (ASTM, 1981). Variation of the proportion of stone in samples may cause problems, in which case the water content of the < 2 mm fraction, $\theta_{<2}$, and the volume of the stone (> 2 mm) fraction, S, are determined (Reinhart, 1961). The water content of the whole soil is

$$\theta = \theta_{<2}(1 - S) \tag{7}$$

The water content of the stone fraction, θ_s, is often considered to be negligible (Hanson and Blevins, 1979) but may not be, in which case it should be determined by oven-drying as for the soil and included in the calculation of θ.

When dealing with organic soils, some inaccuracy in water content determination may occur due to the oxidation and decomposition of organic matter at 105°C, causing weight loss other than that due to water evaporation. In certain soils, volatilization of substances other than water may occur at temperatures below 105°C, causing similar problems. Lower drying temperatures may be considered when working with soils where this occurs but can lead to determination of significantly lower water contents.

Because of its simplicity, the oven-drying method is easily abused. In particular, oven temperatures may not be checked and neither they nor the drying time are usually reported. Common problems include drying of the soil during transit before weighing, loss of soil in transit, water uptake from the atmosphere during cooling because no desiccator was used, and poor determination of the volume of the core or the dry bulk density. The use of thermogravimetric determinations as a reference against which to calibrate and investigate the accuracy of other methods of water content measurement requires special care in its application. The advantages of this method are its simplicity, reliability, and low cost in terms of equipment requirements. The disadvantages are the need for destructive sampling, the time required for drying, and the staff time needed to deal with large numbers of determinations.

Drying time may be reduced to ≤ 20 min with the use of microwave ovens, but there are two problems inherent in this approach: drying time increases with initial water content; and, if a dry sample is left in a microwave oven, its temperature will continue to rise beyond 105°C which may cause weight changes other than those due to evaporation of water. Consequently, drying times must be estimated initially. Microwave drying can give water content determinations within 0.5 to 1.0% of those obtained using conventional oven-drying, if trials are conducted to determine appropriate drying times (Gee and Dodson, 1981; Tan, 1992). For some purposes the method may be suitable, but for best accuracy the use of a conventional oven is recommended (Standards Association of Australia, 1986).

IV. THE NEUTRON METHOD

The neutron method uses the ability of hydrogen to slow down fast neutrons much more efficiently than other substances. In any soil, most of the hydrogen is present in water molecules, and therefore changes in hydrogen concentration occur mainly due to changes in water content. A radioactive source, continually emitting fast neutrons, and a slow neutron detector, are housed within a probe that is lowered into the soil down an access tube. The fast neutrons are slowed as they move through the soil. The number of slow neutrons detected is recorded as a count rate and is converted to volumetric water content using a calibration relationship. For depth measurements in soil, an access tube is installed semipermanently and readings are made at successive depths by lowering the probe within the tube. Measurements can be made with ease to depths of 5 m or more in many soils, once the effort of access tube installation has been completed. Neutron meters of different design for use at the soil surface are also available.

The neutron method was first proposed in the 1940s (Brummer and Mardock, 1945; Pieper, 1949) and field tests soon followed (Belcher, 1950). By the mid-1950s, portable instruments for field use had been developed in North

America (Underwood et al., 1954; Stone et al., 1955) and Australia (Holmes, 1956). Equipment soon became available commercially. Instruments available today are considerable refinements of the early designs. Technological developments have permitted use of less hazardous neutron sources, reducing the amount of shielding required and allowing smaller, lighter yet safer designs. The electronics are more reliable and data can now be stored and processed on board.

The emphasis here is on neutron depth probes; surface meters are only considered briefly. Dual-purpose depth probes that measure soil bulk density by gamma ray attenuation (see Chapter 8), and water content by the neutron method, are also available. Three reports, although published some years ago, still represent the most comprehensive accounts of the theoretical and practical aspects of using neutron depth probes (IAEA, 1972; Greacen, 1981; Bell, 1987) and are recommended for further detail. Use of neutron depth probes is now well established, and standard procedures have been agreed upon (ISO, 1996).

A. Neutrons and Neutron Moderation

Neutrons are uncharged particles of mass very slightly greater than a proton. They are classified according to their kinetic energy measured in electron volts (eV). Fast neutrons have kinetic energies exceeding 1 keV. Thermal neutrons have energies of 0.025 to 0.5 eV and are close to thermal equilibrium with the molecules of the surrounding medium; their movement through the medium is controlled by the gas diffusion laws.

Because they have no charge, neutrons are not influenced by electric fields. They are therefore able to penetrate through the electron cloud of an atom to reach the nucleus. When a neutron comes close to, or collides with, a nucleus, a variety of interactions may occur depending on the energy of the neutron and the characteristics of the nucleus. The probability that collisions resulting in a given interaction will occur when a substance is irradiated with neutrons of a given energy is defined by the interaction cross-section of the isotope, measured in units of area called barns; 1 barn is 10^{-28} m^2. The greater the cross-section, the greater is the probability of interactions. The macroscopic interaction cross-section of a unit volume of soil is calculated as the weighted sum of the values for the individual elements present. There are two types of neutron–nucleus interaction: neutron scattering and neutron capture.

1. Neutron Scattering

Scattering occurs when the collision of a fast neutron with a nucleus causes the neutron's direction of travel to change and its velocity, and so kinetic energy, to reduce. Such collisions may be elastic, i.e., kinetic energy and momentum are

Table 2 The Effect on Fast Neutrons of Collisions with Nuclei of the Commonest Elements in Soils

Nucleus	% energy lost in head-on collision	Average number of collisions to slow 2 MeV neutron to <0.5 eV
Hydrogen	100	18
Oxygen	22.1	152
Silicon	13.8	252
Aluminum	13.3	279
Iron	6.8	519

Source: Hodnett, 1986.

conserved, or inelastic, i.e., some of the neutron's energy is transferred to the nucleus, resulting in the emission of gamma radiation. Inelastic scattering is unimportant in the present context. The elastic scattering cross-section of most elements is small, less than 5 barns, and relatively constant at neutron energies between 2 eV and 2 MeV.

The loss of energy by a neutron in the course of elastic scattering is inversely related to the mass of the nucleus with which it collides. When a head-on collision takes place with a hydrogen nucleus, the neutron loses all of its energy. In practice, collisions occur at all angles, and many are required to slow a fast neutron (Table 2). Heavier nuclei are most likely to deflect a neutron through a greater angle from its original path without significant loss of energy. Collisions with heavy nuclei therefore reduce the distance that fast neutrons move from a source before they are slowed to thermal energies.

2. *Neutron Capture*

Some collisions between a neutron and a nucleus result in the neutron being absorbed (captured) by the nucleus. The capture cross-section depends on both the type of nucleus and the energy of the neutron. For most elements, it is negligible for neutron energies greater than 1 eV, so only slow neutrons are likely to be captured. The capture cross-section for most soil constituents is between 0.1 and 1 barn, but some elements have much larger values. Important examples are gadolinium (46,000 barn), cadmium (2,450 barn), and boron (755 barn). A trace of one of these in soil will greatly increase the soil's macroscopic capture cross-section and so reduce the slow neutron count rate markedly, thus affecting the calibration curve. Other more common elements, such as manganese (33 barn), chlorine (33 barn), and iron (2.6 barn), may have a significant effect if present in sufficient quantity. Capture reactions with certain elements result in emission of alpha particles or protons, and this is the basis on which slow neutron detectors operate.

B. Neutron Sources, Detectors, and Instrument Design

Fast neutron sources usually contain two elements: one emits alpha particles in the course of radioactive decay; the other is beryllium, which absorbs the alpha particles and in the process emits fast neutrons. The reaction is

$$^{9}_{4}\mathrm{Be} + ^{4}_{2}\mathrm{He} \rightarrow ^{1}_{0}\mathrm{n} + ^{12}_{6}\mathrm{C} + 5.74\ \mathrm{MeV}$$

The neutron emitted gains some of the reaction energy plus some of the alpha particle's energy. Most probes use sources with an isotope of americium, ^{241}Am, as the alpha emitter. It has a half-life of 458 years. Source activity in modern probes is usually 1.85 GBq (50 mCi) or less. The sources are constructed to strict safety standards: finely powdered beryllium and sintered americium oxide are contained within a double-walled capsule of stainless steel that is cylindrical or annular in shape. Their working life is at least 20 years, but regular tests for leakage should be conducted (Lorch, 1980).

Improvements in the detection efficiency of thermal neutron detectors have enabled use of lower activity sources in probes. The isotopes ^{10}B, ^{3}He, and ^{6}Li have very high capture cross-sections for neutrons of energy less than 1 eV and are relatively insensitive to high-energy neutrons. Boron trifluoride and helium-3 filled metal tube detectors are most common. Both require a stable 1 to 2 kV supply to operate. Lithium-enriched glass scintillation counters can give 100% detection efficiency but are more complex and delicate than gas counters. They can monitor gamma radiation separately from thermal neutrons and so are useful in dual-purpose probes. The efficiency of a detector declines slowly with time but the useful life is at least 15 years.

The arrangement of the source and detector within the probe contributes to its sensitivity to water content change. Certain geometries result in a linear calibration for the range of water contents commonly encountered. Ideally both source and detector would be placed at the same point, to give a symmetrical distribution of thermal neutrons about the detector. Some designs use an annular source fitted around the midpoint of the detector to achieve a symmetrical arrangement. If the detector is remote from the center of the neutron cloud, a nonlinear calibration results, and the influence of interfaces in the soil and at the surface is exacerbated.

Most neutron depth probes comprise six parts: the probe (containing the source and detector), which is connected by cable to the counting unit; the cable; the counting unit; the power supply; the probe carrier; and a system for lowering the probe into an access tube and locating it at given depths (Fig. 2). The counter unit measures the electronic pulses transmitted from the detector and displays the result. Most instruments count for a preset time, typically between 4 and 64 seconds. Longer count times can be selected on some instruments for high-precision

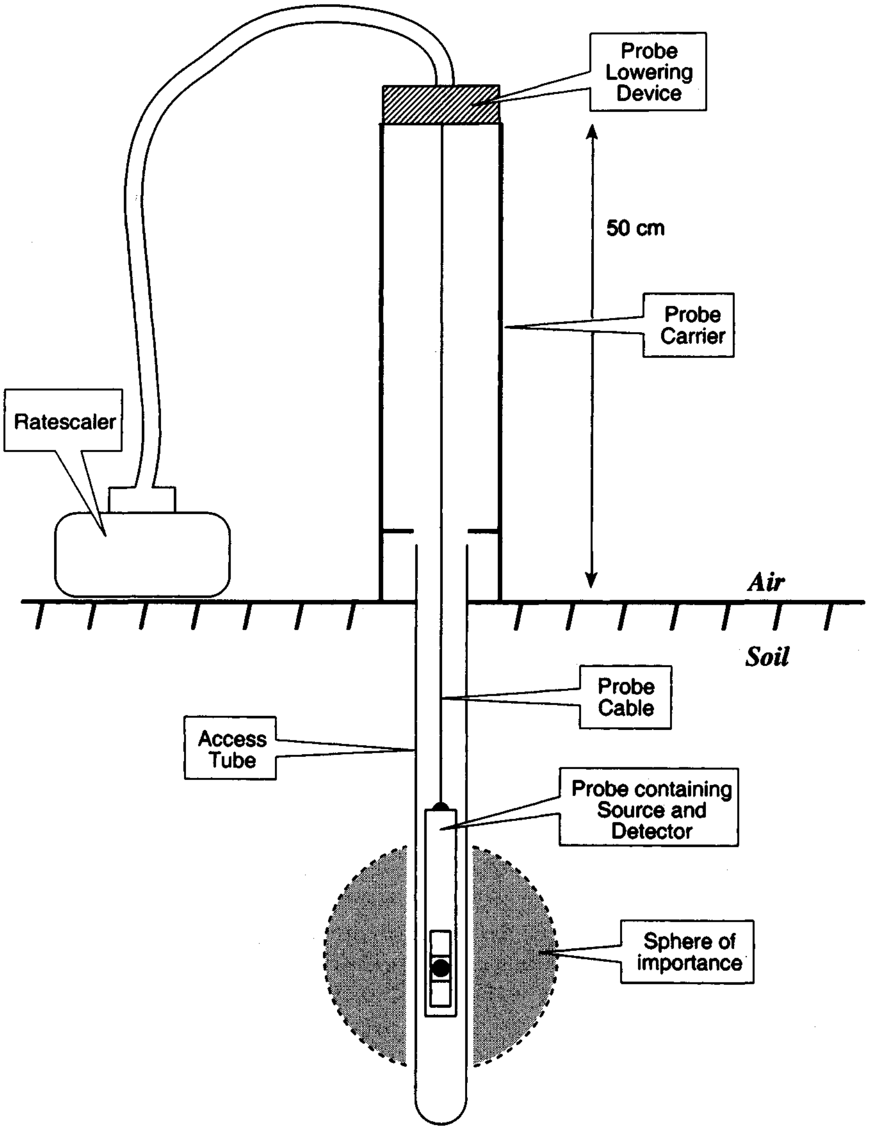


Fig. 2 Principal components of the neutron depth probe. The sphere of importance designates the volume of soil that contributes to the reading.

measurements. Nicolls et al. (1981) provide a useful account of instrument design in relation to sensitivity, accuracy, precision, and convenience of use.

C. Standard Neutron Count Rates

As indicated above, neutron depth probes of different design have different calibrations. However, the sensitivity of instruments of the same design is not identical either, due to differences in source strength and detector efficiency. To ensure data compatibility if slow aging of components occurs, if a component is replaced or a probe is otherwise repaired, or if more than one probe of the same type is in use, neutron count rates in a standard medium should be made at regular intervals. Calibrations should be made in terms of count rate ratio R/R_s, where R is the count rate in soil and R_s the standard neutron count rate. Data from probes of different designs cannot be normalized in this way, but intercalibration is possible (Nakayama and Reginato, 1982). Weekly standard counts are recommended, but if a probe is used less frequently, a standard count should be made before or after each reading occasion. A count time of 1 h minimizes the random error of the standard count, and so of water content measurements obtained with that count.

The use of a water standard is preferred to that of other hydrogen-rich media, such as plastics, because the count rate is almost independent of temperature and there is no possibility of water absorption from the atmosphere (Hodnett and Bell, 1990). A water standard can be cheaply constructed by fixing a water-tight access tube axially in the center of a drum that is then filled with water. The drum should be at least 0.6 m deep and 0.5 m in diameter to ensure that the water surrounding the source, when it is lowered into the access tube, effectively represents an infinite volume.

Some manufacturers suggest taking standard counts in the probe transport shield. This is not advisable, because the shield is not large enough to represent an infinite medium and therefore the counts are easily influenced by surrounding neutron moderators. In addition, temperature and humidity also affect the count rate. Precautions to overcome these shortcomings have been described (Hauser, 1984) but serve more to emphasize the simplicity and reliability of using a water standard.

D. Neutron Movement in Soil—The "Sphere of Importance"

A neutron emitted from the source of a probe travels outward into the soil until it collides with an atomic nucleus. Some energy is lost in the collision and the direction of travel altered. The neutron continues in the new direction until another collision occurs. Most neutrons migrate away from the source, but a proportion return, having been slowed in the process. The further a neutron gets from the source, the smaller its chance of returning; this is particularly so once thermal energies have been attained, as the probability of absorption is then greatly increased. The soil closest to the probe therefore has the greatest influence on the count rate measured. For working purposes a "sphere of importance" can be defined. The center of the sphere of importance lies between the source and the center of the detector. If the source is placed at the center of the detector, these are coincident. The sphere of importance is defined as that which, if the soil and water surrounding the sphere were removed, would result in a thermal neutron count that was a given fraction, usually 0.95, of the count if the medium were infinite in extent (IAEA, 1972).

The size of the sphere of importance depends on

1. The energy spectrum of the neutrons emitted from the source (the type of radionuclide in the source but *not* the source strength)
2. The neutron scattering and capture cross-sections of the soil and its bulk density
3. Soil water content

While the effects of 1 and 2 are constant for a given probe and soil, the influence of soil water content changes with time. The sphere's radius decreases as water content increases, because the greater hydrogen content causes more neutron scattering close to the probe, restricting movement of neutrons away from it. The radius of the sphere of importance of most depth probes with americium–beryllium sources is about 0.15 m in wet soil, increasing to about 0.5 m in very dry soil.

Since water content measurements are thus made on a sizeable volume of soil, there is little advantage to be gained from making readings at depth intervals of less than 0.1 m. When measurements are made through an interface between wet and dry soil, the measurements in the wet soil close to the interface will indicate that the soil is drier than is actually the case. Conversely, the water content of the dry soil near the interface is overestimated, but to a lesser degree than the underestimation for the wet soil (Hodnett, 1986). This effect increases with the difference in water content between the layers. The shape of the measured water content profile is smoothed, and so neutron probes are not suitable if measurements with good depth resolution are required. The slight underestimation of the total soil profile water content is usually disregarded. However, Van Vuuren (1984) found that the bias so introduced can be significant and advocated use of field calibrations to allow for site-specific properties such as the presence of a water table. Wilson (1988a) analyzed the phenomenon and demonstrated theoretically that it would be unwise to rely on measurements closer than about 0.25 m to a marked discontinuity such as a water table.

E. Random Counting Errors

Both radioactive decay and thermal neutron counting are random processes. When repeated neutron counts are made using the same time interval, the number of counts recorded varies. This is an important source of random error in the measurement. (Other errors may arise from changes in the placement depth, calibration uncertainties, thermal effects on the electronics, and warm up.) Repeated counts fit a Poisson distribution. For this distribution, if N is the total number of counts recorded over a time, t, the standard deviation of the mean value of N is $\sqrt{N}$. It is usual to work with a count rate, R, where $R = N/t$, and so the standard deviation of R is

$$\sigma_R = \left(\frac{R}{t}\right)^{0.5} \tag{8}$$

Therefore, if the time taken to obtain a count is increased, the standard deviation of the mean decreases. The absolute error accompanying greater count rates obtained in wet soils is always greater than in dry soils, because if counts are made over a fixed time interval, R is greater, whereas if N is fixed, t is reduced.

The standard deviation of a standard count determination is $(R_s/t_s)^{0.5}$, and that of a water content determination is

$$\sigma_\theta = a\frac{R}{R_s}\left(\frac{1}{Rt} + \frac{1}{R_s t_s}\right)^{0.5} \tag{9}$$

where a is the slope of the calibration curve, R_s is the standard count rate (s^{-1}), and t_s is the standard count time (s). Since the standard count itself introduces a small error, long standard count times of an hour or so should be used, if possible, to minimize that source of error. The depth of water in a layer of soil is obtained by multiplying θ by the layer depth. Similarly σ_θ is multiplied by the layer depth to give the standard error of the layer value. The error associated with the profile water content value is the square root of the sum of squares of the errors attached to the individual layer values.

For field measurement purposes, the advantages of the greater precision obtained at one location associated with longer count times (Fig. 3) needs to be balanced against uncertainties arising from spatial variability of soil water content. Because of the latter, it is usually preferable to conduct measurements in many tubes using a short count time. This provides a better estimate of both the mean water content and its variability than more precise data from fewer tubes.

F. Field Measurements

Before measurements can be made with a depth probe at a new site, access tubes must be installed, measurement depths must be selected, and decisions regarding soil calibration and how to deal with measurements close to the soil surface are necessary. Measurement intervals of 0.1 or 0.2 m, perhaps increasing to 0.3 m at greater depth, are generally appropriate. Once a set of measurement depths has

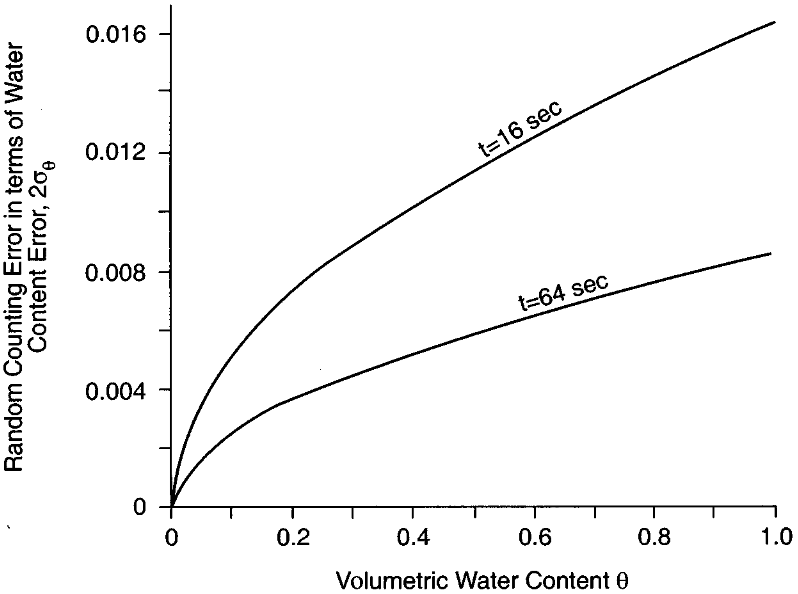


Fig. 3 Relationships between water content error, $2\sigma_\theta$, resulting from the random counting error, and water count, for counting periods of 16 and 64 seconds. (After Bell, 1987.)

been established, it is important to adhere to it. If the depths are changed, the two sets of data will not be strictly comparable because different parts of the soil have been measured. For the same reason, it is important that the chosen measurement depths are accurately maintained on every measuring occasion.

1. Access Tubes

The factors to consider in selecting material for access tubes are transparency to neutrons, mechanical strength and resistance to corrosion in the soils to be investigated, as well as cost and availability. Aluminum, aluminum alloy, stainless steel and some plastics are all suitable; their relative merits are given in Table 3. Aluminum alloy tubing is usually preferred. Polyvinyl chloride (PVC) is not recommended because the chlorine content considerably reduces the neutron count. The iron content of stainless steel has a similar, though less serious, effect, but for some applications the strength is required.

The internal diameter should be sufficient to allow free movement of the probe; a difference of 2 to 4 mm between the outside diameter of the probe and the inner diameter of the tubing normally ensures this. A tubing wall thickness of 1.5 to 5 mm is appropriate. Most equipment is designed for use with 44.5 mm (1.75 inch) or 50.8 mm (2 inch) outer diameter tubing, and the probe carrier fits

Table 3 Relative Advantages of Different Types of Access Tubing

Material	Effect on neutron count	Strength	Resistance to corrosion	Cost
Aluminum	Transparent	Weak	Poor	Expensive
Aluminum alloy	Transparent	Moderate	Poor	Moderate
Stainless steel	Lowers count by 10–15%	Strong	Good	Expensive
Plastic	Increases (PVC decreases)	Moderate	Good	Cheap

on to the top of the access tube while the probe is lowered within it. If tubing of appropriate diameter is not available, an adaptor can be made to allow the probe carrier to be fitted on to the top of larger tubing. Suitable tubing can normally be obtained from stock from suppliers, as can rubber stoppers to close the exposed end. A stopper may be used to close the bottom end, but a turned or cast end-piece of the same material as the tubing, sealed with waterproof adhesive into the end of the access tube, is preferable.

Whichever tubing is selected, it is important that all calibration work and all standard counts are made using tubing of the same material and diameter as used in the field.

2. *Access Tube Installation*

During installation, disturbance to the soil, the soil surface, and vegetation in the vicinity must be minimized to ensure that subsequent measurements are representative of the surrounding area. The access tube must fit tightly into the soil. Biased measurements will be obtained if there are voids adjacent to the tube or if preferential movement of water occurs beside it (Amoozegar et al., 1989). If there is doubt as to how well a tube has been installed, it is best to re-site it nearby. The extra effort is preferable to collecting suspect data over a long period.

Plenty of time should be allowed for installation work. Two people working in favorable conditions can be expected to install only three or four 2 m access tubes per day, using the method given below. Longer tubes or difficult soils may only permit complete installation of one per day. Installation in heavy clay soils is often difficult both when the soil is wet (due to soil sticking to equipment) and when it is dry (because of hardness). Dry sand makes augering difficult and the sides of the reamed hole may collapse.

The installation method described here has been used successfully to install tubes to 3 m and greater depth in many different soils developed on clays, chalk, silts and sandstones, without resort to power-driven implements. A hole is made for the access tube by using a steel guide tube of the same outer diameter as the access tube. The lower end of the guide tube is sharpened by an internal bevel to give a cutting edge of the same diameter as the external diameter of the access tube. A screw auger that moves easily within the guide tube is used through it to drill out soil to about 0.1 m below the cutting end; the guide tube is then hammered in 0.1 m using a sliding hammer. If this procedure is followed, the guide tube will not be hammered down until a hole of slightly smaller diameter has been augered below it, thus disturbance to the soil surrounding the tube is minimized. The process is repeated until the required depth is reached. The guide tube is then withdrawn and the access tube slid into the reamed hole; gentle tamping may be necessary to drive it fully home. The access tube should then be cut off so that the desired length protrudes from the ground. It should be fitted with a stopper so that the tube remains dry and clean.

If access tubes are to be installed to more than 1 m depth, a series of guide tubes 1.15, 2.15, 3.15 and even 4.15 and 5.15 m in length is used successively with an auger having an extendable shaft. Alternatively, an extendable guide tube with 1 m extensions which can be screwed on to the first tube of 1.15 m length can be employed. A removable collar is necessary to protect the top of the screw thread while hammering. A sharpener, and a file to remove any buckling of the cutting edge caused by stones, should be part of the installation kit.

The top of the guide tube should not be driven in too far, in case it is necessary to fit a clasp if mechanical means are required to extract it. Automobile jacks can be used, and powerful rod-pullers are available from drilling equipment suppliers. It is essential that the pull be exerted along the axis of the tube both to reduce effort and to avoid deforming the hole during extraction. Use of a base plate with a central hole for the guide tube is recommended unless it is likely to damage the crop. This presents a firm base when using tube extractors and minimizes surface soil compaction and enlargement of the neck of the hole.

This installation method can be adapted for use in situations where the soil is unstable, or saturated due to a shallow water table, by using the access tube itself to ream the hole, so avoiding the need to withdraw the tube. The greater strength of a stainless steel access tube may be required, however. Sealing the bottom end of a tube installed in this fashion, particularly below a water table, is not easy; bungs and adhesive, bentonite and other materials have been used (Prebble et al., 1981). This installation method may also be preferred in heavy clay soils if considerable effort is required to extract the guide tube, leading to over-enlargement of the hole near the surface. The timing of installation in swelling clays may affect subsequent cracking adjacent to access tubes and should be considered when planning installation in such soils (Jarvis and Leeds-Harrison, 1987).

A power-driven hammer may be used to drive tubes into very dense or stony soils. The power device should only be used to drive the tube down about 0.1 m after augering. Several attempts at installation may be necessary in stony soils.

Unfortunately there is a tendency for greater success in less stony places, which may result in measurements that are not representative of the soil as a whole. Prebble et al. (1981) addressed this problem and described a variety of installation methods that may be required in other situations. Once installation is complete, precautions should be taken to prevent damage to the surrounding soil and vegetation in the course of making measurements.

3. Measurements Near the Soil Surface

The most satisfactory method of overcoming the influence of the soil–air interface on near-surface measurements is to conduct calibrations specifically for the surface soil layers. Many approaches to deal with the effect (some very elaborate!) have been devised, including use of neutron reflectors placed on the soil surface, use of soil-filled trays placed on the surface to extend the soil medium artificially, correction methods, and use of the probe horizontally on the soil surface. Chanasyk and Naeth (1996) provide a comprehensive review of these. However, a calibration or calibrations for the upper 0.2 to 0.3 m are simple to obtain, as core sampling to such shallow depths is straightforward, and provide the most accurate means of determining water content from neutron counts at shallow depth. Accurate depth placement of the probe for measurements close to the soil surface is particularly important, as Fig. 4 illustrates.

G. Calibration

There are three techniques for calibrating soil water content against count rate ratio: theoretical calibrations, drum calibrations, and field calibrations. A linear relationship between count rate ratio and soil volumetric water content is obtained with most neutron depth probes:

$$\theta = a\frac{R}{R_s} + b \tag{10}$$

where R is the count rate (s^{-1}) in soil and R_s is the standard count rate (s^{-1}). Calibrations are specific to the design of neutron probe used. As described in Sec. C, the use of standard counts to normalize count rate measurements results in a soil-specific calibration that can be used with any probe of the same design. However, it is important to use the same type of access tubing for routine field measurements and calibration purposes because of its influence on count rates. The calibration depends on the soil's neutron scattering and capture cross-sections and bulk density. It is important to be aware of particularly high concentrations of neutron absorbers such as iron and of the presence of any very strong absorbers such as gadolinium and cadmium. For instance, the effect on calibrations of gadolinium concentrations of only 1 to 36 mg kg^{-1} in Tasmanian soils is considerable (Nicolls et al., 1977).

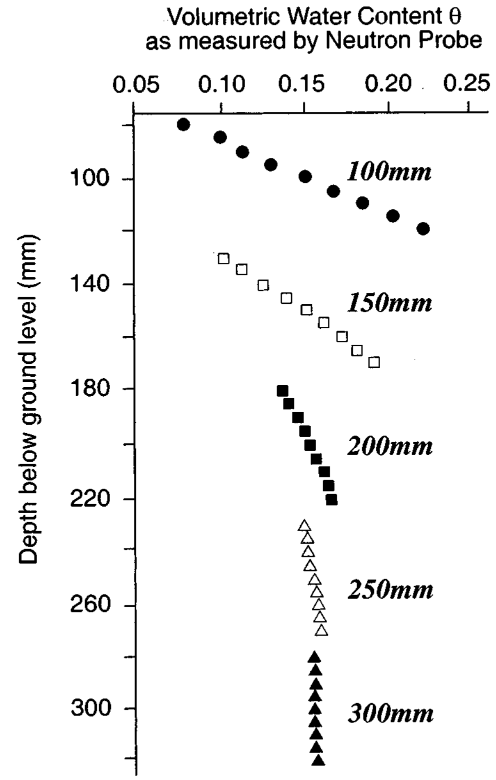


Fig. 4 Effect of depth location on water content measurement at shallow depth. Calibrations for measurements with the probe located at 100, 150, 200, 250, and 300 mm depths were prepared, and measurements precisely at these depths show that the water content of the upper 300 mm of the profile is uniform at 0.15. However, even a small error in the depth location of the probe can cause a significant error in the measured water content. (After Karsten and Van der Vyver, 1979.)

The neutron count rate is influenced by all the hydrogen present in the soil, including that in free and bound water, as well as in other compounds. The hydrogen in adsorbed and structural water and the nonwater hydrogen has the same influence on neutron thermalization as that in free water. Its presence can be expressed in terms of an equivalent water content. Since it does not change with time and is not removed during oven-drying, its effect is incorporated into the intercept term, *b*, of the calibration equation. Greacen (1981) advocated calibration in terms of total water content (i.e., the sum of the free and equivalent (θ_e) water content); both a laboratory method for determining θ_e and a means of estimating it from clay content are described. For some soils, this permits use of a single calibration for different soil layers, providing θ_e has been determined for each one individually.

An increase in bulk density causes an increase in the number of nuclei close to the source, resulting in more neutron scattering close to it and so an increase in the number of slow neutrons detected. This increase in count rate with increase in density is reinforced if the equivalent water content of the dry soil is large, because of the greater concentration of hydrogen close to the source. However, the concentration of neutron absorbers in the vicinity of the source is also increased, and this counteracts the tendency towards a higher count rate. There is disagreement as to the net effect of bulk density on neutron count rates (Greacen and Schrale, 1976; Rahi and Shih, 1981). If soil-specific field calibrations are used, they will incorporate the effect of bulk density. Otherwise it is important to measure field soil bulk density, ρ, and adjust calibrations to this using

$$R = R_i \left(\frac{\rho}{\rho_i} \right)^{0.5} \tag{11}$$

where R_i is the count rate in soil of density ρ_i, and R is the adjusted count rate (Greacen and Schrale, 1976).

1. Theoretical Calibrations

Theoretical models based on diffusion theory have been developed to simulate neutron flux in soils for which the neutron interaction cross-sections are known. The interaction cross-section of a soil may be determined by direct measurement or by detailed chemical analysis and use of published cross-sections (Mughapghab et al., 1981). Assumptions about soil density are made in the theoretical calibration, which is then adjusted to allow for field soil bulk density.

Determination of soil neutron interaction cross-section by chemical analysis requires detailed analysis of the concentration of at least 20 elements in representative samples of the soil (Olgaard, 1965). Omission of the analysis of a crucial neutron absorber such as gadolinium or boron would have a substantial effect on the resulting calibration. Because of a tendency for overestimation of the neutron absorption effect, the procedure is most satisfactory for light-textured soils with low neutron capture cross-sections, <0.004 barn (Greacen and Schrale, 1976). Wilson (1988b) found that the likely minimum error to be achieved in practice with this calibration method ranged from about $\pm 1.6\%$ to $\pm 3.5\%$ volume fraction, with larger errors occurring in drier soils.

Direct measurement of neutron interaction cross-sections requires access to appropriate specialized equipment, a large neutron source, or even a reactor (Couchat et al., 1975, McCulloch and Wall, 1976). A comparison of calibrations obtained by Couchat et al. (1975), who used a large source in a graphite block, with those determined by the conventional field method for sand, chalk, silt, and chalky clay soils, found good agreement (Vachaud et al., 1977). The method was particularly recommended for use in heavy soils, where obtaining samples over a full range of water content is difficult, and for soils with marked layering, as it enables isolation of the layers from one another for calibration purposes.

2. Drum Calibration

This requires the uniform packing of soil of known water content into a large drum of about 1.5 m diameter and 1.2 m depth. An access tube is installed so that neutron counts can be made within the soil-filled drum. The process is repeated with the soil at a different water content. In principle, as the relationship between soil water content and neutron count is known to be very nearly linear, only two points are required, but it is preferable to obtain several over a range of water contents and bulk densities. The method is very laborious, requiring collection of large quantities of soil from the field and care in wetting up and packing to ensure uniformity in the drum. Use of the bulk density correction (Eq. 11) removes the need to pack the soil to the field bulk density. With care, good calibrations with high correlation coefficients can be obtained for a wide variety of soils (Greacen, 1981).

3. Field Calibration

In this method, a calibration is derived by simultaneous measurement of the neutron count rate and sampling of soil for determination of the volumetric water content of each layer on several occasions, so as to cover the range of hydrological conditions characteristic of the site. The theoretical and drum calibration methods assume a homogeneous soil, whereas field calibrations allow for the presence of site-specific features such as textural boundaries or the fluctuations of a shallow water table. Field calibrations usually result in greater scatter in the calibration points due to soil heterogeneity and sampling errors, but if conducted with care may represent the absolute water content of soil at a site better than the alternative methods.

There are two approaches. Simultaneous neutron counts and samples for volumetric water content determination may be achieved by installation of a temporary access tube in the area used for monitoring the soil of interest. Neutron counts are recorded in the temporary tube at the required depths and then five or six undisturbed samples are taken from immediately around it at each depth by coring and, if necessary, excavating around the tube. The temporary access tube is then removed to be used later. The process is repeated for different depths and times of the year to obtain a calibration over the range of water contents found at the site for each soil layer. Alternatively, neutron counts may be recorded in the access tube used for monitoring and samples collected by coring close to (within 2 m of) the tube. This is suitable in soils where samples can be readily collected by coring; otherwise damage to the vegetation and soil around the access tube may render subsequent measurements in it unrepresentative of the wider area. Again, the process is repeated on several occasions. Irrigation of the area, or encouraging drying with a shelter to keep off rainfall, is acceptable to extend the range of hydrological conditions covered by calibration. It is important to avoid times when a wetting front is moving rapidly through the soil (i.e., immediately after rainfall or irrigation).

The first approach is particularly useful where many access tubes are used to monitor a fairly well defined soil (e.g., in the course of field trials or experiments). The second is appropriate where access tubes are located in differing soils, as in a catchment experiment, and a calibration for the soil at each tube is required. However, if obtaining volumetric samples by coring is difficult, use of a temporary access tube at greater distance from the semipermanent tube will be preferable.

The volumetric water content of the samples is determined by oven-drying; then the paired neutron count and water content data are used to determine a calibration for each soil layer by linear regression. The count rate ratio is considered as the independent variable (x) and the water content as the dependent variable (y). The data from different depths should be analyzed separately, even if the soil appears homogeneous, until the calibrations can be reviewed. Pooling data to reduce the number of calibrations may then be appropriate.

Stones can present a problem in deriving calibrations but cannot be ignored. Stocker (1984) described a method using an access tube and sand to measure the volume of soil samples collected from around the temporary access tube in stony soils.

An alternative procedure for in situ calibration, which is applicable in dry, homogeneous, light-textured soils with a high infiltration rate, is described by Carneiro and De Jong (1985). Known amounts of water are allowed to infiltrate the soil between recording neutron counts. The method assumes that there is no loss of water by evaporation or drainage from the profile during the calibration process.

H. Surface Neutron Meters

Surface neutron meters are used widely in civil engineering and soil mechanics for monitoring the water content of earthworks but have other applications where measurements at a smooth, bare soil surface are required. Ahuja and Williams (1985) used a surface gamma-neutron meter to characterize surface soil properties. Measurements represent a layer about 0.35 m deep in dry soil but only 0.15 m deep in wet soil. Farah et al. (1984) showed that only two calibrations were necessary to represent satisfactorily all or part of the layers 0–0.10 and 0–0.30 m deep. However, if a shallow wetting front is present, measurements are difficult to interpret.

I. Radiological Safety

The acquisition, use, transport, storage and eventual disposal of neutron probes is subject to regulation because of the potential hazard to human health and the environment posed by the neutron source. Most governments have legally enforceable radiological safety regulations that must be followed when using neutron probes. The recommendations of the International Atomic Energy Agency (IAEA, 1972, 1990) and the International Commission on Radiological Protection (ICRP, 1990) should be consulted in the absence of specific regulations.

With sensible usage, the radiation hazard to a trained neutron probe operator is only a little greater than that permitted for members of the public. Precautions such as maximizing one's distance from the source when carrying a probe, or transporting one in a vehicle, are straightforward. A probe should never be left unattended except when locked in its designated storage place.

Regular tests to check for leakage from the source are advisable and mandatory in some countries (e.g., in the U.K., tests must be conducted once every two years). Americium–beryllium sources have a half-life of 458 years, much longer than the useful life of the probe, and longer than the time over which the integrity of the source capsule can be expected to be maintained (up to 30 years). When a source is no longer required it must be disposed of at a designated repository for radiological waste and this cost can add significantly to the lifetime cost of the probe.

V. DIELECTRIC METHODS

Dielectric methods for soil water content measurement exploit the strong dependence of soil dielectric properties on water content. These dielectric properties affect the velocity of an electromagnetic wave (used in TDR), the characteristic impedance of a transmission line (used in the Theta probe), and the capacitance of two electrodes embedded in the soil (used in capacitance techniques).

Smith-Rose (1935) explored the electrical properties of soil as a function of water content, and Thomas (1966) used capacitance instruments, but developments were limited by the lack of an accurate method of measuring high-frequency capacitance. TDR was first applied to dielectric measurement by Fellner-Feldegg (1969) and was soon used to investigate the dielectric properties of soils (Hoekstra and Delaney, 1974; Topp et al., 1980). TDR equipment is now available commercially (Table 4). Interest in capacitance techniques revived in the mid-1980s when developments in electronics meant that capacitance in the 100 MHz frequency range could be measured much more readily, and the method is used in a wide variety of applications.

Early work by Topp et al. (1980) suggested that, for most purposes, a

Table 4 Equipment Manufacturers/Suppliers

Equipment name	Address	Principle
TDR Soil Moisture Measurement System (based around the Tektronix 1502C)	Campbell Scientific Ltd., 815W 1800N Logan, UT 84321-1784, USA	TDR
CS615 Water Content Reflectometer	Campbell Scientific Ltd., 815W 1800N Logan, UT 84321-1784, USA	TDR
Easy Test	Easy Test Ltd., Solarza 8b, 20-815 Lublin 56, PO Box 24, Poland	TDR
Moisture Point	Environmental Sensors, Inc. 100-4243 Glanford Ave, Victoria, BC, Canada V8Z 4B9	TDR (with shorting diodes)
HP 54120	Hewlett-Packard Company, 5161 Lankershim Blvd, No. Hollywood, CA 91601, USA	TDR
Trime	IMKO GmbH, Im Stock 2, D-76275 Ettlingen Germany	TDR
Tektronix 1502B/C	Tektronix, PO Box 1197, 625 S.E. Salmon Street, Redmond, OR 97756-0227, USA	TDR
TRASE	Soil Moisture Equipment Corp., PO Box 30025, Santa Barbara, CA 93105, USA	TDR
Theta Probe	Delta-T Devices Ltd., Burwell, Cambridge, UK	Impedance
EnviroSCAN	Sentek Pty Ltd., 69 King William Street, Kent Town, S. Australia 5067, Australia	Capacitance
IH Capacitance probe	Soil Moisture Equipment Corp., PO Box 30025, Santa Barbara, CA 93105, USA	Capacitance
Humicap 9000	SDEC France, 19 rue E. Vaillant, 37000 Tours, France	Capacitance
Troxler Sentry 200 AP	Troxler Electronic Laboratories, Inc., 3008 Cornwallis Road, PO Box 12057, Research Triangle Park, NC 27709, USA	Capacitance

This list is not exhaustive. Sources are given for the convenience of the reader only, and imply no endorsement on the part of the authors.

universal relationship between dielectric measurements and θ would be applicable to the majority of soils, and so calibration would often be unnecessary. However, further studies have shown that the dependence of soil dielectric properties on water content is more complex and that calibration for individual soils is necessary. Much effort has gone into defining precisely the relationship between water content and soil dielectric properties, using physically based models. Progress is being made, but assessment of results is complicated by the fact that various groups are working with different soils and equipment. At the same time, others are attempting to validate the performance of new designs of equipment. The focus in this chapter is on the practical use of dielectric methods, but a brief explanation of dielectric theory and soil dielectric properties is appropriate. The principles and practice of TDR are described in detail. One impedance technique is described. The theory of capacitance measurements is explained, but as different measurement techniques can be used, only one instrument system is discussed in any detail. The principles governing installation and calibration are the same for all of these instruments and are considered together.

A. Dielectrics

A dielectric is an electrical insulator. When a dielectric is placed in an electrical field, the positive and negative charges within it are pulled in opposite directions, producing a polarization of the dielectric and storing energy in it. Every dielectric is capable of storing electrical energy in this way; this is described by the material's permittivity, ϵ, and is measured in picofarads per meter (pF m^{-1}). As the permittivity of any dielectric is always greater than that of a vacuum, ϵ_0 (8.854 pF m^{-1}), it is convenient to work with the relative permittivity, ϵ_r, which is the ratio of the permittivity of the material to that of a vacuum, ϵ/ϵ_0. (ϵ_0 is also known as the electric constant.) ϵ_r is often called the dielectric constant, but the term relative permittivity is preferred, since ϵ_r varies between materials and depends on temperature and pressure and the frequency of the applied field.

Some substances have individual molecules that possess a permanent electrical dipole. They can therefore store greater amounts of energy than other materials and consequently have high relative permittivities. Water is a prime example of such a polar dielectric. When a molecule with a permanent dipole is placed in an electric field, it will attempt to orientate itself with the field. If the electric field is alternating, the molecule will attempt to rotate with the field, but its rotation will be constrained by the presence of adjacent molecules and by collisions with other molecules.

Whether a substance is polar or nonpolar, when the applied electric field is removed, it takes a short time for the molecules to "relax" to random positions and orientations and the polarization to decay. The time required for this relaxation is characteristic of the material. The same relaxation time governs the response to any change in field strength, so that as the field frequency increases, a point is reached where the polarization cannot change direction as fast as the field. Consequently the permittivity of the substance decreases; the frequency threshold at which this occurs is characteristic for any given substance and is known as the relaxation frequency.

In practice, most substances are imperfect dielectrics and exhibit electrical conduction over a wide range of frequencies. This is often because the substance possesses some ionic conductivity. Soil is such a medium, the soil solution providing an electrically conducting pathway. Soils which have high salinity, contain a lot of clay, or receive regular fertilizer applications exhibit the greatest conductivity. The effect of this conduction may be described in the form of a complex relative permittivity, ϵ_r^*, which has a "real" part, ϵ', describing energy storage and an "imaginary" part, ϵ'', describing energy losses:

$$\epsilon_r^* = \epsilon' - j\epsilon'' \tag{12}$$

where

$$\epsilon'' = \frac{\sigma}{\epsilon_0 \omega} + \text{any other loss mechanisms} \tag{13}$$

σ is the low-frequency electrical conductivity, ϵ_0 is the permittivity of free space, ω is angular frequency ($= 2\pi F$, where F is the ordinary frequency), and j is $\sqrt{-1}$. The effect of this conductivity on relative permittivity measurements depends on which measurement method is used. The aim of most soil water content measuring devices is to measure the real permittivity, ϵ', which is related to volumetric water content, without interference caused by losses due to soil electrical conductivity. Additional measurement of the imaginary part of the permittivity can be used to estimate soil solution conductivity and hence to infer the solute content.

B. Dielectric Properties of Water and Soil

At frequencies below 10 GHz the relative permittivity of pure water at 25°C is 78.38 and increases by ca. $0.36°\text{C}^{-1}$ (0–50°C) as temperature falls. When water freezes, the permittivity falls to about 4 (Fig. 5). Within soil, water molecules in the proximity of colloidal surfaces are influenced by the electrical charge on the surface and lose some of their rotational freedom. The permittivity of bound water in soils is therefore less than that of free water. Research has indicated that values of 4 to 40 for bound water are appropriate at frequencies greater than about 100 kHz (Sposito, 1984). The value varies since the dielectric behavior and relaxation frequency of bound water is influenced by the chemistry of the soil solution and the character of the surface. The other constituents of soil have much lower permittivities than free water; the value for air is 1 and that of most soil solids is usually less than 6.

To make progress in deriving calibration equations to relate permittivity to soil water content, a conceptual framework is required. Much theoretical work has been directed at producing models of the permittivity of mixtures for ordered and

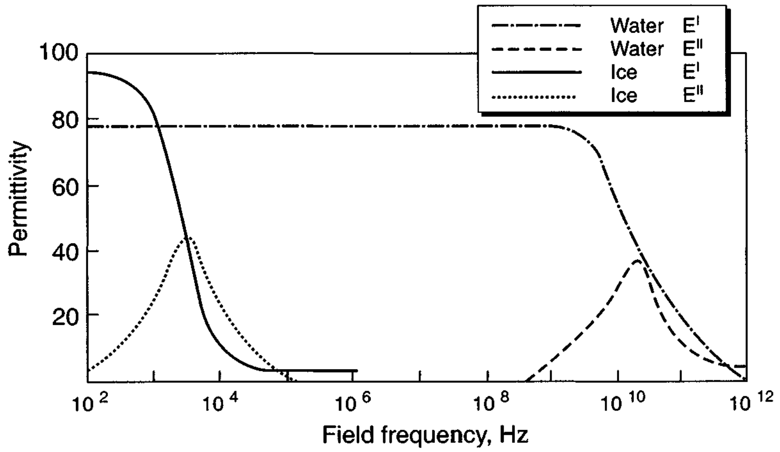


Fig. 5 Change in the real and imaginary permittivity of water and ice, with field frequency.

disordered systems. No real soil conforms to all the assumptions used in deriving these, and indeed, the arrangement of the components in one soil is often quite different from that in another. It is probable that the relationship between permittivity and the concentration of different soil components is similar to that predicted by the models, but the exact values of constants in any one model are unlikely to be realized.

The manner in which soil components contribute to bulk soil permittivity can be illustrated using a straightforward mixing model. Bulk soil is considered as a mixture of four phases: air, solids, free water, and bound water, thus

$$\epsilon^{\alpha} = \epsilon_{\mathrm{a}}^{\alpha} f_{\mathrm{a}} + \epsilon_{\mathrm{s}}^{\alpha} f_{\mathrm{s}} + \epsilon_{\mathrm{w}}^{\alpha} f_{\mathrm{w}} + \epsilon_{\mathrm{bw}}^{\alpha} f_{\mathrm{bw}} \tag{14}$$

where $\epsilon_{\mathrm{a}}^{\alpha}$, $\epsilon_{\mathrm{s}}^{\alpha}$, $\epsilon_{\mathrm{w}}^{\alpha}$, and $\epsilon_{\mathrm{bw}}^{\alpha}$ are the permittivities of air, soil solids, free water, and bound water, respectively, and f_{a}, f_{s}, f_{w}, and f_{bw} are their volume fractions. The total water content, θ, is the sum of f_{w} and f_{bw}. The bound water is often ignored, however. Experimental and theoretical work have shown that a value of about 0.5 for α (Birchak, 1974; Roth et al., 1990; Whalley, 1993; Jacobsen and Schonning, 1994) is appropriate for many soils. Since

$$f_{\mathrm{a}} + f_{\mathrm{s}} + \theta = 1 \tag{15}$$

and

$$f_{\mathrm{s}} = \frac{\rho}{\rho_{\mathrm{p}}} \tag{16}$$

where ρ is soil bulk density and ρ_p particle density, Eq. 14 can be expressed in terms of dry bulk density and particle density:

$$\epsilon^{\alpha} = \epsilon_a^{\alpha}\left(1 - \frac{\rho}{\rho_p} - \theta\right) + \epsilon_s^{\alpha}\frac{\rho}{\rho_p} + \epsilon_w^{\alpha}\theta - (\epsilon_w^{\alpha} - \epsilon_{bw}^{\alpha})f_{bw} \tag{17}$$

If the volume fraction of the bound water, f_{bw}, is assumed to be so small that it can be ignored, then, assuming that α equals 0.5, the permittivity of air is 1, and that of water is 81, Eq. 17 becomes

$$\begin{aligned}\sqrt{\epsilon} &= 1 + \frac{(\sqrt{\epsilon_s} - 1)\rho}{\rho_p} + (\sqrt{81} - 1)\theta \\ &= 1 + \frac{(\sqrt{\epsilon_s} - 1)\rho}{\rho_p} + 8\theta \end{aligned} \tag{18}$$

It is clear that θ makes a very big contribution to the bulk soil permittivity due to the large permittivity of free water. However, it is also notable that dry bulk density has a role, and that its influence will be greater at greater water contents (solving Eq. 18 for ϵ rather than $\sqrt{\epsilon}$ results in $\theta\rho$ terms). More complex dielectric mixing models are available in the literature (e.g., de Loor, 1968) and have been applied to soils (e.g., Dobson et al., 1985).

C. Time Domain Reflectometry

The principle behind TDR is that a fast rise-time electromagnetic pulse is fed into the soil between two or more metal rods, which act as a waveguide. The soil acts as a dielectric between the conductors of this transmission line. The velocity of propagation of the pulse depends only on the permittivity of the soil between the rods. The applied pulse will be reflected either from the end of the transmission line or from impedance mismatches along it (e.g., connectors). The time interval between the incident and reflected pulses is measured. Cable testers use this principle to locate faults and breaks in cables. The cable tester measures the travel time of the pulse to and from any discontinuity and so the distance to it can be determined easily.

The propagation velocity, v, of a transverse electromagnetic (TEM) wave is related to the permittivity of the material by

$$v = \frac{c}{\sqrt{\epsilon_r}} \tag{19}$$

where c is the velocity of light (3×10^8 m s^{-1}). The time, t, taken for a wave to propagate down the transmission line and return is

$$t = \frac{2L}{v} = \frac{2L\sqrt{\epsilon_r}}{c} \tag{20}$$

where L is the length of the line. Topp et al. (1980) used the term apparent relative permittivity of the soil (K_a) in place of ϵ_r to indicate that other factors, principally the imaginary part of the permittivity, influence the measurement. The effect is negligible except when the imaginary part of the permittivity is very large, as in strongly conducting soils.

Because the square root of permittivity is almost linearly related to water content (Eq. 18), the time taken for the pulse to propagate along the line (Eq. 20) is proportional to the square root of permittivity. Thus, the propagation time varies linearly with total water content along the line, even when there are water content variations along it. This makes TDR a good method for estimating total water storage over an extended depth range.

1. TDR Systems

Figure 6 is a block diagram of a TDR instrument. A timer provides synchronizing information to a pulse generator and a receiver. The pulse generator supplies a voltage step with a very fast rise time, effectively feeding a train of high-frequency (predominantly in the range 100 MHz to 1 GHz) electromagnetic waves with a

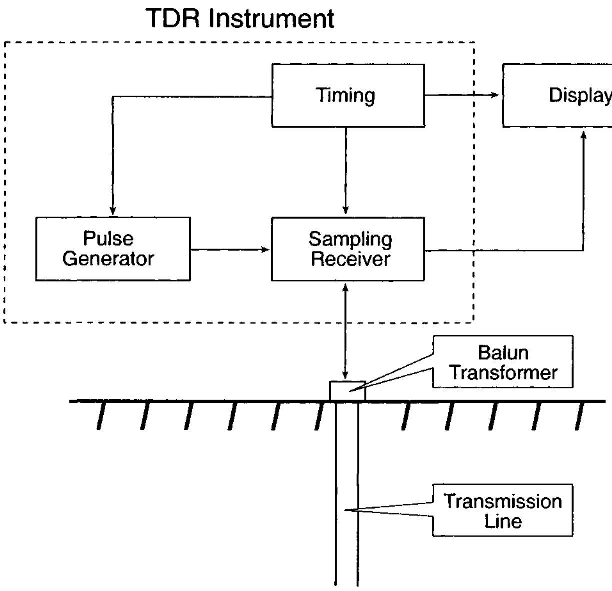


Fig. 6 Block diagram of a TDR instrument.

wide frequency distribution into the sample. The detector circuit measures the sum of the input voltage and the reflected pulse. Because the times involved are very short, a few nanoseconds, the time dependence of the output voltage is determined by sampling the voltage at a series of times after the initial pulse. Pulses are sent repeatedly, every millisecond or so, and one voltage sample is measured after each pulse cycle. Thus a voltage–time curve (the waveform) can be reconstructed from these measurements and used to determine *t*. It is important to realize that the resultant waveform is the sum of a step input and the reflected voltage.

It is possible to assemble a TDR system for soil water content measurement quite easily, if a cable tester is available. Topp et al. (1980) used a Tektronix 1502B cable tester, which can be linked to a PC using a RS-232 interface. This instrument, or the 1502C model, is commonly used in TDR research because of its adaptability. A number of companies provide systems incorporating Tektronix cable testers, with their own waveguides and software. However, such setups are less convenient than the off-the-shelf systems now available (Table 4). For example, the TRASE system (Fig. 7) incorporates a TDR plus a data logger and

Fig. 7 A TRASE TDR system.

interpretation software. Waveguides are available for TRASE that can be used for measurements at the surface or buried for continuous monitoring. Stored data is easily downloaded into a PC via a RS-232 connection. For routine measurement of soil water content, it is a well integrated user-friendly system.

Commercial TDR systems are supplied with in-built software that analyzes each waveform. Such software works well with waveforms produced in homogeneous media. However, dielectric discontinuities along the waveguide may create reflections other than from the end, and if the soil is particularly conductive, the waveform may be attenuated. Automatic analysis of the waveform may then be unreliable. More specialized software can recognize difficult waveforms and tag them so that the user can examine the waveform to determine the end point reflection manually (Heimovaara and de Water, 1993).

A major advantage of TDR is that readings can be logged automatically, and several waveguides can be attached to a multiplexer, which switches between channels to make a measurement on each (Baker and Allmaras, 1990; Heimovaara and Bouten, 1990; Herkelrath et al., 1991). Up to 70 locations in the soil may be monitored, but as channels cannot be read simultaneously, the reading cycle takes longer the more waveguides are monitored; cycles may take 10 to 15 minutes for a lot of sensors.

2. *Waveguides*

The waveguide is the TDR sensor that is inserted into the soil. Waveguides are also referred to as "guides," "probes," "rods," or "wires." Several designs are illustrated in Fig. 8. There has been much discussion about the design of waveguides, in particular their length, width, and number of electrodes (Heimovaara, 1993; Whalley, 1993; Selker et al., 1993; Baumgartner et al., 1994; Noborio et al., 1996). The minimum requirement is two electrodes for each waveguide, one attached to the central conductor of the coaxial cable and one or more attached to the sheath.

TDR provides a measurement of the integrated water content along the full length of the waveguide. Waveguides of up to about 1 m length can be used in favorable conditions. Use of short waveguides installed horizontally from the walls of a pit may be preferable to vertical installation of long waveguides, if measurements at discrete depths are required. Alternatively, vertically installed waveguides of different lengths may be used to derive water content in different depth ranges by difference.

The Easy Test TDR system differs from others in having very small waveguides (rods <6 mm length, <2 mm diameter and separated by <2 mm) (Malicki et al., 1992). For field use, these are attached to a cylindrical body and so can be installed vertically at the base of a preaugered hole, in a manner similar to that used for tensiometer installation. Their short length means that a needle voltage

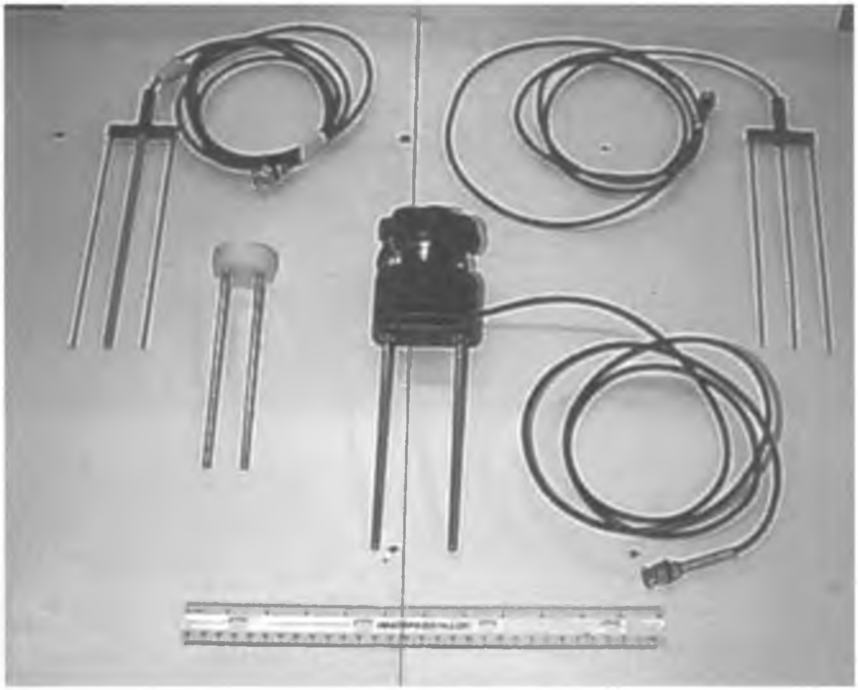


Fig. 8 Different designs of TDR waveguides.

pulse with a very short duration (200 ps) is required, rather than a single step voltage.

Attachment of a coaxial cable to a waveguide results in some reflection of the applied pulse. This is used to identify the position corresponding to the start of the waveguide on the TDR trace. However, too large an impedance mismatch causes only a small proportion of the applied voltage pulse to enter the waveguide, with consequent small signal levels and multiple reflections, making interpretation of the trace difficult (Spaans and Baker, 1993). Two-wire probes normally use a "balun" (an impedance matching transformer) to reduce this problem. Three- and four-wire guides do not normally require the use of a balun. If resistance is also to be measured, a balun cannot be used.

a. Waveguide Sampling Volume

De Clerk (1985) showed that for a waveguide with a rod spacing of 25 mm, 94% of the energy was contained within a cylinder of 50 mm diameter; thus a 20 cm long waveguide has a sampling volume of some 98 cm^3. Whalley (1993) demonstrated that TDR is most sensitive to the soil close to the rod connected to the central conductor of the transmission line. Thus the sampling volume is more concentrated around the central rod of 3- and 4-wire waveguides than around the conductors of a two wire sensor. In addition, the smaller the diameter of the conductors, the smaller the volume of soil to which the measurement is most sensitive. For detailed discussion of waveguide sampling volume see Knight (1992, 1995).

b. Constraints on Waveguide and Cable Length

The length of waveguide used will be dictated by two main factors: the volume of soil to be measured and its electrical characteristics. 15 cm is recommended as the minimum waveguide length for routine field work with most systems. The error in measurement increases as the sensors become shorter, because the accuracy with which the returning pulse can be timed is fixed, and so the proportional accuracy increases as the length of the waveguide increases. However, the shorter the waveguide, and the greater the distance between the electrodes, the smaller the influence of electrical conductivity. In soils with a high electrical conductivity, the length of waveguide that can be used effectively is limited to 50 cm or less. Thus before deciding on a field installation, it is advisable to assess the soil's attenuation characteristics. This can be as simple as taking the TDR to the field site, wetting the soil, and installing a waveguide to see if an interpretable waveform is generated. The effect of attenuation due to conductivity can be reduced using rods coated with heat shrink Teflon to ensure the return of a strong reflection (Kelly et al., 1995). An epoxy-coated waveguide is offered by Soil Moisture Equipment Corp. for use with the TRASE system and has a similar effect.

Cable length also influences the magnitude of the returning step pulse; the longer the cable, the greater is the attenuation of the signal (Heimovaara, 1993). Herkelrath et al. (1991) recommended that coaxial cable runs should be no longer than 30 m. Use of low-loss cable will increase the working distance from the TDR pulser.

3. Waveforms

The output from TDR equipment is a waveform, a graph of voltage versus time. Figure 9 illustrates how the shape of the waveform is made up of voltages from successive reflections at the junctions between the coaxial connector and the waveguide and at the end of the waveguide. The time measured to determine permittivity using Eq. 20 is that between points A and B in Fig. 9a. Figure 9b illustrates the waveforms produced when measuring the permittivity of air and tap water. The travel time for the pulse along a 20 cm waveguide in air is 0.67 ns and 5.97 ns in water; the time increases proportionally with longer waveguides.

Locating the end point, B, of the waveform is fundamental to the measurement of permittivity. In Fig. 10a the position of the reflection from the end of the waveguide is readily distinguished. However, it is not sharp but distributed over

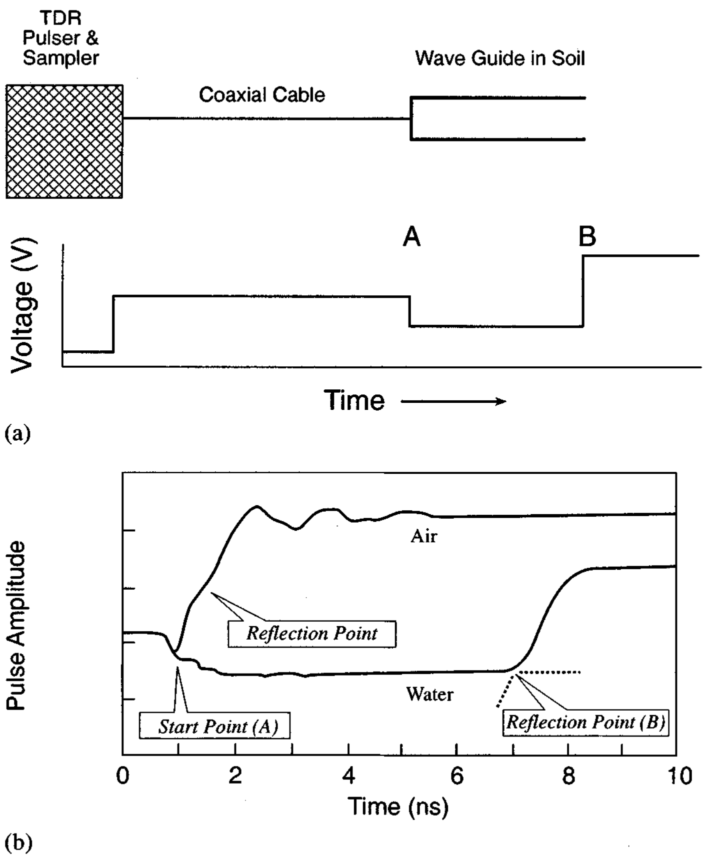
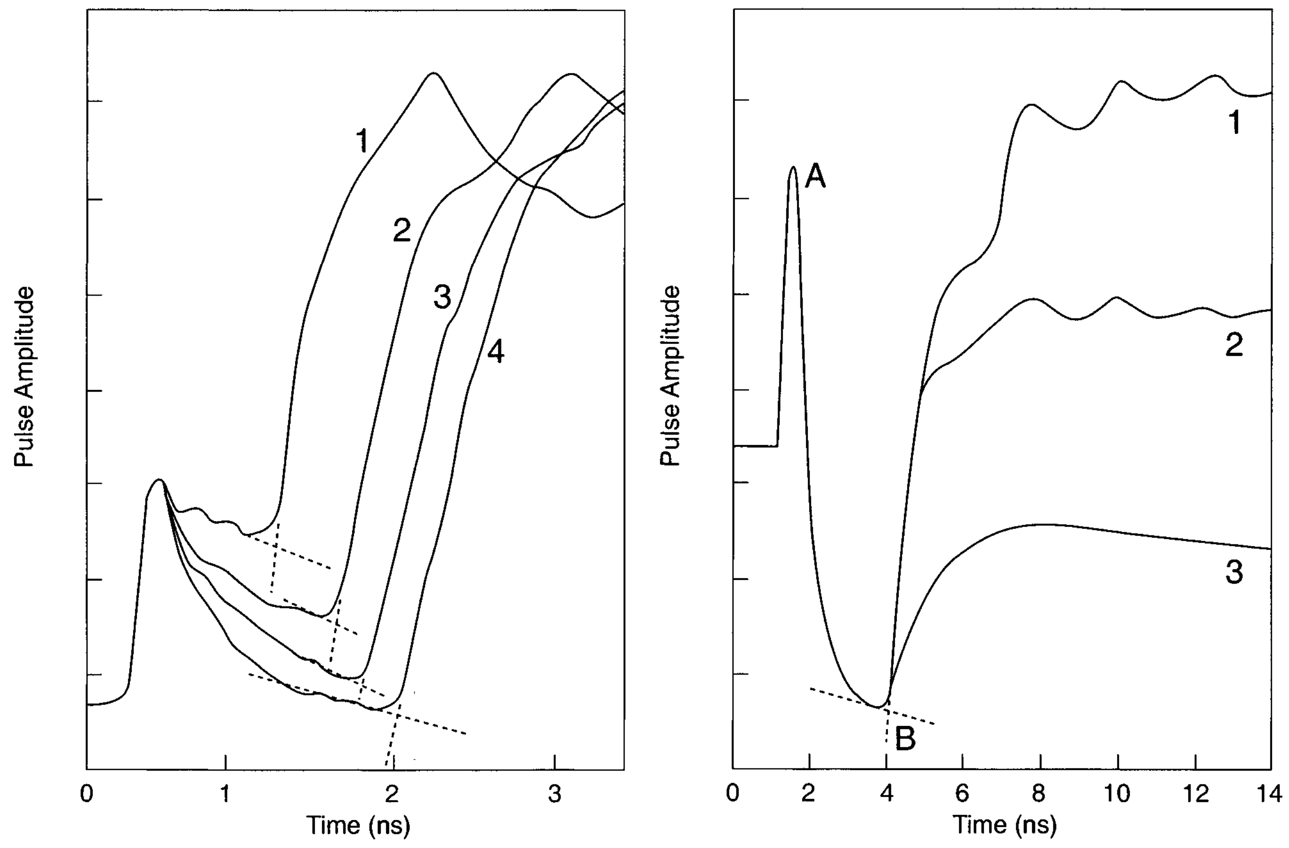


Fig. 9 (a) Relationship between the waveform shown on the TDR screen and the TDR/waveguide setup. Usually only the right-hand part of the waveform is displayed, i.e., from just before A to after B. (b) TDR waveforms produced with a 20 cm waveguide in air and in water.

a range of times. This is due to some dispersion of the pulse (i.e., some frequencies of the wave propagating at slightly different speeds), greater attenuation of some frequency components than others, and penetration of part of the pulse beyond the end of the waveguide. The position of the reflection point can be reliably estimated from the intersection of two tangents to the line (Fig. 10a) and enables estimation of the time of propagation to within 80 ps (Topp et al., 1980). This or similar approaches are used in software for analyzing TDR waveforms. However, in the case of a 20 cm waveguide, the 80 ps results in an uncertainty in water content of about 0.013 by volume.



(a) (b)

Fig. 10 (a) TDR waveforms produced in a wetting homogeneous soil (water content increasing 1–4), showing the method of fitting tangents to determine the reflection point. (b) TDR waveforms produced in solutions of increasing salinity (1–3), illustrating the attenuation of the waveform.

a. *Waveforms in Electrically Conducting, Lossy, Dielectrics*

TEM waves travelling through electrically conducting media are liable to attenuation. The higher frequency components of the waveform are usually lost first. As a result, the amplitude of the reflected portion of the pulse is reduced (Fig. 10b). Locating the reflection becomes more difficult, and the errors in the measurement of the travel time increase. In very conductive media, the waveguide is effectively short circuited and permittivity cannot be measured. Advantage can, however, be taken of the attenuation effect and the waveform analyzed to give the low frequency resistance and hence the bulk soil electrical conductivity of the medium through which it has travelled (Dalton et al., 1984; Topp et al., 1988; Dalton, 1992; Kachanoski et al., 1992; Heimovaara et al., 1995).

b. *Waveforms From Soils*

The waveforms obtained depend on the soil and the manner of installation of the waveguide: horizontal or vertical. Horizontally installed waveguides provide easier traces to work with because they are not usually influenced by water content gradients or other soil changes along the length of the guide. A vertically installed waveguide is more likely to pass through soil density boundaries and wetting or drying fronts that may cause additional reflections, resulting in waveforms that are difficult to interpret (Fig. 11) and that challenge the ability of software to locate the correct end point. If the reflection point can be located, the resulting measurement will represent the integrated water content over the length of the waveguide. Hook et al. (1992) designed TDR waveguides with shorting diodes that make waveform analysis easier for a vertically installed sensor.

D. Impedance Technique

Another property of transmission lines, their impedance, is used in the Theta Probe, developed at the Macaulay Land Use Research Institute (Aberdeen, U.K.). The instrument measures impedance at a fixed frequency of 100 MHz. The technique compares the impedance of a section of fixed transmission line with that of a set of stainless steel electrodes embedded in the soil, whose impedance varies with soil water content (Gaskin and Miller, 1996). The compact buriable sensor produces a voltage output and so can be interrogated with a voltmeter or connected to any logger that takes a dc input. The voltage output can be calibrated directly against water content, or alternatively calibrated to obtain relative permittivity, from which water content can be determined. The suppliers provide two calibration equations, one for mineral and one for organic soils. The volume measured by the probe is much the same as that of the corresponding configuration of TDR probe, where the sampling volume is strongly biased towards the central conductor. The sampling volume of the instrument is ca. 50 cm^3 and gives good

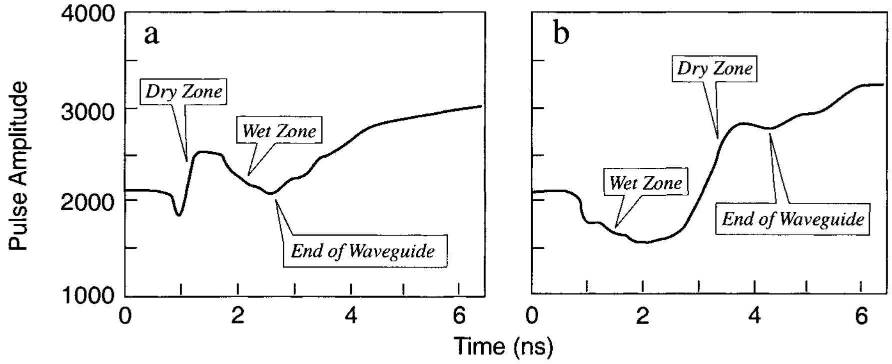


Fig. 11 TDR waveforms produced with waveguides installed vertically in soil with (a) a dry zone overlying a wet layer; (b) a wet zone over a dry layer.

averaging along the 60 mm rod length. Sensors cost about $600 each, so the system is attractive for portable and laboratory use and setups requiring several sensors.

E. Capacitance Techniques

Soil capacitance sensors measure the capacitance between two electrodes whose dielectric is partly or completely the soil to be measured. Capacitance is defined as

$$C = \epsilon_r \epsilon_0 g \tag{21}$$

where g is a geometric constant dependent on the size and arrangement of the electrodes. This measurement is difficult at low frequency unless the material is pure. Impurities lead to complications such as electrical conduction in the material and polarization of colloidal material or at interfaces. As a result, the measured capacitance is different from that of the pure material, and the calculated permittivity is incorrect. To overcome these problems, measurement at frequencies greater than 50 MHz is necessary. High-frequency capacitance can be measured in various ways, and several contrasting soil water sensors are available (Table 4).

It is important to be aware that capacitance sensors may be influenced by soil electrical conductivity, particularly those operating at <50 MHz. However, Gaudu et al. (1993) found that the effects of electrical conductivity were negligible with their system, which operates at about 40 MHz. Eller and Denoth (1996) reported a similar result with an instrument operating at about 32 MHz, except in wet organic soil, when slightly reduced accuracy, due to electrical conductivity, was evident. The IMAG DLO probe, designed to be buried or used for point measurements at the soil surface, operates at 20 MHz and measures the real (capacitive) and imaginary (conductive) parts of the permittivity independently (Hilhorst et al., 1993).

1. IH Capacitance Instruments

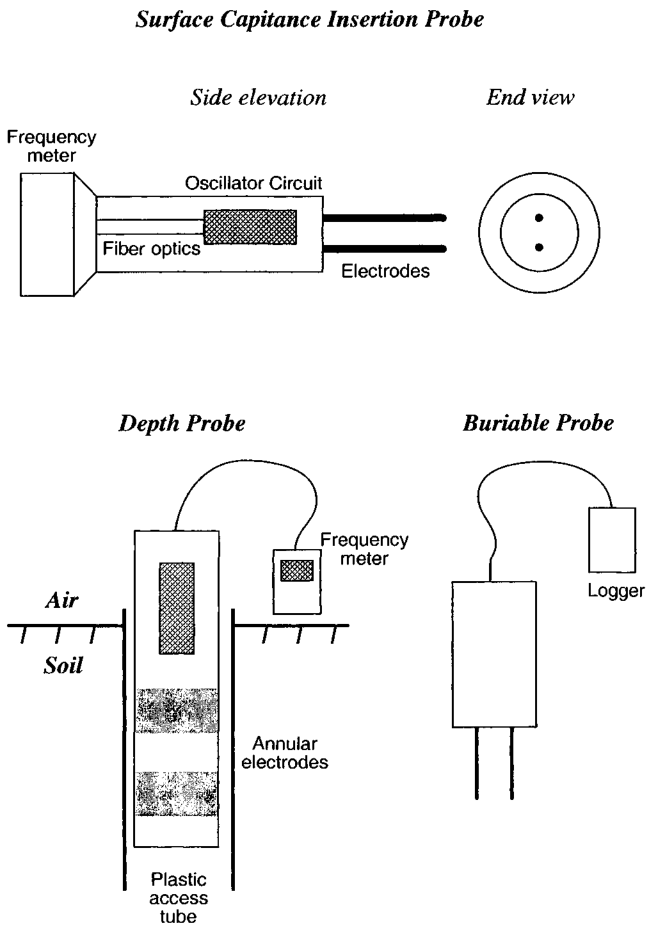
The IH capacitance systems, designed at the Institute of Hydrology (Wallingford, U.K.), give an instantaneous measurement of frequency which is a function of the electrode capacitance, from which soil permittivity can be calculated. Several instruments have been developed using the same sensor electronics (Fig. 12). A sensor that can be inserted into the soil via a plastic access tube, much as a neutron probe, is available (Dean et al., 1987). An insertion probe with two rod-shaped electrodes has been developed that can be used at the soil surface or buried (Dean, 1994), and a tine arrangement that can be towed behind a tractor has been tested by Whalley et al. (1992). The principle of operation is to use the capacitor formed by the electrodes in the soil as part of an oscillator circuit comprising capacitors, an inductor, and a driver transistor. The frequency of oscillation (F) of such a circuit is

$$F = \frac{1}{2\pi\sqrt{LC}} \tag{22}$$

where L is the circuit inductance and C its capacitance. The circuit capacitance, C, is determined mainly by the capacitance of the electrodes, which is the only variable element in the circuit. Calibration of the sensor is necessary to relate oscillation frequency to permittivity (Robinson et al., 1998). A frequency of ~150 MHz is obtained in air and ~75 MHz in water for all electrode configurations.

The design of the instrument gives the electrical field good penetrability into the material under test. The depth probe has a sampling volume of about 800 cm^3 with the field penetrating ~7 cm from the sensor body (Dean et al., 1987). The insertion probe has a sampling volume of about 500 cm^3 for 10 cm rods and 250 cm^3 for 5 cm rods and shows good averaging along the length of the rods (Dean, 1994). In soil, the frequency of oscillation is determined by a combination of the capacitance and the parallel conductance caused by electrical conduction. Ionic conductivity of the soil reduces the frequency of oscillation, but the effect is relatively small for bulk soil electrical conductivity of less than 0.05 S m^{-1} (Robinson, 1998). For higher conductivities the effect can often be compensated for (Robinson et al., 1998).

Several studies using IH sensors have related the instrument frequency reading directly to field soil water content. Robinson and Dean (1993), using the surface probe for measurements to 0.1 m depth, developed an inverse square root model to relate water content to oscillation frequency in a clay soil. Bell et al.



Surface Capitance Insertion Probe

Side elevation

End view

Depth Probe

Buriable Probe

Fig. 12 The Institute of Hydrology surface, depth, and buriable capacitance probes.

(1987) found that linear calibrations satisfactorily represented the water content–frequency relationship measured with the depth probe in four soils, over the normal range of soil water content. Evett and Steiner (1995), using a capacitance depth probe of similar design, also found linear calibrations to be most satisfactory, but Tomer and Anderson (1995), with the same type of equipment, preferred a second order polynomial to represent water content in a fine sand soil. These calibrations are all specific to both the soil and the particular instrument used. Initial calibration of the instruments, using liquids of known permittivity, allows permittivity to be determined from the frequency measurement. This allows more flexibility, permitting soil water content calibration in terms of permittivity; it also enables comparison with other dielectric methods and soil dielectric models. Laboratory trials with the surface probe have shown that well-defined relationships relating water content and permittivity are obtained for individual soils (Gardner et al., 1998). Differences between soils could be described by the parameters of a three-phase mixing model that included a bulk density term and gave results comparable to those obtained by TDR.

F. Field Installation of Dielectric Equipment

As with neutron probe access tubes, the aim during installation must be to minimize disturbance to the surrounding soil and vegetation, so that the water content measurements made are representative of the hydrology of the soil as a whole. The rod-shaped electrodes of most capacitance sensors can be treated similarly to short TDR waveguides and buried at the required depth, from the side of a pit if necessary. The access tube version of the IH capacitance probe requires installation of plastic access tubing, which can be achieved using methods similar to those used for neutron probe access tubing (Bell et al., 1987). However, the volume measured by the depth capacitance probe is smaller than that for the neutron probe, and so the effect of cavities around the tube is more serious.

The physical nature of the soil and its water content at the time of installation are important factors to be taken into account when installing both TDR and capacitance sensors. It is preferable to install sensors into wetted soil if they are to be left for any considerable length of time. Stony soils prevent the use of long TDR waveguides and make installation of depth capacitance access tubes difficult. Very stony soils may preclude any form of installation without completely disturbing the soil around the sensor.

TDR waveguides may be installed horizontally or vertically; the choice depends on the data required. Vertical installation from the surface creates the minimum soil disturbance. Probes of increasing length can be used to give soil profile water contents by subtracting the volumetric water content measured by the shorter sensors, from that measured by the longer ones. Sometimes waveguides may pass through soil horizons and/or density boundaries, giving rise to waveforms that are difficult to interpret and presenting calibration difficulties. The sensors may also act as a focal point for infiltrating water, hence giving unrepresentative field data. Horizontal installation is advantageous for measuring the water content of specific horizons and avoids the problem of channeling water down the waveguide. However, installation requires the digging of a pit, causing major soil disturbance. Hokett et al. (1992) examined the influence of soil cracks next to waveguides and found that an air-filled crack between the rods in an otherwise saturated soil could reduce the measurement of water content by as much as 46%, but water- and air-filled cracks in dry soils had little influence. The evidence suggests that in soils prone to shrinkage, where the rods may act as a focus for cracking, horizontal rather than vertical installation will give more representative results.

G. Calibration

TDR does not require calibration to measure soil permittivity if the length of the waveguide is known accurately, since electromagnetic theory relates the two as in Eq. 20. The calibration of other dielectric sensors in terms of relative permittivity can be achieved using fluids of known permittivity. Tables of the permittivity and temperature coefficients of a large range of fluids are given by Lide (1992). It is important to choose only liquids whose relaxation frequency is much greater than the operating frequency of the equipment.

Soil is inherently a complex material, and yet calibrations between soil permittivity and volumetric water content have been remarkably consistent. The initial suggestion that the relationship between permittivity and soil water content was "universal," so that once established it could be applied to all soils, is too simplistic. However, the Topp et al. (1980) calibration for TDR (Table 5) has been found to be valid for many soils and serves as a good benchmark for comparisons between TDR calibrations and those of other instruments. Different instruments operate at different frequencies, making direct comparisons between calibrations difficult. As the frequency rises, so more components such as bound water will attain their relaxation frequency, resulting in a lowered soil permittivity. In practice this means that instruments such as the IMAG-DLO capacitance probe, operating at 20 MHz, are likely to give greater permittivity measurements for the

Table 5 Empirical Calibration Equations for Obtaining θ from TDR-measured $\in_r$

Soils	Empirical formulae derived for TDR	Source
4 mineral soils	$\theta = (A + B \times \epsilon_r + C \times \epsilon_r^2 + D \times \epsilon_r^3) \times 10^{-4}$ A = −530, B = 292, C = −5.5, D = 0.043	Topp et al. (1980)
Organic soil	$\theta = (A + B \times \epsilon_r + C \times \epsilon_r^2 + D \times \epsilon_r^3) \times 10^{-4}$ A = −252, B = 415, C = −14.4, D = 0.22	Topp et al. (1980)
Loam	$\theta = 0.1138\sqrt{\epsilon_r - 3.38\rho_b - 0.1529}$	Ledieu et al. (1986)
10 mineral soils	$\theta = (A + B \times \epsilon_r + C \times \epsilon_r^2 + D \times \epsilon_r^3 - 370\rho_b + 7.36 \times \%\, clay + 47.7 \times \%\, org.mat.) \times 10^{-4}$ A = −341, B = 345, C = −11.4, D = 0.171	Jacobsen and Schjonning (1994)
62 mineral/organic soils and porous media	$\theta = \dfrac{\sqrt{\epsilon_r} - 0.819 - 0.168\rho_b - 0.159\rho_b^2}{7.17 + 1.18\rho_b}$	Malicki et al. (1996)

same water content than the Topp et al. (1980) calibrations determined using TDR (~ 200 MHz), as found by Perdok et al. (1996). However, although the calibrations may differ, the influential soil factors will, for the most part, be the same.

The number of published calibration models is growing as more measurements are taken, but most apply to TDR. The applicability of any model should be verified where possible by conducting at least a limited calibration for the soil concerned. Calibrations for systems other than TDR are limited, so these instruments will normally require calibration. There is as yet no standard method for calibrating dielectric instruments in terms of soil water content. Some calibration methods are more representative of field conditions than others, but the choice of method will also be based on other factors, including time available and the range of water content required.

1. Field Calibration

The principle of field calibration is the same as for deriving calibrations for the neutron method. Measurements are made, and immediately undisturbed soil samples of known volume are collected from the measurement point, for water content determination by oven-drying. Depending on the type of equipment, and the depth of the soil, it may be possible to sample the volume of soil where the instrument measurement was made. Such an approach assumes temporary installation of equipment and is destructive. Sampling at a greater distance from a permanent equipment installation may be preferred. Alternatively, for the depth capacitance probe, samples can be taken from the access tube at the time of installation (Bell et al., 1987). Covers and irrigation may be used to extend the range of water content involved.

2. Laboratory Calibration

Laboratory methods offer the advantage of being in a controlled environment. The most rapid method is to wet air-dried sieved soil with deionized water using a mist spray while mixing continuously (Malicki et al., 1996). The soil is then packed into a known volume and weighed; the electrodes or waveguides are inserted and measurements taken immediately. A small sample of the soil, ~50 g, is then removed for oven-drying and water content determination as a mass ratio. Volumetric water content is calculated knowing the weight and volume of the packed soil. The soil can be packed to different bulk densities and measurements for a wide range of water content achieved by gradual wetting. Perdok et al. (1996) used a triaxial soil press to provide soil cores with different bulk densities in which to calibrate the IMAG DLO capacitance probe. A complete calibration curve can be derived in two days, allowing overnight drying of the samples for water content determination.

Undisturbed cores from the field can be used (Heimovaara et al., 1994) so that the complete range of soil water content can be achieved on cores that are as close to their field condition as possible. For most equipment, a core of about 10 cm diameter and 15 to 20 cm length is large enough. Cores need to be encased and a perforated base should be fitted, so that in the laboratory they can be wetted from the base upwards, preventing air entrapment. Cores are saturated using deionized water, and then the electrodes/waveguides are inserted and measurements begun. On each measurement occasion the core is also weighed. The cores will dry out in the laboratory from the open top and through the perforated bottom. Drying can take up to two or three months. Finally, the soil core is removed for oven-drying, and the water content on each measurement occasion is calculated from the corresponding weights. At least two cores must be taken for comparison, as natural inhomogeneities such as stones may cause unrepresentative calibrations. Shrink/swell soils are difficult to deal with in this manner. An alternative approach along similar lines is to sieve soil and pack a core and then to treat the core as above. This homogenizes the soil and eliminates the possibility of large stones, cracks, or pores influencing the calibration.

H. Influence of Soil Properties on Calibrations

1. *Soil Temperature*

The relative permittivity of water decreases almost linearly by 0.36 per °C as temperature rises between 5 and 50°C (Lide, 1992). The permittivity of the solid components is likely to change very little with temperature, and so the average change in soil permittivity with temperature will be less than that for pure water. Experiments by Topp et al. (1980) demonstrated that, for the soils used in their experiment, there was a negligible temperature effect in the range of 10–36°C. Halbertsma et al. (1995) showed that the incorporation of temperature compensation for the permittivity of water into a mixing formula replicated data for sand, but in a clay soil no noticeable change of permittivity occurred with an increase in temperature, and so application of the model overestimated the soil water content. For most purposes, with temperature-stable equipment, it is likely that the effect of temperature on permittivity will be small compared with the other errors in the calibration process.

2. *Bulk Density and Soil Mineralogy*

Bulk density, directly or indirectly, has a significant influence on the calibration of dielectric techniques. Topp et al. (1980), using a limited number of soils, found that bulk density was not an important factor in the calibration they produced. Subsequent work on a wider range of soils found that incorporation of bulk density into calibrations improved results (Ledieu et al., 1986; Jacobsen and Schjonning, 1994). The semiphysical mixing model presented by Whalley (1993) gives a physical explanation of the effect of bulk density. The linear model (Eq. 18) shows that the intercept is a function of the permittivity of the solid and its dry bulk density. This approach has proved useful for exploring the dielectric properties of soil in a physical rather than an empirical way (Robinson, 1998). Work with capacitance instruments has also found that bulk density should be incorporated into calibrations (Perdok et al., 1996; Gardner et al., 1998).

The most likely effect of an increase in soil bulk density is to increase the permittivity of the soil. Jacobsen and Schjonning (1994) suggested that the effect of change in bulk density was more than could be accounted for by a change in the ratio of solids to voids and their respective permittivities. As the effect is most noticeable in certain heavier textured soils, it is likely that this is associated with the clay content. As a clay soil becomes more dense, the quantity of bound water increases, and therefore one might expect a decrease in soil permittivity at the same water content, as bulk density increases. The four-phase mixing formula, Eq. 17, gives, using $\epsilon_a = 1$:

$$\theta = \frac{\epsilon^{\alpha} + (\epsilon_{w}^{\alpha} - \epsilon_{bw}^{\alpha})f_{bw} - (\epsilon_{s}^{\alpha} - 1)(\rho/\rho_{p}) - 1}{\epsilon_{w}^{\alpha} - 1} \tag{23}$$

where $\theta > f_{bw}$. Typically, $\epsilon_s = 3.5$, $\epsilon_{fw} = 81.0$, $\epsilon_{bw} = 3.2$, $\rho_p = 2.56$, and values for α range from 0.46 to 0.70 (Dirksen and Dasberg, 1993; Roth et al., 1990). This equation combines the effect of both bulk density and surface area changes (Fig. 13). However, changes in bulk density produce a proportionate change in surface area per unit volume and hence in the amount of bound water, which may be a large fraction of the total water in a clay soil. Peplinski et al. (1995) suggested a refinement of the methodology by incorporating the known surface properties of specific clay minerals into the calibration relationship.

Certain minerals may influence soil dielectric properties and thus calibrations because the solid itself has a high permittivity (Roth et al. 1992; Dirksen and Dasberg, 1993; Robinson et al., 1994; Peplinski et al., 1995). Robinson et al. (1995) demonstrated that iron minerals such as haematite and magnetite had higher permittivities than the values of 4 to 6 normally found in common soil minerals. Some titanium and aluminum hydroxides may also fall into this category and might influence calibrations performed in tropical soils.

3. Organic Soils

Topp et al. (1980) demonstrated, using TDR, that the calibration relationship for an organic soil with a bulk density of 0.422 Mg m^{-3} was significantly different from the calibration found for mineral soils. This finding was supported by Stein and Kane (1983), Pepin et al. (1992), and Roth et al. (1992) for peat soils with

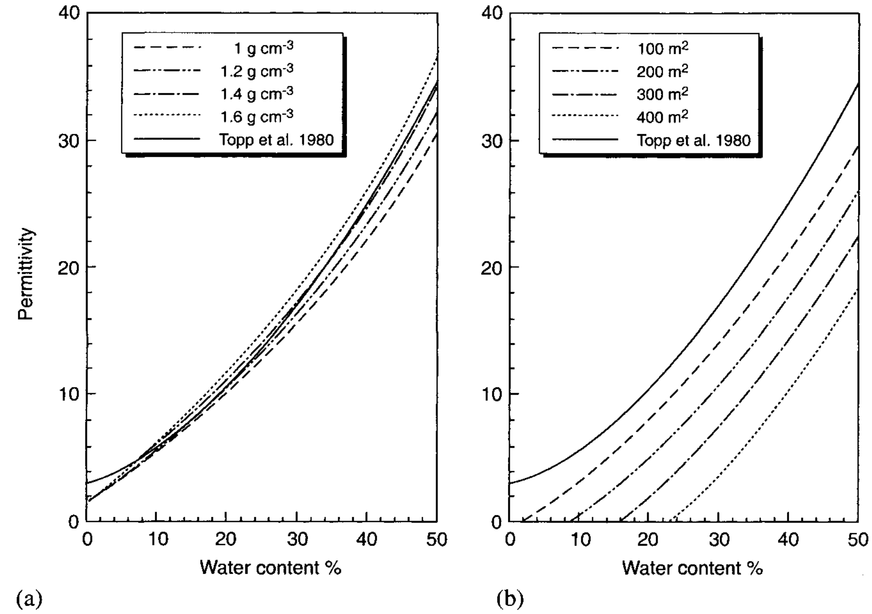


Fig. 13 The effect on the permittivity/water content relationship of (a) increasing bulk density; (b) increasing surface area per unit of soil. (After Dirksen and Dasberg, 1993.)

bulk densities ranging from 0.06 to 0.25 Mg m^{-3}. A calibration derived from measurements in several peat substrates was found to be similar to that of Pepin et al. (1992) by Paquet et al. (1993).

VI. APPLICATIONS OF NEUTRON AND DIELECTRIC METHODS

The examples reviewed briefly in this section illustrate how the neutron and dielectric measurement methods have been used in practical applications. Because neutron probes have been available for so much longer, there are many more reports in the literature of their use. Examples of the application of dielectric methods, particularly capacitance methods, rather than publications on the calibration or evaluation of sensors, are as yet less usual.

Neutron probes have been used most often to measure water content change to depth in the field at weekly, or sometimes more frequent, intervals. Water content distribution has been measured beneath crops (e.g., Bautista et al., 1985), and the soil water regime of different soils and vegetation types, varying from arid rangelands (Nash et al., 1991) to equatorial forest and cleared areas (Hodnett et al., 1996), has been characterized. Soil water content data are frequently collected to measure crop or soil water balances, where the focus of interest may be the soil evaporation and/or plant transpiration components, or the subsurface and deep drainage (recharge) components. McGowan and Williams (1980) used the depth of the drying front, measured by neutron probe, to define the depth above which water content loss was due to evaporation and transpiration, and below which water content change could be ascribed to drainage, and hence derived a catchment water balance (McGowan et al., 1980). Often additional measurements, particularly of soil matric potential, are made to enable partitioning of water content change in the profile into evaporation (including transpiration) and drainage (e.g., Sophocleus and Perry, 1985; Cooper et al., 1990). Neutron probe measurements have been particularly useful in the study of the hydraulic properties of the unsaturated zone of deep aquifers such as the English Chalk and sandstones (Gardner et al., 1990; Cooper et al., 1990) because it is possible to make measurements to depths of 4 m or more.

In many cases, dielectric monitoring methods could have been used to obtain much the same information, with the advantage that more frequent and automated monitoring, if required, would have been feasible. However, measurements at depths greater than about 1 m using TDR or buried capacitance sensors would have necessitated excavation of pits from which to install equipment, entailing some disturbance to the soil's hydrology. The essential difference between the neutron probe and dielectric methods is that neutron probes permit measurement at many depths (to ≥5 m) infrequently, whereas most dielectric methods permit measurement at relatively few depths (due to cost), but with high temporal frequency. TDR has been used successfully in various field studies to obtain frequent measurements of water content, though generally not to depths much below 0.5 m. The aim of these studies has varied from characterizing soil water regimes in time and space (Van Wesenbeeck and Kachanoski, 1988; Herkelrath et al., 1991; Nyberg, 1996) to determining soil evaporation and transpiration rates (Zegelin et al., 1992; Plauborg, 1995). These studies used vertically installed waveguides of different length to monitor water content distribution by layer in the soil profile, but others have used horizontal installations in similar work. Nielsen et al. (1995) set out to study the immediate surface soil and used horizontally installed waveguides for measurements at just 25 mm depth. Measurement at shallower depth, 13 mm, proved unreliable, however.

Other examples of in situ use of TDR include work in peats, including very low density ones (Pepin et al., 1992). Parkin et al. (1995) measured unsaturated hydraulic conductivity using TDR to 0.4 m depth in field plots irrigated using a rainfall simulator. Temporal variations in soil water composition have been investigated by Heimovaara et al. (1995), both in the field and in laboratory cores, using TDR to monitor both water content and bulk soil electrical conductivity, in combination with soil solution sampling.

The neutron method is much less versatile than dielectric methods for container, glasshouse, and laboratory work, but equipment to permit such experimental work has been designed, e.g., Klenke and Flint (1991) described a neutron collimator for use with a CPN 503 probe. The good space and time resolution of TDR measurements has been used effectively in container studies of water uptake by roots (e.g., Wraith and Baker, 1991; Heimovaara et al., 1993). Topp et al. (1996) were able to record the diurnal uptake of water from, and its release to, relatively dry soil in which maize roots were growing. The Easy Test miniprobe, because of its small size, lends itself to this type of study and has been used, with minitensiometers, to obtain soil water release and hydraulic conductivity functions in undisturbed soil cores 100 mm high and 55 mm in diameter, as the cores dried from saturation (Malicki et al., 1992).

Neutron probes are being used increasingly in work associated with potential environmental pollution due to leakage from landfills and accidental spillage of contaminants. Prospective landfill and hazardous waste sites have been characterized for their suitability prior to use and monitored thereafter (Unruh et al., 1990). For example, Kramer et al. (1995) used a 670 m access tube installed horizontally beneath the leachate collection system of a municipal landfill to detect leachate leaks. No attempt at calibration was made; changes in neutron count with distance along the tube, and with time, were interpreted in terms of water content.

Provision of irrigation scheduling advice on the basis of both neutron probe and dielectric measurements is a service industry in high-value crop growing areas of several countries. Remote interrogation of TDR or capacitance sensors installed in farmers' fields will permit the same information to be gained more cheaply and open up the possibility of using more sensors to define crop water requirements better. Design of intelligent irrigation systems incorporating dielectric sensors to monitor water content, and hence water need, are well underway (e.g., Miller and Ray, 1985). Connecting TDR or capacitance sensors to systems that measure soil temperature, rainfall, soil matric potential, and any other parameters that may be required opens up the possibility of studying soil hydrology and crop water use to a level of detail not previously feasible. The U.K. Institute of Hydrology has an operational Automatic Soil Water Station that combines these sensors, using buried capacitance probes for the water content measurements. The possible uses for such systems in research and commercial applications are only just being explored. The revolution in soil water content measurement that dielectric methods have sparked is already having an impact in soil and environmental work beyond the dreams of most earlier neutron probe users.

VII. REMOTE SENSING OF SOIL WATER CONTENT

The development of remote sensing, which was given considerable impetus by the Soviet and U.S. space programs in the early 1960s, is now a flourishing subdiscipline with a wide range of applications in the monitoring of many aspects of the environment. In remote sensing, several methods are used to convey data about