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Soil Water Measurements Relevant to Agronomic and Environmental Functions of Chemically Treated Soil

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ABSTRACT: Modern agricultural, turf, and landscape management routinely apply and depend upon chemical applications to optimize system function for specific outcomes. The effectiveness of these applied chemicals to achieve desired outcomes usually depends upon their interaction with and transport by water. To fully and accurately assess the role of water as a chemical delivery and activation system requires a good understanding of how the applied chemicals, soil, and water interact, the scale at which a phenomenon is important, the nature of soil variability, and which of the three dominant soil water properties (content, movement, or potential energy) is most suited to assessing water's role. The science of this assessment process is extensive and its literature is voluminous. For the uninitiated, however, it is worth being aware at least of the basics of soil water assessment and where some of the pitfalls lie. This presentation describes soil as a three-phase system (solids, liquid, and gases) and highlights some of the key measurements and measurement considerations necessary to appropriately characterize treatment efficacy for specific, and especially, non-intuitive effects. The presentation cannot be comprehensive or substitute for requisite university-level courses in soil physics and soil chemistry, but can, perhaps, alert applicators to situations and considerations that demand more than mere cursory assessment for proper evaluation and interpretation.

KEYWORDS: water potential, infiltration, hydraulic conductivity, miscible displacement, psychrometer, tensiometer, TDR, neutron attenuation, surface seal

Nomenclature

Indices

Ψ_g = gravitational potential
Ψ_m = matric potential
Ψ_p = pressure potential
Ψ_o = osmotic potential
K = unsaturated hydraulic conductivity
K_{fs} = field-saturated hydraulic conductivity
K_{sat} = saturated hydraulic conductivity
R = retardation factor
RO = runoff volume
J_w = water flux density
T_c = water volume transpired by crops
T_w = water volume transpired by weeds
P = crop production
W = water volume applied
WUE_{ag} = agricultural water use efficiency
DP = water volume drained to subsoil
E_d = water volume evaporated in delivery
E_s = water volume evaporated from soil
C_o = solute concentration in incoming fluid
C/C_o = relative solute concentration
D = dispersion coefficient

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Greek Symbols

Ψ = total water potential
θ = volumetric water content
ν = pore water velocity

Abbreviations

SOM = soil organic matter
TDR = time domain reflectometry
AC = alternating current
EC = electrical conductivity
SAR = sodium adsorption ratio
WDPT = water drop penetration time
WUE = water use efficiency
CDE = convective-dispersive equation
ADE = advective-dispersive equation
t = time
BMP = best management practice
ODR = oxygen diffusion rate

Introduction

When chemicals are applied to managed turf, to agricultural fields, landscaping, construction sites, or road ways, they usually interact with applied irrigation water or rainfall. The chemical application is usually to deliver the chemical for the purpose of affecting soil nutrient status, pH, to deliver dissolved or suspended herbicides or pesticides, or to alter soil physical properties — either solid phase properties or hydraulic properties. The importance of the interaction with water can be related to proper delivery of the chemical, or to the subsequent behavior of water in the soil due to the applied chemical, or sometimes both.

To best understand these phenomena, their interactions, and to become adept at the variety of measurements required to characterize them, would require several courses in soil chemistry and soil physics. Our aim in this article is to help those unfamiliar with these concepts and measurements to at least be aware of the fundamentals and, hopefully, avoid the most egregious pitfalls. For more understanding of the concepts, we recommend the texts by Scott [1] and Hillel [2]. For compendia of methods, we direct the reader to Dane and Topp [3] and Carter [4]. Some excellent resources can also be found on the web. Three web sites that we recommend are listed in the Appendix.

After a short introduction to the three-phase soil conceptual model, we will look at the individual phases: soil solids, soil water, and soil air. The key concepts that define the individual phases will be discussed, and test methods/apparati used to measure these phases will be highlighted. The most extensive coverage is given to the water phase, where the concepts of soil water content, soil water potential, soil water movement, infiltration, hydraulic conductivity, field water balance, solute transport, and miscible displacement will be discussed.

Three-Phase Soil Conceptual Model

Soil is a three-phase system that includes the soil solid phase, the soil liquid phase, and the soil gaseous phase (Fig. 1). The solid phase consists of minerals and organic materials. The liquid phase, sometimes referred to as soil water or the soil solution, contains mineral salts and soluble organic compounds dissolved in water. The gaseous phase, sometimes called the soil atmosphere or soil air, contains a dynamic mixture of gases found among the voids between soil particles that are not filled with water. Important gaseous components may also exist in the dissolved state in soil water. Although dissolved gases occur in very small amounts, they can be important to chemical reactions. Physically, however, dissolved gases perform as part of the soil solution and are usually only important as a gas component when changes in temperature or pressure bring about a change of state via off-gassing from the liquid. The proportions and constituents of these three phases change with time due to tillage, traffic, and other management; with the

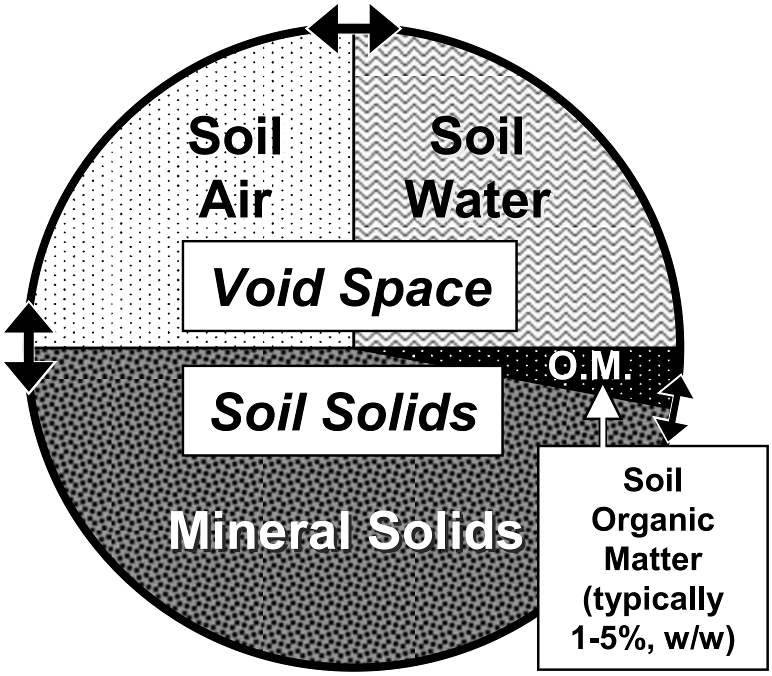


FIG. 1—*Soil as a three-phase system, with solids, liquids, and gases present in varying proportions. Soil organic matter is considered part of the soil solid phase.*

addition of water via rainfall or irrigation; with internal redistribution of water in response to a variety of forces including gravity, pressure, temperature, and particle surface attractions, as well as vapor transfers and condensation; and the loss of water via drainage and evapotranspiration. As soil water content changes, the gas volume changes through physical displacement (Fig. 2). Gas concentrations also change as a result of chemical reactions and biological activity affected by water. Because of hydration effects

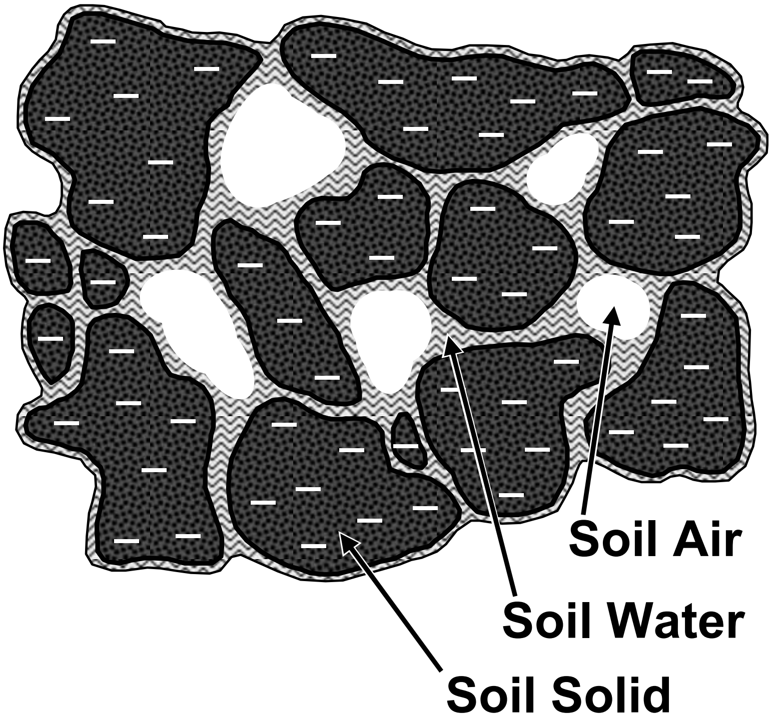
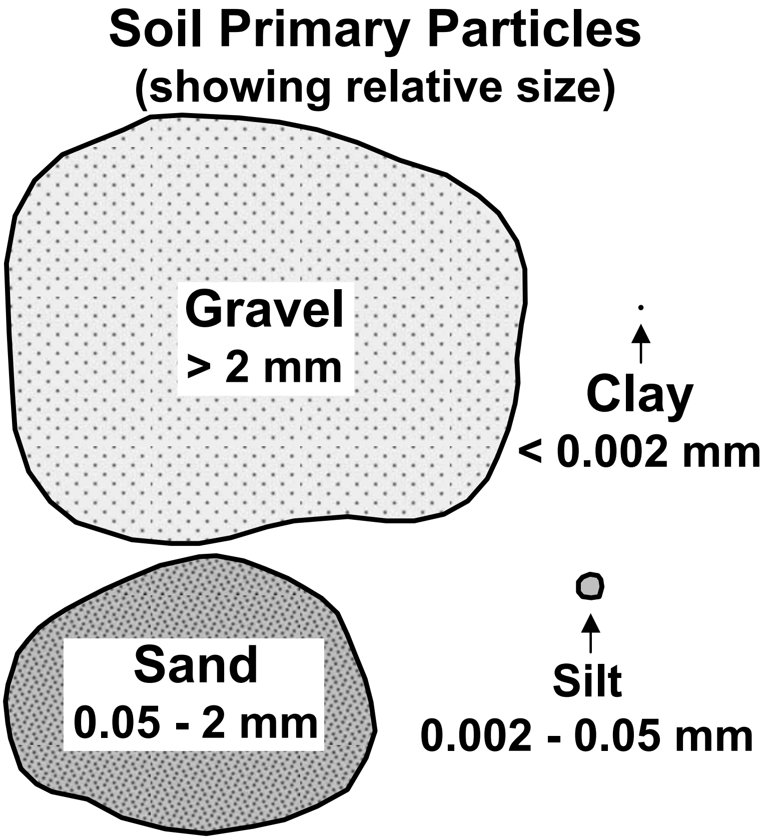


FIG. 2—*Physical arrangement of soil solids, water, and air. Air and water are present predominantly in soil voids or pores. As the soil's water content increases, the volume occupied by soil air decreases and vice versa. The surfaces of soil solids are negatively charged.*



Sand Fraction

Classification	Diameter (mm)
Very Coarse	1.0 – 2.0
Coarse	0.5 – 1.0
Medium	0.25 – 0.5
Fine	0.10 – 0.25
Very Fine	0.05 – 0.10

FIG. 3—*Soil texture is defined by the proportions of sand, silt, and clay in soil. The sand fraction can be further classified, if desired. Gravel is usually omitted when soil texture is measured.*

(and soil solution electrolyte-specific effects) on organic matter and clay minerals, water content can also affect the physical volume and configuration (pore size and pore distribution) of the solid phase.

The Solid Phase

The solid phase, often referred to as the soil matrix in order to draw attention to the arrangement of solids and voids, is described in soil science by its texture, mineralogy, organic matter content, and structure. Texture is defined by the proportion of clay-sized, silt-sized and sand-sized mineral primary particles (Fig. 3), as described in the soil textural triangle (Fig. 4) [5]. Texture is the most stable physical property of soil and a useful index in determining the soil's agricultural potential. Fine- and medium-textured soils, such as clays, clay loams, and silt loams, are generally most desirable for crop production because of their superior

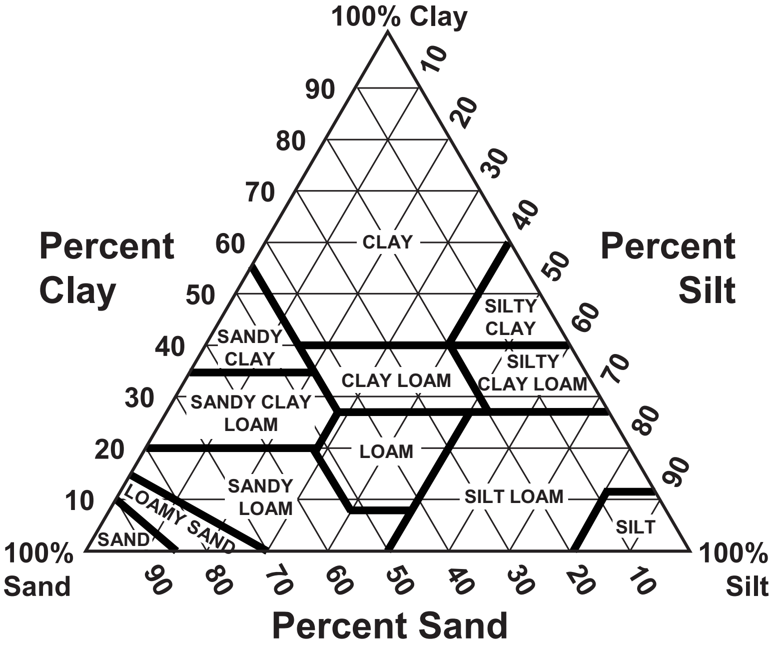


FIG. 4—*The soil textural triangle, indicating the proportions of sand, silt, and clay in each of twelve soil types.*

retention of water and nutrients. Conversely, somewhat coarser textures such as sandy loams and sands can be beneficial in many irrigation schemes where high infiltration rates and good drainage are required. Texture influences the water-holding capacity of the soil which, in turn, affects soil temperature. Clays hold more water per unit volume at a given energy status (soil water potential) than sandy soils. Therefore, the temperature of a wet clay soil, compared to that of a drier sandy soil, responds more slowly to changes of air temperature in the spring and autumn due to the high heat capacity of water. The hydrometer method, described in Dane and Topp [3], is often used to measure soil texture. Because texture is defined by proportions of primary mineral particles, texture cannot change unless there is a removal or enrichment of a size class from the mix. This does not usually occur except in geologic time spans. Soil amendments are often touted as being able to change texture. If any change is occurring, it is usually a change in structure, which, to the uninformed, may look or feel like a change in texture.

Soil organic matter (SOM) is important in the formation and stabilization of soil structure and typically accounts for 1 % to 5 % of soil by weight (Fig. 1). SOM arises from both above-ground and below-ground plant tissues, animal residues and excreta that are deposited on the surface and mixed with the mineral component. During decomposition, which also includes processing by soil mesofauna such as earthworms, ants, etc., organic matter is reduced to chemically more stable forms, with some organic compounds converted to inorganic compounds through mineralization. The mineralized elements that are essential for plant growth are incorporated into the biomass of higher plants and soil microorganisms and enter a cycle of dynamic nutrient release and incorporation.

During decomposition, SOM provides the organic adhesives that are essential to the formation and stability of soil aggregates, which impart a structural attribute termed soil tilth. Soil tilth is the highly desirable soil physical condition that permits free entry and movement of water and air through soil pores (Fig. 2), easy cultivation and planting, and healthy plant growth [2]. There are three broad categories of soil structure: single-grained, massive, and aggregated [5]. Coarse granular soils that are completely loose with primary particles that are unattached to each other are considered single-grained. Soils that are tightly packed in large cohesive blocks or layers are considered massive. Aggregated soils have primary particles clustered in naturally occurring quasi-stable crumbs, known as aggregates or peds. Aggregated soils generally have the most favorable soil structure for plant growth. The stability of soil aggregates is often measured by wet sieving [6]. Tensile strength of aggregates, the force per unit area required to cause structural failure of the aggregate, is also a useful measure of aggregate durability [7,8]. The friable condition of agricultural soil (the ability of aggregates to retain their individual integrity while still permitting easy aggregate separation via tillage or root penetration) is a desirable feature for establishing seedbeds during tillage.

Porosity and bulk density of the soil matrix are features of the soil profile affected by the size and arrangement of aggregates. Total porosity and pore size distribution are important factors affecting water movement. Porosity is defined as the ratio of the total volume of pores to the total soil volume (that is, the volume occupied by both the pores and soil solids) and is often calculated from the bulk density of the soil [9]. Soil bulk density, defined as the mass of oven-dry soil per unit volume of soil, is also used as an indicator of soil compaction. Soil compaction is a process that changes the arrangement of soil solids resulting in altered structure and reduced total porosity (and a pore size distribution that contains fewer large pores and fewer continuous pores than were present initially). Compaction can eventually reduce porosity enough to restrict aeration, impede root penetration, water infiltration and drainage, as well as change water retention characteristics and capillary movement of water. At different times and for different purposes, the optimal bulk density may not be the same as the minimal attainable bulk density [2]. Scott [1] and Hillel [2] present a number of simple equations to characterize the mass and volume relationships of the soil three-phase model.

The status of the solid phase is important, not only because of the direct solid phase properties per se, but also because of the influence of solid phase attributes on the liquid and gaseous phases. The three phases interact, with feedbacks to each separate phase. This is particularly true of the relationship between the solid and liquid phase. Water softens soil (reduces soil strength), increasing its proneness to compaction. Water can also disperse fines and redistribute the fines, filling pores, which in turn can increase soil strength in depositional areas upon drying (common in soil crusting), and cause the formation of surface-soil hydraulic "seals" that restrict infiltration of water, increasing the likelihood and amount of surface runoff.

The Liquid Phase

Water is vital to life in the soil and possesses several unique properties. Water has the greatest specific heat capacity, latent heat of vaporization, surface tension, and dielectric constant of any known liquid. Soil water status is generally characterized in terms of either soil water content or soil water potential (soil water's energy status).

Soil Water Content

Soil water content is characterized either as gravimetric water content (mass of water divided by the mass of soil oven-dried at 104°C) or as volumetric water content (volume of water divided by the total volume of the soil sample). Numerous considerations go into the decision of how best to evaluate soil water content and which sensors or evaluations systems (e.g., water content vs water potential) are the most appropriate for the needed assessment [10]. Volumetric water content is usually the preferred measurement when interpreting water content for practical field considerations because it directly accounts for the effects of structure and bulk density, whereas gravimetric content obscures them. Volumetric water content can be converted reliably to gravimetric content by drying a known volume of soil (e.g., a bulk density sample), but the reverse is not true. This is because the relationship of soil water content to soil volume at higher matric potentials (those from 0 to -300 kPa [4]) is dependent upon structure, which affects water retention properties. If gravimetric water content is determined on a disturbed soil sample, extrapolation of the relationship to a field-relevant volumetric water content will vary with the assumptions made about structure, bulk density, and water release properties. One can determine a soil sample's volumetric water content by multiplying a gravimetric water content by a known bulk density.

There are several common methods for measuring soil water content. Only a few are direct measures. Most use inferences from other measured phenomena that are correlated with water content, or if preferred, or needed, the phenomena can be correlated with water potential. Weighing soil samples before and after oven drying gives gravimetric content (and if the sample volume is known, also gives volumetric content). This is the simplest of soil water measurements, and perhaps the most reliable, since it has the least number of intermediate interpretive steps, and thus fewer sources of error. All other measures of water content must be directly or indirectly correlated to gravimetrically determined water content.

In all the methods we describe, proper sample handling to prevent evaporation and condensation anomalies when collecting, transferring, or storing soil samples prior to oven drying is important for accuracy and precision. All indirect measures require separate calibration for soil horizons with differing soil water retention properties. Neutron and capacitance probes lose reliability close to the soil surface, where the neutron cloud or signal energy are not completely captured in the soil volume. Consistent intimate soil contact of access tubes is important for neutron and capacitance methods. Consistent probe orientation from one insertion to the next is also important. Tensiometers, resistance blocks, and thermocouple psychrometers also require intimate contact with the soil matrix for proper instrument function. Shrink-swell soils pose significant problems for any procedure requiring consistent intimate soil contact with sensors or access tubes.

A water retention curve gives the soil's water content at a given matric potential (suction). These curves are most useful when water content is expressed on a volumetric rather than gravimetric basis. Water retention curves are usually determined with a pressure-plate apparatus. Pressure plates are ceramic plates having a rubber bladder on the bottom that collects and channels water expelled by increased air pressure from soil samples positioned on the ceramic upper face. The ceramic and soil samples are sealed in an air-tight chamber subjected to positive pressure. The positive pressure neutralizes the "suction" (surface adsorptive forces) holding the water in the soil and thus is equivalent to the negative of the soil's matric potential. The bladder is tapped to allow collection of incremental amounts of water forced out of the soil samples as air pressure in the chamber is increased. Each increment of water expelled is quantified. At intermediate points and at the end of the pressurization series, the soil sample's weight is determined via oven drying. The quantity of water remaining in the soil after equilibration at each pressure increment is the water retention at that matric potential. If the sample is either an intact or re-packed core of known volume, then the gravimetric values can be converted to volumetric values. Water retention curves are normally determined in small increments in the wet range, where drainage tends to be large per unit potential change, and determined at larger increments in the drier range. The water retention curve, thus determined, becomes essential to interpreting soil water content from the host of other measurement techniques that indirectly determine water content, by relating an indirect physical phenomenon to water potential, or to water content.

Neutron attenuation (or thermalization) uses a small radioactive source of high energy (fast) neutrons that, through repeated collisions, transfer much of their kinetic energy to the protons in the nuclei of hydrogen atoms (predominantly in water molecules) in nearby soil, creating a detectable backscatter of low energy (slow or thermal) neutrons [11]. The source is located in a cylindrical probe that is lowered into the soil via pre-installed access tubes. The probe also contains a detector that is only sensitive to low energy neutrons (the ones returning to the detector via backscatter). The rate of slow neutron return to the detector is monitored electronically via a meter connected to the detector and correlated with volumetric water content in specific horizons or at specific depths. A calibration relating slow neutron counts per unit time to measured volumetric water content is prepared for each horizon (or depth) to maximize precision and accuracy.

Neutron moisture meters have advantages and disadvantages [11,12]. They are simple and easy to use, accurate (<1 % error) when properly calibrated, can quickly monitor water content non-destructively and repeatedly in exactly the same series of positions to great depth in soil profiles above a water table. They are safe to use with only minimal training. However, they carry a considerable regulatory burden, including required licensing. Meters (probes) cannot be used or left unattended, and they require doubly locked storage. They must be calibrated for each depth position monitored. They do not give accurate soil water estimates for the surface 15 cm to 20 cm of soil, due to incomplete and inconsistent containment of the neutron cloud emanating from the source. Access tubes must keep in consistent orientation and intimate contact with the soil profile (difficult in shrink-swell soils). The access tubes are not easy to properly install and they interfere with field operations. Many technicians express concern regarding radiation exposure (regardless of the small source and strict dose monitoring) and are reluctant to make these measurements, especially on a frequent basis. Nonetheless, many researchers feel that the neutron attenuation technique can yield soil water content estimates second in precision only to volumetric soil sampling.

Time domain reflectometry (TDR) senses soil volumetric water content. The technique measures the travel time of an electronic pulse sent through a wave guide (probe) inserted into soil. Pulse travel time responds to changes in the apparent permittivity of the soil surrounding the probe [13,14]. Since the permittivity of soil mineral matter ranges from 3 to 5 and that of water is about 80, measured soil apparent permittivity correlates with water content [15]. TDR is commonly used in research because its travel time is little affected by soil bulk density, salinity, or clay content. Because of the popularity of TDR, some other systems are touted as TDR-like systems even though they do not directly measure pulse travel time or output a trace termed a waveform, which is central to a true TDR system [12,13]. Most of these other systems measure frequency shifts rather than pulse time and are more properly described as capacitance systems.

TDR has advantages and disadvantages [12]. TDR usually needs no calibration, is accurate, rapid, and non-destructive, readings can be automated, multiple probes can be multiplexed, portable systems are available, and can be used remotely and unattended. However, TDR equipment is expensive ($3k to $7k) and data retrieval and processing require technical expertise [16]. Probes require intimate and consistent probe-to-soil contact and their installation may disturb some soils. With either two- or three-rod probes, TDR senses volumetric water content in soil, on average, in a band about 20 mm above and 20 mm below the plane of the rods, and extending about 20 mm to the outside of the outermost rods [12,17]. Whether the size of the sensed volume is an advantage or disadvantage depends on the system and research need [17].

Capacitance systems operate in the frequency domain, rather than in TDR's time domain. In a capacitance sensor-soil assemblage, changes in the resonant frequency of an electrical signal sent through the system's embedded circuit depend upon changes in the capacitance of the sensor-soil assemblage [17,18]. Capacitance changes in the sensor-soil assemblage, in turn, depend upon not only the nearby soil's water content, but also upon the soil's temperature, bulk density, and porosity [13].

On the positive side, capacitance sensors are often portable, some can be used with access tubes, some can be used unattended, and some can be automated and multiplexed [12]. The initial costs of some capacitance sensors may be less than TDR systems or neutron meters. However, capacitance sensor data are significantly affected by soil temperature and, to a lesser degree, by soil clay content, bulk density, and salinity. Moreover, capacitance sensors are more sensitive to these factors than TDR or neutron systems [13]. Other disadvantages include calibration need for different soil types and horizons within each soil type. Good contact with surrounding soil is essential [13,17]. Capacitance sensors can also measure water content in a very small soil volume, often smaller than the soil's representative elementary volume for water content [19]. Capacitance sensors in access tubes have very narrow radial and shallow vertical sensitivity. Calibration equations relating sensor output to soil volumetric water content can be curvilinear, with steeper slopes for wetter conditions, increasing the variation in estimated water contents [13].

Soil Water Potential

Water potential is defined as the potential energy per unit mass of water with reference to pure water at zero potential. The total water potential (Ψ) is defined as:

$$\Psi = \Psi_g + \Psi_m + \Psi_p + \Psi_o \tag{1}$$

where Ψ_g is the gravitational potential due to the water's position in a gravitational field, Ψ_m is the matric potential due to the attraction of the soil matrix for water molecules, Ψ_p is the pressure potential due to hydrostatic or pneumatic pressure, with the pressure potential equal to zero if the water is positioned above a water table, and Ψ_o is the osmotic potential due to dissolved solutes in the presence of a semi-permeable membrane [20].

Tensiometers measure soil water potential from zero potential (0 kPa or saturation) to about −0.8 atmosphere or −80 kPa (slightly damp). The instrument is a mechanical device that holds water in a reservoir (typically a pipe or plastic tube) connected to a ceramic cup that allows very slow transfer of liquid water from one side of the ceramic to the other. The tube has a gauge that registers water tension (negative of the water's matric potential) via mechanical distortion of a thin hollow coil connected to the water column. The tube is filled with water and stoppered to create a water-tight system except for the capacity of the ceramic cup to allow very slow water migration. The tube is inserted into the soil via a pilot hole, centering the ceramic cup at the depth of interest. Then the hole is back-filled with a soil slurry to ensure intimate contact of the cup with the surrounding soil. As water tension develops in the soil, a tiny volume of water migrates into or out of the tube through the ceramic cup until the water in the tensiometer attains potential energy equilibrium with the water in the surrounding soil. That equilibrium potential (or tension) is registered on the tension gauge. The reading, after adjustment for the height of the gauge above the ceramic cup, is the matric potential at the center of the ceramic.

Tensiometers are simple, inexpensive, and reliable instruments if used by experienced personnel. Care is required at installation to assure intimate soil-ceramic contact and to pack soil around the tube above the cup to prevent preferential flow to the proximity of the ceramic during rainfall or irrigation. Tensiometers require daily reading and recharge of the water column. Their readings can be automated, but the daily manual servicing is still required, preferably with de-gassed water. The greatest limitation of tensiometers is their narrow operating range. This is limited (in theory) by simple physics to one atmosphere tension. However, water column cavitation from the vapor pressure of water and the evolution of dissolved gases under tension usually occurs at a practical limit of about 0.8 atmosphere of tension. Even when using de-gassed water to recharge the columns during daily servicing, the dissolved gases in soil water quickly diffuse into the water in the tensiometer's reservoir. With prolonged use, algae and fungal growth inside the tensiometer can reduce the sensitivity and porosity of the ceramic cup, as can illuviation of microscopic fines into the ceramic from soil water during prolonged use in clay soils. Nonetheless, tensiometers are one of the most commonly used soil water monitoring devices because of their simplicity and low cost (typically $50 per tensiometer).

Granular matrix sensors are, in essence, updated versions of electrical resistance (gypsum) blocks used to measure soil water matric potential [21,22]. Each sensor has two electrodes embedded within a small cylinder of a relatively stable, porous granular matrix that (a) is not affected by freezing temperatures and (b) dissolves but slowly in soil [12]. Before use, each sensor is separately calibrated to read soil water matric potential. To install, the sensor is buried in soil at some desired depth, with the electrode leads extending above the soil surface. With the leads of a sensor attached to a resistance (that is, alternating current or AC bridge) meter (or alternatively, a data logger), one reads the electrical resistance between the electrodes in the sensor, with the resistance being a non-linear function of the water in the matrix in potential energy equilibrium with the water in the surrounding soil. The measured resistance is most directly related to the soil water potential and is used to determine the soil's volumetric water content from a predetermined soil water characteristic (or water retention curve) relationship. Granular matrix sensors are often referred to as watermark sensors, the trade name of the most common type. The primary difference between granular matrix and gypsum block sensors is the tendency of resistance blocks to dissolve over extended time in soil, leading in some cases to measurement anomalies at the end of the sensor duty cycle.

Granular matrix sensors have advantages and disadvantages [12]. Granular matrix sensors are more easily and dependably calibrated than gypsum blocks, yield data that are more reproducible, and last longer in soil. They can be multiplexed and conveniently used to schedule irrigations. As disadvantages, matrix sensors require calibration, measure water content only in relatively wet soil, that is, from 0 to about −200 kPa (but somewhat drier soil than tensiometers), sense water content only at a single point, produce data representative of only a relatively small soil volume, and do not work well in coarse soil (sands) or in soils that shrink and swell.

Thermocouple psychrometers are the only sensors that measure total water potential. Psychrometers use the dewpoint principle. Each psychrometer sensor is a dime-sized hollow ceramic with a hair-thin bi-metallic wire junction sealed inside. The hollow ceramic is placed in the soil and is allowed to come to equilibrium. The relative humidity and temperature inside the ceramic eventually come to equilibrium with the soil atmosphere, at a value that is related to the total potential of the water source [23]. The bi-metal uses the Peltier effect to cool the junction, condensing water. The difference in temperature of the junction at equilibrium with the soil relative humidity *in situ* versus the dewpoint temperature for that water potential, is related to the relative humidity, which is directly related to water potential [24]. Psychrometers are less accurate in the wet range (approximately the range of tensiometers and gypsum blocks), but are well suited to and very accurate in the range of soil water content approaching air-dry water content (usually defined as the water content of soil in equilibrium with air at 25°C and 98 % relative humidity). They are suited to automated reading, and because they integrate the osmotic potential, are among the best choice for soil water characterization in high salinity situations. The instruments are, unfortunately, fragile and utilize a very small sensing volume. Electrode leads should be buried parallel to the sensor depth for 20 cm to 30 cm to prevent heat transfer from other depths to the sensor cup. Cups should be thoroughly cleaned between insertions to prevent carryover of salts from one use to the next. Meters that read psychrometer output require very high precision thermo-electric circuitry for accurate water potential

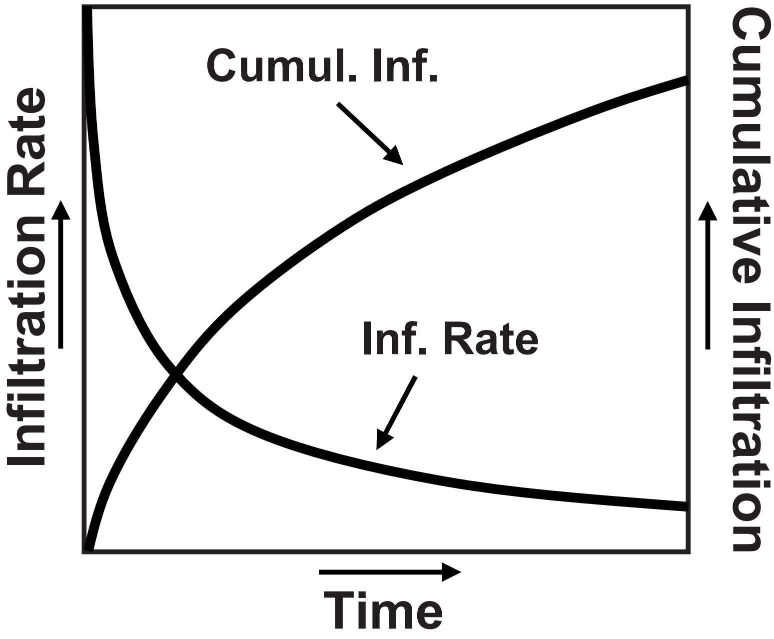


FIG. 5—*Infiltration rate and cumulative infiltration as a function of time.*

readings, and it is strongly recommended that individual calibration curves be developed for each psychrometer sensor.

Soil Water Movement

Water movement is another important attribute of soil water. Because water movement in soil is quite slow, its kinetic energy is considered to be negligible. Soil water flow properties are usually characterized in terms of water intake properties (infiltration), or as conductivity. Both infiltration and conductivity are a function of soil water potential and are characterized at specified water potentials. The soil water's potential energy, which is mainly due to gravity and surface interactions with the solid phase, is of primary importance in determining the state and movement of water in soil [1,2]. In highly saline soils, the osmotic potential can be significant in determining total water potential, especially in the very dry range. The direction of water movement in soil is governed by the net soil water potential difference between observed positions. The gradient causing soil water flow in the system is equal to the difference in total potential divided by the distance of the flow path. Under saturated conditions, downward flow of water is primarily the result of the gravitational potential and the depth, if any, of water ponded above the soil surface. However, until the soil pores are filled with water, most soil water flow in the field occurs while the soil is unsaturated. Even when applying water via flooding, some of the soil profile is unsaturated until the pores become filled with water. Thus, the gradient driving flow is actually the integral of water potential difference across the flow path of interest.

Surface sealing can occur rapidly when water is introduced into irrigation furrows as a result of the instantaneous hydration of hot dry soil, which explodes aggregates and disperses fines in the furrow stream. Segeren and Trout [25], using micro-tensiometers buried just below the seal-forming zone, showed that this phenomenon quickly lowers the infiltration rate by reducing the hydraulic conductivity of the near-surface soil.

Infiltration

Infiltration into an initially dry soil changes markedly with time (Fig. 5). At first, the infiltration rate is relatively high but, as time progresses, the rate drops due to a decrease in the potential gradient at the soil surface that drives the infiltration process. The infiltration rate continually decreases at a decreasing rate, approaching but never quite reaching the soil's saturated hydraulic conductivity. Cumulative infiltration increases quickly at early times but slows as time passes due to the decreasing infiltration rate.

TABLE 1—*Relationship between water potential and maximum pore diameter through which flow occurs.*

Water supply potential		Flow occurs through pores with diameters less than or equal to
mm	kPa	
+5	+0.05	(all pores)
−30	−0.30	1.0 mm
−60	−0.60	0.5 mm

Because of the initially strong dependence of infiltration upon time (Fig. 5), infiltration rate is usually characterized as the steady-state ponded infiltration rate (with a known and constant depth of water ponded on the soil surface), or as the steady-state unsaturated infiltration rate at a specified matric potential [26–28]. In either case, infiltration rate is measured at a location that is "preconditioned" by an application of water (as free water in the case of steady-state ponded infiltration, or as water under tension, for steady-state unsaturated infiltration) at the desired water potential until the potential gradient within the measurement volume stabilizes. Tension infiltrometers are routinely used to evaluate the near-surface hydraulic conductivity as influenced by soil tillage, irrigation amount, water quality, surface slope, and surface sealing [29].

Tension and ponded infiltrometers are used to measure unconfined, three-dimensional, steady-state infiltration rates that are used to calculate unsaturated hydraulic conductivities and other structure- and texture-related parameters affecting water flow [30–32]. A tension infiltrometer releases water only when the soil's attraction for the water overcomes the force with which the infiltrometer holds water — the device's water supply potential. Tension infiltrometers supply water at a slightly negative supply potential (e.g., from −20 mm to −150 mm of water). These negative potentials eliminate flow through large soil pores but permit it through smaller pores in the soil matrix (Table 1). A ponded infiltrometer, in contrast, maintains a constant shallow head of water (typically 5 mm in depth) on the soil surface. Using a ponded infiltrometer at +5 mm water supply potential, one can estimate field-saturated hydraulic conductivity K_{fs}. The difference between K_{fs} and unsaturated hydraulic conductivity at −30 mm or −60 mm potential is the macropore conductivity. Tension and ponded infiltrometers are used to characterize the flow properties of macropores, temporal changes in pore geometry, sorptivity (the capacity of a porous medium to absorb liquid by capillary action), and soil structural changes induced by tillage, irrigation, precipitation, waste applications, and biological activity [29,31,33]. Tension and ponded infiltrometers are also used to characterize the hydraulic conductivity-pressure head relationship near saturation needed to predict contaminant transport.

Determining infiltration rate on a field basis or a landscape basis is problematic because it is generally impossible to have a uniform potential gradient over the entire observed surface area during the period of observation. In large-scale field settings, net infiltration amount is often characterized as the difference between net inflow and net outflow per unit surface area. Thus, in practical applied settings, net infiltration can be used to characterize field effects of a management practice or treatment, but insights gained from infiltration rate can only be arrived at through multiple spatially distributed and controlled point measurements of steady-state ponded infiltration rate or steady-state unsaturated infiltration rate. Although additional information can be gained from the length of time until attainment of steady-state infiltration rate, this measurement is still dependent upon point-scale measurements. When assessing treatment effects through measurements of net infiltration on a field basis, it is very important to minimize all potential sources of systematic error among treatment sites (e.g., water application spatial and temporal uniformity, slope, surface roughness, compaction, etc.).

Site conditions significantly affect infiltration. These are conditions separate from the soil profile per se. In rain-fed, sprinkler-irrigated, and furrow- or basin-irrigated situations, greater plant residue or duff, warmer water temperature, and surface roughness, will generally increase the infiltration rate in a field. These increases occur by protecting the abundance and continuity of macropores from the soil surface to the subsurface, increasing infiltration opportunity time by slowing the runoff stream, and reducing the extent of particle dispersion associated with any runoff. The latter reduces the likelihood of transport of mineral fines that could plug soil pores during infiltration, causing surface sealing. Warmer water temperature reduces water viscosity, which increases hydraulic conductivity and thus infiltration. Conversely, steeper slope, compaction, surface sealing, interrupted flow, a recent history of vegetation burning at relatively high temperatures at the soil surface, or water with very low electrical conductivity (EC) and/or very high sodium adsorption ratio (SAR) will reduce infiltration. It should be emphasized that, in field settings, the land surface is variable. When assessing water-driven phenomena, such as infiltration, one must keep in mind that measurements made on a point scale may not represent the integrated outcome of wide-scale variability. Sometimes, less precise field-scale tests provide more accurate overall assessment of a situation than a few spot tests with high precision, but which are not spatially representative.

In rain-fed and sprinkler-irrigated situations, the infiltration and runoff outcome is also affected by water drop size, energy, and intensity. The applied water, depending upon its drop size, energy, and intensity, can disrupt surface soil aggregates, disperse fines, and induce erosion and seal formation [31,34]. High energy rainfall events, characteristically composed of larger drops and higher rainfall rates, may result in more net infiltration because of more water applied, but may create surface seals that reduce the saturated infiltration *rate* and lead to a greater tendency for runoff, including in subsequent events. In the case of sprinkler irrigation, the capacity for the drops to break surface aggregates, destroy soil structure, and disperse fines is exacerbated by lower EC and/or higher SAR water. It is a common practice in many parts of the world to add gypsum to irrigation water to increase EC and lower SAR in order to reduce the erosivity of the water and to keep infiltration rates relatively high.

Essentially the same water quality and surface condition considerations apply to infiltration in furrow irrigation or other forms of surface irrigation, except that they do not have the kinetic energy component due to droplet impact or splash to exacerbate soil aggregate destruction, particle dispersion, transport, and deposition. Thus, soil surface conditions affect infiltration from surface irrigation primarily through their action upon and interaction with stream flow and shear.

Furrow inflow/outflow measurements (that allow one to estimate net field infiltration) and furrow stream advance time are common measures of soil hydraulic properties on a field-integrated basis. The techniques, accuracy, and precision, of these approaches were described by Trout and Mackey [35–37]. Blair et al. [38] also described the use of recirculating furrow infiltrometers, to better assess hydraulic phenomena *in situ* on segments of furrows. The use of small, calibrated single-furrow flumes to monitor hydraulic processes [39,40] and Imhoff cones to assess sediment losses have also been described in detail [41,42].

From a practical standpoint, runoff prevention is an important water management strategy. In rain-fed situations, it is aimed at capturing precipitation and storing it in the soil profile for crop use. The vast majority of arable and forest land in the world receive seasonally averaged precipitation below the optimal level for crop or tree production. Profile flooding is a rare concern in these areas, and negative crop or forest growth effects rarely result unless the full profile is saturated for periods longer than 48 h. Ecosystems prone to frequent flooding for longer than 48 h tend to be dominated by flood tolerant species. Managed lands in these areas usually produce flood-tolerant crops, such as rice, or are engineered to facilitate internal drainage. Where water capture and runoff prevention are the main concerns, the single most important consideration is to keep water application rate below the steady-state saturated infiltration rate, which is often slightly less than the saturated hydraulic conductivity (K_{sat}). Under irrigated conditions, this can be managed via matching the irrigation application rate to the soil properties. In rain-fed conditions, one minimizes runoff and maximizes infiltration by utilizing soil management practices, such as field leveling, subsoiling (deep ripping), contour farming, use of berms and vegetative filter strips, and preservation of crop residues on the soil surface (e.g., via no-till farming). Under furrow irrigation, there is sometimes the need to reduce infiltration preferentially in the upper portion of a field, to improve the infiltration uniformity between the upper reaches (which have longer infiltration opportunity time) and the lower reaches (which have shorter infiltration opportunity time and smaller average irrigation streams, due to upstream water intake). Interrupted furrow irrigation (sometimes called surge irrigation) is often used to intentionally create a surface seal at the upper end of the field, reducing its average infiltration rate compared to that of the lower reaches of the field [43].

Water drop penetration time [44] is often measured in soils that exhibit hydrophobic properties. Water drop penetration time (WDPT) is another point-scale measurement. WDPT relates qualitatively to infiltration but it cannot be easily related quantitatively to measurements of infiltration rate and net infiltration. This is because of temporal variability, antecedent moisture and humidity effects, and even very small soil disturbance effects, coupled with extreme three-dimensional spatial heterogeneity [45–48]. While in a general sense high WDPT indicates poor infiltration, the high sensitivity of the phenomenon to random and

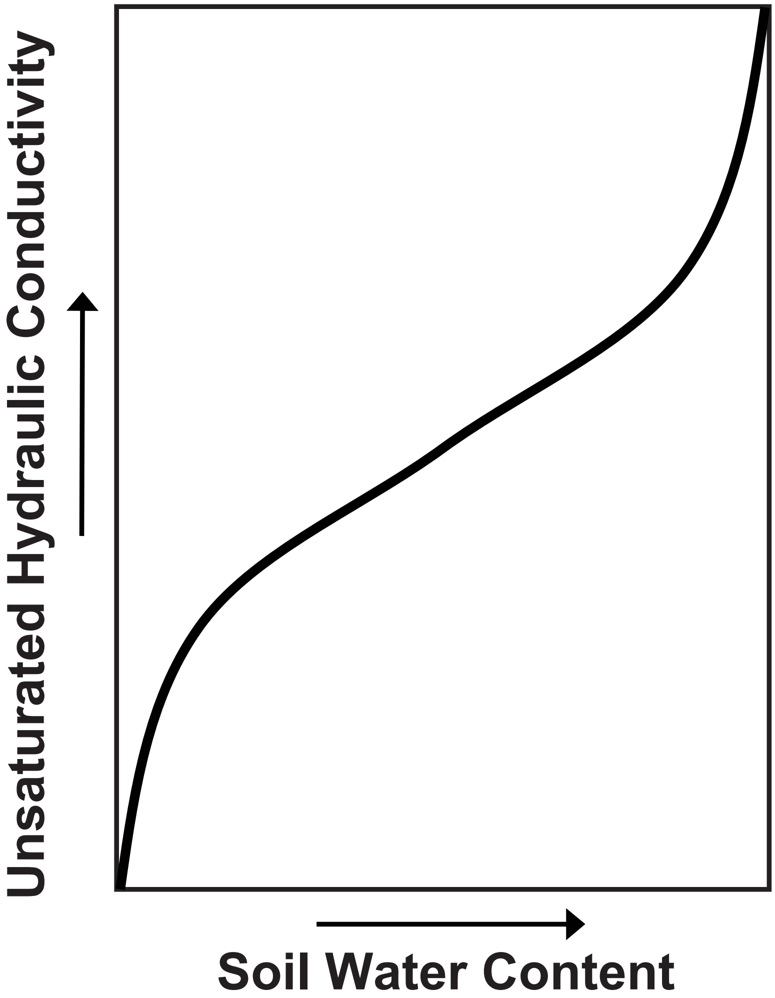


FIG. 6—*Hydraulic conductivity as affected by soil water content. In soil with low water contents, conductivity is minimal because water flows via tortuous pathways through only thin water films surrounding soil solid particles. In contrast, at high water contents, water flows through saturated or nearly saturated pores, greatly increasing hydraulic conductivity.*

often very small environmental variables makes correlations of the two measurements quantitatively inconsistent from one characterization to the next, with interpretations being highly dependent upon the observational scale — e.g., point-scale versus watershed-scale [49,50]. Hence, WDPT is an unreliable, quantitative surrogate for measurement of infiltration rate or amount.

Hydraulic Conductivity

Hydraulic conductivity is a measure of soil water transmission rate and is dependent upon the soil's water content (Fig. 6). Water in unsaturated soil is subject to subatmospheric pressure (sometimes described as tension or matric suction), which can be expressed mathematically as negative pressure potential. Although most soil-water interactions occur in unsaturated soil, the hydraulic conductivity of saturated soil (K_{sat}) is, nonetheless, of great importance for both agricultural production and environmental protection. Saturated hydraulic conductivity controls the downward movement of free water through soil. When a field soil is saturated, all or nearly all of the pores are water-filled and, if open at both ends, conduct water. Flow thus occurs through a relatively large area and consequently conductivity is maximal (Fig. 6). Water flow in unsaturated soil, in contrast, is through water films that surround all soil particles, or as tube flow through water-filled pores with very small diameters.

The hydraulic conductivity of soils depends greatly on texture and structure [51]. At saturation, the most conductive soils are those in which large and continuous pores constitute most of the overall pore volume, whereas the least conductive soils have a pore volume that mostly consists of micropores, being pores with diameters <0.5 mm (Table 1). Thus, at saturation, a sandy soil conducts water more rapidly than a clayey soil, and a well-aggregated soil conducts water more rapidly than a poorly aggregated or dispersed soil. In contrast, soils with high hydraulic conductivity when saturated often have low hydraulic conductivity when unsaturated. In a soil with large pores, those pores empty quickly and become nonconductive as water potential drops (becomes more negative, i.e., as suction develops). In a soil with small pores, many of the pores retain and conduct water even at appreciably low water potentials (i.e., more negative potential, greater suction, drier), so hydraulic conductivity does not decrease as steeply and may actually exceed that of a soil with large pores subjected to the same suction. The presence of viscosity-changing or surface tension-changing solutes or surfactants in the soil solution can alter the natural liquid interfacial properties, as soil hydraulic properties are directly related to surface tension and viscosity [52].

Water moves within the soil profile mostly as a liquid. The equation describing one-dimensional liquid flux can be written as:

$$J_w = -K\frac{d\Psi}{dx} \tag{2}$$

where J_w is the water flux density ($kg \cdot m^{-2} \cdot s^{-1}$), K is the unsaturated hydraulic conductivity ($kg^2 \cdot m^{-1} \cdot s^{-1} \cdot J^{-1}$), Ψ is the total water potential ($J \cdot kg^{-1}$), and x is distance (m). The water's potential gradient, $d\Psi/dx$, is the driving force for water movement in the liquid phase.

Field Water Balance

The interactions between rainfall, irrigation, soil water, runoff, leaching, plant ecology, plant physiology, canopy architecture, transpiration, and evaporation are important in maintaining ecosystem processes. Evapotranspiration is the sum of evaporation and transpiration. In the presence of plants, it is often difficult to distinguish between transpiration and evaporation. In the absence of plants, evaporation of water from the soil surface is dominated by radiation and wind effects when the surface is wet, and by soil properties limiting water vapor transfer to the surface when the soil surface is dry. A plant's evapotranspiration rate depends on the soil water content, soil and air temperature, net radiation, canopy architecture and air movement (wind). Estimates of evapotranspiration can be made using the Penman method [2,20]. Wright [53] describes how weighing lysimeters are used to develop crop coefficients for the Penman equation for use in estimating evapotranspiration from irrigated crops.

Crop water-use efficiency (WUE) can be defined in several ways. Perhaps two of the most common are: the dry matter yield divided by transpiration or, in irrigated agriculture, as harvested yield divided by irrigation water supplied [54]. Overall agronomic water-use efficiency, WUE_{ag}, is:

$$WUE_{ag} = \frac{P}{W} \tag{3}$$

where P is crop production and W is the volume of water applied. The water applied is given by:

$$W = RO + DP + E_d + E_s + T_w + T_c \tag{4}$$

where RO is the volume of water lost by runoff from the field, DP is the volume of water drained below the root zone (deep percolation), E_d is the water volume lost by evaporation during delivery and application to the field, E_s is the volume of water evaporated from the soil, T_w is the water volume transpired by weeds, and T_c is the volume of water transpired by crops.

The water applied can be minimized by avoiding losses due to runoff, drainage, evaporation, and transpiration by weeds. Production or growth can be maximized by using high-yield varieties that are well-adapted to the soil and climate. Properly timing planting, harvesting, tillage, and fertilization operations, along with pest control applications, also maximize growth, thereby increasing WUE.

Solute Transport and Miscible Displacement

Solute transport occurs when water containing a different solute content than the soil solution passes through a body of soil, gradually changing the soil solution's composition [21,55]. When the solutions mix freely with one another, such solute transport is often called miscible displacement. Co-movement of water and non-soluble liquids (e.g., oil) is often termed immiscible displacement. For information on immiscible transport phenomena (not discussed hereafter), the reader should consult texts in chemical engineering, petroleum engineering, or oil and gas exploration and recovery.

Solute transport occurs via convective transport (due to mass flow; passive movement in flowing water), diffusive transport (due to concentration differences), and dispersive transport (due to differences in fluid velocities: slower near the pore wall but faster away from the pore wall) [56,57]. The combination of molecular diffusion and mechanical dispersion is often termed hydrodynamic dispersion [58]. In addition, solute transport may be affected by chemical reactions or exchange as the solute moves through soil and by adsorption of the solute onto either inorganic or organic soil constituents [21].

Fluid moves faster through flow paths with larger cross sections. Consequently, an incoming fluid moves faster in saturated soil through the larger pores and, in unsaturated soil, flow decreases as water films become thinner with drying [21]. The faster fluid moves, the faster displacement and solute transport occur. Without significant water movement, solutes tend to stay where they are [57].

Solute movement is a function of convection and the combination of diffusion and dispersion, often simply called dispersion. Since convection is the main driving force for solute field movement, accurately predicting water movement in either one or two dimensions is very important [57]. Field water flow is either steady or transient, but usually transient. Transient flow occurs when either water contents (θ's) or potentials (ψ's) change with time. Solute flux, mass or volume moving through a unit cross-sectional area per unit time, is a combination of convective flux (requiring the water flux, usually calculated using Darcy's Law; hence, Darcy flux) and dispersive flux. Convective flux is easily determined once water flux is known. Water flux is a function of both time and position [58]. Solute flux is a function of the magnitude and direction of water flux [58].

A key hydraulic parameter for estimating solute transport is pore water velocity ν, defined as

$$\nu = \frac{J_w}{\theta} \tag{5}$$

where J_w is the water flux density defined in Eq 2 and θ is the volumetric water content ($m^3 \cdot m^{-3}$). Solute transport is dominated by convective transport when ν is high but by dispersive transport when ν is low [57]. To mathematically describe solute movement, one must solve coupled flow and transport problems, commonly by using the convective-dispersive equation (CDE), also called the advective-dispersive equation (ADE) [56,58]. Solute transport models employ a numerical solution of the Richard's equation in one, two, or three dimensions to estimate water flux J_w, which is, in turn, coupled with the CDE [57]. The CDE does not require steady water flow or uniform water content but does require input of the flowing soil-solution's solute concentration [56,58].

To use the CDE, one must first know the Darcy flux (the volume of water flowing through a unit cross-sectional area per unit time), being constant with steady flow but varying with changing θ's or ψ's; that is, transient flow [57]. For the latter case, one first solves the Richard's equation for pore water velocity and water content, then continues by solving the transport problem [58].

Plots of relative solute concentration C/C_0 as a function of effluent volume are referred to as breakthrough curves [21]. With miscible solutions, an incoming fluid with a solute concentration C_0 will yield an effluent with a concentration of C/C_0 at time t. Effluent volume is often given in units of pore volumes (the volume of water stored in a soil column) [58]. Only in a specific volume of saturated soil is one pore volume equal to the sample's pore space [56].

Breakthrough curves provide the data that, when fitted via an inverse estimation procedure to a transport model such as the convective-dispersive equation or CDE, yield numerical estimates of three solute velocity and dispersion parameters: the solute pore water velocity (ν), the dispersion coefficient (D), and the retardation factor (R) [56]. The dispersion coefficient is a measure of the spreading of the solute pulse as it moves through the soil, but not a measure of the location of the solute peak [59]. Spreading of the solute pulse is caused primarily by differences in the pore water velocity and secondarily by dispersion [59]. The dispersion coefficient is commonly a fitted parameter, whereas the retardation factor is usually set equal to 1 [56]. Solute transport parameters are determined from solute displacement experiments in laboratory soil columns during steady-state flow using analytical solutions of the CDE [58].

Often a solute's peak concentration decreases and its movement slows as the pulse of solute-containing solution passes through a soil profile. Soils are more similar in the horizontal than vertical direction. In some situations, solute movement through entire soil profiles with quite distinct horizons may resemble movement through more uniform soil profiles [57]. Under steady but unsaturated flow, the one-dimensional convective-dispersive equation at times models solute transport better through layered rather than uniform soil profiles [60].

Management practices can increase or decrease solute transport. By applying a fertilizer that contains a mobile solute, then waiting a day or two before applying water, solutes can diffuse into the soil solution in small intra-aggregate pores where they are less available for transport by preferential flow [61]. Delaying sprinkling after applying nitrogen fertilizer is a best management practice (BMP) to minimize nitrate leaching [62]. The soil solution in small pores resides in what has been termed an immobile (or stagnant, non-flowing) region, whereas the soil solution in larger pores resides in a mobile (or flowing) region [58].

Solutes in "relatively" immobile regions in small pores, however, can still be transported. When preferential flow occurs, some solute can slowly diffuse out of the relatively small pores of soil aggregates into larger pores or preferential flow paths, contributing to the solute transported by bypass flow.

The Gaseous Phase

In soil, the exchange of gases such as oxygen and carbon dioxide is vital to root growth, water and mineral exchange, and respiration [63]. Adequate root respiration requires that there be sufficient, continuous air-filled pores to enable diffusion of gases between soil and the ambient atmosphere in the gaseous phase, and sufficiently thin water films to allow diffusion of dissolved oxygen and carbon dioxide between the soil atmosphere and active root respiration sites. This diffusion must occur at a rate that will sustain respiration and aerobic metabolism as well as prevent accumulation of excess carbon dioxide in the root zone. Microorganisms in the soil also respire and, in conditions of restricted aeration (wet soil), they compete with the roots of higher plants for scarce oxygen.

Gases in the soil can move in either the air phase or in dissolved form through the water phase. Gas transfer is primarily via the process of diffusion. The rate of oxygen diffusion in the gas phase is approximately 10 000 times greater than in the water phase. Thus, soil aeration is primarily dependent upon the volume fraction of air-filled pores, which in turn depends primarily on soil water content (Fig. 2). The minimum or critical air-filled porosity of soils to prevent oxygen stress is typically about 10 %. Optimal air-filled porosity is typically about 25 %. The optimal soil water content for supplying water to prevent stress while still allowing good soil aeration is generally regarded as the soil water content at "field capacity." Field capacity water content is an inexact term, variously defined as the water content in medium-textured soils at $-1/3$ bar (-3.33 MPa) water potential ($-1/10$ bar in sands) or the water content of recently flooded soil after 24 h to 48 h of free drainage. The optimal volume of air-filled pore space maintained in soil is a balance between supplying sufficient water to prevent plant water stress, while assuring sufficient aeration to prevent plant oxygen stress. Since plants use oxygen in a supply-rate-dependent fashion, the aeration measurement best (most quantifiably) correlated with plant physiological and yield performance, is generally regarded as the soil oxygen diffusion rate (ODR) measured using the platinum microelectrode technique or similar approaches [64–66].

Strongly aggregated soils, with macroaggregates of the order of 0.25 mm or more in diameter, generally have a considerable volume of macroscopic pores that drain very quickly and remain air-filled most of the time. In contrast, in relatively unstable soils when the clay fraction is dispersed or aggregates are broken down, many macropores become filled or partially filled with primary particles or aggregate fragments. Thus, severely compacted soil may contain less than 5 % air by volume at its characteristic field capacity soil water content.

Hypoxic and anaerobic conditions in the soil induce a series of chemical and biochemical reduction reactions. Included in these reactions are denitrification, and the reduction of manganese, iron, and sulfates. Soil organisms obtain the energy they need for their vital processes by means of a series of chemical reactions involving the transfer of electrons from substances that serve as sources of energy to substances that may become products of respiration. In aerobic respiration, the final receptor of electrons is oxygen, which combines with hydrogen to form water. The tendency of a solution to donate electrons to a reducible substance (oxidation) or to accept electrons from an oxidizable substance (reduction) can be measured in terms of its oxidation-reduction potential or redox potential. Measurement methods and data interpretation for redox potential, soil oxygen content, and soil ODR are described in detail in Glinski and Stepniewski [67].

Summary

Assessing soil water itself, or soil-water-dependent phenomena in soil can be a complicated and highly nuanced process. Users of these techniques and of the measurement results need to be aware of the techniques' limitations and the various factors that can impinge upon the interpretations of the results [68]. Perhaps the strongest caveat is the need to be cognizant of the transferability of point-scale measurements made in the laboratory or field, to interpretation on a landscape basis, where issues of scale and variability can at times outweigh the measured point-scale treatment effects. This summary should in no way be regarded as sufficiently authoritative to preclude further investigation of the literature that relates to the individual circumstances and measurement protocols an investigator is contemplating. Hopefully, we have helped point the reader toward key literature that can provide a sound basis for experimental design and choice of assessment protocols.

Appendix: Websites of interest with additional information

http://edis.ifas.ufl.edu/AE266 (Verified October 1, 2007)
http://srwqis.tamu.edu/downloads/Field%20devices%20for%20monitoring%20soil%20water%20content.pdf (Verified October 1, 2007)
http://www.cprl.ars.usda.gov/Soil%20Water%20Papers_Presentations/saas2001/sld001.html (Verified October 1, 2007)

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