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Review

Soil water sensing for water balance, ET and WUE☆

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ABSTRACT

The soil water balance can be solved for evapotranspiration (ET) using soil water change in storage data from either weighing lysimetry or soil water sensing and measurement, along with data on the other components of the water balance. Weighing lysimeters are expensive and, although accurate, are difficult to manage and afford little replication. Direct soil water measurement by coring is accurate enough, but plagued by spatial variability that induces unwanted variability in the change in soil water storage between dates, and is destructive and time/labor consuming. Here we focus on soil water sensing using the neutron probe and various electromagnetic (EM) sensors (capacitance, time domain reflectometry (TDR) and quasi-TDR) with respect to the relative levels of uncertainty in profile water content, change in soil water storage, and estimates of deep flux; and their impact on estimated ET and water use efficiency (WUE). Studies consistently showed errors up to and >0.05 $m^3\ m^{-3}$ for capacitance sensors used in access tubes, which implied errors in soil water flux estimation of up to 50 mm day^{-1}, and calibrations that were so sensitive to soil bulk electrical conductivity (σ_{dc}) and temperature that water content and change in storage estimates were rendered unreliable. Also, larger spatial variability of water contents reported by capacitance sensors was tied to the EM field penetration in structured soils around access tubes being non-uniform and influenced by the random arrangement of soil micro-scale water content, σ_{dc} and bulk density distribution. Thus, we recommend that profiling sensor systems based on capacitance technology not be used for studies of water balance, ET and WUE, nor for irrigation scheduling. Recommended methods include the neutron probe, direct volumetric soil sampling and, in some cases, conventional time domain reflectometry with waveform capture and analysis. New sensor development efforts should focus on waveguide approaches using TDR technology.

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1. Evapotranspiration and crop water productivity (CWP)

Water use efficiency is also known as crop water productivity (CWP), often defined as the economic yield (Y) divided by the crop water use or ET:CWP = Y/ET. While economic yield at the field or plot scale is readily measured at harvest, the field or plot scale ET is a season-long function that requires accurate determination of the change in soil water storage over the season (ΔS), irrigation (I), and precipitation (P), as well as the control or determination of the sum of run on and runoff (R), and any deep soil water flux (F):

$$\mathrm{ET} = -\Delta S + P + I - R + F + \varepsilon_{\mathrm{ET}} \quad (1)$$

The error in ET, $\varepsilon_{\mathrm{ET}}$, is a function of the errors in determination of ΔS, I, P, R, and F. Values of P and I may be adequately measured with commercial rain gages, when appropriately sited and replicated, and with meters or weirs, respectively. Runoff and run on may be controlled to near zero amounts by berming and diking, including furrow diking. There are several ways to control error in soil water fluxes (F): (a) arranging plots so that lateral fluxes are as small as possible (adequate plot width and length, avoidance of high slopes, (b) including buffer areas between plots receiving different irrigation amounts), (c) avoiding shallow water tables that could provide water to plants by upward flux, and (d) sensing or measuring the soil water content deeply (e.g., 3 m depth or in any case well below the root zone) so that the bottom measurements are in drier soil and so that irrigation and precipitation penetration and root water uptake processes all occur well above the deepest soil water measurements.

Measuring water content to well below the root zone is critically important to obtaining accurate ET values by soil water balance. Many crops can take up appreciable water from soil to depths much greater than can be easily measured using the newer electromagnetic (EM) soil water sensors that work in plastic access tubes. For example, Phene et al. (1991) measured uptake to depths >2 m for corn (*Zea mays* L.) roots in the Panoche soil at the West Side Field Station (drip irrigation). Musick et al. (1994) measured winter wheat (*Triticum aestivum* L.) water uptake to 2.4 m in a Pullman clay loam. Cai et al. (2005) measured winter wheat water uptake in a deep silt loam soil in the southern Chinese loess plateau to 2.5 m. Grimes et al. (1975) measured uptake to 1.83 m for cotton (*Gossypium hirsutum* L.) rooting in the Panoche soil at the West Side Field Station. And uptake to 2.25 m in the same soil was recorded for sugarbeet (*Beta vulgaris* L. subsp. *vulgaris*) (Howell et al., 1987), while uptake to 3.0 m was recorded for sugarbeet on the Pullman clay loam at Bushland, TX (Winters, 1980). Assuming that we have to measure to 3-m depth to close the soil water balance, an overall bias error of only 0.02 $\mathrm{m^3\,m^{-3}}$ would cause an error of 60 mm in the profile water content.

Other than the soil water balance, many other methods of determining ET exist. The most important of these are discussed in detail by Evett et al. (2012) and they include eddy covariance (EC), the Bowen ratio method (BR), other aerodynamic methods, and the estimation of ET by multiplication of a crop coefficient by a reference ET (e.g., the FAO or ASCE Penman–Monteith reference ET methods). Crop coefficients are most commonly determined originally from ET measured using the soil water balance, either via weighing lysimetry or soil water measurement or sensing. Both the eddy covariance and Bowen ratio methods involve errors that render them unsuitable for season-long determination of ET. These errors include common daily energy balance closure errors of 10–30% for EC and singularity errors with Bowen ratio near sunset when ET rates may still be important in semi-arid and arid regions, both of which impede long-term integration of ET data to determine seasonal ET accurately. For these reasons, this paper focuses on the soil water balance, specifically as determined using soil water sensing, as the most common and potentially accurate method of determining season-long crop ET.

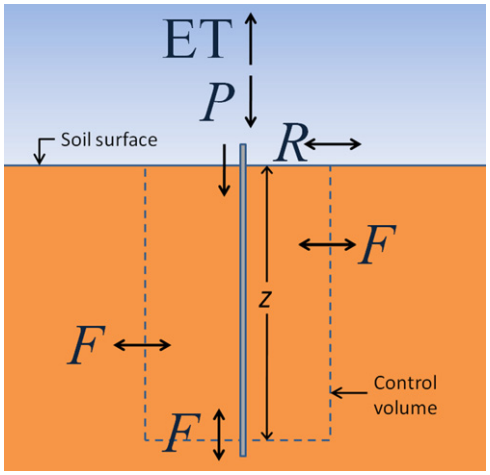

Fig. 1. Schematic of the soil water balance for a control volume of depth z. The change in soil water storage over the depth z, needed to calculate ET for a period of time using Eq. (1), could be calculated using a soil water sensor calibration equation that converts the measured value (C) of a surrogate property to water content in $\mathrm{m^3\,m^{-3}}$, for example, Eq. (2).

1.1. Soil water balance

The soil water balance Eq. (1) is written for a control volume of given surface area and depth z within which water storage changes due to fluxes (F) of water to the atmosphere as evaporation and plant transpiration (ET) and lateral or vertical fluxes into or out of the control volume (Fig. 1). The fluxes and change in storage may be considered as instantaneous rates, or as integral values over some period of time, e.g., the cropping season. Measurements of soil water content throughout the depth of the control volume, at the beginning and end of an arbitrary time period, allow the computation of the change in soil water storage (ΔS_z) for that period. For example, calculating ΔS_z for the control volume illustrated in Fig. 1 in terms of a generalized calibration equation for a soil water sensor, we have

$$\Delta S_z = z(a + bC_i^n) - z(a + bC_f^n) = zb(C_i^n - C_f^n) \quad (2)$$

As with all soil water sensors, the sensor does not measure water content, but measures some property related to water content, shown here as a measured value, C, with the subscript i indicating the value at the beginning of the period and the subscript f denoting the ending value. The term a is the intercept coefficient, the term b is the slope coefficient, and the term n is the exponent of the calibration for water content, θ ($\mathrm{m^3\,m^{-3}}$), where $\theta = a + b(C)^n$ is the calibration equation. Multiplying the depth of the control volume by the difference in water contents over the time period gives ΔS_z for this simplified case where the water content is assumed uniform throughout the control volume at any time. More realistically, water content will vary with depth and ΔS would be calculated for each layer in which water content was measured and ΔS_z calculated as the sum of these values. Eq. (2) demonstrates that the only coefficient of the calibration equation that does not affect the calculation of ΔS_z is the intercept term a. Errors in any other coefficient or exponent (e.g., b and n in this case) will cause errors in ΔS_z.

All soil water sensor calibrations include at least one factorial coefficient and most include at least one exponent. If a soil water sensor is providing inaccurate water content data and if the inaccuracy is not a simple offset bias, then it follows that errors in the factorial coefficient(s) and/or exponent(s) are involved. The

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Table 1
Errors in estimated flux rate (mm d^{-1}) due to errors in water content (θ) of 0.01 and 0.05 m^3 m^{-3} for three soil textures at both permanent wilting point (PWP) and field capacity (FC) water contents.

Texture	Flux rate error (mm d^{-1})			
	For θ error = 0.01 m^3 m^{-3}		For θ error = 0.05 m^3 m^{-3}	
	PWP	FC	PWP	FC
Sandy loam	9	11	43	57
Silt loam	0.01	9	0.6	50
Clay	0.4	4	1.5	21

important interferences on EM soil water sensor calibrations involve the effects of soil properties on the imaginary component of the complex permittivity, which includes the effect of mineral surfaces on dipole relaxation (bound water) and the effect of the loss tangent, σ_{dc}/ω, where ω is the angular frequency of the EM field oscillation. Both the fraction of bound water and the value of σ_{dc} are positively related to soil temperature. The fraction of bound water increases with clay content since it is related to soil surface area. The σ_{dc} is also related to clay content and type and increases with water content and with, of course, soil soluble salt content. Water content, temperature, clay content and type, bound water fraction, soluble salt content and σ_{dc} all vary with depth in many, and perhaps most, soils. It follows that the interferences on EM sensor calibration equation coefficient(s) and exponent(s) will vary with depth leading to the inevitable conclusion that errors in ΔS will vary with depth as may the sign of ΔS errors.

1.2. Deep soil water flux

In situations where the soil water content at the bottom of the control volume is large enough that hydraulic conductivities and flux rates are non-negligible, errors in computed F resulting from errors in sensed water content can be important. For example, we calculated for three soil textures the errors in tension resulting from water content errors of 0.01 and 0.05 m^3 m^{-3} using a Van Genuchten (1980) type water release curve (e.g. for silt loam in Fig. 2). Using the related function of soil hydraulic conductivity as a function of water content, $K(\theta)$, due to Mualem (1976), we calculated the errors in soil water flux rate induced by those same water content errors (Table 1). Errors in flux rate increased with water content, ranging from 4- to 57-mm d^{-1} at field capacity for water content errors of 0.01 and 0.05 m^3 m^{-3}, respectively, for the three textures (Table 1). Errors in flux rate were larger for lighter soil textures. This problem is compounded by the fact that errors in the water content reported by EM sensors tend to also increase with water content (Evett et al., 2006).

2. MAD irrigation scheduling

Irrigation scheduling is also affected by accuracy of soil water sensing. Most soils drain to field capacity quickly enough that irrigation management is, for practical purposes, constrained to working with water contents between field capacity (FC) and permanent wilting point (PWP) values, i.e. the range of plant available water (PAW). Since allowing the soil to dry to PWP causes irreversible plant damage, irrigation scheduling typically works with a management allowed depletion (MAD) range, which is taken as a fraction of PAW. For a MAD fraction of 0.6, this range varies from 0.02 to 0.13 m^3 m^{-3} depending on the soil type (Table 2 and Fig. 3), which can be smaller than the water content errors associated with many soil water sensors. Also important is to note that the water content range between MAD and PWP can be even smaller (Table 2), so that relatively small errors in sensed water content can cause

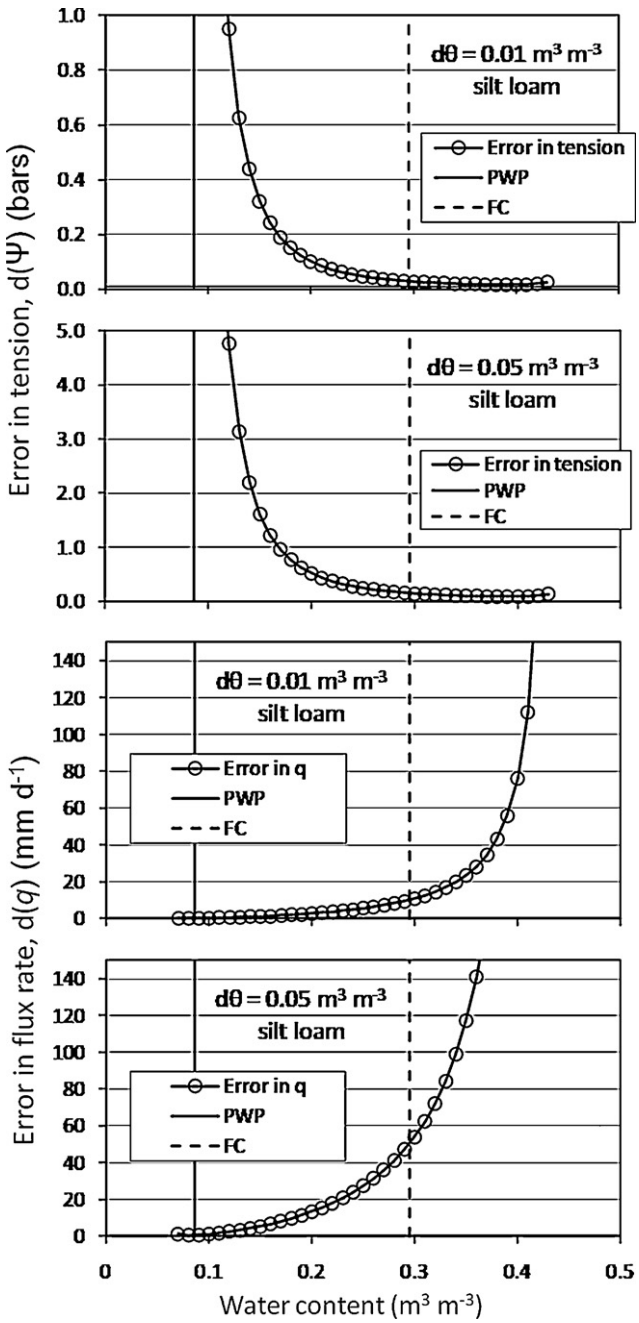

Fig. 2. Effects of water content errors (dθ) of 0.01 and 0.05 m^3 m^{-3} on soil water tensions (top) and flux rates (bottom) for a silt loam soil. Soil water tension and hydraulic conductivity were estimated using the Rosetta pedo-transfer function software (Schaap, 2002). The permanent wilting point (PWP) and field capacity (FC) water contents are indicated by vertical lines.

irrigation to be delayed to the extent that the PWP is unintentionally reached. This problem becomes particularly important when attempting regulated deficit irrigation in order to boost CWP since the MAD fractions associated with regulated deficit irrigation may exceed 0.6.

Exacerbating this problem is the nonlinear relationship between soil water content and soil water potential. As the soil dries, the dependence of soil water potential on water content becomes greater such that small errors in water content can cause large errors in estimated soil water potential. As illustrated in Fig. 2, large errors in soil water potential can occur at water contents greater than permanent wilting point and within the range that might be

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Table 2
Example calculation of management allowed depletion (MAD, $m^3\ m^{-3}$) in three soils with widely different textures. The small range of MAD severely tests the abilities of most soil water sensors, particularly for the loamy sand soil. The range of water content between the water content at the MAD level, θ_{MAD}, and the wilting point, $\theta_{MAD} - \theta_{PWP}$, is even smaller.

Soil	θ_{FC} ($m^3\ m^{-3}$)		θ_{PWP} ($m^3\ m^{-3}$)		θ_{PAW} ($m^3\ m^{-3}$)		MAD (Fraction)		MAD ($m^3\ m^{-3}$)	$\theta_{MAD} - \theta_{PWP}$ ($m^3\ m^{-3}$)
Silt loam	0.295	–	0.086	=	0.209	×	0.6	=	0.126	0.083
Loamy sand	0.103	–	0.066	=	0.037	×	0.6	=	0.022	0.015
Clay	0.332	–	0.190	=	0.142	×	0.6	=	0.085	0.057

θ_{FC}, θ_{PWP}, and θ_{PAW} are soil water contents at field capacity and at the permanent wilting point, and the plant-available water. The water content at the MAD level is $\theta_{MAD} = \theta_{FC} - 0.6\ \theta_{PAW}$.

achieved in deficit irrigation practice. Although presented previously in terms of soil water content and the MAD, the objective of irrigation is to control the soil water potential within the range in which crop water uptake can occur without unacceptable impacts on yield and quality.

3. Comparisons of soil water sensors

There are many soil water content sensors, all of which work by measuring a surrogate property that is empirically or theoretically related to the soil water content. But only a few are capable of the deep measurements needed for soil water balance work. The sensors that work from within access tubes installed vertically in the soil profile are the best suited for water balance work since access tubes can be installed without disturbing the soil profile outside the tube, except in gravelly or stony soils, and since the tubes can be arbitrarily long unless the sensor design otherwise prevents deep installation (e.g., the Diviner 2000 that measures only to 1.6-m depth, or the PR2/6 that measures to only 1-m depth). Some other sensors, such as the MoisturePoint (ESI, Vancouver, BC, Canada) TDR system and conventional TDR probes (Evett, 2000) can be installed to 1.5-m depth or greater, but have not been widely accepted due to installation and soil disturbance difficulties and/or cost related to the fact that, once installed, these sensors are typically left in place such that replication of measurements requires additional sensors.

Of the sensors that work from within access tubes there are three kinds. The neutron probe is well described in the literature (e.g., Evett, 2008) and is known to be practically insensitive to soil bulk electrical conductivity (σ_{dc}) and temperature although calibrations are affected by relatively non-labile soil properties such as texture, carbonate and organic matter contents and bulk density. The surrogate measurement is of slow neutrons, which are detected in linear (for modern neutron probes) proportion to the soil water content. The other sensors all rely on measurements of the electromagnetic properties of soils and so are termed the EM sensors, of which there are two types that work from within access tubes. The most common type measures the resonant frequency of an oscillating electronic circuit that is configured such that one capacitor in the circuit is composed of two electrodes (usually cylindrical rings, one placed above the other as in Fig. 4) situated near the inside surface of a plastic access tube so that the soil outside the access tube is invaded by the fringing field of this capacitor. Once resonance is achieved (milliseconds), the EM field distribution in the soil is unchanging. The resonant frequency is related to the soil apparent permittivity, which is strongly affected by water content but also by other soil properties such as σ_{dc} and the bound water content. More details of these capacitance sensors are given in the following discussion. An early example was described by Dean et al. (1987). Recent discussions of the capacitance sensors are given in Evett et al. (2008), Kelleners et al. (2005), Mazahrih et al. (2008) and Robinson (2001). The second type of EM sensor used in access tubes has only one example to date, which is the Trime tube probe (IMKO Micromodultechnik, GmbH, Ettlingen, Germany). The system injects an electronic square wave pulse into a coaxial cable that carries the pulse to a two-electrode waveguide system in direct contact with the inside of a thin plastic access tube. The pulse moves along the waveguide and reflects from the end of the waveguide. The travel time of the pulse is related to the soil water content. Because the EM pulse is a moving EM field, it reacts to all soil along the waveguide outside the access tube. This system is termed a quasi-TDR system since it does not capture a waveform and interpret it as does a conventional TDR system. This results in errors in travel time determination and thus in water content values inferred from the travel time as discussed by Evett et al. (2006).

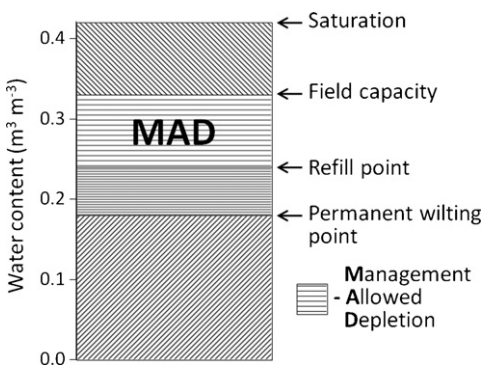

Fig. 3. Diagram of the water contents associated with irrigation scheduling according to maximum allowed depletion for a Pullman series soil. In this case the MAD fraction was 0.6, i.e. irrigation is triggered when soil water content decreases to the refill point, which is 0.6 of the range between field capacity and permanent wilting point.

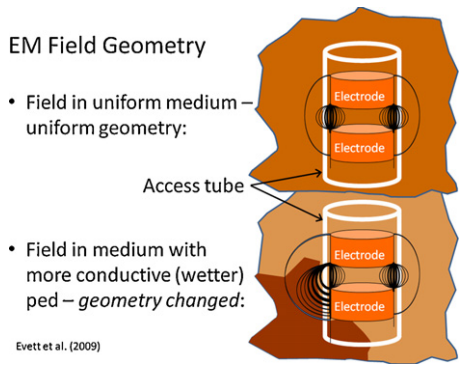

Fig. 4. Illustration of the symmetrical electromagnetic (EM) field of a capacitance sensor in a uniform, structureless soil (upper) and the asymmetrical EM field in a structured soil in which one ped is wetter and thus more conductive than the others, drawing the EM field preferentially into the wetter ped and causing sensor readings to indicated greater permittivity and water content even though the mean water content of the entire volume illustrated is the same in both cases. Note that almost all the field strength lies directly in the vertical place between the two electrodes. Only a small part of the field strength, called the fringing field, invades the soil outside the access tube.

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An early commercially available capacitance sensor was the Sentry 200AP[1] capacitance probe, which was marketed as a replacement for the neutron probe. This probe was lowered inside a plastic access tube to any user-chosen depth and a reading called the *D* value (related to frequency) was taken. Because it was similar to the neutron probe in allowing readings to be taken at multiple depths in an access tube, the NRCS was interested in replacing neutron probes with it. A joint USDA–ARS and NRCS study showed that the *D* value correlated poorly with water content as determined by volumetric sampling, but that the *D* values from four different Sentry 200AP probes used in six access tubes correlated with each other very well (Evett and Steiner, 1995). This was the first indication that the capacitance sensors responded reproducibly to some aspect of the soil–water system around the access tube that was not well correlated with water content alone. By contrast, four neutron probes were very well correlated to measured water content with calibration accuracies of <0.01 $m^3\ m^{-3}$.

Later, the International Atomic Energy Agency (IAEA) became interested in comparing the down-hole capacitance sensors and time domain reflectometry (TDR) to the neutron probe. This task was carried out over four years (1999–2003) in four continents; and the results were written as an IAEA Guidelines (Evett et al., 2008).

To put this into perspective, the down-hole capacitance sensors all employ a capacitive sensor consisting of two cylindrical electrodes arranged along a common axis and separated by a space that is typically filled with a plastic. This arrangement is placed in an access tube composed of some sort of plastic or plastic–fiberglass composite, which itself is inserted into the soil. The electrodes are connected to an oscillating circuit that settles to a resonant frequency that will change if the dielectric properties of the soil material around the access tube change. Since the two electrodes are in the access tube, most of the EM field developed between the electrodes is itself confined to within the access tube. Only a small part of the EM energy, called the fringing field, enters and interacts with the soil. The volume of soil penetrated by the EM field is static, although we now know that its exact shape and volume are dependent on the arrangement of the resistive and capacitive elements of the soil outside the access tube. The capacitance of the soil-access tube system, $C(F)$, is given by (Dean et al., 1987):

$$C = g\varepsilon_a\varepsilon_0 \tag{3}$$

where ε_a is the system apparent relative permittivity (–), ε_0 is the permittivity of free space (F/m) and the geometric constant, g (m), has a value dependant on the geometry of the system. The resonant frequency, F (Hz), is given by (Dean et al., 1987):

$$F = [2\pi(L)^{0.5}]^{-1}(C^{-1} + C_b^{-1} + C_c^{-1})^{0.5} \tag{4}$$

where C_b and C_c are the electrode capacitances including the capacitances of internal circuit elements to which the electrodes are connected, C is the capacitance of the soil-access tube system defined in Eq. (3), and L is the inductance (henries) of the coil in the LC oscillating circuit. As soil water content increases, C increases in concert with the increasing soil apparent permittivity, and F decreases. In contrast, conventional TDR measures the travel time of an EM pulse that travels along electrodes buried or inserted in the soil and that is reflected back to its source.

Multiple field and laboratory studies have shown that capacitance sensors used in access tubes were much less well correlated with field measured water contents than was the neutron probe (Evett and Steiner, 1995; Evett et al., 2006, 2009; Mazahrih et al., 2008). Also, standard deviations of profile soil water content determined by capacitance sensors and by a down-hole quasi-TDR sensor were larger than those determined using the neutron probe or gravimetric sampling by hydraulic push probe, so much so that numbers of access tubes and sensors required to obtain reasonable field mean profile water contents was unaffordable (Table 3). In every case, the standard deviation increased in drier soil with obvious implications for use of soil water sensors in regulated deficit irrigation management. Many of the results of the multi-national IAEA study were published in a guide to field estimation of soil water content (Evett et al., 2008).

Table 3
Number of access tubes (n) needed to find mean water content to a precision d ($m^3\ m^{-3}$) at P = 95% for a given field-measured standard deviation (s, $m^3\ m^{-3}$) of water content for four capacitance (cap.) sensor systems and one quasi time domain reflectometry (TDR) system compared with gravimetric/volumetric direct sampling and the neutron probe.

Method	Soil condition	s ($m^3\ m^{-3}$)	n	
			d = 0.01	d = 0.02
Diviner 2000 (cap.)	Irrigated	0.0131	7	2
	Dry	0.0242	23	6
EnviroSCAN (cap.)	Irrigated	0.0152	9	2
	Dry	0.0266	27	7
Delta-T PR1/6 (cap.)	Irrigated	0.0272	28	7
	Dry	0.1216	568	142
Sentry 200AP (cap.)	Overall	0.0378	55	14
Trime T3 (quasi-TDR)	Irrigated	0.0075	2.2	1
	Dry	0.0238	22	5
Gravimetric by push probe	Irrigated	0.0045	1	1
	Dry	0.0070	2	1
Neutron probe	Irrigated	0.0015	1	1
	Dry	0.0027	1	1

The capacitance sensors, though relatively poorly correlated with field measured soil water content, were very well self-correlated when measurements at the same depth and access tube were compared among sensors (Evett and Steiner, 1995; Evett et al., 2009). This indicated that capacitance sensors responded to some property of the soil–water system around the access tube that was not water content alone. Evett and Steiner (1995) hypothesized that this property was related to soil structure and non-uniform penetration into the soil of the EM field of the sensor. Later studies of EM field penetration in heterogeneous materials showed over estimation of permittivity and uneven EM field penetration in those materials (e.g., Panteny et al., 2005), supporting the inference that the EM field from a capacitance sensor is distorted by the individual arrangement of soil peds and pattern of water content in the peds around each access tube at each depth (e.g., Fig. 4) rather than being responsive to the mean water content of the soil around each access tube at each depth. This means that the geometric constant (fundamental to capacitance measurement theory) changes according to the small scale heterogeneity of soil properties at each measurement depth and access tube, which results in a different resonant frequency and water content estimate even if mean water content around the access tube is the same. Using a different EM sensor, Logsdon (2009) confirmed that uneven water contents in proximity to the sensor caused the sensor to overestimate water content.

Spatial variation of field data from EM sensors was contrasted with that from the NP and volumetric gravimetric sampling by Evett et al. (2009) through calculation of the mean relative difference, $\bar{\delta}_i$, of soil water storage to 1-m depth at each access tube location (10–20 locations) for the six soil water assessment methods studied. The relative difference, δ_{ij}, for location i and time j was defined by (Vachaud et al., 1985):

$$\delta_{ij} = \frac{S_{ij} - E[S_{ij}]}{E[S_{ij}]} \tag{5}$$

[1] Mention of trade names or commercial products in this report is solely for the purpose of providing specific information and does not imply recommendation or endorsement by the U.S. Department of Agriculture.

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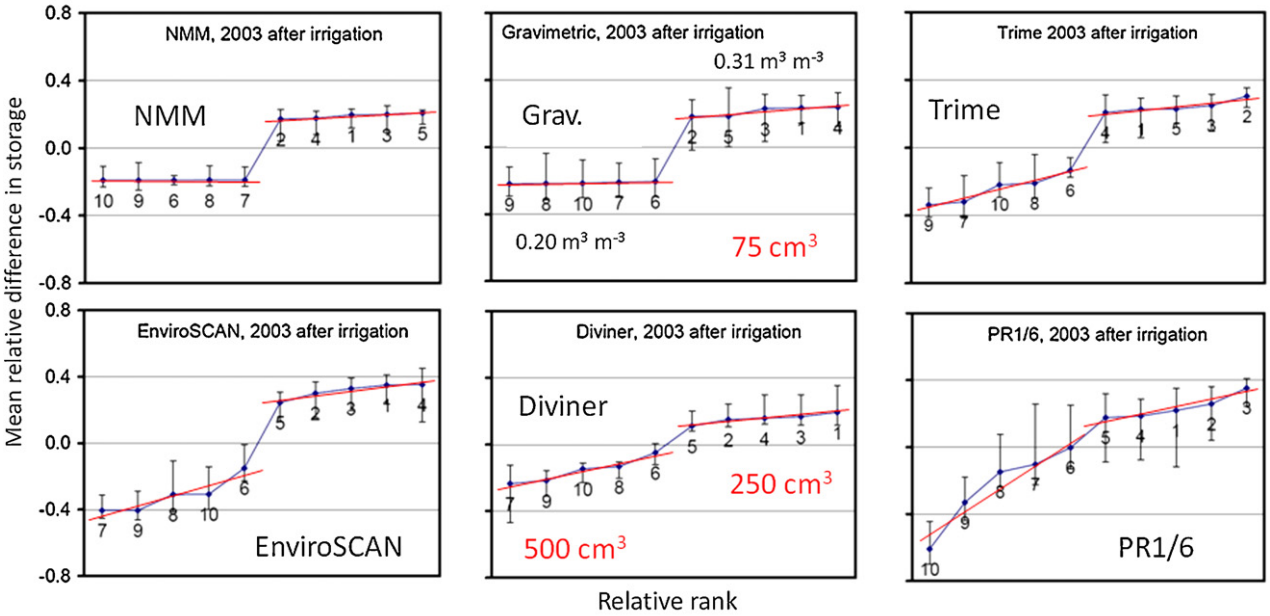

Fig. 5. Plots of the mean relative difference in storage for several sampling dates in 2003 for the neutron probe (NP), volumetric–gravimetric sampling (Grav.), the Trime T3 tube probe (Trime), and the EnviroSCAN, Diviner and PR1/6 capacitance sensor systems. Samples are for profile water contents in the top meter of soil from ten access tubes spaced 10-m apart in linear E–W transects for each sensor system. Half the tubes were in the irrigated half of a winter wheat field (numbered 1–5) and half were in dryland (6–10) with 1 furthest east and 10 furthest west. Volumes in cm^3 are the volumes of the volumetric–gravimetric samples (75) and the Diviner capacitance sensor in wet soil (250) and in dry soil (500). The Diviner sensing volumes are reported in Evett et al. (2006) as are the soil-specific calibrations for each sensor.

where $E[\ ]$ is the expected value operator and S_{ij} is the profile water content at location i and time j. Calculating δ_{ij} for all times and locations allows calculation of the mean relative difference over time, $\bar{\delta}_i$, for each location. Plotting of $\bar{\delta}_i$ vs. rank, with limit bars for the maximum and minimum relative difference at each location, allows easy identification of locations that represent the mean, of locations with extreme values, and of locations that maintained their relative rank with the least variation with time, including locations in a field where the profile water content was usually very representative of the mean for the field. Particularly relevant for studies of soil water sensors, ranked $\bar{\delta}_i$ plots also indicate the overall spatial variability of the data in the area under study.

In studies over three seasons, plots of $\bar{\delta}_i$ for the neutron probe (NP) and volumetric–gravimetric sampling were nearly identical (e.g., Fig. 5) (and Fig. 11 in Evett et al., 2009). Both showed a clear difference in profile water content between the irrigated side of the field and the non-irrigated side, and both showed little spatial variation in values within either the irrigated side or the non-irrigated side. Results from the Trime T3 probe (a quasi-TDR sensor) showed more variation in the non-irrigated side of the field, variation that did not exist at the scale of NP or gravimetric determinations. Evett et al. (2005) explained how inaccurate travel time determinations in the face of electrical losses common in soils compromised the ability of the Trime sensor to accurately reflect water content. Two capacitance sensors (Diviner and EnviroSCAN) showed similar results, again with somewhat more variation in the drier side of the field. The PR1/6 capacitance sensor data showed even more variation, such that the irrigated side of the field was not clearly distinguished from the dry side. Again with the PR1/6, the variation in the dry side appeared to be larger. The consistent increase in apparent variability in the dry side of the field compared with the irrigated side for the EM sensors is consistent with several prior studies that showed that soil water content variability increased as soils dried; however, the effect in this case is not realistic. The effect is strong enough that it overrides the effect of sampling volume, which is larger for the capacitance probes in dry soil than in wet soil (e.g., approximately 250 cm^3 in wet soil vs. 500 cm^3 in dry soil for the Diviner; Evett et al., 2006). In fact, the volume for the gravimetric samples was only 75 cm^3, much smaller than that of the EM sensors, yet spatial variability of gravimetric readings was very small in both wet and dry sides of the field compared with the variability of any of the EM sensors. Again, this is evidence that the EM sensors are sensitive to variations in soil water content at scales smaller than the sensor sampling volume, and it is fairly direct evidence that the EM fields of these sensors must preferentially permeate the wetter parts of the soil at small scales near the access tube at each depth and access tube. Repetition of this experiment in 2004 and 2005 showed similar results (Evett et al., 2009). Unfortunately, few side-by-side comparisons of this sort have been made using multiple access tubes in transects to compare reported variability over space with directly measured variability. In one of the few such studies, Heng et al. (2002) reported greater variability from the Diviner 2000 compared with neutron probe and TDR in a gravelly clay loam soil, as well as soil water storage under and over estimation by the Diviner of up to 50 mm in a 60-cm profile.

Non-uniform soil wetting may occur due to, for example, non-uniform infiltration rate, air entrapment and preferential flow (Clothier et al., 2008; Flury et al., 1994; Ghodrati and Jury, 1990, 1992; Hardie et al., 2011; Hendrickx and Flury, 2001). We note that these are sources of soil water content variability, not error; but because the EM fields of capacitance type sensors will preferentially invade wetter, more conductive soil volumes, such soil water content variability may induce error in capacitance-based soil water sensing. The data discussed here show that neutron probe and gravimetric sampling both resulted in data that showed equivalent and relatively small spatial variation except for that imposed by different irrigation rates, while the EM methods produced data that showed much larger spatial variation. This excess variation is irrelevant to estimation of soil water at scales explored by whole plant root systems and leads to errors in soil water balance calculations as discussed by Evett et al. (2009).

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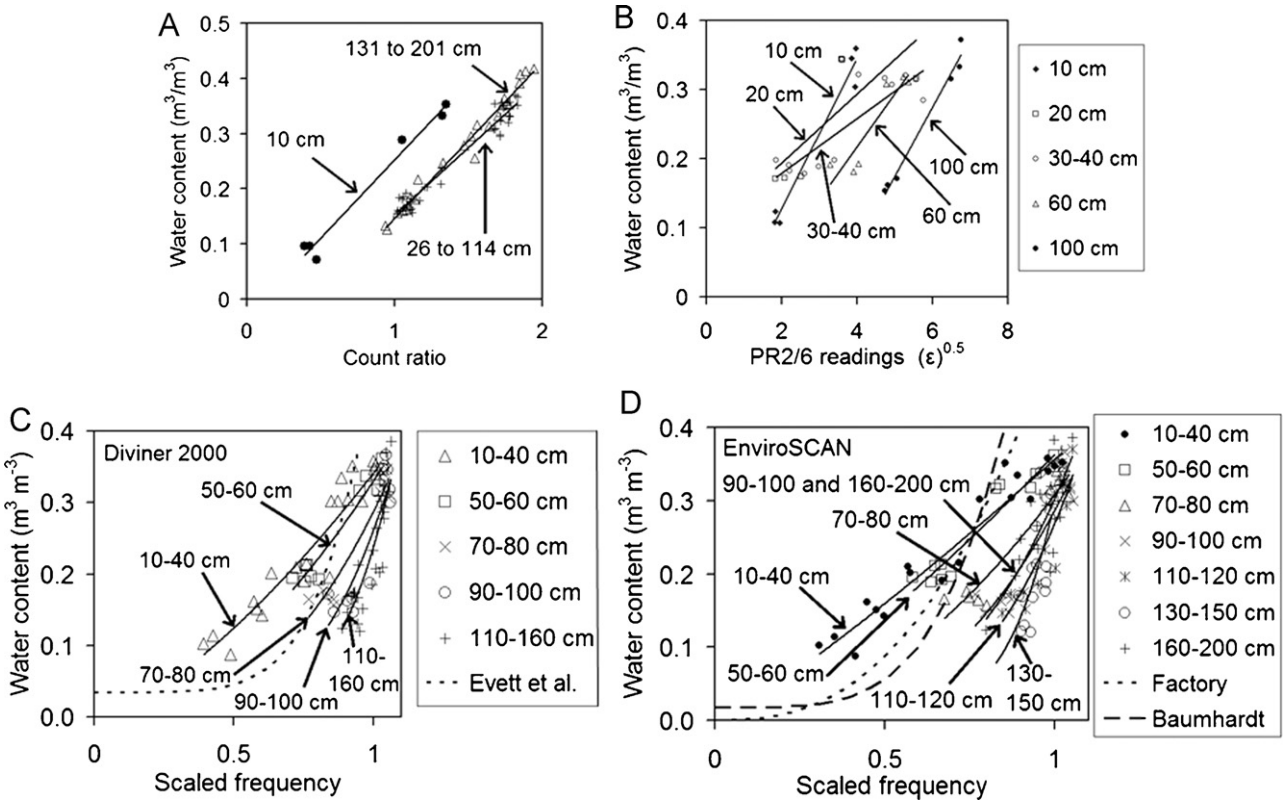

Fig. 6. Field calibrations reported by Mazahrih et al. (2008) in a deep soil in the Central Valley of California. Shown are calibrations for the neutron probe (A) and three capacitance sensing systems; the PR2/6 (B), the Diviner 2000 (C) and the EnviroSCAN (D). For plotted symbols, the vertical axis values are water contents measured using volumetric samplers.

4. Calibration accuracy of capacitance sensors

Laboratory calibrations of capacitance sensors were conducted in large soil columns (three replicates each of ground, sieved and repacked soil from the Ap, Bt, and calcic Bt horizons) by Evett et al. (2006). Calibration accuracy ranged from 0.018 to 0.058 $m^3\,m^{-3}$ (RMSE of regression), comparable to values reported by Baumhardt et al. (2000) and Paltineanu and Starr (1997) for laboratory calibrations, but larger than the calibration accuracy of $\leq$0.01 $m^3\,m^{-3}$ for conventional TDR and NP in the same study. Laboratory calibrations did not, however, provide accurate soil water contents when used in the field (Evett et al., 2009). In the field, water contents were overestimated on the wet end and underestimated on the dry end; and values of change in profile water storage were relatively inaccurate for the capacitance sensors when compared with NP or gravimetric sampling. Subsequently, side-by-side field calibrations of the NP and three capacitance sensor systems were conducted at the West Side Field Station in the San Joaquin Valley, California (Fig. 6; Mazahrih et al., 2008). Calibrations for the NP had high accuracy compared with those from the three capacitance sensors; and only three separate calibrations were needed to cover the 2-m deep profile over which the NP was calibrated. Also, the calibration for the depth range from 26 to 114 cm was only slightly different from that for the 131–201 cm depth range. In contrast, calibrations for the capacitance sensors were affected by increasing bulk electrical conductivity (σ_{dc}) with depth, despite the fact that the sweet pepper crop in the drip irrigated field showed no salinity symptoms. Capacitance sensor calibration slopes increased strongly with depth; and the sensor frequency response to increasing water content became very small as depth increased, making the sensor output highly variable. Since the profile pattern of σ_{dc} is expected to vary greatly across the field and with time (e.g., Burt et al., 2003; Hanson et al., 2003), these calibrations are essentially unusable for accurate soil water content estimation using the capacitance sensors. Other examples of calibration errors increasing with soil σ_{dc} were given by Baumhardt et al. (2000) and Evett and Schwartz (2009).

Fares and Polyakov (2006) reviewed several other calibrations, but did not report on the accuracy (RMSE) of calibration. Unfortunately, failure to report accuracy of calibration is common in studies of soil water sensors and studies employing soil water sensors. Typically, separate calibrations are not done for soil layers that differ in properties that affect EM sensors. The commonly reported coefficient of determination (r2) is useless in assessing accuracy. Geesing et al. (2004) reported RMSE values of 0.03–0.04 $m^3\,m^{-3}$ for field calibrations of the Diviner 2000 in loam and silt loam soils, respectively; and demonstrated the necessity of soil-specific calibration as have other authors (Baumhardt et al., 2000). For field calibrations at two sites, Polyakov et al. (2005) reported RMSE = 0.031 and 0.048 $m^3\,m^{-3}$ for calibration of a capacitance sensor system (EasyAg 50, Sentek, Pty., Ltd., Stepney, South Australia) in a kaolinitic silty clay loam in Hawaii. Although soil classification was the same, calibrations for the two sites (one was a hillslope and the other a nearby cultivated terrace) were clearly different. A laboratory calibration for the same soil using re-packed soil columns resulted in RMSE = 0.039 $m^3\,m^{-3}$. Results with capacitance sensors have tended to be better in sandy soils for which calibration coefficients tend to be similar (Robinson, 2001). However, sandy soils tend to have FC water contents <0.10 $m^3\,m^{-3}$ (Morgan et al., 1999) and available water holding capacity of <0.04 $m^3\,m^{-3}$ (Table 2), which places great demands on accuracy. Although Morgan et al. (1999) reported an RMSE of calibration of 0.0085 $m^3\,m^{-3}$, other calibration studies have reported larger values for coarse textured soils. For example, data from a sandy to sandy loam soil in California resulted in RMSE = 0.031 $m^3\,m^{-3}$; and data from a loamy sand to sandy loam soil in Australia resulted in RMSE = 0.016 $m^3\,m^{-3}$

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(Paltineanu and Starr, 1997). Both were laboratory calibrations using re-packed soil columns.

5. Sources of capacitance sensor imprecision and inaccuracy

Several facets of EM field behavior impact the way in which a capacitance sensor responds to temporal and spatial variations of water content and σ_{dc} within the measurement volume of the sensor. Theory dictates that the EM field volume decreases as frequency increases; and this is borne out by measurement (e.g., Evett et al., 2006). The EM field is radiated as from an antenna and the volume of soil penetrated by the field is static, although not necessarily uniform, for a given realization of the spatial pattern of θ_v and σ_{dc} within the sensing volume. The EM field follows conductive paths (e.g., Schwank et al., 2006) and thus radiates preferentially into more conductive (wetter) peds, causing readings to be biased to higher water content and to be more adversely affected by structured soils with appreciable σ_{dc}. Clay content and type, salinity, and temperature all affect σ_{dc}. Kelleners et al. (2005) suggested that capacitance sensor measurement frequencies should be greater than 500 MHz. However, they did not address the fact that as frequency increases the sensed volume decreases, which could lead to even greater variability in reported water content than we observed. Finally, capacitance sensor measurements are based on Gauss' Law, which relates apparent permittivity to the charge movement or current flow that accompanies an electric field change in a dielectric medium. The resulting value of permittivity is inextricably linked to the field volume (geometric effect), which we have seen is highly affected by σ_{dc} and soil structure. In this regard, the ring oscillator and standing wave oscillator sensors are also subject to Gauss' Law and fall in the category of capacitance sensors even though some are described as TDR sensors by vendors.

All capacitance EM sensors respond to the frequency dependent complex permittivity, $\varepsilon^*_\gamma(\omega)$, which has both real and imaginary components (Kelleners et al., 2005):

$$\varepsilon^*_\gamma(\omega) = \varepsilon'_\gamma(\omega) - j\varepsilon''(\omega)_\gamma = \varepsilon'_\gamma(\omega) - j\frac{\sigma_{dc}}{\omega\varepsilon_0} - j\varepsilon''_{r,\,relax}(\omega) \quad (6)$$

where ε'_γ is the real component of the complex dielectric permittivity (mostly related to the free water content), $j^2 = -1$, and the imaginary component includes the permittivity due to relaxation losses, $\varepsilon''_{r,\,relax}$ (e.g., bound water effects), and the loss tangent, $\sigma_{dc}/(\omega\varepsilon_0)$, and ε_0 and ω are defined above. Capacitance sensor operating frequencies (ω), although greater currently than in the past, are still well below 200 MHz, low enough that there is substantial interference from bound water relaxation, interfacial polarization and σ_{dc}/ω effects on the sensed permittivity (Schwartz et al., 2009a,b; Kelleners and Verma, 2010; Wagner et al., 2011). Estimation of water content under such scenarios requires accounting for the influence of these interrelated factors on both the real and imaginary permittivity and their combined effect on sensed apparent permittivity, ε_a. Capacitance sensors work by measuring the change in frequency of an oscillator, with lower frequencies being related to larger water contents. Because permittivity values can be strongly dependent on measurement frequency in the range employed by capacitance sensors, there is an inherent confounding of the change in frequency and the resulting estimate of permittivity and thus water content.

6. Implications for irrigation scheduling and determination of crop water productivity

The soil water sensing systems based on capacitance measurements are inherently ill suited to accurate measurement. Of the several important problems, the most fundamental and intractable one is that the static EM field does not uniformly permeate the soil around the access tube in structured soils, which includes most agricultural soils. This results in spatial variation in reported profile water contents that does not exist at the measurement scales of gravimetric sampling or neutron probe methods. Such variation is irrelevant to the estimation of soil water at scales explored by whole plant root systems. As shown in the preceding sections, the lack of accuracy extends to differences in water content as well as for absolute values. The reasons for this are complex but include (i) both under estimation and over estimation of water content can occur depending on the actual soil water content since σ_{dc} increases with water content and calibration is affected by σ_{dc}, (ii) soil temperature changes over time strongly affect σ_{dc}, resulting in calibration changes over diel and seasonal periods, and (iii) wetting and drying events over a season will cause different patterns of small scale soil wetness, which will in turn cause different patterns of EM field penetration in the soil at the same location (x,y,z) with time.

Calibrations for capacitance systems will be highly affected by spatial and temporal variations in soil salinity, clay content and type as related to soil electrical conductivity, and soil temperature effects on bulk electrical conductivity and bound water. Since soil salinity varies with time and space under most irrigation systems, particularly drip, large errors in capacitance sensor measurements can occur and are fairly common since 20% of the world's irrigated soils are salt affected (Evett et al., 2006; Evett, 2007; Hachicha and Abd El-Gawed, 2003). From the data on bulk electrical conductivity values in irrigated fields and the sensor dependencies on bulk EC reviewed by Evett (2007), errors larger than 50% in water content could occur at a single location, inducing errors similarly large in profile water content. Thus there is little chance that capacitance systems will prove accurate enough for regulated deficit irrigation practice using management allowed depletion concepts or for determining ET and CWP. For these objectives, we recommend that soil water content be determined using the neutron probe, gravimetric sampling and conventional time domain reflectometry (TDR) methods. Although TDR is difficult to use deeply, the effects of bound water and σ_{dc}/ω for TDR are smaller (due to the larger frequencies involved in TDR) and better understood; and calibration methods exist that practically eliminate temperature and conductivity effects for TDR (Evett et al., 2005; Schwartz et al., 2009a,b). And with TDR, the EM pulse is not a static field but a moving field that is forced to pass along electrodes and be affected by both drier and wetter soil peds. That this results in a true average response to permittivity variations along the electrodes has been well established (Ferré et al., 1996; Hook and Livingston, 1995). Because TDR measures the propagation time of a moving EM field, it is not subject to Gauss' Law but rather to Maxwell's equations. The TDR method is thus insensitive to geometric effects on the sensed permittivity so long as travel time is correctly determined. If TDR methods can be made easier to apply, more reliable and amenable to deep measurement without soil disturbance (e.g., Casanova et al., 2011), then TDR methods may eventually supplant the neutron probe for determinations of ET and CWP.

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