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ENCYCLOPEDIA OF EARTH SCIENCES, VOLUME XII

The ENCYCLOPEDIA of SOIL SCIENCE PART 1

Physics, Chemistry, Biology, Fertility, and Technology

EDITED BY

Rhodes W. Fairbridge

Columbia University

Charles W. Finkl, Jnr.

Ocean Sciences Center
Nova University

Dowden, Hutchinson & Ross, Inc.
Stroudsburg Pennsylvania

SURFACE SOIL WATER CONTENT

Water contained in the surface soil is of great practical importance for agriculture, forestry, and rangeland management (Skidmore, Dickerson, and Schimmelpfennig, 1975; Ulaby, Cihlar, and Moore, 1974). Idso et al (1975*b*) emphasized the importance as follows:

> The presence of an adequate water supply in the upper-most few centimeters of the soil is essential for proper seed germination and crucial to early development of emerging crops. Its presence is also a factor in partitioning water income via rainfall and irrigation into runoff, deep percolation and storage. Disposition of stored water through evaporation is further dependent upon soil water content, as is erosion of the soil itself by wind action. Even populations of many economically important insect pests rise and fall with variations in the water content near the soil surface, since it is here that their eggs are deposited. Also, surface water content is a valuable boundary condition for meteorologists modeling the general circulation of the atmosphere.

Referring to the significant role of water in determining the nature and properties of soils, Low (1973) reported that in its absence, life in the soil comes to a virtual standstill and that all the physical, chemical, and biological reactions and processes that occur in the soil depend on its presence. Surface soil water has not only great practical importance but it also has theoretical implications. It enters as a basic component into the hydrological cycle, into energy exchanged near the surface, and into modeling of various ecosystem processes. It is also highly dynamic.

The water content at the interface of the soil and the atmosphere is constantly changing. It may change rapidly because of precipitation, irrigation (q.v.), soil drainage (q.v.), and/or evaporation (q.v.). Precipitation may occur violently in the form of wind-driven rain and/or hail, or gently in the form of dew or soft-landing snowflakes. The water thus received at the soil surface may run off or infiltrate into the soil and be distributed in the soil volume according to the laws that govern saturated and unsaturated moisture flow (see *Flow Theory*), or it may evaporate from the surface.

Surface Soil Water Content and Soil Detachment

The amount of water in surface soil particles greatly influences their detachment and transport by wind (see *Wind Erosion*). Chepil (1956) found that cohesion of water films between erodible-size soil particles varies directly with water content. He found that resistance of soil particles was equal to $6W^2$ where W, equivalent moisture, is a ratio of the water content in question to water content at -15 bar pressure potential. When the equivalent moisture is 1.0 (-15 bar pressure potential), the force of attraction between soil particles is about 6 dyn/cm^2. A friction velocity of 70 cm/s is required to produce that much surface drag.

Most natural winds are not strong enough to exert that much force to overcome the cohesive force of water at -15 bar pressure potential. However, as noted earlier, water at the surface is subject to drying rapidly. The pressure potential of the water in the atmosphere above the soil at 30°C and 50% relative humidity would be only -970 bars. Therefore, there is usually a steep water-potential gradient (conducive to rapid drying) between a moist soil and the atmosphere above it.

Water and Crust Formation

Water content and the wetting method affect the physical conditions of the soil surface. Surface flooding (Kemper, Evans, and Hough, 1974) and driving rain (Lyles et al, 1969) both destroy soil aggregates. Breakdown and dispersion of soil aggregates create a condition conducive to crust formation.

A saturated or near-saturated state facilitates particles' coming close to each other, a condition necessary for them to become cemented together (Uehara and Jones, 1974). Then, as stated by Uehara and Jones (1974):

> As water is removed by drainage and/or evaporation, pore water pressure increases negatively. If the surface layer consists of water-stable aggregates, water drains rapidly from the large inter-aggregates' pores, and subsequent evaporation results in increased negative pore water pressure within the aggregate, which in turn brings particles in aggregates closer together. If, on the other hand, the surface layer consists of aggregates which slake in water, large pores disintegrate upon wetting and pore size distribution is narrowed and shifted to the fine range.

The effects of reduced pore size (bringing the particles closer to each other) and therefore higher attainable negative pore-water pressure enhance bond formation or crusting.

As the soil drys, bonds form between soil particles, and crust develops. Cracks develop in the crust as the soil shrinks from drying.

If the soil is wetted gently, the aggregates may remain intact through wetting. However, when they are subsequently dried by evaporation, uneven forces develop. The aggregates usually dehydrate nonuniformly so that unequal strains develop. Strains and stresses that develop from wetting and drying break down soil aggregates.

Changeableness of Surface Soil Water Content

The surface or thin layer of porous soil, bounded by the atmosphere above and more soil below, is subject to extreme variations in several parameters, including water content. How many of these parameters vary depends on water content; thus various methods have been devised to detect and measure water content (see *Water Content and Retention*).

A unit volume of most mineral soils is about half solid and half pore space with the pores nonuniform and ranging in diameter from <0.1 to >1000 μm. The total pore volume is the potential volume of water that can be stored in the soil. When the soil is dry, the pores contain no water and are filled with air.

Water received at the soil surface displaces the air in the soil pores. Gravity causes some water to drain from the large pores. The distribution of water in soil depends on a balance between forces that pull the water toward the soil particles and surface-tension forces at the liquid-vapor interface, which keep the surface area of the liquid water at a minimum. The effective diameter of the largest water-filled or smallest air-filled pore in a system can be calculated from

$$D = \frac{\sigma}{RT \ln P/P_0} \qquad (1)$$

where σ is water-surface tension; R is universal gas constant; and P/P_0 is relative humidity. The denominator on the right side of equation 1 represents free energy or water potential. At water potentials of 10^4 (0.01 bar), 10^5, 10^6, 10^7, 10^8 ergs/g, the diameters of the largest water-filled pores are 288, 28.8, 2.88, 0.288, and 0.0288 μm, respectively.

As the diameter of the largest water-filled pore decreases with lesser water contents, the mobility of the water also decreases (Table 1). Hydraulic conductivity (q.v.) may change over

TABLE 1. Some Soil Properties Influenced by Amount of Water Contained in the Soil.

Parameter	Dimension	Soil Material	Wet	Dry	Reference
Albedo	–	Avondale loam	0.14	0.30	Idso et al, 1975*c*
Thermal conductivity	Cal/cm/s/C	Fairbanks sand	5.49×10^{-3}	0.79×10^{-3}	deVries, 1966
Thermal conductivity	Cal/cm/s/C	Healy clay	3.68×10^{-3}	0.38×10^{-3}	deVries, 1966
Heat capacity	$Cal/cm^3/C$	Mineral soil	0.60	0.21	deVries, 1966
Soil temperature*	C	Silty clay	22	54	Idso et al, 1975*b*
Water potential	ergs/g	–	1.4×10^6	960×10^6	Calculated**
Hydraulic conductivity	cm/day	Fine sand	188 @ sat.	0.01 @ θv = 0.10	Bruce, 1972
Hydraulic conductivity	cm/day	Yolo loam	50 @ sat.	0.01 @ θv = 0.30	Davidson et al, 1969
Hydraulic conductivity	cm/day	Slate dust	86 @ θv = 0.10	0.086 @ θv = 0.40	Youngs, 1964
Polarized light	–	–	0.89	0.16	Doll, 1973
Microwave emissivities	–	Avondale loam	0.50	0.90	Poe et al, 1971
Dielectric constant	–	Vernon clay loam	30 @ 21 cm	3 @ 21 cm	Lundien, 1971
Dielectric constant	–	Silty clay loam	16 @ 1.55 cm	3 @ 1.55 cm	Schmugge, 1974

*Diurnal amplitude of surface-soil temperature wave.

**Water potential equals RT 1n P/P_0; where R is universal gas constant 0.462 ergs/°K/g; T is temperature in degrees Kelvin; and P/P_0 is relative vapor pressure or humidity. P/P_0 was 0.999 and 0.50 for wet and dry, respectively.

several orders of magnitude with change in water content.

When the evaporation rate from the soil exceeds the soil's ability to transmit water to that surface, the amount of water in the surface soil can change rapidly. Idso et al (1975*c*), using albedo measurements, found that surface volumetric water of a bare Avondale loam soil at Phoenix, Arizona, decreased from 0.20 to 0.07 cm^3/cm^3 between 1300 and 1500 h 2 days after the soil was irrigated in July. This process was slower in December when the surface soil did not dry until 10 days after irrigation; the volumetric water content decreased from 0.16 to 0.06 cm^3/cm^3 between 1000 and 1500 h.

Jackson (1973), who gravimetrically measured soil water in samples taken at 0- to 0.5-cm increment at 0.5-h intervals, showed that the amount of water in that layer varied widely during a day (Figure 1). He found that water content decreased from 0.23 cm^3/cm^3 at sunrise to 0.10 cm^3/cm^3 at 1400 h, five days after irrigation; that during the night the volumetric water content increased to 0.19 cm^3/cm^3, then decreased rapidly again during the day; and that water content and amplitude of change in content decreased gradually with time after irrigation. This diurnal pattern continued for all days of his experiment.

Measurement

The standard method for measuring soil water content is gravimetric. A soil sample is obtained, weighed, dried at about 105°C, and then reweighed. Then gravimetric water content is calculated by first subtracting the dry-soil weight from the wet-soil weight and dividing the result by the dry-soil weight. To convert that gravimetric or mass ratio (mass of water per unit mass of soil) to volumetric water content, multiply it by the soil bulk density. When the soil bulk density is not known, a separate bulk-density determination is required. (Usually for mineral soils the bulk density (q.v.) is between 1.0 and 1.4 g/cm^3.)

Soil samples to be used in determining gravimetric and volumetric water content must be obtained over some finite depth, usually several centimeters. For detailed measurements, samples have been carefully obtained at 1-cm intervals (Jackson, 1973). Of the several methods described in recent literature, some are for sampling at the near surface only; others for sampling a thin layer of soil. Several involve noncontact remote sensing, which with further refinement and evaluation, may be useful in detecting water amounts in soil by satellite or airplane overflight (Idso, Jackson, and Reginato, 1975*a*).

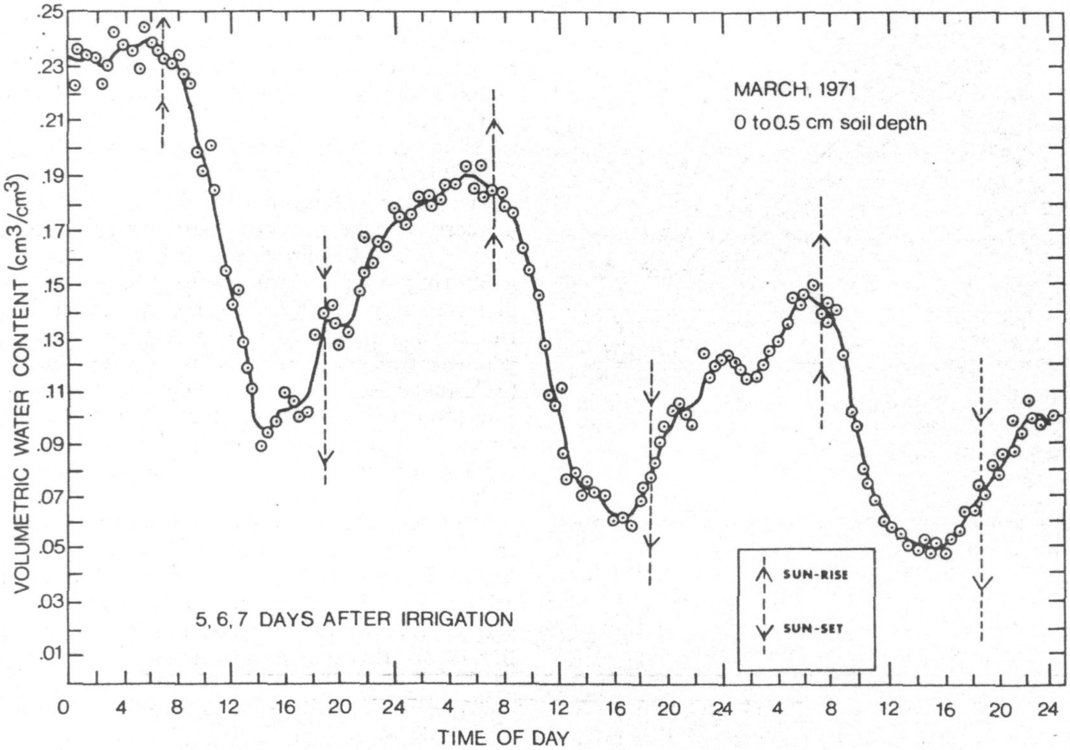

FIGURE 1. Volumetric water contained in the 0- to 0.5-cm increment versus time for 3 days during March 1971. (After Jackson, 1973)

Infrared Reflectance. Using water's property to absorb certain wavelengths in the near infrared as a basis, Skidmore, Dickerson, and Schimmelpfennig (1975) evaluated surface soil water content by measuring reflectance of near-infrared radiation from a soil surface. Previously Bowers and Hanks (1965) had demonstrated the potential of using reflectance measurements to determine soil water content by showing that reflectance at a 1.9-μm wavelength, for a Newtonia silt loam, decreased as water increased. Initial testing of the reflectometer indicated that water 10 μm thick (10 mg/cm^2) absorbed most electromagnetic radiation at the 1.95 μm wavelength. With as little as 1.0 mg/cm^2 of water, the reflectance attenuated >25%. Since infrared radiation does not penetrate deeply, the amount of IR energy reflected is mainly a surface phenomenon, thus indicating the water content only very near the surface.

The shape of the reflectance versus soil water content curve (Figure 2) closely resembles a log-linear relationship for water contents between oven-dried soil and the point of small change in reflectance as water content is increased. The least-squares fit for a log-linear relationship for soil water contents between oven-dried and 0.3 bar tension gave R^2 values of 0.99, 0.96, 0.99 for Carr sandy loam, Farnum sandy clay loam, and Smolan silty clay loam, respectively. At low water contents soil properties (other than water) strongly influenced soil reflectance.

Albedo. In a series of experiments Idso et al (1975*c*) intensively and concurrently measured

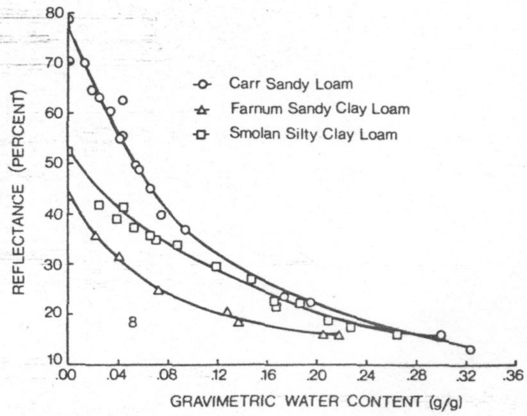

FIGURE 2. Reflectance at 1.95 μm of three soils as influenced by their water content. (After Skidmore, Dickerson, and Schimmelpfennig, 1975)

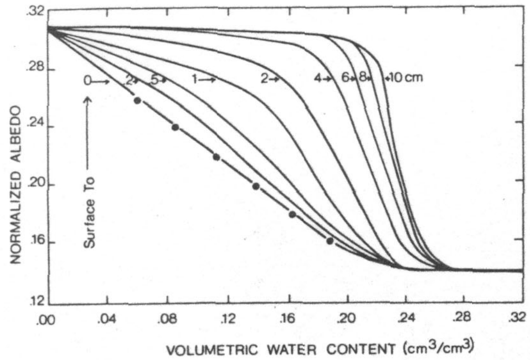

FIGURE 3. Normalized albedo versus average volumetric water content of nine different soil layers. (After Idso et al, 1975c)

the albedo (ratio of reflected to incoming solar radiation) and soil water content of a drying, bare Avondale loam soil. They normalized albedo for sun-zenith-angle effects and combined (in one set) several curves depicting normalized albedo versus soil water content (Figure 3). The volumetric soil-water content values were extracted from these plots for several constant albedo values at the various depth-interval lines and plotted against this latter parameter (Figure 4). Smooth lines were drawn connecting those points and extrapolated to yield the volumetric soil water contents at the soil surface for the chosen constant, normalized, albedo values. These albedo surface water content points (Figure 3) were then replotted to give the relationship between normalized albedo and water content near the surface of the Avondale loam soil. This relationship was linear over a water content range from 0 to 0.18 cm³/cm³.

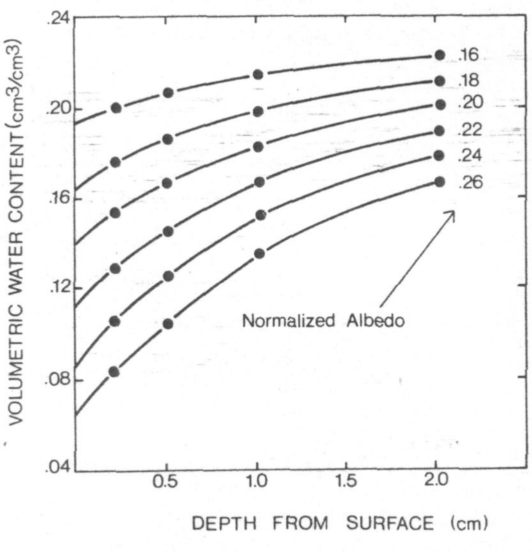

FIGURE 4. Average volumetric water content versus depth at which six selected constant values of normalized albedo were averaged. (After Idso et al, 1975c)

Microwave Radiometry. Since the soil's dielectric properties vary with water content, these properties provide another method of measuring surface soil water content. The dielectric constant of water at microwave frequencies can be as large as 80 for wet soils, whereas for dry soil it is usually less than 5 (Schmugge et al, 1974). Poe, Stogryn, and Edgerton (1971) found that the microwave emissivities for a bare, smooth soil varied from 0.5 for very wet soil to more than 0.9 for a dry soil.

Schmugge et al's (1974) calculations indicated that the microwave emission at 1.55-cm wavelength from soils in a layered model is determined by the dielectric properties of the surface layer, and hence its moisture content. Based on this phenomenon, Schmugge et al (1974) investigated sensing of soil water with microwave radiometers. They found the emission from sandy loam and clay loam soils were a linear function of soil water content over the 0–35% range at a wavelength of 21-cm (Figure 5).

Radar. Radar or active microwave is also used to detect soil water content. Water in the surface soil strongly affects the microwave return, which is not directly influenced by the soil's chemical and mineralogical properties (Lundien, 1971; Edgerton et al, 1971). However, Ulaby's (1974) results indicated that the radar response to soil moisture content depends highly on surface roughness, microwave frequency, and look angle. In a subsequent study, on active microwave response to soil moisture based on the soil skin-depth (attenuation) concept, Ulaby, Cihlar, and Moore (1974) reasoned that because the soil thickness interacting with the electromagnetic radiation varied with moisture content and incident angle, microwave response to soil moisture should include these variations. Hence, they proposed a skin-depth model for active microwave response to soil moisture, defining skin depth as the reciprocal of attenuation coefficient (in terms of wavelength and dielectric constant). They also indicated that presently moisture dependence of soil dielectric properties has not been well defined.

Skin depth ranges from less than 1 cm for wet soils to several centimeters for dry soil (Figure 6). Ulaby, Cihlar, and Moore (1974) found reasonable correlation between radar backscat-

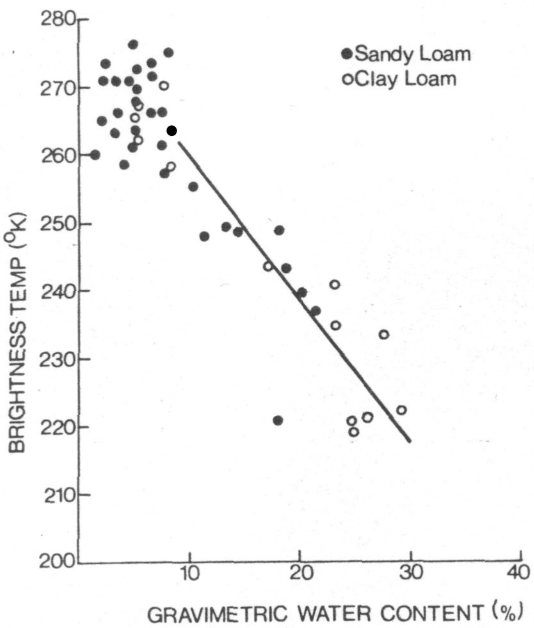

FIGURE 5. Plot of the 21-cm wavelength brightness temperatures versus soil water content. (The brightness temperature is the product of the temperature and emissivity of the surface.) (After Schmugge et al, 1974)

ter from bare soil and moisture content expressed in terms of the mean attenuation coefficient over a distance of one skin depth below the surface. Their results indicated that using radar techniques for determining soil moisture are promising but far from proven.

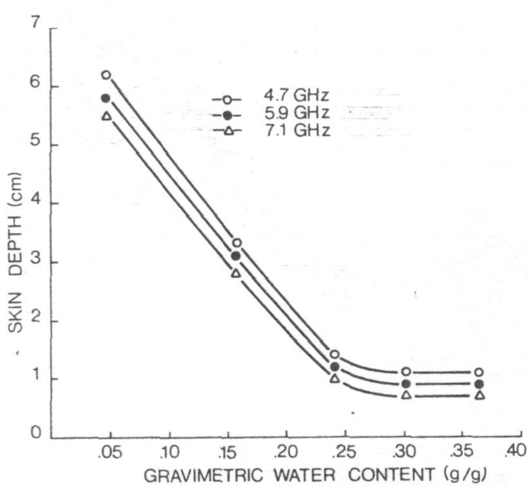

FIGURE 6. Skin depth as influenced by water content at incidence angle of 0. (After Ulaby, Cihlar, and Moore, 1974)

Radar techniques also have the potential for much better spatial resolutions, but unlike the capabilities of radiometric techniques, they have not been demonstrated from an aircraft platform. Also radar techniques would be independent of soil temperature, while the radiometric techniques may yield information on soil temperatures. However, both techniques respond to moisture variations in surface layers of the same thickness.

Thermal Inertia. The thermal inertia concept of measuring soil moisture, which relates the amount of soil water to the amplitude of the diurnal surface soil temperature wave, was evaluated by Idso et al (1975*b*). They found on an Avondale loam soil at Phoenix, Arizona, that the volume of water contained in the surface soil layer (2 to 4 cm thick) was a linear function of the amplitude of the diurnal surface-soil temperature wave for clear day-night periods and also of the daily maximum value of the surface soil-air temperature differential. Diurnal plots of measured surface soil (minus air-temperature differential) on selected clear day-night periods after the plots had been irrigated (Figure 7) showed that the amplitude of the temperature wave increased as the soil water decreased. Figure 8 shows the maximum value of the surface soil minus air-temperature differential as a function of volumetric soil water content.

With the thermal inertia concept, like some other methods, measuring the amount of water in the soil is difficult because different soils have unique relationships between their water content and the measured parameter. Thus different soils have different relationships between amplitude of diurnal temperature and volumetric water content. However, Idso et al (1975*b*) found that soils had a closer relationship between amplitude of diurnal surface soil temperature and pressure potential. They did not evaluate the dependence on climate of the amplitude of diurnal surface soil temperature wave.

Light Polarization. Another method for determining surface-soil water content is to measure light polarization when it is reflected from the soil surface. Any natural surface alters the polarizing properties of light reflected from it. Most dry surfaces reflect weakly polarized light, whereas light reflected from a wet surface may be nearly 100% polarized (Doll, 1973). Thus as water is added to dry soil, the percentage of light polarization increases steadily as it increasingly adopts the reflecting properties of the water. Doll (1973) found that at the optimum reflection angle (about 60°), polarization of the reflected light ranged from

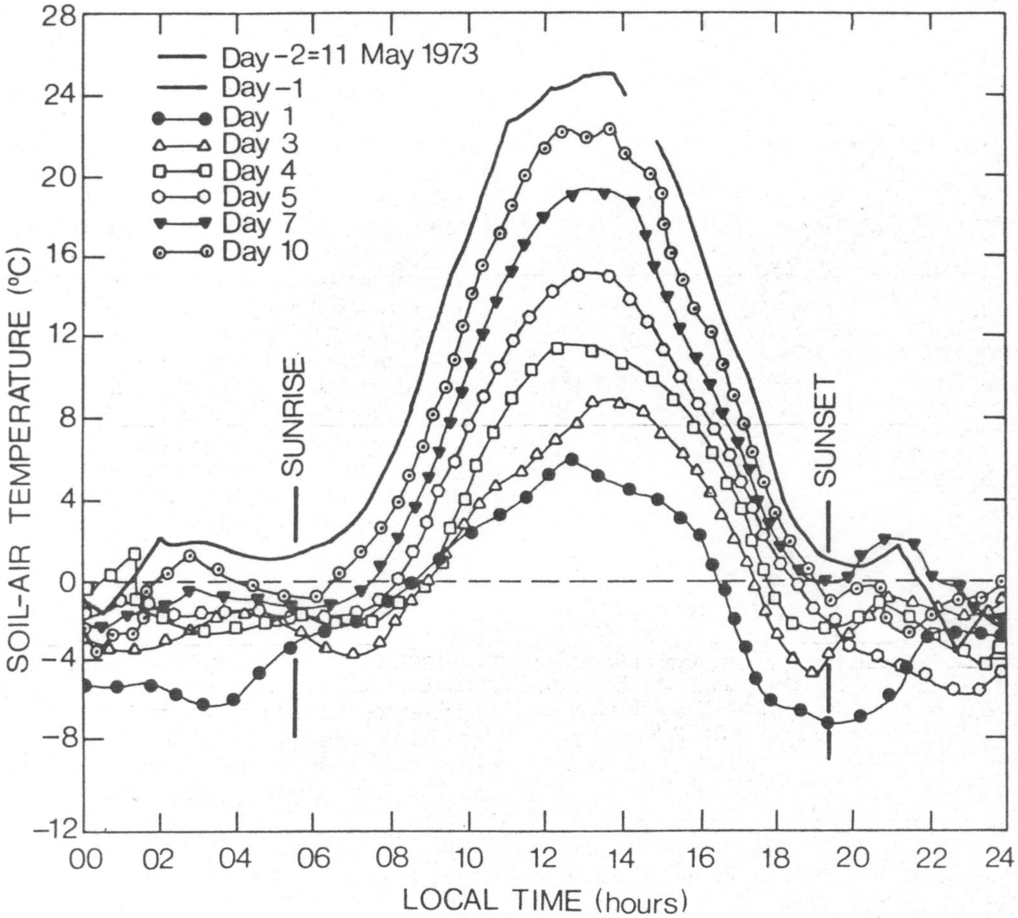

FIGURE 7. Diurnal plots of measured surface soil minus air-temperature differential on selected clear day-night periods after plots had been irrigated with 10 cm of water on day zero. (After Idso et al, 1975b)

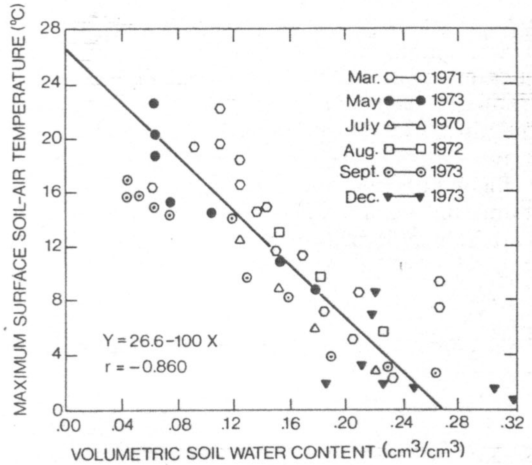

FIGURE 8. The maximum value of the surface soil minus air-temperature differential versus the mean daylight volumetric soil water content of 0- to 2-cm increments. (After Idso et al, 1975b)

15.5% for dry soil to 89% for saturated soil, and increased nearly linearly. Doll suggested that by using range-polarization values, the amount of water in the soil could be determined accurately.

EDWARD L. SKIDMORE

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Cross-references: *Aggregate Stability; Conductivity, Hydraulic; Energy Balance; Evaporation; Flow Theory; Heat Capacity; Infiltration; Pore Space; Spectral Characteristics; Thermal Regimes; Thermoluminescence; Water Content and Retention.*