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Survey of Methods for Soil Moisture Determination

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SURVEY OF METHODS FOR SOIL MOISTURE DETERMINATION

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ABSTRACT: With increased interest in soil moisture information for applications in such disciplines as hydrology, meteorology and agriculture, an overview is needed of both existing and proposed methods for soil moisture determination. This paper discusses the methods of in-situ soil moisture determination including gravimetric, nuclear, and electromagnetic techniques; remote sensing approaches that use the reflected solar, thermal infrared, and microwave portions of the electromagnetic spectrum; and soil physics models that track the behaviour of water in the soil in response to meteorological inputs (precipitation) and demands (evapotranspiration). The capacities of these approaches to satisfy various user needs for soil moisture information varies from application to application, but a conceptual scheme for merging these approaches into integrated systems to provide soil moisture information is proposed that has the potential for meeting various application requirements.

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INTRODUCTION

The moisture content in the surface layers of the soil is an important parameter for many applications in the disciplines of agriculture, hydrology and meteorology. In the field of agriculture the recent paper by Idso et al. (1975) describes the need for soil moisture information for improved yield forecasting and irrigation scheduling, among others. In hydrology, the moisture content of the soil's surface layer is important for partitioning rainfall into its runoff and infiltration components. In meteorology, soil moisture determines the division of net solar radiation into latent and sensible heat components. Recent model studies indicated the importance of soil moisture in such diverse phenomena as desertification (Charney, et al. 1977) and the central Florida sea breeze (Gannon, 1979).

The soil layer that we are considering in all these disciplines is that which can interact with the atmosphere through evapotranspiration (ET), i.e. the soil root zone. The moisture content of this soil layer fluctuates in response to precipitation and potential evapotranspiration (PET). The thickness of this layer will depend upon the type and stage of the soil's plant cover, but it is typically about 1 or 2 meters. Thus, we will call the moisture stored in this layer soil moisture. This moisture is only 0.005% of water on the earth's surface (Nace, 1964) but its seasonal variation accounts for a 1.4 cm variation in sea level (Mather, 1976).

In this paper we present a survey of the general methods for determining soil moisture. The three general approaches which we will consider are in-situ or point measurements, soil water models, and remote sensing.

IN-SITU METHODS

Several in-situ measurement techniques will be described in this paper -- gavimetric, nuclear, electromagnetic, tensiometric, and hygrometric.

Gravimetric Techniques

The oven-drying technique is probably the most widely used of all the gravimetric methods for measuring soil moisture, and it is often used to calibrate other soil moisture determination. This method consists of over-drying a soil sample at 105°C until a constant weight is obtained. Usually, this weight is obtained within 12 hours, but for large samples the drying time increases. The wet weight of the soil sample is taken before oven-drying. The amount of water in the sample can be determined and the moisture content calculated and expressed on a percentage by dry soil weight basis. If the volumetric water content is required, the gravimetric value is multiplied by the bulk density of the soil:

$$\theta = \frac{W_w}{W_d} \cdot \frac{Y_d}{Y_w} \cdot 100 \qquad (1)$$

where θ = volumetric water content (%)

W_w = weight of water (g)

W_d = dry weight of soil (g)

Y_d = oven-dry bulk density (g/cm^3)

Y_w = density of water (g/cm^3)

There are several advantages and disadvantages to the oven-drying gravimetric procedure. Some advantages are:

a. Samples can be taken with an auger or tube sampler.

b. Sample acquisition is inexpensive.

c. Soil moisture content is easily calculated.

Some disadvantages are:

a. Obtaining representative soil moisture values in a heterogeneous soil profile is difficult.

b. Because samples are required over long periods of time to monitor moisture movement or amount over time and space, this method can be very destructive to the site.

Additional information on this procedure, as well as most of the others discussed, can be found in Brakensiek et al (1974).

Nuclear Techniques

Neutron Scattering: The neutron scattering method is an indirect way of determining soil moisture content. This method estimates the moisture content of the soil by measuring the thermal or slow neutron density. Initial development of the neutron probe began in 1950 (Belcher et al., 1950, 1952). Gardner and Kirkham (1952) defined the principles on which the method is actually based. Neutrons with high energy (a million electron volts or more), are emitted by a radioactive source into the soil and are slowed down by elastic collisions with nuclei of atoms and become thermalized. The average energy loss is much greater from neutrons colliding with atoms of low atomic weight (in soils, this is primarily hydrogen), than from colliding with heavier atoms. As a result, hydrogen can slow fast neutrons much more effectively than any other element present in the soil. The density of the resultant cloud of slow neutrons is a function of the soil moisture content in the liquid, solid or vapor state. The number of slow neutrons returning to the detector per unit time over a known volume of influence or soil volume are counted, and the soil moisture content is determined from a standard curve of counts vs volumetric water content.

Two types of neutron probes have been developed. One is a depth probe that is lowered into the soil through an access tube to the depth at which the moisture content is desired. The other is a surface probe that gives the moisture content of the top few centimeters of soil.

Several sources of high energy neutrons have been used. The Americium-Beryllium (Am-Be) source seems to be the one most widely used (Bell and McCulloch, 1966). Older units used a radium-beryllium (Ra-Be) source. Van Bavel and Stirk (1967) found that this source eliminated gamma radiation, decreased the probe weight, increased the count rate and possibly increased the depth resolution of the soil moisture measurement.

The strength of the source varies with the type and manufacturer. Van Bavel (1962) found that 1 or 2 millicuries (mc) of a Ra-Be source were adequate. The strength of the source of Am-Be that Van Bavel and Stirk (1967) used was 150 mc. Others (Long and French, 1967; Bell and McCulloch, 1966) reported using Am-Be sources of 10, 30, 50, and 300 mc.

If subsurface measurements are required, the neutron probe must be placed in an access tube that is usually closed at the bottom. The size and composition of the tube offset the resultant neutron density (Stolzy and Cahoon, 1957). The most acceptable access tube seems to be made of aluminum. The method most used to install the tube in the field is to drill a slightly undersized hole and tamp the access tube into the drilled hole to ensure a tight fit.

The accuracy of the neutron probe can be determined from the deviation calculated by regression analysis where neutron counts are converted to volumetric moisture content (Visvalingam and Tandy, 1972). The calibration depends upon the source strength, the nature of the detector, the geometry of

the source and detector in the probe, the materials used to construct the probe, the size and composition of the access tube, and the physical and chemical properties of the soil (Wilson, 1971). Visvalingam and Tandy (1972), also found that vehicular ignition noise greatly influenced the neutron probe readings, likely by influencing the readout electronics, rather than the principle.

In laboratory calibration, the volume of soil used should be large enough to be considered effectively infinite relative to the neutron flux. Manufacturers of probes supply a generalized calibration curve with each unit. However, if an accurate moisture content determination is desired, the probe should be calibrated for each soil type. Procedures have been developed for laboratory and field calibration (Douglass, 1966; King, 1967; Luebs et al., 1968).

The moisture content value represents an average over a known volume of soil. Therefore, in laboratory calibration the soil used should be homogeneous in texture, structure, density and moisture content (Belcher et al., 1950; Douglass, 1966; Van Bavel, 1961, 1962). Field calibration of the neutron probe is reportedly extremely difficult (Lawless et al., 1963; Stewart and Taylor, 1957); (Rawls et al., 1973).

No matter what type of calibration is used, all electrical equipment has the potential to drift. Therefore, primary standards should be used to periodically recalibrate the probe. Various recalibration procedures have been reported (Churayev and Rode, 1966; Long and French, 1967; Marais and Smit, 1958, 1962; Holmes, 1966; Olgaard and Haahr, 1968; Luebs et al., 1968; Stone et al., 1966; Stewart and Taylor, 1957; Ursic, 1967; Bowman and King, 1965; Bell and Eeles, 1967).

The sphere of influence of the neutron probe measurement is the volume over which the average moisture content is calculated and depends on the amount of moisture in the soil. Van Bavel et al. (1956) and Glasstone and Edlund (1957) defined the sphere of influence as that volume which contains 95% of all the thermal neutrons. This concept has been criticized by Mortier et al. (1960) and Olgaard (1965). They suggest that the sphere of importance is the one which, if all the soil and water outside the sphere were removed, would yield a neutron flux at the source that is 95% of the flux obtained in an infinite medium.

The volume of soil measured is very important when measuring soil moisture with depth. In many studies, the diameter of the sphere of influence cannot be easily related to resolution because of the heterogeneity with soil depth due to pedogenesis or the depositional trends that occurred over a long geologic time period. The vertical resolution is critical to many studies, especially those dealing with monitoring soil moisture in time and space.

The advantages of the neutron probe are:

a. Moisture can be measured regardless of its physical state.

b. Average moisture contents can be determined with depth.

c. The system can be interfaced to accommodate automatic recording.

d. Temporal soil moisture changes can be easily monitored.

e. Rapid changes in soil moisture can be detected.

f. Readings are directly related to soil moisture.

The disadvantages are:

a. Inadequate depth resolution makes measurement of absolute moisture content impossible and limits its use in studying evaporation, infiltration, percolation and placement of the phreatic water surface.

b. The moisture measurement depends on many physical and chemical properties of the soil which are, in themselves, difficult to measure.

c. Care must be taken to minimize health risk.

d. The sphere of influence of the depth probe does not allow for an accurate measurement of soil water at or near the soil surface.

Stone et al. (1966) stated that the accuracy of neutron probe measurements exceeds that of standard techniques, but Stewart and Taylor (1957) argued that it is slightly inferior. If the neutron probe is used, the purchaser should look for a stable, portable, durable model with stable electronics and power components, compatible with available equipment (Bell, 1969; Bell and McCulloch, 1966; Zuber and Cameron, 1966).

Gamma-Ray Attenuation: The gamma-ray attenuation method is a radioactive technique that can be used to determine soil moisture content value within a 1 to 2 cm soil layer. This method assumes that scattering and absorption of gamma-rays are related to the density of matter in their path and that the specific gravity of a soil remains relatively constant as moisture content increases or decreases changing wet density. Changes in wet density are measured by the gamma transmission technique and the moisture content determined from this density change.

Gamma-rays may be collimated to a narrow beam, which permits obtaining a representative reading at any position in the soil. The method first became known in the early 1950's. Work by Gurr (1962), Ferguson and Gardner (1962), Davidson et al. (1963), and Dmitriyev (1966) was instrumental in developing the theoretical basis and procedure for its use.

The basic equipment includes a gamma source surrounded by a collimator, a detector with a collimator, and a scaler. Gurr (1962) used a 25-mc cesium 137 source with a lead collimator, with the beam emerging from a circular hole 4.8 mm in diameter. A scintillation counter was used as a detector, shielded by a lead collimator containing a 12.5-mm-diameter hole. Mansell et al. (1973) stated that collimated radiation from 300 mc each of ^{241}Am and ^{137}Cs provided a high intensity beam comprising 60 and 662 KeV photons. Count rates measured by a single detector and a two-channel gamma spectrometer were corrected for coincidence losses due to pulse-resolving time. They concluded that error in soil water content measurement by the dual energy gamma attenuation method will probably not exceed a standard deviation of 1%.

The gamma ray attenuation technique has the same advantages as items b, c and d listed under neutron meters, as well as the following:

a. Data can be obtained over very small horizontal or vertical distances.

b. The measurement is nondestructive.

Its disadvantages are:

a. Large variations in bulk density and moisture content can occur in highly stratified soils and limit spatial resolution.

b. Field instrumentation is costly and difficult to use.

c. Extreme care must be taken to ensure that the radioactive source is not a health hazard.

Soane (1967) and Corey et al. (1971) also used dual energy, collimated beam gamma-rays to simultaneously measure density and moisture contents of soil columns. Others who have investigated the technique include Gardner and

Roberts (1967) and Gardner et al. (1972). In their studies they used two collimated beams of monoenergetic gamma-rays from ^{241}Am and ^{137}Cs but moved the soil column from one beam to another. In their study, the error in Y_d and θ resulted from the randomness of the emission from the sources, random error in attenuation coefficients and soil column thickness measurements, presence of a small higher energy peak in the ^{241}Am spectrum, and counting dead time.

Goit et al. (1976) showed experimentally that the variability due to differences in Y_d and θ of a soil within the beam of a dual-energy system caused large measurement errors. Nofziger (1978) concluded from his studies that, indeed, large error in the measurement of Y_d and θ can occur in highly stratified material when using the dual gamma technique. Generally, small errors occur if Y_d and θ change linearly in the collimated beam. He also confirmed that both the dual gamma and single systems accurately measure the average water content in the collimated beam if the bulk density of the soil is constant. However, the average water content in the beam may not represent the water content at the middle of the collimated beam and in the middle of the present time period. From this study, graphs were prepared to estimate the error due to inhomogeneity of the soil.

A major problem in many cold regions is the inability to measure in situ water conditions in the freezing, thawing or frozen state. Goit et al. (1976) conducted studies to evaluate attenuation of a dual gamma beam and found that it was a powerful technique for investigating the swelling phenomena associated with freezing soil. They found that errors resulted when attenuation equations developed for homogeneous mixtures were applied to

stratified media. Nofziger (1978) determined that the errors in θ and Y_d due to nonuniform soil systems must be considered to establish the overall accuracy of gamma ray measurements.

Since attenuation of gamma rays is independent of the state of the water in the material tested, the measurement of attenuation is unaffected by the transition of liquid water to ice. Therefore, the use of gamma attenuation has an advantage since measurements of dry bulk density and total water content (including ice), in grams per cubic centimeter can be made simultaneously.

In-Situ Electromagnetic Techniques

Electromagnetic techniques include those methods which depend upon the effect of moisture on the electrical properties of soil. The magnetic permeability of soils is very nearly that of free space and, hence, the approach reduces to methods of exploiting the moisture dependence of the dielectric properties of soil.

The dielectric properties of the moist soil may be characterized by a frequency dependent complex dielectric response function (Bottcher, 1952):

$$\xi(w) = \xi_r(w) = j\xi i(w) \qquad (2)$$

where

$\xi_r(w)$ = the real part of ξ

$\xi_i(w)$ = the imaginary part of ξ

i = square root of -1

and w is the (angular) frequency.

The function $\xi_r(w)$ is about constant from $w=0$ out to the neighborhood of the relaxation frequency w_R of dipoles in the medium. The time w_R^{-1} is the time constant for the decay of polarization, when the electric field

is removed. Beyond w_R, the function ξ_r decreases until in the visible region of the spectrum, and it is equal to the index of refraction squared. The real part of the dielectric response function is a measure of the energy stored by the dipoles aligned in an applied electromagnetic field. When the frequency is greater then w_R, the dipoles can no longer follow the field and the ability of the medium to store electric field energy decreases.

The function $\xi_i(w)$ is a measure of the energy dissipation rate in the medium. Viewed as a function of frequency, and starting from low , it rises to a peak at w_R and, thereafter, decreases. The behavior described is due to the permanent dipoles in the soil medium. In complicated heterogeneous media, there may be more than one relaxation mechanism and more than one absorption peak. Furthermore, at frequencies above, the medium may show further dispersion and absorption regions due to direct molecular excitations. The frequency w_R will generally lie in the microwave range (18 Ghz in H_2O), whereas the latter molecular excitations will be in the submillimeter or infrared regions of the spectrum (Bottcher, 1952; Hasted, 1974). In soils ξ_R is reduced to around 1 GH_z due to the binding of the water molecules to the soil particles (Hoekstra and Delaney, 1974).

The preceding description generally applies to all dispersive media. In a soil, the values of ξ_r are typically between 3 and 5, whereas the value of ξ_r for water is about 80. Hence, relatively small amounts of free water in a soil will greatly affect its electromagnetic properties. This dependence is shown in Figure 1, which presents the results of laboratory measurements at wavelengths of 21 and 1.55 cm (frequencies of 1.4 and 19.4 Ghz). The wavelength dependence is due to the difference in the dielectric properties of water at these two wavelengths.

At low wavelength levels there is a slow increase with soil moisture, but, above a certain point, the slope of the curve sharply increases which is due to the behavior of the water in the soil. When water is first added to a soil it is tightly bound to the soil particles. In this state the water molecules are not free to become aligned and the dielectric properties of this water resemble those of ice. As the layer of water around the soil particle becomes larger, the binding to the particle decreases and the water molecules behave as they do in the liquid; hence, the greater slope at the higher soil moisture values. The transition moisture depends on the soil texture, i.e., particle-size distribution is less for a sand and larger for a clay and has been found to be linearly dependent on the wilting point for the soil. This effect has been demonstrated in laboratory measurements of the dielectric constant (Lundien, 1971; Newton, 1977).

This dependence of the dielectric properties of a soil on moisture content can be used for either an in situ sensor or a remote sensor. In this section in situ devices, measuring either soil resistivity or capacitance, will be discussed; the remote sensor approaches will be presented later.

A variety of implantable sensors, responsive either to resistivity (ξ_i), polarization (ξ_r), or to both have been constructed (Wexler, 1965; Roth, 1966; Thomas, 1963; Gagne and Outwater, 1961; DePlater, 1955; Silva et al., 1974). Traditionally, these have been designed for operation at frequencies below 1MHz. Recently, however, due to a steady decrease in the physical size of high quality, high frequency components, implantable sensors have become a practical reality (Selig et al., 1975; Walsh et al., 1979; Layman, 1979; Wobschall, 1978).

The resistivity of soils depends on moisture content and, hence, can serve as the basis for a sensor. It is possible either to measure the resistivity between electrodes in a soil or to measure the resistivity of a material in equilibrium with the soil. Sensors of either kind can be very compact and an array of them can be connected to standard data collection platforms. The difficulty with resistive sensors is that the absolute value of soil resistivity depends on ion concentration as well as on moisture concentration (Bouyoucos and Mick, 1948). Therefore, careful calibration is required for this technique. Even with careful calibration, the instrument may require frequent recalibration due to changes in organic or salt concentrations. The calibration problem becomes less severe as the operating frequency is increased, since the relative contribution of ion motion decreases.

Implantable sensors, which are sensitive to polarization, ϵ_R, in essence measure capacitance (Thomas, 1963; Gagne and Outwater, 1961; DePlater, 1955: Selig et al., 1975; Walsh et al., 1979; Layman, 1979). This parameter is the electrical quantity which is the most direct indicator of moisture concentration. When the moisture held in the soil can be regarded as free, as it is in most sandy soils, the relationship between ϵ_r and moisture is linear. Furthermore, even in more complicated materials, where the water is relatively tightly bound, such as a montmorillonitic clay, it is possible to determine moisture content by measuring the capacitance of an implanted sensor. Because of this, several capacitive sensors have been constructed. Most of these have been designed for operation below 1 MHz, although more recently some work has also been done up to 100 MHz (Wobschall, 1978; Walsh et al., 1979; Layman, 1979; Silva et al., 1974). The motivation for increasing the operation frequency is again to minimize the contribution of

ionic conductivity, which if it is large, can make accurate measurement of the capacitance difficult. One other promising technique is to work at 10 to 100 MHz frequency, and to utilize a bridge technique that allows a determination of both ε_r and ε_i. These can be used separately as moisture indicators (Walsh et al., 1979; Layman, 1979).

The main advantages of either resistor or capacitor type devices are that they are capable in principle of providing absolute values for soil moisture with calibration, and they can be implanted at any depth. This means that moisture profile data can be obtained by this method. A wide variety of sensor configurations varying from very small to quite large are possible and, hence, there is some control over the sensor volume of influence. The precision of both the resistive and capacitive sensors is high. The first of these is also relatively accurate when other parameters are adequately controlled, whereas the second has a relatively high intrinsic accuracy which is more nearly independent of parameters other than moisture. This follows from the fact that the capacitive sensors are directly responsive to the amount of polarized energy stored in the region of the sensor and this quantity is normally dominated by the water present.

The moisture sensor must be implanted properly to minimize disturbances to the soil. In addition, there are questions of long term reliability, maintenance of the calibration, and the interface with remote collection platforms. Overall, it would seem that the relative advantages in some applications would warrant serious consideration of the implantable sensors.

Tensiometric Techniques

The term "tensiometer" was used by Richards and Gardner (1936) as an unambiguous reference to the porous cup and vacuum gauge combination for

measuring capillary tension or the energy with which water is held by the soil. However, tensiometers were used to measure soil water tension in unsaturated soils as early as 1922 (Gardner et al., 1922). Richards (1949) and others have made extensive developments and improvements in the tensiometers used in the field and laboratory soil water studies.

The energy term can be expressed as pF which is defined as the common log of the height of a water column in centimeters equivalent to the soil moisture tension, or it can be expressed as a suction (negative pressure) or a potential (energy per unit mass). Elrick (1967) recognized six components of the total energy of soil water, of which matric suction is one. Matric suction is the pressure difference across a boundary permeable only to water and solutes, which separates bulk water and soil water in hydraulic, chemical and thermal equilibrium. Dissolved salts or chemicals in the soil water contribute to solute suction. Baver et al (1972) suggested using the term "capillary potential" to denote the total potential, which includes not only surface tension forces but also the osmotic and adhesion forces.

The most widely known method for measuring the capillary or moisture potential is based upon the so-called suction force of the soil for water (Baver et al., 1972, Richards, 1965). Tensiometers are used to measure the suction and consist of a liquid- (usually water) filled porous ceramic cup connected by a continuous liquid column to a manometer or vacuum gauge. In some designs, the liquid is an ethylene glycol-water solution and the measuring gauge a transducer with electrical output. The transducer output can be interfaced to near real time data acquisition systems. The use of an ethylene glycol-water solution as a replacement for water in the tensiometer allows the use of a tensiometer/transducer system in cold areas

(McKim et al., 1976). Since the tensiometer/transducer system has a millivolt output, it is well suited for automatic (including satellite relay) recording systems (McKim et al., 1975; Elzeftawy and Mansell, 1975; Gillham et al., 1976).

Essential steps in the technique include de-airing the water or solution in the tensiometer, placing the tensiometer system in the soil, and allowing it to come to equilibrium with the soil water. The ceramic cup is porous to water and solute but not to air, so that water can flow, and soil water conditions or change in moisture content can be determined. Basically, tensiometers measure the curvature of the water meniscus in the pores of the ceramic cup which, at equilibrium, is related to the force with which the water is held by the soil. As the soil water content increases, it is held at a lower tension; when the tensiometer reads zero, the soil is saturated, and there is zero water tension. The highest tension reading that can be obtained with a tensiometer is about 1 bar (1 atmosphere). In most instances, data cannot be obtained beyond 0.8-0.9 bar because the air entry value of the ceramic cup is exceeded. This means that the moisture content range over which the tensiometer can be used is limited. Richards (1949) stated that for coarse, sandy soils the range of the tensiometer may cover more than 90% of the available moisture content range. Clay soils pose a different problem. For example, for soils containing over 42% montmorillonite clay, the tension can change from 200 to 800 cm H_2O with a 1% change in volumetric water content (Abele, 1979).

Soil moisture measurement procedures using tensiometer/transducer systems are ways to monitor water movement in the field. Recent studies by Bianchi (1962), Klute and Peters (1962), Watson (1967), Rice (1969), Anderson and Burt, (1977) have shown the advantages of using pressure transducers to

produce a fast response, low volume displacement tensiometer system. These types of systems are capable of monitoring moisture movement that occurs rapidly in infiltration, irrigation, groundwater recharge, and evapotranspiration.

Tensiometers have been used for years to measure soil tension. During recent years advancements in system design and performance have made possible the implementation of soil moisture field monitoring programs. However, care still needs to be taken in assessing the use of the system. Listed below are some of the advantages and disadvantages of using tensiometers.

Advantages

a. Systems are easy to design and construct.

b. The system costs relatively little.

c. Information can be obtained on moisture flow under saturated and unsaturated conditions in near real time.

d. The tensiometer can usually be placed in the soil easily and usually with minimal disturbance.

e. The system can operate over long time periods.

f. Using the tensiometer/transducer system the response time is very rapid.

g. Different types of liquids can be used like ethylene glycol solution to obtain data during freezing and thawing conditions.

h. Systems can use with $\pm$ reading tensiometers to read both water table elevation and soil moisture tension, depending on the soil water status.

Disadvantages

a. The tensiometer can be broken easily during installation.

b. Can only be determined within the 0 to 800-cm water tension range.

c. Field installations drift electronically.

Hygrometric Techniques

The relationship between moisture content in porous materials and the relative humidity (RH) of the immediate atmosphere is reasonably well known. Therefore, several relatively simple apparatus for measuring RH have been designed. Basically, the sensors can be classified into seven types of hygrometers -- electrical resistance, capacitance, piezoelectric sorption, infrared absorption and transmission, dimensionally varying element, dew point, and psychrometry.

Electrical resistance hygrometers utilize chemical salts and acids, aluminum oxide, electrolysis, thermal, and white hydrocol to measure RH. Bouyoucos and Cook (1965) considered the white hydrocol hygrometer as the best available. The measured resistance of the resistive element is a function of RH. They stated that casting the stainless steel electrodes in white hydrocol (a form of plaster of Paris) causes greater accuracy because the cement sets hard, is pure, has a low solubility and contains no added salts.

Phene et al (1971, 1973) developed a heat dissipation sensor that was used to measure the soil moisture potential. The accuracy of the matrix potential sensor proved to be as good as or better than that of the thermocouple psychrometer or salinity measurements. The sensor, which had high sensitivity

in the 0 to -2 bar matrix potential range, had an accuracy of $\pm$ 0.2 bar. The accuracy decreased progressively to $\pm$1 bar at a matrix potential of -10 bar.

The primary advantages of using the hygrometer techniques are the simplicity of the apparatus and the low cost. Basic disadvantages include the soil components deteriorating the sensing element and the special calibration required for each material to be tested. The main use for this technology seems to be in applications where RH in the material is directly related to other properties. One example would be drying and shrinkage of cements.

SOIL WATER MODELS

Recent developments of soil water models based on column mass balance gives us an alternative to directly or indirectly measuring of soil moisture in the field. Figure 2 is a schematic diagram of the physical system and driving forces that must be considered in modeling the system.

Based upon conservation of mass, the soil moisture in the system at any time can be determined using the relationship

$$SM_t = SM_{t-1} + P - R + L + E - T + C - Q \qquad (3)$$

where:

SM_t = soil moisture volume at time t

SM_{t-1} = soil moisture volume at previous time

P = precipitation

R = surface runoff

L = net lateral subsurface flow

E = evaporation or condensation

T = transpiration

C = capillary rise from lower levels

Q = percolation

This generalized model represents only a single column that is homogeneous horizontally at all levels. Actual systems will be heterogeneous. Heterogeneous systems can be represented by spatial averages or by linking columns to account for the spatial variability.

Published soil moisture models vary in the level of detail they use in representing the physical system and the temporal variations of the driving forces. Some of the important differences between models are listed below:

- Method used for computing the potential evapotranspiration.
- Method used for computing infiltration and runoff.
- Temporal definition of evaporative demand and precipitation.
- Consideration of saturated and unsaturated layers.
- Number of soil layers used.
- Method used for computing soil evaporation and plant transpiration.
- Consideration of the thermal properties of the soil system.

Many of the published models which simulate soil moisture were developed for agricultural applications. A very simple model, described by Holmes and Robertson (1959), treats the soil as a single homogeneous layer. Potential evapotranspiration is computed empirically and the actual evaporation is set equal to this as long as moisture is available. All precipitation becomes infiltration and groundwater interactions are ignored. All computations are performed on a daily basis.

Jensen et al. (1971) developed an irrigation scheduling model that takes into account the soil moisture. Evapotranspiration is computed on a daily basis using one of several alternative procedures. Actual evapotranspiration is not affected by moisture deficits. Infiltration must be computed externally. Percolation is computed using an empirical relationship. In this model, the soil is treated as a single layer.

Holmes and Robertson (1959) presented and described another model, the Modulated Soil Moisture Model which is slightly more sophisticated than those described above. This model utilizes a two layer soil system and considers the fact that actual evaporation will generally not equal the potential due to moisture deficits. The two are set equal until the moisture in the upper zone is depleted. Thereafter, moisture is extracted from the lower zone at a reduced rate proportional to the moisture level. This model also allows a simple runoff computation.

Baier and Robertson (1966) improved on these models with one called the Versatile Soil Moisture Budget Model. In this model the soil is divided into several layers and the available water for each is taken to be the difference between its field capacity and wilting point. Evapotranspiration can occur simultaneously from each layer and depends on the soil moisture present and the particular soil and plants involved, which are represented by coefficients. Flow between layers is considered; however, the technique used is empirical as is the procedure used for computing infiltration.

Saxton et al (1974) developed a much more comprehensive model that will simulate soil-plant-atmosphere-water systems in greater detail than the models described above. A flow chart of the model is shown in Figure 3. As illustrated, this model considers all of the factors influencing the system. Some processes, like soil moisture redistribution, are modeled using a physics-based approach whereas others, like plant transpiration, are semiempirical.

Many other models have been developed which resemble those mentioned above. Additional information can be found in Saxton and McGuinness (1979),

Feddes et al. (1978), Hildreth (1978), Singh (1971), Kanemasu et al. (1976), Stuff and Dale (1978), Shaw (1963), Goldstein et al (1974), Ritchie (1972), and Youker and Edwards (1969).

All of these models were developed primarily for agricultural applications. Hydrologists have also developed water balance models that include a soil moisture component. State-of-the-art examples of the approaches used in hydrologic modeling can be found in the U. S. Department of Agriculture Hydrograph Laboratory (USDAHL) Model (Holtan et al., 1975) and the National Weather Service River Forecast Model (NWSRFS) (Peck, 1976). Other models are reviewed in Fleming (1975). Most of these models will perform a continuous simulation of the volumes and rates of water movement occuring in each component of the watershed.

In the USDAHL Model, illustrated in Figure 4, the spatial variability of soils and vegetation is accounted for by using zones within which the hydrologic parameters are averaged. Within each zone the soil is subdivided into several homogeneous layers determined from hydraulic properties. Evapotranspiration is computed daily using an empirical equation which considers the crop and soil characteristics, as well as the current soil moisture. Evapotranspiration is drawn from the first two layers, which are considered to be the root zone. These computations are performed daily. Infiltration is also based on soil and crop characteristics and the current soil moisture. A 1-hour time step is used for these computations. The procedure used for soil moisture redistribution and percolation only considers gravity flow.

In the NWSRFS Model, illustrated in Figure 5, two zones are used to simulate soil water storage and movement. The upper layer is that which responds quickly to rainfall and controls overland flow. It is usually very

shallow. The lower layer is the balance of the soil column extending to the water table. Soil hydraulic properties are averaged within each layer. Moisture is stored as either tension or free water. Infiltration, percolation and soil moisture redistribution involve the free water and are computed using empirical equations that use as a controlling factor the ratio of the free water present to the field capacity of the layer involved . Evapotranspiration is also computed using an empirical procedure. Actual evapotranspiration is set equal to the potential until all moisture in the upper layer is depleted. When this occurs, moisture is extracted from the lower zone using an equation which considers the moisture deficit and the crop characteristics. A 6-hour time step is used for simulation.

Most of the models mentioned in preceding sections were developed for practical application to such problems as crop yield estimation, irrigation planning, and runoff forecasting. They were developed to use readily available meteorological data for inputs and, therefore, they usually use a 1-day time step. Goldstein et al (1974) pointed out that models which employ a 1-day averaging will produce accurate weekly average results; however, the daily results will show some deviation on any given day. Jensen et al (1971) made the same point. They noted that expected daily errors between 10 and 15 % should become negligible over 10 to 20 days.

In some situations, especially research, information of greater time resolution and accuracy than can be provided by these models is required. Several complex models capable of simulating soil-plant-atmosphere-water systems have been proposed. Generally, these models are more physically based than the models described above. Increased detail is usually costly and these complex models usually have high data and computer requirements, especially if simulation over extended periods is desired.

Much research has been done on the physics of soil water movement and storage under bare soil conditions. These models usually involve solving the equations describing one dimensional vertical unsaturated flow and horizontal saturated flow. Most of the published models differ on the boundary conditions imposed and the numerical approximations used for solution.

Hillel (1977) described several physically based models designed to simulate soil water conditions under bare soil conditions. All of the models were programmed using a versatile simulation language called Continuous Simulation Modeling Programs (CSMP) (Speckhart and Green, 1976). Other examples of this type of model are presented in Hanks et al. (1969), Warrick et al. (1971), Bresler (1973), Remson et al. (1971), and Pikul et al. (1974).

Some progress has been made at linking these one dimensional models to represent spatially varied systems. Figure 6 is a schematic of one such model presented by Hillel (1977). Others, such as that of Freeze (1978), have extended the solutions to two-dimensional problems.

Including the effects of plants in detailed models greatly increases their complexity. Published models that simulate these effects include those presented by Van Bavel and Ahmed (1976), Lemon et al. (1973), Makkink and vanHeemst (1975), Hanson (1975), Slack et al. (1977), Feddes et al. (1976), Neuman et al. (1975), and Nimah and Hanks (1973).

The reader can gather from the number of references cited in this section that there are a multitude of models which have been developed and documented. A comprehensive description of each, along with their respective advantages and disadvantages, is beyond the scope of this paper. In general, a model can be found in the literature that is adaptable to almost every problem. The principle advantage of models is that they can provide soil

moisture information on a timely basis without the necessity of field visits. A general disadvantage to models is the error of their estimates.

REMOTE SENSING APPROACHES

The remote sensing of soil moisture depends upon the measurement of electromagnetic energy that has either been reflected or emitted from the soil surface. The variation of the intensity of this radiation with soil moisture depends on either the dielectric properties (index of refraction), its temperature or a combination of both. The particular property that is important depends on the wavelength region that is being considered as indicated in Table 1.

Table 1

Electromagnetic Properties for Soil Moisture Sensing

Wavelength region	Property observed
Reflected solar	soil albedo/index of refraction
Thermal infrared	surface temperature
Active microwave	backscatter coefficient/ dielectric properties
Passive microwave	microwave emmision/dielectric properties & soil temperature

The use of reflected solar energy is not a very promising tool for soil moisture determination because the soil spectral reflectance vs water content relationship depends on several other variables, like spectral reflectance of the dry soil, surface roughness, geometry of illumination, organic matter

and soil texture (Jackson, et al 1978). These complicating factors plus the fact that it responds to only a thin surface layer limit the utility of solar reflectance measurements for soil moisture determinations, and thus will not be discussed further here.

Water is unique in that it is near the extremes in its thermal and dielectric properties. As a result, the corresponding properties in the soil are highly dependent on its moisture content. These properties are accessible to remote sensing through measurements at the thermal infrared (10μm) and microwave (1cm to 50cm) wavelengths. The approaches are (Schmugge, 1978):

1. Thermal Infrared

   Measurement of the diurnal range of surface temperature

   Measurement of the crop canopy temperature

2. Microwave

   Active: Measurement of the radar backscattered coefficient

   Passive: Measurement of the microwave emission or brightness temperature

Thermal methods

The amplitude of the diurnal range of surface temperature for the soil is a function of both internal and external factors. The internal factors are thermal conductivity (K), and heat capacity (C), where $P = (KC)^{1/2}$ defines what is known as "thermal inertia." The external factors are primarily meteorological--solar radiation, air temperature, relative humidity, cloudiness, and wind. The combined effect of these external factors is that of the driving function for the diurnal variation of surface temperature. Thermal inertia then is an indication of the soil's resistance to this driving

force. Since both the heat capacity and thermal conductivity of a soil increase with an increase of soil moisture, the resulting thermal inertia will increase.

A complicating factor is the effect of surface evaporation in reducing the net energy input to the soil from the sun. Evaporation complements the other effects of water in soil by reducing the amplitude of the surface diurnal temperature cycle. As a result the day-night temperature difference is an indicator of some combination of soil moisture and surface evaporation.

The basic phenommena has been studied in experiments at the U.S. Water Conservation Laboratory in Phoenix, Arizona, where soil temperatures were measured with a thermocouple versus time, before and after irrigation (Idso, et al., 1975). They observed a dramatic difference in the maximum temperatures before and after an irrigation. On succeeding days the maximum temperature increases as the field dries out.

The summary of results from many such experiments is shown in Figure 7 where the amplitude of the diurnal range is plotted as function of the soil moisture as measured at the surface and at 0 to 1, 0 to 2, and 0 to 4 cm layers. There is a good correlation with the soil moisture in the 0 to 2 and 0 to 4 cm layers of the soil, and this response is related to the thermal inertia of the soil. Initially, when the surface is moist, the temperatures are more or less controlled by evaporation. Once the surface layer dries below a certain level, the temperature will be determined by the thermal inertia of the soil. These results indicate that for this particular soil, the diurnal range of surface temperature is a good measure of its moisture content in the surface layer (2 to 4 cm) layer.

These temperature measurements were repeated for different soil types that ranged from sandy or light soils to heavy clay soils. It is clear that for a given diurnal temperature difference, there can be a wide range of moisture content for these soils (Idso et al., 1975).

However, the T values observed as the soils dried through the transitions between the stages of drying were about the same for all of the soil types studied. Thus, they concluded (Idso et al., 1975) that the relationship between T and moisture content depends on soil type. The relationship between T and pressure potential (the tension with which water is held by soil particles) is independent of soil type. This is the basis for expressing moisture values as a percent of field capacity (FC), where field capacity is the moisture content at the -1/3 bar pressure potential.

It should be emphasized that these experiments were all made in the field, using thermal-couples, and were not remotely sensed. In March 1975, an experiment was performed in which remotely sensed thermal infrared temperatures from an aircraft platform were compared with the in situ thermocouple measurements over a 5-day period. Figure 8 presents the results from both the field experiments (from Figure 7) and the aircraft experiments (Reginato et al., 1976; Schmugge et al., 1978). The field results are expressed as a percent of field capacity so they can be compared with the aircraft results obtained over a wide range of soil textures. There was good agreement between the thermocouple measurements and the remotely sensed radiation measurements made from the aircraft, indicating that the conclusions based on the thermocouple field measurements would also be valid for radiation temperature observations.

This technique is not applicable to fields with a vegetative canopy. However, the difference between canopy temperature and ambient air temperature has been shown to be an indicator of crop stress. Thus, a cropped surface is viewed, if the vegetation is reflecting the soil moisture status, a potential exists for monitoring effective soil moisture over the rooting depths of the particular crop. Following this argument, Jackson et al. (1977) established that a running sum of daily values called "Stress Degree Days" (SDD) can potentially be used for irrigation scheduling. Millard et al. (1977b) confirmed feasibility of this approach for fully grown wheat on the basis of airborne data. Similarly, stress degree days have been successfully correlated with the yield of wheat (Idso et al., 1975).

Microwave Methods

As discussed previously, the dielectric properties of a soil are strong functions of its moisture content. Since the dielectric properties of a medium determine the propagation characteristics for electromagnetic waves in the medium, they will effect the emissive and reflective properties at the surface. As a result, these latter two quantities for a soil will depend on its moisture content, which can be measured in the microwave region of the spectrum by radiometric (passive) and radar (active) techniques. This physical relationship between the microwave response and soil moisture plus the ability of the microwave sensors to penetrate clouds makes them very attractive for use as soil moisture sensors. In the following sections results will be presented demonstrating this sensitivity to soil moisture, along with a discussion of some of the noise factors, e.g. vegetation and surface roughness, which affect the relationship between the sensor response and soil moisture.

Passive microwave

A microwave radiometer measures the thermal emission from the surface and at these wavelengths the intensity of the observed emission is essentially proportional to the product of the temperature and emissivity of the surface (Rayleigh-Jeans approximation). This product is commonly referred to as brightness temperature. All our results will be expressed as brightness temperatures (T_B). The value of T_B observed by a radiometer at a height h above the ground is

$$T_B = \tau \left(r\, t_{sky} + (1 - r)\, T_{soil} \right) + T_{atm} \qquad (4)$$

where r is the surface reflectivity and the atmospheric transmission. The first term is the reflected sky brightness temperature, which depends on wavelength and atmospheric conditions; the second term is the emission from the soil (1-r=e, the emissivity); and the third term is the contribution from the atmosphere between the surface and the receiver. At the longer wavelengths, i.e., those best suited for soil moisture sensing, the atmospheric effects are minimal and will be neglected in this discussion.

Thermal microwave emission from soils is generated within the soil volume. The amount of energy generated at any point within the volume depends on the soil dielectric properties (or soil moisture) and the soil temperature at that point. As energy propagates upward through the soil volume from its point of origin, it is affected by the dielectric (soil moisture) gradients along the path of propagation. In addition, as the energy crosses the surface boundary it is reduced by the effective transmission coefficient (emissivity), which is determined by the dielectric characteristics of the soil near the surface.

The emission from the soil surface can be expressed as:

$$T_B = e \int_{-\infty}^{0} T(z) \alpha(z) \exp\left(-\int_{z}^{0} \alpha(z')dz'\right)dz \qquad (5)$$

where $T(z)$ is the temperature profile and $\alpha(z)$ is the absorptivity as a function of depth which depends on moisture content. Results from numerical solutions to this equation have been presented by Njoku and Kong (1977) Wilheit (1978) and Burke et al. (1979). These papers have included results which indicate that the models do a good job of predicting T_B for a smooth surface. One of the most significant results from these models is that the effective sampling depth is on the order of only a few tenths of a wavelength (Wilheit, 1978). Thus, for a 21-cm-wavelength radiometer this is about 2 to 5 cm.

The range of dielectric constant presented in Figure 1 produces a change in emissivity from greater than 0.9 for a dry soil to less than 0.6 for a wet soil, assuming an isotropic soil with a smooth surface. This change in emissivity for a soil has been observed by truck mounted radiometers in field experiments (Poe et al., 1971; Blinn and Quade, 1972; Schanda et al., 1978; Newton, 1977) and by radiometers in aircraft (Schmugge, 1974; Burke et al1978; Choudhury, et al, 1979) and satellites (Eagleman and Lin, 1976; Schmugge et al, 1977). In no case were emissivities as low as 0.6 observed for real surfaces. This is primarily due to the effects of surface roughness which generally has the effect of increasing the surface emissivity.

As can be seen in Figure 1, there is a greater range of dielectric constant for soils at the 21-cm wavelengths. This fact, combined with a larger soil moisture sampling depth and better ability to penetrate a vegetative canopy, makes the longer wavelength sensors better suited for radiometric soil-moisture sensing.

In Figure 9a and b, the field measurements of Newton (1977) are plotted versus angle of observation for various moisture contents and for three levels of surface roughness. The horizontal polarization is that for which the electric field of the water is parallel to the surface and the vertical polarization is perpendicular to it. These results indicate the effect of moisture content on the observed values of T_B and the effect of surface roughness which is to increase the effective emissivity at all angles and to decrease the difference in T_B for the two polarizations at the larger angles.

For the smooth field there is a 100 K change in T_B from wet to dry soils and clearly this range is reduced by surface roughness. The effect of the roughness is to decrease the reflectivity of the surface and thus to increase its emissivity. For a dry field the reflectivity is already small ($\lesssim 0.1$) so that the resulting increase in emissivity is small. As seen in Figure 10b, surface roughness has a significant effect for wet fields where the reflectivity is larger ($\gtrsim 0.4$). Thus, the range of T_B for the rough field is reduced to about 60 K. The smooth and rough fields represent the extremes of surface conditions that are likely to be encountered, e.g. the rough surface was on a field with a heavy clay soil (clay fraction 60%) that had been deep plowed, which produced large clods. Therefore, the medium rough field, with a T_B range of 80 K, is probably more representative of the average surface roughness condition that will be encountered. Another important observation from Figures 9a and b is that the average of the vertical and horizontal T_B's is essentially independent of angle out to 40°. This indicates that the sensitivity of this quantity, $1/1(T_{BV}+T_{BH})$, to soil moisture will be independent of angle. This result will be useful if the radiometer is to be scanned to provide an image.

When the brightness temperatures for the medium rough field are plotted vs soil moisture in the 0 to 2-cm layer, there is an approximate linear decrease of T_B (Figure 9c). As the thickness of the layer increases, both the slope and intercept of the linear regression also increase. This is because the moisture values for the high T_B cases increase, whereas they remain essentially the same in the low T_B or wet cases. This type of behavior was also seen in the results obtained from aircraft platforms and has led us to conclude that the soil moisture sampling depth is within the 2 to 5-cm range for the 21-cm wavelength. This agrees with the predictions of theoretical models of radiative transfer in soils (Wilheit, 1978; and Burke et al, 1979).

Results from a 1975 aircraft experiment over irrigated agricultural fields are presented in Figure 10 (Choudhury et al, 1979). These results were obtained over fields with a range of soil textures from sandy loam to heavy clays. In the analysis it was observed that there was a dependence of the T_B response to soil moisture on the soil texture, i.e. the slope was greater for sandy soils which had a narrower range of soil moistures (0 - 20%) compared with the clay soils (0 - 35%). To take this texture dependence into account the soil moisture values presented in Figure 11 are normalized to the field capacity (FC) value for the particular soil which were estimated from the measured soil textures.

The solid symbols in Figure 10 are calculated values of T_B obtained with the Wilheit (1978) model using the measured moisture and temperature profiles for the fields. The solid line connects the values determined assuming a smooth surface, and the dashed line connects the values adjusted for surface

roughness using a one parameter model. The dashed line fits the observed values quite well. The values of the parameter were selected empirically to give a best fit to the data and it is clear that the same value works well for both the dawn and midday flights. The effects of soil temperature are seen in the T_B differences between the dawn and midday flights.

There is little change in T_B for soil moisture values for the 0 to 2 cm layer out to about 30% of field capacity. When the data are replotted vs the 0 to 5 cm layer, the flat region extends out to about 50% of FC with a steeper slope beyond that value. In Figure 1, the dielectric curve broke at a moisture level of about 10%, which for the soil involved would be 40 to 50% of FC. Thus, these aircraft results support the conclusion that the sampling depth is about 2 to 5 cm.

A vegetative canopy will act as an absorbing layer whose effect will depend on the amount of vegetation and the wavelength of observation. Basharinov and Shutko (1978) and Kirdiashev et al, (1979) have reported on observations made in the USSR over the 3 to 30-cm wavelength range for a variety of crops. Their results are summarized in Figure 11, where the vegetation factor is the effective transmissivity of the vegetation. Thus, for small grains the sensitivity is 80 to 90% of that expected for bare ground at wavelenghts greater than 10 cm. Broad leaf cultures, like mature corn or cotton, transmit only 20 - 30% of the radiation from the soil at wavelengths shorter than 10 cm and about 60% at the 30-cm wavelength. They observed 30 to 40% sensitivity for a forest at the 30-cm wavelength, although they did not mention the type or height of trees. These results are encouraging for the use of long wavelength radiometric approaches.

The Earth Resources Experiment Package (EREP) on board Skylab contained a 21-cm radiometer. This sensor was non scanning with a 115 km field of view between half power points. With this coarse spatial resolution, it would be difficult to directly compare sensor response and soil moisture measurements. However, there have been two reports of indirect comparisons. McFarland (1976) showed a strong relationship between the Skylab 21-cm brightness temperatures and the Antecedent Precipitation Index (API) for data obtained during a pass starting over the Texas and Oklahoma panhandles and proceeding southeast toward the Gulf of Mexico.

Eagleman and Lin (1976) carried the analysis of the Skylab data a step further and compared the brightness temperature with estimates of the soil moisture over the radiometer footprint. The soil moisture estimates were based on a combination of actual ground measurements and calculations of the soil moisture using a climatic water balance model. They obtained a correlation of 0.96 with data obtained during five different Skylab passes over Texas, Oklahoma and Kansas. This result is very good considering the difficulty of obtaining soil moisture information over a footprint of such a size and considering the fact that the brightness temperature was averaged over the wide range of cultural conditions that occurred over the area.

These results from space supported by the more detailed aircraft and ground measurements presented earlier strongly support the possibility of using microwave radiometers for soil moisture sensing. A difficulty with this approach is that the spatial resolution is limited by the size of the antenna which can be flown. For example, at a wavelength of 21 cm, a 10 m X 10 m antenna is required to yield 20-km resolution from a satellite altitude of

800 km. It is possible to make use of the coherent nature of the signal in active microwave systems (Synthetic Aperture Radar, SAR) to obtain better spatial resolutions (Moore, 1975) and it is this approach which we discuss next.

Active Microwave

The backscattering from an extended target, such as a soil medium, is characterized in terms of the target's scattering coefficient σ_c . Thus, σ_c represents the link between the target properties and the scatterometer responses. For a given set of sensor parameters (wavelength, polarization and incidence angle relative to 0^o), σ_c of bare soil is a function of the soil surface roughness and dielectric properties which depends on the moisture content. The variations of σ_c with soil moisture, surface roughness, incidence angle, and observation frequency have been studied extensively in ground-based experiments conducted by scientists at the University of Kansas (Ulaby, 1974; Ulaby et al, 1974; Batlivala and Ulaby, 1977) using a truck mounted 1 to 18 GHz (30-1.6 wavelengths) active microwave system.

To understand the effects of incidence angle and surface roughness consider the plots of σ_c versus angle presented in Figure 12 for five fields with essentially the same moisture content but with considerably different surface roughness. At the longest wavelength (1.1 GHz, Figure 12a), σ_c for the smoother fields is very sensitive to incidence angle near nadir, while for the rough field σ_c is almost independent of angle. At an angle of about 5^o, the effects of roughness are minimized. As the wavelength decreases, Figures 12b and 12c, all the fields appear rougher, especially the smooth field, and as a result the five curves intersect at larger angles. At 4.25 GHz, they intersect at 10^o, and it was this combination of angle and

frequency that yielded the best sensitivity to soil moisture independent to roughness (Ulaby and Batlivala, 1976; and Ulaby et al, 1978).

These experiments were performed in both 1974 and 1975, the first experiment was performed on a field with high clay content (62%), whereas for the second, the clay content was lower. Although both experiments provided the same specifications of the radar parameters for soil moisture sensing, i.e., frequency around 4.75 GHz and a 7-17° nadir angle, the observed sensitivity of σ_c to soil moisture in the 0 to 1cm layer was different for the two experiments, Figure 13b. When the soil moisture content is expressed as a percent of field capacity to account for textural differences, the sensitivities became almost identical (Figure 13a) with a correlation of 0.84. This dependence on the percent of field capacity resembles to that observed with the thermal inertia and passive microwave techniques. Similarly the sampling depth for active microwave sensors also seems to be limited to the surface few centimeters of the soil for the wavelengths considered in the Kansas study (Ulaby et al, 1978).

Although no detailed airborne investigations have yet been reported on the active microwave response to the soil moisture content beneath a vegetation canopy, we observed the difference between dry soil and soil undergoing irrigation in 1971 while conducting radar observations of agricultural fields. During a flight by the NASA/JSCSPP3A aircraft over a test site near Garden City, Kansas, (Dickey et al) using a 13.3. GHz scatterometer measured several fields, each of which (from aerial photography and field crew's reports) contained sections into which irrigation water was flowing and sections ready for irrigation but not yet wetted. For one of these fields,

a corn field, the effect of the irrigation on the radar return seemed to produce a difference of about 7 dB at angles within 40° from nadir between the irrigated and non-irrigated sections. Since all ground conditions, except soil water content, were similar over the entire field, the differences in σ_c can only be attributed to the effect of moisture.

The presence of a vegetation canopy over the soil surface reduces the sensitivity of the radar backscatter to soil moisture by: a) attenuating the signal as it travels through the canopy down to the soil and back and by b) contributing a backscatter component of its own. Moreover, both factors are, in general, a function of several canopy parameters, including plant shape, height and moisture content, and vegetation density. The effect of the vegetation cover on the radar response to soil moisture is to reduce the sensitivity by about 40% when the bare soil and vegetation-covered responses are compared as a function of percent of FC in the top 5 cm. The vegetation-covered response represents data for several crops, wheat, corn, soybeans, and milo, covering the wide range of growth conditions (Ulaby, et al, 1977)

There are many similarities in the two microwave approaches to soil moisture sensing, e.g. ability to penetrate clouds and moderate amounts of vegetation and the limitation to sampling only the surface 2 - 5cm of the soil. The major difference is that of spatial resolution, for passive systems the resolution is limited by the size of the antenna, and this for practical reasons will be limited to 5 to 10km. On the other hand, using the synthetic apperature techniques, spatial resolutions of 100 m or less are possible from

space, e.g. 25-m resolutions was obtained at the 18 cm synthetic aperature radar on the Seasat Satellite. The problems with the latter approach is the difficulty in getting an absolute calibration for the SAR, the strong sensitivity to surface roughness and look angle, and the large amount of data that would have be handled in any operational context.

The sensitivity to soil moisture of the three remote sensing approaches discussed here has been demonstrated in field or aircraft experiments and to a certain extent from spacecraft platforms. These experiments have also indicated some of the problems associated with each approach. These problems are summarized in Table 2, which presents a comparison of the remote-sensing approaches. Some of the limitations listed are of a fundamental nature, like cloud cover effects at thermal infrared, whereas others could be reduced or eliminated by more advanced technology, like larger antennas to achieve improved radiometer resolution or the development SAR calibration techniques. There is a fundamental limitation which applies to all of the approaches, i.e. they seem to be sensing the moisture content in a layer only 5 - 10 cm thick at the surface. This limitation implies that remote sensing approaches will not be able to satisfy those applications which require knowledge of the moisture conditions in the root zone of the soil.

TABLE 2

Comparison of Remote Sensing Approaches

APPROACH	ADVANTAGES	LIMITATIONS	NOISE SOURCES
Thermal Infrared (10-12 m)	High resolution possible (400 m)	Cloud cover, limits frequency of coverage	Local Met conditions
	Large swath		Partial vegetative cover
	Basic physics well understood		Surface topography
Passive Microwave	Independence of atmosphere	Poor spatial resolution (5-10 km at best)	Surface roughness
	Moderate vegetation penetration	Interference from manmade radiation sources, limits operating wavelengths	Vegetative cover
			Soil temperature
Active Microwave	Independence of the atmosphere	Limited swath width	Surface roughness
	High resolution possible	Calibration of SAR	Surface slope
			Vegetative cover

DISCUSSION

Most hydrologic, agricultural and meterological applications that could benefit from soil moisture measurements have three general requirements; frequent observations, an estimate of moisture within the top 1 to 2 of the soil and, generally, a description of moisture variations over large study areas, - i.e., a county or state. In preceding sections, we have pointed out some of the advantages and disadvantages of the three approaches than can be used to collect soil moisture data. None of the individual techniques completely satisfies the requirements of the applications in a cost-effective manner.

In situ methods can accurately estimate soil moisture throughout the profile. However, the information is reliable only at the point of measurement. To achieve a specified level of accuracy in estimating the areal average for most applications, a large number of point samples will be required. For example, in a study of a large number of intensively sampled fields (20 or more samples per field), Bell et al. (1979) found that for 90% of the cases, the standard deviation (σ) of soil moisture in the surface soil layers was less than 4%. Snedecor and Cochran (1967) presented the following relationship for estimating the required sample size:

$$n = 4 \left(\frac{\sigma}{L} \right)^2 \qquad (6)$$

where n is the sample size and L is the desired level of accuracy. Using this equation and specifying L = 2% and σ = 4% (from Bell et al. (1979)), the required number of samples would be 16. Thus, if a large number of individual field estimates are required, the costs can become prohibitive. In addition, the current state-of-the-art methods generally require on site observations since reliable remote devices have not been developed. Thus, a significant commitment of manpower is required.

Soil water models provide fast answers, as well as predictions, of the soil water regime in a field or over a region. These models require large amounts of meteorological input data that can be difficult and costly to obtain. Model parameters and functions are also difficult to determine. In addition, various sources introduce error into the model predictions that can lead to significant deviations.

Remote-sensing methods offer rapid data collection over large areas on a repetitive basis. Several questions still need to be answered concerning the dependence of sensor observations on soil moisture and other parameters, like vegetation. The major problems related to remote sensing seems to be the spatial resolution, depth of penetration, and cost. For the spatial resolution, the passive microwave sensors will measure the areal soil moisture over ground areas about 1 km^2 in size from satellite altitudes. Presently, we do not know how useful such a measurement would be. The limited depth of penetration of these systems is also a severe drawback. Even if such data were available, how could it be used?

These observations lead us to concluded that a cost-effective soil moisture monitoring program must utilize all three of the approaches and not just one. Each has its advantages and disadvantages, an integrated system should be designed to capitalize on the advantages and minimize the disadvantages. In-situ methods which we consider the most accurate could be used sparingly in such a system for calibration and verifiction of models and remote-sensing measurements. These other two appraches could be used to interpolate between the point measurements for estimating areal averages. Remotely sensed measurements could provide an estimate of the surface conditions, which could then be extrapolated via a model or used to update a

model simulation. These remotely sensed measurements could also be processed rapidly to give us a quick look at the general condition over large areas.

In summary, we reviewed a wide variety of methods for estimating soil moisture in this paper. Each has its own advantages and disadvantages related to large scale soil moisture monitoring. If a successful monitoring system is to be developed, it must incorporated all of these approaches. A very brief description of such an integrated system was presented; however, considerable research is needed to develop an optimal system.

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FIGURE CAPTIONS

1. Real and imaginary parts of the dielectric constant for clay loam soils at wavelength of 21 cm (1.41 GHz) (Lundien, 1971) and 1.55 cm (19.35 GHz) (Wang, 1978).

2. Schematic diagram of Soil-Plant-Atmosphere-Water (SPAW) system.

3. Flow chart of SPAW Model (Saxton, et al., 1974).

4. The USDAHL Model (Holtan and Yaramanoglu, 1977).

5. The NWSRFS Model (Peck, 1976).

6. A distributed soil moisture simulation model.

7. Summary of results for the diurnal temperature variation versus soil moisture (Idso, et al., 1975).

8. Plot of ΔT versus soil moisture in the 0-2 cm layer. The symbols represent the different types of temperature measurement; ● - □ surface thermocouple, ○ - hand held radiometer, △ - aircraft data over test plot, x - aircraft data over the general agricultural fields. (●, □, ○, △ from Reginato, et al., 1976; x from Schmugge, et al., 1978)

9. Results from field measurements performed at Texas A & M University: (a) T_B versus angle for different moisture levels; (b) T_B versus angle for different surface roughness at about the same moisture level; (c) T_B versus soil moisture in different layers for the medium rough field (Newton, 1977).

10. Aircraft observations of T_B over agricultural fields around Phoenix, Arizona from March 1975 flights for both early morning and midday flights.

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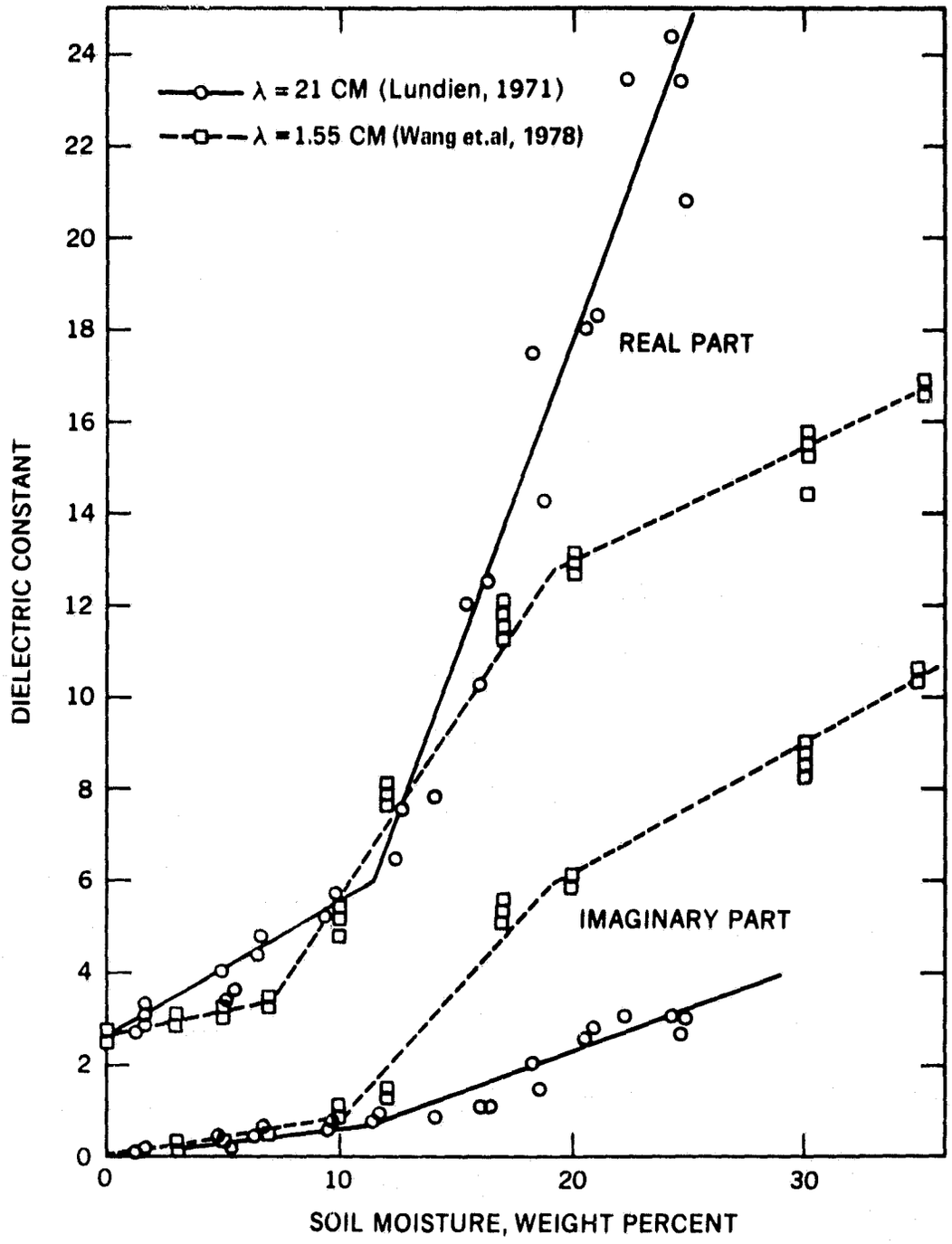


Figure 1. Real and imaginary parts of the dielectric constant for clay loam soils at wavelength of 21 cm (1.41 GHz) (Lundien, 1971) and 1.55 cm (19.4 GHz) (Wang, 1978).

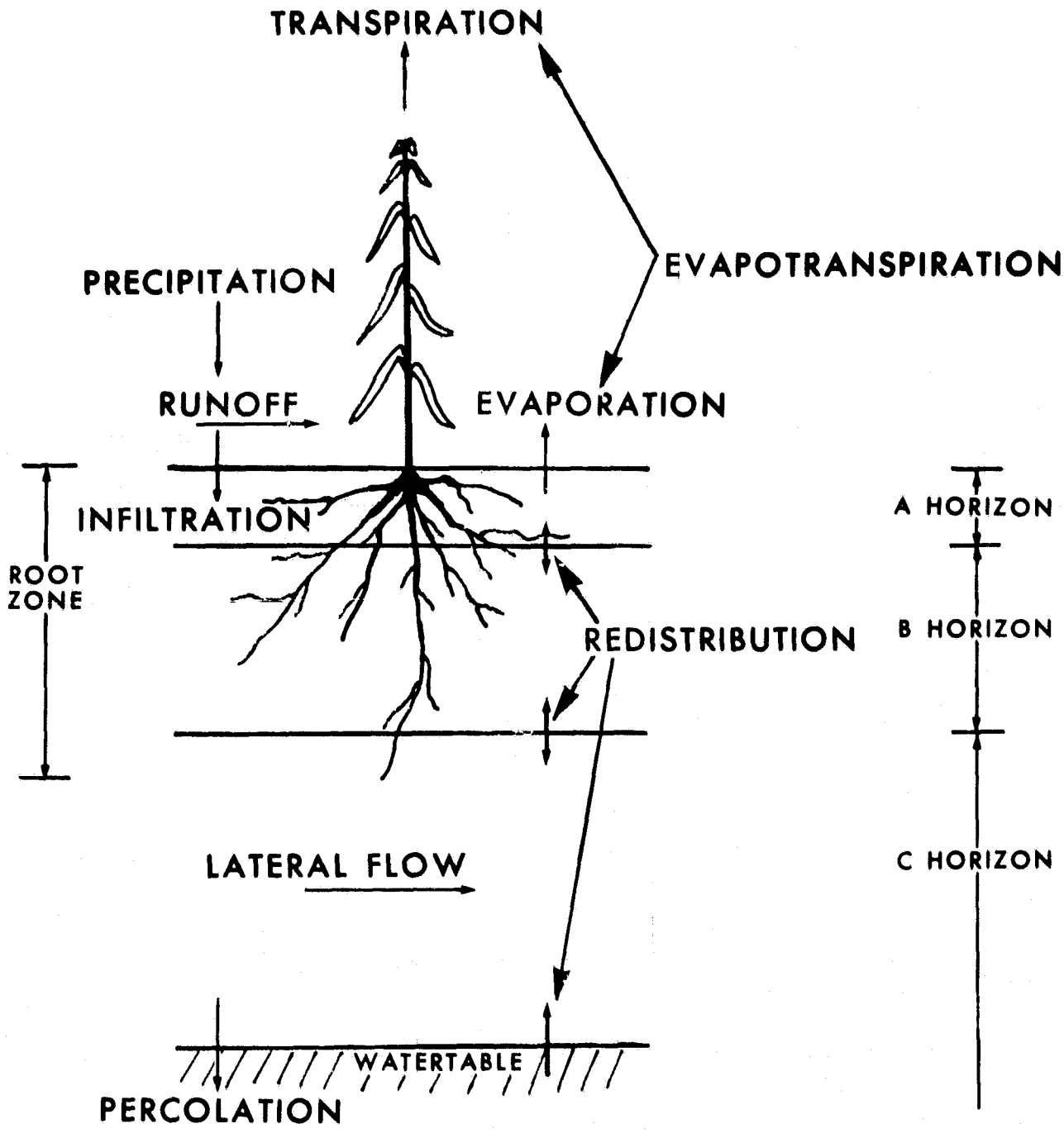


Figure 2. Schematic diagram of Soil-Plant-Atmosphere-Water (SPAW) system.

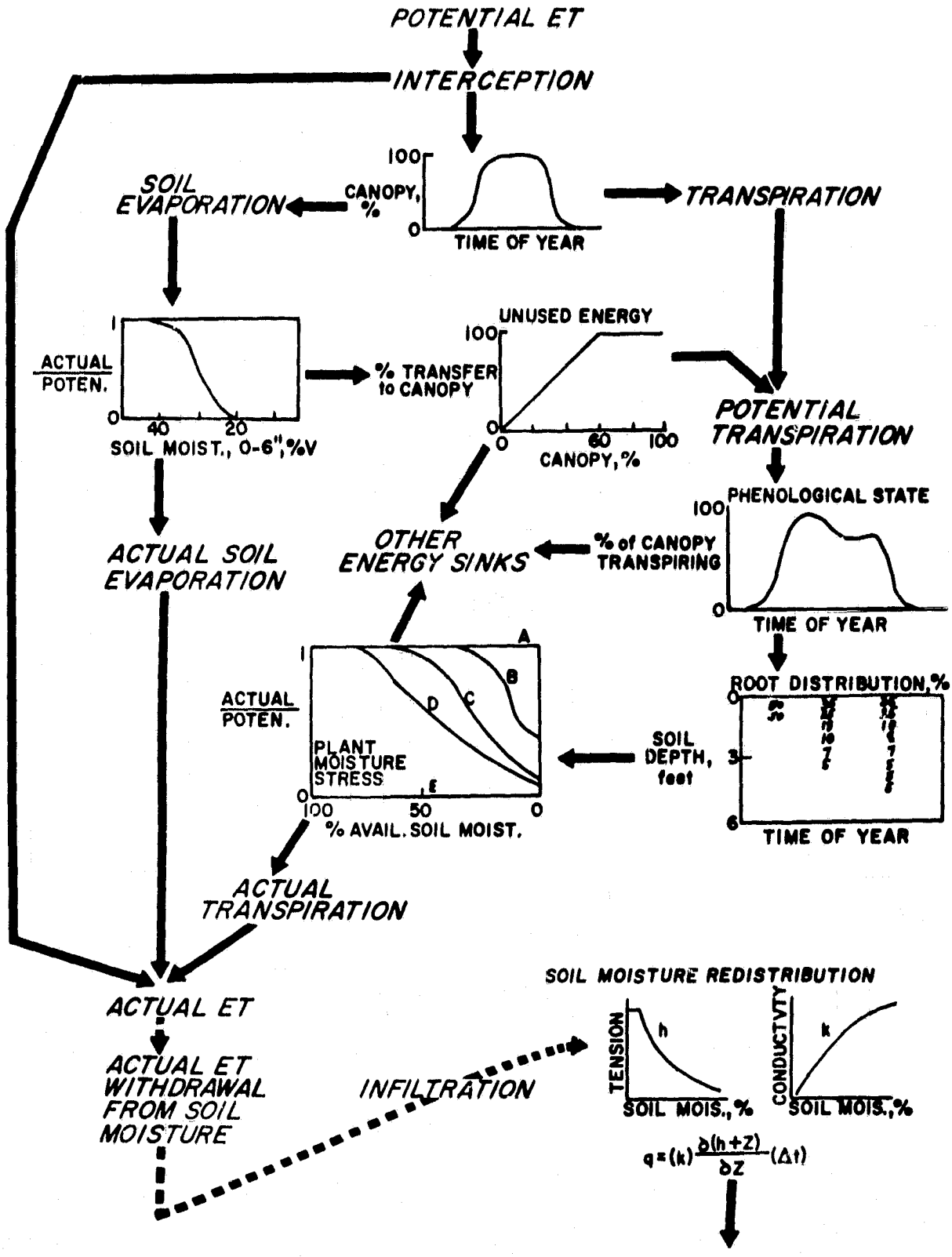


Figure 3. Flow chart of SPAW Model (Saxton, et al., 1974).

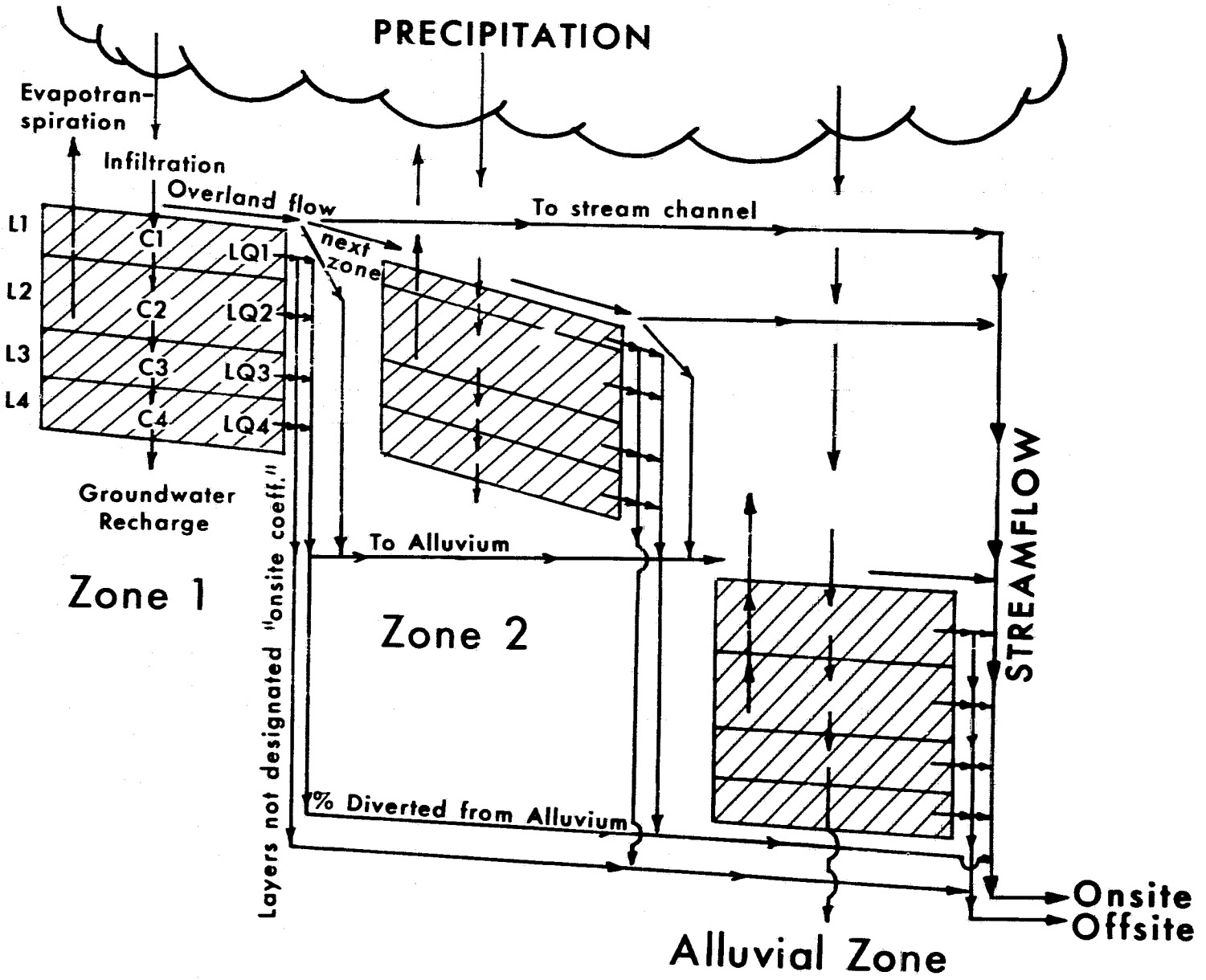


Figure 4. The USDAHL Model (Holtan and Yaramanoglu, 1977).

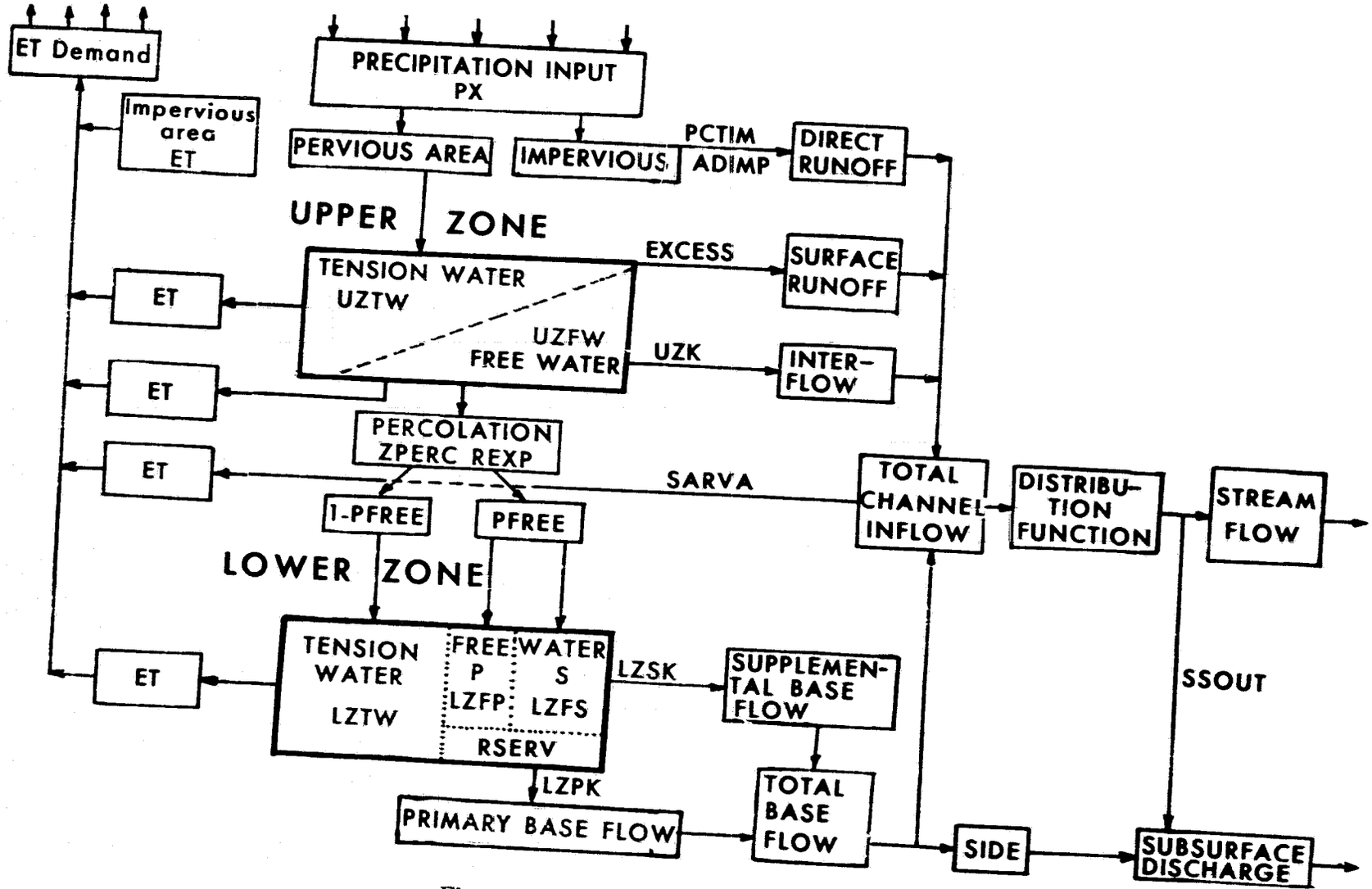
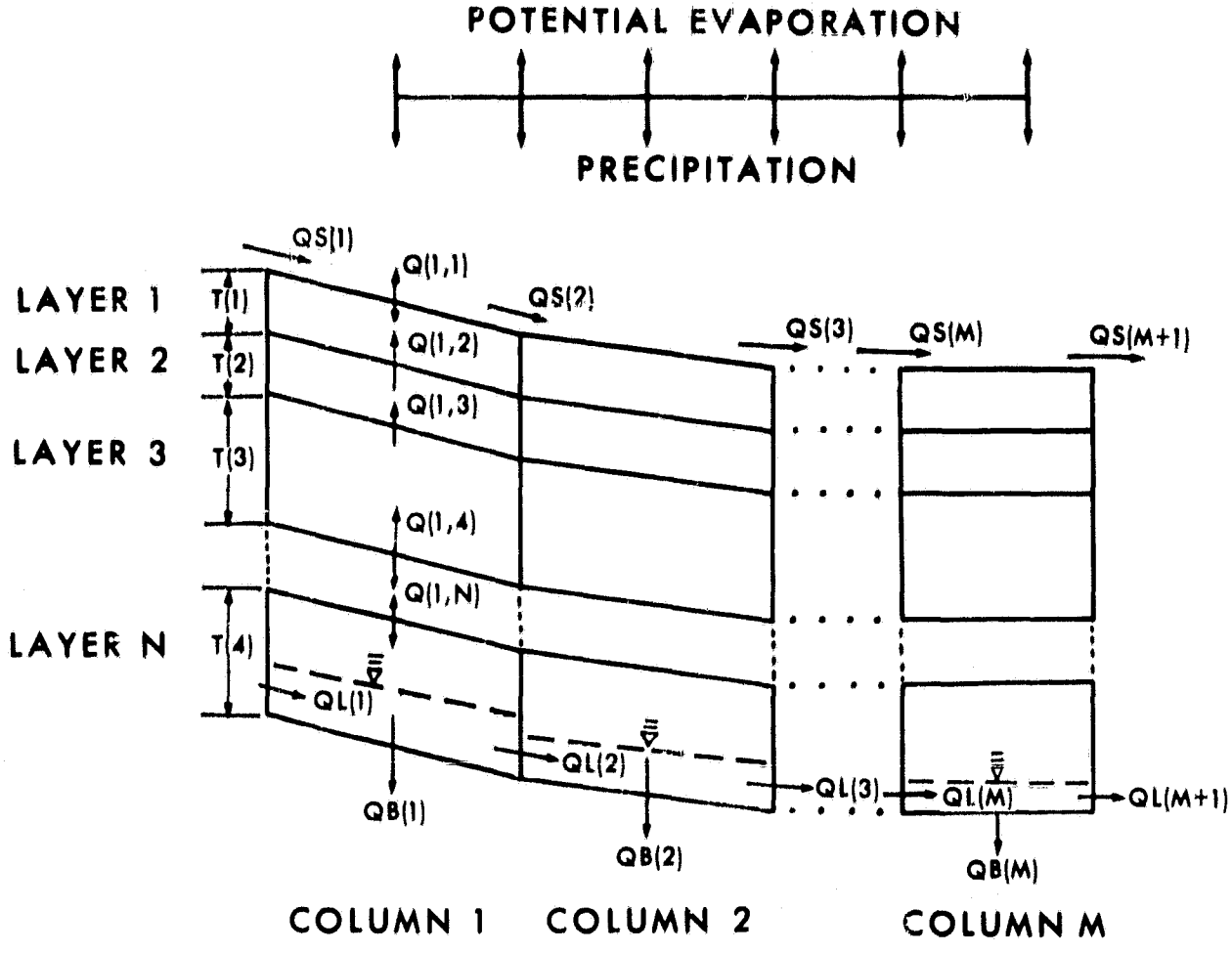


Figure 5. The NWSRFS Model (Peck, 1976).



SYMBOLS

I = LAYER INDEX
J = COLUMN INDEX
Q(I,J) = FLOW BETWEEN TWO LAYERS.
QB(J) = FLOW RATE FOR DEEP PERCOLATION OR A ZERO FLOW BOUNDARY FOR COLUMN J.
QL(J) = SATURATED FLOW TO COLUMN J BELOW THE WATER TABLE.
QS(J) = SURFACE RUNOFF TO COLUMN J.
T(I) = THICKNESS OF LAYER I.

Figure 6. A distributed soil moisture simulation model.

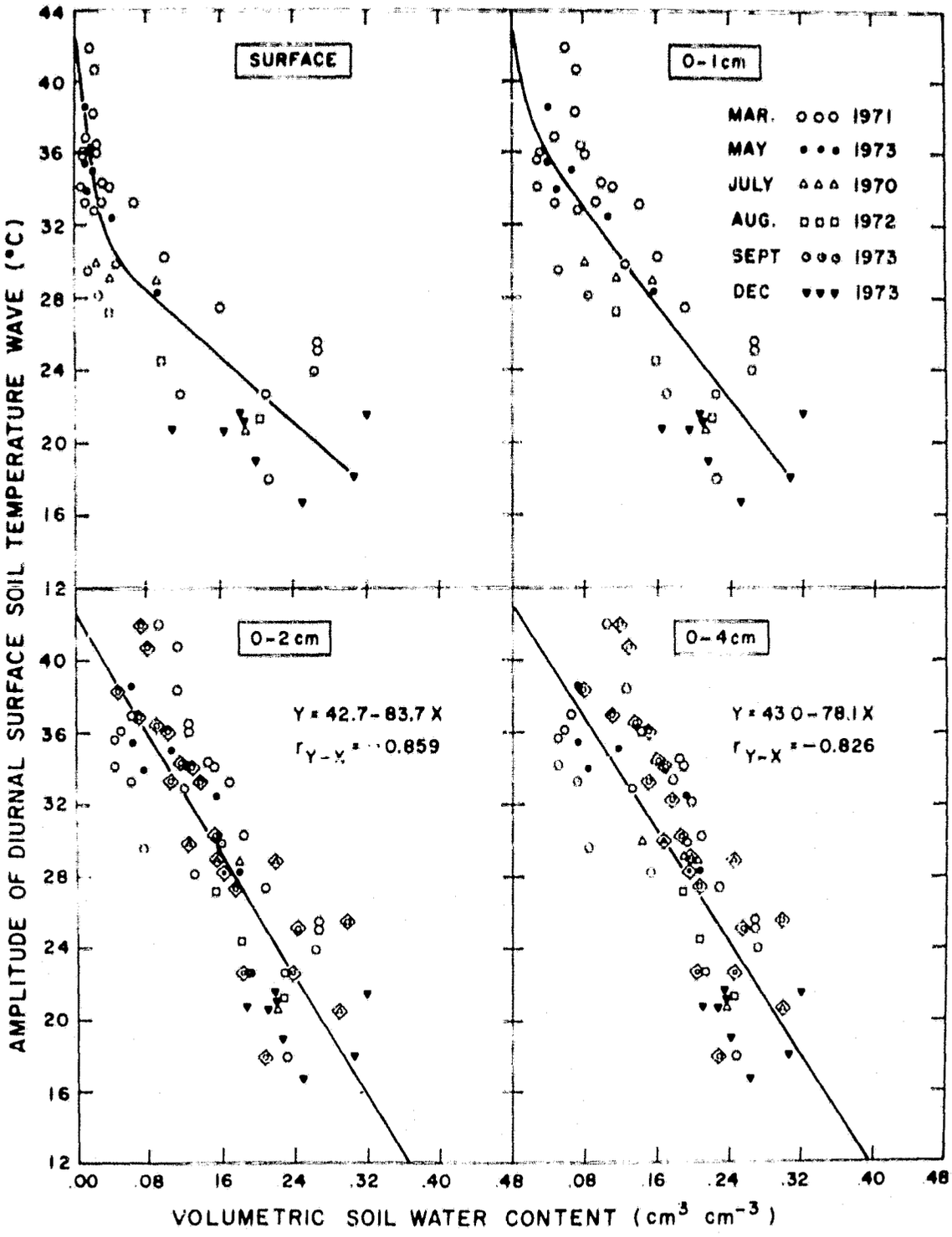


Figure 7. Summary of results for the diurnal temperature variation versus soil moisture. (Idso, et al., 1975).

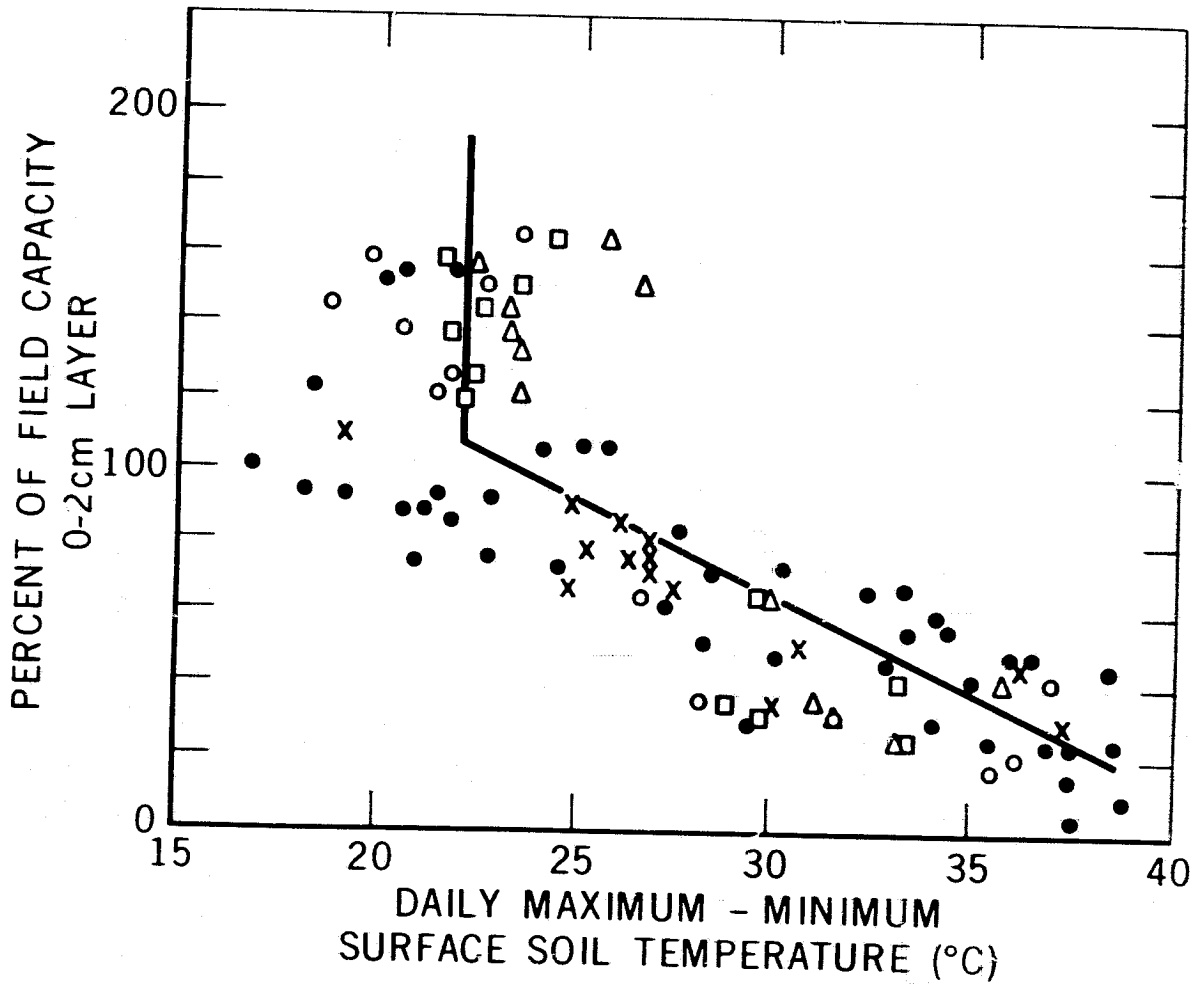
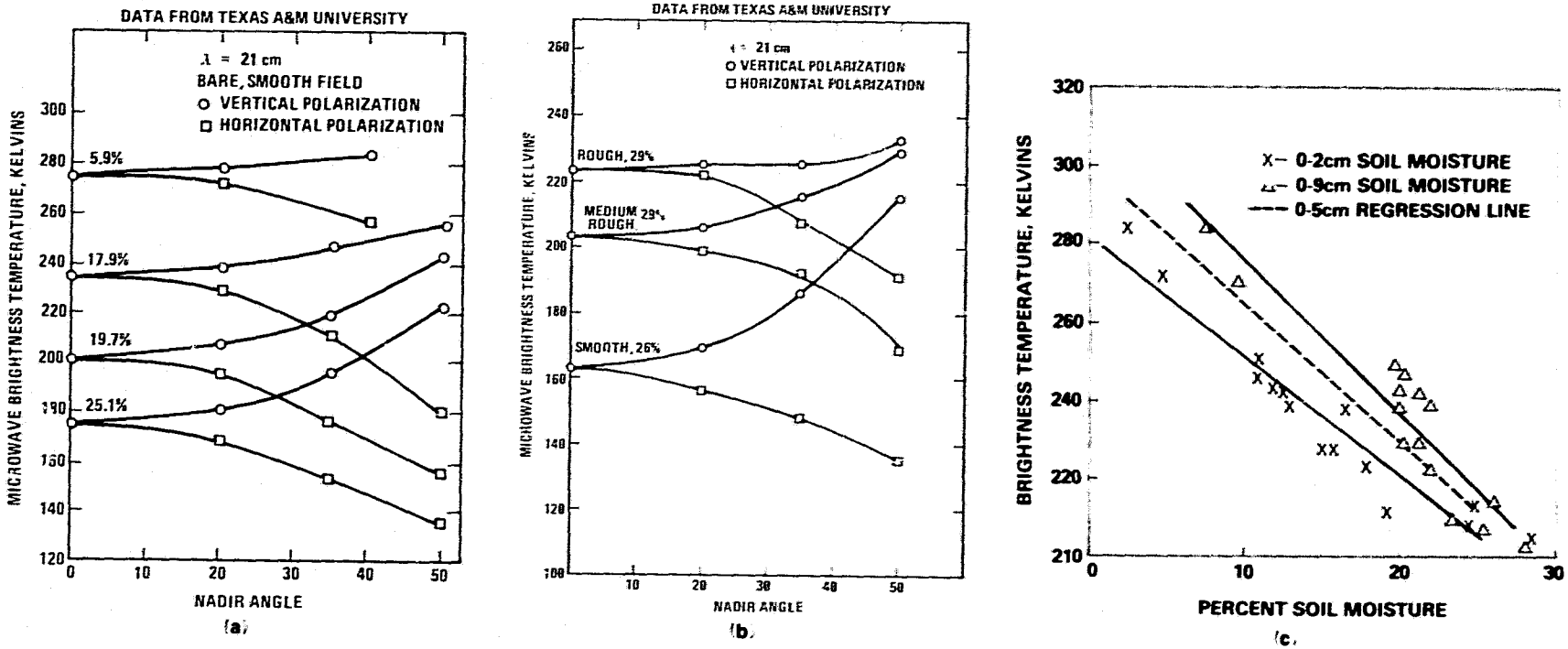


Figure 8. Plot of ΔT versus soil moisture in the 0-2 cm layer. The symbols represent the different types of temperature measurement; ● – □ surface thermocouple, ○ – hand held radiometer, △ – aircraft data over test plot, x – aircraft data over the general agricultural fields. (●, □, ○, △ from Reginato, et al., 1976; x from Schmugge, et al., 1978).



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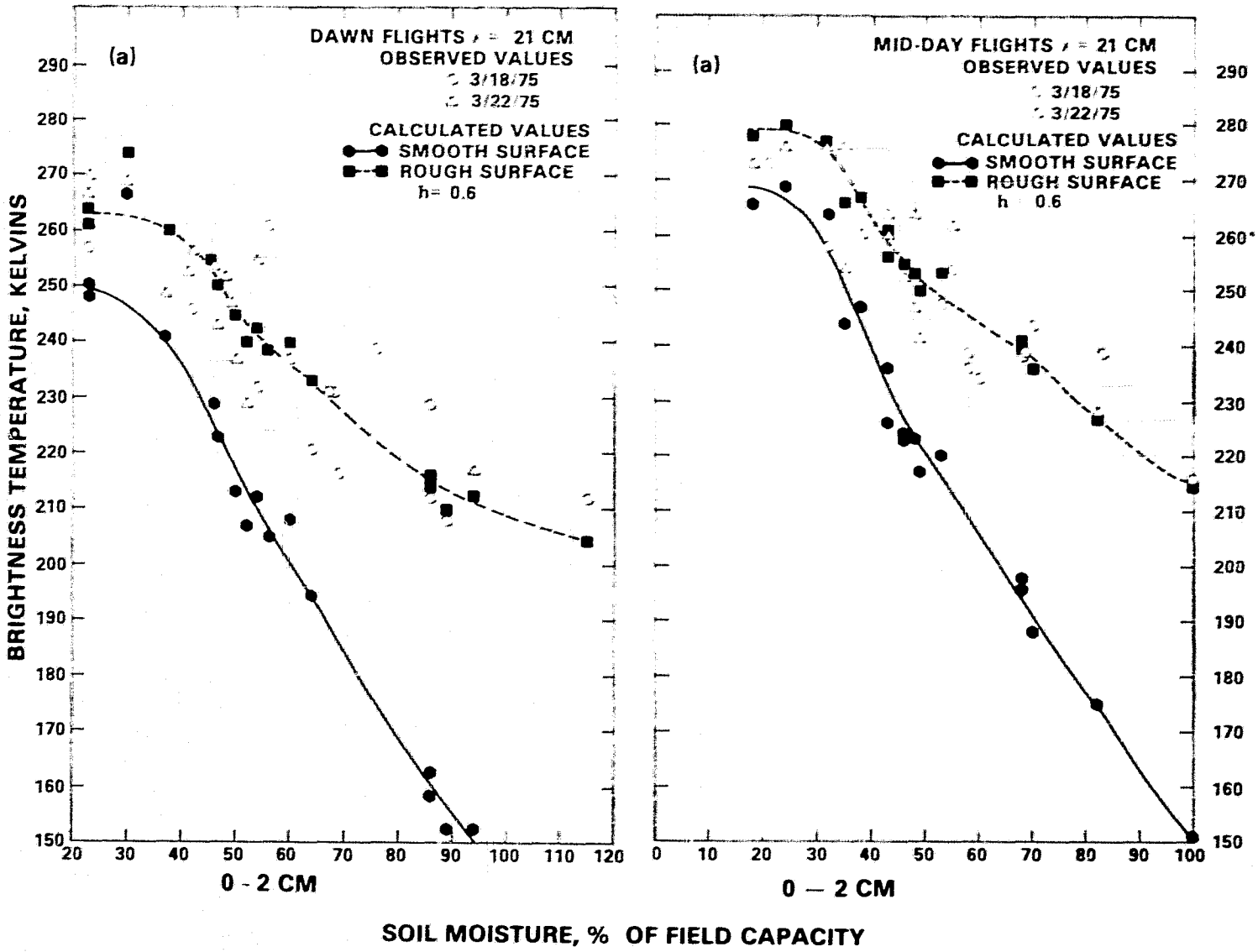

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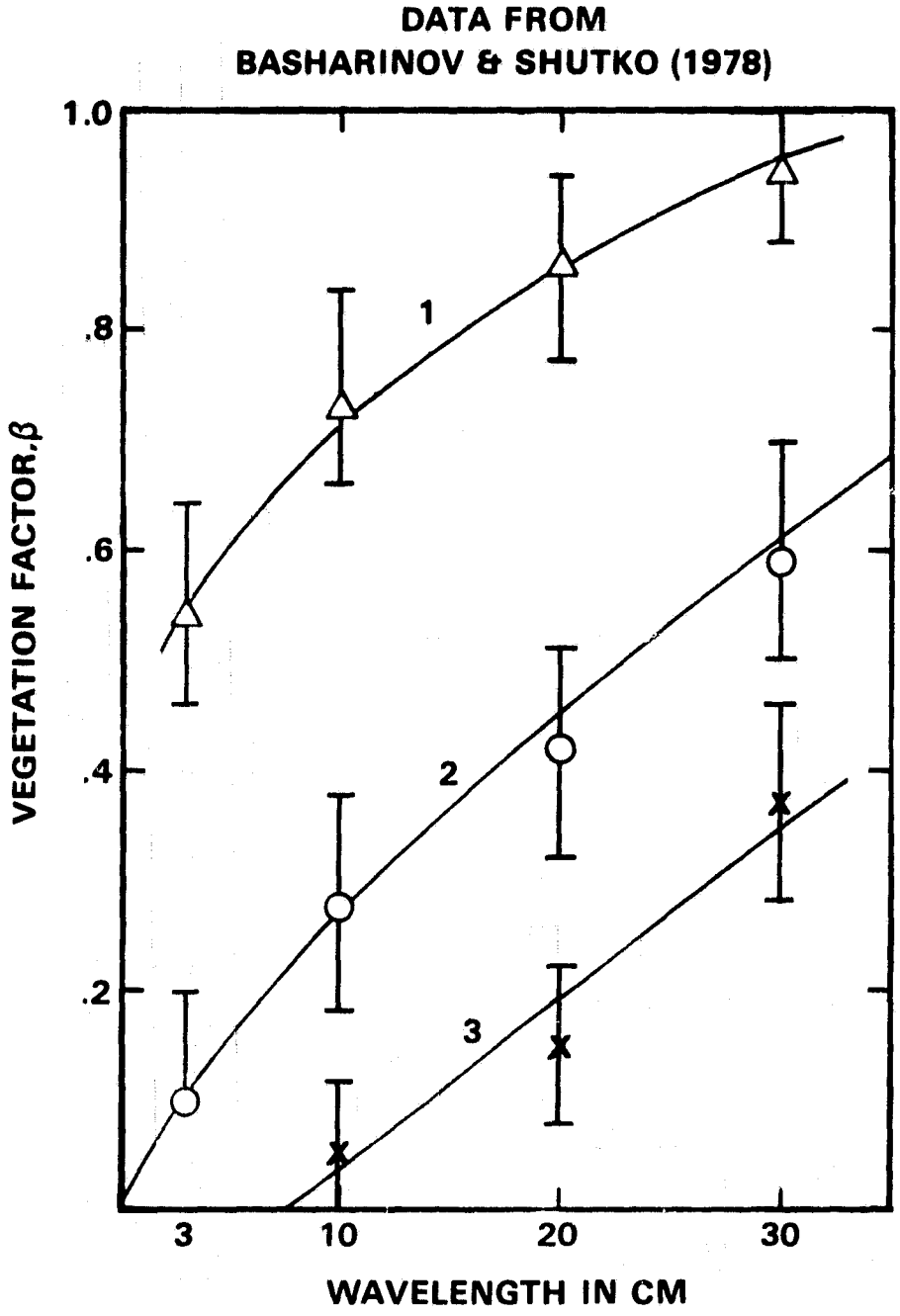

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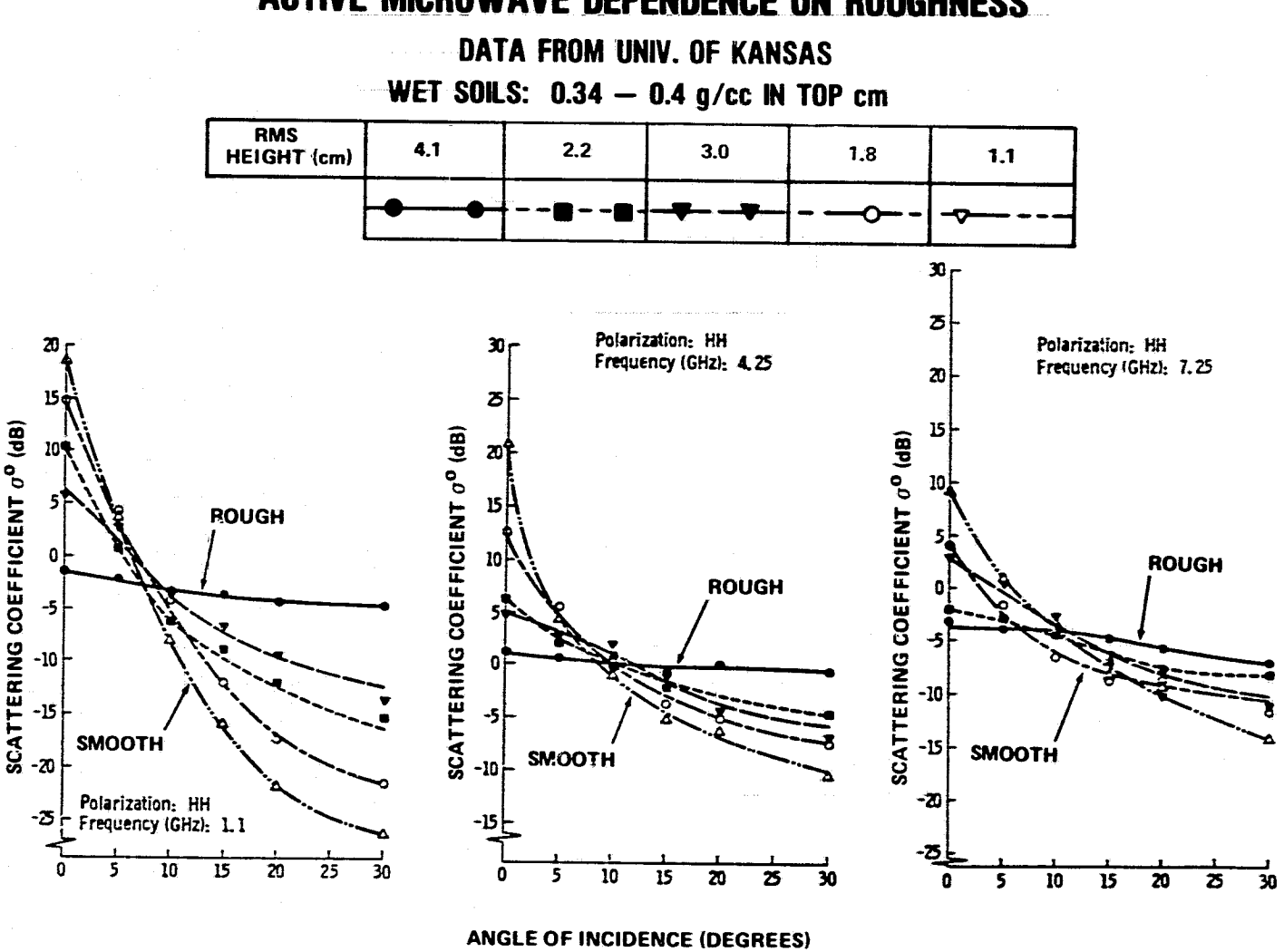

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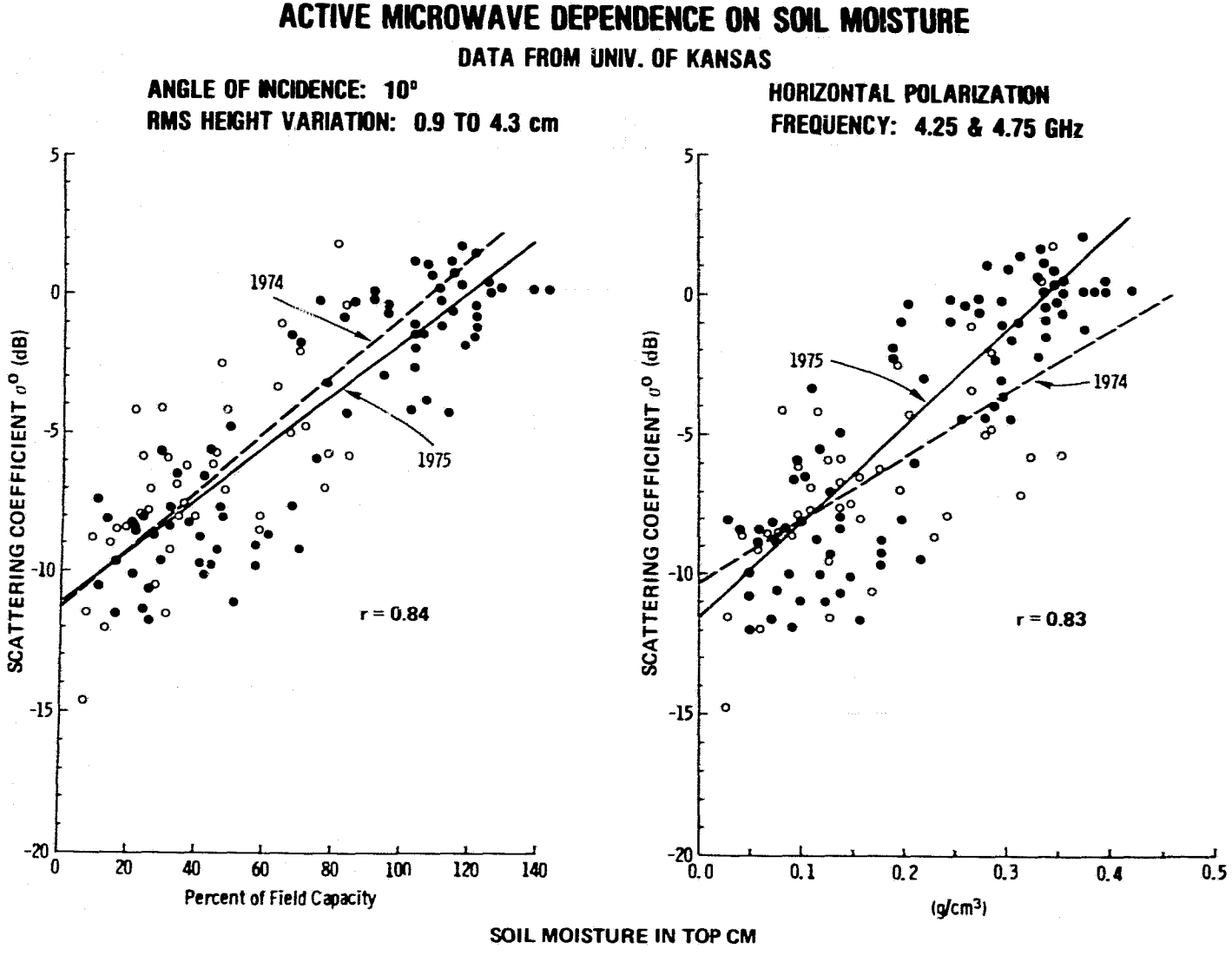

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