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ESTIMATION OF SOIL MOISTURE VARIATION USING RESISTIVITY TECHNIQUE

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1986-87

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ABSTRACT

Relation between bulk resistivity and amount of pore water as well as its properties has been known in form of empirical formulae. The electrical properties of soils in unsaturated zone above water table are dependent on a number of parameters. Important among these are porosity, degree of saturation, grain size and shape, and conductivity of water/ electrolyte saturating the soil. The reported study aims at using the surface resistivity measurements in determination of soil moisture status and its temporal variation.

Introductory section includes general description of soil properties, especially electrical properties. Some recent developments in non-conventional methods of soil-moisture measurements, e.g. time domain reflectometry, optical transmission, geophysical methods etc., have been reviewed in order to familarize readers with the latest state-of-art techniques available in the subject field.

Various factors affecting resistivity of soils have been briefly described. It is observed that soil water content and its salinity has largest effect on soil resistivity. Effects of porosity and porewater conductivity alongwith variation of soil moisture content on soil resistivity have been computed. Resistivities were computed at various porosities (30-50%), porewater conductivities (16-26 ohm-m), and soil water content (10-60%). Relative changes in bulk

resistivity and computed soil moisture show a distinct pattern of relation between soil resistivity and water content.

The study shows that higher soil moisture content (50% and above) has marginal effect on soil resistivity, especially for medium porosity ranges (30-40%). Effect is appreciably large for low soil moisture contents (5-50%) and for higher porosities (40% and above). It is, therefore, possible to use field resistivity measurements to monitor soil water variation for irrigation scheduling etc

1.0 INTRODUCTION

1.1 General

The term 'soil' refers to the weathered and fragmented outer layer of the earth's land surface. It is found in varying thicknesses at various places, typically it's thickness lies upto 3-5 m from the ground surface. It is formed initially from disintegration and decomposition of rocks by physical and chemical processes, and is influenced by the activity and accumulated residues of numerous biological species.

1.2 Soil As A Complex System

The soil is a heterogenous, multi-phase, disperse and porous system (Fig. 1). Three phases in the soil are the solid phase, consisting of soil particles, the liquid phase, consisting of soil water, which always contains dissolved substances, and the gaseous phase, consisting of soil air. In a heterogenous system, the properties differ not only between one phase and another, but also between the internal parts of each phase and the boundaries or interfaces of the phase with its neighbouring phase or phases. Interfaces exhibit specific phenomena, such as absorption, surface tension, and friction, which result from the interaction of adjacent phases and therefore do not exist within the homogeneous phases themselves. These phenomena affect considerably when size of the interfacial area per unit volume of the system is large. Sometimes soils comprise of phases one or more of which can be subdivided into minute particles thereby exhibiting a large surface area. The disperse nature of the soil and its consequent interfacial activity gives rise to such phenomena as swelling, shrinkage, dispersion, agrregation, adhesion, adsorption etc.

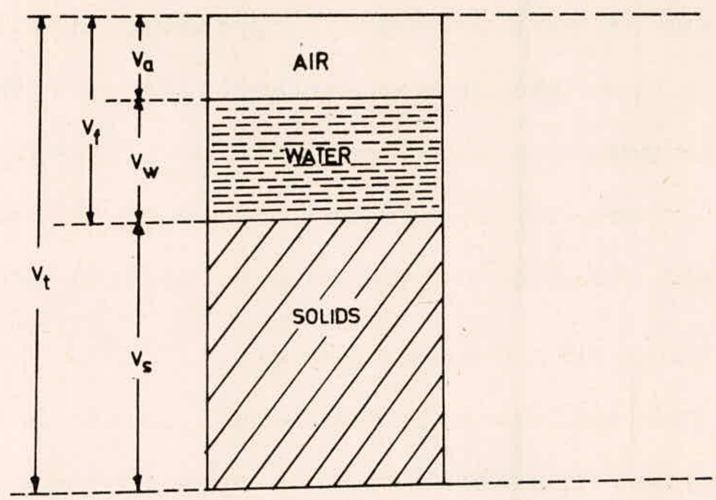

FIG. 1-SCHEMATIC DIAGRAM OF THE SOIL AS A THREE-PHASE SYSTEM (REDRAWN FROM HILLEL, 1971)

The soil is thus seen to be an exceedingly complex system. Its solid matrix consists of particles differing in chemical and minerological composition as well as in size, shape and orientation. The mutual arrangement or organisation of these particles in the soil determines the characteristics of the pore spaces in which water and air are transmitted or retained. The water and air also vary in composition, both in time and in space. Moreoever, it is not always easy to separate these phases, as they interact very strongly upon one another.

Considering fig. 1, various parameters characterising the soil properties are defined as follows:

(i) Porosity $\emptyset$:

$$\emptyset = \frac{V_f}{V_t} = \frac{V_a + V_w}{V_s + V_a + V_w} \quad (1.1)$$

where, V_a is the volume of air, V_w is the volume of water, V_s is the volume of solids, V_t is the total volume of the soil, and $V_f = V_a + V_w$ is the volume of pores.

The porosity is an index of the relative pore volume in the soil. Its value generally lies in the range 0.3 - 0.6 (30 - 60%). Coarse textured soil tend to be less porous than fine-textured soils, though the mean size of individual pores is greater in the former than in the latter. In clayey soils, the porosity is highly variable as the soil alternatively swells, shrinks, aggregates, disperses, compacts and cracks.

(ii) Volume (soil) wetness θ :

$$\theta = \frac{V_w}{V_t} = \frac{V_w}{V_s + V_f} \quad (1.2)$$

The volume wetness (commonly termed 'volumetric water content') is generally computed on the basis of total volume of the soil rather than on the basis of particles alone. In sandy soils, the value of 0 at saturation is of the order of 40 - 50%; in medium textured soils, it is approximately 50%, and in clayey soils, it can be of the order of 60%. In the latter type of soils, the relative volume of water at saturation can exceed the porosity of the dry soil, since clayey soils swell upon wetting.

iii) Degree of saturation S :

$$S = \frac{V_w}{V_f} = \frac{V_w}{V_a + V_w} \quad (1.3)$$

This index (commonly called simply 'saturation') expresses the volume of water present in the soil relative to the volume of pores. The index S ranges from zero in dry soil to 100% in a completely saturated soil. However, 100% saturation is seldom attained, since some air is nearly always present and may become trapped in a very wet soil. This is not a good index for swelling soils, in which porosity changes with wetness.

The relation between degree of saturation and volume wetness is expressed as :

$$S = \frac{\theta}{\emptyset} \quad (1.4)$$

1.3 Characteristics of Water Flow in Unsaturated Soils

In hydrology, unsaturated flow is important for downward vertical flow (natural and artificial recharge) (Smith, 1967), upward vertical flow (evaporation and transpiration), movement of pollutants from ground surface, and horizontal flow (Stallman, 1967) in the capillary zone above the water table.

The unsaturated hydraulic conductivity K_u is a function of the water content as well as negative pressure head (tension). Because part of the pore space is filled with air, the available cross-sectional area available for water flow is reduced, consequently, K_u is always less than the

the saturated value K.

The empirical relation between K_u and water content is of the form (Irmay, 1954) :

$$\frac{K_u}{K} = \left(\frac{S_w - S_{wc}}{1 - S_{wc}}\right)^3 \qquad (1.5)$$

where S_w is the degree of saturation and S_{wc} is the critical saturation - that part of the voids filled with nonmoving water held primarily by capillary forces.

1.4 Soil As A Conductor of Electricity

The earth is not a good conductor of electricity and, in comparison with the normal conductors, like metals, it is extremely poor. But it so happens that the cross section of the path taken by the current is very large, which results in lowering the actual resistance. This is mainly because of the fact that soils at the earth's surface are porous; and under any reasonable circumstances these pores are partly or completely filled with water. This water usually carries some salt in solution, so that the water content of a soil has a far great capacity for carrying current than does its solid matrix, unless highly conducting materials are present.

The relative abilities of materials to conduct electricity when a voltage is applied are expressed as conductivities. Conversely, the resistance offered by a material to current flow is expressed in terms of resistivity. The resistivity of a material is defined by the mathematical expression of ohm's law, which states that the electric field strength at a point in a material is proportional to the current density passing that point.

$$E = R\,j \tag{1.6}$$

where E is the electric field strength, in volts per meter and j is the current density in amperes per square meter and R is the resistivity in ohm-m. This definition is valid for use of direct currents (zero frequency).

The ohm-m may also be defined in terms of a hypothetical experiment in which the electrical resistance through a cube of material with dimensions of one meter on a side is measured. The resistance between opposite faces of such a cube is equal numerically to the resistivity of the material.

In most rocks, electricity is conducted electrolytically by the interstitial fluid, and resistivity is controlled more by porosity, water content and water quality than by the resistivity of the rock matrix. The resistivity of a water bearing soil decreases with increasing water content. In fully saturated formations, water content may be equated with porosity, but in partially saturated rocks, the effect of desaturation on resistivity need be considered.

Clay minerals are capable of conducting electricity electronically, and the flow of current in a clay layer is both electronic and electrolytic. Resistivity values for unconsolidated sediments commonly range from less than 1 ohm-m for certain clays or sands saturated with saline water, to several thousand ohm-m for dry basalts, dry sand, and gravel. The resistivity of sand and gravel saturated with fresh water ranges from about 15 to 600 ohm-m.

In the preceeding paragraphs, the case of saturated formations was discussed. However, pore spaces of a formation need not necessarily be completely filled with an electrolyte. In unsaturated zone, i.e. soil column above water table, part of the pore spaces may be filled with air. In this zone the percentage of water contained in the interstices changes with time and seasons, and in a different manner with different rock types.

The variation in distribution of water in the unsaturated zone is governed by circulation of water coming through atmospheric processes like precipitation, snow fall etc. Various steps involved in this process are (Keller and Frischknecht, 1970): (i) the infiltration of rain water from the surface into the soil immediately beneath the surface during rainy periods; (ii) the downward or lateral movement of this water through the unsaturated zone, (iii) return of the water to the atmosphere during dry periods or transpiration through plants. This circulation of water in the zone of aeration is an important factor in determining near-surface resistivities.

The amount of water held in the unsaturated zone changes slowly with time because evaporation or drainage of water through most soils is not rapid. Pore structures in soil may be very fine, so that capillary forces holding water in the fine pores of a soil are much larger than the gravitational force causing the moisture to drain downward. Frequently, it is found that the water content of the soil depends more on grain size than the distance above the water table. Fine grained zones hold more water than coarse-grained zones. For example, sandstones are partially desaturated and have a high resistivity while shales retain full water saturation by capillarity, and so, have a low resistivity.

A quantitative relationship between resistivity and the degree of saturation in a rock is of the form (Keller and Frischknecht, 1970):

$$\frac{R}{R_{100}} = S_w^{-n1} \; ; \; S_w > S_{wc} \qquad (1.7)$$

where R is the bulk resistivity of a partially saturated soil, R_{100} is the resistivity of the same formation when completely saturated with electrolyte, and n_1 is a parameter determined experimentally, and which usually has a value of approximately 2. Equation (1.7) holds provided the water content is greater than some critical value, S_{wc}, which depends on the texture of the soil. This critical saturation represents the least saturation for which there is a continuous film of water over all the surfaces in a soil. At high saturations, desaturation proceeds through the removal of small amounts of water from

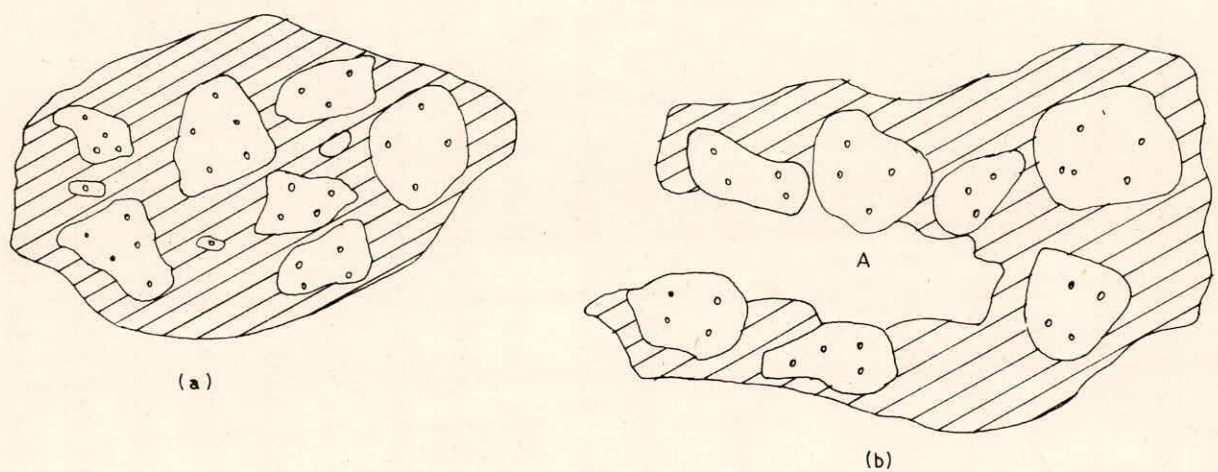

FIG. 2 — DISTRIBUTION OF MOISTURE BETWEEN SOIL PARTICLES: (a) CONTINUOUS FILM OF WATER UNTIL CRITICAL SATURATION, (b) EFFECT OF FURTHER DESATURATION AFTER CRITICAL SATURATION

the centres of large pores, as shown in fig. 2(a). This increases the overall resistance to current flow only moderately since the resistance through connecting pores (where most of the resistance is met) is not affected. Once the critical saturation has been reached, further desaturation will break the continuous film of water over the grains, as at point A in fig. 2(b), meaning that a small loss of water is accompanied by a large increase in resistance. At saturations below the critical saturation, the equation (1.7) is modified as follows:

$$\frac{R}{R_{100}} = aS_w^{-n2}, \quad S_w < S_{wc} \tag{1.8}$$

where a and n2 are parameters determined experimentally; a varies from about 0.05 for sandstones to about 0.5 for igneous rocks, and n2 has a value between 4 and 5.

The critical water saturation, S_{wc}, is about 25 percent of total pore space for permeable rocks and sands, but may be a large as 70 to 80 percent in compact rocks.

In areas of much rainfall, the water in the zone of aeration may have a very low conductivity since over long periods of time the circulating water will leache the exchange ions from the clay minerals, leaving behind ions which do not readily dissociate from the clay particles. In arid climates, the amount of evaporation from the ground surface may be greater than the recharge from rainfall, the extra water coming from migration of water from the permanent water table. This upward migration of water results in an increase in the salinity of the near surface water and in extreme cases, in the formation of caliche zones. In such areas, the increase in the salinity of water in the unsaturated zone may more than offset the effect of removal of water through evaporation etc., with the result that soils in arid areas are moderately conductive. This, therefore, sometimes leads to a paradox, i.e. in arid areas, soils tend to be more conductive than soils from humid areas.

With all this background, about soil and soil properties it is clear that soil moisture studies play important role in the fields of agriculture, meteorology and hydrology. It is useful for hydrologists and meteorologists in estimating infiltration evapotranspiration runoff relationships. Agriculturists use this information for yield estimation, erosion forecasting and irrigation scheduling.

Knowledge of soil moisture content, and its variation in space and time, is essential in conducting scientific studies for determining availability of water for various uses of the mankind. In order to conserve for future requirements, judicious use of available water resources requires precise knowledge of the existing and developable water resources.

Resistivity studies of unsaturated zone, i.e. subsurface above groundwater table, can provide information about soil moisture content in the subsurface by conducting field surveys on surface of the earth.

2.0 REVIEW

Obtaining reliable measurements of soil moisture is one of the most difficult problems in hydrology. Not only are there considerable difficulties in determining the water content of the soil at a particular point but also the spatial variability of soil properties makes it difficult to obtrain reliable as well as unique measurement. possibly, because of these reasons there has been a tendency to avoid making soil moisture measurements and instead rely on its estimates.

Some of the difficulties as referred above, are more concerned with the nature of the soil itself than with the actual techniques employed. For example, in case of methods utilizing buried sensors, there are problems of ensuring proper contact between the sensor and the soil, proper placement of sensor to avoid preferential movement of water into the hole in which it is buried, and to avoid its deterioration with time. There are also the difficulties of producing a satisfactory calibration.

Broadly there are two class of methods for measuring soil moisture. In one, soil samples are taken in fields and subsequently analysed in a laboratory. This causes disturbance in natural state of the soil; but laboratory analyses may provide more accurate results. These are known as direct methods. The other class of methods, knwon as indirect method, do not require extraction of soil samples; instead they measure in-situ the soil water content. In this case, accuracy of the results might be less than that of direct methods but undisturbed state of soil ensures higher reliability.

Sampling and drying (gravimetric) is the most common technique among direct methods. Electrical resistance blocks, neutron-probes, tensiometers are some of the conventional equipment among indirect methods. Some other techniques, e.g. gamma ray attenuation, use of ultrasonic waves, microwave moisture meters, hydrophotographic method, optical transmission method etc. have been used for the measurement of soil water content. However, they are not routinely used as compared to the other conventional techniques.

Conventional methods of soil water measurements, as referred above, have been described in number of text books and research papers; a review of these methods is given by Chand et al. (1985-86). This section is intended to review literature on use of electrical methods for determining soil moisture content. Some recent developments, mostly non-conventional methods, in techniques of soil water measurements are also included for sake of completeness of the review.

2.1 Optical Transmission Method

During experimental studies of rapid transient flows in unsaturated porous media, using two dimensional models, new methods are needed for measuring rapid changes in water-content.

The method is based on the fact that light transmission of quartz sand increases with water content in narrow samples. The increase in light intensity is dependent on light absorption moduli of sand, water and air, and also on the number of pores full of water.

Since relationship between number of pores full of water and the water content is not uniquely known a calibration procedure is needed for every photo sensitive probe and the sample

This method gives comparable results to the gamma-ray absorption method (Hoa, 1981). However, with the present state-of-art, the technique cannot be used with porous materials because of their very high light absorption modulus.

2.2 Microwave Scattering Method

In order to provide a 'non-destructive' alternative to the conventional methods, disturbing in-situ state of soil in one form or other, microwave remote sensing technique has been applied for more than a decade. Almost all initial attempts reported studies based on what is known as microwave backwardscattering and radiometry techniques. These were used to measure the surface soil moisture (upto approx penetration depth of 5 cm). Generally, the measured signal was calibrated by comparing it with gravemetric measurements.

Several methods to obtain in-situ measurements of soil moisture are based on the soil dielectric properties variations with its water content. In this method, need for calibration using gravemetric measurements. is eliminated because these proves make in-situ measurements. The dielectric probes give results which are comparable to the gamma neutron probe ones (Bernard et al., 1984). This technique is entirely automatic and the sampling period may be as small as desired.

A recent publication (Wallender et al., 1985) reports application of 'microwave forward scattering' method for measuring soil moisture on soil surface. Forward scattering of microwaves is a function of the soil water dielectric constant, via the reflection coefficient.

The basis of this method depends on the fact that the dielectric constant of water is very high (approx. 80) compared to that of dry soil (3 to 5). The dielectric constant of soil can be increased to more than

20 by adding water to dry soil. The increase translates into a change in some parameter which helps radar to sense the variation in moisture content.

Studies on effects of surface roughness, vegetation, moisture content, and grzing angle shows that reflection decreased with increased soil moisture and surface roughness. It was greater for horizontal than vertical polarized electric fields and decreased as the grazing angle and the soil dielectric constant increased. Vegetation cover resulted in less predictable results; but it was indicated that reflection was more characteristic of the vegetation than of the underlying soil. For near surface soil moisture measurements, forward scatter method of microwave transmission and reception can also be a promising tool, provided that user has enough funds for applying this method.

2.3 Time Domain Reflectometry System

In irrigation scheduling when to start and stop irrigation is based on monitoring the soil, which is the primary recipient of the applied water. Older methods, such as soil sampling and gravimetric determination of water, are so time consuming that the information is seldom available in time to control both the turn-on and the tunn off of the water supply.

Time domain reflectometry (TDR) is a technique operating over a range of radio frequencies, which can be used to measure the high frequency electrical properties of materials. In soil applications, TDR is used to measure the dielectric constant, which increases with increased moisture content in the soil.

In the TDR technique a step voltage pulse or signal is propogated along a transmission line. The signal's propagation velocity and the amplitude as well as the polarity of the reflected signal are dependent upon the electrical properties of the material making up the transmission line. Parallel pair transmission lines, as shown in fig.3, are normally used for measuring soil water contents. The parallel rods or wires serve as conductors and the soil, in which the rods are installed, serves as the dielectric medium. The pair of rods acts as a wave guide and the signal is reflected from the end of the transmission line in the soil and returns back to the TDR receiver. The timing device in the time-domain reflectometer measures the time between sending and receiving the reflected signal. This time interval relates directly to the propagation velocity of the signal in the soil since the line length is known. The propagation velocity is indicative of the volumetric water content, being smaller as the water content increases.

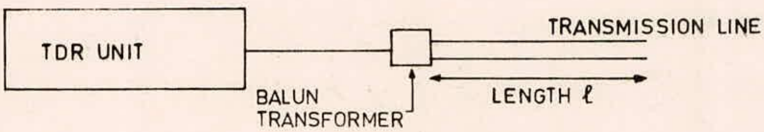

FIG. 3 - BLOCK DIAGRAM OF TDR UNIT FOR SOIL WATER CONTENT MEASUREMENT (AFTER TOPP AND DAVIS, 1985)

TDR method is capable of giving a precise and early determination of soil water content in fields with an accuracy of 2% (Topp and Davis, 1985). Moreover, because of strong dependence of the dielectric constant on water only, there is no need to calibrate the TDR for different soils. Measurement of soil salinity and detection of forzon layers within the soil, useful for irrigation as well as cultivation scheduling, could also be done with the TDR technique.

Although measurement of dielectric constant is possible using traditional frequency-domain method also, an appreciable amount of time is required for the measurements. Moreover, number of systems are required to obtain the measurements over a broad range of frequencies, resulting in expensive instrumentation, especially in the microwave region Time domain measurement methods (e.g. TDR) offer the possibility of overcoming the above limitations of frequency domain methods. Recent developments (Bellany et al., 1985) elaborate the TDR techniques for application in efficient measurements of soil water content.

2.4 Role of Soil Properties on Water Retention Cdharacteristics

Soil water retention characteristics are needed to describe availability of soil water to plants and to model movement of water and solutes in unsaturated soils. Measuring soil water characteristics curves is expensive and time consuming. However, routinely obtainable textural and structural soil properties can be used to estimate and predict water retention charcteristics. Regression analysis performed on experimental data comprising of particle size distribution, bulk density, and percentages of organic matter, sand, silt and clay gives predicted water content for a given soil matric potential (Gupta and Larson, 1979). Of late, there have been developments in instrumentation resulting in automatic laboratory measurement of soil organic matter and related constituents. Electronic sensors using silicon photo transistors and light emitting diodes have been reported (Griffis, 1985). The sensors are useful in predicting carbon content of the soil.

Such regression models may be used to estimate with reasonable accuracy water retention characteristics from particle size distribution, percentage of organic matter, and bulk density. This will be of particular help in modelling salt and water flow in soils and in estimating available water capacities. Water retention curves obtained from these regression equations may also be used to approximate hydraulic conductivity-water content relationship.

2.5 Geophysical Methods

Geophysical methods measure natural or induced physical phenomena in the earth's crust and interpret the measurements to obtain information on the subsurface. The complete spectrum of application of geophysical methods in hydrological problems was outlined by the author in another publication (Goyal et al. 1985-86). Few other publications elaborating such applications are Worthington (1975), Mandel and Shiftan (1981), Marino and Luthin (1982) etc. besides standard text-books available on the subject.

Geophysical methods, in general, are broadly classified as surface methods, applied from the earth's surface; and borehole methods, applied through drilled boreholes. For the purpose of investigating vadose (i.e. unsaturated) zone, surface geophysical methods are considered to be convenient and economical than the borehole methods.

Surface geophysical methods are based on measurements at the earth's surface of anomalies in physical force which must be interpreted in terms of subsurface lithology and variation in physio-chemical properties. Success in applying these methods depends on the existence of sufficient contrasts in the physical properties (e.g. electrical conductivity/resistivity) of subsurface formations. Although geophysical methods have been mostly applied in groundwater exploration, potential for their application in soil moisture studies was recently described by Goyal et al. (1986).

2.5.1 Electrical resistance units

The electrical resistance of a soil volume depends not only upon its water content, but also upon its composition, texture and soluble salt concentration. However, the electrical resistance of porous bodies placed in the soil and left to equilibrate with soil moisture can be calibrated against the water content. Such units, generally called electrical resistance blocks, contains a pair of electrodes embedded in gypsum, nylon, or fibre glass. The porous material, embedded with a pair of electrodes takes up water from the soil until a state of equilibrium is reached. The resistance between electrodes is governed by the moisture present in the porous material.

Porous blocks embedded in the soil tend to equilibrate with the soil moisture (matric) suction, rather than the soil moisture content directly. Different soils can have greatly differing wetness Vs suction relationships (e.g., a sandy soil may retain less than 5% moisture at, say, 15-bar suctions, whereas a clayey soil may retain three or four times as much). Hence, calibration of porous blocks against suction (tension) is basically preferable to calibration against wetness, particularly when the soil used for calibration is a disturbed sample differing in structure from the soil in-situ.

The equilibrium of porous blocks with soil moisture may be affected by the hysteresis i.e., the direction of change. Also, the hydraulic properties of the blocks (or inadequate contact with the soil) may impede the rapid attainment of equilibrium and cause a time lag between the state of water in the soil and the state of water being measured in the block.

The electrical conductivity of most porous blocks is due primarily to the permeating fluid rather than to the solid matrix. Thus, it depends upon the electrolytic solutes present in the fluid as well as upon the volume content of the fluid. Blocks made of inert materials, such as fibreglass, are highly sensitive to even small variations in salinity of the soil solution. On the other hand, blocks made of plaster of paris (gypsum) maintain a nearly constant electrolyte concentration corresponding to that of a saturated solution of calium sulphate. This tends to mask or buffer, the effect of small or even moderate variations in the soil solution. However, since gypsum is soluble, these blocks eventually deteriorate in the soil.

For these reasons, the evaluation of soil wetness by means of electrical resistance blocks is likely to be of limited accuracy. Soil moisture blocks have been found to be more dependable in the drier than in the wetter range (Johnson, 1962). An advantage of these blocks is that they can be connected to a recorder to obtrain a continuous record of soil-moisture variations in-situ. A.C. resistivity technique, described in following paragraphs, may prove to be able to overcome these limitations.

2.5.2 Resistivity metnod

In the electrical resistivity method, the distinctness of surface indications (signals) depends on the contrasts in the physical properties of geologic formations and their vicinity. An essential characteristic for the usuability of any geophysical method is continuity of physical properties. For resistivity 'sounding' procedure, as given in Chand et al. (1985-86) these, physical properties must remain continuous in a horizontal direction since the spacing of transmitting and receiving units is changed horizontally to obtain increased depth penetration.

The resistivity of earth materials is defined by the ohm-m, the electrical resistance of a cube of material with dimensions of 1m on a side. It varies widely with the material and its porosity, grain packing, water content, and conductivity. Dependence of resistivity on these parameters has been described in a number of text books, e.g. Keller and Frischknecht (1970) and Mandel and Shiftan (1981). A comprehensive discussion of the relationships is given in Qian (1985). Worthingtan (1975) has also elaborated the application of geophysical methods for quantitative investigations of aquifers.

Based on this scanty literature available, the author had initiated studies on application of resistivity technique for soil moisture estimation and measurement (Goyal et al., 1986).

3.0 PROBLEM DEFINITION

Most soils and rocksare non-conductors of electricity when they are in completely dry-state. Only exceptions are cases where soils contain certain minerals which are conductors because of their metallic content. When these formations contain water, the resistivity drops considerably and passage of current becomes possible. The resistivity of soil would be determined by the quantity of water held in the soil, and on the resistivity of the water itself. This implies that the conduction through soil is mainly electrolytic in nature i.e. ions present in the 'soil solution' (the water in pore-spaces) are responsible for passage of current.

The main factors which determine the resistivity of soil are :

i. Type of soil,

ii. Chemical composition of salt dissolved in the contained water,

iii. Concentration of the salts dissolved in the contained water,

iv. Moisture content,

v. Temperature of water,

vi. Grain size of the material and distribution of grain size,

vii. Packing arrangement of grains.

3.1 Effect of the Type of Soil

The type of soil is important in determining resistivity of the soil. Generally type of soil are not too clearly defined, for example, the word clay can cover a wide variety of soils. For this reason it is quite difficult to state that clay, or any other soil for that matter, has a resistivity of so many ohm-m. Again, the same general type of soil occurs in various localities and it is often found that the resistivity in one locality is different from that in another. The resistivity of various types of soil has been measured both by taking samples and measuring them in laboratory and by measurement of the undisturbed mass of the soil. Neither of the measurement is easy but the latter is more likely to give accurate results. Table I gives an idea of order of resistivities expected from various types of soils.

Table I : Typical Values of Resistivity Some Soils (after Tagg, 1964)

Type of soil	Resistivity in ohm-m
Loams, garden soils etc.	5 - 50
Clays	8 - 50
Clay, sand and gravel mixtures	4C - 250
Sand and gravel	6C - 100
Slates, shale, sandstone etc.	10 - 500
Crystalline rocks	200 - 10,000

3.2 Effect of Moisture Content

It is because of electrolytic conduction of current that the quantity of water and the nature and amount of the dissolved salts are the most important parameters in determining the resistivity of soil. The actual amount of water is dependent on a number of factors and is likely to be a variable quantity. It varies with the weather conditions, the time of year (affecting seasonal variation), the nature of the sub-soil and the depth to the permanent water table. It is rare that soil is really dry, though possibly desert sands would be the nearest approximation

On the other hand, soil with moisture content greater than about 50 percent is also not found very often.

Moisture content is likely to increase with increasing depth in most localities. However, it can not always be said that because a lot of water is present i.e. the moisture content is high, the resistivity is necessarily low. Distilled water filling the pore spaces will give a high resistivity value. Similarly frozen ground will necessarily have a high resistivity since ice has a high value.

3.3 Effect of Salinity of Water

Amount and nature of salts dissolved in the water govern its conductivity (reciprocal of resistivity). Since the amount of water present in the soil is a major factor in determining the resistivity, it follows that this resistivity is dependent on that of the water itself.

Natural groundwaters vary widely in composition and it is found that different salts have different effects on soil resistivity. Generally, the most common constituent in all such solutions is sodium chloride, alongwith other salts and oxidates (carbonates, sulphates phosphorates, etc.). Relatively low concentrations of dissolved salt are found in surficial waters, while connate waters may be very saline. The resistivity of natural water is determined mainly by the salinity. The concentration of salt in solution varies over wide limits, from 0.1mg/lt to 109 mg/lt and more. The resistivity of a solution may be calculated from the following formula:

$$\rho_v = \frac{10}{(Calafa + Ck\ lk\ fk)} \quad \text{ohm-m} \qquad (3.1)$$

where C_a and C_k are concentrations, in gram equivalents; l_a and l_k are mobilities, and f_a and f_k are the activity coefficients, of the anions and cations in solutions respectively.

The relationship between water resistivity and Nacl concentrations at a temperature of $20^{\circ}C$ is given in Table II.

Table II : Variation of Resistivity with Various Concentrations of Nacl

Concentration g/lt	Resistivity ohm-m	Concentration g/lt	Resistivity ohm-m
.005	1.05×10^3	1.0	5.8
.05	1.10×10^2	10.0	0.65
.50	1.2×10^1	50.0	0.15

Thus the above facts partly explain the difference in resistivities of similar soils from different localities. First the moisture content may be different, and also the quantity and nature of the dissolved salts may not be the same.

3.4 Effect of Grain Size and its Distribution

The grain size and its distribution has an effect on the manner in which the moisture is held. With large grains, moisture is held by surface tension at the points of contact with the grains. If grains of various sizes are present, the spaces between the large grains may be filled with smaller ones and as a result the resistivity is reduced.

Grains of soils are of all shapes and sizes, but it is possible to obtain some idea of the order of volume of the free space which can be filled with water, if the grains are assumed to be spherical in shape.

The most compact arrangement of these is obtained when the lines joining the centres of the spheres form an equilateral parallelopiped having face angles of 60^0 and 120^0. For such an arrangement the pore volume amounts to 25.95 percent of the total and is independent of the grain size. If the spheres are arranged in the least compact manner, the lines joining the centre form cubes and in this case the pore volume is 47.64 percent.

3.5 Effect of Temperature and Pressure

In water bearing earth materials, resistivity decreases with increase in temperature. In formations saturated with electrolyte, increased temperature lowers the viscosity of the electrolyte and the decreased viscosity helps the ions more freely. Electrical resistivity method is reported to be one of the most effective tools for geothermal energy prospecting (Jin, 1977).

At moderate temperatures, the dependence of resistivity on temperature for a rock saturated with an electrolyte is defined by the equation (Glass tone, 1937):

$$Rt = \frac{R15}{1 + \alpha_t\ (t - 15)} \tag{3.2}$$

where R_t is the resistivity at the ambient temperature t^0C, R_{15} is the resistivity measured at a reference temperature of 15^0C (any other reference temperature may be used), and α_t is the temperature coefficient of resistivity.

Regarding effect of pressure, various experimental works conducted on soil/rock samples have shown that increase in pressure results in decrease of resistivity. Increase in pressure results in increased compaction of the pore spaces in the soil mass and hence the

the effect. From a practical point of view the effect of pressure can safely be neglected.

3.6 Variation of Resistivity with Location

As discussed above, it is not always possible to allocate a definite value of resistivity to a given type of soil, as this varies with the location in which the soil is found. Moreover, the resistivity can change rapidly with location, due partly to change of moisture content and partly to change in the type of soil.

During measurement of resistivity, effects of natural and stray currents in the ground should be considered and properly accounted for.

Considering all the above effects, it is seen that the resistivity of the soil is a very variable quantity and, if it is desired to know the value at any given location, the only safe way is to measure it. This will give the value at the time of the measurement under the conditions prevailing at that time.

Apart from using the soil resistivity measurements for other engineering purposes, its main application would be in determining the variation of moisture content in soils. For latter application, we need to consider the effects of moisture content and other related soil parameters on the bulk resistivity of the soil.

3.7 Use of Resistivity Technique for Soil Moisture Variation

A number of empirical formulae relating bulk resistivity and porewater properties are available (e.g. Archie, 1942; Ransom, 1984; Qian, 1985, etc.) . The electrical properties of soils in unsaturated zone above water table are dependent on a number of parameters. Important among these are soil parameters like porosity, grain size and shape and, saturation and conductivity of water present in the soil Surface resistivity measurements using different techniques, for example, Schlumberger and Wenner methods can be utilised in determination of soil moisture status and its spatial as well as temporal variation.

However, application of this technique for soil moisture estimation requires knowledge of porosity and conductivity of soil water in the area under study. Before processing surface resistivity data for estimating soil water status, these two parameters should be determined. For applications concerning studies of temporal variation of soil moisture, these parameters can be determined initially for the specific site and then used in all subsequent processing.

Any change of apparent resistivity observed at the surface is due to the change of true resistivity within the medium. Archie's law is an empirical formula which describes the relationship between the bulk resistivity and the porosity of sedimentary rock in a form:

$$R = R_o \phi^{-m} S^{-n} \tag{3.3}$$

where, R is the bulk resistivity of the rock, ϕ is the porosity of the rock, S is the degree of saturation of the rock, R_o is the resistivity of electrolyte in pores/cracks, m and n are empirical constants, depending upon the grain shape, size, their packing arrangement and compaction properties of the rock.

As a complement to equation (3.3), formation factor as soil parameter is defined by:

$$F = \phi^{-m} S^{-n} \tag{3.3a}$$

As is seen in the equation, formation factor unlike bulk resistivity is independent of electrolyte resistivity.

Equation (3.3) can also be utilized in studying relationship between the bulk resistivity, porosity and saturation in unconsolidated formation, e.g. sands. Values of empirical constants m and n for such unconsolidated formations are reported to be around 1.5 and 2 respectively. Therefore, Archie's formula for unconsolidated formations is represented by (Kelly, 1985).

$$R = R_o \phi^{-1.5} S^{-2} \tag{3.4}$$

and corresponding relation for formation factor is represented by:

$$F = \phi^{-1.5} S^{-2} \tag{3.4a}$$

In equation (3.3), ϕ is defined by

$$\phi = V_p/V_T \tag{3.5}$$

where V_T is the total volume of rock, V_p is the pore volume inside the rock, which forms an interconnected system, and S is defined by

$$S = V_w/V_P \tag{3.6}$$

where V_w is the volume of electrolyte in the pore volume.

From equations 3.3 - 3.6, it can be seen that if the environmental conditions change then variations in the parameters V_T, V_P, V_w and/or R_o may occur, and as a consequence the resistivity R of the rock will change. To discuss the vibrationof bulk resistivity of rock by applying Archie's law, three classes of effects may be considered. They are(i) the effect of a change in total volume of rock, (ii) the effect of change in water bearing volume of rock pores, which governs the saturation, and (iii) the effect of change in electrolyte resistivity. However, in the present study we have been interested in studying the effect of soil saturation on the bulk resistivity. Some variation in the value of electrolyte (water) conductivity has also been considered.

A simple and general way of discussing effect of saturation on the bulk resistivity is to partially differentiate both sides of equation (3.3) with respect to Δv_w. This results in

$$\frac{\Delta R}{R} = -n \frac{\Delta Vw}{Vw}$$

Using equation(2),

$$\frac{\Delta R}{R} = - 2 \frac{\Delta Vw}{Vw} = -2 \frac{\Delta \theta}{\theta}$$

where $\frac{\Delta \theta}{\theta}$ represents the relative change in volumetric water content present within the pores in the soil.

Using equation (3.7) for a particular site, relative change in volumetric water content at various depths of the soil can be estimated given the resistivity at the corresponding depths.

4.0 METHODOLOGY AND DATA

Soil wetness is the parameter used in estimating soil moisture using electrical resistivity technique. As defined earlier, it is usually expressed as a dimensionless ratio of water volume to total soil volume or water mass to dry soil mass. These ratios are usually multiplied by 100 and reported as percentages by volume or by mass. For the purpose of this report, the unit of soil wetness used is the percent by volume (% Vol.).

In order to study variation of resistivity with soil moisture, effect of two parameters i.e. soil - porosity and resistivity (or conductivity) of electrolyte was considered important for reasons discussed earlier. Using equation (3.4a), representing relation between formation factor and porosity as well as moisture content (θ) , formation factors were computed for widely varying soil moisture contents, at various porosities of the soil. As formation factor is independent of electrolyte resistivities, corresponding resistivities were obtained by multiplying it with the electrolyte resistivity (RW). Also, effect of electrolyte resistivity on bulk resistivity of the soil was considered by computing resistivities for varying soil moisture content, at various RWs.

Then, as given in equation (3.7), relative change in bulk resistivity should be twice the relative change in moisture content of the soil. In order to estimate the relative change in soil moisture content with respect to relative change in bulk resistivity, the following equation can be applied.

$$\frac{\Delta\theta}{\theta} = -\frac{1}{2}\,\frac{\Delta R}{R} \qquad \ldots(4.1)$$

The parameter $\frac{\Delta R}{R}$ on right hand side of the equation is derived from the resistivity measurements. Resistivity surveys provide data in terms of apparent resistivities at various electrode spacing for a particular survey technique (Wenner method in the present study). These apparent resistivity values need be converted to true or bulk resistivities at various depths in the subsurface. A number of methods were available (Zohdy et al., 1980) for deriving true resistivities from the apparent resistivity data. The method used for the present study utilizes a computerised curve matching procedure (Schmschal, 1981) for interpretation of field data in terms of layer parameters i.e. the true resistivities and the corresponding layer thicknesses. The series $\Delta R/R$ was calculated for the complete set of true resistivity values for each resistivity sounding.

Resistivity data for computing the relative changes in soil moisture contents has been taken from Goyal et al. (1986). Apparent resistivity data at various electrode spacings for the four typical sets of soundings is reproduced in Table III.

Sl. No.	Electrode Spacing (m)	30.9.85 App.Resistivity (ohm-m)	3.10.85 App.Resistivity (ohm-m)	18.12.85 App.Resistivity (ohm-m)	21.12.85 App.Resistivity (ohm-m)
1.	.25	47.88	48.83	129.99	129.99
2.	.50	50.87	55.89	125.91	143.81
3.	.75	41.92	48.51	96.55	105.50
4.	1.00	38.31	40.19	70.96	75.99
5.	1.25	32.18	34.54	57.38	59.66
6.	1.50	27.32	30.14	46.34	45.69
7.	1.75	25.28	27.47	40.66	39.67
8.	2.00	22.61	26.38	35.67	35.04
9.	2.25	22.61	25.43	32.52	31.79
10.	2.50	21.98	25.12	30.77	29.20

Table III : Apparent Resistivity data for four typical sets of resistivity-soundings (reproduced from Goyal et.al., 1986).

5.0 APPLICATION

Use of resistivity technique in ground water exploration is very well established. For studies in unsaturated zone, the technique has been applied for monitoring variation in soil moisture content. Preliminary studies have shown promising results on correlating soil resistivity variations with its water content (Goyal et al., 1986 and Chand et al. 1985-86).

Various forms of Archie's formula have long been used for estimating effect of saturation on bulk resistivity of a formation. Effect of quality, more specifically salinity of saturating water is also included in the formulae. Porosity governs saturation in a formation and therefore this parameter is also incorporated in the formulae.

Using one such empirical formula, soil moisture was estimated for various bulk resistivities. Bulk resistivities were determined from field resistivity data, using computerised curve matching technique. Field surveys were conducted in the National Institute of Hydrology Campus, and four sets of Wenner resistivity sounding data utilised for computation purposes.

Results show that resistivity technique is effective in monitoring variation of soil moisture content, especially during dry periods Further studies have already been initiated to study variation of resistivity during relatively wet conditions of soil formations.

6.0 RESULTS

Variation of formation factor and resistivity with soil moisture content for various porosities at a typical electrolyte resistivity (RW-20 ohm-m) was computed and results given in Table IV. Five porosities were chosen between 0.30- 0.50, considering typical values of soil porosity. Soil moisture varied between 10-60 percent by volume, covering almost complete range as generally encountered in fields. Plot between formation factor and moisture content for various porosities is shown in figure 4.

Table V shows results on variation of resistivity with soil moisture for various electrolyte resistivities (RW) at a typical porosity (N=0.35). Six resistivities typical for electrolyte were chosen (16-26) to represent values for a site encountering uncontaminated groundwater. Soil moisture variation was kept same as in the previous case. The two sets of results show that variation in resistivity is quite high for low saturations of soils. As water content increases, variation in resistivity is reduced, and for saturations more than .50, it becomes less prominent.

Using apparent reesistivity data from Table III, true resistivities and layer thicknesses were computed; these are given in table VI. For the four sets of data, soil moisture content corresponding to these bulk resistivities was computed using equation (3.4) and tabulated against resistivities in table VI.

Finally, relative change in moisture content, $\frac{\Delta\theta}{\theta}$, was computed using equation (3.7). Relative change in resistivities was determined using resistivity data from table VI. Negative sign in the equation suggests the fact that soil moisture decreases with increase in resistivity value and vice-versa. Results of computations of relative changes in resistivity and moisture content are given in table VII.

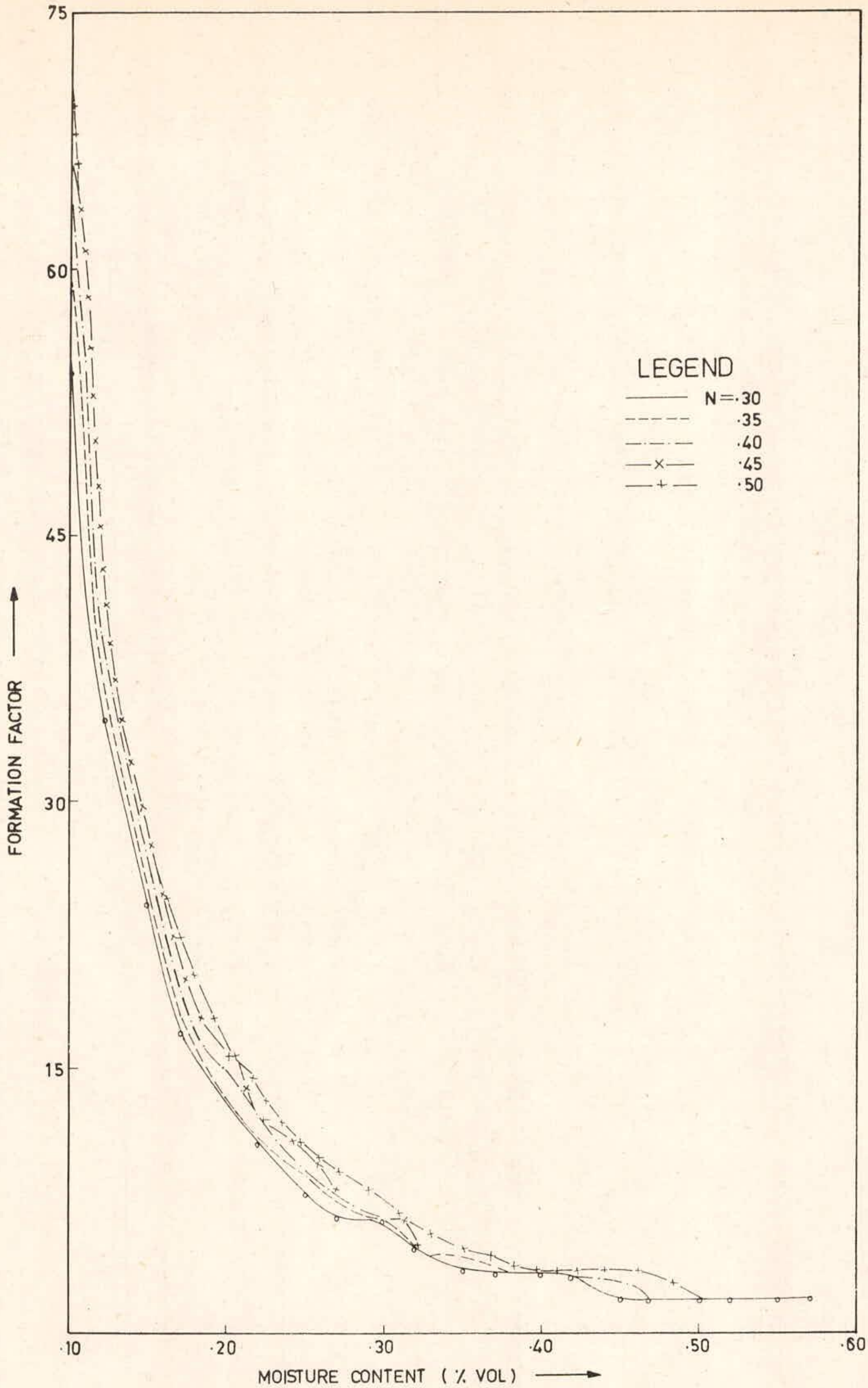

FIG. 4 — PLOT BETWEEN FORMATION FACTOR AND MOISTURE CONTENT FOR VARIOUS POROSITIES

Table IV : Variation of Resistivity with Moisture Content for Vorious (Bulk) Porosities (RW=20 ohm.m)

S. NNo.	Soil Moisture Thets (% of Vol.)	F.factor	R.(ohm.m)	F.factor	R.(ohm.m)	F.factor	R.(ohm.m)	F.factor	R.(ohm.m)	F.factor	R.(ohm.m)
!.	.100	54.772	1095.4	59.161	1183.2	63.246	1264.9	67.082	1341.6	70.711	1414.2
2.	.125	35.053	701.1	37.863	757.3	40.477	809.5	42.933	858.7	45.255	905.1
3.	.150	24.343	486.9	26.294	525.9	28.109	562.2	29.814	596.3	31.427	628.5
4.	.175	17.885	357.7	19.318	386.4	20.652	413.0	21.904	438.1	23.089	461.8
5.	.200	13.693	273.9	14.790	295.8	15.811	316.2	16.771	335.4	17.678	353.6
6.	.225	10.819	216.4	11.686	233.7	12.493	249.9	13.251	265.0	13.968	279.4
7.	.250	8.764	175.3	9.466	189.3	10.119	202.4	10.733	214.7	11.314	226.3
8.	.275	7.243	144.9	7.823	156.5	8.363	167.3	8.870	177.4	9.350	187.0
9.	.300	6.086	121.7	6.573	131.5	6.027	140.5	7.454	119.1	7.857	157.1
10.	.325	5.186	103.7	5.601	112.0	5.988	119.8	6.351	127.0	6.695	133.9
11.	.350	4.471	89.4	3.829	96.6	5.163	103.3	5.176	109.5	5.772	115.4
12.	.375	3.895	77.9	3.207	84.1	4.97	89.9	4.770	95.4	4.028	100.6
13.	.400	3.423	68.5	3.898	74.0	3.953	79.1	4.193	83.9	4.419	88.4
14.	.425	3.032	60.6	3.275	65.5	3.501	70.0	3.714	74.3	3.915	78.3
15.	.450	2.705	54.1	2.922	58.4	3.123	62.5	3.313	66.3	3.492	69.8
16.	.475	2.428	48.6	2.622	52.4	2.803	56.1	2.973	59.5	3.134	62.7
17.	.500	2.191	43.8	2.366	47.3	2.530	50.6	2.683	53.7	2.828	56.6
18.	.525	1.987	39.7	2.146	42.9	2.295	45.9	2.434	48.7	2.565	51.3
19.	.550	1.811	36.2	1.756	39.1	2.091	41.8	2.218	44.4	2.338	46.8
20.	.575	1.657	33.1	1.769	35.8	1.913	38.3	2.029	40.6	2.139	42.8
21.	.600	1.321	30.4	1.643	32.9	1.757	35.1	1.863	37.3	1.961	39.3

F = Formation, R = Resistivity

Table V : Variation of Resistivity with Soil Moisture for various RW (Porosity, N = .35)

S.No.	Soil Moisture Theta (% Vol.)	Resistivity AT RW =					
		16	18	20	22	24	26
1.	.100	946.6	1064.9	1183.2	1301.5	1419.9	1538.2
2.	.125	605.8	681.5	757.3	833.0	908.7	984.4
3.	.150	420.7	473.3	525.9	578.5	631.0	683.6
4.	.175	309.1	347.7	386.4	425.0	463.6	502.3
5.	.200	236.6	266.2	295.8	325.3	355.0	384.5
6.	.225	187.0	210.3	233.7	257.1	280.5	203.8
7.	.250	151.5	170.4	189.3	208.2	227.2	246.1
8.	.275	125.2	140.8	156.5	172.1	187.7	293.4
9.	.300	105.2	118.3	131.5	144.6	157.8	170.9
10.	.325	89.6	100.8	112.0	123.2	134.4	145.6
11.	.350	77.3	86.9	96.6	106.2	115.9	125.6
12.	.375	67.3	75.7	84.1	92.6	101.0	109.4
13.	.400	59.2	66.6	74.0	81.3	88.7	96.1
14.	.425	52.4	59.0	65.5	72.1	78.6	85.2
15.	.450	46.7	52.6	58.4	64.3	70.1	76.0
16.	.475	42.0	47.2	52.4	57.7	62.9	68.2
17.	.500	37.9	42.6	47.3	52.1	56.8	61.5
18.	.525	34.3	38.6	42.9	47.2	51.5	55.8
19.	.550	31.3	35.2	39.1	43.0	46.9	50.8
20.	.575	28.6	32.2	35.8	39.4	42.9	46.5
21.	.600	26.3	29.6	32.9	36.2	39.4	42.7

Table VI: Estimated moisture contents in clean sands, for four sets of resistivity data

Date :		30.9.85		31.10.85		18.12.85		21.12.85	
S.No.	Layer Thickness (m)	Bulk Resistivity (ohm.m)	Soil Moist. (% Vol.) for RW=20 ohm.m	Bulk Resistivity (ohm.m)	Soil Moist. (% Vol.) for RW=20 ohm.m	Bulk Resistivity (ohm.m)	Soil Moist. (% Vol.) for RW=20 ohm.m	Bulk Resistivity (ohm.m)	Soil Moist. (% Vol.) for RW=20 ohm.m
1.	0.25	40.00	53.95	40.00	54.17	130.00	28.74	130.00	28.79
2.	0.10	56.88	45.25	57.83	45.05	120.99	29.79	125.99	29.24
3.	0.15	52.88	46.93	61.00	43.87	123.50	29.48	128.50	28.96
4.	0.20	41.87	52.74	46.89	50.04	116.91	30.30	134.81	28.27
5.	0.28	35.00	57.68	41.00	53.51	91.00	34.35	103.50	32.26
6.	0.39	29.00	63.37	32.00	60.57	63.00	41.28	69.00	39.52
7.	0.55	21.00	74.46	23.00	71.44	42.00	50.56	43.00	50.06
8.	0.78	13.00	94.64	17.50	81.90	28.00	61.92	27.00	63.17
9.	1.08	16.00	85.31	18.00	80.76	20.50	72.36	19.00	75.30

Table VII : Relative changes in bulk resistivity ($\frac{\Delta R}{R}$) and computed moisture-content ($\frac{\Delta\theta}{\theta}$)

Date :	30.9.85		3.10.85		18.12.85		21.12.85	
S.No.	$\frac{\Delta R}{R}$	$\frac{\Delta\theta}{\theta}$	$\frac{\Delta R}{R}$	$\frac{\Delta\theta}{\theta}$	$\frac{\Delta R}{R}$	$\frac{\Delta\theta}{\theta}$	$\frac{\Delta R}{R}$	$\frac{\Delta\theta}{\theta}$
1.	0.397	-0.161	0.446	-0.168	-0.069	0.036	-0.031	0.016
2.	-0.070	0.037	0.055	-0.026	0.021	-0.010	0.020	-0.010
3.	-0.208	0.124	-0.231	0.141	-0.053	0.028	0.049	-0.024
4.	-0.164	0.094	-0.126	0.069	-0.222	0.134	-0.232	0.141
5.	-0.171	0.099	-0.220	0.132	-0.308	0.202	-0.333	0.225
6.	-0.276	0.175	-0.281	0.179	-0.333	0.225	-0.377	0.267
7.	-0.381	0.271	-0.239	0.146	-0.333	0.225	-0.372	0.262
8.	0.231	-0.099	0.029	-0.014	-0.268	0.169	-0.296	0.192

7.0 CONCLUSION

Effects of variation of soil moisture on resistivity measurements during dry and wet weather conditions, as shown by corresponding soil moisture contents, has been studied.

Results show that for high soil water contents (greater than 50%) variation in resistivity is negligible. For lower water contents, variation is uniformly high leading to better correlation between soil moisture and resistivity.

Relative change in resistivity is approximately twice the relative change in soil moisture (table VII). Thus, it is possible to monitor soil moisture variation using resistivity data. Porosity of a formation normally does not vary with time; once a particular site is defined in terms of porosity and other physical properties, measurements of resistivity become independent of these parameters.

Also, it is seen that resistivity of saturating electrolyte has considerable effect on formation resistivity. But, it is fairly reasonable to assume that electrolyte resistivity does not change at a particular site, at least within the zone of interest for soil moisture studies. However, for sites with possible contamination from land fill, industrial, sewage wastes, etc., effect of variation in electrolyte resistivity on resistivity of the formation need be considered.

It is hoped that variations of soil moisture on long term basis can be predicted using long period measurements. Effect of salinity, presence of organic matter and evapotranspiration on resistivity need be considered for getting insight details on behaviour of soil resistivity variations.

On the basis of this study, it is seen that soil resistivity is affected by soil moisture content in the top surface layers. Field measurements of soil resistivity on daily basis are necessary to find the effects of soil evapotranspiration, soil temperature and rainfall on resistivity-moisture relationship.

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