Technical Note TN-5

ELECTRICAL CONDUCTIVITY

OF

SOILS AND ROCKS

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I. INTRODUCTION

It was near the beginning of this century that Conrad Schlumberger first employed the technique of mapping sub-surface geology by injecting electrical currents into the ground and mapping the resulting potential field distribution. Since that time measurement of terrain resistivity has been applied to a variety of geological problems. A partial list of applications includes the determination of rock lithology and bedrock depth; the location and mapping of aggregate and clay deposits; mapping groundwater extent and salinity; detecting pollution plumes in groundwater; mapping areas of high ice content in permafrost regions; locating geothermal areas; mapping archaeological sites, etc.

In many instances resistivity mapping provides definite geological information; however there are also cases where the results are uninterpretable since the "geological noise" is too high. A limitation of resistivity surveying is that the actual value of terrain resistivity itself is seldom diagnostic. As a result of this ambiguity we generally examine the variations of resistivity, either laterally or with depth, to outline the geological features of interest. But a problem arises in that conventional surveys are time-consuming to carry out and the area actually surveyed is often smaller than one might wish in order to fully ascertain the background against which the anomalous feature is to be defined. Furthermore, although conventional resistivity techniques sense to a characteristic depth (determined by the interelectrode spacing) resistivity inhomogeneities much smaller than that depth can, if they are located near the potential electrodes, yield large errors in the measurement and thus a noisy survey profile.

For these reasons application of resistivity surveys to engineering problems is not as common as it might be, particularly in North America. Such surveys have achieved success in Europe and are used more routinely.

It was an awareness both of the potential of resistivity measurements for solving geological problems and equally of some of the drawbacks of conventional resistivity mapping which lead Geonics Limited to develop two new lines of instrumentation employing electromagnetic techniques to measure terrain conductivity. In the first of these a sinusoidally varying magnetic field electromagnetically induces currents in the ground in such a manner that their amplitude is linearly proportional to the terrain conductivity (reciprocal of resistivity). The magnitude of these currents is determined by measuring the magnetic field which they in turn generate. Through the use of electromagnetic techniques, ground contact is avoided and with these patented instruments it is possible to map terrain conductivity virtually as fast as the operator(s) can walk; furthermore the sample volume is averaged in such a manner as to yield excellent resolution in conductivity. Two instruments have been developed by Geonics to cover the range of depths generally useful for engineering geophysics: (i) the EM31, one-man portable, has an effective penetration depth of 6 meters and (ii) the EM34-3, two-man portable, has stepwise selectable depths from 7.5 meters to 60 meters.

In the second approach the current flowing in a loop situated on the ground is abruptly terminated, inducing eddy currents in the ground which diffuse away from the transmitter loop in a manner controlled by the ground conductivity. In this case the dispersal of the currents is determined by measuring the transient decay of their magnetic field. Based on this principle, the Geonics EM37 can be used to determine the electrical properties of the earth to depths of several hundred meters.

These devices are assisting in the solution of many geological problems. With the renewed interest in resistivity it has become apparent that there is a requirement for a short note which discusses, from the point of view of survey interpretation, the various factors that influence terrain resistivity. For example, typical questions raised include (1) When are gravels more resistive than finer material? (2) What is the influence of the depth to water table? (3) It rained heavily last week; will this affect the measurements? (4) What are typical resistivities for the following soil types?, etc., etc.

By describing the various factors that control the electrical conductivity/resistivity of soils and rocks under typical in-situ conditions this note will attempt to provide the technical background against which these questions can be answered and to thus give the operator greater confidence in his survey interpretation. The emphasis throughout this technical note is on those factors that influence the near-surface ground resistivity, particularly of soils.

The various topics are dealt with in the following sequence.

Section II: definition of conductivity & resistivity
Section III: description of relevant physical properties of soils and rocks
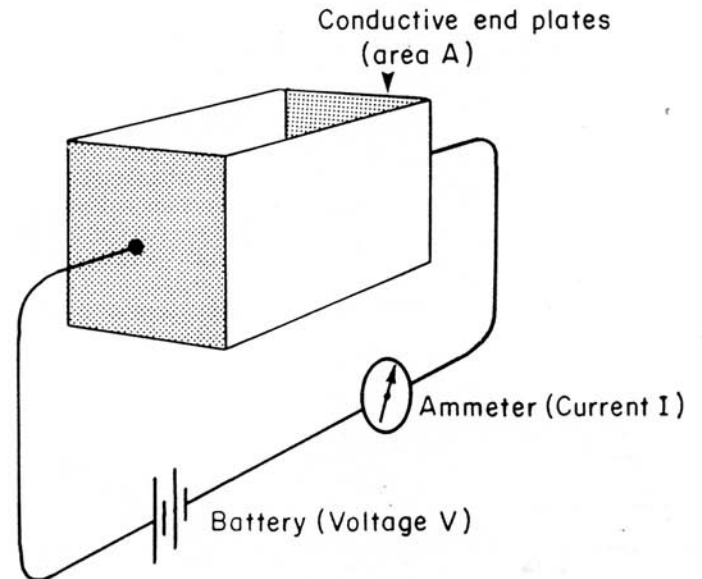Section IV: relation of physical properties to the electrical conductivity
Section V: discussion of conventional resistivity measurements of soils in different climatic zones.

As mentioned above, this technical note is directed towards the user. Many readers will be familiar with the physical properties of soils and we ask their indulgence. None of the material on the electrical properties of soils is original; some has been taken verbatim from the references. However the compilation has been derived from many sources and we hope that it will prove useful.

II. RESISTIVITY/CONDUCTIVITY

Basically the electrical resistivity/conductivity of a substance is a measure of the difficulty/ease with which an electrical current can be made to flow through it. Suppose that we construct a tank with insulating sides and two conductive end plates as shown in Figure 1. The end plates are connected to a battery through an ammeter with which we measure the current flow through the sample. If the tank is empty there is of course no path by which the current can flow from one end plate to the other and the ammeter reads zero.

Suppose we fill the tank with a mixture of clean gravel and tap water. The ammeter will read a finite current and, depending on

RESISTIVITY (ρ)	CONDUCTIVITY (σ)
$\rho = \dfrac{RA}{L}$ ohm·metres	$\sigma = \dfrac{GL}{A}$ mhos/meter
where $R = \dfrac{V}{I}$ ohms	where $G = \dfrac{I}{V}$ mhos

FIGURE 1. Resistivity/conductivity tank.

various parameters to be discussed in Section IV, might indicate a current flow of a few thousandths of an ampere i.e. a few milliamperes, for a voltage of one volt. If the ammeter reads two milliamperes the electrical resistance of the material in the tank, given by the ratio of the voltage divided by the current, is 500 ohms.

If we now fill a number of different rectangular tanks, of different dimensions, with the same mixture we would find that the electrical resistance is proportional to the length of the tank and inversely proportional to the area of the conductive plate electrodes. The constant of proportionality, a property of the mixture only, independent of the tank dimensions, is defined as the electrical resistivity. In the MKS system it has the units of ohm-meters and is the electrical resistance measured on a cubic sample whose dimensions are all one meter. In the CGS system of units the resistivity is defined as the resistance across two opposite faces of a cubic sample one centimeter on each side and the units are ohm-centimeters. From the defining relation we see that, given the resistivity in ohm-centimeters, we must divide by 100 in order to get the resistivity in ohm-meters.

Suppose that we empty our tank (now assumed to be one meter on a side) and fill it with a mixture of clay saturated with water. We might find that the ammeter read a few tenths of an ampere, perhaps as much as half an ampere. The resistance would then be two ohms, the resistivity two ohm-meters. The range of resistivities displayed by unconsolidated materials at temperate ambient temperatures usually lies between one ohm-meter and one thousand ohm-meters; the resistivity of rocks can vary from a few tens of ohm-meters to as high as 100,000 ohm-meters, discussed in further detail in Section IV.

The reciprocal of the electrical resistivity of our sample is defined as the electrical conductivity. In the MKS system the unit of conductivity is the mho per meter and a resistivity of one ohm-meter exhibits a conductivity of one mho per meter, 100 ohm-meters is equivalent to a conductivity of 0.01 mhos per meter, etc. The electromagnetic instruments actually measure terrain *conductivity* rather than *resistivity* and for this reason much of the remainder of this technical note will be concerned with the conductivity of various terrain materials rather than the resistivity.

It was stated above that a resistivity of 100 ohm-meters corresponds to a conductivity of 0.01 mhos per meter. To avoid the inconvenience of having zeros immediately following the decimal point all conductivities will be expressed in millimhos per meter: a conductivity of .01 mhos per meter corresponds to 10 millimhos per meter, etc. This has the advantage that the range of resistivities from 1 to 1,000 ohm-meters is covered by the range of conductivities from 1,000 to 1 millimhos per meter and such numbers are easily handled. Table 1 lists the conversion from resistivity and conductivity in various units to conductivity in millimhos per meter.

In the experiments described above a direct current was employed. Had we used an alternating current and varied the frequency we might have discovered that the electrical properties of the sample varied with frequency. In reality soils and rocks are complex substances in which there are many (some poorly understood) mechanisms which govern the mode of current flow through the sample [1–5].

Fortunately, for materials with conductivity of the order of one to 1.000 millimhos per meter the electrical properties which control the current flow are relatively independent of frequency and the DC or low frequency conductivity measured with conventional resistivity equipment will be essentially the same as that measured using low frequency electromagnetic techniques.

Finally it should be noted that measurement of the electrical conductivity or resistivity of geological samples is, in reality, a very difficult procedure requiring much more complex equipment than the simple tank referred to above.

III. FACTORS AFFECTING TERRAIN CONDUCTIVITY

Most soil and rock minerals are electrical insulators of very high resistivity. However on rare occasions conductive minerals such as magnetite, specular hematite, carbon, graphite, pyrite and pyrrhotite occur in sufficient quantities in rocks to greatly increase their overall conductivity. This note assumes that such minerals are absent.

In general the conductivity is electrolytic and takes place through the moisture-filled pores and passages which are contained within the insulating matrix. The conductivity is therefore determined for both rocks and soils by
(1) porosity; shape and size of pores, number, size and shape of interconnecting passages
(2) the extent to which pores are filled by water i.e. the moisture content
(3) concentration of dissolved electrolytes in the contained moisture
(4) temperature and phase state of the porewater
(5) amount and composition of colloids
Since the constituents, structure, and included moisture of soil or rock are of great importance in determining the conductivity, this section will discuss the various physical and chemical properties in sufficient detail to illustrate how they affect the material conductivity.

III. 1. Soil Constituents
Soils consist basically of four components namely (i) mineral material (ii) organic material (iii) water and (iv) gases.

Figure 2 shows schematically the relative proportion of these components in an unsaturated loam soil (to be defined later). The illustration refers to a temperate zone soil; for humid region soils about 45% of the solid space may be occupied by clay minerals and 5% by organic matter.

The mineral fraction of a soil can be extremely variable, however some generalizations can be made based on grain size [6]:
sand size range – mostly quartz, small amounts of feldspar and mica present, all grains coated to some extent with iron and aluminum oxide (the rusty reddish-brown colour observed is due to the iron).
silt size range – quartz still dominant, less feldspar and mica than in the sand fraction, more iron and aluminum

TABLE 1. Resistivity/conductivity unit conversion factors

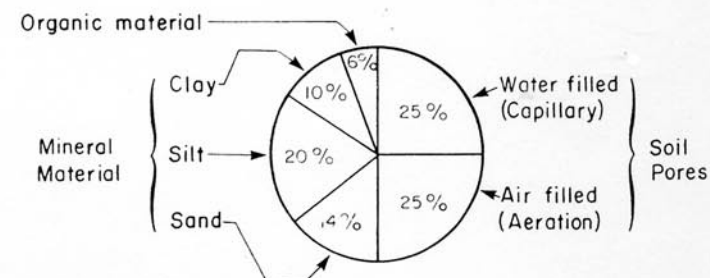|	
mhos/meter × 1000	⟶ millimhos/meter
$\dfrac{1000}{\text{ohm-meters}}$	⟶ millimhos/meter
$\dfrac{100,000}{\text{ohm-centimeters}}$	⟶ millimhos/meter
$\dfrac{\text{cgs electrostatic units}}{9 \times 10^{6}}$	⟶ millimhos/meter

Note: 1 Siemen (S) = 1 mho

FIGURE 2. The volumetric composition of a loam soil when excess water has been removed. On a weight basis the percentage composition of the dry soil would be: organic matter 4%, clay 22%, silt 44% and sand 30% (after L.R. Webber [6]).
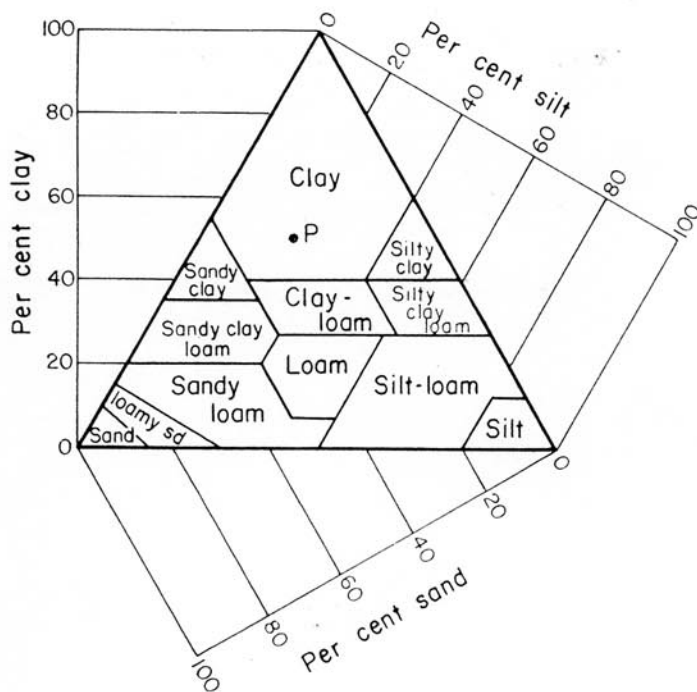
FIGURE 3. U.S. Department of Agriculture textural classification triangle with axes added. The point P represents a clay (soil) containing 50 per cent clay, 20 per cent silt, and 30 per cent sand (after D. Kirkham [7]).

oxides in the form of coatings due to the greater particle surface area on which the coatings may reside.

clay size range – finely divided quartz, feldspar, mica, iron and aluminum oxides, all of which appear in the coarser part of the clay fraction (0.002 millimeters – 0.001 millimeters in diameter); the finer part (less than 0.001 millimeters) is colloidal and consists mainly of layer silicates with smaller amounts of iron and aluminum oxides.

As indicated above soils are classified on the basis of texture or grain size independently of the mineralogical content of each particle size component. Sand is defined as particles with diameters between 0.05 millimeters and 2 millimeters, silt has diameters between 0.002 millimeters and 0.05 millimeters and clay has diameters less than 0.002 millimeters. Figure 3 shows the classification triangle for various types of soil and is useful as it allows an estimate of the clay content which often consists essentially of clay minerals which affect the soil conductivity.

The minerals in the sand and silt fractions of the soil are electrically neutral and are generally excellent insulators. Completely dry clay is also an insulator but the introduction of moisture changes the situation radically. Clay consists of microscopically fine particles

TABLE 2. Exchange capacity of common clays (after Keller and Frischknecht [5])

Clay	Exchange Capacity
Kaolinite	3 to 15 m-equiv/100 g
Halloysite . 2H$_2$O	5 to 10
Halloysite . 4H$_2$O	40 to 50
Montmorillonite	80 to 150
Illite	10 to 40
Vermiculite	100 to 150
Chlorite	10 to 40
Attapulgite	20 to 30

TABLE 3. The cation exchange capacity (CEC) and colloid content of five soils of different textures (after L. R. Webber [6]).

Soil Texture	Organic Matter %	Clay %	CEC* me/100 g
Sand	1.7	7	6.3
Sandy loam	3.2	13.2	13.7
Loam	4.9	16.8	20.2
Silt loam	5.4	18.4	24.0
Clay loam	5.5	31.2	27.2

*The cation exchange capacity of a soil is expressed in terms of milliequivalents per 100 grams of soil (meq/100 g). A milliequivalent is defined as 1.0 milligram of hydrogen or the amount of any other element that will combine with or displace it. The milliequivalent weight of any element may be found as follows:

$$\frac{\text{atomic weight of element}}{\text{valence of element} \times 1000} = \text{milliequivalent weight}$$

exhibiting a sheet-like structure, for which reason clays are often called "layer silicates". Composed of stable secondary minerals that have formed as a result of weathering of primary minerals such as feldspar, mica, etc, the particles are so fine-grained that they are described as micro-crystals.

Their crystalline structure is such that, as a result of crystal imperfections, the surface appears to be negatively charged [5]. During the formation of the clay through weathering, positive charges (cations) are adsorbed to the surface. These cations (typically Ca, Mg, H, K, Na, NH$_3$) are loosely held to the surface and can subsequently be exchanged for other cations or essentially go into solution should the clay be mixed with water. For this reason they are called exchangeable ions and the cation exchange capacity (CEC) of the soil is a measure of the number of cations that are required to neutralize the clay particle as a whole i.e. the weight of ions in milliequivalents adsorbed per 100 grams of clay.

The exchange capacities of some common clays are given in Table 2 and of some different soil textures in Table 3. It is seen from Table 3 that the cation exchange capacity increases with clay content and from Table 2 that this will depend on the type of clay.

Clay minerals are not the only materials which have cation exchange capacity; indeed any fine-grained mineral including quartz displays this property. The special significance of clay in this respect is that, because of the extremely small particle size, the surface area per unit volume of clay is very large and a great many ions are adsorbed. It will be seen in the next section that these adsorbed ions can contribute appreciably to the soil conductivity which thus becomes a function of the clay content.

The organic matter includes the remains of plant and animal life in the soil. The end products of decay accumulate as a blackish, finely-divided colloidal substance known as humus [6] which has large surface area per unit volume, takes up large amounts of water and can develop a negative electrical charge of varying intensity. Little is known of the effects of such colloidal characteristics on the electrical conductivity of humus but they may be significant.

The important influence of soil moisture on the electrical conductivity of soils is discussed at length in a later section.

The direct effect of soil gases on conductivity is negligible, however the indirect effect of CO$_2$ is important and is also described in a later section.

III. 2. Formation of Soils [8] and Soil Profiles [9]

Soils are the result of mechanical, chemical, and biological weathering processes acting on surficial materials in such a way as to grossly alter their physical and chemical properties. In the process of decomposition of parent materials new and stable substances such as clay minerals and humic materials are formed.

As an example of the weathering process consider the chemical

weathering of orthoclase feldspar (an important constituent of granites and other intrusive rocks) to kaolinite, a clay.. Atmospheric carbon dioxide is slightly soluble in water so as to form carbonic acid.

$$CO_2 + H_2O \rightleftharpoons H_2CO_3 \qquad (1)$$

Carbon   Water   Carbonic
dioxide   acid

A small amount of the carbonic acid dissociates so as to produce a hydrogen ion and a bicarbonate ion

$$H_2CO_3 \rightleftharpoons H^+ + HCO_3^- \qquad (2)$$

Hydrogen   Bicarbonate
ion   ion

The bicarbonate ion and water combine with the feldspar to form kaolinite, dissolved silica, free potassium ions and free bicarbonate ions

$$2K\,Al\,Si_3O_8 + 2H^+ + 2HCO_3^- + H_2O \longrightarrow \qquad (3)$$

Orthoclase
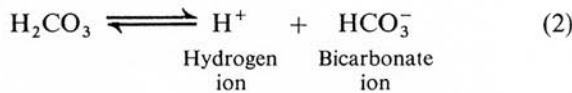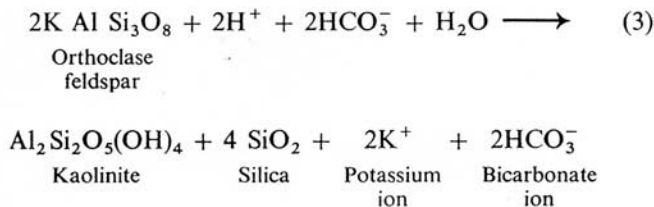feldspar

$$Al_2\,Si_2O_5(OH)_4 + 4\,SiO_2 + 2K^+ + 2HCO_3^-$$

Kaolinite   Silica   Potassium   Bicarbonate
ion   ion

The formula for kaolinite can also be written in the following manner

$$Al_2O_3 \cdot 2SiO_2 \cdot 2H_2O$$

to show that it consists of alumina, silica, and water.

These equations illustrate three important points: (i) The potassium and silica produced by dissolving the feldspar appear as dissolved material. Note that the potassium, converted to a potassium ion, goes into solution and may or may not be present depending on the drainage. (ii) Water is used up in the reaction; it is absorbed into the kaolinite structure. (iii) Hydrogen ions are used up in the reaction and the solution becomes more basic (i.e. less acidic) as the reaction proceeds. All of the reactions described herein affect the electrical properties of the soil by varying either the clay content or the ionic type and concentration in the soil water.

Now let us examine how various factors control this weathering process. The production of soil is a positive feedback process in that once a thin veneer of soil forms, the parent material weathers more rapidly, and more soil is formed. Suppose for example that we have a thin layer of soil on granite. This layer retains moisture which supplies both water that converts carbon dioxide to carbonic acid and water that is hydrated to form kaolinite. Other acids present in the soil contribute additional supplies of hydrogen ion to convert more feldspar to kaolinite as the equations indicate. Plant roots and the process of bacterial decay produce quantities of carbon dioxide to yield more carbonic acid (the amount of carbon dioxide in soil can be as much as ten times greater than that of rainwater, which makes soil water a particularly efficient dissolver of feldspars).

The speed of weathering is a function of climatic type since the rate at which chemical reactions proceed increases with temperature and, more importantly, biological process also proceed faster with higher temperature. Furthermore, since water is needed for the weathering reactions and vegetation grows more lushly in humid climates weathering is most intense in tropical climates which are wet and warm. The surface water resulting from rainfall percolates downwards through the soil, which is more or less permeable, and as it passes reacts chemically with minerals within the soil. The higher the rainfall the more water is available and the more the
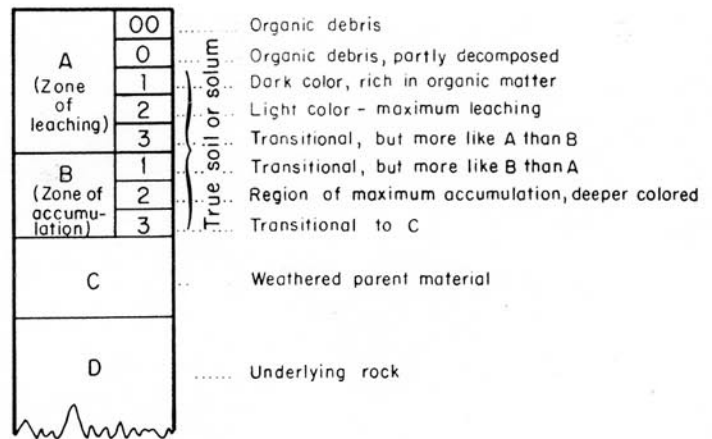

FIGURE 4. Normal or zonal soil profile (after G.B. Maxey [9]).

minerals will react with the undersaturated water. The longer the time the water percolates through, the more time it has to react chemically and again the higher the temperature, the faster the reaction rates.

As a result of these factors tropical soils tend to be thick. Well drained soils tend to be devoid of most unstable minerals. Arid soils contain a variety of minerals. The clay minerals which form are a function of temperature and humidity and there is a tendency for different clays to form in the various climatic belts [10].

As long as there is rainfall and the temperatures are not too low there is a tendency for soils to form a layered structure. In general three different regions or horizons are recognized as we move down through the soil profile, usually designated by the letters A, B, and C. These horizons can vary in thickness from a centimeter to several meters or tens of meters and differ in colour, texture, structure, and other properties.

The chief characteristics of a profile are illustrated in Figure 4. The A horizon, closest to the surface, is the most intensely weathered and has the soluble minerals leached out and most other minerals altered. This layer usually contains much humus, which contributes to its dark colour. Structurally it is friable, granular, or platey. When either friable or granular this horizon is permeable, much more so than the underlying B horizon which is generally a zone of clay accumulation, some of which was formed in-situ and some of which was transported downwards from the A zone by the soil water. The B horizon generally displays a vertical structure in widely or closely spaced joints and it is this horizon that exerts the greatest influence on water movement vertically downwards. When the clay is dry these joints allow rapid downwards movement but when the clay is wet it expands and can close the joints to make the layer impermeable, which may in turn cause the A horizon to become saturated for appreciable periods of time. The C horizon consists of less-weathered parent material and is usually relatively permeable [9].

This layering, with its relatively clay-free A horizon and clay-rich B horizon, greatly influences the vertical profile of electrical conductivity which will be seen to be a function of both clay and water content.

III. 3. Soil Moisture

It was noted at the beginning of this section that the electrical conductivity of soils and rocks was primarily electrolytic and took place through the moisture filled pores and passages which lie within the matrix of insulating minerals. For this reason a knowledge of the way in which soil moisture is distributed in typical terrain is important in understanding terrain conductivity.

When rainwater or irrigation falls on the surface of the soil a fraction runs off directly as surface runoff and the remainder percolates directly into the soil. A fraction of this moisture is retained by the soil, the remainder moves vertically downwards under the force

of gravity until it reaches the water table. In a soil moisture profile it is generally possible to distinguish four stages of moisture occurrence depending on the relative continuity of the moisture films across the soil grains [11] as shown in Figure 5. In the uppermost region we have the pendular state in which the pore space is largely filled with water vapour. The actual liquid water exists only in very small isolated rings around the grain contacts and a continuous path does not exist between the various moisture occurrences. At greater depth we have the funicular stage in which the pendular rings have coalesced to the point where the liquid films have just become continuous throughout the pore space and entirely enclose or encapsulate the vapour phase. Across the sample the moisture path is now continuous. Again further down, in the capillary stage, all pore spaces are occupied by liquid but the liquid pressure within the pores is less than the total pressure caused by gravity since capillary action within the fine pore spaces has caused the moisture to ascend into these pore spaces. Capillary rise, determined effectively by pore size and type, seldom exceeds several meters. Finally we arrive at the phreatic surface (water table) at which the atmospheric pressure is in equilibrium with the hydrostatic pressure. All pores within the phreatic surface are completely filled with liquid under hydrostatic pressure and this is the region of groundwater, also known as the zone of saturation. The three regions above the phreatic surface are collectively referred to as soil moisture, suspended water, vadose water, or zone of aeration.

Figure 5 also illustrates typically the fraction of pore space filled by liquid in the different zones.

A term that is often used in discussing the moisture content of soils is field capacity. Suppose that the water table is very deep below a surface soil of homogenous nature. After a heavy rainfall or irrigation the moisture content reaches a quasi-equilibrium condition in one to three days. As the water percolates down through the soil on its way to the water table capillary forces will retain a certain fraction of the moisture in the small pore spaces. This is the field
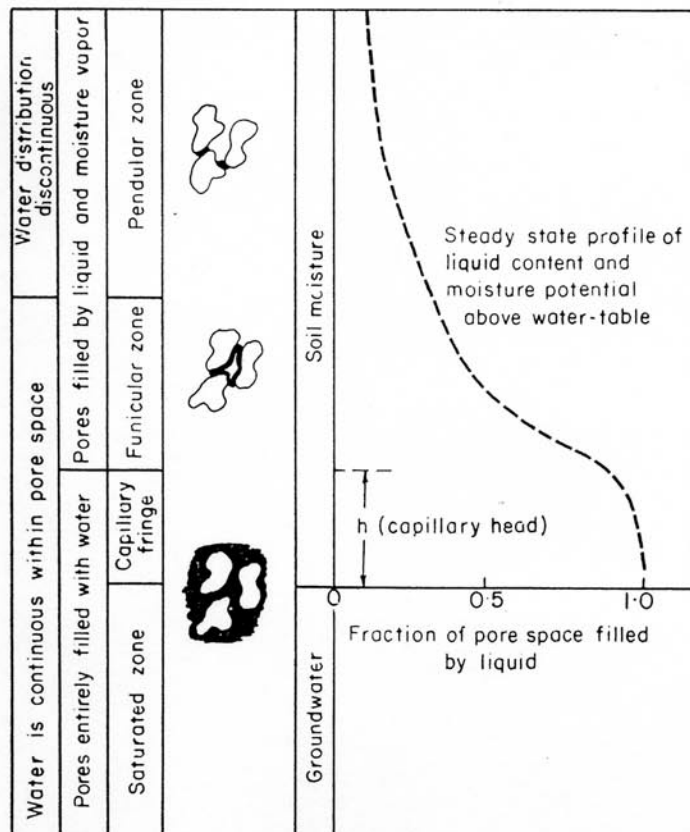
TABLE 4. Permeability of soils (after D. K. Todd [12])

SOIL CLASS	Clean gravel	Clean sands, mixture of clean sands & gravels	Very fine sands, silts, mixtures of sand, silt and clay, glacial till, stratified clays etc	Unweathered clays
FLOW CHARACTERISTICS	Good aquifers		Poor aquifers	Impervious
PERMEABILITY (gal/day/ft²)	10^6 .... 10^4 .... 10^2 .... 10 .... 10^{-2} .... 10^{-4}			

capacity and it is the moisture content of the soil after the gravitational water has been removed by deep seepage.

Note that in the event that we have a layered situation in which a soil with very fine pores overlies more granular material, both being far removed from the water table, moisture content of the upper layer will be many times that of the lower which, as will be seen later, affects the electrical conductivity.

Groundwater is free to move laterally with velocities ranging from a meter or more per day to less than a meter per year, depending on the hydraulic pressure differential and the permeability of the material. Table 4 illustrates the permeability of various materials in gallons per day per square foot of cross-section per foot of hydraulic head per lateral foot of distance. It is seen from the Table that for soil materials a factor of 10^{10} separates permeable from impermeable materials. The saturated hydraulic conductivity decreases with clay content, increasing compaction, and decreasing radius of the soil pores.

The laws of hydraulic movement of water produce a water table which in general is not horizontal and in fact is often a subdued version of the local topography, all other factors being equal. Two examples are shown in Figure 6; for the humid zone the moisture moves from topographic high regions down to the draining streams whereas in an arid zone the moisture moves downwards away from the streams.

The wide ranges of permeability greatly influence the final profile of the water table or phreatic zone sometimes leading, for example, to the occurrence of a perched water table as shown in Figure 7. Again such layering will be seen subsequently to influence the electrical properties.

It will be shown in Section IV that the electrical conductivity of soils and rocks depends on the porosity and on the degree to which the pores are filled with moisture. Figure 2 showed the volumetric composition of a loam soil from which the excess water has been removed i.e. the soil is undersaturated. We see that in this temperate zone example approximately 50% of the soil is occupied by either gas or capillary water. This division into about 50% solids is fairly typical of most soils.

If all of the moisture in a soil sample is removed by drying, the ratio of the empty volume to the total volume of the soil matrix is known as the soil porosity, a parameter that is relevant to the soil conductivity. Table 5 illustrates the porosity for a variety of typical terrain materials.
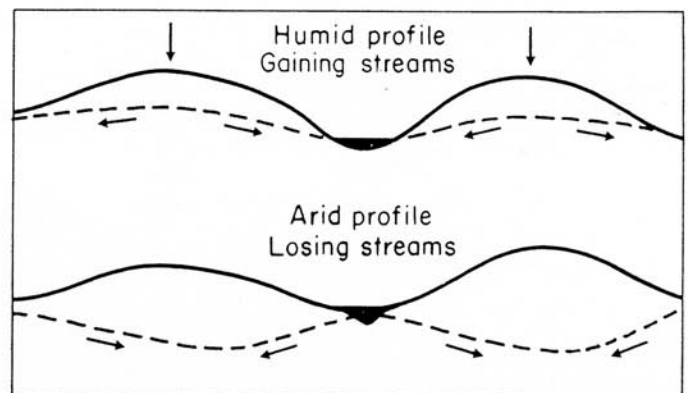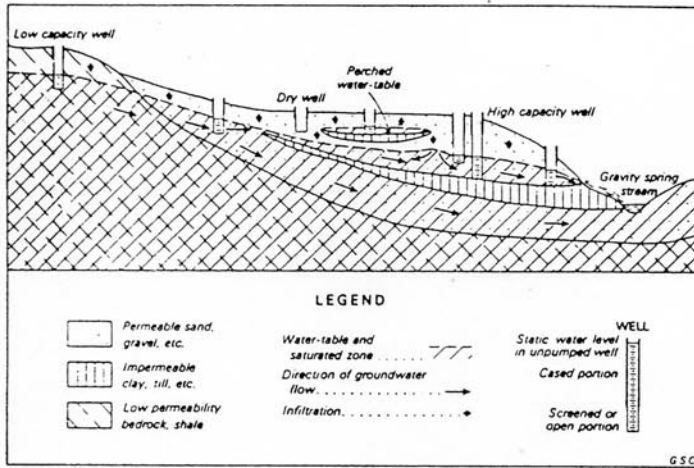

FIGURE 5. Liquid occurrence in soils (after P. Meyboom [11]).

FIGURE 6. Profiles of water tables in arid and humid zones (after G. B. Maxey [9]).

9

LEGEND

Water-table conditions (unconfined aquifers).

FIGURE 7. Water-table conditions – unconfined aquifers (after I.C. Brown [13]).

TABLE 5. Representative porosity ranges for sedimentary materials (after D. K. Todd [12])

Material	Porosity, %
Soils	50–60
Clay	45–55
Silt	40–50
Medium to coarse mixed sand	35–40
Uniform sand	30–40
Fine to medium mixed sand	30–35
Gravel	30–40
Gravel and sand	20–35
Sandstone	10–20
Shale	1–10
Limestone	1–10

Table 6 gives further data on porosity and also indicates the value of the ratio ρ_x (mixture resistivity) divided by ρ_1 (electrolyte resistivity) about which more will be said later in Sections IV and V.

It is obvious that the degree of compaction of a soil will affect the magnitude of the soil porosity.

IV. ELECTRICAL CONDUCTIVITY OF SOILS AND ROCKS

In Section III it was noted that the conductivity of soils and rocks was principally electrolytic. This section will discuss the parameters that determine the conductivity of electrolytes and the effects that arise when electrolytes are present in both clay-free and clay-rich soils and rocks.

IV. 1. Properties of Electrolytes

As an introduction suppose we empty our hypothetical tank of Section II and refill it with distilled water. The measured conductiv-

TABLE 6. Porosity of soils and rocks (after C. A. Helland [14])

Rock or Formation	Porosity	Ratio $\dfrac{\rho_x}{\rho_1}$	
Igneous and metamorphic rocks	½– 2	100	
Dense limestones and sandstones	3– 4	50	–100
Clays and sands in general	8–15	20	– 40
Porous clays, sands, sandstone, cellular limestones, and dolomites	15–40	3	– 20
Marl, loess, clay, and sandy soil	40–75	1.5–	4
Peat, diatomaceous earth	80–90	1.0–	1.5

TABLE 7. Mobility of common ions at 25°C (after Keller and Frischknecht [5])

ion	Mobility (m²/sec V)
H^+	36.2×10^{-8}
OH^-	20.5×10^{-8}
SO_4^-	8.3×10^{-8}
Na^+	5.2×10^{-8}
Cl^-	7.9×10^{-8}
K^+	7.6×10^{-8}
NO_3^-	7.4×10^{-8}
Li^+	4.0×10^{-8}
HCO_3^-	4.6×10^{-8}

ity is very low. If however a small amount of table salt (NaCl) is dissolved in the distilled water the conductivity increases substantially.

The conductivity of an electrolyte is proportional both to the total number of charge carriers (ions) in the solution and their velocity. In distilled water there are few ions and the conductivity is low. The dissolved sodium chloride molecules dissociate to form both positively charged sodium ions and negatively charged chloride ions which greatly increase the conductivity.

When a voltage is applied between the two end plates an electric field is established in the tank; the positively charged ions are attracted towards the negative plate, the negatively charged ions to the positive plate. The velocity of the ions is effectively controlled by the viscosity of the fluid. This velocity is slightly different for different ions since it depends upon their effective diameter, as illustrated in Table 7, where it can be seen that chloride ions move slightly more rapidly than the sodium ions.

In a sodium chloride solution the amount of current that flows, and therefore the electrical conductivity, is proportional to the sum of the number of sodium ions multiplied by their mobility (velocity per unit electric field) and the number of chloride ions multiplied by their mobility.

The further addition of different salts to our tank would increase the electrical conductivity independently of the presence of the sodium and chlorine ions as long as the concentrations remain reasonably dilute. The following equation is often used to calculate the approximate electrical conductivity in mhos per meter of a dilute solution of various salts at normal ambient temperatures:

$$\sigma = 96500[C_1M_1 + C_2M_2 + ...] = 96500 \sum C_1M_1 \quad (4)$$

where C_i = no. of gram equivalent weights of ith ion per 10^6 cm³ of water

M_i = mobility of ith ion in meters per second per volt per meter.

For example, suppose that one gram of salt is dissolved in our tank which holds 10^6 cm³ of water. The atomic weight of sodium is 23, of chlorine is 35, so the atomic weight of sodium chloride is 23 + 35 = 58. Since we have introduced one gram of sodium chloride we have introduced

$\dfrac{23}{58}$ grams of sodium and $\dfrac{35}{58}$ grams of chlorine.

The gram equivalent weight of an ion is the atomic or molecular weight of the ion divided by its valence. In the case of both sodium and chlorine the valence is one so the gram equivalent weight of sodium is 23 grams, of chlorine 35 grams.

Then the number of gram equivalent weights of sodium per 10^6 cm^3 of water is given by

$$C_{Na} = \frac{\text{weight of Na per } 10^6 \text{ cm}^3 \text{ H}_2\text{O}}{\text{gram equivalent weight}} = \frac{23/58}{23} = \frac{1}{58}$$

and that for chlorine is given by

$$C_{Cl} = \frac{\text{weight of Cl per } 10^6 \text{ cm}^3 \text{ H}_2\text{O}}{\text{gram equivalent weight}} = \frac{35/58}{35} = \frac{1}{58}$$

Using the data from Table 7 for the mobilities at 25° C we find that the conductivity for one gram of sodium chloride in 10^6 cm^3 of water is given by

$$\sigma = 96500 \left[\frac{1}{58} \times 5.2 \times 10^8 + \frac{1}{58} \times 7.9 \times 10^8 \right]$$

$$= 0.00022 \text{ mho/m.} \quad (5)$$

The addition of only one part per million of sodium chloride by weight has produced the appreciable conductivity of 0.22 millimhos per meter.

The concentration of dissolved salts in natural groundwaters is substantially higher than one part per million, as a result of which their conductivity is much greater than 0.22 millimhos per meter. For example Table 8 illustrates the contribution of various ions to the measured conductivity of three of the Great Lakes. Lakes Erie and Huron which both occur in regions of Paleozoic carbonate rocks contain more dissolved salts than Lake Superior which is situated largely in Precambrian crystalline rocks, and this is reflected in the higher conductivities of the first two lakes. The bottom line of the table gives the measured conductivity: agreement with the calculated values using the above equations is good.

The following data from Heiland [14] illustrates typical values for the conductivity of various natural waters.
1. Meteoric waters, derived from precipitation; 1 to 30 millimhos per meter.
2. Surface waters (lakes, rivers) vary from 0.3 millimhos per meter for very pure water to as large as 10,000 millimhos per meter for salt lakes. Surface waters in districts of igneous rocks are estimated to range from 2 to 30 millimhos per meter; surface waters in areas of sedimentary rocks vary from 10 to 100 millimhos per meter (compare the Great Lakes above).
3. Soil waters (discharged into the atmosphere by evaporation) may be as large as 10,000 millimhos per meter but their average is around 10 millimhos per meter.
4. Normal groundwater in areas of igneous rock ranges from 6 to 30

millimhos per meter and in areas of sedimentary rocks to as large as 1,000 millimhos per meter.
5. Mine waters (copper, zinc, etc., sulfates) are of high conductivity, generally not less than 3,000 millimhos per meter.

The temperature dependence of the electrical conductivity of the electrolyte is almost entirely due to the temperature dependence of the viscosity of the liquid, which in turn directly affects the ionic mobility. The variation of either quantity with temperature is approximately linear over normal ambient temperatures. The temperature coefficient for a sodium chloride solution is 0.022 which value applies approximately to most other ions so that the electrolyte conductivity for a temperature other than 25°C is given by

$$\sigma(T) = \sigma(25°C)[1 + \beta(T - 25°)]$$

where $\beta = 2.2 \times 10^{-2}$ per °C

T = temperature (°C) at which conductivity is to be calculated.

A change of conductivity of 2.2% per degree centigrade implies that a change in temperature of 40°C will cause the conductivity to nearly double. The effect of temperature is illustrated in Figure 8 which shows calculated conductivity of four of the Great Lakes as a function of season.

The change of conductivity with temperature and therefore with season is not negligible and this applies equally well to ground conductivity over the normal range of ambient temperatures.

IV. 2. Conductivity of Saturated Clean (Clay-Free) Mixtures

We have examined how the electrical conductivity of the electrolyte in our tank varies with the concentration of dissolved salts and with temperature. Let us now start to fill the tank with perfectly insulating spheres of uniform radius (for example uniform pebbles). Assume for the time being that we can maintain these spheres uniformly dispersed throughout the solution. Since the spheres are insulating, the electrical current will find it more difficult to cross the tank and the conductivity will be reduced. A relationship was derived by Maxwell for the conductivity σ_x of a mixture consisting of a medium with conductivity σ_1 in which spherical grains of conduc-

TABLE 8. Conductivity of Great Lakes with contribution of the various ions (after L. H. Doherty [15])

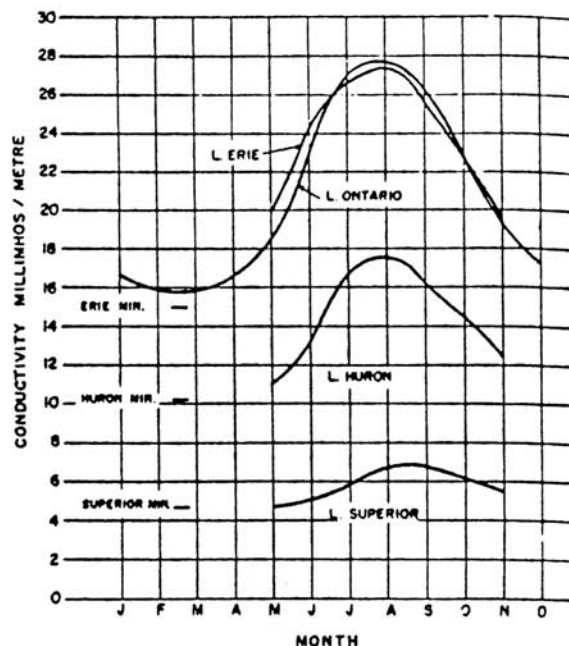| Ion	Lake Erie	Lake Huron	Lake Superior
HCO$_3$	5.8	4.1	2.6
Ca	10.1	7.3	3.8
Mg	3.6	3.2	2.4
Na	1.8	0.6	0.2
Cl	3.9	1.2	0.2
SO$_4$	3.4	1.8	0.3
Calculated Conductivity (mmhos/m)	28.6	18.2	9.5
Measured Conductivity (mmhos/m)	26.7	18.2	8.4

FIGURE 8. Seasonal variation of conductivity of Lakes Ontario, Erie, Huron and Superior (after L.H. Doherty [15]).

tivity σ_2 (here assumed to be zero) are imbedded in regular arrangement and in such a manner that their spacing is large compared with their radius [14]. Letting n be the porosity of the mixture (defined as the ratio of the volume of solution only, divided by the total volume of the solution including the spheres) Maxwell showed that the conductivity of the mixure was given by

$$\frac{\sigma_x}{\sigma_1} = \frac{2n}{3 - n} \qquad (7)$$

In Section II it was shown that the conductance of a current path is directly proportional to the cross-sectional area and inversely proportional to the length. The addition of insulating spheres to the tank tends to reduce the cross-sectional area available for current flow and to increase the effective length of the current paths, reducing the overall conductivity.

This dilute mixture does not resemble a soil so we continue to add insulating spheres until the tank is full. The condition for Maxwell's derivation, that the inter-particle distance be large compared with the particle radius, is invalid and the above equation can no longer be expected to apply. It turns out however that an almost equally simple empirical relationship called Archie's Law [5] is applicable to clean (i.e. clay-free) saturated mixtures. This equation is

$$\frac{\sigma_x}{\sigma_1} = n^m \qquad (8)$$

where n is the fractional porosity defined above and m is a constant. The reciprocal quantity ρ_x/ρ_1 is often called the formation factor (FF) of the rock or soil sample.

Originally derived from resistivity measurements on samples of consolidated porous rock, this equation also applies to a variety of unconsolidated materials. That such a simple law should give excellent results for a variety of both consolidated and unconsolidated materials with widely differing porosities is a surprise and indeed the underlying reasons for its success are not well understood [4]. It has been established [16] that in the case of marine sands the exponent m is dependent on the *shape* of the particles, increasing as they become less spherical (i.e. more platey) and that variations in the *size* of the particles and in the *dispersion* of sizes appear to have a very small effect.

Values of m are listed in Table 9 where it is seen that m varies from 1.2 for insulating spheres to 1.85 for very platey fragments of shell. Samples of natural sands have values in the range 1.4 to 1.6. The value of 1.85 for shell fragments is in good agreement with earlier measurements made on kaolinite particles and marine illite clays. This fact may or may not be significant for clays in normal soils depending on the extent to which ion exchange effects also contribute to the conductivity as discussed further on in this section.

Archie's Law for various exponents is plotted in Figure 9 along with Maxwell's Law. Over the range of porosities of most unconsolidated terrain materials (20% to 70% – see Tables 5 and 6) the different exponents do not greatly affect the mixture conductivity. Interestingly enough, Maxwell's Law gives excellent agreement with Archie's Law with exponent m = 1.3 over all ranges of porosity.

It would appear that for relatively clay-free substances located beneath the water table (so that the mixture is completely saturated) the primary matrix property measured through the electrical conductivity is the porosity of the matrix, essentially independent of the particle size or the particle size distribution. This explains why it is a relatively difficult matter using resistivity techniques to distinguish between sand and gravel. As long as the porosity is the same for both sand and gravel the resistivity or conductivity contrast may be quite small.

Another point is that if the porosity is assumed to vary from 20% to 70% the conductivity of the mixture varies by a factor of approximately 8, depending somewhat on which value of exponent is

TABLE 9. The effect of particle shape on the FF/n relation using artificial samples of decreasing sphericity. (after P. D. Jackson et al. [16])

Sample no.	Mean size ϕ^*	Spread of sizes	Sphericity	Best fit Archie line FF = n^{-m} m ± 0.01
Spheres	0.38	0.17	1.0	1.20
Rounded sand	0.38	0.17	0.83	1.40
Shaley sand	0.50	0.34	0.78	1.52
Shell fragments	0.38	0.17	0.5	1.85

$^*\phi = -\log_2$ (diameter in mm).

adopted. Even in clean mixtures we do not expect a large range of conductivities and experimentally this has been confirmed.

Figure 9 is useful for the following type of calculation: in many glacial deposits gravel appears as pebbles (of the order of a few centimeters in diameter) dispersed throughout a mixture of finer relatively clean sand. How does the presence of this gravel modify the mixture conductivity? To obtain an approximate answer let us assume that the sand itself has a porosity of 30%. Using Archie's Law with m = 1.6 the electrical conductivity of the sand/water mixture will be 15% of the conductivity of the water. Assuming a water conductivity of 20 millimhos per meter gives a conductivity for the mixture of 3 millimhos per meter, a not uncommon value. We use this value in conjunction with Figure 9 to determine the effect of introducing pebbles into the mixture. If for example the pebbles occupy 50% of the volume the conductivity of the mixture will fall to one-third or approximately 1 millimho per meter, also a value that commonly occurs in actual practice. As will be seen later the addition of a relatively small amount of clay can increase these numbers by virtue of increasing the conductivity of the pore water but the relative values may well be similar.

IV. 3. Conductivity of Unsaturated Clean Mixtures

For Archie's Law to apply the material must be fully saturated with fluid. If the mixture is partially saturated the conductivites will be decreased since gas or air bubbles act as insulating particles to further impede the current flow. In the funicular stage of soil moisture the pendular rings have coalesced so that the liquid films are continuous throughout the pore space but only a fraction of the available pore space is filled with water. This moisture varies with time and temperature as a result of drainage, evaporation, and loss of water to plant roots.

In the event that the soil is partially desaturated the following approximate expression applies [5]

$$\frac{\sigma_x}{\sigma_1} = s^n \qquad (9)$$

where s is the fraction of total pore volume filled with electrolyte and n is a parameter experimentally determined to be approximately 2. This expression is equivalent to Archie's Law with m = 2 and indicates as seen from Figure 9 that if a small fraction of the total pore volume is filled with water the conductivity can be very low.

More recent work on soils has shown that the electrical conductivity varies as follows with moisture content when the soil is partially desaturated [17]

$$\sigma_x = \sigma_1 \theta(a\theta + b) + \sigma_s \qquad (10)$$

where θ is the volumetric water content (cm^3 of water per cm^3 of soil), a and b are constants which depend on the soil texture, and σ_s is a contribution from ''surface conductivity'' which will be discussed under Colloidal Conductivity. The values of the empirically determined constants a and b for a variety of soils are shown in

FIGURE 9. Graph of Archie's Law and Maxwell's Law.

In the figure:

n (fractional porosity)

Insulating particles in conducting fluid

$$r = \frac{volume\ of\ solid}{total\ volume}$$

$$n = \frac{volume\ of\ fluid}{total\ volume}$$

$$n + r = 1$$

ARCHIE'S LAW

$$\frac{\sigma_x}{\sigma_1} = n^m$$

m = 2·0
m = 1·6
m = 1·3

MAXWELL'S LAW

$$\frac{\sigma_x}{\sigma_1} = \frac{2(1-r)}{2+r}$$

r (fractional volume of solid)

TABLE 10. Measured moisture content constants and surface conductivities (after J. D. Rhoades et al. [17])

Soil type	a	b	σ_s
			mmho/cm
Pachappa fsl	1.382	−0.093	0.18
Indio vfsl	1.287	−0.116	0.25
Waukena l	1.403	−0.064	0.40
Domino cl	2.134	−0.245	0.40

increasing depth the conductivity rises rapidly as the moisture films become continuous, then more slowly as the available pore volume starts to fill, eventually saturating with complete filling of the pores. This situation will be rendered even more complex if the porosity itself is, as is often the case, a function of depth as suggested by the profile of Figure 4.

IV. 5. Colloidal Conductivity

It will be recalled that for clays the cation exchange capacity (CEC) was a measure of the number of ions adsorbed to the surface of clay particles. When the clay particles are immersed in a liquid there is evidence to show that these adsorbed ions can partially dissociate themselves from the clay particles and become available for ionic conductivity. Since the ion exchange capacity of clays can be great due to their large surface area many ions may be supplied for electrical conductivity; the addition of a small amount of clay to an otherwise clean mixture can substantially increase the electrical conductivity.

The addition of clay appears to affect the electrical conductivity of the mixture in two ways. Repeating equation (10)

$$\sigma_x = \sigma_1 \theta(a\theta + b) + \sigma_s \qquad (10)$$

the added ions increase the value of σ_1 above the value that the porewater would have in the absence of the clay. Furthermore although dry clay is highly resistive, as soon as a thin layer of moisture (perhaps only a few molecules thick) surrounds the clay particles ion movement across the surface of the clay particle within the cloud of adsorbed ions may occur. This surface contribution to the conductivity, essentially independent of the moisture content, is the second term in the equation. It will be most significant at low moisture contents.

The contribution to σ_1 from clay content will be most evident for soils in which the porewater is relatively pure (and therefore has low conductivity) and will be least effective in soils having highly saline porewater. For either contribution we should expect that clays with higher CEC will produce more conductive soils and a comparison of Table 11 with Table 2 shows that this is the case

In summary, in areas where the soil porewater is not particularly saline the electrical properties of the soil may be strongly influenced by and indeed in some cases completely dominated by the amount and type of clay minerals present. The possible influence of clay materials should always be kept in mind. It is also possible that a similar effect arises as a result of the colloidal properties of humus which might be important in tropical climates where the humus layer is well leached and contains few clay minerals.

IV. 6. Effects of Freezing on Soil Conductivity

Suppose that we take our tank, filled with a clean sand/water mixture, and start to lower the temperature. The electrical conductivity of the mixture will decrease in exactly the same way that the conductivity of the electrolyte does. When the temperature reaches 0°C the water freezes and since the conductivity of ice is extremely low the conductivity of the mixture falls essentially to zero.

Now suppose that in addition to the sand some clay is introduced to the mixture and the temperature is again reduced. As the temperature passes through 0°C some of the water freezes; however the

Table 10 from which it is seen that except for very low moisture the relationship is essentially a square law with mosture content.

It has been suggested by Rhoades et al [17] that it might be possible to estimate the values of σ_s, a, and b on the basis of soil texture and mineralogy and, given a measurement of the bulk soil conductivity and the moisture content, to determine the conductivity of the fluid and thus salinity.

IV. 4. Electrical Layering Arising from Soil Moisture

From the preceding paragraphs we observe that, even if the soil material is physically homogeneous, for example consists of clean well-sorted silt, as soon as we introduce soil moisture of finite electrical conductivity we have established a medium in which the electrical conducvitity can vary strongly with depth. Near surface where the moisture content is low the conductivity is also low. With

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TABLE 11. Physical properties of typical Upland and Lowland soils in western Puerto Rico (after J. W. Walker et al. [18])

Topographic location	Sample no.	Moisture (%)	Conductivity (mmho/m)	Clastic material (%)	Clay fraction (%)	Clay type(s) (%)
Uplands	1	42	16.0	21	79	100% kaolinite
Uplands	2	46	1.2	42	58	100% kaolinite
Uplands	3	25	1.0	5	95	100% kaolinite
Lowlands	4	21	65.0	66	34	100% montmorillonite
Lowlands	5	32	169.0	39	61	60% montmorillonite 40% kaolinite
Lowlands	6	25	269.0	15	85	90% montmorillonite 10% kaolinite

electrical field of the adsorbed ions on the clay particles locally orients the neaby water molecules to prevent their freezing [17]. Furthermore as the solvent freezes there is a tendency for the impurity ions to stay within the liquid fraction and the actual electrical conductivity of the remaining liquid water increases with decreasing temperature. The net result is that mixtures containing clay or silts tend to have an electrical conductivity which decreases relatively slowly with temperature as the temperature passes through 0°C and indeed to retain a moderate conductivity even at temperatures well below freezing as illustrated in Figure 10.
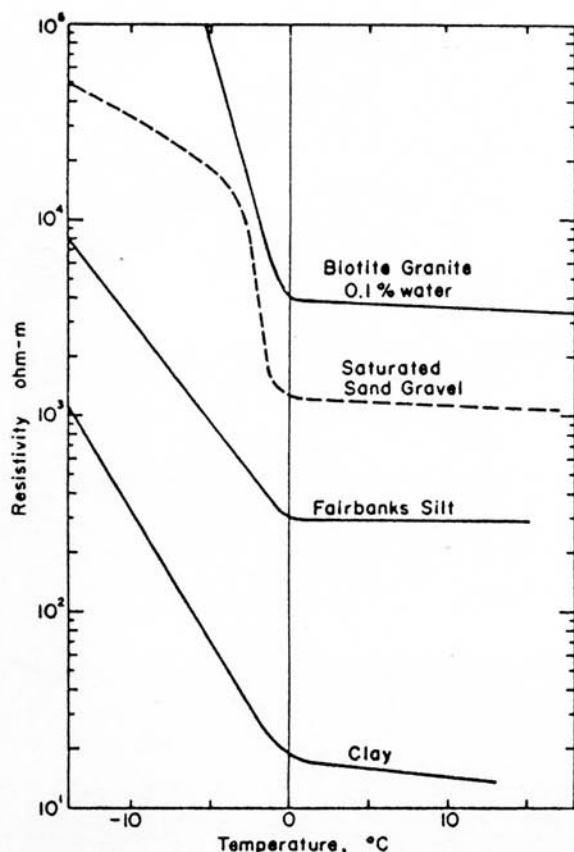

FIGURE 10. Resistivities for several soils and one rock type as a function of temperature (after Hoekstra and McNeill [20]).

IV. 7. Electrical Properties of Rocks

As with unconsolidated materials the matrix of most water-bearing rocks is insulating and the electrical conductivity is also electrolytic. A major difference between unconsolidated sediments and rocks lies in the types of pore geometries. For example in sedimentary rocks the porosity is generally inter-granular and consists of voids still remaining from the compaction process. In igneous rocks the porosity of the rock itself may be extremely small, however moisture circulates through fractures in the rock which are the result of mechanical breakage. These cracks are called joints and may be sufficiently large and/or numerous that their presence completely dominates the electrical conductivity. Such jointing may also play a major role in the conductivity of the more impermeable sedimentary rocks. When large joints are present the local conductivity may be expected to vary strongly with position. A third form of porosity which may not make a major contribution to electrical conductivity as normally measured is vugular porosity. It consists of cavities (as might be caused by solutions circulating in limestone) interconnected by small pores which dominate the D.C. electrical behavior. It should be noted that in the case of electromagnetic excitation of currents direct electrical (ohmic) inter-cavity connection is not necessary and appreciable response to this excitation may come from the fluid in each cavity separately.

Table 12 illustrates normal porosity for various rock types. It is seen that the total porosity can be very small, of the order of a few percent, and we would therefore expect, particularly for unfractured Precambrian igneous rocks and high-ranking metamorphosed rocks, the electrical conductivity to be very low, as is generally the case. On the other hand shales may be relatively porous and in addition the conductivity of the pore water may be high as a result of ion exchange effects. For this reason shales can be as conductive as 30 to 40 millimhos per meter and in some cases even higher.

Since in general the porosities are smaller for consolidated materials and furthermore at this end of the scale the various exponents in Archie's Law have a large effect, it is difficult to predict the electrical properties of any given rock type.

There is another feature which occurs in sedimentary rocks and which is of importance in determining their electrical characteristics. Being depositional in nature these rocks are layered and the electrical conductivity perpendicular to the bedding planes may be less than the conductivity parallel to the planes.

This feature is illustrated in Table 13 which lists the coefficient of anisotropy for various layered rocks. The coefficient is defined in Figure 11.

Another type of rock conductor called a structural conductor [21] occurs as a result of fracturing and is often linear in shape. Such conductors arise in the interior of faults, shear zones, contact frac-

TABLE 12. Normal ranges in porosity for rocks (after Keller and Frischknecht [5])

Rock type	Intergranular porosity (%)	Joint porosity (%)	Vugular porosity* (%)
Paleozoic sandstones and shale	5–30	0–1	0
Paleozoic limestones	2–10	0–2	0
Paleozoic clastic volcanics	5–30	0–2	0
Post Paleozoic sandstones and shale	10–40	0	0
Post-Paleozoic limestones	4–20	0–2	0
Post-Paleozoic clastic volcanics	10–60	0	0
Precambrian sediments and low-rank metamorphosed sediments	1–8	0–2	0
Precambrian igneous rocks and high-rank metamorphic rocks	0–2	0–2	0
More recent igneous rocks	0–10	0–2	0

*Vugular porosity accounts for an appreciable total porosity only in rare cases.

TABLE 13. Coefficients of anisotropy for layered rock (after Keller and Frischknecht [5])

Rock type	Coefficient of anisotropy
Volcanic tuff, Eocene and younger, from Nevada	1.10–1.20
Alluvium, thick sections from the southwestern United States	1.02–1.10
Interbedded limestones and limey shales from northeastern Colorado	2.0 –3.0
Interbedded anhydrite and shale, northeastern Colorado	4.0 –7.5
Massive shale beds	1.01–1.05
Interbedded shale and sandstone	1.05–1.15
Baked shale or low-rank slate	1.10–1.60
Slates	1.40–2.25
Bitumenous coal and mudstone	1.7 –2.6
Anthracite coal and associated rocks	2.0 –2.6
Graphitic slate	2.0 –2.8

ture zones, etc., where the rock material has been ground into small particles which allows increased circulation of groundwater resulting in enhanced weathering to produce clay minerals. Structural conductors occur in a wide range of sizes.

V. EXAMPLES OF CONVENTIONAL MEASUREMENTS OF SOIL AND ROCK RESISTIVITIES

The material in this section gives a broad indication as to the range of resistivity or conductivity which might be encountered in various terrain materials in various climatic zones. Extreme caution must be exercised in employing these values for anything other than a rough guide as to anticipated survey results.

As an introduction Table 14 illustrates the resistivities of igneous and metamorphic rocks as given by Telford et al [22]. Table 15 lists resistivities of sediments from the same source. Table 6, from Heiland [14], lists the ratio ρ_x/ρ_0 as defined in Section IV for various consolidated and unconsolidated materials.

V. 1. Temperate Zones

Table 16 records the ranges of resistivity compiled for different terrain materials from a variety of survey and laboratory measurements, from Culley et al [23].

Table 17 from Sellman et al [24] shows survey data over different soil types made with three different measurement techniques (i) radio-frequency magneto-telluric at approximately 300 kHz, (ii)
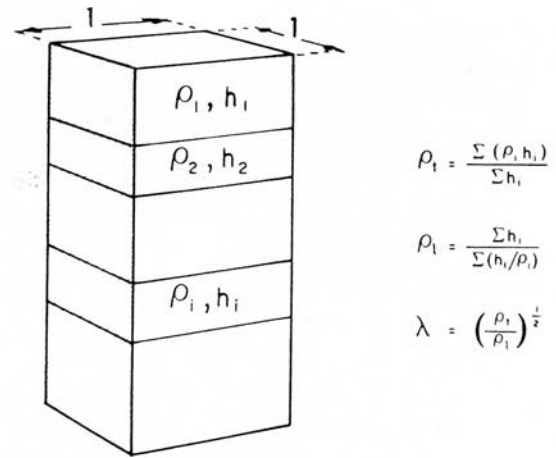

FIGURE 11. Definition of coefficient of anisotropy (λ) for layered sediments (after Keller and Frischknecht [5]).

$$\rho_t = \frac{\Sigma \, (\rho_i h_i)}{\Sigma h_i}$$

$$\rho_l = \frac{\Sigma h_i}{\Sigma (h_i/\rho_i)}$$

$$\lambda = \left(\frac{\rho_t}{\rho_l}\right)^{\frac{1}{2}}$$

low-frequency magnetically induced currents using dipole transmitter and receiver at 10 kHz, and (iii) standard Wenner array at DC. The measured resistivities are relatively independent of the method of measurement; the general trend of increasing resistivity with particle size should be noted.

A complete set of measurements by Smith-Rose [25] on a variety of soils from different depths, with different moisture contents, and measured at various frequencies, shows that at the lower frequencies (i.e. between 1 and 100 kHz) the conductivity of soils is essentially independent of frequency. At higher frequencies the conductivity rises and the increase in conductivity is generally greatest for the most poorly conducting samples. The results of his laboratory measurements on soil samples taken from various parts of England are illustrated in Table 18 and profiles of conductivity with depth
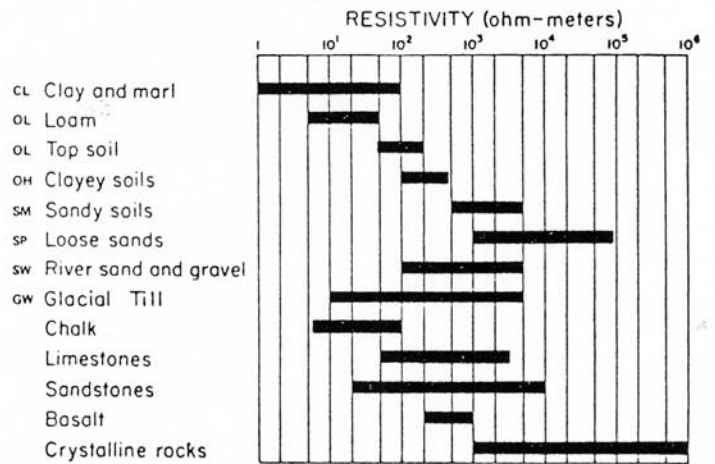
TABLE 14. Resistivities of igneous and metamorphic rocks (after W. M. Telford et al. [22])

Rock type	Resistivity range (Ωm)
Granite	3×10^2–10^6
Granite porphyry	4.5×10^3 (wet)–1.3×10^6 (dry)
Feldspar porphyry	4×10^3 (wet)
Albite	3×10^2 (wet)–3.3×10^3 (dry)
Syenite	10^2–10^6
Diorite	10^4–10^5
Diorite porphyry	1.9×10^3 (wet)–2.8×10^4 (dry)
Porphyrite	10–5×10^4 (wet)–3.3×10^3 (dry)
Carbonatized porphyry	2.5×10^3 (wet)–6×10^4 (dry)
Quartz porphyry	3×10^2–9×10^5
Quartz diorite	2×10^4–2×10^6 (wet)–1.8×10^6 (dry)
Porphyry (various)	60–10^4
Dacite	2×10^4 (wet)
Andesite	4.5×10^4 (wet)–1.7×10^2 (dry)
Diabase porphyry	10^3 (wet)–1.7×10^5 (dry)
Diabase (various)	20–5×10^7
Lavas	10^2–5×10^4
Gabbro	10^3–10^6
Basalt	10–1.3×10^7 (dry)
Olivine norite	10^3–6×10^4 (wet)
Peridotite	3×10^3 (wet)–6.5×10^3 (dry)
Hornfels	8×10^3 (wet)–6×10^7 (dry)
Schists (calcareous and mica)	20–10^4
Tuffs	2×10^3 (wet)–10^5 (dry)
Graphite schist	10–10^2
Slates (various)	6×10^2–4×10^7
Gneiss (various)	6.8×10^4 (wet)–3×10^6 (dry)
Marble	10^2–2.5×10^8 (dry)
Skarn	2.5×10^2 (wet)–2.5×10^8 (dry)
Quartzites (various)	10–2×10^8

TABLE 15. Resistivities of sediments (after W. M. Telford et al. [22])

Rock type	Resistivity range (Ωm)
Consolidated shales	$20-2 \times 10^3$
Argillites	$10-8 \times 10^2$
Conglomerates	$2 \times 10^3-10^4$
Sandstones	$1-6.4 \times 10^8$
Limestones	$50-10^7$
Dolomite	$3.5 \times 10^2-5 \times 10^3$
Unconsolidated wet clay	20
Marls	3–70
Clays	1–100
Alluvium and sands	10–800
Oil sands	4–800

TABLE 16. Resistivity ranges for various terrain materials (after Culley et al. [23])
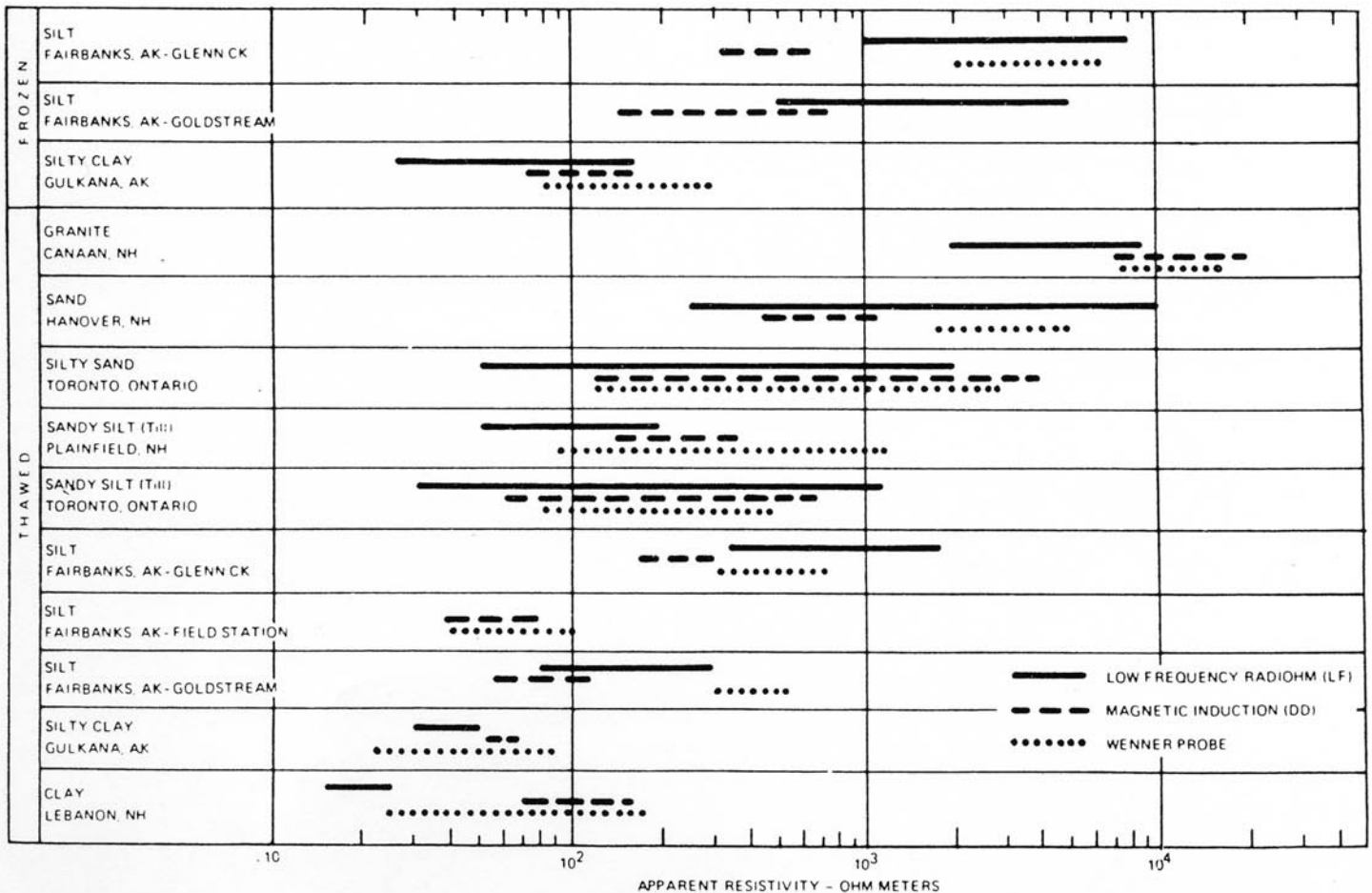

from these measurements are shown in Figure 12. It is seen from Table 18 that the soil type varies rapidly with depth, as does the soil moisture content; both of these influence the conductivity profiles. Smith-Rose concludes that clays have the highest conductivities, greater than 10 millimhos per meter, loams and chalks are of the order or 10 millimhos per meter and sandy or gritty soils are appreciably less. He also points out that a diurnal temperature range of 20°C at the surface of the soil represents a temperature change of approximately 1.4°C at a depth of one foot. Measurements made on a soil sample at 1.2 MHz as a function of moisture content by weight show a conductivity that increases approximately as the square of the moisture content.

V. 2. Tropical Humid Zones [26]

The examples given above have been taken from temperate zone soils and seem to be fairly representative. In the case of soils occurring in tropical humid climates (annual temperatures of the order of 25°C; annual precipitation of greater than a meter, generally falling during a part of the year) the weathering can be very deep. Indeed, unweathered parent rock may not occur until 30 meters in flat country, 60 meters in hilly country and in rare occasions depths

TABLE 17. Apparent resistivity data obtained for various material types using three measuring techniques (after Sellman et al. [24])
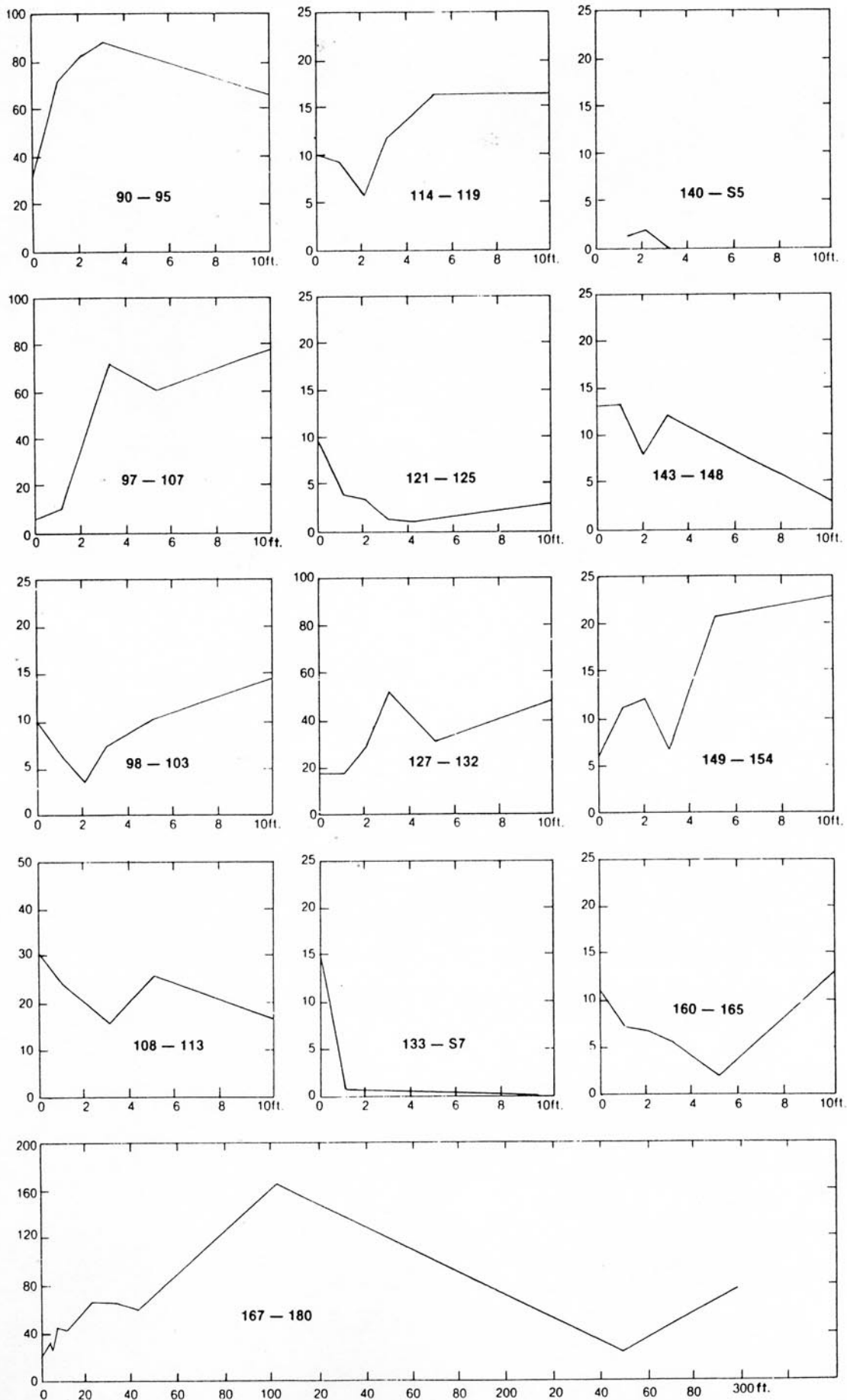
FIGURE 12. Plots of conductivity (mmhos/m) versus depth (feet) for data in Table 18.

17

Sample No.	Site	Geological classification	Depth	Description of sample	Moisture content	Conductivity at 20°C Millimho/m	
						1 kHz	100 kHz
			ft.		per cent		
90	Rugby Radio	Lower lias	Surface	Dark fibrous loam	60	33	38
92	Station (1)		1	Loam and clay	33	72	78
93			2	Clay and sand	26	83	89
94			3	Blue clay	25	89	100
95			10	Blue clay	23	67	72
97	Rugby Radio	Lower lias	Surface	Loam	22	6.1	7.2
104	Station (2)		1	Loam and clay	13	9.4	9.4
105			3	Blue clay	27	72	83
106			5	Clay and sand	21	61	61
107			10	Blue clay	25	78	89
98	Baldock, Herts	Chalk	Surface	Fibrous loam	21	9.4	10.0
99	(P.O. Receiving Station)		1	Chalky loam	21	6.1	6.1
100			2	Chalk	24	8.1	2.9
101			3	Chalk	27	7.2	7.8
102			5	Chalk	26	10.1	15.6
103			10	Chalk	27	14.4	15.6
108	Tatsfield, Kent	Upper greensand	Surface	Fibrous loam	37	30	38
109	(B.B.C. Receiving Station)		1	Brown, sandy clay	1	24	27
110			2	Brown sand	15	20	22
111			3	Light brown sand	13	15.6	10.7
112			5	Light brown sand	20	26	26
113			10	Yellow sand	15	16.7	19
114	Brookmans Park, Herts	London clay	Surface	Fibrous loam	19	10.0	10.6
115	(London Regional Station)		1	Stony loam	18	9.4	10.0
116			2	Light sandy clay	22	5.6	7.2
117			3	Sandy clay	22	12.2	13.3
118			5	Sandy clay	21	16.7	18.8
119			8 to 10	Clay and shingle	10	16.7	17.8
121	Daventry	Upper lias	Surface	Fibrous loam	28	9.4	10.6
122	Northants		1	Sandy loam	16	3.8	3.8
123			2	Brown sand	14	3.2	3.2
124			3	Brown sand	5.0	1.1	1.2
126			5	Sand and sandstone	8.5	0.8	1.0
125			10	Sand and sandstone	24	2.9	3.7
127	Washford Cross, Somerset	Red Marls	Surface	Reddish-brown loam	23	16.7	18.9
128	(West Regional Station)		1	Reddish-brown clay	20	16.7	18.9
130			2	Reddish-brown clay	18	28.9	31
129			3	Reddish-brown clay	21	52	54
131			5	Reddish-brown clay	19	31	34
132			10	Reddish-brown clay	15	48	50
133	Brendon Hills, Somerset	Devonian	Surface	Black fibrous loam	210	14.4	16.7
134			1	Loam and slate	9.0	0.3	0.3
135			2	Loam and slate	9.0	0.2	0.2
136			3	Loam and slate	8.0	0.1	0.1
137			5	Loam and slate	5.5	0.0	0.0
S.7			10	Slate	—	0.0	0.0
140	Merrivale, Dartmoor, Devon	Granite	1	Gritty loam	18	1.3	1.3
141			2	Gritty loam	13	1.6	1.6
S.1			3 to 10	Granite	—	0.0	0.1
S.2			3 to 10	Granite	—	0.0	0.1
S.5			3 to 10	Granite	—	0.0	0.0
139	Dousland, Dartmoor, Devon	Devonian	Surface	Loam	47	6.1	6.7
142			1	Dark brown loam	41	2.7	2.7
S.4			Below 1	Slate	—	0.0	0.0
S.6			Below 1	Granite	—	0.0	0.0
143	Moorside, Edge, Yorks	Millstone grit	Surface	Fibrous loam	130	13.3	17.8
144	(North Regional Station)		1	Dark grey clay	60	13.3	15.6
145			2	Dark grey clay	35	7.8	10.6
146			3	Dark grey clay	39	12.2	16.7
147			5	Dark grey clay	19	9.4	11.1
148			10	Yellow and grey clay	15	3.4	3.7
149	Westerglen, Falkirk	Boulder clay	Surface	Fibrous loam	38	6.1	7.2
150	(Scottish Regional Station)		1	Fibrous loam	30	11.1	11.1
151			2	Clay and loam	19	12.2	12.2
152			3	Dark grit and clay	18	6.7	7.8
153			5	Dark grit and clay	18	21	24
154			10	Dark grit and clay	15	23	26

TABLE 18. (concluded)

Sample No.	Site	Geological classification	Depth	Description of sample	Moisture content	Conductivity at 20°C Millimho/m	
						1 kHz	100 kHz
			ft.		per cent		
160	Teddington, Middlesex	London	Surface	Fibrous loam	26	11.1	12.2
161	(N.P.L.)	clay	1	Sandy loam	20	7.2	7.8
162			2	Sandy loam	13	6.7	7.2
163			3	Fine gravel	6.5	5.6	6.7
164			5	Coarse gravel	2.9	1.9	2.0
166			7	Fine sand	2.6	1.6	1.6
165			10	Sand and shingle	20	13.3	15.6
167	Wychbold, Droitwich	Red Marls	1	Red clay and loam	15	16.7	20
168	(Midland Regional Station)		2	Red clay	13	34	32
169			3	Red clay	14	23	24
170			5	Red clay and stones	15	43	51
171			10	Red clay and stones	21	41	44
172			20	Red clay suspension	31	67	78
173			30	Red clay suspension	41	67	78
174			40	Red clay suspension	25	61	72
175			50	Red clay	27	78	83
176			100	Grey clay and salt	28	177	233
177			150	Red clay and salt	27	—	—
178			200	Red clay and salt	24	—	—
179			250	Red clay and salt	22	24	26
180			300	Red clay and salt	31	73	89

of 250 meters have been observed. The resulting soils are red to yellow, soft with a high clay content, and with a specific gravity approximately one-half of that of the parent rock. Compared with the soils of temperate climates they are thick, humus poor, permeable, and have a high clay-silt ratio. Although the clay content is high this is somewhat compensated for by the fact that in well drained soils the clays are kaolinitic and the cation exchange capacity is less than in temperate zones (see Table 11). Iron-rich concretionary horizons called laterites form as a result of reprecipitation. These are hard, permeable, and if dry are usually very resistive.

The nature and the extent of the soil formation is a function of the rock type, texture, jointing, surface relief, vegetation, water-table, and micro-climate. For example, the relative absence of silt is due to the fact that fine-grained rocks weather faster than coarse-grained rocks; the abundance of clay results from the fact that it is a stable mineral.

An excellent series of measurements of the electrical properties of the weathered zones in tropical climates has been carried out by Palacky and Kadekaru in Brazil [27]. The measurements were done with conventional resistivity techniques and many soundings were taken. The results gave good agreement with a three-layered earth model and are summarized in Table 19. In general a resistive soil, relatively thin, is situated on top of a thick quite conductive weathered zone, in turn situated on the relatively resistive parent rock.

V. 3. Tropical Arid Zones

It is in tropical arid zones that the most conductive soils are encountered. As for humid zones the weathered layer can be many tens of meters in thickness, however in arid climates this material often contains a high salt concentration due to evaporation. Whereas the surface materials are usually dry, possibly lateritic, and resistive, deeper material may approach resistivities of the order of 1 ohm-meter. Furthermore substantial lateral variations of resistivity are not uncommon.

In the humid tropical zones referred to above the conductivity of the intermediate zone was due to the presence of an abundance of colloidal particles with moderate ion exchange capacities: abundant rainfall means that drainage patterns are well established and soil salts have long since been leached into major river channels and

TABLE 19. Tropical resistivity profiles (after Palacky and Kadekaru [27])

	Site	Soil		Weathered Zone		Bedrock	
		Thickness (meters)	Resistivity (ohm-meters)	Thickness (meters)	Resistivity (ohm-meters)	Type	Resistivity (ohm-meters)
Humid Tropical	Nova Lima	5–10	2000–15000	20–80	50–100	schists	∞
					200–400	granodiorite	∞
	Canabrava	5	90	3–18	15–27	basic	∞
		5	360	12–30	3–12	ultrabasic	∞
	Santa Fé	3	11000–18000	35	80–100	dunite	∞
	Quatipuro	0–1	200	15–35	10–30	phyllites	∞
			300–700	3–8	35–20	serpentinites	> 300
Arid Tropical	Santa Luz	–		3.5	65	granite	910
		1	120	10–20	13	basic volcanic	∞
	Curaça	–		6–10	20–35	amphibolites	1200
				15–20	80–130	gneiss	2000

TABLE 20. Diagnostic criteria for distinguishing between unaffected soil sites and encroaching and developed saline seeps (after Rhoades and Halvorson [28])

Site type	Salt content	Water content	Soil electrical conductivity
Unaffected	Low, increasing with depth	Low, increasing with depth	Low, increasing with depth
Encroaching saline seep	Low, increasing to a peak at a relatively shallow depth, then decreasing with further depth	Moist surface, becoming wet with depth	Intermediate, increasing to a peak at a relatively shallow depth, then decreasing with further depth
Developed saline seep	High, decreasing with depth	Relatively uniformly wet to the water table	High, decreasing with depth

thence into the oceans. In hot arid climates drainage patterns are poorly established and drainage basins may have no outlet to permanent streams. Salt-bearing waters drain from topographically high regions into lower areas where the water evaporates, leaving a high residual salt content, largely a reflection of the imperfect drainage channels.

The occasional addition of water might be due either to precipitation or irrigation. In the cultivated dry land soils of the northern Great Plains the near surface (plant root-zone) is leached with essentially pure water derived from rain and snow melt during the wet months. Rhoades and Halvorson [28] discuss how this results in a soil salinity concentration which increases with depth. However during the summer months when a saline water table in this climate is situated within approximately one meter from the surface an upwards flow of moisture is caused by evaporation at the soil surface and transpiration within the root zone. This upwards flow reverses the original leaching which transported salts downwards and causes them to ascend so that a concentration peak can form in the soil salinity profile. The soils can remain excessively wet as a

result of the hygroscopic nature of the salts and because the salts reduce the effective use of water by plants via evapo-transpiration. Rhoades defines such a condition as a "saline seep" and notes that it can form without the water table actually emerging at the surface. As this seep develops the peak in the soil salinity profile moves upwards until it finally appears at or near the surface as indicated in Table 20 which illustrates the various conditions that can exist in a region which sees seasonal precipitation followed by seasonal drought. Figure 13 illustrates vertical profiles of resistivity versus depth as determined by an expanding Wenner array for the three conditions described in Table 20.

For saline soils, the contribution to conductivity from salt concentration generally outweighs that from cation exchange capacity and the conductivity is relatively independent of clay content.

The effect of weathered-zone salinity is probably at its worst in western Australia where the zone is often 100 meters in thickness and has been known to exceed 300 meters. Hygroscopic salts maintain soil moisture at levels substantially above the water table and resistivities of the order of 1 ohm-meter have been observed. Large lateral variations result from the imperfect drainage patterns; slight changes in surface topography often show up as large changes in subsurface conductivity. A lateritic hard pan is not unusual on the surface. This layer is extremely resistive and completes the complexity and difficulty of making conventional electrical measurements in this environment.

V. 4. Arctic Zones

Finally we turn our attention to arctic regions in which permafrost is a major consideration. It should be noted that the definition of permafrost requires only that the mean annual temperature of the ground be less than 0°C for several years; the definition is completely independent of the nature of the material, the amount of moisture, and indeed whether or not the moisture is frozen. In
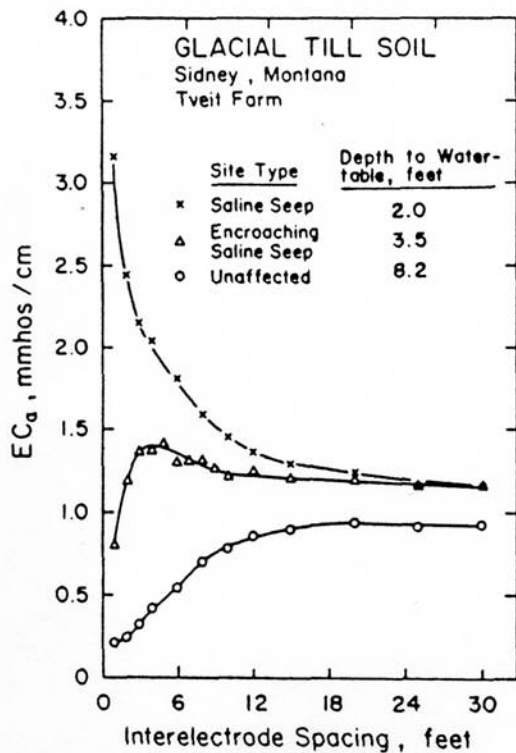

FIGURE 13. Relation between soil electrical conductivity, EC$_a$, and interelectrode spacing for a saline seep, an encroaching saline-seep site, and an unaffected site for glacial till-clay loam soil near Sidney, Montana (after Rhoades and Halvorson [28]).
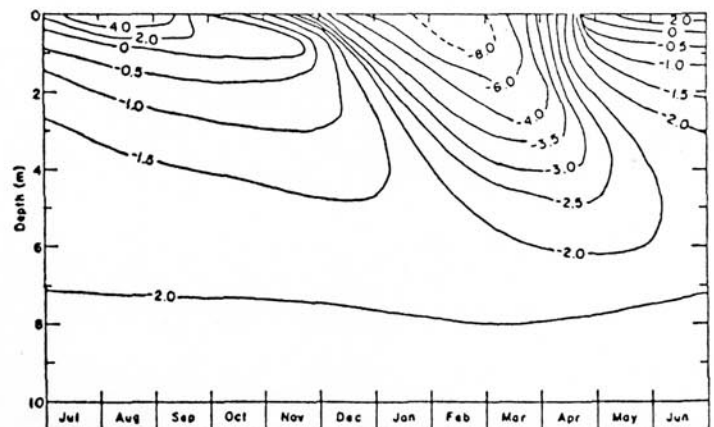

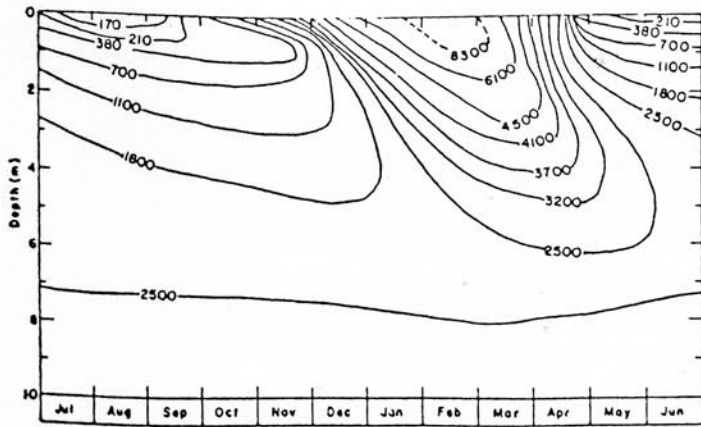FIGURE 14. Vertical and temporal temperature variations – contoured in °C (after Arcone et al [29]).

FIGURE 15. Computed vertical and temporal resistivity variations from data of Figure 14 – contoured in ohm-meters (after Arcone et al [29]).

northern climates we find that mineralogy, porosity, free water content, and ionic concentration within the free water are important resistivity factors as they are for temperate zones. In permafrost regions we have the additional complexity that small changes in temperature of the ground near 0°C may exert a large influence on terrain conductivity. The effect depends on the nature of the material and also the moisture content since as we have seen earlier even at temperatures substantially below 0°C considerable moisture may remain unfrozen in the case of clays whereas in the case of coarser material virtually all will be frozen. In short, ice content is a complicated function of many variables.

As an example of electrical layering arising due to temperature only consider Figure 14 which illustrates annual variations in ground temperature recorded near Fairbanks, Alaska. The region is in a discontinuous permafrost zone and subsurface material consists primarily of perennially frozen organic silt containing a varying amount of ground ice. From laboratory measurements on saturated organic silt to determine the variation of electrical resistivity with temperature, a plot of electrical resistivity with depth and time was derived [23] and is shown in Figure 15. We see from this figure that a resistivity range of 50 to 1 can occur as a result of temperature changes alone.

In the discontinuous zone of permafrost subsurface temperatures are subtly influenced by many variables and even in ground which is

laterally uniform insofar as material type is concerned lateral temperature variations can bring about large changes in electrical resistivity as shown in Figure 16 where the survey traverse was taken over glacial lake basin sediments. The frozen areas outlined on the figure were derived from resistivity measurements and confirmed by drilling [24].

VI. SUMMARY

In this technical note we have discussed the concept of terrain conductivity, examined some of the properties of soils and rocks that affect their conductivity, and reviewed some measurements on typical terrain materials under a variety of climatic conditions.

To summarize, terrain conductivity is usually determined by one or more of the following parameters:
(1) clay content, clay type
(2) moisture profile with depth
(3) moisture salinity
(4) moisture temperature
Of these the most complex is usually the moisture profile, by which is meant the way in which (i) the porosity, (ii) the extent to which the pores are filled with water, and, (iii) the number, size and shape of interconnecting passages all vary with depth. The moisture profile is affected by material type (directly influencing porosity etc., or indirectly influencing permeability and water table location), topography (influencing the water table location), compaction (influencing porosity), and season (rates of precipitation, evaporation).

It is evident that many parameters may affect the ground conductivity. Fortunately at any given location relatively few are usually dominant however the survey interpreter should be aware of the possible alternatives. It is hoped that this brief treatment will be useful in assessing those parameters that are influential in the survey area and will thereby lead to a more accurate interpretation of survey data.

The bibliography gives a good indication of current interest in the electrical properties of rocks and soils and the curious reader will find much valuable material in the cited references.

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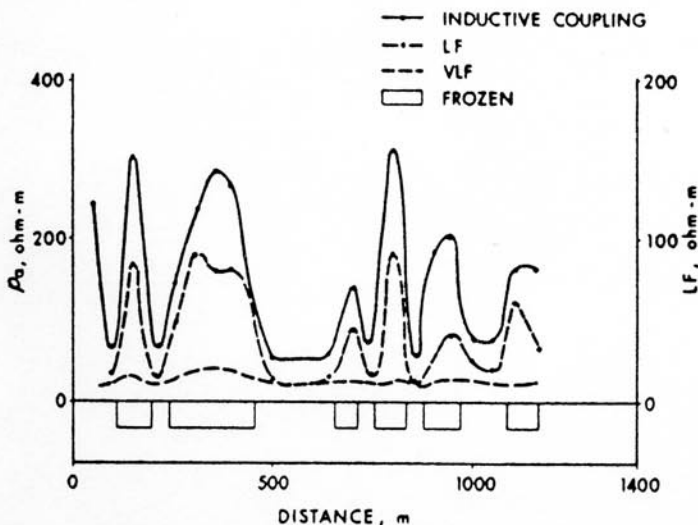[16] Jackson, P.D., Taylor Smith, D., Stanford, P.N. (1978) "Resistivity –

FIGURE 16. Results of the apparent resistivity measured with VLF (18.6 kHz), LF (375 kHz) radiohm and a magnetic induction instrument (EM31) over a section of shallow discontinuous permafrost (after P. Hoekstra [30]).

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