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GEOLOGICAL SURVEY

ELECTRICAL AND MAGNETIC PROPERTIES OF
ROCK AND SOIL

by

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UNITED STATES
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ABSTRACT

Field and laboratory measurements have been made to determine
the electrical conductivity, dielectric constant, and magnetic
permeability of rock and soil in areas of interest in studies of
electromagnetic pulse propagation. Conductivity is determined by
making field measurements of apparent resisitivity at very low
frequencies (0-20 cps), and interpreting the true resistivity of
layers at various depths by curve-matching methods. Interpreted
resistivity values are converted to corresponding conductivity
values which are assumed to be applicable at 10^2 cps, an assumption
which is considered valid because the conductivity of rock and soil
is nearly constant at frequencies below 10^2 cps. Conductivity is
estimated at higher frequencies (up to 10^6 cps) by using statistical
correlations of three parameters obtained from laboratory measurements
of rock and soil samples: conductivity at 10^2 cps, frequency and
conductivity measured over the range 10^2 to 10^6 cps. Conductivity
may also be estimated in this frequency range by using field measure-
ments of water content and correlations of laboratory sample measure-
ments of the three parameters: water content, frequency, and
conductivity measured over the range 10^2 to 10^6 cps. This method
is less accurate because nonrandom variation of ion concentration
in natural pore water introduces error.

Dielectric constant is estimated in a similar manner from
field-derived conductivity values applicable at 10^2 cps and
statistical correlations of three parameters obtained from lab-
oratory measurements of samples: conductivity measured at 10^2 cps,
frequency, and dielectric constant measured over the frequency range
10^2 to 10^6 cps. Dielectric constant may also be estimated from field
measurements of water content and correlations of laboratory sample
measurements of the three parameters: water content, frequency, and
dielectric constant measured from 10^2 to 10^6 cps, but again, this method
is less accurate because of variation of ion concentration of pore water.

Special laboratory procedures are used to measure conductivity and
dielectric constant of rock and soil samples. Electrode polarization
errors are minimized by using an electrode system that is electro-
chemically reversible with ions in pore water.

Magnetic permeability is calculated from measurements of magnetic susceptibility made at 10^3 cps. These values are applicable over this frequency range 10^2 to 10^6 cps because magnetic permeability is nearly constant in this range.

INTRODUCTION

Three properties of rock and soil are important in studies of the propagation and attenuation of the electromagnetic pulse: electrical conductivity, dielectric constant, and magnetic permeability. The frequency range of interest in electromagnetic pulse problems is generally considered to be 10^2 to 10^6 cps. This report describes field and laboratory methods of measuring the three parameters in this frequency range.

ELECTRICAL CONDUCTIVITY

Standard geophysical equipment may be used to measure electrical resistivity of rock and soil in place. A photograph of typical field equipment is shown in figure 1. Equipment of this type is designed to operate at very low frequencies, generally between 0 and 20 cps, in order to avoid skin effect problems. The flow of current would be limited at the depths of interest if higher frequencies were used. Four electrodes, usually metal stakes, are driven a few inches into the soil in an inline array. A metered source of electric current is connected to the two outer electrodes, and the resulting potential difference between the two inner electrodes is measured. A number of different electrode configurations may be used, but the most common ones for shallow measurements (depths ranging from a few inches to

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Figure 1.--Typical field equipment for measuring electrical
resistivity at low frequencies. This particular equipment
operates at 3 cps.

a few thousand feet) are the Wenner and the Schlumberger configurations shown in figure 2. Apparent resistivity is computed as a function of measured current and potential and electrode spacing by use of the formulas shown in figure 2. In practice a resistivity depth profile is obtained by making a series of measurements with the electrodes expanded symmetrically about the center point of the array, causing the average depth of current flow to increase. Apparent resistivity is computed from measurements made at each electrode spacing, and results are plotted against the electrode spacing on double logarithmic graph paper as shown in figure 3.

True resistivity is interpreted from graphs of apparent resistivity plotted against electrode spacing by curve-matching techniques. This interpretation is necessary because apparent resistivity is not equal to true resistivity, except in the theoretical case of the semi-infinite isotropic conducting medium, a case which is never found in nature. In making interpretations of true resistivity it is generally assumed that the earth beneath the electrodes consists of flat-lying electrically homogeneous layers, each characterized by a discrete resistivity. This assumption is usually sufficiently valid to obtain meaningful interpretations, although in certain geologic environments, such as those where faults or steeply dipping strata occur, serious errors may result. In these cases it is necessary to use more sophisticated interpretation procedures such as those described by Vozoff (1960) and Alfano (1959). A description of these methods is beyond the scope of this report, however, so the discussion will be limited to the simple case of flat-lying layers.

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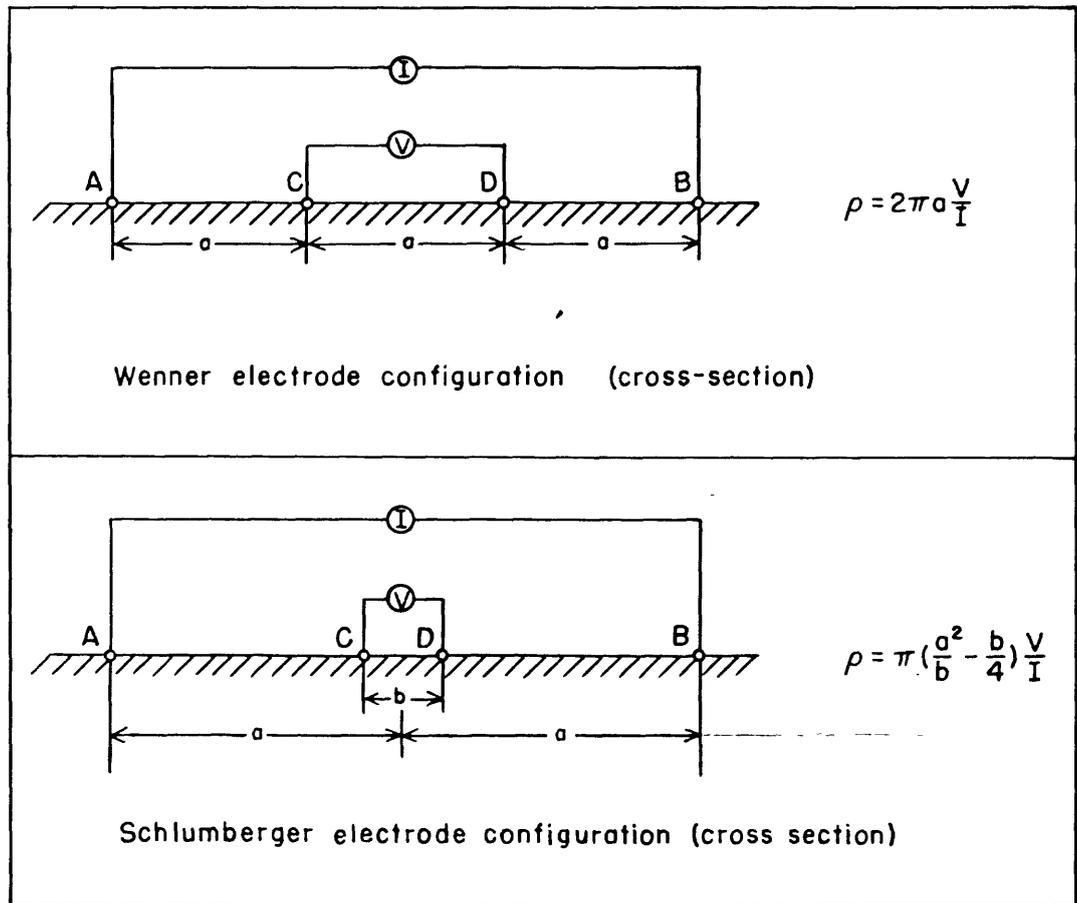
Figure 2. Electrode configurations and formulas used to determine
apparent resistivity. In the formulas V and I represent electrical
potential and current, and a and b represent distances between
electrodes indicated in the diagrams.

APPARENT RESISTIVITY, OHM-METERS

ELECTRODE SPACING (a), FEET

STATION D-2, DIPOLE ANTENNA SITE

YUCCA FLAT, NTS

INTERPRETATION		
DEPTH	RESISTIVITY	ohm-meters
0 TO 0.5 feet	190	610
0.5 TO 3.5	220	
3.5 TO 17	220	
17 TO 100	350	
100 TO > 380	250	
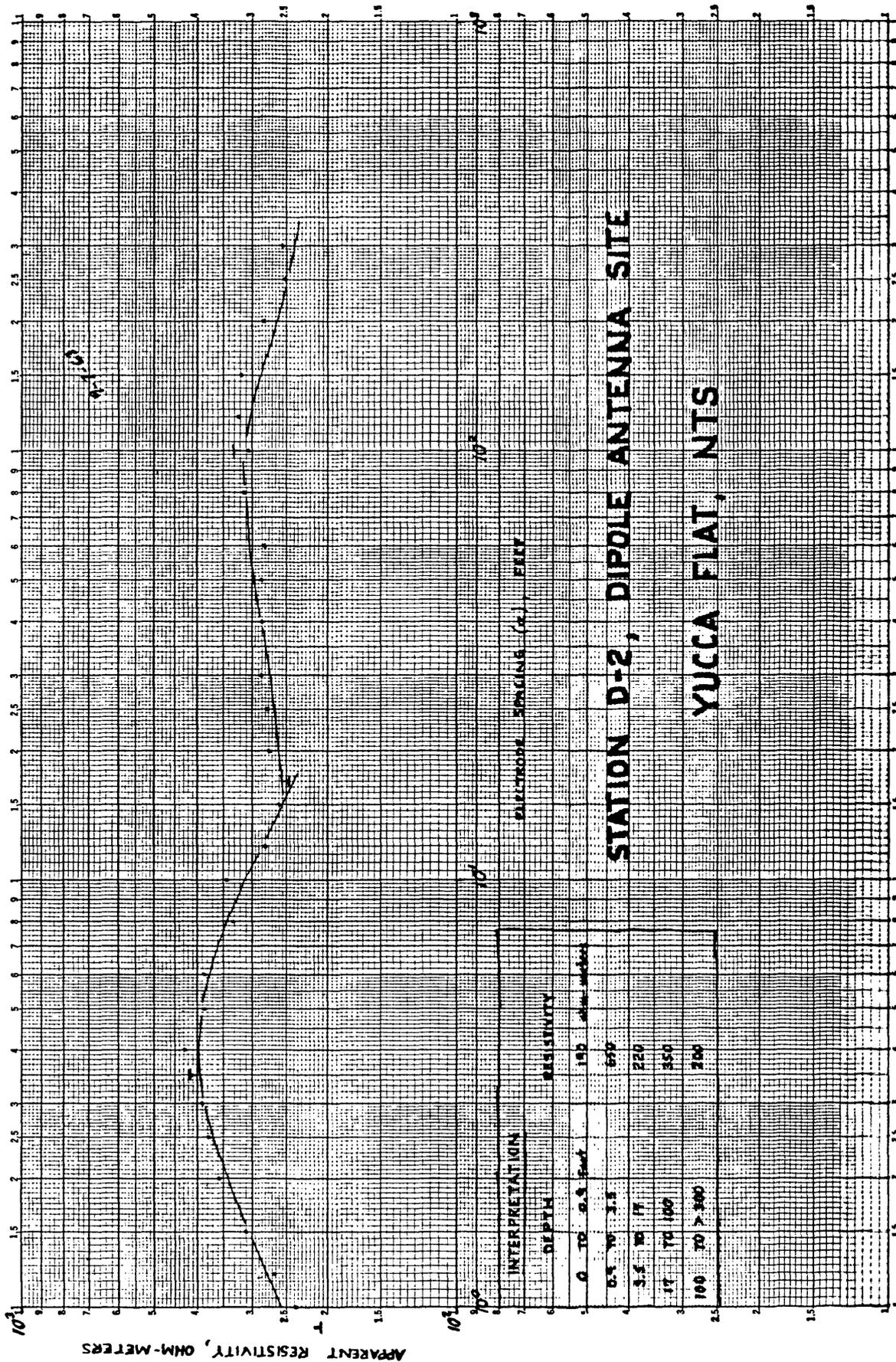
Figure 3.--Apparent resistivity values obtained from field measurements plotted against electrode spacing. Interpretation was made by the curve-matching method.

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Various theoretically derived curves are available for use in making layered earth interpretations by the curve-matching method. Two-layer curves for interpreting measurements of apparent resistivity made with the Schlumberger electrode configuration are shown in figure 4. These curves may be constructed by methods described by Stefanesco (1930). Methods used to construct similar two-layer curves for the Wenner configuration are described by Roman (1960).

The theoretically derived two-layer curves are drafted on tracing paper, having double logarithmic scales, with the same modulus as the graph paper used to plot the field data. In practice the theoretically derived curves are placed underneath the graph of field data on a light table, and the curves are carefully adjusted until a good fit is obtained between one of them. The left-hand branch of the graph of field data represents the upper two layers. The true resistivities of the upper two layers and the depth of the boundary between them is then determined as follows. The true resistivity of the uppermost layer, ρ_1, is obtained by noting where the "resistivity index" of the theoretically derived curve intersects the apparent resistivity scale of the graph of the field data. The true resistivity of the second layer, ρ_2, is determined by multiplying ρ_1 by the ratio ρ_2/ρ_1 represented by the matched curve. The boundary between the two layers is determined by noting where the "depth index" of the theoretically derived curve intersects the electrode spacing scale of the graph of the field data.

TWO-LAYER CURVES
FOR INTERPRETING
SCHLUMBERGER CONFIGURATION
FIELD CURVES
$\rho_1 < \rho_2$

$\rho_2/\rho_1 = \infty$ K = 1.0

19/1	0.9
9/1	0.8
17/3	0.7
4/1	0.6
3/1	0.5
7/3	0.4
13/7	0.3
3/2	0.2
11/9	0.1

DEPTH INDEX

RESISTIVITY INDEX

RESISTIVITY INDEX

DEPTH INDEX

$\rho_2/\rho_1 = 9/11$	K = 0.1
2/3	0.2
7/13	0.3
3/7	0.4
1/3	0.5
1/4	0.6
3/17	0.7
1/9	0.8
1/19	0.9

1/∞
1.0

TWO-LAYER CURVES
FOR INTERPRETING
SCHLUMBERGER CONFIGURATION
FIELD CURVES
$\rho_1 > \rho_2$
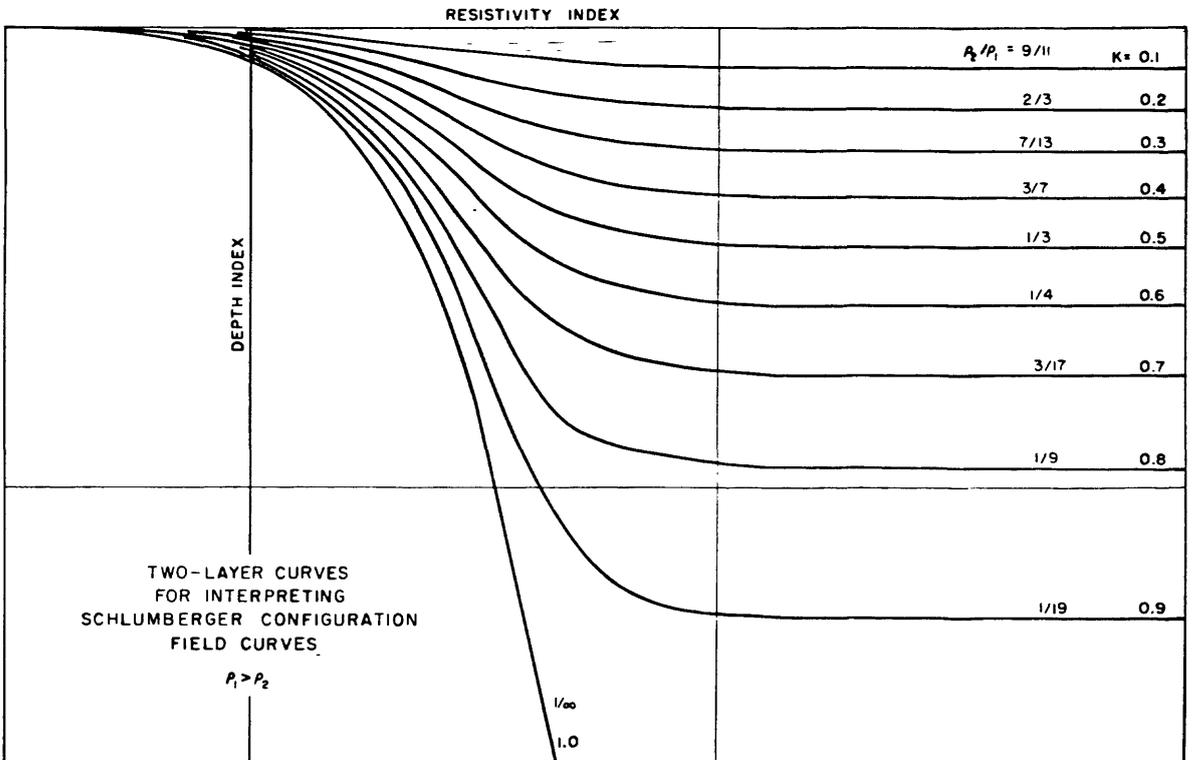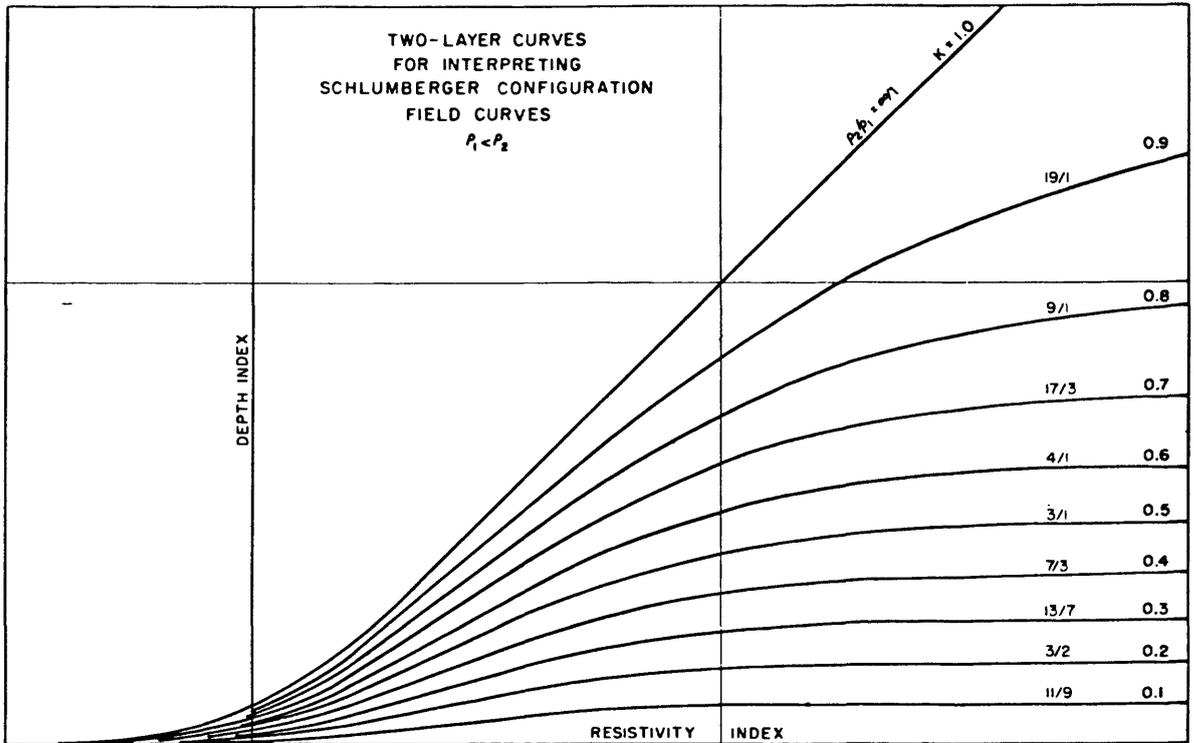
Figure 4.--Theoretically-derived two-layer curves for interpreting true
resistivity from graphs of field resistivity data. Field data and
theoretically-derived curves are plotted on graph paper with double
logarithmic scales having the same modulus.

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For field data obtained with the Schlumberger configuration, true resistivities and depths of boundaries of layers beneath the first two layers are interpreted by using auxiliary curves, shown in figure 5, together with the theoretically derived two-layer curves shown in figure 4. The auxiliary curves are first used to lump together the resistivities and thicknesses of the two upper layers so that they may be considered a single layer for interpretive purposes. Then the true resistivity of the third layer is determined by matching one of the theoretically derived two-layer curves to this composite layer and to the branch of the field curve representing the third layer. When the match has been made, the true resistivity, ρ_3, is obtained by multiplying the composited resistivity of the upper two layers by the ratio ρ_2/ρ_1, represented by the matched curve. The true depth of the boundary between the second and third layers is determined by using depth index of the two-layer curves and the dashed lines on the auxiliary curves. This procedure of lumping the resistivities of upper layers and determining the true resistivity of the next deeper layer is continued downward on the graph of field data until resistivities and boundary depths for all layers have been interpreted. Four types of auxiliary curves are available for the four possible combinations of resistivities of three-layer groups (fig. 5). The type H auxiliary curves are used when $\rho_1 > \rho_2 < \rho_3$, type K curves when $\rho_1 < \rho_2 > \rho_3$, type Q when $\rho_1 > \rho_2 > \rho_3$, and type A when $\rho_1 < \rho_2 < \rho_3$. A detailed description of use of the auxiliary curves and theoretically derived two-layer curves is given by Keller and Frischknecht (1966) and by Zohdy (1965) who also discuss the theoretical basis of the auxiliary curve method of interpretation.

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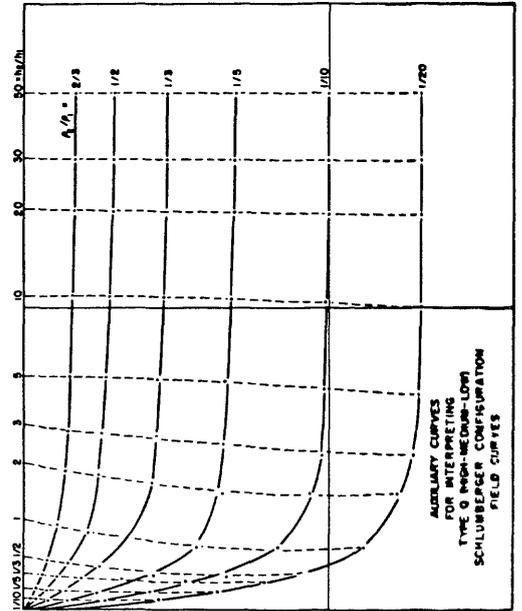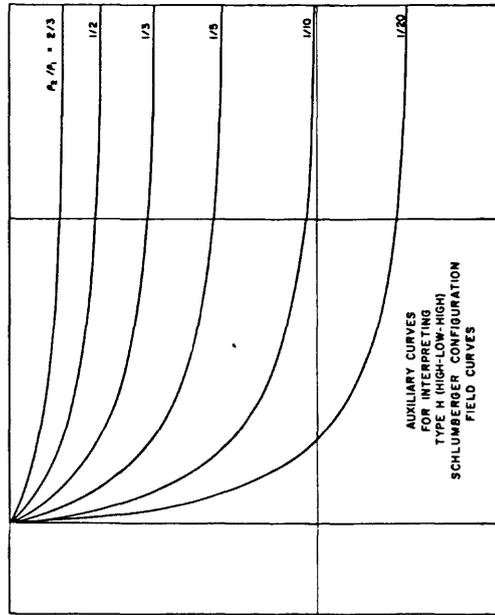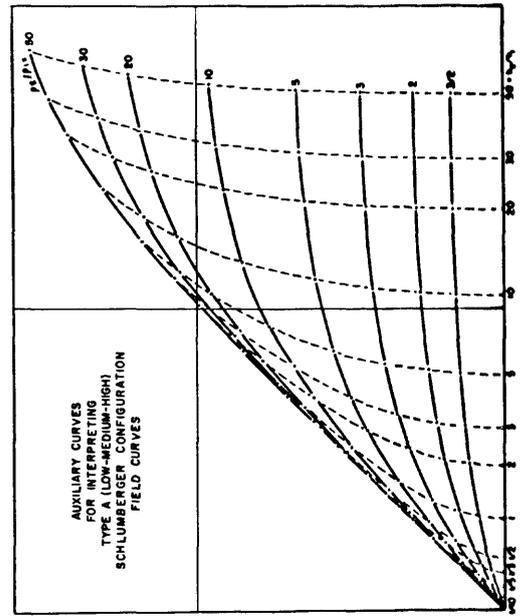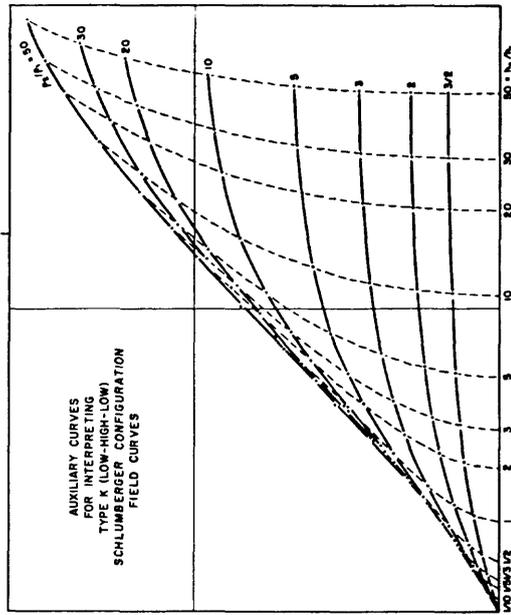
Figure 5.--Auxiliary curves for interpreting Schlumberger configuration field curves. Curves are plotted using double logarithmic scales having the same modulus as graphs of field data.

The auxiliary curve technique has also been used to interpret Wenner configuration field resistivity data (Ono, 1959); however, theoretically derived curves for 3 and 4 layers are used more commonly to interpret field measurements made with this configuration (Mooney and Wetzel, 1956).

It is possible to make resistivity interpretations automatically, using a digital computer to fit field-derived values with a theoretically-derived curve. However, experimental applications of this method have indicated that interpreted results are widely divergent depending on what particular mathematical technique of curve fitting is used (Vozoff, 1958). Until these numerical difficulties are overcome, it is generally agreed that better interpretations can be obtained by using curve-matching methods, together with judgement, based on experience and knowledge of geology and characteristic resistivities in the area of interest. Accuracy of these interpretations is usually considered to be approximately 10-30 percent, although in certain difficult cases, for example a thin layer of high resistivity between two thick layers of low resistivity, errors of 100 percent or more are possible.

After a resistivity interpretation has been made by the curve-matching method, it is advantageous to use a computer program to check the accuracy and modify the interpretation if necessary so that a more perfect fit to the field data can be achieved. A program that operates in this manner is being developed by the U.S. Geological Survey at the present time.

An interesting question arises regarding the ability to detect
the top of the ground-water table by resistivity field measurements.
Since resistivity is strongly dependent on water content, one would
expect that this would be rather easy to do. In areas where the water
table exists within well drained material, such as coarse sand or
gravel, it is usually possible to determine the depth of the water
table quite accurately. However, more commonly the water table
occurs in fine-grained soil or rock which is highly saturated above
the water table. In these cases it is usually difficult or impossible
to detect the top of the water table by resistivity field measurements.
An example of this is the water table in tuff at the Nevada Test Site.
The tuff above the water table commonly has a saturation of over
90 percent, and there is no detectable difference in the resistivity
above and below the water table, even when sensitive electric logs
are made in drill holes.

Rough estimates of resistivity may be made for areas where no field resistivity data are available by use of tables, approximate formulas or maps. Table 1 may be used to estimate the low-frequency resistivity and conductivity of various types of rock and soil of different geologic ages.

If the porosity, the degree of saturation, and the resistivity of the pore fluid of a rock or soil are known, the following approximate formula may be used to estimate its resistivity at low frequencies (Keller, 1962):

$$\rho \eqsim a\, \rho_w\, (S\emptyset)^{-n} \tag{1}$$

where

ρ is the estimated resistivity of rock or soil, ohm-meters,

ρ_w is the resistivity of water in rock or soil, ohm-meters,

S is the fraction of total pore space filled with water,

$\emptyset$ is the fractional volumetric porosity,

a and n are empirically determined constants for particular rock or soil types. The value for $\underline{a}$ generally falls in the range $0.6 < a < 1.2$, and that of $\underline{n}$ in the range $1.6 < n < 2.2$.

At high frequencies conductivity increases slightly, the increase being greater for low-conductivity than for high-conductivity rock and soil. Very rough estimates of the conductivity of surface material at standard broadcast frequencies ($\sim 10^6$ cps) for various parts of the United States may be made from the map shown in figure 6. This map is based on measurements of ground-wave field strength for commercial

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Table 1.--Typical values of electrical conductivity and resistivity of various types of rock and soil of different geologic ages. Values are applicable in the frequency range 0 to 10^2 cps

[Modified after Sunde (1949).]

Electrical conductivity (millimho/meter)	Electrical resistivity (ohm-meters)	Geologic period and rock type				
		Quaternary	Quaternary Tertiary Cretaceous	Jurassic Triassic Carboniferous	Devonian Silurian Ordovician Cambrian	Cambrian Precambrian
3333 (Seawater)	0.3					
100	10	Shallow playa deposits	Loam Clay			
33 (Yucca Flat ground water)	30		Chalk	Chalk Trap		
10 (Very fresh water)	100		Alluvium	Alt. basalt Shale		
3.3	300			Limestone Sandstone	Shale Limestone	
1.0	1,000	Coarse sand and gravel in surface layers			Sandstone Dolomite	
0.33	3,000					Sandstone
0.10	10,000					Quartzite Slate Granite Gneiss

NUMBERS ON MAP REPRESENT ESTIMATED EFFECTIVE
GROUND CONDUCTIVITY IN MILLIMHOS PER METER

CONDUCTIVITY OF SEAWATER IS NOT SHOWN ON MAP
BUT IS ASSUMED TO BE 5000 MILLIMHOS PER METER
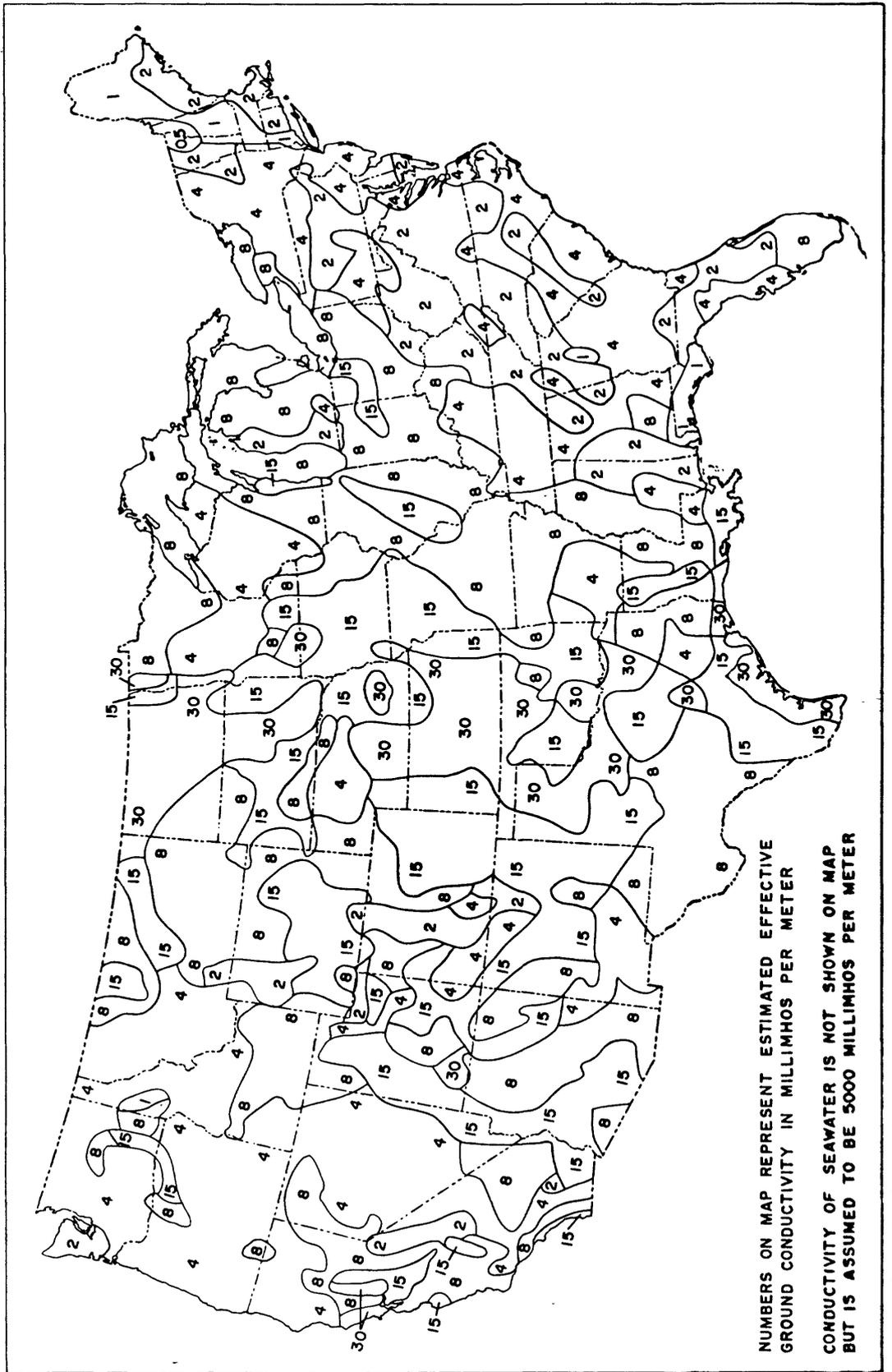
Figure 6.--Estimated effective ground conductivity in the United States applicable at frequencies in the standard broadcast band (after Fine, 1954).

broadcast stations (Fine, 1954). The values apply to very shallow

surface material, because skin effect limits depth penetration

severely at these frequencies.

More accurate estimates of conductivity over the range 10^2 to

10^6 cps may be made by using values of conductivity interpreted

from field resistivity measurements, together with statistical

correlations of laboartory measurements of conductivity made over

this frequency range. Since the conductivity of rock and soil

containing natural pore water is nearly constant between 0 and

10^2 cps, conductivity values obtained from interpretations of field

resistivity measurements made in the range 0-20 cps may be assumed

to be applicable at 10^2 cps. Errors due to this assumption are

usually less than 1 percent.

The procedure recommended for estimating conductivity over the

range 10^2 to 10^6 cps is outlined as follows.

1. Make field measurements of resistivity at low frequencies

and use the curve-matching method of interpretation to determine the

true resistivity and thickness of each electrical layer at the field

site. Convert the interpreted resistivity values to conductivity and

assume that these values are applicable at 10^2 cps.

2. Use formulas or curves representing statistical correlations

of data obtained from laboratory measurements of rock and soil samples

to estimate conductivity at any frequency between 10^2 and 10^6 cps. The

three parameters that are correlated are: conductivity measured at

10^2 cps, frequency, and conductivity measured at 10^2, 10^3, 10^4, 10^5,

and 10^6 cps.

Correlations of all laboratory conductivity data available at the present time at sites of interest in electromagnetic pulse studies have been used to obtain the following formula for estimating conductivity by the method described above. The formula represents a second degree surface fitted by the method of least squares to the logarithms of the correlated parameters.

$$K = 0.028 + 1.098K_{100} - 0.068F + 0.036K_{100}^2 - 0.046FK_{100} \qquad (2)$$
$$+ 0.018F^2$$

Standard error $= 0.058$

Where K is $\log_{10}$ of conductivity (millimhos/meter) at a specified

frequency in the range 10^2 to 10^6 cps,

K_{100} is $\log_{10}$ of conductivity (millimhos/meter) determined from

interpretations of field resistivity measurements which

are assumed to be applicable at 100 cps,

F is $\log_{10}$ of frequency (cps) at the specified frequency in

the range 10^2 to 10^6 cps.

A graph showing a family of curves representing equation (2) at 10^2, 10^3, 10^4, 10^5, and 10^6 cps is shown in figure 7. Data points shown in the figure represent conductivity measurements at only one fo the five frequencies, 10^4 cps, plotted against conductivity measured at 10^2 cps. The data points are presented as an example of the degree of scatter of the measured values which is about the same at all frequencies.
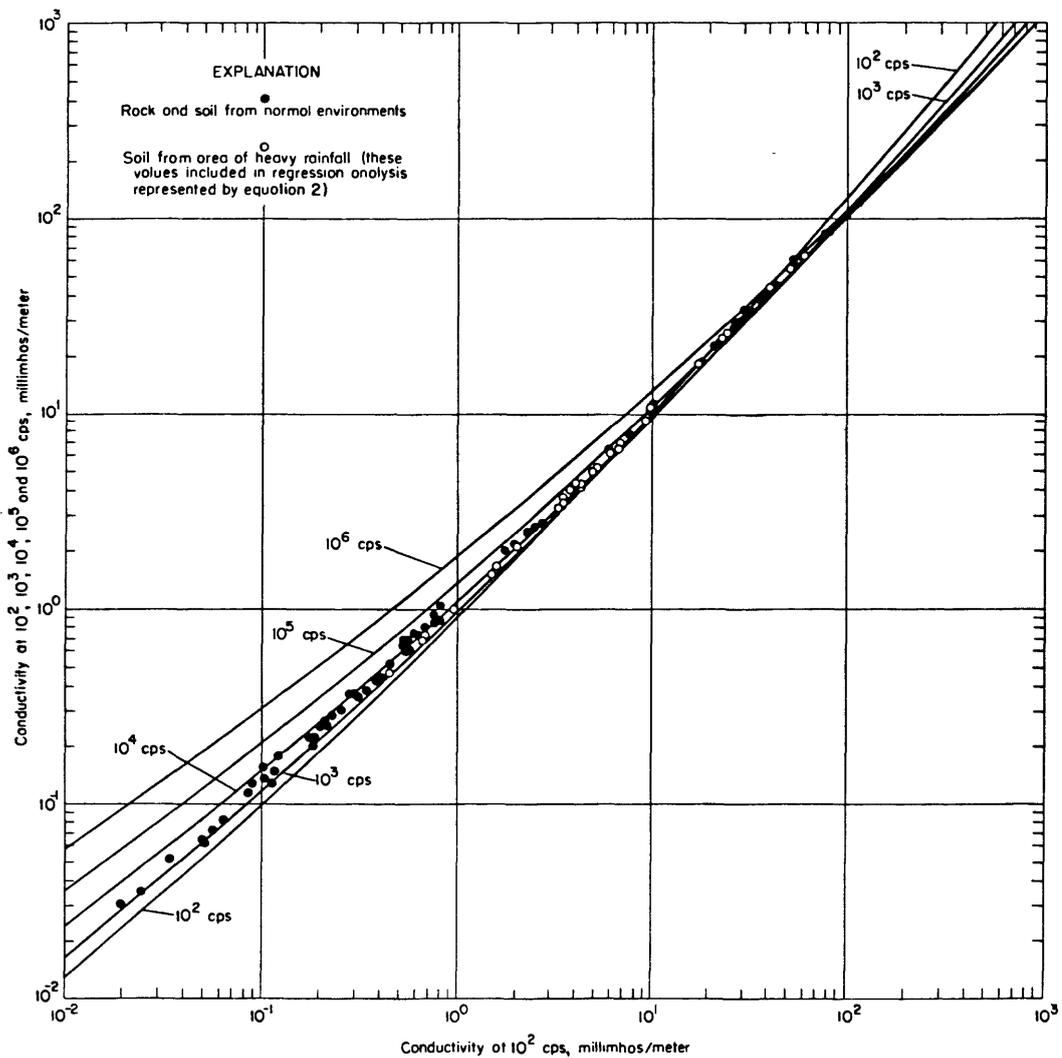
Figure 7.--Family of curves that may be used to estimate conduc-
tivity of rock and soil in the frequency range 10^2 to 10^6 cps
from known conductivity at 10^2 cps. Data points represent con-
ductivity of samples measured in the-laboratory at 10^4 cps
plotted against conductivity measured at 10^2 cps.

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If field resistivity measurements are not available in the area of interest, conductivity may be estimated from measurements of water content of soil and rock at the field site. This is accomplished by using statistical correlations of laboratory measurements of conductivity, frequency and water content of rock and soil samples. However, the estimate made by this method is less accurate than the one previoulsy described, because natural variations of ion concentration of pore water in rock and soil in different environments also influence the conductivity values. The following formula for estimating conductivity as a function of water content and frequency was obtained by fitting a second degree mathematical surface by the method of least squares to laboratory data for the same set of samples used to obtain equation (2).

$$K = -0.604 + 1.640W - 0.062F + 0.062W^2 - 0.070FW + 0.021F^2 \quad (3)$$

Standard error = 0.242

where K is $\log_{10}$ of conductivity (millimhos/meter) at a specified

frequency in the range 10^2 to 10^6 cps,

W is $\log_{10}$ of water content (percent by volume) or rock or

soil for which estimate is being made,

F is $\log_{10}$ of frequency (cps) at the specified frequency in

the range 10^2 to 10^6 cps.

A graph showing the family of curves representing equation (3) at 10^2, 10^3, 10^4, 10^5, and 10^6 cps is shown in figure 8. Data points shown in the figure represent conductivity measurements at only one of the five frequencies, 10^4 cps, and are presented as an example of the degree of scatter of the measured values.

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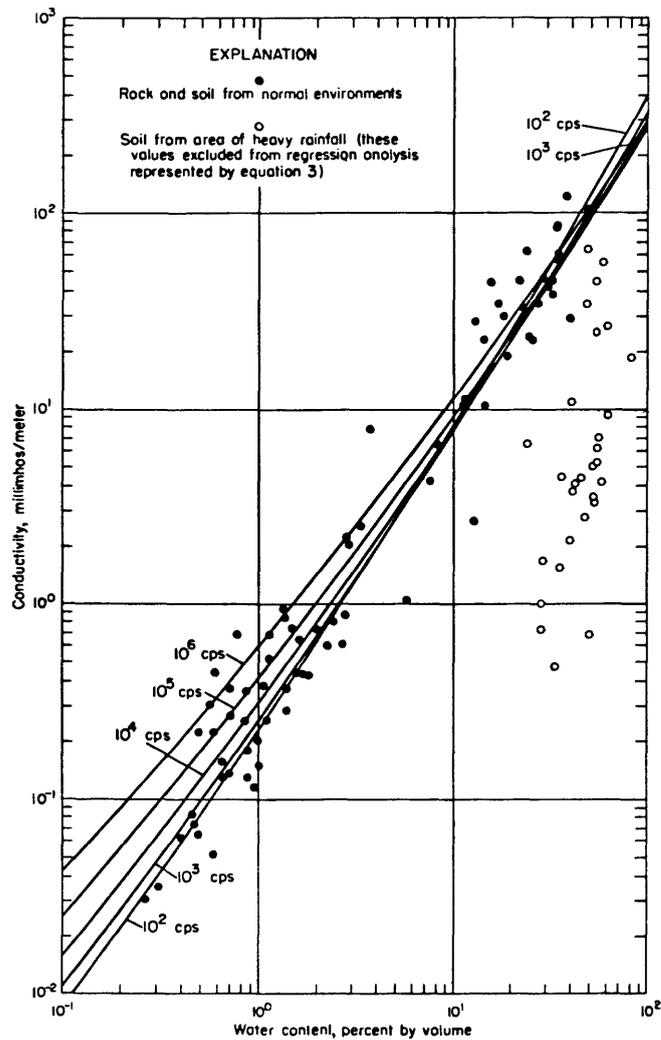
Figure 8.--Family of curves that may be used to estimate conductivity
of rock and soil in the frequency range 10^2 to 10^6 cps from known
water content. Data points represent conductivity of samples
measured in the laboratory at 10^4 cps plotted against water content.

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Since laboratory measurements are made at room temperature, a temperature correction should be applied to estimates made by this method. A useful rule-of-thumb for making the temperature correction is: as temperature increases, resistivity decreases at a rate of about 2 percent per degree centigrade in the range of normal rock temperatures above the freezing point (Richards, 1954).

It is interesting to notice, in figure 8, the group of samples designated by open circles that stands apart from the rest. These samples represent near-surface soil in an area where the rainfall is unusually high; approximately 35 inches/year. The low conductivity of these samples is attributed to the leaching action of rain water which has removed some of the natural salt content, causing the conductivity of pore water to be anomalously low. Since data for these samples were anomalous, they were not included in the regression analysis represented by equation (3). The graph indicates that rather large errors would result if equation (3) were used to estimate the conductivity of material represented by these samples. It should be noted that this problem does not exist when estimates are based on field resistivity measurements by the use of equation (2), because for that correlation the data points representing the leached samples do not form a separate group (fig. 7).

DIELECTRIC CONSTANT

At the present time there is no available field method for measuring dielectric constant of rock or soil _in situ_, although research is being conducted by the U.S. Geological Survey to determine the feasibility of a four-electrode system from which dielectric constant would be determined at low frequencies by sensitive phase-shift measurements.

Since no field method is available now, it is necessary to make use of statistical correlations of laboratory measurements of rock and soil samples. The following procedure is recommended for estimating dielectric constant over the frequency range 10^2 to 10^6 cps.

1. Use interpretations of field resistivity measurements made at low frequencies to estimate conductivity at 10^2 cps for various electrical layers present at the field site.

2. Use formulas or curves representing statistical correlations of data obtained from laboratory measurements of rock and soil samples to estimate dielectric constant at frequencies between 10^2 and 10^6 cps. The three parameters that are correlated are: conductivity measured at 10^2 cps, frequency, and dielectric constant measured at 10^2, 10^3, 10^4, 10^5, and 10^6 cps.

The following formula was obtained by fitting a second degree mathematical surface by the method of least squares to laboratory measurements of conductivity, frequency, and dielectric constant for

the same set of samples used to obtain equation (2).

$$D = 5.491 + 0.946K_{100} - 1.097F + 0.069K_{100}^2 - 0.114FK_{100} + 0.067F^2 \quad (4)$$

Standard error $= 0.190$

where D is $\log_{10}$ of dielectric constant (ϵ/ϵ_o) at a specified

frequency in the range 10^2 to 10^6 cps,

K_{100} is $\log_{10}$ of conductivity (millimhos/meter) determined

from interpretations of field resistivity measurements

which are assumed to be applicable at 10^2 cps.

F is $\log_{10}$ of frequency (cps) at the specified frequency in

the range 10^2 to 10^6 cps.

A graph showing a family of curves representing equation (4)
at 10^2, 10^3, 10^4, 10^5, and 10^6 cps is shown in figure 9. Data
points shown in the figure represent dielectric constant measure-
ments at only one of the five frequencies, 10^4 cps, plotted against
conductivity measured at 10^2 cps. The data points are presented as
an example of the degree of scatter of the measured values.

If field resistivity values are not available in the area of
interest, dielectric constant may be estimated from measurements of
water content of soil and rock at the field site. This is accomplished
by using statistical correlations of laboratory measurements of dielec-
tric constant, frequency, and water content of rock and soil samples.
As was true in the case of conductivity estimated from water content,
the results of this method are less accurate than those based on
field resistivity measurements because of natural variations of ion
content of pore water in rock and soil in different environments.
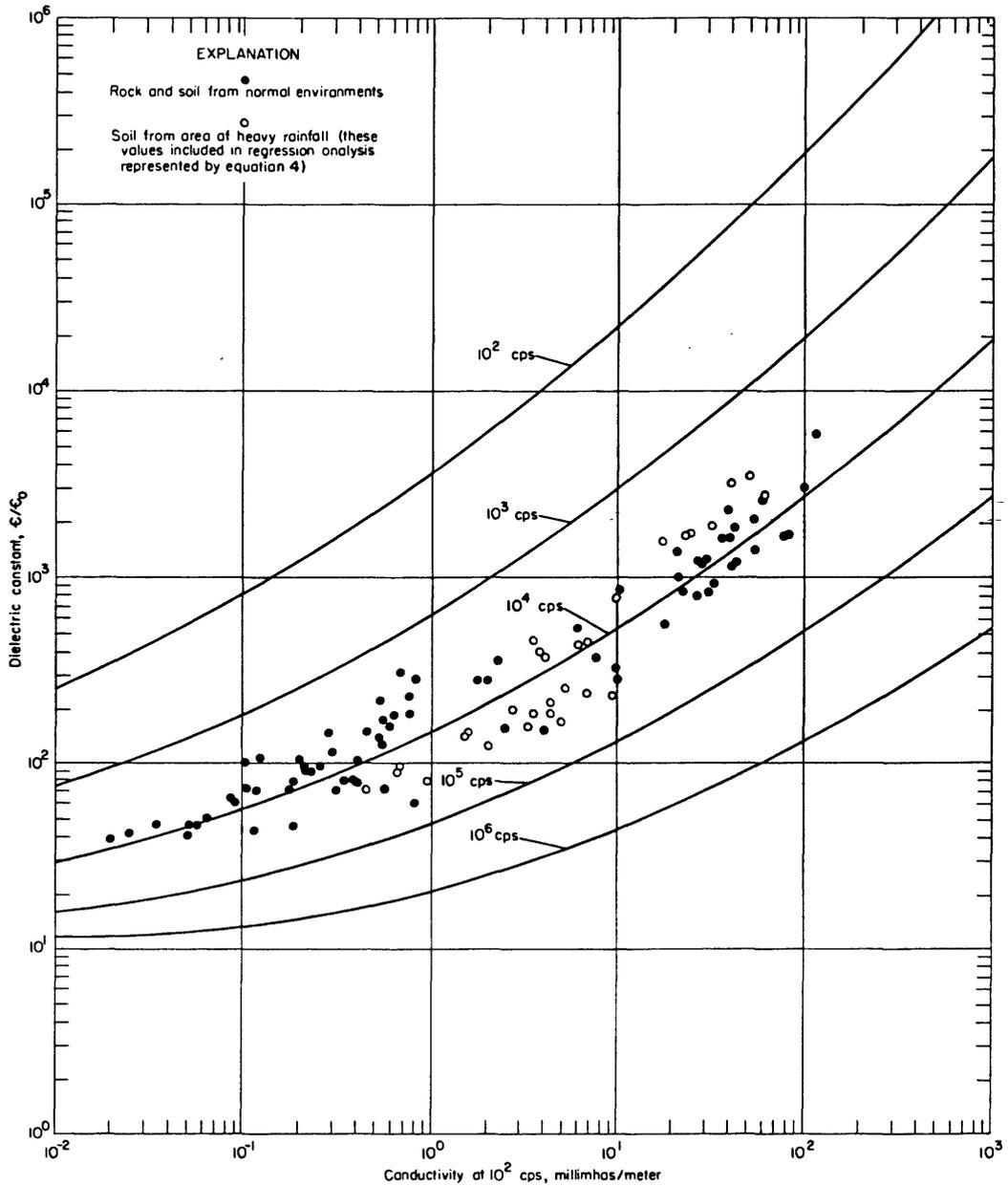
Figure 9.--Family of curves that may be used to estimate dielectric constant of rock and soil in the frequency range 10^2 to 10^6 cps from known conductivity at 10^2 cps. Data points represent dielectric constant of samples measured in the laboratory at 10^4 cps plotted against conductivity at 10^2 cps.

The following formula for estimating dielectric constant as a function

of frequency and water content was obtained by fitting a second degree

mathematical surface by the method of least squares to laboratory data

for the same set of samples used to obtain equation (3).

$$D = 4.905 + 1.308W - 0.971F + 0.111W^2 - 0.168FW + 0.059F^2 \quad (5)$$

$$\text{Standard error} = 0.233$$

where D is $\log_{10}$ of dielectric constant (ϵ / ϵ_o) at a specified

frequency in the range 10^2 to 10^6 cps,

W is $\log_{10}$ of water content (percent by volume) of rock or

soil for which estimate is being made,

F is $\log_{10}$ of frequency (cps) at the specified frequency in

the range 10^2 to 10^6 cps.

A graph showing the family of curves representing equation (5)

at 10^2, 10^3, 10^4, 10^5, and 10^6 cps is shown in figure 10. Data points

shown in the figure represent dielectric constant measurements at only

one of the five frequencies, 10^4 cps, and are presented as an example

of the degree of scatter of the measured values.

It is interesting to notice that in figure 10 as in figure 8,

the group of samples designated by open circles stands apart from the

rest. These samples are the ones that represent soil in an area of

high annual rainfall. Again, the anomalously low dielectric constant

of these samples is attributed to the leaching action of rain water.

Data for these samples were not used in obtaining equation (5) because

they are considered anomalous. Large errors would result if equation

(5) were used to estimate the dielectric constant of material represented
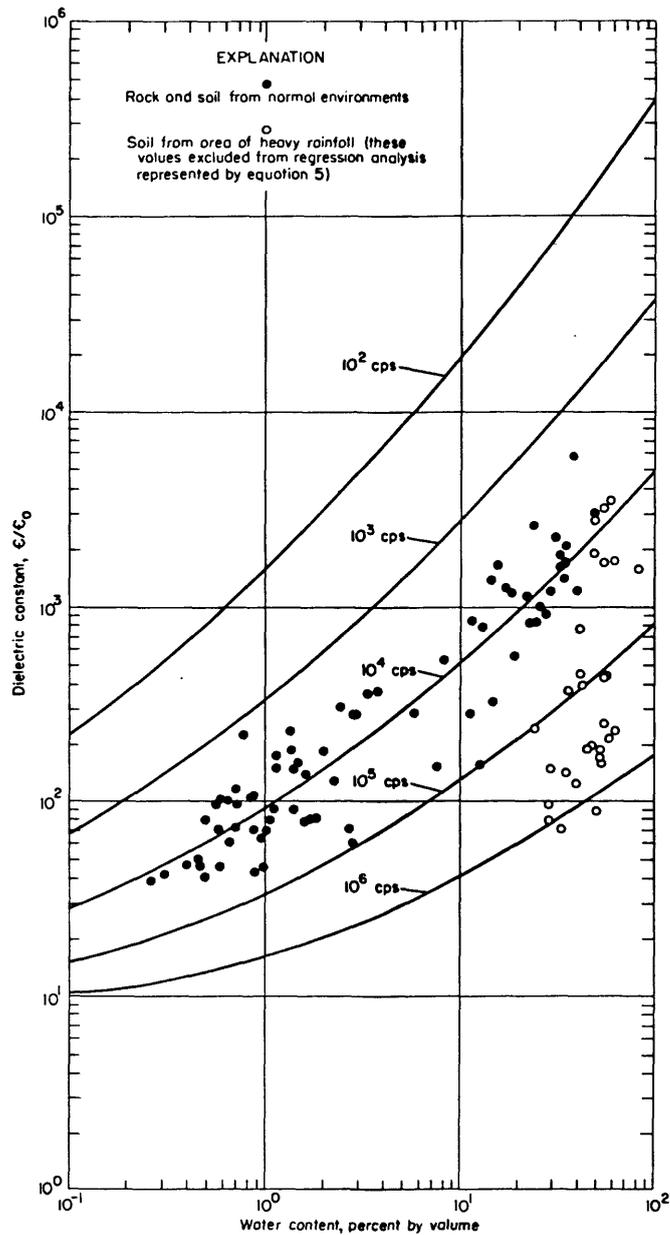
Figure 10.--Family of curves that may be used to estimate dielectric constant of rock and soil from known water content. Data points represent dielectric constant of samples measured in the laboratory at 10^4 cps plotted against water content.

by these samples. It should be noted that this problem does not exist when estimates are based on resistivity field measurements by the use of equation (4), because in this correlation the leached samples do not form a separate group (fig. 9).

LABORATORY METHODS USED TO DETERMINE CONDUCTIVITY AND DIELECTRIC CONSTANT

Conductivity and dielectric constant are measured in the laboratory over the frequency range 10^2 to 10^6 cps using natural-state rock and soil samples. In the case of rocks, samples are prepared by cutting a 1-inch diameter core out of a larger sample, and cutting the core to a length of 1 inch. In the case of soils, it is usually impossible to collect a sample at the field site and preserve it intact. Therefore, when the soil sample is collected it is sealed in a plastic bag to prevent water loss, and then density is measured in a shallow augered hole at the field site, using a gamma-gamma density probe such as the one described by Cameron and Bourne (1958). In the laboratory the soil sample is repacked to the field-measured density in a plastic cylinder having a length of 1 inch and an inside diameter of 1 inch.

Equipment used to make the laboratory measurements is shown in figure 11. The equipment consists of the following: a sample holder, oscillators, capacitance-resistance bridges, a tunable null indicator (10^2 to 10^5 cps), a radio used as a tuner at 10^6 cps, and a frequency counter. The most difficult problem to overcome in making laboratory measurements of conductivity and
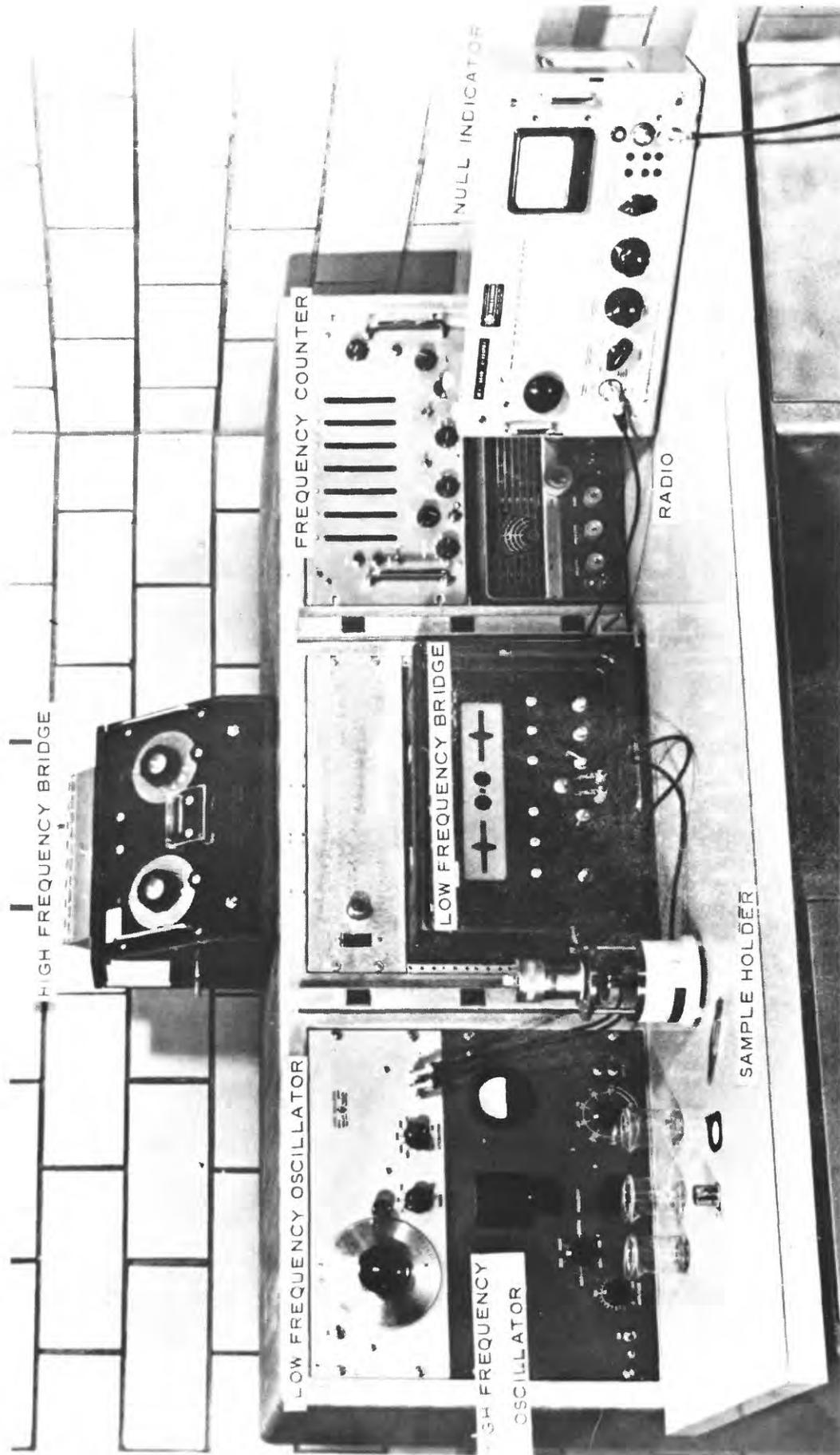
Figure 11.--Laboratory equipment used to make measurements of dielectric constant and conductivity of natural-state samples of soil and rock.

dielectric constant of moist soil and rock samples is electrode polarization, which causes measured values of capacitance to be too large, sometimes by as much as an order of magnitude. To overcome this problem a special electrode arrangement was devised as shown in figure 12. The sample is enclosed between two blotter pads saturated with an aqueous suspension of finely divided particles of metallic silver and silver chloride. Platinized platinum electrodes are placed against the blotter pads, and the whole assemblage is clamped between the plates of the sample holder.

Electrode polarization is reduced so that the effect is insignificant by the action of the saturated blotter pads and the platinized platinum electrodes. The blotter pads act as electrodes that are electrochemically reversible with chloride ions in pore water in the sample. When electric current flows in one direction, the chloride ions move out of the rock sample and unite chemically with metallic silver particles in the blotter to form silver chloride. When the current is reversed, chloride ions are released from the silver chloride in the blotters and move toward the sample. The reversible electrodes prevent ions from accumulating in high concentrations at the sample-electrode interface, thereby minimizing electrode polarization. The platinized platinum electrodes placed against the blotter pads have a large effective surface area so that ions carrying current within the blotters are prevented from accumulating in high concentrations at the blotter-metal interface, and thus a low-impedence contact is achieved.

Figure 12.--Photograph of electrode arrangement used to make laboratory measurements of dielectric constant and conductivity. Rock sample is shown between the electrodes.

A number of different electrode materials and electrolytes were tested in the search for a nonpolarizing electrode system. Figure 13 shows the results of these tests which were made at the lowest frequency of interest, 10^2 cps, at which electrode polarization problems were most severe. Measurements of a sample of topsoil containing 50 percent water by volume indicated that electrode polarization errors existed with all systems except the platinized platinum electrodes and blotters filled with the aqueous suspension of silver and silver chloride particles. Largest errors were a little more than an order of magnitude, and occurred with stainless steel and polished platinum electrodes and blotter pads filled with sodium chloride solution. Figure 13 shows that for measurements of a sample of limestone containing 0.45 percent water by volume, no significant electrode polarization error occurred with any of the electrode systems.

Since water content has such an important influence on electrical properties of soil and rock, it is informative to plot relationships between conductivity, dielectric constant, and frequency as families of curves representing various values of water content. Figure 14, based on equation (3), shows conductivity plotted against frequency for various water content values. The curves show that conductivity is not strongly dependent on frequency, especially at low frequencies.
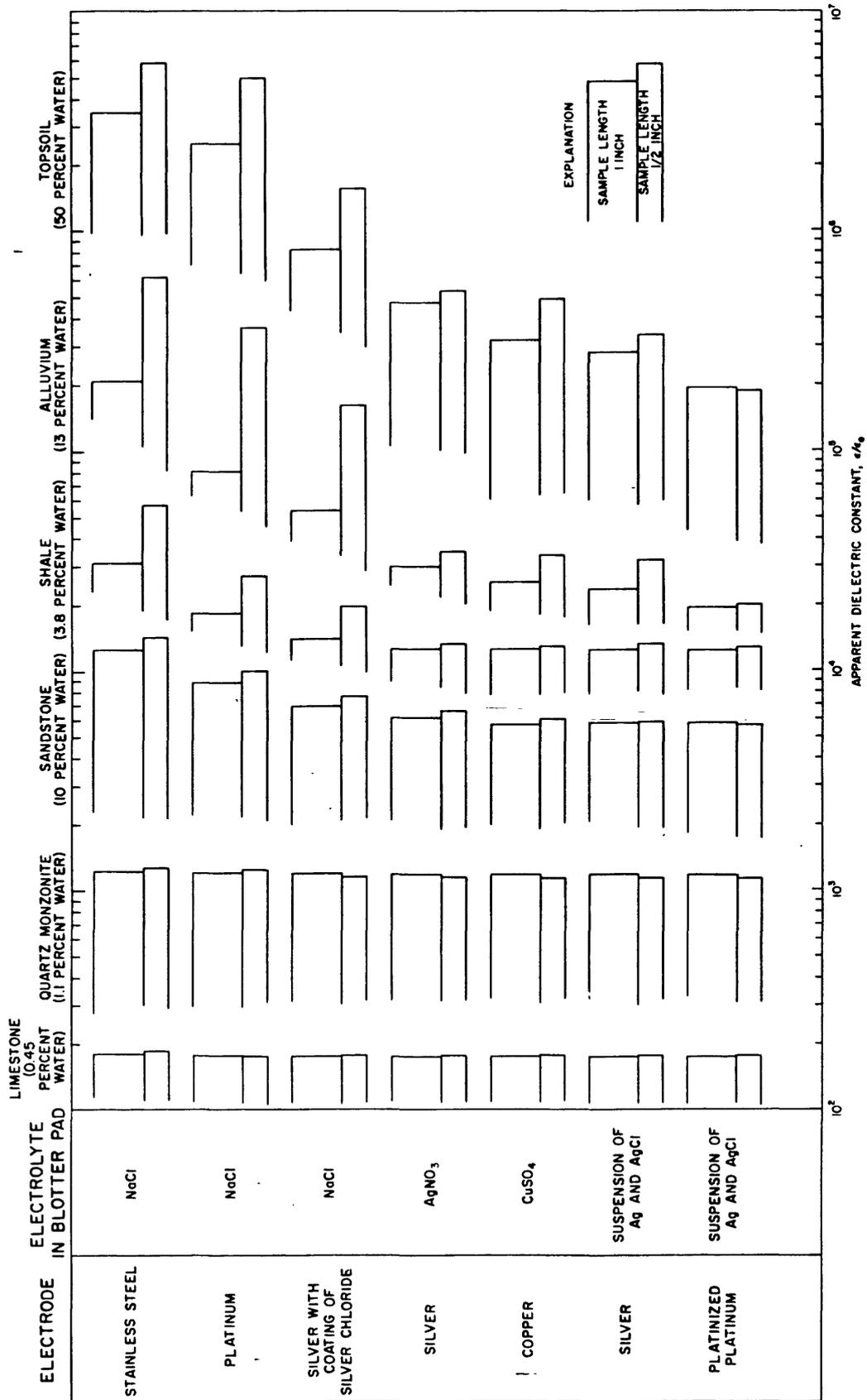
Figure 13.--Bar graph showing apparent dielectric constant measured in the laboratory at a fixed frequency of 102 cps using various electrode-electrolyte systems and several rock types. Measurements were made on 1-inch-long samples first, then repeated after ends were cut off, making samples 1/2-inch long. Wide bars represent measurements of 1-inch samples, and narrow bars represent measurements of 1/2-inch samples. When wide and narrow bars are about the same length, polarization error is minimum.

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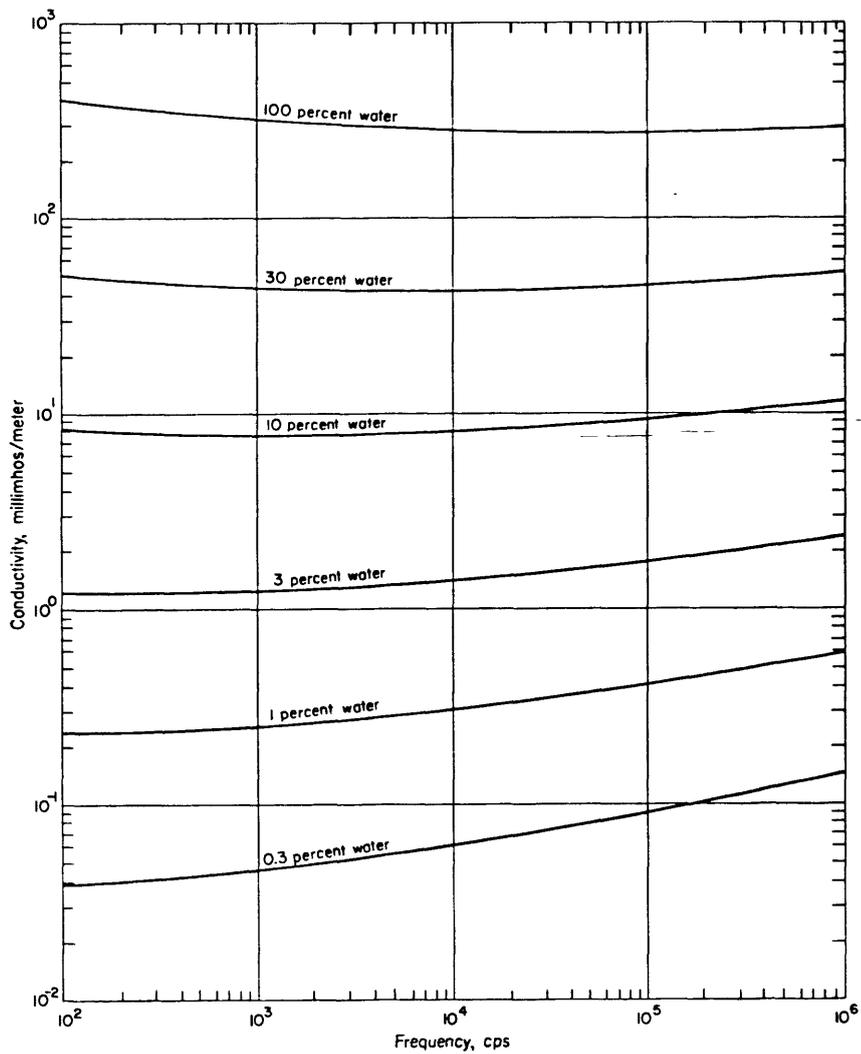
Figure 14.--Graph of conductivity plotted against frequency for laboratory measurements of samples having different water content. Curves represent equation (3).

Figure 15, based on equation (5), shows dielectric constant plotted against frequency for various water content values. Trends of the curves suggest that at some frequency above 1 megacycle the curves for all values of water content become asymptotic to a value of dielectric constant between 1 and 10.

MAGNETIC PERMEABILITY

Relative magnetic permeability (μ/μ_o) of rock and soil is less important than conductivity and dielectric constant in studies of electromagnetic pulse propagation, because for most earth materials μ/μ_o is only slightly greater than unity. Table 2 gives typical values and ranges of relative magnetic permeability for a number of types of soil and rock. Rock in iron-mining areas appears to be the only natural material for which magnetic permeability is very significant.

Magnetic permeability is determined by measuring magnetic susceptibility of rock and soil samples under a weak magnetic field (order of magnitude of 1 gauss) having a frequency of 10^3 cps (Mooney, 1952). For the purposes of this study magnetic susceptibility may be considered constant over the frequency range 10^2 to 10^6 cps; variations with frequency are barely detectible, being only a few percent (Vincenz, 1965). Commercially available bridges are calibrated so that magnetic susceptibility is measured in cgs units. Relative permeability is computed from measured susceptibility by the formula:

$$\mu/\mu_o = 1 + 4\pi k_{cgs}, \tag{6}$$

where μ/μ_o is relative magnetic permeability,

k_{cgs} is magnetic susceptibility (cgs units).
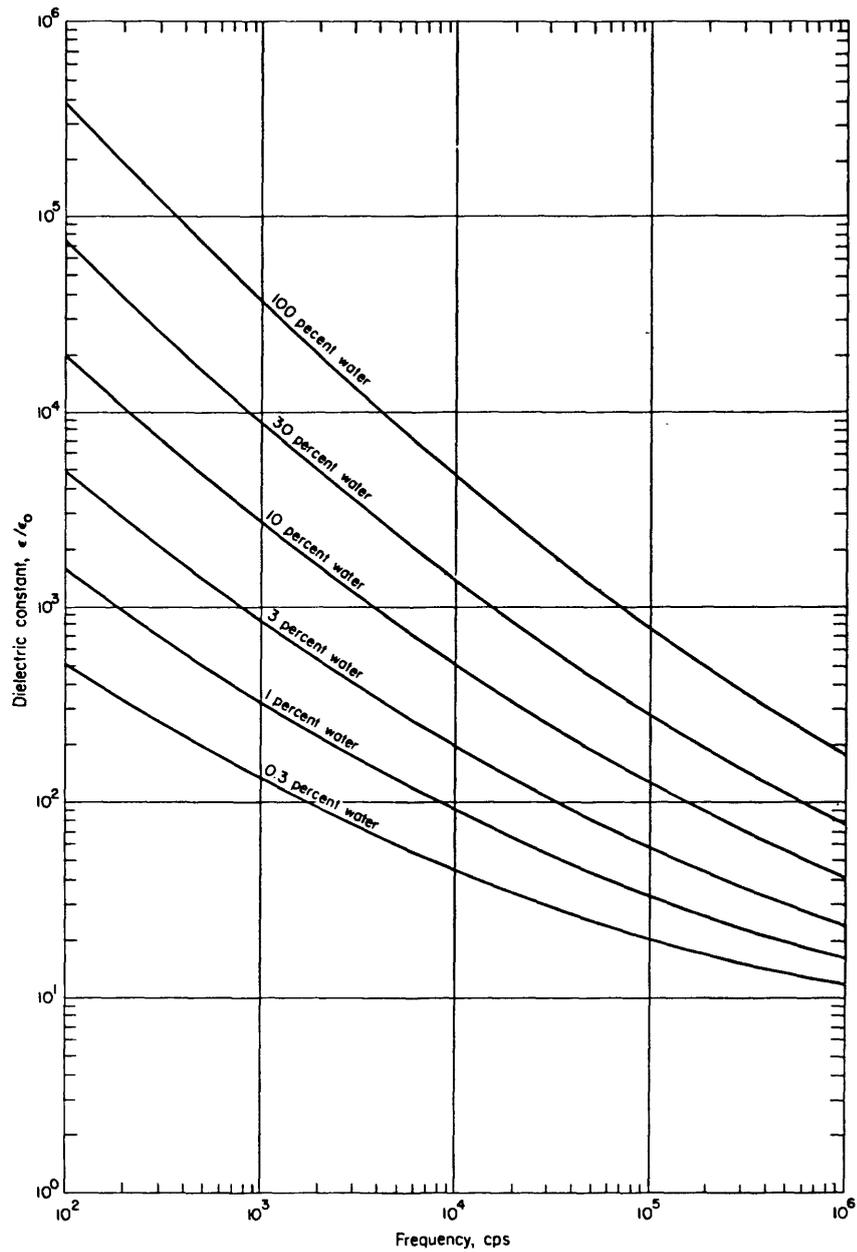
Figure 15.--Graph of dielectric constant plotted against frequency
for laboratory measurements of samples having different water
content. Curves represent equation (5).

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Table 2.--Typical values and ranges of relative magnetic
permeability of various types of rock and soil.
Values are applicable over the frequency
range 10^2 to 10^6 cps

Rock or soil type	Relative magnetic permeability (μ/μ_o)		Source of information
	Typical value	Range	
Soil and sedimentary rock (general)	1.0006	1.00001 to 1.001	Nettleton (1940)
Alluvium (NTS, Nevada)	1.004	1.0005 to 1.014	Monk (1965) (oral commun.)
Sandstone with unusually high magnetite content (Neroly Fm., California)	1.010	1.002 to 1.025	Bath (1965) (oral commun.)
Volcanic rock and soil (Amchitka Island, Alaska)	1.021	1.0001 to 1.053	Scott and Cunningham (1965) (written commun.)
Granite (Cheyenne Mtn., Colo.)	1.076	1.017 to 1.136	Scott (1965) (written commun.)
Rock in iron-mining areas (Sweden)	1.1	1.00001 to 14	Werner (1945)

SUMMARY OF RECOMMENDED METHODS

Electrical and magnetic properties of soil and rock in areas of interest in electromagnetic pulse studies may be determined by making field and laboratory measurements. The following approach is recommended.

1. Make field measurements of apparent resistivity at the site of interest.

2. Interpret true resistivity values for earth layers by the curve-matching method, and convert to corresponding conductivity values.

3. Estimate conductivity and dielectric constant at frequencies in the range 10^2 to 10^6 cps from field-derived conductivity values (which are assumed to be applicable at 10^2 cps) using statistical correlations of laboratory measurements of conductivity and dielectric constant at 10^2, 10^3, 10^4, 10^5, and 10^6 cps.

4. If field resistivity measurements are not available, measure water content of soil and rock at the field site, and estimate conductivity and dielectric constant from correlations of laboratory measurements of water content, conductivity (10^2 to 10^6 cps), and dielectric constant (10^2 to 10^6 cps).

5. Determine relative magnetic permeability by making laboratory measurements of magnetic susceptibility of rock and soil samples.

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