APPLICATION OF GEOELECTRIC METHODS FOR DEMARCATING THE CLAY BED IN AND AROUND VILLAGE SIVNI DISTT. JABALPUR (M.P.)

A DISSERTATION
Submitted in Partial fulfilment of the requirements
for the award of
M. Tech. DEGREE IN APPLIED GEOPHYSICS

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CERTIFICATE

Certified that the dissertation entitled "APPLICATION OF GEOELECTRIC METHODS FOR DEMARCATING THE CLAY BED IN AND AROUND VILLAGE SIWNI, DISTT. JABALPUR (M.P.)" being submitted by Ramesh Chand in partial fulfilment for the award of the degree of M.Tech. in APPLIED GEOPHYSICS of the University of Roorkee is a record of the student's own work carried out by him under my supervision and guidance. The matter embodied in this dissertation has not been submitted for the award of any other degree or diploma.

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ROORKEE
DATED JUNE 3, 1974

(O. P. VERMA)

ACKNOWLEDGEMENTS

The author expresses profound gratitude to DR. O.P. VERMA for his invaluable guidance; constant help and Encouragement at various stages.

The author thanks DR. R.S. MITHAL, Professor and Head, Department of Geology and Geophysics, University of Roorkee for the facilities provided. He is thankful to DR. D.B. SIKKA for providing facilities in carrying out field work and for the critical discussion about the Geology of the area.

It is a great pleasure to acknowledge thanks to Professor V.K. GAUR, DR. B.B.S. SINGHAL and DR. H. SINVHAL for giving the fruitful suggestions.

Thanks are also due to his friends Sri Dinesh Chandra, Sri H.K. Mittal, for their help most willingly rendered in different ways.

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CHAPTER I

INTRODUCTION

INTRODUCTION

Geophysical exploration may be considered as an application of principles of geophysics to geological exploration. The first objective of the geophysical exploration is the location of geologic structure, the information regarding the occurance of specific minerals is obtained in an indirect manner. The Geophysicist measures at the earth's surface anomalies in physical forces which must be interpreted in terms of geology. The geophysical methods are applicable only if the detectable differences in physical properties exist. Geophysical methods may be broadly classified under two heads major and minor. There are four major geophysical methods, Gravitational, Magnetic, Seismic and Electrical. In gravitational methods, measurements are made of anomalies in gravity attraction produced by differenco in densiti of formations and structures. In magnetic method measurements are made of anomalies in the earth's magnetic field due to geologic bodies of different degree of magnetism. In either case, the reaction of geological bodies are permanent, spontaneous and unchangeable, the operator can not control the depth precisely. In the other two major methods, energy is applied to the ground for the purpose of producing a measurable reaction of geologic bodies. In the seismic methods, energy is supplied by explosives and travel time of reflected and refracted waves are measured. While in electrical methods ground is energised either inductively by passing a time varying current in loop of wire or conductively passing the current by making galvanic contact with ground. It was formerly believed that geoelectric methods were suitable only for shallow exploration and such methods were therefore primarily used for mining and engineering geophysical problems. Today, however, modern developments and refined techniques of interpretation have considerably increased the depth of investigation. It is now claimed, particularly by Soviet geophysicists, that exploration by geoelectric methods may be carried out with reasonable accuracy down to depths of 8-10 Km.

Geoelectric exploration consists of exceedingly diverse principles and techniques, and utilizes both stationary and variable currents produced either artificially or by natural processes. In one group of electrical methods energy is applied galvanically and the distribution of the potential resulting from ore bodies is measured. In the resistivity method, which is one of the most widely used methods of geoelectric exploration, a direct or very low frequency alternating currents, is introduced into the ground by two electrodes and the potential difference is measured between two points suitably chosen with reference to the current electrodes. The potential difference for unit current sent through the ground is a measure of the electrical resistance of the ground between the points. The resistance is a function of the geometrical configuration of the electrodes and the electrical parameters of the ground. Broadly speaking, we can distinguish two types modus operandi of resistivity measurements. In the first, known as geoelectric profiling or mapping, the electrodes and probes are shifted without changing their relative configurations. This gives us an idea of the surface variation of resistance values within a certain depth. In the second method, known as geoelectric sounding, the positions of the electrodes are changed with reference to a fixed central point called sounding point. In this way, the measured resistance values at the surface reflect the vertical distribution of resistivity values in a geological section.

The two types of electrode configurations which are most frequently used in resistivity sounding are called the Wenner and Schlumberger arrays. The Wenner configuration is used in the United States and Canada, and in other English speaking countries, and the Schlumberger Configuration is almost exclusively used in European countries and in the U.S.S.R.

In the galvenic Electromagnetic methods, the distribution of Electromagnetic field resulting from energized ore bodies is measured, while in inductive Electromagnetic methods, the primary energy is applied inductively to the ground and the distortions of electromagnetic field are determined.

Electrical methods constitute an important tool in exploration of ore bodies and burried structures. It is because of the fact that all the Electrical properties of rocks and minerals vary between widest limits.

If all the electrical properties of rocks, the electrical conductivity is the most important, because it controls the passage of low frequency artificial and natural currents. The conductivity of the rock formations, however, depends mainly upon the porosity, fractional saturation and the conductivity of the fluid therein.

Electrical methods have been employed successfully for the location of native metals, sulphide ore bodies and other metallic bodies such as amunition dumps, reservoir for tapping water and for the determination of depth to bed rock at Engineering sites.

THE PROBLEM :

The problem, taken up and studied by the author was to demarcate the clay bed in and around village Sivni, Distt. Jabalpur (M.P.). The investigations were carried out by Geophysical methods namely the Electrical resistivity method. The method was chosen due to the fact, that clay has good contrast in conductivity in comparision to the overburden.

LOCATION AND ACCESSIBILITY :

The land is located on the east side of village SIVNI-TOLA, Tehsil Jabalpur and bounded by Narmada river to the west and Nala to the eastern side. The village is 12 miles from Jabalpur Railway Station. The area (about 98 acre) belong to Sri Chamru Patel. It can be reached by a side road from National High Way No. 6 near Tilwara Ghat. The nearest railway station is Madan Mahal on broad Gauge. The topography is undulating with maximum relief of about

7 meters from the bottom of Narmada river. Much of the terrain is ancient meander of the Narmada river.

GEOLOGY OF THE AREA IN BRIEF :

The rocks exposed in the area are Deccan trap lavas, Jabalpur (?) Lameta succession. These consist of clays overlain by arkosic to fine grain sand stones. The sand stone is 1 to 3 meter thick. The thickness of the clay bed exposed in Nala is about 3 meters and of varrying colour. The lavas are exposed in the Western part of the area investigated. The lavas cut across the clays and underlie the sandstones. Precambrian rocks are exposed in the Narmada river. The sand stones and clays are horizontal to about 2^{o} dipping to the west. The overburden which is black coloured soil, varies from 0-6 meter in thickness and at places kanker is found.

GEOPHYSICAL INVESTIGATION :

Following geoelectric methods were used for demarcating the clay beds in the region:

(i) Resistivity surveys

(ii) Mise-à-la masse method.

Resistivity soundings with Schlumberge configuration were taken at 50 stations, though soundings with Wenner array were taken on 15 stations to check thickness and resistivity of layers. The results obtained with two configurations were quite similar, therefore, only the Schlumberger array was used as the ~~resistivity meter~~ at other stations. The resistivity meter was quite sensitive, thus, increasing the reliability of the field observations and the results.

The area was divided in a grid pattern to carry out soundings. The map of the area showing the location of various sounding points is attached, Fig. 11. The spacing between two sounding stations vary from 30 meters to 110 meters and the current electrode spread from 120 m to 240 meters according to the topography. To facilitate interpretation of the soundings, insitu resistivity of the formations were measured at different places with the help of megger using Wenner configuration and small electrode separation (a= 2.5 cm). Subsequent to resistivity survey "Mise-A-la-Masse" method was used in order to confirm the extension of the clays.

The observations, thus collected by using the Mise-A-la-Masse method, were plotted and the regular pattern of contours was found showing the presence of clay in the area. Readings obtained by using Schlumberger method were plotted on Log-Log graph paper of 62.5 mm modulus. Most of the curves obtained were showing presence of three layers. So these field curves were then matched with theoretical curves for Schlumberger method by using partial and full curve matching techniques. Thus the depth to interfaces and the resistivities of the formations were determined as mentioned in the table [illegible] 1

Since the topography was undulated and the topographical map was not available, therefore, elevations of the sounding points were measured using brunton campass, ~~[illegible]~~ but the inherent limitations of method still exist. The geoelectric sections are attached herewith showing the thicknesses and depths of clay bed. At places electrical soundings indicate unduly thick bed of low resistivity which is likely due to the presence of an incient burried nala containing ground water.

SCHEME OF THE THESIS :

The main text is divided into 6 chapters, which are briefly described below :

Chapter 2 deals with the electrical properties of earth materials, in which the effect of moisture, dissolved salt, temperature and pressure etc, is studied on resistivity.

In Chapter 3 theoretical foundations of geoelectric sounding is discussed in order to understand the basic principles and interpretation techniques. Instrumentation and field operation is described in Chapter 4, while Chapter 5 processing of the data is discussed and the results presented. The work is summarised and concluded in Chapter 6.

CHAPTER II

ELECTRICAL PROPERTIES

OF

EARTH MATERIALS

Electrical Properties of Earth Materials

The flow of current through the earth materials i.e., rocks and soils, are controlled by their electrical properties. The electrical properties of the earth materials can be defined by the three parameters:

(1) Electrical conductivity

(2) dielectric constant, and

(3) The magnetic permeability

However the conductivity is the most important property in determining the flow of current at very low frequencies.

The abilities of materials to conduct electricity when a voltage is applied are expressed as 'conductivities'. It is the reciprocal of resistivity which is defined as the resistance offered between the opposite faces of a unit uniform cube. The unit to express resistivity is thus ohm-meter in M.K.S. system. While the unit of conductivity is mhos per met er.

The electric conduction in minerals may take pl ace by electronic or ionic processes. Rocks are not formed of minerals alone and the electrical properties of a rock are not necessarily determined by the properties of mineral constituents alone. All rocks at the earth surface are porous. Under reasonable circumstances these pores are partly or completely filled with water. This water usually carried some salt in solution so that the water content of a rock has a far greater capacity for carrying current than does the solid matrix of the rock, unless highly conducting minerals are present.

Mechanism of the flow of current involve conduction through the solid matrix or grains and through the liquid electrolyte. However solid conduction mechanism can be expected to be important in comparison with electrolyte conduction through pore water in three cases -

1. In a rock containing a high percentage of conducting minerals.
2. In which fluid present in the pore spaces are completely frozen.
3. In a rock which is far enough below the surface of the earth where all pore spaces can be considered closed by overburden pressure.

However most rock-forming minerals can be considered solid electrolytes. Electrolytic conduction can thus take place in ionic bonded crystals. But the rocks and other earth materials near the surface possess a reasonable good porosity, therefore, the conduction will be electrolytic. Near the earth surface the conducting medium being an aqueous solution of common salts distributed in a complicated manner through the pore structure. The resistivity of a water bearing rock will depend on the amount of water, the way in which the water is distributed in the rock.

EFFECT OF MOISTURE CONTENT

As the electrical conduction through soil and rocks is mainly electrolytic, it follows that the quantity of water and nature and amount of salt are the important factors in determining the resistivity. The actual amount of water present in soils is variable, It varies with weather, time of year and the nature of subsurface soil and depth of the permanent water table. The variation of soil resistivity, with the moisture content is shown in fig.2. As the moisture contents increases with depth, the resistivity first decreases rapidly but after some value (14 to 18 percent) the moisture contents, the rate of decrease becomes much less.

EFFECT OF THE DISSOLVED SALT

The conduction in most of the near surface rocks will be electrolytic. The medium of conduction is the solution of salts, distributed through the pore spaces of the rock.

When a salt goes into solution of water, the constituent ions in the solid salt separate and these ions are free to move in the solution. When an electrical field is applied across an electrolytic solution, Cations will go towards the negative pole and the anions to the positive pole. As the ions are accelerated, a drag force starts acting on them, and as a result of which their velocity is decreased.

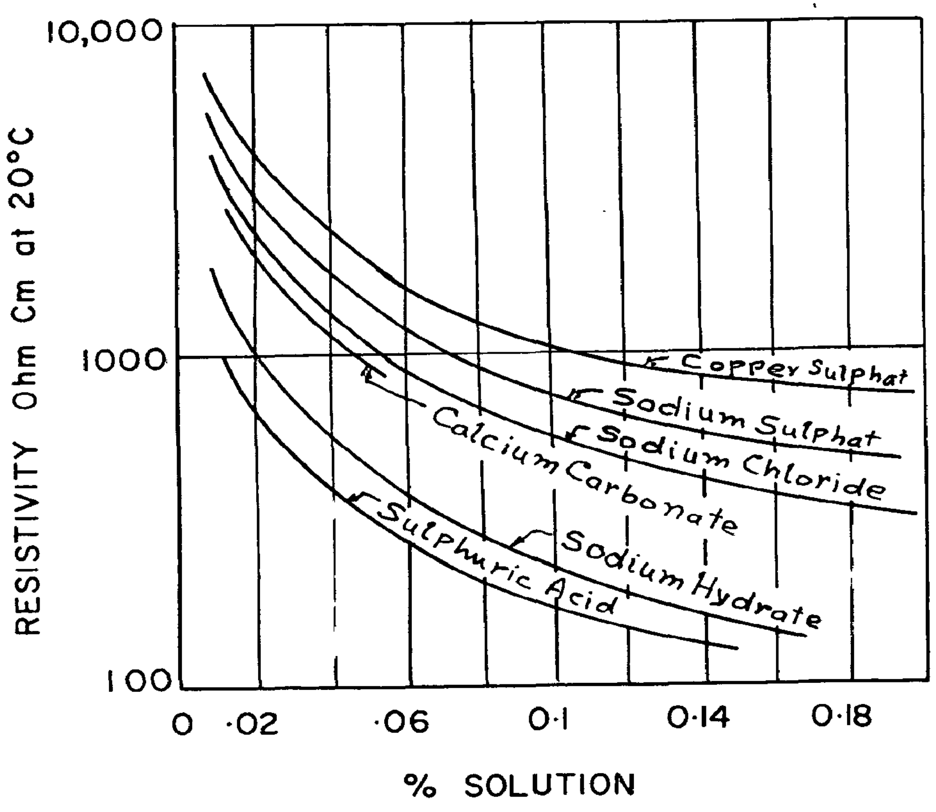


FIG. I TYPICAL RESISTIVITY CURVES OF SOLUTION

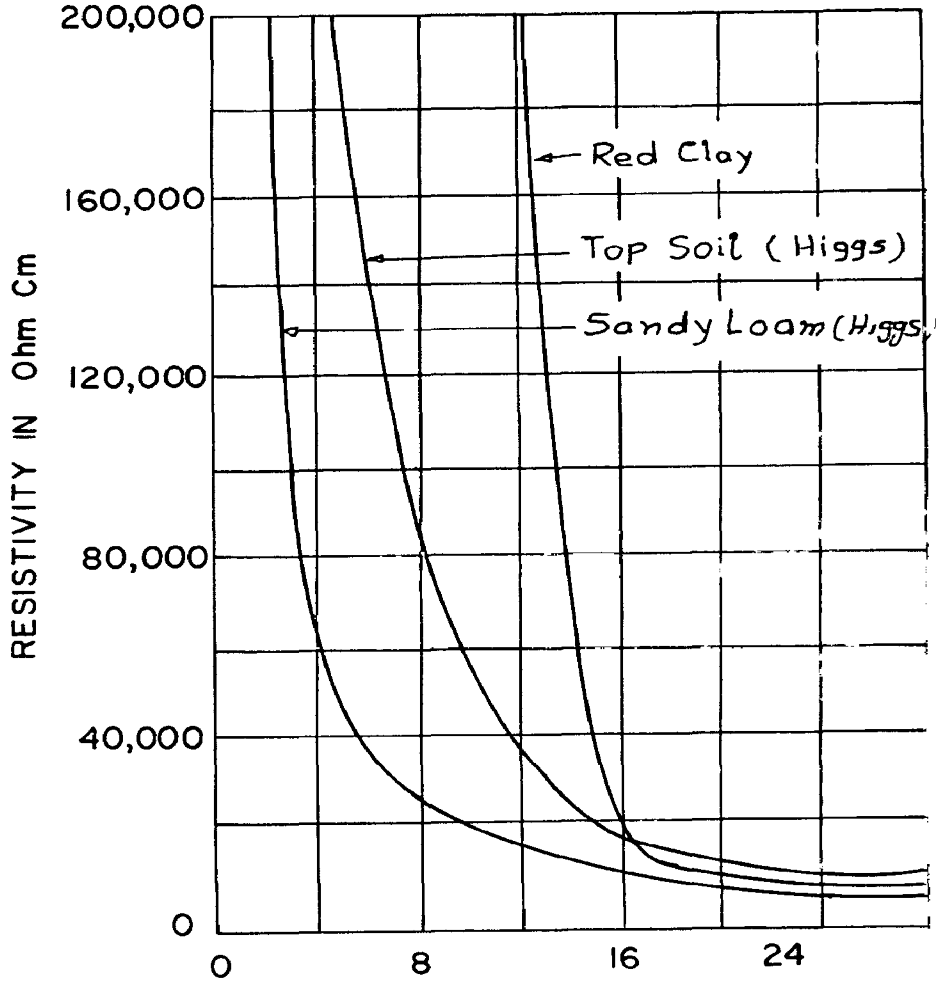


FIG. 2 MOISTURE AS PECENTAGE OF DRY SOIL (By weight)

VARIATION IN SOIL RESISTIVITY WITH MOISTURE CONTENT.

This velocity with which it moves is known as mobility of ions and is measured in meters/sec., when a voltage gradient of one volt/meter is applied. If a solution contains a high concentration of ions, the motion of one ion will be influenced by the motion of ions close to it, and thus reducing their mobility. The mobilities of a few of the commoner ions at 25°C in dilute solutions are:

H	36.2	x	10^8	m^2/Sec V
OH	20.5	x	10^8	
SO_4	8.3	x	10^8	
Na	5.2	x	10^8	
Cl	7.9	x	10^8	
K	7.6	x	10^8	
No_3	7.4	x	10^8	
Li	4.0	x	10^8	
HCo_3	4.6	x	10^8	

(after Keller, G.V.)

When an electrical field is applied to an electrolyte the amount of current which flows is found by multiplying the number of ions present by the velocity with which they move. The current flowing through an electrolyte per one volt per meter applied field is:

$$I = AF \; (V_1C_1 + V_2C_2 + V_3C_3 \; .... + V_nC_n)$$

where V_n and C_n represent the mobility and concentration of ions, A cross sectional area through which the current flows, F the Faraday number.

Thus the resistivity of water is governed by the amount of salts dissolved, it, therefore, follows that resistivity of soil which itself depends upon the water contents; it must be indirectly dependent on the salt dissolved content of water. Even a small amount of salt dissolved in water can reduce the resistivity considezably. As the different salts have different effect on the (Fig No1) resistivity and this is perhaps the reason for the resistivity variation of some type of soil, from one locality to another.

EFFECT OF GRAIN SIZE AND ITS DISTRIBUTION

The effect of the grain size on the resistivity lies in the fact that how the grain size and its distribution effects the manner in which the moisture is held by soils and rocks. The moisture is probably hold by surface tension at the points of contact of grains. Also if grains of various size are present, the space between the large grain may be filled with smaller grains thus reducing the porosity and hence the resistivity will be reduced. The free space between the grains depends upon the arrangement of the grains.

The quantity of exchangeable ions attached to clays is usually expressed in terms of the weight of ions in milliequivalents absorbed per 100 gm of clay. The exchange capacities of some common clays are:

Kaolinite	3 to 15 in equiv/100 gm
Halloysite 2 H_2o	5 to 10

Halloysite $4H_2O$	40 to 50
Montmorrillonite	80 to 150
Illite	10 to 40
Vermiculite	100 to 150
Chlorite	10 to 40
AHapulgite	20 to 30

RELATIONS BETWEEN RESISTIVITY, POROSITY AND TEXTURE

The resistivity of a water-bearing rock decreases with increasing water content. In fully saturated rocks, water content may be equated with porosity. The texture of a rock has some effect on the resistivity . Rocks may be grouped into three general categories on the basis of their pore geometries. In case of sedimentary rocks, porosity is intergranular in nature, consisting of the space left over after the rock grains were compacted. In Igneous rocks porosity occurs in the form of joints. A third form of porosity common in limestones and in some volcanic rocks is Vugular porosity, consisting of large, irregular cavities formed either by solution or by large gas bubbles. Pore spaces must be interconnected and filled with water in order that a rock may conduct electricity. In all three types of porosity, the pore volume may consist of two parts - the larger voids which are called storage pores and the finer connecting pores. Most of the resistance to current flow is met in the connecting pores. All the three types of porosity are usually present in any rock in varying proportions. Sedimentary sandstone and shales

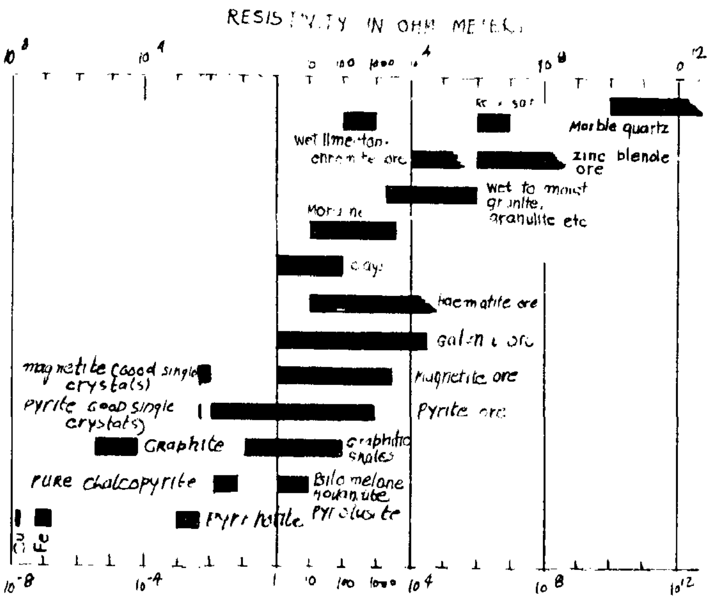


FIG. 3 APPROXIMATE RESISTIVITY RANGES OF SOME ROCKS, MINERALS, ORES AND COMMON METALS. THE RESISTIVITY SCALE (HORIZONTAL) IS LOGARITHMIC. (After Parasnis)

may have 5 to 60% intergranular porosity. Archie's L aw relates resistivity and porosity as:

$$\rho_{rock} = a \rho_w \phi^{-m}$$

where ρ_{rock} is the bulk resistivity of the rock

ρ_w is the resistivity of the water contained in the pore structure

ϕ is the fractional porosity

a and m are emprically determined parameters. Approximate values for a and m are 1 and 2 respectively.

Archie's Law indicates that the ratio of Bulk-resistivity to water resistivity should be a constant for a given porosity, that it should not depend upon the resistivity of the water in the rocks. This ratio is called the formation factor. This ratio is less when a rock is saturated with dilute solution than when the rock is saturated with a highly saline water. This may be explained by considering that the conductivity of the water distributed through the pore space is usually increased by two phenomenon -

1. Ionization of clay minerals.
2. Surface conductance.

Clay minerals such as kaolinite, halloysite, montmorillonite, vermiculite, illite, chlorite and others have the property of adsorbing certain anions and cations and retaining these in an exchangeable state. The common exchangeable ions adsorbed on clay are Ca, Mg, H, K, Na and NH_3, in order of decreasing abundance.

Archie's equation indicates that bulk resistivity of a rock is proportional to the resistivity of water, contained in it. The ratio of rock resistivity to water resistivity is termed as the formation factor which is a measure of porosity

$$F = \frac{\rho_{rock}}{\rho_w} = \bar{a}s^n \, \phi^m$$

<u>Resistivities determined in field</u>

Resistivity of water of Narmada river	= 1.5 ohm-meter
" " " Nala	= 1.04 ohm-meter
" " " Well	= 0.755 ohm-meter
Resistivity of silica sand	= 12.56 ohm-meter
" " sand mixed clay	= 3.4 ohm-meter
" " clay (In pit)	=1.5 to 2.14 ohm-m

EFFECT OF TEMPERATURE AND PRESSURE

If the temperature is high enough to derive water from the rock as steam or low enough to freeze the water in the pores of a rock, the resistivity will be affected very much. Thus large changes in the temperature effect the resistivity of a water bearing rock markedly. But at moderate temperatures, a change in temperature changes the conductivity of rocks only when the conductivity of the electrolyte changes. When there is an increase in temperature the viscosity of water is decreased and the mobility of the ions increases and hence the conductivity also increases. The dependence of resistivity in (FIG. 5)

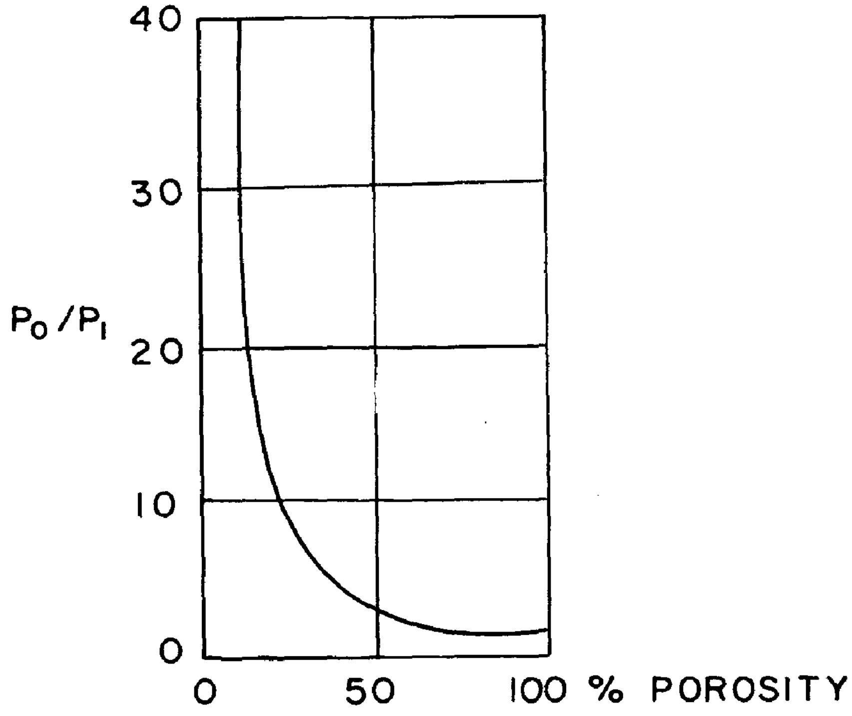


FIG. 4 RELATION BETWEEN RESISTIVITY RATIO P_0/P_1 AND POROSITY (After Sundberg)

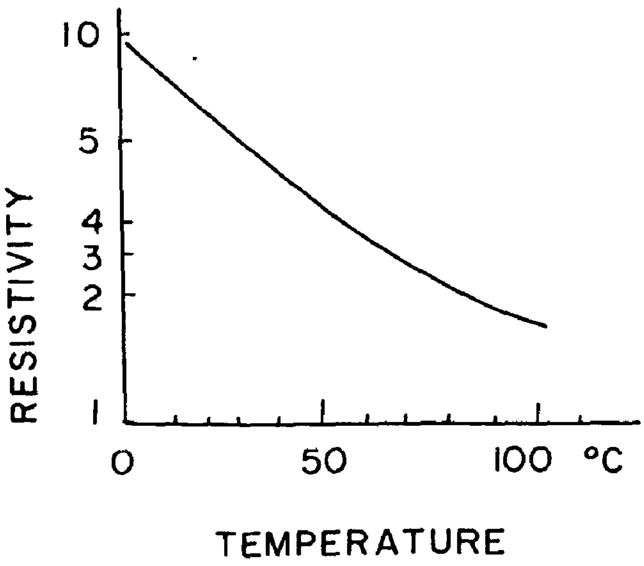


FIG. 5 DECREASE OF RESISTIVITY AT Nacl SOLUTION WITH TEMPERATURE.

given by the relation -

$$\text{Resistivity}_{(t)} = \frac{\text{Resistivity } (18^\circ C)}{1 + \alpha_t\ (t-18)}$$

Where α is the temperature coefficient of resistivity which has a value of about 0.025 per degree centigrade for most electrolytes

Keeping the gradual rise of temperature with depth in mind, then the temperature at a depth of 8000 ft in sedimentary rocks will be about 40° C higher than the temperature at the surface, and it would mean that the rocks at 8000 ft have a resistivity only half of what would have been at the surface for the same rock.

As resistivity is inversely proportional to temperature, that is why freezing increases the resistivity.

Resistivity of rocks decreases with increasing pressure, which is caused due to compactness of material lying above the layer.

EFFECT OF ANISOTROPY

By the word anisotropy of rocks we mean that their properties are different in different directions. This type of character is generally found to be associated with some rock formations and it is natural to expect it to occur in stratified formation. This anisotropy plays an important role in all stratified formations and under such conditions the resistivity along the bedding plane is quite different from those at right angles to it the anisotropic coefficient " λ " of a formation is defined as the ratio of the apparent parallel resistivity to apparent normal resistivity. The value of anisotropy is found vary from one to ten. In homogenous isotropic conditions it is unity.

Such conditions of anisotropy met in the field make the interpretation difficult.

TO DEFINE THE AVERAGE DENSITY OF AN ASSEMBLAGE OF RELATIVELY THIN BEDS

The large number of individual layers apparent or an electric log can not be resolved with any of the surface based electrical prospecting methods. It is necessary to device a realistic way of defining the average resistivity of an assemblage of thin beds.

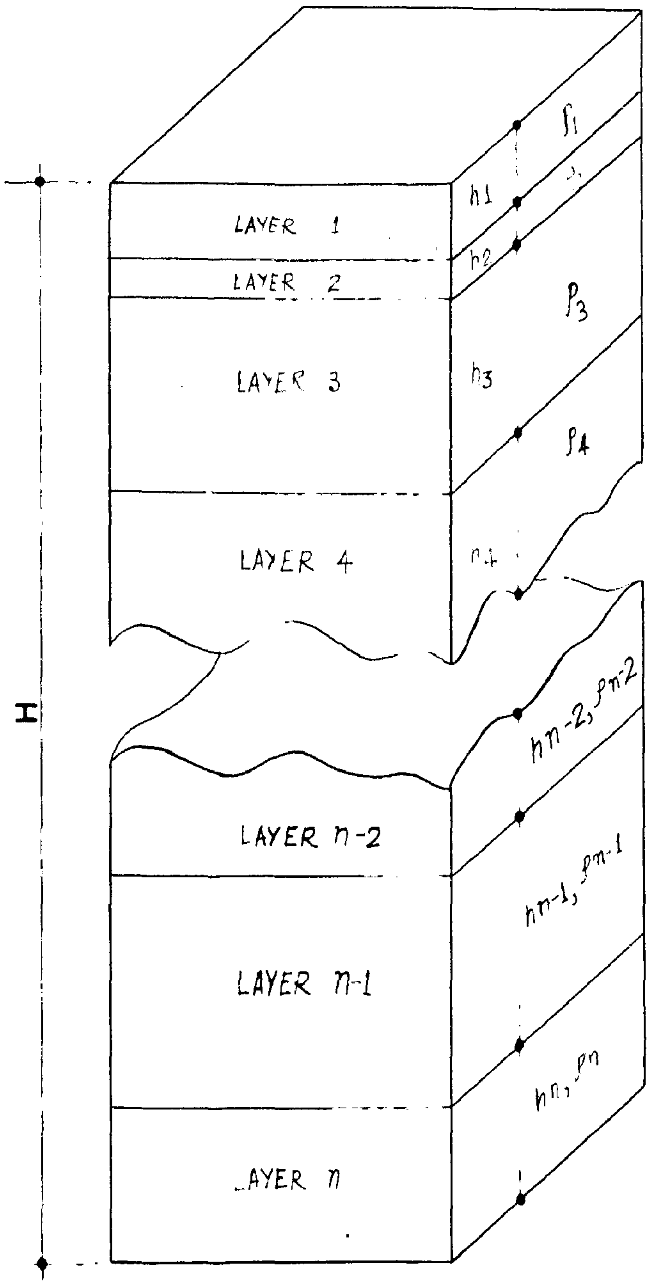
It has been shown (Schlumberger and other, 1934) that the average electrical properties for a finely layered sequence may be described with a cost of five parameters, these parameters being defined in terms of a column of rock one meter square cut from the sequence of layers. The column consists of n horizontal beds, each with its own characteristic resistivity, and thickness ti as shown in figure No 6.

COLUMN FROM A LAYERED SEQUENCE OF ROCKS:

These parameters are defined by considering the resistance to current flowing either vertically or horizontally through the column.

For current flowing vertically through the length

FIGURE 6



COLUMN FROM A LAYERED SEQUENCE OF ROCKS, USED IN DEFINING AVERAGE LONGITUDINAL RESISTIVITY, AVERAGE TRANSVERSE RESISTIVITY AND ANISOTROPY ARISING FROM LAYERING.

of the column, the Transverse-Resistance 'T', is the sum of the resistance met in each of the individual layers.

$$\text{Thus } T = \sum_{i=1}^{n} \rho_i \cdot t_i$$

If the bed resistivities 'ρ_i' and thickness 't_i' are given in ohm meters and meters respectively, the transverse resistance is expressed in ohms.

If the column is macroscopically uniform. Then average transverse resistivity 'ρ_{tr}' will be given by

$$\rho_{tr} = \frac{T}{H} = \frac{\Sigma \rho_i t_i}{\Sigma t_i}$$

where H brings the total thickness of the assemblage of fine layers. For current flowing laterally through the column, the longitudinal resistance, is that of each of the layers considered to be connected in parallel. If S is the conductance (reciprocal of the resistance)

$$S = \sum_{i=1}^{n} \frac{t_i}{\rho_i} \qquad \text{(S = Longitudinal conductance)}$$

If the column is macroscopically uniform, the average longtitudinal resistivity: ρ_e is given by

$$\rho_e = \frac{H}{S} = \frac{\Sigma t_i}{\Sigma t_i/\rho_i}$$

Unless the resistivities of the individual layers are exactly same, the ρ_e is smaller than ρ_{tr}. This dependance of resistivity on the direction of current flow constitutes anisotropy. The coefficient of anisotropy for a layered sequence of rocks is defined as

$$\lambda = \sqrt{\frac{\rho_{tr}}{\rho_{e}}} = \sqrt{\frac{T/H}{H/S}} = \sqrt{\frac{ST}{H^2}}$$

Thus an asemblage of thin layers each of which is isotropic, will appear to be anisotrppic when considered as a macroscopically uniform medium. This type of anisotropy might be termed as macro-anisotropy.

Here individual layer might be anisotropic, if the direction of maximum conduction is parallel to the bedding planes, then the general anisotropy will be the product of the mac ro-anisotropy and the average micro-anisotropy.

CHAPTER - III

THEORETICAL FOUNDATION

OF

ELECTRICAL RESISTIVITY METHOD

THEORETICAL FOUNDATIONS OF ELECTRICAL RESISTIVITY METHOD

Theory of Current Flow:

The ground is regarded as consisting of regions of approximately constant resistivity separated from others of differing resistivity by plane interfaces. The interfaces correspond generally to boundaries between horizontal layers of different electrical properties or to faults and vertical contacts.

For better interpretation it is necessary to understand the behaviour of current flow in layered media and the distribution of potentials. In dealing with this the starting point is the Ohm's law :

$$\frac{V}{I} = R \qquad \text{.....(1)}$$

Where

I = Current flowing in the conductor

V = Potential difference between two surfaces of the conductors.

R = Resistance of the conductor

If a conductor carries a current with parallel lines of flow over a cross sectional area A, then its resistivity is defined as

$$\rho = \frac{R \times A}{L} \qquad \text{.....(2)}$$

Where R is the resistance measured between two equipotential surfaces separated by a distance L .

For examples (1) and (2) we obtain :

$$\rho = \frac{A}{L} \cdot \frac{\nabla V}{I}$$

or $$\rho \frac{I}{A} = \frac{\nabla V}{L}$$

$$\rho J = \frac{\nabla V}{L}$$

$$J = \frac{1}{\rho} \cdot \frac{\nabla V}{L} \qquad \ldots\ldots(3)$$

where J is the current density defined as the flow of the current per unit area.

If the lines of current flow are not parallel so that the current density varies over the conductor, the ratio $\nabla V/L$ becomes in the limit the maximum potential gradient $\frac{dV}{dL}$ and the expression of current density changes to

$$J = \frac{1}{\rho} \cdot \frac{\partial V}{\partial L}$$

The component of current density in a direction 'r' is

$$J_r = -\frac{1}{\rho} \cdot \frac{\partial V}{\partial r}$$

The negative sign has been introduced here to express the fact that the potential increases in the opposite direction to the current flow. Here potential gradient in the direction of r is used instead of maximum gradient.

Theory of Current Flow In Homogeneous Earth :

The flow of current in a medium is based on the principle of conservation of charge and is expressed by the relation :

$$\text{div}\, \bar{J} = -\partial q/\partial t$$

Where $\bar{J}$ = current density (A/m^2) and
q = The charge density (c/m^3)

This is equation of continuity, for stationary current it reduces to:

$$\text{div}\, \bar{J} = 0 \quad \text{i.e.}\ \nabla \cdot \bar{J} = 0$$

But Ohm's law states that $\bar{E} = \bar{J}$

Where E is field intensity (V/m)

or $\bar{J} = \sigma \bar{E}$

$\bar{J} = -\sigma\, \text{grad}\ V$

For an isotropic medium, ρ is a scaler function of the point of observation and J is in the same direction $\bar{E}$

$\therefore$ $\text{div}\, \bar{J} = \text{div}\,(-\frac{1}{\rho}\ \text{grad}\ V) = 0$

or $\text{div}\,(\frac{1}{\rho})\ \text{grad}\ V + \frac{1}{\rho}\ \text{div grad}\ V = 0$

8 This is a fundamental equation of electrical prospecting with direct current. If the medium is homogeneous, ρ is independent of the co-ordinate

Therefore we get

$\text{div Grad}\ V = 0$

or $\nabla^2 V = 0$

Thus, the electric potential distribution for current flow is a homogeneous isotropic medium satisfies Laplace's equation and potential is in this case a harmonic function.

Consider a point source of current of strength 'I' embeded in a homogeneous medium of infinite extent and of uniform resistivity. The equipotential surfaces in this case will be spherical, and the current lines radial. Considering two such equipotential surfaces at distance "dr" apart and a distance "r" from the point source of current. The potential difference across these two surfaces is given by

$$dv = - \frac{I \rho \, dr}{4 \pi r^2}$$

The potential at a distance r from the point source is given by

$$V = - \int_{\infty}^{r} \frac{I \rho \, dr}{4 \pi r^2}$$

$$= \int_{r}^{\infty} \frac{\rho I \, dr}{4 \pi r^2}$$

$$= \frac{\rho I}{4 \pi r}$$

At the surface of a semi infinite earth this expression for the potential at a distance r will becomes:

$$V = \frac{\rho I}{2 r \pi}$$

Potentials about a Point Electrode at the Surface of a Layered Ground:

Now let us introduce some heterogeneity into the ground by bringing in a number of horizontal layers, each having different but homogeneous and isotropic electrical properties. Because of the presence of these layers, the potential will no longer have spherical symmetry about P, and we must now look for solutions of $\nabla^2 V = 0$ which can be made to satisfy the continuity conditions at a number of horizontal boundaries. Such solutions are found by separating Laplace's equation in the cylindrical coordinate system whose origin is at P and whose z axis is vertical and positive in the downward sense. Since symmetry with respect to the coordinate θ still exists, we may write :

$$\nabla^2 V = \frac{\partial^2 V}{\partial r^2} + \frac{1}{r}\frac{\partial V}{\partial r} + \frac{\partial^2 V}{\partial z^2} = 0$$

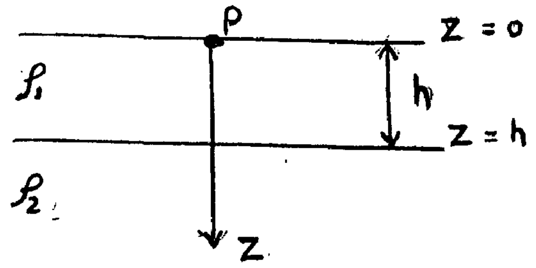


Fig.: 7
A point source of current at the surface of a uniform layer

The complementary solution of this equation is formed from the characteristic functions of the separated variables. By choosing only those which are well behaved when $r \to 0$, we obtain for the complementary function :

$$V(r,z) = \int_0^\infty \left[A(\lambda)e^{-z\lambda} + B(\lambda)e^{\lambda z}\right] J_0(\lambda r)\, d\lambda$$

to which we must add the particular solution $V(r,z) = I\rho_1/2\pi R$ which applies within the close vicinity of P. If no boundaries exist other than the surface $z = 0$, then $A(\lambda) = B(\lambda) = 0$. If there are boundaries, then the coefficients $A(\lambda)$ and $B(\lambda)$ will be determined from the boundary conditions.

To illustrate, let us assume a single homogeneous layer whose resistivity is ρ_1 and whose thickness is h lying on top of a uniform half-space whose resistivity is ρ_2 (Fig. No.7). For the potential in the upper medium we may write :

$$V_1(r,z) = \frac{I\rho_1}{2\pi R} + \int_0^\infty \left(A(\lambda)\, e^{-\lambda z} + B(\lambda)\, e^{z\lambda} \right) J_0(\lambda r)\, d\lambda \quad 0 \leqslant z \leqslant h$$

In a substatum there is no external source of current, and therefore the particular integral is not required. Moreover, we reject terms involving $e^{\lambda z}$, since the potential must remain finite when $z \to \infty$. Consequently, we may write for the potential in $z \geqslant h$

$$V_2(r,z) = \int_0^\infty C(\lambda) e^{-\lambda z} J_0(\lambda r)\, d\lambda \qquad z \geqslant h$$

The boundary conditions which these solutions must satisfy are the following :

$$\frac{\partial V_1}{\partial z} = 0, \quad z = 0$$

$$V_1 = V_2, \quad z = h$$

$$\frac{1}{\rho_1}\frac{\partial V_1}{\partial z} = \frac{1}{\rho_2}\frac{\partial V_2}{\partial z}, \quad z = h$$

Thus three equations are available to determine the three unknown functions A, B and C. To apply these conditions, we make use of the Lipschitz integral identity (Watson (a), Chapter 13).

$$\frac{1}{R} = \int_0^\infty e^{-\lambda z} J_0(\lambda r)\, d\lambda$$

Then, on substituting for V_1 and V_2 into these expressions and absorbing $2\pi/I\rho_1$ into the constants, we arrive at the following equations :

$$A - B = 0$$

$$(1 + A)\, e^{-\lambda h} + B\, e^{\lambda h} = C\, e^{-\lambda h}$$

and

$$-\frac{1}{\rho_1}\left[(1 + A)\, e^{-\lambda h} - B\, e^{\lambda h}\right] = -\frac{1}{\rho_2}\, C\, e^{-\lambda h}$$

Solving, we get

$$A = B = \frac{I \rho_1 k}{2\pi(e^{2\lambda h} - k)}, \quad k = \frac{\rho_2 - \rho_1}{\rho_2 + \rho_1}$$

and therefore, on $z = 0$

$$V_1(r) = \frac{I \rho_1}{2 r \pi}\, G(r; k)$$

where $G(r,k) = 1 + 2kr \int_0^\infty (e^{2\lambda h} - k)^{-1} J_0(\lambda r)\, d\lambda$

This solution was derived originally by Stefanescu. Normally we take the layer thickness as the unit of length by setting $h = 1$, and the function $G(r;k)$ may then be evaluated numerically for given values of k. It is clear that the analysis can be extended to any number of layers, and in fact extensive tables for evaluating potentials over one, two and three layers which rest conformably upon a uniform substratum are available.

Concept of Apparent Resistivity :

The expression for resistivity is true under homogeneous conditions of the ground. In actual practice, however, sub-surface conditions are not homogeneous and if the same electrode arrangement is now used to take measurements over such a medium the observed potential difference will be different from the measured and homogeneous medium and the resistivity value determined shall vary with different electrode arrangements. Then we get a fictitious value of resistivity called as "apparent resistivity" designated by ρ_a. For this non homogeneous case, the apparent resistivity may be defined as the resistivity of an equivalent homogeneous medium in which the current "I" would produce the potential drop of the same value "ΔV". The importance of this concept lies in the fact that it provides a measure of deviation from homogeneous and isotropic surface.

For four electrode systems, let current enters the earth through A and leaves through electrode B, the measuring electrodes are M and N, Then potential at M is given by

$$V_M = \frac{I\rho}{2\pi}\left(\frac{1}{AM} - \frac{1}{BM}\right)$$

Similarly, the potential at the point N will be given by

$$V_N = \frac{I\rho}{2\pi}\left(\frac{1}{AN} - \frac{1}{BN}\right)$$

Now, the potential differences between the electrodes M and N is :

$$\Delta V = V_M - V_N$$

$$= \frac{I\rho}{2\pi}\left(\frac{1}{AM} - \frac{1}{BM}\right) - \frac{I\rho}{2\pi}\left(\frac{1}{AN} - \frac{1}{BN}\right)$$

$$\Delta V = \frac{I\rho}{2\pi}\left(\frac{1}{AM} - \frac{1}{BM} - \frac{1}{AN} + \frac{1}{BN}\right)$$

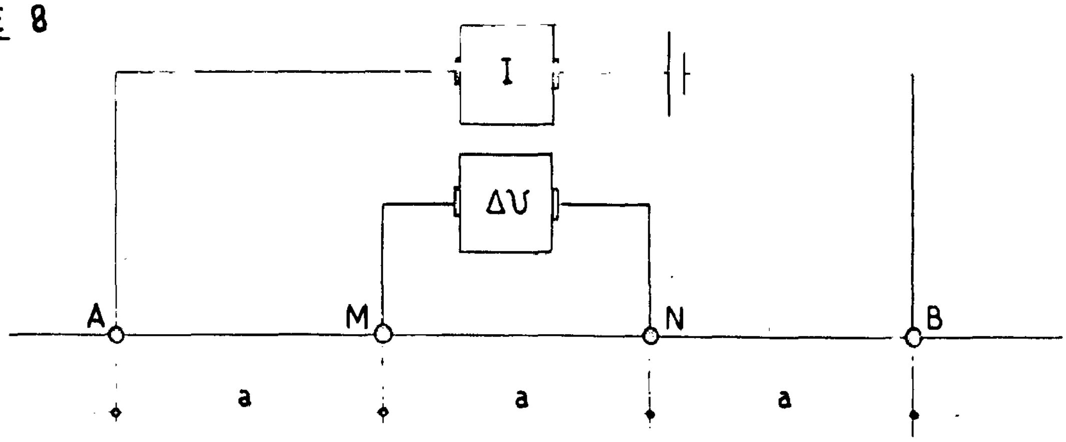
$$\therefore \rho = 2\pi\frac{\Delta V}{I} \frac{1}{(1/AM - 1/BM - 1/AN + 1/BN)}$$

This is the basic formula for calculating the apparent resistivity for any electrode configuration.

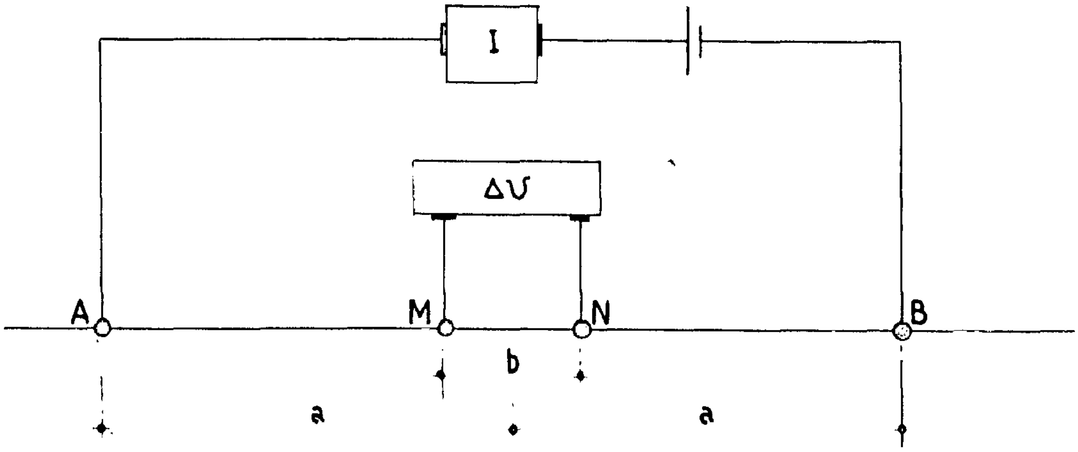
Different Configurations and Derivation of Formulae :

By convention, electrical sounding methods have been classified as "Dc" methods or "AC" methods. The DC methods for measuring earth resistivity have been used most widely. Generally, four terminal arrays are used in order to minimize the effect of material near the current electrodes. Current is driven through one pair of electrodes, the potential established in the earth by this current is measured with the second pair of electrodes. Strictly direct current is

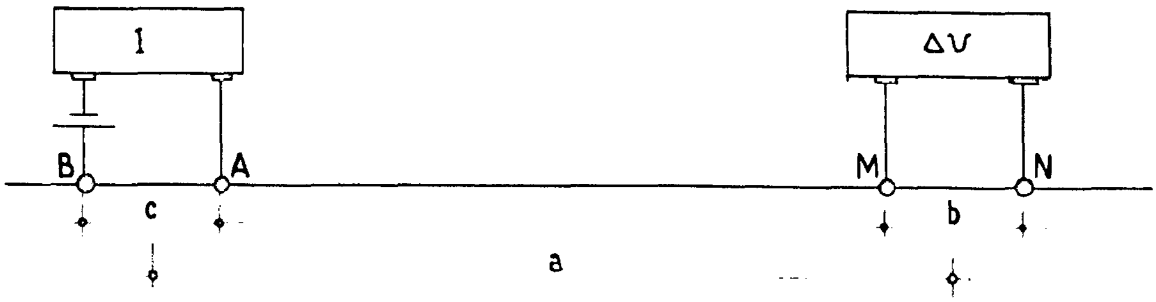
FIGURE 8



WENNER ARRAY



SCHLUMBERGER ARRAY



POLAR DIPOLE ARRAY

ELECTRODE ARRAYS COMMONLY USED IN THE DIRECT CURRENT RESISTIVITY METHOD. A AND B ARE CURRENT ELECTRODES. M AND N ARE MEASURING ELECTRODES AND a, b and c ARE ARRAY SPACING FACTORS.

not used, but rather a low frequency alternating current is used so that the voltages developed in the earth by this current can be easily recognized in the presence of the other, miscellaneous voltages (or self potentials) which arise at electrode contacts. However, the frequency of the current is made sufficiently low that the assumption may be made that the flow of the current in the earth can be completely described by a solution to Laplace's equation.

The electrode Arrangements may be of Three types : (FIG. No. 8)

1. Arrangements in which the potential difference between two widely spaced measuring electrodes is said to be measured. An example is the wenner-array, in which four electrodes are equally spaced along a straight line.

2. Arrangements in which the gradient of potential (electric field intensity) is said to be measured, using a closely spaced pair of measuring electrodes. An example is the Schlumberger Array, in which two closely spaced measuring electrodes are placed mid way between two current electrodes.

The measuring electrodes are placed closely enough together that the ratio of voltage observed between them to their separation approximately equals the potential gradient at the mid point of the current spread.

3. Arrangements in which a second spatial derivative of the potential is said to be measured using a closely spaced current electrode pair. An example is the polar dipole array. The voltage measured this way is the approximately equal to the second derivate of the potential, after it has been divided by the distances AB and MN, provided these distances are small compared to the separation between dipole centres.

The Wenner array, in which a potential difference is m measured, is one of the most commonly used electrode arrays for determing resistivity. In the Wenner array, four electrodes are equally spaced along a straight line, as shown in Fig The distance between any two adjacent electrodes is called the array spacing, a. The geometric factor for the Wenner array is:

$$K = \frac{2\pi}{\frac{1}{a} - \frac{1}{2a} - \frac{1}{2a} + \frac{1}{a}} = 2\pi a$$

The Lee modification of the Wenner array uses a third measuring electrode at the midpoint, O, of the ordinary Wenner array. A potential difference is then measured between both M and N and the centre electrode O, and apparent resistivities calculated for each half of the array. The advantages of such a system is said to be that horizontal changes in resistivity may be recognized by comparing the apparent resistivities measured with each half of the array. The geometric factor for one half of the Lee array is:

$$K = \frac{2\pi}{\frac{1}{AM} - \frac{1}{BM} - \frac{1}{AO} - \frac{1}{BO}} = \frac{2\pi}{\frac{1}{a} - \frac{1}{2a} - \frac{1}{2a} - \frac{1}{2a}} = 4\pi a$$

The Schlumberger array, which also is widely used in measuring earth resistivities, is designed to measure approximately the potential gradient.

~~We are now in a position to answer some of these questions but not all.~~

DIRECT CURRENT RESISTIVITY SURVEYS:

Comparison of Arrays:

Many techniques have been used in measuring earth resistivity by giving various relative positions for the electrodes. Since high power and accordingly heavy equipment is required in making deep resistivity surveys, both efficiency and safety are factors to be considered more seriously. The lowest signal which can be detected is limited either by the sensitivity of the equipment or by the level of telluric noise between the measuring electrodes. Let the threshold voltage which can be measured in the absence of noise is Vc, and let the average telluric noise delivered from the measuring electrodes is VT. The telluric noise level varies widely with time, location and local resistivity, these factors are common for all electrode arrays. However the telluric noise will also be proportional to the separation between measuring electrodes.

Let primary telluric field strength is E_T, such that

$$V_T = \overline{MN} \quad E_T$$

Where $\overline{MN}$ is the spacing between the measuring electrodes. BERDICHEVSKIY (1965) has shown that the electric field is proportional to the square root of resistivity.

$$E_T = K_T \ \rho^{1/2}$$

Let Vs is the signal voltage developed between the measuring electrodes by the current supplied with the current electrodes. We can compute this voltage for various arrays for a uniform earth.

1 For the Wenner Array

$$V_{s,\,w} = \frac{I \rho}{2 \pi a_w}$$

where I = Current supplied to the ground

ρ = Resistivity of the ground

a_w = Spacing between any two adjacent electrodes.

2. For the Schlumberger Array :

$$V_{s,\,s} = \frac{b I \rho}{\pi a_s^3}$$

b = Spacing between measuring electrodes

a_s = Half the spacing between the current electrodes.

ARRANGEMENTS:

In case of Schlumberger arrangement, the four electrodes are placed along a common line, with the outer two serving as current electrodes and the inner two as measuring points. The inner pair are located at the center of the array and the separation between them is small compared to the total array length, usually less than one fifth the total length. In studying resistivity as a function of depth, the current electrode separation is increased in a series of steps but the measuring electrode separation is increased only when the observed voltage becomes too small to measure.

a = for current electrode

Kg = Geometrical factor

$Kg = \pi a^2/b$

In case of Wenner array, four equally spaced and colinear electrodes are used. The outer two electrodes are normally used to provide current to the ground, while the inner two are used to measure the voltage drop caused by this current. If resistivity is being investigated as a function of depth, the center point of the array is held fixed, and the array spacing is expanded about the array mid point, all four electrodes being separated by equal distances at all times. If the measurements are made over a completely uniform earth, the resistivity of the earth may be computed very simply from the measured voltage current and array spacing.

$$\rho = Kg \frac{V}{I}$$

V → Voltage measured between two inner electrodes

I → Current supplied to the outer two electrodes

Kg → Geometric factor for the array ($2\pi a$) when a is the spacing.

INTERPRETATION OF RESISTIVITY DATA

The following methods have been employed to interpret the resistivity data:

1. Tagg's First Method

The method is based on the theory of images. The effect of various images on the value of the potential at a point is considered. If configuration used is wenner, the surface resistivity is denoted by ρ_1 and apparent resistivity by ρ_a, if a is the electrode separation. The relation between ρ_1 and ρ_a will be as:

$$\frac{\rho_a}{\rho_1} = 1 + 4\sum_{n=1}^{\infty} \frac{K^n}{\sqrt{1+\left(\frac{2nh}{a}\right)^2}} - 4\sum_{n=1}^{\infty} \frac{K^n}{\sqrt{4+\left(\frac{2nh}{a}\right)^2}}$$

K - Electrification constant

h - The depth of interface

$$K = \frac{\rho_2 - \rho_1}{\rho_2 + \rho_1}$$, ρ_2 = Resistivity of the 2nd infinite layer

In this method apparent resistivity ρ_a is plotted along Y axis against electrode separation 'a' on X axis. The vertical electrical sounding curve is then drawn. The VES curve is extended upto Y axis keeping in view the trend of the curve for less electrode separations. The surface resistivity ρ_1 is the value of resistivity on the apparent resistivity axis at the point where the curve cuts the y axis. Three different electrode separation a_1, a_2, a_3 are chosen on the curve and corresponding values of ρ_a noted. The ratio $\frac{\rho_{a1}}{\rho_1}$ or $\frac{\rho}{\rho_{a1}}$ is calculated depending on the variation of conductivity of the second layer with reference to first layer.

Tagg's positive or negative curves drawn between $\frac{h}{a}$ and $\frac{\rho_1}{\rho_{a1}}$ or $\frac{\rho_{a1}}{\rho_1}$ for different values of k are used to find out a series of values of $\frac{h}{a}$ for the calculated ratio between surface resistivity and the apparent resistivity at an electrod separation "a" for different values of k. Each value thus obtained is multiplied by a to yield h. This results on a set.of value of h and corresponding values of K for an electrod separation 'a'. A graph is plotted between h and K for a particular value of a. The procedure is repeated twice again with 'a' different values of electrode separation a_2 and a_3 The three curves between h and K under ideal condition should intersect at one point. The value of h and K at the point of intersection are the depth of interface between the Ist and 2nd Layer and the value of electrification constant. When the curves instead of intersecting at a point form a triangle, the centre of gravity of the triangle is used to determine the value of h and K. The resistivity of the 2nd layer "ρ_2" is determined by the following relation:

$$K = \frac{\rho_2 - \rho_1}{\rho_2 + \rho_1}.$$

This method can be extended to three layer problem.

Knowing the resistivity of the Ist and 2nd layer and thickness of the Ist layer as h_1, let h_2 be the thickness of the 2nd layer and 'd' being the distance of maximum curvature, then depth of the 2nd layer can be estimated by the relation:

$$\frac{h_1 + h_2}{\rho'} = \frac{2}{3} d$$

If ρ' is the replacement resistivity for the combined effect of layers of resistivity ρ_1 and ρ_2 then from the Hummels Formula we get

$$\frac{h_1 + h_2}{\rho'} = \frac{h_1}{\rho_1} + \frac{h_2}{\rho_2}$$

The resistivity ρ' is now used as surface resistivity and the depth of the 2nd Interface can be calculated as before by plotting the curves between K and $h_1 + h_2$. The electrification constant K_2 in this can be related with ρ_3 (the resistivity of the 3rd layer) by the relation

$$K_2 = \frac{\rho_3 - \rho_2}{\rho_3 + \rho_2}$$

where ρ_2 is the resistivity of the 2nd layer.

The layer of $(h_1 + h_2)$ determined by the intersection of graphs is used for an accurate determination of resistivities.

MOORE'S CUMMULATIVE CURVE METHOD

This method suggested by Moore(1945) consists essentially in plotting the integral of the apparent resistivity values against the electrode separation of a Gish Rooney system, and observing the breaks in the integral curves. The point of intersection of straight lines drawn to intersect at zones of maximum curvature in a cummulative (integral curve) indicates the depths of underlying materials.

The data for analysis are obtained by taking an initial electsode spacing of some convenient value say 3 ft for shallow work and the electrode spacing is then increased regularly, by increments of 3 ft for each successive determination. The initial value of apparent resistivity is plotted as the initial ordinate of the cummulative curve. Each subsequent value of apparent resistivity is added to the sum of all preceeding values and each total, thus obtained is plotted as ordinate of another point in the cummulative curve. By using regularly increased electrode spacing it would appear that, a substantially straight line with a given slope is obtained. As the electrode spacing approaches to a value corresponding to the depth of the surface layer, the plotted cummulative curve tends to change its direction, the new slope being a function of the two resistivities.

The line drawn tangent to the cummulative curve and intersecting in the region, where the change in slope occurs will give a good approximation of depth to the interface of the two materials if we project the point on the horizontal depth axis.

CURVE MATCHING

In this method we compare the field graph plotted on a double log paper with the therretical curves plotted on Log-Log papers with same modulus for different resistivity and quickness ratios. The bi -log paper permits a wide range of values for the variable to be presented on a single graph and the resistivity for smaller electrode separation can be accurately plotted. The apparent resistivity can be expressed in terms of dimensionless ratios as is evident from the relation

$$\frac{\rho_a}{\rho_1} = \left[1 + 2 \sum_{n=1}^{n=\infty} \frac{K^n}{\left(1 + \left\{\frac{2nh}{a}\right\}^2\right)^{1/2}} \right]$$

This means that the theoretical curves computed with equation can be plotted without regard to the system of units used and without regard to the absolute magnitude of their resistivities, electrode spacing or thickness of the beds. The apparent resistivity is calculated in terms of the resistivity of the overburden and the spacing in terms of the thickness of the overburden. In normal field survey the values of ρ and h are unknown, being the object of the survey. The dimensionless ratios $\left(\frac{\rho_a}{\rho_1}\right)$ and $\frac{a}{h}$ expressed in the terms of overburden resistivity and thickness thus can not be used in plotting field data. The slope of the curve plotted in the Logarithmatic coordinates is preserved even when the ordinate and abscissa of each point alongthe curve are multiplied by arbitrary constants. The preservation of curve shape in longarithmic coordinates is the basis for curve matching method of interpretation.

A field curve which is plotted between apparent resistivity and electrode separation will have the same shape as a curve computed from the theoretically derived expression provided both are plotted on the same logarithmic scale. The field curve may be compared direct with a set of theoritical curves by superposition. The restriction in moving the field curve around is that coordinate axis of both sides of curves must be kept parallel.

The curve matching may be either complete or partial. In case of partial matching a part of the field curve is matched with standard curve. The restriction being that the thickness and resistivity of the common layers for two partial matchings should come out to be same.

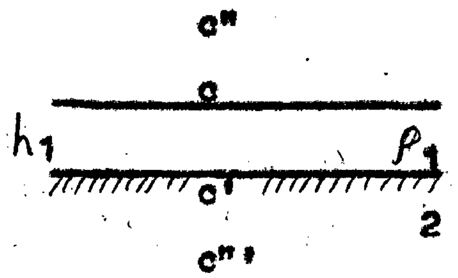
The distribution of the Conductivity (σ) within ground

It was shown by Slichter and Langer that if 'σ' is a function of depth only, the equation div $(\sigma \nabla V) = 0$ possesses a unique solution. Then the subsurface distribution of σ can be calculated from a knowledge of the surface potential produced by a single point electrode without any further physical or geologic data.

In the general case when σ is not a function of z only, Stevenson proved that the equation div $(\sigma \nabla V)$ possesses no unique solution. In this case if the surface potential is measured everywhere for all positions of a point electrode on the surface, the distribition of σ

can, in principle, be uniquely determined.

In the simplest case we have a horizontal layer of thickness h_1 and resistivity ρ_1 overlying a second homogeneous medium of resistivity ρ_2. The potential of a point electrode C, through which a current I is passing into such an earth was first calculated by Hummel by the method of electric images.



The potential is given by the sum of

(i) The potential of c in a semi infinite medium of resistivity $\rho_1 = \dfrac{I\rho_1}{2\pi CP}$

(ii) The potentials of fictitious current sources c', c", and c"' ... etc.

where c' is the image of c in the plane $z = h$

c" is the image of c' in the plane $z = 0$

Thus the potential at the surface point P is

$$V(r) = \frac{I\rho}{2\pi}\,\frac{1}{r}\left\{1 + \sum_{n=1}^{\infty}\frac{2k^n r}{(r^2 + 4n^2h^2)^{1/2}}\right\}$$

where $r = CP$

$$K = \frac{\rho_2 - \rho_1}{\rho_2 + \rho_1}$$

The equation can be expressed in the closed form

$$V(r) = \frac{I\rho}{2\pi} \frac{1}{r} \left\{ 1 + 2r \int_0^\infty K(\lambda, k, h) Jo(\lambda r) d\lambda \right\}$$

where $k(\lambda) = K \exp(-2\lambda h)/(1-K \exp(-2\lambda h)$

and Jo is the Bessel function of order zone a(apparent resistivity) is determined for different values of electrode separation(a). Thus the value of $\frac{\rho a}{\rho 1}$ is plotted against $\frac{a}{h_1}$ on a double log paper for different values of $\frac{\rho 2}{\rho 1}$, ranging from 0 to ∞

i.e. $\frac{\rho 2}{\rho 1} = 0$ i.e perfectly conducting substratum

$\frac{\rho 2}{\rho 1} = \infty$ i,e perfectly insulating substratum

This plot shows that ρa approaches to ρ_1 when the current electrode separation is small compared to the thickness of the top layery and ρa approaches to ρ_2 when current electrode separation is large.

With the addition of a third layer (h_2, $\rho 2$) sandwiched between the top layer (h_1, $\rho 1$) and the substratum ($\rho 3$) the problem becomes much more complicated. The apparent resistivity curve can then take four basic shapes depending upon the relative magnitudes of ρ_1, ρ_2, $\rho 3$

1. Q type (descending)
2. A type (ascending)
3. K type (displaced Anisotropic)
4. H type (Hummel type with minimum)

In every case, however, $\rho a \to \rho_1$ for small values of a
$\rho a \to \rho_3$ for large values of a

At intermediate values of L it is influenced by the resistivity of the middle layer. The observed ρa is plotted against "a" on a transparent double log paper with the same modulus as the master curves paper. Keeping the respective axis parallel, the transparent paper is Slid on various master curves in succession until a satisfactory match is obtained with some curve (if necessary an interpolated one). The value of $C_1C_2 = 2a$ coinciding with the point 1.0 on the X-axis of the matching master curve gives h_1 and the value of a coinciding with the point 1.0 on the X axis gives $\rho 1$. The values of $h2, \rho 2$ ... etc. are obtained from the appropriate parameters belonging to the matching master curves.

The potential distribution on the surface of an earth composed of four or more layers is often needed. It can be determined by electric images but the numerical computations using conventional methods are laborious. A rigorous and accurate method for rapidly calculating the resistivity curve for any combination of thickness and resistivities in a stratified earth having a perfectly conducting or insulating substratum, was developed by FLATHE. But due to inherent limitation, this method is now of historical importance.

If in equation Div $(\rho \nabla V)$, $\sigma = \frac{1}{\rho}$ is constant we get Laplace's equation $\nabla^2 V = 0$ which is true within each layer of the stratified earth. The potential distribution on the surface is given by

$$V(r) = \frac{I \rho}{2\pi} \frac{1}{r} \left\{ 1 + 2r \int_0^\infty K(\lambda, K, h) \, Jo(\lambda r) \, d\lambda \right\}$$

$K(\lambda)$ is a function of all the strata thickness and resistivities, known as Kernel function of resistivity Its value for the two layer earth is

$$K(\lambda) = K \exp(-2\lambda h) / (1 - K \exp(-2\lambda h))$$

For three layer earth, its value will be:

$$K(\lambda) = \frac{K_1 \exp(-2\lambda h_1) + K_2 \exp(-2\lambda(h_1 + h_2))}{1 - K_1 \exp(-2\lambda h_1) - K_2 \exp(-2\lambda(h_1 + h_2)) + K_1 K_2 \exp(-2\lambda h_2)}$$

For an arbitrary horizontally stratified earth $K(\lambda)$ is obtained by solving a certain system of linear simultaneous equations resulting from the boundary conditions.

<u>Determination of Kernel function</u>

For Schlumberger array we have

$$\rho a = \frac{\pi a^2}{I} \frac{dv}{dr} = \frac{\pi a^2}{2b} \frac{\Delta V}{I} \qquad \ldots (1)$$

where $V(r) = \frac{I\rho}{2} \frac{1}{r} \left(1 + 2r \int_0^\infty K(\lambda, k, h) \, Jo(\lambda r) d_\lambda \right)$ .... (2)

From these two above equations (1) and (2) we get

$$\rho a = \rho_1 (1 + 2 r^2) \int_0^\infty K(\lambda) J_1 (\lambda r) \, d\lambda \qquad \ldots\ldots (3)$$

Since $Jo'(x) = -J_1(x)$

J_1 is the Bessal function of order 1.

Now according to Hankel's transformation in Bessal function theory, if we have a function f(r) such that

$$f(r) = \int_0^\infty K(\lambda) Jn (\lambda r) \, d\lambda$$

$$\text{then } K(\lambda) = \int_0^\infty f(r) J_n (\lambda r) \, r dr$$

Applying the transformation to equation (3) we find that

$$K(\lambda) = (\tfrac{1}{2}) \rho_1^{-1} (\rho a - \rho 1) \, r^{-1} J_1 (\lambda r) dr$$

The equation first observed by KING shows that the Kernel function can be computed. It plays a central role in the modern theory of the interpretation of resistivity data.

Direct derivation of the layered earth parameters

The electric conductivity of a layered earth is a function of the depth only. According to the Slichter - Langer theorem, therefore, the knowledge of the surface potential due to a point electrode should suffice to determine the thickness and resistivities of the various layers.

Two Interpretation methods based on this approach are available -

The Ist method can be traced back to Hummel (although it has been subsequently refined and improved by several workers). Let us take the three layer case (ρ_1, ρ_2, ρ_3) with $h_2 > h_1$, clearly as long as the current electrode separation does not exceed a certain value, the apparent resistivity curve will not differ appreicially from a two layer case with the same $\frac{\rho_2}{\rho_1}$ as that in the three layer case under consideration. At larger electrode separations the third layer, that is, the infinite substratum (ρ_3) will influence the measurements.

Hummel showed that for sufficiently large separations the apparent resistivity curve obtained is virtually the same as that for a two layer case with the substratum but with a top layer of thickness $\bar{h} = h_1 + h_2$ and a resistivity given by $\frac{\bar{h}}{\bar{\rho}} = \frac{h1}{\rho} + \frac{h2}{\rho_2}$

$\bar{\rho}$ is derived by applying Kirchoffes law for resistances in parallel. Thus by matching the initial branch of a measured resistivity curve with an appropriate curve (two layer case) we obtain ρ_2 and h_1 (ρ_1 is known from the asymptote of ρ_a) for very small electrode separations.

Similarly by matching the branch obtained with large electrode separations we get $\frac{\rho_3}{\bar{\rho}}$ and $\bar{h}$ again with the help of two layer master curves. Then ρ_3 and h_2 can be separately evaluated.

The above procedure is facilitated by the use of an auxiliary diagram in which $1 + \frac{h_2}{h_1}$ is plotted along the x axis and on the y axis, both on a logarithmic scale, and for this reason known as the auxiliary curve (or auxiliary point) method. The method is easily extended to any number of layers with thicknesses h_1, h_2, h_3 ... and resistivities ρ_1, ρ_2, ρ_3, It is of course, valid only if $h_1 < h_2 < h_3$ .... but can also be validly employed in many other cases by constructing special auxiliary diagrams.

The second direct method of interpretation is due to Pekris and is based on Slichter's analysis. Practical procedure is as follows:

We measure the surface potential V(r) of a single point electrode by removing the other current electrode to great distance. We complete the function

$$K(\lambda) = \lambda \int_0^\infty V(r)\, Jo(\lambda r) r dr,$$ by numerical or mechanical integration. Plot $|f_1(\lambda)|$ against λ where $$f_1(\lambda) = \frac{K(\lambda) + 1}{K(\lambda) - 1}$$

for large λ the points will lie on a straight line with a slope $2h_1$ and an intercept $\log\left(\frac{1}{K_1}\right)$ with the y axis where $$K_1 = \frac{\rho_2 - \rho_1}{\rho_2 + \rho_1}$$. If all the points lie on a straight line $h_2 = \infty$ (two layer case) otherwise with h_1 and K_1 thus determined, we proceed to calculate another closely similar function $f_2(\lambda)$, The plot of whose logarithmic gives in turn h_2 and K_2 $\frac{\rho_3 - \rho_2}{\rho_3 + \rho_2}$

The process is continued by calculating a third function f_3 if all the points do not lie on a straight line, that is if $h_3 \neq \infty$ and so on.

This method implies no restriction on the relative magnitudes of h_1, h_2, h_3, ... and is thus quite general. Unfortunately it requires considerable computation.

Recently Developed Method

Koefoed has recently developed the method into a rapid practical procedure. He starts by constructing $K(\lambda)$ from the observed ρ a curve by means of a small number of standard curves. Next he defines a modified Kernel function $Gn(\lambda) = K(\lambda) / (1 + K(\lambda))$ and this he treats in essentially the same manner as $f(\lambda)$ above to obtain successive strata parameters.

AMBIQUITY IN THE INTERPRETATION

A relatively thin layer sandwiched between two layers whose resistivities are much higher than that of the sandwiched layer. The current flow in the earth will then tend to concentrate into the middle layer and evidently the fraction of the total current carried by it will be unaltered if we increase its resistivity (ρ) but at the same time increase the thickness (h) in the same proportion. Thus all the midle layers for which the ration $\frac{h}{\rho}$ is the same are electrically equivalent.

On the other hand if the resistivity of the middle layer is much larger than that of the layers on either side, the electric current density in it will tend to be less than in the layer below. The current flow will evidently be unaltered if we increase the thickness of the middle layer provided we at the same time decrease its resistivity in the proportion. Thus in this case all middle layers for which the product $h\ \rho$ is the same are electrically equivalent. Thus h and ρ can not be separately determined. If in the first case the thickness of a layers is very small compared to its depth, the effect on the ρ a curve is so small that the presence of the layer will be suppressed.

CHAPTER - IV

INSTRUMENTATION

Resistivity survey can be carried out either by using low frequency alternating current or by direct current. However there are some limitations and advantages in each of these types of energizations. There being two reasons why direct current is not normally used. The first is that electrochemical e.m.f.'s produced between the metal electrodes and the ground would be a source of error in the readings. The second is that direct current measurements are also affected by natural earth currents which produce a slowly varying potential difference across the electrodes. Electrochemical effects can be avoided by using non-polarizing electrodes but these are troublesome and reduce the operating speed. An electrode of this type consists of a porous pot containing a metal electrode immersed in an electrolyte of one of its own salts.

Like the potential differences of electrochemical origin earth current tend to be unidirectional and both are eliminated if alternating current is used in the ground and rectified before measurement. It is very desirable that low frequency be used, this is because ground inductance and capacitance and more complex frequency effects such as the induced polarization becomes increasingly important at frequencies above a few ten cycles per second.

Advantages of direct current

There are some definite advantages of using direct current. These are:

(1) The skin depth is infinite.

(2) Instrumentation is comparatively simple.

ADVANTAGES OF ALTERNATING CURRENTS :

Everything discussed above relates to direct current. However, alternating current offers definite advantages from several points of view. These advantages have often led to a preference for the method in many countries, for example, Sweden Advantages include ease of power production and of measurement, facility to amplify potentials, and the ability to filter. This last possibility in particular facilitates the distinction between the useful signal and nonwanted electrical perturbations which can be natural or man-made, such as polarization of the electrodes or telluric currents which in general are slowly variable.

Admittedly, the problem of alternating current distribution in a heterogeneous medium is more difficult. However, alternating current carries an additional independent parameter in its frequency; also, one can measure both the electric field and its phase, and even the components of the induced magnetic field. There thus results a more flexible operation, as much for the execution of measurements as for production of currents In particular, it is easy to induce current in the earth without the necessity of electrodes, and to measure the results in the same way; this technique permits a continuous measurement as one advances, either by vehicle or by plane.

DISADVANTAGES OF ALTERNATING CURRENTS : - THE SKIN EFFECT

All these advantages however are largely offset by the major difficulty met in trying to penetrate a conducting earth with alternating current. This phenomenon, called the skin-effect, consists in the concentration of alternating current near the contact between materials of different resistivities. These concentration of current is more pronounced at higher frequencies and for greater differences in resistivity and becomes particularly important at the surface of the earth. The skin-effect means a rapid decrease of current density with depth, and consequently a decreased depth of investigation.

The depth of penetration depends on the manner in which the current is introduced in the earth, but one can set an upper limit with respect to a plane wave, that is to say, to a source sufficiently removed that for direct current the current density would be constant at all depths. This is the case for certain telluric currents, for example.

The so-called "depth of penetration" is given in kilometers by the expression :

$$\delta = \frac{1}{2}\sqrt{\rho T}$$

and is the depth at which the current density is reduced to about one-third its value at the surface. The period T is given in seconds and the resistivity ρ in ohm-meters.

It follows that a 1000 cycles-per-second current in an earth of 10 ohm-meters resistivity would be reduced to one-third its surface density at a depth of 50 meters. Variations of the resistivity of the near-surface formations thus have a preponderant effect on the distribution of the electric field at the surface, rendering a severe limitation on alternating current methods in sedimentary sections of relatively high conductivity.

POSSIBILITY OF A COMPROMISE :

However, it is still possible to profit the advantages on the condition that frequencies be used according to the resistivity of the beds and the depth of investigation desired. For example, if the resistivity is 10 ohm-meters as above, a current of one cycle per second would have a useful depth of penetration greater than 1000 meters. The same useful penetration could be attained in a medium of 1000 ohm-meters with a current of 100 cycles per second.

In passing, it is noted that sinusoidal alternating current is not the only way in which a variable current may be used. The study of the behaviour of short pulses, or of the transient behaviour when a current is applied, which is theoretically equivalent, is able to afford certain practical advantages. These techniques are sometimes used today.

THE ROLE OF SKIN EFFECT IN PROSPECTING WITH DIRECT CURRENTS :

The skin-effect is also important from a practical viewpoint in prospecting with direct current. In effect, it appears when the current circuit is closed or opened. The current attains its steady-state distribution only after a certain period of time has elapsed. Inasmuch as certain techniques to overcome variable natural potentials involve a series of opening and closing the current circuit, it is necessary to investigate the time constants and to assure ourselves that the steady-state is really attained.

INSTRUMENTS OF RESISTIVITY SURVEY

Based on the above discussion the instruments available for carrying out resistivity field operations can thus be divided into two types.

1. A.C. Instruments

Following are the A.C. Instruments:

(a) Tellohm Soil Resistivity Meter

(b) Geophysical Megger

(c) A B E M Terra meter

2. D.C. Resistivity Meter

In D.C. resistivity measurements, instrumentation is usually simple. Current is generally provided from dry batteries. Current electrodes are generally steel or copper-clad steel stakes driven a few inches into the ground. In dry areas, the soil around the electrodes may have to be moistened to improve contact. Where bare rock is exposed at the surface, it may not be possible to drive a stake into the ground and in such a case, a current electrode may be formed by building a small mud puddles around a piece of copper screening.

A most useful and protable instrument is the Tellohm soil Resistivity meter, has a simplified circuit. Alternating current is supplied to the two outer electrodes the potential differences across the inner two being measured by a potentio meter. Variation in the supply voltage affects both potentio -

meter and current electrode equally, and therefore does not upset the balance point. Thus it is not necessary to measure current and potential difference separately to obtain a value for resistance and the potentiometer can be calibrated in ohms. The advantage of using a potentiometric method of measurements is that by so doing no current is drawn from the ground through the potential electrodes, and therefore the resistance between these and the ground is not incorporated in the total measured resistance.

The current is supplied by five 1.5 Volt batteries and converted into alternating current at a nominal 110 cycles per sec. by vibrator. To operate the instrument the potentiometer is adjusted until a null reading is obtained on the galvenometer. The potentiometer then gives directly the ground resistance between the inner electrodes.

The another instrument is Geophysical Megger uses a hand cranked generator and can generate about three times as much power as the Tellohm under normal operating conditions. It has also the advantage of working at a lower frequency, this varying between about 10 and 20 c/s depending upon the speed of rotation of the handle. The principle of the instrument is somewhat similar to that of the Tellohm, but here a synchronous commutator is used for rectification and the resistance is read on an ohm meter directly and not from a calibrated potentiometer. The operating range is from 0.3 to 30 ohm full scale with an accuracy of $\pm$ 1 percent of full scale on each range. Provided that ground resistivities are not too low, the Geophysical Megger can be used for investigations down to ahdepth of several hundreds of feet. Its disadvantages when compared with the Tellohm are its greater weight and shorter scale range.

A newer instrument, the ABEM Terrameter, also operates at a low frequency (4 c/s) and is therefore suitable for greater depths of investigation than the Tellohm, but is based on a transistorized electronic oscillator and is therefore far more portable than the Megger. The maximum output is 6W and it covers a resistance range from 0.01 to 10,000 ohm.

The another instrument is Resistivity meter, the resistivity measurements in the field can be made either at very low frequency or by direct current.

THE FOLLOWING ~~INSTRUMENTS~~ EQUIPMENTS WERE USED IN THE RESISTIVITY SURVEY

1. Resistivity meter (Manufactured by N.G.R.I. Hyderabad)
2. Electrodes
3. Power source
4. Insulated cables
5. Surveying Equipments.

1. RESISTIVITY METER

The resistivity meter used in the survey was manufactured by the National Geophysical Research Institute, Hyderabad. The circuit diagram of the instrument used is shown in the figure No. 9 .

PRINCIPLE OF THE RESISTIVITY METER

The resistivity meter is a null detector and utilised two potentiometer circuits for measurements. The voltage

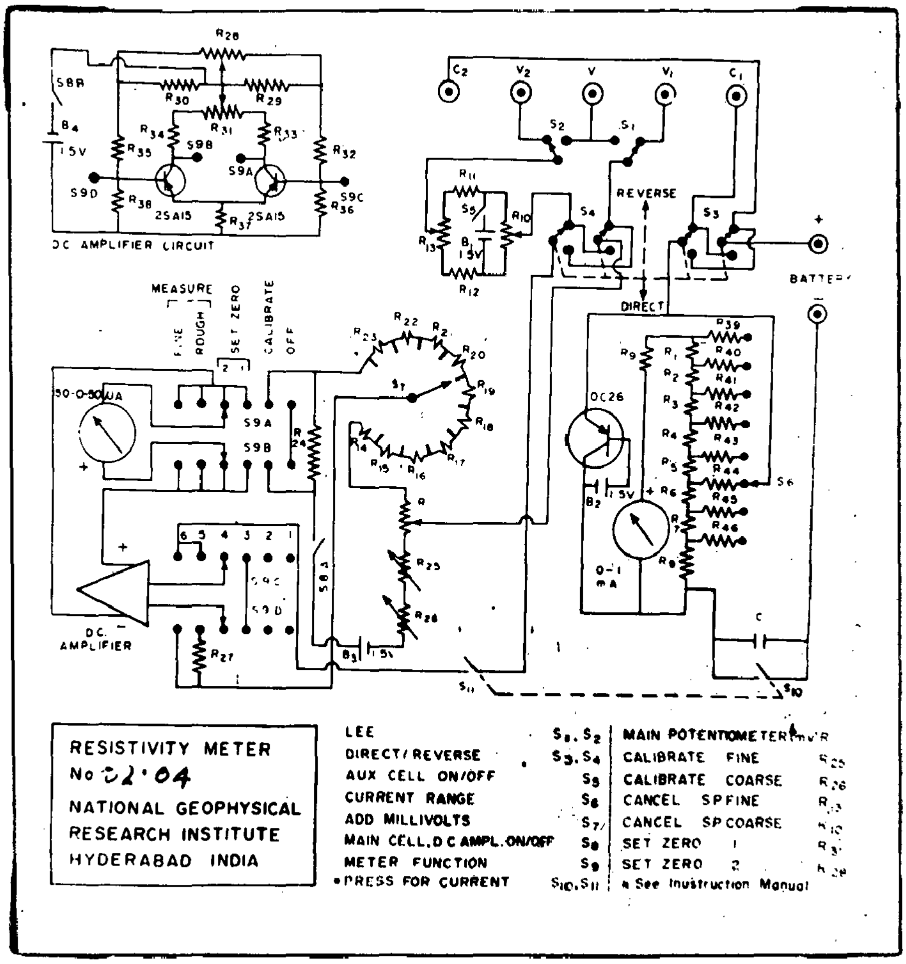
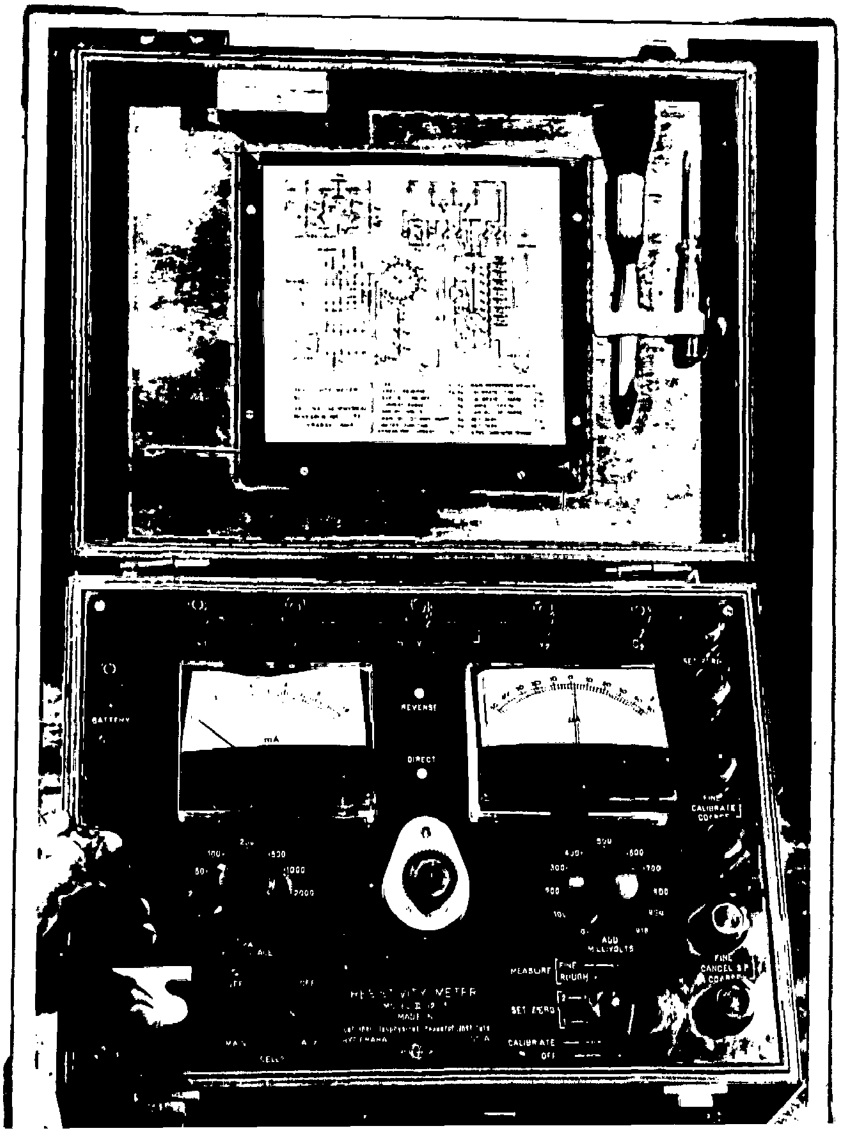

DC AMPLIFIER CIRCUIT
S8B
B4
1.5V
S9D
S9C
2SA15
2SA15
S9A
S9B
C2
V2
V
V1
C1
S2
S1
REVERSE
S4
S3
S5
B1
1.5V
BATTERY
+
DIRECT
MEASURE
FINE
ROUGH
SET ZERO
CALIBRATE
OFF
30-0-30μA
S9A
S9B
S7
OC26
1.5V
B2
0-1 mA
S6
S9C
S9D
DC AMPLIFIER
S8A
B3
1.5V
S11
S10
C
RESISTIVITY METER
No 02·04
NATIONAL GEOPHYSICAL RESEARCH INSTITUTE
HYDERABAD INDIA
LEE S1, S2
DIRECT/REVERSE S3, S4
AUX CELL ON/OFF S5
CURRENT RANGE S6
ADD MILLIVOLTS S7
MAIN CELL, D.C AMPL. ON/OFF S8
METER FUNCTION S9
*PRESS FOR CURRENT S10, S11
MAIN POTENTIOMETER
CALIBRATE FINE R25
CALIBRATE COARSE R26
CANCEL SP FINE R13
CANCEL SP COARSE R10
SET ZERO 1 R31
SET ZERO 2 R28
* See Instruction Manual

FIG. 9


REVERSE
DIRECT
BATTERY
mA
RESISTIVITY METER
MEASURE
FINE
ROUGH
SET ZERO
CALIBRATE
OFF
ADD MILLIVOLTS
FINE
CALIBRATE
COARSE
CANCEL SP
COARSE

picked up between the two electrodes for potential measurements is neutralised by feeding an equal and opposite voltage through a pre-calibrated circuit which gives the value of potential difference between the two different potential electrodes. The current through the ground is measured by milliammeter placed in the current circuit. The resistivity can be calculated by measuring the potential difference Δ V and current I. through the current electrodes.

The instrument is very portable and assembled with a wooden cabinet. There are two potentiometer circuits, one potentiometer is used for the cancellation of self potential while the other is used for the measurement of potential difference between potential Electrodes. The potential difference is measured with the help of a circular disc graduated in millivolts and can measure the range of 0 to 600 millivolts. There is a d.c m.ammeter of 700 ampere range, which is graduated in both the sides of the zero value of the scale i.e. (50 -0- 50). This ammeter is placed in the potentiometer circuit for measuring the current passing through the potentiometer resistance wire. The same ammeter functions as a galvenometer for cancelling self-potential and for the balance of the potential due to ground.

There are two switches for fine and coarse cancellation of s-p and potential differences due to the ground. In order to cancell out the minor inhomogenities the resistivity meter is provided with a current reversing key between the two current electrodes. Two sets of readings are taken one in the direct and other in reverse direction.

OPERATIONS

1. The electrode connections are made depending upon the polarity of the current. The main and auxiliary switches are then switched on.

2. When the function switch is off the null indicating meter should show zero. If not, the mechanical zero is adjusted with a screw driver.

3. Function switch is turned to calibrate position and calibrate control is adjusted to get a specified deflection.

4. The function switch is turned to "set zero 1 position and set zero 1 control is adjusted so that needle is brought to zero.

5. Again the function switch is turned to "set zero 2" position and 'set zero 2' control is adjusted to bring the needle to zero.

6. The function switch is then turned to "coarse" measuring position current range is put to 2000 mA and 'Add mv' switch is adjusted to zero. Potentiometer circuit is set to zero. Also the direct/reverse switch is put to "Direct".

7. The double button switch is pressed lightly with left thumb so that only the potential Electrodes are connected. The shift in the needle of galvenometer arised due to s.p. in the circuit is balanced by adjusting 'cancel s.p' control.

8. The double button switch is pressed again slightly harder in order that current electrodes are also connected. The deflection of the galvenometer is balanced by adjusting main potentiometer and add mv control and then the function switch is turned to fine measuring position.

9. An accurate null is obtained by adjusting the potentiometer. The total potential difference is noted by reading the potentiometer plus "add mv" control. Also the current is noted and the thumb is lifted off.

10. Direct/reverse switch is put to "Reverse" and step No.9 is repeated.

11. For every new observations, step 6 to 10 are repeated.

ELECTRODES

For Wenner configuration four electrodes are used during the resistivity survey. Two electrodes known as current electrodes are used to pass the current into the ground while the other two are used to measure the potential difference between two points.

(a) Current Electrodes

They are metal (Cu) rods about 18" long having a diameter of about an inch. The current electrodes are pointed at the base so that they can be easily driven in the ground. The electrodes are hammered with the help of a heavy hammer so that the electrodes reach to a depth of about 6". This helps in making a proper contact with the moist soil.

(b) Potential Electrodes

Even with the best possible electrodes a contact potential of the order of 100 mv is produced due to the existence of contact polarization. This is a serious drawback with the direct contact potential electrode. Non-polarizable electrodes are used to overcome this difficulty.

The non-polarizing electrodes are made up of identical metal rods put in identical solutions. The potential electrodes are flat bottomed porous pots filled with a saturated solution of $CuSO_4$. A perforated copper rod is put in the solution. The flow of the electric current through earth in vicinity of pot causes the electrolysis of the $CuSO_4$ solution

$$SuSO_4 \quad \text{—} \quad Cu^{++} + SO_4^{--}$$

The Cu^{++} ions moves towards the central copper rod and get deposited there

$$Cu^{++} \text{———} Cu + 2e$$

The flow of these copper ions sets up a potential at each electrode. This potential devéloped at each eléctrode depends upon the current flow in the vicinity of electrodes which itself depends on the resistivity of subsurface formation, lesser will be the current flowing through surface layer of earth and hence smaller will be the potential developed at the electrode. The differences of potential at two electrodes so developed is measured with the help of d.c. compensator arrangement in resistivity meter.

Sometimes rods made up of zinc and solution of $ZnSO_4$ is also used in the porous pots to make the potential electrodes. The rod is hollow and the surface area of contact is large this helps in the maintainence of potential equilibrium. To avoid the internal polarization of the potential electwodes the following precautions should be taken:

1. The rod must be cut from the same piece of metal to ensure that the physical and chemical characters of the two rods in the two electrodes are the same.
2. The two porous pots should have the same degree of porosity.
3. The chemicals used should be pure and in recrystalized form. Distilled water should be used for the preparation of the solution.
4. The solution should be fully saturated and the porous pot should have a few crystals left in the bottom.
5. To prevent the loss due to evaporation the porous pot is sealed at the top with the help of a cork.

PRECAUTIONS

In order to get correct readings following points should be kept in mind.

1. Calibration should be checked after few observations and not after every observation on frequent use of the standard cell will result in rundown of the voltage.
2. Current range should not be changed when the current switch is on as it may destroy the ameter.

3. Drift of zero settings are easily seen as change in the null indicator position when the function switch is in fine and the thumb is off.

4. The null indicator should be adjusted, being very sensitive and delicate, a small error in balancing amounting less than half a small division on the meter amount to an error of about 0.1 mv.

5. The polarity of the battery cells inside the instrument should be checked carefully. Connecting the cells with wrong polarity may permanently damage the instrument.

6. The ammeter should never be touched while the instrument is in us e. This may lead to generation of static charges.

7. In cloudy weather too much of variation in the measurements are observed. They are due to electricity in the atmosphere. Hence no observation should be taken when the weather is cloudy.

POWER SOURCE :

A large number of flash light cells of 1.5 volts are connected in a series to form the source of power for the survey. Generally, 64 such cells are connected in series and kept in a wooden box to provide 96 volts. Seven to eight such boxes may be connected in series to form the source of current. The boxes are tapped at regular intervals so that various voltages can be drawn for the survey. Thus desired voltage can be obtained.

INSULATED CABLES :

Flexible "PVC" cables with thick copper core are used for the survey. The cables are rolled on four different wheels. The length of the cable is governed by the length of the profile. Since the current electrodes are at a greater distance than the potential electrodes from the observer, much longer cables are needed for current electrodes than for potential electrodes.

SURVEYING INSTRUMENTS :

Prismatic campass or thedolite is generally used for the purpose of laying profile lines and finding out the dissection of the profile line with respect to North. Measuring tape and meter chains are used for fixing the electrode positions.

REASON FOR SELECTING THE PARTICULAR ARRANGEMENT :

The selection of particular arrangement depends upon the following factors :

(i) Ease in carrying out in field

(ii) Availability of Master curves for interpretation.

Seeing these two points Schlumberger arrangement is best, as it takes less time in comparision to Wenner arrangement.

(1) In case of Wenner - In sounding method all the four electrodes are shifted for getting each reading, while in Schlumberger arrangement, potential electrodes are not changed regularly.

(2) In case of Wenner - We do not have master curves which are necessary for interpretation.

As two types of resistivity measurements are possible, the selection depends upon the problem, whether one want to study lateral variation in resistivity or vertical distribution of resistivity, for that reason one can use mapping (geoelectrical profiling) and Vertical Electrical Sounding respectively.

COMPARISION OF SCHLUMBERGER AND WENNER METHODS :

The Schlumberger theoretical master curves are based on the value of electric field intensity at the sounding stations, whereas the Wenner curves utilize the potential difference between M and N. Hence the Schlumberger and Wenner methods are sometimes known as gradient and potential methods, respectively.

As the names suggest, the anomaly in the Schlumberger or gradient method is rounded off, and the minor distortions are not indicated. Wenner curves, in contrast, retain these distortions and undulations, which are due to local inhomogeneities; as these are normally irregularly scattered in the ground, they are not important from a practical point of view.

While the electrical properties of the various layers are reflected to a remarkable degree in the Schlumberger curve, the theoretical condition that (MN/AB) → 0, required for apparent resistivity calculation, is never fulfilled in practice. In Wenner's method, such difficulties are not met with, since (MN/AB) is constant (1/3).

In the gradient method, MN is changed after a number of changes in AB; this cuts the working hours by about 50%. The cable length is reduced, and cable continuity as well as leakage locations in potential cables can be easily checked and detected.

The introduction of personal error in the Schlumberger method is small, since only two electrodes are changed at a time (rather than all four).

The Wenner method is sensitive to near-surface inhomogeneities, but cannot always clearly resolve deeper events (DEPPERMANN, 1954). The master curves (MOONEY and WETZEL, 1956) available for the potential method have been plotted for shallower events only.

FIELD PROCEDURE :

In field while laying out profile one must always keep in mind some of the factors which improve the results and increase the efficiency of the field work. Selection of a suitable place is one, most important factor to be considered. Its importance lies in the fact that the electrical resistivity method is very much sensitive to topographic effects which is due to local squeezing together or drawing apart of the E.P. lines and therefore effect the potential readings giving

FIG.10.

incorrect potential difference reading and hence fictitious results. Efforts, therefore should always be made to select fairly plain area for profile layout.

SURVEYING :

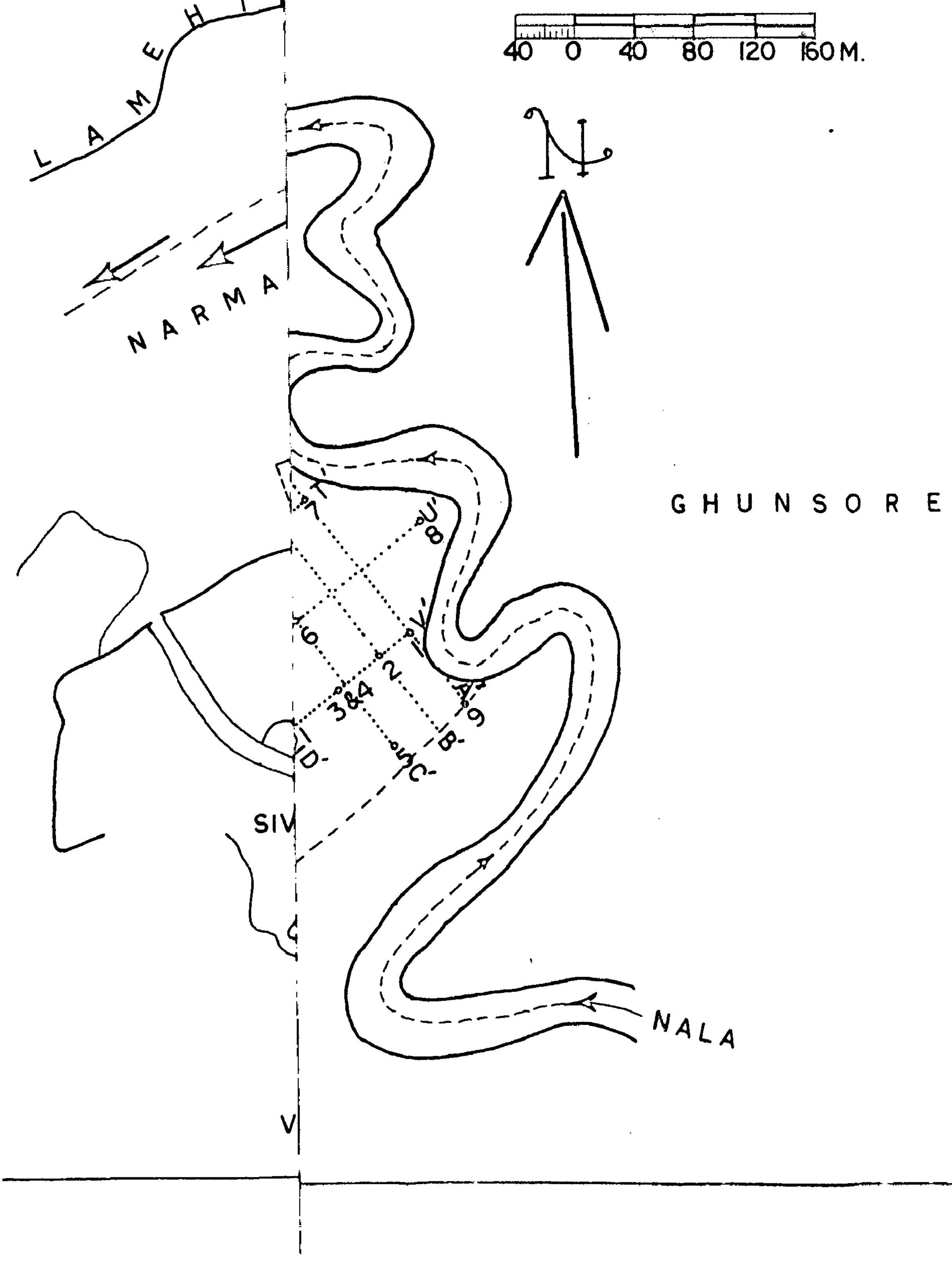
For laying out profile and marking various points for current and potential electrodes, the survey is started with prismatic campass. Base point was selected such that fairly plain area is available on the either side of it. Profile line is then marked with the help of prismatic campass kept at the base point and fixing a ranging rod on the either side of it. Bearing of this line is noted down positions for putting potential and current electrodes are marked, with the help of peg drawn into ground at predetermined distance from the base station on either side of it.

In order to ensure good contact between earth and electrodes and hence to reduce the contact resistance, holes are prepared and watered in advance. As the contact of current electrodes with ground can directly be made by drawing them into the ground through a distance of about 10 cm, so the preparation of holes is only found necessary for potential electrodes. By preparing the suitable holes and watering them in advance, the resistance offered by the contact between the pot and ground, which is variable from place to place and point to point can be made small. Normally holes of about 6 inch depth and diameter slightly greater than one diameter of the porous pots are prepared and watered in such a way that neither the hole is left dry nor it is over flood l with water.

MAP OF THE AREA SHOWING THE LOCATIONS OF VARIOUS SOUNDING POINTS

SCALE :— 16" = 1 MILE

1 CM = 40 METERS



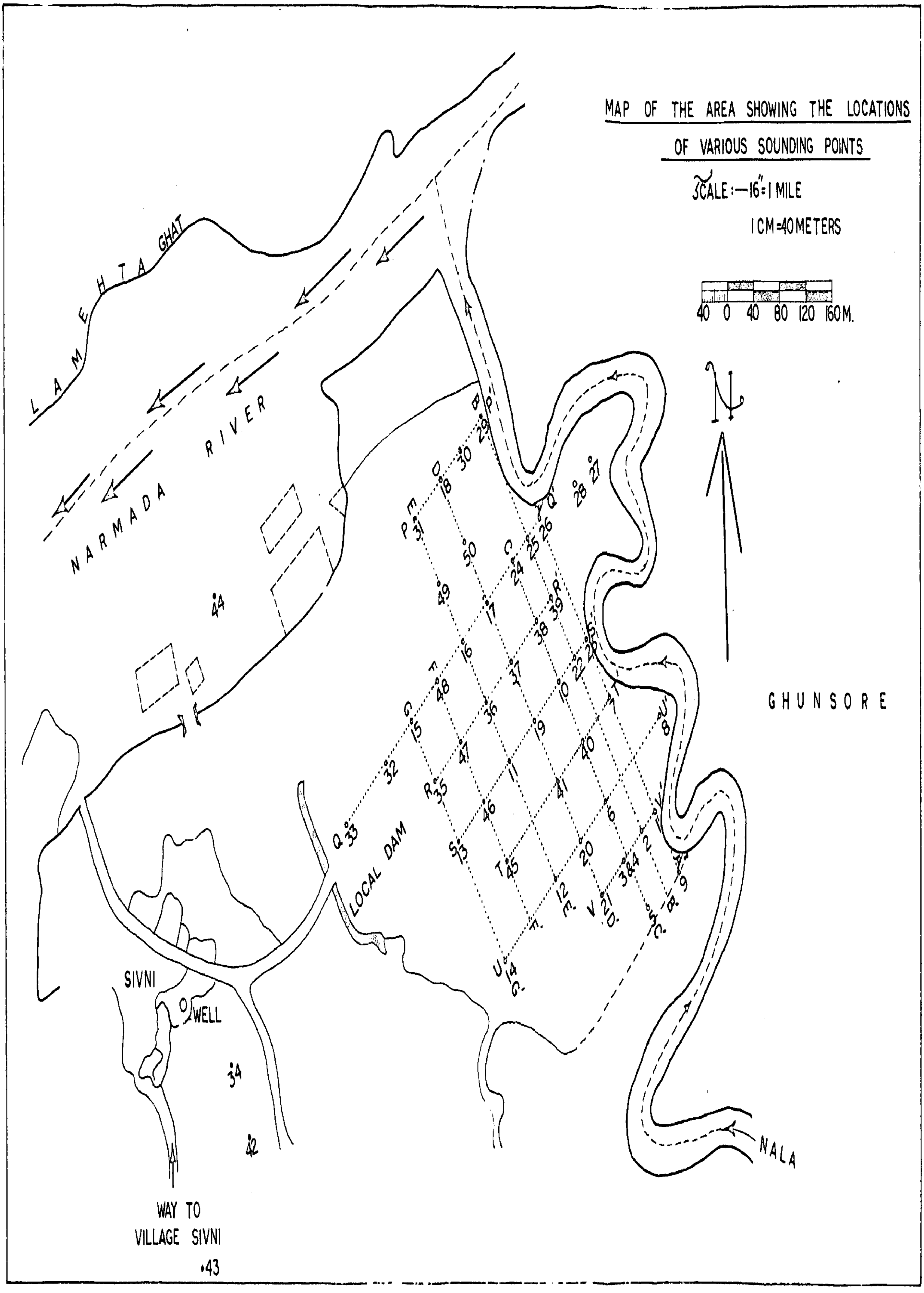
THE "MISE - À - LA - MASSE" METHOD

The name for a technique of electrical mapping was suggested by Schlumberger (1920) and may be translated on excitation of the mass. The idea is to use a subsurface conductive mass itself as one current electrode of a pair by connecting it directly to one pole of a battery, the second eletrode being as usual a metal rod but placed on the earth's surface at a great distance. A current is passed and the voltage at points on the surface are mapped by means of a volt-meter with respect to some base station.

It will be noticed that the operation is exactly the same as in the measurement of self potentials. If we consider the mass to be an electrically-charged body, the shape of the equipotential lines (Lines joining points with equal voltages) will in some measure reflect the geometry of the conducting body and should be expected to give clues as regards its extent, dip, pitch etc. It is obvious of course that atleast a part of the mass concerned must be accessive so that an electrical earthing can be made in it.

The mise-à-la-masse technique is of great help in testing wheather a small sulphide mineral show (in out crop or in a bore hole) is an isolated occurance or whether it is a part of a large electrically contineous mass.

FIG II



MAP OF THE AREA SHOWING THE LOCATIONS
OF VARIOUS SOUNDING POINTS
SCALE:—16"=1 MILE
1 CM=40 METERS
40 0 40 80 120 160 M.
N
LAMEHTA GHAT
NARMADA RIVER
GHUNSORE
LOCAL DAM
SIVNI
WELL
NALA
WAY TO
VILLAGE SIVNI
•43

THE " MISE - À - LA - MASSE" METHOD

The name for a technique of electrical mapping was suggested by Schlumberger (1920) and may be translated on excitation of the mass. The idea is to use a subsurface conductive mass itself as one current electrode of a pair by connecting it directly to one pole of a battery, the second eletrode being as usual a metal rod but placed on the earth's surface at a great distance. A current is passed and the voltage at points on the surface are mapped by means of a volt-meter with respect to some base station.

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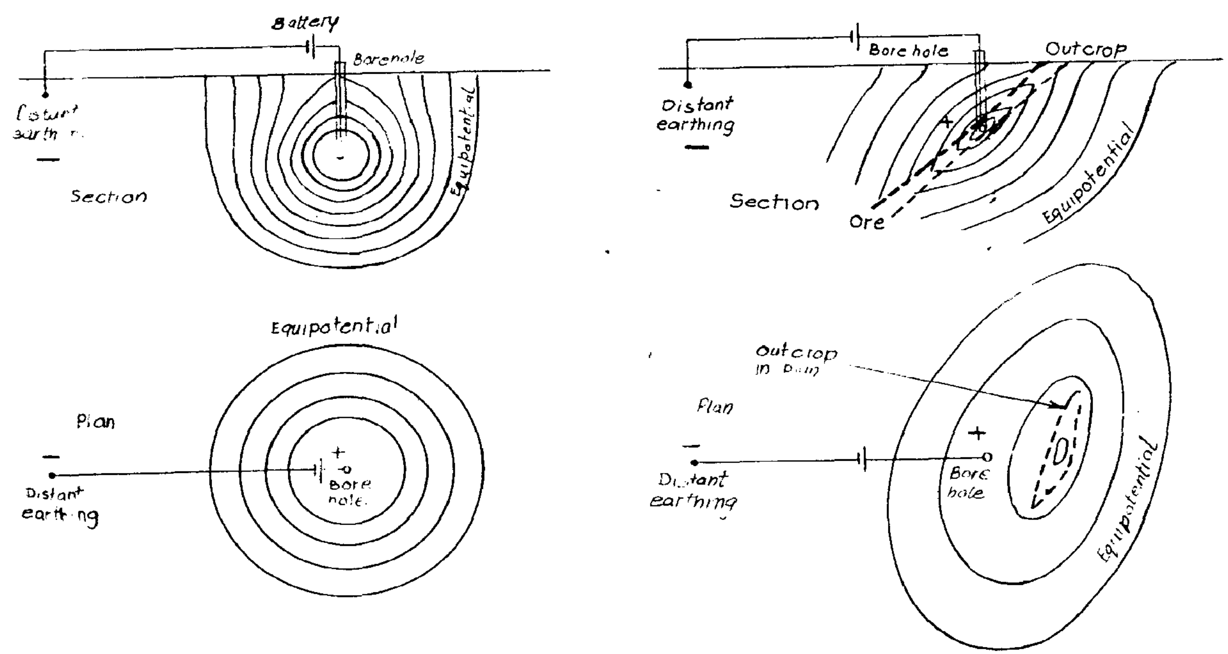


FIG. 12 PRINCIPLE OF THE MISE-À-LA-MASSE METHOD. RIGHT-HAND PART OF THE FIGURE SHOWS THE DISTORTION OF THE EQUIPOTENTIAL LINES DUE TO AN ORE.

C H A P T E R - V

PROCESSING, PRESENTATION AND DISCUSSION OF RESULTS

DATA PROCESSING, PRESENTATION AND DISCUSSION OF RESULTS

Data Processing and Presentation:

The readings taken in the field as described in Chapter IV have been processed to obtain apparent resistivities for different current electrode spreads. For this purpose computor programme was designed and the data computed. Some of the sheets are given in Appendix. These apparent resistivity values were plotted against the half the current electrode spreads on the transparent double logarithmic paper of modulus 62.5 mm. These curves were then matched with the Master curves published by European Association of Exploration Geophysics (1963) 2nd revised edition utilising the principle of complete curve matching as well as partial curve matching. The depth of interfaces and the resistivities thus obtained by this interpretation is shown in Fig 13 to 19 and in table No. 1.

Geoelectric sections were then drawn utilizing the interpreted data i.e. the resistivities and the depth to the interfaces and is shown in fig No. 20.

DISCUSSION OF RESULTS

The values of resistivities were determined and it was found that resistivity of the top layer varies from 6.2 to 35 ohm meter, while the value of resistivity determined with the help of Megger in field was found 8 to 40 ohm meter which are nearly same. The resistivity of the 2nd layer ranges from 1.5 to 12 ohm meter except few soundings obtained by interpretation of data, the determined value of resistivity in field with the help of Megger is 1.5 to 4.14 ohm meter in pit which is also coinciding. The places where the value of ρ_1 is informed to be high, the value of resistivity of 2nd layer is correspondingly high. Substratum has been considered to be of very high resistivity.

At places, the depth of interfaces is found unduly high as in line A'A, B'B, C'C etc..... at point VES No. 7, 22, 10 etc. which are likely due to burried nala, ρ_2 varying from 1.5 to 11.3 ohm meter.

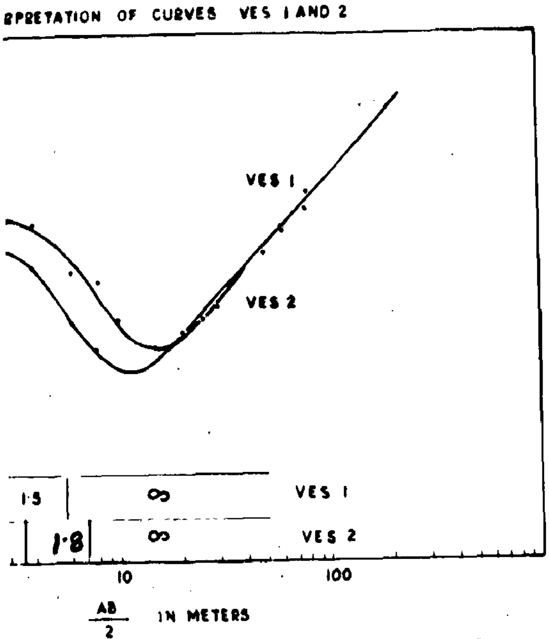
The course of the burried nala data is shown in fig 21. However this has also been reflected in the results of Mise-À-La-Masse method discussed in the next paragraph.

The material in the burried nala can likely be clayey soils along with boulders and water may be present in the pore spaces. At other places, very low resistivity values of the 2nd layer is found to correspond to the clay bed. The course of the burried nala is shown in the map (fig No. 21).

The equipotential map obtained in the Mise-À-La-Masse method is shown in fig No 22. However the course of burried nala has affected the disposition of equipotential contours. Thus from the equipotential contour, the course of the burried nala could be deciphered and is shown by dotted line in the fig. No 22. On the other hand the otherwise regular pattern of the equipotential lines indicate the regularity of the clay bed in the area investigated except in the nala portion.

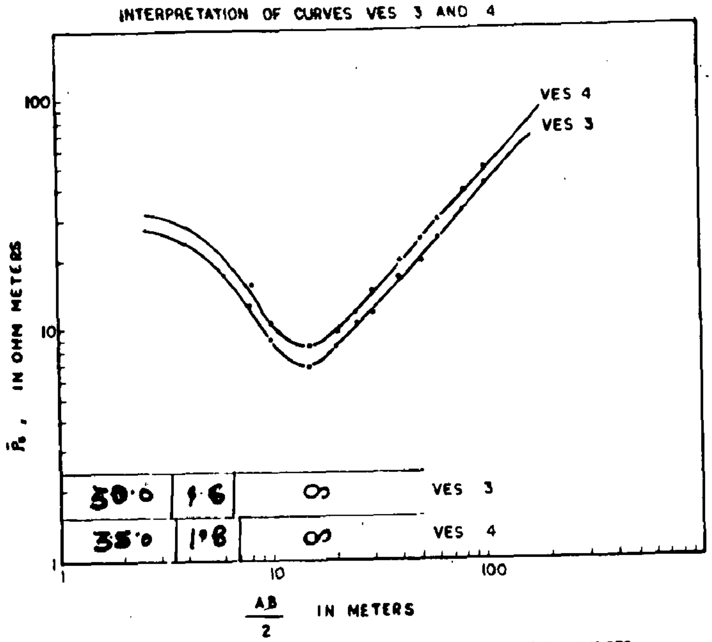
FIG 13

RPRETATION OF CURVES VES 1 AND 2



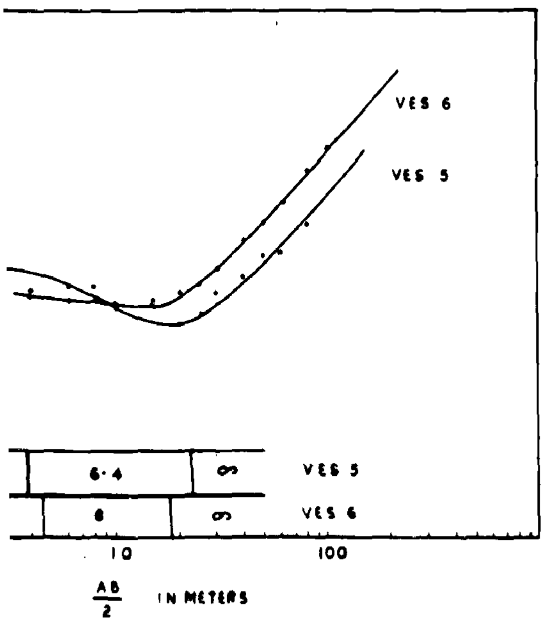
COLUMNS DESIGNATE TRUE RESISTIVITY IN OHM-METER.
LAYERS IN METER ARE PLOTTED USING THE LOGARITHMIC
'HE ABSCISSA AXIS.

INTERPRETATION OF CURVES VES 3 AND 4



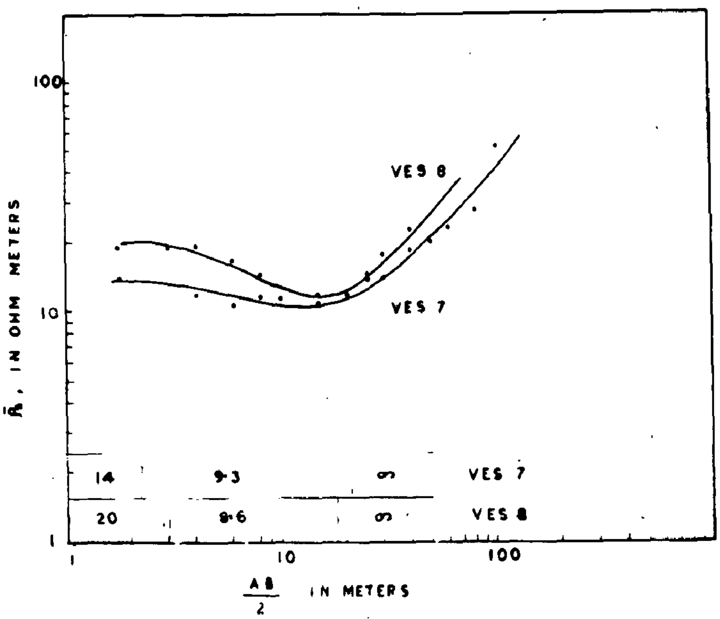
NUMBER IN COLUMNS DESIGNATE TRUE RESISTIVITY IN OHM-METER.
DEPTH TO LAYERS IN METERS ARE PLOTTED USING THE LOGARITHMIC
SCALE OF THE ABSCISSA AXIS.

ITERPETATION OF CURVES VES 5 AND 6



COLUMNS DESIGNATE TRUE RESISTIVITY IN OHM-METER.
YERS IN METERS ARE PLOTTED USING THE LOGARITHMIC
HE ABSCISSA AXIS.

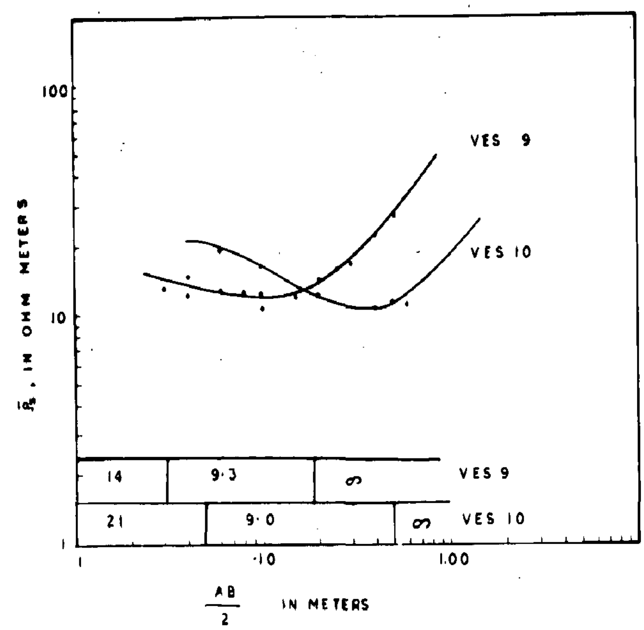
INTERPRETATION OF CURVES VES 7 AND 8



NUMBER IN COLUMNS DESIGNATE TRUE RESISTIVITY IN OHM-METER.
DEPTH TO LAYERS IN METERS ARE PLOTTED USING THE LOGARITHMIC
SCALE OF THE ABSCISSA AXIS.

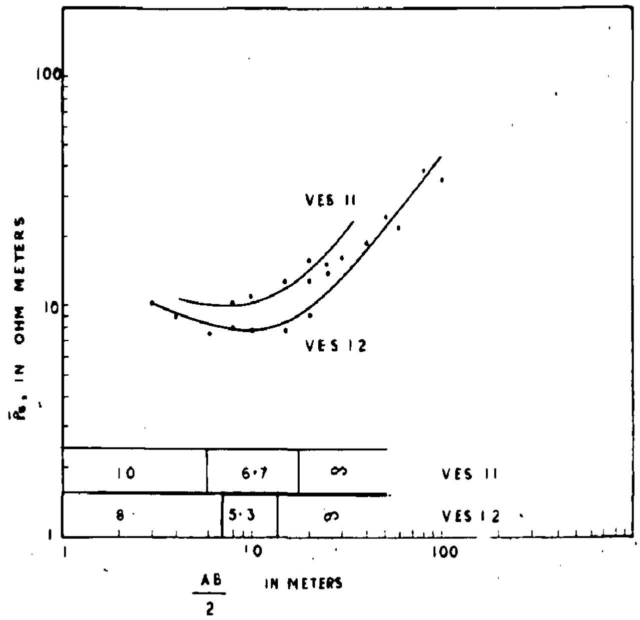
FIG 14

INTERPRETATION OF CURVES VES 9 AND 10



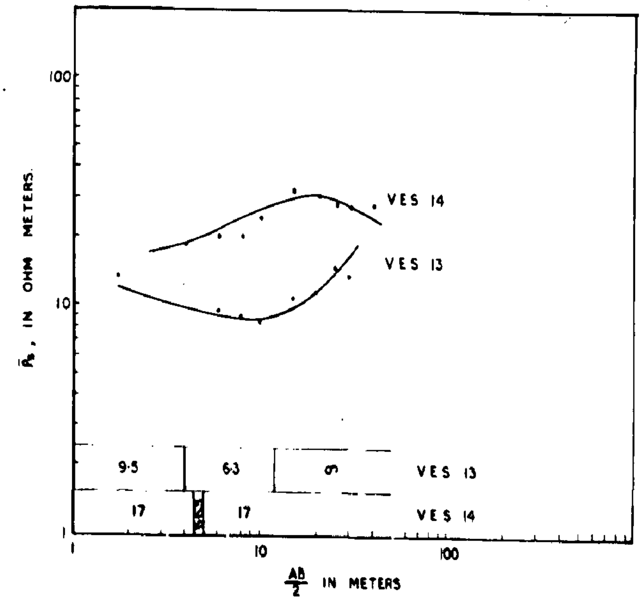
NUMBER IN COLUMNS DESIGNATE TRUE RESISTIVITY IN OHM-METER. DEPTH TO LAYERS IN METERS ARE PLOTTED USING THE LOGARITHMIC SCALE OF THE ABSCISSA AXIS.

INTERPRETATION OF CURVES VES 11 AND 12



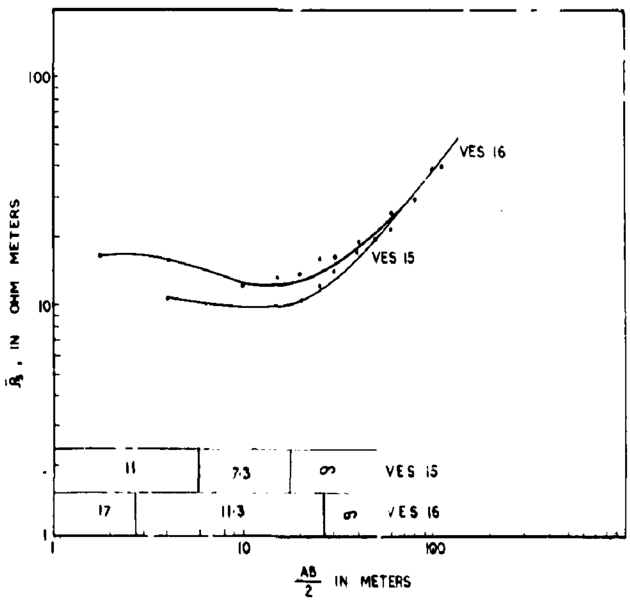
NUMER IN COLUMNS DESIGNATE TRUE RESISTIVITY IN OHM-METER. DEPTH TO LAYERS IN METERS ARE PLOTTED USING THE LOGARITHMIC SCALE OF THE ABSCISSA AXIS.

INTERPRETATION OF CURVES VES 13 AND 14



NUMBER IN COLUMNS DESIGNATE TRUE RESISTIVITY IN OHM-METER. DEPTH TO LAYERS IN METERS ARE PLOTTED USING THE LOGARITHMIC SCALE OF THE ABSCISSA AXIS

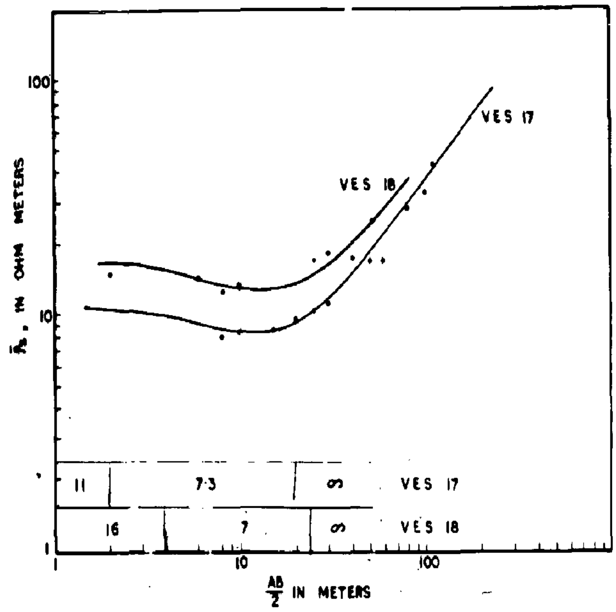
INTERPRETATION OF CURVES VES 15 AND 16



NUMBER IN COLUMNS DESIGNATE TRUE RESISTIVITY IN OHM-METER DEPTH TO LAYERS IN METERS ARE PLOTTED USING THE LOGARITHMIC SCALE OF THE ABSCISSA AXIS.

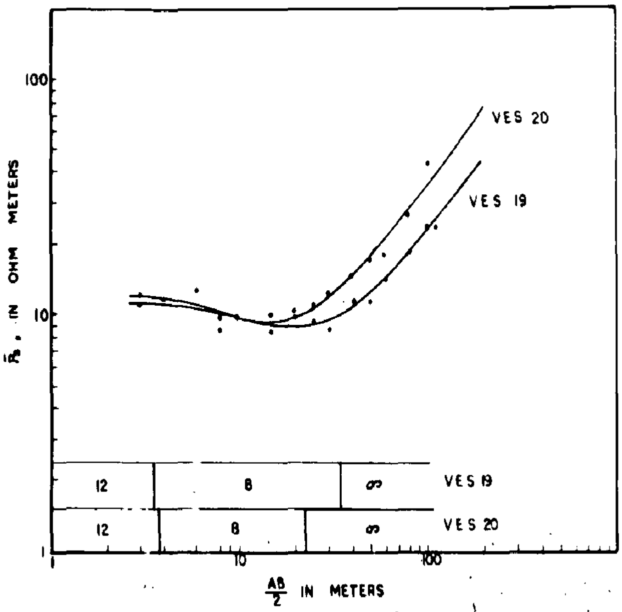
FIG. 15

INTERPRETATION OF CURVES VES 17 AND 18



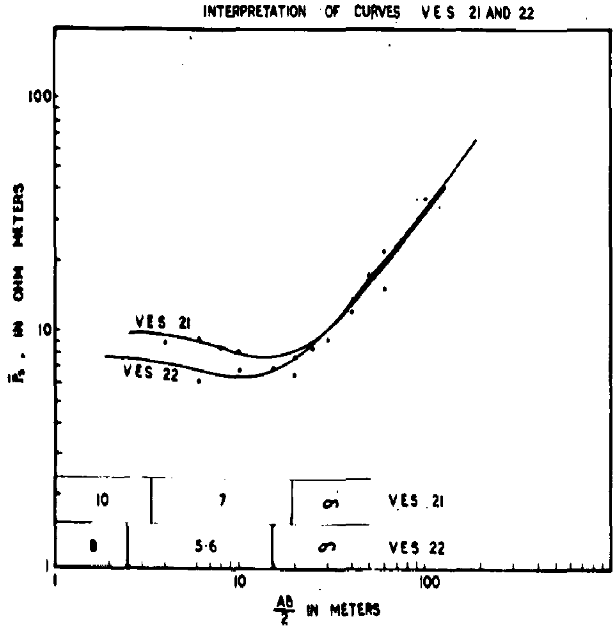
NUMBER IN COLUMNS DESIGNATE TRUE RESISTIVITY IN OHM-METER. DEPTH TO LAYERS IN METERS ARE PLOTTED USING THE LOGARITHMIC SCALE OF THE ABSCISSA AXIS.

INTERPRETATION OF CURVES VES 19 AND 20



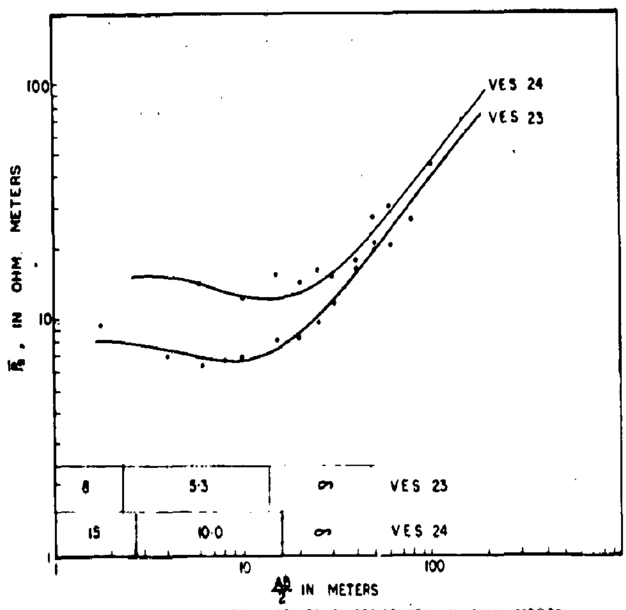
NUMBER IN COLUMNS DESIGNATE TRUE RESISTIVITY IN OHM-METER. DEPTH TO LAYERS IN METERS ARE PLOTTED USING THE LOGARITHMIC SCALE OF THE ABSCISSA AXIS.

INTERPRETATION OF CURVES VES 21 AND 22



NUMBER IN COLUMNS DESIGNATE TRUE RESISTIVITY IN OHM-METER. DEPTH TO LAYERS IN METERS ARE PLOTTED USING THE LOGARITHMIC SCALE OF THE ABSCISSA AXIS.

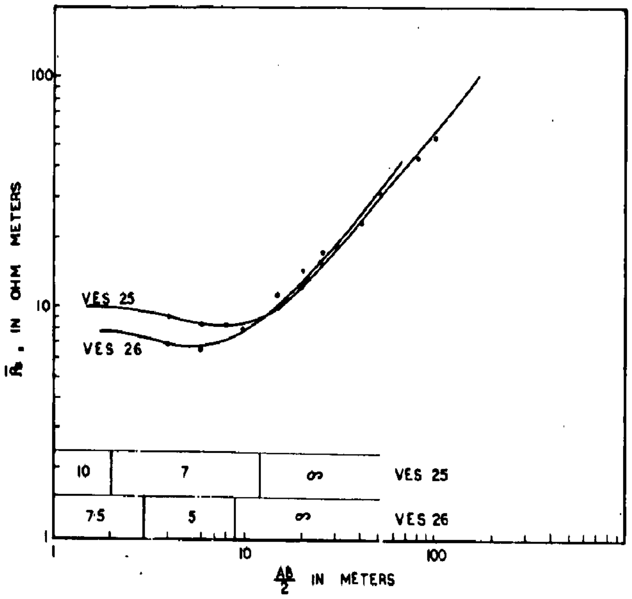
INTERPRETATION OF CURVES VES 23 AND 24



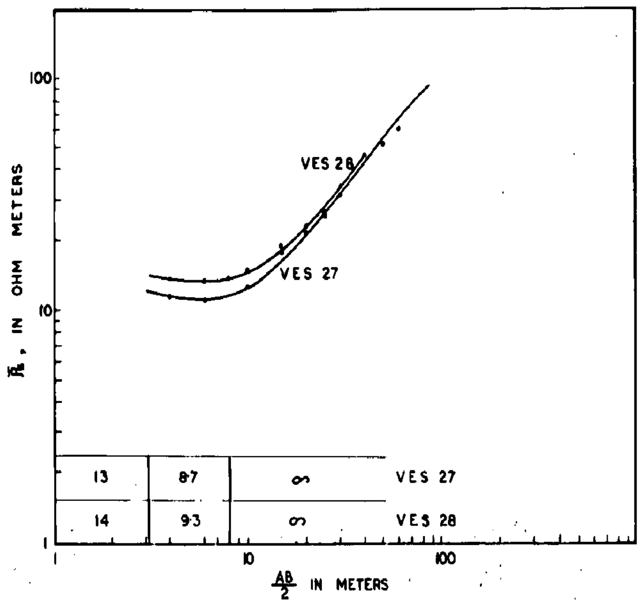
NUMBER IN COLUMNS DESIGNATE TRUE RESISTIVITY IN OHM-METER. DEPTH TO LAYERS IN METERS ARE PLOTTED USING THE LOGARITHMIC SCALE OF THE ABSCISSA AXIS.

FIG 16.

INTERPRETATION OF CURVES VES 25 AND 26

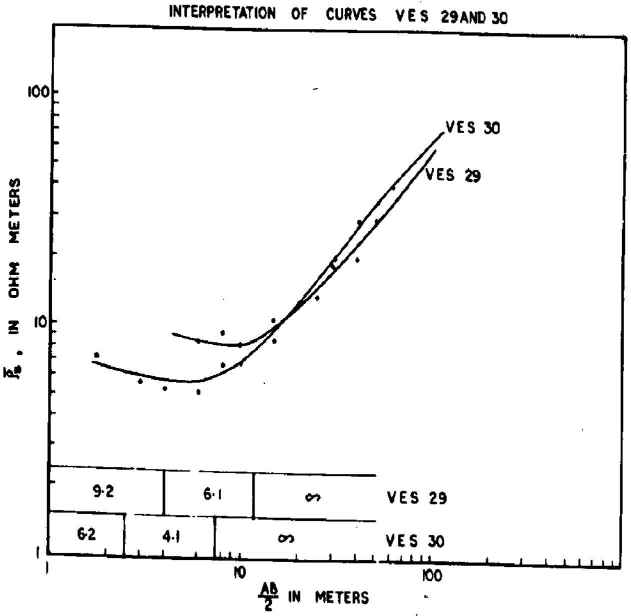


NUMBER IN COLUMNS DESIGNATE TRUE RESISTIVITY IN OHM-METER. DEPTH TO LAYERS IN METERS ARE PLOTTED USING THE LOGARITHMIC SCALE OF THE ABSCISSA AXIS.



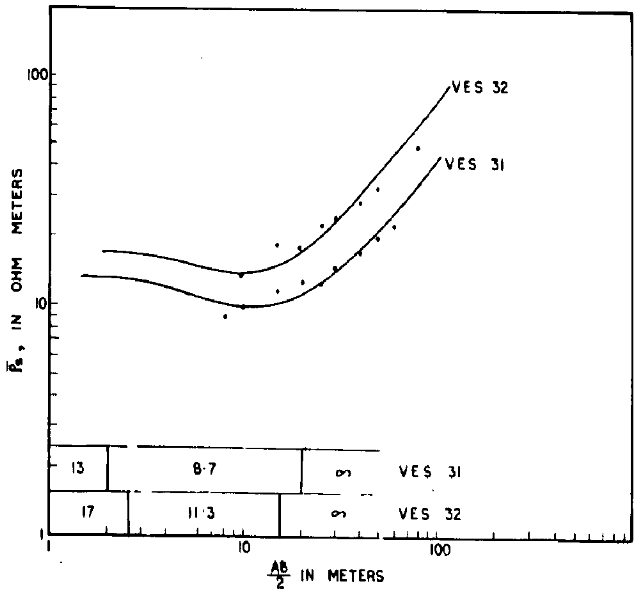
NUMBER IN COLUMNS DESIGNATE TRUE RESISTIVITY IN OHM-METER. DEPTH TO LAYERS IN METERS ARE PLOTTED USING THE LOGARITHMIC SCALE OF THE ABSCISSA AXIS.

INTERPRETATION OF CURVES VES 29 AND 30



NUMBER IN COLUMNS DESIGNATE TRUE RESISTIVITY IN OHM-METER. DEPTH TO LAYERS IN METERS ARE PLOTTED USING THE LOGARITHMIC SCALE OF THE ABSCISSA AXIS.

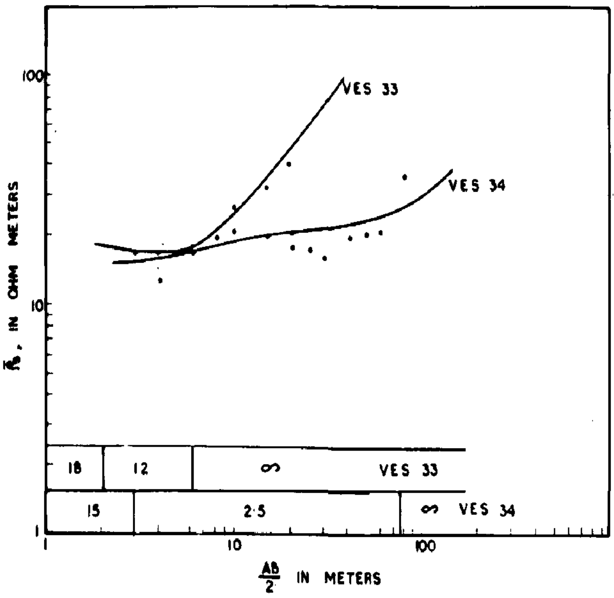
INTERPRETATION OF CURVES VES 31 AND 32



NUMBER IN COLUMNS DESIGNATE TRUE RESISTIVITY IN OHM-METER. DEPTH TO LAYERS IN METERS ARE PLOTTED USING THE LOGARITHMIC SCALE OF THE ABSCISSA AXIS.

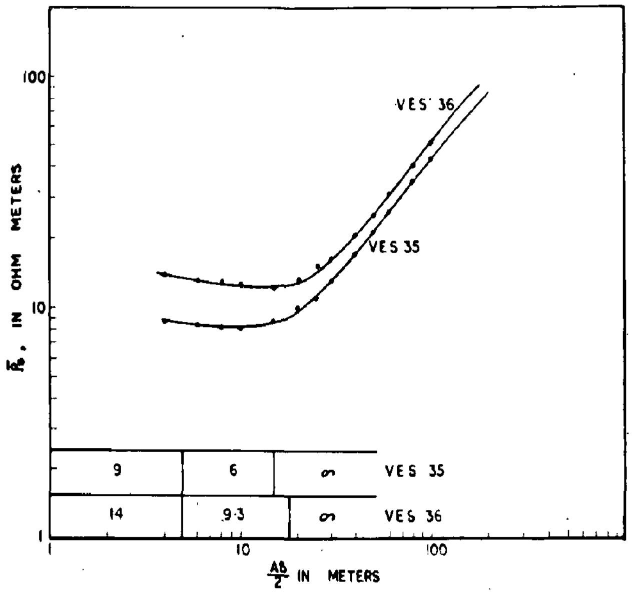
FIG 17

INTERPRETATION OF CURVES VES 33 AND 34



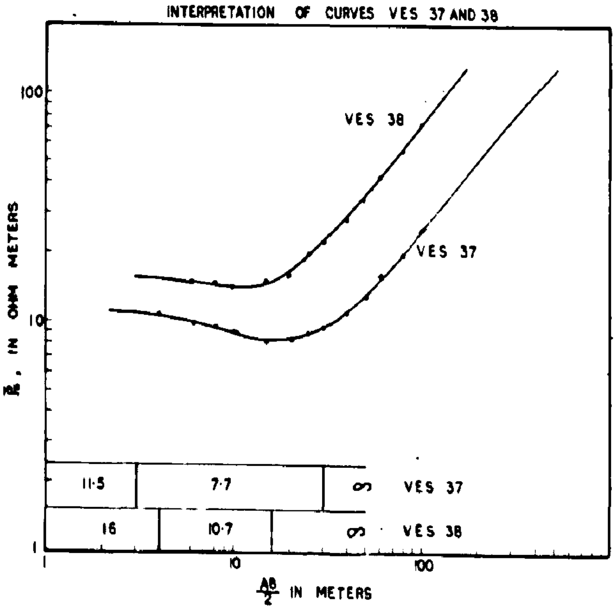
NUMBER IN COLUMNS DESIGNATE TRUE RESISTIVITY IN OHM - METER. DEPTH TO LAYERS IN METERS ARE PLOTTED USING THE LOGARITHMIC SCALE OF THE ABSCISSA AXIS.

INTERPRETATION OF CURVES VES 35 AND 36



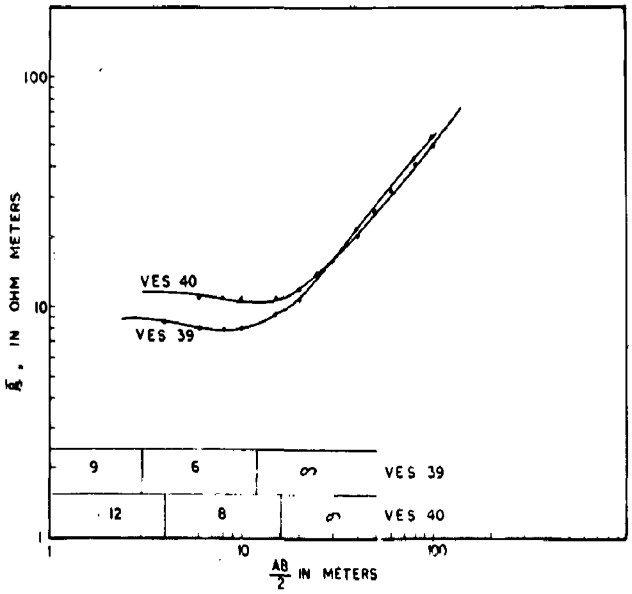
NUMBER IN COLUMNS DESIGNATE TRUE RESISTIVITY IN OHM - METER. DEPTH TO LAYERS IN METERS ARE PLOTTED USING THE LOGARITHMIC SCALE OF THE ABSCISSA AXIS.

INTERPRETATION OF CURVES VES 37 AND 38



NUMBER IN COLUMNS DESIGNATE TRUE RESISTIVITY IN OHM - METER. DEPTH TO LAYERS IN METERS ARE PLOTTED USING THE LOGARITHMIC SCALE OF THE ABSCISSA AXIS.

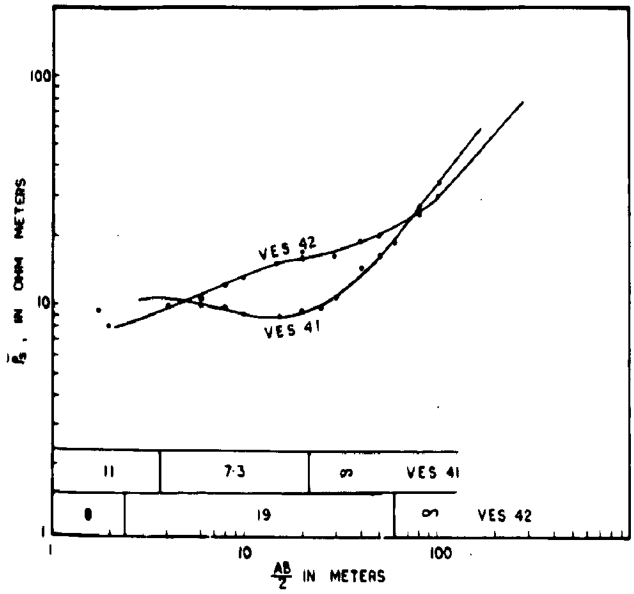
INTERPRETATION OF CURVES VES 39 AND 40



NUMBER IN COLUMNS DESIGNATE TRUE RESISTIVITY IN OHM - METER. DEPTH TO LAYERS IN METERS ARE PLOTTED USING THE LOGARITHMIC SCALE OF THE ABSCISSA AXIS.

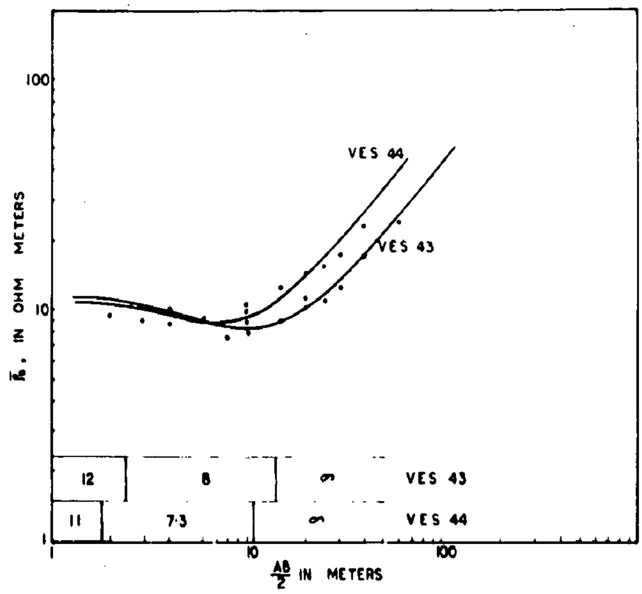
FIG 18

INTERPRETATION OF CURVES VES 41 AND 42



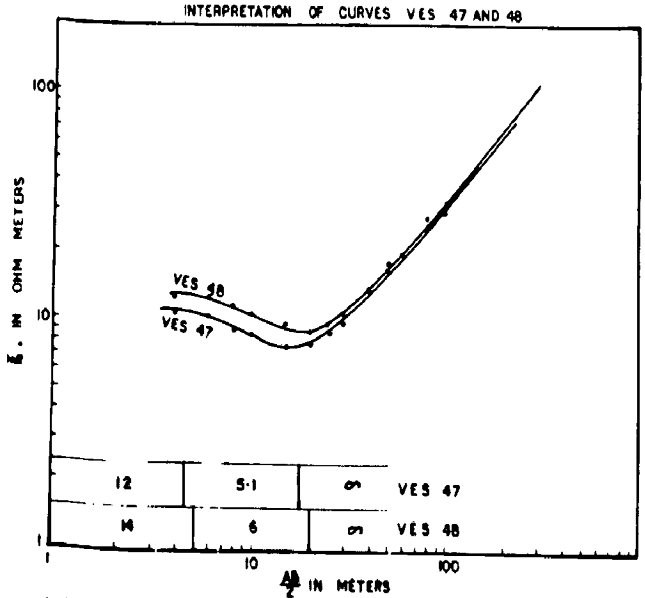
NUMBER IN COLUMNS DESIGNATE TRUE RESISTIVITY IN OHM - METER. DEPTH TO LAYERS IN METERS ARE PLOTTED USING THE LOGARITHMIC SCALE OF THE ABSCISSA AXIS.

INTERPRETATION OF CURVES VES 43 AND 44



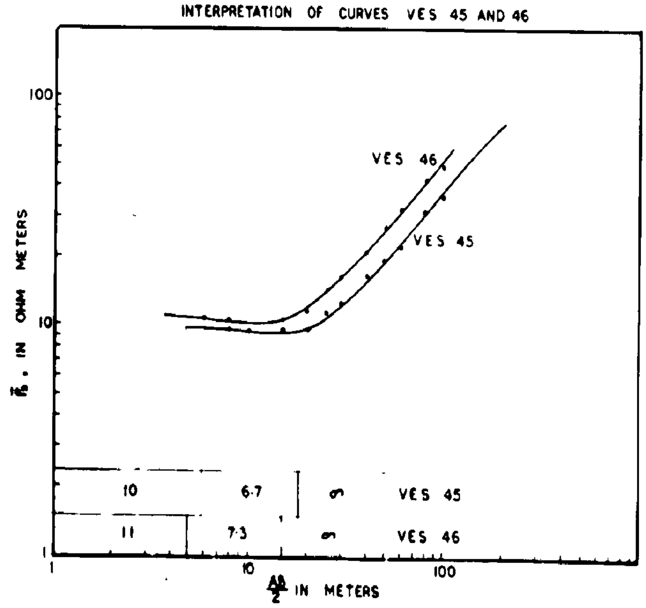
NUMBER IN COLUMNS DESIGNATE TRUE RESISTIVITY IN OHM-METER DEPTH TO LAYERS IN METERS ARE PLOTTED USING THE LOGARITHMIC SCALE OF THE ABSCISSA AXIS.

INTERPRETATION OF CURVES VES 47 AND 48



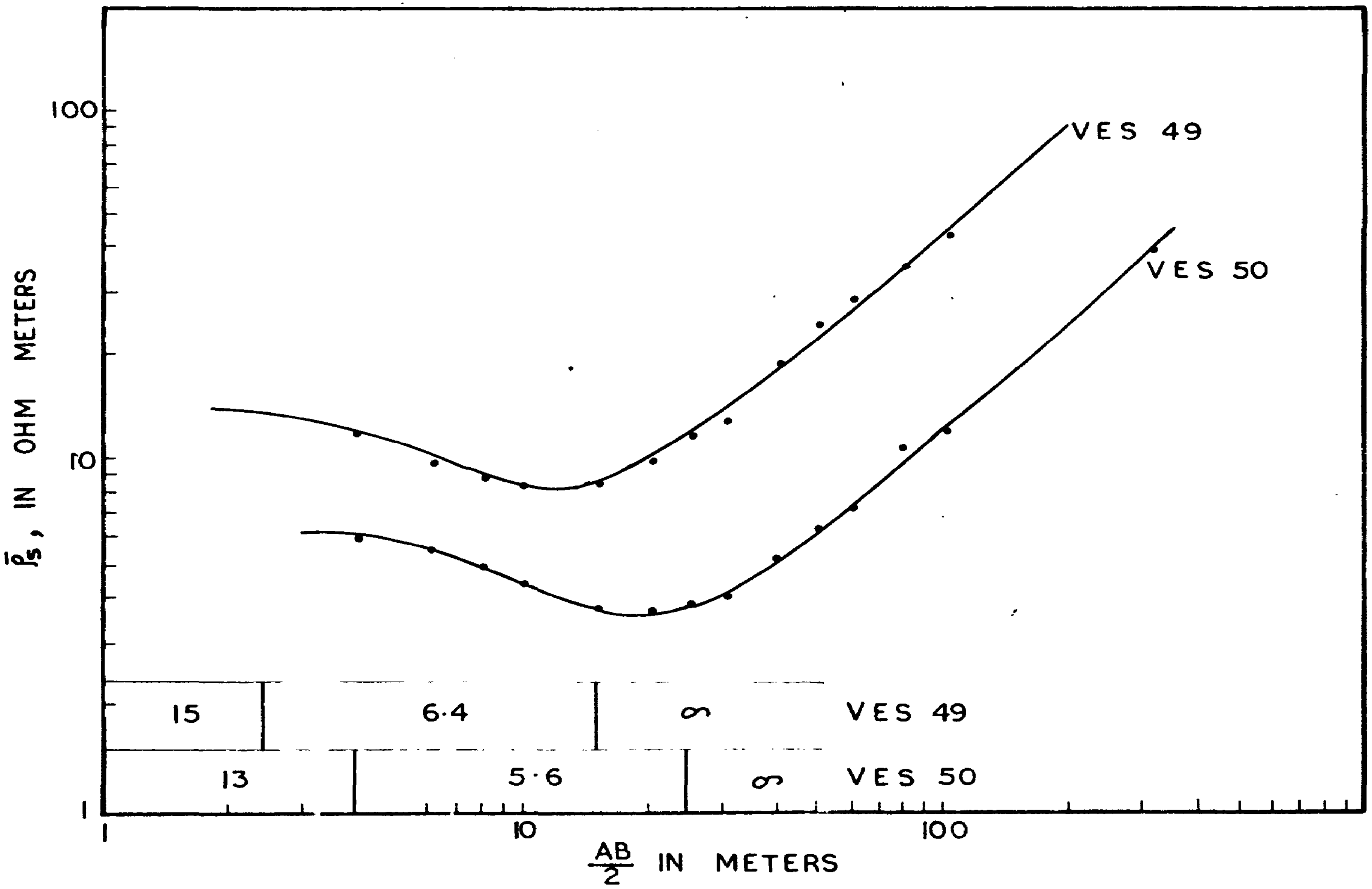
NUMBER IN COLUMNS DESIGNATE TRUE RESISTIVITY IN OHM - METER. DEPTH TO LAYERS IN METERS ARE PLOTTED USING THE LOGARITHMIC SCALE OF THE ABSCISSA AXIS.

INTERPRETATION OF CURVES VES 45 AND 46



NUMBER IN COLUMNS DESIGNATE TRUE RESISTIVITY IN OHM - METER. DEPTH TO LAYERS IN METERS ARE PLOTTED USING THE LOGARITHMIC SCALE OF THE ABSCISSA AXIS.

INTERPRETATION OF CURVES VES 49AND 50



NUMBER IN COLUMNS DESIGNATE TRUE RESISTIVITY IN OHM–METER. DEPTH TO LAYERS IN METERS ARE PLOTTED USING THE LOGARITHMIC SCALE OF THE ABSCISSA AXIS.

FIG . 19

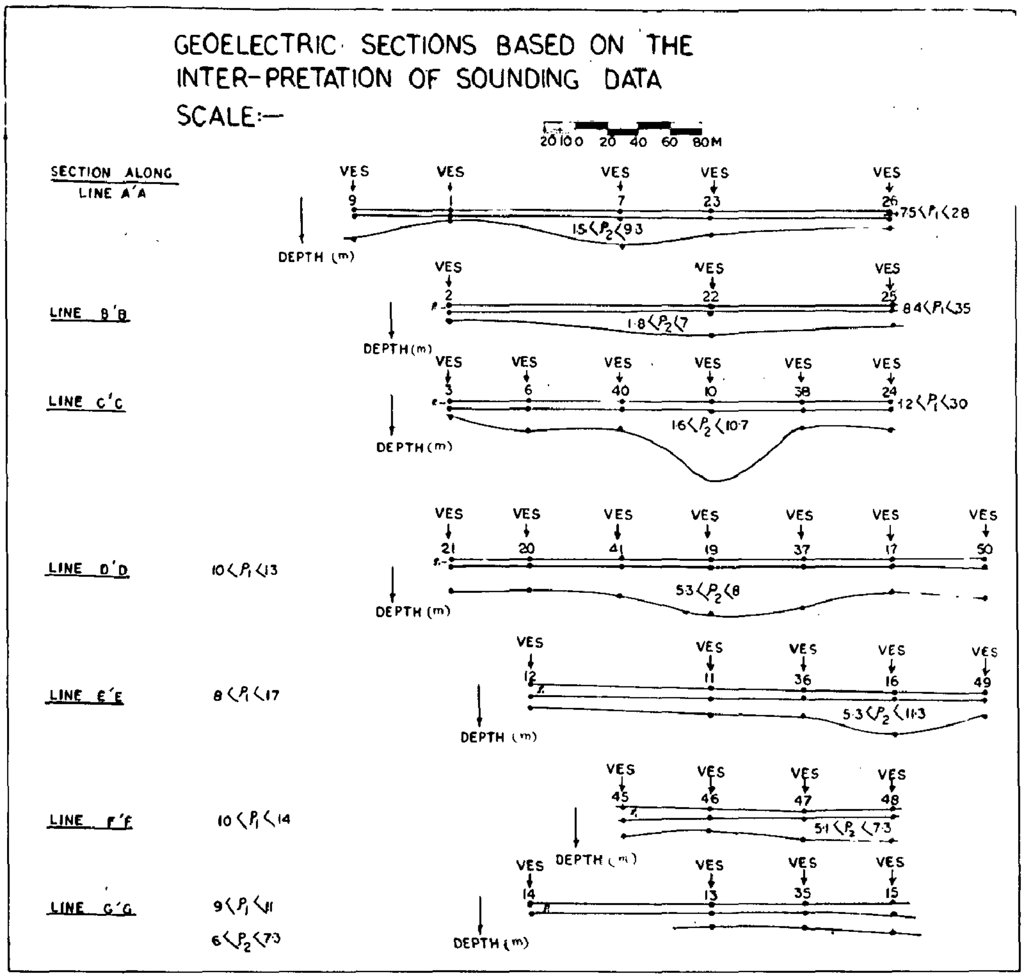
TABLE: 1

The following table shows resistivities and depth of Interfaces and elevation of sounding stations. P_1 and P_2 are the true resistivities in Ohm-meter of overburden and clay formation respectively, while h_1 and h_2 are the depth of the layers in meters.

V.E.S. No.	Elevation with reference to V.E.S. 1"level	P_1 (Ohm-m)	P_2 (Ohm-m)	h_1 (Meters)	h_2 (meters)
1	2	3	4	5	6
1	0	28	1.5	2.8	5.6
2.	0.5	35	1.8	3.5	7.0
3.	0.5	30	1.6	3.4	6.8
4.	0.5	35	1.8	3.5	7.0
5.	1.0	15	6.4	3.8	22.8
6.	1.5	12	8.0	4.5	18.0
7.	-0.5	14	9.3	2.2	21.0
8.	0	20	8.6	3.0	18.0
9.	0	14	9.3	3.0	18.0
10.	1.5	21	9.0	5.0	50.0
11.	3.0	10	6.7	6.0	18.0
12.	4.0	8	5.3	7.0	14.0
13.	1.0	9.5	6.3	4.0	12.0
14.	1.0	17.0	663	4.5	5.0
15.	3.0	11.0	7.3	6.0	18.0
16.	-0.5	17.0	11.3	2.7	27.0
17.	-1.0	11.0	7.3	2.0	20.0
18.	1.0	16.0	10.7	4.0	24.0
19.	0.5	12.0	8.0	3.5	35.0
20.	1.0	12.0	8.0	3.7	22.2
21.	1.0	10.0	7.0	3.5	20.0

1	2	3	4	5	6
22.	0	8.4	5.6	2.6	15.6
23.	0	8.0	5.3	2.3	13.8
24.	0	15.0	10.0	2.6	15.6
25.	0	10.0	7.0	2.0	12.0
26.	1.0	7.5	5.0	3.0	9.0
27.	1.0	13.0	8.7	3.0	9.0
28.	1.0	14.0	9.3	3.0	9.0
29.	1.0	9.2	6.1	4.0	12.0
30.	-0.5	6.2	4.1	2.5	7.5
31.	-0.5	13.0	8.7	2.0	20.0
32.	0	17.0	11.3	2.6	15.6
33.	-0.5	18.0	12.0	2.0	6.0
34.	0	15.0	22.5	3.0	75.0
35.	2.5	9.0	6.0	5.0	15.0
36.	2.5	14.0	9.3	5.0	18.0
37.	0	11.5	7.7	3.0	30.0
38.	0	16.0	10.7	4.0	16.0
39.	0	9.0	6.0	3.0	12.0
40.	1.0	12.0	8.0	4.0	16.0
41	0.5	11.0	7.3	3.6	21.6
42.	0	8.0	19.0	2.4	60.0
43.	-1.0	12.0	8.0	1.4	14.0
44.	0	11.0	7.3	1.8	10.8
45.	3.0	10.0	6.7	6.0	18.0
46.	2.0	11.0	7.3	5.0	15.0
47.	1.0	12.0	5.1	4.5	18.0
48	2.0	14.0	6.0	5.0	20.0
49.	-0.5	15.0	6.4	2.5	15.0
50	1.0	13.0	5.6	4.0	24.0

FIG 20


GEOELECTRIC SECTIONS BASED ON THE INTER-PRETATION OF SOUNDING DATA
SCALE:—
20 10 0 20 40 60 80M
SECTION ALONG LINE A'A
LINE B'B
LINE C'C
LINE D'D
10 < ρ1 < 13
LINE E'E
8 < ρ1 < 17
LINE F'F
10 < ρ1 < 14
LINE G'G
9 < ρ1 < 11
6 < ρ2 < 7.3
DEPTH (m)
.75 < ρ1 < 28
1.5 < ρ2 < 9.3
8.4 < ρ1 < 35
1.8 < ρ2 < 7
12 < ρ1 < 30
1.6 < ρ2 < 10.7
5.3 < ρ2 < 8
5.3 < ρ2 < 11.3
5.1 < ρ2 < 7.3

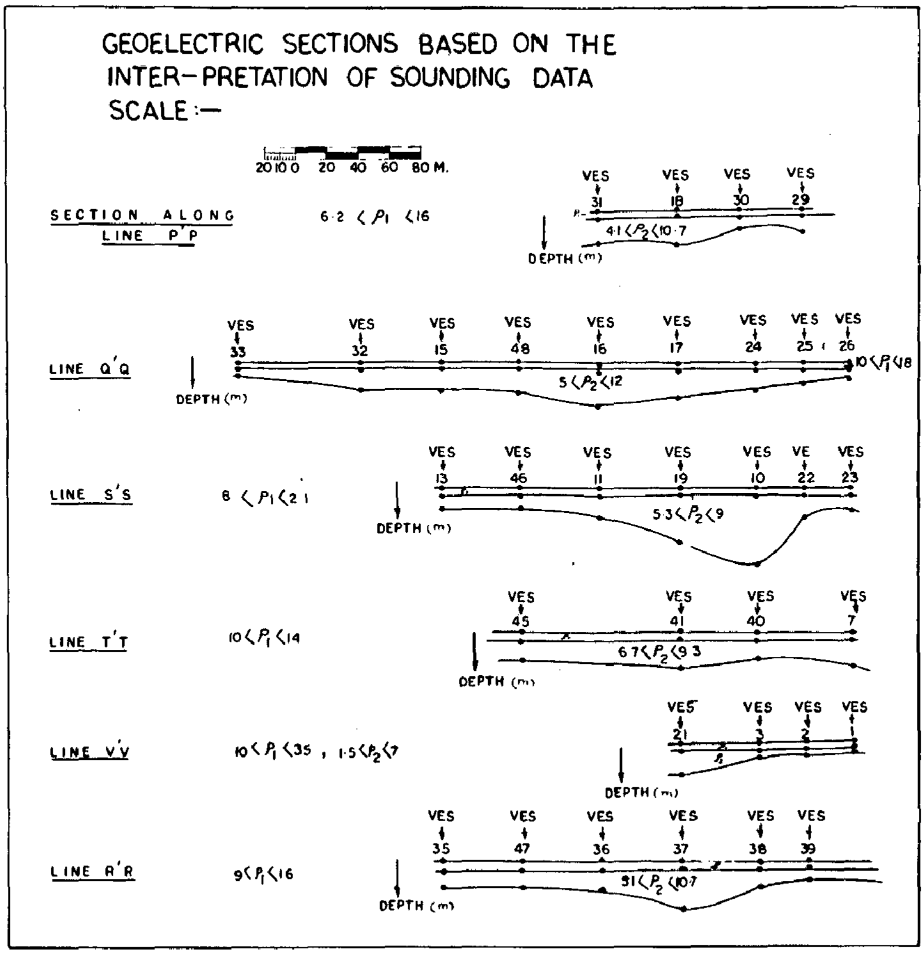
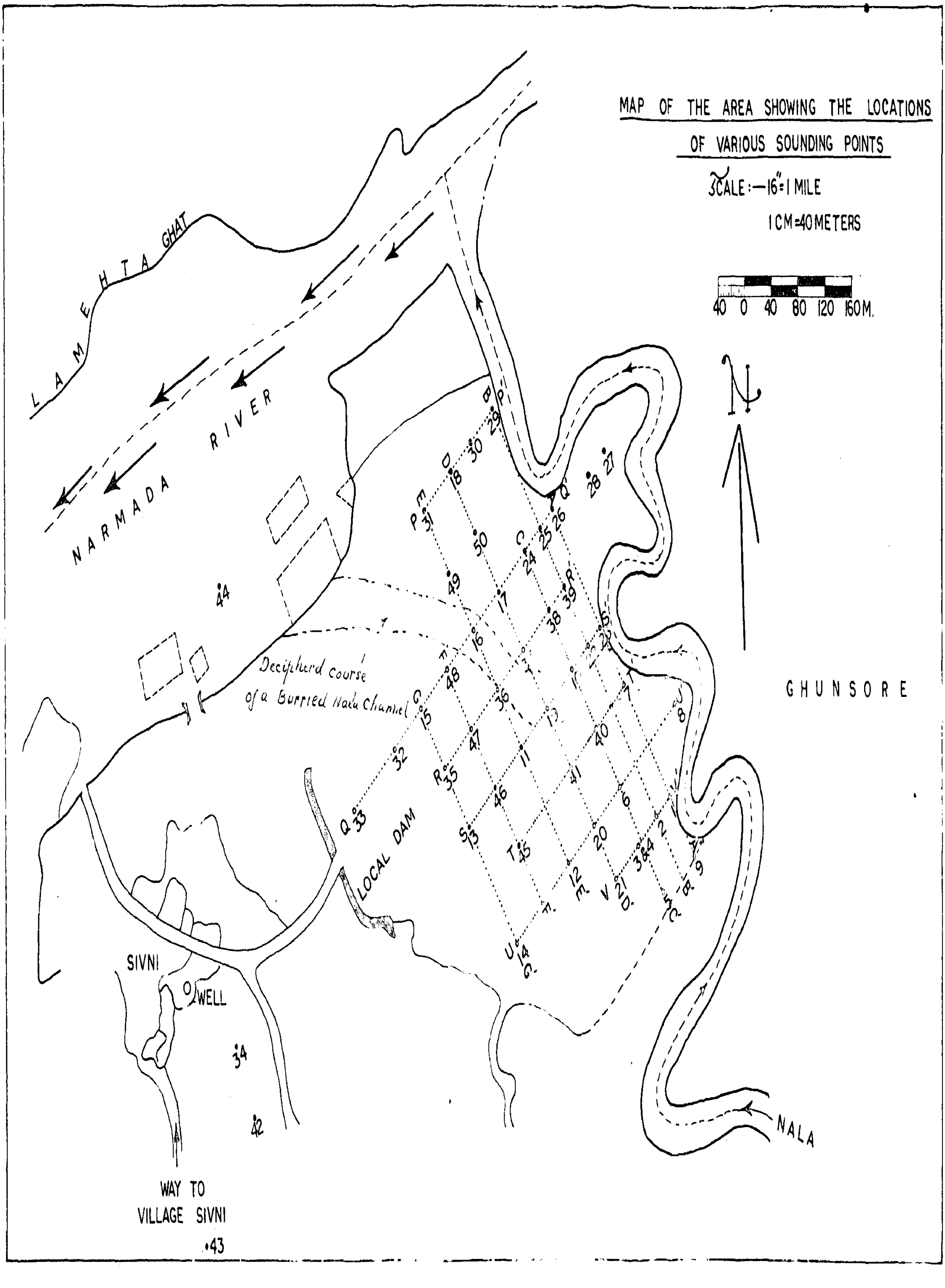

GEOELECTRIC SECTIONS BASED ON THE INTER-PRETATION OF SOUNDING DATA
SCALE:—
20 10 0 20 40 60 80 M.
SECTION ALONG LINE P'P
6.2 < ρ1 < 16
4.1 < ρ2 < 10.7
LINE Q'Q
10 < ρ1 < 18
5 < ρ2 < 12
LINE S'S
8 < ρ1 < 21
5.3 < ρ2 < 9
LINE T'T
10 < ρ1 < 14
6.7 < ρ2 < 9.3
LINE V'V
10 < ρ1 < 35 , 1.5 < ρ2 < 7
LINE R'R
9 < ρ1 < 16
3.1 < ρ2 < 10.7
DEPTH (m)

FIG 21

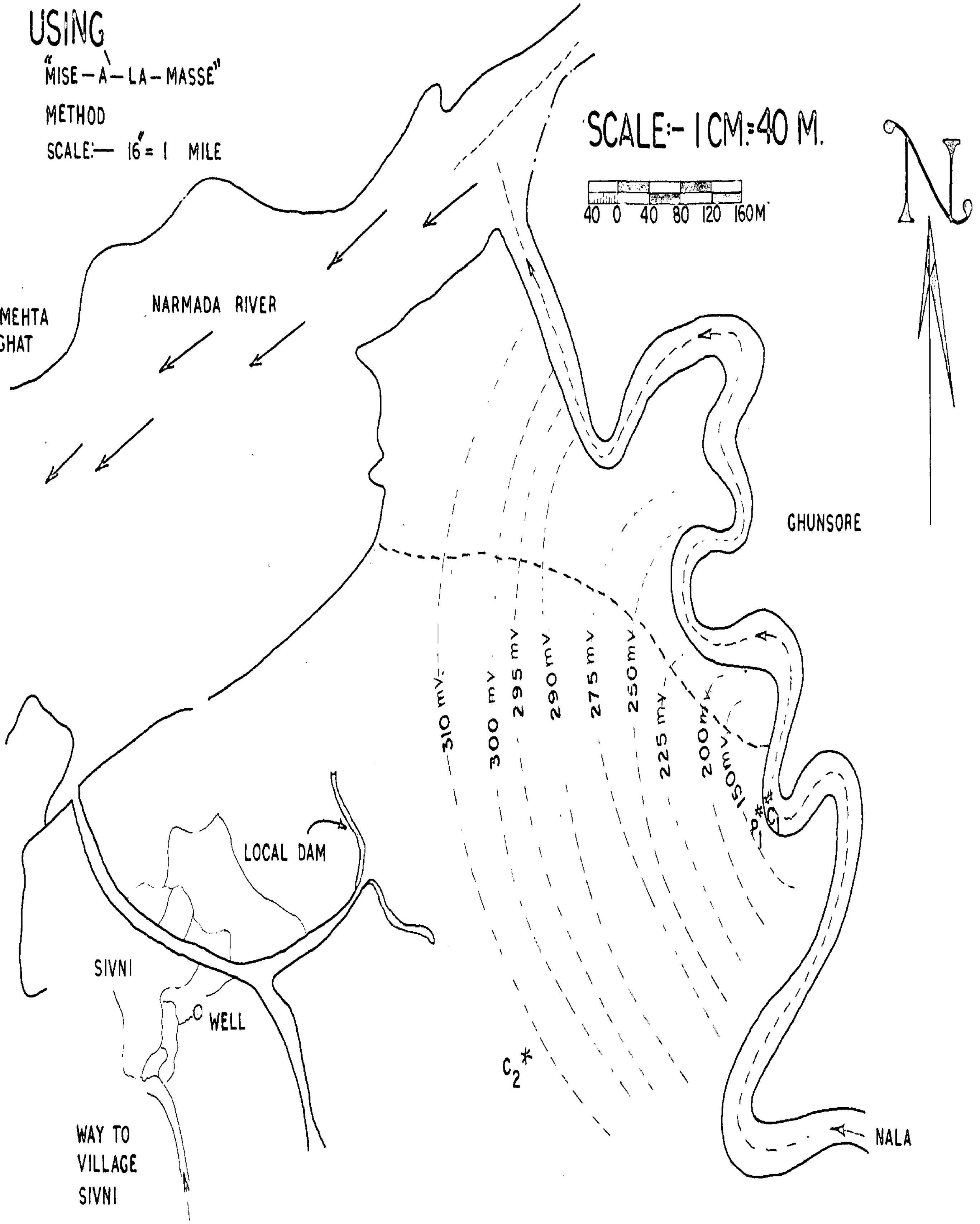


MAP OF THE AREA SHOWING THE LOCATIONS
OF VARIOUS SOUNDING POINTS
SCALE :— 16" = 1 MILE
1 CM = 40 METERS
40 0 40 80 120 160M.
N
LAMEHTA GHAT
NARMADA RIVER
GHUNSORE
Deciphered course of a Burried Nala Channel
LOCAL DAM
SIVNI
WELL
WAY TO VILLAGE SIVNI
NALA

FIG. 22

EQUIPOTENTIAL MAP OF THE AREA USING "MISE-À-LA-MASSE" METHOD

SCALE:— 16" = 1 MILE



C_1 AND C_2 - POSITON OF CURRENT ELECTRODES

P_1 — POSITION OF THE FIXED POTENTIAL ELECTRODE

-.-.-.- Alignment of the deciphered burried nala channel.

CHAPTER - VI

SUMMARY AND CONCLUSION

SUMMARY AND CONCLUSION

Resistivity survey was carried out to demarcate commercial clays. For this purpose depth sounding were taken at predetermined locations. These sounding curves were interpreted by the technique of partial and complete curve matching. Then geogical section were drawn. These geologic sections were found matching with the Geology in the region.

A burried river valley has also been demarcated on the basis of resistivity data (fig. 21)

Prospecting by Mise-A-La-Masse Method was also carried out with the aim of demarcating the boundaries of clay beds. The uniform pattern (as shown in fig. 22) of the equipotential lines suggest that the boundaries lies out side the area of measurements.

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APPENDIX

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C  C  TYAGI ARRANGEMENT SCHIUMBERGER 27474 CCR
C     A=AB(M)/2.
C     SK=SPECING FACTOR
C     DRI=DIRECT READING I(MA)
C     DRV=DIRECT READING V(MV)
C     RRI=REVERSE READING I(MA)
C     RRV=REVERSE READING V(MV)
C     DVI=DIRECT V/I (DRV/DRI)
C     RVI=REVERSE V/I (RRV/RRI)
C     R=MEAN V/I (DVI+RVI)/2.
C     ROH=SK*R OHM.M.
      DO 11 K=1,5
      PUNCH 9
9     FORMAT(10X,59HDEPARTMENT OF GEOLOGY AND GEOPHYSICS, UNIVERSITY OF
     1ROORKEE)
      PUNCH 16
16    FORMAT(36X,7HROORKEE)
      PUNCH 17
17    FORMAT(4X,14HDATA SHEET NO.,25X,5HDATED)
      PUNCH 12
12    FORMAT(43X,8HLOCATION)
      PUNCH 13
13    FORMAT(15X,24HARRANGEMENT SCHLUMBERGER)
      PUNCH 14
14    FORMAT(/,2HNO,3X,1HA,6X,2HSK,5X,3HDRI,4X,3HDRV,4X,3HRRI,4X,3HRRV,4
     1X,3HDVI,4X,3HRVI,5X,1HR,5X,3HROH,/)
2     READ1,I,A,SK,DRI,DRV,RRI,RRV
      DVI=DRV/DRI
      RVI=RRV/RRI
      R=.5*(DVI+RVI)
      ROH=SK*R
      PUNCH 1,I,A,SK,DRI,DRV,RRI,RRV,DVI,RVI,R,ROH
1     FORMAT(I2,10F7.3)
      IF(I-19)10,11,11
10    GO TO 2
11    CONTINUE
      STOP
      END
```

DEPARTMENT OF GEOLOGY AND GEOPHYSICS, UNIVERSITY OF ROORKEE
ROORKEE

DATA SHEET NO. 1 DATED

LOCATION

ARRANGEMENT SCHLUMBERGER

NO	A	SK	DRI	DRV	RRI	RRV	DVI	RVI	R	ROH
1	1.500	6.280	27.000	324.500	27.000	311.000	12.019	11.519	11.769	73.906
2	2.000	11.770	24.000	90.200	24.000	93.000	3.758	3.875	3.817	44.922
3	3.000	27.470	38.500	31.000	39.500	54.500	.805	1.380	1.092	30.010
4	4.000	49.450	54.000	18.000	55.000	23.000	.333	.418	.376	18.581
5	4.000	23.350	54.000	38.600	54.000	40.000	.715	.741	.728	16.994
6	6.000	54.050	48.000	9.400	48.000	10.000	.196	.208	.202	11.104
7	8.000	98.910	50.000	4.300	50.000	4.300	.086	.086	.086	8.506
8	10.000	155.470	55.000	3.000	55.000	5.000	.055	.091	.073	11.307
9	10.000	75.360	54.000	6.500	54.000	10.000	.120	.185	.153	11.513
10	15.000	173.480	75.000	3.900	62.000	3.100	.052	.050	.051	8.847
11	20.000	310.860	124.000	3.500	124.000	5.800	.028	.047	.037	11.657
12	20.000	117.750	124.000	8.200	124.000	9.500	.066	.077	.071	8.404
13	25.000	188.500	140.000	9.000	140.000	7.500	.064	.054	.059	11.108
14	30.000	274.250	164.000	6.500	164.000	8.100	.040	.049	.045	12.207
15	40.000	494.550	116.000	3.000	116.000	3.000	.026	.026	.026	12.790
16	50.000	777.150	72.000	2.000	72.000	2.000	.028	.028	.028	21.587
17	50.000	376.500	148.000	7.200	148.000	11.200	.049	.076	.062	23.404
18	60.000	549.500	200.000	7.700	193.000	10.500	.039	.054	.046	25.525
19	80.000	989.100	230.000	6.300	230.000	10.000	.027	.043	.035	35.049

DEPARTMENT OF GEOLOGY AND GEOPHYSICS, UNIVERSITY OF ROORKEE ROORKEE

DATA SHEET NO. 2

DATED

LOCATION

ARRANGEMENT SCHLUMBERGER

NO	A	SK	DRI	DRV	RRI	RRV	DVI	RVI	R	ROH
1	1.500	6.280	65.000	347.000	63.000	347.000	5.338	5.508	5.423	34.058
2	2.000	11.770	61.000	181.700	61.000	186.000	2.979	3.049	3.014	35.474
3	3.000	27.470	110.000	127.600	110.000	125.500	1.160	1.141	1.150	31.603
4	4.000	49.450	132.000	82.000	133.000	82.000	.621	.617	.619	30.603
5	4.000	23.350	134.000	178.000	134.000	178.000	1.328	1.328	1.328	31.017
6	6.000	54.950	136.000	43.000	136.000	49.400	.316	.363	.340	18.667
7	8.000	98.910	130.000	21.000	130.000	21.000	.162	.162	.162	15.978
8	10.000	155.470	120.000	8.000	120.000	9.300	.067	.078	.072	11.207
9	10.000	75.360	120.000	20.000	120.000	16.200	.167	.135	.151	11.367
10	15.000	173.480	116.000	6.100	116.000	6.100	.053	.053	.053	9.123
11	20.000	310.860	130.000	40.000	128.000	40.000	.308	.313	.310	96.396
12	20.000	117.750	128.000	1.400	126.000	11.000	.011	.087	.049	5.784
13	25.000	188.500	156.000	9.500	156.000	9.300	.061	.060	.060	11.358
14	30.000	274.250	118.000	5.200	124.000	6.800	.044	.055	.049	13.563
15	40.000	494.550	138.000	5.000	138.000	5.800	.036	.042	.039	19.352
16	50.000	777.150	122.000	3.000	120.000	4.500	.025	.038	.031	24.127
17	50.000	376.500	270.000	12.400	275.000	17.500	.046	.064	.055	20.625
18	60.000	549.500	225.000	8.000	225.000	14.200	.036	.063	.049	27.109
19	80.000	989.100	215.000	6.400	215.000	11.000	.030	.051	.040	40.024

DEPARTMENT OF GEOLOGY AND GEOPHYSICS, UNIVERSITY OF ROORKEE
ROORKEE

DATA SHEET NO. 3 DATED

LOCATION

ARRANGEMENT SCHLUMBERGER

NO	A	SK	DRI	DRV	RRI	RRV	DVI	RVI	R	ROH
1	1.500	6.280	138.000	243.000	132.000	243.700	1.761	1.846	1.804	11.326
2	2.000	11.770	172.000	122.700	172.000	142.700	.713	.830	.772	9.081
3	3.000	27.470	144.000	153.000	140.000	150.300	1.063	1.074	1.068	29.339
4	4.000	49.470	160.000	28.400	160.000	31.400	.178	.196	.187	9.245
5	4.000	23.350	158.000	63.600	158.000	61.000	.403	.386	.394	9.207
6	6.000	54.950	122.000	20.500	122.000	27.400	.168	.225	.196	10.787
7	8.000	98.910	128.000	14.500	124.000	13.500	.113	.109	.111	10.987
8	10.000	155.470	140.000	10.700	140.000	9.400	.076	.067	.072	11.161
9	10.000	75.360	140.000	24.000	140.000	20.000	.171	.143	.157	11.842
10	15.000	173.480	152.000	10.800	152.000	10.000	.071	.066	.068	11.870
11	20.000	310.860	136.000	5.400	136.000	6.100	.040	.045	.042	13.143
12	20.000	117.750	280.000	26.700	275.000	25.500	.095	.093	.094	11.073
13	25.000	188.500	300.000	16.800	295.000	15.200	.056	.052	.054	10.134
14	30.000	274.250	300.000	12.680	300.000	9.500	.042	.032	.037	10.138
15	40.000	494.550	330.000	8.000	330.000	8.000	.024	.024	.024	11.989
16	50.000	777.150	250.000	3.600	250.000	4.000	.014	.016	.015	11.813
17	50.000	376.500	250.000	10.500	250.000	6.000	.042	.024	.033	12.425
18	60.000	549.500	240.000	6.200	240.000	6.200	.026	.026	.026	14.195
19	80.000	989.100	245.000	6.000	245.000	5.200	.024	.025	.025	24.627

DEPARTMENT OF GEOLOGY AND GEOPHYSICS, UNIVERSITY OF ROORKEE

ROORKEE

DATA SHEET NO. 4 DATED

LOCATION

ARRANGEMENT SCHLUMBERGER

NO	A	SK	DRI	DRV	RRI	RRV	DVI	RVI	R	ROH
1	1.500	6.280	72.000	140.000	72.000	142.200	1.944	1.975	1.960	12.307
2	2.000	11.770	70.000	71.000	70.000	74.000	1.014	1.057	1.036	12.190
3	3.000	27.470	52.000	21.000	53.000	25.000	.404	.472	.438	12.026
4	4.000	49.450	46.000	8.500	48.000	12.100	.185	.252	.218	10.802
5	4.000	23.350	47.000	21.600	48.000	25.000	.460	.521	.490	11.446
6	6.000	54.950	53.000	13.000	52.000	14.800	.245	.285	.265	14.559
7	8.000	98.910	57.000	19.800	61.000	13.200	.347	.216	.282	27.881
8	10.000	155.470	50.000	6.000	51.000	7.400	.120	.145	.133	20.607
9	10.000	75.360	50.000	18.000	51.000	25.000	.360	.490	.425	32.035
10	15.000	173.480	52.000	4.400	52.000	5.700	.085	.110	.097	16.848
11	20.000	310.860	124.000	5.100	124.000	6.200	.041	.050	.046	14.164
12	20.000	117.750	128.000	12.000	124.000	14.000	.094	.113	.103	12.167
13	25.000	188.500	112.000	6.300	112.000	7.700	.056	.069	.063	11.781
14	30.000	274.250	108.000	5.000	108.000	5.000	.046	.046	.046	12.697
15	40.000	494.550	96.000	3.600	96.000	2.800	.038	.029	.033	16.485
16	50.000	777.150	72.000	2.700	72.000	4.400	.038	.061	.049	38.318
17	50.000	376.500	72.000	3.700	72.000	3.000	.051	.042	.047	17.518
18	60.000	549.500	96.000	3.000	96.000	4.400	.031	.046	.039	21.179
19	80.000	989.100	132.000	3.300	132.000	3.200	.025	.024	.025	24.353

DEPARTMENT OF GEOLOGY AND GEOPHYSICS, UNIVERSITY OF ROORKEE
ROORKEE

DATA SHEET NO. 5

DATED
LOCATION

ARRANGEMENT SCHLUMBERGER

NO	A	SK	DRI	DRV	RRI	RRV	DVI	RVI	R	ROH
1	1.500	6.280	152.000	384.400	150.000	378.300	2.529	2.522	2.525	15.860
2	2.000	11.770	150.000	378.300	160.000	189.100	2.522	1.182	1.852	21.797
3	3.000	27.470	126.000	69.200	130.000	69.200	.549	.532	.541	14.855
4	4.000	49.450	130.000	36.200	130.000	38.300	.278	.295	.287	14.169
5	4.000	23.350	130.000	70.000	130.000	70.100	.538	.539	.539	12.582
6	6.000	54.950	126.000	28.200	126.000	28.800	.224	.229	.226	12.429
7	8.000	98.910	116.000	13.500	114.000	14.500	.116	.127	.122	12.046
8	10.000	155.470	190.000	13.100	185.000	13.100	.069	.071	.070	10.864
9	10.000	75.360	185.000	28.400	185.000	23.600	.154	.128	.141	10.591
10	15.000	173.480	100.000	4.800	25.000	6.800	.048	.272	.160	27.757
11	20.000	310.860	170.000	3.800	160.000	5.600	.022	.035	.029	8.914
12	20.000	117.750	165.000	11.100	158.000	11.100	.067	.070	.069	8.097
13	25.000	188.500	211.000	10.700	205.000	10.700	.051	.052	.051	9.699
14	30.000	274.250	126.000	4.500	125.000	6.300	.036	.050	.043	11.808
15	40.000	494.550	210.000	7.700	205.000	3.600	.037	.018	.027	13.409
16	50.000	777.150	215.000	4.400	200.000	5.300	.020	.027	.023	18.249
17	50.000	376.500	220.000	3.800	225.000	3.500	.017	.016	.016	6.180
18	60.000	549.500	185.000	5.400	185.000	6.400	.029	.035	.032	17.525
19	80.000	989.100	205.000	4.900	205.000	4.900	.024	.024	.024	23.642