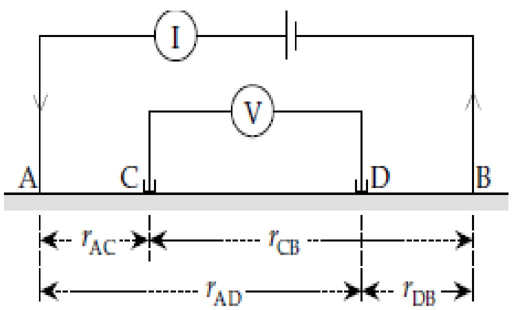
Ministry of Higher Education and Scientific Research

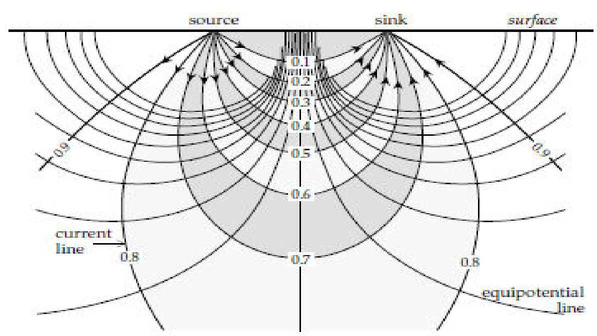
Al-Karkh University of Science

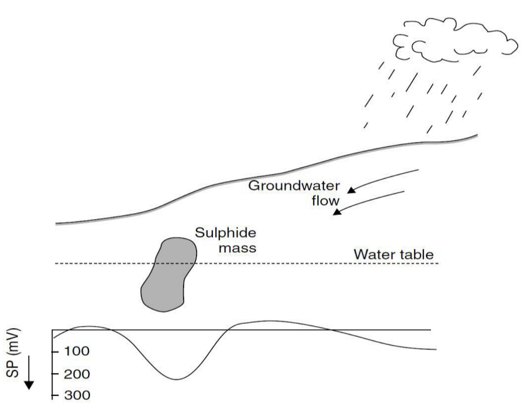
College of Geophysics and Remote Sensing

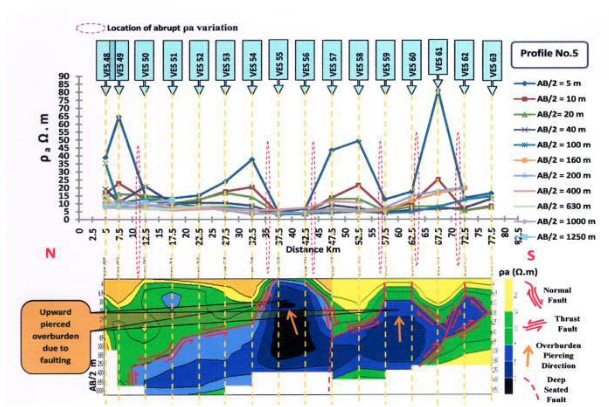
Department of Geophysics

Electrical Geophysical Methods









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Ph.D. In Geophysics

Introduction

Electrical methods represent one of four principal groups of geophysical exploration techniques. The other three are: seismic, gravimetric and magnetic. The latter two differ in that they depend on naturally occurring physical fields. The seismic and most of electrical methods make use of artificial sources; the added control over source position and characteristics offers important advantages. Since each of the four techniques measure the effect of different physical properties of the subsurface materials, each has its field of application. Often a combination of techniques is more effective than one alone (Orellana and Mooney, 1966). Resistivity method has its origin in the 1920's due to the work of Schlumberger brothers. In this method the midpoint of the electrode array remains fixed, but the spacing between the electrodes increased to obtain more information about deeper sections of Subsurface. The most commonly used methods for measuring earth resistivity are those of four electrodes. Current is driven through one pair of electrodes and the potential is measured with second pair of electrodes (Keller and Frischknecht, 1970).

There are a number of ways in which electric current can be employed to investigate subsurface conditions in a certain area. In the most commonly used method the current is driven through the ground by using a pair of electrodes, and the resulting distribution of the potential in the ground is read by using another pair of electrodes connected to a sensitive voltmeter.

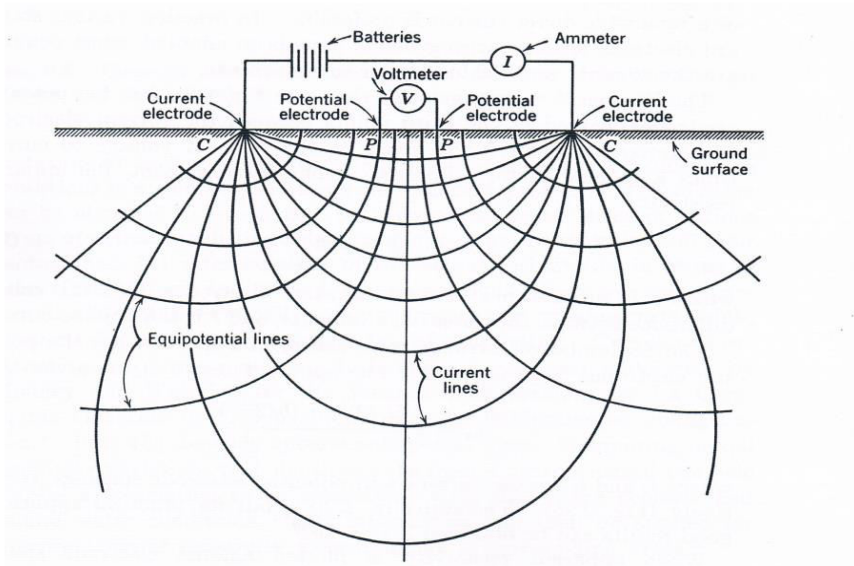


Figure (1): The geometry of current distribution within homogeneous and isotropic subsurface media (Todd, 1959).

From the magnitude of the current applied and from the knowledge of the current electrode separation it is possible to calculate the potential distribution and the path of the current flow if the underground materials were homogeneous. Anomalous conditions or inhomogeneities within the ground, such as electrically better or poorer conducting layers, are inferred from the fact that they deflect the current and distort the normal potentials. This represents briefly the principle of measuring subsurface variation in the electrical resistivity, which is the (reciprocal of conductivity) within the earth. In 1920's the technique of the method was perfected by Conrad Schlumberger, who conducted the first experiments in the field of Normandy (Sharma, 1986). In some surveys a combination of drilling and geophysical measurements may provide the optimum cost benefit ratio. The proper design of a geophysical survey is important not only in insuring that the needed data will be obtained but also in controlling costs, as the expense of making a geophysical survey is determined primarily by the detail and accuracy required (Zohdy, et.al.,1990).The electrical methods of geophysical exploration include a variety of techniques employing both natural and artificial sources, of which the latter has wider application. Within the artificial source group a distinction may be made between inductive and conductive methods. The inductive methods uses frequencies up to a few thousands cycles per second and the measurement of the electromagnetic field set up by the induced earth currents. The conductive methods involve the use of direct current (DC), or alternating current (AC) with frequencies up to few tens of cycles per second to study the electrical field (Orellana and Mooney, 1966).

Major Applications of Electrical methods

1- Detecting Lateral and vertical changes of soil and subsurface rock properties.
2- Detecting faults, weakness zones and cavities and any other subsurface geological structures.
3- Detecting groundwater table, aquifers thickness and its physical properties and groundwater pollution with salts and hydrocarbons.
4- Exploring subsurface sources of mineral ores.
5- Detecting underground seepages near earth filled dams and sewer leakages and groundwater pollution with chemical fluid seepages and achieving Earthling surveys related to civil engineering and building projects.
6- Detecting subsurface archeological remains.
7- Monitoring subsurface seepages for environmental applications.

Rocks Electrical Conductivity

With the exception of clays and certain metallic ores, the passage of electricity through rocks takes place by way of the groundwater contained in the pores and fissures, while the rock matrix being non conducting. All other factors being constant an increase in the concentration of dissolved salts in the groundwater leads to decrease in resistivity. In general way the resistivity is also controlled by the amount of water present. The more porous or fissured a rock the lower the resistivity. Degree of saturation also affects resistivity which increases with decrease in the amount of water in the pore spaces and fissures (Griffiths and King, 1981).

Most rocks conduct electricity only because of mineralized water in pores and fissures (Brine). This property is called (electrolytic conductivity). Their conductivity depends on the conditions of contained water, the amount of water that is contained, and the manner in which the water is distributed (Kunetz, 1966).

Rocks Bulk or Total Resistivity (ρ_s)

Rocks bulk or total resistivity of rocks (ρ_s) could be expressed as a function of many variables as following:

$$\rho_s = f(c, n, S_w, T_p, Q, \rho_m, \rho_w)$$

Where: c=clay content, n = porosity , S_w=degree of rock saturation with brine, T_p= temperature, Q= Ionic exchange, ρ_m= solid part or rock grains resistivity, ρ_w= Brine resistivity.

Current Behavior In Homogeneous Ground

The property of the electrical resistance of a material is usually expressed in terms of its resistivity. If the resistance between opposite faces of a conductive cylinder of length (L) and cross sectional area (A) is (R), (Figure (2)), the resistivity (ρ) is expressed as:

$$\rho = R\ A/L$$

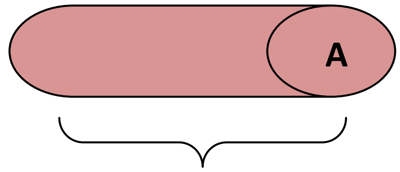


Figure (2): The resistivity of a conductive cylindrical body.

The simplest approach to the theoretical study of the current flow in the earth is to consider first the case of completely homogeneous isotropic earth layer of uniform resistivity. For a quantitative treatment, let us consider a homogeneous layer of length (L) and resistance (R) through which a current (I) is flowing. The potential difference across the ends of the resistance is given by Ohm's law and is:

$$\Delta V = RI$$

While: $\rho = R\ A/L$

The resistance, R, of the layer is specified by its length (L), of cross section area (A), and the resistivity (ρ). By definition: $R = \rho\ L / A$, and, therefore the equation can be written as the following:

$$\Delta V / L = \rho\ I / A$$

Where: I= current in Ampere (A) or mille Ampere (mA), L= length of the bead or layer, ΔV= potential difference in Volts (V)

The current density for each cross-sectional unit could be defined as (i), and the potential gradient could be expressed as (grad V), where:

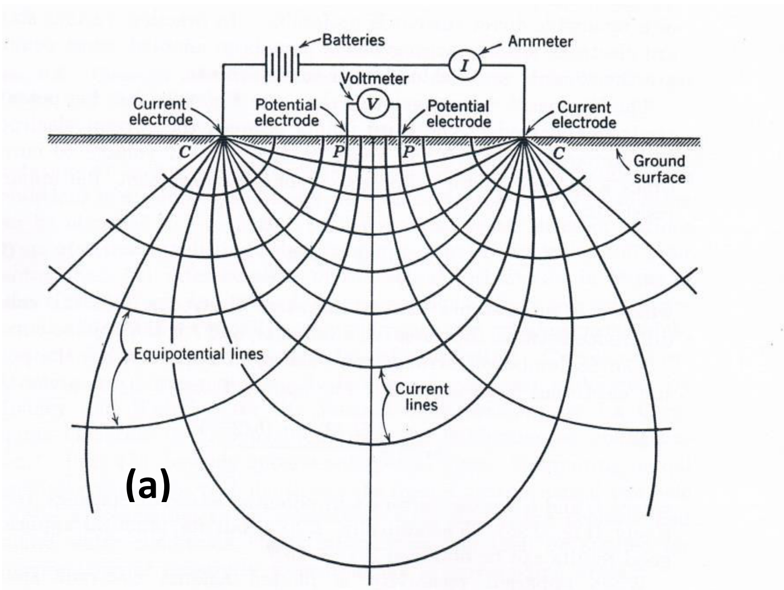
$$\text{Grad } V = \rho i$$

If a semi-infinite conducting layer of uniform resistivity bounded by the ground surface and a current strength (+I) enters at point (A) on the ground surface,

(Figure 3).The current will flow away radially from the point of entry and at any instant its distribution will be uniform over a hemispherical surface of the underground resistivity (ρ).

At a distance (r), away from the current source, the current density (i) would be (Sharma, 1986):

$$i = I / 2\pi r^2$$



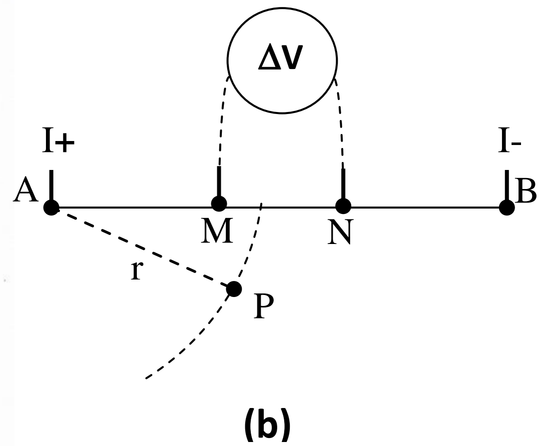


Figure (3): (a) The geometry of current distribution within homogeneous and isotropic subsurface media (Todd, 1959).

(b) Method of calculating potential distribution due to a current source in a homogeneous medium (Sharma, 1986).

$$\text{as: Grad } V = \rho i$$

The potential gradient $-\frac{\delta V}{\delta r}$ associated with the current is given by:

$$-\frac{\delta V}{\delta r} = \rho i = \rho I / 2\pi \ r^2$$

By integration we get:

$$V = \rho I / 2\pi \ r$$

The potential at distance (r) (e.g., at point (P) in figure (3,(b)) , is obtained by integrating the previous equation and is :

$$V_M^A = \rho I / 2\pi r$$

This is the basic equation which enables the calculation of the potential distribution in a homogeneous conducting semi-infinite medium.

The potential difference (ΔV) between the potential electrodes M and N , Figure(3,(b)) which caused by current (+I) at the source entry point (A) is:

$$\Delta V_{MN}^A = \frac{\rho I}{2\pi} \left(\frac{1}{AM} - \frac{1}{AN}\right)$$

In the same manner, the potential difference (ΔV) between the points M and N , caused by (-I) current at the (sink) or exit point (B) is :

$$\Delta V_{MN}^B = \frac{-\rho I}{2\pi} \left(\frac{1}{BM} - \frac{1}{BN}\right)$$

The total potential difference between M and N is therefore, given by the sum of the right-hand sides of the previous two equations, and is:

$$\Delta V_{MN}^{AB} = (\Delta V_{MN}^A + \Delta V_{MN}^B)$$

Or

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

$\left(\frac{1}{AM} - \frac{1}{AN} - \frac{1}{BM} + \frac{1}{BN}\right)$ is called the geometrical factor (G).

Then the potential difference between M and N could be written as:

$$\Delta V_{MN}^{AB} = \frac{\rho I G}{2\pi}$$

A more simple way to write the previous equation is:

$$\Delta V \, 2\pi = \rho I G$$

Or:

$$\rho = 2\pi \frac{\Delta V}{I} * \frac{1}{G}$$

(ρ) Will be constant in homogenous and isotropic Medias even if the electrode array or the geometrical factor (G) is changed.

It's common in the geoelectrical – hydrogeological studies to use the Ohm resistivity meter as a field instrument and Schlumberger configuration as a ground electrodes array, (Figure (4)).

The Vertical Electrical Sounding (VES) by using Schlumberger array

In the generalized Schlumberger array the distance between the potential electrodes (MN) is small as compared to the distance between current electrodes (AB) , where : AB ≥ 5MN, (Kunetz,1966).The (VES) points apparent resistivity (ρ_a) field readings were obtained as the half current electrodes separation (AB/2) which for example increased in steps starting from 1 to 200 m gradually, while the half distance between potential electrodes MN/2 was gradually increased in steps starting from 0.2 m to a 40 m, according to the geometrical factor (G) of Schlumberger configuration in order to obtain a measurable potential difference.the apparent resistivity is measured at every step of changing electrode spacing. The current gain (output current) of the resistivity meter is increased gradually from 1 to 1000 mAmp. , to yield a sufficient current penetration to the required depths according to the subsurface lithology. The Schlumberger array, (Figure (4)), is used by keeping the potential electrodes at a closer distance. The apparent resistivity (ρ_a) is determined by using the Equation:

$$\boldsymbol{\rho_a} = \pi \left\{ \frac{(\frac{AB}{2})^2 - (\frac{MN}{2})^2}{MN} \right\} * \frac{\Delta V}{I}$$

Where (AB/2) is the half distance of current electrodes, (MN/2) is the half distance of potential electrodes and (ΔV/I) is the potential difference over the applied current.

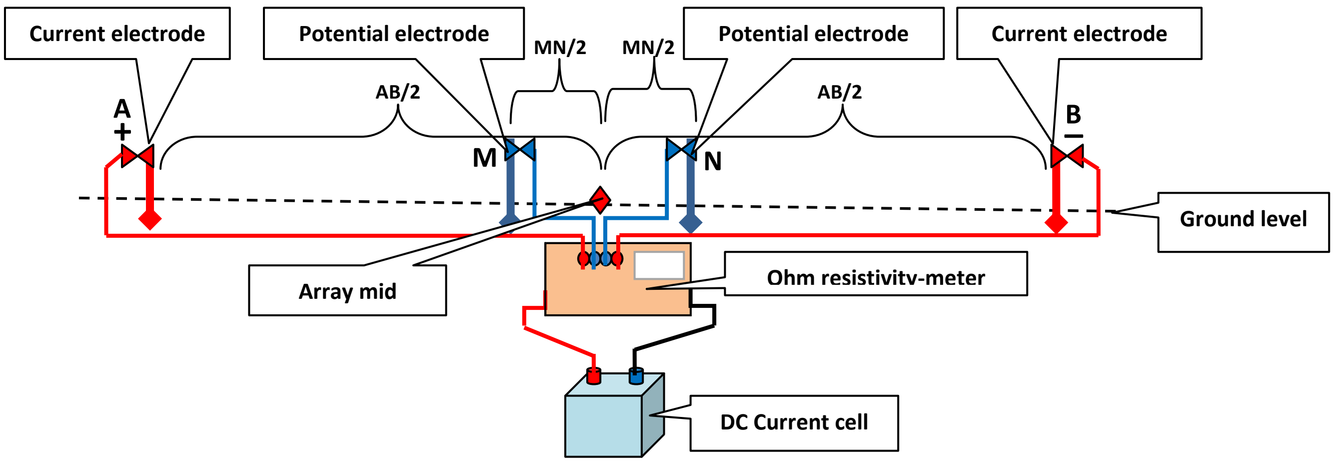


Figure (4): The electrode array for Schlumberger configuration at field resistivity survey, (AL-Khafaji 2014).

Apparent resistivity (ρ_a) with AB\2, MN\2 field readings is arranged in a table like the one of figure (5).

Vertical electrical Sounding (VES) Resistivity Curves

The next step is to draw the apparent resistivity curve on a log-log paper designed especially for this reason, each MN\2 spacing drawn as a separated segment then these segments will form the whole shape of the curve, figure 6. In this figure the sounding resistivity curve is called unsmoothed curve as it composed of segments and these curve segments may show vertical and horizontal shifting from each other. This is mainly produced by the effect of heterogeneity or inhomogeneity of subsurface rock layers. The curves must be smoothed to solve resistivity curve discontinuities, Figure 6 , which produced by different (MN) spacing's of the same(AB) spacing's and mainly caused by the lateral heterogeneity and anisotropy.

VES No.				
AB	MN	AB/2	MN/2	ρa Ω.m
6.4	2	3.2	1	268.5
8	2	4	1	220.7
10	2	5	1	173.4
12.6	2	6.3	1	118.5
16	2	8	1	70.7
20	2	10	1	40.4
25	2	12.5	1	24.8
32	2	16	1	16.97
40	2	20	1	14.2
40	20	20	10	22.1
50	20	25	10	13.2
50	2	25	1	13.2
64	20	32	10	10.7
80	20	40	10	10.3
100	20	50	10	10
126	20	63	10	10.2
160	20	80	10	10
200	20	100	10	10.3
250	20	125	10	10.62
320	20	160	10	11.03
320	20	160	10	10.13
400	100	200	50	10.3
400	100	200	50	11.05
500	20	250	10	10.6

Figure (5): a table arrangement to record the information of one VES point by using Schlumberger electrodes configuration.

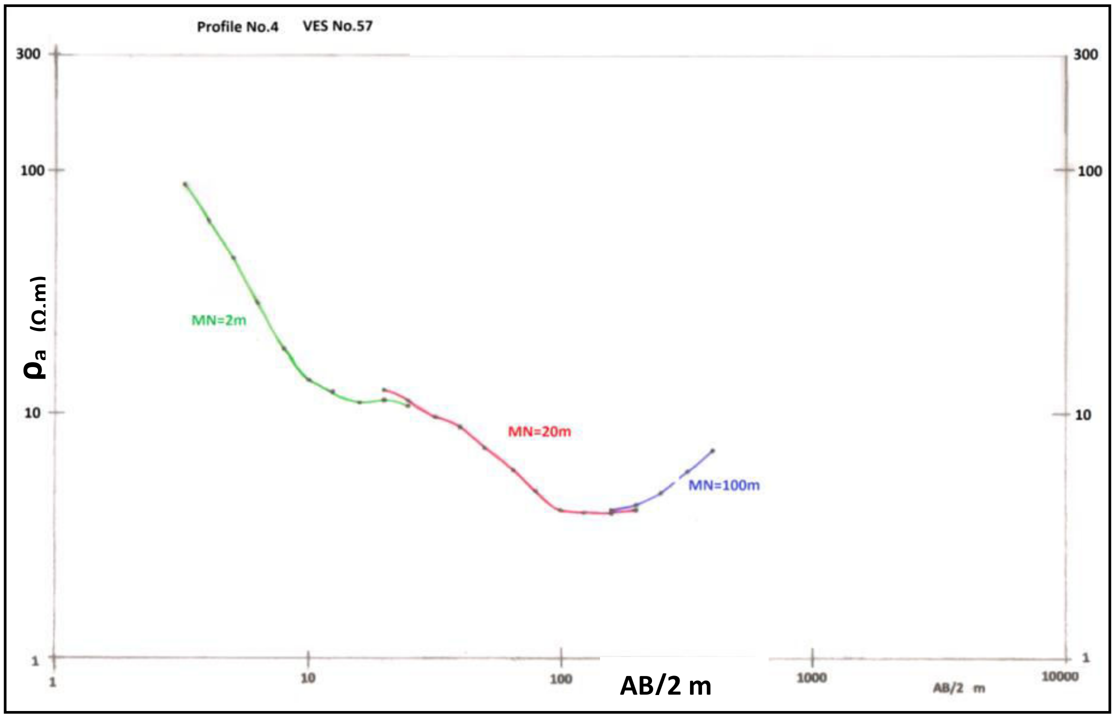


Figure (6): The unsmoothed VES resistivity curve showing the MN segments.

The scales of the both curve axes are logaretmic. After the above procedures the curve must be smoothed by eliminating the shifted segments in order to prepare it for interpretation, figure 7. The smoothing attended by shifting the discontinued curve segments of the same AB/2 spacings but with different MN/2 spasings upward or downward in order to obtain a smoothed curve like the one appears in Figure 7. This procedure also corrects curve deformations that may produced by heterogeniety.

The manual interpretation of resistivity curve is achieved by the assistance of standard curves which provided by some text books of geophysics or catalogs of resistivity curve interpretation like (Orellana and Mooney, 1966) complete and partial matching standard curves.

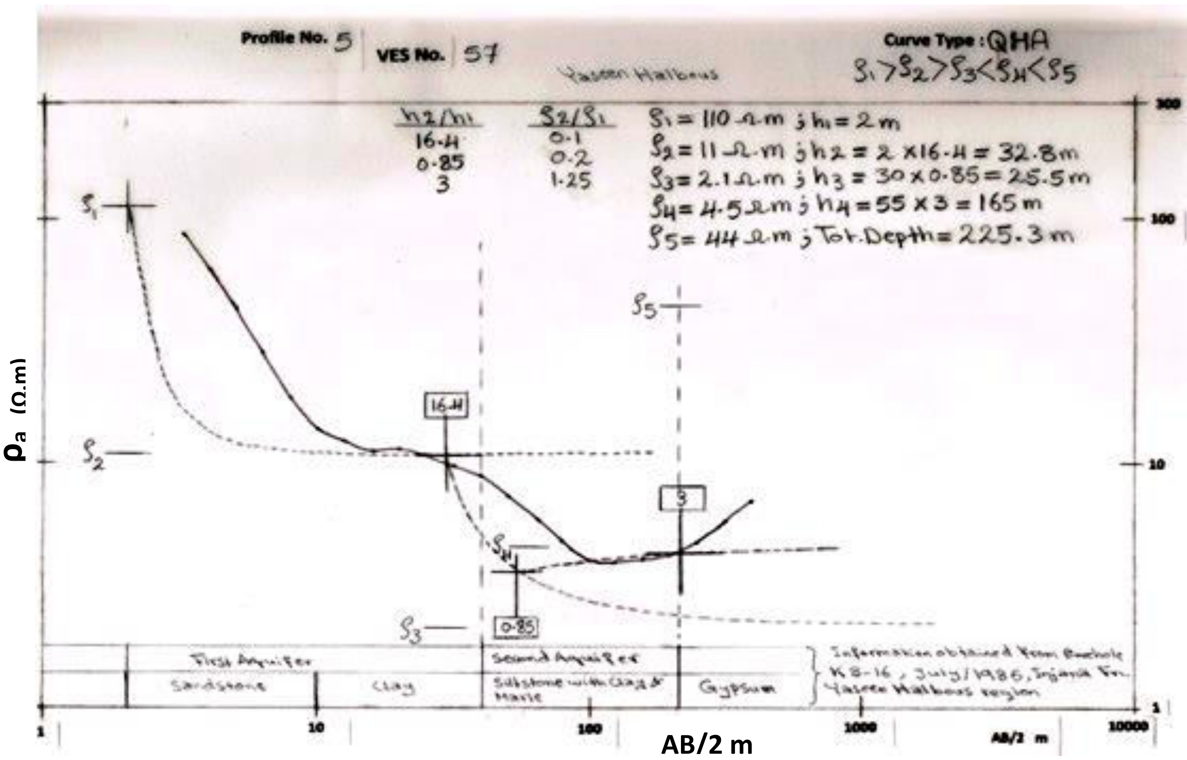


Figure (7): The smoothed VES resistivity curve interpreted manually by using the auxiliary point method.

Generally the partial curve matching by using the auxiliary point method is very common in the manual interpretation of VES curves. The two layer resistivity standard curve of (Orellana and Mooney, 1966), is shown in figure 8.

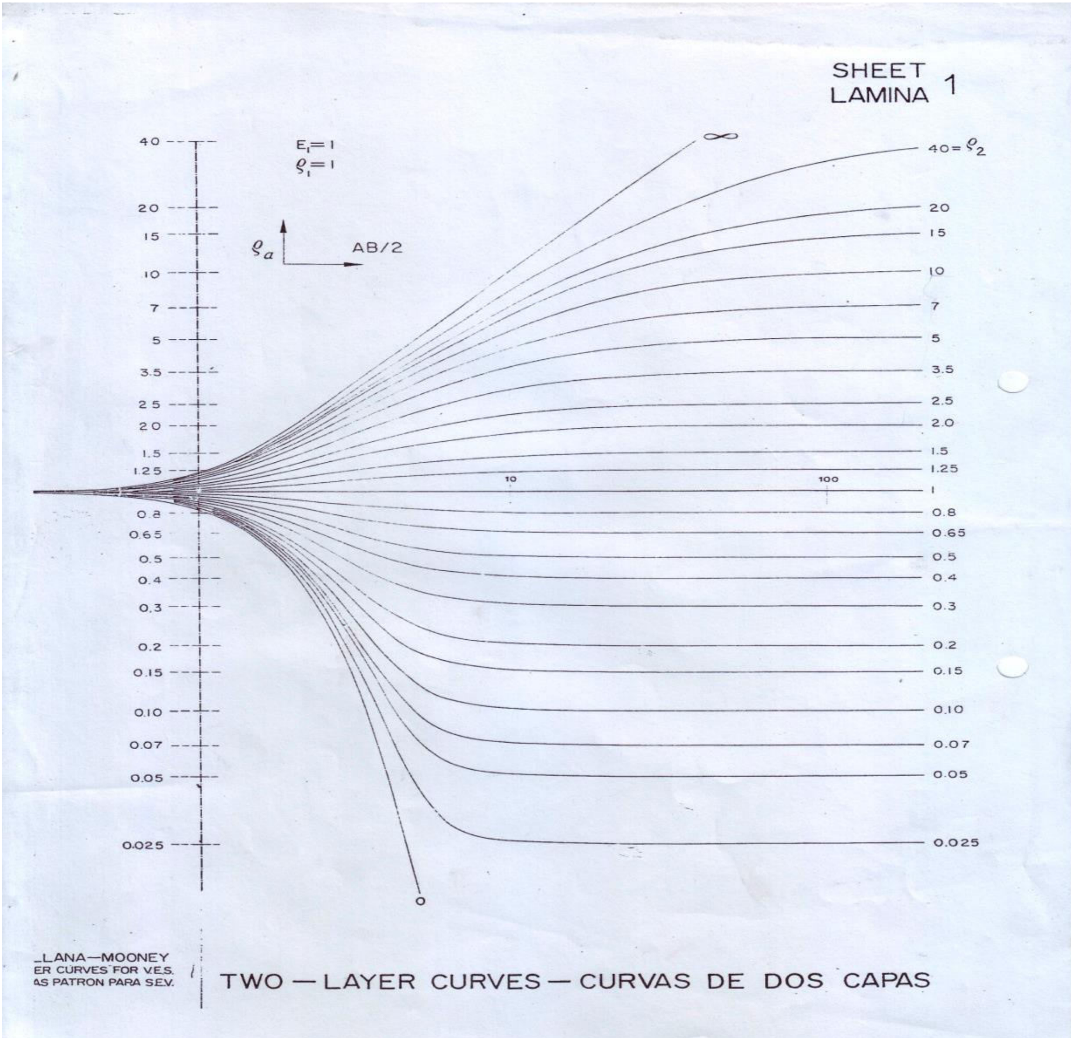


Figure (8): The two layers standard curves of (Orellana and Mooney, 1966) which used in interpreting the Schlumberger VES resistivity curve by adopting the auxiliary point method.

The limitations of Resistivity method

Resistivity surveying is an efficient method for delineating shallow layered sequences or vertical discontinuities involving changes of resistivity. It does, however, suffer from a number of limitations:

1. Interpretations are ambiguous. Consequently, independent geophysical and geological controls are necessary to discriminate between valid alternative interpretations of the resistivity data.

2. Interpretation is limited to simple structural configurations. Any deviations from these simple situations may be impossible to interpret.

3. Topography and the effects of near-surface resistivity variations can mask the effects of deeper variations.

4. Depth of penetration of the method is limited by the maximum electrical power that can be introduced into the ground and by the physical difficulties of laying out long lengths of cable. The practical depth limit for most surveys is about 1km (Kearey et.al. 2002).

<u>Resistivity Qualitative Field Techniques</u>

<u>Qualitative interpretation of resistivity method surveying data</u>

1- (VES) curve type interpretations in which depend on the number of subsurface detected layers. The three layers sounding curves represents (A, H, K, and Q) types. Composite Sounding curve types for more than three layers curve types , for example : HKH type for layering resistivity of : $\rho1>\rho2<\rho3>\rho4<\rho5$

2- (VES) Curve type mapping: in this method similar curve types are zoned for a map across the area. This zone provides a primary evidences about the geoelectrical zoning columns under each VES point in the area. One of these maps appears in figure (9).

3- Equi-apparent resistivity surfaces: an iso-apparent resistivity map could be constructed for every Pseudo- depth or Sounding current electrodes separation (AB\2). Such maps or surfaces may provide a primary vision about the subsurface apparent resistivity variations (figure 10).

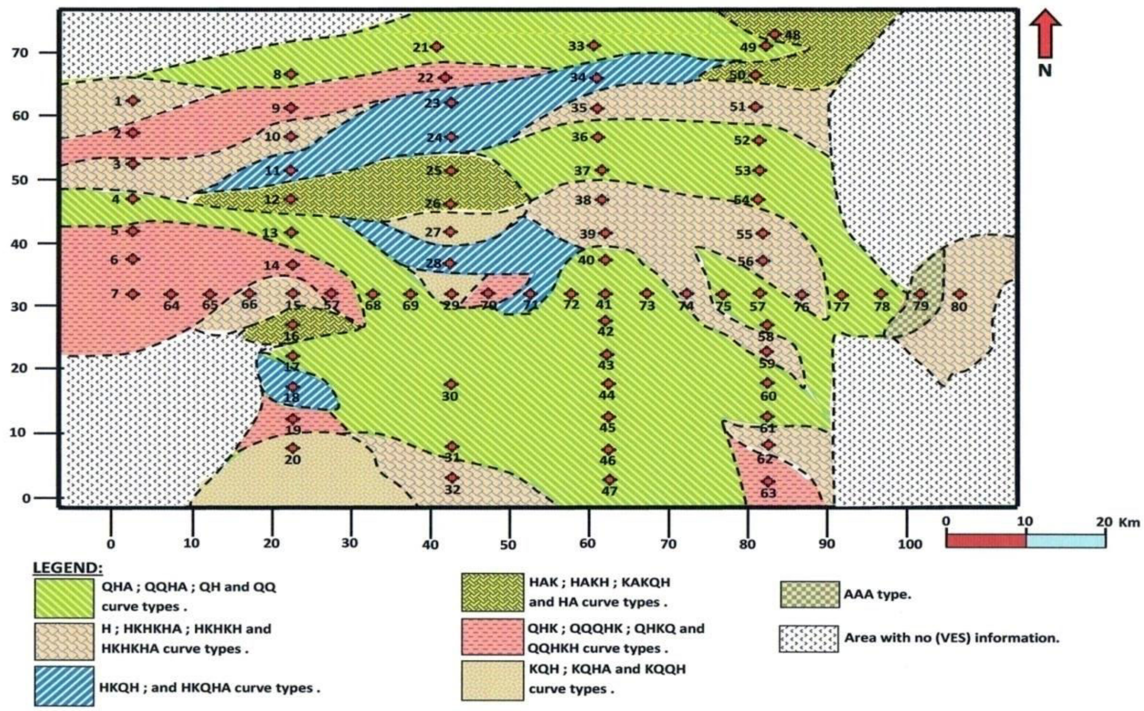


Figure (9): VES Curve types zoning map at South t o Sinjar anticline region \ Iraq , (Al-Khafaji , 2014)

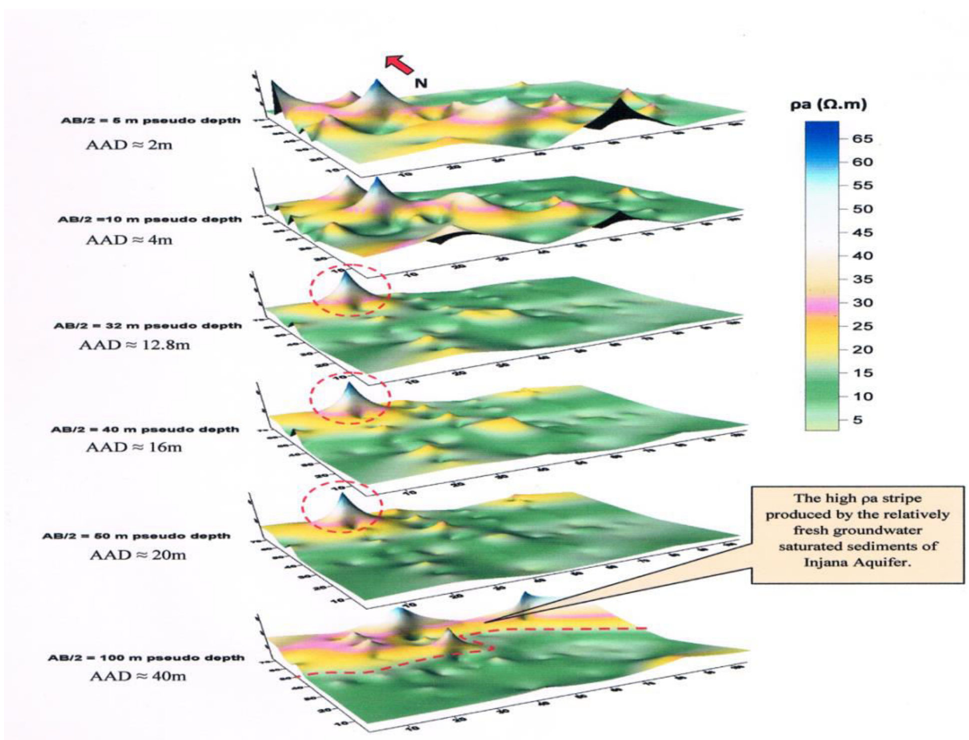


Figure (10): Equi-apparent resistivity maps at South t o Sinjar anticline region \ Iraq , (Al-Khafaji , 2014)

4- Apparent resistivity Pseudo sections: in this type of qualitative interpretation a section is drawn by making the AB/2 spacing as a y- axis and distance between VES points as x-axis. Then the measured apparent resistivity values contoured to visualize lateral and vertical variation of apparent resistivity along the section. In this section the AB/2 spacing considered as a function to the depth, therefore it's not an actual depth, but it represent a pseudo depth, and that's why these sections are called pseudo sections. Figure 11 shows one of these pseudo sections.

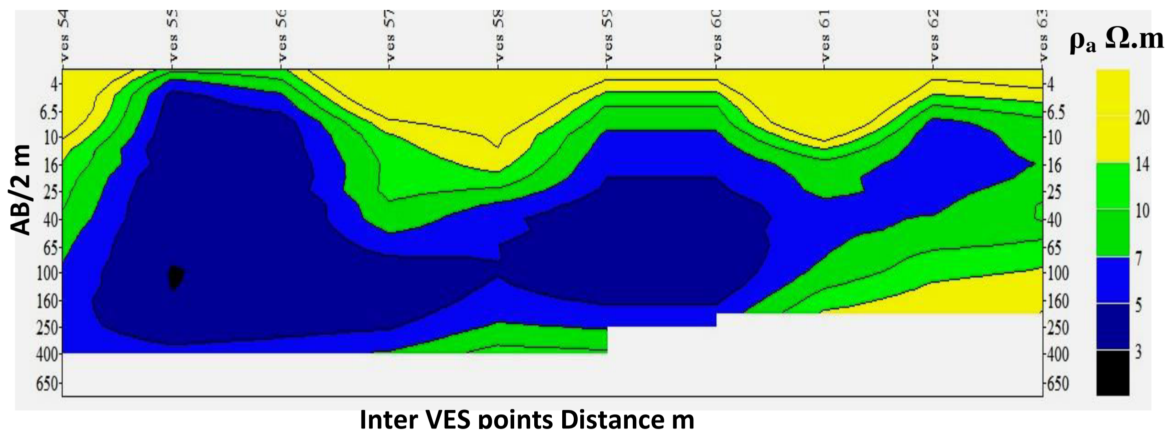


Figure (11): apparent resistivity ρ_a pseudo section along a profile of 10 VES points showing the vertical and horizontal variation according to subsurface lithology.

5- Constant Separation Traversing (CST) and Horizontal electrical Profiling (HEP) for the apparent resistivity data: a plot is drawn between the AB/2 spacing as a y-axis and the inter VES distance as a x-axis , then connecting values of apparent resistivity which belongs to same AB\2 spacing's along the one profile. Figure 12 showing a horizontal electrical profile. This type of profiles used to detect the sudden or abrupt lateral variations of apparent resistivity with pseudo depth which expressed as the AB\2 spacing. The sudden lateral variations in ρ_a are useful in detecting fault locations and near surface lenses like gravel lenses and other lateral variations in lithology facies or clay and groundwater content.

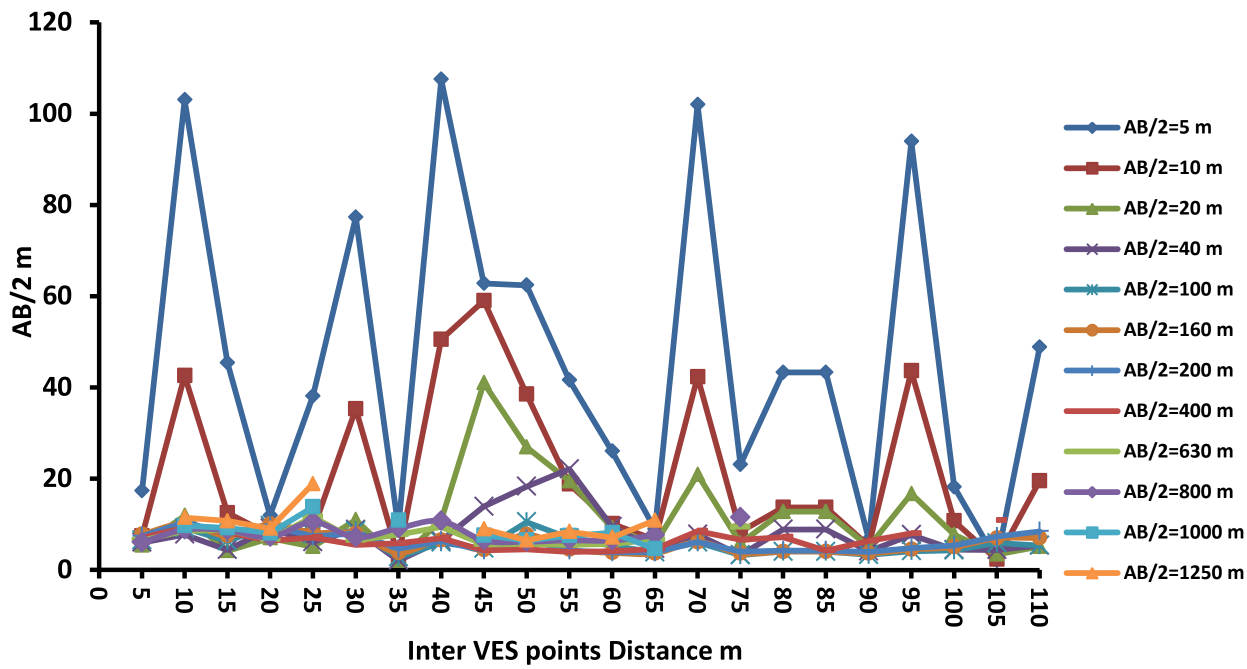


Figure (11): apparent resistivity ρ_a Horizontal Electrical Profile along a profile composed of 20 VES points showing the lateral variation with pseudo depth.

VES Curve types

The curve type represents a description of how layers change their resistivity values with depth under the array mid point. Therefore , curve type represents a part of the descriptive interpretation. Curve types must be determined for all of the (VES) points in the surveyed area during the manual interpretation. Its important to mention here that the curve type is not important as a descriptive tool only, but it also represents the key that leads the way during the quantitative interpretation, especially if the partial curve matching method with the standard resistivity curves adopted . Figure (12), shows a sample of a smoothed and manually interpretted resistivity curve of (QHA) type.

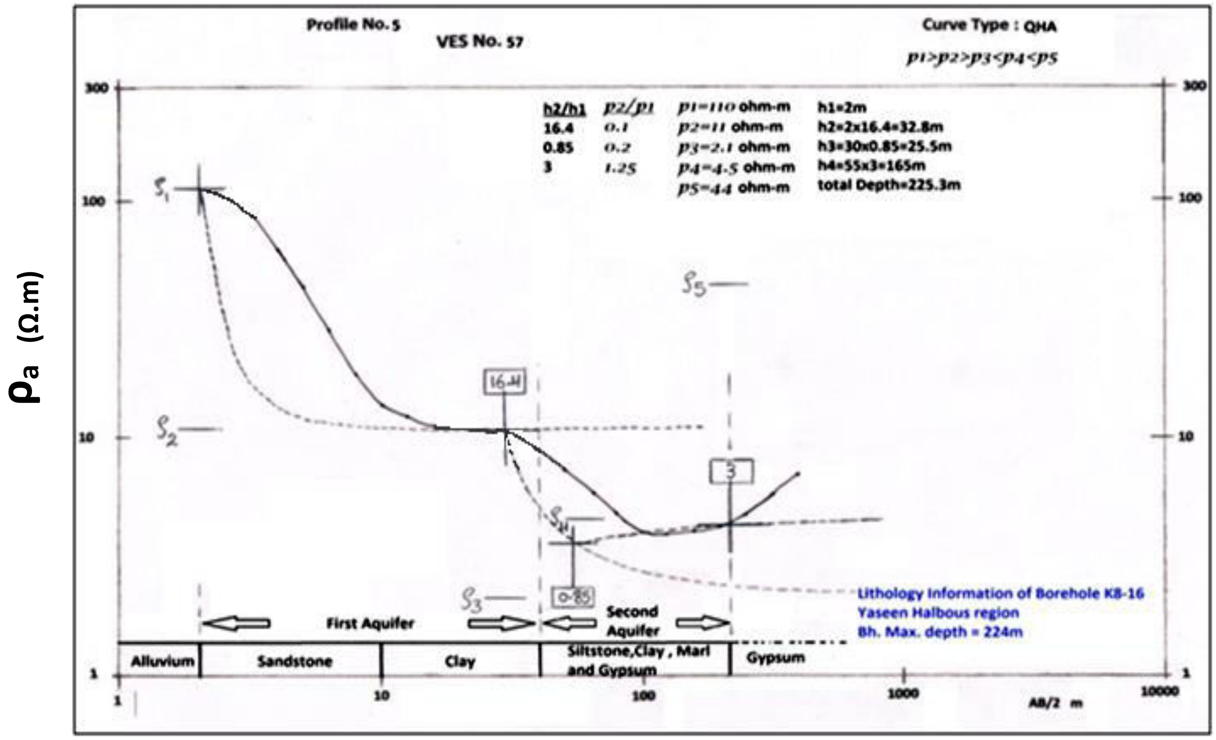


Figure (12): (QHA) resistivity curve type after smoothing drawn and interpreted manually, (Al-Khafaji, 2014).

The whole set of three – layer sounding curves can be devided into four groups, depending on the relative values of ρ_1 , ρ_2 and ρ_3 , (Bhattacharya & Patra , 1968):

1- Minimum type : when $\rho_1 > \rho_2 < \rho_3$. this is also referred as (H-type) (associated with the name of Hummel).
2- Double assending type : when $\rho_1 < \rho_2 < \rho_3$. this is also known as (A-type) (corresponding to the term anisotropy).
3- Maximum type : when $\rho_1 < \rho_2 > \rho_3$, this is known as (K-type) or is sometimes referred as DA-type (meaning displaced or modified anisotropy).
4- Double descending type : when $\rho_1 > \rho_2 > \rho_3$. this is known as (Q-type) and is sometimes referred to as DH-type (meaning displaced Hummel or modified Hummel).

The apparent resistivity curve for a three-layer structure generally has one of four typical shapes, determined by the vertical sequence of resistivities in the layers

(Figure 13). The type K curve rises to a maximum then decreases, indicating that the intermediate layer has higher resistivity than the top and bottom layers. The type H curve shows the opposite effect; it falls to a minimum then increases again due to an intermediate layer that is a better conductor than the top and bottom layers. The type A curve may show some changes in gradient but the apparent resistivity generally increases continuously with increasing electrode separation, indicating that the true resistivities increase with depth from layer to layer. The type Q curve exhibits the opposite effect; it decreases continuously along with a progressive decrease of resistivity with depth. Once the observed resistivity profile has been identified as of K, H, A or Q type, the next step is equivalent to one-dimensional *inversion* of the field data. The technique involves iterative procedures that would be very time-consuming without a fast computer. The method assumes the equations for the theoretical response of a multi-layered ground. Each layer is characterized by its thickness and resistivity, each of which must be determined. A first estimate of these parameters is made for each layer and the predicted curve of apparent resistivity versus electrode spacing is computed.

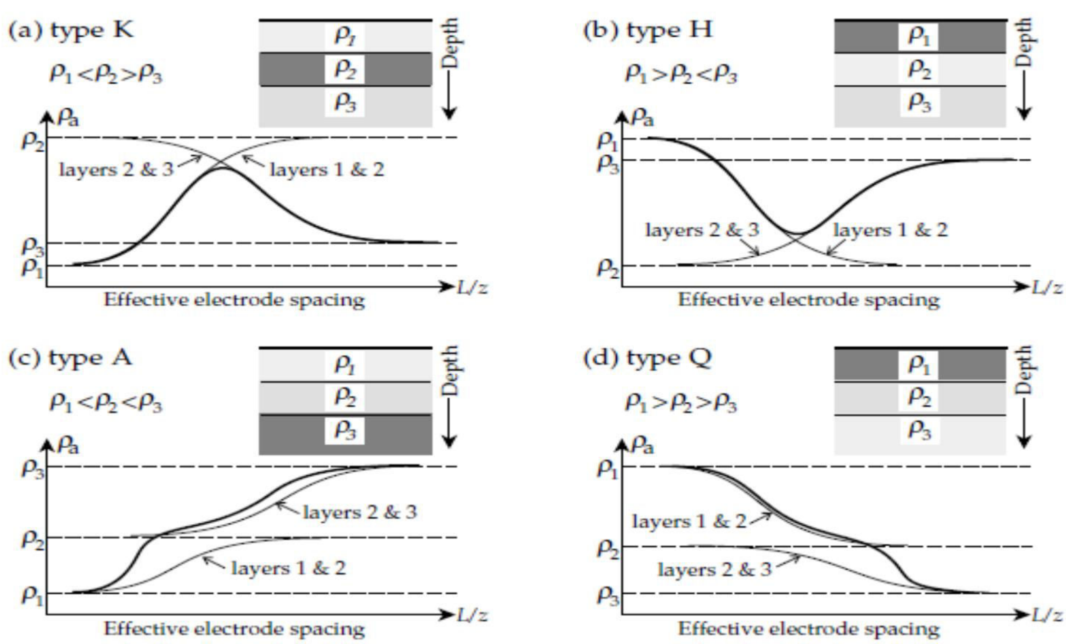


Figure 13: The main four VES curve types after smoothing.

(CST) Constant Separation Traversing Approach.

The object of electric horizontal profiling or combined Traversing approach is to detect the lateral variations in the resistivity of the ground. In Schlumberger method of electrical profiling, the current electrodes (AB) remains fixed at a relatively large distance, for instance, a few hundred meters, and the potential electrodes (MN) with a small constant separation (Sharma , 1986).

The appearance of the resistivity profile obtained by horizontal profiling will depend not only on the positions of the potential electrodes M and N, but also on the positions of the current electrodes A and B with respect to the inhomogeneities in the earth and the reason behind that is that all electrodes are moved after each measurement (Kunetz, 1966).

The effect of vertical structures e.g. (faults, fissures, dikes, veins, and shear zones) is lateral. If these features crop out, abrupt discontinuities in the slope of (ρa) curves are obtained as the mobile electrode configuration crosses the vertical resistivity boundary. Many of key features found in the resistivity anomalies over a vertical fault are found also in anomalies over near vertical structures. The fault represents a vertical contact problem between two media of differing resistivity. Therefore, the calculated curves for (ρa) are discontinuous at the vertical boundary. The discontinuity will be evident in practice as a steep gradient in the resistivity curve (Sharma, 1986).

The traversing obtained by moving an electrode spread with fixed electrode separation along a traverse line, the array of electrodes being aligned either in the direction of the traverse (longitudinal traverse) or at right angles to it (transverse traverse). The former technique is more efficient as only a single electrode has to be moved from one end of the spread to the other, and the electrodes reconnected, between adjacent readings. The Vertical discontinuity distorts the direction of current flow and thus the overall distribution of potential in its vicinity. (VES) data from several soundings can be presented in the form of a pseudosections and it is now possible to invert the data into a full, two-dimensional geoelectric model rather than a sequence of discrete, uni-dimensional geoelectric sections. This technique is known as electrical imaging or electrical tomography (Kearey et.al, 2002).

The advantage of using several line lengths (several AB spacing's) in the horizontal profiling or is undertaken either to study layers at several different depths, or, more often , to facilitate the distinction between structures that are indistinguishable because they produce overlapping effects at the surface (Kunetz, 1966).

The interpretation of resistivity profiles is usually qualitative in nature. The effects of subsurface structures are located by the abrupt lateral changes in (ρa) profiling curves. The (ρa) varies over a vertical boundary differs considerably between the various electrode arrays (Keller, 1970).

The lateral inhomogeneities in the ground affect resistivity measurement in different ways. The effect depends on: (1) the size of the inhomogeneity with respect to its depth of burial, (2) the size of the inhomogeneity with respect to the size of electrode array, (3) the resistivity contrast between the inhomogeneities and the surrounding media, (4) the type of electrode array used, (5) the geometric form of the inhomogeneities, and (6) the orientation of the electrode array with respect to the strike of inhomogeneities (Zohdy et al , 1990).

Iso-Apparent Resistivity Maps.

The apparent resistivity map has numerous applications since the advent of electrical prospecting and considered very useful on large scales. Its principal advantages lie in the ease of making field measurements and simplicity in qualitative interpretation of results. Once they are free from various electrode effects, the apparent resistivities reflect the corresponding variation in the true resistivities in a zone has a fairly well known (nearly constant) depth. One of electrical sounding advantages is that it permits continuous coverage which makes them preferable for detailed surveys of features such as semi vertical faults. The fact, the possibilities of the resistivity map are fairly limited, when it comes to defining precisely the nature and the form of structures, which is not a major inconvenience. In preliminary surveys subsurface object reveal different anomalies when studied in details by other methods. Therefore resistivity mapping attempts to localize shallow contacts or facies variations such as faults and constructive pockets. In the interpolation of a certain geophysical parameter, such as (resistivity), the depth to a given bed is determined precisely at some certain points in the investigated area by using expensive methods such as well drilling. While in simple cases, the apparent resistivity could be calibrated by drawing structural contours that gives predicted depths directly (without additional well drilling). This has been done, for example, in the Joplin district (USA) and near Hettenschlag (Alsace), (Kunetz, 1966).

Space Sections.

The space section represents an apparent resistivity section which considered by plotting the apparent resistivity (ρa) , as observed , along vertical lines located beneath the sounding stations on the chosen profile. The apparent resistivity values are then contoured; Figures (14 and 15). Generally a linear vertical scale is used to suppress the effect of near- surface layers. It represents one of the important constituents of qualitative interpretation (Zohdy et al., 1990).

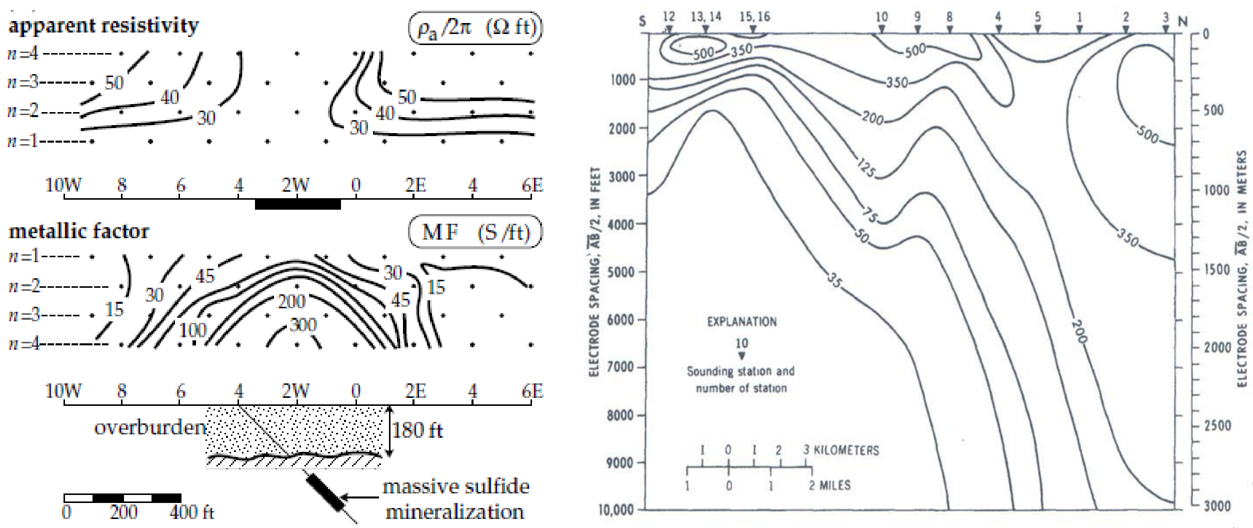


Figure (14): Pseudo depth section near Minidoka , Idaho. Values on contour lines designate apparent resistivities in ohm.m, Snake River basalt thickness toward the north. After (LOWRIE W,2007), (Zohdy et al., 1990).

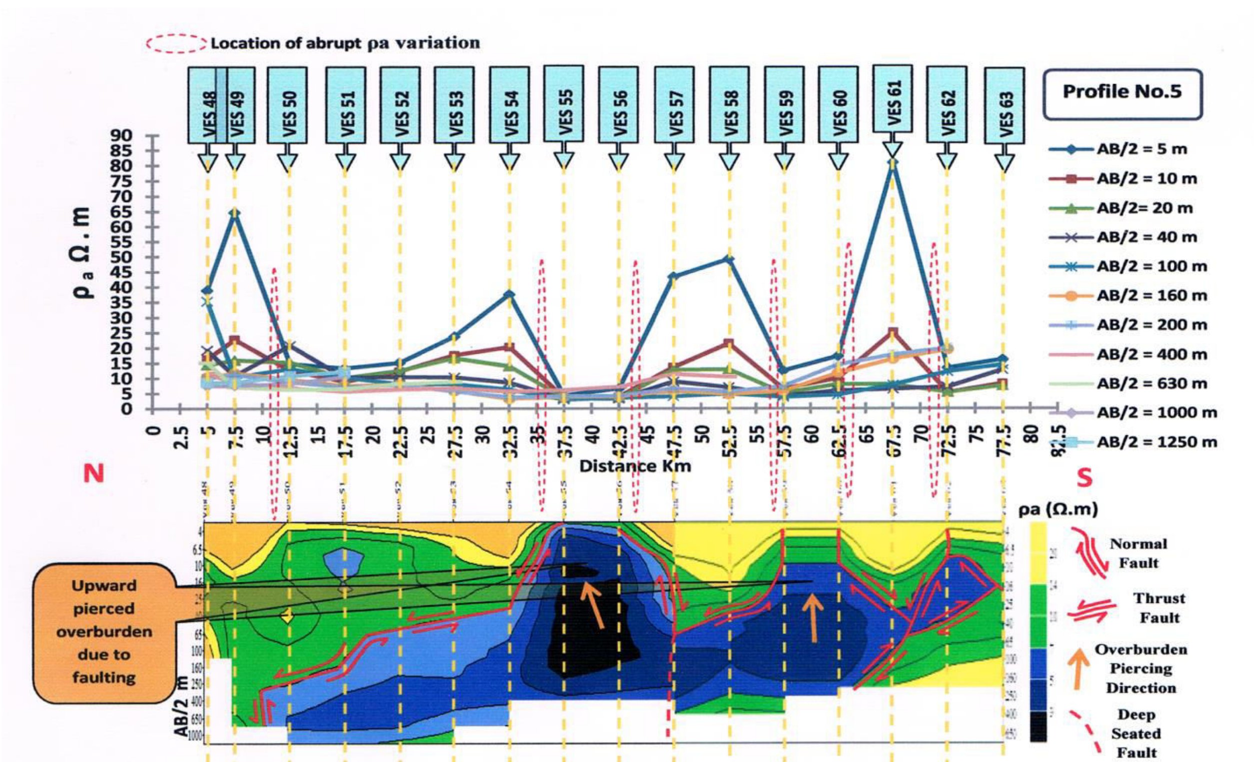


Figure (15): Spacing section and Pseudo depth section for the same profile line which composed of 16 VES points, (AL-Khafaji, 2014).

Quantitative Interpretation

Resistivity Curves Interpretation by using the Auxiliary point Method.

The sounding resistivity curve quantitative interpretation requires that the field curves be drawn on transparent bilogarithmic paper of the same modulus as used in the master curves. An improvement on this procedure is to plot the field data onto plain transparent paper which can then be superposed onto the bilogarithmic paper as necessary. This facilitates curve matching and tracing. Before proceeding with the interpretation by using Ebert method, the interpreter must decide that his field curve represents a structure of two, three, four or more layers. Occasionally the appearance can be misleading; some three-layer curves look superficially like two-layer curves and a similar statement applies to multi-layer curves. Curves representing five or more layers can be interpreted by means of auxiliary point method. When the number of layers reaches to five or six, interpretation becomes complicated (Orellana and Mooney, 1966).

In the partial curve matching each part of the curve will be interpreted relatively to the part before. The first curve part is the most important one in interpretation where a special care must be taken during matching to insure best matching resolution to find the ratio $\rho 2/\rho 1$. Because the matching resolution for the following parts of the curve will depend on the first part matching resolution. In other words the positions of crosses during interpretation are relative to each other and they are drawn relatively to the first cross. Figures (16 and 17), shows manual drawn sounding field curves before and after smoothing. The detailed steps for interpreting a resistivity sounding curve by using Ebert method could be found in (AL-Khafaji, 1999).

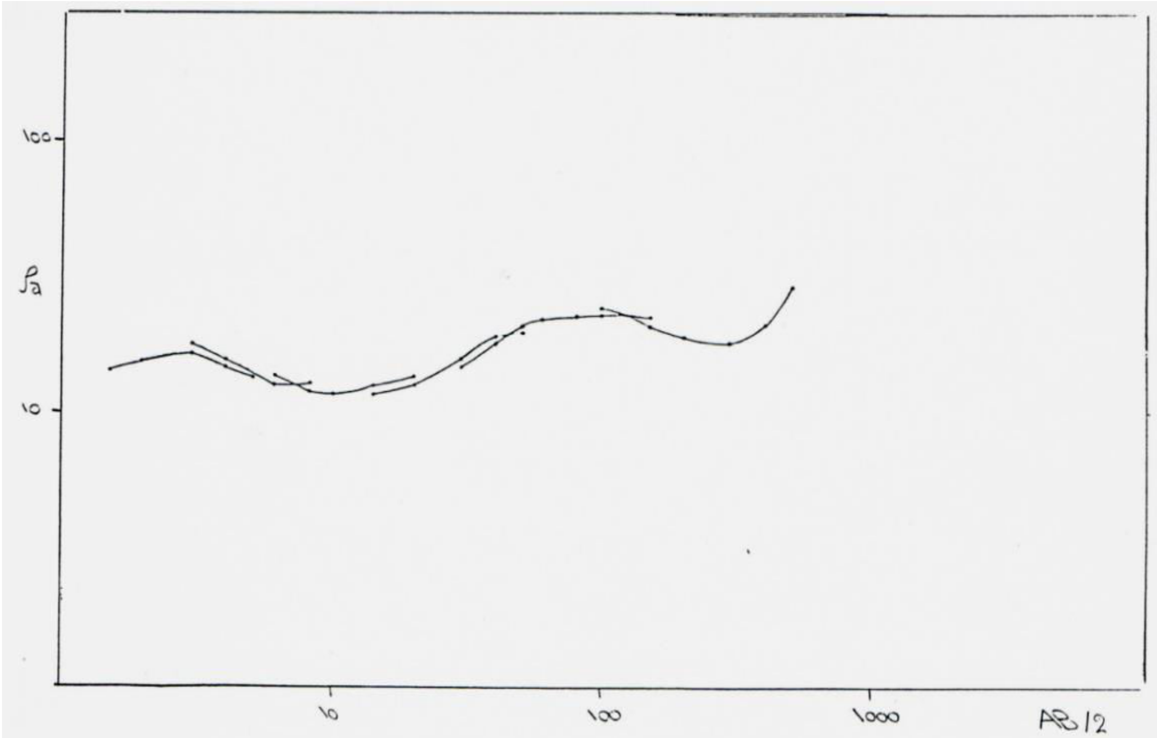


Figure (16): Schlumberger Sounding curve before smoothing shows the discontinuous MN segments, After (AL-Khafaji, 1999)

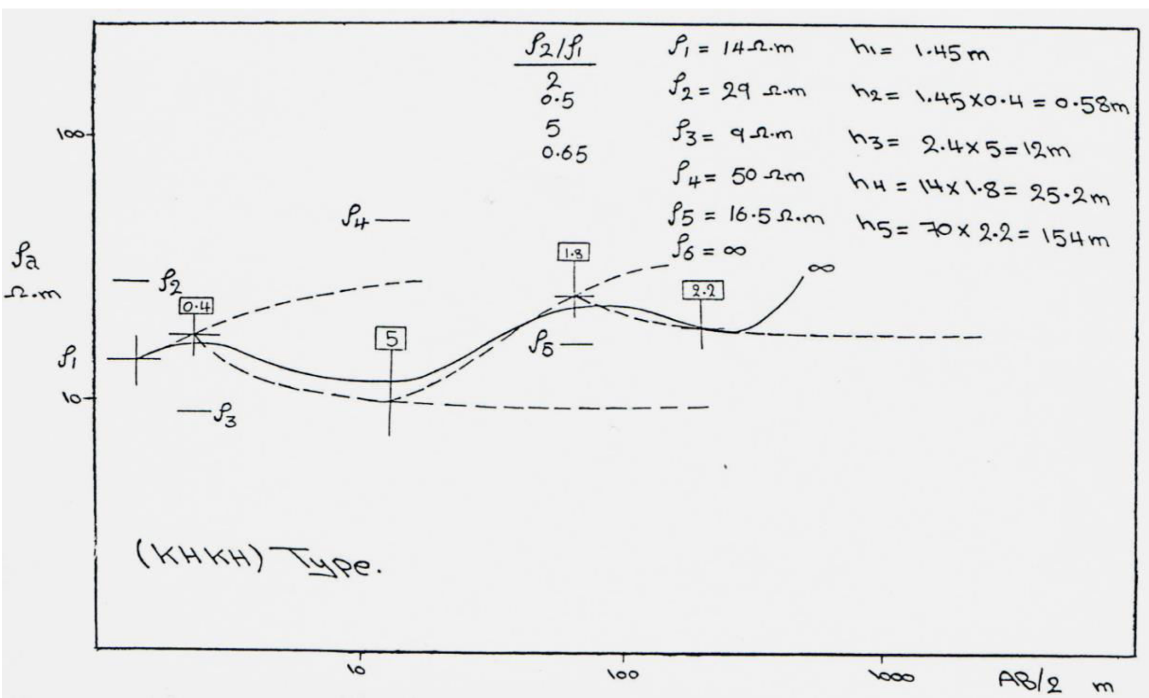


Figure (17): Schlumberger Resistivity sounding curve after smoothing, (AL-Khafaji, 1999).

The field data of VES point could be fed to computer software. This procedure require a special care that interpretation results must agree with the nearby subsurface lithology information which obtained from boreholes or well logging graphs if available. An example of such reliable computer software is the (IPI2Win) software.

The data input to a computer software specialized for (VES) processing, interpretation and results enhancement. The software usage is useful in enhancing the manual interpretation results during curve interpretation. This has been noticed during software usage:

1- The software is reliable during curve smoothing and discontinuity treatment.
2- It is able to treat ambiguity that related to principal of equivalence and suppression by reducing the predicted resistivity layering number as much as possible.
3- The reduction of (r m s %) between the calculated and the field curve as much as possible.

The software uses the common forward and inversion technique (IPI2Win, 2001). Figure (18 and 19), shows one of the processed and interpreted VES points , first manually and ,second by attending (IpI2Win) computer software, for the same VES point.

It's important to mention that the enhancement of (VES) results by using such computer software should be attended very carefully without effecting the actual thickness values of layers which assisted by boreholes information. Therefore, it would be better to interpret the sounding curve manually at first then enhancing this interpretation by using the computer software.

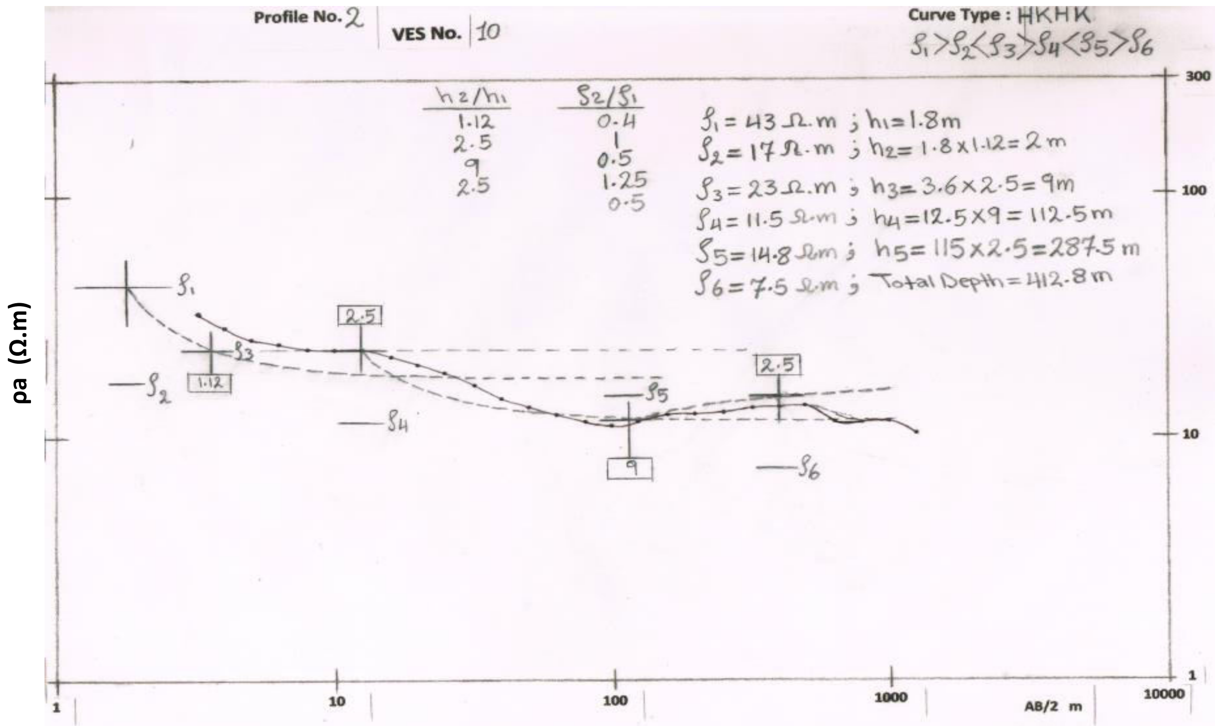
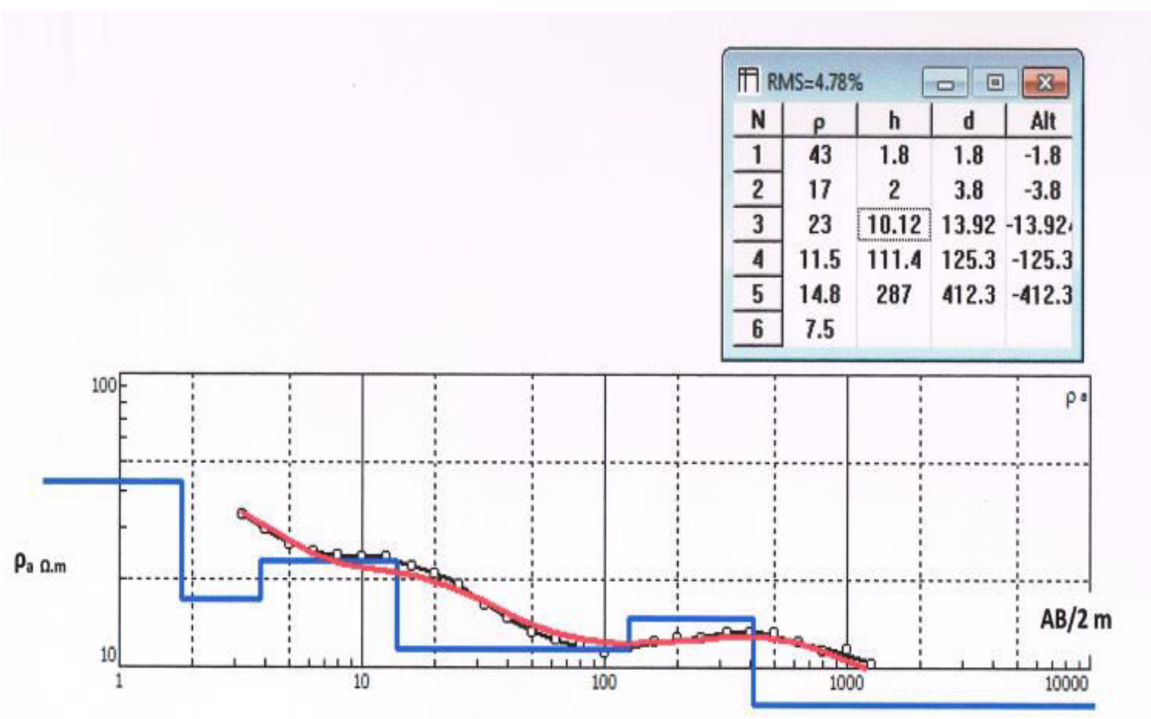


Figure (18): Profile No.2; VES No.10 manual interpretation using the auxiliary point method technique.



N	ρ	h	d	Alt
1	43	1.8	1.8	-1.8
2	17	2	3.8	-3.8
3	23	10.12	13.92	-13.92
4	11.5	111.4	125.3	-125.3
5	14.8	287	412.3	-412.3
6	7.5			

Figure (19): Profile No.2; VES No.10 interpretation using computer software.

It’s important to mention that a special care must be taken when enhancing the curve interpretation using this software and this could be briefed by the following:

1- The reduction of rms% could be achieved by changing the field curve very slightly to make it more matched with the calculated or theoretical curve during what is called forward and inversion method. Therefore its preferable to keep the layers thickness (h) values which is assured by manual interpretation and borehole information constant. The common change will happen slightly on (ρ) values in a logical amount that agree with the geological conditions.
2- The curve type must remain the same of the manual interpretation, so that the interpreter who uses such software must always check out the changes that might happen on the curve type during his enhancement operation. If the curve type changed then the interpretation will be inaccurate.
3- During the drawing of resistivity and pseudo depth sections, it is very important to check out the depth of investigation because the software has the ability to extend a proposed depth on the section which is sometimes not desired by the interpreter. So that, sections depth must be logical according to the array current electrodes (AB) distance which is the main balance of investigation depth and these depths must be always compared with the lithology and hydrology of the boreholes which are close to the sounding points.
4- Its preferable to draw the geoelectrical sections after studying the resistivity ranges of rock layers in the study area with the assistance of the software drawn resistivity sections. The reason behind this is that software depends

on the absolute layers resistivity values when connecting the geoelectrical zones. While the interpreter may use the rock resistivity ranges during the drawing of the geoelectrical sections. So that a combination between software displays and interpreter improvisations must be attended to give the best subsurface geological picture about the rocks electrical zones.

The three major steps of interpreting the VES resistivity curve are:

1- The manual drawn sounding curve before smoothing that shows discontinuous MN segments of the curve.
2- The manual smoothed and interpreted sounding resistivity curve by using the auxiliary point partial curve matching method. This method achieved by using the two layers standard curves set of Orellana and Mooney, 1966.
3- The enhanced resistivity curve by using the (IPI2Win) computer software. This enhancement represented by more accurate smoothing, treatment of equivalence and suppression and the reduction of r m s% to give more accurate results.

The final sounding curves interpretation results after enhancement could be tabeled in the form which appear in figure 20.

The (VES) interpretation results represent the thickness (h) in meter and resistivity(ρ) in ohm.m for each of the electrical zones within each of the (80) geoelectrical columns located under the midpoints of the (VES) points in the study area. Table (2-3), shows a sample of the results that obtained after interpretation.

VES No.	ρ1	h1	ρ2	h2	ρ3	h3	ρ4	h4	ρ5	h5	ρ6	Total Depth m	Curve Type
1	76.8	2.3	16.8	254	32.5							256.3	H
2	375	2.77	20.1	1	9.82	280	16.3	178	8.37			461.77	QHK
3	302	1.43	5.96	6.79	11.5	7.13	8.83	73.5	12.8	375	5.07	463.85	HKHK
4	103	1.72	46.3	3.93	18.6	48.9	8.27	192	12	405	10000	651.55	QQHA

Figure (20): A sample of interpretation results for the (VES).

Geoelectrical Sections

The (VES) curves are generally used to determine electrical resistivity variations as a function of depth. The inverse problem in resistivity interpretations states that for a given vertical electrical sounding curve, obtained from field observations, a geo-electrical section, which could produce can be constructed. Several possible solutions can be admitted unless other geophysical or geological information is available (Fadhil, 2009). Figure (21), shows example of a geoelectrical section at northeastern al-madinah almunawarah (harrat al-aqul) central arabian shield (Loni , 2005) .

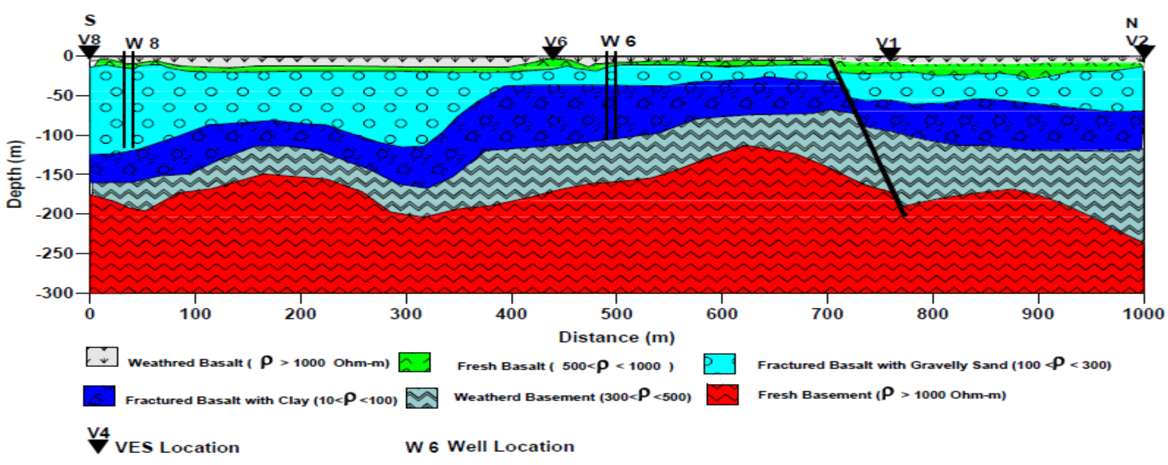


Figure (21): A Geoelectrical section in northeastern Al-Madinah AlMunawarah (Harrat Al-Aqul), central Arabian shield (Loni, 2005).

It is also possible to implement the computer software in constructing the geoelectrical sections, figure 22, showing one of computer drawn geoelectrical sections.

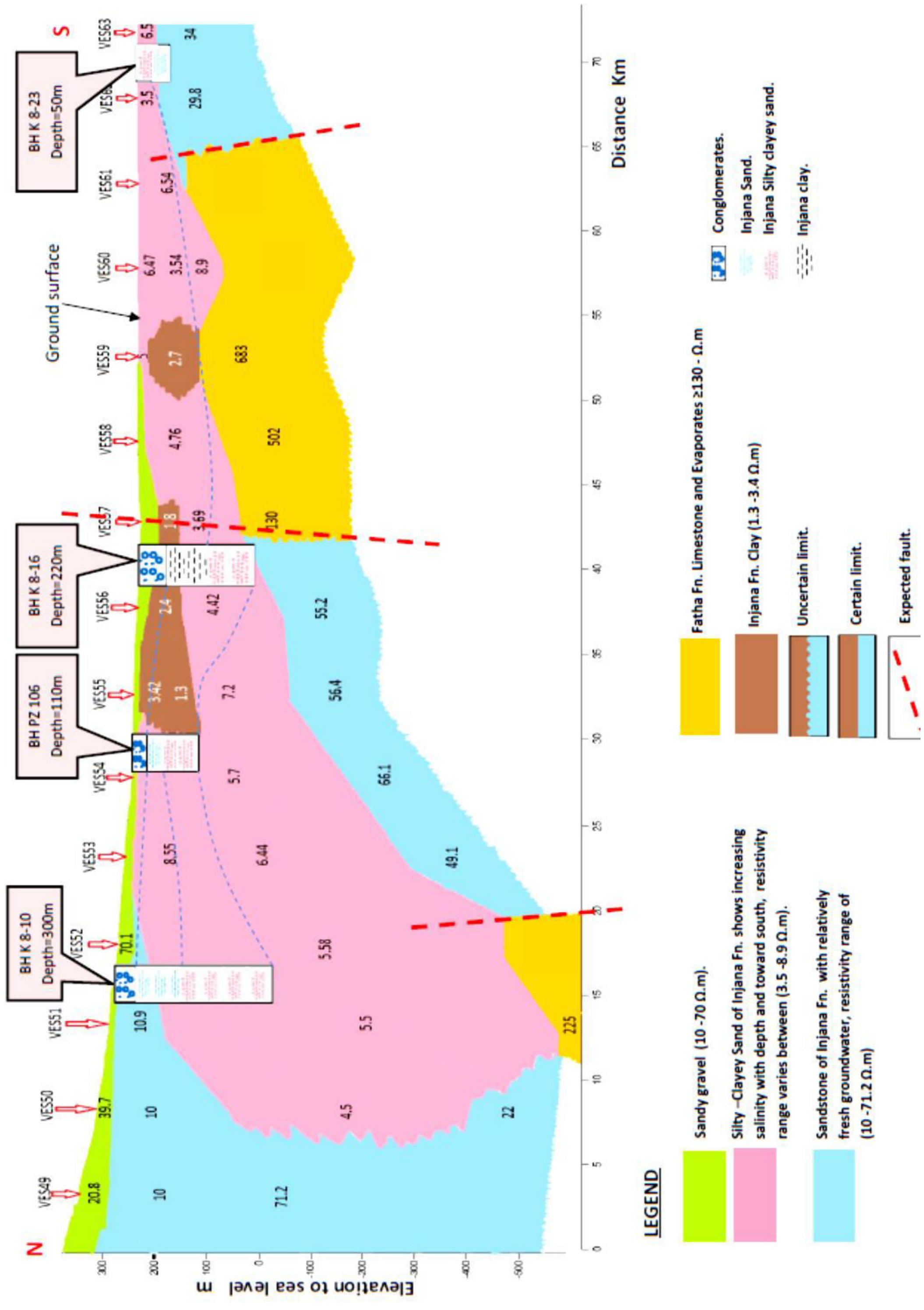


Figure (22): A computer drawn Geoelectrical section for 15 VES points by using Shlumberger array, (AL-Khafaji , 2014).

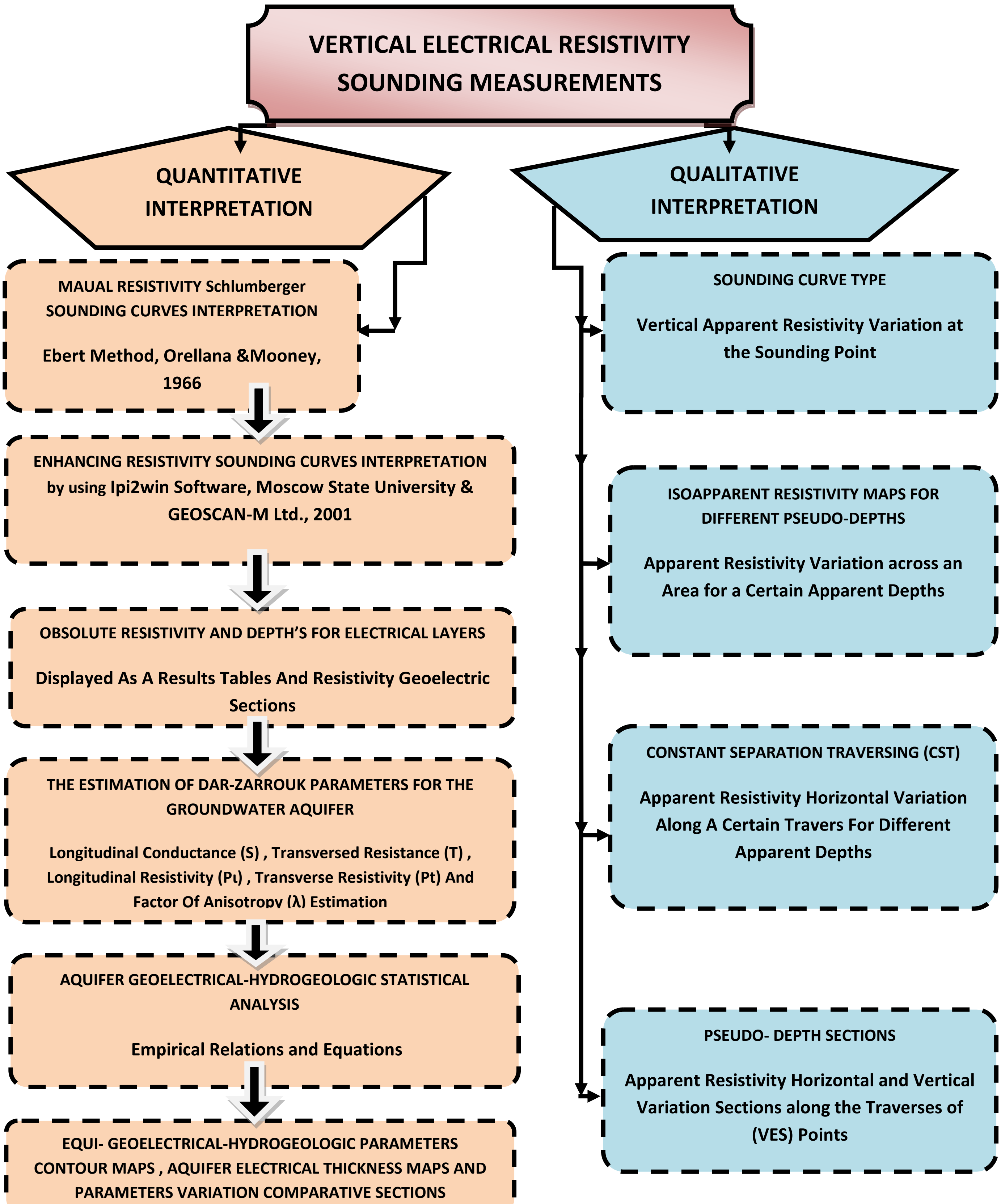


Figure (23): A flow chart showing the processing and interpretation procedures of (VES) field data.

It’s rather easier to describe the procedures of interpreting Vertical electrical Sounding data qualitatively and quantitatively by following the flow chart which shown in the figure 23.

The Main Types Of Resistivity Surveying Electrode Arrays

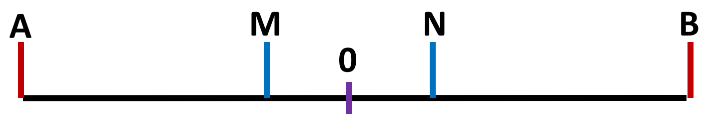
There are several forms of electrod arrays used in resistivity method , but they all used to measure the electric field (E) or the potential difference (ΔV), and they are all depends on the distances which separate between the potential electrodes MN and the distance which separate between the current current electrodes AB. The electrode arrays could be classified into the following:

A- Bipole-Dipole arrays
B- Dipole-Dipole arrays
C- Bipole-Bipole arrays

A- Bipole-Dipole arrays

In such electrode arrays the distance MN remains smaller that the distance AB, in other words (MN=1/5 – 1/10AB) and it includes the following sub-arrays:

1- Symmetrical and linear Schlumberger:
AB ≥ 5-10 MN



$$\rho_a = \pi \frac{(AB/2)^2 - (MN/2)^2}{MN} * \frac{\Delta V}{I}$$

$\pi \frac{(AB/2)^2-(MN/2)^2}{MN}$ Is called the geometrical factor (K).

Were: 0 is the array midpoint or the measurement point.

AB is the distance between the current electrodes.

MN : is the distance between the potential electrodes.

ΔV : is the potential difference.

I : the applied current .

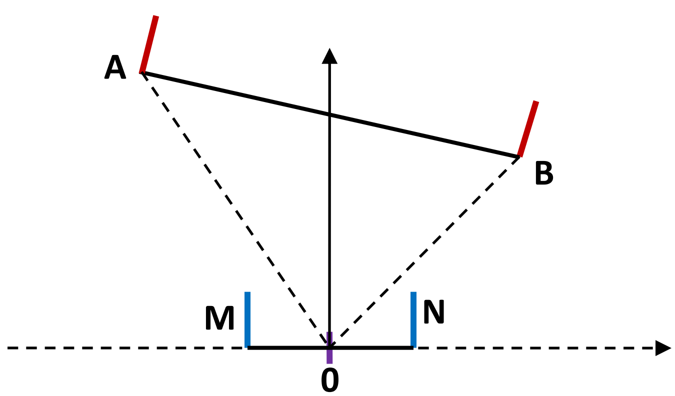
And when MN is close to zero then the ρ_a equation will be:

$$\boldsymbol{\rho_a = \pi (AB/2)^2 * \frac{E}{I}}$$

Were E: is the electric field.

2- Asymmetrical or non linear Schlumberger array:

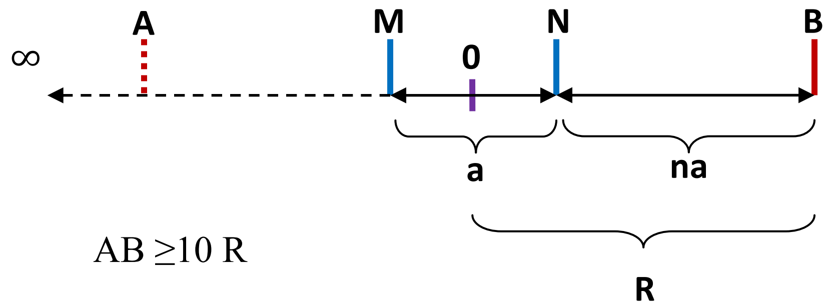
This array gives the same ρ_a value of the symmetrical - linear Schlumberger array, but with more precise geometrical factor.



In this array the distance A0 is equal to distance B0.

A0=B0 , and, 0.75≤ A0/ B0 ≤ 1.3

3-Half Schlumberger or (pole-Dipole) array:



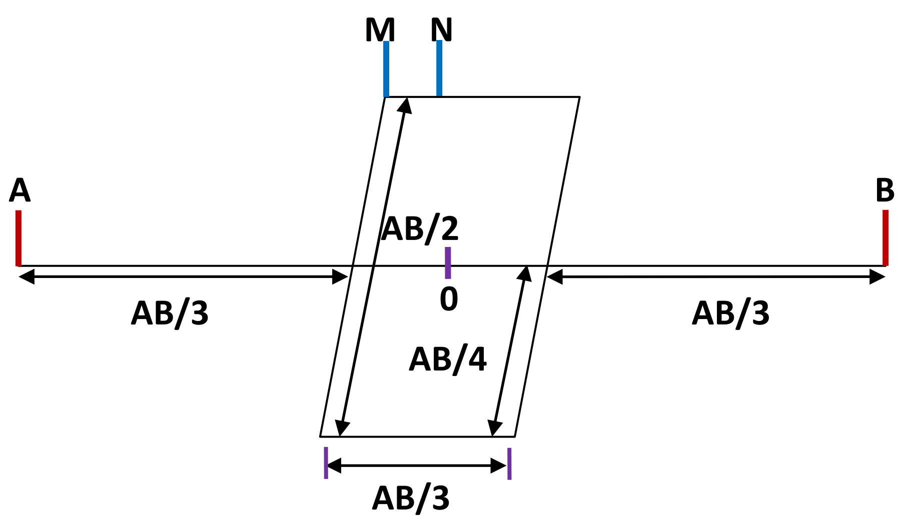
AB ≥10 R

n=BM/BA

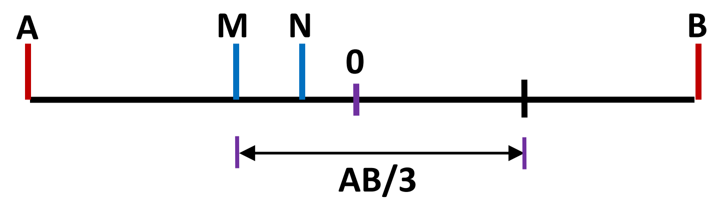
R is the distance from the midpoint to the current pole

$$\mathbf{K= 2\pi n\ (n+1)\ a}$$

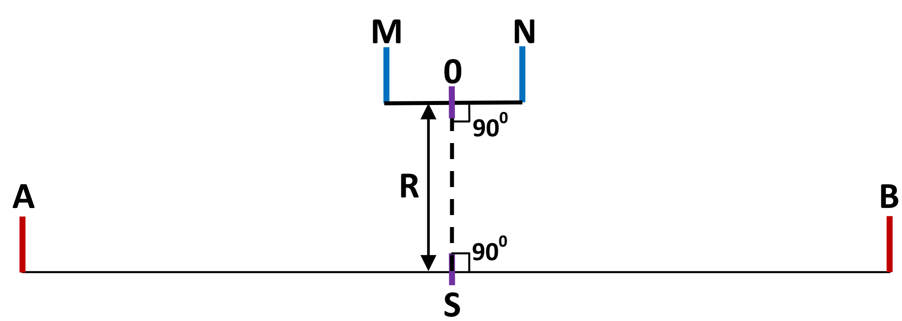
4- AB- Rectangular array:



5- AB- profile array:

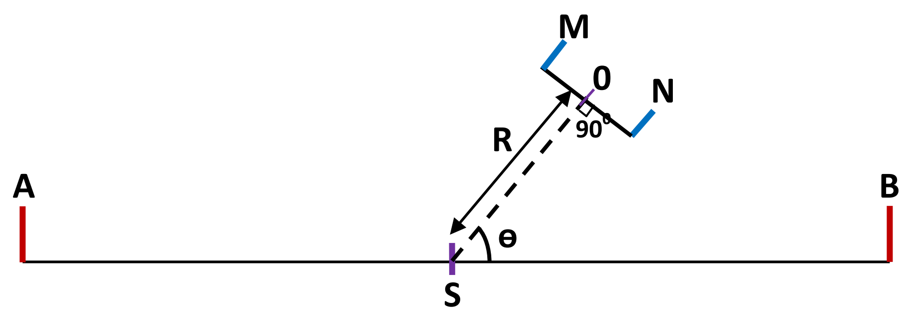


6- Bipole-Dipole Equatorial:



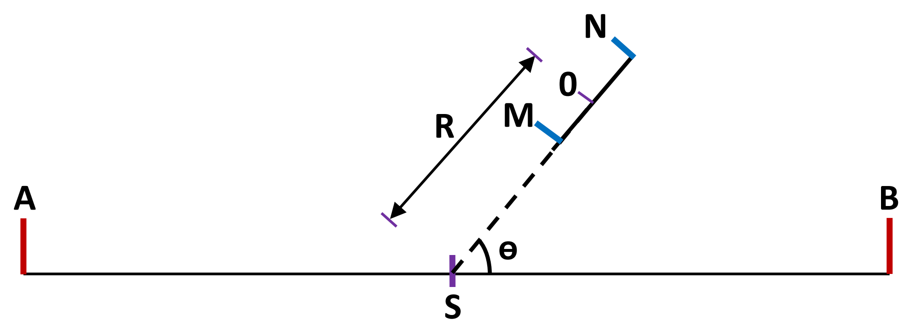
$R \leq 3\text{-}AB$

7-Bipole-Dipole Azimuthal:



R ≤ 3-5AB

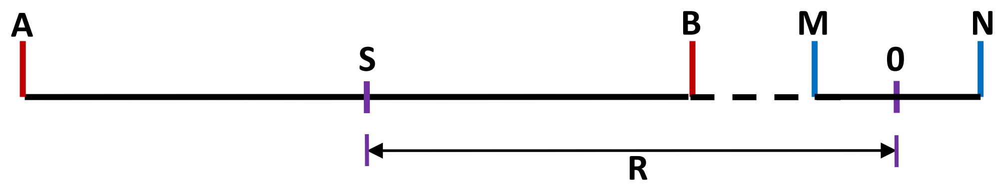
8-Bipole-Dipole Radial:



R ≤ 3-5AB

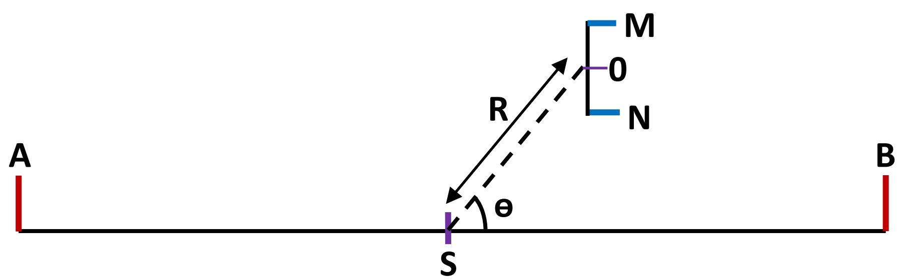
MN=1/5 – 1/10 AB

9-Bipole-Dipole Polar:



R ≥ 3-5AB

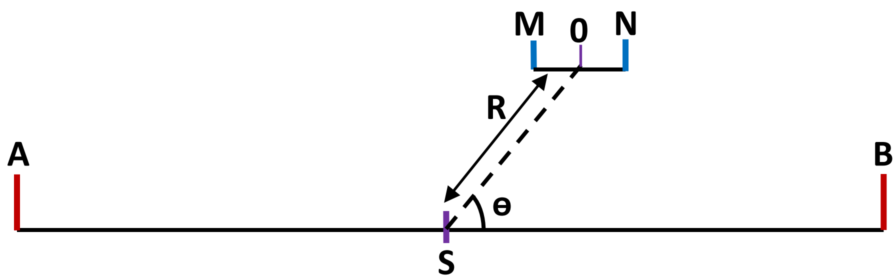
10-Bipole-Dipole Perpendicular:



R ≥ 3-5AB

MN=1/5 – 1/10 AB

11-Bipole-Dipole Parallel:

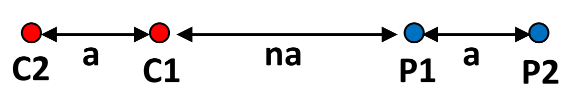


$R \geq 3\text{-}5AB$

B-Dipole-Dipole arrays

In these types of electrode arrays the distance between the current electrodes (AB) or (C1C2) is almost equal to the distance between the potential electrodes(MN) or (P1P2), but the distance between the midpoint of current electrodes and the midpoint between the potential electrodes (R) or (na) is always larger or equal to 1/3AB to 1/5 AB. The dipole –dipole arrays could be classified to the following arrays:

1- Ordinary Dipole-Dipole array (Axial or Polar):



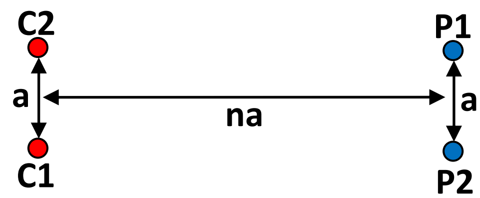
R=3-5AB

$$K=\pi n(n+1)(n+2)a$$

$$\rho_a = K\,(\Delta V/I)$$

Where: n = C1P1/C1C2, and called the spacing factor.

2- Equatorial Dipole-Dipole array :



AB=MN

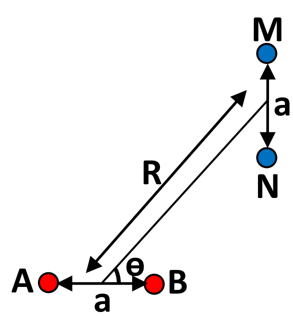
na=b=R

R or b ≥3-5AB

$$L=\sqrt{a^2+b^2}$$

$$K=\frac{2\pi Lb}{L-b}$$

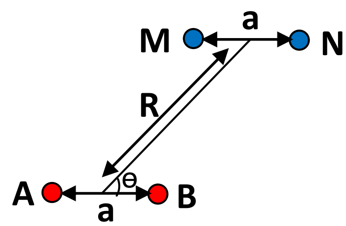
3-Perpendicular Dipole-Dipole array:



AB=MN

R ≥3-5AB

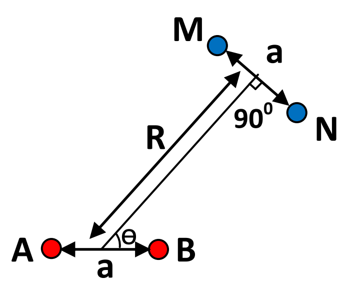
4-Parallel Dipole-Dipole array:



AB=MN

R ≥3-5AB

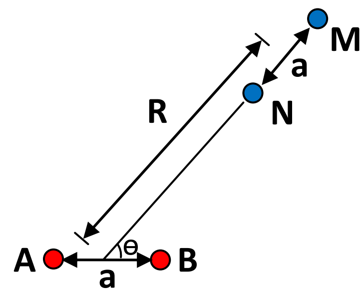
5-Azimuthal Dipole-Dipole array:



AB=MN

R ≥3-5AB

6-Radial Dipole-Dipole array:



AB=MN

R ≥3-5AB

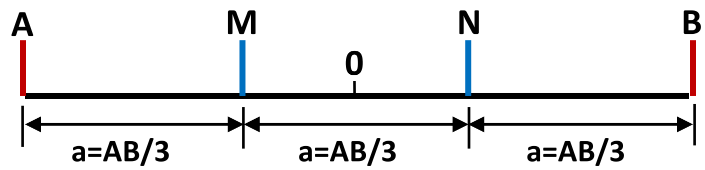
C-Bipole-Bipole arrays

Used to measure the potential difference ΔV between MN, where: MN ≥ 1/3AB .

These arrays could be classified as the following:

1- Wenner array

MN = 1/3 AB

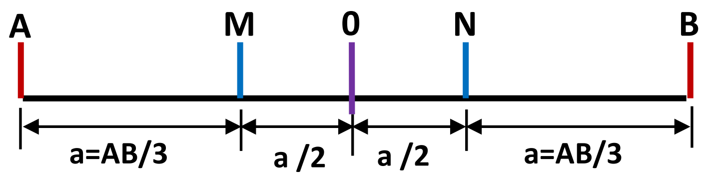


$$K=2\pi a$$

$$\rho_{a\ winner} = 2\pi a \frac{\Delta V}{I}$$

2- Lee Partitioning array:

This array is similar to Wenner array but with an additional electrode at the midpoint 0 .

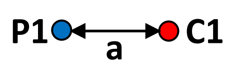


$$K=4\pi a$$

$$\rho_{a\ L.P.}= 4\pi a\ \frac{\Delta V}{I}$$

3- Pole- Pole array:

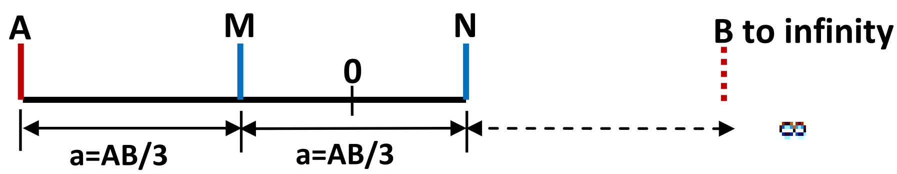
In this array two electrodes from the four electrodes, one for potential and other for current are stacked in a very far distance or to infinity.



$$K=2\pi a$$

$$\rho_a= 2\pi a\ \frac{\Delta V}{I}$$

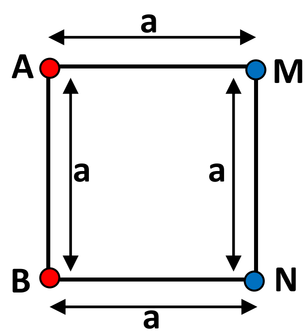
4- Half-Wenner or Three electrode array:



$$K=4\pi a$$

$$\rho_a= 4\pi a\ \frac{\Delta V}{I}$$

5- Square array:



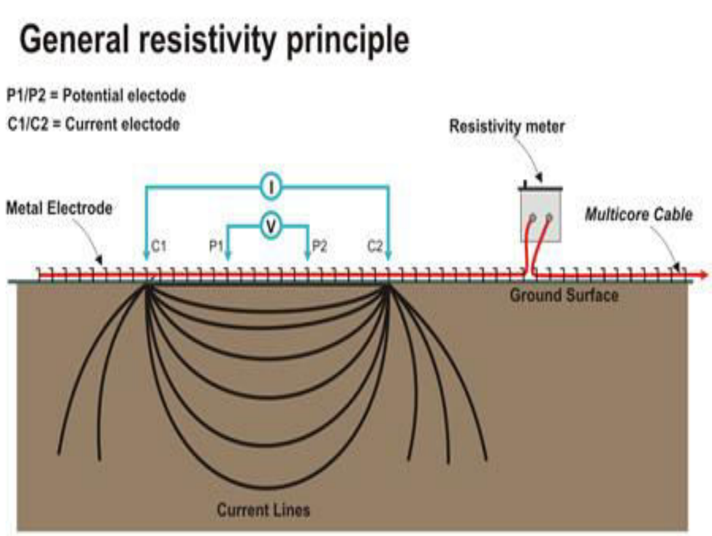
$$K_{Sq.} \frac{2\pi a}{2 - 2\sqrt{2}}$$

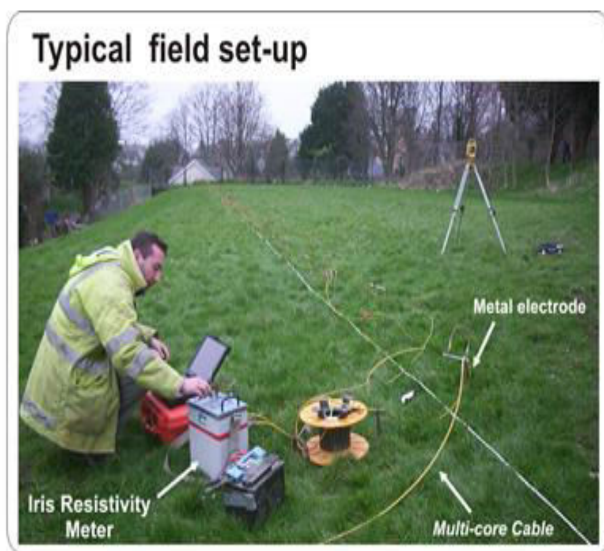
$$\rho_{a\ Sq.} = \frac{2\pi a}{2 - 2\sqrt{2}} \frac{\Delta V}{I}$$

Electrical Resistivity Tomography (ERT)

The Resistivity technique is a useful method for characterizing the sub-surface materials in terms of their electrical properties. Variations in electrical resistivity (or conductivity) typically correlate with variations in lithology, water saturation, fluid conductivity, porosity and permeability, which may be used to map stratigraphic units, geological structure, sinkholes, fractures and groundwater.

The acquisition of resistivity data involves the injection of current into the ground via a pair of electrodes and then the resulting potential field is measured by a corresponding pair of potential electrodes. The field set-up requires the deployment of an array of regularly spaced electrodes, which are connected to a central control unit via multi-core cables. Resistivity data are then recorded via complex combinations of current and potential electrode pairs to build up a pseudo cross-section of apparent resistivity beneath the survey line. The depth of investigation depends on the electrode separation and geometry, with greater electrode separations yielding bulk resistivity measurements from greater depths.



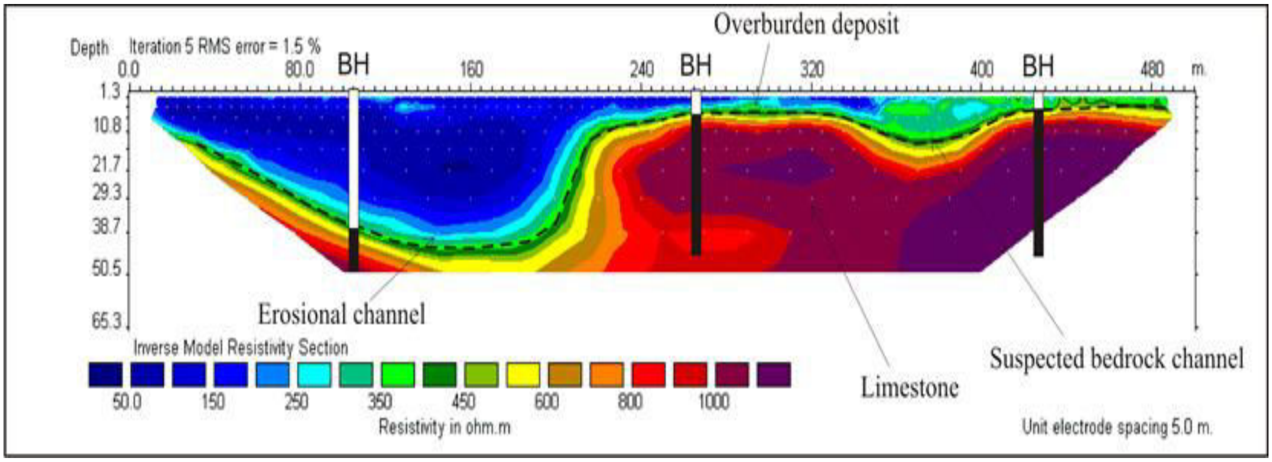


The recorded data are transferred to a PC for processing. In order to derive a cross-sectional model of true ground resistivity, the measured data are subject to a finite-difference inversion process via RES2DINV (ver 5.1) software.

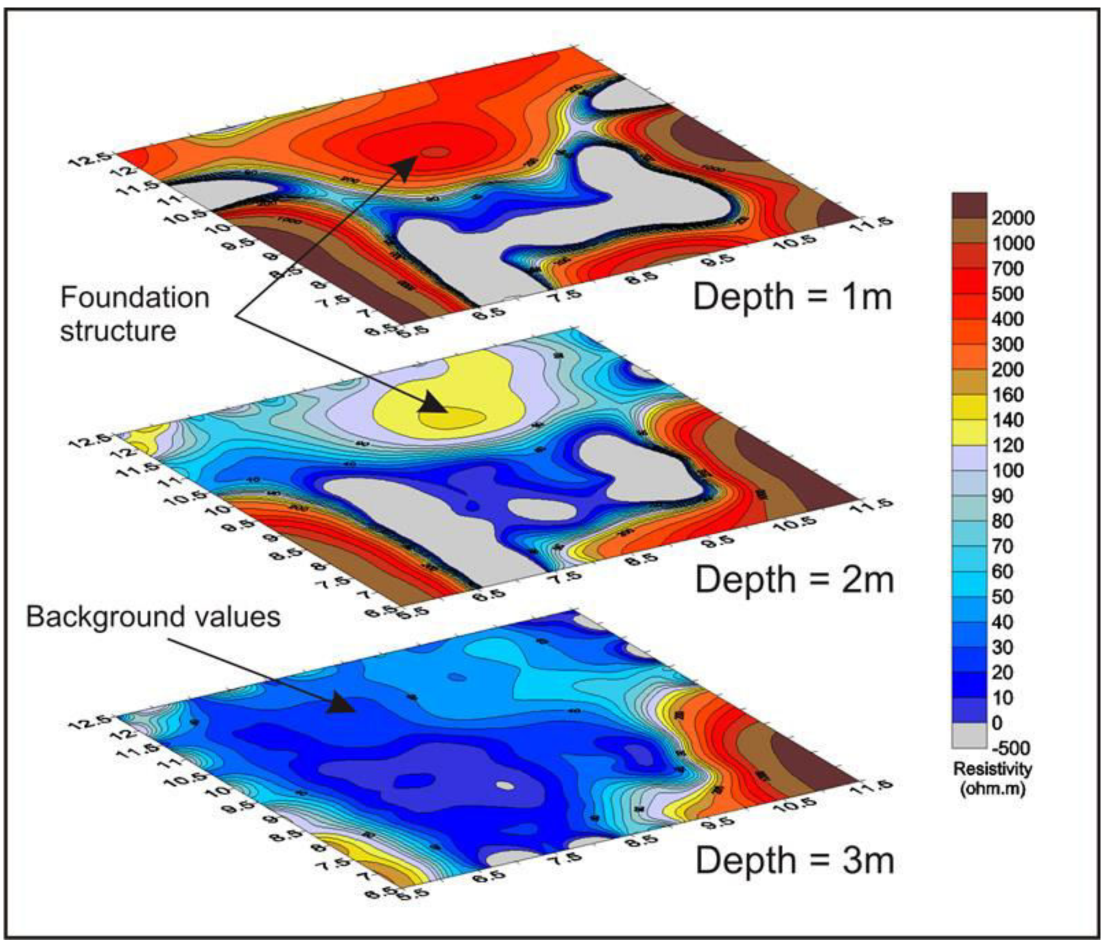
Data processing is based on an iterative routine involving determination of a two-dimensional (2D) simulated model of the subsurface, which is then compared to the observed data and revised. Convergence between theoretical and observed data is achieved by non-linear least squares optimization. The extent to which the observed and calculated theoretical models agree is an indication of the validity of the true resistivity model (indicated by the final root-mean-squared (RMS) error).

The true resistivity models are presented as color contour sections revealing spatial variation in subsurface resistivity. The 2D method of presenting resistivity data is limited where highly irregular or complex geological features are present and a 3D survey maybe required. Geological materials have characteristic resistivity values that enable identification of boundaries between distinct lithologies on resistivity cross-sections. At some sites, however, there are overlaps between the ranges of possible resistivity values for the targeted materials which therefore necessitate use of other geophysical surveys and/or drilling to confirm the nature of identified features.

Constraints: Readings can be affected by poor electrical contact at the surface. An increased electrode array length is required to locate increased depths of interest therefore the site layout must permit long arrays. Resolution of target features decreases with increased depth of burial.



As part of a hydrological study, a series of resistivity tomography profile lines were acquired to map variations within the overburden thickness. The example section above displays an extensive erosional channel feature together with more subtle overburden thickness variations.



The figure above shows a 3D resistivity survey which carried out to map the lateral and vertical extent of buried foundations in engineering site application. The grey zones represent noisy data due to buried services and the high resistivity values (red) reflect the foundation material. The resistivity suggests that the foundations extend to a maximum depth of 2m.

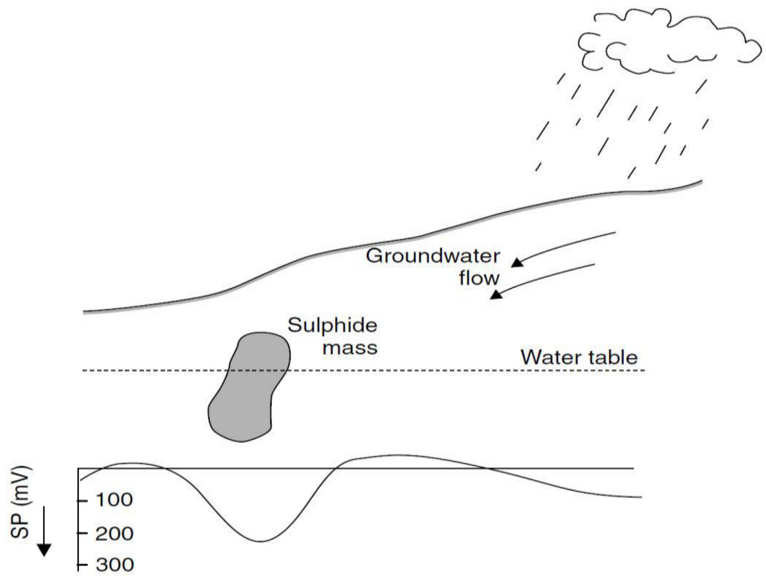
Electrical Self Potential (SP) Surveys

SP surveys were at one time popular in mineral exploration because of their low cost and simplicity. They are now little used because some near-surface ore bodies that are readily detected by other electrical methods produce no SP anomaly.

Origins of natural potentials

Natural potentials of as much as 1.8 V have been observed where Alunite (**Alunite** is a hydroxylated aluminium potassium sulfate mineral, formula $KAl_3(SO_4)_2(OH)_6$), weathers to sulphuric acid, but the negative anomalies produced by sulphide ore bodies and graphite are generally less than 500 mV. The conductor should extend from the zone of oxidation near the surface to the reducing environment below the water table, thus providing a low-resistance path for oxidation–reduction currents (see the following figure).

From the following figure Sources of SP effects. The sulphide mass straddling the water table concentrates the flow of oxidation–reduction currents, producing a negative anomaly at the surface. The downslope flow of groundwater after rain produces a temporary SP, in this case inversely correlated with topography.



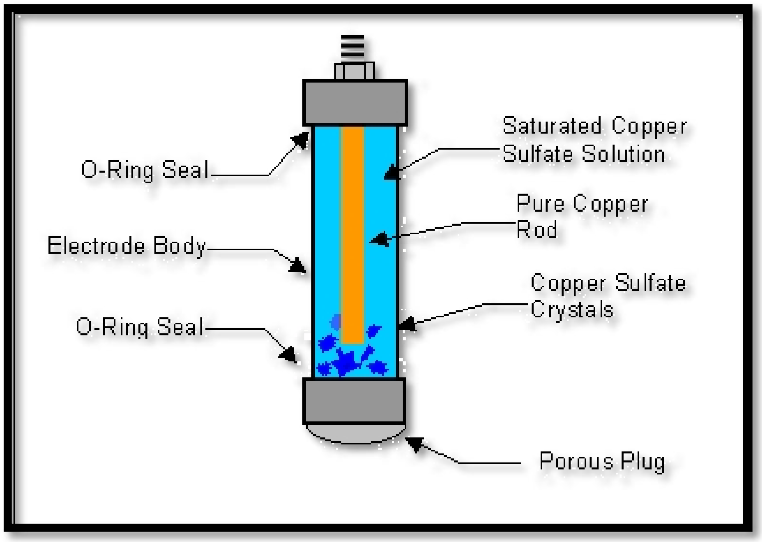
sometimes SP surveys are useful in hydrogeology but can make mineral exploration surveys inadvisable for up to a week after heavy rain.

Movements of steam or hot water can explain most of the SPs associated with geothermal systems, but small (<10 mV) voltages, which may be positive or negative, are produced directly by temperature differences. Geothermal SP anomalies tend to be broad (perhaps several kilometres across) and have amplitudes of less than 100 mV, so very high accuracies are needed.

Small alternating currents are induced in the Earth by variations in the ionospheric component of the magnetic field and by thunderstorms. Only the long-period components of the associated voltages, seldom amounting to more than 5 mV, are detected by the DC voltmeters used in SP surveys. If, as is very occasionally the case, such voltages are significant, the survey should be repeated at different times of the day so that results can be averaged.

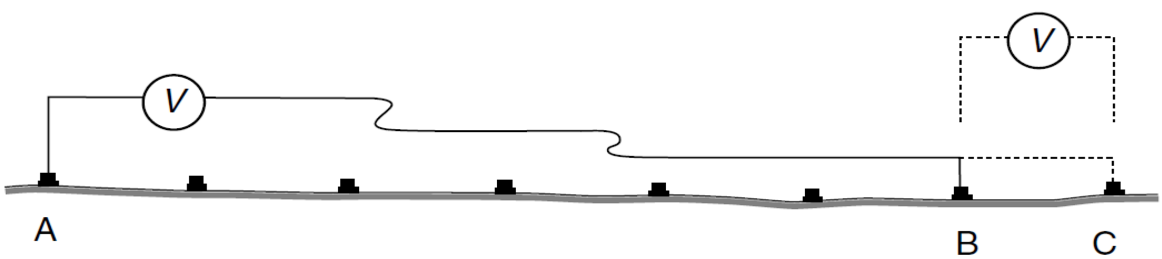
SP surveys

Voltmeters used for SP work must have millivolt sensitivity and very high impedance so that the currents drawn from the ground are negligible. A pair of Copper/ copper-sulphate 'pot' electrodes are almost universal is used; see the following figure, and linked to the meter by lengths of insulated copper wire.



An SP survey can be carried out by using two electrodes separated by a small constant distance, commonly 5 or 10 m, to measure average field gradients. The method is useful if cable is limited, but errors tend to accumulate and coverage is slow because the voltmeter and both electrodes must be moved for each reading. More commonly, voltages are measured in relation to a fixed base. One electrode and the meter remain at this point and only the second electrode is moved. Sub-bases must be established if the cable is about to run out or if distances become too great for easy communication. Voltages measured from a base and a sub-base can be related provided that the potential difference between the two bases is accurately known.

The following figure shows how a secondary base can be established. The end of the cable has almost been reached at field point B, but it is still possible to obtain a reading at the next point, C, using the original base at A. After differences have been measured between A and both B and C, the field electrode is left at C and the base electrode is moved to B. The potential difference between A and B is thus estimated both by direct measurement and by subtracting the B to C voltage from the directly measured A to C voltage. The average difference can be added to values obtained with the base at B to obtain values relative to A.



The figure above shows how to move a base station in an SP survey. The value at the new base (B) relative to A is measured directly and also indirectly by measurements of the voltage at the field point C relative to both bases. The two estimates of the voltage difference between A and B are then averaged.

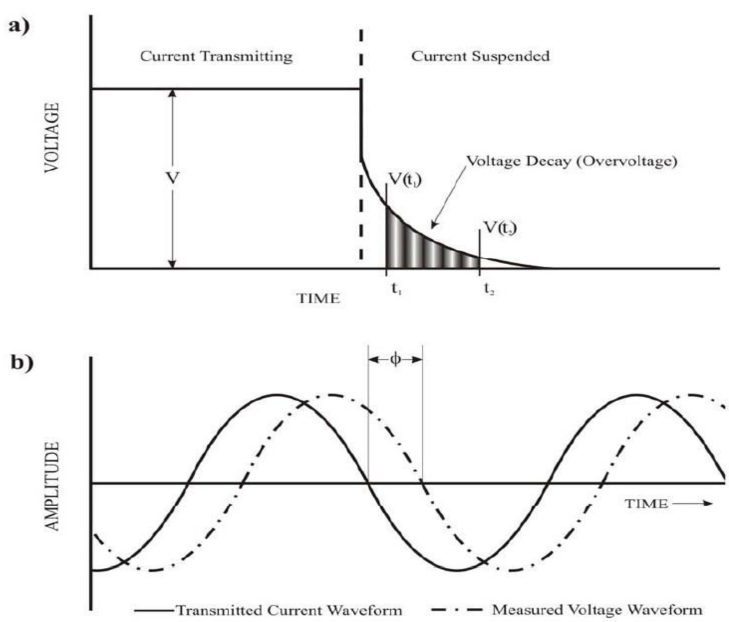
Constraints

If two estimates of a base/sub-base difference disagree by more than one or two mill volts, work should be stopped until the reason has been determined. Usually it will be found that copper sulphate solution has either leaked away or become under saturated. Electrodes should be checked every two to three hours by placing them on the ground a few inches apart. The voltage difference should not exceed 1 or 2 mV. Accumulation of errors in large surveys can be minimized by working in closed and interconnecting loops around each of which the voltages should sum to zero .

Induced Polarization (IP) Method

Conrad Schlumberger (Dobrin 1960) probably was first to report the induced polarization phenomenon, which he called "provoked polarization." While making conventional resistivity measurements, he noted that the potential difference, measured between the potential electrodes, often did not drop instantaneously to zero when the current was turned off. Instead, the potential difference dropped sharply at first, then gradually decayed to zero after a given interval of time. Certain layers in the ground can become electrically polarized, forming a battery when energized with an electric current. Upon turning off the polarizing current, the ground gradually discharges and returns to equilibrium.

The study of the decaying potential difference as a function of time is now known as the study of induced polarization (IP) in the time domain (see the figure below).

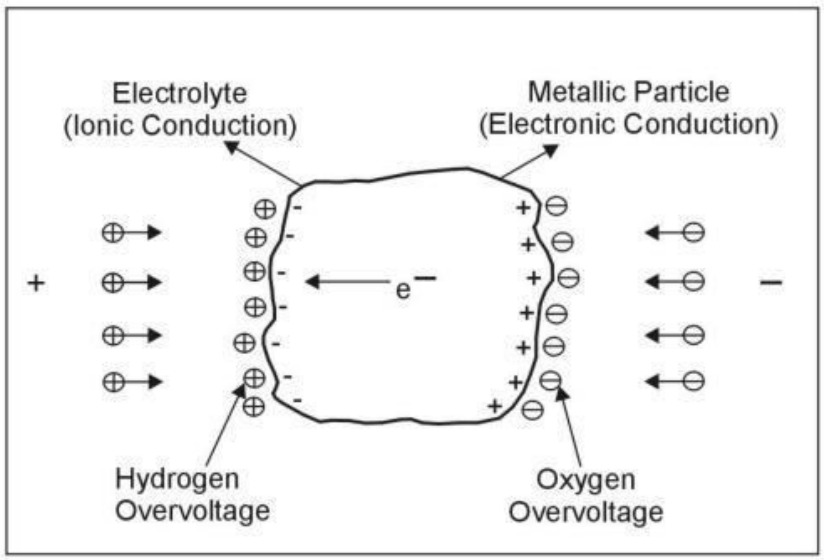


In this method the geophysicist looks for portions of the earth where current flow is maintained for a short time after the applied current is terminated. Another technique is to study the effect of alternating currents on the measured value of resistivity, which is called IP in the "frequency domain" (see the previous figure - b). In this method the geophysicist tries to locate portions of the earth where resistivity decreases as the frequency of applied current is increased. The induced electrical polarization method is widely used in exploration for ore bodies, principally of disseminated sulfides. Use of IP in geotechnical and engineering applications has been limited, and has been used mainly for groundwater exploration. Groundwater IP studies generally have been made with time-domain IP.

General Theory of the IP Effect

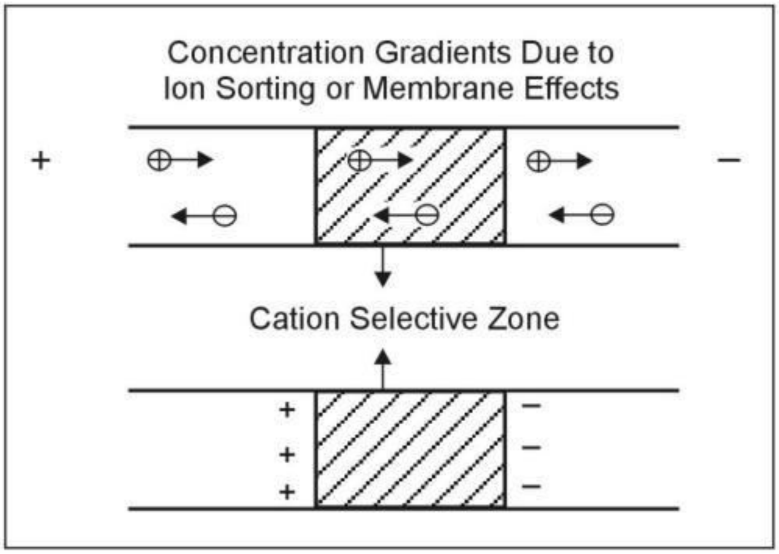
The origin of induced electrical polarization is complex and is not well understood. This is primarily because several physio-chemical phenomena and conditions are likely responsible for its occurrence. Only a fairly simple discussion will be given here. According to Seigel (1970), when a metal electrode is immersed in a solution of ions of a certain concentration and valence, a potential difference is established between the metal and the solution sides of the interface. This difference in potential is an explicit function of the ion concentration, valence, etc. When an external voltage is applied across the interface, a current is caused to flow, and the potential drop across the interface changes from its initial value. The change in interface voltage is called the "overvoltage" or "polarization" potential of the electrode. Overvoltages are due to an accumulation of ions on the electrolyte side of the interface waiting to be discharged. The time constant of buildup and decay is typically several tenths of a second.

Overvoltage is therefore established whenever current is caused to flow across an interface between ionic and electronic conduction. In normal rocks, the current that flows under the action of an applied emf does so by ionic conduction in the electrolyte in the pores of the rock. There are, however, certain minerals that have a measure of electronic conduction (almost all the metallic sulfides (except sphalerite) such as pyrite, graphite, some coals, magnetite, pyrolusite, native metals, some arsenides, and other minerals with a metallic lustre). The figure below represents a simplified representation of how over voltages are formed on an electronic conducting particle in an electrolyte under the influence of current flow.



Overvoltage on a metallic particle in electrolyte. (Seigel 1970; Geological Survey of Canada)

The most important sources of nonmetallic IP in rocks are certain types of clay minerals (Vacquier 1957, Seigel 1970). These effects are believed to be related to electrodialysis of the clay particles. This is only one type of phenomenon that can cause "ion-sorting" or "membrane effects." For example, the figure below shows a cation-selective membrane zone in which the mobility of the cation is increased relative to that of the anion, causing ionic concentration gradients and therefore polarization.



Nonmetallic induced polarization agent. (Seigel 1970; Geological Survey of Canada)

A second group of phenomena includes electro kinetic effects that produce voltage gradients through the `streaming potential' phenomenon. These voltage gradients will have the same external appearance as polarization effects due to separation of g Electrokinetic effects seem less important than membrane effects in the overall polarization picture.

IP surveys are perhaps the most useful of all geophysical methods in mineral exploration, being the only ones responsive to low-grade disseminated mineralization. There are two main mechanisms of rock polarization and three main ways in which polarization effects can be measured. In theory the results obtained by the different techniques are equivalent but there are practical differences. These ways of measuring polarization effects are :

1- Membrane polarization:

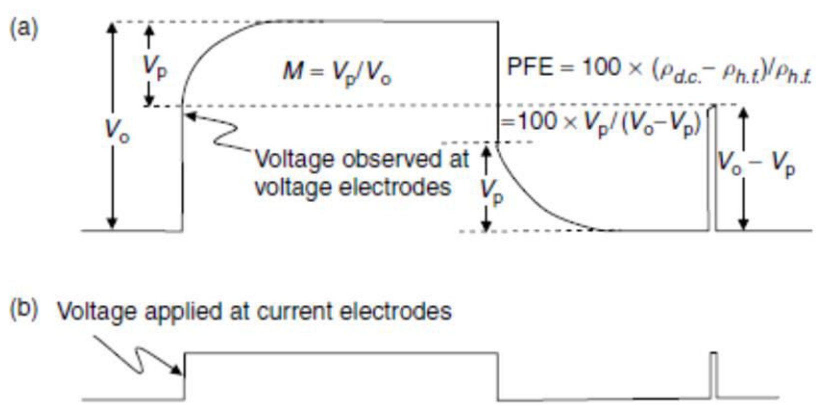
The surfaces of clays and some other platy or fibrous minerals are negatively charged and cause *membrane polarization* in rocks with small pore spaces. Positive ions in the formation waters in such rocks congregate near the pore walls, forming an *electrical double layer*. If an electric field is applied, the positive ion clouds are distorted and negative ions move into them and are trapped, producing concentration gradients that impede current flow. When the applied field is removed, a reverse current flows to restore the original equilibrium.

2- Electrode polarization:

This *electrode polarization* occurs not merely at artificial electrodes but wherever grains of electronically conducting minerals are in contact with the groundwater. The degree of polarization is determined by the surface area, rather than the volume, of the conductor present, and polarization methods are thus exceptionally well suited to exploration for disseminated *porphyry* ores. Strong anomalies are also usually produced by massive sulphide mineralization, because of surrounding disseminated haloes. Although, for equivalent areas of active surface, electrode polarization is the stronger mechanism, clays are much more abundant than sulphides and most observed IP effects are due to membrane polarization.

3-The square wave in chargeable ground:

When a steady current flowing in the ground is suddenly terminated, the voltage V_o between any two grounded electrodes drops abruptly to a small *polarization voltage* V_p and then declines asymptotically to zero.



The figure above shows ground response to a square-wave signal and to a spike impulse. The ratio of V_o to V_p is seldom more than a few percent. Input voltage waveform is for reference only. In practice its amplitude will be many times greater than the measured voltage, the exact values depending on the array being used.

Similarly, when current is applied to the ground, the measured voltage first rises rapidly and then approaches V_o symptotically (see the previous figure). Although in theory V_o is never reached, in practice the difference is not detectable after about a second.

Chargeability is formally defined as the polarization voltage developed across a unit cube energized by a unit current and is thus in some ways analogous to magnetic susceptibility. The *apparent chargeability* of an entire rock mass is defined, in terms of the square wave shown in the previous figure, as the ratio of V_p to V_o. This is a pure number but in order to avoid very small values it is generally multiplied by a thousand and quoted in milli-volts per volt. The ratio of V_p to V_o cannot be measured directly since electromagnetic transients are dominant in the first tenth of a second after the original current ceases to flow. The practical definition of time-domain chargeability, which is in terms of the decay voltage at some specified delay time, is only tenuously linked to the theoretical definition. Not only do different instruments use different delays, but also it was originally essential and is still quite common to measure an area under the decay curve using integrating circuitry, rather than an instantaneous voltage. The results then depend on the length of the integration period as well as on the delay and are quoted in milliseconds.

4-Frequency effects:

The previous figure also shows that if a current were to be terminated almost immediately after being introduced, a lower apparent resistivity, equal to $(Vo - Vp)/I$ multiplied by the array geometrical factor, would be calculated. The IP frequency effect is defined as the difference between the ‘high frequency ‘and ‘DC’ resistivities, divided by the high-frequency value. This is multiplied by 100 to give an easily handled whole number, the *percent frequency effect* (PFE). The origin of the theoretical relationship between the PFE and the chargeability could be given by:

$$\mathbf{M = [PFE] / (100 + [PFE])}$$

and illustrated in the previous figure.

Because of electromagnetic transients, the theoretical PFE cannot be measured and the practical value depends on the frequencies used. To cancel telluric and SP noise, ‘DC’ measurements are taken with current reversed at intervals of the order of a few seconds, while the ‘high’ frequencies are usually kept below 10 Hz to minimize electromagnetic induction.

5-Metal factors:

A PFE can be divided by the DC resistivity to give a quantity which, multiplied by 1000, 2000 or 2000π, produces a number of convenient size known as the *metal factor*. Metal factors emphasize rock volumes that are both polarizable and conductive and which may therefore be assumed to have a significant sulphide (or graphite) content. Although this may be useful when searching for massive sulphides, low resistivity is irrelevant and can be actually misleading in exploration for disseminated deposits. As usual when factors that should be considered separately are combined, the result is confusion, not clarification.

6-Phase:

The square-wave which appears in page 53, can be resolved by Fourier analysis into sinusoidal components of different amplitudes and frequencies. The asymmetry of the voltage curve implies frequency-dependent phase shifts between the applied current and the measured voltage. In *spectral* IP surveys, these shifts are measured, in milliradians, over a range of frequencies.

Time-domain IP Surveys

Large primary voltages are needed to produce measurable IP effects. Current electrodes can be plain metal stakes but non-polarizing electrodes must be used to detect the few millivolts of transient signal.

Time-domain transmitters

A time-domain transmitter requires a power source, which may be a large motor generator or a rechargeable battery. Voltage levels are usually selectable within a range of from 100 to 500 V. Current levels, which may be controlled through a current limiter, must be recorded if apparent resistivities are to be calculated as well as IPs.

Current direction is alternated to minimize the effects of natural voltages, and cycle times can generally be varied from 2 to 16 seconds. One second each for energization and reading is not generally sufficient for reliable results, while cycles longer than 8 seconds unreasonably prolong the survey.

Time-domain receivers

A time-domain receiver measures primary voltage and one or more decay voltages or integrations. It may also be possible to record the SP, so that chargeability, resistivity and SP data can be gathered together. Early *Newmont* receivers integrated from 0.45 to 1.1 seconds after current termination. The SP was first balanced out manually and the primary voltage was then *normalized* by adjusting an amplifier control until a galvanometer needle swung between defined limits. This automatically ratioed V_p to V_o for the M values recorded by a second needle. Experienced operators acquired a 'feel' for the shape of the decay curve from the rates of needle movement and were often able to recognize electromagnetic transients where these persisted into the period used for voltage sampling.

With purely digital instruments, the diagnostic information provided by a moving needle is lost and enough cycles must be observed for statistical reduction of noise effects. Digital systems allow more parameters to be recorded and very short integration periods, equivalent to instantaneous readings. Natural SPs are now compensated (*backed-off* or *bucked-out*) automatically rather than manually. Memory circuits store data and minimize note taking.

The receiver must be tuned to the cycle period of the transmitter so that it can lock on to the transmissions without use of a reference cable (which could carry inductive noise). Cycle times of 4, 8 or 16 seconds are now generally favoured. Changing the cycle time can produce quite large differences in apparent chargeability, even for similar delay times, and chargeabilities recorded by different instruments are only vaguely related.

Decay-curve analysis

With readings taken at several different delay times, curve analysis can be attempted. A method suggested for use with Huntec receivers assumed that each decay curve was a combination of two exponential decays, corresponding to electrode and membrane polarizations, which could be isolated mathematically. This is far too drastic a simplification and the separation, using a limited number of readings, of two exponential functions that have been added together is in any case virtually impossible in the presence of even small amounts of noise. Nonetheless, research continues into the controls on decay-curve shapes, and chargeabilities should be recorded at as many decay times as are conveniently possible in areas of interesting anomaly. In non-anomalous areas a single value generally suffices.

IP Data

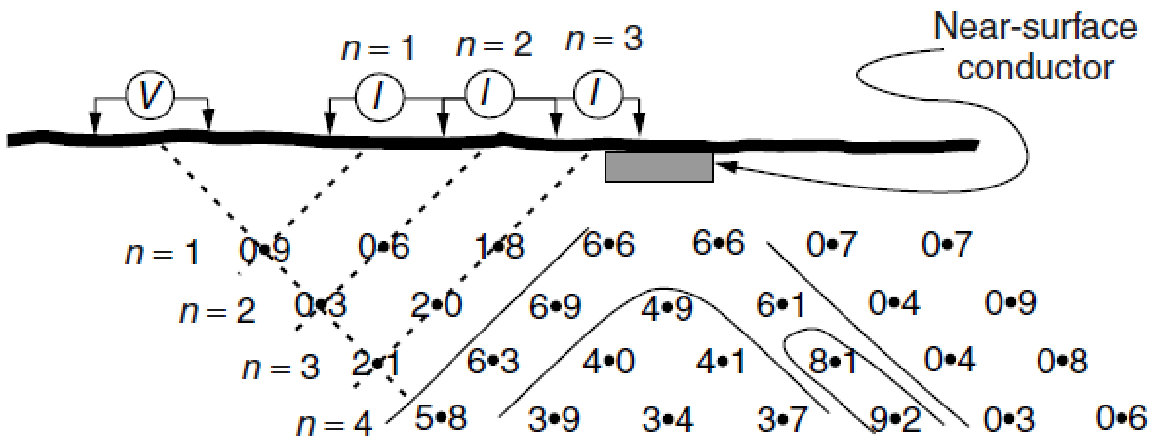
The methods used to display IP data vary with the array used. Profiles or contour maps are used for gradient arrays, while dipole–dipole data are almost always presented as pseudo-sections. In surveys with either array, the spacing between the voltage electrodes should not be very much greater than the width of the smallest target that would be of interest.

Gradient array data

Current paths are roughly horizontal in the central areas investigated using gradient arrays, and chargeable bodies will be horizontally polarized. Profiles can be interpreted by methods analogous to those used for magnetic data, with approximate depths estimated by adopting the straight slop and Peters methods.

Dipole–dipole data

Dipole–dipole traverses at a single *n* value can be used to construct profiles but multispaced results are almost always displayed as pseudo-sections , (see the next figure).



The figure above showing the Pseudo-section construction. The three different positions of the current dipole correspond to three different multiples of the basic spacing. Measured values (of IP or resistivity) are plotted at the intersections of lines sloping at 45◦ from the dipole centers. The plotting 'point' often doubles as a decimal point for IP values. The pant's leg anomaly shown is typical of those produced by small, shallow bodies.

The relationships between the positions of highs on pseudosections and source body locations are even less simple with dipole–dipole than with Wenner arrays. In particular, the very common *pant's leg* anomaly , see the figure above is usually produced by a near-surface body with little extent in depth; every measurement made with either the current or the voltage dipole near the body will record high chargeability. Anomaly shapes are thus very dependent on electrode positions, and the directions of apparent dip are not necessarily the directions of dip of the chargeable bodies.

Even qualitative interpretation requires considerable experience as well as familiarity with model studies. Pseudo-sections are nearly always plotted in relation to horizontal baselines, even in rugged terrain. Referencing them to topographic profiles (using construction lines similar to those of the previous figure, but at 45◦ to the actual ground surface) has its dangers, since it might be taken as implying much closer correlation with true sub-surface distributions of resistivity and chargeability than actually exist. However, steep and varied slopes do influence dipole–dipole results and it is better that they be displayed than ignored.

Negative IPs and masking

Negative IP effects can be caused by power or telephone cables or by signal contribution sections, or by detecting lateral inhomogeneities.

Layering can also produce negative values, and can conceal deeper sources, most readily if both the surface and target layers are more conductive than the rocks in between. In these latter circumstances, the penetration achieved may be very small and total array lengths may need to be 10 or more times the desired exploration depth.

Interactions between conduction and charge in the earth are very complex, and interpreters generally need more reliable resistivity data than is provided by the dipole–dipole array, which performs poorly in defining layering. A small number of Wenner or Schlumberger expansions, carried out specifically to map resistivity, may prove invaluable. Also, any changes in surface conditions that might correlate with changes in surface conductivity should be noted. The detectability of ore is likely to be quite different beneath bare rock ridges and under an intervening swamp.

Spectral Induced Polarization

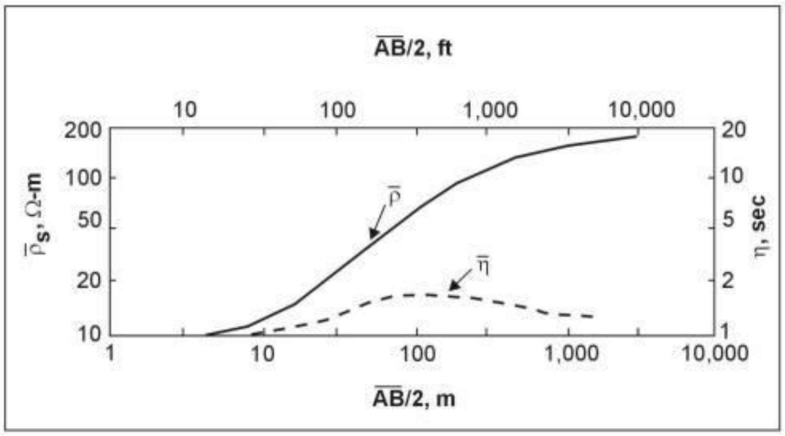
Spectral Induced Polarization (SIP) measures the variation of resistivity with frequency. The method, therefore, requires several resistivity measurements at different frequencies. These can be recorded in the frequency domain or in the time domain. In the time domain, where voltage measurements are recorded after the transmitting current has been turned off, the decaying voltage is sampled several times as it decays. Using the Fourier Transform, these data can be transformed into the frequency domain providing resistivity values at different frequencies. SIP measurements are occasionally used in mineral exploration to assist in identifying graphite and clay from sulfide mineralization. In addition, some information about the habit of the polarizable minerals could be obtained. It can also be used to map clay and, in some cases, contamination.

Sounding and Profiling

The techniques of sounding and profiling, used in resistivity measurements, are also used in the IP method. IP soundings are most commonly made using the Schlumberger array, pole-dipole array, or Wenner array, and usually in the time domain. The apparent chargeability η_a versus the electrode spacing a is plotted on logarithmic coordinates. The IP sounding curve is an interpreted curve matching procedures, either graphically, using sets of IP sounding master curves, or by computer. At present, only a few two-layer master curves (for the Wenner array) have been published in the United States (Seigel, 1959; Frische and von Buttlar, 1957). Three- and four-layer curves have been published in the Soviet Union.

An IP sounding curve can be of significant value in complementing a resistivity sounding curve. For example, the resistivity and IP sounding curves for the following four-layer geoelectric section are shown in the figure below:

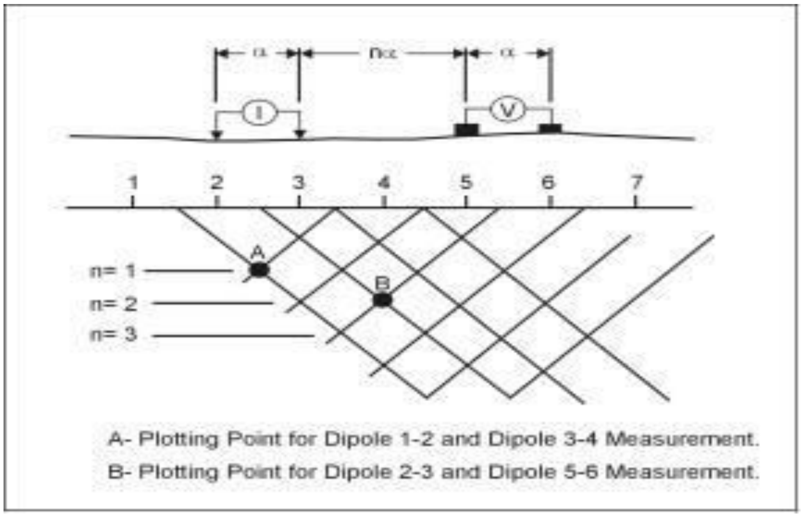
Layer No.	Thickness (m)	Resistivity (Ωm)	Chargeability (s)
1............	10	10	1
2............	10	160	1
3............	5	40	10
4............	infinity	infinity	1



Apparent resistivity and apparent chargeability (IP) sounding curves for a four-layer model. (Zohdy 1974)

From the figure above It is obvious that layer 3 cannot be distinguished on the four-layer resistivity curve (which resembles a two- or three-layer curve). But layer 3 is characterized by a different chargeability from the surrounding layers, and its presence is indicated clearly by the IP sounding curve.

When profiling, the pole-dipole or dipole-dipole, (see the figure below), arrays are used almost exclusively.



It can be easily employed in the field using short lengths of wire or multi-conductor cables allowing several values of the spacing multiplier (n) to be measured from one current dipole location. For one or two values of n, the IP and resistivity results are plotted as profiles. For more than two values of n, the profile method of presentation becomes confusing. A two-dimensional (usually called pseudosection) format has been developed to present the data (see the previous figure). This form of presentation helps the interpreter separate the effects of IP and resistivity variations along the line from vertical variations. The 45° angle used to plot the data is entirely arbitrary. The pseudosection plots are contoured, and the resulting anomalous patterns can be recognized as being caused by a particular source geometry and/or correlated from line to line. However, the contoured data are not meant to represent sections of the electrical parameters of the subsurface (Hallof, 1980). The pseudosection data plots are merely a convenient method for showing all of the data along one given line in one presentation.

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