Electrical Methods.

In this section we will look at the third potential field method, DC resistivity, and the related methods induced polarization (IP) and self potential (SP). The DC resistivity method is a way of estimating the electrical conductivity of geological formations, and so we will be looking at what makes rocks more or less conductive. One of the most important factors in rock conductivity is water content, and so DC resistivity methods are used extensively in groundwater exploration. However, DC methods also have applications in geothermal exploration (hot water is more conductive, steam is resistive) and mineral exploration (ore minerals are conductive).

Electromagnetic methods can also be used to study ground conductivity, but that will be the subject of the next section.

1. Introductory Theory.

Although in geophysical prospecting we are really interested in the effects of electric currents flowing in the Earth, electrical methods are founded in a fundamental property of matter called charge, which is measured in coulombs (C) and comes in two varieties, positive and negative. An electron/proton has a charge of 1.6×10^{-19} C. *Coulomb's Law* gives the force on a charge q_1 due to a charge q_2 at a distance R:

$$\mathbf{F} = \frac{q_1 q_2}{4\pi\epsilon_o R^2}\mathbf{r}$$

where ϵ_o is the permittivity of free space; $\epsilon_o \approx 8.85 \times 10^{-12}\mathrm{C}^2\mathrm{N}^{-1}\mathrm{m}^{-2}$ (or F/m; in cgs units $1/(4\pi\epsilon_o) = 1$). $\mathbf{r}$ is the unit vector in the direction of q_2 from q_1. Note that by this definition, like charges repel and unlike charges attract. The electric field, **E**, is defined as the force per unit charge and so the field due to a charge q is

$$\mathbf{E} = \frac{1}{4\pi\epsilon_o}\frac{q}{R^2}\mathbf{r}$$

and has units of N/C but more usefully V/m .

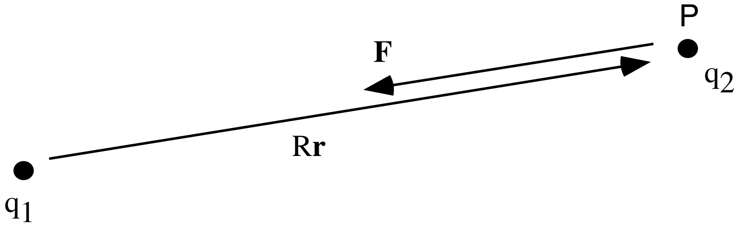


Fig. E.01: Geometry for Coulomb's Law.

Potential is defined as the work required to move a point charge from an infinite distance to the point P, and so a potential field may be related to **E**. The work required to move a test charge from infinity to a point P is given by

$$V_P = -\int_{\infty}^{P} \mathbf{E} \cdot d\mathbf{r}$$

The minus sign indicates that we must do work against the electric field to approach a like charge. Electric potential is measured in NmC^{-1}, or volts. Note that force is a **vector** (i.e. it has a magnitude and a direction) but that potential is a **scalar** (i.e. it has magnitude only). Both are functions of three dimensional space. Dealing with potential is often easier than dealing with the field directly. The electric field is obtainable from the potential by differentiation:

$$\mathbf{E} = -\nabla V$$

so it is easy to see that the units of **E** are V/m.

The gradient operator, ∇ , equals

$$\mathbf{x}\frac{\partial}{\partial x} + \mathbf{y}\frac{\partial}{\partial y} + \mathbf{z}\frac{\partial}{\partial z}$$

where **x, y, z** are the cartesian unit vectors. Applied to a scalar, ∇ becomes gradient and results in a vector:

$$\nabla U = \left(\frac{\partial U}{\partial x}, \frac{\partial U}{\partial y}, \frac{\partial U}{\partial z} \right)$$

Dotted into a vector, ∇ becomes divergence and is a scalar:

$$\nabla \cdot \mathbf{V} = \frac{\partial \mathbf{V}_x}{\partial_x} + \frac{\partial \mathbf{V}_y}{\partial_y} + \frac{\partial \mathbf{V}_z}{\partial_z} \quad .$$

Crossed into a vector, ∇ becomes the curl or rotation and is another vector:

$$\nabla \times \mathbf{V} = \begin{vmatrix} \mathbf{x} & \mathbf{y} & \mathbf{z} \\ \frac{\partial}{\partial x} & \frac{\partial}{\partial y} & \frac{\partial}{\partial z} \\ V_x & V_y & V_z \end{vmatrix} = \mathbf{x}\left(\frac{\partial V_z}{\partial y} - \frac{\partial V_y}{\partial z}\right) + \mathbf{y}\left(\frac{\partial V_x}{\partial z} - \frac{\partial V_z}{\partial x}\right) + \mathbf{z}\left(\frac{\partial V_y}{\partial x} - \frac{\partial V_x}{\partial y}\right)$$

where the second term is the determinant of that funny matrix.

The potential is **conservative**, that is, it doesn't matter what path the charge takes to go from point to point, the total work done is the same. Potential fields are also additive, or linear, so the effect of several charges may be obtained by integrating the effect of all the constituent charges.

When we considered gravity and magnetics we considered distributed regions of mass or magnetization, but this is not our interest here because we will be concerned with electric current, which is the flow of charge. Current in a wire is measured in amperes, 1 A being 1 C/s, but in a three dimensional medium, charge flow has to be measured by current density, **J**, having units of $\mathrm{Cm}^{-2}\mathrm{s}^{-1}$ or Am^{-2}. Note that **J** has a direction as well as magnitude.

Charged particles in a material are in constant thermal motion, colliding with each other and moving in all directions, resulting in no net current flow. However, in an electric field the charges will have a slight preferred direction, resulting in a drift velocity. The charges will usually come to rest in a very short time after the field has been removed, because the time between collisions is very short, on the order of 10^{-14} s for free electrons in a metal (see below). The drift velocity is proportional to **E**, so current density is also proportional to **E**;

$$\mathbf{J} = \sigma \mathbf{E}$$

This expression is Ohm's Law. The constant of proportionality, σ, is called conductivity, and depends on the number of charge carriers per unit volume, their charge magnitude, and their mobility. Conductivity has the unit S/m (a siemen is a reciprocal ohm) and is a measure of how easily an electric current is generated by an electric field. It is common, especially in geophysics, to consider resistivity, ρ, the reciprocal of conductivity. We see that resistivity has units of Ωm.

$$\rho = \frac{1}{\sigma}$$

If we re-arrange Ohm's Law to use resistivity, you see that it is similar to the circuit form of Ohm's Law that you probably have already seen, with ρ mapping to resistivity, V mapping to electric field, and I mapping to current density:

$$V = IR \rightarrow \mathbf{E} = \mathbf{J}\rho$$

In effect, $V = IR$ is an integral form while the $\mathbf{J} = \sigma\mathbf{E}$ is the differential form.

2. Resistivity of rocks and minerals.

We have seen that electrical conduction in a material is accomplished by a movement of charge, which in practice means the movement of electrons, holes (the absence of an electron), ions or ion vacancies. Materials may be classified on the basis of the nature of the charge carriers and the type of movement.

Metallic conduction is due to 'free' electrons, that is electrons which do not take part in crystal bonding, in a metal. This form of conduction is very efficient, resulting in room temperature resistivities for metals of about 10^{-7} Ωm. The resistivity of metallic conductors decreases with temperature because the thermal agitation of the conduction electrons, which impedes their movement in response to an electric field and therefore lowers the drift velocity. In the extreme case of superconductors, at sufficiently low temperatures resistivity goes to zero. Massive deposits of native metals are extremely rare in rocks, and so metallic conduction is not of direct importance to prospecting. (Of course, the technology used to build prospecting apparatus depends very much on metallic conduction.) However, fine grains of metallic conductor dispersed through a porous rock play an important role in the induced polarization effect, described later.

Semiconduction is the result of behavior between metallic conduction and conduction in "insulators". There are no free electrons as such in a semiconductor, but an applied electric field can supply enough energy to move electrons into higher energy levels, where they are free to conduct. As a result of these few electrons, and the 'holes' they leave behind in the lower energy levels, moderate conductivities are achieved, between about 10^{-5} and 10^{5} Ωm. Although rock-forming minerals are not semiconductors at surface temperatures, semiconduction is important in geophysical prospecting because many of the sulphide minerals are semiconductors with quite low resistivities, making ore bodies detectable using electrical methods.

Solid electrolytic conduction is the result of moving ions and ion vacancies through a crystal structure. As may be imagined, it is very difficult to move ionically bonded ions through a crystal lattice, so the resistivities of silicate minerals are very high; quartz has resistivities between 10^{10} and 10^{14} Ωm. Halite has a resistivity of 10^{12} Ωm or so. It would, in fact, be close to impossible to move ions through a perfect lattice, but defects in the lattice allow the motion of ions and holes. Three types of defects are illustrated in the diagram below. Schottky defects are lattice vacancies and Frenkel defects are ions in interstitial positions. Impurities also constitute defects in the lattice.

Shottky and Frenkel defects are thermally induced, so the number of defects, and hence conductivity, increases with increasing temperature. Impurities do not depend on temperature, but rather account for the variation in resistivity between different samples of the same mineral. Because smaller ions such as Mg^{++} and $Fe^{++,+++}$ are more mobile than larger ions such as Na^{+} and K^{+}, mafic rocks are slightly more conductive than felsic rocks, but for most geophysics rock conductivity is dominated entirely by water content.

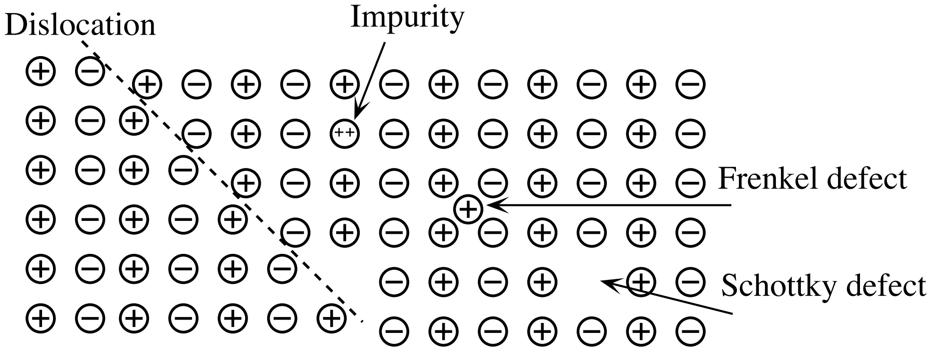


Fig. E.02: Three different types of defects found in ionic crystals.

Electrolytic conduction is the more familiar movement of dissociated ions through a liquid or solution. Melting of a solid electrolyte vastly increases the mobility of the ions, and so the resistivity drops by an order of magnitude or more. However, unless you are engaged in the search for geothermal resources, this will not be important, as most near-surface rocks are well below their melting points. Of much greater significance are the low resistivities of salt solutions. Ions in solution are very mobile so while the resistivity of pure water is very high, the addition of a soluble ionic salt lowers the resistivity greatly. One molar NaCl solution has a resistivity of 0.22 Ωm; the resistivity of sea water is between 0.25 (near surface) and 0.30 Ωm (below the thermocline) depending on the temperature; to a good approximation the conductivity is seawater is

$$\sigma(t) = 3 + t/10 \quad \mathrm{S/m}$$

where t is temperature in Celsius. Increasing the concentration of salt also increases the conductivity of salt solution. The resistivity of ground water rarely get higher than 10 to 100 Ωm but can be much less resistive if salty.

Because the resistivity of silicate minerals at surface temperatures is very high and the resistivity of ground water is, on the other hand, quite low, electrical conduction in most surface rocks is through water contained in pores, cracks and fissures. Archie's law gives an empirical relationship between water content and rock resistivity:

$$\sigma_r = \sigma_w \phi^{+a} \quad or \quad \rho_r = \rho_w \phi^{-a}$$

where ρ_w is the resistivity of the water (σ_w the corresponding conductivity, which increases with salinity and temperature) , ϕ is the fractional content of water in the rock (equal to the porosity if the rock is saturated). The exponent, a, is related to how well the pores are interconnected, and may take values between 1 and 3, but is usually about 2. We see that as water content goes up, rock conductivity goes up, and as water conductivity goes up, so does the rock conductivity. This simple form of Archie's law with an exponent of 2 can be a useful guide to porosity given resistivity, but if reliable results are required, the exponent (and another multiplicative constant called formation factor) must be established by making laboratory studies of a representative suite of rocks from the area being considered.

The resistivity of wet rocks is extremely variable, varying over many orders of magnitude as the porosity and groundwater conductivity changes. In general, igneous rocks are more resistive than sedimentary rocks, whose resistivity goes up with consolidation and which are generally more resistive than unconsolidated sediments. Clays are almost always relatively conductive, around 10 Ωm at least, because they have unpaired electric charges on the crystal surfaces which in turn create charges in the groundwater. One needs to beware of extrapolating from the porosity and resistivity of hand samples to the resistivity of the bulk rock. For porous sediments this extrapolation may be quite good, but for rocks of lower porosity most of the water may be contained in porosity of a larger scale than may be sampled with a specimen, that is cracks, joints, fissures, faults and so on.

The range in the resistivity of rocks is from about 1 Ωm for clays and porous, salt water saturated sandstones to about 10^6 Ωm for relatively unfractured igneous rocks.

Rock type	**porosity**	**resistivity**
Soil, unconsolidated rock	0.1–0.5	1–1000 Ωm
Consolidated sedimentary rock	0.01–0.4	10–10000 Ωm
Igneous rocks	0.001–0.1	100 – 10^6 Ωm
Metamorphic Rocks	0.001–0.1	100 – 10^6 Ωm
Metalliferous ores	–	0.1–100 Ωm

Table 2.01: Typical resistivities and porosities. More comprehensive tables can be found in Telford *et al.*; these are coarse guidelines only. Apart from ore bodies, the main factor is porosity and groundwater salinity, and so initial porosity, intergranular mineral growth, and weathering are all contributing and cumulative factors.

3. Measuring rock resistivity in the laboratory.

For a simple series circuit, such as a piece of wire, Ohm's law becomes

$$V = IR,$$

where V is voltage (actually a potential difference ΔV), I is electric current and R is resistance. However, it is resistivity, ρ, rather than resistance, that is the intrinsic property of a material. If we consider a regular solid, resistivity and resistance are related by

$$R = \rho L/A$$

where L is the length of the prism and A is the cross sectional area of the base.

This forms the basis of measuring rock resistivity: cut a regular solid from the rock, such as a core or a cube, apply current electrodes to the ends and see how much current may be passed for a

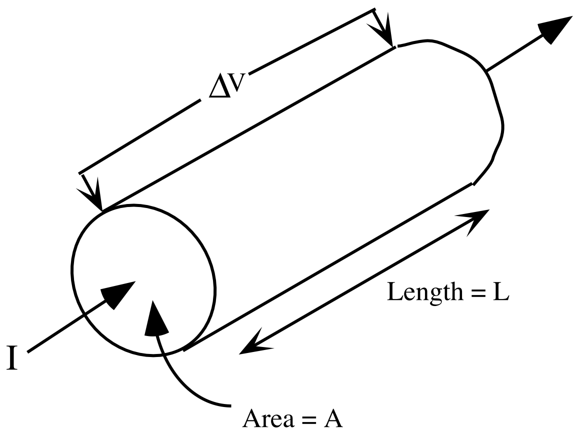


Fig. E.03: Measuring resistivity of a regular prism.

given voltage, that is,

$$\rho = \frac{A\Delta V}{IL}$$

This is called the two electrode method, and works well if the contact resistance of the current electrodes can be kept low. If the rock is resistive, however, it is quite likely that the contact resistance of the electrodes will be higher than the resistivity of the rock, giving erroneous results. The solution is to devise a method which ignores the contact resistance, and so the four electrode method was devised, whereby the potential difference is measured across the center of the rock rather than the ends. For a high impedance voltmeter the electrode resistance is not critical, and the resistivity can be computed from $\rho = \Delta V.A/(I.s)$ where s is the potential electrode spacing. In Figure E.04, R_r represents the resistance of the rock specimen and R_c represents an electrode contact resistance, either current or potential.

Although the D.C. resistance is usually required, the current source must be low frequency (1 Hz to 1 kHz) A.C. to avoid *polarization* (note that when we consider field experiments, this frequency will be lower because the length scale is larger). Polarization results from a build-up of charge carriers on the ends of the sample or within the sample itself, inhibiting further flow of current. This property is actually exploited as an exploration method (induced polarization) described later.

Anisotropy. It is possible that the resistivity of a mineral, rock, or rock formation depends on the direction of measurement. Thus the resistivity measured along the c crystal axis may be different to that measured along the a crystal axis. The coefficient of anisotropy, λ, is the ratio of maximum to minimum resitivity, thus

$$\lambda = \frac{\rho_{max}}{\rho_{min}}.$$

Mineral (micro) anisotropy is not likely to be of any importance in prospecting, but (macro) anisotropy due to bedding structure or metamorphic foliation is common, where the alignment of non-spherical grains or microcracks produces a preferred direction for current flow. This alignment is most often controlled by gravity, and therefore produces the situation in which $\rho_z > \rho_x, \rho_y$ (i.e. horizontal resistivity in either direction is less than vertical resistivity). This is horizontal anisotropy. Interbedding between otherwise isotropic sedimentary units of different resistivity will also produce electrical anisotropy on a much larger scale. The terms macroanisotropy and microanisotropy are often used to indicate the scale on which the anisotropy is caused.

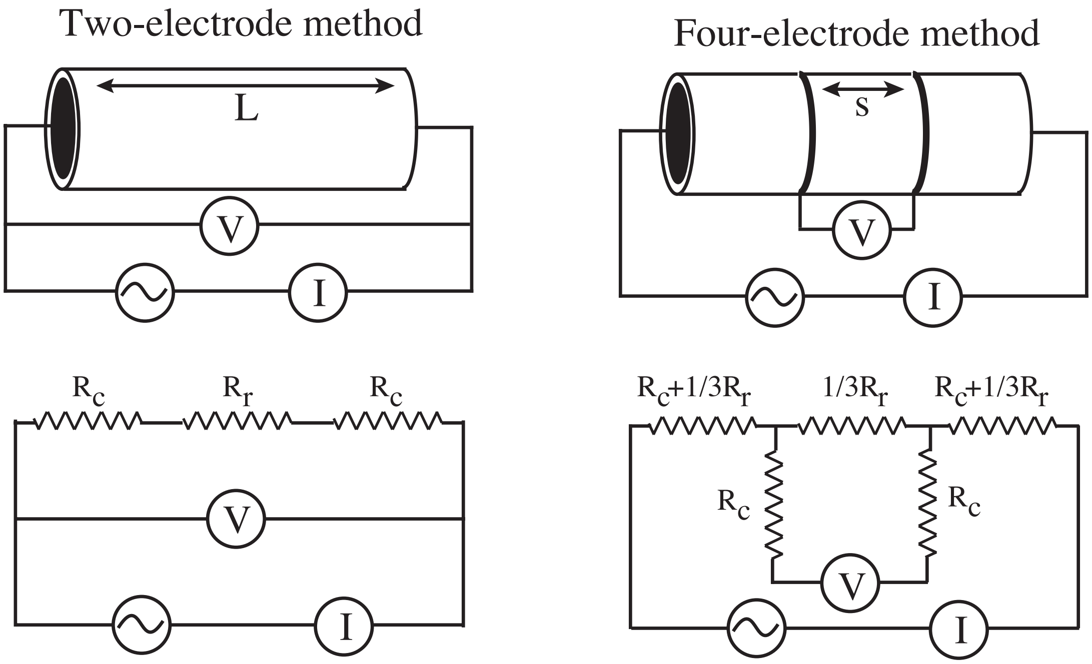


Fig. E.04: Two- and four-electrode methods of measuring rock resistivity, showing equivalent circuit elements. In the two-electrode method, the potential drop across the current electrodes is measured along with the potential drop across the sample. In the four-electrode method, the potential drop across the current electrodes is not part of the voltage measuring circuit, and the potential electrode impedance simply adds to the already high input impedance of the voltmeter.

4. DC resistivity method

The measurement of the Earth's resistivity is very similar in concept to the laboratory resistivity measurement of rock samples. A DC electric current is passed through the ground *via* a pair of *current* electrodes and a resulting potential difference is measured between a second pair of *potential* electrodes. It is common to designate the current electrodes A and B and the potential electrodes M and N. Although the two-electrode method is occasionally practiced in the laboratory, it is never used in the field because the resistance of the current circuit is always dominated by the electrode contact and the material immediately surrounding the electrodes.

Theory. As in all potential methods, we consider a point source first and later combine point sources to model more realistic situations.

Consider a single current electrode, considered to be a point source of current, on the surface of a homogeneous, isotropic halfspace, injecting a current I into the Earth (Figure E.05). The flow of electric current will be radially symmetric in the halfspace. This should be intuitively obvious, but is also a consequence of the two boundary conditions which must be satisfied at a contact between materials of differing conductivity, namely that the tangential electric field must be continuous and the normal electric current density must be continuous. Since there is no current flowing in the atmosphere, there must be no normal component of **J** at the Earth's surface, i.e. **J** is tangential at the surface.

To compute the E-field inside the earth we will balance the current flowing into the earth at the

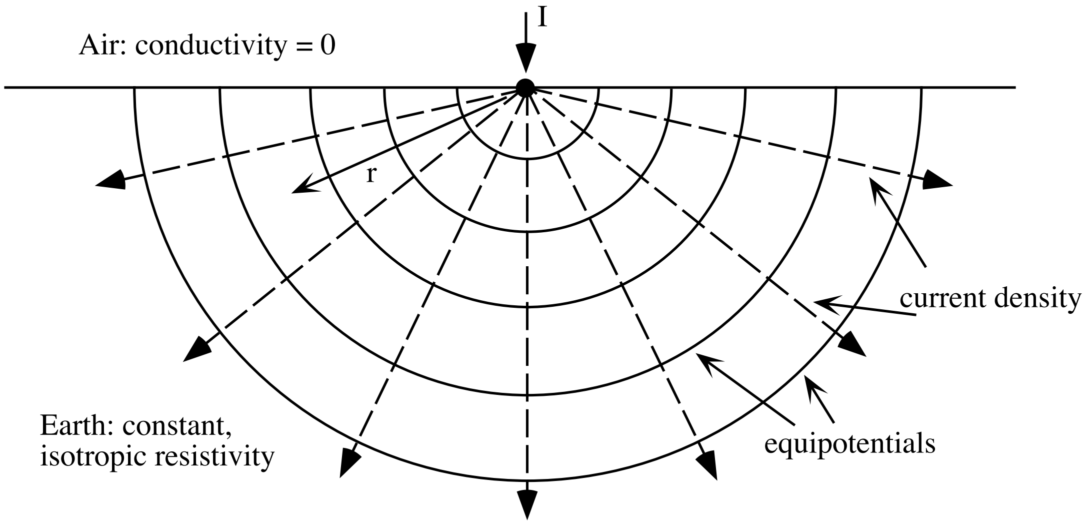
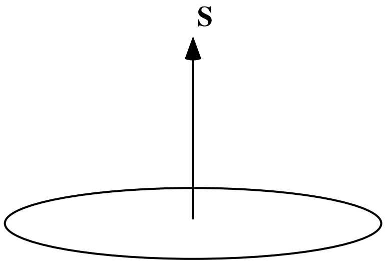


Fig. E.05: Current I being injected into homogeneous ground from a single electrode. Current density is directed radially out from the electrode and equipotentials are hemispherical.



(A planar surface may be described by a vector, **S**, whose magnitude is the area of the surface and whose direction is the (outward) normal to the surface. Thus in the above case $d\mathbf{s}$ describes the surface element.

electrode with the total current flow out of a hemispherical surface. The total current across the hemisphere must be equal to I because there are no sources or sinks of current other than I (i.e. charge is conserved). Because of the radial symmetry, the current density will be constant at a distance r from the current electrode, so the total current flow across the hemispherical surface will be

$$I = \int_{hemisphere} \mathbf{J} \cdot d\mathbf{s} = 2\pi r^2 J$$

($d\mathbf{s}$ is defined in the box.) The integration is simply the surface area of the sphere times the (constant) current density which is always normal to the hemisphere. So, we have that at any distance r, the current density is

$$\mathbf{J} = \frac{I}{2\pi r^2}\mathbf{r}$$

where **r** is the outward normal to the hemisphere defined by r. Now substituting the above expression into Ohm's law gives

$$\frac{I}{2\pi r^2}\mathbf{r} = \sigma\mathbf{E}$$

or using the definition of resistivity and rearranging

$$\frac{\rho I}{2\pi r^2}\mathbf{r} = \mathbf{E} \quad .$$

To obtain the potential at a distance R we integrate the electric field from infinity to R:

$$V_R = -\int_{\infty}^{R} \mathbf{E}.d\mathbf{r} = -\int_{\infty}^{R} \frac{\rho I}{2\pi r^2}dr = \left.\frac{\rho I}{2\pi r}\right|_{\infty}^{R}$$

$$= \frac{\rho I}{2\pi R} \quad .$$

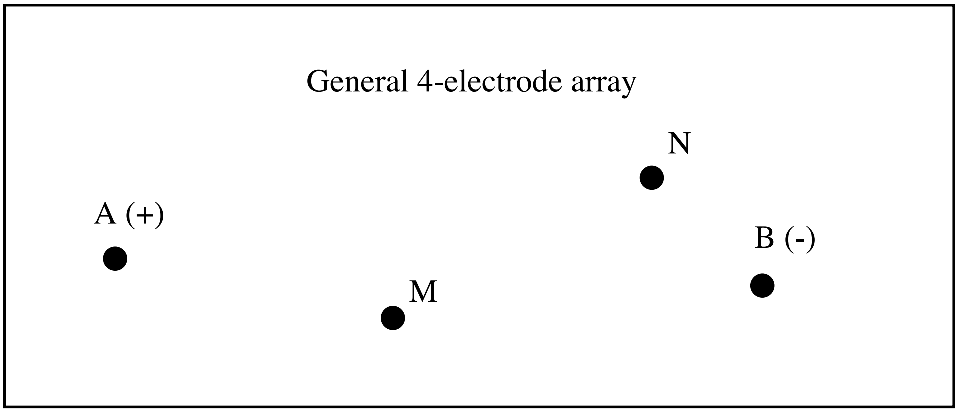


Fig. E.06: The general 4-electrode array. M and N are potential electrodes, A and B are current injection electrodes.

Now for the general four-electrode array: we have that the potential at electrode M is simply the sum of the effects of the two current electrodes:

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

and similarly the potential at N is

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

so the potential difference measured across MN is

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

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

where k is called the geometric factor (we will see why $1/k$ is used soon):

$$k = 1/\left(\frac{1}{AM} - \frac{1}{BM} - \frac{1}{AN} + \frac{1}{BN}\right)$$

Apparent resistivity: What we have derived above is often called a *forward* problem in geophysics; that is, given a possible structure for the Earth (in this case an isotropic, homogeneous halfspace of resistivity ρ), we can compute what measurement would be made over that Earth (in this case ΔV for a given combination of I and electrode geometry). The forward problem tells us how to go from an Earth structure to a theoretical measurement. However, in real life we will have the measurement and want to know what the structure of the Earth is (after all, that is the point of making the measurement in the first place). The mechanism for going from a measurement to a theoretical structure is called the inverse problem. In the above case the inverse problem is very easy and unique:

$$\rho = 2\pi \frac{\Delta V}{I} k \ \ .$$

That is, given a measurement of ΔV the above expression would correctly yield the resistivity of a homogeneous, isotropic halfspace. *If the Earth were not a homogeneous, isotropic halfspace the above expression would not yield the resistivity of the Earth.* However, to aid interpretation, an *apparent* resistivity is defined for any measurement over any Earth structure as

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

so although this resistivity is not likely to represent the truth, it is a sort of 'average' resistivity. The apparent resistivity is the first step in the analysis of electrical sounding data; ρ_a is computed for every measurement and is then further interpreted for more complicated structures than halfspaces.

Sounding and Profiling. There are two possible objects to using electrical methods. Electrical sounding is an attempt to determine the resistivity of the Earth as a function of depth. Sounding makes the assumption that the structure is locally one dimensional (1D), that is resistivity varies only with depth and does not change significantly in lateral directions on the scale of the sounding array (i.e. 'layer cake geology'). It is accomplished by making the electrode array successively larger between measurements, making the depth of penetration successively deeper. Profiling, on the other hand, has the explicit object of determining lateral changes in resistivity and hence structure. It is accomplished by moving a fixed size array across the ground.

It is clear that ρ_a can easily be computed for any arrangement of electrodes, but the further interpretation is complicated and depends heavily on previous workers' efforts at solving difficult mathematical problems, so in practice only a few specific geometries are used:

The *Schlumberger array* is the most commonly used arrangement, and was developed by Conrad Schlumberger in the 1930's. Like most of the standard arrays it is colinear and symmetrical. The particular feature of the Schlumberger array is that the potential electrode spacing is very much smaller than the current electrode spacing.

To compute the expression for apparent resistivity for this array we firstly note that for a symmetrical array $AM = BN$ and $BM = AN$ so that

$$\frac{1}{k} = (\frac{2}{AM} - \frac{2}{BM})$$

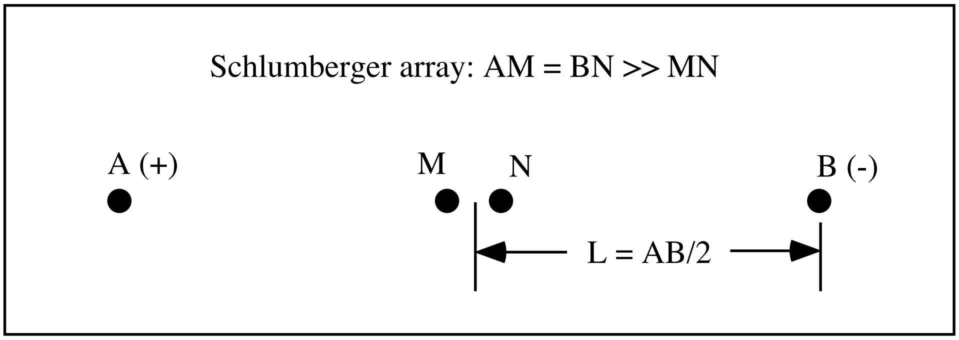


Fig. E.07: The Schlumberger array. MN is small compared with AB and the array is symmetrical and colinear.

Now for the Schlumberger array we write

$$BN = AM = AB/2 - MN/2 = (AB - MN)/2$$

and

$$AN = BM = (AB + MN)/2$$

so

$$\frac{1}{k} = \frac{4}{AB - MN} - \frac{4}{AB + MN}$$
$$= 4\frac{(AB + MN) - (AB - MN)}{(AB - MN)(AB + MN)} = \frac{8MN}{(AB - MN)(AB + MN)}$$

or

$$k = \frac{(AB - MN)(AB + MN)}{4MN}$$

yielding an expression for apparent resistivity:

$$\rho_a = \frac{\pi \Delta V}{4.I.MN}(AB - MN)(AB + MN)$$

Because $AB >> MN$ we may set $AB \pm MN \approx AB$ and also write $E \approx \Delta V/MN$ so

$$\rho_a \approx \frac{\pi E}{4.I}(AB)^2 = \frac{\pi E}{I}\left(\frac{AB}{2}\right)^2$$

The *Wenner array* was developed by Frank Wenner in the U.S. at about the same time that Schlumberger developed his techniques in France. Wenner worked in the national standards laboratory on material properties, and realized that he could apply the same four-point method used in the laboratory in the field to measure bulk Earth resistivity. Consequently, the array that bears his name has equal spacings between all electrodes.

From our expression for a symmetric array we have very simply that $1/k = 2/a - 2/2a = 1/a$ so that $k = a$ and

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

Finally we have the *dipole-dipole* array, which comes in two varieties, equatorial and polar. We have for the polar array

$$\rho_a = \frac{\pi r^3}{I.AB} \cdot \frac{\Delta V}{MN}$$

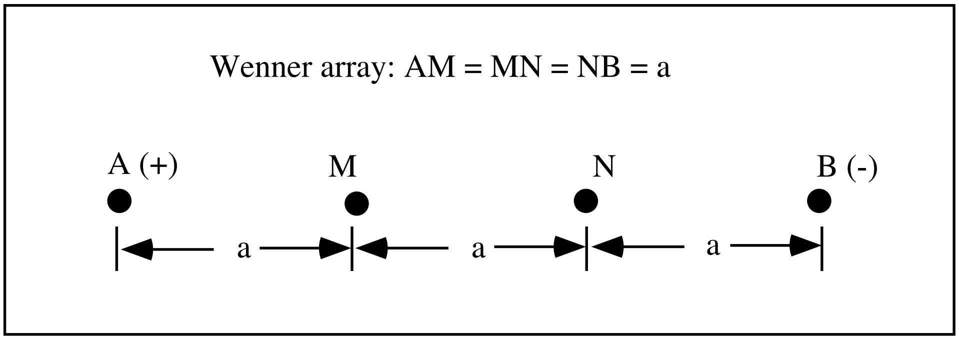


Fig. E.08: The Wenner array. The array is symmetrical and colinear but now the electrodes are equally spaced.

and for the equatorial array

$$\rho_a = 2\frac{\pi r^3}{I.AB} \cdot \frac{\Delta V}{MN}$$

Observe that as in magnetism, the dipole field falls of as r^3 and the polar field is twice as strong as the equatorial field. $\Delta V/MN$ is approximately E, the electric field, and I.AB is the dipole moment of the transmitter, and is a meausure of power supplied to the Earth.

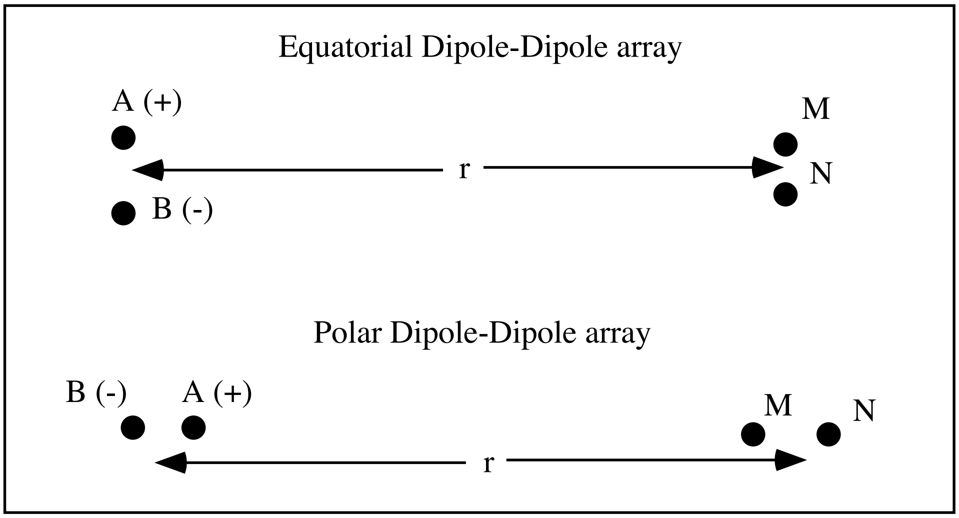


Fig. E.09: The dipole-dipole arrays. Both the current electrode pair and potential electrode pair form dipoles which are separated by a distance which is large compared to the dipole length.

5. Electrical sounding.

In an electrical sounding the electrode array is systematically made larger while the center of the array remains fixed over the area of interest. As the array gets larger, the electric currents flowing deeper and deeper in the Earth are sensed, and so the resistivity of deeper and deeper structure is measured.

An apparent resistivity is computed for each electrode spacing and plotted as a function of the electrode spacing, both the resistivity and spacing scales being logarithmic. This plot is the *sounding*

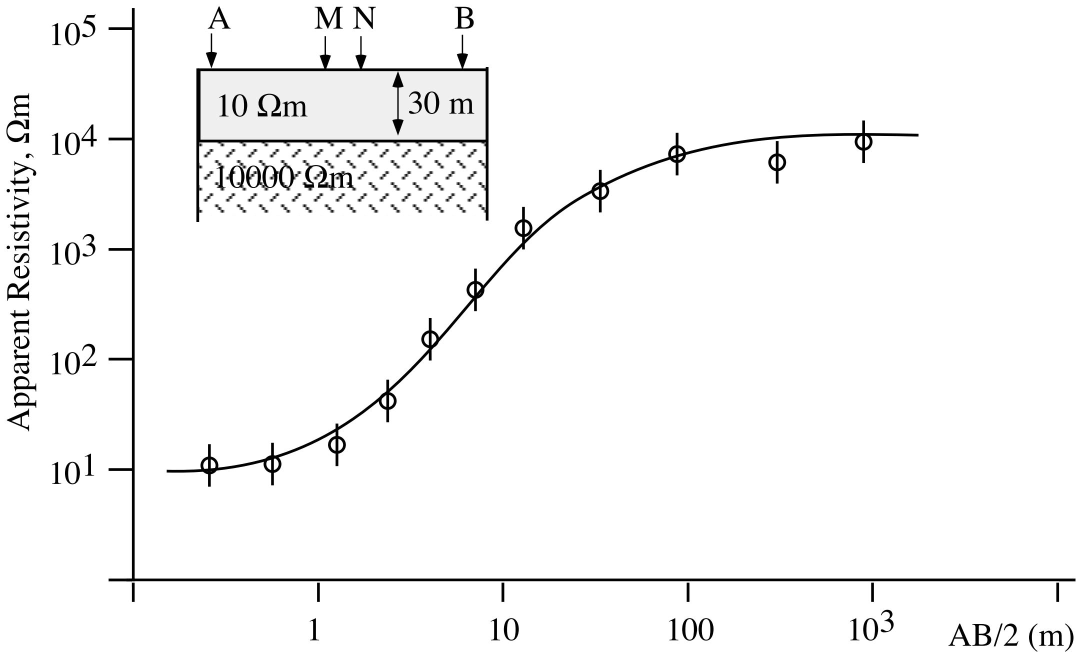


Fig. E.10: Electrical sounding. The electrode spacing is made progressively large, sensing progressively deeper structure. Apparent resistivity is plotted against electrode spacing to give a sounding curve.

curve. Logarithmic scales are used because a) the range in resistivity of Earth materials is more than 5 orders of magnitude and b) the resistivity method is only sensitive to structure which is of comparable size to its depth of burial.

Roughly speaking, the apparent resistivity curve represents the true resistivity as a function of depth, smoothed by the response of the experiment. Thus, if the Earth were composed of a 10 Ωm layer 30 m thick overlying a 1000 Ωm layer, when the electrode spacing is much less than 30 m the Earth looks like a 10 Ωm halfspace to the experiment and the apparent resistivity would truely represent the resistivity of the top layer. When the electrode spacing is much greater than 30 m the apparent resistivity is the resistivity of the lower layer. At electrode spacings of about 30 m both layers influence the apparent resistivity and the sounding curve makes a smooth transition between the two layer resistivities.

For the Schlumberger array, ρ_a is plotted as a function of $AB/2$, for the Wenner array, a, and the dipole-dipole array, r.

6. Electrical profiling

Electrical profiling consists of moving an array of fixed geometry across the surface of the Earth in an attempt to resolve lateral changes in resistivity. As in sounding, the depth of penetration of the electric current is comparable to the electrode spacing, but now the electrode spacing is chosen to be sensitive to the sort of structure being investigated. Thus, if the resistivity of an aquifer at a depth of 20 m were to be mapped, an AB/2 or a of 30 - 50 m would be used. For profiling the apparent resistivity is plotted as a function of the lateral position of the center of the array. By its nature, resistivity profiling is less quantitative than sounding.

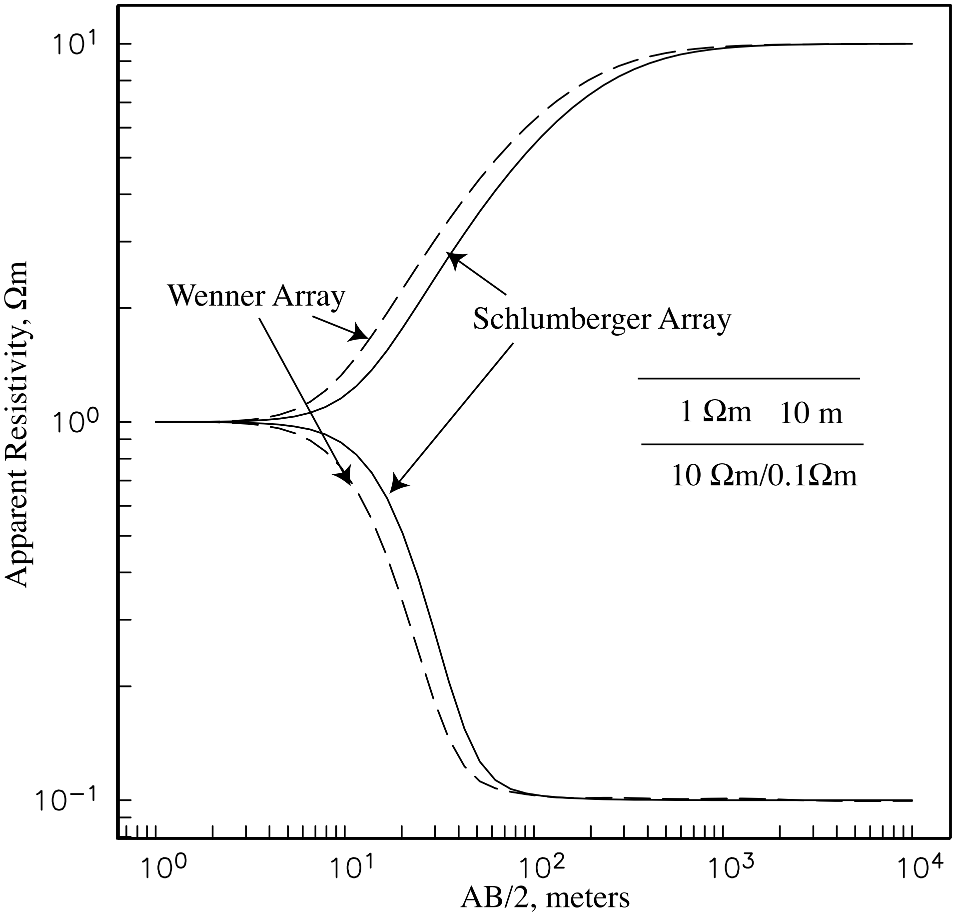


Fig. E.10b: Electrical sounding. Sounding curves of the Schlumberger and Wenner arrays are different over the same structure.

A combination of sounding and profiling can be conducted to produce a *pseudo-section*. A dipole-dipole array is moved laterally, but the source-reciever spacing is also changed. The apparent resistivities are plotted on a cross section at a depth of half the electrode spacing and beneath the center of the array. A sort of two-dimensional picture is built up which qualitatively indicates resistivity structure as a function of both depth and lateral position. Great care must be exercised when interpreting data presented as pseudo-sections, because there is no guarantee that the structure of the real Earth will be reflected when the data are plotted this way.

Electrode effects in electrical sounding. The object of electrical sounding is to estimate resistivity as a function of depth. For the results to be realistic the true structure must be, for the most part, only a function of depth, or one-dimensional (1D). We know, of course, that this will never be perfectly true, but in many circumstances the 1D approximation is good enough, for example when mapping sedimentary strata with little deformation and relatively small dips, or mapping the depth to basement, where again the structure of the sediment/basement interface is mostly horizontal. Remember that the 1D approximation only has to hold for scales on the order of the largest electrode spacings. Thus variations in the depth, say, to a horizon may be mapped by multiple soundings spaced well apart.

While electrical sounding is quite tolerant to deviations from the 1D approximation that occur at depth (it is possible to map beds with slopes of up to 15^o), variations in resistivity near the surface,

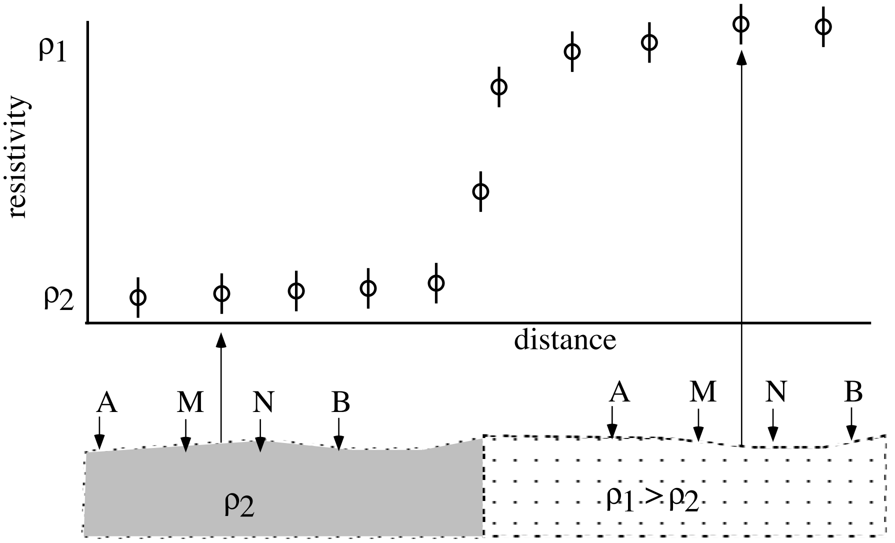


Fig. E.11: Electrical profiling. An resistivity array is moved along the surface while holding the dimensions fixed. The apparent resistivity is plotted as a function of distance to give a resistivity profile or map.

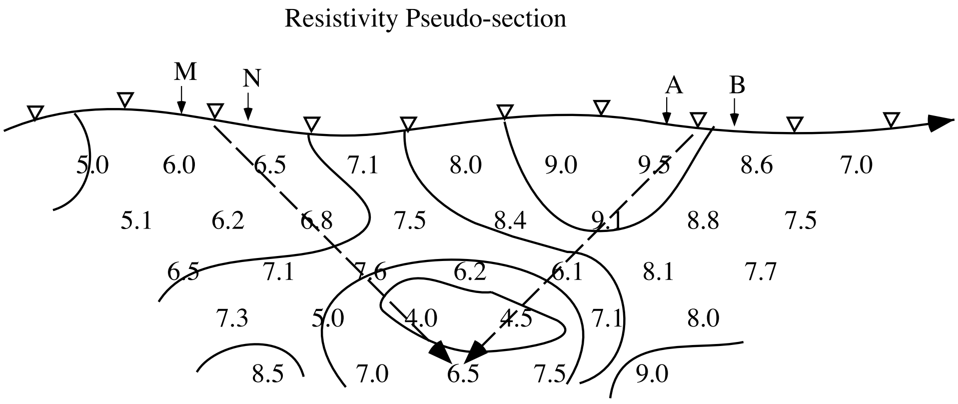


Fig. E.11A: Electrical pseudo-section. The equally spaced stations across the surface of the earth are all occupied by dipole transmitter and receiver arrays. For each transmitter/receiver pair, the apparent resistivity is plotted mid-way between the stations and at a depth proportional to the station spacing. The apparent resistivities can be contoured and interpreted graphically, or modeled quantitatively.

where, of course, the electrodes are, are capable of producing large effects in the sounding curve.

When an electrode is moved and placed over a region of near-surface heterogeneity, a distorted measurement will be made. The distortion depends on whether a current electrode or potential electrode is placed on an inhomogeneity. For obvious reasons distortions are called current (or AB) electrode effects and potential (or MN) electrode effects. Current electrode effects are not as severe as potential electrode effects, because small inhomogeneities only serve to alter the effective shape of the current electrode, and as long as the distorting structure is small compared with the electrode spacing the point source approximation will hold. On the other hand, a distortion of the electric field near the potential electrodes will have a large effect, no matter how far apart the current electrodes

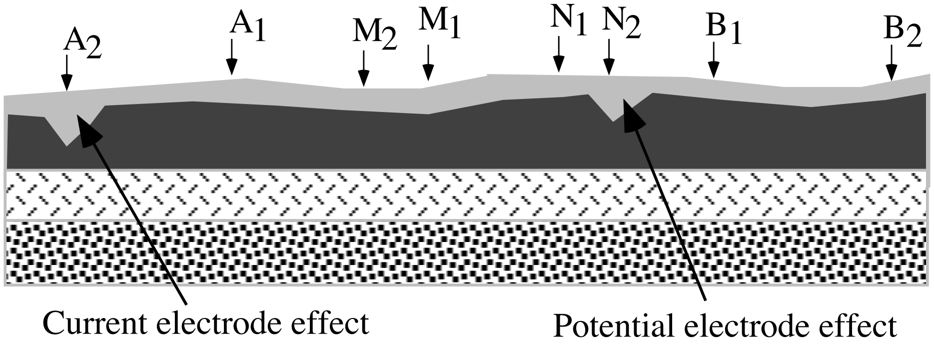


Fig. E.12: Electrode effects during a Wenner sounding. While the deeper structure is 1D, variations in the surface layer (say, in the thickness of the weathered layer) cause effects when both the current and potential electrodes are moved. The effect on the data is smaller for the current electrodes, because they are spaced further apart and the distortion is a small part of the total current path.

are.

How may electrode effects be recognized? A sounding curve for a truly 1D earth must be smooth; just as in gravity and magnetics, variations due to distant structure are broad on nature. Thus scatter between adjacent points on a sounding curve, is not due to sloppy measurement procedure, is due to electrode effects. This type of 'noise' is easy to handle because it is random in nature and we can draw a smooth curve through the scattered data for interpretation (which is not, in fact, the best way to handle such noise but gives an idea of its manageability nevertheless). For a Wenner or dipole-dipole sounding, where the potential electrodes are moved at every measurement, scatter due to MN electrode effects is common and there is little that may be done to eliminate it. For the Schlumberger array, however, the MN electrodes do not have to be moved, and so MN electrode effects are not a problem. Well, nearly not a problem. The trouble is, one may start a sounding with a MN spacing of about 0.5 m and an AB spacing of a few meters, but by the time the AB spacing is up to a few hundred meters the signal on the 0.5 m dipole is very small. Now, the potential gradient in the central 1/3 of an array is very nearly linear (see diagram on p. 637 [525] of Telford et al.), so all that needs to be done to increase the signal strength is to increase the potential electrode spacing, because even with MN equal to a few tens of meters it will still be small compared to the larger AB, as required by the Schlumberger array. But once the potential electrodes are moved there can be distortion due to MN effects. The solution, always carried out for Schlumberger sounding, is to make repeat measurements at 2-4 current electrode spacings every time the MN dipole needs to be enlarged. The result will then be a sounding curve which looks like that in figure E.13.

Sections of sounding curve where the MN electrodes did not move are free of MN electrode effects, but between these sections any distortion due to moving the MN electrodes is clearly visible. The procedure now is to simply slide the sections of the curves vertically so that they all line up to produce one sounding curve. The section of curve with the largest MN spacing is usually taken to be the most reliable, since for the larger spacing any small variations in surface resistivity have proportionately least effect. Also, what is the mathematical effect of sliding the data vertically on a log-log plot? The ability to remove MN electrode effects in this way accounts for the popularity of the Schlumberger array.

More severe trouble arises when the Earth departs from the 1D approximation in a more sys-

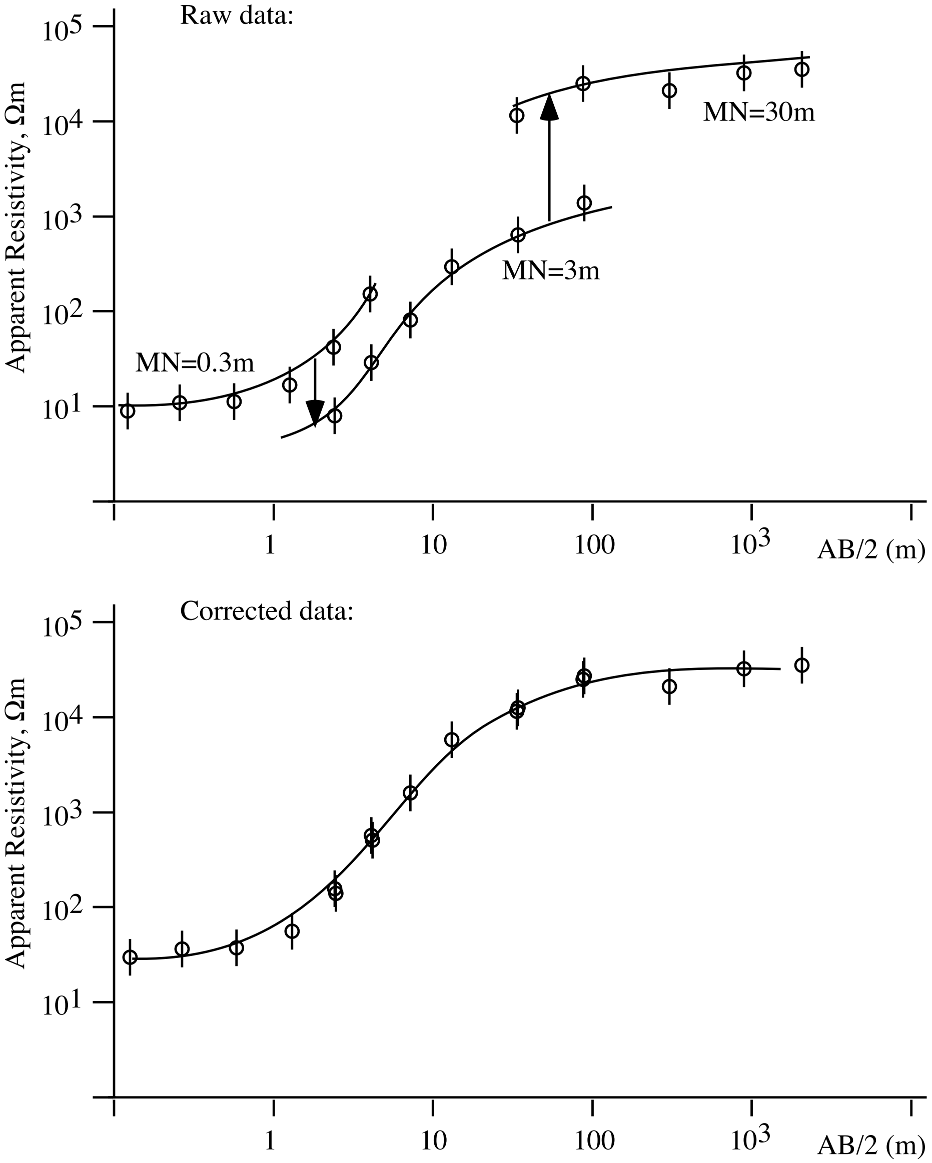


Fig. E.13: Correction for electrode effects in a Schlumberger sounding; three different MN spacings have been used (0.3, 3 and 30 m), with an overlap between each (i.e. both MN spacings measured with the same AB). The offsets between the different segments of the curves are removed by sliding the curves vertically. Usually, the largest MN is taken to be the most accurate.

tematic, or large-scale fashion, because then the source of the distortion may affect the current electrodes at several adjacent spacings. The danger then is that the electrode effect, which has now graduated into what may be better called a 3D effect, will be incorrectly interpreted as 1D structure. One danger sign is that for a truly 1D earth, an *ascending* part of the sounding curve may not exceed 45° in slope (on our log-log plots). To see why this must be so consider the most extreme transition; a perfectly conductive layer underlain by a perfect resistor. The current will be trapped in the conductive layer, and so instead of the electric field falling of as the inverse square of distance, as it does in a 3D medium, we essentially have a 2D medium and the electric field falls off as the inverse of linear distance, that is $\Delta V \propto 1/AB$. When this is substituted into our formula for apparent resistivity for the Schlumberger array, for example, which says $\rho_a \propto \Delta V.AB^2$, we get that $\rho_a \propto AB$. That is, double AB and ρ doubles too, and so there is the origin of the 45° slope. (What would the slope be on a linear-linear plot?) Although not very common, an ascending slope of $> 45^\circ$ on a sounding curve is a sure indication of 3D structure. A descending part of the curve may be very steep, however.

A very good practice is to test for 1D structure during the actual sounding. This may be done in one of two ways. A second electrical sounding may be completed at 90° to the first, and of course the two soundings should give the same result for a 1D earth. However, apart from requiring a lot of work, the 1D approximation holds better when a sounding is carried out with the electrode array along the strike of any 2D structure than across it. Thus the crossed sounding, if it crosses the structural strike also, may show the effects of 2D structure while the first sounding is interpretable using the 1D approximation. The second method to test for inhomogeneity is to move only one current electrode at a time as the array expands. Thus, say that the current electrodes are at an AB/2 spacing of 100 m, and that the next measurement is to be made at a spacing of 150 m. Instead of moving both current electrodes to 150 m as would normally be done, move only one to 150 m and make a measurement using this asymmetrical array. Then move the second electrode and make a normal measurement. The apparent resistivity of the 100/150 m measurement should fall between ρ_a for the 100/100 m and 150/150 m spacings; in fact it is possible to combine the 100/100 m and 150/150 m measurements mathematically to predict the 100/150 measurement. Any structure which is affecting only one end of the sounding array (i.e., non 1D structure) will show up as an inconsistency in the measurements (see Zohdy, 1978, Geophysics 43, 748-766).

7. Equipment.

The equipment for the electrical sounding or profiling may in principle be very simple; a set of dry cell or lead-acid batteries provides a current source, insulated wire connects the electrodes, which may be merely metal pegs driven into the ground, an ammeter measures the current and a voltmeter measures the potential differences. Very useful work may be done with such equipment, but there are refinements which improve the precision and accuracy of the measurements.

Current circuit. The bulk and cost of the current source for the AB circuit will be measured by the power of the transmitting unit ($V.I$), yet the parameter that determines the size of the signal across the potential electrodes is only current (I). Thus we have an easy optimization problem: maximize I for a given power ($V.I$). If the resistance of the AB circuit is R, then the power will be simply $P = I^2R$ so $I = (P/R)^{1/2}$. We see that the resistance of the current circuit must be minimized. Unless the cable connecting the AB electrodes is very long, most of the AB resistance will actually be at the grounding electrodes. Because of the inverse square law we have already examined, the grounding resistance will be determined by the resistivity of the material in immediate contact with the electrode and with what amounts to the surface area of the electrode. AB electrodes may be (a) metal (aluminum, steel, brass) pegs driven into the earth, (b) metal foil buried in a shallow pit, or (c) an extensive antenna of buried copper braid. The latter is only used for a relatively fixed installation such as a very large dipole-dipole spread. The choice between (a) and (b) depends on several factors. The foil is potentially a better ground, but digging pits is harder work, and a peg may be more effective if the nearsurface material is dry and resistive because a 1 m peg is more likely to reach down to a zone of capillary water. $10 - 100\ \Omega$ represents the typical range of grounding resistance. It is perfectly normal to water the electrode with a salt solution to lower the resistance of the immediate area, but then the grounding resistance is likely to change with time as the water seeps into the soil. The one restriction to be observed with current electrodes is that they do not violate the point source approximation used in the computation of apparent resistivity. That is, the size of the electrode should be small compared with the current electrode separation AB. Thus, a 1 m long stake would not be driven fully into the ground until AB was about 10 m or more.

Because the transmitter circuit is likely to have a resistance of 100 Ω or more, a fairly high transmitter voltage is required to drive a significant current through the ground. 100 V would be typical and up to 1000 V possible. Voltage higher than this would too hazardous to use and produce problems with leakage. The power source may be dry cell or lead-acid batteries or a portable generator connected to a DC power supply. Although the DC behavior of the ground is required, in practice the current is low frequency AC or switched to reverse polarity every few seconds. This prevents polarization of the current electrodes and removes any DC offset between the potential electrodes (see below).

Potential circuit. Stainless steel pegs are commonly used as potential electrodes for simple electrical soundings, but they are quite noisy. This is because metal in contact with damp ground corrodes, and corrosion is an oxidation reaction which has associated electrical potentials which are fairly large. The potential difference across MN may be quite small, and mostly this corrosion noise makes accurate measurement difficult. The solution is to use *non-polarizable* electrodes; a solution of metal salt held in a porous container to which contact is made *via* a metal rod:

Most commonly copper-copper sulphate is used, but lead-lead chloride are also employed. The ‘porous pot’ electrode is emplaced into the soil so that the porous ceramic is in good electrical contact. Watering potential electrodes to lower grounding resistance is a risky business because excessive drift (see below) may develop.

It is important that the device used to measure the potential difference (PD) across MN does not draw too much current (if it does the potential will be less than the correct value). That is, its input impedance must be much bigger than the resistance across MN. In bygone days this was ensured by using a potentiometer, but today voltmeters, with input resistance of 10 MΩ or greater are readily available.

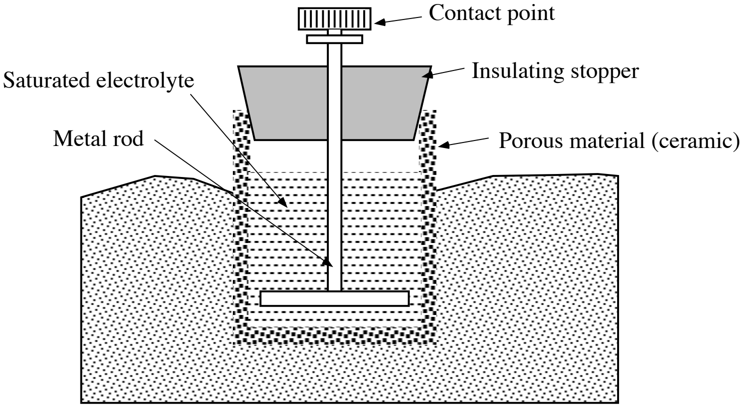


Fig. E.14: Non-polarizable electrode, or ’porous pot’. Electrical contact between the metallic parts of the AB circuit and electrolytes in the ground is made *via* a metal-metal salt system, for example copper and copper sulphate. The compatible metal in the salt, and a saturated salt solution, ensure that electrochemical reactions on the metal surface are kept to a minimum. Small amounts of salt solution leaking through the ceramic case provides an electrical contact with the ground.

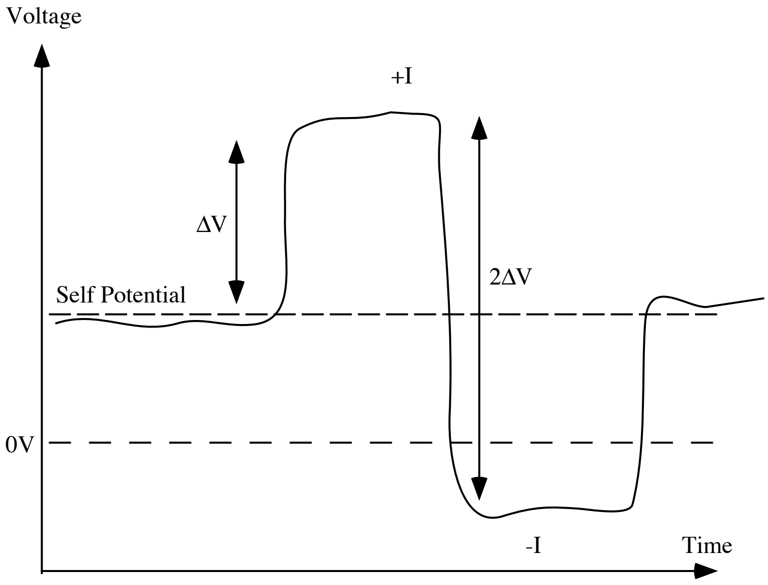


Fig. E.15: There will be some voltage on the MN circuit even with no current flowing through AB ('a self potential'). To remove this, and double the signal size, the current is reversed in the AB circuit, to measure $(-\Delta V)$ to $(+\Delta V)$ rather than 0 to $(+\Delta V)$.

It will be found that a potential difference across MN will be several mV to several V without any current in AB. This is because of self potentials (see next section) in the ground, low frequency currents induced by variations in the Earth's magnetic field, and DC offsets between electrode pairs. Using an AC or reversed AB current has the effect of averaging away this DC potential.

Figure E.15 shows a switched or square wave current, but AC has much the same effect. The $+\Delta V$ and $-\Delta V$ values may be measured manually and subtracted to obtain $2\Delta V$, but many modern instruments will do this manipulation automatically and moreover average ('stack') several cycles transmitted waveform to improve signal to noise ratio.

8. Interpretation

There are three approaches to the quantitative interpretation of both profiling and sounding resistivity results; matching pre-computed curves to the data (curve matching), comparing the computed response of a guessed model to the field data and then adjusting the model to obtain a fit to the data (interactive forward modeling), and mathematical/numerical inversion schemes to automatically obtain a subsurface resistivity structure from the data. With profiling data it is often not necessary to make a quantitative interpretation; the detection of an anomalous region or plotting a psuedo-section is usually sufficient. With sounding data, however, it is always necessary to make some kind of quantitative estimation. The theory allowing the computation of synthetic apparent resistivity curves for a given model consisting of discrete layers in the earth is complicated but has been developed for a long time. However, the equations relating ρ_a to a layered structure (the forward problem) are complicated and it is not easy to produce an expression relating a layer structure to ρ_a (the inverse problem). Thus many workers rely on curve matching and iterative forward modeling.

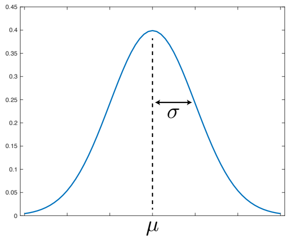
Curve matching. Pre-computed sounding curves are plotted on transparent paper of the same log-log scale as the sounding data (or *vice-versa*). Two-layer curves plot ρ_a/ρ_1 (apparent resistivity

'Stacking'. A *signal to noise ratio* (SNR) is simply how many times bigger the signal is than the noise obscuring the measurement process. That is, if the signal ($2\Delta V$) is 1 mV and the natural background and electrode noise is 0.2 mV the SNR is 5. For one measurement the error will therefore be 20%. If better than 20% accuracy is required several measurements may be made and averaged. If e is the error in one measurement then the error in an average of N independent such measurements is $e/\sqrt{N}$ and so noise decreases as the square root of the number of measurements.

This all comes from our friend the normal distribution, or Gaussian noise, which is given by the probability function on x by

$$f(x) = \frac{1}{\sigma\sqrt{2\pi}} e^{-\frac{1}{2}\left(\frac{x-\mu}{\sigma}\right)^2}$$

where μ is the mean and σ is the standard deviation. If we take an average of N samples of x



$$\sum_{i}^{N} x_i/N$$

this average is also normally distributed (this comes from a thing called *the central limit theorem*), with an *expectation value* of μ and a *variance* of σ^2/N, which gives a standard deviation of $\sigma\sqrt{N}$.

Another way to look at the problem is to realize that as measurements are summed, the signal grows linearly with N (that is, if 10 measurements of 1 mV are summed then the result is 10 mV). The noise, on the other hand, is as often negative as positive (i.e. in the above example the noise will fluctuate around ±0.2 mV) and so will tend to cancel.

Stacking is of great importance in many aspects of quantitative science. By its use a signal well below the noise level can be recovered with as much precision as the experimenter has time for. The use of stacking is prevalent in all aspects of exploration geophysics, particularly as modern microcomputers can do the necessary signal processing in real time.

divided by the resistivity of the top layer) as a function of A/h (electrode spacing divided by the thickness of the top layer) for many different ratios of ρ_1/ρ_2. Because of the logarithmic axes, the type curves may be slid horizontally and vertically to obtain a match between one of the type curves and the field data. The position of the point ($A/h = 1,\ \rho_a/\rho_1 = 1$) on the field plot then gives the resistivity and thickness of the first layer. The ratio ρ_1/ρ_2 for the matching curve yields the resistivity of the second layer.

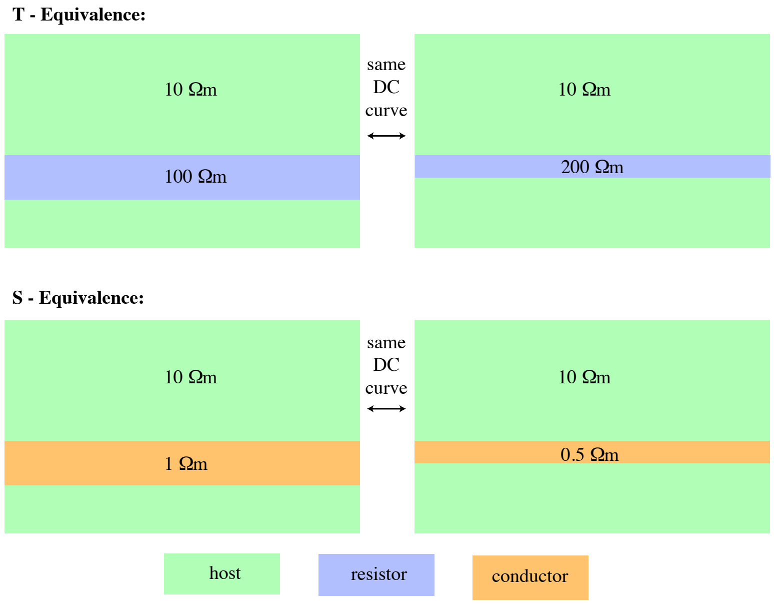
For soundings reflecting 3 layers, the same exercise may be performed with a suite of 3 layer curves, or repeated application of the 2 layer curves to both the top and bottom parts of the sounding data can produce satisfactory results.

Since the advent of fast computers and fast, simple algorithms for generating 1D sounding curves, curve matching has become obsolete.

Forward modeling. Modern computers and the development of efficient algorithms for predicting the resistivity response of layered structures makes trial and error or automatic curve matching using computers quite easy. A guess is made for a layered structure to fit the data and the resistivity response of the structure computed and plotted on the same graph as the data. The regions where the computed response fails to fit the data are noted, and with experience or an efficient computer algorithm the model can be corrected to bring the fits closer. This process is repeated (i.e. one *iterates*) until a satisfactory fit is obtained, leaving the final model as the estimate of the true structure of the Earth. If the Earth really is 1D with the same number of layers as was used in the interpretation, this method produces very accurate results. If the assumptions are wrong, of course, the final model may be meaningless.

Inversion. There are schemes where the computer can automatically generate new models to test against the data, and make estimates of how well the model response fits the data. These inversion schemes are moderately complicated, and vary both in their efficiency and the type of models they generate. As we proceed through the geophysics course we will discover that there is rarely a unique solution available to describe a geophysical measurement. Particular trouble in resistivity occurs if a thin resistive (or conductive) layer is sandwiched between thicker conductive (or resistive) layers (*equivalence*), or if there exists several layers of steadily increasing (or decreasing) resistivity (*supression*).

In the case of equivalence, it is very difficult to tell apart models with beds of similar transverse resistance, $T = \rho t$ where ρ and t are the resitivity and thickness of a thin resistive layer within a conductive host, and models with layers of similar longitudinal conductance, $S = t/\rho = t\sigma$ where now the thin conductive bed is within a resistive host. These are called T and S equivalence respectively. Supression occurs in sections of steadily increasing or decreasing resistivity. The intermediate layers are not resolvable and are usually lost in the interpretation.



If a layer is *anisotropic*, with transverse resistvity ρ_t greater than longitudinal resistivity ρ_l, as is usually the case, we have the coefficient of anisotropy

$$\lambda = \sqrt{\rho_t/\rho_l}$$

So how do you compute the resistivity over a layered earth? This isn't examinable, but just so you know...

For a Schlumberger array over an earth composed of layers with resistivities $\rho_1, \rho_2 ... \rho_N$ and thicknesses $t_1, t_2, ... t_{N-1}$ (counting from the top down), the apparent resistivity is given by a thing called a Hankel transform:

$$\rho_a(AB) = \left(\frac{AB}{2}\right)^2 \int_0^\infty T_1(\lambda) J_1\left(\frac{AB}{2}\lambda\right) \lambda \, d\lambda$$

where J_1 is the first-order Bessel function of the first kind and $T_1(\lambda)$ is the *resistivity transform*, given for the top of each layer by

$$T_i = \frac{T_{i+1} + \rho_i \, tanh(\lambda t_i)}{1 + T_{i+1} \, tanh(\lambda t_i)/\rho_i} \quad .$$

Notice how the transform at the top of each layer depends only on the resistivity and thickness of that layer and the transform of the layer below. By starting at the bottom of the heap with $T_N = \rho_N$, we can compute $T_{N-1}, ... T_2, T_1$ in succession. This is called a *recurrence* relation and is how one gets T_1 at the surface of the earth.

This used to be a very difficult computation to make, because the Bessel function oscillates as it decay for large λ, making the integral have chunks of positive and negative contribution, but in 1971 'fast Hankel transforms' were discovered.

If we make the change of variables $x = ln(AB/2)$ and $y = ln(1/\lambda)$, then

$$\rho_a(x) = \int_{-\infty}^{\infty} T_1(y) J_1(e^{x-y}) e^{2(x-y)} dy$$

and if we define $F(x-y) = J_1(e^{x-y}) e^{2(x-y)}$ then

$$\rho_a(x) = \int_{-\infty}^{\infty} T_1(y) F(x-y) dy$$

which is a convolution integral with $T_1(y)$ as the input function, $\rho_a(x)$ as the output function and $F(x-y)$ at the filter function. Now, $F(x-y)$ is independent of the layer parameters t_i and ρ_i which appear only in $T_1(y)$.

The convolution may be approximated by a *linear filter*:

$$\rho_a = \sum_{k=k_{min}}^{k_{max}} T_1(\lambda_k) f_k$$

where f_k are the coefficients of a moving average filter. The values for this filter are computed by other people using P2C2E, but are published and available in lengths from 7 points to hundreds of points. So what was previously a fiendishly complicated problem has been reduced to computing the resistivity transform (a few hyperbolic tangents or exponentials) and a simple sum.

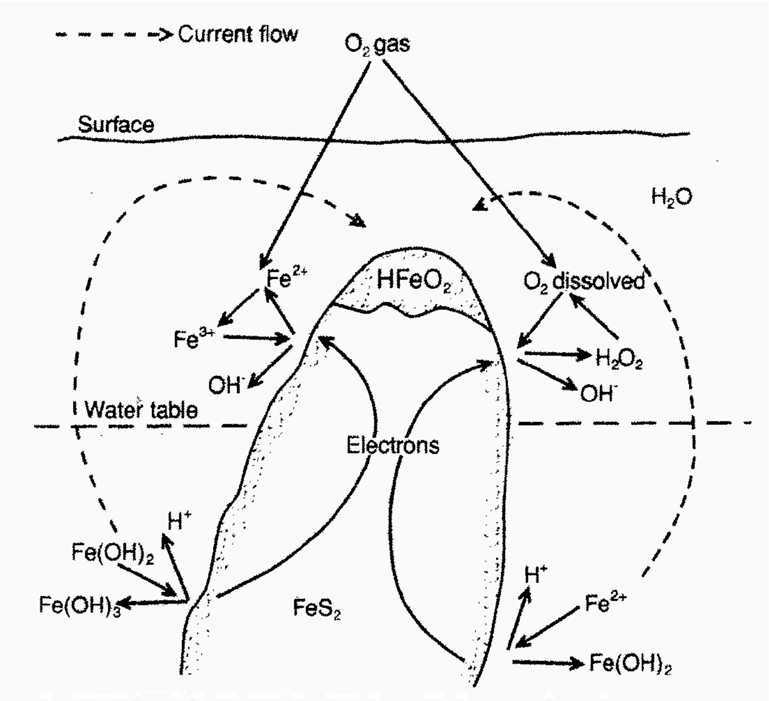
and mean resistivity

$$\rho_m = \sqrt{\rho_t \rho_l}$$

Anisotropic layers appear as layers of resistivity ρ_m and thickness λt, that is, depths are overestimated and resistivities are underestimated.

9. Self potential method

The natural potentials in the ground which produce noise for resistivity sounding are themselves capable of telling us something about subsurface strucure. In particular, there are often large potentials associated with conductive ore bodies. The nature of the production of the electric currents responsible for these potentials is not thoroughly understood, but this has not diminished the value of mineral resources discoved using the method. The field procedure for the self potential (SP) method is simply to take a sensitive voltmeter and a pair of non-polarising electrodes and, using one electrode as a reference, map the potential on the ground surface. Negative anomalies of potential characteristically occur over mineralized zones.



You also get self potentials when groundwater flows though soil that has clay minerals, another effect associated with the surface charge on these minerals, and with temperature gradient around geothermal areas.

10. Induced polarization method

We mentioned polarization in our discussion of laboratory measurements of electrical conductivity. It was pointed out that if charge builds up on the surface of a specimen the apparent conductivity will decrease. If a the voltage driving the DC current is then turned off, a residual voltage will be seen across the sample for a short time as the charge re-equilibrates. This same phenomenon is seen in rocks *in situ*; if a current is passed through the ground while the potential difference is monitored across a second pair of electrodes, after the current is interrupted there is an initial drop in voltage followed by a slow decay.

The cause of the polarization is twofold. The charged surfaces of minerals attract ions from the electrolyte in pores and thereby produce concentrations of charge, which can impede the applied current flow if the pore spaces are small enough. This is called *membrane polarization* and is found everywhere, especially in clayey materials. *Electrode polarization*, on the other hand, is caused by conductive mineral grains blocking the pores of the host rock. Charge builds up on the surfaces of these grains, which also inhibit the movement of ions in the electrolyte filling the pores. Electrode polarization is of greater magnitude than membrane polarization, and is obviously the reason IP is used for exploration, as oxides, sulphides and native metals all exhibit this sort of IP effect (as does, unfortunately, graphite).

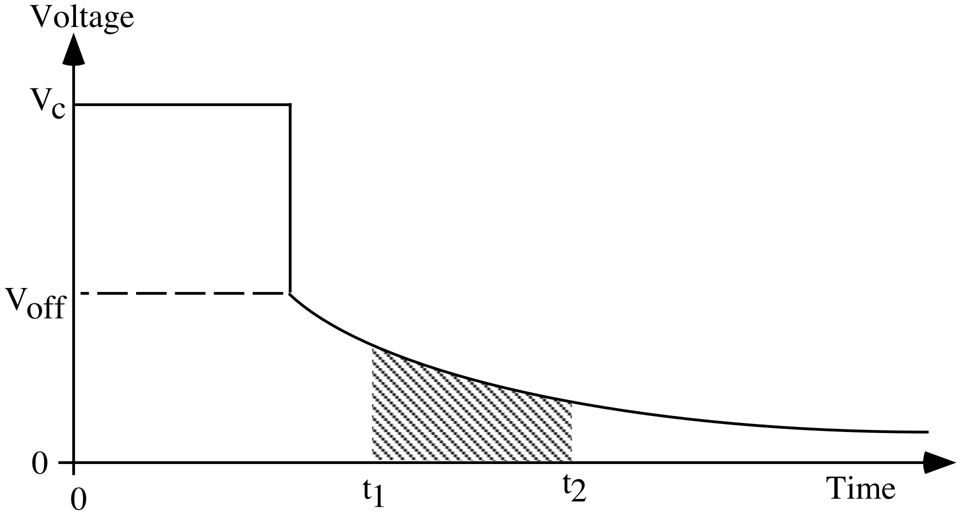


Fig. E.16: Induced polarization. After the current is suddenly switched off in the AB circuit, the voltage in the MN circuit drops, but not to zero. The residual voltage decays to zero over a period of time. The chargeability is measured by integrating the residual voltage between two time t_1 and t_2.

The IP meausurement must somehow quantify the extent of the polarization. This is done either in the *time domain* by sharply interrupting a DC current, as illustrated above, or in the *frequency domain* by making resistivity measurements at two frequencies, one low enough to make a DC meausurement and the other high enough to be influenced by polarization effects.

Chargeability, a time domain measure, is the integral of the decay voltage as a proportion of the continuous voltage before cutoff:

$$M = \frac{1}{V_c} \int_{t_1}^{t_2} V(t)\, dt$$

M has units of seconds (usually ms), and clearly depends on t_1 and t_2, which must be fixed to obtain comparable results.

Percent frequency effect is a frequency domain measurement, obtained by estimating the apparent

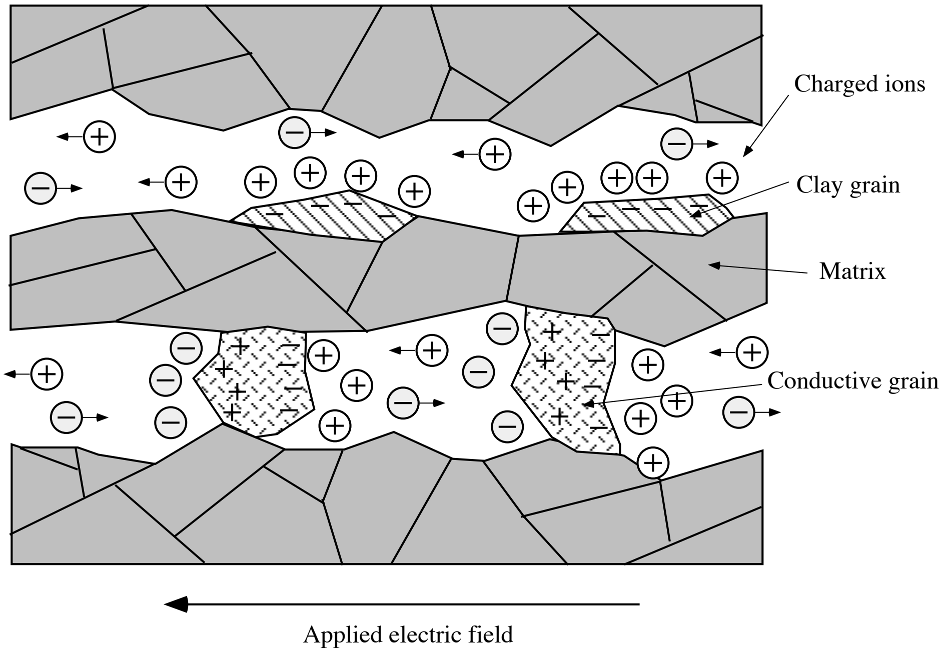


Fig. E.17: The storage of charge occurs in the ground for two reasons. Firstly, surface charges on clay minerals attract dissolved ions and impede their flow in response to the applied electric field. Secondly, conductive grains (metal, sulphides or graphite) blocking pores conduct electricity but do not allow the flow of electrolytic charge, causing charge buildup near their surfaces, or polarization.

resistivity at low and high frequencies:

$$PFE = 100(\rho_{dc} - \rho_{ac})/\rho_{ac}$$

Again, the result will depend on the actual frequencies used, which will usually be in the range of 0.1 - 10 Hz.

Metal factor, another frequency domain measure, attempts to normalise the IP measurement by the DC resistivity:

$$\begin{aligned} MF &= 2\pi \times 10^5(\rho_{dc} - \rho_{ac})/\rho_{ac}\rho_{dc} \\ &= 10^3 PFE/(\rho_{dc}/2\pi) \end{aligned}$$

Equipment for IP work is very similar to that for resistivity. Indeed, resistivity measurements are almost always made at the same time as IP measurements. However, the decay voltages are smaller than the V_c voltage which gives the DC value, so transmitter currents tend to be larger and stacking times longer. A motor-driven generator is usually required, whereas resistivity work may often be accomplished with batteries. The dipole-dipole array is preferred for IP work, for two main reasons. One is that larger currents are needed, so more time is spend on reducing the AB grounding impedance and then one does not wish to move the AB electrodes. The other is that psuedo-sections are a common way of presenting the results.

Results are presented as profiles, but psuedo-sections are favoured in IP studies. Quantitative modelling is rare, but can be accomplished in conjuntion with 2D resistivity modelling.