Geophysics

Seismic Method

* Basic principles

* The seismic method makes use of the properties of the velocity of sound. This velocity is different for different rocks and it is this different which is exploited in the seismic method.
* When we create sound at or near the surface of the earth, some energy will be refracted back (bounced back). They can be characterized as echoes. From these echoes we can determine the velocities of the rocks, as well as the depths where the echoes came from .
* When we use the seismic method, we usually discuss two types of seismic methods, depending on whether the distance from the sound source to the detector (the "ear") is large or small with respect to the depth of interest: the first is known as refraction seismic, the other as refraction seismic.

* **Physical-base**

1.Theory of Elasticity

* The seismic method utilize the propagation of waves of the earth, since this propagation depends upon the elastic properties of the rocks. The size and shape of a solid body can be changed by applying forces to the external surface of the body.
* These external forces are opposed by internal forces which resist the changes in size and shape. As a result the body tends to return to its original condition when the external forces are removed.
* This property of resisting changes in size or shape and of returning to the un deformed condition when the external forces are removed is called elasticity.
* Elasticity is the property of solid materials to return to their original shape and size after the forces deforming them have been removed. Recall Hooke's law — first stated formally by Robert Hooke .

* When the seismic waves propagate in solids caused deformation, we can describe these deformations in terms of the forces, defining two useful concepts stress and strain **.**
* **Stress (σ).** Stress is defined as force per unit area : $\sigma = \dfrac{F}{A}$

Where: σ=Stress F=Force A=Area

* The Stress may be divided in two type:
* **Normal (vertical) stress**
* When the force is perpendicular to the element.
* **2 -Shearing Stress**
* When the force is neither parallel nor perpendicular to the element

* **Strain (ε)**. When an elastic body is subjected to stresses, changes in shape and dimensions occur. These changes are called strains .

* **1- Longitudinal Strain(ε_l)**

* Defined as the ratio of change between length (ΔL) to origin length (L) that is a result to vertical stress. $\varepsilon_l = \frac{\Delta L}{L}$

* **2- Transversal Strain (ε_w)**

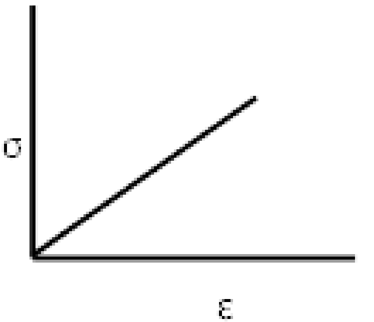
* The ratio between the change of width (Δw) to the origin width (w), that is a result to vertical stress. $\varepsilon_w = \frac{\Delta w}{w}$

* **3- Shear Strain**

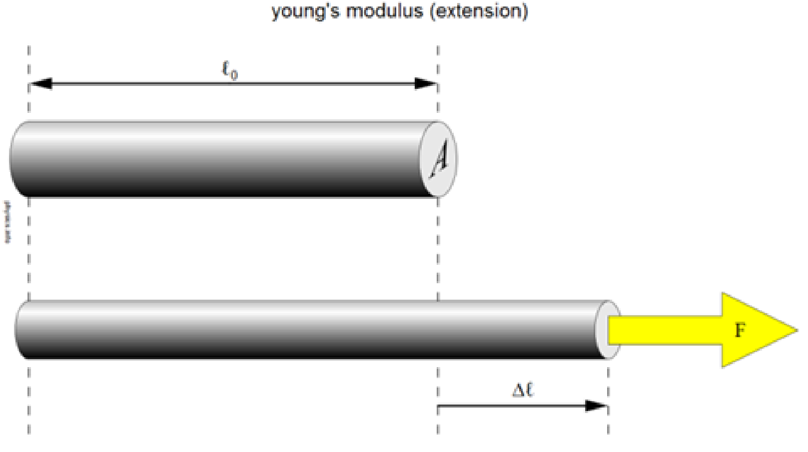
* This is measured for tan of deformation angle (tan Ø) that is result to shearing stress.

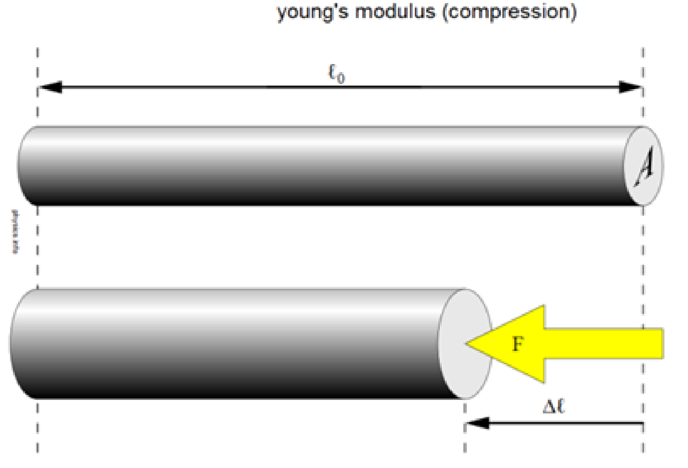
* Stress-Strain and elastic Moduli

* According to Hook's law that the linear relationship between the stress and strain is show in (Fig.1).



* **1- Young's Modulus (E)**:- The ratio of the stress to the deformation is a measure of the property of the rock to resist deformation. $E = \frac{\left(F/A\right)}{\left(\Delta L/L\right)} = \frac{FL}{A\Delta L}$





* **2-Bulk Modulus (B)**:-The Bulk Modulus is defined as the compressional stress by change in volume. $B = \frac{\Delta P}{\Delta V / V}$

* **3- Compressibility modulus (**K**)**:- The compressibility is the reciprocal of the bulk modulus. $K = \frac{1}{B}$

* **4- Shear (Rigidity) modulus (G)**:- That is relative consistency between shearing stress and shearing strain, also it's resistance measurement to shearing strain for materials. $G = \frac{\left(F / A\right)}{\phi}$

* 5- **Poisson's Ration (*μ*)**:- That is Relationship between width strain and length strain while the body under compressional and tensile stress. $\mu = \frac{\left(\Delta W / W\right)}{\left(\Delta L / L\right)} = \frac{L \Delta W}{W \Delta L}$

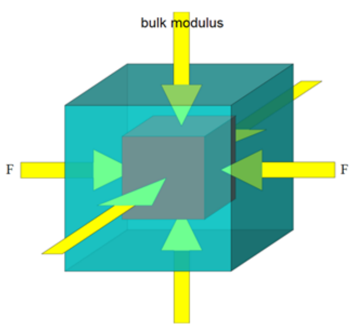


Fig.(3).Types of Bulk modulus

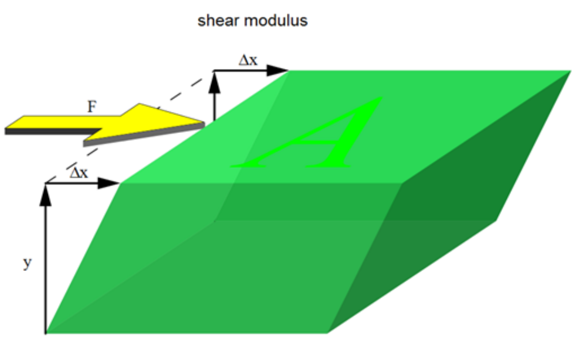
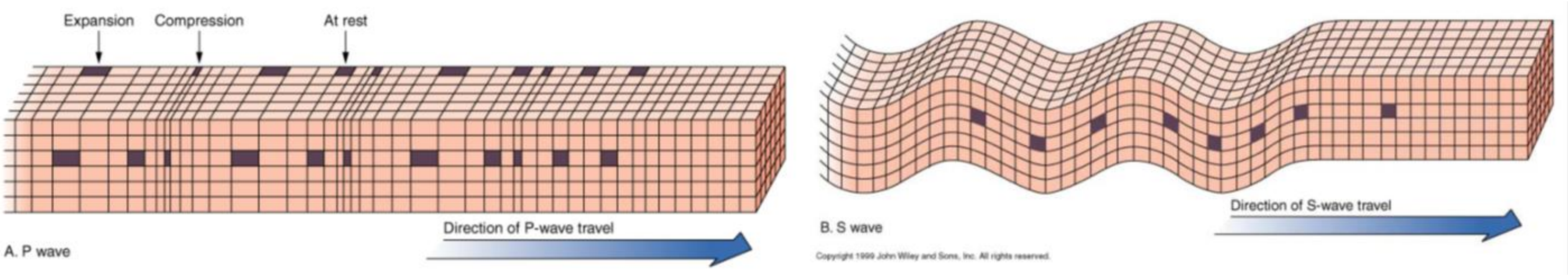


Fig.(4).Types of shear modulus

Seismic Waves

- Seismic waves propagate in solid as pattern of particle deformation traveling through the materials with velocities that depend upon their plastic module and density, that can be expressed by: $velocity = \sqrt{\frac{elasticModuli}{density}}$
- There are two mainly types of elastic seismic waves:
- **1. Body Waves**
- The body waves propagate through subsurface layers, and are divided into:
- **a- Compressional wave (P-wave)**
- This type of wave variously known as a dilatational, longitudinal, irrotational compressional or P-wave, is usually the first (primary) event. The P-wave is common in seismic exploration. The motion of the particles is always in the direction of wave propagation (left).
- **Shear Wave**
- The second event observed on seismic records, is known as shear, transverse, rotational, or S-wave.The motion of individual particles is always perpendicular to the direction of wave propagation (right).



4. Relationship between Vp and Vs

* Compressional waves travel fastest at speeds between (1.5) to (8) kilometers per second in the Earth's crust. Shear waves, also known as secondary or S waves, travel more slowly, usually at 60% to 70% of the speed of P waves. P waves shake the ground that the particles motion in the direction of propagating, while S waves shake perpendicularly or transverse to the direction of propagation.

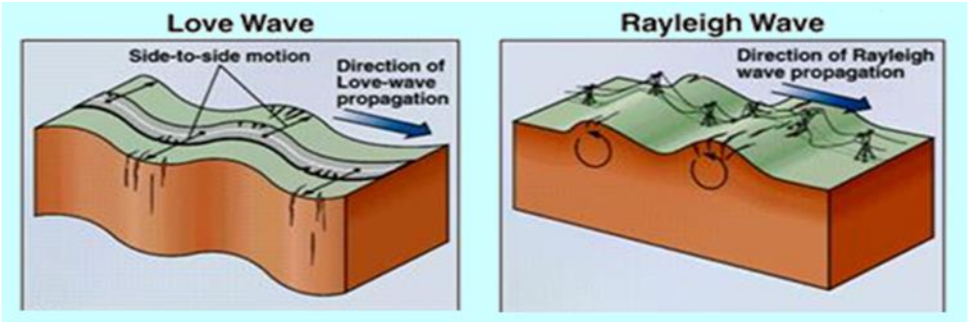
*$V_s = 0.7\ V_p$ .... (18)

* For most consolidated rock materials V_p/V_s is between (1.5-2). Shear waves will not propagate in liquid materials because for liquids, rigidity (G) is zero .

* 2. Surface Waves

* **a- Rayleigh Wave:-** This wave travels along the surface of the solid materials. The amplitude of this wave motion decreases exponentially with depth. The particle motion, always in a vertical plane, is elliptical and retrograde with respect to the direction of propagation. $V_R=0.9\ Vs$, that V_R is Rayleigh wave velocity

* **b- Love Wave:-** A love wave involves transverse motion parallel to the surface of the ground is observed only when there is a low-speed layer overlying a higher-speed substratum. The wave motion is horizontal and transverse .



* There are several factors affected on the seismic wave's velocities which are :

1. Petrophysics property of rocks (lithology, density, porosity, pores size and shape, fluids content, minerals content and saturation). Porosity and fluid saturation , increasing porosity reduces velocity, Filling the porosity with fluid increases the velocity.

* $$\frac{1}{V_{sat}} = \frac{\phi}{V_F} + \frac{1-\phi}{V_M}$$

1. Tectonic and compaction movements.
2. Depth of subsurface rocks and age of rocks.
3. Cracks and joints.
4. Anisotropy.
5. Pressure and Temperature.

* Relationship between siesmic waves and elastic module

$$V_p = \sqrt{\frac{\lambda + 2G}{\rho}} = \sqrt{\frac{B + \frac{4}{3}G}{\rho}} \quad \ldots\ldots\ldots\ldots\ldots \quad (15)$$

$$V_p = \sqrt{(E/\rho)(1-\mu)/(1-2\mu)(1+\mu)} \ldots\ldots\ldots\ldots$$

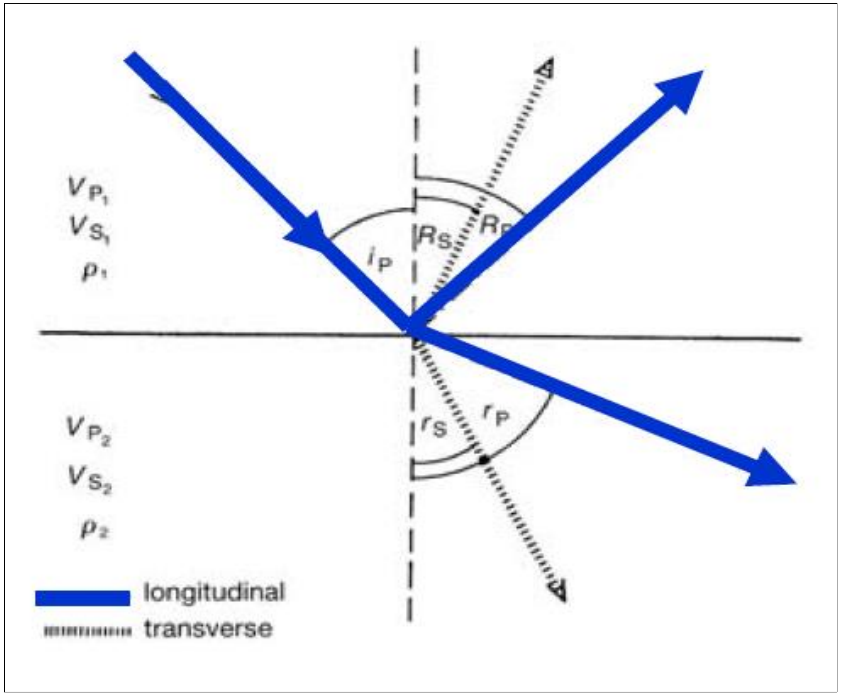
$$Vs = \sqrt{\frac{G}{\rho}} = \sqrt{\frac{E}{2\rho(1+\mu)}} \quad \ldots\ldots\ldots\ldots\ldots \quad (17)$$

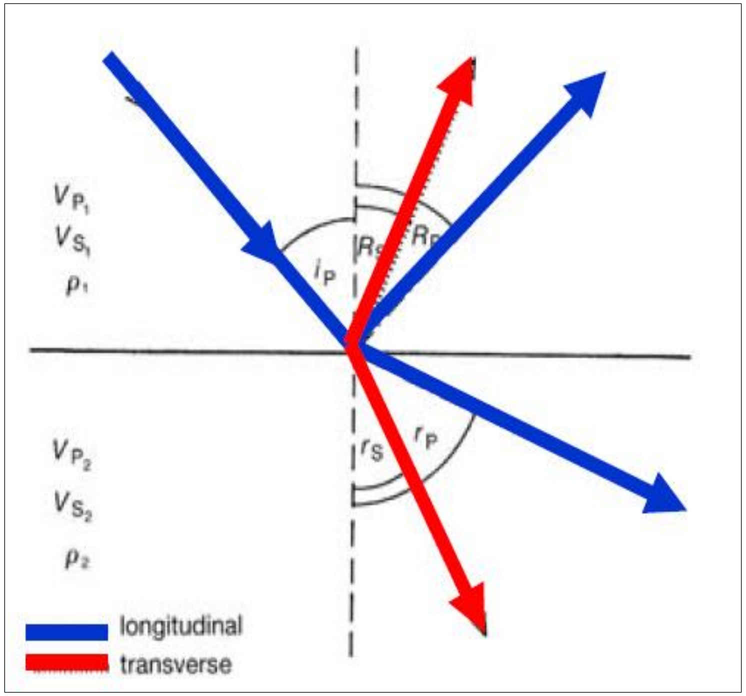
Seismic methods

* Seismic method is divided into two types depending on the power source of seismic waves generating , reflection and refraction. The seismic waves have been studied by capturing on the surface of the Earth by:
* 1. Natural seismology (Seismology Earthquake): to identify the physical qualities and underground installation.
* 2. Explosive seismology (Seismology Explosion). Artificial explosions (nuclear) at selected sites on the surface of the Earth for the purpose of obtaining information on superficial underground geological structures.

Reflection and transmission

* Seismic rays obey Snell's Law (just like in optics), the angle of incidence equals the angle of reflection, and the angle of transmission is related to the angle of incidence through the velocity ratio.
* The angle of incidence equals the angle of reflection, and the angle of transmission is related to the angle of incidence through the velocity ratio. But a conversion from P to S or vice versa can also occur. Still, the angles are determined by the velocity ratios





* **Seismic rays and reflection – refraction transmission**

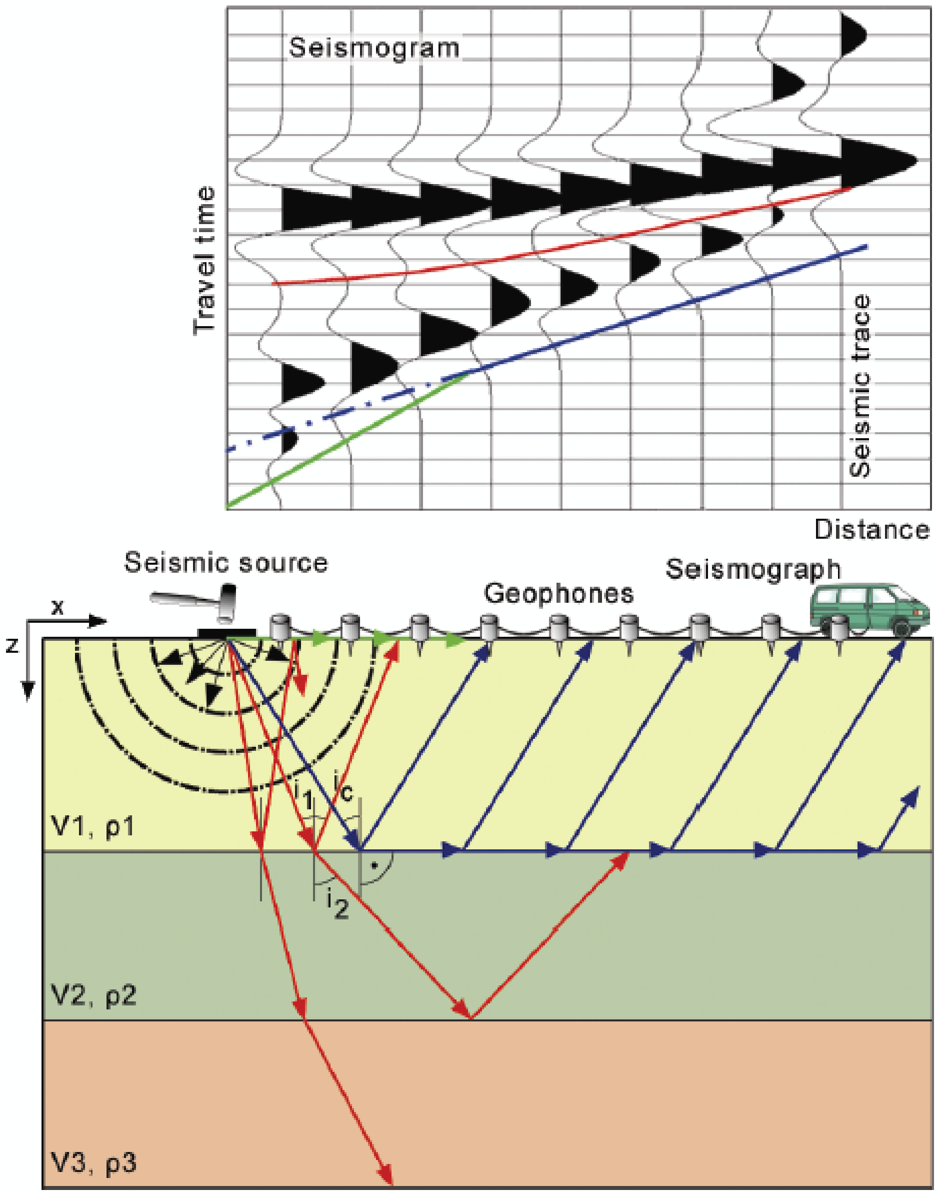
* Seismic refraction

* Seismic refraction is a geophysical principle governed by Snell's Law. Used in the fields of engineering geology, geotechnical engineering and exploration geophysics, seismic refraction traverses (seismic lines) are performed using a seismograph(s) and/or geophone(s), in an array and an energy source.
* The seismic refraction method utilizes the refraction of seismic waves on geologic layers and rock/soil units in order to characterize the subsurface geologic conditions and geologic structure.
* The methods depend on the fact that seismic waves have differing velocities in different types of soil (or rock): in addition, the waves are refracted when they cross the boundary between different types (or conditions) of soil or rock. The methods enable the general soil types and the approximate depth to strata boundaries, or to bedrock, to be determined.

* Field work

* The seismic refraction method is based on the measurement of the travel time of seismic waves refracted at the interfaces between subsurface layers of different velocity.
* Seismic energy is provided by a source ('shot') located on the surface. For shallow applications this normally comprises a hammer and plate, weight drop or small explosive charge (blank shotgun cartridge).
* Energy radiates out from the shot point, either travelling directly through the upper layer (direct arrivals), or travelling down to and then laterally along higher velocity layers (refracted arrivals) before returning to the surface.
* This energy is detected on surface using a linear array (or spread) of geophones spaced at regular intervals. Beyond a certain distance from the shot point, known as the cross-over distance, the refracted signal is observed as a first-arrival signal at the geophones (arriving before the direct arrival). Observation of the travel-times of the direct and refracted signals provides information on the depth profile of the refractor

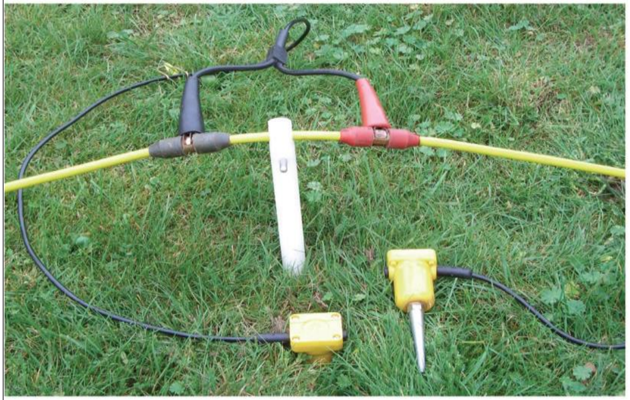
* Sketch of seismic survey

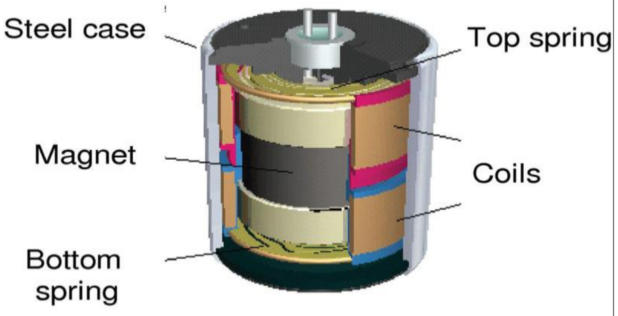


*Seismic sources

- Rifles and guns (Cheap , Repeatable – fire into water filled hole, Shallow targets 0-50m)
- Sledge hammer (Cheap, Repeatable once plate is stable, Targets 15-50m)
- Weight drops(Cheap, Repeatable – automated, Targets > 50m)
- Vibroseis(No pulse, frequency sweep, Significant signal with stacking /deconvolution)
- Explosives(Various sizes – target depth, Safety and expense can be an issue)
- Air guns(At sea, Very repeatable, Large array for big signal)

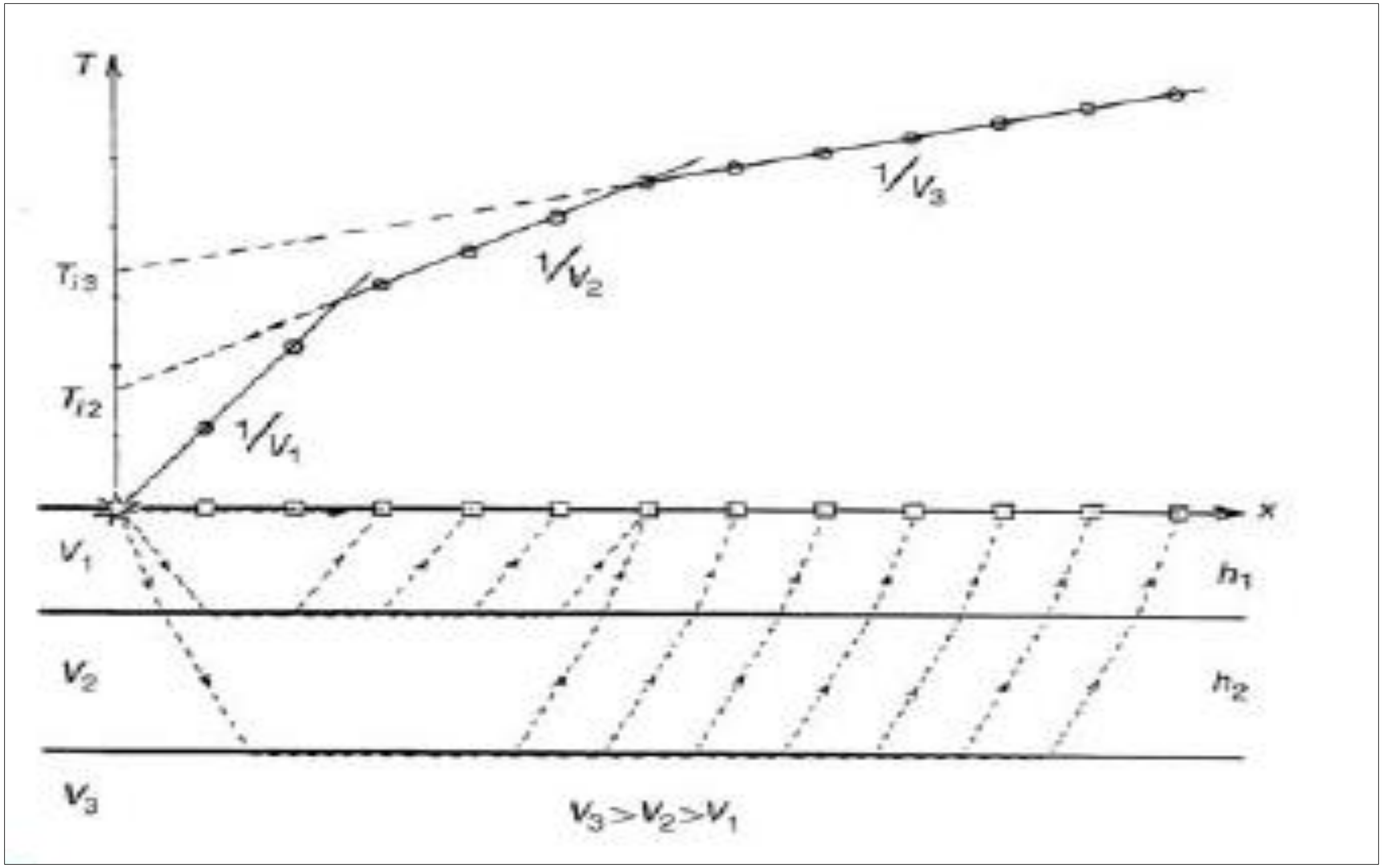
* **Seismic receiver**
* Geophones, Cylindrical coil suspended in a magnetic field, the inertia of the coil causes motion relative to the magnet generating a electrical signal, geophones are sensitive to velocity



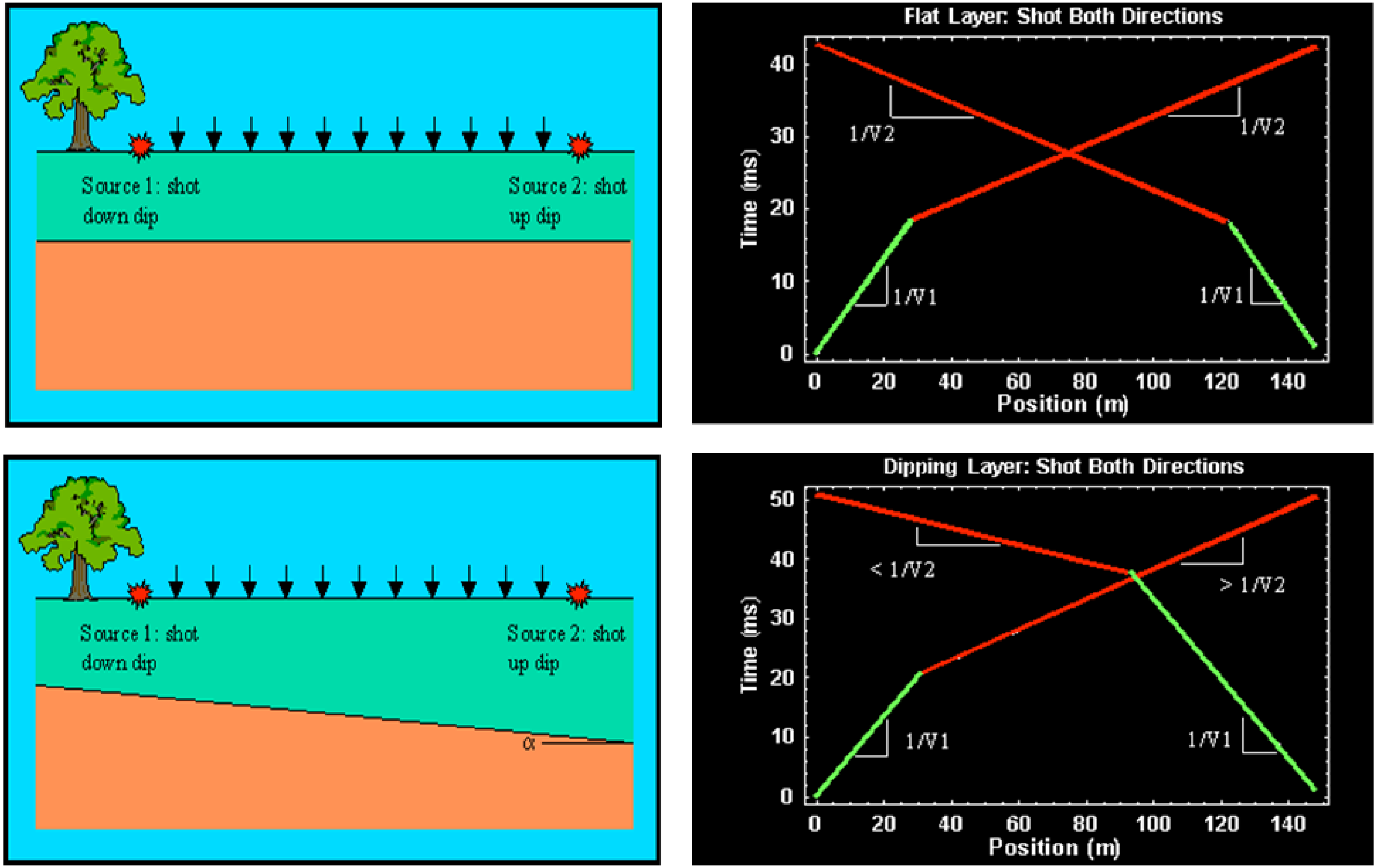


* Interpretation

* 1- Shots are deployed at and beyond both ends of the geophone spread in order to acquire refracted energy as first arrivals at each geophone position. Data are recorded on a seismograph and later downloaded to computer for analysis of the first-arrival times to the geophones from each shot position
* 2- Travel-time versus distance graphs are then constructed and velocities calculated for the overburden and refractor layers through analysis of the direct arrival and T-minus graph gradients.
* 3- Depth profiles for each refractor are produced by an analytical procedure based on consideration of shot and receiver geometry and the measured travel-times and calculated velocities.
* 4- The final output comprises a depth profile of the refractor layers and a velocity model of the subsurface.



*Sketch of seismic survey show the time vs distance



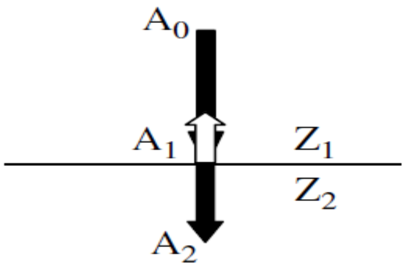
* Seismic path with shot both direction in horizontal and tilt layers.

* Seismic reflection

- In sedimentary material, elasticity and density strongly depend on porosity. At a layer boundary, e.g. between sand and clay or till, a porosity change normally occurs, leading to contrasting densities and seismic velocities.
- A seismic wave impinging on this layer boundary will be partly reflected and partly refracted. The intensity of the reflected wave depends on the magnitude of the contrast between seismic velocities and densities at the boundary, regardless of the sign of the contrast.
- The product of velocity V and density ρ is the acoustic impedance $I = V\rho$ of a medium. The strength of a reflection from an acoustic contrast interface is defined by the *reflection coefficient R*

$$R = \frac{I_2 - I_1}{I_2 + I_1}$$

* with I_1 = acoustic impedance of the first layer and I_2 = acoustic impedance of the second layer. Equation 21 is valid for normal incident rays or waves (ray path perpendicular to layer boundary).



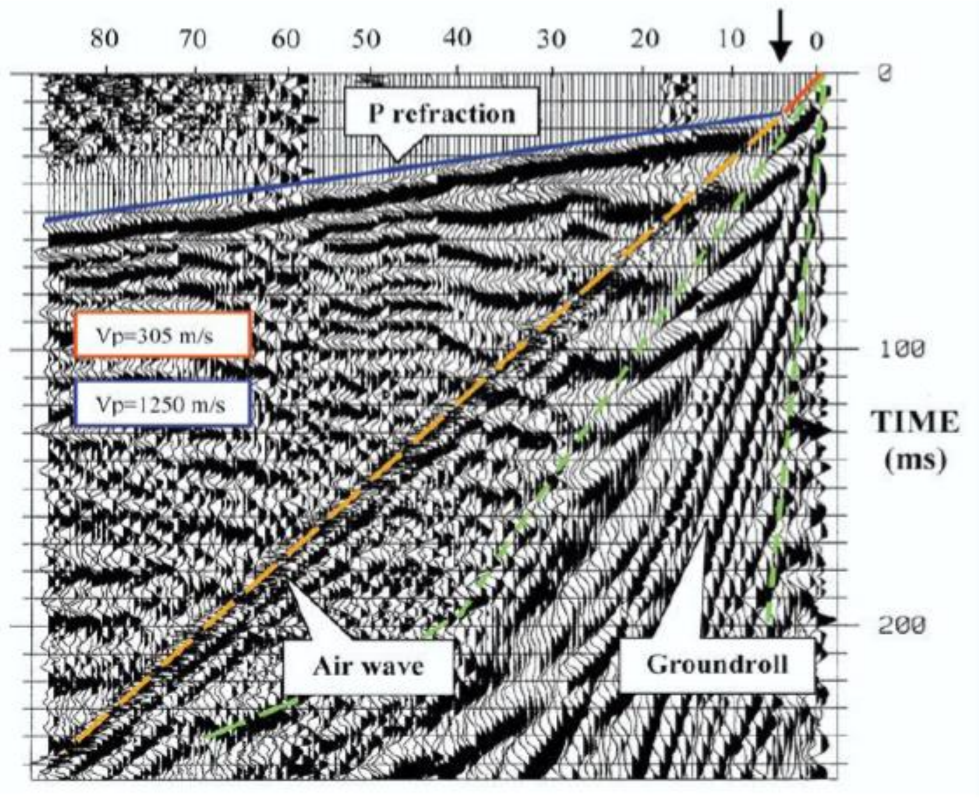
* **Z(I):** acoustic impedance

* Normal reflection and transmission

* Field work

* Land seismic surveys tend to be large entities, requiring hundreds of tons of equipment and employing anywhere from a few hundred to a few thousand people, deployed over vast areas for many months.
* There are a number of options available for a controlled seismic source in a land survey and particularly common choices are Vibroseis and dynamite.
* A land seismic survey requires substantial logistical support. In addition to the day-to-day seismic operation itself, there must also be support for the main site or camp (for catering, waste management and laundry etc.), smaller camps (for example where the distance is too far to drive back to the main camp with vibrator trucks), vehicle and equipment maintenance, medical personnel and security.
* Unlike in marine seismic surveys, land geometries are not limited to narrow paths of acquisition, meaning that a wide range of offsets and azimuths is usually acquired and the largest challenge is increasing the rate of acquisition

* The seismic waves reflected have been picked up by geophones which distributed on the earth surface, then its sent to the recording unit to record on magnetic tape, then a primary and advance processing are being to strengthen useful signals (Signals) and reducing the noise, the final result as a seismic section is ready for final interpretation



Typical data plots (seismic recording)

* Important differences between refraction and reflection seismic

REFRACTION SEISMICS	REFLECTION SEISMICS
Based on contrasts in : seismic wave speed (c)	Based on contrasts in : seismic wave impedances (ρc)
Material property determined : wave speed only	Material properties determined: wave speed and wave impedance
Only traveltimes used	Traveltimes and amplitudes used
No need to record amplitudes completely : relatively cheap instruments	Must record amplitudes correctly : relatively expensive instruments
Source-receiver distances large compared to investigation depth	Source-receiver distances small compared to investigation depth

<u>Geophysical methods</u> respond to the physical properties of the subsurface media (rocks, sediments, water, voids, etc.) and can be classified into two distinct types: **<u>Passive methods</u>** are those that detect variations within the natural fields associated with the Earth, such as the gravitational and magnetic fields. In contrast are the **<u>Active methods</u>**, such as those used in exploration seismology and **DC Electrical Resistivity** in which artificially generated signals are transmitted into the ground, which then modifies those signals in ways that are characteristic of the materials through which they travel. The altered signals are measured by appropriate detectors whose output can be displayed and ultimately interpreted.

Overview of Applied Geophysics

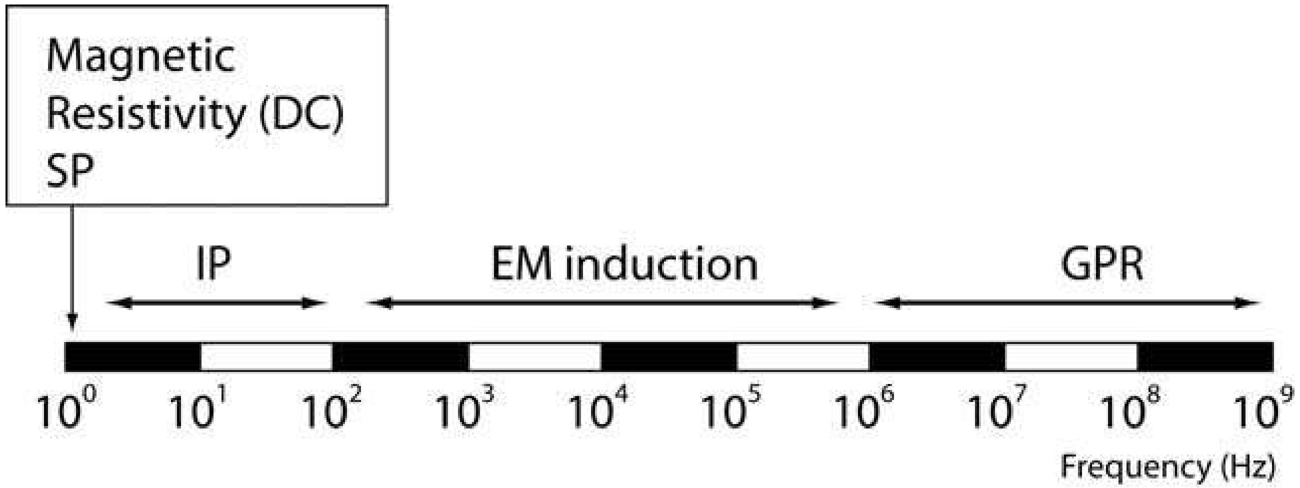
Applied geophysics provides a wide range of very useful and powerful tools which, when used correctly and in the right situations, will produce useful information about the subsurface.

What main physical properties do we measure?

- Density
- Magnetic properties
- **Resistivity, conductivity**
- Acoustic Velocity
- Electromagnetic properties

To work – the target body must have sufficiently different physical properties to the surrounding rock and Materials

Electrical surveying methods…The electrical methods is one of the geophysical techniques, and it is basing on passing natural electrical currents or artificial in the Earth.



- **DC Resistivity method**
- Induced polarization method (IP)
- Self-potential (SP) method

Higher frequency methods (Electromagnetic surveys):

- ✓ Electromagnetic induction methods
- ✓ Ground penetrating radar (GPR)

Electrical Resistivity Method

- Resistivity surveying investigates variations of electrical resistance, by causing an electrical current to flow through the subsurface using wires (electrodes) connected to the ground. The fundamental physical law used in resistivity surveys is Ohm's Law that governs the flow of current in the ground.

➢ **Resistivity = 1 / Conductivity**

❖Link resistivity (i.e., the ability of the earth to prevent the conduction of an electric current) to the subsurface structure.

❖Useful because Resistivity is one of the most variable physical properties, i. e., resistivity of earth materials varies by around 10 orders of magnitude.

$$1.6\times10^{-8}\ \Omega m < \rho < 10^{16}\ \Omega\ m$$

native silver **pure sulfur**

Resistivity method

The Electrical resistivity method is used in the study of horizontal and vertical discontinuities in the electrical properties (resistivity/conductivity) of the subsurface

Application of Elec. Resistivity…

- Exploration of bulk mineral deposit (sand, gravel and metals)
- Exploration of underground water supplies (hydrogeology)
- Engineering/construction site investigation
- Environmental: e.g., Waste sites and pollutant investigations
- Cavity, karst detections
- Archaeological investigations
- Geology problems
- Glaciology, permafrost

1- The Fundamental Of DC Resistivity (Ohm's Law):

The main fundamental concepts of DC electrical resistivity method depend on ohm's law which back credit for the German scientist George ohm in 1827 who explained that, the electrical current I in a conductive wire represents the proportional to the potential deference ΔV across it (see Fig. 1). Equation (1) is expressed this linear relationship (Lowrie, 2007).

$$\Delta V = IR \;\ldots\ldots\ldots 1$$

Where I is the electrical current,
Δ V is the potential deference and
R represents the resistance of the conductor.
The unit of resistance is measured in ohm (Ω).

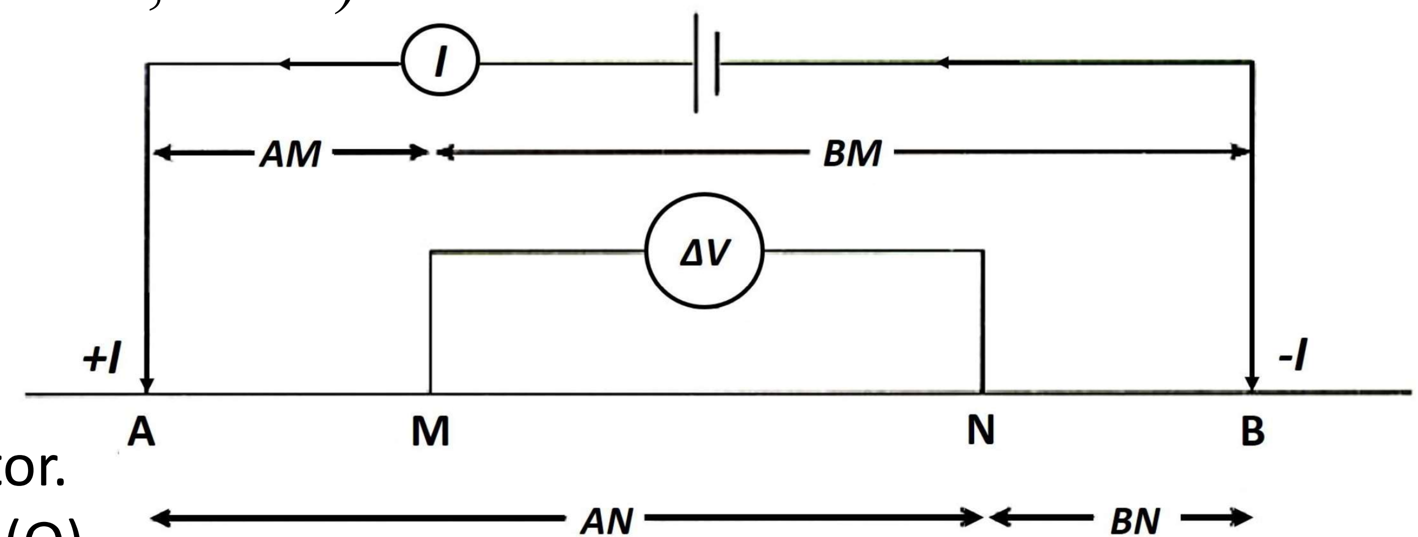


Figure 1

- However, no units of length in this form of Ohm's law.

For more precisely, for a specific material the resistance is proportional to the length L and reversely proportional to the cross-section area A of the conductor (Figure 2)

The relationship between the parameters (I, V, and geometry) could be expressed in the Equation 2:

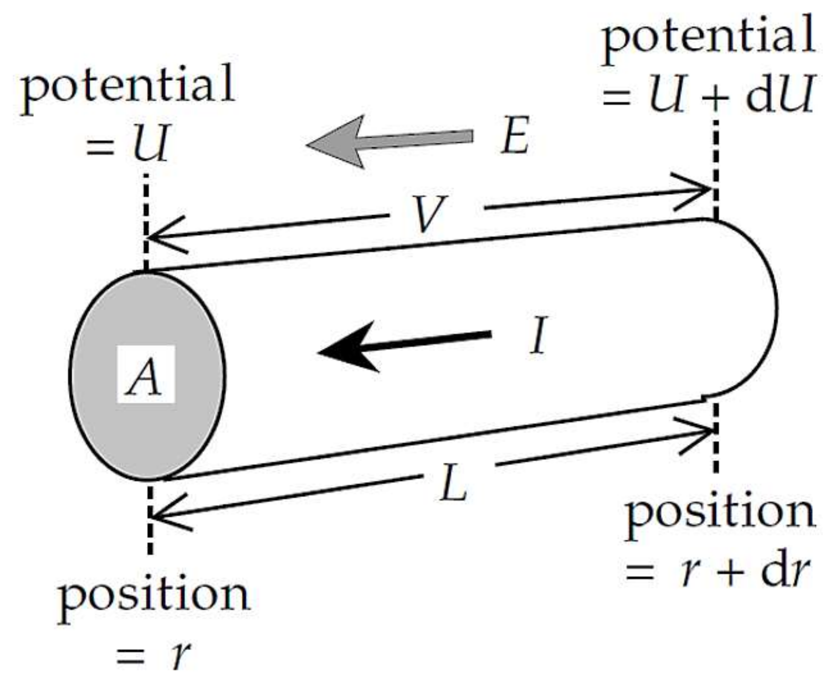
$$R = \rho * l/A \quad ....2$$

Which the proportional constant R is the resistance of the conductor. The resistivity (ρ) is a physical attribute of the material conductor expresses its capability to oppose a flow of charge. The unit of resistivity measured in ohm.m (Ωm). From substituting eq. 1 for R in eq. 2 and re-position the terms we get the following Expression 3:

$$V/L = \rho * I/A \quad ....3$$

From Fig. 2, the ratio V/L represents the electric field ***E****, and the ratio I/A represents the current density* ***J****, so we can re-write ohm's law as:* $\mathbf{E} = \rho * \mathbf{J}$ *.........4*

This eq. 4 is just the generalization for Ohm's Law in vector form for current flow in a continuous medium.



$$\text{electric field } E = \frac{-\,dU}{dr} = \frac{V}{L}$$

$$\text{current density } J = \frac{\text{current}}{\text{area}} = \frac{I}{A}$$

Fig. 2: The parameters used in defining resistivity

2- Current Flow And Voltage:

In practice, using ohm's law parameters to determine resistivity (or conductivity) of ground material, it supposed to inject the electric current by one or pair of electrodes located on the surface of half-space acting the ground, and measure the voltage in another dipole, as shown in **Figure (3)**.

Taking into account that the electric current cannot flux through the non-conducting air, and inflows radially outward over a hemisphere of radius r and surface area $\mathbf{2\pi r^2}$.

In this state current density (Equation 5) is:

$$J = Ir/2\pi r^2 \quad \text{……….5}$$

So that using equation(3) and (4), the measured voltage at point M (in Fig. 3) is:

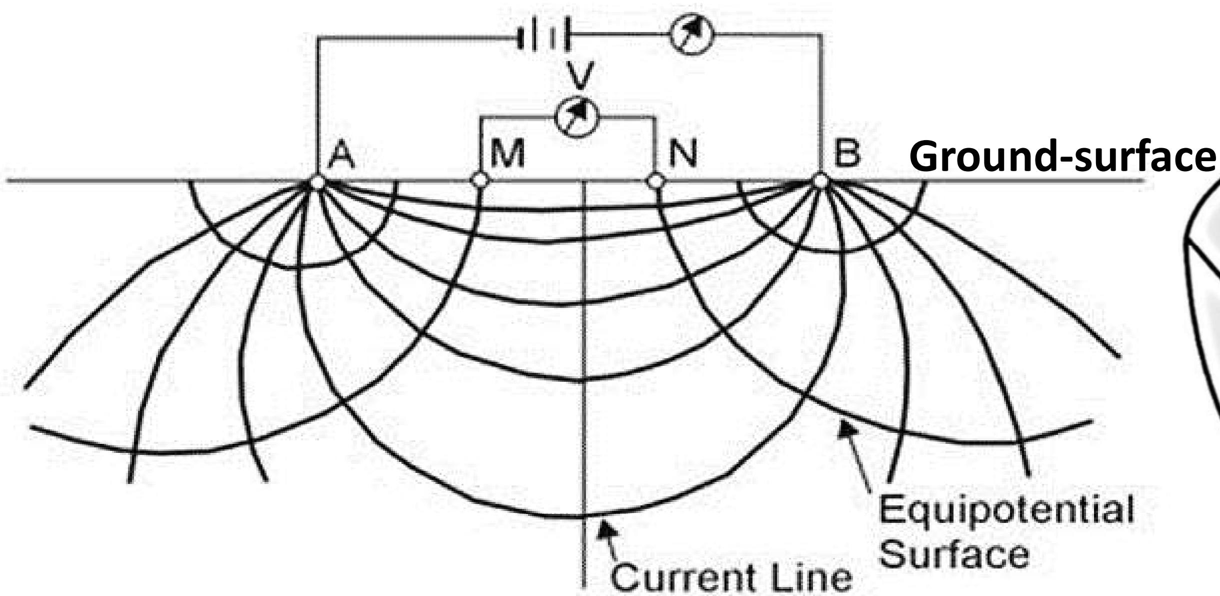
$$V = I\,\rho\,/2\pi r_M \quad \text{………6}$$

Where $\mathbf{r_M}$ represents the distance from the current inject pole to the potential one (e.g., M).
The equation (6) represents the basic equation to calculate resistivity.

The measured potential across the dipole M & N of the voltmeter in figure 3 is the potential difference.

$$V_{MN} = V^{AB}_{M} - V^{AB}_{N} = I\,\rho\,/2\pi[1/r_{M} - 1/r_{N}] \;\ldots\ldots 7$$

where r_N is the distance from inject current point to the potential pole N.



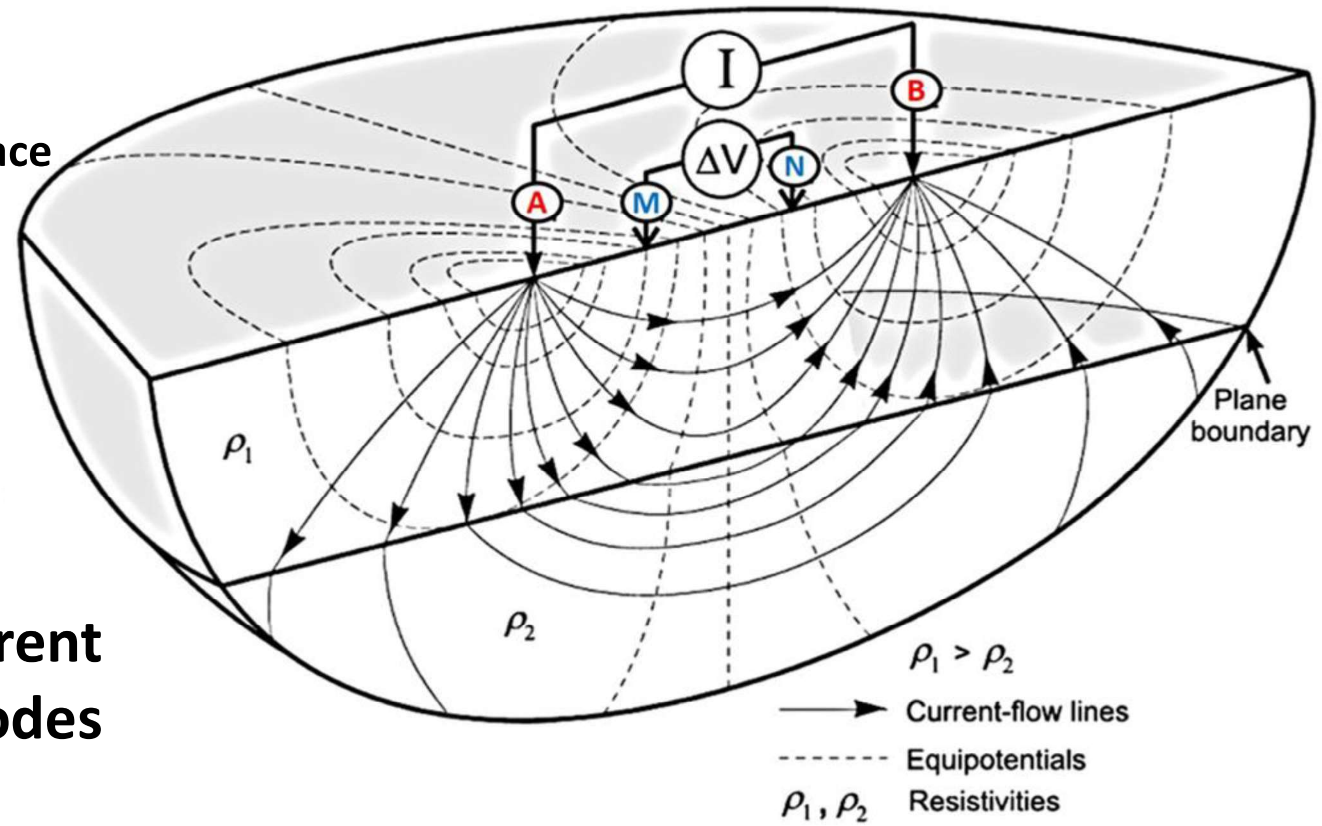
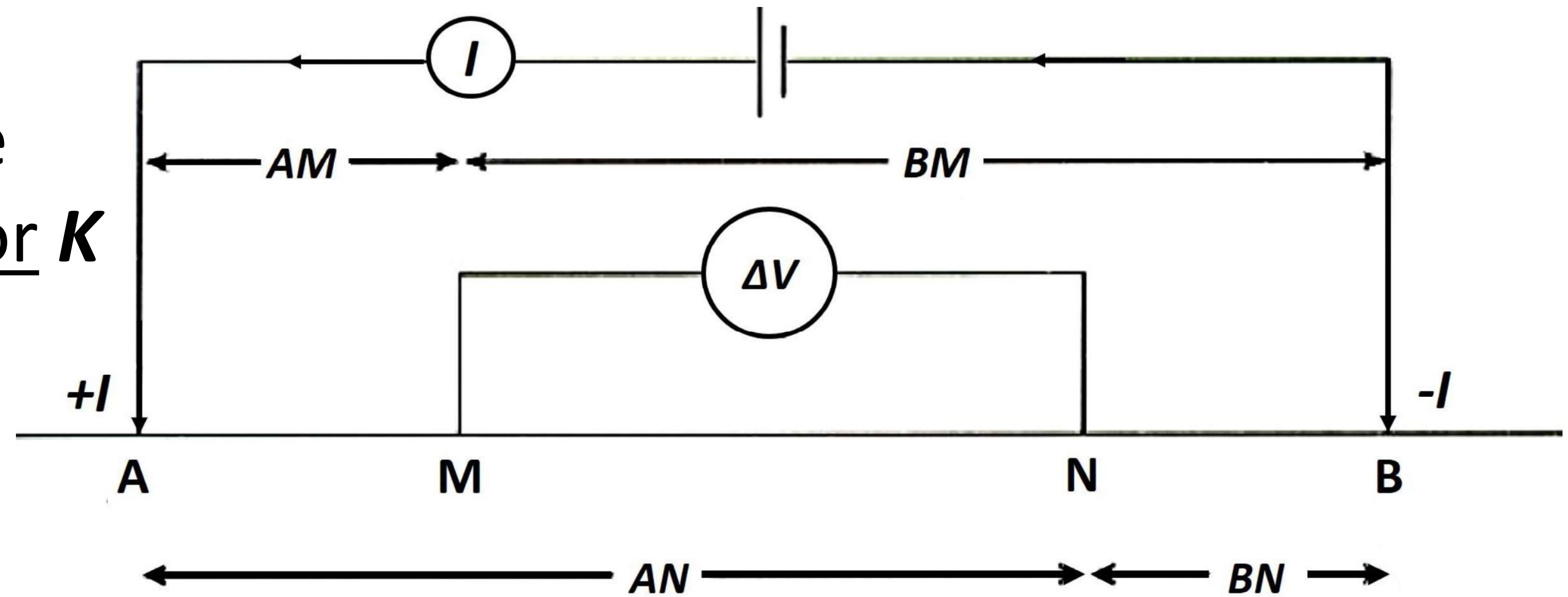


Fig. 3a & b: Four electrodes layout of current (A & B) and potential (M & N) electrodes with flow lines in hemi-sphere media.

3- Resistivity of four electrode array and the Geometric factor *K*



As shown in Figure (4), the four electrodes setup is comprise from two current electrode, a point source and the reversion sink current point. And another two potential electrodes to measure the potential deference. So, $V_M = (\rho I/2\pi)[\frac{1}{AM} - \frac{1}{BM}], V_N = (\rho I/2\pi)[\frac{1}{AN} - \frac{1}{BN}]$ ……***9***

$$\triangle V_{MN} = V_M - V_N = \frac{I\rho}{2\pi}\{[\frac{1}{AM} - \frac{1}{BM}] - [\frac{1}{AN} - \frac{1}{BN}]\} \quad \text{......}\mathbf{10}$$

Rearranging Eq. (**9**) in terms of resistivity $\boldsymbol{\rho_a}$ yields

$$\rho = \frac{\triangle V_{MN}}{I} \frac{2\pi}{[\frac{1}{AM} - \frac{1}{BM}] - [\frac{1}{AN} - \frac{1}{BN}]} \quad \text{......}\mathbf{11}$$

So, the geometric factor ***K*** in this case is:

$$K = \frac{2\pi}{[\frac{1}{AM} - \frac{1}{BM}] - [\frac{1}{AN} - \frac{1}{BN}]} \quad \text{......}\mathbf{12}$$

4- Apparent Resistivity:

The measured true resistivity value from equation (11) is originated under hypothesis that the earth has homogeneous resistivity **ρ**. In fact, the earth is an **inhomogeneous** medium and it has heterogeneous resistivity distribution inside e. g., various lithologies and geological features. In other words, the all field resistivity measurements are for apparent resistivity, while the true resistivities can be acquired from the inversion and interpretation programs (Reynolds, 2011). Actually, the relation between the "true" resistivity and the "apparent" resistivity is a complicate relationship (Loke, 2016). This leads to re-arrange **Eq. (11)** in terms of apparent resistivity ρ_a which is explained to be the resistivity that would have been calculated if the ground were indeed uniform and isotropic (Everett, 2013).

$$\rho_a = \left[\frac{\Delta V_{MN}}{I}\right] K = KR \quad ....12$$

From eq. **(12)** we can notice that the apparent resistivity ρ_a is mainly depends on ground resistance $R = \Delta V/I$ and a geometric factor K that relies simply on the configuration (arrangement) of the current and potential electrodes. This equation is a **fundamental** equation in the **direct - current (dc)** method for **electrical prospecting**.

5- Arrays of the Electrodes (Electrodes Configuration)

In general, four electrodes (in arbitrary or collinear layout) are employed for each individual resistivity measurement. The term of electrode arrays is related to the specific arrangement of the current and potential dipoles in electrical resistivity surveying (Xu, 1994). More than 90 surface electrode arrays have been proposed to date (Reynolds, 2011). However, the most commonly employed arrays are the **Wenner** (alpha), Wenner (beta), Wenner (gamma), **dipole-dipole**, **Wenner-Schlumberger**, **pole-dipole**, **equatorial dipole-dipole** and **pole-pole arrays**.

The conventional arrays with their corresponding geometry factor are shown in **Fig. 5**. Each electrode-array type has specific advantages and disadvantages (Loke, 2020).

The selection of a specific array depends on a number of parameters, e.g**., site condition & ease of deployment, subsurface heterogeneities, geological feature to be delineated, depth of the target, sensitivity of the resistivity meter & the background noise level.** Furthermore, the **sensitivity of** a particular **array** to **vertical & lateral variations** in subsurface, the **penetration depth**, the **signal-noise ratio**, and the **horizontal data coverage** are factors that determine the **choice of the configuration.**

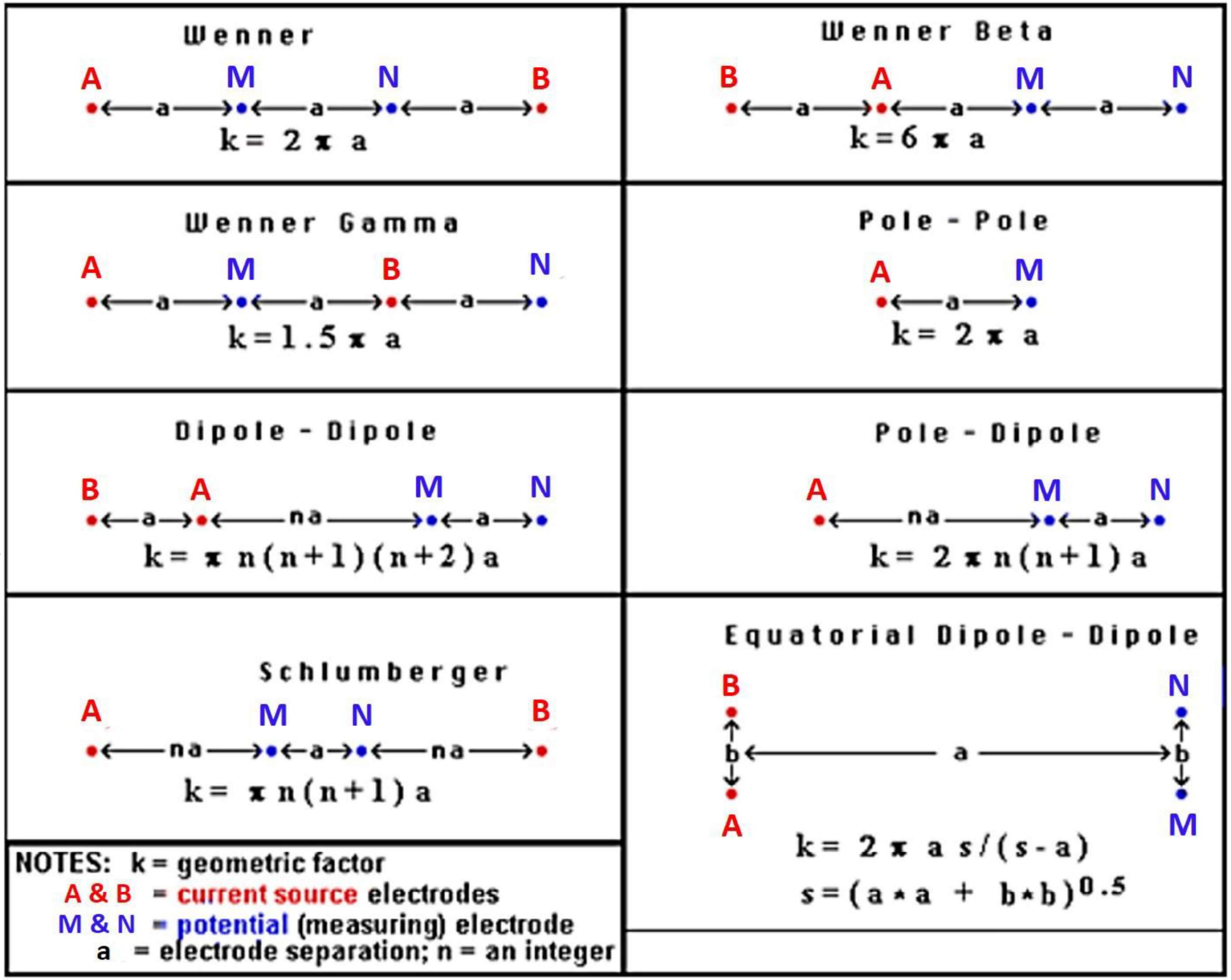


Figure 5: The most common electrode arrays utilized in resistivity prospections with their corresponding geometric factor. Note that the **Wenner-Schlumberger, dipole-dipole** and **pole-dipole** arrays have **two parameters**, **the dipole length a** and **the dipole separation factor n**. Commonly integer values are used for **n**.

5.1- Common Array-types

We will focus on some array-types which are commonly used in **DC resistivity** surveying

Wenner-array : This array consists of four electrodes (A, B, M, N) placed at the - surface of the ground along a straight line. Where (A, B) are current electrodes, and (M, N) are potential electrodes, fig. (1-2) .The electrode distance (MN) is equal to electrode spacing (a) or
. AB). These distances used as a faction depth of (**ρ**a) measurements 1/3)
.(The (**ρ**a) values for this array can be calculated using equation (12 $\boxed{\rho a = 2\pi a\ \Delta V/I}$

Schlumberger Array: In the electrical surveys, that is the Schlumberger array is the most widely used for deep sounding, especially in investigation of groundwater aquifers. Four electrodes (AM & NB) are placed along a straight line on the earth surface in the same direction, as shown in figure (8), when taking measurements the potential electrodes remain fixed, while the current electrodes expanding to various .distances

$$\rho_a = K\frac{\Delta V}{I}$$

Magnetic method

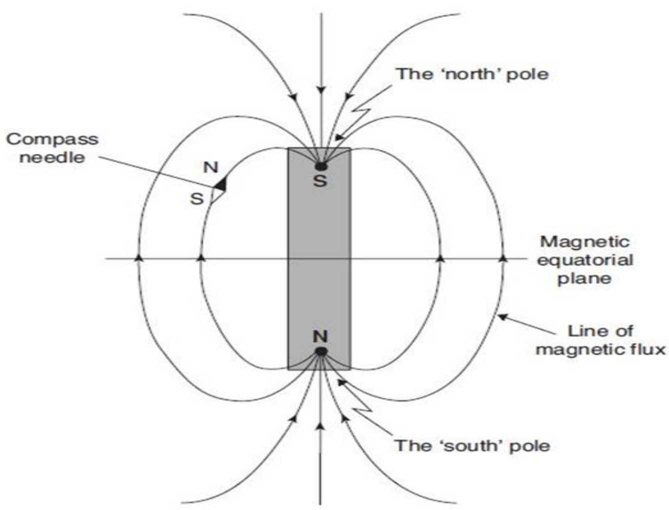
1-1 Introduction

In 1915, Adolf Schmidt made a balance magnetometer that enabled more wide spread magnetic surveys to be undertaken. As with many geophysical methods, advances in technology were made during the Second World War which enabled more efficient, reliable and accurate measurements to be made thereafter. In the 1960s, optical absorption magnetometers were developed which provided the means for extremely rapid magnetic measurements with very high sensitivity, ideally suited to airborne magnetic surveys. Since the early 1970s,magnetic gradiometers have been used which measure not only the total Earth's magnetic field intensity, but also the magnetic gradient between sensors. This provides extra information of sufficient resolution to be invaluable in delimiting geological targets.

Geomagnetic methods can be used in a wide variety of applications (Table 3.1) and range from small-scale investigations to locate pipes and cables in the very near surface, and engineering site investigations, through to large-scale regional geological mapping to determine gross structure, such as in hydrocarbon exploration. Commonly in the larger exploration investigations, both magnetic and gravity methods are used to complement each other. Used together prior to seismic surveys, they can provide more information about the subsurface, particularly the basement rocks, than either technique on its own.

1-2 Basic theory and units

Around a bar magnet, a magnetic flux exists, as indicated by the flux lines in Figure 3.1, and converges near the ends of the magnet, which are known as the magnetic poles. If such a bar magnet is suspended in free air, the magnet will align itself with the Earth's magnetic field with one pole (the positive north-seeking) pointing towards the Earth's north pole and the other (the negative south seeking) towards the south magnetic pole. Magnetic poles always exist in pairs of opposite sense to form a dipole. When one pole is sufficiently far removed from the other so that it no longer affects it, the single pole is referred to as a monopole.



If two magnetic poles of strength m_1 and m_2 are separated by a distance r, a force exists between them (Box 3.1). If the poles are of the same sort, the force will push the poles apart, and if they are of opposite polarity, the force is attractive and will draw the poles towards each other .Note the similarity of the form of the expression in Box 3.1 with that for the force of gravitational attraction in Box 2.1; both gravity and magnetism are *potential fields* and can be described by comparable potential field theory.

$$F = \frac{m_1 m_2}{4\pi\mu r^2}$$

where μ is the magnetic permeability of the medium separating the poles; m_1 *and* m_2 are pole strengths and r the distance between them.

The closeness of the flux lines shown in Figure 3.1, the flux per unit area, is the flux density B (and is measured in weber/m^2 = teslas). B, which is also called the 'magnetic induction', is a vector quantity (the former c.g.s. units of flux density were gauss, equivalent to 10^{-4} T.) The units of teslas are too large to be practical in geophysical work, so a subunit called the nanotesla (nT = 10^{-9} T) is used instead, where 1 nT is numerically equivalent to 1 gamma in c.g.s. units (1 nT is equivalent to 10^{-5} gauss). The magnetic field can also be defined in terms of a force field which is produced by electric currents. This magnetizing field strength H is defined, following Biot-Savart's Law, as being the field strength at the center of a loop of wire of radius r through which a current is flowing such that H = 1/2r. Consequently the units of the magnetizing field strength H are amperes per meter (A/m).

The ratio of the flux density B to the magnetizing field strength is a constant called the absolute magnetic permeability (μ). Practically, the magnetic permeability of water and air can be taken to be equal to the magnetic permeability of free space (a vacuum), denoted μ_0 which has the value $4\pi \times 10^{-7}$ Wb A^{-1} m^{-1}. For any medium other than a vacuum, the ratio of the permeabilities of a medium to that of free space is equal to the relative permeability μ_r, such that $\mu_r = \mu/\mu_0$ and, as it is a ratio, it has no units.

1-3 Magnetic susceptibility

It is possible to express the relationship between B and H in terms of a geologically diagnostic parameter, the magnetic susceptibility κ Susceptibility is in essence a measure of how susceptible a material is to becoming magnetized. For a vacuum, $\mu_r = 1$ and $\kappa = 0$. Although susceptibility has no units, to rationalize its numerical value to be compatible with the SI or rationalized system of units, the value in c.g.s. equivalent units (e.g. Un rationalized units such as e.m.u., electromagnetic units) should be multiplied by 4π. Some materials have negative susceptibilities

1-4 Induced and remnant magnetization

So far the discussion has centered upon a magnetization that is induced by an applied field H where the induced intensity of magnetization is denoted by J_i. In many cases, in the absence of an applied field (H), there is still measurable intensity of magnetization which is sustained by the internal field strength due to permanently magnetic particles. The intensity of this permanent or remnant magnetization is denoted by J_r.

A rock mass containing magnetic minerals will have an induced as well as a remnant magnetization. These magnetizations may have different directions and magnitudes of intensity (Figure 3.3). The magnitude and orientation of the resultant J dictate both the amplitude and shape of a magnetic anomaly, respectively. Consequently, interpretation of magnetic data is complicated by having greater degrees of freedom of the magnetic parameters and physical properties compared with gravity, which is largely dependent upon only rock density.

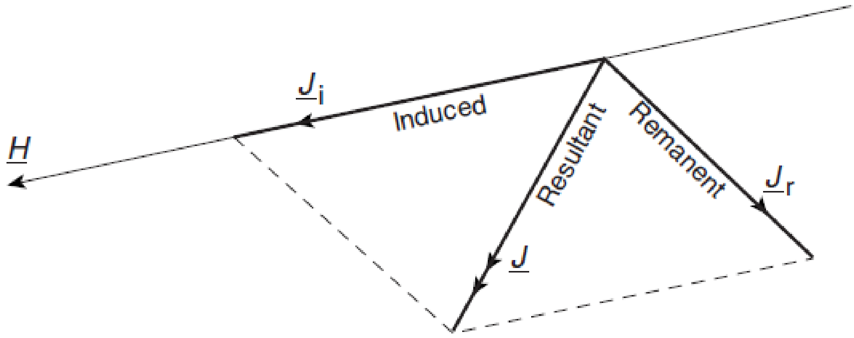


Figure 3.3 Vectorial summation of induced and remanent intensities of magnetisation.

1-5 Susceptibility of rocks and minerals

Magnetic susceptibility is an extremely important property of rocks, and is to magnetic exploration methods what density is to gravity surveys. Rocks that have a significant concentration of Ferro- and/or ferromagnetic minerals tend to have the highest susceptibilities. Consequently, basic and ultrabasic rocks have the highest susceptibilities, acidic igneous and metamorphic rocks have intermediate to low values, and sedimentary rocks have very low susceptibilities in general ,In this compilation of data,

Mineral or rock type	Susceptibility*
Sedimentary	
Dolomite (pure)	−12.5 to +44
Dolomite (impure)	20,000
Limestone	10 to 25,000
Sandstone	0 to 21,000
Shales	60 to 18,600
Average for various	0 to 360
Metamorphic	
Schist	315 to 3000
Slate	0 to 38,000
Gneiss	125 to 25,000
Serpentinite	3100 to 75,000
Average for various	0 to 73,000
Igneous	
Granite	10 to 65
Granite (m)	20 to 50,000
Rhyolite	250 to 37,700
Pegmatite	3000 to 75,000
Gabbro	800 to 76,000
Basalts	500 to 182,000
Oceanic basalts	300 to 36,000
Peridotite	95,500 to 196,000
Average for acid igneous	40 to 82,000
Average for basic igneous	550 to 122,000

1-6 Magnetic instruments

Magnetometers used specifically in geophysical exploration can be classified into three groups: the torsion (and balance), fluxgate and resonance types, of which the last two have now completely superseded the first. Torsion magnetometers are still in use at 75% of geomagnetic observatories, particularly for the measurement of declination.

1-6-1 Torsion and balance magnetometers

Historically the first to be devised (1640), these comprise in essence a magnetic needle suspended on a wire (torsion type) or balanced on a pivot. In the Earth's magnetic field the magnet adopts an equilibrium position. If the device is taken to another location where the Earth's magnetic field is different from that at the base station, or if the magnetic field changes at the base station, the magnet will align itself to the new field and the deflection from the rest position is taken as a measure of the Earth's magnetic field. A development of the Schmidt variometer was the compensation variometer. This measured the force required to restore the beam to the rest position. In exploration work, the greatest precision with a balance magnetometer was only 10 nT at best.

1-6-2 Fluxgate magnetometers

The fluxgate magnetometer was developed during the Second World War to detect submarines. It consists of two parallel cores made out of high-permeability ferromagnetic material. Primary coils are wound around these cores in series but in opposite directions (Figure 3.18). Secondary coils are also wound around the cores but in the opposite sense to the respective primary coil. The fluxgate magnetometer can be used to measure specific magnetic components with the same attitude as the sensor cores.

As the fluxgate magnetometer is relatively insensitive to magnetic field gradients, it has the advantage that it can be used in areas where very steep gradients would militate against the use of resonance-type devices, which are affected. Some portable fluxgate magnetometers suffer from temperature effects owing to inadequate thermal insulation, which can reduce the resolution to only ±10 to 20 nT, this being inadequate for ground exploration surveys.

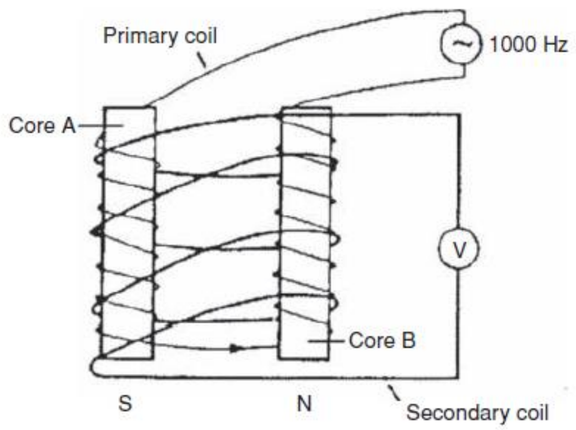


Figure 3.18 Basic operating principle of the fluxgate magnetometer.

1-6-3 Resonance magnetometers

There are two main types of resonance magnetometer: the proton free-precession magnetometer, which is the best known, and the alkali vapour magnetometer. Both types monitor the precession of atomic particles in an ambient magnetic field to provide an absolute measure of the total magnetic field, F.

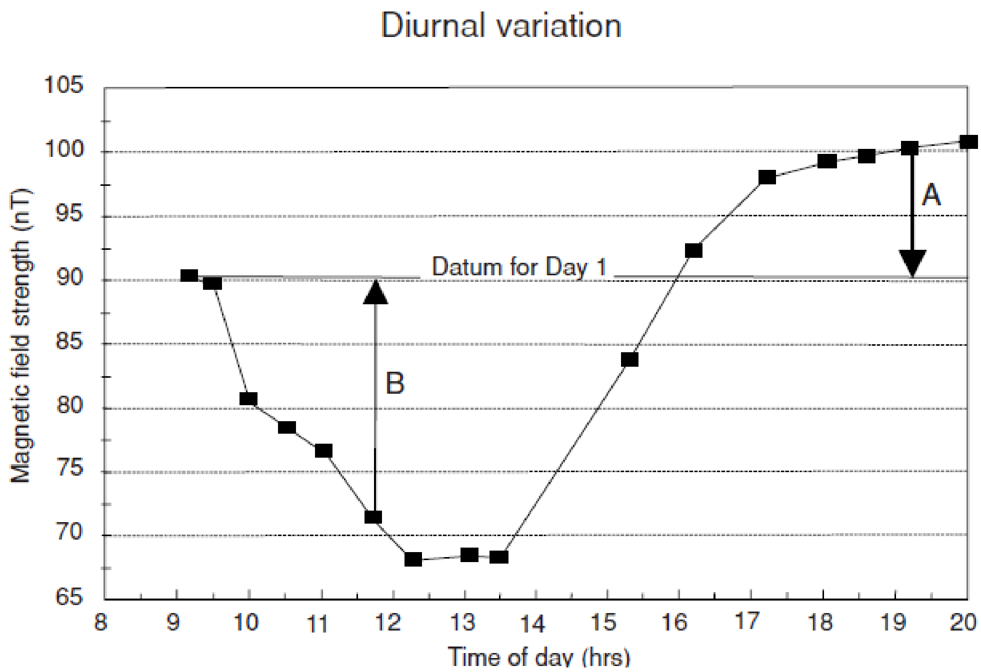
1-7 Airborne magnetometer systems

Magnetic surveys are commonly acquired using aircraft, either fixed wing or helicopters. Magnetometers can be fitted to aircraft either as single or multiple sensors depending upon the type of measurements to be made. The sensors can be rigid mounted as booms (from the front of the aircraft nose) or stingers (from the aircraft's tail) (Figure 3.24A) or at the end of the aircraft's wing tips (especially useful in horizontal gradiometer mode to provide as large a horizontal separation of the sensors as possible). Where the aircraft to be used is not specifically adapted for magnetic surveying, temporary mounts can be added or the sensors can be deployed from helicopters using a towed 'bird'

1-8 Magnetic corrections

All magnetic datasets contain elements of noise and will require some form of correction to the raw data to remove all contributions to the observed magnetic field other than those caused by subsurface magnetic sources.

1-The most significant correction is for the diurnal variation in the Earth's magnetic field. Base station readings taken over the period of a survey facilitate the compilation of the diurnal 'drift' as illustrated in Figure 3.25. Measurements of the total field made at other stations can easily be adjusted by the variation in the diurnal curve. For example, at point A in Figure 3.25, the ambient field has increased by 10 nT and thus the value measured at A should be reduced by 10 nT. Similarly, at B, the ambient field has fallen by 19 nT and so the value at B should be increased by 19 nT.



2-Rarely, a terrain correction may need to be applied when the ground over which a survey is conducted is both magnetic and topographically rough. Unlike the gravity case where terrain corrections, although laborious, are relatively easy to calculate, corrections for terrain for magnetic data are extremely complex. If the rough terrain is made up largely of low-susceptibility sedimentary rocks, there will be little or no distortion of the Earth's magnetic field. However, if the rocks have a significant susceptibility, a terrain factor may have to be applied. Anomalous readings as large as 700 nT have been reported by Gupta and Fitzpatrick (1971) for a 10mhigh ridge of material with susceptibility $\kappa \approx 0.01$ (SI).

3-Another way of correcting for the effects of topography, or of reducing the data to a different reference plane, is by upward continuation. This permits data acquired at a lower level (e.g. on the ground) to be processed so that they can be compared with airborne surveys. The effect of this is to lessen the effects of short-wavelength high-amplitude features, as the magnetic force is indirectly proportional to the square of the distance between source and sensor

4-Survey data at any given location can be corrected by subtracting the theoretical field value Fth, obtained from the International Geomagnetic Reference Field, from the measured value, Fobs This works well in areas where the IGRF is tied-in at or near to geomagnetic observatories, but in many places the IGRF is too crude. Instead, it is better to use a local correction, which can be considered to vary linearly over the magnetic survey area.

5- Another method of calculating the anomalous field δF is to determine statistically the trend of a regional field to isolate the higher frequency anomalies, which are then residualised in the same way that gravity residuals are calculated. The regional field is subtracted from the observed field to produce a residual field