Chapter 11*

GEOPHYSICS

11.1 INTRODUCTION

This Chapter is intended to be a guide to the applications of seismic refraction and resistivity on engineering sites. Many civil engineers and geologists have some acquaintance with these basic geophysical tools, but few apply them frequently. The primary purpose of this Chapter is to provide the reader with a working knowledge of the methods and a basis to judge the applicability of these methods and the results to his particular exploration problem. In common with other indirect methods of sub-surface exploration, there are no rigid and inflexible approaches to making sense of the data nor are there any handbooks that infalliably lead the engineer, geologist, or geophysicist to the correct answer. The general case will require thought and care, ambiguities and uncertainties are not uncommon. Some foreknowledge of the site conditions and an understanding of what is geologically plausible will always assist in resolving the raw data into meaningful information.

Two methods are based on the measurement of different physical properties of the sub-surface material, i.e., the **resistivity method** and the **seismic refraction method.** The resistivity method shows up variations in the resistivity of each material and this is largely dependent upon the amount of salinity in the water contained in the material. This method is capable of detecting water tables, of distinguishing between porous and non-porous materials, or of distinguishing fresh from salty water. The seismic refraction method displays the variations in **seismic velocity** of earth materials, and these are largely dependent upon hardness and degree of consolidation. With this method it is possible to distinguish between rock and soil, between consolidated and unconsolidated materials, between compacted soils and loose, and between fresh rock and weathered rock.

For example, both dry sand and bedrock show high resistivity but the seismic velocity of bedrock is usually considerably greater than that of sand. A resistivity survey could therefore be used to outline the limits of sand deposit and a few seismic lines could separate sand from bedrock. As another example, badly weathered bedrock may show the same seismic velocity as a gravel lens. In this case, a distinction can be made between them with the help of their differences in resistivity.

Exploratory drilling is almost always done in a site investigation; its value, in terms of the quality of information gathered, will be enhanced if geophysical surveys are carried out first and the results used to guide the drilling operations. To this end, results obtained from shallow investigations are useful because they provide rapidly a picture of the underlayers. Seismic refraction and resistivity methods offer rapid, inexpensive, and accurate methods of sub-surface exploration. Their application to site investigation should be routine rather than the exception. When seismic refraction and resistivity methods are used widely, and particularly when allied with the exploratory drill, it will invariably speed the recovery of sub-surface information and reduce site investigation costs.

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* Tables and Figures without credit lines, included in this chapter, have been compiled by the author from comparative sources for this Handbook.

11.2 SEISMIC REFRACTION METHOD

11.2.1 *Uses of Seismic Refraction*

Fundamentally the seismic method relies on the fact that earth materials are commonly sub-horizontally layered and that strength normally increases with depth. The most common applications for seismic engineering surveys are:

- depth of bedrock,
- shape of bedrock surface,
- depth of water table,
- fault location,
- rippability assessment,
- blasting assessment,
- sand and gravel assessment,
- cavity detection, and
- determination of dynamic elastic constants,

11.2.2 *Definitions*

Elastic Constants

The elastic properties of substances are characterized by elastic moduli that specify the relation between 'stress' and 'strain'. The two moduli of immediate interest for the study of elastic waves are:

K = bulk modulus = **incompressibility**, and
μ = shear modulus = **rigidity**

Elastic Waves

In an elastic isotropic medium two kinds of elastic waves propagate:

P. Waves: primary waves, compressional waves, and longitudinal waves. The motion of the medium is in the same direction as the direction of wave propagation (Fig. 11.1). Their velocity is given by

$$V_p = \sqrt{\frac{K + 4/3\mu}{d}}$$

Where,

d is the density of the medium.

Fig. 11.1 P Waves

S. Waves: Secondary waves, shear waves, and transverse waves. The particles of the medium move at right angles to the direction of wave propagation (Fig. 11.2). Their velocity is given by:

$$V_s = \sqrt{\frac{\mu}{d}} \ .$$

Shear waves do not propagate through liquids and gases. V_p is greater than V_s.

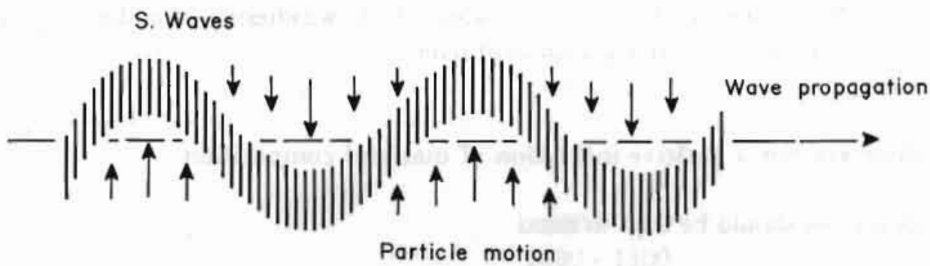


Fig. 11.2 S. waves

Velocities in Rocks

Generally only P waves are of importance as an exploration tool. Typical values for the velocity of P waves in some rocks are given in Table 11.1.

Effect of Porosity and Saturation

The velocities generally increase as porosity decreases. This is expressed by the law:

$$\frac{1}{V_r} = \frac{\phi_p}{V_f} + \frac{[1-\phi_p]}{V_{ma}}$$

where, V_r = rock velocity,
V_f = fluid velocity, and
V_{ma} = matrix velocity.

The velocities also increase, for a given porosity, as saturation increases. (V_{air} = 330 m/s, V_{water} = 1500 m/s). Table 11.2 provides velocities of propagation of seismic waves in some sub-surface materials.

Effect of Depth and Age

The velocities increase with depth of burial and geologic ages. Faust (1951), finds that for shale and sand:

V_r = $46.5\ Z^{1/6}\ T^{1/6}$ m/s,
Z = depth in metres, and
T = age in years.

Before attempting to estimate the composition from the computed velocities, the seismic analyst should familiarize himself with the general nature of the terrain under study. He should know :

- where the water table is,
- whether the overburden is a product of the weathering of underlying bedrock, and
- whether it is mostly glacial drift over.

Velocities alone are not a positive indication of material composition

Several general rules should be kept in mind

I - velocity is roughly proportional to the degree of consolidation of rock or soil.
II - in unconsolidated materials, velocity increases with water content.
III - weathering of a rock will greatly reduce its velocity.
IV - a particular rock type will include a range of velocities and these ranges **may overlap for different rock types**.
V - correlation of velocity with the type of earth material, to a great extent, will depend upon the overall geological characteristics of the area under study.
VI - velocity measurements are very sensitive to the dip of the interface. If high precision measurement velocities are required, always assume that a dip exists and follow the procedure for dipping discontinuity.

Seismic refraction techniques are generally used for site investigation for civil engineering purposes and reflection is a powerful tool in petroleum exploration.

11.2.3 *Data Acquisition*

The aim of seismic exploration is to determine the depth and characteristics of the near surface layers. There are two methods of seismic exploration; **reflection** and **refraction.** The two methods basically involve the production of **energy** that is transmitted into the ground. After a time interval this energy, having been **reflected** and **refracted** from one or more sub-surface physical discontinuities, returns to **detectors** spread on the surface of the ground.

Seismic Sources

The standard method of producing seismic waves on land is to explode a dynamic charge. Individual explosions from the energy source are called **shots** and their locations are called **shot points.** A hammer striking a steep plate is generally used for shallow refraction measurements. Seismographs actually contain special circuitry that allows successive sets of waves from successive impacts to be added together. The purpose of summing is to strengthen weak signals and to increase the signal-to-noise ratio through cancelling random, background seismic noise. This process is called **stacking** because successive waves are stacked or added together.

The Detectors

The sensing devices used are called **geophones.** A geophone is a type of microphone (transducer) that converts seismic vibrations, or motions of the earth, to electric impulses (Fig. 11.3).

Table 11.1 Velocities of propagation of seismic waves in some subsurface materials

Materials	feet/sec.	metre/sec.
TOP SOILS:		
light and dry	600 - 900	200 - 300
moist	1000 - 1300	200 - 420
clayed	1300 - 2000	420 - 640
semi-consolidated sandy clay	1250 - 2150	400 - 700
wet loam	2500	800
Rubble or gravel (dry)	1970 - 2600	650 - 850
Cemented sand	2800 - 3200	900 - 1050
Cemented sandy clay	3800 - 4200	1200 - 1400
Water-saturated sand	4600	1500 -
Glacial till	5600 - 7400	1800 - 2400
Glacial moraine dry	2500 - 5000	800 - 1600
Glacial moraine saturated	5000 - 7000	1600 - 2250
Loose rock talus	1250 - 2500	400 - 800
WEATHERED AND FRACTURED ROCKS		
GRANITE		
- friable and highly decomposed	1540	500
- badly fractured and partly decomposed	2200	700
- softened and partly decomposed	10500	3300
- solid, monolithic	18500	6000
- badly broken and weathered	3000 - 8000	1000 - 2600
- little sign of weathering	10000 - 13000	3200 - 4200
entirely unweathered	16000 - 20000	5100 - 6400

Geophones used in land surveys are electromagnetic geophones. A coil is placed between the poles of a magnet. The coil acts as an internal element while the magnet moves with the earth. The relative motion of the coil and the magnet produces an electric impulse proportional to the velocity of the motion of the earth.

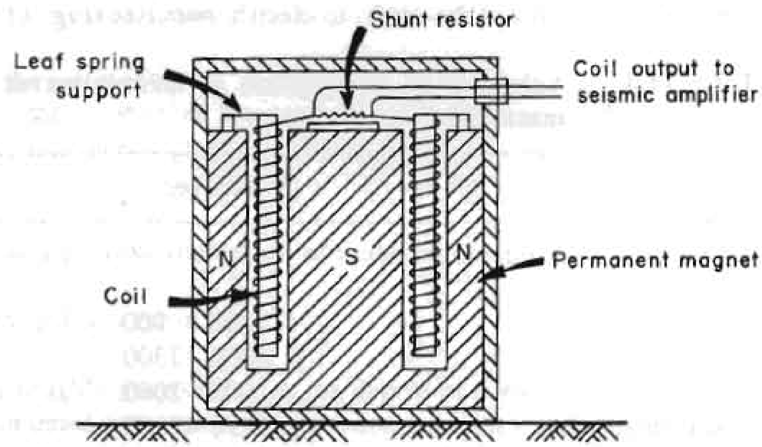


Fig. 11.3 Geophone

In most of the work, the geophones are placed along a straight line. The geophones are connected to the recording equipment by long cables.

Seismic Amplifier and Recorder

The geophone signals are transmitted by this cable to the **seismograph** where they are amplified and recorded by special equipment designed for this purpose. The record of the arriving pulse is called a **seismogram.** In addition to the simultaneous recording of the output signals, the recorder also marks the **zero time** instant of the explosion and a series of timing lines. Determining the time interval between the explosion and the arrival of the energy at the geophones, the raw data are transformed into a **graph** also called **dromochronique** (Figure 11.4). Generally, a large proportion of refraction surveys for engineering purposes are concerned with depths in the order of 10 to 30 metres. It is apparent that arrival times must be picked up with reasonable accuracy. This means that :

- the source energy,
- the amplifier gains, and
- the placement of the geophones,

are all important factors in obtaining the sharp breaks required for accurate timing of the first arrivals.

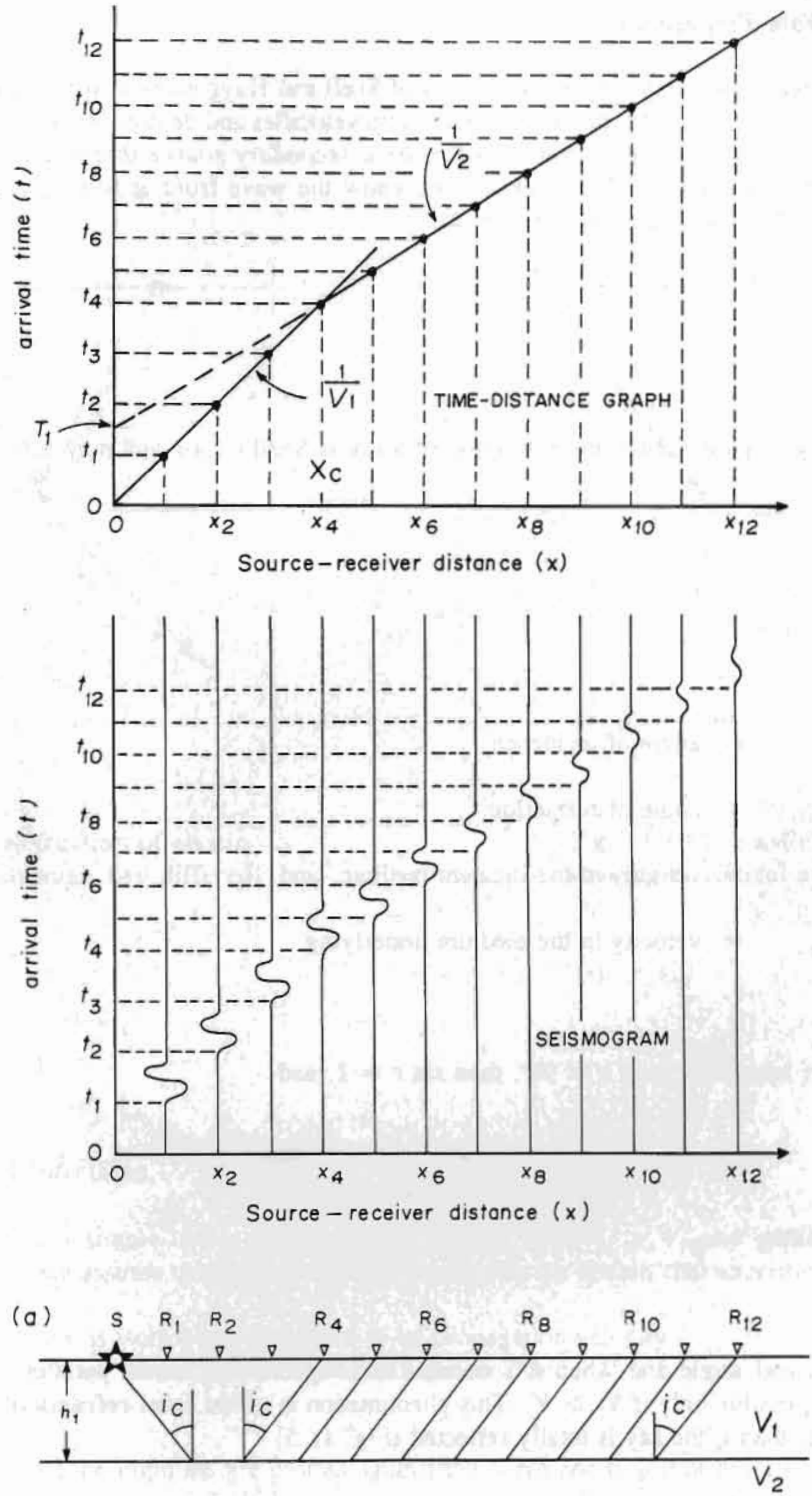


Fig. 11.4 Time-distance graph and seismogram

11.2.4 *Seismic Wave Propagation*

Light and seismic waves are very similar. The Laws of Snell and Huygens both apply making it possible to mathematically interpret travel times and distances into velocities and depths. Huygen's principle says: *Every point on a wave front may be considered to be a secondary source that emits waves travelling radially outward from the point"*. This means if we know the wave front at time t, we can define the position of the wavefront at a later instant $t + \Delta t$.

Snell's Law

The fundamental law that describes the refraction of a ray is Snell's Law and may be expressed by the equation:

$$\frac{\sin i}{\sin r} = \frac{V_1}{V_2}$$

where,

i = angle of incidence,

r = angle of refraction,

V_1 = velocity in the incident medium, and

V_2 = velocity in the medium underlying,

when i increases, r increases, until r = 90°, then sin r = 1, and

$$\sin i_c = \frac{V_1}{V_2} \; .$$

i_c is called the **critical angle** and when this occurs the refracted rays travel parallel to the interface. Obviously, this is possible only if $V_2 > V_1$. This phenomenon is called **total refraction**. If the angle of incidence is greater than i_c the ray is totally reflected (Fig. 11.5)

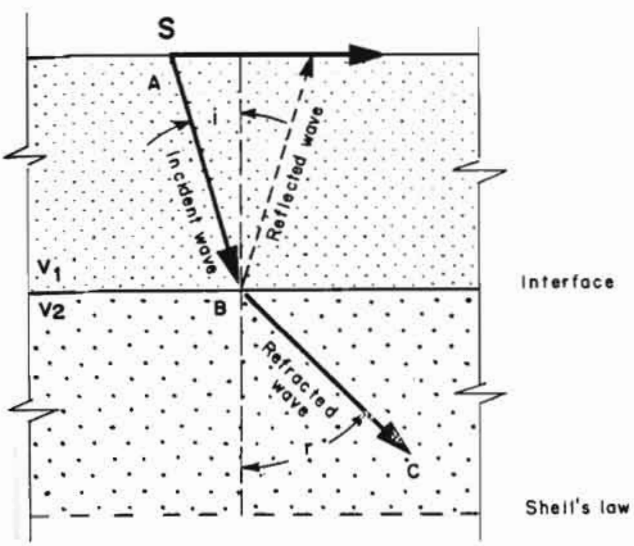


Fig. 11.5 (a) Refraction of elastic wave passing through two different velocity media

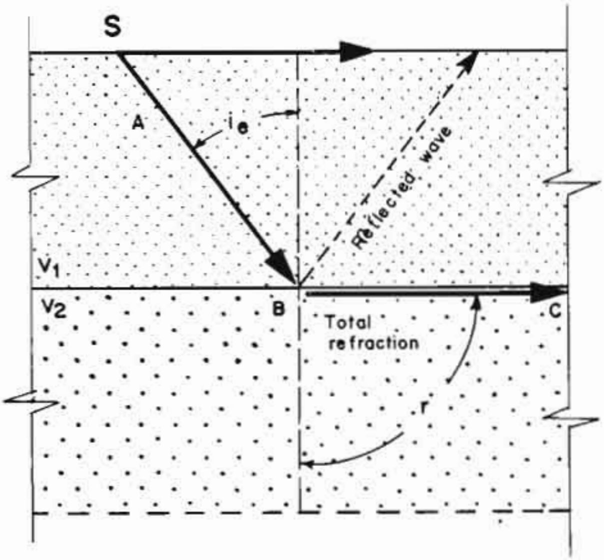


Fig. 11.5(b) Elastic wave striking at critical angle, i_e

$$\frac{\sin i}{V_1} = \frac{\sin r}{V_2}$$

$$\sin r = 1$$

$$\sin i_e = \frac{V_1}{V_2}$$

11.2.5 Parallel Interfaces

Let us begin with a simple case : two layers V_1 and V_2 with $V_2 > V_1$ and with plane and parallel boundary. Let us we assume that the sub-surface layers possess certain characteristics :

- each layer is isotropic with regard to its propagation velocity
- ray paths are made up of straight line segments, and
- $V_2 > V_1$.

The arrival times of the impulses are plotted against the corresponding shot to detector distances :

- the first arrival times are those of direct arrivals through the first layer, and
- if we draw a line through these points its slope will be $1/V_1$.

At some distance from the shot, a distance called **critical distance,** it takes less time for the energy to travel down to the top of the second layer, refract along the interface at the higher velocity V_2, and travel back to the surface, than it does for the energy to travel directly through the top layer. The energy that arrives at the detectors beyond the critical distance will plot along a line with a slope $1/V_2$ (Fig. 11.6).

Critical Distance, X_b

The critical distance is the distance from the shot to the point at which the refracted energy arrives at the same time as the direct one. Beyond this critical distance the refracted wave arrives first to the detector.

Intercept Time

The line through these refracted arrivals, $1/V_2$, will not pass through the origin but will intersect the time axis at a time called the **intercept time.** Because both the intercept time and the critical distance are directly dependent upon the velocities V_1 and V_2 and the thickness of the top layer, they can be used to determine the depth to the top of the second layer.

11.2.6 *Analysis of Time-distance Graphs*

a. *Parallel Interfaces*

i) Two Layers

Velocities V_1 and V_2 may be read directly. They are given by the slopes of two straight lines. Numerical values of V_1 and V_2 may be used to identify the material.

Depths:- if we wish to determine the depth to the single discontinuity that separates the top layer from the underlaying material, this may be computed from the critical distance Xb , using the formula:

$$D_1 = h_1 = \frac{X_b}{2} \cdot \sqrt{\frac{V_2-V_1}{V_2+V_1}} \; .$$

This may be computed also from the intercept time T_i using the formula :

$$D_1 = h_1 = \frac{T_i}{2} \cdot \frac{V_1 \cdot V_2}{\sqrt{V_2^2-V_1^2}} \; .$$

ii) Three Layers

The time distance graph will have three straight lines.

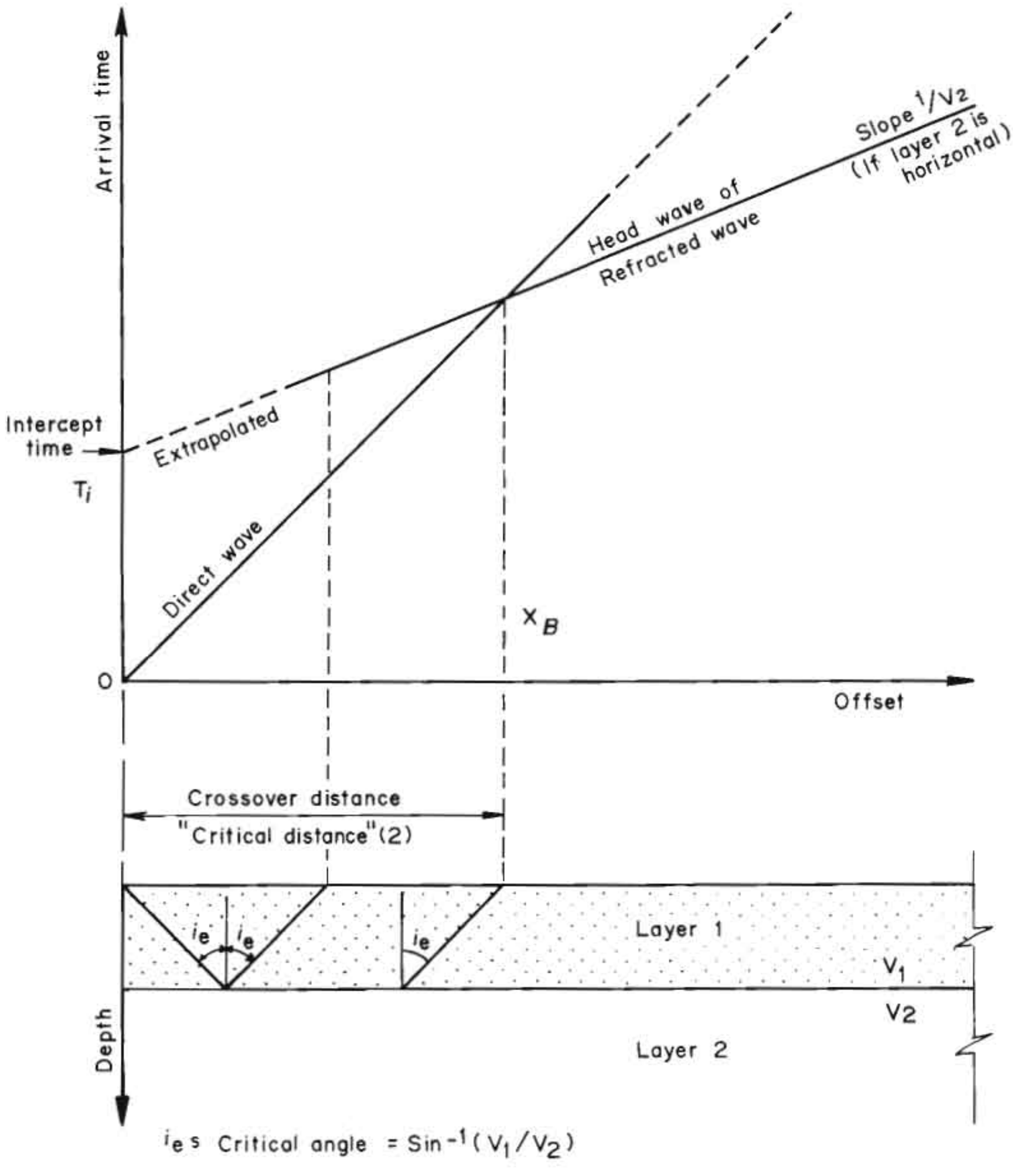


Fig. 11.6 Critical distance and intercept time for parallel interfaces

Velocities:- three straight lines on the seismic graph indicate three types of material. The top layer has a velocity V_1, the middle V_2, and the underlayer a velocity represented by slope V_3.

Depths:- the depth of the bottom of the first layer may be computed by the two expressions:

$$D_1 = h_1 = \frac{Xb_1}{2} \cdot \sqrt{\frac{V_2 - V_1}{V_2 + V_1}} \quad = \frac{Ti_1}{2} \cdot \frac{V_1 \cdot V_2}{\sqrt{V_2^2 - V_1^2}}$$

The depth of the bottom of the second layer, which has the velocity V_2, may be computed using:

$$D_2 = h_1 + h_2 = 0.8D_1 + \frac{Xb_2}{2} \cdot \sqrt{\frac{V_3 - V_2}{V_3 + V_2}}$$

or

$$D_2 = h_1 + h_2 = \frac{Ti_2}{2} \cdot \frac{V_3 \cdot V_2}{\sqrt{V_3^2 - V_2^2}} - 2D_1$$

Non-Parallel Interfaces (Fig. 11.7)

In this case a single traverse, in one direction only, cannot determine whether or not the rock surface is horizontal. An additional traverse must be run in the **reverse** direction.

By inspection of the time - distance graphs **direct** and **reverse**, we may draw the following conclusions:

$T_{AB} = T_{BA}$ - **these times must be equal**, if this is not the case, there is something wrong with the data.

X_b **down is < than X_b up,** a large difference means a large dip.
T_i **down is < than T_i up.**
V'_2 **apparent down is < than V_2.**
V'_2 **apparent up is > than V_2.**

For an approximate interpretation we compute:

$$V_2 = 2 \cdot \frac{V'_2\, down \cdot V'_2\, up}{V'_2\, down + V'_2\, up}$$

V_1 down and V_1 up must be the same:

$$D\ down = \frac{X_b\ down}{2} \cdot \frac{\sqrt{V_2 - V_1}}{V_2 + V_1}$$

$$D\ up = \frac{X_b\ up}{2} \cdot \frac{\sqrt{V_2 - V_1}}{V_2 + V_1} \ .$$

The approximate interpretation is usually sufficiently accurate, the error is less than 5 to 10 per cent.

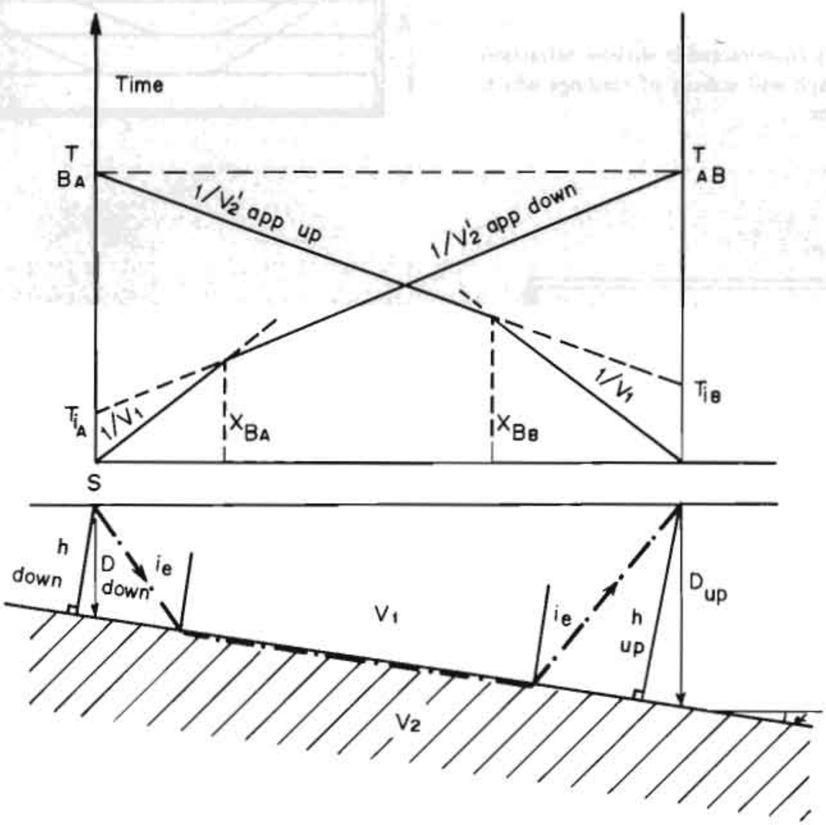


Fig. 11.7 Time-distance for non-parallel interfaces

11.2.7 **Examples of Geological Models Inferred from Seismic Refraction** (Fig. 11.8)

The topographic and layering conditions, in reality, are generally quite different from the idealized ones considered above.

Fig. 11.8(a) Multi-layered model

This model is very rarely encountered in shallow refraction survey. The seismic graph will consist of readings which lie on many straight lines.

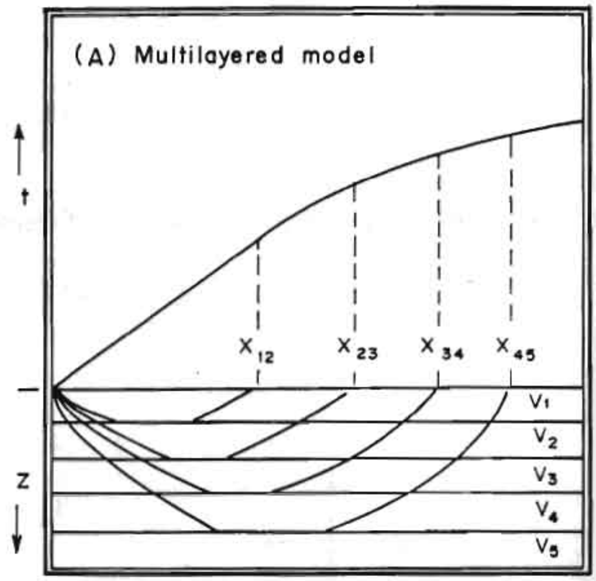


Fig.11.8(b) Continuous velocity increase

Probable interpretation is that the near-surface material is not completely uniform but becomes gradually harder with increasing depth. If the curvature is not great, draw an average straight line through it and treat it as uniform.

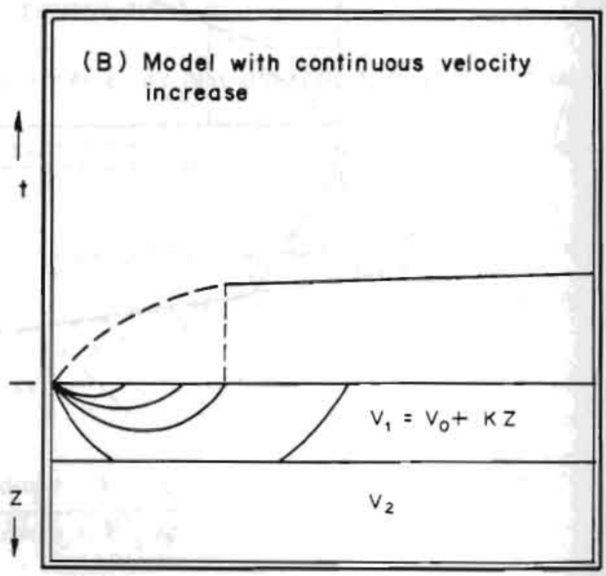


Fig. 11.8(c) Multi dipping layer model

The formulae needed for an exact interpretation are quite complicated because there are many unknown quantities to be determined.

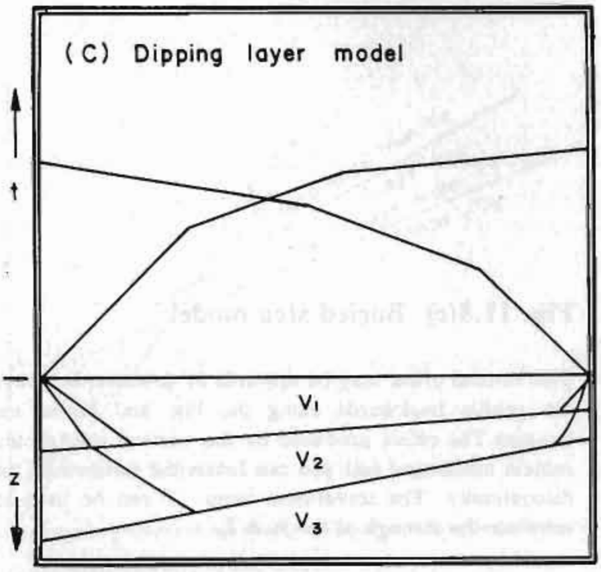


Fig. 11.8(d) Sloping surface model

If the lines run up or down the side of a hill, then the computed depths and dip will be in reference to the ground surface rather than the horizontal. Since the dip of the ground surface is easily measured, the correction can be applied after the interpretation has been completed.

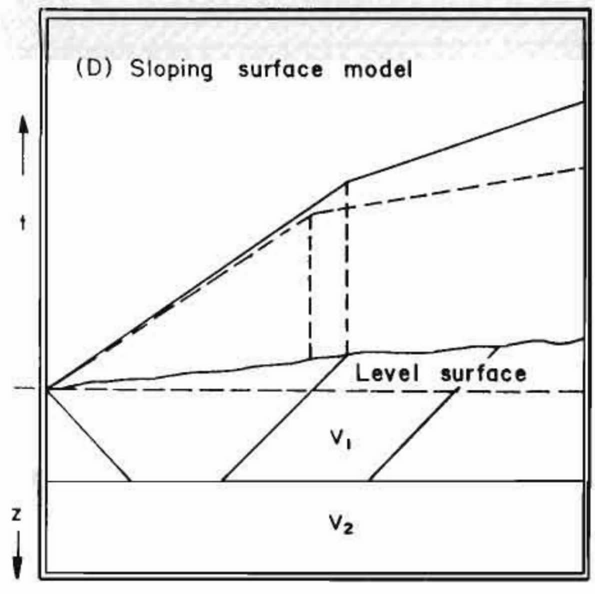


Fig. 11.8(e) Buried step model

The vertical offset may be upwards or downwards. Move the profile backwards along the line and repeat the traverse. The effect produced by the vertical step should remain unchanged and you can locate the position of the discontinuity. The travel-time jump *T* can be used to calculate the through of the fault Z_t.

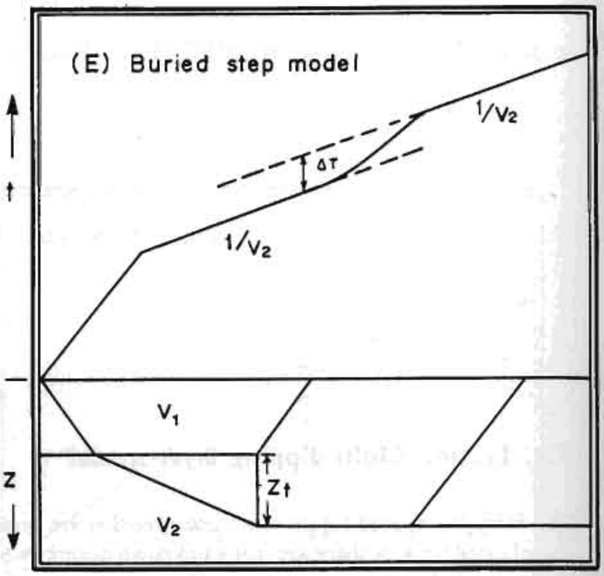


Fig. 11.8(f) Discordant body model

Generally the discordant body has a velocity, *V4*, greater than the others, so the seismic rays that cross it arrive sooner.

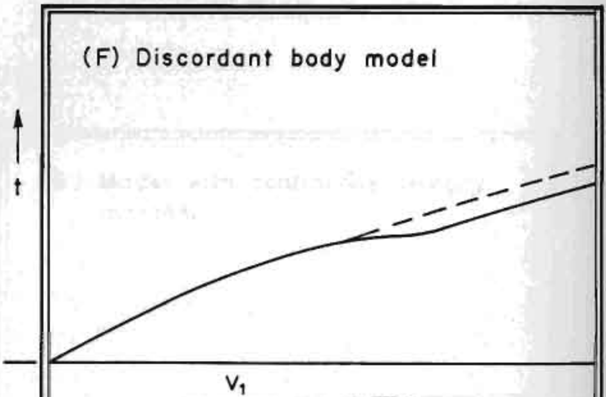
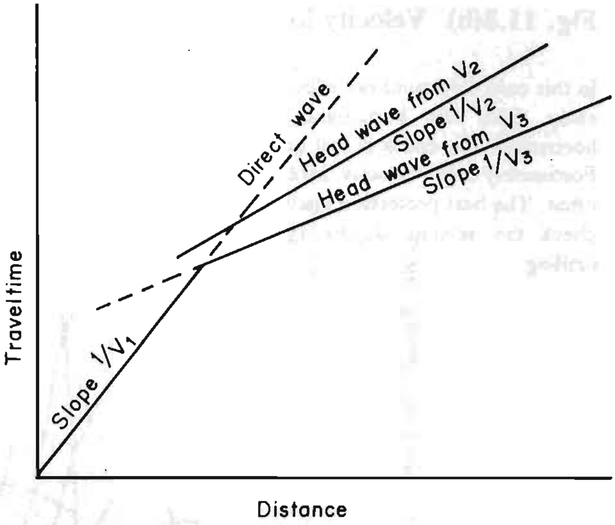


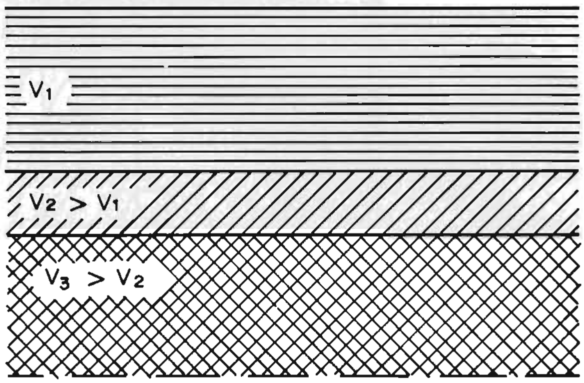
Fig. 11.8(g) Blind zone

The blind zone is one of the major problems in refraction studies. The blind zone refers to the portion of the underground that lies between the surface layer and a layer, *V3*, and is not represented in the graph of the 'first break'.

The existence of this blind zone leads to errors in the interpretation of the measurements. This volume of material is not suggested by the first breaks because of the **relative thickness** of the layers. The speed of the seismic wave in the blind zone is generally intermediate in value between that of the surface layer and the layer *V3* (Fig. 11.9).

In ordinary practice it is often represented by the zone of the water table above a high speed bedrock surface.

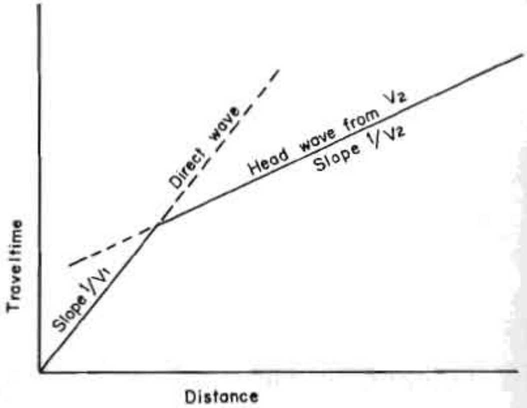




V_2 does not show as first arrival

Fig. 11.8(h) Velocity inversion

In this case the sound ray which enters is bent downwards, away from the horizontal, rather than towards the horizontal. The result is that no ray returns to the surface. Fortunately this 'velocity inversion' does not occur very often. The best protection against it, when suspected, is to check the seismic depths against known depths from drilling.



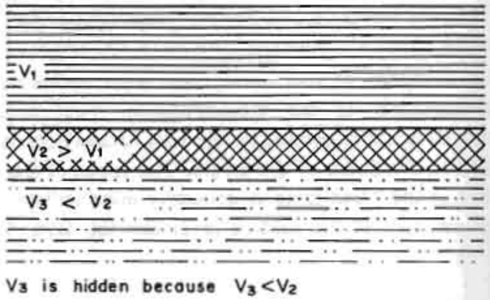
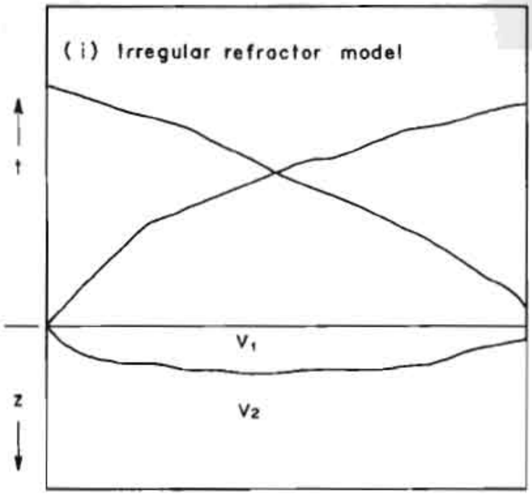


Fig. 11.8(i) Irregular refractor model

The readings will scatter about a straight line. We can interpret the straight line portion, this will give us the average depth along the line. Or we can use a more complicated method for mapping the bedrock surface by finding the depth under each geophone.



In every case, one should keep in mind that for a given graph there is a most probable interpretation but alternative interpretations are usually possible.

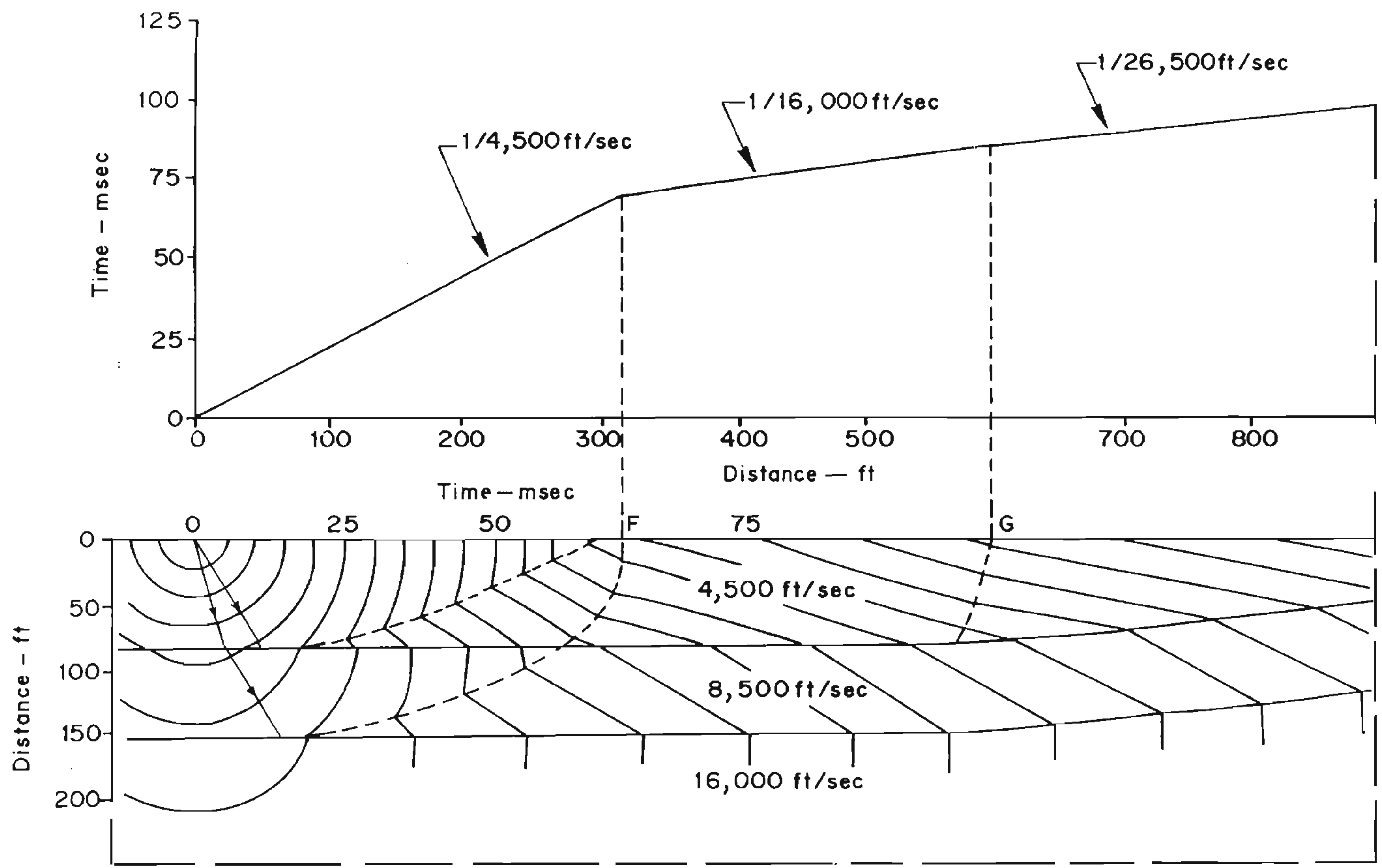


Fig. 11.9 Wave-front diagram and maximum 'undetectable' thickness of "blind zone"

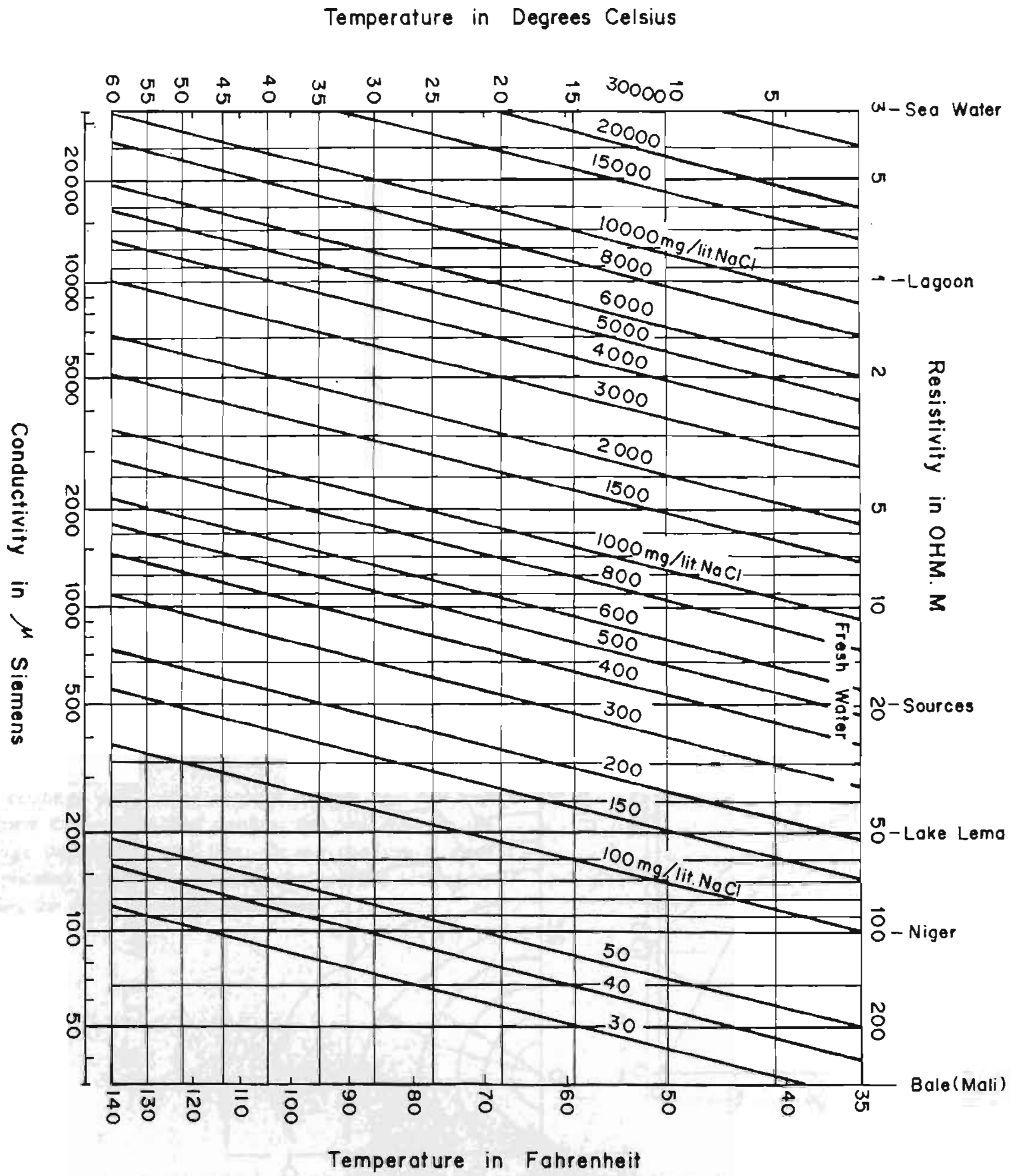


Fig. 11.10 Resistivity vs temperature and salinity

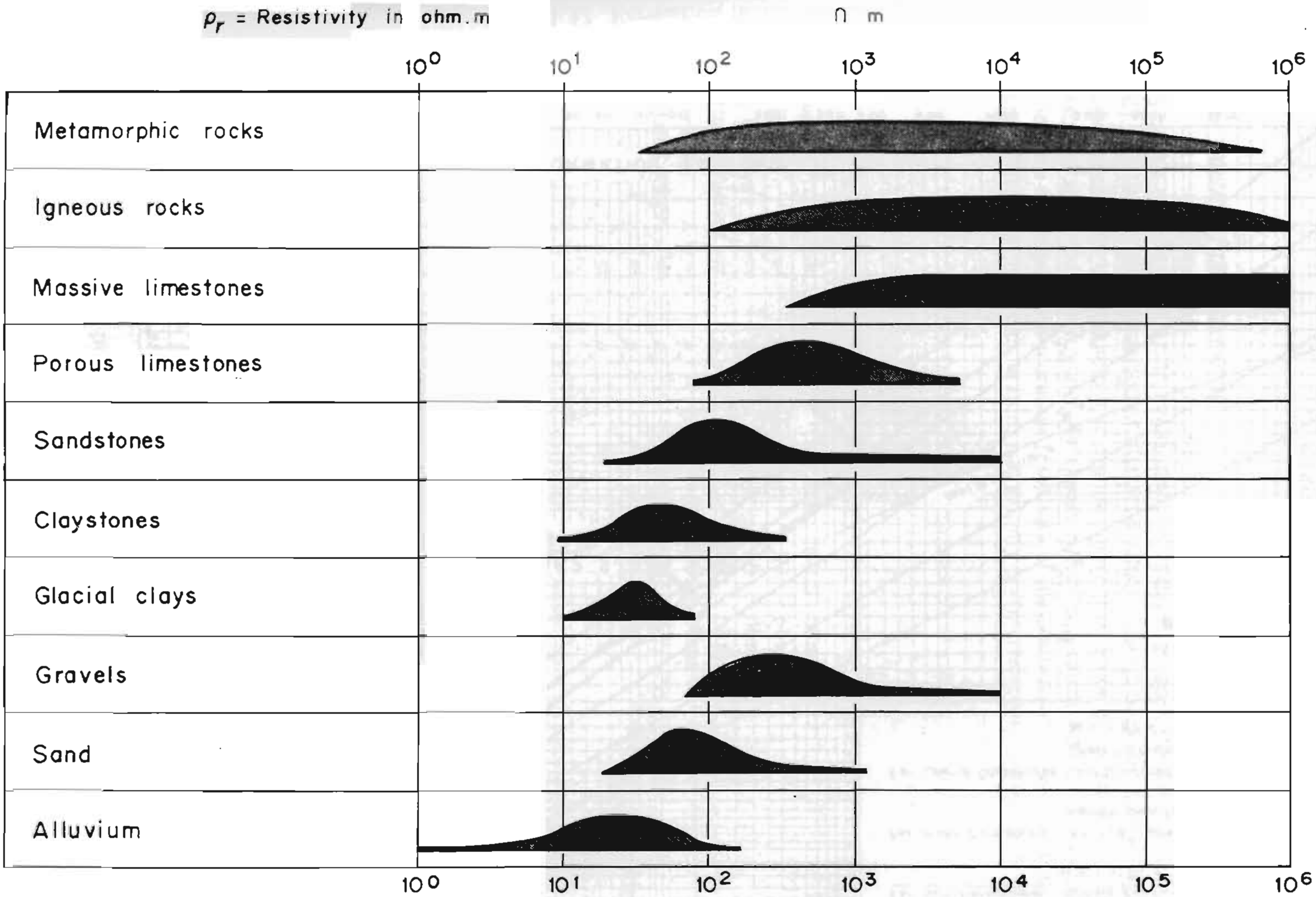


Fig. 11.11 Resistivity for different rock types

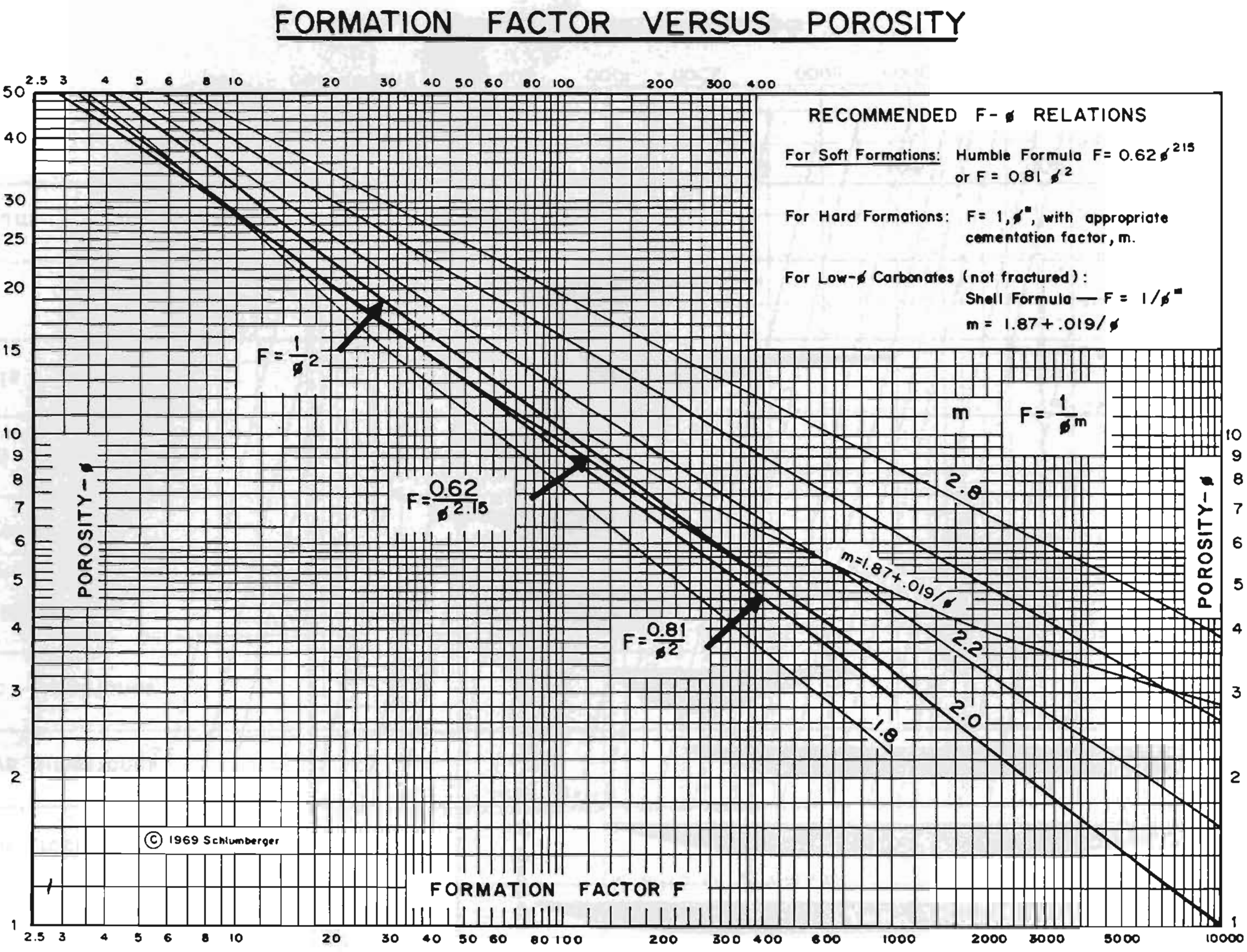


Fig. 11.12 Formation factor versus porosity

11.3 ELECTRICITY RESISTIVITY METHOD

11.3 Resistivity of Rocks

Electrical resistivity is a physical property that characterizes a material almost as definitely as its density. For most rocks the resistivity depends upon the quality and the quantity of the water filling the open spaces in the rock.

11.3.1a *The Quality of the Electrolyte*

The quality of the electrolyte, i.e., its resistivity , ρ_w , depends upon the salinity and the temperature (Fig. 11.10). Salinity is the percentage of dissolved salts, it is expressed in ppm, 1 mg/litre = 1 ppm. Water is considered to be undrinkable if it contains more than 8,000 ppm but it also depends upon the kind of dissolved salts. Salinity is generally expressed in "equivalent NaCl". Temperature also has an influence, because as temperature increases resistivity of the electrolyte decreases. Figure 11.10 (Chart 1) shows the influence of both temperature and salinity on the resistivity of the electrolyte.

Example:

Sea water	=	ρ_w	=	90.3	ohms.m
The Rhone	=	ρ_w	=	80	ohms.m
Lake Geneva	=	ρ_w	=	50	ohms.m
Charnawati River	=	ρ_w	=	200	ohms.m
Rainwater	=	ρ_w	=	30-1000	ohms.m
distillated water	=	ρ_w	=	∞	
oil	=	ρ_w	=	∞	
air	=	ρ_w	=	∞	

11.3.1b *The Quantity of Electrolyte*

Rocks are insulators in the dry state and their resistivities depend upon the water content. Water content in rocks is related to porosity ϕ and saturation S_w.

$$\textbf{porosity} \quad \phi \text{ is defined as } = \frac{\text{Void volume}}{\text{Total volume}} .$$

Porosity is expressed often as a percentage although it is used as a decimal $0 \leq \phi \leq 1$.

Table 11.2 Porosity of different materials

Materials	Grain Diameter	Porosity
Gravel	2.5 mm	45 %
Sand	0.125	40 %
Silty sand	0.005	32 %
Silt	0.003	36 %
Clayed silt	0.001	38 %
Clay	0.0002	47 %

It can be seen that porosity does not depend upon the grain size but it depends strongly upon the grain distribution. A well-graded, non-uniform soil will have a low porosity, a soil consisting of one or two grain sizes, i.e., a poorly-graded or uniform soil will have a high porosity of up to 50 per cent (Table 11.2).

The degree of **saturation**, Sw, is defined as

$$Sw = \frac{Vw}{Vv} \cdot 100 \qquad 0 < Sw < 1$$

This equation expresses the ratio of water present in the soil pores to the total amount which could be present if all the pores were full of water. It is the percentage of total void volume that contains water.

11.3.2 *Darcy's Law*

The resistivity of porous, water-bearing rocks follows Darcy's Law.

In saturated rocks:

$$\rho_r = a \cdot \phi_p^{-m} \cdot \rho_w$$

where,

ρ_r = resistivity of rock,
ρ_w = resistivity of interstitial water,
ϕ_p = porosity, and
a, m = certain parameters.

In unsaturated rocks:

in this case saturation must be taken into account, and Darcy's Law becomes

$$\rho_r = F \cdot \rho_w \cdot S_w^{-n} .$$

The value of n is usually close to 2 if more than about 30 per cent pore space is water filled but can be much greater for lesser water content. Fig 11.11 provides the general resistivity values for different rock types:

F is called **formation factor**, $F = a \cdot \phi^{-m}$ (Fig. 11.12).

11.3.3 *Point Current Electrode on Homogeneous Earth*

Consider a point electrode on the surface of a homogeneous isotropic earth, extending to infinity in the downward direction, and having a resistivity. The current flowing into the earth spreads out radially and hemispherical equipotential develops (Fig. 11.13). The potential at a distance r from the point current source is:

$$V_r = \frac{\rho \cdot I}{2\Pi} \cdot \left(\frac{1}{r}\right) .$$

In practice there are two electrodes, one positive A+, sending current into the ground, and, the other negative B - collecting the returning current.

The total potential at any point P in the ground will be :

$$V_p = \frac{\rho \cdot I}{2\Pi} \cdot \left(\frac{1}{r} - \frac{1}{r'}\right)$$

r and r' are the distances of P from the two electrodes.

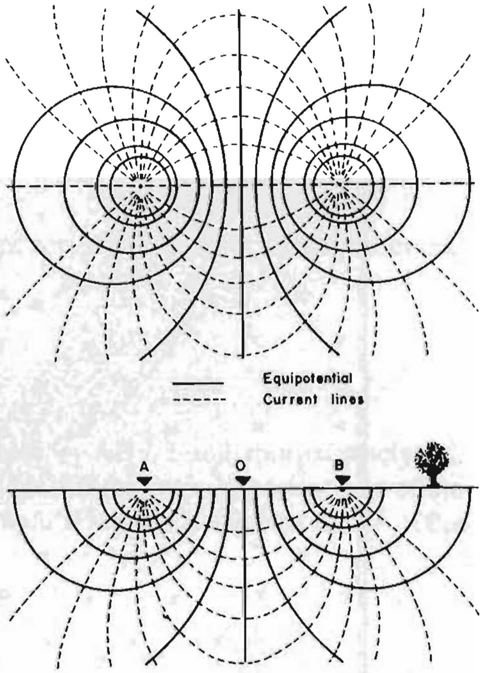


Fig. 11.13 Current flow through point electrodes

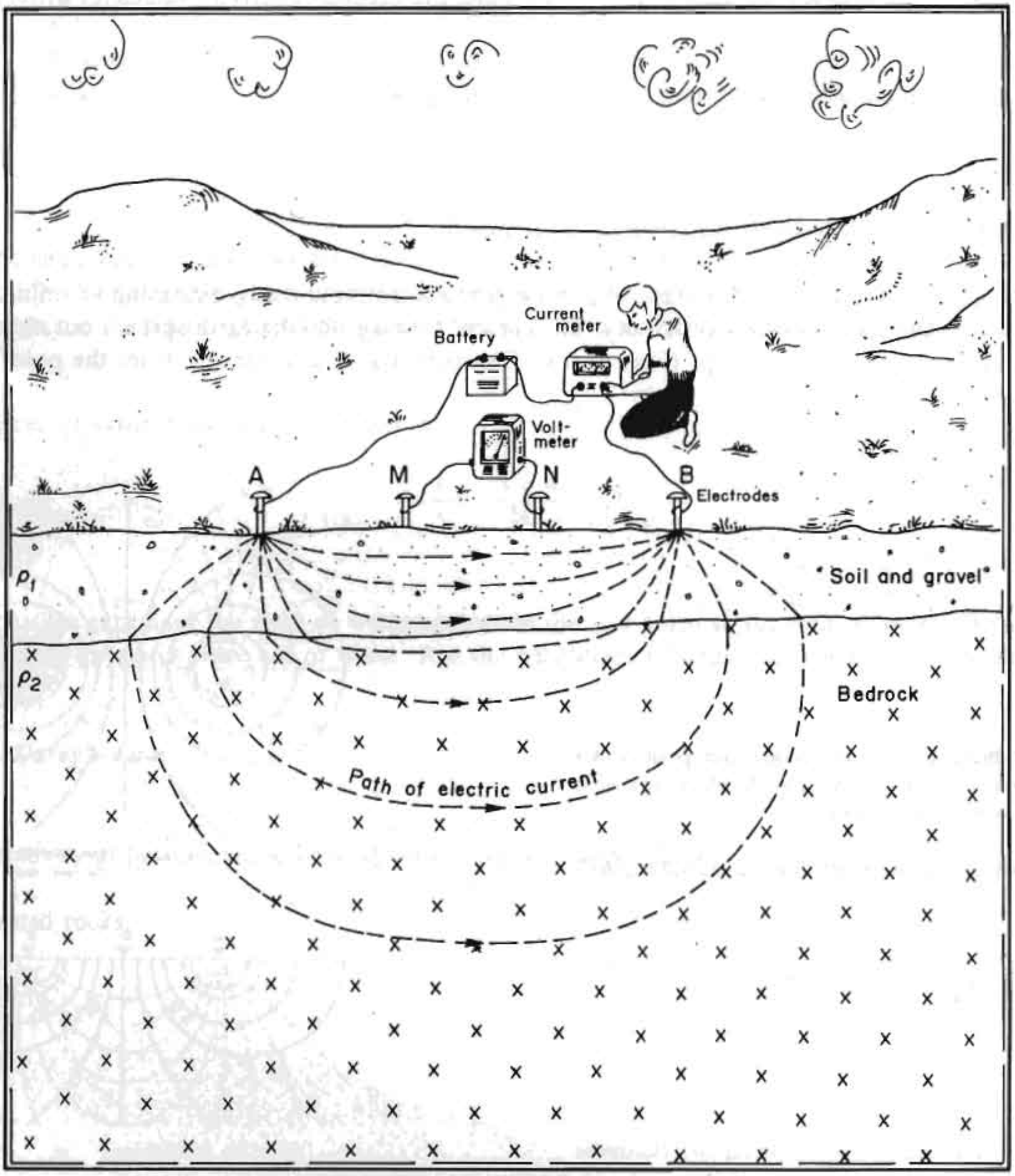


Fig. 11.14 Electrical resistivity survey

11.3.4 *Apparent Resistivity*

Let A +, B - be the current electrodes placed on the ground surface, and M, N two potential probes.

$$VM = \frac{\rho \; . \; I}{2\Pi} \left(\frac{1}{AM} - \frac{1}{BM}\right)$$

$$VN = \frac{\rho \; . \; I}{2\Pi} \left(\frac{1}{AN} - \frac{1}{BN}\right)$$

so,

$$\Delta V = VM - VN = \frac{\rho \; . \; I}{2\Pi} \left(\frac{1}{AM} - \frac{1}{BM} - \frac{1}{AN} + \frac{1}{BN}\right)$$

and

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

$$\rho = \Delta \, v_I \; . \; K \; .$$

ρ is called the apparent resistivity, K is a **geometric constant**, so called because it depends only upon the geometric configuration.

This is the fundamental equation of the **resistivity method**. It gives resistivity in terms of quantities that can be measured ΔV, I and the electrode separation distance.

11.3.5 *Current Penetration*

The distribution of current lines shows that 1/3 penetrate to a depth = AB / 2 and approximately 1/2 penetrate to a depth equal to the distance AB. In fact we can consider that the most important part of the current lays down a parallel piped AB/4, AB/2, 3AB/2. For example if we use a distance AB = 100m more than 200'000 m^3 will be taken into account.

11.3.6 *Depth of Investigation*

The **depth of investigation** is a function of the spacing between the two outer electrodes A and B. We can assume that the depth of investigation varies from 1/10 to 1/4 AB, but the heterogeneity of the ground can influence the depth of investigation.

11.3.7 *Heterogeneous Medium*

For a uniform sub-surface the apparent resistivity is equal to the true resistivity. If the earth is non-homogeneous, the current distribution will be affected and so is the potential and this produces a change in the resistivity measured. The measured **apparent resistivity** depends upon the resistivity of the various materials through which the current passes; it is an average of all those resistivities.

11.3.8 *Electrical Profiling or Mapping*

There are two basic field procedures used in the fields; **electrical profiling** in which the electrode separation remains constant during the survey, and **electrical sounding in which the centre of the electrode spread** is maintained at a fixed location and the electrode spacing AB is increased.

This method is normally employed in the rapid survey of an area. The object of mapping is to determine the lateral variations in the conductivity of the ground. Profiling or mapping is primarily useful for detecting local non-homogeneities and is employed typically in delineating geological boundaries, fracture zones, or steeply dipping contacts between different types of earth material. Figure 11.14 shows the lay-out of the electrical survey method.

Mapping Procedures

A number of different configurations of current and potential electrodes exists. In all arrangements, the electrodes are laid out along a line with the current electrodes generally placed on the outside of the potential electrodes. The opposite is theoretically equivalent. In the engineering world, three arrays have found wide usage (Fig. 11.15).

a. The Schlumberger Configuration

The Schlumberger Configuration is symmetric and colinear. The two potential electrodes M N are closely placed midway between the two current electrodes so that: MN < AB / 5.

b. The Wenner Array

The Wenner Configuration uses four electrodes, equally spaced along a straight line AM = MN = NB.

c. Dipole-Dipole Array

In this arrangement the current electrodes are on one side of the array and the potential electrodes on the other side. There is the same distance between the two current electrodes and between the two potential electrodes. The usual routine in this procedure is to move the chosen array with a fixed separation, as a whole, in suitable steps along the line of the array itself. When one traverse is finished the array is moved to the next parallel line and so on until the area of investigation is covered. Usually the array is moved so that each of the four electrodes advances through a distance equal to array separation AB, but it is also possible to advance through any other distance.

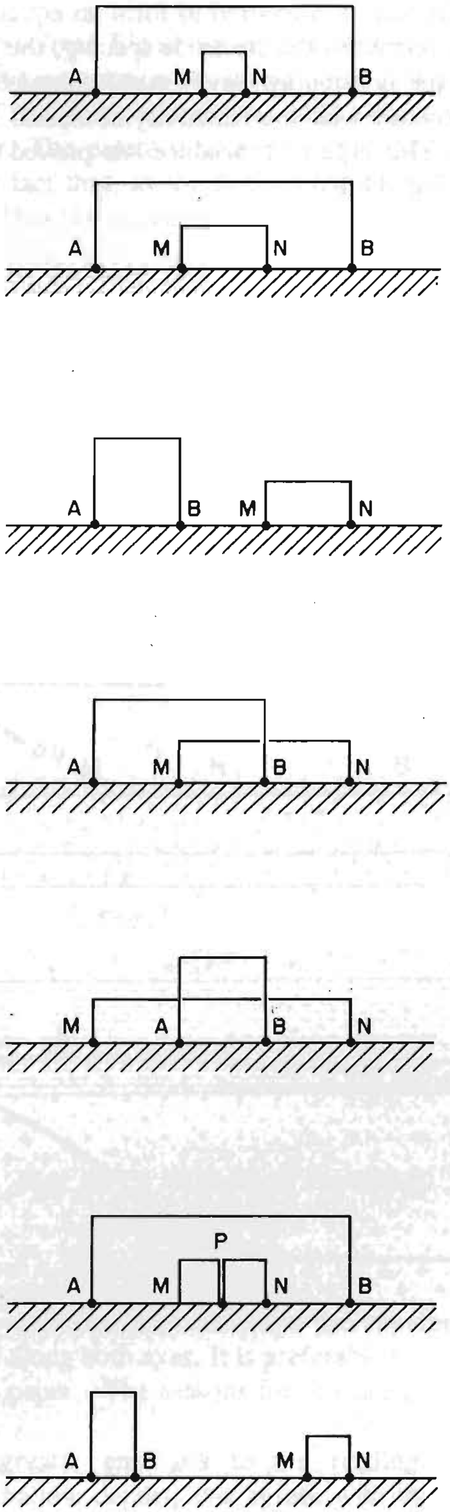


Fig. 11.15 Different configurations of current and potential electrodes

The apparent resistivity values calculated at each station are plotted on a map at positions corresponding to the centre of the array, and the map is contoured to form an equal resistivity contour map. Since the depth of investigation is roughly related to the electrode spacing, the depth of investigation, essentially, will be constant for all the readings. A mapping survey may be thought of as an **electrical trench 2.** The geologic boundaries can be correlated with the resistivity contours. A resistivity map must always be related to the spacing AB used. This type of procedure has proved helpful in delineating fault zones, contacts, and lateral variations (Fig. 11.16).

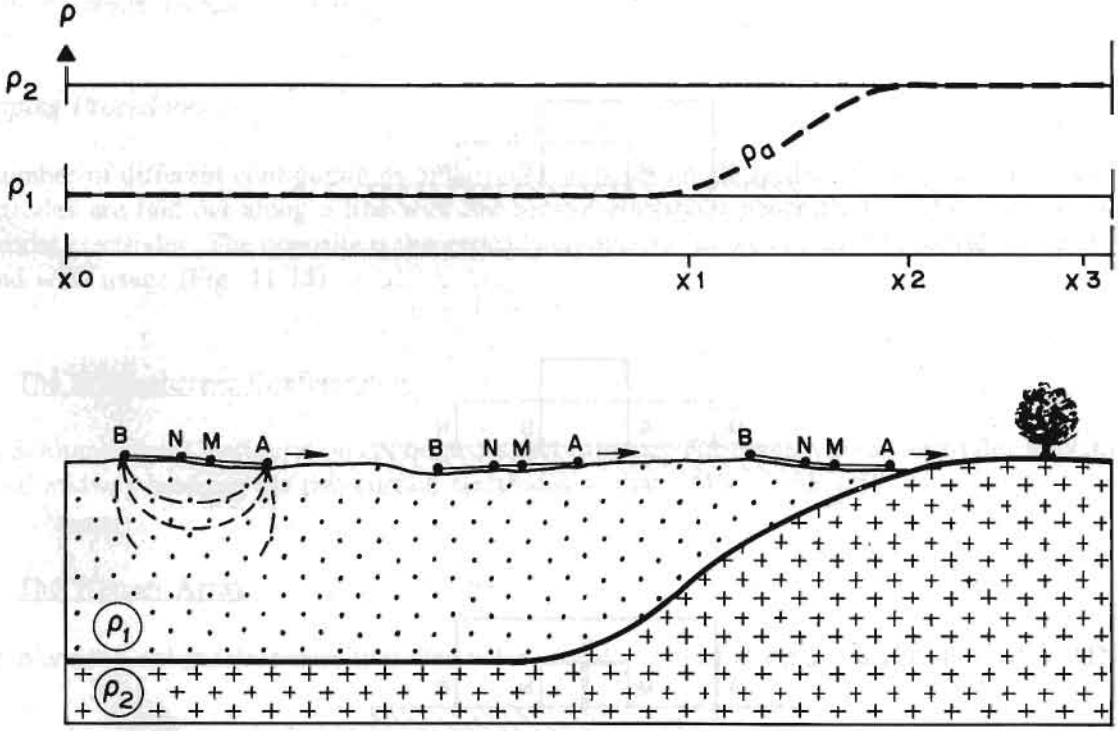


Fig. 11.16 Electrical profiling

11.4 ELECTRICAL SOUNDING

Electrical sounding, also called **Vertical Electrical Sounding** (VES), is the second basic procedure designed to provide information on the variation in earth conditions **with depth**. Among the applications are : estimation of the variation of resistivity with depth, estimation of depth to water-bearing layer, and estimation of thickness of a layer. The essential idea behind electrical sounding, assuming resistivity variations with depth only, is the fact that, as the distance between AB is increased, the current lines reach increasingly deeper levels. Thus the apparent resistivity will be more and more influenced by the resistivity at deeper levels.

11.4.1 *Field Procedure*

The MN electrodes are kept fixed and AB moved outward symmetrically about the centre position O. At some stage the MN voltage V will in general fall below the reading accuracy of the voltmeter in which case the distance MN is increased maintaining of course the condition MN < AB (Fig. 11.17).

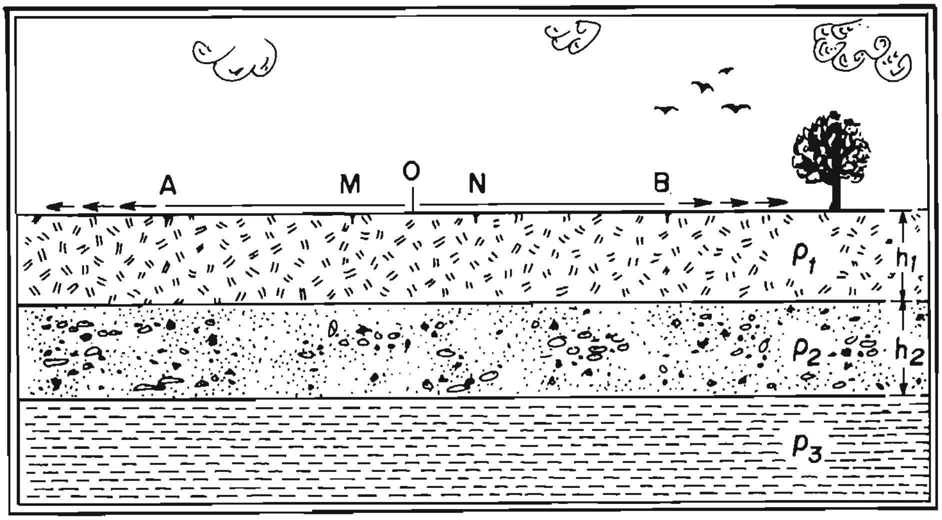


Fig. 11.17 Electrical sounding

11.4.2 **Plotting**

The data are plotted with OA = AB/2 on the x axis and the apparent resistivity on the Y axis (Fig. 11.18). **Logarithmic scale** is used along both axes. It is preferable to use double-logarithmic graph paper rather than ordinary cross-section paper. The reasons for this are given below.

1. The logarithmic plot gives greater emphasis to the readings at small electrode separations, corresponding to material at shallow depths, this agrees with the fact that the measured quantities (apparent resistivities) represent a weighted average of all the true resistivities in a fairly large volume and that the material close to the surface is always weighted more heavily than the material at depth.

2. The quantitative methods of interpretation require that the data be plotted on double logarithmic paper.

3. Use of logarithmic scales makes the shape and size of the curve independent of units of measurements so that comparison becomes easier.

4. The precision required, 10 per cent minimum, is relative to 10 or 11 ohms.m, 100 or 110, 1000 or 1100 and so on for the resistivities, the same for the depths.

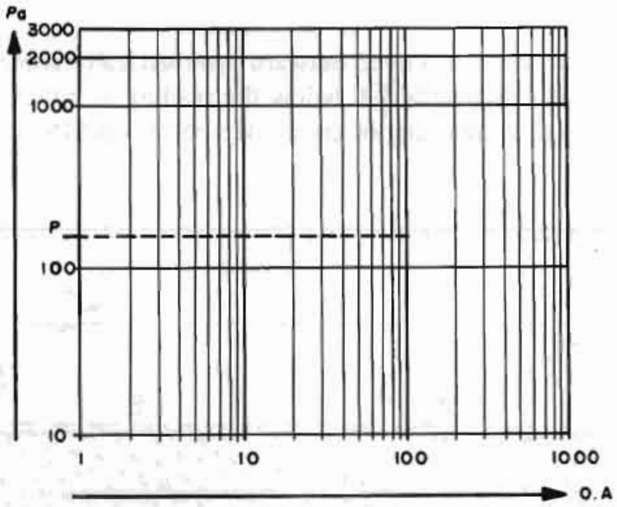


Fig. 11.18 Distance versus resistivity plot in a double logarithmic graph

11.4.3 ***Quantitative Interpretation***

Homogeneous Earth

In this case, the apparent resistivity measured is in fact the true resistivity, no matter the distance AB. The sounding curve in this case will be a straight line crossing the Y axis at the value of the true resistivity.

Two Layer Structures

As a first example, consider a low resistivity layer, such as a wet soil, overlying a thick high resistivity layer such as a dense rock. The true resistivity of the surface material is not changed by the presence of the rock, since the rock does not extend to the surface (Fig. 11.19).

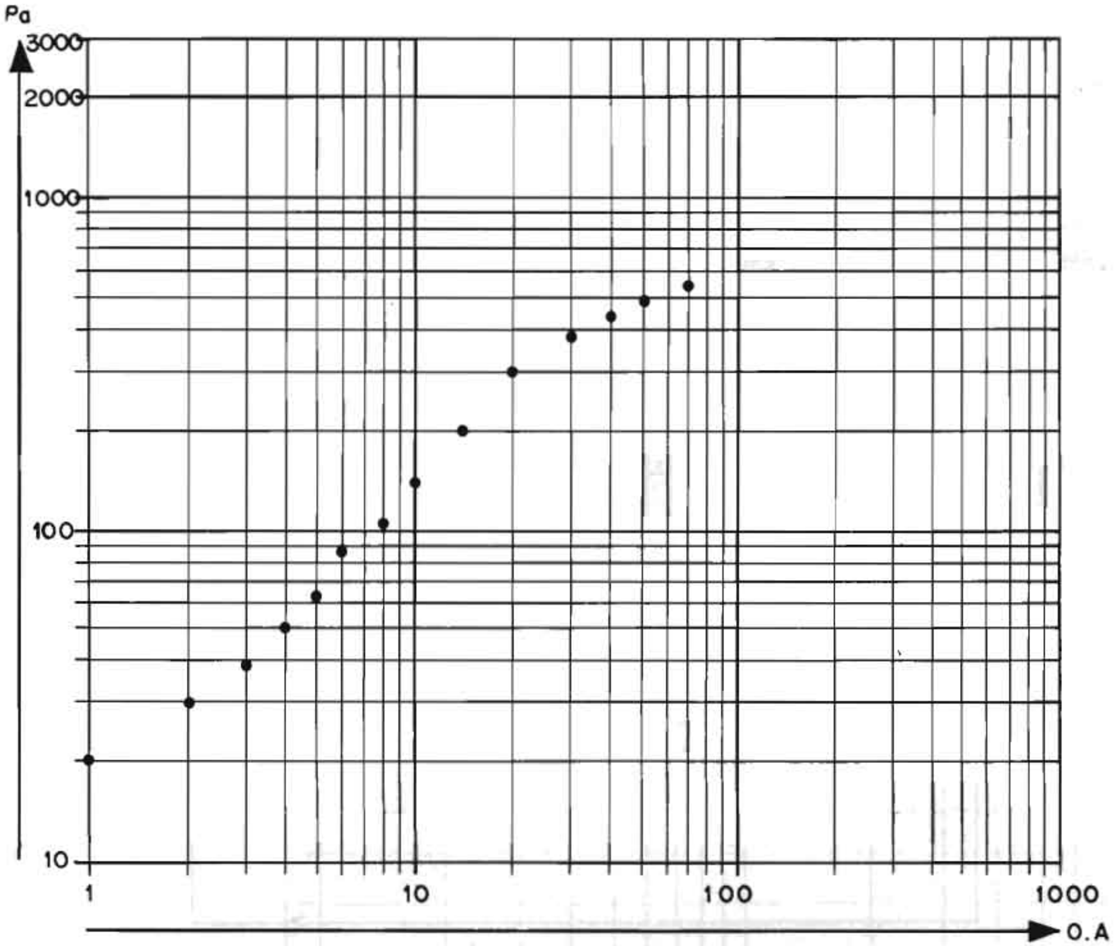


Fig. 11.19 Distance - resistivity plot

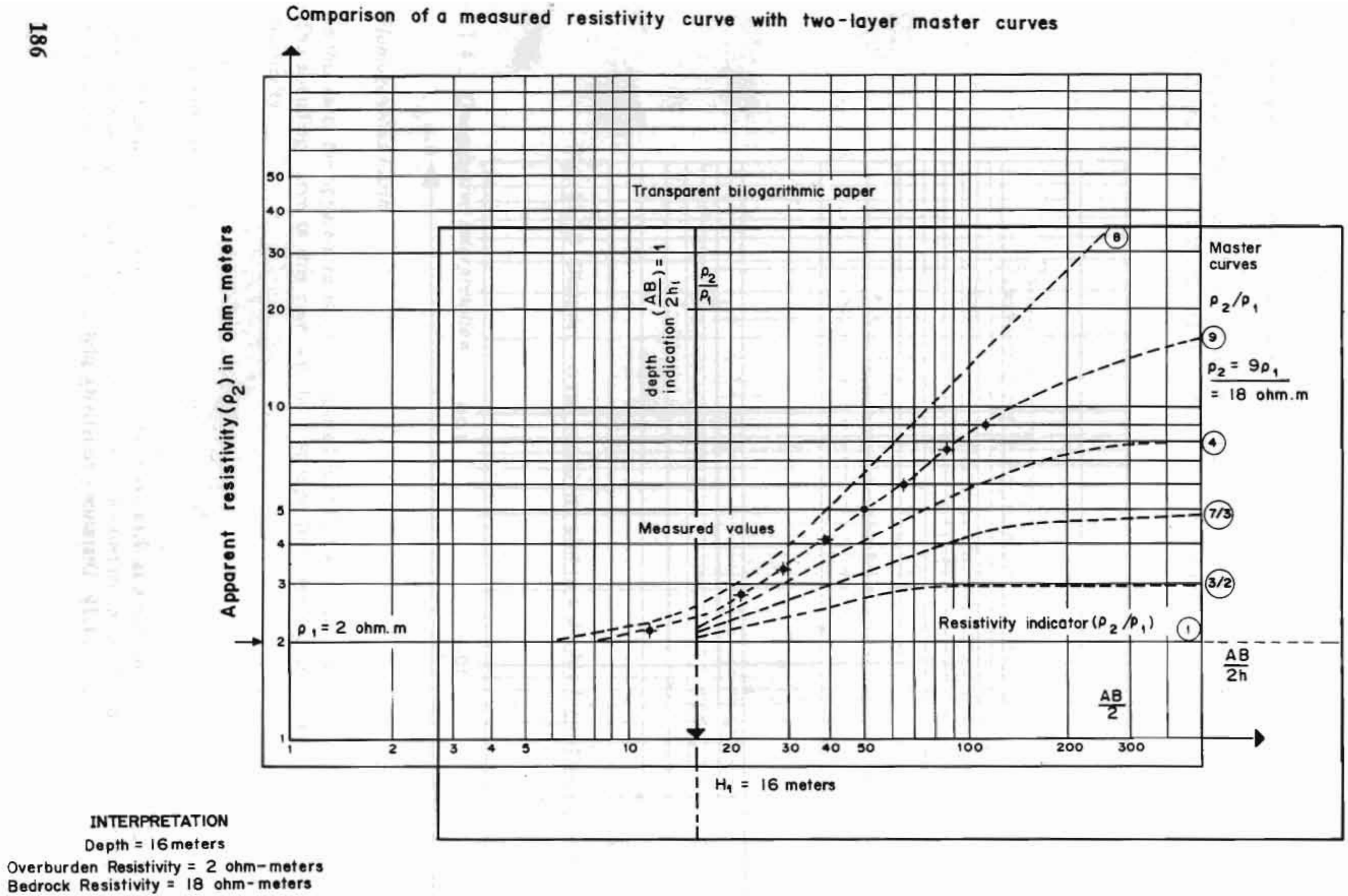


Fig. 11.20 Using curve matching to interpret a curve graphed in ohm-meters

The bedrock effect upon the distribution of current will depend more precisely upon the ratio of the electrode spacing to the depth of the bedrock; so when the electrode is small compared to the depth, the current density will be largely unaffected by the rock. As the electrode spacing is increased, the effect of the lower-lying high resistivity stratum increases. Hence, the apparent resistivity curve will rise smoothly to gradually approach the **true resistivity** of the bedrock layer as the electrode spacing becomes very large relative to the depth of the rock. The true resistivity of the near surface soil layer can be easily determined. It is simply the left hand limit of the sounding curve. If the sounding curve is extrapolated back to the limit of zero electrode separation, this apparent resistivity will be equal to the true resistivity of the surface layer.

11.4.4 *Interpretation by Curve Matching*

The first step is to plot ρ apparent against OA on transparent double logarithmic paper. The method involves a comparison of the measured curve with a set of theoretically calculated master curves (with the same double logarithmic modulus). Keeping the respective axes parallel, the paper is slid on various master curves in succession until a satisfactory match is obtained with some curve; if necessary an interpolated one (see Fig. 11.20). The value of OA, coinciding with the point 1 on the X axis of the matching curve, gives H1 and the measured value of ρ apparent coinciding with the point 1 on the axis gives ρ 1. The value of ρ is obtained from the appropriate parameter belonging to the matching master curve.