GEOPHYSICAL INVESTIGATION

ELECTRICAL METHODS

Electrical Method

- Electrical resistivity method is based on the difference in the electrical conductivity or the electrical resistivity of different soils. Resistivity is defined as resistance in ohms between the opposite phases of a unit cube of a material.
    - $\rho = (\frac{RA}{l})$
    - ρ is resistivity in ohm-cm,
    - R is resistance in ohms,
    - A is the cross sectional area (cm 2),
    - L is length of the conductor (cm).

- The resistivity values of the different soils are listed in table 1.4

Material	Resistivity (Ω-cm)
Massive rock	> 400
Shale and clay	1.0
Seawater	0.3
Wet to moist clayey soils	1.5 - 3.0

Table 1.4 : Resistivity of different materials

Principle

- All electrical methods are based on the fundamental fact that different materials of earth's crust methods posses widely different electrical properties. Resistivity, Electrochemical activity and dielectrical constant are some of these properties that are generally studied through this methods. Result obtained from such studies when interpreted properly give sufficiently useful clues regarding the nature and make up of the subsurface materials.

Two common methods

- ## 1.Equipotential Methods:
- In this method 2 primary electrodes are inserted into the ground,6-7 metres apart from each other, across which current is introduced. The position of these primary electrodes remains fixed in the subsequent investigations.
- Potential between these primary electrodes is determined with the help of two search electrodes and points of equal potential found out along the entire region under investigation, which are joined to get equipotential lines.

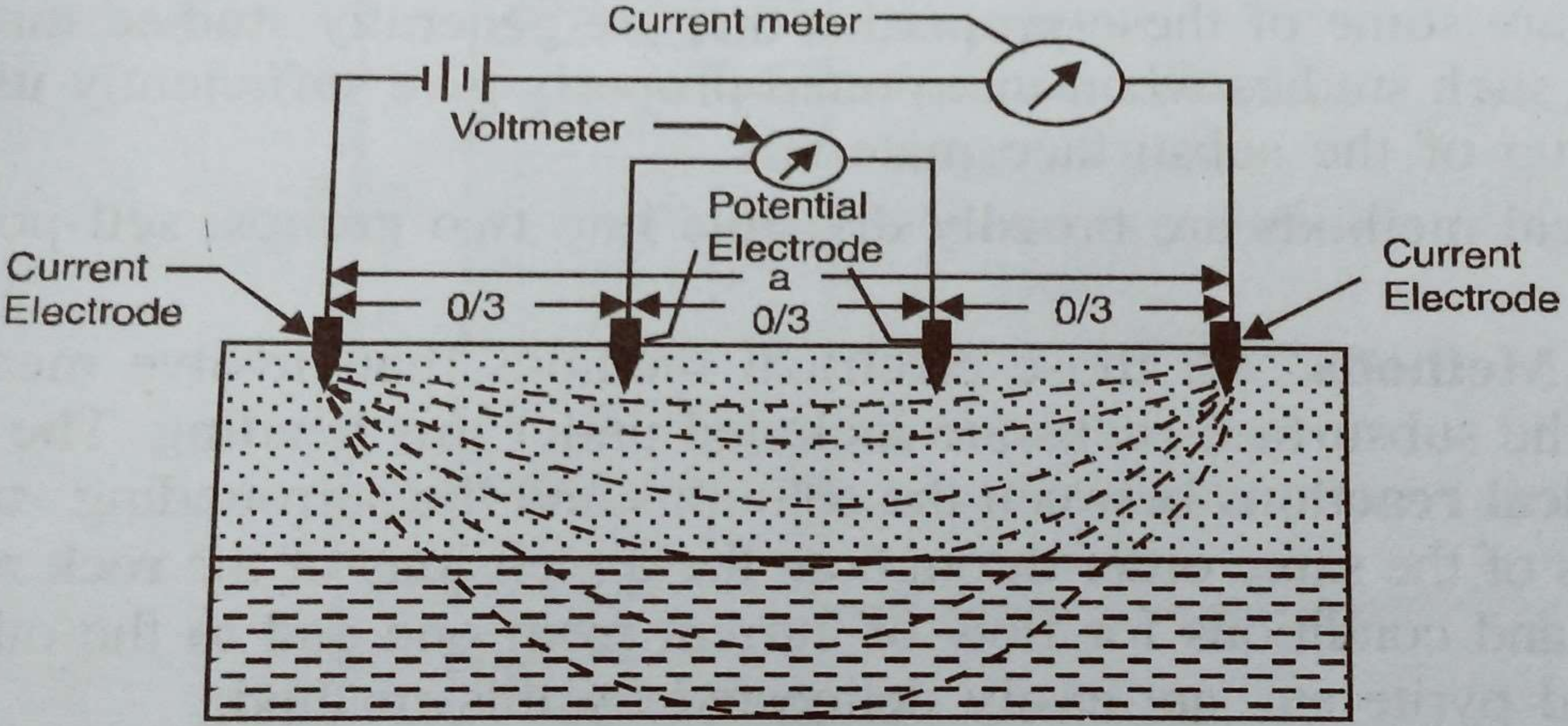
- Under normal conditions, i.e when the material below is of uniform nature, the equipotential lines would be regular in character. But in cases when the material below is not of uniform character i.e ,it contains patches of high or low conductivity, the equipotential lines would show clear distortions or irregularities which would include the probable locations of rock masses of different characteristics.

2.Resistivity Method:

- It is similar to equipotential method but in this case it is the resistivity of the material of the subsurface which is determined from which important interpretations are made. Here also, a known current is introduced through two electrodes- the current electrodes, which are inserted at some distances apart from each other. Potential gradient is then measured directly with the help of two or more potential electrodes placed at proper distances within the two outer current electrodes.

A typical resistivity meter

- A resistivity meter consist of both a voltmeter an a currentmeter (ammeter).
- Most system report the ratio V/I instead of each one seperatly.
    - The resistance can then be converted into resistivity using geometrical parameters based on the type of array.
    - Most modern resistivity system typically utilize atleast 4 electrodes.
    - About 70% of the current applied by two electrodes at the surface stays within a depth equal to the seperation of the electrodes.
- In this method, the electrodes are driven approximately 20cms in to the ground and a dc or a very low frequency ac current of known magnitude is passed between the outer (current) electrodes, thereby producing within the soil an electrical field and the boundary conditions.



Resistivity Method for Subsurface Investigation

Fig. 22.2.

RESISTIVITY-HOW DO WE MEASURE IT???

- Apply a non potential difference(measured with voltmeter) to a circuit with a resistive material of known length and cross sectional area.
- Then measure the current with ammeter.
- This gives the resistance R.

The spacing of this current and potential electrodes is of vital importance in this method. A number of arrangements have been suggested of which WENNER'S arrangement is followed quite commonly. In this arrangement , the potential electrode are placed at a distance 1/3 'a', where 'a' is the total distance between the current electrodes.

Resistivity is then calculated by the formula:

$$\rho = 2\pi \frac{dv}{I}$$

Where ρ=resistivity in ohm-m, v=Potential difference, I=current in ampere, d=$\frac{1}{3}$ 'a', where 'a' is total distance between outer electrodes.

- For simple sounding, a *Wenner array* is used as shown in fig. 1.16. Then, the resistivity is given as,
- $$\rho=\left(\frac{2\pi Ra}{l}\right)$$
  - a is the spacing between the electrodes.

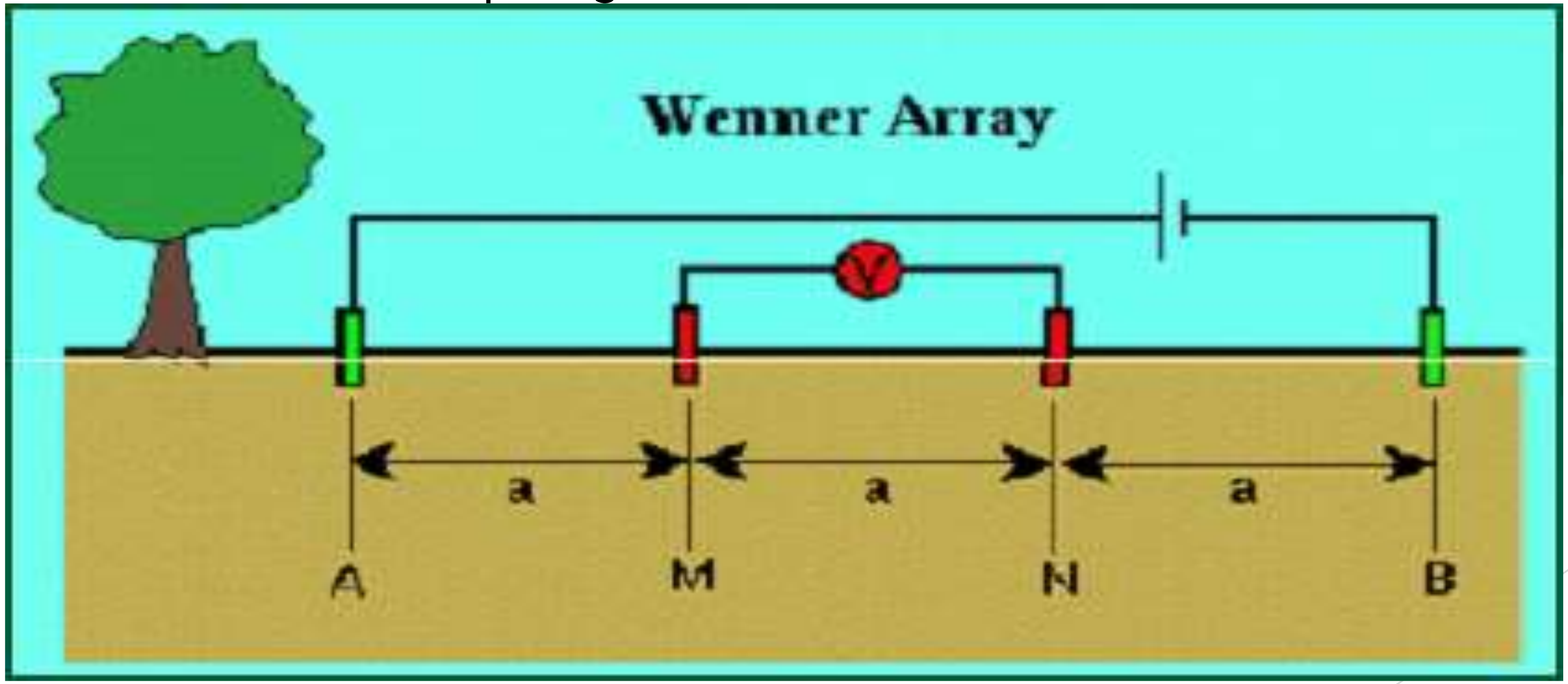


Fig. 1.16 Wenner arrangement

Advantages of Electrical Method

- It is a very rapid and economical method.
- It is good up to 30m depth.
- The instrumentation of this method is very simple.
- It is a non-destructive method.

Disadvantages of Electrical Method

- It can only detect absolutely different strata like rock and water.
- It provides no information about the sample.
- Cultural problems cause interference, e.g., power lines, pipelines, buried casings, fences.
- Data acquisition can be slow compared to other geophysical methods, although that difference is disappearing with the very latest techniques.

Resistivity used in:

- Map faults
- Map lateral extent of conductive contaminant plumes
- Locate voids
- Map heavy metals soil contamination
- Delineate disposal areas
- Map paleochannels
- Explore for sand and gravel
- Map archaeological sites

Methods

- 1.Self Potential methods:
- All those electrical methods that involve measurement of natural electrical potential of the subsurface rocks are included under this heading. The natural potential may be due to electrochemical reactions between the solutions and the surrounding subsurface rocks. These reactions are not always of the same order throughout the dimensions of the rock masses thereby creating a potential difference and conditions for flow of current from one end to the other end. Elongated ore bodies of magnetite and pyrite etc are easily delineated by this method.

- 2.Potential Drop methods:
- This include a variety of methods in which electrical current is artificially introduced from an external source at certain points and then its flow through subsurface materials recorded at different distances. The behaviour of the flow lines of the electrical current is directly related to the nature of the subsurface materials.

The depth of penetration of electrical current in these investigations is broadly equal to d although there are many attached to this generalization.

The resistivity method envisages interpretation of the qualitative as well as quantitative characters of the sub surface materials which are governed by two basic principle:

(i) If material below is of uniform nature , the resistivity values would be of regular character.

(ii) If the material is non-uniform, that is ,it consist of layers or masses of different character ,then these would be indicated irregularities or anomalies in the resistivity values. The depths of at which these anomalies occur can be calculated and also the nature of the sub surface material broadly understood.

- In common practice, the resistivity of a given area is measured by gradually increasing the distance between the electrodes and by changing the directions of the profiles . Results are plotted on a graph with resistivity and depth as main factors. An abrupt change in the resistivity curve(rise or fall) would often suggest a change in the nature of the material corresponding to that depth. The existence of specially low zones such as aquifers, fractured zones, buried valley etc. may be indicated even in linear traverse keeping the spacing between the electrodes constant along the entire traverse.

Applications

- (a)In Prospecting: The electrical methods have been successfully employed in delineation of ore bodies occurring at shallower depths. For such surveys at great depths, these are not of much help.
- In table 1 some typical value-ranges of resistivity are given. As may be seen, rocks exhibit a great variation ranging from as high resistivity as > 10^5 ohms-meters in igneous rocks to as low as less than 1 ohm-m for clayey marls.
- (b)In Civil Engineering: Resistivity methods have been widely used in engineering investigation for determination of
- (i)Depth to the bed rock: as for instance, in important projects like dams, buildings and bridge foundations, where it would be desirable that the structure should rest on sound hard rocks rather than on overburden or soil.

- (ii) Location of geological structures: like folds; buried valleys, crushed and fractured zones due to shearing and faulting (Fig 22.3)
- (iii) Location of Aquifers: and other water bearing zones which could be easily interpreted on the basis of known resistivity values of moistures rich rocks and dry rocks.

Some more Applications

- Characterize subsurface hydrogeology
- Determine depth to bedrock/overburden thickness
- Determine depth to groundwater
- Map stratigraphy
- Map clay aquitards
- Map salt-water intrusion
- Map vertical extent of certain types of soil and
- groundwater contamination
- Estimate landfill thickness

ELECTRICAL RESISTIVITY METHOD FOR GROUNDWATER INVESTIGATION

- Electrical resistivity methods of geophysical prospecting are well established and the most important method for groundwater investigations. Groundwater, through the various dissolved salts it contains, is ionically conductive and enables electric currents to flow into the ground. Consequently, measuring the ground resistivity gives the possibility to identify the presence of water, taking in consideration the following properties:
- • a hard rock without pores or fracture and a dry sand without water or clay are very resistive: several tens thousands ohm.m
- • a porous or fractured rock bearing free water has a resistivity which depends on the resistivity of the water and on the porosity of the rock (see below): several tens to several thousands ohm.m
- • an impermeable clay layer, which has bound water, has a low resistivity: several units to several tens ohm.m
- • mineral orebodies (iron, sulphides, …) have very low resistivities due to their electronic conduction: usually lower or much lower than 1 ohm.m

GROUNDWATER DETECTION

- To identify the presence of groundwater from resistivity measurements, one can look to the absolute value of the ground resistivity, through the Archie law: for a practical range of fresh water resistivity of 10 to 100 ohm.m, a usual target for aquifer resistivity can be between 50 and 2000 ohm.m. Most of the time it is the relative value of the ground resistivity which is considered for detecting groundwater: in a hard rock (resistant) environment, a low resistivity anomaly will be the target, while in a clayey or salty (conductive) environment, it is a high resistivity anomaly which will most probably correspond to (fresh) water. In sedimentary layers, the product of the aquifer resistivity by its thickness can be considered as representative of the interest of the aquifer. However, electrical methods cannot give an estimation of the permeability but only of the porosity. The contrast of resistivity between a fresh water and a salted water (coming from a sea intrusion for instance) is high and the depth of the water wedge is usually well determined with electrical methods.

ELECTRICAL RESISTIVITY METHOD USING SCHLUMBERGER , WENNER AND DIPOLE – DIPOLE ARRAY

Introduction

1. Surface electrical resistivity surveying is based on the principle that the distribution of electrical potential in the ground around a current-carrying electrode depends on the electrical resistivity's and distribution of the surrounding soils and rocks.
2. The usual practice in the field is to apply an electrical direct current (DC) between two electrodes implanted in the ground and to measure the difference of potential between two additional electrodes that do not carry current.
3. Usually, the potential electrodes are in line between the current electrodes, but in principle, they can be located anywhere.
4. The current used is either direct current, commutated direct current (i.e., a square-wave alternating current), or AC of low frequency (typically about 20 Hz).
5. All analysis and interpretation are done on the basis of direct currents.
6. The distribution of potential can be related theoretically to ground resistivity's and their distribution for some simple cases, notably, the case of a horizontally stratified ground and the case of homogeneous masses separated by vertical planes (e.g., a vertical fault with a large throw or a vertical dike).
7. For other kinds of resistivity distributions, interpretation is usually done by qualitative comparison of observed response with that of idealized hypothetical models or on the basis of empirical methods.

RESISTIVITY SURVEY USES

1. Mineral grains comprised of soils and rocks are essentially nonconductive, except in some exotic materials such as metallic ores, so the resistivity of soils and rocks is governed primarily by the amount of pore water, its resistivity, and the arrangement of the pores.
2. To the extent that differences of lithology are accompanied by differences of resistivity, resistivity surveys can be useful in detecting bodies of anomalous materials or in estimating the depths of bedrock surfaces.
3. In coarse, granular soils, the groundwater surface is generally marked by an abrupt change in water saturation and thus by a change of resistivity.
4. In fine-grained soils, however, there may be no such resistivity change coinciding with a piezometric surface.

RESISTIVITY DATA INTERPRETATION

1. Generally, since the resistivity of a soil or rock is controlled primarily by the pore water conditions, there are wide ranges in resistivity for any particular soil or rock type, and resistivity values cannot be directly interpreted in terms of soil type or lithology. Commonly, however, zones of distinctive resistivity can be associated with specific soil or rock units on the basis of local field or drill hole information, and resistivity surveys can be used profitably to extend field investigations into areas with very limited or nonexistent data.
2. Also, resistivity surveys may be used as a reconnaissance method, to detect anomalies that can be further investigated by complementary geophysical methods and/or drill holes.
3. The electrical resistivity method has some inherent limitations that affect the resolution and accuracy that may be expected from it.
4. Like all methods using measurements of a potential field, the value of a measurement obtained at any location represents a weighted average of the effects produced over a large volume of material, with the nearby portions contributing most heavily.
5. This tends to produce smooth curves, which do not lend themselves to high resolution for interpretations.

DIFFICULTIES

1. Another feature common to all potential field geophysical methods is that a particular distribution of potential at the ground surface does not generally have a unique interpretation.
2. Although these limitations should be recognized, the non-uniqueness or ambiguity of the resistivity method is scarcely less than with the other geophysical methods.
3. For these reasons, it is always advisable to use several complementary geophysical methods in an integrated exploration program rather than relying on a single exploration method.

Theory

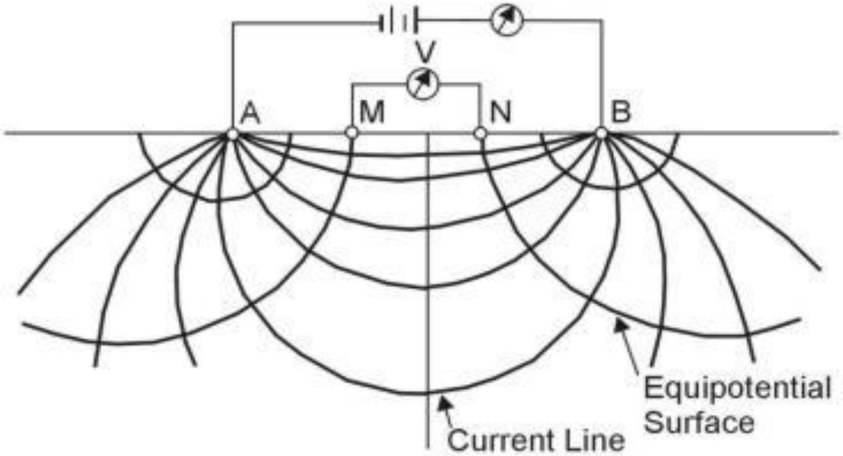
1. Data from resistivity surveys are customarily presented and interpreted in the form of values of apparent resistivity ρ_a.
2. Apparent resistivity is defined as the resistivity of an electrically homogeneous and isotropic half-space that would yield the measured relationship between the applied current and the potential difference for a particular arrangement and spacing of electrodes.
3. An equation giving the apparent resistivity in terms of applied current, distribution of potential, and arrangement of electrodes can be arrived at through an examination of the potential distribution due to a single current electrode.
4. The effect of an electrode pair (or any other combination) can be found by superposition.
5. Consider a single point electrode, located on the boundary of a semi-infinite, electrically homogeneous medium, which represents a fictitious homogeneous earth.
6. If the electrode carries a current I, measured in amperes (a), the potential at any point in the medium or on the boundary is given by:

$$U = \rho \frac{I}{2\pi r},$$

where

U = potential, in V,
ρ = resistivity of the medium,
r = distance from the electrode.

The mathematical demonstration for the derivation of the equation may be found in textbooks on geophysics, such as Keller and Frischknecht (1966).

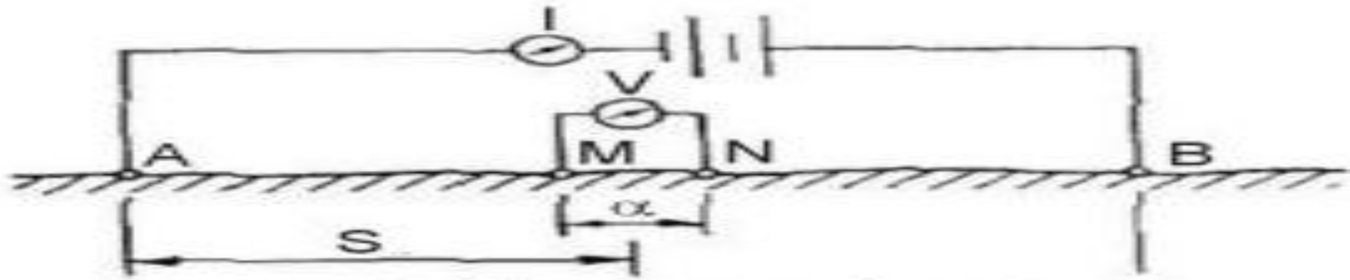


- Figure illustrates the electric field around the two electrodes in terms of equipotentials and current lines. The equipotentials represent imagery shells, or bowls, surrounding the current electrodes, and on any one of which the electrical potential is everywhere equal. The current lines represent a sampling of the infinitely many paths followed by the current, paths that are defined by the condition that they must be everywhere normal to the equipotential surfaces.

Apparent Resistivity

1. Wherever these measurements are made over a real heterogeneous earth, as distinguished from the fictitious homogeneous half-space, the symbol ρ is replaced by ρ_a for apparent resistivity.
2. The resistivity surveying problem is, reduced to its essence, the use of apparent resistivity values from field observations at various locations and with various electrode configurations to estimate the true resistivity's of the several earth materials present at a site and to locate their boundaries spatially below the surface of the site.
3. An electrode array with constant spacing is used to investigate lateral changes in apparent resistivity reflecting lateral geologic variability or localized anomalous features.
4. To investigate changes in resistivity with depth, the size of the electrode array is varied.
5. The apparent resistivity is affected by material at increasingly greater depths (hence larger volume) as the electrode spacing is increased.
6. Because of this effect, a plot of apparent resistivity against electrode spacing can be used to indicate vertical variations in resistivity.
7. The types of electrode arrays that are most commonly used (Schlumberger, Wenner, and dipole-dipole) are illustrated in upcoming slides.
8. There are other electrode configurations that are used experimentally or for non-geotechnical problems or are not in wide popularity today.
9. Some of these include the Lee, half-Schlumberger, polar dipole, bipole dipole, and gradient arrays.

Schlumberger Array



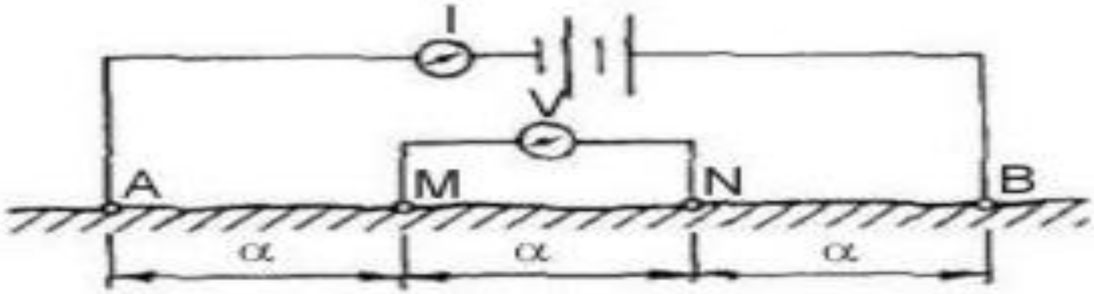
a. Schlumberger Array

1. For this array , in the limit as *a* approaches zero, the quantity V/*a* approaches the value of the potential gradient at the midpoint of the array. In practice, the sensitivity of the instruments limits the ratio of *s* to *a* and usually keeps it within the limits of about 3 to 30. Therefore, it is typical practice to use a finite electrode spacing and equation 2 to compute the geometric factor (Keller and Frischknecht, 1966). The apparent resistivity (⍰) is:

$$\rho_a = \pi\left[\frac{s^2}{a} - \frac{a}{4}\right]\frac{V}{I} = \pi a\left[\left(\frac{s}{a}\right)^2 - \frac{1}{4}\right]\frac{V}{I},$$

2. In usual field operations, the inner (potential) electrodes remain fixed, while the outer (current) electrodes are adjusted to vary the distance *s*.
3. The spacing *a* is adjusted when it is needed because of decreasing sensitivity of measurement.
4. The spacing *a* must never be larger than 0.4*s* or the potential gradient assumption is no longer valid.
5. Also, the *a* spacing may sometimes be adjusted with *s* held constant in order to detect the presence of local inhomogeneities or lateral changes in the neighborhood of the potential electrodes.

Wenner Array



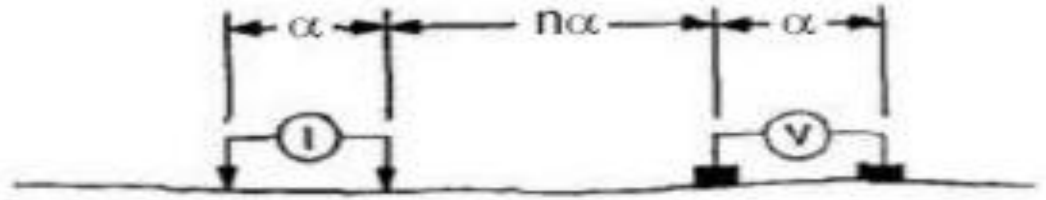
b. Wenner Array

1. This array consists of four electrodes in line, separated by equal intervals, denoted *a*. Applying equation 2, the user will find that the geometric factor K is equal to *a* , so the apparent resistivity is given by:

$$\rho_a = \pi\left[\frac{s^2}{a} - \frac{a}{4}\right]\frac{V}{I} = \pi a\left[\left(\frac{s}{a}\right)^2 - \frac{1}{4}\right]\frac{V}{I},$$

2. Although the Schlumberger array has always been the favored array in Europe, until recently, the Wenner array was used more extensively than the Schlumberger array in the United States.
3. In a survey with varying electrode spacing, field operations with the Schlumberger array are faster, because all four electrodes of the Wenner array are moved between successive observations, but with the Schlumberger array, only the outer ones need to be moved. The Schlumberger array also is said to be superior in distinguishing lateral from vertical variations in resistivity.
4. On the other hand, the Wenner array demands less instrument sensitivity, and reduction of data is marginally easier.

Dipole-dipole Array



c. Dipole - Dipole Array

1. The dipole-dipole array is one member of a family of arrays using dipoles (closely spaced electrode pairs) to measure the curvature of the potential field.
2. If the separation between both pairs of electrodes is the same *a,* and the separation between the centers of the dipoles is restricted to *a(n+1),* the apparent resistivity is given by:

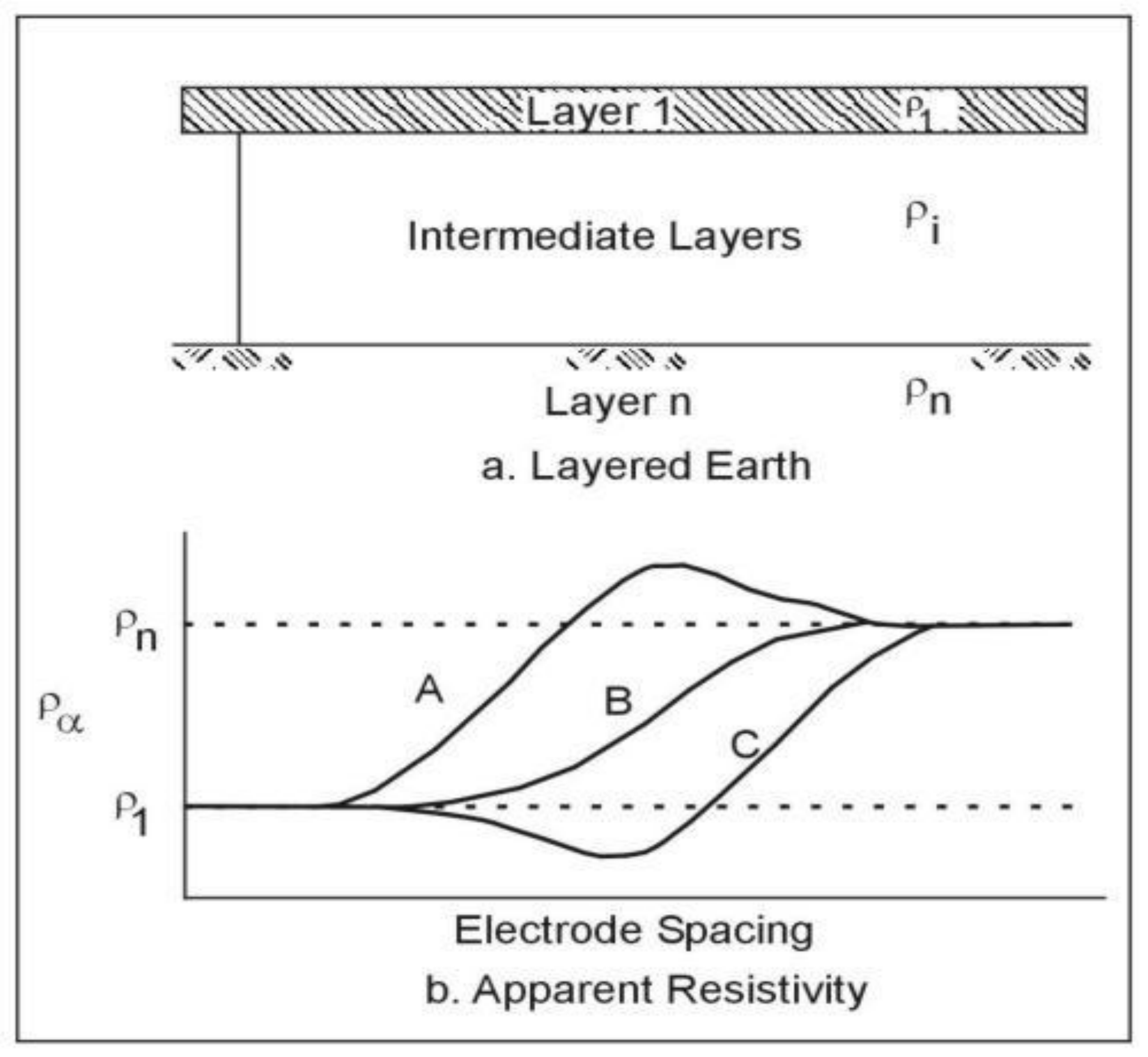
$$\rho_a = \pi a n(n+1)(n+2)\tfrac{V}{I},$$

3. This array is especially useful for measuring lateral resistivity changes and has been increasingly used in geotechnical applications.

Depth of Investigation

- To illustrate the major features of the relationship between apparent resistivity and electrode spacing, figure 3 shows a hypothetical earth model and some hypothetical apparent resistivity curves.
- The earth model has a surface layer of resistivity ρ1 and a basement layer of resistivity ρn that extends downward to infinity (figure 3a).
- There may be intermediate layers of arbitrary thicknesses and resistivities.
- The electrode spacing may be either the Wenner spacing *a* or the Schlumberger spacing *a*; curves of apparent resistivity versus spacing will have the same general shape for both arrays, although they will not generally coincide.
- For small electrode spacings, the apparent resistivity is close to the surface layer resistivity, whereas at large electrode spacings, it approaches the resistivity of the basement layer.
- Every apparent resistivity curve thus has two asymptotes, the horizontal lines $\rho_a = \rho_1$ and $\rho_a = \rho_n$, that it approaches at extreme values of electrode spacing.
- This is true whether ρ_n is greater than ρ_1, as shown in figure 3b, or the reverse.
- The behavior of the curve between the regions where it approaches the asymptotes depends on the distribution of resistivities in the intermediate layers.
- Curve A represents a case in which there is an intermediate layer with a resistivity greater than ρ_n.
- The behavior of curve B resembles that for the two-layer case or a case where resistivities increase from the surface down to the basement.
- The curve might look like curve C if there were an intermediate layer with resistivity lower than $\rho_{1.}$
- Unfortunately for the interpreter, neither the maximum of curve A nor the minimum of curve C reach the true resistivity values for the intermediate layers, though they may be close if the layers are very thick.

Asymptotic behavior



- Figure 3. Asymptotic behavior of the apparent resistivity curves at very small and very large electrode spacings.

Instruments and Measurements

- In concept, a direct current (I), or an alternating current of low frequency, is applied to the current electrodes, and the current is measured with an ammeter.
- Independently, a potential difference *V* is measured across the potential electrodes, and, ideally, there should be no current flowing between the potential electrodes.
- This is accomplished either with a null-balancing galvanometer (old technology) or very high input impedance operational amplifiers.
- A few resistivity instruments have separate "sending" and "receiving" units for current and potential; but in usual practice, the potential measuring circuit is derived from the same source as the potential across the current electrodes, so that variations in the supply voltage affect both equally and do not affect the balance point.
- Power is usually supplied by dry cell batteries in the smaller instruments and motor generators in the larger instruments.
- From 90 V up to several hundred volts may be used across the current electrodes in surveys for engineering purposes.
- In the battery-powered units, the current usually is small and is applied only for very short times while the potential is being measured, so battery consumption is low.
- Care should be taken to NEVER energize the electrodes while they are being handled, because with applied potentials of hundreds of volts, DANGEROUS AND POTENTIALLY LETHAL shocks could be caused.

Measurements

- Current electrodes used with alternating current (or commutated direct current) instruments commonly are stakes of bronze, copper, steel with bronze jackets, or, less desirably, steel, about 50 cm in length.
- They must be driven into the ground far enough to make good electrical contact.
- If there is difficulty because of high contact resistance between electrodes and soil, it can sometimes be alleviated by pouring salt water around the electrodes.
- Many resistivity instruments include an ammeter to verify that the current between the current electrodes is at an acceptable level, a desirable feature.
- Other instruments simply output the required potential difference to drive a selected current into the current electrodes.
- Typical currents in instruments used for engineering applications range from 2 mA to 500 mA. If the current is too small, the sensitivity of measurement is degraded.
- The problem may be corrected by improving the electrical contacts at the electrodes.
- However, if the problem is due to a combination of high earth resistivity and large electrode spacing, the remedy is to increase the voltage across the current electrodes.
- Where the ground is too hard or rocky to drive stakes, a common alternative is sheets of aluminum foil buried in shallow depressions or within small mounds of earth and wetted.

Data Acquisition

- Resistivity surveys are made to satisfy the needs of two distinctly different kinds of interpretation problems:
- (1) the variation of resistivity with depth, reflecting more or less horizontal stratification of earth materials; and
- (2) lateral variations in resistivity that may indicate soil lenses, isolated ore bodies, faults, or cavities.
- For the first kind of problem, measurements of apparent resistivity are made at a single location (or around a single center point) with systematically varying electrode spacings.
- This procedure is sometimes called vertical electrical sounding (VES), or vertical profiling.
- Surveys of lateral variations may be made at spot or grid locations or along definite lines of traverse, a procedure sometimes called horizontal profiling.

Vertical Electrical Sounding (VES) - 1D Imaging

- Either the Schlumberger or, less effectively, the Wenner array is used for sounding, since all commonly available interpretation methods and interpretation aids for sounding are based on these two arrays.
- In the use of either method, the center point of the array is kept at a fixed location, while the electrode locations are varied around it.
- The apparent resistivity values, and layer depths interpreted from them, are referred to the center point.
- In the Wenner array, the electrodes are located at distances of a/2 and 3a/2 from the center point.
- The most convenient way to locate the electrode stations is to use two measuring tapes, pinned with their zero ends at the center point and extending away from the center in opposite directions.
- After each reading, each potential electrode is moved out by half the increment in electrode spacing, and each current electrode is moved out by 1.5 times the increment.
- The increment to be used depends on the interpretation methods that will be applied. In most interpretation methods, the curves are sampled at logarithmically spaced points.
- The ratio between successive spacings can be obtained from the relation

$$\frac{a_i}{a_{i-1}} = 10^{1/n},$$

where

n = number of points to be plotted in each logarithmic cycle.

Continue...

- VES surveys with the Schlumberger array are also made with a fixed center point.
- An initial spacing *s* (the distance from the center of the array to either of the current electrodes) is chosen, and the current electrodes are moved outward with the potential electrodes fixed.
- According to Van Nostrand and Cook (1966), errors in apparent resistivity are within 2 to 3 percent if the distance between the potential electrodes does not exceed $2s/5$.
- Potential electrode spacing is, therefore, determined by the minimum value of *s*.
- As *s* is increased, the sensitivity of the potential measurement decreases; therefore, at some point, if *s* becomes large enough, it will be necessary to increase the potential electrode spacing.
- The increments in *s* should normally be logarithmic and can be chosen in the same way as described for the Wenner array.
- For either type of electrode array, minimum and maximum spacings are governed by the need to define the asymptotic phases of the apparent resistivity curve and the needed depth of investigation.
- Frequently, the maximum useful electrode spacing is limited by available time, site topography, or lateral variations in resistivity.
- For the purpose of planning the survey, a maximum electrode spacing of at least three times the depth of interest may be used, but the apparent resistivity curve should be plotted as the survey progresses in order to judge whether sufficient data have been obtained.
- Also, the progressive plot can be used to detect errors in readings or spurious resistivity values due to local effects. Sample field data sheets are shown in figures 4 through 6.

Figure 4. Example data sheet for Schlumberger vertical sounding.

SCHLUMBERGER ELECTRICAL RESISTIVITY DATA SHEET

STATION NO. ____________ DIRECTION ____________ DATE ____________

PROJECT ________________________ LOCATION ______________________

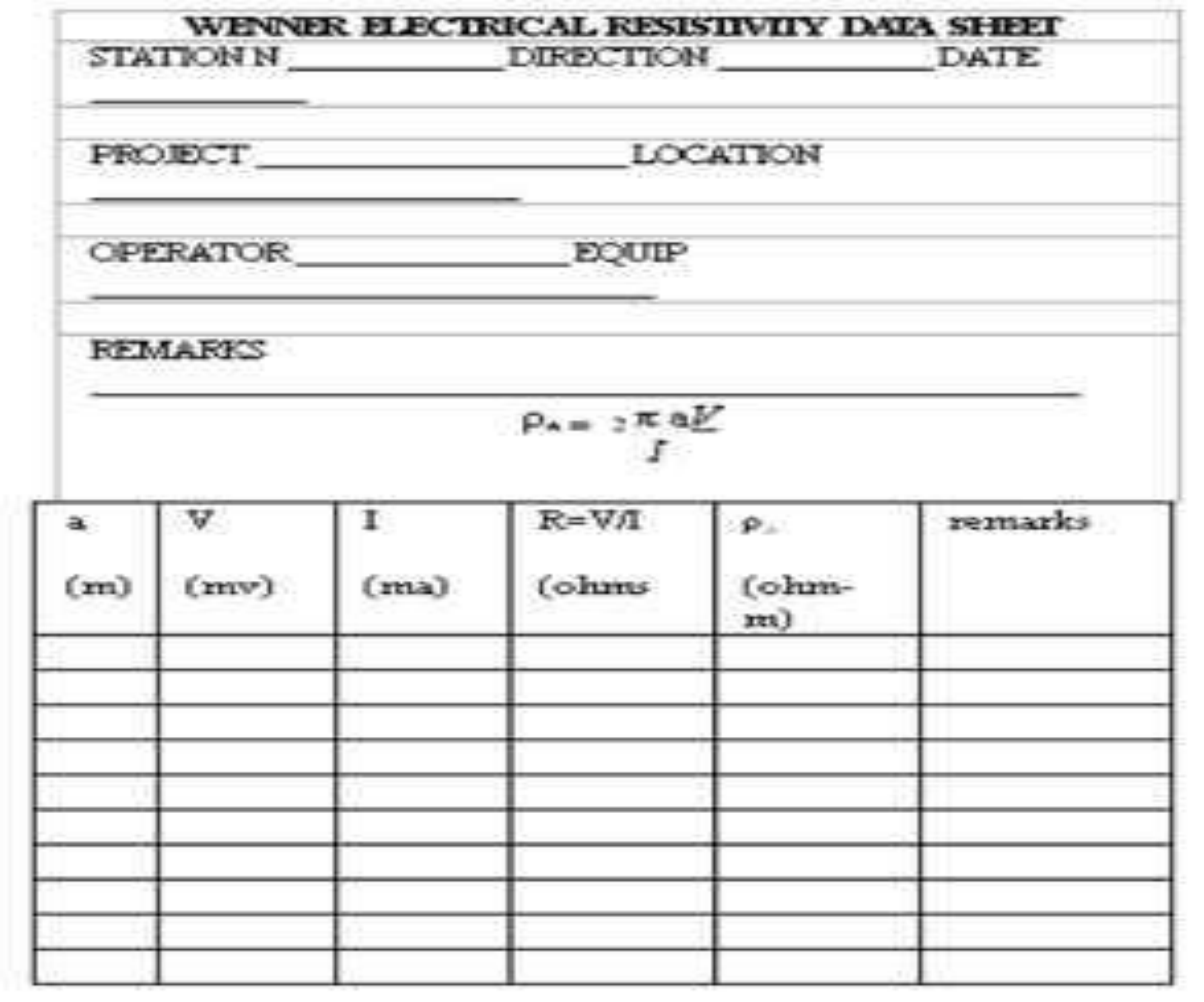
OPERATOR ____________ EQUIP ____________________________

REMARKS ____

$$\rho_a = \pi b\left[\left(\frac{a}{b}\right) - \frac{1}{4}\right]\frac{V}{I}$$

a (AB/2 (m)	b MN (m)	v (mv)	I (ma)	R=V/I (ohms)	ρa (ohm-m)	remarks

Figure 5. Example data sheet for Wenner array.



WENNER ELECTRICAL RESISTIVITY DATA SHEET

STATION N ______________ DIRECTION ______________ DATE ______________

PROJECT ________________________ LOCATION ____________________________

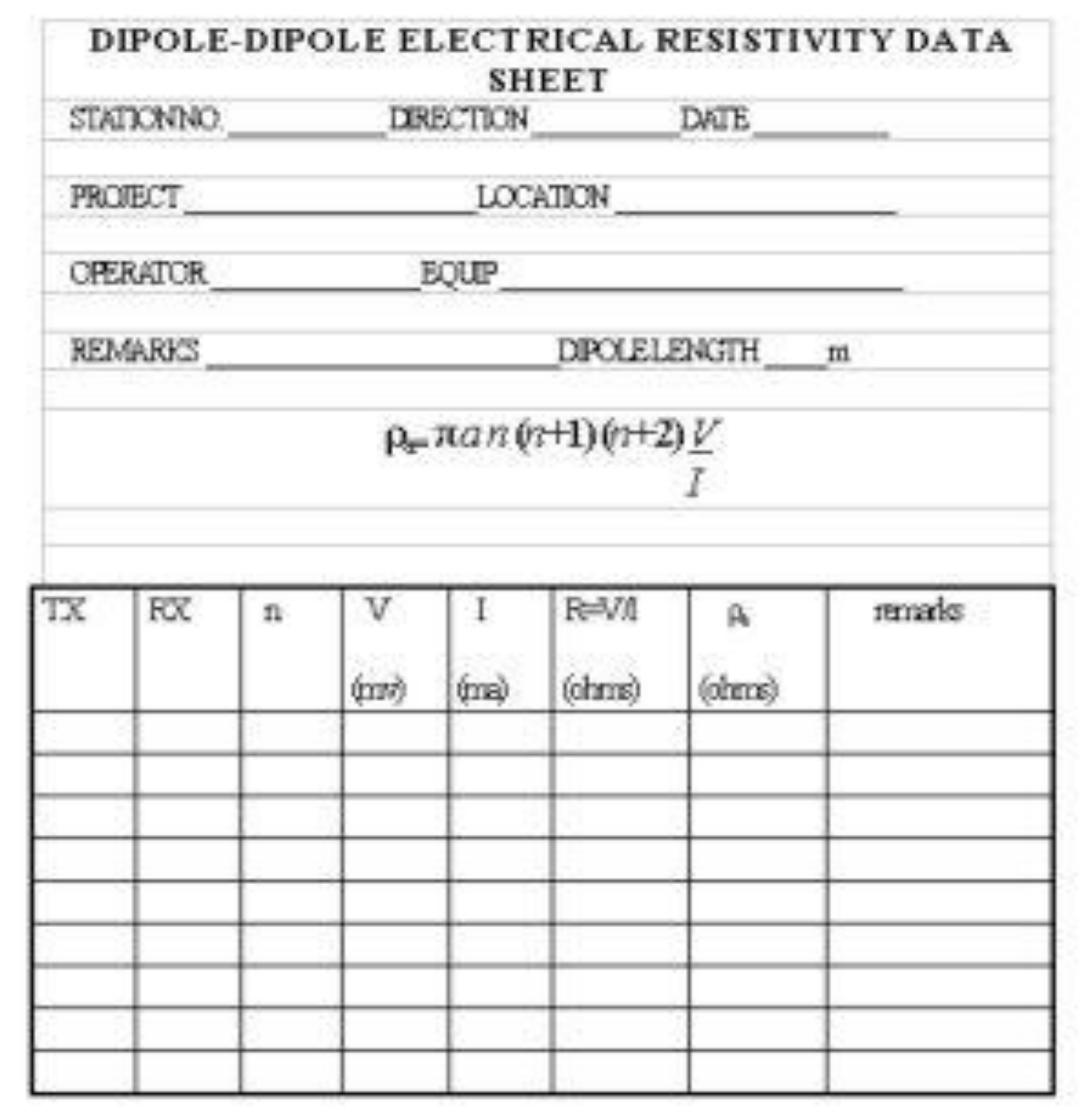
OPERATOR __________________ EQUIP ______________________________________

REMARKS ____

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

a (m)	V (mv)	I (ma)	R=V/I (ohms	ρ_a (ohm-m)	remarks

Figure 6. Example data sheet for dipole-dipole array



DIPOLE-DIPOLE ELECTRICAL RESISTIVITY DATA SHEET

STATION NO. ____________ DIRECTION ____________ DATE ____________

PROJECT ____________________ LOCATION ____________________

OPERATOR ______________ EQUIP ____________________________

REMARKS _________________________ DIPOLE LENGTH _____ m

$$\rho_a = \pi a n (n+1)(n+2) \frac{V}{I}$$

TX	RX	n	V (mv)	I (ma)	R=V/I (ohms)	ρ_a (ohms)	remarks

Interpretation of Vertical Electrical Sounding Data

- The interpretation problem for VES data is to use the curve of apparent resistivity versus electrode spacing, plotted from field measurements, to obtain the parameters of the geoelectrical section: the layer resistivities and thicknesses.
- From a given set of layer parameters, it is always possible to compute the apparent resistivity as a function of electrode spacing (the VES curve).
- Unfortunately, for the converse of that problem, it is not generally possible to obtain a unique solution.
- There is an interplay between thickness and resistivity; there may be anisotropy of resistivity in some strata; large differences in geoelectrical section, particularly at depth, produce small differences in apparent resistivity; and accuracy of field measurements is limited by the natural variability of surface soil and rock and by instrument capabilities.
- As a result, different sections may be electrically equivalent within the practical accuracy limits of the field measurements.

Horizontal Profiling - 1D Imaging

- Surveys of lateral variations in resistivity can be useful for the investigation of any geological features that can be expected to offer resistivity contrasts with their surroundings.
- Deposits of gravel, particularly if unsaturated, have high resistivity and have been successfully prospected for by resistivity methods.
- Steeply dipping faults may be located by resistivity traverses crossing the suspected fault line, if there is sufficient resistivity contrast between the rocks on the two sides of the fault.
- Solution cavities or joint openings may be detected as a high resistivity anomaly, if they are open, or low resistivity anomaly if they are filled with soil or water.
- Resistivity surveys for the investigation of aerial geology are made with a fixed electrode spacing, by moving the array between successive measurements.
- Horizontal profiling, per se, means moving the array along a line of traverse, although horizontal variations may also be investigated by individual measurements made at the points of a grid.
- If a symmetrical array, such as the Schlumberger or Wenner array, is used, the resistivity value obtained is associated with the location of the center of the array.
- Normally, a vertical survey would be made first to determine the best electrode spacing.
- Any available geological information, such as the depth of the features of interest, should also be considered in making this decision, which governs the effective depth of investigation.
- The spacing of adjacent resistivity stations, or the fineness of the grid, governs the resolution of detail that may be obtained.
- This is very much influenced by the depths of the features, and the achievable resolution diminishes with depth.
- As a general rule, the spacing between resistivity stations should be smaller than the width of the smallest feature to be detected, or smaller than the required resolution in the location of lateral boundaries.

Interpretation of Horizontal Profiling Data

- Data obtained from horizontal profiling for engineering applications are normally interpreted qualitatively.
- Apparent resistivity values are plotted and contoured on maps, or plotted as profiles, and areas displaying anomalously high or low values or anomalous patterns are identified.
- Interpretation of the data, as well as the planning of the survey, must be guided by the available knowledge of the local geology.
- The interpreter normally knows what he is looking for in terms of geological features and their expected influence on apparent resistivity, because the resistivity survey is motivated by geological evidence of a particular kind of exploration problem (e.g., karst terrain).
- The survey is then executed in a way that is expected to be most responsive to the kinds of geological or hydrogeological features sought.
- A pitfall inherent in this approach is that the interpreter may be misled by his preconceptions if he is not sufficiently alert to the possibility of the unexpected occurring.
- Alternative interpretations should be considered, and evidence from as many independent sources as possible should be applied to the interpretation.
- One way to help plan the survey is to construct model VES sounding curves for the expected models, vary each model parameter separately by say 20%, and then choose electrode separations that will best resolve the expected resistivity/depth variations.
- Most investigators then perform a number of VES soundings to verify and refine the model results before commencing horizontal profiling.
- The construction of theoretical profiles is feasible for certain kinds of idealized models, and the study of such profiles is very helpful in understanding the significance of field profiles.
- Van Nostrand and Cook (1966) give a comprehensive discussion of the theory of electrical resistivity interpretation and numerous examples of resistivity profiles over idealized models of faults, dikes, filled sinks, and cavities.

MAGNETIC PROSPECTING

Contents:

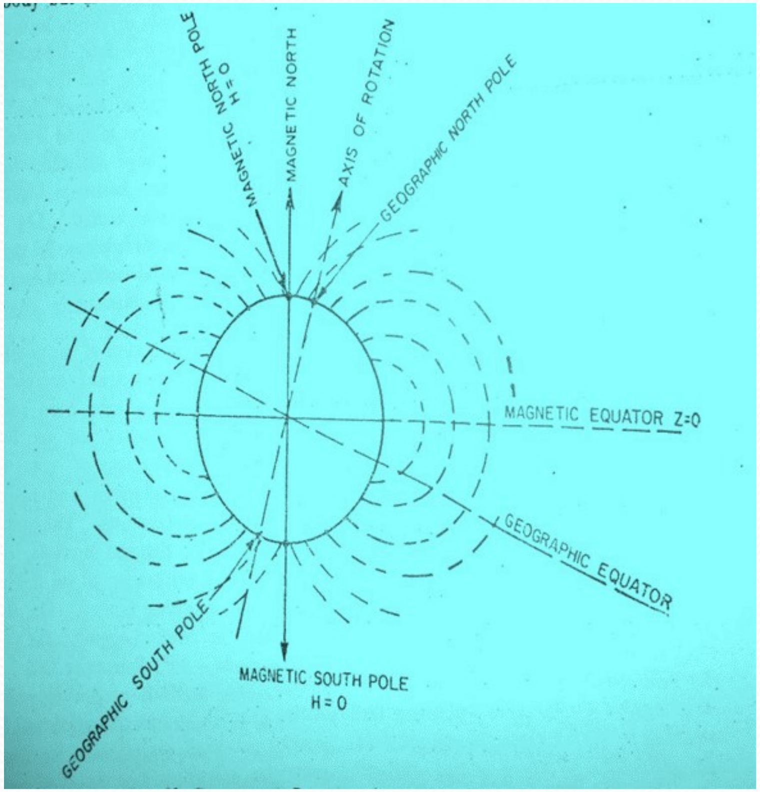
- Introduction
- Earth's Magnetism
- Components of earth's magnetic field
- Magnetic Poles
- Magnetism of rocks and minerals
- Instruments
- Field Procedures
- Applications
- Conclusion
- References

Introduction:

- Magnetic prospecting is a geophysical method based on the examination of magnetic field anomalies of the Earth caused by non-similar magnetization intensity of rocks.
- The magnetic field of the Earth magnetizes rocks to a variable degree which is determined by their magnetic susceptibility, intensity and magnetizing force.
- It is based on magnetic anomaly observed in the field on paramagnetic/ferromagnetic minerals like magnetite, pyrrhotite, franklinite, cobaltite and diamagnetic minerals like rock salt , quartzite etc.

Earth’s Magnetism:

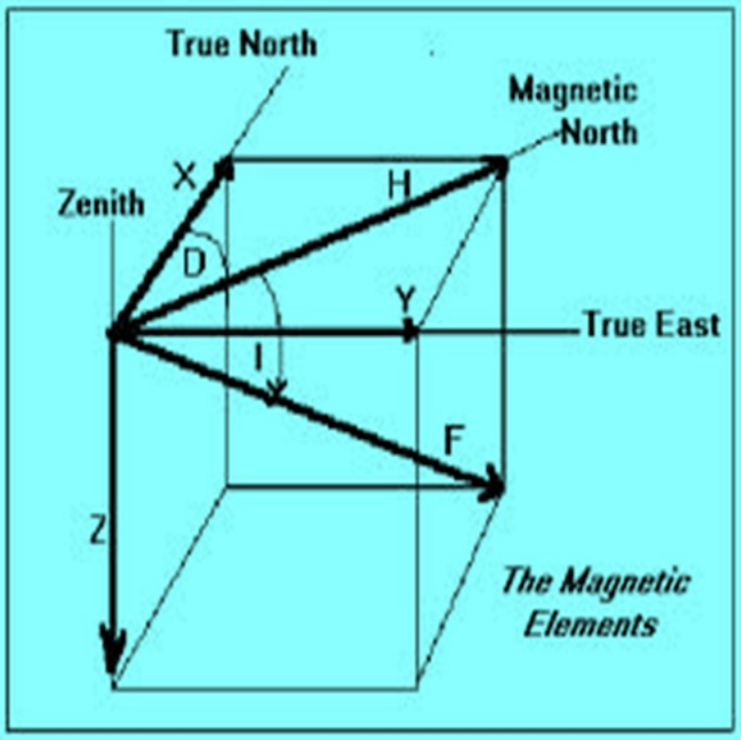
- The earth behaves like a large magnet and a magnetic field not only exists in its body but also surrounds it far in space.
- The intensity of the field varies on the surface of the earth form place to place with latitude, and geological factors.
- The total intensity of the earth’s field is about 70,000 gammas at the magnetic poles and 39,000 gammas at the magnetic equators.
- The intensity of the field also varies with respect to time.



Diagrammatic representation of the Earth’s Magnetic field

Components of Earth’s Magnetic Field:

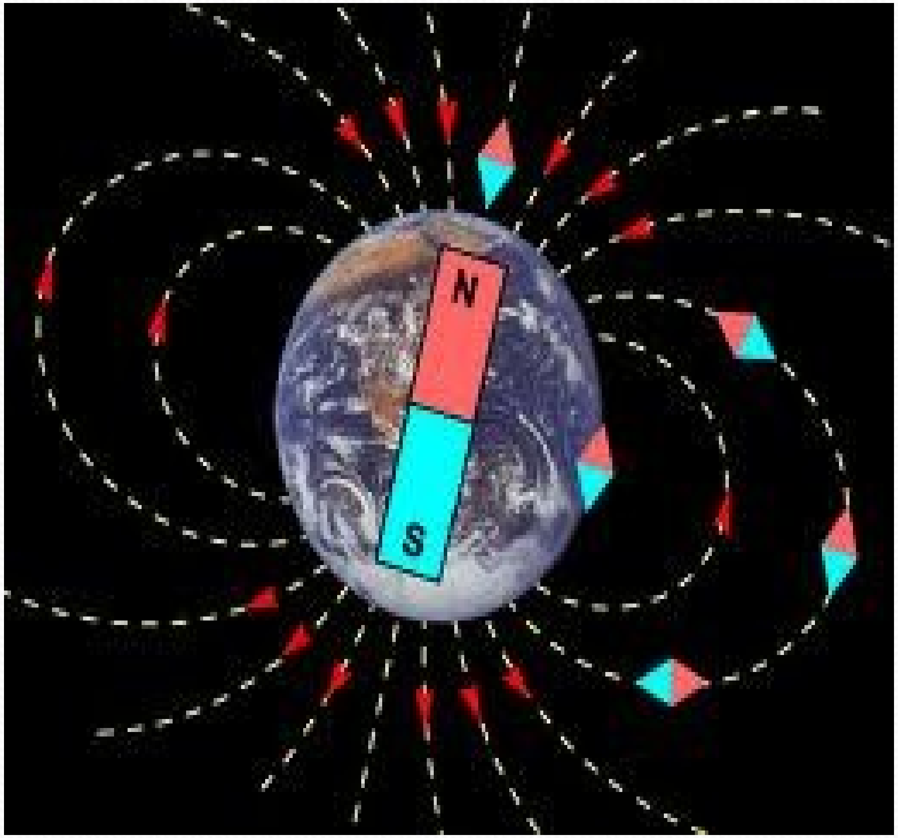
- **Declination (D)**:-It indicates the difference in degrees, between the true north and magnetic north.
- **Inclination(I):-** It is the angle, in degrees, of the magnetic field above or below horizontal.
- **Horizontal intensity(H):-** It defines the horizontal component of the total field intensity.
- **Vertical intensity (Z):-** It defines the vertical component of the total field intensity.
- **Total intensity(T):-** It is the strength of the magnetic field, not divided into its component parts $T^2=Z^2+H^2$.



Vector diagram of the Earth’s magnetic field components (in Northern Hemisphere).

Magnetic Poles:

- The magnetic force surrounding a magnet is not uniform.
- There exists a great concentration of force at each end of the magnet and a very weak force at the center.
- The two ends which are the regions of concentrated lines of force, are the poles of the magnet.
- Magnet have two magnetic poles and both have equal magnetic strength.

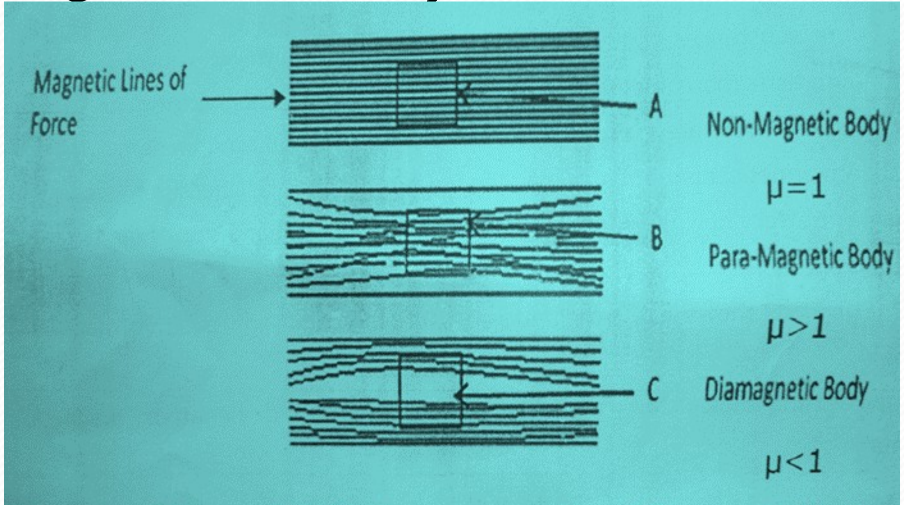


A bar magnet showing the two magnetic poles and the lines of flux

Magnetism of Rocks and Minerals:

- Rocks and minerals exhibit two types of magnetism.
  - **Induced**:- Induced magnetism is that which is acquired by a body when subjected to a magnetizing force. The intensity of magnetism acquired depends not only on the nature of the material composing the body, but also on the strength of the magnetizing force.
  - **Remnant:-** Some substances containing Iron, Nickel, etc are however known to retain their magnetism more or less permanently, even after magnetizing force has been removed. They therefore said to posses remnant magnetism.

- Based on magnetic susceptibility magnetic materials are classified into:-
  - **Ferromagnetic:-**shows very high magnetic susceptibility and very high positive magnetic anomaly.
  - **Paramagnetic:-** shows high magnetic susceptibility and high positive magnetic anomaly.
  - **Diamagnetic:-** shows low magnetic susceptibility and negative magnetic anomaly.



Distortion of Magnetic fields by paramagnetic and diamagnetic bodies.

- **Magnetic permeability(μ):-** is a measure of modification by induction of the force of attraction or repulsion between two magnetic poles.
- **Magnetic susceptibility(k):-** is a measure of the magnetic response of a material to an external magnetic field.

Magnetic susceptibilities of some minerals:

Mineral	Susceptibilities k*10(c.g.s.units)
Paramagnetic	
Magnetite	200,000 to 1,300,000
Illimnite	135,000 to 252,000
Pyrrhotite	125,000
Franklinite	36,000
Jacobsite	2,000
Hematite(specular)	426-3,200
Hematite(amorphous)	40-528

Contd…

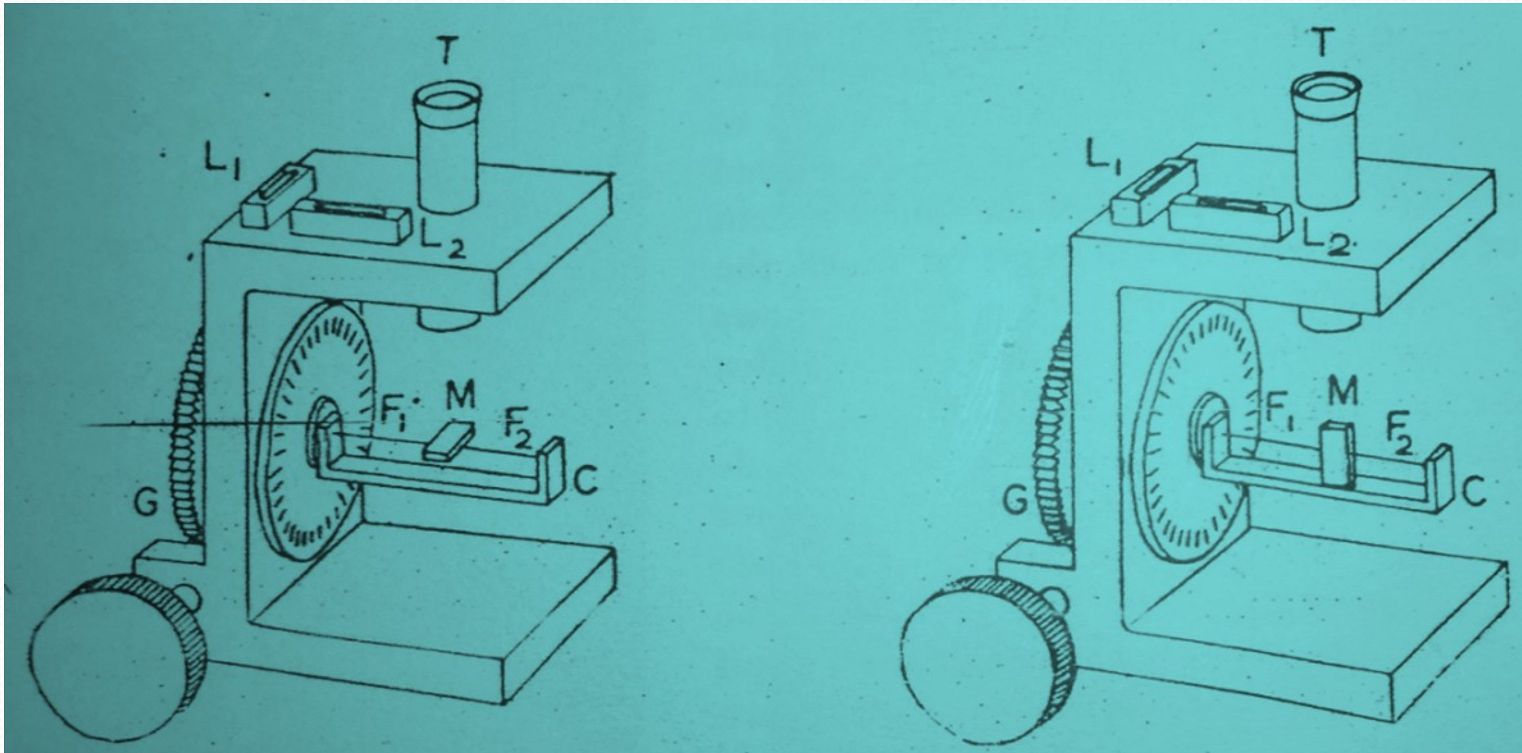
limonite	**57-220**
Serpentine	1,300
Chromite	244-9,400(depending on fe content)
Hausmannite	132-318
Pyrite	4-420
Chalcopyrite	32

Diamagnetic:

Calcite	**-0.6 to -1.04**
Diamond	-1.80
Graphite	-8.00
Quartz	-1.07 to-1.2
Rock salt	-0.82 to -1.3
Anhydrite	-1.1 to -1.2

Instruments:

- A magnetometer is a scientific Instrument used to measure the strength and direction of the magnetic field in the vicinity of the instrument. Magnetic field varies from place to place and differences in earth's magnetic field.
- Magnetometer are often a frequent component instrument on spacecraft that explore planets.
- Schmidt type magnetometer
- Torsion magnetometer
- Flux gate magnetometer
- Nuclear or the proton magnetometer
- Rubidium vapors magnetometer



Torsion Magnetometer (Vertical force)

(b) Torsion Magnetometer (Horizontal force)

FIG. 91. Torsion Magnetometers
(Askania Werke)

*From Notes kindly furnished by Shri G. Varadarajan.

Torsion magnetometer

Contd…

- Highly sensitivity Torsion magnetometers for measuring the vertical and the horizontal components of the earth’s magnetic field.
- Magnet is suspended by a torsion fibre. The earth’s magnetic field deflects the magnet form its null position and a restoring torque applied to the same torsion fiber helps to bring the magnet back into the null position.
- The extend of the restoring torque applied gives a measure of the intensity of the earth’s magnetic field.

Field Procedures:

- In the case of vertical magnetometers the instrument is set at right angle to the magnetic meridian and at least two reading are taken for each setting, by rotating the system through 180^0.
- In the case of horizontal magnetometers , the instrument is set parallel to the magnetic meridian, so that the poles of magnet system are free to move .the reading is taken only once in a position . Because rotating through 180^0 will not give any variation in reading.

Applications:

- Exploration of magnetic ores of iron
- In mining field magnetic prospecting may be applied.

  1. directly in the search for deposits of highly magnetic minerals such as magnetite, pyrrhotite and some of the manganese ores.
  2. indirectly for locating bodies of other minerals which themselves may not be magnetic minerals.
- In petroleum exploration, for determination of thickness of sediments .
- In engineering projects for locating construction materials such as granite, basalt and other building stones

Conclusion:

- Magnetic prospecting is the oldest geophysical methods employed for exploration of magnetite deposits . It has come a long way in terms on accuracy and precision, from the dip needle to the latest rubidium and cesium vapour magnetometer .
- Magnetic prospecting is very successful method in directly locating iron, manganese and pyrrhotite deposits, and indirectly in locating places of gold, diamond, quartz and bauxite .
- It is also useful in ascertaining the depth of sedimentary basins and the thickness of sedimentary strata.

References:

- Milton B D (1998), Introduction to Geophysical Prospecting, 4th edition, Mc Graw-Hill International Editions- Geology series, pg. 476-519
- Ramakrishna T S (2006), Geophysical Practice in Mineral Exploration and Mapping, Memoir 62, Geological Society of India, Bangalore, pg 34-46.
- Ramachandra Rao M.B. (1975), Outlines of Geophysical Prospecting – A Manual for Geologist, pg. 120-169
- http://wikipedia.org
- http://wikimediafoundation.org

Gravity Method

Contents

- Definition
- Branches of Geophysics
- Scope of Geophysics
- Geophysical Methods
    - Gravity Method
        - Gravity Anomalies
        - Gravity Reductions
- Applications of Gravity Method

Geophysics (Definition)

- Geophysics is the application of method of physics to the study of the Earth.
- On the other sense, it is a subject of natural science concern with the physical processes and the physical properties of the earth and it's surrounding space environment and the use of co-ordinary methods for the analysis.
- It involves the application of physical theories and measurements to discover the properties and processes of the earth.

Geophysics

- Geophysics has contributed significantly in the understanding of many physical processes that lead to scientific and economic contribution to our society.

Branches of Geophysics

- **Solid Earth Geophysics :**
  - Earthquakes, Tsunamis, Tectonics
  - Geodynamics

- **Exploration Geophysics :**
  - Oil and Gas exploration
  - Minerals exploration

- **Environmental & Engineering Geophysics :**
  - Groundwater exploration
  - Contaminant delineation
  - Utility or object detection

Scopes of Geophysics

➢ **Natural hazard studies:**

- Earthquake
- Landslide

➢ **Resource explorations:**

- Oil and gas exploration
- Mineral prospecting
- Geothermal exploration
- Groundwater exploration

Scopes of Geophysics(Cont.)

- **Engineering:**
  - Underground utility locating
  - Concrete inspection
  - Rebar locating
  - Pavement evaluation
  - Underground void locating
  - Ground strength testing

- **Environmental application:**
  - Underground storage tank locating
  - Contamination delineation
  - Landfill delineation
  - Bedrock depth mapping

Geophysical Methods

➢**Potential Fields:**

- Gravity
- Magnetics

➢**Diffusive Fields:**

- Electrical
- Heat Flow
- Electromagnetics(EM)

➢**Wave Propagation:**

- Seismic (Sound Waves)
- Radar (EM Waves)

Geophysical methods

- **Gravity method:**

- Gravity method is a non-destructive geophysical technique that measures differences in the earth's gravitational field at specific locations.
- The gravity method is a relatively cheap, non-invasive, non-destructive remote sensing method.

Gravity method

- In gravity surveying, subsurface geology is investigated on the basis of variations in the Earth's gravitational field arising from differences of density between subsurface rocks.
- An underlying concept is the idea of a **causative body,** which is a rock unit of different density from its surroundings.

Gravity method

- The basis on which the gravity method depends is encapsulated in two laws derived by Newton, namely his **Universal Law of gravitation** and his **Second Law of Motion**.

$$F = G\frac{Mm}{R^2} \;\ldots\ldots\ldots\ldots(1) \qquad F = mg \;\ldots\ldots\ldots(2)$$

$$g = \frac{GM}{R^2} \;\ldots\ldots\ldots\ldots(3)$$

This shows that the magnitude of acceleration due to gravity on Earth (g) is directly proportional to the mass (M) of the Earth and inversely proportional to the square of the Earth's radius (R)

Gravity method

Units of gravity

- The mean value of gravity at the Earth's surface is about 9.8ms-2.Variations in gravity caused by density variations in the subsurface are of the order of 100 mms-2.
- This unit of the micrometer per second per second is referred to as the *gravity unit (gu). In gravity surveys on land an* accuracy of ±0.1 gu is readily attainable, corresponding to about one hundred millionth of the normal gravitational field.
- At sea the accuracy obtainable is considerably less, about ±10 gu.The c.g.s. unit of gravity is the *milligal* (1 mgal = 10-3 gal = 10-3cms-2), equivalent to 10 gu.

Gravity method (Process)

- Gravity techniques measure ***minute variations in the earth's gravity field***. Based on these variations, subsurface density and thereby composition can be inferred.
- These variations can be determined by measuring the earth's gravity field at numerous stations along a traverse, and correcting the gravity data for ***elevation, tidal effects, topography, latitude***, and ***instrument drift***.

Gravity method (Process)

- The gravity field on the surface of the Earth is not uniformly the same everywhere. It varies with the distribution of the mass materials below. A Gravity survey is an direct means of calculating the density property of subsurface materials.
- ***The higher the gravity values, the denser the rock beneath.***

Gravity method (Equipment)

- Modern instruments capable of rapid gravity measurements are known as ***gravity meters*** or ***gravimeters***. Gravimeters are basically spring balances carrying a constant mass. Variations in the weight of the mass caused by variations in gravity cause the length of the spring to vary and give a measure of the change in gravity.

Gravity method (Equipment)

- There are ***two types*** of gravimeters:
  1. ***Relative*** and
  2. ***Absolute***.

Absolute gravimeters measure the ***local gravity in absolute units***, ***gals***.

Relative gravimeters compare the value of gravity at one point with another. They must be calibrated at a location where the gravity is known accurately, and then transported to the location where the gravity is to be measured. They measure the ratio of the gravity at the two points.

Gravity Anomalies

- A causative body represents a subsurface zone of anomalous mass and causes a localized perturbation in the gravitational field known as a gravity anomaly.
- Gravity anomaly map yield the difference between the observed gravity values and the theoretical gravity values for a region of interest.

Gravity Anomalies(Cont.)

- A very wide range of geological situations give rise to zones of anomalous mass that produce significant gravity anomalies.
- On a small scale, buried relief on a bedrock surface, such as a buried valley, can give rise to measurable anomalies.

Gravity Anomalies(Cont.)

- Depending on what we want to emphasize there are **3** anomaly map:
- Free-air or Faye anomaly
- Bouguer anomaly and
- Isostatic gravity anomaly maps

☐ **Faye anomaly:**

- Is defined by applying only normal, free-air, terrain and tidal corrections to the measured gravity value.

Gravity anomalies(Cont.)

- ☐ **Bouguer anomaly:**
- Is defined by applying normal, free-air, terrain and tidal correction to the measured gravity value.
- The difference between the Bouguer and the Faye anomaly arises from the Bouguer plate correction.
- Bouguer anomalies are usually negative in region of large elevation and are mainly in positive in oceanic regions.

Gravity anomalies(Cont.)

- ☐ **Isostatic gravity anomaly:**
- Is defined by applying isostatic correction to the Bouguer anomaly.

Gravity reduction

Before the results of a gravity survey can be interpreted it is necessary to correct for all variations in the Earth’s gravitational field which do not result from the differences of density in the underlying rocks. This process is known as ***gravity reduction***.

Gravity reduction

- **Drift correction**

Correction for instrumental drift is based on repeated readings at a base station at recorded times throughout the day. The meter reading is plotted against time and drift is assumed to be linear between consecutive base readings.

Drift Correction

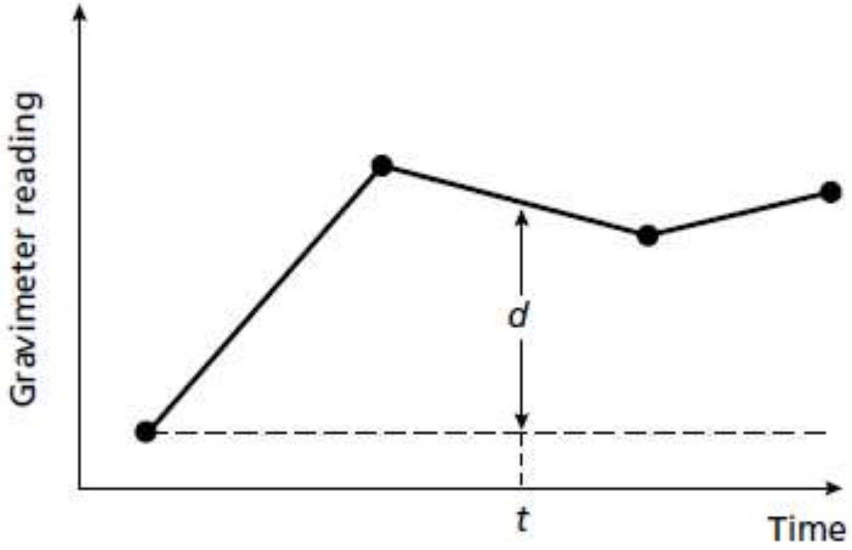


Fig. 6.10 A gravimeter drift curve constructed from repeated readings at a fixed location. The drift correction to be subtracted for a reading taken at time *t* is *d*.

From the figure drift is assumed to be linear between consecutive base readings. The drift correction at time *t is d, which is subtracted from the observed value.*

Eötvös correction

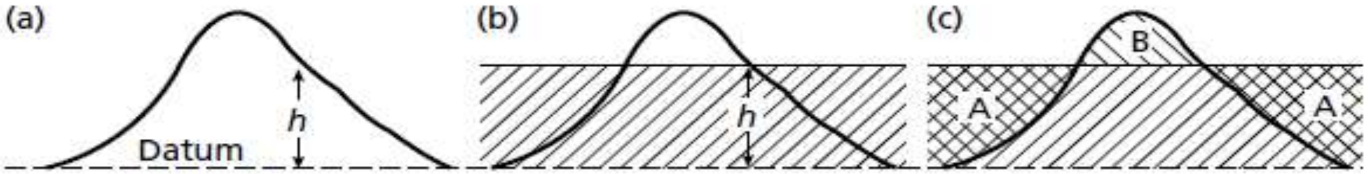
The Eötvös correction (EC) is applied to gravity measurements taken on a moving vehicle such as a ship or an aircraft. Depending on the direction of travel, vehicular motion will generate a centripetal acceleration which either reinforces or opposes gravity.

Gravity reduction

Elevation corrections

Correction for the differing elevations of gravity stations is made in three parts :

- Free air Correction
- Bouger Correction
- Terrain Correction



. 6.12 (a) The free–air correction for an observation at a height *h* above datum. (b) The Bouguer correction. The shaded region responds to a slab of rock of thickness *h* extending to infinity in both horizontal directions. (c) The terrain correction.

Elevation Correction

- **The free-air correction (FAC)** corrects for the decrease in gravity with ***height in free air*** resulting from increased distance from the centre of the Earth. The FAC is positive for an observation point above datum to correct for the decrease in gravity with elevation. The free-air correction accounts solely for variation in the distance of the observation point from the centre of the Earth; no account is taken of the gravitational effect of the rock present between the observation point and datum.

Elevation Correction

Bouguer correction

On land the Bouguer correction must be subtracted, as the gravitational attraction of the rock between observation point and datum must be removed from the observed gravity value. The Bouguer correction of sea surface observations is positive to account for the lack of rock between surface and sea bed. The correction is equivalent to the replacement of the water layer by material of a specified rock density r.

Elevation Correction

Terrain corrections

A correction applied to observed values obtained in geophysical surveys in order to remove the effect of variations in the observations due to the topography near observation sites.

Latitude Correction

Gravity varies with latitude because of the non-spherical shape of the Earth and because the angular velocity of a point on the Earth’s surface decreases from a maximum at the equator to zero at the poles. The centripetal acceleration generated by this rotation has a negative radial component that consequently causes gravity to decrease from pole to equator. Consequently, points near the equator are farther from the centre of mass of the Earth than those near the poles, causing gravity to increase from the equator to the poles.

Latitude Correction

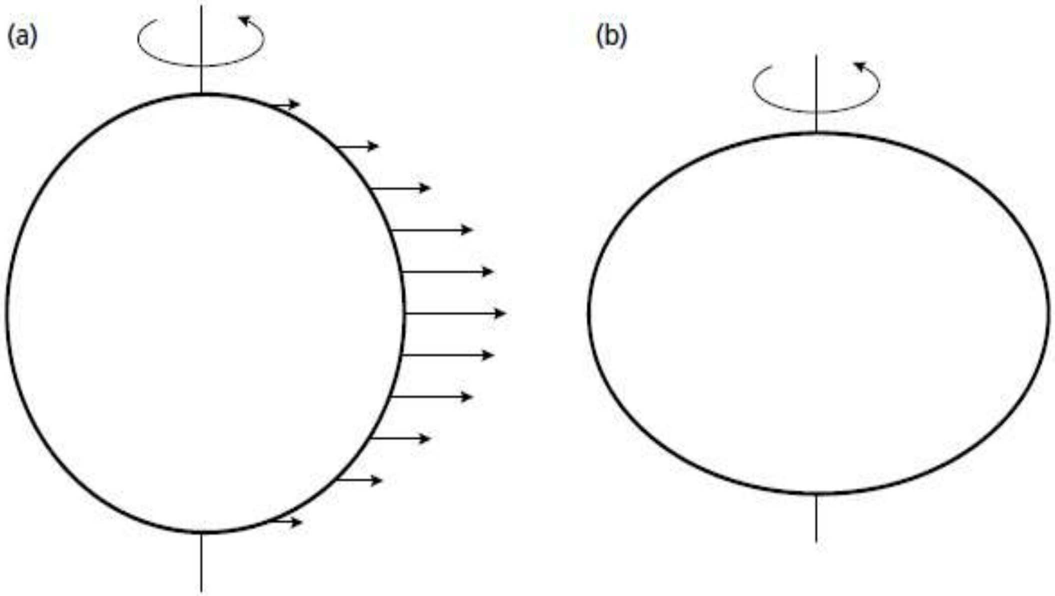


Fig. 6.11 (a) The variation in angular velocity with latitude around the Earth represented by vectors whose lengths are proportional to angular velocity. (b) An exaggerated representation of the shape of the Earth. The true shape of this oblate ellipsoid of revolution results in a difference in equatorial and polar radii of some 21 km.

Tidal Correction

Gravity measured at a fixed location varies with time because of periodic variation in the gravitational effects of the Sun and Moon associated with their orbital motions, and correction must be made for this variation in a high precision survey. In spite of its much smaller mass, the gravitational attraction of the Moon is larger than that of the Sun because of its proximity.

Tidal Correction

- These *solid Earth tides* are considerably smaller than oceanic tides and lag farther behind the lunar motion. They cause the elevation of an observation point to be altered by a few centimeters and thus vary its distance from the centre of mass of the Earth. The periodic gravity variations caused by the combined effects of Sun and Moon are known as *tidal variations*.
- They have a maximum amplitude of some 3 gu and a minimumperiod of about 12 h.

Application of Gravity method

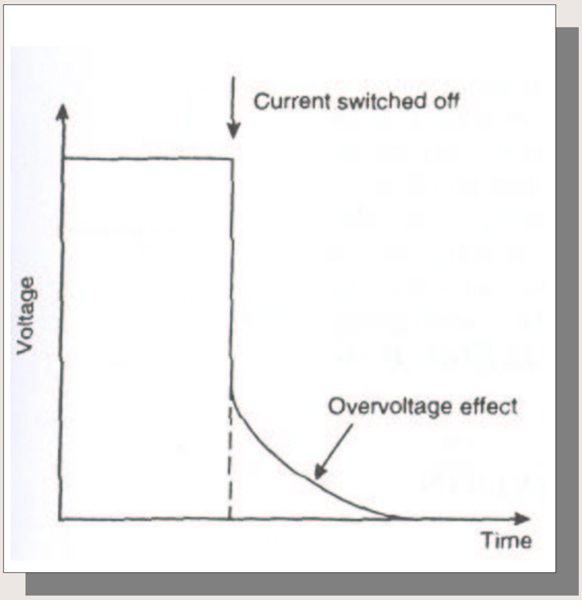
- Determine shape of the Earth
- Hydrocarbon exploration
- Regional geological studies
- Iso-static compensation determination
- Detection of sub-surface cavities (microgravity)
- Location of buried rock-valleys
- Determination of glacier thickness
- Tidal oscillations
- Basin Geometry

Induced Polarization Method

- Principles;
- Areas of application;
- Measurement;
- Equipment and layout;
- Interpretation;
- Case histories.
- <u>Reading:</u>
  - Reynolds, Chapter 9.
  - Telford et al., Chapter 9.

Induced Potentials

- After current is switched off (or turned on), the voltage between potential electrodes takes 1s - 1 min to decay (or build up)
  - The ground acts somewhat like a capacitor.
- Overvoltage decay times and rise times are measured and are diagnostic of the nature of the subsurface.
- Applications:
  - Metallic deposits with low EM anomalies and high resistivity;
  - *Disseminated* Cu, Pb-Zn ores, Au;
  - Pyrite, chalcopyrite, magnetite, clay, graphite.



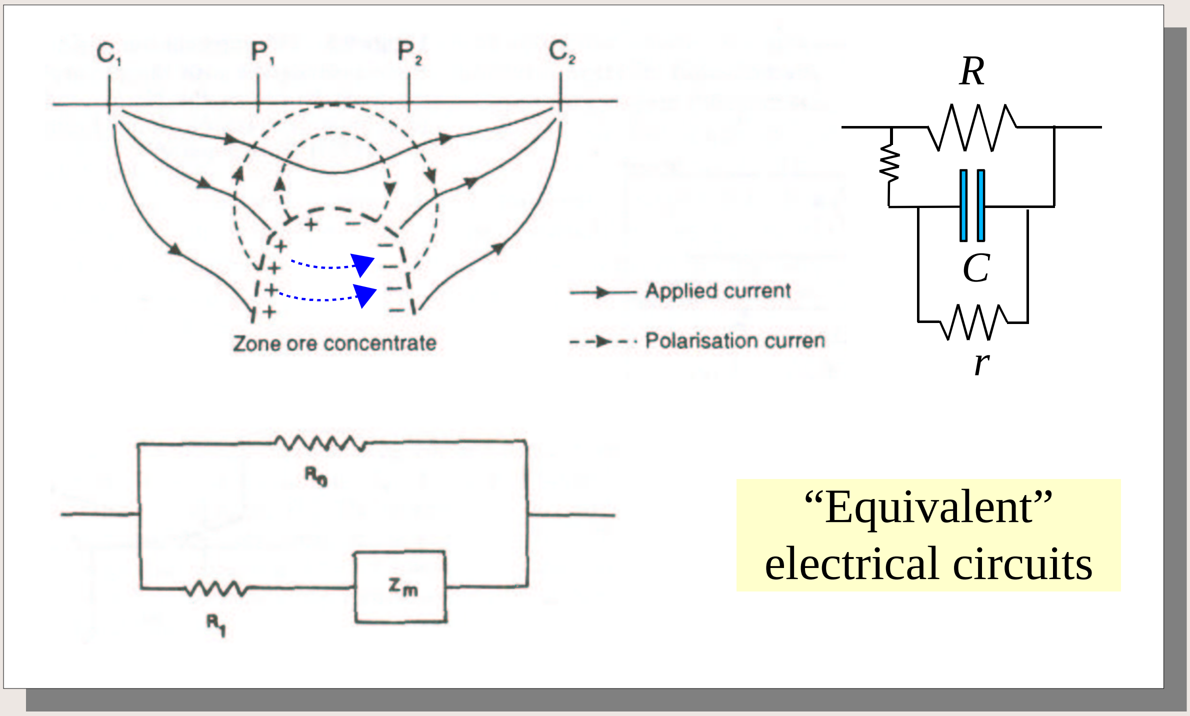
IP Techniques

- *Time domain* (pulse transient);
- *Frequency domain* (using harmonic signals):
  - *Traditional* variable-frequency IP (using two or more frequencies of < 10 Hz);
  - *Phase domain* (measure phase delays between current and voltage);
  - *Spectral IP* (measure phases and amplitudes at frequencies 10^{-3} to $4 \cdot 10^{3}$ Hz).
- Using conventional resistivity arrays
  - Most commonly *double-dipole* configuration;
  - Schlumberger arrays for broad reconnaissance surveys.

Origin of IP

Macroscopic

- IP s sensitive to *dielectric* rater than conductivity characteristics.
- Disseminated (poorly conductive) ore body is polarized (develops *surface charges*) by the imposed current;
- When the current is switched off, the charges cause transient current through the conductive overburden.
  - These currents flow *in the same direction* and cause the overvoltage effect.

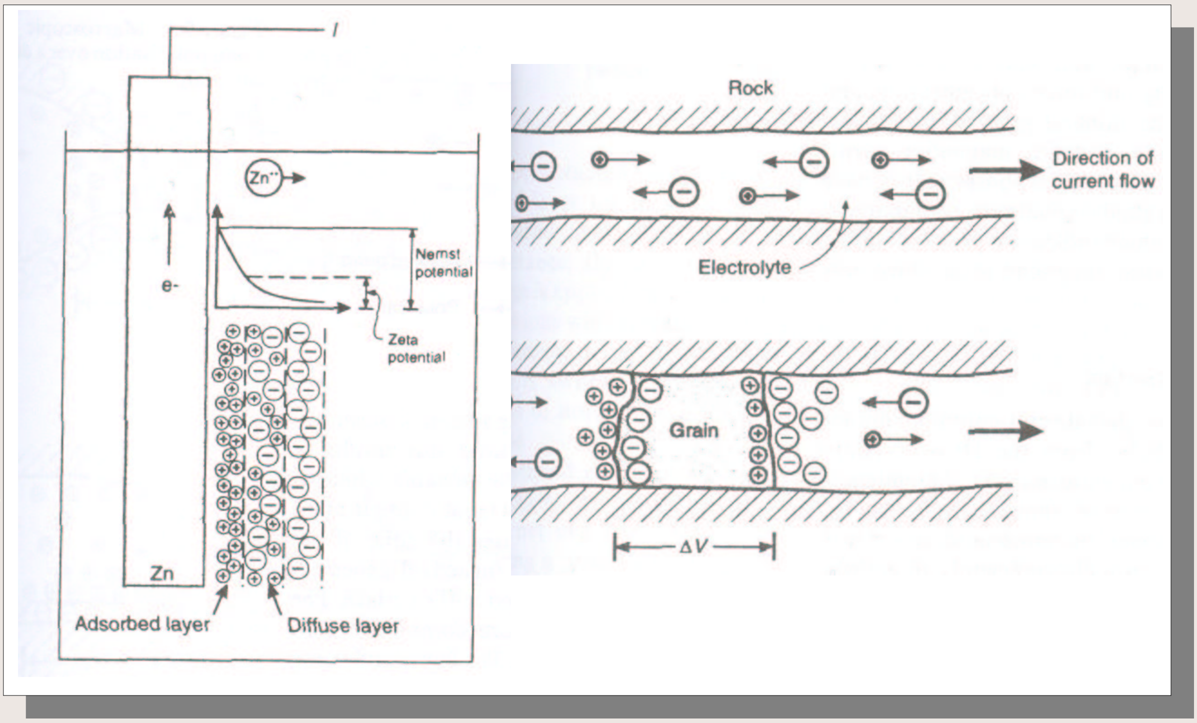


“Equivalent” electrical circuits

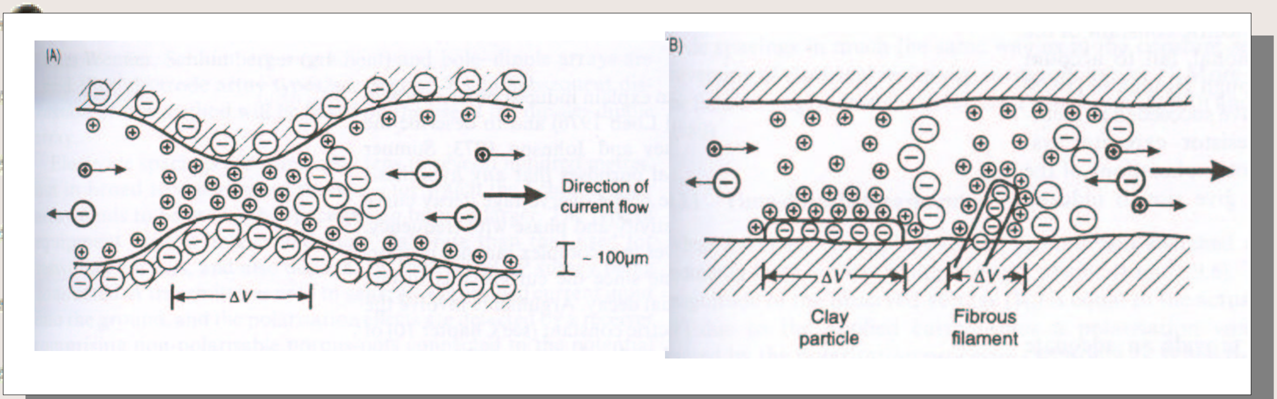
Origin of IP

Microscopic

- Grain (and electrode) polarization:

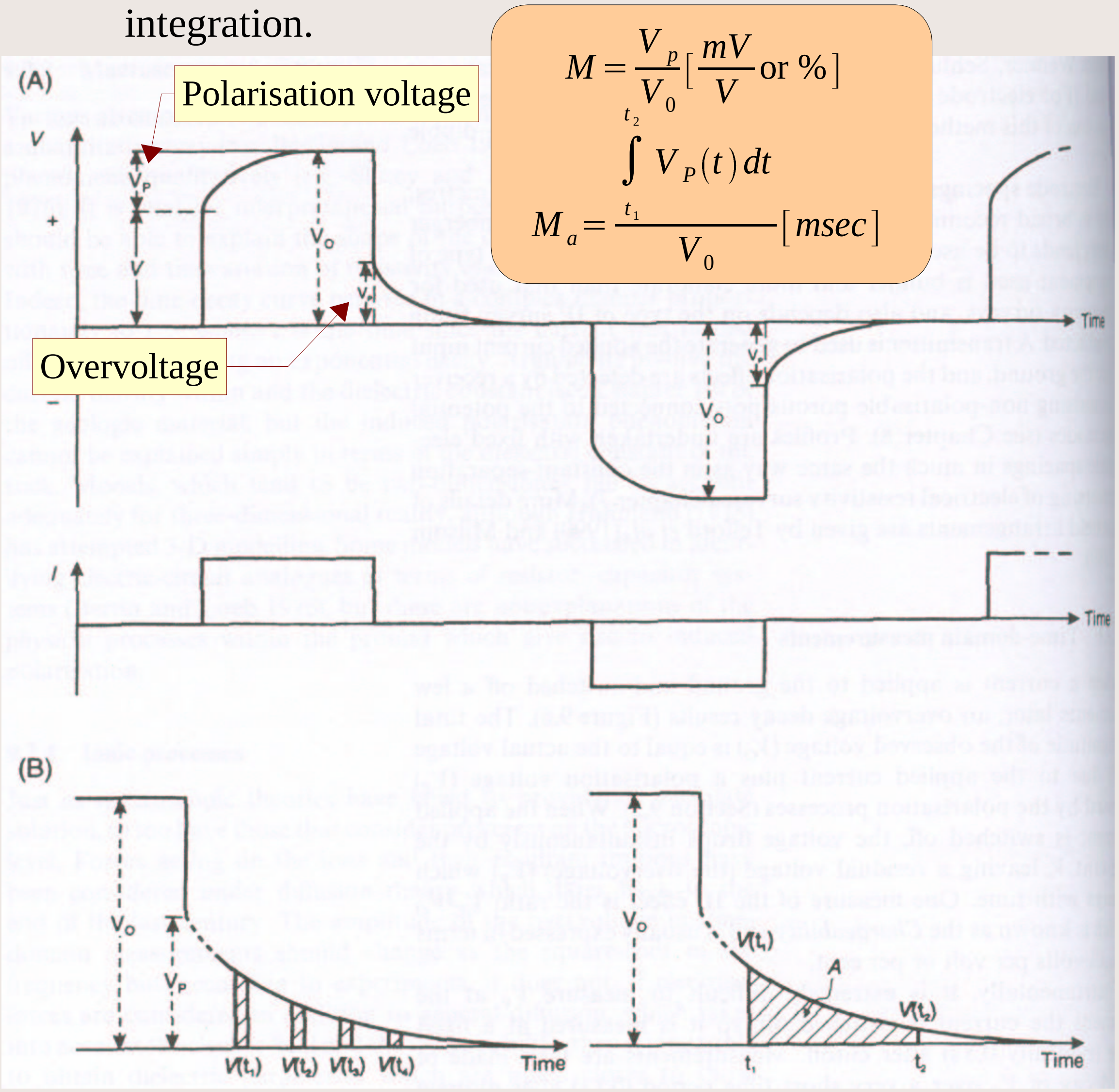


- Electrolytic (membrane) polarization:



Time-domain IP

- Measuring *apparent chargeability (M)*
  - Apparent chargeability (M_a) increases with increasing duration of the pulses (~ 3-5 s);
  - Graphite has M_a.=11.2 ms, magnetite - 2.2 ms at 1 s integration.



$$M = \frac{V_p}{V_0}\left[\frac{mV}{V}\text{ or }\%\right]$$

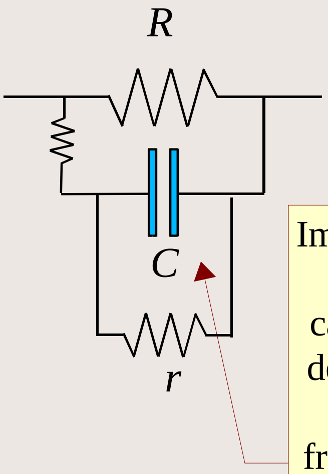
$$M_a = \frac{\int_{t_1}^{t_2} V_P(t)\,dt}{V_0}[msec]$$

Polarisation voltage

Overvoltage

Variable-frequency IP

- Using the same array as in DC resistivity measurements but driving AC current at several frequencies.
- Measuring ρ_a*(frequency)*:
    - ρ_a *decreases* with frequency;
    - This decrease is measured as the *Frequency Effect* (FE):



Impedance of the capacitor decreases with frequency: $Z=1/i\omega C$; hence the total resistance *decreases.*

$$Frequency\,Effect = \frac{\rho_a(f_0)-\rho_a(f_1)}{\rho_a(f_1)} \quad [\text{unitless or }\%].$$

- FE can also be expressed as the *Metal Factor* (variation of apparent *conductivity*):

$$Metal\,Factor = 2\times10^5\pi\,\frac{\rho_a(f_0)-\rho_a(f_1)}{\rho_a(f_0)\rho_a(f_1)} = 2\times10^5\pi\,(\sigma_a(f_1)-\sigma_a(f_0))$$

$[\text{siemens/m}]$.

Spectral (complex resistivity) IP

- Using AC current at *a range* of frequencies from 30 to 4000 Hz.
- Measuring complex *impedance*:

$$Z(\omega)=\frac{U(\omega)}{I(\omega)}K.$$

Geometric factor of the array

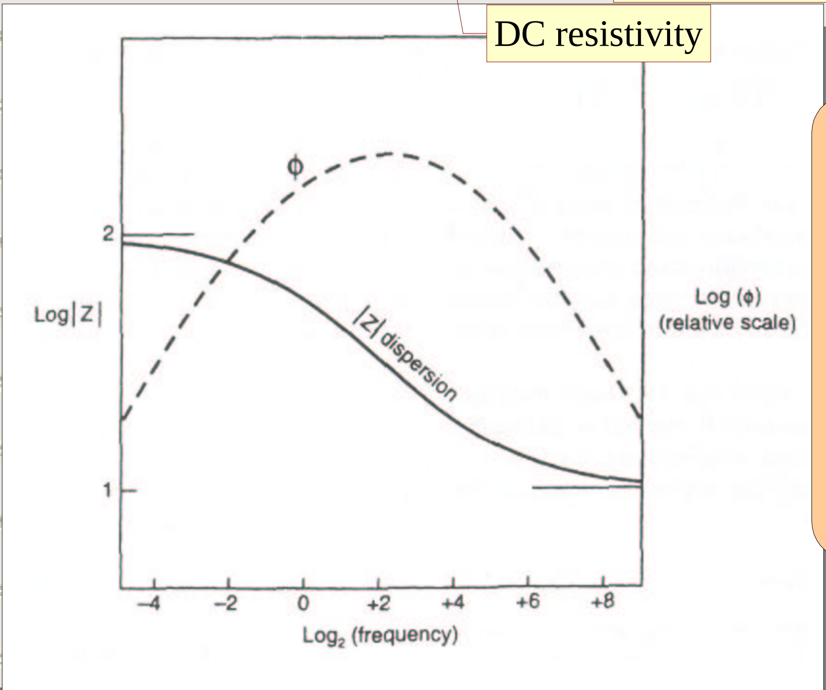
- The *Cole-Cole* model for complex resistivity:

$$Z(\omega)=\rho_0\left[1-M\left(1-\frac{1}{(1+i\omega\tau)^c}\right)\right]=|Z|e^{i\phi}$$

Chargeability

Relaxation time

DC resistivity

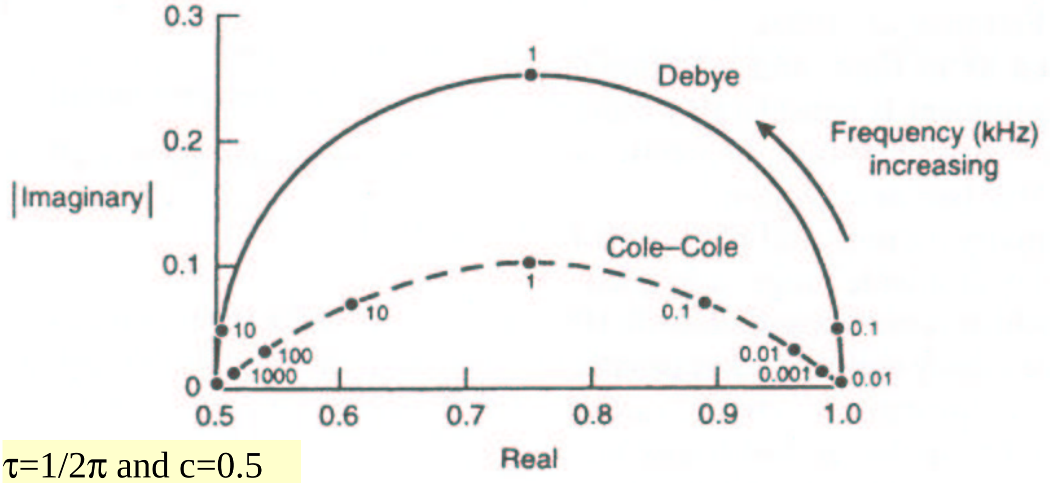


Typical values:
M: 0-1 (depending on mineral content);
τ: 10^{-4} - 10^{4} (depending on grain size);
c: 0.2 - 0.6 (depending on grain size distribution)

Cole-Cole relaxation spectra

- For varying frequencies, complex resistivity describes a semicircle in (ReZ, ImZ) plane:

$$Z(\omega)=\rho_0[1-M(1-\frac{1}{(1+i\omega\tau)^c})]=\Re(Z)+i\Im(Z)$$



τ=1/2π and c=0.5

- The *critical frequency* at which the maximum phase shift is measured is indicative of τ:

$$F_c=\frac{1}{2\pi(1-M)^{1/2c}}$$

Independent of resistivity

Chargeability of various materials

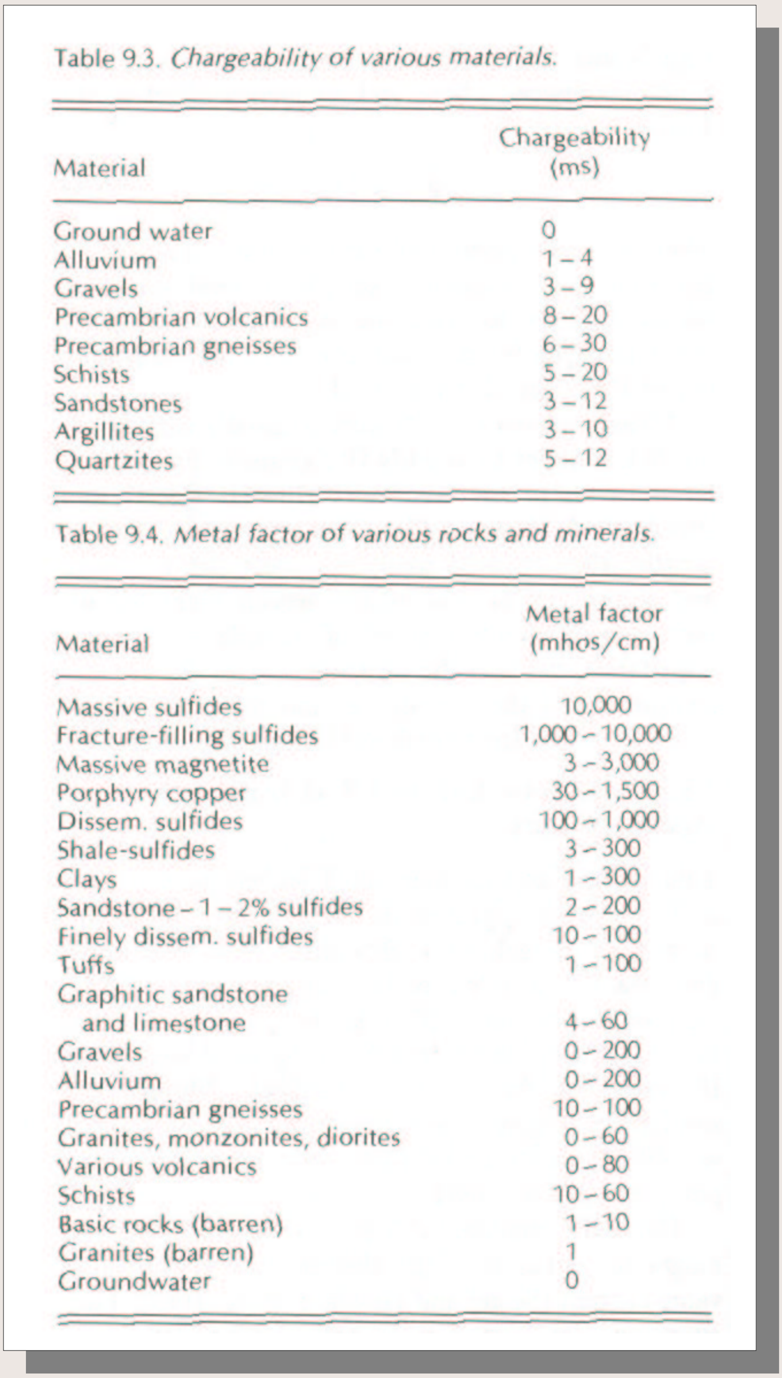


Table 9.3. *Chargeability of various materials.*

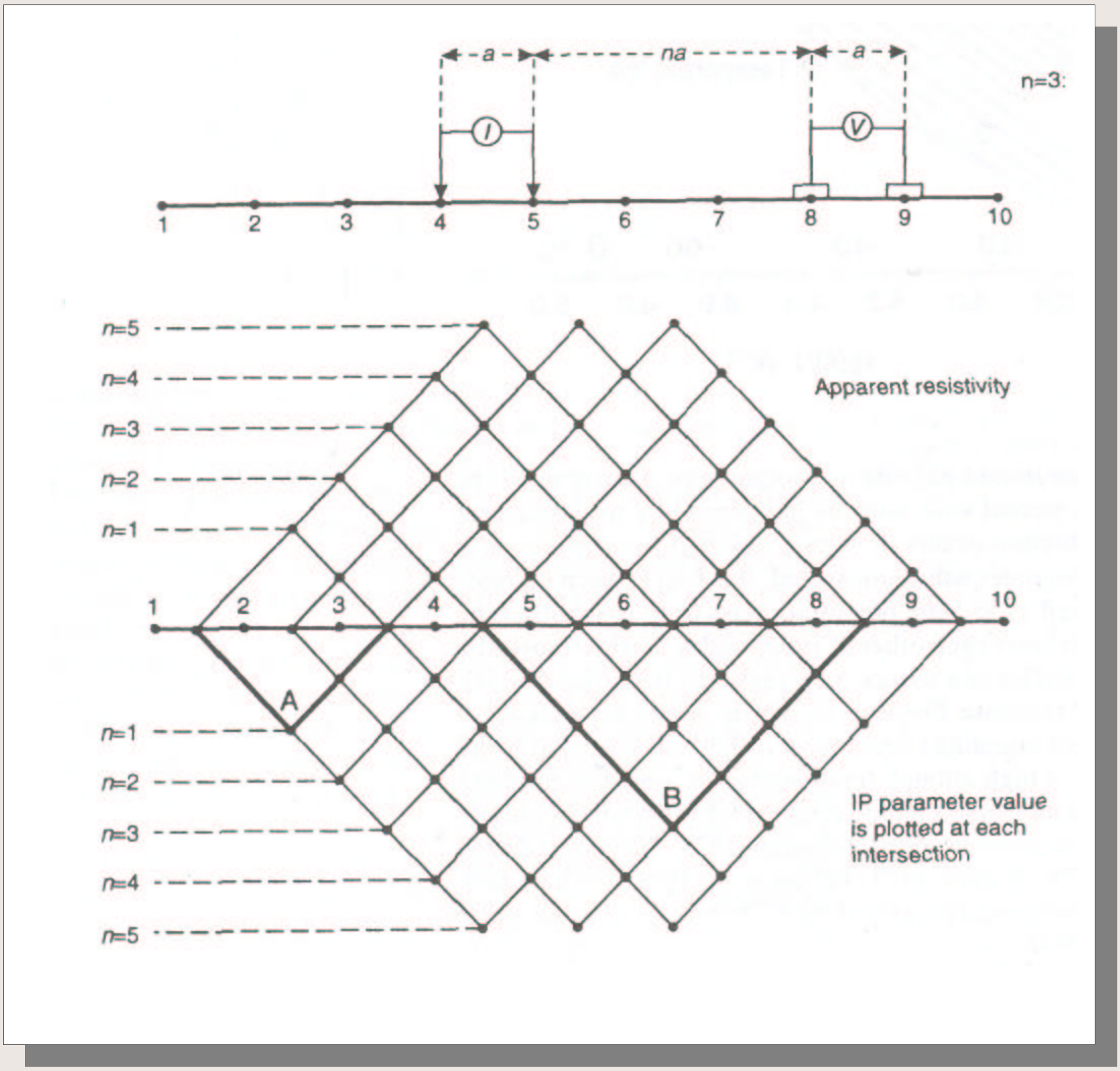
Material	Chargeability (ms)
Ground water	0
Alluvium	1 – 4
Gravels	3 – 9
Precambrian volcanics	8 – 20
Precambrian gneisses	6 – 30
Schists	5 – 20
Sandstones	3 – 12
Argillites	3 – 10
Quartzites	5 – 12

Table 9.4. *Metal factor of various rocks and minerals.*

Material	Metal factor (mhos/cm)
Massive sulfides	10,000
Fracture-filling sulfides	1,000 – 10,000
Massive magnetite	3 – 3,000
Porphyry copper	30 – 1,500
Dissem. sulfides	100 – 1,000
Shale-sulfides	3 – 300
Clays	1 – 300
Sandstone – 1 – 2% sulfides	2 – 200
Finely dissem. sulfides	10 – 100
Tuffs	1 – 100
Graphitic sandstone and limestone	4 – 60
Gravels	0 – 200
Alluvium	0 – 200
Precambrian gneisses	10 – 100
Granites, monzonites, diorites	0 – 60
Various volcanics	0 – 80
Schists	10 – 60
Basic rocks (barren)	1 – 10
Granites (barren)	1
Groundwater	0

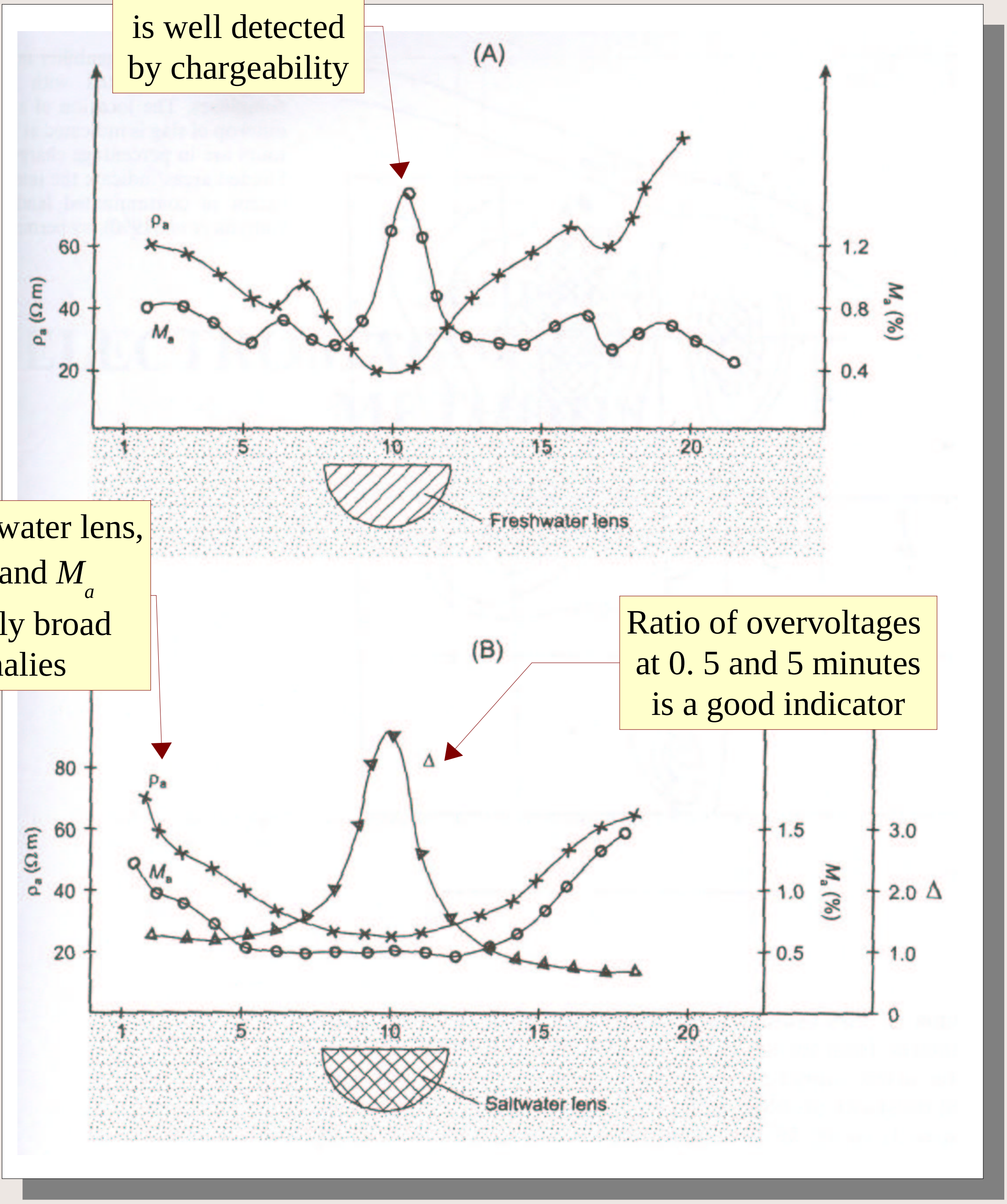
Displays of IP data

- Profiles and maps of apparent chargeability (time-domain IP);
- Pseudo-sections (combined with ρ_a)



Lab Case History

(Ogilvy and Kuzmina, 1972)



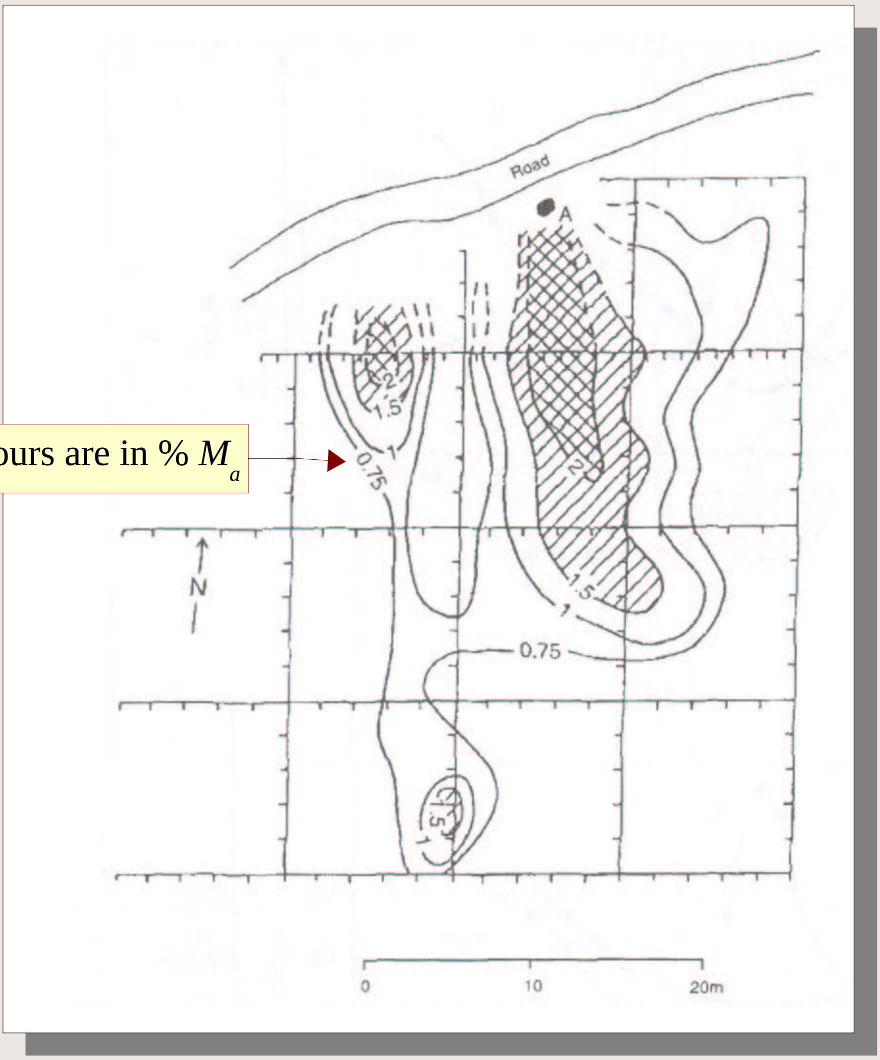
Freshwater lens is well detected by chargeability

For saline water lens, both ρ_a and M_a show only broad anomalies

Ratio of overvoltages at 0. 5 and 5 minutes is a good indicator

IP Case History

- Identification of a contamination with cyanide complexes (slags from plating works; Cahyna et al., 1990);
    - Resistivity survey failed to detect th contamination;
    - IP chargeability identified both the known and unknown slag deposits.



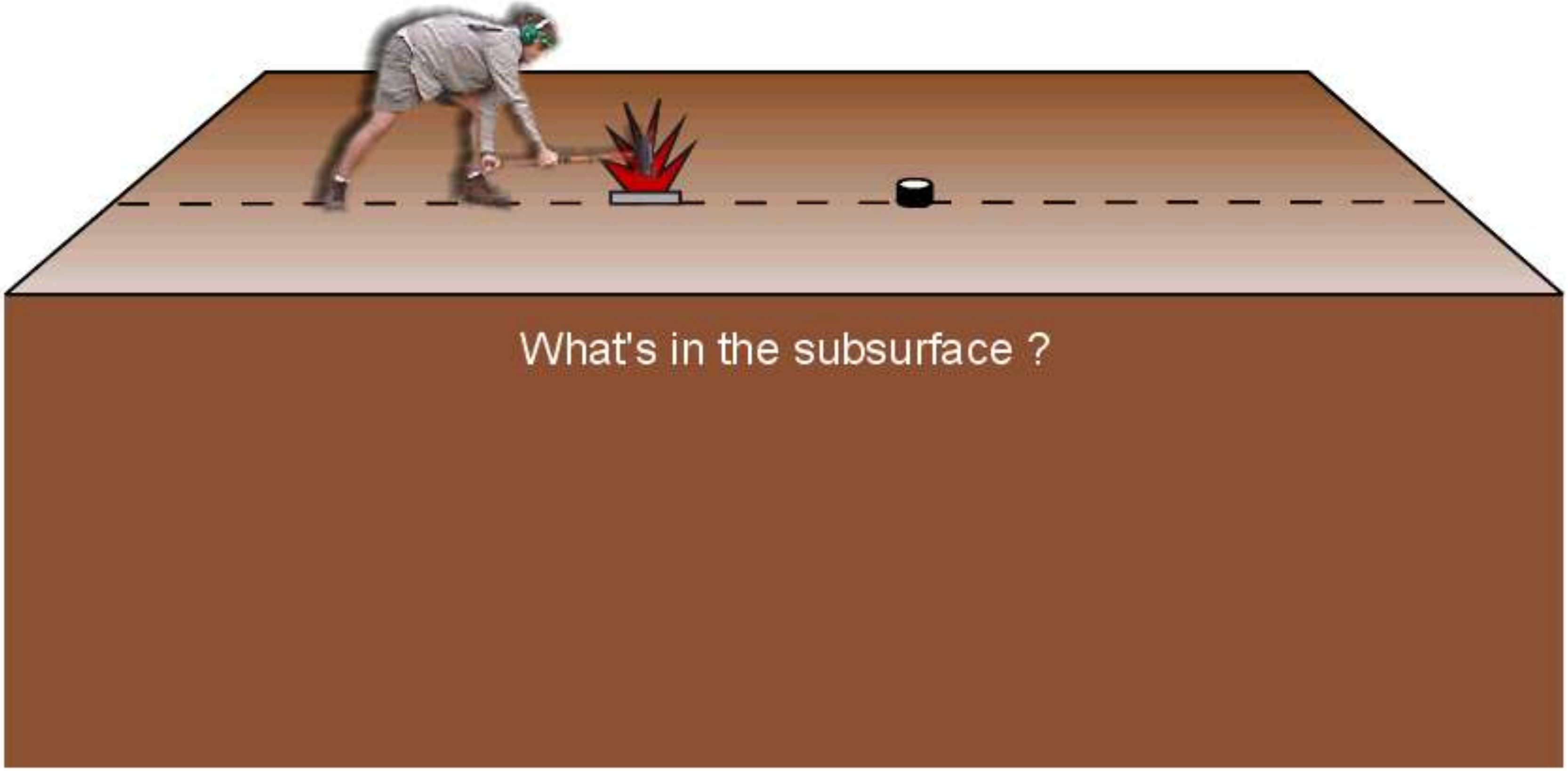
Contours are in % M_a

Using Geophysics to Characterize the Subsurface:

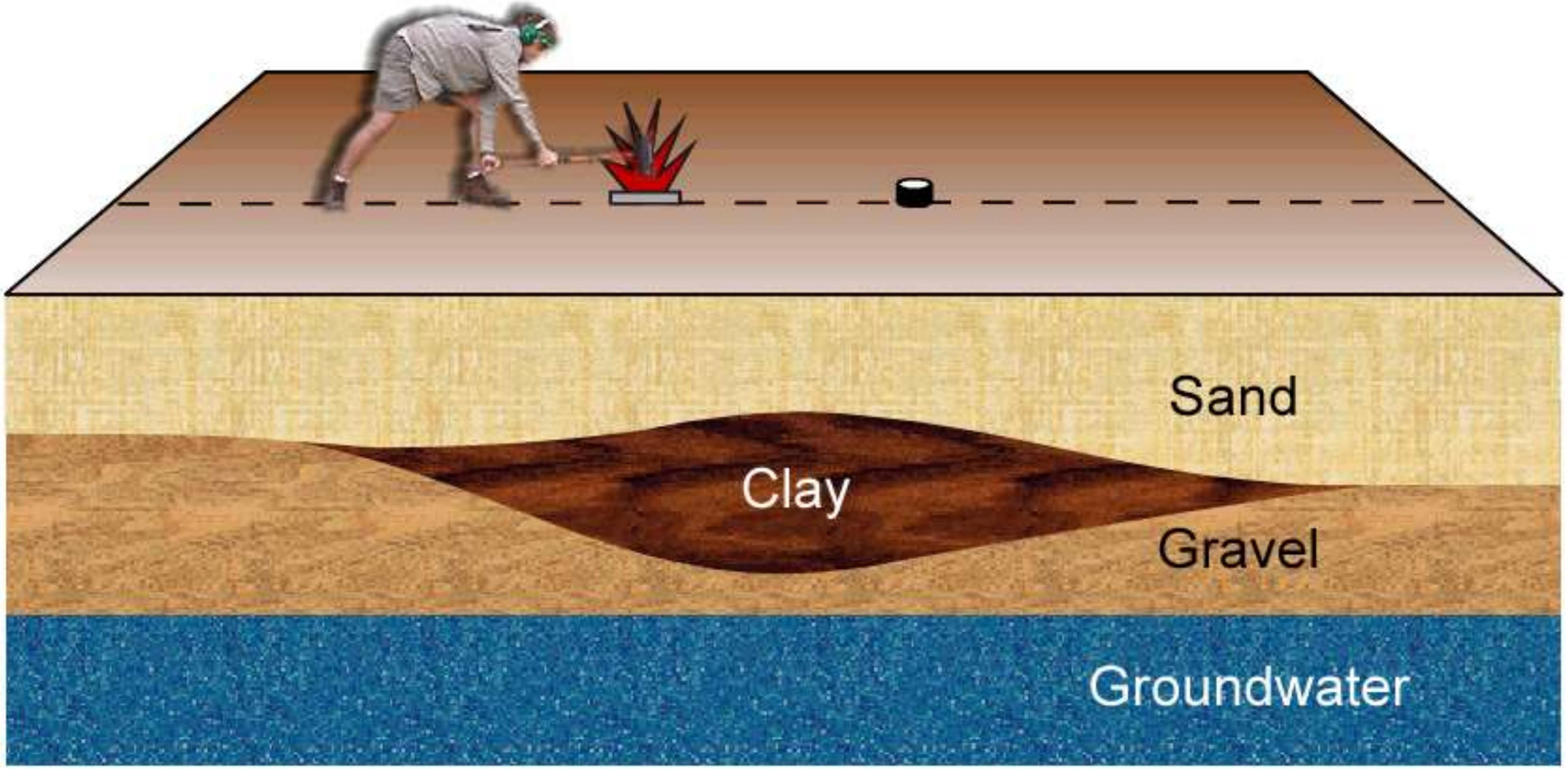
“The Principles”

BERAN GÜRLEME
OCAK/2011

Using Geophysics to Characterize the Subsurface: The Principles



Using Geophysics to Characterize the Subsurface: The Principles



Seismic Methods in Environmental & Engineering Investigations

Objectives

- Subsurface material at a site: unconsolidated sand & gravel, or bedrock.
- Depth to water table.
- Lateral changes in the mechanical/ hydraulic properties of an aquifer.
- Local dip of the water table (the hydraulic gradient) or lateral changes in its depth.
- Depth to bedrock beneath unconsolidated sediments.
- Mechanical competency (or rippability) of subsurface materials.

We determine subsurface conditions by remotely sensing physical properties of materials in situ.

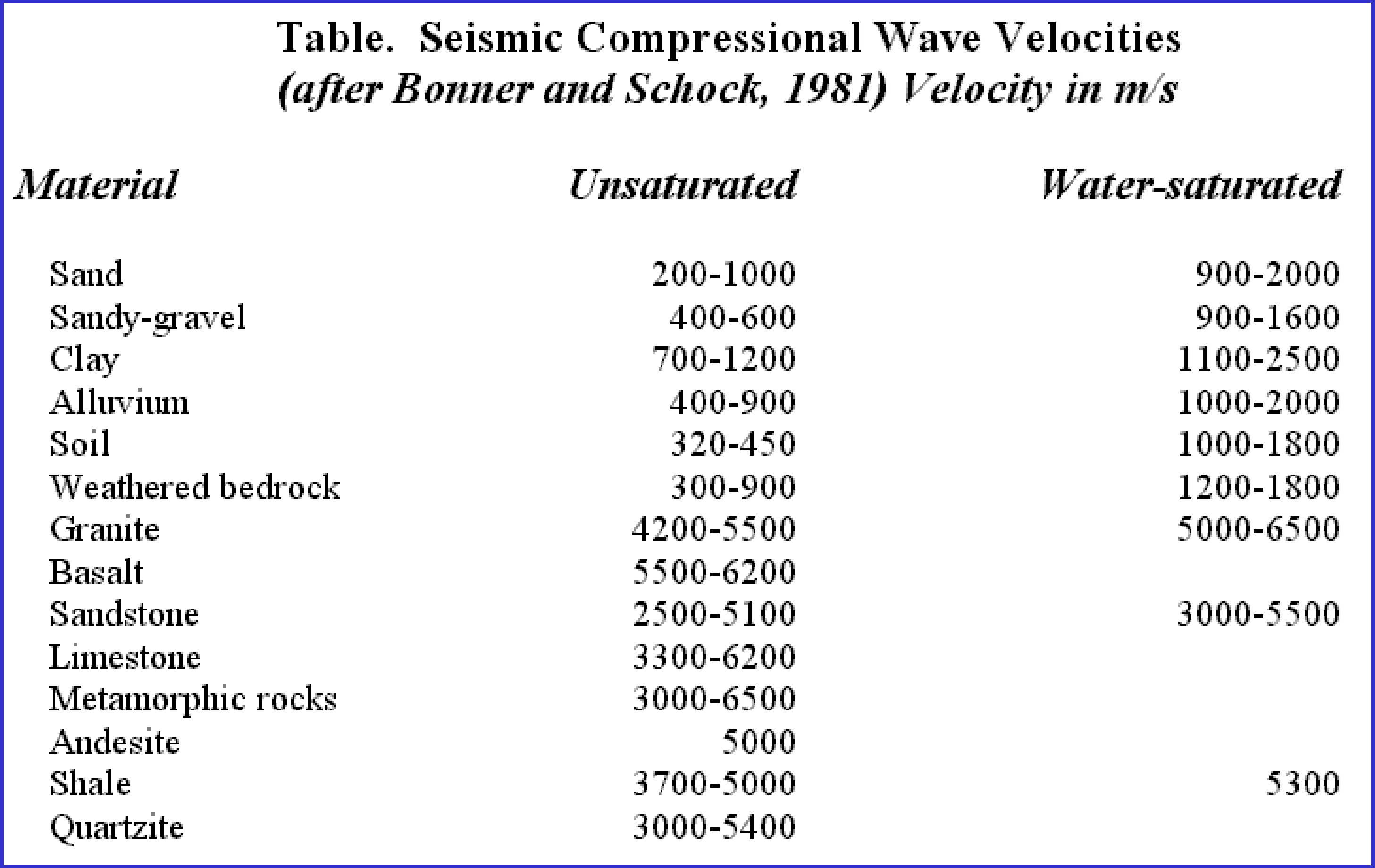
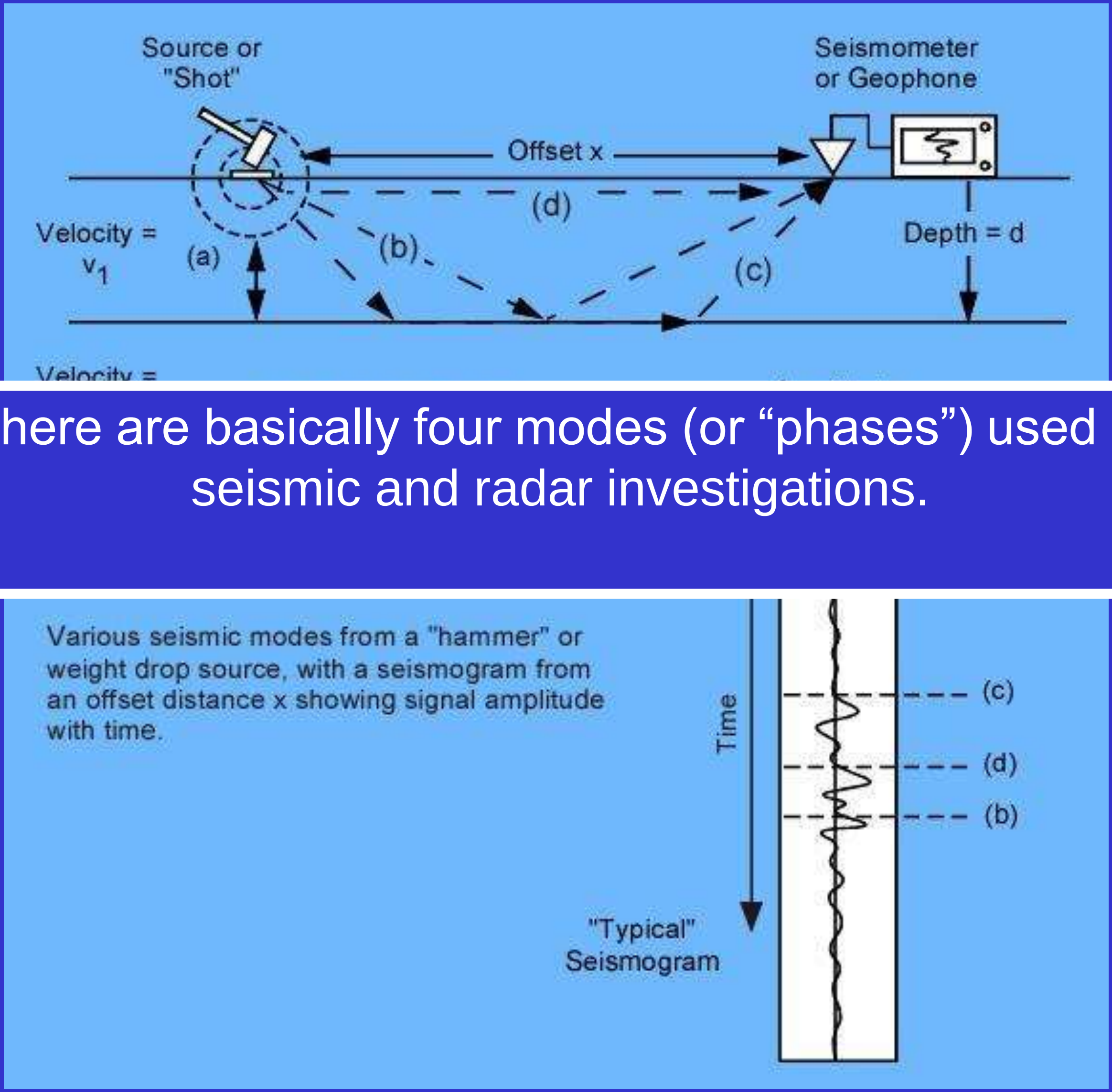


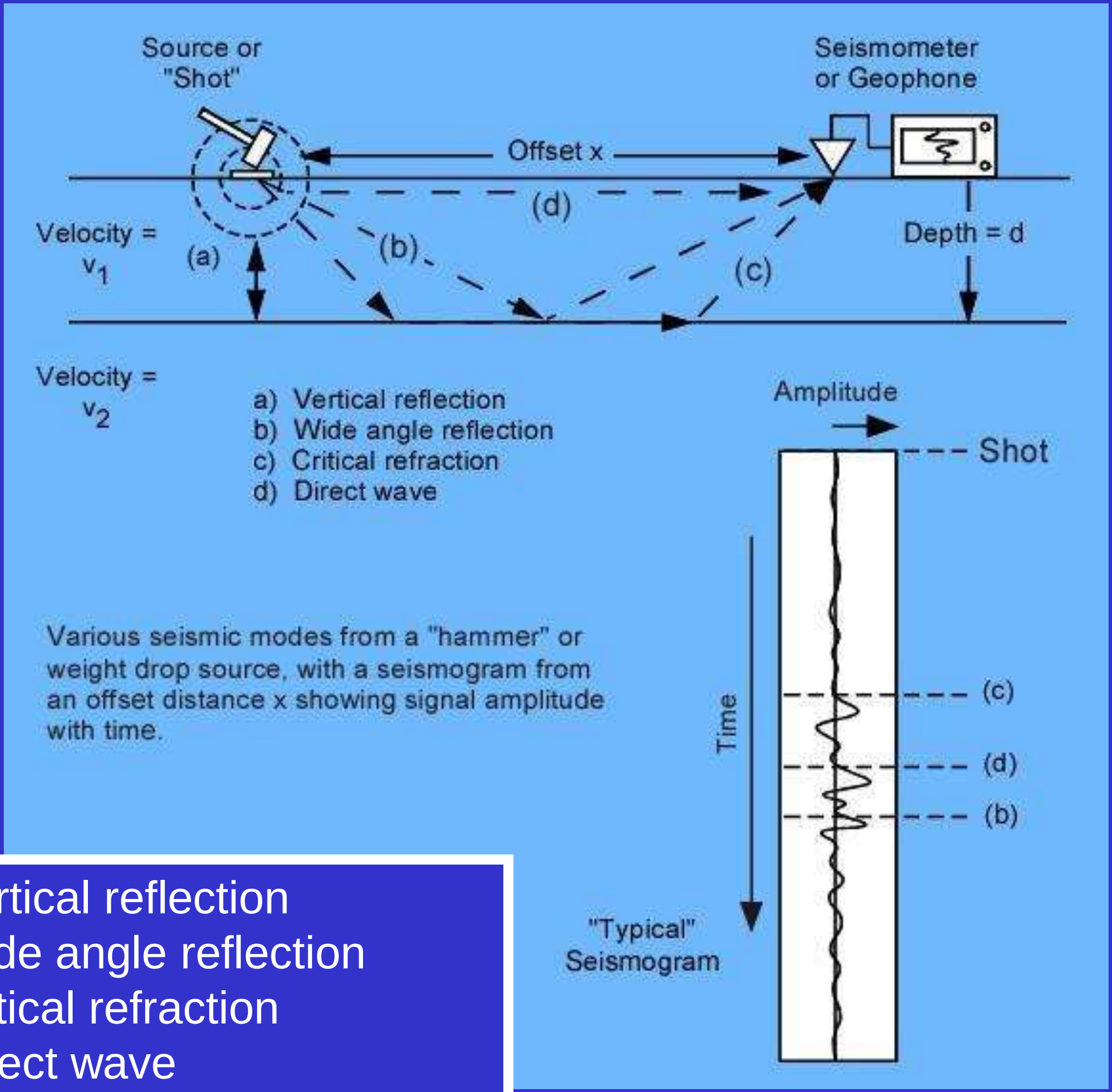
Table. Seismic Compressional Wave Velocities
(after Bonner and Schock, 1981) Velocity in m/s

Material	*Unsaturated*	*Water-saturated*
Sand	200-1000	900-2000
Sandy-gravel	400-600	900-1600
Clay	700-1200	1100-2500
Alluvium	400-900	1000-2000
Soil	320-450	1000-1800
Weathered bedrock	300-900	1200-1800
Granite	4200-5500	5000-6500
Basalt	5500-6200	
Sandstone	2500-5100	3000-5500
Limestone	3300-6200	
Metamorphic rocks	3000-6500	
Andesite	5000	
Shale	3700-5000	5300
Quartzite	3000-5400	

There are basically four modes (or “phases”) used in seismic and radar investigations.

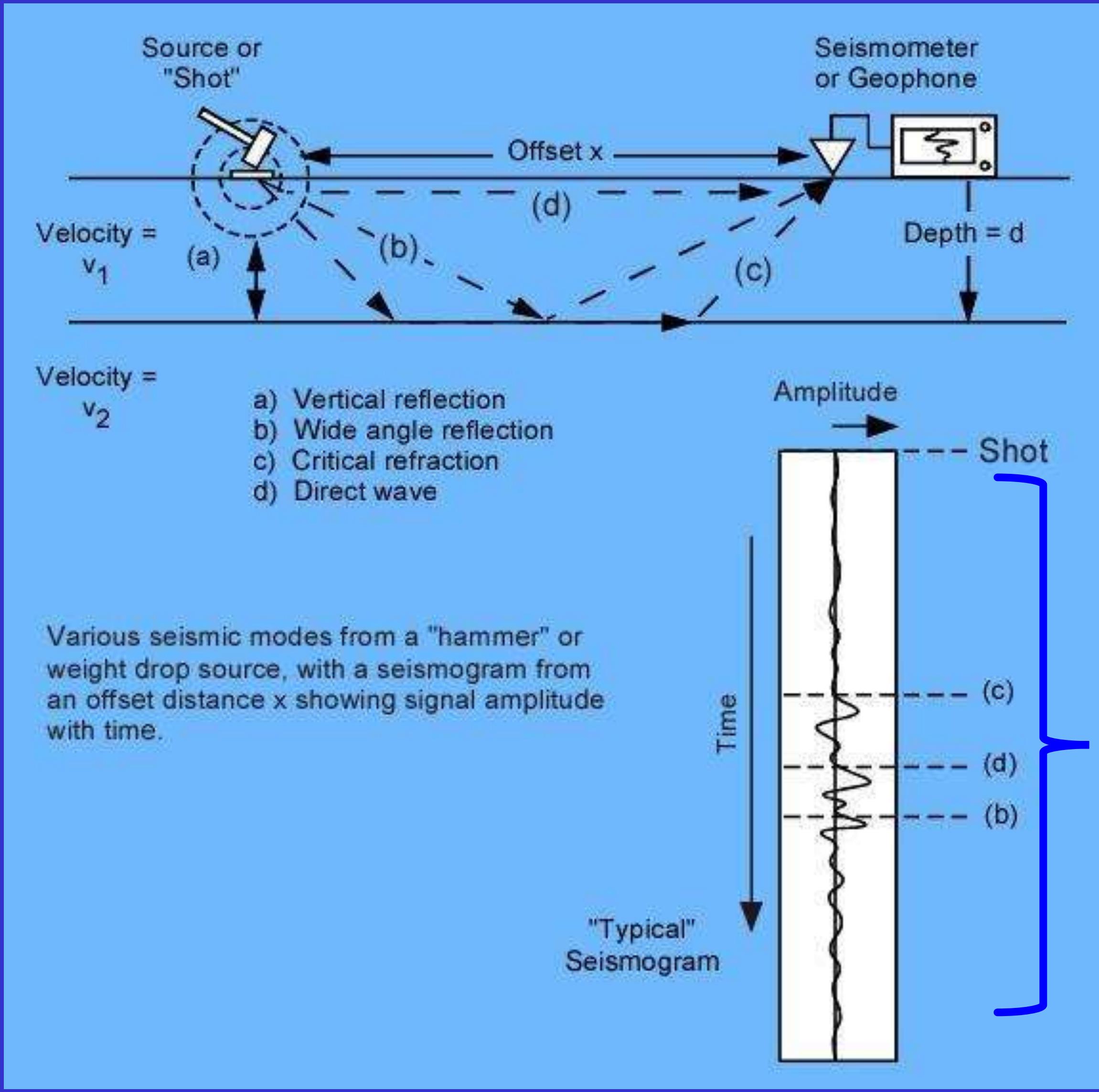


Various seismic modes from a "hammer" or weight drop source, with a seismogram from an offset distance x showing signal amplitude with time.



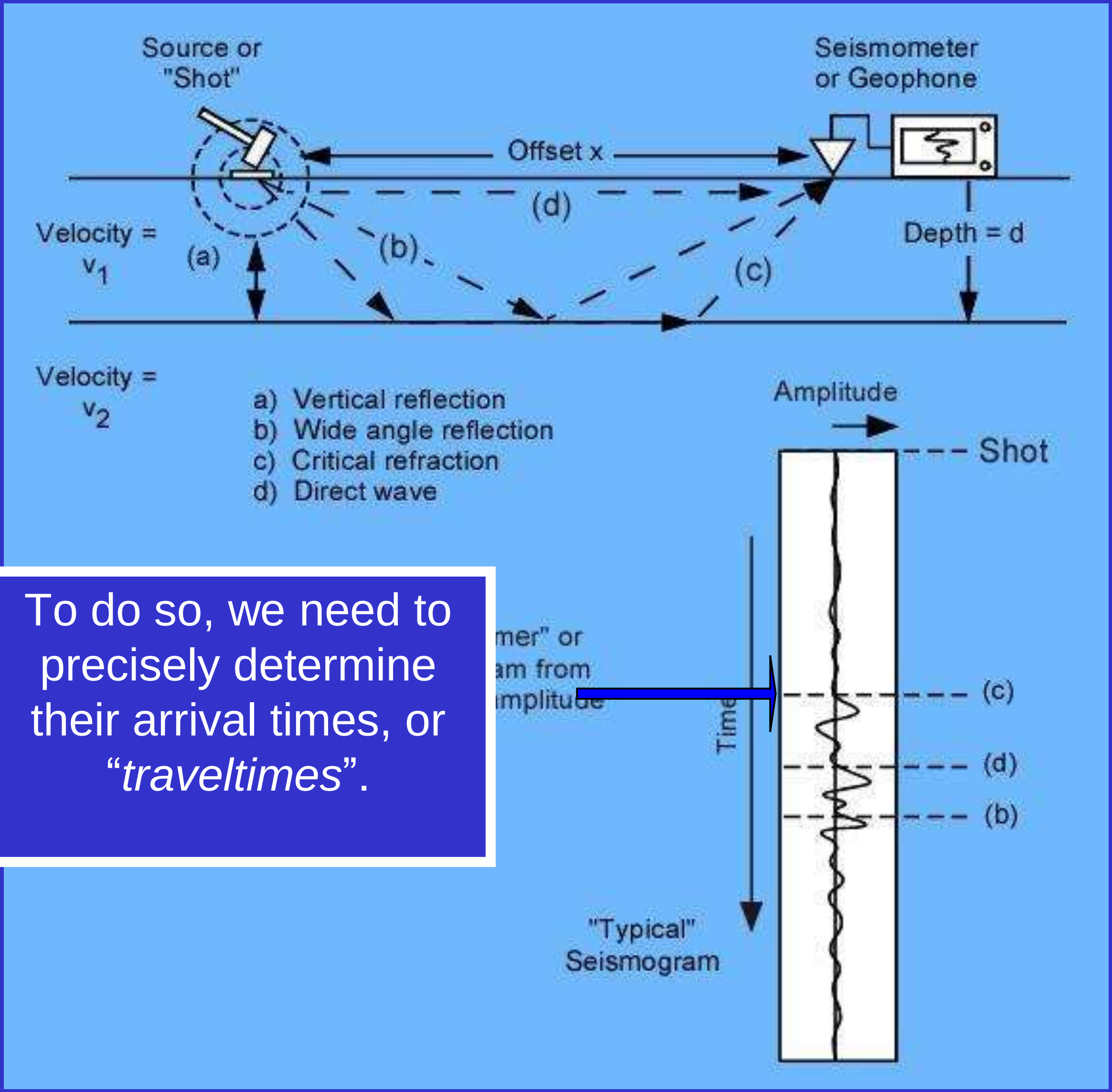
Various seismic modes from a "hammer" or weight drop source, with a seismogram from an offset distance x showing signal amplitude with time.

a) Vertical reflection
b) Wide angle reflection
c) Critical refraction
d) Direct wave

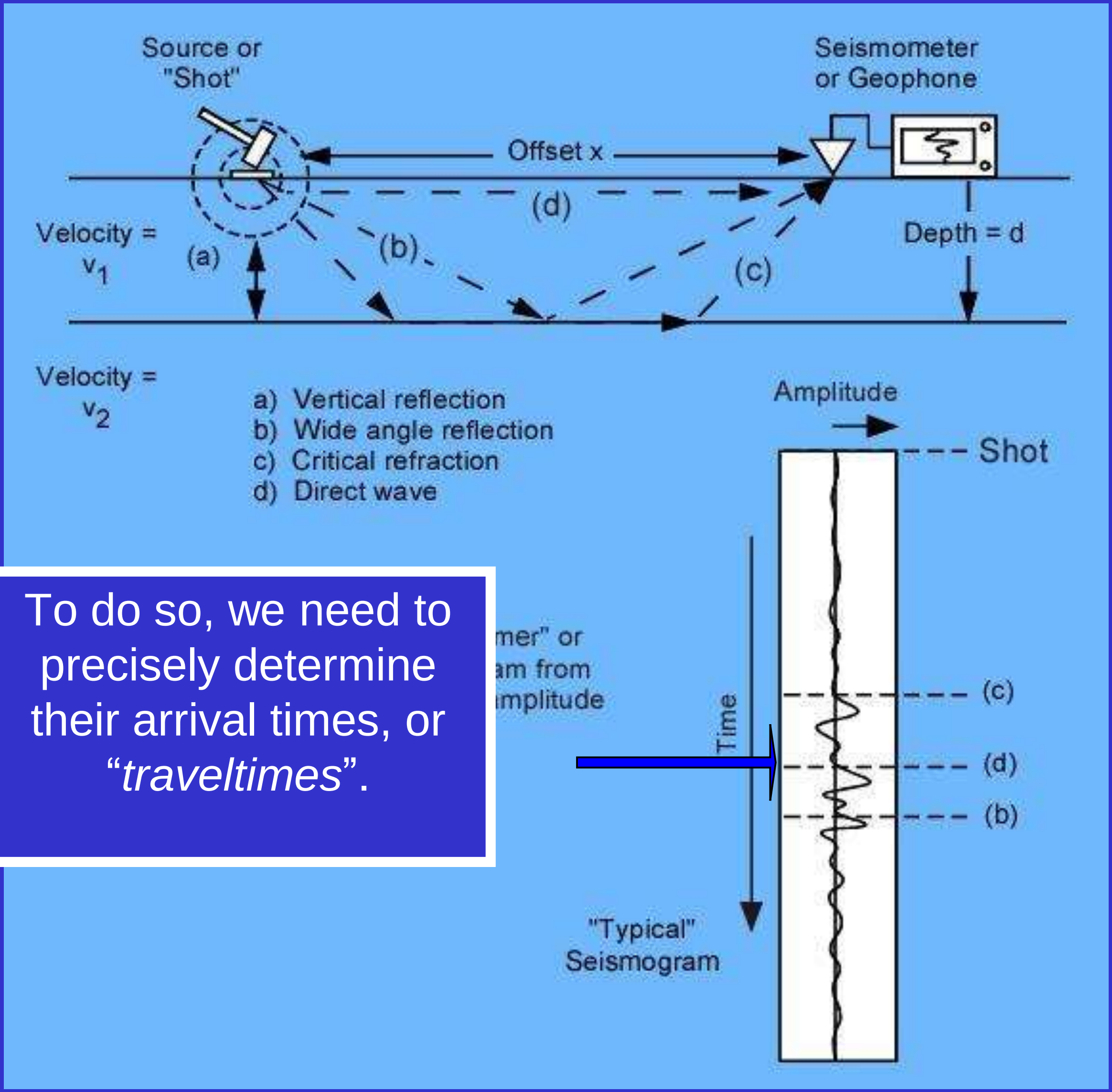


Various seismic modes from a "hammer" or weight drop source, with a seismogram from an offset distance x showing signal amplitude with time.

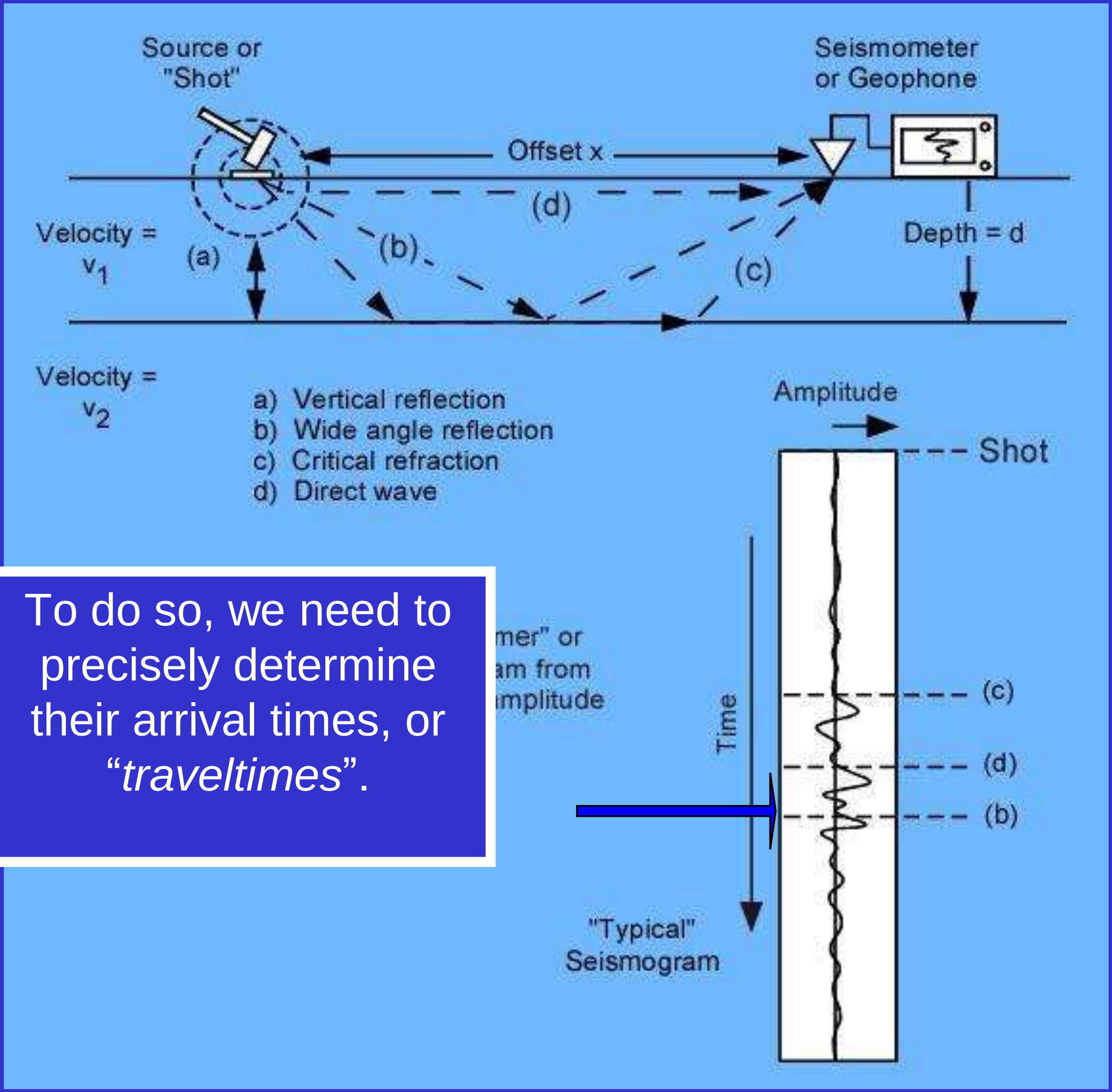
The seismic method records these signals and analyses their relationship.



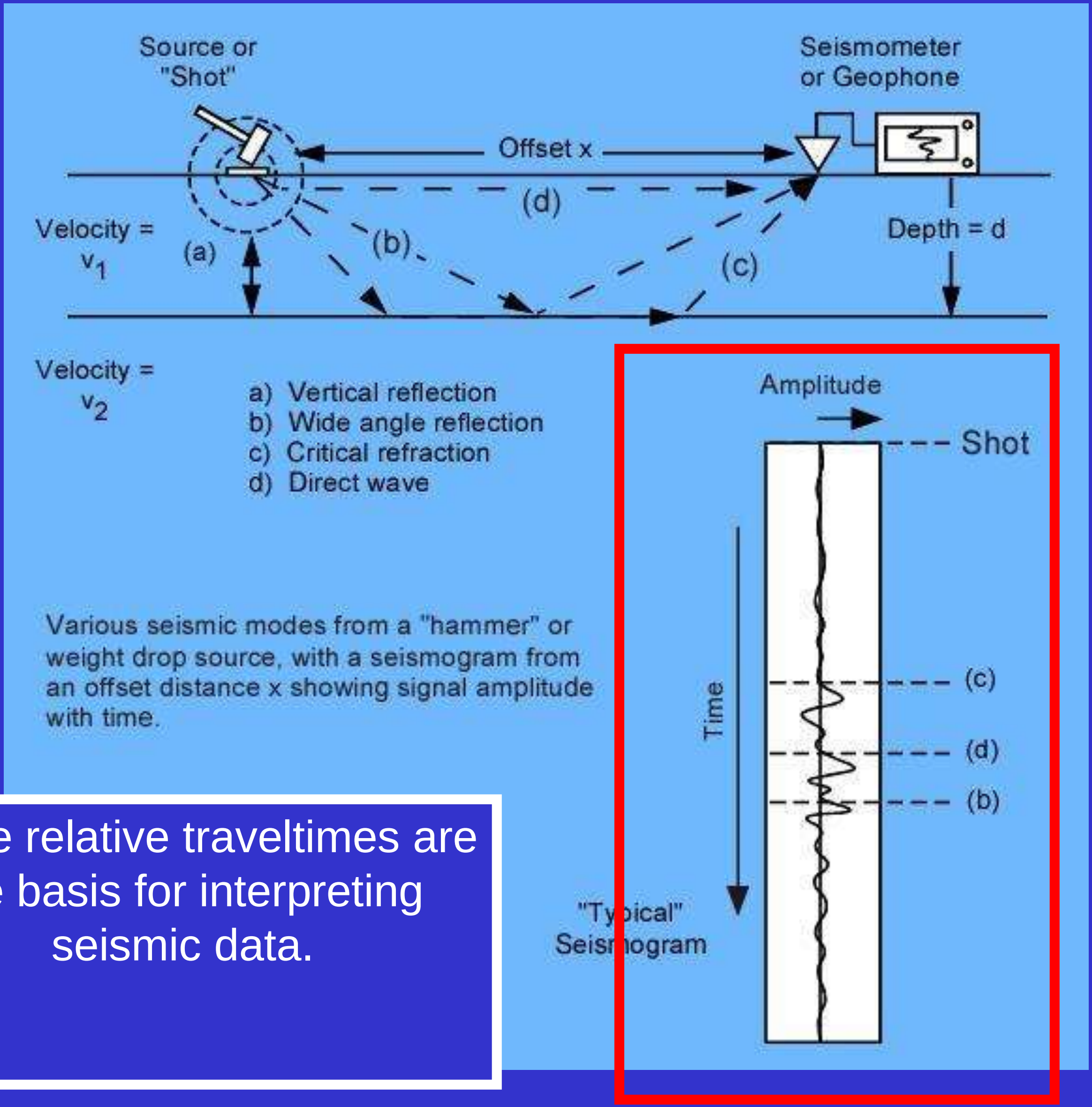
To do so, we need to precisely determine their arrival times, or "*traveltimes*".



To do so, we need to precisely determine their arrival times, or "*traveltimes*".



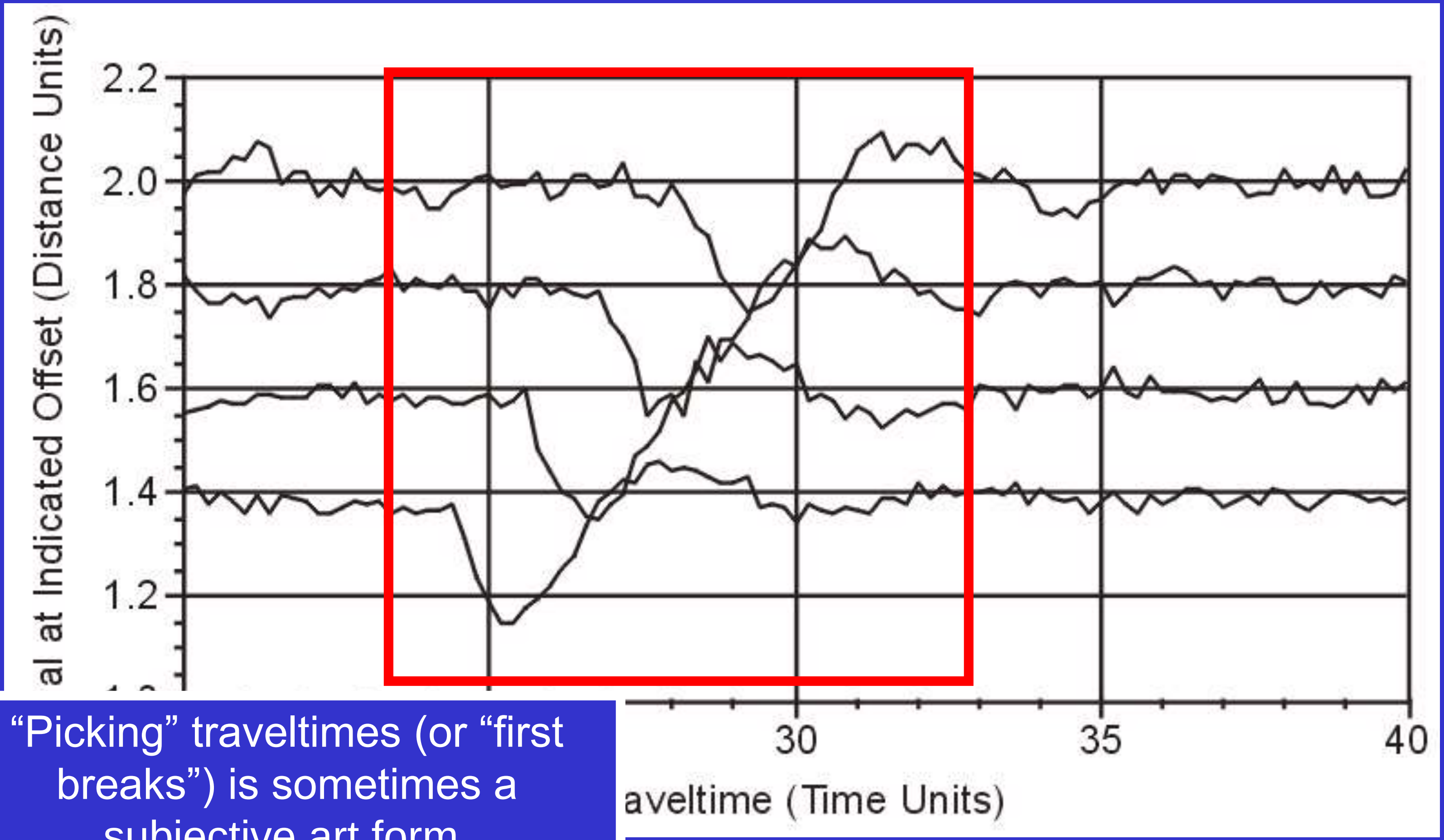
To do so, we need to precisely determine their arrival times, or "*traveltimes*".



Various seismic modes from a "hammer" or weight drop source, with a seismogram from an offset distance x showing signal amplitude with time.

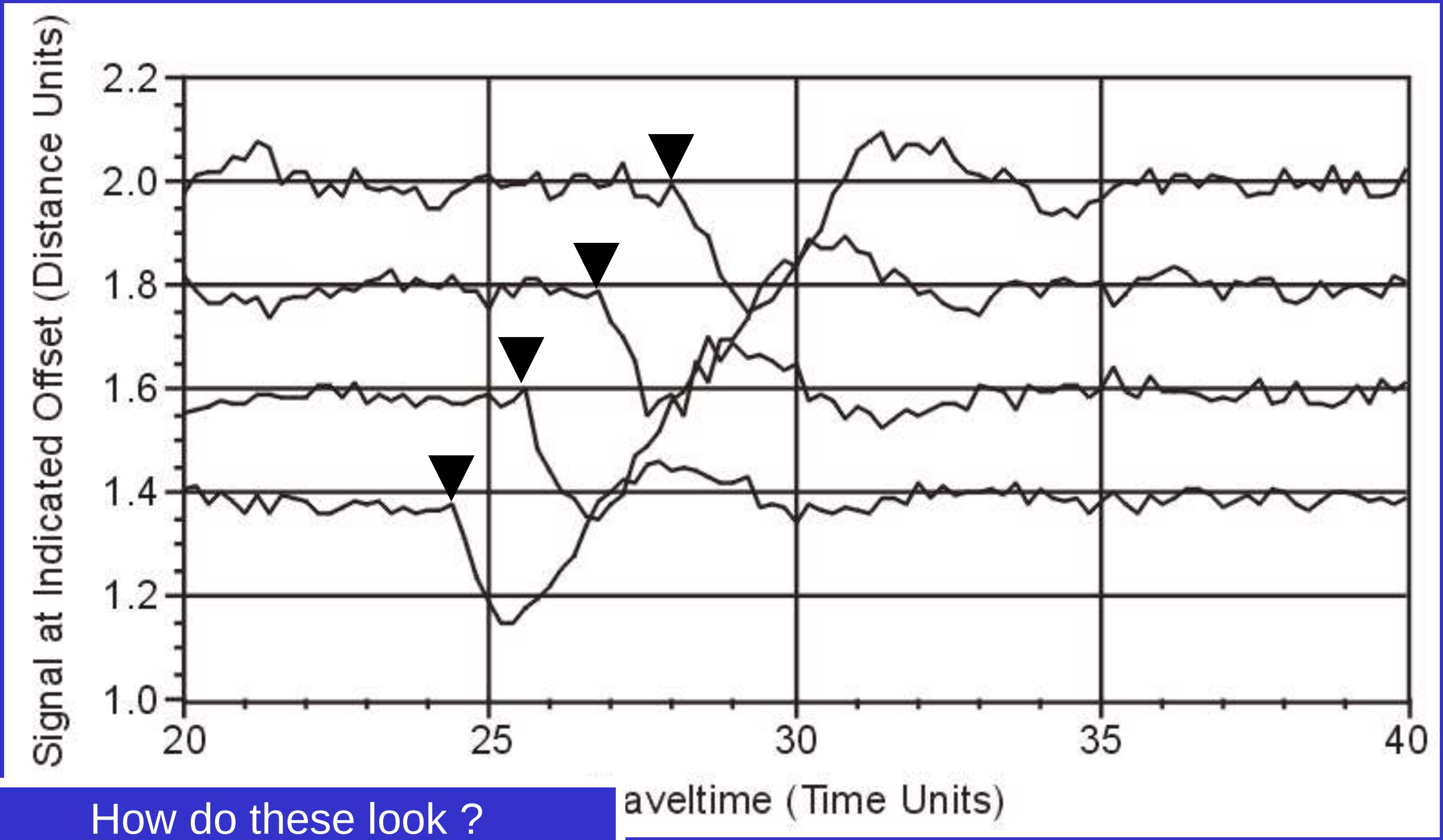
These relative traveltimes are the basis for interpreting seismic data.

An example of 4 traces recorded at adjacent position offsets.



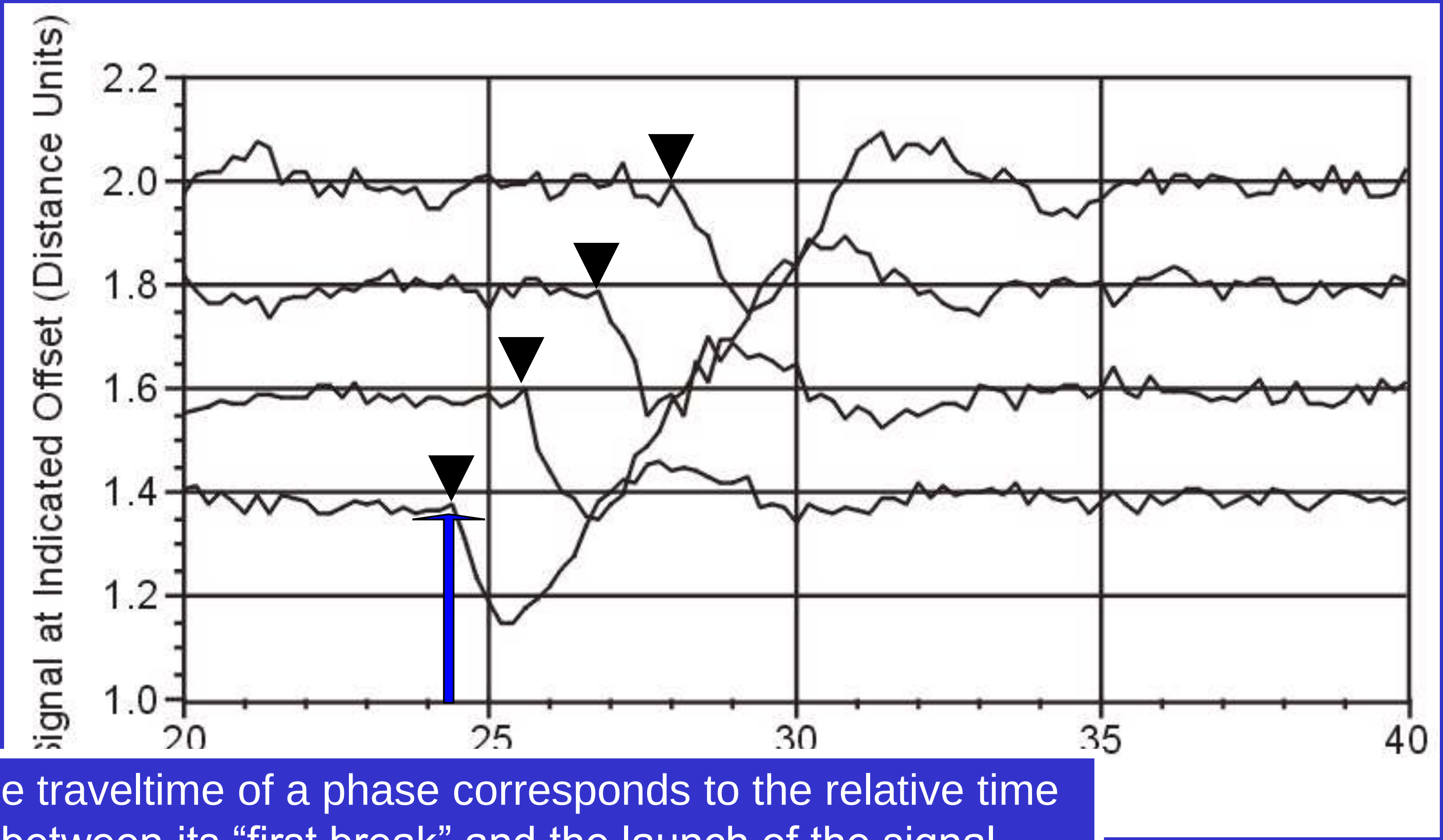
"Picking" traveltimes (or "first breaks") is sometimes a subjective art form.

An example of 4 traces recorded at adjacent position offsets.



How do these look ?

An example of 4 traces recorded at adjacent position offsets.



The traveltime of a phase corresponds to the relative time between its "first break" and the launch of the signal.

General

Seismic *refraction* techniques are the most common seismic method used for environmental site assessments.

While seismic *reflection* techniques are the overwhelming methods of choice for oil exploration at depths of 100's to 1000's of meters or more, their utility is restricted at depths of interest to environmental assessment (from 1 to 10 meters).

Seismic Sources

Natural sources (earthquakes, microearthquakes etc.)

- Useful for local structural studies and natural hazards
- Passive “vibration” studies are similar in application

Seismic Sources !!

Artificial Sources

- Sledge hammers, post-hole drivers & weight drops (preferred for environmental site assessments)
- Explosives (blasting caps, shot-gun shells, quarter sticks of dynamite)
- Vibroseis (A platform firmly implanted on the ground is vibrated at progressively increasing frequency in a "chirp" mode.

The transmitted chirp is cross-correlated with the composite signal received at the geophones to detect specific reflections, refractions, etc.

Size varies:

- One person vibrators or compactors;
- Articulated earth movers.

(*University of Bergen*.)

(*U British Columbia: Lithoprobe Project*.)

(Network for Earthquake Engineering Simulation; U Texas.)

Vibrating type sources

Seismic wave types

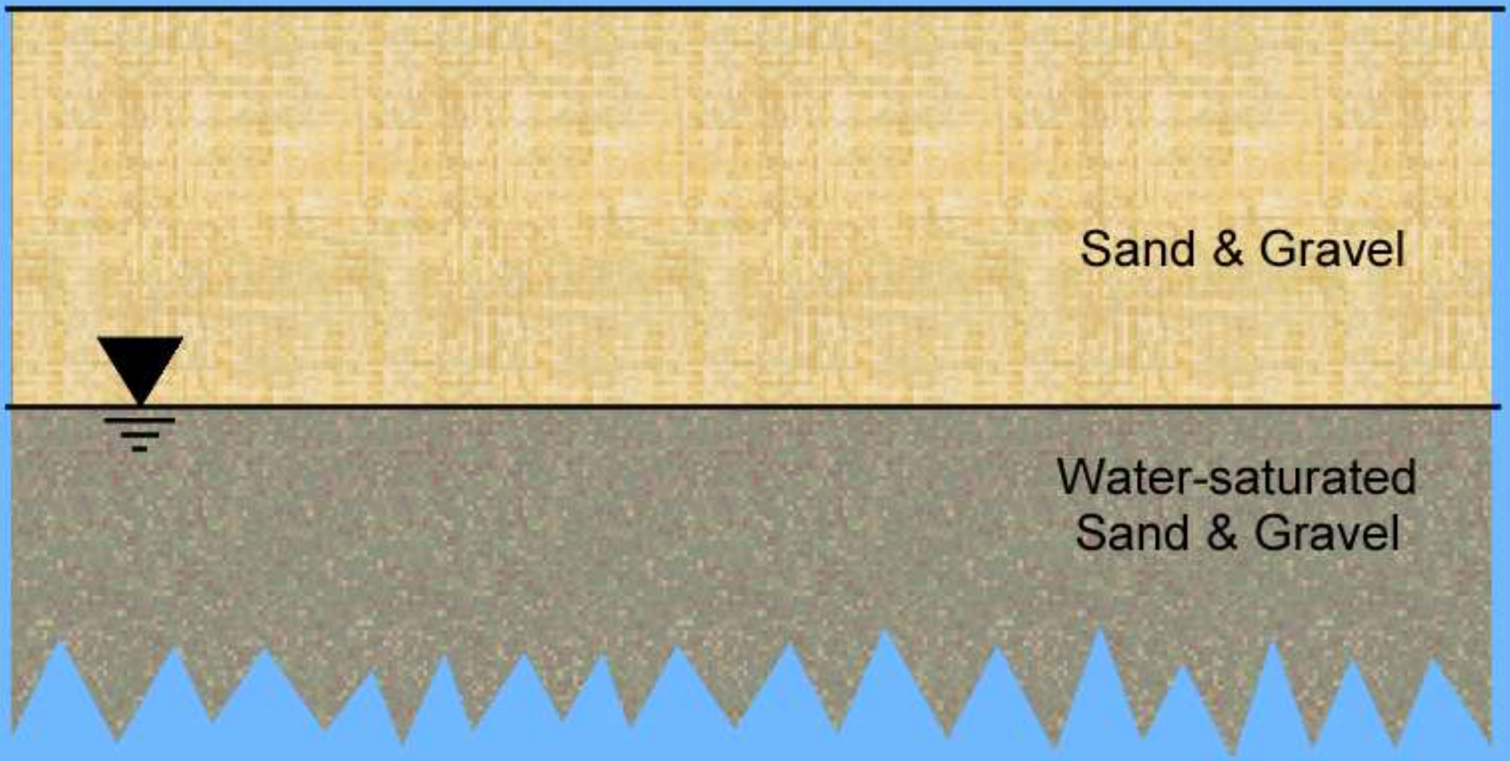
P Waves. Compressional waves. (This mode travels at the highest velocity, arriving at the geophone first (the primary arrival), thus is responsible for the clean "first break" on a seismic trace. It is therefore the usual mode that has been traditionally employed.)

S Waves. Shear waves. (Excellent attributes for "seeing through" the water-table in unconsolidated sediments to bedrock below, but require special, non-standard field procedures).

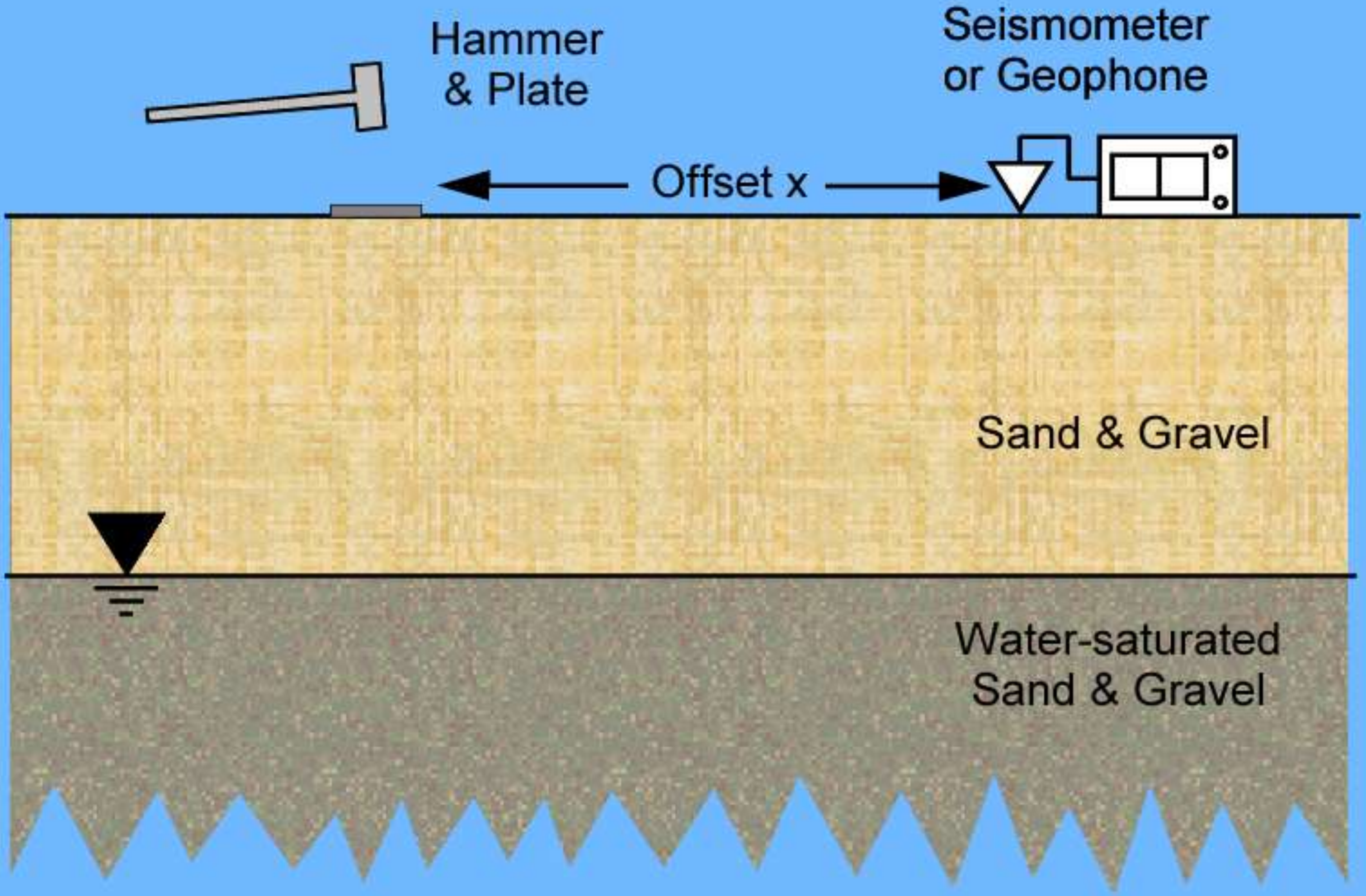
Surface Waves (Love and Rayleigh Waves: largely "noise" on standard surveys, but contain "signal" for special applications).

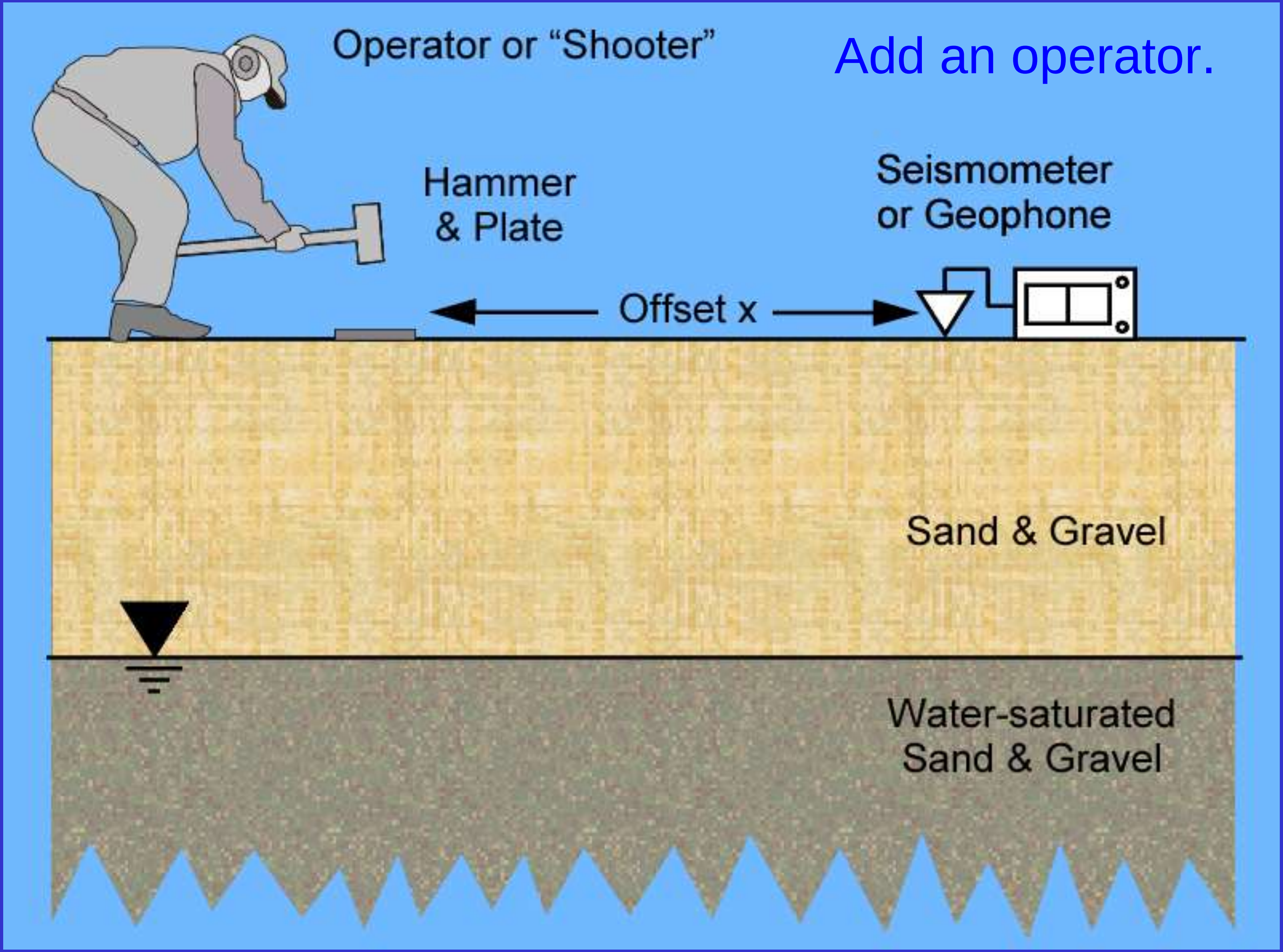
Implementing the Seismic Method.

Consider a two layered earth model.



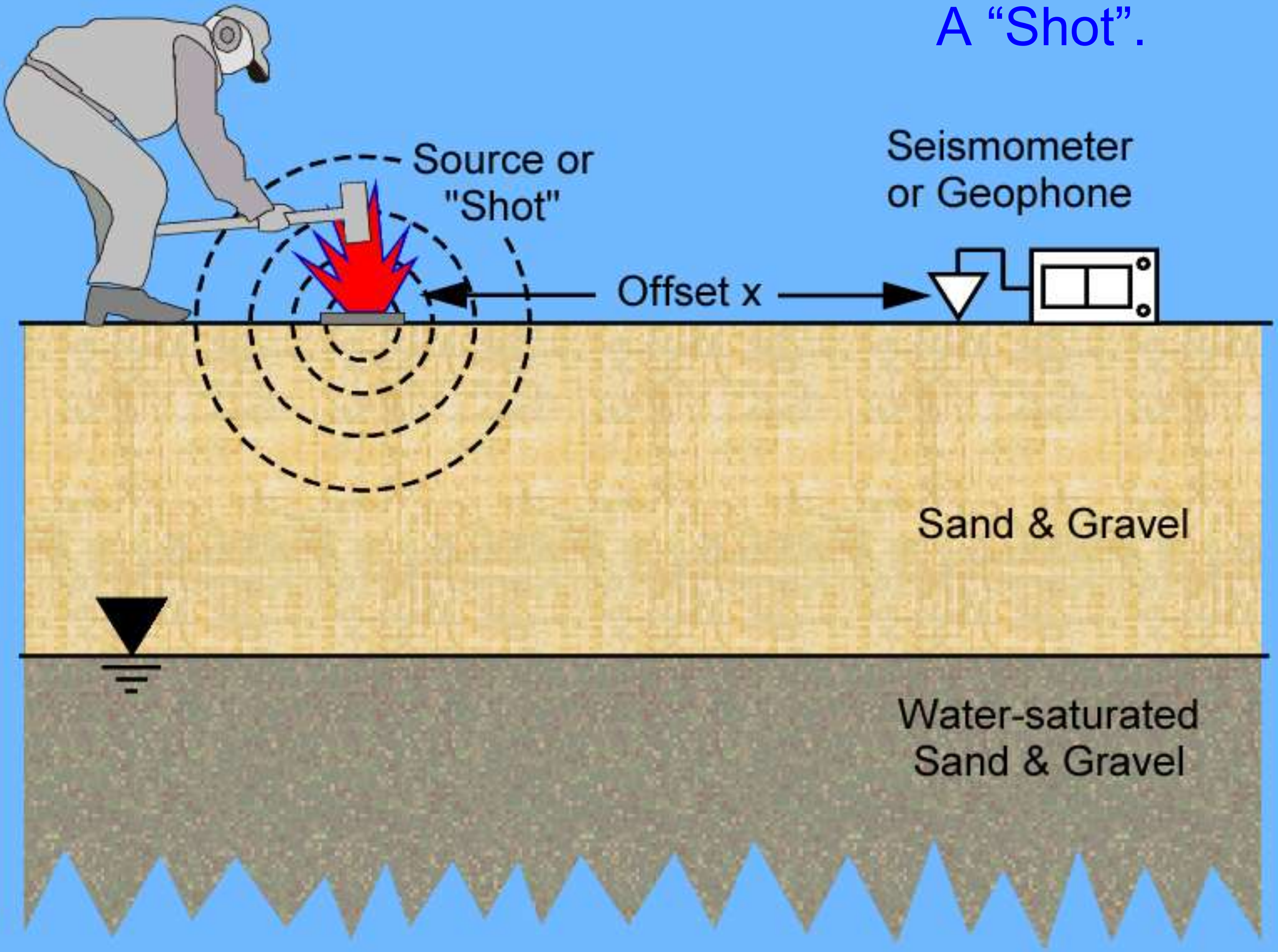
Principal instruments.



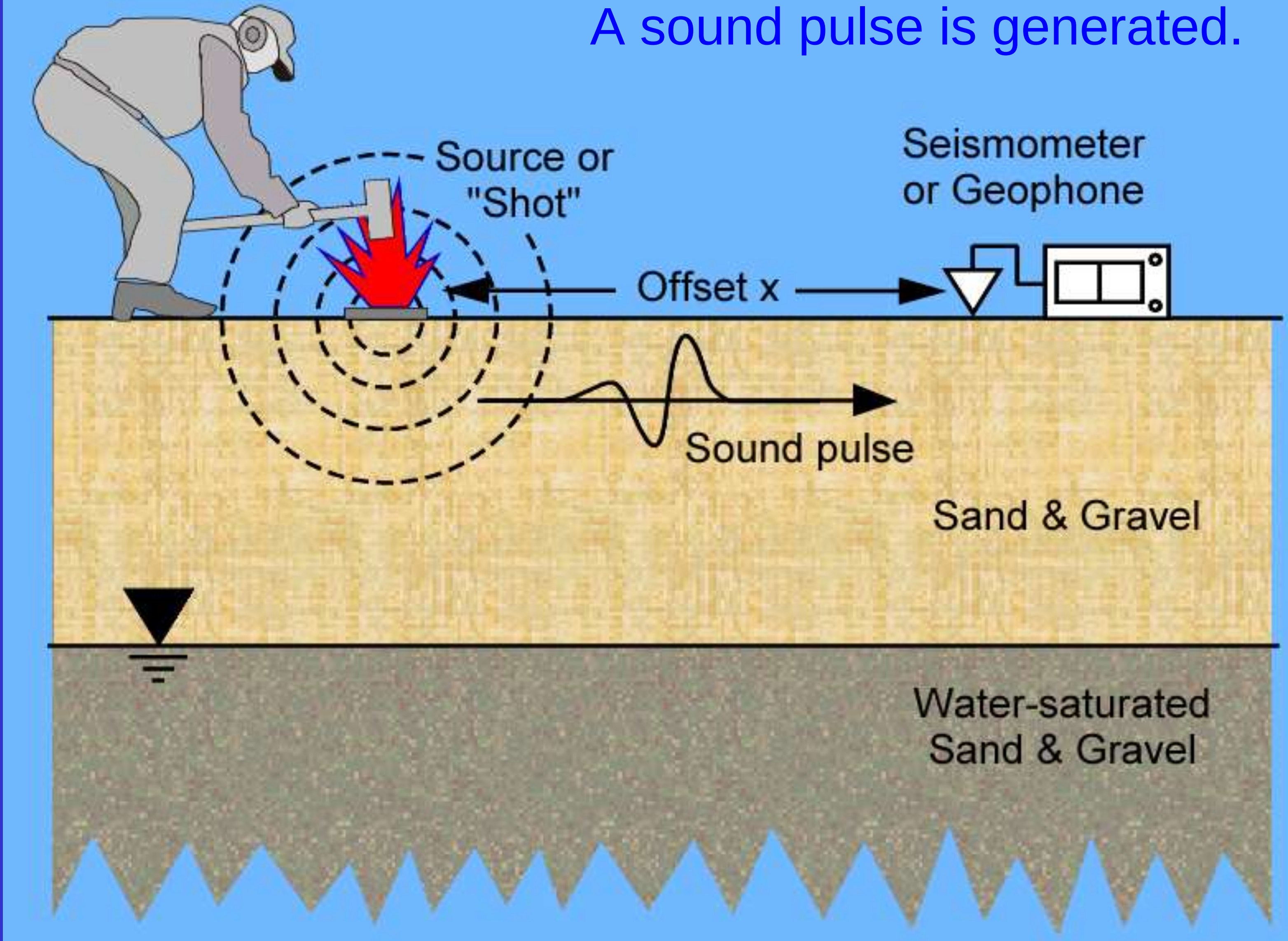


Add an operator.

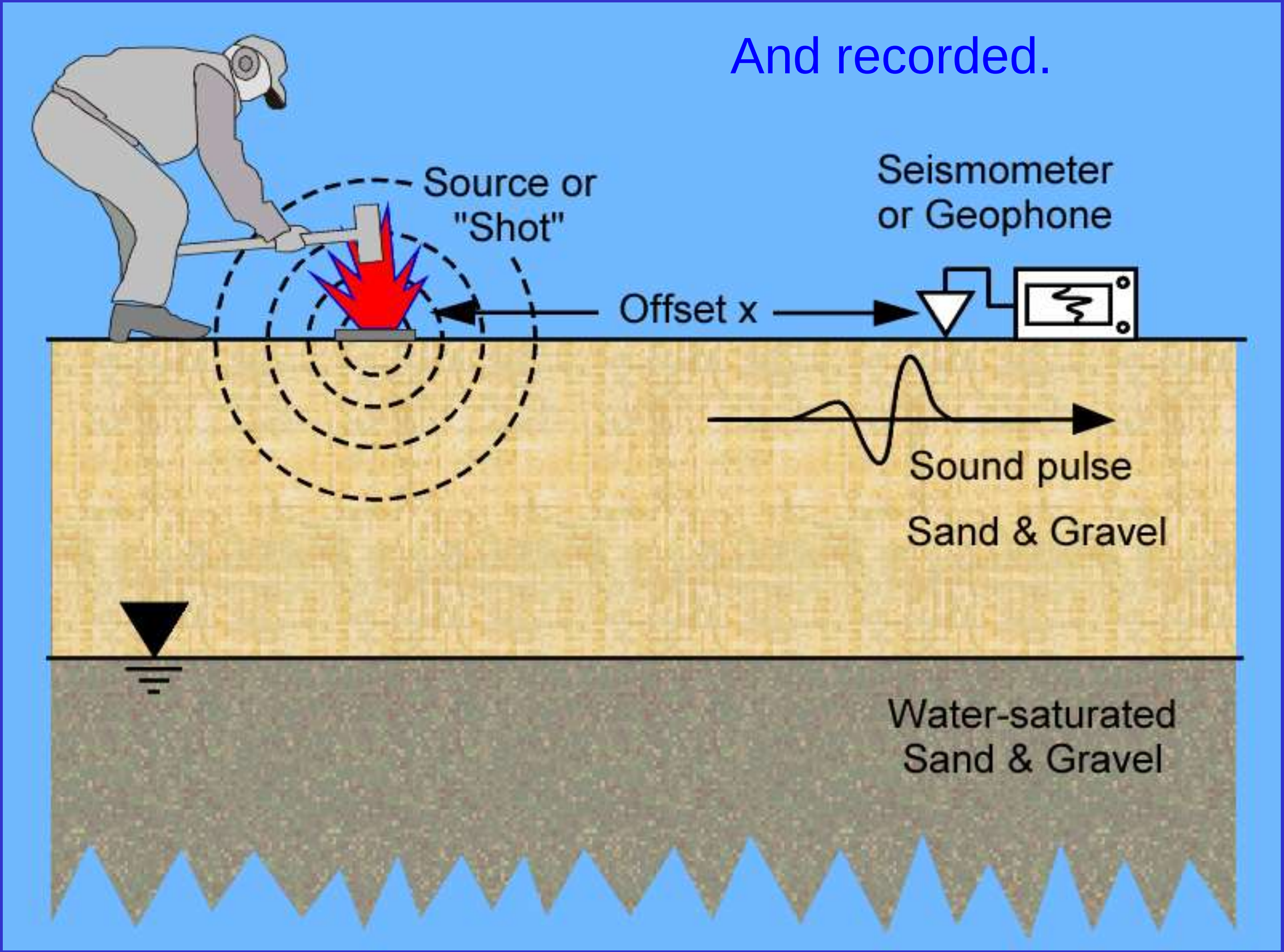
A “Shot”.


Source or
"Shot"
Seismometer
or Geophone
Offset x
Sand & Gravel
Water-saturated
Sand & Gravel

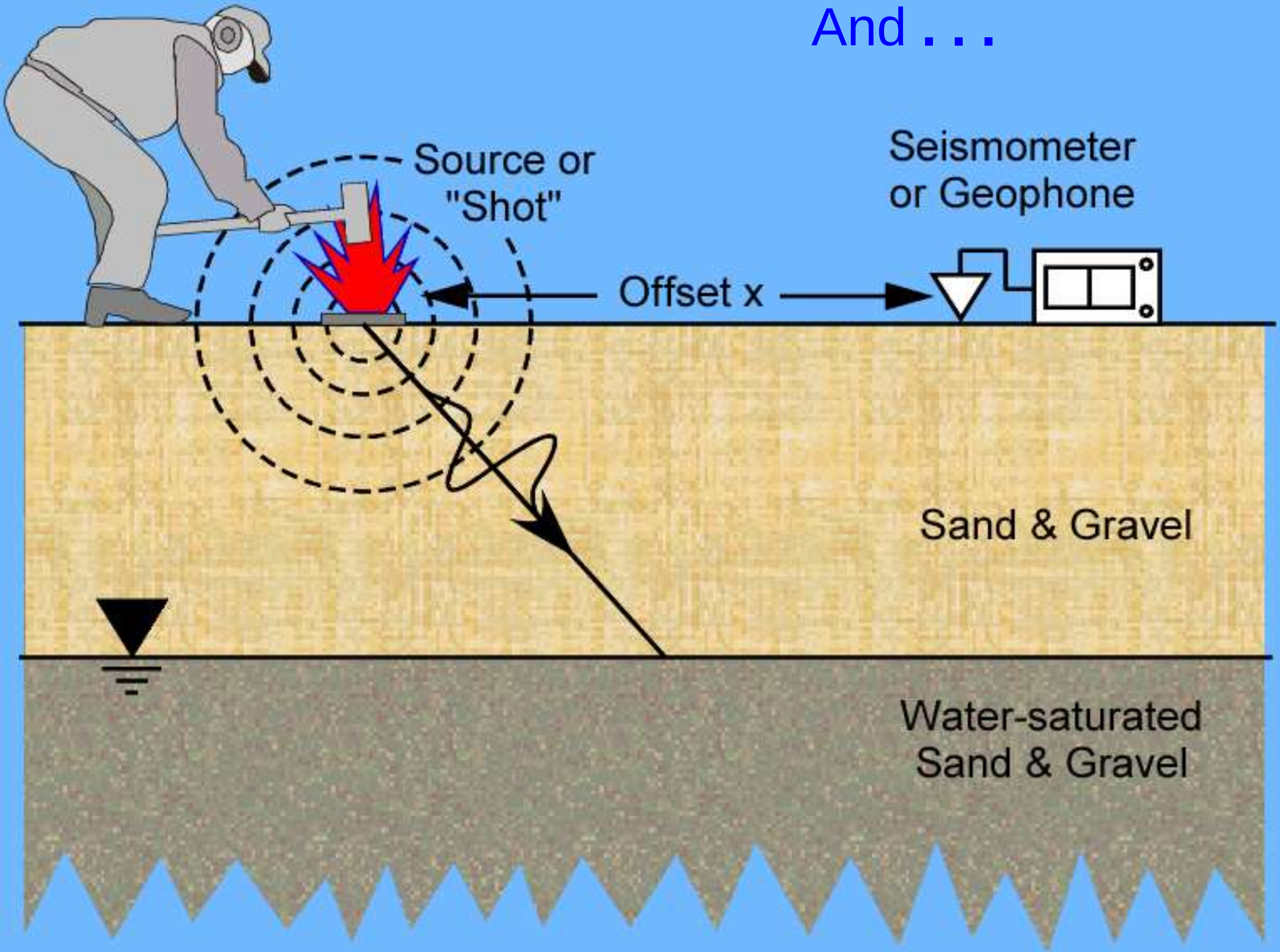
A sound pulse is generated.



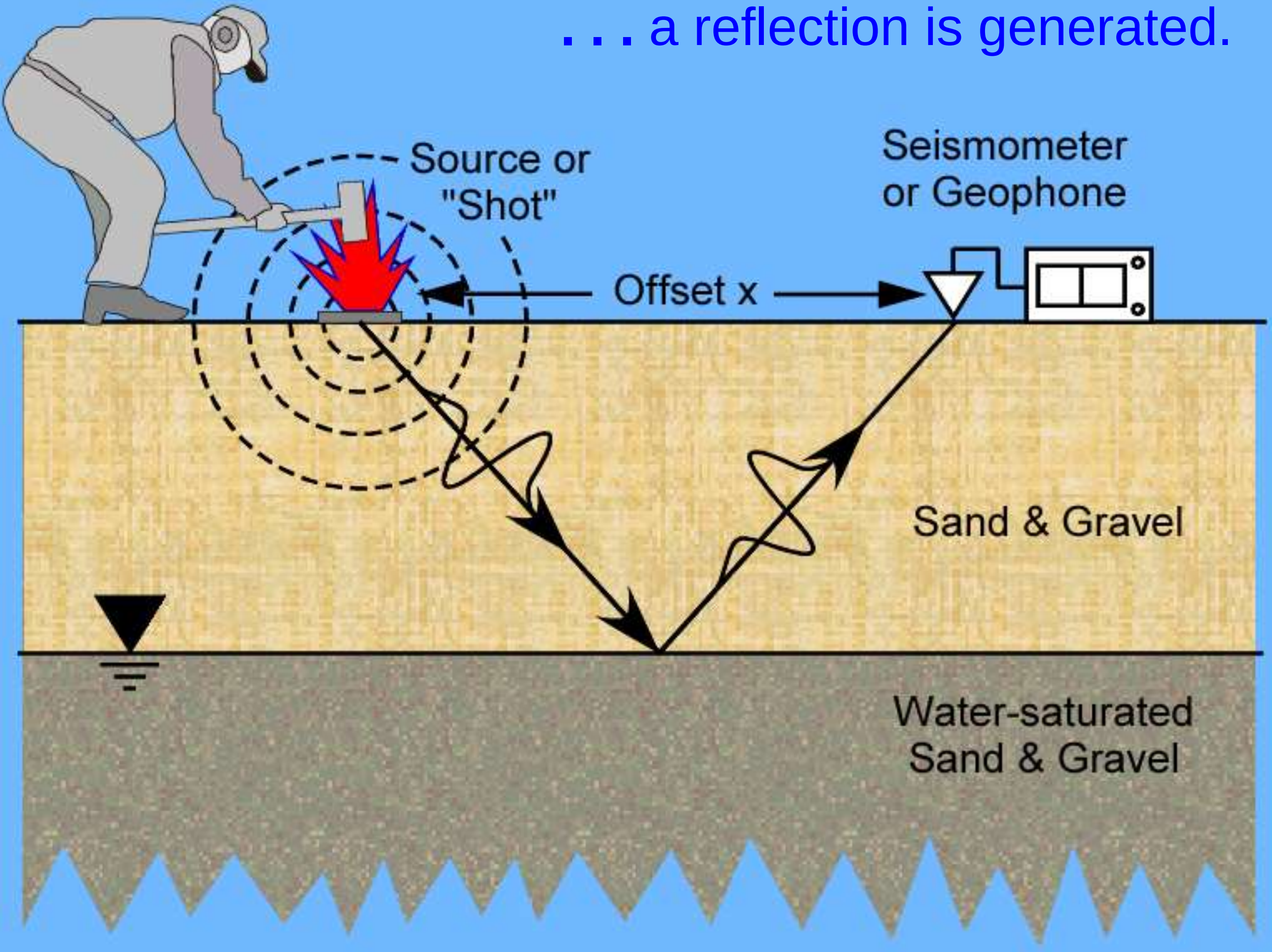
And recorded.



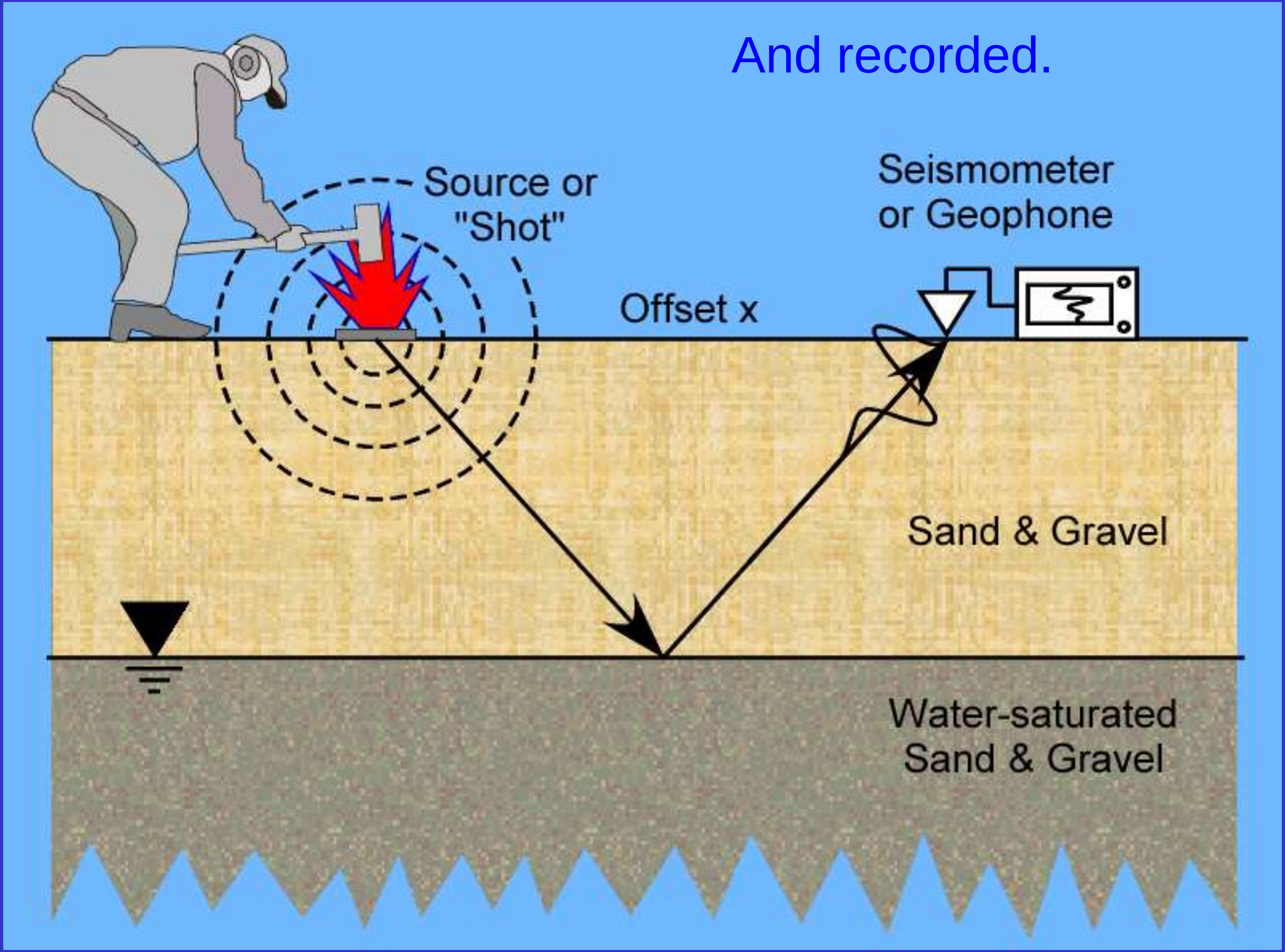
And . . .

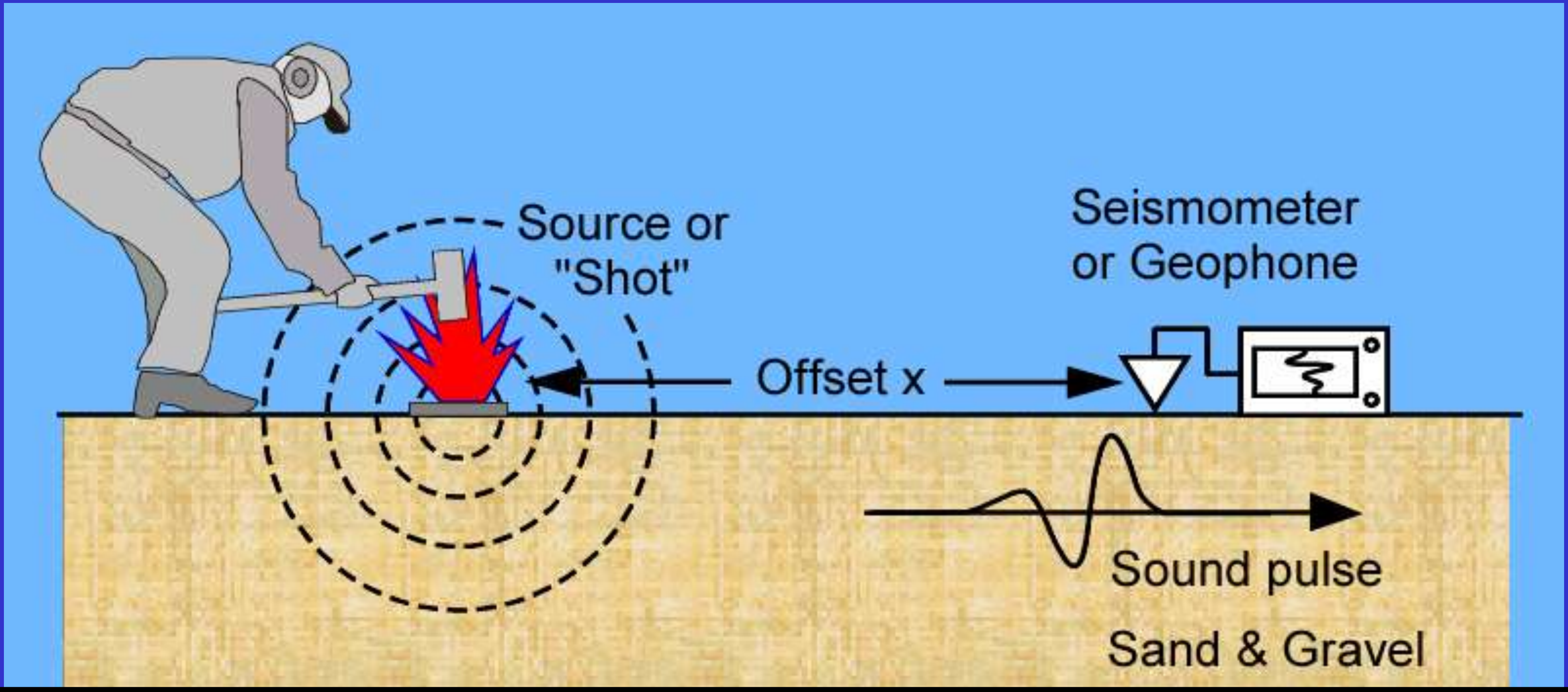

Source or
"Shot"
Seismometer
or Geophone
Offset x
Sand & Gravel
Water-saturated
Sand & Gravel

. . . a reflection is generated.

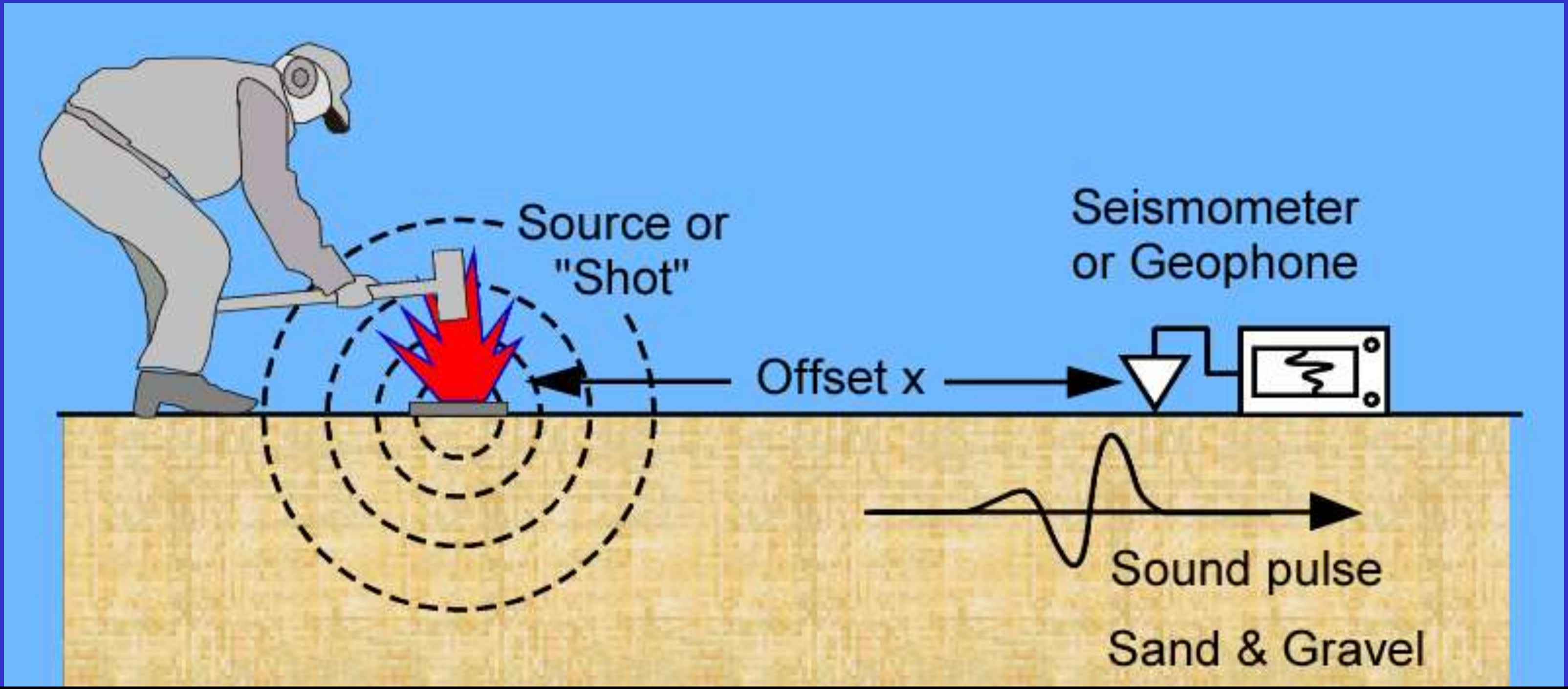


And recorded.

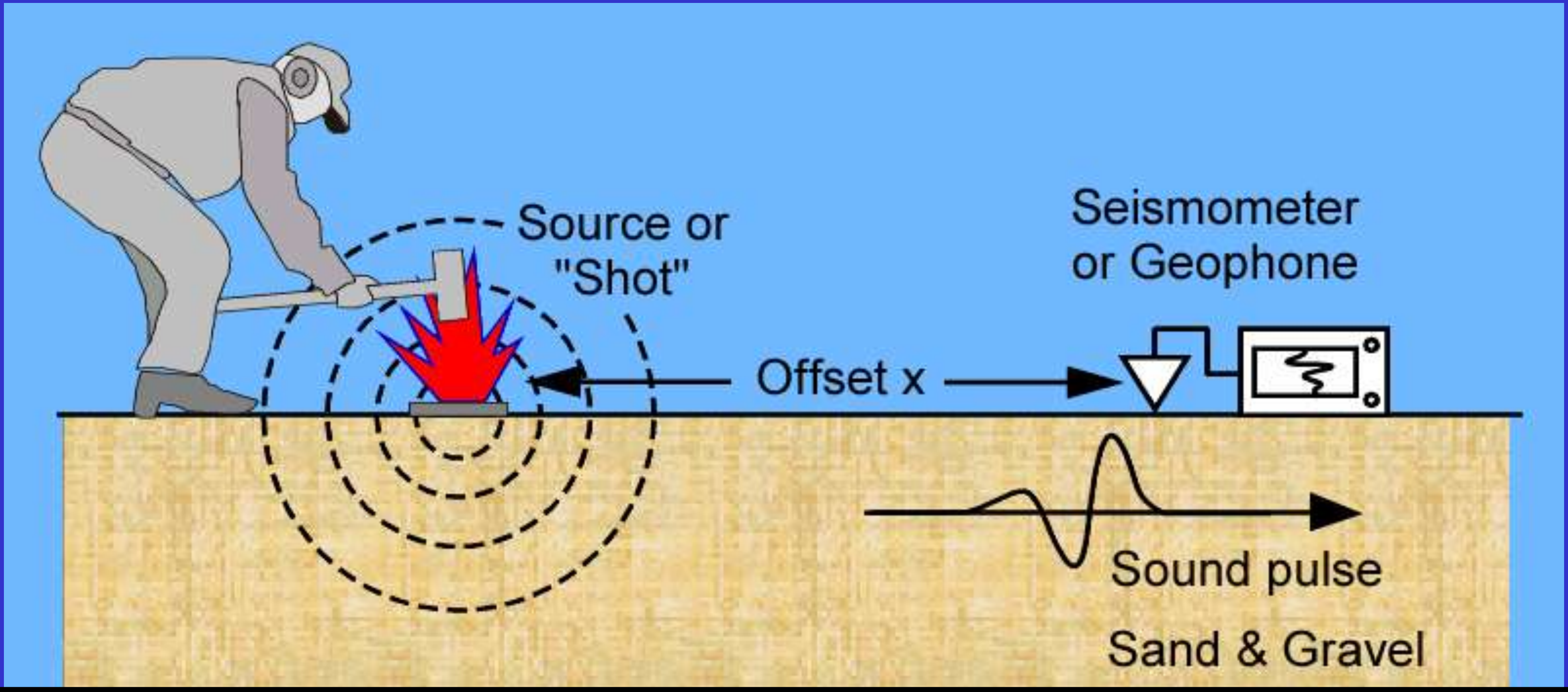




To summarize: An impulsive source (a sledge hammer blow to a steel plate) generates a sound wave that travels through the subsurface. . . .



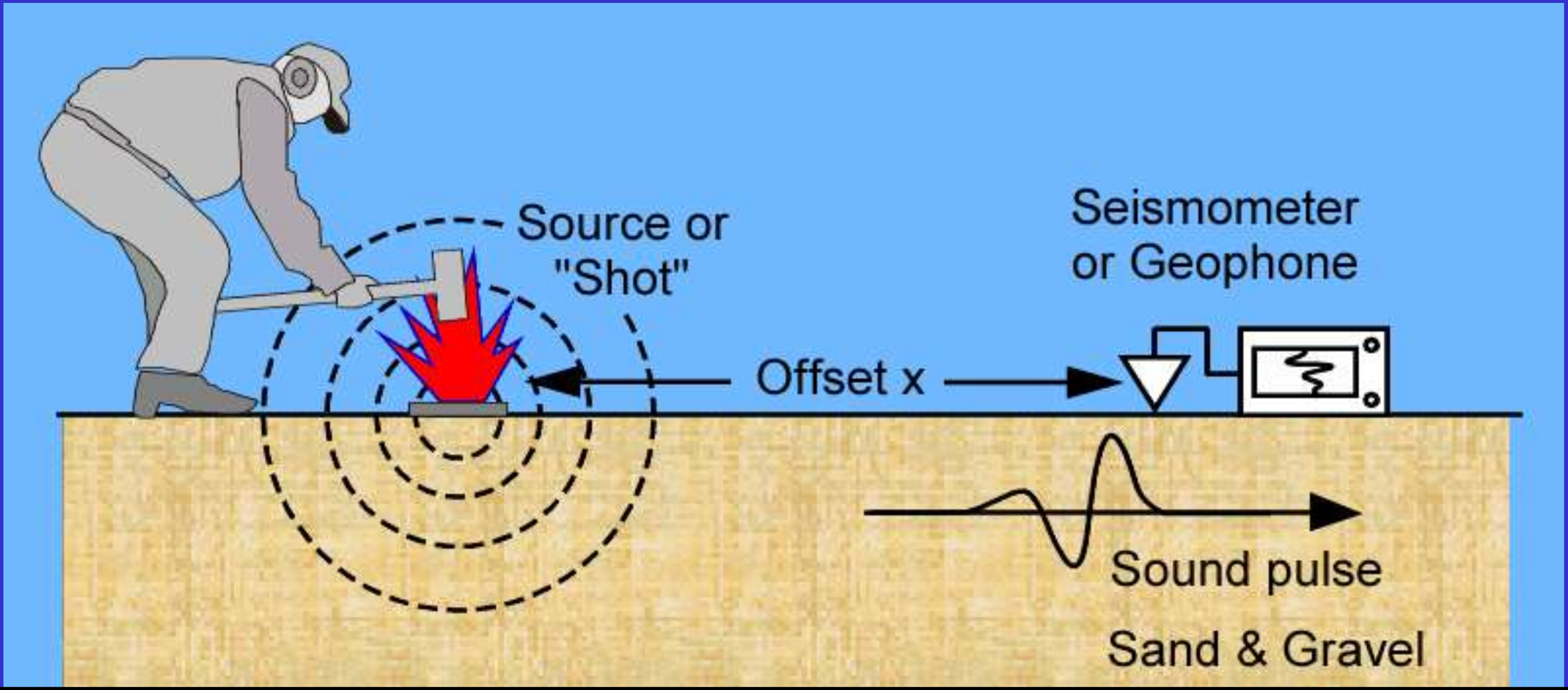
. . . If one knows the distance (x) between the “shot” and the sensor, and the time (T) it takes the wave to travel this distance, . . .



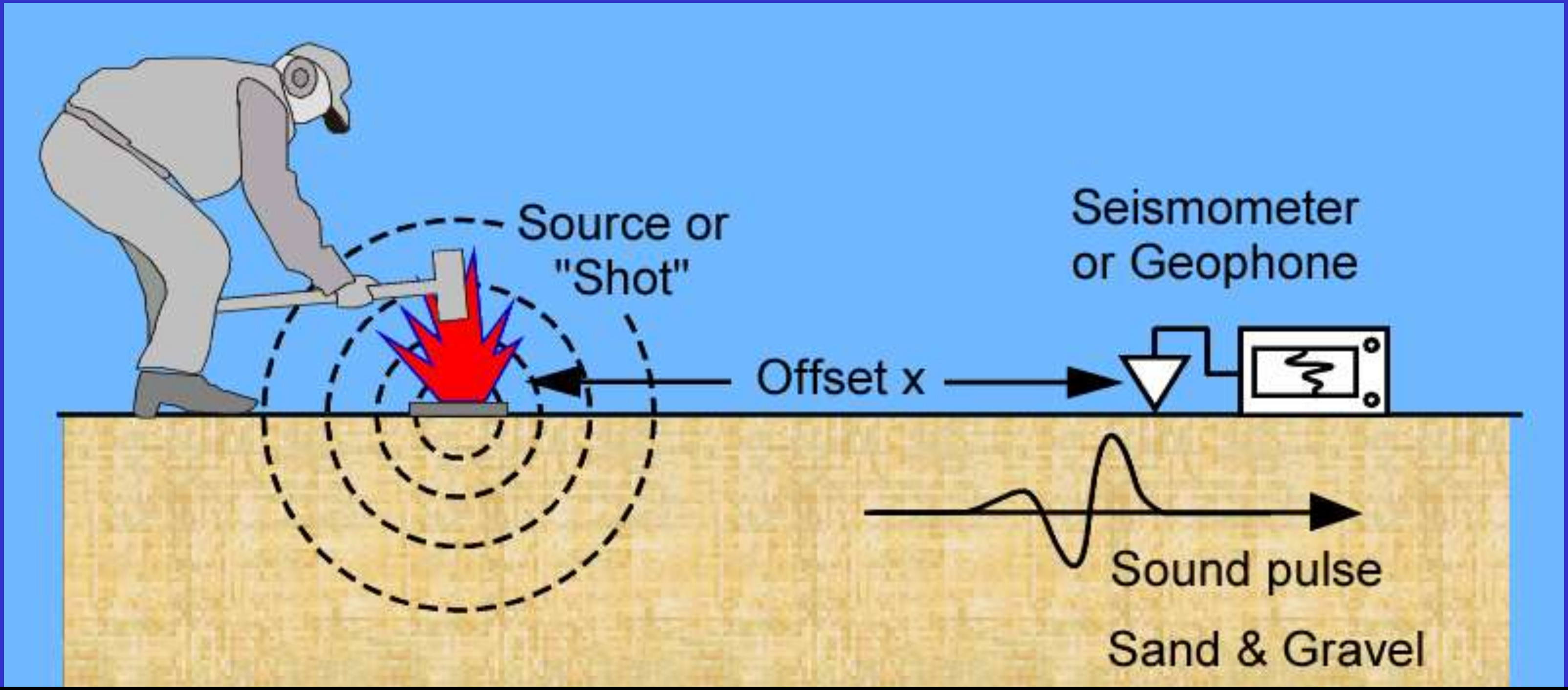
. . . one can determine the velocity (V) of the material,

$$\boxed{V = x / T}$$

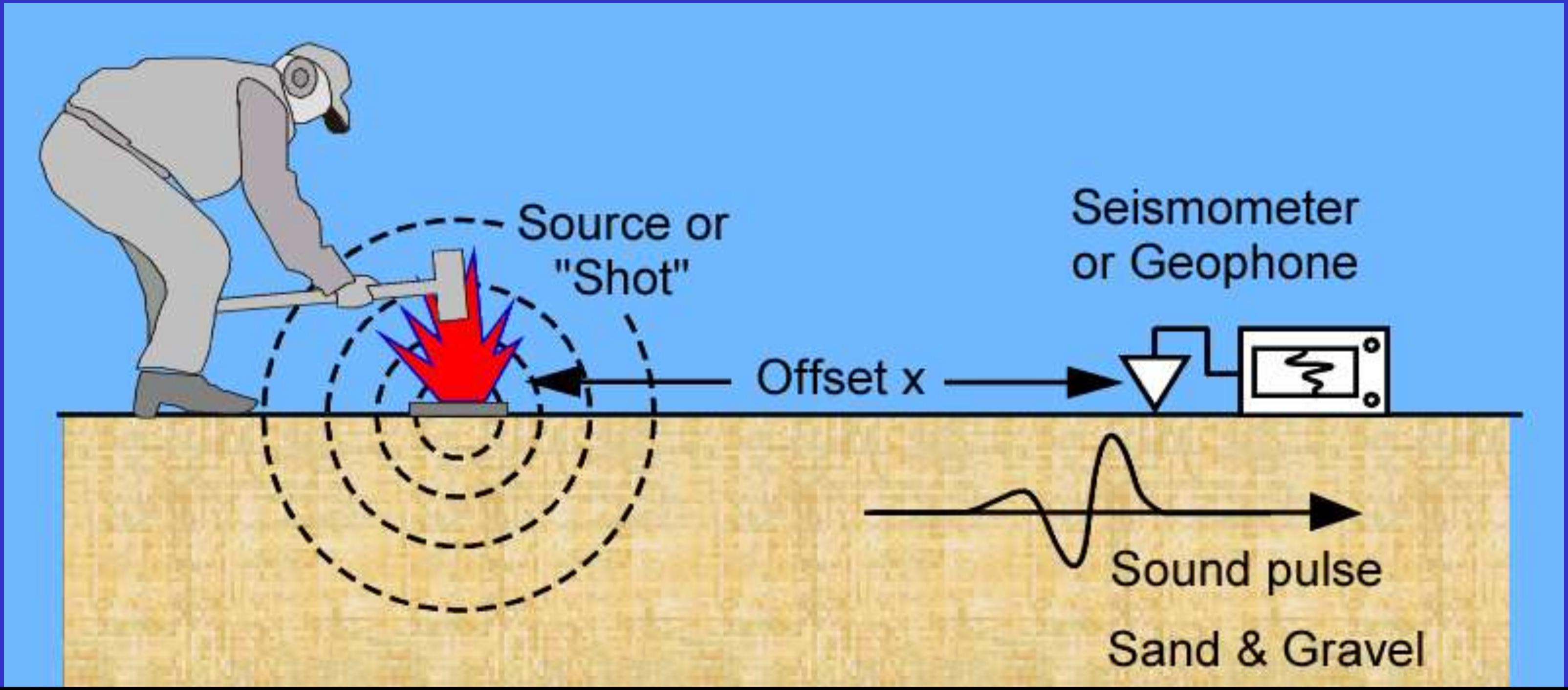
and tell, for example, . . .



. . . whether the medium is
bedrock,
dry soil, or
saturated soil,
among other possibilities.



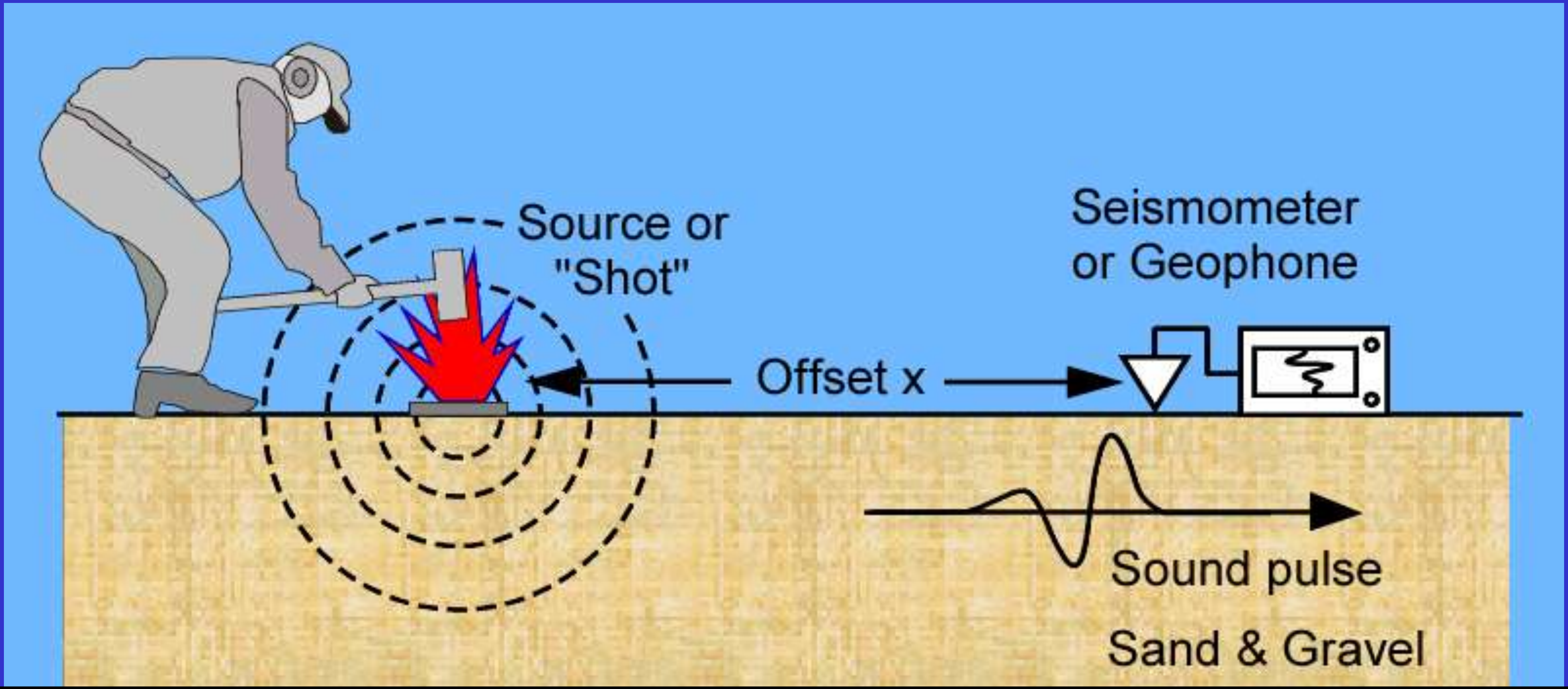
In this way, we determine the <u>material properties</u> of the subsurface.



Alternatively, if one knows the velocity (V) of the material and the time (T) it takes the wave to get to a sensor, then rearranging

$$V = x / T$$

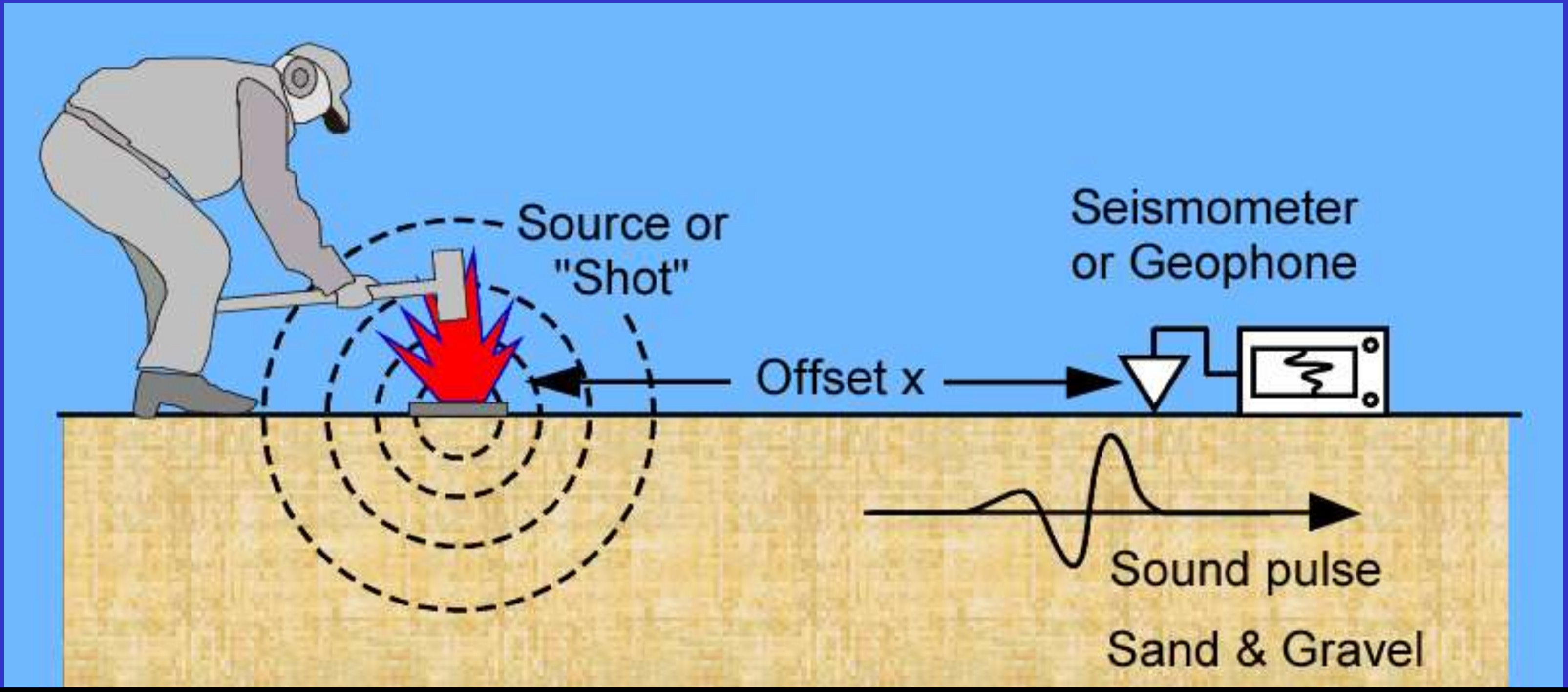
. . .



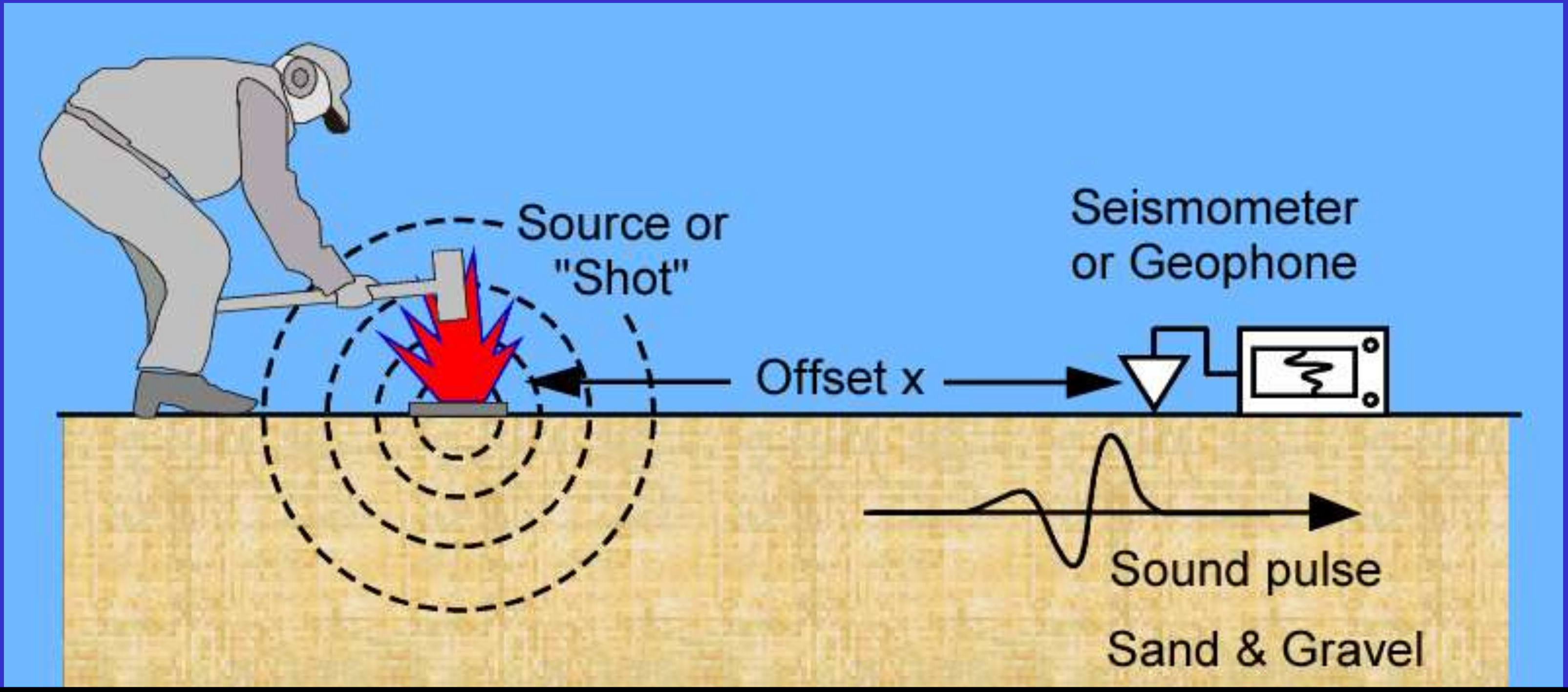
. . . to the form

$$\boxed{x = V / T ,}$$

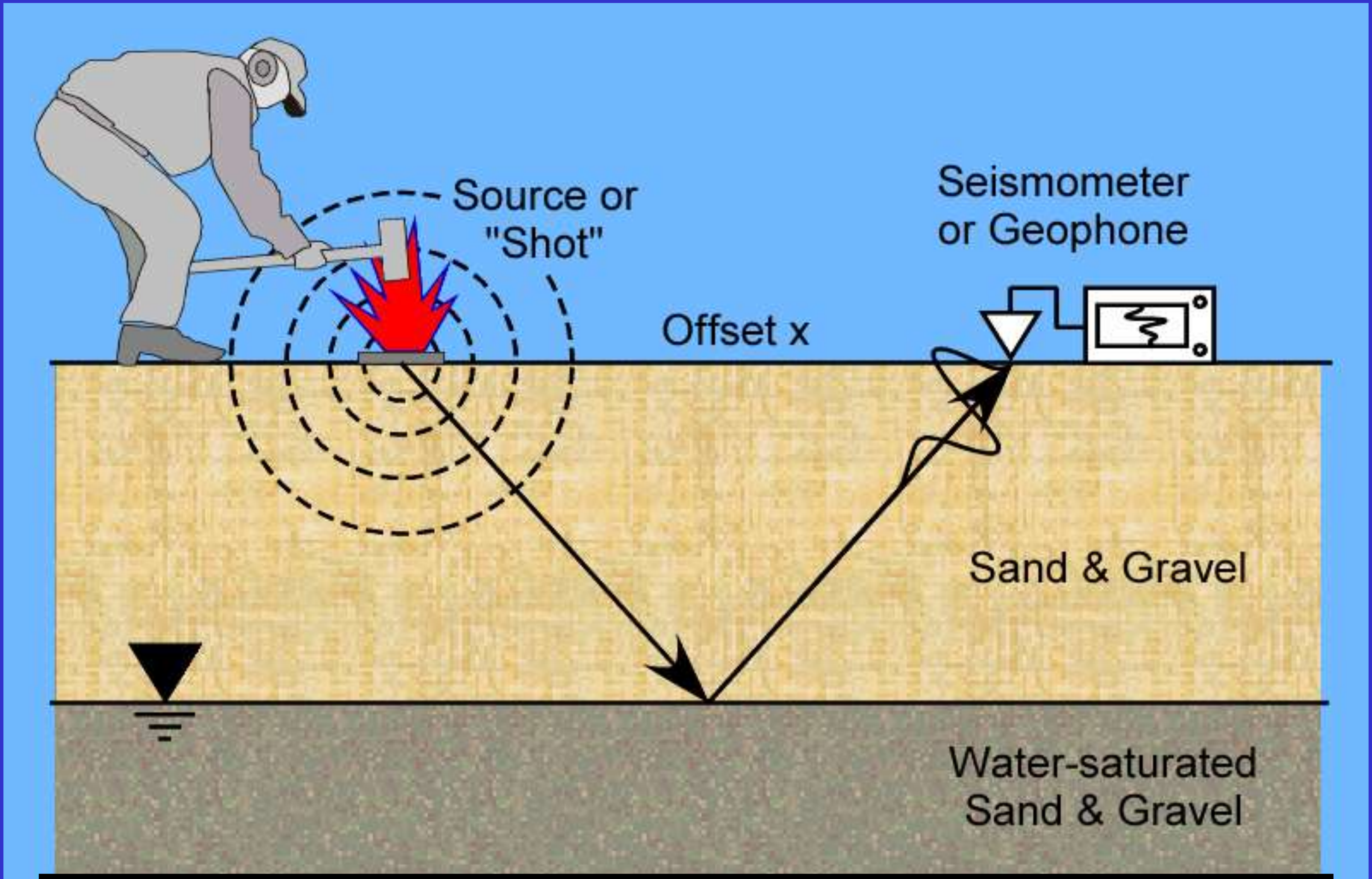
one can determine the distance (x) from the shot to the sensor.



While not particularly useful for the case above, . . .



. . . the latter concept is critical for determining the nature of structures at depth below the surface. For example, when there are <u>layers</u> at depth.



Determining the depth when V and T are known is the principle of the reflection method.

Theory: Behavior of Waves in the Subsurface

In order to understand how to extract more detailed subsurface information from geophysical measurements at the surface, we first analyze the behavior of waves (seismic or radar) in the subsurface.

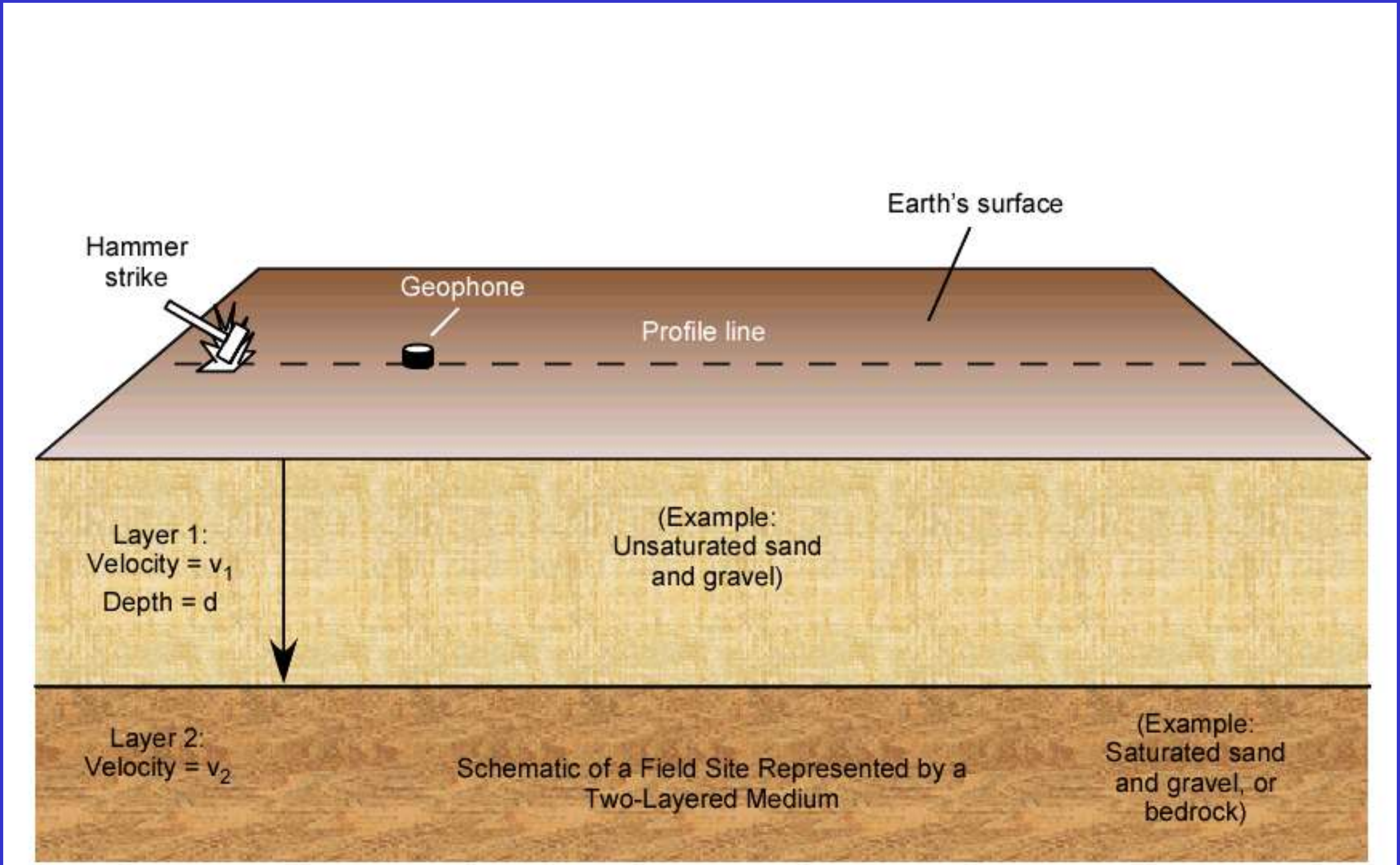
Mathematical Underpinnings: Traveltimes of Principal Phases

1) Direct Wave

$$T(x) = x / V_1$$

2) Reflected Wave

$$T(x) = \sqrt{(x / V_1)^2 + (2d / V_1)^2}$$



This is the field situation to be considered.

Please review the animation sequence for
Reflected Phases at this time.

Please minimize this application, the animation
sequence is found on the index page.

Maximize this application when ready to continue.

Essential points for discussion.

1) The relative difference in arrival times of the ‘direct’ and ‘reflected’ phases as offset increases.

Essential points for discussion.

1) The relative difference in arrival times of the ‘direct’ and ‘reflected’ phases as offset increases.

2) The synchrony of the two phases along the lower interface.

Essential points for discussion.

1) The relative difference in arrival times of the ‘direct’ and ‘reflected’ phases as offset increases.

2) The synchrony of the two phases along the lower interface.

3) The difference in the ‘apparent’ velocity of the two phases along the surface

a) The direct (primary) wave travels @ v_1.
b) The reflected wave @ $v_1 / \sin \theta_i$
(where θ_i is the incident angle).

Essential points for discussion.

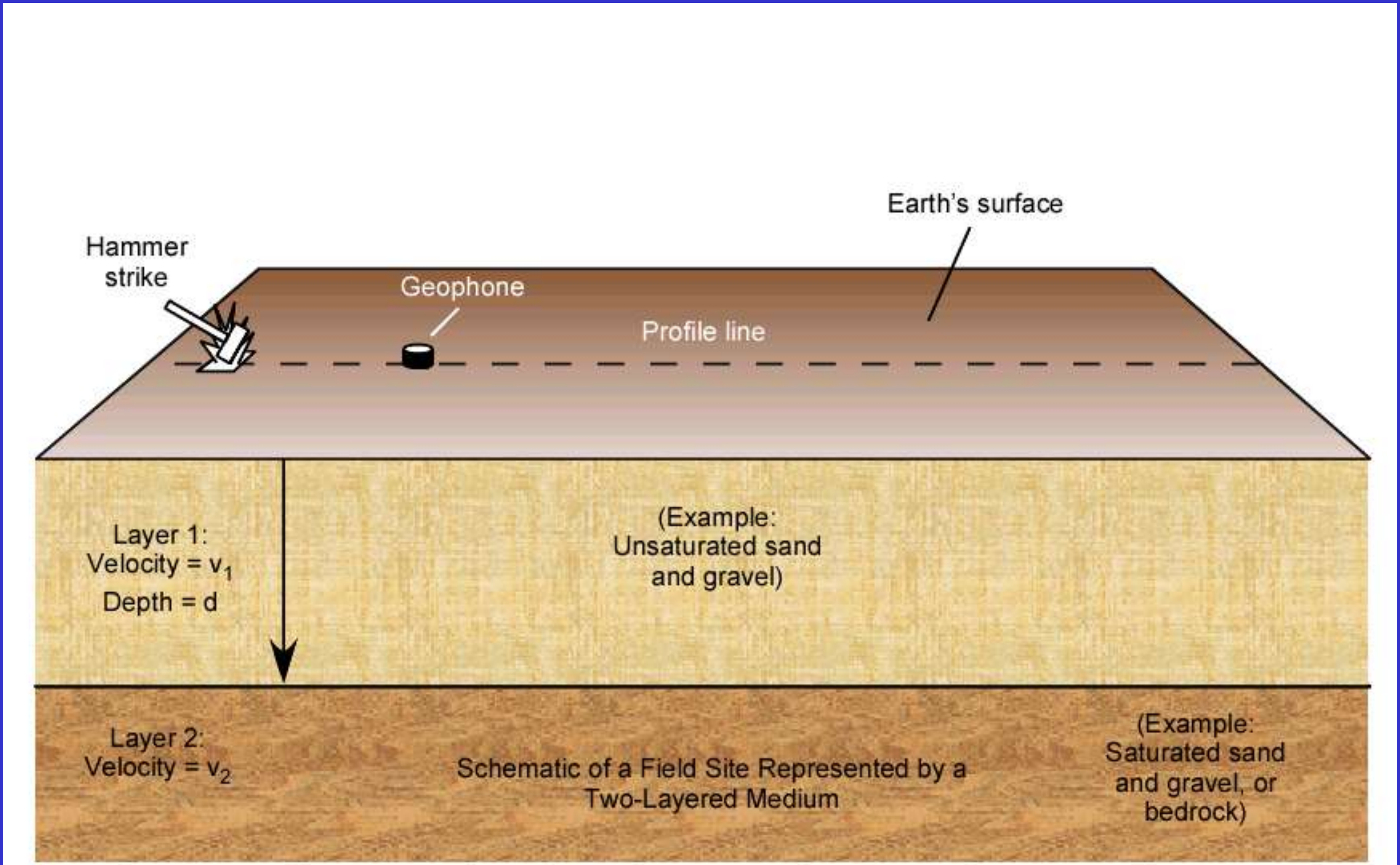
1) The relative difference in arrival times of the ‘direct’ and ‘reflected’ phases as offset increases.

2) The synchrony of the two phases along the lower interface.

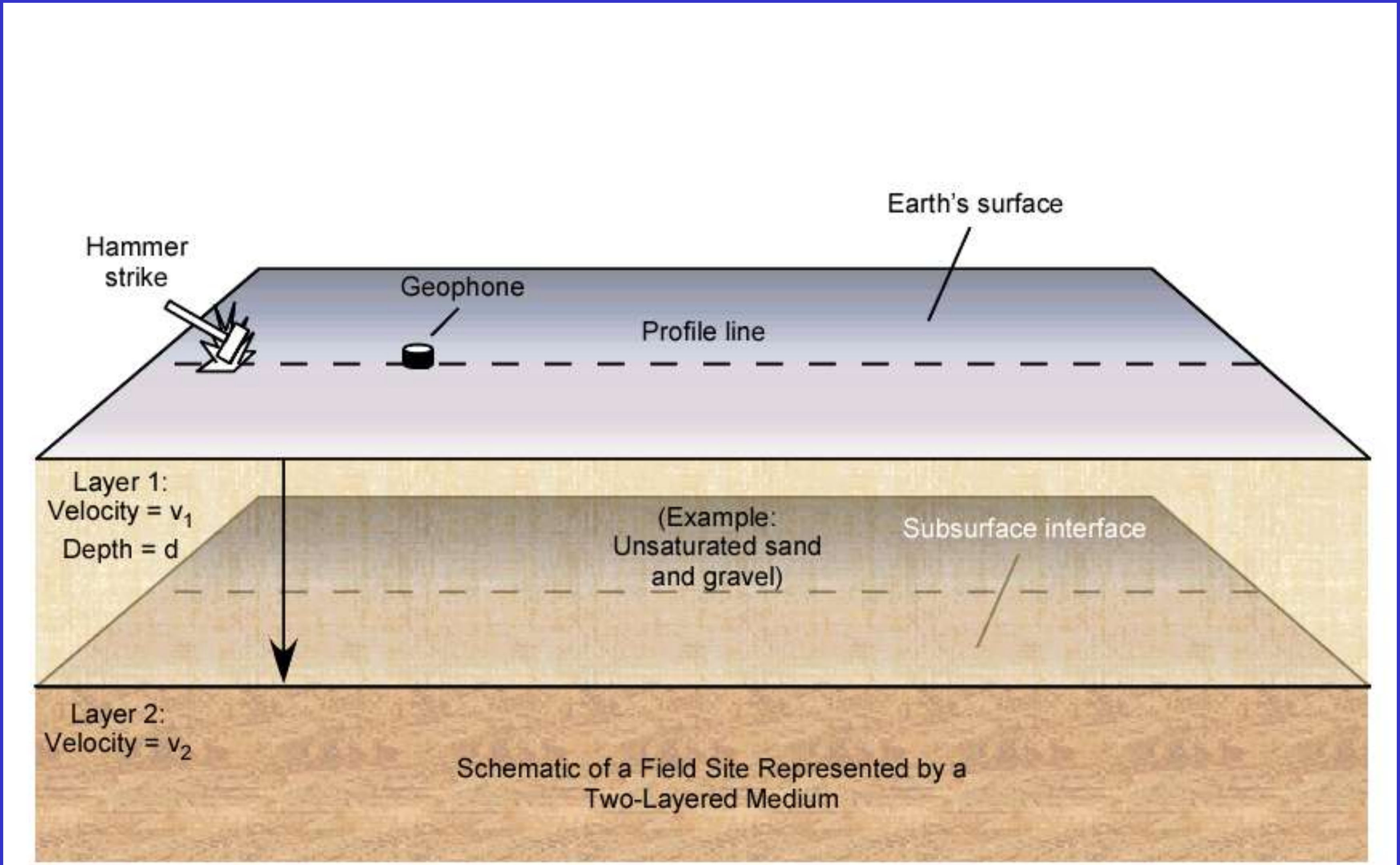
3) The difference in the ‘apparent’ velocity of the two phases along the surface

- a) The direct (primary) wave travels @ v_1.
- b) The reflected wave @ $v_1 / \sin \theta_i$ (where θ_i is the incident angle).

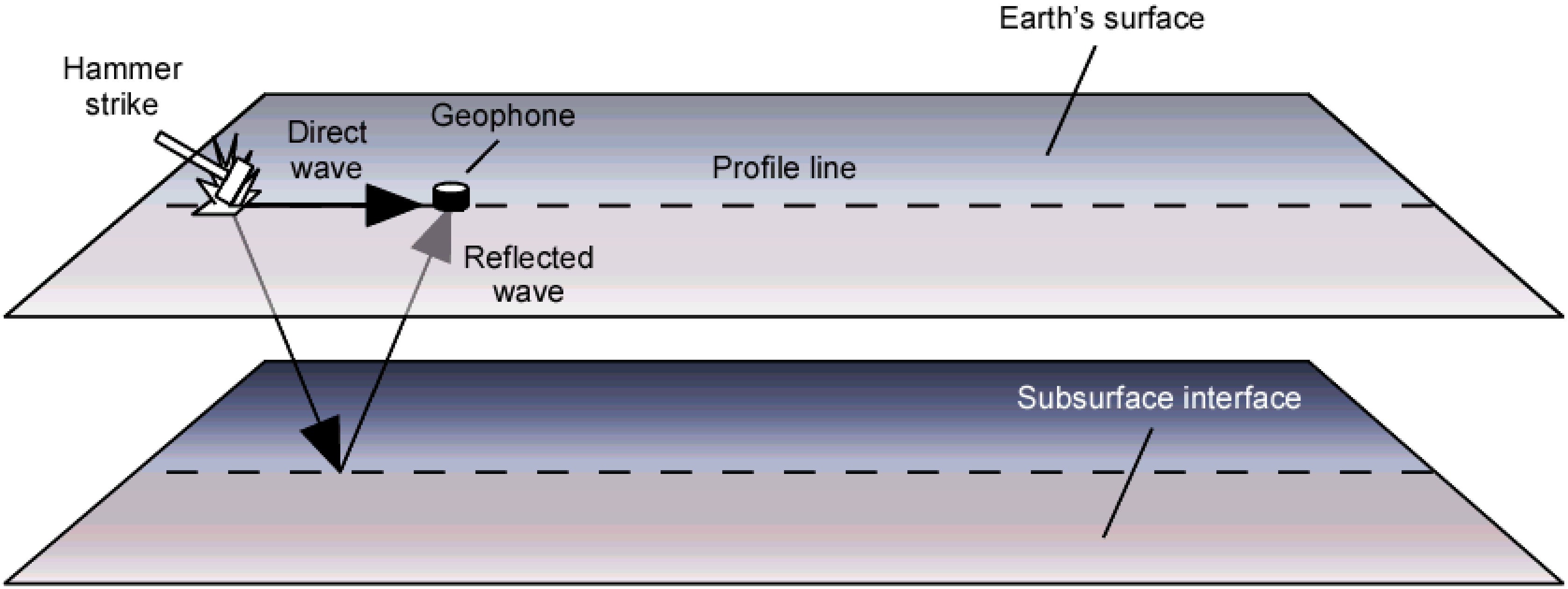
Traveltime Relations for Direct and Reflected Phases



The field situation.

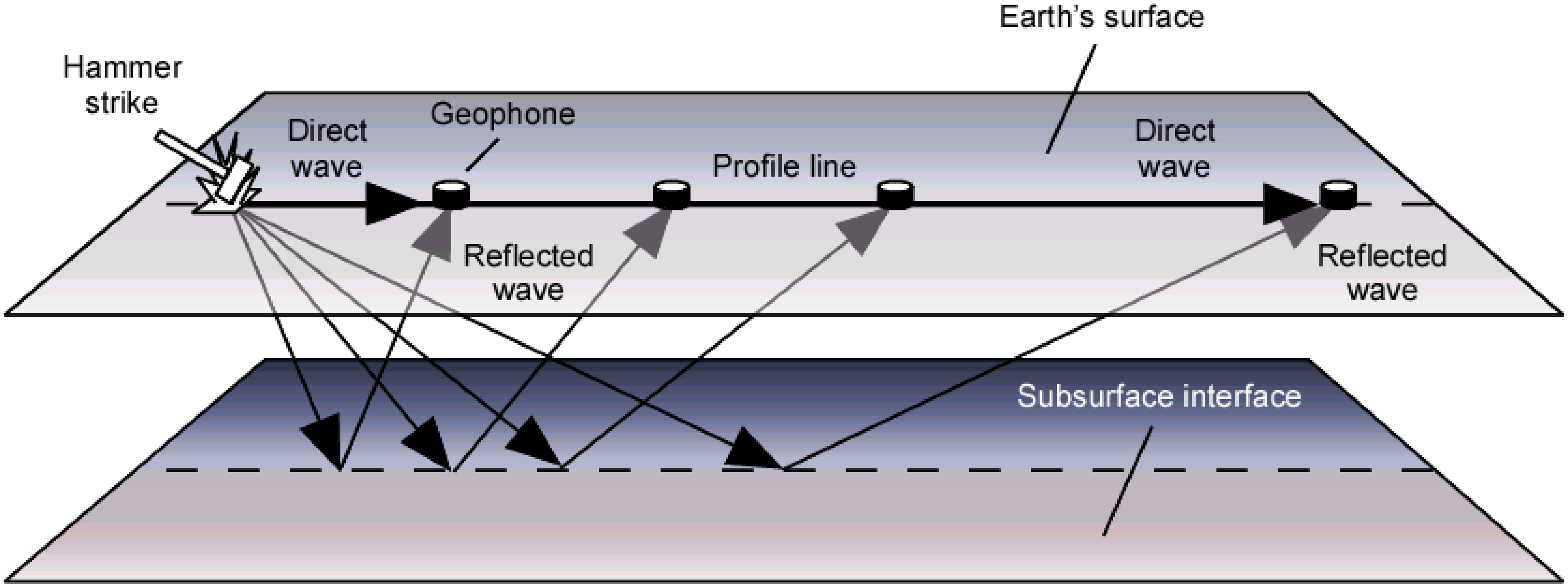


The field situation showing cutaway.



Direct Wave and Wave Reflected
From Subsurface Interface

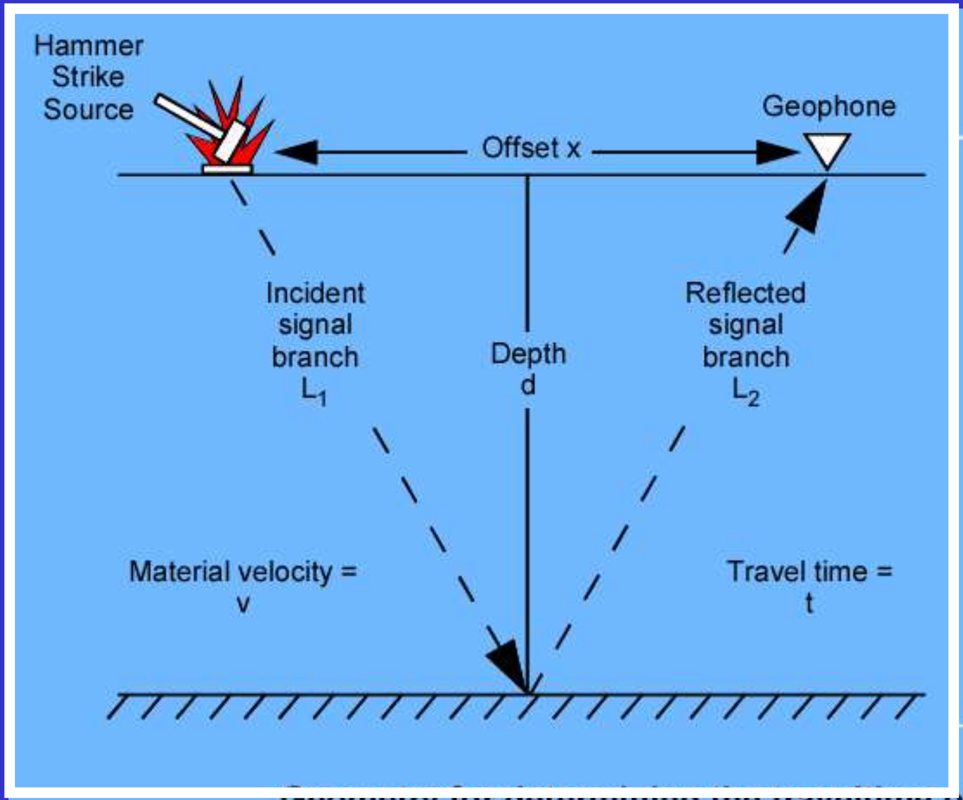
The field situation with geophone.



Direct Wave and Wide-Angle
Reflected Wave

Direct and reflected ray paths.

How can we use the ‘reflected’ phase to determine the depth to the respective horizon (or layer) ?



Derivation:

Distance the reflected signal travels is

$$L_{tot} = L_1 + L_2 = v\ t$$

From the Pythagorean theorem

$$(vt) = 2\ [(x/2)^2 + (d)^2]^{1/2}$$

or

$$(vt)^2 = (x)^2 + (2d)^2$$

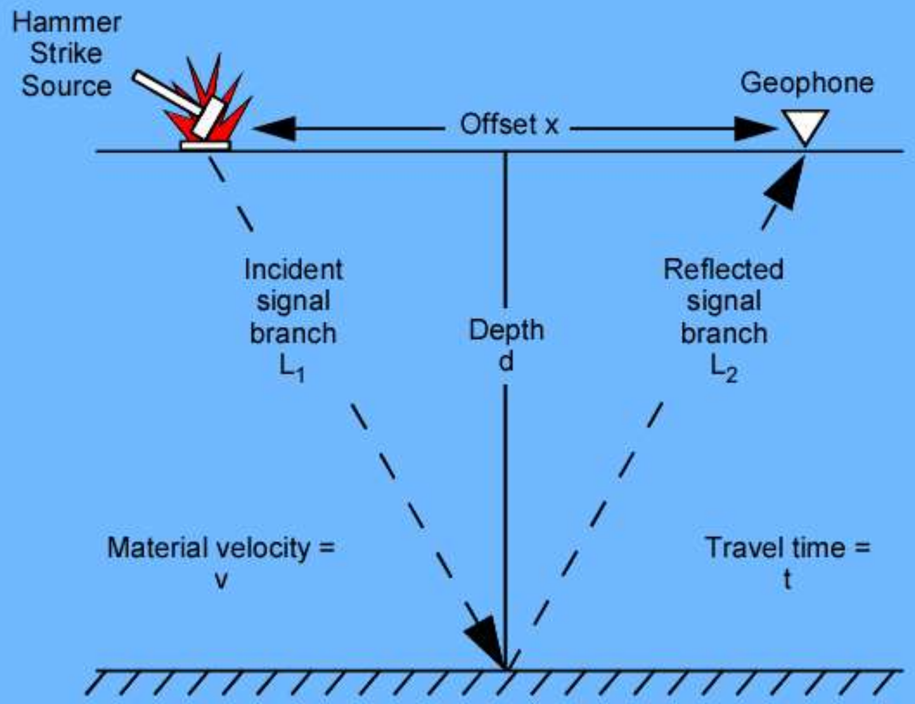
This can be rearranged for the travel time

$$t = (1/v)\ [\ (x)^2 + (2d)^2\ ]^{1/2}$$

Geometry for determining the traveltime of reflections from a horizontal interface at depth.
From the travelime, we determine the reflector depth:

$$\boxed{d = (1/2)\ [\ (vt)^2 - (x)^2\ ]^{1/2}}$$

Geometry for a reflection.



Derivation:

Distance the reflected signal travels is

$$L_{tot} = L_1 + L_2 = v\ t$$

From the Pythagorean theorem

$$(vt) = 2\ [(x/2)^2 + (d)^2]^{1/2}$$

or

$$(vt)^2 = (x)^2 + (2d)^2$$

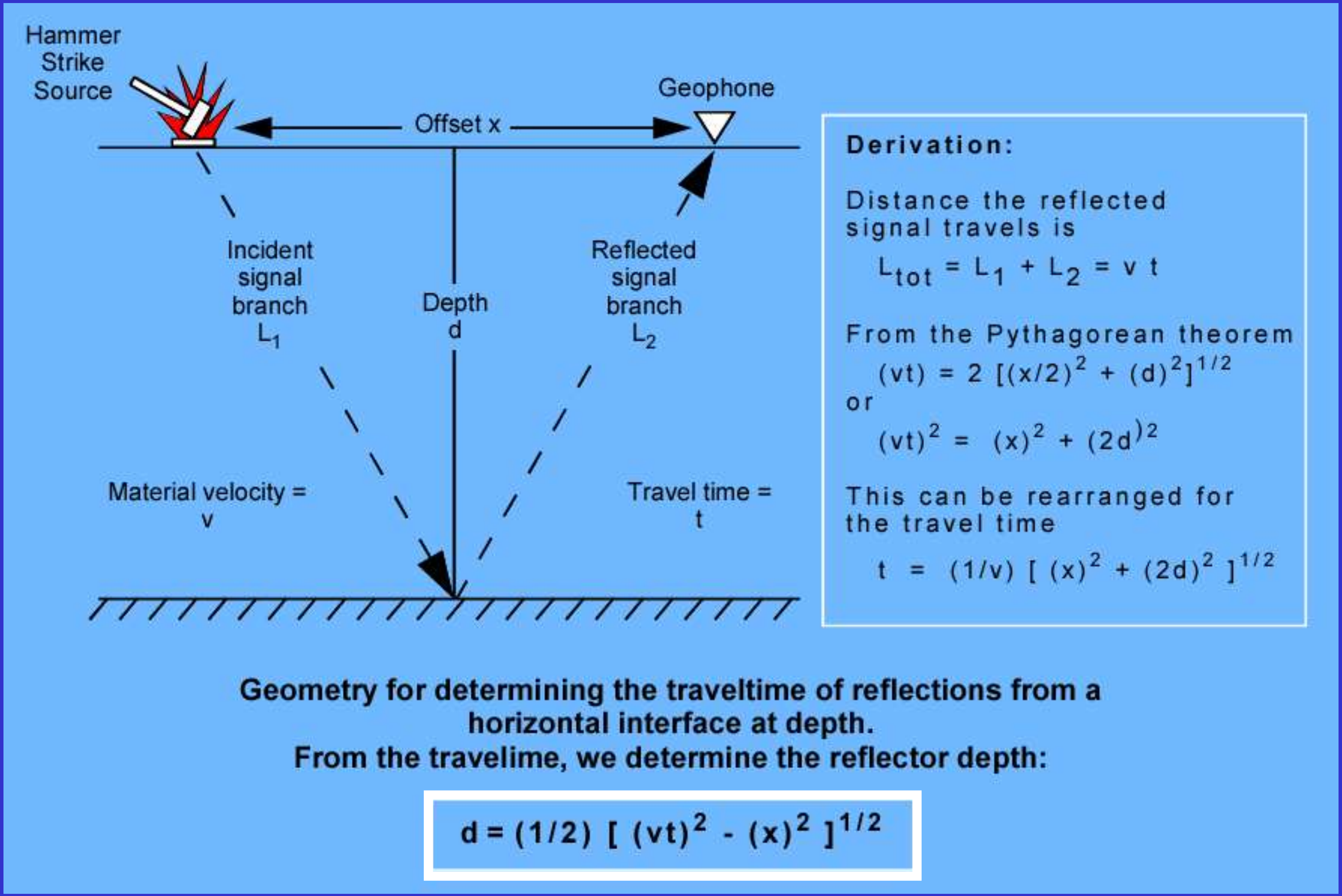
This can be rearranged for the travel time

$$t = (1/v)\ [\ (x)^2 + (2d)^2\ ]^{1/2}$$

Geometry for determining the traveltime of reflections from a horizontal interface at depth.
From the travelime, we determine the reflector depth:

$$\boxed{d = (1/2)\ [\ (vt)^2 - (x)^2\ ]^{1/2}}$$

Traveltime for a reflection.



Derivation:

Distance the reflected signal travels is

$$L_{tot} = L_1 + L_2 = v\ t$$

From the Pythagorean theorem

$$(vt) = 2\ [(x/2)^2 + (d)^2]^{1/2}$$

or

$$(vt)^2 = (x)^2 + (2d)^2$$

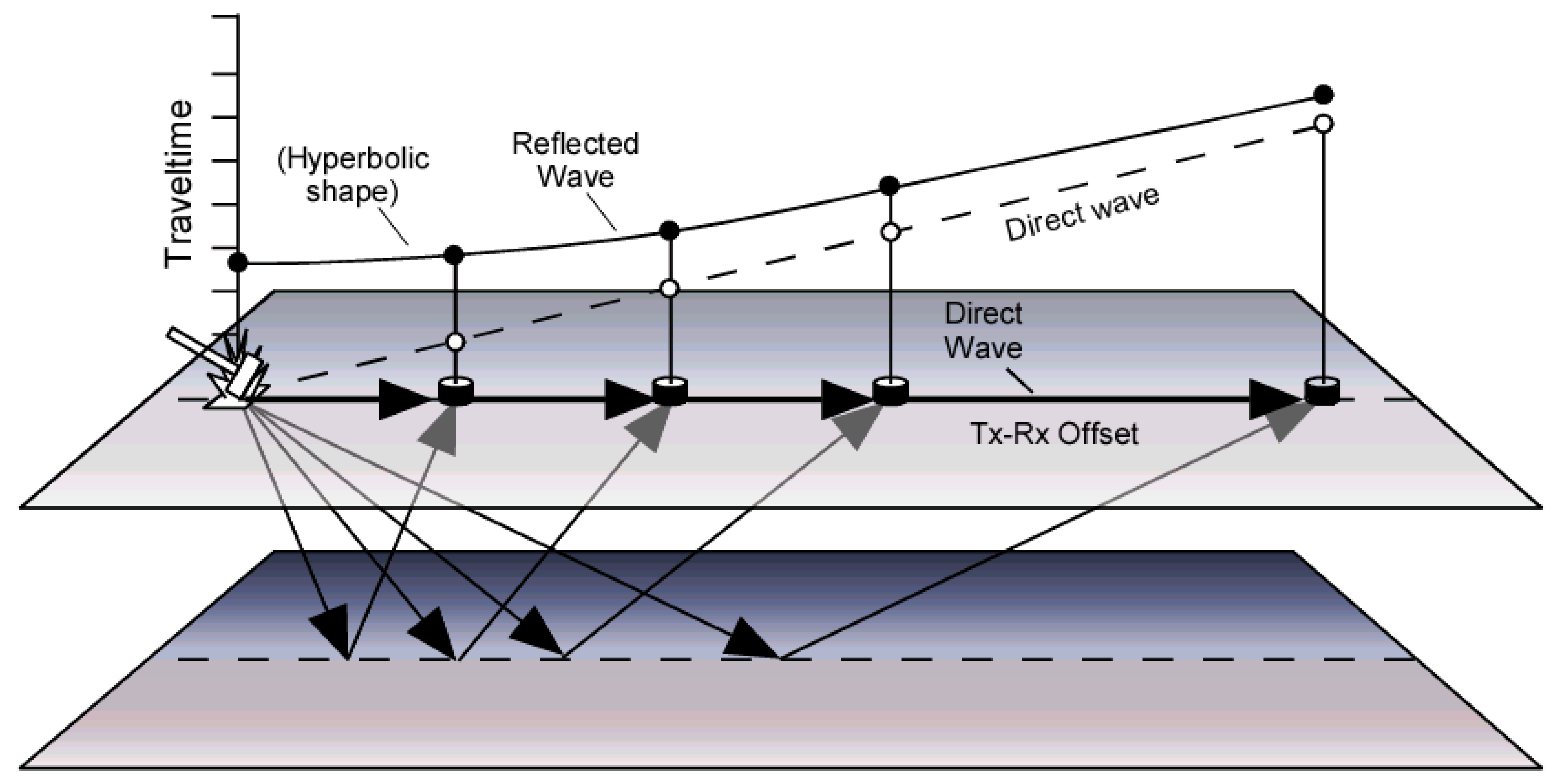
This can be rearranged for the travel time

$$t = (1/v)\ [\ (x)^2 + (2d)^2\ ]^{1/2}$$

Geometry for determining the traveltime of reflections from a horizontal interface at depth.
From the travelime, we determine the reflector depth:

$$\boxed{d = (1/2)\ [\ (vt)^2 - (x)^2\ ]^{1/2}}$$

If we “know” x and v, we can determine d, the depth to the reflector.



Travel time of reflected wave compared to direct wave.

Direct and reflected ray paths with traveltimes.

Please review tutorial on **Analyzing Direct and Reflected Phases** at this time.

Please minimize this application, the tutorial is found on the index page.

Maximize this application when ready to continue.

Sources of "noise"

Direct air wave
Ground-roll
Wind noise
Rain noise
Traffic
Heavy machinery

Specific concerns with seismic methods

- Geologic attenuation
  - Lossy materials
  - Anomalous random scatterers
- High reflectivity masking zones
  - High velocity layers at the surface significantly curtail the transmission of seismic energy to greater depth.
  - In addition, high velocity materials at the surface can lead to strong lateral reflections and refraction.

    Thus, frozen ground, macadam and concrete paved roadways, parking lots, building foundations, etc., can lead to significant aberrations of the seismic signal, often negating the results.

Details on Seismic Reflection Procedures

[Before attempting to implement ***reflection*** methods, one needs to seriously consider the feasibility of doing so.]

"Walk-away" experiments

- The travel time curve: Equation of hyperbola
- Asymptotic behavior at large and small offset distances
- Need to separate depth to reflector from interval velocity

Standard Field and Interpretation Procedures

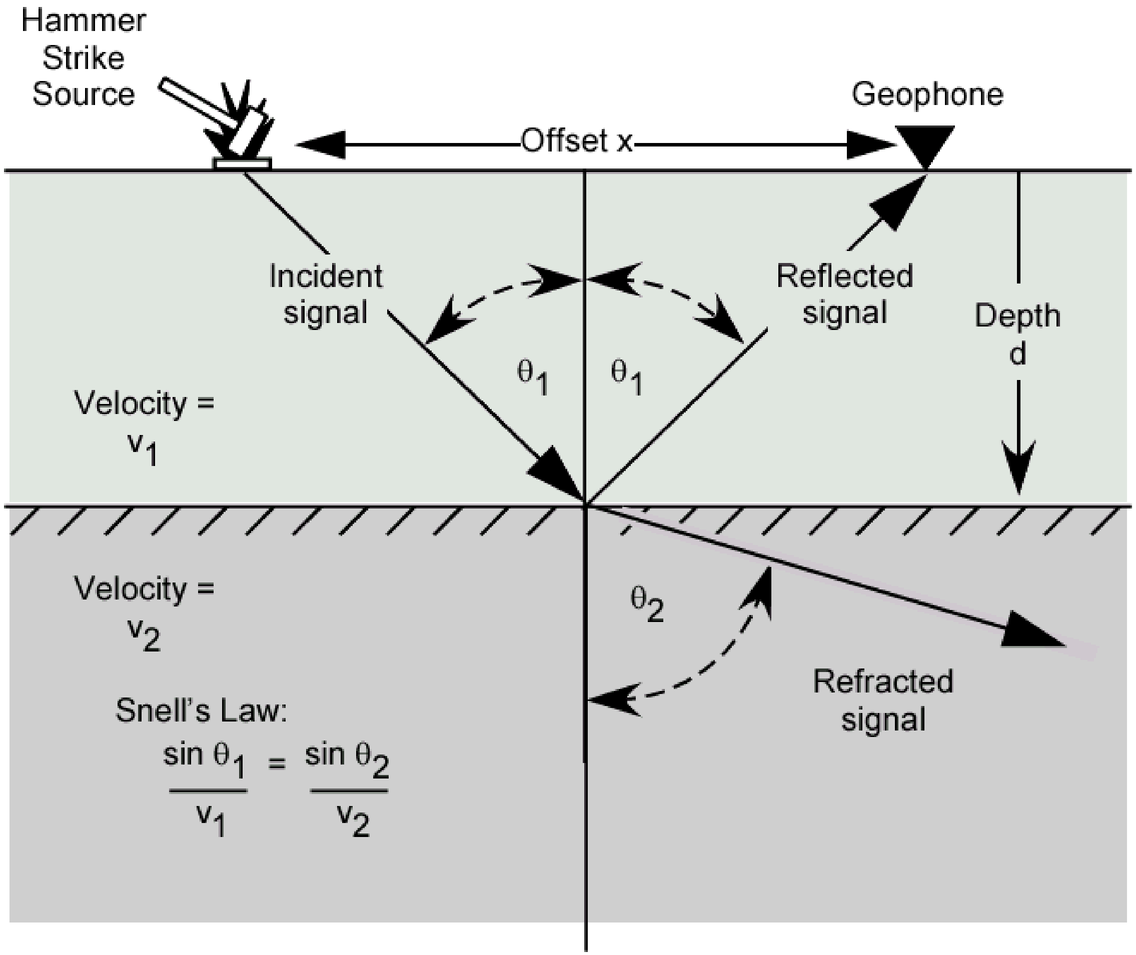
- Cross-correlation of source and reflected signal
- Compensation for normal "step-out"
- Stacking: Common Depth Point etc.
- Migration

Next, consider the ‘refracted’ phase.

Principles of Seismic Refraction Method

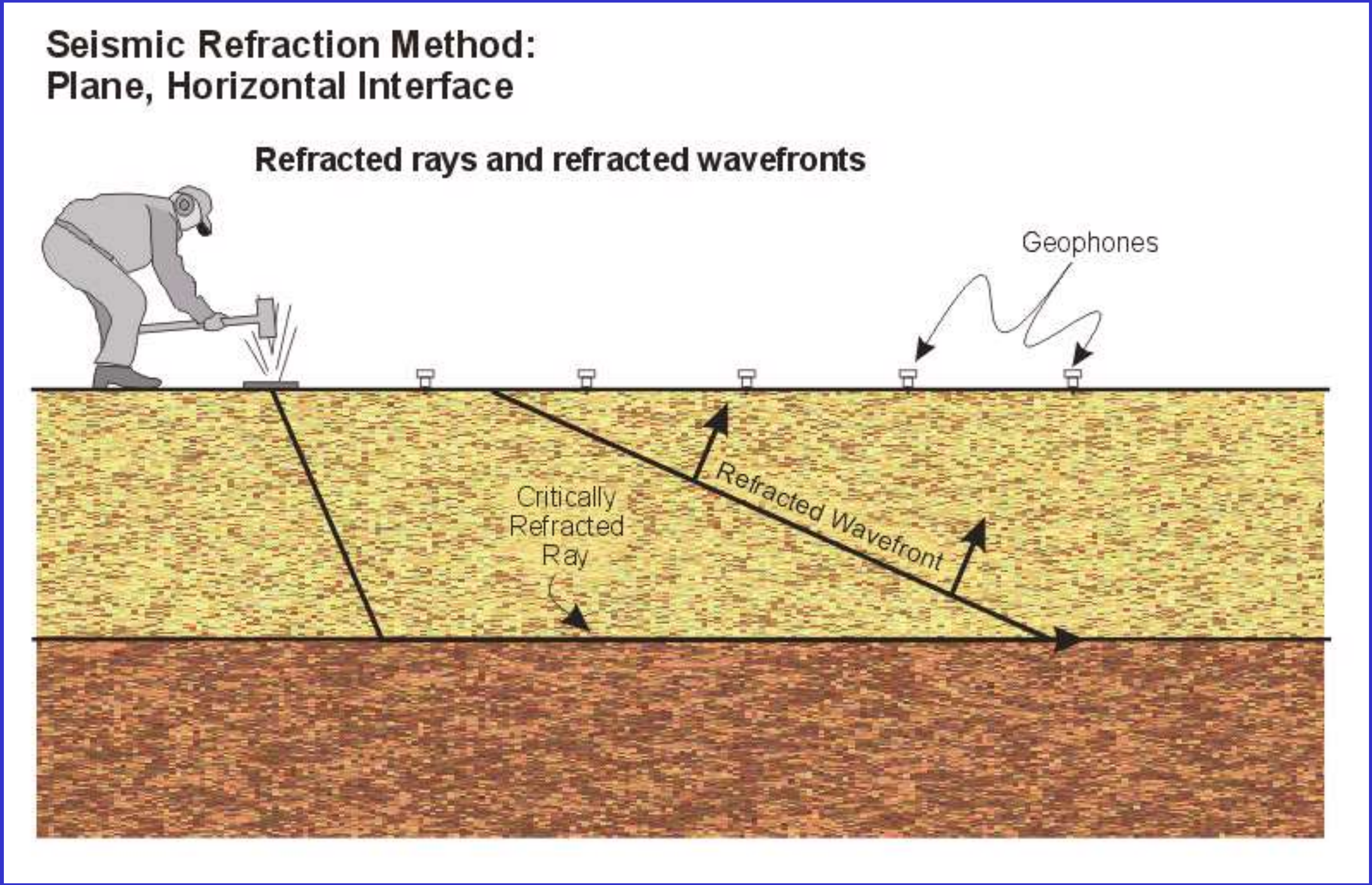
Topics

- Refraction in a medium having continuously varying properties
  - Refraction at an interface; Snell's law
  - Critical refraction and the Head wave
- Travel time curve for refracted wave: constant slope with offset at origin (t = 0)
- Determining depth-of-refractor d_1, and velocities: v_1 and v_2
  - How long does a seismic line have to be for a single depth determination?
- Dipping refractors
  - Concept of "apparent velocity"
  - Reversed profiles
- The delay time or time-term method
- Examples



Snell's Law for Reflection and Refraction

Snell's Law for Reflection and Refraction.



Refraction at the Critical Angle.

Please review the animation sequence for Refracted Phases at this time.

Please minimize this application, the animation sequence is found on the index page.

Maximize this application when ready to continue.

Essential points for discussion.

1) The relative difference in arrival times of the ‘direct’, ‘reflected’ & ‘refracted’ phases as offset increases.

Essential points for discussion.

1) The relative difference in arrival times of the ‘direct’, ‘reflected’ & ‘refracted’ phases as offset increases.

2) The synchrony of the direct and reflected phases along the lower interface.

Essential points for discussion.

1) The relative difference in arrival times of the ‘direct’, ‘reflected’ & ‘refracted’ phases as offset increases.

2) The synchrony of the direct and reflected phases along the lower interface.

3) The refracted wavefront is tangential to the reflected wavefront at the critical angle.

Essential points for discussion.

1) The relative difference in arrival times of the ‘direct’, ‘reflected’ & ‘refracted’ phases as offset increases.

2) The synchrony of the direct and reflected phases along the lower interface.

3) The refracted wavefront is tangential to the reflected wavefront at the critical angle.

4) The difference in the ‘apparent’ velocity of the three phases along the surface

   a) The direct (primary) wave travels @ v_1.
   b) The reflected wave @ $v_1 / \sin\theta_i$
   (where θ_i is the incident angle).
   c) The refracted wave @ $v_2 = v_1 / \sin\theta_c$
   (where θ_c is the ’critical’ angle).

Essential points for discussion.

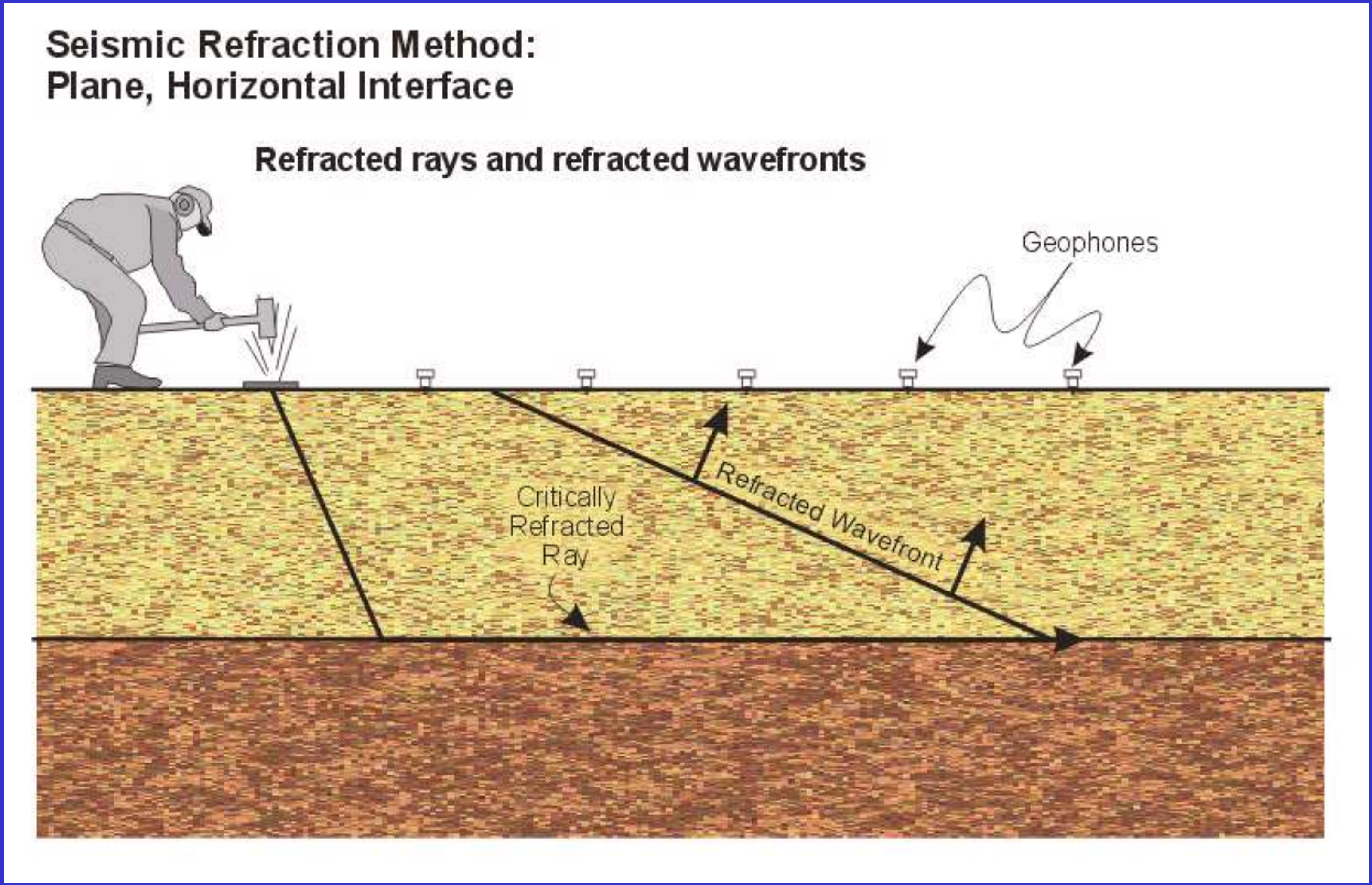
1) The relative difference in arrival times of the ‘direct’, ‘reflected’ & ‘refracted’ phases as offset increases.

2) The synchrony of the direct and reflected phases along the lower interface.

3) The refracted wavefront is tangential to the reflected wavefront at the critical angle.

4) The difference in the ‘apparent’ velocity of the three phases along the surface

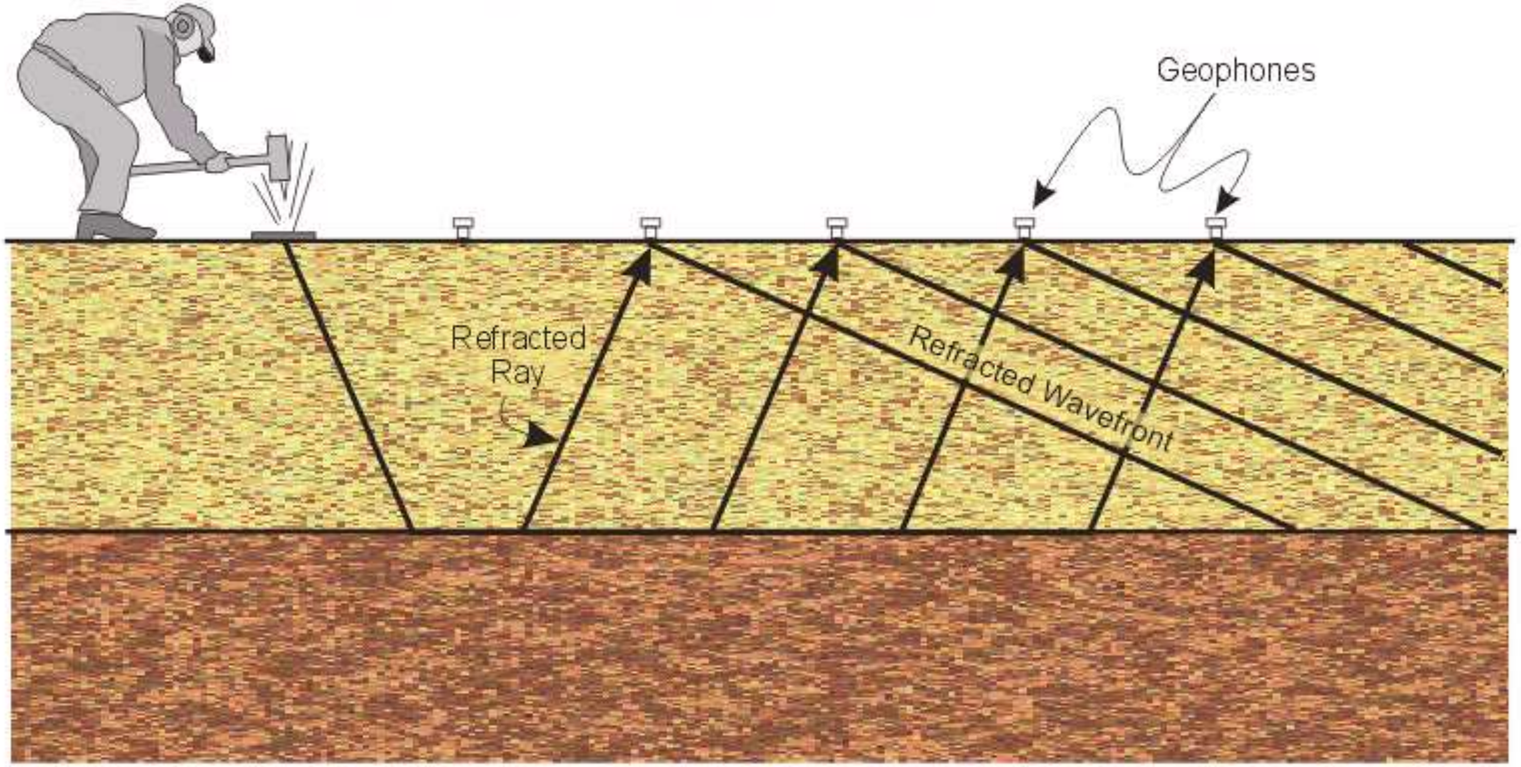
   a) The direct (primary) wave travels @ v_1.
   b) The reflected wave @ $v_1 / \sin \theta_i$
   (where θ_i is the incident angle).
   c) The refracted wave @ $v_2 = v_1 / \sin \theta_c$
   (where θ_c is the ’critical’ angle).



Refraction at the Critical Angle.

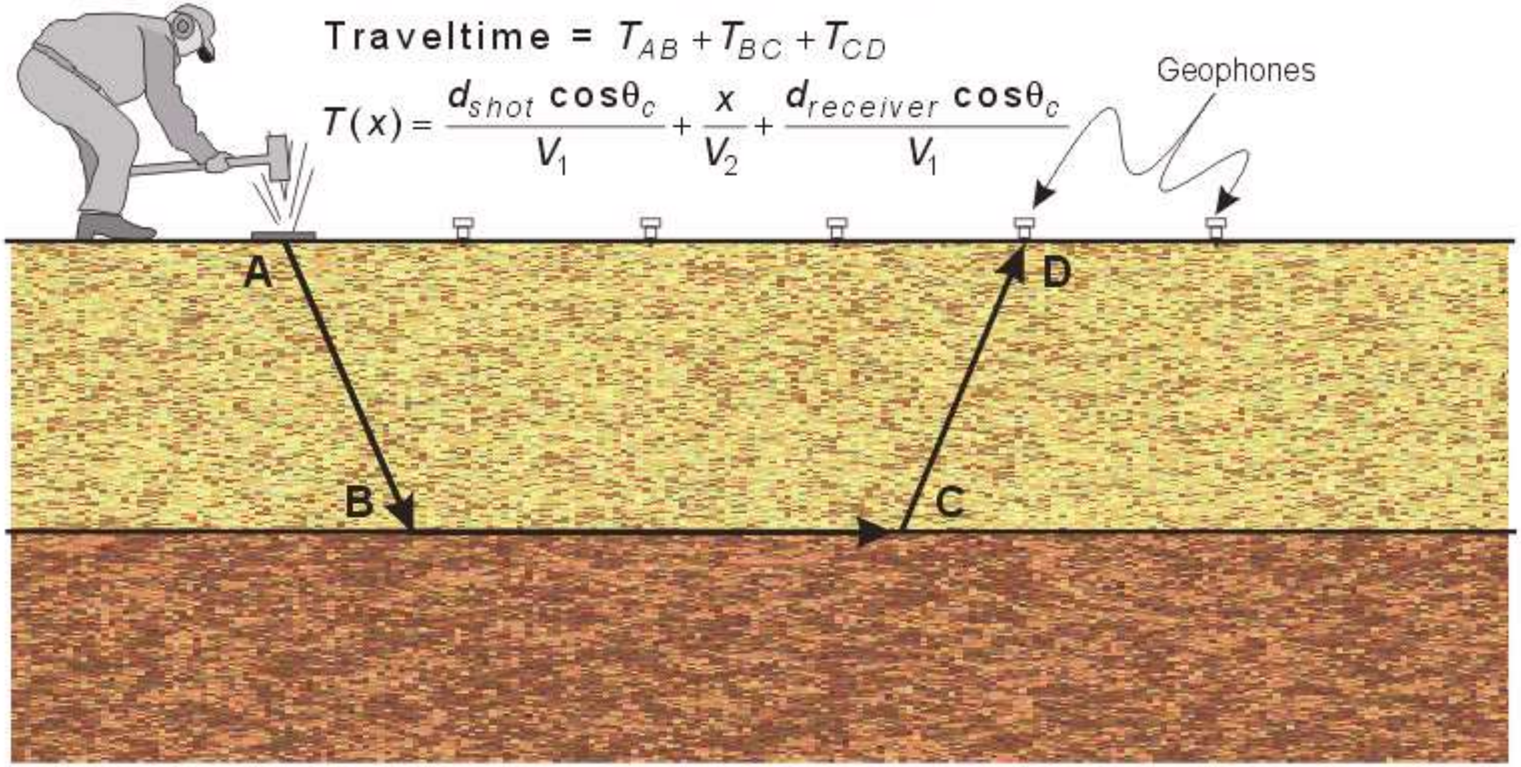
Seismic Refraction Method: Plane, Horizontal Interface

Refracted rays and refracted wavefronts



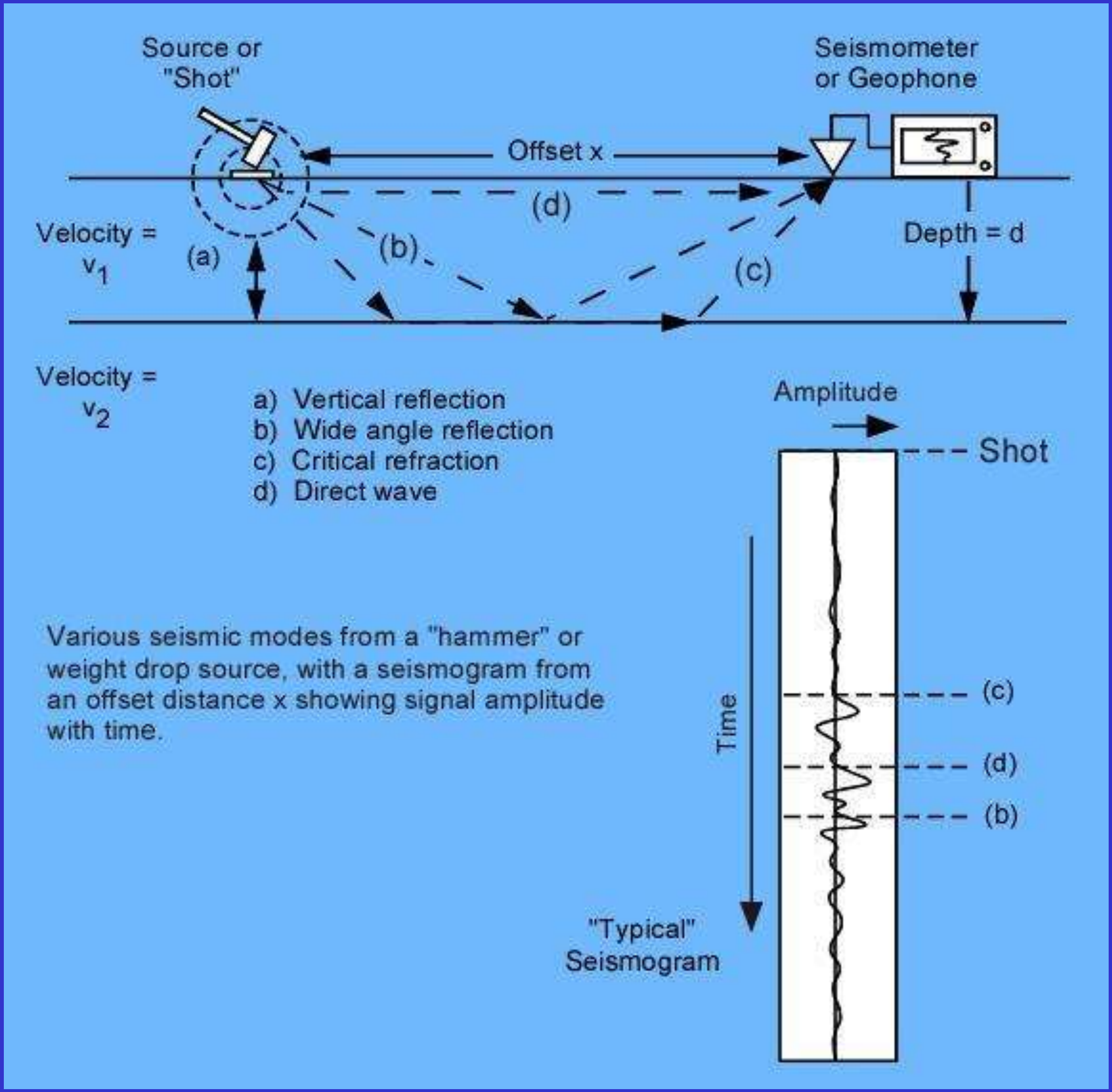
Seismic Refraction Method: Plane, Horizontal Interface

Refracted rays and refracted wavefronts



$$\text{Traveltime} = T_{AB} + T_{BC} + T_{CD}$$

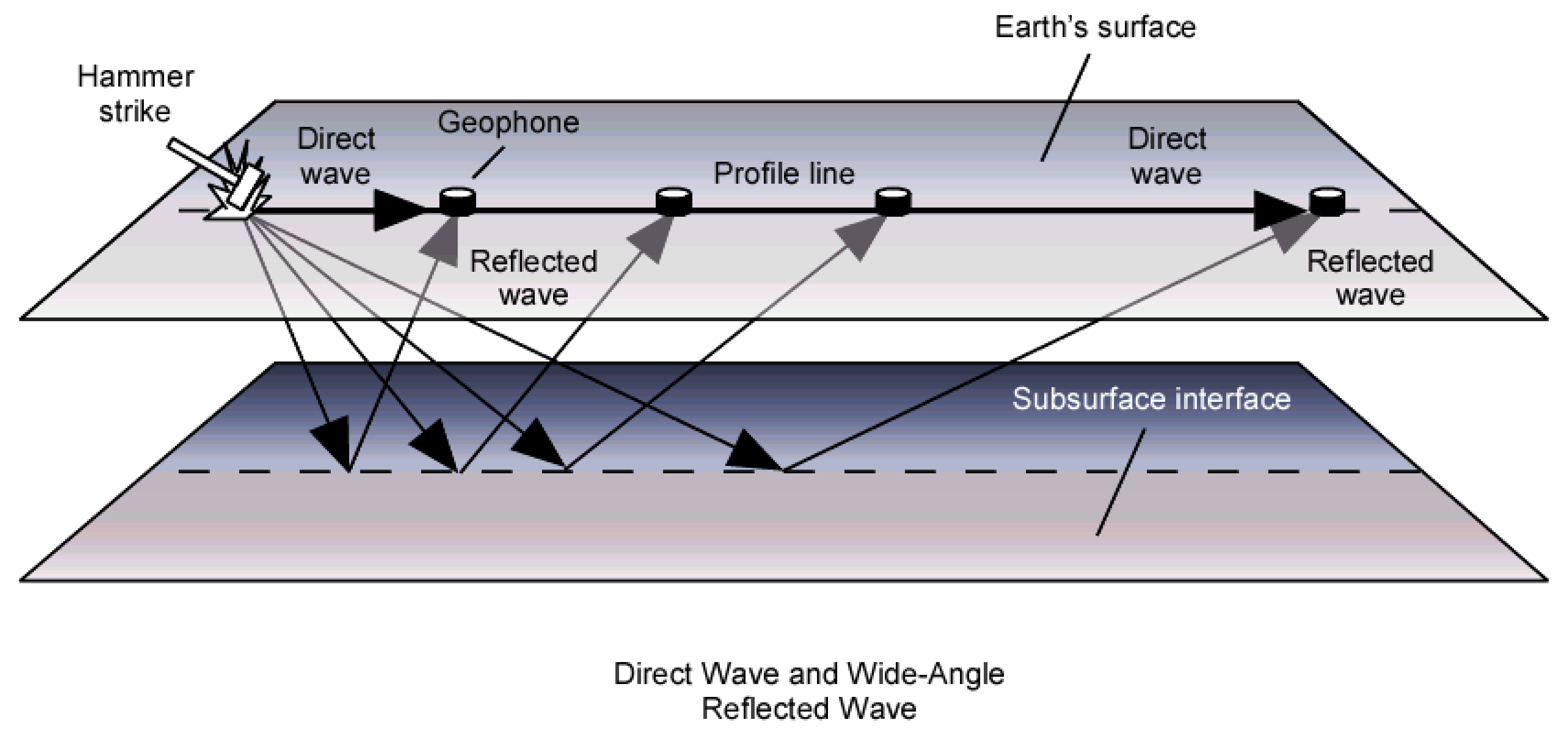
$$T(x) = \frac{d_{shot}\cos\theta_c}{V_1} + \frac{x}{V_2} + \frac{d_{receiver}\cos\theta_c}{V_1}$$



Various seismic modes from a "hammer" or weight drop source, with a seismogram from an offset distance x showing signal amplitude with time.

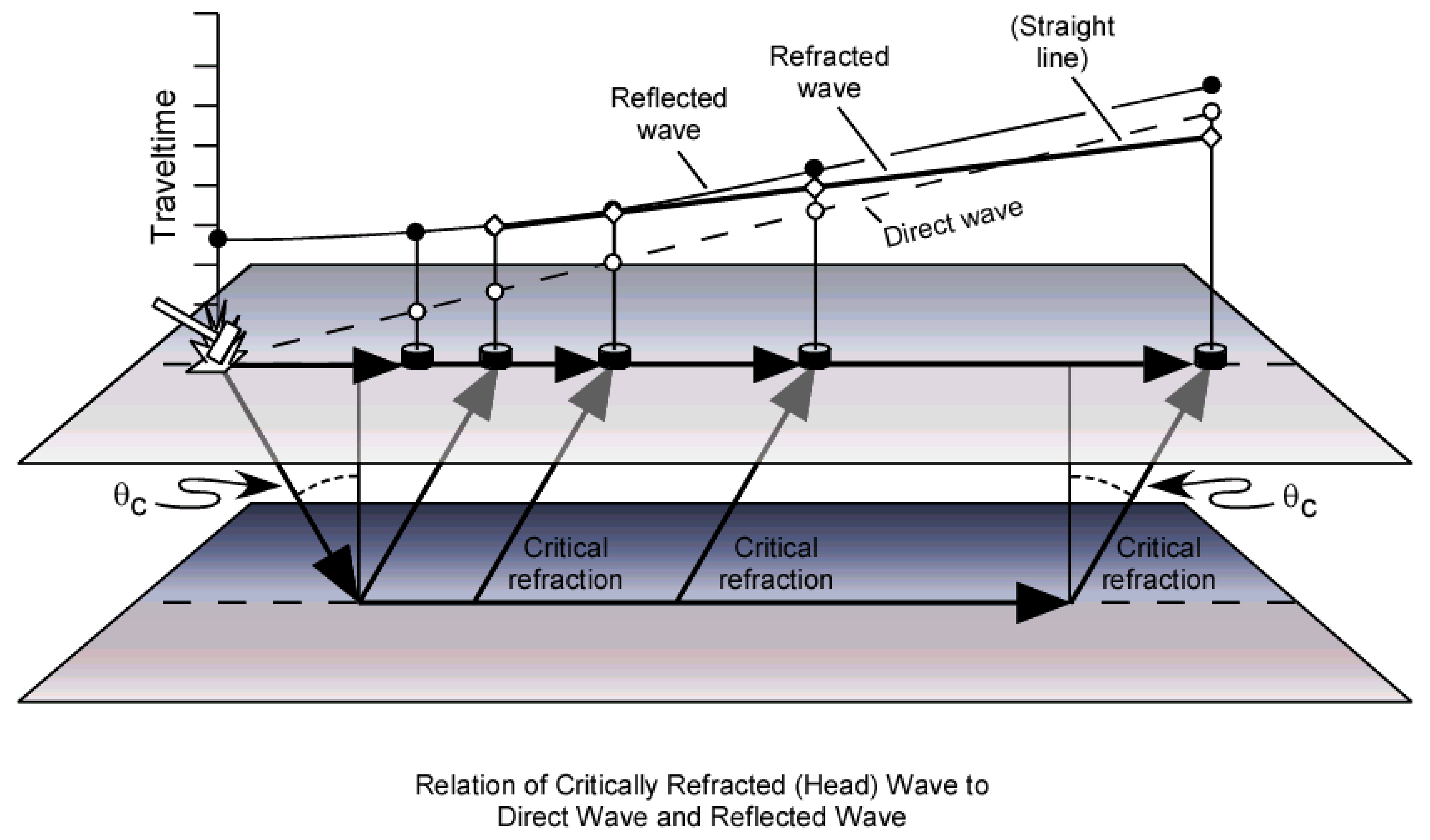
In summary: A seismic (or radar) signal generates a number of modes.

How can we use the ‘refracted’ phase to determine structure in the earth ?



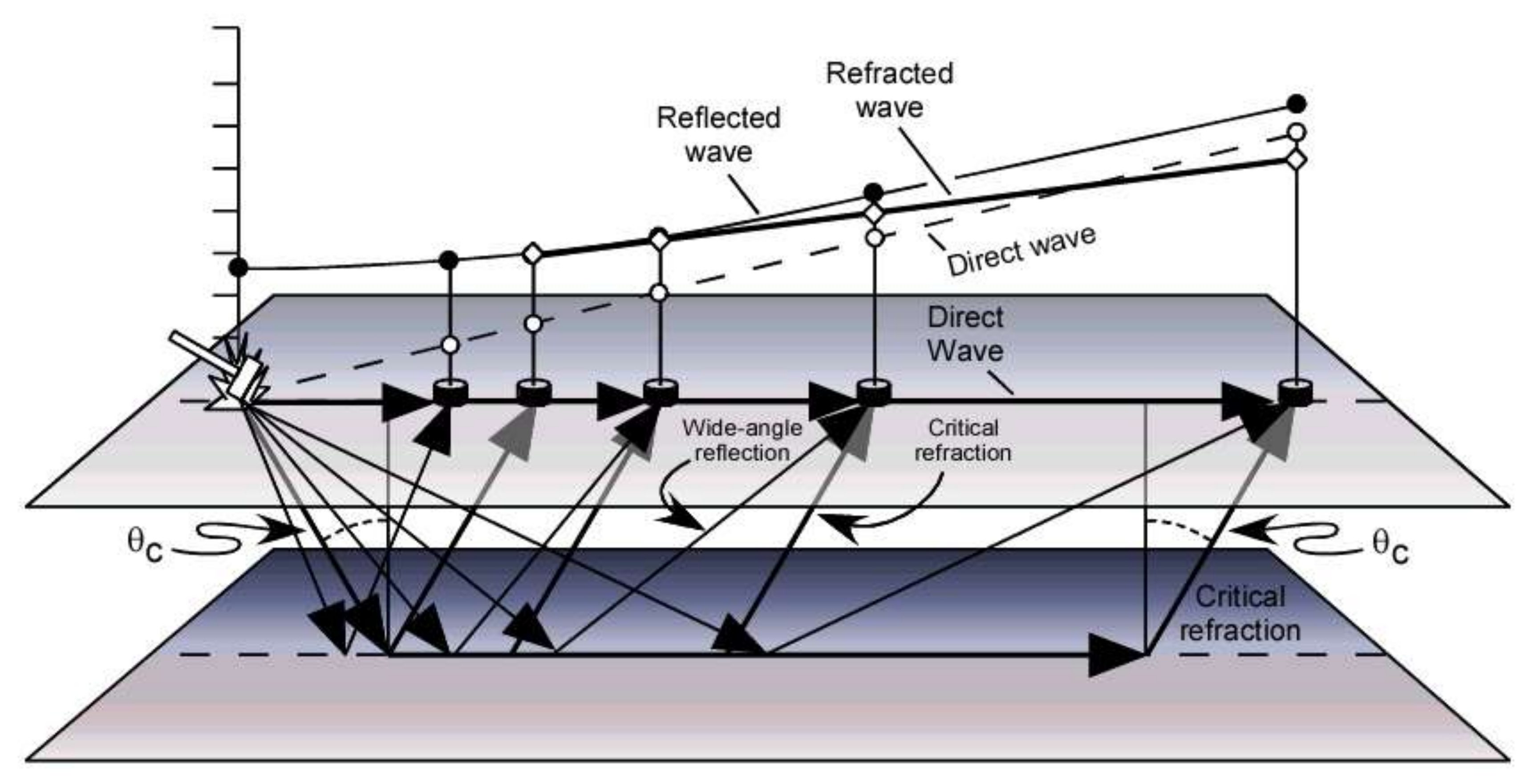
Direct Wave and Wide-Angle Reflected Wave

Recall the *direct* and <u>reflected</u> ray paths.



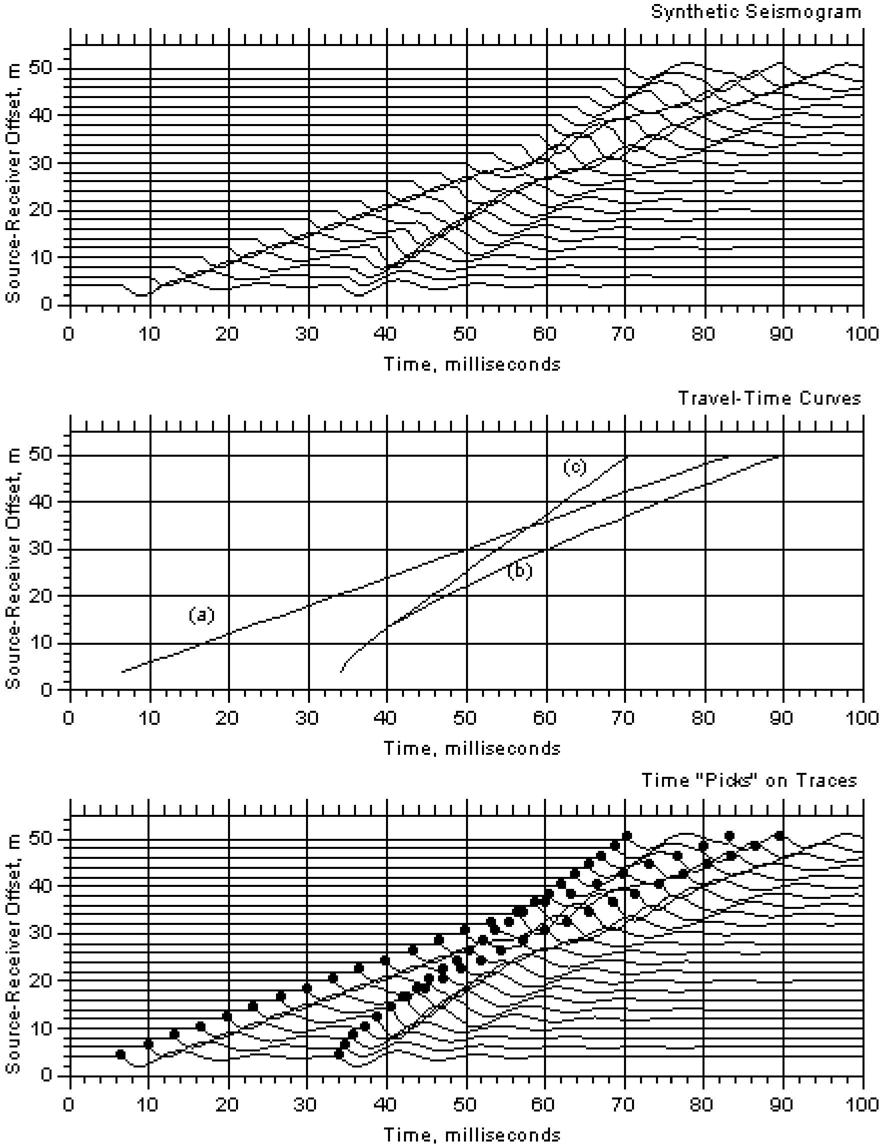
Relation of Critically Refracted (Head) Wave to
Direct Wave and Reflected Wave

The direct and refracted ray paths.



Composite of All Principal Phases: Direct Wave, Wide-Angle (Large Offset) Reflected Wave, Critically Refracted (Head) Wave

Composite of direct, reflected and refracted ray paths.



Synthetic Seismogram

Travel-Time Curves

Time "Picks" on Traces

Example of a synthetic seismogram showing (a) the direct wave, (b) the reflected wave, and (c) the critically refracted wave. Parameters of the model are that V_1 = 600 m/s; V_2 = 1200 m/s; d = 10 m.

Set of synthetic seismograms.

Details on “Picking” and Interpreting Seismic Data

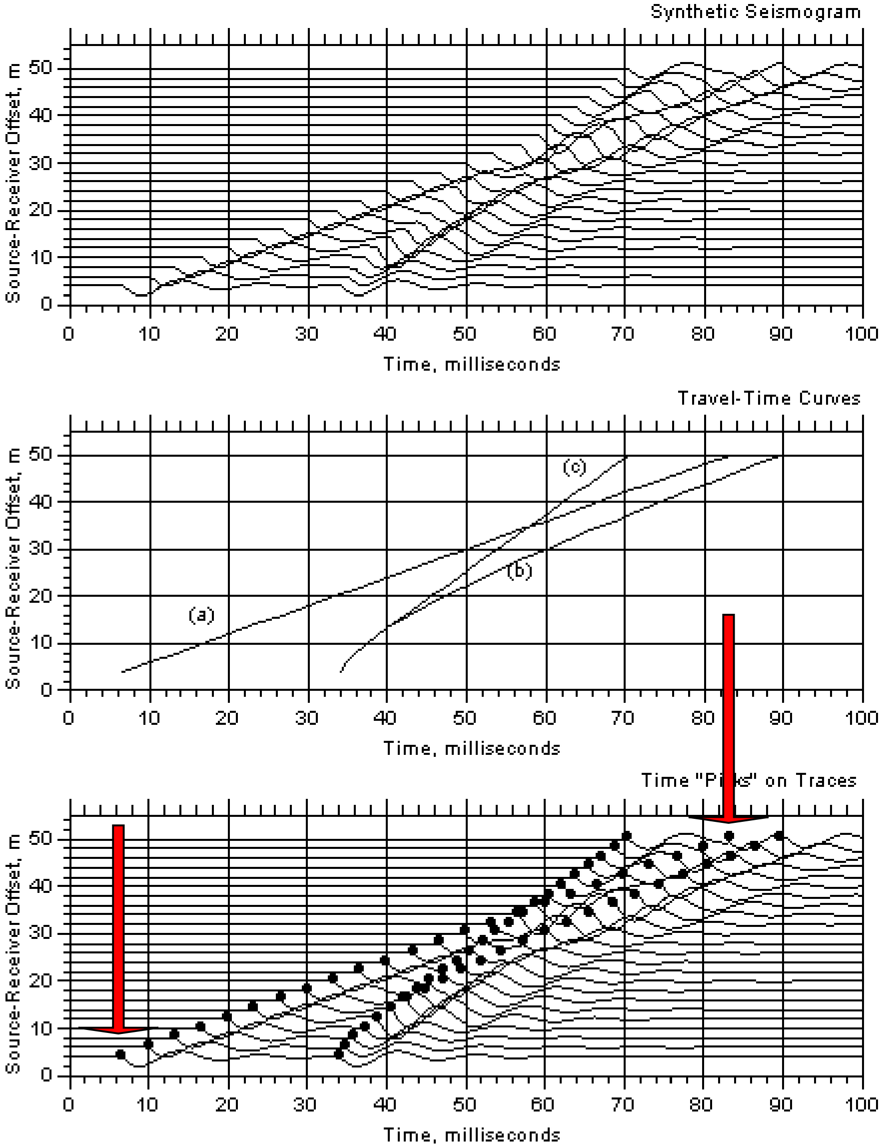
Mathematical Underpinnings: Traveltimes of Principal Phases

1) Direct Wave

$$T(x) = x / V_1$$

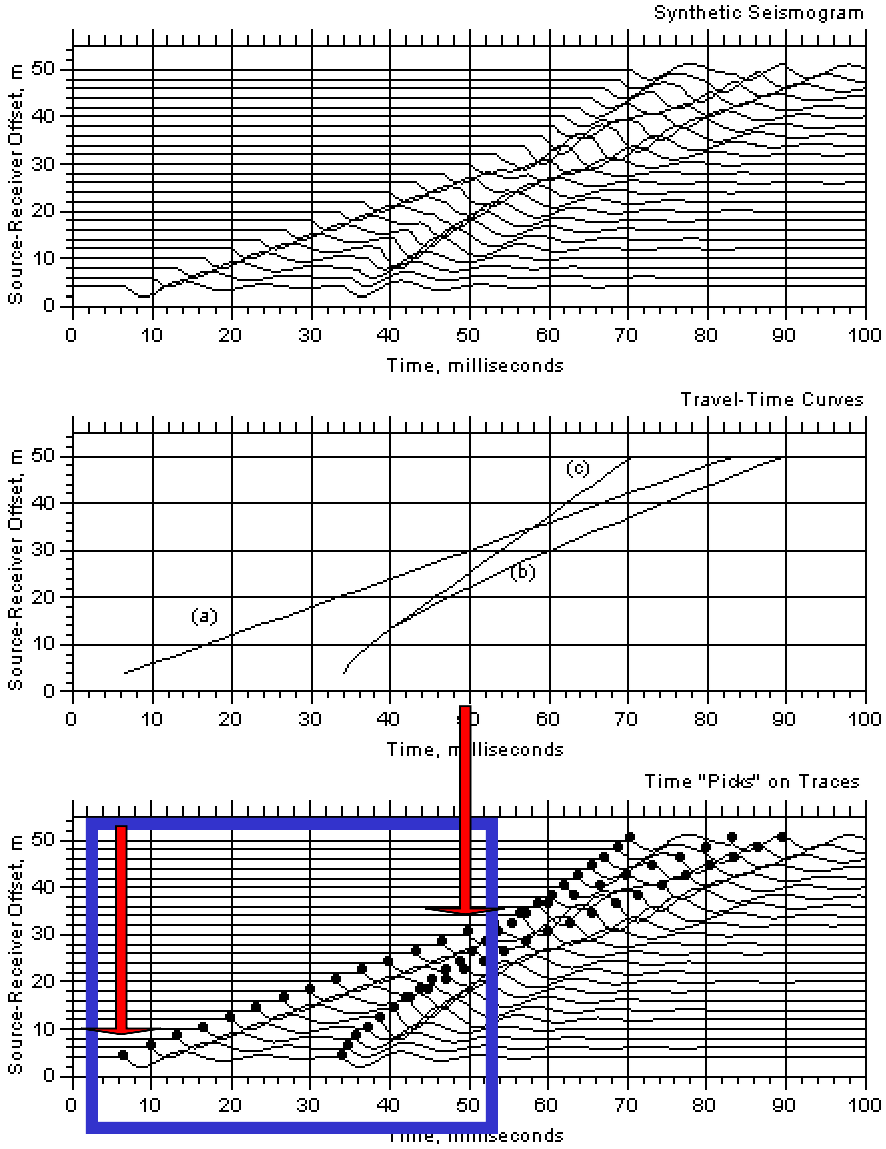
2) Reflected Wave

$$T(x) = \sqrt{(x / V_1)^2 + (2d / V_1)^2}$$



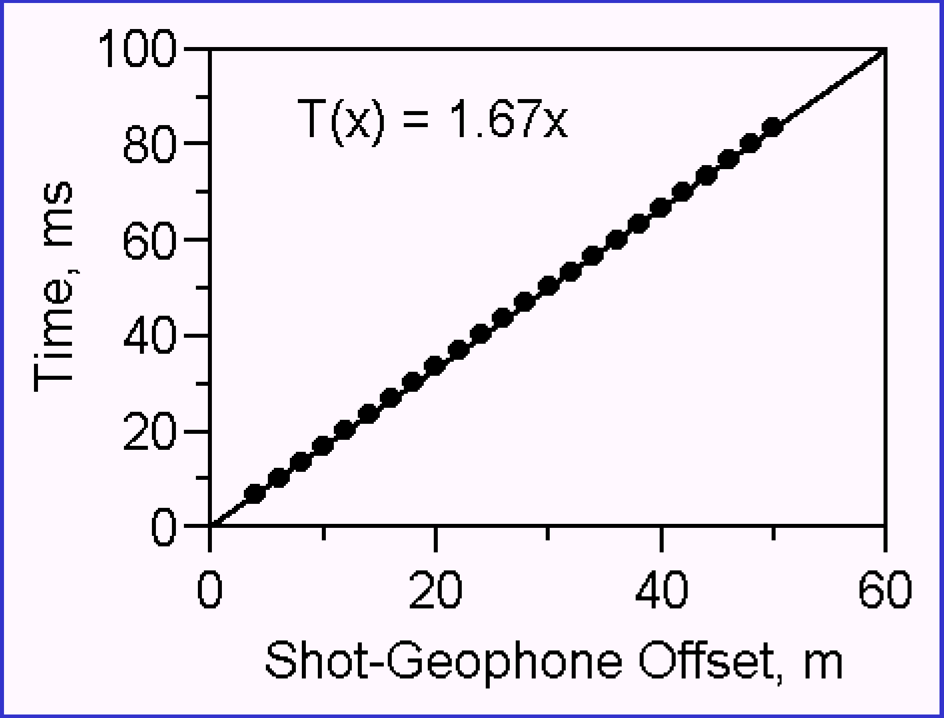
Example of a synthetic seismogram showing (a) the direct wave, (b) the reflected wave, and (c) the critically refracted wave. Parameters of the model are that V_1 = 600 m/s; V_2 = 1200 m/s; d = 10 m.

Set of synthetic seismograms. (Direct phase.)

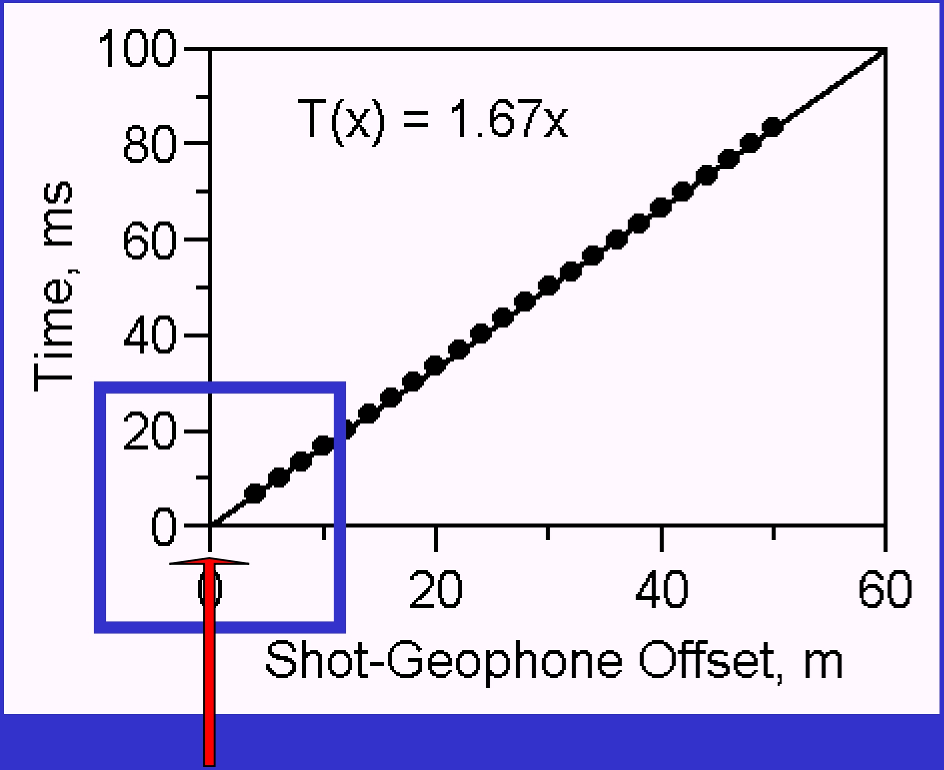


Example of a synthetic seismogram showing (a) the direct wave, (b) the reflected wave, and (c) the critically refracted wave. Parameters of the model are that V_1 = 600 m/s; V_2 = 1200 m/s; d = 10 m.

(Direct wave “picks” are best in here.)



Analysis of direct ground wave.



Direct ground wave traveltime should (?) go through origin.

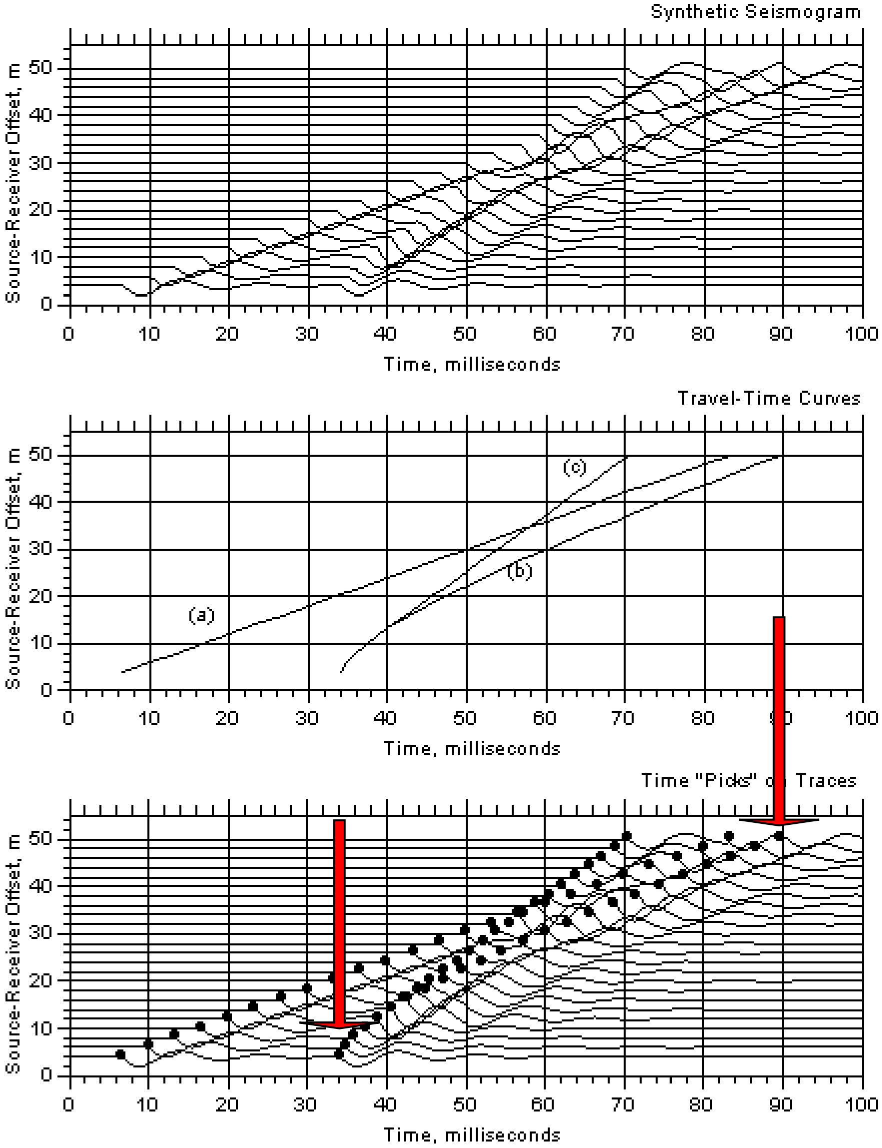
Mathematical Underpinnings: Traveltimes of Principal Phases

1) Direct Wave

$$T(x) = x / V_1$$

2) Reflected Wave

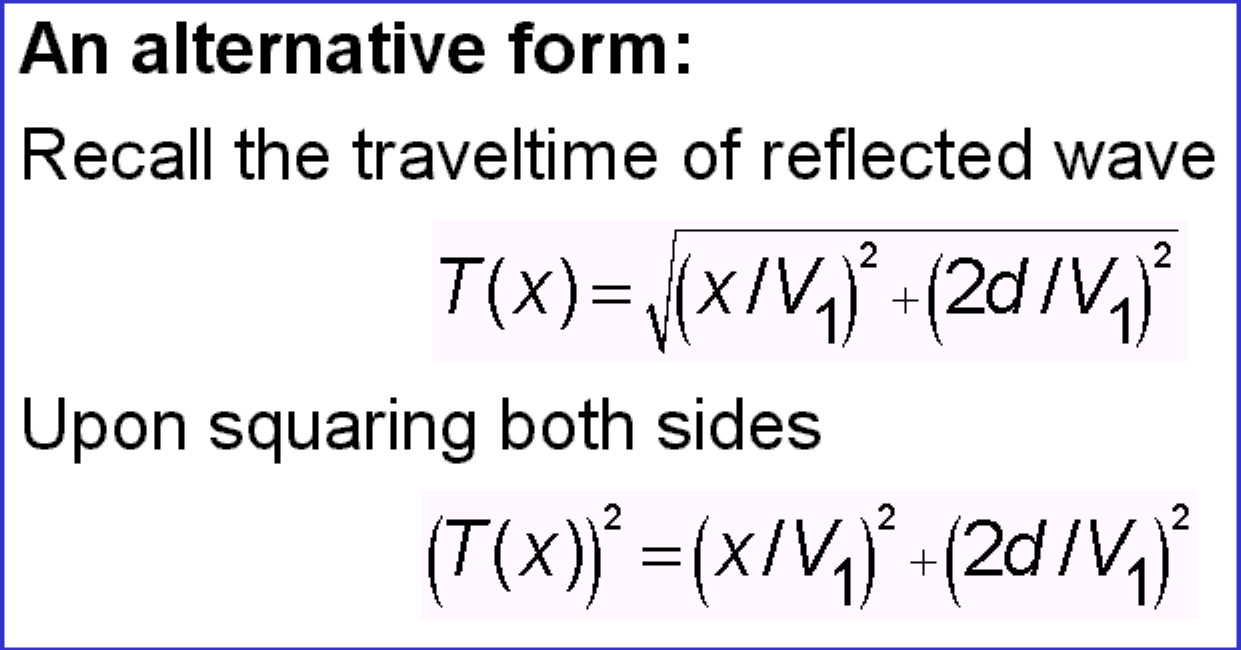
$$T(x) = \sqrt{(x / V_1)^2 + (2d / V_1)^2}$$



Example of a synthetic seismogram showing (a) the direct wave, (b) the reflected wave, and (c) the critically refracted wave. Parameters of the model are that V_1 = 600 m/s; V_2 = 1200 m/s; d = 10 m.

Set of synthetic seismograms. (Reflected phase.)

Analyzing reflected phases: *An alternative expression for the traveltime*.



An alternative form:

Recall the traveltime of reflected wave

$$T(x)=\sqrt{(x/V_1)^2+(2d/V_1)^2}$$

Upon squaring both sides

$$(T(x))^2=(x/V_1)^2+(2d/V_1)^2$$

Analyzing reflected phases: *An alternative expression for the traveltime*.

Plotting T^2 versus X^2 results in a straight line

$$T^2 = \frac{X^2}{V^2} + \left(\frac{2d}{V}\right)^2$$

$$(\eta = m\xi + b)$$

where

$$m = 1/V^2; \quad b = (2d/V)^2$$

Analyzing reflected phases: *An alternative expression for the traveltime*.

Plotting T^2 versus X^2 results in a straight line

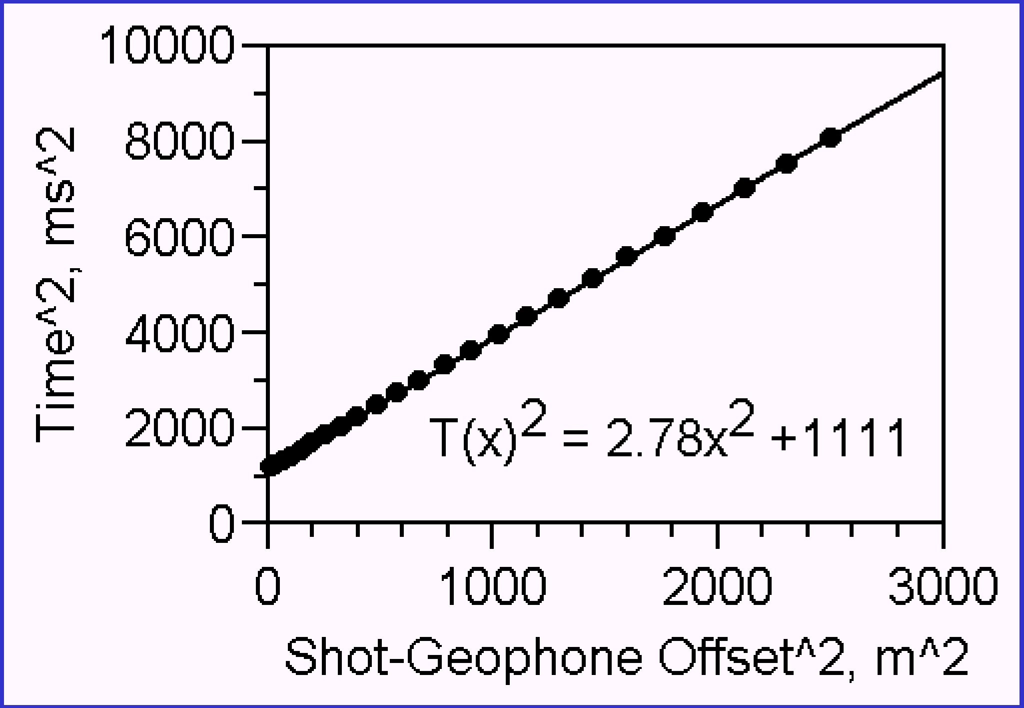
$$T^2 = \frac{X^2}{V^2} + \left(\frac{2d}{V}\right)^2$$

$$(\eta = m\xi + b)$$

where

$$\boxed{m = 1/V^2; \quad b = (2d/V)^2}$$

The slope m, and the intercept b, provide the essential parameters for interpretation.

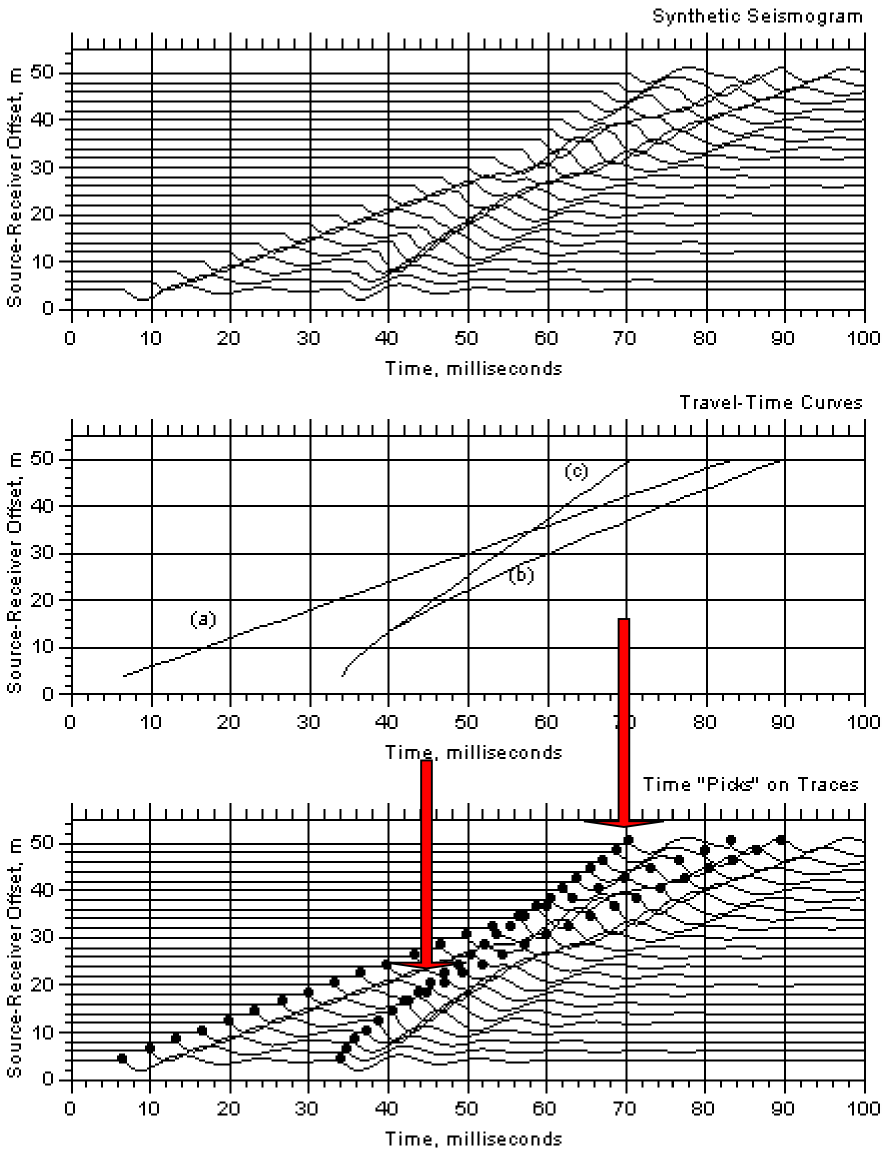


Analysis of reflected phase using T^2 - X^2 method.

3) Critically Refracted or Head Wave

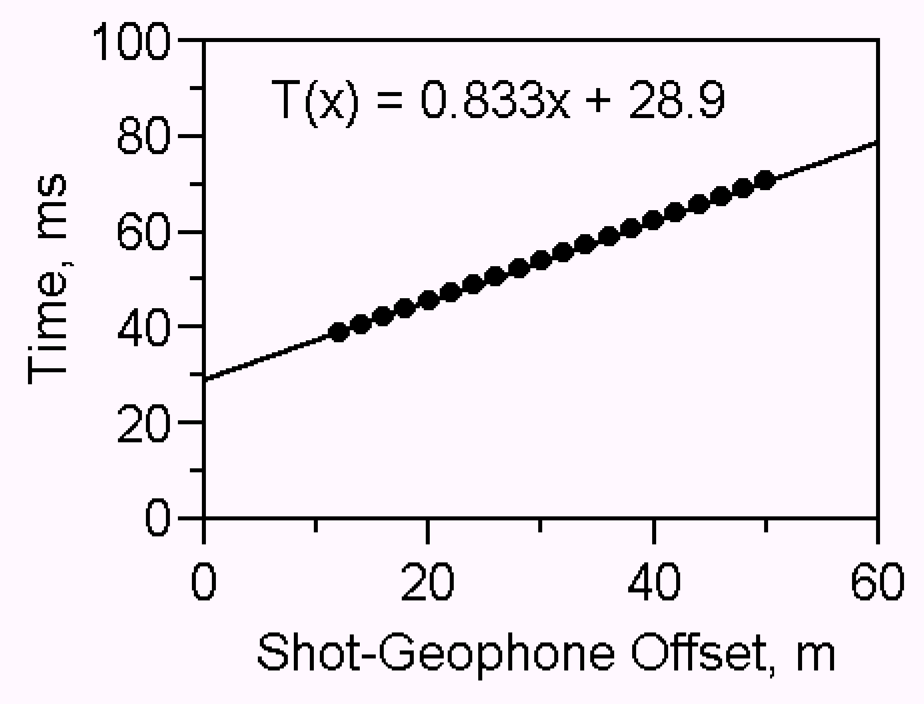
3a) Plane horizontal interface

$$T(x) = \frac{x}{V_2} + \frac{2d\cos(\theta_c)}{V_1}$$

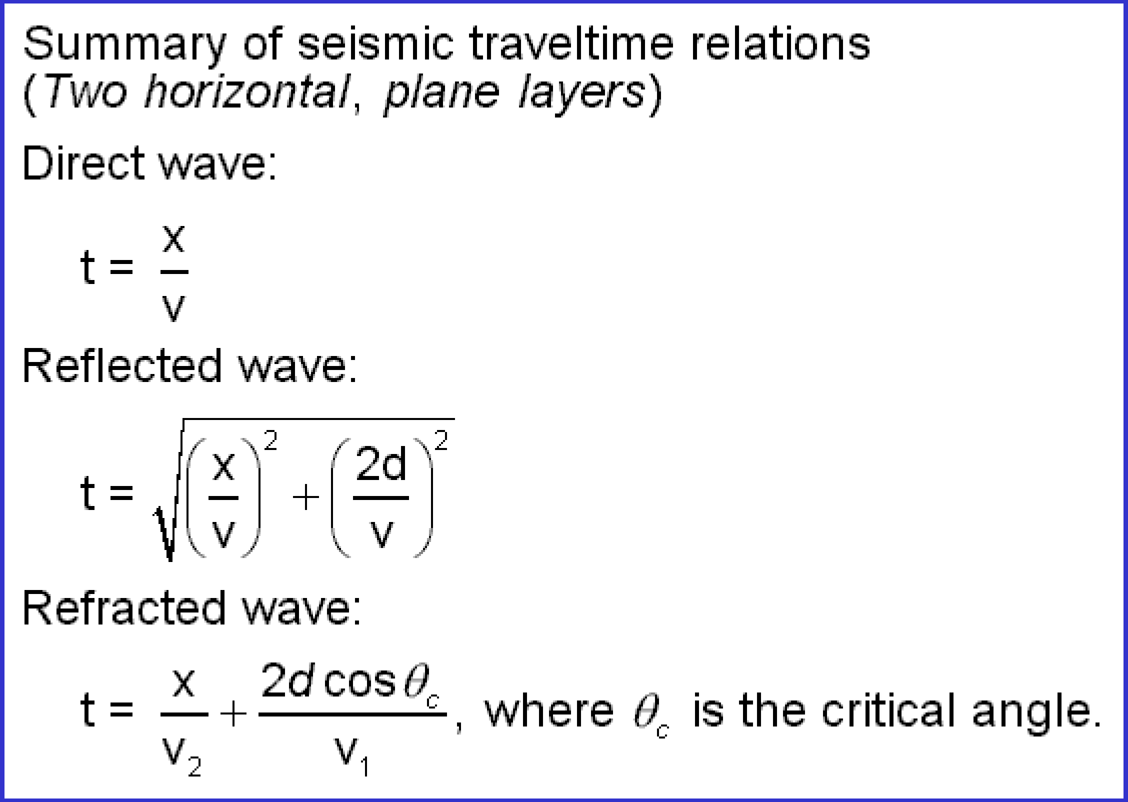


Example of a synthetic seismogram showing (a) the direct wave, (b) the reflected wave, and (c) the critically refracted wave. Parameters of the model are that V_1 = 600 m/s; V_2 = 1200 m/s; d = 10 m.

Set of synthetic seismograms. (Refracted phase.)



Analysis of critically refracted phase.



Summary of seismic traveltime relations
(*Two horizontal*, *plane layers*)

Direct wave:

$$t = \frac{x}{v}$$

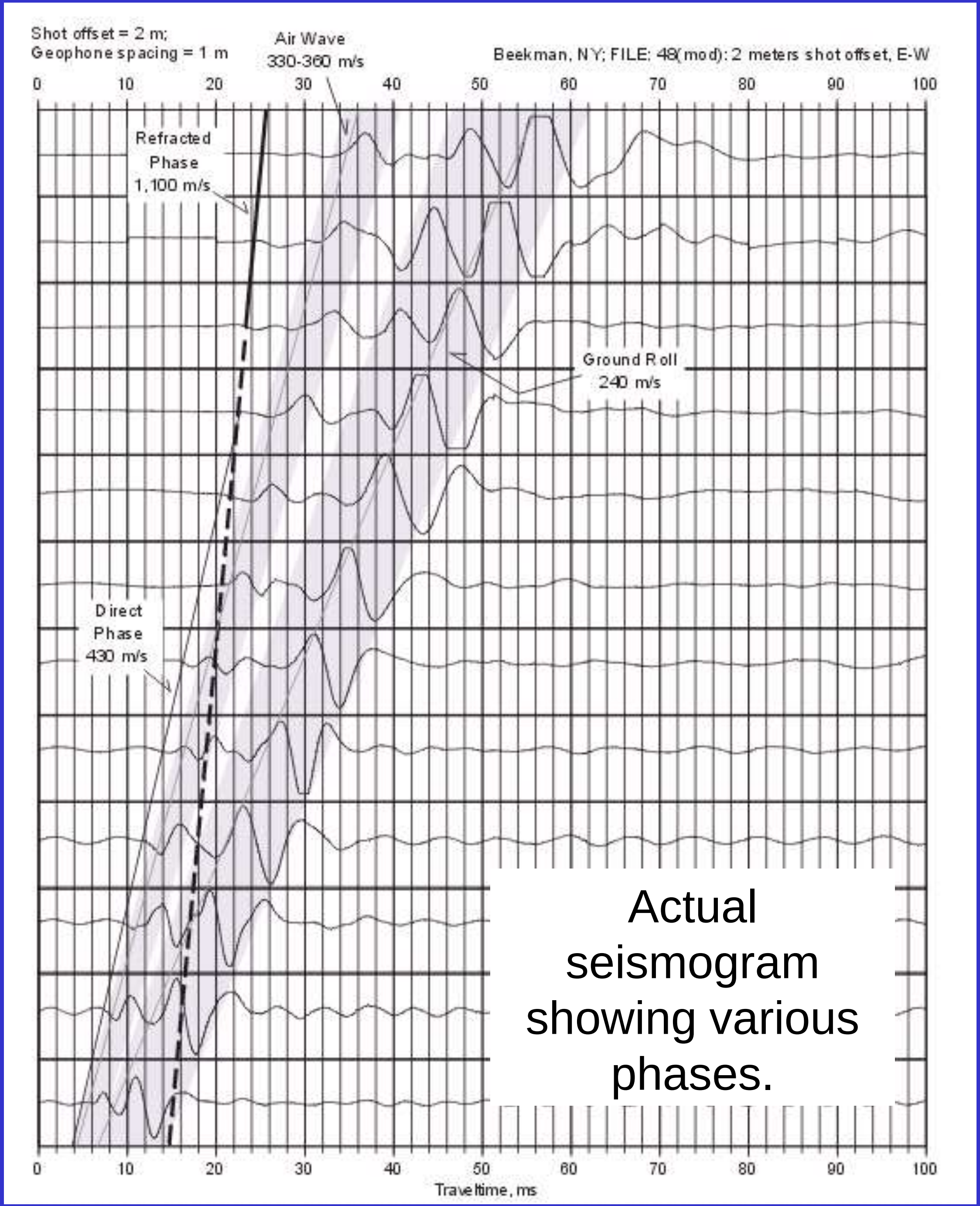
Reflected wave:

$$t = \sqrt{\left(\frac{x}{v}\right)^2 + \left(\frac{2d}{v}\right)^2}$$

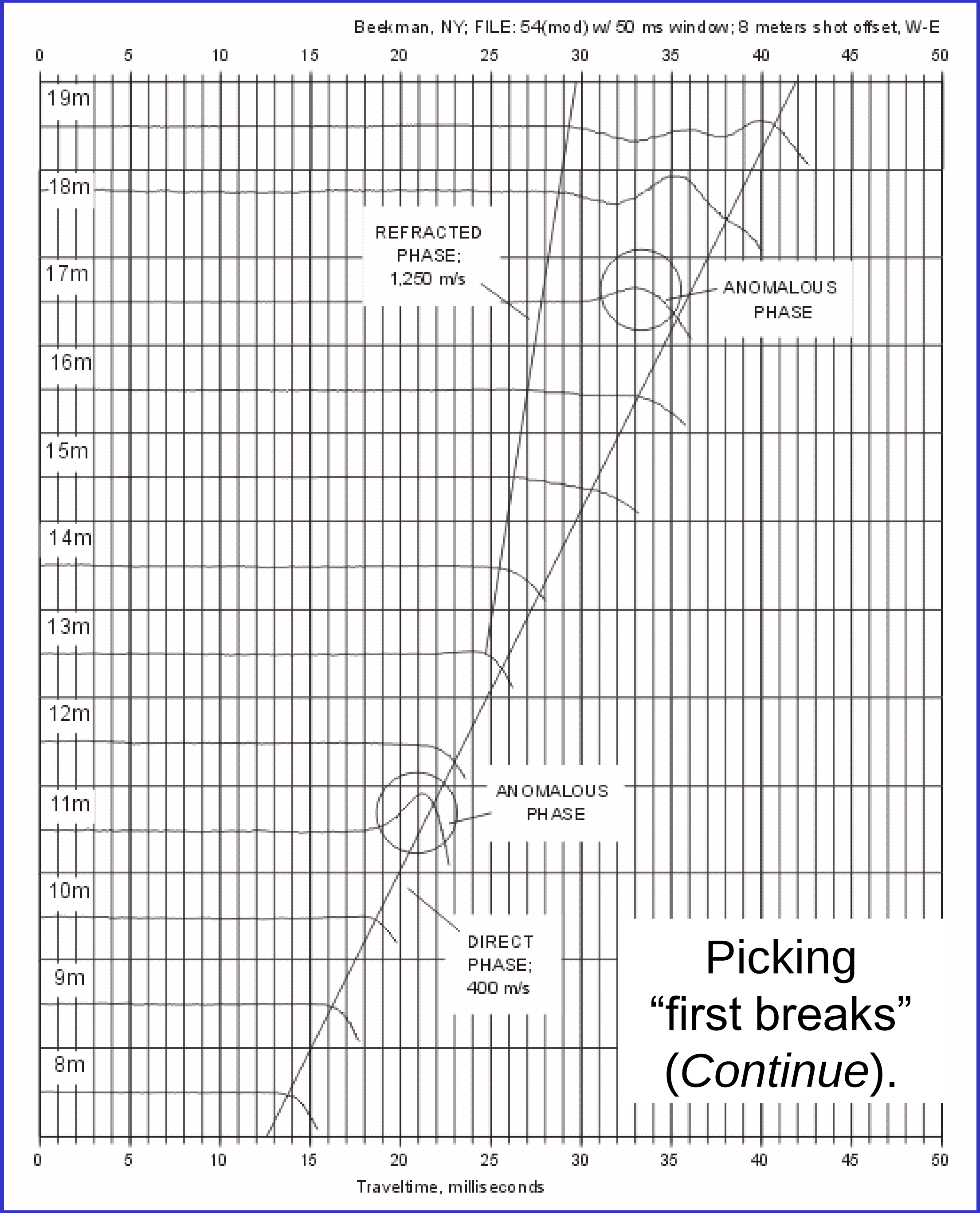
Refracted wave:

$$t = \frac{x}{v_2} + \frac{2d\cos\theta_c}{v_1},$$ where θ_c is the critical angle.

Traveltime relations.

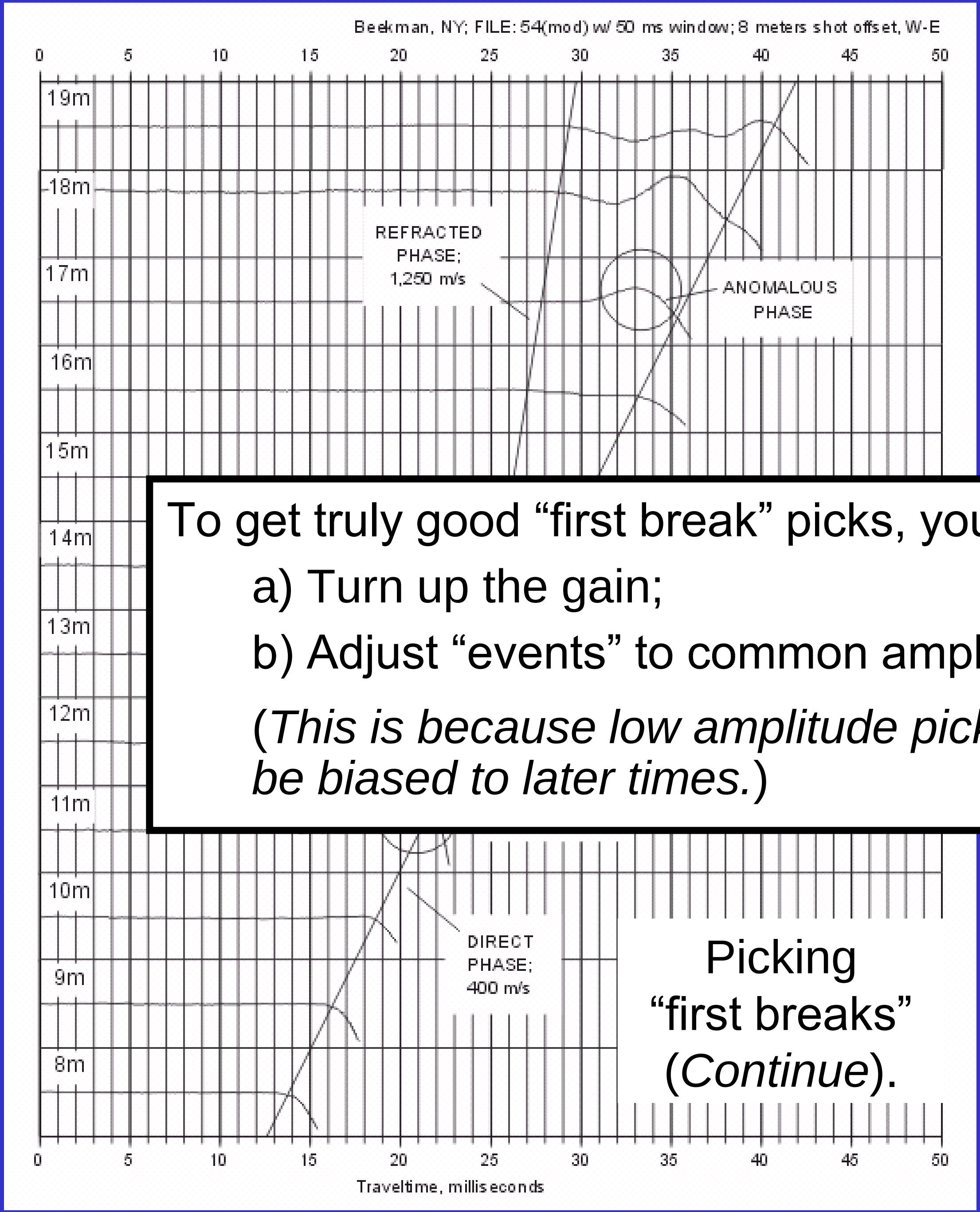


Actual seismogram showing various phases.



Picking
“first breaks”
(*Continue*).

Picking “first breaks” (*Continue*).



To get truly good “first break” picks, you need to

a) Turn up the gain;

b) Adjust “events” to common amplitude.

(*This is because low amplitude picks tend to be biased to later times.*)

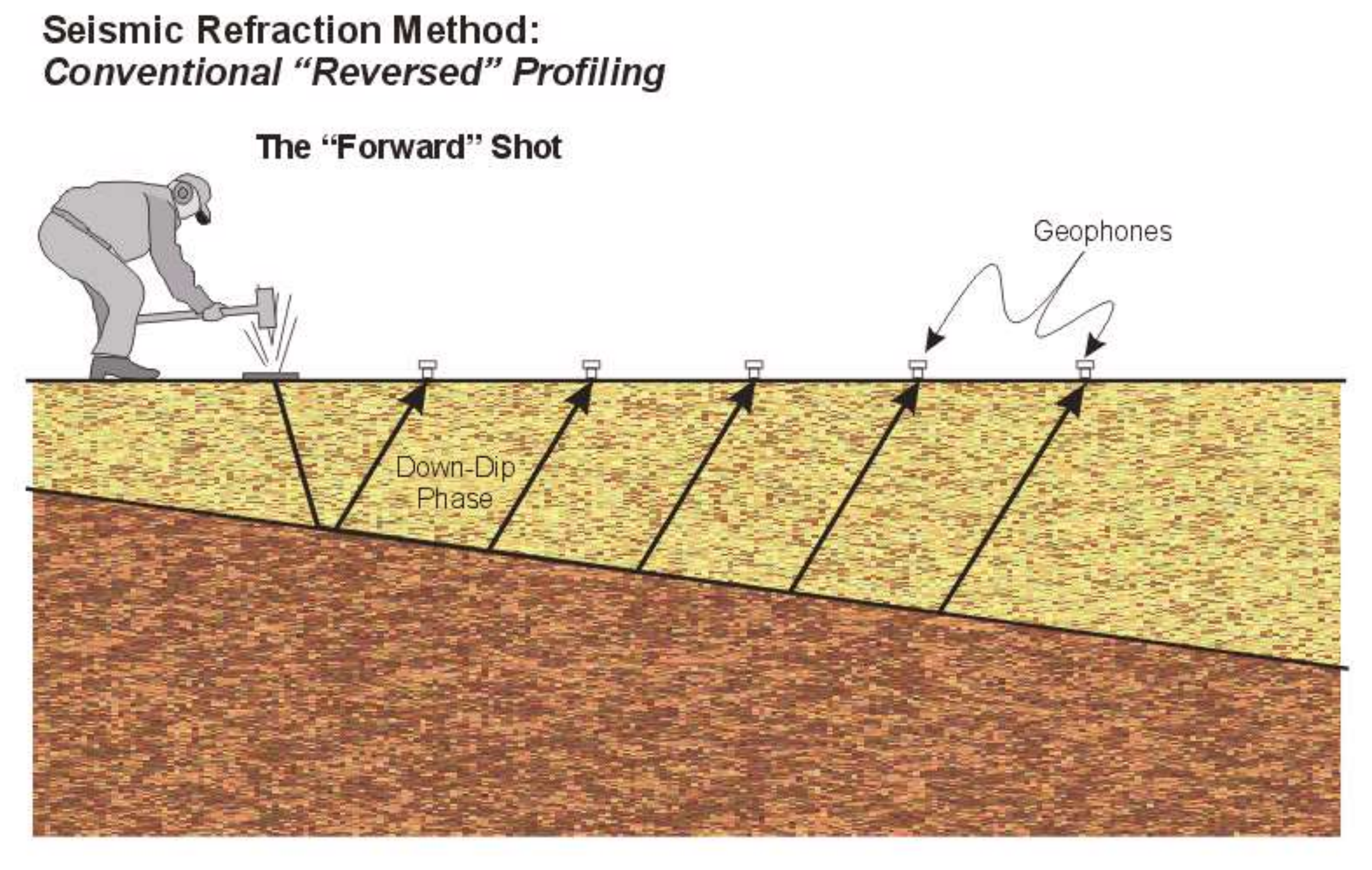
Using the refraction method for more complicated field situations.

Consider “dipping” interfaces.

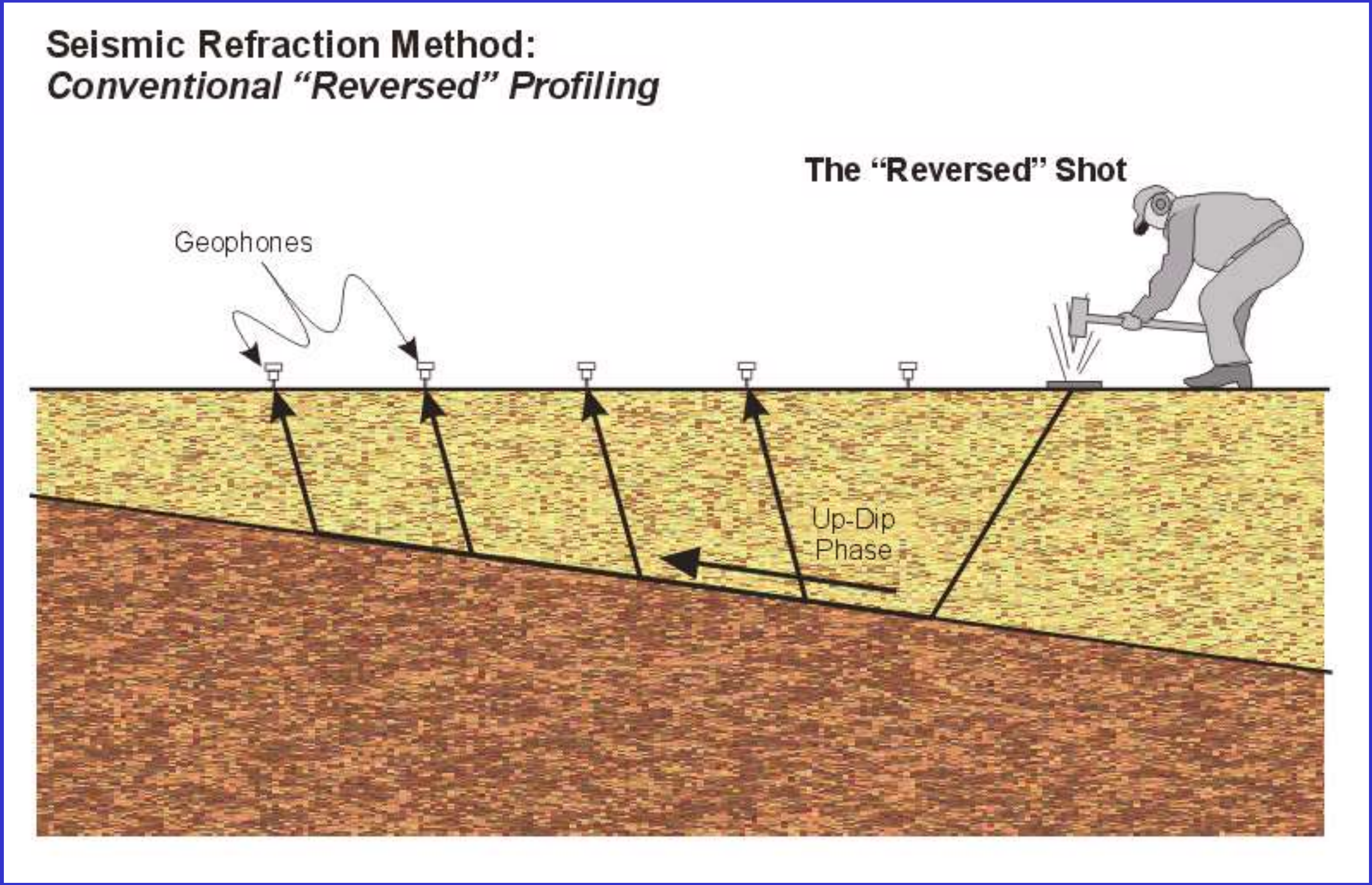
Consider “dipping” interfaces.

We employ “reversed” refraction profiling.

Procedure: Step 1; The “Forward” Shot.

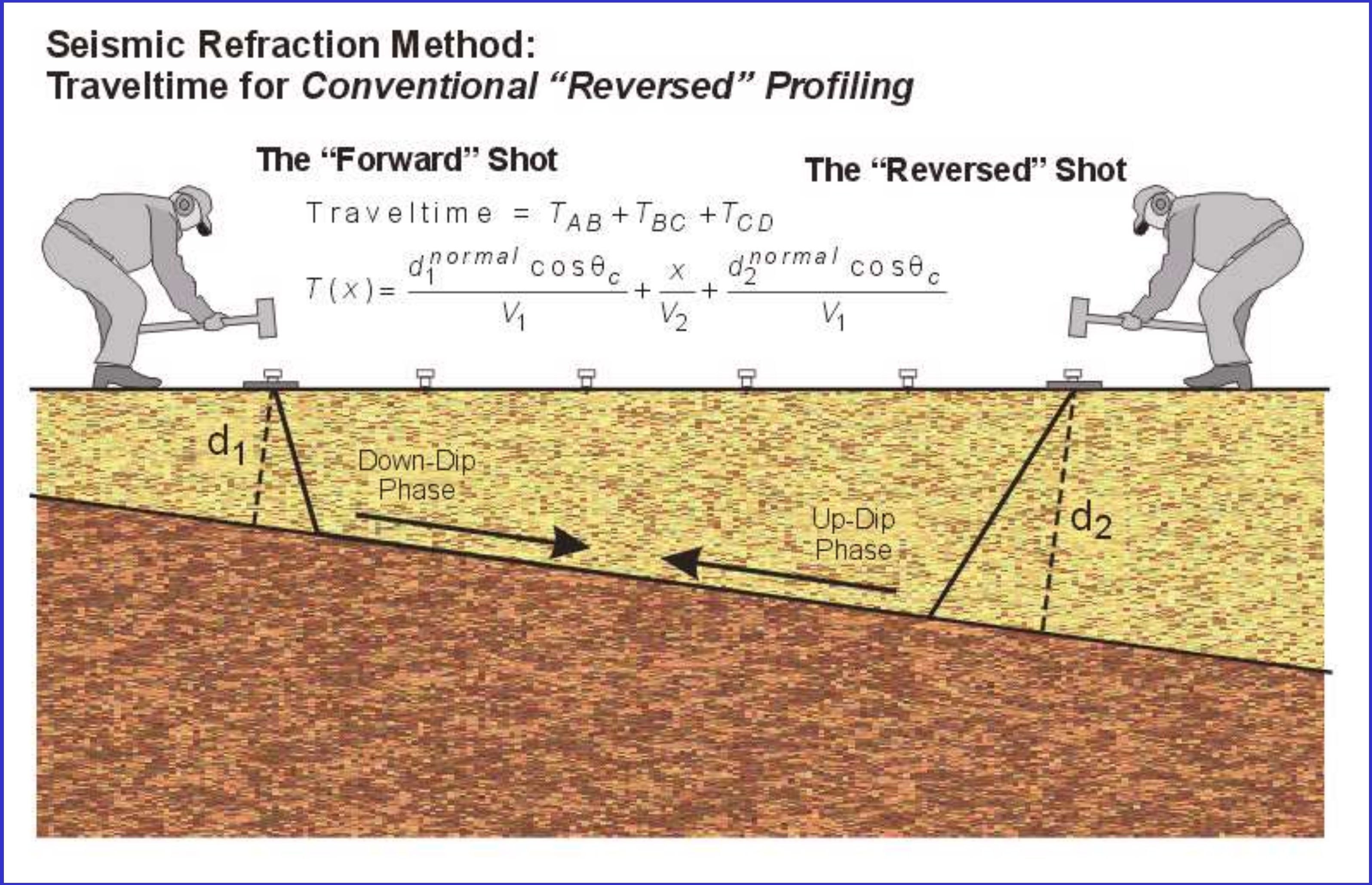


Procedure: Step 2; The "Reverse" Shot.



We use the theoretical traveltime of the respective refracted phases.

The Theoretical Traveltime for a Refracted Phase on a Dipping Interface.



Seismic Refraction Method:
Traveltime for *Conventional "Reversed" Profiling*

The "Forward" Shot

The "Reversed" Shot

$$\text{Traveltime} = T_{AB} + T_{BC} + T_{CD}$$

$$T(x) = \frac{d_1^{normal}\cos\theta_c}{V_1} + \frac{x}{V_2} + \frac{d_2^{normal}\cos\theta_c}{V_1}$$

3b) Dipping interface

3b.1) General form

$$T(x) = \frac{x}{V_2} + \frac{d_A^{normal} \cos(\theta_c)}{V_1} + \frac{d_B^{normal} \cos(\theta_c)}{V_1}$$

3b.2) Traditional form for reversed profiles

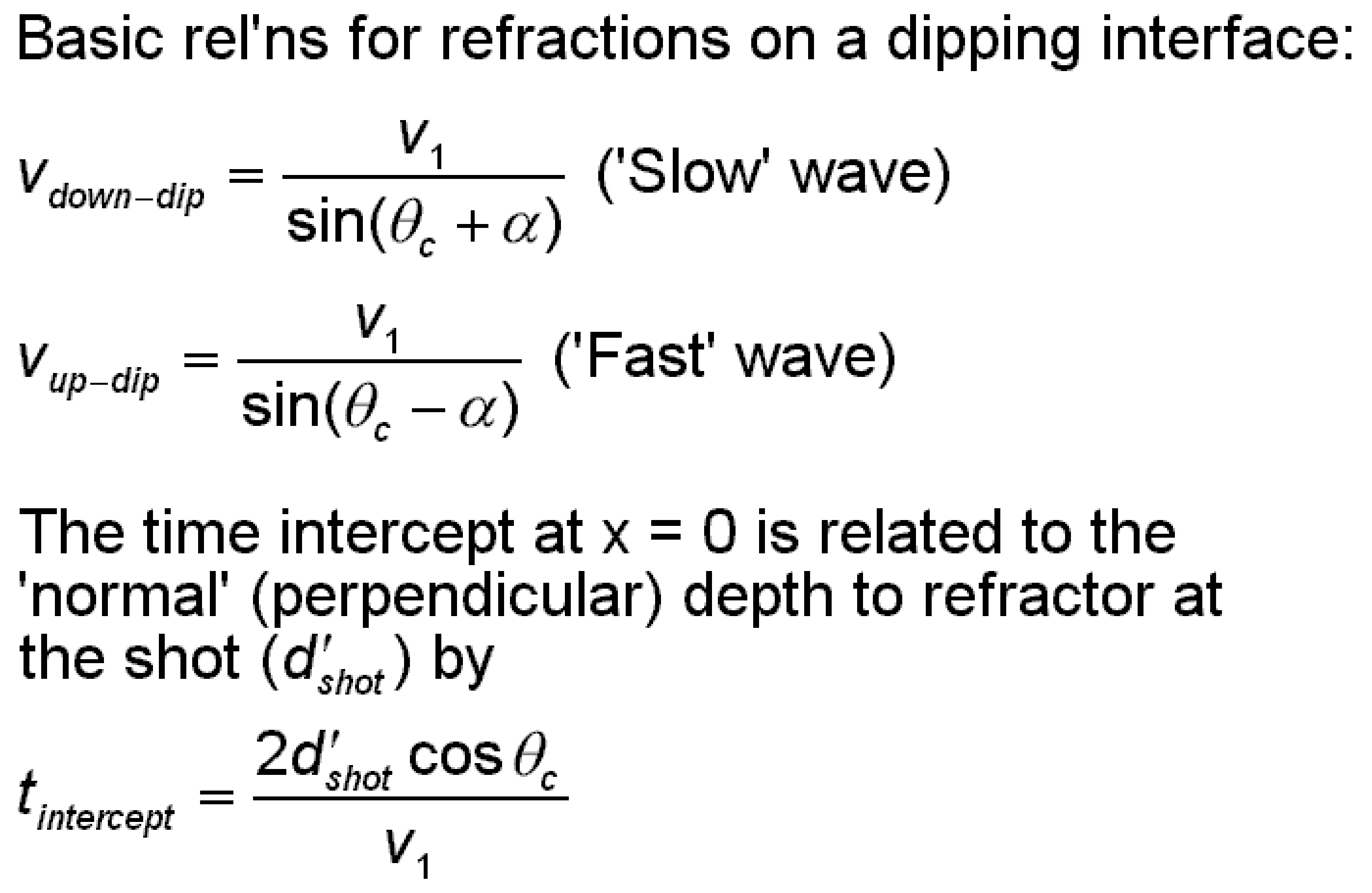
$$T_{Forward}(x) = \frac{x_{AB}\sin(\theta_c + \alpha)}{V_1} + \frac{2d_A^{normal}\cos(\theta_c)}{V_1}$$

$$T_{\mathrm{Re}\,versed}(x) = \frac{x_{BA}\sin(\theta_c - \alpha)}{V_1} + \frac{2d_B^{normal}\cos(\theta_c)}{V_1}$$

where

$$\left. Slope \right|_{\text{Forward}}^{\text{T-Refraction}} = \frac{1}{V_{\text{Forward}}^{\text{Refractor}}} = \frac{\sin\left(\theta_c + \alpha\right)}{V_1}$$

$$\left. Slope \right|_{\text{Reversed}}^{\text{T-Refraction}} = \frac{1}{V_{\text{Reversed}}^{\text{Refractor}}} = \frac{\sin\left(\theta_c - \alpha\right)}{V_1}$$



Basic rel'ns for refractions on a dipping interface:

$$V_{down-dip} = \frac{V_1}{\sin(\theta_c + \alpha)}$$ ('Slow' wave)

$$V_{up-dip} = \frac{V_1}{\sin(\theta_c - \alpha)}$$ ('Fast' wave)

The time intercept at x = 0 is related to the 'normal' (perpendicular) depth to refractor at the shot (d'_{shot}) by

$$t_{intercept} = \frac{2d'_{shot}\cos\theta_c}{V_1}$$

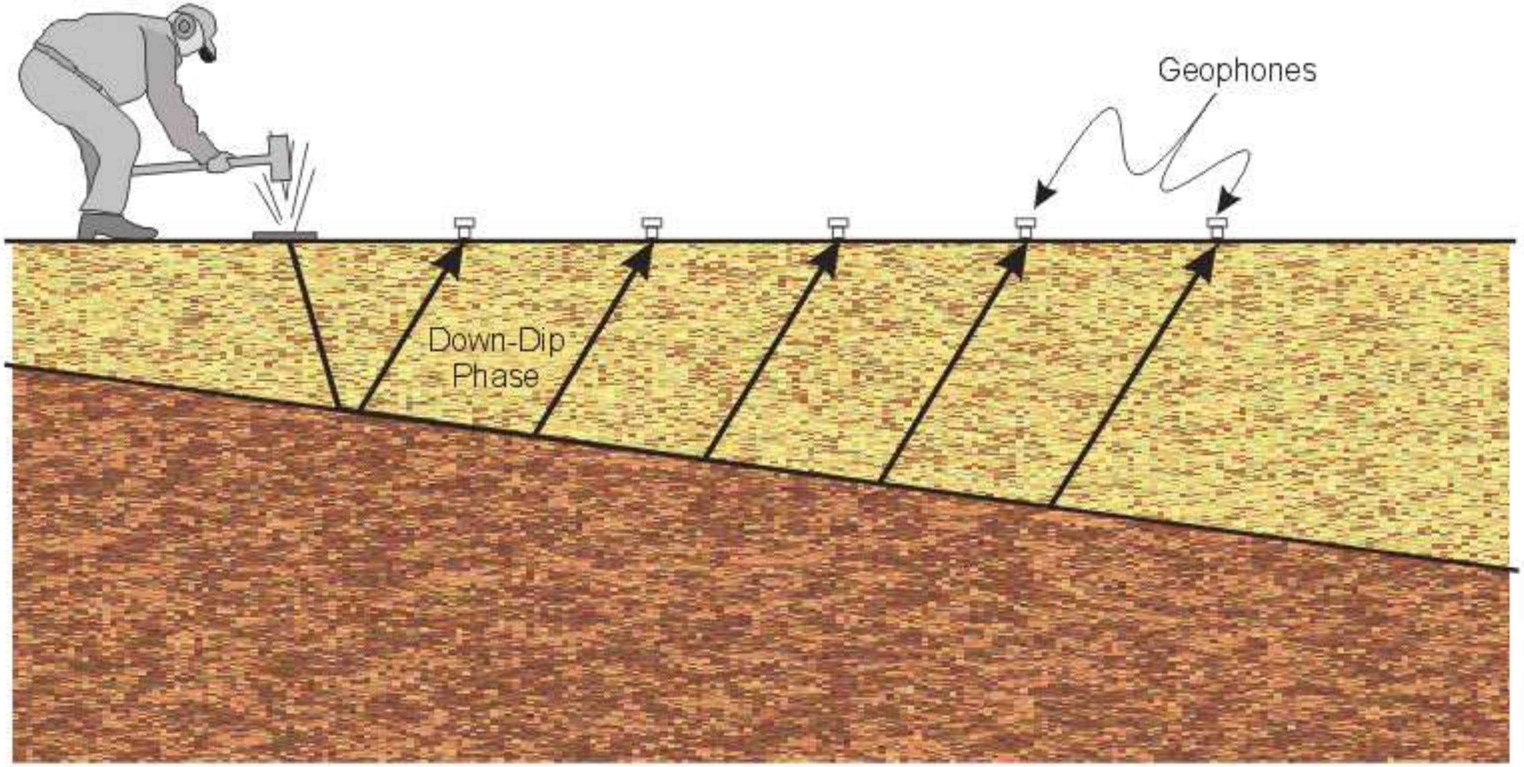
Summary of apparent velocities and intercept times for a dipping interface.

How do we gather and interpret field data ?

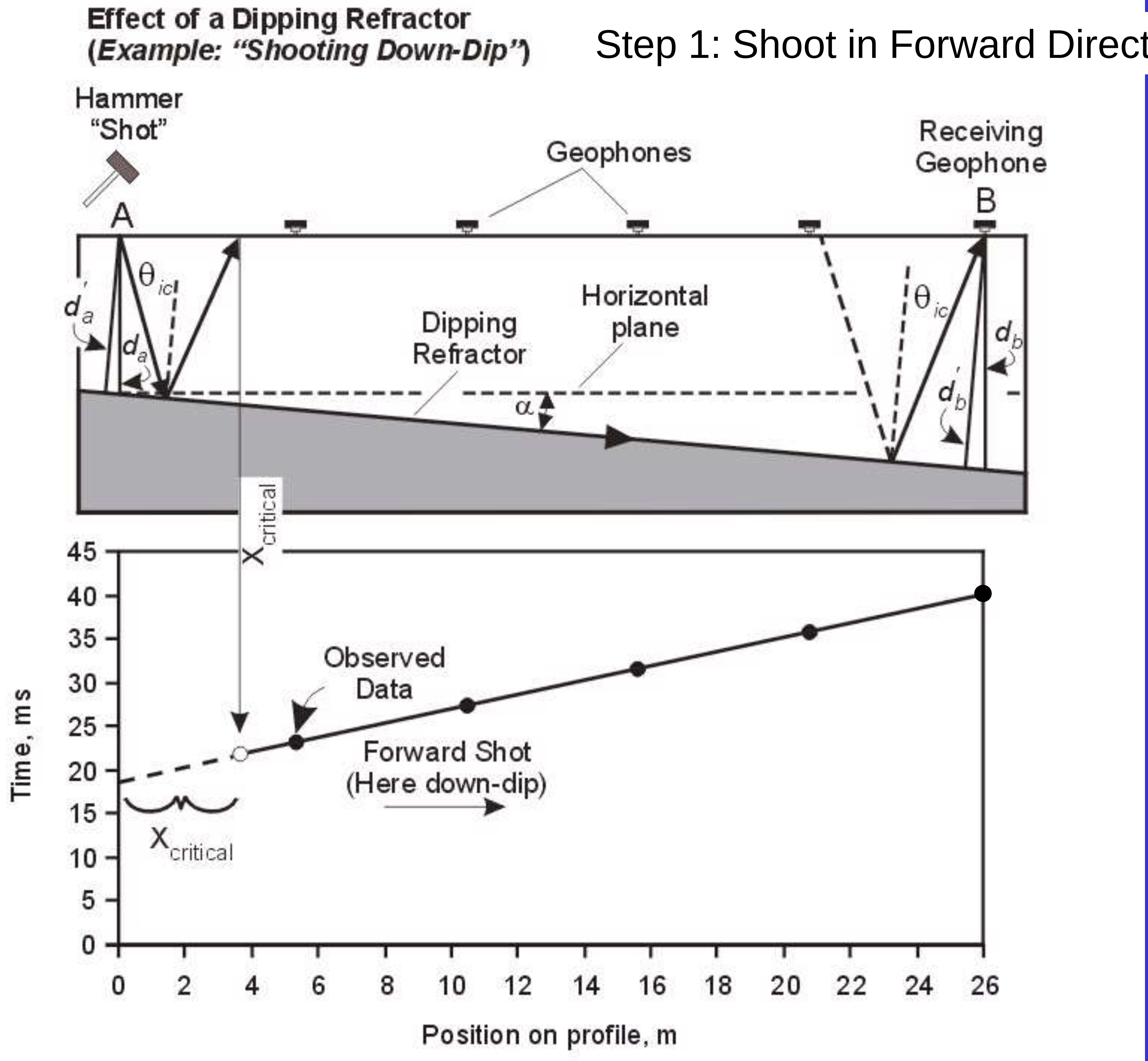
Procedure: Step 1; The “Forward” Shot.

Seismic Refraction Method:
Conventional “Reversed” Profiling

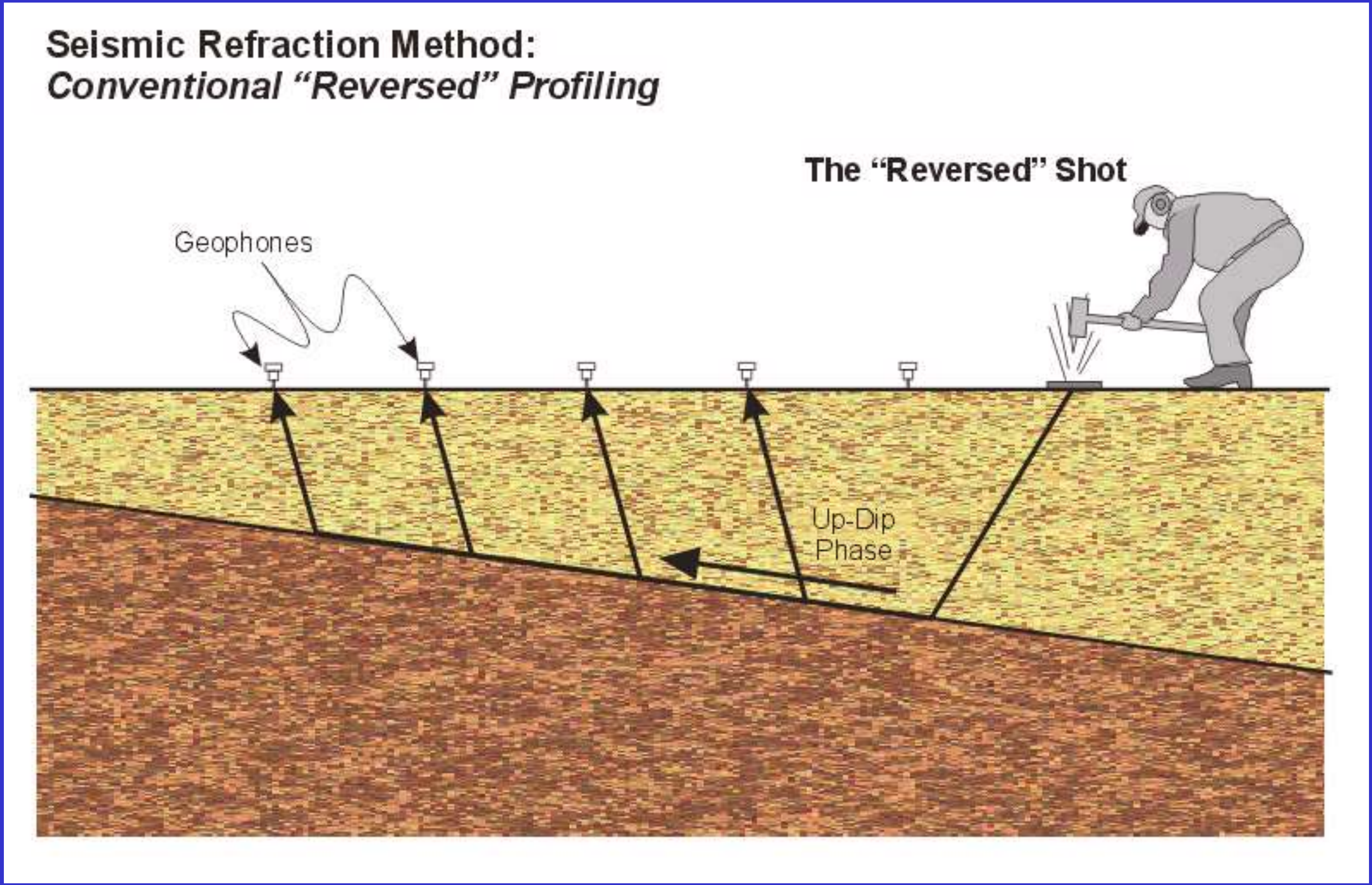
The “Forward” Shot



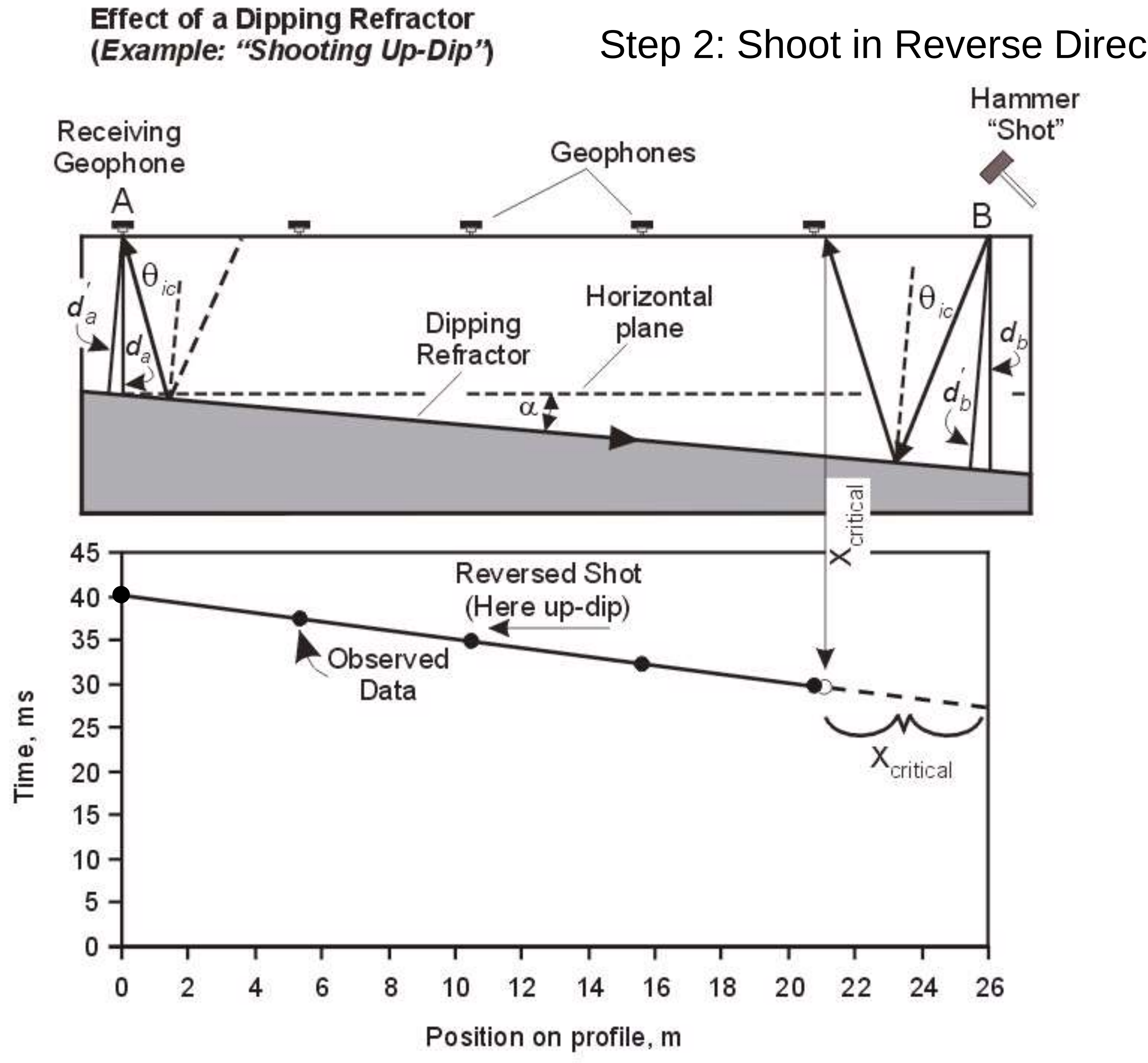
Step 1: Shoot in Forward Direction.



Procedure: Step 2; The “Reverse” Shot.

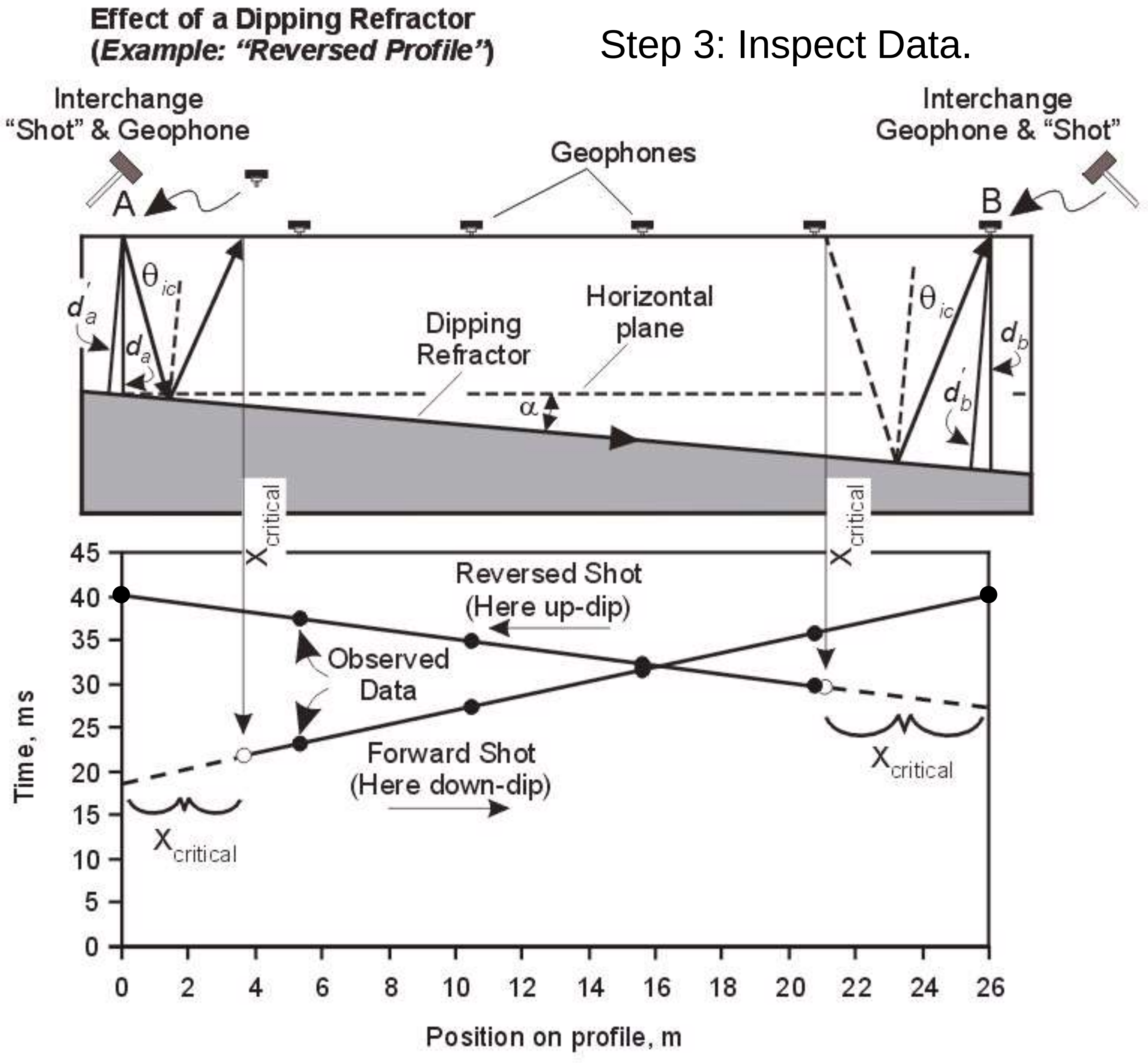


Step 2: Shoot in Reverse Direction.

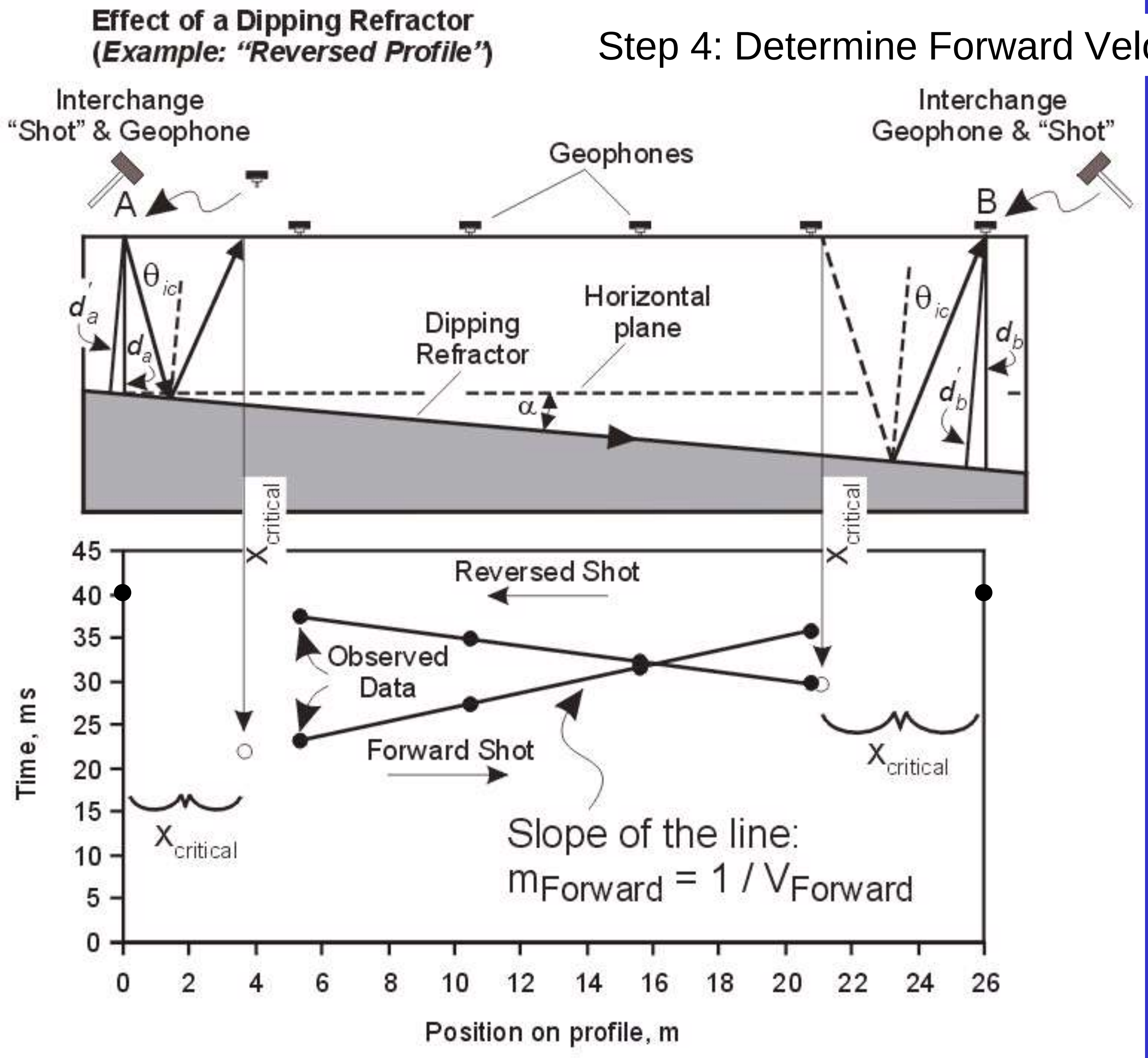


Effect of a Dipping Refractor
(*Example: "Shooting Up-Dip"*)

Step 3: Inspect Data.



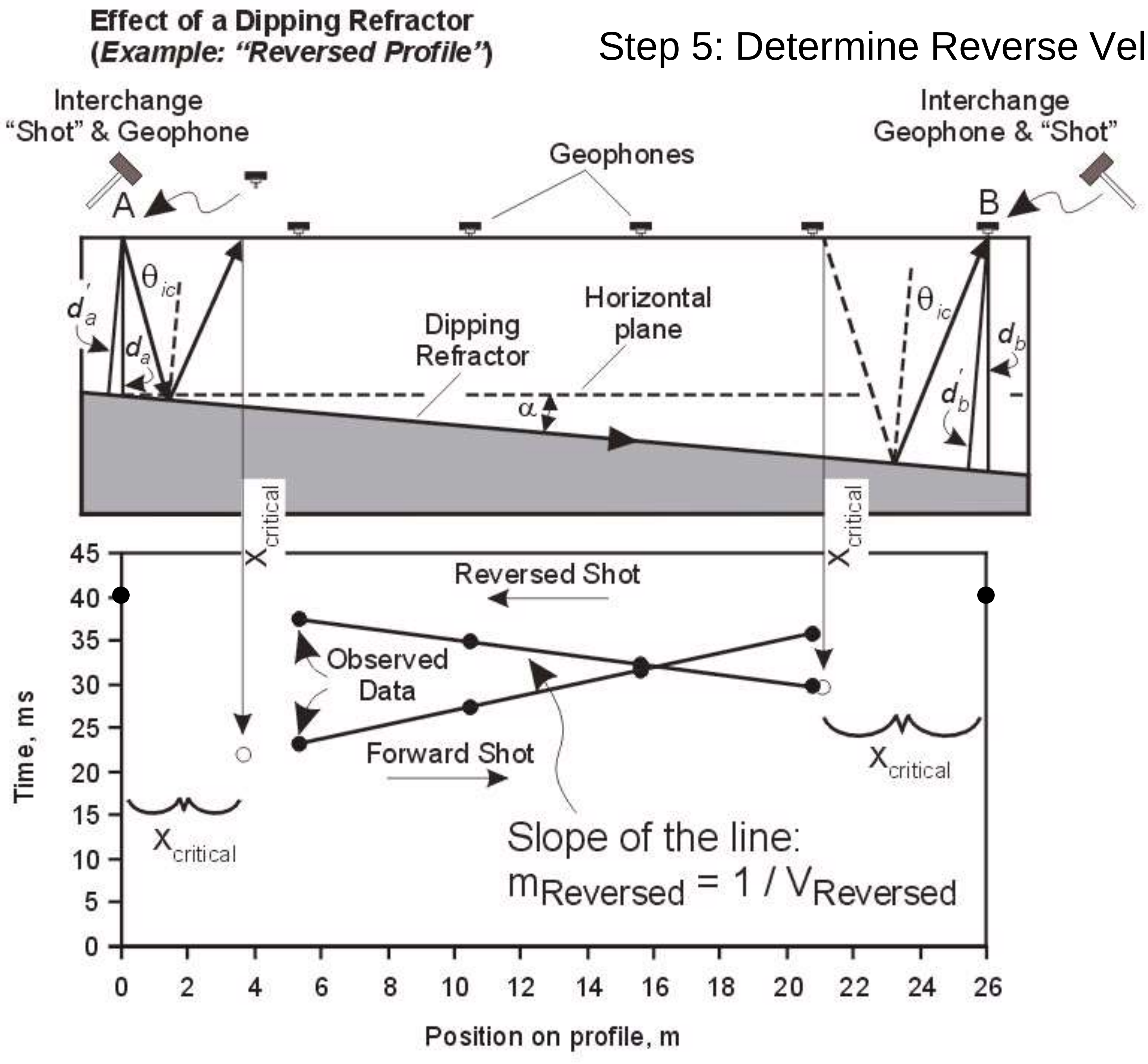
Step 4: Determine Forward Velocity.



Slope of the line:

$m_{Forward} = 1 / V_{Forward}$

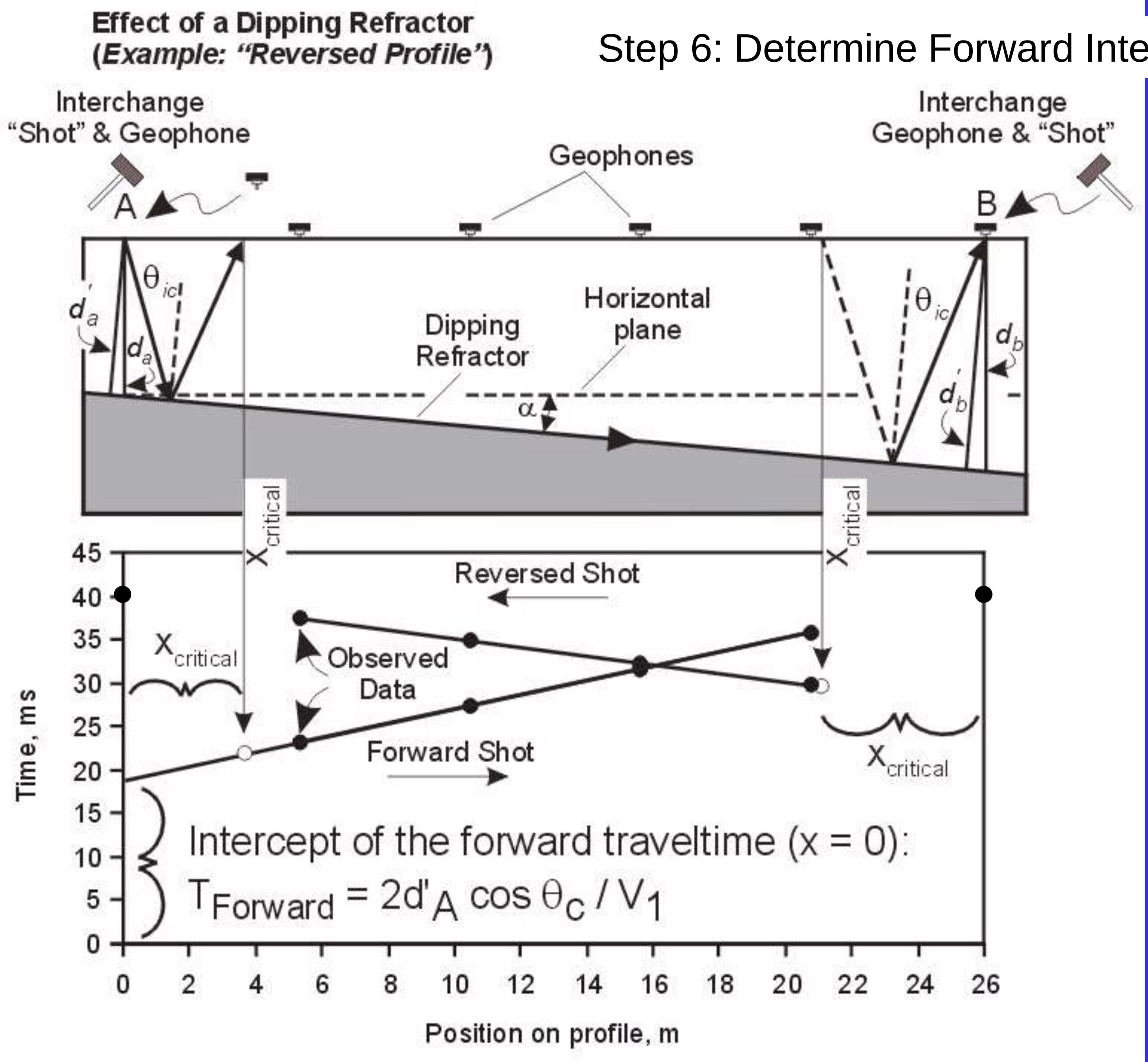
Step 5: Determine Reverse Velocity.



Slope of the line:

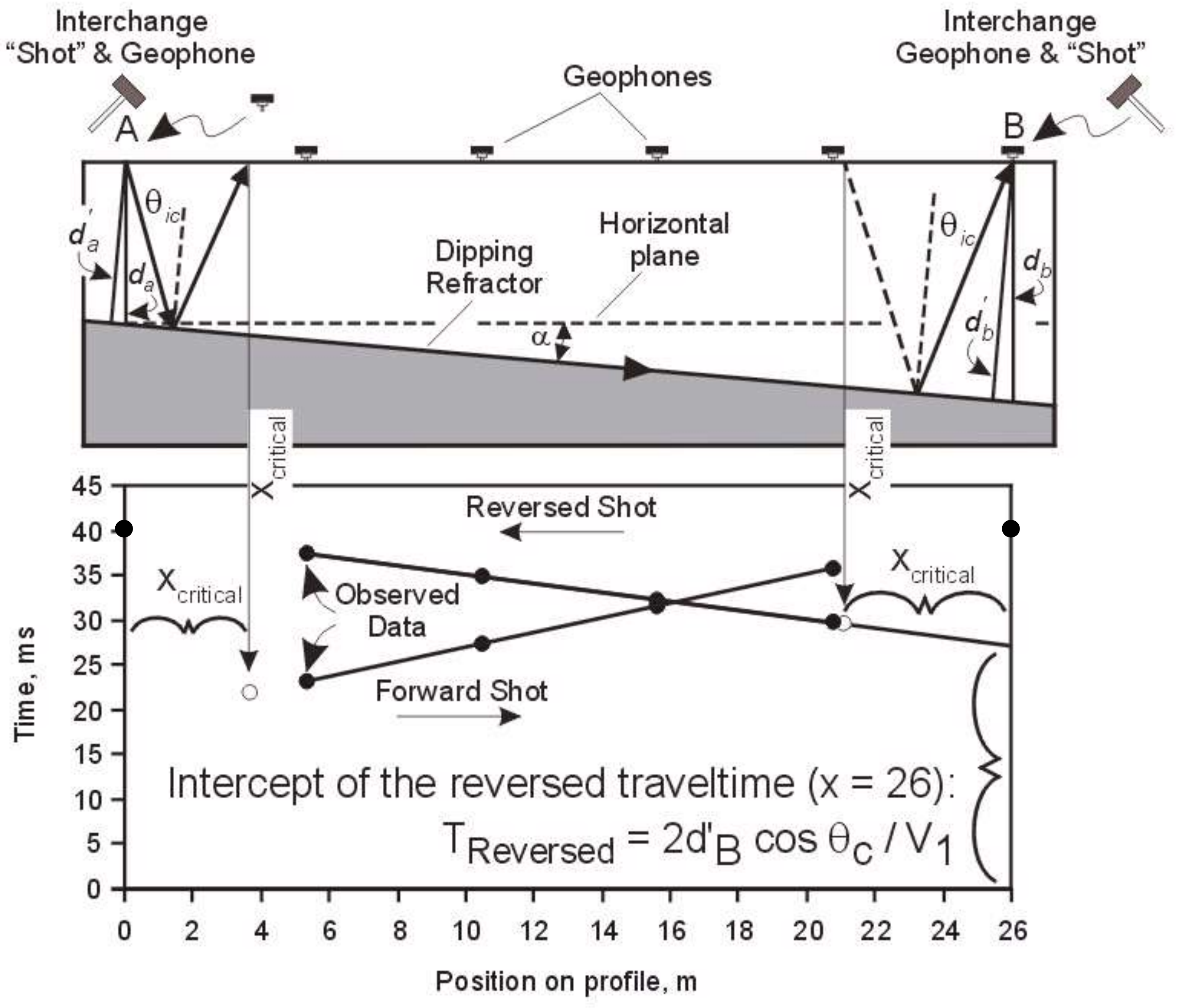
$$m_{Reversed} = 1 / V_{Reversed}$$

Step 6: Determine Forward Intercept.



Step 7: Determine Reverse Intercept.

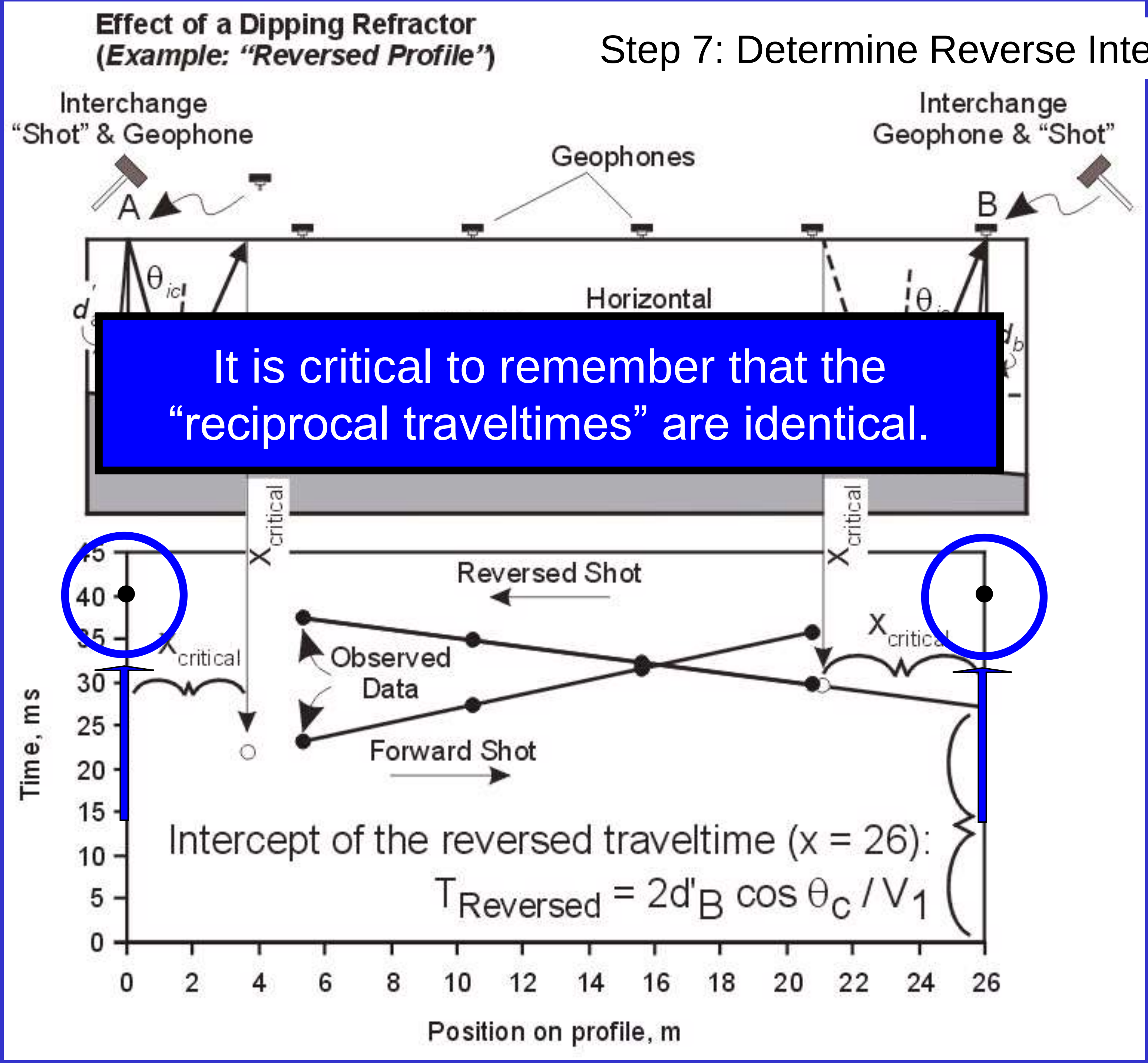
Effect of a Dipping Refractor
(*Example: "Reversed Profile"*)



Intercept of the reversed traveltime (x = 26):

$$T_{Reversed} = 2d'_B \cos\theta_c / V_1$$

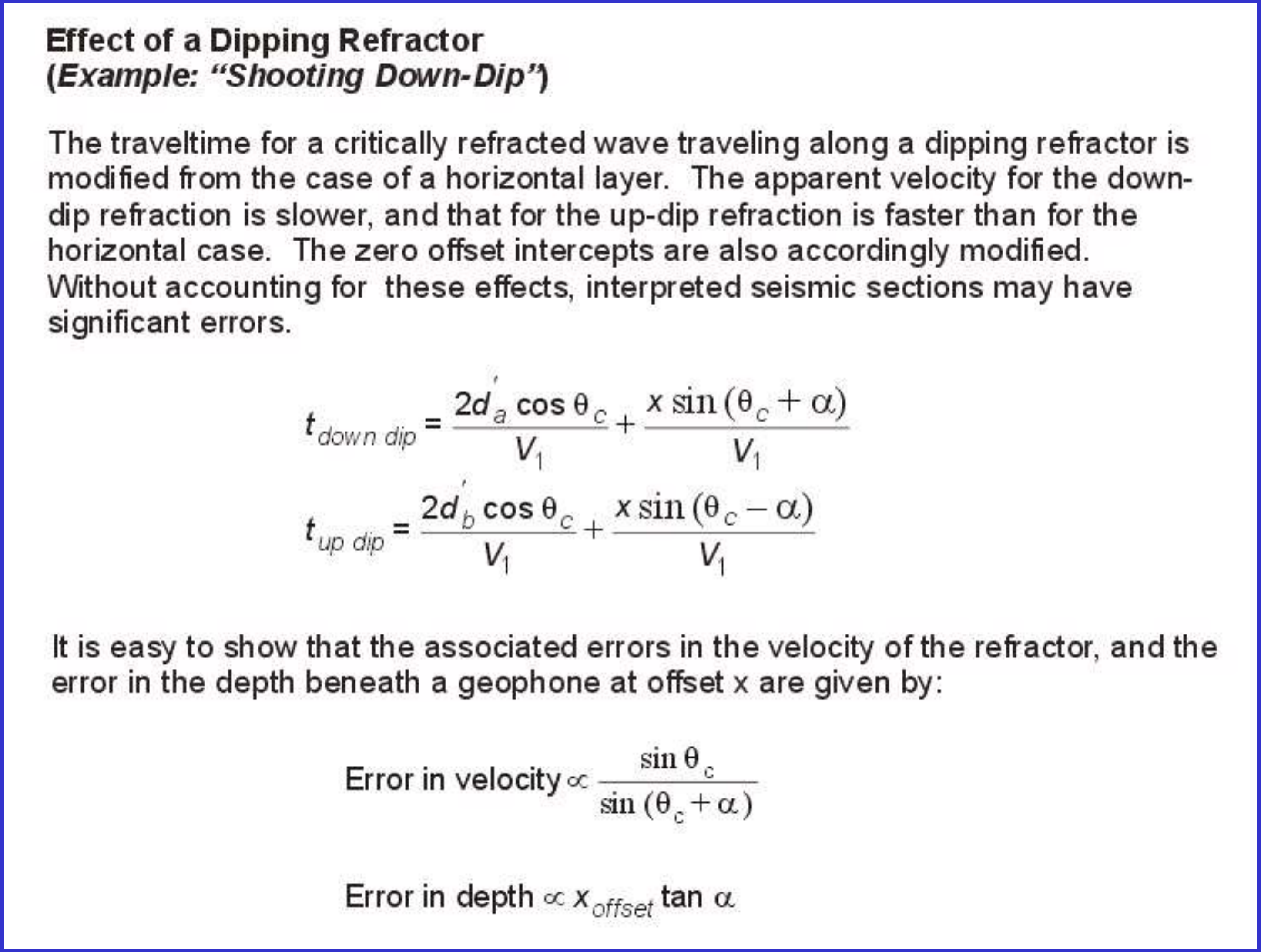
Step 7: Determine Reverse Intercept.



Intercept of the reversed traveltime (x = 26):

$T_{Reversed} = 2d'_B \cos \theta_C / V_1$

It is critical to remember that the "reciprocal traveltimes" are identical.



Effect of a Dipping Refractor
(*Example: "Shooting Down-Dip"*)

The traveltime for a critically refracted wave traveling along a dipping refractor is modified from the case of a horizontal layer. The apparent velocity for the down-dip refraction is slower, and that for the up-dip refraction is faster than for the horizontal case. The zero offset intercepts are also accordingly modified. Without accounting for these effects, interpreted seismic sections may have significant errors.

$$t_{down\ dip} = \frac{2d_a' \cos\theta_c}{V_1} + \frac{x\sin(\theta_c + \alpha)}{V_1}$$

$$t_{up\ dip} = \frac{2d_b' \cos\theta_c}{V_1} + \frac{x\sin(\theta_c - \alpha)}{V_1}$$

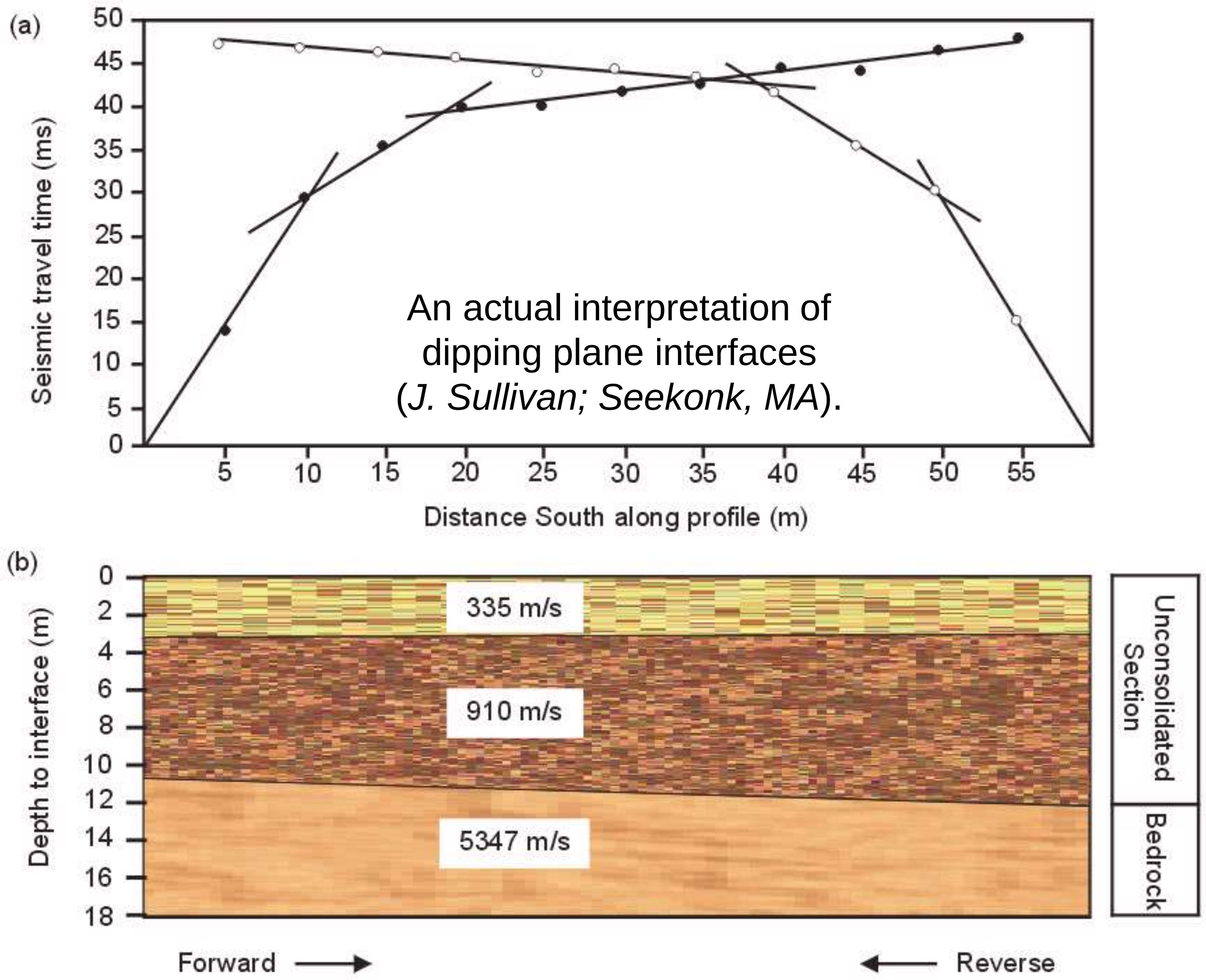
It is easy to show that the associated errors in the velocity of the refractor, and the error in the depth beneath a geophone at offset x are given by:

$$\text{Error in velocity} \propto \frac{\sin\theta_c}{\sin(\theta_c + \alpha)}$$

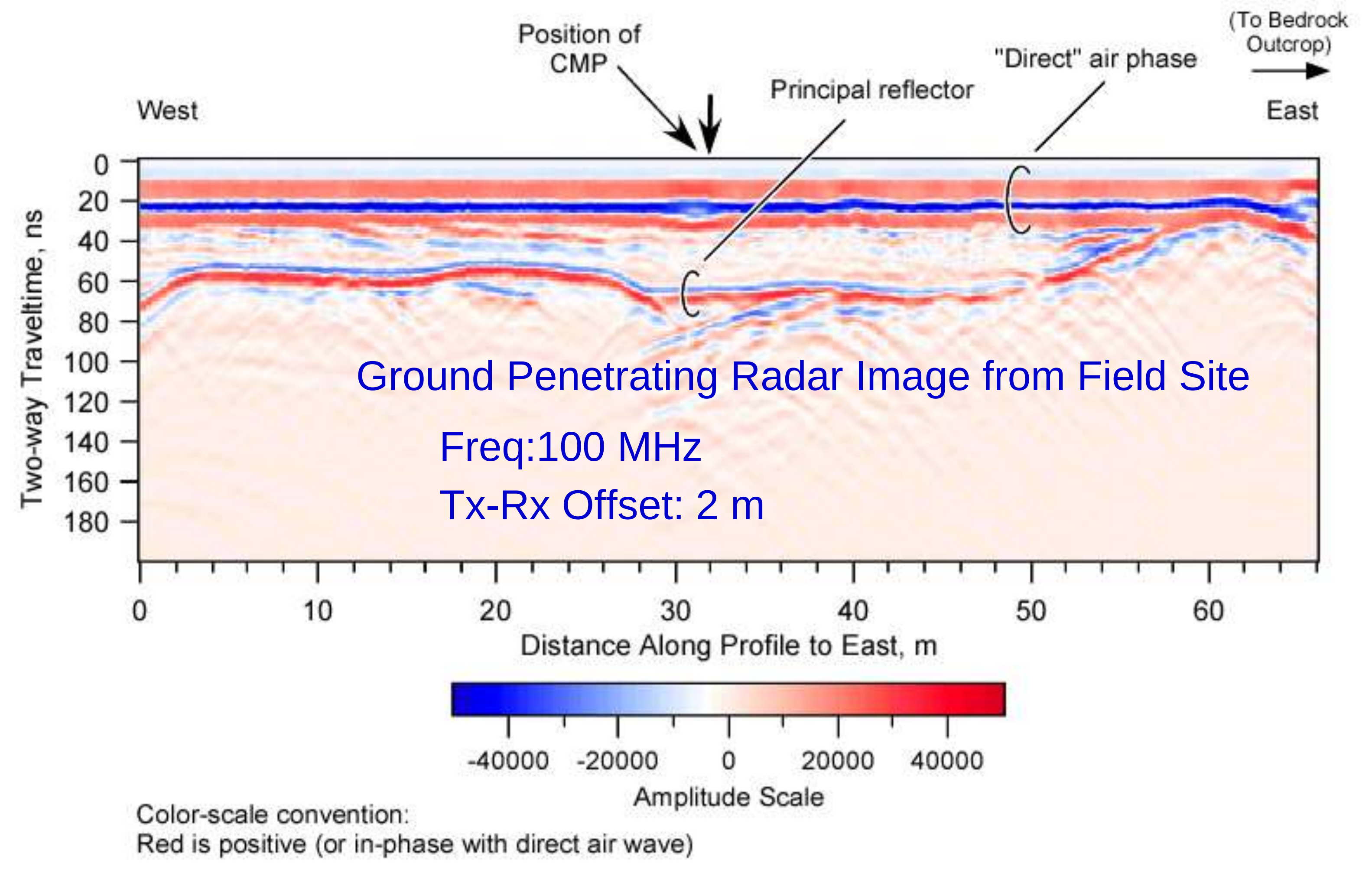
$$\text{Error in depth} \propto x_{offset} \tan\alpha$$

Traveltime relations: Dipping refractor.

Some Examples.



An actual interpretation of dipping plane interfaces (*J. Sullivan; Seekonk, MA*).



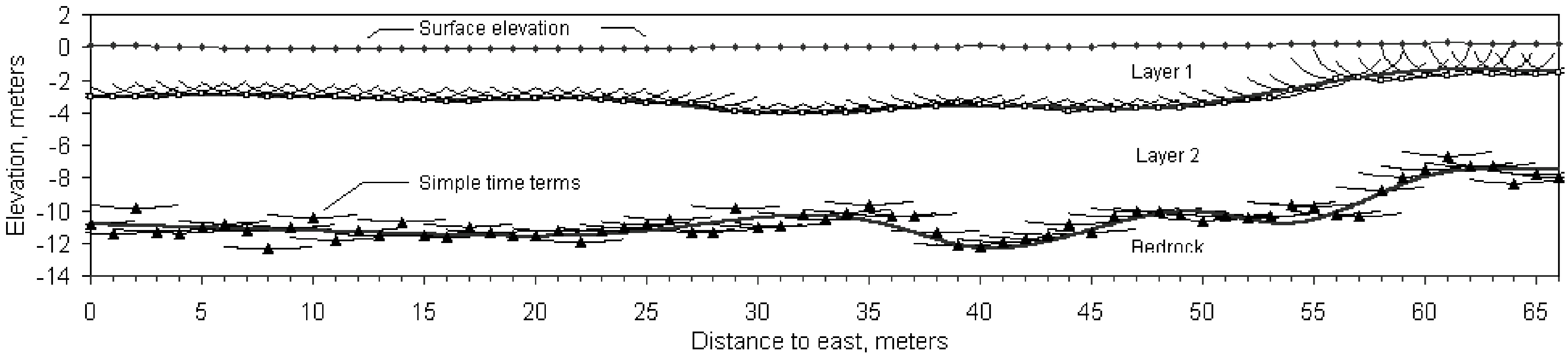
Ground Penetrating Radar Image from Field Site

Freq:100 MHz

Tx-Rx Offset: 2 m

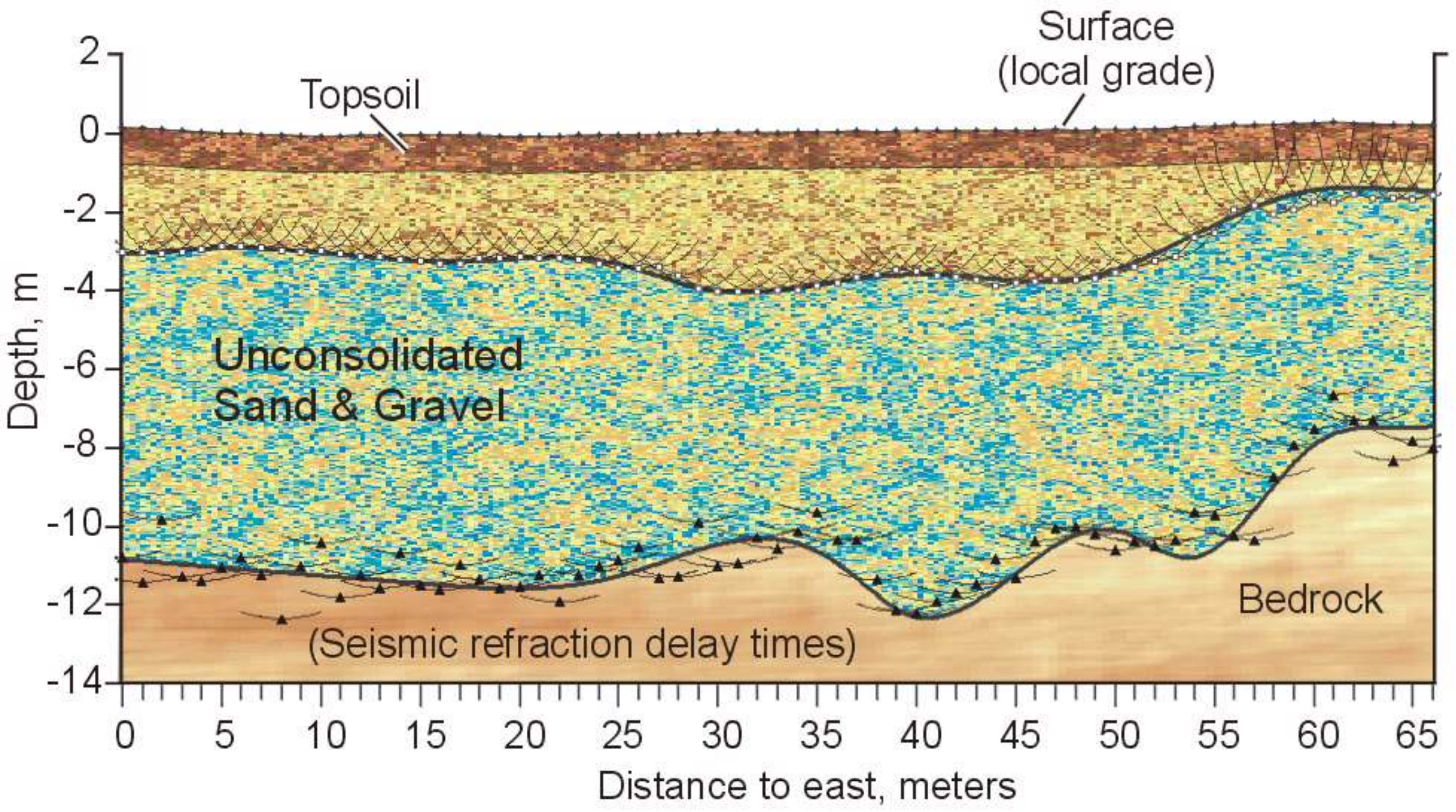
Color-scale convention:
Red is positive (or in-phase with direct air wave)

Application of Seismic Delay Time Method:
Geologic Interfaces Inferred using a Migration Technique
on Seismic Refraction Delay Time Data and GPR Data

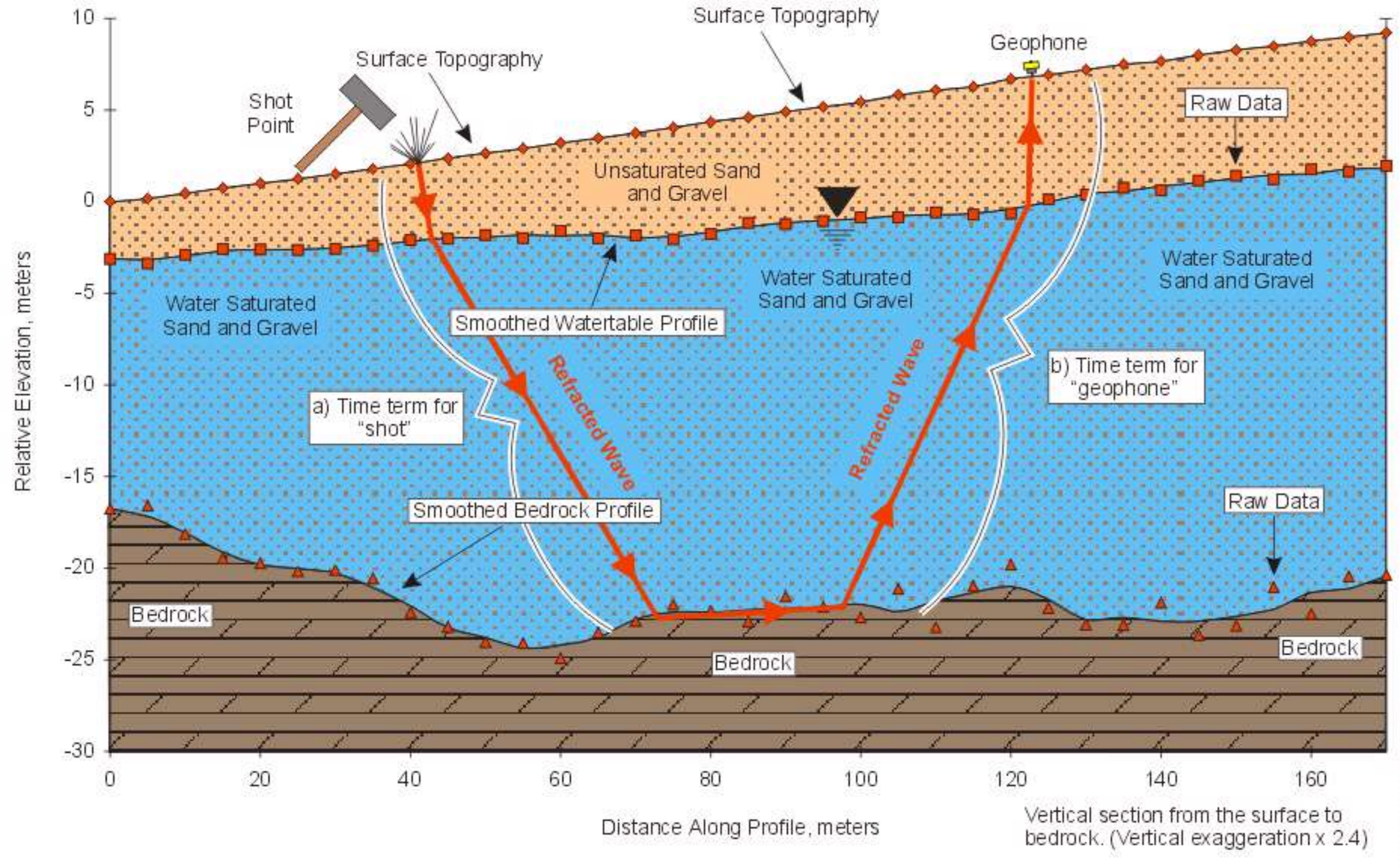


This interpretation uses a procedure which is an alternative to conventional "reversed" profiling methods. Layer 1 is defined using ground penetrating radar (GPR) profile. The inferred structure of Layer 1 is then used with seismic delay times associated with the bedrock refractor to determine undulations in depth to bedrock. The smooth line representing the bedrock interface is the "best fitting" surface that minimizes the time delay residuals for the migrated "depth to the refractor".

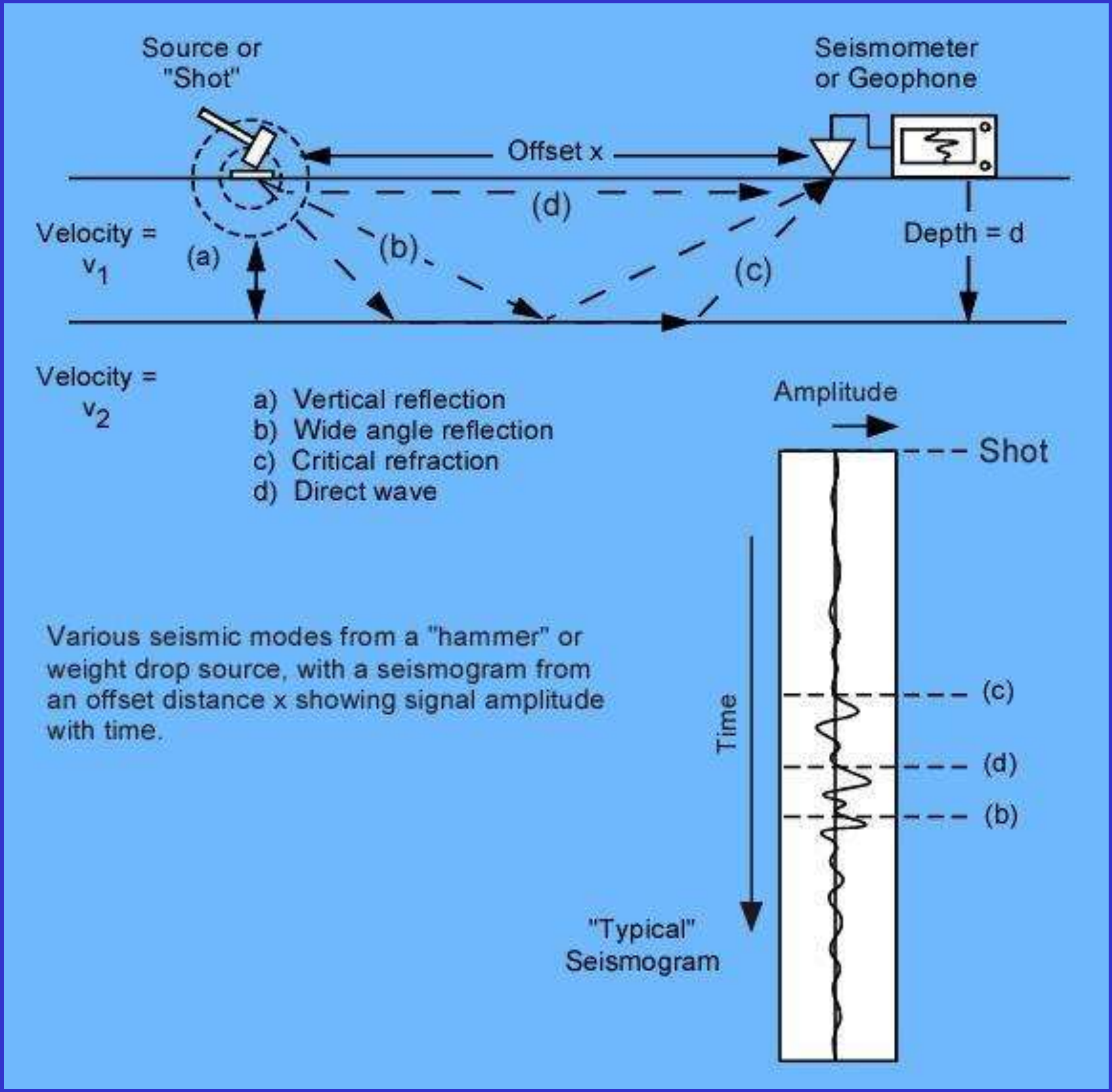
Example of a refined interpretation using a combination of seismic refraction methods and ground penetrating radar.



Composite interpretation using seismic refraction, DC resistivity, EM, GPR and gravity.



Example of refraction study: Palmer River Basin.



Various seismic modes from a "hammer" or weight drop source, with a seismogram from an offset distance x showing signal amplitude with time.

In summary, a seismic interpretation depends on properly identifying and time-picking appropriate phases.

Field Procedure for Seismic Refraction Surveys

(*A checklist for a "typical" seismic refraction sounding*.)

- Begin by deploying a 12 channel recording system w/ 40 Hz geophones at predetermined (1 m?) spacing.

Field Procedure for Seismic Refraction Surveys

(*A checklist for a "typical" seismic refraction sounding*.)

- Begin by deploying a 12 channel recording system w/ 40 Hz geophones at predetermined (1 m?) spacing.
- Perform a walkaway calibration experiment w/ shot points (hammer blows) at offset distances of 1, 5, 10, 15, 20, 25 & 30 meters from the first geophone. This procedure provides 100% redundancy for any set of shot point-geophone offsets.

Field Procedure for Seismic Refraction Surveys

(*A checklist for a "typical" seismic refraction sounding*.)

- Begin by deploying a 12 channel recording system w/ 40 Hz geophones at predetermined (1 m?) spacing.
- Perform a walkaway calibration experiment w/ shot points (hammer blows) at offset distances of 1, 5, 10, 15, 20, 25 & 30 meters from the first geophone. This procedure provides 100% redundancy for any set of shot point-geophone offsets.
- Identify direct wave and refracted wave "first breaks".

Field Procedure for Seismic Refraction Surveys

(*A checklist for a "typical" seismic refraction sounding*.)

- Begin by deploying a 12 channel recording system w/ 40 Hz geophones at predetermined (1 m?) spacing.
- Perform a walkaway calibration experiment w/ shot points (hammer blows) at offset distances of 1, 5, 10, 15, 20, 25 & 30 meters from the first geophone. This procedure provides 100% redundancy for any set of shot point-geophone offsets.
- Identify direct wave and refracted wave "first breaks".
- Reverse profile to identify dip on refractor.

Field Procedure for Seismic Refraction Surveys

(*A checklist for a "typical" seismic refraction sounding*.)

- Begin by deploying a 12 channel recording system w/ 40 Hz geophones at predetermined (1 m?) spacing.
- Perform a walkaway calibration experiment w/ shot points (hammer blows) at offset distances of 1, 5, 10, 15, 20, 25 & 30 meters from the first geophone. This procedure provides 100% redundancy for any set of shot point-geophone offsets.
- Identify direct wave and refracted wave "first breaks".
- Reverse profile to identify dip on refractor.
- Based on these “calibration” runs, design an optimal field plan.

Field Procedure for Seismic Refraction Surveys

(*A checklist for a "typical" seismic refraction sounding*.)

- Begin by deploying a 12 channel recording system w/ 40 Hz geophones at predetermined (1 m?) spacing.
- Perform a walkaway calibration experiment w/ shot points (hammer blows) at offset distances of 1, 5, 10, 15, 20, 25 & 30 meters from the first geophone. This procedure provides 100% redundancy for any set of shot point-geophone offsets.
- Identify direct wave and refracted wave "first breaks".
- Reverse profile to identify dip on refractor.
- Based on these “calibration” runs, design an optimal field plan.
- Execute the optimized survey plan assuring adequate reciprocal shot point-geophone data for both conventional reversed profiling as well as a delay time analysis.

Field Procedure for Seismic Refraction Surveys

(*A checklist for a "typical" seismic refraction sounding*.)

- Begin by deploying a 12 channel recording system w/ 40 Hz geophones at predetermined (1 m?) spacing.
- Perform a walkaway calibration experiment w/ shot points (hammer blows) at offset distances of 1, 5, 10, 15, 20, 25 & 30 meters from the first geophone. This procedure provides 100% redundancy for any set of shot point-geophone offsets.
- Identify direct wave and refracted wave "first breaks".
- Reverse profile to identify dip on refractor.
- Based on these “calibration” runs, design an optimal field plan.
- Execute the optimized survey plan assuring adequate reciprocal shot point-geophone data for both conventional reversed profiling as well as a delay time analysis.
- Separate shot point time-terms from receiver time-terms.

Field Procedure for Seismic Refraction Surveys

(*A checklist for a "typical" seismic refraction sounding*.)

- Begin by deploying a 12 channel recording system w/ 40 Hz geophones at predetermined (1 m?) spacing.
- Perform a walkaway calibration experiment w/ shot points (hammer blows) at offset distances of 1, 5, 10, 15, 20, 25 & 30 meters from the first geophone. This procedure provides 100% redundancy for any set of shot point-geophone offsets.
- Identify direct wave and refracted wave "first breaks".
- Reverse profile to identify dip on refractor.
- Based on these “calibration” runs, design an optimal field plan.
- Execute the optimized survey plan assuring adequate reciprocal shot point-geophone data for both conventional reversed profiling as well as a delay time analysis.
- Separate shot point time-terms from receiver time-terms.
- Shoot in orthogonal direction to determine dip and strike of refractor in three dimensions.

Field Procedure for Seismic Refraction Surveys

(*A checklist for a "typical" seismic refraction sounding*.)

- Begin by deploying a 12 channel recording system w/ 40 Hz geophones at predetermined (1 m?) spacing.
- Perform a walkaway calibration experiment w/ shot points (hammer blows) at offset distances of 1, 5, 10, 15, 20, 25 & 30 meters from the first geophone. This procedure provides 100% redundancy for any set of shot point-geophone offsets.
- Identify direct wave and refracted wave "first breaks".
- Reverse profile to identify dip on refractor.
- Based on these “calibration” runs, design an optimal field plan.
- Execute the optimized survey plan assuring adequate reciprocal shot point-geophone data for both conventional reversed profiling as well as a delay time analysis.
- Separate shot point time-terms from receiver time-terms.
- Shoot in orthogonal direction to determine dip and strike of refractor in three dimensions.

Field Procedure for Seismic Refraction Surveys

(*A checklist for a "typical" seismic refraction sounding*.)

- Begin by deploying a 12 channel recording system w/ 40 Hz geophones at predetermined (1 m?) spacing.
- Perform a walkaway calibration experiment w/ shot points (hammer blows) at offset distances of 1, 5, 10, 15, 20, 25 & 30 meters from the first geophone. This procedure provides 100% redundancy for any set of shot point-geophone offsets.
- Identify direct wave and refracted wave "first breaks".
- Reverse profile to identify dip on refractor.
- Based on these “calibration” runs, design an optimal field plan.
- Execute the optimized survey plan assuring adequate reciprocal shot point-geophone data for both conventional reversed profiling as well as a delay time analysis.
- Separate shot point time-terms from receiver time-terms.
- Shoot in orthogonal direction to determine dip and strike of refractor in three dimensions.

Each of these wave modes (or ‘phases’) provide useful, oftentimes essential, information on the subsurface

In addition, strong analogies exist between
- Seismic (acoustic or mechanical) phenomena and
- Ground penetrating radar (electromagnetic) signals.

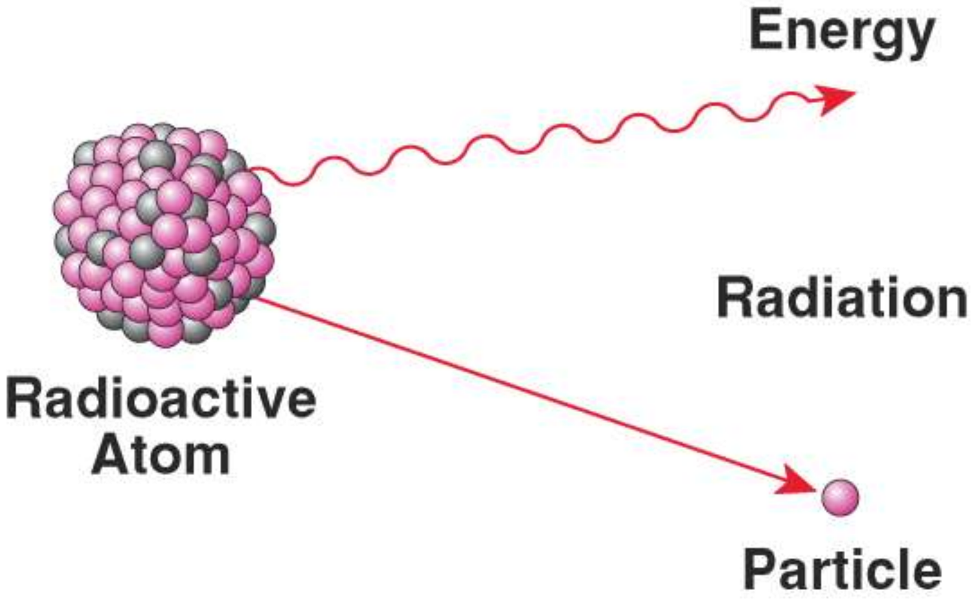
RADIOMETRIC METHODS FOR AGE DETERMINATION

CONTENTS

1. Introduction
2. Characteristics
3. Radiometric dating
4. Uranium-Lead method
5. Potassium-Argon method
6. Rubidium-strontium method
7. Carbon Dating
8. Conclusion

INTRODUCTION

- The spontaneous disintegration of a heavy nuclei with the emission of certain harmful radiations is called RADIOACTIVITY.



CHARACTERISTICS

- Excess or lacking of neutrons, relative to the protons, affects the stability of the nucleus.
- Heavy nuclei are generally unstable(83-92).
- Rate of disintegration of radioactive sample at any instant of time is directly proportional to the number of atoms present in the sample at that instant of time- LAW OF RADIOACTIVE DECAY.

$$dN/dt = -\lambda N$$

- The time during which half the original number of atoms disintegrate – HALF LIFE

$$T= 0.693/\lambda$$

RADIOMETRIC DATING

- Technique used to date materials such as rocks by observing the abundance of naturally occurring radioactive isotope and it’s decay products.

RADIOMETRIC DATING METHODS

1. URANIUM-LEAD
2. POTASSIUM ARGON
3. RUBIDIUM STRONTIUM
4. CARBON DATING

URANIUM – LEAD DATING METHOD

- Uranium- Lead dating is one of the oldest and if done properly one of the most accurate.
- Uranium comes as two common isotopes; U235 and U238.
- Both are unstable and radioactive, shedding nuclear particles in a cascade that doesn't stop until they become lead (Pb).
- The two cascades are different—U235becomes Pb207(half life-704 million years) and U238 becomes Pb206 (half life- 4.47 billion years).

- Lead atoms created by uranium decay are trapped in the crystal and build up in concentration with time; helping us in dating.
- The favourite mineral among U-Pb daters is zircon ($ZrSiO_4$), for several good reasons.
- Some zircons are obviously disturbed and can be ignored, while other cases are harder to judge. In these cases, the concordia diagram is a valuable tool.
- Uranium- lead dating works only for metamorphic and igneous rocks.

POTASSIUM ARGON METHOD

- The potassium-argon (K-Ar) isotopic dating method is especially useful for determining the age of lavas.
- Potassium has one radioactive isotope (^{40}K).
- Potassium-40 decays with a half-life of 1250 million years.
- What simplifies things is that potassium is a reactive metal and argon is an inert gas.
- The mineral Sanidine, the high-temperature form of Potassium Feldspar, is the most desirable.
- Meteoric and volcanic rocks have been analysed by this method.

RUBIDIUM- STRONTIUM METHOD

- The utility of Rubidium Strontium isotope system results from the fact that ^{87}Rb (one of the isotopes of Rubidium) decays to ^{87}Sr with a half-life of 49 billion years.
- The method is applicable to very old rocks because the transformation is extremely slow: the half-life, or time required for half the initial quantity of rubidium-87 to disappear; is approximately 50 billion years.
- It is used to date igneous and metamorphic rocks.

CARBON DATING

- Carbon dating is a variety of radioactive dating; applicable only to matter which was once living.
- Neutron produced by cosmic ray bombardment produces radioactive isotope Carbon-14
- Carbon-14 decays with a half-life of about 5730 years by the emission of an electron, disintegrating to nitrogen 14.
- Carbon dating only works on fossils that used to be alive. You can not carbon date a rock or sedimentary layer.

Parent Isotope and mode of Decay	Daughter isotope	Half life (m.y)	Geological age range	Occurrence and useful geological materials
U-238, alpha & beta	Pb-206	4,510	older than about 10 m.y	Accessory minerals, e.g. zircon and sphene, in igneous and metamorphic rocks
U-235, alpha & beta	Pb-207	713		common igneous and metamorphic whole rocks. Many have sufficient U and Pb for age determination
Th-232, alpha & beta	Pb-208	13,900		U & Th minerals. Rare and restricted to uncommon geological environments.
K-40, electron capture and beta	Ar-40 Ca-40	11,850 1,470	older than about 0.1 m.y	Many common rock forming minerals, e.g.biotite, muscovite, sanidine, hbl, glauconite, basic igneous whole rocks, slates
Ru-87, beta	Sr-87	50,000	older than about 10 m.y	Most K-minerals contain sufficient Rb for age work, e.g. biotite, muscovite, k-felds, glauconite, Many common igneous and metamorphic rocks, some sedimentary whole rocks, e.g. shales, yeild time of diagenesis
Sm-147, alpha	Nd-143	106,000	older than about 10 m.y	Mafic and ultramafic igneous rocks

CONCLUSION

- Radiometric methods are one of the most effective tools in age determination.
- They are even used to date fossils.
- Radiometric methods other than the ones listed above are: samarium-147 to neodyium-143, cosmogenic nuclide dating and so on.

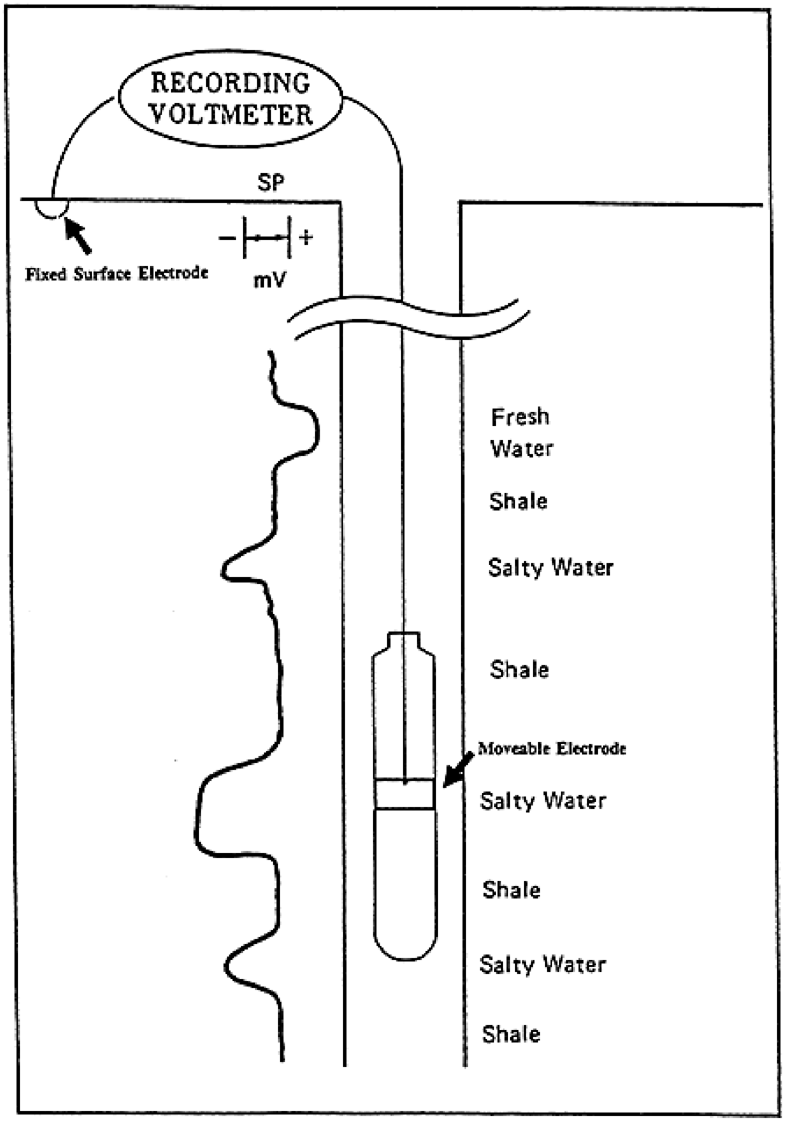
REFERENCE

- Harry Y Mc Sween, Steven M.Richardson and Maria E Uhle(2009). Geochemistry, PVT LTD, New Delhi, pp 290-298.
- John V Walther(2010). Essentials of Geochemistry, Jones and Barhett, New Delhi,pp 383-400
- Brian Mason, Carleton B. Moore(1982). Principles of Geochemistry, John Wiley and Sons, pp 198-208
- C.S. Pichmuthu (1985), Archaean Geology, Oxford & IBH pub.Co. p.125
- http://creation.mobi/radiometric-dating-age-of-earth
- http://geology.about.com/od/geotime_dating/a/uraniumlead.htm

Borehole Geophysical Techniques

Spontaneous Potential (SP) Log

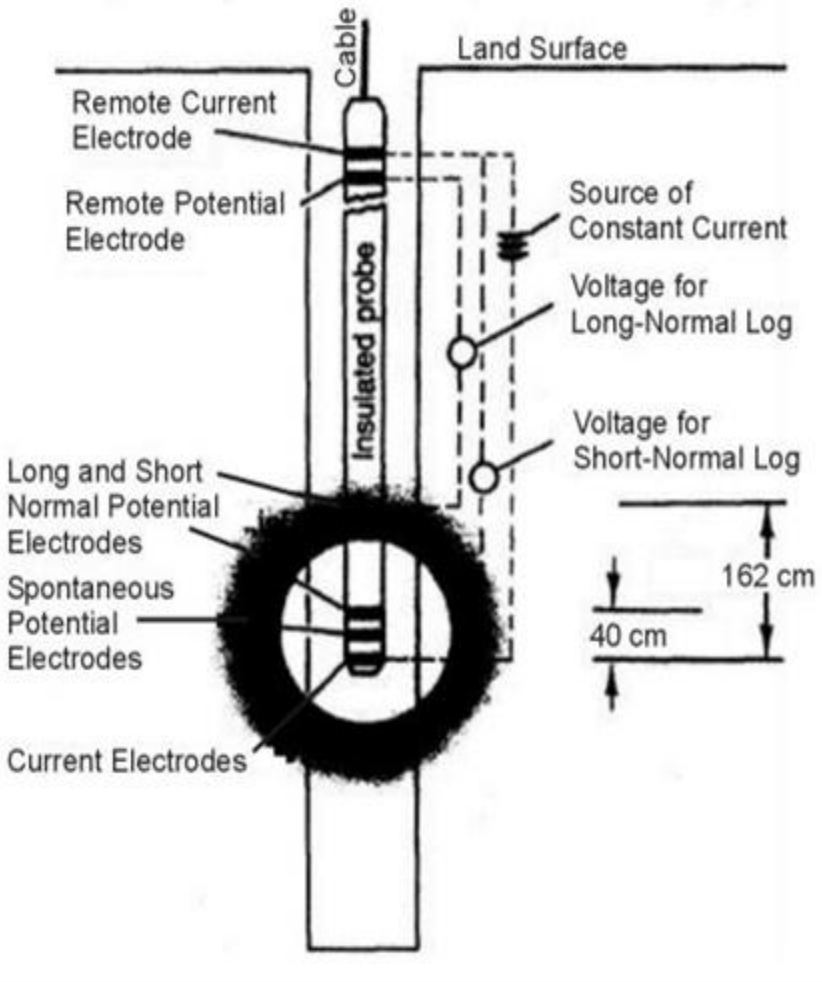
- **Measures natural potential difference between a movable electrode in the borehole and the fixed potential of a surface electrode**
- +ve deflection- Cations- Mud - *Membrane Potential*
- -ve deflection-Anions-SSt, Porous Lst- *Liquid Junction Potential*
- **Applications**:
  -correlation
  -lithology indication
  -porosity and permeability
  -resistivity
  -salinity



Resistivity Log

- **Measures the resistivity of a formation by passing electricity through electrodes placed at standard spacings**
- Shale, Clay & Saltwater –> Low
- Freshwater sand –> Moderate to high
- Cemented sandstone & Nonporous limestone –> High
- Various resistivity tools in use

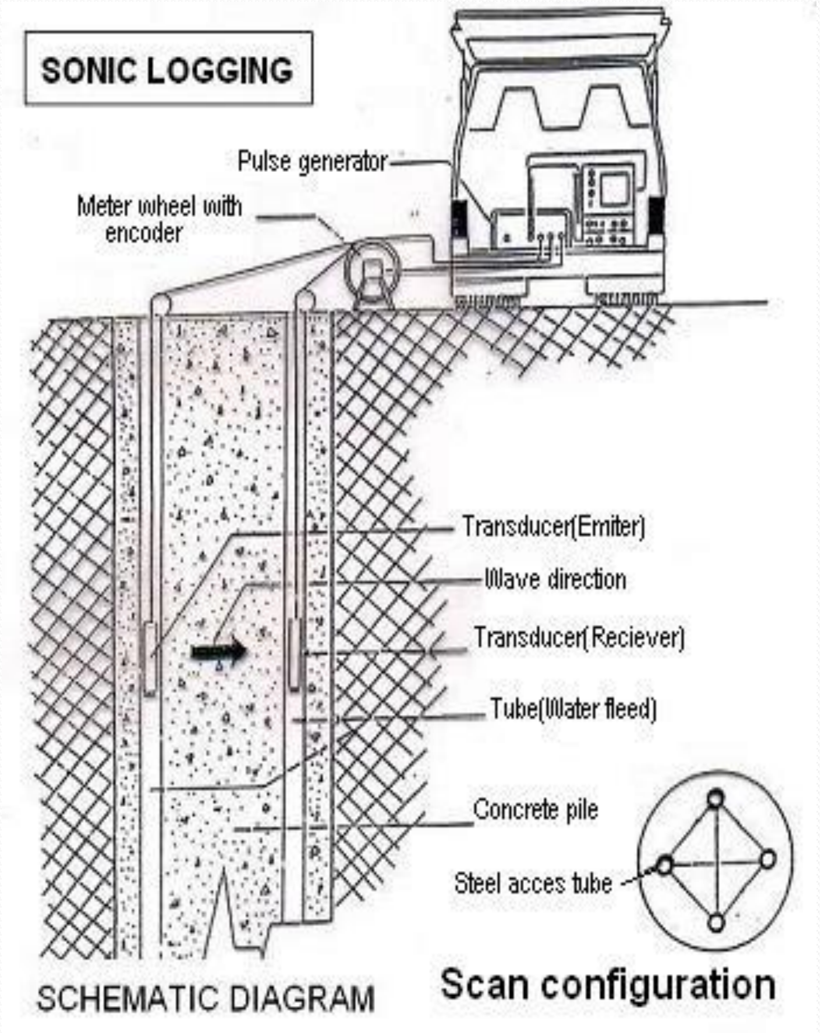
- **Applications:**
  -porosity and permeability
  -fluid characteristics
  -oil detection
  -metallic mineral exploration



Sonic Log

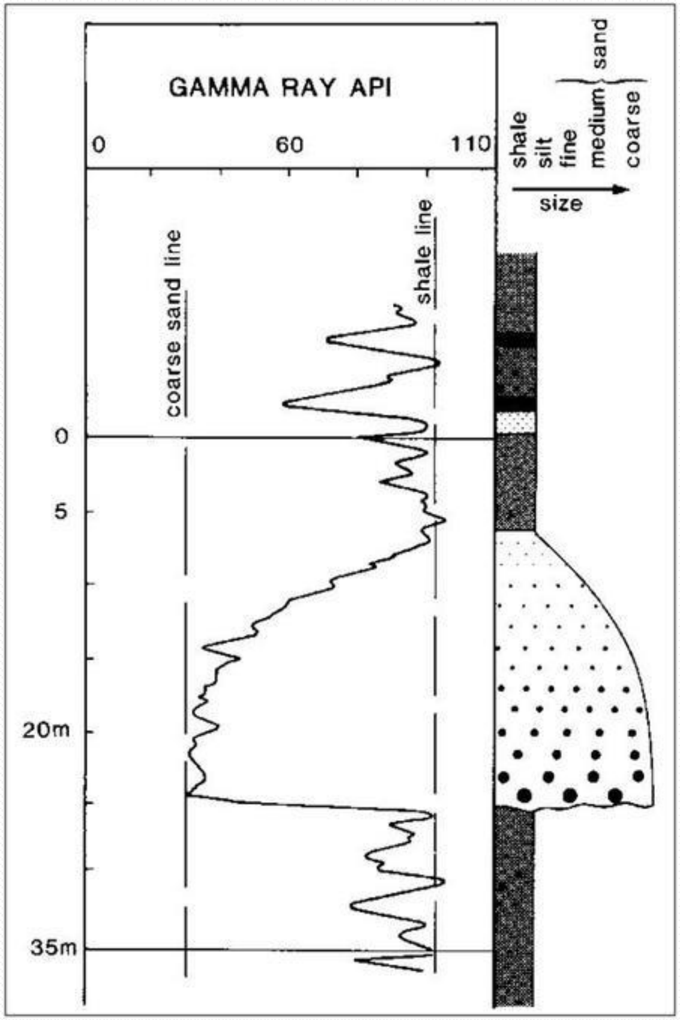
- **Measures the time of first arrival of sound pulses from a transmitter at a receiver**
- The arrival time depends on the density of the formation, which in turn depends on the porosity and permeability

- **Applications:**
  -Density of the formation
  -Porosity calculation for known lithology
  -Calibration of seismic data



Gamma Logging

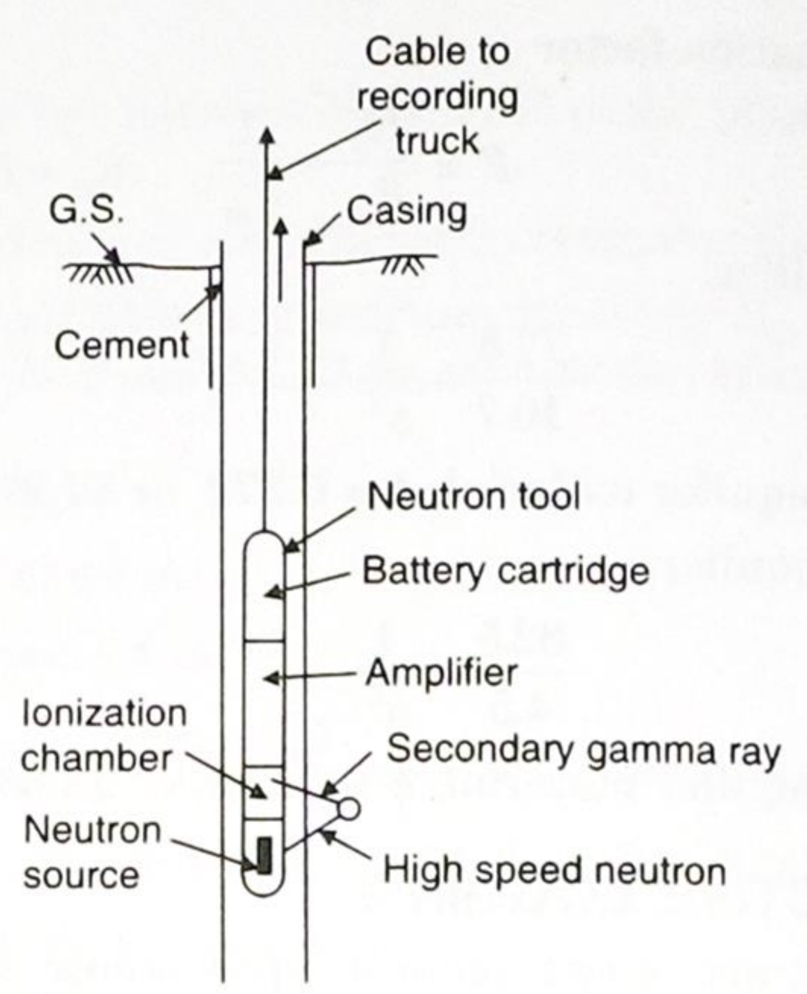
- **It measures the natural radioactivity of the formation.**
- It provides a measurement of the muddiness of a unit.
- Quartz arenite and clean carbonate sediments -> low log response.
- Mudstone, volcanic ash and granite wash -> high log response.
- It can be run in open as well as cased holes.
- **Applications:**
  -Identification of Lithology
  -Facies correlation.
  -Evaluation of radioactive minerals.
  -Identify shale volumes.



Neutron Logging

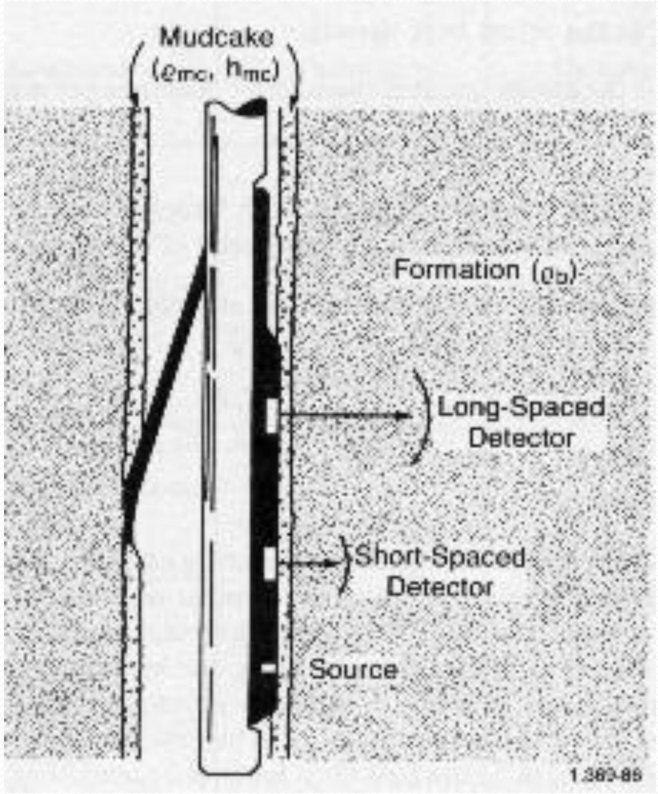
- **It measures H^+ content of the borehole environment.**
- A fast neutron source is used to bombard the rock. When any individual neutron collides with a H^+ ion, some of the neutron's energy lost and it slows down.
- A large number of slow neutrons recorded indicates a large amount of fluid i.e., high porosity
- **Applications:**

  -measures formation water, hydrocarbons, and bound water in clay minerals, gypsum, etc.



Density Logging

- **This tool operate by emitting gamma radiation and detecting the proportion of the radiation that returns to detectors on the tool.**
- Denser the formation, more electrons are present, and more energy is lost due to collisions.
- If matrix density is known, then energy loss is directly related to porosity.
  (density decreases with increasing porosity)
- **Applications:**
  -Identification of lithology.
  -Measurement of bulk density & porosity.



AIRBORNE ELECTROMAGNETIC SURVEY

ABSTRACT

An attractive aspect of Electromagnetic Induction survey is the opportunity for the Air borne electromagnetic survey. In this report we will explain the airborne Electro-magnetic (AEM) techniques which deals with a number of topics relating to airborne EM survey systems and methods. These AEM topics include: Basic Principles, Factors Affecting Detect ability, Survey Data Presentation, Applications of AEM and the advantages of using it.

Table Of Content

CHAPTER 01

1. Introduction 04
2. Objective 04
3. History 05
4. Basic Principles 06
5. Systems 07
6. Factors Affecting Detect-ability 07
   6.1 Signal to Noise Ratio 07
   6.2 Penetration 08
   6.3 Discrimination 08
   6.4 Resolution 08
   6.5 Conductivity-Width Aperture 08
   6.6 Lateral Coverage 08
7. Applications 09
   7.1 Structural Mapping 09
   7.2 Mineral Exploration 09
   7.3 Environmental 09

7.4 Oil and Gas Exploration 10

7.5 Freshwater Mapping 10

8. Suvey Data Presentation 10
9. Advantages Of AEM
10. Conclusion 12
11. References 13

1. INTRODUCTION

Airborne geophysical surveying is a process of measuring the variation in several key physical or geochemical parameters of the earth. The most important parameters measured are conductivity (which is the inverse of resistivity), magnetic susceptibility, density, and radioactive element concentration. Any change in the earth's near surface that causes a measurable change in these parameters presents a potential application for airborne geophysics.

Electromagnetic surveys map the three dimensional variation in conductivity, caused by changes in mineralogy, intensity of alteration, water content or salinity. Airborne magnetic surveys map the variation of the magnetic susceptibility, generally due to changes in the magnetite content of the rock. Gamma-ray spectrometric surveys measure the radiation of one or more of the natural radio-elements: potassium, uranium, or thorium, or a specific man-made radioelement. Airborne gravity surveys map density variations within the earth.

Airborne electromagnetic surveys are used to conduct a rapid survey at a relatively low cost of a broad area to search for metallic conductors. Graphite, pyrite and or pyrrhotite are present in the bedrock and are most commonly what provides the electromagnetic response. The conductivities which is determined through the electromagnetic method varies over seven orders of magnitude between various geologic materials. The strongest responses come from massive sulphides. Fresh water on the other hand is highly resistive and will provide responses very different from salt water.

2. OBJECTIVE

The general objective of our report on **AEM** (Airborne ElectroMagnetic) surveys is to explain the importance of AEM and how it helps in rapid and relatively low-cost search for metallic conductors, e.g. massive sulphides, located in bed-rock and often under a cover of overburden and/or fresh water. This method can be applied in most geological environments except where the country rock is highly conductive or where overburden is both thick and conductive. It is equally well suited and applied to general geologic mapping, as well as to a variety of engineering problems (e.g., fresh water exploration.)

Semi-arid areas, particularly with internal drainage, are usually poor AEM environments. Tidal coasts and estuaries should be avoided. Weathered maific flows can provide strongly conductive backgrounds, particularly flows of Tertiary or Quaternary age.

Conductivities of geological materials range over seven orders of magnitude, with the strongest EM responses coming from massive sulphides, followed in decreasing order of intensity by graphite, unconsolidated sediments (clay, tills, and gravel/sand), and igneous and metamorphic rocks. Consolidated sedimentary rocks can range in conductivity from the level of graphite (e.g. shales) down to less than the most resistive igneous materials (e.g. dolomites and limestones). Fresh water is highly resistive. However, when contaminated by decay material, such lake bottom sediments, swamps, etc., it may display conductivity roughly equivalent to clay and salt water to graphite and sulphides.

Typically, graphite, pyrite and or pyrrhotite are responsible for the observed bedrock AEM responses. The following examples suggest possible target types and we have indicate the grade of the AEM response that can be expected from these targets.

1) Massive volcano-sedimentary stratabound sulphide ores of Cu, Pb, Zn, (and precious metals), usually with pyrite and/or pyrrhotite. Fair to good AEM targets accounting for the majority of AEM surveys.
2) Carbonate-hosted Pb-Zn, often with marcasite, pyrite, or pyrrhotite, and sometimes associated with graphitic horizons. Fair to poor AEM targets.
3) Massive pyrrhotite-pentlandite bodies containing Ni and sometimes Cu and precious metals associated with noritic or other mafic/ultramafic intrusive rocks. Fair to good AEM targets.
4) Quartz veins containing Au with pyrite, sometimes also with Sb, Ag, Bi, etc., in volcanic or sedimentary (and possibly intrusive) rocks. Poor AEM targets.
5) Skarn deposits of Cu, Zn, Pb, and precious metals, usually with pyrite and magnetite, around igneous intrusions. Fair to poor AEM targets.

Conductive targets can be concealed by other geological conductors, "geological noise", such as:

1) Lateral variations in conductive overburden.
2) Graphitic bands in metamorphosed country rock.
3) Altered (to clay facies) mafic-ultramaific rocks.
4) Faults and shear-zones carrying appreciable groundwater and/or clay gouge.
5) Magnetite bands in serpentinized ultramafics.

3. HISTORY

The first successful test flight of an airborne electromagnetic (AEM) system was in 1948. Since then many systems of various design have been built, mounted on fixed wing aircraft and on helicopters.

The photograph at the bottom shows the 'INPUT' time domain airborne electromagnetic system, developed in 1959.

4. BASIC PRINCIPLES

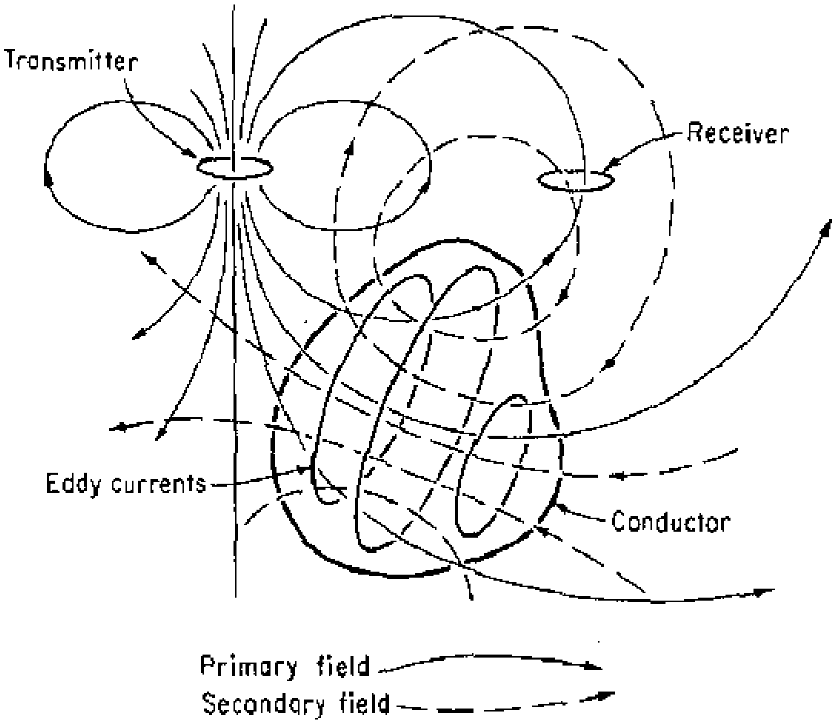
Electromagnetic-induction prospecting methods, both airborne and (most) ground techniques, make use of man-made primary electromagnetic fields in, roughly, the following way: An alternating magnetic field is established by passing a current through a coil, (or along a long wire). The field is measured with a receiver consisting of a sensitive electronic amplifier and meter or potentiometer bridge. The frequency of the alternating current is chosen such that an insignificant eddy-current field is induced in the ground if it has an average electrical conductivity.

If the source and receiver are brought near a more conductive zone, stronger eddy currents may be caused to circulate within it and an appreciable secondary magnetic field will thereby be created. Close to the conductor, this secondary or anomalous field may be compared in magnitude to the primary or normal field (which prevails in the absence of conductors), in which case it can be detected by the receiver. The secondary field strength, Hs, is usually measured as a proportion of the primary field strength, Hp, at the receiver in percent or ppm (parts per million).

Anomaly = Hs / Hp

Increasing the primary field strength increases the secondary field strength proportionally but the "anomaly" measured in ppm or percent remains the same.

.



A generalized picture of electromagnetic induction prospecting.

5. SYSTEMS

1) The transmitter coil may be fixed to aircraft and the receiver mounted in a bird which is trailed some distance behind.
2) Both transmitter and receiver may be mounted in the aircraft.
3) Both are mounted in the bird.

6. FACTORS AFFECTING DETECTABILITY

There are at least six factors that determine whether or not a particular conductor will be detectable with any EM system.

6.1. Signal-to-noise ratio:

In practice, because of "system noise" (Ns) and "geological noise" (Ng), the ability of a system to recognize and measure an anomaly is limited by the "signal-to-noise" ratio:

Signal-to-noise = Hs / (Ns + Ng)

Because Hs and Ng are proportional to the primary field strength Hp, and Ns, in frequency-domain systems, usually contains elements proportional to Hp, there is little to be gained by increasing the primary field power. In time domain systems Ns is not greatly affected by Hp, so extra power does result in increased signal-to-noise. Attempts to increase the signal-to-noise are sometimes made by increasing the distance between the transmitter and receiver. This results in roughly the same Hs and Ng but often a lower system noise Ns. However the longer bird required to achieve this is more prone to flex, and thus may actually display increased system noise Ns. In addition, the larger bird is heavier and more difficult to handle and thus may reduce survey productivity, increasing cost. In conductive areas Ng may be higher, thereby offsetting any advantage of lower Ns.

6.2. PENETRATION:

The penetration of an AEM system is its effective depth of exploration. Commonly, this is taken to include the elevation of the system above ground, as this is also affected by local environment and flying conditions.

In general, systems with large transmitter-receiver coil separation, usually referred to as Tx-Rx, have greater penetration than those with small separations. Penetration is closely related to signal-to-noise, as the system that produces a larger anomaly from a given conductor can, of course, look further into the ground. Penetration is usually defined as the maximum depth at which a large vertical sheet will produce a recognizable anomaly of at least twice the amplitude of the system noise.

6.3. DISCRIMINATION:

The discrimination of an AEM system is the ability of the system to differentiate between conductors of different physical properties or geometric shapes. Discrimination, particularly between flat lying surficial conductors and steeply dipping conductors, is vitally important. Good discrimination can be achieved in HEM systems by using several frequencies and both co-axial and co-planar coil pairs.

6.4. RESOLUTION:

Resolution refers to the ability of an AEM system to recognize and separate the interfering effects of nearby conductors. A system that does this well also produces sharp anomalies over isolated or discrete conductors. Resolution generally increases with decreasing flight elevation and coil separation. Typically the HEM systems have better resolution than the fixed wing time domain systems.

6.5. CONDUCTIVITY-WIDTH APERTURE:

All AEM systems are, to some extent, aperture-limited. Below a certain "response factor", which includes the conductivity and dimensions of the conductor, the anomaly produced by the system will be below the recognition level. At the upper end of the response factor, some systems are limited and others are not. The ones that are not limited sometimes cease to be multi-channel systems and lose their discrimination. Time domain systems like INPUT are aperture limited.

6.6. LATERAL COVERAGE:

In addition to penetration, the lateral coverage of an AEM system is important because it dictates, to some extent, the maximum distance between survey lines, which in turn affects the cost of exploration. Alternatively, at a given survey line spacing, a system with good lateral coverage will have a better chance of detecting a conductor that lies between two survey lines. Like penetration, lateral coverage generally increases with increasing coil separation.

7. APPLICATIONS

Some of the major applications of Air borne Electromagnetic survey are:

7.1 STRUCTURAL MAPPING

Airborne magnetic surveying has been used extensively in oil exploration for mapping bedrock structure and depth to basement. Gravity and gravity gradiometry are also commonly used for a variety of oil and gas exploration and development applications.

7.2 MINERAL EXPLORATION

Airborne electromagnetic surveying originally grew out of the field of mineral exploration, in particular the search for conductive metallic sulphides and oxides in resistive geology. These applications are still common, but have been extended to almost every kind of economic mineral deposit.

7.3 ENVIRONMENTAL

Environmental monitoring is a relatively new application of airborne geophysics, requiring precise measurement of the geophysical parameters (conductivity, etc.) and high resolution of anomaly size and depth. While EM is used most often for mapping groundwater location and salinity, magnetic surveys have been used to detect buried metallic objects and pipelines, and gamma-ray spectrometry can be used to detect man-made radioactivity.

7.4 OIL AND GAS EXPLORATION

Airborne geophysics has been used for petroleum exploration for many years. Recent years have seen the application of airborne electromagnetic data to hydrocarbon exploration and development in the areas of oil sands exploration and engineering, groundwater management, shallow drilling hazards and Quaternary gas exploration.

7.5 FRESHWATER MAPPING

Airborne electromagnetic surveys can be used to detect freshwater-filled fractures in rock, or sandy paleochannels in the overburden. Water-filled fractures are generally more conductive than the surrounding rock.

8. SURVEY DATA PRESENTATION

In addition to a digital data file, the results of an AEM survey, the data, is presented in a variety of formats. Some contractors only present the EM anomaly locations plotted on the flight path maps, together with a coding indicating anomaly strengths and certain parameters derived by computer-modeling the anomaly sources as vertical sheets. Before the advent of personal computers with their interactive display capabilities, stacked profiles of the EM, altimeter, magnetic, and sometimes, spheric noise data used to be a common form of data presentation. However, because handling the large amount of paper involved was always an onerous task and most explorationists can now display profiles, using their computer, directly from the digital data base, it is no longer common to produce hard copy profile displays.

Typically contractors present EM data in two principle formats:

1) As profile maps showing the in-phase and quadrature components of complimentary co-axial and co-planar frequency pairs plotted as coloured profiles on the flight path. This map also shows the locations of significant EM anomalies displayed using an icon code to indicate the calculated conductivity-thickness product of the source assuming that it is a vertical sheet. The process of "picking" and modeling these anomalies will be described in more detail in the interpretation section.

2) As a coloured map of the apparent resistivity with embedded contours calculated from the coplanar or coaxial EM data. This map shows the apparent ground resistivity assuming the ground to be of uniform conductivity both laterally and vertically. These maps are helpful in outlining conductive overburden and showing discrete bed-rock conductors. Actual values of resistivity bear little relation to the true resistivities of the overburden or bed-rock features.

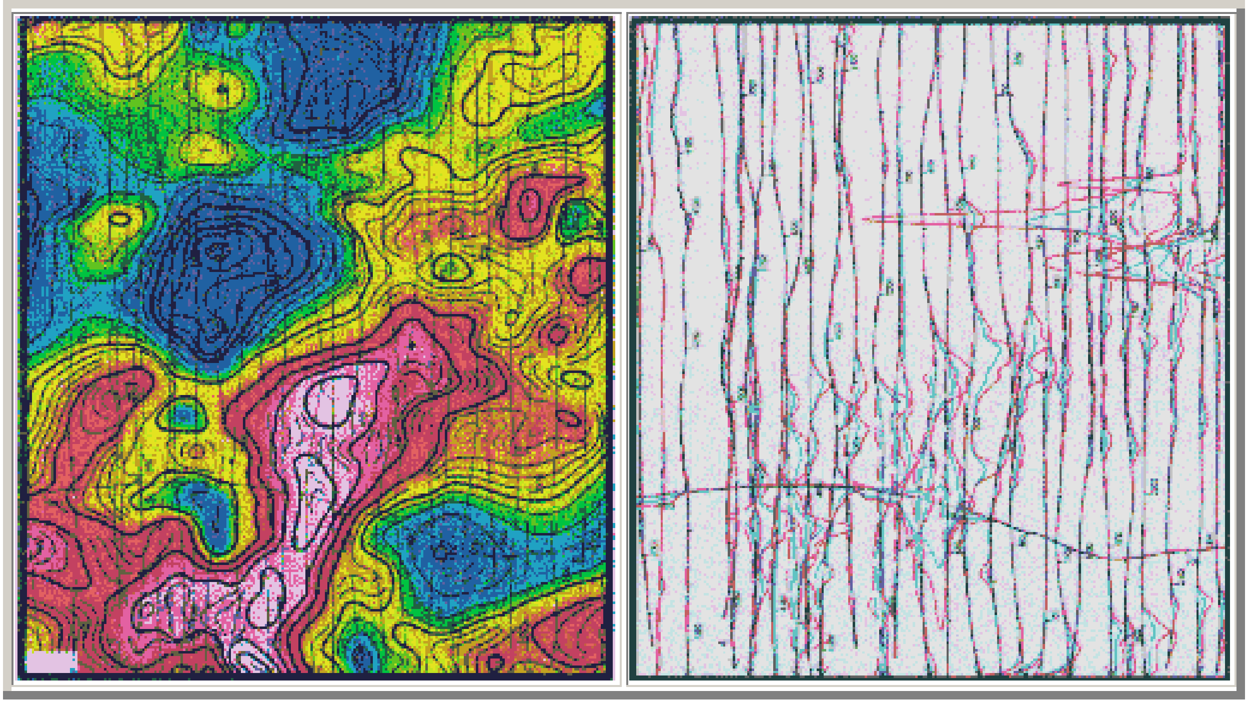


Figure illustrates a typical suite of final maps of both the magnetic data and the EM data, including the interpretation map, that survey contractor would deliver after the completion of a combination Magnetic-HEM survey operations and the required compilation and interpretation phases of data analysis. Moving the mouse over the pictures will allow you to see different presentations. "Clicking" on the visible version will produce an enlargement.

9. ADVANTAGES OF AEM

The advantages of airborne surveying include the rapid acquisition of multiple datasets, access over difficult terrain and non-invasive ground investigations which allow the localization of potential target areas for ground surveying, and are therefore highly cost effective.

10. CONCLUSION

Air borne electromagnetics is a very useful method for surveying large areas in order to support hydrogeological investigations. Due to the dependency of Geophysical parameter electrical conductivity from both the mineralization of ground water and the clay content, information about water quality and aquifer characteristics, respectively, can be derived from AEM data. The results however are sometimes ambiguous- as a consequence, additional information e.g: from drillings, are requires for a solid hydorgeological interpretation of the AEM data.

12. REFERENCES

http://www.geoexplo.com/airborne_survey_workshop_EM.html

Book: Groundwater Geophysics- Kirsch (Ed.)

http://www.geoexplo.com/airborne_survey_workshop_EM.html

http://www.fugroairborne.com/services/geophysicalservices/byapplication/