KING SAUD UNIVERSITY

Faculty of Sciences

Dept. of Geology & Geophysics

SEISMIC EXPLORATION
GPH 221

2012 / 2013 (1433/1434)

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Seismic Exploration (GPH 221)

When & Where

Saturday	10-10:50	G B 80/1
Monday		

Lessons start on 26/01/2013 (14/3/1434)

Lessons end on 20/05/2013 (10/7/1434)

Second semester midterm holiday 21 – 29 /3/2013 (9-17/5/1434)

Midterm Exam: Monday, 01 April, 2013 (20/5/1434)

King Saud University
College of Science
Geology and Geophysics Department.

جامعة الملك سعود
كلية العلوم
قسم الجيولوجيا والجيوفيزياء

Academic Year 1433- 1434H (2012 – 2013)
Second Semester

Seismic Exploration (GPH 221)

Lecture's Time: **Saturday & Monday: 10:00 -10:50** Lecture's Room: **G B 81/1**

Instructor: Dr. Mahmoud ELWAHEIDI

Office Hours: Saturday, Sunday & Monday: 11:00 am -12:00 am

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I. COURSE OUTLINES

Activity	No of Weeks	No. of hours
1. GENERAL INTRODUCTION • Historical development s • Importance of seismic methods • Seismic methods fields of applications 2. SEISMIC METHODS: REFLECTION AND REFRACTION • Elastic constants • Types of Seismic Waves and their propagation • Signal to noise ratio (SNR) • Basic theoretical principles	**3**	**6**
3. SEISMOLOGY AND GEOLOGY • Wave Interactions with Boundaries • Velocities and Rock Properties • Seismic Velocities of Common Earth Materials	**1**	**2**

4. REFRACTION METHOD • Field Procedures • First Arrivals • Data Corrections • Determining Earth Structure from Travel Times	**3**	**6**
5. REFLECTION METHOD • Field Procedures • Data Processing • Interpretation: Qualitative & Quantitative (isochrones and isopach maps) • Seismic stratigraphy • Case studies	**5**	**10**
6. SEISMIC INSTRUMENTATION 7. SEISMIC SOURCES	**3**	**6**

II. GRADING SYSTEM

Assessment	Assessment task	Week due	Proportion of Final Assessment
1	**Lab (12 sessions) & field trips (2)**		20 %
2	**Mid-term exam**	7 (**Monday, 01 April, 2013**)	25%
3	**Attendance, Quizzes & Assignments**		15 %
4	**Final exam**		40 %

III. TEXT BOOKS- REFERENCES

- Lectures' notes .
- J.M. Reynolds, 2011, An Introduction to Applied and Environmental Geophysics
- Lowrie, W., 1997. Fundamental of geophysics. Cambridge University Press.
- Telford, W., Geldart, L., and Sheriff, R., 1990. Applied geophysics, second edition. Cambridge University Press.
- http://utam.geophys.utah.edu/stanford/node2.html

GENERAL I N T R O D U C T I O N

Applied geophysics is a branch of Geology that has the characteristics of an Applied Science.

•**The General Scope of Applied Geophysics:** dealing with the physical fields of the earth which are important for determining the geological structures of the crust and the upper part of the mantle, mineral explorations and other geological activities (Engineering geology, Environmental Geology, hydrogeology etc.).

Classifications of Geophysical Methods

1. According to the *source of signal*:

- **Natural Fields:** GRAVITY, MAGNETIC, GEOTHERMAL, RADIOACTIVE, TELLURIC
- **Artificial Fields:** ELECTRIC, SEISMIC, ELECTROMAGNETIC

2. According to the *physical properties of the investigated target*:

- **Static Methods:** The distortions in the static field are measured, and we try to generate a model that best represents the phenomena that caused these. (Magnetic and gravimetric methods).
- **Dynamic Methods:** A signal is introduced in the earth and the response signal is measured at various points in terms of amplitude or arrival time. For example: the use of the arrival time concept in seismic methods, and the frequency or phase difference in electromagnetic methods.

- **Relaxation methods:** the measured time is the time necessary for the investigated media to return to its initial status (before the application of any signal). (Induced Polarization method)
- **Methods based on the Integration of Effect: the measured signals** represent the statistical mean over an area or a volume (Radioactive methods).

Interpretation of Geophysical Data

- **Direct Method:** The effect of an anomalous body on a certain physical field is determined. In this case the following parameters of the body are **KNOWN:** Dimension, Form and Depth. The solution of this method is always unique.

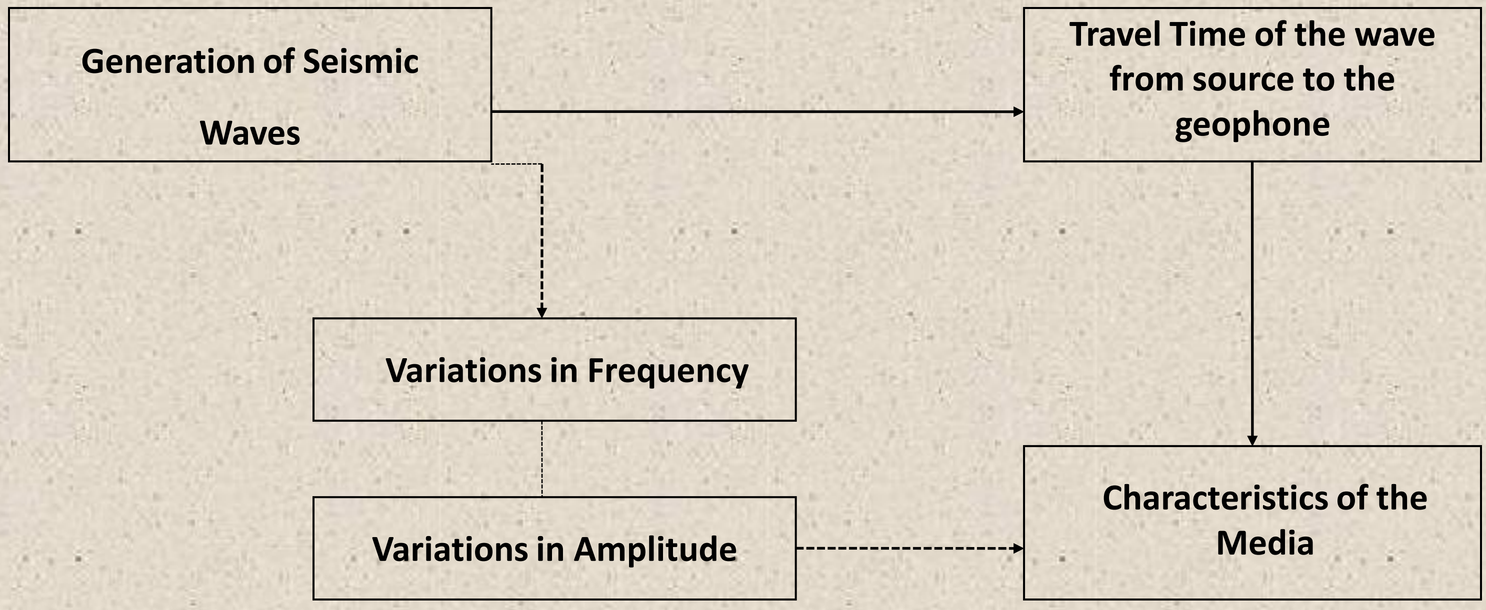
- **Inverse Method:** The characteristics of an anomalous body, responsible for a specific anomaly in the physical field is determined. This method is almost always ambiguous. To overcome this problem it is necessary to utilize several geophysical methods in combination; otherwise direct geological data are necessary (borehole data).

SEISMIC METHODS

- **Seismic methods are considered as the most important geophysical prospecting technique.**
- **Major advantages are:**
  - ✓ Great accuracy
  - ✓ High resolution
  - ✓ Great Depth of penetration
- **Major Fields of applications :**
  - ✓Oil Explorations
  - ✓Ground water Explorations
  - ✓Civil engineering

- **A seismic method is based on the same principles of seismology (seismic of earthquakes). However, it differs in the following aspects :**
  - The Source and the Location of the energy are controlled and can be chosen according to the scope of investigation.
  - The distance between the source and the receivers is relatively small.
  - Field procedure is based on taking measurements along profiles.

- **The fundamental Concept of Seismic Methods is given by the following drawing:**



➢ Historical Development

1845 - First experiments with artificial earthquakes in order to measure velocity of seismic waves

1899 -The development of the theory of reflection and refraction of seismic waves at boundaries.

1913 - Utilizing of reflection method for determining the depth of water.

1914-1918 - During the first world war, enormous development in both: instrumentation and techniques of interpretation.

1919- The born of the refraction seismic method

1922 -1925- Development of instruments for the seismic reflection techniques.

1922- Development of instruments for the seismic refraction. Those characterized by their low sensibility and their mechanical nature.

1924- First success of the refraction technique resulted in discovering the Orchard Salt in Texas.

1927- First commercial success of the reflection method.

1941-The start of using instruments with 24 channels.

1956- The development of the Common Depth Point Technique in seismic reflection method.

1956-1960- The development of 2D and 3D modeling of seismic data.

1960-1965- The revolution in the field of computers resulted in enormous development in the techniques of measurements, processing and interpretation of seismic data.

1972 - A system based on 200 channels is developed by Shell Oil Company.

Fundamental Concepts

- Seismic waves produce deformations in the medium in which they pass. These deformations can be described based on the assumption of **elastic** and **homogeneous** medium.

For the utilization of such deformations for the scope of geological explorations, it is necessary to recognize the following physical concepts which govern these events.

- **Generation**
- **Transmission**
- **Absorption**
- **Attenuation**

→ **IN the materials of the earth's crust**

- **Reflection**
- **Refraction**
- **Diffraction**

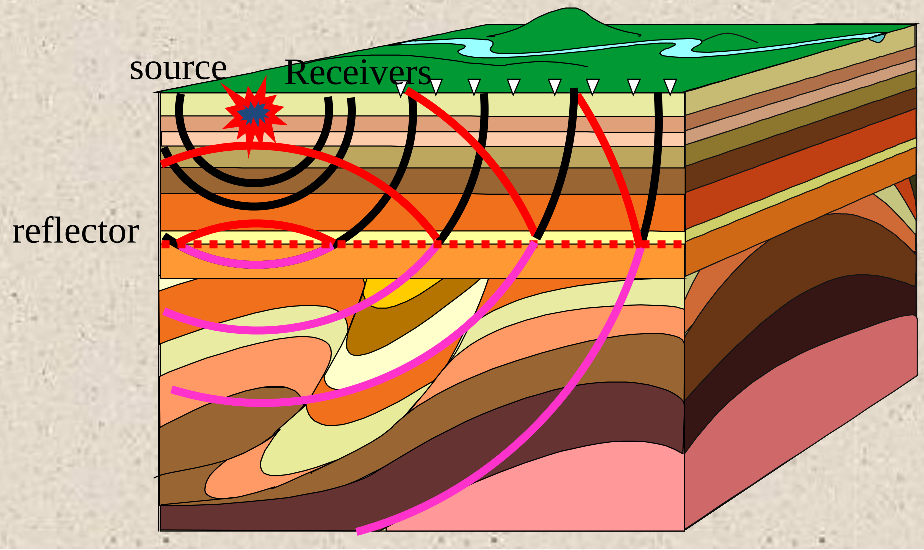
→ **AT the lithological discontinuity**

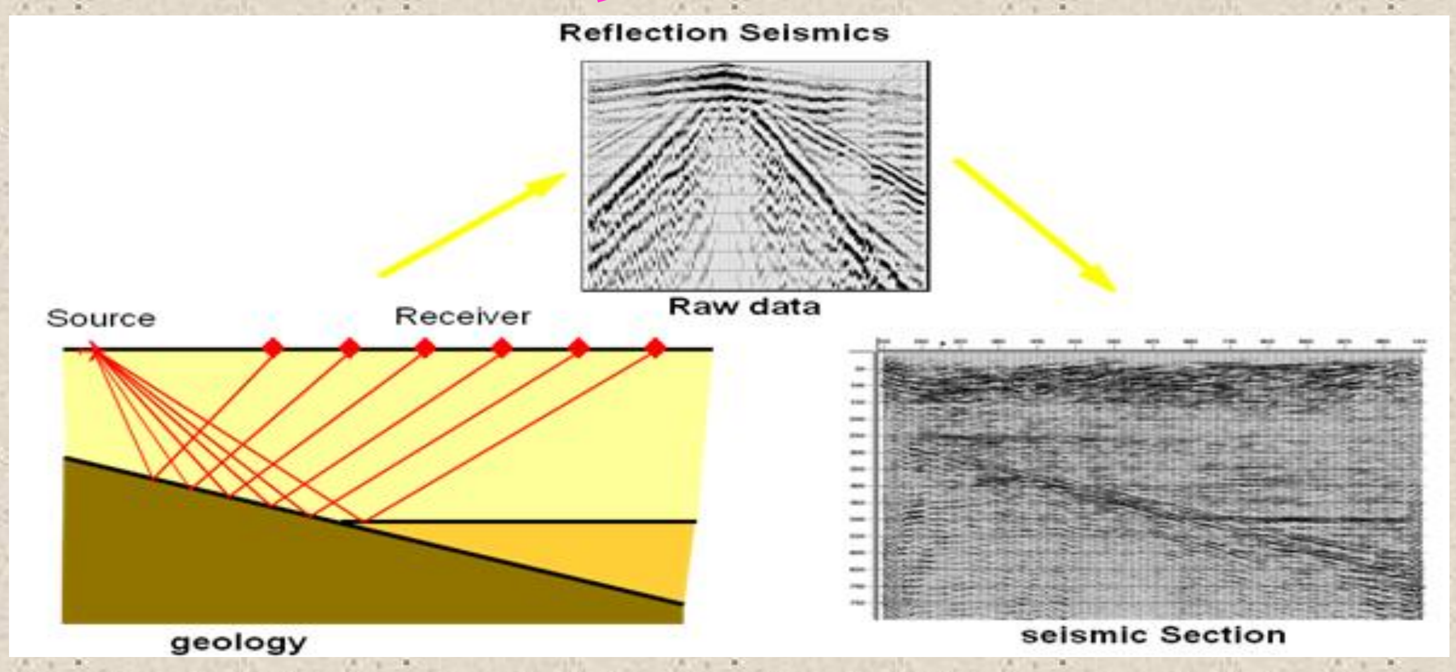
- The **velocity of seismic waves** which propagate in solids depends on **the elastic property** and **density** of the medium in which they pass. This relation can be described in terms of two types of forces:
    - Stress = the applied force
    - Strain = the derived deformation resulted from the application of a stress.
- **The theory of Elastic Waves states** that: in case of an ideal elastic medium, once the force causing the deformation is over, the medium restores its initial conditions (form and volume).

SEISMIC EXPLORATION

I. Introduction

- In seismic surveying, seismic waves are created by controlled sources and propagate through the subsurface.
- These waves will return to the surface after reflection or refraction at geological boundaries.
- Instruments distributed along the surface (called geophones/hydrophones) detect the ground motion caused by these returning waves and measure the arrival times of the waves at different ranges from the source.
- Seismic surveying provides a detailed picture of subsurface geology.
- Artificial sources, such as explosions, are used in the seismic methods.
- Location, timing and source characteristics are, unlike earthquakes, under the direct control of the geophysicists.


source
Receivers
reflector


Reflection Seismics
Raw data
Source
Receiver
geology
seismic Section

II. Stress and Strain

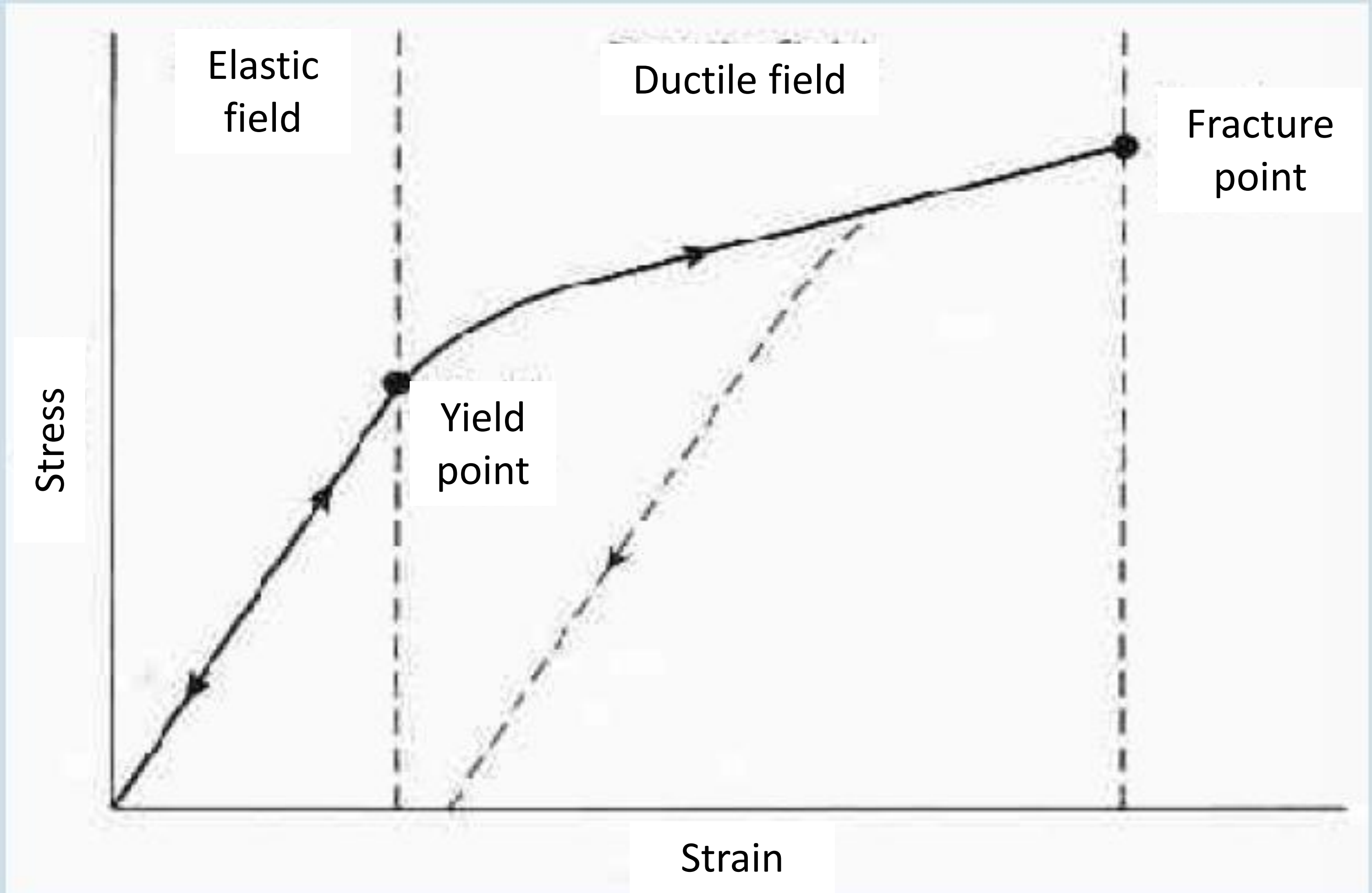
If an external force F is applied across an area A of a surface of a body, the ratio of the applied force to the area (F/A) is known as stress.

Stress can be resolved into two components:

- one at right angle to the surface (normal or dilatational stress)
- one in the plane of the surface (shear stress)

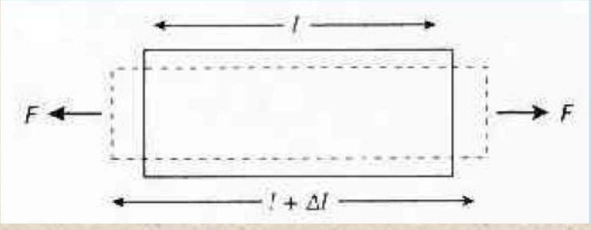
A body subjected to stress undergoes a *change of shape and/or size*, this change is known as **strain**.

- According to **Hooke's law**, stress and strain are linearly dependent and the body behaves elastically until the yielding point is reached.
- This elastic strain is reversible so that removal of stress leads to removal of strain.
- At stresses beyond the yield point, the body behaves in a plastic or ductile manner, and permanent damage results. If further stress is applied, the body is strained until it fractures.


Elastic field
Ductile field
Fracture point
Yield point
Stress
Strain

The linear relationship between stress and strain in the elastic field is specified by its various **elastic moduli.** These are:

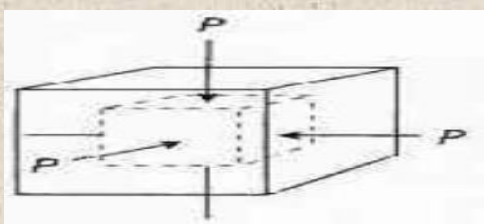
- ❑ Young modulus, **E:**



$$E = \frac{\text{longitudinal stress } F/A}{\text{longitudinal strain } \Delta l/l}$$

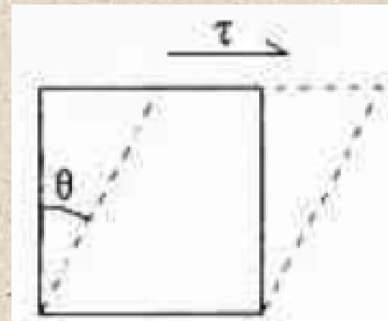
- ❑ Bulk modulus, **K**

Known also as the *incompressibility* of the medium. It describes the ratio of the pressure applied to a volume to the amount of change in volume.



$$K = \frac{\text{volume stress } P}{\text{volume strain } \Delta v/v}$$

- Shear modulus, μ: (know also as Lame's second parameter)



$$\mu = \frac{\text{shear stress } \tau}{\text{shear strain } \tan\theta}$$

For gases & fluids, $\mu = 0$. This implies that gases and fluids don't allow propagation of S-waves.

- Lame's first parameter, λ

- Poisson's ratio, δ:

$$\delta = \frac{\textbf{change in width/width}}{\textbf{Change in length/length}}$$

The Poisson's ratio is also given by following in terms of Lame's parameters: $\delta = \dfrac{\lambda}{2(\lambda + \mu)}$

It's values range from 0.05 (for very hard rocks) to 0.45 (for loose sediments). It has a maximum value of 0.5.

III. Types of Seismic Waves

Seismic waves are *parcels of elastic strain energy that propagate from a seismic source such as an earthquake or an explosion*.

The strains associated with the passage of a seismic pulse may be assumed to be elastic (except in the vicinity of the source).

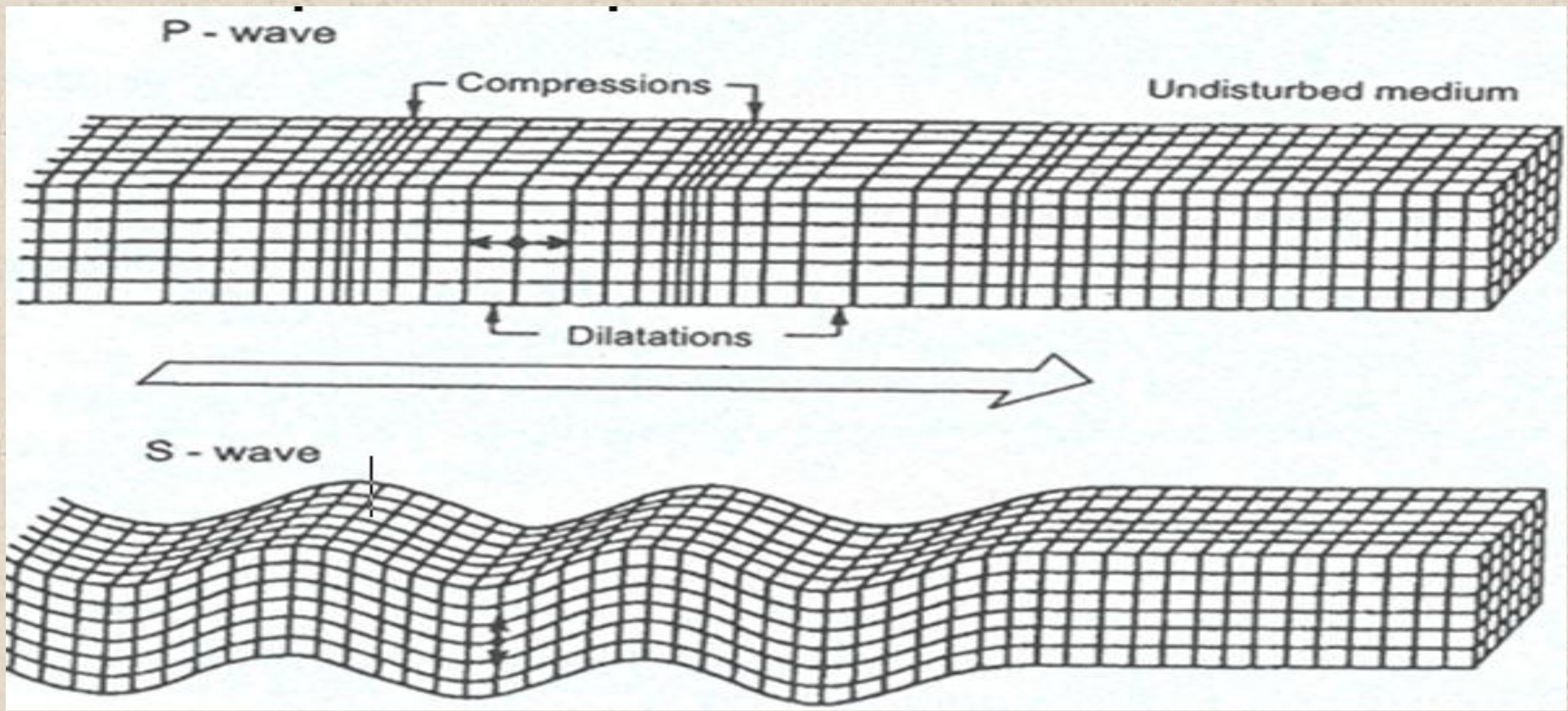
The propagation of seismic pulses is determined by the ***elastic moduli*** and ***densities*** of the materials through which they pass.

There are two groups of seismic waves: **body waves** and **surface waves**.

III. 1. Body Waves: P-waves and S-waves

❑ P-wave (longitudinal, primary or compressional wave). Material particles oscillate about a fixed point in the direction of wave propagation by compressional and dilatational strain.

❑ S- wave (transverse, secondary or shear wave). Particle motion is at right angles to the direction of wave propagation and occurs by pure strain.



III. 2. Surface waves:

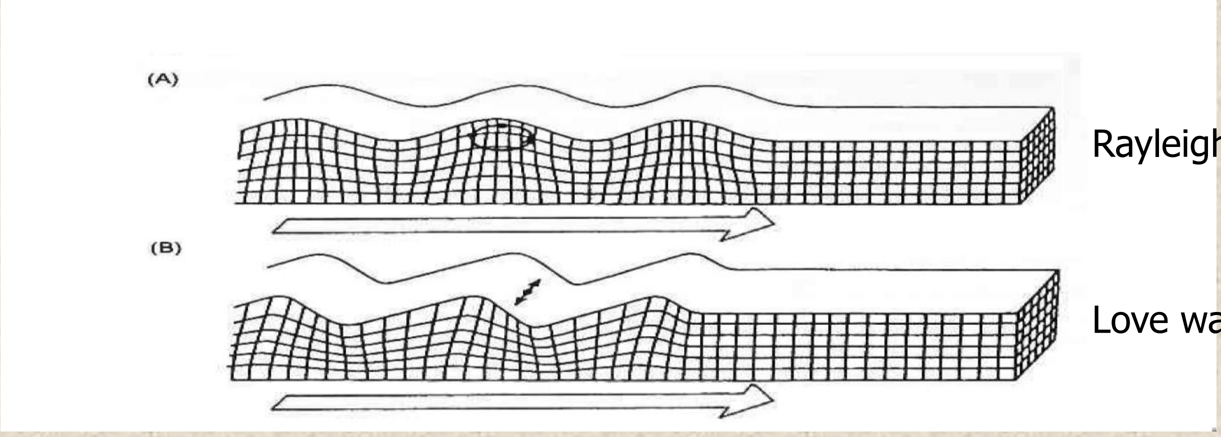
Waves that do not penetrate into the subsurface media are known as surface waves, these include:

❑ Rayleigh waves:

Travel along the free surface of the earth with amplitudes that decrease exponentially with depth; particle motion is in an elliptical sense in a vertical plane with respect to the surface. Rayleigh waves travel only through a solid medium.

❑ Love waves:

Occur only where a medium with a low S-wave velocity overlies a layer with a higher S-wave velocity; particle motion is at right angles to the wave propagation but parallel to the surface.



III. Seismic Waves Velocities of Earth Materials

III.1. Propagation of Seismic Waves

In homogeneous isotropic media, the velocities of the P and S waves through a medium is given by equations:

$$V_p = \left(\frac{k + 4/3\,\mu}{\rho} \right)^{1/2}$$

$$V_s = \left(\frac{\mu}{\rho} \right)^{1/2}$$

Where:

K and μ are known as elastic moduli.

Vp: velocity of P-wave

Vs: velocity of S-wave

ρ: density of the medium

P and S-velocities

P-velocity

$$V_P = \sqrt{\frac{\kappa + \frac{4}{3}\mu}{\rho}}$$

change of shape and volume

S-velocity

$$V_S = \sqrt{\frac{\mu}{\rho}}$$

change of shape only

For liquids and gases $\mu = 0$, therefore

➔ $V_S = 0$ and V_P is reduced in liquids and gases

➔ Highly fractured or porous rocks have significantly reduced V_P

The bulk modulus, κ is always positive, therefore $V_S < V_P$ always

P-waves are the most important for controlled source seismology

- They arrive first making them easier to observe
- It is difficult to create a shear source, explosions are compressional

Young's modulus

E

Poisson's ratio

σ

Lamé's constants

$$\lambda = \frac{\sigma E}{(1+\sigma)(1-2\sigma)}$$

$$\mu = \frac{E}{2(1+\sigma)}$$

***P*–wave velocity**

$$\alpha = \sqrt{\frac{\lambda + 2\mu}{\rho}}$$

***S*–wave velocity**

$$\beta = \sqrt{\frac{\mu}{\rho}}$$

Ratio of *P*–wave velocity to *S*–wave velocity

$$\frac{\beta}{\alpha} = \sqrt{\frac{1-2\sigma}{2(1-\sigma)}}$$

III.2. Velocity as a function of rock type:

In principle, the velocity of P-wave (Vp) and S- wave (Vs) depend on the lithological characteristics of the rocks. Other important factors include: deposition environment, origin and evolution of rocks.

For <u>igneous and metamorphic rocks</u>: porosity is low →Vp *depends mainly on the elastic properties*

For <u>sedimentary rocks</u>: porosity is medium to high →Vp *depends on the porosity and the nature of the materials in the pores*.

	v_p (km s^{-1})
Unconsolidated materials	
Sand (dry)	0.2–1.0
Sand (water-saturated)	1.5–2.0
Clay	1.0–2.5
Glacial till (water-saturated)	1.5–2.5
Permafrost	3.5–4.0
Sedimentary rocks	
Sandstones	2.0–6.0
Tertiary sandstone	2.0–2.5
Pennant sandstone (Carboniferous)	4.0–4.5
Cambrian quartzite	5.5–6.0
Limestones	2.0–6.0
Cretaceous chalk	2.0–2.5
Jurassic oolites and bioclastic limestones	3.0–4.0
Carboniferous limestone	5.0–5.5
Dolomites	2.5–6.5
Salt	4.5–5.0
Anhydrite	4.5–6.5
Gypsum	2.0–3.5

	v_p (km s^{-1})
Igneous/Metamorphic rocks	
Granite	5.5–6.0
Gabbro	6.5–7.0
Ultramafic rocks	7.5–8.5
Serpentinite	5.5–6.5
Pore fluids	
Air	0.3
Water	1.4–1.5
Ice	3.4
Petroleum	1.3–1.4
Other materials	
Steel	6.1
Iron	5.8
Aluminium	6.6
Concrete	3.6

As can be seen from the above table, a considerable overlap in the seismic velocities exist → A knowledge of seismic velocity alone is not sufficient to determine rock type.

III.3. Velocity of seismic waves as a function of geological age and depth:

$$v = 1.47\,(z\,T)^{1/6}\ \mathrm{km/s}$$

z: depth in km

T: geological age in millions of years

III.4. Attenuation of seismic waves:

Amplitude and energy of a seismic wave are related by the relation:

$$\mathbf{E = \rho W^2 a^2}$$

where,

ρ is density

W is the angular frequency

E is the energy, and

A is the amplitude

- **Causes of seismic waves attenuation:**
  - The produced energy is constant. This is being distributed over increasing areas
  - Absorption of energy by rocks →Kinetic energy is being transformed into heat
  - Partition of energy at discontinuities

BEHAVIOR OF ELASTIC WAVES AT DISCONTINUITIES

At an interface between two rock layers there is generally a change in propagation velocity resulting from the difference in physical properties of the two layers.

At such an interface, the energy within an incident seismic pulse is partitioned into transmitted and reflected pulses.

The relative amplitudes of the transmitted and reflected pulses depend

on the velocities (v) and densities (ρ) and the angle of incidence.

Elastic Waves

Reflection:
Waves remain in the same media

Refraction:
Waves pass through the second media and change direction

Diffraction:
Waves follow certain paths.

Basic Condition for Generating

Reflection
Refraction

Sharp changes in the elastic properties of the discontinuity.

Diffraction

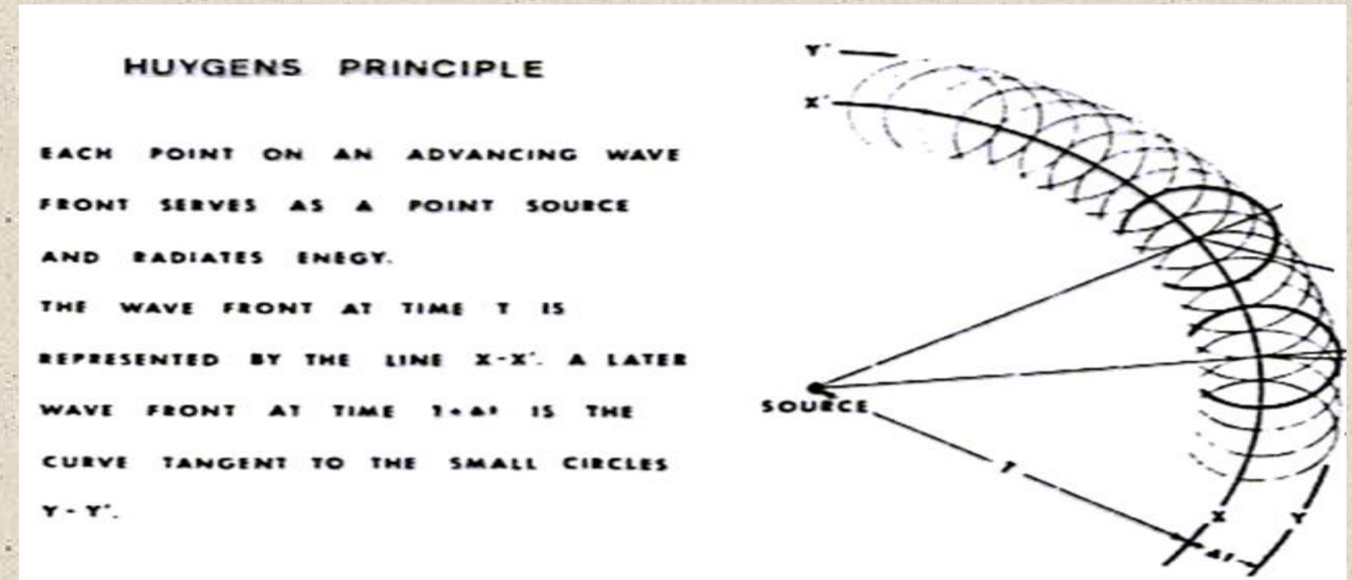
Sharp changes in the morphology of the discontinuity.

BASIC THEORETICAL PRINCIPLES OF SEISMIC METHODS

I. Huygens Principle:

"Every point on the wave front is a source of a new wave that travels out of it in the form of spherical shells."

Seismic rays are used instead of the wave front to describe the wave propagation.



☞**Note:**

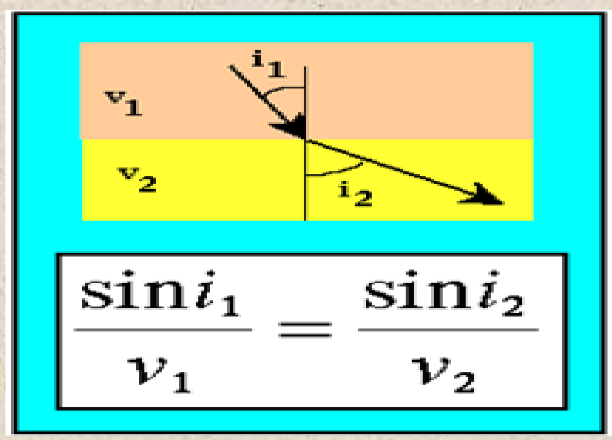
❑ **Raypaths:** *Raypaths are lines that show the direction that the seismic wave is* propagating. For any given wave, there are an infinite set of raypaths that could be used.

❑ **Wavefront:** *Wavefronts connect positions of the seismic wave that are doing the same thing at the same time.*

II. Fermat Principle:

“The wave will travel from the source at a minimum time; *the wave path is not necessary* a straight line.”

III. Snell’s Law :



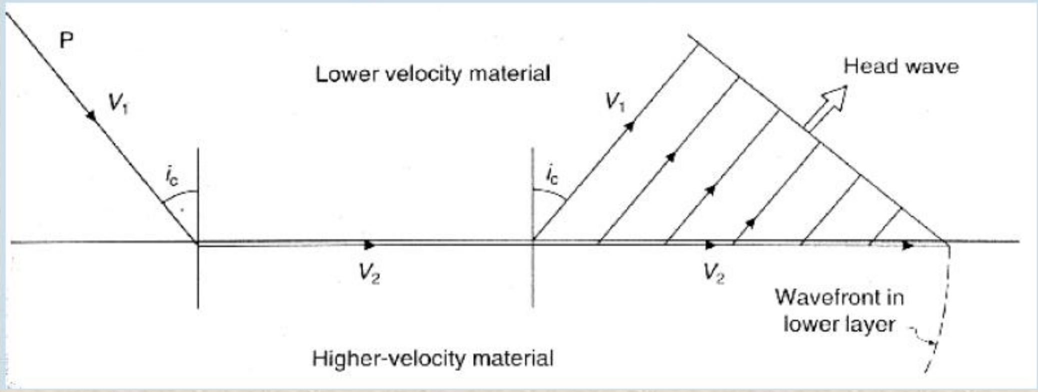
In seismic refraction technique we deal with *Direct* and *refracted* waves. The travel time for the direct waves is calculated simply by dividing the distance by the velocity:

Distance / Velocity

Critical refraction concept:

When the velocity in the upper layer is lower than in the underlying layer, there is a particular angle of incidence, for which the angle of refraction is 90°.

$$\frac{\sin i_c}{\sin 90°} = \frac{v_1}{v_2} \Rightarrow \sin i_c = \frac{v_1}{v_2}$$



☞ **Critical angle** = the incident angle for which the refraction angle is 90°**.**

☞ **<u>Note:</u>**

❑ In refraction technique we use the concept of critical angle. In particular, because the wave reflected at the critical angle simply propagates along the refractor about which we would like to obtain information. The waves produced in this way are called *HEAD WAVES*.

❑ Although a *<u>Head Wave</u>* must travel along a longer path than the direct arrival before it could be recorded at the surface, it travels along the bottom of the layer at a faster speed than the direct arrival. Therefore, Head Waves can be recorded prior to the time of arrival of the direct wave at certain distances.

IV. Law of Reflection:

This law is utilized in the seismic reflection method. It states that "the angle of incidence is equal to the angle of reflection".

In case of I=0, the ratio of the reflected energy of P-wave, E_r, to the incident energy, E_i , is given by:

$$E_r / E_{i|I=0} = \frac{(\rho_2 \mathbf{v}_2 - \rho_1 \mathbf{v}_1)^2}{(\rho_2 \mathbf{v}_2 + \rho_1 \mathbf{v}_1)^2}$$

The square root of the above relationship is called ***Reflection Coefficient, R***. This coefficient gives the ratio between the amplitudes of the incident (A_i) and reflected waves (A_r). It's given by:

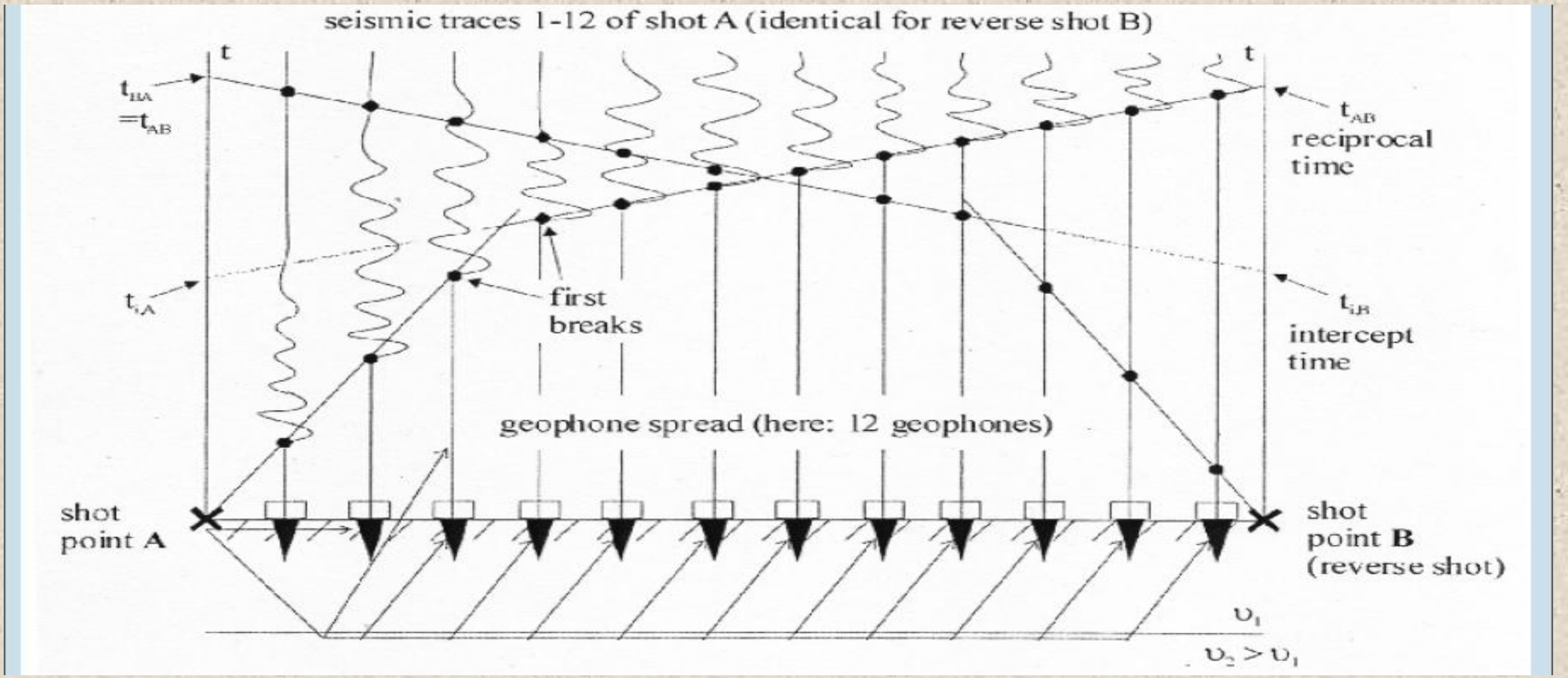
$$R = A_r / A_i = \frac{(\rho_2 \mathbf{v}_2 - \rho_1 \mathbf{v}_1)}{(\rho_2 \mathbf{v}_2 + \rho_1 \mathbf{v}_1)}$$

The Reflection Coefficient, R, shows that the quantity of the reflected energy is based on the contrast between the ***acoustic impedance***, *defined as* *multiplication of velocity by density*, along the opposite side of the reflector surface. In this case, three situations can be recognized:

1. If $\rho 1*V1 < \rho 2*V2$ → no change in the phase of the reflected wave
2. If $\rho 1*V1 > \rho 2*V2$ → shift in the phase of the reflected wave with 180°
3. If $\rho 1*V1 = \rho 2*V2$ → the reflection coefficient is zero: R = 0.

☞ **Note:** Since the variation in the density of different types of rocks is relatively small, the reflection coefficient depends mainly on the contrast in velocities at both sides of the reflecting surface.

Seismic Refraction Method



- The first seismic method utilized in the field of exploration.
- It was used in seismology for determining the Mohorovicic discontinuity, and to discover the nuclei of earth.
- The Description of the geometry of refracted waves is more complex than that of reflected ones.

- The Velocity and thickness of layers are described in terms of TIME. This time is the time required by the refracted wave to travel from the source (at surface) to the receiver (also at the surface), taking in consideration the principle of Fermat.
- The Distance between the Receiver and the Source must be very much larger than the depth of the investigated discontinuity.
- Because of this large distance, the frequencies of interest in the refraction is lower than those in reflection.

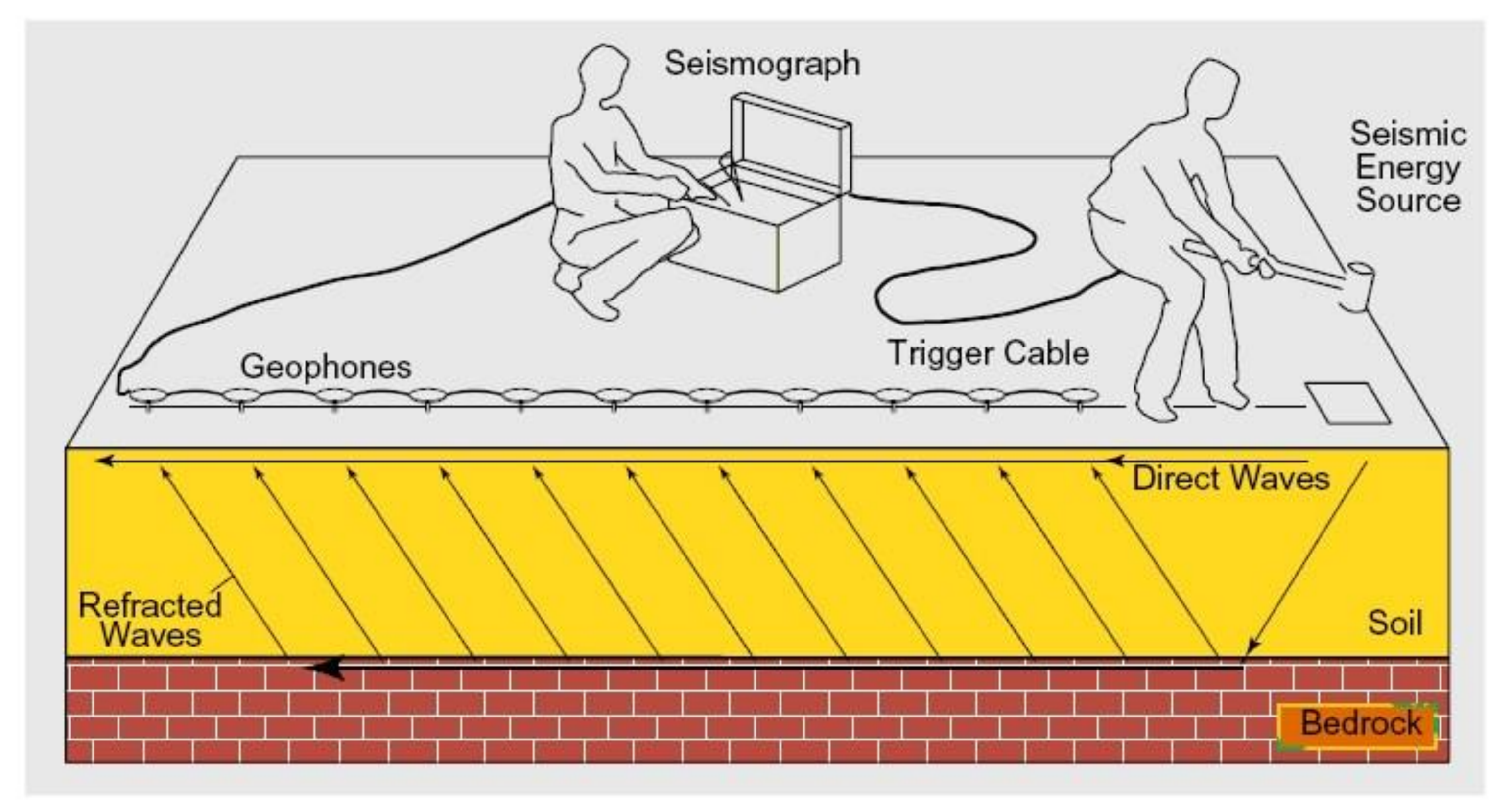
Fields of Applications of Seismic Refraction Method

- Determining lateral extensions of layers.
- Mapping of sedimentary basins.
- Determining the physical properties of the bed rock.
- Detecting buried structures of small dimensions.
- Detecting salt domes.

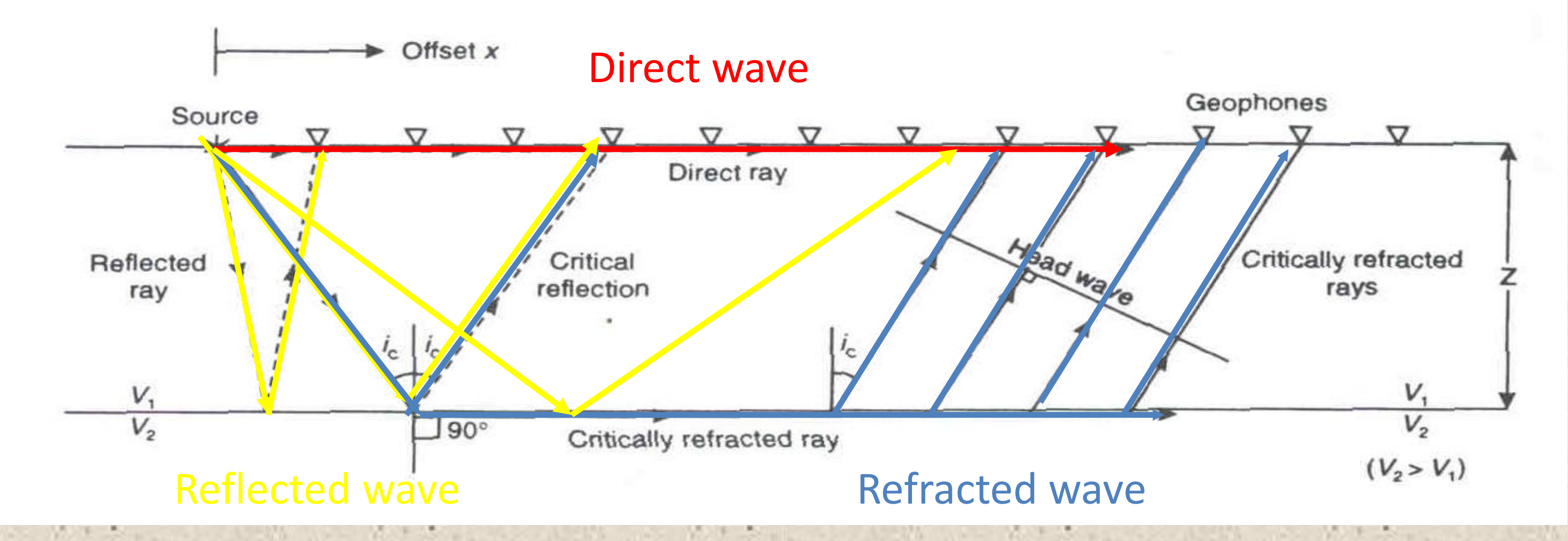
Ideal Conditions of Application

- Coincidence between seismic interfaces and stratigraphical or lithological ones.
- Extended interfaces, homogeneous with small dip angles (less than 15°- 20°).
- Small thickness of layers that are characterized by low velocity.
- Not complicated topography of the investigated area.

Seismic Refraction



❑ Refraction surveys use the process of critical refraction to determine interface depths and layer velocities. Critical refraction requires an increase in velocity with depth. If not, then there is no critical refraction; Hidden layer problem will be faced.



❑ Geophones laid out in a line to record arrivals from a shot. Recording at each geophone is a waveform called a **seismogram.**

❑ Direct signal from shot travels along top of first layer.

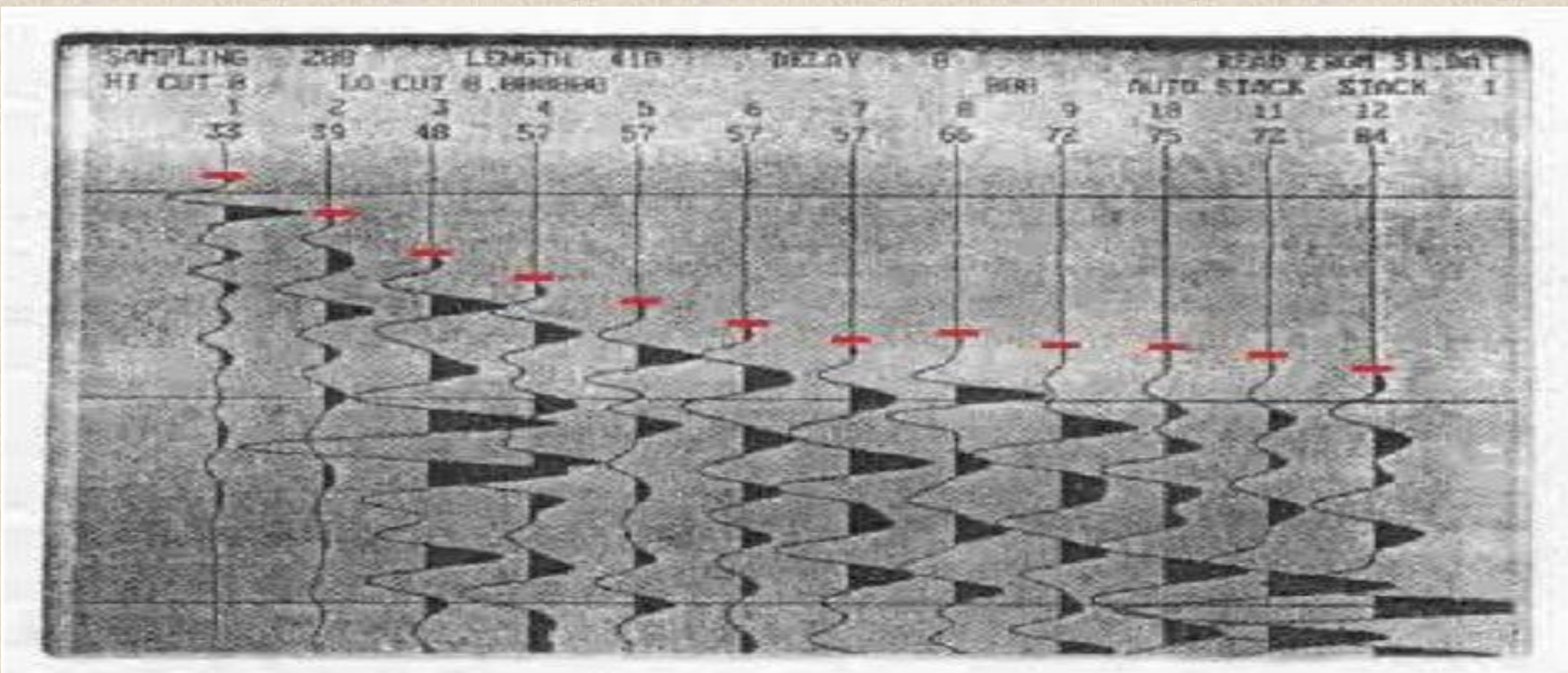
❑ Critical refraction is also recorded at distance beyond which angle of incidence becomes critical.

First Arrival Picking

In most refraction analysis, we only use the travel times of the first arrival on each recorded seismogram. As velocity increases at an interface, critical refraction will become first arrival at some source-receiver offset.

First Break Picking

The beginning of the first seismic wave, the first break, on each seismogram is identified and its arrival time is picked. An example of first break picking process is shown in the figure below:



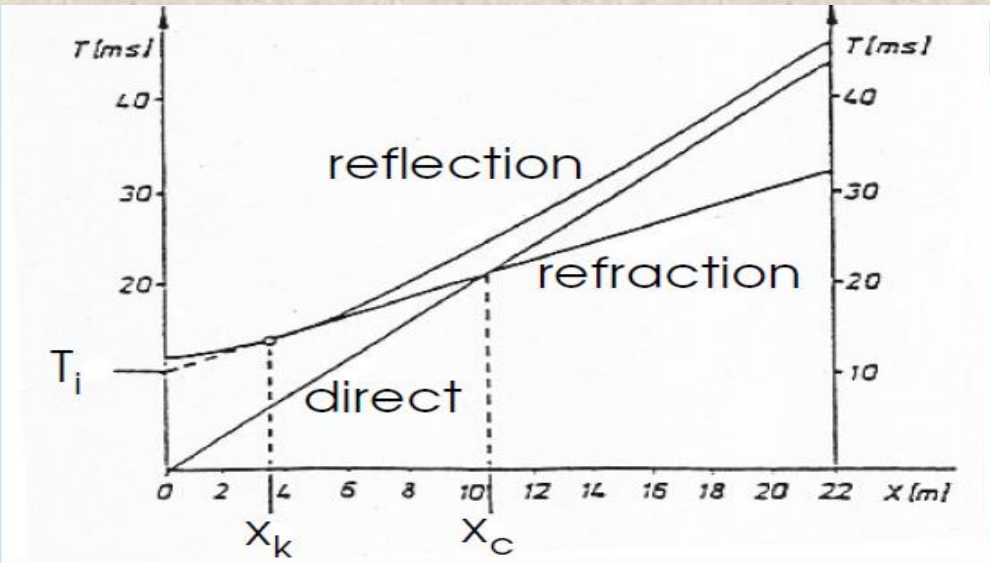
Travel Time Curves

Analysis of seismic refraction data is primarily based on interpretation of critical refraction travel times. Usually we analyze P wave refraction data, but S wave data occasionally recorded.

<u>Plots of seismic arrival times vs. source-receiver offset </u>are called *travel time Curves*.

In the figure below, travel time curves for three arrivals can be noted:

- Direct arrival from source to receiver in top layer
- Critical refraction along top of second layer
- Reflection from top of second layer



Traveltime curves for reflected, refracted and direct waves.

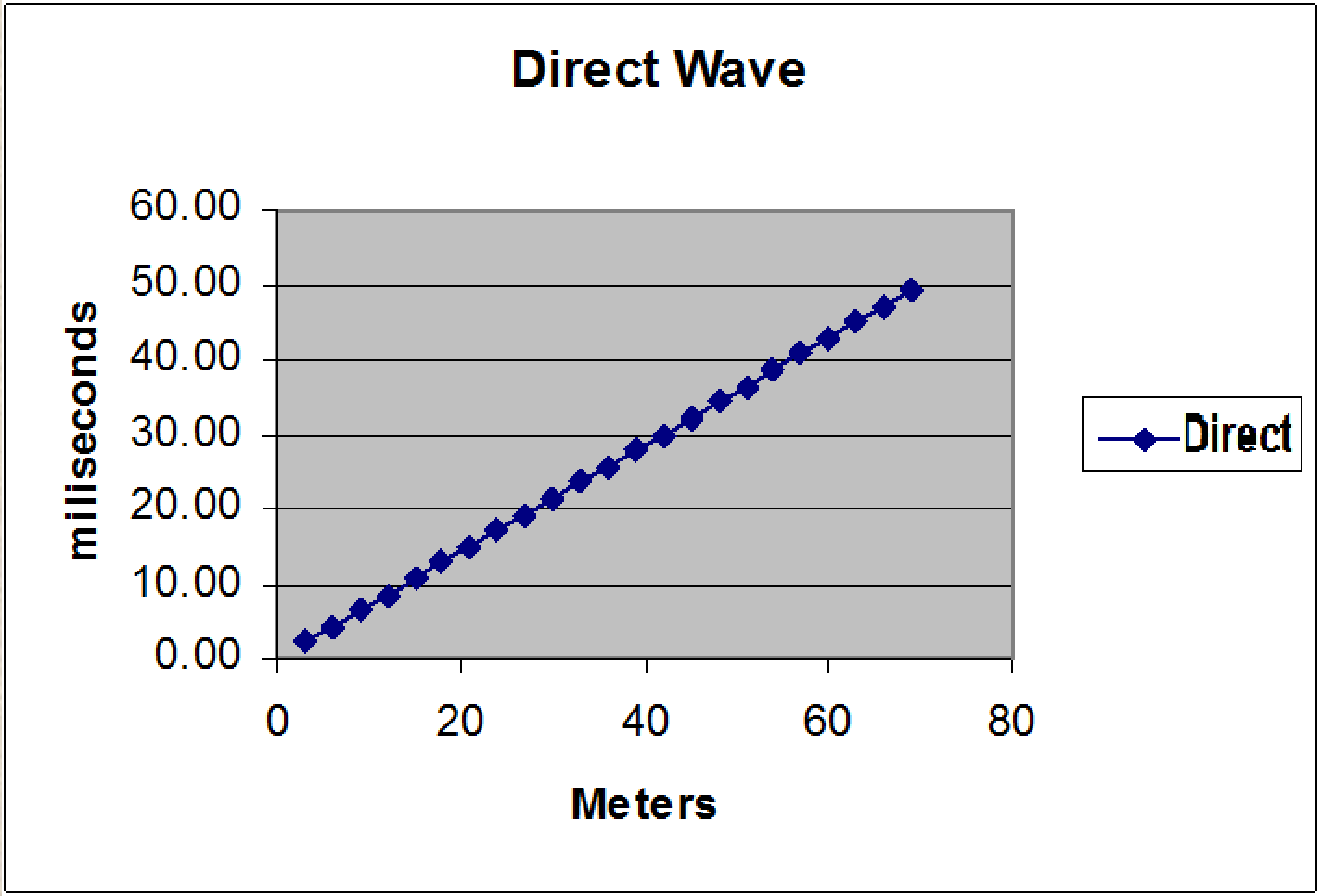
Refraction for $x > x_k$. At a distance x_k called critical distance the reflected arrival is coincident with the first critically refracted arrival and the travel times of the two are identical.

Critical Distance is:
the offset at which critical refraction first appears. In this case:

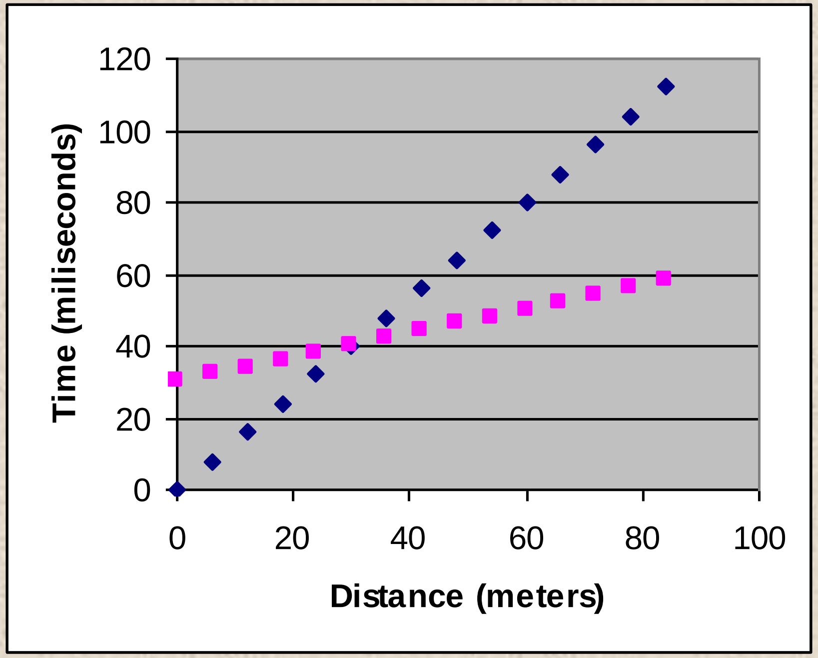
- Critical refraction has same travel time as reflection
- Angle of reflection same as critical angle

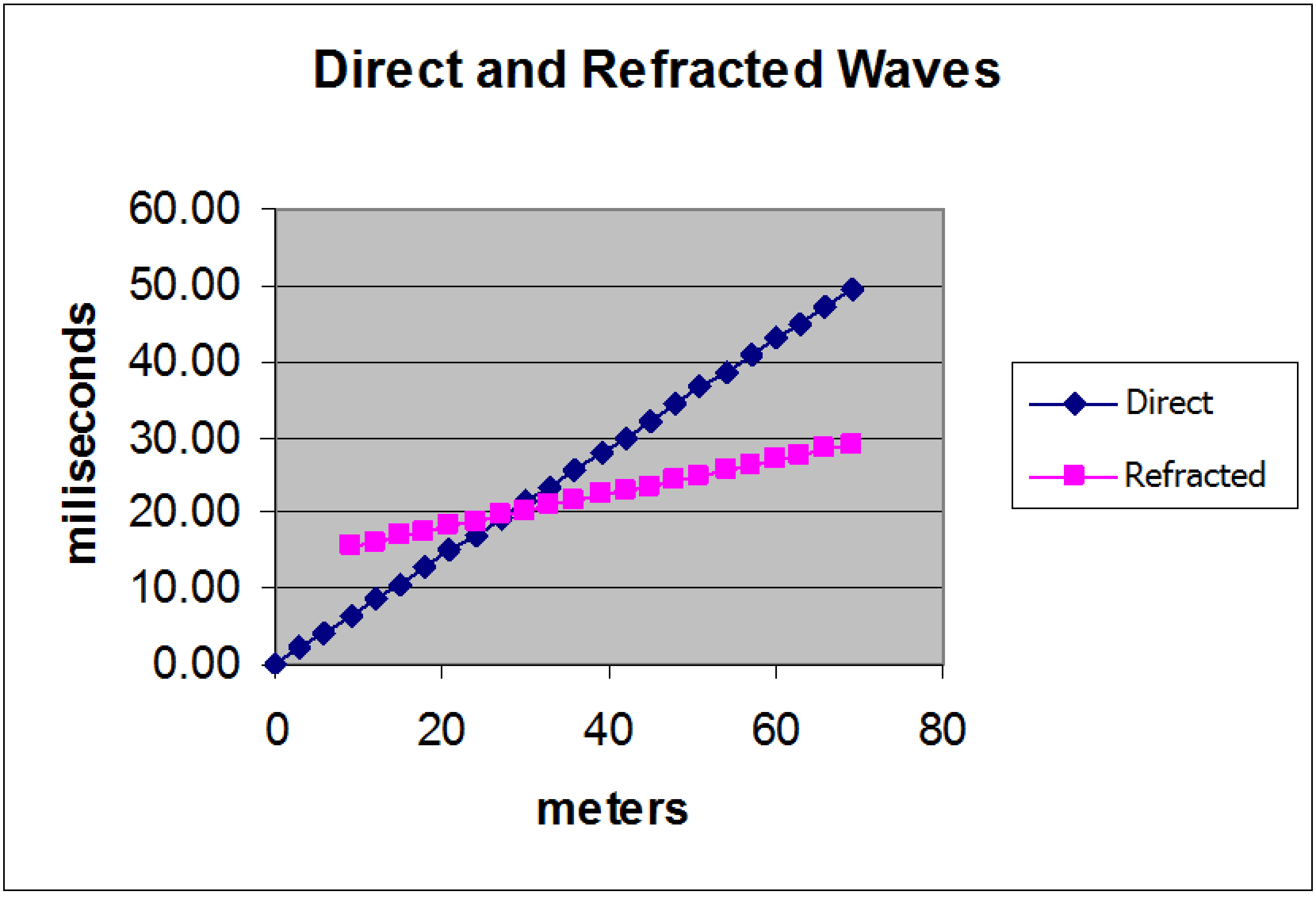
Crossover Distance is
The offset at which critical refraction becomes first arrival.

Plotting Travel Time Curves

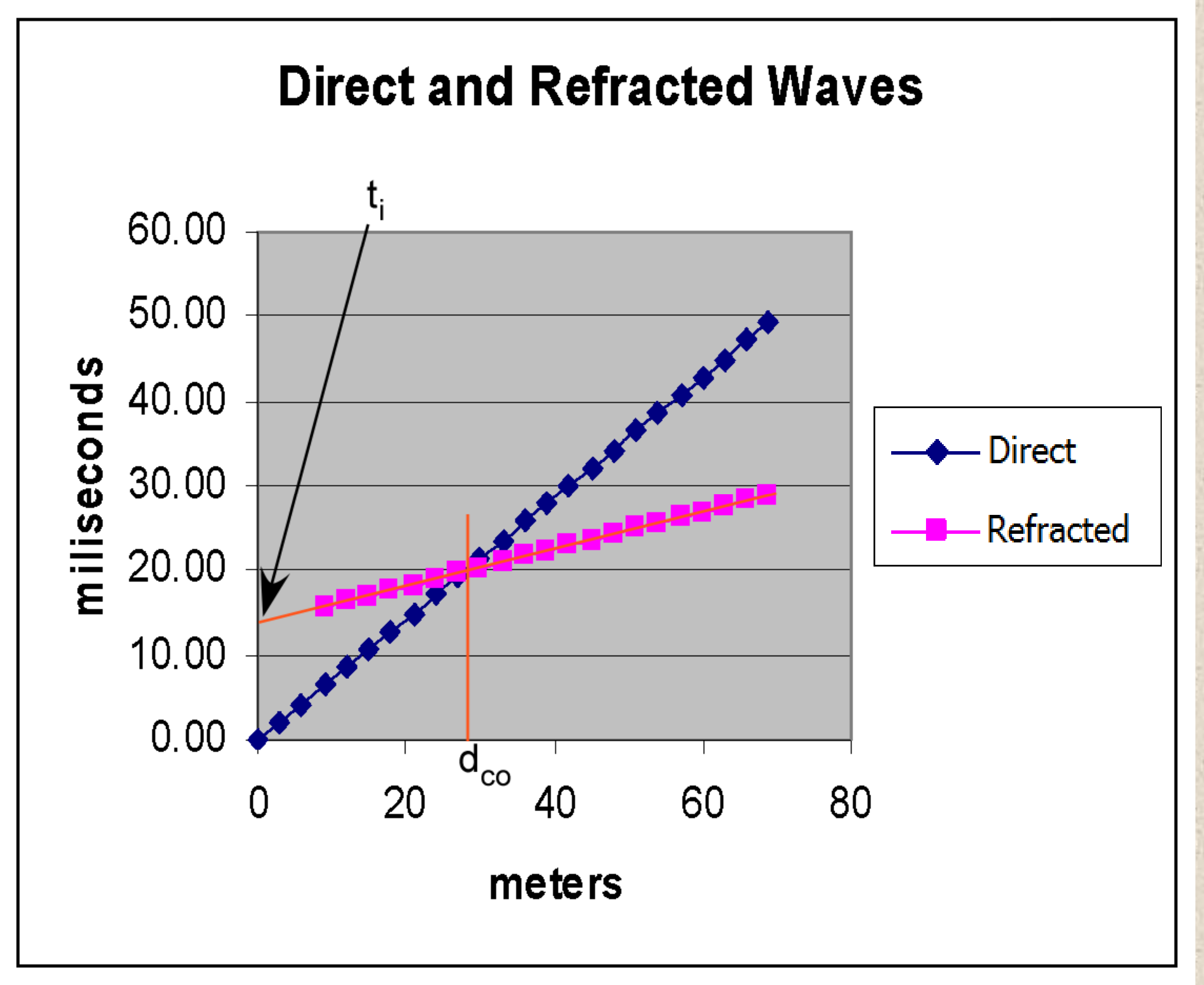


Distance	TimeD	Time R
0	0	30
6	8	32
12	16	34
18	24	36
24	32	38
30	40	40
36	48	42
42	56	44
48	64	46
54	72	48
60	80	50
66	88	52
72	96	54
78	104	56
84	112	58





Direct and Refracted Waves
60.00
50.00
40.00
30.00
20.00
10.00
0.00
miliseconds
0
20
40
60
80
meters
Direct
Refracted


Direct and Refracted Waves
t_i
60.00
50.00
40.00
30.00
20.00
10.00
0.00
miliseconds
d_co
0
20
40
60
80
meters
Direct
Refracted

Interpretation of Refraction Travel Time Curves

Interpretation objective is to determine interface depths and layer velocities.

Data interpretation requires making **assumptions** about layering in subsurface: shape and number of different first arrivals.

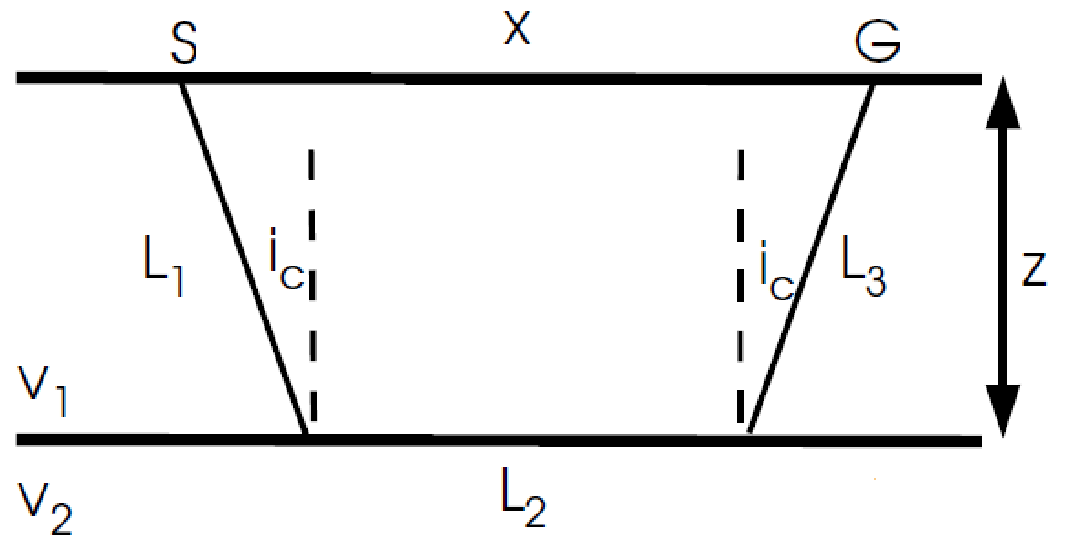
The Assumptions are:

- Subsurface is composed of stack of layers, usually separated by plane interfaces
- Seismic velocity is uniform in each layer
- Layer velocities increase in depth
- All ray paths are located in vertical plane, i.e. no 3-D effects with layers dipping out of plane of profile.

Two layer case

Travel way of the refracted wave

$$L_R = L_1 + L_2 + L_3$$



Travel time of the refracted wave

$$T_R = \frac{L_1}{v_1} + \frac{L_2}{v_2} + \frac{L_3}{v_1}$$

$$\boxed{\Rightarrow T_R = \frac{x}{v_2} + \frac{2z}{v_1}\frac{\sqrt{v_2^2 - v_1^2}}{v_2}}$$

Travel time of the direct wave

$$\boxed{T_D = \frac{L_D}{v_1}}$$

From travel times of direct arrival and critical refraction, we can find *velocities of two layers* and *depth to interface* :

1. Velocity of layer 1 given by slope of direct arrival
2. Velocity of layer 2 given by slope of critical refraction
3. Depth of the refractor

Derivation of the travel time equation for the case of two horizontal layers

Starting with: $T_R = \frac{L_1}{V_1} + \frac{L_2}{V_2} + \frac{L_3}{V_1}$

$T_R = Z / V_1 \cos i_c + (X - 2Z\tan i_c)/V_2 + Z / V_1 \cos i_c$

$T_R = X/V_2 + 2Z / V_1 \cos i_c - 2Z \tan i_c / V_2$

Substituting for V_2 and for $\tan i_c$ according to relations (4) and (5) above, we obtain:

$T_R = X/V_2 + 2Z / V_1 \cos i_c - 2Z \sin^2 i_c / V_1 \cos i_c$

$T_R = X/V_2 + 2Z (1 - \sin^2 i_c) / V_1 \cos i_c$

Using relation (6)

$\boxed{T_R = X/V_2 + 2Z \cos i_c / V_1,}$ Or:

$T_R = X/V_2 + 2Z \sqrt{(1-(V_1/V_2)^2)} / V_1$

Finally, we obtain:

$$\boxed{T_R = X/V_2 + 2Z \sqrt{V_2^2 - V_1^2} / V_1 * V_2}$$

The following relations need to be taken in consideration :

$L_1 = L_3$ ……………………………………………,(1)

$\cos i_c = Z/L_1 \rightarrow L_1 = Z/\cos i_c$ …….. ………(2)

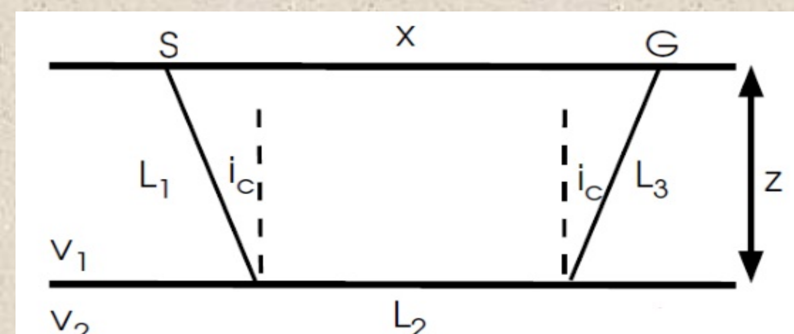
$L_2 = X - 2(Z\tan i_c)$ ………………. …………(3)

$\sin i_c = V1/V2 \rightarrow V2 = V1/\sin i_c$ ……….(4)

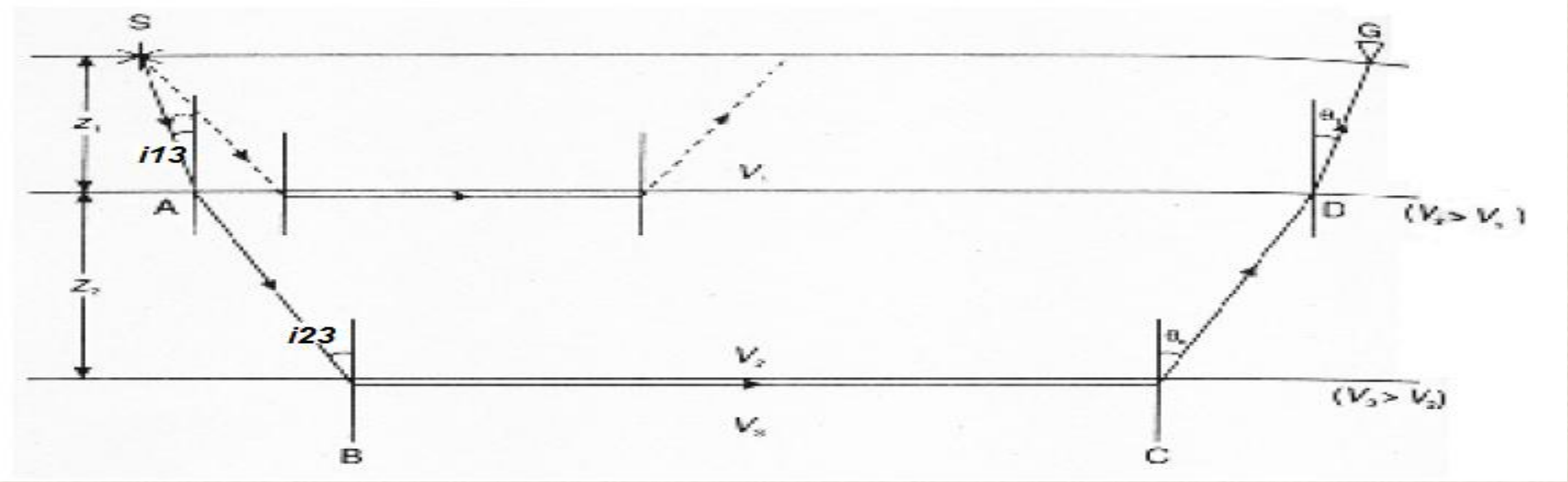
$\tan i_c = \sin i_c / \cos i_c$ ………….. ………….(5)

$\sin^2 i_c + \cos^2 i_c = 1 \rightarrow \sin^2 i_c = 1 - \cos^2 i_c$…(6)

$\rightarrow \cos i_c = \sqrt{1 - \sin^2 i_c} = \sqrt{V_2^2 - V_1^2} / V_2$



Three Horizontal Layers Case



Based on the figure above, the refracted travel time can be written as:

$T_R = 2SA/V_1 + 2AB/ V_2 + BC/V_3$

$T_R = 2Z_1 / V_1 \cos i_{13} + 2Z_2 / V_2 \cos i_{23} + (X – 2Z_1 \tan i_{13} - 2Z_2 \tan i_{23})/V_3$

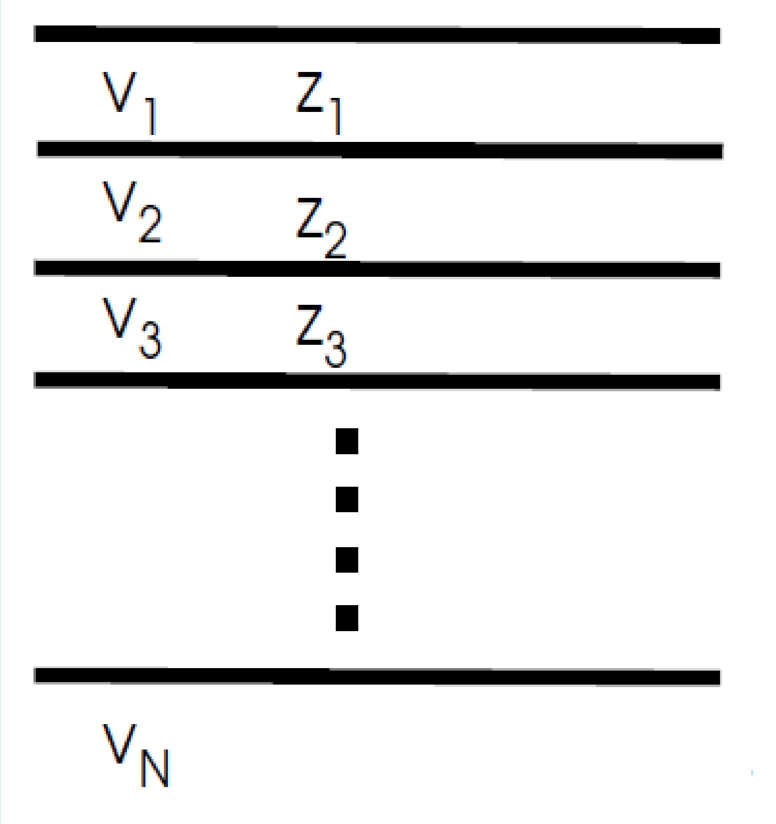
$T_R = X/V_3 + (2Z_2 /V_2)* (1/\cos i_{23} - V_2 \tan i_{23} / V_3) + (2Z_1 /V_1)* (1/\cos i_{13} – V_1 \tan i_{13} / V_3)$

Noting that: $V_2/ V_3 = \sin i_{23}$, and $V_1/ V_3 = \sin i_{13}$, we obtain:

$\boxed{T_R = X/V_3 + 2Z_2 \cos i_{23}/ V_2 + 2Z_1 \cos i_{13}/ V_1}$ Or:

$\boxed{T_R = X/V_3 + 2Z_2\sqrt{V_3^2 – V_2^2}/ V_3 {*} V_2 + 2Z_1\sqrt{V_3^2 – V_1^2}/ V_3 {*} V_1}$

Multilayer case

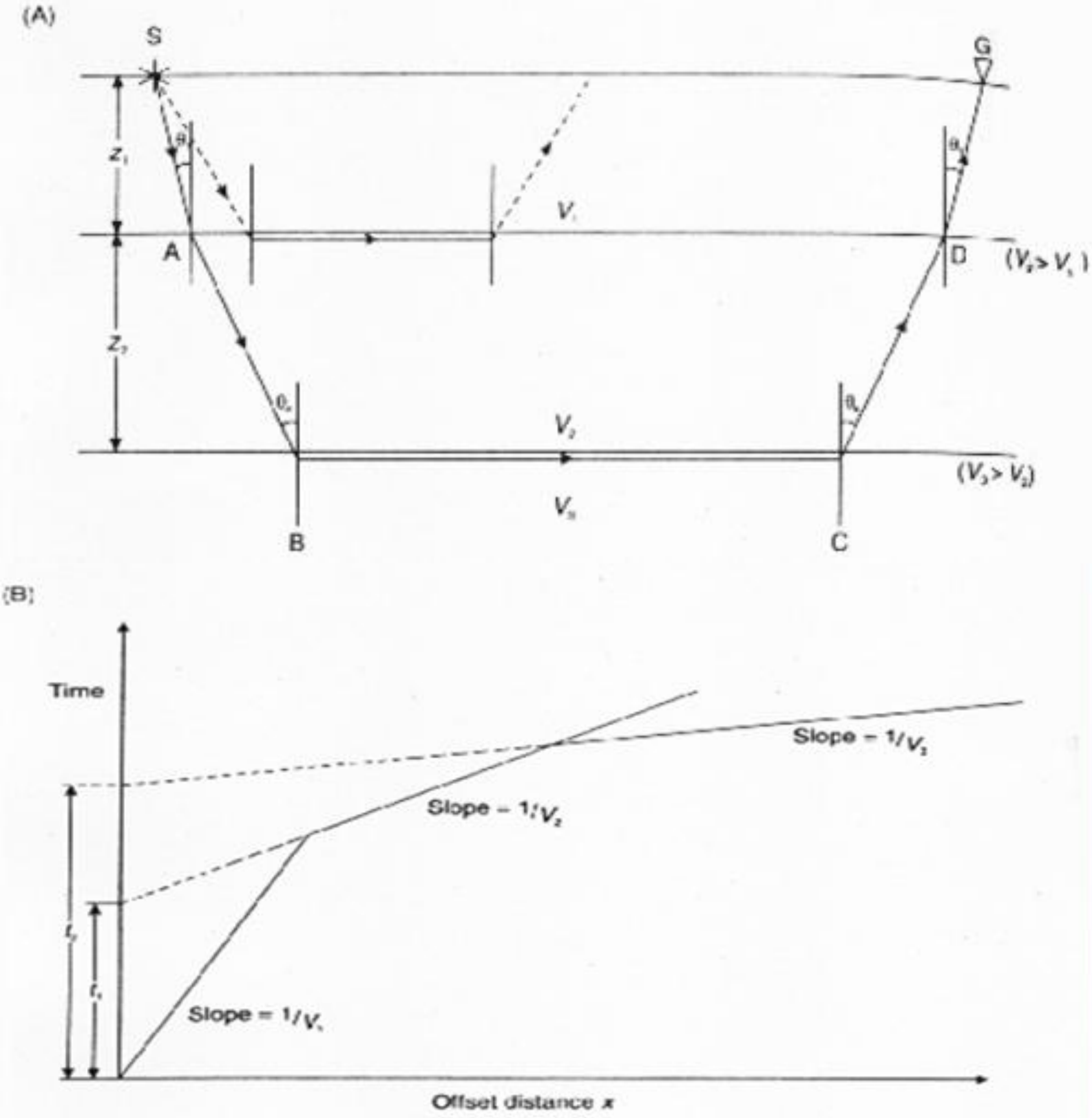


N layers

$v_1 < v_2 < v_3 < \ldots < v_N$

Travel time

$$T_N = \frac{x}{v_N} + \frac{2}{v_N} \sum_{k=1}^{N-1} (z_k - z_{k-1}) \sqrt{(v_N / v_k)^2 - 1}$$



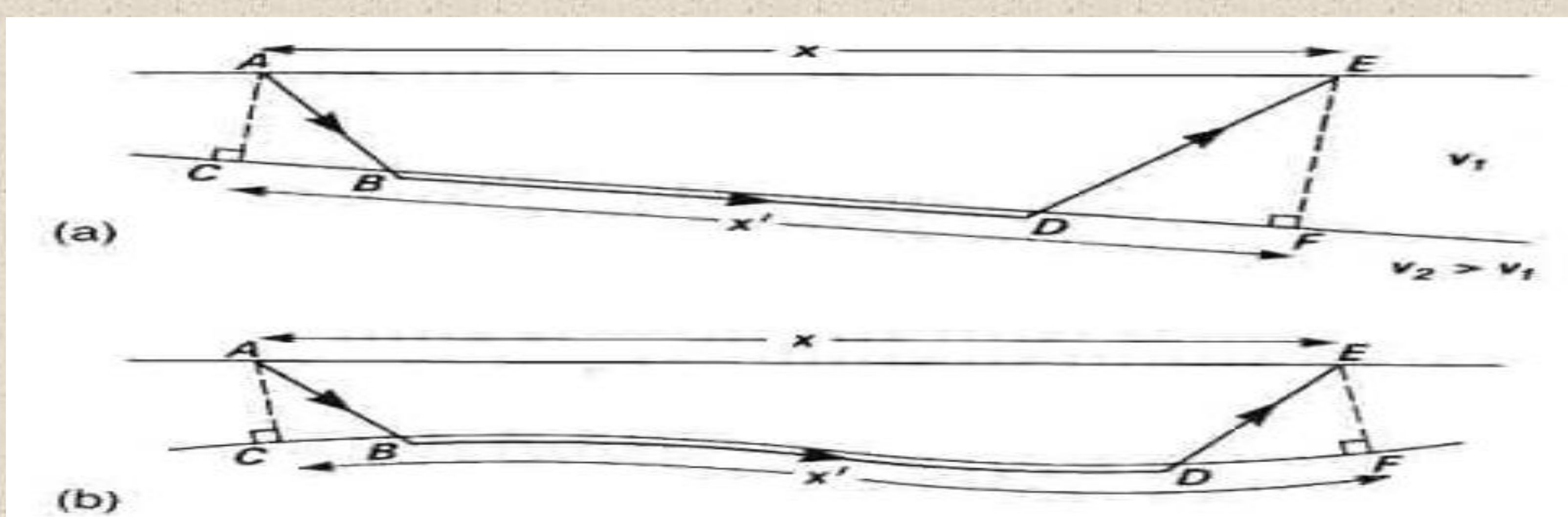
(A) Simple raypaths diagram for refracted rays, and (B) their respective travel time-distance graphs for a three-layer case with horizontal planar interfaces.

Delay Time Concept

For irregular travel time curves, e.g. due to bedrock topography or glacial fill, much analysis is based on delay times.

Total Delay Time is defined as *the difference in travel time along actual ray path and projection of ray path along refracting interface*:

$$\delta t = T_{AB} - T_{CF}$$



$$T_{CF} = \frac{CF}{V_2}$$

<u>Total delay time</u> is the delay time at shot plus delay time at geophone:

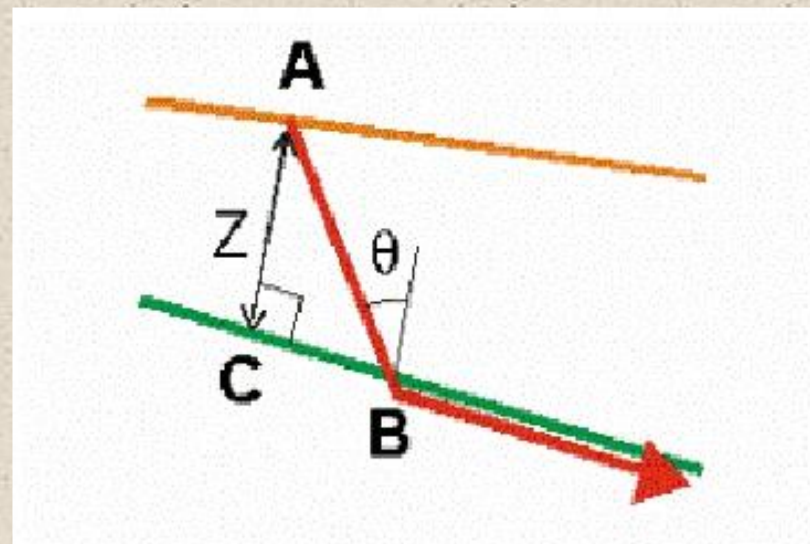
$$\delta t = \left(\frac{AB}{V_1} - \frac{CB}{V_2}\right) + \left(\frac{DE}{V_1} - \frac{DF}{V_2}\right) = \delta t_S + \delta t_G \approx T_{AE} - \frac{x}{V_2}$$

For small dips, can assume x=x′, and:

$$\delta t = T_{AE} - \frac{x'}{V_2}$$

Calculation of Refractor Depth from Delay Time

If velocities of both layers are known, then refractor depth at point A can be calculated from delay time at point A:



$$\delta t_A = \frac{AB}{V_1} - \frac{BC}{V_2}$$

Using the triangle ABC to get lengths in terms of Z:

$$\delta t_A = \frac{Z}{V_1 \cos\theta} - \frac{Z \tan\theta}{V_2}$$

$$= \frac{Z}{V_1 \cos\theta}\left(1 - \frac{V_1 \sin\theta}{V_2}\right)$$

Using Snell's law to express angles in terms of velocities:

$$\delta t_A = \frac{Z}{V_1\left(1 - \frac{V_1^2}{V_2^2}\right)^{\frac{1}{2}}}\left(1 - \frac{V_1^2}{V_2^2}\right)$$

Simplifying:

$$\delta t_A = \frac{Z\left(V_2^2 - V_1^2\right)^{\frac{1}{2}}}{V_1 V_2}$$

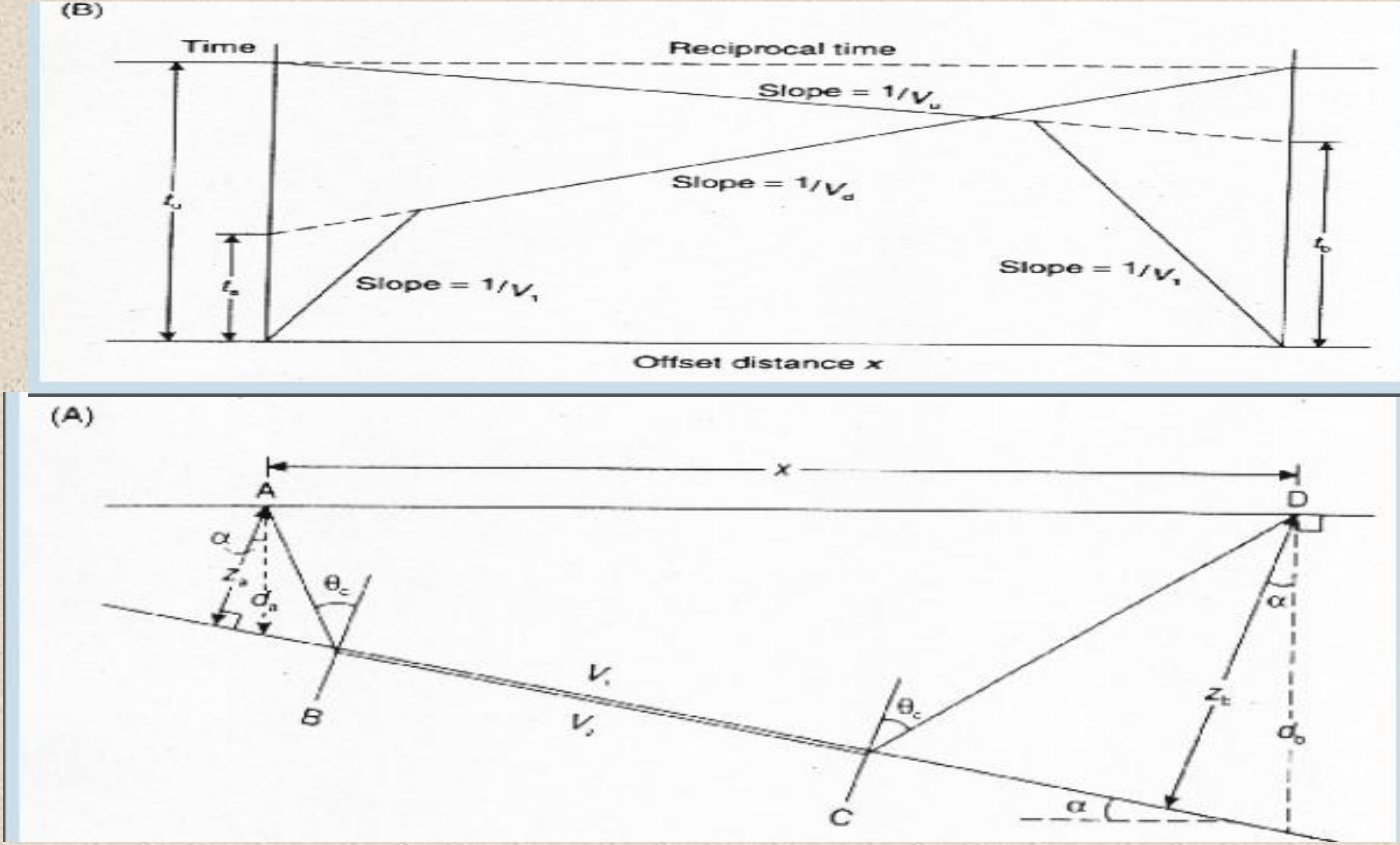
So, refractor depth at A is:

$$Z = \frac{\delta t_A V_1 V_2}{\left(V_2^2 - V_1^2\right)^{\frac{1}{2}}}$$

Dipping layer case

When a refractor lies at an angle to the horizontal, it is no longer adequate to undertake only one direction of (forward) shooting. It becomes necessary to carry out both forward and reverse shooting in order to determine the parameters.

The refractor velocities determined in the case of dip are referred to as apparent velocities;
v_u: upslope direction
v_d: downslope direction



A. Raypath geometry over a refractor dipping at an angle α, and
B. the respective travel time-distant graph for the forward (down-dip) and reverse (up-dip) shooting directions.

Travel time calculations for a dipping refractor

Total travel time over a refractor dipping at an angle is given by:

$$T_{ABCD} = (x\cos\alpha)/v_2 + [(z_a + z_b)\cos i_c]/v_1 \quad (1)$$

where v_2 is the refractor velocity, and z_a and z_b are the distances perpendicular to the refractor.

The down-dip travel time t_d is given by:

$$t_d = x[\sin(\theta_c + \alpha)]/v_1 + t_a \quad (2)$$

where $t_a = 2z_a(\cos\theta_c)/v_1$.

The up-dip travel time t_u is given by:

$$t_u = x[\sin(\theta_c - \alpha]/v_1 + t_b$$

Where $t_b = 2z_b(\cos\theta_c)/v_1$

Equations (1) and (2) above can be written in terms of the apparent up-dip velocity v_u and down-dip velocity v_d such that:

$$t_d = x/v_d + t_a, \quad \text{where} \quad v_d = v_1/\sin(\theta_c + \alpha)$$

$$t_u = x/v_u + t_b, \quad \text{where} \quad v_u = v_1/\sin(\theta_c - \alpha)$$

An approximate relationship between true and apparent velocities for shallow angles of dip (<10°) is given by:

$$v_2 \approx (v_d + v_u)/2$$

CONCLUSION NOTES ON DIPPING LAYERS

- Dipping layers can be detected only by conducting both forward and reverse shootings
- The intercept times for forward and reverse shooting in case of dipping layers are not equals
- In both cases, horizontal and dipping layer cases,:
  - the velocity of the first layer is equal in forward and reverse shootings
  - the total travel time – reciprocal time – of the refracted wave is equal

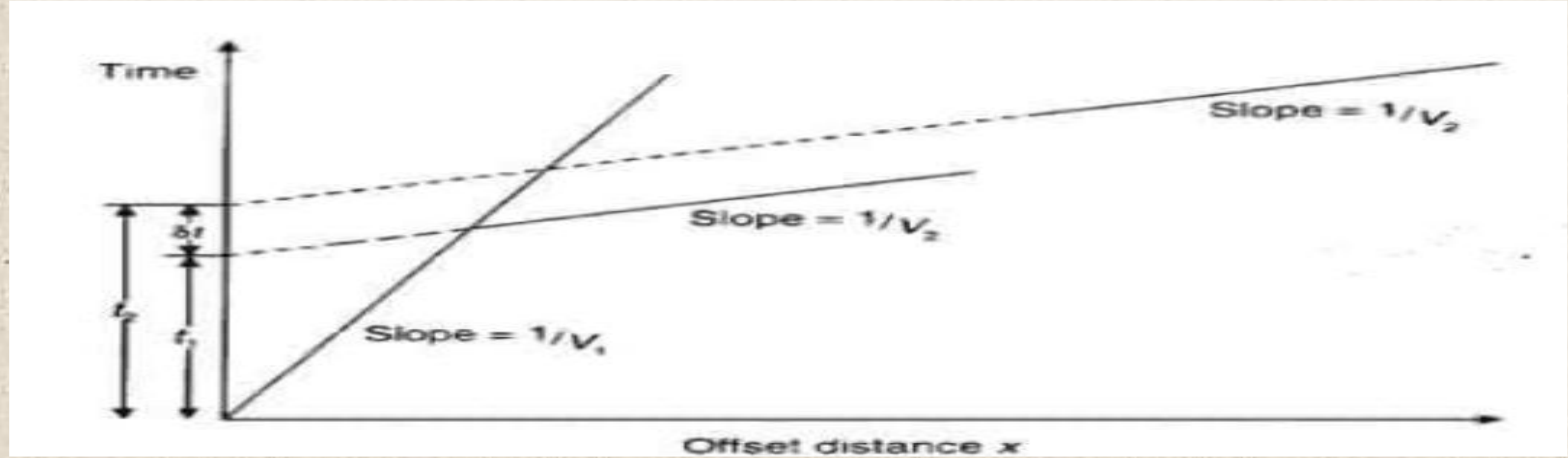
- **in dipping layers case, apparent velocities and thicknesses are obtained**
- **The depths d_a (updip) and d_b (downdip) –both perpendicular to the surface- can be calculated from the following relations:**

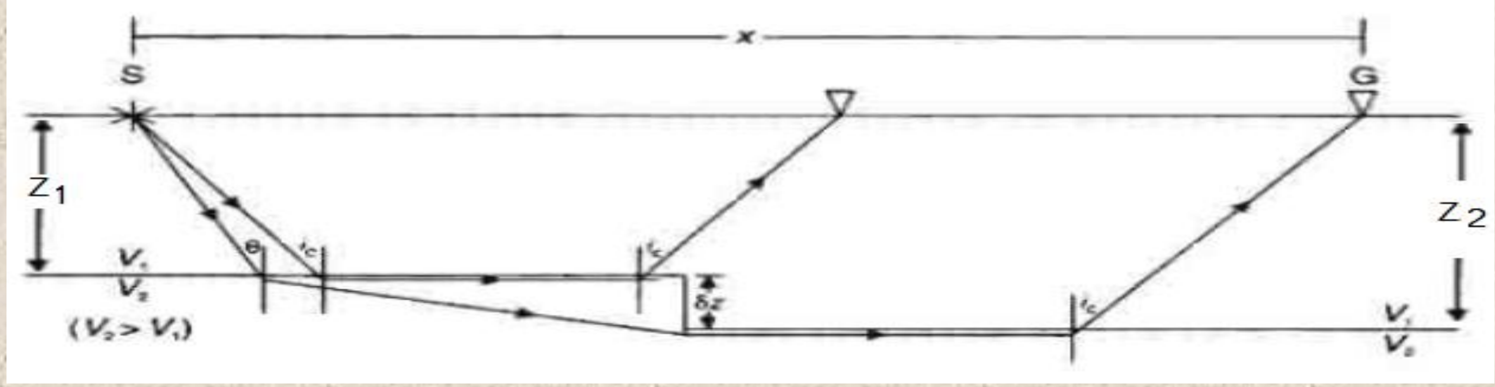
$$d_a = z_a / \cos\alpha$$

$$d_b = z_b / \cos\alpha$$

NORMAL FAULT CASE (Diffraction)

In the presence of a normal fault affected the refractor, there will be a sharp displacement in the travel time curve. See the figure below.





Taking in consideration the profile in the previous figure directed from S (shoot point) to the G (geophone):

The travel time for the refracted wave before arrival to the fault boundary is:

$$t_b = X/V_2 + 2Z_1\cos i_c / V_1$$

The travel time for the refracted wave after arrival to the fault boundary is:

$$t_a = X/V_2 + (Z_1+Z_2)\cos i_c / V_1$$

The difference in the intercept times for the above two equations is given by:

$\delta_t = \delta z \cos i_c / V_1$. Finally, we obtain:

$$\boxed{\delta_z = \delta t\, V_1/\cos i_c,}$$

OR:

$$\boxed{\delta_z = \delta_t\, V_{1*} V_2 / \sqrt{V_2^2 - V_1^2}}$$

Seismic Refraction:
Additional Interpretation Methods

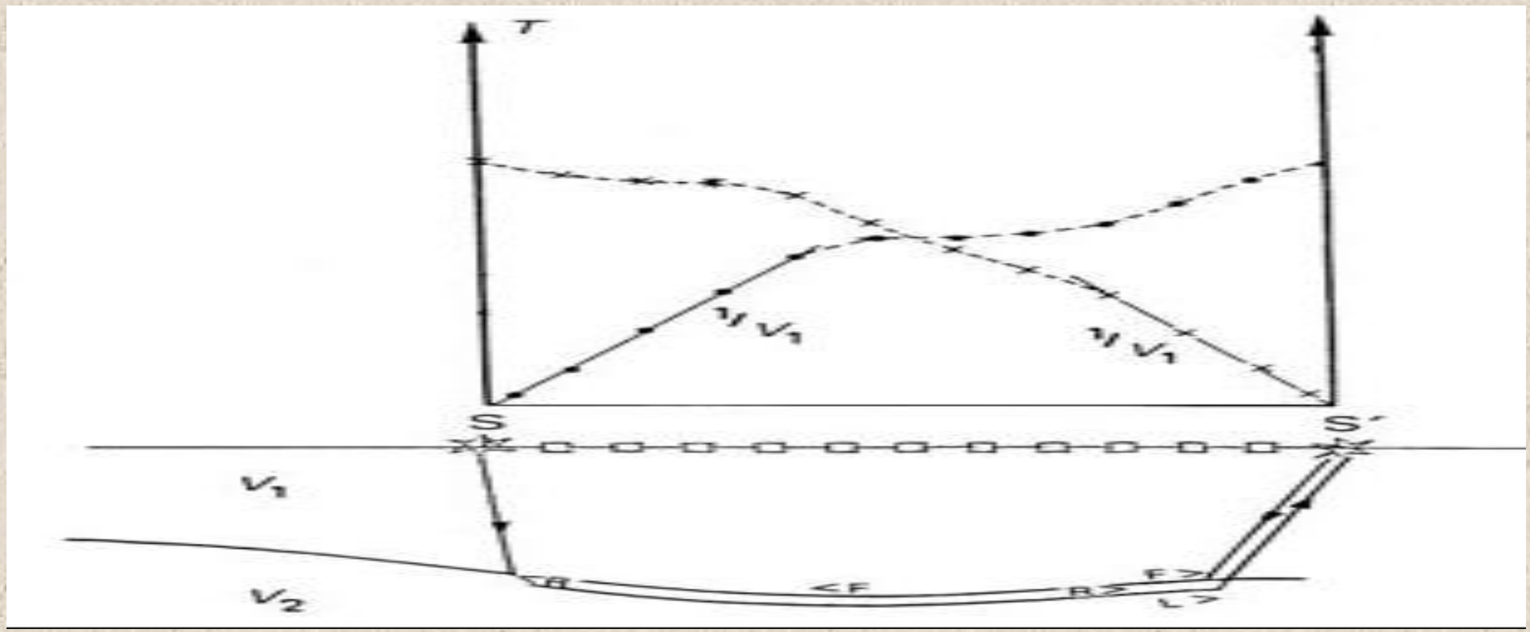
I. Phantom Arrivals
II. Plus minus-method
III. GRM Method

Seismic Refraction:
Additional Interpretation Methods

I. Phantom Arrivals

In case of irregular interfaces, it is not possible to extrapolate the head wave arrival time curve back to the intercept. Therefore, there is a need to find a way to calulate the layer thickness beneath the shotpoint, S. The phantom arrivals technique is used in such cases.

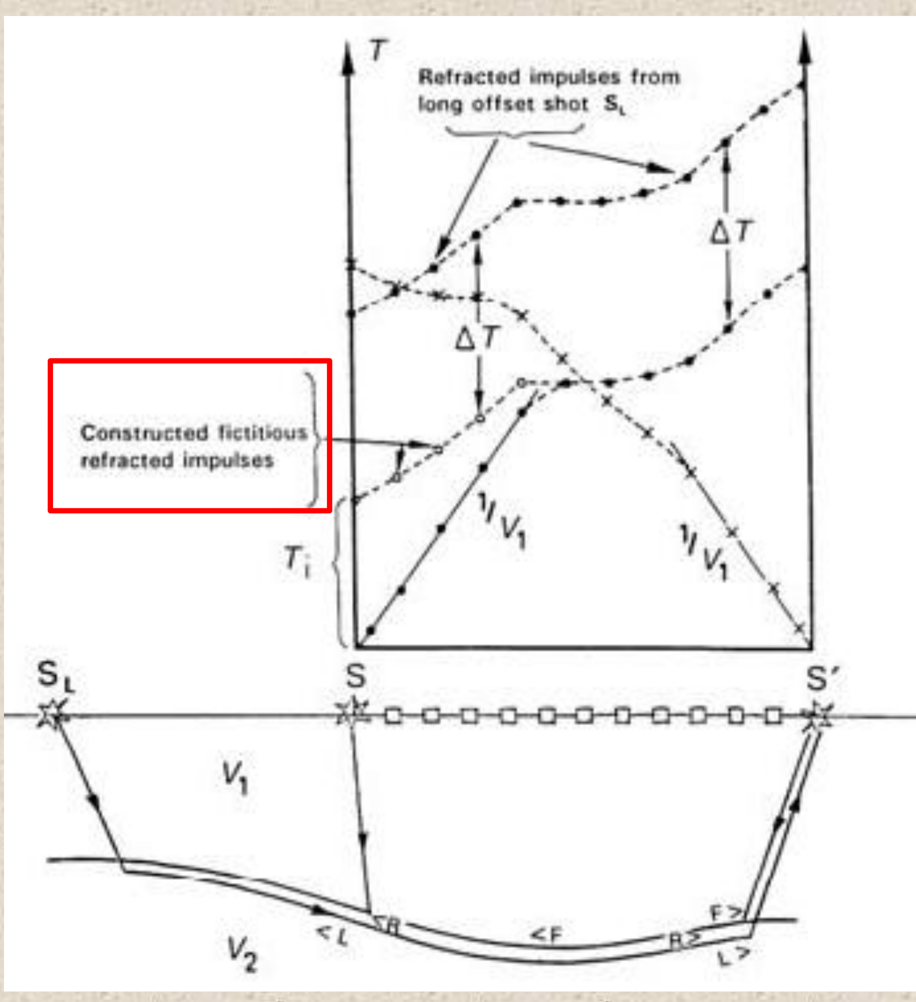
The main advantage of this technique: it removes the necessity to extrapolate the travel time graph from beyond the crossover point back to the zero-offset point.



Phantom arrivals

To obtain the phantom arrivals, the following procedure is followed:

- A long-offset shot, SL, is conducted
- The head wave travel time curves for both shots (S and S_L) will be parallel, offset by time ΔT
- ΔT is Subtracted from the SL arrivals to *generate fictitious 2nd layer arrivals close to the shot point S* – these are called the phantom arrivals.
- The intercept point at shotpoint, S can then be determined: Ti
- Finally, the perpendicular layer thickness beneath the shotpoint, S, can be now determined using the known formula:



$$T_i = \frac{2h_s\sqrt{V_2^2 - V_1^2}}{V_2 V_1}$$

II. Plus minus-method

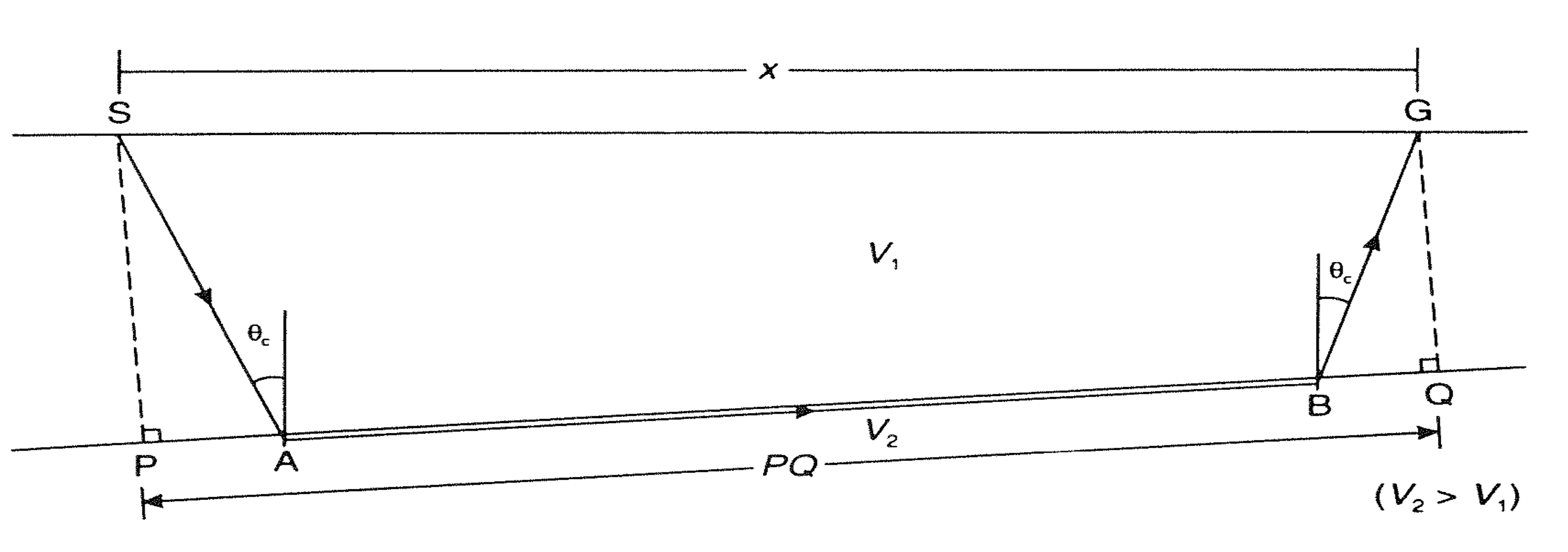
This method uses intercept times and delay times in the calculation of the depth to the refractor <u>below any geophone location</u>.

<u>Assumptions to use the method:</u>

-Present layers are homogeneous

-Large velocity contrast between the layers

-Angle of dip of the refractor is less than 10 degrees

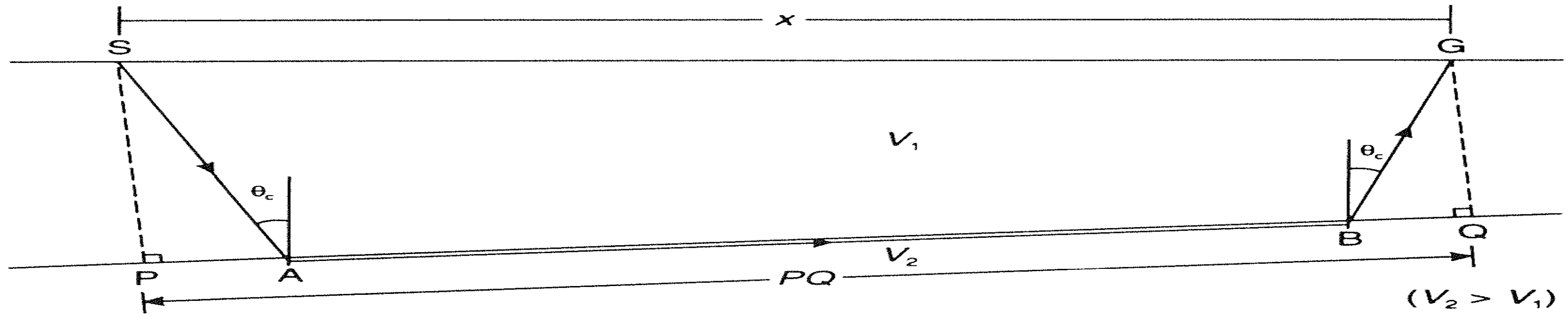


The delay time (δt) is the <u>difference</u> in time between:

1) T(SG) along the track SABG and,

2) T(PQ)

➢ The total delay time is effectively the sum of the "shot-point delay time", δt_s, and the "geophone delay time", δt_g



The total delay time is given by:

$$\boxed{\delta t = T_{SG} - T_{PQ}}$$

and

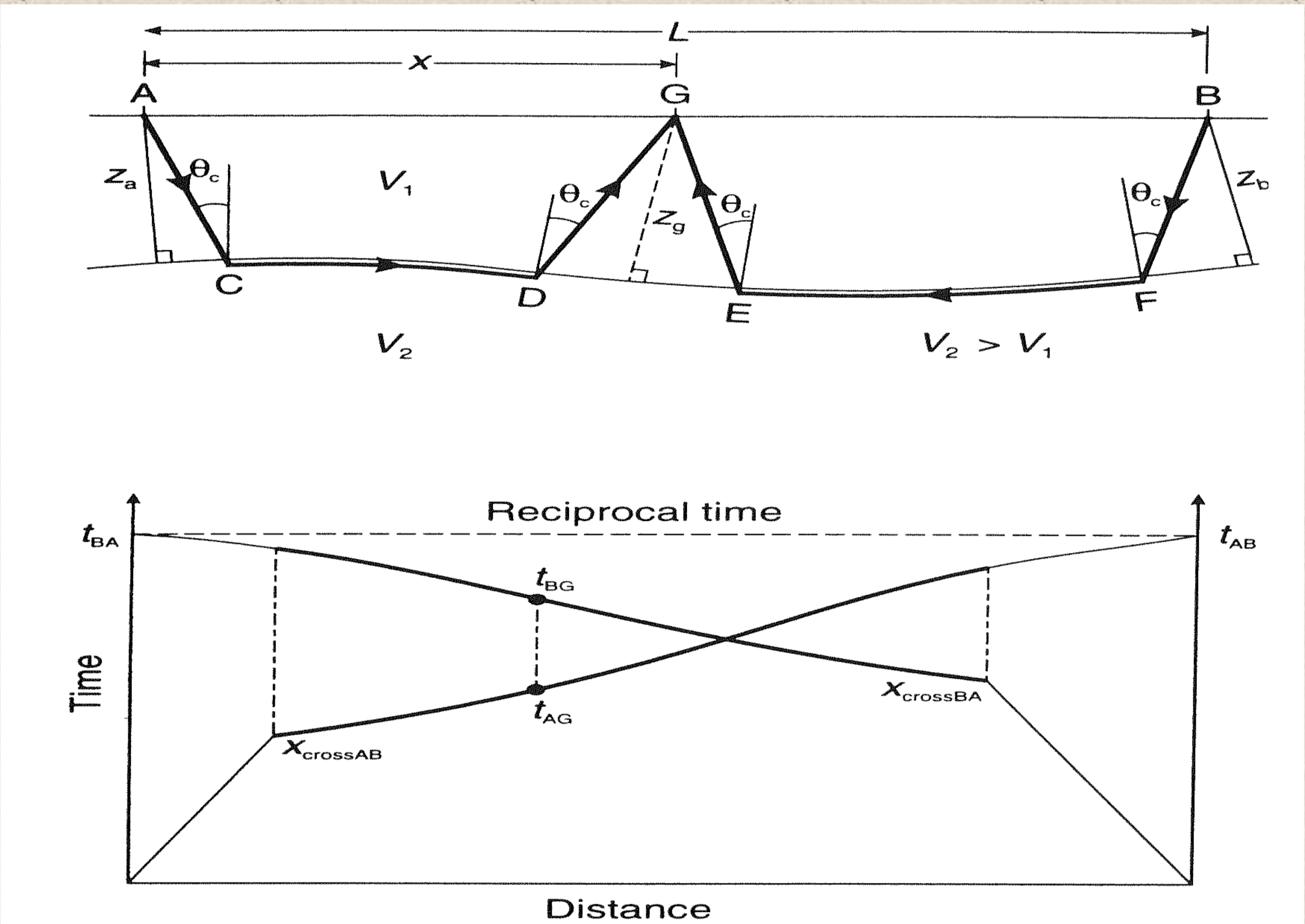
$$T_{SG} = (SA + BG)/V_1 + AB/V_2 \quad \text{and} \quad T_{PQ} = PQ/V_2 .$$

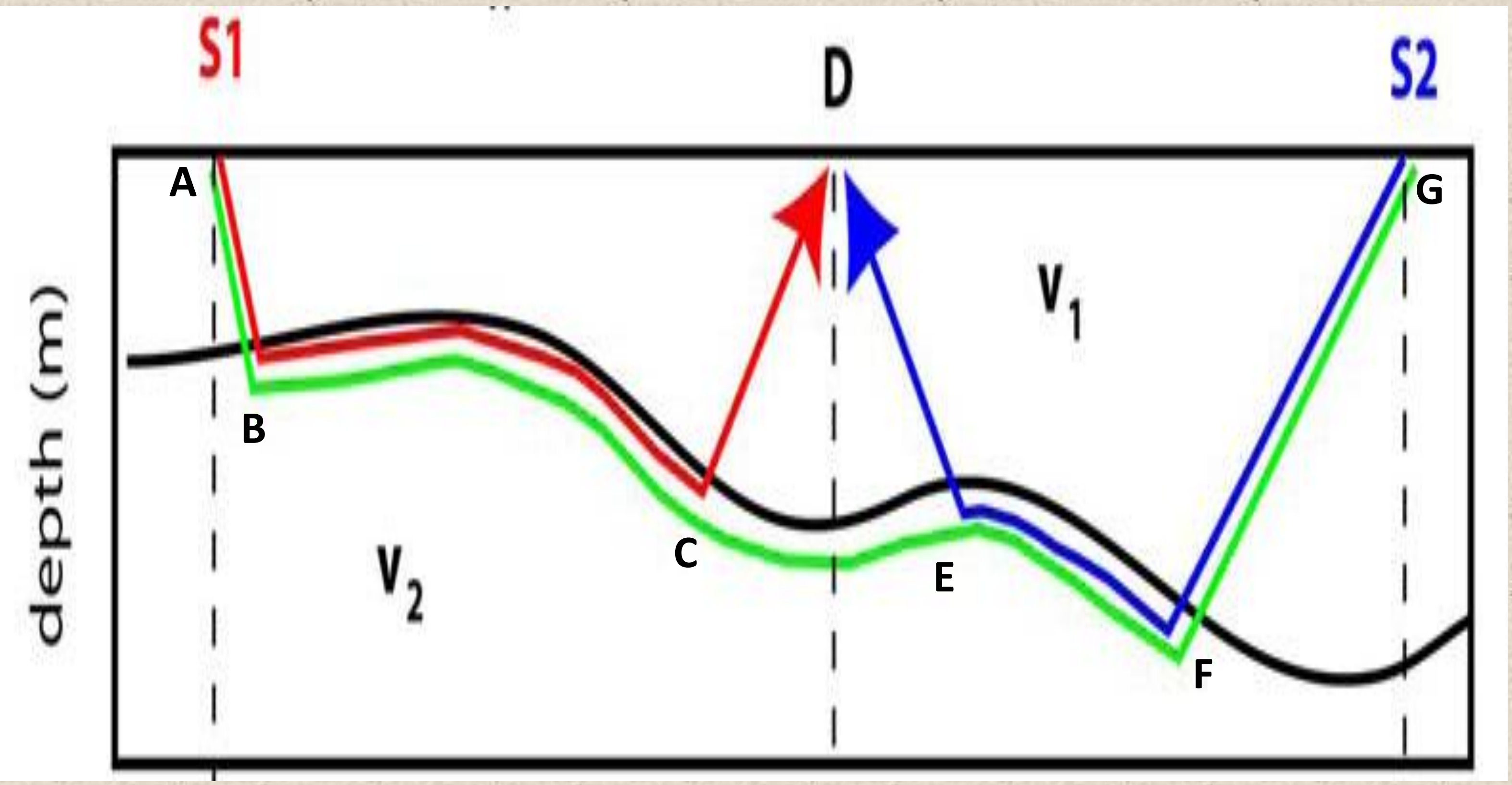
Thus:

$$\begin{aligned}\delta t &= (SA + BG)/V_1 - (PA + BQ)/V_2 \\ &= (SA/V_1 - PA/V_2) + (BG/V_1 - BQ/V_2) \\ &= \delta t_s + \delta t_g \approx T_{SG} - x/V_2 .\end{aligned}$$

Alternatively:

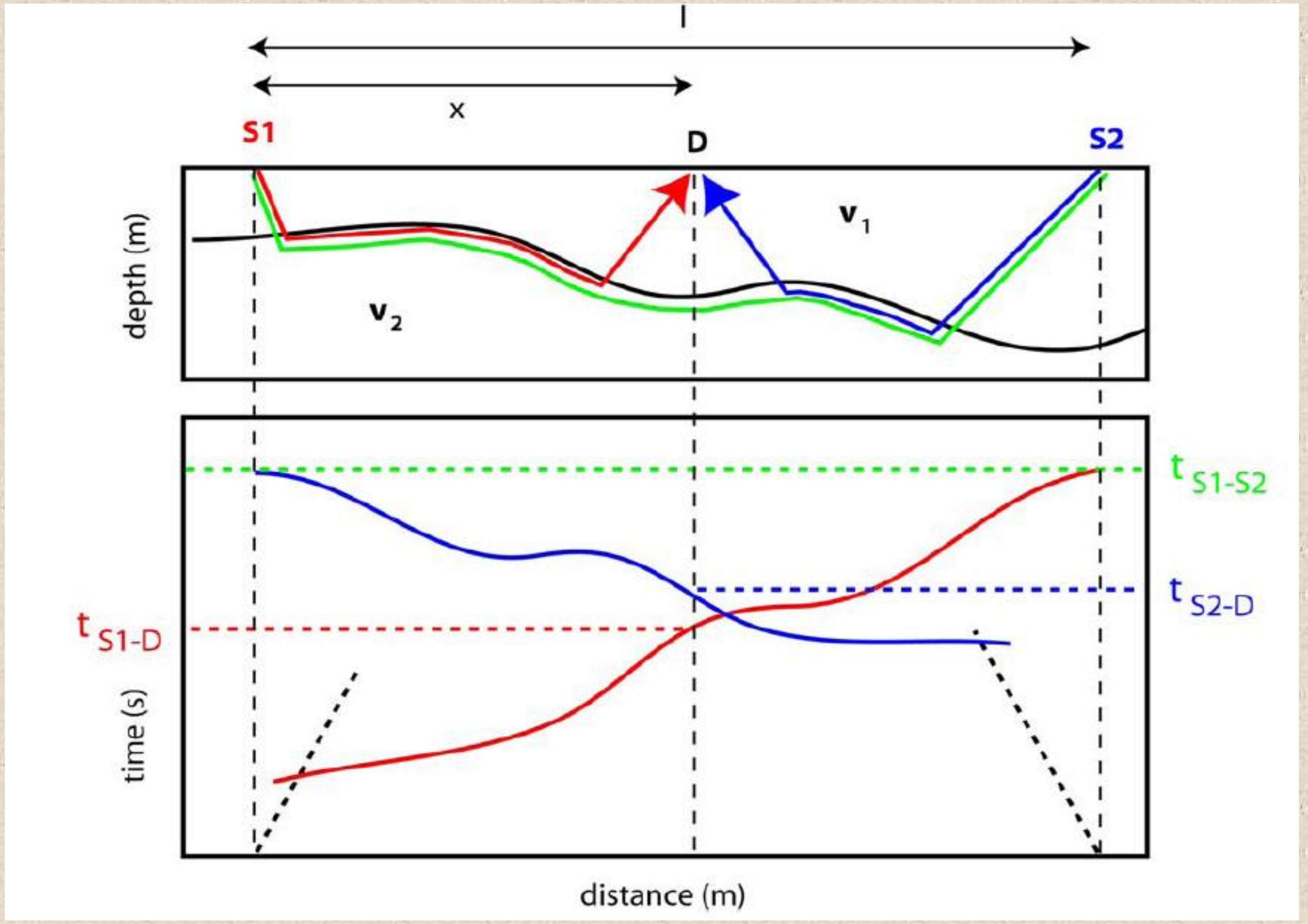
$$\boxed{T_{SG} = x/V_2 + \delta t_s + \delta t_g}$$

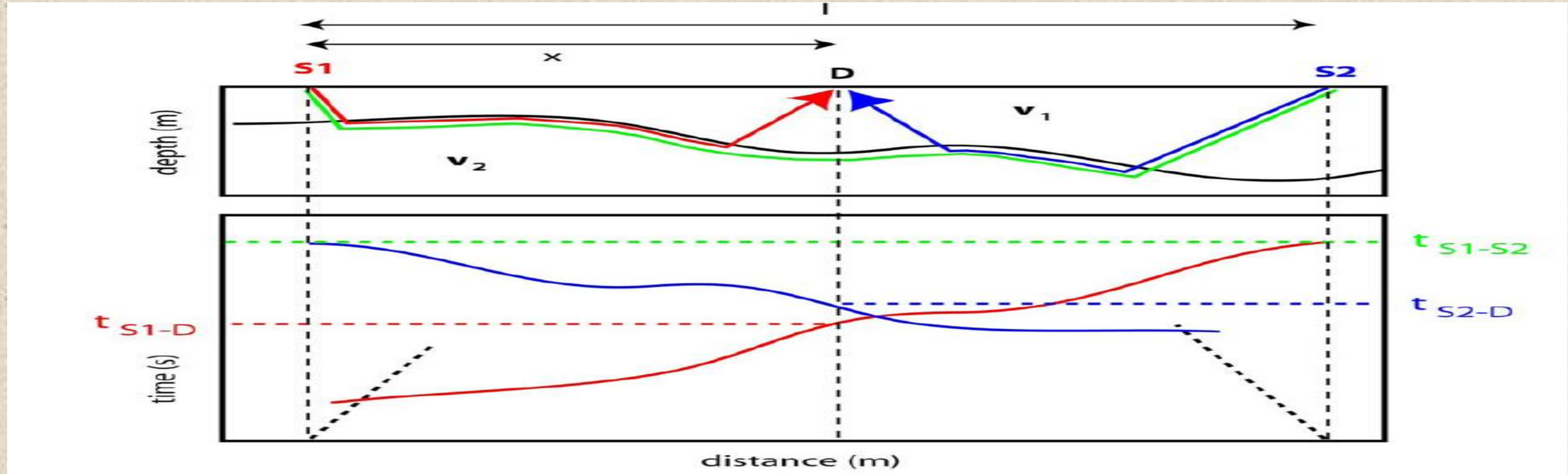

L
x
A
G
B
z_a
θ_c
V_1
θ_c
z_g
θ_c
θ_c
z_b
C
D
E
F
V_2
$V_2 > V_1$
Reciprocal time
t_{BA}
t_{AB}
t_{BG}
t_{AG}
$x_{crossBA}$
$x_{crossAB}$
Time
Distance



Time CDE = Time ABCD + Time DEFG – Time ABCEFG

Total time


l
x
S1
D
S2
v1
v2
depth (m)
time (s)
t S1-S2
t S2-D
t S1-D
distance (m)



Consider the model with two layers and an irregular interface. The refraction profile is reversed with two shots (S1 and S2) fired into each geophone (D).

Consider the following three travel times:

(a) The reciprocal time is the time from *S1 to S2*

$$t_{S_1S_2} = \frac{l}{v_2} + \delta_{S_1} + \delta_{S_2} = t_{S_2S_1}$$

(b) Forward shot into the geophone

$$t_{S_1D} = \frac{x}{v_2} + \delta_{S_1} + \delta_D$$

(c) Reverse shot into the geophone

$$t_{S_2D} = \frac{(l-x)}{v_2} + \delta_{S_2} + \delta_D$$

The goal is to find V_2 and the delay time at the detector, δD. From the delay time, δD , we can find the depth of the interface.

(a) The reciprocal time is the time from S^1 to S^2

(b) Forward shot into the detector

(c) Reverse shot into the detector

$$t_{S_1S_2} = \frac{l}{v_2} + \delta_{S_1} + \delta_{S_2} = t_{S_2S_1}$$

$$t_{S_1D} = \frac{x}{v_2} + \delta_{S_1} + \delta_D$$

$$t_{S_2D} = \frac{(l-x)}{v_2} + \delta_{S_2} + \delta_D$$

Minus term to estimate velocity (v_2)

(b)-(c) will eliminate δ_D

$$t_{S_1D} - t_{S_2D} = \frac{(2x-l)}{v_2} + \delta_{S_1} - \delta_{S_2}$$

$$t_{S_1D} - t_{S_2D} = \frac{2x}{v_2} + C$$

where C is a constant. A plot of $t_{S_1D} - t_{S_2D}$ versus $2x$ will give a line with slope = $1/v_2$

(a) The reciprocal time is the time from S^1 *to* S^2

(b) Forward shot into the detector

(c) Reverse shot into the detector

$$t_{S_1S_2} = \frac{l}{v_2} + \delta_{S_1} + \delta_{S_2} = t_{S_2S_1}$$

$$t_{S_1D} = \frac{x}{v_2} + \delta_{S_1} + \delta_D$$

$$t_{S_2D} = \frac{(l-x)}{v_2} + \delta_{S_2} + \delta_D$$

Plus term to estimate delay time at the detector

(b)+(c) gives

$$t_{S_1D} + t_{S_2D} = \frac{l}{v_2} + \delta_{S_1} + \delta_{S_2} + 2\delta_D$$

Using the result (a) we get

$$t_{S_1D} + t_{S_2D} = t_{S_1S_2} + 2\delta_D$$

Re-arranging to get an equation for δ_D

$$\delta_D = \frac{1}{2}(t_{S_1D} + t_{S_2D} - t_{S_1S_2})$$

This process is then repeated for all detectors in the profile

Calculation of the **depth to the refractor beneath any geophone** (z) from the delay time:

$$T^{+} = t_{S1D} + t_{S2D} - t_{DS1S2} = 2\delta t_{D}$$

$$2\delta t_{D} = \frac{2\, z \cos(i)}{V1}$$

, where *i* is the critical angle

$$\boxed{z = \frac{T^{+}V1}{2\cos(i)} = \frac{T^{+}V1V2}{2\sqrt{(V_2^2 - V_1^2)}}}$$

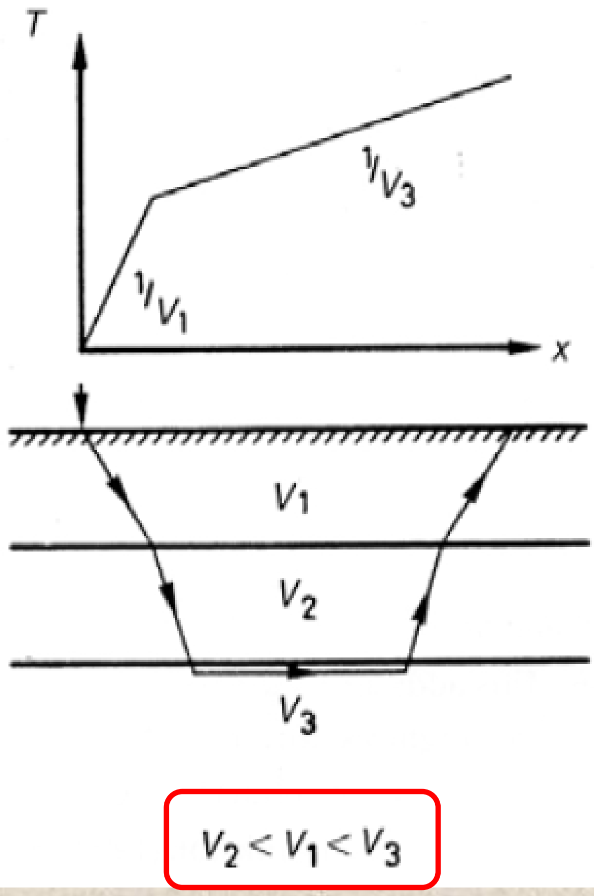
Limitations of Seismic Refraction Method

I. Low Velocity Layers

- **They are completely invisible to the refraction method**
- **They will cause miss-interpretation of the depth of lower lying layers**

  The intercept of the refraction from layer 3 will be dependent on the thickness and velocity in layer 2

- **Lower layers appear deeper than they are**



How to resolve low velocity layer?

The detection of the presence of a low velocity layer can be done either by:

- direct geological information (such as drilling),

or

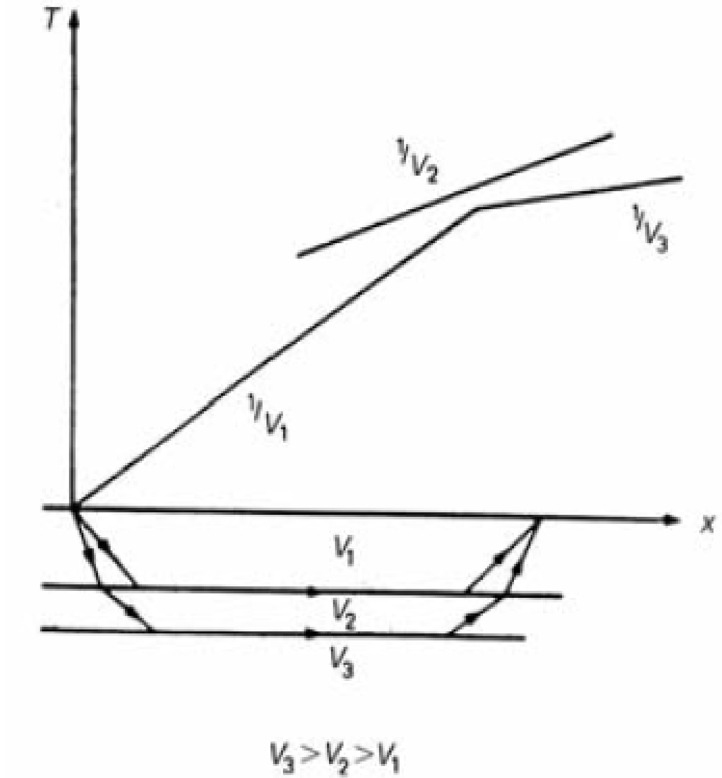
- by using other geophysical exploration methods, such as seismic reflection.

II. Hidden Layers

- **If a layer is thin it may never produce a first arrival**

  Either the direct or refraction from a lower (much higher velocity layer) is always first

- **Lower layers always appear too shallow as a layer has been missed**



<u>How to resolve the hidden layer problem?</u>

It is necessary to drill a test well in areas where the presence of this problem is expected. This way, corrections to depth values obtained by seismic refraction can be made.

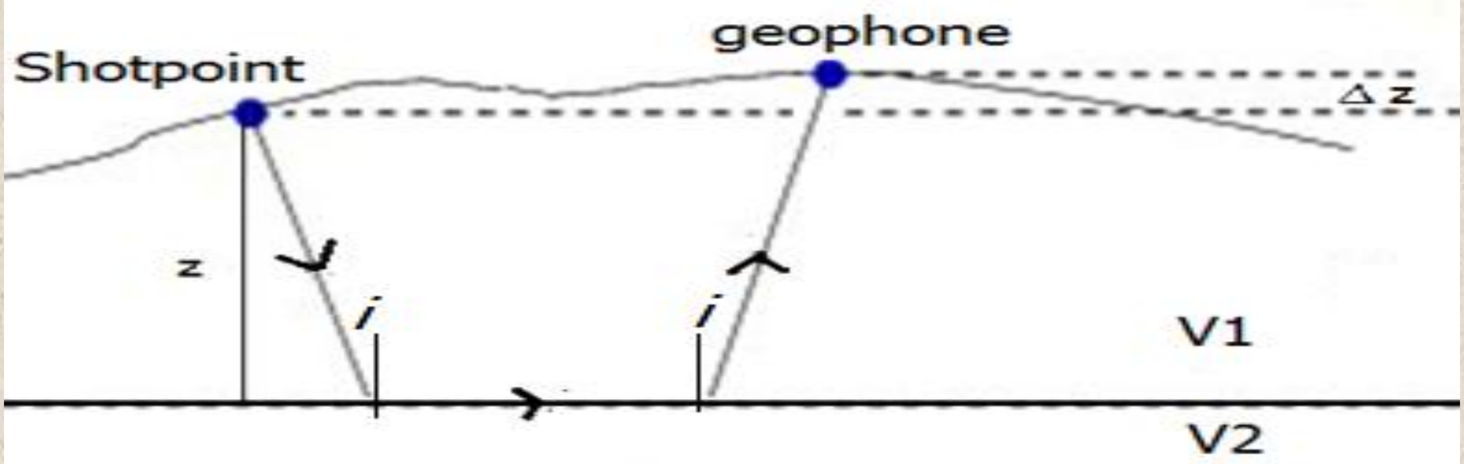
Corrections of Seismic Refraction Data

- In certain cases, it is necessary to apply corrections to refraction data in order to remove the difference in travel time that is **not produced** by the existence of irregularities in the subsurface refractor. There are two types of corrections:

I. Topographic correction

The purpose of this correction is to place the shotpoint and the receiving geophone at the same level. This can be done by:

(a) adding the time if the geophone (and/or the shotpoint), chosen level is higher.



$$T = \frac{X}{V2} + \mathrm{D.T},$$

$$D.T = \frac{2Z1 + \Delta Z}{V_1} . \mathrm{Cos\,i},$$

where:
D.T is the delay time,
i is the critical angle

(b) removing the time necessary for the wave to travel from the chosen level to the receiving geophone (and/or the shotpoint) if this level is lower.

➢Derivation of the equations for other situations (shotpoint above the chosen level) can be done using the same concept of delay time, as done in the above case.

When to apply the topographic corrections?

The application of topographic correction depends on the value of Δz

☞ If the delay time calculated for Δz is within the accuracy of measurement, then Δz can be ignored and there will be no need to apply topographic correction,

☞ If the delay time calculated for Δz is out of the range of accuracy of measurements, then there will be a need to apply topographic correction.

Example:

If the accuracy of measurement is ± 0.01 sec, then we have:

If ($D.T = \dfrac{2Z1 + \Delta}{V_1}$) < 0.01 → no need to make corrections

If ($D.T = \dfrac{2Z1 + \Delta Z}{V_1} . \text{Cos}\, i ,$) > 0.01 → corrections must be done

II. Correction for the zone of alteration

The zone of alteration is characterized by irregularities and variations in velocity and thickness.

The correction for this problem is based on calculating its thickness at several points along the seismic profile.

Calculation of the thickness of the zone of alteration can be obtained by conducting short seismic profiles in order to obtain two layer case. In this model, the first layer represent the alteration zone and has the velocity of V1.

By constructing Travel-Distance graph for these short profiles, the thickness of this zone can be calculated.

Seismic Refraction Field Procedures

➢It means the relative position of the shot point to the geophone.

I. Profile Shooting

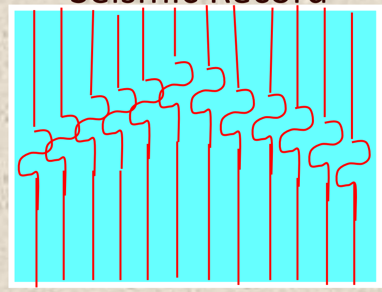
The shotpoint and the receiving geophones are located on one line.

Test profiles are usually conducted in the study area in order to:

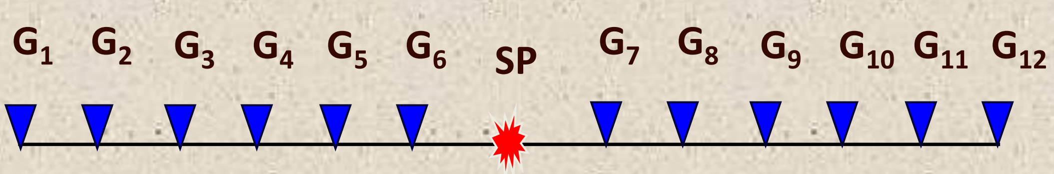
- choose the most adequate distance between the shopoints, and
- determine profile length most adequate for the investigated depth of the target.

There are several configurations for profile shooting:

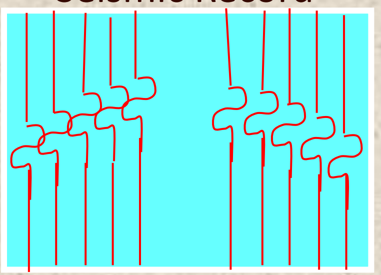
Seismic Record



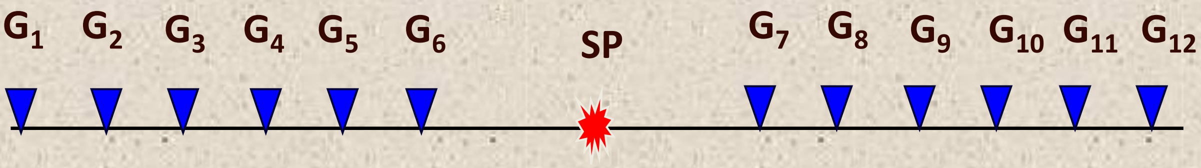
1- Split Spread: (SP) is placed at the center of spread.



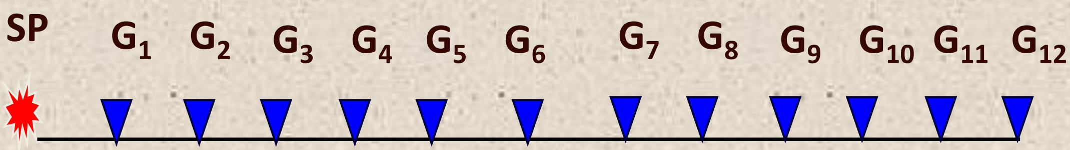
Seismic Record



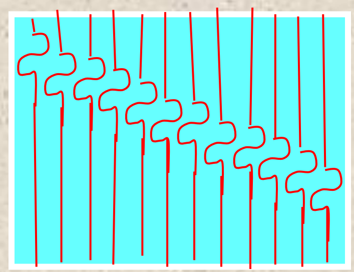
2- Split Spread with a gap: (SP) is placed at the center of spread with plotting a gap.



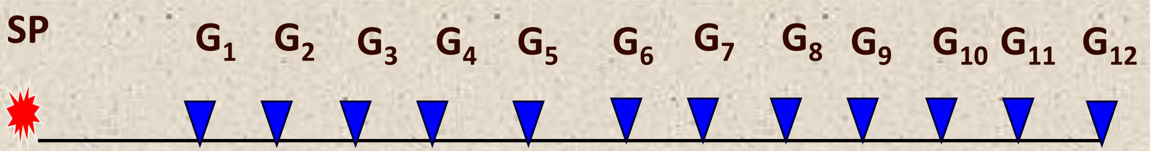
3- End of Spread: (SP) is placed at the end of spread.



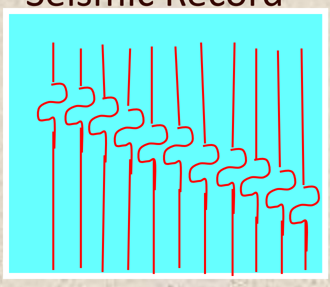
Seismic Record



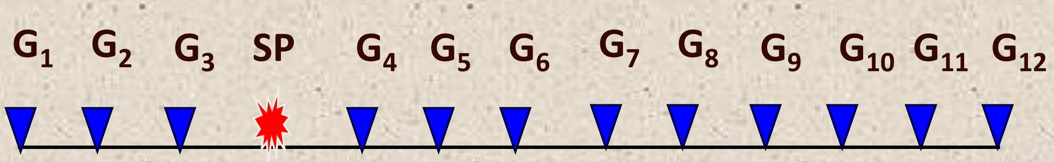
4- End of Spread with a gap: (SP) is placed at the end of spread with plotting a gap.



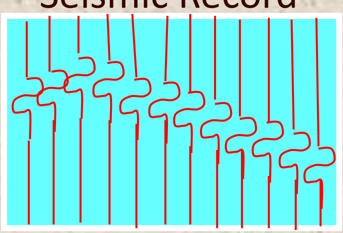
Seismic Record



5- Unbalanced Spread : (SP) is not placed at the center



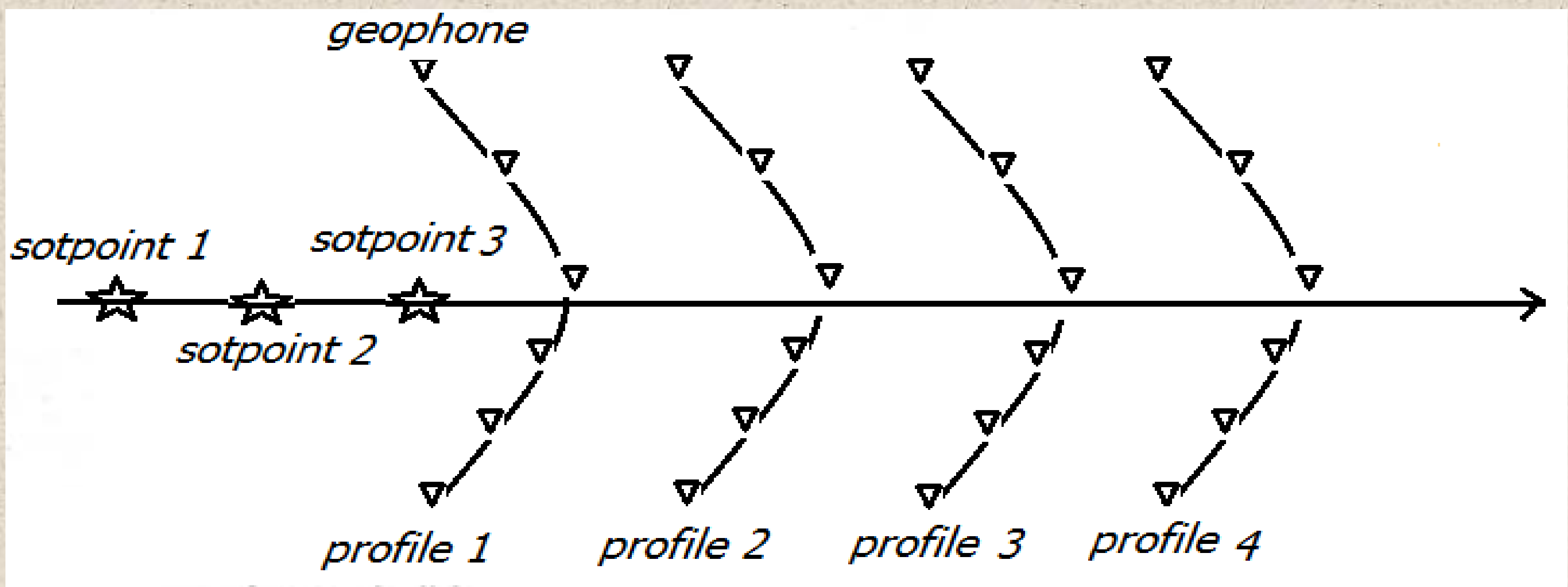
Seismic Record



II. Arc Shooting

In this arrangement, the shotpoints represent the centers of arcs along which the geophones are planted.

The arc shooting arrangement is used for the purpose of mapping lateral variations in the subsurface.



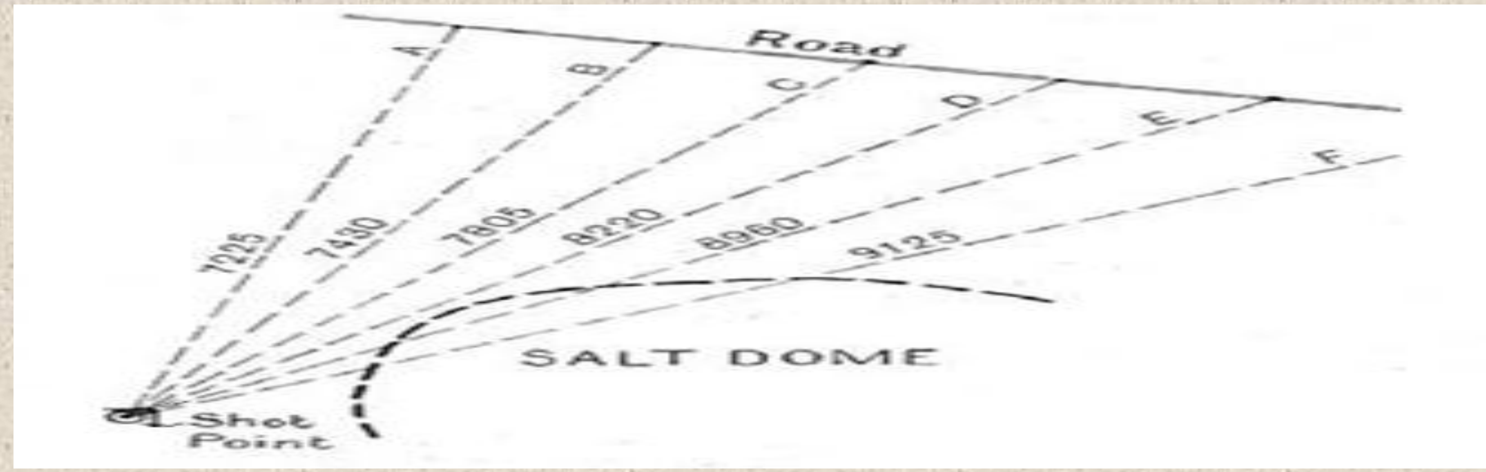
III. Fan Shooting

This is the first field arrangement used in the early seismic exploration words. In this arrangement, the geophones are placed along circles with shotpoints at the centre of the related circles.

In particular, the arrangement was effective in detecting salt domes:

In the absence of salt domes, the arrival times for the seismic wave will be the same for all geophones. While, in the presence of salt dome, the arrival times for the waves that cross the salt dome will be shorter (the wave travels faster) than other arrivals.

Therefore, by plotting first arrivals (the faster waves), it will be possible to locate the anomalous zone.



General Considerations on Seismic Refraction Method

I. RELATED TO SURVEY DESIGN

- Length of seismic profiles: the profile length should be 3-5 times the investigated depth.

- Distance between successive geophones: controls
  1) depth of penetration; 2) mapping of the surface layer; 3) details about the topography of the refractor

II. <u>RELATED TO THE INTERPRETATION PROCESS</u>

- Travel time – distance graph:
  1) same scale to be used for both time and distance axis;
  2) same symbols for time arrivals from the same shotpoint

- Use of specialized software packages:
  1) reduces considerably the time needed for the interpretation;
  2) provides excellent graphical outputs

REVIEW:
SEISMIC REFRACTION METHOD

Seismic Reflection Method

I. Introduction and General considerations

- Seismic reflection is the most widely used geophysical technique. It can be used to derive important details about the geometry of structures and their physical properties.

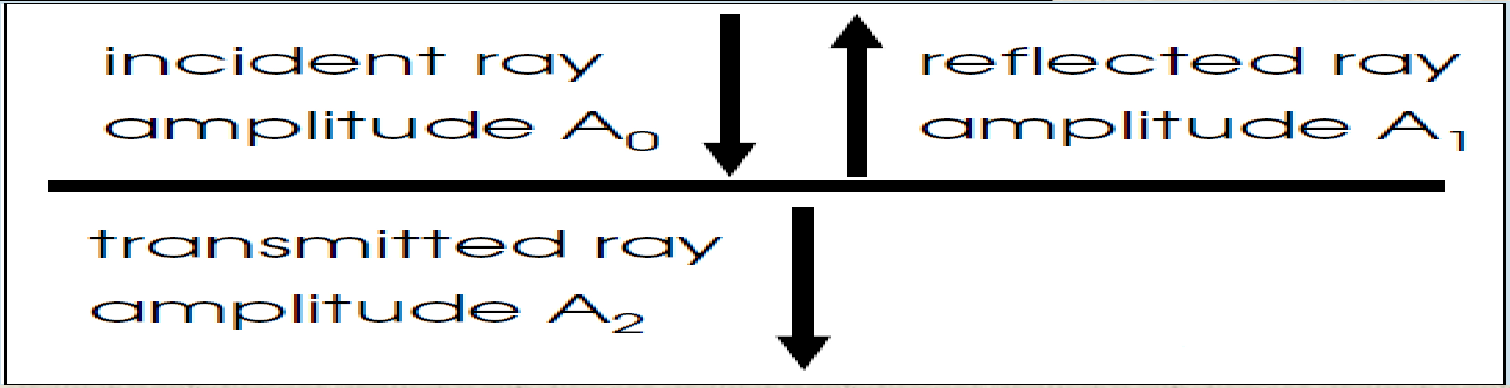
- Major fields of application of Seismic reflection include:
  - ✓ hydrocarbon exploration,
  - ✓ research into crustal structure with several kilometers of depths of penetration,
  - ✓ Engineering and environmental investigations (depth <200m),
  - ✓ mapping structural features such as shallow faults, buried valleys and Quaternary deposits,
  - ✓ Hydrological studies of aquifers.

- The basic principle of the seismic reflection technique application is to measure the time taken for a seismic wave that travels from a source down into the ground where it is reflected back to the surface where it can be detected by a receiver (geophone):

  - The measured time is known as the two way time (TWT).
  - The basic issue in seismic reflection interpretation is the conversion of the measured two way time into depth. Although the two way time (TWT) is known (measured), still there are two unknown parameters; these are: depth and velocity. Velocity is considered as the parameter that most affects the conversion of the two way time into depth.

II. Reflection and Transmission of seismic waves in layered media

- At an interface between two rock layers there is generally a change in propagation velocity resulting from difference in physical properties of the two layers. At such an interface, the energy within an incident seismic wave is partitioned into transmitted and reflected waves.
- The relative amplitudes of the transmitted and reflected waves depend on: the velocities (V), densities (ρ) and the angle of incidence.
- The total energy of the transmitted and reflected waves must be equal to the energy of the incident ray.



- The amount of energy transmitted through the interface is <u>inversely</u> proportional to the acoustic impedance defined by:

$$Z = \rho V \qquad = \text{acoustic impedance}$$

Where:
Z: acoustic impedance
V: velocity
ρ: density

This means that the smaller the contrast in acoustic impedance across the rock interface the greater is the portion of the transmitted energy.

- The more energy is reflected, the greater is the contrast. This is expressed by the Reflection Coefficient, R, given by:

$$R = A_1 / A_0$$

$$R = \frac{(\rho_2 v_2 - \rho_1 v_1)}{(\rho_2 v_2 + \rho_1 v_1)} = \frac{(Z_2 - Z_1)}{(Z_2 + Z_1)} \qquad -1 < R < 1$$

Where:

R: reflection coefficient

A_1: amplitude of the reflected wave

A_0: amplitude of the incident wave

Z: acoustic impedance

V: velocity

ρ: density

Negative values of the reflection coefficient indicate a phase change of 180° in the reflected wave.

- The transmission coefficient, T, is given by:

$$T = A_2 / A_0$$

$$T = 2\frac{Z_1}{(Z_2 + Z_1)}$$

Where:

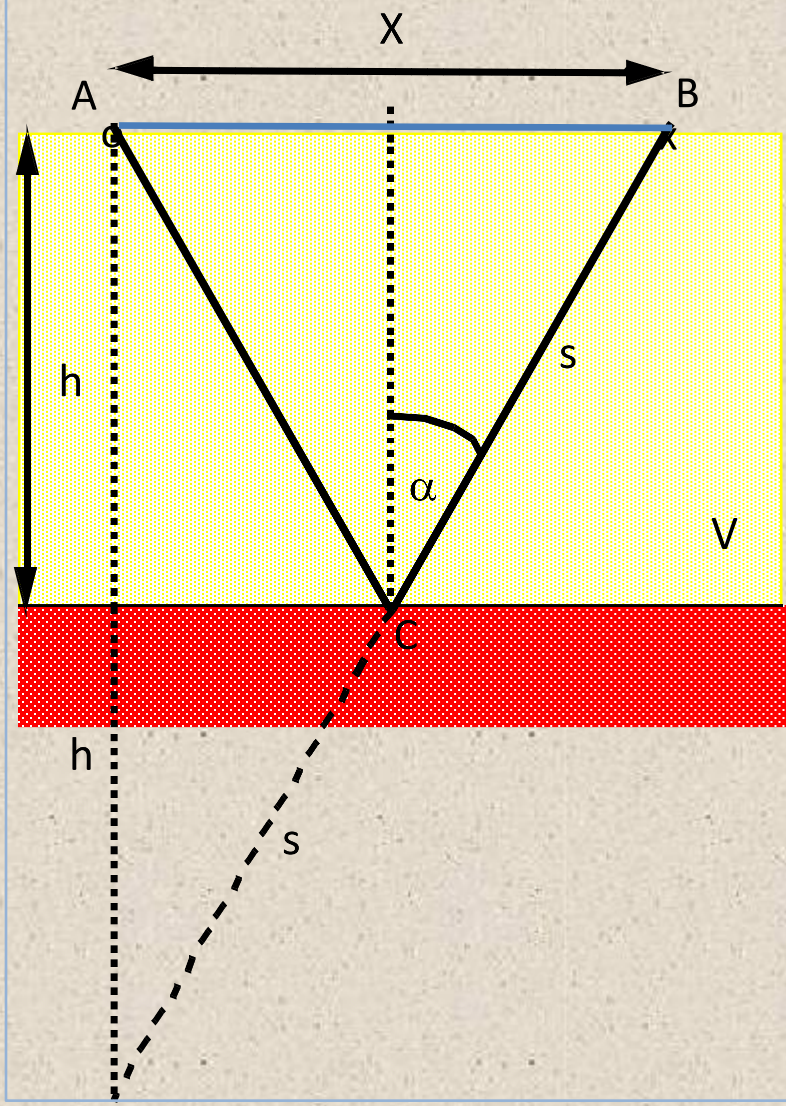
T: transmission coefficient

A_2: amplitude of the transmitted wave

A_0: amplitude of the incident wave

Z: acoustic impedance

III. The Case of Single Horizontal Reflector



The travel time equation of a reflected wave from a shot point to a receiver (geophone) located at a horizontal offset X can be derived as follows:

$$4S^2 = 4h^2+x^2 = t^2v^2$$

$$t^2=(4h^2+x^2)/v^2$$

$$\boxed{\mathbf{t = (4h^2+X^2)^{1/2}/V} \quad \ldots\ldots\ldots(1)}$$

There are two unknown parameters in the above equation, these are: Velocity (V) and depth (h).
Now, Eq. 1 can be written in the form:

$$\boxed{\frac{t^2v^2}{4h^2} - \frac{x^2}{4h^2} = 1}$$

Hyperbola

The Intercept Time

By measuring many reflection times, t, at different offsets, x, it will be possible to calculate the depth, h, and the velocity, V (See the figure below).

By substituting X= 0 in Eq. 1:

$$\boxed{t = (4h^2+X^2)^{1/2}/V \;........(1)}$$

we obtain:

$$\boxed{t_\circ = 2h/V \;....................(2)}$$

Eq. 2 is the travel time equation of a vertically reflected wave (intercept on the time axis of the time – distance curve).

By squaring Eq. 1 and substituting Eq. 2 in Eq. 1, we obtain: Velocity can be determined using Eq. 3.

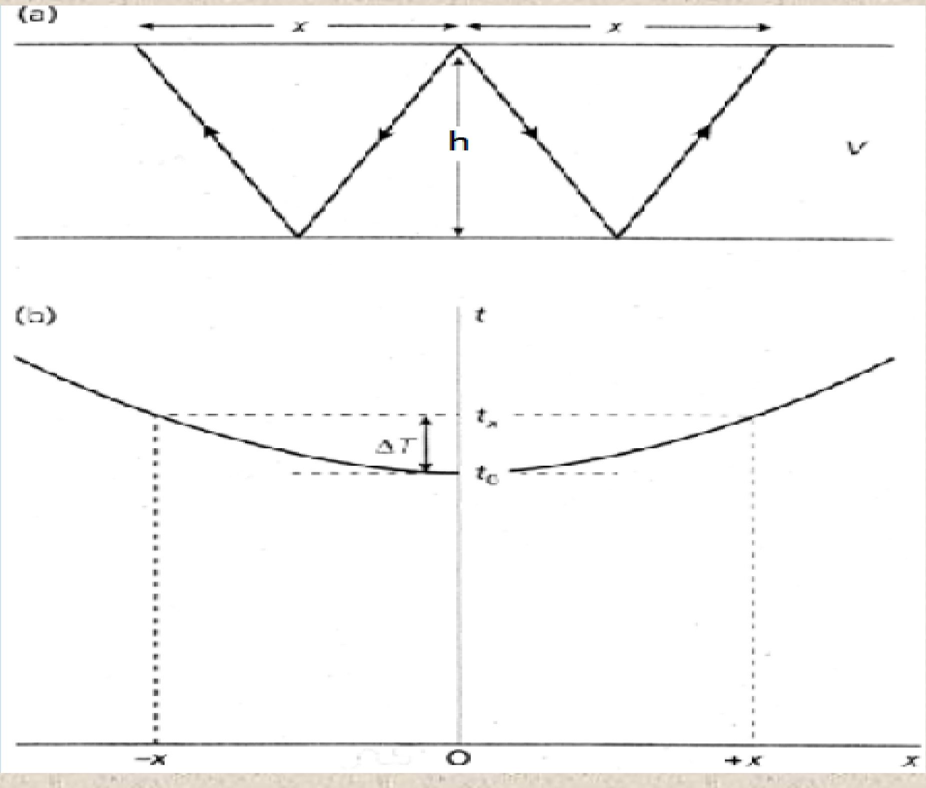


Fig. : (a) Section through a single horizontal layer showing the geometry of reflected ray paths and (b) time-distance curve for reflected rays from a horizontal reflector.
ΔT= normal moveout (NMO).

By squaring Eq. 1 and substituting Eq. 2 in Eq. 1, we obtain:

$$t^2 = 4h^2/ V^2 + X^2/V^2$$

$$\boxed{t^2 = t_\circ^2 + X^2/V^2 \text{ ....}\mathbf{(3)}}$$

<u>Velocity can be determined using Eq. 3.</u>

CALCULATION OF THE VELOCITY

- Plot t^2 against x^2

The graph will produce a straight line of slope $1/v^2$. The intercept on the time axis will give the vertical two way travel time, t_0 , from which the depth to the reflector can be found.

This method is unsatisfactory, since the values of x are restricted.

- A much better method of determining velocity is by considering the increase of reflected travel time with offset distance, the moveout.

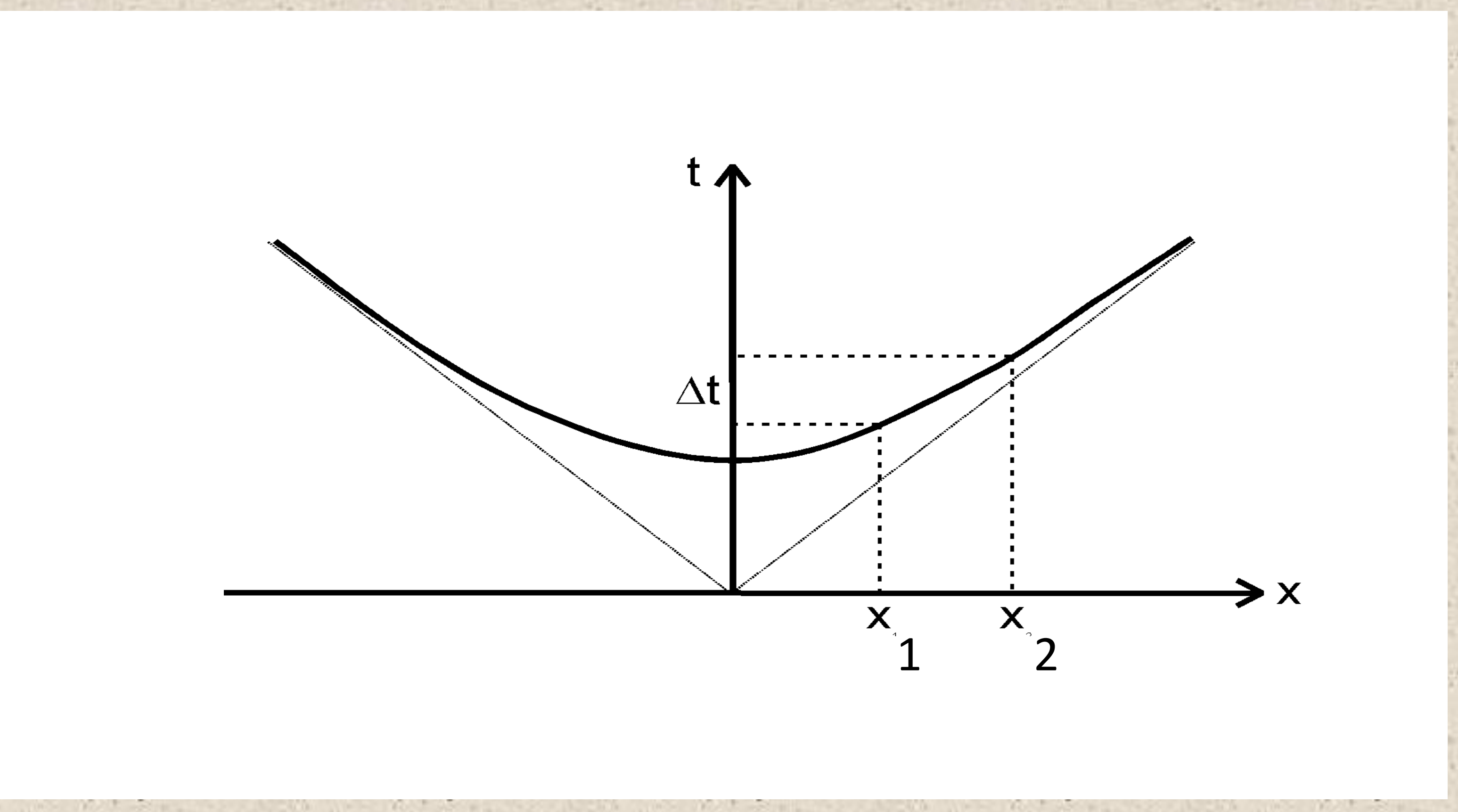
MOVEOUT

- Moveout is defined as the difference between travel times t_1 and t_2 of reflected-ray arrivals recorded at two offset distances x_1 and x_2.

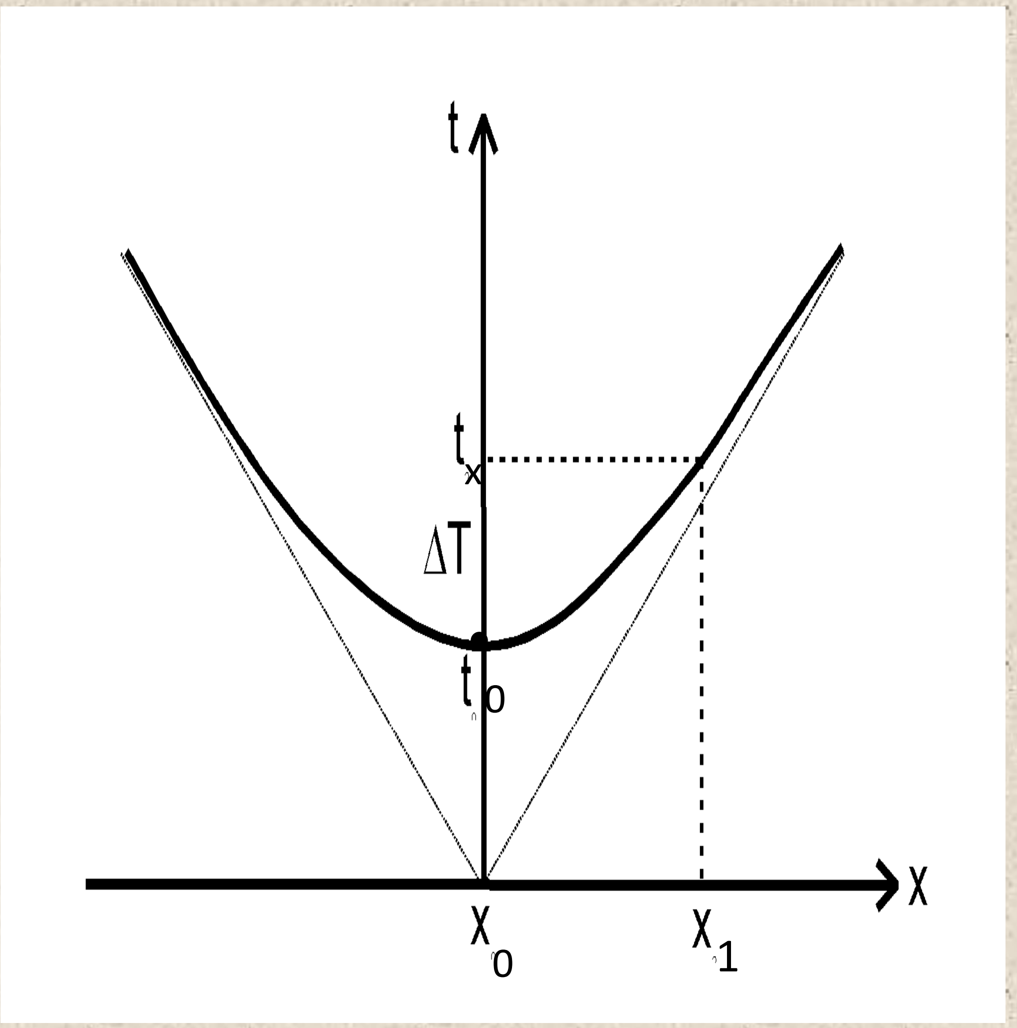
$$t_2 - t_1 = \frac{x_2^2 - x_1^2}{2 v^2 t_0}$$

- Normal moveout (NMO) at an offset distance x is the difference in travel time ΔT between reflected arrivals at x and at zero offset. (see Figure)

$$\Delta T = t_x - t_0 \approx \frac{x^2}{2 v^2 t_0} \Rightarrow v = \frac{x}{(2 t_0 \Delta T)^{1/2}}$$



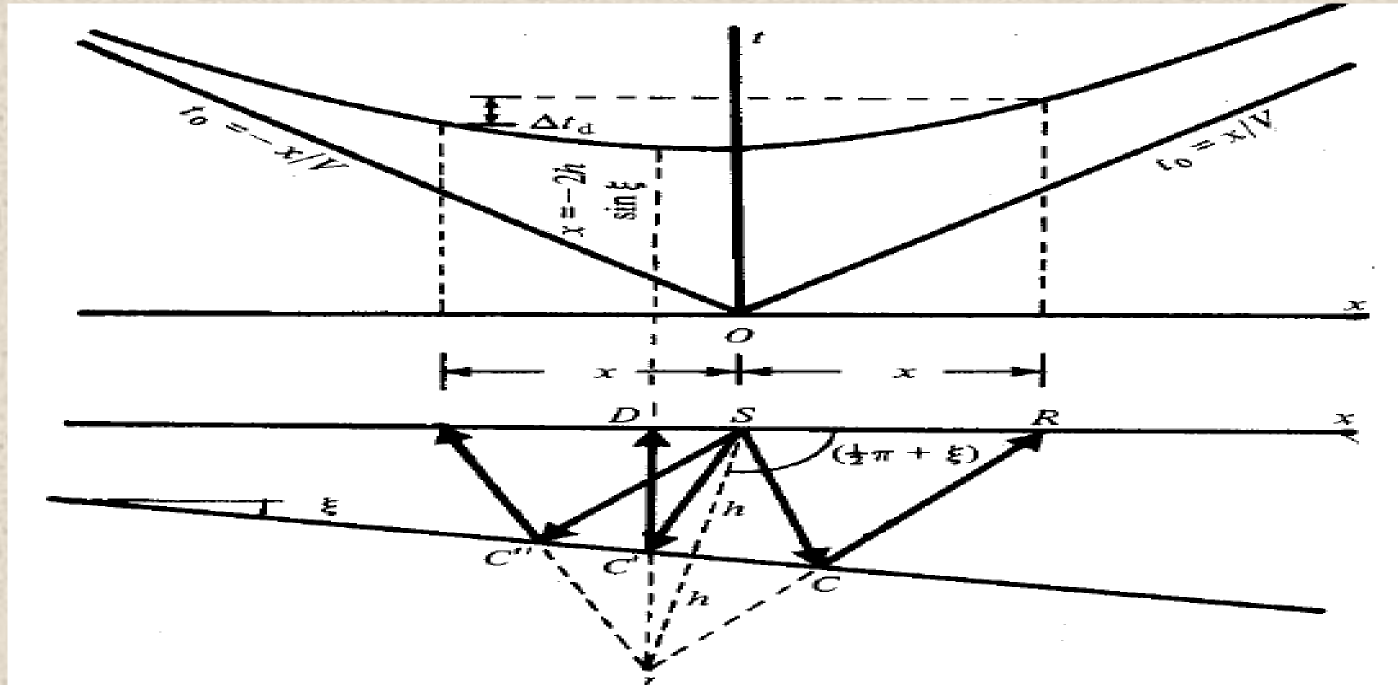
MOVE OUT: Difference in travel time $t(x_1)$ and $t(x_2)$: $t_2 - t_1 \approx \dfrac{x_2^2 - x_1^2}{2v^2 t_0}$



Normal Moveout: Difference in travel time t_0 and t(x): $\Delta T = t_x - t_0 \approx \dfrac{x^2}{2v^2 t_0}$

Travel Time for Single Plane Dipping Layer

The length of the reflection path, SCR, from Source to Receiver is the same as path IR, from an imaginary source, I, obtained by reflecting the surface source, S, in the interface.



The Cosine rule in the triangle IRS gives the travel time, T, as VT, which is the distance travelled by the seismic reflection:

$$\cos(90+\xi) = -\sin\xi$$

$$c^2 = a^2 + b^2 - 2ab\cos(C)$$

$$(IR)^2 = V^2T^2 = x^2 + 4h^2 - 4hx\cos(90+\xi)$$
$$= x^2 + 4h^2 + 4hx\sin\xi$$

By combining X terms using the complete square rule, we obtain:

$$V^2T^2 = (x + 2h\sin\xi)^2 + 4h^2 - 4h^2\sin^2\xi$$

$$V^2T^2 = (x + 2h\sin\xi)^2 + 4h^2\cos^2\xi$$

$$\boxed{\frac{V^2T^2}{(2h\cos\xi)^2} - \frac{(x + 2h\sin\xi)^2}{(2h\cos\xi)^2} = 1}$$ **..Eq. (1)**

Equation 1 represents a asymmetrical (not symmetrical) hyperbola with it's apex shifted from X = 0. Apex is displaced towards UPDIP direction to:

X = - $2h\sin\xi$

Estimation of Reflector Dip

By measuring travel times at two locations offset by same distance ΔX to either side of the source, an estimation for the dip can be obtained by:

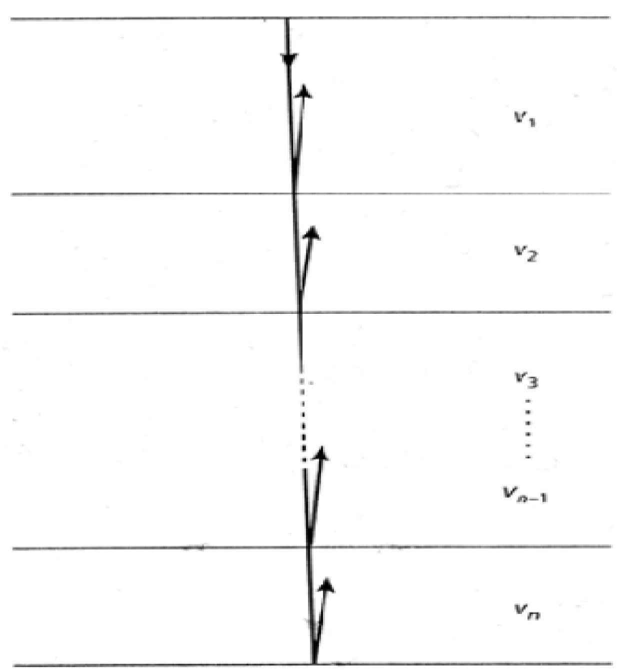
$$\sin \xi \approx \frac{1}{2} V \frac{\Delta t_d}{\Delta x}$$

Where:

Δt_d is the difference in observed travel times.

TYPES OF VELOCITIES

I. Interval Velocity:



$\mathbf{v}_i$ = interval velocity
$\mathbf{z}_i$ = thickness of the interval
τ_i = one way travel time

$$\mathbf{v}_i = \frac{\mathbf{z}_i}{\tau_i}$$

II. Average Velocity:

The interval velocity may be averaged over several depth intervals to yield an average velocity $\bar{v}$.

$$\bar{v} = \frac{\sum_{i=1}^{n} z_i}{\sum_{i=1}^{n} \tau_i} = \frac{\sum_{i=1}^{n} v_i \tau_i}{\sum_{i=1}^{n} \tau_i} \qquad \text{or} \qquad \bar{v} = \frac{Z_n}{T_n}$$

Z_n = total thickness of the top n layers
T_n = total one way travel time through the n layers

III. ROOT MEAN SQUARE (rms) Velocity:

$$V_{rms} = \left[\frac{\sum_{i=1}^{n} V_i^2 t_i}{\sum_{i=1}^{n} t_i}\right]^{\frac{1}{2}}$$

Where:

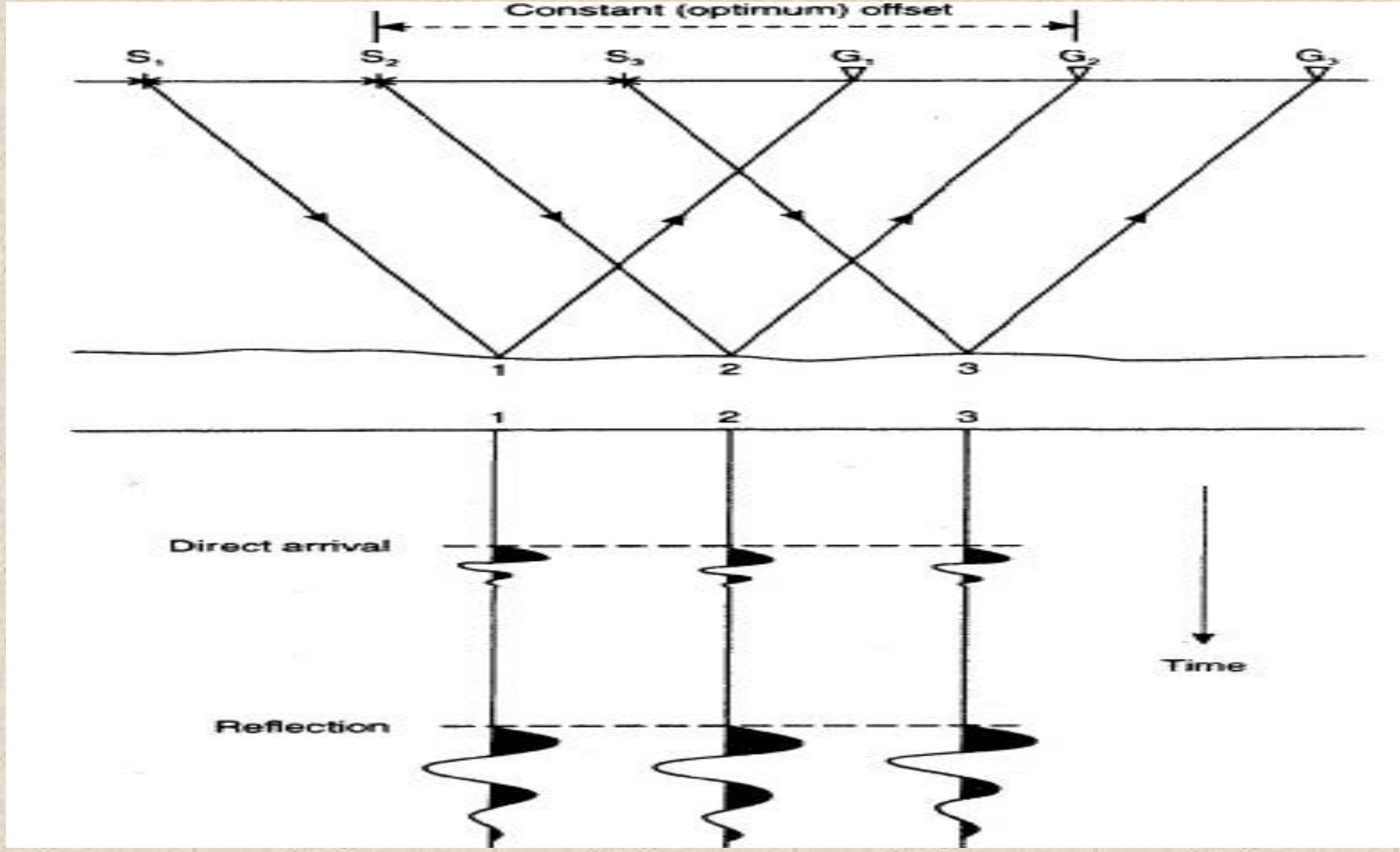
$t_i = 2\, Z_i / V_i$ → the vertical travel time in the *i* th layer

V_i = interval velocity

SEISMIC REFLECTION FIELD PROCEDURES & DATA ACQUISITION

I. Field Procedures

Seismic reflection profiling obtains a cross-section through subsurface by recording data continuously along a surface profile.



- *Single-channel Seismic Profiling (constant offset)*: In case of shallow reflection profiles, the simplest survey is to use a source and a single geophone (receiver). In this procedure:
  - source and receiver are both moved along profile by same amount between shots
  - plotting successive shots side by side creates a reflection profile of the subsurface.
  - used in engineering and hydrogeological surveys.

- *Multichannel Seismic Profiling (single fold)*:

Multichannel Recording: Single-Channel reflection surveys are subject to noise. Combining multiple reflections from single subsurface location allows attenuation of this noise.

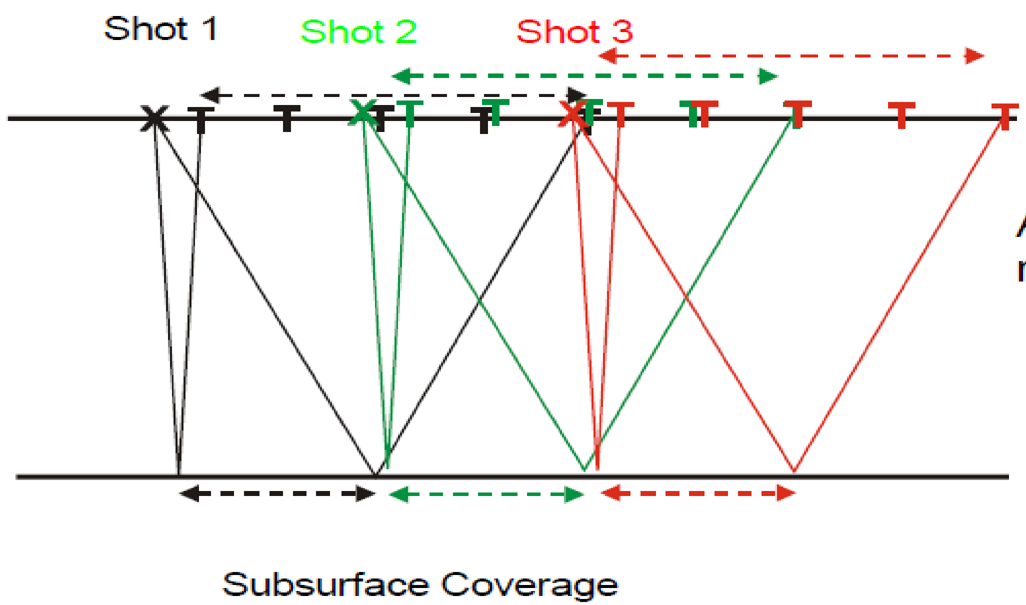
What recording channel means? One seismogram recorded by the instrumentation. But the channel may be connected to one or more geophones (receivers) on the ground.

Recording a shot with multiple channels produces many seismograms per shot and allows attenuation of many noise types during processing. 12 – 2000 channel recording is common.

In the *Multichannel Seismic Profiling,* entire recording configuration, shot and receivers, are moved along the profile a distance, ΔS, between shots.

- Shot and receivers are all moved same distance along a profile.
- In practice, and to save time, twice as many receivers as required are laid out. Only those needed for each shot are made alive
- Recording cable and receivers are brought to front from rear during survey

Example: Single-Fold Reflection Profile

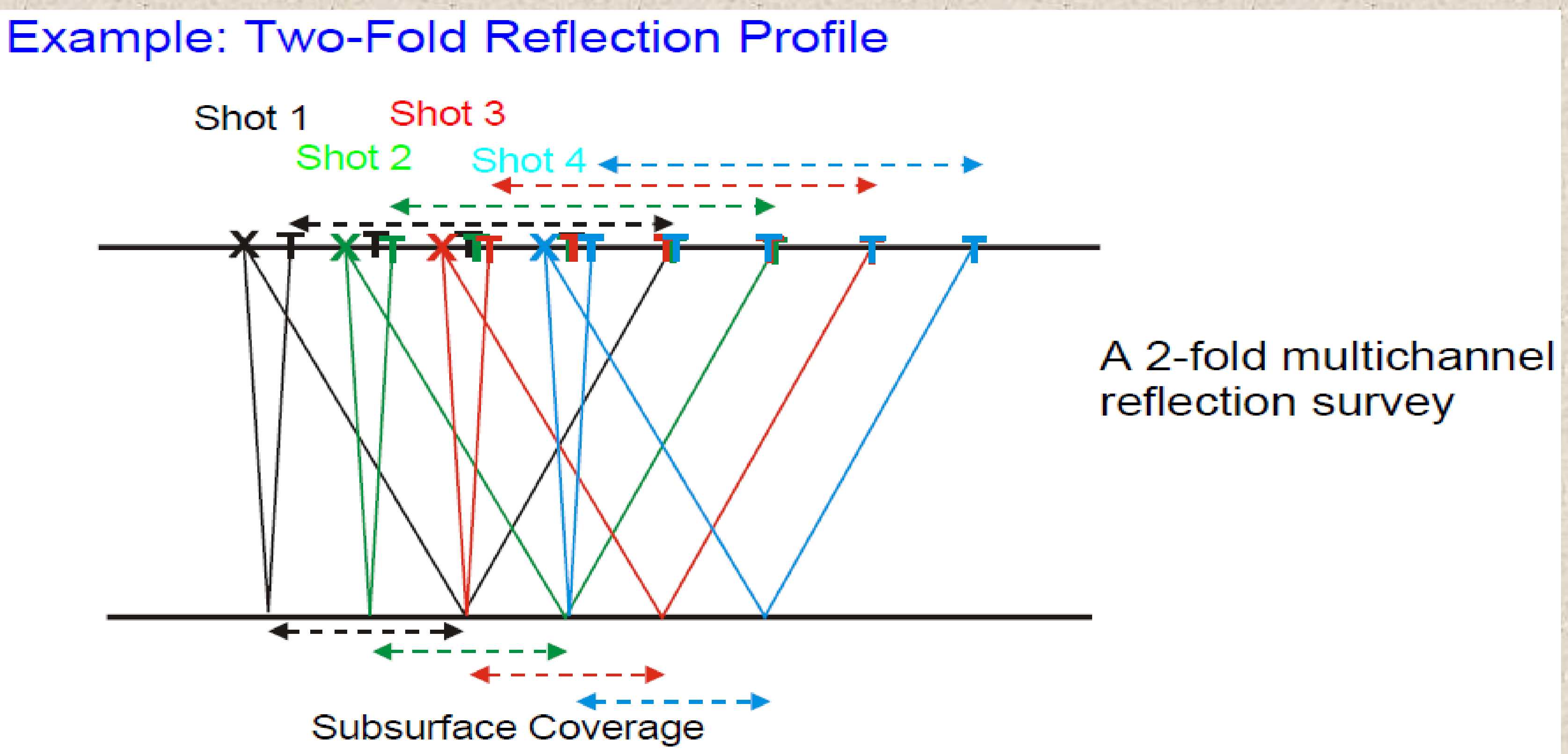


A 1-fold multichannel reflection survey

- Each shot reflects signal from half the spread length on a plane horizontal interface in the subsurface
- Spacing between reflection points on interface is half the receiver spacing, Δr, at the surface.
- Complete coverage of the subsurface interface can be achieved by moving shot and receivers up by half the spread length plus geophone spacing.

What is a FOLD?

Fold is the multiplicity of subsurface reflection coverage, i.e. the number of reflections recorded from each subsurface point in a multichannel survey. In the example above, fold is one because only ONE reflection from each subsurface position, i.e. one reflection every $\Delta r/2$.



➢Average fold along a multichannel profile is called NOMINAL FOLD, and is given by:

$$f = \frac{n\Delta r}{2\Delta s}$$

where: n is the number of recording channels, Δs and Δr are the shot and the receiver spacing respectively.

➢Multifold reflection profiling was developed in the early 1960s, and has been used extensively by the oil industry.

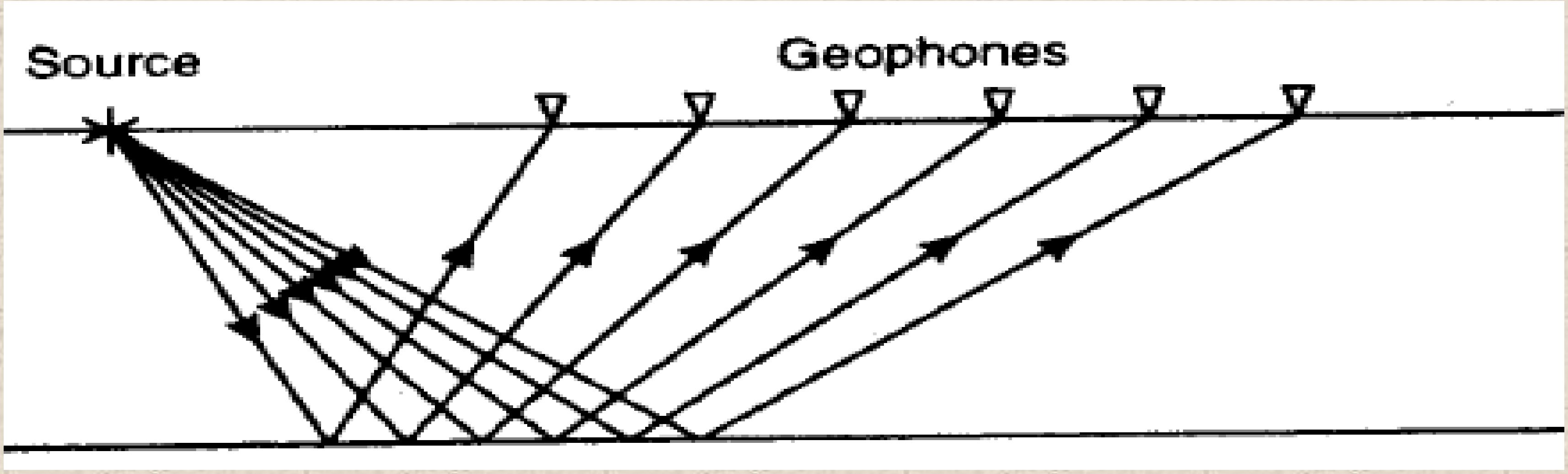
➢ Surveys up to 100 fold or more are recorded today.

<u>II. Gathering of Seismic Data</u>

Large volume of seismic data can be recorded and be organized in different ways. A GATHER is the name for a <u>collection</u> of <u>seismic traces</u>.

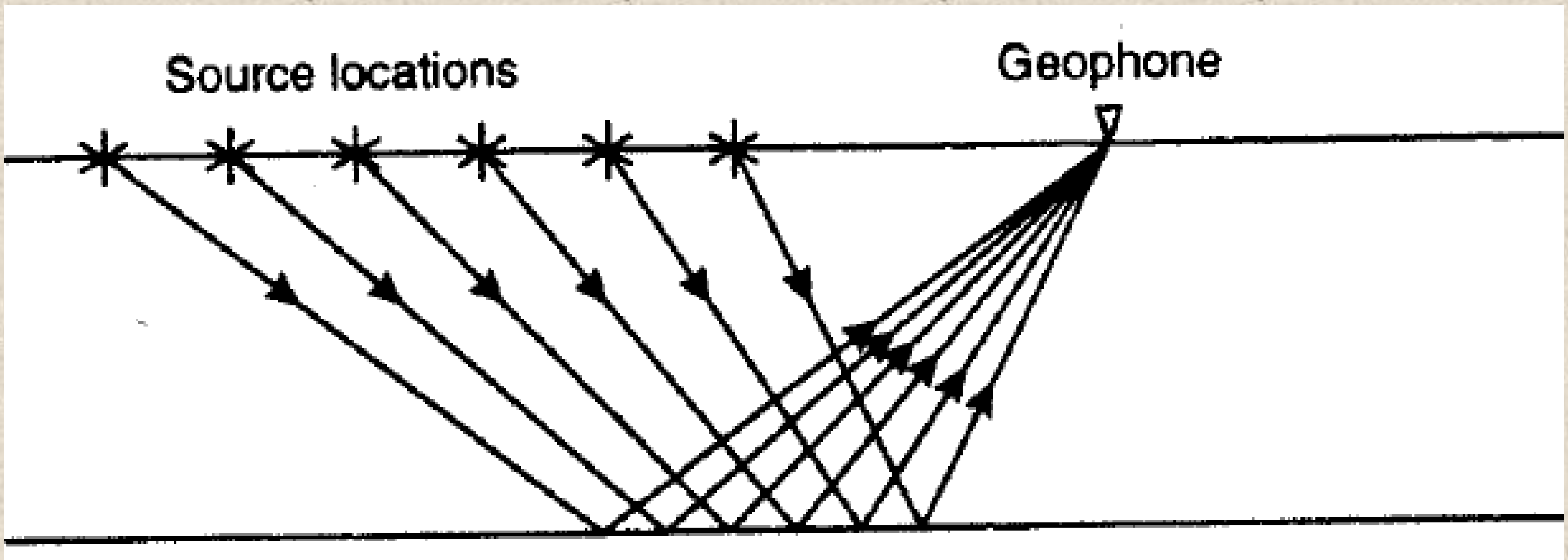
There are several trace gathers, these are:

- ***Common Shot Gather***: a collection of seismic traces recorded at several receivers (geophones) from single shot. This is the configuration in which seismic data are acquired in the field.



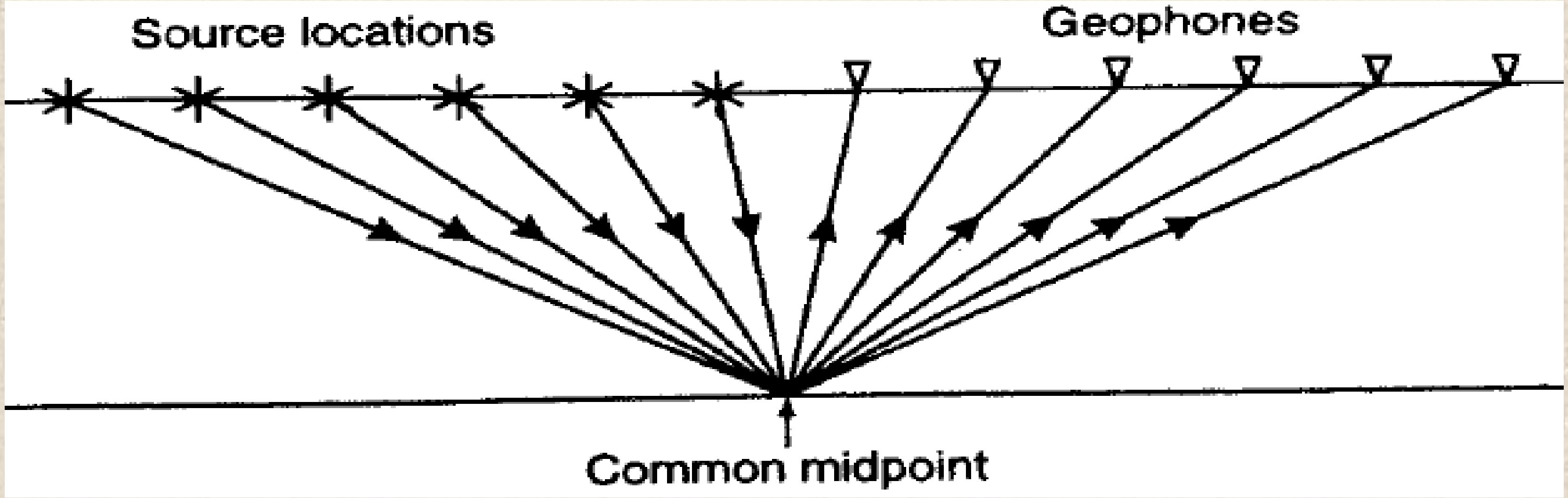
❑ ***Common Receiver Gather***:

A collection of seismic traces corresponding to several shots recorded at a single receivers (geophone).



❑ *Common Midpoint (CMP) Gather*:

A collection of seismic traces in which the shot and the receiver are symmetrically distributed about the same midpoint location.



The <u>*common Depth Point (CDP)*</u> is the point on a <u>plane horizontal interface</u> from which all the reflections in a *Common Midpoint (CMP) Gather* are generated.

➢ **The *Common Midpoint (CMP) Gather* is fundamental in seismic reflection data processing for two reasons:**

1) The variation of travel time with offset, the moveout will depend only on the velocity of the subsurface layers (horizontal uniform layers).

➔ The subsurface velocity can be derived.

2) The reflected seismic energy is usually very weak. It is imperative to increase the signal-noise ratio of most data.

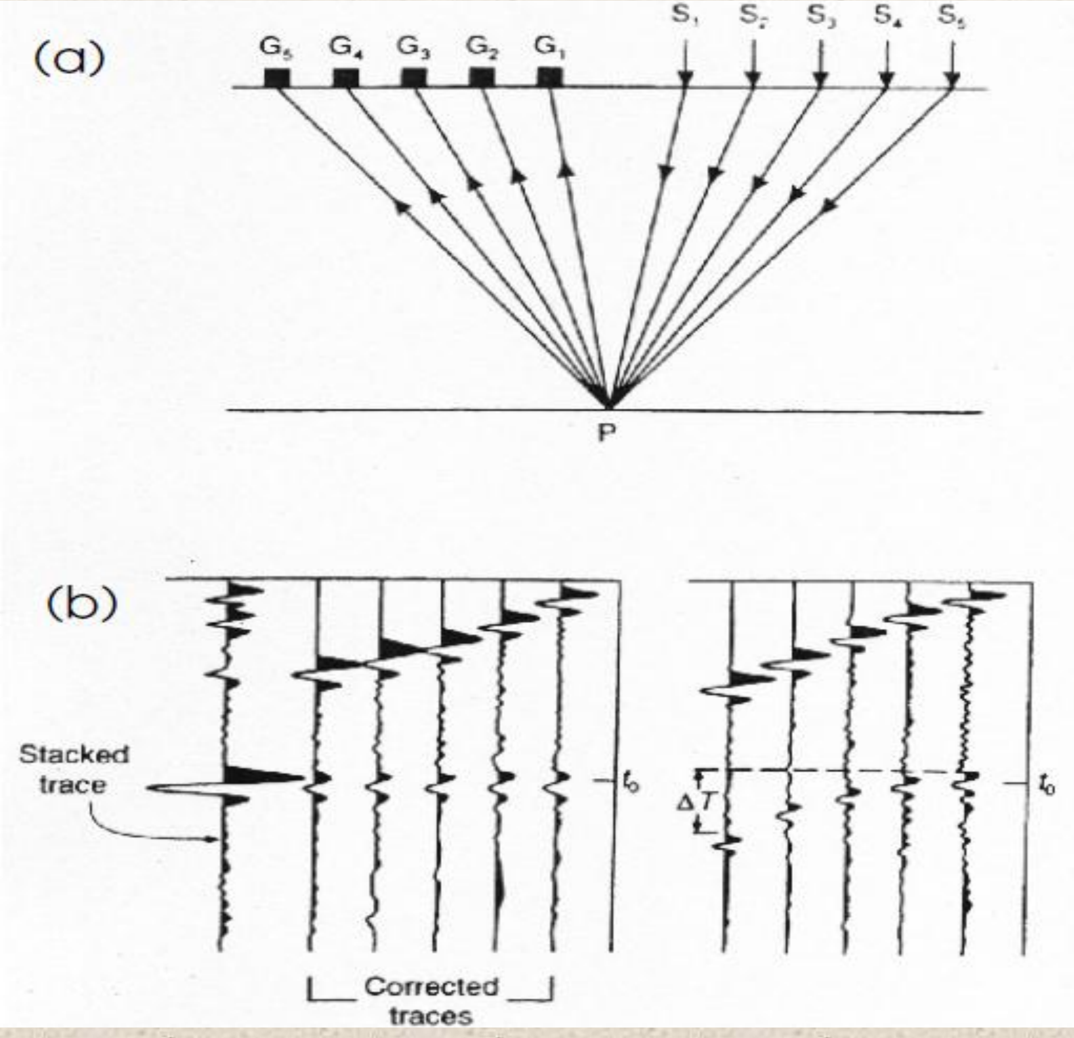
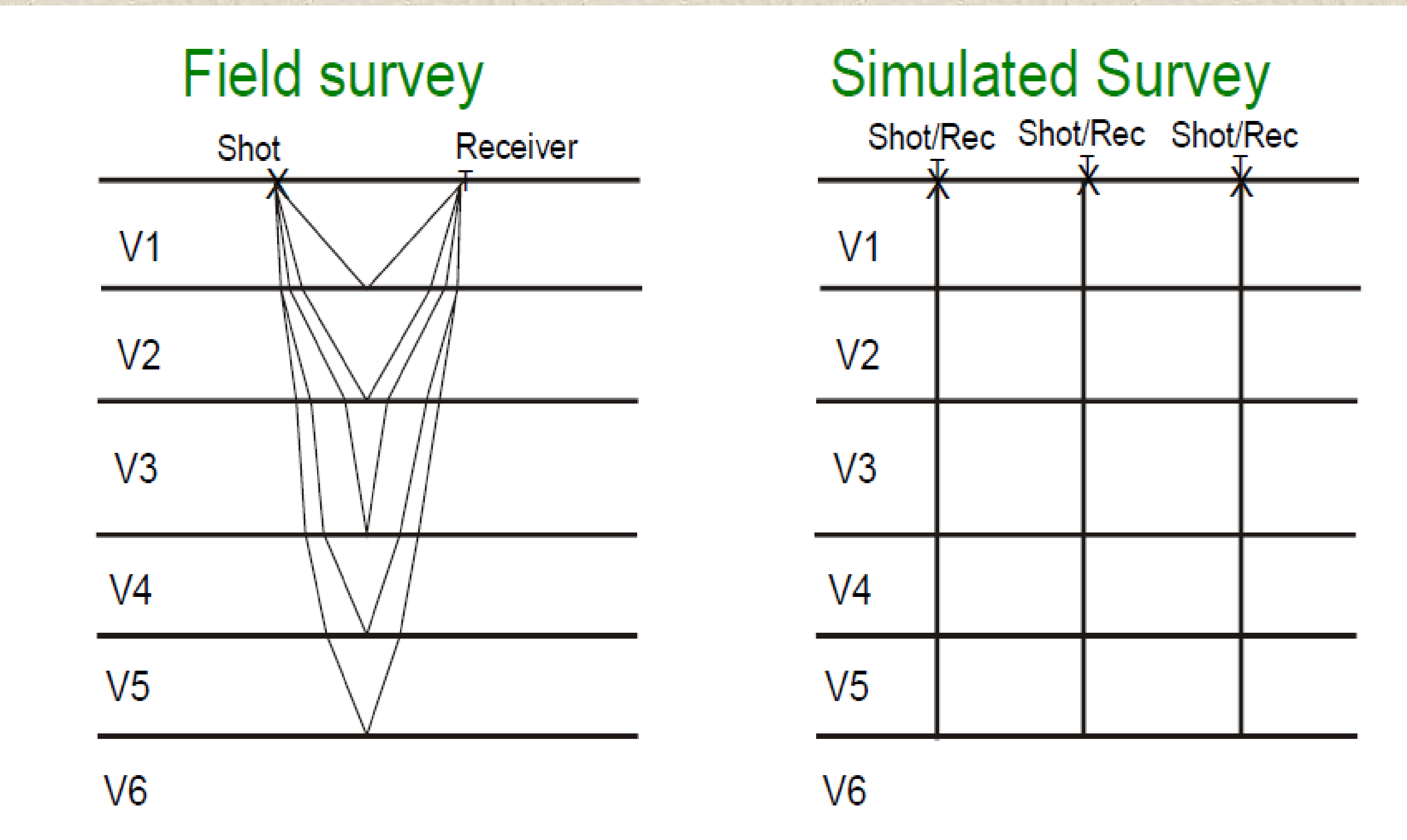


Fig. : Given the source-receiver layout and corresponding ray-paths for a common depth point spread, shown in (a), the resulting seismic traces are illustrated in (b), uncorrected (on the right), (corrected on the left) – note how the reflection events are aligned – and the final stacked trace.

Display & Processing of Seismic Reflection Data

GENERAL CONSIDERATIONS

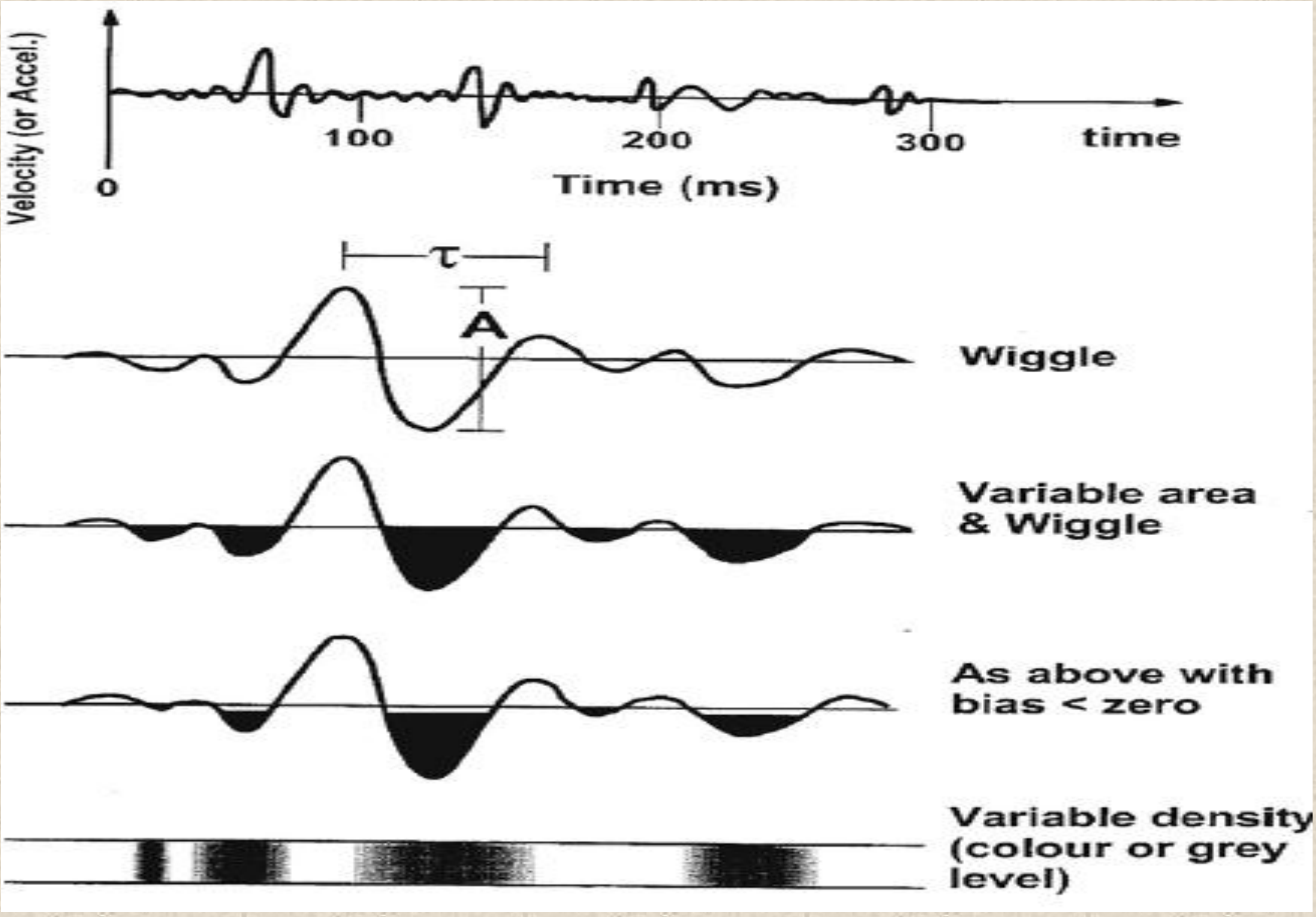
- Uses secondary reflection arrivals, which must be extracted by processing of raw seismic recordings
- Reflections generated by both increase and decrease in seismic impedance
- Maximum source-receiver offset usually less then target depth with angles of incidence less than 30^o
- Reflection data processed to simulate a survey recorded with coincident sources and receivers: subsurface "cross-section"


Field survey
Shot
Receiver
V1
V2
V3
V4
V5
V6
Simulated Survey
Shot/Rec
Shot/Rec
Shot/Rec
V1
V2
V3
V4
V5
V6

DISPLAY OF SEISMIC DATA

- Geophone measures velocity of ground vibration
- Hydrophone measures pressure variations in water
- Vibration is from some rest value, zero for ground velocity, and seismogram shows this variation from rest

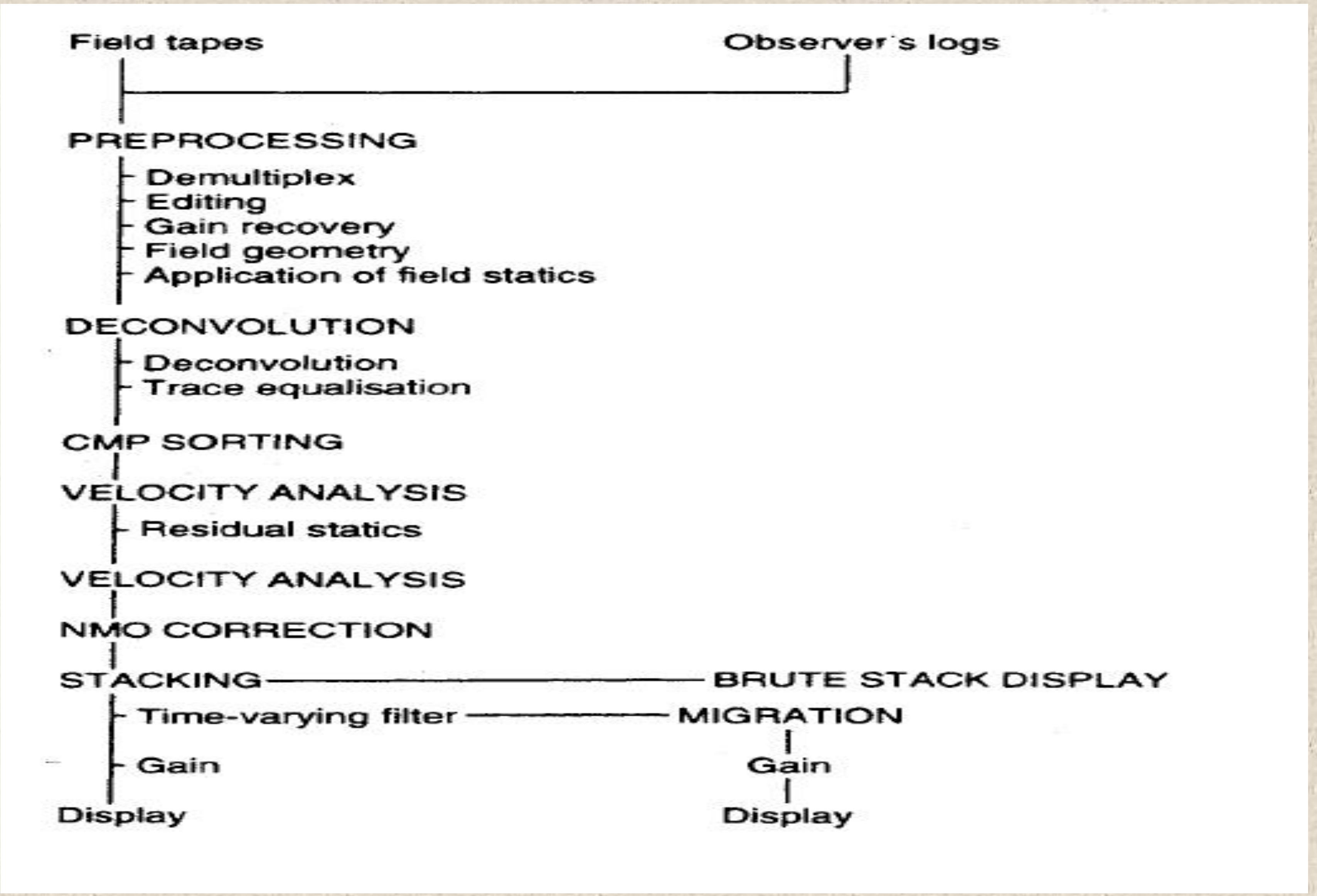
❑ Seismograms can be displayed is various ways:



Why do we need data processing?

- **To solve acquisition problems**
- **To enhance the appearance of the data**
- **To give a true view of the subsurface**

PROCESSING OF SEISMIC DATA



PRE-PROCESSING OF REFLECTION DATA

I. Demultiplexing

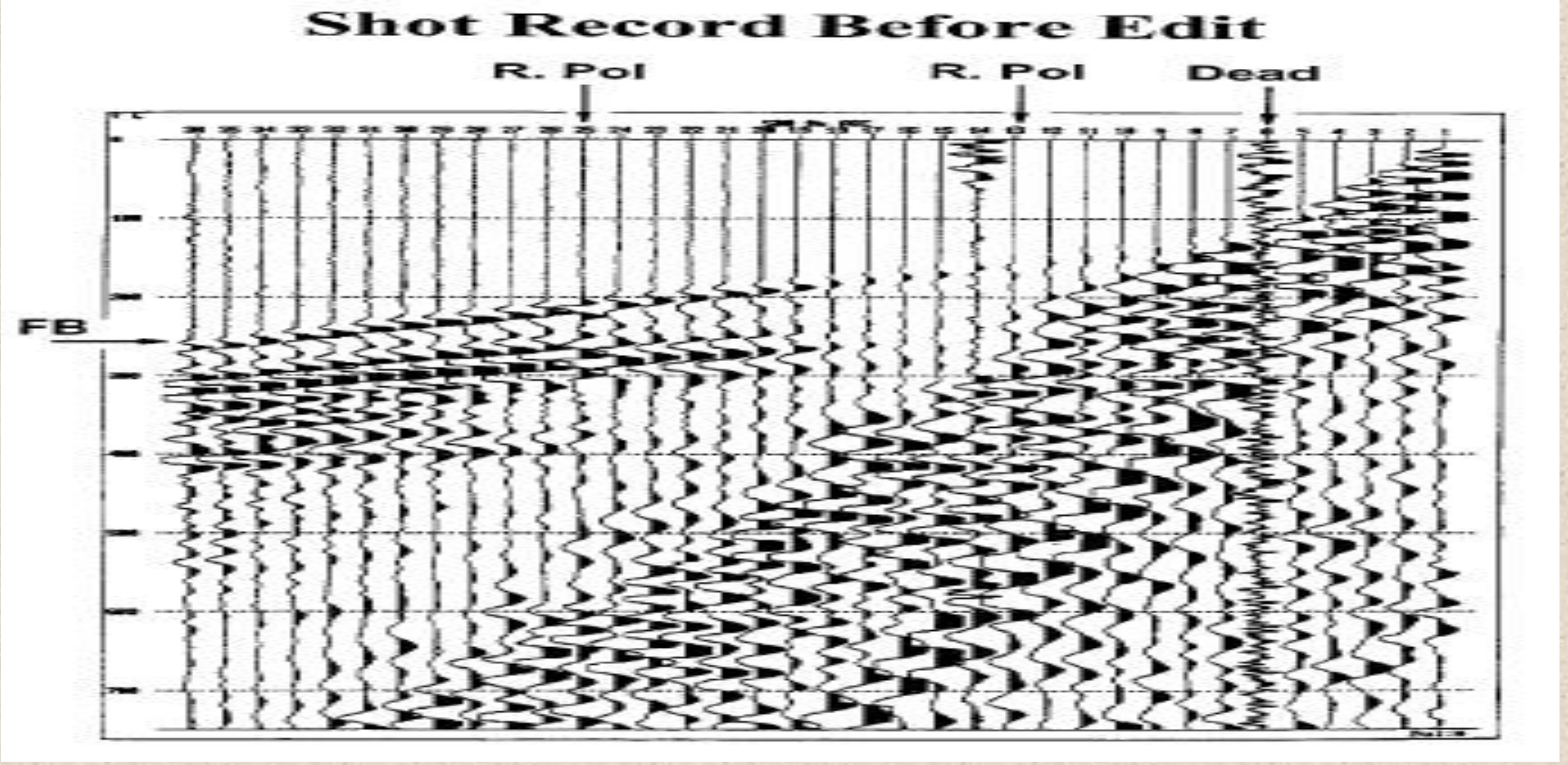
Prior to mid-1980s, a shot gather was recorded *multiplexed* → adjacent samples came from adjacent traces at the same time.

Demultiplexing means reorganizing the data such that successive samples on tape represent successive time samples for each seismogram.

Now most oil industry seismic data is recorded demultiplexed.

	Sample 1	Sample 2	Sample 3		Sample 4
Channel 1	a_1	a_2	a_3	$\cdots$	a_n
Channel 2	b_1	b_2	b_3	$\cdots$	b_n
Channel k	k_1	k_2	k_3	$\cdots$	k_n

II. Trace Editing

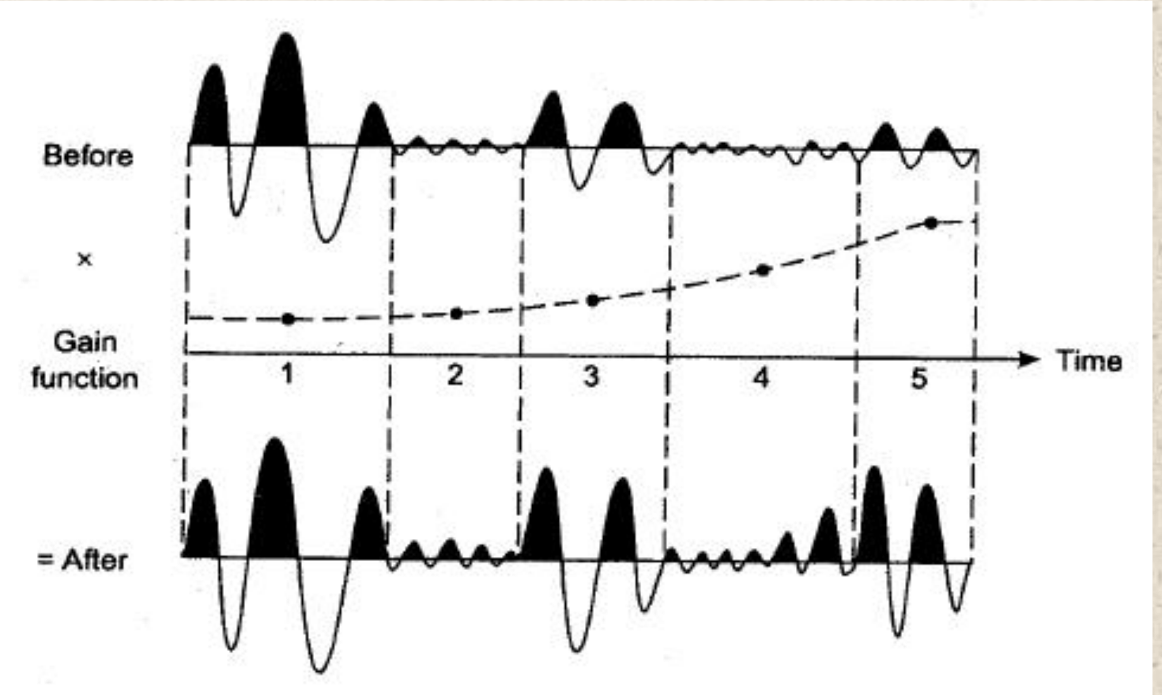


Removal of traces contaminated with high amplitude noise, e.g. due to geophones close to machinery, or that did not record data.

III. Gain Recovery

Seismic waves decrease in amplitude as they propagate further into the Earth:

Reflections recorded at late times have lower amplitude than those at early times.



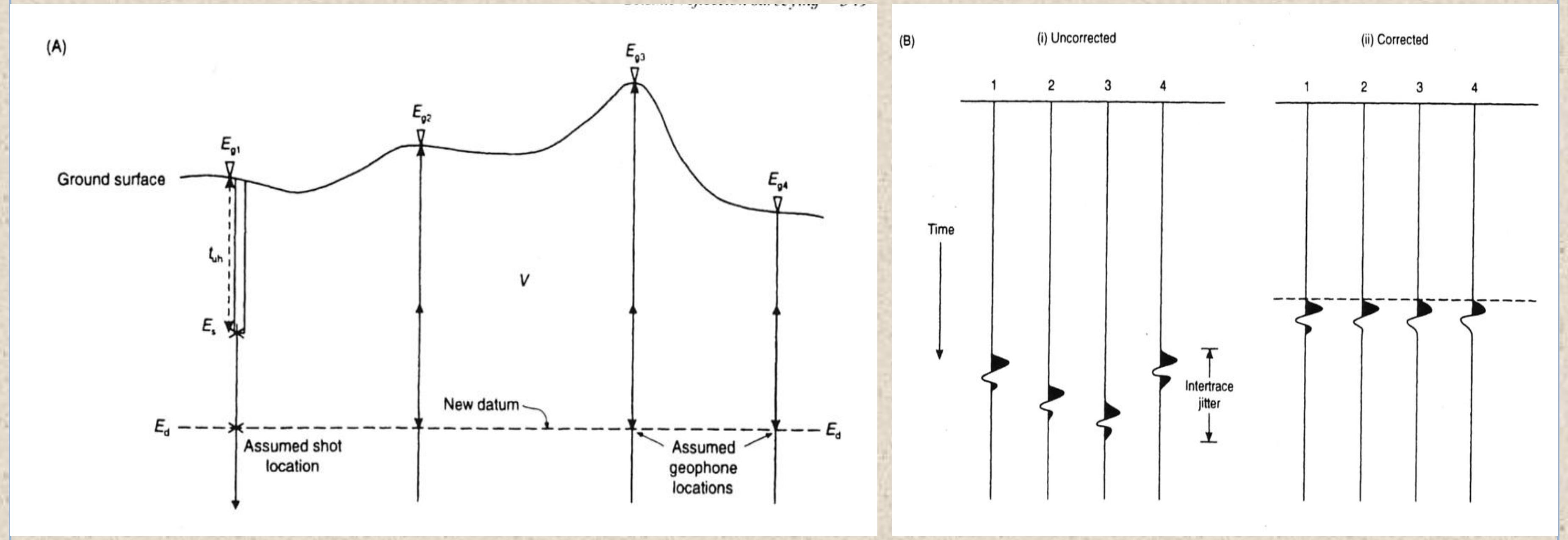
IV. Elevation Statics

Vertical travel time through the near-surface layer varies due to changes in elevation of shot or receiver; it causes:

- misalignment of reflections
- after NMO correction, reflections do not stack correctly.

Statics

- The time shift is added to a trace to correct it to a recording made with source and receiver at a specified datum elevation.
- It is constant for each trace.



$$t_s = (E_s - E_d)/V$$
$$t_g = (E_g - E_d)/V$$
$$t_e = t_s + t_g$$

Where:
t_g: Time correction to each trace
t_s: Time correction to source depth
t_e: Total time correction

- Each trace to be shifted by the amount t_e to line up reflectors.

- To make corrections we need to know the velocity of the surface weathered material. Velocity can be determined in two ways:

1. **Uphole traveltime:** Using: $V = d/t_{uh}$

2. **Refraction survey:**

Used to determine near surface velocities and variations in the thickness of the weathered layer.

Static corrections: An Example

(a) Pre – static correction

(b) Post – static correction

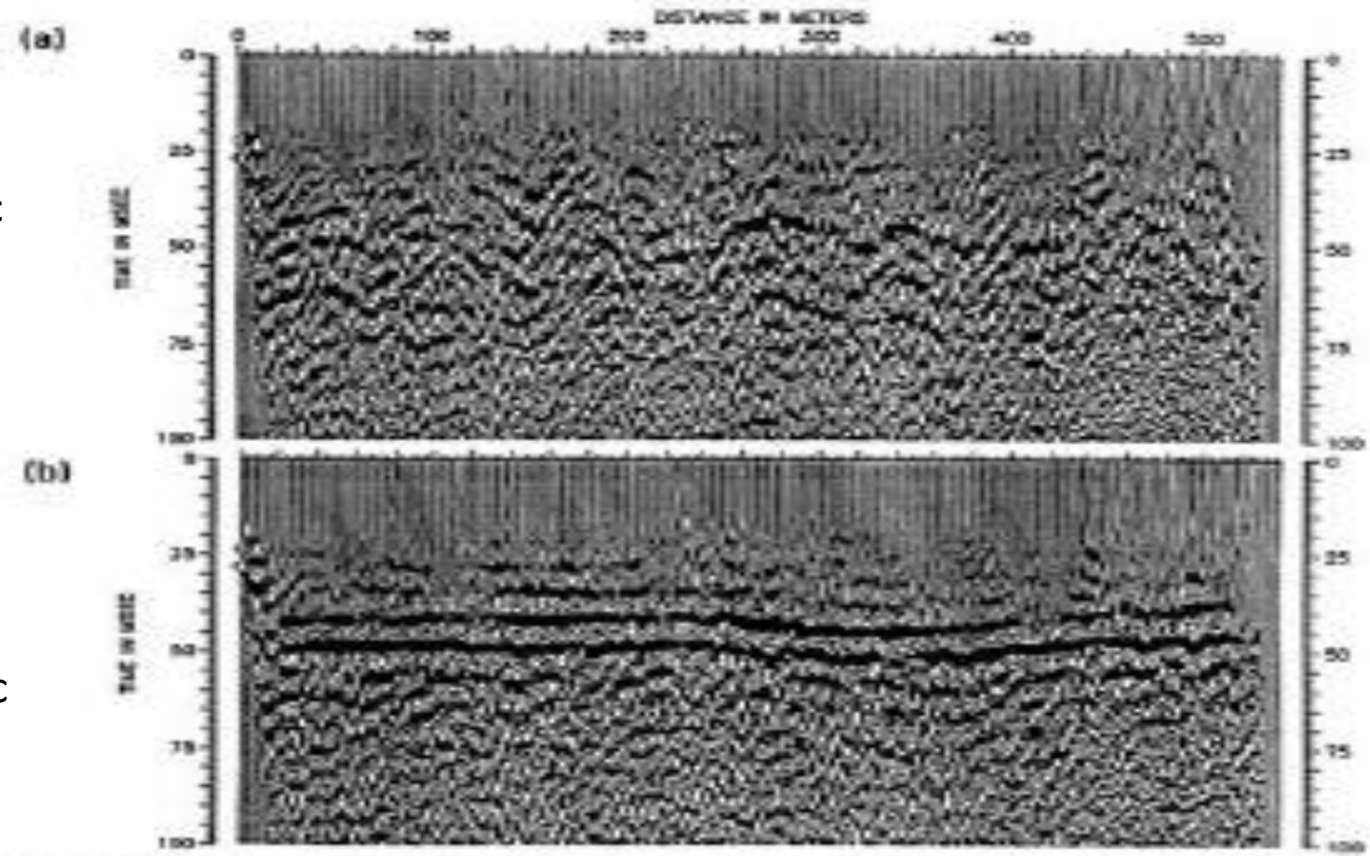
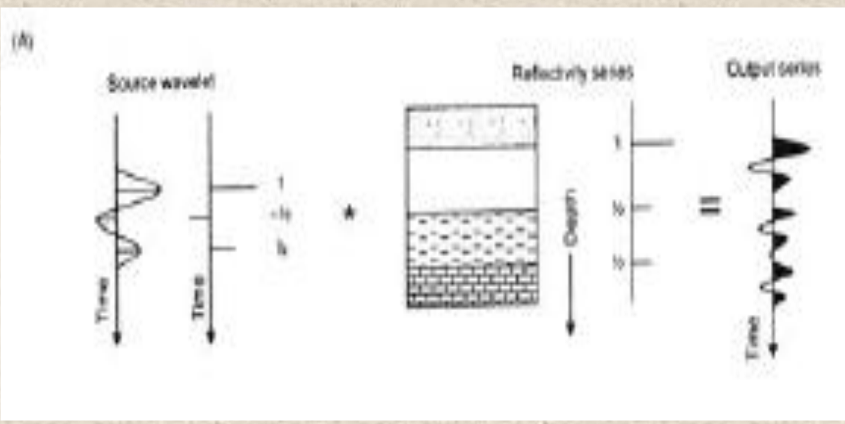


Fig. 4.17 Effect of applying residual static corrections on CDP reflection data recorded at a waste-dump site in Zealand, Denmark. Time section (a) before application of statics and (b) after application of statics. (After Ploug, 1991.)

V. Reflectivity and Convolution

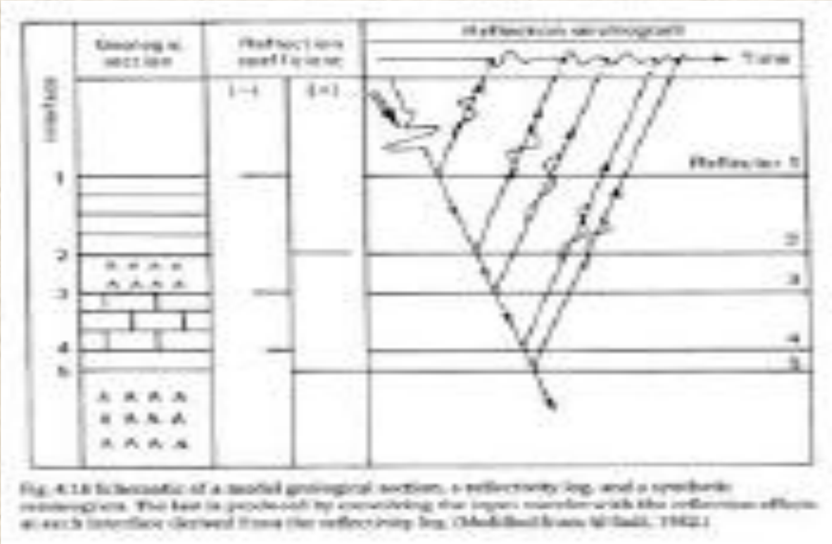
The seismic wave is sensitive to the sequence of impedance contrasts:

The **reflectivity series** (R)



We input a source wavelet (W) which is reflected at each impedance contrast. The seismogram recorded at the surface (S) is the convolution of the two:

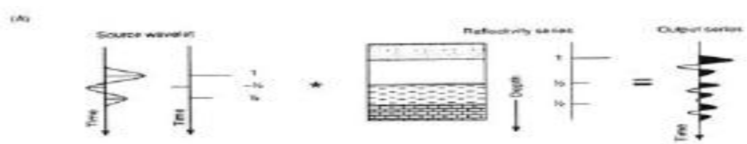
S=W*R



Convolution

Reflectivity series				1	½	½
Source wavelet	½	-½	1			
Output			0			

Recorded waveform



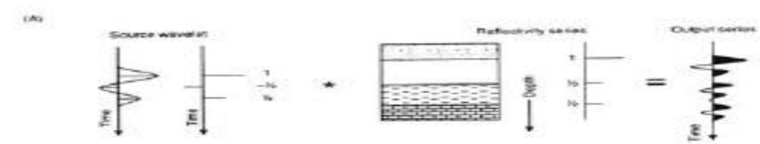
Reflectivity series			1	½	½
Source wavelet	½	-½	1		
Output		0	1		

Recorded waveform

Convolution

Reflectivity series		1	½	½
Source wavelet	½	-½	1	
Output	0	1	0	

Recorded waveform



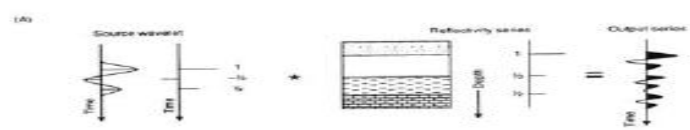
Reflectivity series		1	½	½
Source wavelet		½	-½	1
Output	0	1	0	¾

Recorded waveform

Convolution

Reflectivity series		1	½	½	
Source wavelet			½	-½	1
Output	0	1	0	¾	0

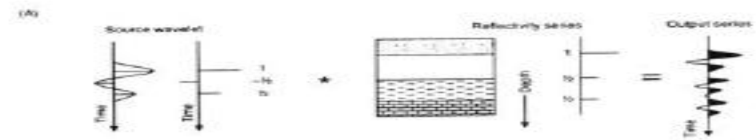
Recorded waveform



Reflectivity series		1	½	½		
Source wavelet				½	-½	1
Output	0	1	0	¾	0	¼

Recorded waveform

Convolution

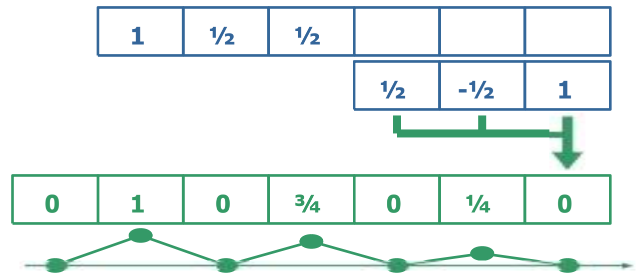


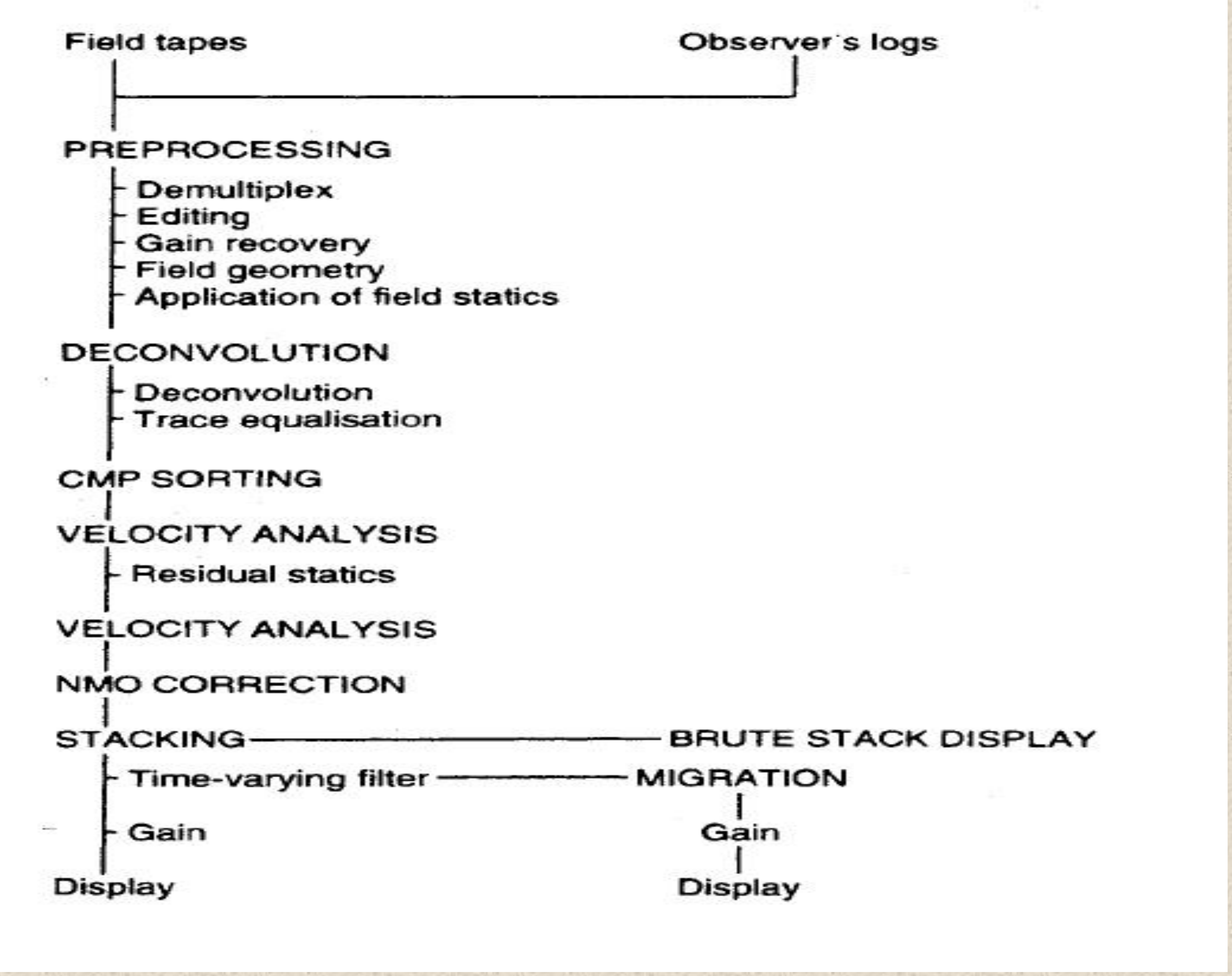
Reflectivity series

Source wavelet

Output

Recorded waveform




Field tapes
Observer's logs
PREPROCESSING
Demultiplex
Editing
Gain recovery
Field geometry
Application of field statics
DECONVOLUTION
Deconvolution
Trace equalisation
CMP SORTING
VELOCITY ANALYSIS
Residual statics
VELOCITY ANALYSIS
NMO CORRECTION
STACKING
BRUTE STACK DISPLAY
Time-varying filter
MIGRATION
Gain
Gain
Display
Display

VI. Deconvolution:

Undoing the convolution to get back to the reflectivity series – what we want.

- A seismic trace is considered to be the convolution of:
  - a seismic wavelet, which represents the propagating pulse, and,
  - a reflectivity time series, which represents the subsurface layering
- Wavelet may vary from shot to shot, and includes reverberations as the signal bounces around between layers before continuing
- Wavelet can be long making the identification of individual boundaries difficult.

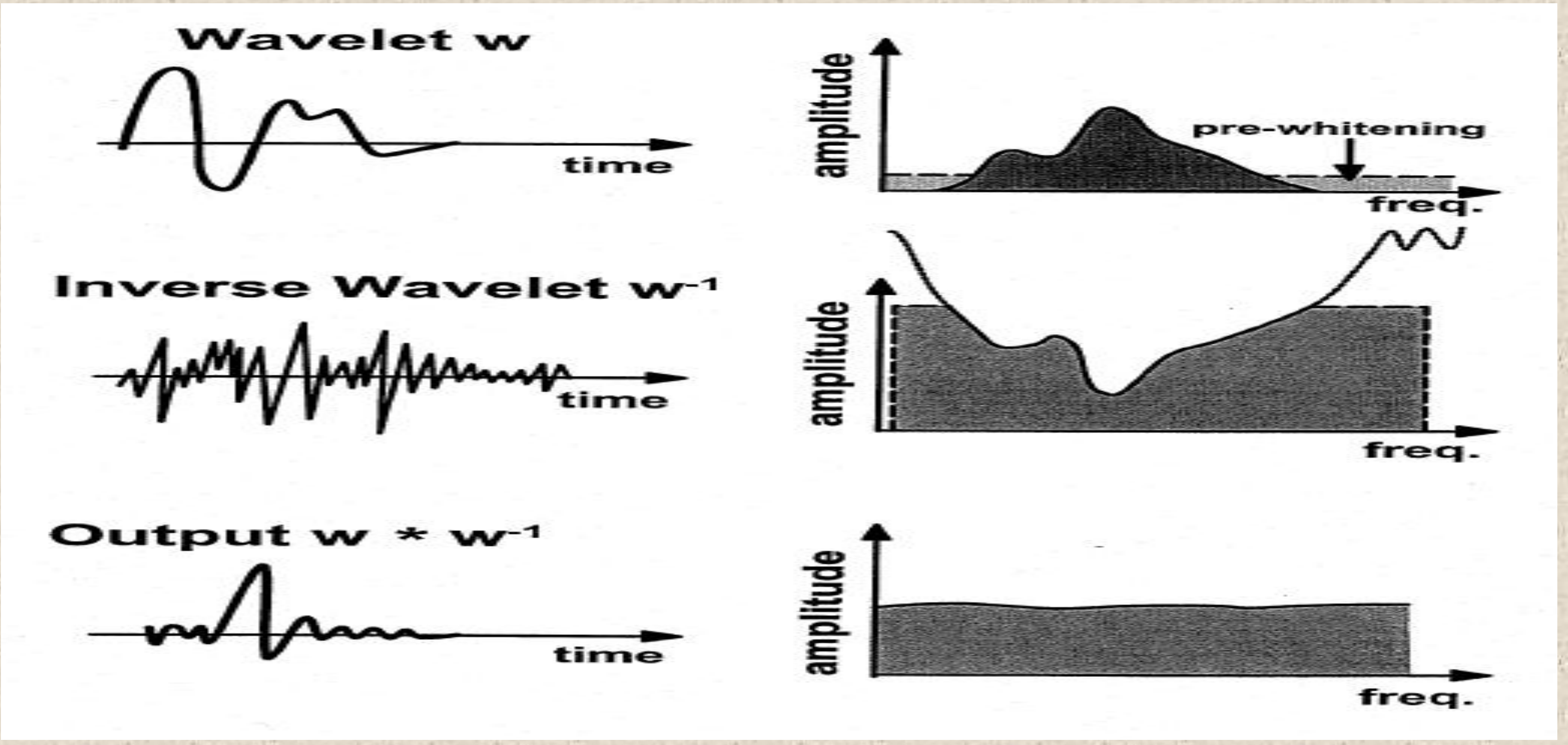
- Deconvolution processing is used to:
  - Correct for changes in source wavelet
  - Remove multiples and reverberations
  - Compress the source wavelet to a short symmetric pulse, which is as close to a spike as you can get with real data

Inverse Wavelet

The inverse of the seismic wavelet is a wavelet which when convolved with the seismic wavelet just gives a spike.

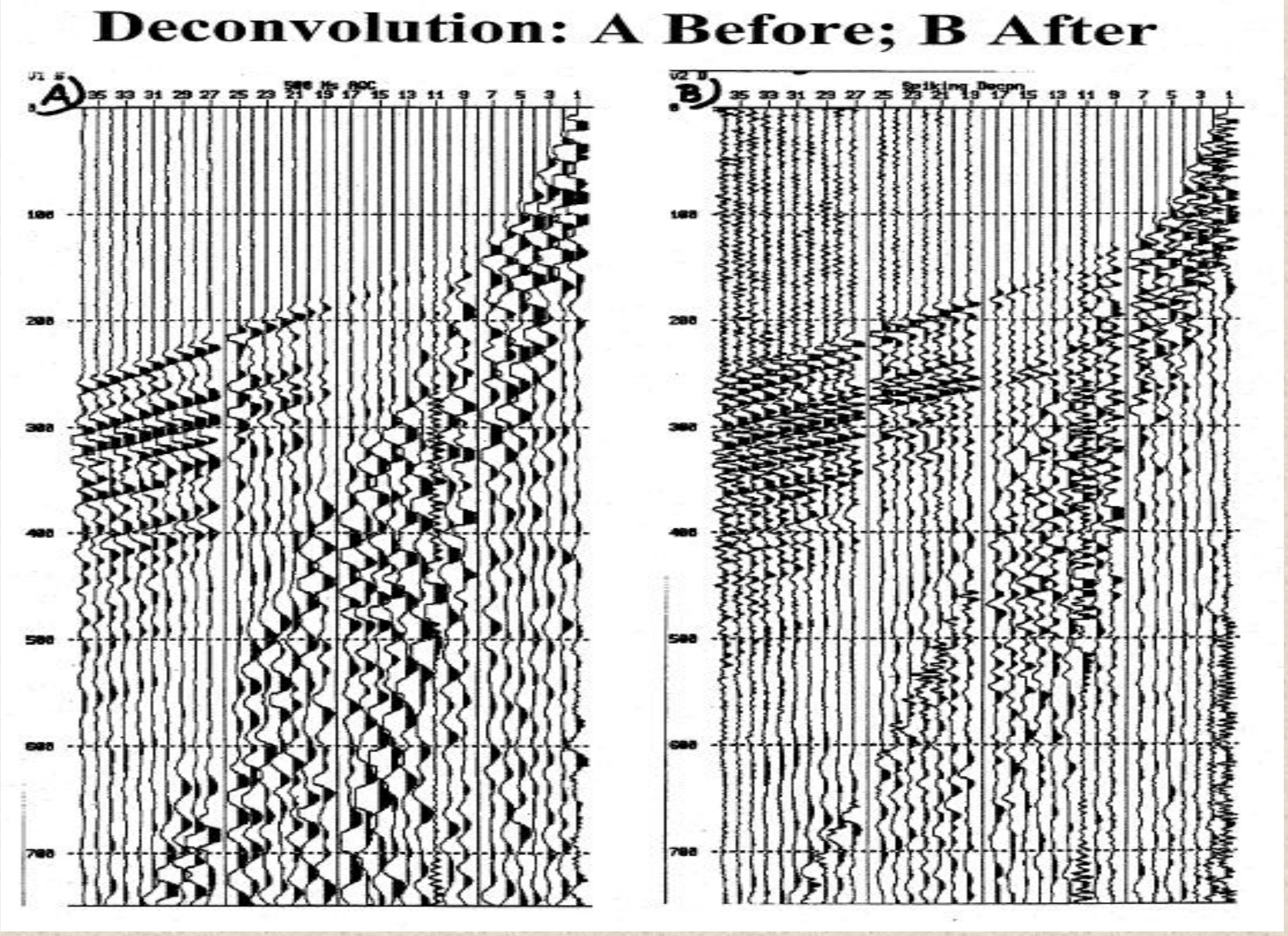
In deconvolution, we estimate an inverse seismic wavelet and convolve it with the recorded seismogram to produce a new seismogram that is closer to the Earth reflectivity series.

- In practice, the exact inverse can never be found owing to the difficulty in estimating the wavelet in the seismic data.



- To avoid amplifying noise, the action of the deconvolution operator is limited by a pre-whitening, typically from 0.01 to 1.0% of the maximum amplitude.
- By compressing the wavelet, deconvolution increases the bandwidth, and hence the resolution of the data.

- Deconvolution is applied to each shot gather recorded in a survey.



Spiking or whitening deconvolution

- Reduces the source wavelet to a spike. The filter that best achieves this is called a **Wiener filter.**

Seismogram **S = R*W** (reflectivity*source)

- Deconvolution operator, D, is designed such that:

$$\mathbf{D*W} = \delta$$

So,

$$\mathbf{D*S = D*R*W = D*W*R} = \delta\ \mathbf{*R = R}$$

Time-variant deconvolution

The operator D changes with time to account for the different frequency content of energy that has traveled greater distances.

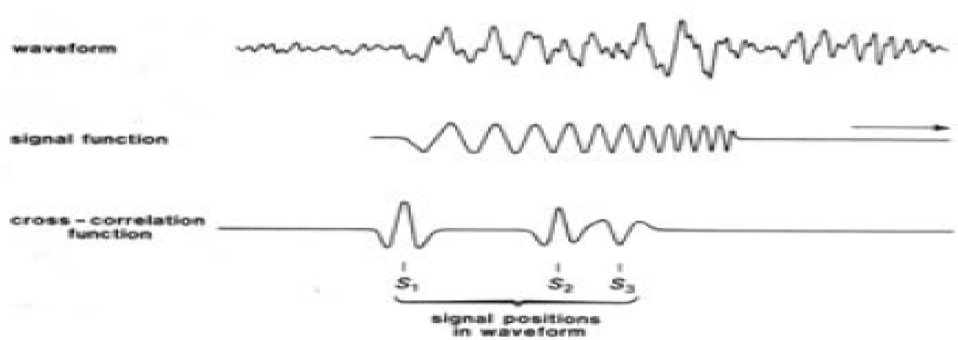
Predictive deconvolution

The arrival times of primary reflections are used to predict the arrival times of multiples which are then removed

Predictive deconvolution

A multiple reflection is a duplication of the primary reflection and should have similar characteristics as the primary event.

Deconvolution using correlation



If we know the source pulse

Then cross-correlating it with the recorded waveform gets us back (closer) to the reflectivity function

If we don't know the source pulse

Then autocorrelation of the waveform gives us something similar to the input plus **multiples**.

Autocorrelation → correlating a signal against a time shifted copy of itself.

Cross-correlation → Comparing one time series with another time series to measure similarity.

VII. Velocity Analysis:

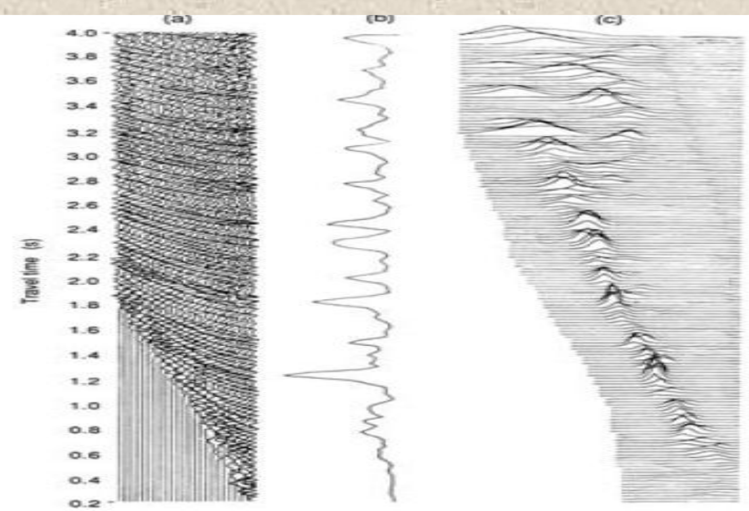
Velocity analysis

Determination of seismic velocity is key to seismic methods

Velocity is needed to convert the time-sections into depth-sections i.e. geological cross-sections

Unfortunately reflection surveys are not very sensitive to velocity

Often complimentary refraction surveys are conducted to provide better estimates of velocity



Normal move out (NMO) correction

The reflection traveltime equation predicts a hyperbolic shape to reflections in a CMP gather. The hyperbolae become fatter/flatter with increasing velocity

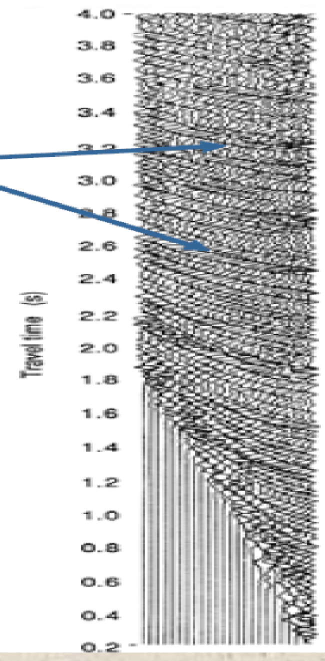
$$T_x^2 = T_0^2 + \frac{x^2}{V_1}$$

We want to subtract the NMO correction from the common depth point gather

$$\Delta T_{NMO} \approx \frac{x^2}{2T_0V_1^2}$$

But for that we need velocity...

reflection hyperbolae become fatter with depth (i.e. velocity)



Stacking Velocity

- The stacking velocity is simply the velocity of the hyperbola that fits best the form of a reflection in a CMP gather.

- Stacking velocities must be estimated from CMP gathers of field data to provide velocity values for NMO correction:

  When a CMP is stacked after NMO correction, each time with a different velocity, and the results plotted side-by-side, the optimal stacking velocity can be identified.

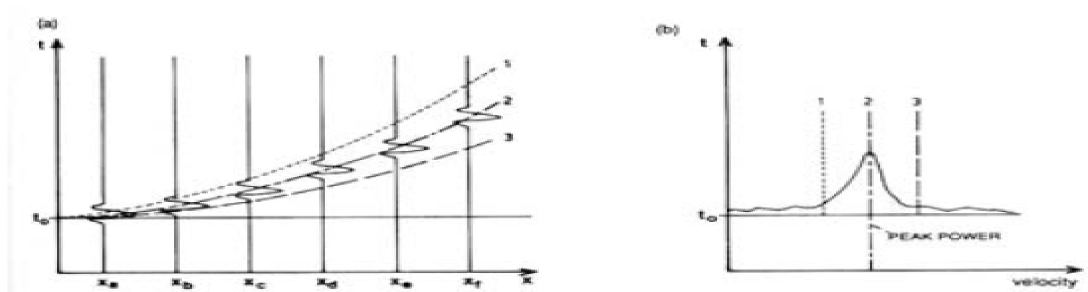
Stacking velocity

In order to stack the waveforms we need to know the velocity. We find the velocity by trial and error:

$$\Delta T_{NMO} = \frac{x^2}{2T_0 V_1^2}$$

- For each velocity we calculate the hyperbolae and stack the waveforms
- The correct velocity will stack the reflections on top of one another
- So, we choose the velocity which produces the most power in the stack

V_2 causes the waveforms to stack on top of one another



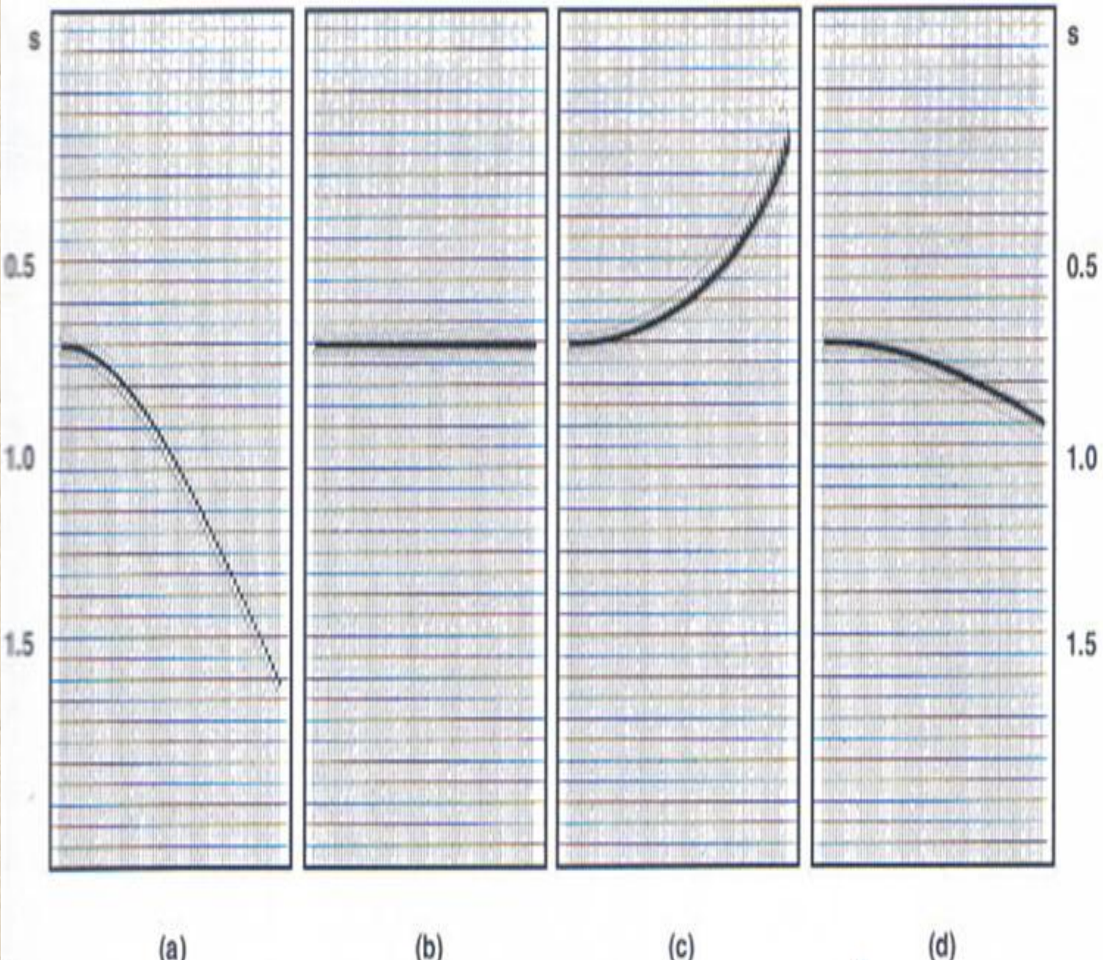


FIGURE (a) CMP gather containing a single event with a moveout velocity of 2264 m/s. (b) NMO-corrected gather using the appropriate moveout velocity. (c) Overcorrection because the velocity (2000 m/s) used was too low. (d) Undercorrection because the velocity (2500 m/s) used was too high. (Adapted from Yilmaz, 1987.)

VIII. Stacking:

Stacking

Each CMP gather contains several traces corresponding to a single position on a subsurface interface (assuming horizontal layers).

In CMP gathers, many seismic arrivals appear approximately hyperbolic with their apex at zero offset

Stacking Velocity

The stacking velocity of an arrival is the velocity value that characterises the hyperbola that is the best fit to the arrival

Creating a Reflection Profile

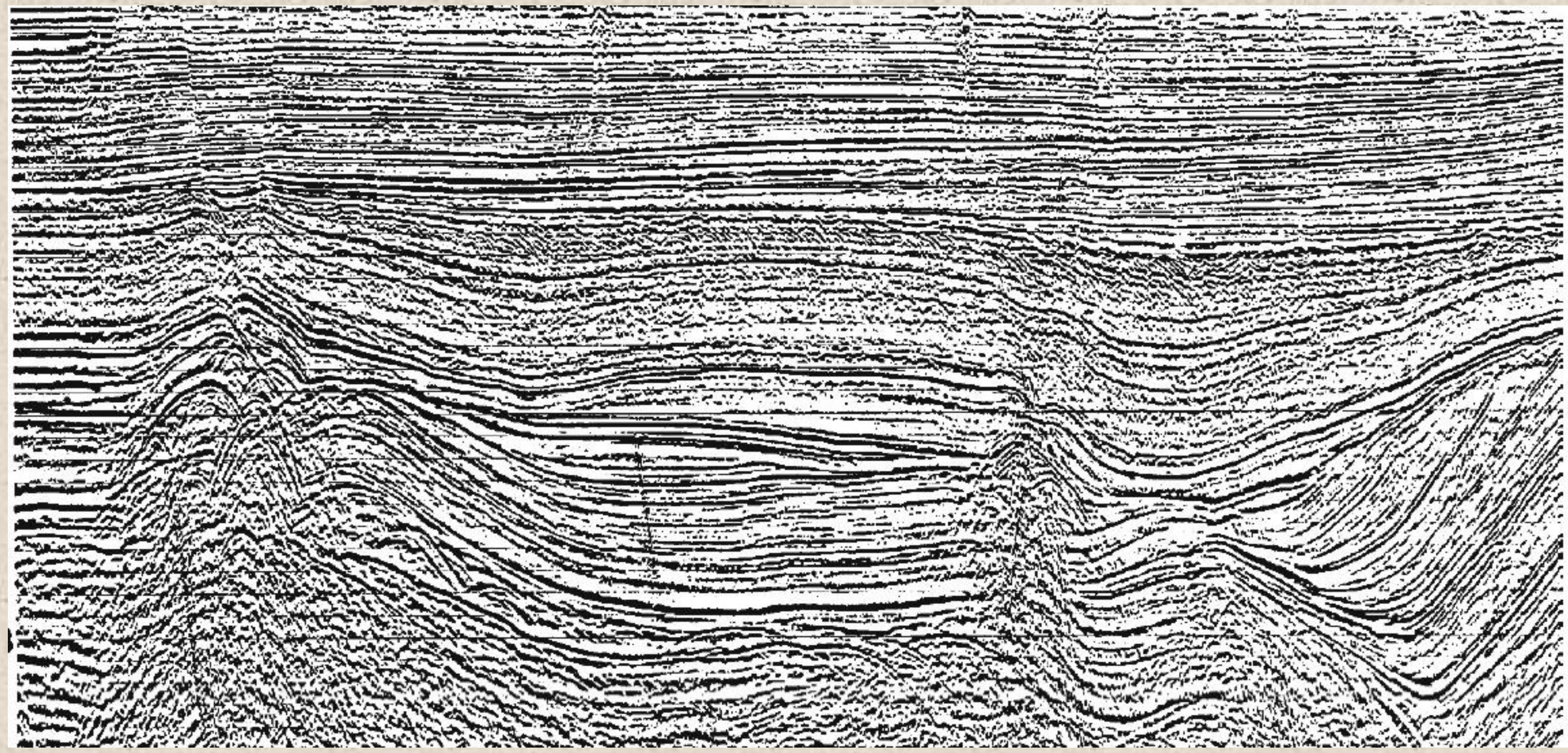
The traces in each CMP gather are combined in a 3-stage process to produce a single trace that represents the reflections recorded at that CMP position.

- NMO Correction: Aligns each primary reflection at its zero-offset reflection time using primary reflection stacking velocities
- Muting: Parts of traces stretched greatly by NMO correction are muted, i.e. set to zero
- Stacking: All traces in a CMP gather are summed to create a single seismogram that simulates a recording made with a coincident source and receiver at the CMP location

Stacking enhances primary reflections and suppresses random noise and unaligned coherent arrivals such as multiples.

Stacked Reflection Profile

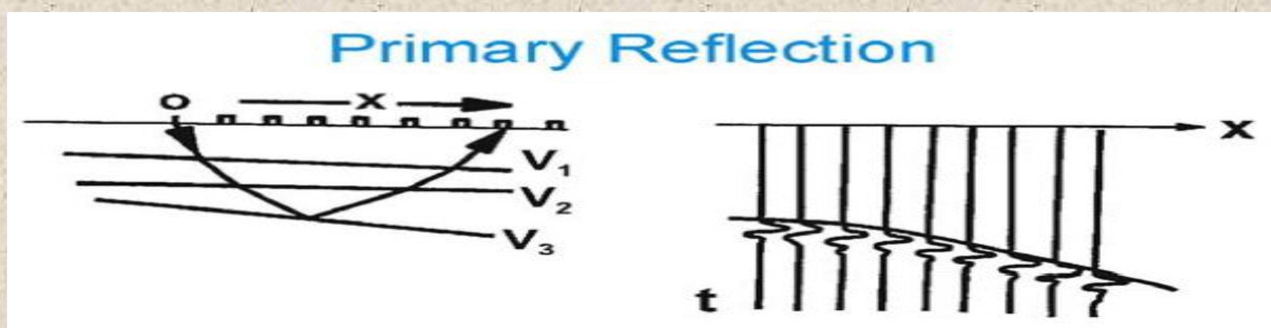
A seismic reflection profile is created by plotting side-by-side each stacked trace from the CMP gathers along a seismic profile.

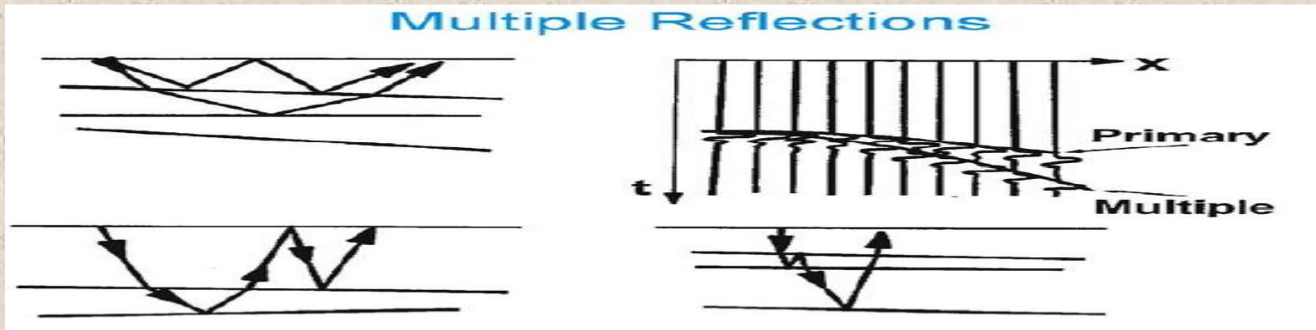


Stacking Suppresses Multiples

- **Primary Reflection:** Wave that travels directly down to an interface, is reflected, and then travels directly back to surface

- **Multiple Reflection:** Wave that is reflected more than once, and so has two or more up and downgoing legs





- Seismic velocity usually increases with depth, so multiple reflections have more moveout than primary reflections, i.e. lower stacking velocities

- When the primary reflection is aligned after NMO, the multiple will be misaligned and so will be partly cancelled by stacking

❖ Reflections on a stacked seismic section are considered to have been produced by normal incidence reflection:

- Upgoing and downgoing paths are identical

- Angle of reflection is zero

IX. Frequency Filtering:

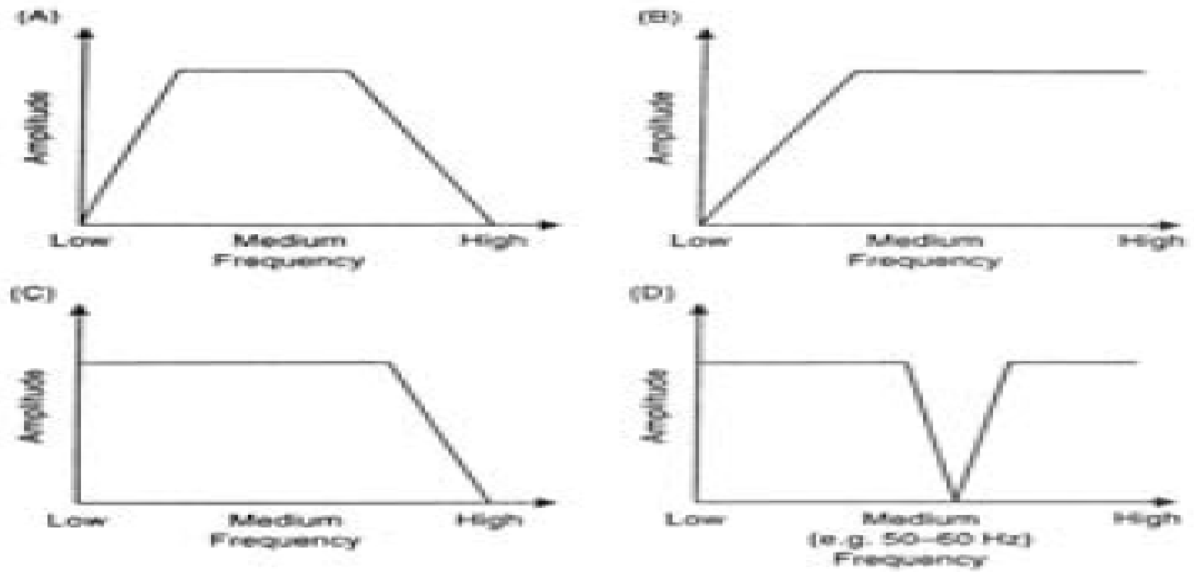
- High and low frequency noise can be present in the final stack and migrated sections

- Bandpass filtering is applied to remove unwanted noise.

Hi-pass: to remove ground roll

Low-pass: to remove high frequency jitter/noise

Notch filter: to remove single frequency



X. Migration:

- A dipping reflection is NOT at its true subsurface position.
- Migration moves reflections on a stacked seismic section to their true subsurface position

A reflection is in its true subsurface position when the angle between the normal incidence raypath and reflector is 90°.

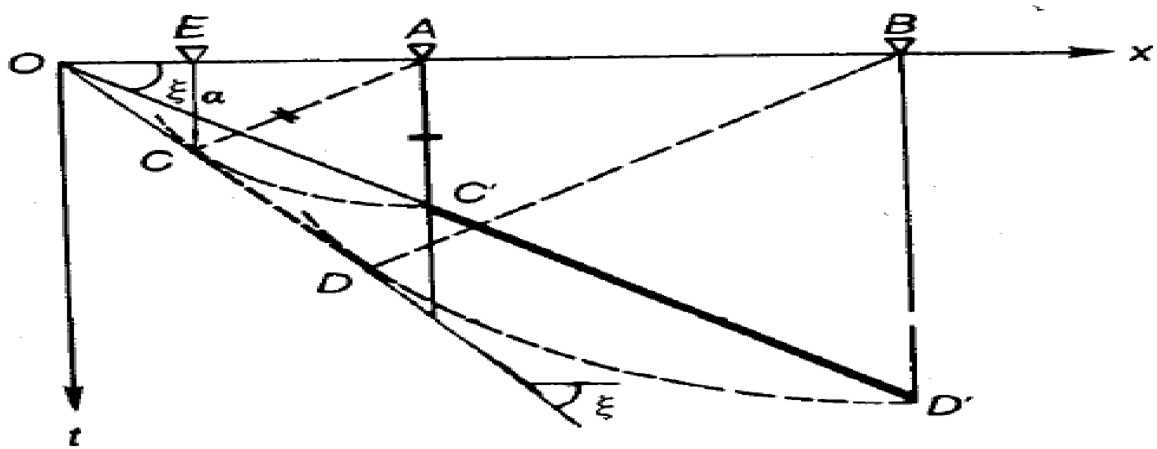


Fig. Migration principle. Migration of segment $C'D'$ into CD increases the dip from ξ_a to ξ.

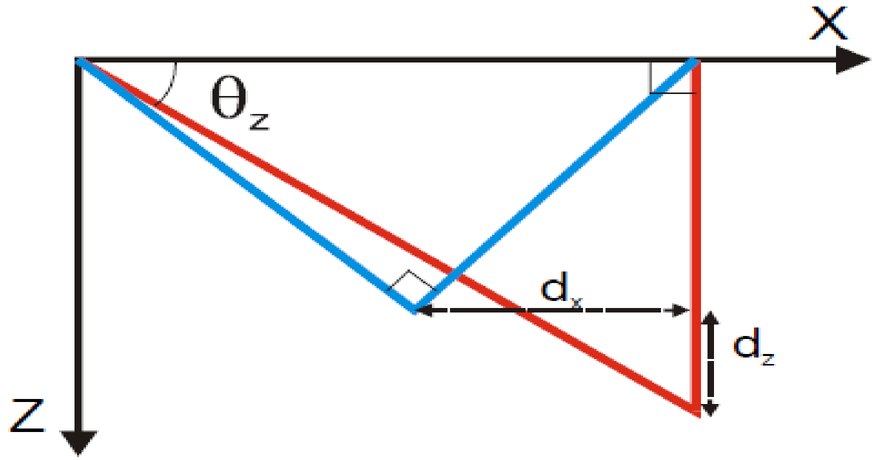
$$\sin\xi = \frac{BD}{OB} = \frac{BD'}{OB} = \tan\xi_a$$

The true position of a dipping reflection is updip at shallower time.

Maximum dip on unmigrated section is 45°.

Migration of a Dipping Reflection

In constant velocity, its possible to calculate the distance a dipping reflection will have moved after migration.



The lateral displacement is given by:

$$d_x = \frac{V^2 T}{4} \tan \theta_t \; .$$

The apparent dip on a section with the vertical axis in time is:

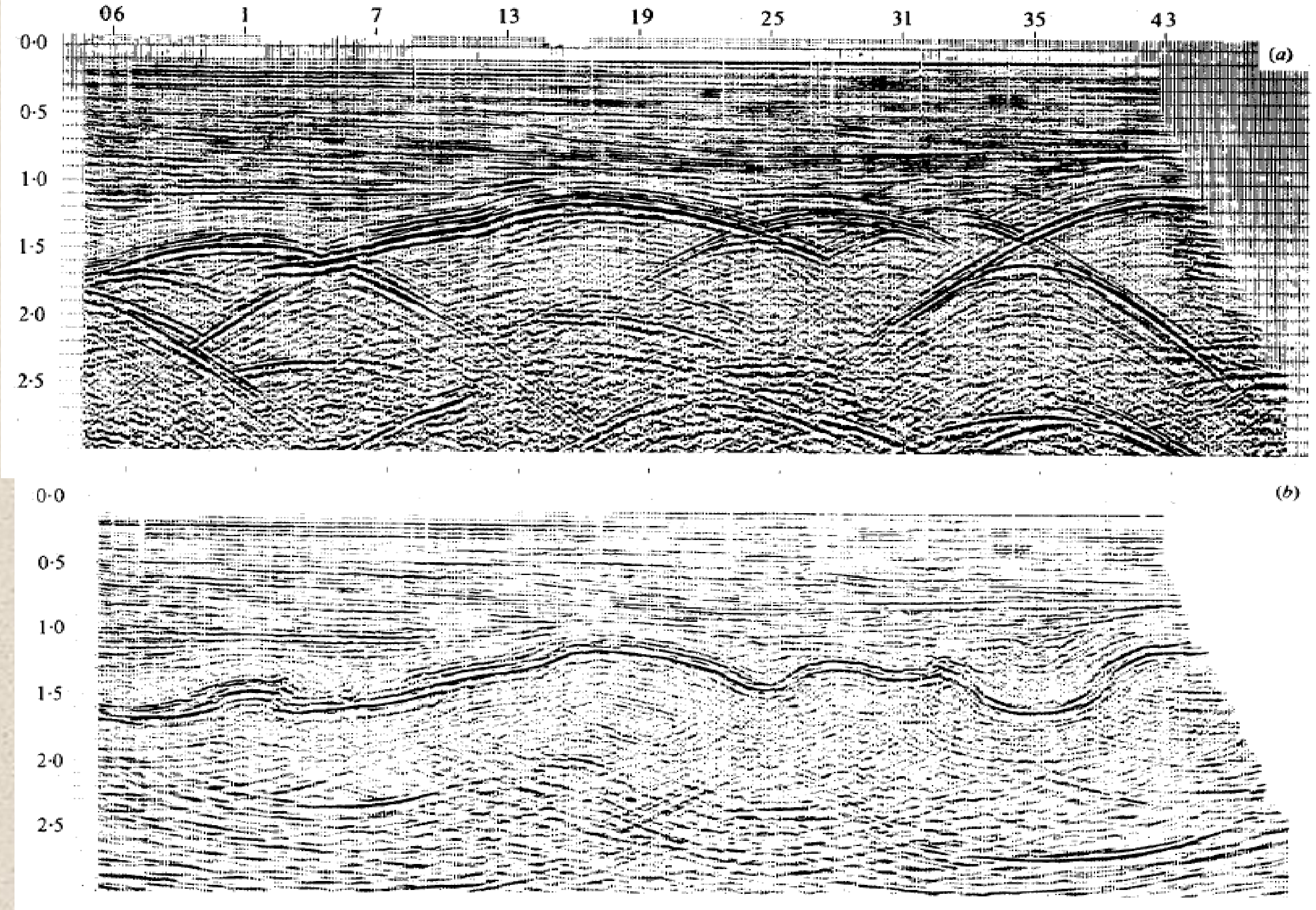
$$\frac{dt}{dx} = \tan \theta_t = \frac{2}{V} \tan \theta_z$$

where θ_z is the apparent dip on a section that has been converted to depth by scaling the vertical axis by V/2.

The vertical displacement is given by:

$$d_z = \frac{VT}{2} \left(1 - \left(1 - \frac{V^2}{4} \tan^2 \theta_t \right)^{\frac{1}{2}} \right)$$

Example: Migration of reflection from a rugged interface



(a) Unmigrated section; (b) Migrated section.

Resolution of structure

Consider a vertical step in an interface

To be detectable the step must cause an delay of ¼ to ½ a wavelength

This means the step (h) must be 1/8 to ¼ the wavelength (two way traveltime)

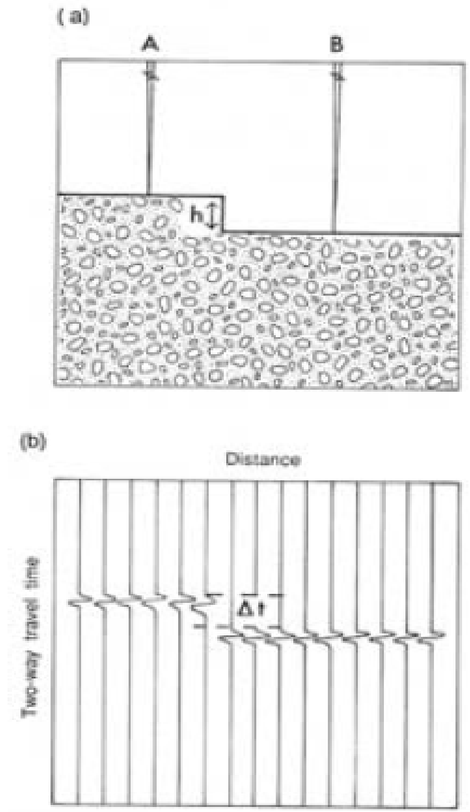
Example:

20 Hz, α = 4.8 km/s then λ = 240 m

Therefore need an offset greater than 30 m

Shorter wavelength signal (higher frequencies) have better resolution.

What is the problem with very high frequency sources?



Fresnel Zone

Tells us about the horizontal resolution on the surface of a reflector

First Fresnel Zone

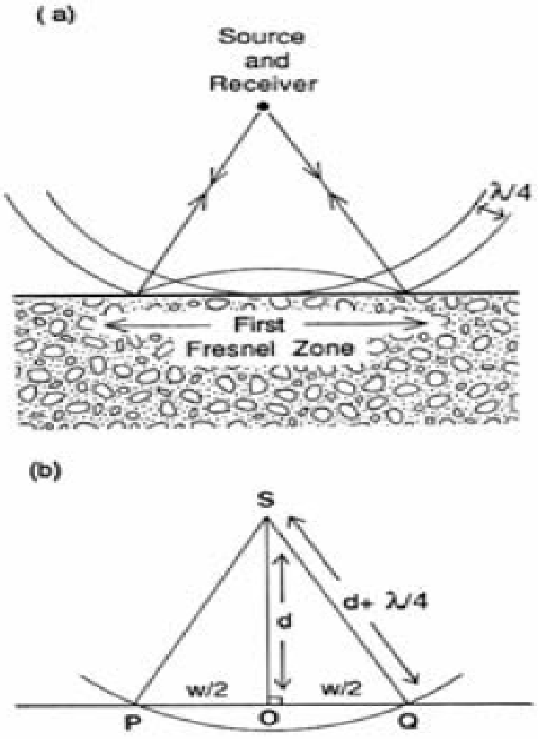
The area of a reflector that returns energy to the receiver within half a cycle of the first reflection

The width of the first Fresnel zone, w:

$$\left(d+\frac{\lambda}{4}\right)^2 = d^2 + \left(\frac{w}{2}\right)^2$$

$$w^2 = 2d\lambda + \frac{\lambda^2}{4}$$

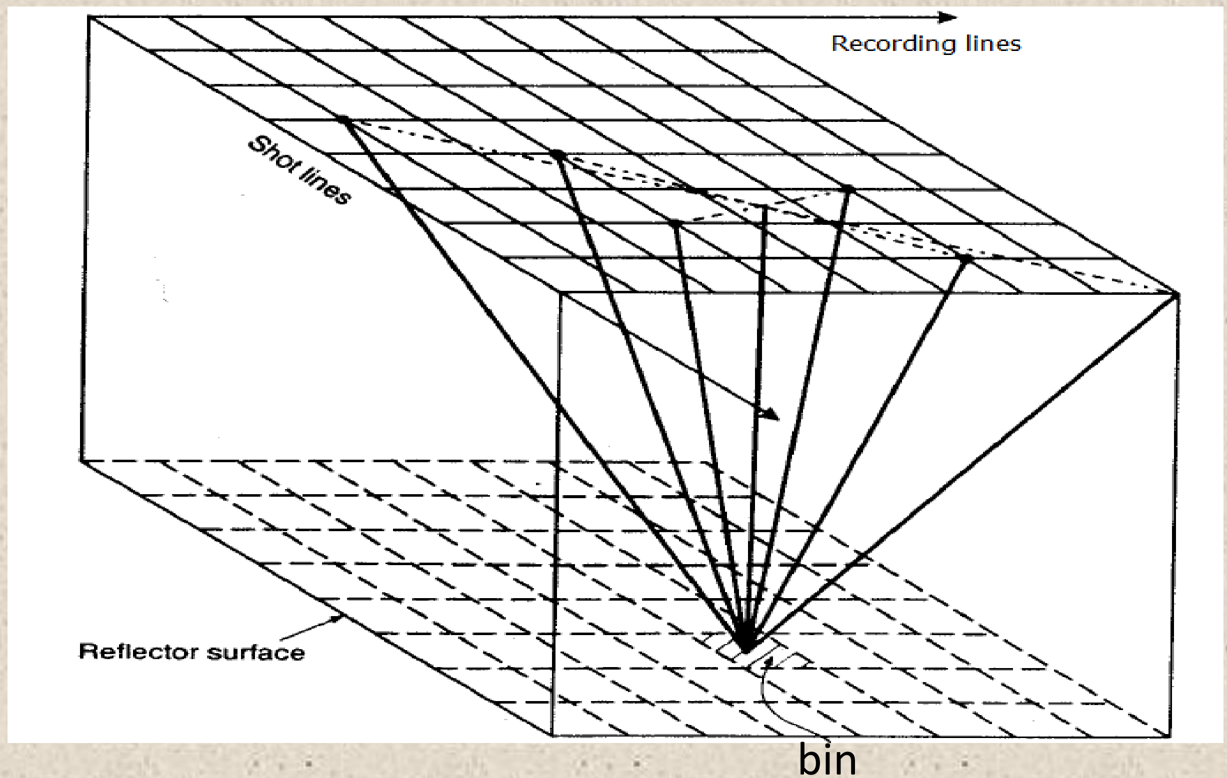
If an interface is smaller than the first Fresnel zone it appears as a. point diffractor, if it is larger it appears as an interface



Example:
30 Hz signal, 2 km depth where α = 3 km/s then λ = 0.1 km and the width of the first Fresnel zone is 0.63 km

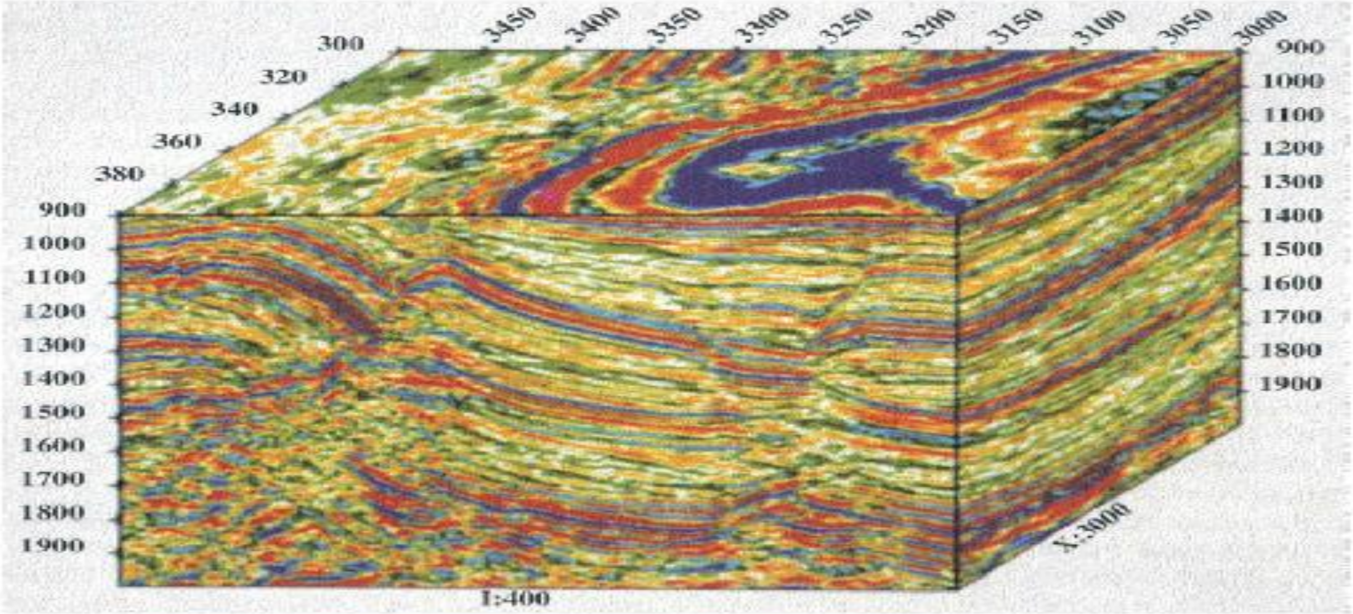
3-D Seismic Surveying

- In 2-D surveys, shots and receivers lie along the same surface line.
- 3-D surveys can be acquired with shots fired along a line orthogonal to the recording spread, so midpoints are areally distributed



3-D Seismic Volume

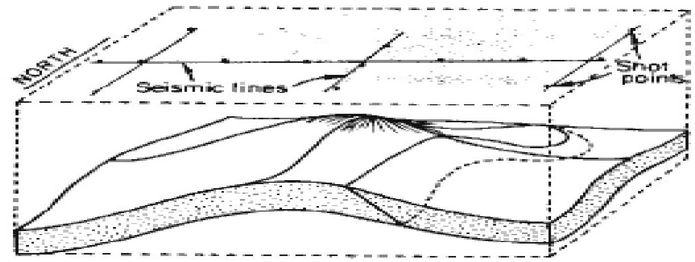
Final result of a 3-D survey is a data cube that provides a continuous image of a subsurface volume.

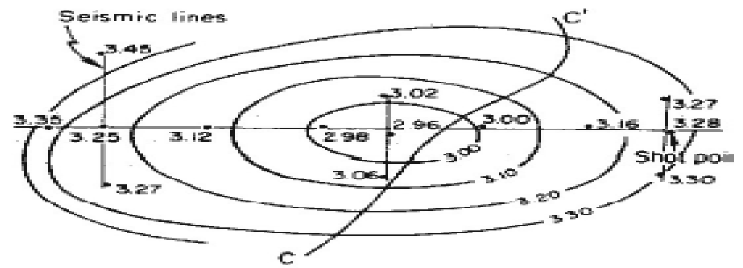


- Interpretation is made by mapping horizons and faults in 3-D
- Horizons are differenced to emphasise geological changes in a specific time interval (geologic or recorded).

Structural Interpretation

Grid of seismic lines or 3-D survey are necessary to make a reasonable interpretation



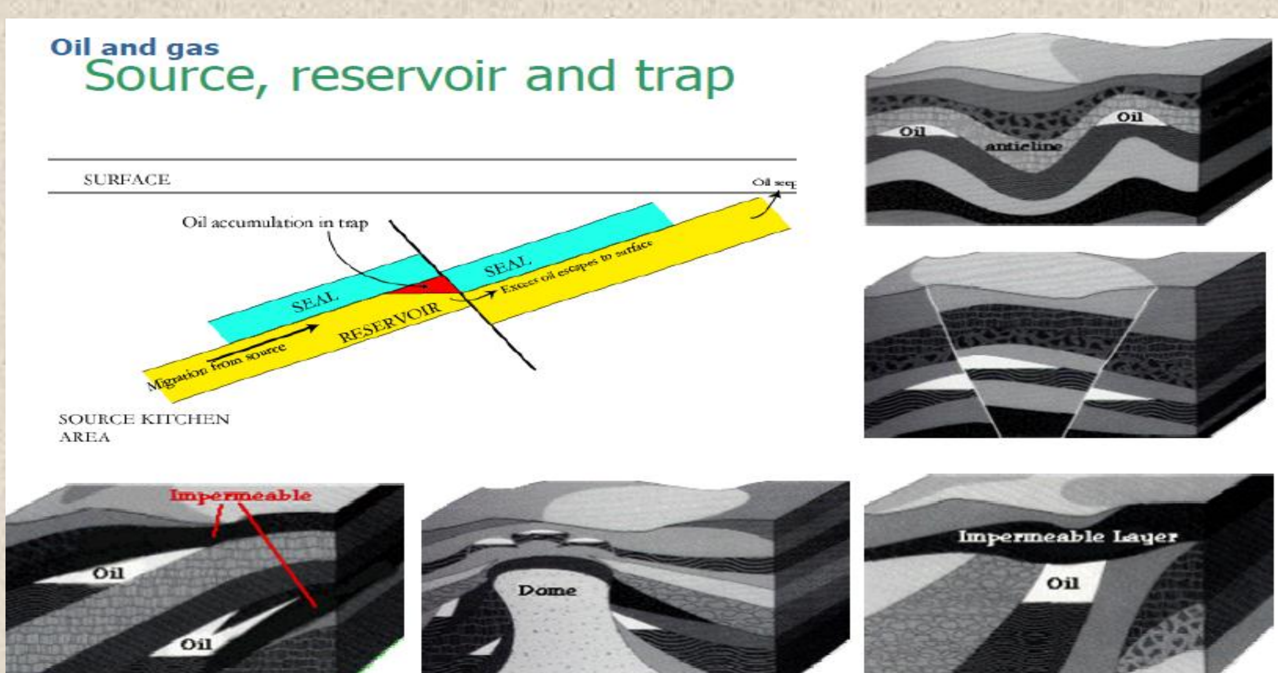


- To map a horizon, pick its reflection on each seismic line and check tie at line intersections
- Horizon times are contoured to produce a <u>time structure map</u>
- Faults interpreted from offsets in reflections
- Horizon is contoured separately on opposite sides of fault

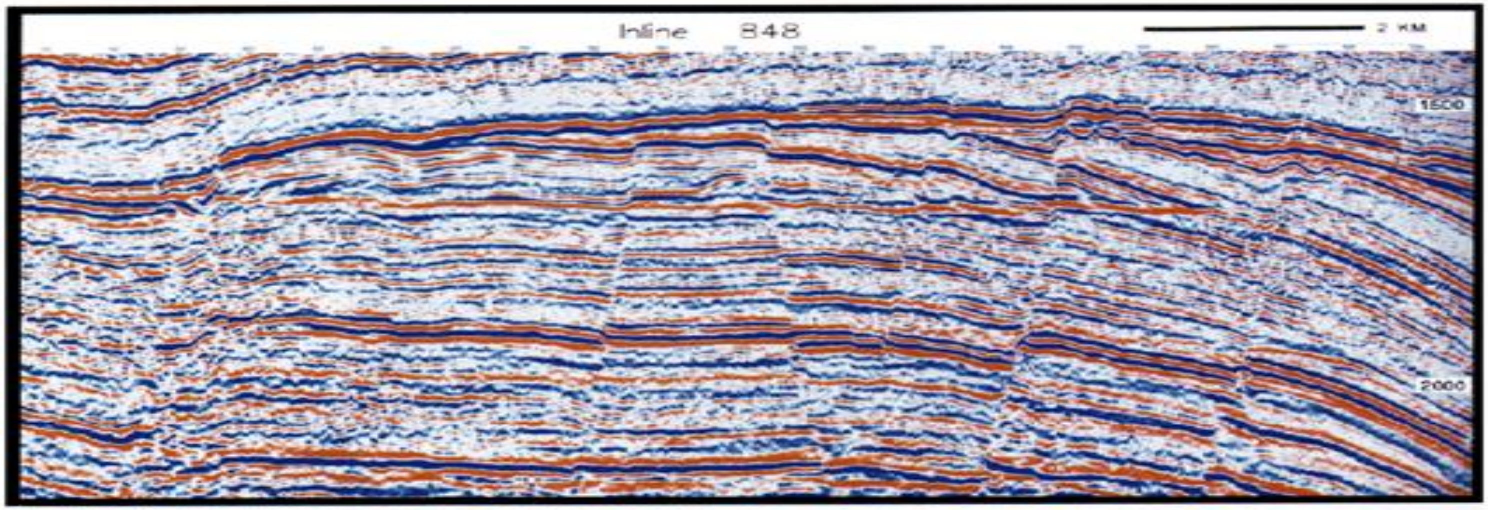
Isochron Map
Map showing time difference between two interpreted seismic horizons

- Depth and isopach maps can be made if velocities available for time to depth conversion of horizon times

Examples of Structural Features Responses



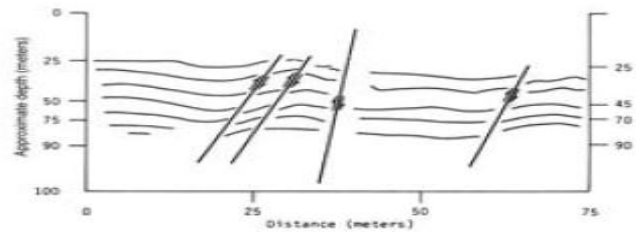
Faults

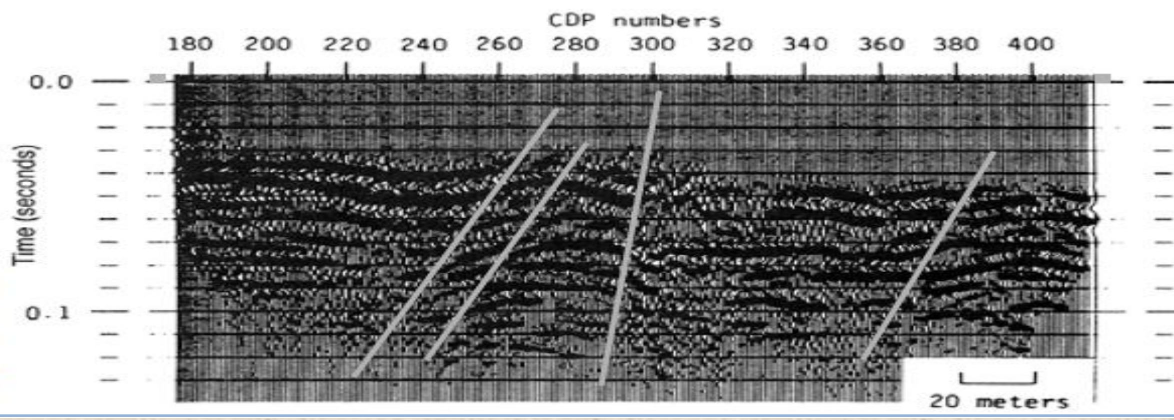


Locating faults

- ➔ **Migrating fluids**
- ➔ **Seismic hazard**

Identified as discontinuities in reflection surfaces





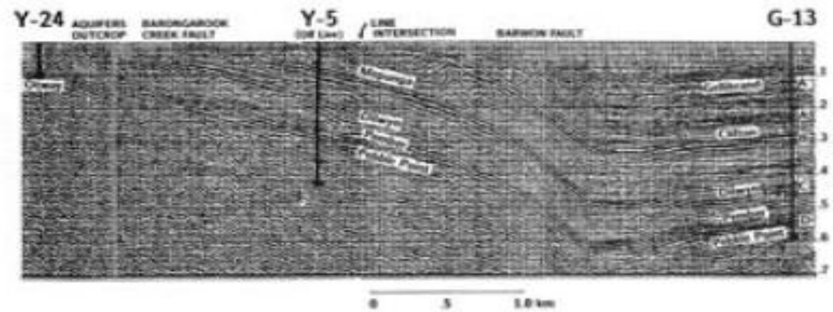
None of these faults visible at the surface – suggests recent inactivity

Groundwater

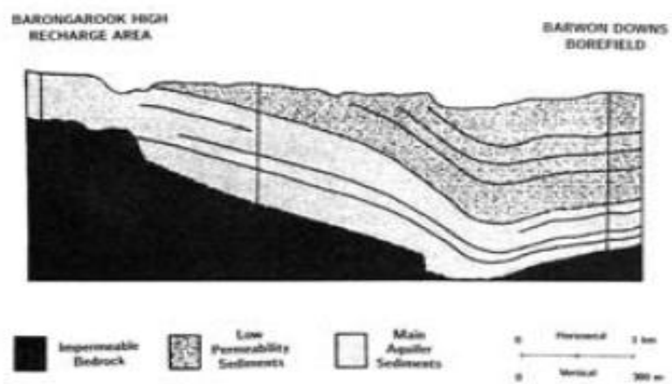
Objective:

Map subsurface location of aquifer for the purpose of drilling a well

(a)

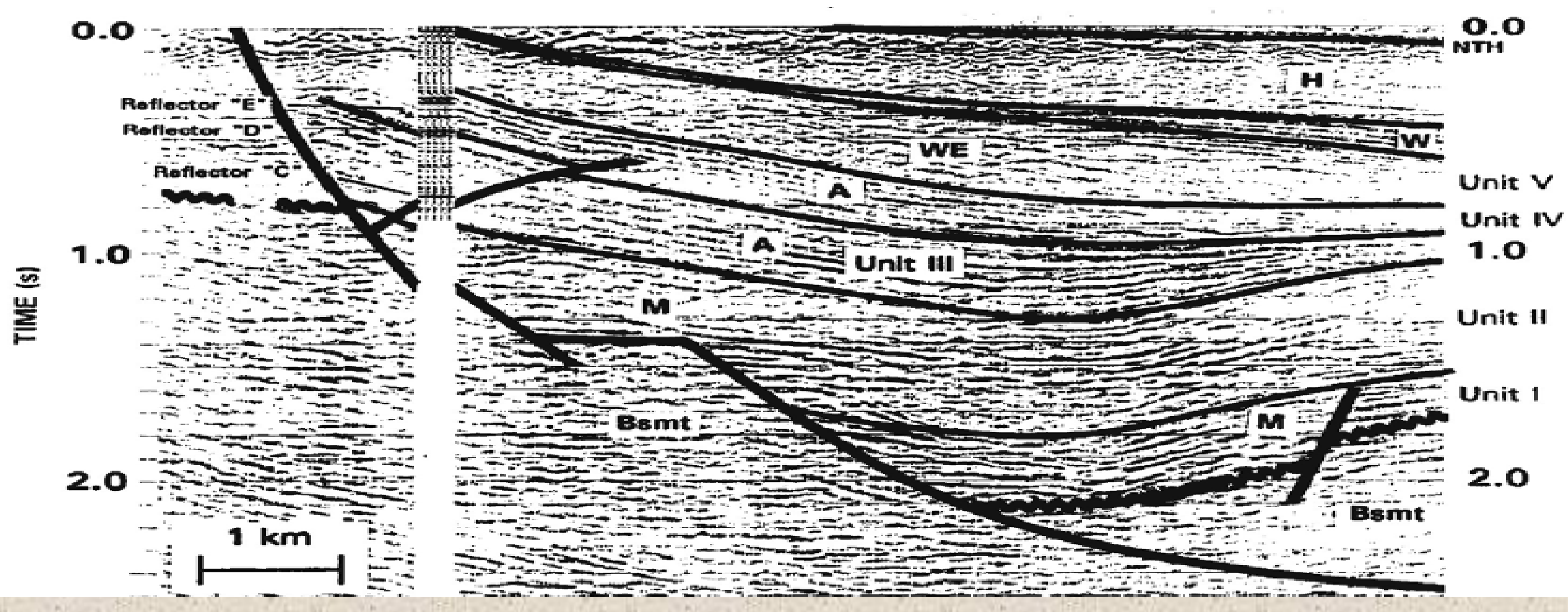


(b)



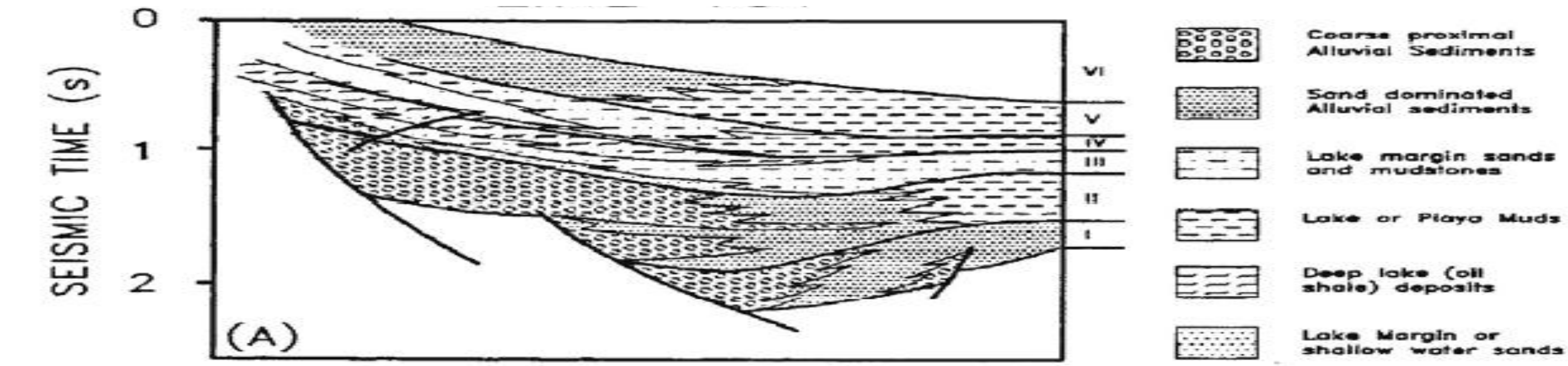
Seismic Stratigraphy

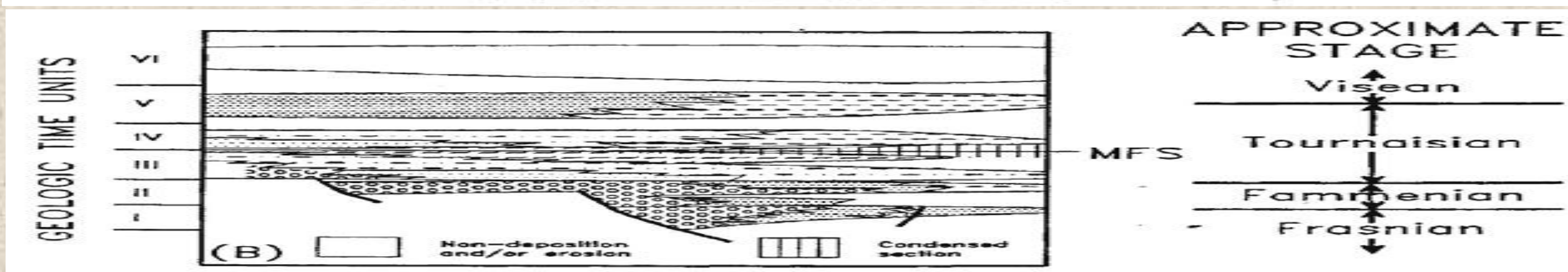
In sedimentary basins, seismic reflections arise from the interference in reflections from many thin geological strata



- Seismic reflections considered as isotime markers
- Seismic reflections occur together in stratigraphic units separated by discontinuities, e.g unconformity
- Stratigraphic units can be classified on basis of internal character

SEISMIC FACIES UNIT	TOP BOUNDARY	BOTTOM BOUNDARY	EXTERNAL FORM	REFLECTION CONFIGURATION	REFLECTION CONTINUITY	ENVIRONMENT
I	E:CONCORDANT	F: ONLAP	WEDGE?	DIVERGENT	DISCONTINUOUS HIGH AMPLITUDE	PROXIMAL FAN? SYN-RIFT CLASTICS
II	D:CONCORDANT WITH DISTAL EROSIONAL TRUNCATION	E: LOCAL DOWNLAP OTHERWISE CONCORDANT	FAN	PROXIMAL, CHAOTIC OR REFLECTION-FREE; DISTAL, SUBPARALLEL	DISCONTINUOUS HIGH AMPLITUDE	ALLUVIAL FAN
III	C:CONCORDANT TO TOPLAP TRUNCATION DISTALLY	D: CONCORDANT EXCEPT DOWNLAP TO SW ON LINE 48X	SHEET OR WEDGE WITH PROXIMAL MOUND	UNIFORM PARALLEL; CHAOTIC IN MOUND	DISCONTINUOUS	FAN DELTA COMPLEX
IV	B:CONCORDANT	C: PROXIMAL CONCORDANT, DISTAL DOWNLAP	BASIN FILL	UNIFORM PARALLEL	CONTINUOUS	LACUSTRINE
V	A: ONLAP OF OVERLYING UNITS	B: PROXIMAL CONCORDANT, DISTAL DOWNLAP	BASIN FILL (NE-SW TROUGH)	PROXIMAL, CHAOTIC OR REFLECTION-FREE; DISTAL, PROGRADING OBLIQUE CLINOFORMS	LOW AMPLITUDE DISCONTINUOUS	ALLUVIAL PLAIN
VI	EROSIONAL (PRESENT BEDROCK SURFACE)	A: ONLAP	BASIN FILL (SLOPE FRONT)	SUBPARALLEL	UPPER DISCONTINUOUS; LOWER CONTINUOUS HIGH AMPLITUDE	





SEISMIC INSTRUMENTATION
&
SOURCES

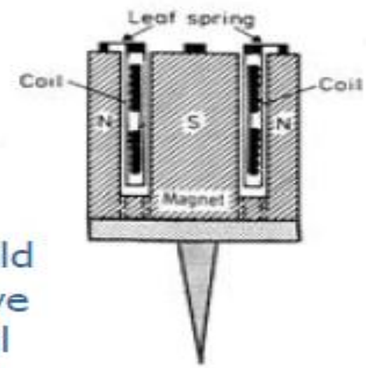
SEISMIC INSTRUMENTATION

1. Seismograph

- 24 – 120 channels or more
- Digital
- Screen control

2. Geophones (Seismometers):

A geophone is used to transform seismic energy into electrical voltage.



Geophones

- Cylindrical coil suspended in a magnetic field
- The inertia of the coil causes motion relative to the magnet generating a electrical signal
- Geophones are sensitive to velocity

Hydrophones

- Used at sea
- Use piezoelectric minerals to sense pressure variations

Instrument response

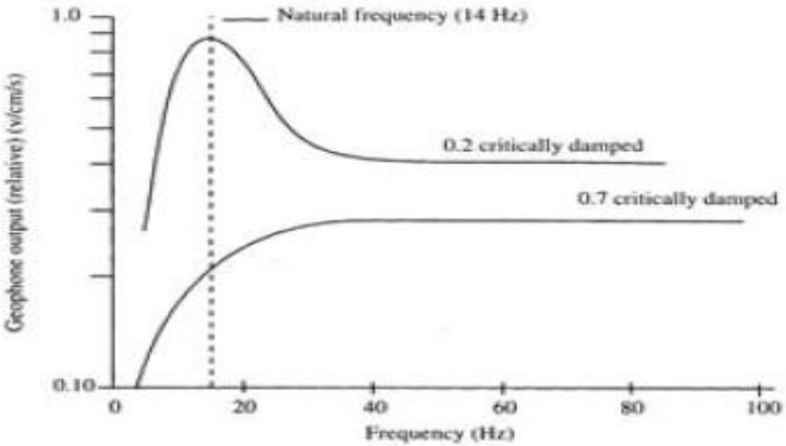
- The relation between the input ground motion and the output electrical signal

Natural frequency

- The frequency which produces the maximum amplitude output

Damping

- Reduces the amplitude of the natural frequency response and prevents infinite oscillations
- Want a **flat response**



- At critical damping, coil will return to rest most quickly.
- If damping very small, coil will oscillate at the natural frequency of the electromechanical system.
- Normal damping is 70% critical.

Natural Frequency

Natural frequency and damping affect the range of frequencies the geophone can record:

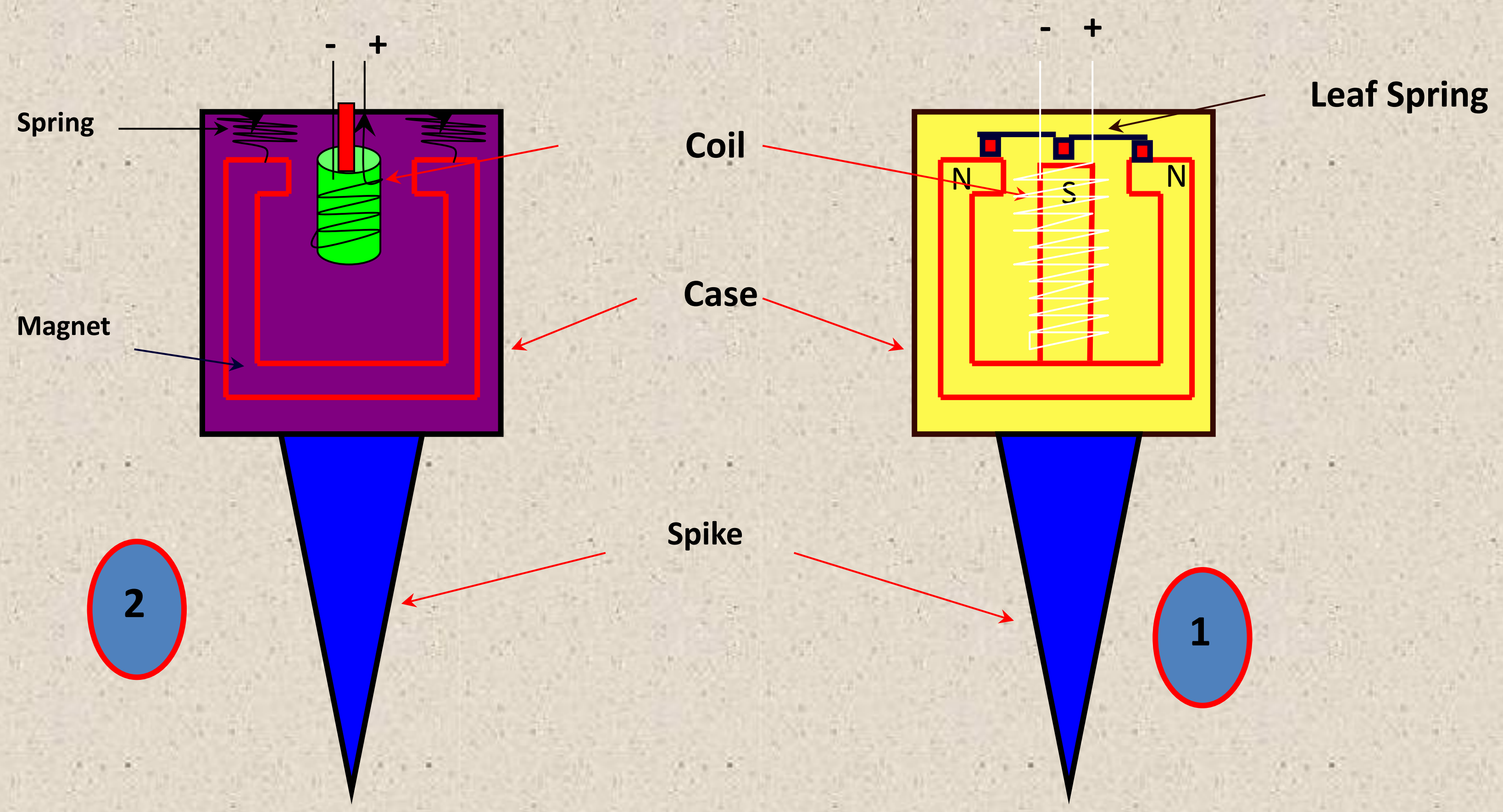
- 14 Hz geophones used in oil exploration
- 30 Hz geophones used in high resolution studies
- 100 Hz geophones used in very shallow work

Type of Geophones:

A- Land survey:

Two types of geophones are used:

- Moving coil geophone (1).
- Moving magnet geophone (2).



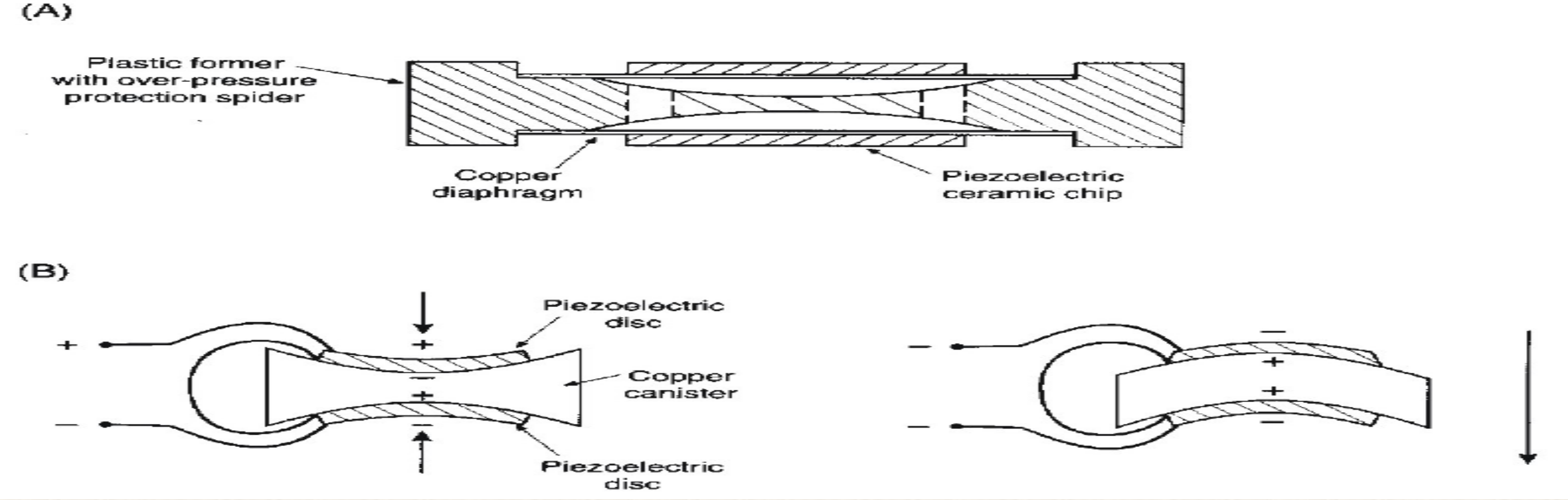
B- Sea survey:

Hydrophone :

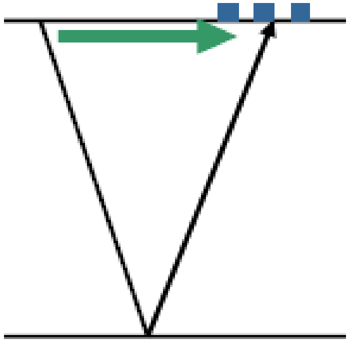
Hydrophones used to detect the pressure variations in water due to a passing seismic wave.

A hydrophone comprises two piezoelectric ceramic discs cemented to a sealed hollow canister.

- A pressure wave squeezes the canister, bending the ceramic and generating a voltage.
- The two discs are connected in series so that the output generated by acceleration of the hydrophone cancels
- Pressure will squeeze ceramics and so produce output.



Deployment



Important considerations

- Need good coupling to the ground – spike
- Mini-arrays to reduce surface wave noise

Offset of geophones

Small offsets
- Near-vertical incidence retains P-energy
- High resolution of subsurface reflectors

➔ **Seismic reflection analysis**

Large offsets
- Improves velocity sensitivity
- Provides horizontal averages only

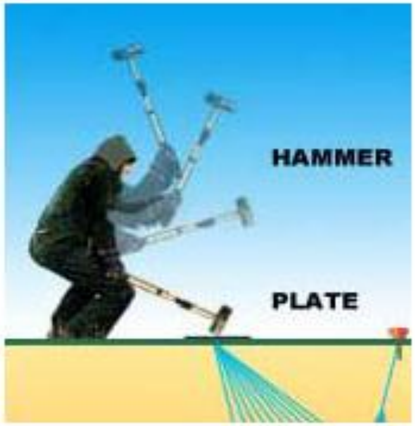
➔ **Seismic refraction analysis**

SEISMIC SOURCES

Seismic sources

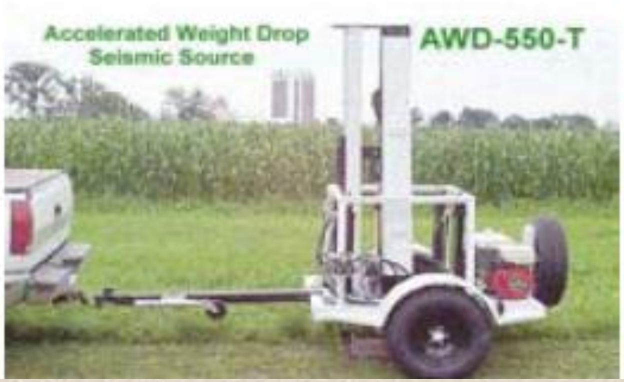
Consider
- **Energy input**
- **Repeatability**
- **Cost**
- **Convenience**

Rifles and guns
- Cheap
- Repeatable – fire into water filled hole
- Shallow targets 0-50m



Sledge hammer
- Cheap
- Repeatable once plate is stable (and with training!)
- Targets 15-50m

Weight drops
- Cheap
- Repeatable – automated
- Targets > 50m

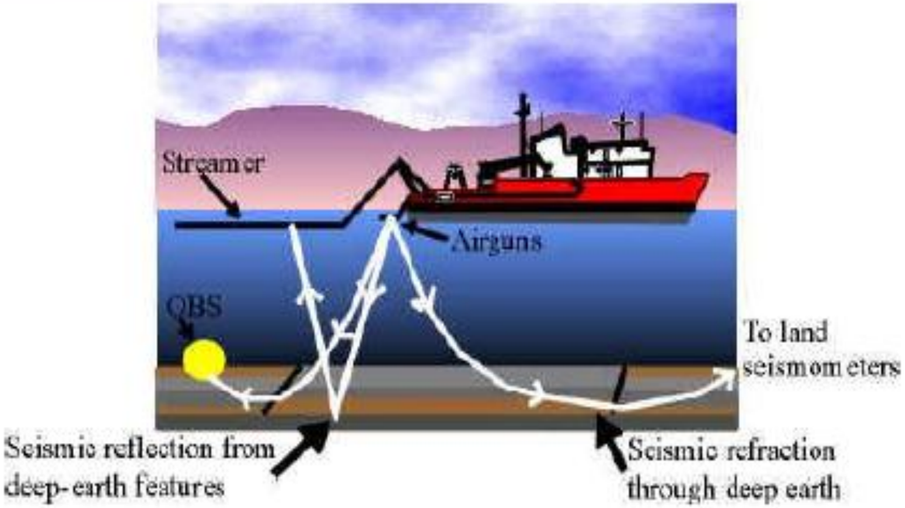


Vibroseis

- No pulse, frequency sweep
- Significant signal with stacking/deconvolution

Explosives

- Various sizes – target depth
- Safety and expense can be an issue



Air guns

- At sea
- Very repeatable
- Large array for big signal

1. Explosives:

- %60 of seismic works use dynamite. It is a mixture of Nitroglycerin + Gelatin + Binding material
- The explosive sources that were placed in the shot holes are similar to those used for blasting during road construction.

Advantages

- has sharp peak
- Decrease surface noise
- has wide range of frequency

Disadvantages

- requires drill hole
- dangerous
- 3- Stack of charge

2- Weight drops:

- composed of a rectangular steel plate, 2 to 3 tons usually, and dropped 3 to 4 times at height of 3 m
- used in desert areas.

Advantage

- Not dangerous
- Easley used , cheap
- Not distort the surface

Disadvantage

- Record is weak, so dropped 3-4
- Low frequency
- Not used in mountainous area

3- Vibroseis:

- It is used to generate seismic wave in the form of a wave train of control energy.
- Always 4 vibrators are used

Advantage

- controlled frequency is
- The energy is controlled
- Not distort the surface

Disadvantage

- Not suitable when soil is too thick
- Expensive
- Not used in mountainous area

4- Sledge Hammer:

- used to generate seismic wave of small energy,
- used for shallow depths of investigation

Advantage

- very cheap
- easily used
- Not distort the surface

Disadvantage

- Give small energy
- used only for shallow depths
- Low frequency

ADVANTAGES & DISADVANTAGES OF SEISMIC METHODS

Advantages and Disadvantages of Seismic Methods

Advantage
Can detect both lateral and depth variations in a physically relevant parameter; seismic velocity.
Can produce detailed images of structural features present in the subsurface.
Can be used to delineate stratigraphic and in some instances depositional features.
Response to seismic wave propagation is dependent on rock density and a variety of physical (elastic) constants. Thus, any mechanism of changing these constants (porosity changes, permeability changes, compaction, etc.) can, in principle, be delineated via the seismic methods.
Direct detection of hydrocarbons, in some instances, is possible.

Disadvantage
Amount of data collected in a survey can rapidly become overwhelming.
Data is expensive to acquire and the logistics of data acquisition are more intense than other geophysical methods.
Data reduction and processing can be time consuming, can require sophisticated computer hardware, and can demand considerable expertise.
Equipment for the acquisition of seismic observations is, in general, more expensive than equipment required for the other geophysical surveys
Direct detection of common contaminants, present at levels commonly seen in hazardous waste spills, is not possible.

COMPARISON BETWEEN REFRACTION & REFLECTION SEISMIC METHODS

REFRACTION METHOD		REFLECTION METHOD	
Advantage	Disadvantage	Advantage	Disadvantage
Refraction observations generally employ fewer source and receiver locations, thus they are relatively cheap to acquire.			Because many source and receiver locations must be used to produce meaningful images of the Earth's subsurface, reflection seismic observations can be expensive to acquire.
Little processing is done on refraction observations with the exception of trace scaling or filtering to help in the process of picking the arrival times of the initial ground motion.			Reflection seismic processing can be very computer intensive, requiring sophisticated computer hardware and a relatively high-level of expertise. Thus, the processing of reflection seismic observations is relatively expensive.
Because such a small portion of the recorded ground motion is used, developing models and interpretations is no more difficult than our previous efforts with other geophysical surveys.			Because of the overwhelming amount of data collected, the possible complications imposed by the propagation of ground motion through a complex earth, and the complications imposed by some of the necessary simplifications required by the data processing schemes, interpretations of the reflection seismic observations require more sophistication and knowledge of the process.

REFRACTION METHOD		REFLECTION METHOD	
Advantage	Disadvantage	Advantage	Disadvantage
	Refraction seismic observations require relatively large source-receiver offsets (distances between the source and where the ground motion is recorded, the receiver).	Reflection seismic observations are collected at small source-receiver offsets.	
	Refraction seismic *only* works if the speed at which motions propagate through the Earth increase with depth.	Reflection seismic methods can work no matter how the speed motions porpagate through the Earth vary with depth.	
	Refraction seismic observations are generally interpreted in terms of layers. These layers can have dip and topography.	Reflection seismic observations can be more readily interpreted in terms of complex geology.	
	Refraction seismic observations only use the arrival time of the initial ground motion at different distances from the source (i.e., offsets).	Reflection seismic observations use the entire reflected wavefield (i.e., the time-history of ground motion at different distances between the source and the receiver).	
	A model for the subsurface is constructed by attempt to reproduce the observed arrival times.	The subsurface is directly imaged from the acquired observations.	