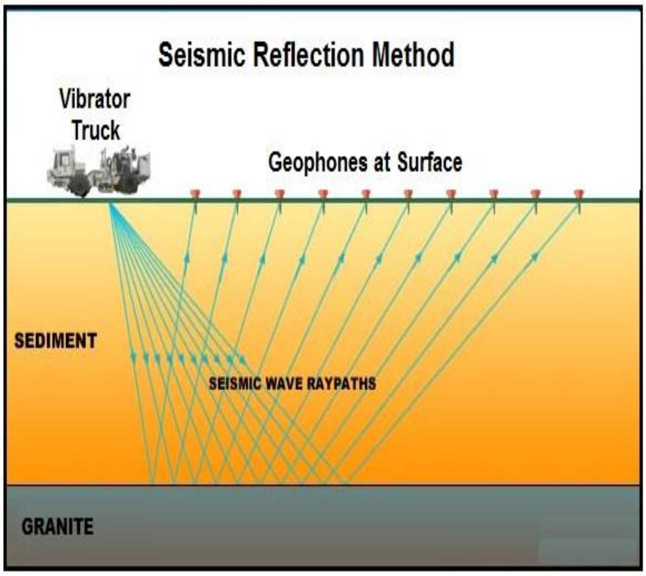
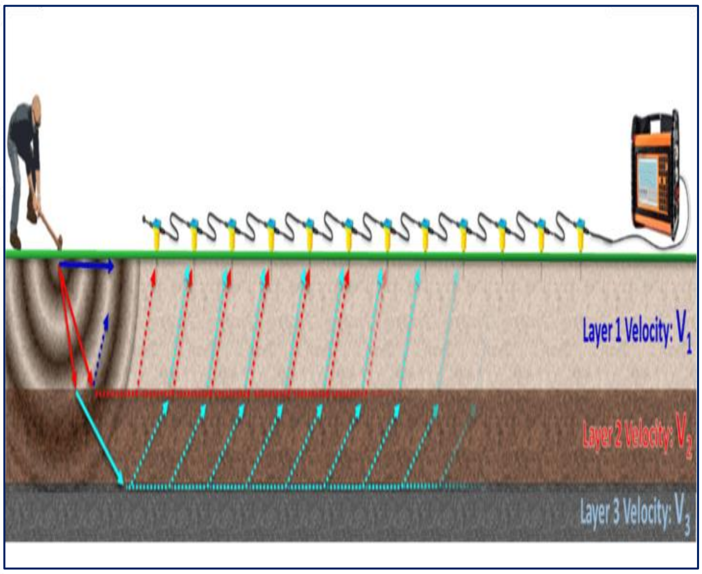
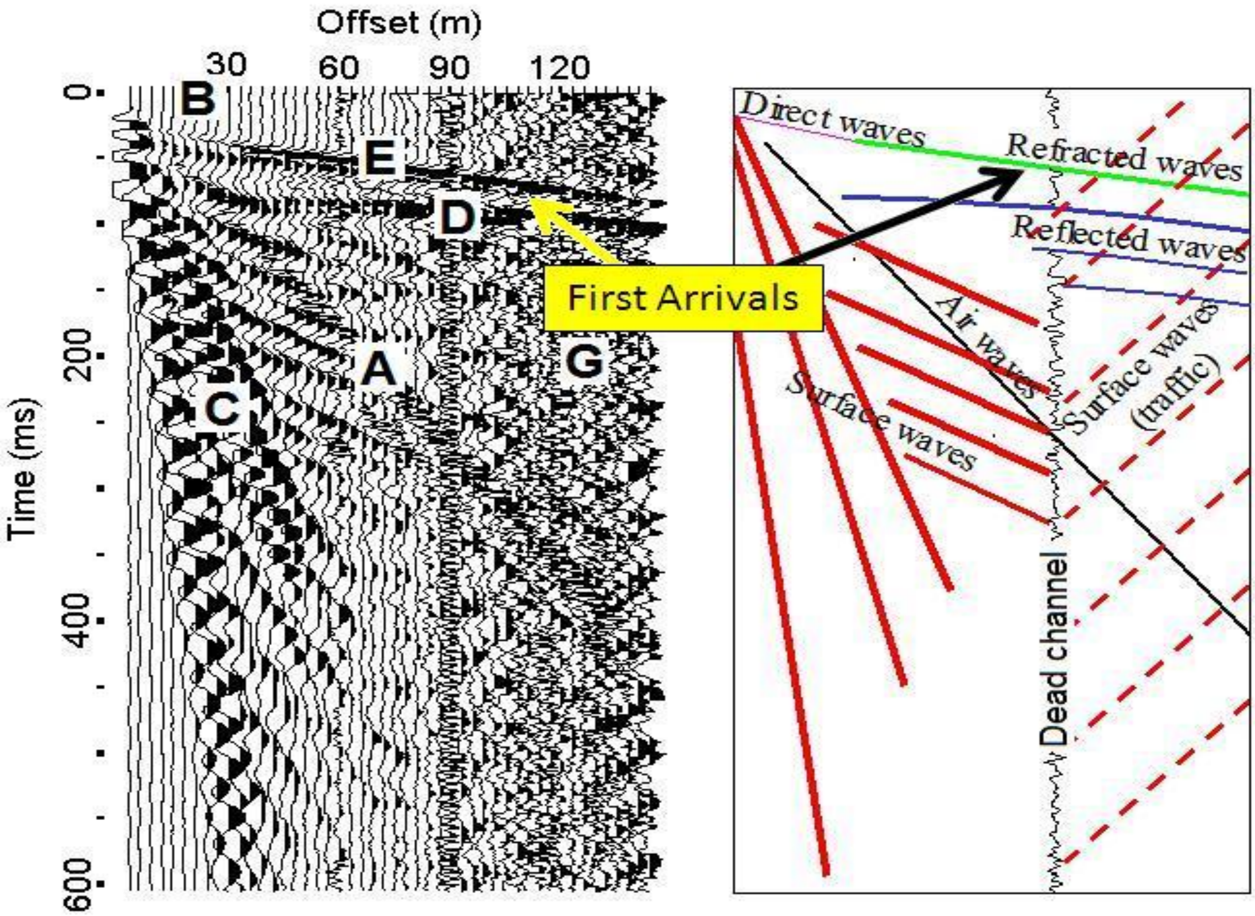
Chapter 3

III. THE SEISMIC METHOD







III.1 Introduction

To determine the geological composition of a region, geophysicists carry out various surface measurements. To define, in particular, the contour lines of the geological layers, they employ seismic prospecting methods: an explosion is triggered near the surface layers of the terrain being studied, and the artificially induced shock wave propagates through the subsoil. Since the geological layers have different densities and compositions, when the wave front crosses the boundary separating two layers, some of the transported energy is reflected and refracted towards, the surface due to the discontinuity in the elastic constants of the layers. At the surface, seismographs record the arrival of the energy released by the explosion during the 4 or 5 seconds that follow it. By pointing these arrivals on the seismographs (time section) and cross-referencing their indicators with those provided by other measurements, the seismologist refines his knowledge of the subsoil, position of discontinuities, speed of propagation, sometimes absorption coefficients.

There are two main methods of seismic prospecting: **Reflection seismics** and **Refraction seismics**; both aim to determine the depth and shape of the geological discontinuities that make up the subsoil.

A. Reflection Seismics

Widely used for industrial exploration, and responsible for the discovery of numerous oil deposits, the principle of this method is simple. Seismic waves **(elastic)** generated by the explosion of charges of a few kilograms of explosives (TNT) propagate through sedimentary or metamorphic terrains and are reflected off the boundaries between formations of different types. The reflected waves are recorded at the Earth's surface by a series of seismographs (most often 24 or 48). If the propagation speed of the seismic waves is known at different levels, measuring the propagation times allows the depth of these levels to be calculated.

B. Seismic Refraction

Its purpose is to determine the propagation speeds of seismic waves as well as the depth of the different interfaces. The seismic wave is refracted at the interface and returns to the seismographs located on the ground surface. The path of the refracted wave allows us to determine the depth of the interfaces.

III.2 Theoretical Overview

3.2.1Types of Seismic Waves

• There are several different kinds of seismic waves, and they all move in different ways.

- ➢ Body waves

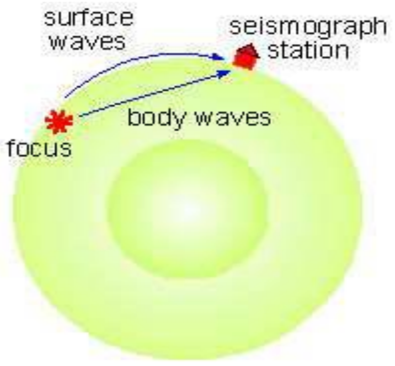
• P or Primary waves

• S or Secondary waves

- ➢ Surface waves

• Love waves

• Rayleigh wave

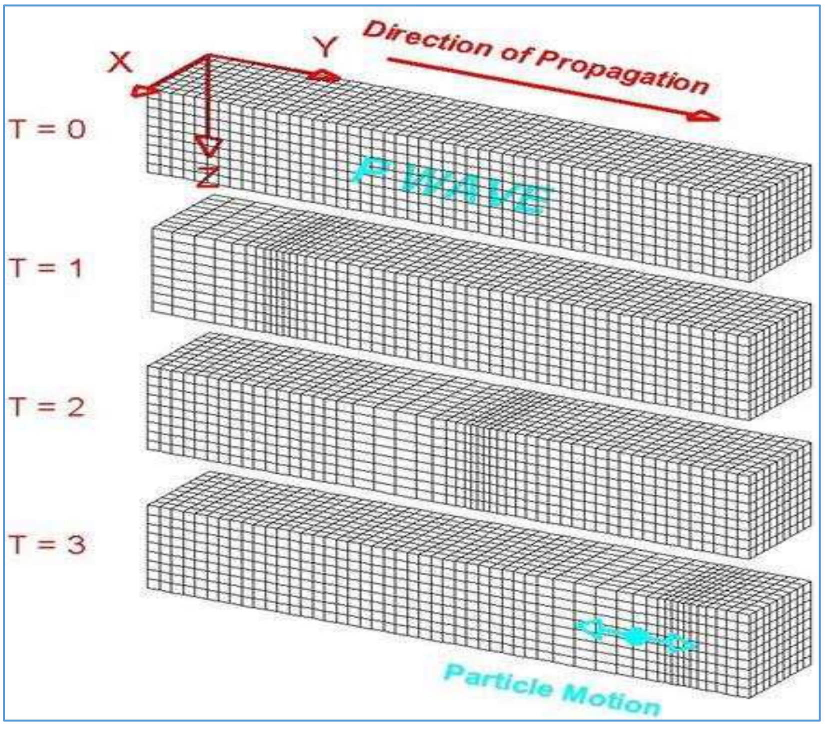


For shallow studies, the use of **S-waves**, like that of surface waves, is not common (even though they are the subject of research programs), and studies are generally limited to **P-waves**.

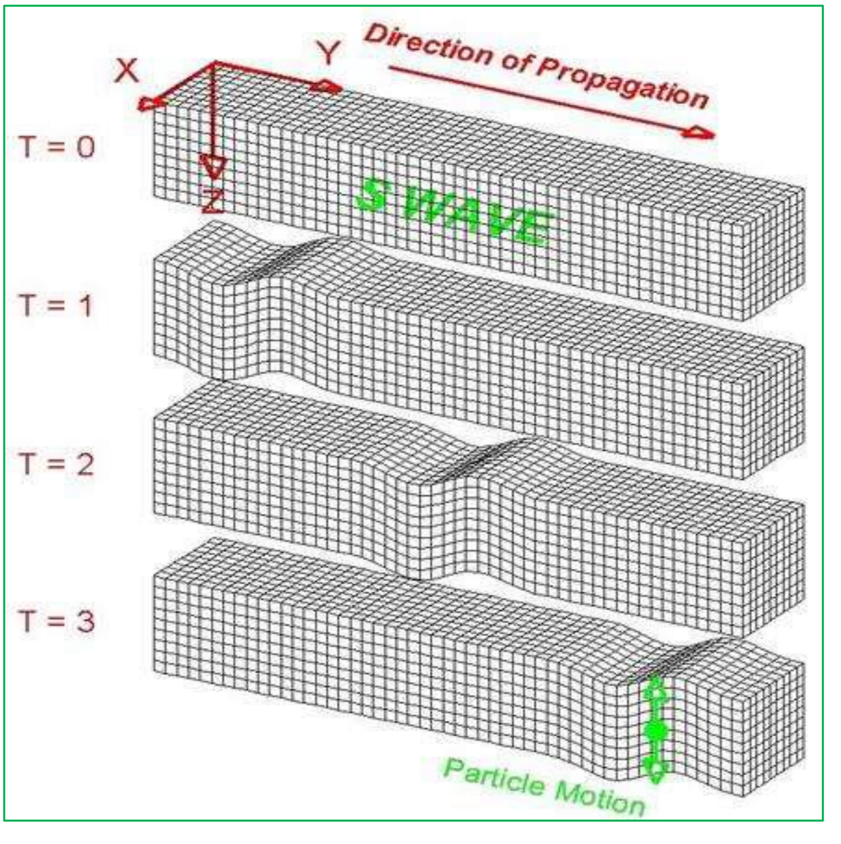
In a homogeneous medium, measuring the speed simply consists of measuring the arrival time, td, of the direct P-wave. At a distance Δ from the point of disturbance, we have:

$$V_p = \frac{\Delta}{t_d}$$

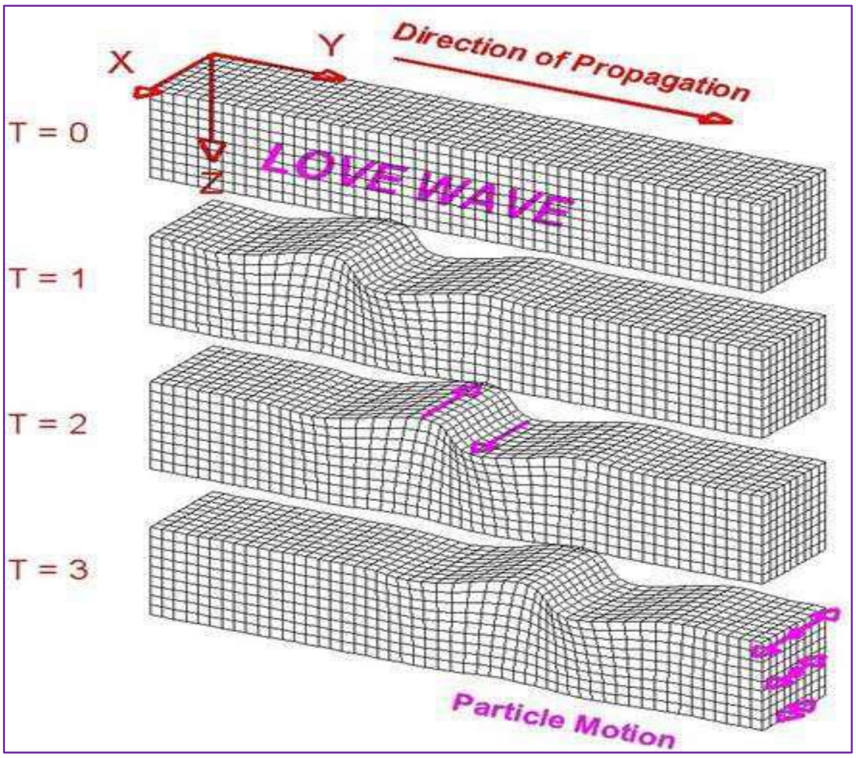
Longitudinal waves (P waves) cause only a change in volume as they propagate through a medium; areas of expansion and compression are formed in that medium in the direction of propagation.



Transverse waves (S waves) cause a change in the shape of the body. As they propagate, the particles of the medium move relative to each other in a direction perpendicular to the direction of propagation.



SURFACE WAVES

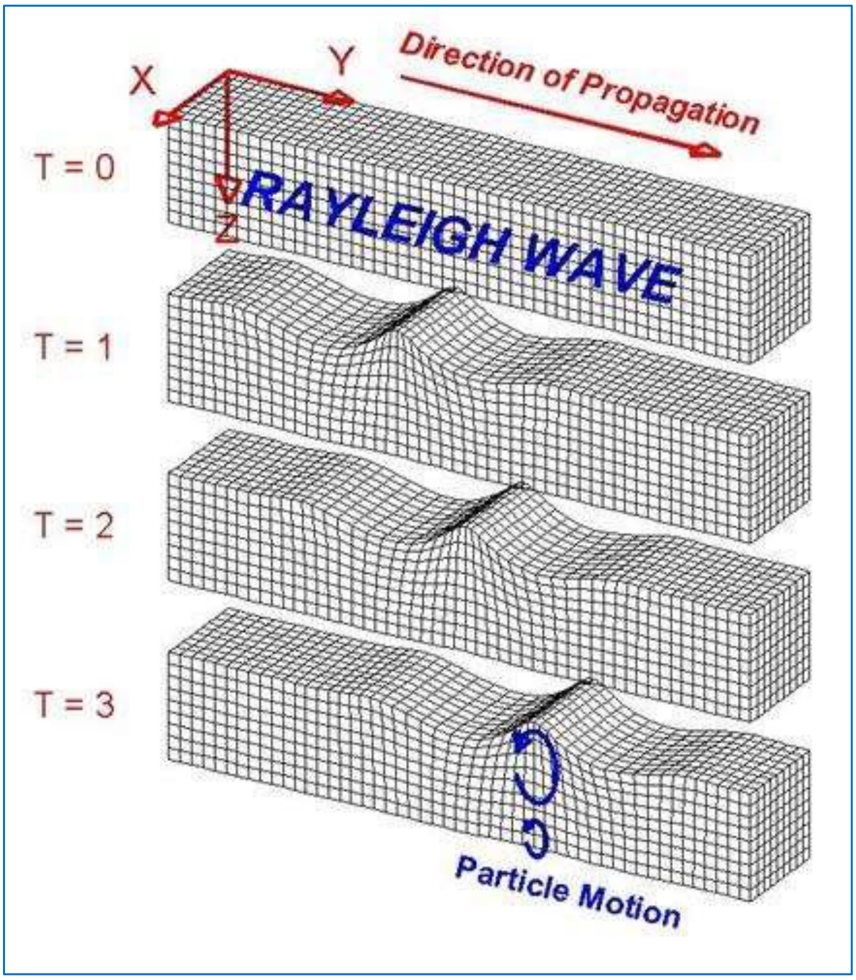


• Travelling through the crust, surface waves are of a lower frequency than body waves, and are easily distinguished on a seismogram as a result.

• Though they arrive after body waves, it is surface waves that are almost enitrely responsible for the damage and destruction associated with earthquakes.

• This damage and the strength of the surface waves are reduced in deeper earthquakes.

RAYLEIGH WAVES



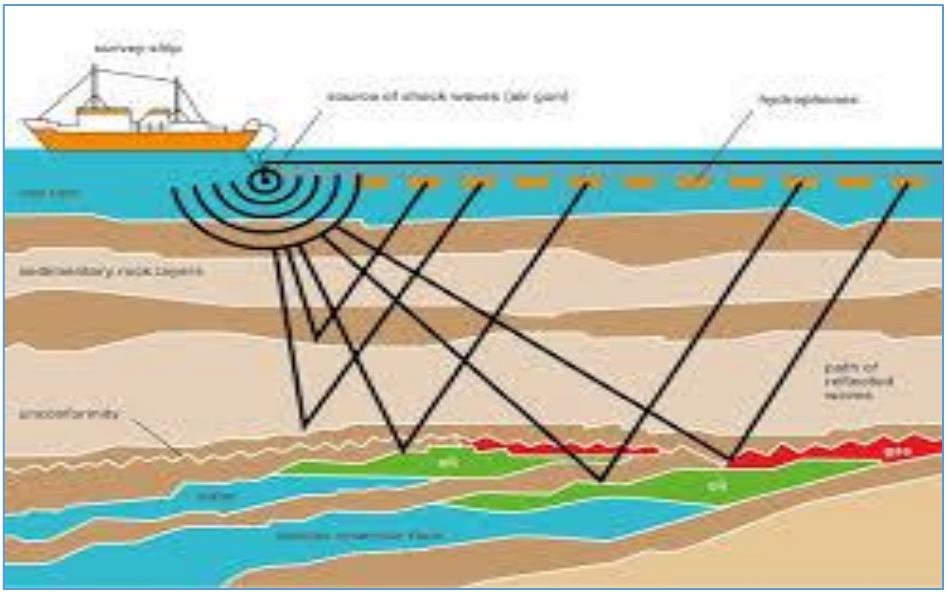
• The other kind of surface wave is the Rayleigh wave, named for John William Strutt, Lord Rayleigh, who mathematically predicted the existence of this kind of wave in 1885.

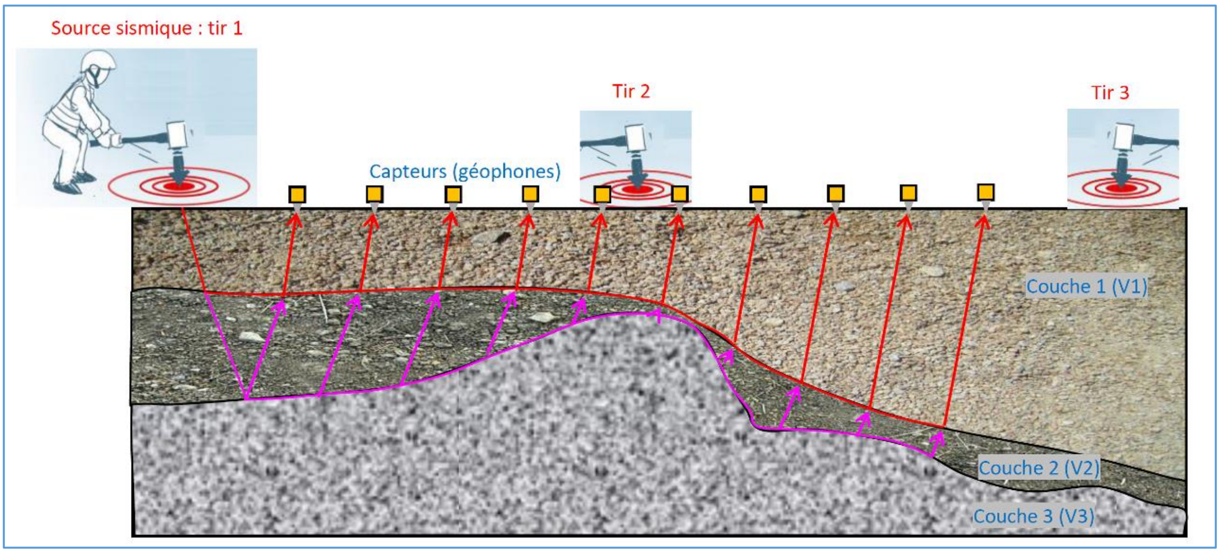
• A Rayleigh wave rolls along the ground just like a wave rolls across a lake or an ocean. Because it rolls, it moves the ground up and down, and sideto- side in the same direction that the wave is moving.

• Most of the shaking felt from an earthquake is due to the Rayleigh wave, which can be much larger than the other waves.

III.2. Reflection and refraction of seismic waves

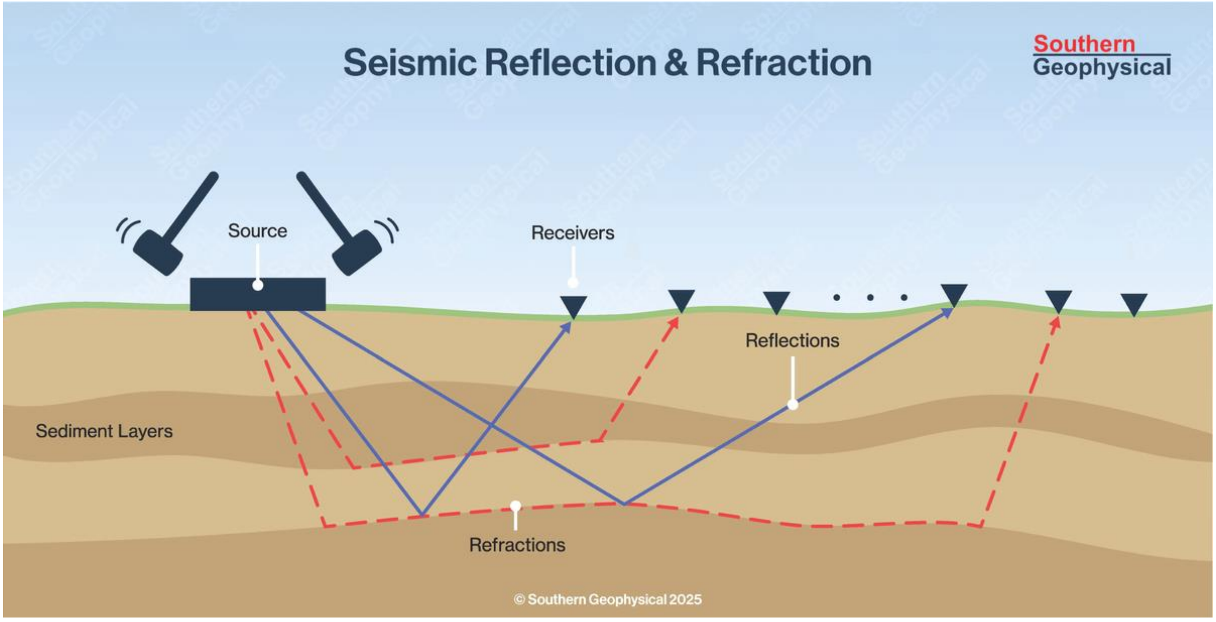
Seismic Reflection





Seismic refraction

Seismic refraction allows us to evaluate the mechanical characteristics, in particular the hardness, of the soils constituting the subsoil, through the modeling of these in the form of a cross-section of the distribution of velocities increasing with depth.



Seismic Reflection and Refraction

Reflection and refraction of seismic waves (Snell's law or Descartes' law)

We consider two **homogeneous media** W1 and W2 with respective seismic velocities VP1, VS1 and VP2, VS2 separated by a plane boundary (Figure.)

$$\frac{\sin\alpha_1}{V_1} = \frac{\sin\alpha_2}{V_2}$$

$$\frac{\sin\alpha}{V_1} = \frac{\sin\beta}{V_2}$$

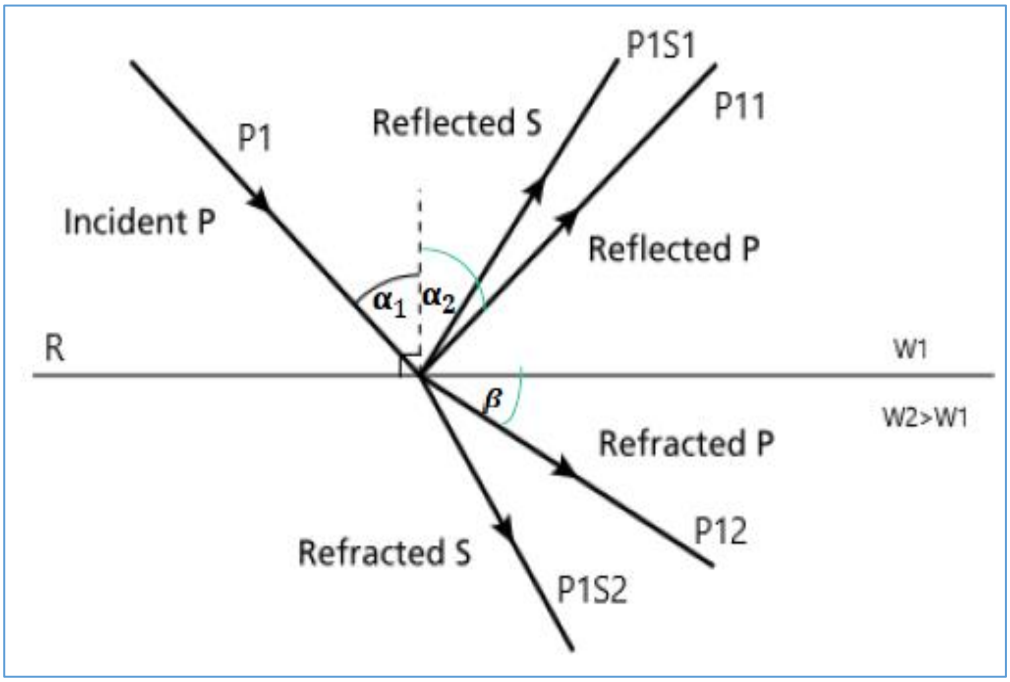
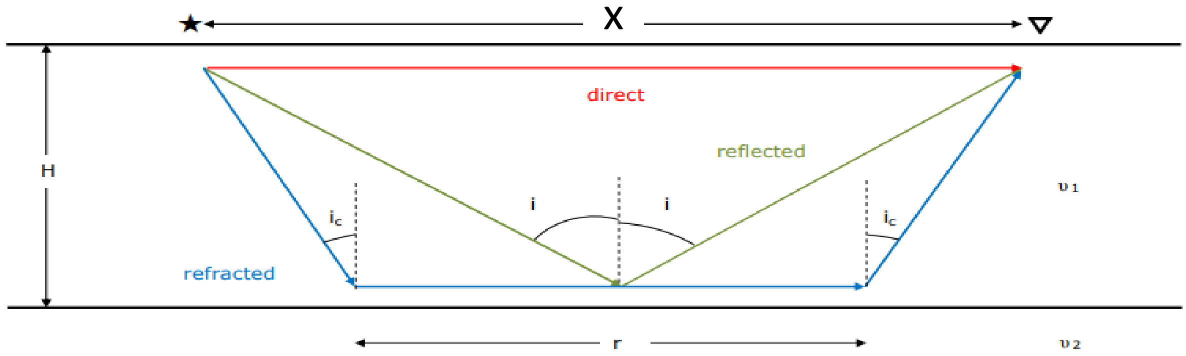


Figure. Reflection and transmission of the wave across a seismic boundary **(Snell's or Descartes' law).**



1.1 The direct wave

This is the easy one! In a layered medium the direct wave travels straight along the surface with velocity **V_1**. At distance **x** clearly the travel time **t_{dir}** will be:

$$t_{dir} = \Delta / v_1$$

1.2 The reflected wave

To calculate the reflected wave we need to do a little geometry. The length of the path the ray travels in layer 1 is obviously related to the distance in a non-linear way. The travel time for the reflection is given by:

travel time reflected wave

$$t_{refl} = \frac{2}{v_1}\sqrt{(\Delta/2)^2 + h^2}$$

$$t = \frac{1}{V_1}\sqrt{x^2 + 4h^2}$$ or

The propagation **speed of seismic waves** depends on **the velocity** within the **matrix**, the **porosity**, and the **seismic velo**city in the **fluid** filling the pores. The empirical relationship:

$$\frac{1}{V} = \frac{\emptyset}{V_f} + \frac{(1-\emptyset)}{V_m}$$

Conditions for Seismic Surveying

The conditions favorable for conducting seismic surveys are created by the existence of:

- seismic boundaries that coincide with geological boundaries,
- extensive seismic boundaries with gentle slopes (2°-10°),
- a thin weathered layer,
- a shallow water table;
- a relatively flat topography in the area of investigation.

III.3 Wave Propagation in Discontinuous Media

III.3.1 Case of a Bilayer

Calculation of the (hodochrone) equation giving the arrival time of the first refracted wave as a function of the distance X between the emission point and the receiver in the case of a bilayer composed of a first slow layer with velocity V1 and finite thickness, resting on a layer of infinite thickness with velocity V2 > V1.

NB Un hodochrone est un graphique qui représente le temps d'arrivée des ondes sismiques en fonction de leur distance par rapport à l'épicentre d'un séisme

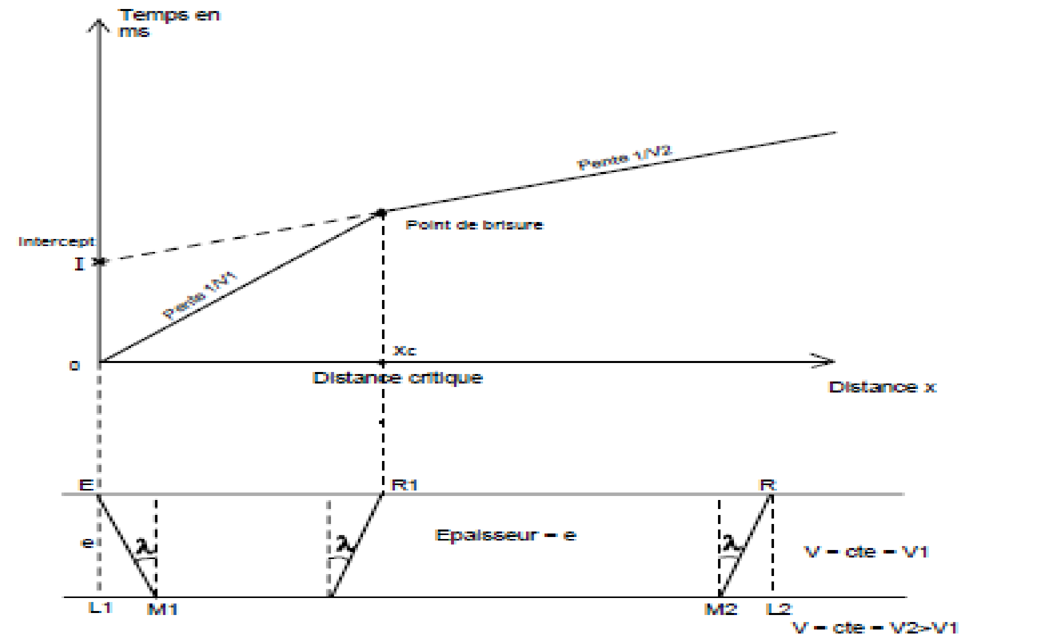


Figure 5

(λ) or (i)

According to Snell's law (ground has two layers), we have:

$$T = \frac{2e}{V_1 cos\lambda} + \frac{X}{V_2} - \frac{2e.tg\lambda}{V_1} sin\lambda = \frac{X}{V_2} + \frac{2e}{V_1}\left(\frac{1}{cos\lambda} - tg\lambda sin\lambda\right) = \frac{X}{V_2} + \frac{2e}{V_1}\left(\frac{1 - sin^2\lambda}{cos\lambda}\right)$$

$$cos^2\lambda = 1 - sin.^2\lambda$$

$$T = \frac{EM1 + M2R}{V_1} + \frac{M1}{V_2} \quad V_2 = \frac{V_1}{sin\lambda}$$

$$L1M1 = e.tg\lambda = M2L2 \;\Rightarrow\; \frac{M1M2}{V_2} = \frac{X}{V_2} - \frac{2L1M1}{V_2} \quad et\; T = \frac{2e}{V_1 cos\lambda} + \frac{X - 2e\, tg\lambda}{V_2}$$

Finally,

$$T = \frac{X}{V_2} + \frac{2e}{V_1} cos\lambda$$

This last expression is the time-distance equation of a wave refracted from a straight line with a slope of 1/V2 and a y-intercept, or **intercept time** t_i or t_0

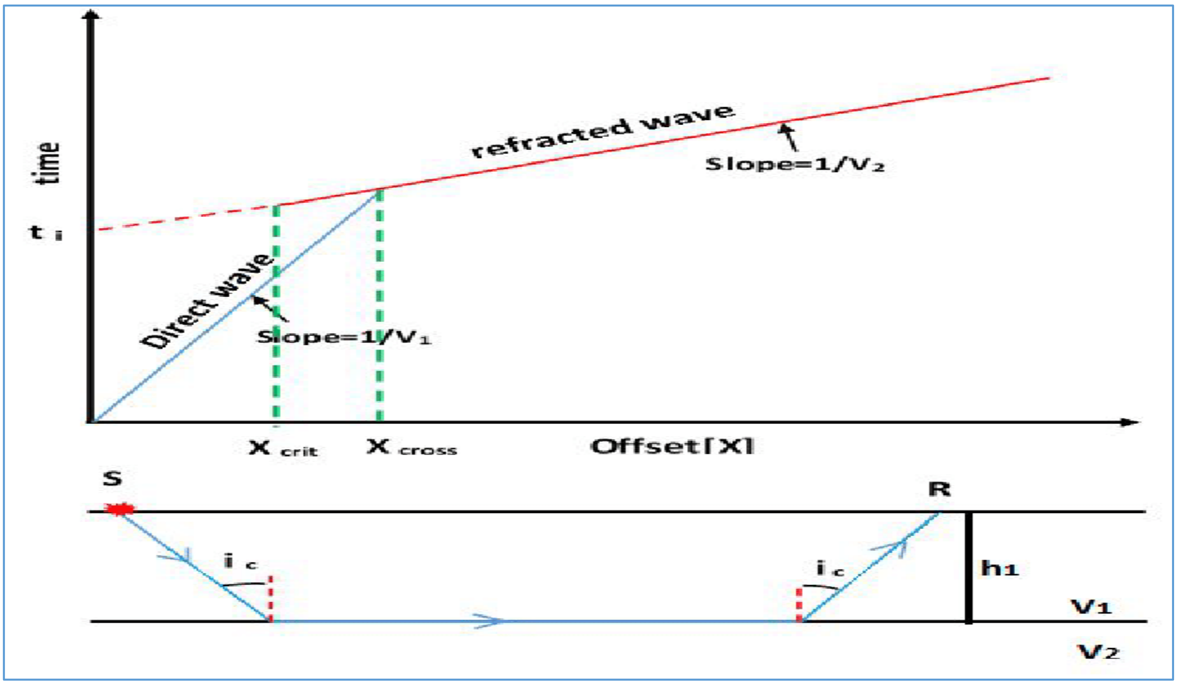
$$t_0 = \frac{2e}{V_1} cos\lambda$$ or $$t_i = \frac{2h\cos i_c}{v_1} \Rightarrow h = \frac{t_i v_1}{2\cos i_c}$$

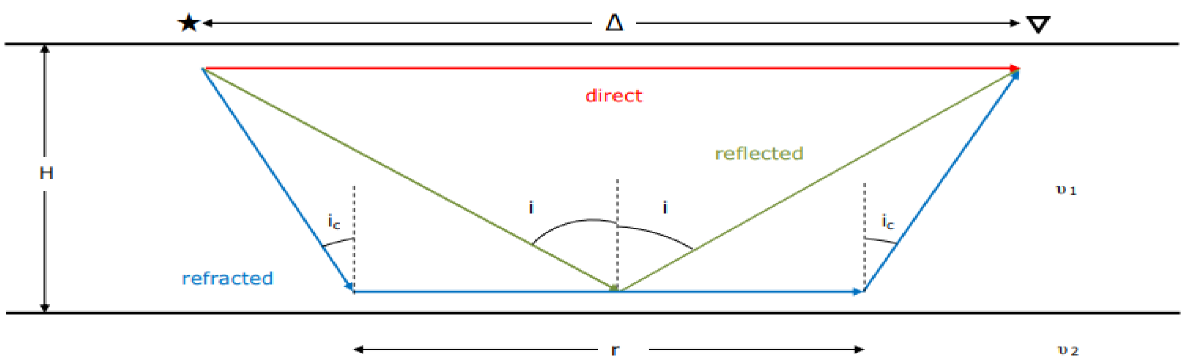
where

- h = depth of refractor
- i_c= critical angle
- V_1 = velocity of upper layer

Intercept time (t_i) is the **extrapolated travel time** of the **refracted (head) wave** back to **zero offset** on a travel-time curve.

On the "time-distance" graph, the lines with slopes of 1/V1 and 1/V2 intersect at a point called the break point. For receptions to the left of this break point, the direct travel time is less than the refracted travel time, and conversely for receptions to the right of the break point.



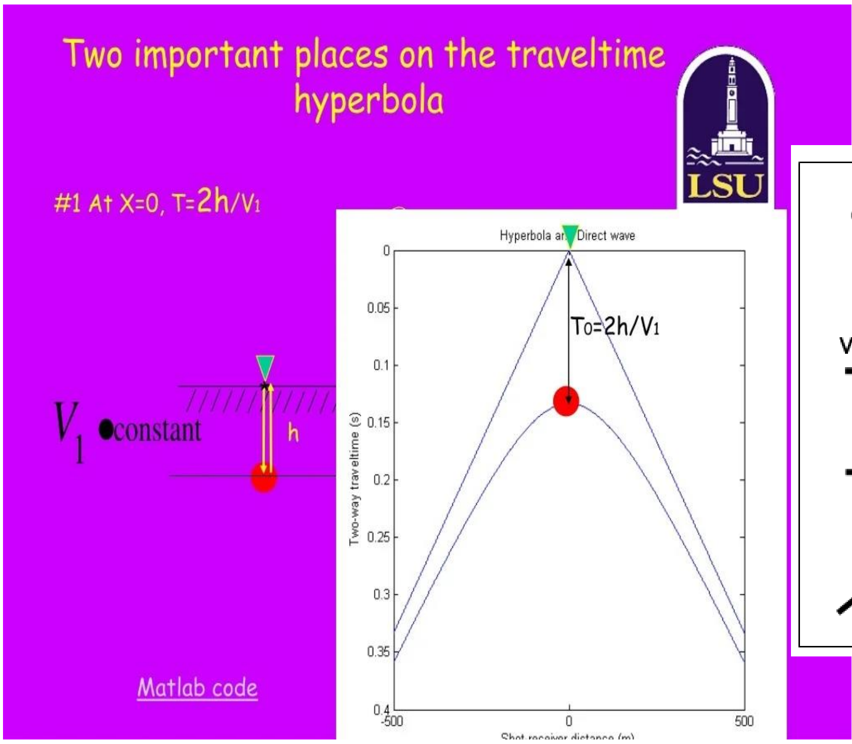


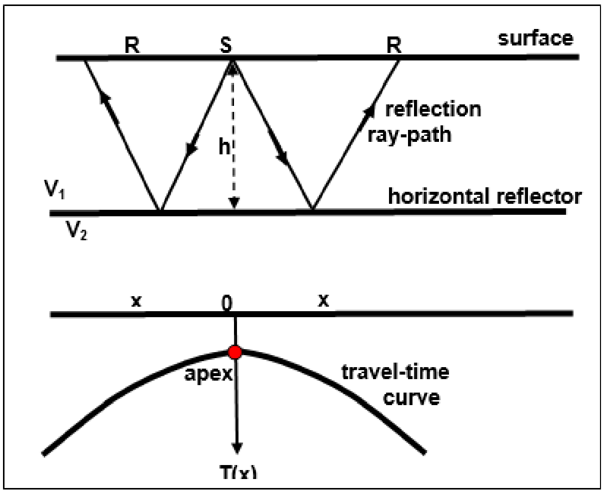
a- The reflected wave

To calculate the reflected wave we need to do a little geometry. The length of the path the ray travels in layer 1 is obviously related to the distance in a non-linear way. The travel time for the reflection is given by:

travel time reflected wave $$t_{refl} = \frac{2}{v_1}\sqrt{(\Delta/2)^2 + h^2} \quad \text{or} \quad t = \frac{1}{V_1}\sqrt{x^2 + 4h^2}$$

In refraction seismology this arrival is often of minor interest, as the distances are so large that the reflected wave has merged with the direct wave. Note that this has the form of a **hyperbola**.





The hyperbolic travel-time = $$T(x) = \sqrt{T_0^2 + \frac{x^2}{V^2}}$$

b- The refracted wave

As we can easily see from the figure above the refracted wave needs a more involved treatment.

Refracted waves correspond to energy which propagates horizontally in medium 2 with the velocity v2. This can only happen if the emergence angle i_2 is 90°, i.e.

critical angle $$\frac{\sin i_c}{v_1} = \frac{\sin 90°}{v_2} = \frac{1}{v_2} \Rightarrow \sin i_c = \frac{v_1}{v_2} \qquad (4)$$

Travel time refracted wave

$$t_{refr} = \frac{2h\cos i_c}{v_1} + \frac{\Delta}{v_2}$$

$$T_1 = \frac{x}{V_2} + \frac{2h_1\cos l_1}{V_1}$$

$$= t^i_{refr} + \frac{x}{v_2}$$

III.3.1.1 Thickness Calculation

The dromochronic branch allows us to define the velocities V1 and V2, inverses of the slopes of the lines, and consequently the value of λ, such that sinλ = V1/V2. The intercept time t0 then allows us

to: **Calculate the thickness (e) or (h) of the layer.**

$$t_0 = \frac{2e}{V_1}\cos\lambda \quad \Rightarrow e = \frac{t_0 V_1}{2\cos\lambda}$$

$t_0 = t_i$

The thickness of the layer can also be calculated from the x-coordinate of the break point Xc or critical distance.

$$e = \frac{X_c}{2}\sqrt{\frac{V_2 - V_1}{V_2 + V_1}}$$

or

overtaking distance $\Delta_u = 2h\sqrt{\frac{v_2 + v_1}{v_2 - v_1}}$

$\Delta = x$

III.3.2 Case of Three-layer

III.3.2.1 The refraction from layer 2

The arrival time t_2 of the refraction from layer 2 is given by: $t_2 = \frac{2h_1\cos i_{12}}{v_1} + \frac{\Delta}{v_2} = t^{i2} + \frac{\Delta}{v_2}$

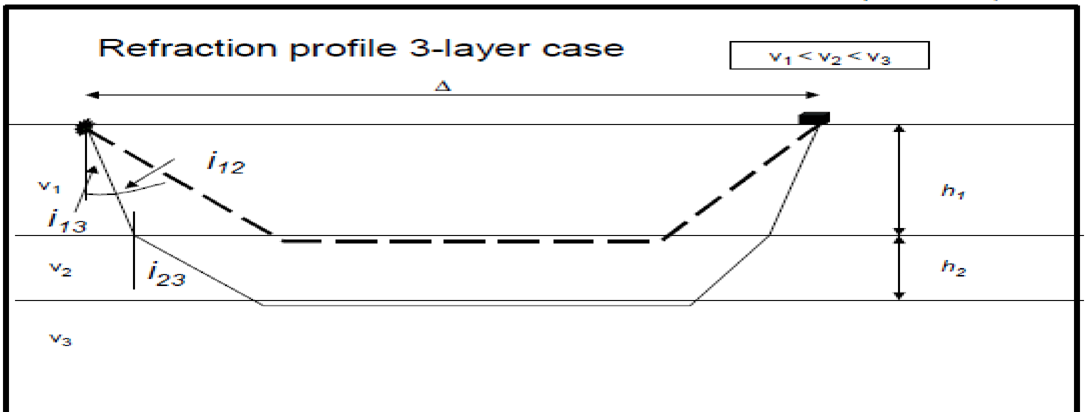


Figure 3: Geometry of 3-layer refraction experiment.

arrival time t3 of the refracted wave in layer 3

$$t_3 = \underbrace{\frac{2h_1\cos i_{13}}{v_1} + \frac{2h_2\cos i_{23}}{v_2}}_{t^{i3}} + \frac{\Delta}{v_3} = t^{i3} + \frac{\Delta}{v_3}$$

III.3.2.2 Determining the velocity depth model for the 3-layer case

$$h_1 = \frac{v_1 t^{i2}}{2\cos i_{12}} \quad \text{, where} \quad i_{12} = \arcsin\frac{v_1}{v_2}$$

an intermediate intercept time t*

$$t^* = t^{i3} - \frac{2h_1\cos i_{13}}{v_1} \quad \text{, where} \quad i_{13} = \arcsin\frac{v_1}{v_3}$$

Using t^* calculate the thickness h_2 of layer 2

$$h_2 = \frac{v_2 t^*}{2\cos i_{23}} \quad \text{, where} \quad i_{23} = \arcsin\frac{v_2}{v_3}$$

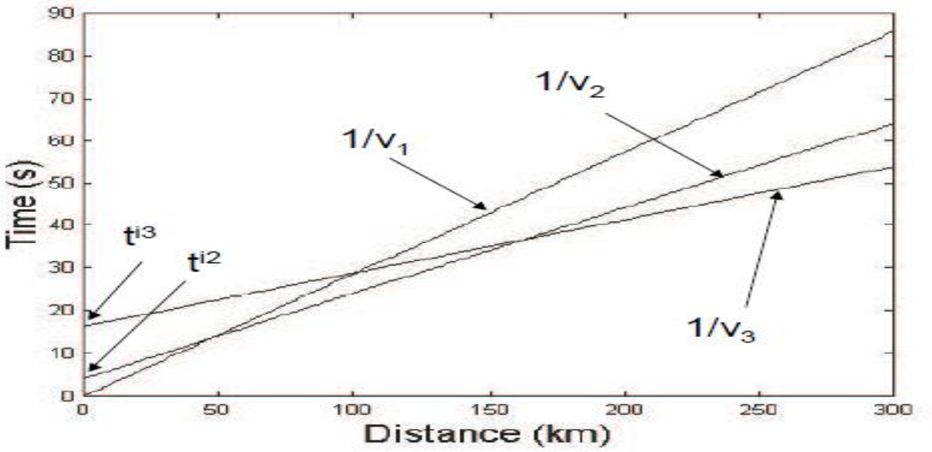


Figure : Travel-time diagram for the 3-layer case

In the model shown in Figure the velocites are v_1=3.5km/s, v_2=5km/s, v_3=8km/s. The layer thicknesses are h_1=10km and h_2=25km.

III.3.3 Case of Inclined 2-layer

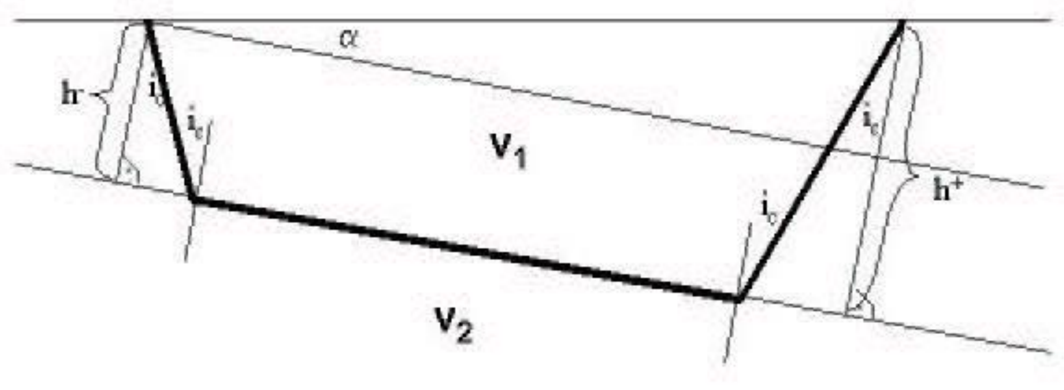


Figure 5: Refracted ray path for the inclined 2-layer case.

Let us develop the *forward problem*, i.e. calculating the travel times of the direct and refracted waves for a given model. With seismic velocities v1 and v2 and inclination angle (Alfa) the travel time of the refracted waves are

$$t^-_{refr} = \frac{2h^- \cos i_c}{v_1} + \frac{\sin(i_c + \alpha)}{v_1}\Delta = t_i^- + \frac{1}{v_2^-}\Delta$$

$$t^+_{refr} = \frac{2h^+ \cos i_c}{v_1} + \frac{\sin(i_c - \alpha)}{v_1}\Delta = t_i^+ + \frac{1}{v_2^+}\Delta$$

Read the intercept times t_i+ and ti- from the travel time diagram. Determine the distances

from the layer interface as

$$h_1 = \frac{t_{i1} V_1}{2 \cos i_c}$$

$$h^- = \frac{v_1 t_i^-}{2 \cos i_c}$$

$$h^+ = \frac{v_1 t_i^+}{2 \cos i_c}$$

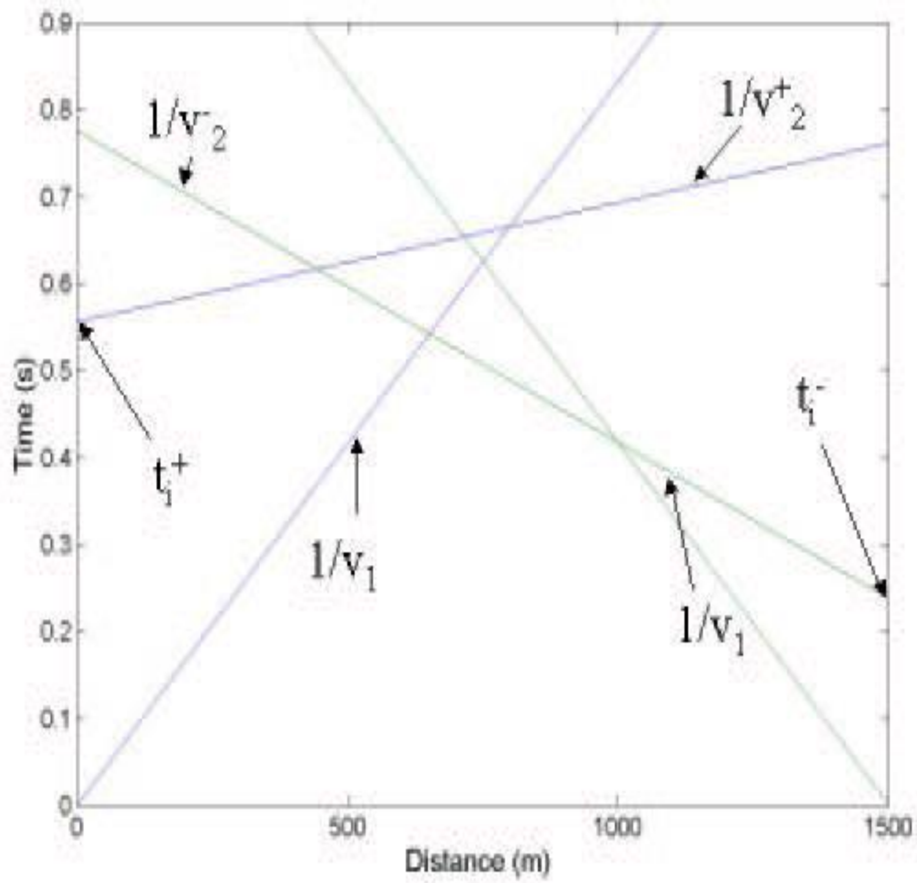


Figure 6: Travel-time diagram for the inclined-layer case for the model in the previous figure.

Case of inclined (Dipping) 2-Layer (other way)

We consider two layers:

- **Layer 1:** velocity V1
- **Layer 2 (refractor):** velocity V2, dipping at angle α

Because the refractor dips, the depth changes along the profile.

a) Critical angle

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

b) Apparent velocities

$$V_{A1} = \frac{V_1}{\sin(i_c - \alpha)}$$

$$V_{A2} = \frac{V_1}{\sin(i_c + \alpha)}$$

c) thickness

$$h_1 = \frac{t_{i1} V_1}{2 \cos i_c}$$

$$h_2 = \frac{t_{i2} V_1}{2 \cos i_c}$$

d) Dip angle α of the refractor

$$\tan \alpha = \frac{V_{A1} - V_{A2}}{V_{A1} + V_{A2}} \cot i_c$$

e) ravel-Time Equations

$$T_1(x) = t_{i1} + \frac{x}{V_{A1}}$$

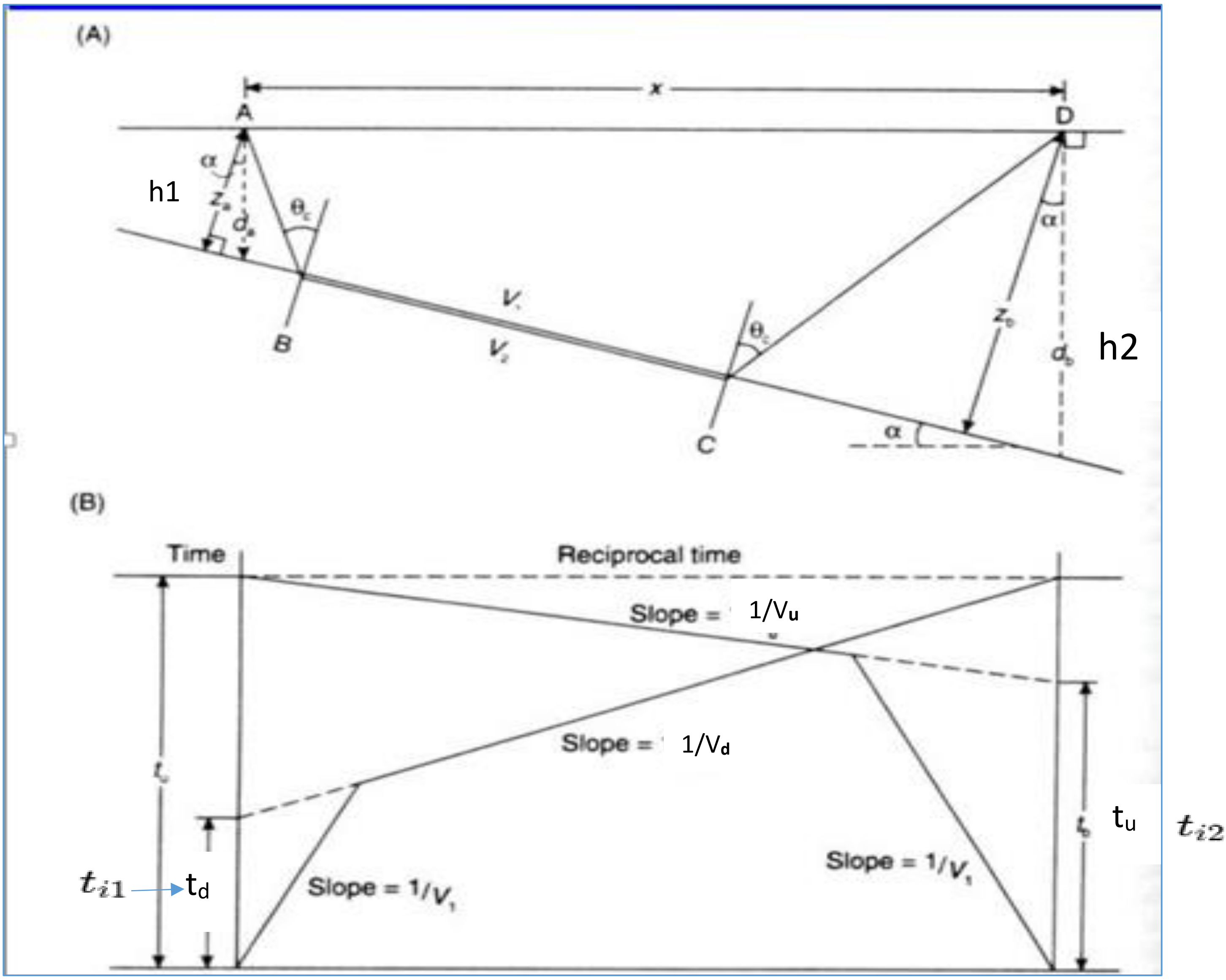
$$T_2(x) = t_{i2} + \frac{x}{V_{A2}}$$

f) Intercept Time Equations

The intercept times differ because the refractor is at different depths beneath each shot point.

$$t_{i1} = \frac{2h_1 \cos i_c}{V_1}$$

$$t_{i2} = \frac{2h_2 \cos i_c}{V_1}$$



III.4 Implementation of a Seismic Device

The standards to which seismic devices must conform are set out below for practical purposes. The detailed theoretical justifications for the need to follow these standards are presented in the section on interpretation.

III.4.1 Number of Sensors and Shots

The simplest system, shown schematically below, appears to be the minimum requirement, both in terms of the number of sensors and the number of shots, below which the information provided risks becoming highly unreliable. Shots O and P, external to the system, are called offset shots or long-range shots. Shots A and B are called end shots, and shot C is called center shot.

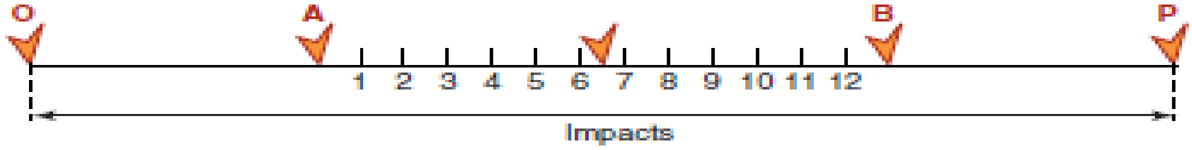
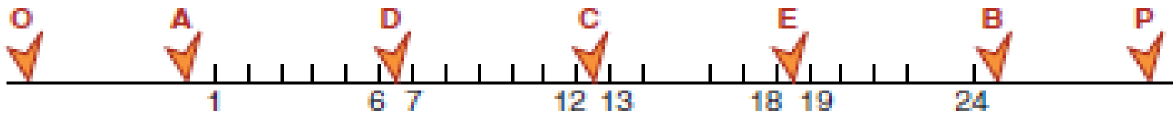


Figure: Minimum acquisition device (12 Trace)

The number of sensors deployed is generally determined by the number of channels available on the recorder, typically 24 channels in seismic refraction laboratories.

For a 24-sensor array, the number of shots depends on the assumed homogeneity of the subsurface and the required accuracy. Based on the minimum array described above, it becomes necessary to use



5 shots for 24 sensors: the two offset shots O and P, the end shots A and B, and one central shot C. When there is a risk of encountering numerous and rapid lateral facies variations, intermediate shots D and E are sometimes added between sensors 6 and 7 on the one hand, and 18 and 19 on the other.

Figure: 24-Trace Device

In the latter case, the dromochromic profile of the device could be as follows:

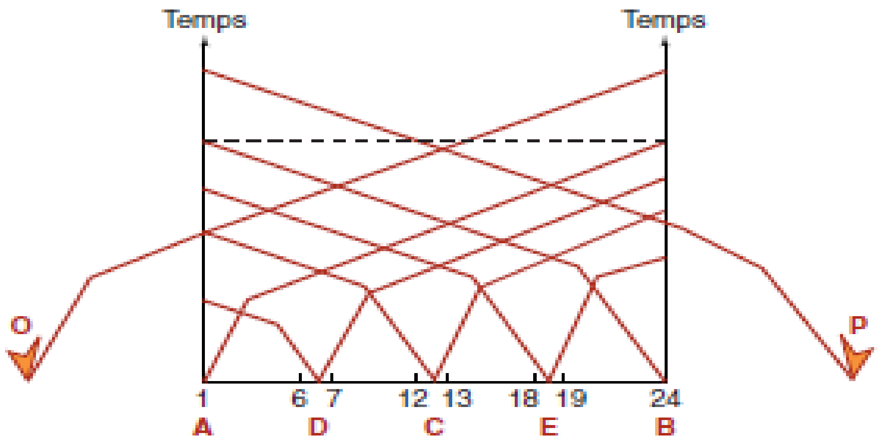


Figure: Dromochronology Associated with a measuring device 24 traces, 7 shots

III.4.1 Device Length

The device length is the distance between the two end shots, or, roughly equivalently, the distance between the two extreme sensors 1-12 or 1-24.

The choice of this length will depend on the depth of the target and the velocity contrast observed between the different intermediate seismic horizons (site velocity law).

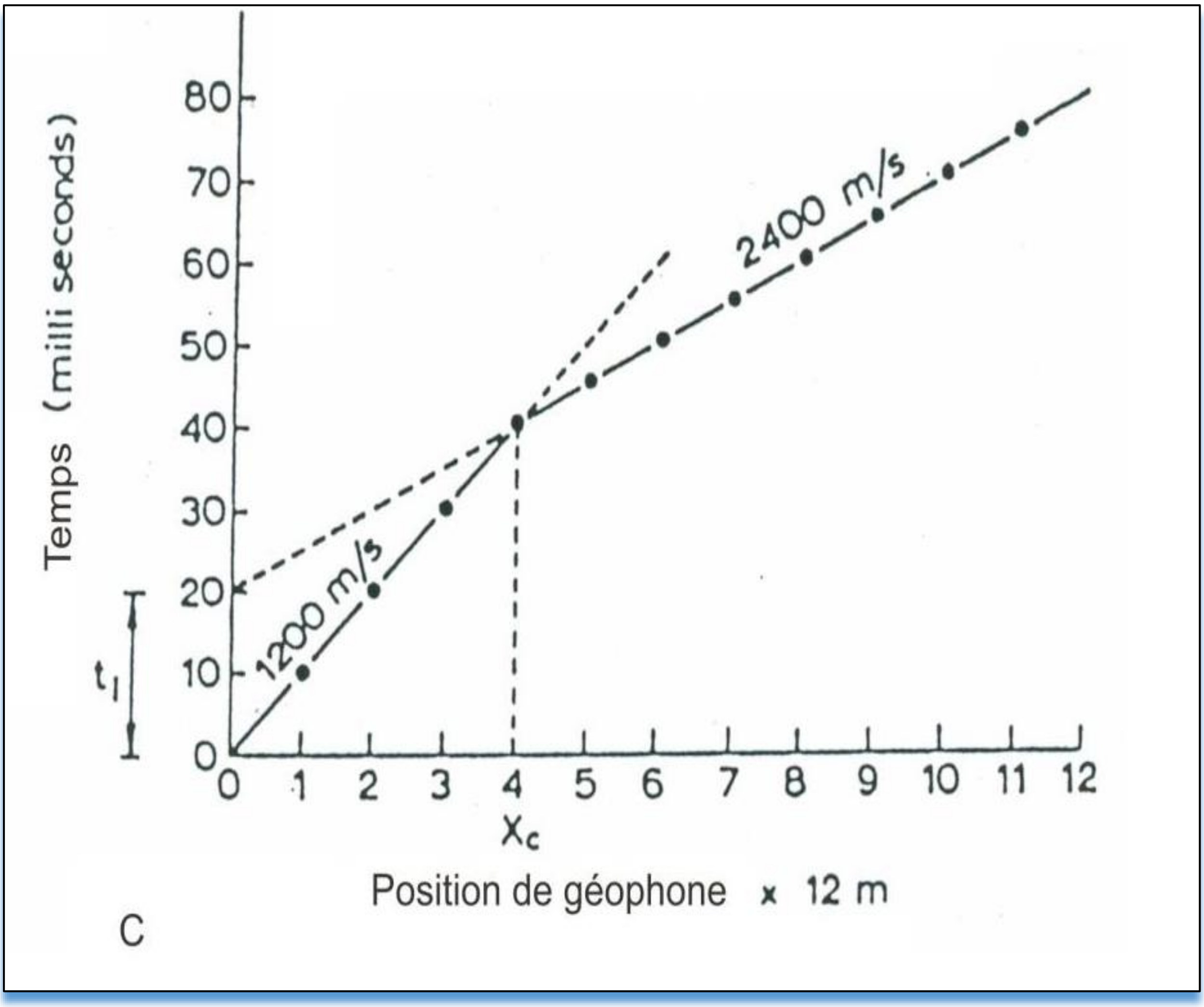
The device length must be long enough to allow for the propagation to the top of the bedrock over approximately half the device length, at least for the shocks from the two end shots A and B.

Assuming that a minimum of information is available on the expected geology of the site, a good approach is to use the formula for calculating the critical distance or break point.

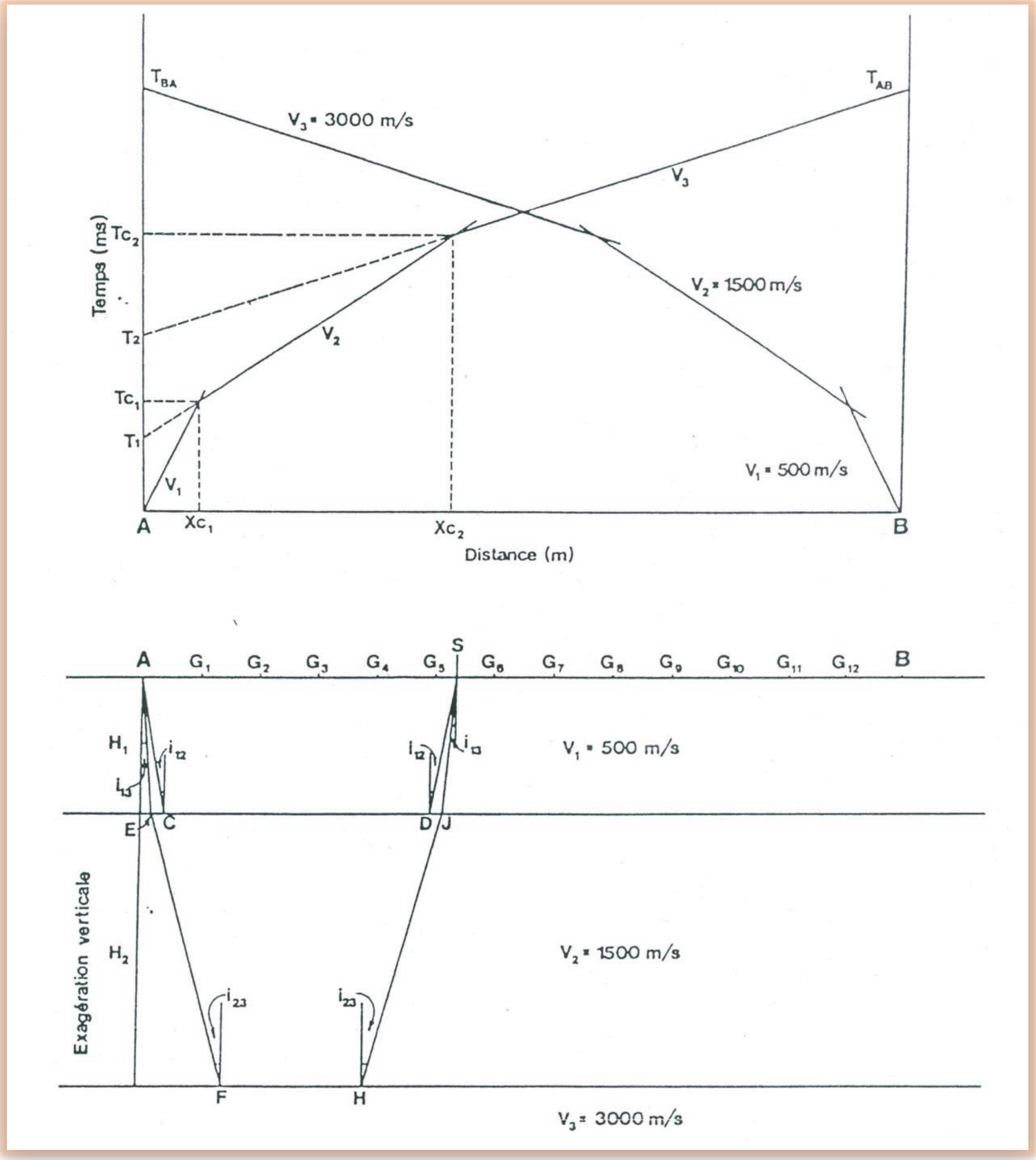
We saw previously (Propagation in discontinuous media) that the critical distance: Xc, that is to say the distance from the point of origin from which the refracted image is pointed at the roof of the medium with velocity V2, was calculated using the formula:

For 2 layers

$$Xc = 2\ h\ \sqrt{(V2 + V1)(V2 - V1)},$$



For 3 layers


T_BA
T_AB
V3 = 3000 m/s
V3
Temps (ms)
Tc2
V2 = 1500 m/s
T2
V2
Tc1
T1
V1 = 500 m/s
V1
A
Xc1
Xc2
B
Distance (m)
S
A G1 G2 G3 G4 G5 G6 G7 G8 G9 G10 G11 G12 B
H1
i12
i13
V1 = 500 m/s
E
C
D
J
Exagération verticale
H2
V2 = 1500 m/s
i23
F
H
V3 = 3000 m/s

e: thickness: h

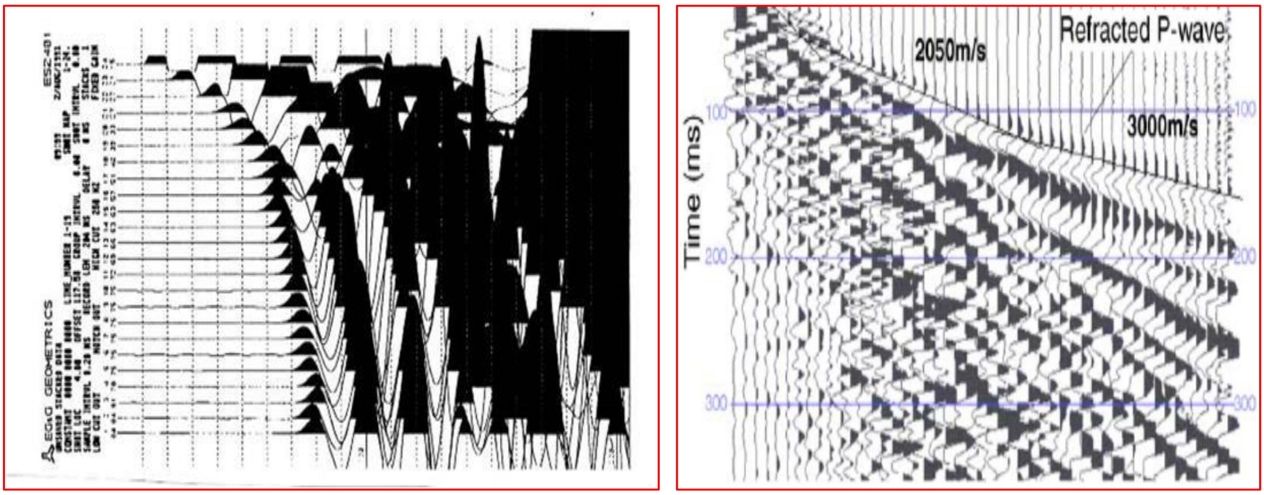
$$X_{c_2} = 2e_2\sqrt{\frac{V_3+V_2}{V_3-V_2}} - 2e_1\frac{V_3\sqrt{V_2^2-V_1^2}-V_2\sqrt{V_3^2-V_1^2}}{V_1(V_3-V_2)}$$

$$H_2 = \frac{Xc_2}{2}\sqrt{\frac{V_3-V_2}{V_3+V_2}} - H_1\frac{\sqrt{1-\left(\frac{V_1}{V_3}\right)^2}-\sqrt{1-\left(\frac{V_1}{V_2}\right)^2}}{\sqrt{\left(\frac{V_1}{V_2}\right)^2-\left(\frac{V_1}{V_3}\right)^2}}$$

$$H_2 = \frac{T_2}{2}\frac{V_2\,V_3}{\sqrt{V_3^2-V_2^2}} - \frac{H_1\cos i_{13}}{\sin i_{12}\cos i_{23}}$$

III.5.3.2 Recording (seismic film)

A seismic signal is recorded at a point on the ground by a sensor (seismograph or geophone). If a number of geophones are placed (generally in a straight line) and the seismic signals obtained from the same impact are recorded on the same document, a seismic film is obtained, from which it is easy to determine the travel times between the origin of the disturbance and the various sensors.



Multi-channel seismic recording