4. Seismic Methods

4.1 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 detect the ground motion caused by these returning waves and measure the arrival times of the waves at different ranges from the source.

Introduction

- These travel times will be converted to depth values.

→ subsurface geological interfaces can be systematically mapped

- Seismic methods represent a natural development of the already long-established methods of earthquake seismology.
- In earthquake seismology travel times of earthquake waves are recorded at seismological observatories. They provide information on the gross internal layering of the earth.

Introduction

- In the same way, but on a smaller scale seismic surveying provides a detailed picture of subsurface geology.
- Artificial sources, such as explosions, are used in the seismic method. Location, timing and source characteristics are, unlike earthquakes, under the direct control of the geophysicists.

Introduction

- Seismic methods represent the single most important geophysical method in terms of the amount of survey activity and the very wide range of its applications.
- The method is particularly well suited to the mapping of layered sedimentaries and is therefore widely used in the search for **oil and gas**.
- The method is also used for the mapping of near surface sedimentary layers, the location of the water table, engineering application, determination of the depth to bedrock.
- Seismic surveying can be carried out at land or at sea.

4.2 Seismic Waves

To understand the different types of seismic waves that can propagate through the ground away from a seismic source some elementary concepts of **stress** and **strain** need to be considered.

4.2.1 Stress and Strain

An external force F is applied across an area A of a surface of a body.

The ratio of force to area (F/A) is known as **stress**.

Stress can be resolved into two components:
a) one at right angles to the surface (normal or dilatational stress)
b) one in the plane of the surface (shear stress)

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

Stress and 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.

Stress and Strain

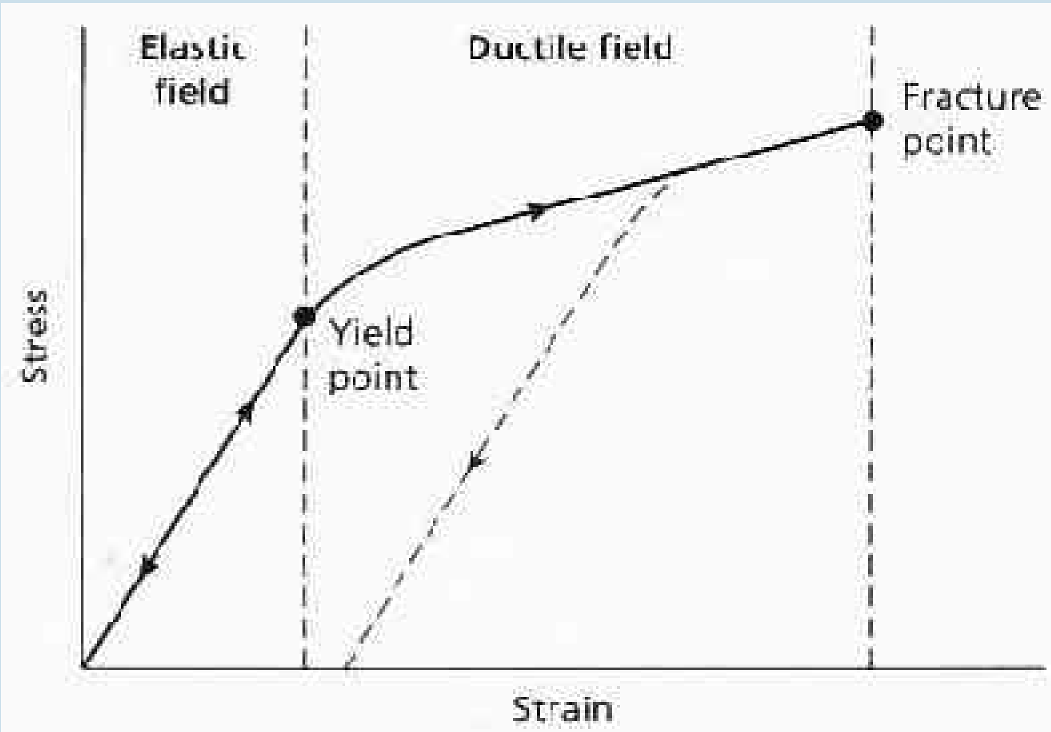


Fig. 1: A typical stress-strain curve for a solid body.[1]

→ Earthquake occurs when rocks are strained until fracture.

→ In exploration seismology the amounts of stress and strain lie well within the elastic behavior of natural materials.

Stress and Strain

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

a) Young modulus E

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

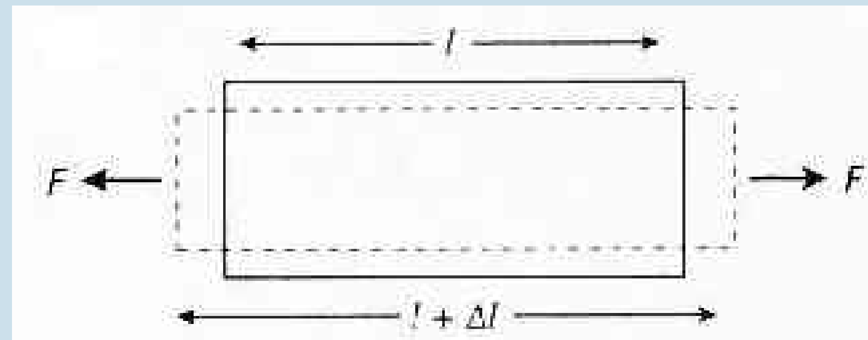


Fig. 2 (a)[1]

b) Bulk modulus K

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

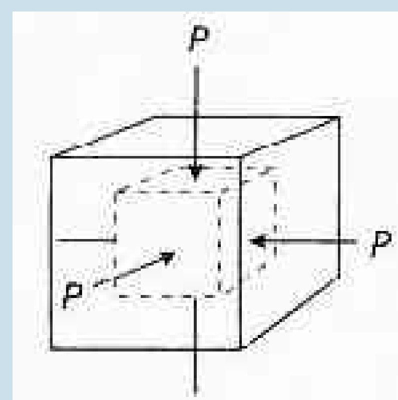


Fig. 2 (b)[1]

Stress and Strain

c) Shear modulus μ

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

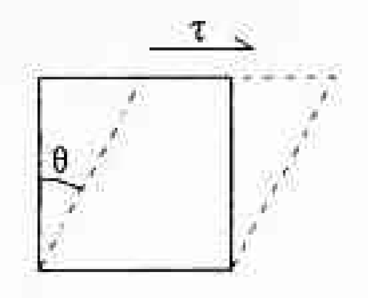


Fig. 2 (c)[1]

d) Axial modulus Ψ

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

(no lateral strain)

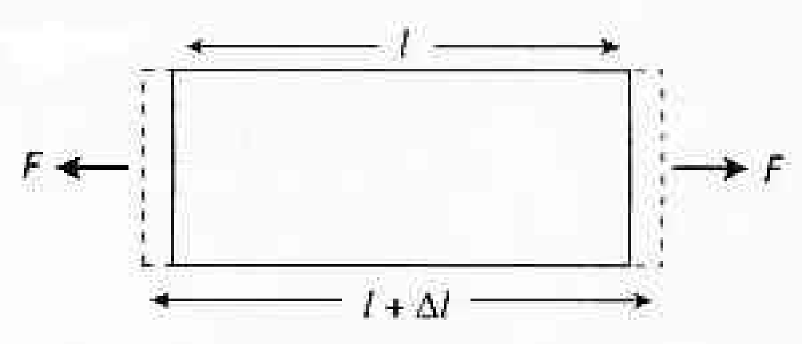


Fig. 2 (d)[1]

4.2.2 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.

4.2.2.1 Body 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.

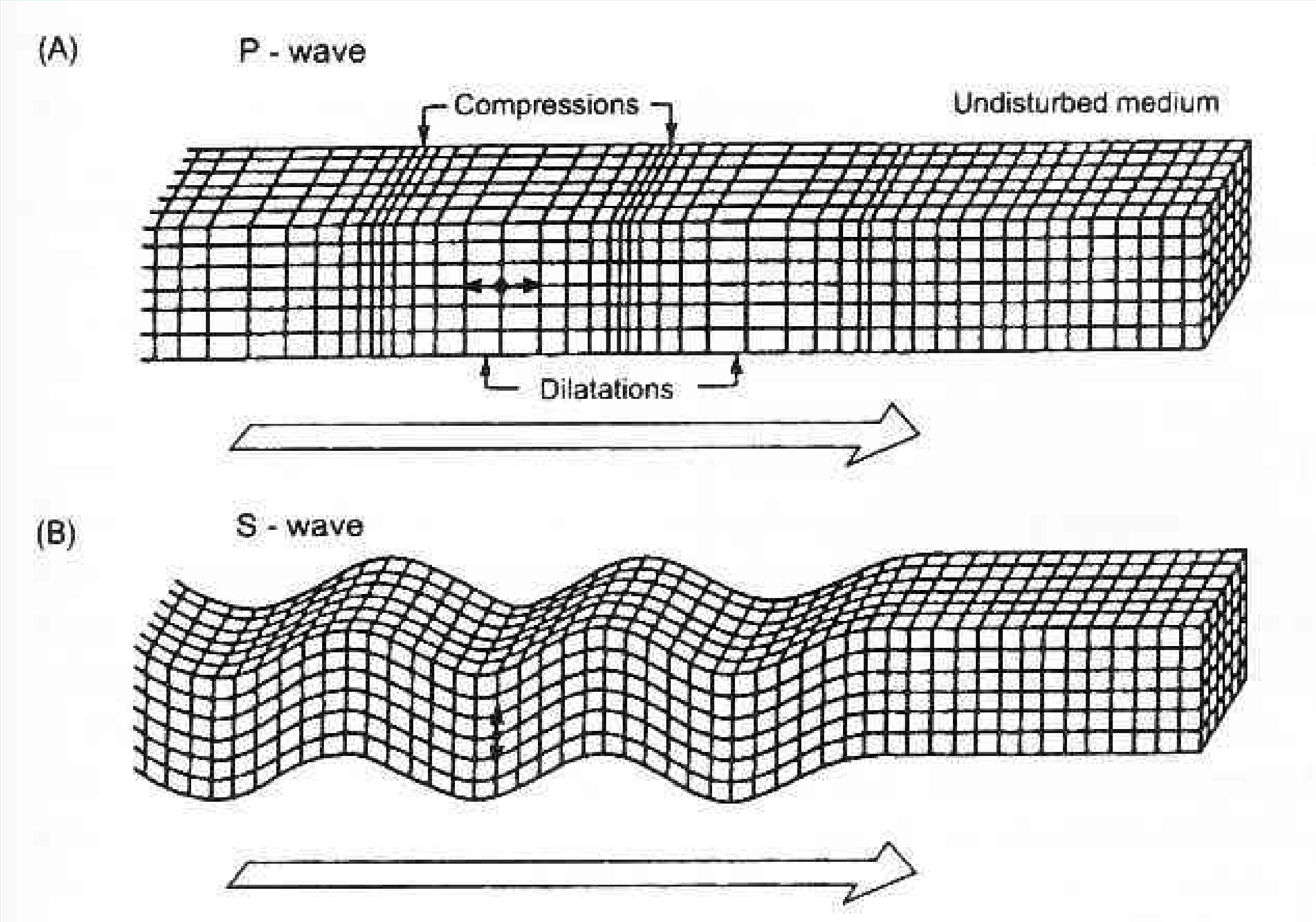


Fig. 3: Elastic deformations and ground particle motions associated with the passage of body waves. (A) A P-wave, and (B) an S-wave.[2]

4.2.2.2 Surface waves

Waves that do not penetrate into the subsurface media are known as surface waves.

- **Rayleigh wave**
  Travels 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 wave**
  Occurs 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.

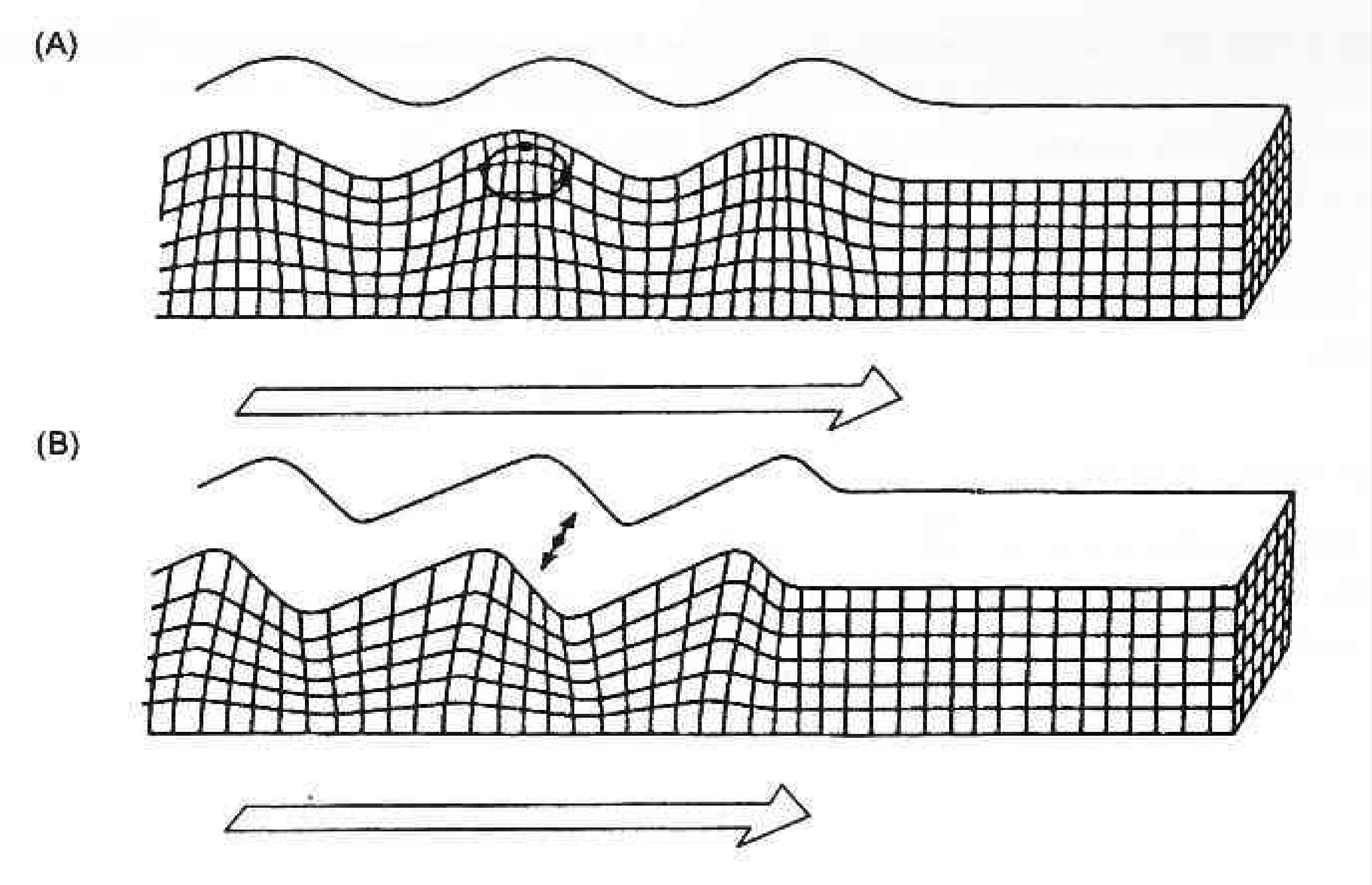


Fig. 4: Elastic deformations and ground particle motion associated with the passage of surface waves. (A) A Raleigh wave, and (B) a Love wave.[2]

4.3 Seismic wave velocities

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

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

4.3.1 Elastic wave velocity 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

Seismic wave velocities

	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

4.3.2 Time-average equation to estimate rock porosity

In porous rocks the nature of the materials within the pores strongly influences the elastic wave velocity.

- Water saturated rocks have different elastic wave velocities compared with gas saturated rocks.

Seismic velocities can be used to estimate porosity using the time average equation.

$$1/v = \phi/v_f + (1-\phi)/v_m$$

where

v	p-wave velocity for a rock
ϕ	fractional porosity
v_f	acoustic velocity in the pore fluid
v_m	acoustic velocity in the rock

4.4 Ray paths in layered media

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.

4.4.1 Reflection and transmission of normally incident seismic rays.

The total energy of the transmitted and reflected rays must be equal to the energy of the incident ray.

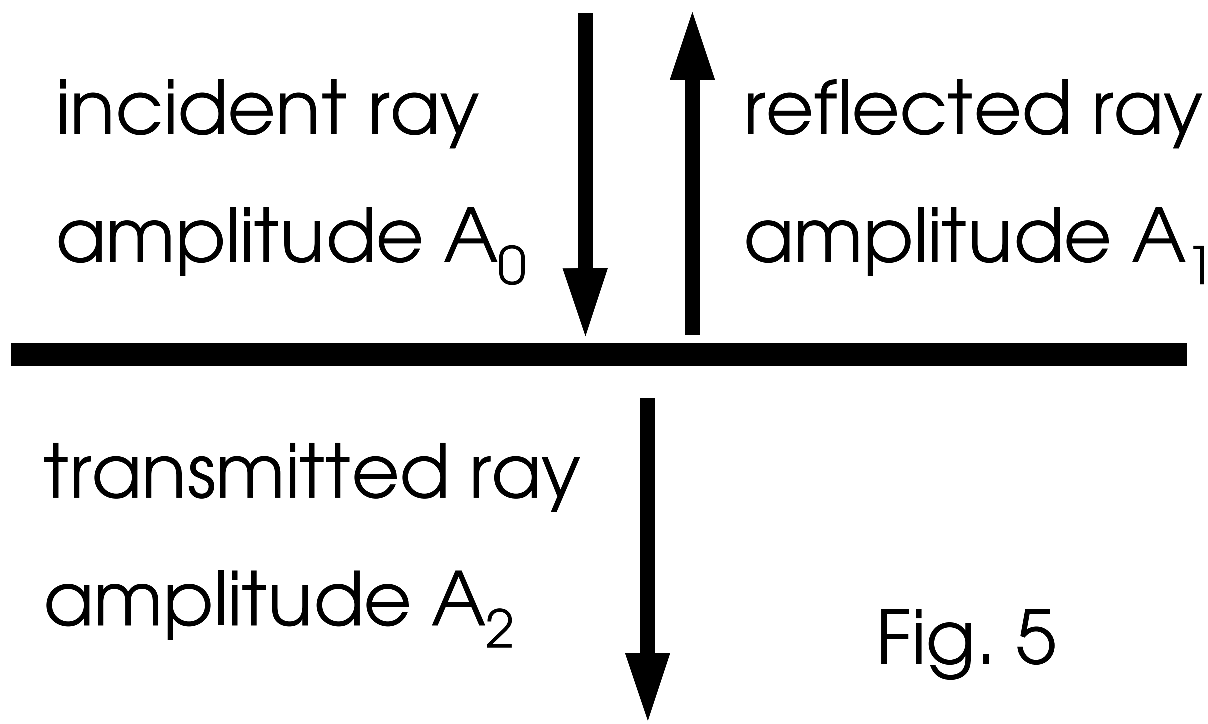


Fig. 5

$Z = \rho v$ = acoustic impedance

The smaller the contrast in acoustic impedance across the rock interface the greater is the portion of energy transmitted through the interface.

Reflection coefficients

The more energy is reflected the greater the contrast.

$$R = A_1 / A_0$$ reflection coefficient

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

A negative value of R signifies a phase change of 180° in the reflected ray.

$$T = A_2 / A_0$$ transmission coefficient

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

4.4.2 Reflection and refraction of obliquely incident rays

In the case of incident waves reflected and transmitted waves are generated as described in the case of normal incidence.

When a P-wave is incident at an oblique angle on a plane surface, four types of waves are generated:

- reflected and transmitted P-waves
- reflected and transmitted S-waves

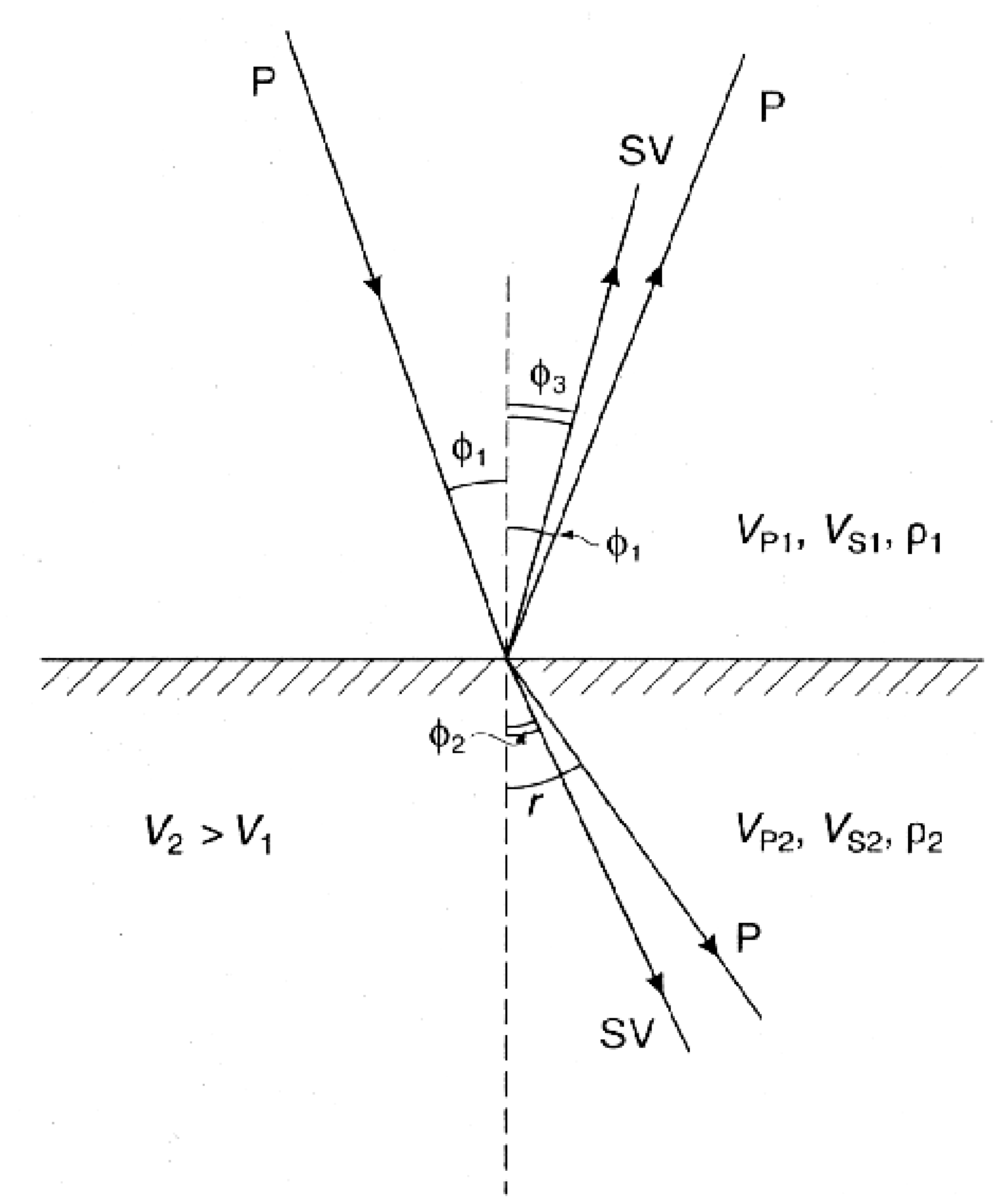


Figure 6: Geometry of rays associated with a P-wave (shown as ray) incident obliquely on a plane interface, and converted vertically polarized S-waves (SV; shown as ray). v_p and v_s are the respective P- and S-wave velocities. Suffixes 1 and 2 depict the layer number.[2]

Snell's law

The geometry of the various reflected and refracted waves relative to the incident waves is directly analogous to light and can be described using Snell's law of refraction.

$$\frac{\sin i}{v_{p_1}} = \frac{\sin r}{v_{p_2}} = \frac{\sin p_1}{v_{s_1}} = \frac{\sin p_2}{v_{s_2}} = p$$

i and r are the angles of incidence and refraction.

p is the ray parameter.

v_1 and v_2 are the speeds of propagation for P- and S-waves.

$$\frac{\sin i}{\sin r} = \frac{v_1}{v_2}$$

Critical refraction

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

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

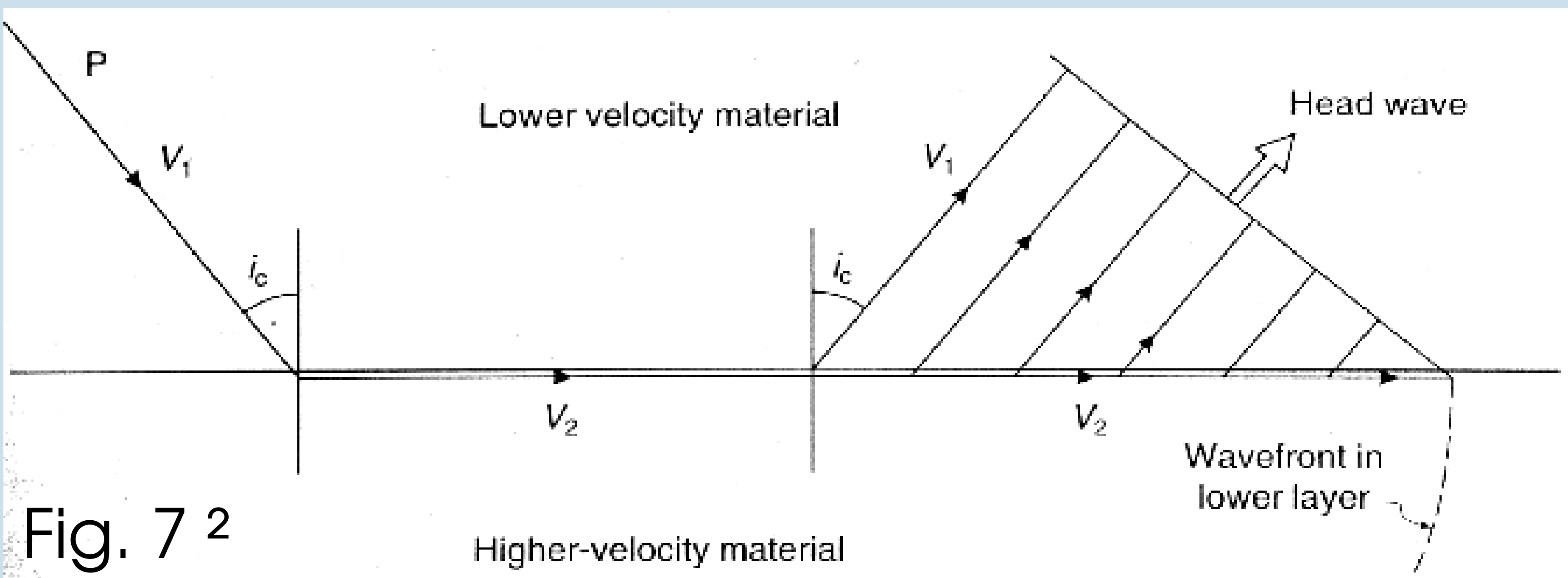


Fig. 7 [2]

4.5 Seismic data acquisition systems

The fundamental purpose of seismic surveys is to accurately record ground motion caused by known sources in a known location.

The record of ground motion with time constitutes a seismogram.

The essential instrumental requirements are to

- generate a seismic pulse with a suitable source
- detect the seismic waves in the ground with a suitable transducer.
- record and display seismic wave forms on a suitable seismograph.

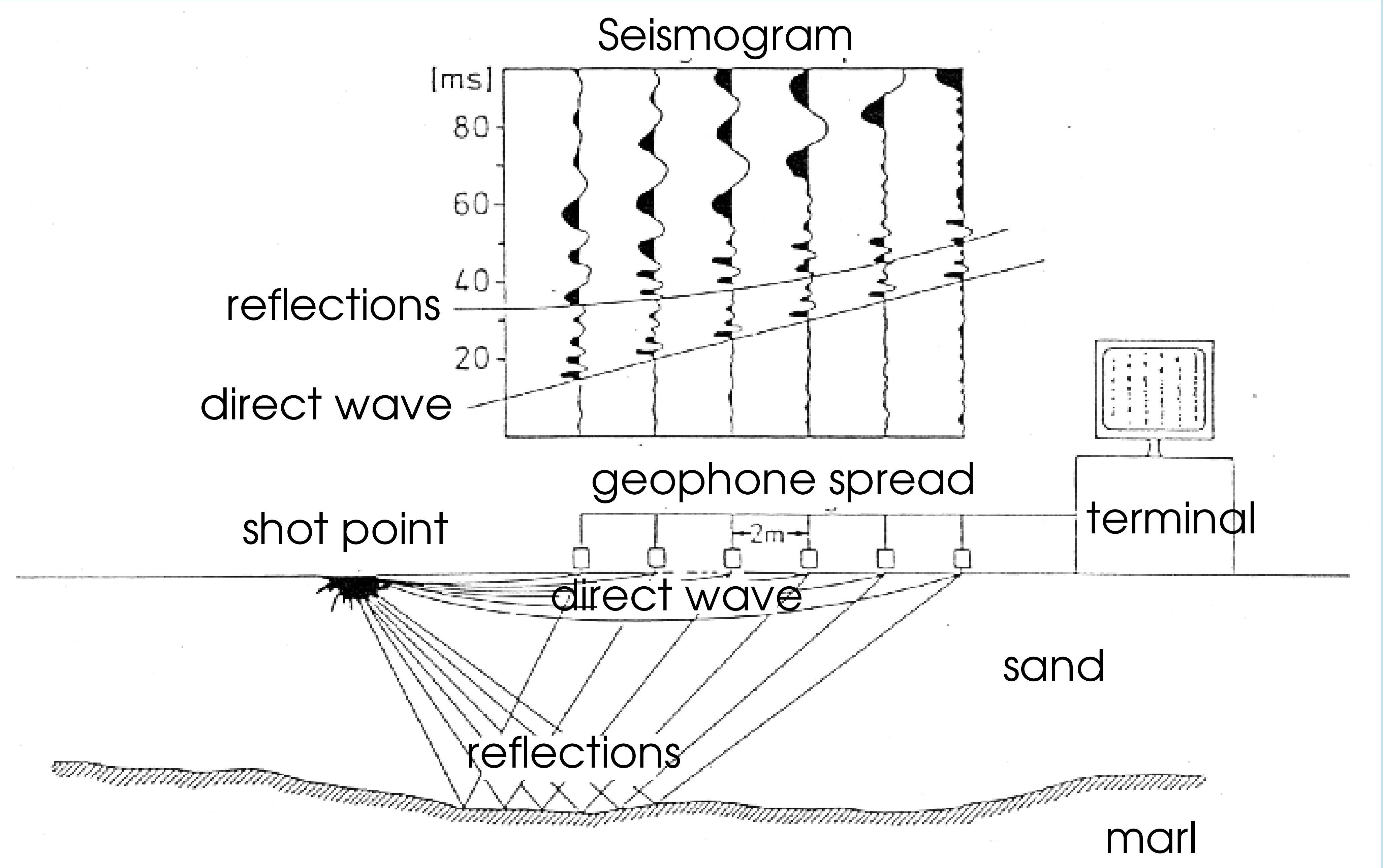


Figure 8: Seismic field work: Source, geophone spread and terminal. [3]

Seismic Sources

Geophysics
Cologne

Seismic sources on land

Impact
Sledge hammer
Drop-weight
Accelerated weight

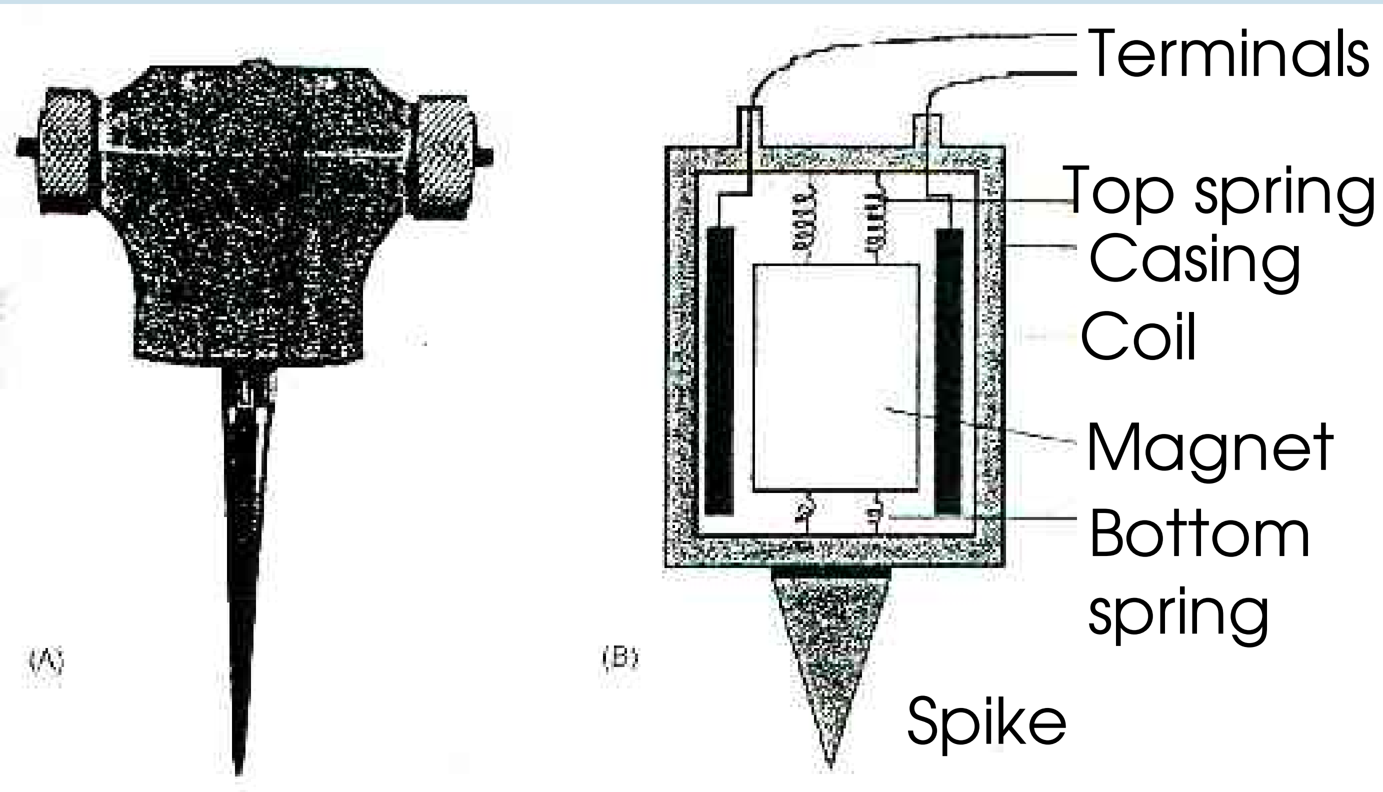
Impulse
Dynamite
Air gun

Vibrator
Vibroseis

Transducers

Conversion of the ground motion to an electrical signal requires a transducer which is sensitive to ground motion.

Devices used on land to detect seismic ground motions are known as seismometers or geophones.



The relative movement of the magnet with respect to the coil results in a small voltage.

Fig. 9: Field geophone [3]

4.7 Seismic reflection surveying

4.7.1 Introduction and general considerations

Seismic reflection is the most widely used geophysical technique. Important details about the geometry of the structure and about the physical properties can be derived.

Its predominant applications are hydrocarbon exploration and research into crustal structure with depths of penetration of several kilometers.

Introduction and general considerations

Since 1980 the method has been applied increasingly to engineering and environmental investigations (< 200 m)

- mapping quaternary deposits, buried valleys, shallow faults
- hydrogeological studies of aquifers
- shallow coal exploration
- pre-construction ground investigations for pipe, cables and sewerage schemes

Introduction and general considerations

The essence of the seismic reflection technique is <u>to measure the time</u> taken for a seismic wave to travel from a source down into the ground where it is reflected back to the surface and then detected at a receiver.

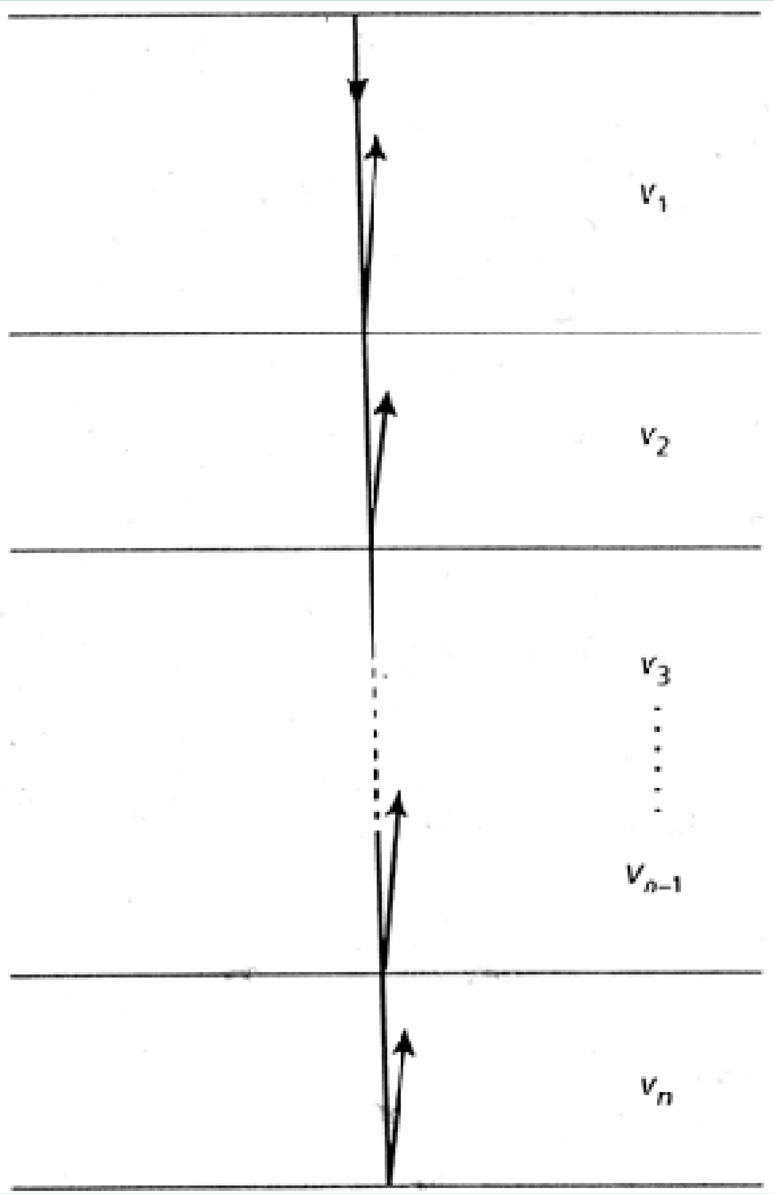
The time is known as the two way travel time (TWTT).

The most important problem in seismic reflection surveying is the translation of <u>TWTT</u> to <u>depth</u>.

While travel times are measured, the one parameter, that most affects the conversation of to depth is seismic velocity.

→ two unknowns (depth + velocity)

4.7.2 Geometry of reflected ray paths



v_i = interval velocity
z_i = thickness of the interval
τ_i = one way travel time

$$v_i = \frac{z_i}{\tau_i}$$

Fig. 10: Vertical reflected ray paths in a horizontally-layered ground. [1]

Average velocity and total one way travel time

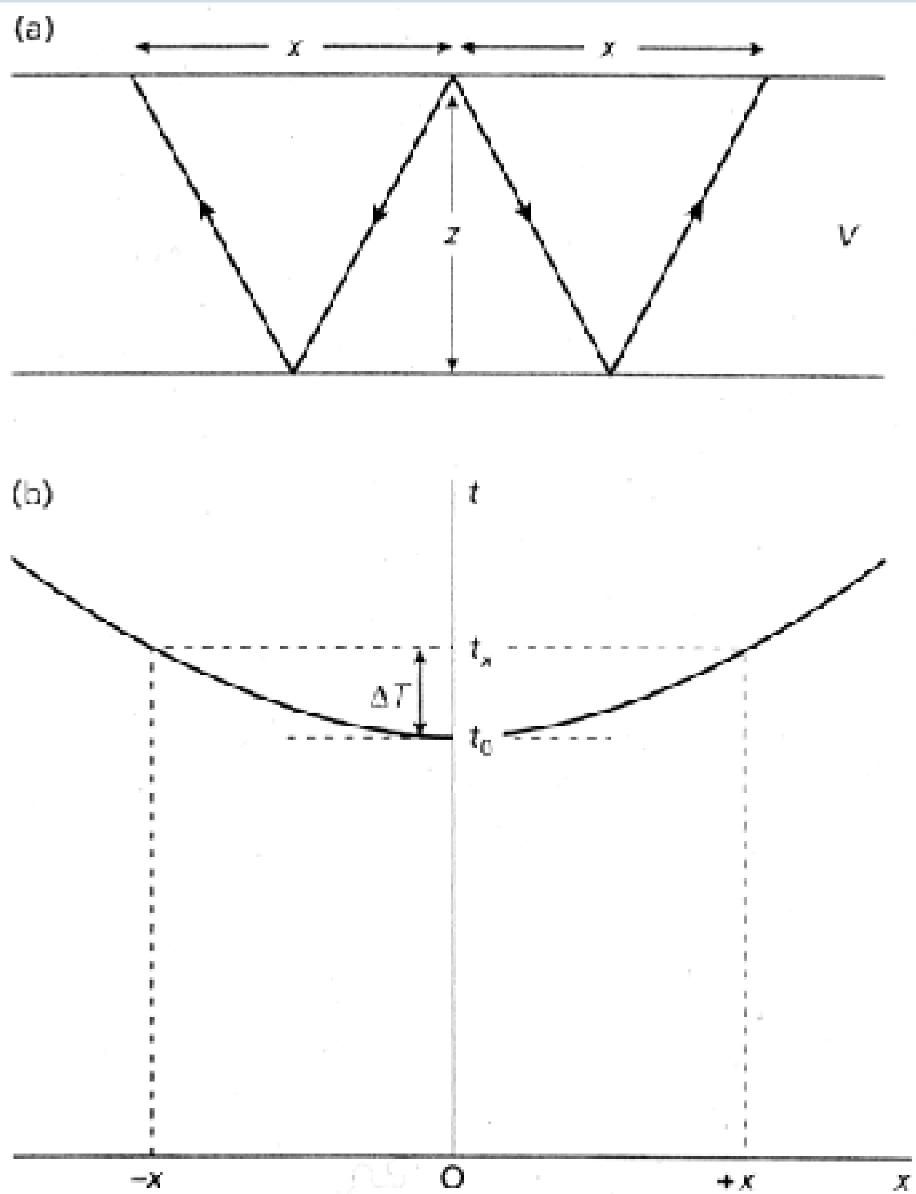
The interval velocity may be averaged over several depth intervals to yield an average velocity $\bar{v}$.

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

Z_n = total thickness of the top n layers

T_n = total one way travel time through the n layers

4.7.2.1 Single horizontal reflector



The equation for the travel time t of the reflected ray from a shot point to a detector at a horizontal offset x is given by:

$$t = (x^2 + 4z^2)^{1/2} / v$$

t is measured at offset distance x.

→ still two unknown values (z + v)

Fig. 11: (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). [1]

Intercept time

However, if many reflection times t are measured at different offsets x, there will be enough information to derive z and v. (Figure 11 b)

Substitute x = 0 in the equation:

$$t_0 = 2\,z/v$$

This is the travel time of a vertically reflected ray (intercept on the time axis of the time distance curve, compare Figure 11 b).

Rewrite the equation

$$t^2 = 4\,z^2/v^2 + x^2/v^2 = t_0^2 + x^2/v^2$$

→ This is the simplest way of determining the velocity.

Velocity determination

➔ 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 <u>depth</u> 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 11)

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

4.7.3 Data acquisition

The initial display of seismic profile data is normally in groups of seismic traces recorded from a common shot, known as <u>shot gathers</u>.

The geophones may be distributed on either side of the shot, or only on one side.

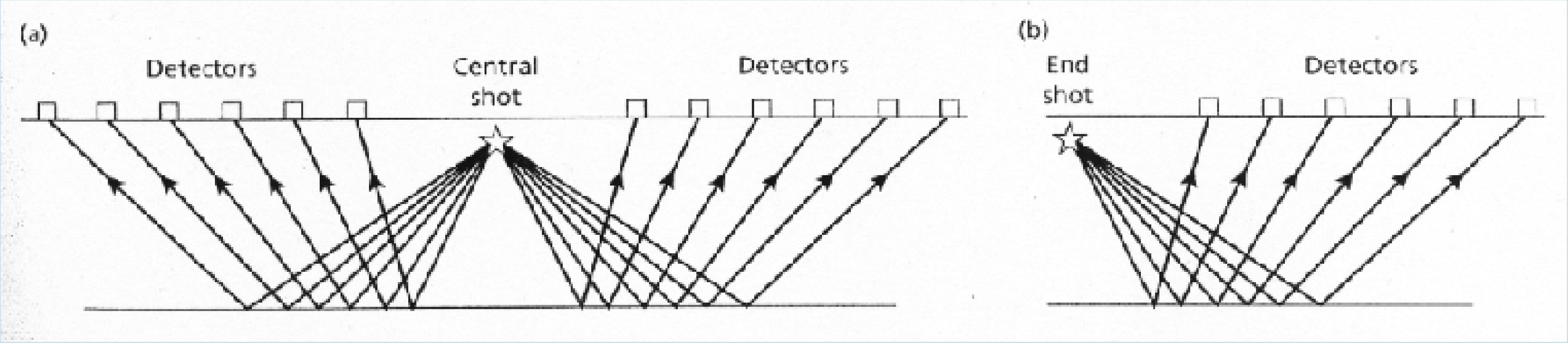


Fig. 12: Shot-detector configurations used in multichannel seismic reflection profiling. (a) Split spread, or straddle spread. (b) Single-ended or on-end spread. [1]

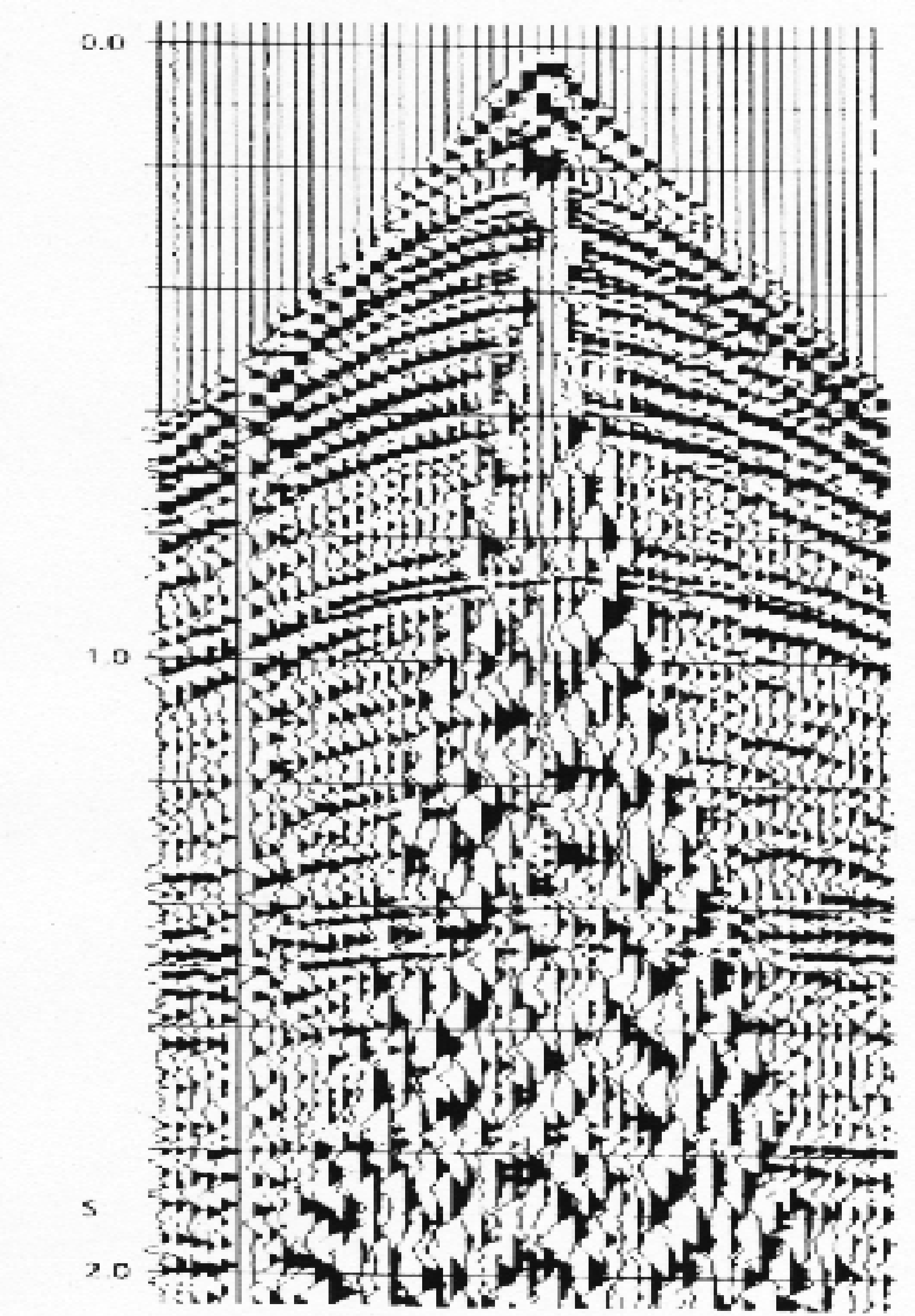


Fig. 13: A draped seismic record of a shot gather from a split spread. Sets of reflected arrivals from individual interfaces are recognizable by the characteristic hyperbolic alignment of seismic pulses. The late-arriving high-amplitude, low-frequency events, defining a triangular-shaped central zone within which reflected arrivals are masked, represent surface waves (ground roll). These latter waves are a typical type of coherent noise. [1]

Multiple shotpoints and common mid-point

If more than one shot location is used, reflections arising from the same point on the interface will be detected at different geophones. The common point of reflection is known as the common mid-point (CMP).

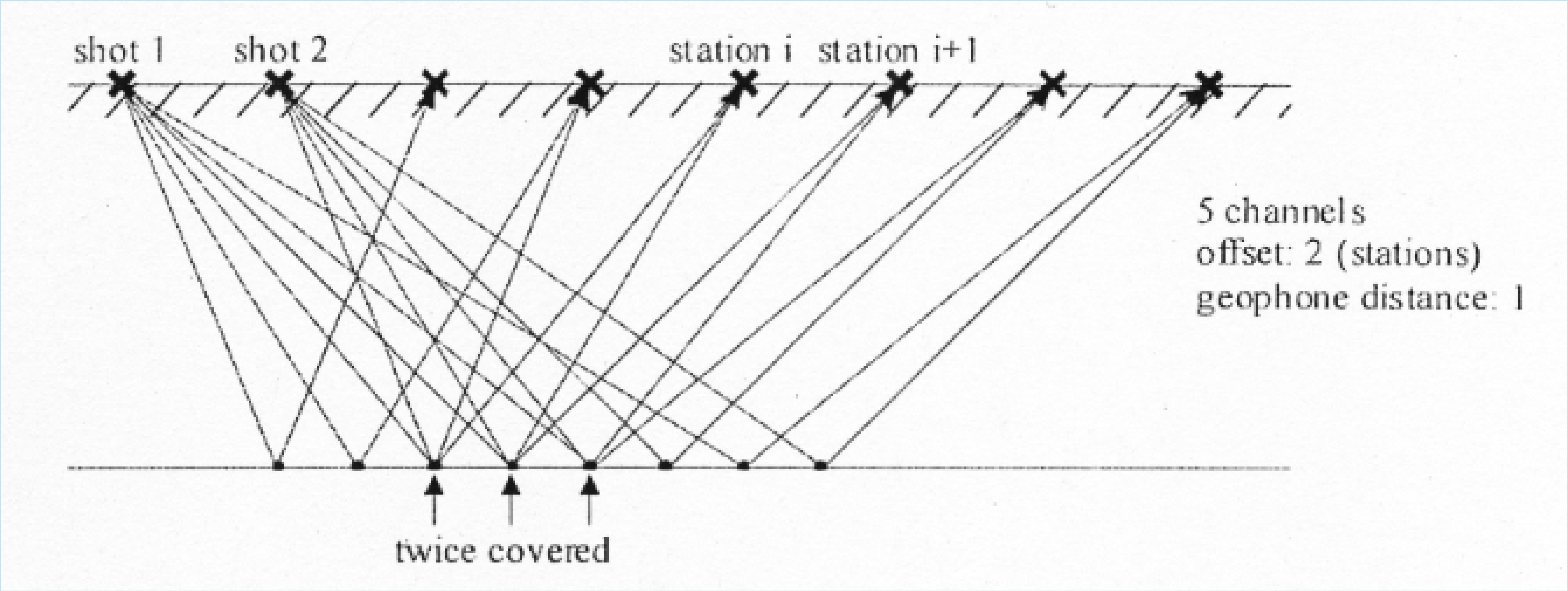


Fig. 14: Data acquisition for reflection seismic. [4]

Common mid-point

The number of times the same point on a reflector is sampled as the fold of coverage.

For example: 12 different shot-geophone locations
→ 12-fold coverage

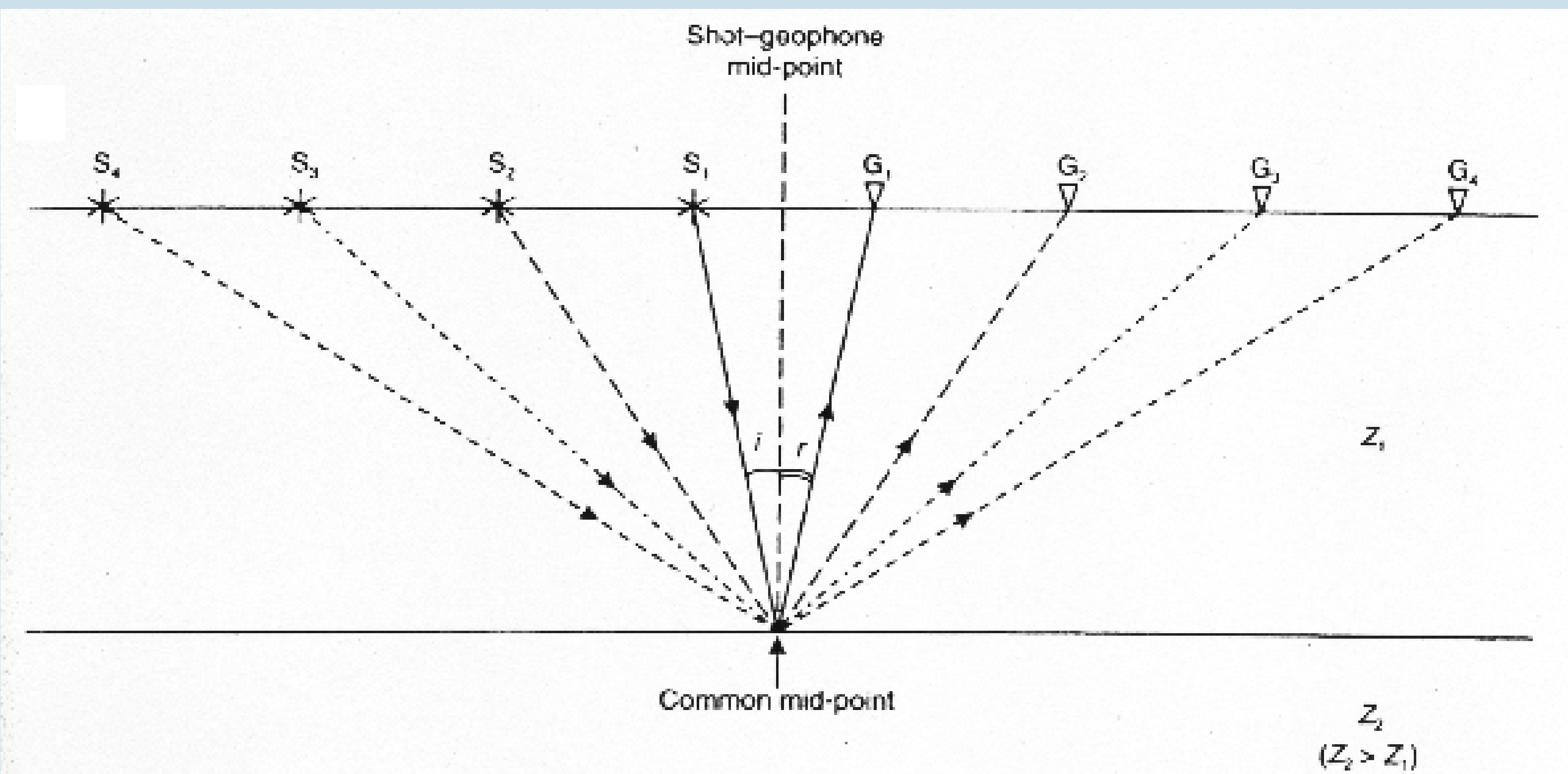


Fig. 15: Principle of the common mid-point over a horizontal interface. [2]

Common mid-point gather

The CMP gather lies in the heart of seismic processing for two main 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.

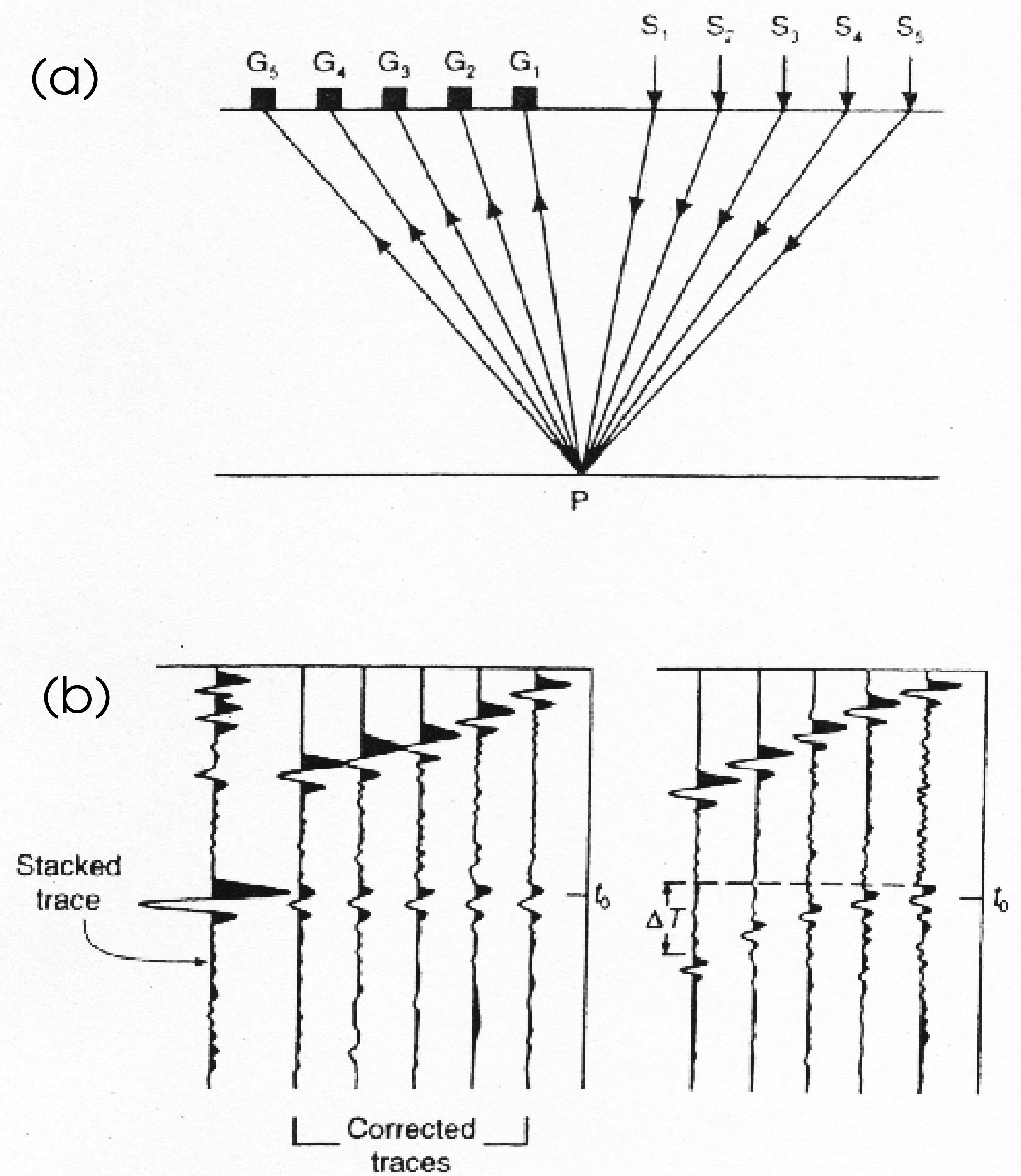


Fig. 16: Given the source-receiver layout and corresponding ray-paths for a common depth point spread, shown in (a), the resulting seismic traces are illustrated in (b), uncorrected (on the right), (corrected on the left) – note how the reflection events are aligned – and the final stacked trace. [2]

4.7.4 Case studies

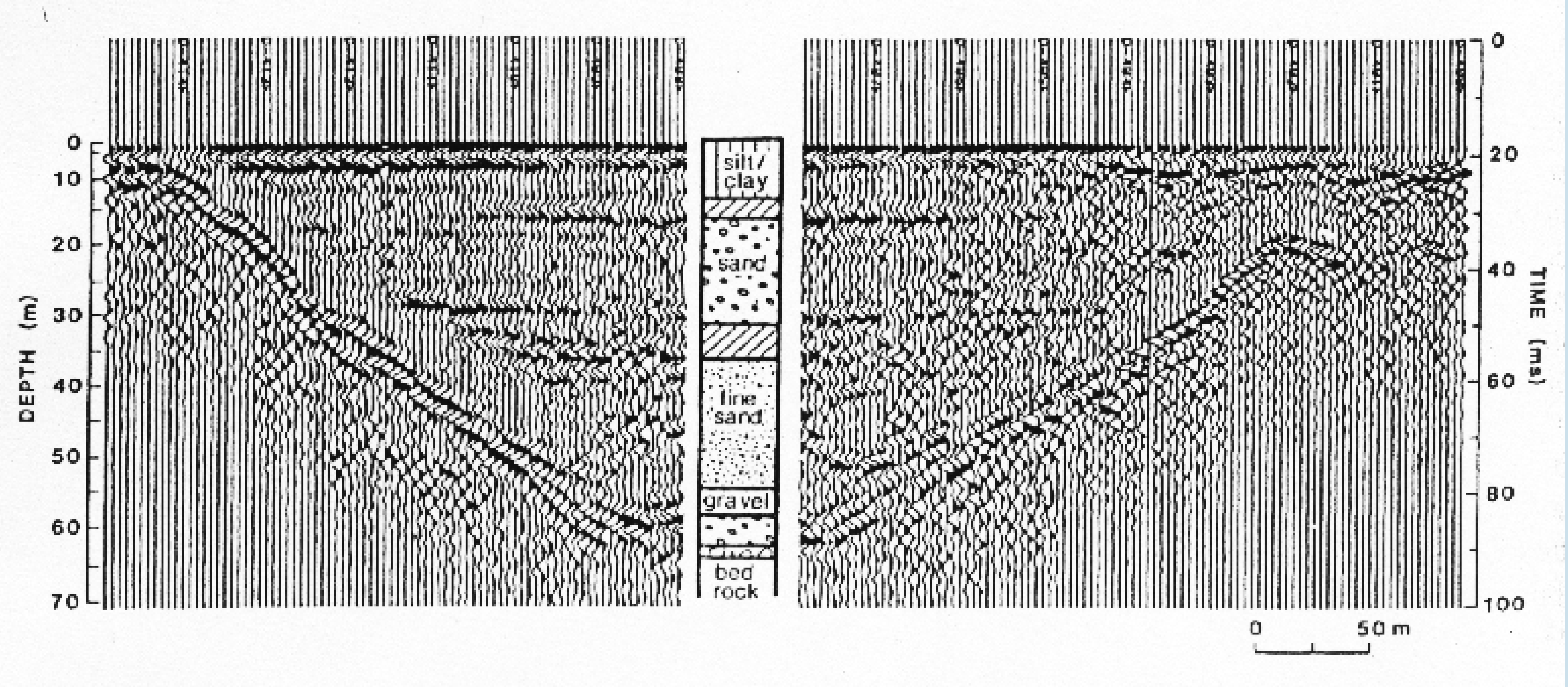


Fig. 17: Unmigrated optimum offset reflection section from Dryden, Ontario, showing a steep sided bed-rock valley filled with clay and sand deposits; the section is around 500m long. [2]

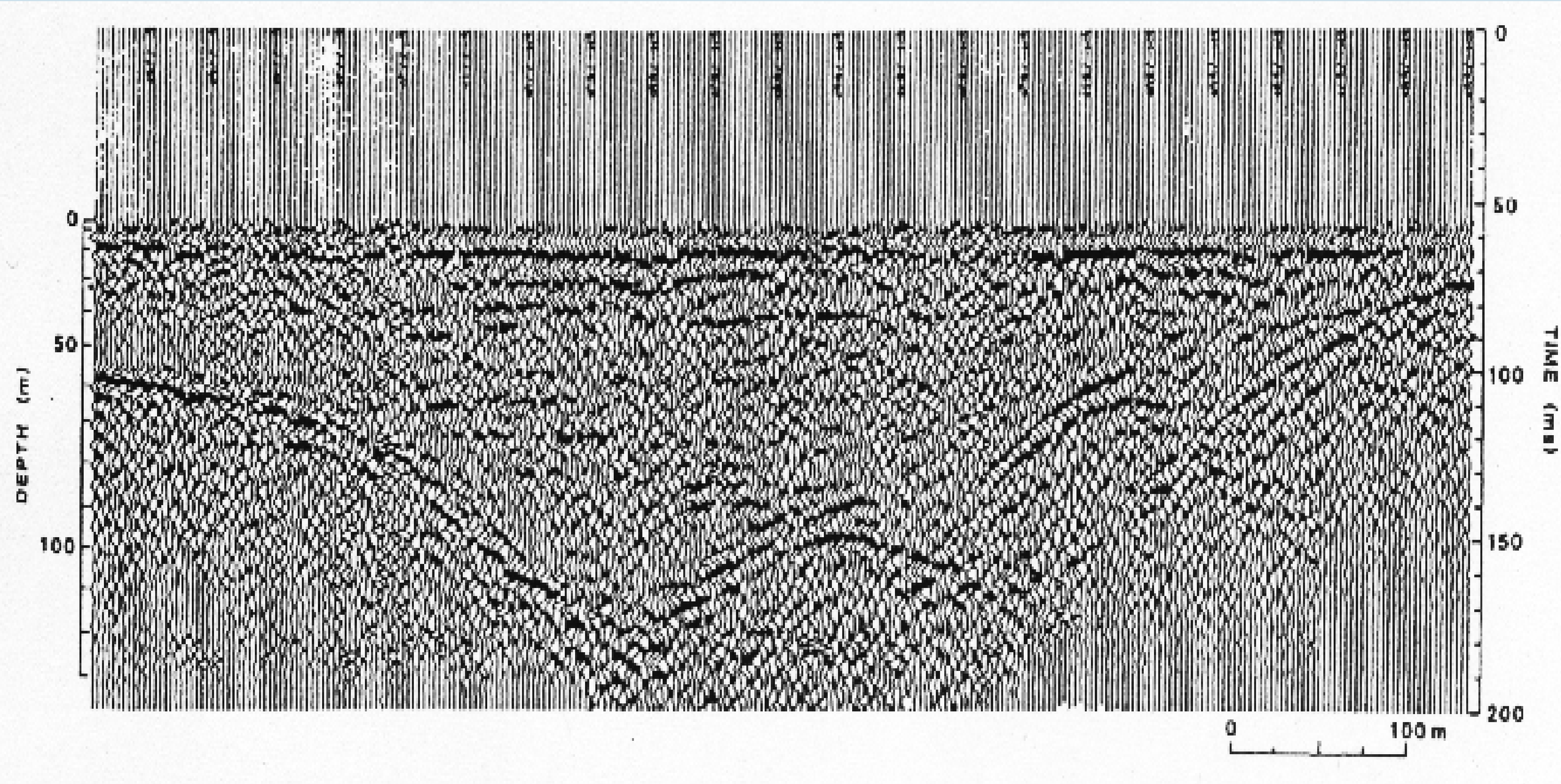


Fig. 18: Optimum offset reflection section from Shawville, Quebec, showing a broad (ca. 500m wide) buried river channel, part of the proto Qtawa River.[2]

4.8 Seismic refraction surveying

The seismic refraction surveying method uses seismic energy that returns to the surface after traveling through the ground along refracted ray paths.

The most commonly derived geophysical parameter is the seismic velocity of the layers present.

A number of geotechnical parameters can be derived from seismic velocity.

In addition to the more conventional engineering applications of foundation studies for dams and major buildings, seismic refraction is increasingly being used in hydrogeological investigations to determine saturated aquifer thickness, weathered fault zones.

4.8.1 General Principles

The refraction method is dependent upon there being an <u>increase in velocity with depth</u>.

The direction of travel of a seismic wave changes on entry into a new medium.

The amount of change of direction is governed by the contrast in seismic velocity across the boundary according to <u>Snell's law</u>.

Snell's law

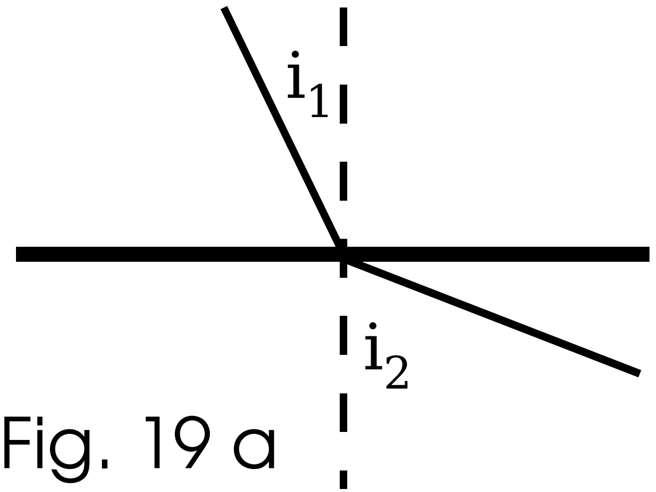


Fig. 19 a

$$\frac{\sin i_2}{\sin i_1} = \frac{v_2}{v_1}$$

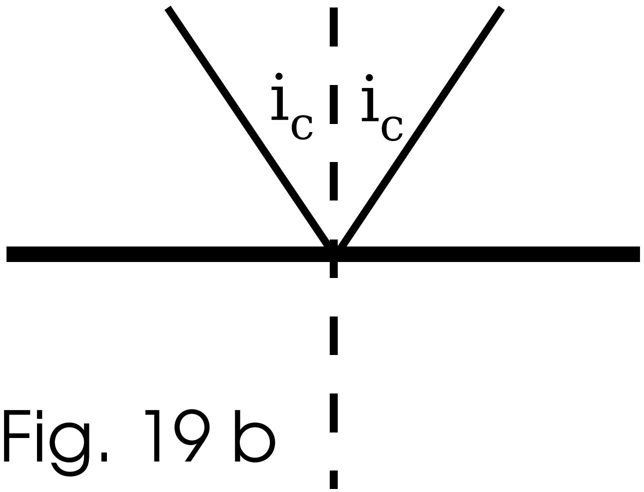


Fig. 19 b

If $i_2 = 90°$ ($\sin 90° = 1$)

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

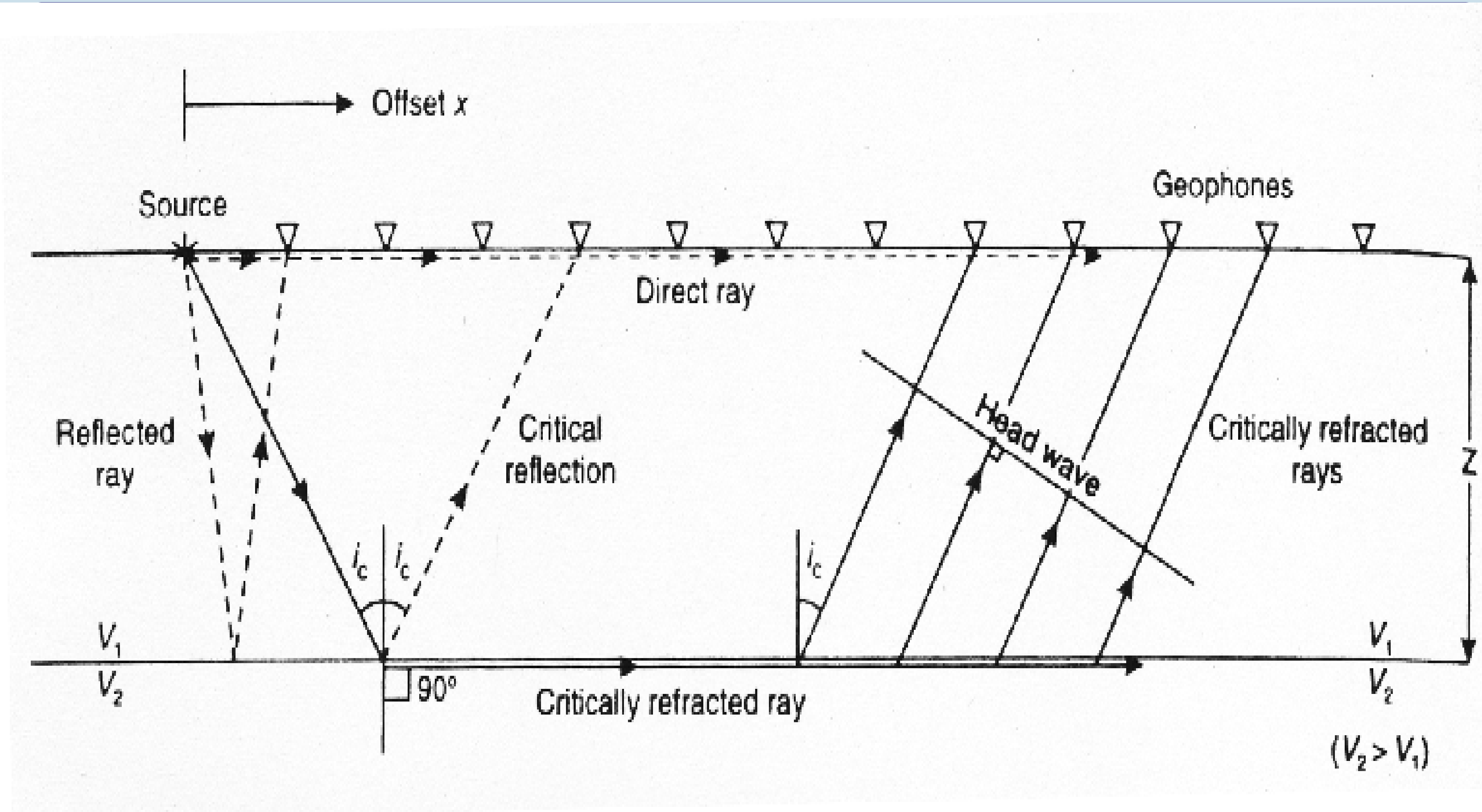


Fig. 20: Raypath diagram showing the respective paths for direct, reflected and refracted rays. [2]

4.8.2 Geometry of the refracted ray paths

Two layer case

Travel way of the refracted wave

$$L_R = L_1 + L_2 + L_3$$

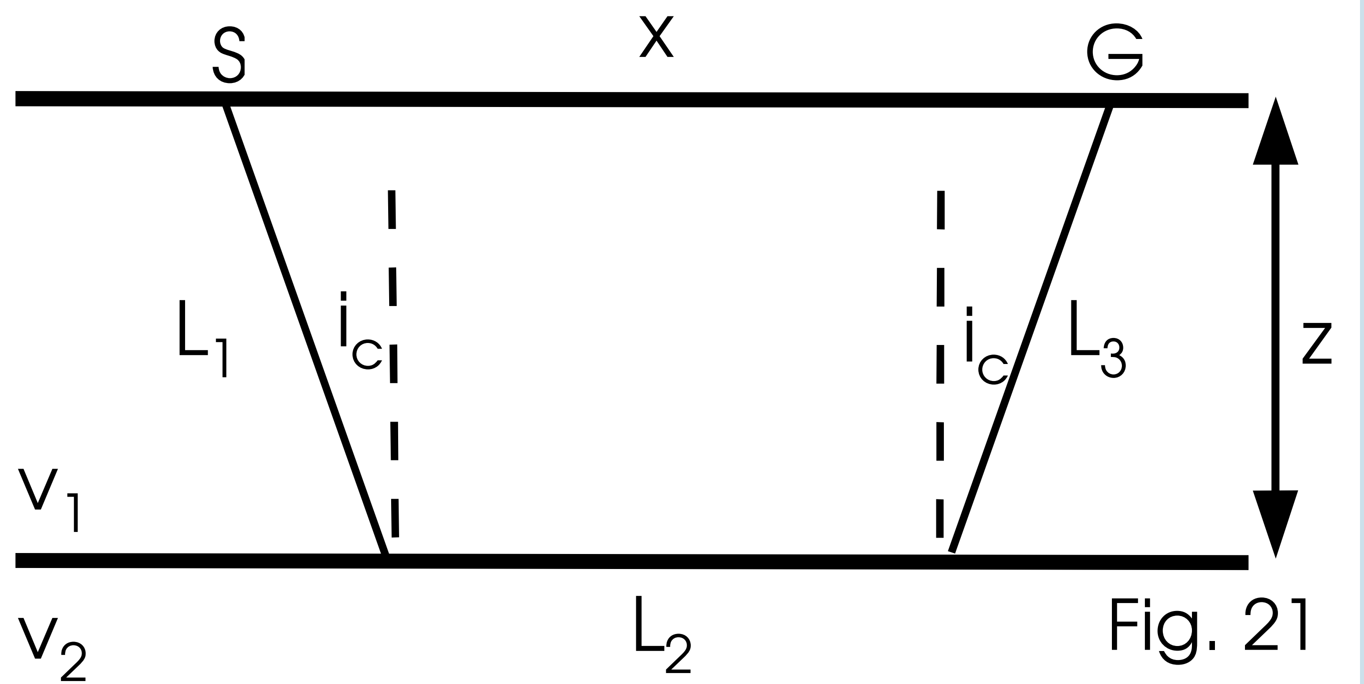


Fig. 21

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 times

Travel time of the direct wave

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

Travel time of the reflected wave (see 4.7.2.1)

$$\boxed{T_i = \frac{2\sqrt{z^2 + (x/2)^2}}{v_1}}$$

<u>Example:</u> $v_1 = 500$ m/s

$v_2 = 1000$ m/s

$z = 3$m

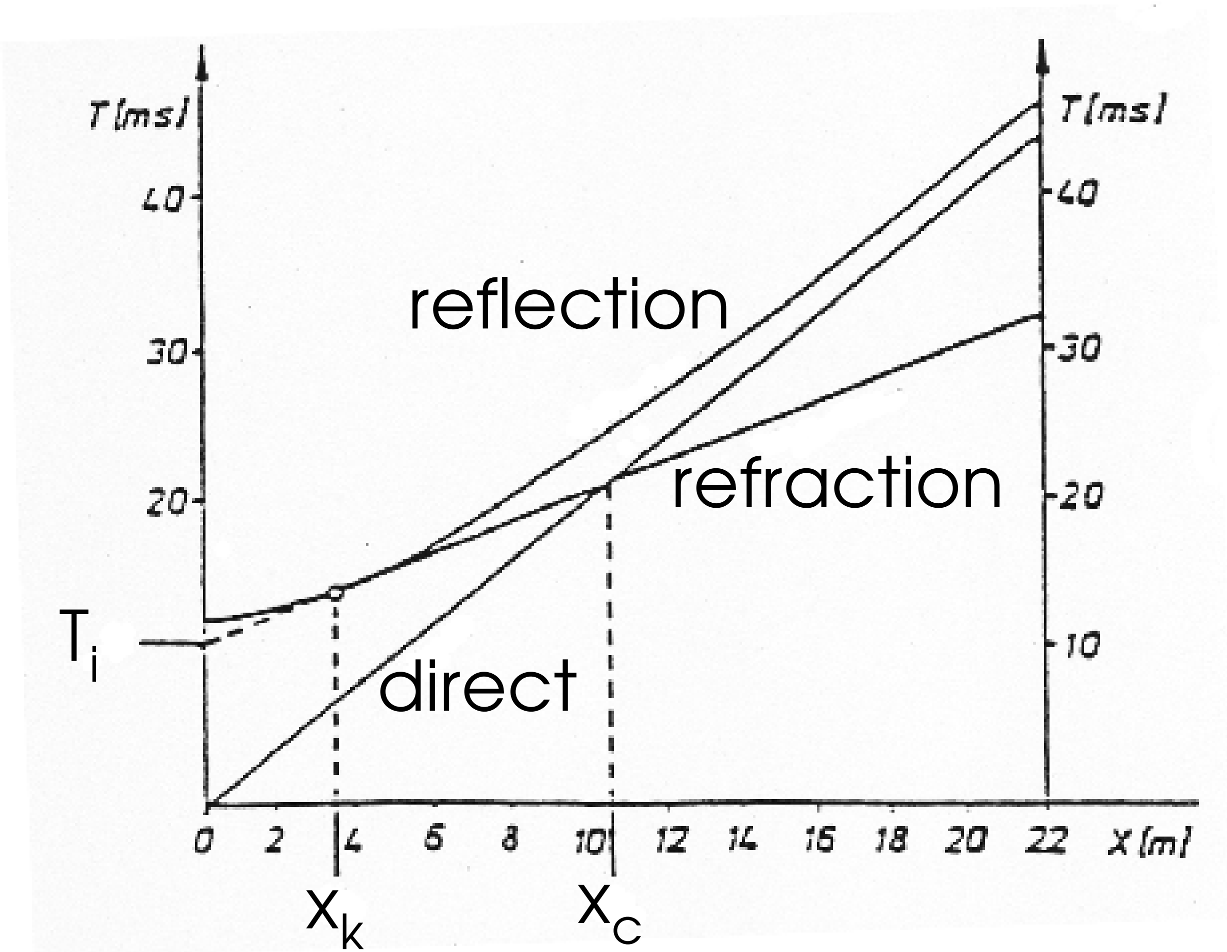


Fig. 22: Traveltime curves for reflected, refracted and direct waves.

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.

The crossover distance x_c is the offset at which the critically refracted waves precede the direct waves.

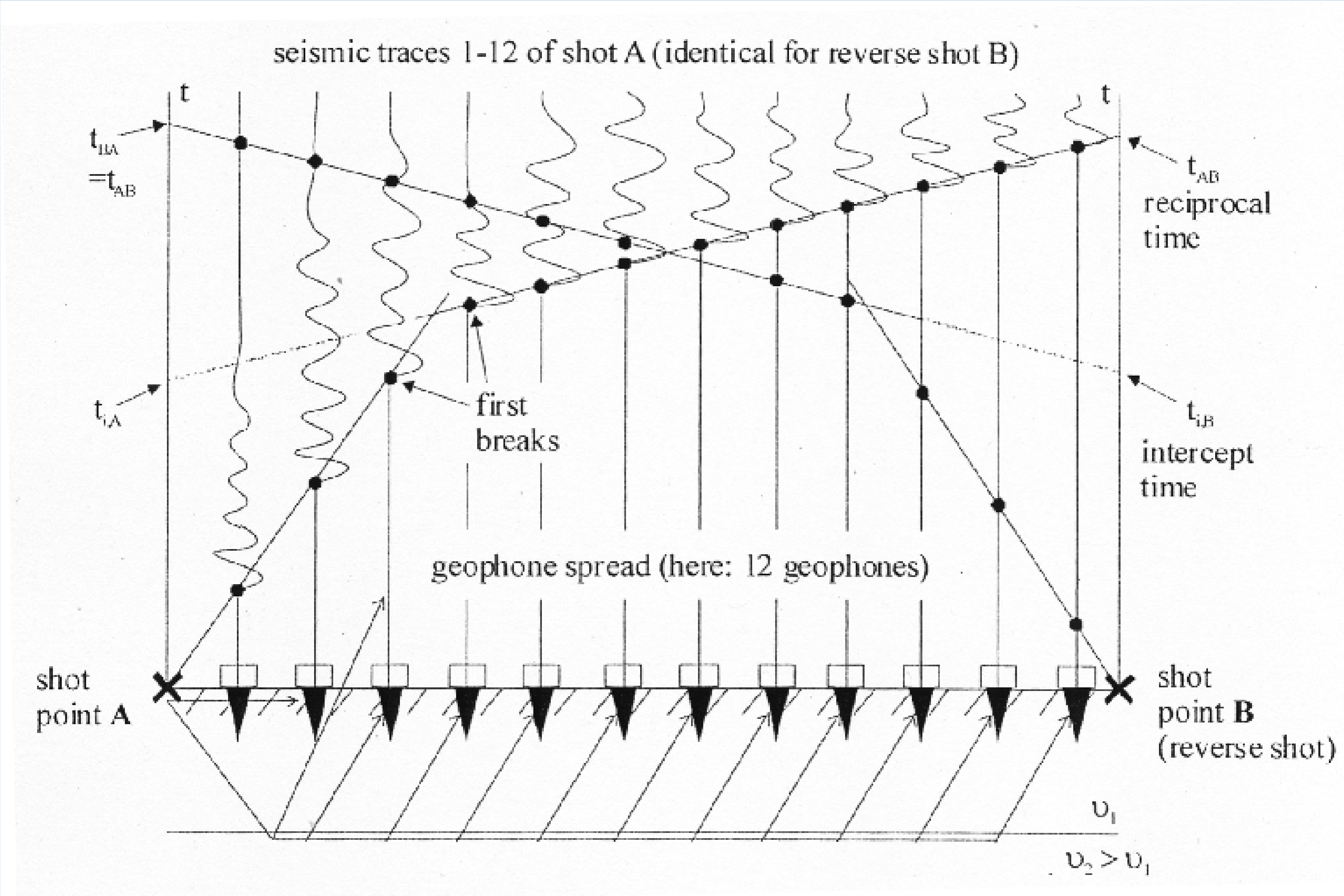


Fig. 23: Principle of refraction seismic. [4]

Calculation of the depth of the refractor

Two methods

A) Using the intercept times

$$T_i = \frac{2z}{v_1}\frac{\sqrt{v_2^2 - v_1^2}}{v_2}$$

(travel time of the refracted ray with x = 0)

$$\boxed{z = \frac{v_1 T_i}{2}\frac{v_2}{\sqrt{v_2^2 - v_1^2}}}$$

B) Using the crossover distance x_c

$$x_c = 2z\sqrt{\frac{v_2 + v_1}{v_2 - v_1}} \qquad \Rightarrow z = \frac{x_c}{2}\sqrt{\frac{v_2 - v_1}{v_2 + v_1}}$$

Multilayer case

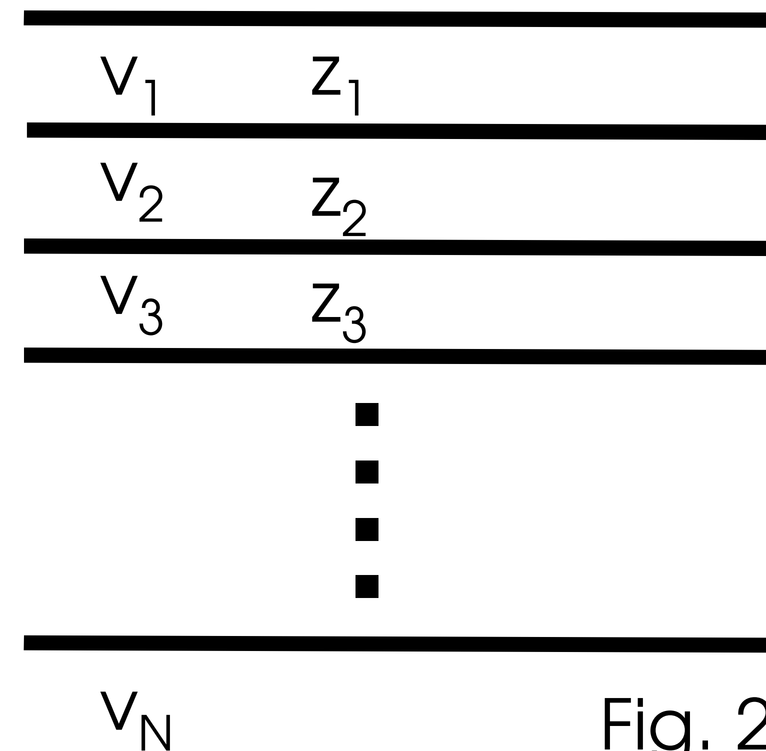


Fig. 24

N layers

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

Travel time

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

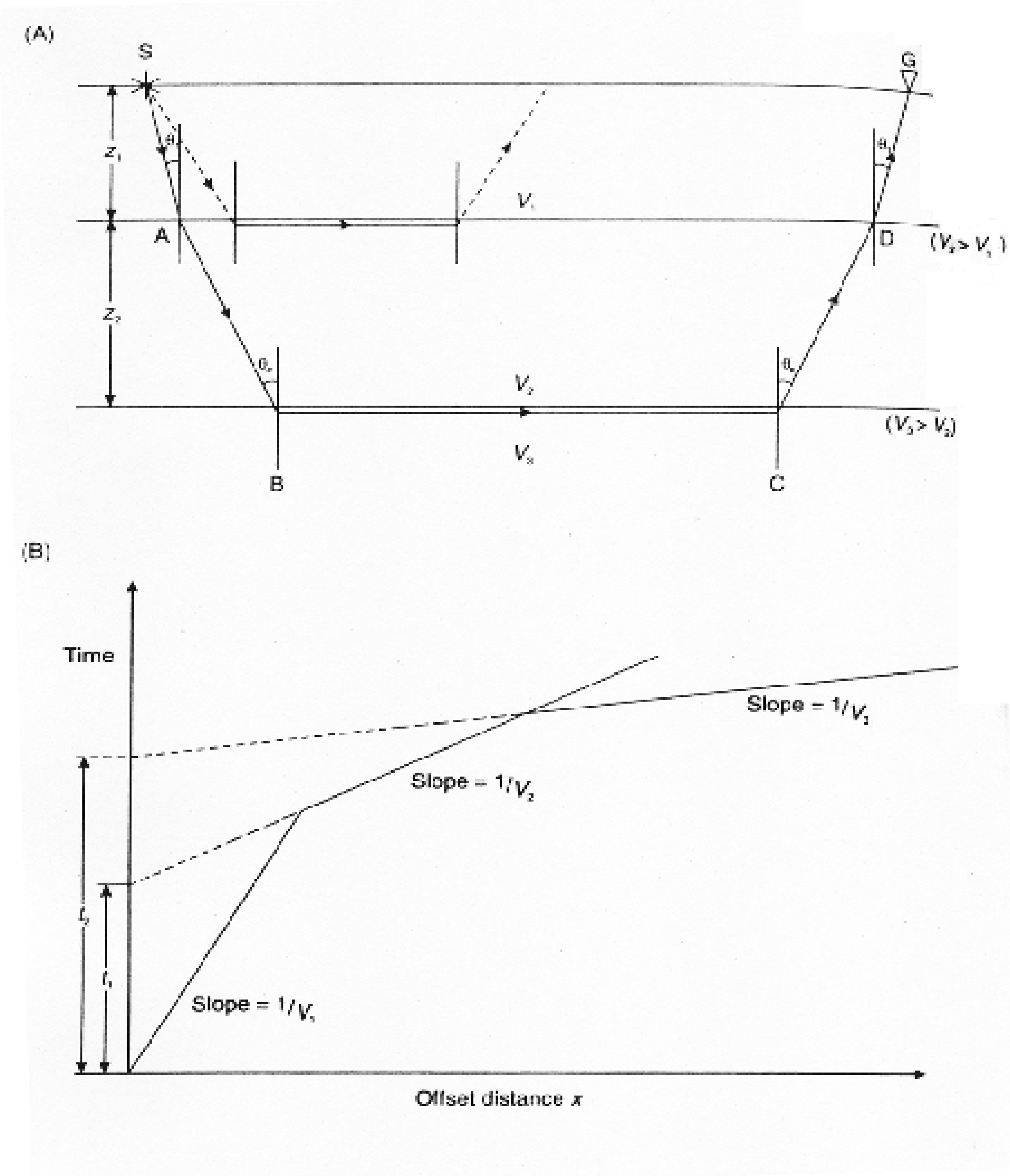


Fig. 25: (A) Simple raypaths diagram for refracted rays, and (B) their respective travel time-distance graphs for a three-layer case with horizontal planar interfaces. [2]

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

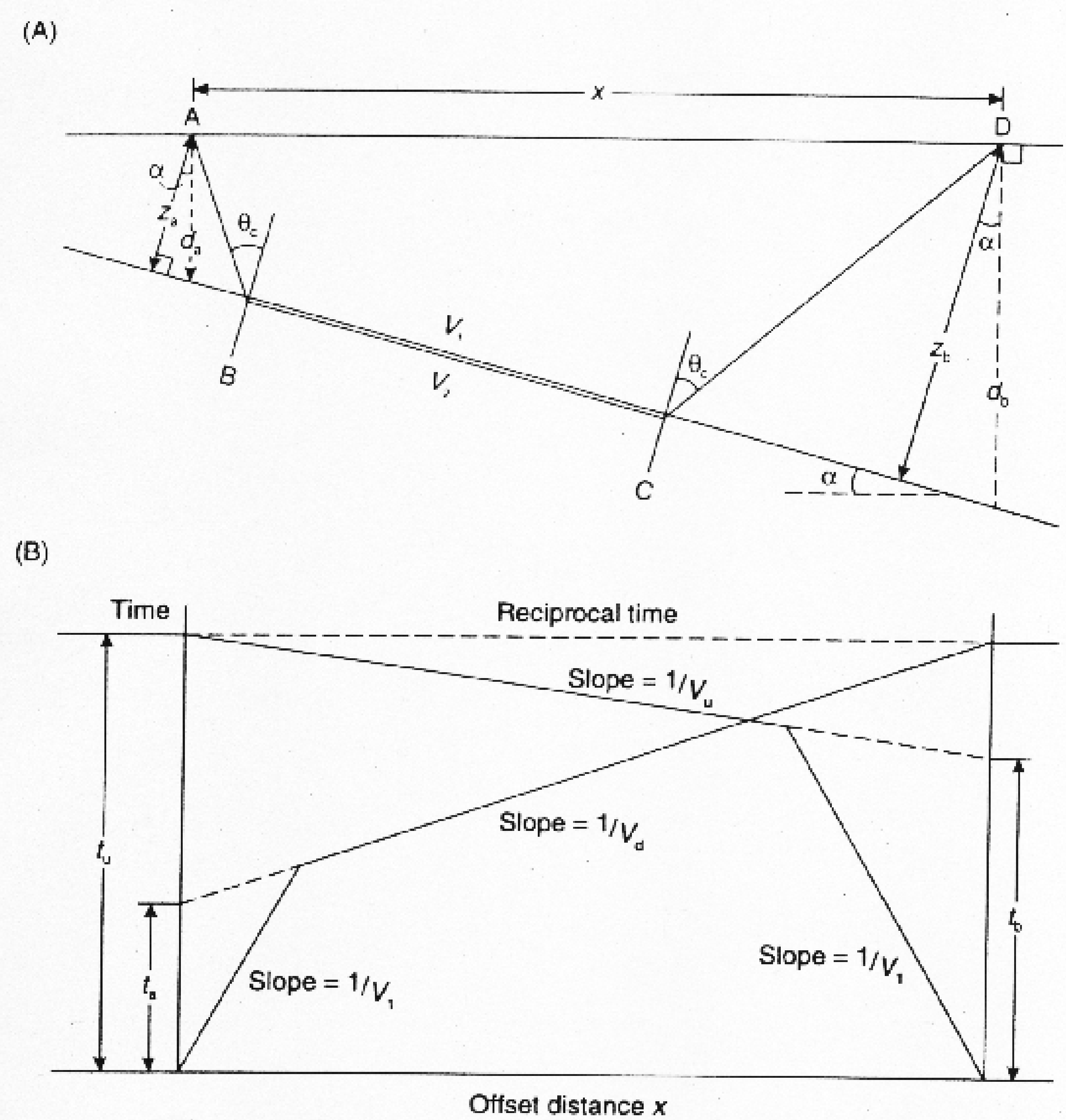


Fig. 26: (A) Raypath geometry over a refractor dipping at an angle α, and (B) the respective travel time-distance graph for the forward (down-dip) and reverse (up-dip) shooting directions. [2]

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

Travel time calculations for a dipping refractor

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:

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

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

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

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

References

1) Kearey, P., Brooks, M.: An Introduction to Geophysical Exploration, Blackwell, 2002

2) Reynolds, J. M.: An Introduction to Applied and Environmental Geophysics, Wiley, 1998

3) Kirsch, R.: Umweltgeophysik in der Praxis: Untersuchung von Altablagerungen und kontaminierten Standorten, Script

4) Dietrich, P.: Introduction to Applied Geophysics, Script, Sept. 2002