KING SAUD UNIVERSITY
COLLEGE OF SCIENCE
Department of Geology and Geophysics

PRINCIPLES OF GEOPHYSICS

(GPH 201)

2018/2019 (1439/1440)

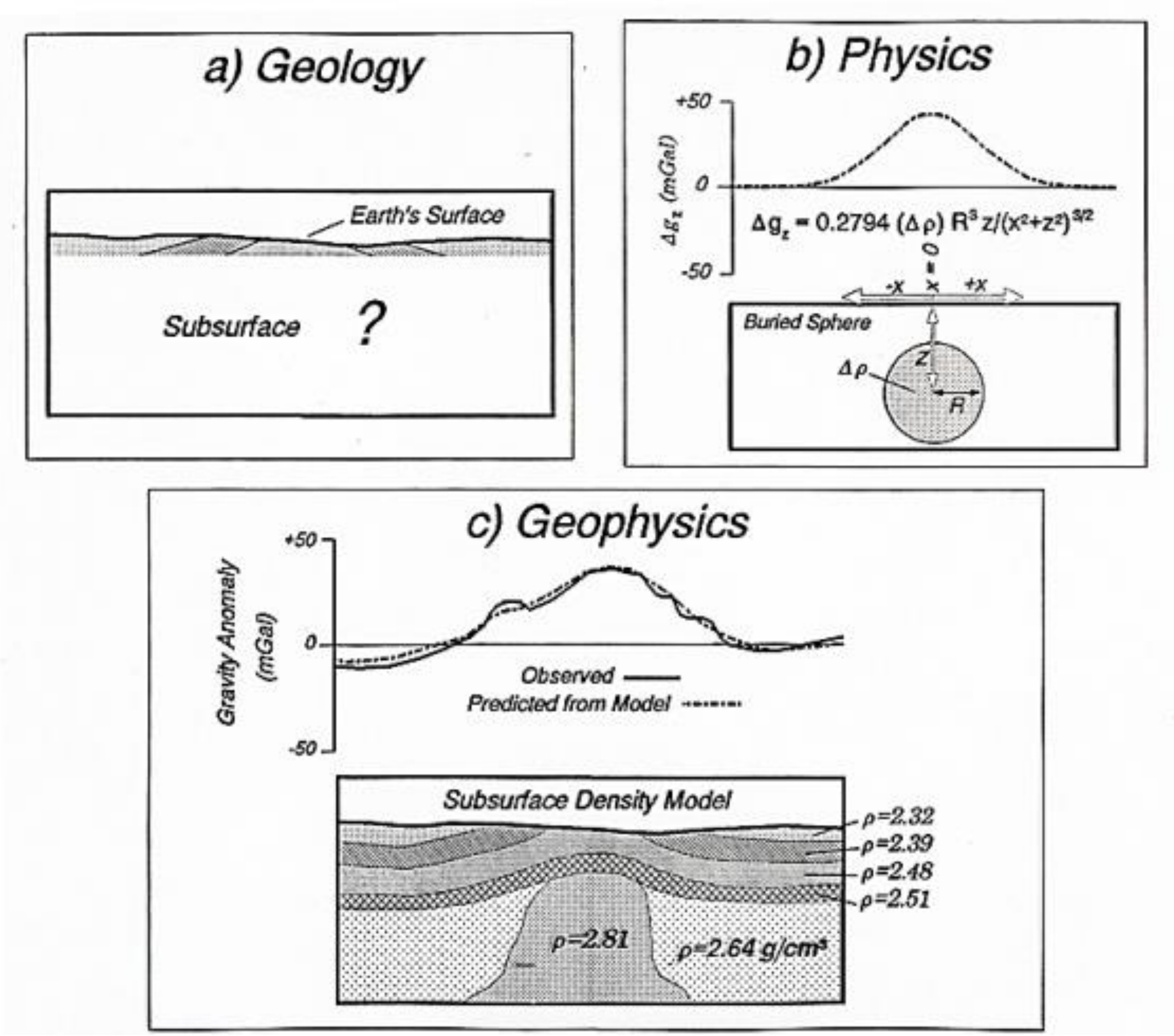
UNIT ONE

INTRODUCTION

Definition, scopes of application and classification of geophysical methods

- **Geophysics** is a multidisciplinary physical science; it is an incorporation of Physics, Mathematics, and Geology.
- **Geophysics** is the science that deals with investigation of the Earth, using the principles and laws of Physics.
- The **physical properties** of earth materials such as *density*, *elasticity*, *magnetization*, and *electrical conductivity* can be retrieved from observational measurements of the corresponding physical fields such as *gravity*, *seismic waves*, *magnetic fields*, and various kinds of electrical fields.

Geophysics = Geological Observations + Physical Laws



Divisions of Geophysics

- **Global Geophysics**
  Study of earthquakes, magnetic field, physical oceanography, Earth's thermal state and meteorology
- **Exploration Geophysics:**
  Physical principles are applied to the search for, and evaluation of, resources such as oil, gas, minerals, water and building stone.
- **Other divisions of geophysics include:**
  oceanography, atmospheric physics, climatology, petroleum geophysics, environmental geophysics, engineering geophysics and mining geophysics.

Classification of Geophysical Exploration Methods

Geophysical methods are divided into two main categories according to the source of signal; **Passive** and **Active:**

- **Passive methods (**Natural Sources):

  Measurements of naturally occurring fields. Ex. Self Potential (SP), Magnetotelluric (MT), Telluric, Gravity and Magnetic.

- **Active Methods** (Induced Sources):

  A signal is injected into the earth and then measure how the earth respond to the signal. Ex. DC Resistivity, Seismic Refraction and Ground Penetrating Radar (GPR).

Fields of Application of Geophysical Methods

- Oil and gas exploration
- Mineral exploration
- Hydrogeological investigations
- Engineering and environmental investigations
- Tectonic studies
- Natural hazards assessment (Earthquakes, landslides etc.)
- Archaeology

Geophysical Methods and their applications

Geophysical method	Dependent physical property	Applications (see key below) 1	2	3	4	5	6	7	8	9	10
Gravity	Density	P	P	s	s	s	s	!	!	s	!
Magnetic	Susceptibility	P	P	P	s	!	m	!	P	P	!
Seismic refraction	Elastic moduli; density	P	P	m	P	s	s	!	!	!	!
Seismic reflection	Elastic moduli; density	P	P	m	s	s	m	!	!	!	!
Resistivity	Resistivity	m	m	P	P	P	P	P	s	P	m
Spontaneous potential	Potential differences	!	!	P	m	P	m	m	m	!	!
Induced polarization	Resistivity; capacitance	m	m	P	m	s	m	m	m	m	m
Electromagnetic (EM)	Conductance; inductance	s	P	P	P	P	P	P	P	P	m
EM–VLF	Conductance; inductance	m	m	P	m	s	s	s	m	m	!
EM – ground penetrating radar	Permitivity; conductivity	!	!	m	P	P	P	s	P	P	P
Magneto-telluric	Resistivity	s	P	P	m	m	!	!	!	!	!

P = primary method; s = secondary method; m = may be used but not necessarily the best approach, or has not been developed for this application; (!) = unsuitable

Applications

1 Hydrocarbon exploration (coal, gas, oil)
2 Regional geological studies (over areas of 100s of km^2)
3 Exploration/development of mineral deposits
4 Engineering site investigations
5 Hydrogeological investigations
6 Detection of sub-surface cavities
7 Mapping of leachate and contaminant plumes
8 Location and definition of buried metallic objects
9 Archaeogeophysics
10 Forensic geophysics

UNIT TWO

Basics of Seismic Methods

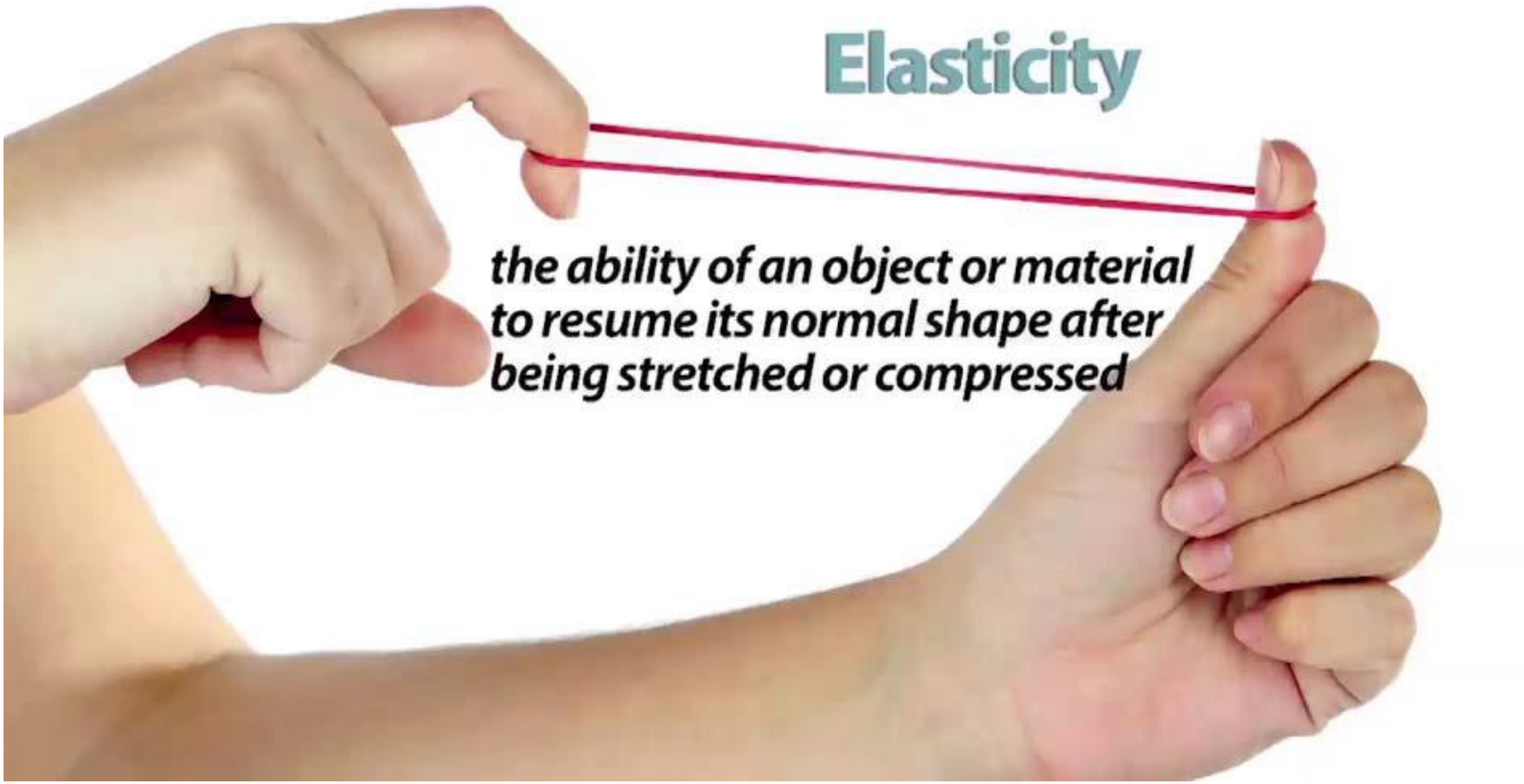
Fundamental Considerations, seismic waves, characteristics of seismic waves propagation

Fundamental Considerations

- In seismic surveying, seismic waves are created by a controlled source and **propagate** through the subsurface.
- Some waves return to the surface after **refraction** or **reflection** at Geological boundaries within the subsurface.
- Instruments distributed along the surface detect the **ground motion** caused by these returning waves and hence **measure the arrival times of the waves** at different ranges from the source.

- Seismic methods are particularly well suited to **mapping of layered sedimentary sequences** and are therefore widely used in the search for oil and gas.
- The methods are also used, on a **smaller scale**, for mapping of near-surface layers, locating groundwater aquifers and in site investigation and determination of depth to bedrock.
- Seismic surveying can be carried out on land or at sea and is used extensively in offshore geological surveys and the exploration for offshore resources.

Elasticity


Elasticity
the ability of an object or material to resume its normal shape after being stretched or compressed

- **Elasticity** is a **physical property** of a material that is defined as: *the ability of an elastic material to restore its original shape after the removal of the deforming force.*
- When an elastic material is deformed due to an external force, it experiences internal forces that oppose the deformation and restore it to its original state if the external force is removed.

- There are various **elastic moduli**, such as Young's modulus, the shear modulus, and the bulk modulus, all of which are *measures of the inherent stiffness of a material as a resistance to deformation under an applied load*. Various moduli apply to different kinds of deformation.

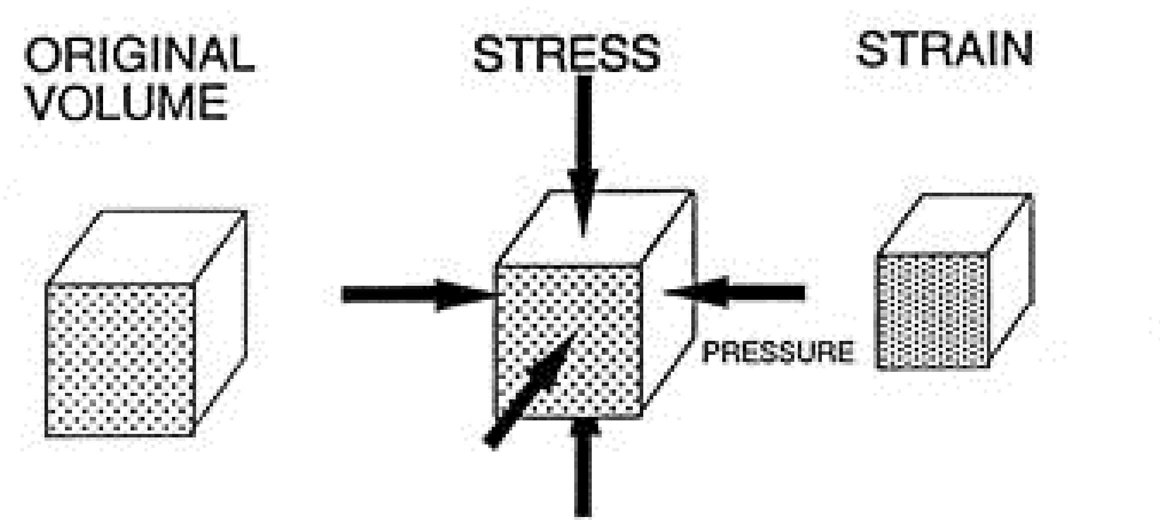
- The **elasticity** of a material is described by a **stress-strain curve**, which shows the relation between stress and strain.

- To understand the propagation of elastic waves we need to describe the deformation of our medium and the acting stress. <u>The relation between stress and strain is governed by</u> ***elastic constants***.

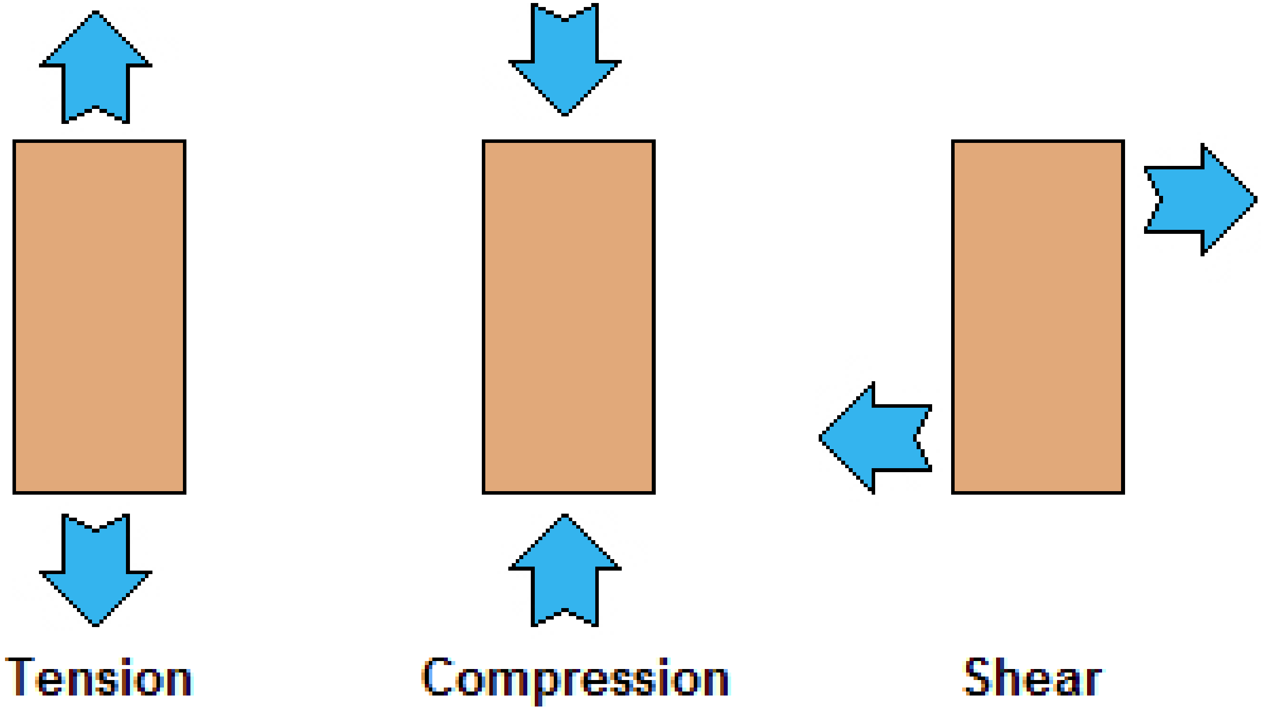
Stress and Strain

- **Stress** is the ratio of applied force F to the area across which it acts.

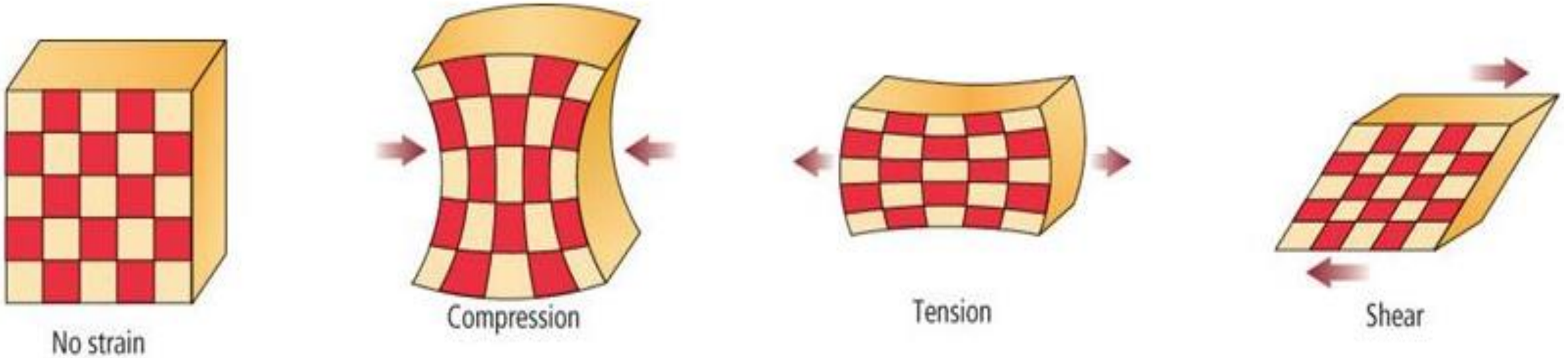
- **Strain** is the deformation caused in the body, and is expressed as the ratio of change in length (or volume) to original length (or volume).



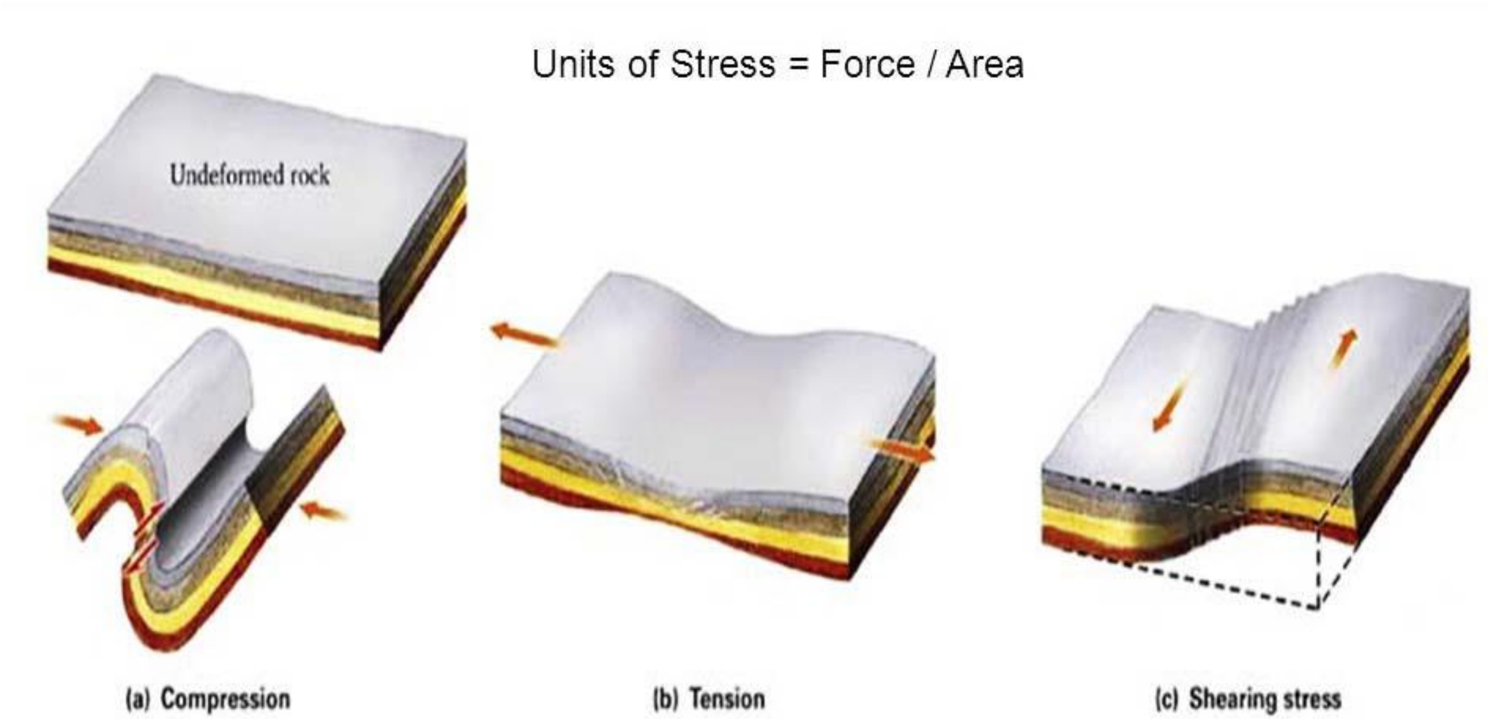
Types of stress



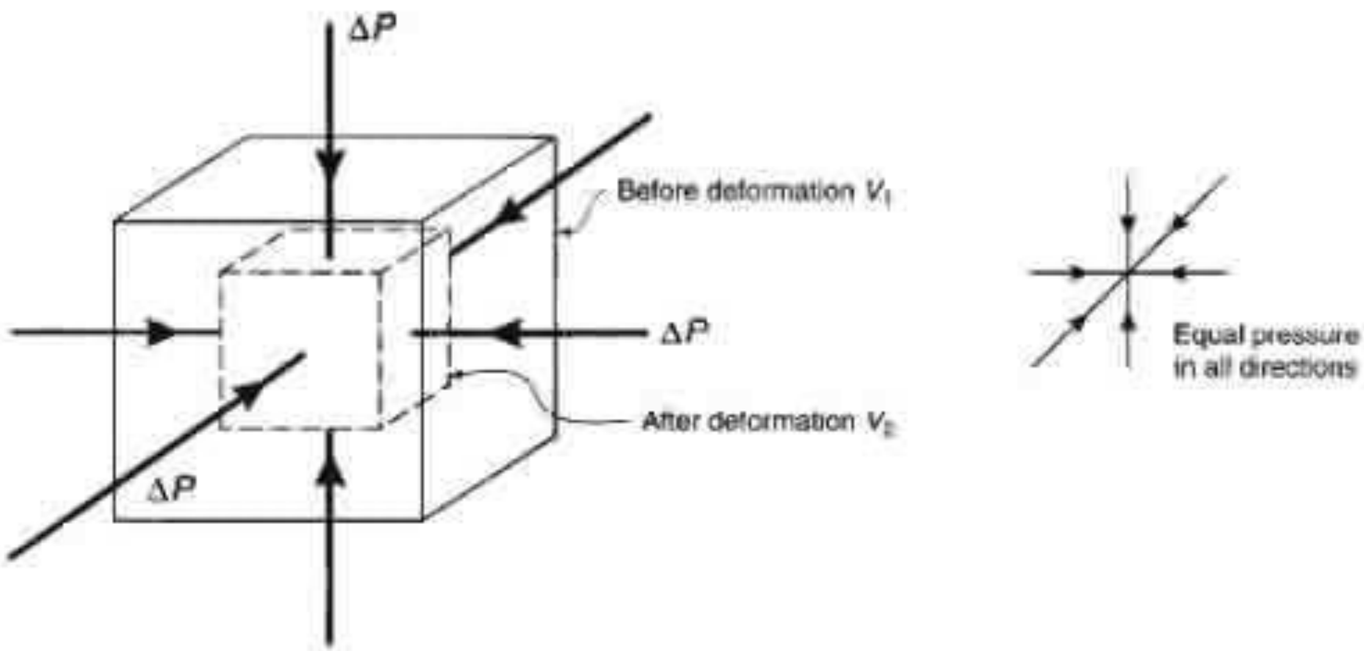
- **Compression** causes a material to shorten.
- **Tension** causes a material to lengthen.
- **Shear** causes distortion of a material.



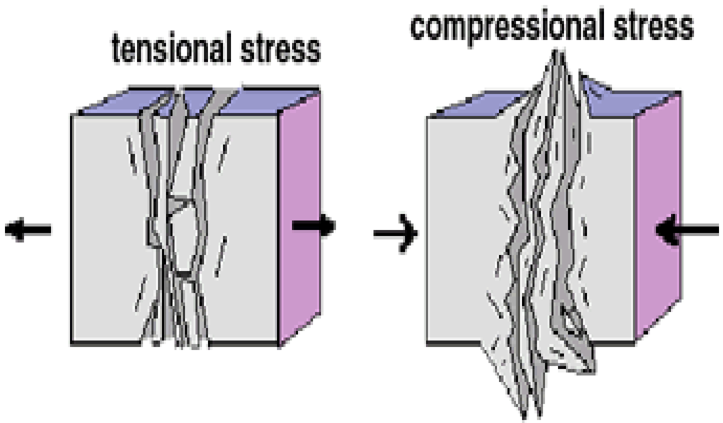
- Stress towards the interior ☞ compression.
- Stress towards the exterior ☞ tension (extension, dilatation).

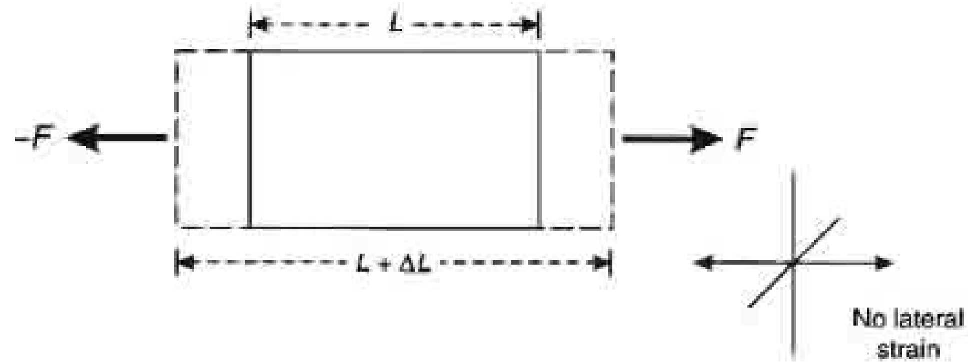


Normal Stress (Pressure): Forces act equally in all directions perpendicular to faces of body, e.g. pressure on a cube in water:



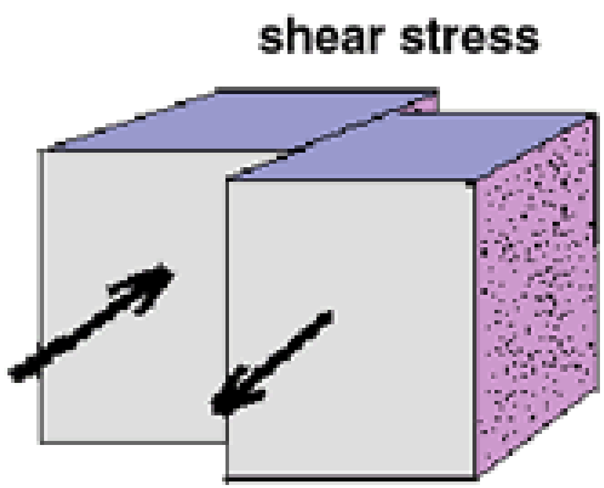
- **Axial Stress**:

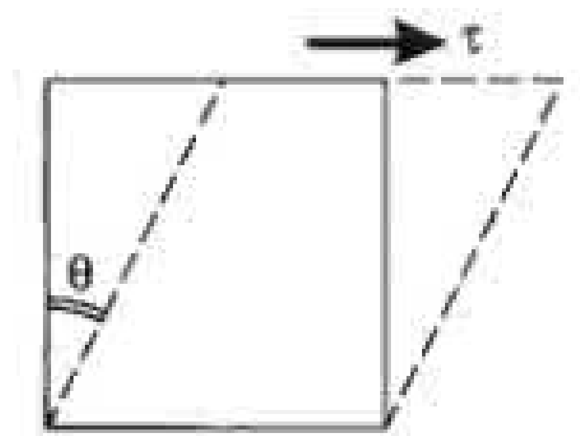


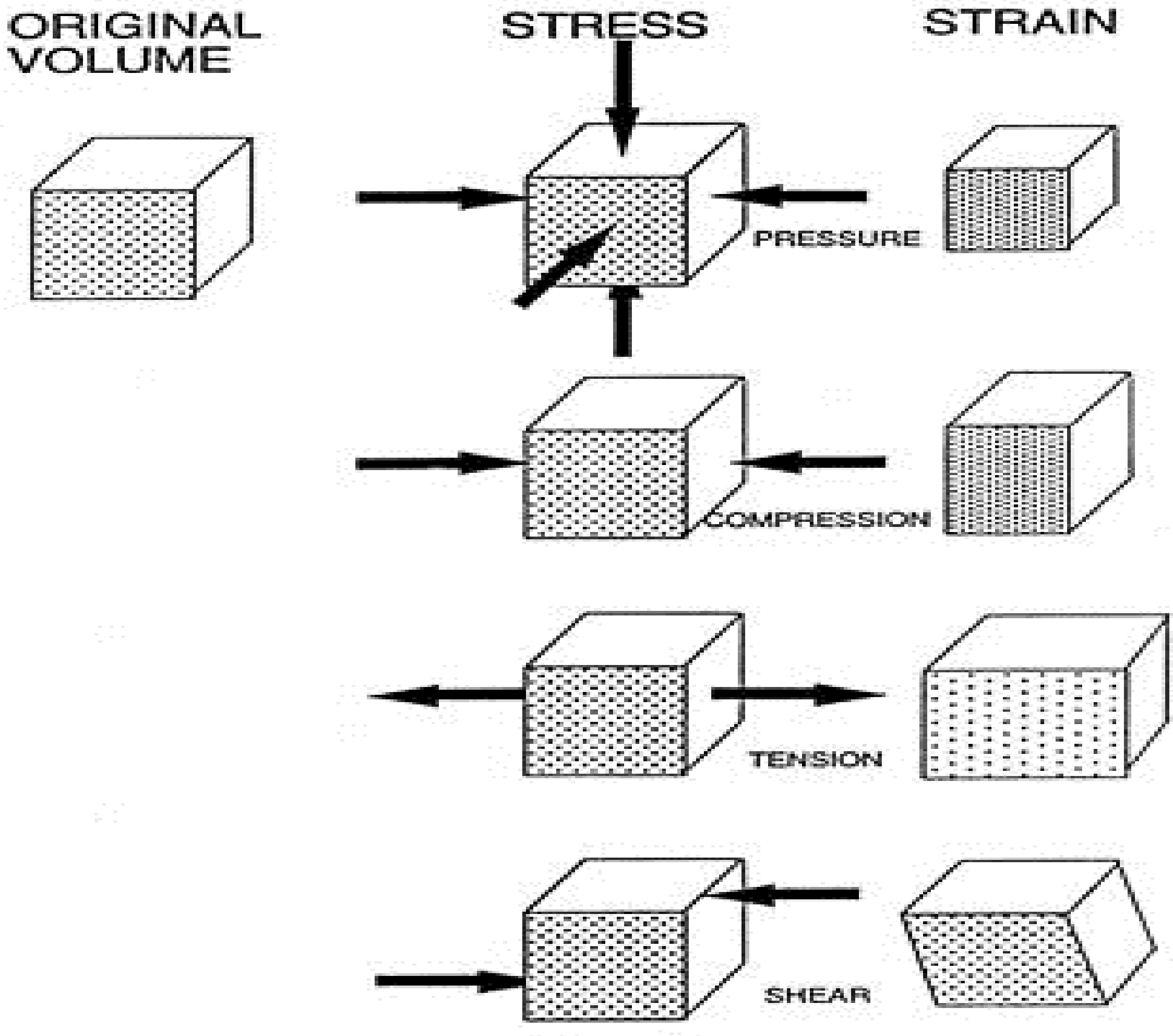


> ➢ **stress acts in one direction only.**
>
> ➢ **change in volume of solid occurs.**
>
> ➢ **associated with P wave propagation.**

- **Shear Stress:** stress acts parallel to a face of a solid, e.g. pushing along a table:
    - No change in volume.
    - Fluids such as water and air **do not support** shear stresses.
    - **Associated with S wave propagation**






ORIGINAL VOLUME
STRESS
STRAIN
PRESSURE
COMPRESSION
TENSION
SHEAR

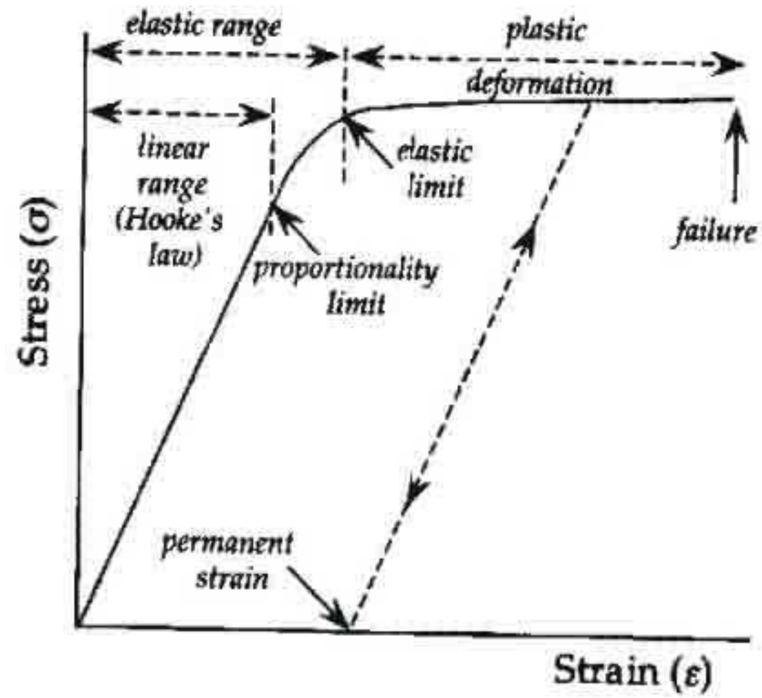
Stress-Strain relation

Hooke’s Law

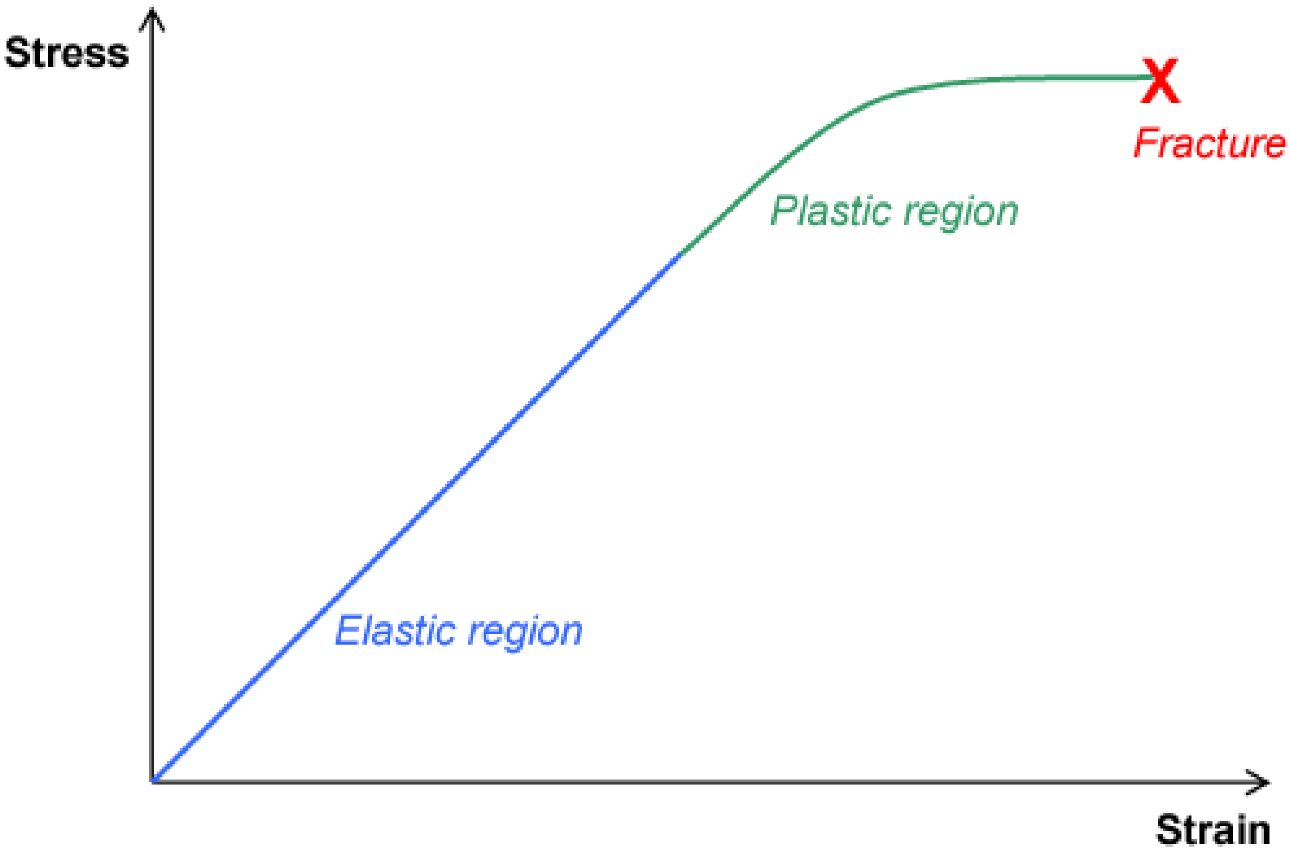
Hooke's Law

Stress is proportional to strain:

- At **low to moderate strains**: Hooke's Law applies and a solid body is said to behave ***elasticaly***, i.e. will return to original form when stress is removed.
- At **high strains**: the elastic limit is exceeded and a body deforms in a plastic or ductile manner: it is unable to return to its original shape, being permanently strained, or damaged.
- At **very high strains**: a solid will fracture, e.g. in earthquake faulting.



Hooke's Law



Hooke's Law

$$F \propto e$$

This is the force applied (N)

This is the extension (m)

Constant of proportionality is *the <u>modulus</u>. It* is the *ratio of stress to strain*.

Elastic Constants

- Elastic constants describes the strain of a material due to an applied stress.

Stress = Elastic Modulus * strain

- **Elastic Modulus** is the *constant of proportionality of Stress vs Strain*.
- The <u>higher the value of the modulus</u>, the <u>stronger the material</u>, the <u>smaller the strain</u> produced by a given stress.

Elastic Constants

- ***Elastic constants*** are different for different kinds of stress (twisting, compressing, stretching) and for different materials.
- Based on the relationships between elastic moduli and Lamé coefficients (λ and μ), the elasticity can be quantified by various elastic moduli:

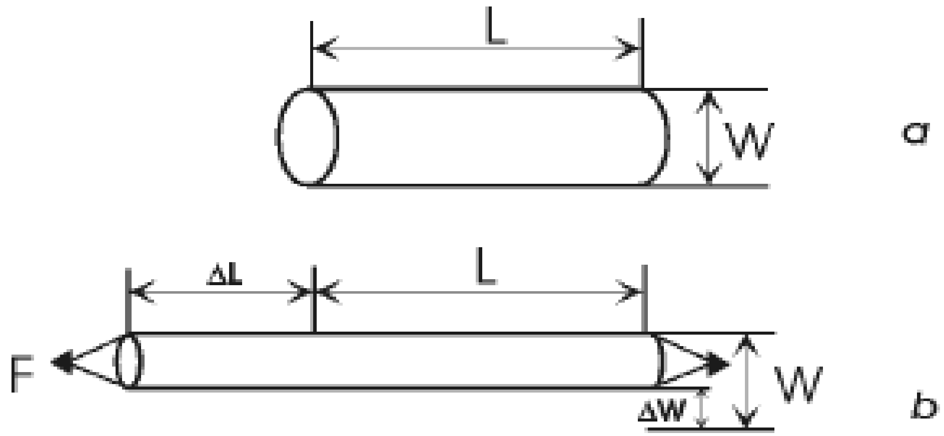
Elastic Constants

- Young's modulus (stretch modulus): **E**
- Bulk modulus (incompressibility): **K**
- Shear Modulus (rigidity): **μ**
- Axial Modulus (**ψ**)
- Poisson's Ratio: **σ**

Young’s Modulus (E)

- **Young’s Modulus (E):** the ratio of extensional stress to the resulting extensional strain for a cylinder being pulled apart at both ends.

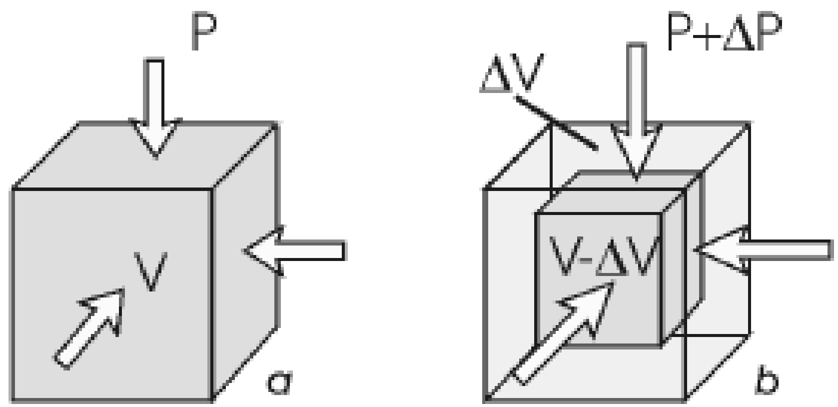
- Longitudinal strain is proportional to longitudinal stress.



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

Bulk modulus: K

- **Bulk Modulus (K):** Measure of the capacity of the material to be compressed. It can be carried out for solid, liquid, and gas.



$$k = \frac{\text{stress}}{\text{strain}} = \frac{\Delta P}{\Delta V / V}$$

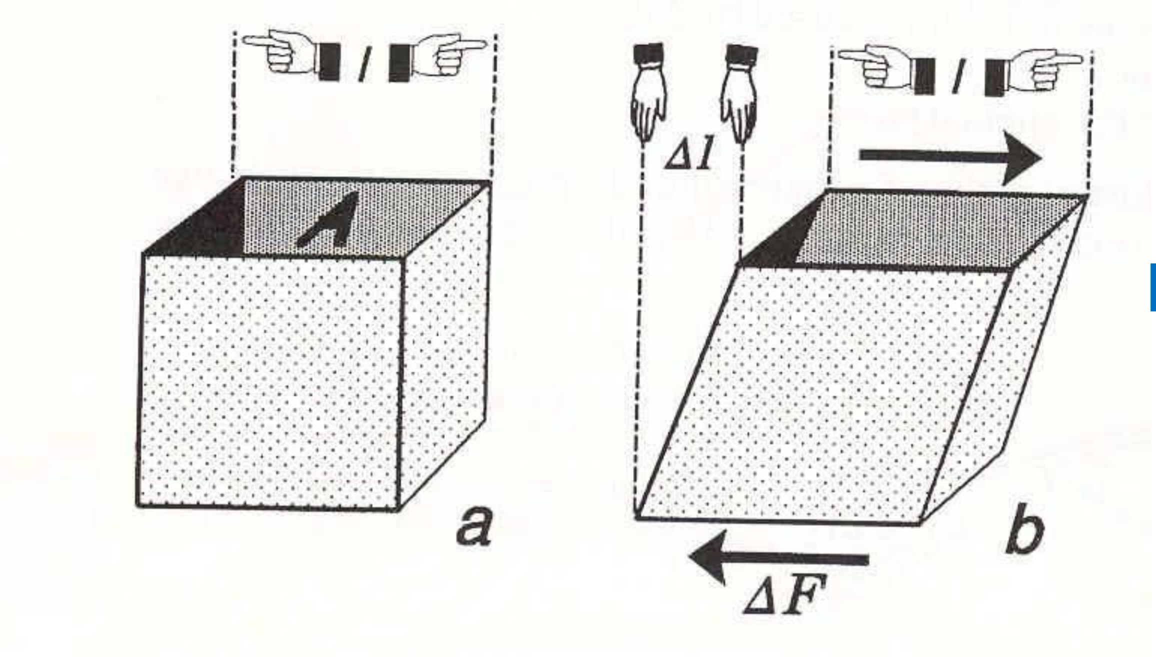
vhere:

$\Delta P = P' - P$ = pressure change (applied stress)
P = original confining pressure
P' = confining pressure under the applied stress
$\Delta V = V - V'$ = change in volume caused by ΔP
V = original volume

Shear Modulus (μ)

- **Shear Modulus (μ):** Measures the amount of angular deformation due to the application of a shear stress on one side of the object.

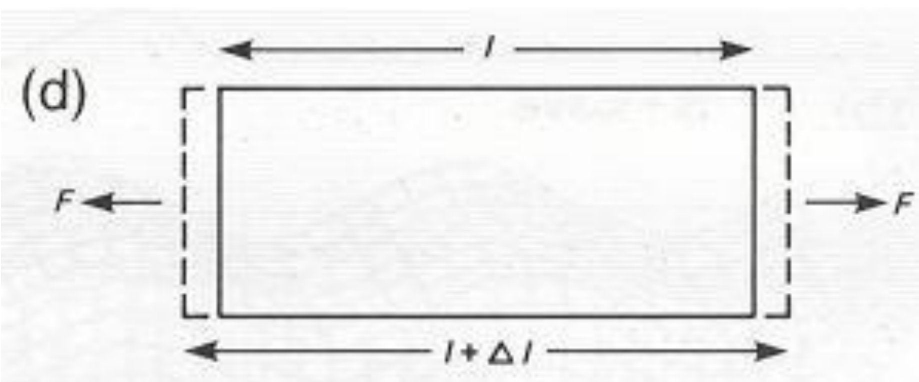
- μ =0 for liquid and gas (no rigidity)



$$\mu = (\Delta F/A) / \tan\theta$$

Axial Modulus (ψ)

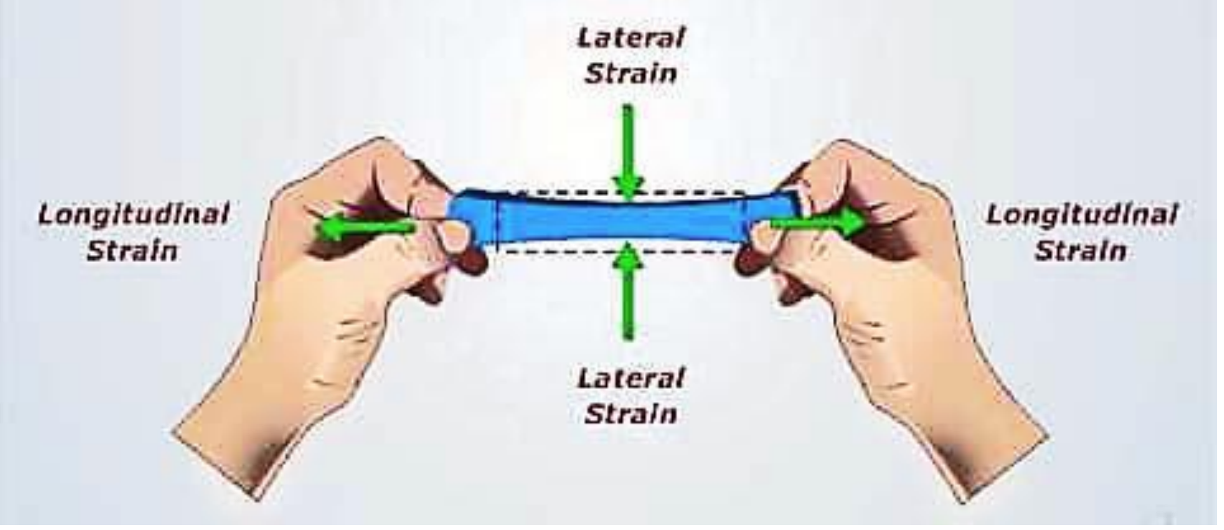
Axial Modulus (ψ): The response to longitudinal stress, similar to Young's Modulus, except that strain is uniaxial – no transverse strain associated with the application of the longitudinal stress.



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

Poisson's Ratio (σ)

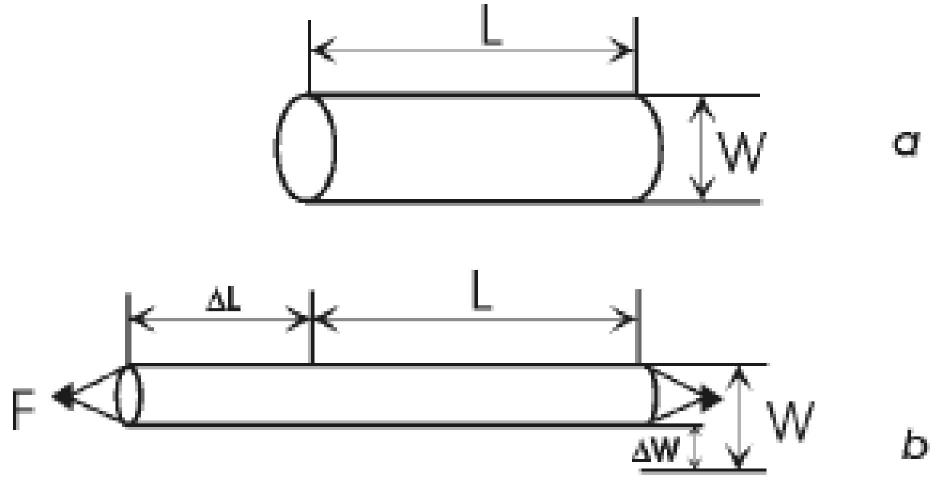
- When a material is compressed in one direction, it tends to expand in the other two directions perpendicular to the direction of compression.



Poisson's Ratio: σ

- This Modulus is **defined as** the *<u>ratio of transverse contraction strain to longitudinal extension strain in the direction of stretching force</u>* .

$$\sigma = (\Delta W/W) \,/\, (\Delta L/L)$$



Elastic Constants

μ = shear modulus (as before)

λ = first Lamé coefficient (no direct physical interpretation)

Young's Modulus: $E = \mu\,(3\lambda + 2\,\mu) \div (\lambda + \mu)$

Bulk modulus: $K = \lambda + 2/3\;\mu$

Poisson's Ratio: $\sigma = \lambda\,/\,2\,(\lambda + \mu)$

Lamé 1 in terms of Poisson & Young

$\lambda = E\,\sigma\;/\;(1 + \sigma)(1 - 2\sigma)$

Seismic waves

Seismic waves

- **Seismic waves** are parcels of elastic strain energy that propagate outwards from a seismic source such as an earthquake or an explosion.
- Seismic waves travel away from any seismic source at speeds determined by the elastic moduli (Young's modulus *E*; Bulk modulus *K*; Shear modulus μ. and Axial modulus Ψ) and the densities of the media through which they pass.

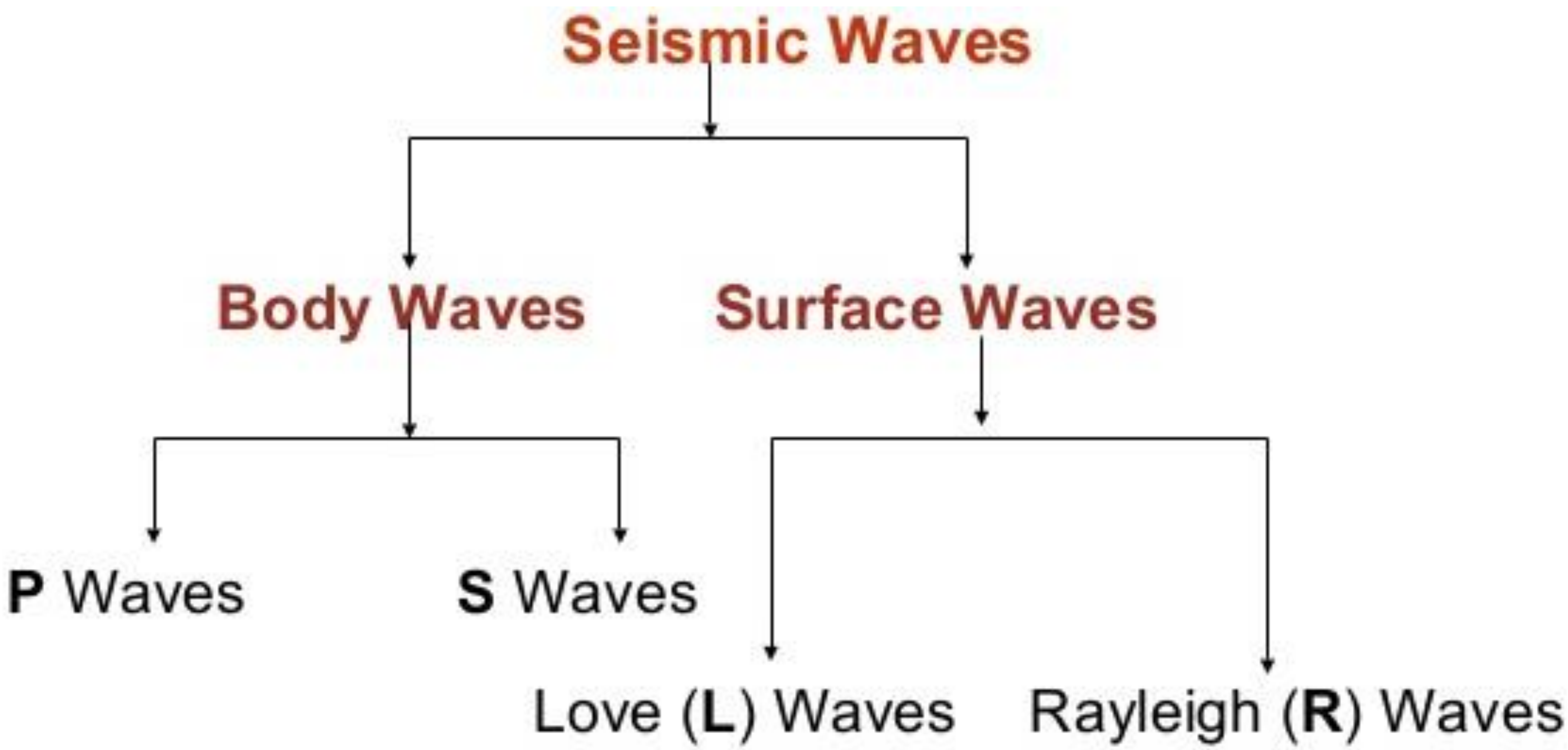
Seismic waves

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

Body waves can propagate through the internal volume of an elastic solid and may be of two types:

- *Compressional waves* (the longitudinal, primary or *P-waves* of earthquake seismology)
- *Shear waves* (the transverse, secondary or *S-waves* of earthquake seismology).

Seismic waves



Body waves, P and S

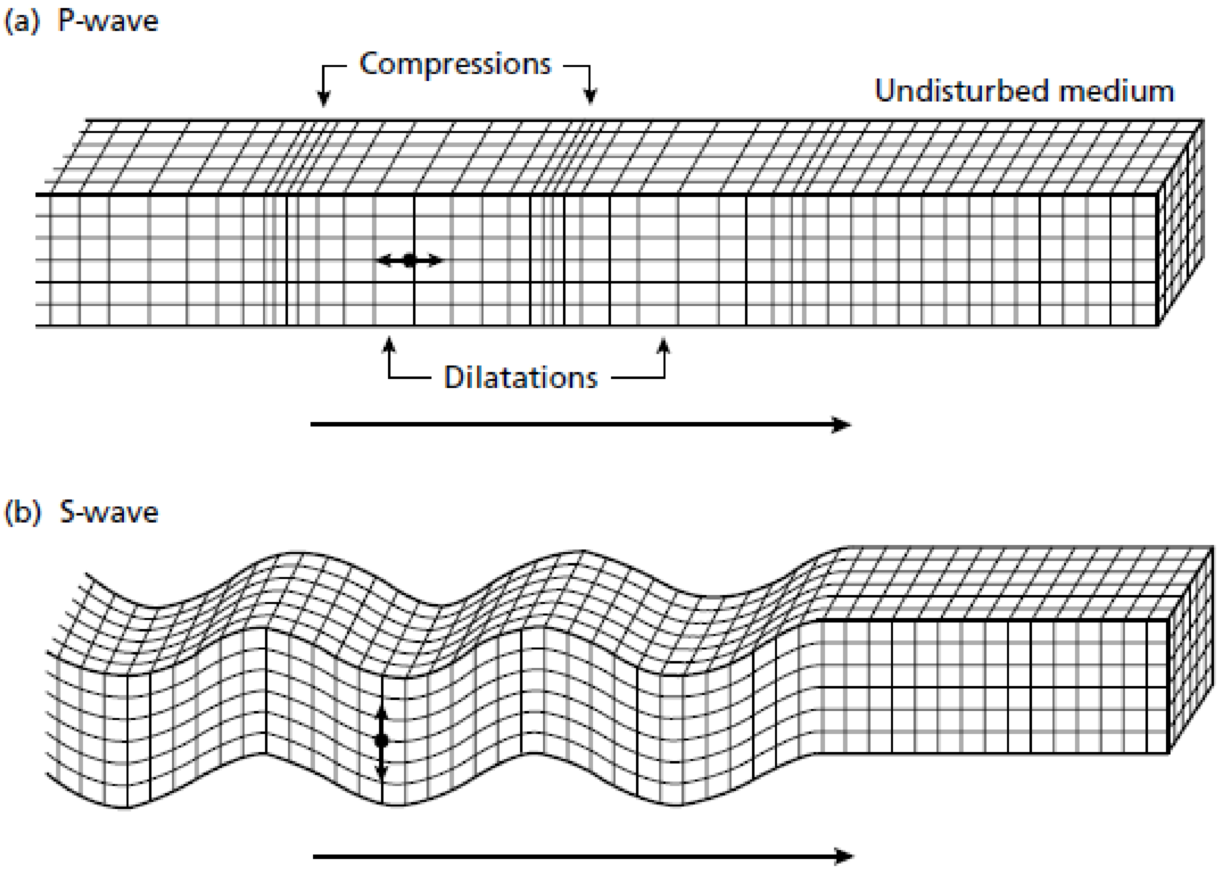


Fig. 3.3 Elastic deformations and ground particle motions associated with the passage of body waves. (a) P-wave. (b) S-wave. (From Bolt 1982.)

Surface waves

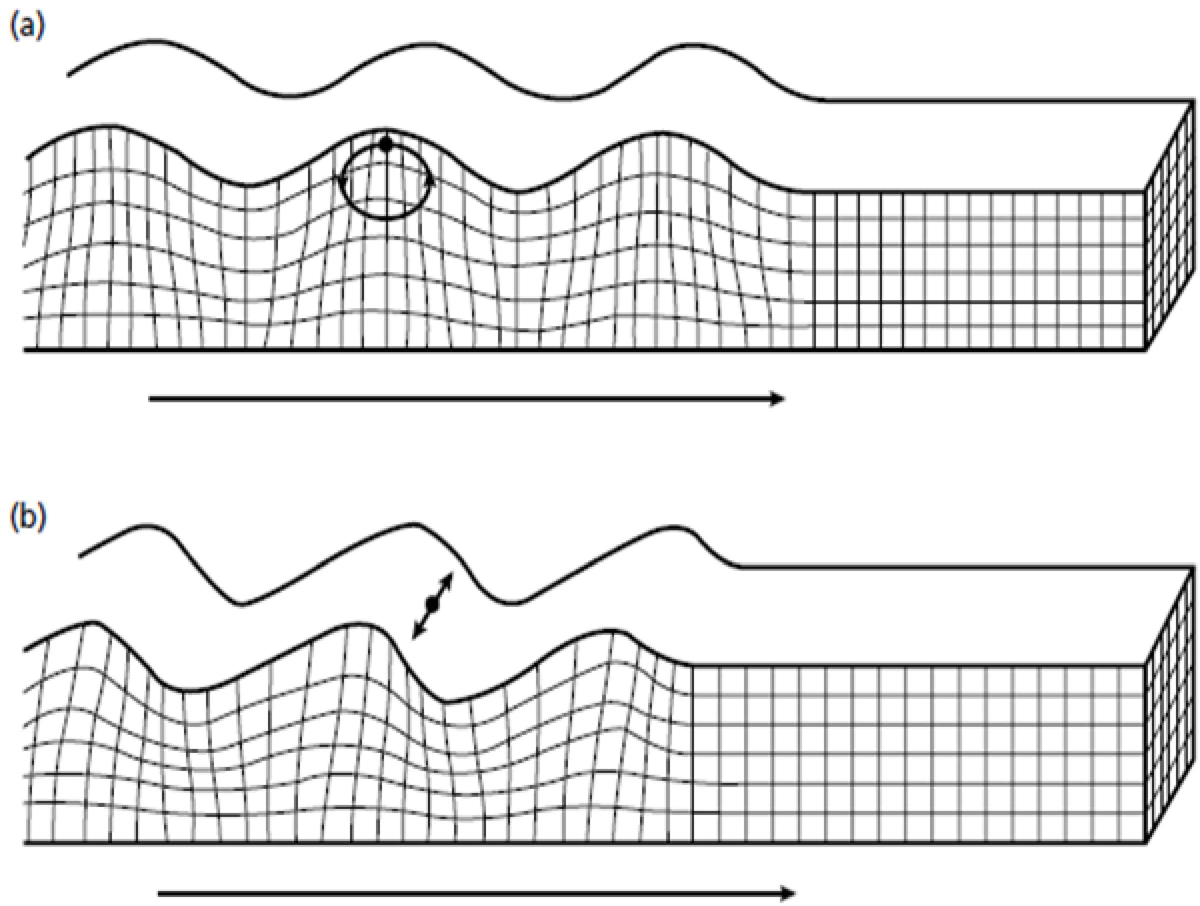


Fig. 3.4 Elastic deformations and ground particle motions associated with the passage of surface waves. (a) Rayleigh wave. (b) Love wave. (From Bolt 1982.)

- One application of **<u>shear wave</u>** is in *engineering site investigation* where the separate measurement of <u>*V*p</u> and <u>*V*s</u> for near-surface layers allows direct calculation of *Poisson’s ratio* and estimation of the elastic moduli, which provide valuable information on the in situ geotechnical properties of the ground. These may be of great practical importance, such as the value of *rippability.*

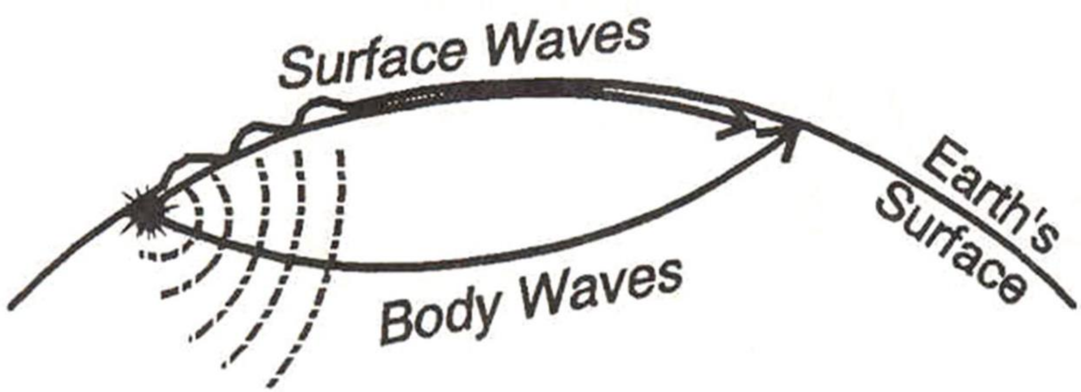
Surface waves

- **Surface waves** can propagate along the boundary of the solid.
- Two types of surface waves: **Rayleigh waves** and **Love wave**
- *Rayleigh waves* propagate along a free surface, or along the boundary between two dissimilar solid media.

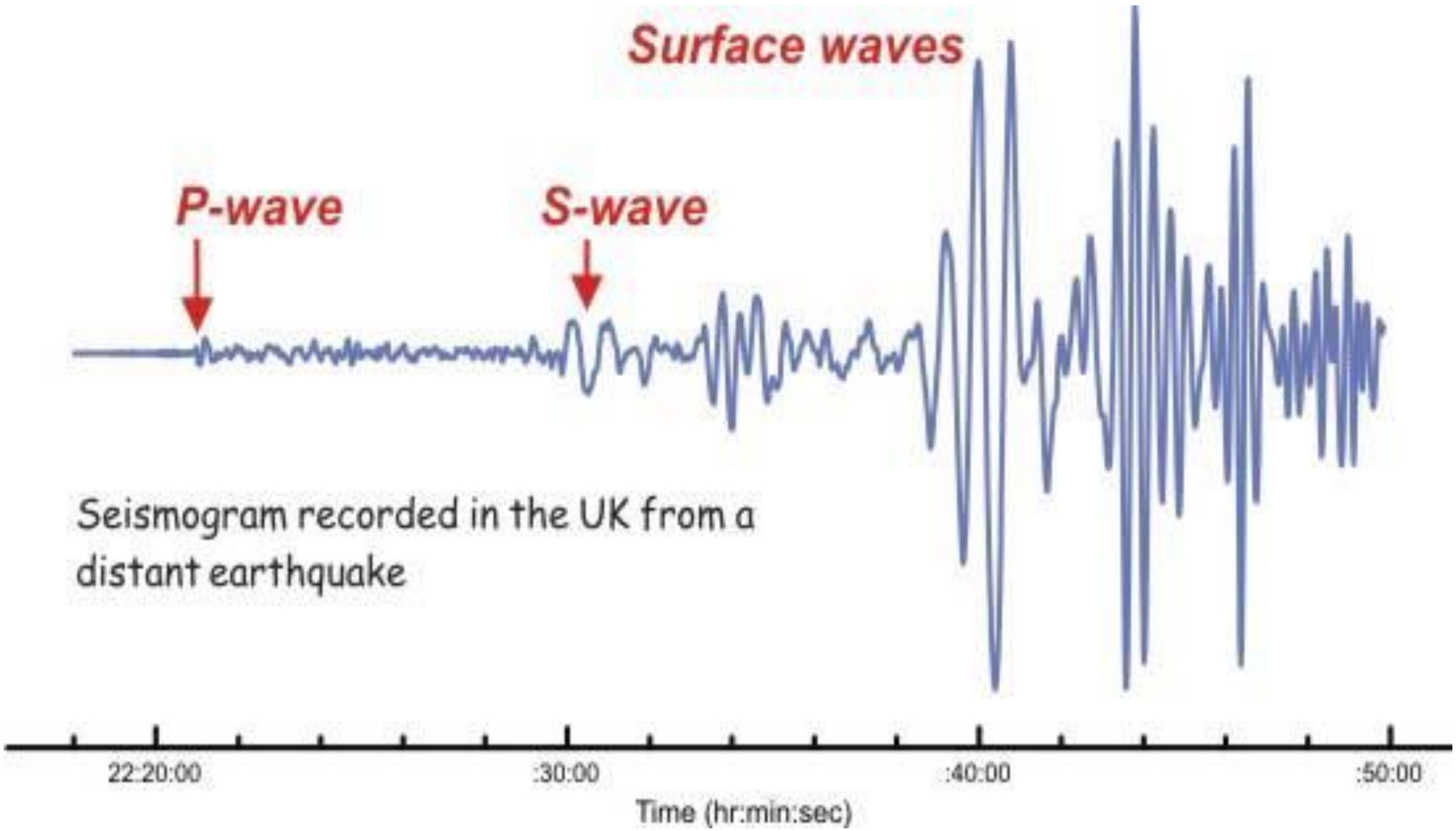
Seismic waves

Types of Seismic Waves

Primary Wave (P-wave)	Secondary Wave (S-wave)	Surface Wave
• Travels through ground	• Travels through ground	• Travels ***only*** on Earth's surface
• Fastest waves	• Medium speed waves	• Slowest waves
• Can travel through solid and liquid	• Only travel through solids	



Velocity of waves


Surface waves
P-wave
S-wave
Seismogram recorded in the UK from a distant earthquake
22:20:00
:30:00
:40:00
:50:00
Time (hr:min:sec)

Velocity of Body waves

- V_p : Velocity of the *compressional* wave
  V_s: Velocity of the *shear* wave
- For the <u>same material</u>, *Vp > Vs*.
- The *more rigid* the material, the *higher* V_p *and* V_s.
- <u>Shear waves cannot travel through fluids</u> (V_s =0).

Velocity of Body waves

<u>Relationship between Vp and Vs</u>

Compressional Waves

$$V_p = \sqrt{\frac{\left(\frac{4}{3}\mu + k\right)}{\rho}}$$

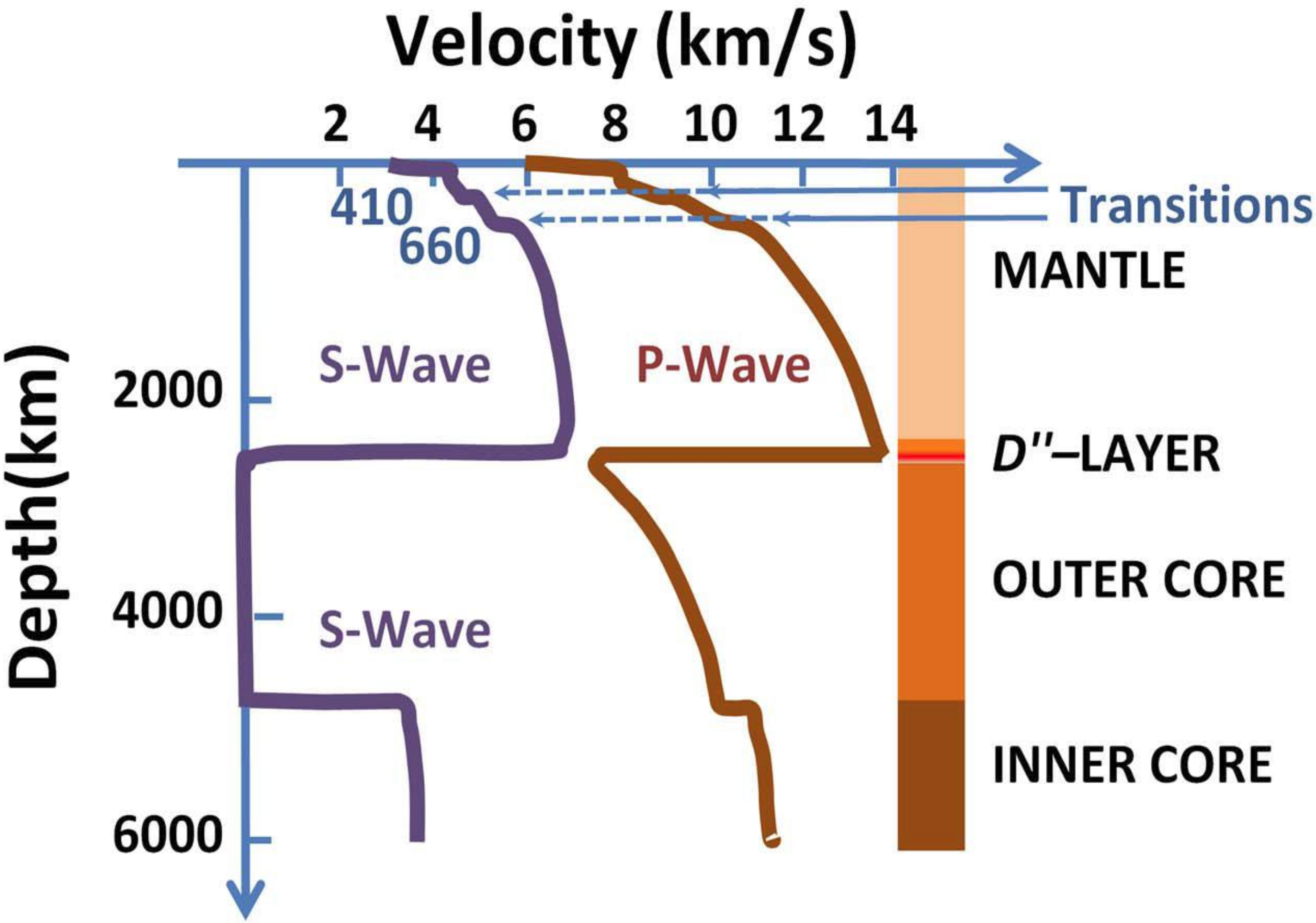
Shear Waves

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

- Averaged Vp/Vs = 1.732 for the crust
- For mafic rocks, Vp/Vs = 1.81
- For felsic rocks, Vp/Vs = 1.70

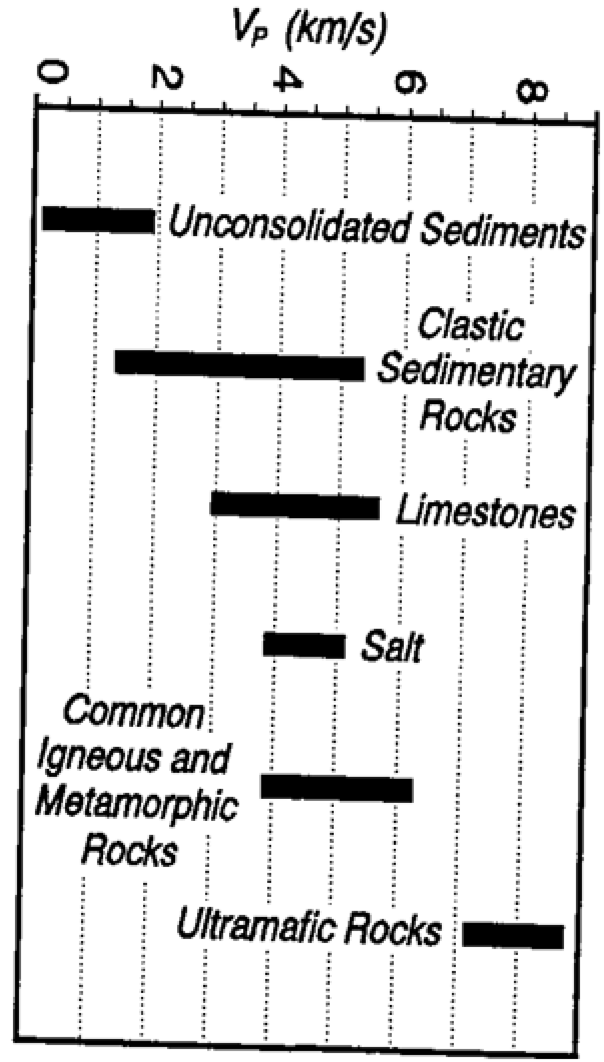
Note that:

Poisson's ratio (σ) = Vp/Vs


Velocity (km/s)
2
4
6
8
10
12
14
410
660
Transitions
MANTLE
S-Wave
P-Wave
D''–LAYER
OUTER CORE
S-Wave
INNER CORE
Depth(km)
2000
4000
6000

Typical values for elastic constants, density, and seismic velocities for selected materials, listed according to increasing compressional wave velocity (V_p). Compiled from Kinsler et al. (1982) and other sources. SI units for density are kg/m^3; the literature, however, commonly gives densities in g/cm^3.

	ELASTIC CONSTANTS 10^9 N / m^2			SEISMIC VELOCITIES km / s	
	Bulk Modulus (k)	Shear Modulus (μ)	Density (ρ) kg / m^3 / g / cm^3	Compres. Wave (V_p)	Shear Wave (V_s)
Air	0.0001	0	1.0 / 0.001	0.32	0
Water	2.2	0	1000 / 1.0	1.5	0
Ice	3.0	4.9	920 / 0.92	3.2	2.3
Shale	8.8	17	2400 / 2.4	3.6	2.6
Sandstone	24	17	2500 / 2.5	4.3	2.6
Salt	24	18	2200 / 2.2	4.7	2.9
Limestone	38	22	2700 / 2.7	5.0	2.9
Quartz	33	39	2700 / 2.7	5.7	3.8
Granite	88	22	2600 / 2.6	6.7	2.9
Peridotite	139	58	3300 / 3.3	8.1	4.2



Approximate ranges of compressional wave velocity (V_P) for some materials encountered at Earth's surface (from Griffiths and King, 1981).

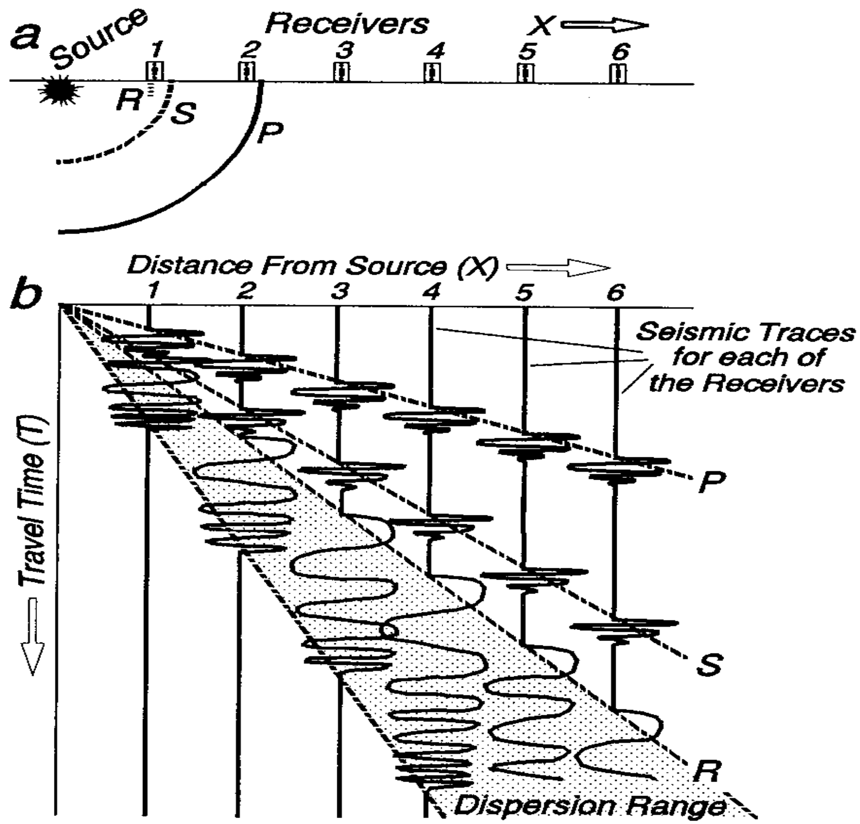


FIGURE 3.17 a) Initial wave fronts for P, S, and R waves, propagating across several receivers at increasing distance from the source. b) Travel-time graph. The seismic traces are plotted according to the distance (X) from the source to each receiver. The elapsed time after the source is fired is the travel time (T).

On a *travel-time graph* seismic traces from several receivers are plotted side by side, according to the horizontal distance (X) from the source to each receiver (Fig. 3.17). Travel time (T) is commonly plotted as *increasing downward* in refraction and reflection studies, because T often relates to depth within the Earth. For each of the initial P-wave, S-wave, or R-wave arrivals, the travel time from the source to a receiver is linear, expressed by the *travel-time curve:*

$$\boxed{T = \frac{X}{V}}$$

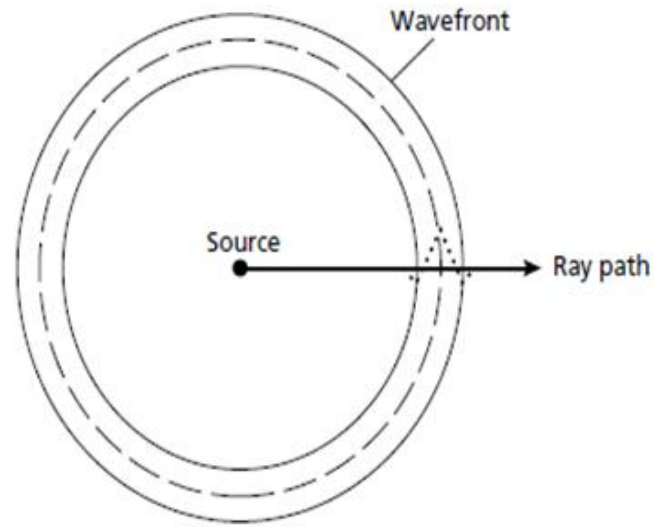
where:

T = total time for the wave to travel from the source to the receiver
X = distance from the source to the receiver, measured along the surface
V = seismic velocity of the P, S, or R arrival.

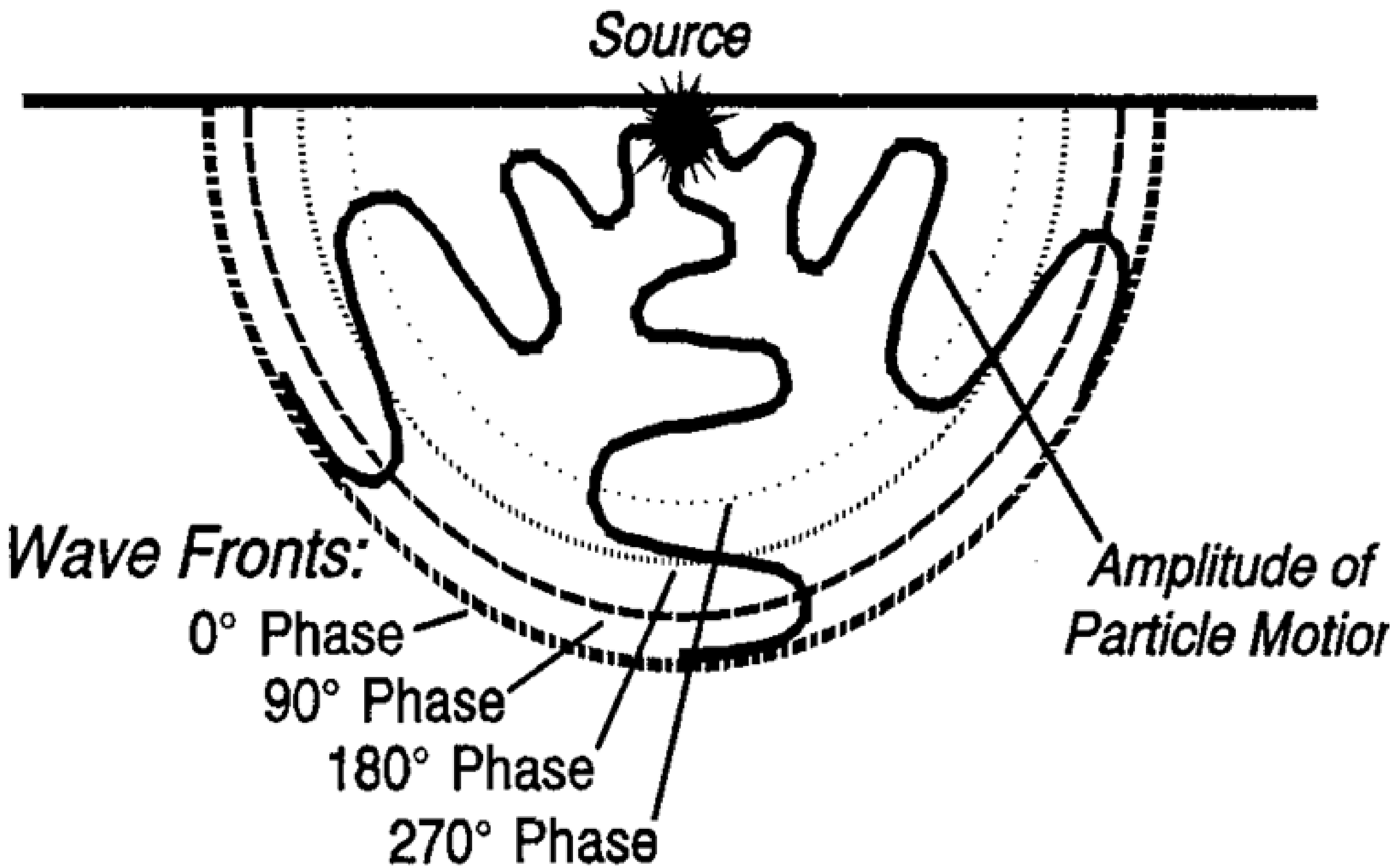
Wave front and Ray path

Waves and Ray

- A seismic pulse propagates outwards from a seismic source at a *velocity determined by the physical properties of the surrounding rocks.*
- If the pulse travels through a homogeneous rock it will travel at the same velocity in all directions away from the source so that at any subsequent time the ***wavefront***, *defined as the locus of all points which the pulse has reached at a particular time*, will be a sphere.
- The propagation velocity of a seismic wave is the velocity with which the seismic energy travels through a medium.

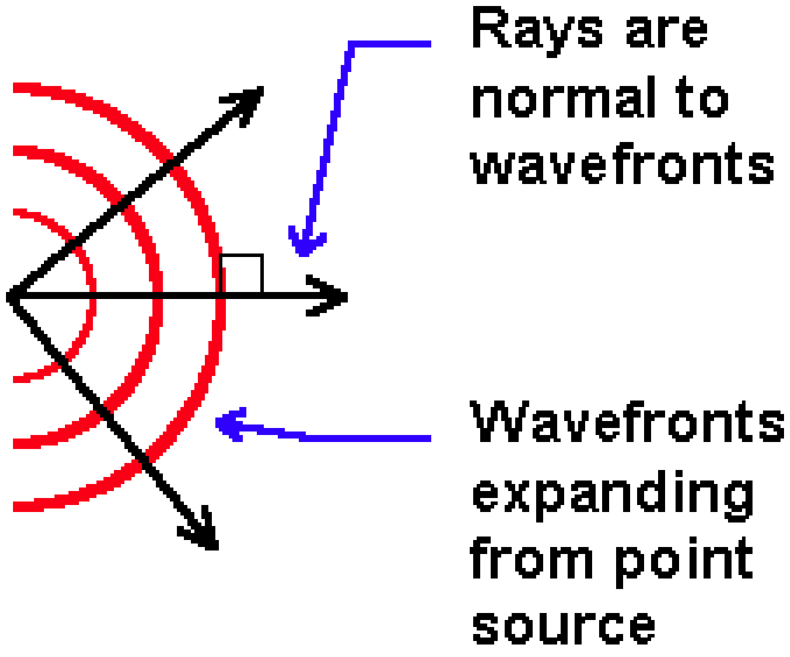


The relationship of a ray path to the associated wavefront.



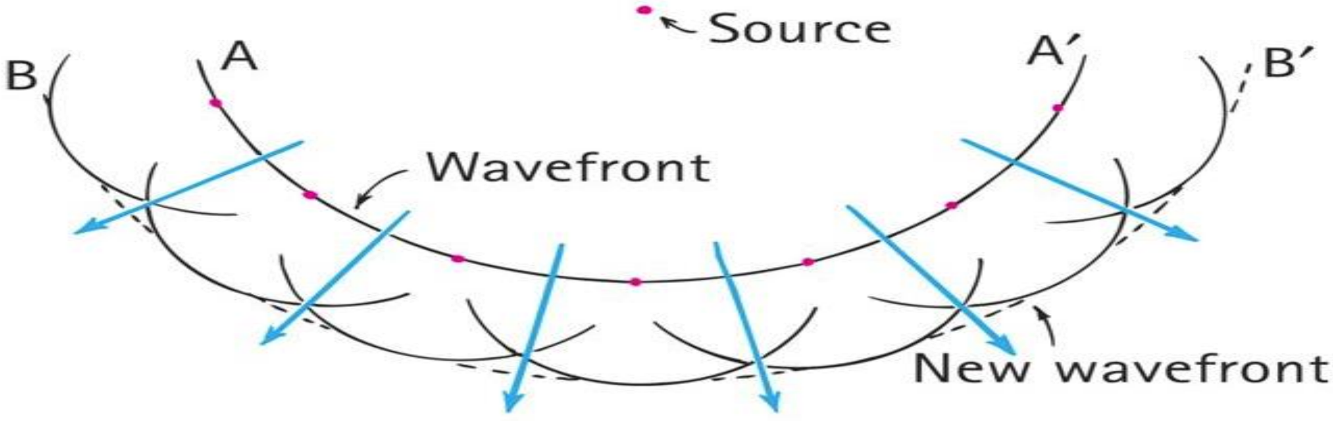
Wave fronts are surfaces along which particle motions of the propagating wave are in phas((one complete oscillation is 360° of phase). For example, a surface where particle motions reach their maximum positive amplitude is 90° phase; where they are maximum negative amplitude is 270° phase.

Wave front and Ray path



Huygens's Principle

- Huygens's Principle is a method of analysis applied to problems of wave propagation.
- It says that “*every point on a wave-front could be considered a source of a secondary spherical wavelets which spread out in the forward direction*”.



Ray paths in layered media

- **At an interface** between two rock layers there is a change of 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 depends on: the **velocities and densities** of the two layers, and the **angle of incidence** on the interface.

Ray paths in layered media

- Seismic energy is partitioned when waves encounter materials of different acoustic impedance (ρ*V).

- For example when a P wave traveling in one material strikes the boundary of another material at an oblique angle, the energy separates into four phases: Reflected P wave, Reflected S wave, Refracted P wave and Refracted S wave.

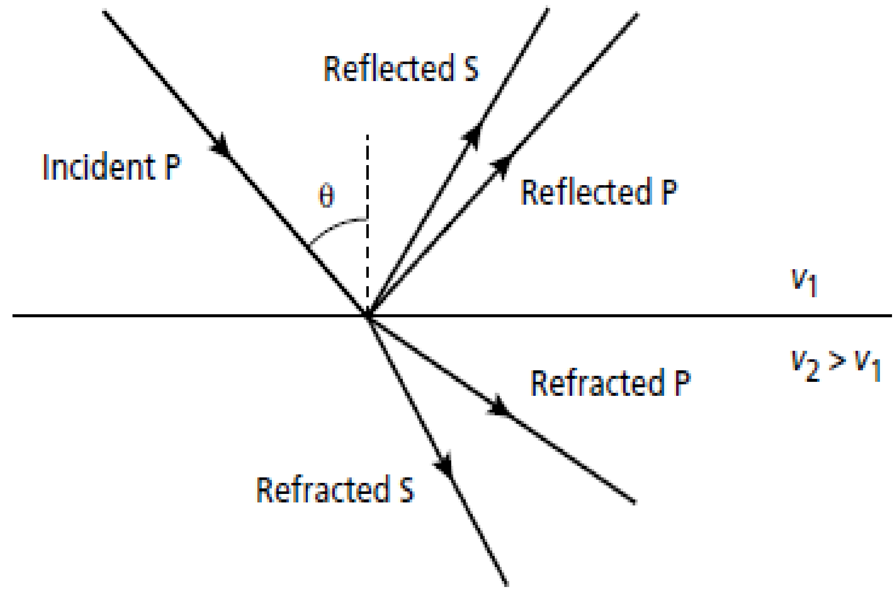


Fig. 3.9 Reflected and refracted P- and S-wave rays generated by a P-wave ray obliquely incident on an interface of acoustic impedance contrast.

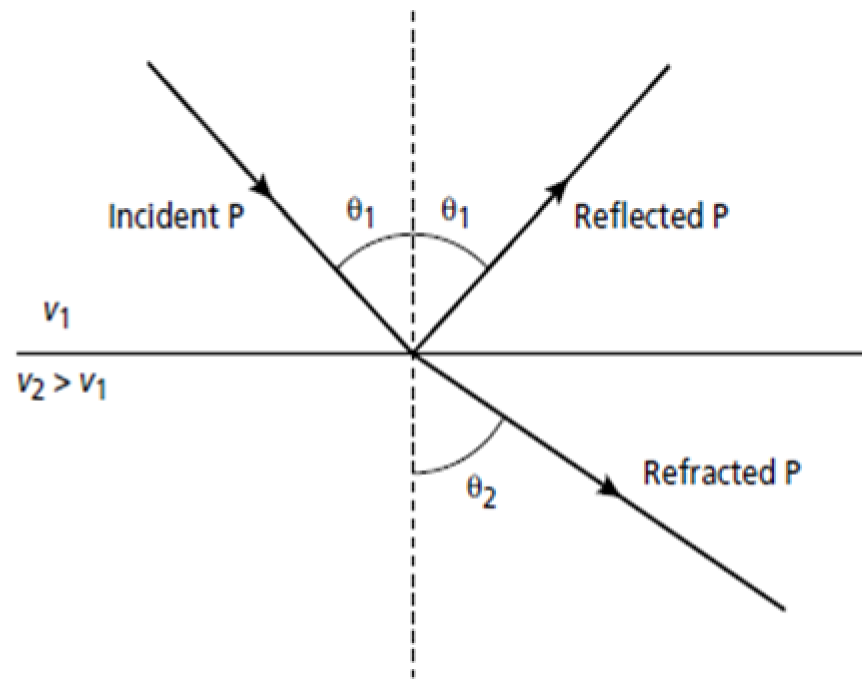
Ray paths in layered media: Snell Law

- Raypaths are refracted according to Snell's Law:

$$\mathbf{Sin\theta_1/V_1 = Sin\theta_2/V_2}$$

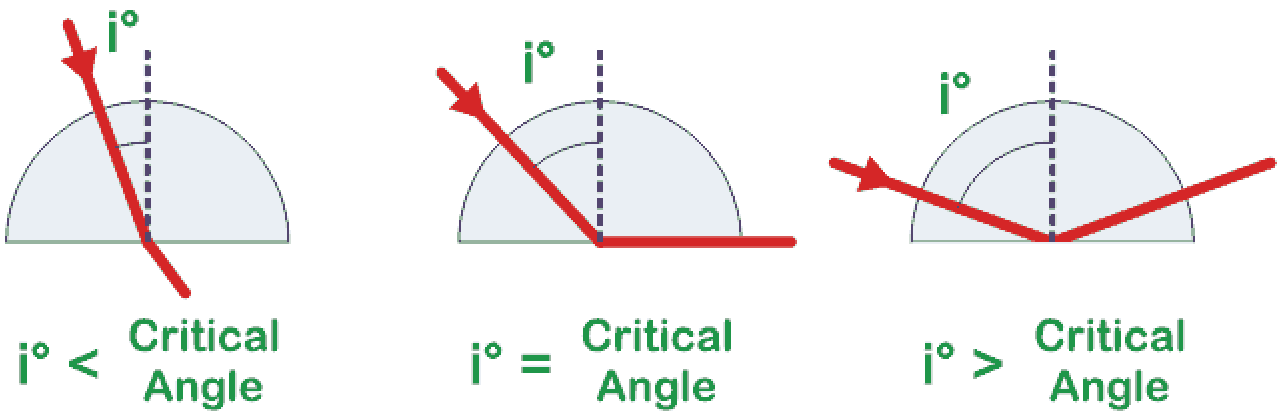
or

$$\mathbf{Sin\theta_1/Sin\theta_2 = V1/V_2}$$



Reflected and refracted P-wave rays associated with a P-wave rays obliquely incident on an interface of acoustic impedance contrast.

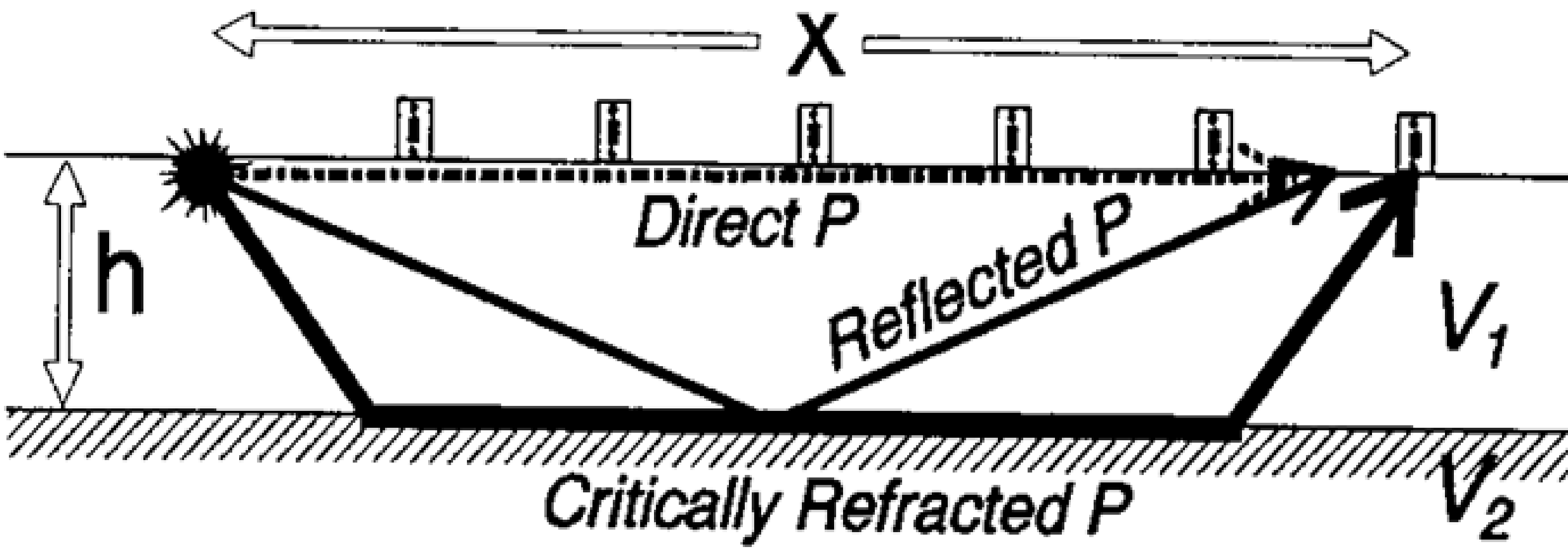
Ray paths in layered media: Critical angle



UNIT THREE

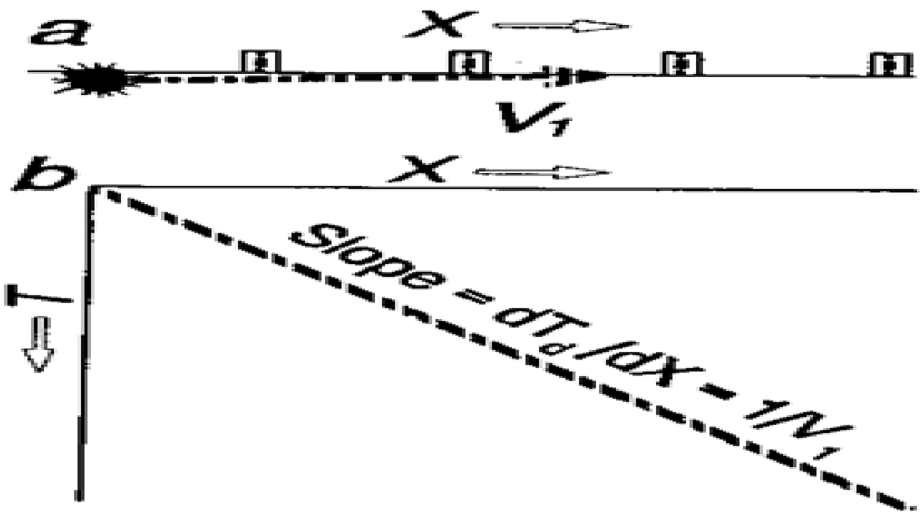
SEISMIC REFRACTION

Direct, Critically Refracted, and Reflected waves



Direct, reflected and refracted ray paths from a near surface source to a surface detector in the case of a simple two-layer model.

The Direct Ray



Selected raypath (a) and travel-time curve (b) for direct wave. The slope, or first derivative, is the reciprocal of the velocity (V_1).

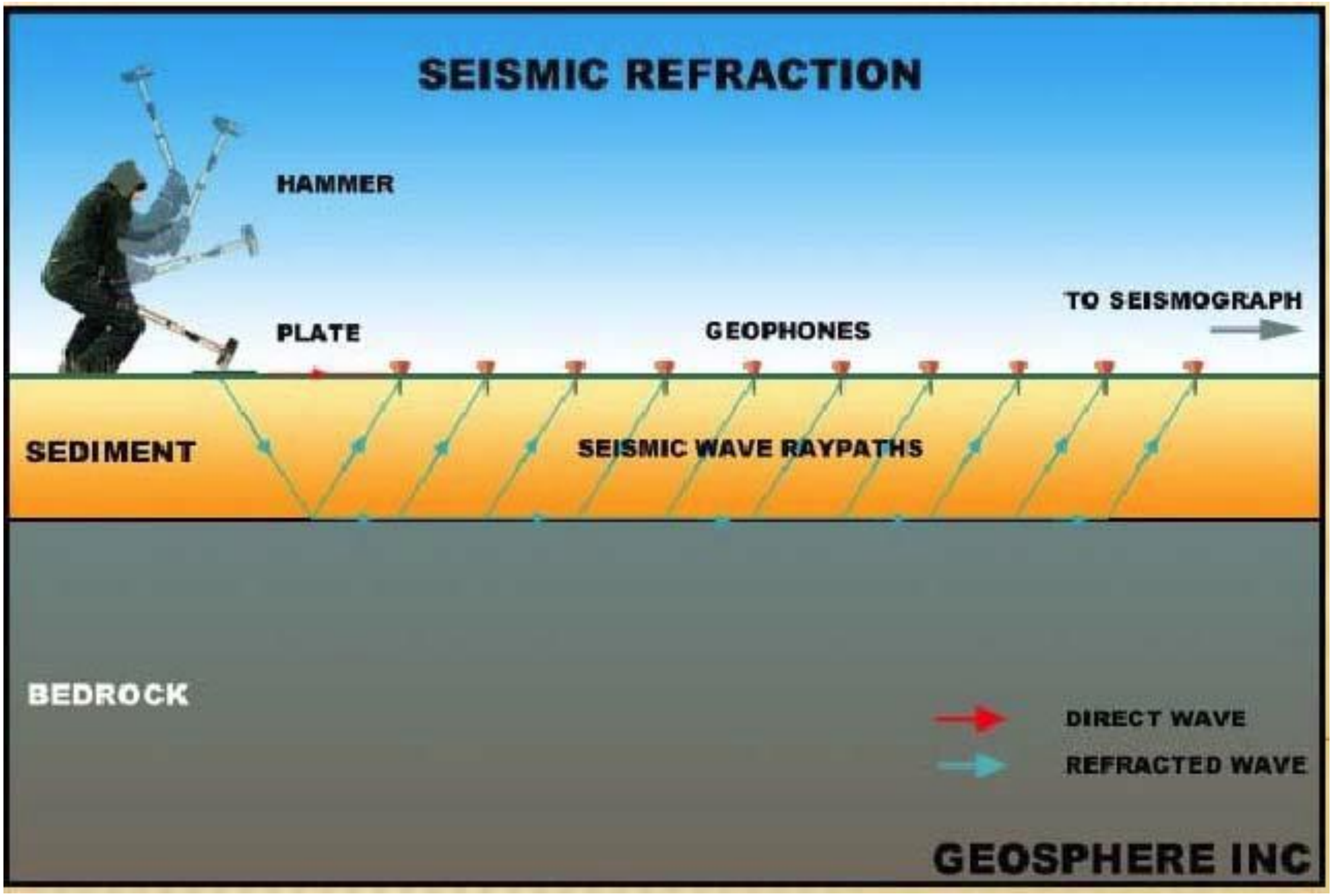
- The travel time of a direct ray is given by:

$$\mathbf{T_d = X/V_1}$$

which defines a straight line of slope = $\mathbf{l/\mathit{V}_1}$ passing through the time–distance origin.

- The velocity $\mathbf{V_1}$ of the wave that goes directly from the source to a receiver is therefore:

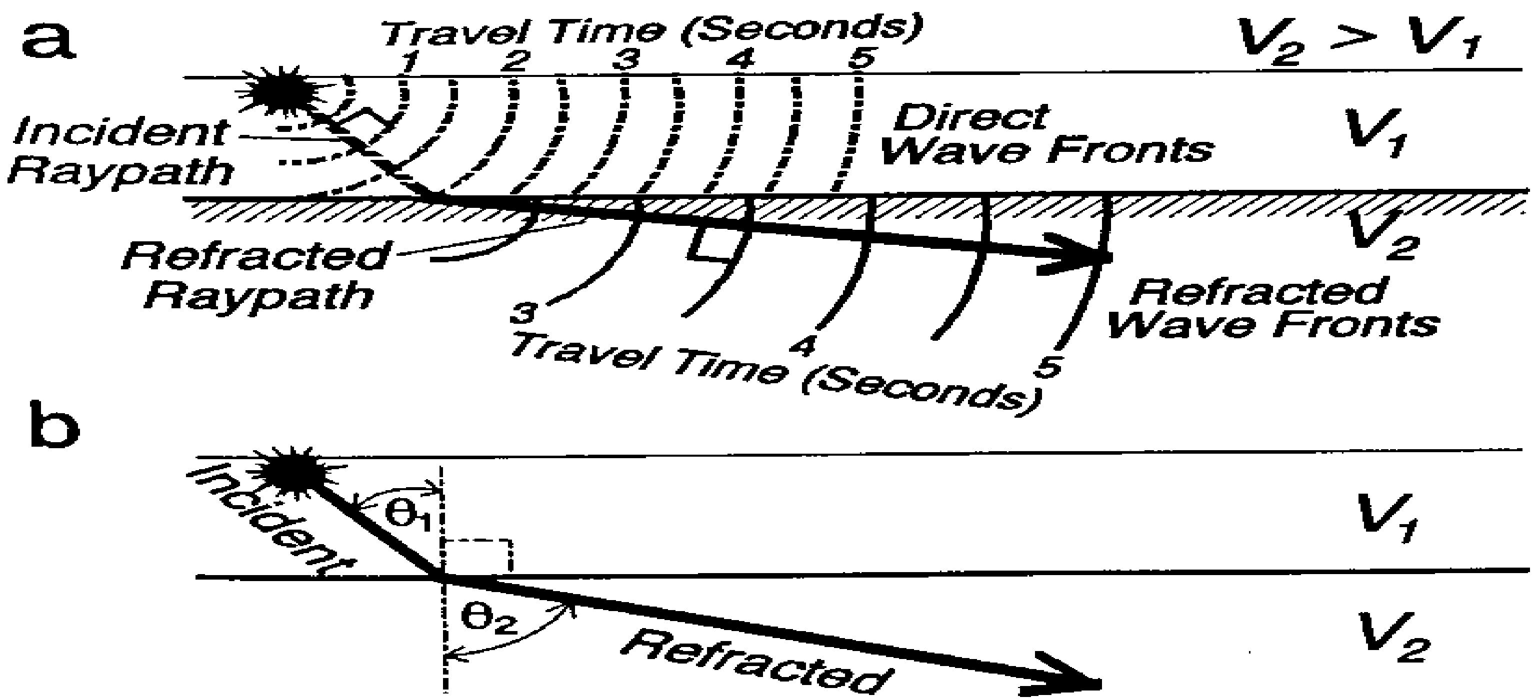
$$\mathbf{V_1 = X/T_d}$$


SEISMIC REFRACTION
HAMMER
PLATE
GEOPHONES
TO SEISMOGRAPH
SEDIMENT
SEISMIC WAVE RAYPATHS
BEDROCK
DIRECT WAVE
REFRACTED WAVE
GEOSPHERE INC

$$\frac{\sin\theta_1}{V_1} = \frac{\sin\theta_2}{V_2}$$

where:

θ_1 = angle of incidence
θ_2 = angle of refraction
V_1 = seismic velocity of incident medium
V_2 = seismic velocity of refracting medium.



- Refraction from a layer of velocity (V_1) to one of velocity (V_2). Note that Ray paths refract across an interface where velocity changes.

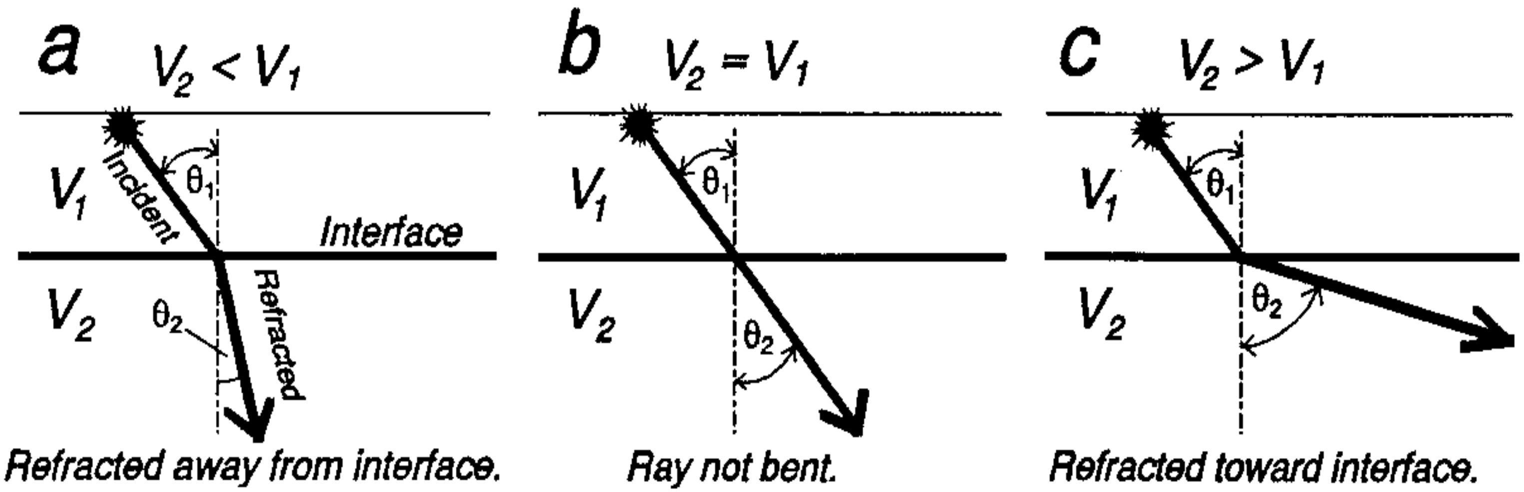


FIGURE 3.23 Behavior of refracted ray when velocity (a) decreases, (b) remains the same, and (c) increases across an interface.

When a ray strikes an interface at an incidence angle θ_1, there will be three situations of refracted waves:

- If the velocity decreases across the interface, the ray is refracted away from the interface.
- If the velocity remains the same, the ray is not bent.
- If the velocity increases across the interface, the ray is bent toward the interface.

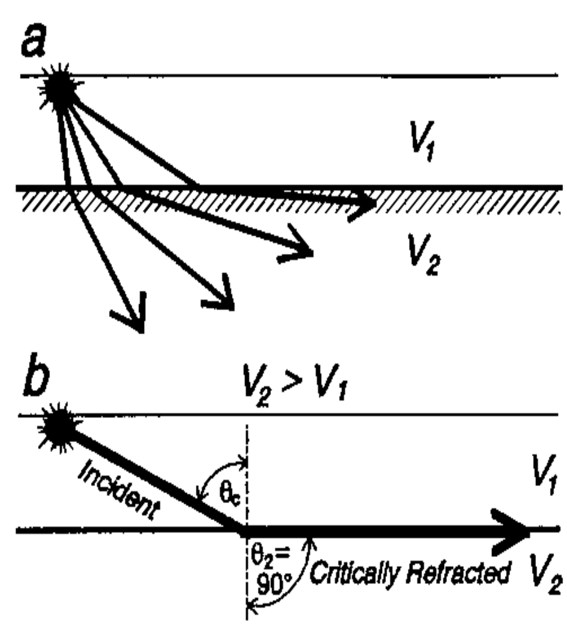
- In case of V2>V1, if the angle of incidence (θ_1) increases, the angle of refraction (θ_2) increases.
- A special situation known as **critical refraction**, occurs *when the angle of refraction* (θ_2) *reaches 90 degrees*.
- The *angle of incidence* (ϑ_1) *that produce a critical refraction is called the critical angle* (θ_c). In this case, the Snell's law shows:

$$\frac{\sin\theta_c}{V_1} = \frac{\sin(90^\circ)}{V_2}$$

$$\frac{\sin\theta_c}{V_1} = \frac{1}{V_2}$$

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

$$\boxed{\theta_c = \sin^{-1}\left(\frac{V_1}{V_2}\right)}$$



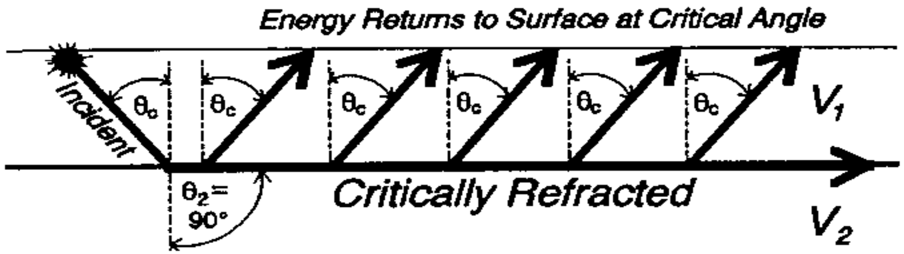


FIGURE 3.25 A critically refracted wave, traveling at the top of the lower layer with velocity V_2, leaks energy back into the upper layer at the critical angle (θ_c).

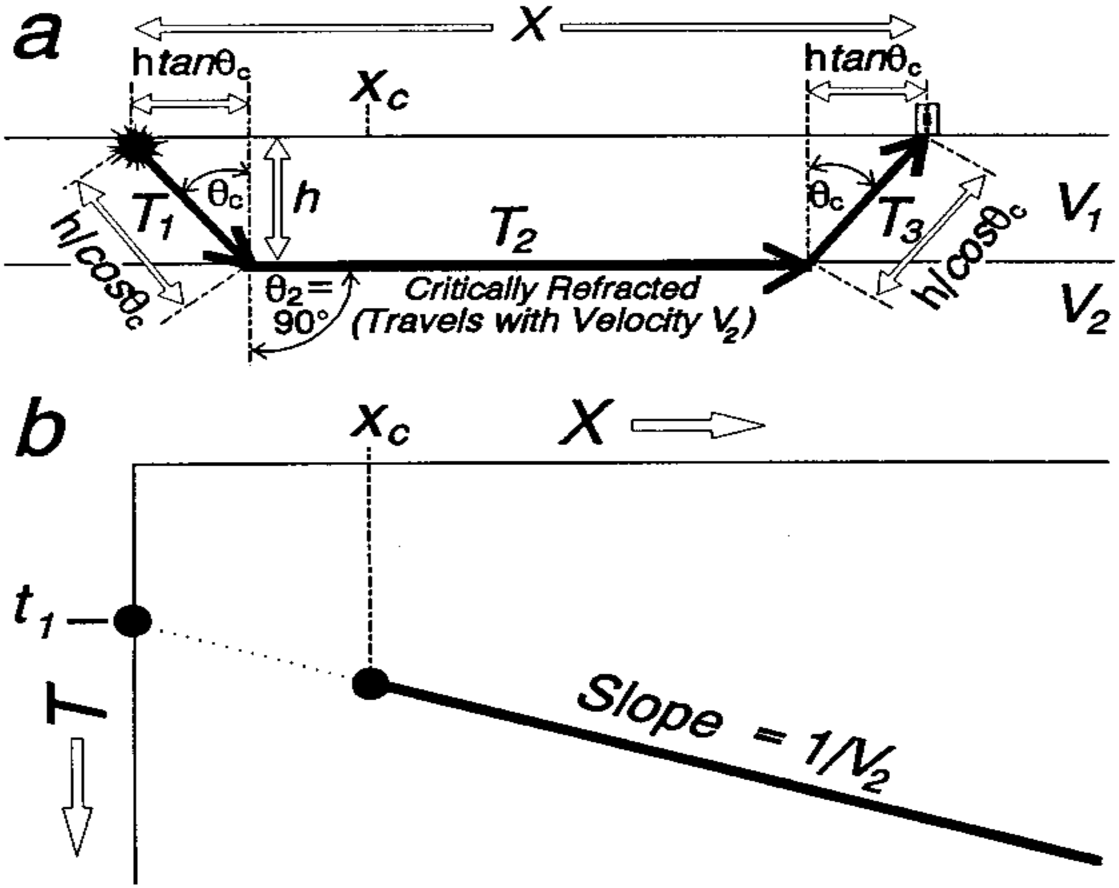
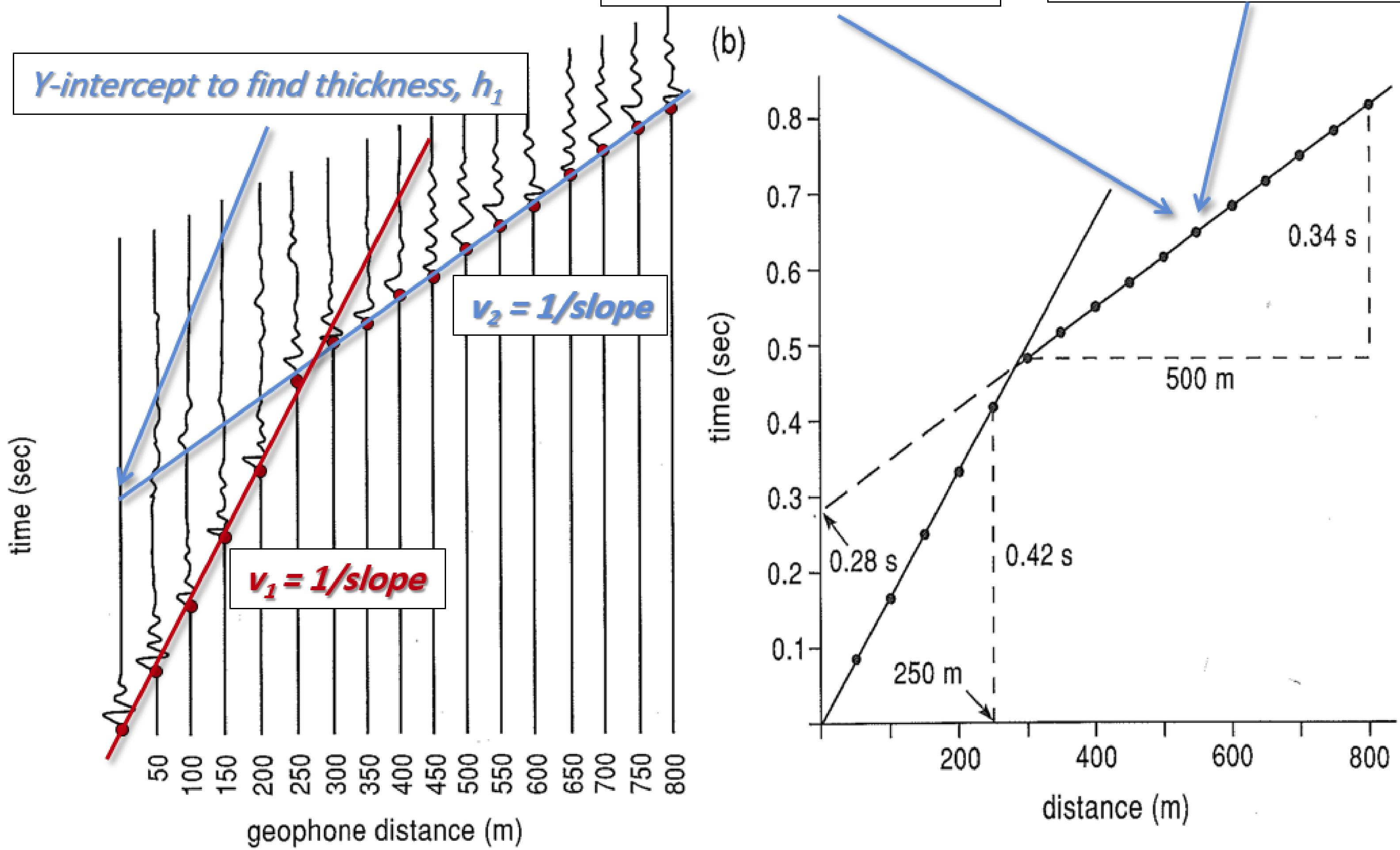


FIGURE 3.26 a) Geometry showing the three segments (T_1, T_2, T_3) comprising the total time path for a critically refracted ray that returns to the surface. b) Travel-time curve for critically refracted wave. The wave arrives at the surface only at and beyond the critical distance (X_c). The intercept time (t_1) is the projection of the curve to the T-axis.

Making a t-x Diagram

Refracted Ray Arrival Time, t $t = \dfrac{x}{v_2} + 2h_1\sqrt{\dfrac{1}{v_1^{\,2}} - \dfrac{1}{v_2^{\,2}}}$ or $t = \dfrac{x \sin i_c}{v_1} + \dfrac{2h_1 \cos i_c}{v_1}$



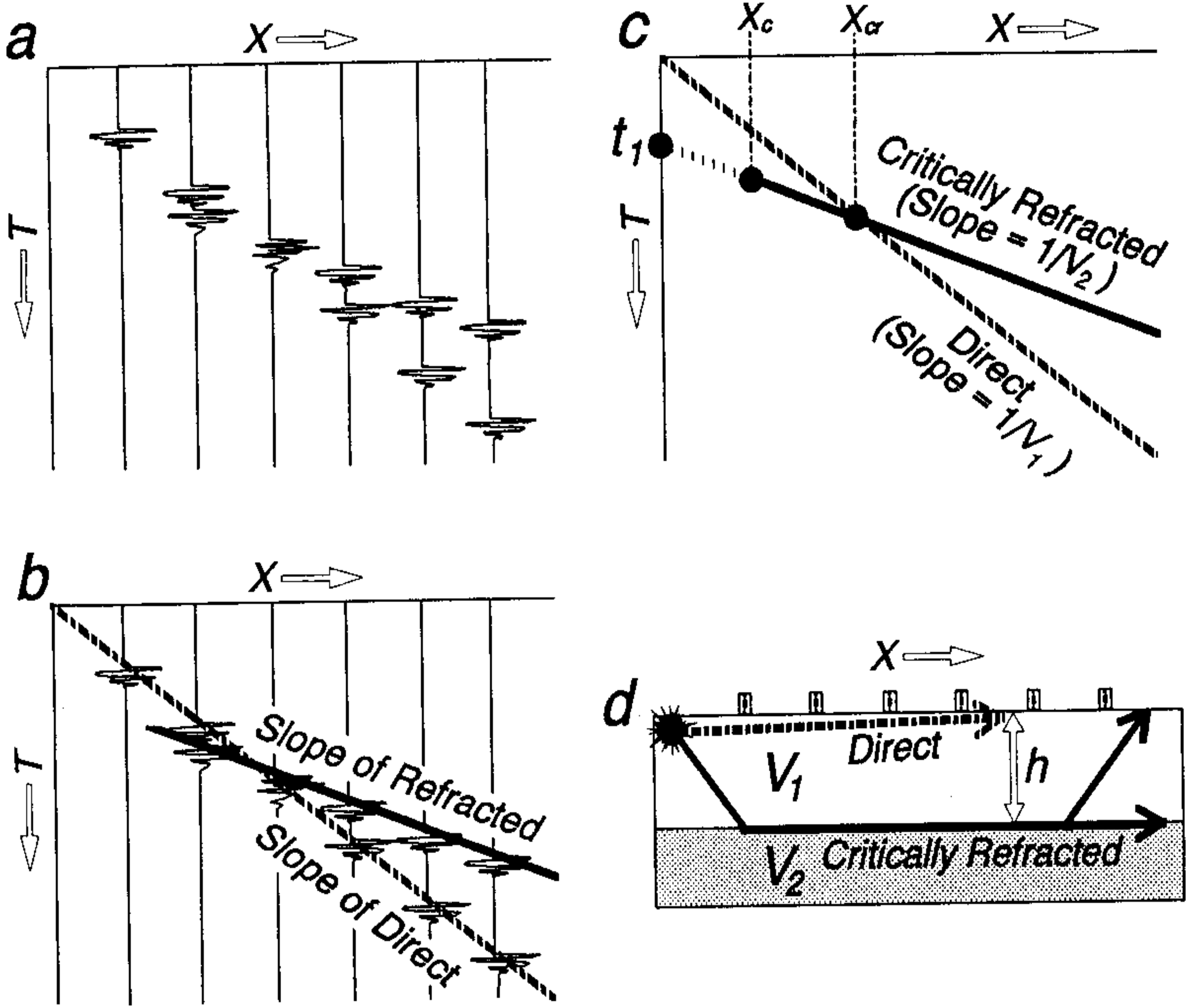


FIGURE 4.3 *Refraction interpretation from travel-time graph.* a) Seismic traces showing events on uninterpreted record. b) Straight lines drawn through events define direct and critically refracted arrivals. c) Events plotted on travel-time graph, with the T-axis intercept (t_1), slopes for direct and critically refracted arrivals, critical distance (X_c), and crossover distance (X_{cr}) identified. d) Simple horizontal interface model used to interpret arrivals.

Single Horizontal Interface

The theory behind a critical refraction from a single, horizontal interface was developed in Chapter 3. The model (Fig. 4.3d) involves an interface at depth (h), separating a lower velocity (V_1) from a higher velocity (V_2). The travel time (T) to a receiver at horizontal distance (X) from the source is:

$$\boxed{\mathbf{T = t_1 + \frac{X}{V_2}}}$$

where:

$$t_1 = \text{T-axis intercept} = \frac{2h\,cos\,\theta_c}{V_1}$$

$$\theta_c = \text{critical angle} = sin^{-1}\left(\frac{V_1}{V_2}\right)$$

$$X_c = \text{critical distance} = 2h\,tan\,\theta_c$$

$$X_{cr} = \text{crossover distance} = 2h\sqrt{\frac{V_2 + V_1}{V_2 - V_1}}$$

The above equations can be *forward modeling equations*; when applied to a hypothetical model (Fig. 4.3d), they yield a predicted travel-time graph (Fig. 4.3c).

Inversion, on the other hand, can be used to interpret the velocity structure from an observed refraction profile (Fig. 4.3a). The intercept time (t_1) and the slopes of the direct and refracted arrivals are read directly from the travel-time plot (Fig. 4.3b,c). The observed slopes and intercept time can then be solved for the true velocities (V_1, V_2) and the depth to the interface (h), using the following *inversion equations*:

$$\text{Slope of Direct} = \frac{1}{V_1} \Rightarrow \mathbf{V_1 = \frac{1}{\textbf{slope of direct}}}$$

$$\text{Slope of Refracted} = \frac{1}{V_2} \Rightarrow \mathbf{V_2 = \frac{1}{\textbf{slope of refracted}}}$$

$$\theta_c = sin^{-1}\left(\frac{V_1}{V_2}\right)$$

$$t_1 = \frac{2h\,cos\,\theta_c}{V_1} \Rightarrow \mathbf{h = \frac{t_1 V_1}{2\cos\theta_c}}$$

Fig. 4.3d, in this case, represents the inversion model that results from the observed refraction profile (Fig. 4.3a).

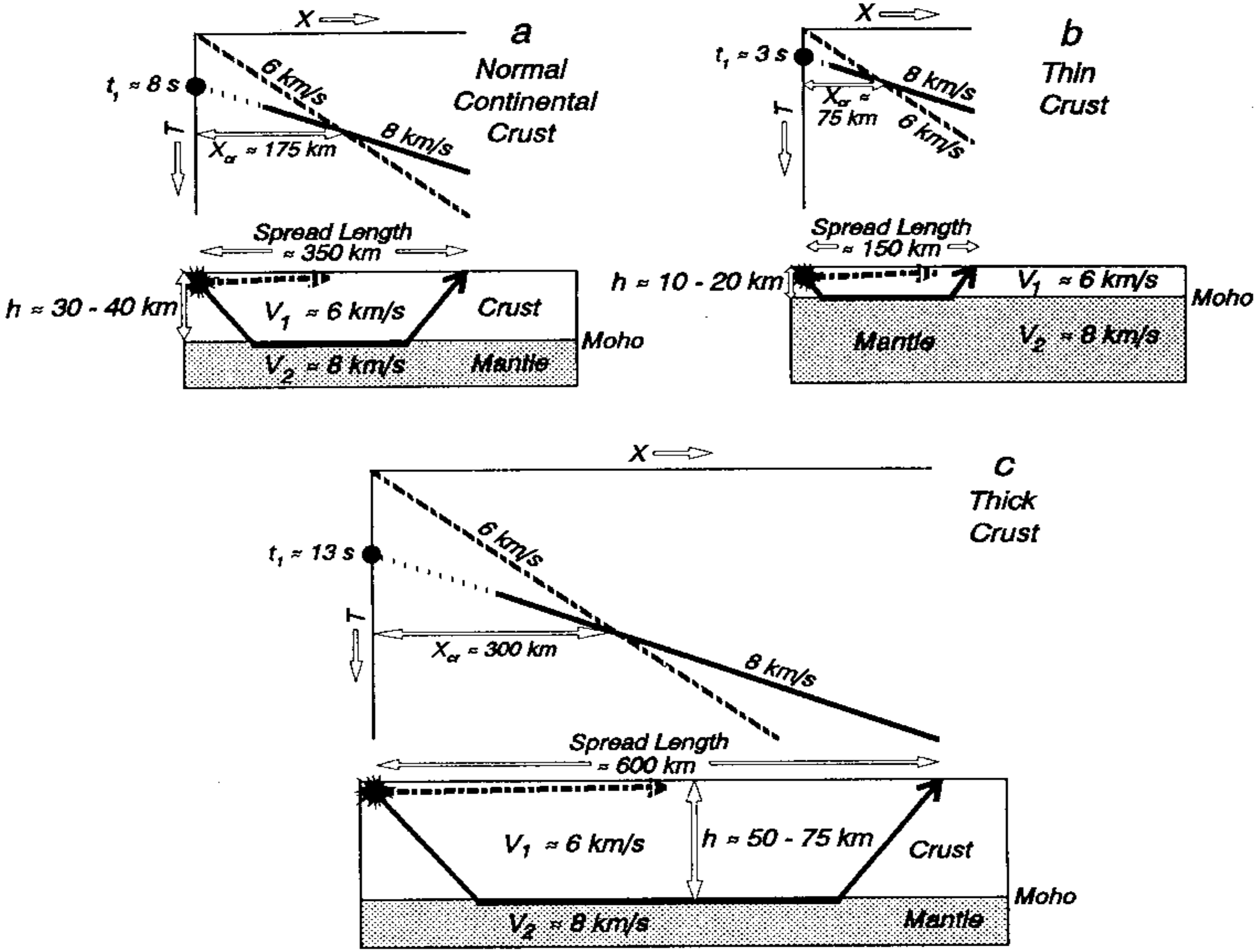


FIGURE 4.4 Comparisons of intercept times (t_1) and crossover distances (X_{cr}) for different crustal thickness. The grossly simplified models illustrate the approximate spread lengths ($2X_{cr}$) necessary to resolve the depth to Moho (h). The travel-time graphs were determined using forward modeling equations presented in text. Inversion equations can be used to interpret crustal thickness if the T-axis intercept (t_1) and apparent velocities are read from observed refraction profiles. a) The distance from the source to the farthest receiver must be about 350 km to resolve the crustal thickness in regions of typical continental crust; the T-axis intercept is about 8 s. b) Oceans and regions of very thin continental crust require about 150 km spread lengths, where a shorter T-intercept of about 3 s might be expected. c) Very deep Moho beneath some mountain ranges necessitates very long spread lengths (≈ 600 km), and results in large T-axis intercept times (≈ 13 s).

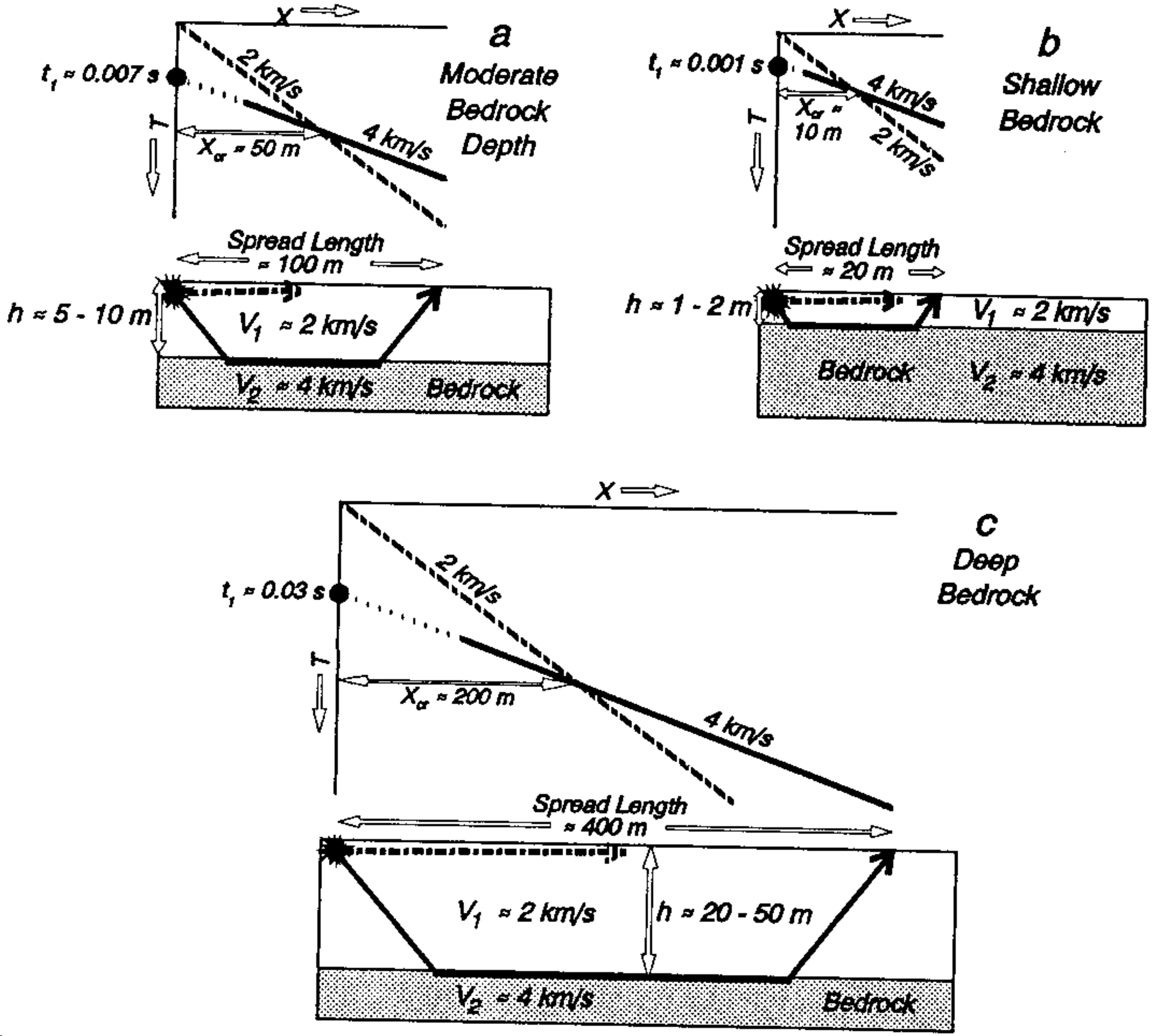
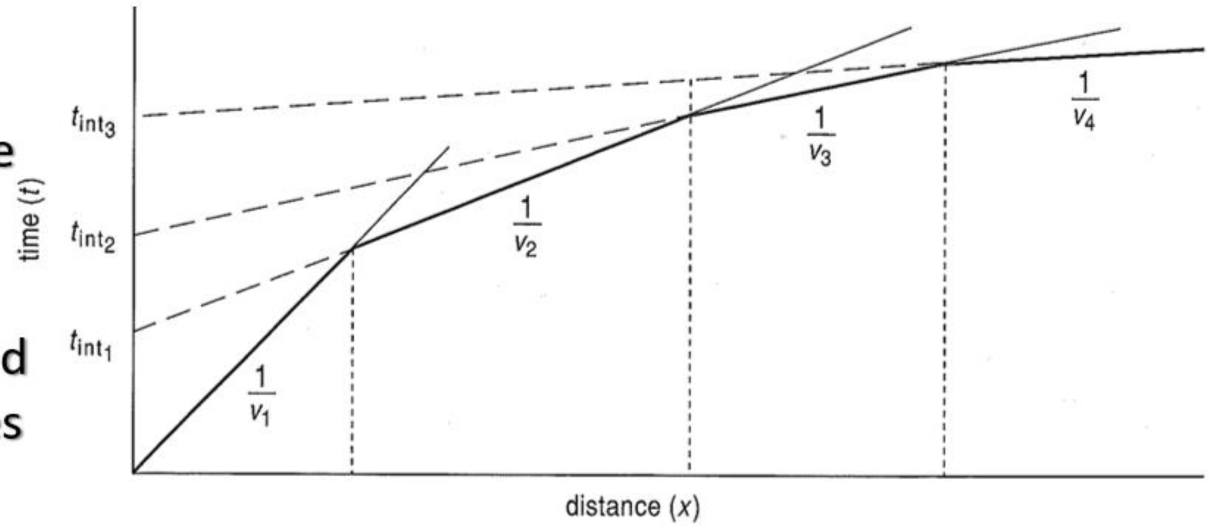
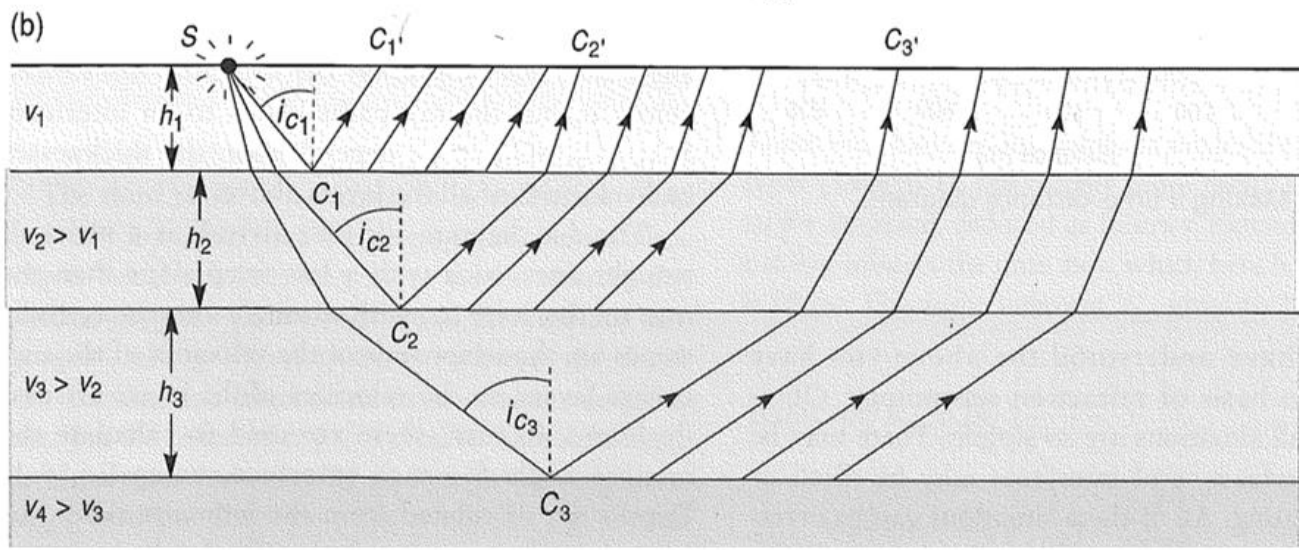


FIGURE 4.5 Approximate T-axis intercept times (t_1), crossover distances (X_{cr}), and required spreadlengths for bedrock depths (h) that are (a) moderate; (b) shallow; and (c) deep. Travel-time graphs were determined using forward modeling equations presented in text.

Multiple Layers

- Seismic refraction can detect multiple layers
- The velocities are easily found from the slopes on the t-x diagram





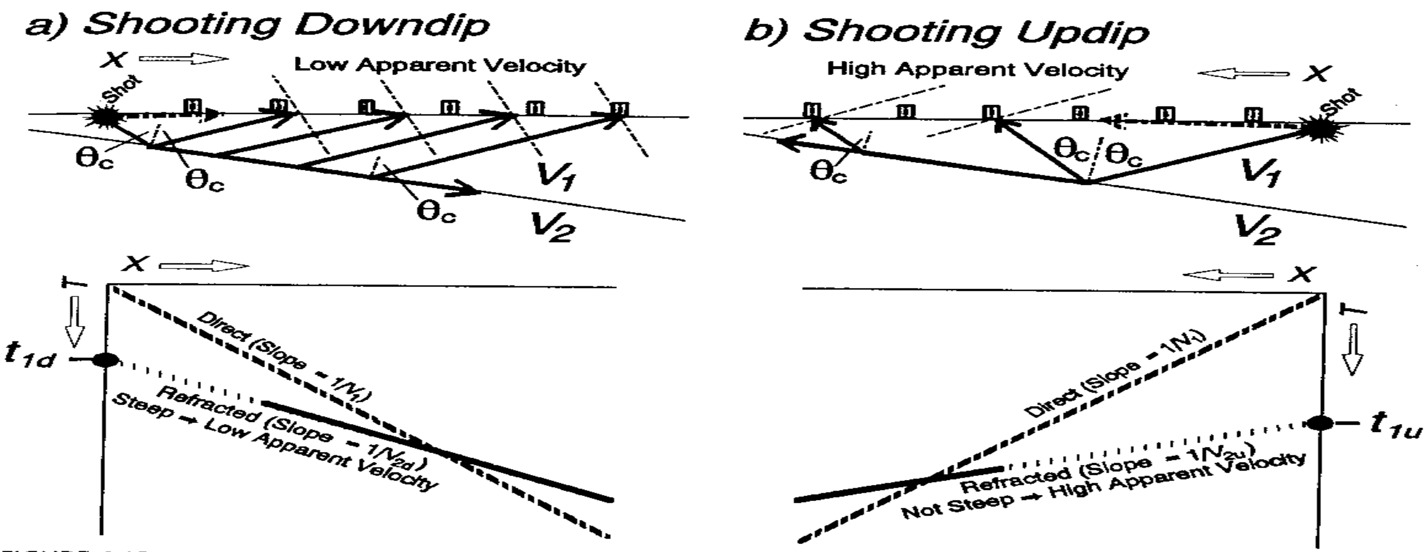


FIGURE 4.13 *Seismic survey of a dipping refractor.* a) Selected raypaths shooting toward receivers in a downdip direction. Raypaths for the critical refraction emerge at a shallow angle, resulting in an apparent velocity (V_{2d}) lower than the true velocity (V_2). b) When the receivers are updip from the source, rays emerge at a steeper angle; an apparent velocity (V_{2u}) higher than V_2 results.

6 km/s refraction might be from crystalline basement. The higher velocity arrival (8 km/s; probably from the Moho) stands out because of its negative slope.

Single Dipping Interface

For a dipping interface (Fig. 4.13), apparent velocities observed at the surface are not equal to the true velocity of the refracting layer. When the source shoots downdip toward the receivers, the apparent velocity is lower than the true velocity (Fig. 4.13a); a velocity higher than the true velocity results from shooting updip (Fig. 4.13b).

The dipping interface can be resolved by recording a *reversed refraction profile*. A profile is shot in one direction (as from a shotpoint at A to receivers extending to B), then in the other direction (from B to A; Fig. 4.14a). The seismic travel-time records (Fig. 4.14b) are superimposed with the same horizontal and vertical scales, then analyzed according to the equations presented below (see also Burger, 1992, p. 80–85; Telford et al., 1976, p. 281–284).

For a dipping interface, the *intercept times* shooting in the downdip and updip directions are not equal:

$$t_{1d} \neq t_{1u}$$

where:

t_{1d} = T-axis intercept when shooting downdip (from A to B)
t_{1u} = T-axis intercept when shooting updip (from B to A).

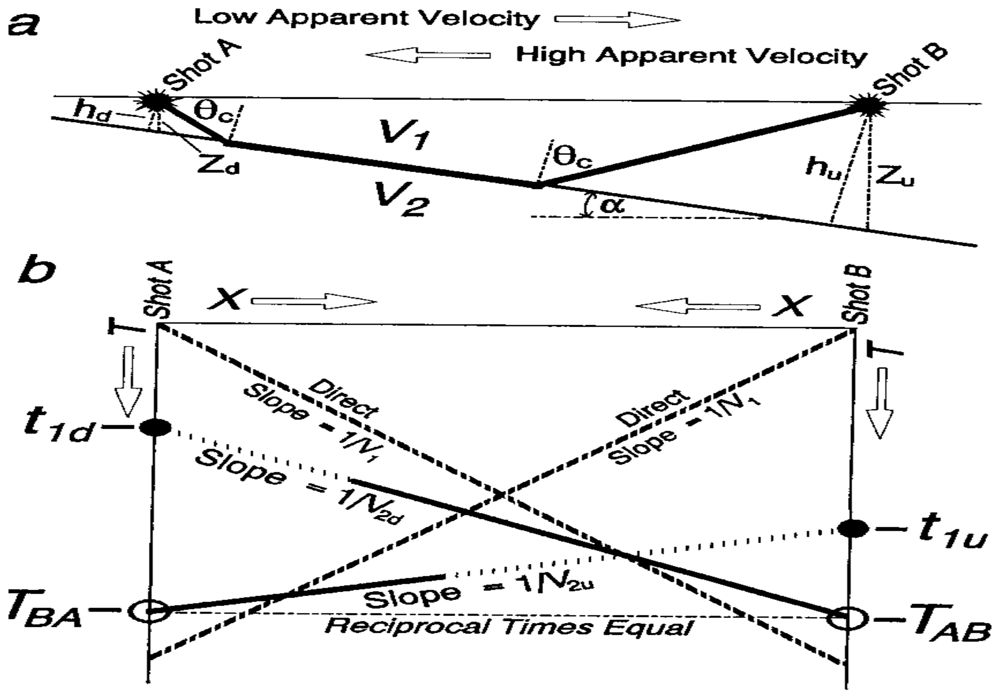


FIGURE 4.14 A reversed refraction profile is the combination of surveys shooting in downdip and updip directions (Fig. 4.13). a) A ray from the source at A, shooting to a receiver at B, traverses the same course as a ray from the shot at B shooting to a receiver at A. The reciprocal times T_{AB} and T_{BA} in (b) are therefore exactly the same. Apparent velocities V_{2d} and V_{2u} in (b) are different, because the rays emerge at different angles. b) Superposition of travel-time curves for the downdip (Fig. 4.13a) and updip (Fig. 4.13b) surveys. Note that, because the interface is deeper beneath Shot B, the intercept time (t_{1u}) for updip shooting is greater than when shooting downdip (t_{1d}).

The travel time from the shot at A to a receiver at B, however, has to be the same as the travel time from the shot at B to a receiver at A, because the exact raypath is utilized. Thus the *reciprocal times* must be equal:

$$T_{AB} = T_{BA}$$

where:

T_{AB} = travel time from shot at A to receiver at B
T_{BA} = travel time from shot at B to receiver at A.

In picking events on a reversed refraction plot (Fig. 4.14b), one can verify if refractions are from the same interface by determining if $T_{AB} = T_{BA}$. If reciprocal times are not the same, the analysis below will be erroneous.

The *apparent velocities* for the refracted arrival when shooting in the downdip and updip directions are:

$$V_{2d} = \frac{V_1}{\sin(\theta_c + \alpha)}$$

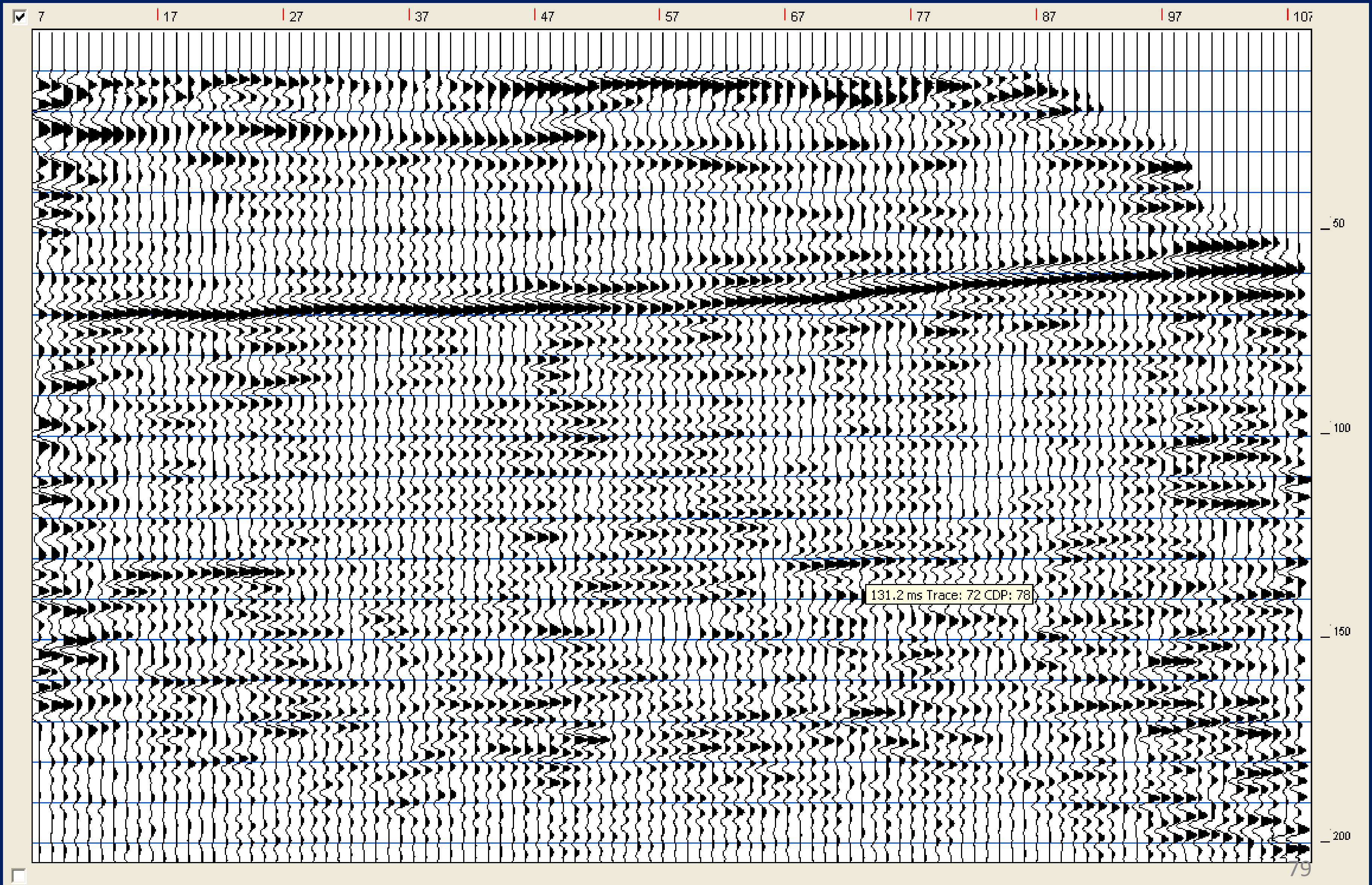
UNIT FOUR

SEISMIC REFLECTION

Seismic Reflection

- Things to know before we start…
  - Seismic reflection is the single most important technique for seeing into the Earth.
    - It is useful for shallow and deep depths
    - Massively used by the oil and gas industry
    - It can detect:
      - Stratigraphy
      - Faults
      - Folds
      - Oil & Gas Reservoirs
      - Groundwater Resources
  - Why so popular?
    - Produces results that actually look a lot like an actual geologic cross section!!

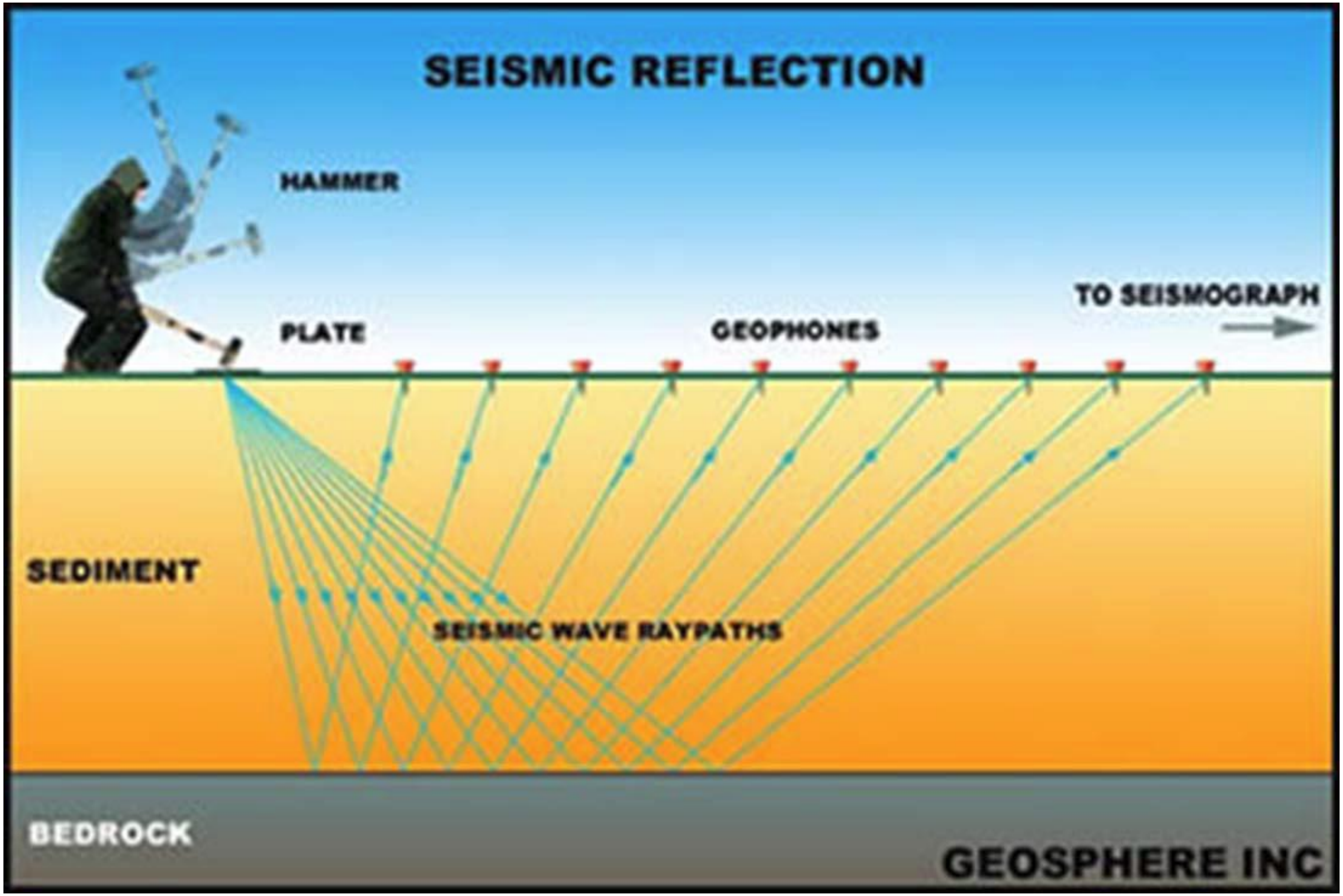
Example of seismic section

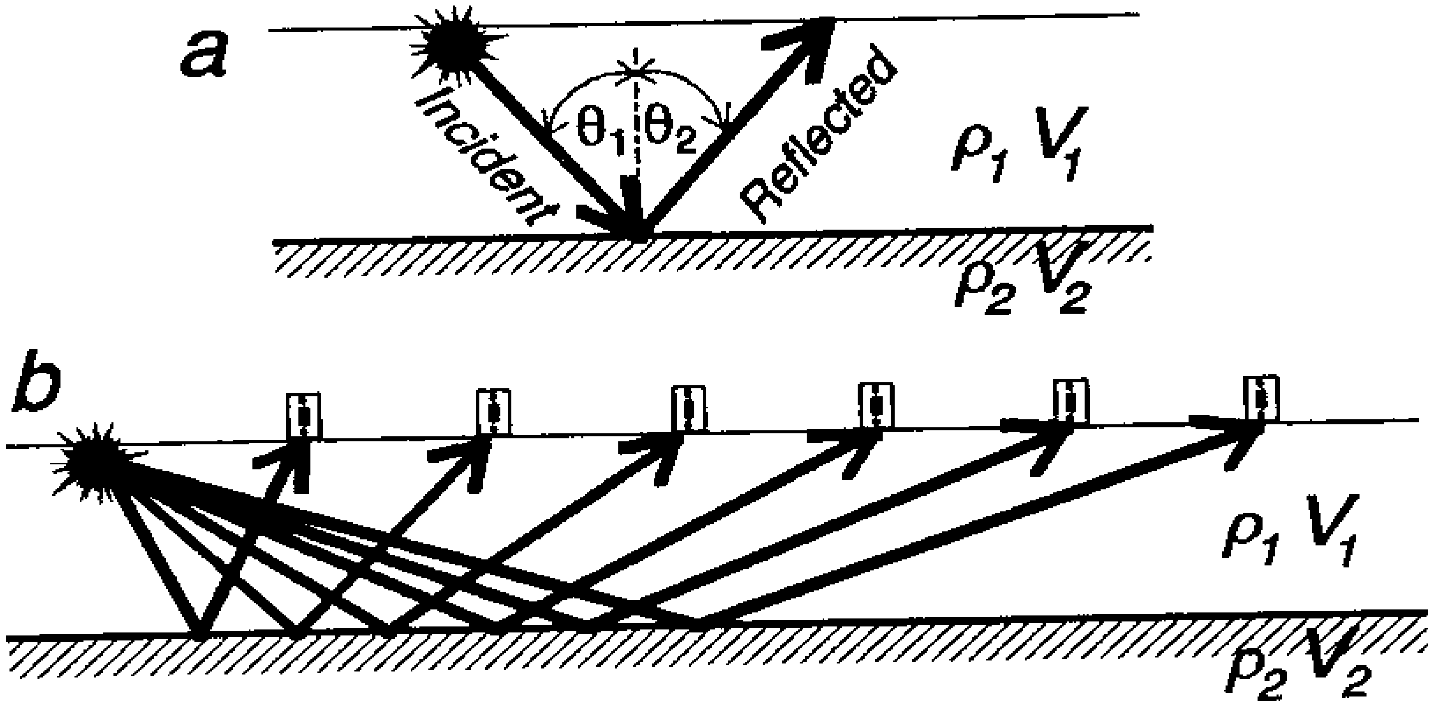


- In seismic reflection surveys, seismic energy pulses are **reflected** from subsurface interfaces and **recorded at near-normal incidence** at the surface.

- The travel times are measured and can be converted into estimates of depths to the interfaces.

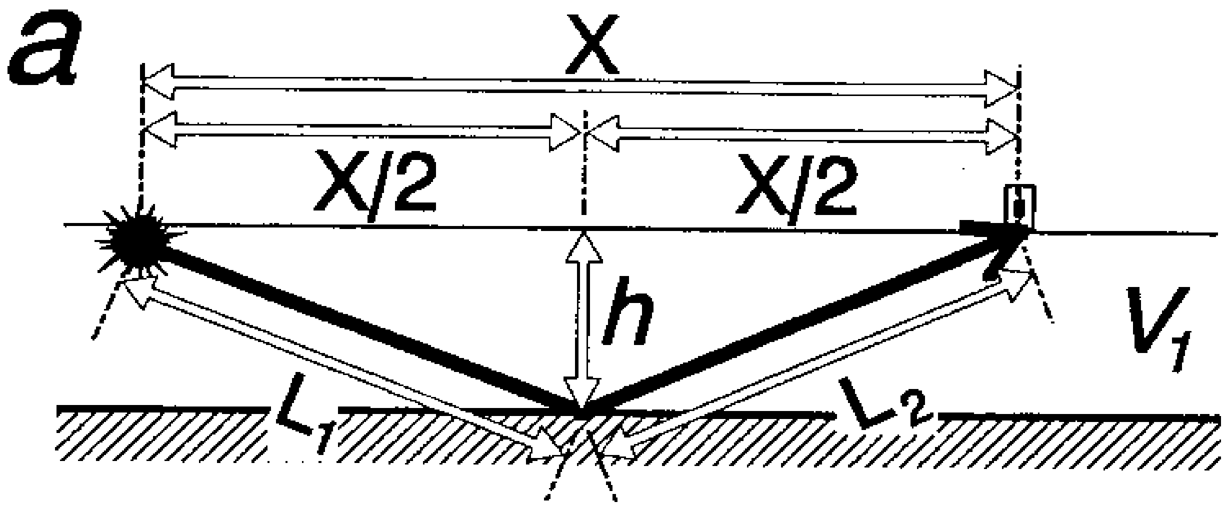
Seismic reflection





- A compressional wave is reflected back at an angle (θ_2) equal to the incident angle (θ_1). Note the V-shaped raypaths from source to receivers.

- Reflection occurs when the ***acoustic impedance*** of the lower layer ($\rho_2 V_2$) differs from that of the upper layer ($\rho_1 V_1$).

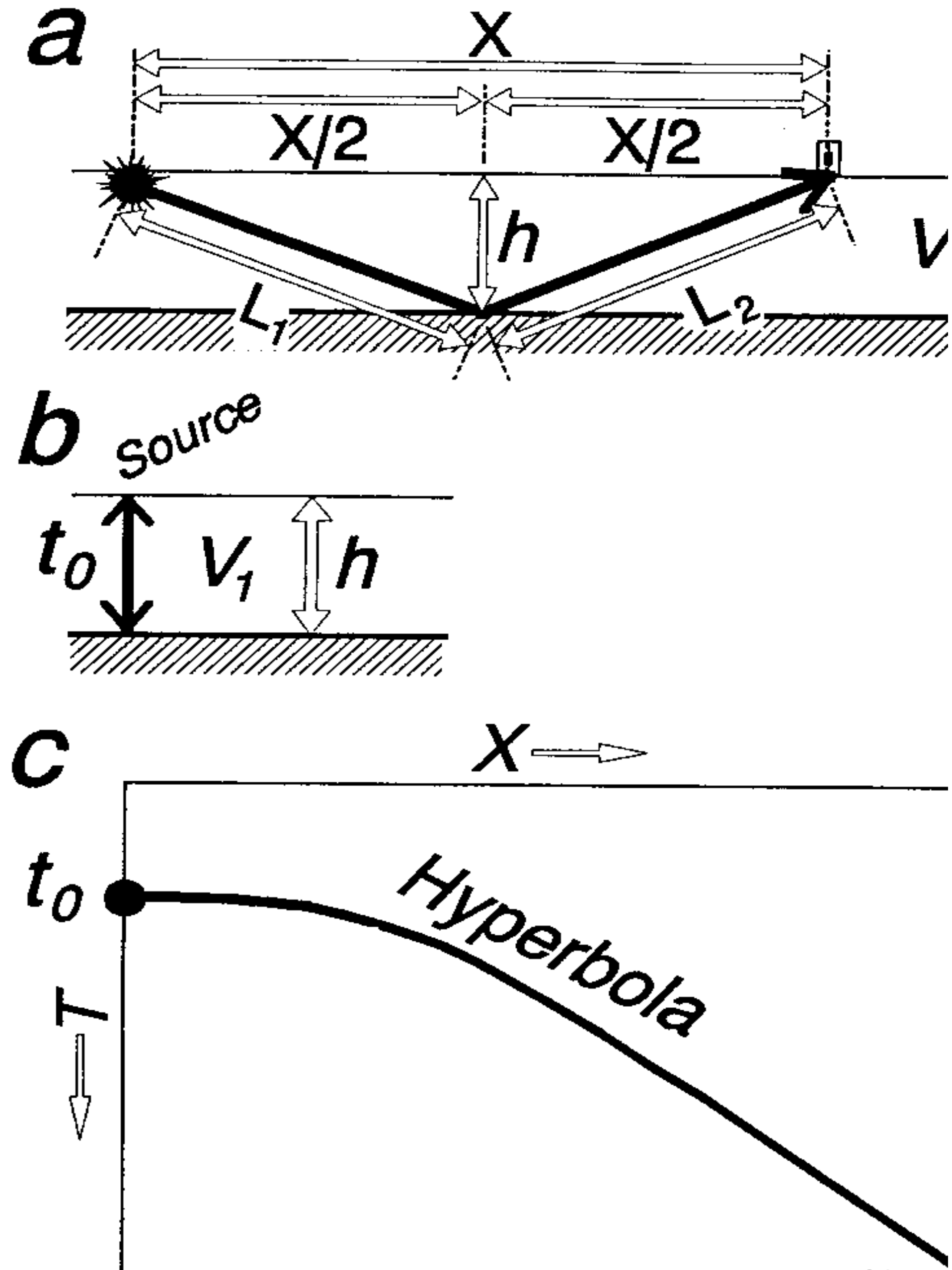


$$
\begin{aligned}
L &= L_1 + L_2 \\
&= \sqrt{h^2 + (X/2)^2} + \sqrt{h^2 + (X/2)^2} \\
&= 2\sqrt{h^2 + (X/2)^2} \\
&= \sqrt{4h^2 + X^2}
\end{aligned}
$$

The total travel time (T_f) from source to receiver is:

$$
\begin{aligned}
T_f &= L/V_1 \\
&= \frac{\sqrt{4h^2 + X^2}}{V_1} \\
T_f{}^2 &= \frac{4h^2 + X^2}{V_1{}^2} \\
&= \frac{4h^2}{V_1{}^2} + \frac{X^2}{V_1{}^2} \\
T_f{}^2 &= \left(\frac{2h}{V_1}\right)^2 + \left(\frac{1}{V_1}\right)^2 X^2
\end{aligned}
$$

$$T^2 = t_o^2 + \frac{x^2}{v^2}$$



- This is an equation of hyperbola, with $\mathbf{t_0}$: T-axis intercept time

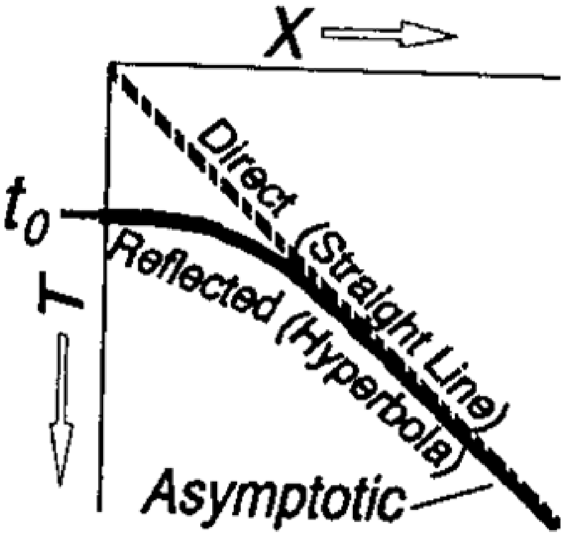
- $\mathbf{t_0}$: is the travel time vertically down to the interface and back up to the source:

  $$\mathbf{t_0 = 2h/V_1}$$

- **For long distance from the source**, as the offset X becomes very large, t_0 becomes insignificant. and therefore,

$$T^2 = t_o^2 + \frac{x^2}{v^2} \approx \frac{x^2}{v^2}$$

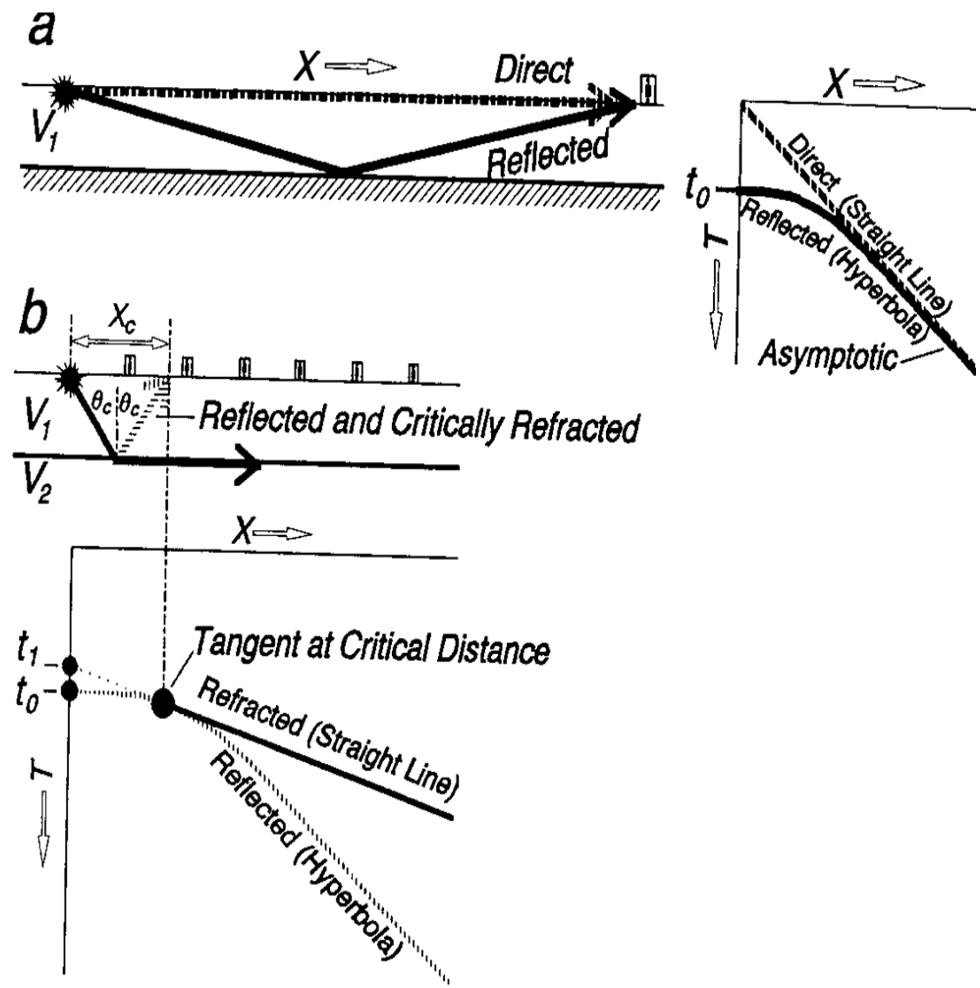
$T_f = X/V1 = T_d$

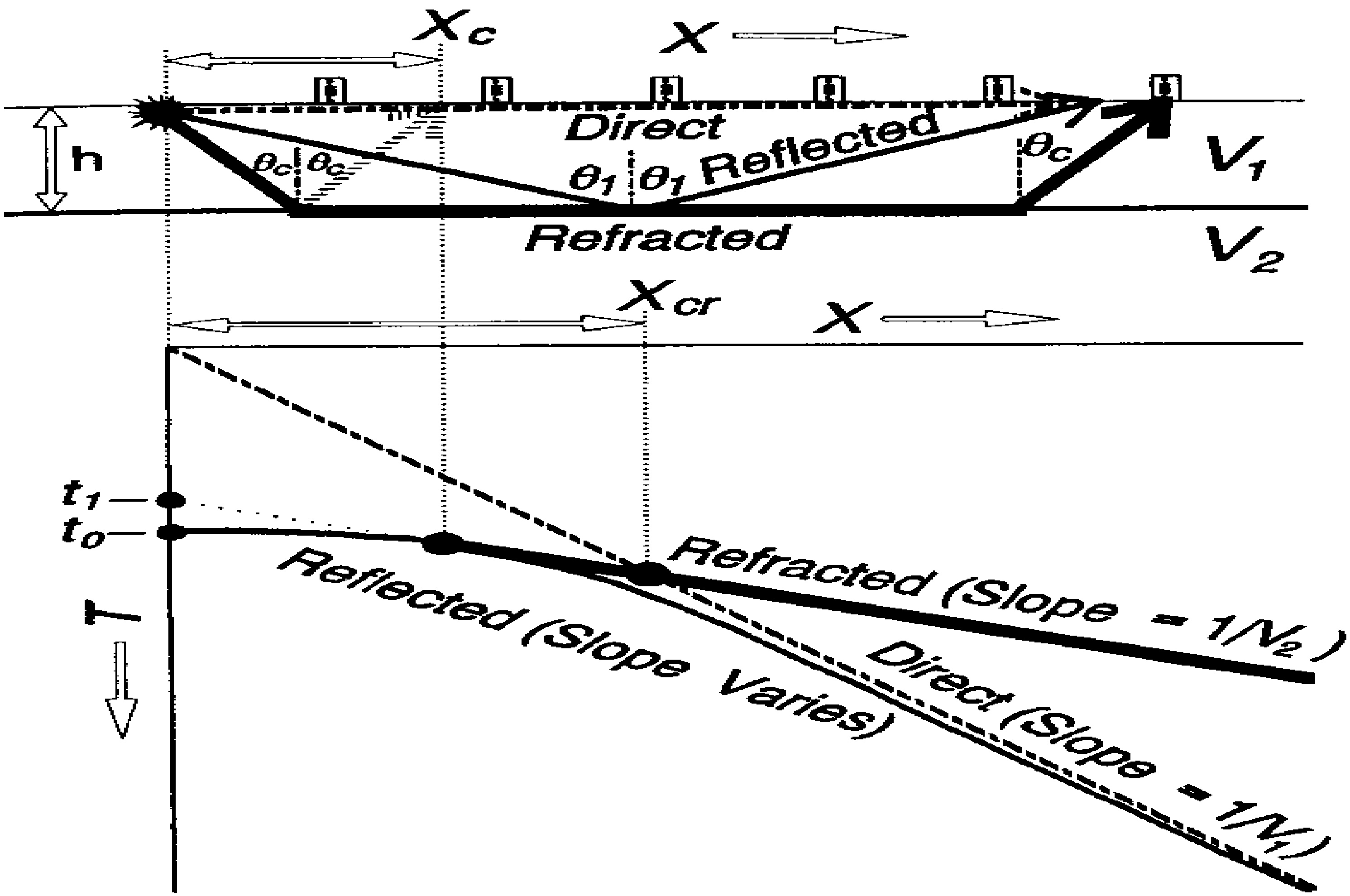


- In this case, the **travel time curve** (T_f) for reflections **recorded at large distances** is therefore approximately the **same as for the direct wave**. The reflected wave is asymptotic to the direct wave.

Seismic reflection

- For ***critical distance* (X$_c$)**, *the reflected and refracted wave have the same arrival time*.
- The straight line for the refracted wave is **tangent** to the hyperbola of the reflection.

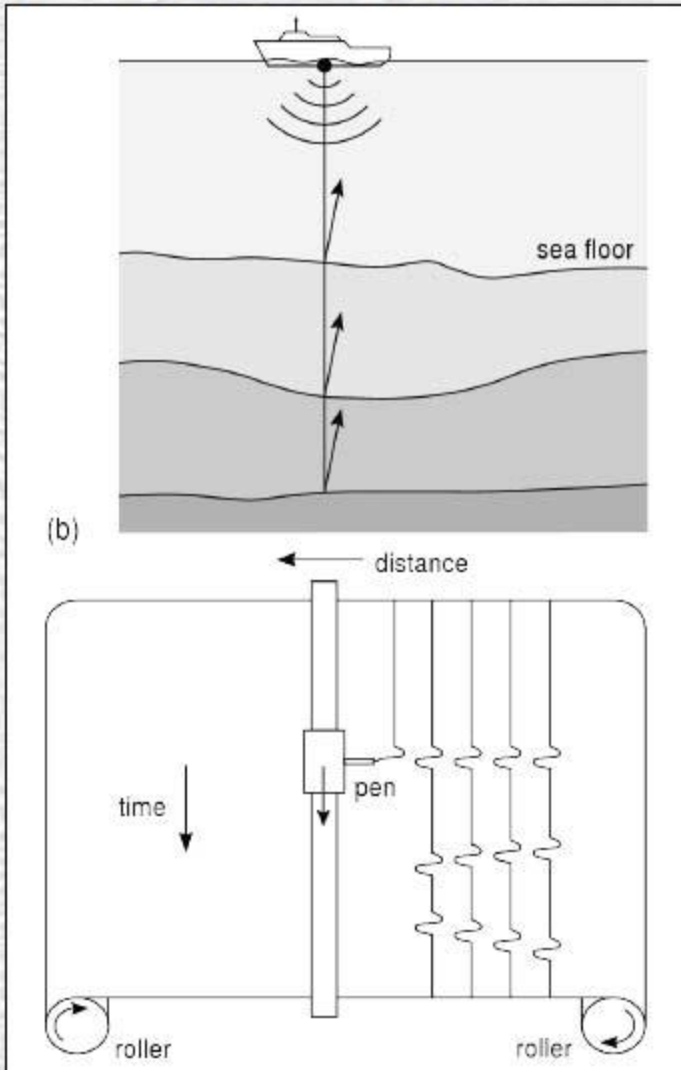




X_c
X
Direct
Reflected
h
θ_c θ_c
θ_1 θ_1
θ_c
V_1
Refracted
V_2
X_{cr}
X
t_1
t_0
T
Refracted (Slope = $1/V_2$)
Reflected (Slope Varies)
Direct (Slope = $1/V_1$)

Seismic Reflection :: The Basics

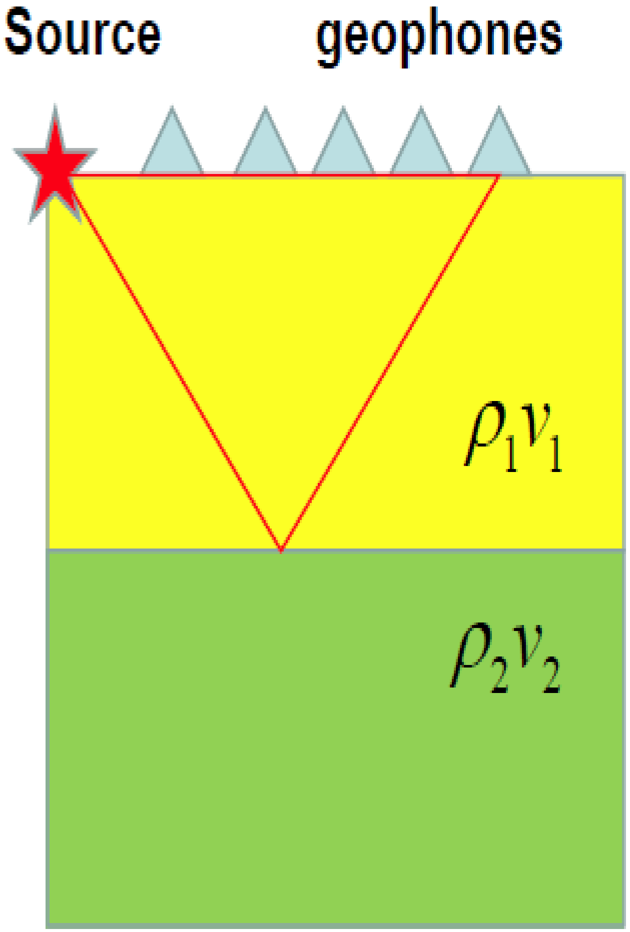
- In the simplest sense seismic reflection is echo sounding.
  - Echoes come from layers in the Earth, not fish or the sea floor
- E.g. a ship sends out a seismic pulse
  - The pulse is reflected back to a receiver on the ship's bottom after some time has passed
  - The various arrivals can be used to map out subsurface "reflectors" or layers



Seismic Reflection

Fundamental considerations

- Reflection coefficient.
- Transmission coefficient
- Acoustic impedance.
- Zoeppritz equations.
- Negative polarity reflection
- Two-Way Time (TWT)



Reflection coefficient $$R = \frac{\rho_2 v_2 - \rho_1 v_1}{\rho_1 v_1 + \rho_2 v_2}$$

A typical value for R is 0.001

Reflectors reflect contrasts of acoustic impedance: $\rho \cdot v_p$

Polarity of reflected wave depends on sign of reflection coefficient

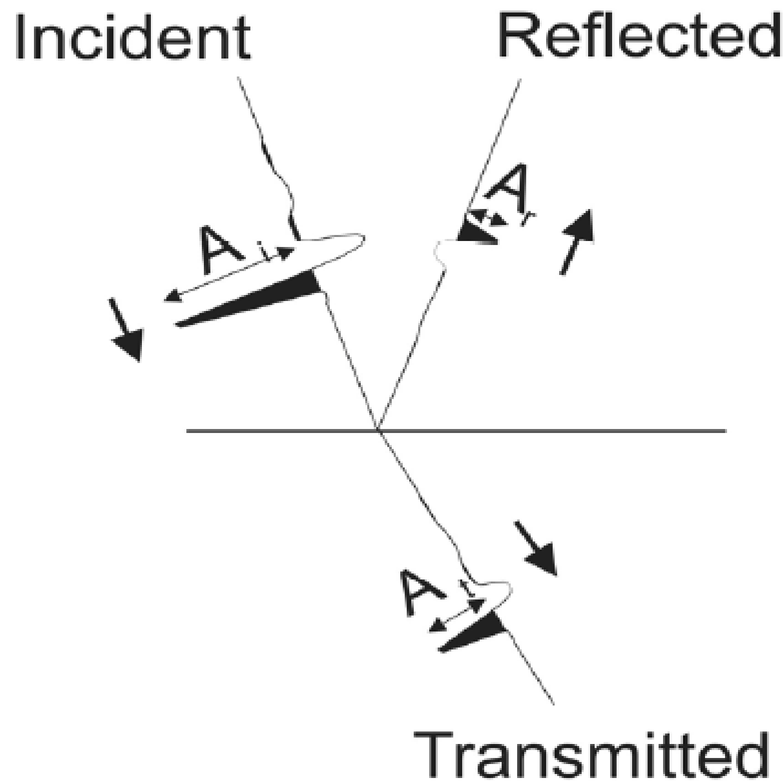
For small angle of incidence

$$R = \frac{A_r}{A_i}$$

$$R = \frac{v_2\rho_2 - v_1\rho_1}{v_2\rho_2 + v_1\rho_1}$$

$$T = \frac{A_t}{A_i}$$

$$T = \frac{2v_1\rho_1}{v_2\rho_2 + v_1\rho_1}$$



R is the reflection coefficient

T is the transmission coefficient.

Seismic Reflection

These equations show that the reflection and transmission coefficients depend on the **difference in impedance** between the two layers.

• If Z1 = Z2, there is no reflection. All energy is transmitted into the second layer. This does not mean that $\rho1=\rho2$ and v1= v2! All that matters is that $\boldsymbol{\rho_1 v_1 = \rho_2 v_2}$

• **R** can have a value of +1 to -1. **R** will be negative when $Z_1 > Z_2$. A negative value means that there will be a phase change of 180° in the phase of the reflected wave (a peak becomes a trough). This is called a **negative polarity reflection**.

• **T** is always positive – transmitted waves have the same phase as the incident wave. **T** can be larger than 1.

• Reflection coefficients for the Earth are generally less than ±0.2, with maximum values of ±0.5. Most energy is transmitted, not reflected.

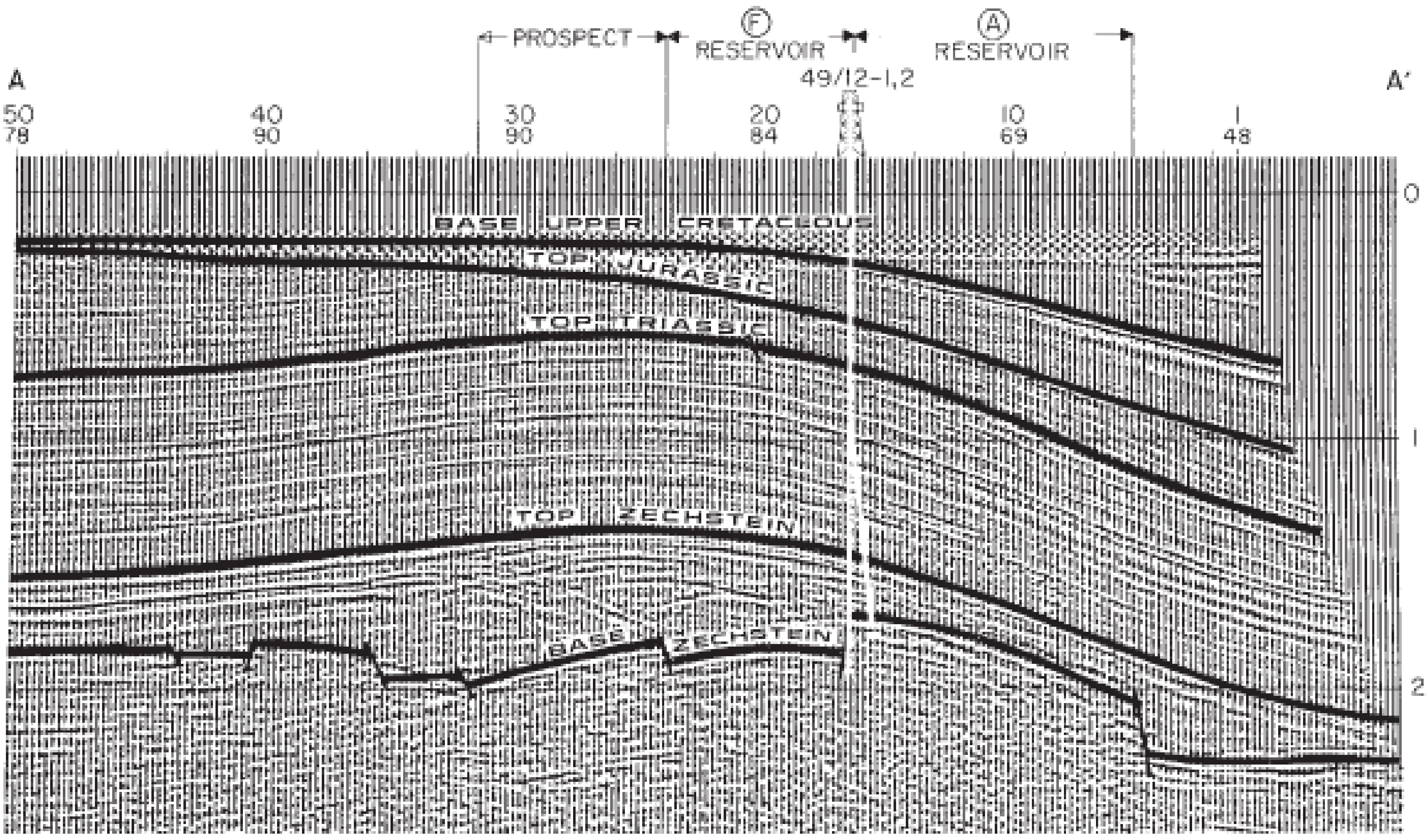


Fig. 4.61 Interpreted seismic section across the North Viking gas field, North Sea. (Courtesy Conoco UK Ltd.)

UNIT FIVE

ELECTRICAL METHOD (DC- RESISTIVITY)

DC RESISTIVITY METHOD

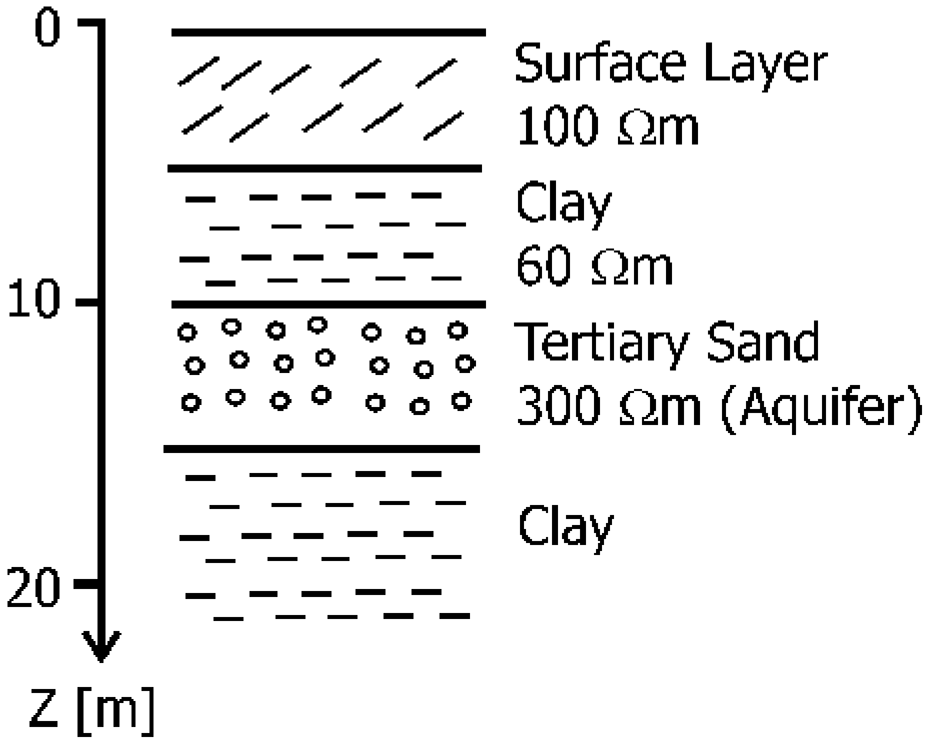
- DC- resistivity method is used in mapping subsurface horizontal and vertical discontinuities.

- It utilizes direct currents or low frequency alternating currents to investigate the electrical properties (resistivity) of the subsurface.

- A resistivity contrast between the target and the background geology must exist.

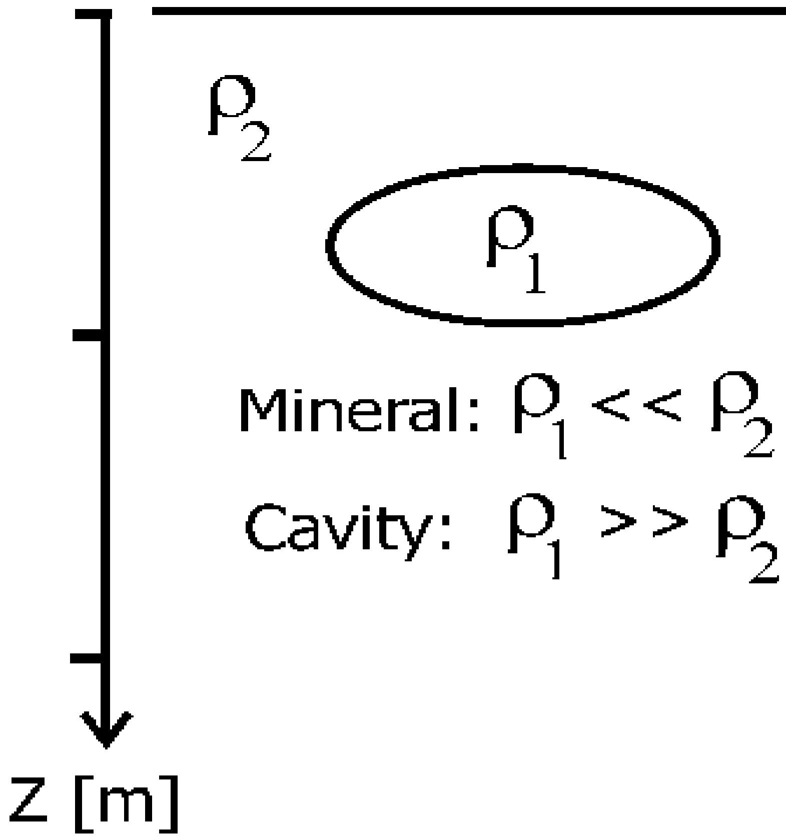
Applications of Resistivity Surveying

- **Groundwater exploration**



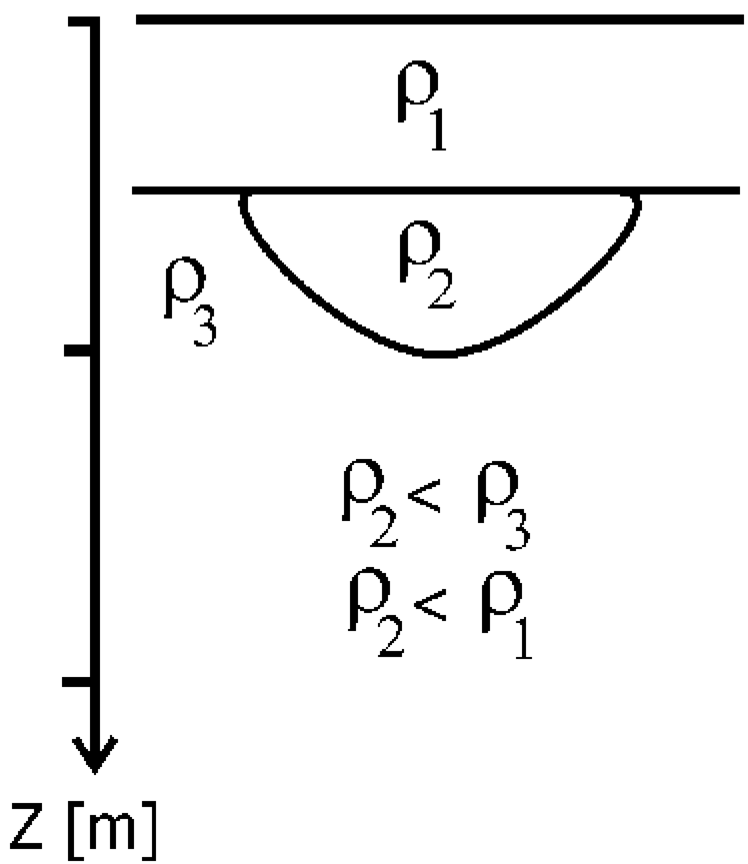
Applications of Resistivity Surveying

- **Mineral exploration and detection of cavities**



Applications of Resistivity Surveying

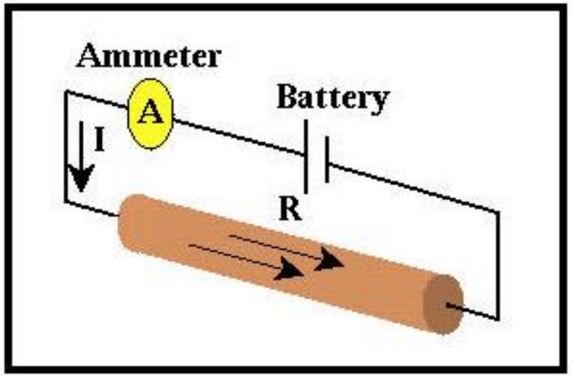
- **Waste site exploration**



Current Flow and Ohm's Law

- In 1827, Georg Ohm defined an empirical relationship between the *current flowing* through a wire and the *voltage potential* required to drive that current:

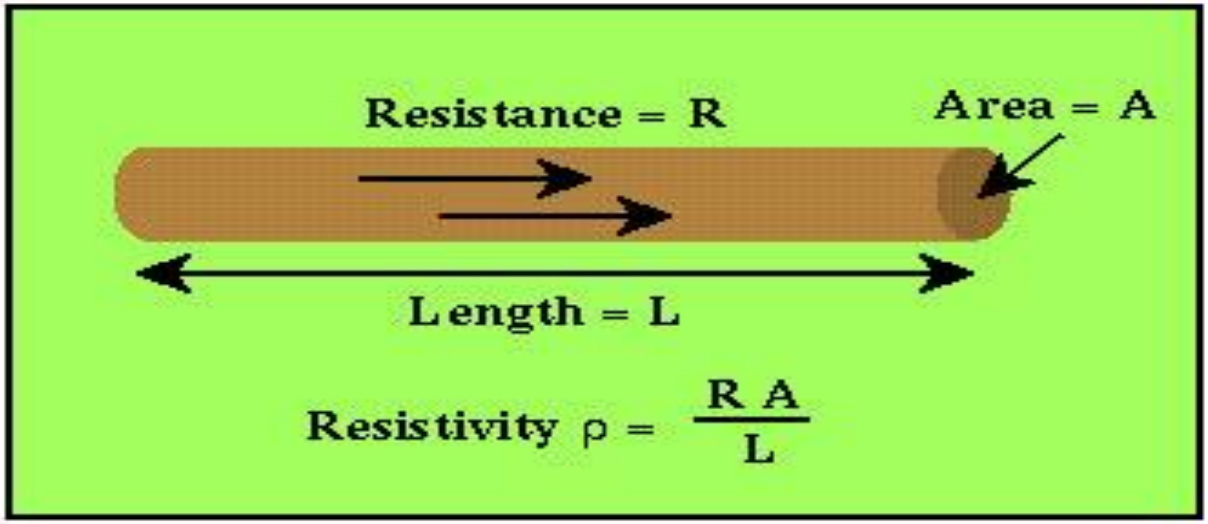
$$\mathbf{V = IR}$$



- Ohm found that the current, I, was proportional to the voltage. The constant of proportionality is called the *resistance of the material (R)* and has the *units of voltage (volts) over current (amperes)*, or Ohm (Ω).

Resistivity VS Resistance

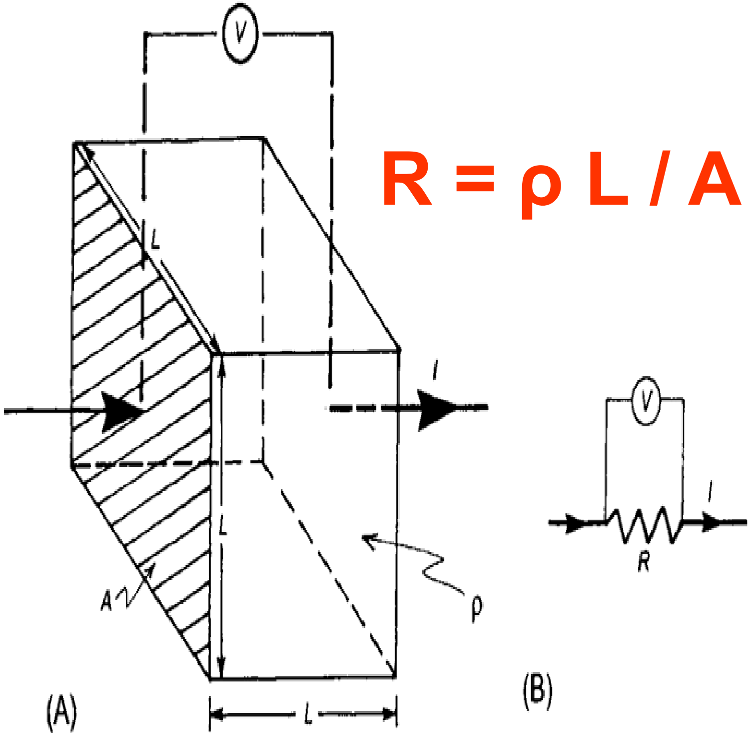
- **Resistance** depends *on the type of material from which the wire is made* and the *geometry of the wire (dimensions)*. Fore example, increasing the <u>length</u> of the wire will increase the measured resistance, while decreasing the <u>diameter</u> of the wire will increase the measured resistance.



Resistivity VS Resistance

- **Resistivity, ρ** is a property that describes a material's ability to transmit electrical current independent of the material's geometrical factors.
- **Resistivity** is the reciprocal of the **conductivity (1/ρ)** of the material. The unit of **Resistivity** is ***ohm-m***.
- The unit of conductivity is Siemens per meter (**S/m**).

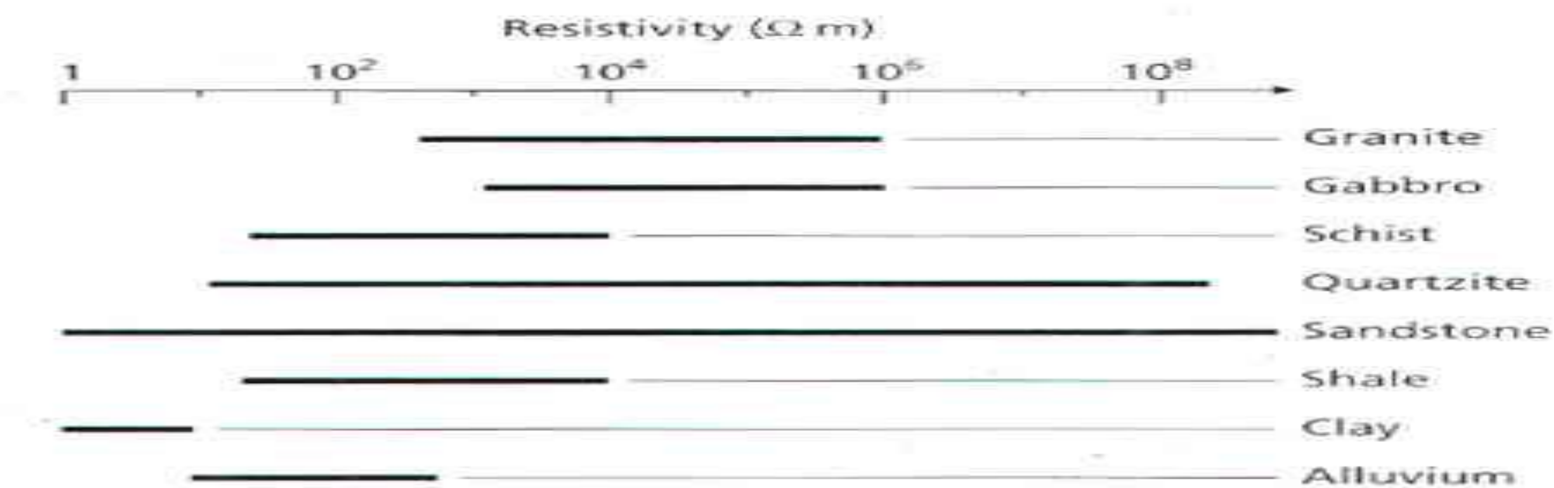
$R = \rho L / A$



Resistivity of Earth's Materials

- **Resistivity values in ohm-m of different rock types and materials:**

Material Value	Resistivity range	Typical
Igneous & Metamporphic rocks	$10^2 - 10^8$	10^4 10^3
Sedimentary rocks	$10 - 10^8$	10^3
Unconsolidated	$10^{-1} - 10^4$	10^3
Groundwater	1 - 10	5
Pure water		10^3



Resistivity of Earth's Materials

- **Resistivity related to rock type:**
  - Igneous rocks → highest resistivities
  - Sedimentary rocks→ the most conductive due to their high fluid content
  - Metamorphic rocks→ intermediate but overlapping resistivities
- **Resistivity related to rock age**:
  - Young volcanic rock (Quaternary) ≈10−200 Ωm
  - Old volcanic rock (Precambrian) ≈100−2000 Ωm

Resistivity of Earth's Materials

- **Important remarks about the resistivity of rocks:**
  - Rocks are usually porous and pores are filled with fluids, mainly water. As a result, **rocks are electrolytic conductors;** *electrical current is carried out through a rock mainly by the passage of ions in pore waters.*
  - There is considerable overlap in resistivity values of different rock types.
  - Identification of a rock type **is not possible** *solely on the basis of resistivity data.*
  - **Resistivity of rocks depends on:** porosity, saturation, content of clay and resistivity of pore water (Archie's formula)

Archie's Law

- **Archie's Law** is an Empirical relationship used to determine the bulk resistivity of a **saturated porous rock.**

$$\boxed{\rho_0 = a\rho_w\phi^{-m}}$$

where:

ρ_0 = bulk rock resistivity

ρ_w = pore-water resistivity

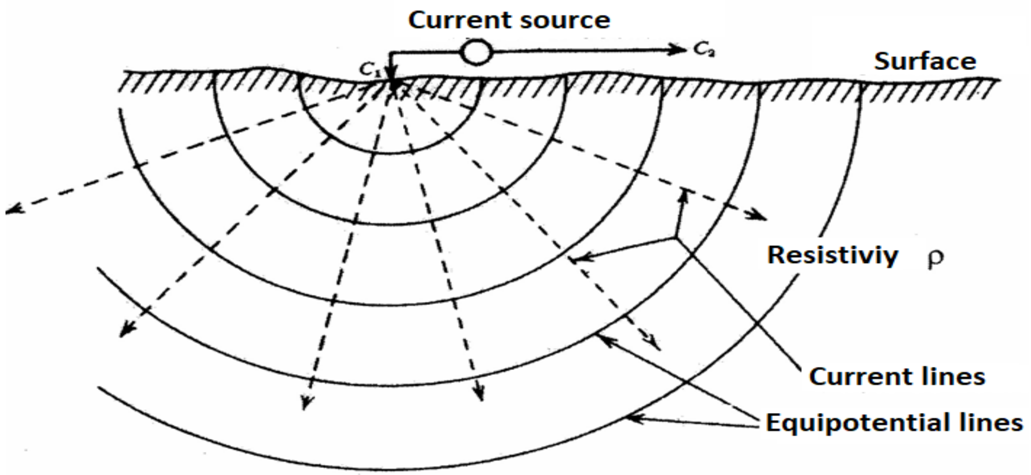
a = empirical constant ($0.6 < a < 1$)

m = cementation factor (1.3 poor, unconsolidated) $< m <$ 2.2 (good, cemented or crystalline)

ϕ = fractional porosity (Volume of liquid/Volume of rock)

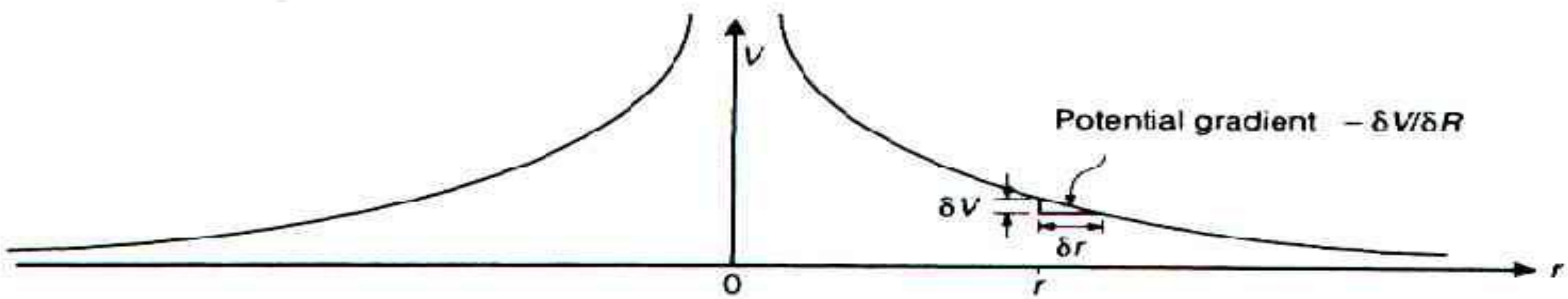
Current Flow in a Homogeneous Earth

- **Current flow for a single surface electrode:**
  - Current flows radially away from the electrode so that the current distribution is uniform over hemispherical shells centered at the source.
  - Lines of equal voltage (equipotential) intersect the lines of equal current at right angles.



Potential Decay Away from the Point Electrode

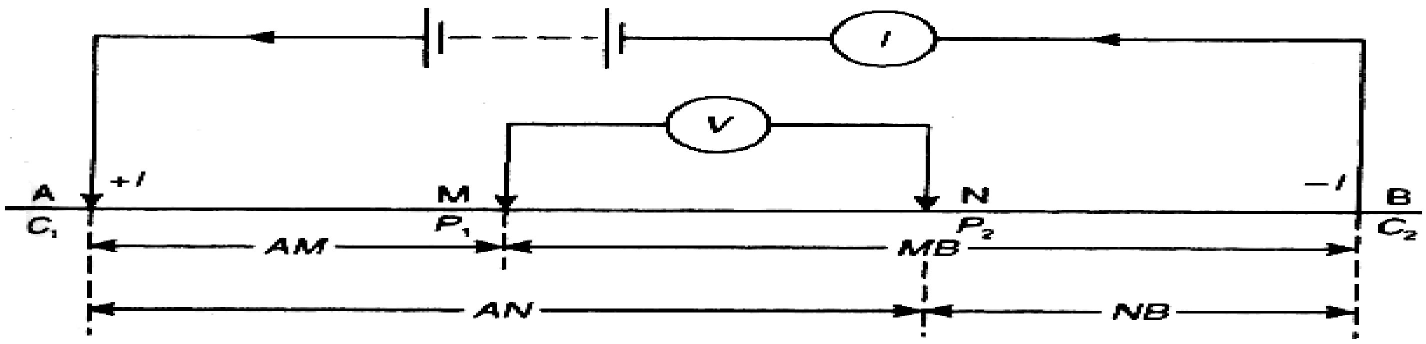
- The voltage drop between any two points on the surface is given by the potential gradient: **dV/dr.**
  dV/dr is negative because the potential decreases in the direction of current flow.



- The **potential V_r** measured at a **distance r** is given by:

$$V_r = I\,\rho\,/\,2\pi\,r$$

Two Current Electrodes



- The potential $\mathbf{V_M}$ at the internal electrode **M** is the sum of the potential contributions $\mathbf{V_A}$ and $\mathbf{V_B}$ from the current source at **A** and the sink at **B**.
- The potentials at electrode M and N are:

$$V_M = V_A + V_B \quad \text{and} \quad V_N = V_A + V_B$$

$$V_M = \rho I / 2\pi\ (1/AM) + \rho(-I) / 2\pi\ (1/MB)$$

$$V_N = \rho I / 2\pi\ (AN) + \rho(-I) / 2\pi\ (1/NB)$$

$$\Delta V = V_M - V_N = \rho I / 2\pi\ (1/AM - 1/MB - 1/AN + 1/NB)$$

Potential for the General Case

$$\Delta V = V_M - V_N = \rho I / 2\pi \left(1/AM - 1/MB - 1/AN + 1/NB \right)$$

Therefore,

$$\rho = (2\pi \Delta V / I) \left(1/AM - 1/MB - 1/AN + 1/NB \right)^{-1}$$

We may write also:

$$\rho = K \, \Delta V_{MN} / I$$

With:

$$K = 2\pi \left(1/AM - 1/MB - 1/AN + 1/NB \right)^{-1}$$

K is the Geometric factor

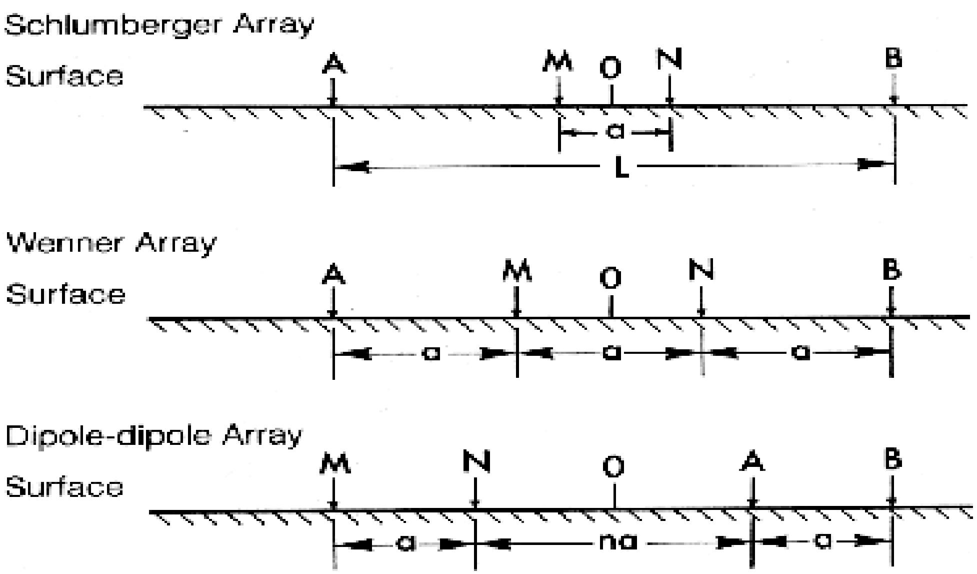
True and Apparent Resistivity

- **ρ** is considered **true resistivity** of the **subsurface if it is homogeneous**.
- If the ground is uniform, the resistivity should be constant and independent of both electrode spacing and surface location.
- If subsurface inhomogeneities exist, the resistivity will vary with the relative positions of electrodes. In this case, the calculated value **(ρ)** is called **apparent resistivity:**

$$\rho_a = K \, \Delta V_{MN} / I$$

- In general, all *field data are apparent resistivity*. *They are interpreted to obtain the true resistivity* of the subsurface layers.

Electrode configurations



L = AB = Separation current electrodes
a = MN = Separation potential electrodes
O = Point of measurement

Derived geometric factors:

$$k=\frac{\pi}{a}\left[\left(\frac{L}{2}\right)^2-\left(\frac{a}{2}\right)^2\right] \quad \text{... Schlumberger}$$

$$k=2\pi a \quad \text{... Wenner}$$

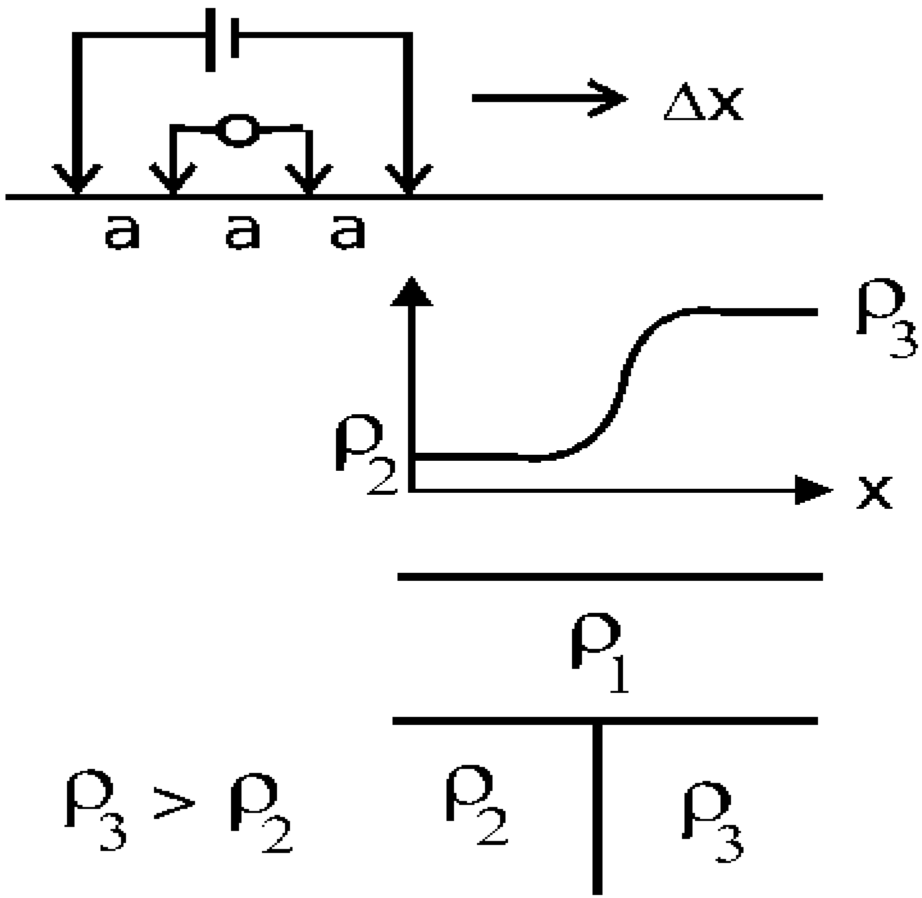
$$k=\pi n(n+1)(n+2)a \quad \text{... Dipole-Dipole}$$

Field Procedures

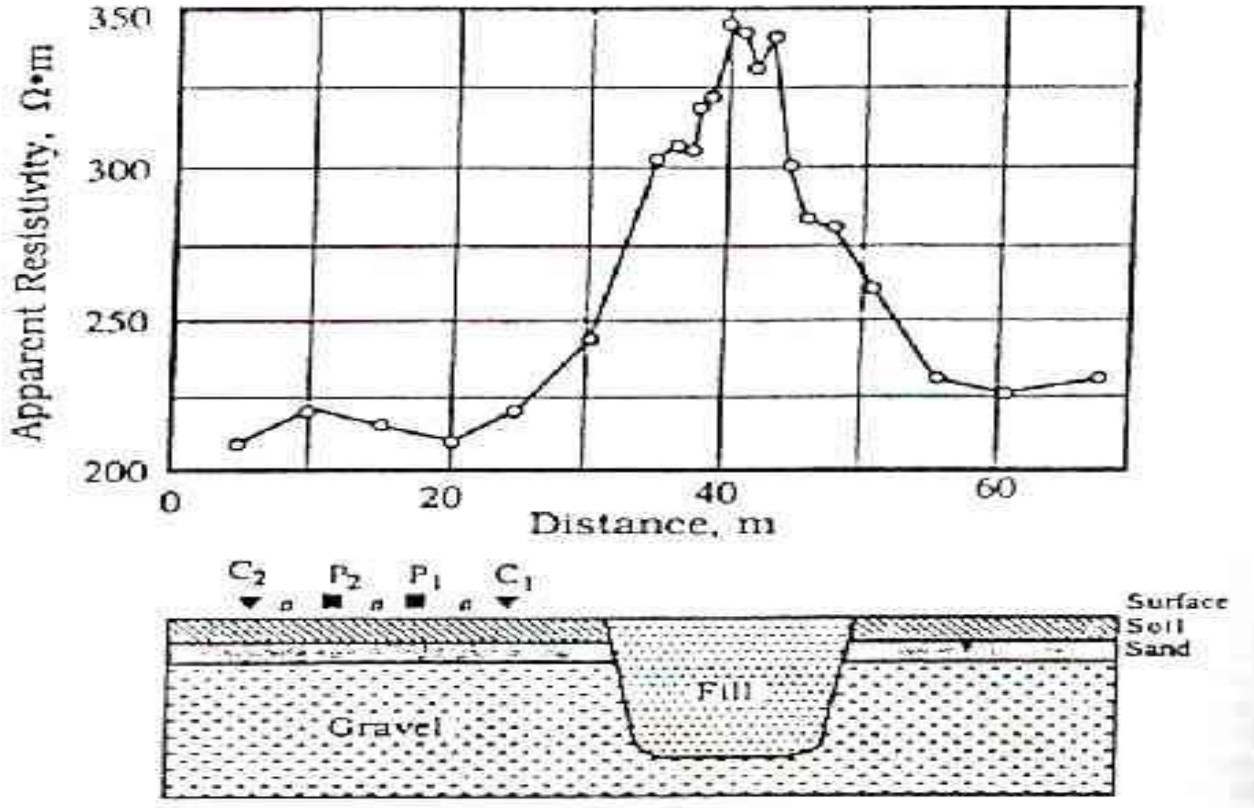
- There are two main filed procedures for the deployment of electrode configurations:
  - **Horizontal Electrical profiling (HEP)**
    - ✓ The object of HEP is to detect lateral variations in the resistivity of the subsurface.
    - ✓ In this case, the *current and potential electrodes* are *maintained at a fixed separation and progressively moved along a profile.*
    - ✓ It is employed in mineral prospecting to locate faults or bodies of anomalous conductivity.
    - ✓ It is used in geotechnical surveys to determine variations in bedrock depth and the presence of steep discontinuities

Field Procedures: HEP

HEP Mapping mode:



Field Procedures: HEP

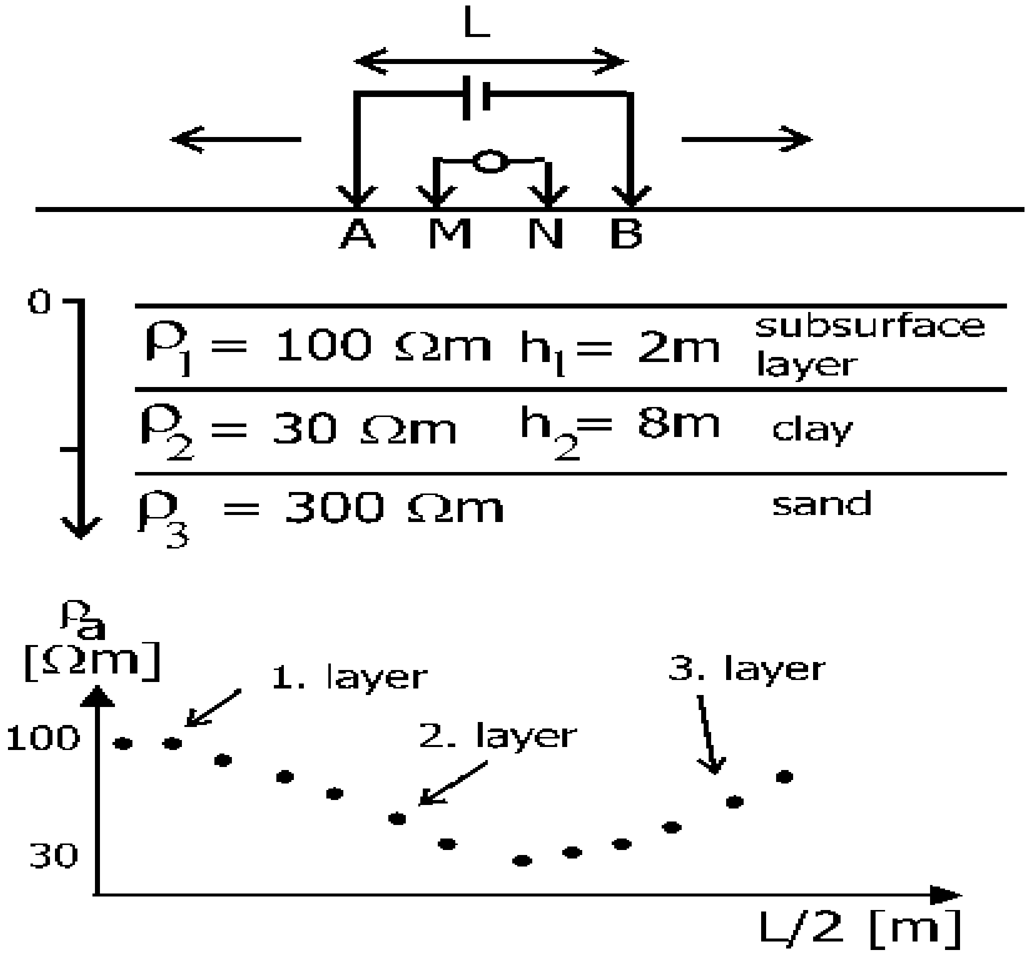


Example: Observed apparent resistivity profile across a resistive landfill using the Wenner configuration.

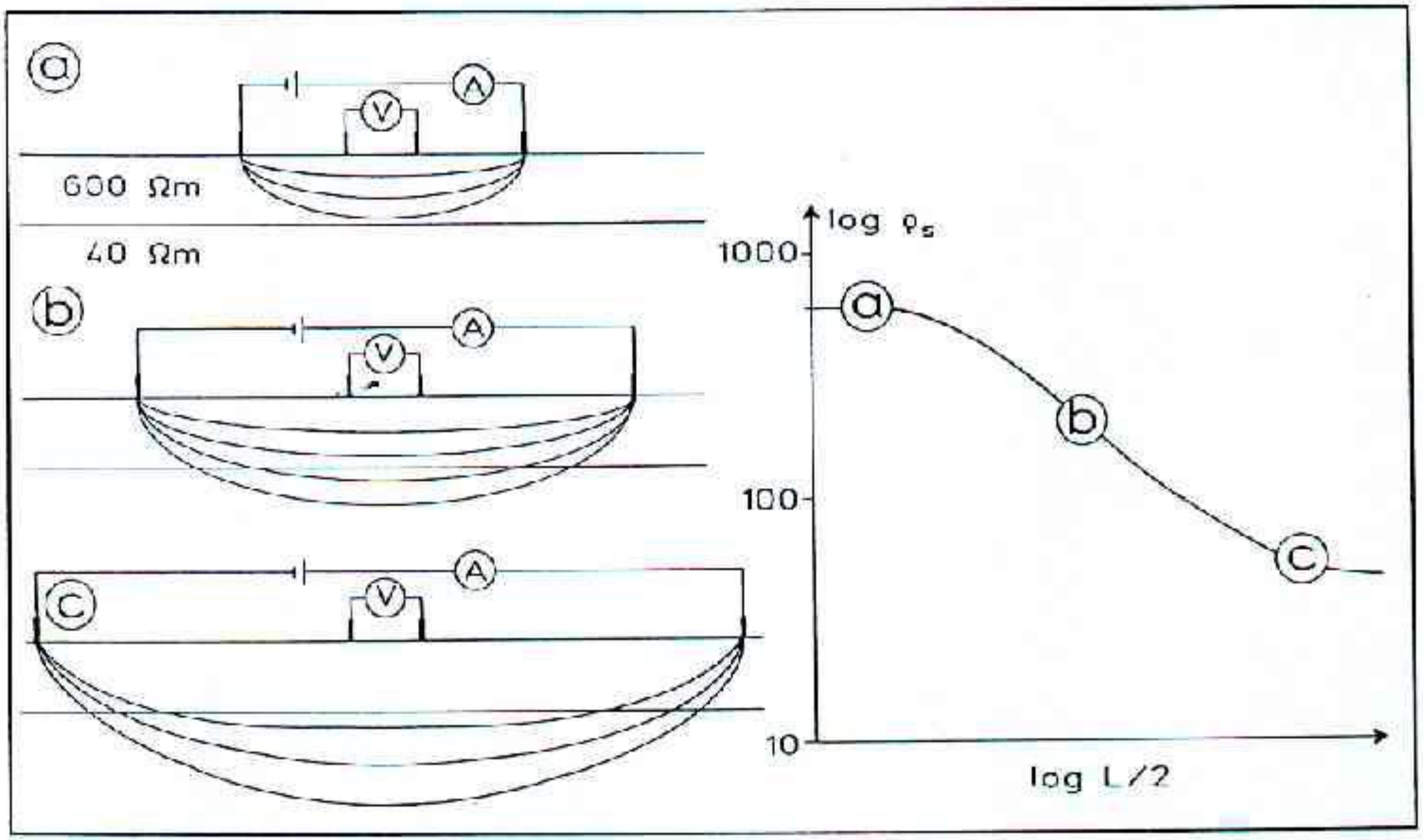
Field Procedures: Sounding (VES)

- **Vertical Electrical Sounding (VES):**
  - VES is used to deduce the variation of resistivity with depth below a given point on the ground surface and to correlate it with the available geological information in order to infer the depths and resistivities of the layers present.
  - Current and potential electrodes are maintained at the same relative spacing and the whole spread is progressively expanded about a fixed central point. As the distance between the current electrodes increases, so the depth to which the current penetrates is increased.

Field Procedures: Sounding (VES)

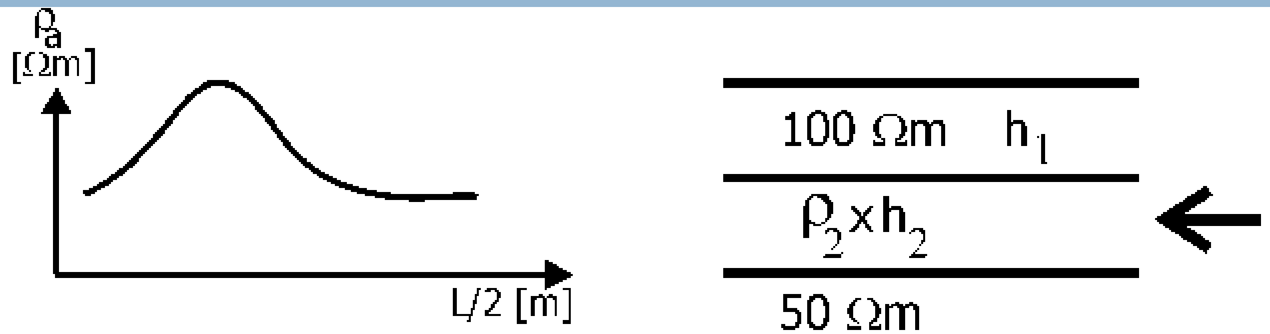


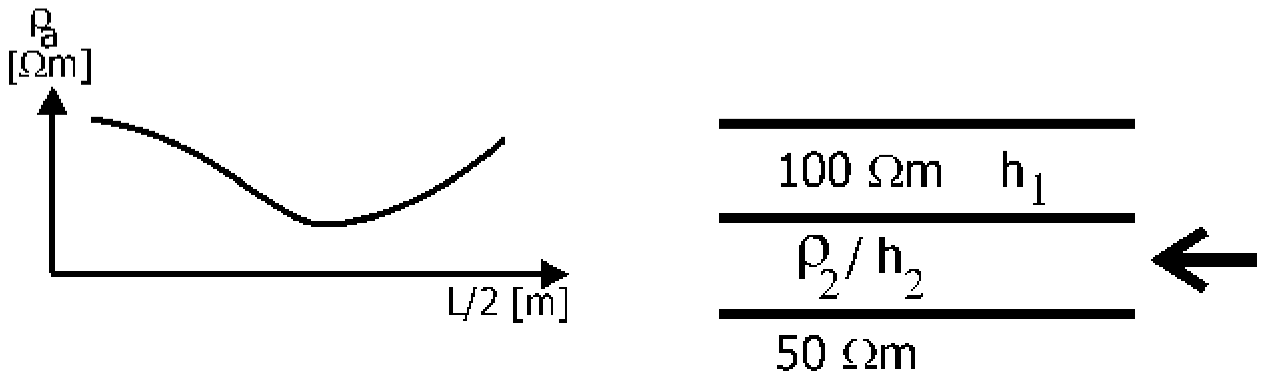
Field Procedures: Sounding (VES)



Development of a sounding curve

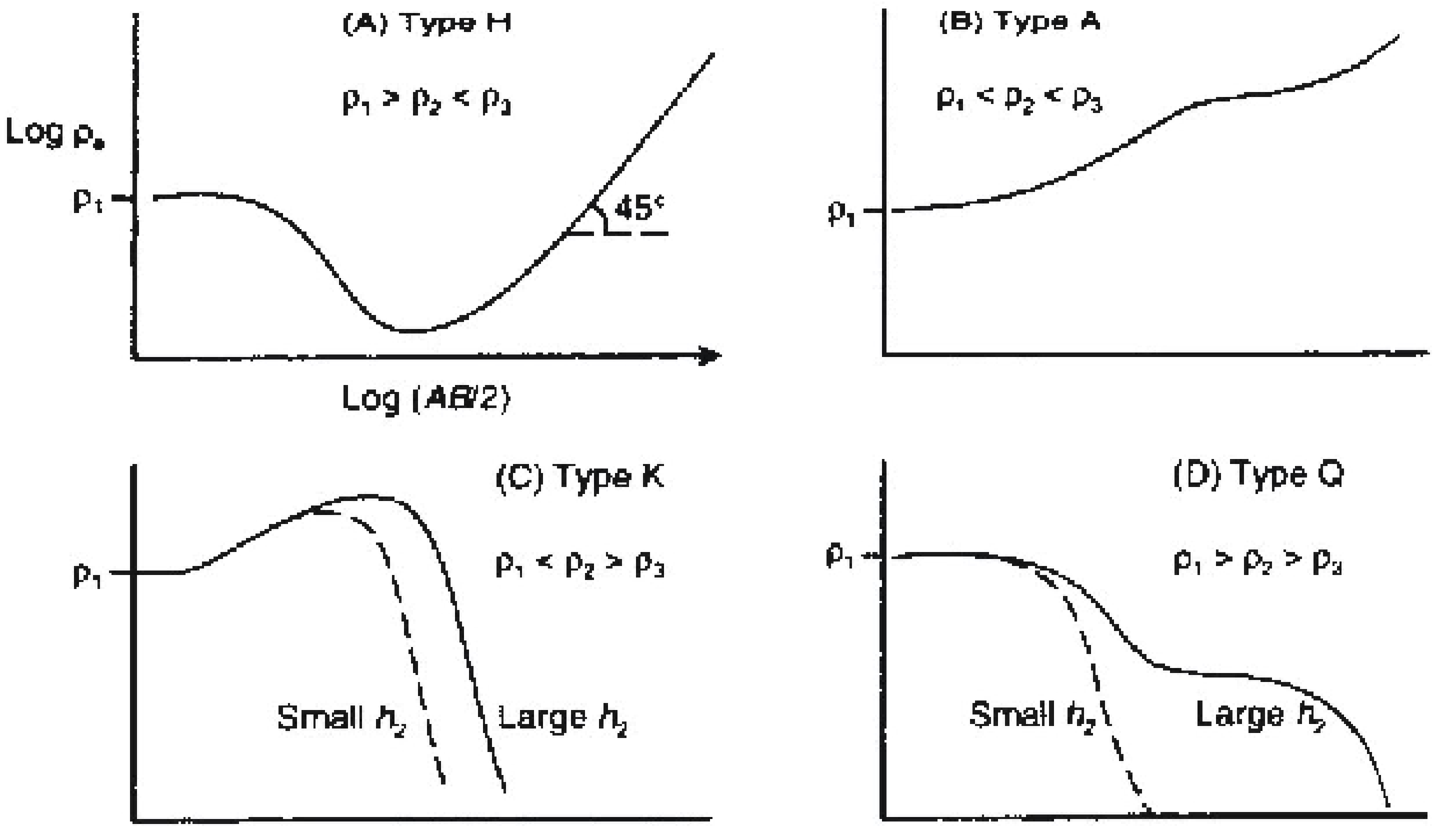
Field Procedures: Sounding (VES)





VES: case of three layers

Types of VES Resistivity Curves



VES: Types of resistivity sounding curves

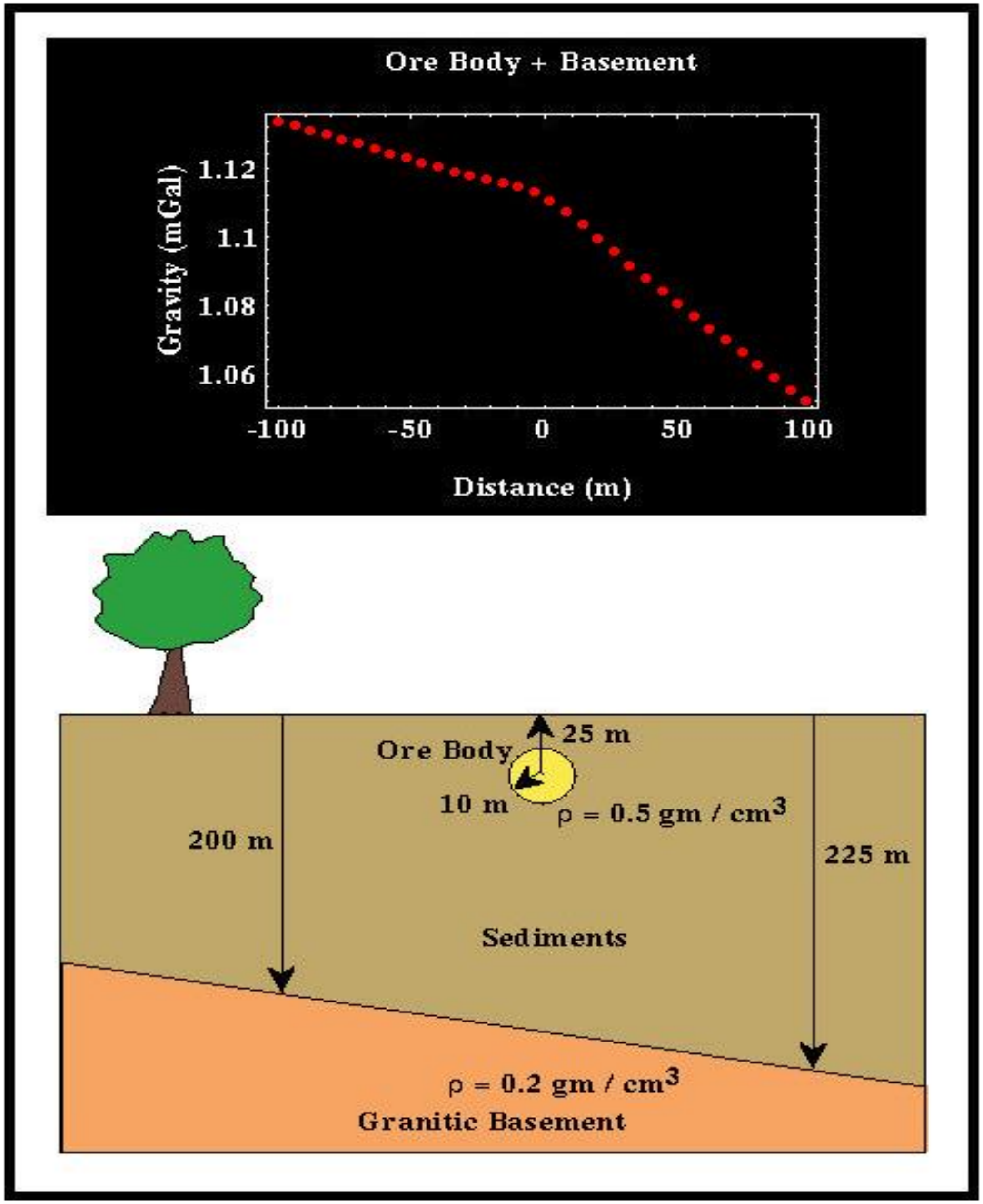
UNIT SIX

GRAVITY METHOD

INTRODUCTION

- Gravity method consists of **measuring, studying and analyzing** *variations*, in space and time, of the gravity field of the Earth. This method is considered one of the fundamental disciplines of geophysics.

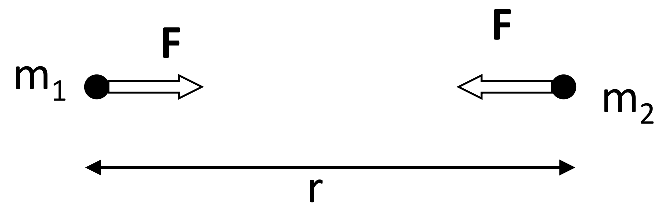
- The objective of exploration work is to **associate** the gravity variations with differences in the distribution of densities and hence rock types.


Ore Body + Basement
Gravity (mGal)
1.12
1.1
1.08
1.06
-100
-50
0
50
100
Distance (m)
25 m
Ore Body
10 m
ρ = 0.5 gm / cm3
200 m
225 m
Sediments
ρ = 0.2 gm / cm3
Granitic Basement

Applications of gravity surveying

- Hydrocarbon exploration
- Geological structures
- Faults location
- Ore bodies exploration
- Cavities detection
- Archaeology

The Gravitational Force:



- **Newton's law of gravitation** states that *the mutual attractive force between two point masses, m_1 and m_2, is proportional to one over the square of the distance* between them. The constant of proportionality is usually specified as *G, the gravitational constant*.
- Thus, the force of one body acting on another is given by *Newton's Gravitational Law :*

$$\mathbf{F = G\, m_1\, m_2 / r^{2}}$$

F *is the force of attraction,*

G *is the gravitational constant.* $G = 6.6725985 * 10^{-11}$ N m^2 / kg^2 (SI)

r *is the distance between* the two masses, m_1 *and* m_2.

Gravitational Acceleration (التسارع الجاذبي)

- When making measurements of the Earth's gravity, we usually don't measure the gravitational force, ***F***. *Rather, we measure the gravitational* acceleration, ***g***.

- *The gravitational acceleration is the time rate of change* of a body's speed under the influence of the gravitational force. That is, if you drop a rock off a cliff, it not only falls, but its speed increases as it falls.

Gravitational Acceleration (التسارع الجاذبي)

- **Newton's second law** states that *force is proportional to acceleration*. *The constant of proportionality is the mass of the object:*

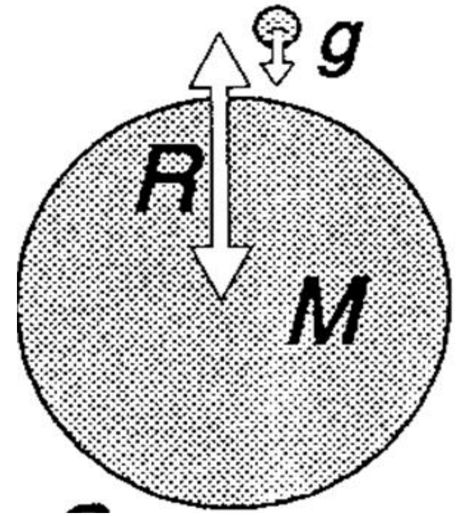
$$\mathbf{F = m_2\, g}$$

- Combining Newton's second law with his law of mutual attraction ($\mathbf{F = G\ m_1\ m_2\ /r^2}$), *the gravitational acceleration on the mass m_2 can be shown to be:*

$$g = G\, m_1 / r^2$$

- **For Earth's Gravity Field,**

$$g = GM / R^2$$



M: Mass of the Earth
R: distance from the observation point to Earth's center.

- The above equation illustrates **two fundamental properties of gravity**:

  - Acceleration due to gravity (g) does not depend on the mass (m) attracted to the Earth.

  - The farther of Earth's center of mass (the greater the R), the smaller the gravitational acceleration.

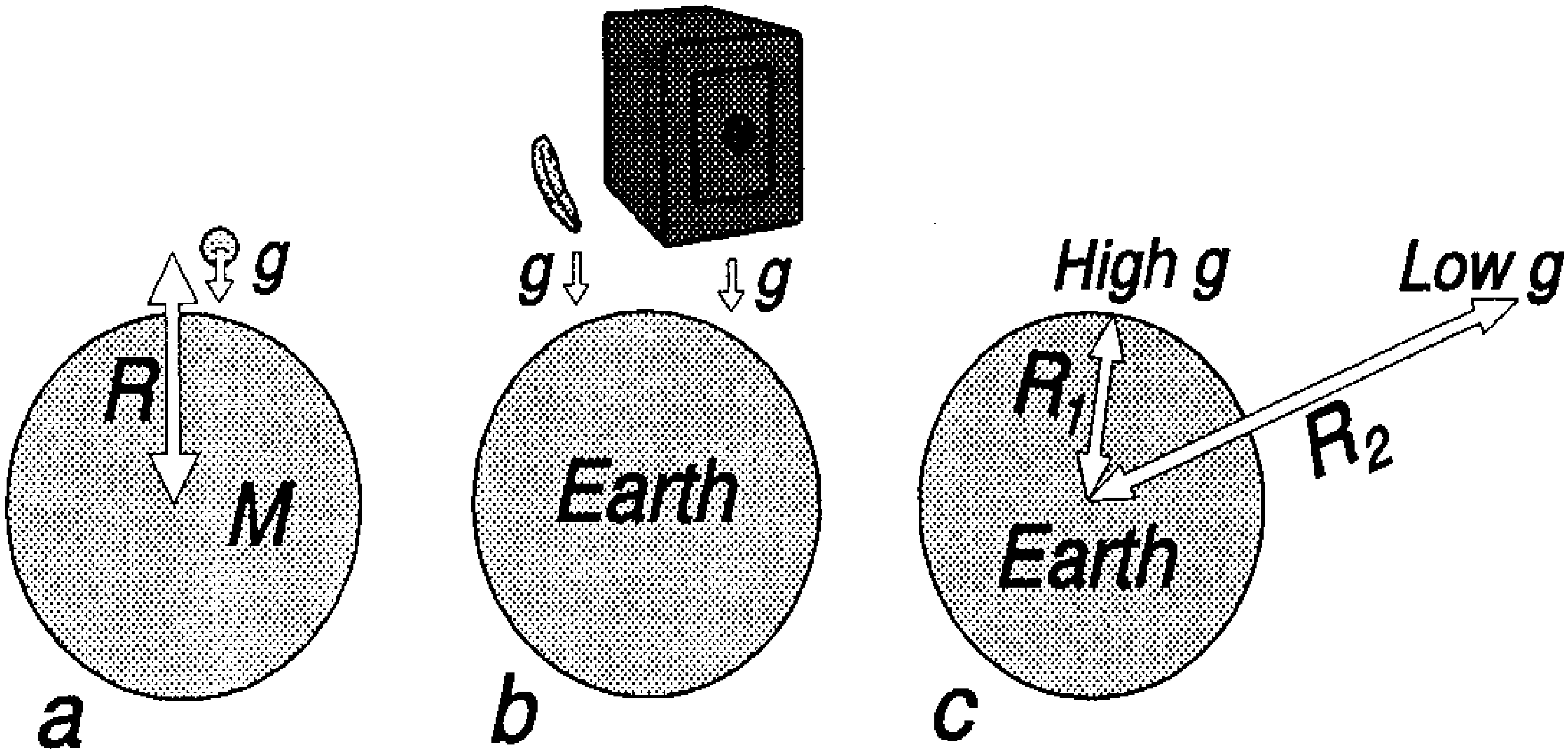


FIGURE 8.3 a) The mass (M) of the Earth and radius (R) to Earth's center determine the gravitational acceleration (g) of objects at and above Earth's surface. b) The acceleration is the same (g), regardless of the mass of the object. c) Objects at Earth's surface (radius R_1) have greater acceleration than objects some distance above the surface (radius R_2).

UNITS of MEASURING GRAVITATIONAL ACCELERATION

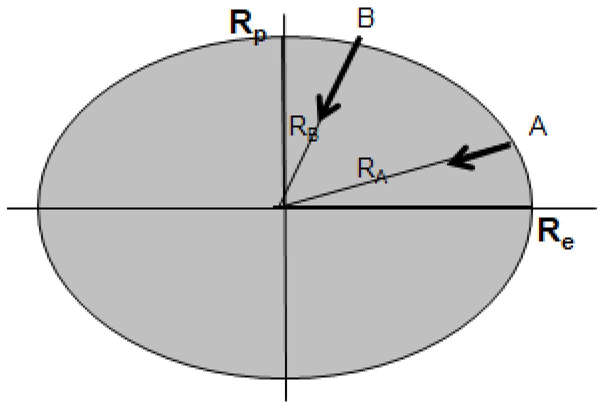
- Gravitational acceleration (gravity) is commonly expressed in units of **milliGals (mGal).**

$$\mathbf{1\ Gal = 1 cm/s^2 = 0.01 m/s^2}$$

$$\mathbf{1 mGal = 10^{-3}\ Gal = 10^{-3}\ cm/s^2 = 10^{-5}\ m/s^2}$$

<u>Latitude Dependent</u> Changes in Gravity

- <u>Two features affect the Earth gravity value</u>: the **shape** and **rotation** of the Earth.
- As an approximation, the shape of the Earth is elliptical, with the widest portion of the ellipse at the equator.
- The **elliptical shape** of the Earth causes the *gravitational acceleration to vary with latitude* because the distance between the gravimeter and the earth's center varies with latitude.
- Thus, we expect the **gravitational acceleration** to be <u>*smaller at the equator than at the poles*</u>, because the surface of the earth is farther from the earth's center at the equator than it is at the poles.



Variation of gravity due to **elliptic shape** of the Earth.

$$g=GM/R^2$$

$R_B < R_A$

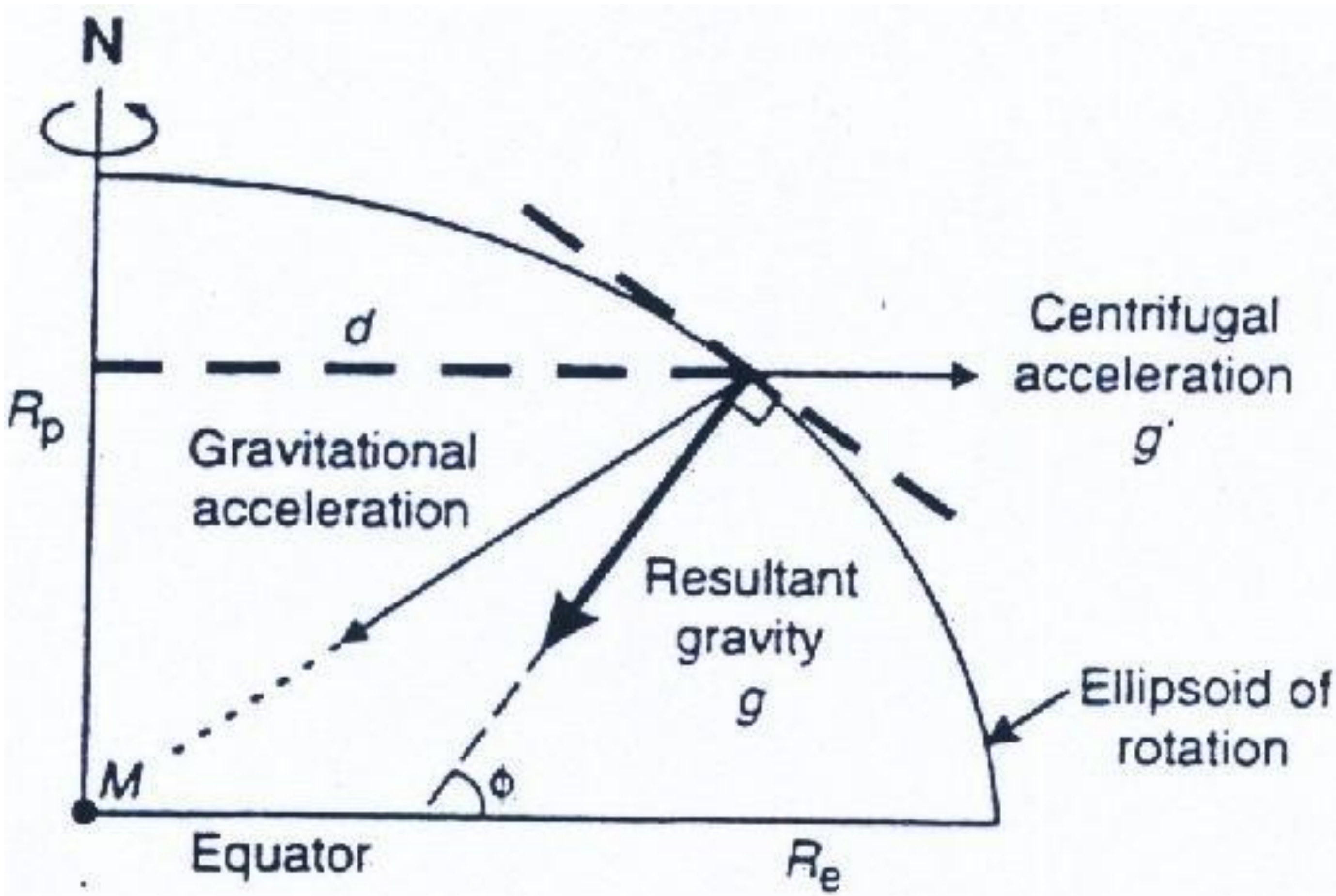
Therefore:

$$g_B > g_A$$

- ***Rotation*** - In addition to shape, the fact that the Earth is *rotating also causes a change in the gravitational acceleration with latitude*.

- We know that if a body rotates, it experiences an outward directed force known as a **centrifugal force**. The size of this force is *proportional to the distance from the axis of rotation and the rate at which the rotation is occurring*.

- The size of the centrifugal force is relatively large at the equator and goes to zero at the poles. This force always acts away from the axis of rotation. Therefore, this force acts to **reduce the gravitational acceleration** .



Gravity (**g**) is the **<u>resultant</u>** of Gravitational acceleration and Centrifugal acceleration.

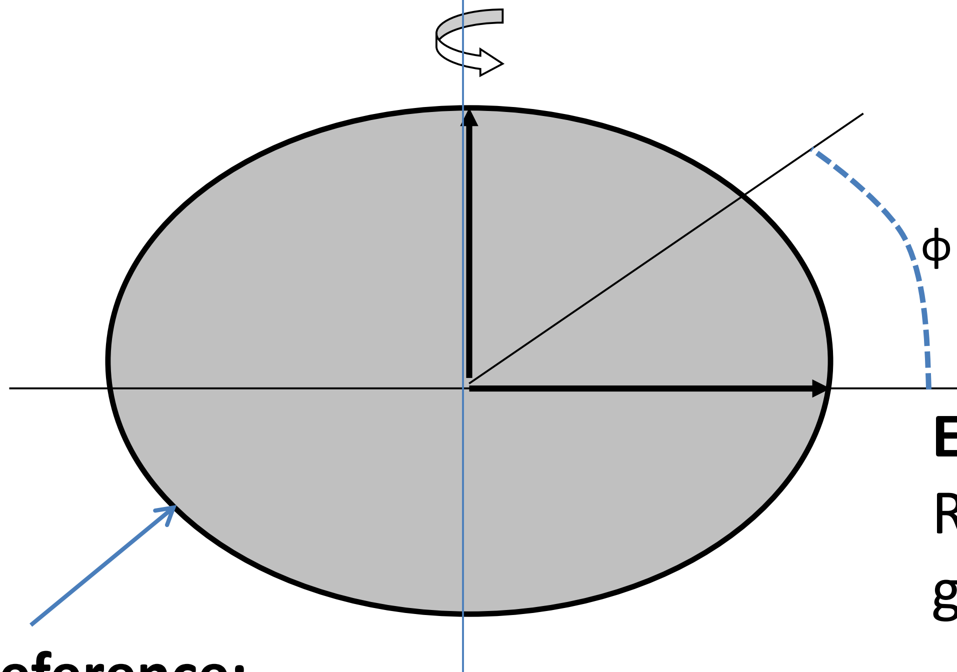
THE REFERENCE GRAVITY FORMULA

- By assuming the Earth is **elliptical** with the appropriate dimensions, is **rotating** at the appropriate rate, and **contains no lateral variations** in geologic structure, we can derive a mathematical formulation for the Earth's gravitational acceleration that depends only on the latitude of the observation.

Pôle, $\phi = 90°$

$R_{pole} = 6356$ Km

$g_{th} = 983\ 217.72$ mGal



Equateur, $\phi = 0°$:

$R_{eq} = 6378$ Km

$g_{th} = 978\ 031.85$ mGal

Ellipsoid reference:

$f = (R_e - R_p)/R_e = 1/298.247$

- The <u>average value of gravity for a given latitude </u>is approximated by the **1967 Reference Gravity Formula**, adopted by the International Association of Geodesy:

$$\mathbf{g_{th} = g_{eq} (1 + 0{,}005278895 \sin^2(\phi) + 0{,}000023462 \sin^4(\phi))}$$

Φ:Latitude of the observation point (degrees)

g_{eq}: theoretical gravity at the equator (978,031.85 mGal).

Ellipsoid reference: R_{eq} = 6378 Km; R_{pole}= 6356 Km

- **This equation takes into account the fact that the Earth is elliptic and rotating about an axis through the poles**.

HOW DO WE MEASURE GRAVITY

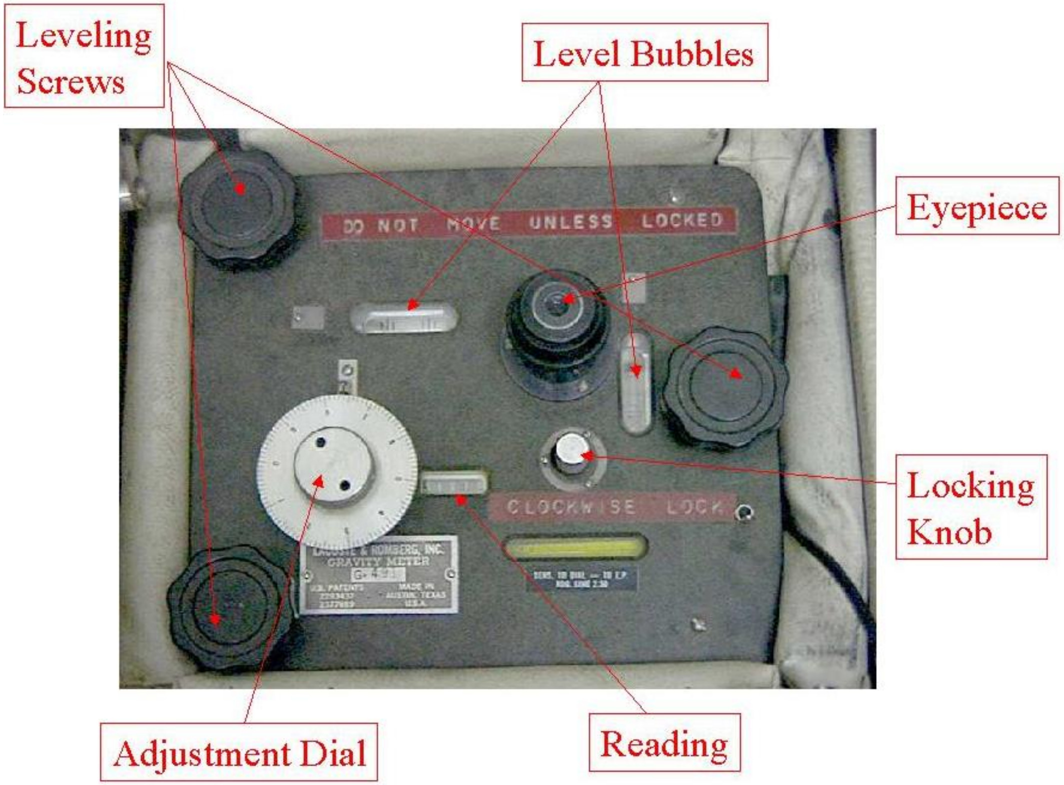
- Two ways are used to measure gravity:
  - **Absolute measurements of gravity (g)**
  - **Relative measurements of gravity (g)**

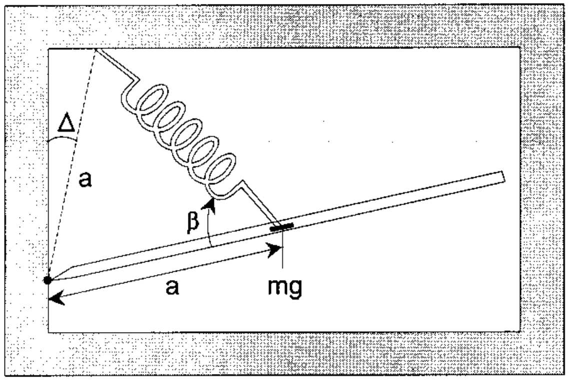
Gravimeters using mass and spring

- Instruments of this type are produced by several manufacturers; ***LaCoste and Lomberg**, **Texas Instruments*** *(Worden Gravity Meter), and* ***Scintrex****. Modern gravimeters are capable of measuring changes in the Earth's* gravitational acceleration with a precision of 0.001 mgal.
- This precision <u>can be obtained only under optimal conditions</u> when the recommended field procedures are carefully followed.

Worden Gravity meter

Lacoste-Romberg Gravity meter





Electronic Gravity Meter
CG-5 AUTOGRAV, SCINTREX

CORRECTIONS of GRAVITY DATA

- **Correction of temporal variations (*time dependent*)**
- **correction of spatial variations (*location dependent*)**

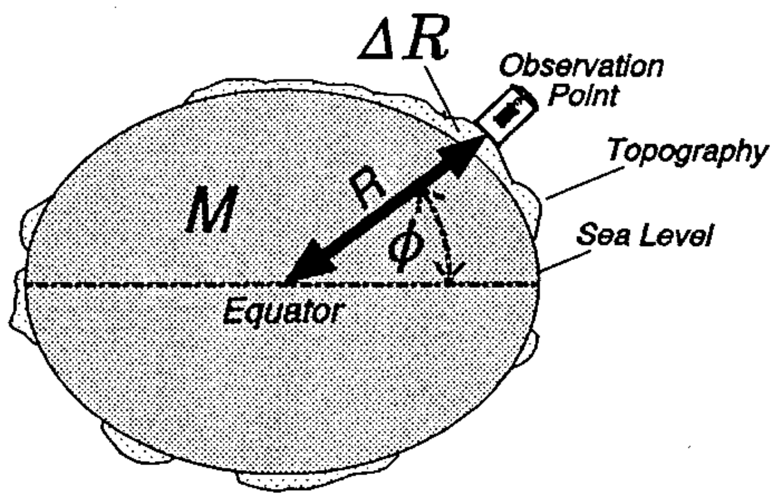
CORRECTION OF *TEMPORAL* VARIATONS

- There are changes in the observed gravity that are *time dependent*. In other words, these factors cause variations in gravity that would be observed even if we didn't move our gravimeter.
- Two factors cause temporal variations:
  - **<u>Instrument Drift</u>**

    Changes in the observed gravity caused by changes in the response of the gravimeter over time.
  - **<u>Tidal Effects (Tides)</u>**

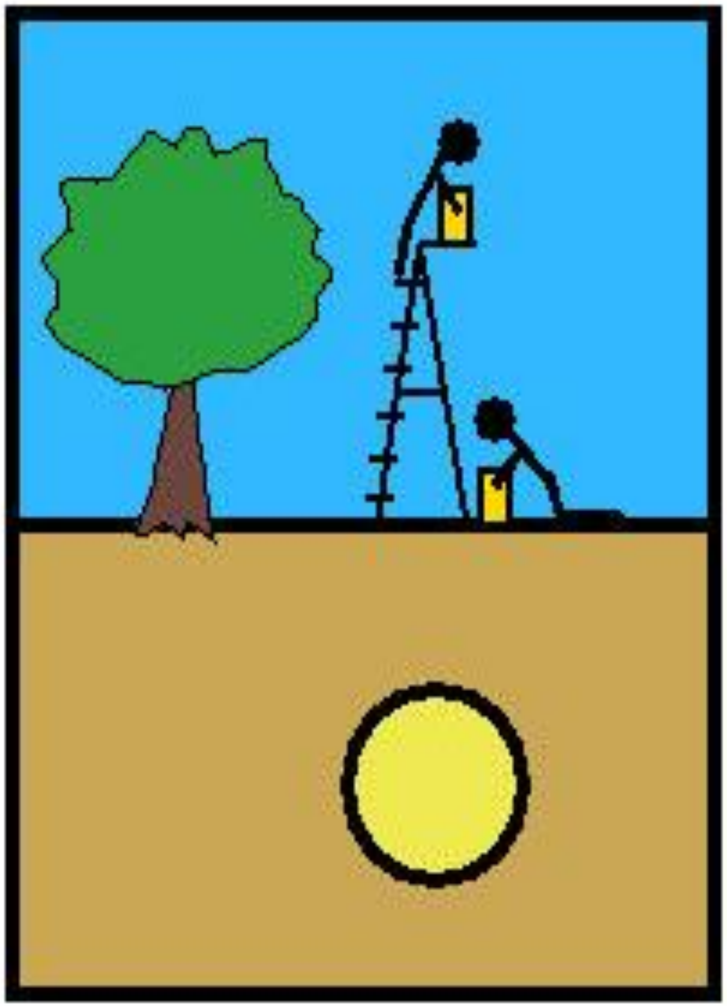
    Changes in the observed gravity caused by the gravitational attraction of the sun and moon.

CORRECTION OF SPATIAL VARIATIONS



- **Observed Gravity** (**g_s**) *at a specific location* on Earth's surface *is a function of three main components*:
  - ➢ The **latitude** of the observation point, accounted for by the theoretical gravity formula.
  - ➢ The **elevation** of station (ΔR), which changes the radius (R).
  - ➢ The **mass distribution** (M) in the subsurface, relative to the observation point.

Variation in Gravitational Acceleration Due to Changes in Elevation



Would the two instruments record the same gravitational acceleration?

The instrument placed on top of the step ladder would record a smaller gravitational acceleration than the one placed on the ground.

FREE AIR CORRECTION

- **Free-Air Correction (FAC)** is used to *account for variations in the observed gravitational acceleration that are related to elevation variations*.
- *In applying this correction, we* mathematically convert our observed gravity values to ones that look like they were all recorded at the **same elevation**.
- To apply an elevation correction to our observed gravity, we need to know the **elevation** of every gravity station. If this is known, we can correct all of the observed gravity readings to a **common elevation**, <u>usually chosen to be sea level</u>.

- Consider the equation for the gravitational acceleration (g) as a function of R:

$$\mathbf{g = GM/R^2}$$

$$\mathbf{dg/dR = -2(GM/R^3) = -2(g)/R}$$

Assuming average value of g=980625mGal and R=6367Km,

$$\mathbf{dg/dR = -0.3086\ mGal/m}$$

<u>dg/dR = average value for the change in gravity with increasing elevation.</u>

- *Stations at elevations high above sea level have lower gravity readings than those near sea level*.

- *To compare gravity observations for stations with different elevations, a Free Air Correction (FAC) must be added back to the observed values:*

$$\mathbf{FAC = h \times (0.3086\ mGal/m)}$$

Where h is elevation of the station above sea level.

FREE AIR ANOMALY

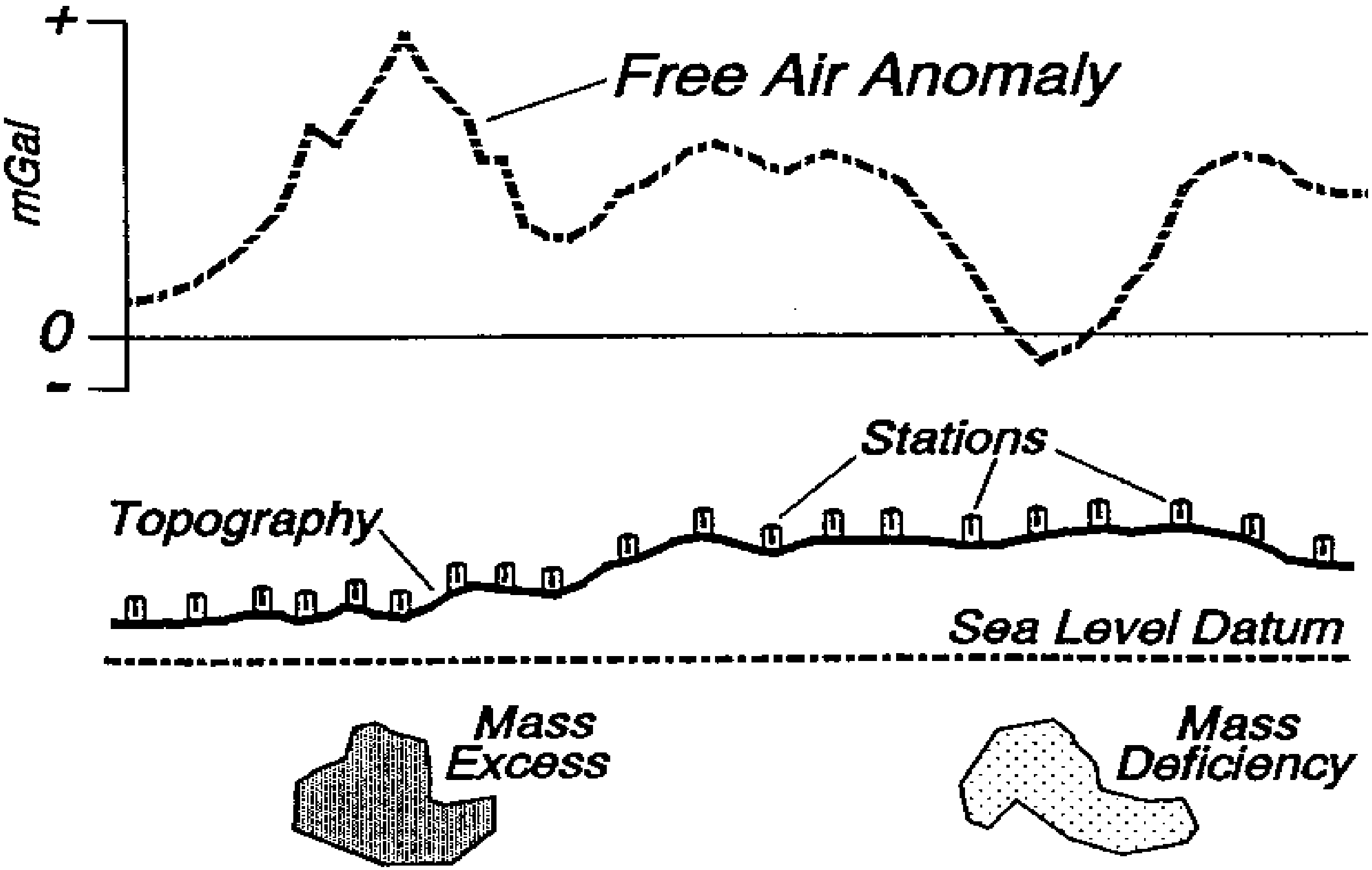
- The **Free Air Gravity Anomaly** is the *observed gravity corrected for the latitude and elevation of station*.

$$\mathbf{\Delta g_{fa} = g_{obs} - g_{th} + FAC}$$

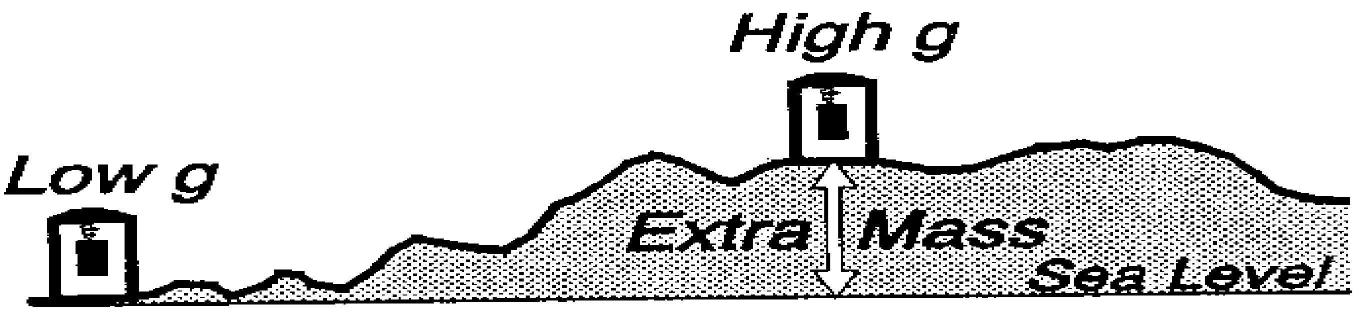
$\mathbf{\Delta g_{fa}}$: Free Air gravity Anomaly.

$\mathbf{g_{obs}}$ **:** Gravitational acceleration observed at the station

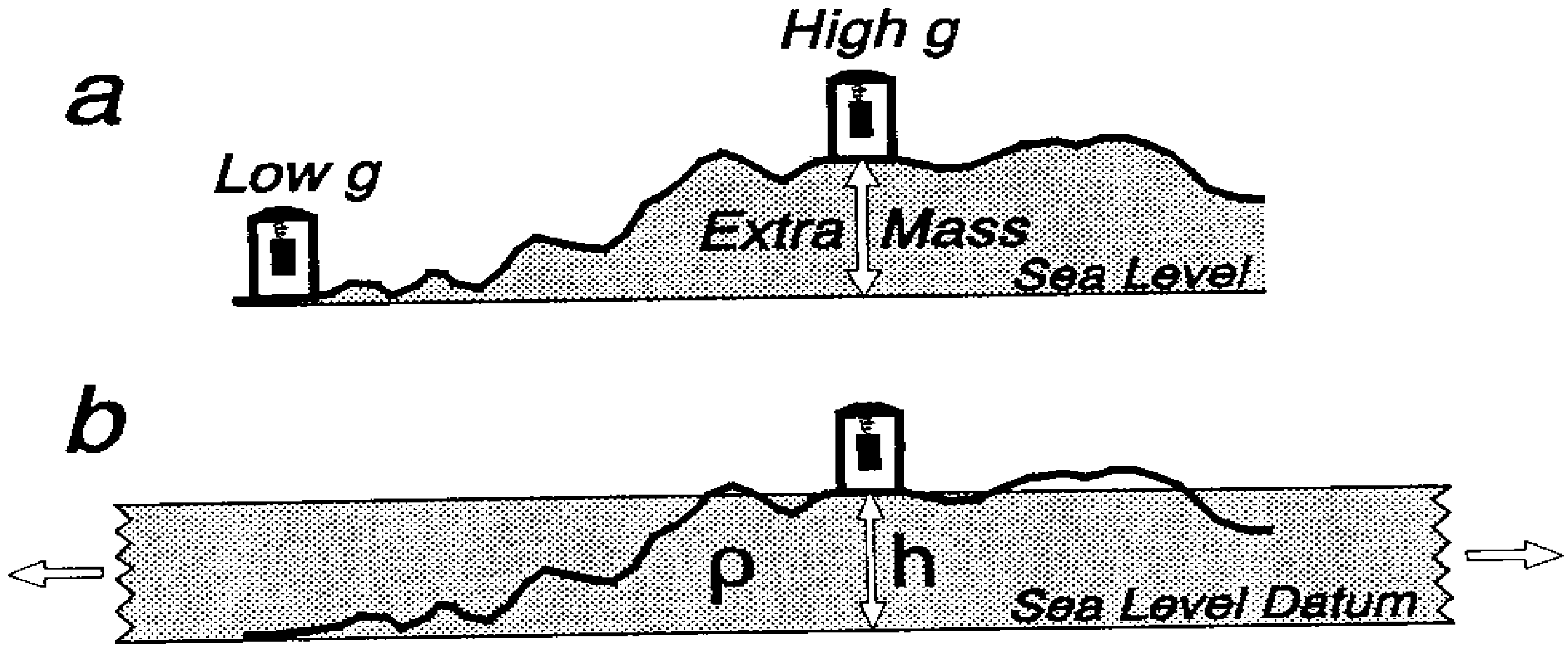
$\mathbf{g_{th}}$: Theoretical gravitational acceleration


+
mGal
0
-
Free Air Anomaly
Stations
Topography
Sea Level Datum
Mass
Excess
Mass
Deficiency

BOUGUER CORRECTION



- In addition to the gravity readings differing at two stations because of elevation differences, the readings will also contain a difference because there is **more mass** below the reading taken at a higher elevation than there is of one taken at a lower elevation.
- *Mountainous areas would have extra mass compared to areas near sea level*, tending to **increase the gravity**.



- To correct the effect of extra mass, we assume that the excess mass underneath the observation point at higher elevation *can be approximated by a* ***slab of uniform density*** and ***thickness***.

- The attraction of such slab is:

$$\mathbf{BC = 2\pi\rho G h}$$

BC: Bouguer correction

ρ: Density of the slab

G: Universal Gravitational constant

h: thickness of the slab (station elevation).

$$\mathbf{BC = 0.0419\ \rho\ h}$$

BC is in mGal; **ρ** in g/cm^3; **h** in meters.

BOUGUER GRAVITY ANOMALY

- The ***Simple Bouguer gravity anomaly** (**Δg_B**)* results from subtracting the effect of the infinite slab (BC) from the Air Free Anomaly (**Δg_{fa}**).

$$\Delta g_B = \Delta g_{fa} - BC$$

$$\Delta g_B = g_{obs} - g_{th} + FAC - BC$$

COMPLETE BOUGUER ANOMALY

$$\Delta g_{BC} = g_{obs} - g_{th} + FAC - BC + TC$$

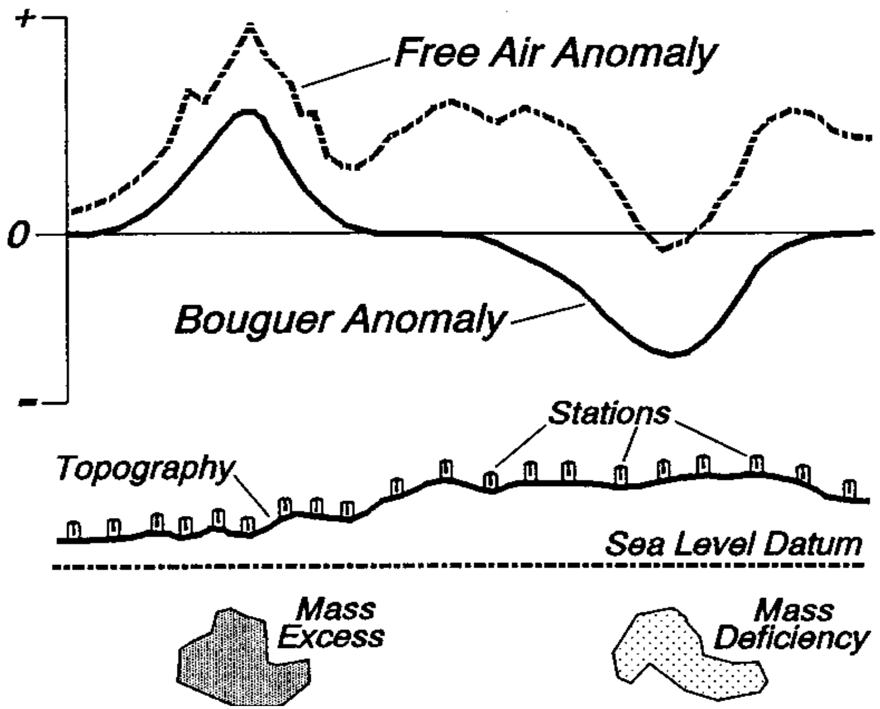
$$\Delta g_{BC} = g_{obs} - g_{th} + 0.3086h - 0.0419\,\rho h + TC$$

ρ: mean density of the extra mass above sea level (reference)

h: elevation

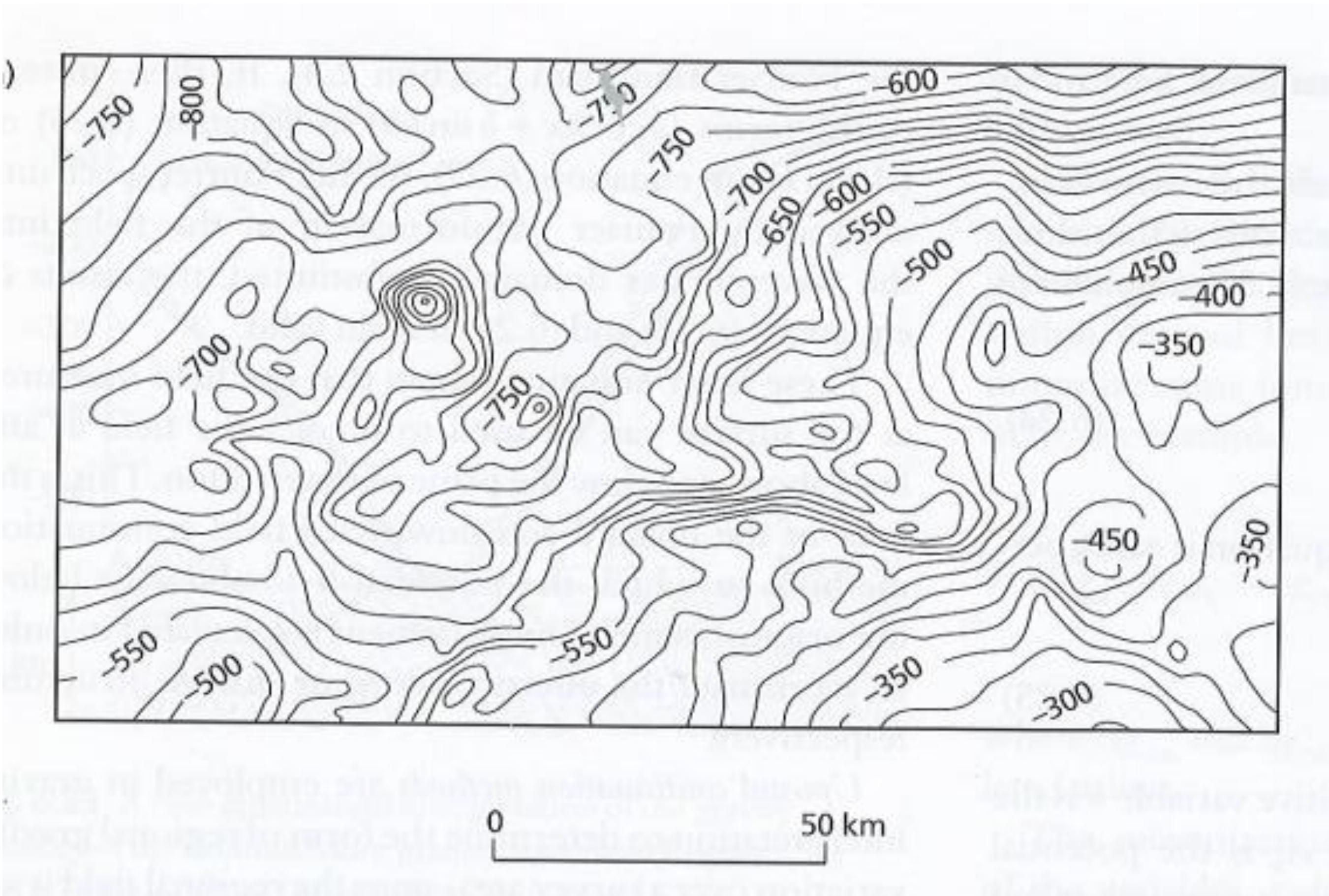
TC is to account for topographic relief in the vicinity of the gravity station. TC>0

COMPLETE BOUGUER CORRECTION



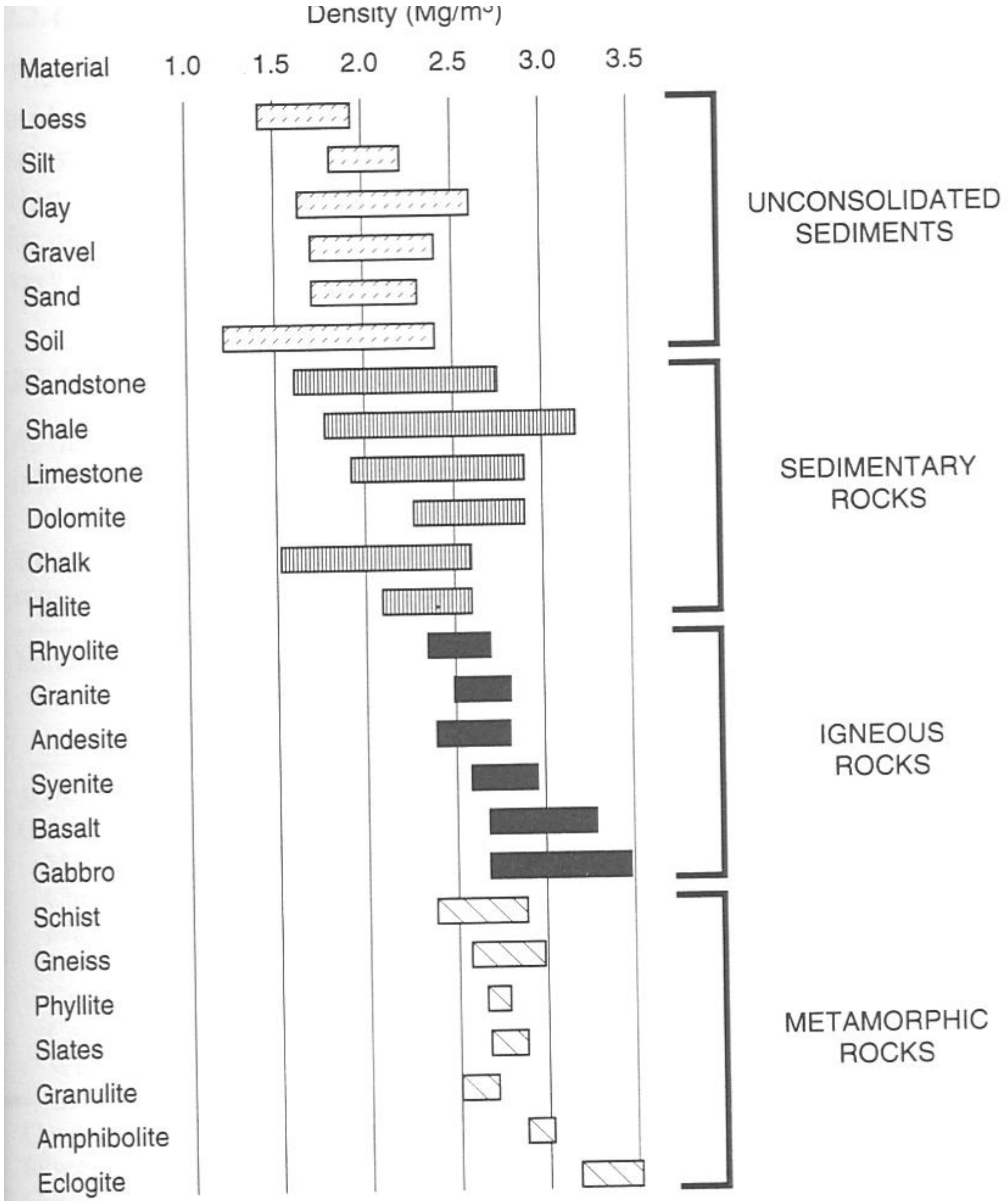
- The Bouguer Anomaly reflects changes in mass distribution below the surface.

- *Mass excess results in positive anomaly*; Mass deficiency result in negative anomaly.

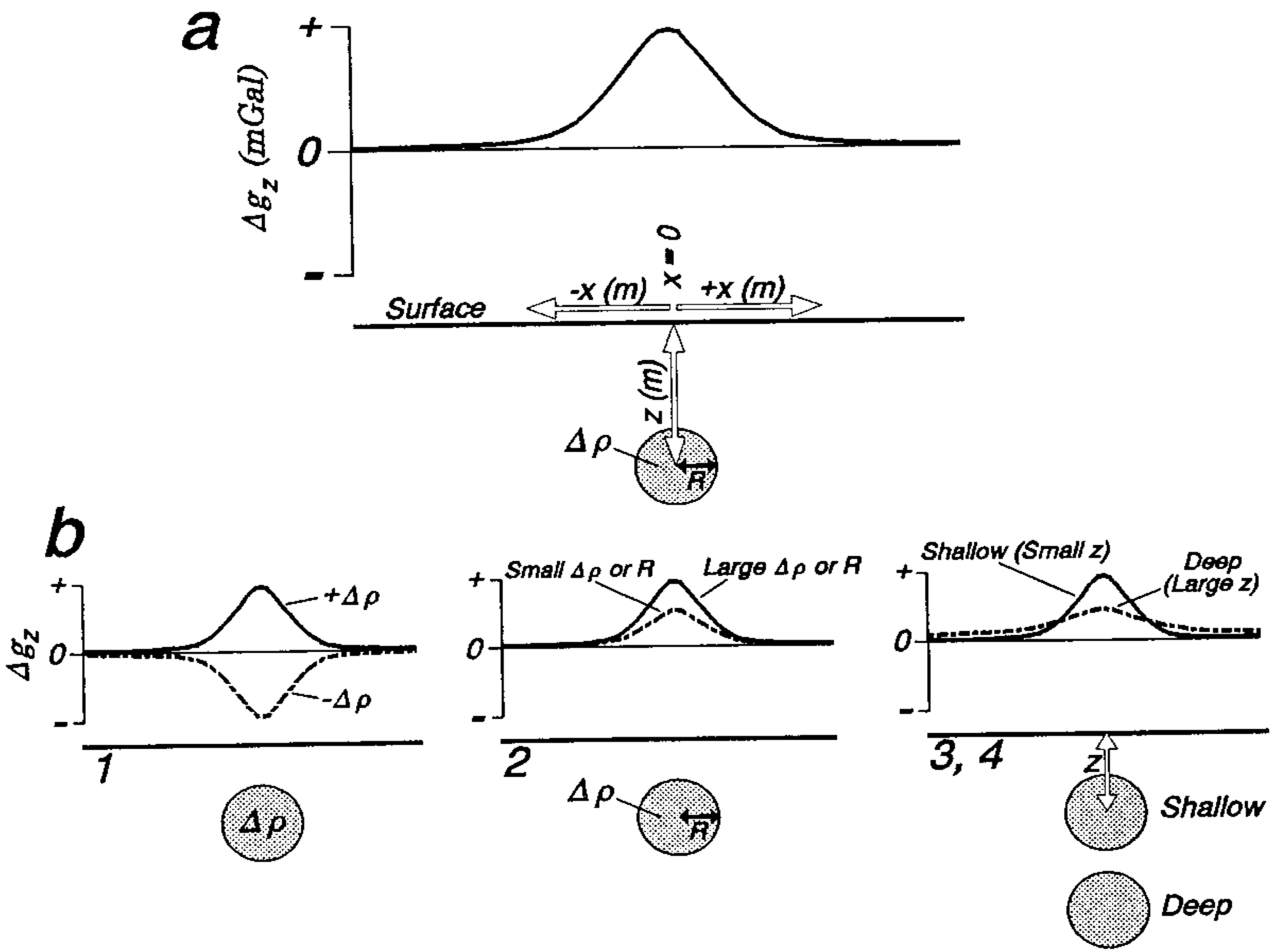


Example of Observed Bouguer Anomaly map. It reflects changes in mass distribution below the surface.

Density of Earth Materials



Gravity Anomaly Over a Buried Sphere



UNIT SEVEN

MAGNETIC METHOD

Basic Definitions

- ***Magnetic survey***: ***Measurements*** *of the magnetic field* or its components at a series of different locations over an area of interest and locating **anomalies** in the Earth's magnetic field.
- *The objective* is **locating** concentrations of magnetic materials or determining depth to basement.
- Most rocks are nonmagnetic, however, certain rock types contain sufficient magnetic minerals to produce significant magnetic anomalies.
- Magnetic methods can be performed on **land**, at **sea** and in the **air**.

Applications of Magnetic Survey

- Archaeological ruins
- Basic igneous dykes
- Metalliferous mineral deposits
- Geological boundaries including faults
- Large-scale geological structures

Magnetic Force

- Charles Augustin de Coulomb, in 1785, showed that the force of attraction or repulsion between electrically charged bodies and between magnetic poles also obey an **inverse square law** like that derived for gravity by Newton.
- The mathematical expression for the **magnetic force** *experienced between two magnetic monopoles* is given by:

$$\boxed{F_m = \frac{1}{\mu} \frac{p_1 p_2}{r^2}}$$

Where: **μ** is a constant of proportionality known as **the magnetic permeability**,

$\mathbf{p_1}$ and $\mathbf{p_2}$ are the *strengths of the two magnetic monopoles*, and r is the *distance between the two poles*.

Magnetic force

- *The expression of* ***magnetic force*** *(F_m) is* identical to the **gravitational force** expression (F_g). **<u>There are two important differences:</u>**
- Unlike the gravitational constant, *G, the* ***magnetic permeability, μ****, is a property of the material in* which the two monopoles, *p_1 and p_2, are located. If they are in a vacuum,* ***μ* is called *the magnetic permeability of free space*.**
- Unlike ***m_1*** *and* ***m_2, p_1 and p_2 can be either positive or negative in sign****. If* ***p_1*** *and* ***p_2*** *have the same sign,* the force between the two monopoles is ***repulsive***. If ***p_1*** *and* ***p_2*** *have opposite signs, the force* between the two monopoles is ***attractive*.**

MAGNETIC FIELD STRENGTH

- The *magnetic field strength, H, is* defined as: *the force per unit pole strength exerted by a magnetic monopole,* p_1.

$$\boxed{H = \frac{F_m}{p_2} = \frac{p_1}{\mu r^2}}$$

- The magnetic field strength *H is analog to the* gravitational acceleration, *g*.

UNITS

- Given the units associated with force, *N, and magnetic monopoles, Amp-m, the unit of the magnetic field strength is Newtons per* Ampere-meter, *N/(Amp - m)*.

N/(Amp - m) is referred to ☞ *Tesla* (T).

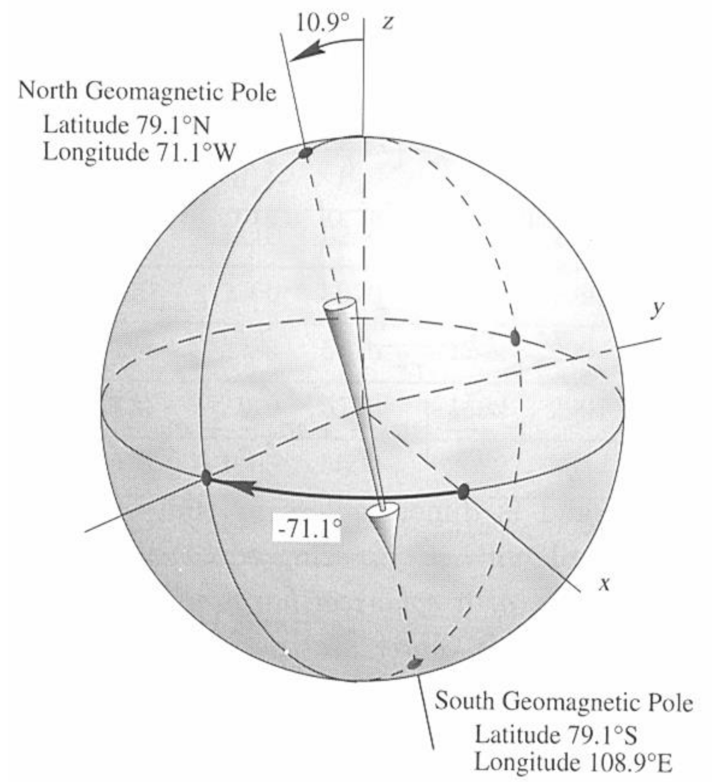
A nanotesla (nT) is referred to ☞ gamma

☞1nT = 10^{-9} T = 1 gamma.

- The average strength of the Earth's magnetic field is about 50,000 nT.

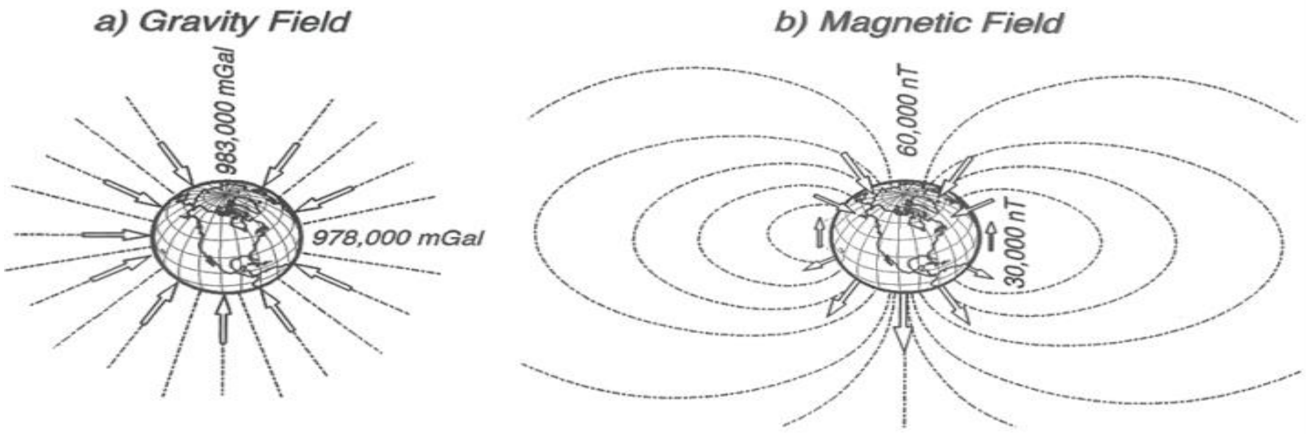
EARTH’S MAGNETIC FIELD

- The *Earth magnetic field originates largely (98%) from within and around the* **Earth's core**. It’s thought to be *caused by motions of liquid metal in the* core.
- The earth's magnetic field can be explained as a dipole at the earth's center, inclined about 10.9° from Earth’s rotational axis dipole.



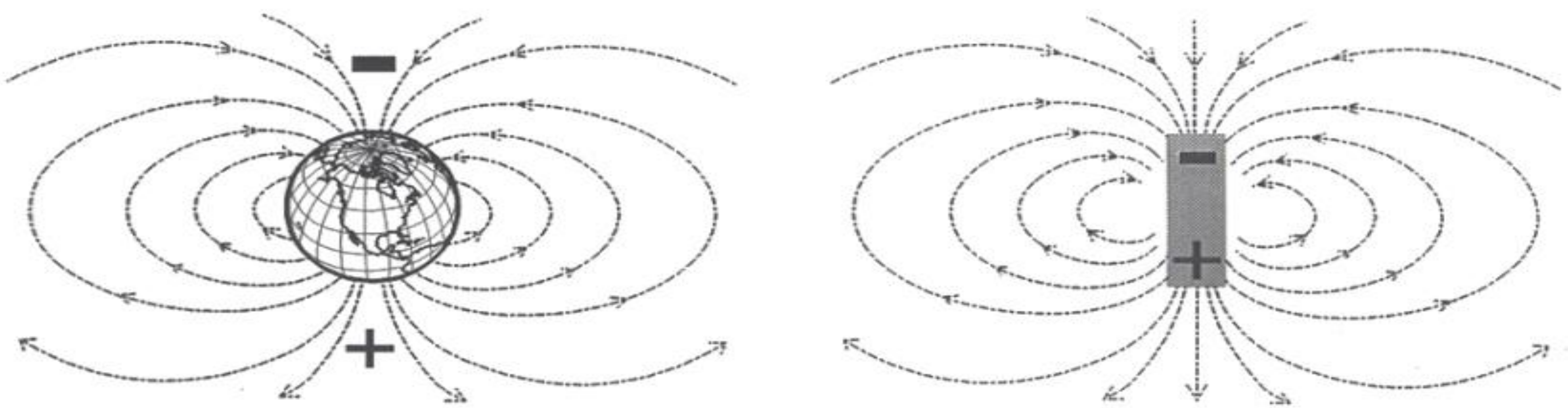
EARTH'S MAGNETIC FIELD

- Relative to the Earth's **gravity field**, the **magnetic field** changes rapidly in both magnitude and direction.
- The **magnetic field** is *Horizontal* near the *equator* and *vertical* near the *poles*. The *strength* at the *poles* is about *twice* as that at the *equator*.



EARTH'S MAGNETIC FIELD

- The earth's magnetic field is similar to that produced by a simple bar magnet placed in the center of the earth.
- The magnetic lines get into the earth from the north pole, and get out of it from the south pole. Thus the positive end of the bar magnet points to the south, and vice versa.



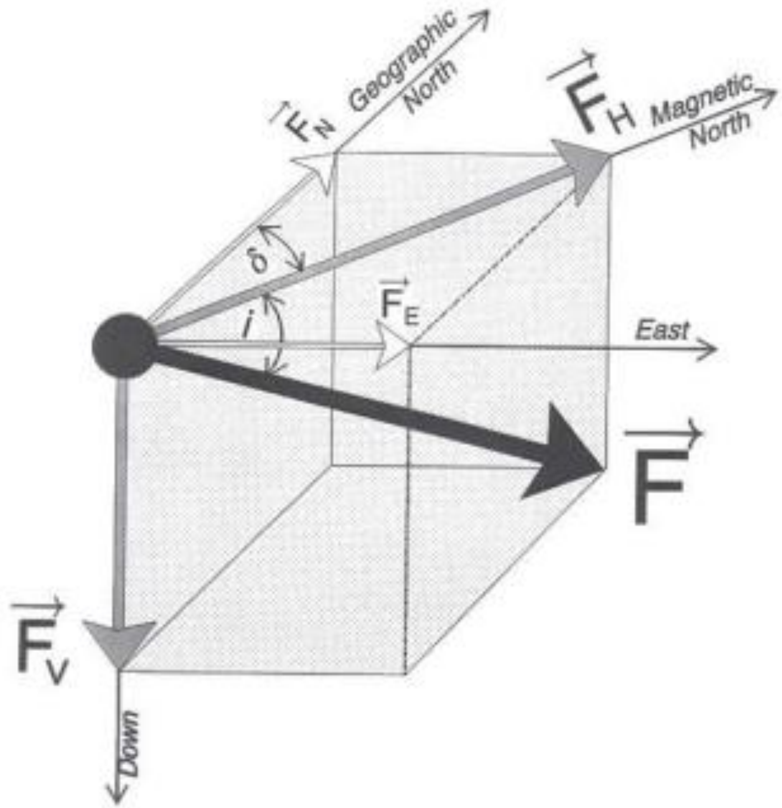
a) Earth b) Bar Magnet

Strength and Direction of Magnetic Field

- The orientation of compass needle indicates the direction of Earth's magnetic field.
- The magnetic field strength at:
  - geomagnetic pole = 60,000 nT
  - equator = 30,000 nT
  - Riyadh = 42,000 nT
- Magnetic field strength varies with latitude.

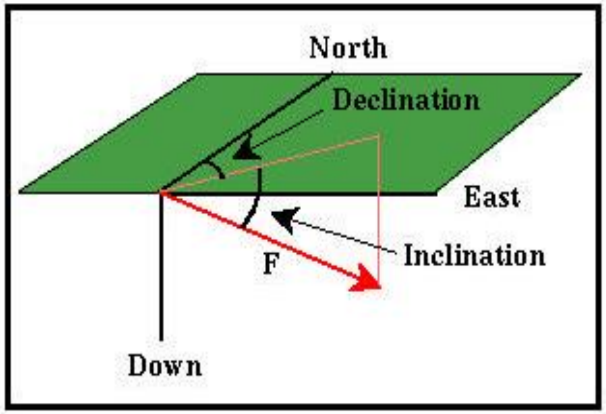
Strength and Direction of Magnetic Field

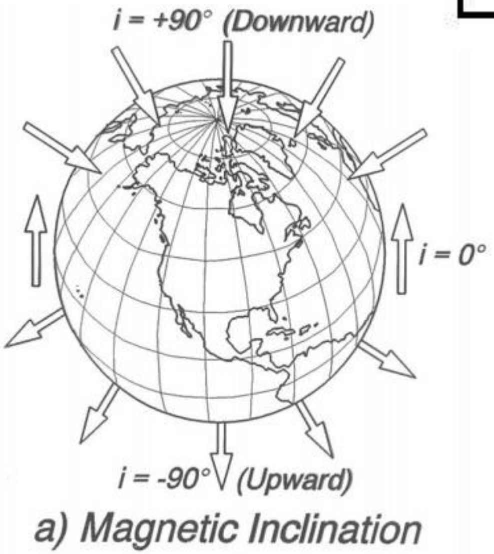
- The geomagnetic field can be described in terms of:
  - ***Inclination, I***
  - ***Declination, δ***
  - ***Total force vector, F***



Magnetic Inclination

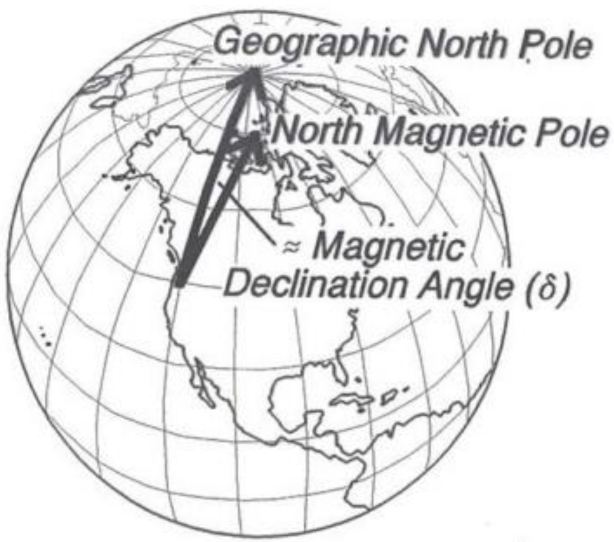
- **Magnetic inclination (i)**:
  The angle between the magnetic line and the horizontal.
- i=0 at the equator, and i=90 at the poles.
- $\tan(i)=2\tan(\Phi)$; where Φ is the geographic latitude



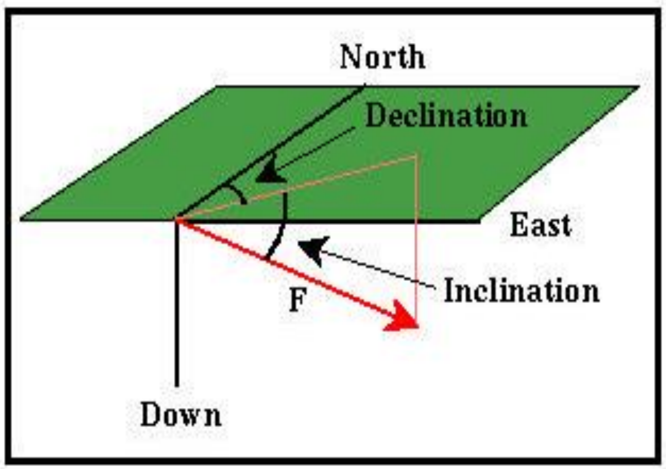


Magnetic Declination

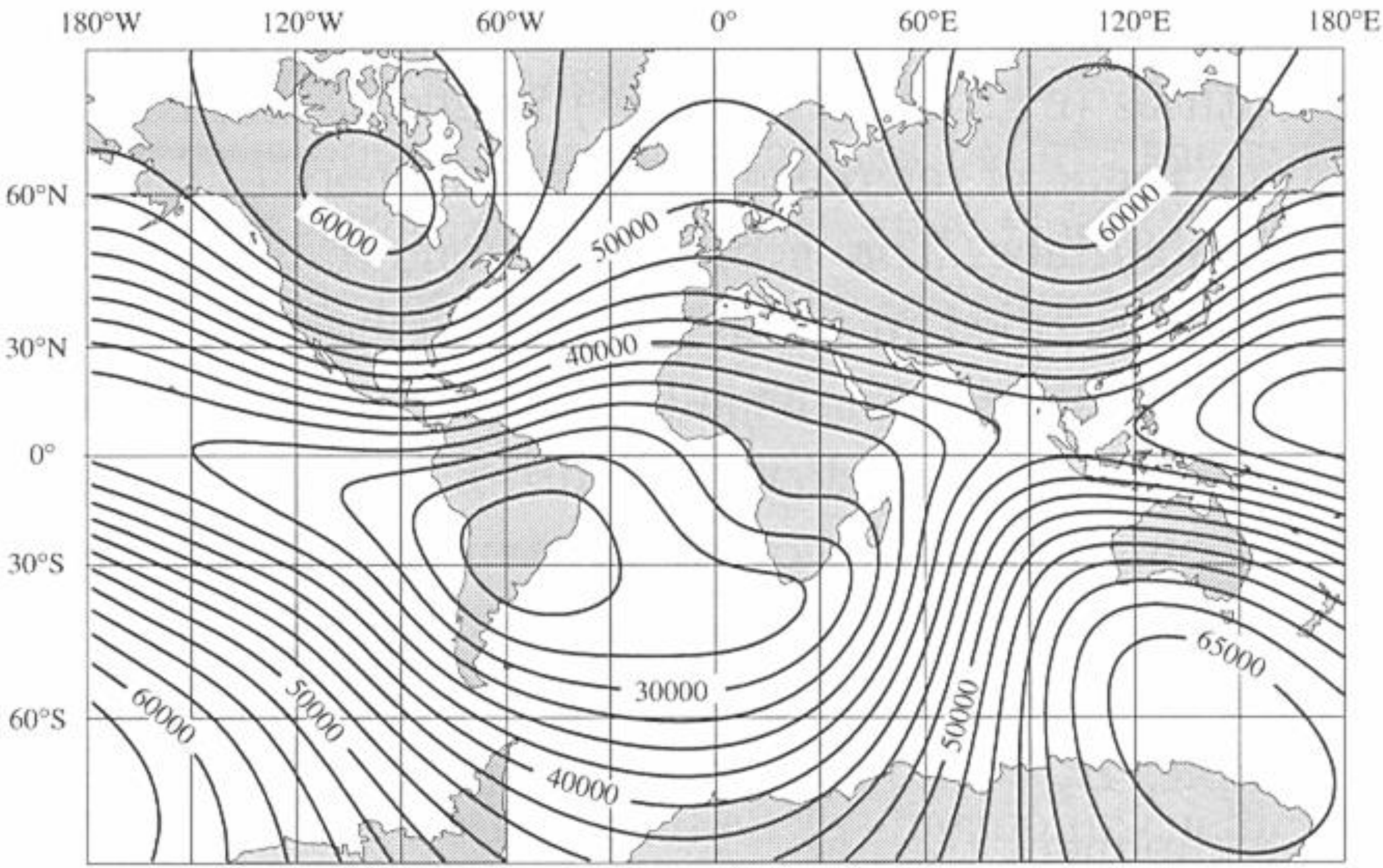
- **Magnetic declination (δ)**:
  The horizontal angle between the local magnetic line and the geographic north.



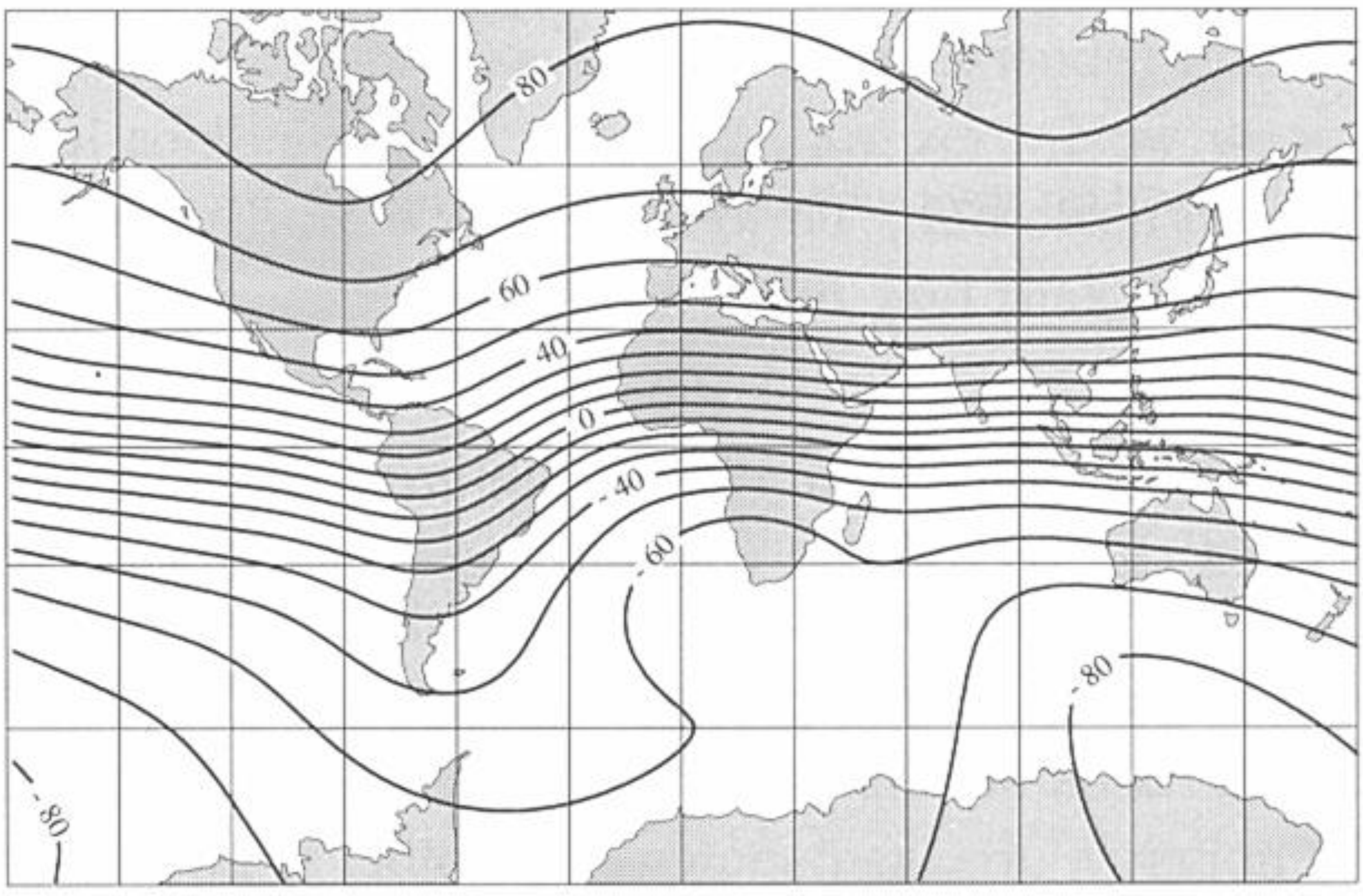
b) Magnetic Declination



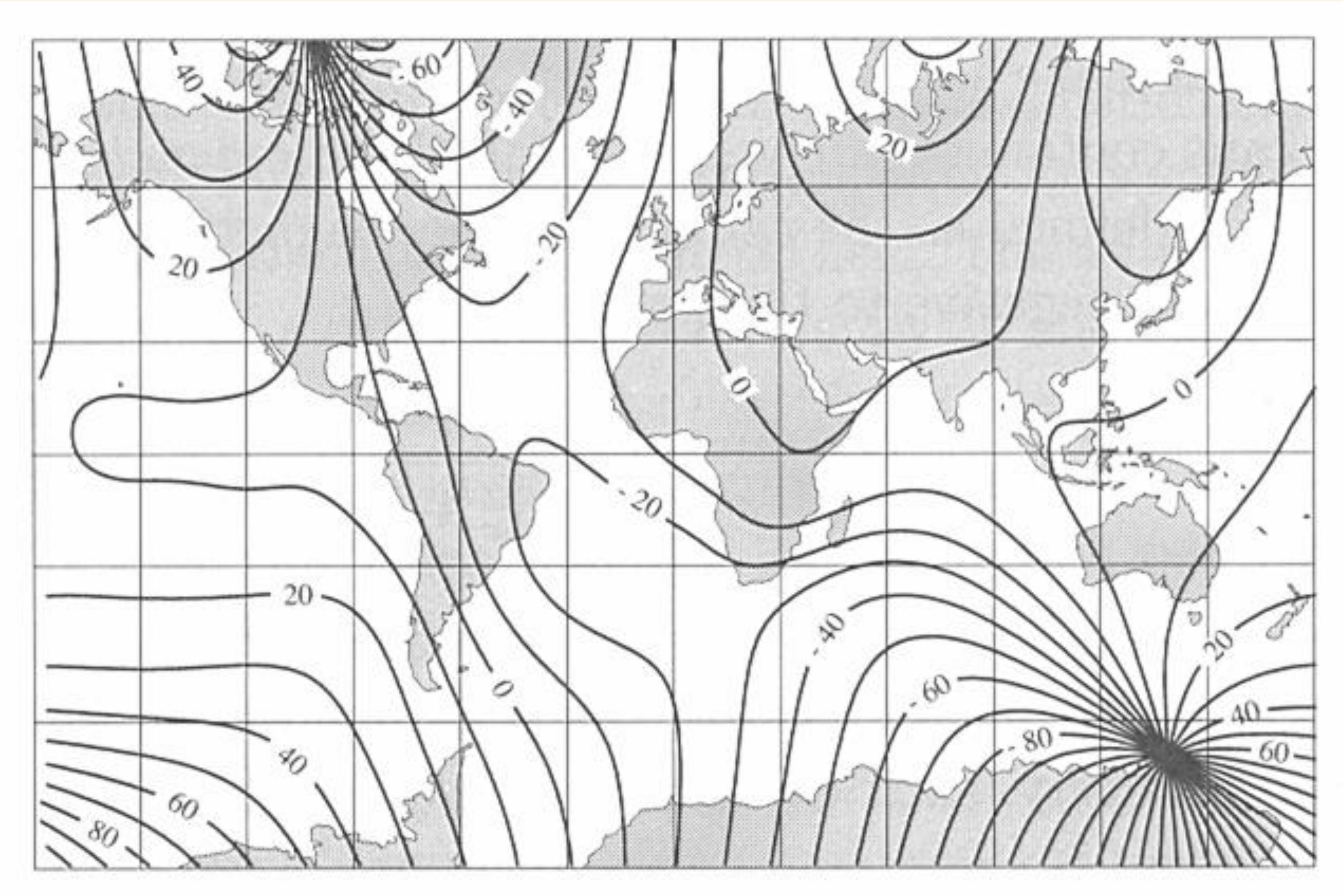
Map of total intensity of Earth's magnetic map based on IGRF 1990, contour: 2,500nT



Map showing constant inclination of total magnetic fiels, contour: 10° (based on IGRF 1990)



Map showing constant declination. Contour interval: 10°



MAGNETIZATION OF ROCKS

Type of magnetization

- Magnetization of rock occurs in two ways:
- It can be induced by Earth's present magnetic field: **Induced magnetization**
- It could have formed some time in the past: **Remnant magnetization**.

MAGNETIZATION OF ROCKS

- $F_{observed} = F_{reference} + F_{local}$

- $F_{local} = F_{induced} + F_{remanent}$

- **Induced part, $F_{induced}$** : proportional to the ambient magnetic field (present Earth's magnetic field) and depends on the susceptibility.
- **Remanant part, $F_{remanent}$** : remains unchanged if there is no field present and is <u>independent of ambient magnetic field</u>. It has formed some time in the past.
- The magnitude is very variable, on the scale of 1000nT.

MAGNETIZATION OF ROCKS
Induced magnetization

- If a body is placed within an external magnetic field (H), the body acquires a magnetization (I), with intensity proportional to the overall magnetic susceptibility (k) of the body.

MAGNETIZATION OF ROCKS

Intensity of induced magnetization

- The strength of the magnetic field induced by the magnetic material due to the inducing field is called the *intensity of magnetization, I.*
- The magnitude and direction of magnetization induced within a material depends on the magnitude and direction of the external (ambient) field (H) and the ability of the material to be magnetized.

$$I = k\,H$$

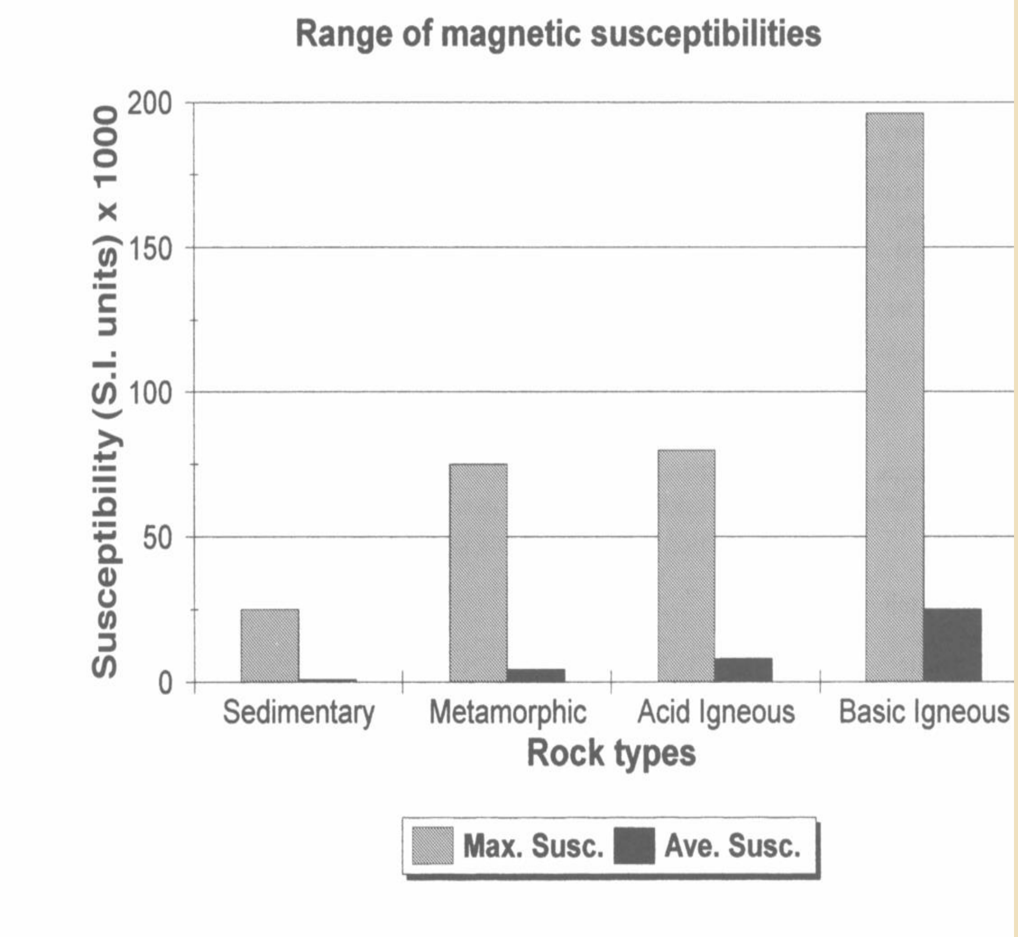
I: Intensity of the magnetization of the material (**Induced magnetization**).

K: magnetic susceptibility of the material

H: magnitude of the ambient field (Earth's field).

MAGNETIZATION OF ROCKS

Magnetic Susceptibility



The intensity of magnetization, *I, is related to the strength of the inducing* magnetic field, *H, through a constant of proportionality,* ***K****, known as the* ***magnetic*** **susceptibility.**

The magnetic susceptibility (k, a dimensionless quantity) is a measure of the **degree to which a substance may be magnetized.**

MAGNETIZATION OF ROCKS

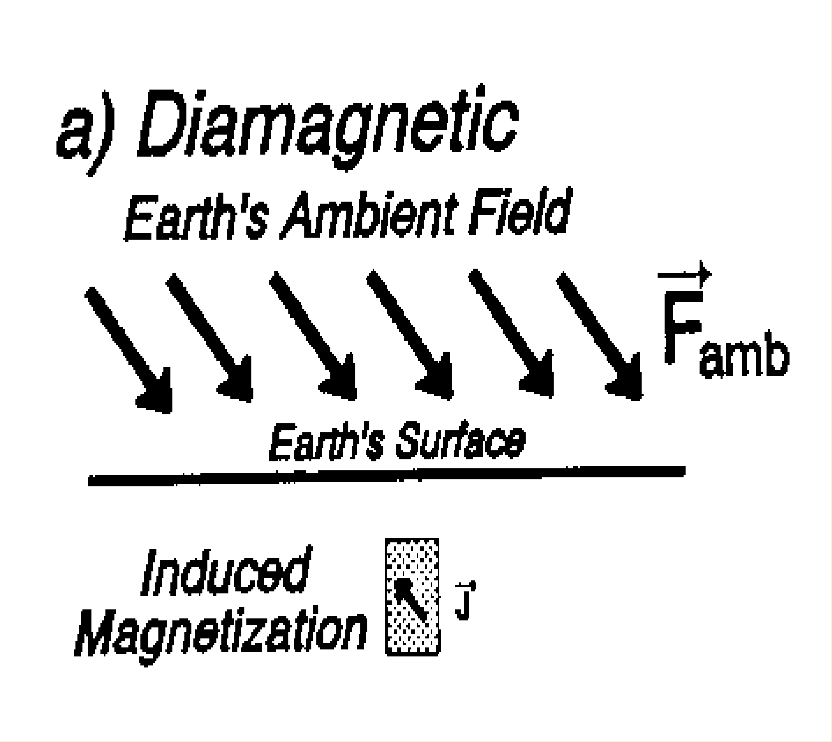
Type of Magnetic Behavior

- The type of magnetism exhibited by a body depends on the mineral's magnetic susceptibility.

  - Diamagnetic material,
  - Paramagnetic material,
  - Ferromagnetic material.

MAGNETIZATION OF ROCKS

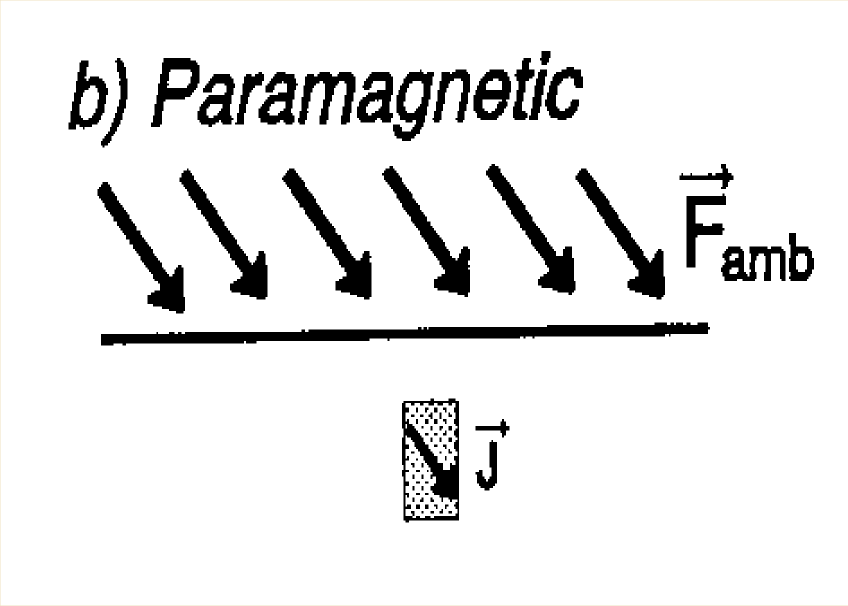
Diamagnetic material, k ~ -10 $^{-4}$



The **Diamagnetic mineral,** such as halite (rock salt) has negative and low magnetic susceptibility. The body acquires a weak magnetization and opposite to the external Field.

MAGNETIZATION OF ROCKS

Paramagnetic material, k ~ +10 $^{-4}$

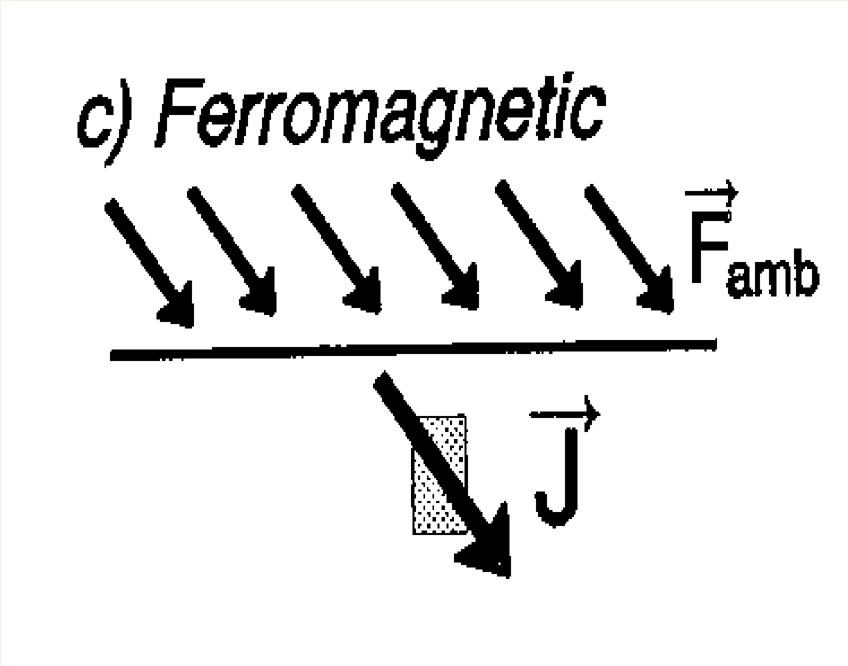


The magnetic susceptibility of **paramagnetic minerals** is positive.

The magnetization in paramagnetic is weak but in the same direction as the external field.

MAGNETIZATION OF ROCKS

Ferromagnetic material, k ~ +10 $^{-1}$



In some metallic minerals rich in iron, cobalt, manganese and nickel, atomic magnetic moments align strongly with external field. Susceptibility on the order of 10^{-1} indicate that the magnetization in the same direction as, and about 1/10 the magnitude of the external field. In this case we have a strong magnetization.

Under some circumstances, induced magnetization may remain in ferromagnetic materials, even after the external field is removed (remnant magnetization).

Magnetic Measurement Instruments

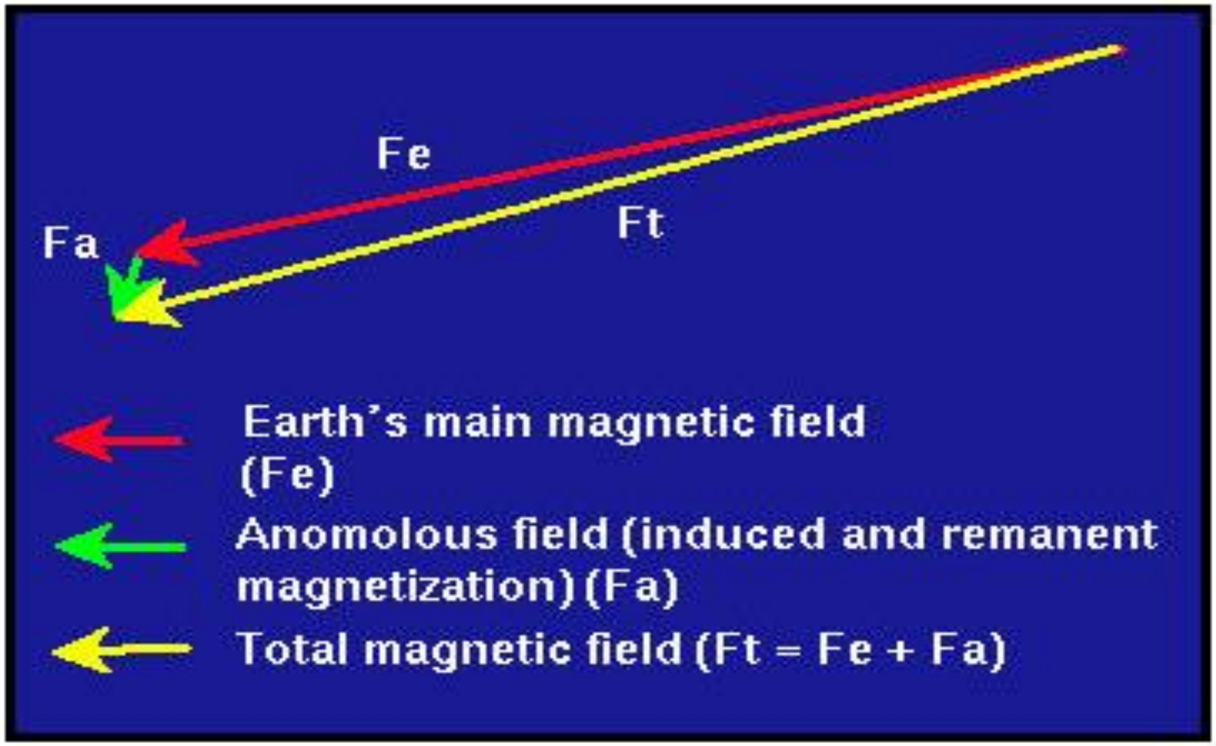
- Three types of magnetometers are frequently used in magnetic surveying. These are:
  - ➢ Proton magnetometer
  - ➢ Cesium vapor magnetometer
  - ➢ Fluxgate magnetometer

TOTAL FIELD MEASUREMENTS

- Using Proton procession magnetometer, we measure only the magnitude of the total magnetic field as a function of position.

- Surveys conducted using the proton precession magnetometer do not have the ability to determine the direction of the total field as a function of location.

Magnetic Anomaly

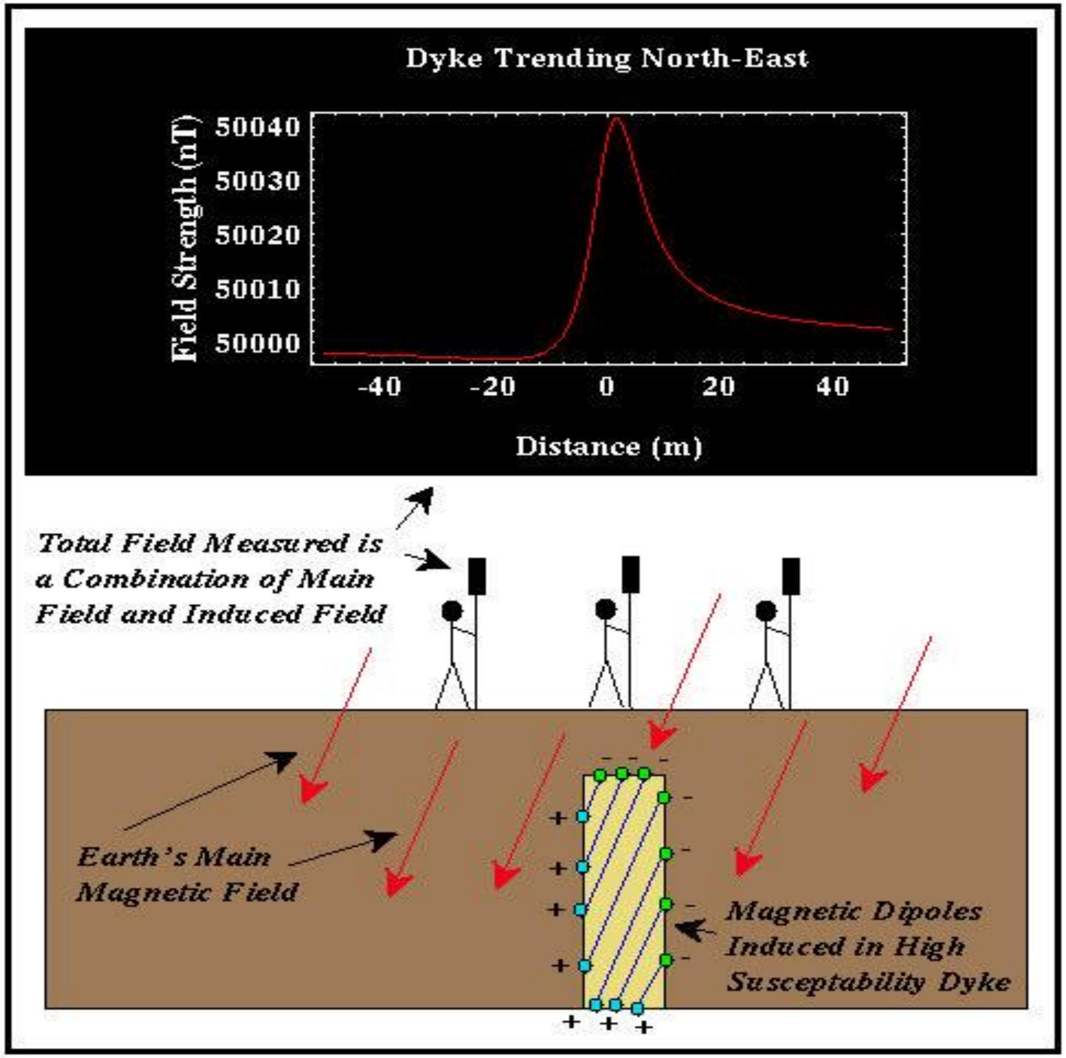


As F_t is almost parallel to F_e, the observed magnetic anomaly is approximated as follow:

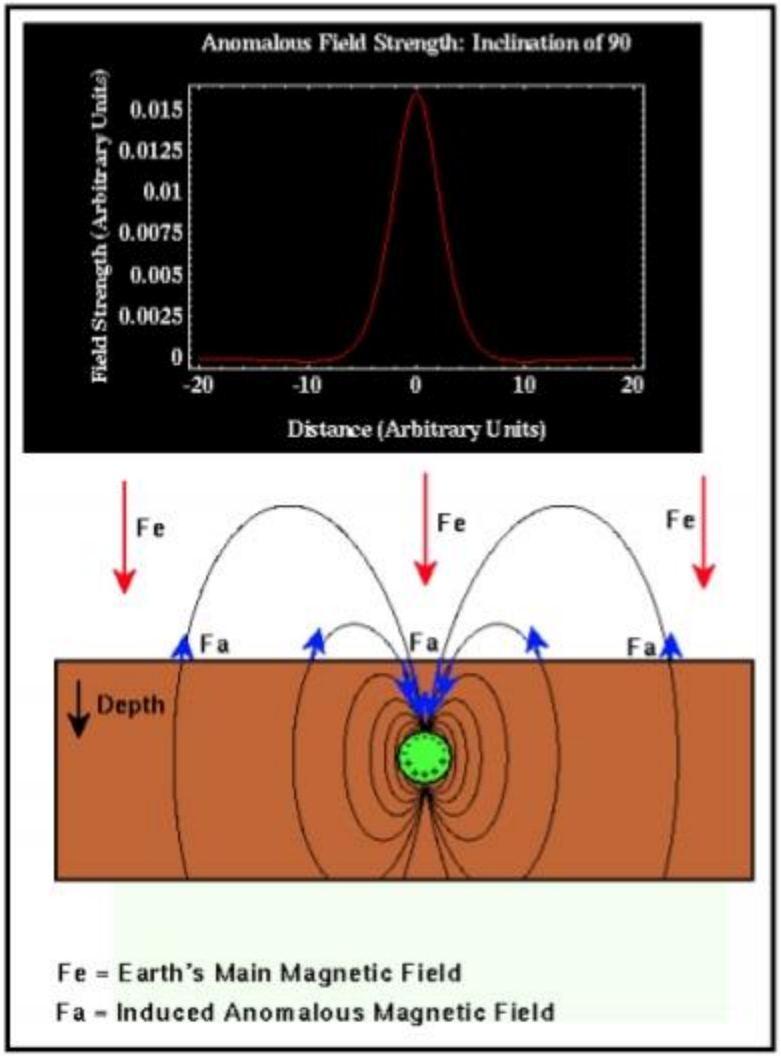
Observed Anomaly $\Delta F = F_t - F_e$

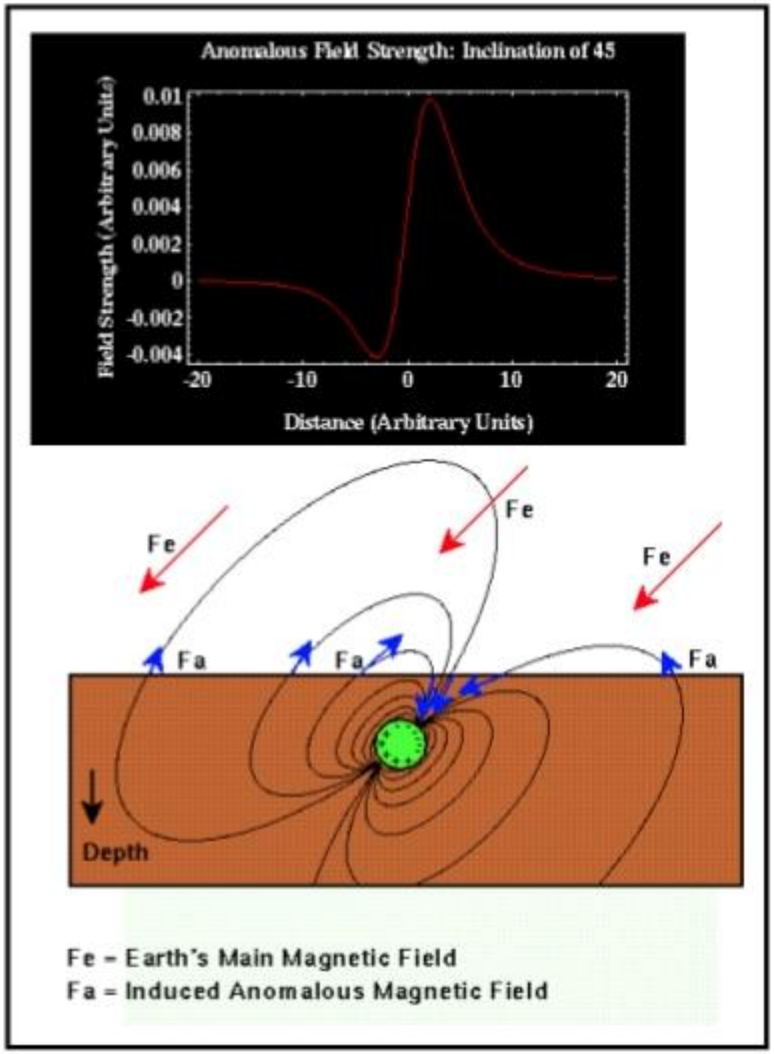
ΔF = the component of F_a parallel to F_e

MAGNETIC SURVEYING & ANOMALIES



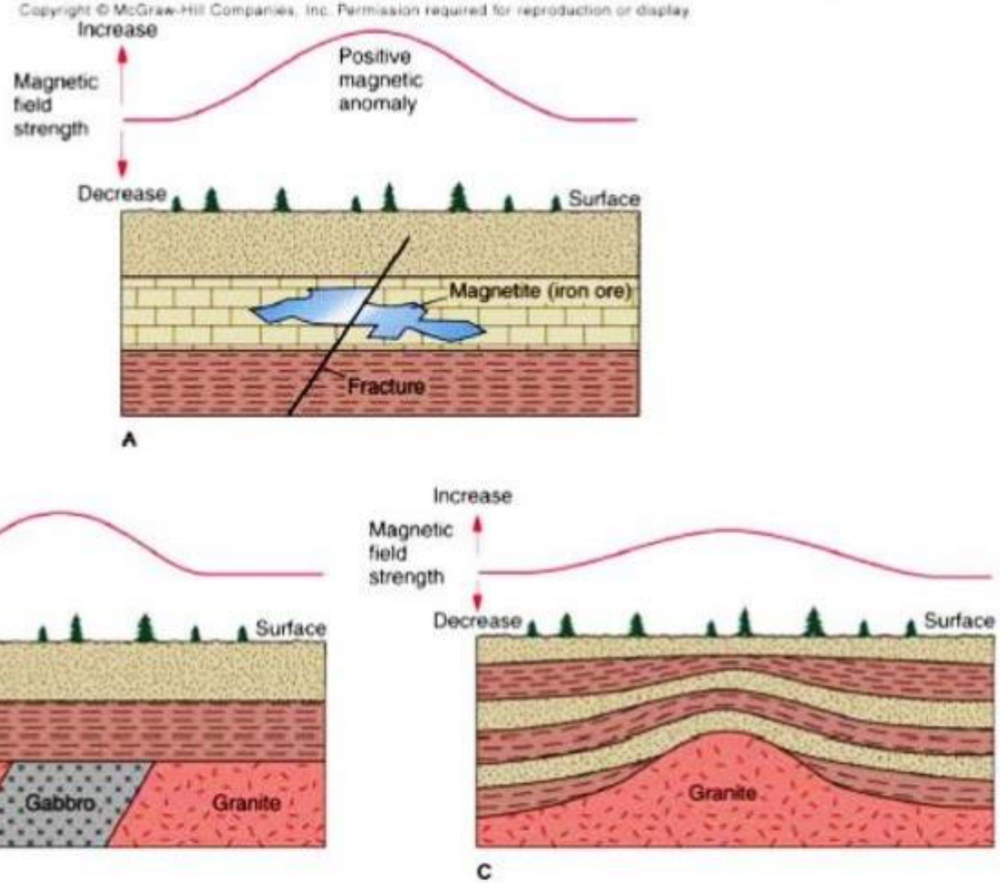
MAGNETIC SURVEYING & ANOMALIES





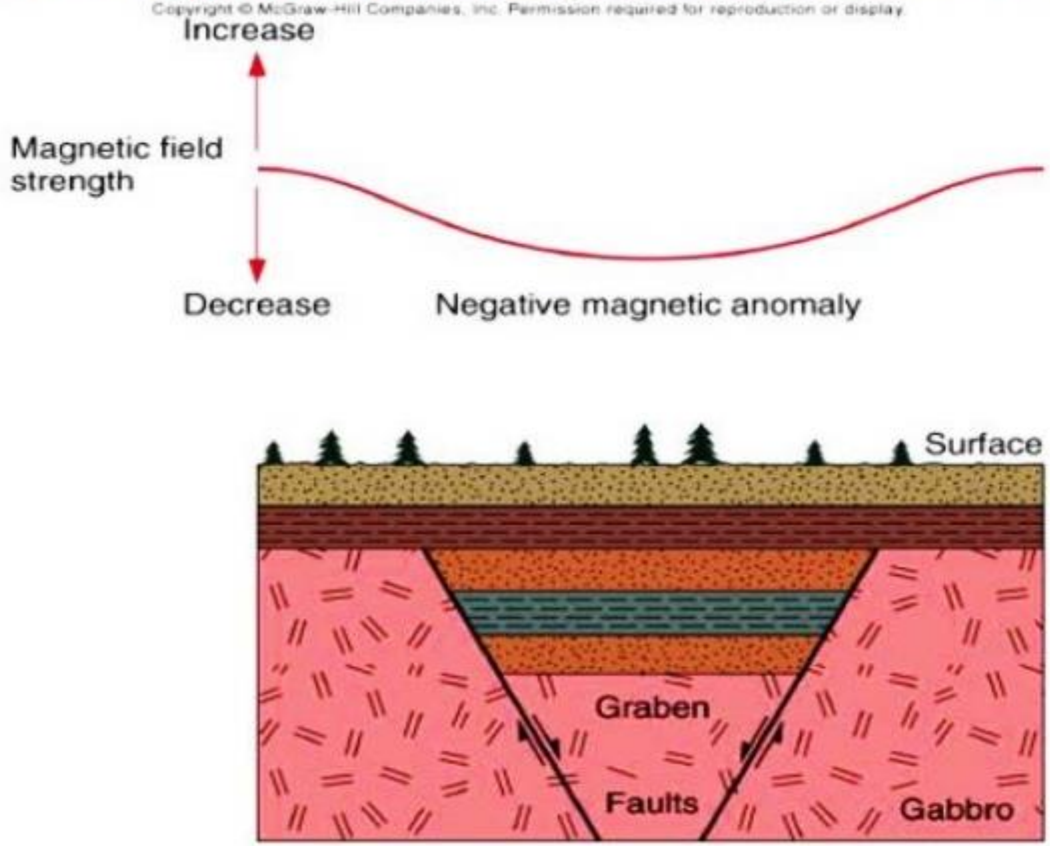
MAGNETIC SURVEYING & ANOMALIES

magnetic anomalies occur in local field from magnetic rock below surface (similar to gravity anomalies)



MAGNETIC SURVEYING & ANOMALIES

removal of magnetic material from near surface causes negative anomaly (example is normal faulting)



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