Iraqi Kurdistan Region
Tishk International University (TIU)
Faculty of Engineering
Department of Petroleum and Mining Engineering
Erbil

Lecture 1

28/1/2024

Applied Geophysics Lecture Notes
Fourth Semester
By:

Dr. Fadhil Ali Ghaib

2023/2024

Contents:

- Course Information and description
- Assignment and Exam. Information
- Introduction to applied geophysics
- Geophysical surveying

Text Books:

1-Keary P. and Brooks, M. (1986, 1991): An Introduction to Geophysical Exploration.
Link: An Introduction to Geophysical Exploration.pdf

2-Reynolds, J. M. (1997): An Introduction to Applied and Environmental Geophysics.
Link: an introduction to applied and environmental geophysics- j.m reynolds.pdf

3-Dobrin, M. (1974, 1983): An Introduction to Geophysical Prospecting.
Link: Introduction To Geophysical Pr-dobrin.pdf

Course Information

Course Code and Name: Applied Geophysics PTR 228

Weekly Hours: Theoretical: 2Hrs, Practical:2Hrs

Lecturer Name And Department: Fadhil Ali- PETM

Office: Room 225

Student’s handbook link
https://tiu.edu.iq/student-handbook/

Course Description:

The course is to obtain a good understanding of the definition and application of geophysics in mineral exploration, petroleum exploration, environmental investigations, groundwater exploration and monitoring, and geotechnical studies. In our case here the science of geophysics will be oriented toward the application in oil explorations. It includes some simple theoretical bases and the application of two methods which are used mainly in oil exploration. The field activity will be discussed and how to correct, process and interpret data in such a way to be in benefits for petroleum engineers

Assignment and Exam. Information

1- Problems solved inside the room and a quiz for theoretical side of problems.
15% (10+5)
2- Quizzes, at least two for theoretical lectures,
20%
3- Midterm (including all pre-exam lectures + problems).
20%
4- Attendance 5%
5- Final (including all given lectures, theoretic and practical "problems"
40% (10 Practical + 30 Theoretical)
TOTAL MARKS 100%

Attention: Cheating may destroy your future

- Exercises will involve a degree of collaboration with fellow students. The policy on this issue is that while you may discuss laboratory problems with other students, the actual material submitted must be your independent work, unless otherwise indicated. We expect each student to complete his own calculations, to draw own graphs *etc.*. We expect student to have developed his/her own answers. **Each student should have his own drawing and calculating equipment such as calculator, ruler, graph paper *etc*.**

Important notes:

a. Be prepared for class

b. The ones that do the best in the class are those that attend regularly

c. Turn off your cell-phone device or silent it and put it in your pocket or bag.

d. Pay attention and write everything in the class. You are responsible of every- thing stated by the lecturer in the class even if it is not present in the lecture slides.

e. Ask questions and participate discussions.

f. Review and compare your notes with your colleagues.

g. Utilize textbooks and web resources.

h. Do your own work by yourself.

i. Use office hours (Not just before the exams).

Objectives applied geophysics

THE objects of a geophysical survey (application of geophysics) are to locate subsurface geological structures or bodies and where possible to estimate their dimensions and relevant physical properties. In oil prospecting, structural information is sought because of the association of oil with features such as anticlines in sedimentary rocks. In mining geophysics, the emphasis is on detection and determination of physical properties. Though minerals ore bodies give distinctive and measurable geophysical indications they are often of irregular shape and occur in rocks of complex structure, making precise quantitative interpretation difficult or impossible.

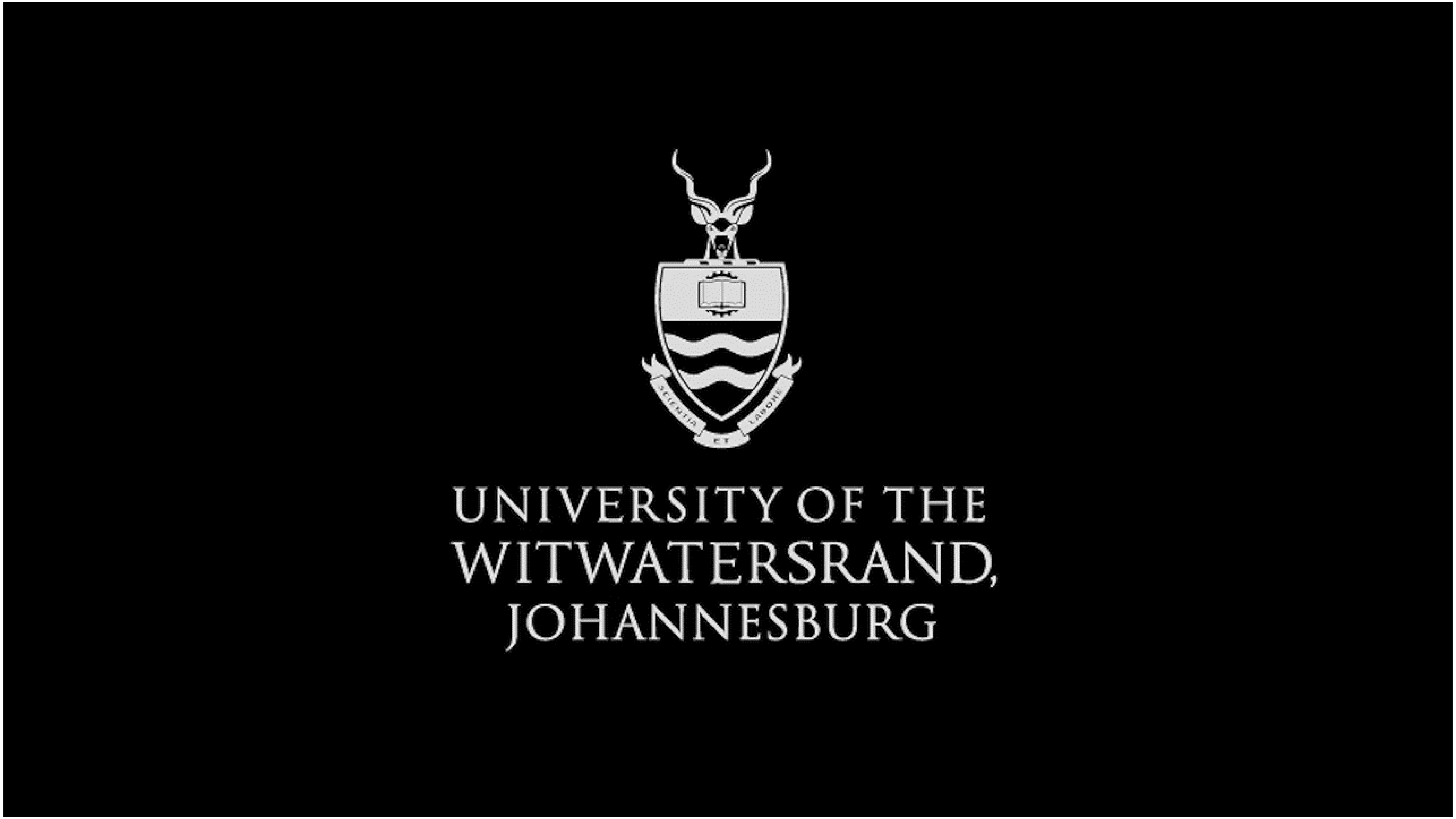
In site investigation, engineers may be interested in both structure and physical properties. Variations in bedrock depth are often needed on major construction sites and the mechanical properties of the overburden may be important when heavy loads must be sustained.

A geophysical survey consists of a set of measurements, usually collected to a systematic pattern over the earth's surface by land, sea or air, or vertically in a borehole.

In many instances more than one method should be used to survey the same ground. The search for oil may start with gravity and airborne magnetic work as a preliminary before seismic exploration.

By working at different scales, geophysical methods may be applied to a wide range of investigations from studies of the entire Earth to exploration of a localized region of the upper crust for engineering or other purposes. In the geophysical exploration methods (also referred to as geophysical surveying) discussed in this course, measurements within geographically restricted areas are used to determine the distributions of physical properties at depths that reflect the local subsurface geology

Importance of geophysics: https://youtu.be/gRramwK6ff8?list=PLYZ6d6uICinPb_F8u7QGMYJMXII3zSB4Z



Geophysical surveying

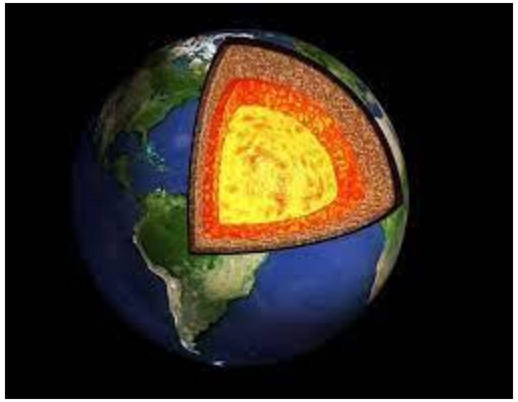
- Geophysical surveying methods are divided into those that make use of natural fields of the Earth (passive methods) and those that require the input into the ground of artificially generated energy (active methods). The first, utilizes the gravitational, magnetic, electrical and electromagnetic fields of the Earth itself. The second involves the generation of local electrical or electromagnetic fields or generation of seismic waves whose propagation velocities and transmission paths through the subsurface are mapped to provide information on the distribution of geological boundaries at depth.
- Generally, natural field methods can provide information on Earth properties to significantly greater depths and are logistically simpler to carry out than artificial source methods. The latter, however, are capable of producing a more detailed and better resolved picture of the subsurface geology.

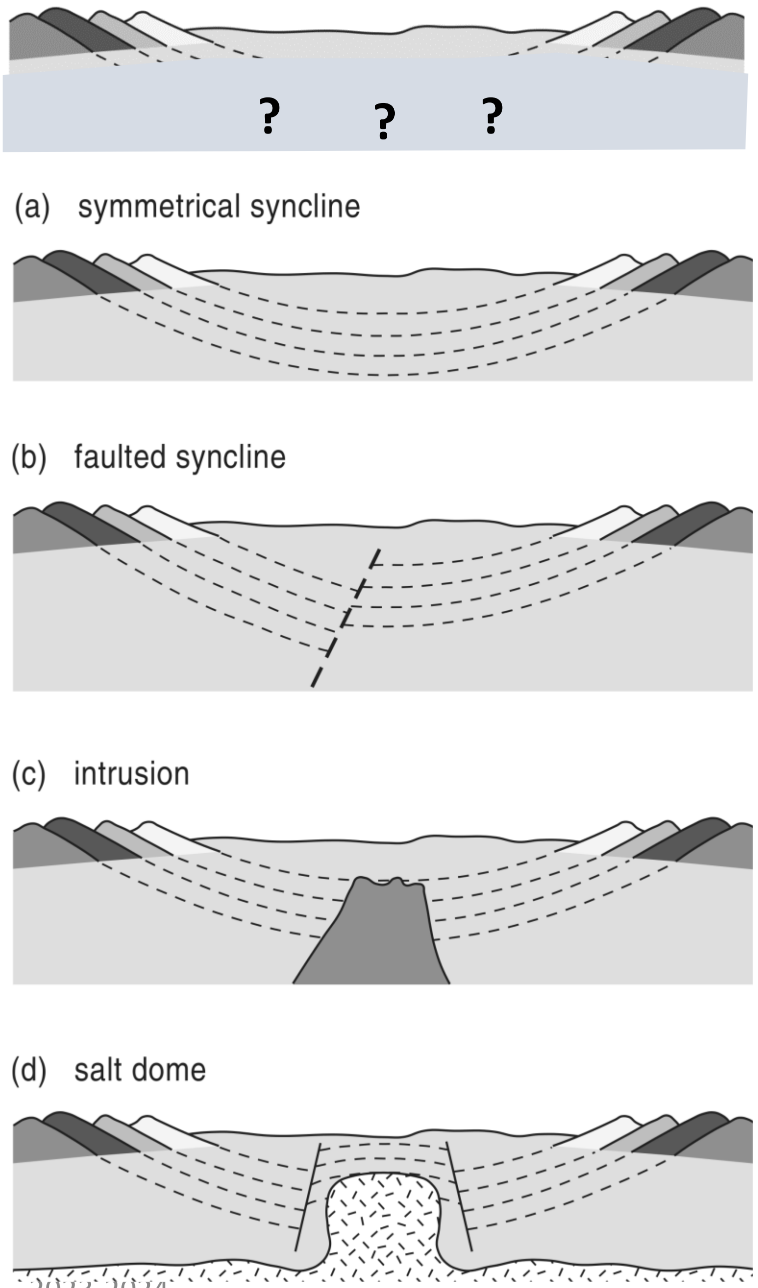
Geology and Geophysics:

Geology involves the study of the earth by direct observations on rocks either from surface exposures or from boreholes and the deduction of its structures, composition and historical evolution by analysis of such observations.

Geophysics involves the study of the inaccessible earth by means of physical measurements, usually on or above the ground surface.

Hence, Geophysics is the third dimension of geology





- **Surface geology to be interpreted for subsurface picture.**

- **Many types of structures can be imagined by geologists.**

- **Geophysics can end the argument**

Physical properties of rocks as used in different geophysical methods are:

Physical property	Geophysical method
- Density	Gravity method
- Magnetic susceptibility	Magnetic method
- Elasticity (wave velocity)	Seismic method
- Electrical resistively or conductivity	Electrical and electromagnetic methods
- Radioactivity	Radiometric method
-Thermal conductivity	Geothermal method

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Lecture 2

8/2/2024

Applied Geophysics Lecture Notes
Fourth Semester
By:

Dr. Fadhil Ali Ghaib

2023/2024

Contents:

- Geophysical surveying applications
- The use of geophysics in mining exploration
- Geophysical methods
  (Magnetic Method)
- Applications
- Elements
- Source of the Geomagnetic field
- Variations of geomagnetic field

Geophysical surveying applications

Applications	Appropriate survey methods*
Exploration for fossil fuels (oil, gas, coal)	S, G, M, (EM)
Exploration for metalliferous mineral deposits	M, EM, E, SP, IP, R
Exploration for bulk mineral deposits (sand and gravel)	S, (E), (G)
Exploration for underground water supplies	E, S, (G), (Rd)
Engineering/construction site investigation	E, S, Rd. (G), (M)
Archaeological investigations	Rd, E, EM, M, (S)

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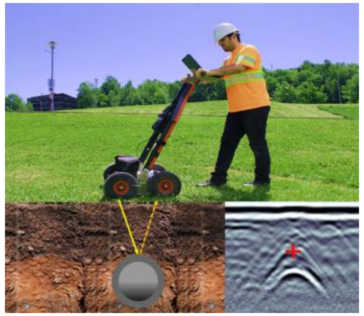
* G, gravity; M, magnetic; S, seismic; E, electrical resistivity; SP, self-potential; IP, induced polarization; EM, electromagnetic; R, radiometric; Rd, ground-penetrating radar. secondary methods in brackets.



- Geophysics definition:
https://youtu.be/n8bE9PPPaok

The use of geophysics in mining exploration:

Geophysical tools were applied to mineral exploration almost three centuries before geophysics was used in oil exploration. The magnetic needle of compass was used since 1640 in Sweden.

The principal metal found in massive sulfide ore bodies are, copper, Nickel, Lead and Zink. They are found in many minerals (there is difference between mineral and metal). These minerals are characterized by high conductivity, high density and sometimes high magnetic susceptibility. That makes Resistivity, Electromagnetic, Gravity and Magnetic methods suitable for exploration for those minerals.

Iron ores are of greatest economic interest contain. Magnetite and Hematite. Magnetite has the highest magnetic susceptibility among all minerals. Hence magnetic method is used in magnetite explorations. Also, gravity method can be used because of their high density relative to surrounding.

geophysical methods

1- Magnetic method:

It is based on the study of the natural magnetic field of the Earth and its relation to the subsurface geology. When materials have magnetic properties, they distort the geomagnetic field. These field distortions are interpreted into geology.

Magnetic method is a potential (passive) method , it resembles gravity. Oftenly, magnetic and gravity methods are applied conjunctionally before seismic methods.
Why?

Applications:

Locating

– Pipes, cables and metallic objects

– Buried military ordnance (shells, bombs, etc.)

– Buried metal drums of contaminated or toxic waste

– Concealed mineshafts and adits

Mapping

– Archaeological remains

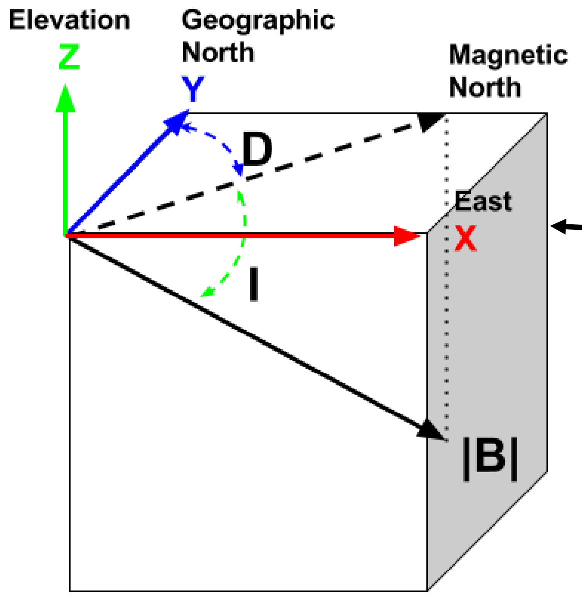
– Concealed basic igneous dykes

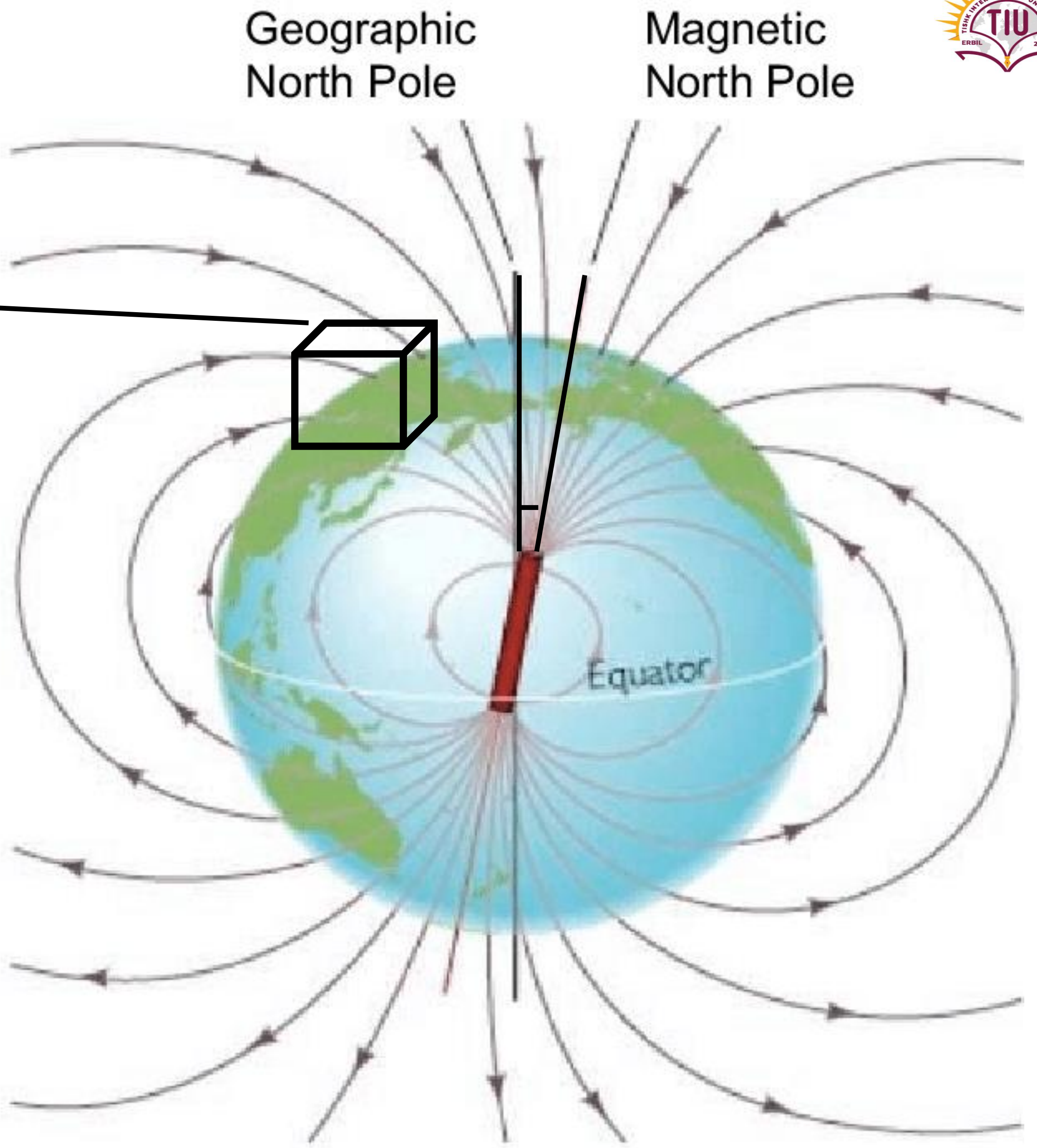
– Metalliferous mineral lodes

– Geological boundaries between magnetically contrasting lithologies, including faults

– Largescale such as regional basins within basement rocks which are important is oil ecploration.

Geomagnetic elements





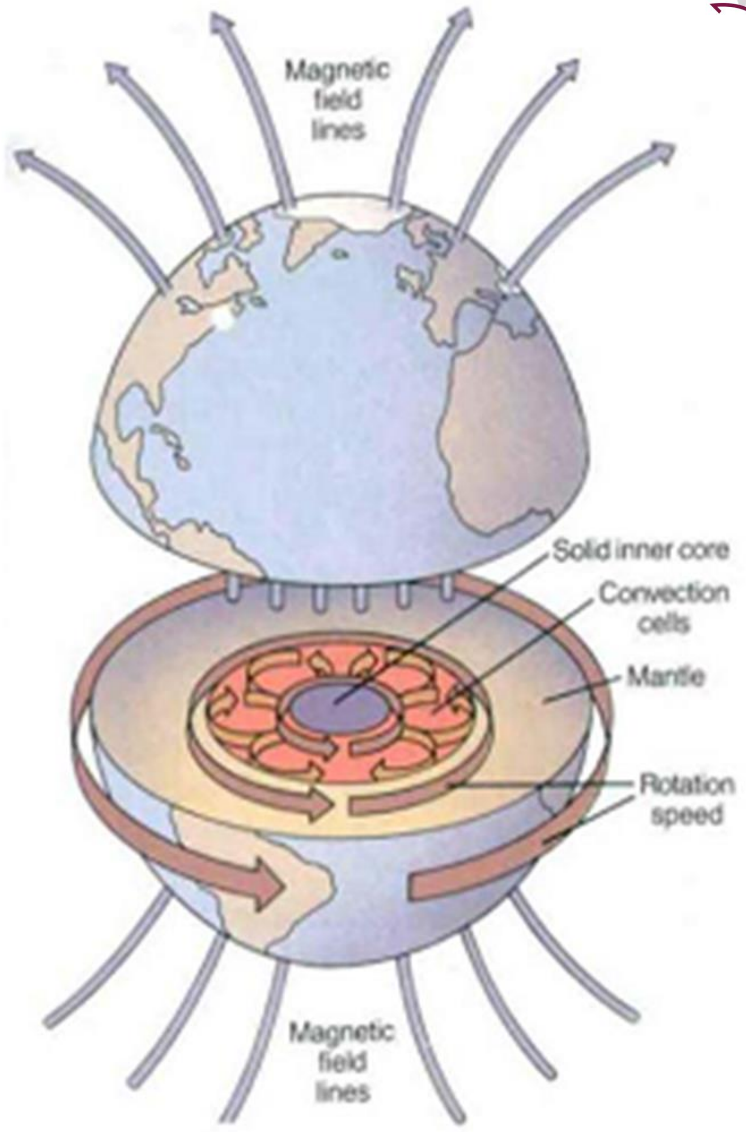
The Earth as a huge magnet

- **magnetic field**
- **direction**
- **Intensity**
- **North-pole**
- **South pole**

Source of the Geomagnetic field:

Geodynamo: the conversion, within the Earth, of mechanical energy (convection of metals) to electrical energy which produces the magnetic field. A magnetic field produced by such fluid motion is inherently unstable and not as uniform as about a simple bar magnet.

Dynamo action produced by the circulation of charged particles in couples convective cells within the outer, fluid, part of the Earth's core.



Variations of geomagnetic field

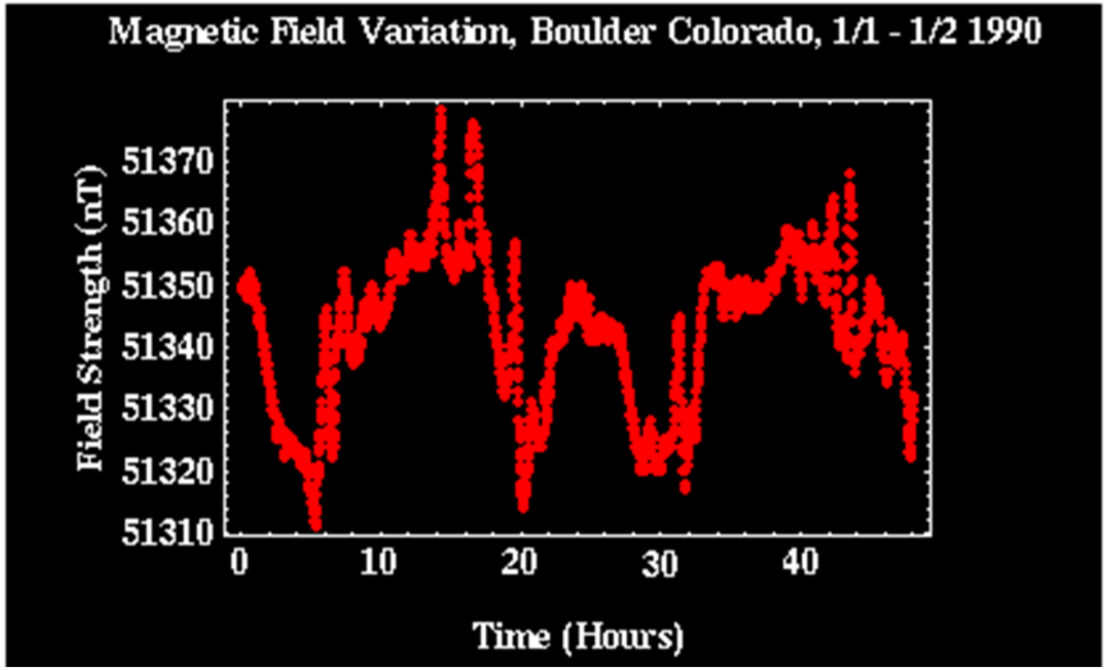
Magnetic field of the earth change during time in both intensity and direction
Magnetic readings taken at the same location at different times will NOT yield the same results.

- Temporal magnetic variations
Temporal variations are classified according to the rate of occurrence and source to:

- Secular variations: through years (tens and hundreds)
Slow changes in magnetic north over time. Shown below are the declination and inclination of the magnetic field around Britain from the years 1500 through 1900.
Hence some corrections should be carried out.

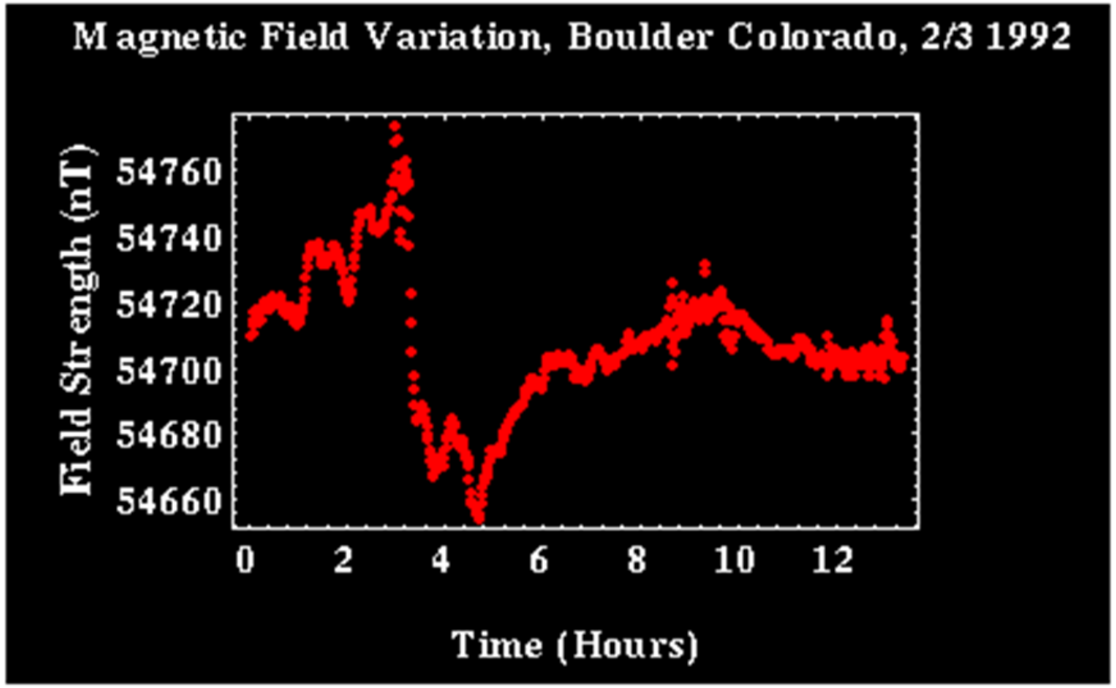
Variations during:

Quiet day



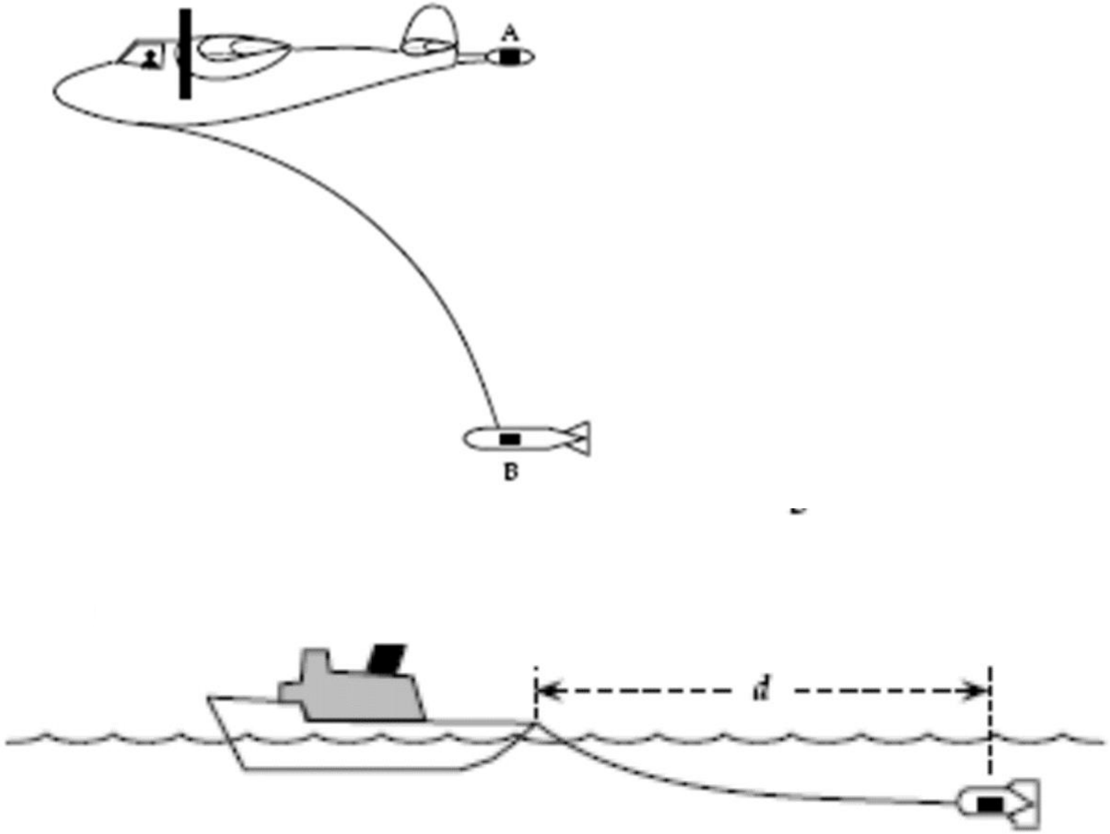
and

Disturbed day

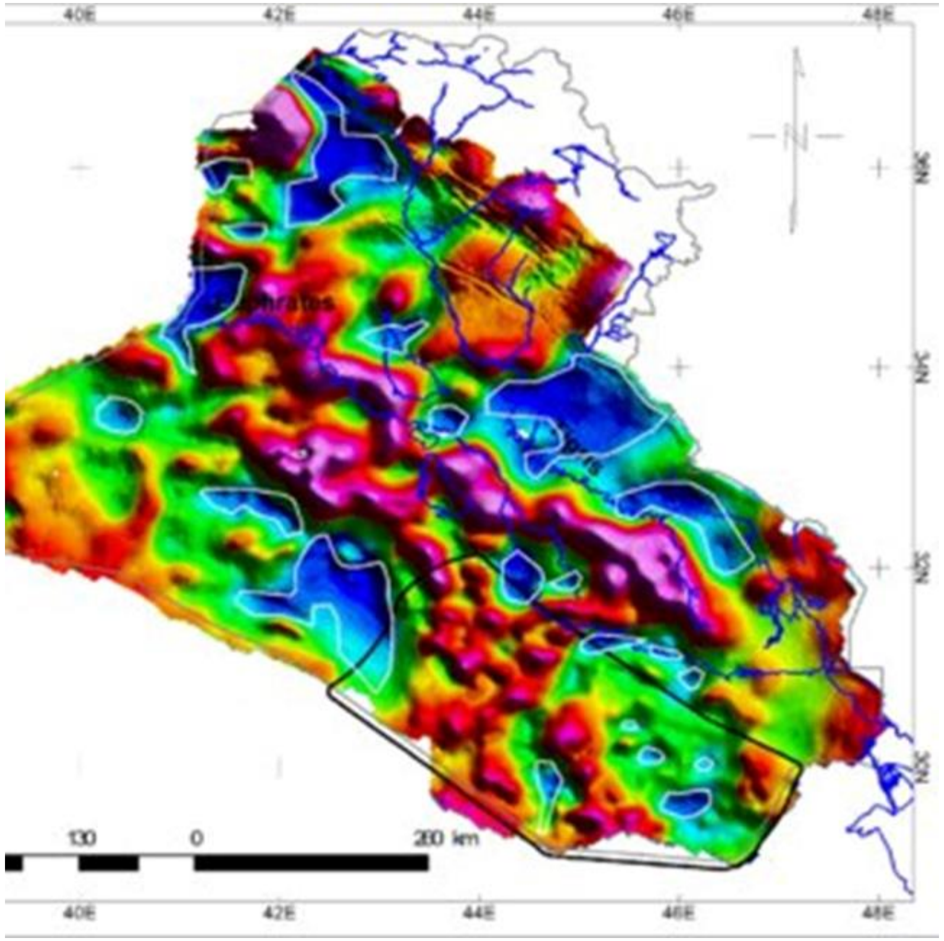


Data Acquisition:

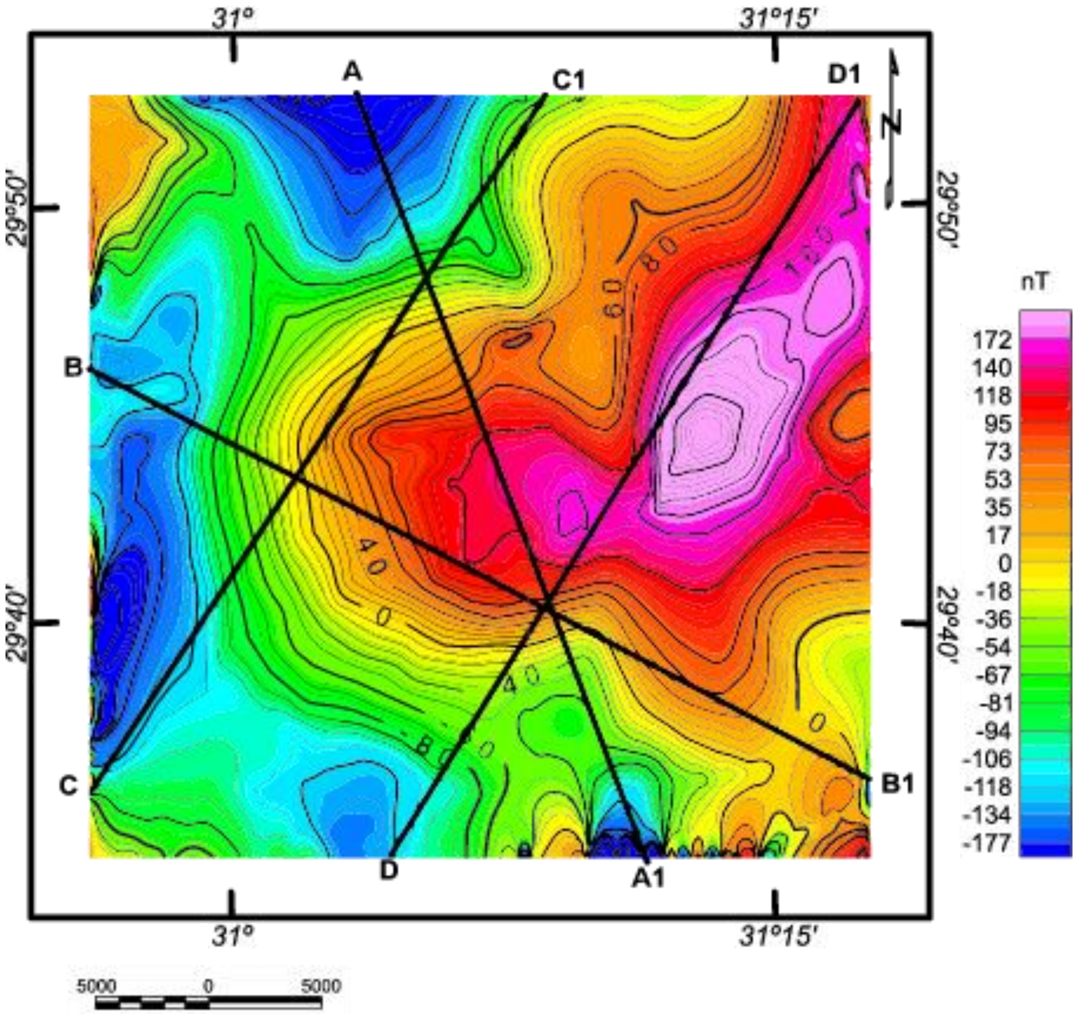
The geomagnetic field is measured by Equipment called Magnetometer on land , sea or from air. Field measured data are plotted as maps or profile and interpreted in view of geology.



Examples of magnetic maps



Aeromagnetic map of Iraq



Land-magnetic map of a certain area

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Lecture 3

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Contents:

- How the magnetic anomaly occurs
- Electrical methods
- A brief about Resistivity methods
- Theory
- General aspects

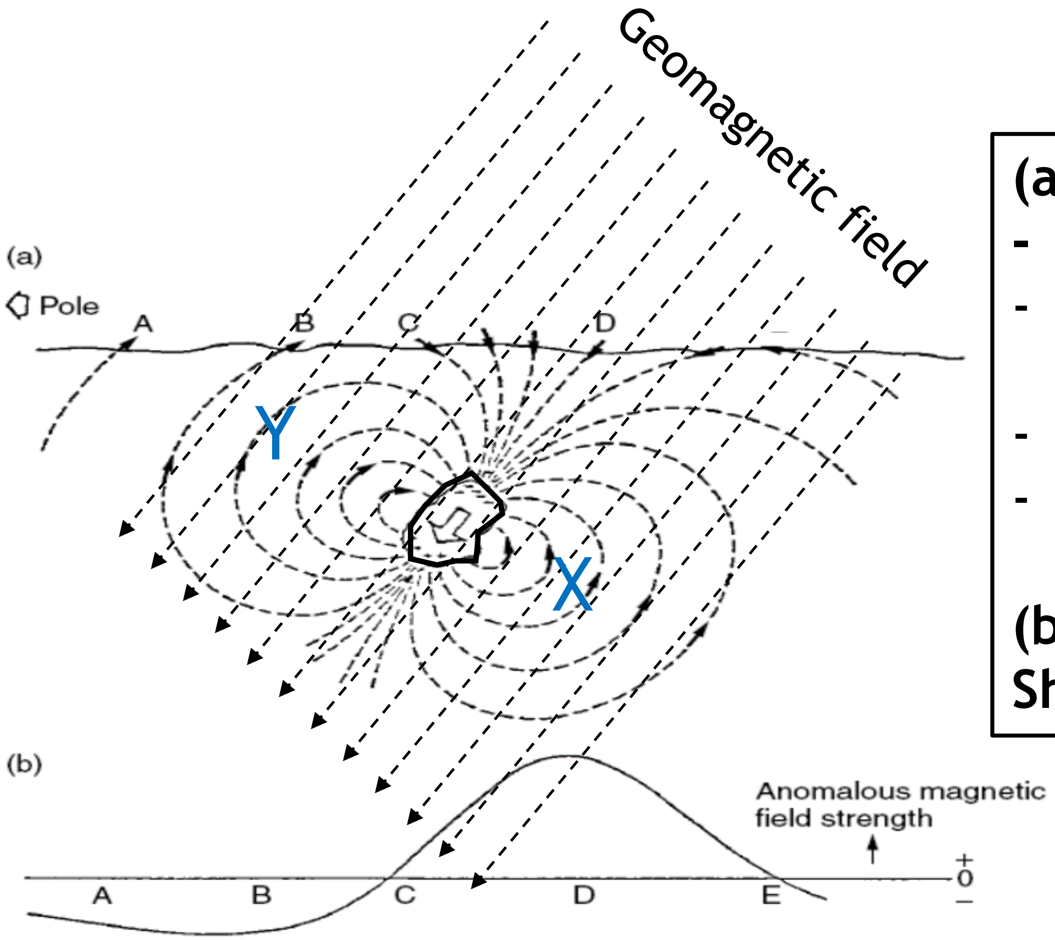
How the magnetic anomaly occurs:

The induced material magnetizes proportionally to the external field and the proportionality constant is the susceptibility K

I = KH, (H is the geomagnetic field (external for a certain body)
While I is the magnetization of that body (also called polarization)

Susceptibility K is the degree of the body to be magnetized.

When H is constant, I depends only on the susceptibility of the material.
Each material has its own susceptibility andIt is dimensionless



(a) A hidden body
- Magnetized by the geomagnetic field.
- It will have its own magnetic field, it has its own susceptibility value.
- In area X; maximum anomaly value
- In area Y minimum anomaly value

(b) The measured magnetic anomaly Showing the effect of a hidden body.

How is a magnetic anomaly formed?

https://youtu.be/cleAxsA1iq0

2- Electrical methods:

Many types of electrical methods are in use, some of them are:

- Self-potential (SP)
- Telluric currents
- Resistivity
- Electromagnetic (EM)
- induced polarization (IP)

Each method can be used to solve a certain geological problem, some of them are passive and others are active. Among them the resistivity method is the most important and widely used in solving engineering, water and environmental problems..

A brief about Resistivity methods

Applications

Electrical resistivity methods are used extensively in the search for:

- suitable groundwater sources

- monitor types of groundwater pollution

- engineering surveys to locate sub-surface cavities, faults and fissures, permafrost, mineshafts, etc.

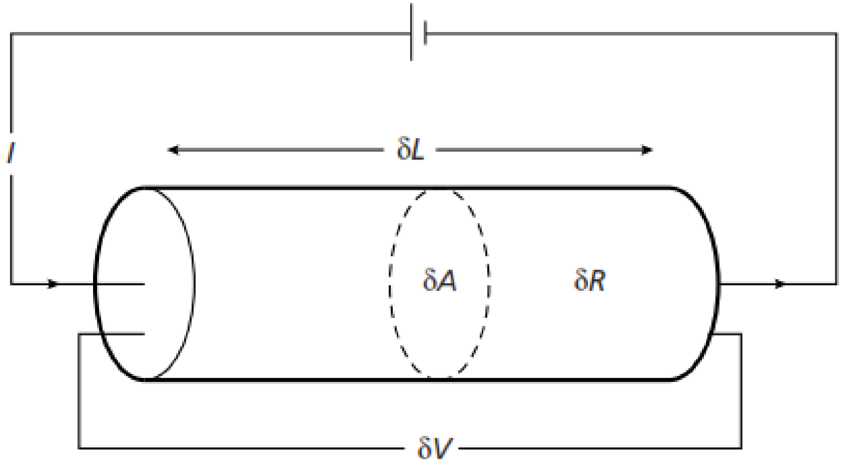
- **and in archaeology for mapping out the areal extent of remnants of buried foundations of ancient buildings**
- **Electrical resistivity methods are also used extensively in downhole logging for both petroleum and water wells.**

Theory:

The *resistivity of a material is defined as the resistance* in ohms between the opposite faces of a unit cube of the material. For a conducting cylinder of resistance *δR, length δL and cross-sectional area δA the* resistivity ρ is given by

$$\rho = \frac{\delta R \delta A}{\delta L}$$

R=V/I = Resistance
V in volts and I in amperes
When we have a uniform body
R α L/A
R = ρ L/A [ρ=R/(L/A)]
ρ is the true resistivity.



Typical Electrical Resistivities of Some Earth Materials

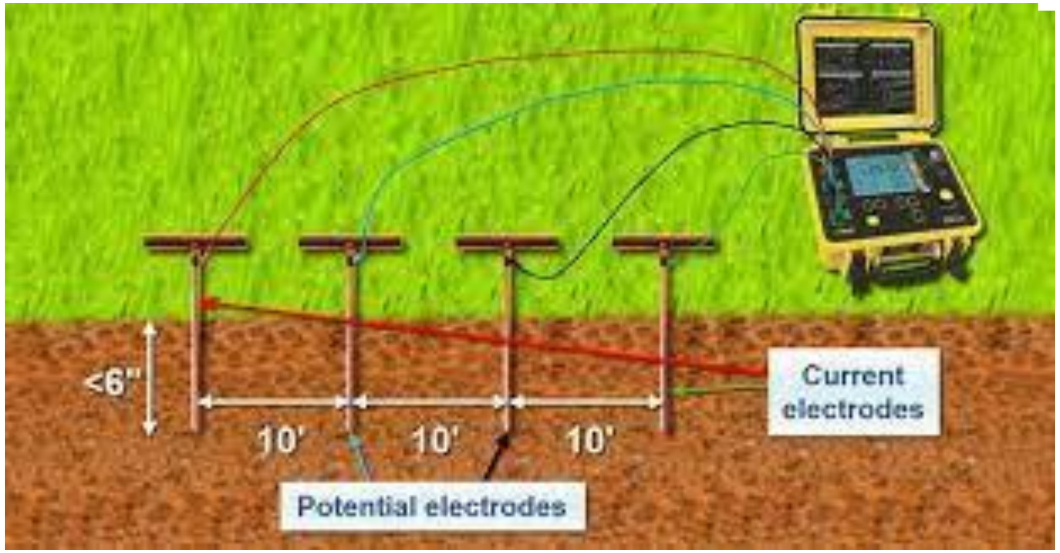
Resistivity of some rockss

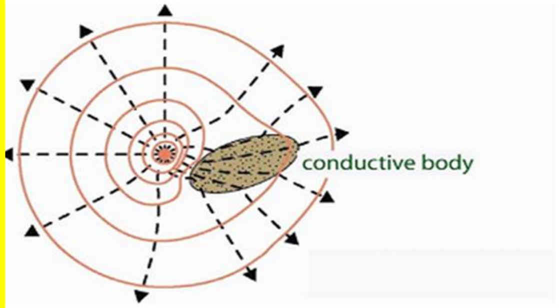
Material Resistivity	(Ωm)
Clay	1-20
Sand, wet to moist	20-200
Shale	1-500
Porous limestone	100-1,000
Dense limestone	1,000-1,000,000
Metamorphic rocks	50-1,000,000
Igneous rocks	100-1,000,000

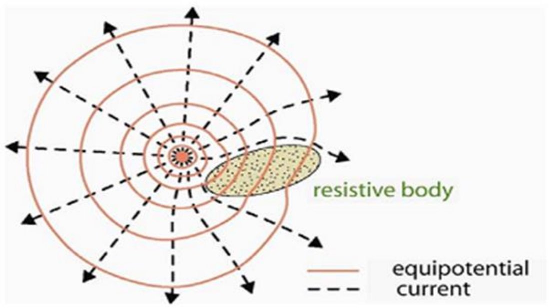
Resistivity of some metals

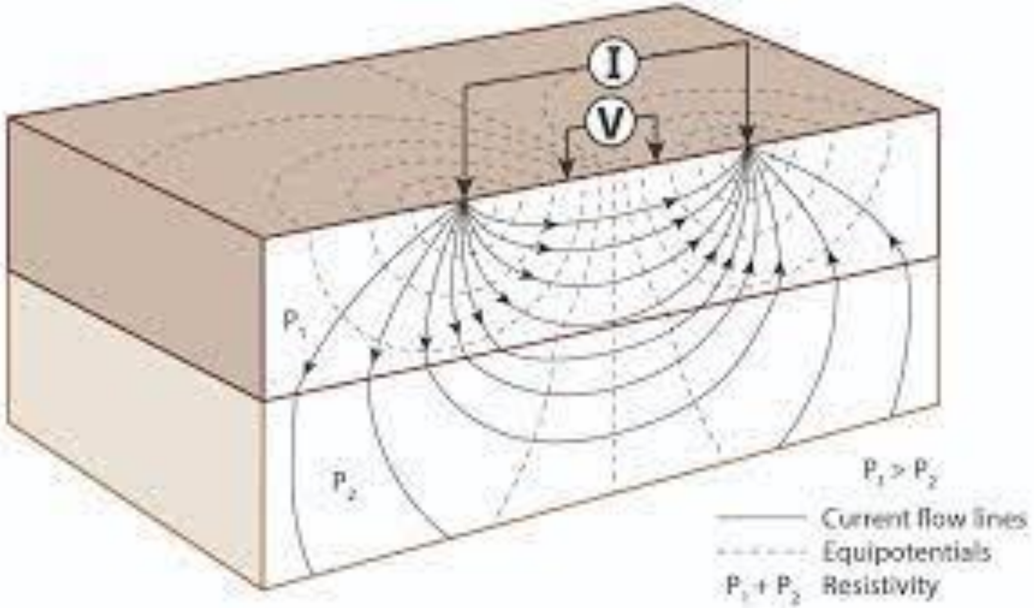
Metals	Resistivity [Ohm-m]
Aluminum	2.8×10^{-8}
Copper	1.7×10^{-8}
Iron	10×10^{-8}
Mercury	95×10^{-8}
Silver	1.6×10^{-8}
Steel	$15\text{-}90 \times 10^{-8}$

▶ <u>Resistivity method</u> involves artificially-generated electric currents introduced into the ground and the resulting potential differences are measured at the surface. Deviations from the pattern of potential differences expected from homogeneous ground provide information on the form and electrical properties of subsurface inhomogeneities.



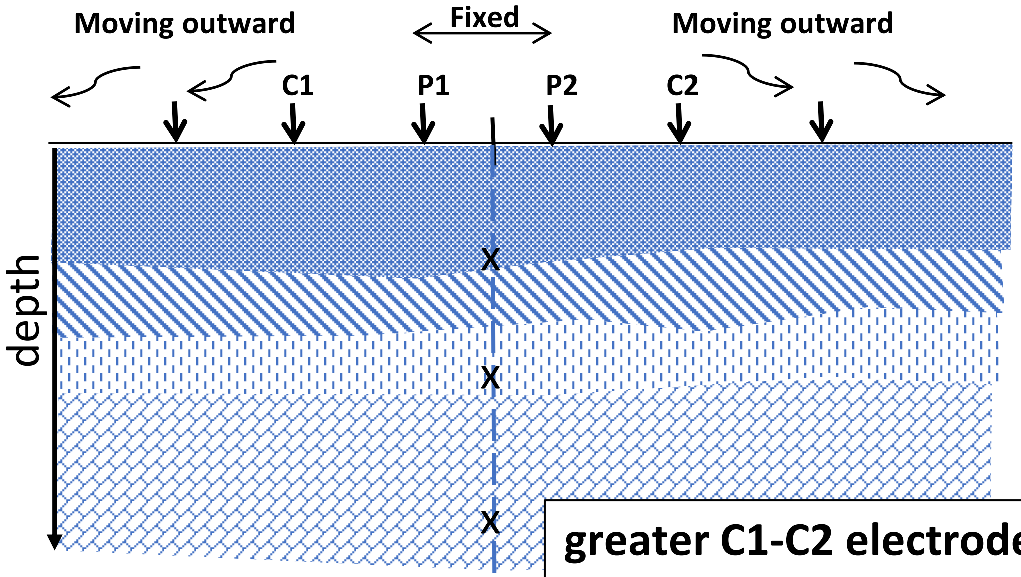




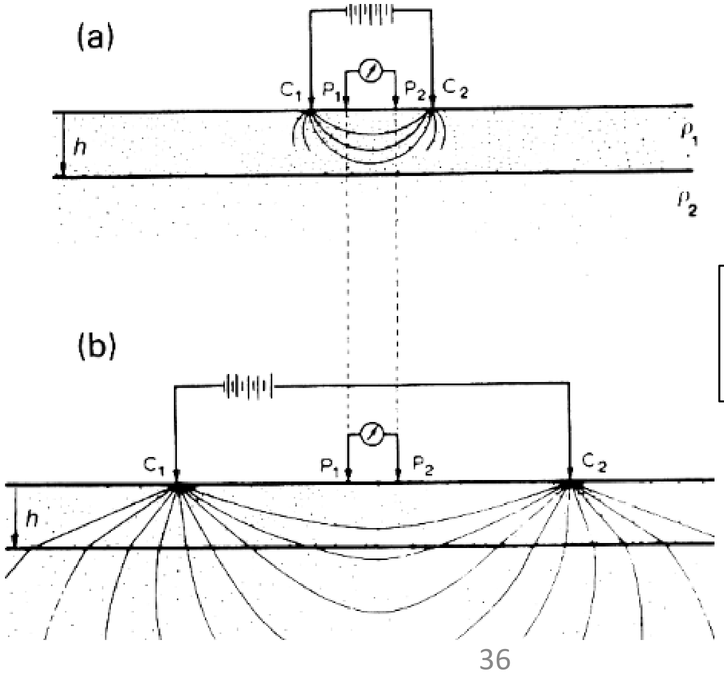


Resistivity ρ is measured at the mid-point (center) of the system of four electrodes, different depths depending on the distance between the outer two electrodes which are current sending.

Electrical resistivity is a fundamental and diagnostic physical property of rocks that can be determined by a wide variety of techniq

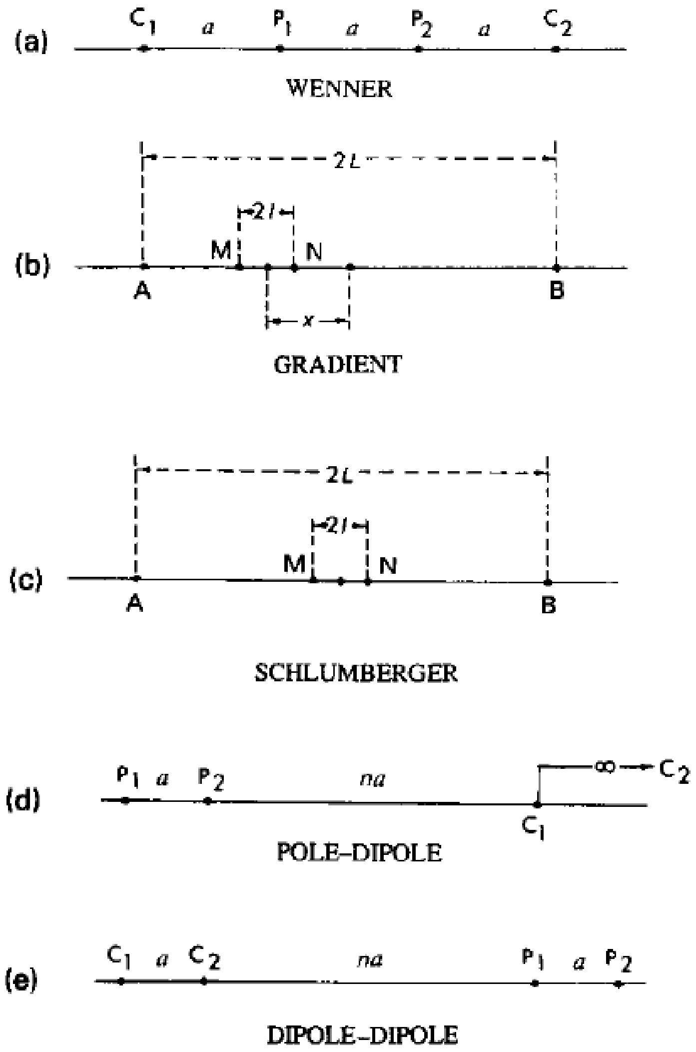


greater C1-C2 electrode separation gives greater depth investigation



What is the difference between (a) and (b)?

- There are several types of electrode configurations.
- Each type is used for solving a certain problem.
- Each type has its advantages and disadvantages.



$$\rho_a = 2\pi a \frac{\Delta V}{i}$$

$$\rho_a \approx \frac{\pi}{2l} \cdot \frac{\left(L^2 - x^2\right)^2}{L^2 + x^2} \frac{\Delta V}{i}$$

for L>10·l

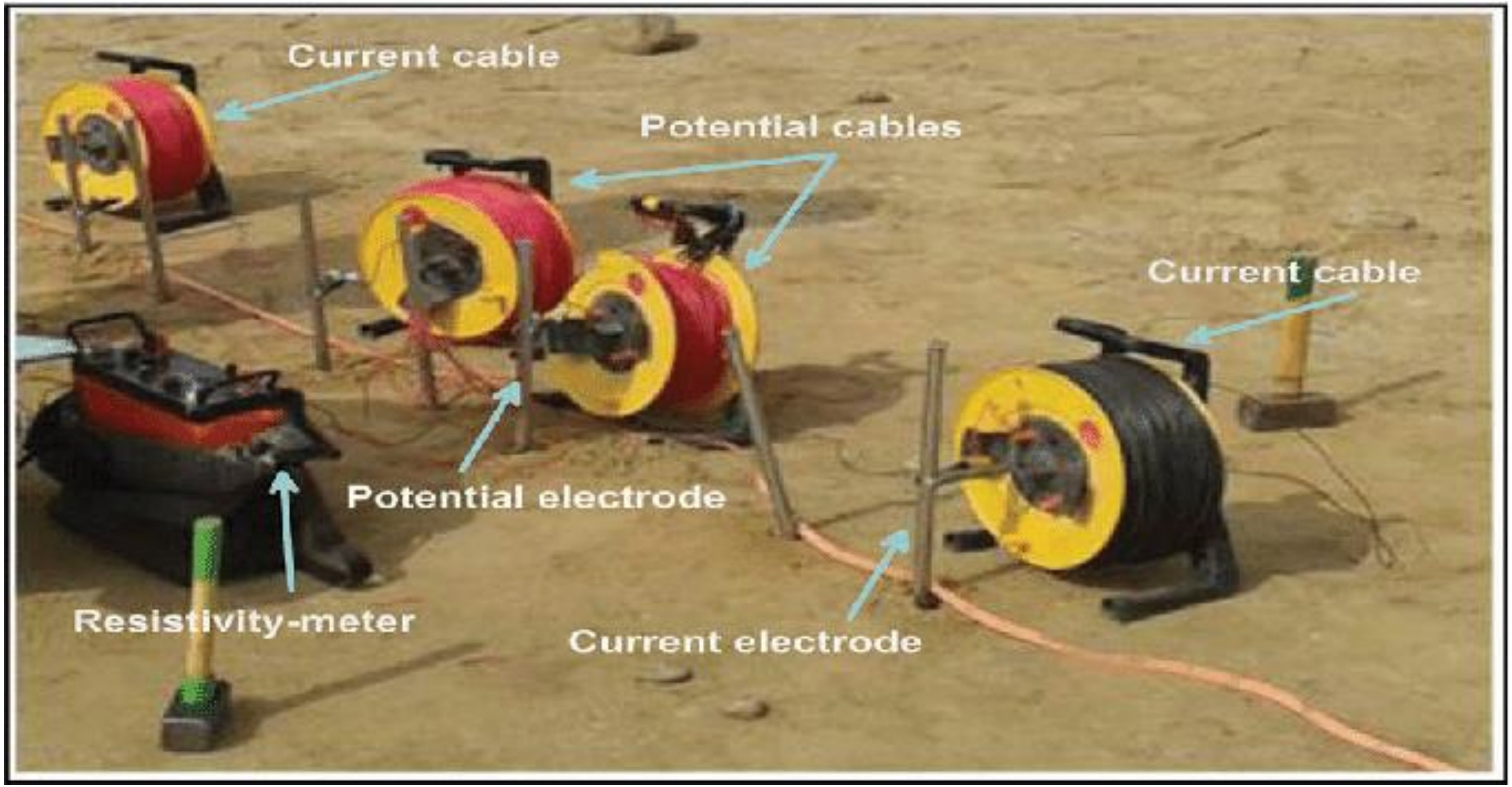
$$\rho_a \approx \frac{\pi}{2} \cdot \frac{L^2}{l} \cdot \frac{\Delta V}{i}$$

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

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

Among these configurations the dipole-dipole arrangement is successfully used in exploration for oil in Rusia

(Telford et al. 1990; Sharma 1997)

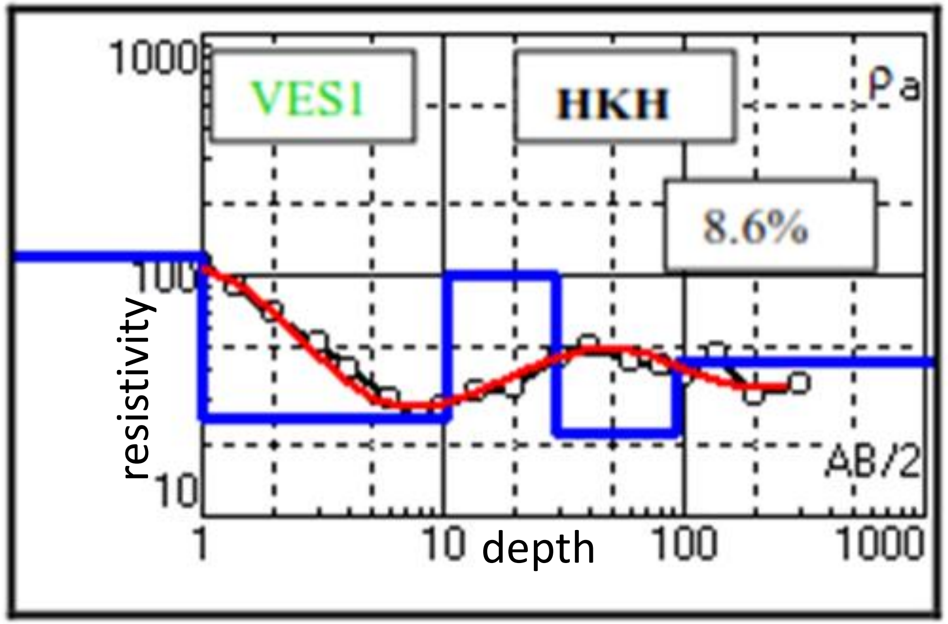

Current cable
Potential cables
Current cable
Potential electrode
Resistivity-meter
Current electrode

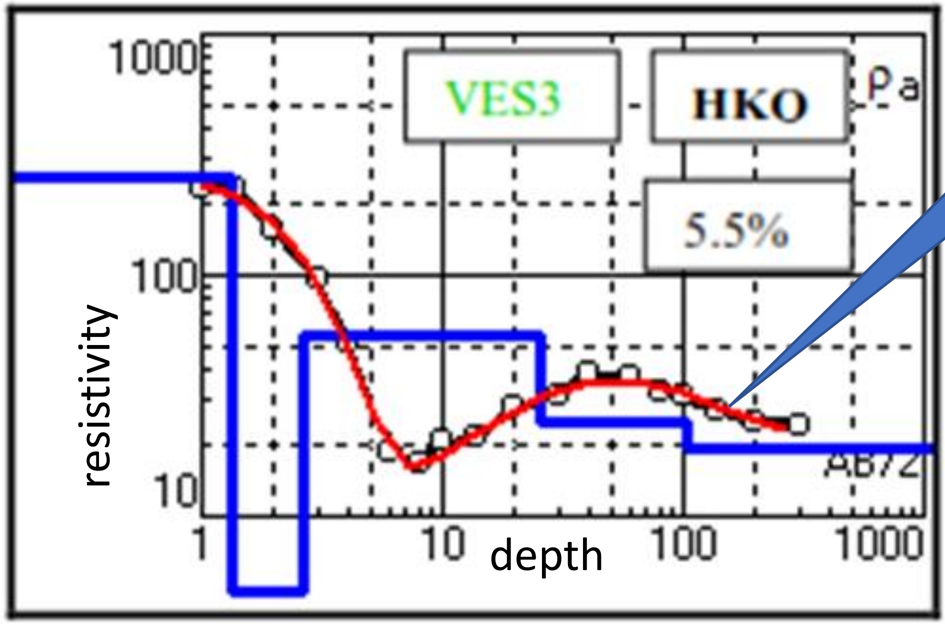
General aspects:

- From the value of resistivity which is measured the interpreter can predict the type of rock, porosity, permeability and type of the fluid which is present in pores at that depth.
- Depth of investigation can go down to a about one mile.
- There are many configurations other than the type given in the previous slides.
- Surveys are carried out in one, two or three dimensional.
- The SI unit of resistivity is the ohm-metre (ohm m) and it is the most used unit.
- An Electrode is made of convenient metallic matter of about 60cm length penetrated 10-20cm into the ground.
- Factors affecting resistivity are salinity of water, type of rocks, degree of cementation and finally type and amount of clay.

A simple 1-dimentional curve of a resistivity field data at two points are given as examples below:

The apparent resistivity values are plotted on a log-log graph paper and processed to give true resitivities and then interpreted in terms of depth.





How we get such a curve? See the next slide

1-Dimensional means; investigating a certain point on the Earth by measuring the resistivity at many depths underneath that point (i.e., getting variation of resistivity with depth)

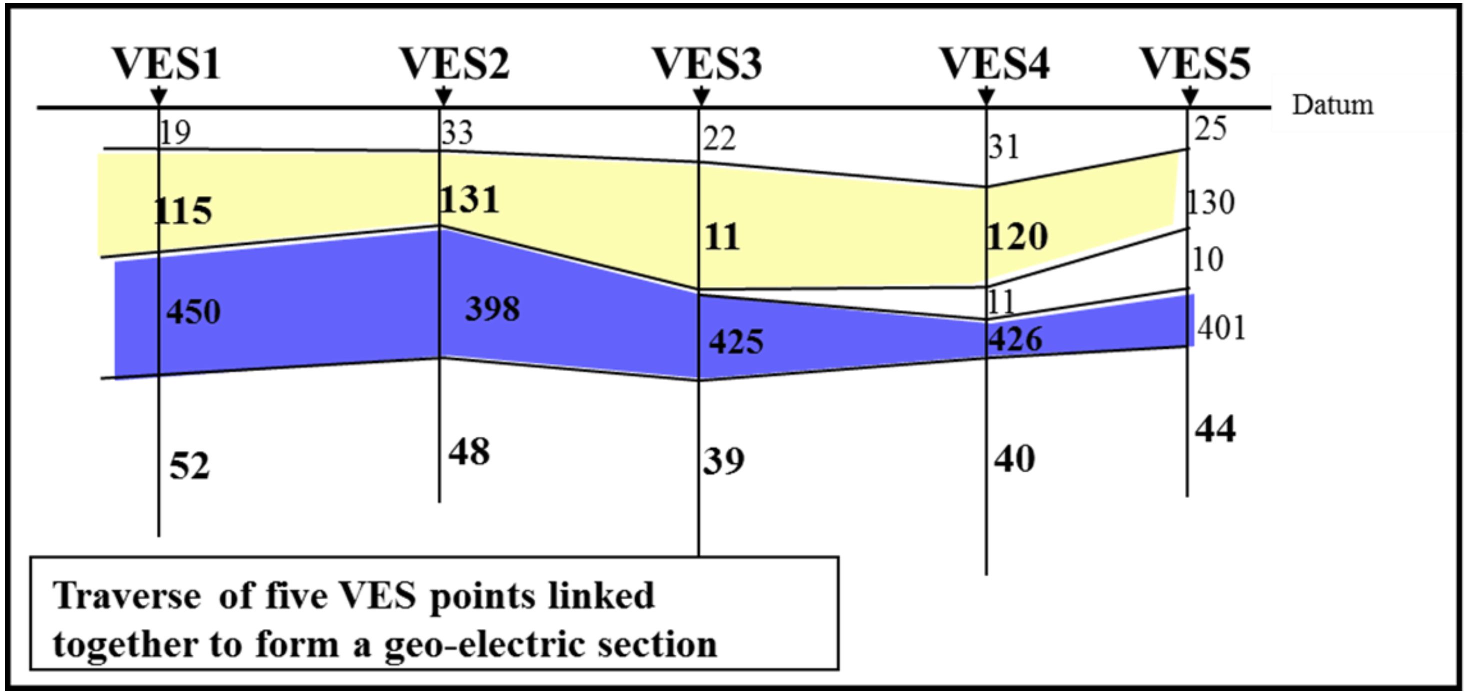
Mark No.	C1C2/2 (m)	P1P2/2 (m)	K	R (ohm)	ρ_a (ohm.m)
1	3	1	12.6		
2	4	1	23.6		
3	5	1	37.7		
4	6	1	55		
5	8	1	98.9		
6	10	1	155		
7	12.5	1	244		
8	12.5	5	41.2		
9	15	1	351.7		
10	15	5	62.8		
11	20	5	118		
12	25	5	188		

To be filled in field

Field Procedure for VES Schlumberger as an example

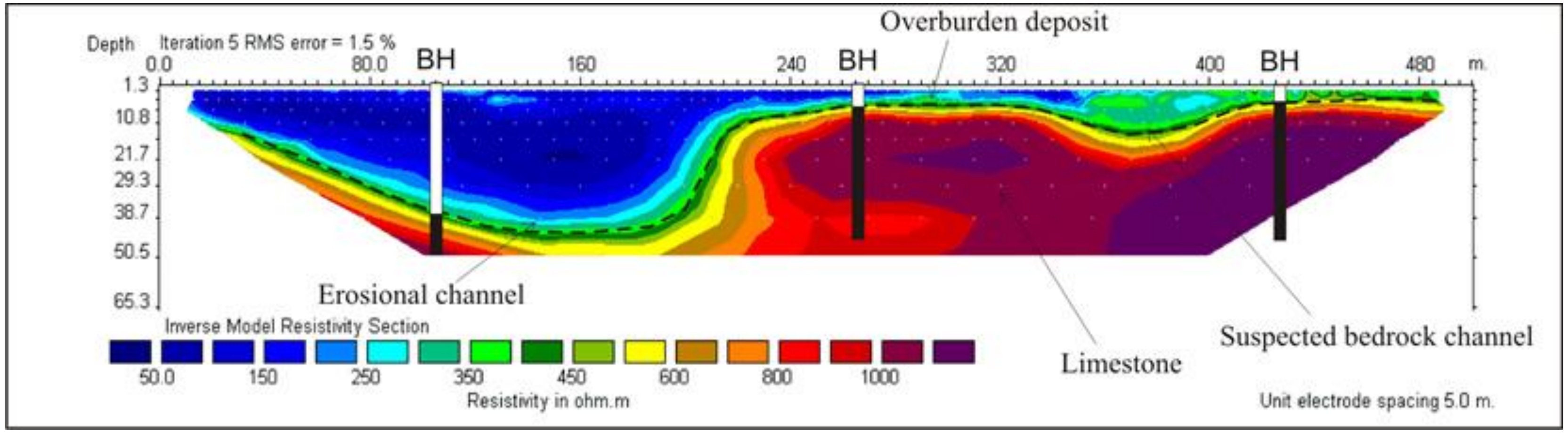
When the data is got from field they are plotted on log-log paper and interpreted by standard curves manually or using special software to obtain curves such as those given in the previous slide.

When more than one test are plotted one beside one the cross section of the ground a one-dimensional section



Traverse of five VES points linked together to form a geo-electric section

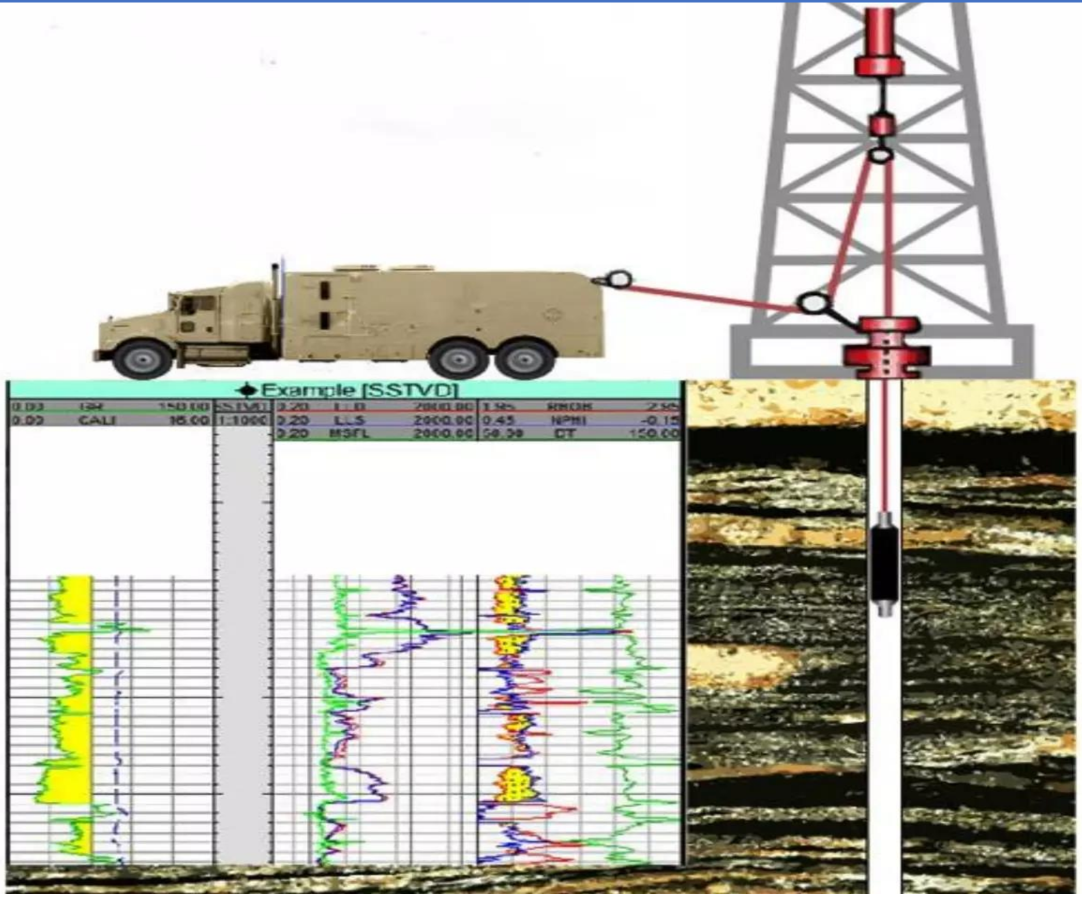
There is also a Two-dimensional sections (more advanced)



One D survey: https://youtu.be/JQ3Q0r26Nr0

For petroleum engineer, electrical resistivity and electrical SP logging are of great importance. They will be given in details in the 3rd grade. These logs are used to:

- **Determine hydrocarbon versus water bearing zone.**
- **Indicate permeable zones.**
- **Determine porosity**



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Lecture 4

Applied Geophysics Lecture Notes
Fourth Semester
By:

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2023/2024

Contents:

- Radioactive methods
- Gravity Method
- Applications of Gravity method
- The Geoid
- The Universal Law of Gravitation
- Measurements of Gravity

-

3- Radioactive methods:

Radiometric surveys on the ground surface are primarily ‘passive,’ measuring natural radioactivity. However, some ‘active’ methods are in use on the ground surface and in boreholes.
The passive methods are the most widely used radiometric techniques, in particular the use of gamma ray scintillometers and gamma ray spectrometers.

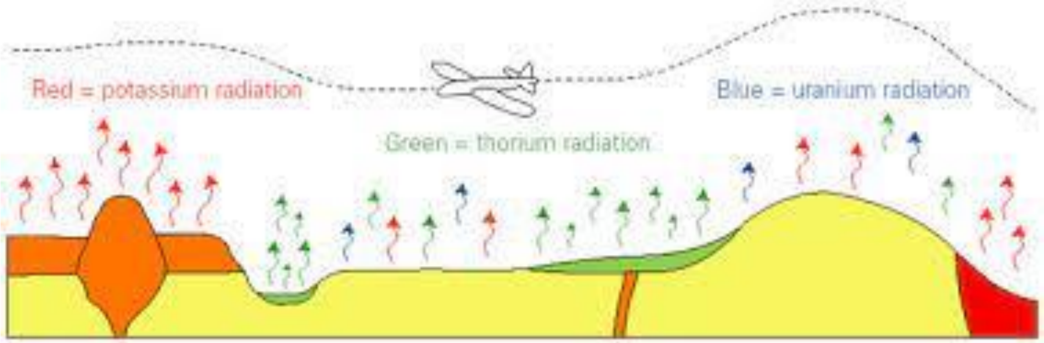
The radiometric survey detects and maps natural radioactive emanations (γ-rays) from rocks and soils. Gamma radiation occurs from the natural decay of elements like U, Th, and K. The radiometric method detects these elements at the surface.

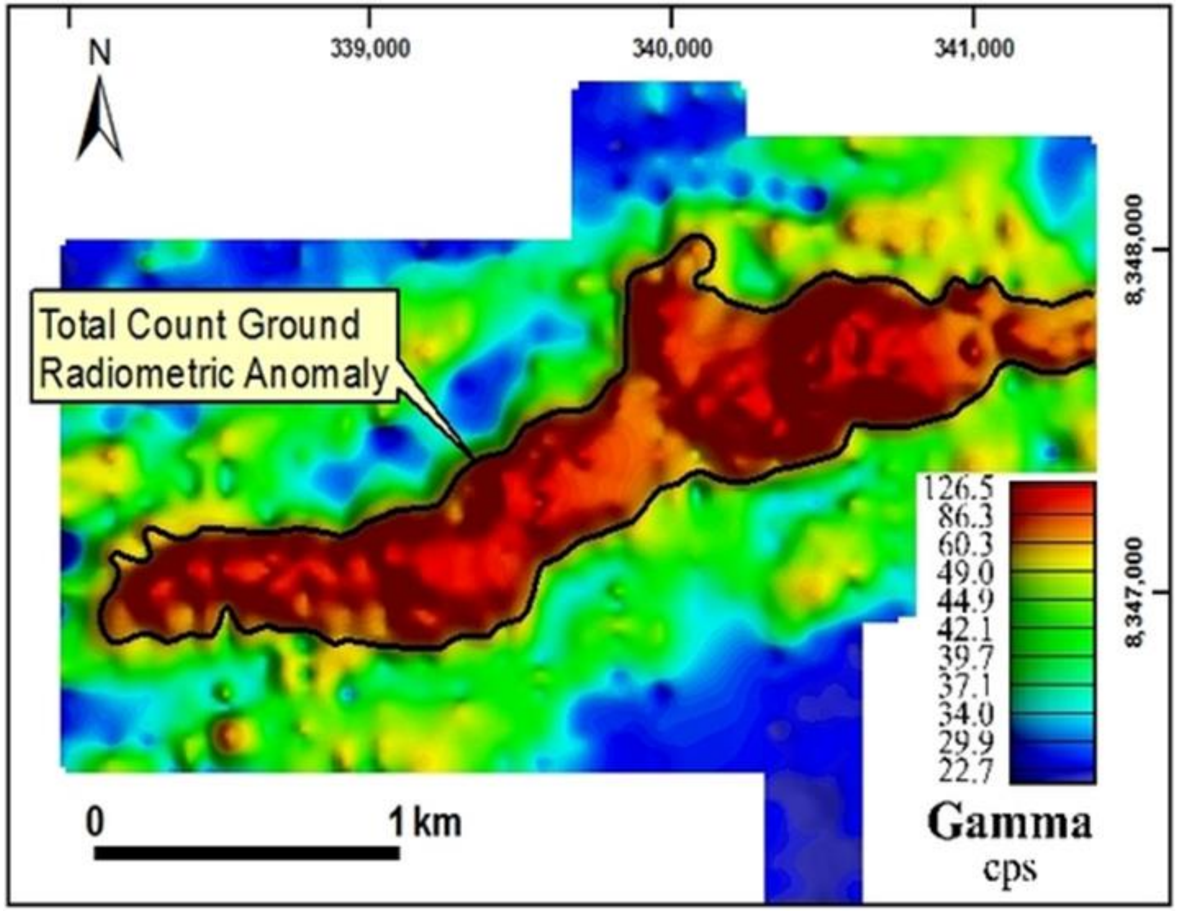
This technique has basically two applications: geological mapping and identification of outcropping bodies with high content of radioactive minerals.

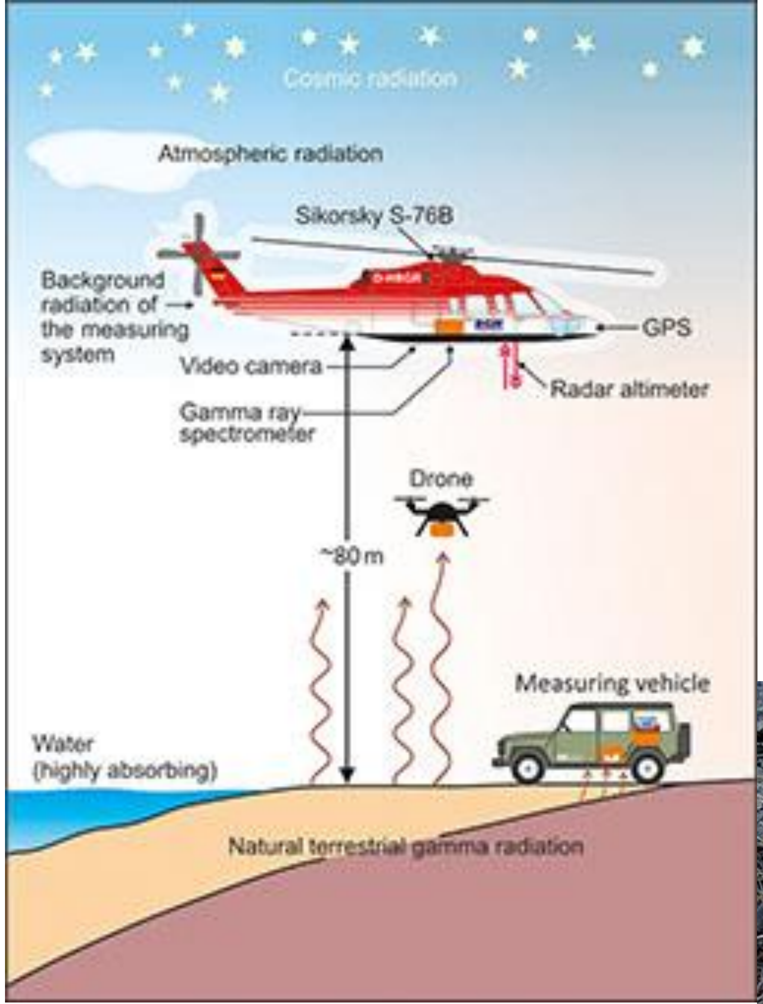
The most important method in geophysical radiometric is the measurement of gamma radiation. Gamma radiation is a high-energy electromagnetic radiation emitted during radioactive decay processes.

Measurements can be carried out from aircrafts (plane, helicopter or drone), on the ground (vehicle based or hand-held instruments) or in boreholes (borehole tool). The principle of measurement is almost the same for each of the setups.

In geophysical applications gamma ray spectrometers are mostly used for geological mapping, for mineral exploration, particularly Uranium exploration, and for environmental problems such as identification and mapping of radioactive contaminations.

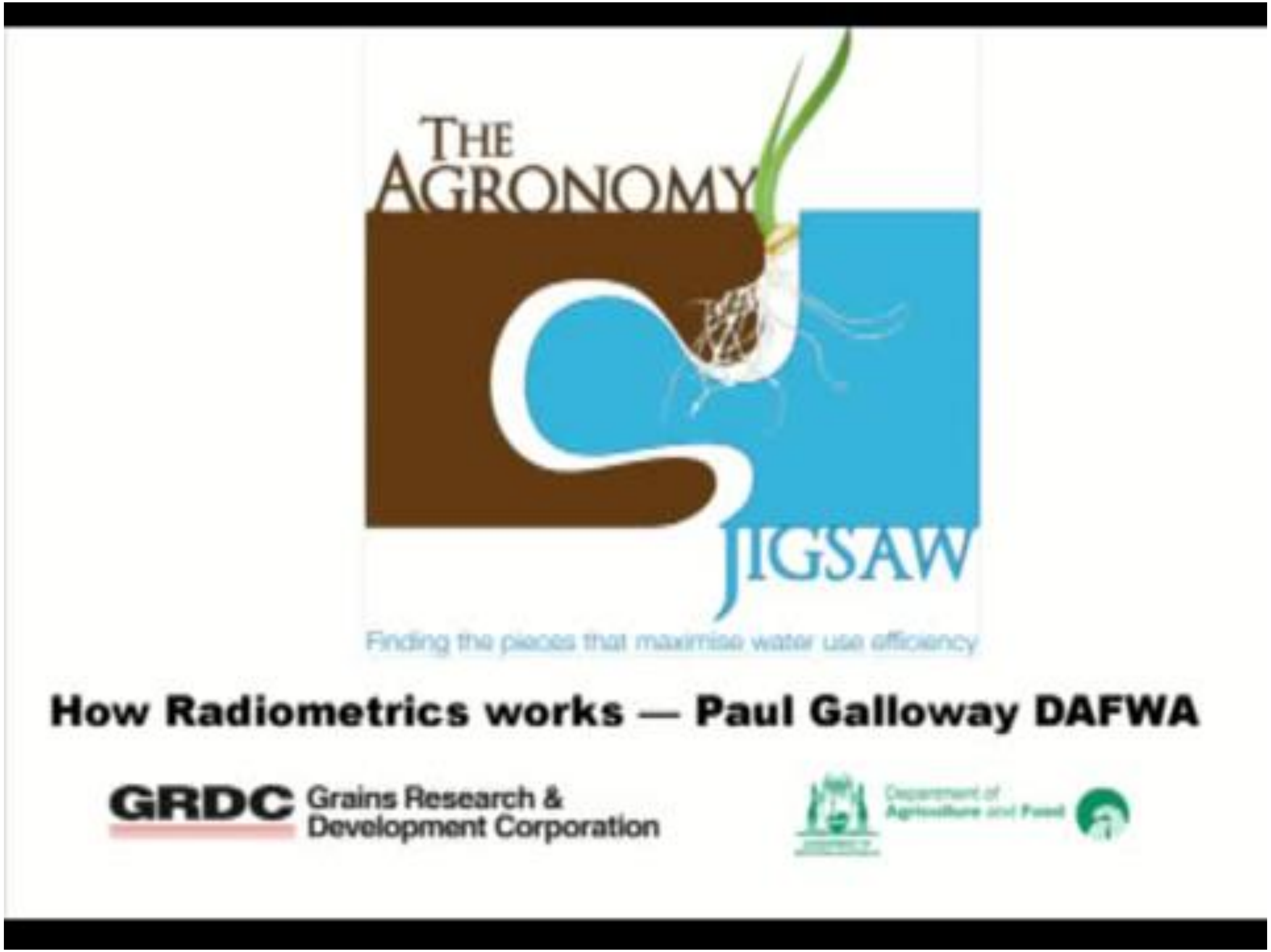

Red = potassium radiation
Green = thorium radiation
Blue = uranium radiation


N
339,000
340,000
341,000
Total Count Ground Radiometric Anomaly
8,348,000
8,347,000
126.5
86.3
60.3
49.0
44.9
42.1
39.7
37.1
34.0
29.9
22.7
Gamma
cps
0
1 km


Cosmic radiation
Atmospheric radiation
Sikorsky S-76B
Background radiation of the measuring system
GPS
Video camera
Radar altimeter
Gamma ray spectrometer
Drone
~80 m
Measuring vehicle
Water (highly absorbing)
Natural terrestrial gamma radiation

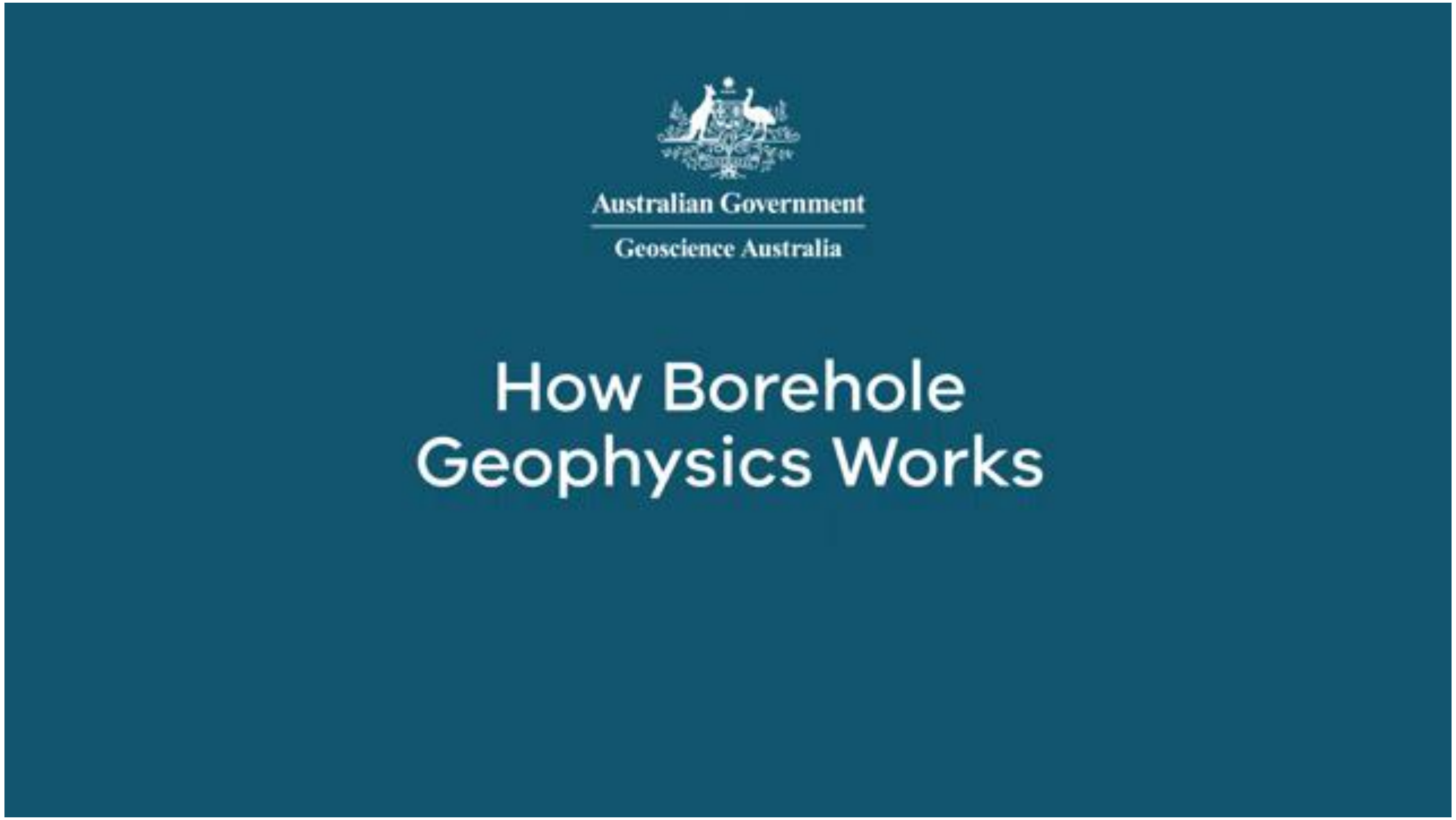

Geiger counter

https://youtu.be/tjp4IlcJyUU



In petroleum engineering the method is mainly used to explore boreholes through a technique known as (Gamma ray logging).

https://youtu.be/-_AbBynqI9E



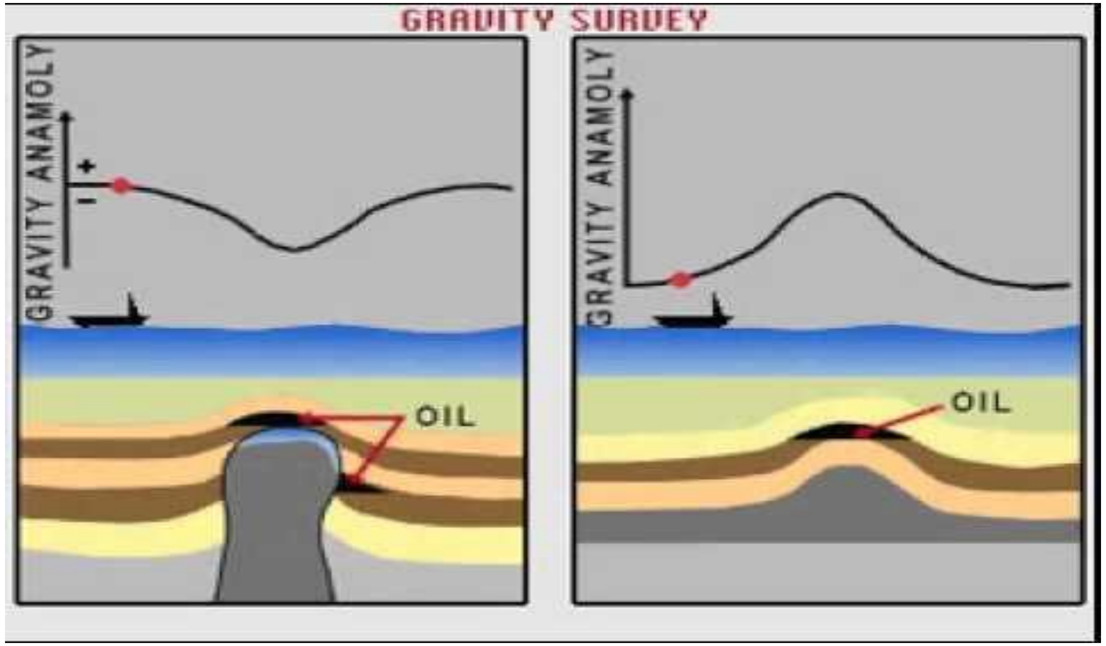
Will be given in details in the 3rd stage of your study program

4- Gravity Method

General Review

The basic concept underlying gravity surveying is the variation of the Earth gravitational field caused by lateral variation of subsurface rock-densities. In other words, a given rock body whose density is different from its surrounding medium (i.e. geological anomaly) produces a corresponding disturbance (gravity anomaly) in the Earth gravity-filed. The form and amplitude of the created anomaly depends on the subsurface geological anomaly such as a salt dome, granite intrusion, buried valley, folded or faulted beds.

<u>Gravity surveying</u> involves measurements of the changes in gravitational acceleration (gravity) at points over a given area distributed randomly or in the form of a grid or along a line. The observation-data are then subjected to a series of corrections and some processes in order to reduce them to gravity values measured relative to a datum-plane in such a way that all gravity values are related to variations in the sub-datum.



The corrected and processed field data is then interpreted to what they mean in geology. This gravity-to-geology process (interpretation) forms the ultimate objective of any gravity-survey project.

Applications of Gravity method:

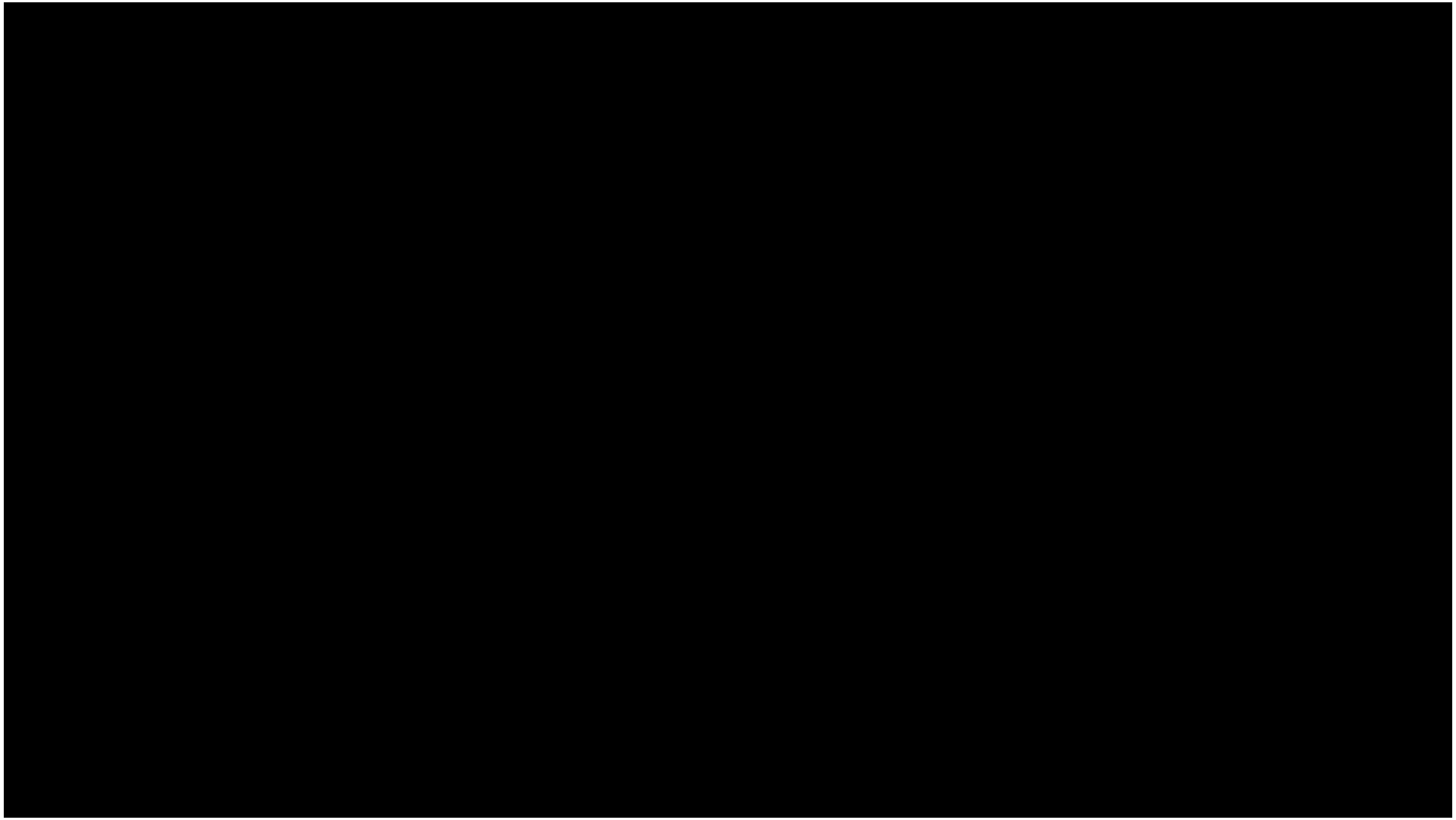
Primary;

hydrocarbon exploration, regional geological studies.

Secondary;

explorations for: mineral deposit, Site investigations, hydrogeology, karsts, geodesy, isostasy, archaeology and volcanic monitoring.

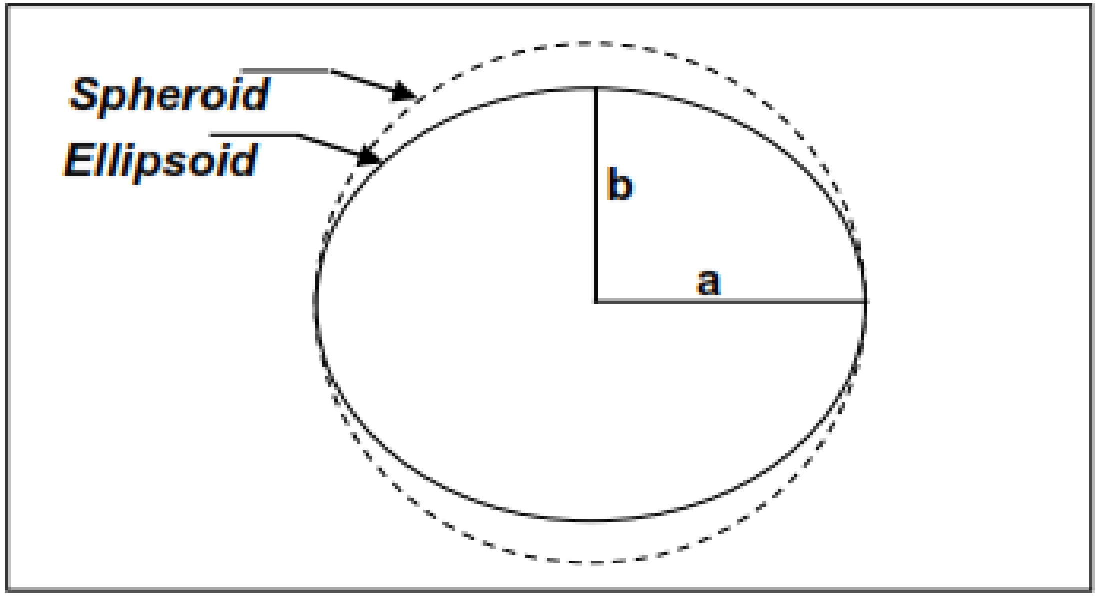
Gravity method---definition: https://youtu.be/M2aSsy5tzI0



The Earth's Shape

The Ellipsoid

The Earth is approximately ellipsoidal rather than being perfectly spherical along the mean sea-level surface. This model is called the reference or normal ellipsoid. It is a theoretical surface.

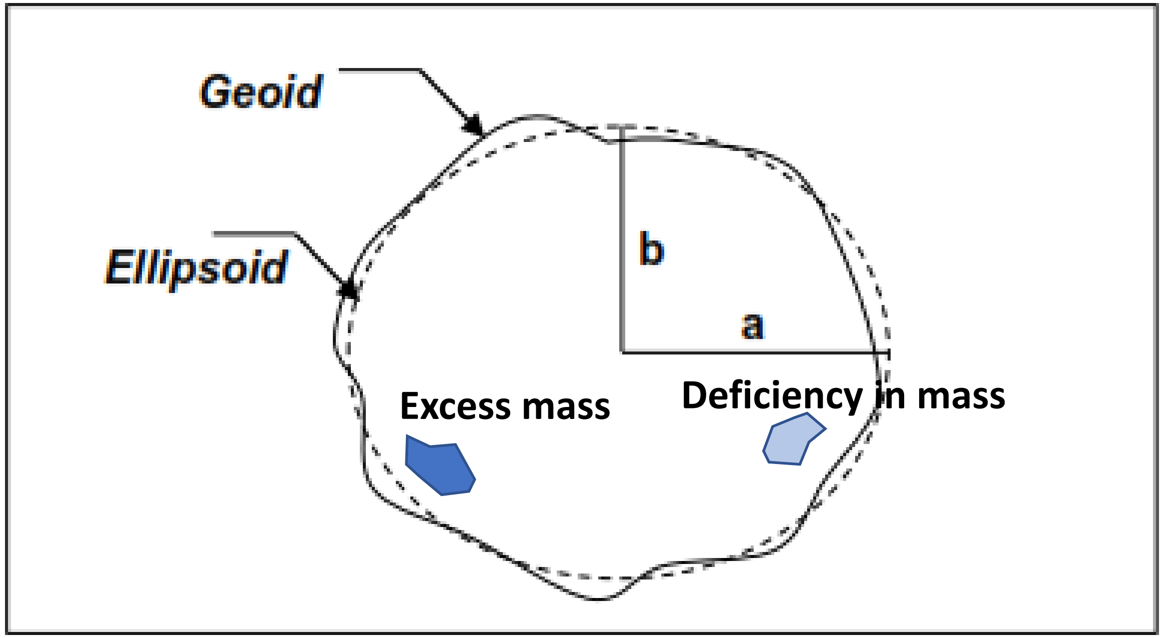


<u>Up to date information</u>
The equatorial radius (a) = 6378.160 km
The polar radius (b) = 6356.775 km
The difference (a-b) = 21.385 km
The angular Rotation Speed (ω) = 7.292 $*10^{-5}$ radian/sec
(About 460 meters per second)

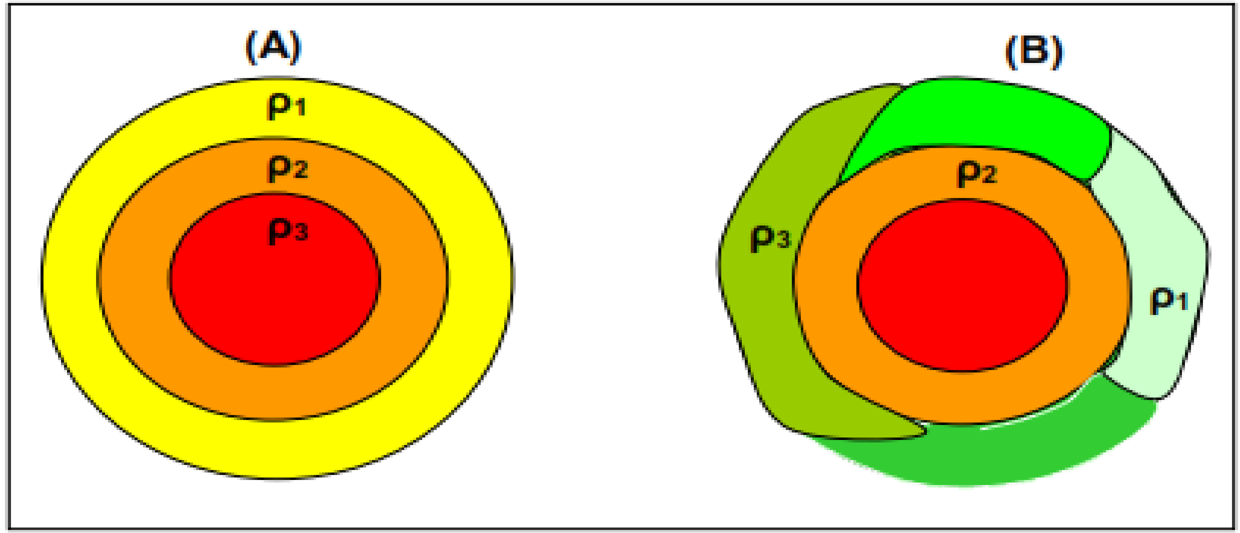
The Geoid

It is the Ellepsoid surface affected by mass distribution beneath the datum level. It is global with the plumb line vertical at any point on its surface.

The difference between Geoid and Ellipsoidal surfaces is called <u>Warp</u> which indicates the presence of either excess or deficiency in mass.
Warp is an anomaly.



Explanation



A: A theoretical model of the earth having vertical variations in density leading to the Ellipsoidal model.

B: An actual Earth having vertical and horizontal variations in density leading to Geoid model.

The corrected and processed field data is then interpreted to what they mean in geology. This gravity-to-geology process (interpretation) forms the ultimate objective of any gravity-survey project.

Applications of Gravity method:

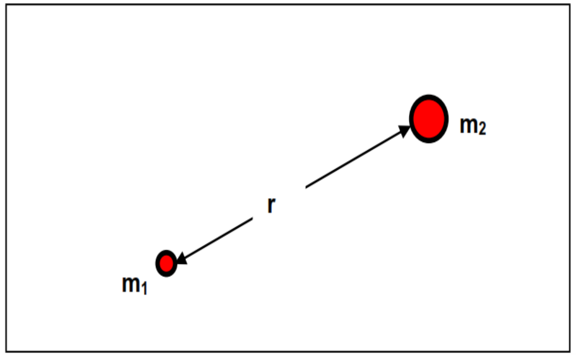
Primary; hydrocarbon exploration, regional geological studies.

Secondary; explorations for: mineral deposit, Site investigations, hydrogeology, karsts, geodesy, isostasy, archaeology and volcanic monitoring.

The Universal Law of Gravitation

Isaac Newton (1643-1727) formulated the universal law of gravitation which evaluates the attraction force F that exists between two particles of masses m_1 and m_2 located at distance r apart.

$$\boxed{F = \frac{G m_1 m_2}{r^2}}$$



The constant G, called the Universal Gravitational Constant, was experimentally determined and found to be of the value:

$G = 6.673 \times 10^{-8}$ (Dyne.cm^2/gm^2) or, (cm^3.gm^{-1}.sec^{-2})

or,

$G = 6.673 \times 10^{-11}$ (N.m^2/kg^2) or, (m^3 .kg^{-1}.sec^{-2})

cgs and SI units ?

How can you derive this unit?

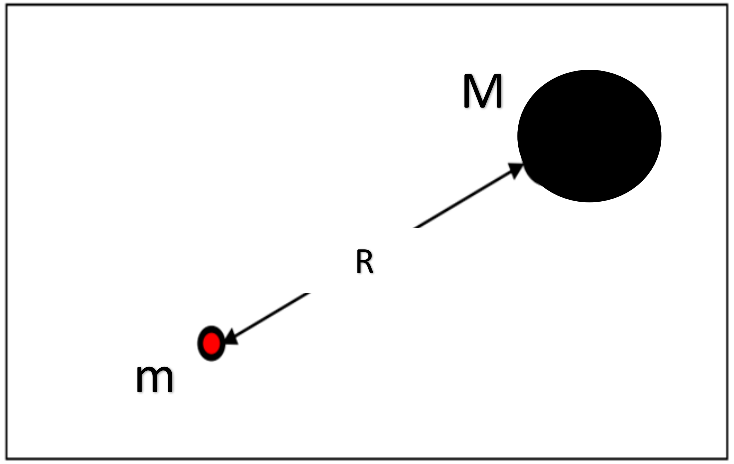
Newton's second law of motion states that any body of mass (m) under the effect of attraction force (F) (say it is the force of attraction of the Earth mass -M- with a radius -R-) moves with acceleration (a) where:

$F = ma$

$G\, m.\, M / r^2 = ma$ ------ or ----- $= mg$

Acceleration a is called gravity (g), then

$g = G\, M/R^2$ ----------------------- in our course, acceleration is called gravity



<u>The units of g </u> that are in common use are:

1 gal = 1 centimeter /sec ^{2}

1 milligal = (1/1000) gal (used in Petroleum surveys)

1 microgal = (1/1000 000) gal (used in small scale surveys)

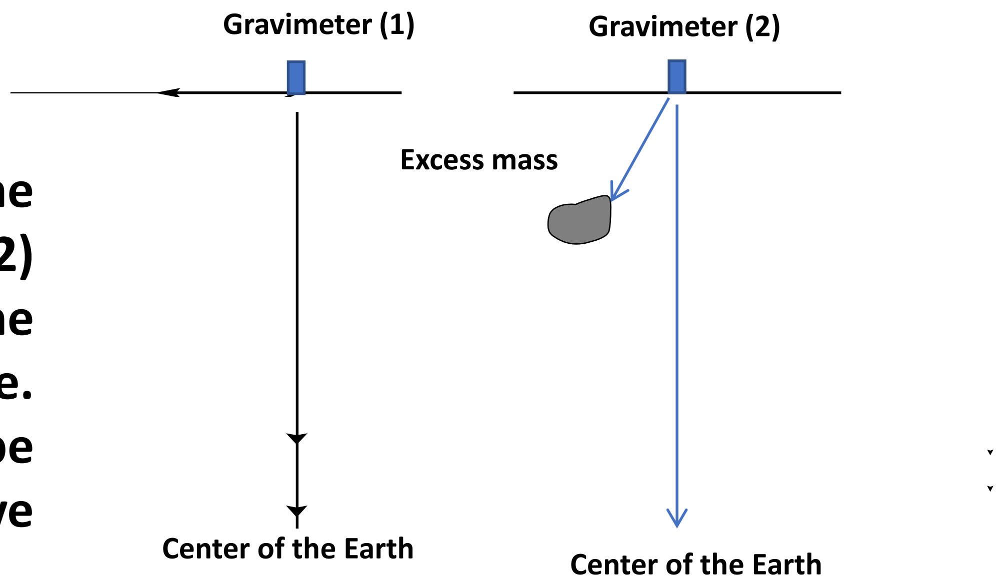
Newtons laws: https://youtu.be/Af9lRX4xsr0

Note that

The acceleration is simply called gravity (g) and it is measured during the gravity survey. According to the relation $g = G\,M/R^2$, g is proportional to *two variables only*; they are density and distance so gravity changes on a horizontal plane (datum plane) in a small district should be due to changes in density beneath datum only. *If the earth was perfectly sphere and having a homogeneous density, gravity value will not change any where on its surface.*

Gravity anomaly explanation

Gravimeter (1) will measure (g), due to the attraction of the Earth while gravimeter (2) measures the (g) of the Earth plus (g) due to the excess mass present. The anomaly will be positive. In the case of a cavity instead of mass there will be an efficiency in mass so giving rise to a negative anomaly relatively.

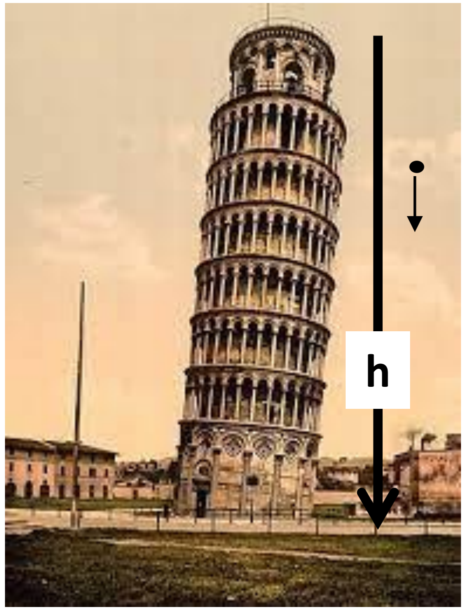


Measurements of Gravity

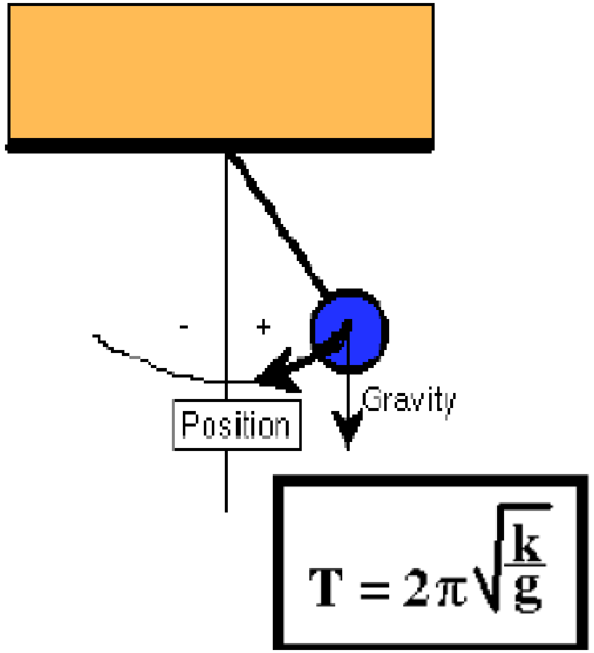
Two types of gravity

1- Absolute gravity

Determination of the acceleration due to gravity in absolute terms requires very careful experimental procedures and is normally only undertaken under laboratory conditions. Two methods of measurement are used, namely the Falling body and Swinging pendulum methods.



$h = (1 / 2)\, g\, t^2$



2- Relative gravity measurements

Because it is difficult to design precise portable absolute measuring instruments, gravity is measured relatively (relative to a reference station called Base Station), these stations are or are not tied to primary base stations (primary base stations have absolute gravity values).

Devices which measure relative gravity values are called Gravimeters. They are accurate balances measure gravity in different points on the surface to show the variation in gravity, hence the variation in density. A net of International Base Stations (primary) are present in all countries.

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Lecture 5

Applied Geophysics Lecture Notes
Fourth Semester
By:
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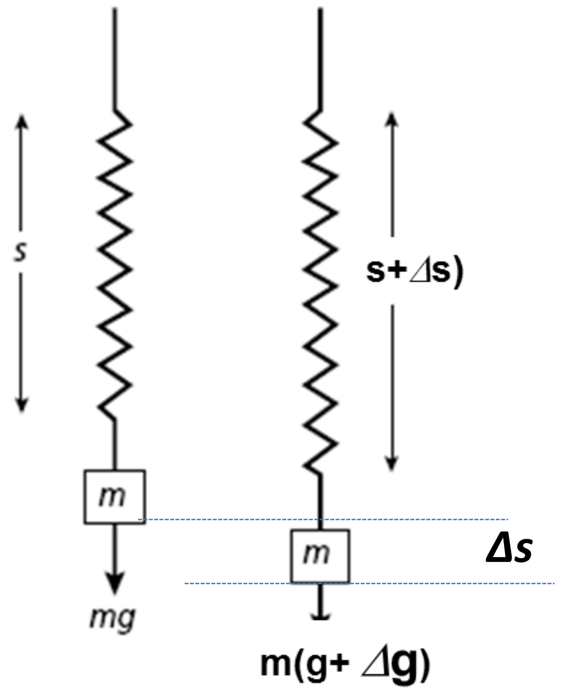
2023/2024

Contents:

<u>The principle of a gravimeter</u> is a spring of initial length *s* has been stretched by an amount d*s* as a result of an increase in gravity d*g* increaseing the weight of the suspended mass *m*. The extension of the spring is proportional to the extending force (Hooke's Law), thus

$m\,\Delta g = k\,\Delta s$
and $\Delta s = (m/k)\,\Delta g$
where k is the elastic spring constant and
Δs must be measured to a precision of
1 : 108 in instruments suitable for gravity surveying on land.



All gravimeters depend upon this principle. Based on that, many types of gravimeters are now present such as those which are shown in the next page .

Commercial types of Gravimeters

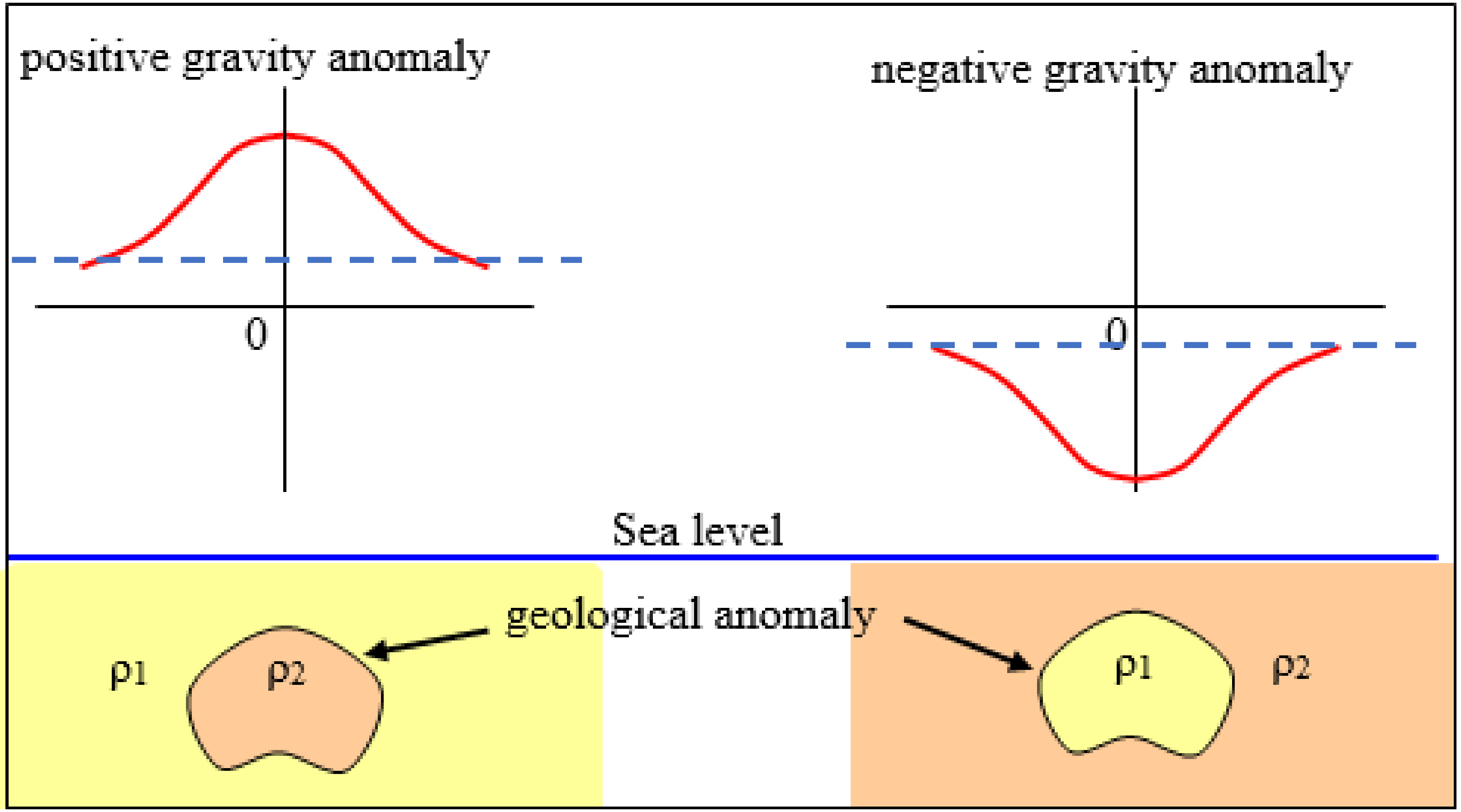
LaCost & Romberge

EG

CG-5 Autograv


GPS system
Trimble

Gravity anomaly idea can be understood from the following figure. It is the abnormal gravity measurements due to the presence of abnormal density of rocks underneath the surface.



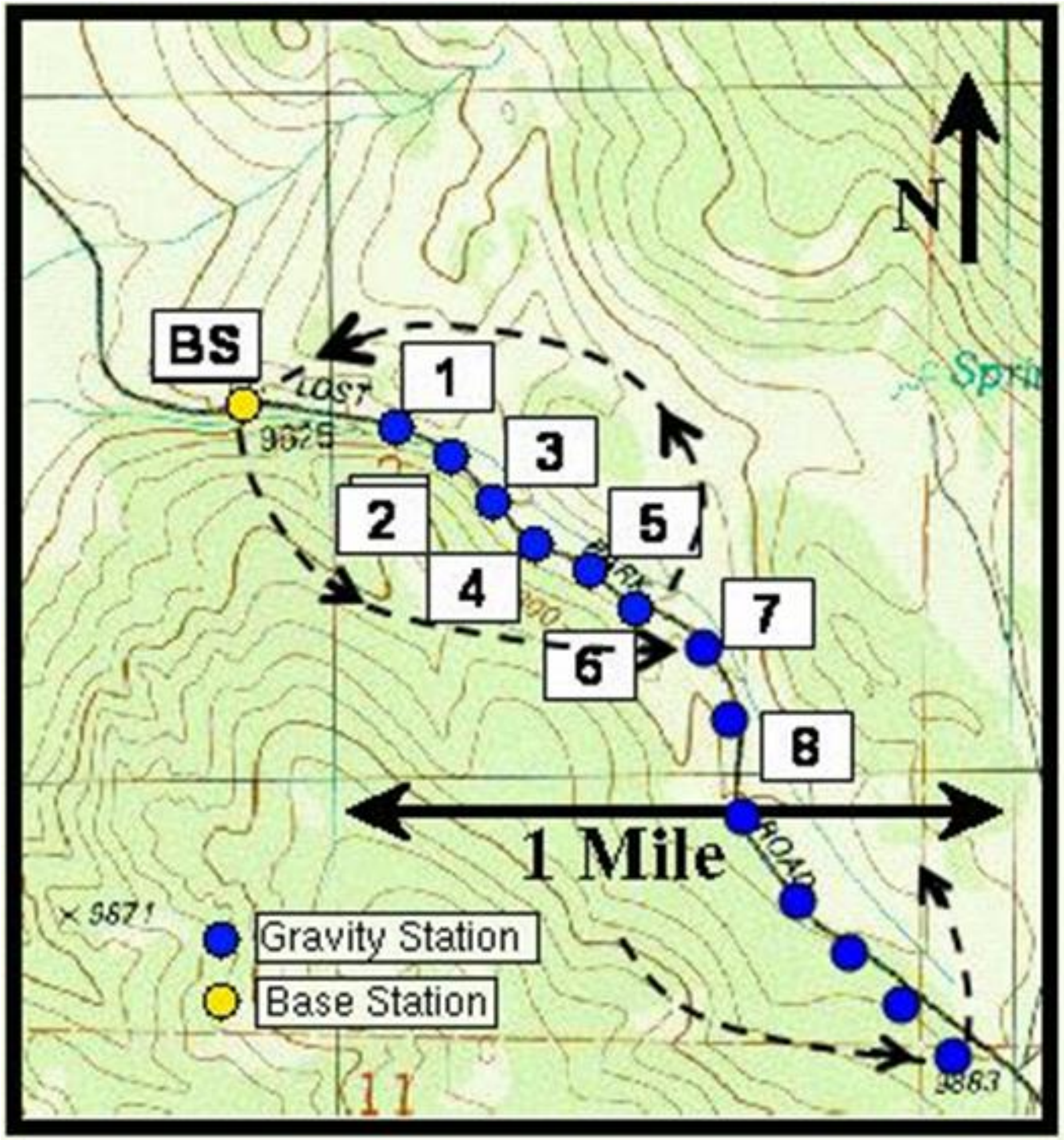
Gravity surveying

It involves measuring the gravity values at defined locations (the survey station-points or stations) which are distributed throughout the survey area. In addition, other measurements data must be made available. These are the supporting data which principally include location coordinates (longitudes, latitudes and elevations) and times at which readings are taken. These data are necessary for computing the gravity anomaly in a later processing stage.

A land gravity survey is normally conducted through two operation phases. These are; establishment of the base-station network and documenting the gravity readings at all of the station-points in the survey area. A location map with an adequate scale must be first made available

A base-station is a point located within, or near, the survey area. Its gravity value and position is precisely known. The gravity value at the base station is used as a reference value for computing gravity at all the survey-points to be relative.

St.	Time	go(SD)	coordinates X Y Z	Remarks
BS	8:05	5684.32	---	---
1	8:18	5688.66	---	---
2	8:31	5679.25	---	---
3	8:39	5978.65	---	---
4	8:49	5992.24	---	---
5	9:00	5983.28	---	---
6	9:25	5894.36	---	---
BS	9:35	5684.63	---	---
7	10:00	5882.95	---	---
-	-	-	-	
-	-	-	-	



Field Procedure

In gravity survey the relative measurements of gravity are taken by measuring gravity values from stations relative to the base station which is located in the study area. The measurements of all stations are relative to this base station which is called (secondary base station). In case we need the absolute gravity value of the secondary base station, we should look for nearest primary base station.

Primary base stations are topographically, and geographically well-defined reference stations present in all or most countries where the absolute value of gravity is established accurately.

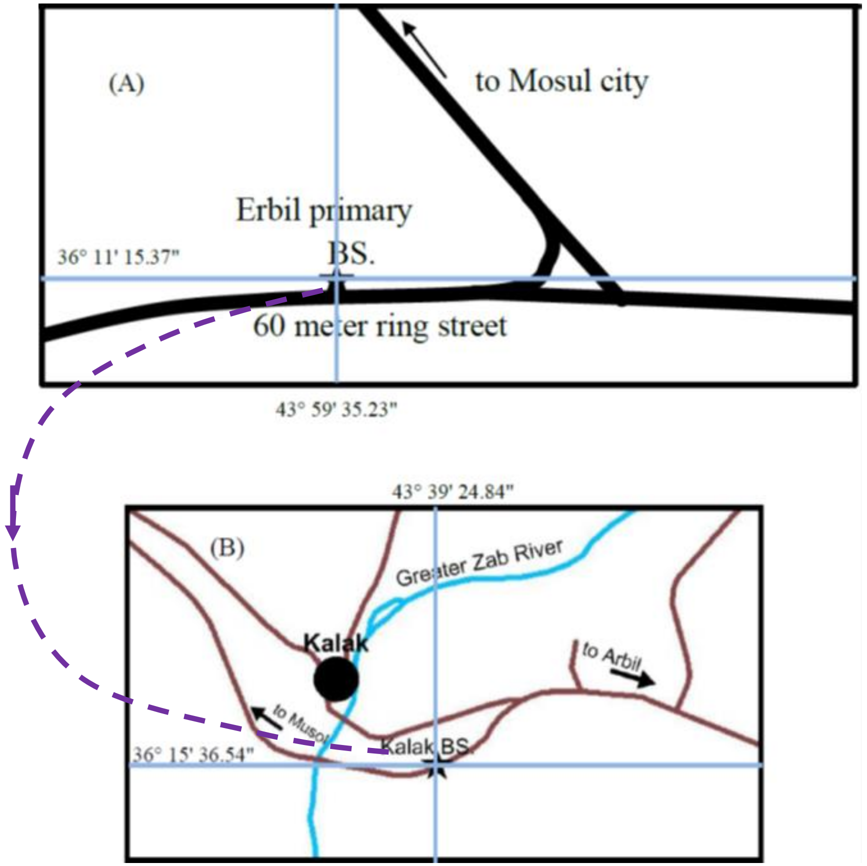
How to transfer absolute gravity value from a primary base station to a secondary base station in the studty area? In the next slide.

Transferring an absolute gravity value

Here is an example how to transfer the absolute vale from a primary base station in Erbil (E) with an absolute gravity value of (979679.434 mGal) to a secondary base station in Kalak Town (K).

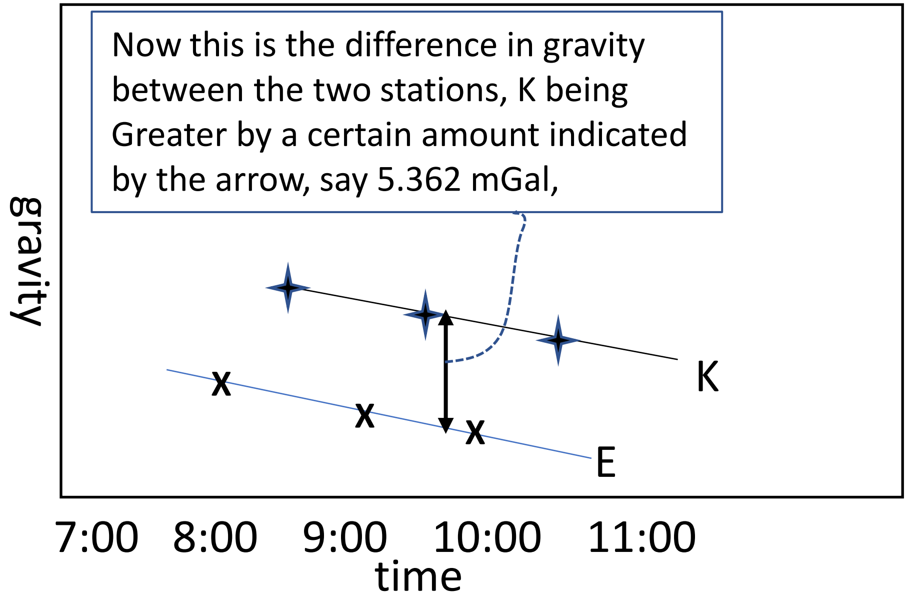
Steps are as follows:

- Read the gravimeter value at E, say at 8 AM, go to K and read the gravimeter value, say 8:30AM. Read many times in this manner by looping between the two stations, say two times for each. Now you will have the following set of data:



Time	station	Reading
8:00	E	512.3
8:30	K	515.5
9:00	E	512.1
9:30	K	515.3
10:00	E	5512
10:30	K	515.1

Plot them on a graph parer



So, since the absolute value of E is (979679.434 mGal), then the absolute value in K will be (979679.434 + 5.362 = 979684.796 mGal)

Factors Affecting Gravity

Factors which affect readings can be subdivided into two categories: those that give rise to temporal variations and those that give rise to spatial variations in the gravity.

A. **Temporal Variations - These are changes in the observed acceleration that are time dependent. In other words, these factors cause variations in acceleration that would be observed even if we didn't move our gravimeter. These are Instrument Drifts.**

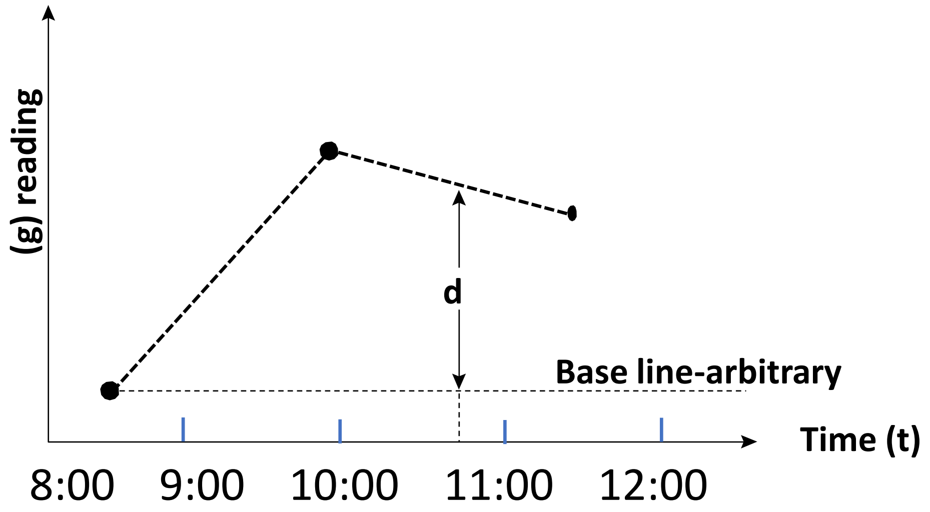
B. **Spatial Variations - are changes in the observed gravity that are space dependent (i.e., gravity change from place to place or due to elevation change. They are not related to subsurface geology, such as moving towards north or south and different altitudes.**

Gravity reductions or Corrections

Before the results of a gravity survey can be interpreted it is necessary to correct for all variations in the Earth's gravitational field which do not result from the differences of density in the underlying rocks. This process is known as gravity reduction or reduction to the geoid, as sea-level is usually the most convenient datum level.

Drift correction

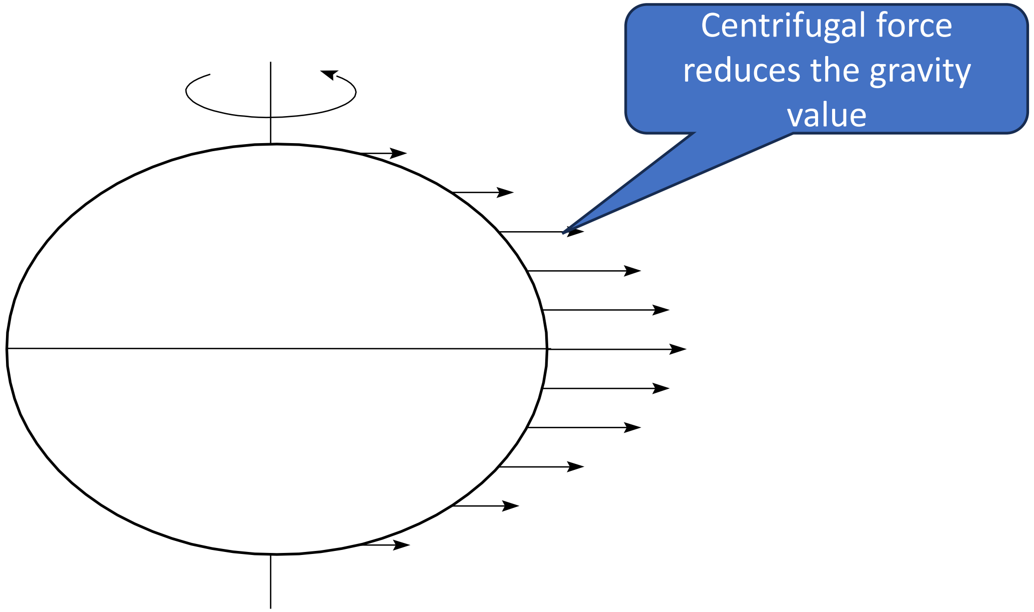
Gravimeter readings change (drift) with time as a result of elastic creep in the springs, producing an apparent change in gravity at a given stations. The instrumental drift can be determined simply by repeating measurements at the same station at different times of the workday, typically every 1 – 2 hours.



Solid dots are measurements at base station in different times

Latitude correction

Gravity varies with latitude because of the non-spherical shape of the Earth and because the angular velocity (rotation velocity) of a point on the Earth's surface decreases from a maximum at the equator to zero at the poles.



Points near the equator are farther from the centre of mass of the Earth than those near the poles, causing gravity to increase from the equator to the poles.
International formulas are present to carry out the correction for all measured stations.

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Lecture 6

Applied Geophysics Lecture Notes
Fourth Semester
By:

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2023/2024

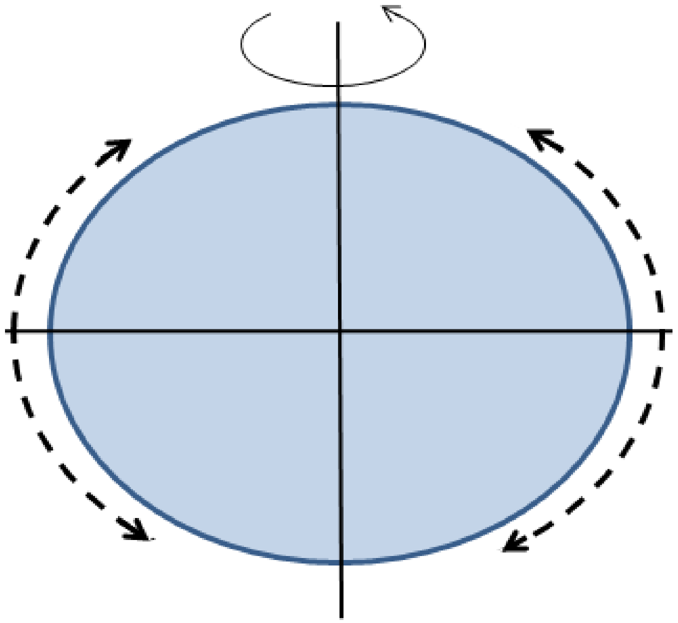
Contents:

- Free Air Correction
- Bouguer Correction
- Terrain (Topographic) Correction
- Bouguer anomaly

The following formula is used to calculate the theoretical value of gravity at any latitude on the Earth.

$[g_{Ø} = g_0(1+C1\sin 2Ø - C2\sin^2 2Ø)]$

Where $g_{Ø}$ is the theoretical value at a latitude Ø, g_0 is the theoretical gravity value at equatorial sea level (=978.0318Gal), C1 and C2 are constants (0.0053024 and 0.0000059 respectively).

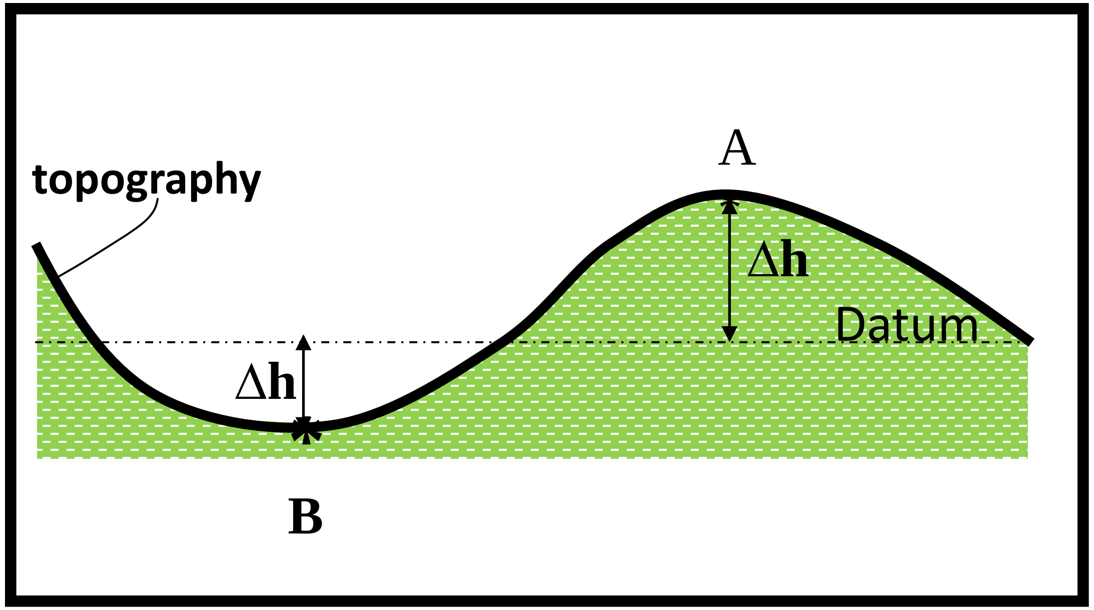


gravity increasing towards poles

This gradient of gravity variation with distance towards north or south is calculated by differentiating the formula:
$(\Delta g\Phi) = 0.812 \sin 2\,\Phi$ mGal. km-1

Free Air Correction:

Accounts for the height difference between the gravity station and a certain datum level. Gravity decreases with height because the gravitational acceleration is proportional to $1/r^2$. Therefore, the free air effect is added to the gravity value of a certain station if it is above datum and subtracted if below. No accounts are taken for rock density



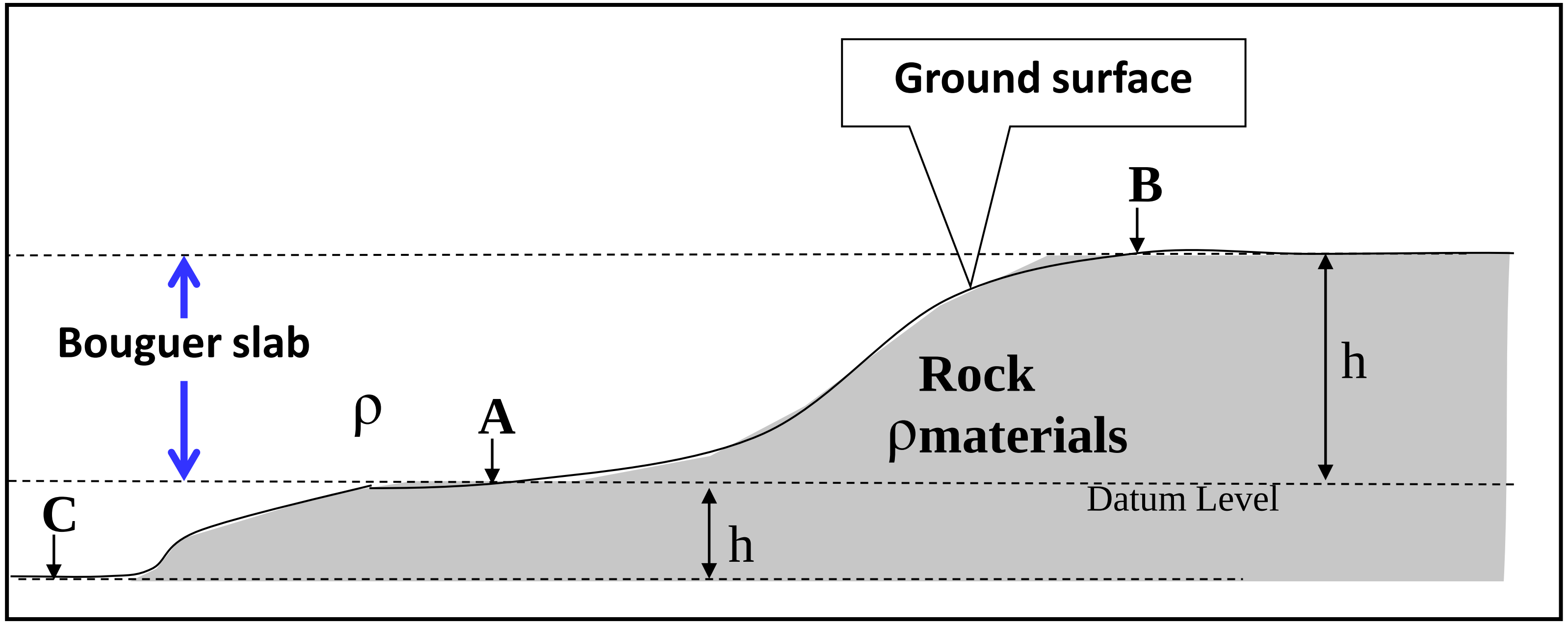
F.A.C.=0.3086mGal x Height difference(Δh)

Bouguer Correction:

Accounts for the effect of the attraction of rock materials between the station and datum levels, by approximating them into an infinite horizontal slab. This slab has a thickness equal to the elevation difference between the station and datum level with a homogeneous density, it is called Bouguer Slab.

$\Delta g = 2\pi G h \rho$
$2\pi G = 0.04191$

h is the Bouguer slab thickness having a density ρ
0.04191x Density x h



Alluvium (wet)	1.96–2.00
Clay	1.63–2.60
Shale	2.06–2.66
Sandstone	
Cretaceous	2.05–2.35
Triassic	2.25–2.30
Carboniferous	2.35–2.55
Limestone	2.60–2.80
Chalk	1.94–2.23
Dolomite	2.28–2.90
Halite	2.10–2.40
Granite	2.52–2.75
Granodiorite	2.67–2.79
Anorthosite	2.61–2.75
Basalt	2.70–3.20
Gabbro	2.85–3.12
Gneiss	2.61–2.99
Quartzite	2.60–2.70
Amphibolite	2.79–3.14
Chromite	4.30–4.60
Pyrrhotite	4.50–4.80
Magnetite	4.90–5.20
Pyrite	4.90–5.20
Cassiterite	6.80–7.10
Galena	7.40–7.60

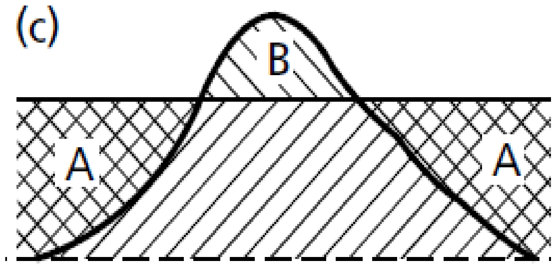
Typical mean values and ranges of density for some common rock types, density of rocks are important for the purpose of Bouguer correction and in interpretations. Unit is in gm/cm^{3}.

Gravity anomalies result from the difference in density, or *density contrast*, between a body of rock and its surroundings. For a body of density ρ_1 embedded in material of density ρ_2, the density contrast $\Delta\rho$ is given by

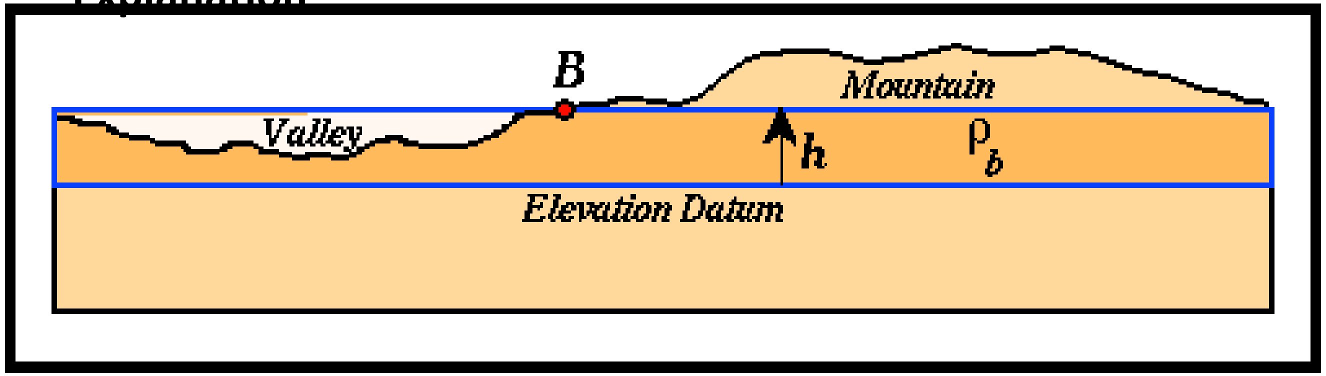
$$\Delta\rho = \rho_1 - \rho_2$$

Terrain (Topographic) Correction:

This correction accounts for extra mass above (hills, etc.), or deficit of mass (valleys, etc.) below a reading's elevation. The Bouguer correction makes the assumption that the topography around the gravity station is flat. This is rarely the case and a further correction, the terrain correction (TC), must be made to account for topographic relief in the vicinity of the gravity station. This correction is always positive as may be appreciated from consideration of the figures. The regions designated A (left figure) form part of the Bouguer correction slab although they do not consist of rock. Consequently, the Bouguer correction has overcorrected for these areas and their effect must be restored by a positive terrain correction. Region B consists of rock material that has been excluded from the Bouguer correction. It exerts an upward attraction at the observation point causing gravity to decrease. Its attraction must thus be corrected by a positive terrain correction.



Explanation



Bouguer Anomaly

The end product of gravity data corrections is the Bouguer Anomaly (B.A.), which should correlates only with lateral variations in density of the upper parts of the crust and which are of most interest to applied geophysicists and geologists.

The B.A. is the difference between the observed gravity value (g_o) adjusted by the algebraic sum of all the necessary corrections, and that at a certain base station (g_{base})

$$\Delta gb = g_o + \sum (\text{all corrections}) - g_{base}$$

$$= (g_o - g_{base}) + [\Delta \text{ Drift C.} + \Delta (\text{FAC} - \text{BC}) +/- \Delta\text{LC} + \Delta\text{TC}] = \text{B.A.}$$

This formula is used for relative gravity measurements.

If we bring to our mind the procedure of transferring the absolute gravity value from a primary base station to the secondary base station at the area of work (an example given ten slides before), i.e., if the absolute value of the secondary base station is known, the formula of the Bouguer anomaly will be:

B.A.= $(g_{abs} - g_{theo.}) + [\Delta \text{ Drift C.} + \Delta \text{ (FAC} - \text{BC)} +/- \Delta LC + \Delta TC]$

Where (g_{abs}) is the absolute value calculated in the secondary base station

And g_{theo} is the theoretical gravity value at other observing stations calculated using the international formula of the theoretical value of gravity at any point on the earth (here will be the station):

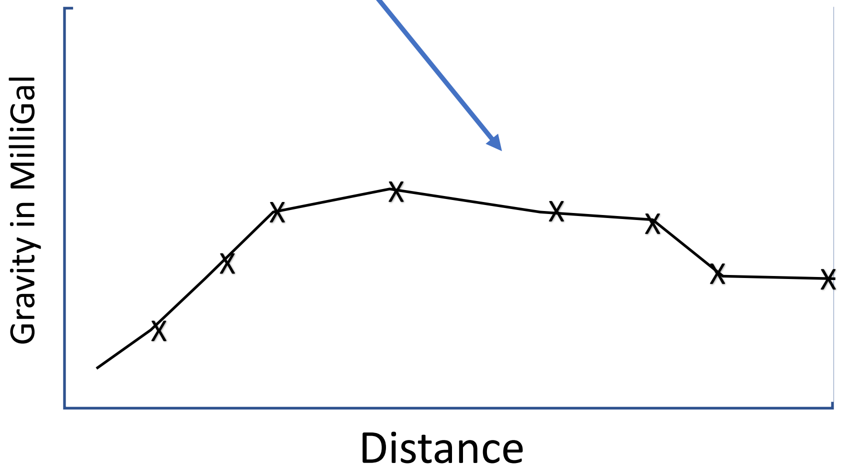
$$g_\phi = g_0(1 + k_1 \sin^2\phi - k_2 \sin^2 2\phi)$$

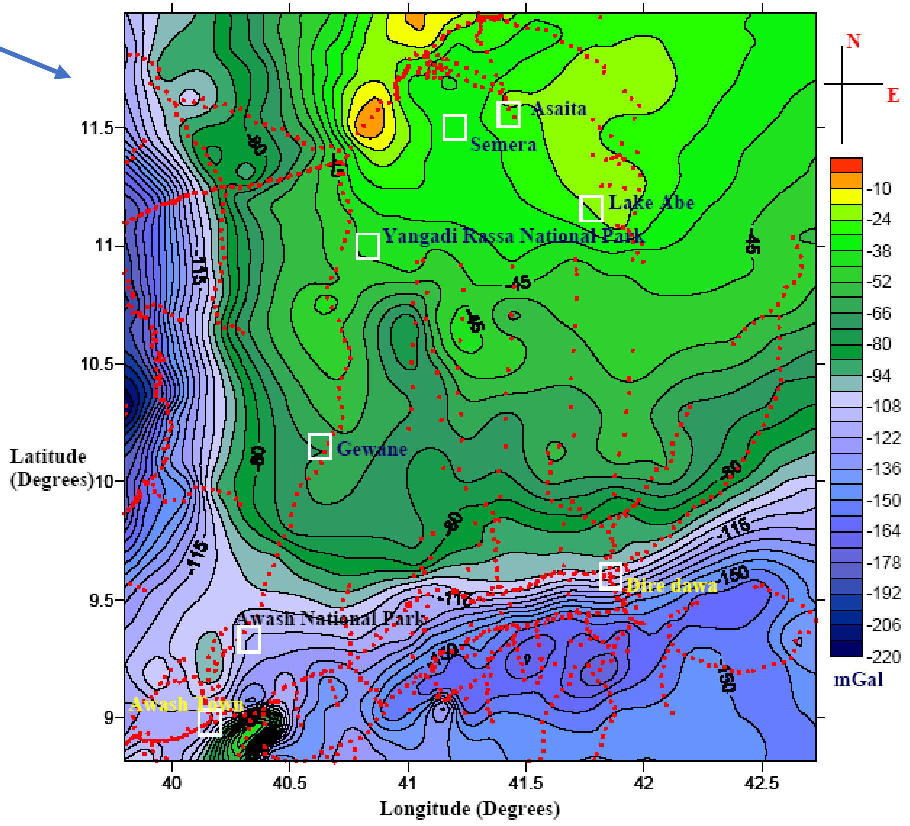
$$g_\phi = 9\,780\,318.5\,(1 + 0.005278895 \sin^2\phi + 0.000023462 \sin^4\phi) \text{ gu}$$

Where ø is the latitude value at the station. (1mGal=10g.u.)

Question: By your mobile see what your place's latitude is and calculate the theoretical value

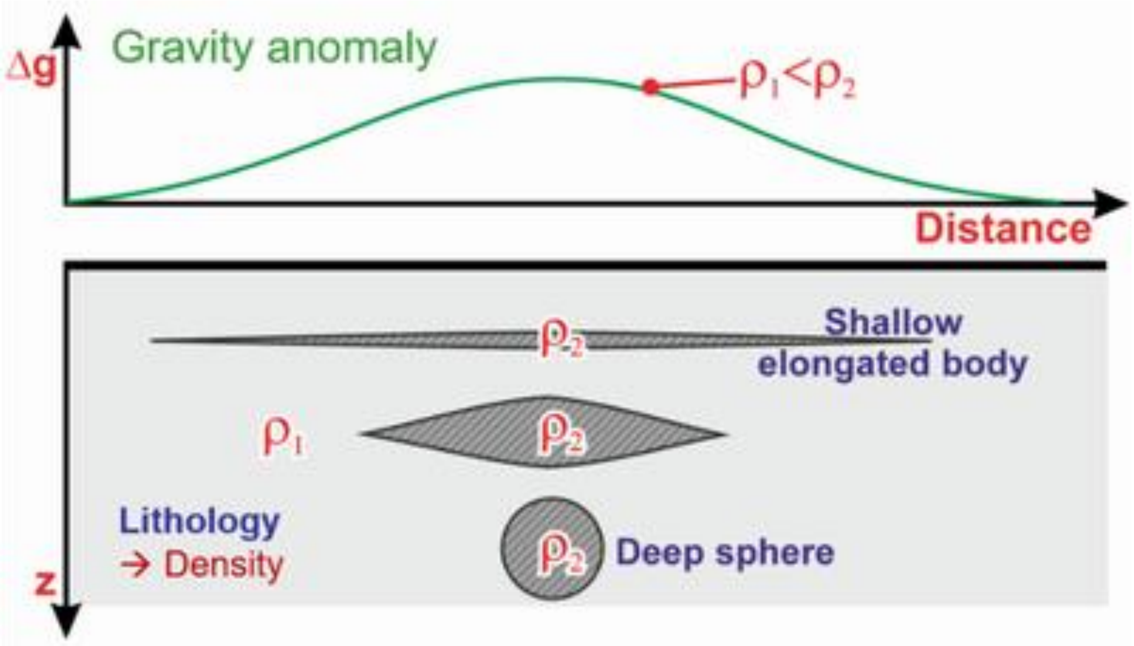
Once the Bouguer Anomaly is calculated for each observation station after corrections, a profile or map can be drawn. They are called Bouguer Anomaly Profile or Map





The problem of ambiguity

Although for any given body, a unique gravity field is predicted, a single gravity anomaly may be explained by an infinite number of different bodies, e.g., spheres and point masses. Because of this, it is important to use additional information from surface outcrops, boreholes, mines and other geophysical methods. The meaning of gravity data is dependent on how much other information is available.



Any of the three bodies may produce the gravity anomaly shown at the top.

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Lecture 7

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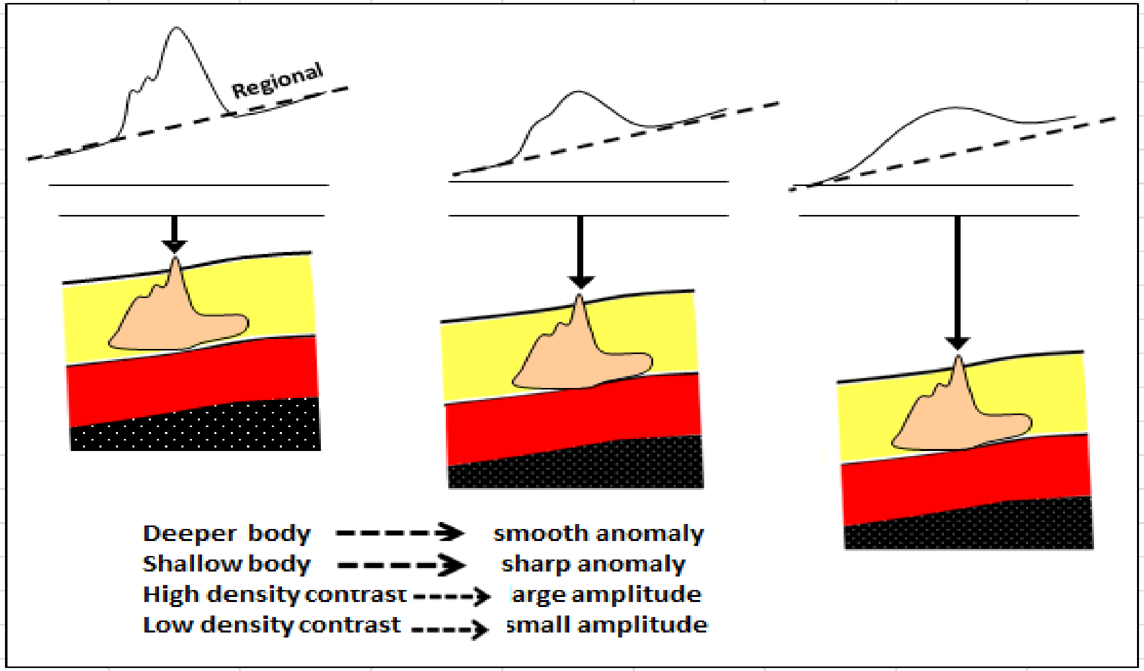
2023/2024

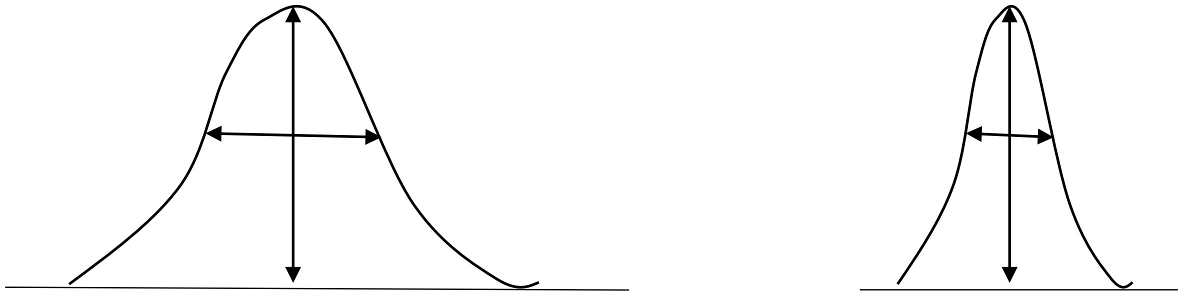
Contents:

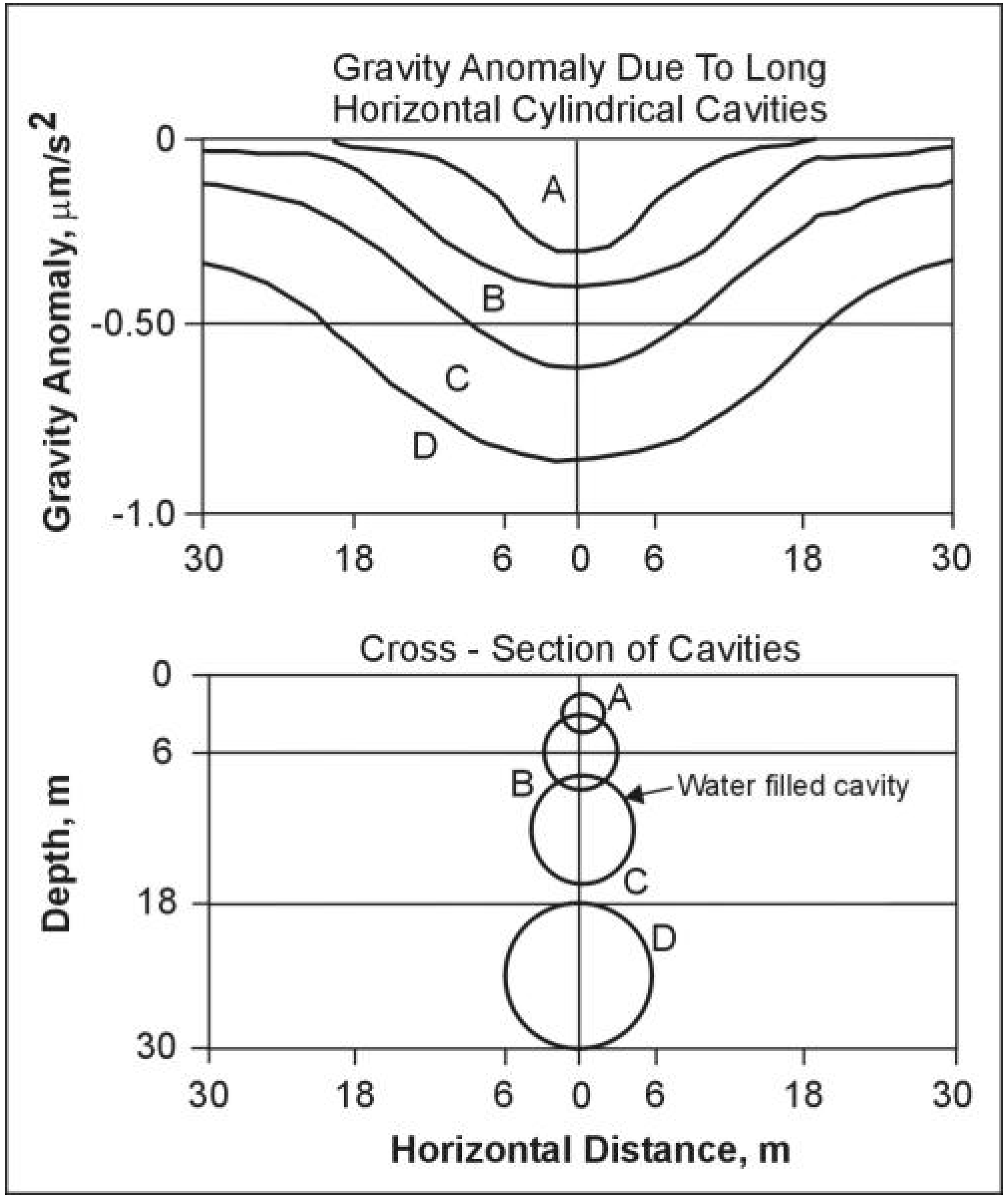
Regional and Residual Gravity

The Bouguer map is the sum of gravity effects resulting from density changes (geological anomalies) existing in the subsurface medium of the surveyed area. These changes, expressed by the gravity anomalies, are produced from lateral changes in density. The amplitude of the anomaly is function of both the density difference (density contrast) and the depth of the responsible geological structure. In fact, the anomaly amplitude gets larger with the increase of density contrast and with the decrease of the depth of the anomalous body.

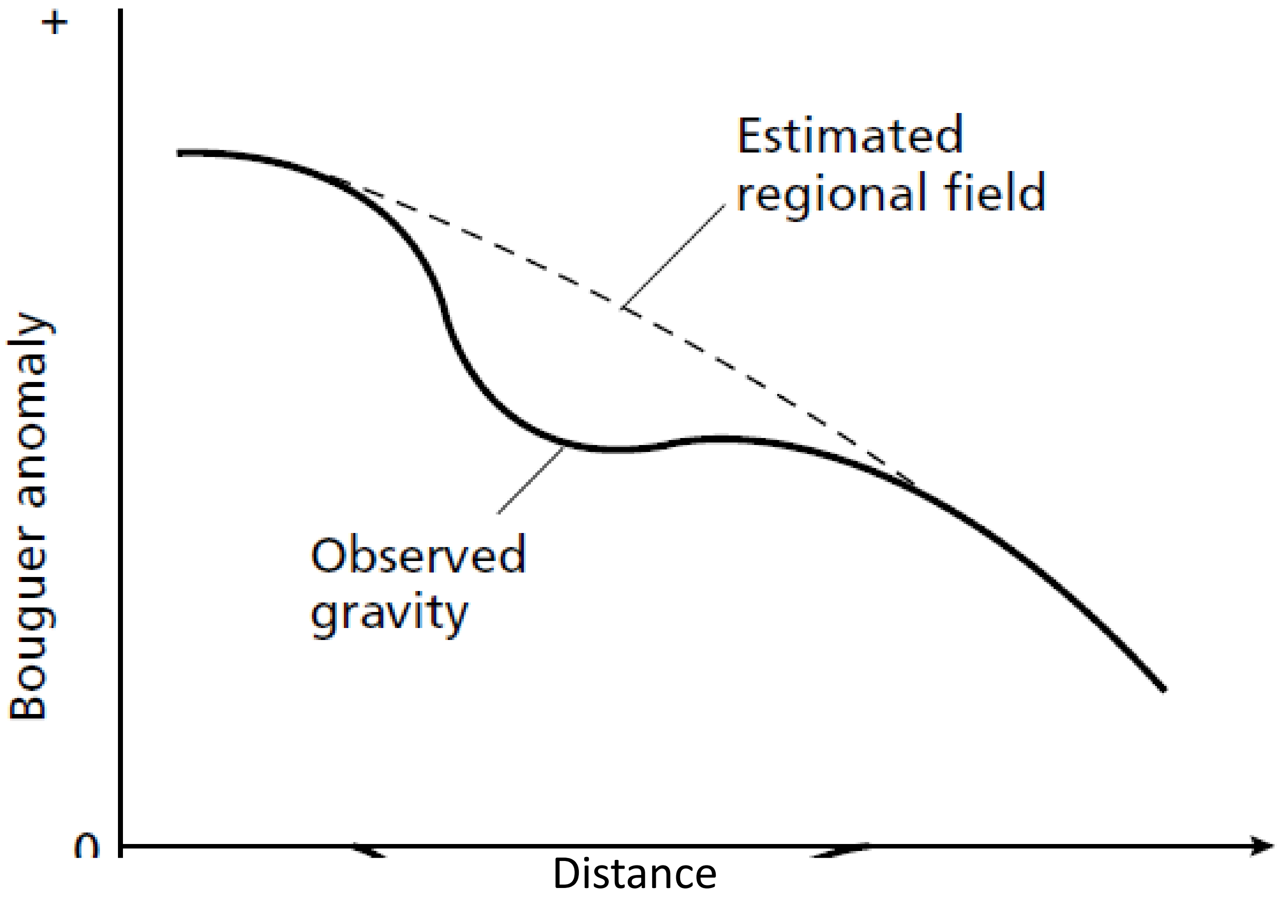


<u>Rule:</u> For a given anomalous mass, the amplitude of the gravity anomaly, its smoothness and width are governed by depth of the mass. As the depth increases, the resulting gravity anomaly gets wider, weaker, and smoother.

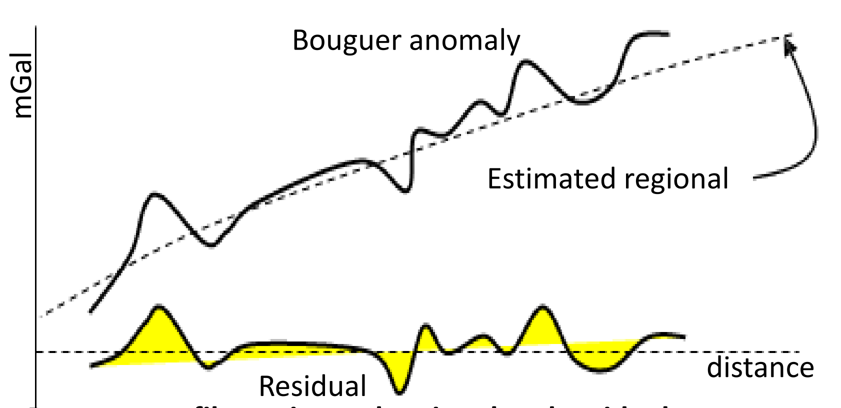




- There may be a gentle trend in the gravity data, reflecting a long-wavelength gravity anomaly attributable to deep-seated crustal features (such as Basement rocks); this is known as a <u>regional anomaly</u>.
- Shorter-wavelength anomalies arising from shal-lower geological features are suprimposed on the regional anomaly, and these anomalies are often isolated for further analysis, they are called <u>residual anomaly</u>.
- Separation of the regional from the Bouguer anomaly will leave a residual anomaly. The deeper the body the broader the anomaly. The interpreter may wish to emphasize some anomalies and suppress others, e.g., shallow anomalies are important to mineral exploration (for Mining purposes), and deep anomalies are important for oil exploration.
- <u>The problem lies in separating out the two effects, and it is not strictly possible to do this without affecting what is left.</u>

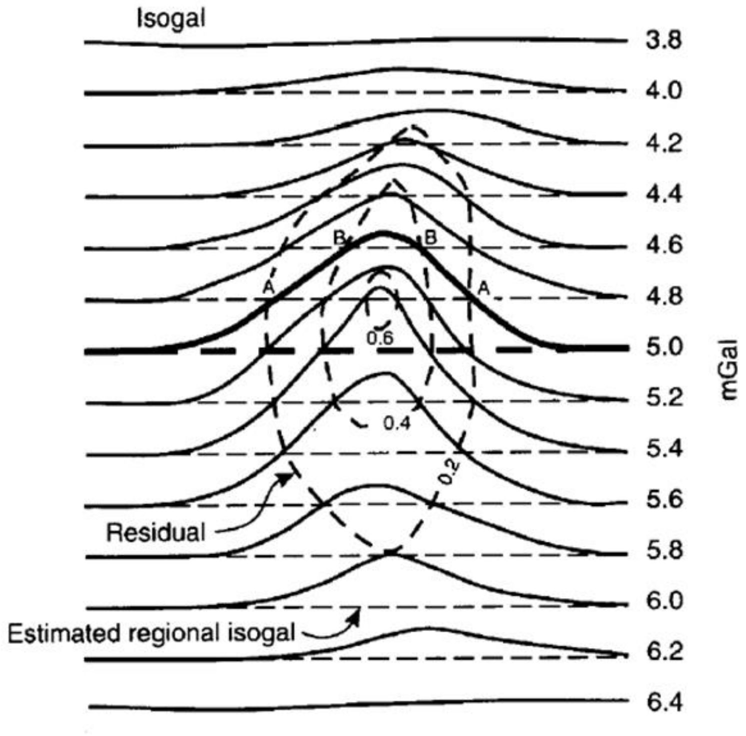


Process of separation may be performed graphically by sketching in a linear or curvilinear field by eye. Such a method is biased by the interpreter, but this is not necessarily disadvantageous as geological knowledge can be incorporated into the selection of the regional field.



Bouguer profile, estimated regional and residual

Whether it is a profile or a contour map, the graphical smoothing technique is basically dependent on personal judgment. For this reason, the computed regional and residual variations may differ from one interpreter to another. The extent of difference depends on the degree of complexity of the given Bouguer gravity data and on the interpreter individual skill.

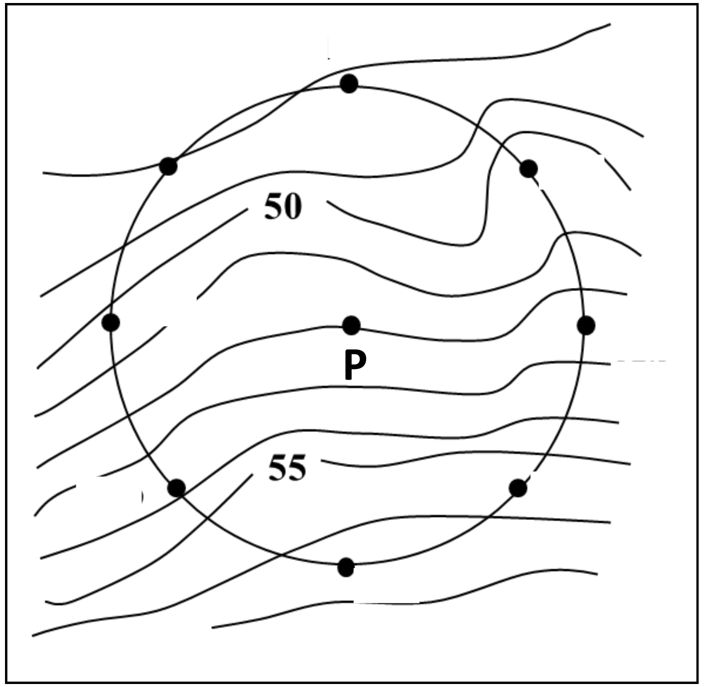


Bouguer anomaly map, regional and residual

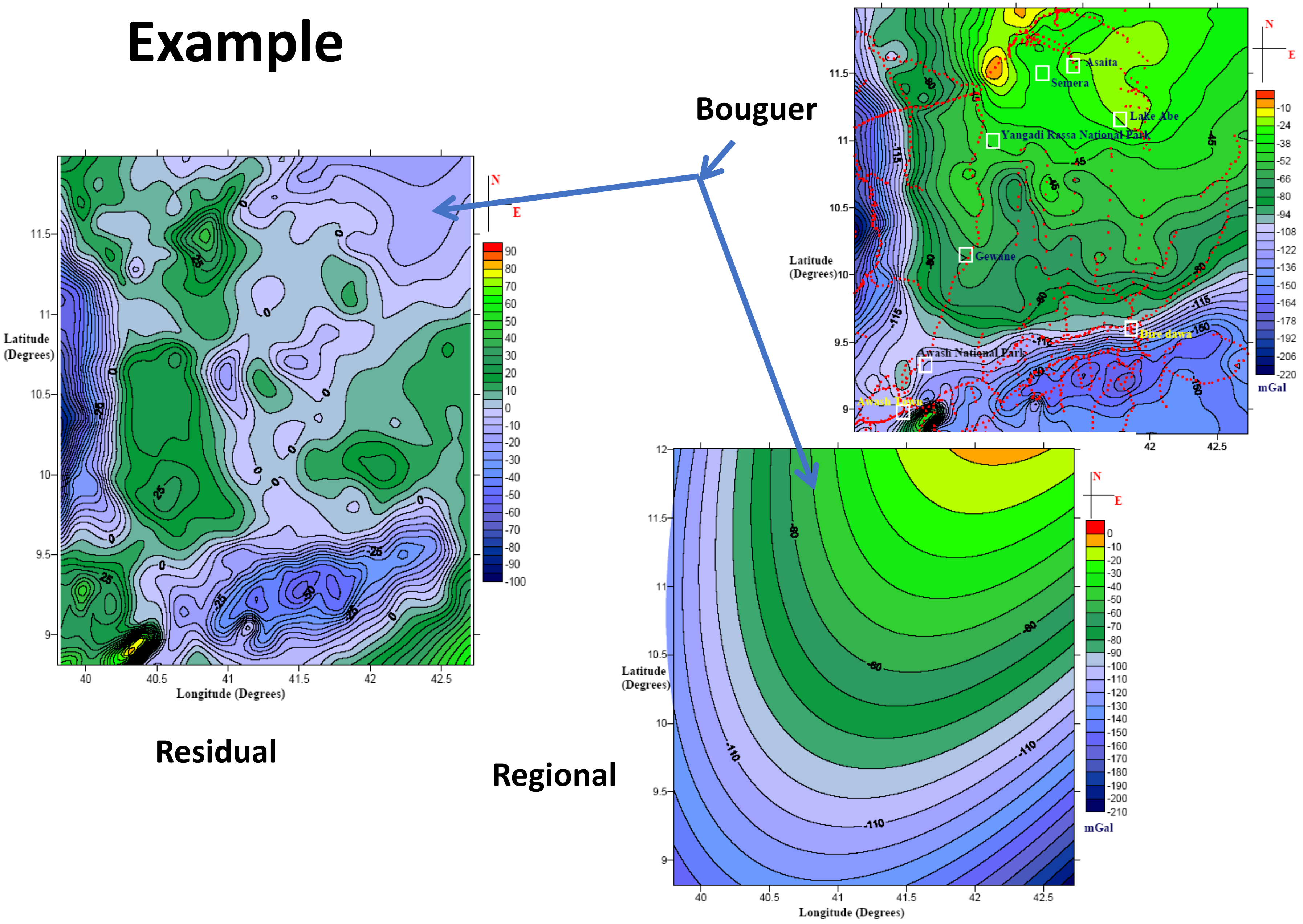
There are many arithmetic and analytical methods, among them is the **Griffin's method**. It serves as a direct way to compute residual gravity from a given Bouguer gravity map. It involves averaging of the gravity values of a set of points on the gravity map which are located at equal distances from the point at which the residual is to be computed. The residual gravity value at a point P in Figure is equal to the observed Bouguer value (g_P) at that point minus the computed average of the gravity values at a set of points which are equally spaced about the circumference of a circle of a suitable radius. The process is repeated for all observation points in the survey area.

Regional = (56.4+55.6+52.5+50.3 +48.1+48+50.2+53.9)/8 = 51.87
Residual = 52-51.87 = 0.13

Explanation; if radius approaches zero, regional will approximate the Bouguer value in the center and residual will approach zero.



Example

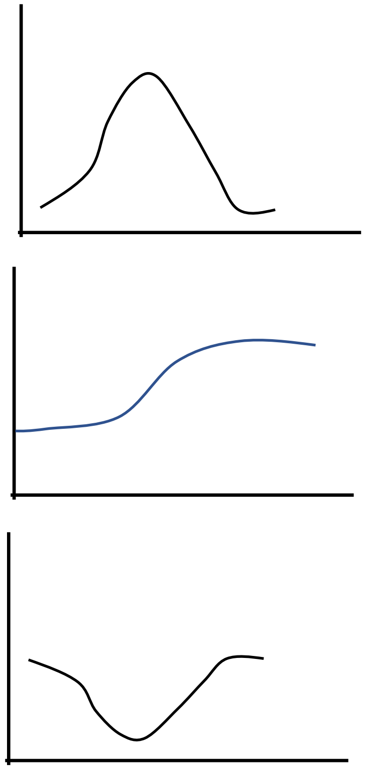


Bouguer
Residual
Regional
Asaita
Semera
Lake Abe
Yangadi Rassa National Park
Gewane
Dire dawa
Awash National Park
Latitude (Degrees)
Longitude (Degrees)
mGal
N
E

Interpretation of Gravity Data:

Interpretation of gravity data is to translate these data obtained after corrections into geology, which means size, shape and the depth of the subsurface structures that give rise to gravity anomaly.

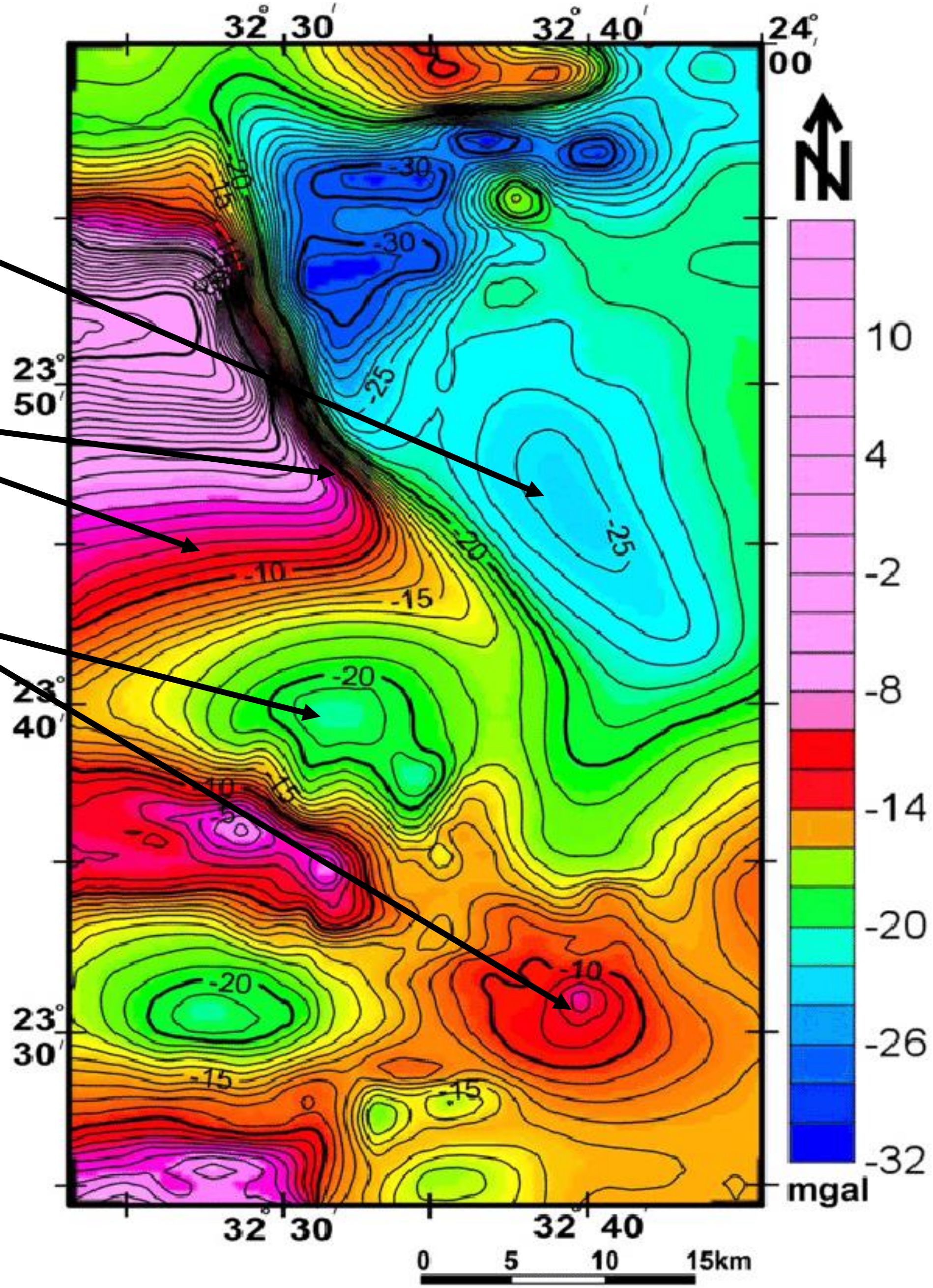
It is necessary before carrying out interpretation to differentiate between two-dimensional and three-dimensional anomalies on Bouguer anomaly maps. Two-dimensional anomalies are elongated in one horizontal direction so that the anomaly length in this direction is at least twice the anomaly width. Such anomalies may be interpreted in terms of structures which theoretically extend to infinity in the elongate direction by using profiles at right angles to the strike. Three-dimensional anomalies may have any shape and are considerably more difficult to interpret quantitatively.



2-Dimensional anomaly

Linear anomaly

3-Dimensional anomaly



From profiles we can not know while the Anomaly is 1-D or 2-D

Iraqi Kurdistan Region
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Lecture 8

Applied Geophysics Lecture Notes
Fourth Semester
By:
Dr. Fadhil Ali Ghaib
2023/2024

Contents:

Bouguer maps are interpreted in two stages; <u>qualitative and quantitative</u>

<u>Qualitative interpretation</u> involves the description of the Bouguer anomaly map or profile and their transformed regional and residuals. The description involves the dimension (i.e., 2D, 3D, Linear or any other shape), trend, amplitude, gradient, shape, location relative to study area, relative density and possible geologic source after integrations with any other information.

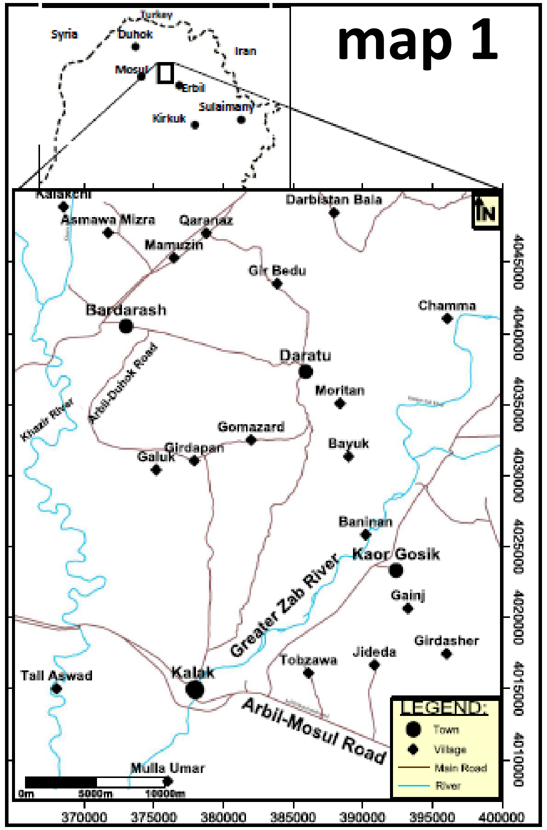
<u>General aspects for qualitative interpretation:</u>

1-Surface geological information should be collected as much as possible.
These Information are; rock types, major structural trends, jointing, faulting, folding…etc.
2- Maps such as topographic, geologic, gravity…etc. are better to be of one scale for the purpose of overlays and correlations.
3- Divide the map into sectors to describe separately and then collectively.
4- Correlate sectors with geology.
5- Delineate important anomalies that should be quantitatively interpreted.

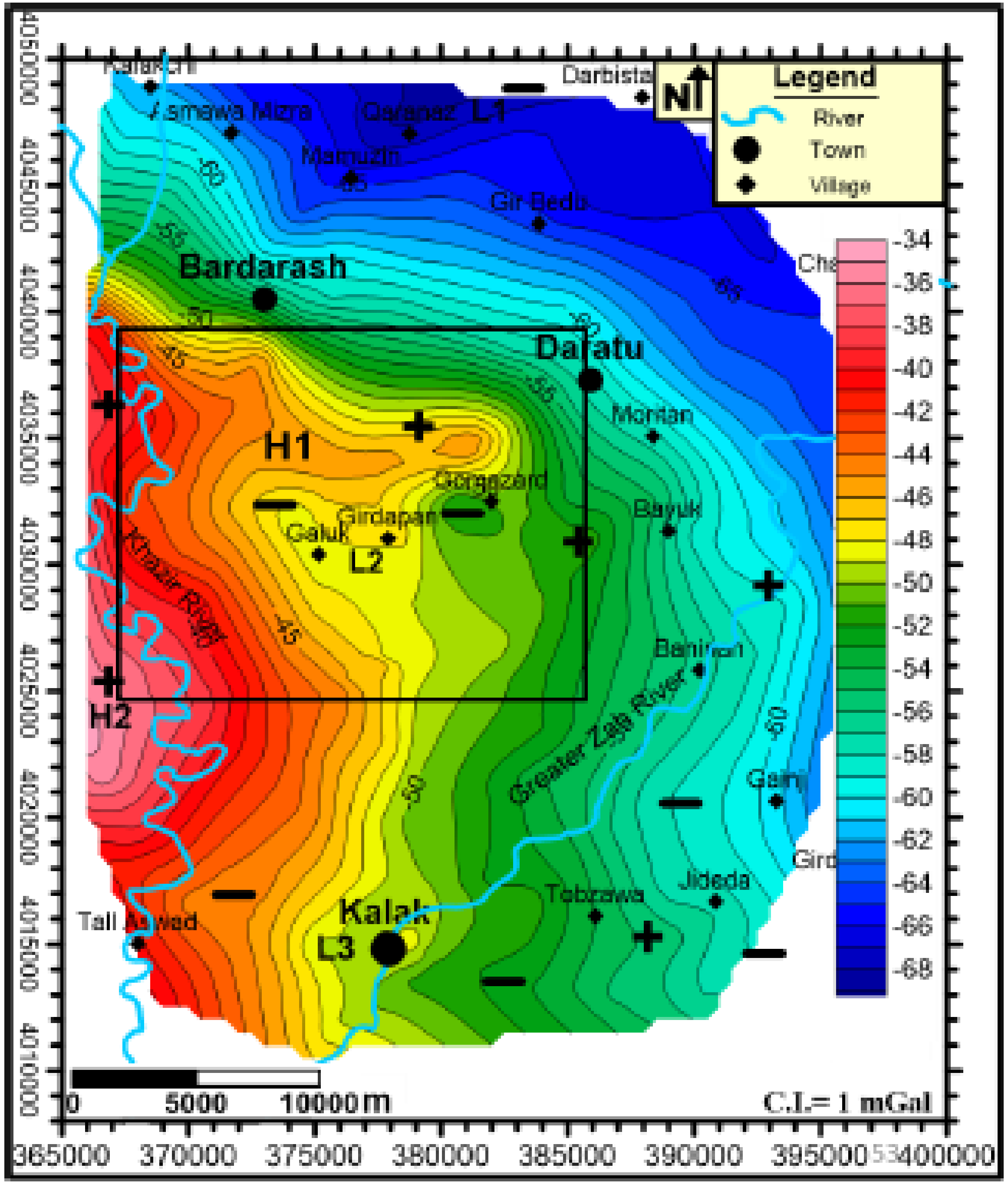
Homework:

Interpret qualitatively the given Bouguer anomaly map in details. Write a report Including:

- Introduction (talking about principles of Gravity method)
- Location of the area (Draw the given location map-describe the locality of the studied area –map 1)
- Procedure of collecting data (from your lecture notes and textbooks, write how we collect data)
- Definition of Bouguer anomalies (talk about the corrections of data to be ready for drawing Bouguer Anomaly Map)
- Qualitative interpretation (Describe the map depending upon points given in the previous slide, map 2).
- Using hand smoothing method separate residual anomalies for the part inside the box, i.e. anomalies named H1 and L2, map 3.

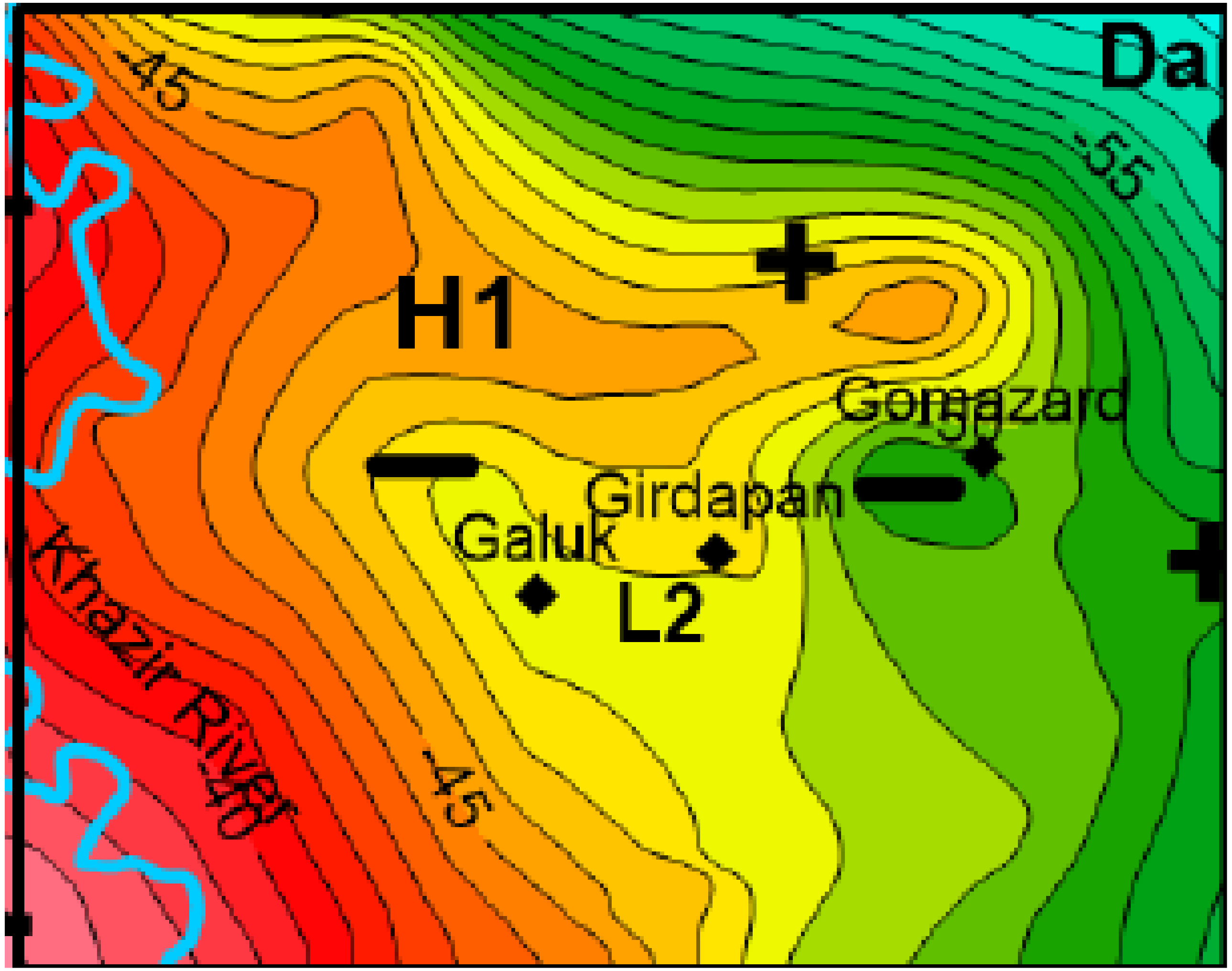


Map 1: Location of the studied area



Map 2: Bouguer Anomalies in Kalak-Bardarash Area

map 3



Map 3: Bouguer Anomalies around Girdapan Village

<u>Quantitative interpretation</u> is the determination of numerical values for the causative bodies in the subsurface.

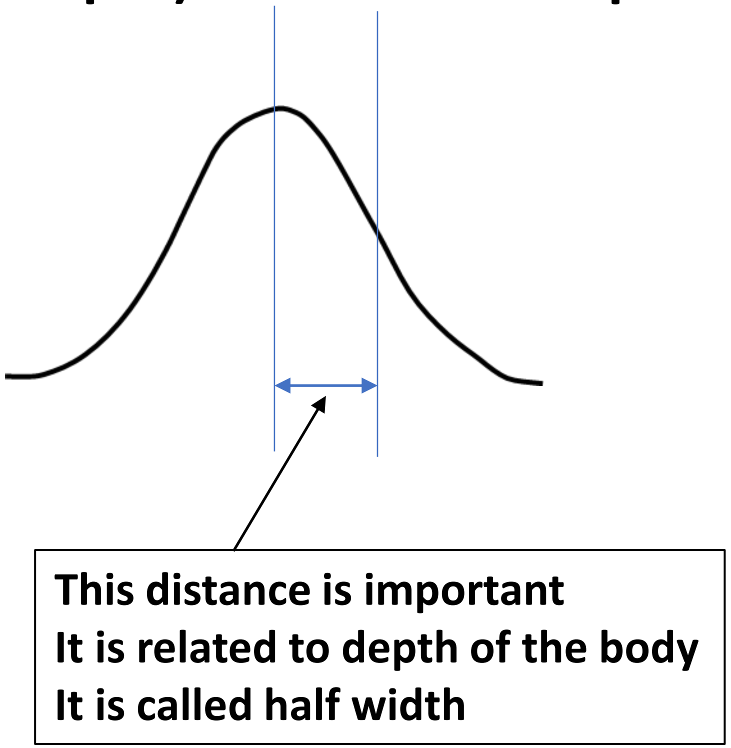
- **The first step in the quantitative interpretation is the visual inspection of the residual map to choose the profile across the anomaly of interest.**

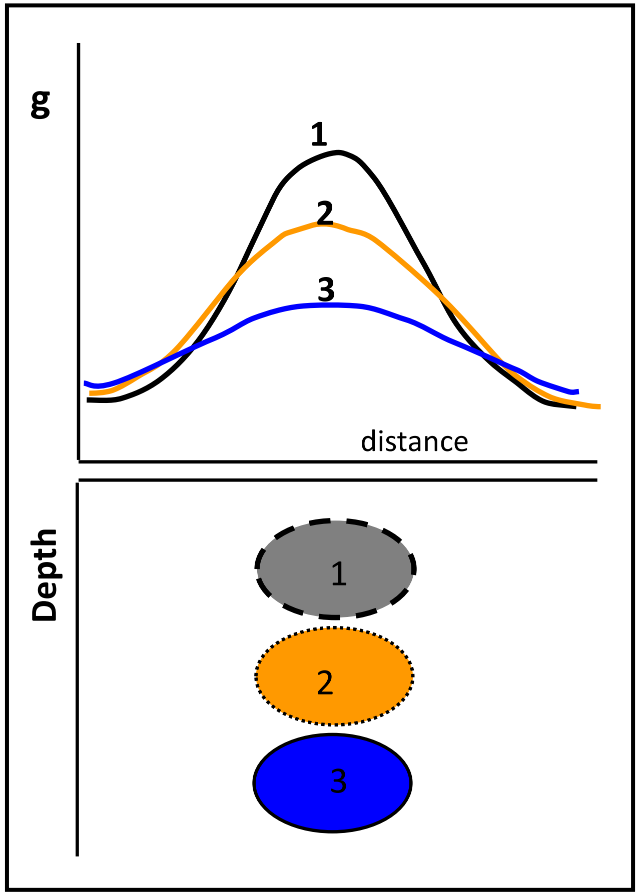
- **The second is to estimate approximately the horizontal extension, depth, shape and thickness of target using a geological background (Well logging, seismic sections, previous studies).**

- **The third step is to construct a geometric model which satisfies the above-mentioned estimations and is consistent with the geologic situation by using recent computer programming**

Define the causative bodies

You should remember and know that Gravity anomalies decay with the inverse square of the distance (depth) from their source so that anomalies caused by deep structures are of lower amplitude and greater extent than those caused by shallow sources. This wavenumber–amplitude relationship to depth may be quantified to compute maximum depth (or limiting depth) at which the top of the anomalous body could be situated.



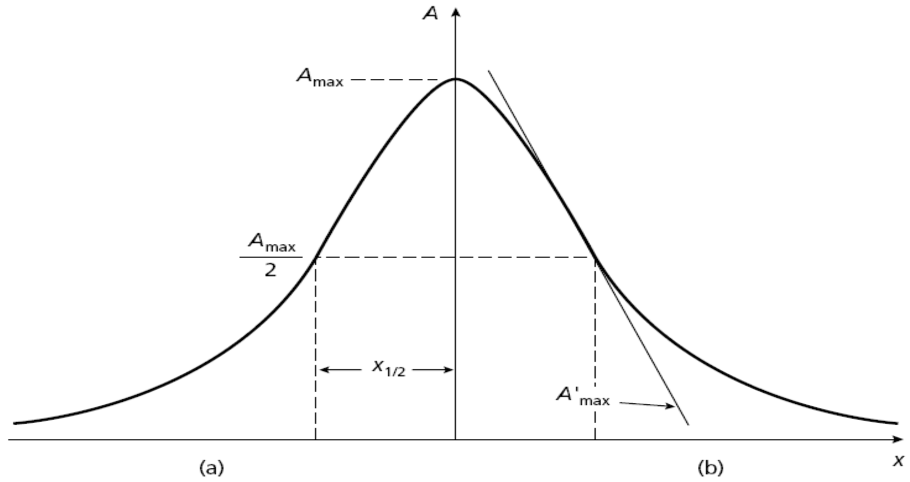


According to the three steps discussed three slides before, we can divide the quantitative interpretation activities to two stages; Direct quantitative interpretation and indirect interpretation

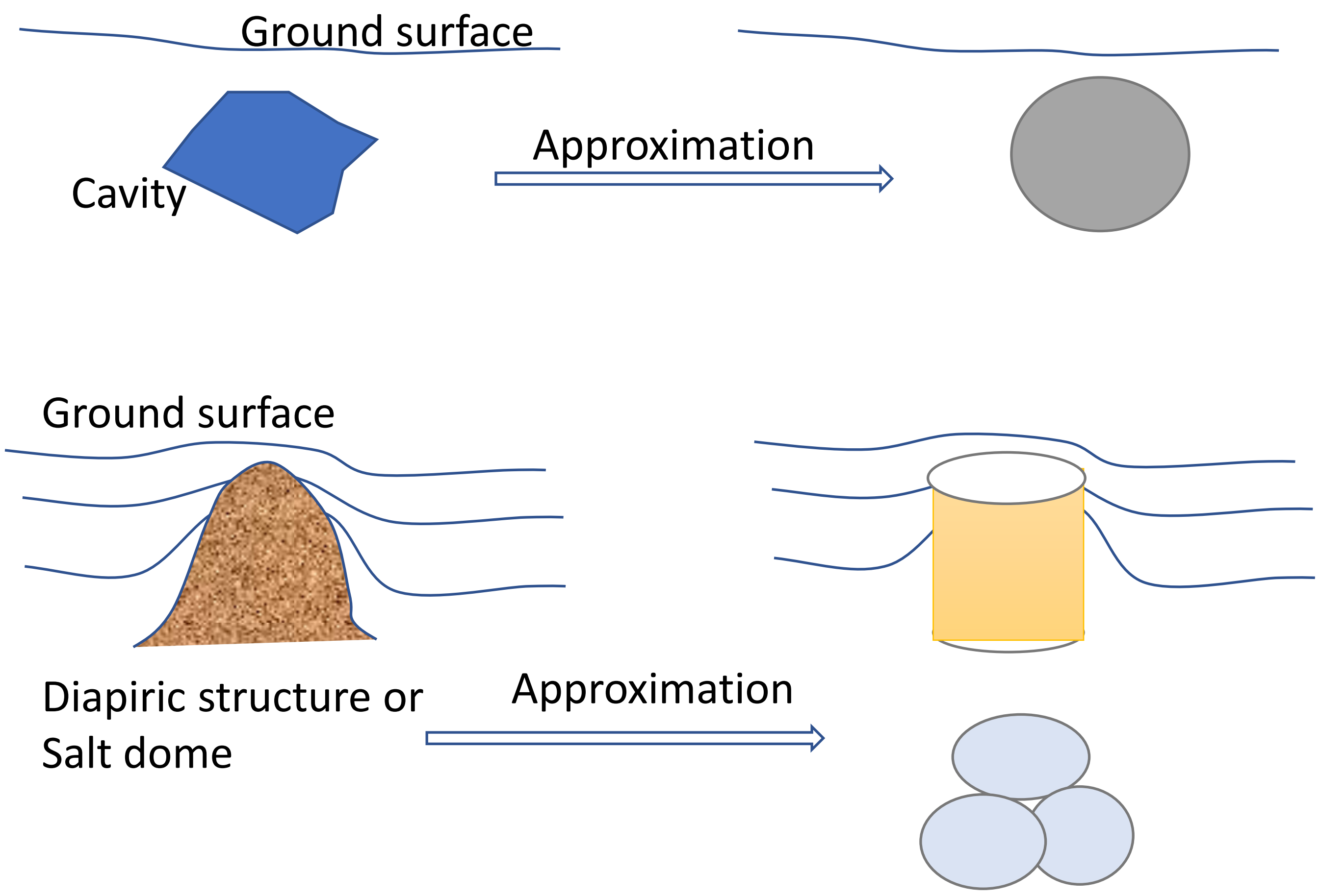
The first is the direct use of the anomaly parameters. The original data are analyzed to produce an interpretation.

Anomaly parameters are:

- **Amplitude (A_{max})**
- **Half width ($X_{1/2}$)**
- **Half-maximum amplitude ($A_{max/2}$)**
- **Maximum slope or gradient (A'_{max})**



Certain geologic structures can be approximated to models with known geometric forms. For example, a buried cavity may be represented by a sphere, a salt dome by a vertical cylinder, a basic igneous dyke by an inclined sheet or prism, etc.

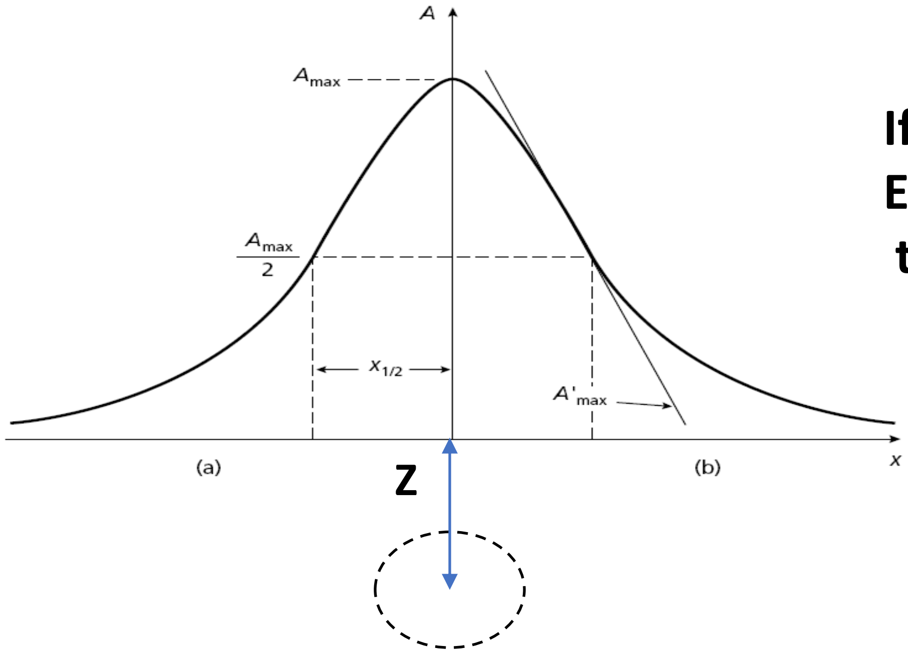


Some methods of direct interpretation

(a)Half-width method: The half-width of an anomaly ($x_{1/2}$) is the horizontal distance from the anomaly maxi- mum to the point at which the anomaly has reduced to half of its maximum value. Because the theoretical assumption is based on center point of a sphere and the center-line of a horizontal cylinder , the depth Z is assumed to be from datum to center of the causative bodies.

If the anomaly is three-dimensional
Ex. A sphere

$$z = \frac{x_{1/2}}{\sqrt{\sqrt[3]{4} - 1}}$$



If two-dimensional body
Ex. A cylinder perpendicular to page

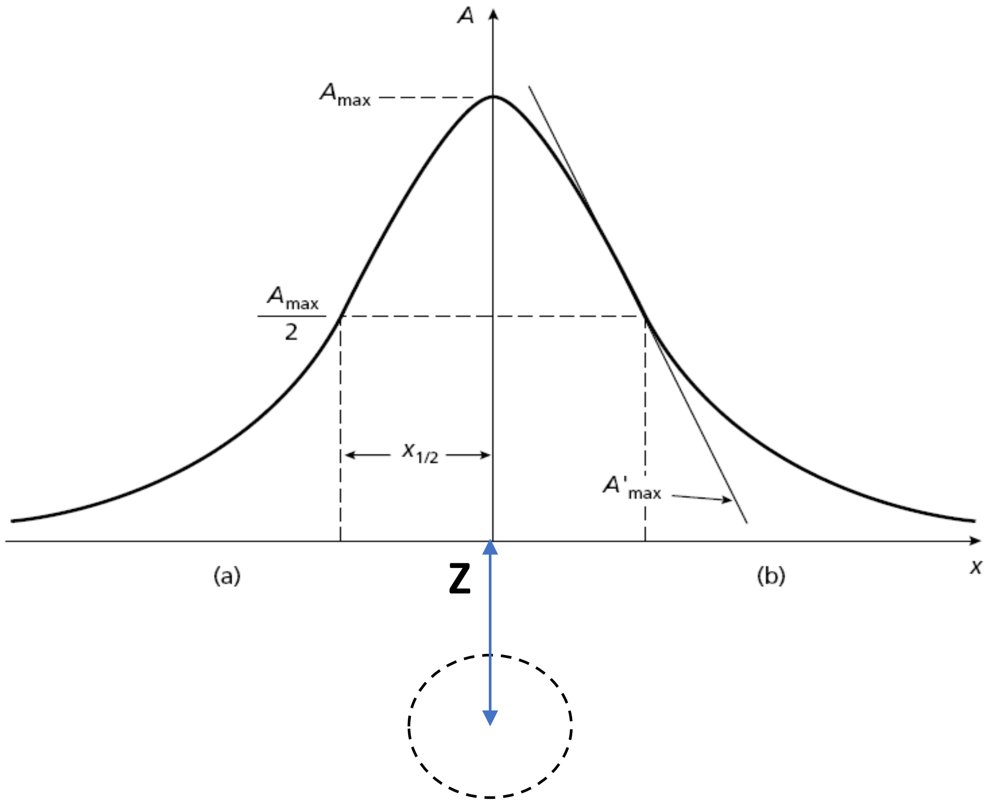
$$z = x_{1/2}$$

(b)Gradient–amplitude ratio method.

This method requires the computation of the maximum anomaly amplitude (A_{max}) and the maximum horizontal gravity gradient (A'_{max})

If three-dimensional body

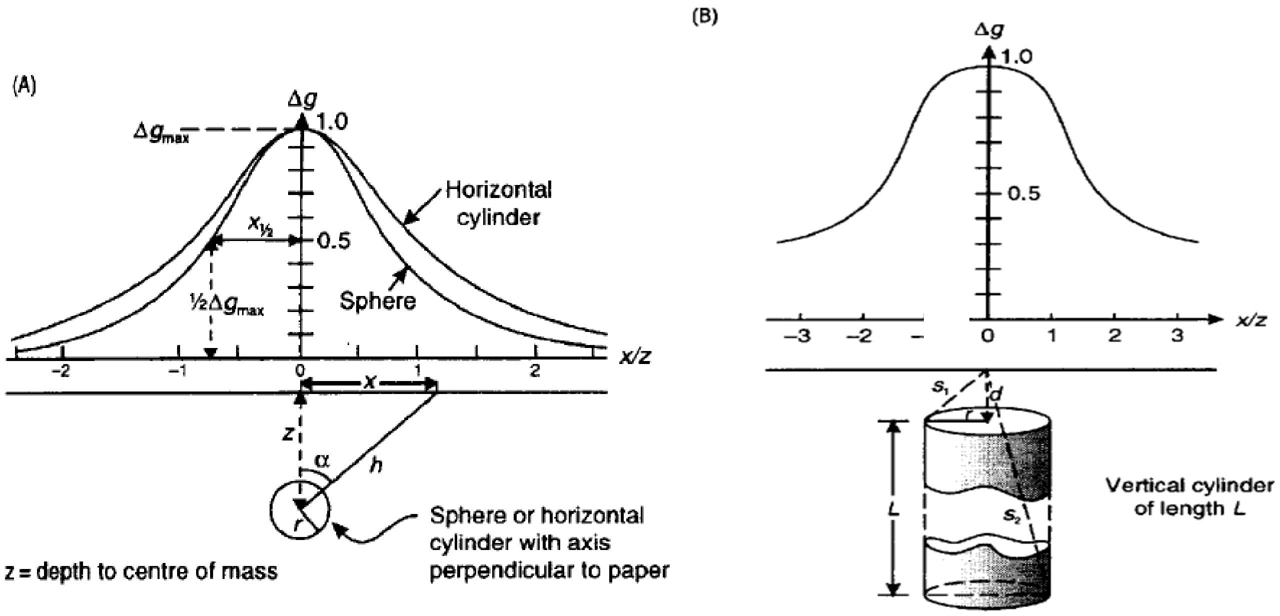
$$z < 0.86 \left| \frac{A_{\max}}{A'_{\max}} \right|$$



If two-dimensional body

$$z < 0.65 \left| \frac{A_{\max}}{A'_{\max}} \right|$$

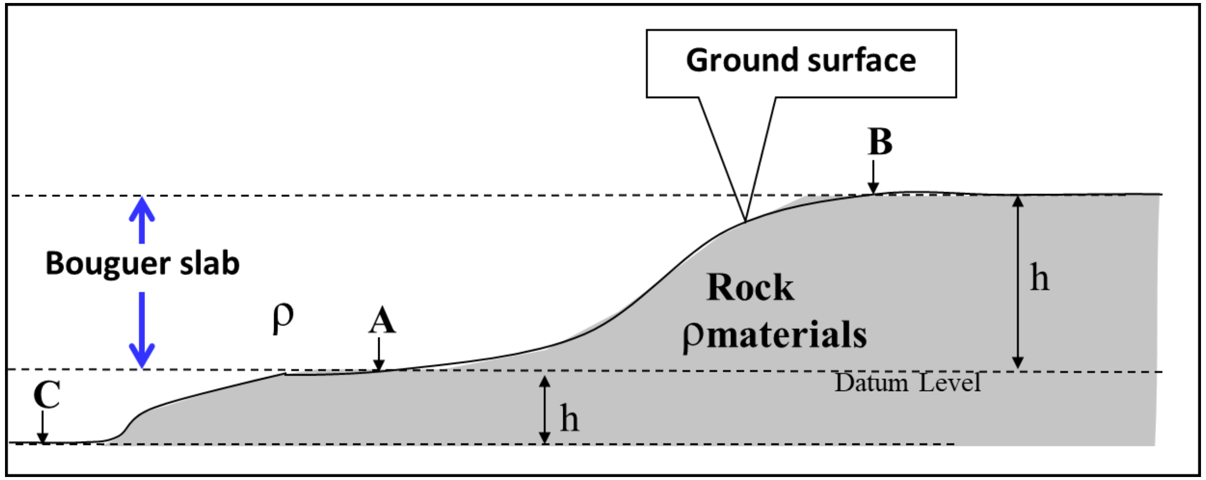
Geometry	*Formula*	*Depth*
Ball	$\Delta g = \dfrac{4\pi G R^3 \Delta\rho}{3z^3} \dfrac{1}{\left[1+\left(x^2/z^2\right)\right]^{3/2}}$	$z = 1.305 x_{1/2}$
Horizontal cylinder	$\Delta g = \dfrac{2\pi G R^2 \Delta\rho}{z} \dfrac{1}{\left[1+\left(x^2/z^2\right)\right]}$	$z = 1.0 x_{1/2}$
Vertical cylinder	$\Delta g = \dfrac{\pi G R^2 \Delta\rho}{\left(x^2+z^2\right)^{1/2}}$	$z = 0.58 x_{1/2}$



Approximate thickness

If the density contrast (σ) of an anomalous body is known, its thickness *t* may be crudely estimated from its maximum gravity anomaly Δg_{max} by making use of the Bouguer slab formula

(remember Bouguer correction).



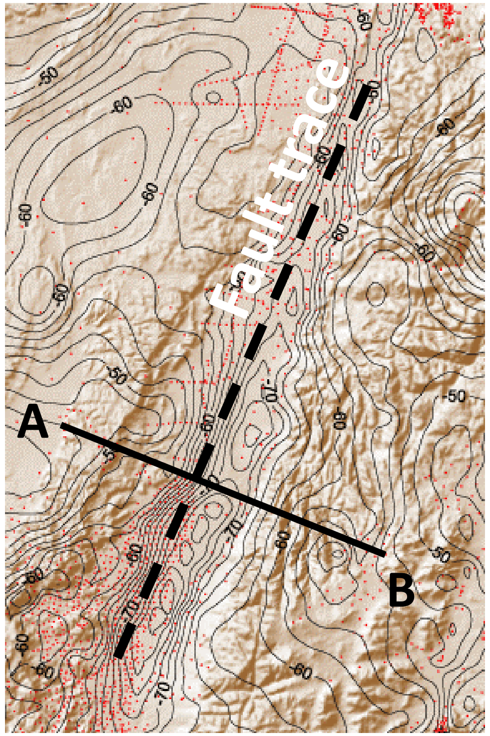
$\Delta g = 2\pi G h \rho$

$2\pi G = 0.04191$ then $\Delta g = 0.04191 h\rho$

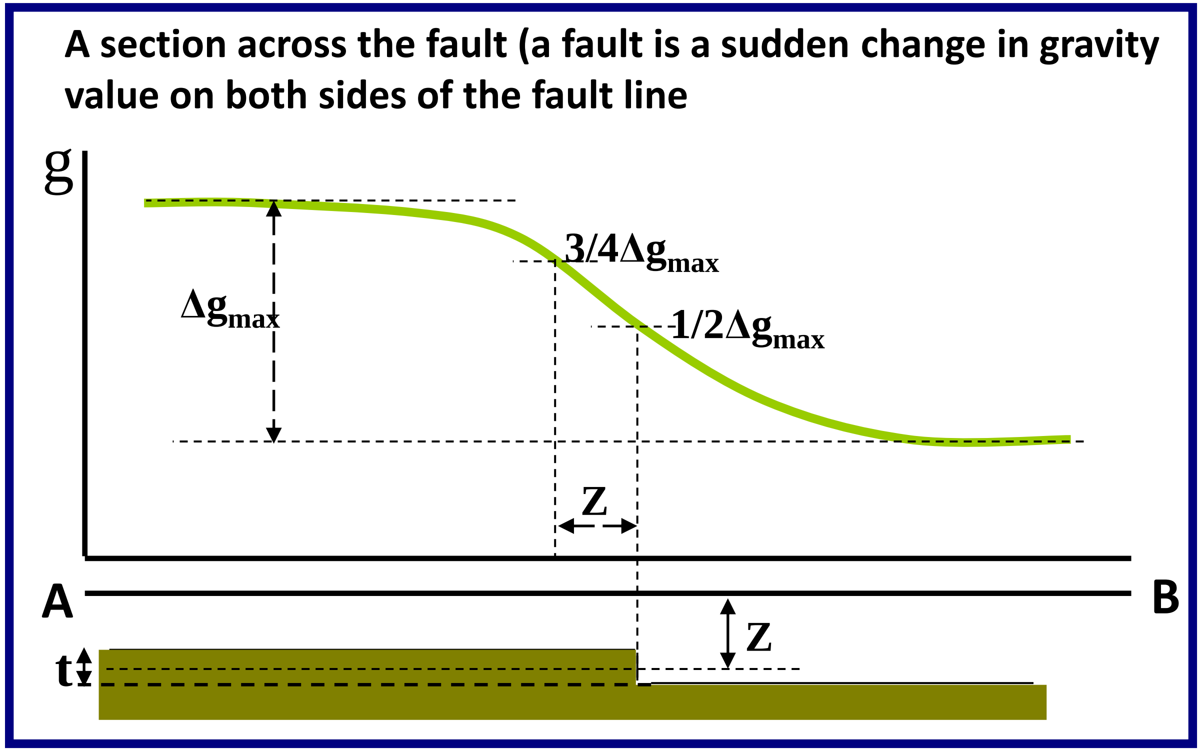
Replace ρ by Δρ which is the density contrast between the causative body and surroundings.

This thickness will always be an underestimate for a body of restricted horizontal extent. The method is commonly used in estimating the throw of a fault from the difference in the gravity fields of the upthrown and downthrown sides.

(c) For a step model anomaly (i.e., fault) also the half width with the maximum gravity value could be used.



Fault shape on a map



The horizontal distance Z is approximately equal to the depth to center of the slab having a thickness t. This thickness could be calculated using the Bouguer slab formula (t = 0.04191σ /Δg_{max}); σ is density contrast.

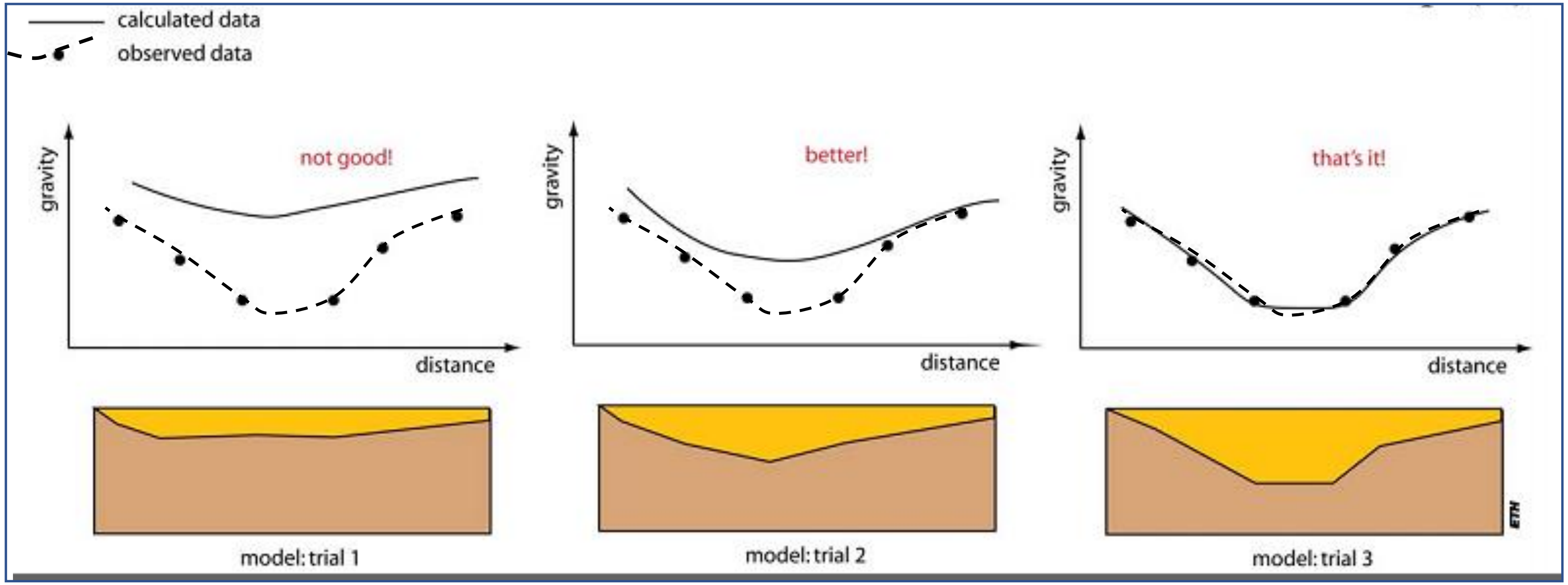
Calculation of gravity anomalies using the above methods should be regarded as a first step in the direct interpretation process. There are other, more sophisticated, and commonly computerized methods of gravity anomaly analysis.

Indirect interpretation

In indirect interpretation, the causative body of a gravity anomaly is simulated by a model whose theoretical anomaly can be computed using formulas which some of them are given in previous slides , and the shape of the model is altered until the computed anomaly closely matches the observed anomaly. Altering models should be in consistence with the geology of the area.

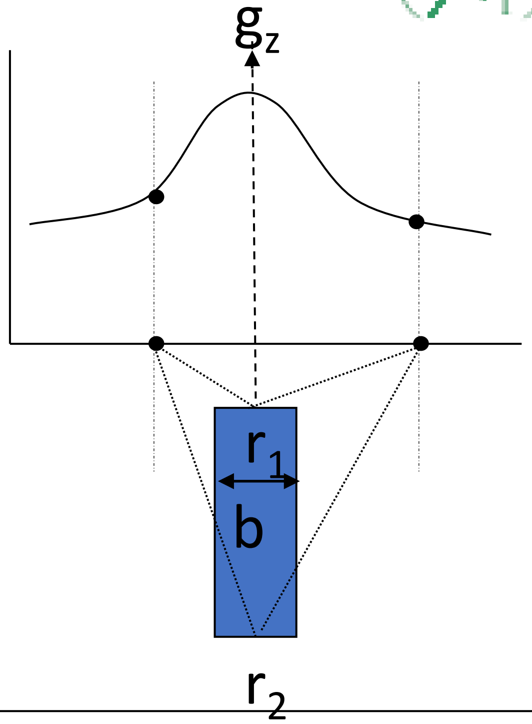
Steps of modeling are:

1- Construction of a reasonable model (model trial 1)
2- Computation of its gravity anomaly using standard formulas (model trial 1)
3- Comparison of the calculated anomaly with the original Bouguer anomaly model trial 2)
4- When there is a poor matching between them alter the shape of the model as accepted geologically and compute again (model trial 3).
5- There might be many steps of trials before getting the most accepted model which should be in consistent with geology.
6- There are a lot of software doing that.



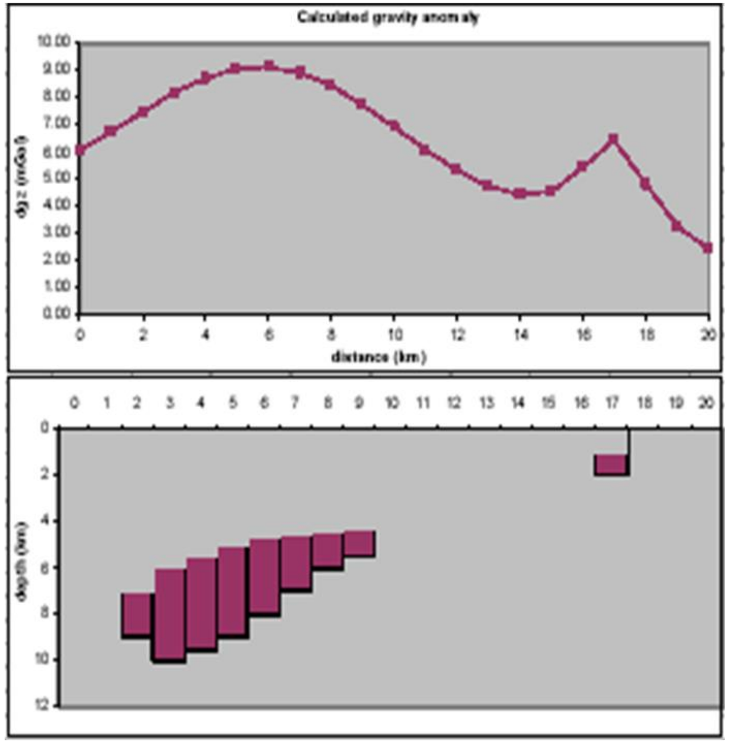
Gravity anomaly due to a two dimensional vertical Column (prism):

$$\Delta g_z = 2G\Delta\rho b \ln\left(r_2 / r_1 \right)$$

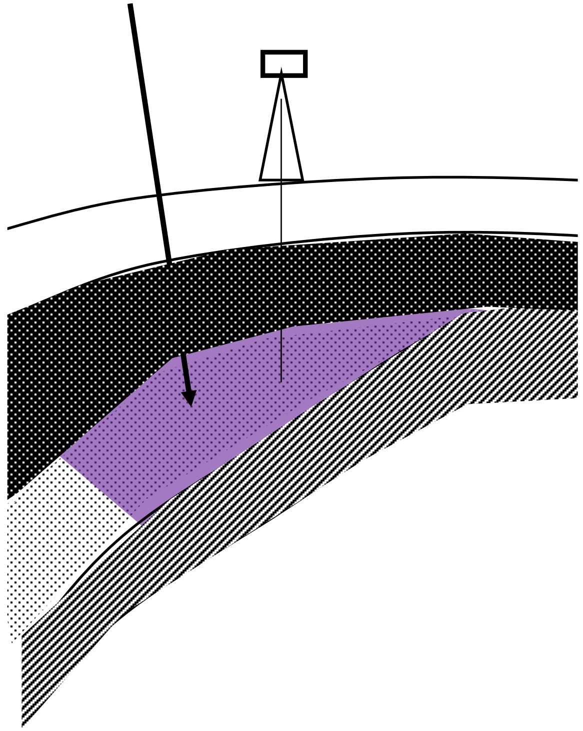


Gravity anomaly due to a series of vertical columns:

$$\Delta g_z = 2G\sum_i \Delta\rho_i b_i \ln\left(r_i^2 / r_i^1 \right)$$

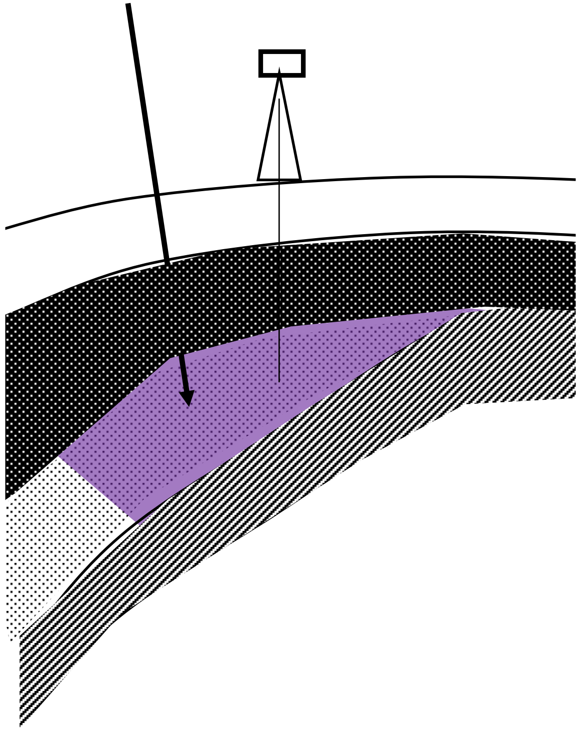


Ore deposits
(density more than surrounding rocks



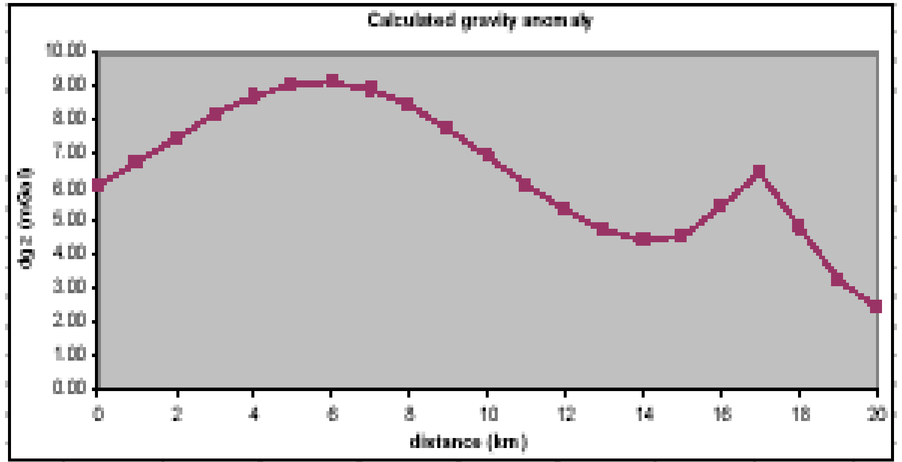
Geological picture

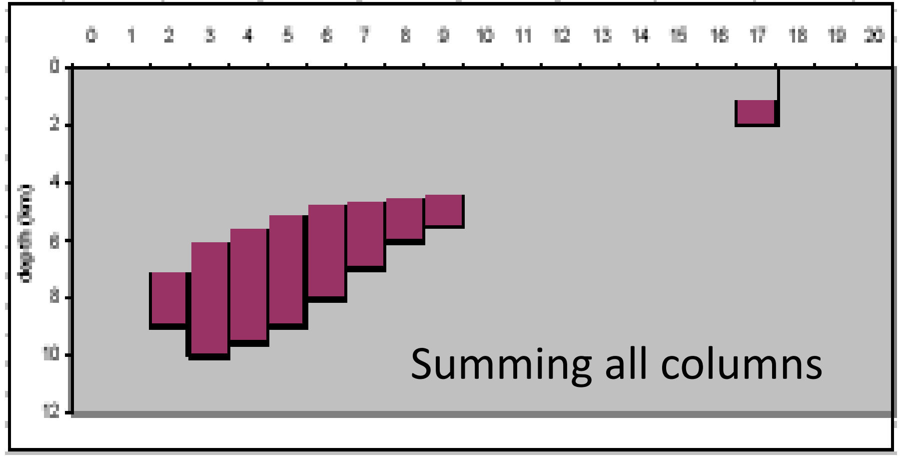
Ore deposits
(density more than surrounding rocks



Gravity anomaly due to a series of vertical columns:

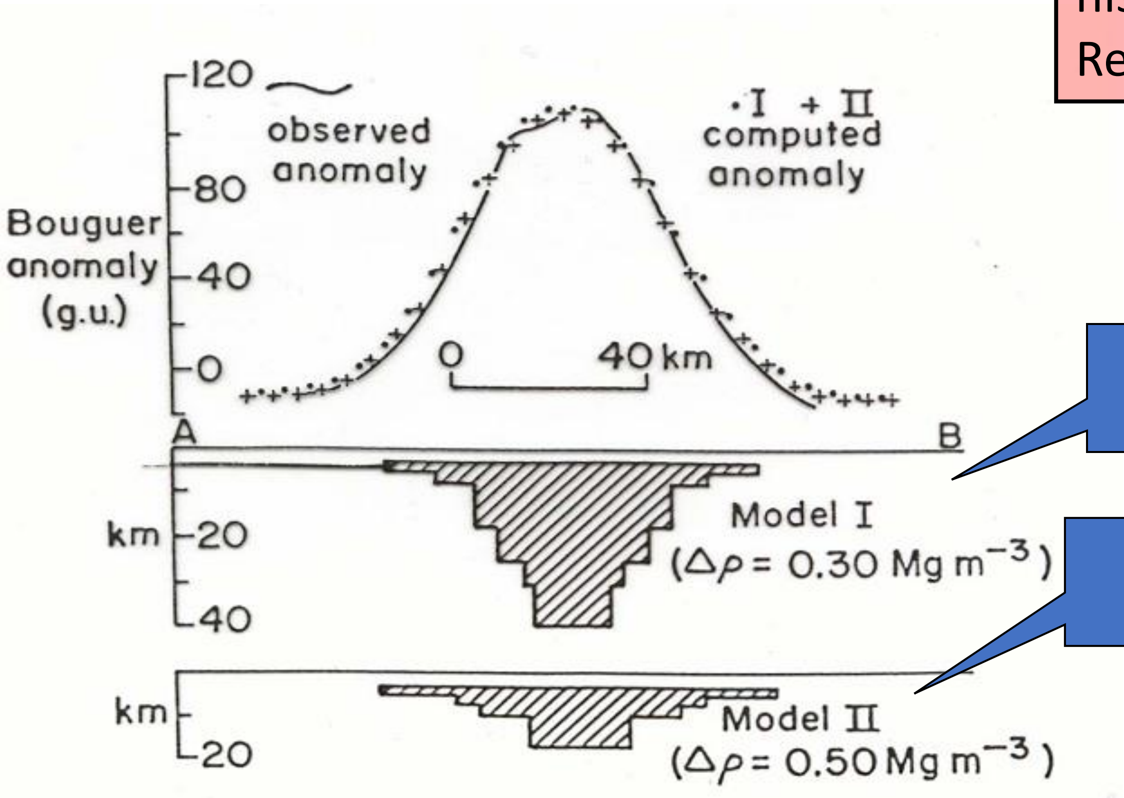
$$\Delta g_z = 2G\sum_i \Delta\rho_i b_i \ln\left(r_i^2 \big/ r_i^1 \right)$$





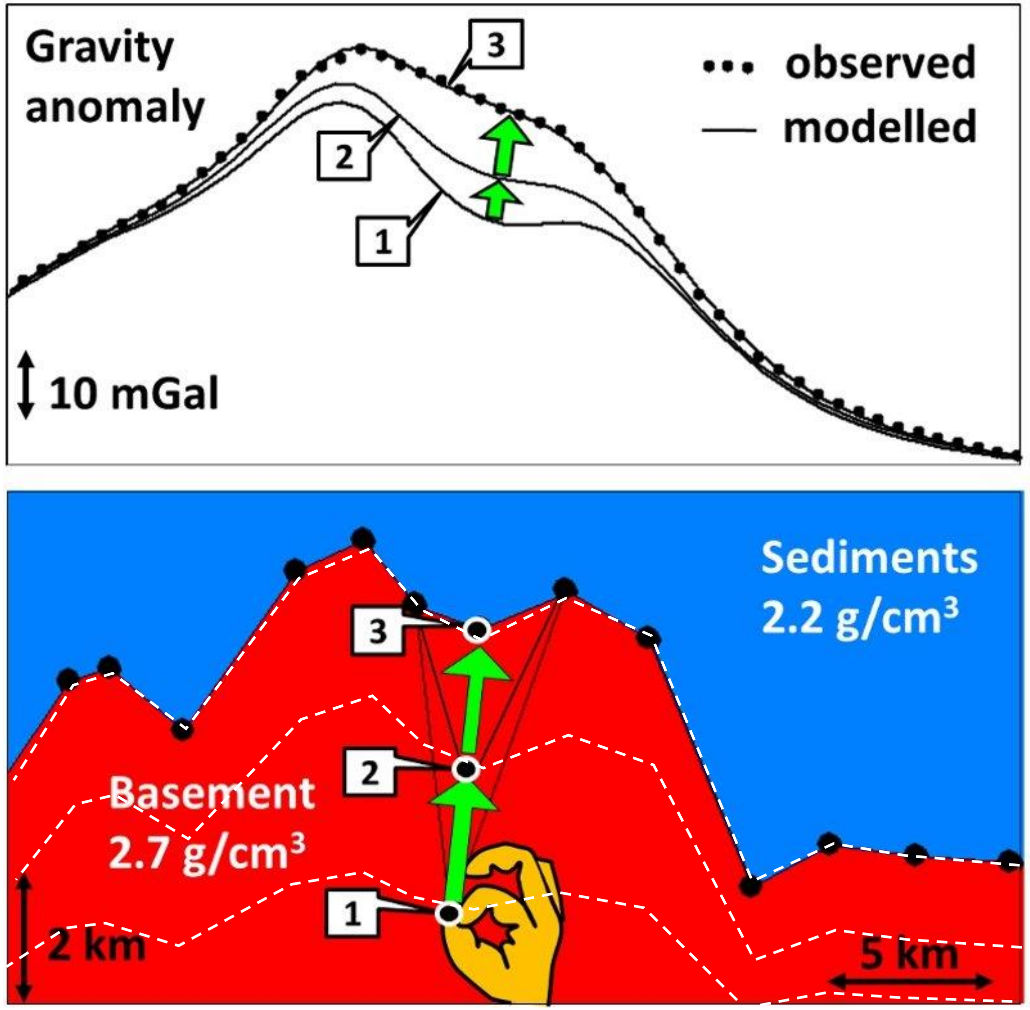
The solution may not be unique and more than one solution for a certain Case can be given.

Note: Read some case histories on page 92 in Reynolds

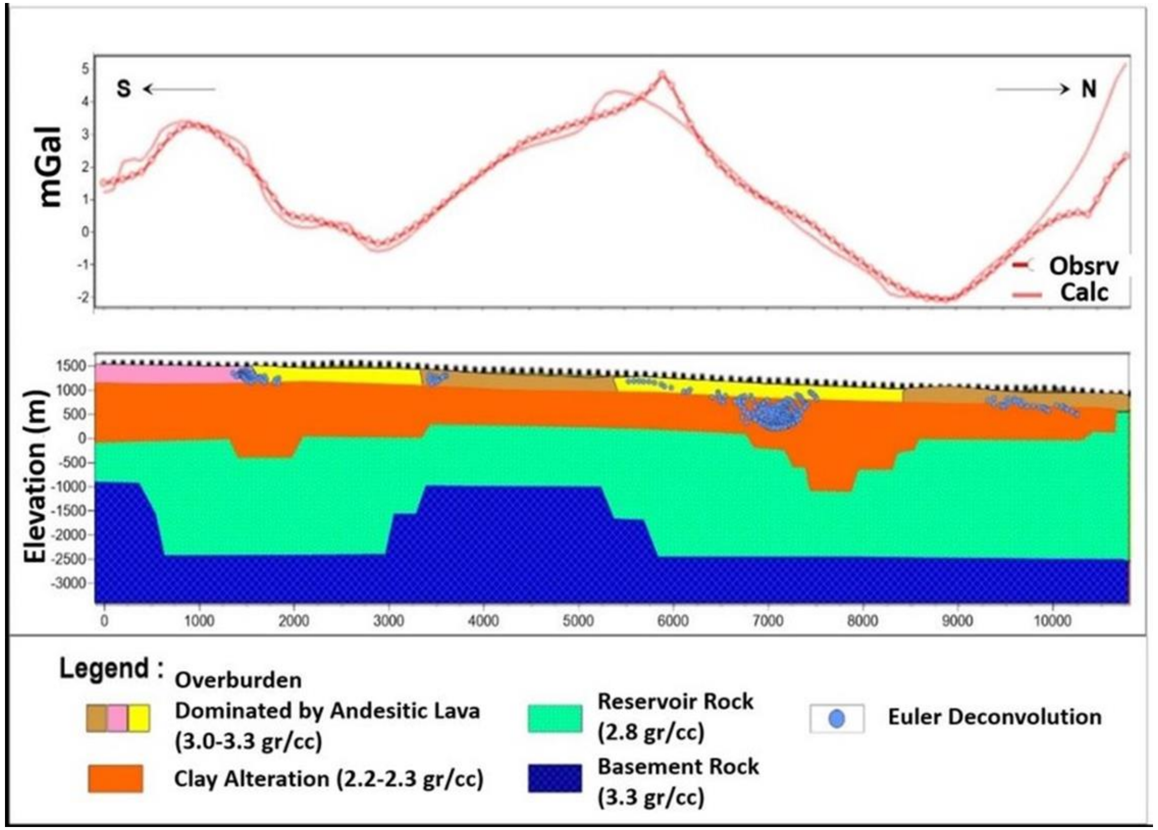


Solution 1

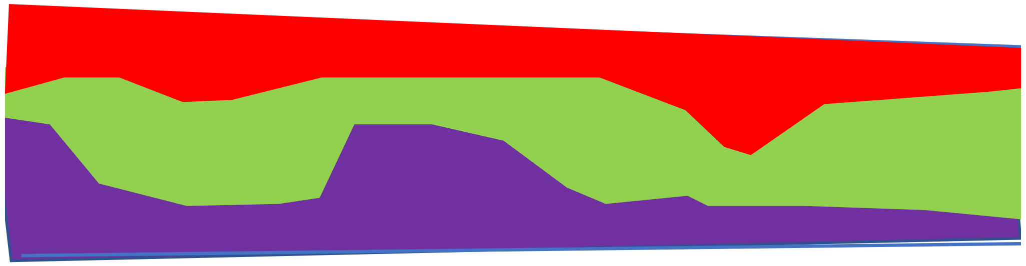
Solution 2



Gravity modelling in practice: The hand-hold pointer is at the boundary between top basement and sediments; move it upwards through positions 1 to 3 and the calculated gravity anomaly is changed. The aim is a match with the observed gravity.



Geophysical model



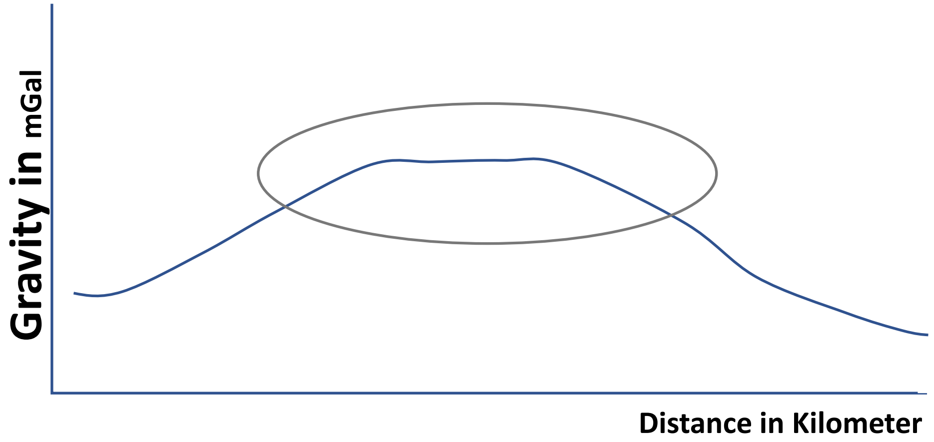
Geological model

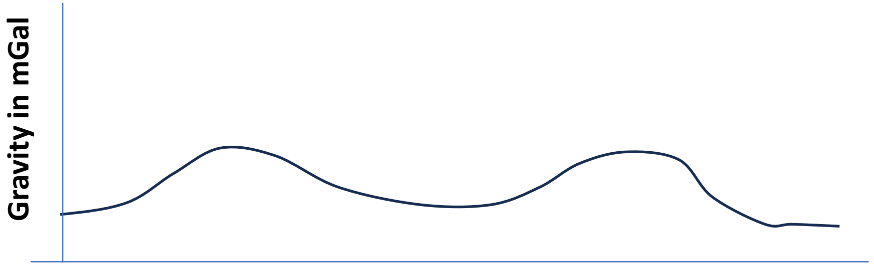
Gravity Geophysical Method in Oil Exploration (Examples)

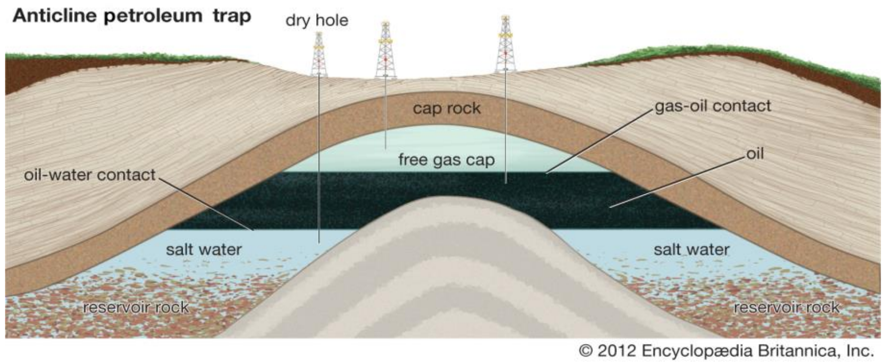
Gravity method is an essential part of oil exploration. It doesn't replace seismic, rather, it add to it. Despite being comparatively low-resolution, gravity method has some very big advantages. At a comparatively low-cost survey can provide coverage of large areas. Allowing quick regional coverage, even gravity surveys can now be recorded from an aircraft with fairly high reliability.

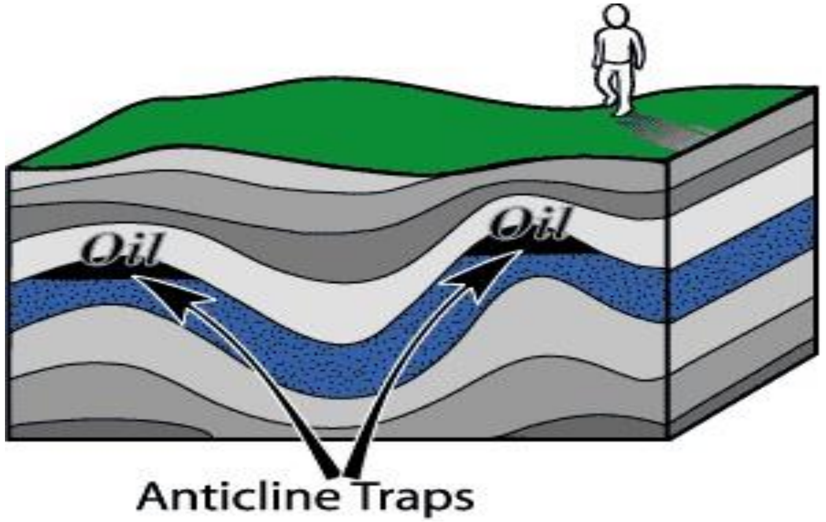
Gravity anomalies give indications for subsurface geological Structures which are favorable for oil tramping.

Anticlinal Trap



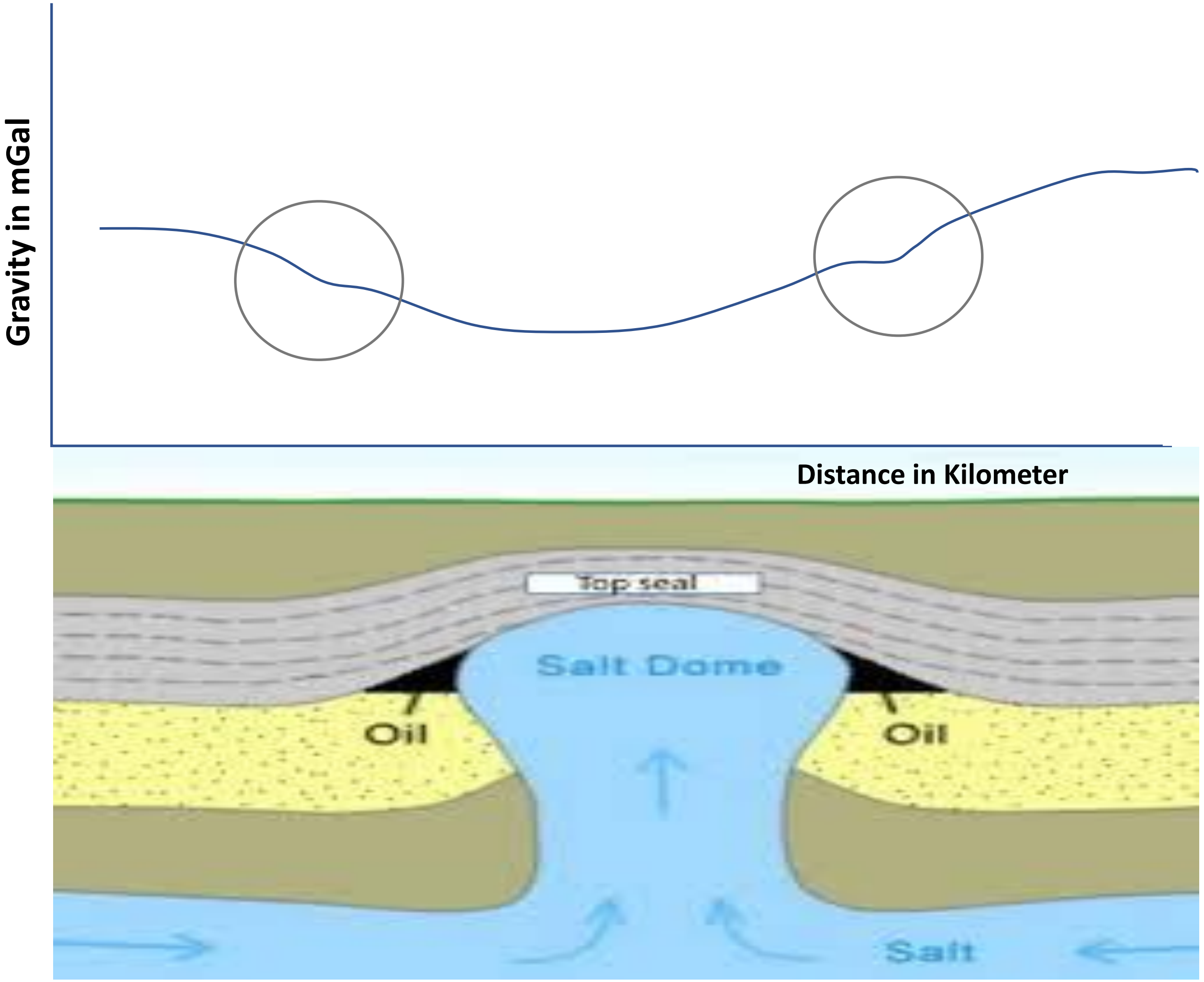






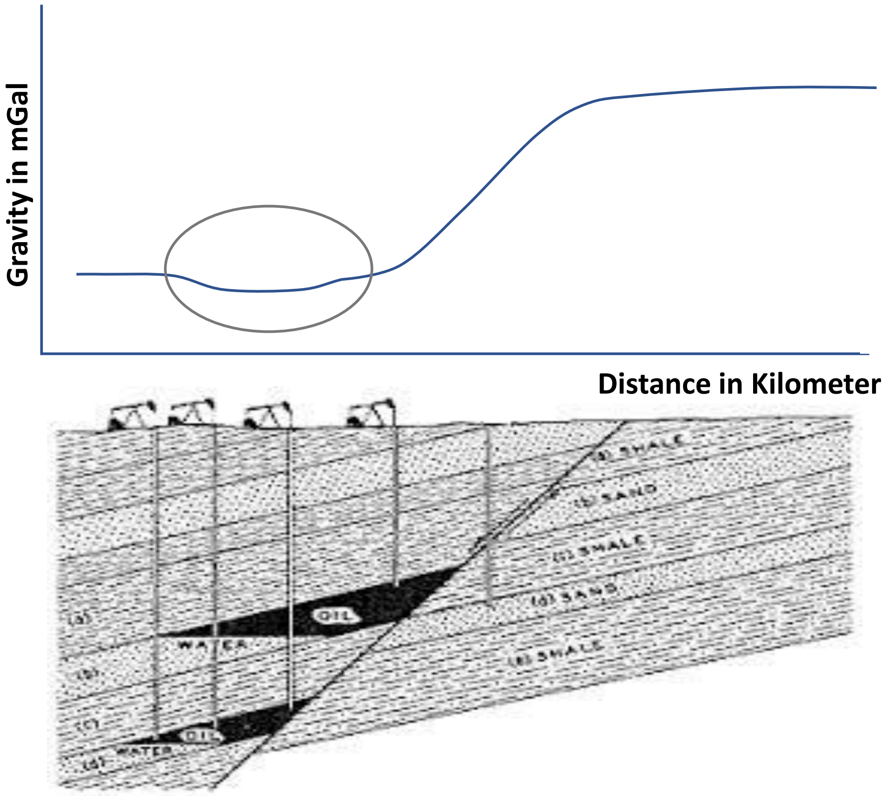
In all cases without additional information, a meaningful interpretation can not be given

Salt Dome Trap



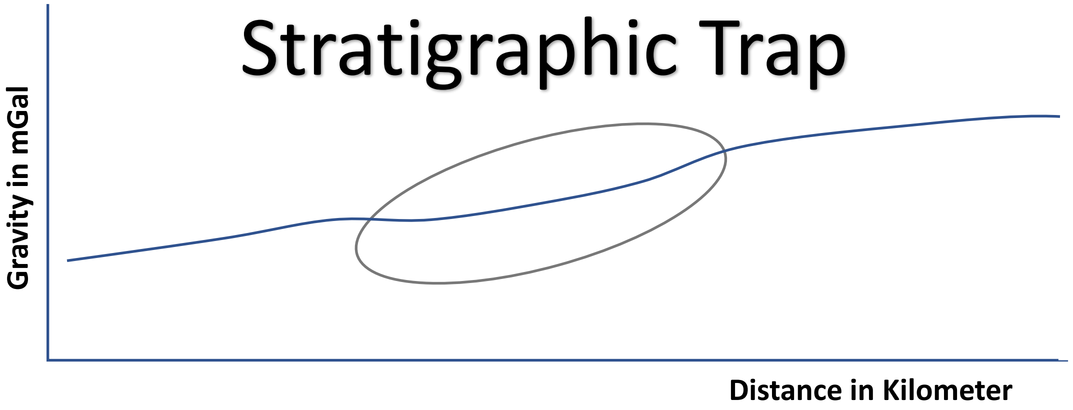
In all cases without additional information,
a meaningful interpretation can not be given

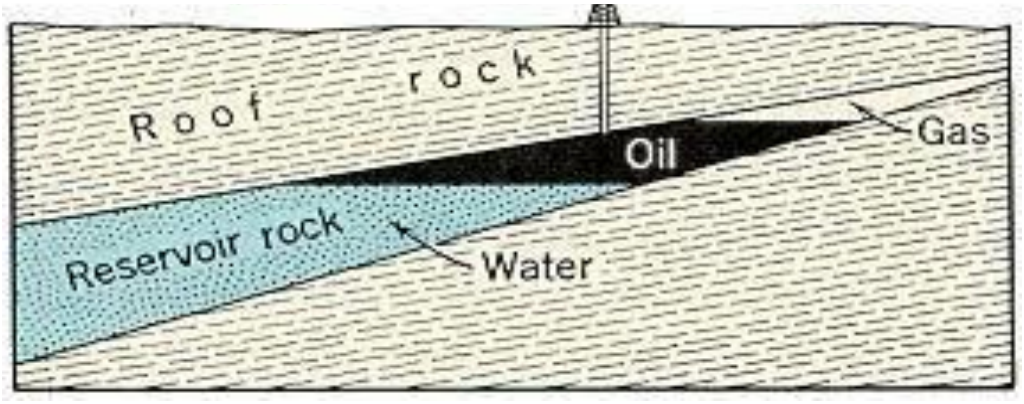
Fault Trap



In all cases without additional information,
a meaningful interpretation can not be given

Stratigraphic Trap





Iraqi Kurdistan Region
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Lecture 9

(Week 10)

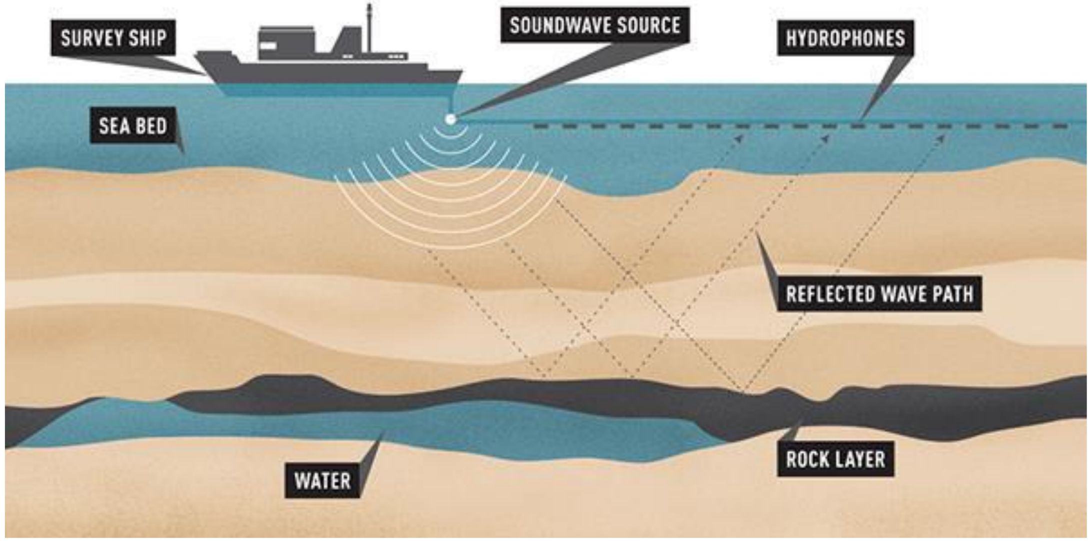
Applied Geophysics Lecture Notes
Fourth Semester
By:

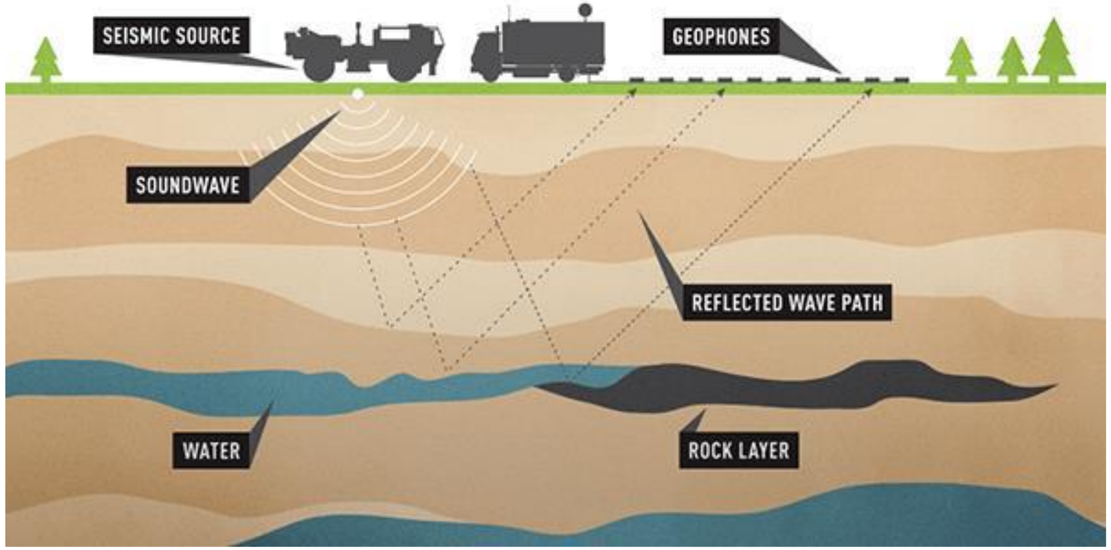
Dr. Fadhil Ali Ghaib

2023/2024

Contents:

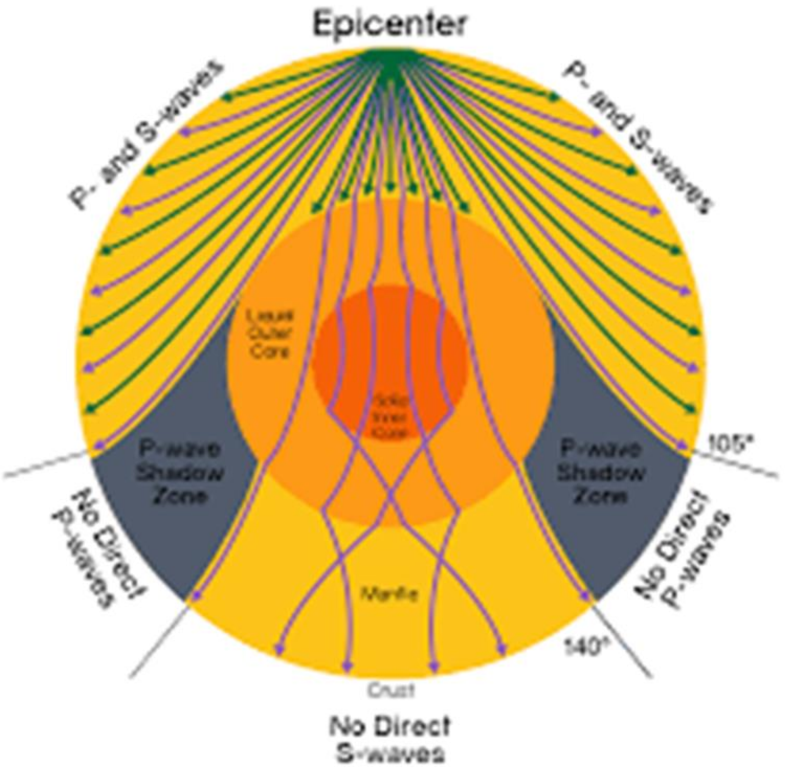
Seismic method





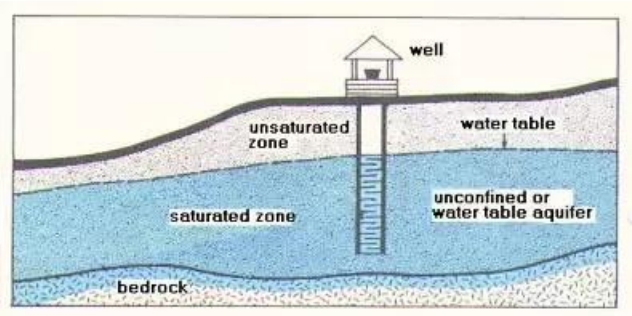
- **In seismic surveying, seismic waves are created by <u>controlled</u> sources and propagate through the subsurface.**
- **These waves will return to the surface after reflection or refraction at <u>geological boundaries</u>.**
- **Instruments distributed along the surface detect the ground motion caused by these returning waves and measure the <u>*arrival times*</u> of the waves at different ranges from the source.**

- Travel times from source to receiver will be ***converted to <u>depth</u>*** values. Subsurface <u>geological interfaces</u> can be systematically mapped.
- Seismic methods represent a natural development of the already long-established methods of <u>earthquake seismology</u>.
- In earthquake seismology travel times of earthquake waves are recorded at seismological observatories.
- They provide information on the gross internal layering of the earth.
- 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 <u>direct control of the geophysicists</u>.***

- <u>Applications:</u> **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 sediments 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**

- **Seismic methods are hardly used for ORE deposit exploration because these deposits are often of local extent and wave velocity has no or low contrast with surrounding host rocks.**

Seismic exploration

Seismic waves

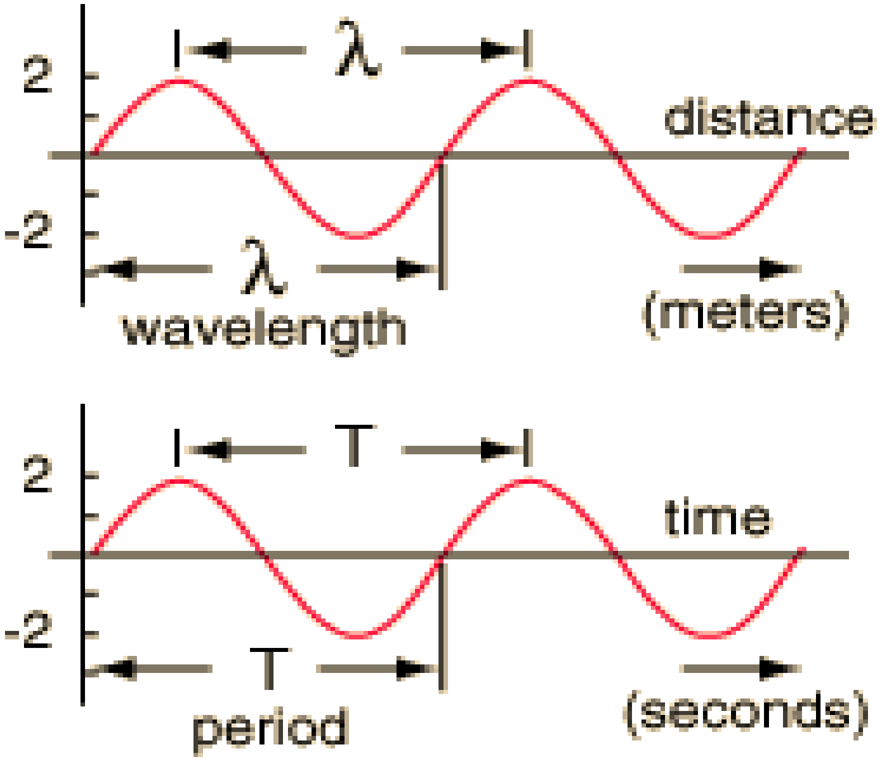
A wave is a periodic disturbance that transmits energy, without causing permanent deformation of the material.**

Seismic waves are parcels of elastic strain energy that propagate outward from a seismic source (e.g., earthquake, explosion).

Note that the particles of the material that the wave travels along vibrate but are not translated.

Wave parameters: velocity, frequency, period, wavelength, amplitude

Waves may be graphed as a function of time or distance. A single frequency wave will appear as a sine wave in either case. From the distance graph the wavelength may be determined. From the time graph, the period and frequency can be obtained. From both together, the wave speed can be determined.



$\lambda : wave\ length\,(m)$

$T : wave\ \ period\,(s)$

$v : velocity\ of\ \ wave\,(m/s)$

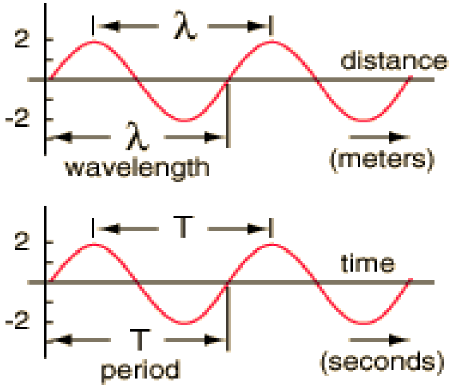
$f = 1/T\ \ (1/s)$

$v = \lambda * f\ \ (m/s)$

$v = \lambda / T\ (m/s)$

$v = \Delta x / \Delta t\ (m/s)$

Definitions:



Characteristics of waves:

Amplitude (A) – the maximum amount of displacement caused by the wave – i.e., from the rest position (0) to a peak or trough. The peak-to-trough displacement is 2A.

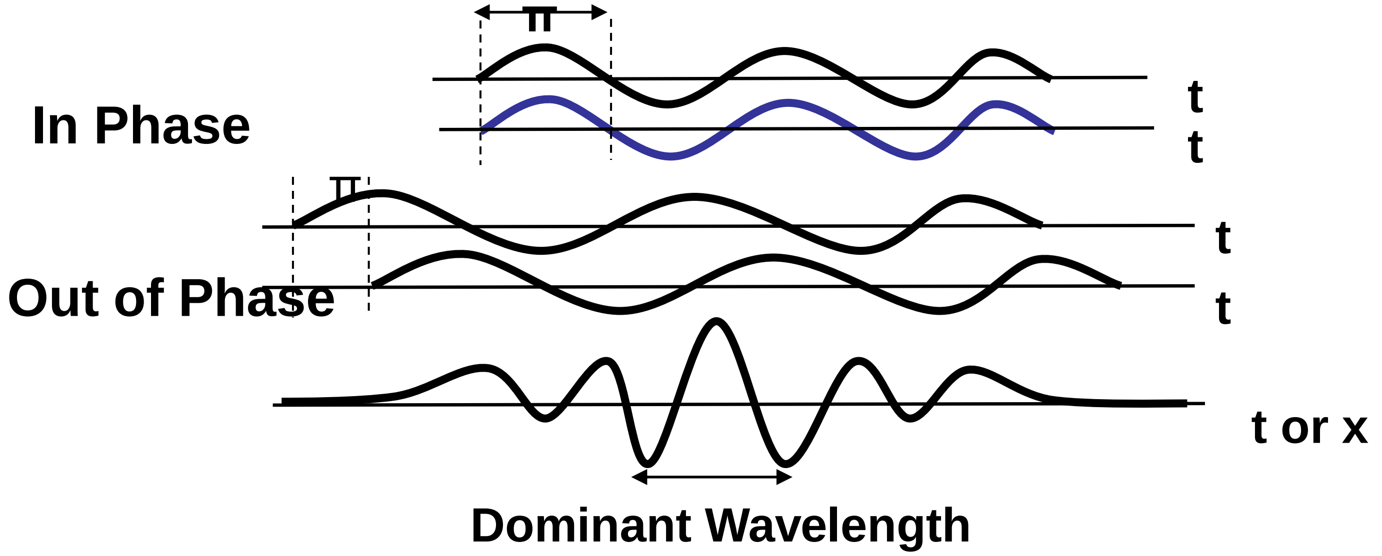
Wavelength (λ) – the distance (in m) between two points on a wave that have the same displacement (or phase). For example, the distance between two peaks or two troughs.

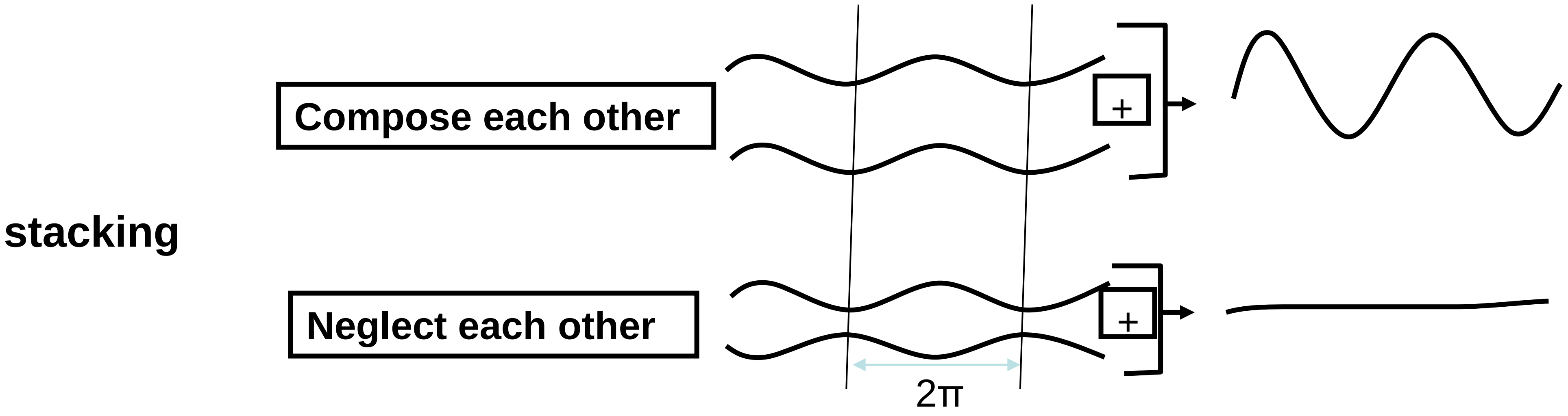
Period (T) – the time taken for a wave to go through one cycle (e.g., peak, trough, peak). The units are time (s).

Frequency (f) – the number of wave cycles that occur in one second. This is the inverse of the period ($f = 1/T$). The units are Hertz (Hz). 1 Hz = 1 s^{-1}.
– son

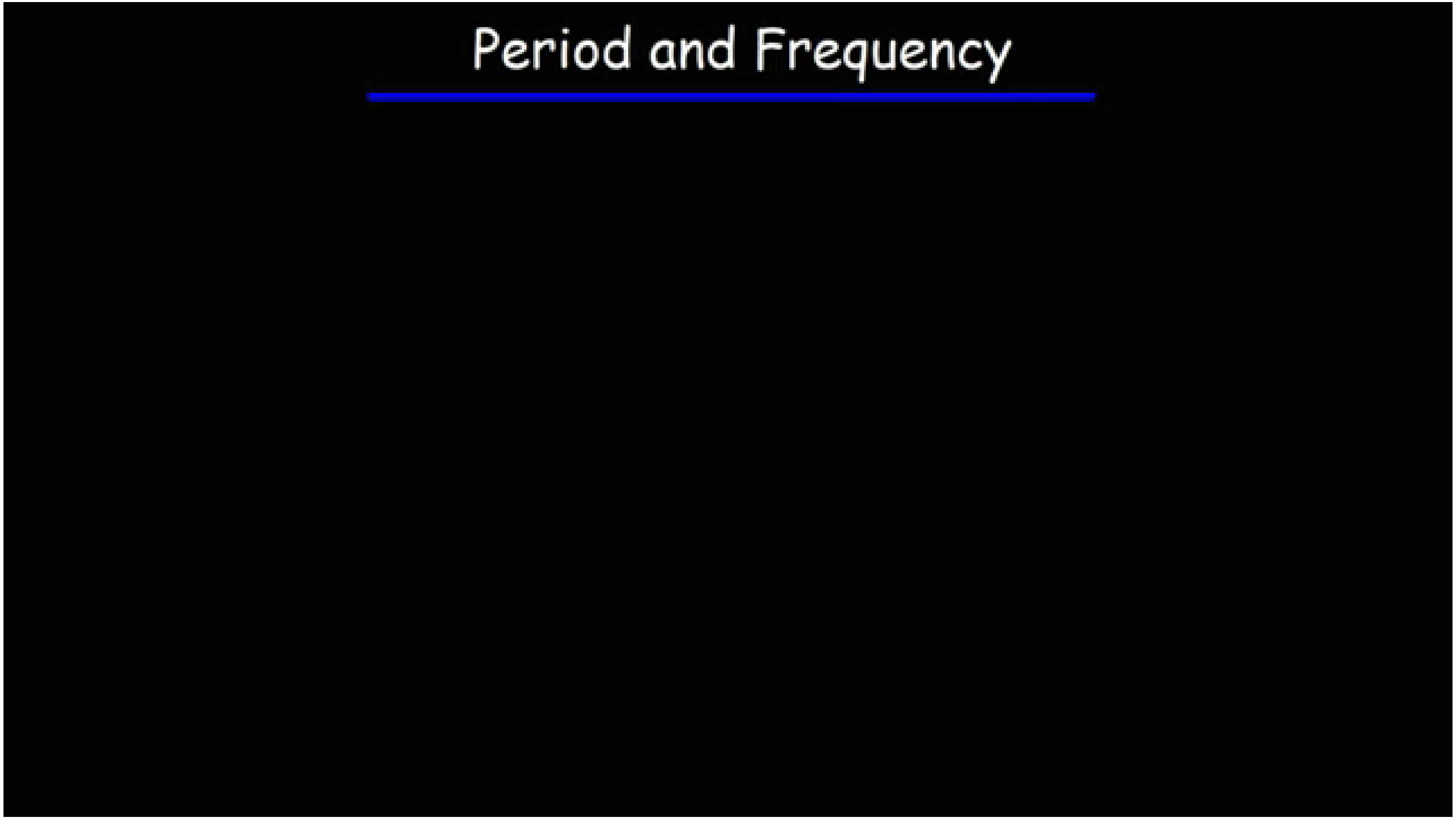
Velocity (v) – the rate at which the wave passes through the material: $v=d/t$ and $v=\lambda f$
note: - velocity is a material property (depends on composition, etc. –
- frequency depends on the source of the seismic waves

Expressions





Wave properties: https://youtu.be/ys0mQbJYQjY



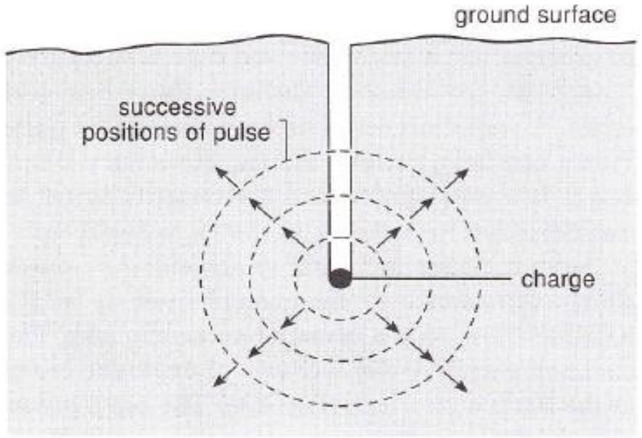
Wavefront and Ray wave

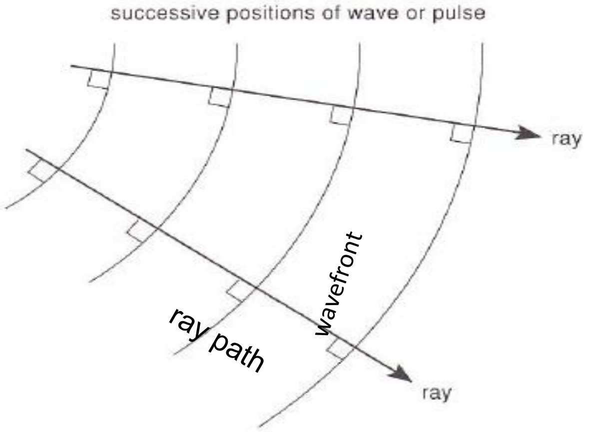
Seismic wave propagation can also be addressed through visualization of wavefronts and rays. Consider an explosion (point source). This will generate seismic waves that propagate outward from the source:

- **The wavefront is the locus of points where the pulse has reached at a given time. All points on the wavefront have the same phase (e.g., peak, trough)**

- **The ray is a vector which shows the direction that the wave is travelling**

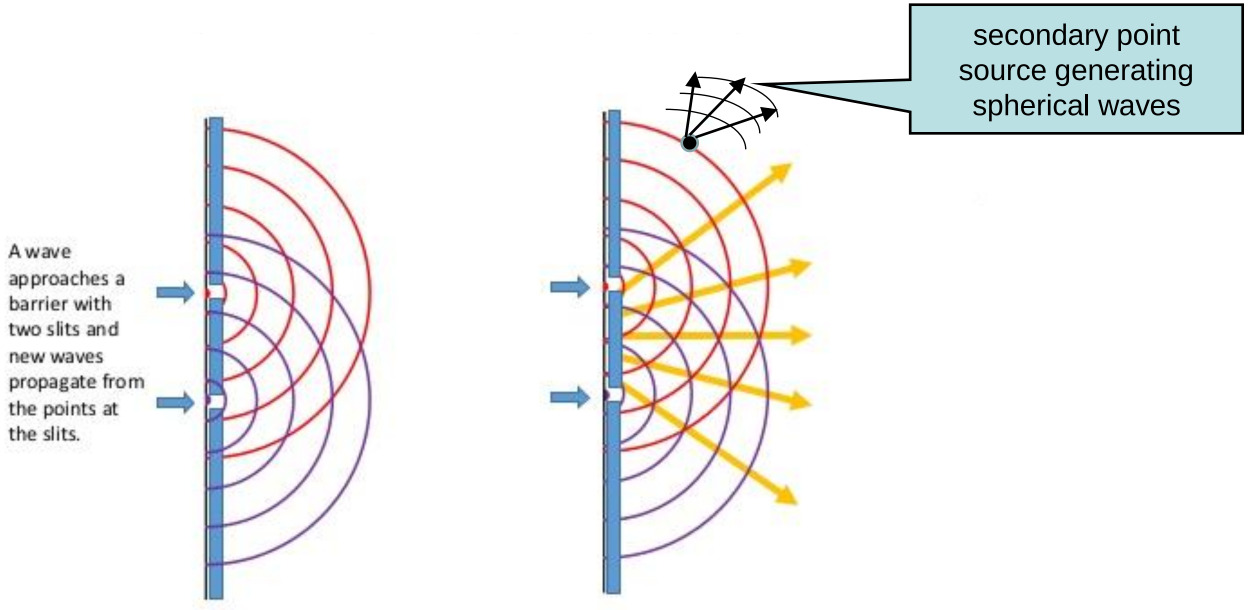
- **The wavefront and ray are orthogonal**



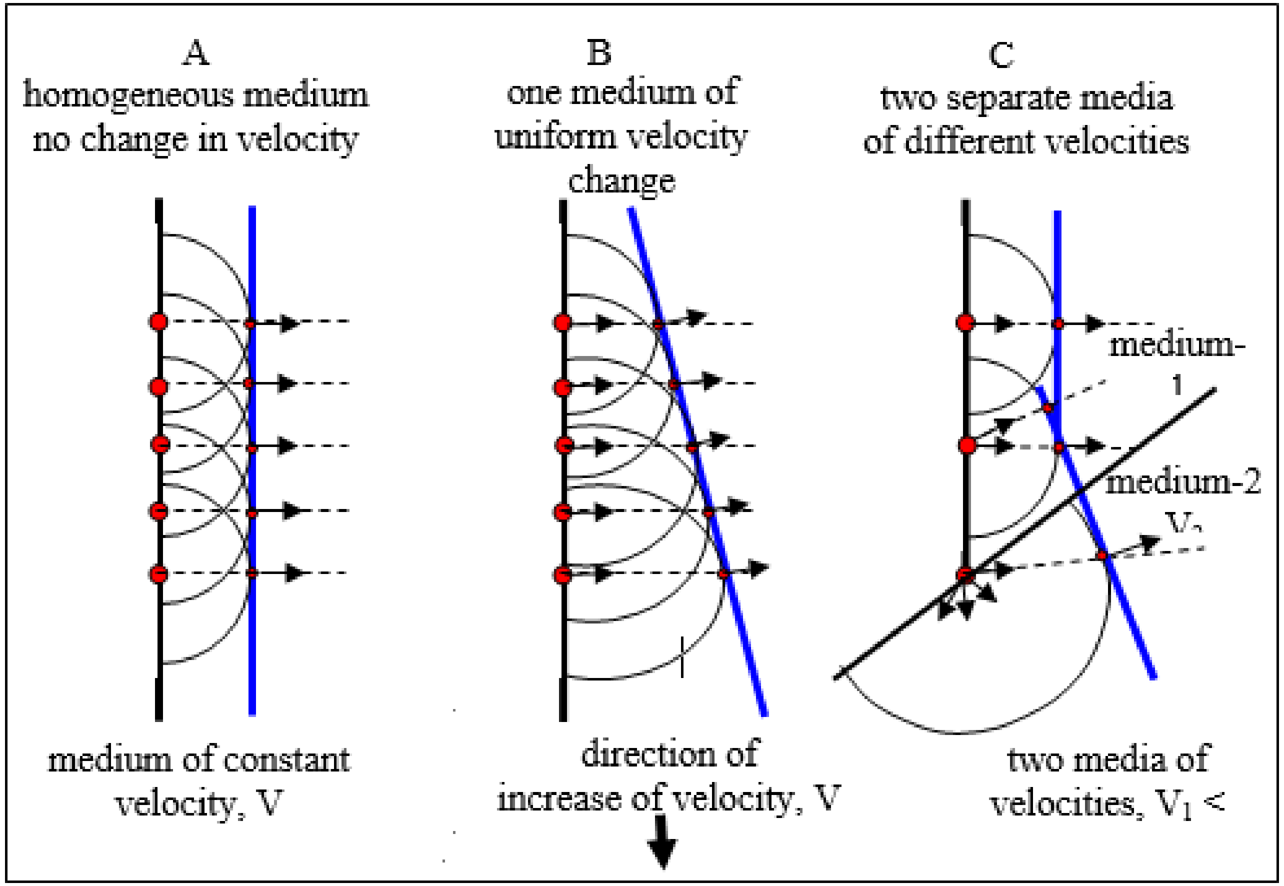


Over time, the wavefront will expand spherically. The velocity at which the wavefront travels away from the source is the seismic velocity of the material.

Propagation of the wave front can also be described using <u>Huyghens' Principle</u>; every point on a wavefront acts as a secondary point source generating spherical waves. The secondary waves propagate outwards the envelope of secondary waves gives the overall wavefront.



Plane-wave propagation according to Huygens' Principle. (A) through a homogeneous medium where velocity is constant. (B) Inhomogeneous medium of velocity which is uniformly changing across the propagation direction. (C) Two media of different velocities.



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

<u>Stress and Strain</u>

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

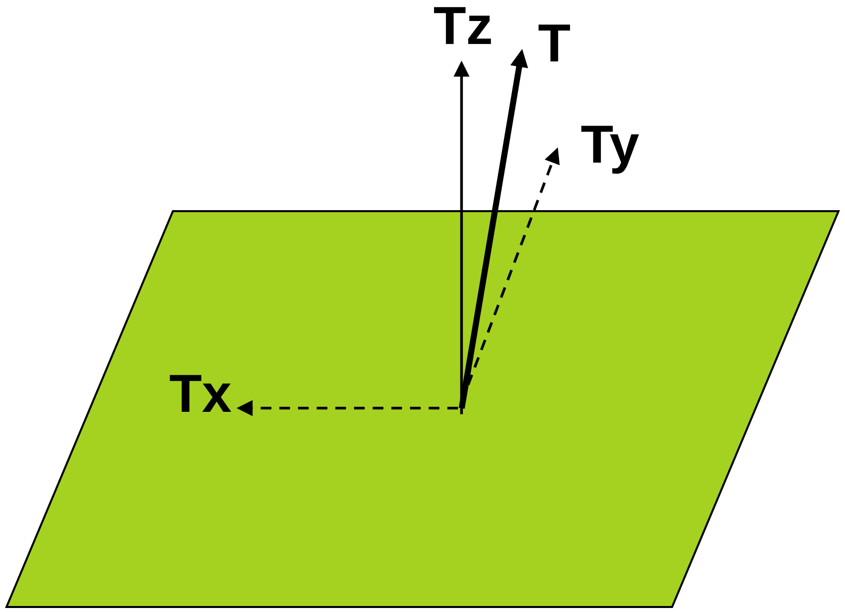
Stress can be resolved into two components:

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.

T = F/A



Tx and Ty are in one plane, they are the stress shearing components

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.

Elasticity

When a wave transmits through a medium, changes take place, depending upon:

- Energy of the wave

- Physical properties of the medium

These changes are:

- Redistribution of the internal forces

- Modifications of the geometrical shape

The analyses of these two types of change and any related features are called the *Theory of Elasticity*.

Elastic deformation and Hooke's Law

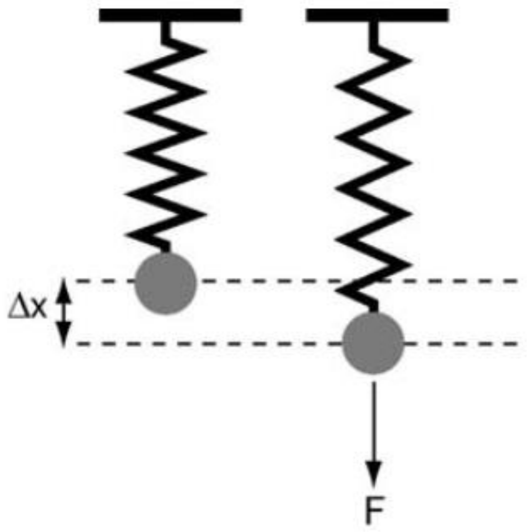
Hooke's law states that any strain is proportional to the stress that produced it (linear elasticity)

For a spring: $F = k\Delta x$

where F is the force used to stretch the spring (stress)

Δx is the amount of stretch (strain)

k is the spring constant (material property that measures the stiffness of the spring)



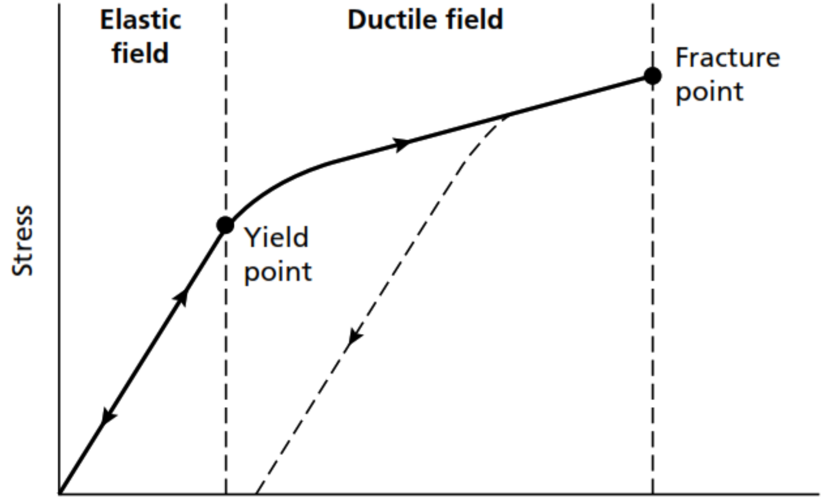
Rearranging the equation (F = kΔx):
We get k = F / Δx = stress / strain

k governs how much stress is needed to produce a given strain.

stress and strain are linearly related

– for an elastic material: the material returns to its original shape when the stress is removed
– above the elastic limit: deformation is permanent (plastic or ductile deformation)

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.



A typical stress-train curve for a solid body.

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

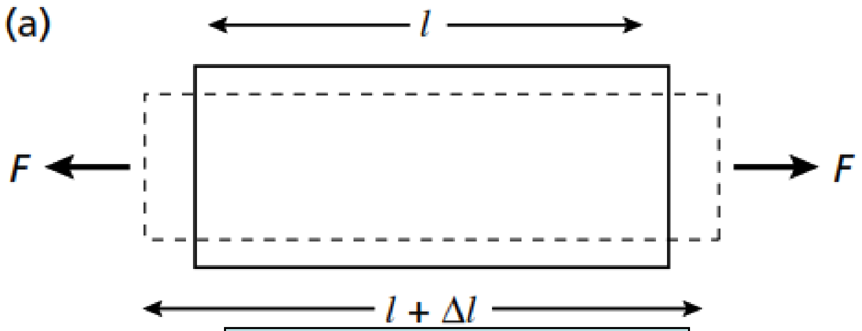
When a rock is subject to stress, a set of elastic parameters (moduli) determine how much strain will occur. The moduli are measures of the strength of the material.

The greater the modulus, the stiffer the material, or the smaller the elastic strain that results from the application of a given stress.

a) Young modulus E

Young's modulus (E) is defined as the ratio of the stress applied to the material along the longitudinal axis of the specimen tested and the deformation or strain, measured on that same axis.

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

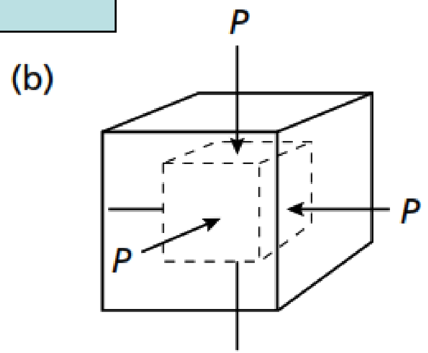


Change in Length Per Unit Length

b- Bulk modulus K

a coefficient of elasticity of a substance, expressing the ratio between a pressure that acts to change the volume of the substance and the fractional change in volume produced

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

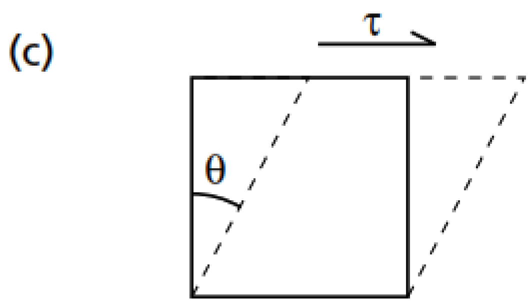


Change in volume Per voleum

c) Shear modulus

The shear modulus is defined as the ratio of shear stress to shear strain. It is also known as the <u>modulus of rigidity</u>

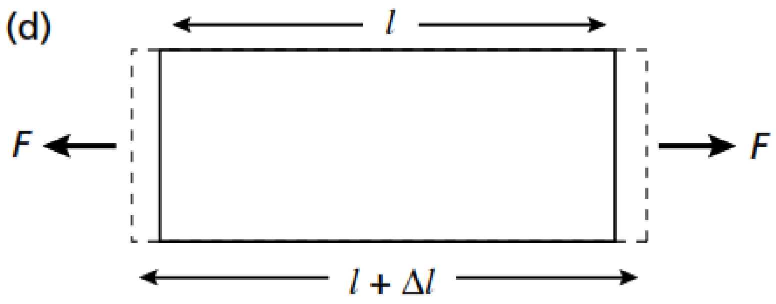
$$\mu = \frac{\text{shear stress}}{\text{shear strain}} = \frac{\tau}{\tan\theta}$$



d) Axial modulus ψ

Axial modulus (φ) is the ratio of longitudinal stress to uniaxial longitudinal strain, i.e. where there is no lateral strain.

$$\psi = \frac{\textit{longitudinal stress } F/A}{\textit{longitudinal strain (uniaxial) } \Delta l/l}$$



Iraqi Kurdistan Region
Tishk International University (TIU)
Faculty of Engineering
Department of Petroleum and Mining Engineering
Erbil

Lecture 10

(Week 11)

Applied Geophysics Lecture Notes
Fourth Semester
By:

Dr. Fadhil Ali Ghaib

2023/2024

Contents:

Types of Seismic Waves

- **Seismic waves are parcels of elastic 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: <u>body waves and surface waves</u>.**

<u>Body waves</u>; They travel through the whole body of the Earth

1- P waves

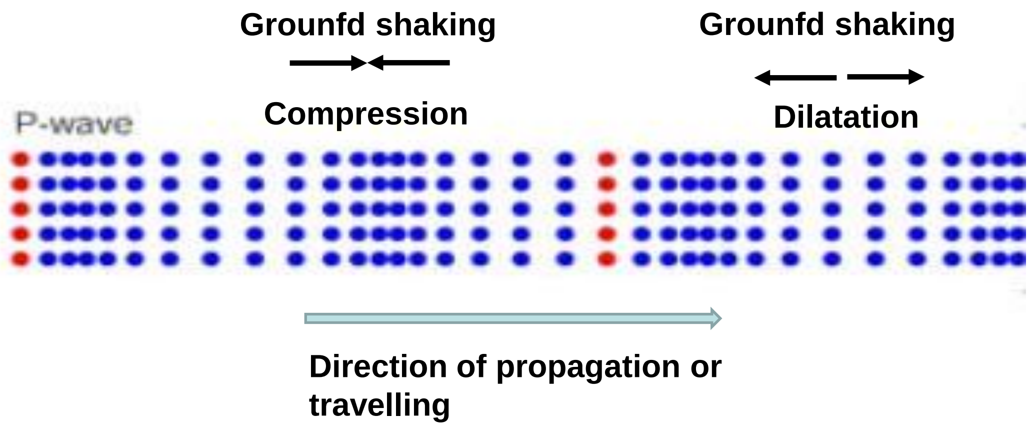
(longitudinal, primary or compressional wave), Material particles oscillate about a fixed point in the direction of wave propagation by *compressional and dilatational strain.*

Type of Strain: ***Volume deformation***

They are *fastest, deeper penetration and least attenuation*.

$$Vp = (K/\rho)^{1/2}$$

$$= [(K + 4/3\mu/\rho)]^{1/2}$$



2- S waves

(transverse, secondary or shear wave)

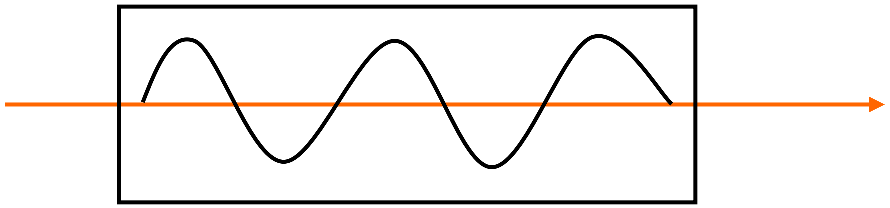
Particle motion is at right angles to the direction of wave propagation and occurs by pure strain.

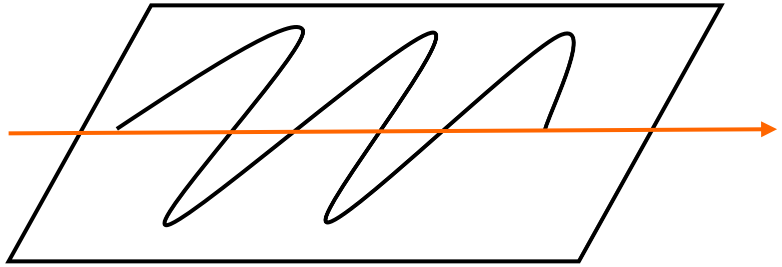
Type of strain: *Shape Deformation*

Two types of S waves:

1- SV

2-SH

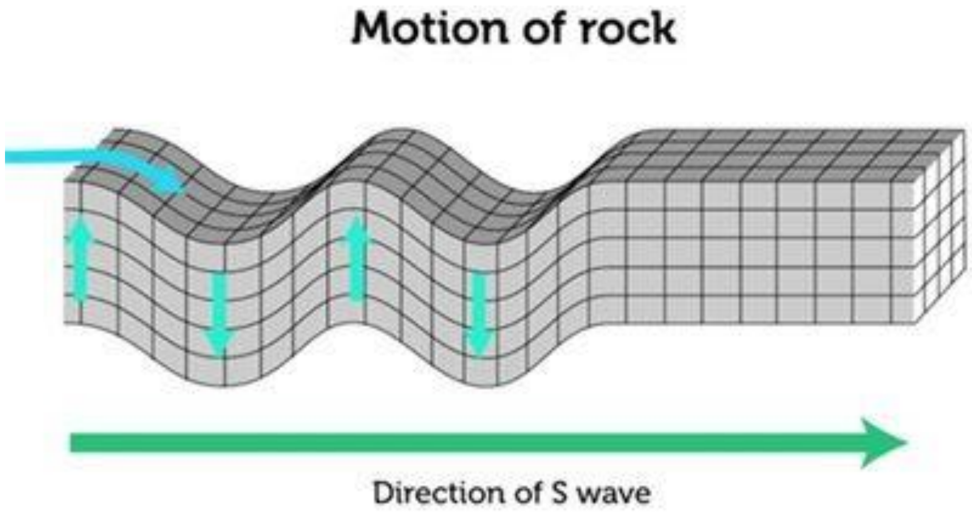




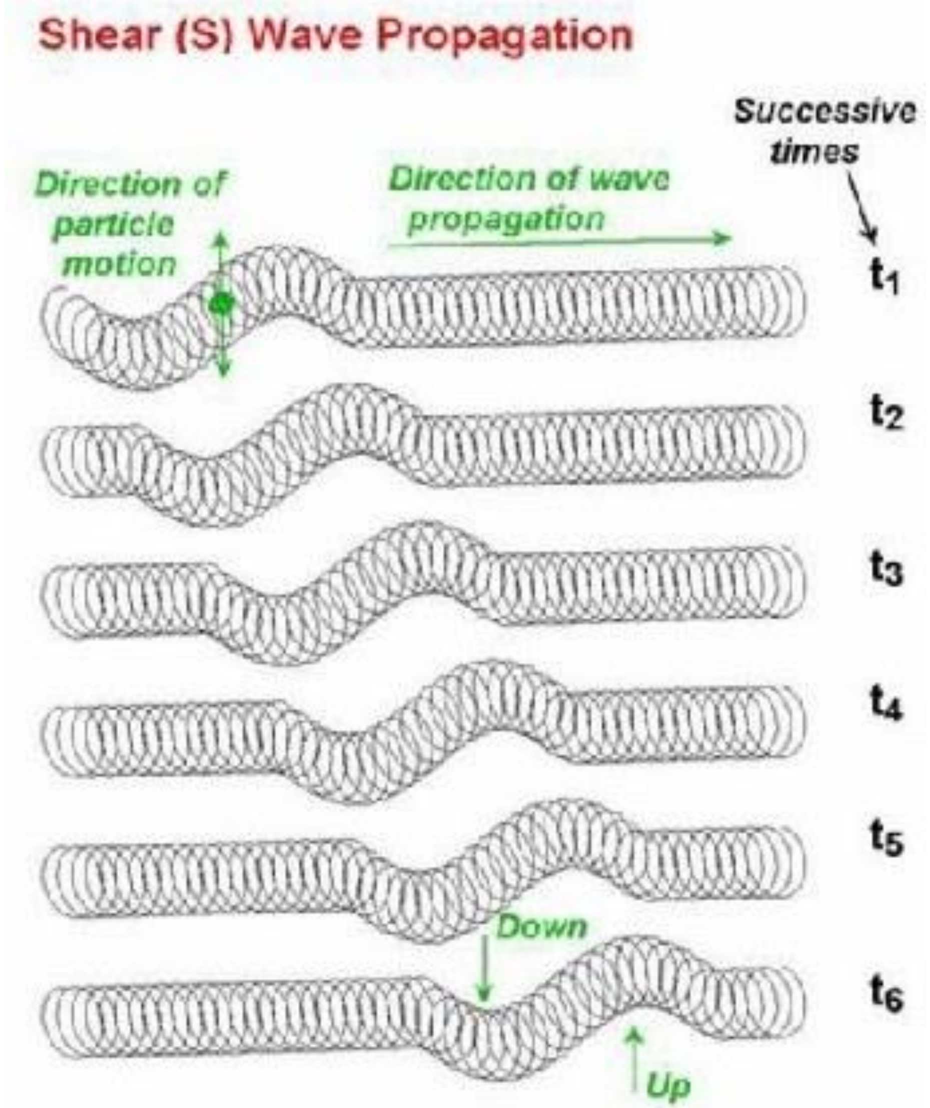
$Vs = (\mu/\rho)^{1/2}$

$Vp/Vs = [2(1-\sigma)/(1-2\sigma)]\ 1/2$

$Vp \sim 1.7\ Vs$



Motion of rock
Direction of S wave



Shear (S) Wave Propagation
Successive times
Direction of particle motion
Direction of wave propagation
t1
t2
t3
t4
t5
t6
Down
Up

Difference between S waves and P waves

P waves	S waves
The first wave to hit seismographs	Second waves to hit seismographs
They are compression waves	They are shear waves
Can move through solids and liquids	Can only move through solids
Shake the medium in the direction in which they are propagating	Shake the medium in the direction perpendicular to which they are moving

P and S GIF

Surface waves; ***They do not penetrate into the deeper subsurface media. They decay rapidly with depth and travel at the free surface of a semi-infinite media.***

$V_R = 0.92\ Vs$

There are two types:

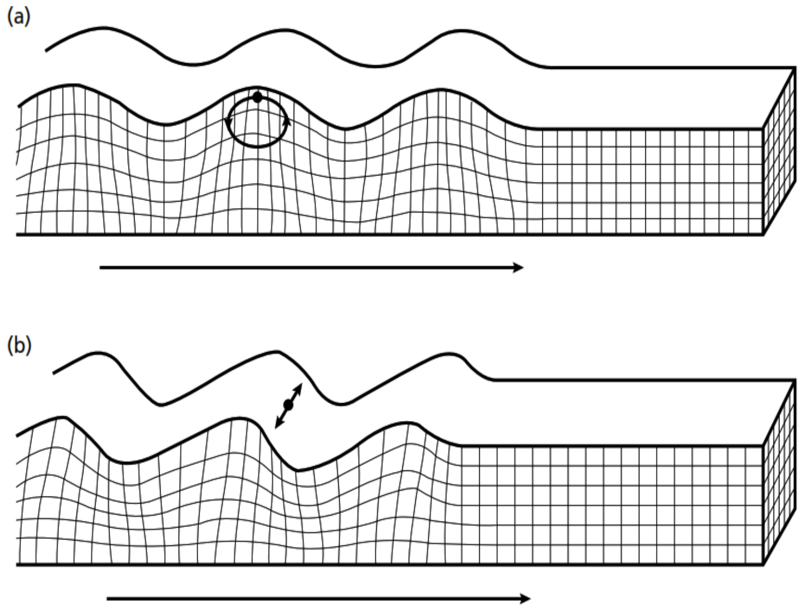
1- Rayleigh waves (also called Ground Roll)

Travel along the free surface of the earth with amplitudes that decrease exponentially with depth; Particle motion is in an elliptical sense in a vertical plane with respect to the surface.

Rayleigh waves travel only through solid media.

2- Love waves

particle motion is at right angles to the wave propagation but parallel to the surface. They look like SH waves and not detected by geophone.



(a) a Raleigh wave

(b) a Love wave.

Brief equations on seismic wave velocities

$$Vp = \{(K+4/3\mu)/\rho\}1/2$$

$$Vs=(\mu/\rho)1/2$$

K is Bulk Modulus, μ is Shear Modulus while ρ is density.

Elastic wave velocity as a function of geological age and depth.

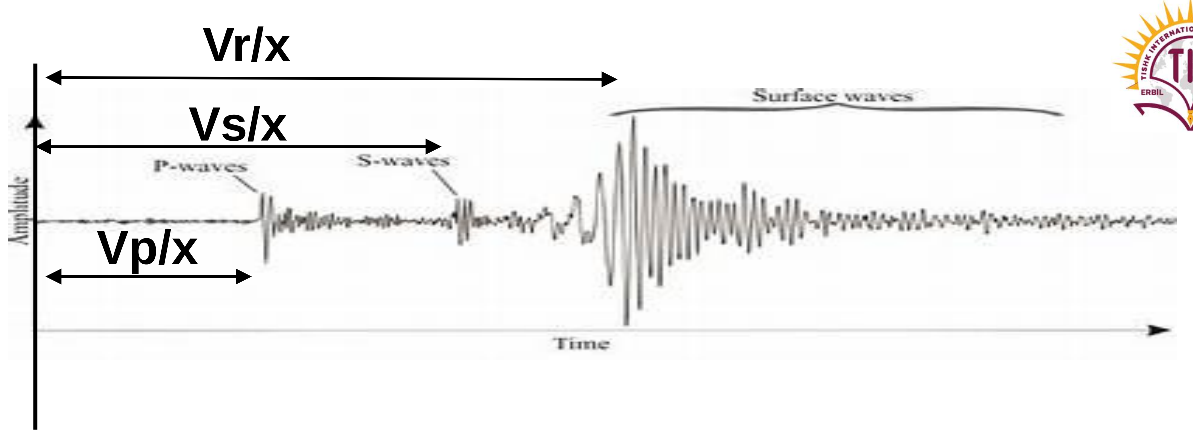
$$V = 1.47(zT)1/6 km/s$$

z: depth in km , T: geological age in millions of years

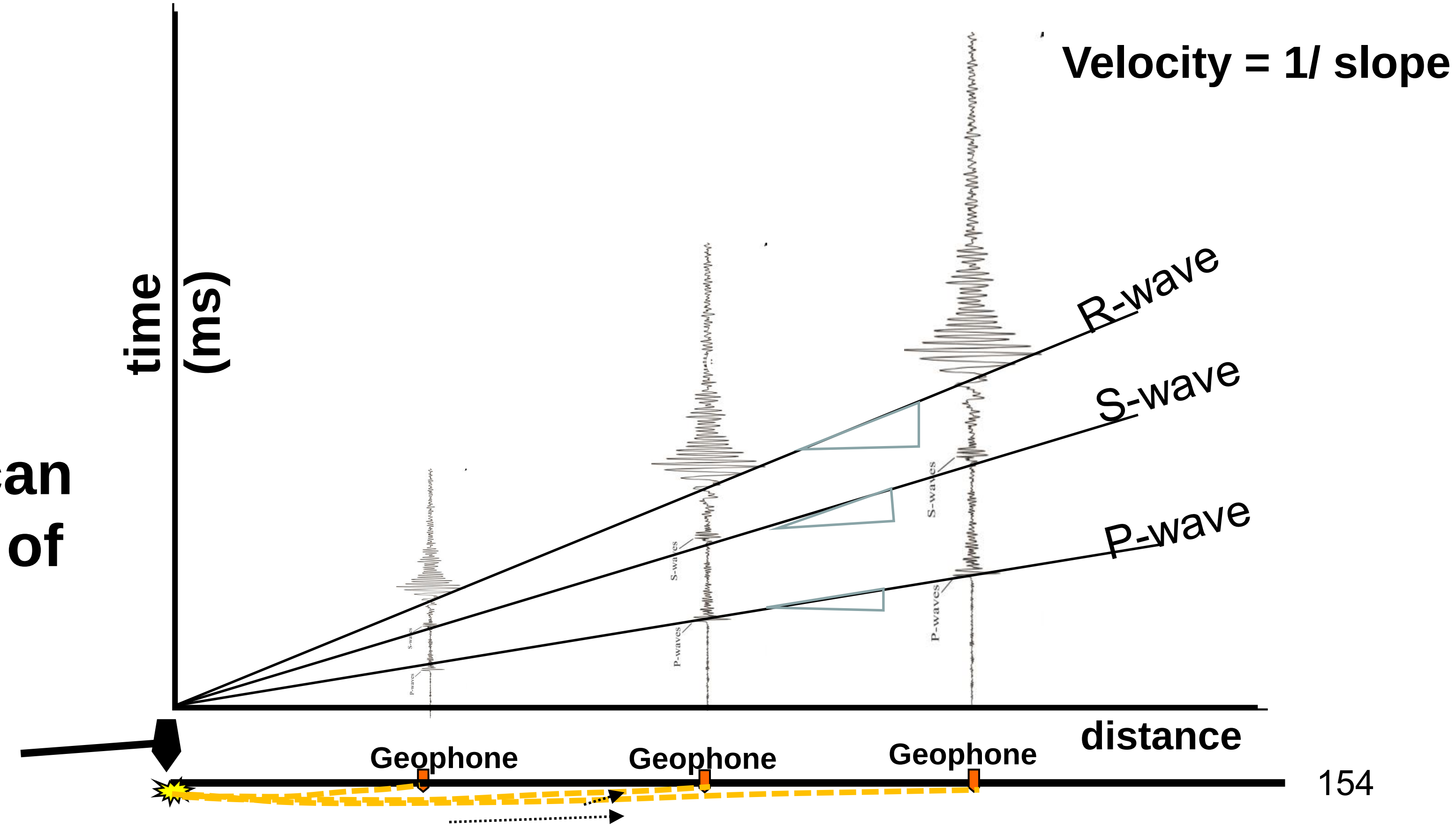
Surface Waves

Seismic wave types: https://youtu.be/H9Ifk0Bde4Q

The three types of waves (P, S and R) are presented on a seismogram respectively according to their velocity.



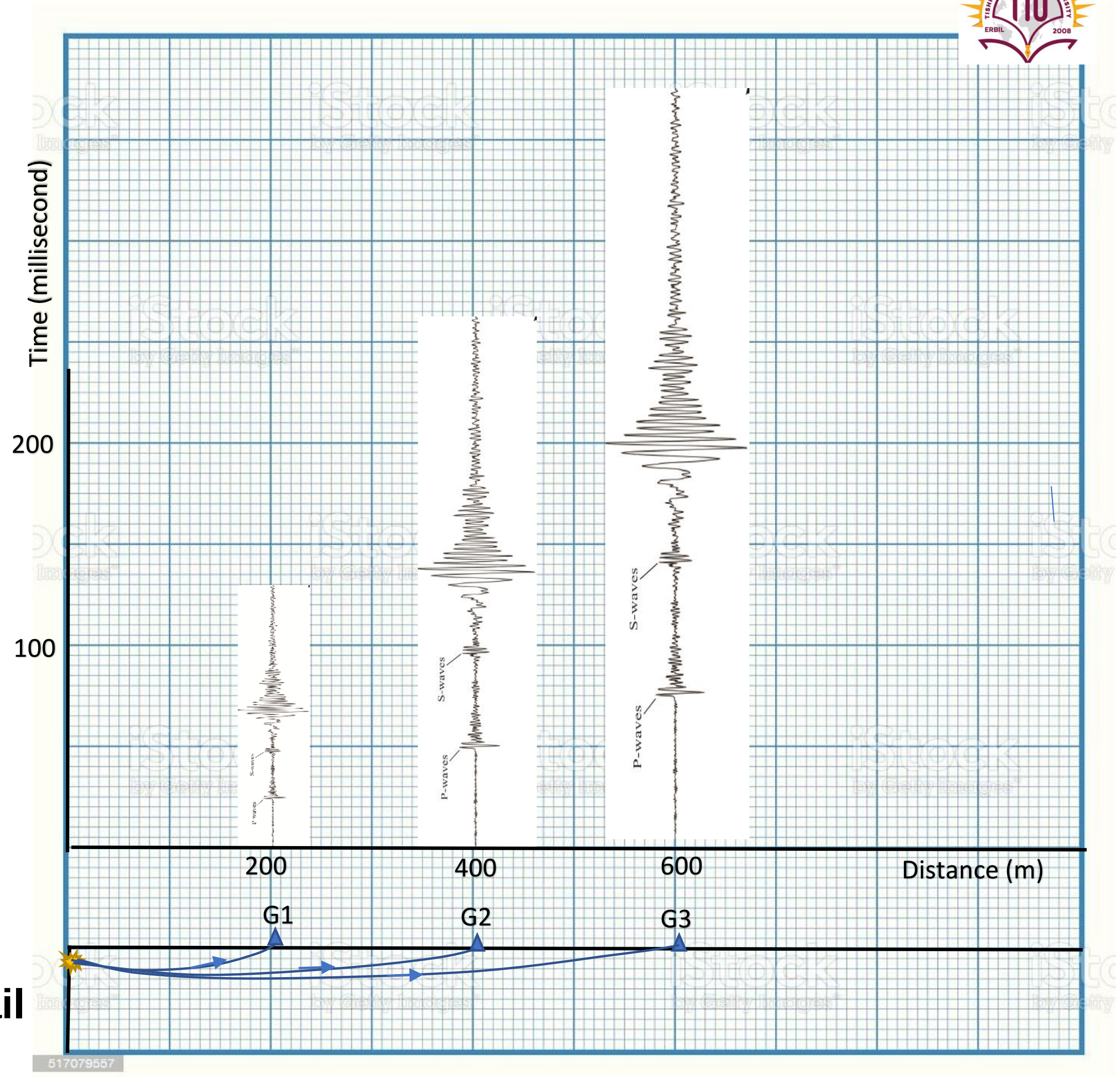
When more than one seismogram recording the same signal along a line, velocity in the upper layers can be calculated from the slope of the lines.



Name of the student:

Date: 28/4/2024

Homework: Three geophones are arrayed along a traverse, the inter-distance from the shot point is 200m as shown in the figure.
Calculate the velocity of P, S and Raleigh waves.



Calculations:

Scan and send your answer to e-mail
famehdilo@gmail.come
Dead line: 7/5/2024

2023-2024

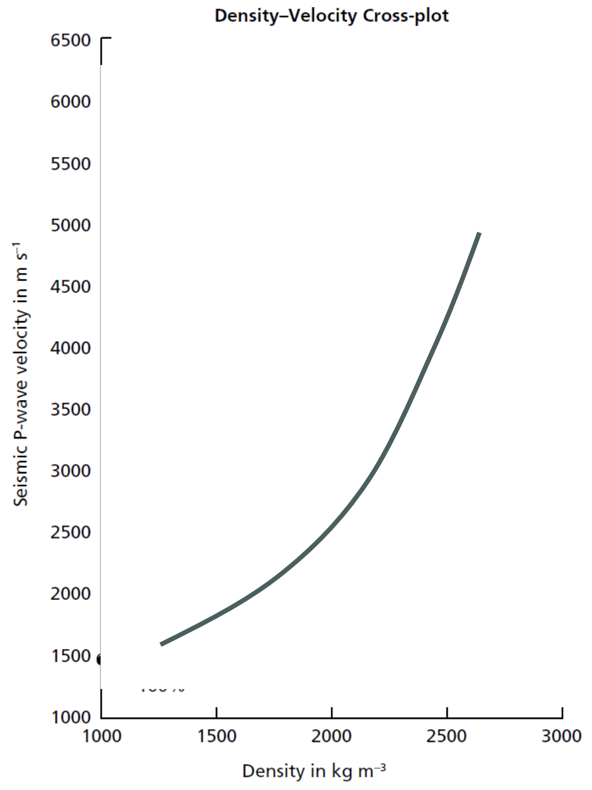
Velocity in rocks is affected by:

1-Texture (grain shape and sorting)

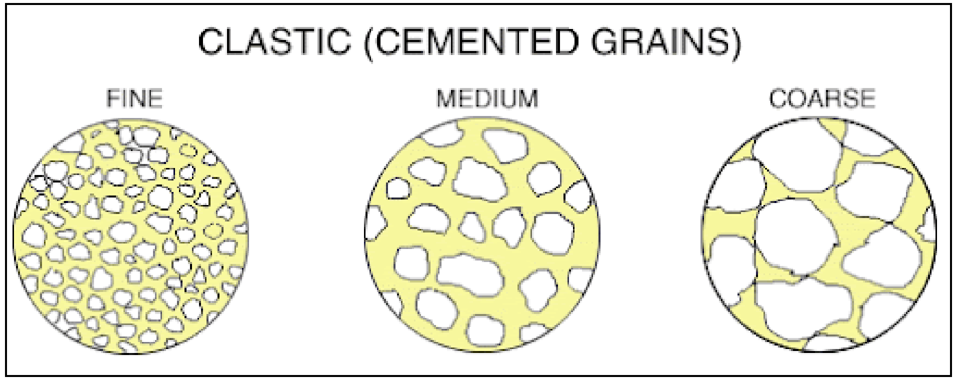
2- Porosity

3- Pore fluids

These three factors form density which is the most important factor affecting the velocity of waves.



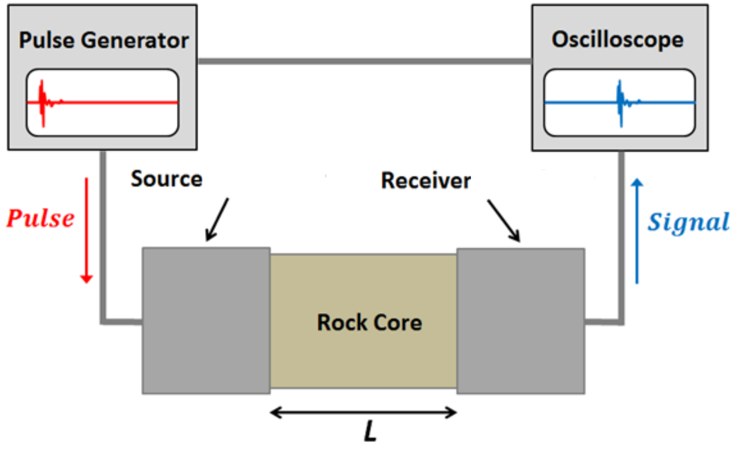
Texture refers to the sizes and shapes of grains, the relationships between neighboring grains, and the orientation of grains within a rock.



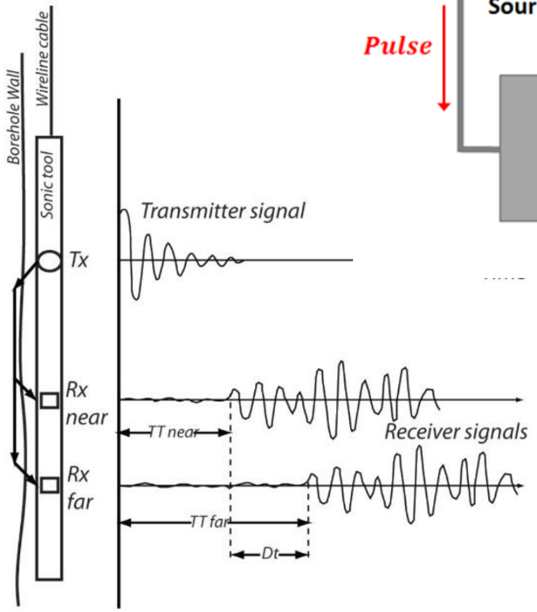
Velocity is measured by one of the following methods:

1-In laboratories

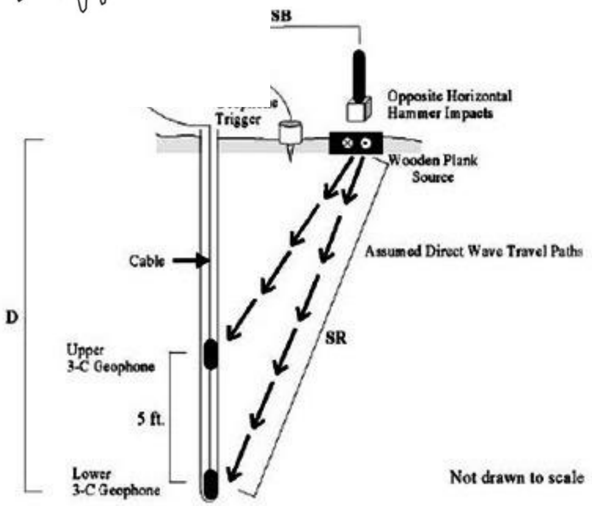
(from cylindrical specimens)



2-Acoustic logs (inside boreholes opposite to a certain rock type)

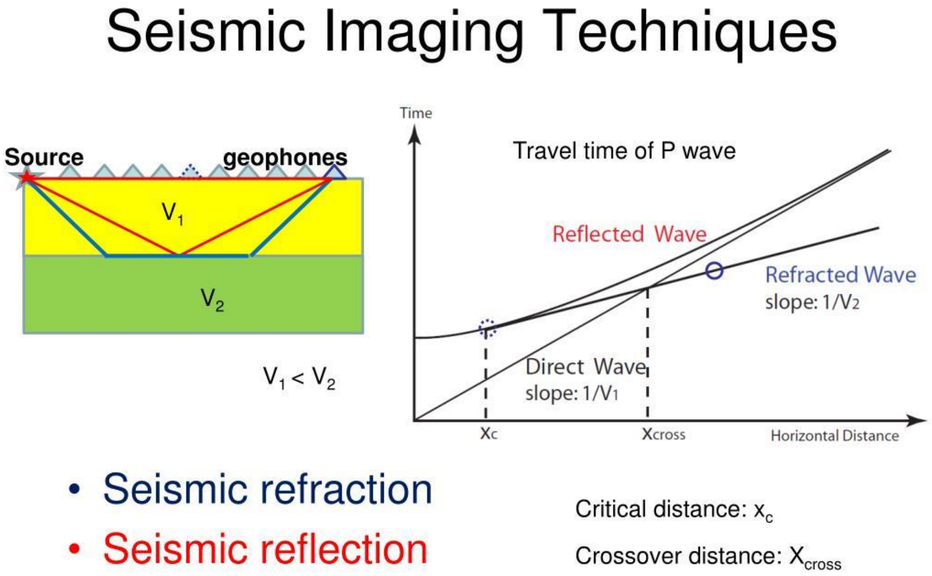


3-Up-hole shooting



4-Schmidt hammer system (remember your homework)

5-On-land seismic surveying (T-X diagram), will be the core material of our course



Velocity of waves in Sedimentary rocks:

Tend to have low velocities relative to Igneous and Metamorphic rocks, due to pore space that is filled with low velocity materials (air, water, hydrocarbons). Velocity will decrease as porosity increases.

Velocity increases with:

- **Depth (pressure) – higher pressures result in compaction which reduces the porosity.**
- **In addition, elastic moduli are larger at higher pressures.**

– Age – older rocks have greater degree of cementation, which results in an increase in rigidity

Sedimentary rocks are formed from pre-existing rocks or pieces of once-living organisms. They form from deposits that accumulate on the Earth's surface. Sedimentary rocks often have distinctive layering or bedding. Many of the picturesque views of the desert southwest show mesas and arches made of layered sedimentary rock. Examples: Limestone, Sandstone, Shale….

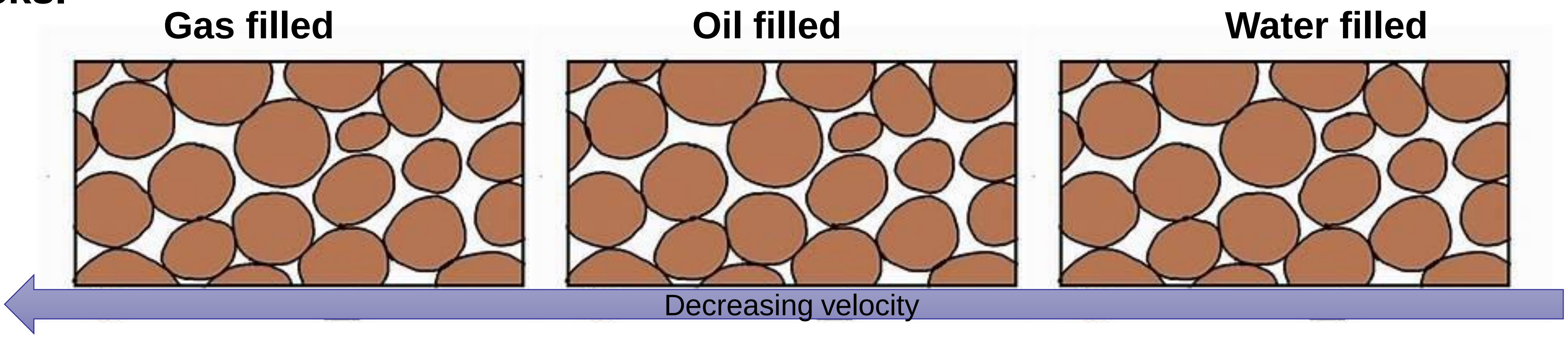
Compressional wave velocities in Earth materials.

	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

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.**

$1/V=Ø/V_f + (1-Ø)/V_m$

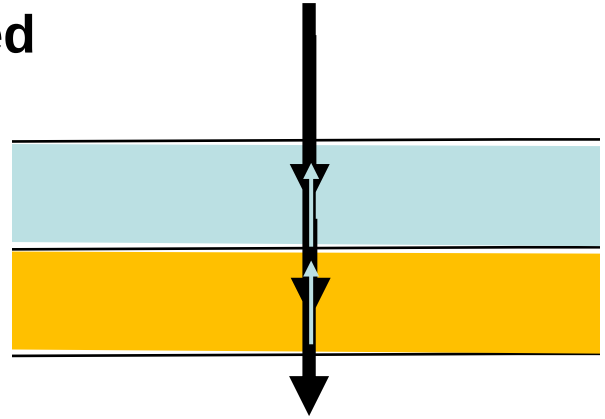
where Vp wave velocity for a rock (as a total), Ø porosity, V_f velocity of the pore fluid and V_m velocity of rock matrix

Details for this topic will be given in acoustic (sonic) well logging, 3rd grade.

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) of the medium and the angle of incidence.**

For normally incident seismic rays, at an interface between two rock layers there is generally a change in propagation velocity (as well as density) resulting from the difference in physical properties of the two layers. When an elastic wave hits such an interface, the energy within the incident wave is partitioned into transmitted and reflected waves. The relative <u>amplitudes</u> of the transmitted and reflected pulses depend on the velocities (V) and densities (ρ) and the angle of incidence. The simplest case is that of a normal incidence wave.



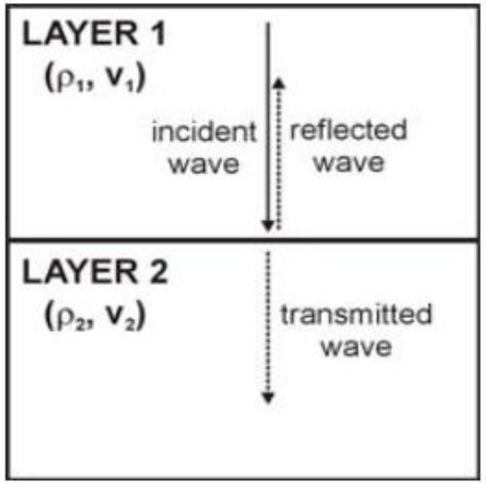
- <u>**Homework:**</u>
- **A P-wave of amplitude 1.0 is produced by an [illegible]thquake at the Earth's surf[illegible] and travels vertically downward where it encounters th[illegible]oho (boundary between cr[illegible] and mantle).**
- **What is the amplitude of the P-wave[illegible]hat are reflected and trans[illegible]ed at the Moho?**
-
- **Assume: v_c= 5000 m/s[illegible] ρ_c= 2850 kg/m^3 , v_m= 8000 [illegible], ρ_m = 3250 kg/m^3**
-

C=crust

M=mantle

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

amplitude A_i



amplitude A_r

amplitude A_t

The acoustic impedance (Z) of a material is defined as the product of its density (ρ) and velocity (V). It controls the partition to reflected and transmitted.

Z = vρ = acoustic impedance

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

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Lecture 11

Applied Geophysics Lecture Notes
Fourth Semester
By:

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Contents:

- Reflection and Transmission coefficients
- Changes of Propagation Direction at Interfaces
- Snell's law
- Seismic data acquisition systems
- Seismic Energy Sources
- Geophones

Reflection and Transmission coefficients

Consider a wave that encounters an interface at normal incidence. The incident wave has an amplitude of A_i. It will produce a reflected wave (amplitude A_r) and transmitted wave (amplitude A_t).

A_r depends on the <u>reflection coefficient</u> (R) of the interface:

$$\frac{A_r}{A_i} = R = \frac{Z_2 - Z_1}{Z_2 + Z_1}$$

$Z_1 = \rho_1 V_1$

$Z_2 = \rho_2 V_2$

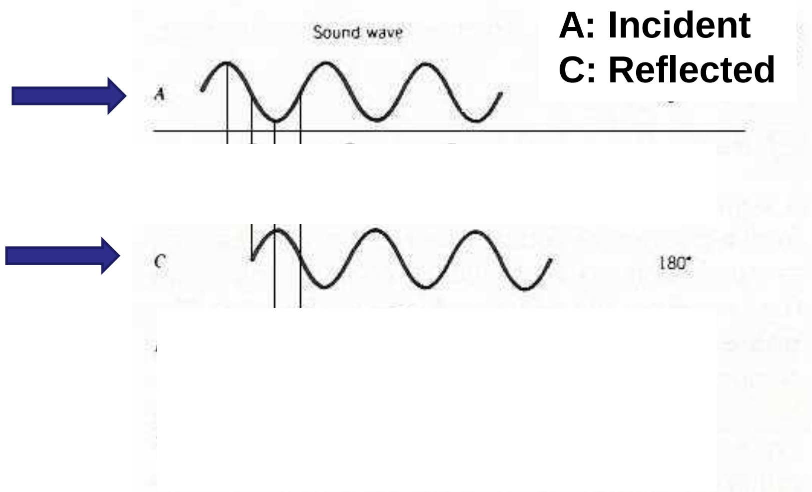
A_t depends on the <u>transmission coefficient</u> (T) of the interface:

$$\frac{A_t}{A_i} = T = \frac{2Z_1}{Z_2 + Z_1}$$

The more energy is reflected the greater the contrast.

$-1<R<1$

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



A: Incident
C: Reflected

Transmission coefficient:
$T=A_t / A_i$

$T=2\, Z_1/(Z_2+Z_1)$…………… $0 \leq T \leq 2$

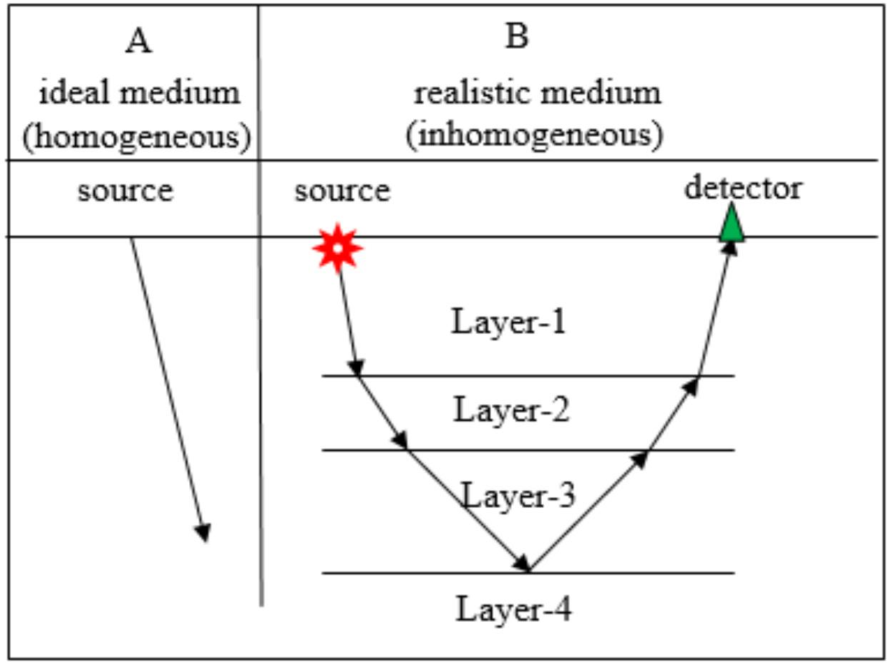
When R=0, there is no reflection

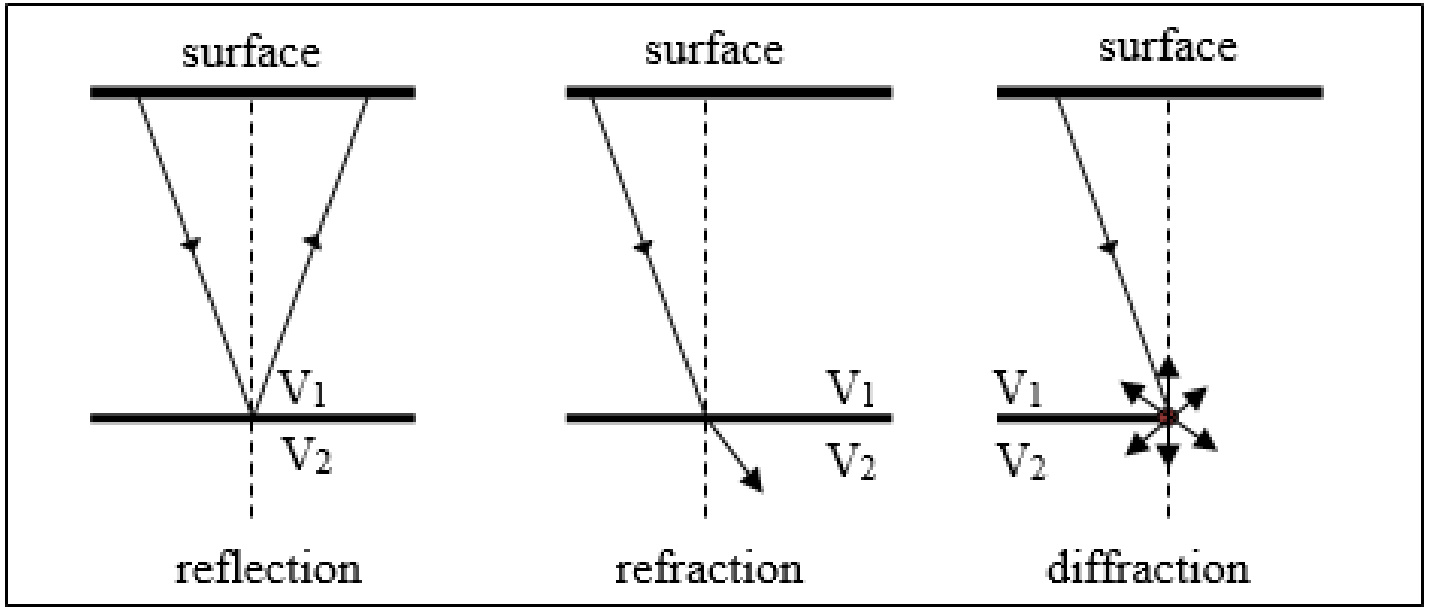
Changes of Propagation Direction at Interfaces

In an idealized homogenous and elastic medium, a seismic wave propagates with no changes taking place on ray-path direction or on the waveform of the travelling seismic pulse. In nature, however, the medium is far from this idealized form. In the solid crust of the Earth, it is commonly made up of rock layers of varying physical properties and varying geometrical forms and sizes.

In such inhomogeneous environments a moving seismic wave would suffer from a number of changes whenever it meets a change in the properties of the medium. <u>In particular, changes in energy content, waveform, propagation velocity, direction of motion, and new waves are generated at the interface planes.</u>

The common changes in ray-path direction, which are of significance to exploration seismology, are reflection, refraction (transmission) and diffraction. These shapes of the moving wave ray-path occur at the boundaries of media having different seismic propagation velocities





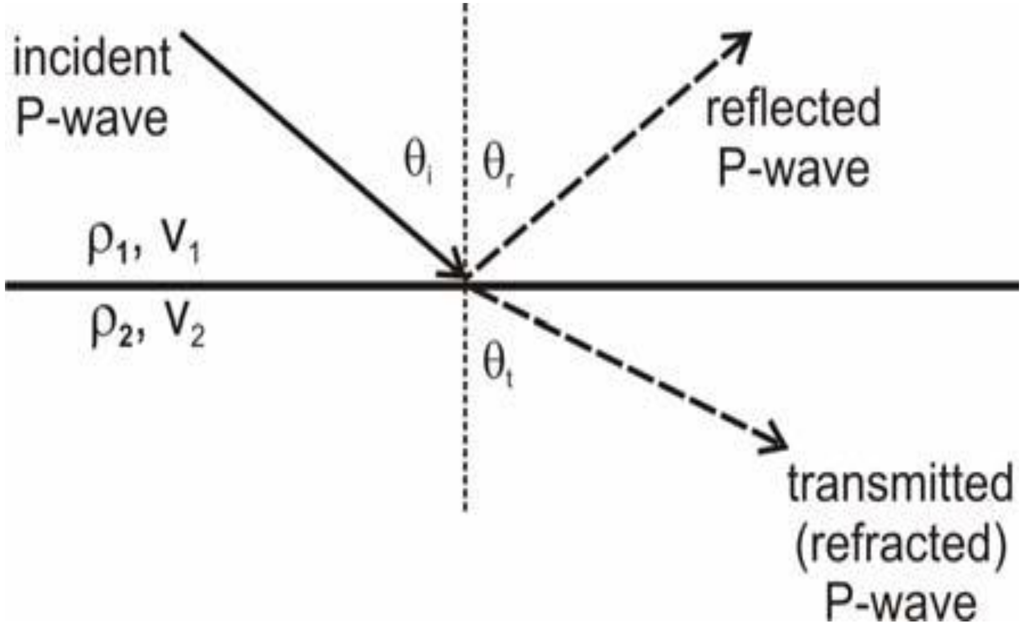
Snell's law
(Reflection and refraction of obliquely incident rays)

As a seismic wave encounters an interface between two different materials, some energy will be reflected off the interface and some will be transmitted through the interface.

Each wave is characterized by a ray parameter:

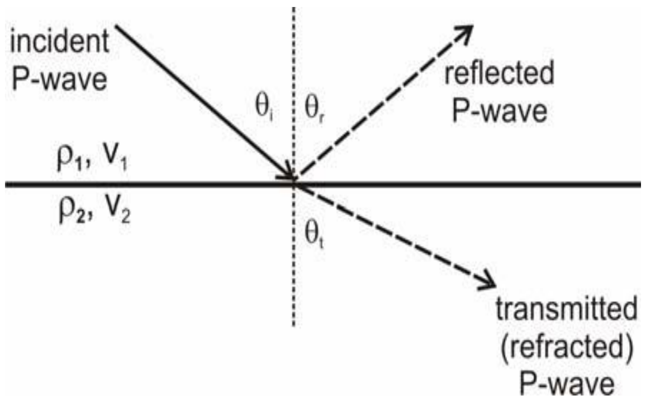
$$p = \frac{\sin\theta}{v}$$

were
θ is the angle between the wave and the normal of the interface, and v is the velocity of waves in the material.

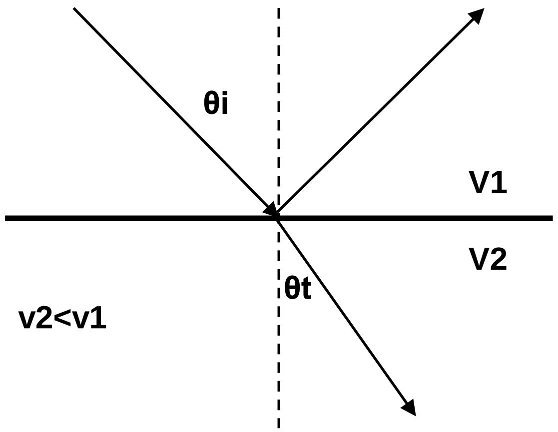


Snell's Law: The ray parameter is the same for all waves at the interface.

$$p = \frac{\sin\theta_i}{v_1} = \frac{\sin\theta_r}{v_1} = \frac{\sin\theta_t}{v_2}$$

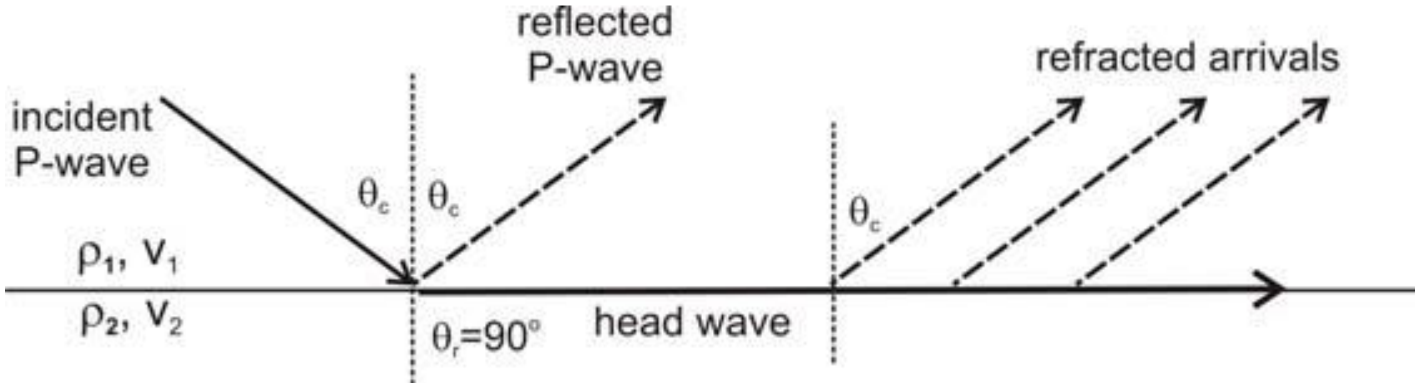


The reflection angle is always the same as the incidence angle: θi= θr
A change in velocity across the interface results in a change in the direction of travel of the wave that passes though the interface – the wave is refracted
If velocity increases (v_2>v_1), the refracted wave is bent away from the normal (θt>θi)
If velocity decreases (v2<v1), the refracted wave is bent toward the normal (θt<θi)



Critical angle:

If $v_2 > v_1$, there will be an angle of incidence that produces a refracted wave with an angle of 90° (i.e., it travels along the interface) – this is the critical angle (θ_c).



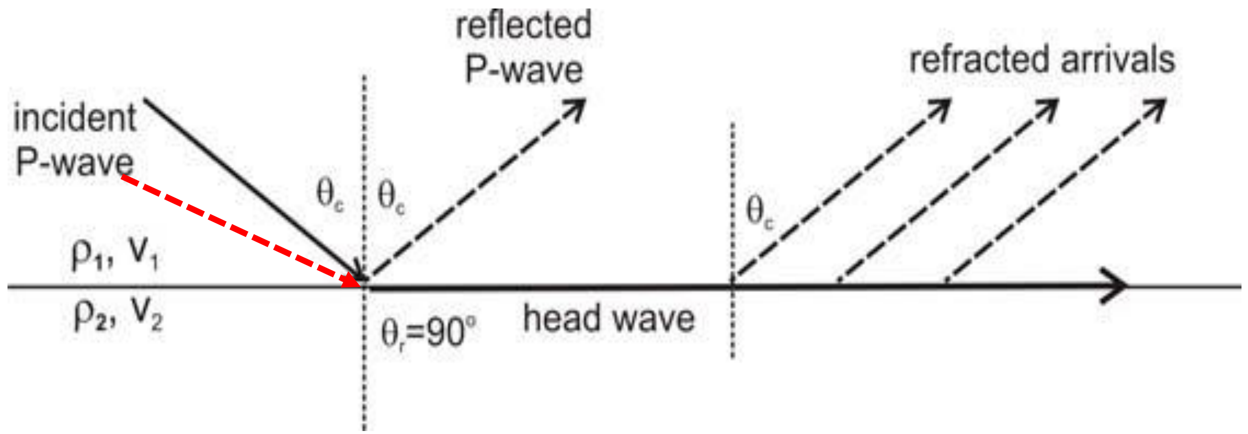
$$\frac{\sin\theta_c}{v_1} = \frac{\sin(90^o)}{v_2} = \frac{1}{v_2}$$

$$\rightarrow \theta_c = \sin^{-1}\left(\frac{v_1}{v_2}\right)$$

The refracted wave is called a **head wave**, and it will travel along the interface with a velocity v_2.

As it travels, secondary waves will be produced (Huyghens' Principle) that travel upward toward the surface at an angle θc. These are called refracted arrivals.

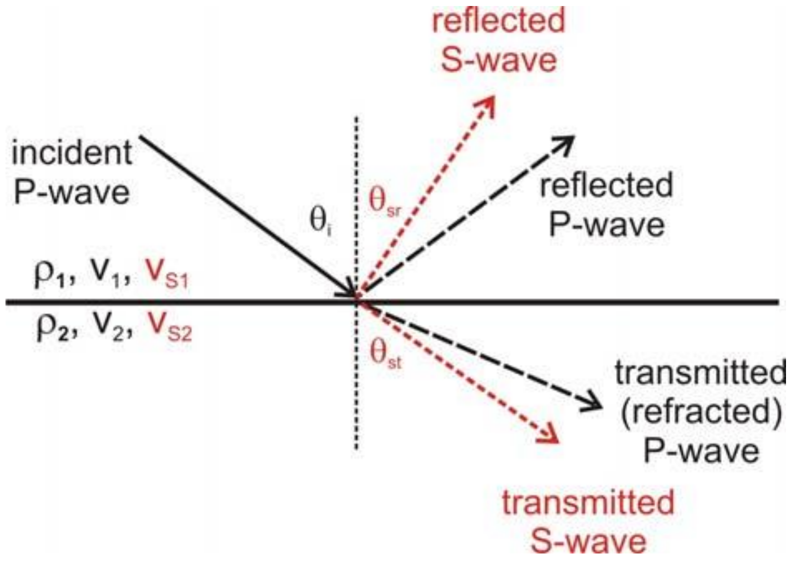
At angles of incidence greater than the critical angle ($\theta i > \theta c$) –the dotted incident ray-, there is no refracted energy. All energy is reflected at the incidence angle θi. This is total internal reflection.



Mode conversion:
In addition to the reflected and refracted P-waves, an incident P-wave will also generate a reflected S-wave and a refracted S-wave. Note that these are SV-waves (partitioning of seismic energy)

The reflection and refraction angles of the S-waves are also given by Snell's law:

$$p = \frac{\sin\theta_i}{v_1} = \frac{\sin\theta_{sr}}{v_{s1}} = \frac{\sin\theta_{st}}{v_{s2}}$$



Geometry of rays associated with wave partition. No Partition in normal incidence

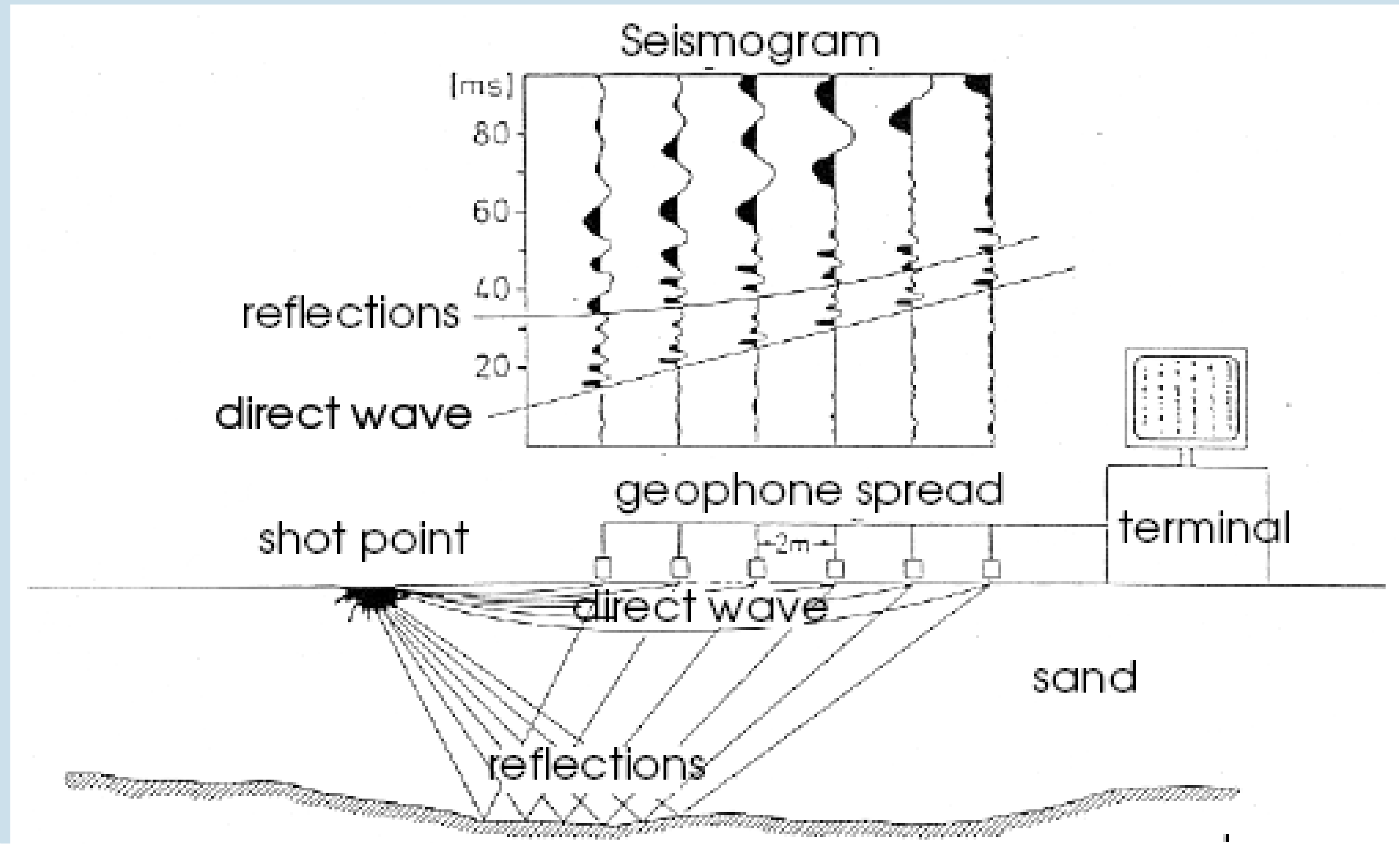
Seismic data acquisition systems

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.



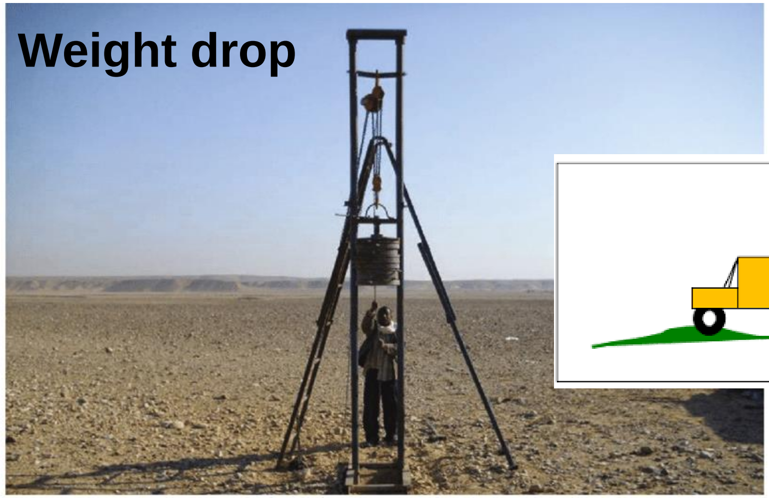
Seismic Field-work
(source, geophone and terminals)

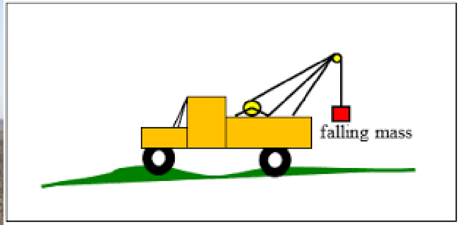
Seismic Energy Sources:

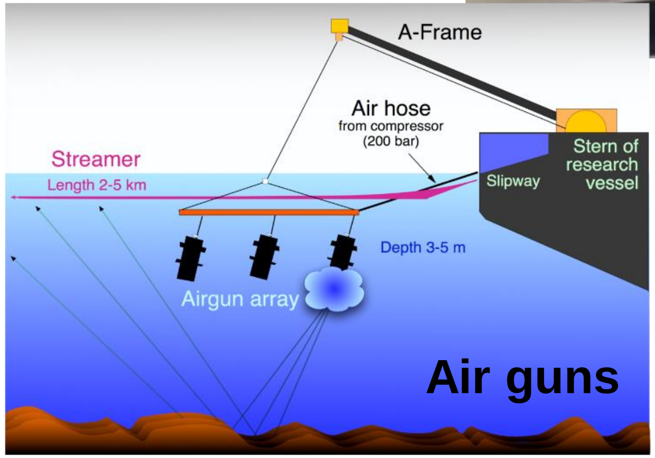
The aim is to produce a large enough signal into the ground to ensure sufficient depth penetration and high enough resolution to image the subsurface. There are a large varieties of energy sources. Choosing one depends upon the aim of the survey and environment for the area of work.

There are three types of seismic sources to be used on land, in water and down borehole.

Seismic sources on land:
1- Impact, involving Sledgehammer and Drop-Weight.
2- Impulse, involving Dynamite and Air guns (air Guns are also used in marine surveys and down borehole).
3- Vibrator, involving Vibroseis.







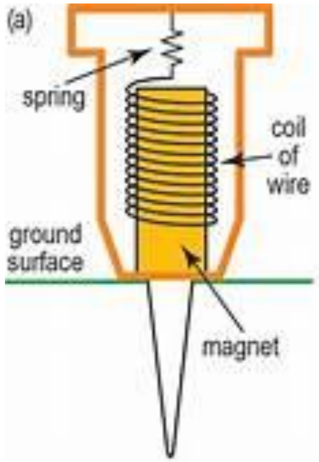
Examples of Energy Sources

https://www.youtube.com/watch?v=4El6U0XTNS0&pp=yg
UTdmlicm9zZWlzIGluIGFjdGlvbg%3D%3D

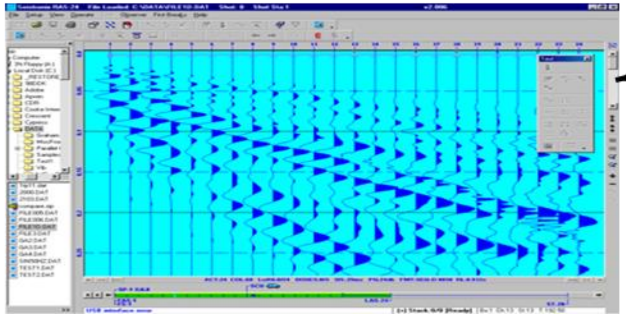
Geophones (seismometers) and Seismograms:

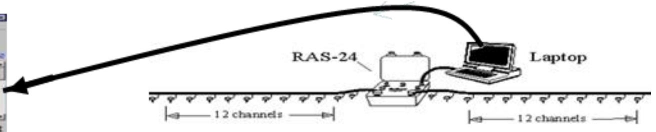
Conversion of the ground motion to an electrical signal requires a geophone. It is sensitive to ground motion. Devices used on land to detect seismic ground motions are known as seismometers or geophones. The relative movement of the magnet with respect to the coil results in a small voltage.

Geophones are implanted into the ground firmly and vertically and used to convert seismic energy (land vibration) into a measurable electrical voltage.



1. The recording must be timed accurately relative to the seismic source. Times need to be accurate to better than one thousandth of a second (one millisecond).
3. The electrical signals must be stored for future use.





Geophone: https: youtu.be/EaQ9cERXBf4

The monument of the oldest seismometer in the world which was built nearly 2000 years ago by a Chinese genius named Zhang Heng. It was used to know the direction of earthquake waves.

Data acquisition: https://youtu.be/bFFnNHm4_QM

Seismic reflection surveying

Seismic reflection surveying is the most widely used and well-known geophysical technique. The current state of sophistication of the technique is largely a result of the enormous investment in its development made by the hydrocarbon industry, coupled with the development of advanced electronic and computing technology.

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

For shallow investigations reflection method is used in:

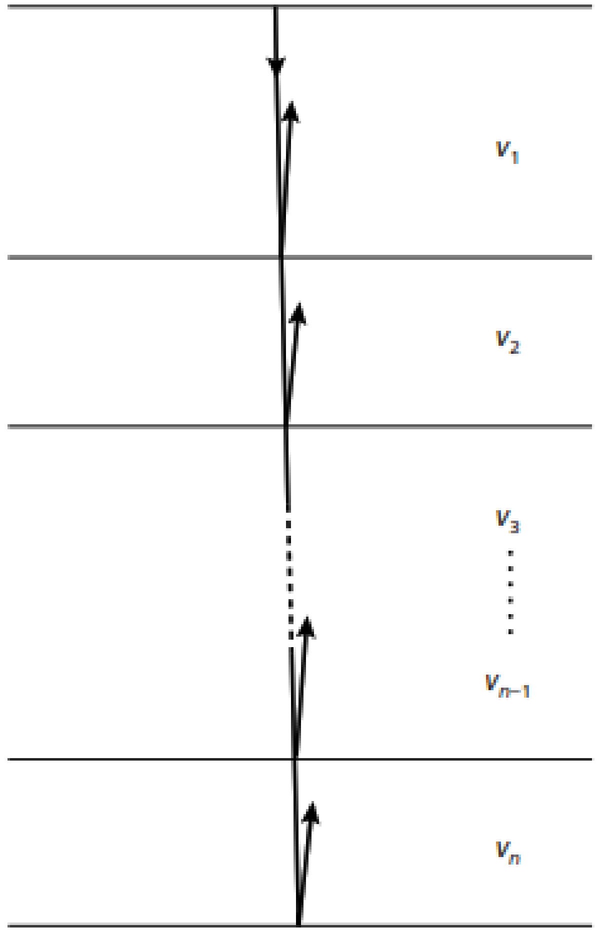
- engineering and environmental investigations (< 200 m)
- mapping quaternary deposits, buried valleys, shallow faults.
- hydrogeological studies of aquifers.
- shallow coal exploration.
- preconstruction.
- ground investigations for pipe.

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

The time is known as the two way travel time *<u>(TWTT).</u>*
The most important problem in seismic reflection surveying is the translation of TWTT to depth.
While travel times are measured, the one parameter, that most affects the conversation of to depth is seismic velocity.

➔ *two unknowns (depth + velocity)*

Geometry of reflected ray paths



Vertical reflection ray path in
A horizontally layered ground

The figure shows a simple physical model of horizontally-layered ground with vertical reflected ray paths from the various layer boundaries. This model assumes each layer to be characterized by an interval velocity V_i, which may correspond to the uniform velocity within a homogeneous geological unit or the average velocity over a depth interval containing more than one unit. If z_i is the thickness of such an interval and t_i is the one-way travel time of a ray through it, the interval velocity is given by

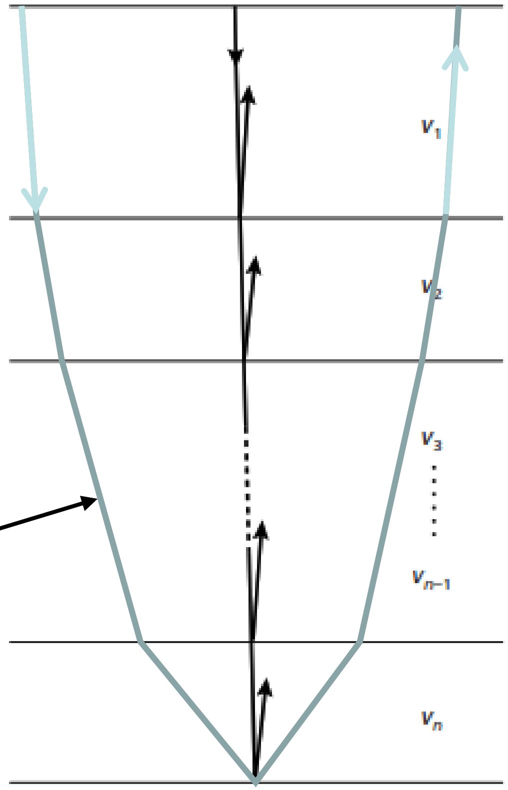
$$v_i = \frac{z_i}{\tau_i}$$

The interval velocity may be averaged over several depth intervals to yield a *time-average velocity* or, simply, *average velocity.Thus the average velocity of the top n layers in the figure* is given by:

$$\overline{V} = \frac{\sum_{i=1}^{n} z_i}{\sum_{i=1}^{n} \tau_i} = \frac{\sum_{i=1}^{n} v_i \tau_i}{\sum_{i=1}^{n} \tau_i}$$

if Z_n is the total thickness of the top *n layers and* T_n is the total one-way travel time through the *n layers*,

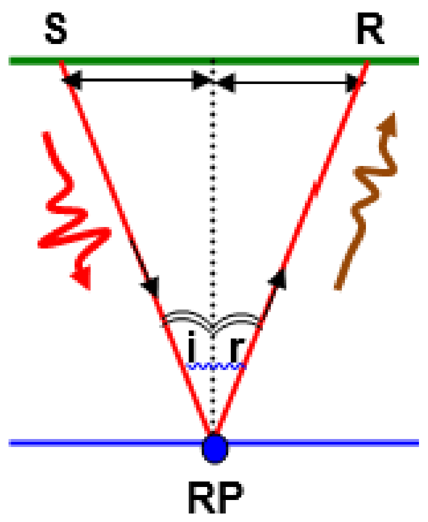
$$\overline{V} = \frac{Z_n}{T_n}$$



The reflection process involves two main types of changes. These are:

(i) Change in Energy content:

The energy content of the incident wave is shared among all of the reflected and transmitted waves. Consequently, the amplitude of any of the reflected waves is always less than that of the incident wave.

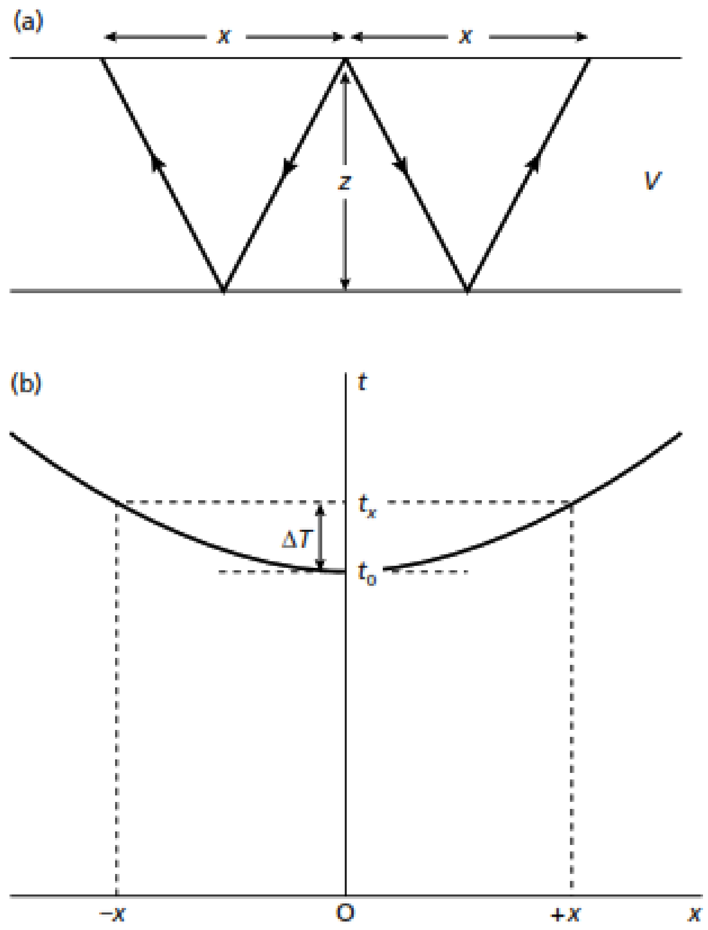


(ii) Change in Propagation Direction:

At the interface, part of the incident seismic energy is reflected following a travel path defined by the law of reflection which states that angles of incidence and reflection are equal provided that these are of the same wave types.

Single horizontal reflector

The basic geometry of the reflected ray path is shown (a) for the simple case of a single horizontal reflector lying at a depth *z beneath a homogeneous top* layer of velocity *V. The equation for the travel time t (TWTT) of* the reflected ray from a shot point to a detector at a horizontal offset, or shot–detector separation, *x is given by* the ratio of the travel path length to the velocity

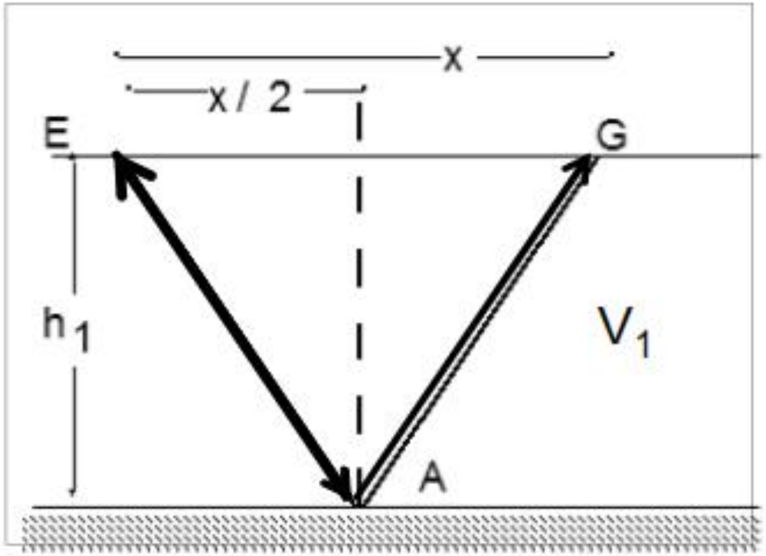


$$t = (x^2 + 4z^2)^{1/2} / V$$

(b) time–distance curve for reflected rays from a horizontal reflector.

Still we have two unknowns; <u>z and V</u>

Explanation



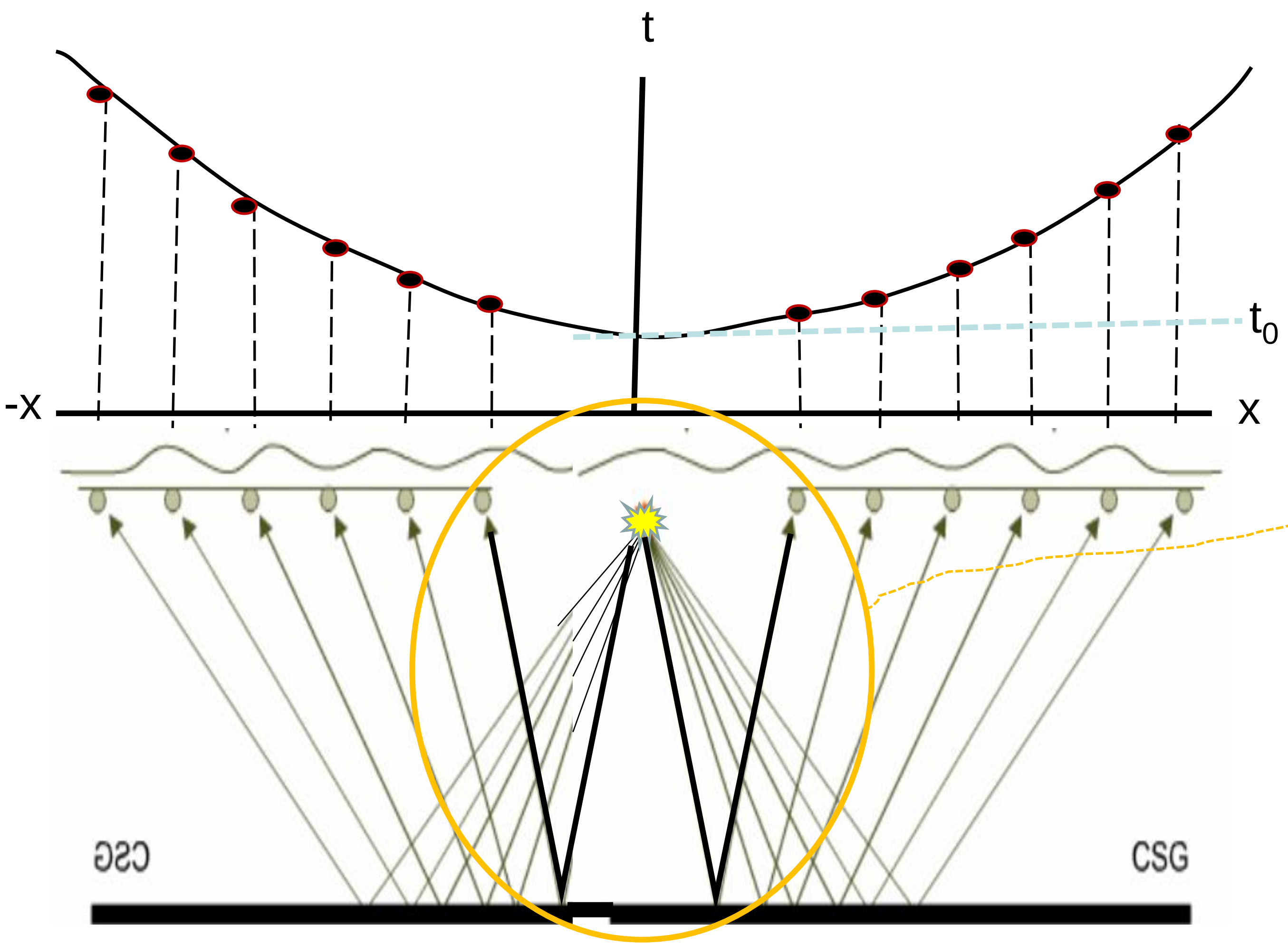
How long does it take a ray to travel from E to A to G?

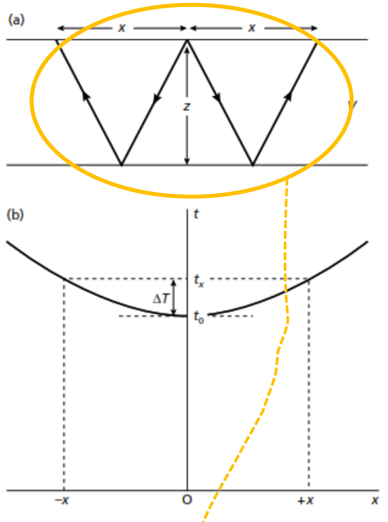
Basic formula:

The reflection path E-A-G, is symmetric, thus the time it takes to travel from the surface to the reflector and back is:

$$time = 2 * EA / V_1 = \frac{2 * \sqrt{\left(\frac{x}{2}\right)^2 + h_1^2}}{V_1} = \frac{\sqrt{x^2 + 4h_1^2}}{V_1}$$

Where has this figure come from?



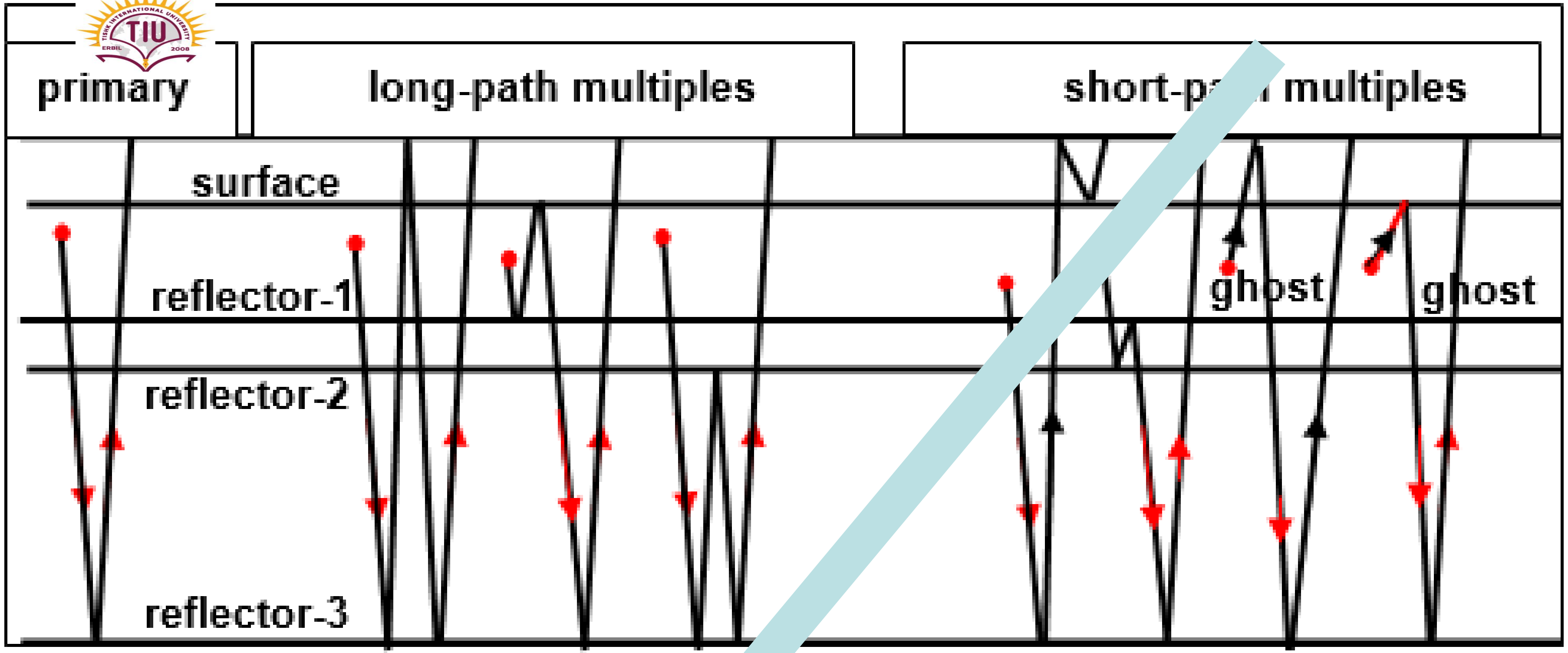


X	t
X1	t1
X2	t2
X3	t3
X4	t4
X5	t6
x6	t7

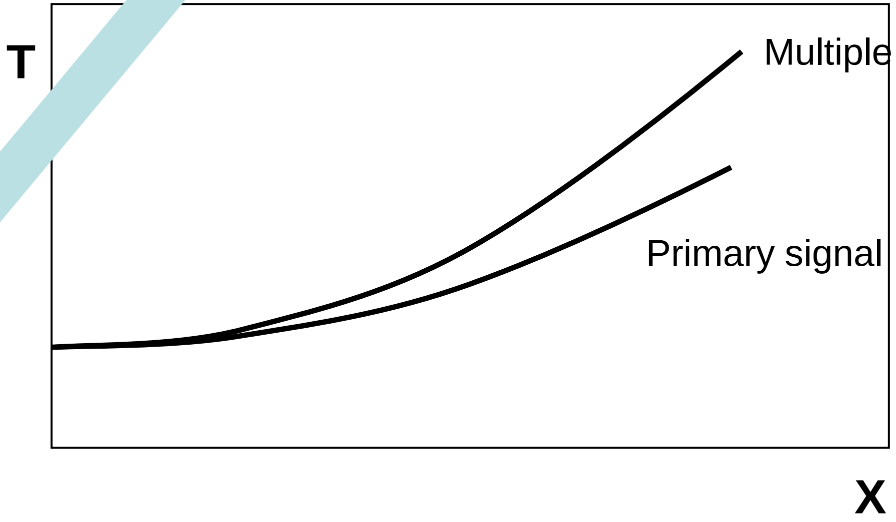
Multiples

It is possible that a seismic wave arrives at a point on the surface after being reflected several times from a number of interfaces. The first arrival (called the primary reflection) is the strongest, followed by other arrivals (the multiple reflections) which are of less energy.

A well-known type of multiple is the ghost reflection which occurs when a wave travels upwards from a source-point located at a certain depth and is reflected by the free earth surface or by the base of a low-velocity surface layer (the weathering zone).



The multiple arrives at a short time-interval after the primary reflection



Intercept time

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

Substitute x = 0 in the equation: $t = (x^2 + 4z^2)^{1/2} / V$

you will get $t_0 = 2z/v$

see the previous page.

This is the travel time

of a vertically reflected ray

(intercept on the time axis of the time distance curve).

Rewrite the equation $t = (x^2 + 4z^2)^{1/2} / V$ **at** X_0

So as to be:

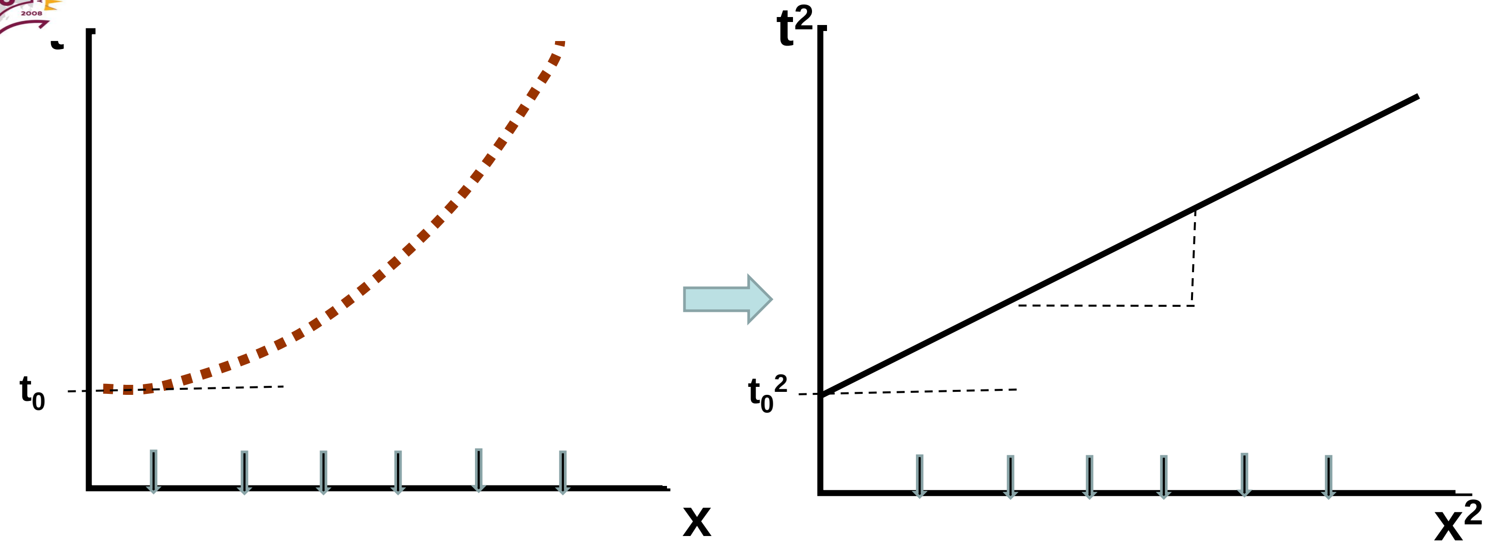
$t_0^2 = 4z^2 / v^2$

→ **This is the simplest way of determining the velocity**

NOW

Plot t^2 **against** x^2 **(this procedure converts hyparbula ti straight line)**

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



$X1^2$	$t1^2$
$X2^2$	$t2^2$
$X3^2$	$t3^2$
$X4^2$	$t4^2$
$X5^2$	$t5^2$
$X6^2$	$t6^2$

slope = $1/v^2$

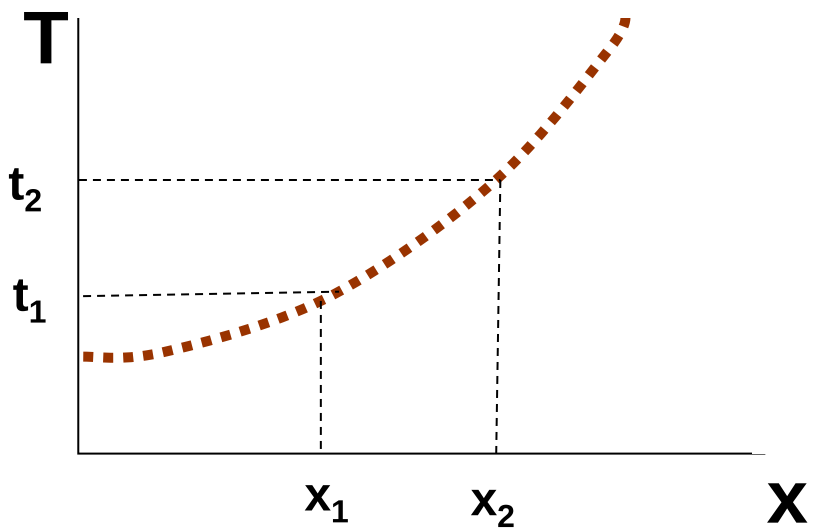
$t_0^2 = 4z^2 / v^2$

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 t1 and t2 of reflected ray arrivals recorded at two offset distances x1 and x2.

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

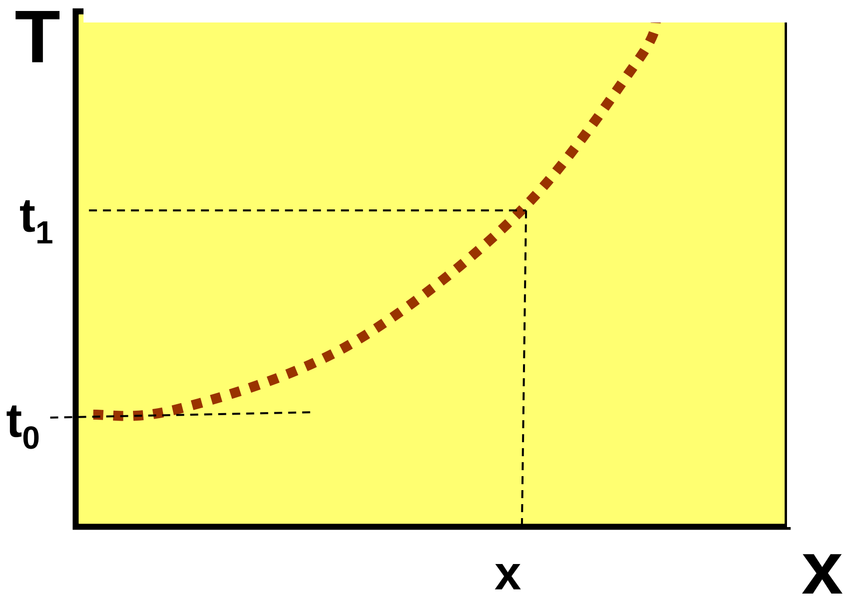


The effect of the separation between receiver and source on the arrival time of a reflection that does not dip, abbreviated NMO. A reflection typically arrives first at the receiver nearest the source. The offset between the source and other receivers induces a delay in the arrival time of a reflection from a horizontal surface at depth. A plot of arrival times versus offset has a hyperbolic shape.

The effect of the separation between receiver and source on the arrival time of a reflection that does not dip, abbreviated NMO. A reflection typically arrives first at the receiver nearest the source. The offset between the source and other receivers induces a delay in the arrival time of a reflection from a horizontal surface at depth. A plot of arrival times versus offset has a hyperbolic shape.

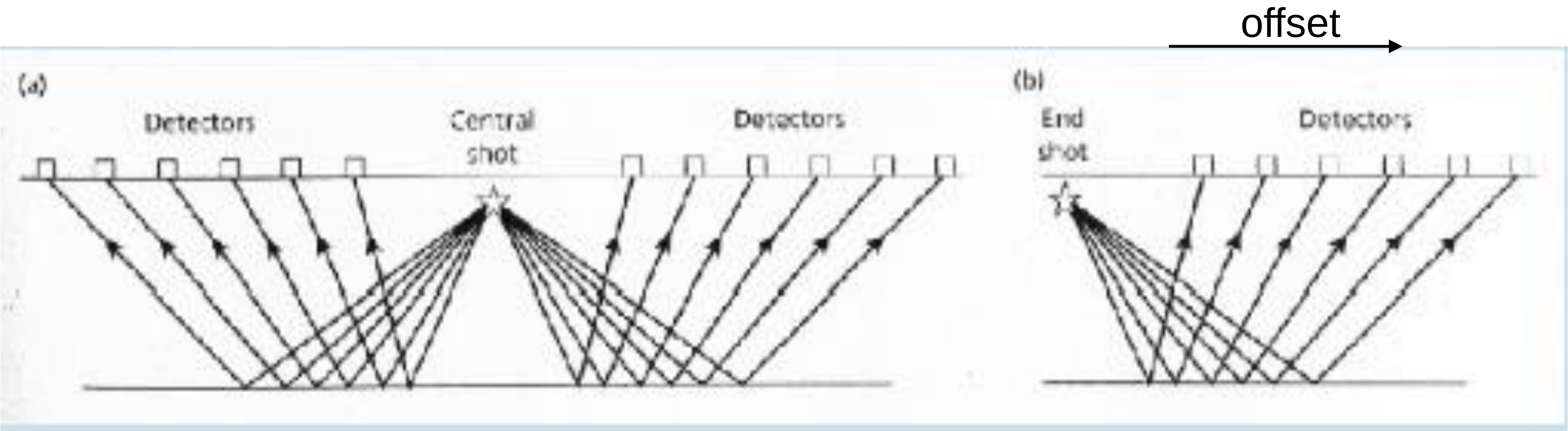
- **Normal moveout (NMO) at an offset distance x is the difference in travel time ΔT between reflected arrivals at x and at zero offset.**

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



Data acquisition

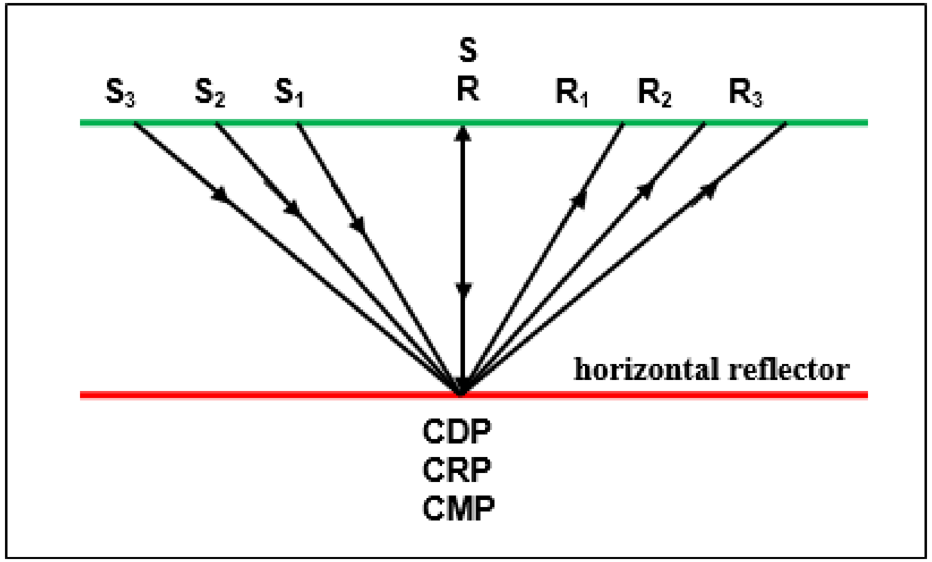
The initial display of seismic profile data is normally in groups of seismic traces recorded from a common shot, known as shot gathers.



Shot-detector configurations used in multichannel seismic reflection profiling. (a) Split spread, (b) Single-ended or one-end spread.

CDP, CRP, and CMP

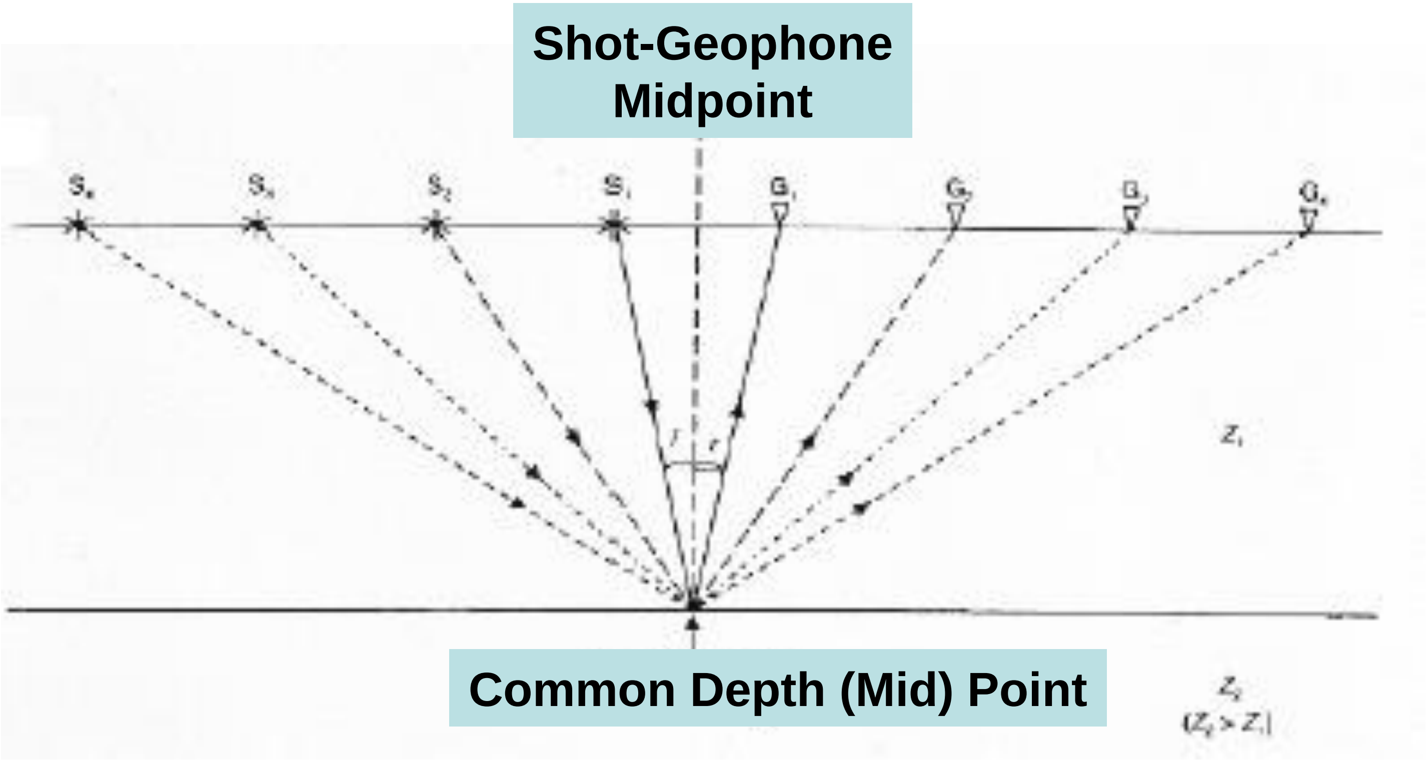
By use of certain source-receiver layout, it is possible to shoot a number of shots such that the reflection points resulting from this number of shots coincide on each other, that is one point-location will serve as common point to all of the implemented shots. In this type of repeated shooting-spread, the reflection point becomes known as common depth point, common reflection point, or common mid-point. These are normally abbreviated as CDP, CRP, CMP respectively



The four reflections ray-paths, shown in this figure, have one common depth point (the CDP), and each produces a *seismic trace*. These traces belong to the same depth point (CDP) and thus they form a group of traces called the (CDP-Gather). In this case (case of horizontal reflector) the CDP and the other points (CRP and CMP) will coincide on each other. However, these points do not fall at one location-point when the reflector is dipping.

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

For example: 4 different shot-geophone locations give 4 fold coverage



Principle of the common Midpoint over a horizontal interface.

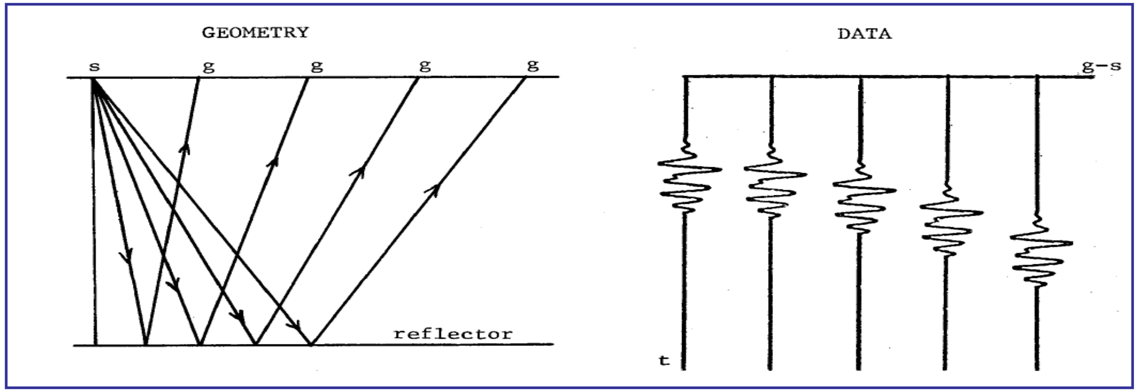
The CMP gather lies in the heart of seismic processing for two main reasons:

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.

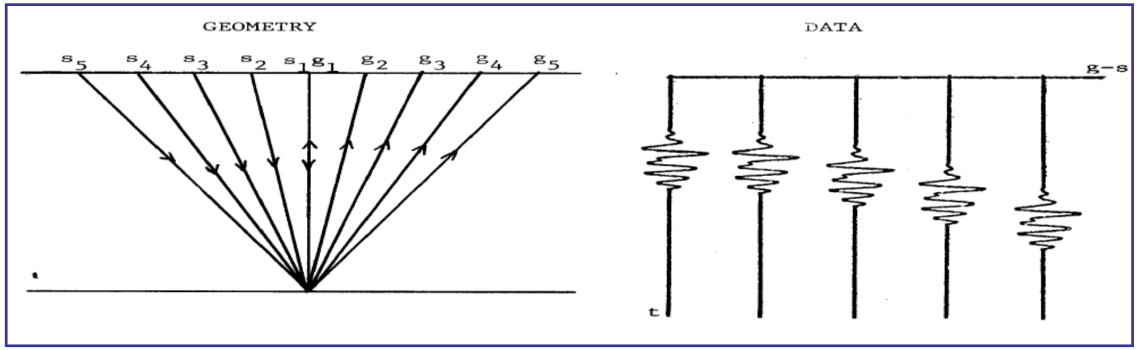
2) The reflected seismic energy is usually very weak. It is imperative to increase the *signal/noise* ratio of most data.

NMO Correction

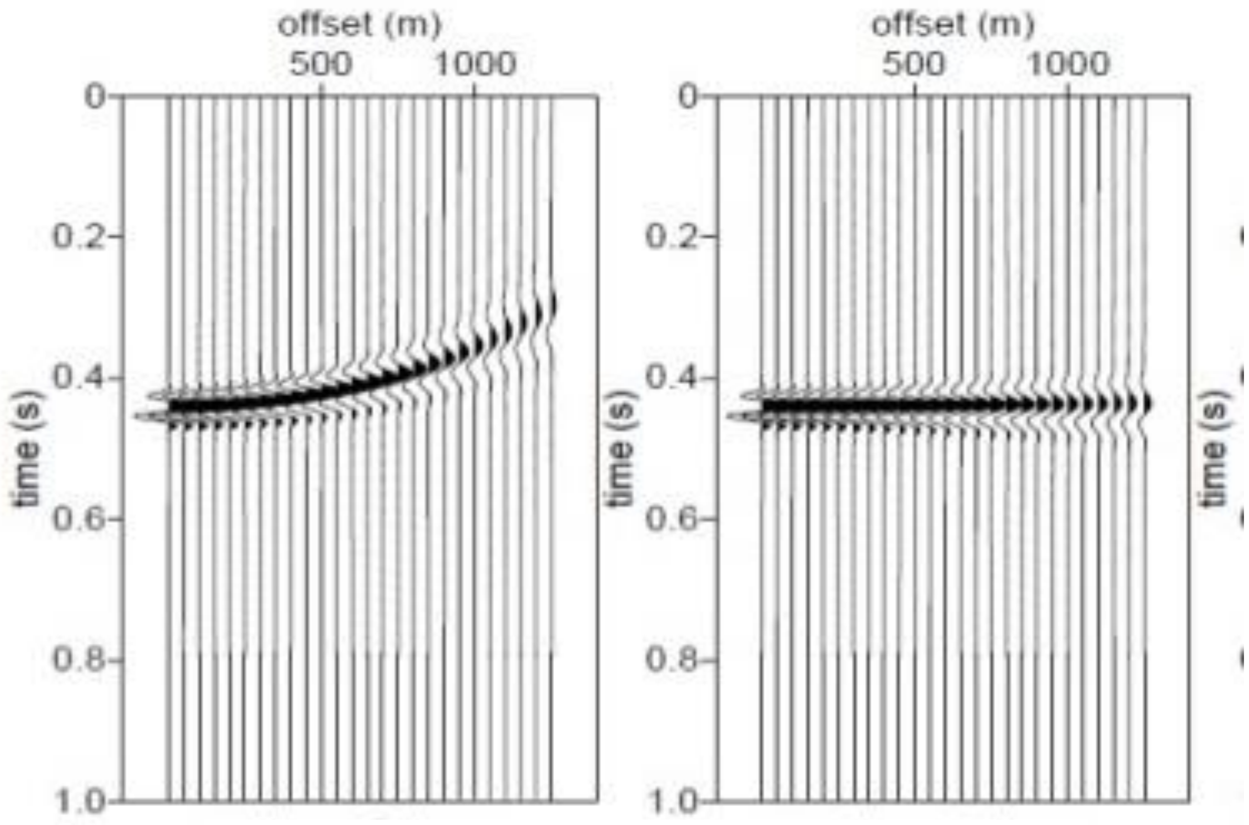
A function of time and offset that can be used in seismic processing to compensate for the effects of normal moveout, or the delay in reflection arrival times when geophones and shotpoints are offset from each other.



What is the difference?

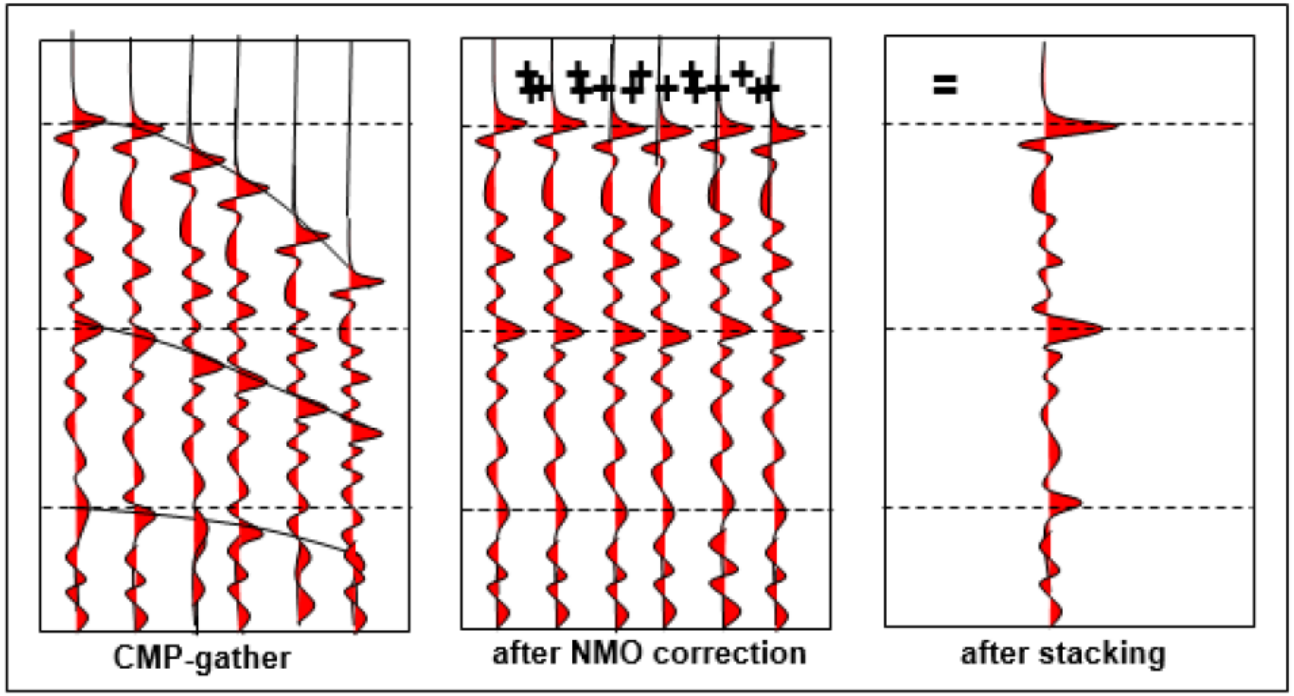


A processed seismic record that contains traces that have been added together from different records to reduce noise and improve overall data quality (i.e. to increase Signs/Noise ratio).



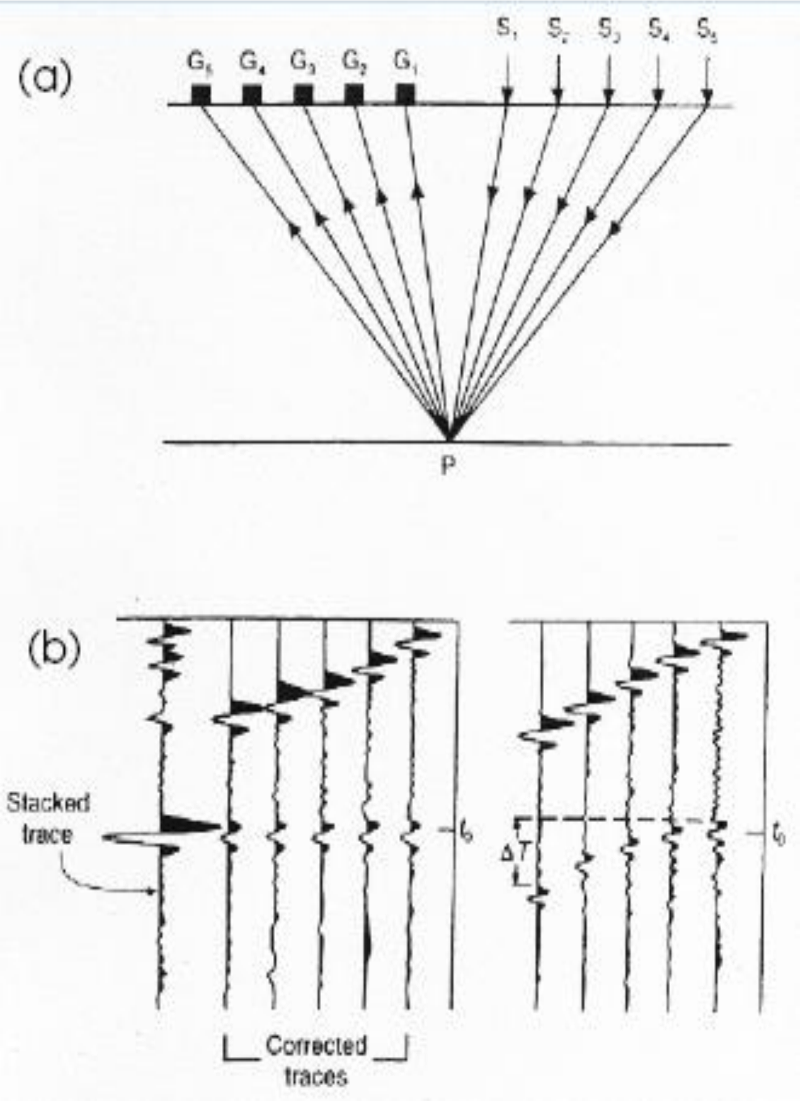
Before After

A processed seismic record that contains traces that have been added together from different records to reduce noise and improve overall data quality (i.e. to increase Signs/Noise ratio).



CMP-stacking of CMP gather-traces after being NMO-corrected

Given the Source receiver layout And corresponding raypaths 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. The number of traces that have been added together during stacking is called the fold.



<u>Migration</u>

is the process of reconstructing a seismic section so that reflection events are repositioned under their correct surface location and at a corrected vertical reflection time; that is the transformation of apparent reflection positions to true positions.

<u>On unmigrated stacked sections, reflections are always plotted below the mid-point, which is only correct for horizontal layers.</u>

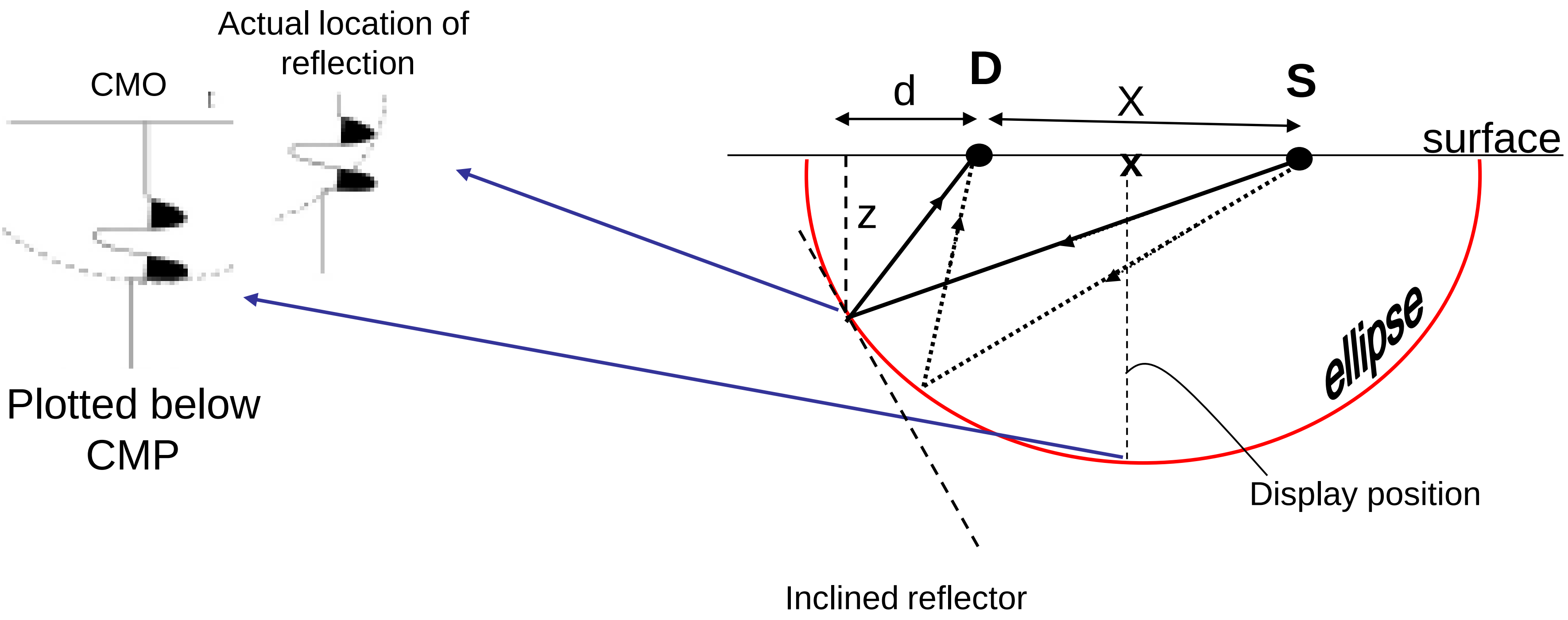
For horizontal reflectors the reflection point is vertically below the source/receiver midpoint.

For dipping layers the reflection comes from a point up dip.

Therefore, a traveltime section will always show a reduced dip

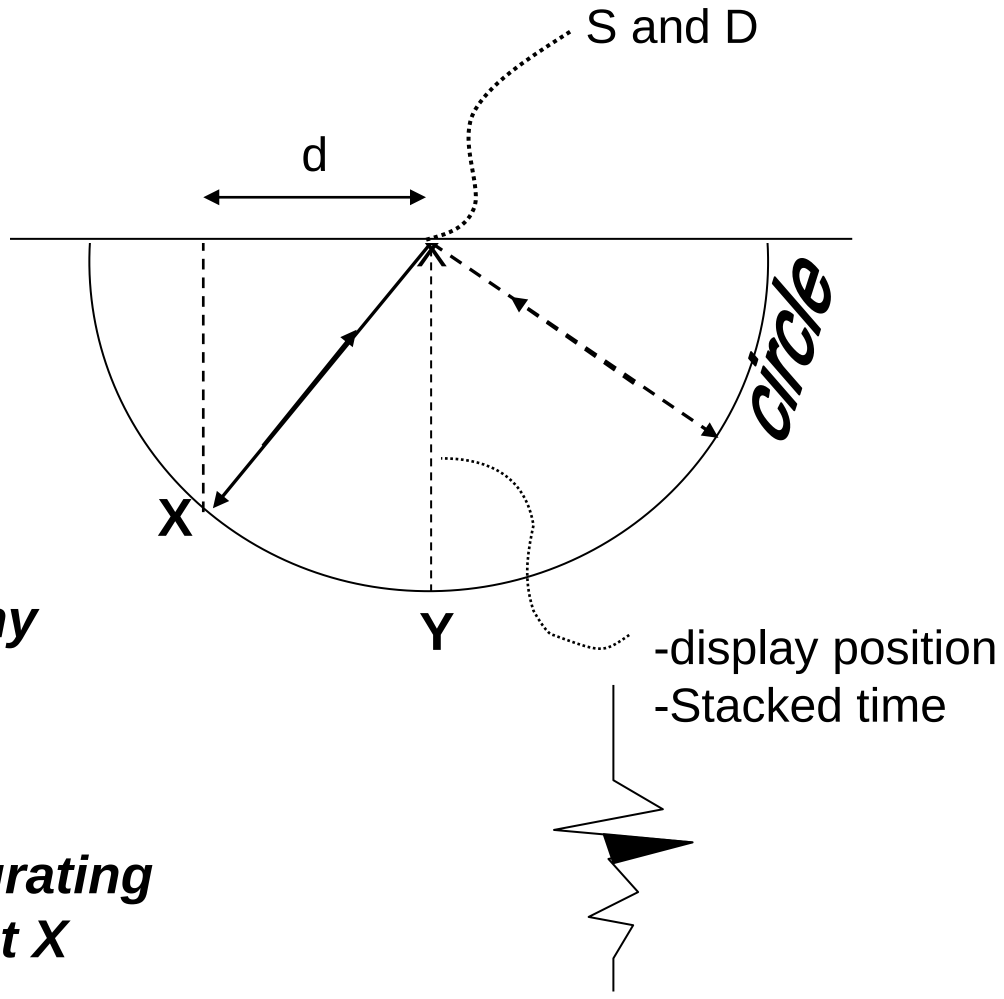
The reflection point is any point on the surface of the ellipse With two foci S and D at depth Z

$$T_Z = 1/V\,\sqrt{Z^2 + d^2} + 1/V\,\sqrt{Z^2 + (d+X)^2}$$



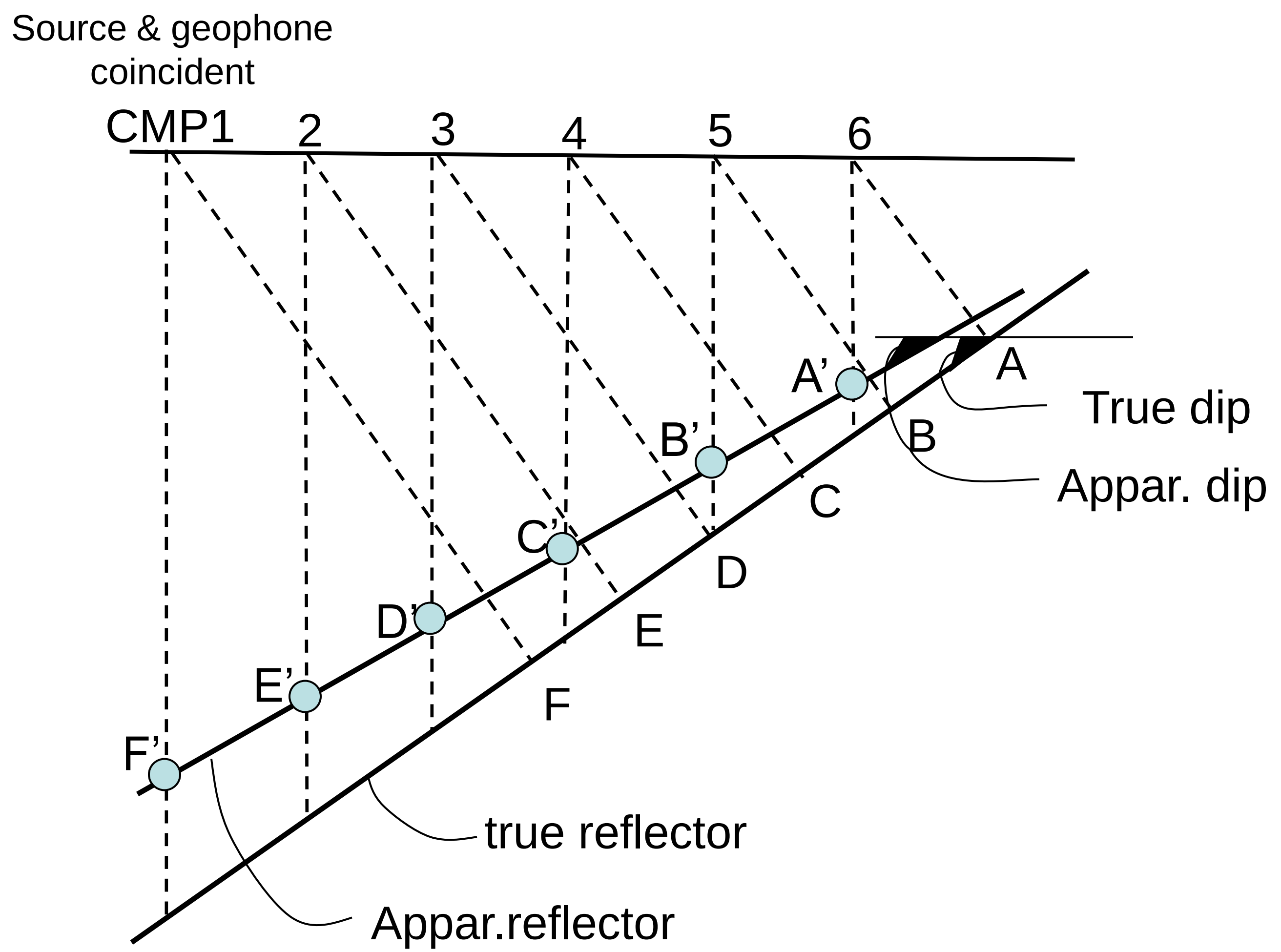
When the S and D are among a Common Mid Point gather, the ellipse is gone to be a circle (i.e. the two points coincide each other)
Then

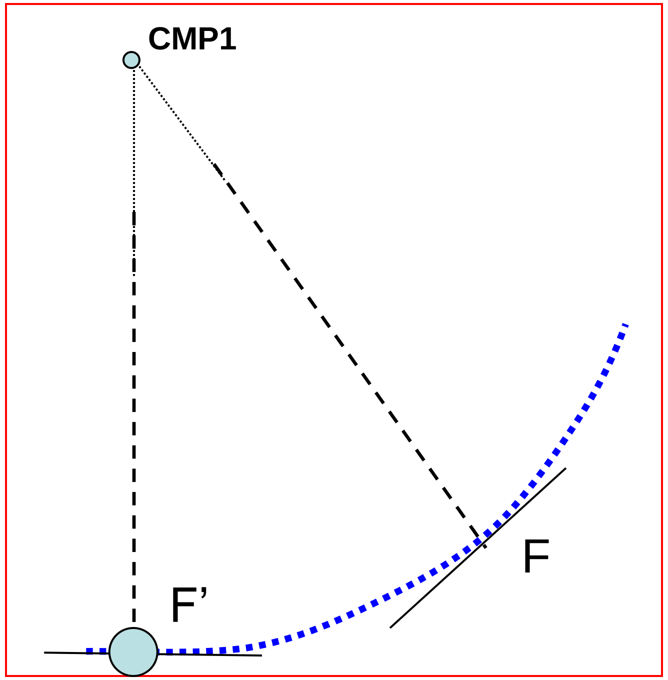
$$T_Z = 2\sqrt{(Z/V)^2 + (d/V)^2}$$

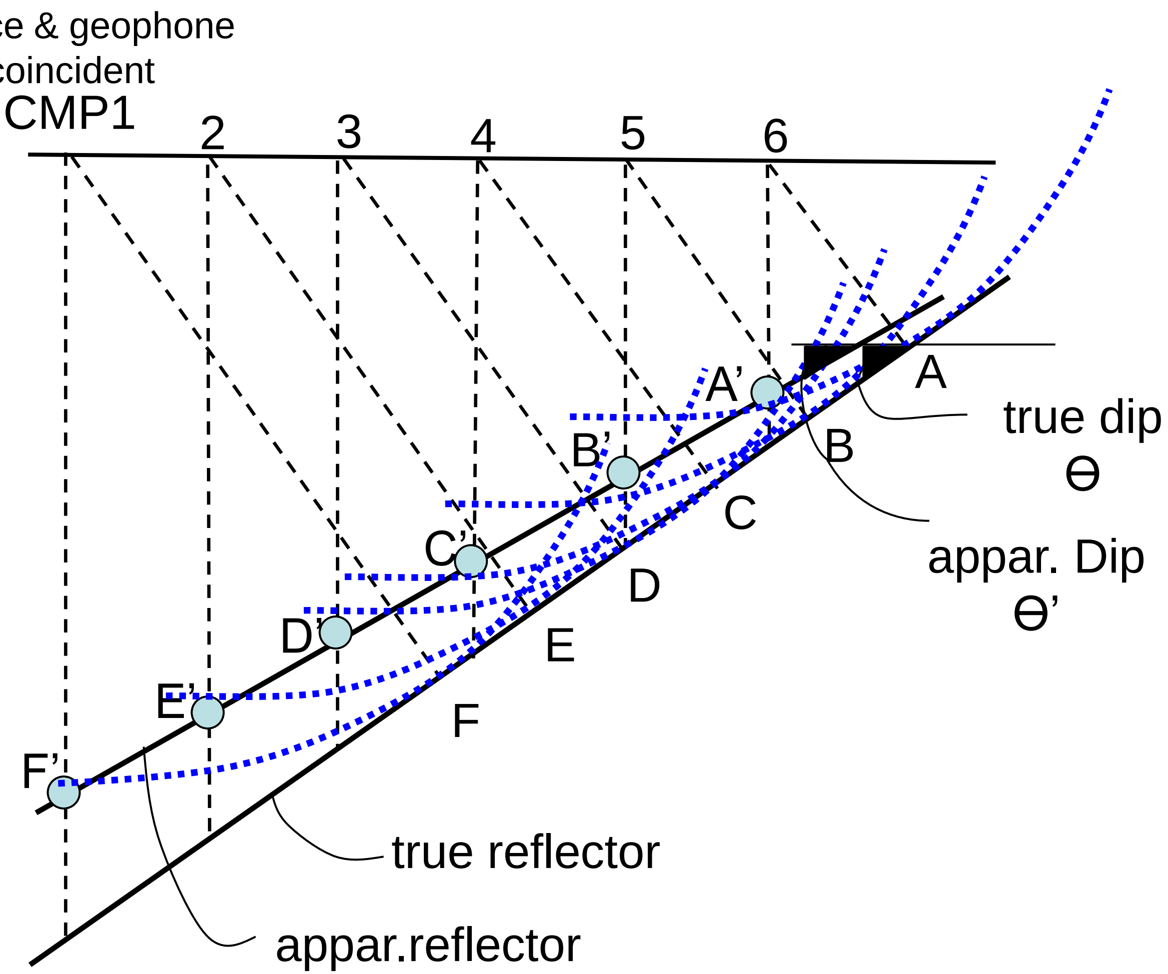


Explanation: the vertical time is any time reflected from the surface of The circle, it will be vertical <u>only</u> when the reflector is horizontal

<u>So</u> migration is the process of migrating the point Y to the actual position at X

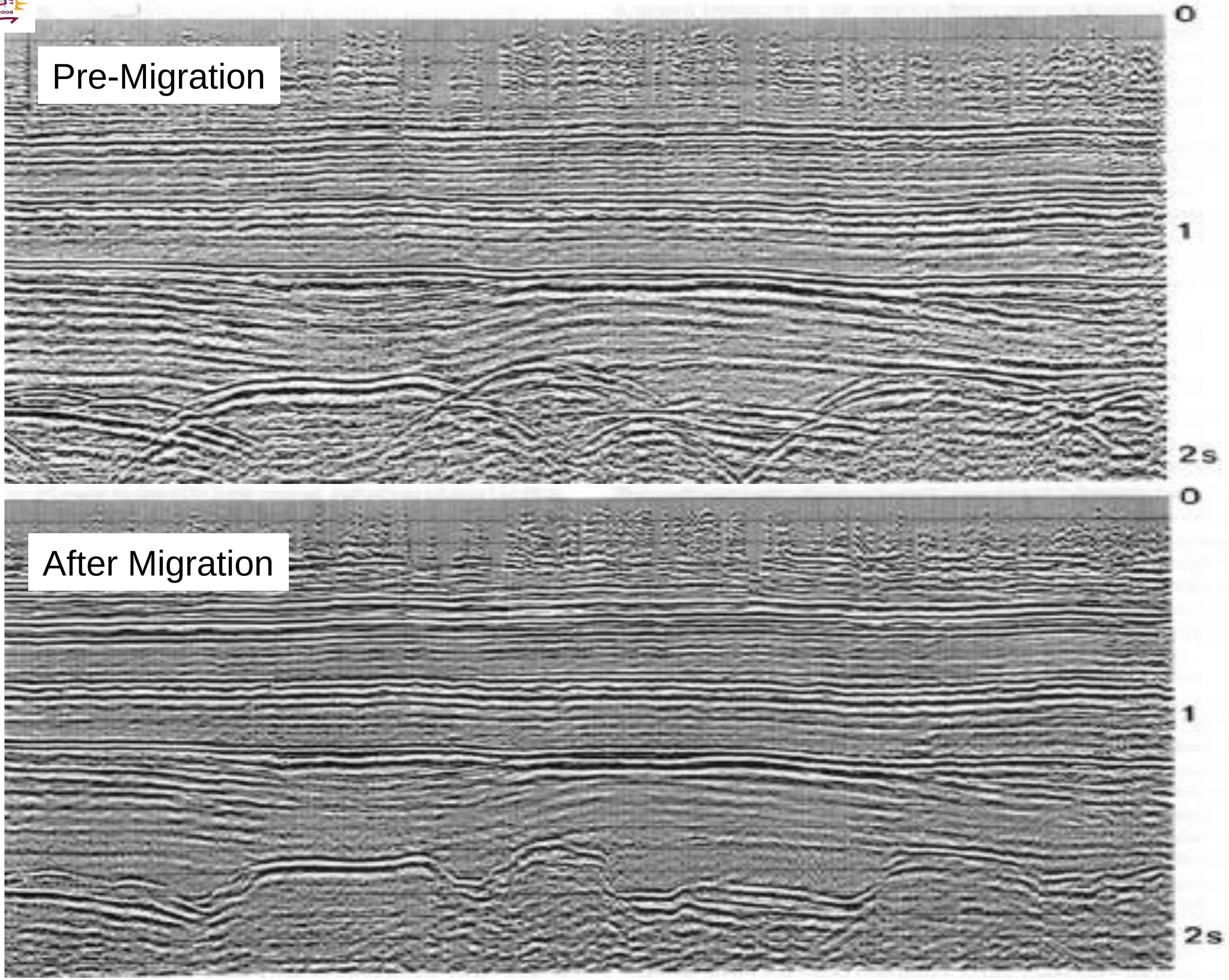

Source & geophone
coincident
CMP1
2
3
4
5
6
A'
A
True dip
B'
B
Appar. dip
C'
C
D
D'
E
E'
F
F'
true reflector
Appar.reflector

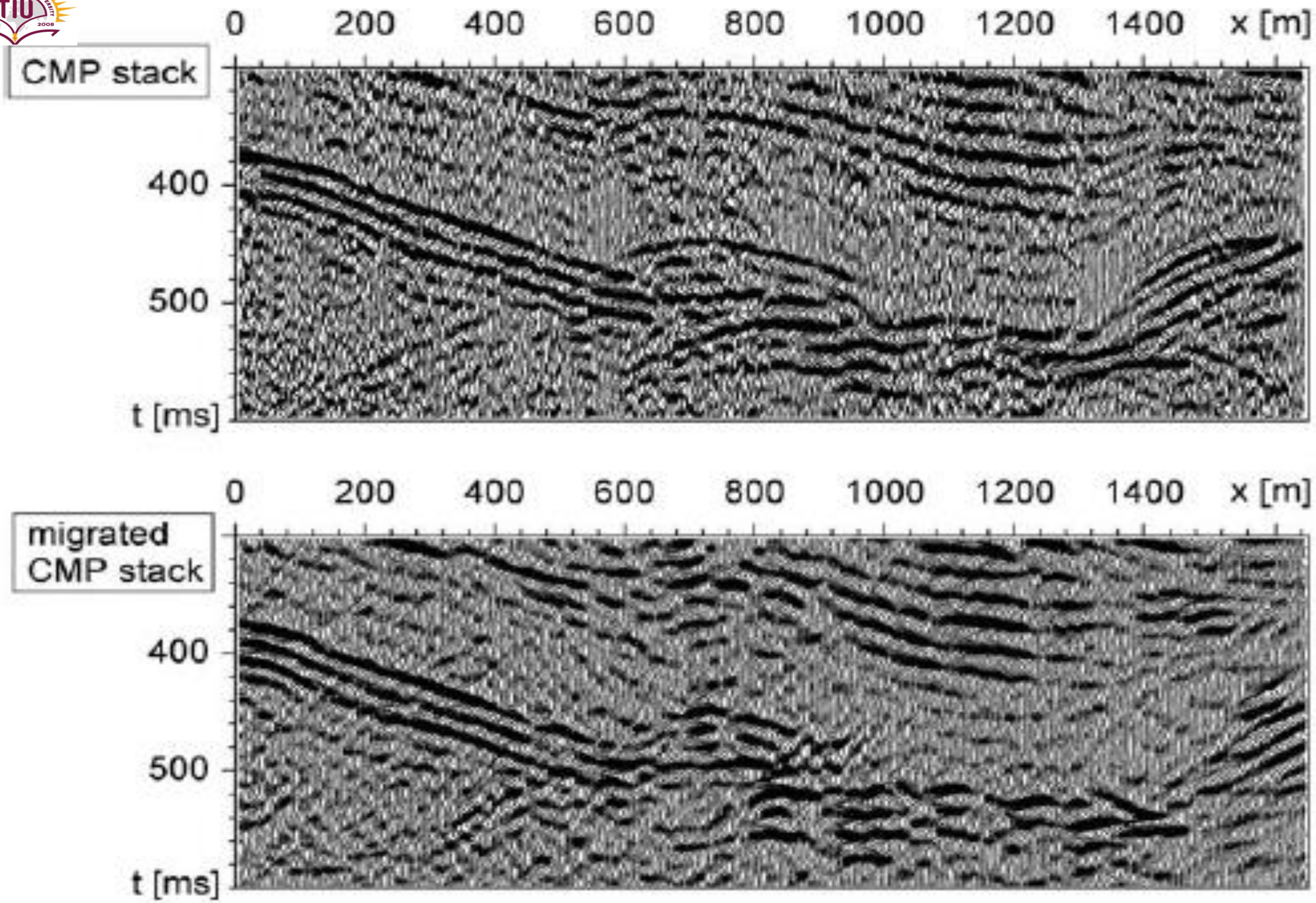




The vertical depth (apparent) is migrated by an angle ϴ' to give the True depth (perpendicular to the reflector).

Two Example (migration and defining bedrocks))





Top: CMP-stacked section of tertiary sediments and cap rock of salt dome with a pseudo depth scale. Bottom: Migrated version of top figure

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.

Principles

The refraction method is dependent upon there being an increase in velocity with depth.

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 Snell's law.

The seismic refraction surveying method uses seismic energy that returns to the surface after travelling through the ground along refracted ray paths. As briefly discussed, the first arrival of seismic energy at a detector offset from a seismic source always represents either a direct ray or a refracted ray.

Applications

- Rock competence for engineering applications
- Depth to Bedrock
- Groundwater exploration
- Correction of lateral, near-surface, variations in seismic reflection surveys
- Crustal structure and tectonics

REFRACTION SEISMICS	REFLECTION SEISMICS
Based on contrasts in : seismic wave speed (c)	Based on contrasts in : seismic wave impedances (ρc)
Material property determined : wave speed only	Material properties determined: wave speed and wave impedance
Only traveltimes used	Traveltimes and amplitudes used
No need to record amplitudes completely : relatively cheap instruments	Must record amplitudes correctly : relatively expensive instruments
Source-receiver distances large compared to investigation depth	Source-receiver distances small compared to investigation depth

Important differences between refraction and refection seismic

Critical Refraction

When seismic velocity increases at an interface (V_2>V_1), and the angle of incidence is increased from zero, the transmitted P wave will eventually emerge at 90°.
•Refracted wave travels along the upper boundary of the lower medium.

Head Waves
The interaction of this wave with the interface produces secondary sources that produce an upgoing wavefront, known as a <u>head wave</u>, by Huygen's principle.

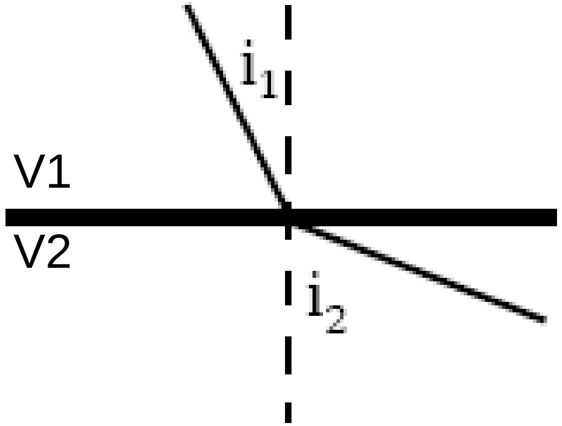
The ray associated with this head wave emerges from the interface at the critical angle.

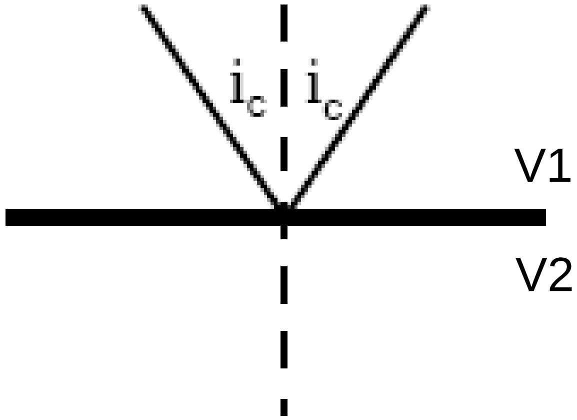
Snell's law again

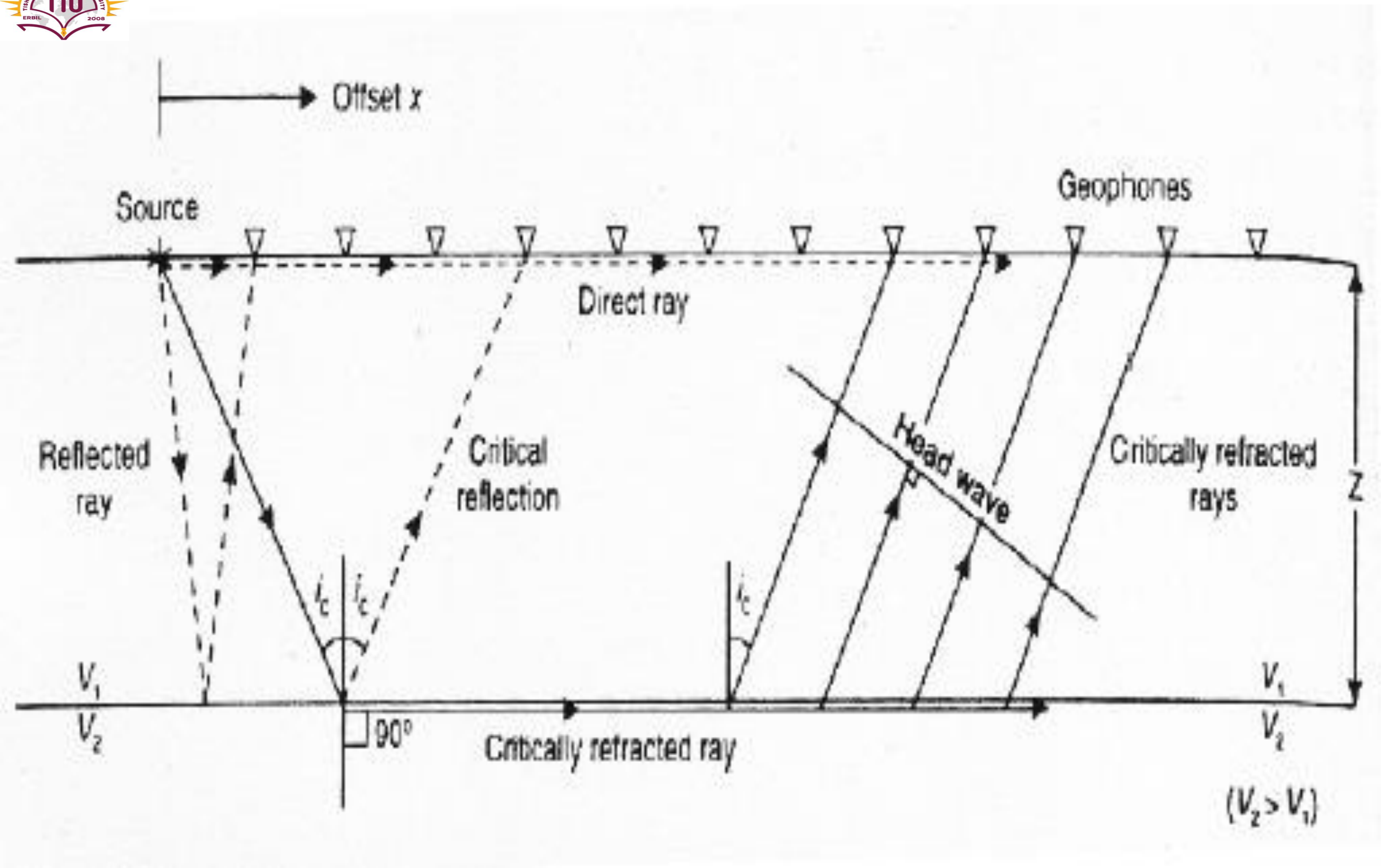
$$\frac{\sin i_2}{\sin i_1} = \frac{v_2}{v_1}$$

If $i_2 = 90°$ ($\sin 90° = 1$)

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







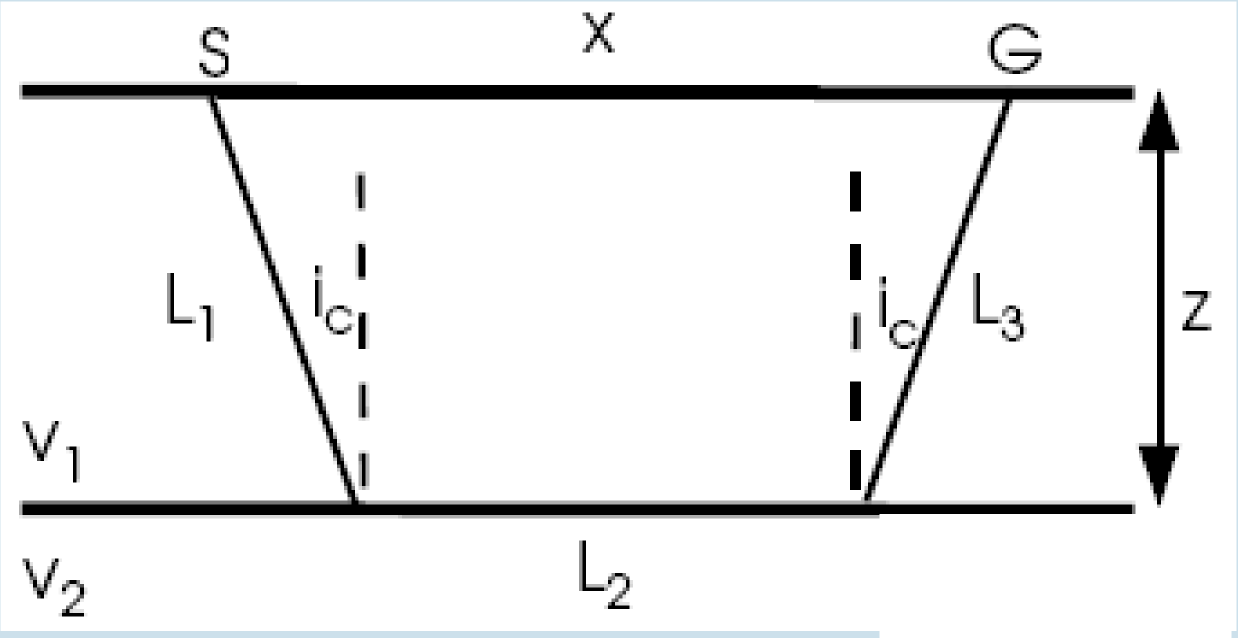
Raypath diagram showing the respective paths for direct, reflected and refracted rays.

Geometry of the refracted ray paths

<u>Two layer case</u>

Travel way of the refracted wave

$$L_R = L_1 + L_2 + L_3$$



Travel time of the refracted wave

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

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

Travel times

Travel time of the direct wave

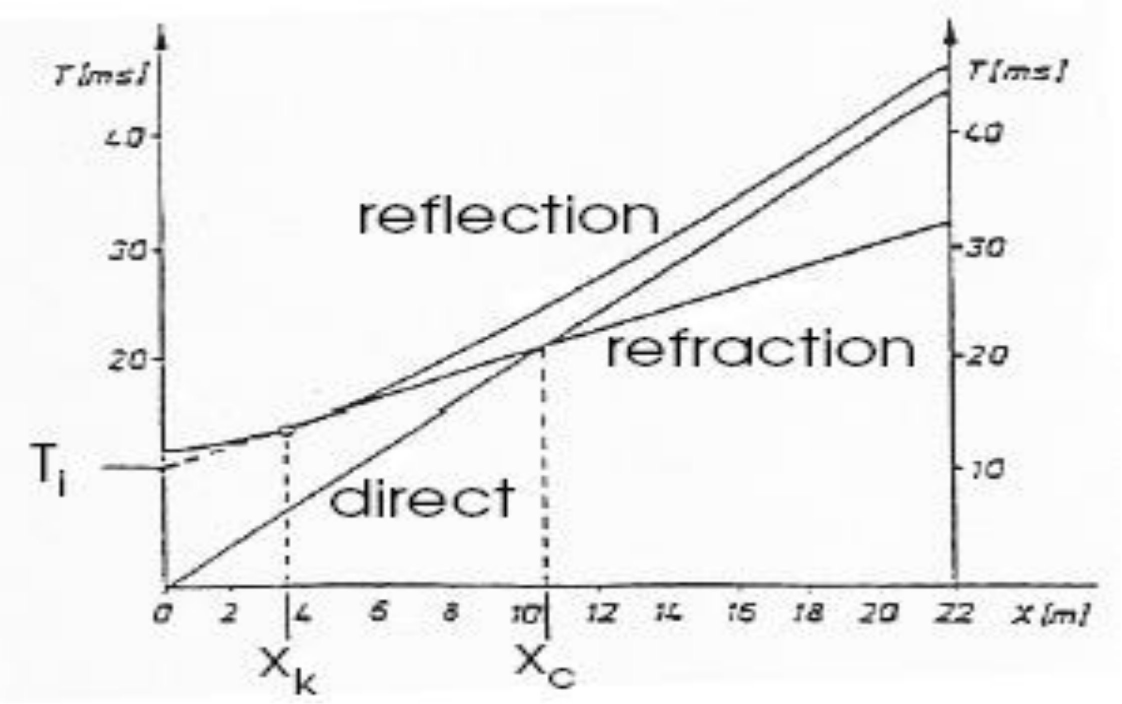
$$T_D = \frac{L_D}{v_1}$$

Travel time of the reflected wave

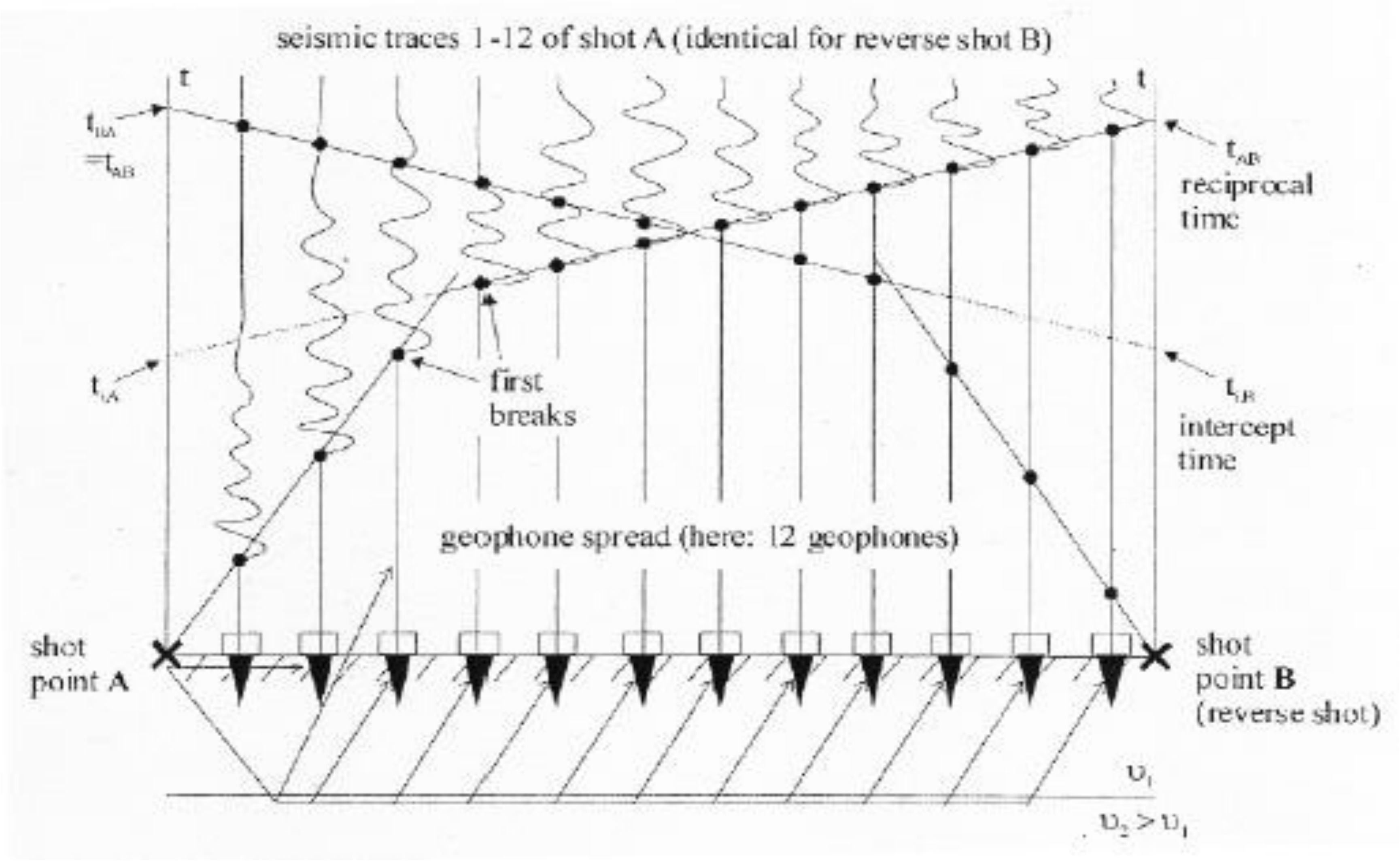
$$T_i = \frac{2\sqrt{z^2 + (x/2)^2}}{v_1}$$

Example: v1 = 500 m/s
v2 = 1000 m/s
z = 3m

- 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.



Traveltime curves for reflected, refracted and direct waves.



Principle of refraction seismic.

Calculation of the depth of the refractor

Two methods

A) Using the intercept times (travel time of the refracted ray with x = 0)

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

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

B) Using the crossover distance x_c

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

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

Multilayer case

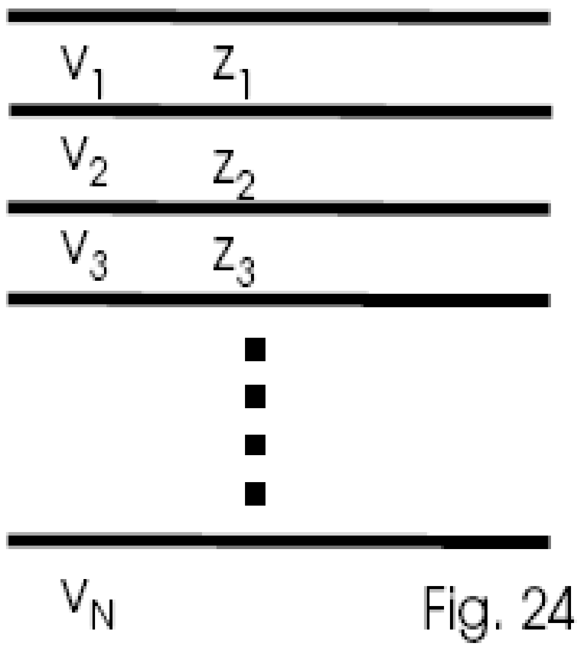


Fig. 24

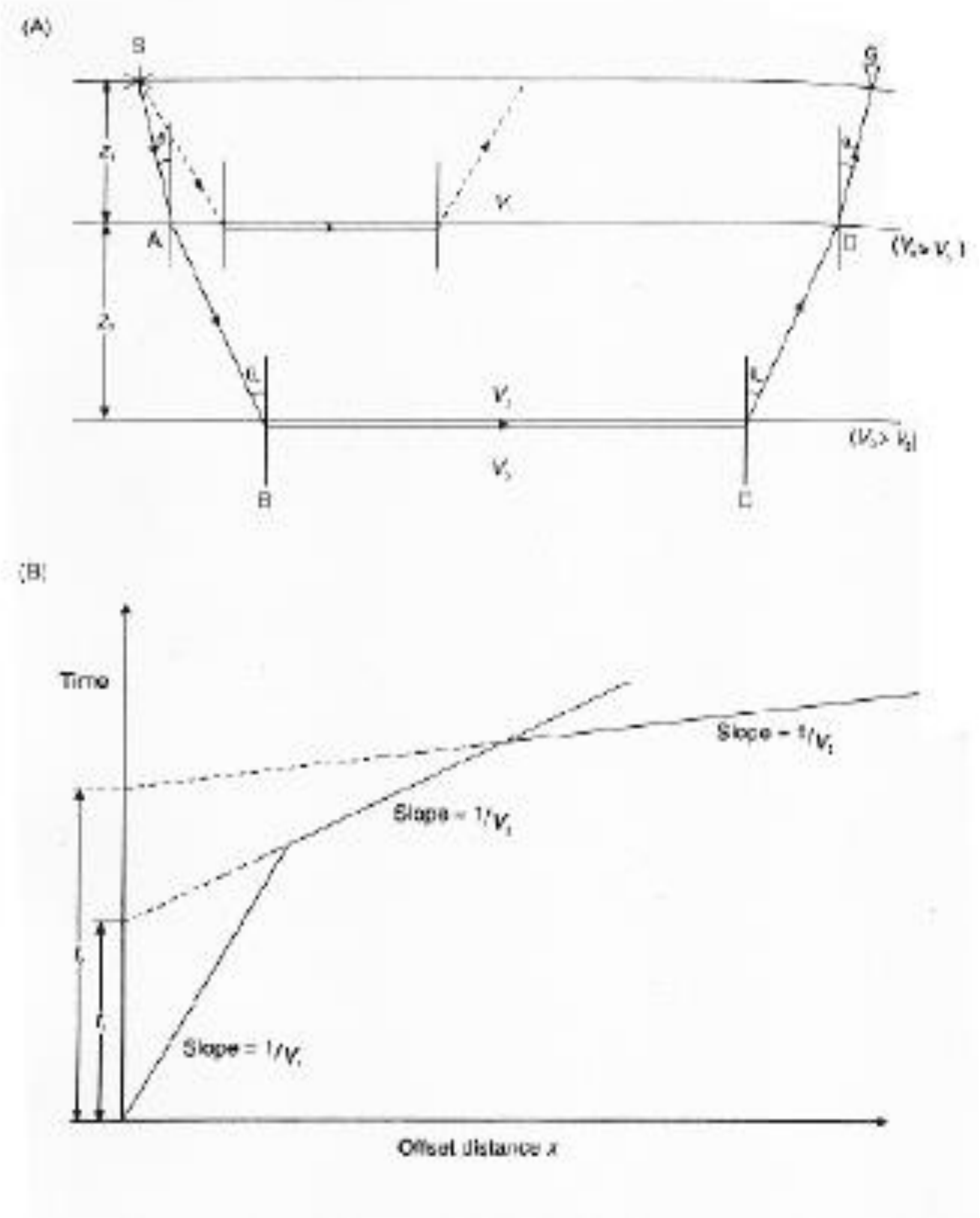
N layers

$V_1 < V_2 < V_3 < \ldots < V_N$

Travel time

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

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



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.

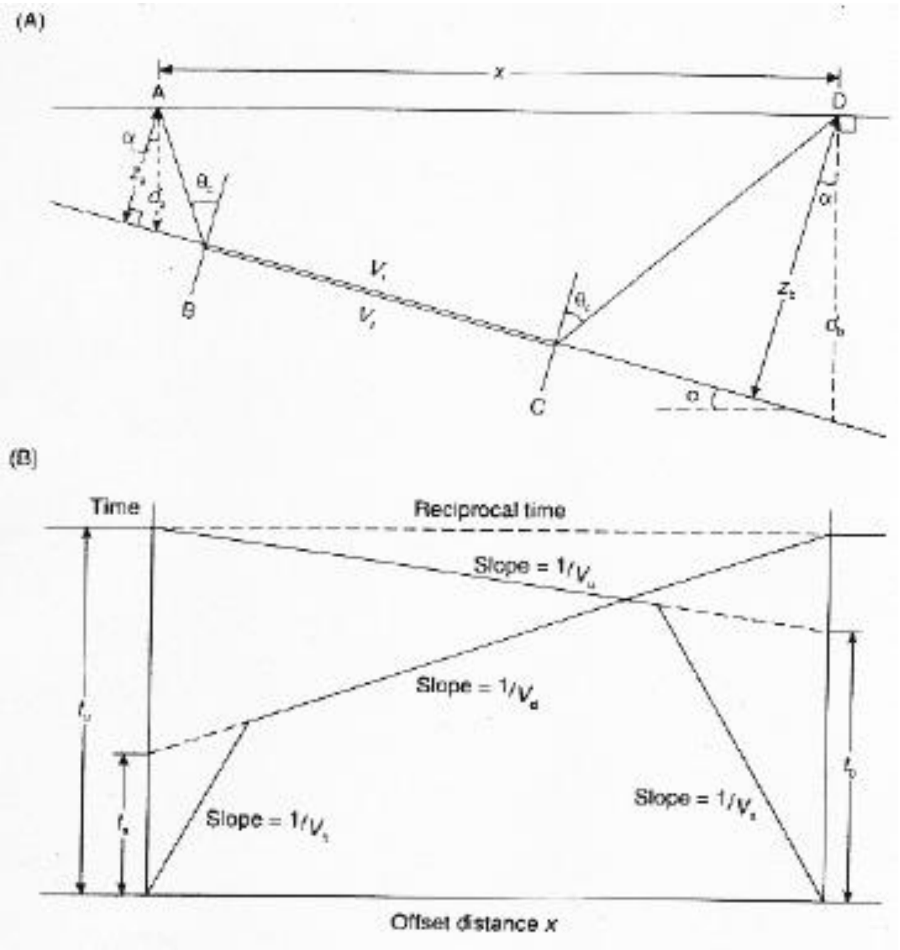
The refractor velocities determined in the case of

dip are referred to as apparent velocities;

v_u: upslope direction

v_d: downslope direction

- Raypath geometry over a refractor dipping at an angle
- and (B) the respective travel time-distance
- graph for the forward (downdip) and reverse
- (updip) shooting
- directions.



Travel time calculations for a dipping refractor

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

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

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

The down-dip travel time t_d is given by:

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

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