เอกสารประกอบการสอน

ว.ธณ. 205482 ธรณีฟิสิกส์

GEOL482 GEOPHYSICS

โดย

อาจารย์ ดร. นิติ มั่นเข็มทอง

ภาควิชาธรณีวิทยา
คณะวิทยาศาสตร์
มหาวิทยาลัยเชียงใหม่

ธันวาคม 2559

คำนำ

เอกสารประกอบการสอนฉบับนี้ ได้จัดทำขึ้นโดยการรวบรวมและดัดแปลงจากภาพนิ่งที่ใช้สอนในห้องบรรยาย เพื่อใช้ประกอบการสอนในกระบวนวิชา ว.ธณ. 205482 ธรณีฟิสิกส์ (GEOL482: Geophysics) ในภาคส่วนของการศึกษาด้านความถ่วง แม่เหล็ก และคลื่นไหวสะเทือน ซึ่งเป็นกระบวนวิชาเลือกสำหรับนักศึกษาสาขาธรณีวิทยา ชั้นปีที่ 4 สาขาวิชาวิชาธรณีวิทยา และเป็นกระบวนวิชาบังคับสำหรับนักศึกษาสาขาฟิสิกส์ โปรแกรมร่วมธรณีวิทยา ชั้นปีที่ 4 หรือชั้นปีที่ 3 ภาควิชาฟิสิกส์และวัสดุศาสตร์ คณะวิทยาศาสตร์ มหาวิทยาลัยเชียงใหม่ ผู้จัดทำหวังเป็นอย่างยิ่งว่า เอกสารประกอบการสอนฉบับนี้ จะเป็นประโยชน์ต่อนักศึกษาเพื่อประกอบการเรียนวิชาธรณีฟิสิกส์ รวมถึงการใช้สำหรับเป็นเอกสารอ้างอ้างอิงสำหรับการวิจัยค้นคว้าอิสระต่อไป

อาจารย์ ดร. นิติ มั่นเข็มทอง
ธันวาคม 2559

ภาควิชาธรณีวิทยา **คณะวิทยาศาสตร์**
ว.ธณ. 482 (205482) ธรณีฟิสิกส์ **4(3-3-6)**

เงื่อนไขที่ต้องผ่านก่อน : ว.ธณ. 324 (205324) ธรณีวิทยาโครงสร้าง

คำอธิบายลักษณะกระบวนวิชา

ธรณีฟิสิกส์และการสำรวจธรณีฟิสิกส์ด้วยวิธีต่างๆ การสำรวจด้านไฟฟ้า: การสำรวจด้านความต้านทานไฟฟ้าจำเพาะ การสำรวจด้านการเหนี่ยวนำโพลาไรซ์ และ การสำรวจด้านศักย์ไฟฟ้าธรรมชาติ การสำรวจด้านแม่เหล็กไฟฟ้า: การสำรวจด้านม่เหล็กไฟฟ้าเหนี่ยวนำและการสำรวจด้านเรดาร์ทะลุพื้นดิน การสำรวจด้านศักย์: การสำรวจด้านความโน้มถ่วงและการสำรวจด้านแม่เหล็ก และการสำรวจด้านคลื่นไหวสะเทือน

วัตถุประสงค์กระบวนวิชา : นักศึกษาสามารถ

1. อธิบายหลักการสำรวจทางธรณีฟิสิกส์
2. วิเคราะห์ข้อมูลทางธรณีฟิสิกส์
3. ประยุกต์ใช้ความรู้ทางธรณีฟิสิกส์เพื่อการสำรวจแหล่งทรัพยากรทางธรณีวิทยาและการศึกษาด้านสิ่งแวดล้อม
4. เก็บข้อมูลภาคสนามด้วยเครื่องมือธรณีฟิสิกส์

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กระบวนวิชานี้ได้ผ่านความเห็นชอบจากที่ประชุมคณะกรรมการบริหารประจำคณะวิทยาศาสตร์ ในคราวประชุมครั้งที่ ..../............... เมื่อวันที่ .... กำหนดเปิดสอนตั้งแต่ภาคการศึกษาที่ 1 ปีการศึกษา 2554 เป็นต้นไป

(ลงนาม) ....
(รองศาสตราจารย์ ดร. สัมพันธ์ สิงหราชวราพันธ์)
คณบดีคณะวิทยาศาสตร์
วันที่........เดือน.........................พ.ศ................

Contents

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Chapter 4: Gravity Method

4.1. Introduction to gravity method

- Gravity method is a gravitational field measurement at a series of different locations over areas of interest.
- The fundamental equation used for mathematical treatment of the data and results is Newton's law of gravitation. Difference of gravitational field can be measured where density contrasts are presented in a geological structure, particularly lateral variations
- Gravity surveying may be conducted on many scales, e.g., small-scale prospecting, regional marine surveys, and global satellite surveys.
- The objective in exploration is to associate variations with differences in the distribution of densities and hence rock and mineral types e.g., chromite bodies have very high densities. Buried paleo-channels may contain gold or uranium.

4.2. Basic concept of gravity theory

Staying Gravity method involves two Newton's laws: Universal Law of Gravitation and the second law of motion.

4.2.1. Universal law of gravitation

"The force of attraction F between two masses, m_1 and m_2, is directly proportional to the product of the masses and inversely to the square of distance between them"

$$F\ \alpha\ \frac{1}{r^2} \quad \text{and} \quad F\ \alpha\ \frac{m_1 \times m_2}{r^2}$$

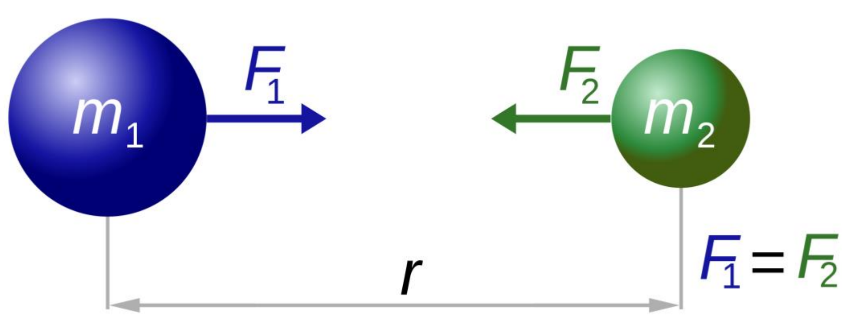


Figure 4.1: *Universal law of gravitation is the relation between force and distance known as "Inverse Square Law" (https://www.muhendisbeyinler.net/newtonun-kutle-cekim-kuvveti-dogru-mu/).*

The relations come up to the equation of force of attraction

- G (big gee) is a gravitational constant equal to the force between two unit masses (1kg.) separated by a distance of 1 m.
- This can be measured in the laboratory and this gravity works everywhere in the universe, not just on the earth.

$$F\ \alpha \frac{m_1 \times m_2}{r^2}$$

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

G value is 6.672 x 10^{-11} $m^3/kg \cdot s^2$ in SI unit

4.2.2. Second law of motion

"The acceleration of an object is dependent upon two variables - the net force acting upon the object and the mass of the object"

$$\mathrm{F} = \mathrm{m_2 g}$$

$$\mathrm{F} = \mathrm{m_2 g}, \mathrm{g} = \frac{\mathrm{F}}{\mathrm{m_2}} = \frac{\frac{\mathrm{Gm_1 m_2}}{\mathrm{r^2}}}{\mathrm{m_2}} = G\frac{m_1}{r^2}$$

$$\mathrm{g} = G\frac{m_1}{r^2}$$

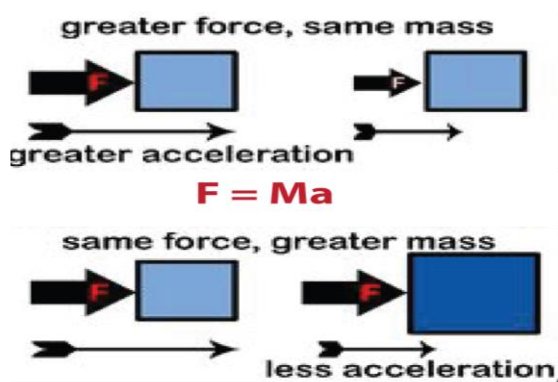


Figure 4.2: *Second law of motion is relation between mass and distance (*http://www.kingnail.info/*newtons-laws-of-motion-poster.html).*

4.2.3. The gravitational acceleration (g)

- The gravitational acceleration (g, little gee) defines the acceleration on an object caused by the Earth.
- When making measurements of the Earth's gravity, we don't measure the gravitational force, we measure the gravitational acceleration.
- We can combine the two laws to obtain the gravitational acceleration of the object (gravity meter), which the gravity meter (m_s) or is a survey station at the Earth surface.

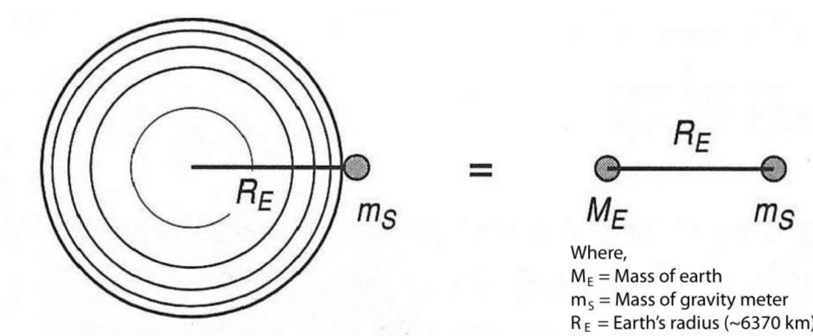


Figure 4.3: *The calculation of the measured gravitational acceleration (g) at the Earth surface (modified from Robinson and Crouch, 1998).*

The g is constant everywhere if the earth is a perfect sphere (radiuses (R_E) are equal), is a homogenous planet (masses (m_E) are equal), and does not rotate. Unfortunately, this never happened on our Earth ever.

4.2.4. Shape of the Earth and gravitational acceleration

The Earth is not a perfect homogeneous sphere, and it rotates.

- The rotation causes the Earth to be an oblate spheroid with an eccentricity 1/298.
- The polar radius is 20 km less than the equatorial radius, which means that g is ~ 0.4% less at equator than at the pole, where g, at the equator, is ~ 5300 mGal.

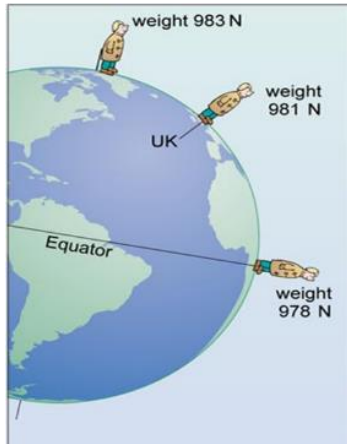


Figure 4.4: *Variation of the Earth's gravitational acceleration due to heterogeneous body and imperfectly spheroidal shape (*http://mrtremblaycambridge.weebly.com/p2-matter-and-forces.html*).*

4.2.5. Units for gravitational acceleration

4.2.5.1.Gravity effect calculation due to buried dense mass (Figure 4.5)

$$\mathrm{g} = G\frac{M_E}{R^2{}_E}$$

$$\delta_{\mathrm{g}} = G\frac{\Delta m}{d^2} = \frac{G}{d^2} = \frac{4}{3}\pi r^3 \cdot \delta_\rho$$

$$\delta_{\mathrm{g}} = \frac{G}{100^2}\frac{4}{3}\pi 50^3 \cdot 0.3 = 1.048 \times 10^{-6} m/s$$

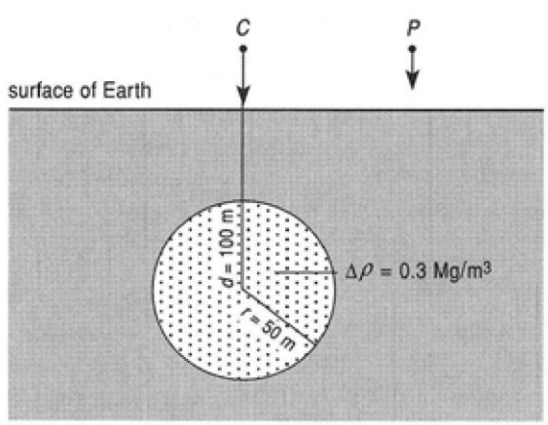


Figure 4.5: *Gravity effect calculation due to buried dense mass (Robinson and Crouch, 1998).*

- The calculation on Figure 4.5 shows gravity variation of 1.048×10^{-6} (m/s^2) due to the buried mass is just 0.000001% of an average of Earth's gravity (9.78 m/s^2) which requires a very sensitive measurement with small units to detect it.
- The SI unit is not a sensitive unit enough and is reasonable to use the mGal unit instead.
- In the example gravity different on Figure 4.5, the buried sphere anomaly defined in mGal unit, which is 0.1 mGal, would be more sensible.

4.2.5.2.The Galileo unit

- The Gal is named after Galileo Galilei when 1 mGal is 0.001 Gal.
- g_E = 9.8 m/s^2 = 980 Gal = 980,000 mGal = 9800 g.u.
- Thus, the Earth's gravitational acceleration is approximately 980 Gals or 980,000 mGal

4.2.5.3.Precision of gravity surveys

- 1 g ~9.8 m/s^2 = 980 Gals
  - 1 milligal is a typical unit on gravity maps.
  - 5 microgals is a precision of the best relative meters
  - 1 microgal is a precision of the best absolute meter (zero drift in principle)
  - 0.01 microgals is a precision of the SG (Superconducting Gravimeter)
- Note that: 1 microgal gravity change is caused by....
  - 2.4 cm layer of water or equivalent.
  - 2.4 millibar air pressure change.
  - 3 millimeter vertical change in elevation at Earth's surface.

4.3. Gravity and geology

4.3.1. Gravitational acceleration and density of material

- If we were to calculate the density of a room filled with people, the density would be given by the average number of people per unit space (cubic meter). The higher the number, the more closely spaced are the people with a typical unit used to describe density of substances are gm/cm^3.
- In relating our room analogy to substances, we can use the point mass described earlier as we did the number of people.

- Consider a simple geologic example of an ore body buried in soil. We would expect the density of the ore body, d_2, to be greater than it of the surrounding soil, d_1. Accepted?
- Density of the material can be thought of as a number that quantifies the number of point masses needed to represent the material per unit volume.
- Thus, to represent a high-density ore body, we need more point masses per unit volume than we would for the lower density soil.
- Now, we describe the gravitational acceleration experienced by a ball as it is dropped from a ladder. This acceleration can be calculated by measuring the time rate of change of the speed of the ball as it falls.
- In acquisition, we measure the gravitational acceleration at the Earth's surface from rocks up to a few kms beneath. In interpretation, we have to understand how subsurface rocks affect the acceleration at the surface.

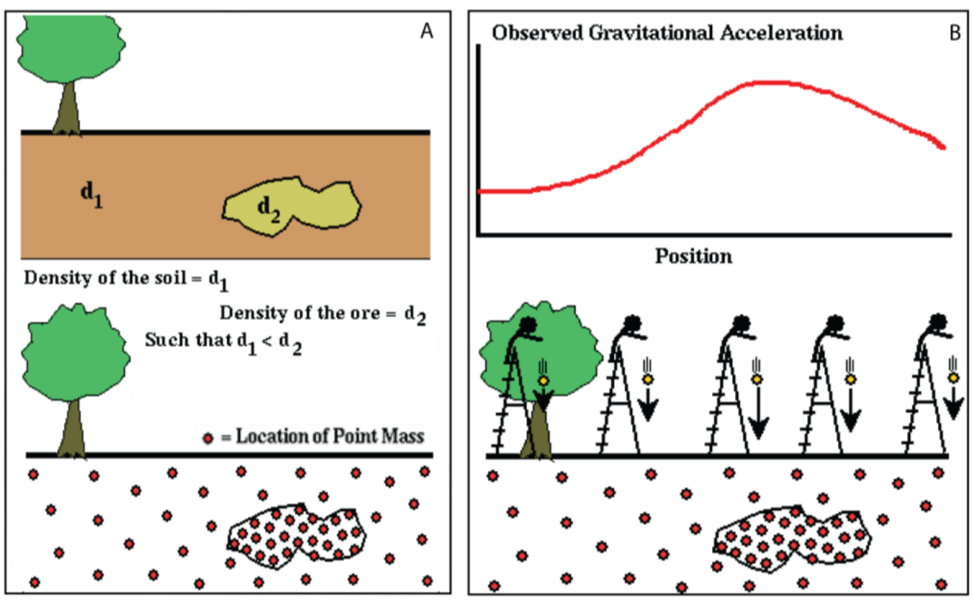


Figure 4.6: *shows how density of rocks defined as mass per unit volume (*http://www.ukm.*my/rahim/Gravity%20method.htm).*

4.3.2. Rock density

Different types of rocks have different densities, and the denser rocks have the greater gravitational attraction. Earth's interior made of denser material; e.g. mostly iron in the core.

	Density (Mg/m^3)
Unconsolidated	
clay	1.5–2.6*
sand, dry	1.4–1.65
sand, saturated	1.9–2.1
Sediments	
chalk	1.9–2.5
coal, anthracite	1.3–1.8
coal, lignite	1.1–1.5
dolomite	2.3–2.9
limestone	2.0–2.7
salt	2.1–2.6
sandstone	2.0–2.6
shale	2.0–2.7
Igneous and metamorphic	
andesite	2.4–2.8
basalt	2.7–3.0
gneiss	2.6–3.0
granite	2.5–2.8
peridotite	2.8–3.2
quarzite	2.6–2.7
slate	2.6–2.8
Minerals and ores	
barite	4.3–4.7
chalcopyrite	4.1–4.3
galena	7.4–7.6
haematite ore	4.9–5.3
magnetite ore	4.9–5.3
pyrite	4.9–5.2
sphalerite	3.5–4.0
Other	
oil	0.6–0.9
water	1.0–1.05

*The ranges of values (taken from a variety of sources) are approximate. Densities depend partly on whether the rock is weathered and the degree of its porosity.

Figure 4.7: *Density variations in g/cm of rocks, minerals, and other materials (Milsom, 2003).*

4.3.3. Earth's density (calculation)

Where, $G = 6.672 \times 10^{-8}\ m^3/Mg{\cdot}s^2$

$g = 9.7803\ m/s^2$

$R_E = 6370$ km

Thus, $M_E \approx 5.97 \times 10^{21}$ Mg or 5.97×10^{24} kg.

Earth's density of 5.5 Mg/m^3 (compared to average density of upper crust which is around 2.67 Mg/m^3. (Lafehr, 1991)

4.3.4. Density contrast

- The density of one rock unit relatives to another, which commonly a surrounding rock (Figure 4.8).
- Density contrasts can be either positive or negative depending on the anomaly properties.

- The density contrast is the most important geologic parameter for interpretation.
- Gravity anomalies are caused by density contrasts within the Earth's sedimentary section, crust and sub-crust can be analyzed and interpreted as lithologic and/or structural anomalies.

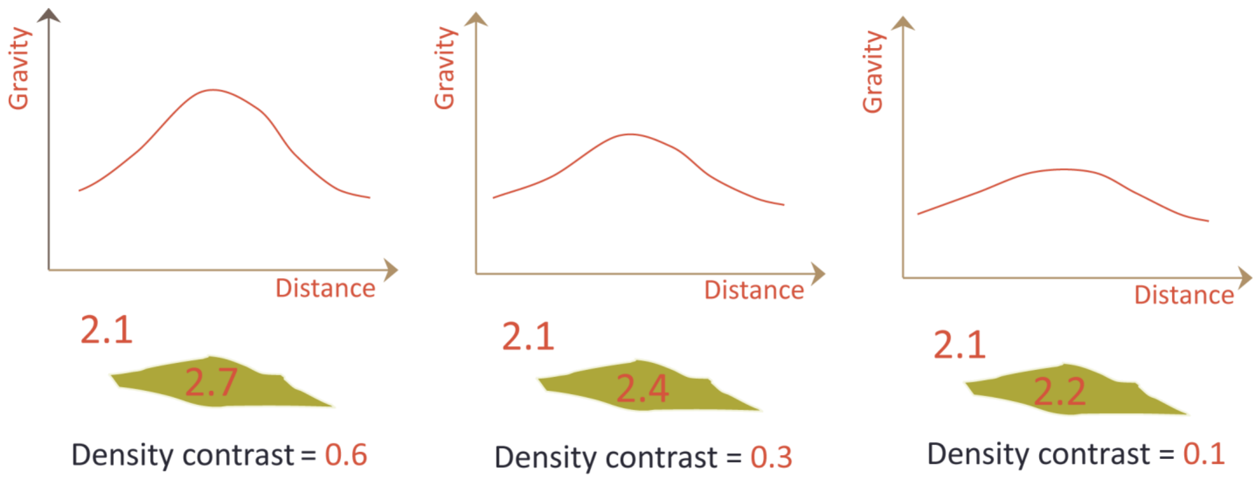


Figure 4.8: *Gravity profiles over the different density contrast between the buried dense and surrounding rocks.*

4.4. Gravity acquisition

- Gravity acquisition measures the changes of subsurface rock density by looking at changes in gravity acceleration. The survey is used for petroleum and mineral prospecting, seismology, geodesy, geophysical surveys and other geophysical research, and for metrology.
- This class will focus on conventional land gravity survey. Special gravity surveys which include, underwater survey, shipborne survey, airborne survey, satellites survey are topics for a graduate class level (212731).
- There are two types of gravity survey:
  - Relative gravity measurement is simpler and is the standard procedure in gravity survey. Thus, field gravity surveying is done using relative gravity measurement.
  - Absolute gravity measurement is difficult and requires complex device and a lengthy period of observation. It would be impossible to get the accuracy required in absolute gravity measurement quickly.

4.4.1. Gravity meter

- Gravity meter or Gravimeter is an instrument used for measuring the local gravitational acceleration (g). It is a type of accelerometer, which varies by about 0.5% over the surface of the Earth.
- Thus, the gravimeters are designed to be much sensitive in order to measure very tiny fractional changes of one mGal. The sensitive gravity meters are enables to be used for both detailed field investigations and large scale regional.
- Note that: even small surrounding gravitational source is able to affect the reading.

Worden Gravimeter

- The Wordon gravimeter is housed in a thermos flask for temperature stability, but it also incorporates a mechanical temperature compensation device. It is evacuated to eliminate errors due to changes in barometric pressure.
  - The meter weighs about 3 kg and the mass weighs 5 mg.
  - An accuracy of gravity reading is less than 0.01 mGal.

LaCoste-Romberg Gravimeter

- LaCoste-Romberg Gravimeter consists in a hinged beam, carrying a mass, supported by a spring attached immediately above the hinge.
  - A reading resolution of 0.01 mGal.
  - Less sensitive to horizontal vibrations.
  - Requires a constant temperature environment.

Scintrex CGx

- Scintrex CGx is a microprocessor based automated gravity meter with self-leveling. The meter uses a mass supported by a sensitive spring and relatively small electrostatic restoring force. The latest model of Scintrex Gravity meter is CG6.
  - A measurement range of 8000 mGal without resetting.
  - A reading resolution of 0.001 mGal.

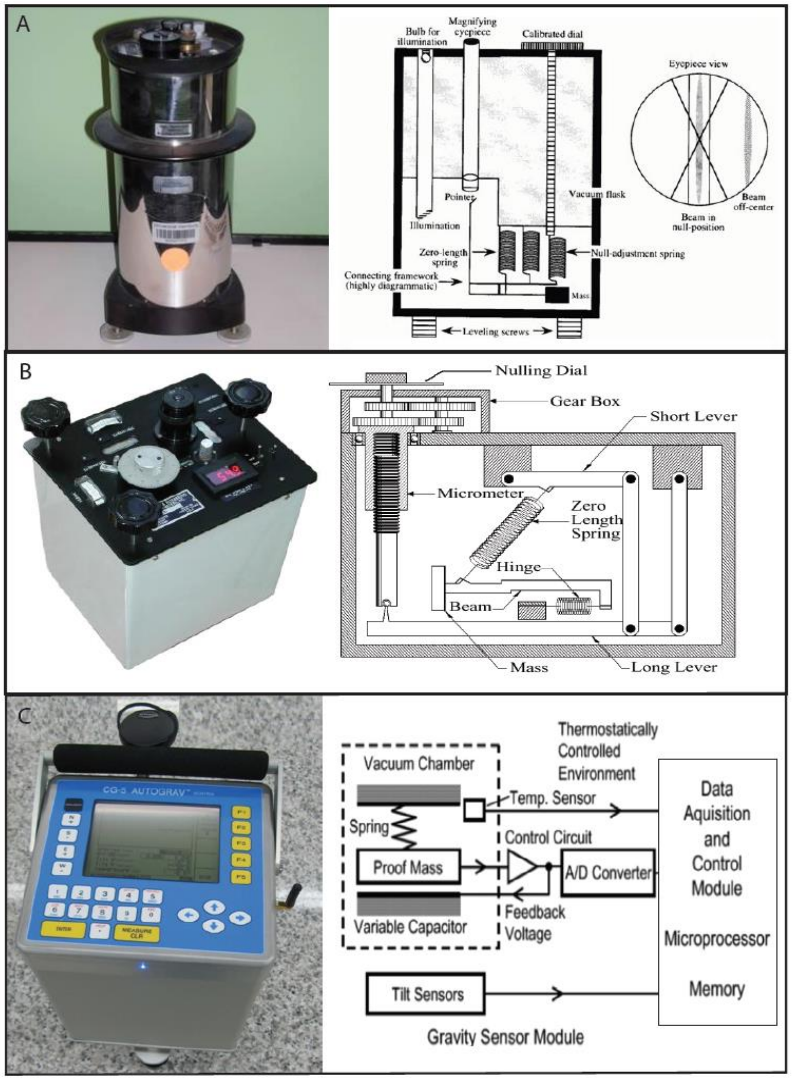


Figure 4.9: *Three different gravity meter models of A) Wordon B) LaCoste-Romberg and C) Scintrex CGx.*

How to take care the gravity meter

- Do not drop or bump the gravity meter.
- Always keep the gravity meter in charge with a constant temperature.
- Always leave the gravity meter alone and in upright position.
- Always keep the gravity meter in a dry condition.
- Have any question, read a tutorial manual first.

4.4.2. Relative gravity measurements

- Relative gravity instruments may be used when either the magnitude of gravitational measures not gravitational force directly, but relative changes in gravity between the known and the point of interest.
- Gravity meter is the measurement of the strength of a gravitational field. Most common relative gravimeters are spring-based (Figure 4.10) with precision of the relative gravity measurements is in the mGal.
- However, the strength of the spring must be calibrated because of stretching of the spring by placing the instrument in a reference point with a known gravitational acceleration.
- Calibration of gravimeters is usually done by the manufacturer. We take a reading at two stations of known g and determine the difference in g per scale division or use a tilt table.

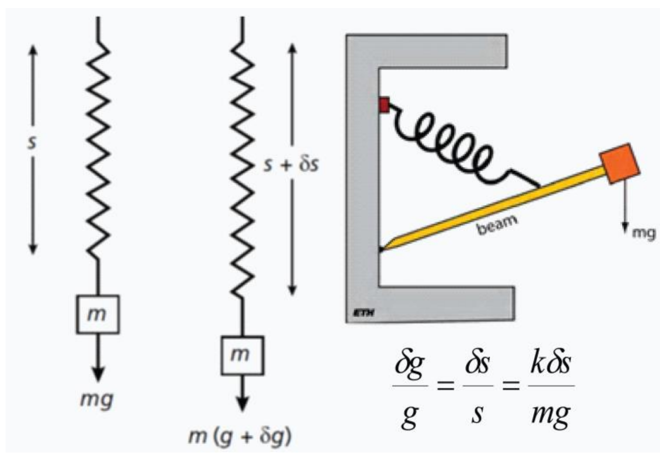


Figure 4.10: *Principle of stable gravimeter operation. Small gravity variation due to the target can be detected by high sensitive measurements (Kearey et al., 2002).*

4.4.3. Relative gravity measurement requirements

- All gravity meters have some drift therefore drift correction is generally needed.
- The gravity meters are sensitive to temperature. All instruments either are within a chamber kept at constant temperature or have a temperature-compensating machine.
- Buoyancy effects from changes in air pressure also affect the reading of the instruments. Therefore the moving system of the meters must be with in a sealed, constant pressure chamber.
- Good gravity meter must be insensitive to magnetic field variation and seismic noise.

4.4.4. Absolute gravity measurements

- In the laboratory, absolute gravity is measured the acceleration of a mass during free fall in a vacuum directly or using the pendulum method.
  - Free falling object (Figure 4.11) is an object that is falling under the sole influence of gravity only. Any object that is being acted upon only by the force of gravity. The free-falling objects do not encounter air resistance nor drift effects.
- In the surveying, absolute gravity may be measured using (relatively) portable sensitive instruments with long period measurement (few days). Accuracy is in few microGal.
  - At a reference gravity station, only one representative reading is used for an absolute gravity value.

Free falling object equation, $\Delta s = V_i \Delta t + \frac{1}{2} a \Delta t^2$

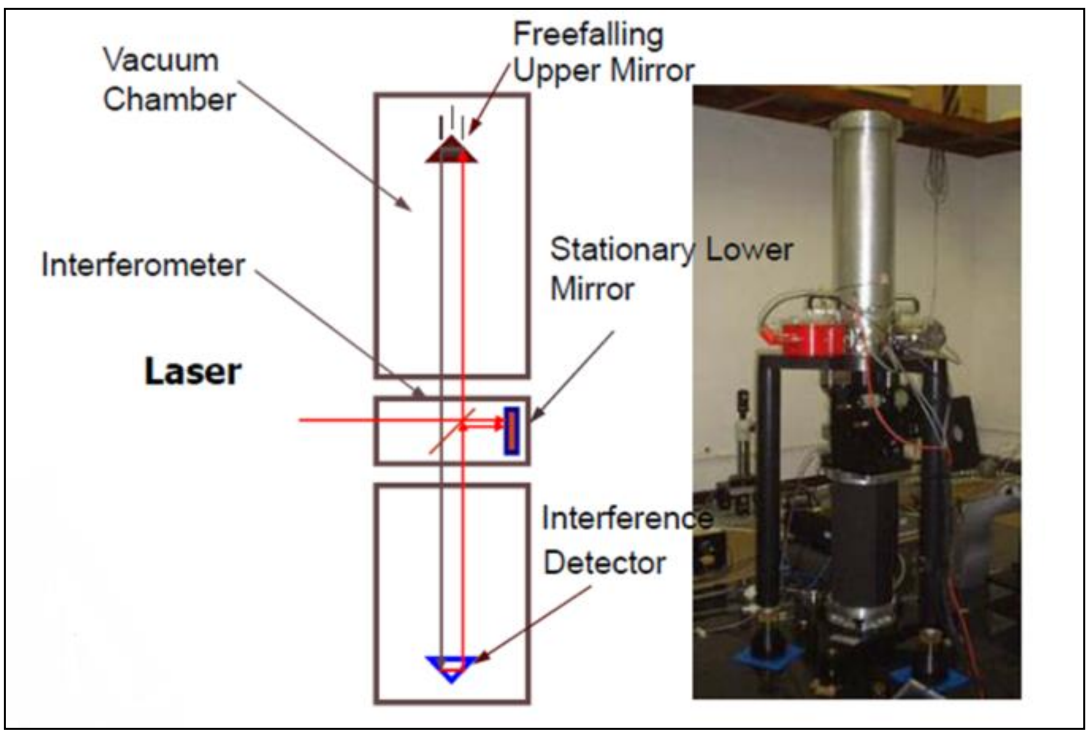


Figure 4.11: *Experiment for free falling object to achieve an absolute gravity values (modified from http://geodaf.mt.asi.it/gravimeter_page.html).*

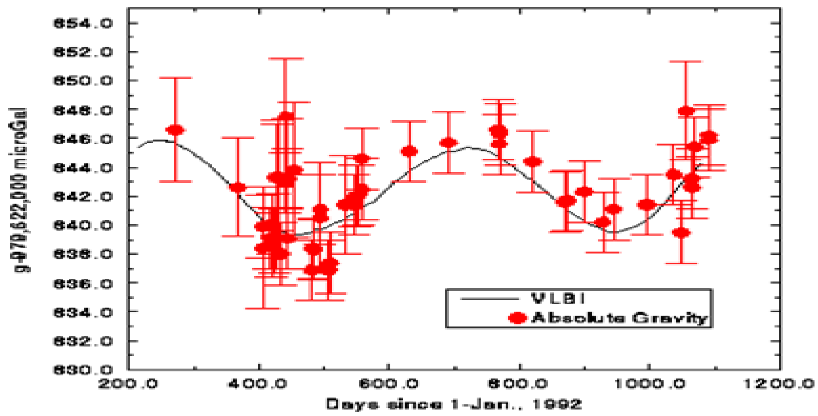


Figure 4.12: *A plot of absolute gravity observation at Table Mountain, Cape Town, South Africa* (http://www.ngs.noaa.gov/GRD/GRAVITY/pm/html).

4.4.5. Establishing reference gravity stations

- Historically, the instruments used to make absolute gravity measurements have had poor accuracies.
- Thus, most instruments used to measure gravity were relative instruments. Relative instruments could survey differences relative to a national base station to establish referenced base stations.
- More reference stations could then be established from these. With each transfer, the measurement error of the transfer increased.
- Internal design elements and the ways they are combined to make a gravimeter cannot guarantee that the published number is a ‘true’ gravity value.

4.4.6. Worldwide absolute gravity Network

- The International Gravity Standardization Net 1971 (IGSN71) remains the official worldwide absolute gravity datum.
- Numerous absolute gravity stations have been established since the 1960s.
- Modern absolute gravimeter measurements loosely confirm the IGSN-71 values, within its error limits.
- About 1,900 worldwide stations, including about 450 U.S. stations, were in this network. Each site had an estimated standard error of less than 0.05 mGals.

Figure 4.13: *Information of absolute gravity base station (Mark No.3/1), Chiang Mai, Thailand established by National Institute of Metrology Thailand.*

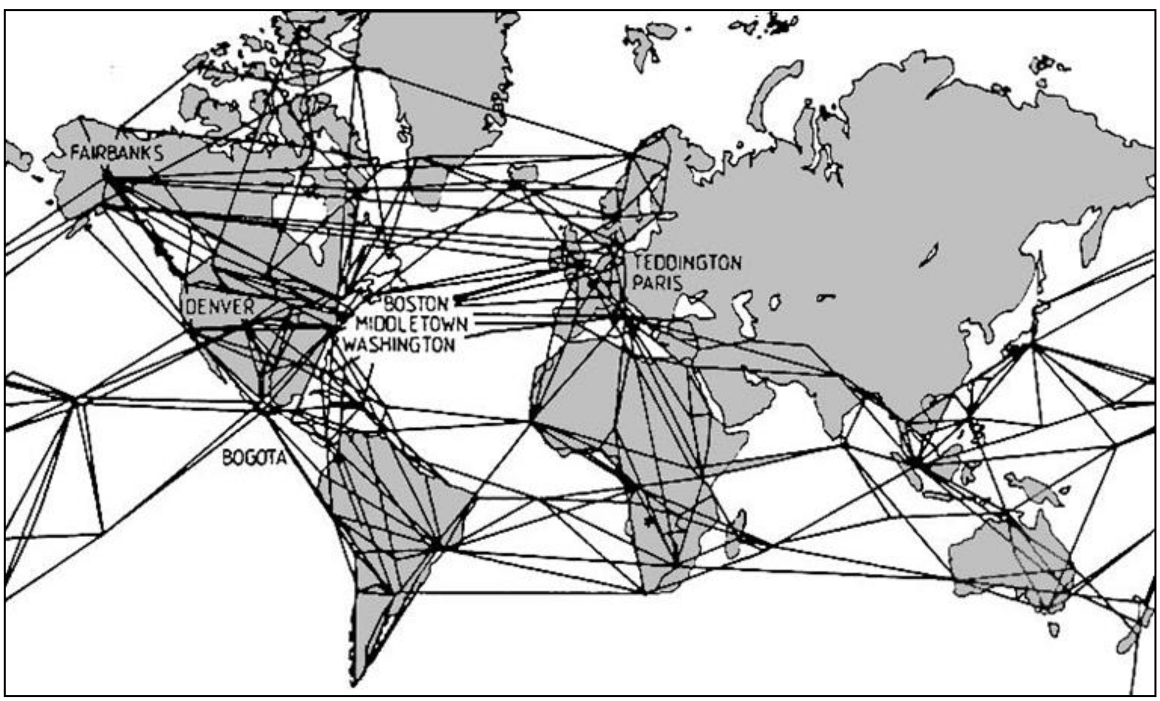


Figure 4.14: *Network of international absolute gravity stations (*http://www.igik.edu.pl/pl/ *geodezja-i-geodynamika-igsn).*

4.4.7. Gravity measurement on land

- The pattern of gravity stations is generally designed to be a rectangular grid.
- In some survey, gravity measurements along existing roads or waterways (in jungle) may be good enough. Sometimes, the quality survey with regular and dense grid is not the best solution for all aspects. It spends much time and budget.
- Both elevation and geographical position must be known accurately.
- All gravity readings require a correction for drift of the gravity meter and earth tide. Therefore, we must make repeated measurements at some control stations several times a day.
- Locations of the referenced stations have to be convenient accessibility.

4.4.7.1.Land gravity survey parameter consideration

- Orientation of 2D profile survey or 3D grid survey
- Regular grid or opportunity grid (along roads) pattern
- Density / spacing of gravity station
- Regional location: Equator or high latitude regions
- Effectiveness of detailed traverses
- The target anomaly in term of size and depth
- Geologic structural control and background of rock density

4.4.7.2.Gravity surveying method

- The following field procedure is usually adopted (Figure 4.15)

    1) Measure a base station,

    2) Measure rover stations,

    3) Premeasure the base station approximately every two – four hours.

- If the survey area is large, time can be saved by establishing a conveniently sited base station to reduce driving time.
- Note that: Collected gravity data from field are raw called gravity reading. The gravity reading is not ready for geological interpretation process.

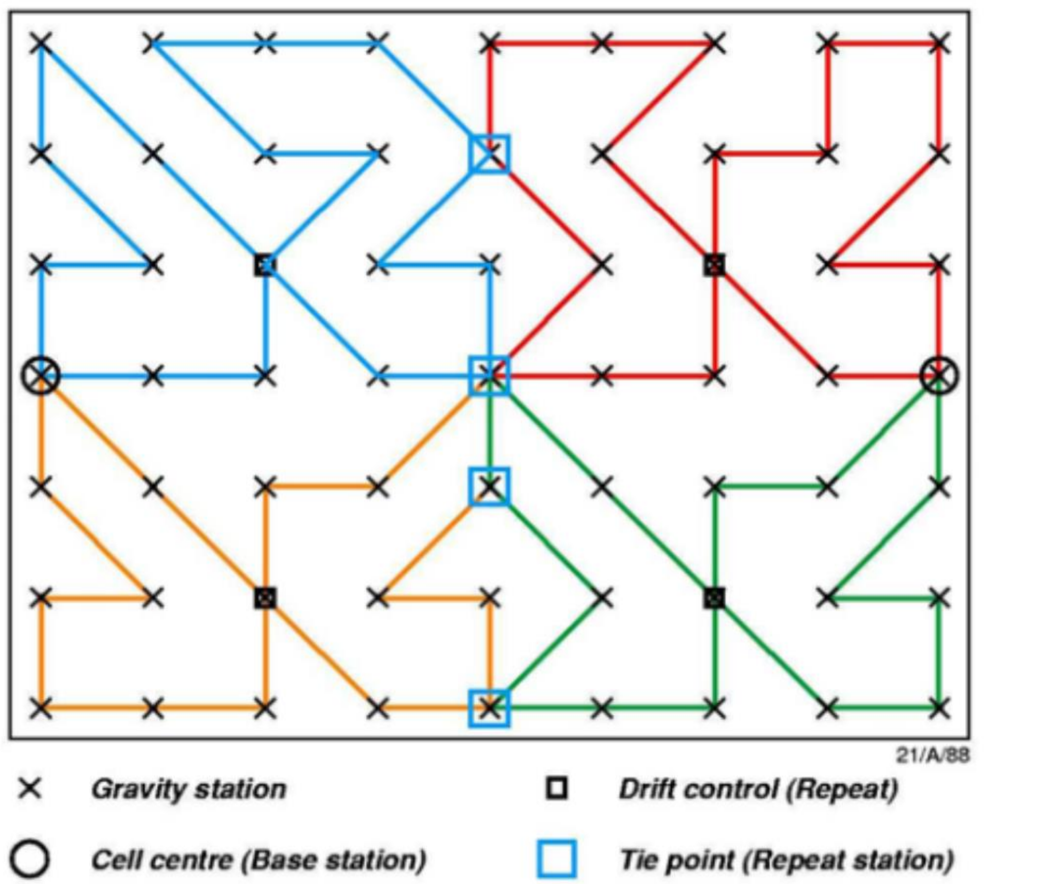


Figure 4.15: *Gravity survey designs. Colored lines represent loops of gravity survey (Murray and Tracey, 2001).*

4.4.7.3.Gravity station selections

Selection of gravity stations must be avoided

- Moving or vibrating locations. (e.g. traffic roads, windy site, or even Earthquake)
- Dense/tunnel trees or under roof structures may block GPS signal.
- Swampy, soft land and loose rocks.
- Power line and radio tower may destruct GPS signal.
- Site near big lake, sea, or river. (Figure 4.16)
- Site near narrow valley, cliff, or on the high slope surface.
- Site has no permission or red zones.
- Site has strong sunlight.

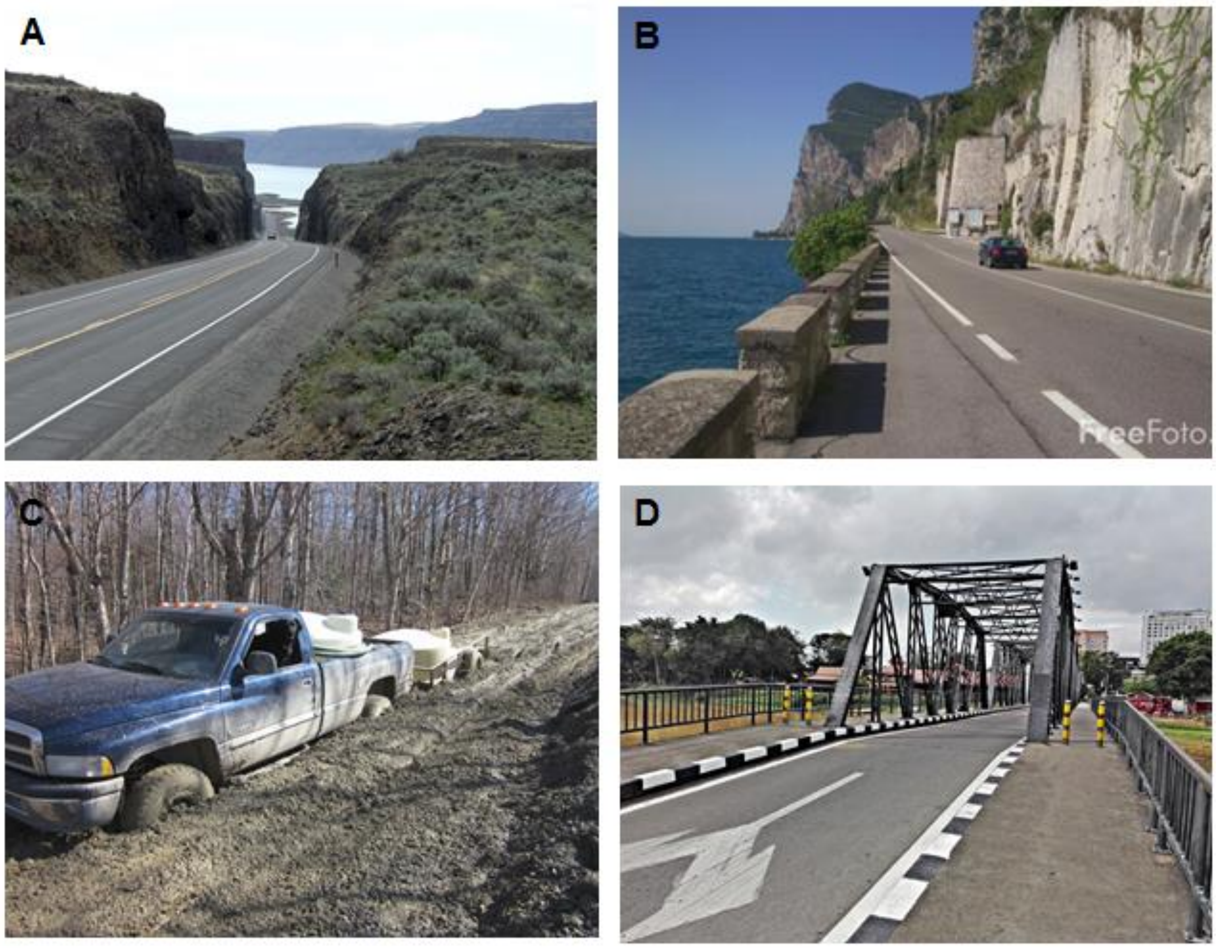


Figure 4.16: *Examples of bad gravity station selections (https://www.wikipedia.org/).*

4.4.7.4.Land gravity survey errors

1) Effect of elevation errors

The exact relation, hence, depends on the reduction density. For a reduction density of 2.67 g/cc, the effect of the elevation error is approximately 0.197 mGal/m.

2) Effect of latitude errors

The typical error of latitude mis-location is approximately 0.8 mGal/km.

3) Relative horizontal position errors

Relative horizontal position error in a strong gravity gradient produces error. The gravity gradient near range-front faults reaches a value of 10 mGal/km. Therefore, horizontal position error of 10 m can produce a gravity error up to 0.1 mGal.

4) Effect of instrument capabilities

Generally, gravimeter drift is around 1.0 mGal per month.

4.4.8. Factor that influence gravity

- Instrument
- Man surveyor
- Earth tide
- Latitude
- Elevation
- Topography
- Curvature
- Density contrast of subsurface rock.

Note that: density contrast is the anomaly key for gravity survey, but sometimes is much smaller than the gravity effects that listed above excluding the density contrast. Geophysicists' duty is to minimize that noised effects and maximize the anomaly caused by the density contrast.

Observed gravity	980658.67	Observed gravity	**980658.67**
Normal gravity	980674.39	Latitude (ϕ)	**45.62**
Free-air correction	30.93	Elevation (m)	**100.24**
Bouguer correction	11.22	Bouguer density (g/cm^3)	**2.67**
Free-air anomaly	15.22		
Bouguer anomaly	4.00		
Elevation error (m)	**0.33**	Latitude error (ϕ)	**0.01**
Bouguer anomaly error	0.06	Bouguer anomaly error	0.90
(All gravity values are in milliGals.)			

Figure 4.17: *Gravity effects that need to be reduced from a gravity reading and gravity anomalies after reduction (Burger et al., 2006).*

4.5. Gravity data reduction and gravity anomaly

- Gravity reduction is necessary to make many corrections for all variations in the Earth.
- Gravitational fields, which do not result from the density contrast in the underlying rocks to the raw gravity readings to obtain the gravity anomalies that are the target of exploration. This process is known as gravity reduction. (LaFehr, 1991)

Standard gravity correction flow

- The instrument scale factor
- Tide and Instrument drift corrections*
- Latitude correction*
- Free air correction*
- Bouguer correction*
- Curvature correction
- Terrain correction
- Isostasy correction

Note that: the correction methods will be done in the lab exercise. The complex anomalies may be calculated with extra corrections using computer software and for various environments.

4.5.1. Instrument scale factor

- Instrument scale factor transforms a dial reading to a gravity measurement in mGal. The instrument scale factor value is different for each gravimeter; and is based on the gravimeter calibration. The formula is:

$$r_c = r \cdot S(r)$$

where; r_c = scale corrected reading in mGal

r = instrument reading in dial units

$S(r)$ = scale factor (dial units/mGal), which may be a function of the reading

4.5.2. Tidal and drift corrections

- Gravitational pull of Sun and Moon varies with time called Tidal gravity. Solid Earth and ocean also move (Figure 4.18A). This leads to daily cyclic variations of ~0.1-0.2 mGal (Figure 4.18B).
- A gravity meter drift is caused by mechanical stresses and strains in the mechanism as the meter is moved or used. This leads to typical variations of ~0.5 mGal per month.
- These effects are not from the shape of the earth.
- The tidal and drift effects vary to times. The long-term tidal and drift can be documented by recording the readings at repeated base stations over a number of weeks or months.

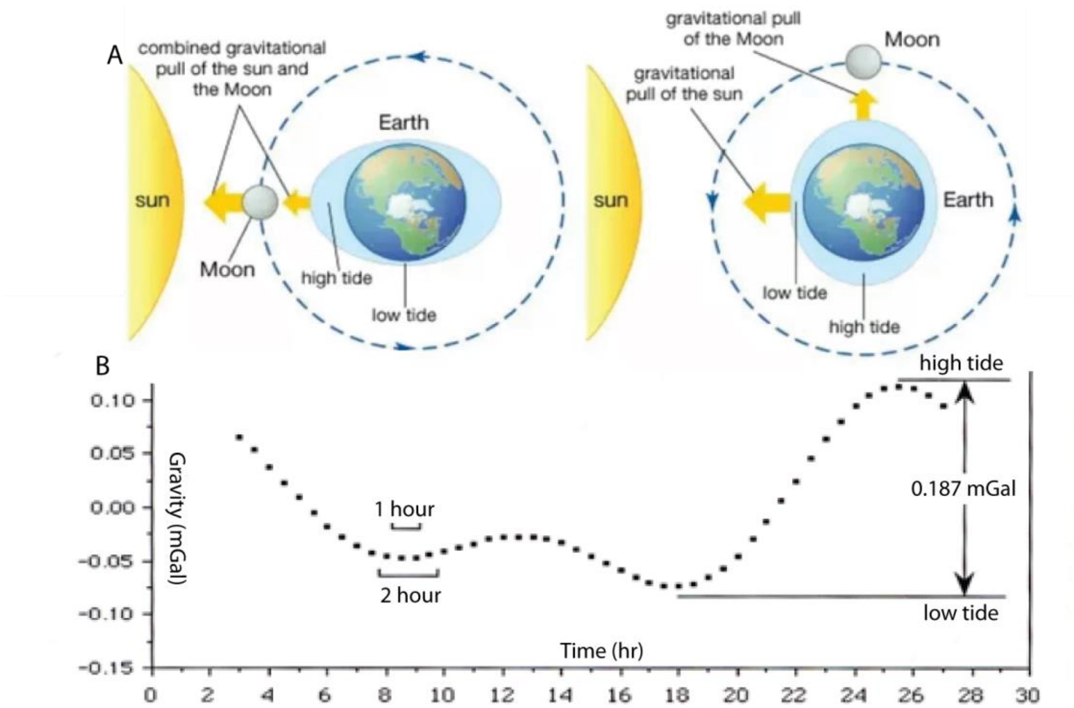


Figure 4.18: *A) Diagram of spring and neap tide and B) graph of daily tides (*https://www.quora.com/*What-is-the-force-that-causes-the-ocean-tides).*

Combination of Tidal and Drift

- Tidal and Drift effects are corrected by setting up loops of survey.
- In practical, re-measurements of gravity at the same position in different time are impossible to be the same shown as a combination effect (Figure 4.19).

4.5.3. Observed gravity values

Observed gravity (g_{obs}) is a representative gravitation acceleration at each station after corrections have been applied for instrument drift and earth tides.

- Relative measurements of gravity collected in a loop. Generally, we collect relative gravity values not absolute gravity values.
- Gravity tie measurements to an absolute gravity value at known absolute gravity station are the practical method to get the observed gravity values at the established base station.
- In theory, the gravity at the same position of base station has to be same. Rate of gravity change at the base is the key to calculate changes of gravity due to instrument drift and Earth tides and get an observed gravity values at any station.

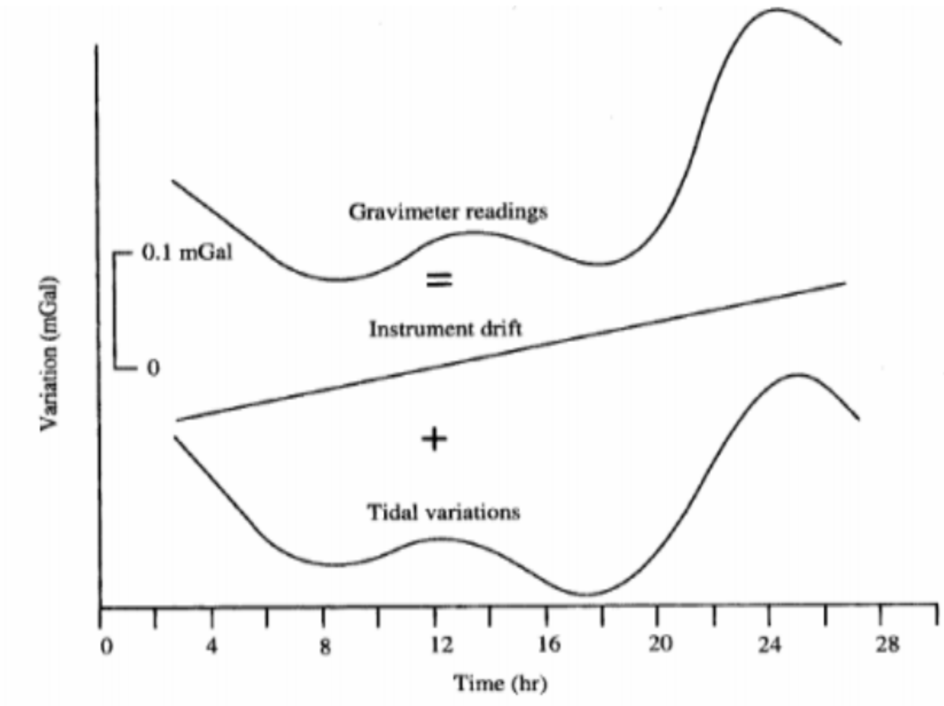


Figure 4.19: *A combination curve (upper) of instrument drift (middle) and Earth tidal (below) variations (Robinson and Crouch, 1998).*

4.5.4. Corrections due to shape of the Earth (Latitude correction)

- Even the earth surface is assumed an ellipsoid without any topographic relief, gravity values still vary because of 2 forces.
- Earth's centrifugal force is a maximum at the Equator.
- The gravitational force at the Equator is less than it is at the Poles.
- On the surface of the ellipsoid, this gravity is defined as the normal gravity.
- Shape of the earth can be described by several surfaces.

Terrain surfaces

- Earth's terrain surface is expressed in terms of the elevation, slope, and orientation of terrain features at the Earth surface (Figure 4.20, red line).
  - Topography describes elevation of landforms above sea level.
  - Bathymetry describes depths of landforms below sea level.

The geoid

- The geoid is a mathematical formula describing a theoretical equipotential surface under the influence of Earth's gravitation and rotation corresponding to mean sea level (Figure 4.20, blue line). All points at mean sea level have the same gravitational acceleration (Figure 4.21).
- The surface of the seas is a shape of the geoid. This surface extends through the continents and corresponds to the level that water would reach in canals connecting the sea.
- The geoid is wrapped due to absence or presence of attracting material. Figure 4.20 shows that the geoid is wrapped up on land and down at sea.

Note that: the concept of the geoid is of fundamental importance to geodetic surveying, or plane surveying, because instruments containing spirit levels measure heights above the geoid, not heights above the reference spheroid.

Reference ellipsoid Earth

- A referenced ellipsoid is a mathematically defined surface of the Earth. It is an ideal idealized Earth model which is a symmetrical homogeneous ellipsoid shape can be calculated using a Normal or Theoretical Gravity (g_N) Formula
- The latest formula is IERS (2003), but the commonly used formula is World Geodetic System (WGS 84).
- Due to ellipsoid Earth model, Normal gravity is ~ 9.780318 m/s^2 at equator and 9.83152 m/s^2 at poles (Figure 4.22).
- Standard gravity is taken as the free fall acceleration of an object at sea level at latitude of 45.5° and is 9.80665 m/s^2 or 980665 mGal.

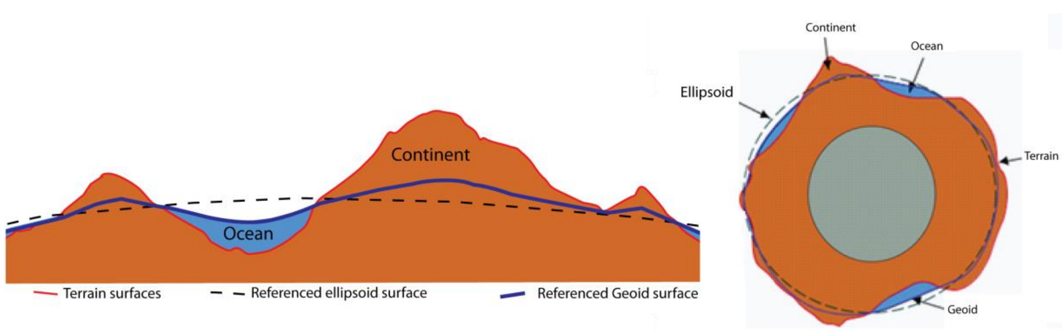


Figure 4.20: *Relationship among the geoid, the spheroid, terrain surfaces, and anomalous mass (Modified fromhttps://www.usgs.gov/media/images/geoid-illustration).*

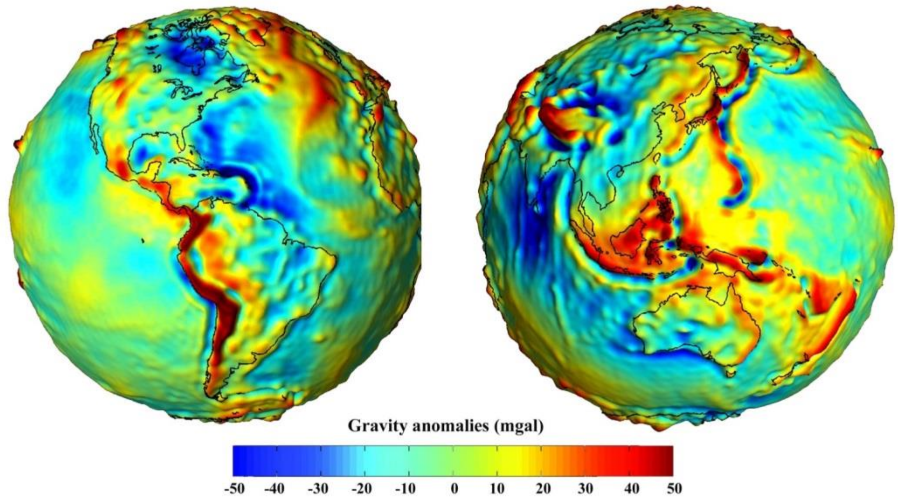


Figure 4.21: *Earth Gravity and Equipotential Surfaces (https://medium.com/planet-os/grace-tellus-monthly-mass-grids-now-available-134ba383bc2e#.l24zr2839).*

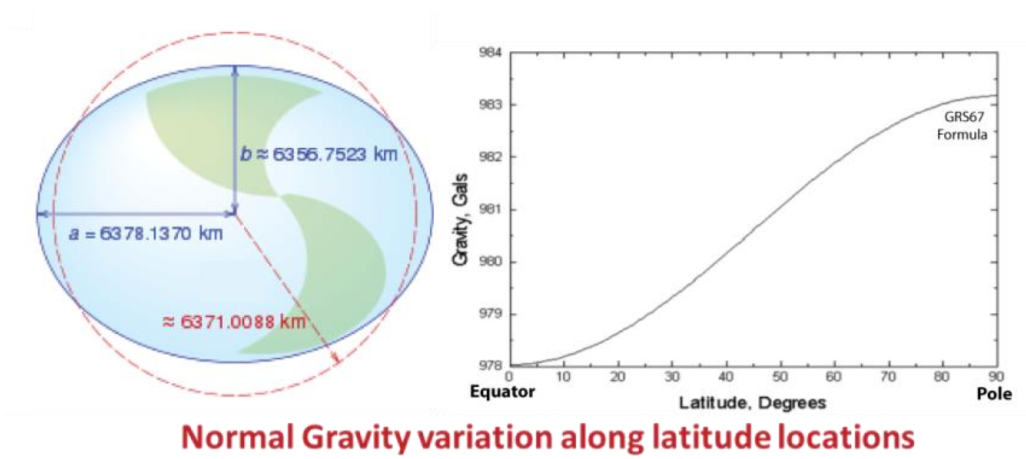


Figure 4.22: *Normal gravity variations along latitude locations (*https://commons.wikimedia.*org/wiki/File:WGS84_mean_Earth_radius.svg).*

Equation of latitude correction

- Latitude Correction is the corrections due to shape of the ellipsoid Earth
- Latitude Correction is the correction subtracted from g_{obs} that accounts for Earth's elliptical shape and rotation.
- Latitude correction equation

$$\Delta g_L = g_{obs} - g_N$$

4.5.5. Gravity anomaly (Δg)

- Gravity anomaly is the difference (Δg or d_g) between the measured gravity which is an observed gravity (g_{obs}) after correcting time variation and an expected value which is an ideal or normal gravity (g_N).

$$\Delta g = g_{obs} - g_N \text{ +/- corrections}$$

$$\Delta g = \Delta g_L \text{ +/- corrections,}$$

- Whenever a measured value departs from an expected value, anomaly exists and this is the anomaly of target.
- Note that in theory, after drift and latitude corrections are applied, the gravity (Δg) at the Geoid or MSL has to be zero.

4.5.6. Free air Corrections

- Free-air correction (FAC) accounts for the difference in elevation between the gravimeter and the referenced level of Geoid (MSL).

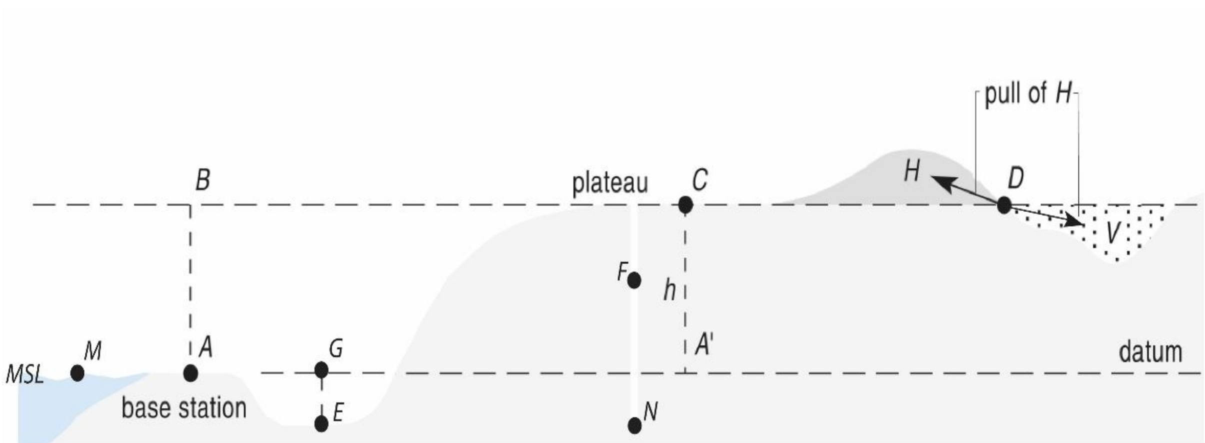


Figure 4.23: *Combination of Free-air, Bouguer slab, and Terrain corrections (modified from Musset and Khan, 2000).*

From free air correction to free air Anomaly

- Free air anomaly is come out after the FA Correction is added when the station is above the datum (MSL) (Figure 4.23 B, C, D).
- If a gravimeter is below sea level (Figure 4.23, E, N), the correction must be subtracted.

In marine survey (on a ship) (Figure 4.23, M) or flying at datum level (Figure 4.23, G), the correction must be zero.

Free air anomaly is the gravity anomaly when FA correction is applied

$$FAC = 0.3086h \text{ mGal}$$

$$\text{Thus,} \quad \Delta g_F = g_{obs} - g_N +/- FAC$$

$$\Delta g_F = g_{obs} - g_N + 0.3086h$$

4.5.7. Bouguer slab correction

From free air Anomaly to Bouguer slab correction

- The meaning of the term "Free-air correction" becomes more apparent in relation to the Bouguer correction.
- Free air correction assumes only air lies between the gravimeter and the referenced datum, but Bouguer correction assumes materials other than air lie between them.
- Bouguer correction accounts for the additional gravitational attraction of the uniform mass between the material of the infinity slab that lies between the gravimeter and the referenced datum (MSL).
- The Bouguer slab is assumed to be an infinite plate of uniform density and thickness.

From Bouguer correction to Simple Bouguer Anomaly

- The Bouguer correction (BC) on land is in the opposite direction of the free-air correction; the added attraction of the extra mass increases the observed gravity. In this case, the Bouguer correction is subtracted. (Figure 4.23, B, C, D, F)
- Simple Bouguer anomaly (Δg_B or called SBA) is the gravity anomaly when FA and Bouguer slab corrections are applied.
- Simple Bouguer Anomaly is the gravity anomaly when Bouguer correction is applied.

$$BC = 2\pi G\rho h = 0.0419\rho h \text{ mGal}$$

$$BC \approx 0.1119h \text{ mGal, when applying } \rho = 2.67 \text{ (crust average)}$$

$$\Delta g_B = (g_{obs} - g_N) + FAC - BC$$

$$\Delta g_B = g_{obs} - g_N + 0.3086h - 0.0419\rho h$$

4.5.8. Terrain correction

- Bouguer gravity anomaly is calculated on the assumption that the gravity station is sitting on a horizontal plane. If the topography differs from a plane this assumption is incorrect and a terrain effect must be applied to compensate.
- The Terrain correction accounts for variations in the observed gravitational acceleration (g) caused by variations in topography near each gravity station.
  - The presence of a mountain (Figure 4.23, H) above the gravity station tends to attract a mass upwards that make the g less.
  - The valley lack of ground (Figure 4.23, V) below the gravity station reduces the downward attraction as well as g value.

From Bouguer correction to Simple Bouguer Anomaly (SBA)

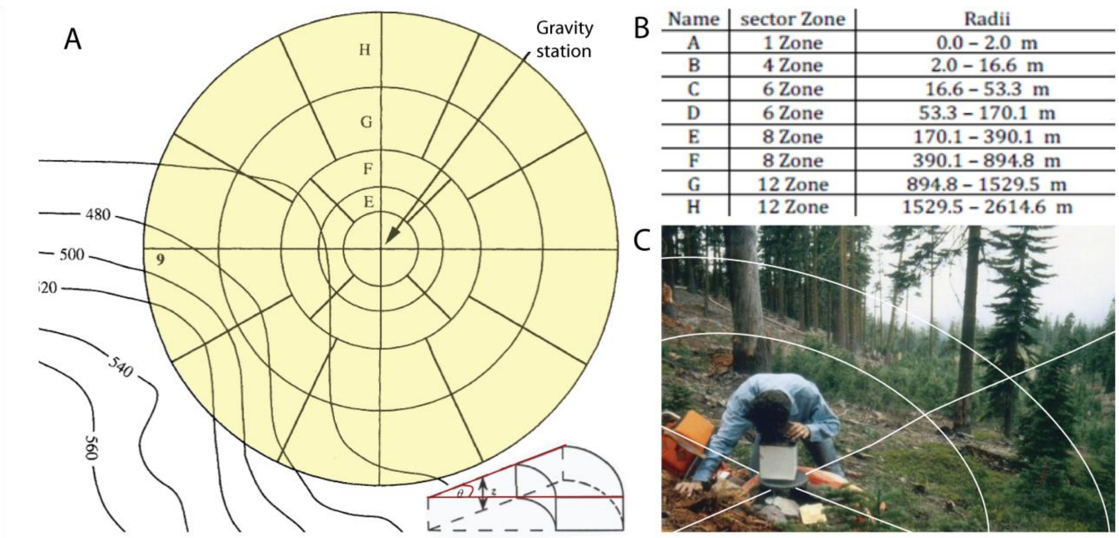
- The terrain correction is positive regardless of whether the local topography consists of a mountain or a valley. The correction is always added in the Bouguer anomaly.
- Gravity anomaly when Bouguer correction is applied is called completed Bouguer Anomaly. (CBA)

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

$$\Delta g_T = g_{obs} - g_N + 0.3086h - 0.0419\rho h + TC$$

Hammer chart method

- Terrain correction can be computed using transparent template, called Hammer chart, which is placed over a topographic map (Figure 4.24A).
- Chart is centered on gravity station and topography read off at center of each sector zone (Figure 4.24B).
- Contribution to terrain correction is obtained from tabulated values for each segment and then summed to obtain total correction. The calculation is based on formula for gravitational attraction of cylindrical segment (Figure 4.24A).
- Note that: No simple calculation for full terrain correction.



Name	sector Zone	Radii
A	1 Zone	0.0 – 2.0 m
B	4 Zone	2.0 – 16.6 m
C	6 Zone	16.6 – 53.3 m
D	6 Zone	53.3 – 170.1 m
E	8 Zone	170.1 – 390.1 m
F	8 Zone	390.1 – 894.8 m
G	12 Zone	894.8 – 1529.5 m
H	12 Zone	1529.5 – 2614.6 m

Figure 4.24: *A) Hammer chart of sector zones overlaying a topographic map. Radii for each sector zones used for Terrain correction computation. C) Gravity measurement at slope surfaces with terrain effects (modified Robinson and Crouch, 1998).*

4.5.9. Curvature correction

- Curvature correction is a step in producing the Bouguer anomaly is to convert the geometry for the Bouguer correction from an infinite slab (Ideal Bouguer slab) to a spherical cap (real lithosphere shape) which the real density is zero (air) not 2.67 g/cc (rocks).
- The radius from the station of 166.735 km is used for computing the curvature effect.

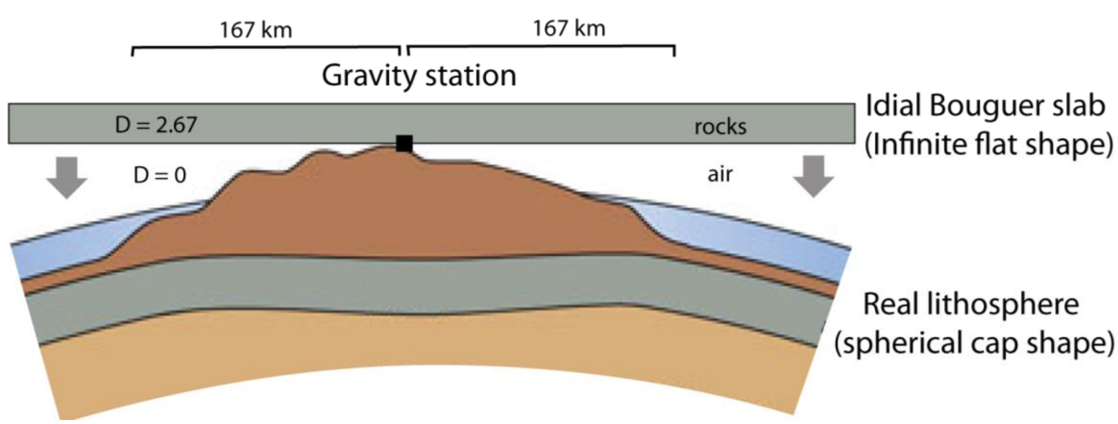
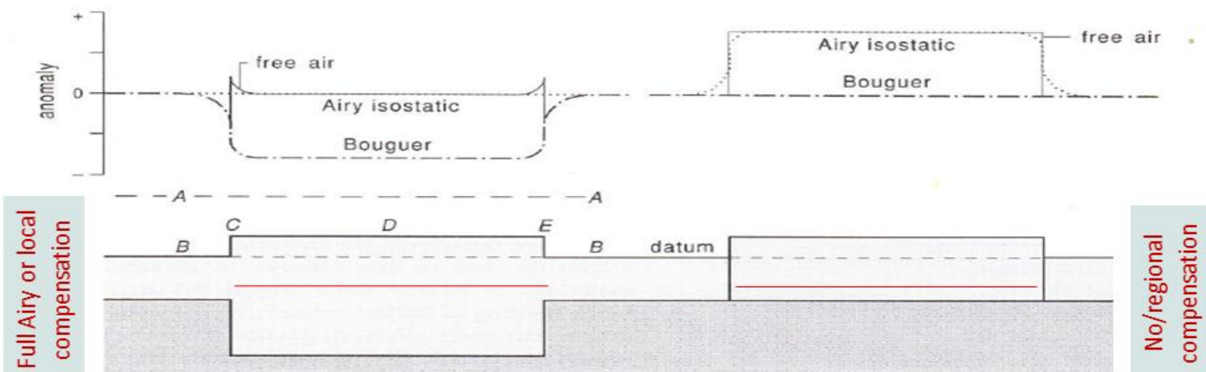


Figure 4.25: *Understanding in curvature effect of the Earth (*modified from Robinson and Crouch, 1998*).*

4.5.10. Isostasy correction

- Isostasy refers to the state of gravitational equilibrium between the lithosphere and asthenosphere.
- Isostasy is the study of how loads (e.g. mountain root) on the Earth’s surface, have to compensated for at depth. The uncompleted compensation stage may affect to gravity measurement.
- Gravity anomalies can be used to test if an area is in isostatic equilibrium, since there the FA should be approximately zero.
- Thus the isostatic anomaly (IA) equation:

$$IA = CBA - \text{predicted effect of the root}$$



Extra mass problem	Free-air	Bouguer	Airy isostatic
Airy comp.	(0) w/ edge effect	(-)	0
No comp. (regional)	(+)	(0)	(+) with edge effect

Figure 4.26: *Gravity anomaly with and without isostatic compensation (*modified from Robinson and Crouch, 1998*).*

4.6. Density determination

4.6.1. Laboratory determination

- Density test from core sample or cuttings from drilling, or sample from outcrops. (based on Archimedes’ principle)

Archimedes’ principle of density determination

Archimedes' principle is that a rock sample totally or partially immersed in a fluid (liquid or gas) is buoyed (lifted) up by a force equal to the weight of the volume of the fluid that is displaced (Figure 4.27).

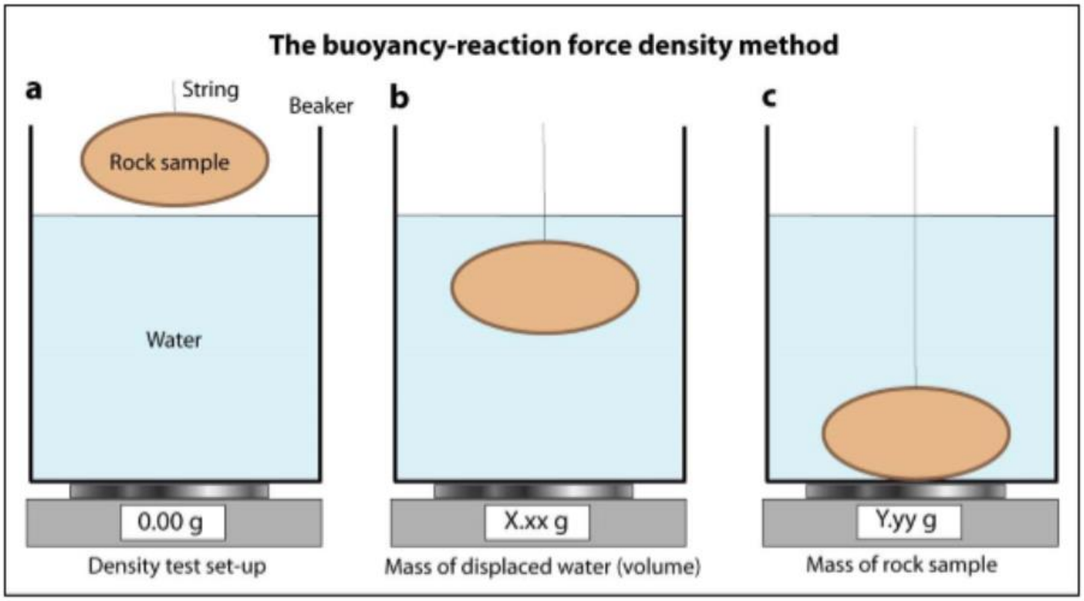


Figure 4.27: *The experiment of rock density estimation. In this experiment, the tension force of the string is not significant.*

4.6.2. Field determination

- Gamma-Gamma density log
- Estimate density from seismic velocity or magnetic anomaly
- Measuring value of “g” in bore-hole
- Measuring value of “g” at the ground surface
- Parasnis’ method
- Nettleton’s method

Nettleton’s method of density determination

- An indirect means of density determination in which a closely spaced gravity traverse is run over some topographic feature e.g. hill or valley of uniform surface lithologic unit.
- Nettleton’s method (Nettleton, 1939 and Papp, 2009) is based on the observation that over an area of corrected constant density no terrain correlated, gravity anomalies should remain after applying the Bouguer correction (Figure 4.28, red line).

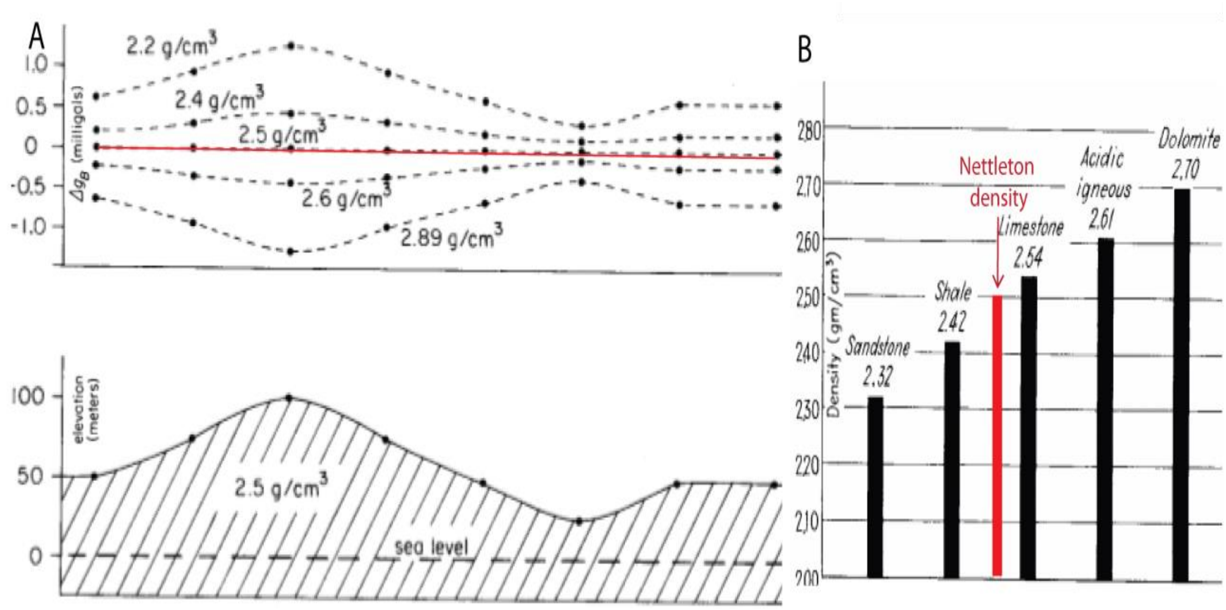


Figure 4.28: *A) Nettleton's method is based on the Bouguer gravity with a constant density should remain after applying the Bouguer correction (red line). B) Rock type is determined from calculated Nettleton's method* density (modified from Nettleton, 1939).

4.7. Gravity data interpretation

4.7.1. There are two approaches to the interpretation of an anomaly

- Qualitative interpretation is a visual inspection based on the plan shape, the general profile, other evidence (records, air photos, excavations etc.) geological models.
- Quantitative interpretation is a numerical modelling based on a presumed structure, which is adjusted until the computed anomaly matches the observed one.

4.7.2. There are two displays of an anomaly

- As profiles (line survey) that has some characteristic which is easily interpreted, but less accurate.
- As map (grid survey) that has some characteristic which need skillness and time to interpret, but more accurate.

4.7.3. The general rules of gravity interpretation

- Higher than average density bodies will cause a positive gravity anomaly with the amplitude being in proportion to the density excess.
- Lower than average density bodies will cause a negative gravity anomaly.
- The areal extent of the anomaly will reflect the dimensions of the body causing it.
- A sharp high frequency anomaly will generally indicate a shallow body.
- A broad low frequency anomaly will generally indicate a deep body.
- The edges of a body will tend to lie under inflection points on the gravity profile.
- The depth of a body can be estimated by half the width of the straight slope.

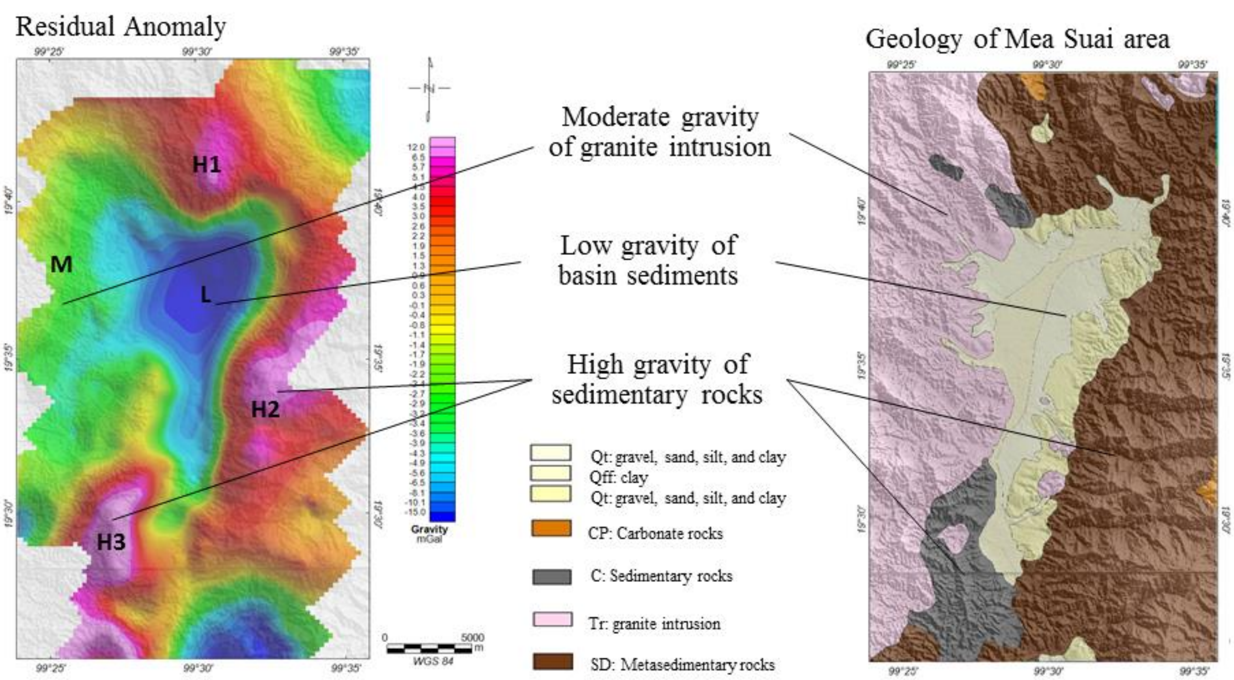


Figure 4.29: *Lithologic interpretation on the residual CBA map (Kanthiya, 2017) compared to geologic information (DMR, 2007) of Mae Suai basin, Chiang Rai Province (Kanthiya et al., 2017).*

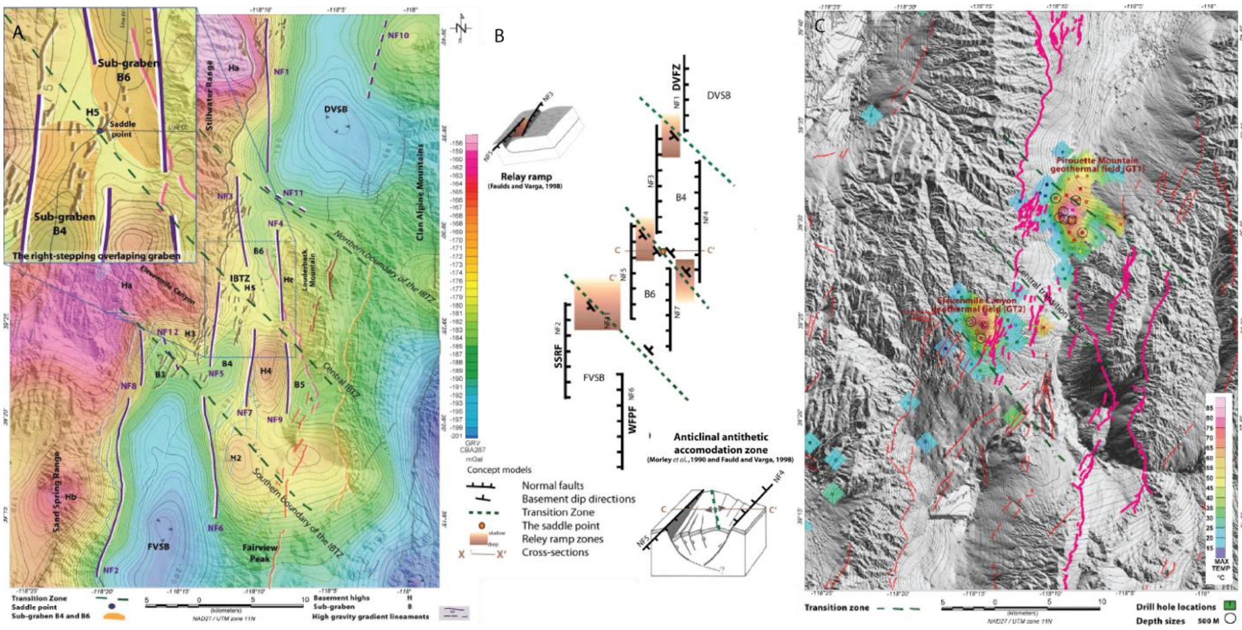


__Figure 4.30:__ (A) Fault interpretations on CBA map in Dixie-Fairview Valley, NV, USA. (B) Terminations in these fault patterns define the Inter-Basin Transition zone (IBTZ) that the linkage structure appears to be consistent with the anticlinal antithetic accommodation model. (C) Consistence between IBTZ and geothermal locations (modified from Mankhemthong et al., 2008).

4.7.4. The edge detection techniques for interpretation

- The structural boundaries can be detected and interpreted using several edge detection techniques for gravitational field.
    - The horizontal derivative (THDR) method produces maximum ridge of anomaly gradients over the contacts between different densities among subsurface rock units that are caused by features such as stratigraphic or structural contacts which juxtapose units of different densities (Figure 4.32).
    - The Euler deconvolution technique of 3 orthogonal gradients along x, y, and z axes to integrally estimate subsurface locations in both horizontal and vertical (depth) dimensions (Figure 4.32).
    - The 2nd horizontal derivative method in the appropriated gradient directions (Whitehead & Musselman, 2006) perpendicular to fault strikes show a zero trending of possible fault lineaments (Figure 4.32).

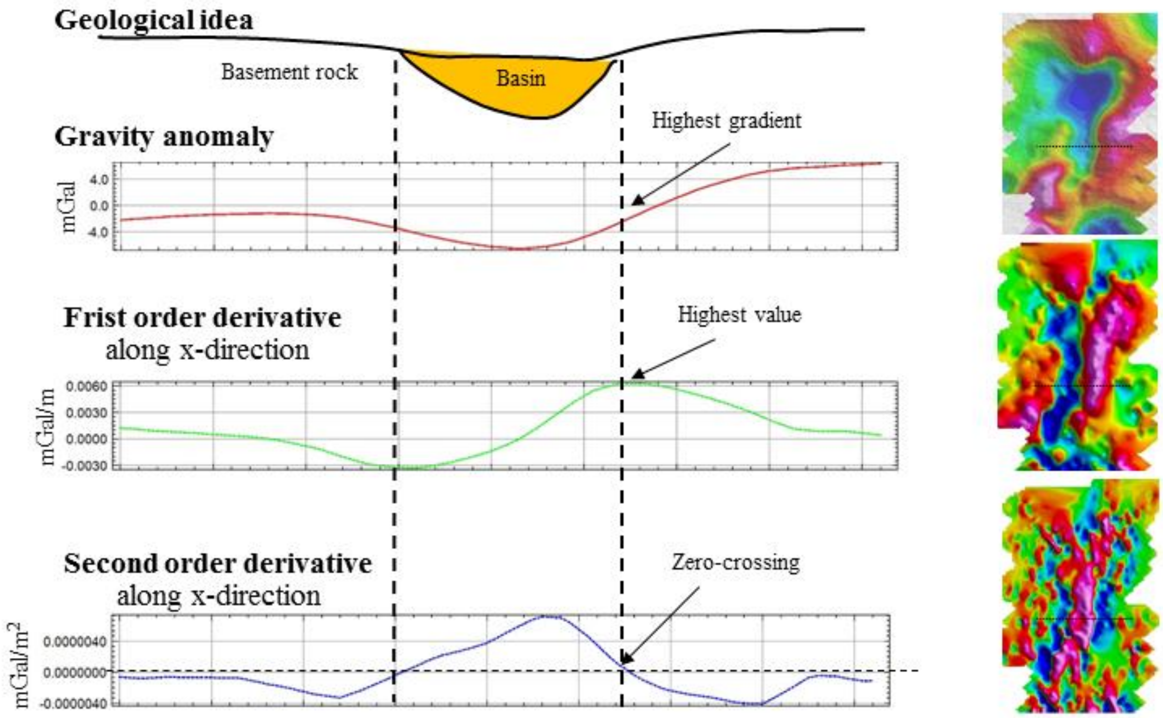


Figure 4.31: *Concept of the first and second order horizontal derivatives of gravity data compared to a geological model (modified from Kanthiya, 2017).*

4.7.5. Regional and residuals

- Gravity anomaly is a gravity value resulted from target. Anomalies from shallow target are important to mineral exploration, and deep anomalies are important for Earth crustal study. One survey's signal is another's noise.
- The regional anomalies are caused by large and deep structures, often larger than our survey area that usually represents the long-wavelength high-amplitude anomalies (Figure 4.33B). Sometimes they are also referred to as a trend.
- These must be removed to enable local residual anomalies to be interpreted. Numerous techniques are used to remove the trend anomaly.
- Thus, the residual anomalies are, in an ideal world, anomalies caused only by our target (Figure 4.33C).

4.8. Gravity data modeling

There are two approaches to interpretation of the Bouguer anomaly:

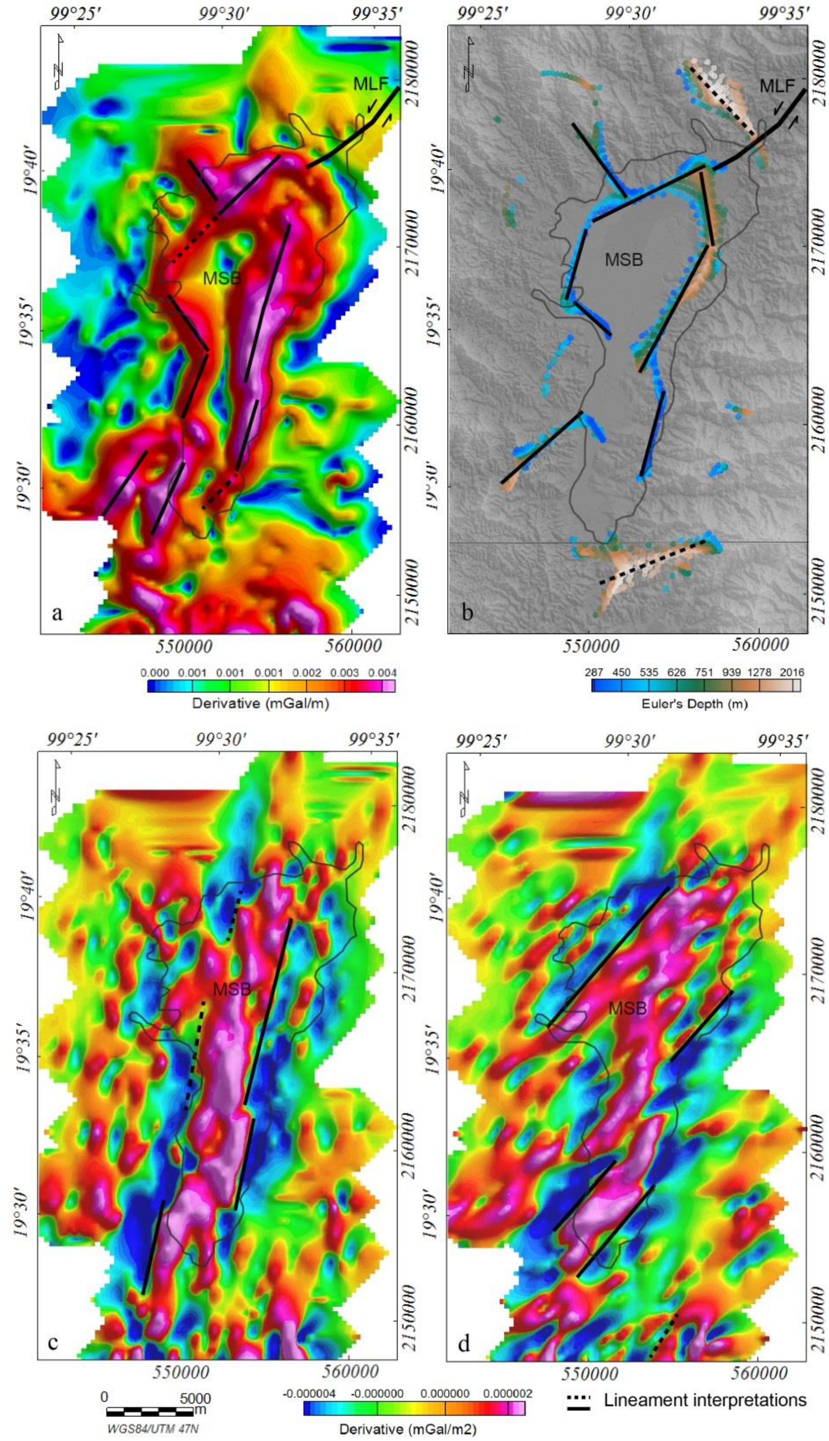


Figure 4.32: *Fault lineament interpretations displaying with edge detection technique anomaly of (a) the total horizontal derivative, (b) the Euler deconvolution's cluster solutions, (c and d) the 2nd horizontal derivative along the ESE and the SE directions respectively (Kanthiya, 2017).*

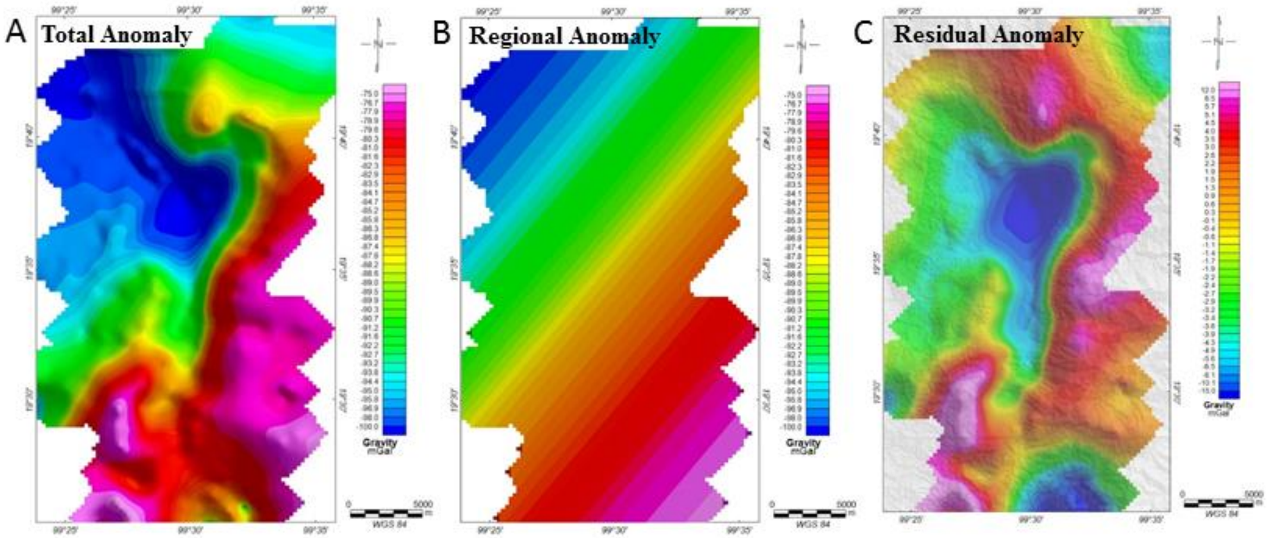


Figure 4.33: A) Total *CBA* B) Regional *CBA* and (C) residual *CBA. The trend method from Whitehead and Musselman (2006) were used to distinguish regional and residual signals (modified from Kanthiya, 2017).*

1) Forwarding method

- Forwarding method is a direct method, which the original data are analyzed (Figure 4.34).
- Start with a density model construction and then compute calculated data.
- Non-unique with many solutions.
- Computation method can be a simple equation or a complex computer algorithm.

2) Inversion method

- Inversion method is an indirect method which models are constructed to compute theoretical gravity anomaly (Figure 4.34).
- Start with measured data set up and compute a model of the Earth e.g. density, depth, thickness.
- Non-unique with many solutions.
- Solving approach for the inverse problem can involve a trial-and-error approach or an automated inversion algorithm.

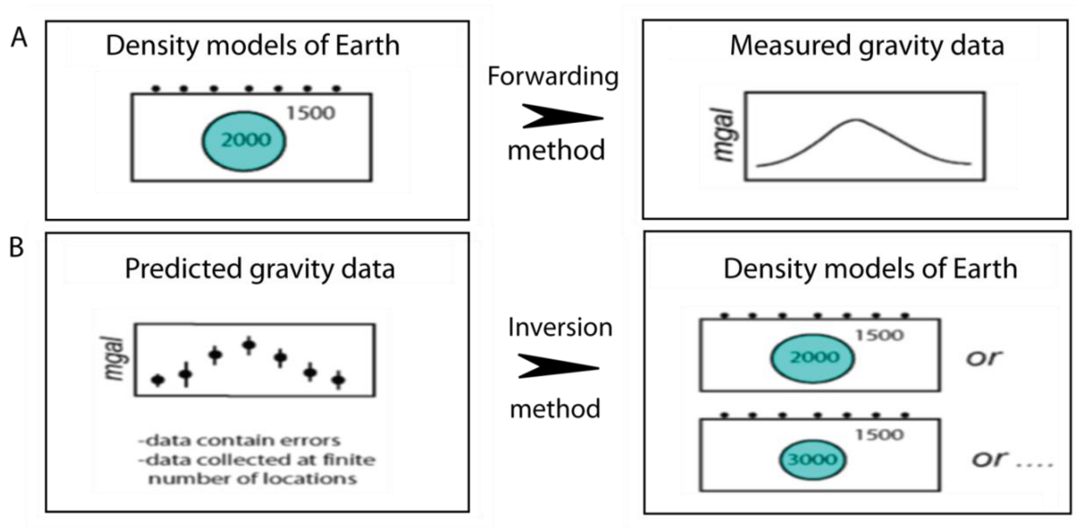


Figure 4.34: *Concept of A) forwarding and B) Inversion methods in gravity data (modified from Kanthiya, 2017).*

4.8.1. Forwarding gravity model construction

1) Construction of a reasonable geologic model.
2) Computation of its gravity anomaly shown as calculated data (Figure 4.35, solid line).
3) Comparison of computed with observed data (Figure 4.35, dashed line).
4) Alteration of the model to improve correspondence of observed and calculated anomalies and return to step 2) until archive the best fit model (Figures 4.35 and 4.36).

Note that: the modeling of inversion method will be focused in a graduated level course (212731).

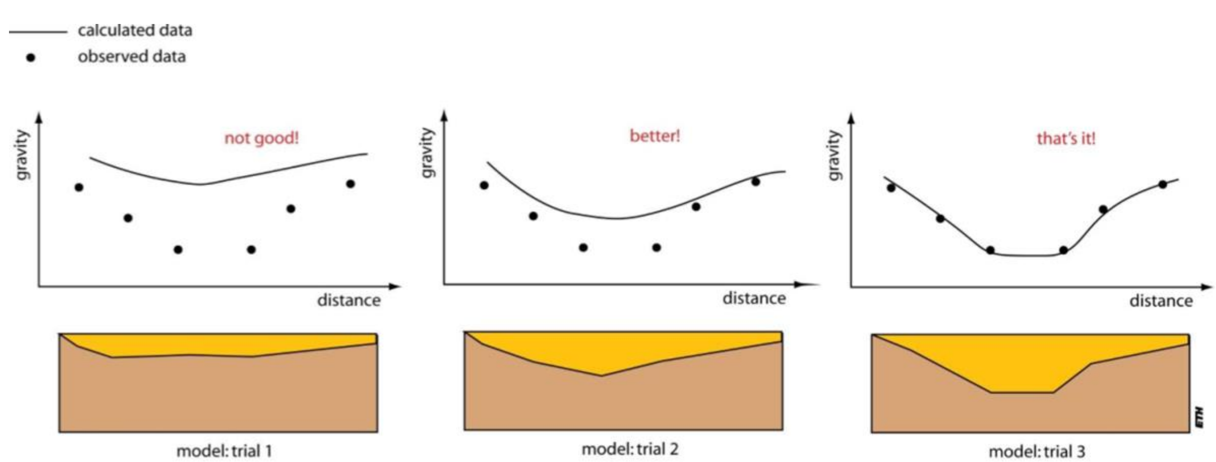


Figure 4.35: *Gravity forward modeling construction with the best-fit solution (right).*

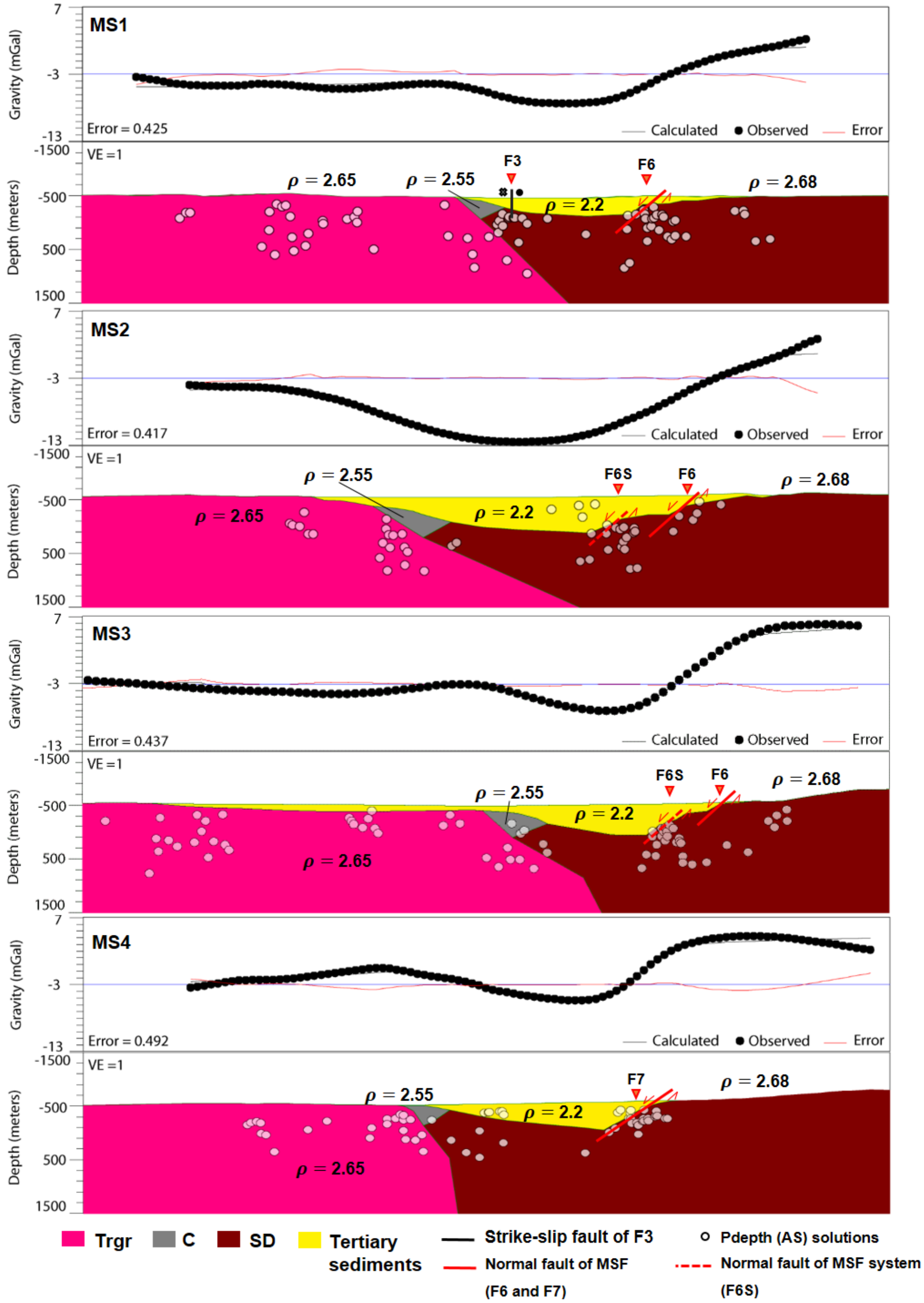
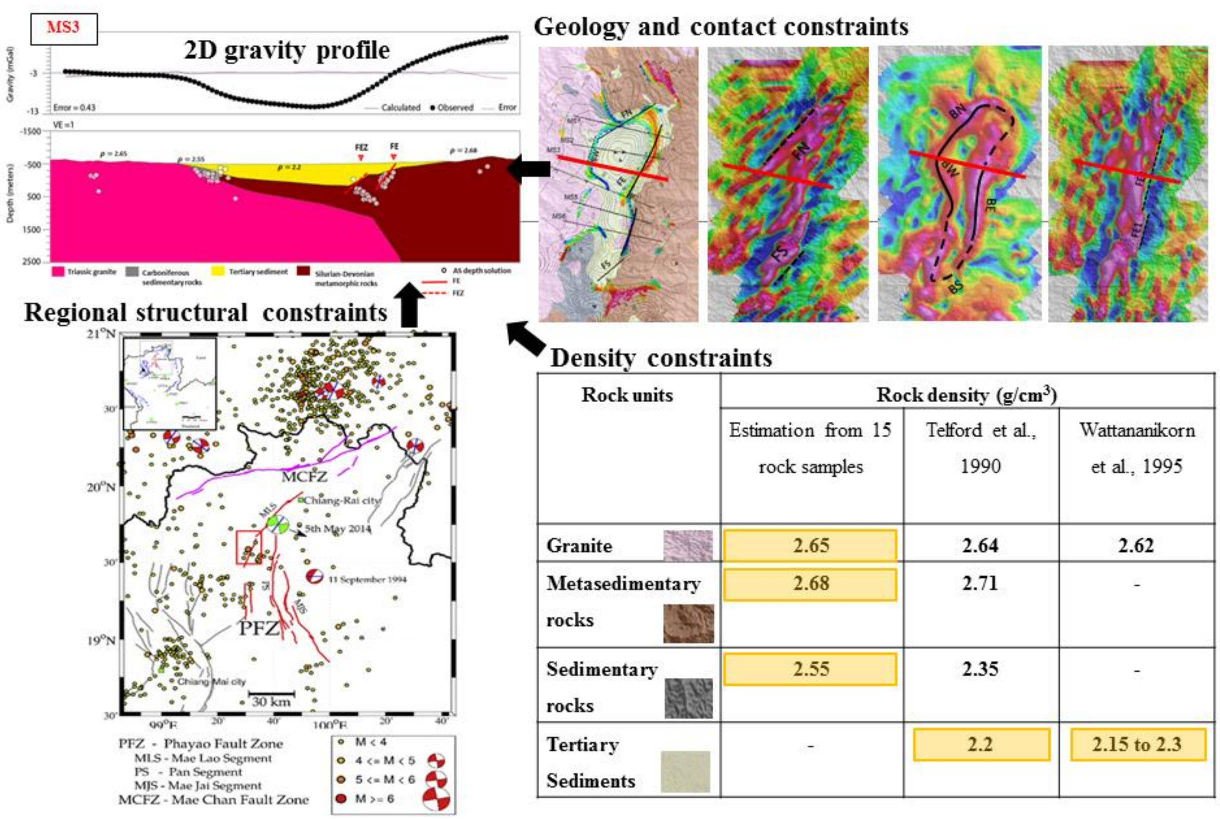


Figure 4.36: *2-D forward gravity profiles with misfits less than 1% representing half-graben geometry of Mae Saui basin (Kanthiya, 2017). Profile locations are shown on Figure 4.37.*

4.8.2. Data constrains for modeling

- Surface and subsurface locations of anomaly sources.
- Earthquake and hot spring locations for fault contact indications (Figure 4.37).
- Geological surface contact released from geologic map, fault map, etc. (Figure 4.37).
- Shapes, size, and orientation of anomaly sources.
- Previous geophysical or geological data in 1D (log/drilling), 2D (cross-section, profile), or 3D models.
- Physical properties such as density for individuals (Figure 4.37).



Rock units	Rock density (g/cm³)		
	Estimation from 15 rock samples	Telford et al., 1990	Wattananikorn et al., 1995
Granite	2.65	2.64	2.62
Metasedimentary rocks	2.68	2.71	-
Sedimentary rocks	2.55	2.35	-
Tertiary Sediments	-	2.2	2.15 to 2.3

__Figure 4.37:__ Surface geological and geophysical data constraints for the 2D gravity modeling (modified from Kanthiya, 2017).

4.8.3. Ambiguity in gravity modeling

- This is the intrinsic problem that gravity interpretation not unique. Although for any given body, a unique gravity field is predicted, a single gravity anomaly may be explained by an infinite number of possible solutions (Figure 4.38).
- It is important to use constraints from surface outcrop, boreholes, mines and other geophysical methods. The value of gravity data is dependent on how much other information is available.

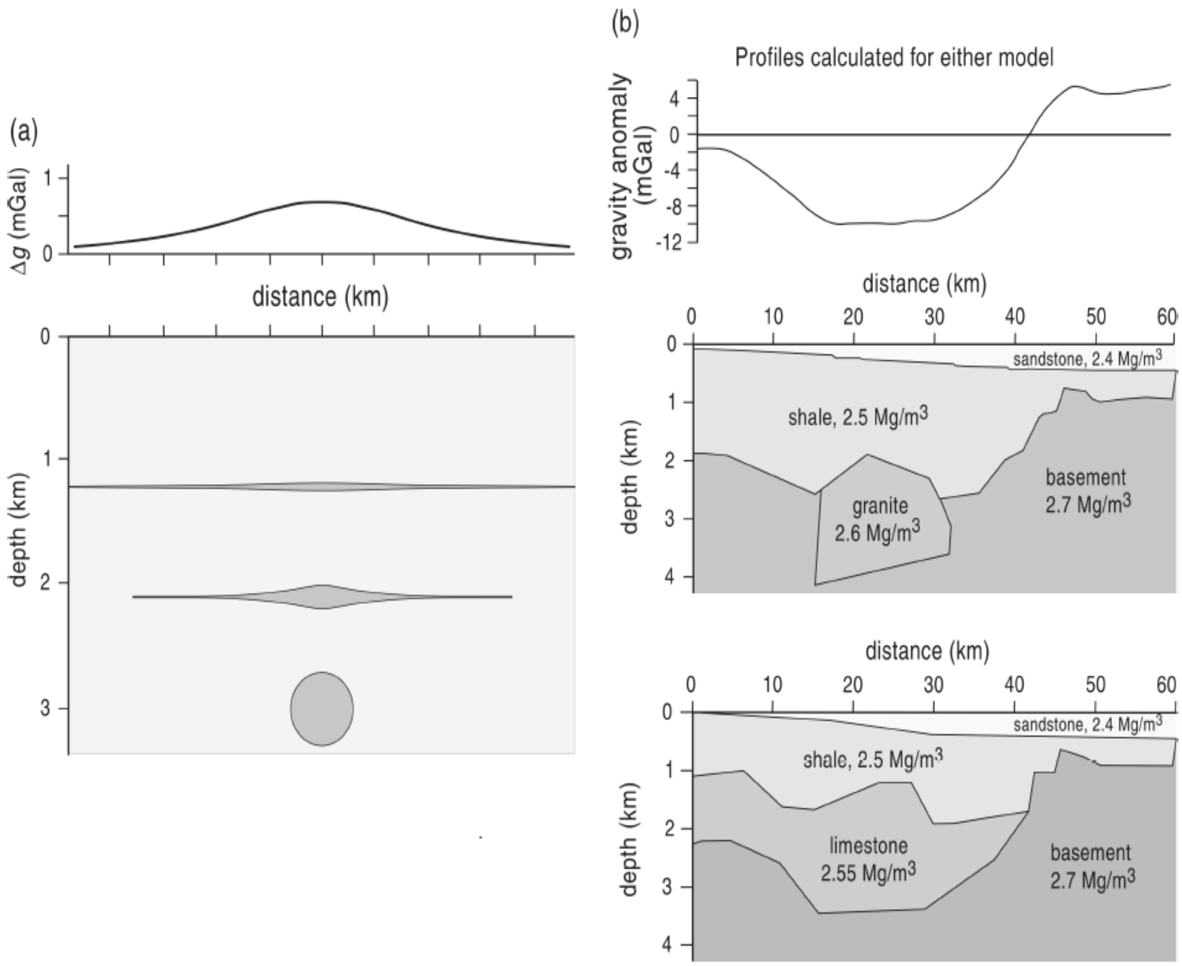


Figure 4.38: *Non-uniqueness of the gravity interpretation. The plotted models produce exactly the same gravity anomalies (Musset and Khan, 2000).*

4.9. Application in gravity method

4.9.1. Archaeological application

Title: Archaeological microgravimetric prospection inside don church, Spain (Jorge et al., 2012). See Figure 4.39.

Purpose: To investigate buried shallow archaeological structure using microgravimetric technique to prospect structure of Don church, in urban area of Alfafar town, Valencia, Spain.

Methodology: Collected data with Lacoste & Romberg D203 gravity meter and process via standard gravity correction with reduction density of 2.1 g/cc.

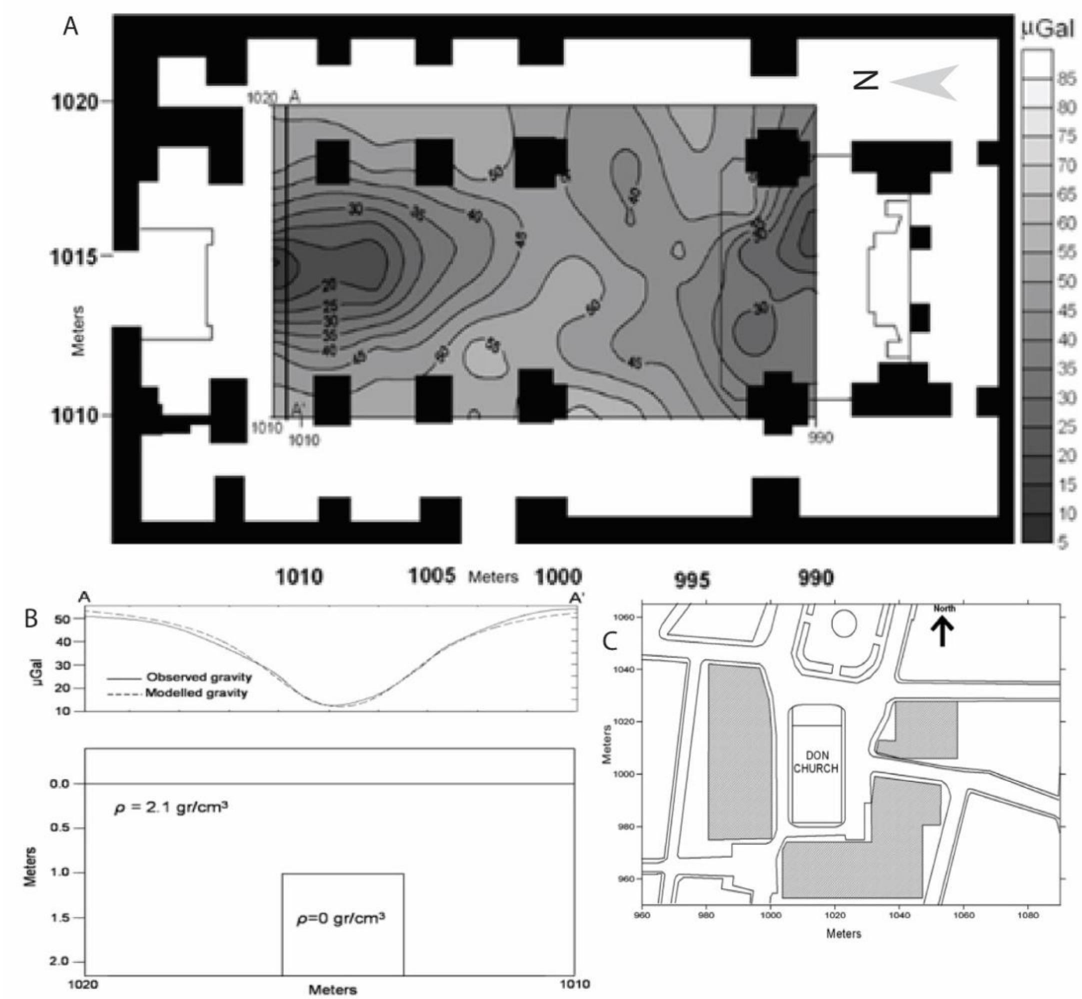


Figure 4.39: *A) The final residual Bouguer gravity anomaly of gravity anomaly lows B) 2D gravity modeling show the cavity rectangle 2.5 m wide and 1.2 m high, located 1 m under the actual floor that can be linked to the expected crypt. C) A location of the Don Church (modified from* Jorge et al., 2012*).*

4.9.2. Geohazard application

Title: Collapse susceptibility map in abandoned mining areas by microgravity survey: A case study in Candado hill, Spain (Martínez-Moreno et al., 2016). See Figure 4.40.

Problem The presence of disused gypsum mine galleries in Candado hill has caused constructions at the western end of the hill to collapse.

P**urpose:** Therefore, the precise dimensions and position of the galleries are unknown, making it essential to undertake a thorough microgravity study to assess the collapse susceptibility.

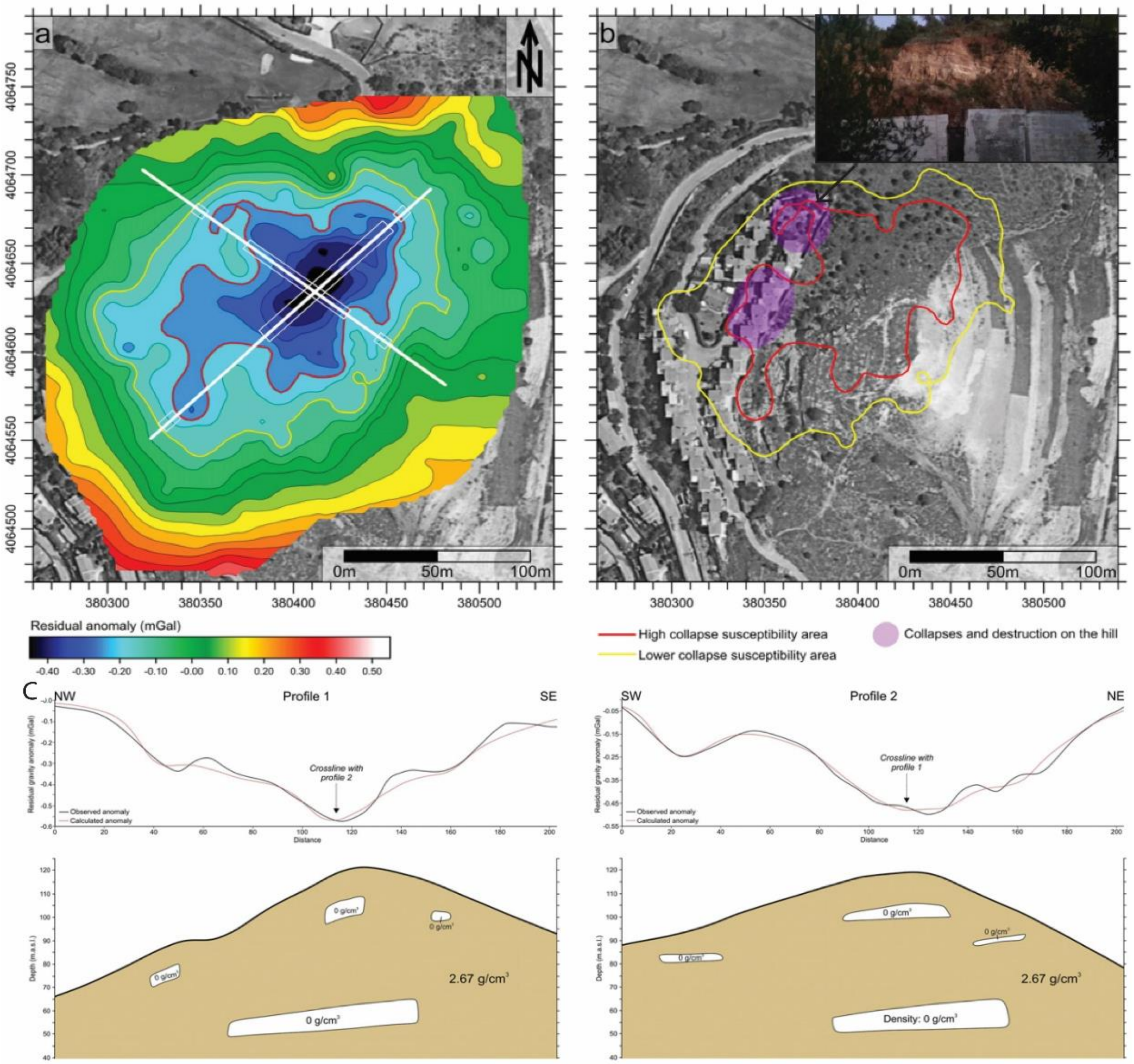


Figure 4.40: *Gravity Interpretation of collapse susceptibility (red and yellow lines) map over (a) residual anomaly map (b) orthophoto (c) Microgravity models showing the galleries of gypsum mine activity, where density of cavity is 0 g/cm^3 (modified from Martínez-Moreno et al., 2016).*

4.9.3. Geothermal application

Title: Geothermal exploration using airborne gravity and magnetic data at Siwa Oasis, Western Desert, Egypt (Zaher, et al., 2018). See Figure 4.41.

Purpose: 3D inversion gravity can improve subsurface image of geothermal source that can help us determined its geothermal potentiality at the Siwa Oasis, sitting 18 m below sea level and stretching for more than 50 km from the East to the West.

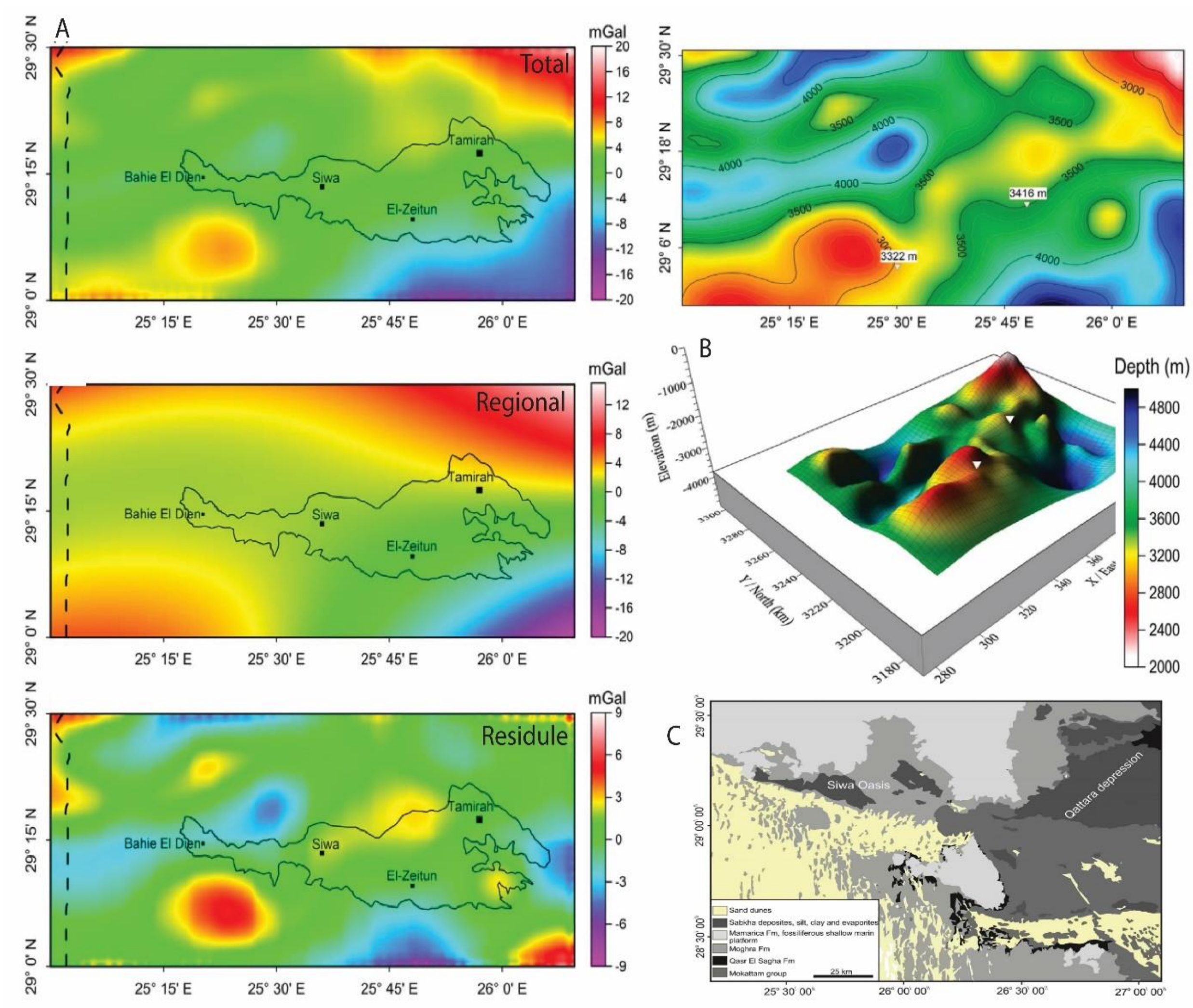


Figure 4.41: *A) Airborne BA map of Siwa Oasis with the onshore Bouguer correction reduction density is 2.67 g/m3. The polynomial regression technique was used to separate regional and residual anomalies B) Depths to the basement model of a greater than 2 km. In the southwest and northeast corners is shallowest depths range from 2 - 2.5 km. The SW-NE trend of the basement Highs (red) separates two major basins (depth>4 km) at the southeast and northwest areas. The inverted white triangles refer to the deep wells used for modeling constrain (*modified from Zaher, et al, 2018*).*

4.10. Exercises for gravity method chapter

1) How is the gravitational acceleration, g, related to geology?
2) Why is the geoid importance to geodetic surveying, or plane surveying?
3) Most free-air gravity anomalies are in the range of a few hundred mGal, while in the measurements at sea, most shipboard corrections are small close to 1 mGal, why?
4) If we require a gravity precision of 0.01 mGal, then relative station elevations need to be known to about 3 cm. To get such a precision requires very careful location surveying to be done, how's about 0.1 mGal precision?
5) Is terrain correction necessary for the survey over flatted areas?
6) If you took a gravity meter 1 km down a mine in rocks of density 2.3 g/cc, how much gravity would change?
7) The mean radius of the Earth is 6371 km. On the taking a gravity meter 1 km up in a balloon you would expect the value of g to decrease by?
8) Figure 4.42 is a Bouguer anomaly map, contoured at an interval of 50 mGal, of a drift-covered area.

   1- On the map, sketch in what you consider the regional field along A-A' profile and then remove it from the observed field to isolate residual anomalies,

   2- What further information/technique would be required before a full interpretation could be made of the Bouguer anomaly?

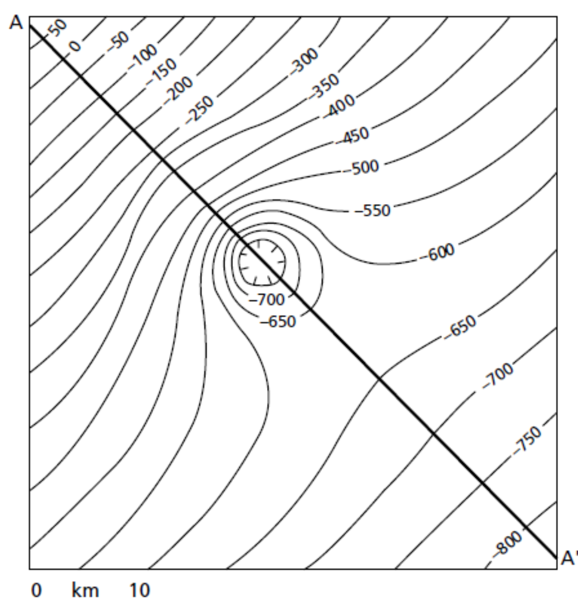


Figure 4.42: *Bouguer anomaly map pertaining to Question 8 Contour interval is in mGal unit (Kearey et al., 2002).*

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Chapter 5: Magnetic Method

5.1. Introduction to magnetic method

- Magnetic survey is one of geophysical method used to investigate subsurface geology on the basis of anomalies in the Earth’s magnetic field resulting from the magnetic properties of rocks.
- Magnetic prospecting is used to search for oil and minerals, archaeology research and hazardous waste. The prime targets are the depth to basement, metal ore bodies, hydrothermal alteration (geothermal) and archaeology, e.g. fire pits, kilns and disturbed Earth.
- Magnetic survey is the oldest method of geophysical prospecting, but has become relegated to a method of minor importance because of the advent of seismic reflection surveying.
- The great problems involve in interpreting magnetic anomalies greatly limits their use and surrounding noise.

5.2. Basic concept of magnetic theory

- The magnetic field of a exposed/buried rocks can create a magnetic force on other rock~~t~~s with magnetic fields (Figure 5.1). That force is what we call magnetism.

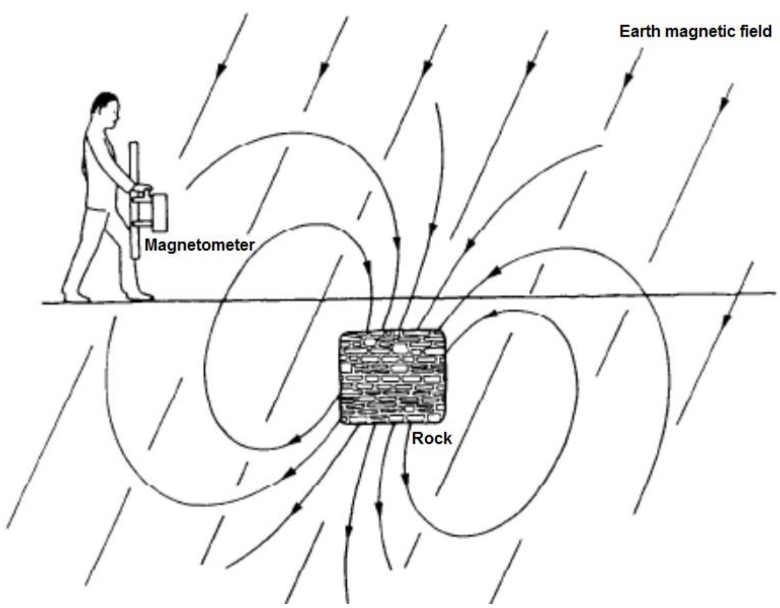


Figure 5.1: *Buried rock distort the Earth's magnetic field, the resulting of a created magnetic force due to the magnetized rock are recorded during magnetic survey (Clark, 1996)*

5.2.1. Magnetic force

- Coulomb's Law states that magnetic force is an attraction or a repulsion that arises between two poles charged particles because of their motion (Figure 5.2).
- It is the basic force responsible for such effects as the action of electric motors and the attraction of magnets for iron.
- Force can be + or – depended on attraction or repulsion of the force.

$$F = \left(\frac{m_1 m_2}{\mu r^2}\right)$$

F = magnetic force

m_1, m_2 = The poles of strengths

r = Distance (cm)

μ = Magnetic permeability constant (a property of the medium which the magnet located). In vacuum μ = 1; atmosphere μ ~ 1

- Note that: a property of the medium is matter to strength of magnetic field.
- Unit is dyne or in Newton seconds.

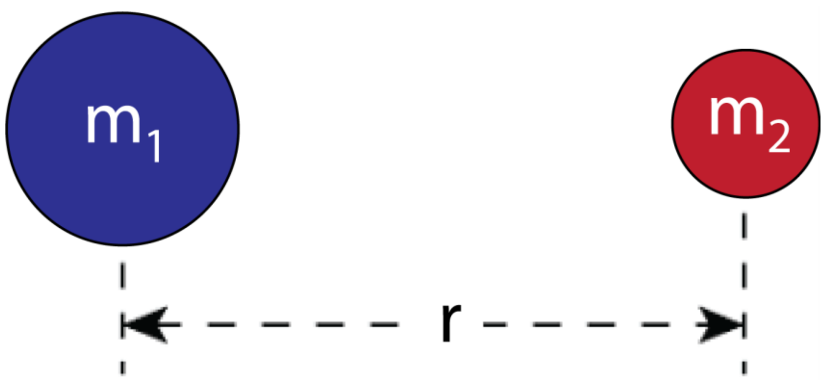


Figure 5.2: *Magnetic force generated by two magnetic poles in a distance of r (https://www. muhendisbeyinler.net/newtonun-kutle-cekim-kuvveti-dogru-mu/).*

5.2.2. Magnetic field

- Magnetic field strength or magnetic field is a magnetic force per one unit pole (m_2).
- Magnetic field (B) present magnetic force (*F*) of large magnet (m_1) that affect to the one unit pole (m_2) which is spotted far from the large magnet at the *r* distance.

- The strength of the force between magnets depends on the distance between them.

$$H = \frac{\left(\frac{m_1 m_2}{\mu r^2}\right) r}{m_2}$$

$$H = \left(\frac{m_1}{r^2}\right)$$

- Note that the magnetic field strength component in any direction is given by the partial derivative of the potential in that direction.

Magnetic field unit

- Assuming a unit pole is positive.
- Units are dyne/unit pole or Newton seconds / unit pole.
- Magnetic measurements use the Tesla unit. It is a large unit, and the smaller unit of Nanotesla is used for small fields like the Earth's magnetic field.

5.2.3. Magnetic dipole moment

- A common bar magnet exhibits a pair of equal force acting to parallel to each other but in opposite direction (dipoles), if the magnet is placed in a magnetic field strength (H).
- Magnetic moment is defined as a charge of the two opposing poles (m) separated by a finite distance (l) (Figure 5.3).

$$M = ml$$

where M: is a vector

Where, M = the magnetic moment

m = magnetic field strength of one unit pole

l = a distance apart

- Note that: the combination of attractive and repulsive forces on the same bar magnet creates a torque.

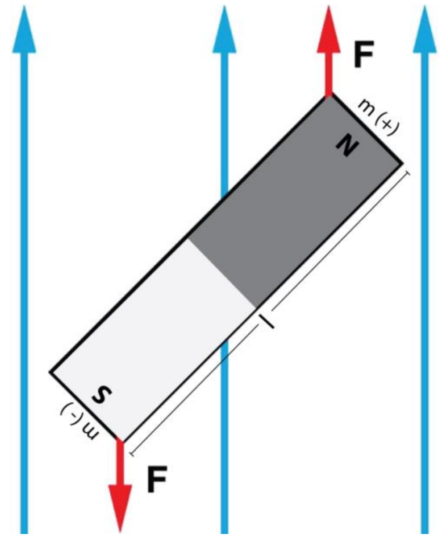


Figure 5.3: *A magnet placed in a magnetic field will experience a magnetic dipole moment (https://physics.stackexchange.com/questions/290585/direction-of-electric-dipole-moment-and-magnetic-dipole-moment).*

5.2.4. Induced magnetization

- If a material is placed in a magnetic field it will acquire magnetization in the direction of the inducing field. The magnetization is lost when the material is removed from the field. This phenomenon is referred to as induced magnetization. In geophysics, intensity shows how good that rocks are induced to be a magnetization (Figure 5.4)
- The intensity of the induced magnetization (I) of material is defined as the magnetic dipole moment (M) per volume (V) or unit pole strength (m) per unit area (A).

$$I = \frac{M}{Volume} \qquad I = \frac{ml}{Volume} \qquad I = \frac{m}{Area}$$

Where. I is induced magnetization in Am^{-1}.

M is the magnetic moment of a sample of length *(l)* and cross-sectional area (A)

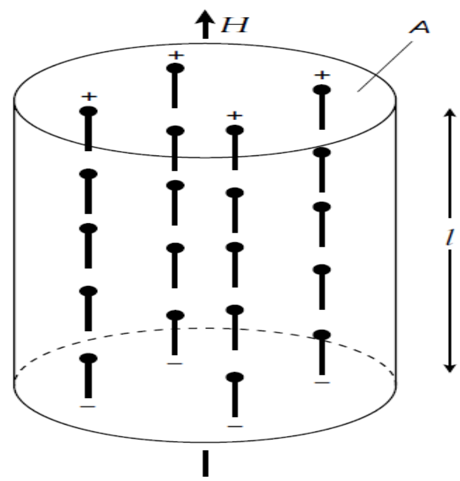


Figure5.4: *Schematic representation of an element of material in which elementary dipoles align in the direction of an external field to produce an overall induced magnetization (Dobrin and Savit, 1988).*

5.2.5. Magnetic susceptibility

- The magnetic susceptibility is the basic rock physical parameter determining the applicability of a magnetic survey. The magnetic susceptibility can also be used to determine geological processes.
- Magnetic susceptibility or magnetization (k) is a temporary magnetized material while a magnetic field (H) applied.
- The direction of magnetization is parallel to Earth magnetic field.
- Thus, induced magnetization (I) would be a constant proportionality due to a magnetic field strength (H) of the inducting field.

$$I = kH$$

- Note that: k is dimensionless due to induced magnetization (I) and magnetic field (H) are in the same unit
- Figure 5.5 shows the most common rock-forming minerals exhibit a very low magnetic susceptibility. There are only two geochemical groups which provide such minerals.
- The iron–titanium–oxygen group possesses a solid solution series of magnetic minerals such as magnetite (Fe_3O_4). Hematite (Fe_2O_3) is also common iron oxide group but it is antiferromagnetic and does not respond any magnetization.

- The iron–sulphur group provides the magnetic mineral. Histogram shows mean values and ranges in susceptibility of common rock types. (After Dobrin and Savit 1988).

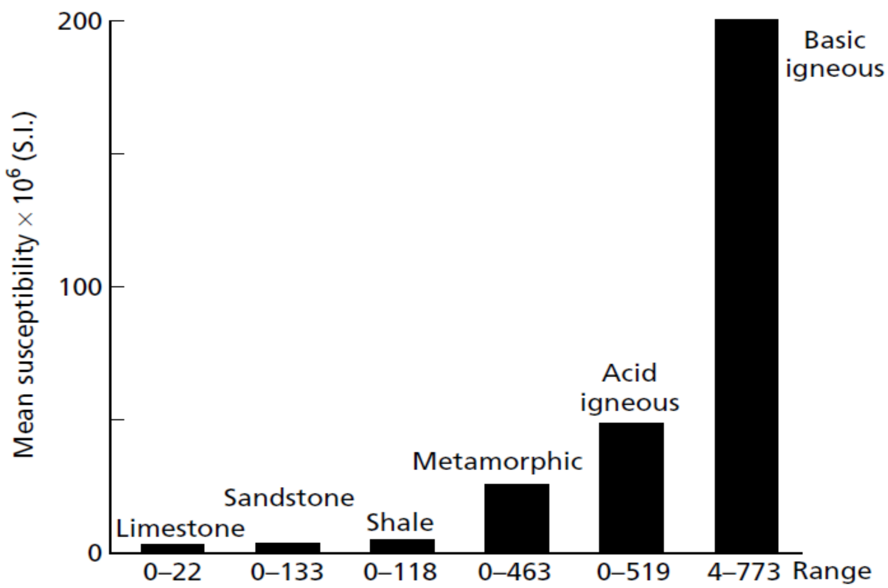


Figure 5.5: *Histogram shows mean values and ranges in susceptibility of common rock types (Dobrin and Savit, 1988).*

5.3. Magnetic properties of rocks on Earth

- All atoms have electrons. So all materials should be magnetic, but there is great variability in the magnetic properties of materials.
- The electrons in some atoms align to cancel out one another's magnetic influence.
- While all materials show some kind of magnetic effect, the magnetism in most materials is too weak to be detected without highly sensitive instruments.

5.3.1. Record of Earth's magnetic field in rocks

- Rocks and minerals at high temperatures (e.g. molten) must cool through the Curie temperatures and stay below that temperature through time. Thus, Earth's magnetic field can be record in crystalline rocks at the time of their cooling.
  - Above Curie temperatures, atoms are random (no magmatism even it is a ferromagnetic type) (Figure 5.6A).

- Below Curie temperature, atoms align in domains that are independent of each other which means magnetic field is absent (e.g. Moon) (Figure 5.6B).
- Below Curie temperature, atoms align with magnetic field if one is present (e.g. Earth) (Figure 5.6C). The alignment directions were depended on the orientation of Earth magnetic field during that times (Figure 5.7).

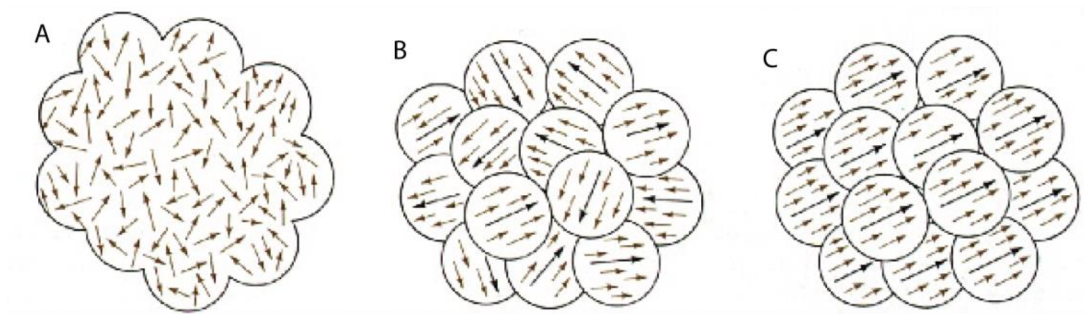


Figure 5.6: *Magnetic domains preserved in crystalline rocks in different situations (modified from http://ees2.geo.rpi.edu/geo1/lectures/lecture15/interior_07.html).*

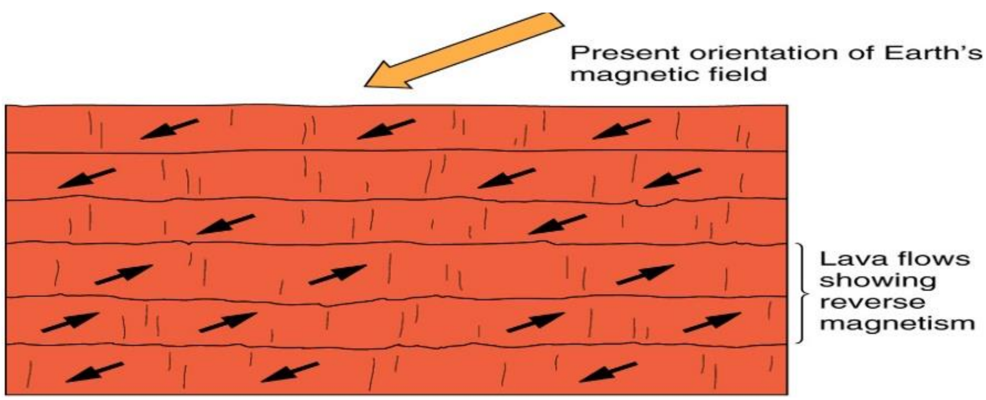


Figure 5.7: *Magnetic domains preserved in volcanic rocks in different time periods and magnetic field patterns (*http://ees2.geo.rpi.edu/geo1/lectures/lecture15/interior_07.html*).*

5.3.2. Magnetic properties of materials

- Atoms with similar magnetic orientations line up with neighboring atoms in groups called magnetic domains. Difference of magnetic domain's orientations gives different magnetic properties of materials.

Diamagnetic materials

- Diamagnetic materials have a negative magnetic susceptibility, but very low (~0) which cannot be measured.
- Electrons are oriented so their individual magnetic fields cancel each other out (Figure 5.8A).
- Examples of Diamagnetic materials are. Quartz, Feldspar, and rock salt.

Paramagnetic materials

- Paramagnetic materials are weak positive magnetic susceptibility but the atoms themselves are randomly arranged so the overall magnetism of a sample is zero (Figure 5.8B).
- Examples of Paramagnetic materials are Fe-Mg silicates (Pyroxene, Amphibole, and Olivine)

Ferromagnetic materials

- Ferromagnetic materials have very strong magnetic properties. These metals are the best known as meteorite materials that do not occur naturally on Earth (Figure 5.8C).
- Examples of Ferromagnetic materials are iron, nickel, and cobalt.

Ferrimagnetic materials

- Magnetic domains of ferrimagnetic materials are oriented in one direction mainly. That material can be strong magnetized and have high susceptibility.
- Magnetic domains in a ferrimagnetic mineral will always orient themselves to attract a permanent magnet (Figure 5.8D).
- Examples of well know anitferromagnetic materials are Magnetite and Ilmenite.

Anitferromagnetic materials

- *M*agnetic domains of anitferromagnetic materials are parallel and antiparallel in equal numbers, that involve the net magnetic susceptibility is very weak (close to 0) (Figure 5.8E).
- An example anitferromagnetic Material is Hematite.

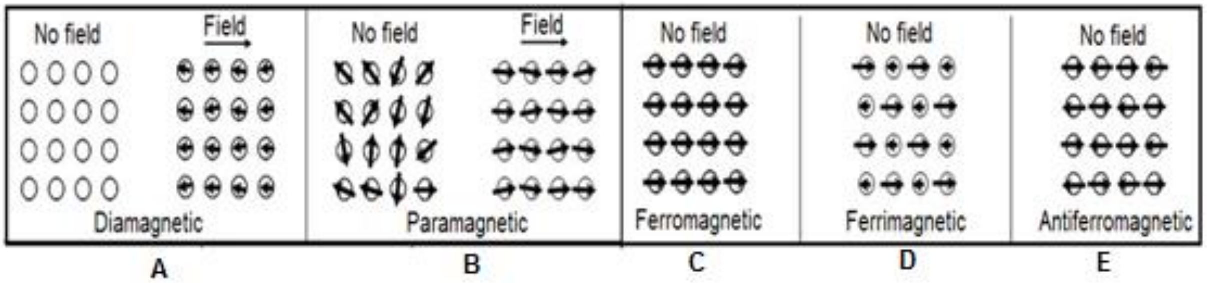


Figure 5.8: *Schematic diagram showing orientation of magnetic moments in the crystal lattice of different materials: A) diamagnetic, B) paramagnetic, C) ferromagnetic, D) ferromagnetic, and E) antiferromagnetic (*https://kaiserscience*. wordpress.com/physics/ electromagnetism/sources-of-magnetism/).*

5.3.3. Rock magnetization

- Rock magnetization is commonly found in lithology of dykes, faulted, folded or truncated sills and lava flows, massive basic intrusions, metamorphic basement rocks and magnetite ore bodies.
- Basic igneous rocks are usually highly magnetic due to their relatively high magnetite content. The proportion of magnetite in igneous rocks tends to decrease with increasing acidity so that acid igneous rocks.
- Metamorphic rocks are also variable in their magnetic character depending on basic chemical composition.

5.3.4. Curie temperature of magnetization

- The Curie temperature (T_c) is the temperature where a material's permanent magnetism (Paramagnetism) changes to induced magnetism (Ferro/ferrimagnetism).
- Rocks at deep crust (> Curie depth, Figure 5.9) cannot be magnetized because those rocks lose all magnetization above the Curie temperature in deeper region. Magnetic model can't sense deep into crust because heat destroys magnetism.

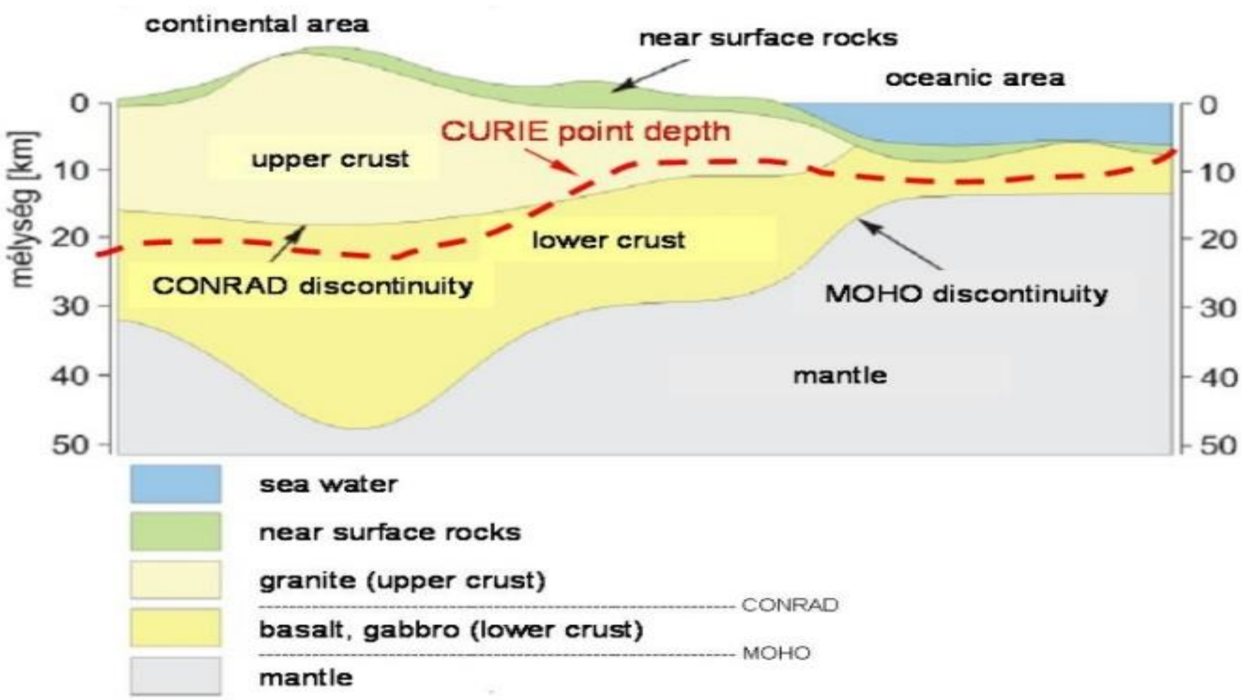


Figure 5.9: *Comparison between The Curie point depth and Moho depth (https://www.camizu.org/interior-of-earth-crust/).*

5.4. Earth magnetic field

5.4.1. Earth magnet

- Due to Earth's magnetic field similar to the bar magnet, Earth magnets create a magnetic field in the space around them (Figure 5.10).
- Earth magnet has north and south magnetic poles.
- Earth magnet and its field are detected by compasses.
- Earth magnet and its field create forces on other magnetic materials. Earth magnet has the north magnetic pole is at the south geographic pole.

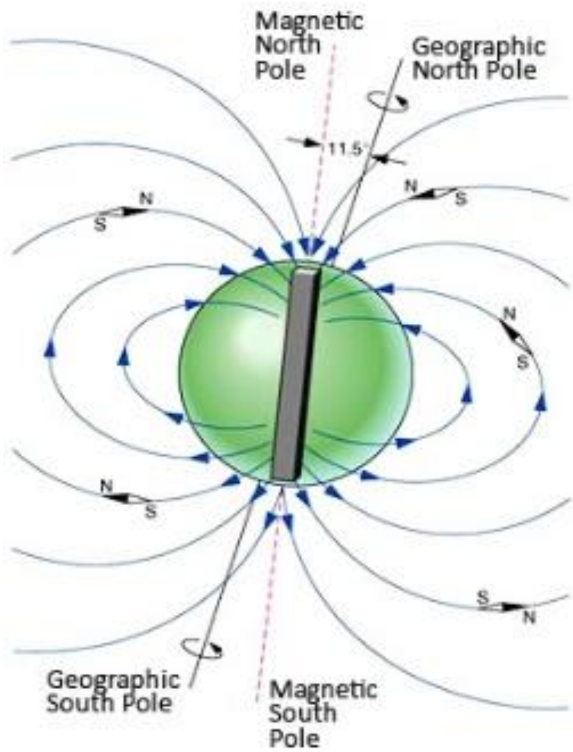


Figure 5.10: *A current Earth magnet and its magnetic field (*https://www.heartmath.*org/ research/global-coherence/)*

5.4.2. Sources of Earth's magnetic field

- The Earth's magnetic field or geomagnetic field is more complicated than a simple dipole field. The Earth's magnetic field consists of:
    - The main field (B) is the magnetic field that extends from the Earth's interior out into space, where it meets the solar wind. This field is generated by convection in the liquid outer core, which drives electric currents. The Earth's field is 99% from Earth's interior.
    - The external field such as diurnal or magnetic storm is caused by electric currents in ionized layers of the outer atmosphere. This field is only 1% of Earth magnetic Field.
    - Local anomalies are caused by magnetic bodies in the crust, where the temperature is higher than the Curie temperature. These bodies are the targets of magnetic survey.

5.4.3. Earth's magnetic field components

- At any location, total Earth's magnetic field (B) can be represented by combination of geometric elements in Horizontal (H) and Vertical (Z) components. It's a vector quantity defined by its total intensity and direction (Figure 5.11).

- Magnetic intensity of total Earth's magnetic field (B) varies in strength from about 25 000 nT in equatorial regions to about 70 000 nT at the poles (Figure 5.12).
- At any point on the Earth's surface a freely suspended magnetic orientation will assume a position in space in the direction of the ambient Earth magnetic field. This will be at an angle to both the vertical (inclination) pole and North geographic pole (declination) (Figure 5.11).
  - Inclination is the dip of B. The Earth's magnetic field lines intersect the ground surface at an angle ranging from zero (at the equator) to 90 degrees (at the poles) (Figure 5.13).
  - Declination (D) is horizontal angle between geographic and magnetic north. Note that presently tilted angle is around 11.5° (Figure 5.13).

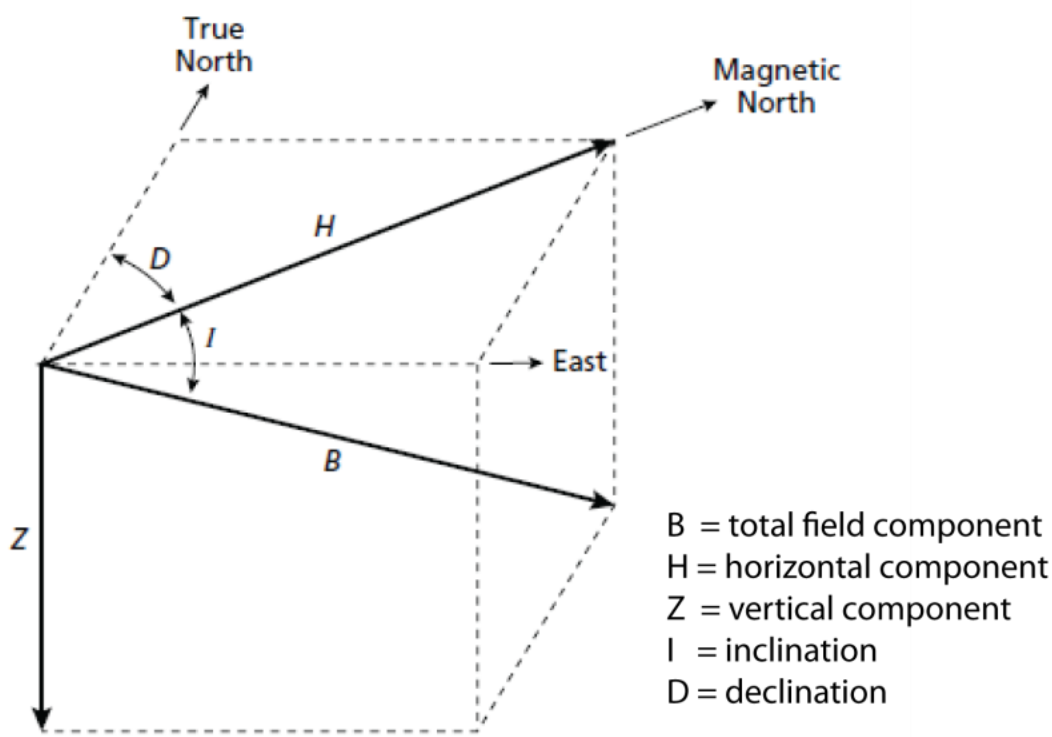


Figure 5.11: *The Earth magnetic field elements (modified from Robinson and Crouch, 1998).*

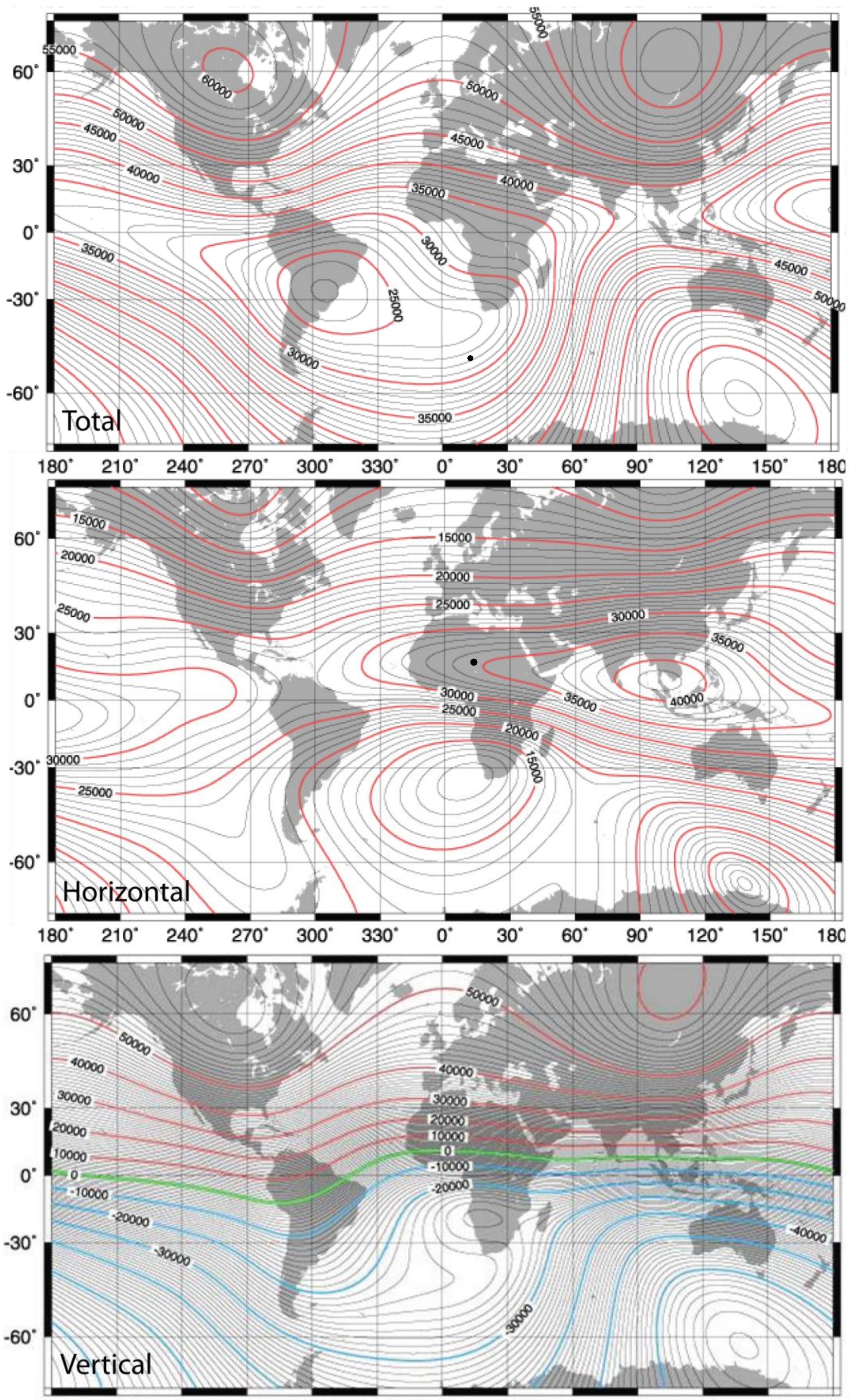


Figure 5.12: *Total, Horizontal, and Vertical Earth Magnetic Intensity Fields (*https://www.ngdc.noaa.gov/geomag/WMM/DoDWMM.shtml*).*

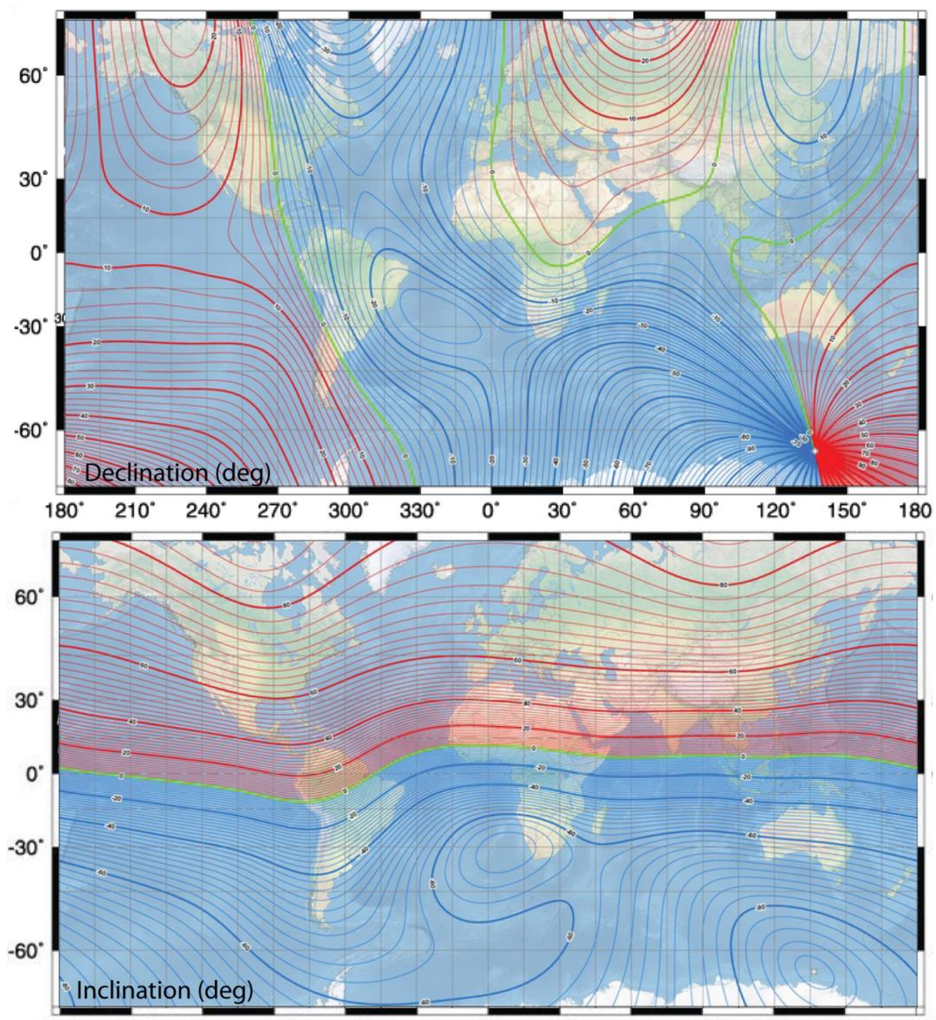


Figure 5.13: *Angles of Earth's magnetic declination and inclination (*https://www.ngdc.noaa.gov/geomag/WMM/DoDWMM.shtml*).*

5.5. Magnetic acquisition

- Magnetic surveys provide a cheap and simple way of investigating the subsurface like gravity.
- Field surveys are relatively simple compared to gravity.
- The magnetic field is measured along a series of traverses. Some points on the traverses are repeated using tie lines to monitor short-term variations.
- The times of the observations are accurately matter.

- During the survey, a separate record of the field is kept at a fixed base station. The timing of the observations is synchronised with those of the main survey.
- Compared to gravity, magnetic survey is much easier due to no accurate elevation control.

5.5.1. Target of magnetic acquisition

- Basement: tends to be igneous or metamorphic, thus greater magnetic properties compared to sediments.
- Soils and other weathered products: because magnetic minerals tend to weather rather rapidly compared to quartz, will get reduction of magnetic materials with weathering.
- Man-made objects: iron and steel. Beware its *noise.*
- Ore deposits: many economic ores are either magnetic, or associated with magnetic minerals. These are structurally controlled.

5.5.2. Magnetometers

5.5.2.1. **The proton magnetometer**

- The proton magnetometer is the most common hand-held field equipment. They are based on the precession of protons in the magnetic field (Figure 5.14).
- The proton magnetometer has a container filled with a hydrogen rich liquid (e.g. water or an alcohol).
- When an electric current is passed into the coil, the magnetic field is generated (about 5 to 10 nT) and the protons are aligned with this field.
- Then the current is switched off and protons start to align with the Earth's magnetic field, processing.
- In the coil the electric current is induced by the electromagnetic induction. The frequency is measured and strength of the magnetic field computed.
- The readings are quick (few seconds) and great sensitivity, but the instrument has to be moved manually between positions.
- The meter can measure only an absolute strength of a total magnetic field.
- They do not work in the presence of AC power interference, e.g., below power lines.

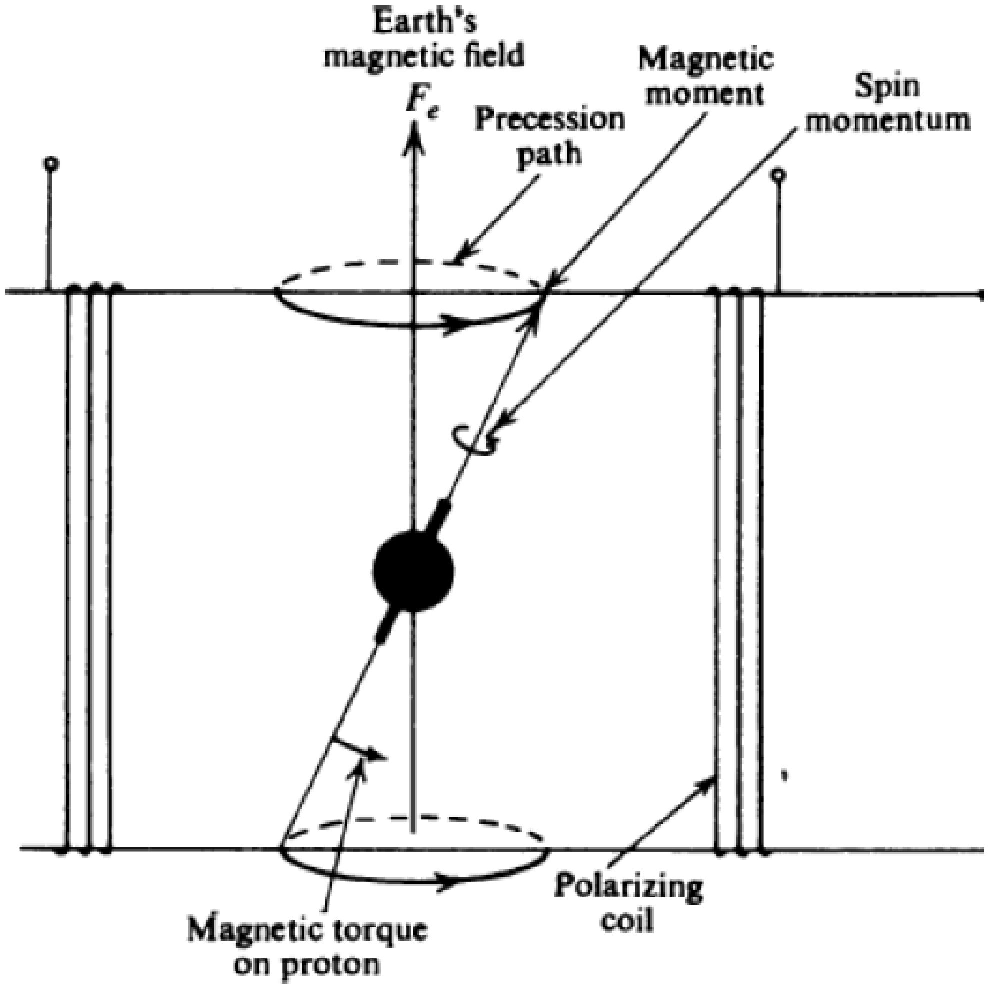


Figure 5.14: *Principle of the proton magnetometer (Telford et al., 1990).*

G-858 Magnetometer

- The G-858 MagMapper is a high sensitivity magnetometer (Figure 5.15).
- G-858 data acquisition offers either continuous or discrete station recording.
- It is designed to interface easily with standard computers and peripherals. Geometrics encourage clients to provide their own processing computer hardware.

Figure 5.15: *G-858 Magnetometer, a commonly used proton magnetometer (Scietrex Limited, 1996).*

5.5.3. Magnetic field design

- These are usually done with portable proton precession magnetometers. Profiles or networks of field design points are measured in the same way as for gravity.
- To achieve accurate measurement, reading of base magnetometer has to set up continually. Reading of rover station can be recorded in continuous or discrete modes depending on survey design (Figure 5.16).
- Survey line is important to perpendicular to the strike of an elongate body or structure.

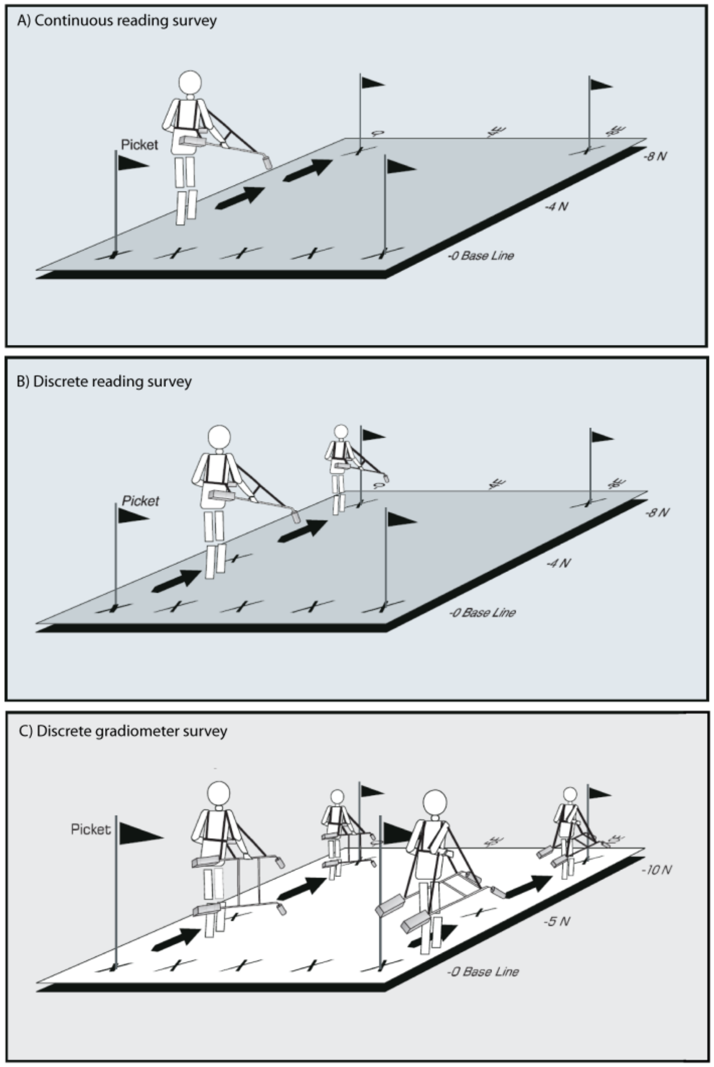


Figure 5.16: *Magnetic field designs of A) Continuous reading, B) discrete reading, and C) discrete gradiometer surveys (modified from Scietrex Limited, 1996).*

5.5.4. Requirement of magnetic survey

- Record the time at which readings were taken, for drift correction. It is necessary to tie back to the base station at 2-3 hour intervals.
- Select the base station in an open area and accessible.
- Toward the arrow on a sensor to North.
- Stay away from interfering objects, e.g., wire fences, railway lines, roads,
- Not carry metal objects e.g. mobile phones or keys.
- Take multiple readings at each station.

5.5.5. Total magnetic field data

- An anomaly is created when the Earth’s magnetic field is disturbed by an object that can be magnetized. The resultant anomaly can also be viewed in terms of vectors as previously described for Figure 5.17.
  - Observed magnetic data at any point on the Earth represent the resultant total magnetic field (B) including an induced magnetization (I) and remnant Earth’s field (H).
    - Induced magnetization, a local anomaly, depends on properties of underlying rock that can have its own magnetic field. Exist only while a magnetic field exist and aligned in the direction of recent field.
    - Remnant magnetization, a normal anomaly, is a presence of remanence makes interpreting magnetic anomalies more uncertain because its direction is generally unknown that related to the inclination and declination of a survey station.

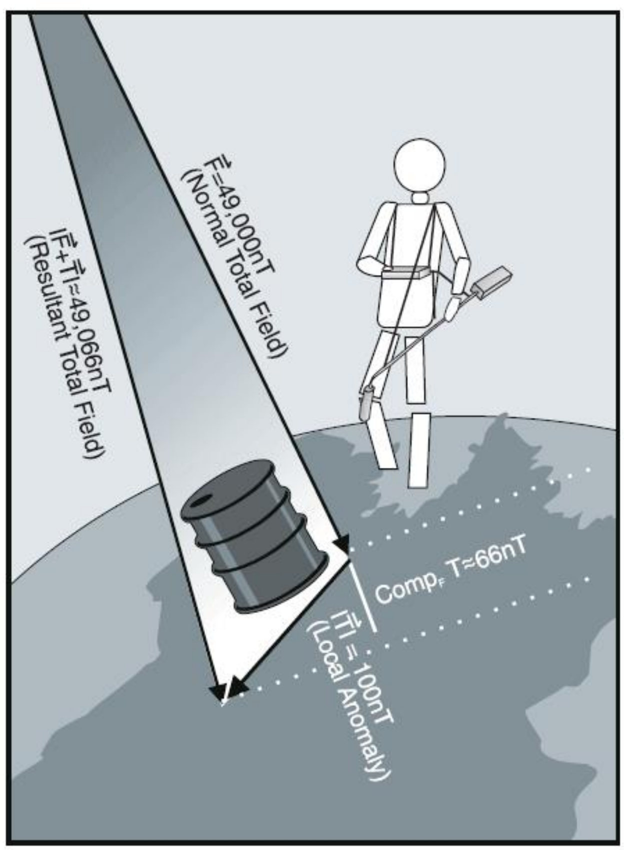


Figure 5.17: *The Earth's magnetic field interacting with a magnetized material (Scietrex Limited, 1996).*

5.6. Magnetic data reduction and magnetic anomaly

Reduction of the observations is much simpler than for gravity, but the filtering technique is less straight forward compared to gravity.

5.6.1. Time variation corrections

- The Earth's magnetic field varies with time. The variation is depended upon the frequency, duration and intensity of these fluctuations. Short term temporal variations in the field can be as large as those caused by underground structures.
- If magnetic measurements are to be interpreted for structure, they must first be corrected for the temporal variations. Once the magnetic field has been corrected, then anomalies can be interpreted.
- This requires the use of a fixed base station against which we can compare the area readings.

Diurnal variation corrections

- Fluctuations with a period lasting of several hours to one day are called diurnal variations.
- Diurnal variation are not predictive and are usually not a serious problem when conducting magnetic surveys that can be corrected via maths.
- This diurnal drift can cause a variation of the order of ±50 nT per hour (Figure 5.18A).
- Only correction has been done for time variation measurement.

Micro-pulsations

- Unpredictable short-term blips or spikes in the magnetic field are called micro-pulsations.
- These can range in intensity from a few through to ±10 nT (Figure 5.18B).
- These variations can present a problem when you are surveying in that they may appear similar to anomalies caused by buried objects.

Magnetic storms

- When the amplitude and duration of micro-pulsations becomes severe it is then called a magnetic storm.
- Magnetic storms are large in amplitude (± 100 – 1000 nT) and change very rapid that hard to be corrected (Figure 5.18C).

- Strong recommendation to do not conduct a total-field survey during a magnetic storm, as you may not be able to remove all of the rapidly changing variations in the magnetic field.
- Several agencies provide magnetic activity forecast information. Better to check the forecast before deciding to collect the magnetic data.

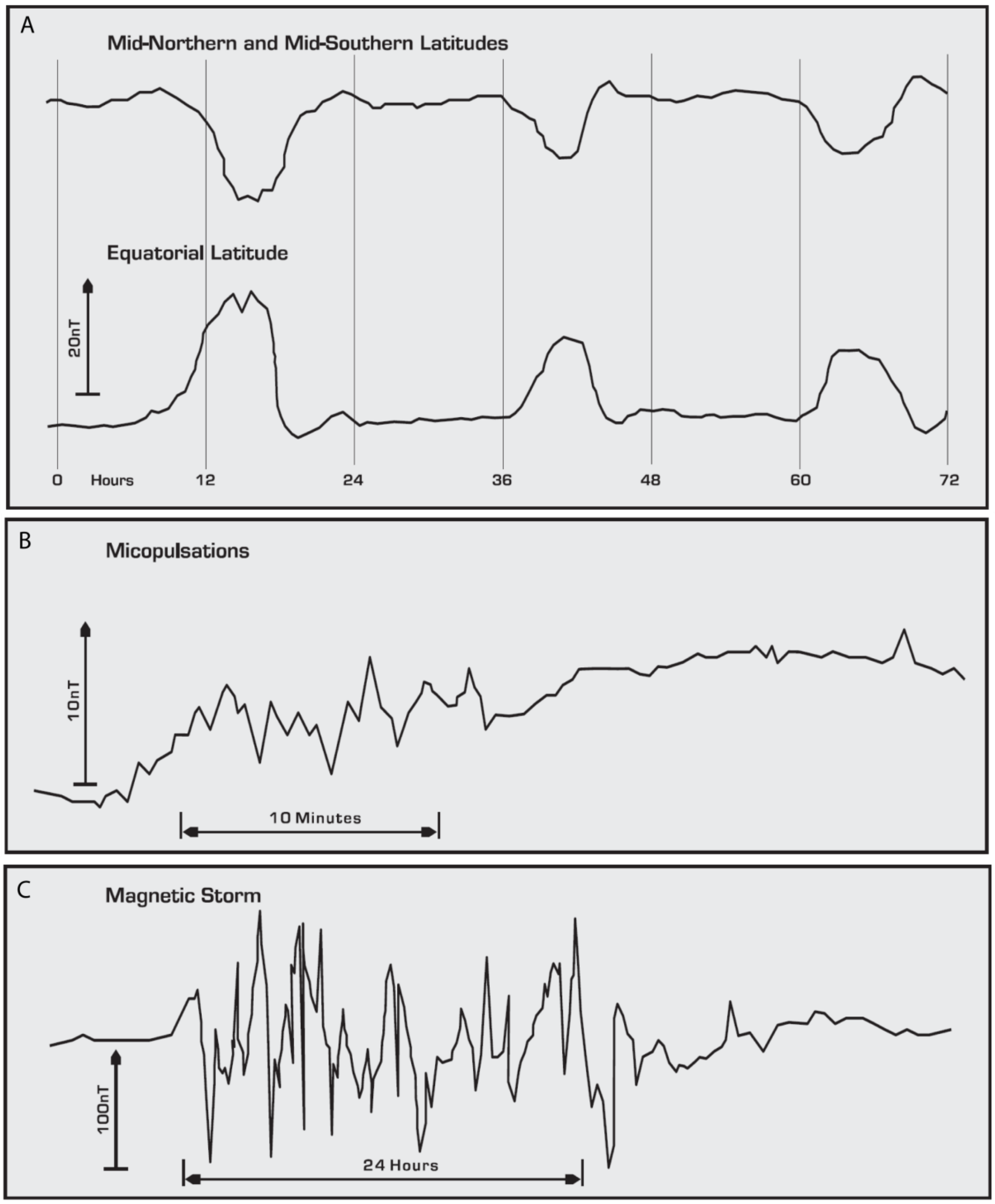


Figure 5.18: *A) Typical variation. B) Typical micro-pulsations. C) Typical magnetic storm (modified from Scietrex Limited, 1996)..*

5.6.2. Elevation and terrain corrections

- The gradient of the magnetic field is around 0.03 nTm^{-1} at the pole and -0.015 nTm^{-1} at the equator. No elevation correction is applied for ground survey.

5.6.3. Latitude corrections

- This correction equivalent of the latitude correction in gravity method.
- We can use the International Geomagnetic Reference Field (IGRF), updated every 5 years, which defines the theoretical undisturbed magnetic field at any point of the Earth surface.
- Note that: the IGRF is imperfect and in areas remote from observatories can be substantially in error.
- Alternative method for small surveys: use a trend analysis, where the regional field is approximated by a linear trend (Figure 5.19).

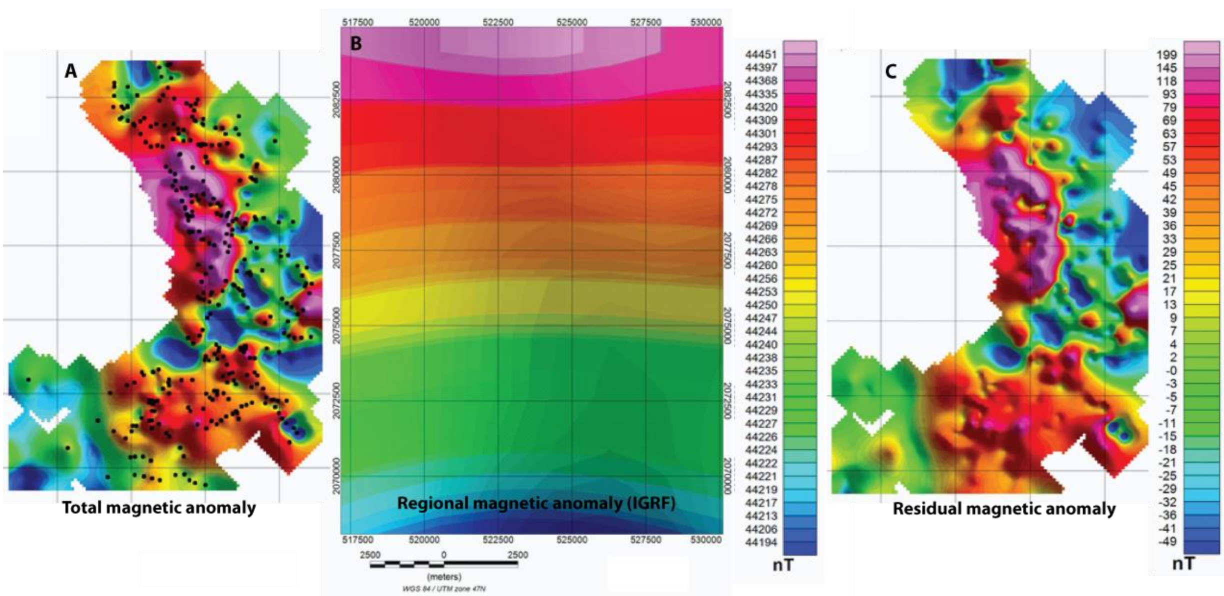


Figure 5.19: *A) Total magnetic intensity map of Mae On area, Chaing Mai. B) The trend analysis calculated from IGRF is applied for Latitude corrections. C) Residual magnetic anomaly after applying Latitude corrections (modified from* Mankhemthong et al., 2016).

5.6.4. Magnetic anomaly

- Normally, magnetic anomalies are residuals anomalies from the regional magnetic field. They may be positive or negative relative to the regional field.

- No simple correlation with lithology can be extremely detailed.
- It is assumed that these anomalies are caused by buried structures with increased magnetic susceptibility due to iron objects.
- The strength and shape of the anomaly is controlled by the shape, composition and depth of the structure.
- Benson et al., 1982 have calculated that the total field response in nT for different target distance and mass (Figure 5.20). It can be slightly misleading due to surrounding all man-made metallic objects.

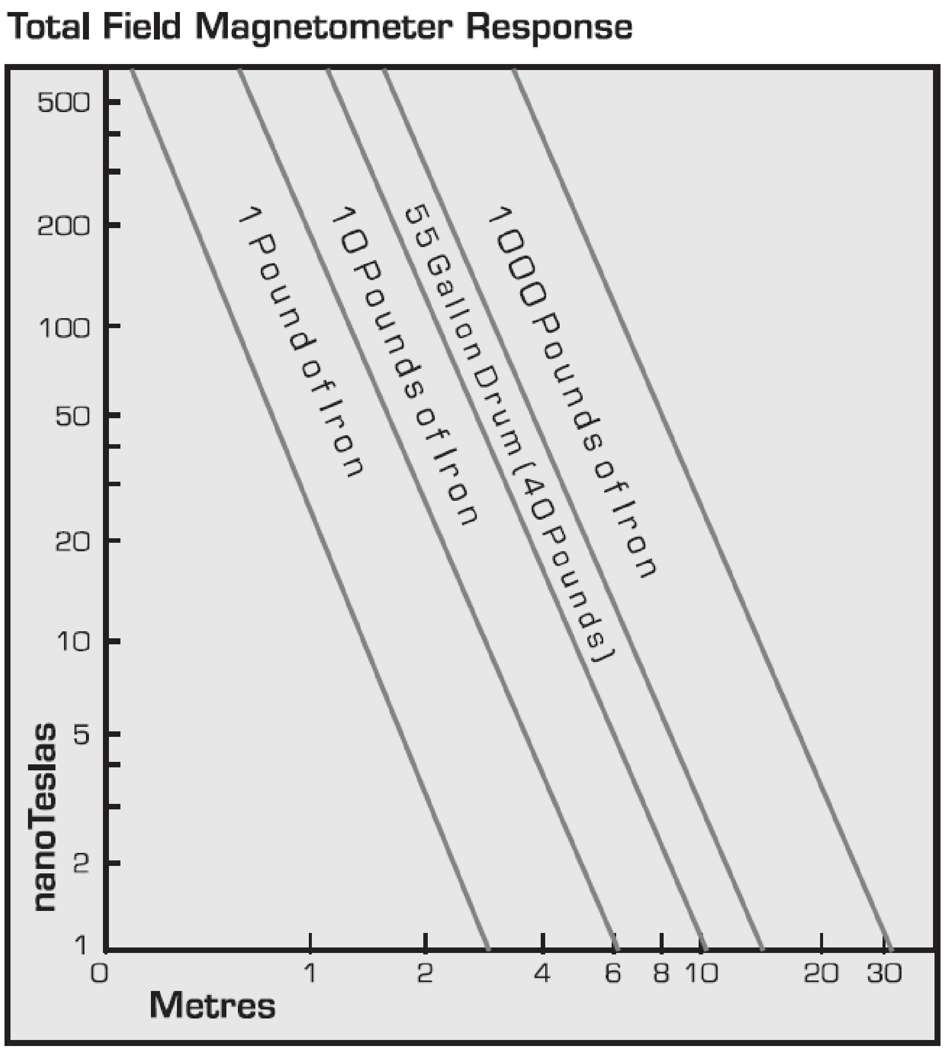


Figure 5.20: *Total magnetic field response for different target distance and mass (Benson et al., 1982).*

5.7. Magnetic data filtering techniques

5.7.1. Residual and regional magnetics

- Magnetic variation or susceptibility may be analyzed using either total intensity or residual maps. Once corrected, the field can then be divided into the average (regional) field and the local (residual) field.
- Residual anomaly is what remains after regional magnetic trends are removed from the total intensity Residual maps show local magnetic variations, which may have exploration significance.
- Residual anomaly reveals more detailed geologic features. It brings out the subtle magnetic anomalies that result from the changes in rock type across basement block boundaries.

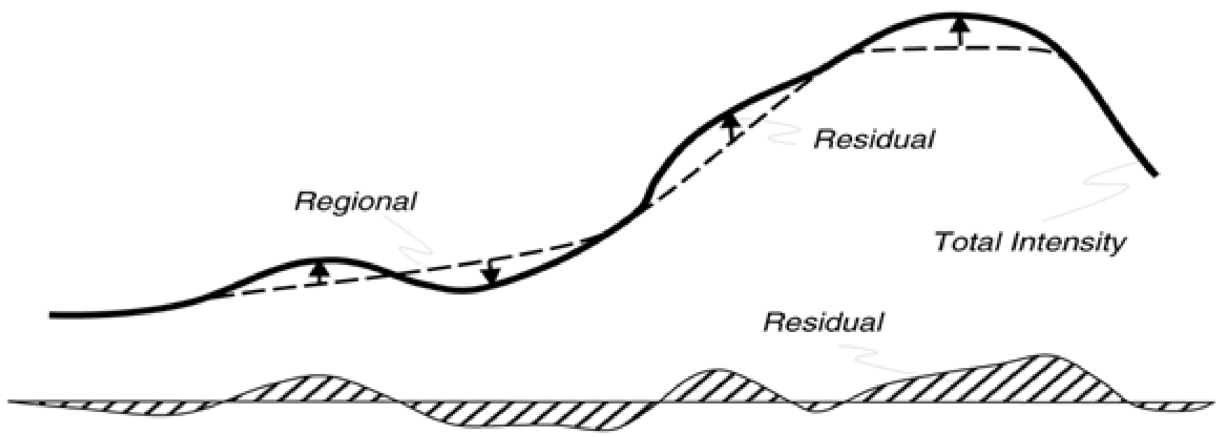


Figure 5.21: *Total magnetic intensity (bold solid), regional (dashed) and residual (regular solid) anomalies of magnetic data (Musset and Khan 2000).*

5.7.2. Effect on anomaly due to orientation not a latitude location

- Unlike gravity, variation with profile direction, body orientation and inclination.
- Note the change in shape of the anomaly with change in profile direction and different inclinations.
- There can be quite large magnetic anomalies at the edge of magnetic bodies.
- Away from the poles, likely to have positive and negative anomaly values (Figure 5.22a, b).
- Thus, latitude is matter in magnetic anomaly pattern.

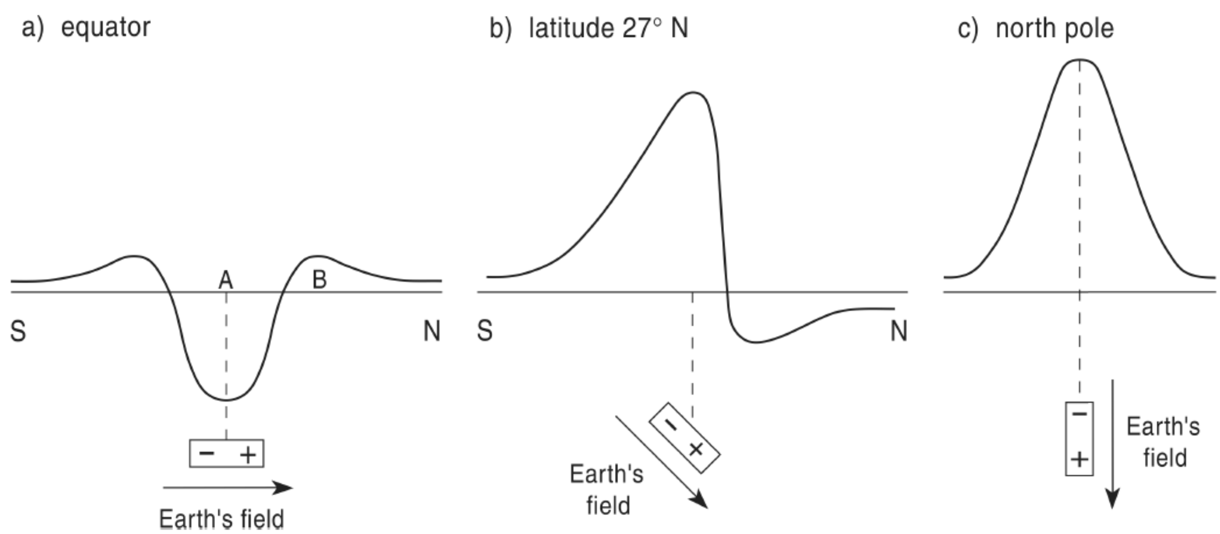


Figure 5.22: *Anomaly of a dipole at different latitudes (Musset and Khan 2000).*

5.7.3. Magnetic reduction to pole

- Magnetic anomalies at any latitude location are much less straight forward than gravity due to the di-pole property.
- The simplification of magnetic anomalies by modifying the anomaly pattern would be in a vertical field, i.e. if locality were at North/South magnetic pole.
- Induced magnetic effects after filtered would then be symmetric anomalies.
- In theory, the anomaly after the magnetic reduction to pole applied is directly analogous to that of a gravity anomaly.

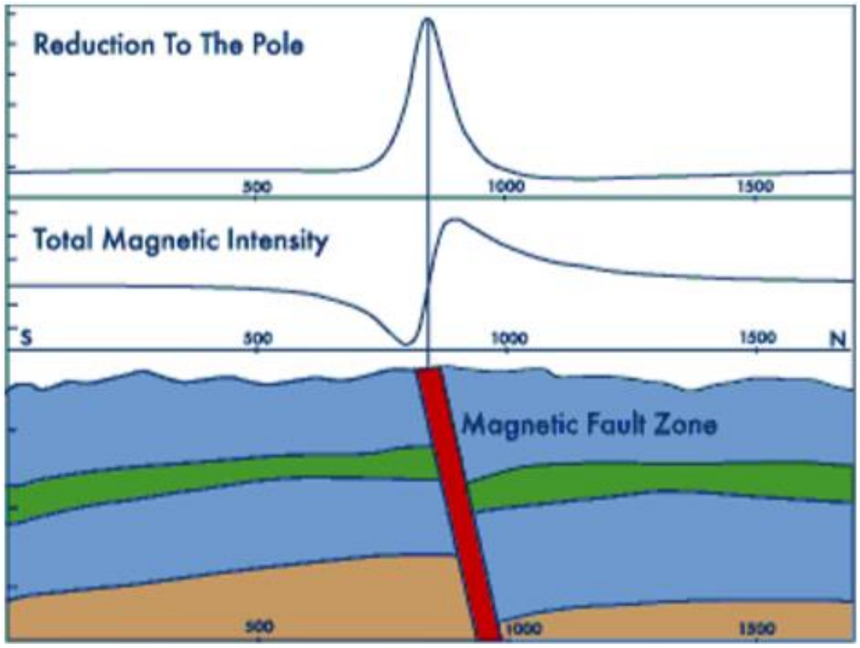


Figure 5.23: Magnetic profile of total magnetic intensity and anomaly after the Reduction to pole correction (www.gravmag.ou.edu).

5.7.4. Magnetic gradients

- Magnetic gradient techniques provide a different way of viewing the data, which has the potential for revealing otherwise, unnoticed features.
- A magnetic anomaly that changes in strength in a given direction. They emphasize the edges of anomalies.
- Gradient techniques may be done in either the spatial or the frequency domain.
- Gradient techniques can be done in any direction. 1D magnetic field gradient is a variation with respect to one direction, while a 2D gradient is a variation with respect to two directions.
- In the case of near-vertical geological boundaries, the maximum horizontal gradient crossing over the boundary would give better solution (Figure 5.24).

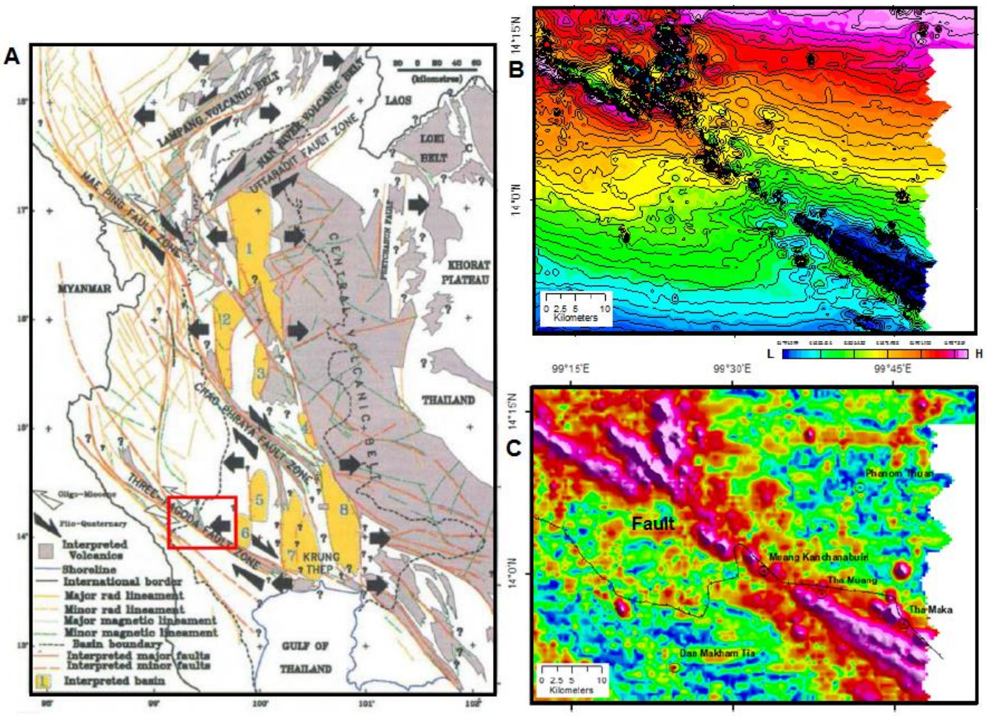


Figure 5.24: *(A) A sketched structural map of the Central Plain area. A red rectangle indicates an area of (B) aeromagnetic anomaly and (C) Analytic signal of fault indication in Kanchanaburi area (modified from Tulyatid and Rangubpit, 2016).*

5.7.5. Pseudogravity transformation

- This method involves calculating the equivalent gravity field for a magnetic field, and then interpreting this gravity field (Figure 5.25).

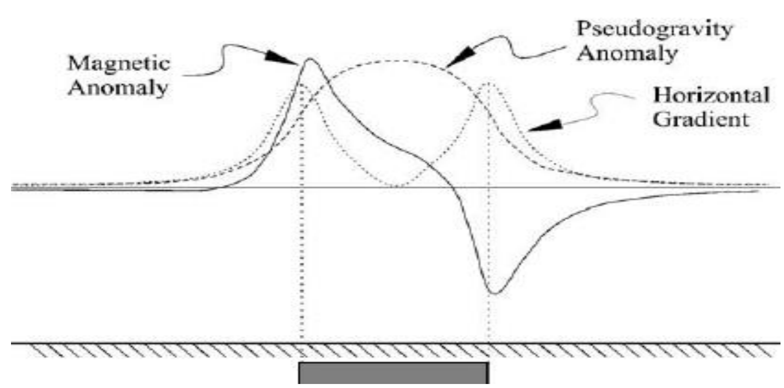


Figure 5.25: *Sketch of the relation between the magnetic anomaly, pseudogravity anomaly and the horizontal gradient anomaly given a magnetic body (grey block) with susceptibility* (https://www.researchgate.net/).

5.7.6. Upward and downward continuation

- Upward continuation suppresses the signals due to small shallow bodies, just as taking the second derivative enhances them.
- The upward continuation is most useful when applied to ground data so they may be compared with aeromagnetic data and to determine the depth to the basement.
- Note that: downward continuation is problematic because noise will blow up exponentially, and if the data are continued down past some body, a meaningless answer will result. Thus, this process must be done carefully, using low-noise data, and in a known situation of target.

5.8. Magnetic data interpretation and modeling

- The shape of the anomaly across a magnetic body varies with:
  - The shape of the body.
  - The magnetic properties of the body.
  - The direction of the survey.
  - The depth of the body.
- The positive anomalies in this case are largely due to the presence of magnetite (iron oxide) as either: A mafic igneous mineral (e.g. basalts etc.) (Figure 5.26).
- Granites, clays and limestones provide negative or no anomalies.
- Several parameters can vary in interpretation – need other constraints (e.g. fault map (Figure 5.26), hot spring locations) to come up with a good interpretation.
- At the elementary level, the detection and inspection of anomaly patterns ~~is~~ are helpful when combined with other data.

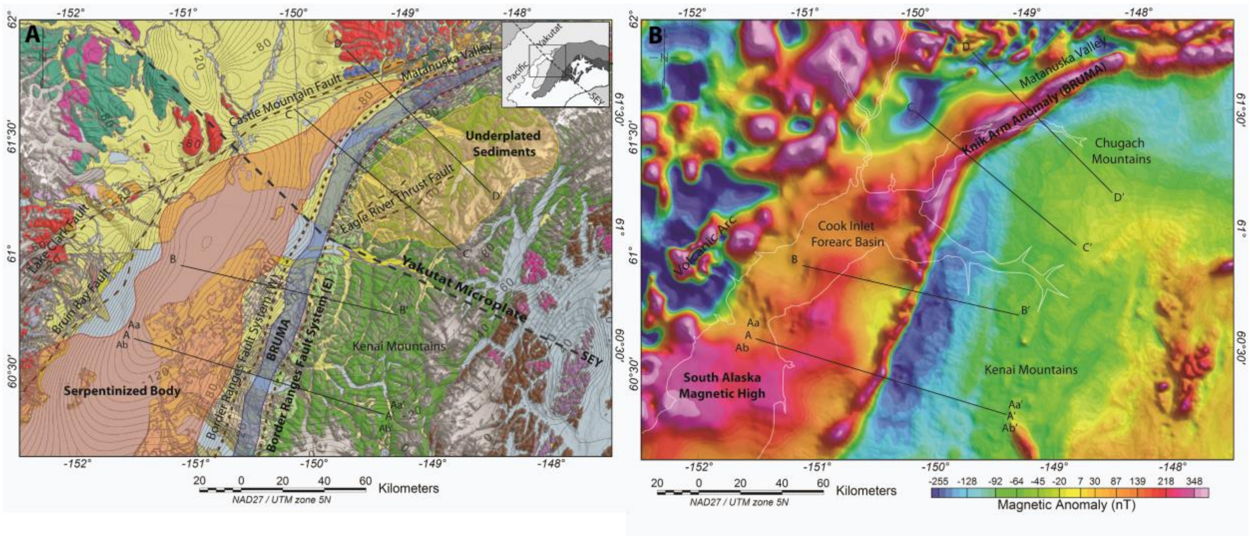
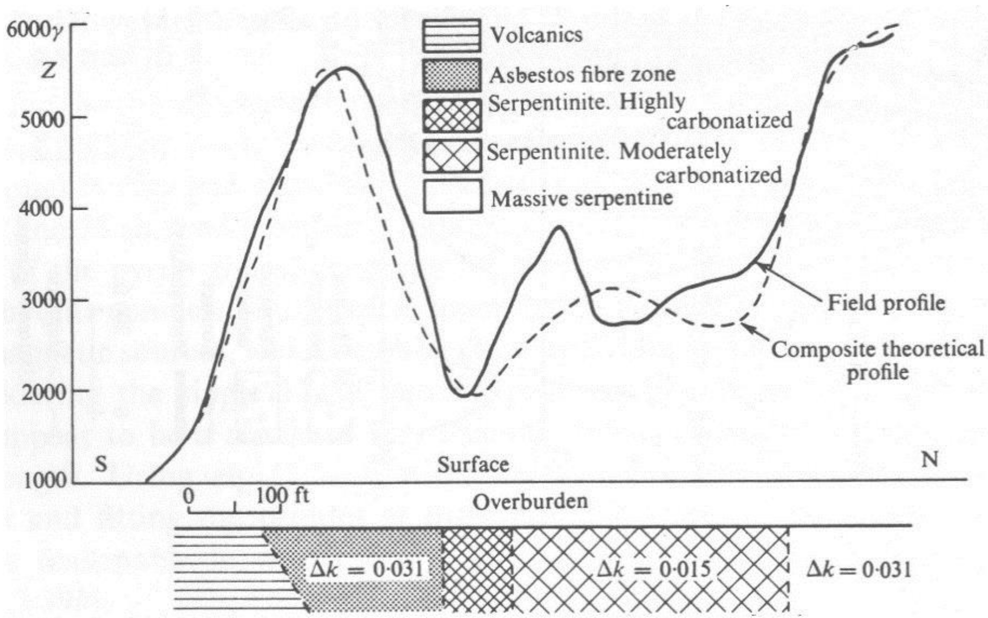


Figure 5.26: *A) Interpretations the Border Range ultramafic and mafic assemblages and associated the Border Range fault system over an area of (B) magnetic anomaly highs in Kenai Peninsula, Alaska, USA (modified from Mankhemthong et al., 2013).*

- The procedure for magnetic modeling is close to that one described in the gravity part.
- The calculation of theoretical anomalies is more ambiguous (due to the number of variables) but can yield some insights, especially when constrained by alternative geophysical techniques.
- The models are usually built as a set of polygonal bodies with constant physical parameters of susceptibility (k) (Figure 5.27).
- Constructed magnetic model can be varied systematically to match the calculated anomaly profile to the observed one, using other data (geological maps, boreholes etc.) as a guide.
- The model is more ambiguous (due to the number of variables) as described in Gravity Chapter but can yield some insights, especially when constrained by alternative geophysical techniques (Figure 5.28).
- The good result would show the best-fit numerical interpretation. It is, however, only feasible for relatively simple structures (Figure 5.29).
- Because of more complex anomalies then a simple homogeneous body are almost exclusively modelled using computers.



Figures 5.27: *Example of 2D magnetic modelling and interpretation of a serpentine ore body (modified from Robinson and Crouch, 1998).*

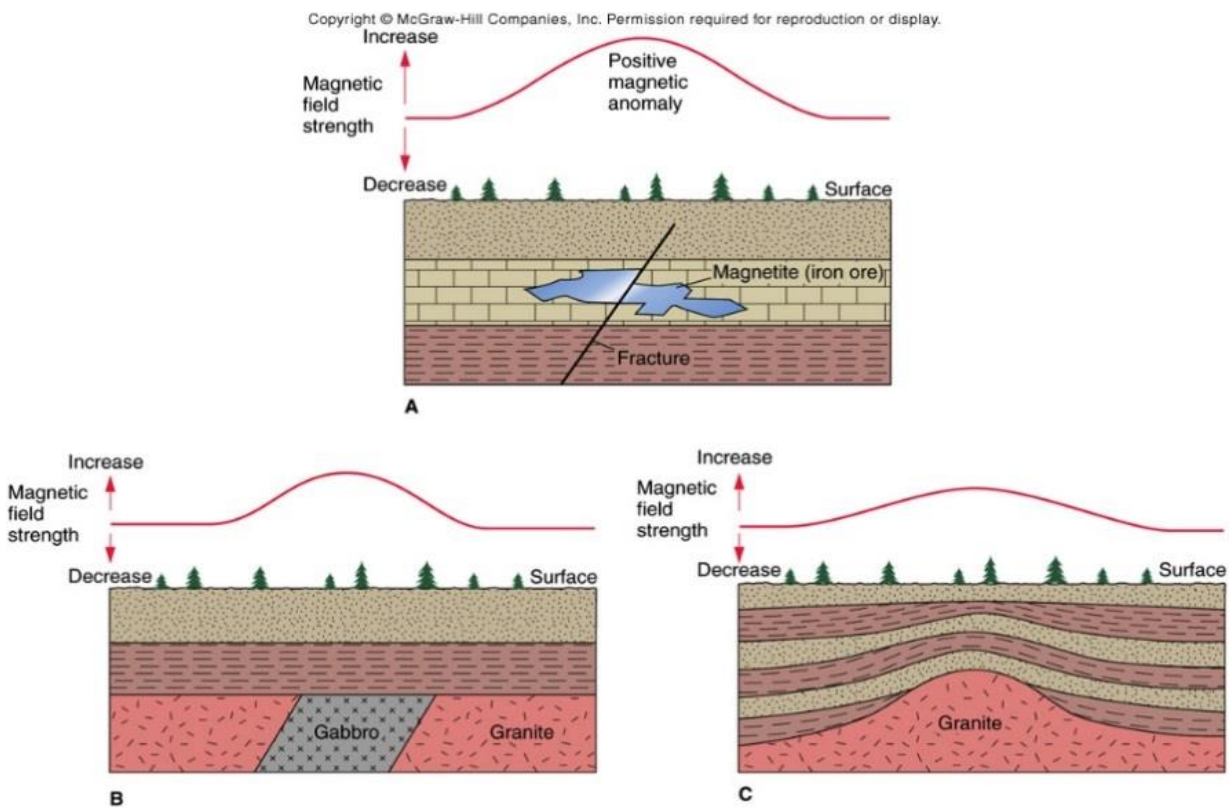


Figure 5.28: *2D magnetic modeling shows ambiguity in geological Interpretations (www. slideshare.net/kreston2/9-magnetic-introduction).*

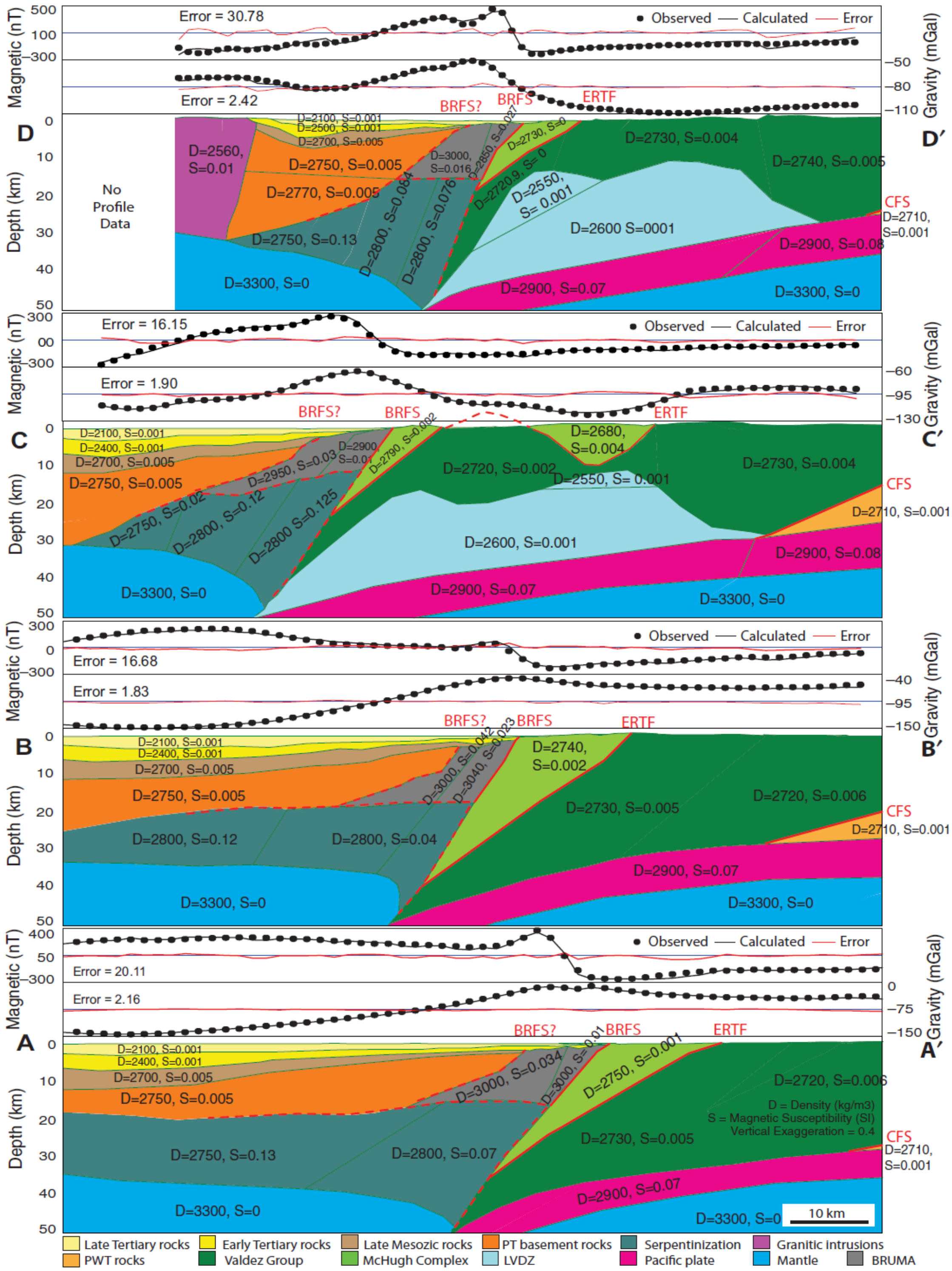


Figure 5.29: *2D integrated gravity and magnetic models across the Cook Inlet forearc basin, Alaska, USA. The profile locations are shown in Figure 5.36A (Mankhemthong et al., 2013).*

5.9. Applications in magnetic method

5.9.1. Archeological application

Title: Magnetic survey of archaeological kiln sites with Overhauser magnetometer (Hatakeyama et al., 2018):

Purpose: to conduct magnetic surveys using an Overhauser magnetometer, a fast-sampling, high-accuracy magnetometer, at two archaeological sites to detect magnetic anomalies associated with buried climbing Sue ware kilns in Japan.

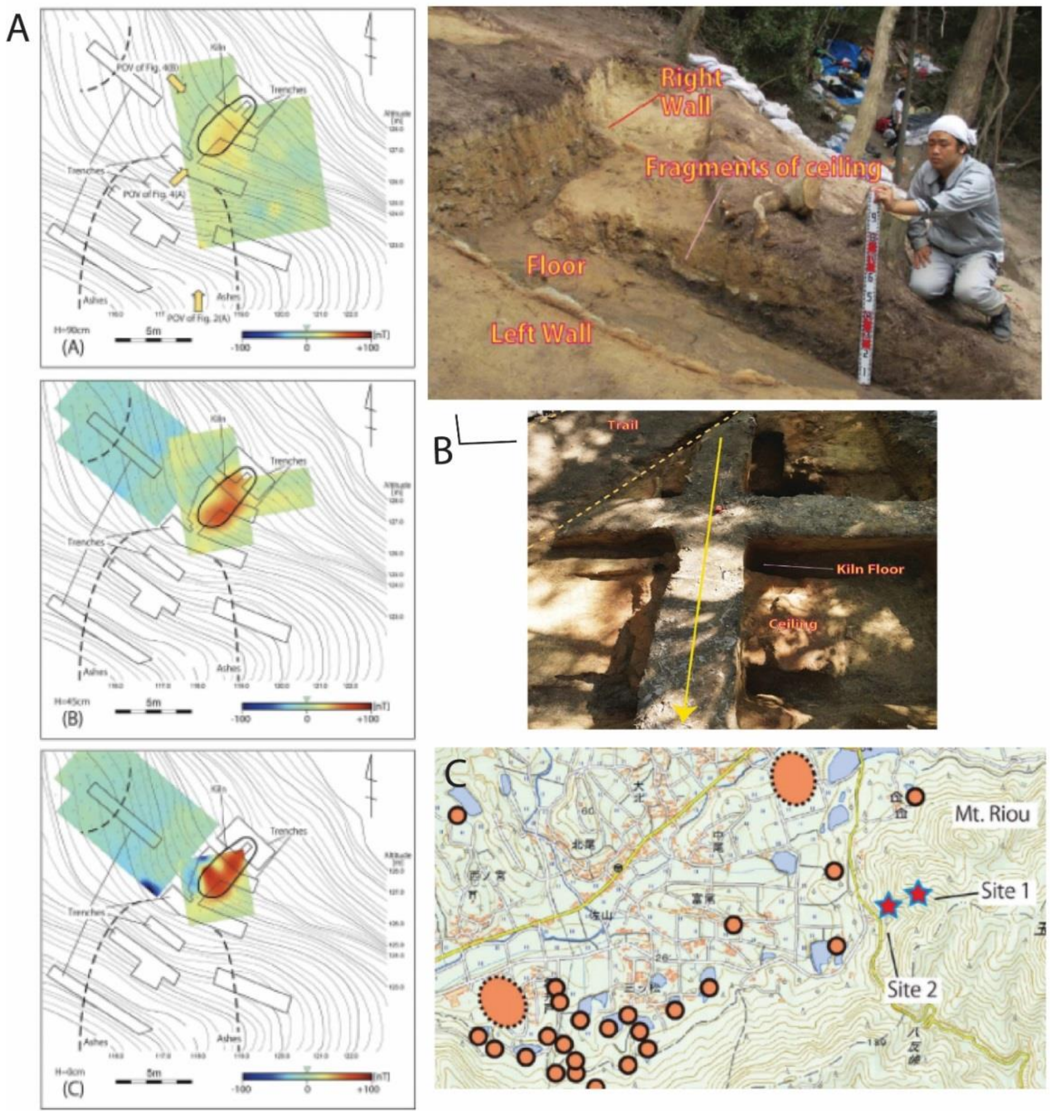


__Figure 5.30:__ A) magnetic anomalies, and excavation maps (rectangle) with the horseshoe-shaped outline shows positions of discovered kilns. B) Photographs of buried old kiln during the excavation show that the kiln floor is located 20–50 cm below the ground surface. C) Location of survey site (1 and 2) with records of old kilns (modified from Hatakeyama et al., 2018).

5.9.2. Geothermal application

Title: 2D resistivity imaging and magnetic survey for characterization of thermal spring: A case study of Gergedi thermal spring in Main Ethiopian Rift, Ethiopia *(*Abdulkadir and Eritro*,* 2017*).*

Purpose: Magnetic surveys were carried out at thermal springs, located in the Main Ethiopian Rift where the SW -NE trending structures of the Wonji Fault Belt System cut across the survey area, to characterize the geothermal condition.

Summary: the magnetic surveys have revealed new potential sites with magnetic anomaly highs that can be used for further exploitation of the thermal resources.

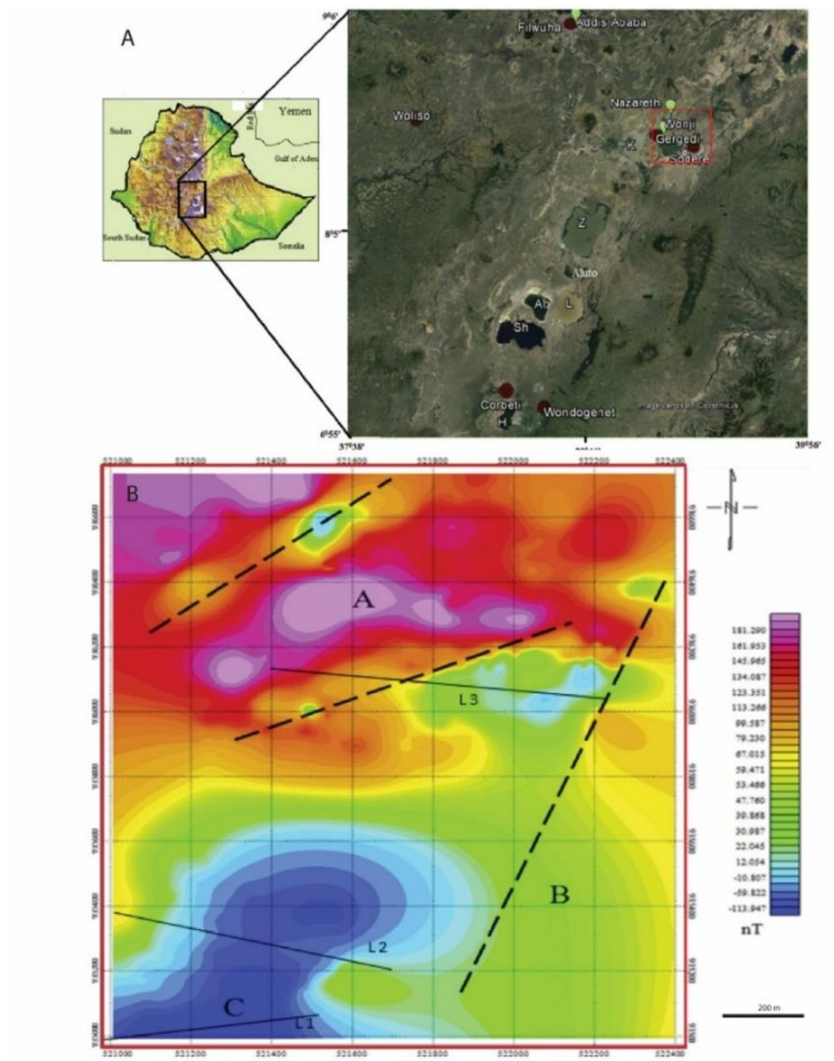


Figure 5.31: *A) A location of known thermal centers, Rift Valley Lake. B) Magnetic intensity highs (A, B, and C) could be the responses from rocks which have a high magnetic susceptibility. This high magnetic anomaly zone trends in the SW to NNE direction following the regional fault patterns (modified from Abdulkadir et al., 2017).*

5.9.3. Hydrogeological application

Title: Investigations of ground water flow associated with the Saratoga warm springs and Tecopa Hot Springs near Death Valley, California, using magnetic and conductivity methods (Wamalwa et al., 2010).

Purpose: To investigate the geologic and structural controls on the springs' locations, using 282 ground magnetic data in an area of Saratoga Springs in south of Death Valley.

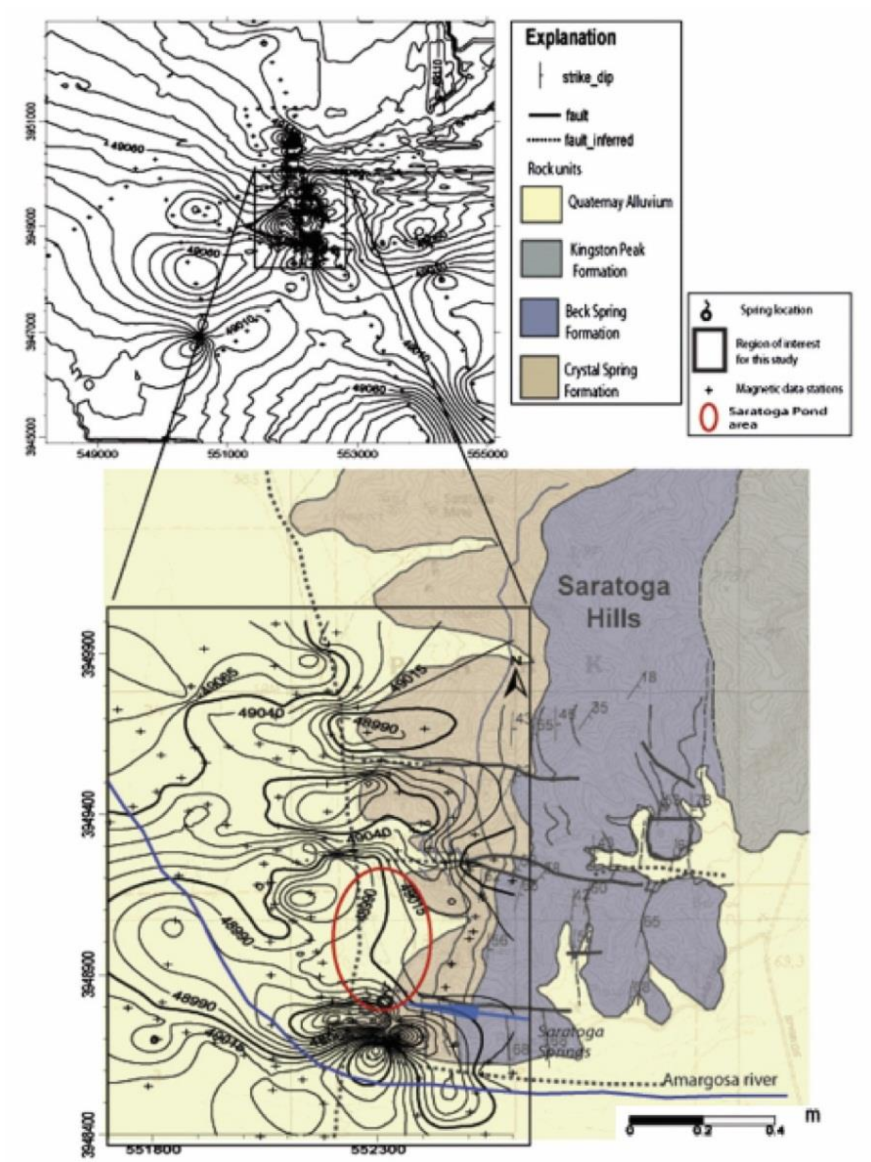


Figure 5.32: *Magnetic map with stations (dot) shows anomaly highs correlates with the lower Crystal Springs Formation, indicating that Crystal springs formation hosts magnetic minerals such as magnetite._The high anomalies south of the spring are due to the presence of a diabase at shallow depth (modified from Wamalwa et al., 2010).*

5.10. Exercises for magnetic method chapter

1) Sketch the anomaly of buried sphere with induced magnetization at the south magnetic pole.

2) Compare and contrast the techniques of interpretation of gravity and magnetic anomalies.

3) How are the magnetic field values at equator compared to that at the pole?

4) Could a total field magnetic survey detect the burial chamber (spherical void) illustrated here in a region? Where Earth magnetic field is 55000 nT and $i = 70^{\circ}$.

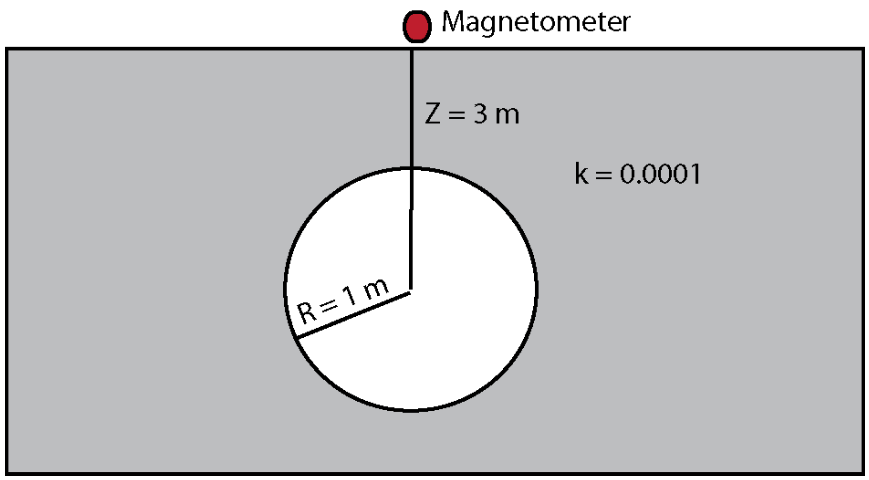


Figure 5.33: *Buried magnetic body.*

5) Describe relationship between the magnetic provinces (Figure 5.34) and its geologic setting.

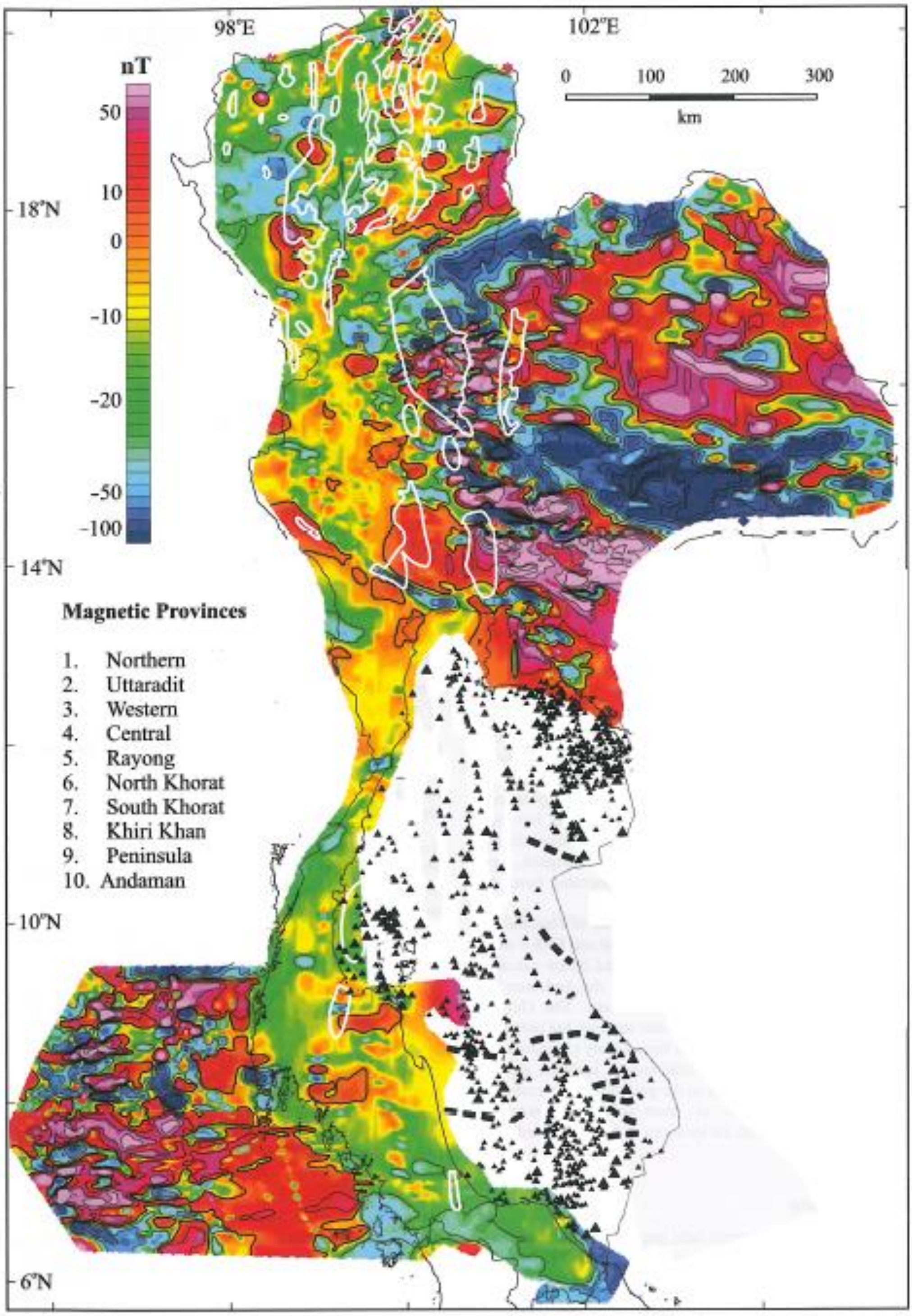


Figure 5.34: *IGRF-corrected aeromagnetic data in Thailand (Milsom, 2011).*

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Chapter 6: Seismic Method

6.1. Introduction to seismic method

- In seismic method, seismic waves are created by a controlled source and propagate through the subsurface.
- Some waves will return to the surface after refraction or reflection at geological boundaries within the subsurface.
- Instruments distributed along the surface detect the ground motion caused by these returning waves and hence measure the arrival times of the waves at different ranges from the source.
- These travel times may be converted into depth values and, hence, the distribution of subsurface geological interfaces may be systematically mapped.

Earthquake seismology

Recordings of distant or local earthquakes are used to infer earth structure and faulting characteristics.

Applied seismology (seismic exploration)

A signal, similar to a sound pulse, is transmitted into the Earth. The signal recorded at the surface can be used to infer subsurface properties.

6.1.1. History of seismology study

Exploration seismic developed from early work of earthquake studies

- 1846: Robert Mallett, made first use of an artificial source in a seismic experiment.
- 1888: August Schmidt used travel time vs. distance plots to determine subsurface seismic velocities.
- 1899: G.K. Knott explained refraction and reflection of seismic waves at plane boundaries.
- 1910: A. Mohorovicic identified separate P and S waves on travel time plots of distant earthquakes, and associates them with base of the crust, the Moho.
- 1916: Seismic refraction developed to locate artillery guns by measurement of recoil.
- 1921: ‘Seismos’ company founded to use seismic refraction to map salt domes, often associated with hydrocarbon traps.

- 1920: Practical seismic reflection methods developed. Within 10 years, the dominant method of hydrocarbon exploration.

6.2. Elastic wave theory

- Elastic wave is a motion in a medium in which, when particles are displaced, a force proportional to the displacement acts on the particles to restore them to original position.
- If a medium material has the property of elasticity and the particles in a certain region are set in vibratory motion, an elastic wave will be propagated.

6.2.1. Stress and strain

- Stress is the ratio of applied force (F) to the area (A) across which it acts which composes of normal stress and shear stress (Figure 6.1).
- Strain is the deformation caused in the body and is expressed as the ratio of change in length (or volume) to original length (or volume) (Figure 6.1).

6.2.2. Hooken Behavior

- Study a linear relation between stress & strain and how solid objects deform and become internally stressed due to prescribed loading conditions.
- At low to moderate strains: Hooke's Law applies and a solid body behaves which will return to original form when stress removed.
- At high strains: the elastic limit is exceeded and a body deforms in a plastic or ductile manner: it is unable to return to its original shape, being permanently strained, or damaged.
- At very high strains: a solid will fracture, e.g. in earthquake faulting.

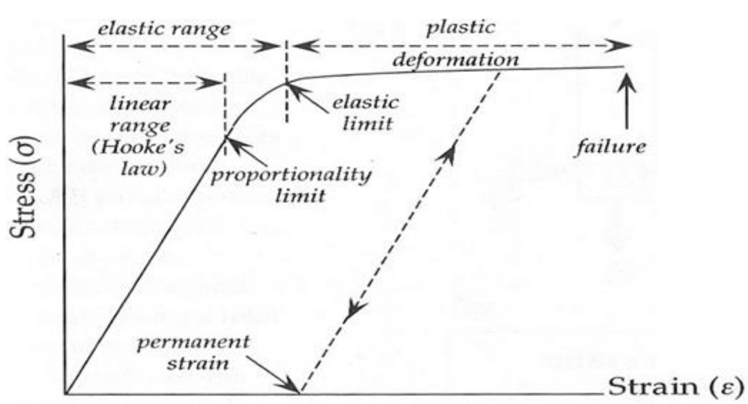
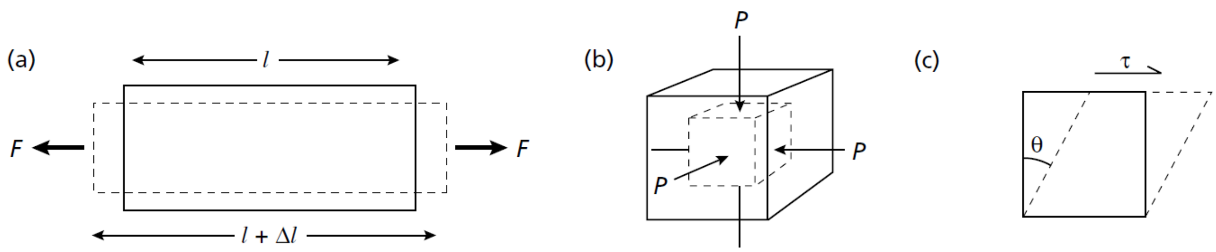


Figure 6.1: *Graph relationship between stress and strain (http://nptel.ac.in/courses/ 105108072/ mod02/lec2.html).*

6.2.3. Elastic constant

Constant of proportionality is called the modulus

- **Young's modulus ($\boldsymbol{\varepsilon}$)** = Proportionality of stress and strain change (how a length change) (Figure 6.2a).
- **Bulk modulus (K)** = Proportionality of stress and strain (how the volume change) (Figure 6.2b).
- **Shear modulus/rigidity ($\boldsymbol{\mu}$)** = How the object is <u>possible to deform </u>by shear stress. *In gas and liquid phases (Figure 6.2c), $\boldsymbol{\mu}$ = 0*
- **Poisson's ratio ($\boldsymbol{\sigma}$)** = Proportionality of <u>max and min strains</u>



$$E = \frac{\text{longitudinal stress } F/A}{\text{longitudinal strain } \Delta l/l} \qquad K = \frac{\text{volume stress } P}{\text{volume strain } \Delta v/v} \qquad \mu = \frac{\text{shear stress } \tau}{\text{shear strain } \tan\theta}$$

Figure 6.2: *The elastic constants of (a) Young's modulus, (b) Bulk modulus, and (c) Shear modulus (modified from Kearey et al., 2002)*

6.2.4. Wave terminology

Figure 6.3 explains important seismic wave terminology including;

- Amplitude is the maximum departure of a wave from the average value.
- Peak is the highest point on a wave and the lowest point on a wave is called trough.
- Phase; the angle of lag or lead of a sine wave with respect to a reference.
- Period (sec/cycle) is the time (t) for one wave cycle.
- Frequency (cycle/sec or Hertz) is the repetition rate of a periodic waveform in one time unit.
- Wave number (cycle/distance) is the number of waves per unit distance perpendicular to a wave front that is reciprocal of the wavelength (for potential field).
- Wave speed/velocity depends on the medium that the wave is travelling through.

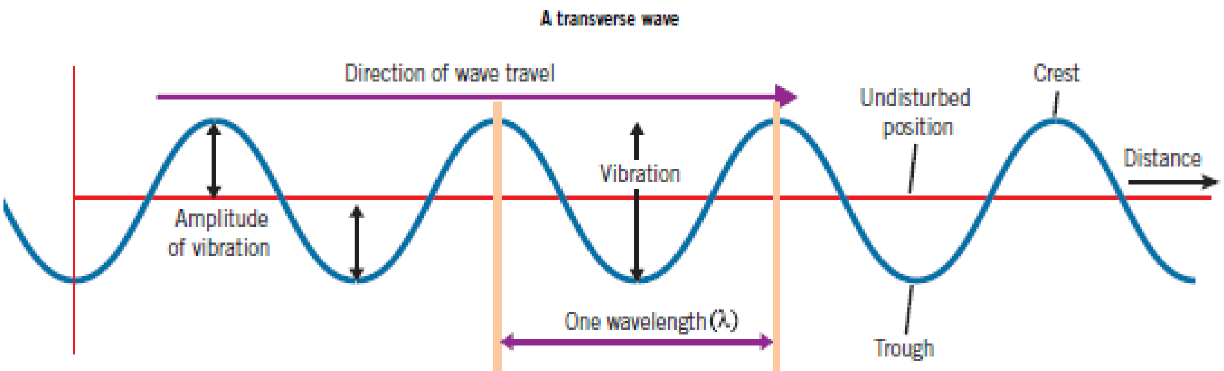


*Figure 6.3: Terms of transverse wave form (*https://revisionworld.com/gcse-revision/physics/waves/describing-waves*).*

- Wavefront is a surface containing points affected in the same way by a wave at a given time (Figure 6.4).
- Raypath is the path or direction along which wave energy propagates through medium (Figure 6.4).

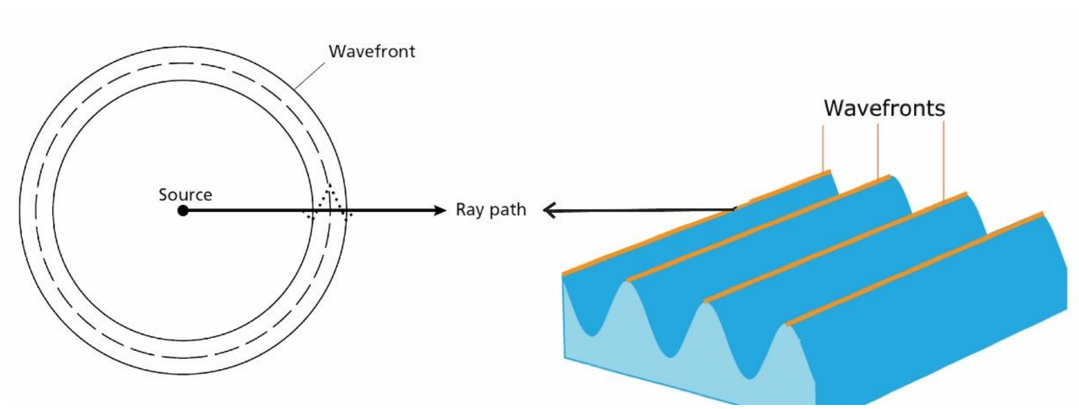


Figure 6.4: The relationship of a ray path to the associated wavefront (modified from Kearey et al., 2002).

6.2.5. Huygen's principle

- The Huygens-Fresnel principle states that every point on a wavefront (Figure 6.5, blue lines) is a source of wavelets (yellow points).
- These wavelets spread out in the forward direction, at the same speed as the source wave. The new wavefront (green lines) is a line tangent to all of the wavelets.

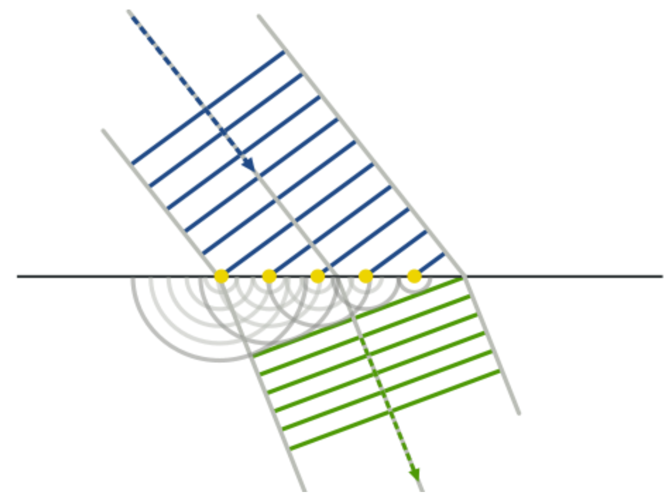


Figure 6.5: *Wave refraction in the manner of Huygen's Principle (https://en.wikipedia.org/wiki/Huygens%E2%80%93Fresnel_principle#/media/File:Refraction_on_an_aperture_-_Huygens -Fresnel_principle.svg).*

6.2.6. Diffraction

- Diffraction is the bending of waves as they pass through a gap or past an edge that similar in size to the wavelength of the waves (Figure 6.6).
- A secondary source follows Huygen's principle causes radial scattering of incident seismic energy.

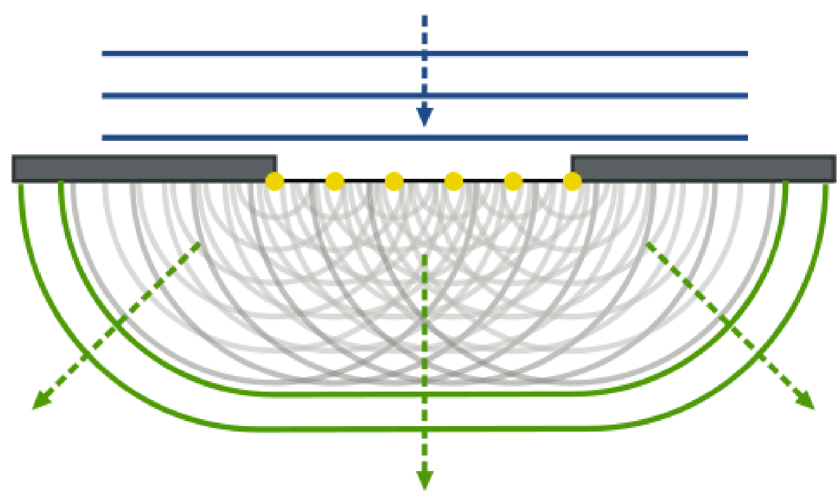


Figure 6.6: *Wave diffraction in the manner of Huygen's Principle (https://en.wikipedia.org/wiki/Huygens%E2%80%93Fresnel_principle#/media/File:Refraction_on_an_aperture_-_Huygens-Fresnel_principle.svg).*

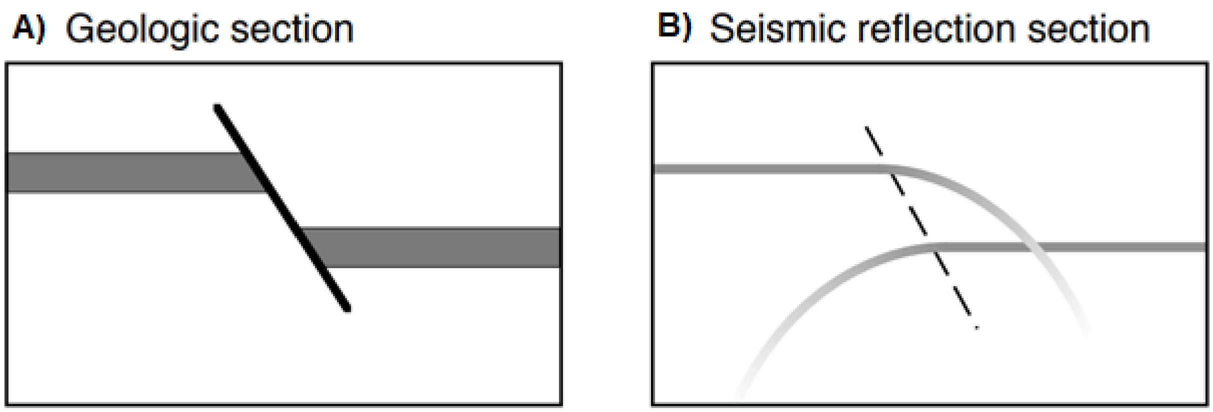


Figure 6.7: *A) geologic section of a high angle normal fault; B) schematic seismic reflection profile at the same location. The fault plane would not be imaged seismically but the truncations of the offset bed would generate the fault (dashed line) (*http://www.geo.cornell.edu/*geology/faculty/RWA/structure-lab-manual/chapter-11.pdf).*

6.3. Seismic wave properties

- Seismic waves are parcels of elastic strain energy that propagate outwards from a seismic source. Generally, seismic waves are sound waves travelling through the earth.
- Seismic waves transfer energy and information from one place to another, but they do not transfer material of the medium.
- Seismic wave is elastic wave because requires a medium to propagate.

6.3.1. Seismic wave types

- Body wave: Seismic waves are pulses of strain energy that propagate in solid Earth (Figure 6.8) comprising of Compressional wave (P-Wave) and Shear wave (S-Wave).
  - Compressional wave (P-Wave) shows a particle motion consists of alternating compression and dilation. Particle motion is parallel to the direction of propagation (longitudinal).
  - Shear wave (S-Wave) shows a particle motion consists of alternating transverse motion. Particle motion is perpendicular to the direction of propagation (transverse).
- Surface wave: Seismic waves that travel across the Earth's surface and have no stresses act on that surface (Figure 6.8). Rayleigh wave (R-Wave) and Love wave (L-Wave) are types of surface wave.

- Rayleigh wave (R-Wave) shows a particle motion consists of elliptical motions (generally retrograde elliptical) in the vertical plane and parallel to the direction of propagation. Amplitude decreases with depth.
- Love wave (L-Wave) shows a particle motion consists of alternating transverse motions. Particle motion is horizontal to ground surface and perpendicular to the direction of propagation (transverse).

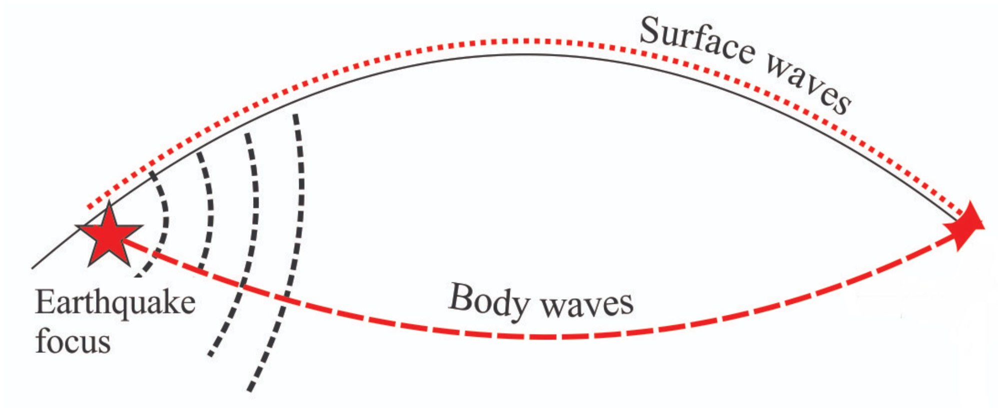


Figure 6.8: Body and surface wave propagations (*http://www.geological-digressions.com/?tag=body-waves*, 2017).

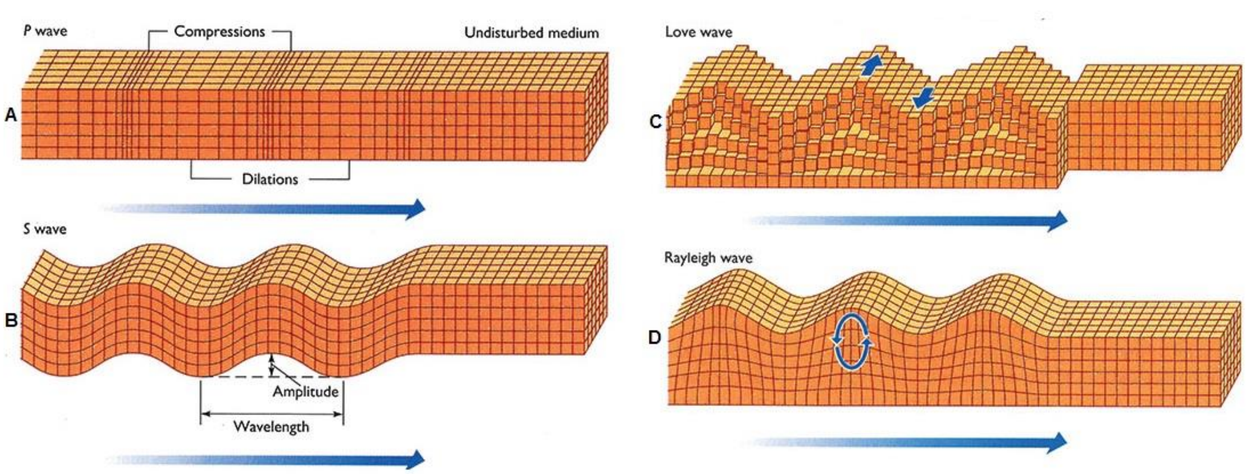


Figure 6.9: *Elastic deformations and ground particle motions associated with wave propagations. (A) P-wave, (B) S-wave, (C) Rayleigh wave and (D) Love wave (Bolt, 1982).*

6.3.2. Seismic wave velocity

- The seismic wave velocities of rocks are the velocities at which wave motions travel through them.

- The seismic wave velocities depend on elastic properties of medium rocks; how easy it is to strain the rock for a given stress.
- Density, porosity, aperture, and pressure are also factors causing change of the velocities.
- Figure 6.10 shows examples of *ranges of P-wave velocities in rocks.*

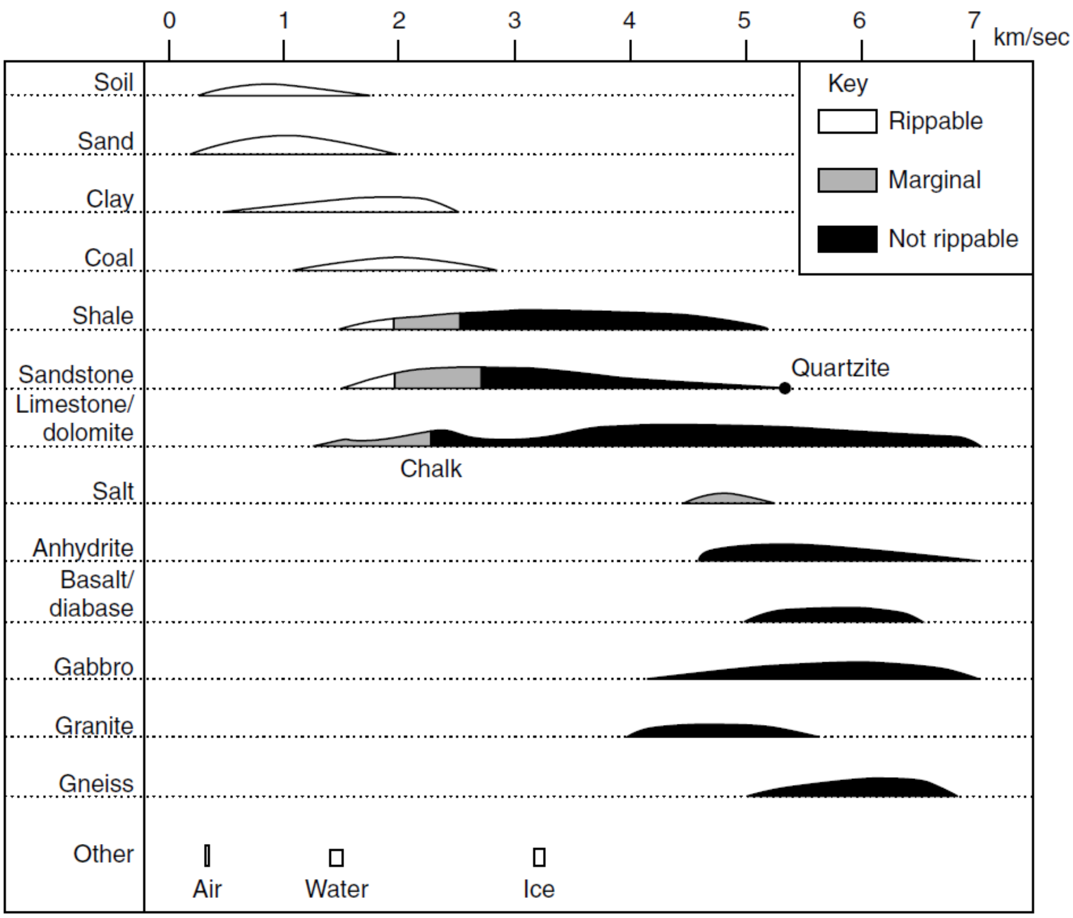


Figure 6.10: *Ranges of compressional wave velocities in typical Earth materials (Milson, 2003).*

6.3.3. Attenuation of elastic wave

- As a seismic wave propagates through a medium, the elastic energy associated with the wave is gradually absorbed by the medium, eventually ending up as heat energy. This is known as an elastic attenuation.
- The elastic attenuation will eventually cause the total disappearance of the seismic wave (Figure. 6.11).

- The attenuation is controlled by the temperature, composition, melt content, and volatile content of the rocks through which the waves travel.

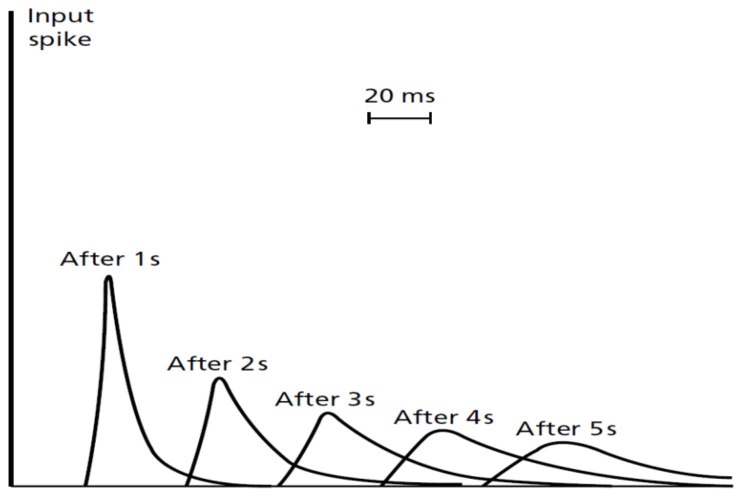


Figure 6.11: *Progressive change of shape of an original spike pulse during its propagation through* a medium *due to the effects of absorption (Anstey, 1977).*

6.3.4. Acoustic impedance and reflection coefficient

- Acoustic impedance is a product of density and seismic velocity, which varies among different rock layers, commonly symbolized by Z. The difference in acoustic impedance between rock layers affects the reflection coefficient (Figure 6.12).
- Reflection coefficient is a ratio of amplitude of the reflected wave to the incident wave, or how much energy is reflected (Figure 6.13). If the wave has normal incidence, then its reflection coefficient can be expressed as

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

Where, R = reflection coefficient, whose values range from −1 to +1
ρ_1 = density of medium 1
ρ_2 = density of medium 2
v_1 = velocity of medium 1
v_2 = velocity of medium 2

- Typical values of R are approximately −1 from water to air, meaning that nearly 100% of the energy is reflected and none is transmitted; ~0.5 from water to rock; and ~0.2 for shale to sand.

- Note that at non-normal incidence, the reflection coefficient defined as a ratio of amplitudes depends on other parameters, such as the shear velocities, and is described as a function of incident angle.

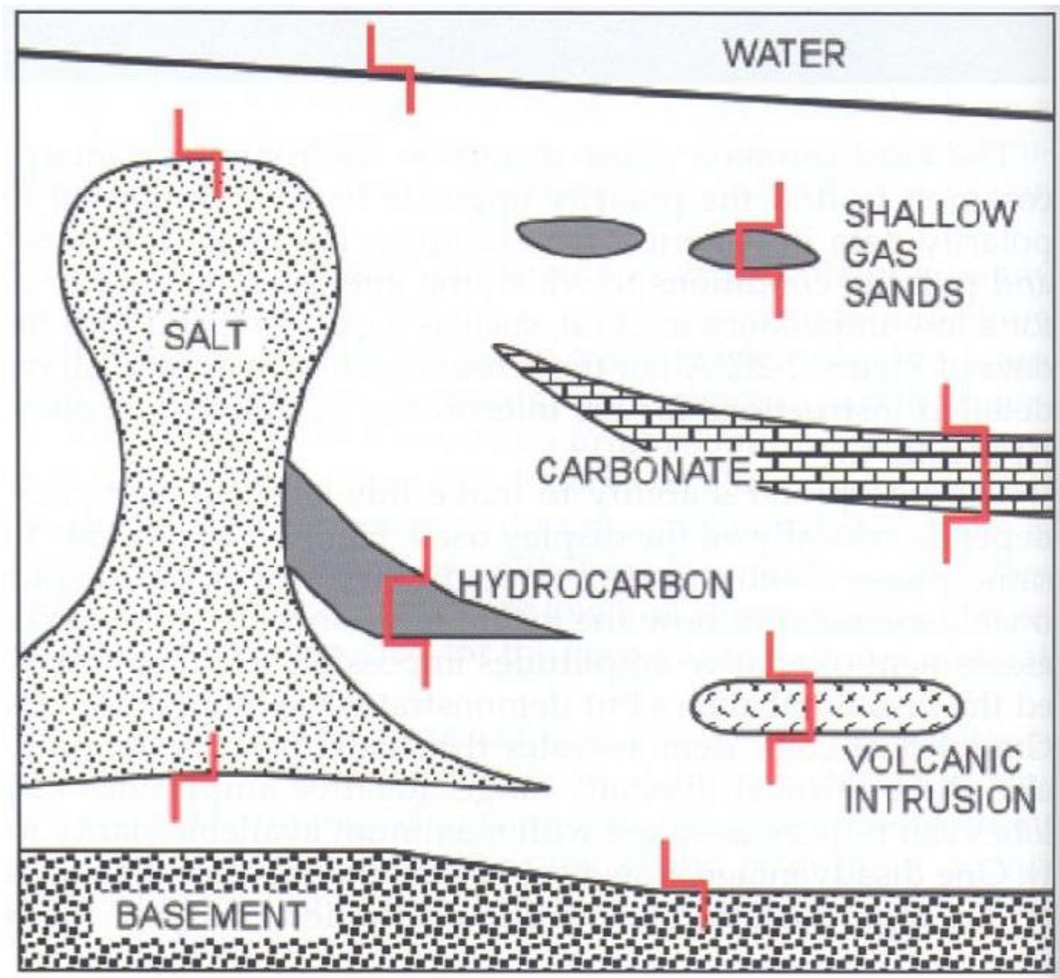


Figure 6.12: *Subsurface rocks with acoustic impedance contrast (Brown, 2010).*

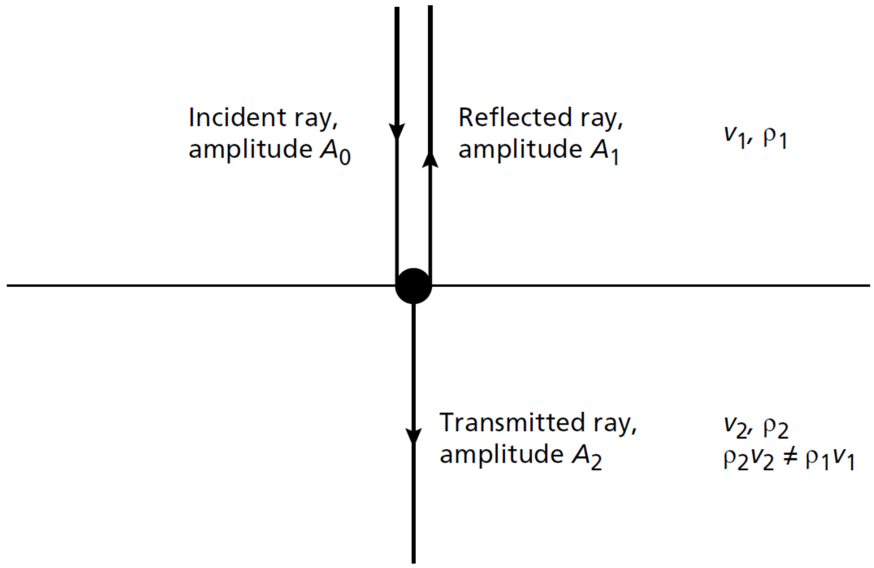


Figure 6.13: *Reflected and transmitted rays associated with a ray normally incident on an interface of acoustic impedance contrast (modified from Kearey et al., 2002).*

- Snell's Law is a formula used to describe the relationship between the angles of incidence and refraction, when referring to light or other waves passing through a boundary between two different isotropic media, such as water, glass, or air (Figure 6.14).
- Critical angle is the angle of incidence for which the angle of refraction is 90°. The angle of incidence is measured with respect to the normal at the refractive boundary. Consider a light ray passing from glass into air. It is at this point no light is transmitted into air (Figure 6.14).
- Total internal reflection is the phenomenon which occurs when a propagated wave strikes a medium boundary at an angle larger than a particular critical angle with respect to the normal to the surface (Figure 6.15)
    - The angle of incidence equals the angle of reflection.
    - The incident and reflected rays lie in the same plane with the normal.

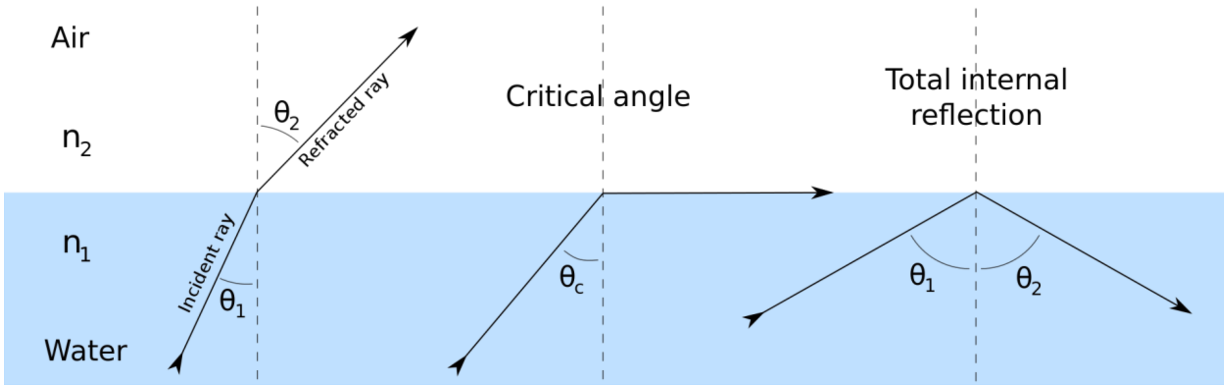


Figure 6.14: *Illustrations of refracted critical angle and total internal reflection https://en. wikipedia. org/ wiki/Total_internal_reflection#/media/File:RefractionReflextion.svg.*

6.4. Seismic acquisition

The fundamental purpose of seismic acquisition is accurately to record the ground motion caused by a known source in a known location. The record of ground motion with time constitutes a seismogram and is the basic information used for interpretation through either modelling or imaging. The essential instrumental requirements are to;

- Generate a seismic pulse with a suitable source.
- Detect the seismic waves in the ground with a suitable seismic transducer.
- Record and display the seismic waveforms on a suitable seismograph.

6.4.1. Seismic acquisition design

Seismic survey (Figure 6.15) is designed based on:

- Imaging objectives: image area, target depth, dips, velocity, size/thickness of bodies to be imaged, etc.
- Survey methods between seismic refraction and reflection (Table 6.1).
- Survey parameters: survey area, fold, offsets, sampling, shooting direction, etc.
- Balance between data quality and budget and facility limits.

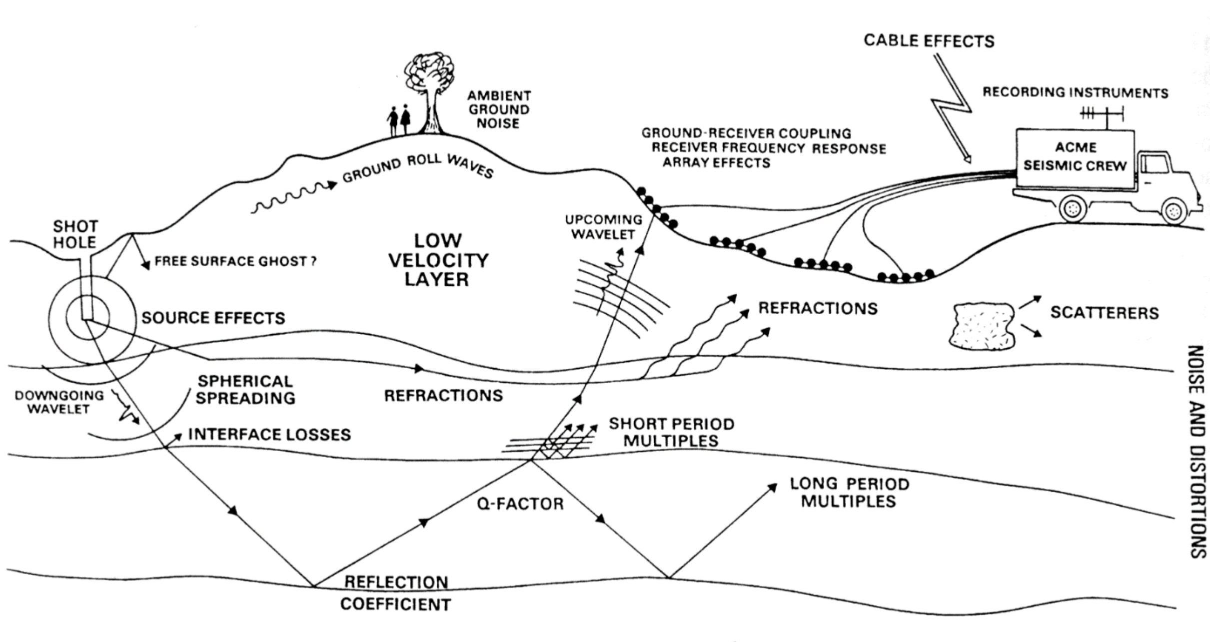


Figure 6.15: *A Seismic survey diagram (http://www.mines.edu/fs_home/tboyd/GP311/ MODULES/SEIS/main.html).*

Table 6.1: *Seismic* refraction and reflection *method comparison (modified from Kearey et al., 2002).*

	Seismic Method Comparison	
	Refraction	Reflection
Typical Targets	Near-horizontal density contrasts at depths less than ~50 m	Horizontal to dipping density contrasts, and laterally restricted targets such as cavities or tunnels at depths greater than ~20 m
Required Site Conditions	Accessible dimensions greater than ~5x the depth of interest; unpaved greatly preferred	None
Vertical Resolution	10 to 20 percent of depth	5 to 10 percent of depth
Lateral Resolution	~1/2 the geophone spacing	~1/2 the geophone spacing
Effective Practical Survey Depth	1/5 to 1/4 the maximum shot-geophone separation	>50 feet
Relative Cost	$N	$3xN to $5xN
higher velocity of lower layer requirement	Yes	No

6.4.2. Selected natural frequency for explorations

Natural frequency and damping affect the range of frequencies the geophone can record (See Figure 6.16).

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

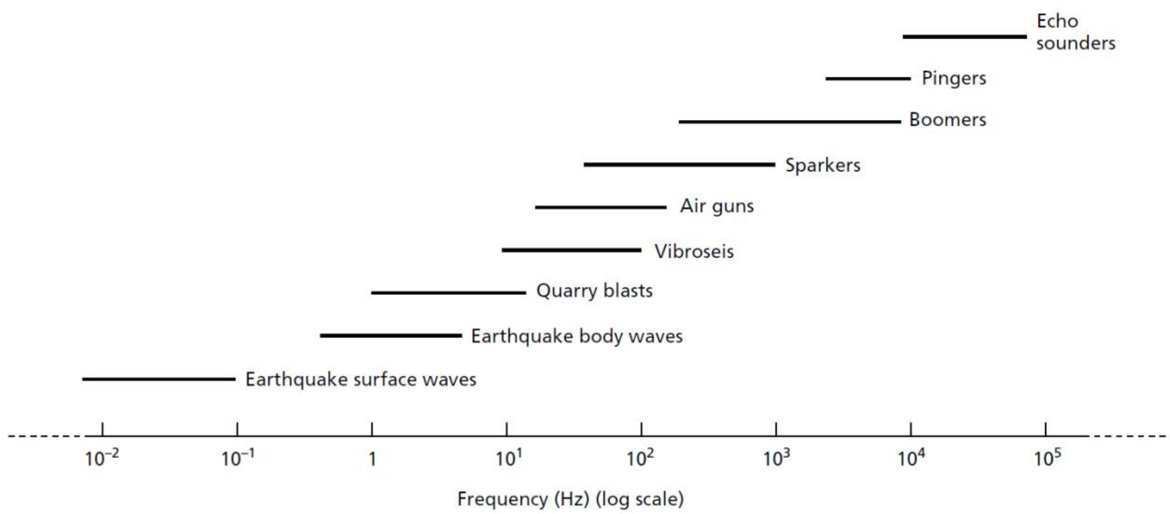


Figure 6.16: *The seismic/acoustic spectrum (modified from Kearey et al., 2002).*

6.4.3. Seismic source

Seismic source is a device that generates controlled seismic energy used to perform reflection and refraction surveys.

- A seismic source can be simple, e.g. dynamite, or it can use more sophisticated technology, e.g. a specialized air gun. Seismic sources can provide single pulses or continuous sweeps of energy. Both types of seismic sources generate seismic waves, which travel through medium of subsurface.

- Seismic source can be generated on onshore and offshore. Land and marine surveys operate on the same basic principles but using different survey configuration (See Figure 6.17A and B).

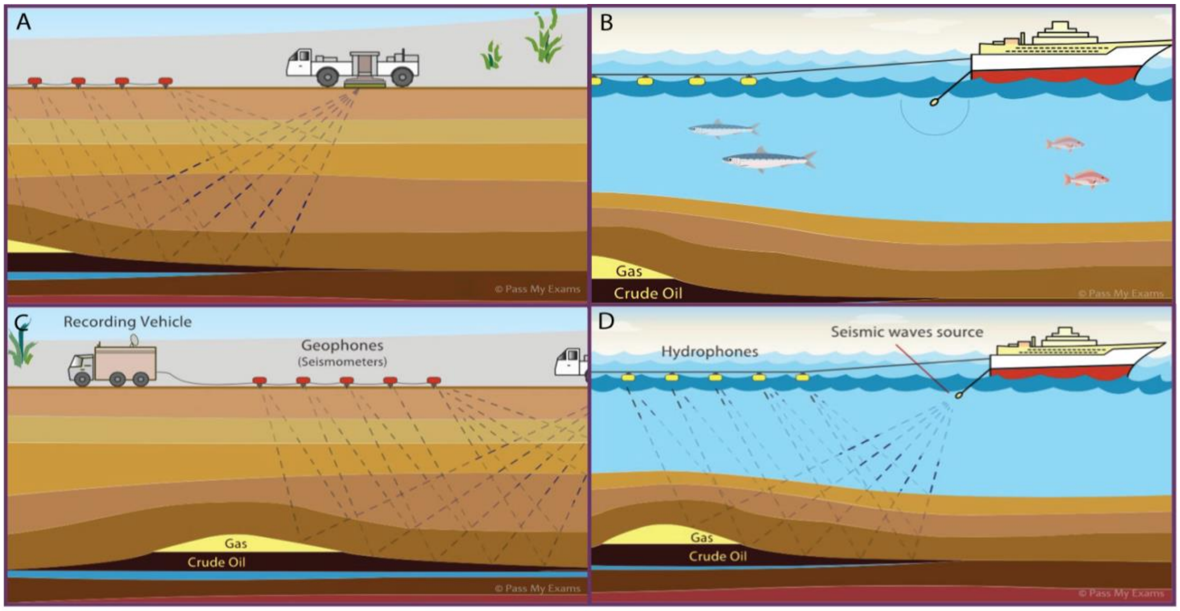


Figure 6.17: *Different on sources and receivers between onshore and offshore seismic surveys (http://www.passmyexams.co.uk/GCSE/chemistry/finding-crude-oil.html).*

6.4.3.1. Seismic source requirements

- Sufficient energy to generate a measurable signal at receiver.
- Short duration pulse, i.e. containing enough high frequencies, to resolve the desired subsurface layering.
- Repeatable source with a known, consistent waveform.
- Minimal mechanical noise and environmental impact.
- Convenience of operation.

6.4.3.2.Land seismic sources

The traditional seismic source is a small charge of explosive source. Impact and vibratory of mechanical sources are now more popular but explosives are still quite commonly.

Mechanical sledge hammer (mechanical source)

- Vertically down on plate to generate P-waves.
- Horizontally against side of plate to produce S waves.
- Problems with repeatability and possible bouncing of hammer.

- Used for refraction spreads up to 200 m.

Accelerated weight–drop (mechanical source)

- Mechanical system, using compressed air or thick elastic slings, forces weight onto base plate with greater force.
- Better repeatability than sledge hammer.

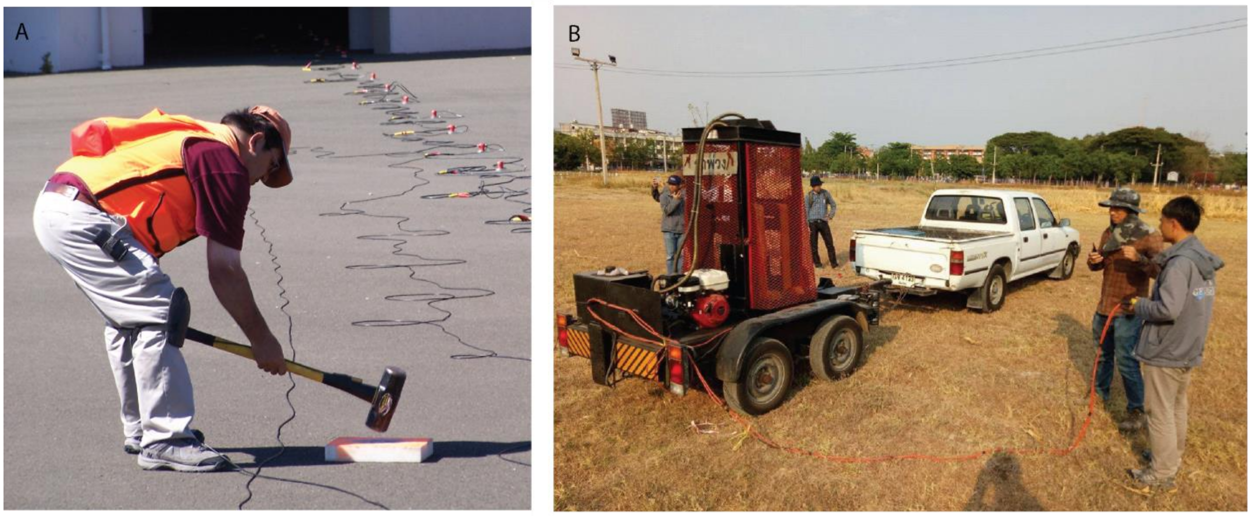


Figure 6.18: *Seismic sources from A) sledge hammer and B) accelerated weight–drop (https://washingtondnr.wordpress.com/2011/04/12/earthquake-awareness-dnr-experts/).*

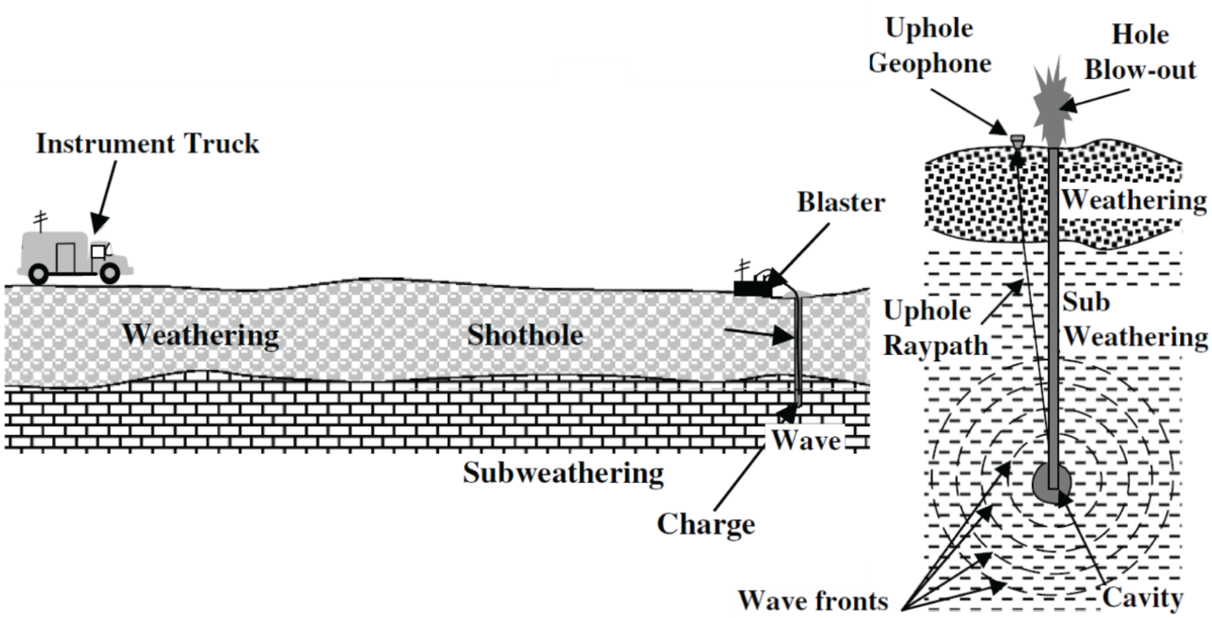


Figure 6.19: *Explosive source operation (Gadallah and Fisher, 2009).*

Buffalo gun (explosive source)

- Metal pipe inserted up to 1 m into the ground, and a blank shotgun cartridge fired.
- Exploding gases from gun impact ground and generate the seismic pulse.

Dynamite

- Shot holes up to 30 m are drilled, and loaded with dynamite, which usually comes in 0.5 m plastic cylinders that can be screwed together.

6.4.3.3.Marine seismic sources

- The airgun is most common seismic source used at sea.
- Essentially, an airgun is a cylinder that is filled with compressed air.
- The sudden release of air creates impulse of strong high frequent acoustic signals in the water (Figure 6.20A)

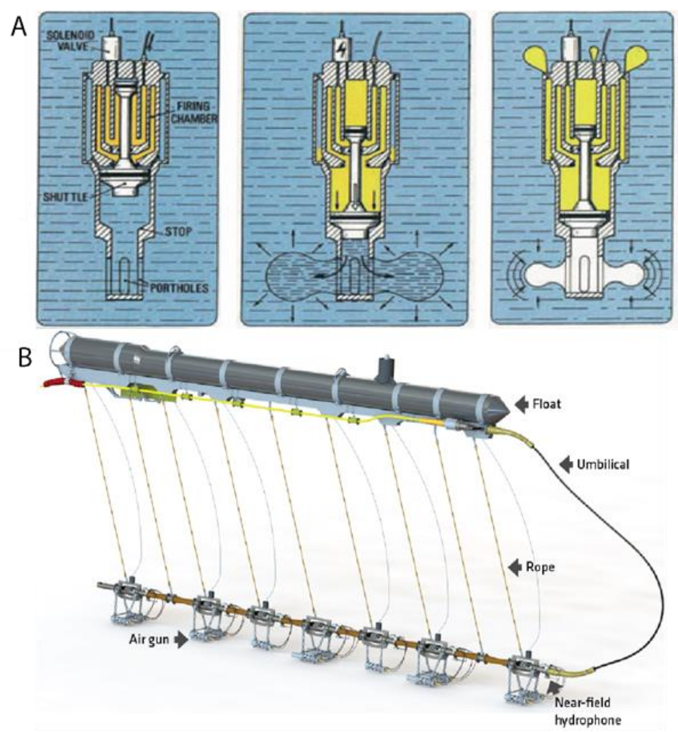


Figure 6.20: *A) Operation of a S15 airguns. Air pushes water through the portholes creating cavities that collapse and create a strong high frequent acoustic signal. B) Array with several airguns suspended below a surface float (https://www.geoexpro.com/articles/2010/04/marine-seismic-sources-part-iii).*

6.4.4. Seismic sensors

The geophone

- Geophone is essentially only type of sensor used on land.
- A geophone comprises a coil suspended from springs inside a magnet.
- When the ground vibrates in response to a passing seismic wave, the coil moves inside the magnet, producing a voltage, and thus a current, in the coil by induction.
- As coil can only move in one direction, usually vertical, the geophone only senses the component of seismic motion along axis of coil.

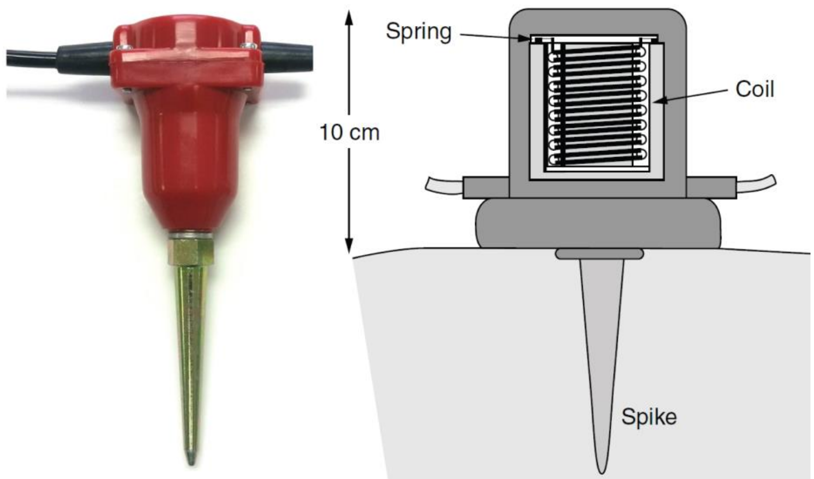


Figure 6.21: *An example of a geophone and internal system.*

The Hydrophone

- Hydrophone is essentially only type of sensor used on marine survey.
- Sound waves in water reflected by different boundaries arrive at different times.
- Hydrophones used to detect the <u>pressure variations</u> in water due to a passing seismic wave (See Figure 6.22).

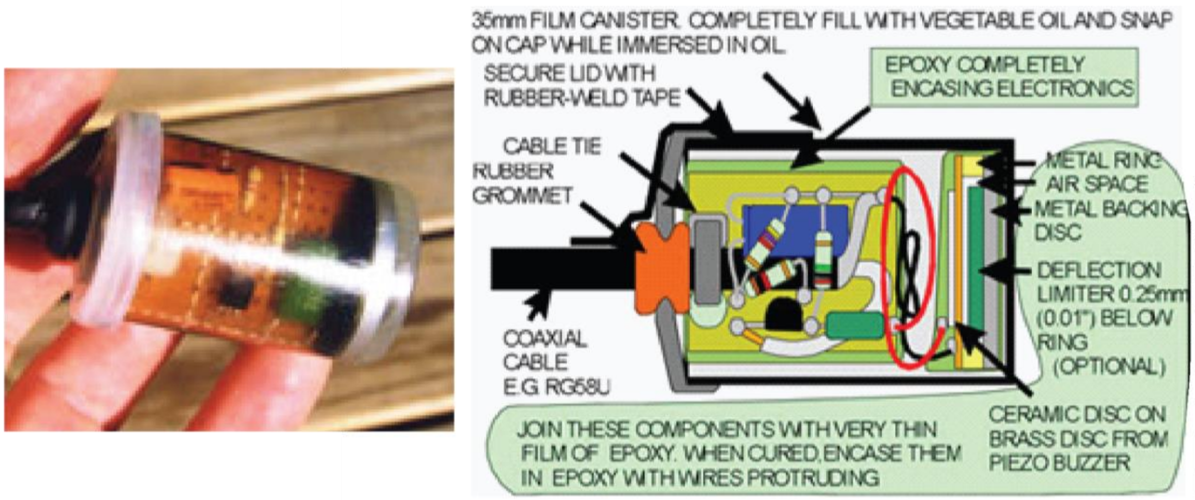


Figure 6.22: *Diagram information of a simple DIY bender hydrophone (*http://www.exfac.com/aussie_bray/hydro_dunking.htm*).*

6.4.5. Geophones spread

- A line of geophones laid out for a seismic survey is known as a spread, the term array being reserved for geophones feeding a single recording channel. Arrays are common in reflection work but are almost unknown in refraction surveys where the sharpest possible arrivals are needed.
- Shot would usually be placed at one end of spread for first recording, then second recording made at other end.
- The spread length should be eight times the expected refractor depth.
- Sufficient information on the direct wave and reasonable coverage of the refractor is obtained if the length of the spread is about three times the crossover distance.
- Off-end, end-on and split-spread shooting are main spread types (Figure 6.23).

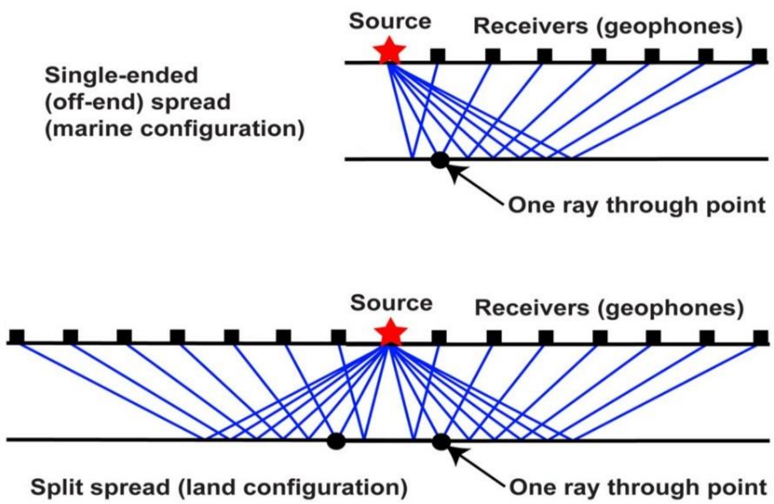


Figure 6.23: *Geophone spread types of off-end and split-spread shooting (*https://agsc34.weebly.com/*).*

6.4.6. 3D seismic survey

- Nearly all seismic surveys are now 3-D. Very little 2-D shooting is done today in a hydrocarbon exploration propose because of the limits of 2-D such as distortion of the image of geologic structure and inadequate subsurface sampling for small-scale geologic features.
- A typical 3-D survey is carried out by shooting closely spaced parallel lines of 2D distribution of shot source and receivers locations (Figures 6.24, 6.25, and 6.36).

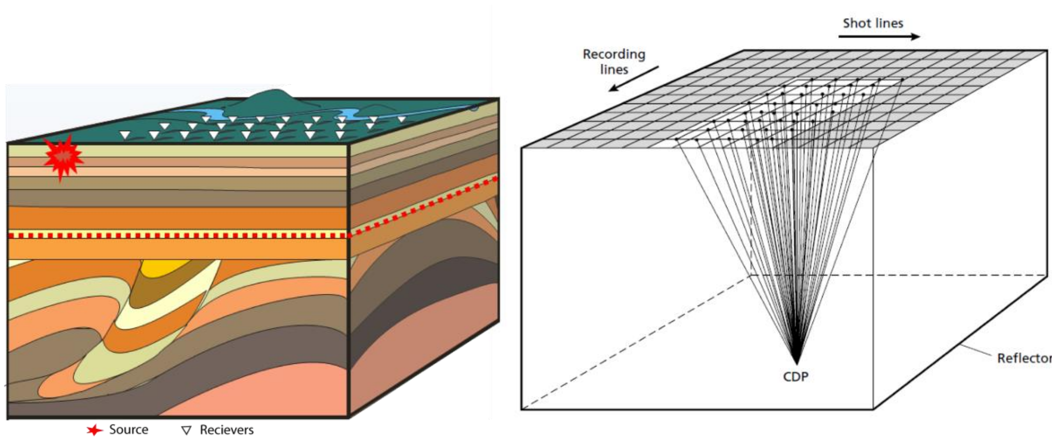


Figure 6.24: *Three-dimensional reflection survey design and reflected ray paths defining a common depth point from an aerial distribution of source and receivers locations (modified from* https://www.slb.com/~/media/Files/resources/oilfield_review/ors89/oct89/53dseismic.pdf*).*

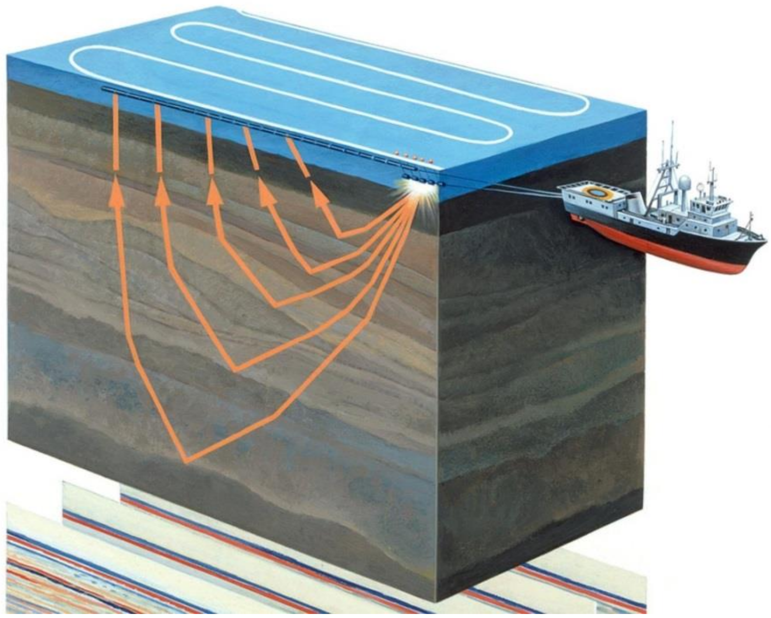


Figure 6.25: *Three-dimensional marine reflection survey and reflected ray paths* (http://en. *stonkcash.com/seismic-energy/).*

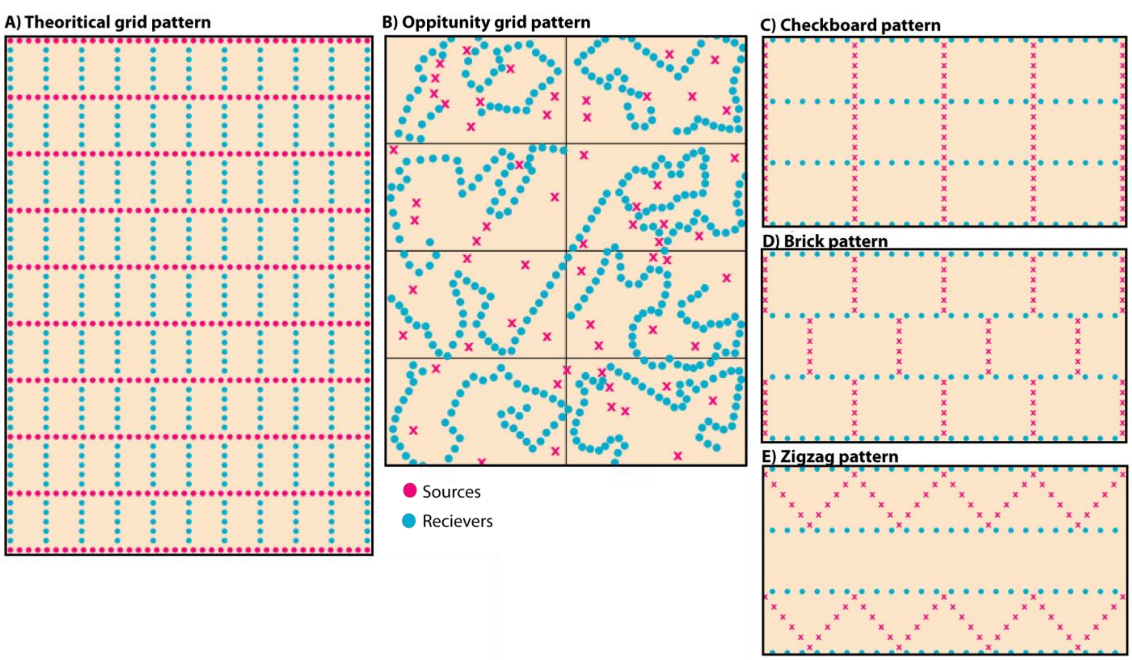


Figure 6.26: *3D seismic grid survey of A) theoretical and B) opportunity patterns. C), D) and E) are checkboard, brick, and zigzag pattern, respectively (modified from https://www.slb.com/~/media/Files/resources/oilfield_review/ors89/oct89/5_3d_seismic.pdf).*

6.4.7. Seismic shot record

- Figure 6.27 is a seismic trace recorded from a single shot point. They are a record~~ing~~ of the Earth's response to seismic energy passing from the source, through subsurface layers, and back to the receivers.
- Numerous seismic records are displayed together in a single shot section.
- The shot records are used to identify principal reflections of interest, including various sources of noise.

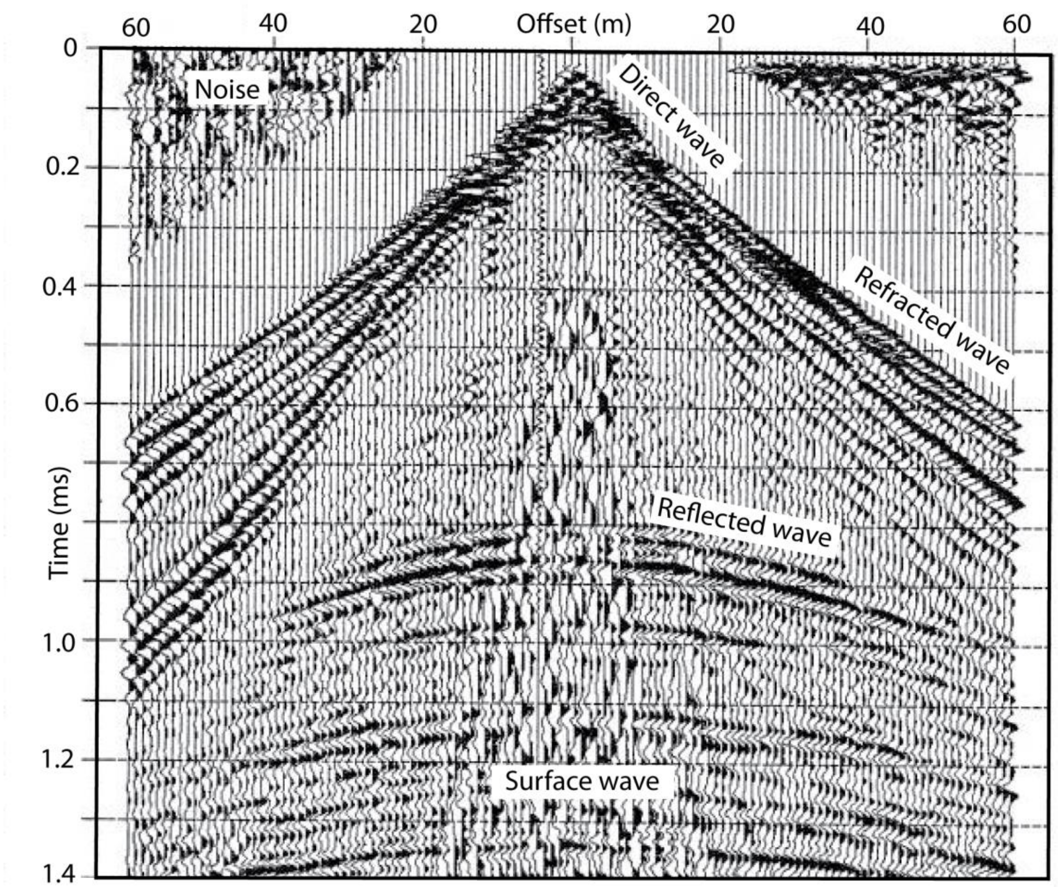


Figure 6.27: *A single shot record as it is recorded in the field. The shot is at the center There were 120 geophones laid out in this "split" spread (modified from* http://www.gebrproject.com */?p=1210).*

6.5. Seismic refraction processing and interpretation

- The seismic refraction surveying method uses seismic energy that returns to the surface after travelling through the ground along refracted ray paths (Figure 6.28).

- Refraction seismograms may also contain reflection events as subsequent arrivals, though generally no special attempt is made to enhance reflected signals in refraction surveys.
- Refraction seismology is applied to a very wide range of scientific and technical problems, from engineering site investigation surveys to large-scale experiments designed to study the structure of the entire crust or lithosphere.

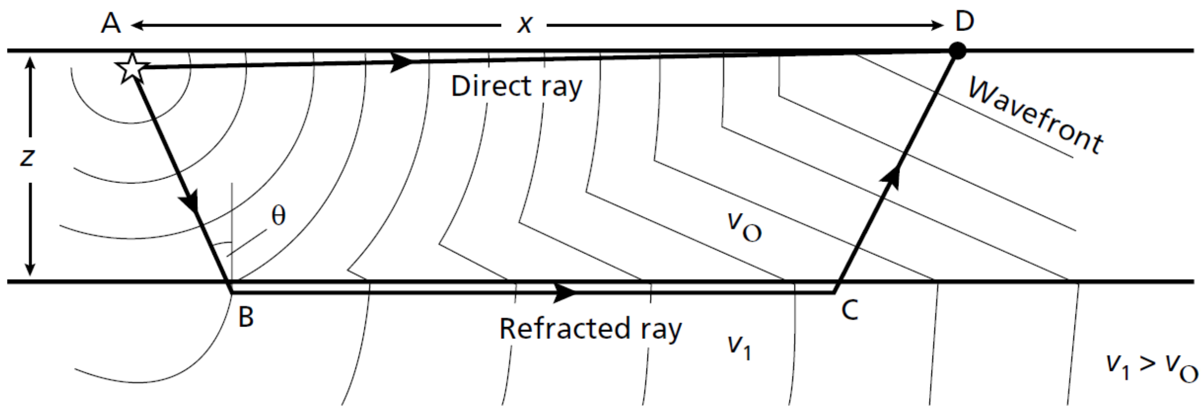


Figure 6.28: *The expanding first arrival wavefronts for direct and refracted waves (solid arrows) through a two-layer model. A is a location of source and D is a location of receiver (modified from Kearey et al., 2002).*

6.5.1. Assumptions for seismic refraction interpretation

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

6.5.2. Two layer model

Two layers model is the simplest acquisition case to consider the subsurface thickness of seismic layer using velocity determination of propagating direct and refracted waves.

6.5.3. First-break picking

- First-break picking detecting or picking the onset arrivals of refracted signals from all the signals received by receiver arrays and produced by a particular source signal generation (Figure 6.29).
- Picking first arrivals on refraction records may be difficult at remote geophones where the signal-to-noise ratio is poor.

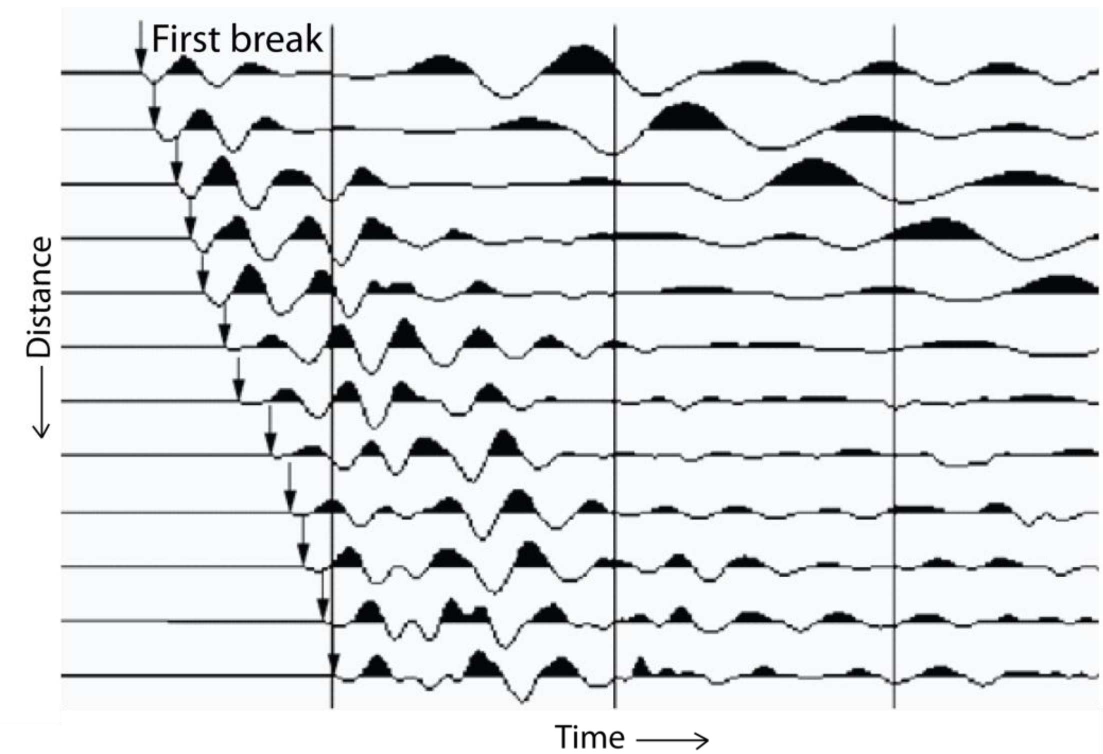


Figure 6.29: *Illustration of the first-break picking on a single seismic trace (Burger et al., 2006).*

6.5.4. A travel time curve

- A travel-time curve is a graph of arrival times, commonly P waves (first wave arrival), recorded at different points as a function of distance from the seismic source.
- A travel-time curve takes for seismic waves to travel from the shot station (distance = 0) to receiver stations varying distances away.
- Seismic velocities within the earth can be computed from the slopes of the resulting curves (Figure 6.30).

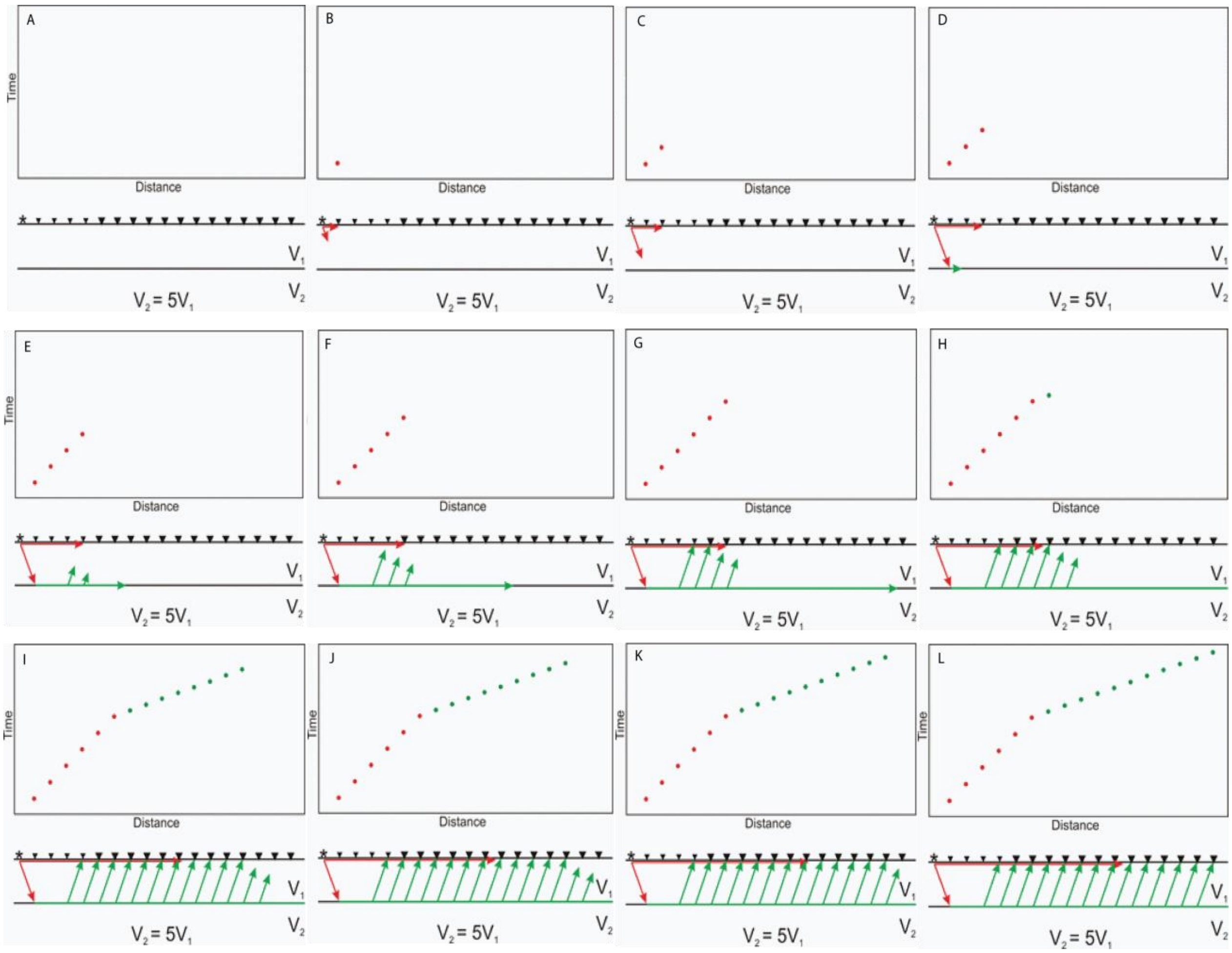


Figure 6.30: *Multiple snapshots showing relation between ray paths of the wave propagation and the travel time curve of the first arrival wave (https://en.wikipedia.org/wiki/Reflection_seismology#/media/File:Seismic_Refraction_Principal.png).*

6.5.4.1.Two layers model of direct wave

- Direct wave is a seismic wave which travels through the ground directly from the source to the receivers without being reflected off or refracted by a subsurface layer.
- Figure 6.31 shows the direct wave travels horizontally through the top of the upper layer from A to D at velocity v_0.

6.5.4.2.Two layer model of heads wave

- Figure 6.31 also shows the critical ray follows the line of the interface and sends a return ray back to the surface. This is detected by the geophones.

- The critical ray (or head wave) moves in a lower layer with a velocity v_1. It thus sends a progressive series of return rays along its path.
- These are detected in turn by each geophone.
- Both the down going and return rays meet the interface at the critical angle of refraction.
- Note that: the first head wave is a reflected wave at the critical angle.

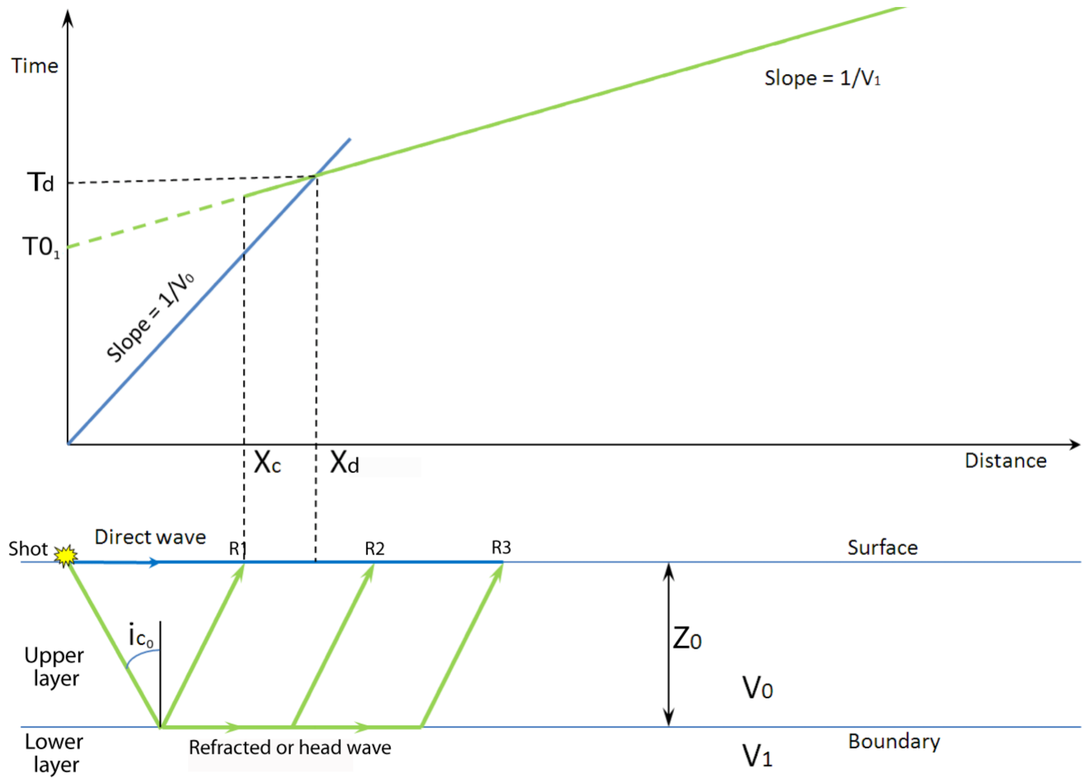


Figure 6.31: *The travel-time curve for the direct wave and the head waves plotted from ray path wave propagations through the bottom layer of the two-layer model (https://en.wikipedia.org/wiki/Reflection_seismology#/media/File:Seismic_Refraction_Principal.png).*

6.5.4.3.Depth (Z_0) termination (step of solving)

- Determine the velocities of both layers (V_0 and V_1) in meters per second (m/s) on the plotted graphs (Figure 6.31, Slope green and blue).
- Determine the crossover distance X_d in meter
- Determine the time intercept ($T_D(x) = T_H(x)$) for V_2
- Determine the depth (Z_0) to the lower layer using both refraction equations

When, i_{c_0} = *critical angle*

V_0 = velocity of the first layer

V_1 = velocity of the second layer

Z_0 = thickness of the first layer

T_{01} = intercept

Energy refracting across the interface, travelling along the underside and then back up to surface, travel time:

$$t = \frac{SA}{\alpha_1} + \frac{AB}{\alpha_2} + \frac{BR}{\alpha_1}$$

With some algebra

$$t = \frac{2Z_1}{\alpha_1}\sqrt{1 - \frac{{\alpha_1}^2}{{\alpha_2}^2}} + \frac{x}{\alpha_2}$$

This fits to the equation of straight line

$$t = a + bx$$

Where, the slope of the line is

$$\frac{1}{\alpha_2}$$

And the intercept is

$$\frac{2Z_1}{\alpha_1}\sqrt{1 - \frac{{\alpha_1}^2}{{\alpha_2}^2}}$$

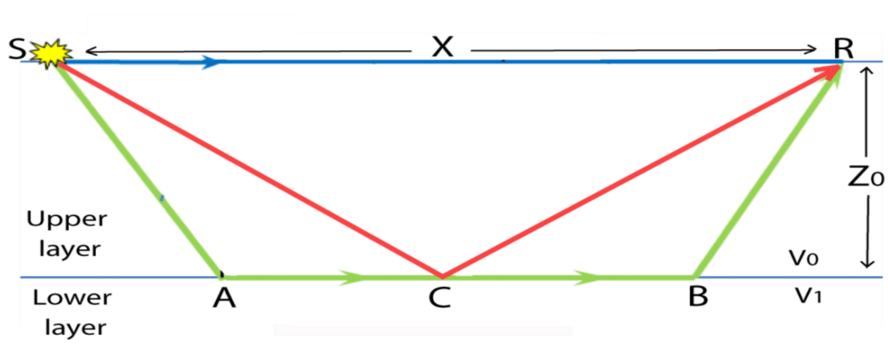


Figure 6.32: *Refracted and reflected wave propagations from source to any receiver (modified from Robinson and Crouch, 1998).*

6.5.4.4.Summary of important equations

For refractor parallel to surface

$$\frac{\sin i}{\sin r} = \frac{V_1}{V_2}$$

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

$$i_c = \sin^{-1}\frac{V_1}{V_2}$$

$$T_2 = \frac{2h\cos(i_c)}{V_1} + \frac{x}{V_2}$$

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

$$h = \frac{T_i V_1}{2\cos(\sin^{-1}\frac{V_1}{V_2})}$$

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

6.5.4.5.Critical distance

- Offset at which critical refraction first appears (Figure 6.31)
- Critical refraction has same travel time as reflection.
- Angle of reflection same as critical angle.

6.5.5. Multiple layer model

- The geometry of the ray path in the case of critical refraction at the second interface is shown in Figure. 6.33.
- The seismic velocities of the three layers are V_0, V_1 ($>V_0$) and V_2 ($>V_1$). The angle of incidence of the ray on the interfaces are critical angle.
- The interpretation of travel-time curves for a three layers model shows that the less gradient curve represents the higher velocity layer.
- Calculation for the thicknesses of three/multiple layers model does not work using simple maths. Better to achieve results using commercial computer software.

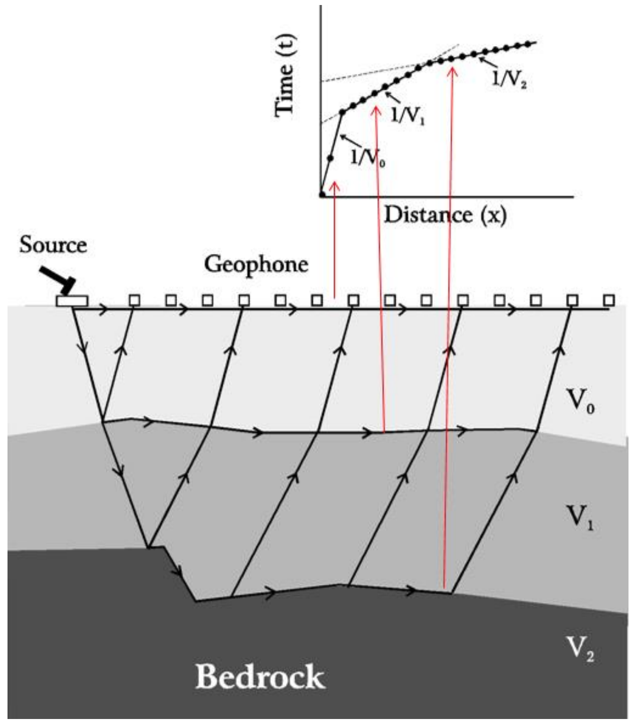


Figure 6.33: *Ray path for a wave refracted through the bottom layer of a three-layer model* (*Robinson and Crouch, 1998*).

6.5.6. Dipping layer model

- The presence of a dipping interface is recognised if the reversed profile of wave propagation is not the mirror image of the forward profile (Figure 6.34).
- The analysis of a dipping interface introduces three new issues:
  - There is an additional unknown (the dip angle).
  - The travel-time curves are no longer symmetrical and so the updip and downdip intercepts are not equal (Figure 6.34).
  - The updip and downdip velocities in layer 2 are not equal.
- When a refractor dips, the slope of plotted the travel time curve does not represent the "true" layer velocity:
- Shooting updip, i.e. geophones are on updip side of shot, apparent refractor velocity is higher and shooting downdip apparent velocity is lower.
- To determine both the layer velocity and the interface dip forward and reverse refraction profiles must be acquired.

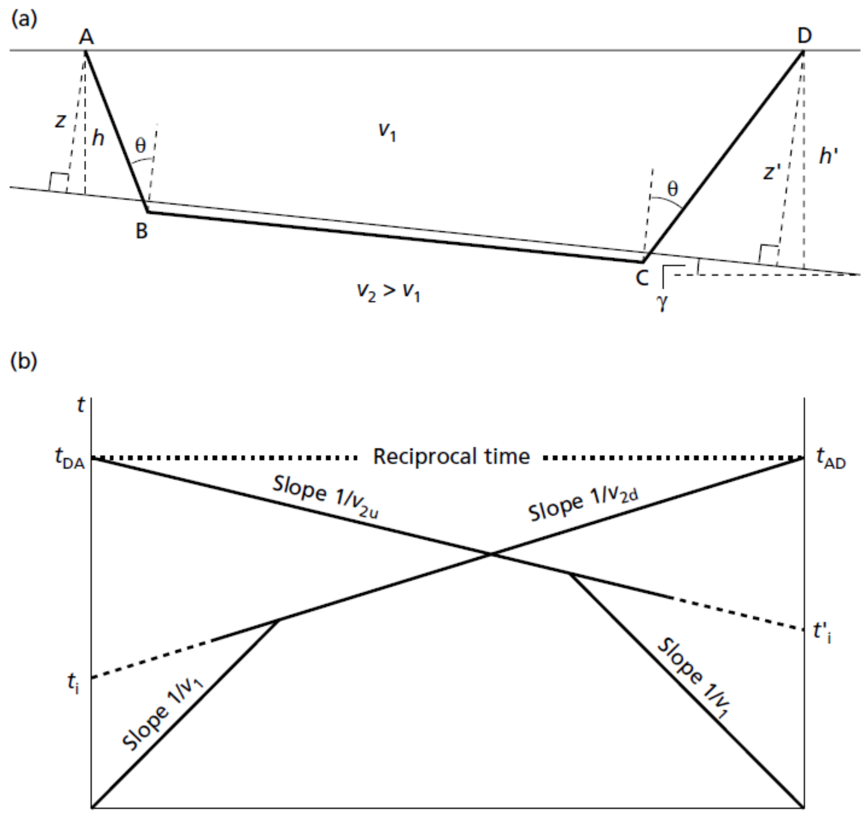


Figure 6.34: *(a) Ray-path geometry and (b) travel-time curves for head wave arrivals from a dipping refractor in the forward and reverse directions along a refraction profile line (Kearey et al., 2002).*

6.5.7. Dipping model calculations. See Figure 6.34 for level explanation.

$$\gamma = \frac{1}{2}\left[\sin^{-1}\left(\frac{V_1}{V_{2d}}\right) - \sin^{-1}\left(\frac{V_1}{V_{2u}}\right)\right]$$

$$\theta_c = \frac{1}{2}\left[\sin^{-1}\left(\frac{V_1}{V_{2d}}\right) + \sin^{-1}\left(\frac{V_1}{V_{2u}}\right)\right]$$

- The assumption of planar refracting interfaces would often lead to unacceptable error or imprecision in the interpretation of refraction survey data.
- For example, a survey may be carried out to study the form of the concealed bedrock surface beneath a valley fill of alluvium or glacial drift.
- Such a surface is unlikely to be modelled adequately by a planar refractor. In such cases the constraint that refracting interfaces be interpreted as planar must be dropped and different interpretation methods must be employed.

6.5.8. Hidden layer problem

- It is possible for layers to exist in the Earth, yet not produce any refracted first-arrival waves. In this case the layers will be undetectable in a simple first arrival refraction survey.
- The observed data could be interpreted using the methods discussed above and yield a self-consistent, but erroneous, solution.
- For this reason, the possibility of undetected layers should always be considered.
- In practice, there are two different types of problem. In order to be detected in a first arrival refraction survey,

  (a) A layer must be underlain by a layer of higher velocity so that head waves are produced.

  (b) A layer must have a thickness and velocity such that the head waves become first arrivals at some range**.**

6.5.9. Faulted planar Interface

- The effect of a fault displacing a planar refractor is to offset the segments of the travel-time plot on opposite sides of the fault (see Figure 6.36).
- There are thus two intercept times t_{i1} and t_{i2}, one associated with each of the travel time curve segments, and the difference between these intercept times t_D is a measure of the throw of the fault**.**

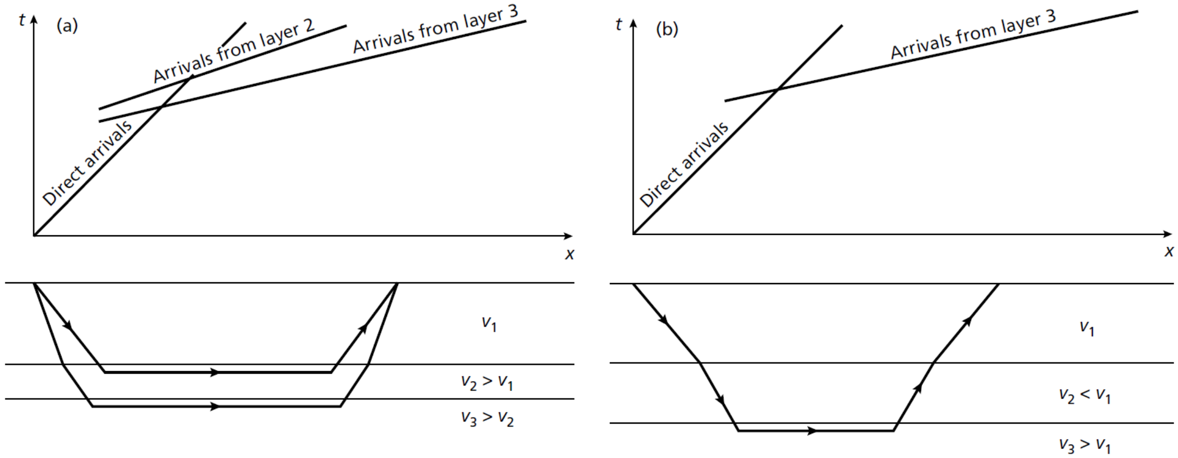


Figure 6.35: *The undetected layer problem in refraction seismology. (a) A hidden layer: a thin layer that does not give rise to first arrivals. (b) A blind layer: a layer of low velocity that does not generate head waves (Kearey et al., 2002).*

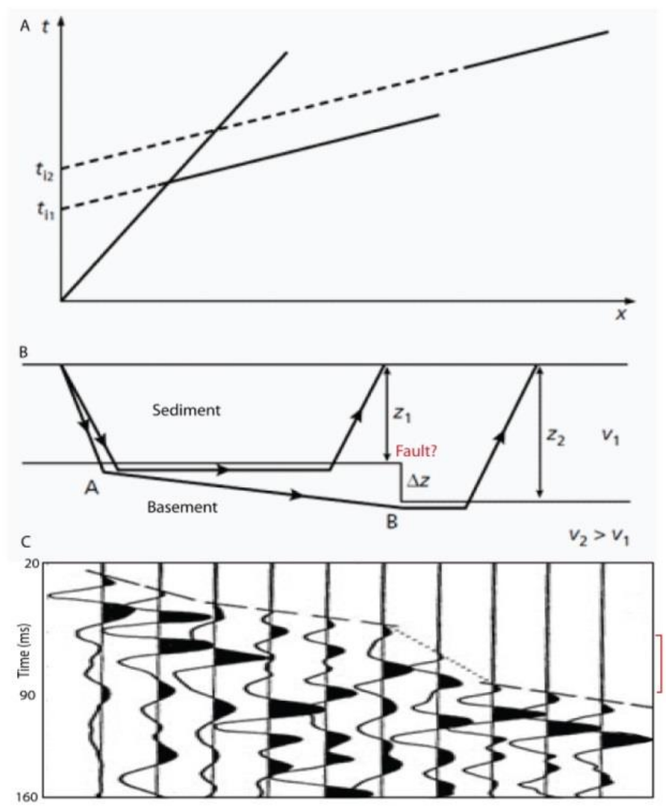


Figure 6.36: *(A) Offset segments of the travel-time curve for (B) refracted arrivals from opposite sides of a fault (C) Arrival time shift because of down dip fault displacement (modified from Burger et al., 2006).*

6.5.10. Advantages and disadvantages of seismic refraction

Advantages

- Simple layout.
- Low manpower requirements.
- Limited equipment requirements.
- Rapid data reduction and analysis (computer not needed).
- Easy interpretation.

Disadvantages

- Relatively large energy input required.
- Relatively long layout (10 times depth).
- Limited number of model layers.
- Limited velocity differences.
- Limited interface geometry (assume smooth).

6.6. Seismic reflection processing

6.6.1. A seismic reflector

- A seismic reflector is a boundary between beds with different properties. There may be a change of lithology or fluid fill from Bed 1 to Bed 2.
- The property changes cause some sound waves to be reflected towards the surface. There are many reflectors on a seismic section.
- Major changes in properties usually produce strong, continuous reflectors as shown by the arrow.
- Due to a reflected signal will always arrive after a direct ray and possibly the refracted ray. Thus they cannot be detected using first arrival methods. The full waveform must be recorded and analyzed.

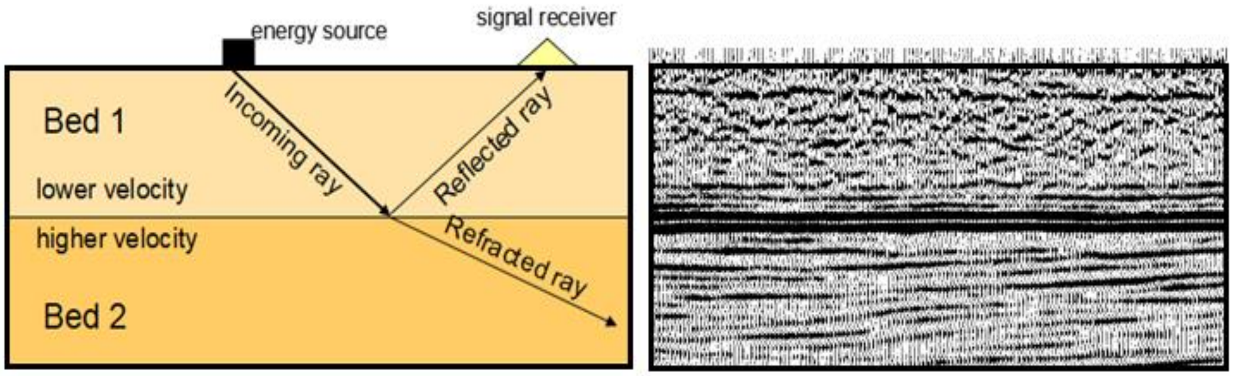


Figure 6.37: *The two layers model of Beds 1 and 2 with different properties and a seismic reflector corresponding a boundary between beds (modified from* https://www.iris.edu/hq/inclass/*lesson/seismic_reflection).*

6.6.2. Seismic convolution

- Convolution: is a mathematical way of combining two signals to achieve modified signal of seismic trace. Convolution in the time domain is represented in the frequency domain by a multiplying the amplitude spectra and adding the phase spectra (Figure 6.38, left).
- Figure 6.38 of seismic trace shows a spike series representing an acoustic impedance response from the earth which is convolved (*) with a source wavelet to produce a resulting seismic signal which is measured.
- The convolutional model of the seismic trace states that the trace we record is the result of the earth's reflectivity (what we want) convolved with the source wavelet (and it's ghosts), multiples, the recording system and some noise.
- Note that: Deconvolution is a filtering process which removes a wavelet from the recorded seismic trace by reversing the process of convolution.
    - In principal by deconvolving the source wavelet we could obtain the earth's reflectivity. However, noise (unwanted signal) and other features are also present in the recorded

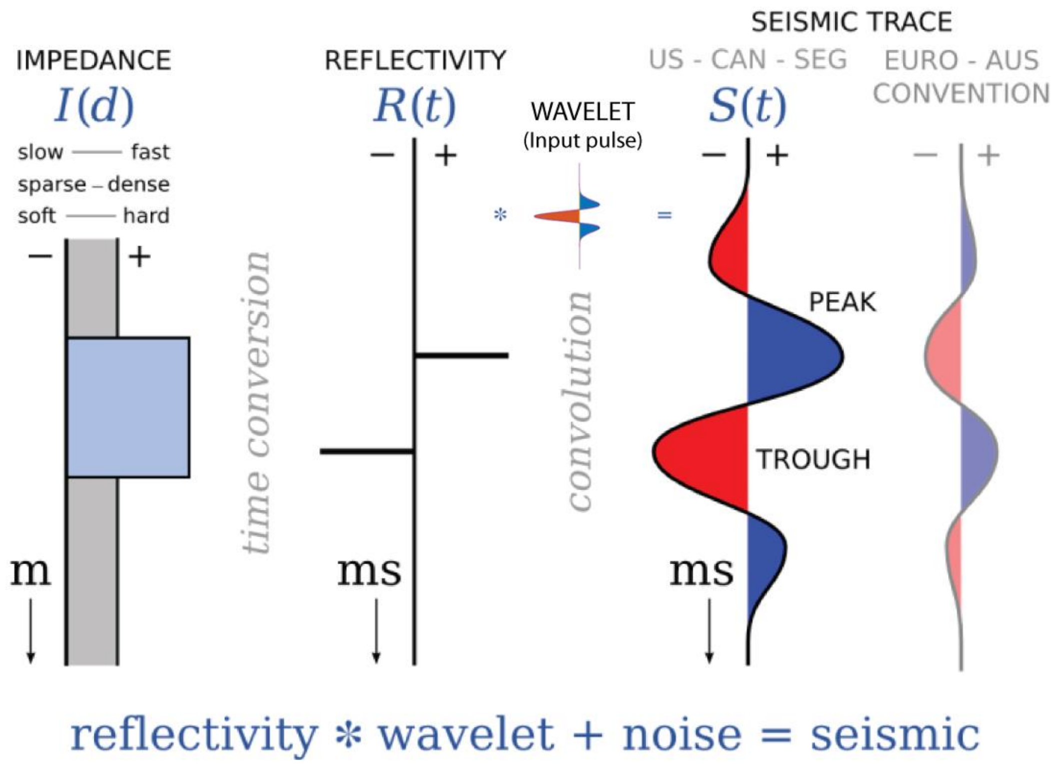


Figure 6.38: *The convolution model of seismic trace which is a result of a between subsurface reflectivity series and wavelet generated by the source (modified from* http://subsurfwiki.org/wiki/Convolutional_model*).*

6.6.3. Seismic reflection processing

The purpose of seismic processing is to manipulate the acquired data into an image that can be used to infer the sub-surface structure. Only minimal processing would be required if we had a perfect acquisition system.

Processing routines generally fall into one of the following categories:

- Enhancing signal at the expense of noise
- Providing velocity information
- Collapsing diffractions and placing dipping events in their true subsurface locations (migration)
- Increasing resolution (wavelet processing)

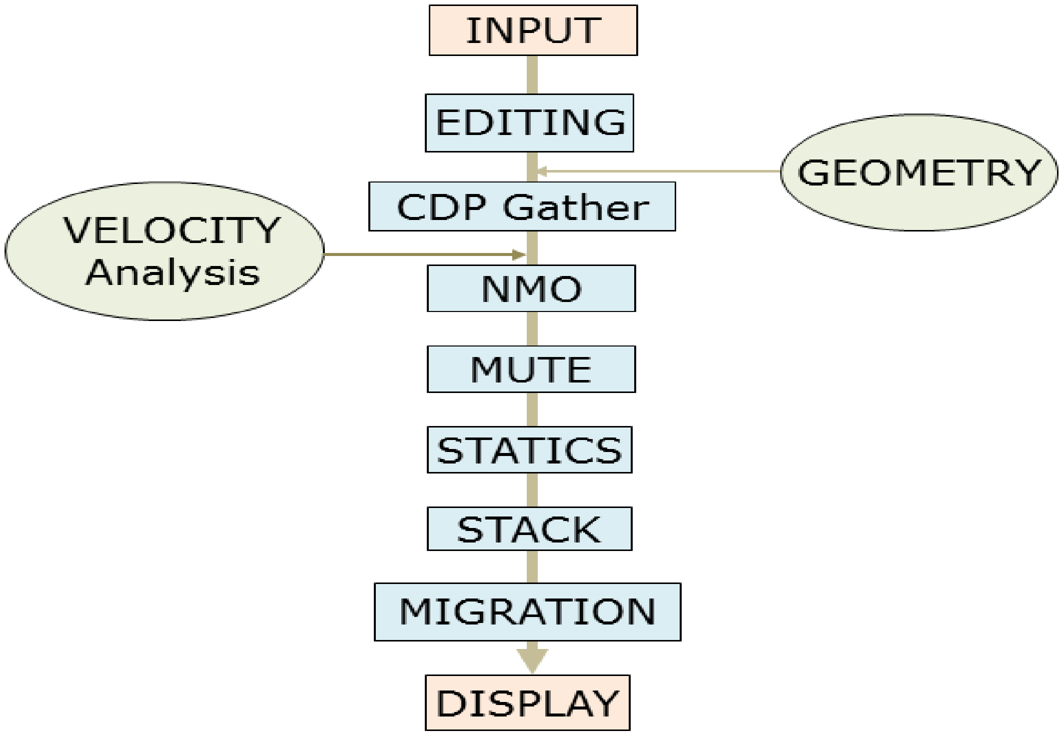


Figure 6.39: *Basic seismic reflection processing.*

6.6.4. *Simple processing sequence flow*

- Reformat
- Geometry Definition
- Field Static Corrections (Land - Shallow Water - Transition Zone)
- Amplitude Recovery
- Noise Attenuation (De-Noise)
- Deconvolution
- CMP Gather
- NMO Correction
- De-multiple (Marine)
- Migration
- CMP Stack

6.6.4.1.Editing dead or corrupted traces

- The data editing is one of the original methods of controlling noise in seismic data by muting bad data. In practical it is the simplest methods of removing noise.
- The emphasis is eliminating isolated high-amplitude unpredictable events. These events are unpredictable in the sense that they leave high-amplitude residuals when predictable events are removed.

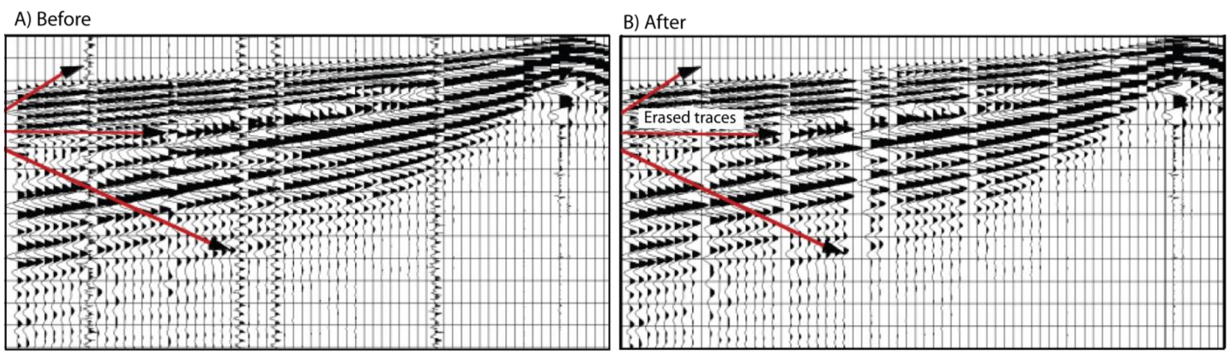


Figure 6.40: *A seismic shot record before and after editing bad seismic traces (modified from https://www.iris.edu/hq/inclass/lesson/seismic_reflection).*

6.6.4.2.Geometry

The association with determination of source and receiver positions for measured data and calculation of CMP position.

6.6.4.3.Common depth point and gather

- CDP is the halfway point in the travel of a wave from a source to a flat-lying reflector to a receiver.
- In the case of flat layers, CDP is vertically below the common midpoint (CMP). Thus, traces constitute both CMP and CDP gathers (Figure 6.41A).
- In the case of dipping beds, there is no CDP, so dip move out processing is necessary to reduce smearing, or inappropriate mixing, of the data (Figure 6.41B).

- The fold of the stacking refers to the number of traces in the CMP gather (Figure 6.42). In exploration aspect, numbers of fold may conventionally be 24, 30, 60 or exceptionally, over 1000.

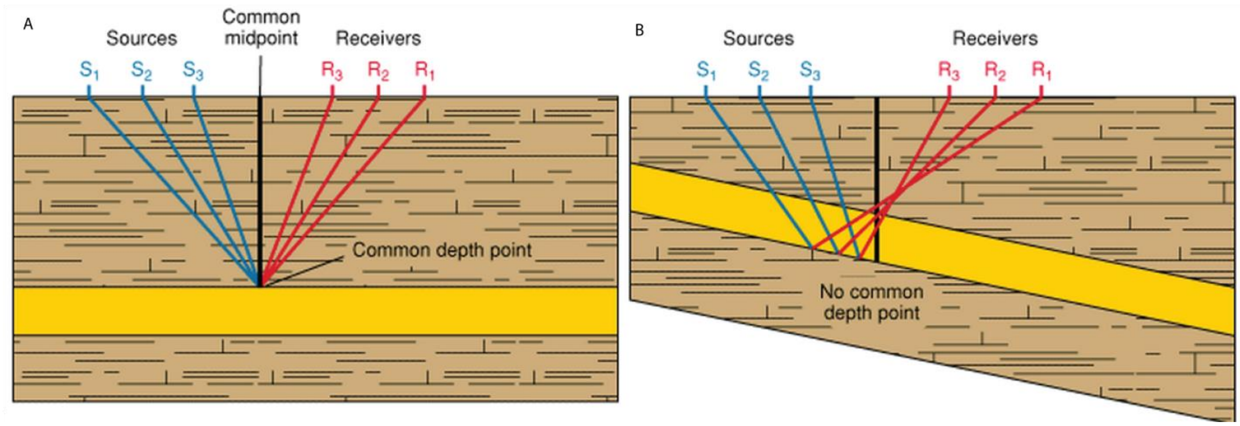


Figure 6.41: *Diagrams of common depth point and common midpoint recorded on A) flat layers and B) dipping layers (modified from* http://www.glossary.oilfield.slb.com/*Terms/c/ common_depth_point.aspx).*

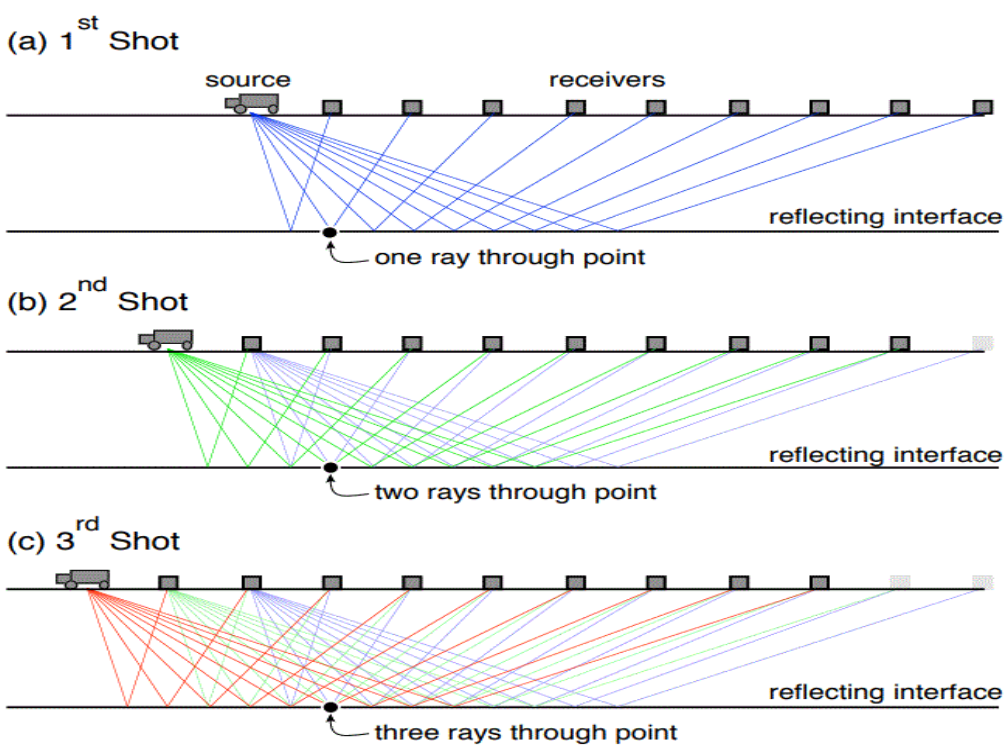


Figure 6.42: *CMP schematic, for 3-fold cover with a 9-channel system from 3 shot points are progressively one geophone group interval further to the right* (http://www.geo.cornell.edu/ *geology/faculty/RWA/structure-lab-manual/chapter-11.pdf).*

6.6.4.4. Velocity analysis

- The velocity analysis is the method to find the corrected velocity.
- Velocity can be calculated from normal moveout or the change in arrival time produced by source-receiver offset.
- A good velocity model is the basis for:
  - Stacking (improvement of S/N-Ratio).
  - Appropriate conversion from travel time into depth.
  - Geometrical correction (migration).

Types of velocities

- Average velocity, V_{AVG}= (Z/T) at which represent depth to bed (from surface to layer). Average velocity is commonly calculated by assuming a vertical path, parallel layers and straight ray paths, conditions that are quite idealized compared to those actually found in the Earth.
- Interval velocity, V_{INT}= ($\Delta Z/\Delta T$) of a specific layer or layers of rock,
- Instantaneous velocity, V_{INST}= $\Delta Z/\Delta T$
- Root Mean Square (RMS) velocity: V_{RMS} is the square root of the average squared velocity
- NMO velocity: used to correct for normal moveout (NMO)
- Stacking velocity, $V_{Stacking}$: derived from seismic velocity analysis, a measurement of average velocity from surface to reflector
- Dix velocity, V_{Dix}: $V_{Stacking}$ converted to an interval velocity
- Migration velocity: V_{Mig}: smoothed and calibrated $V_{Stacking}$

Two way time (TWT)

- Two-way travel time is the time taken for a seismic wave to travel from the shot down to a reflector or refractor and back to a geophone at the surface.
- For finite offsets, the two-way travel times are affected by normal moveout; the normal-incidence two-way travel time is measured at zero offset.

$$t_0 = \frac{2Z}{V}$$

X^2-t^2 *Plot Method*

- The averages of arrival times from the geophones plot as a straight line on x^2-t^2 plot. The slope of the straight line is equal to the inverse of velocity squared (Figure. 6.43).
- Orthogonal distance to the interface or thickness can be determined from the velocity times the intercept time divided by two.

$$t_x = \frac{\sum s}{V} \qquad t = \frac{\sqrt{x^2+4z^2}}{V}$$

$$t_x = \frac{\sqrt{x^2+4z^2}}{V} \qquad t^2 = \frac{(x^2+4z^2)}{V^2}$$

$$t_0 = \frac{2Z}{V} \qquad t^2 = \frac{1}{V^2}X^2 + \frac{4Z^2}{V^2}$$

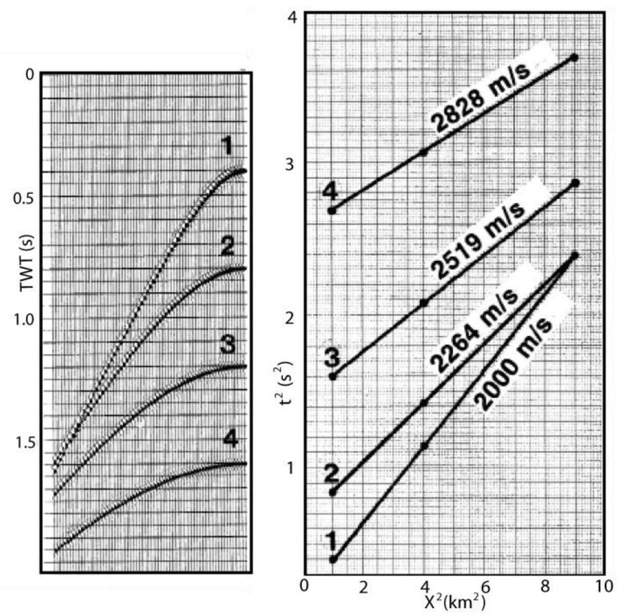


Figure 6.43: *The t^2 – x^2 velocity analysis applied to the synthetic gather derived from the velocity function (*https://wiki.seg.org/wiki/Velocity_analysis*).*

6.6.4.5.Normal moveout

- Normal moveout (NMO) is a difference in reflection travel time response in curved hyperbola from the same horizontal surface due to difference in distance between source and receivers. More time needed to reach distant receivers so the data look like a curve.
- Reflection travel times must be corrected for NMO to display a near horizontal reflector prior to summing the traces in the CMP gather along the offset axis (Figures 6.44 and 6.45).

- NMO depends on velocity above the reflector, offset, two-way zero-offset time associated with the reflection event, dip of the reflector, the source-receiver azimuth with respect to the true-dip direction, and the degree of complexity of the near-surface and the medium above the reflector.
- Typically, NMO decrease with increasing depth and also decrease with increasing velocity (Figure 6.46).

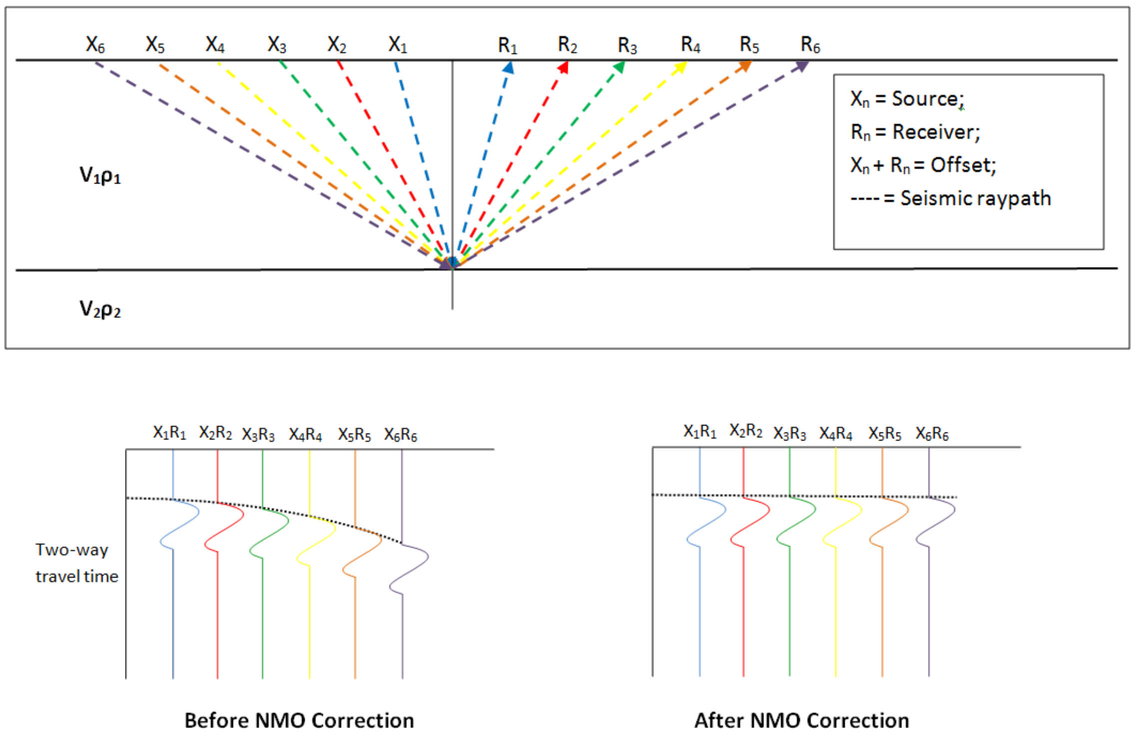


Figure 6.44: *A set of reflection events in a CMP gather is corrected for NMO (*https://en.wikipedia.org/wiki/Normal_moveout*).*

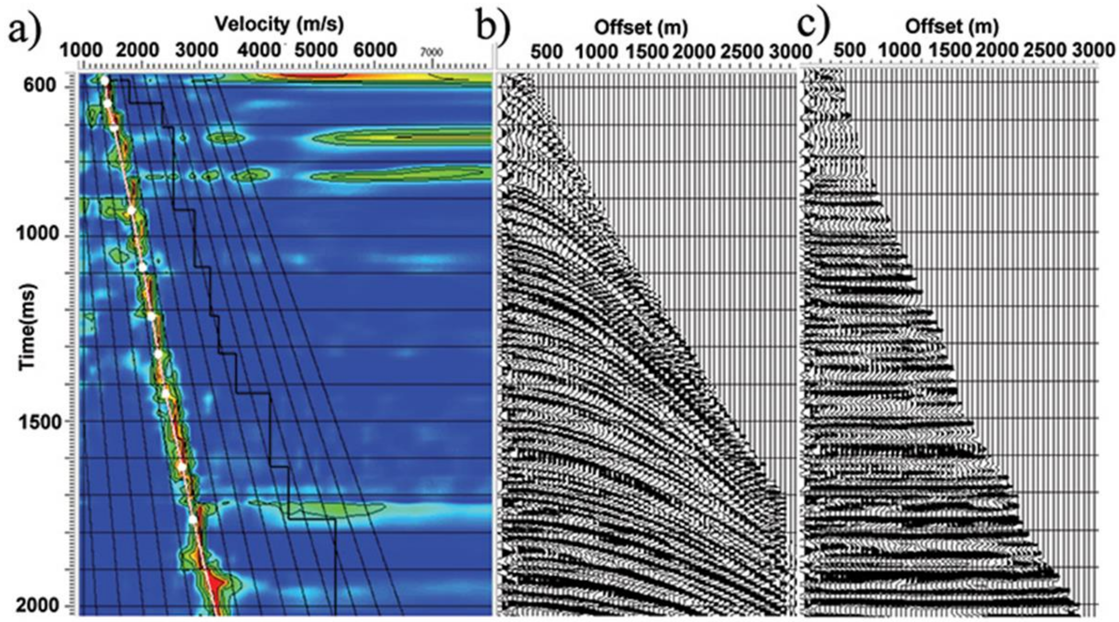


Figure 6.45: *A representative common reflection point gather illustrating (a) velocity analysis of a CMP gather (b) before and (c) after the NMO correction (*http://archives.aapg.org/ explorer/ 2013/01jan/geocorner0113.cfm *,2013).*

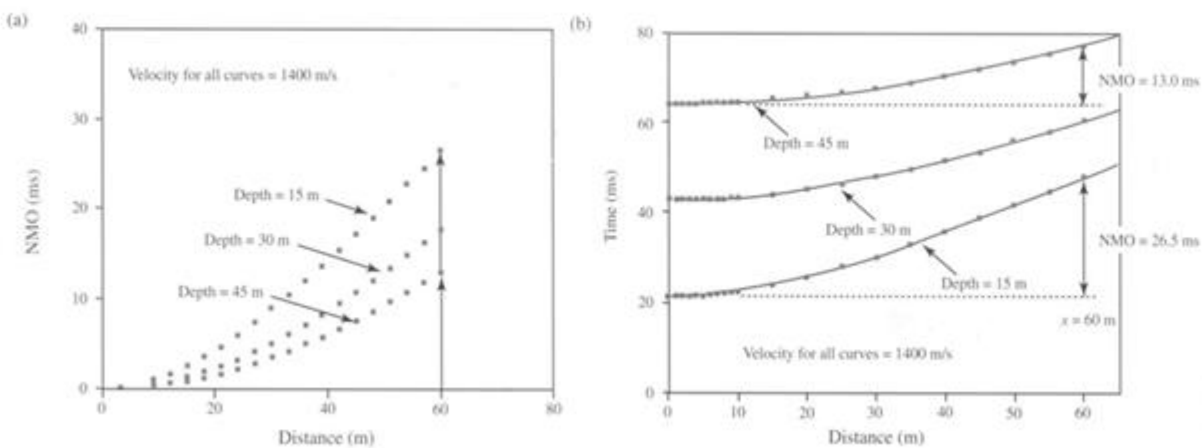


Figure 6.46: *A NMO decreases with increasing depth and with increasing velocity of layers* (Burger et al., 2006*).*

6.6.4.6.CMP gather and corrected velocity

- With the speed of computers, we can iteratively try different velocities and see which value is best (see Figure 6.47).

- We know the velocity is correct when all the reflections are at the same time valve. Reflectors are flat.
- If the velocity is too fast, the reflection curves down. We have not corrected the gather enough (upper right).
- If the velocity is too slow, the reflection curves up. We have over-corrected the gather (lower right).

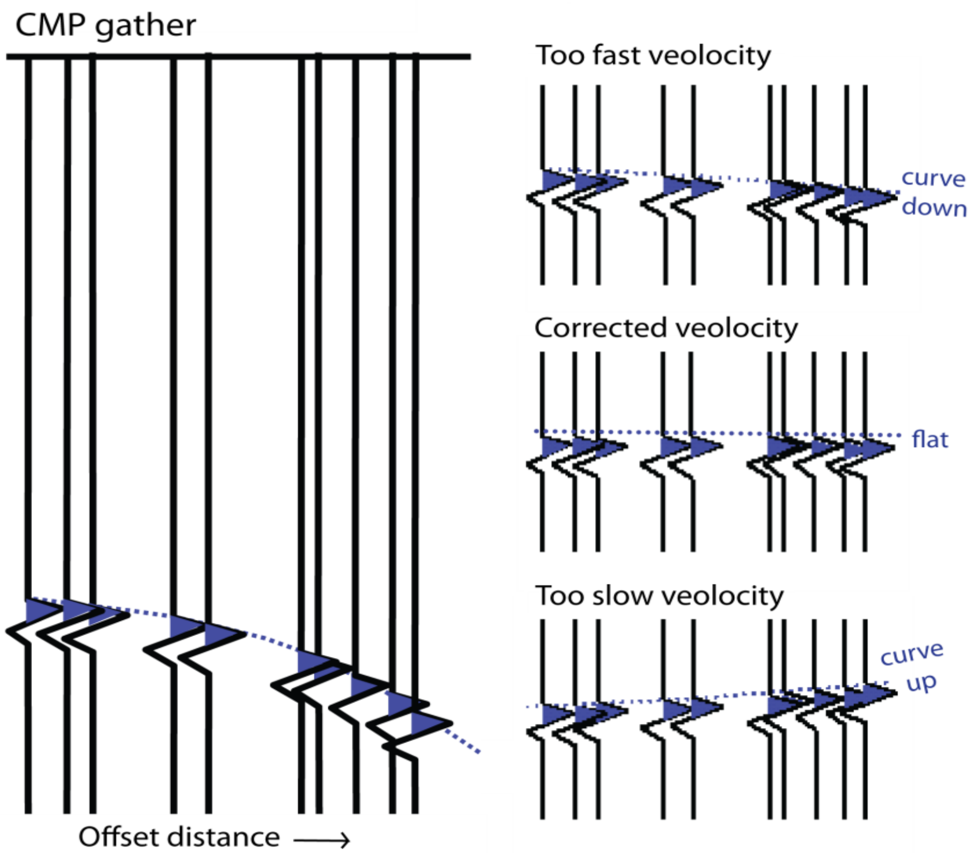


Figure 6.47: *CMP gather and velocity is various values. The corrected velocity will give a flat line with good CMP gather (https://fenix.tecnico.ulisboa.pt/downloadFile/1970943312278156/2_Normal-Moveout%20-Correction.pdf).*

6.6.4.7.Dip moveout

- The difference in the arrival times or travel times of a reflected wave, measured by receivers at two different offset locations, that is produced when reflectors dip. Seismic processing compensates for dip moveout (DMO).
- If we can correct the DMO and achieve the velocity of the dipping layer, the dipping angle can be estimated.

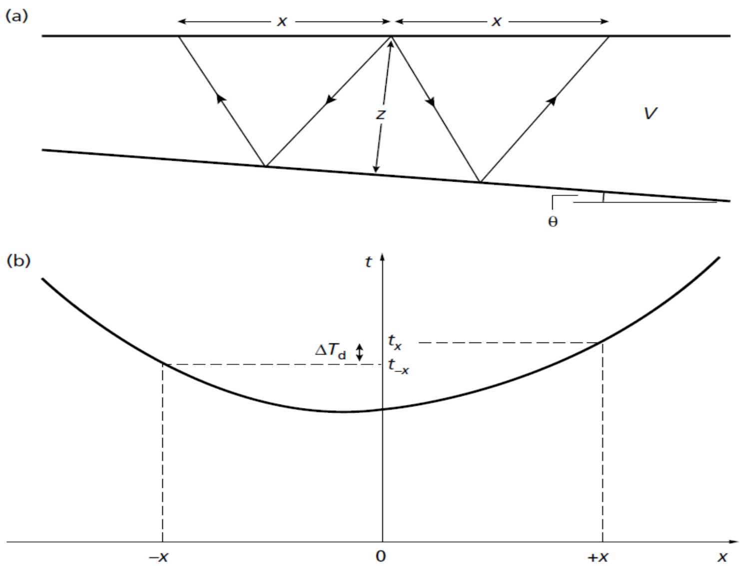


Figure 6.48: *(a) Geometry of reflected ray paths and (b) time–distance curve for reflected rays from a dipping reflector (Kearey et al., 2002).*

6.6.4.8.Static correction

- Static correction is a bulk shift of a seismic trace in time to compensate for the effects of variations in elevation or near-surface weathering thickness (Figures 6.49 and 6.50).
- A common static correction is the weathering correction, which compensates for a layer of low seismic velocity material near the surface of the Earth.

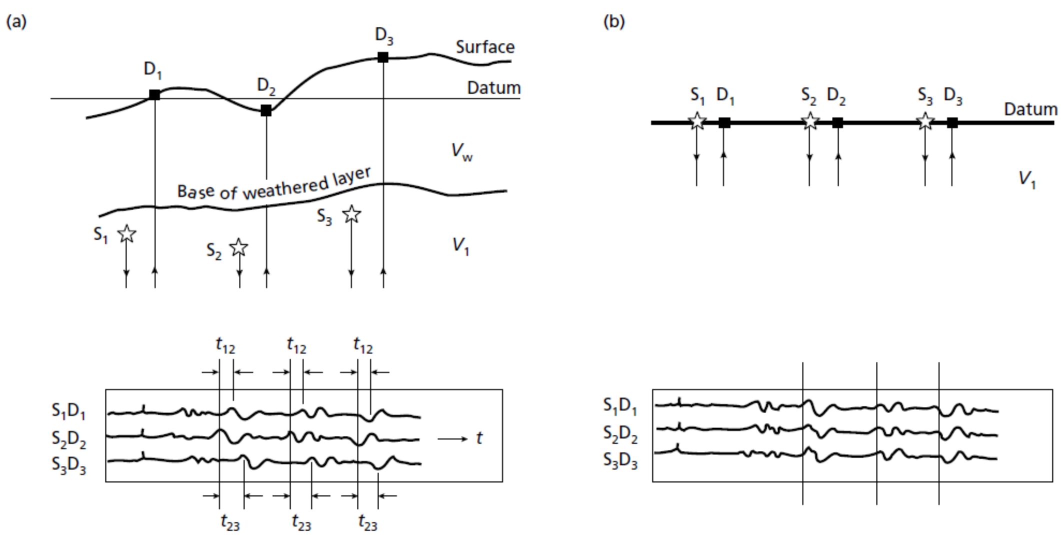


Figure 6.49: *Static corrections. (a) Seismograms showing time differences between reflection events on adjacent seismograms due to the different elevations of shots and detectors and the presence of a weathered layer. (b) The same seismograms after the application of elevation (Kearey et al., 2002).*

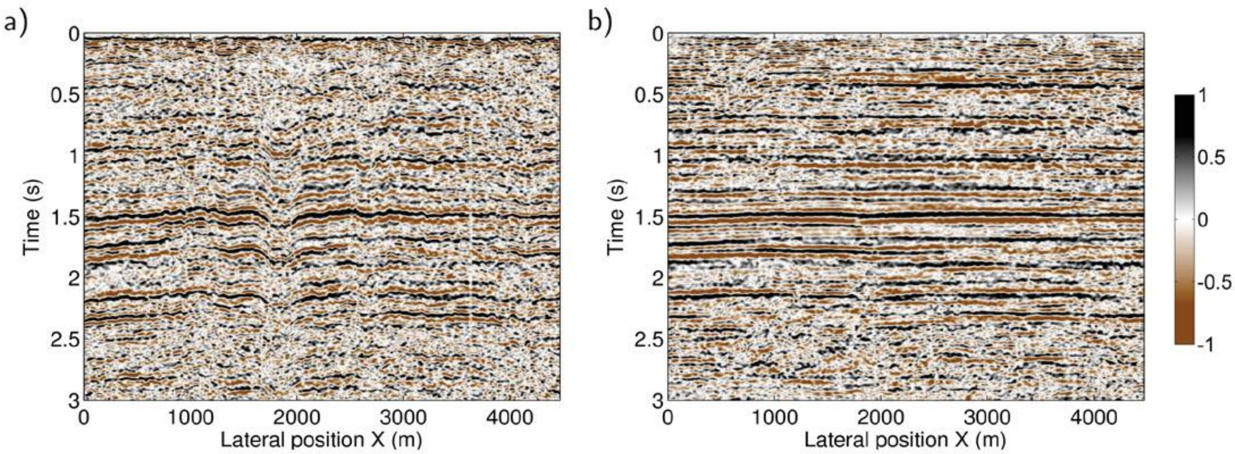


Figure 6.50: *An example of stacked section before (a) and after (b) the static correction. The deformations present on the geometry, especially between x-locations 1800 m, have been removed. (https://csegrecorder.com/articles/view/what-should-multicomponent-near-surface-corrections-look-like).*

6.6.4.9.Stacking

- .A processed seismic record that contains traces that have been added together from different records to reduce noise and improve overall data quality (Figures 6.51 and 6.52).
- Numbers of traces that have been added together during stacking is called the fold.

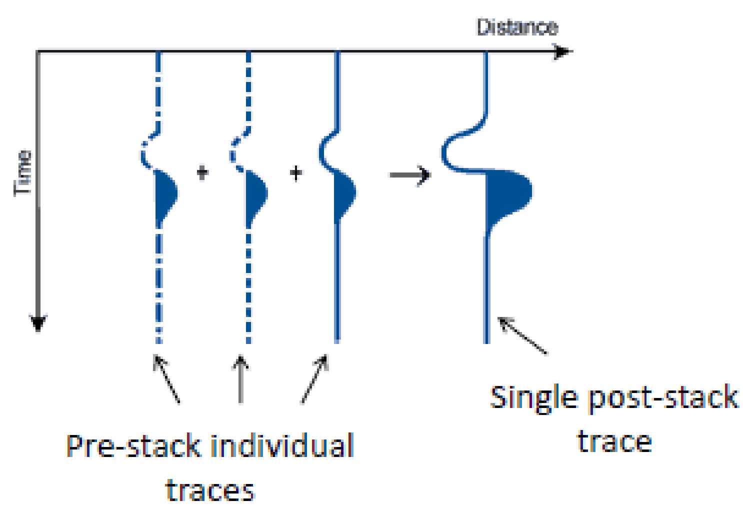


Figure 6.51: *Seismic g stacking result in redundancy of the data that improves the signal-to-noise ratio (https://commons.wikimedia.org/wiki/File:Simple_diagram_illustrating_the_effect_of_seismic_stacking.png, 2012).*

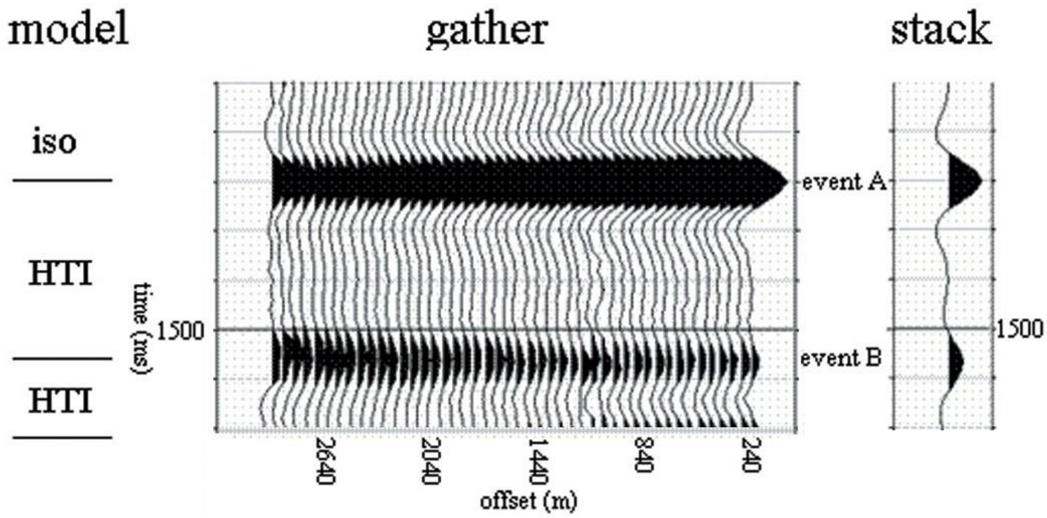


Figure 6.52: *A seismic gather and its stack. There are two events with different amplitude of reflectors (https://csegrecorder.com/articles/view/avo-projected-pilot-trace-for-dynamic-trim-statics)***.**

6.6.4.10. Seismic migration

- Seismic migration is the process by which seismic events are geometrically re-positioned in either space or time to the location the event occurred in the subsurface rather than the location that it was recorded at the surface (Figure 6.53).
- Normally, reflections are not positioned in the subsurface correctly since they have dip.
- Migration moves the reflections deeper and further updip to create a more accurate image of the subsurface.
  - Low dips give slight corrections moving events deeper and updip (Figure 6.54B).
  - High dips give larger corrections moving events deeper and updip (Figure 6.54B).

Bow tie

- A concave-upward event in seismic data produced by a buried focus. The name was coined for the appearance of the event in unmigrated seismic data (Figure 6.55, left).
- Synclines commonly generate bow ties.
- The bowtie can be corrected by proper migration method (Figure 6.55, right).

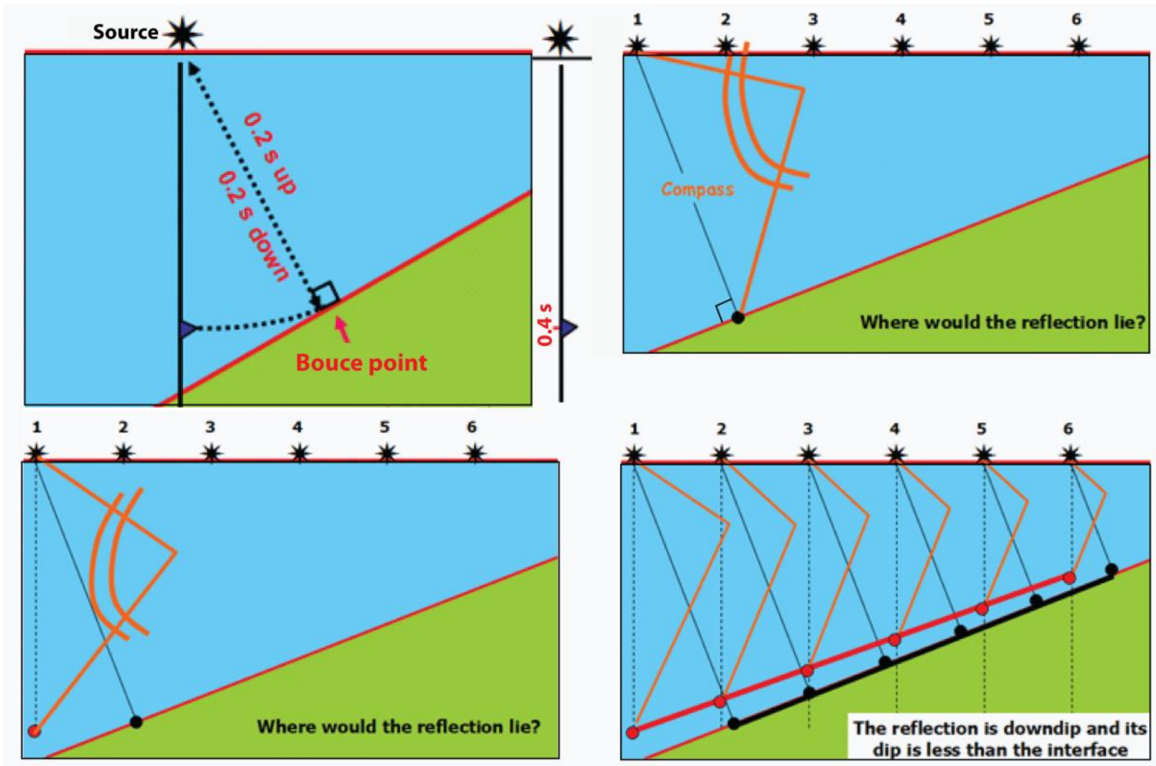


Figure 6.53: *Positioning problems where reflections are not positioned in the subsurface correctly since they have dip. This is what you should have found the real reflection surface is the black line but the seismic reflection is displayed on seismic section where the red line is located (https://en.wikipedia.org/wiki/Seismic_migration).*

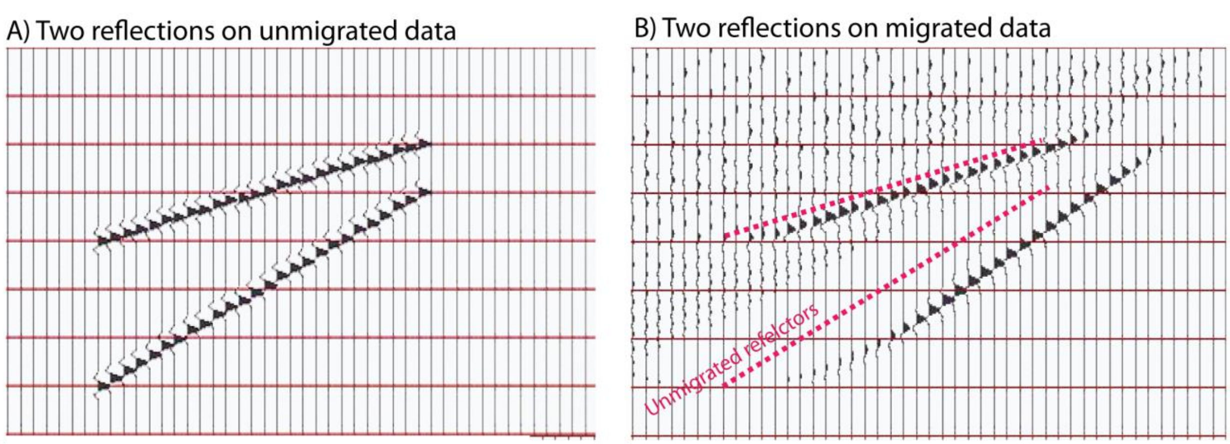


Figure 6.54: *Two reflections on dipping interfaces A) before and B) after migration. Constructive interference occurs where the reflections are properly positioned (black reflectors) and Destructive interference dominates where the reflections are not properly positioned (red dots) (modified from http://slideplayer.com/slide/8567850/).*

6.6.4.11. Multiple reflections

- Seismic return to the surface after reflection at a single interface, known as primary reflections.
- There are many paths in a layered subsurface by which rays may return to the surface after reflection at more than one interface. Such rays are called multiples (Figure 6.56).
- Two types of multiple that are reflected at interfaces of high reflection coefficient with strong amplitudes comparable with primary reflections:
  - **Ghost reflections:** seismics from a buried explosion on land are reflected back from the ground surface or the base of the weathered layer. They arrive a short time after the primary (Figure 6.57).
  - **Water layer reverberations:** seismics from a marine source are repeatedly reflected at the sea bed and sea surface.
- Results of the multiple corrections on a seismic section are shown in Figure 6.58.

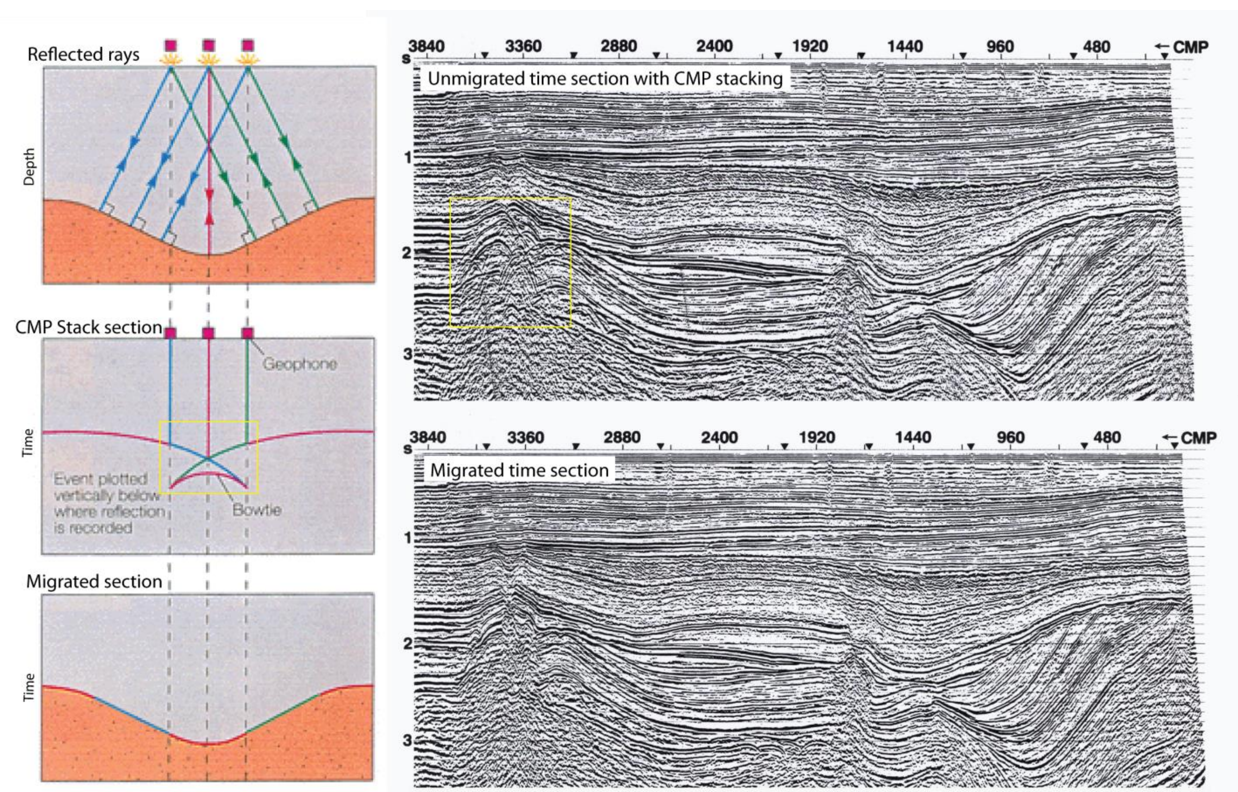


Figure 6.55: *(left) Diagram of bowtie events (yellow rectangle) on unmigrated seismic data. (Right) Comparison of seismic time section associated with the CMP gathers before and after migration (modified from* http://www.glossary.oilfield.slb.com/Terms/b/bow_tie.aspx*).*

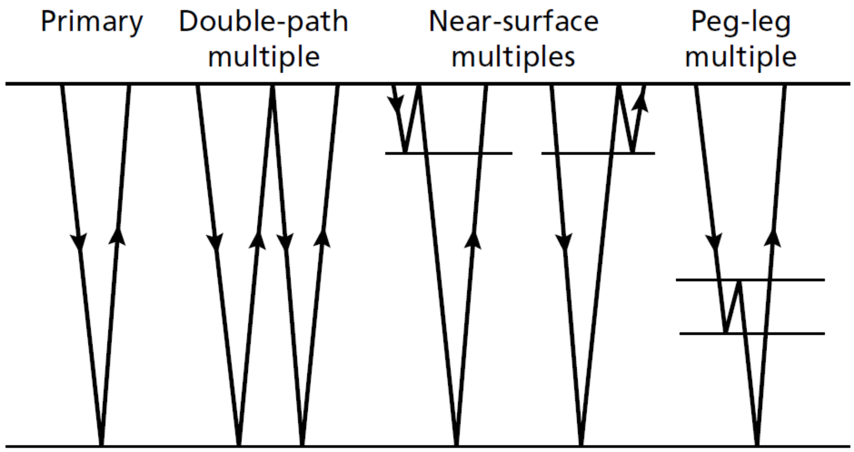


Figure 6.56: *Various types of multiple reflections in a layered ground elevation (Kearey et al., 2002).*

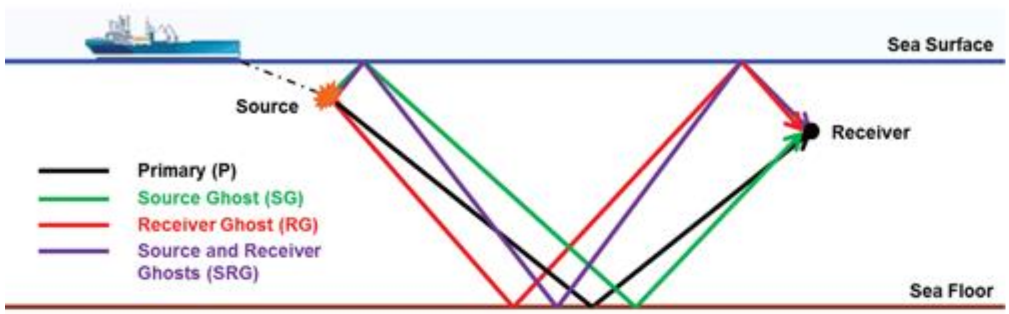


Figure 6.57: *Ghost reflections from both source and receiver sides (https://www.offshore-mag.com/articles/print/volume-75/issue-11/geology-geophysics/processing-based-broadband-enhances-image-quality-in-frontier-areas.htm).*

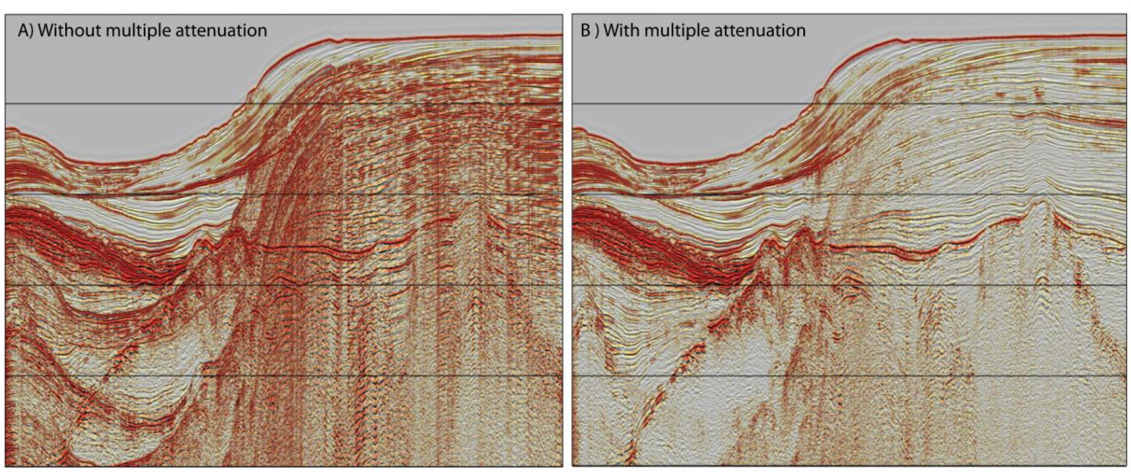


Figure 6.58: *Comparison of seismic section without and with multiple corrections* (https://www.cgg.com/en/What-We-Do/Offshore/Customer-Challenges/ImproveEfficiency/*Dovetail).*

6.7. Seismic reflection interpretation

- Seismic reflections come from interfaces where the acoustic properties of the rocks change, and this fact is the basis of our understanding of the nature of seismic data (Figure 6.59).
- Acoustic impedance of a rock layer is the product of the density and the velocity of that layer, and strictly a reflection is generated by a contrast in acoustic impedance.
- In fact impedance and lithology normally follow each other, so that impedance boundaries and lithologic boundaries normally concur.
- Interpreting seismic data requires an understanding of the subsurface formations and how they may affect wave reception.

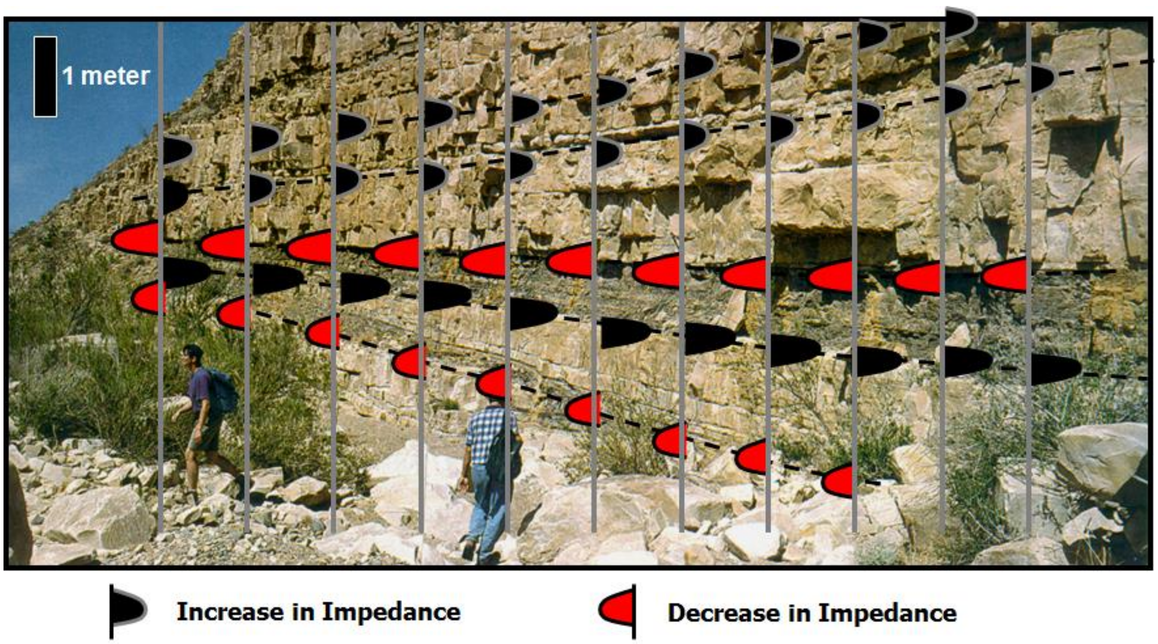


Figure 6.59: *The ideal seismic response (*http://archives.aapg.org/*slide_resources/schroeder/6/ index.cfm).*

6.7.1. Vertical data resolution

- Seismic resolution is a measure of minimum spatial or temporal separation between two reflection events so that they can be distinguished and resolved separately.
- Data resolution in vertical direction, thickness have to be larger than the half of wavelength (See Figure 6.60 for explanation).

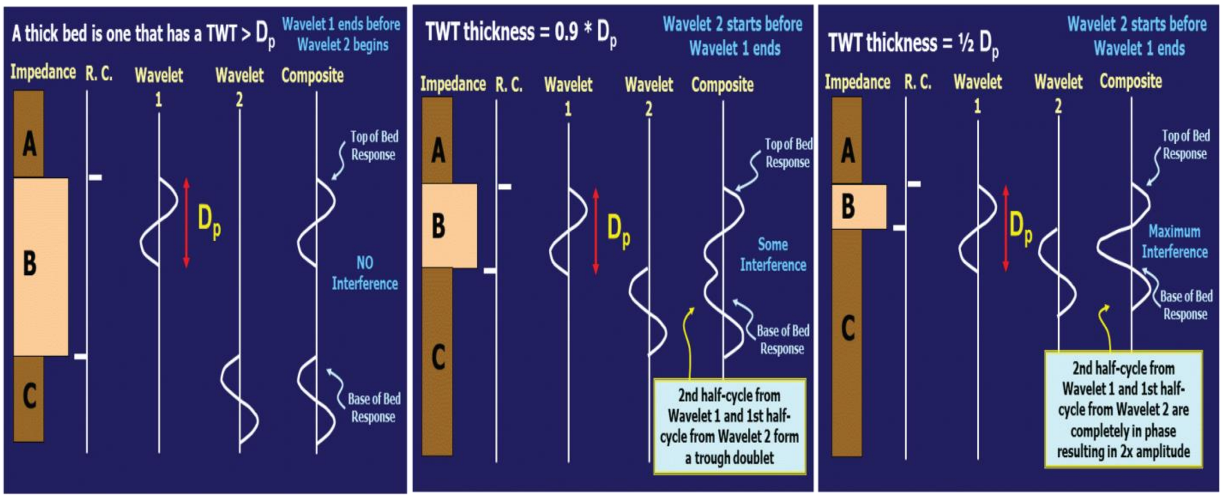


Figure 6.60: *TWT thicknesses indicating a vertical data resolution (*http://archives.aapg.org/ *slide_resources/schroeder/6/index.cfm).*

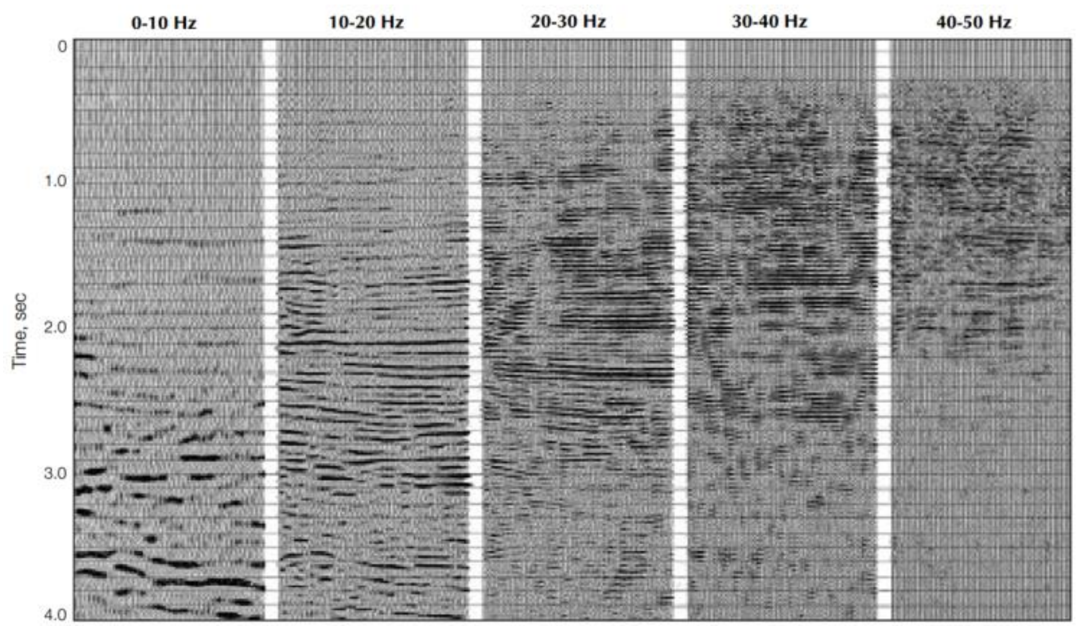


Figure 6.61: *Filtered 2D data showing frequency content variation (0-10 to 40-50 ha) with depth. Lower frequencies penetrate deeper, but higher frequencies give better vertical resolution data (Telford et al., 1990).*

6.7.2. Well to seismic

- The more control the geoscientist has in mapping the subsurface, the greater the accuracy of the depths. Control can be increased by the correlation of seismic data with borehole data. The synthetic seismogram is the primary means of obtaining this correlation.

- Velocity data from the sonic log and the density log are used to create a synthetic seismic trace (Figure 6.62).
- A synthetic seismogram is the result of forward modelling the seismic response of an input earth model (Figure 6.62).
- This trace closely approximates a trace from a seismic line that passes close to the well in which the logs were acquired. The synthetic then correlates with both the seismic data and the well log from which it was generated (Figure 6.62).

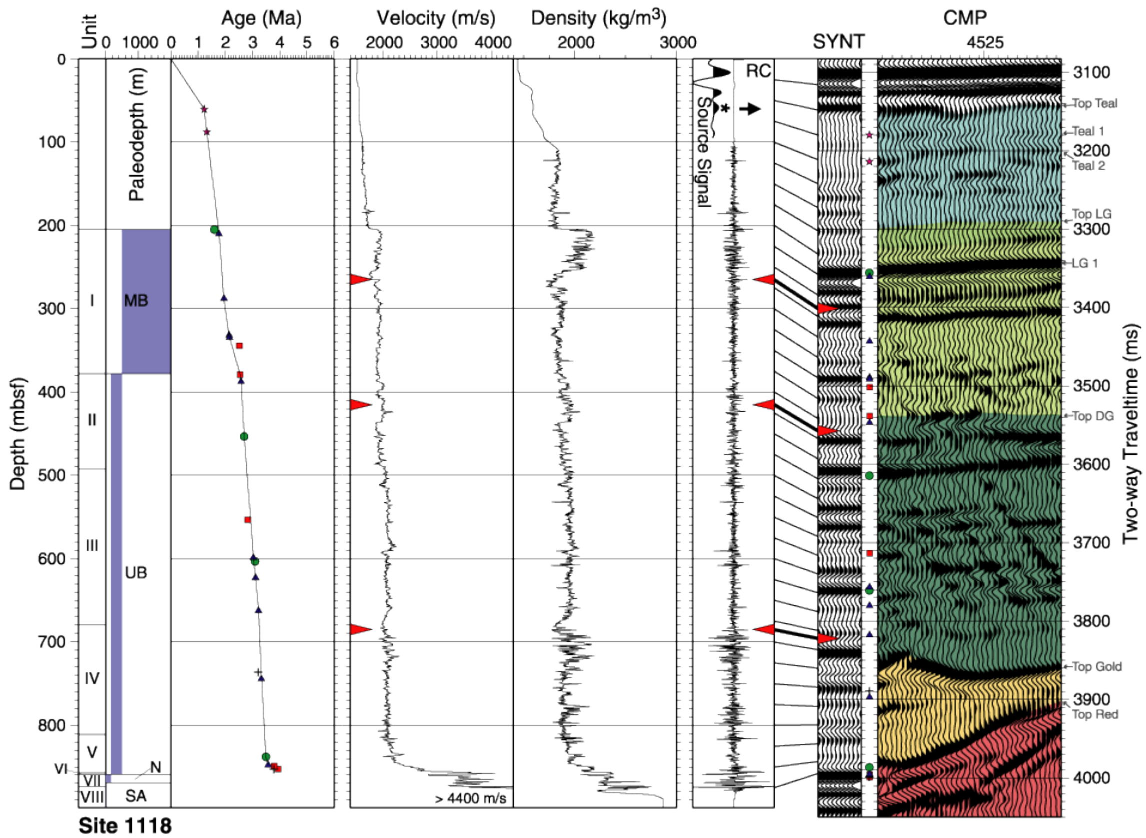


Figure 6.62: *Logging and coring results check shot-corrected depth conversion, and their correlation via synthetic seismogram (*http://www-odp.tamu.edu/publications/180_SR/167/167_f3.htm*, 1999).*

6.7.3. Concept of seismic interpretation

- It is important for the interpreter to have a basic understanding of what tectonic influences and depositional systems occur within the area of the seismic survey to be investigated.

- Although this preconceived Earth model may be vague and incomplete, particularly in frontier basins, it provides interpreters with insight and constraints as to what types of structures, faulting, and stratigraphic geometries may exist.
- The interpretation of fault styles, structural geometries, and facies patterns must be consistent with regional tectonic forces and basin infilling.

6.7.4. Seismic interpretation workflow

- Data gathering and loading.
- Well to seismic calibration/tie.
- Investigation of seismic section and horizon selection.
- Fault determination and correlation.
- Picking of reflector time (TWT) over seismic grid.
- Contouring time grid.
- Time to depth conversion using velocity.
- Identify potential i.e. closures, prospects, leads.

6.7.4.1. Structural interpretation

- Seismic interpretation of structural styles is the first geologic approach to the interpretation of seismic data. The main application of structural analysis of seismic sections is in the search for structural traps containing hydrocarbons.
- Discontinuity of seismic characters is caused by applied stresses in the subsurface system. That stresses control the structural styles appeared on seismic section.
- Major structural styles are fault and folds (Figures 6.63-6.65).

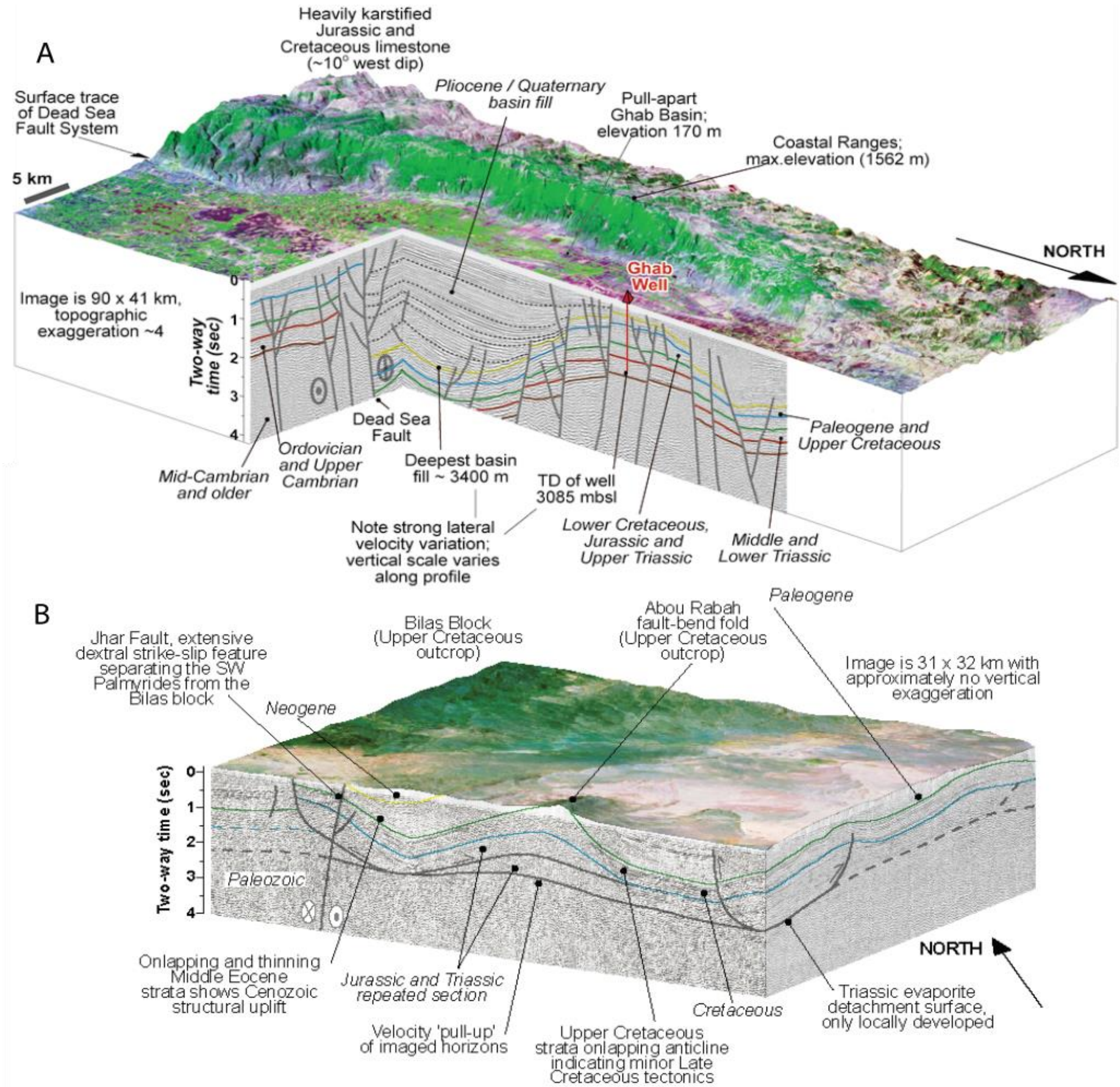


Figure 6.63: *A) Normal and strike-slip fault and B) associated anticlinal structures on 3D seismic block model for Coastal Ranges / Ghab Basin along the Dead Sea Fault System in western Syria (Brew et al., 2001).*

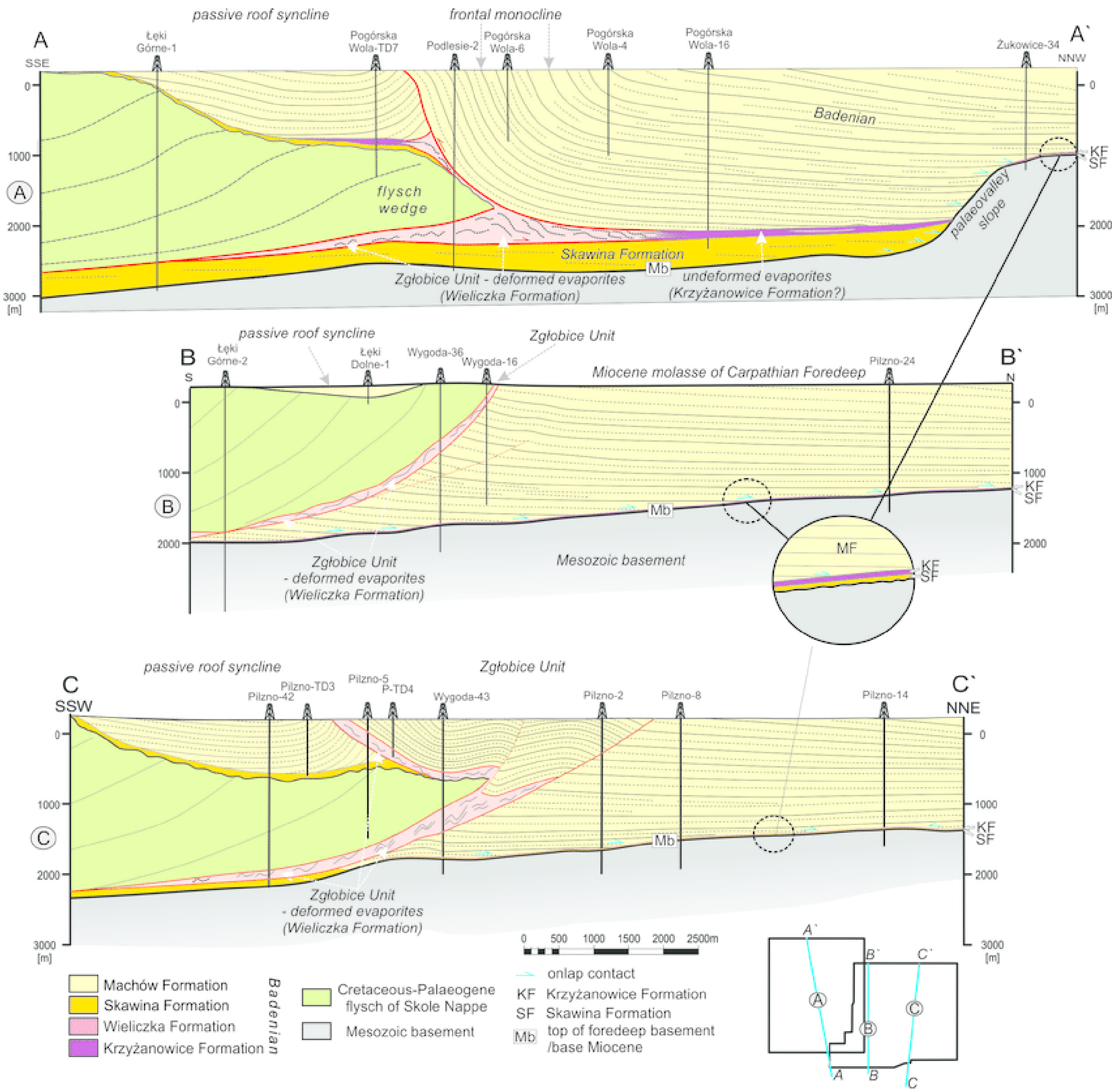


Figure 6.64: *Wedge structure with thrust on an interpreted seismic section. The evaporate succession is split, tectonically thickened and deformed in front of the flysch Skole Nappe (https://www.researchgate.net/publication/304066028).*

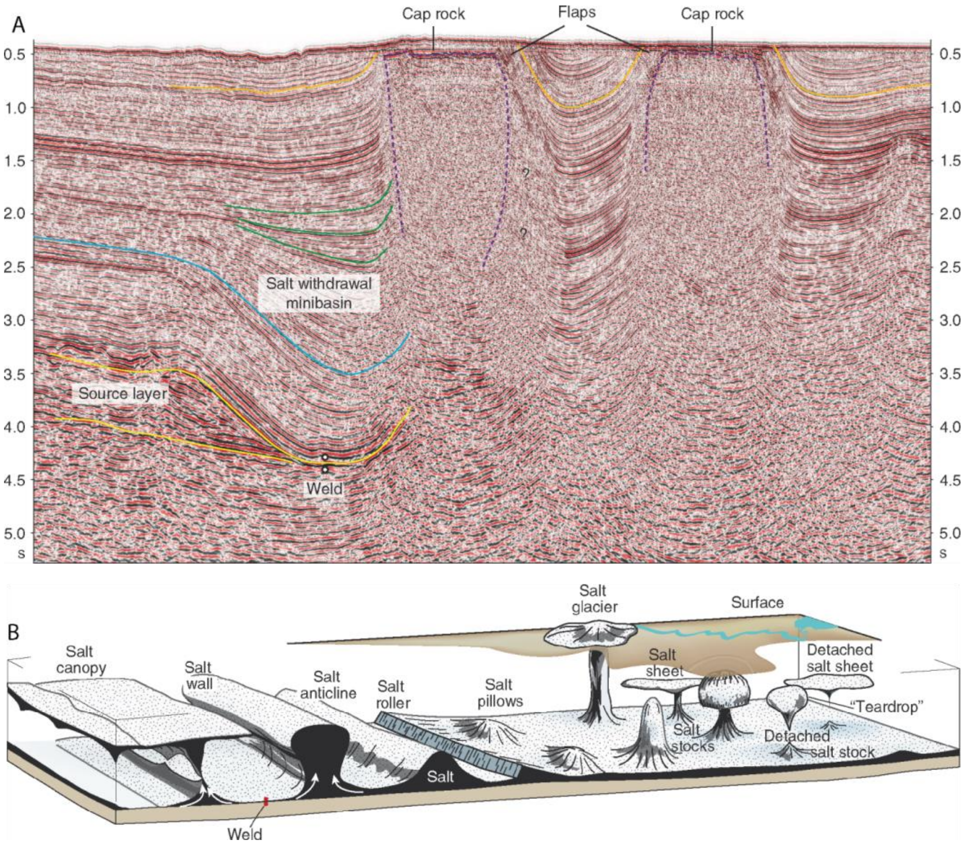


Figure 6.65: *A) Salt anticline interpretation on seismic section. B) Formations of salt rheology (http://sanuja.com/blog/exams/glgy341-final).*

6.7.4.2. Stratigraphic interpretation

- Seismic stratigraphy is basically a geologic approach to the stratigraphic interpretation of seismic data. Seismic reflections allow the direct application of geologic concepts based on physical stratigraphy (See Figure 6.66).
- Stratigraphic Interpretation composes of;
  - Stratal surface: seismic reflections parallel stratal surfaces.
  - Unconformities: Reflection terminations mark unconformities. Igneous and salt intrusions (Figure 6.67) make the unconformity of interest for petroleum prospects.
  - Lithologic facies and sequences: Changes in reflection character indicate facies changes within sedimentary sequences (Figure 6.68).
  - Physical properties associated with sedimentation process such as porosity, fluid, and pressure (Figure 6.69).

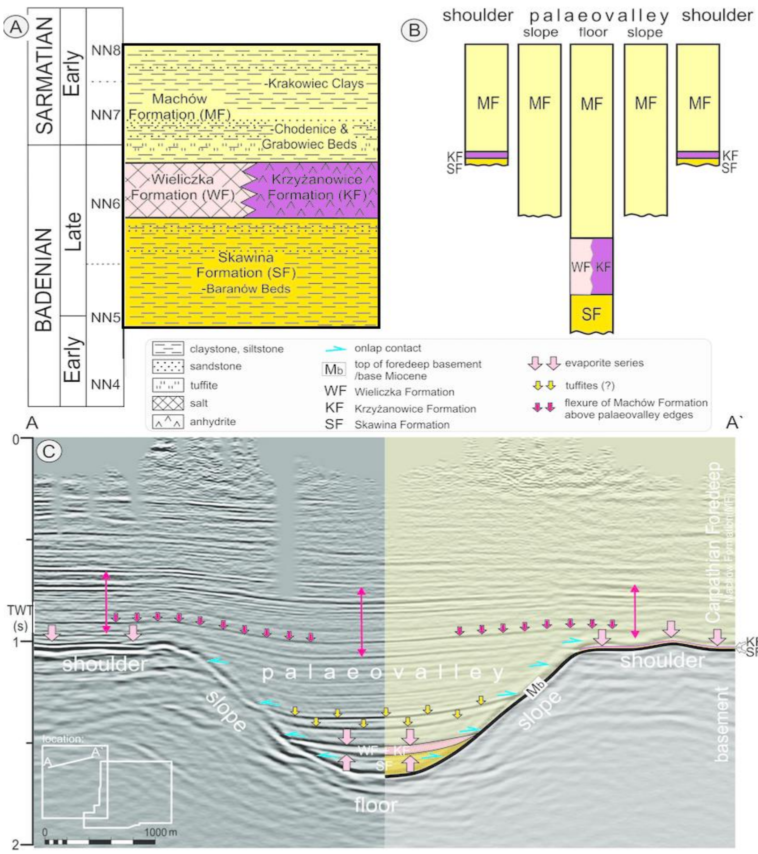


Figure 6.66: *A) Stratigraphy of the Miocene deposits in the Carpathian Foredeep near to Pilzno. B) Variations at different palaeotopographic positions in the Pogórska Wola palaeo valley. C) Seismic section transverse to the palaeo valley axis (*https://www.researchgate.net/publication/304066028*).*

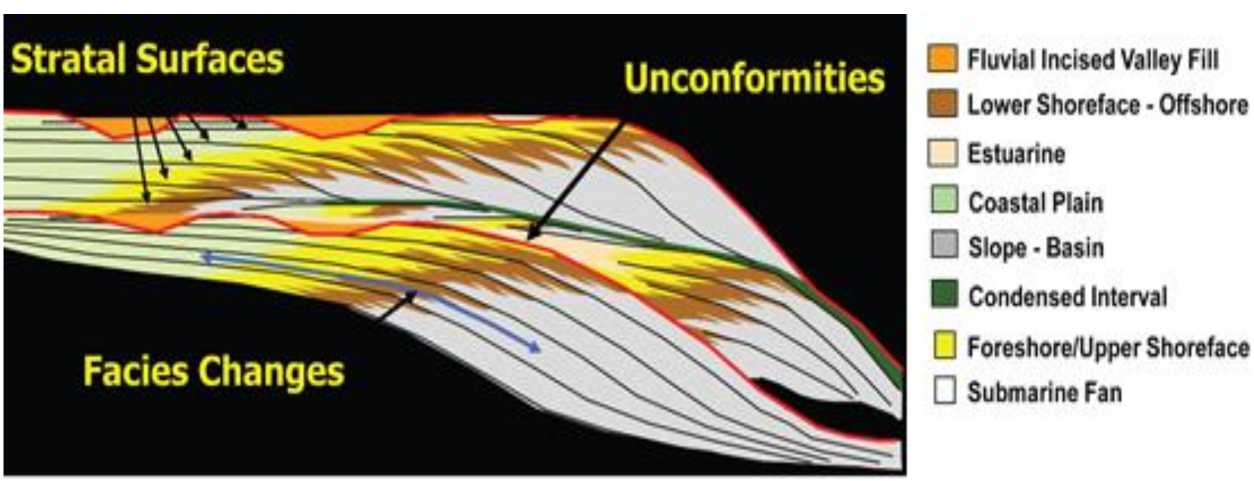


Figure 6.67: *Different types of geological boundary and facie changes defining seismic sequences (*http://archives.aapg.org/slide_resources/schroeder/6/index.cfm*).*

Dark grey marls, black organic-rich mudstones	Siltstones, shales, reef limestones	Calcareous sandstones, oolites, bioclastic limestones	Sandstones, mudstones, dolomitic mudstones, evaporites	**Typical lithologies**
Very thin and continuous units	Thin to intermediate tabular bodies with lensoid reef limestones	Intermediate continuous to lensoid bodies	Irregular to discontinuous units	**Bed geometry**
25–50	150–450	100–50	50–25	**Thickness (m)**
Basinal	Outer shelf	Inner shelf	Coastal	**Environment**

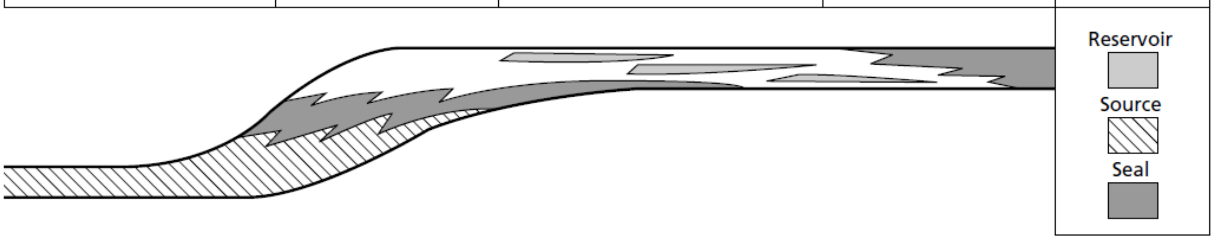


Figure 6.68: *The overall geometry of a typical depositional sequence and its contained sedimentary facies (modified from Kearey et al., 2002).*

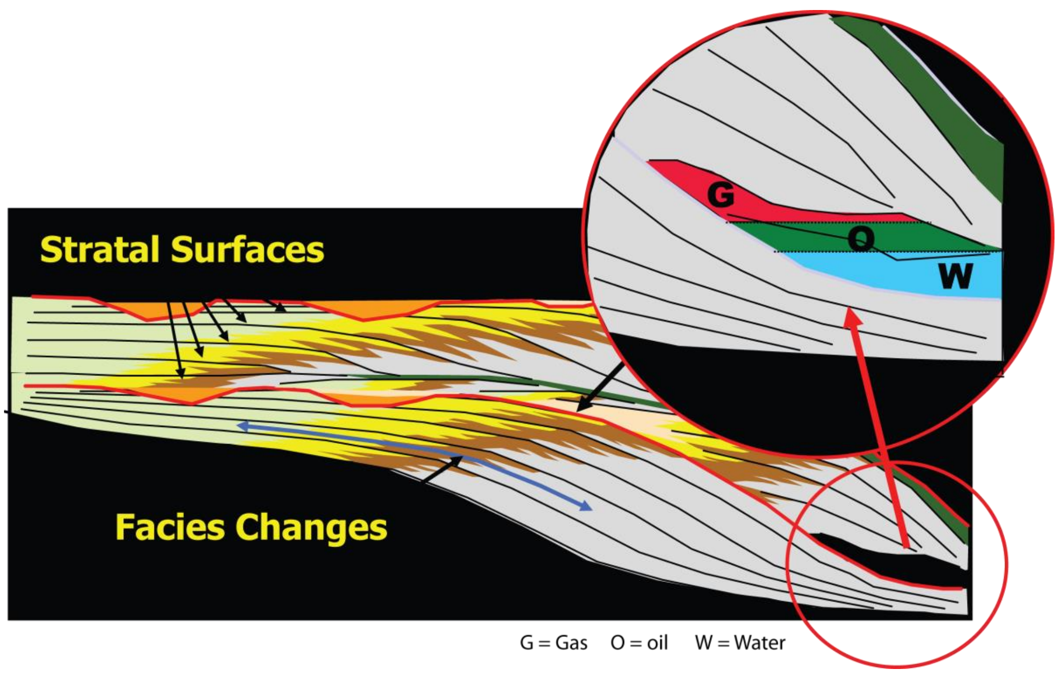


Figure 6.69: *Interpreted seismic sequences showing fluid contacts where the hydrocarbon is trapped (9http://archives.aapg.org/slide_resources/schroeder/6/index.cfm).*

6.7.4.3. Direct hydrocarbon indicator

- Direct hydrocarbon indicator (DHI), is an anomalous seismic attribute value or pattern that could be explained by the presence of hydrocarbons in oil or gas reservoir.
  - **Bright spots:** localized amplitudes of greater magnitude than background amplitude values. Oil/gas increases the acoustic impedance contrast (Figure 6.70A, left).
  - **Flat spots:** nearly horizontal reflectors that cross existing stratigraphy, possibly indicating a hydrocarbon fluid level within an oil or gas reservoir.
  - **Dim spots:** low amplitude anomalies. The oil/gas decrease the acoustic impedance contrast (Figure 6.70A, right).
  - **Polarity reversals** can occur where the capping rock has a slightly lower seismic velocity than the reservoir and the reflection has its sign reversed (Figure 6.70A, middle).
- Some geoscientists regard amplitude versus offset anomalies as a type of direct hydrocarbon indicator. For example, the amplitude of a reflection might increase with the angle of incidence, a possible indicator of natural gas.

6.7.5. Scale of stratigraphic correlation

- Using outcrops or cores.
  - Visually correlate laminae and beds.
  - Units are centimeters thick.
- Using well logs.
  - Pattern correlation of log markers.
  - Units are meters thick.
- Using seismic data.
  - Seismic correlation of bed sets and larger units.
  - Units are tens to hundreds of meters thick.

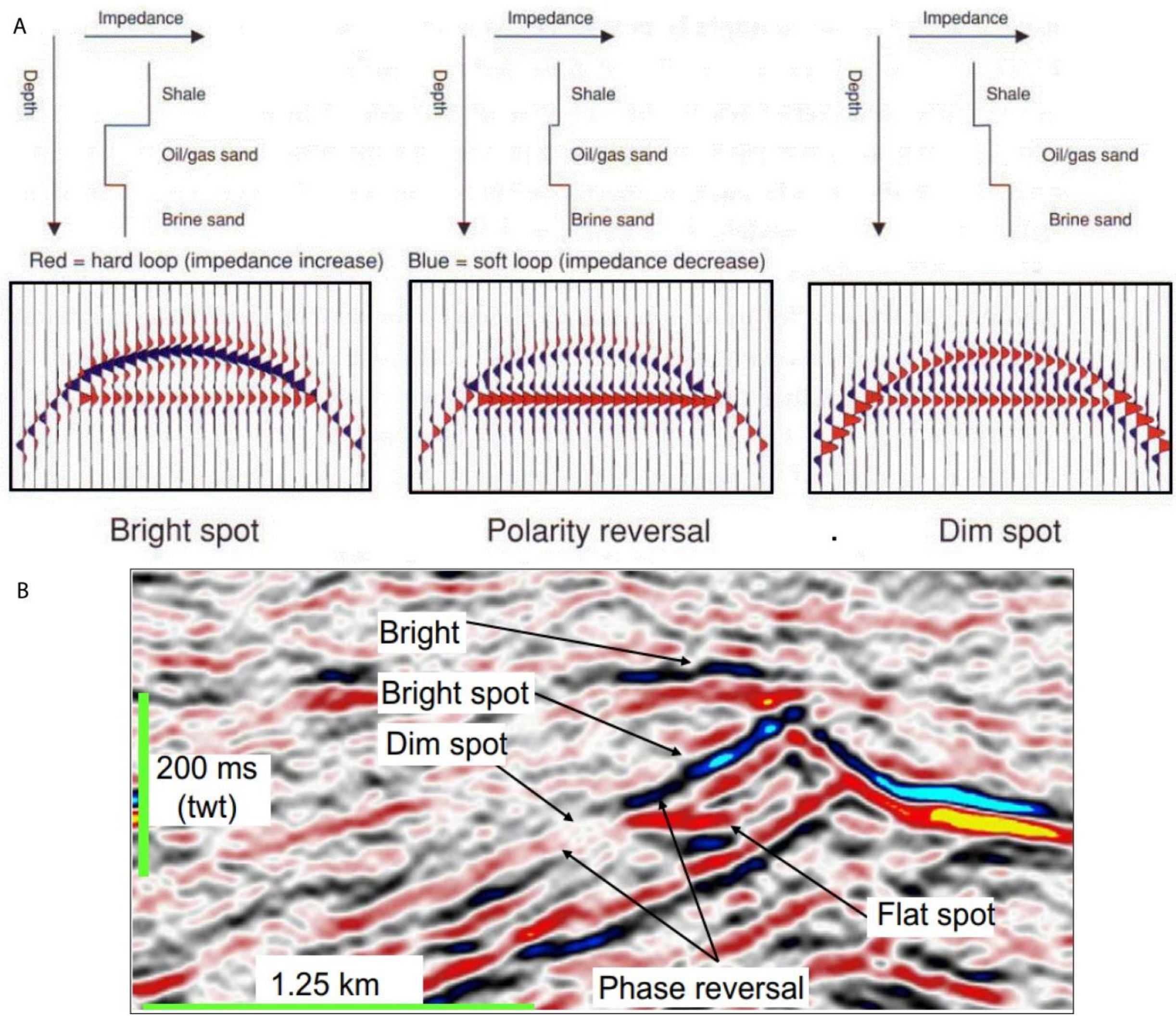


Figure 6.70: *Anomalous seismic attribute patterns of bright spot, polarity reversal, and dim spot on seismic section shows a presence of hydrocarbons in oil or gas (modified from Brown, 2010). B) The anomalies interpreted* as hydrocarbon in a reservoir. Note also the increased amplitude and continuity in a cap rock above the hydrocarbon-filled reservoir *(Løseth et al., 2009).*

6.8. 3D Seismic data

- A set of numerous closely-spaced 2D seismic lines that provide a high spatially sampled measure of subsurface reflectivity. Typical receiver line spacing can range from 300 - 600 m and a typical distance between shot point and receiver groups is 25 m.

- The resultant data set can be "cut" in any direction but still display a well sampled seismic section. The display of 3D seismic data allows for more thorough analysis and attributes functions.

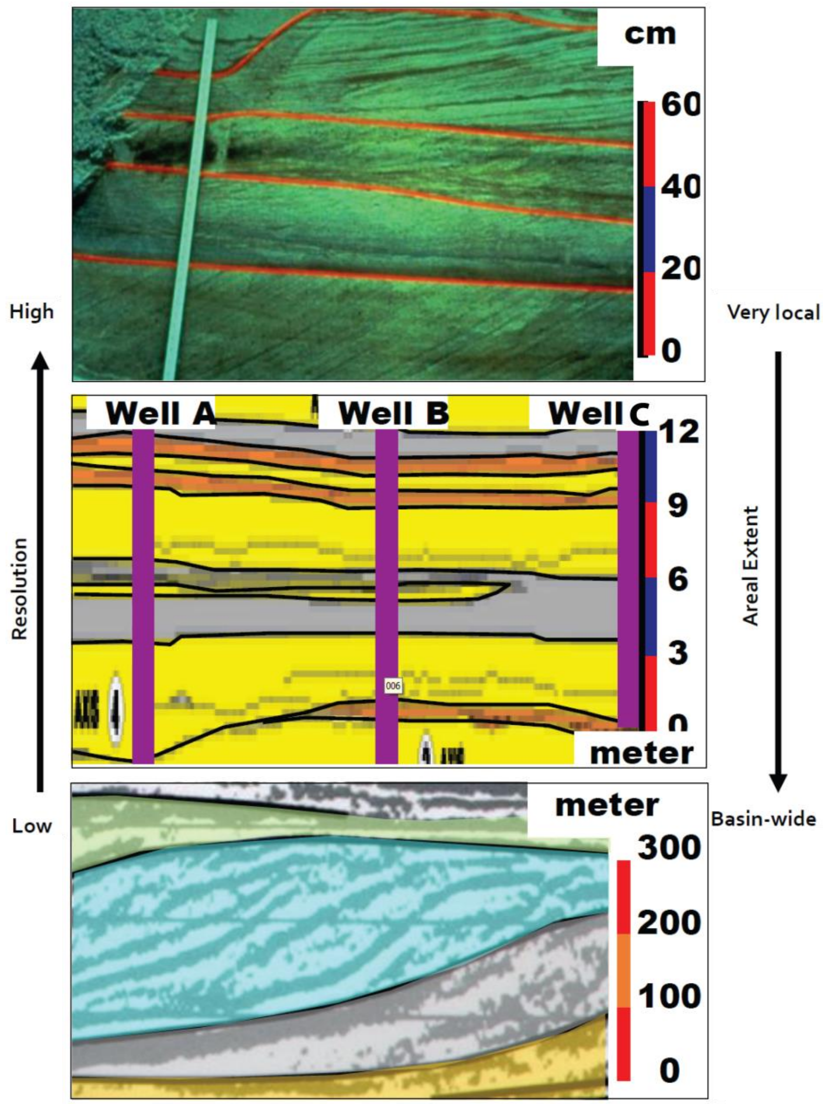


Figure 6.71: *Scale and resolution of stratigraphic correlations based on outcrop, logs, and seismic profile data* gas (*modified from Brown, 2017*).

- 3D seismic data can provide more accurate subsurface maps and more detailed information about subsurface structures and stratigraphic features.

6.8.1. Time slice map

- Time slice map is an improvement over vertical sections to display a horizontal map view at a certain arrival time (Figure 6.72A).

- A time slice is a convenient tool to evaluate changes in amplitude of seismic data.
- A time-slice map is useful for the interpretation of depositional systems (Figure 6.72C) because it provides the opportunity to see a portion of depositional systems in map view. This view is a key to interpreting faulting systems because it allows a view of the offset of the faults (Figure 6.73).

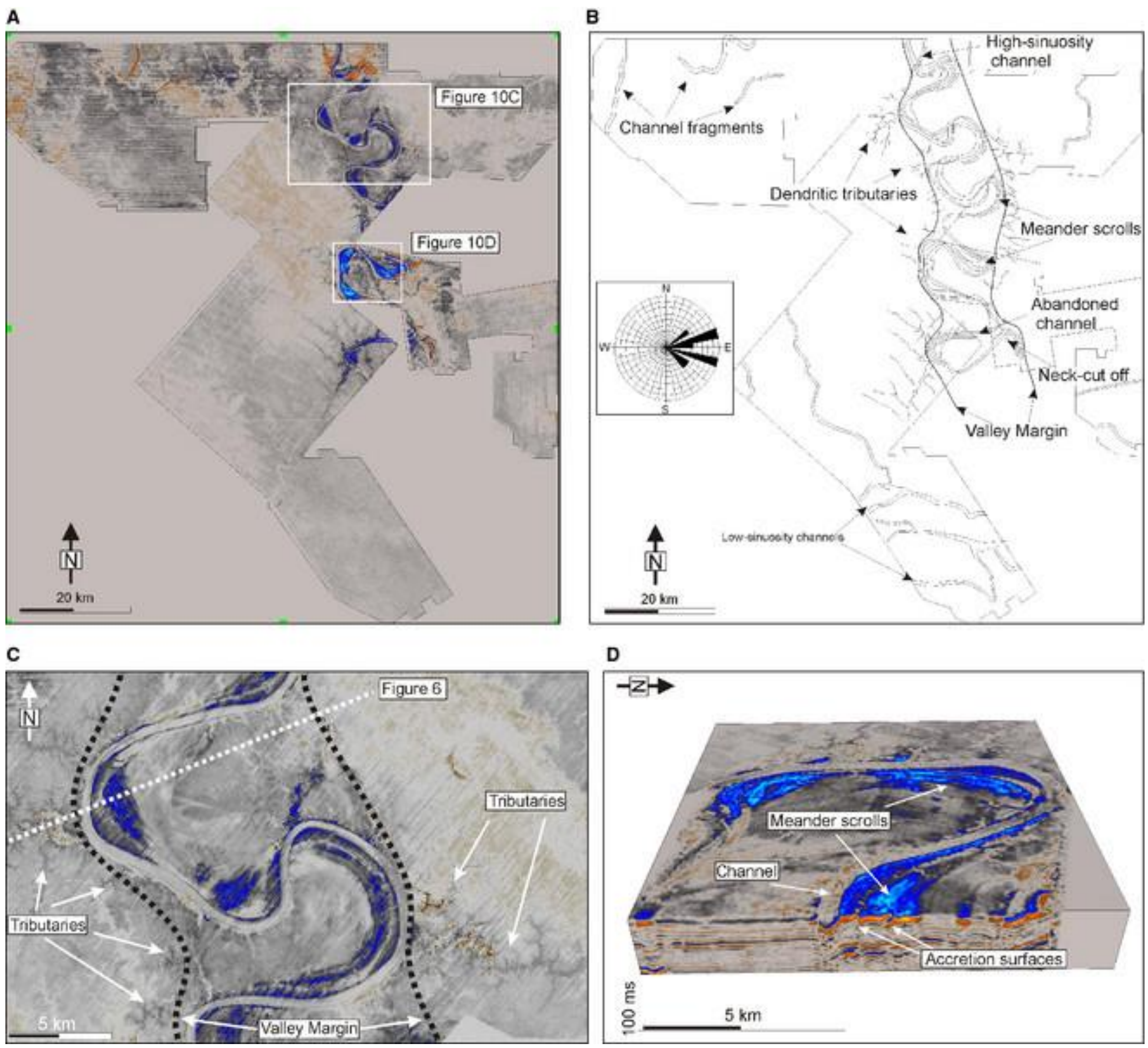


Figure 6.72: *A) Time-slice at 108 TWT and (B) interpreted map-view. These diagrams show a large highly sinuous meandering channel (C) Close-up image showing the tributaries on the margins. (D) Three-dimensional seismic cube showing the progression of the meandering channel (*https://www.researchgate.net/publication/269726545*).*

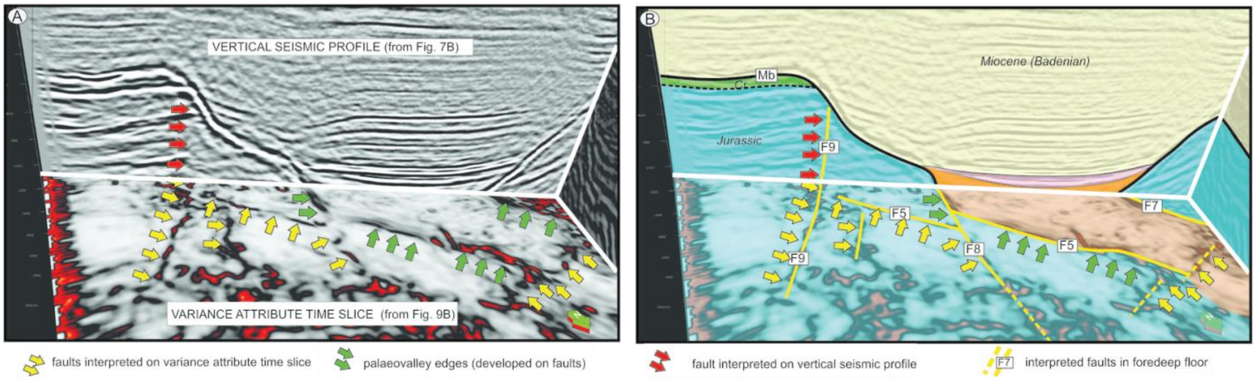


Figure 6.73: *3D diagram combining vertical seismic sections with a horizontal time slice map showing four sets fault interpretations* (https://www.researchgate.net/publication/ 304066028).

6.8.2. Structural maps

- A structural map is a type of subsurface maps whose contours represent the elevation of a particular reflector of subsurface formation or structural markers (e.g. folds, faults, salts, and volcanoes) (Figures 6.74 and 6.75)
- A map identifying the seismic image times at which subsurface structure is located. Time-structure maps can be converted into depth-structure maps if seismic propagation velocity can be defined throughout 3D seismic image space.

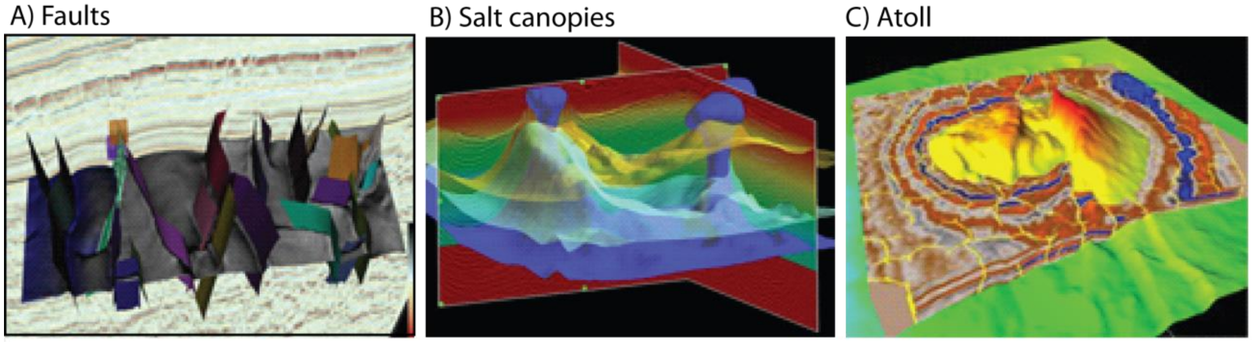


Figure 6.74: *Structural maps and interpretations of normal (A) faults, (B) salt canopies and (C) limestone atoll* (https://www.researchgate.net/publication/ 304066028).

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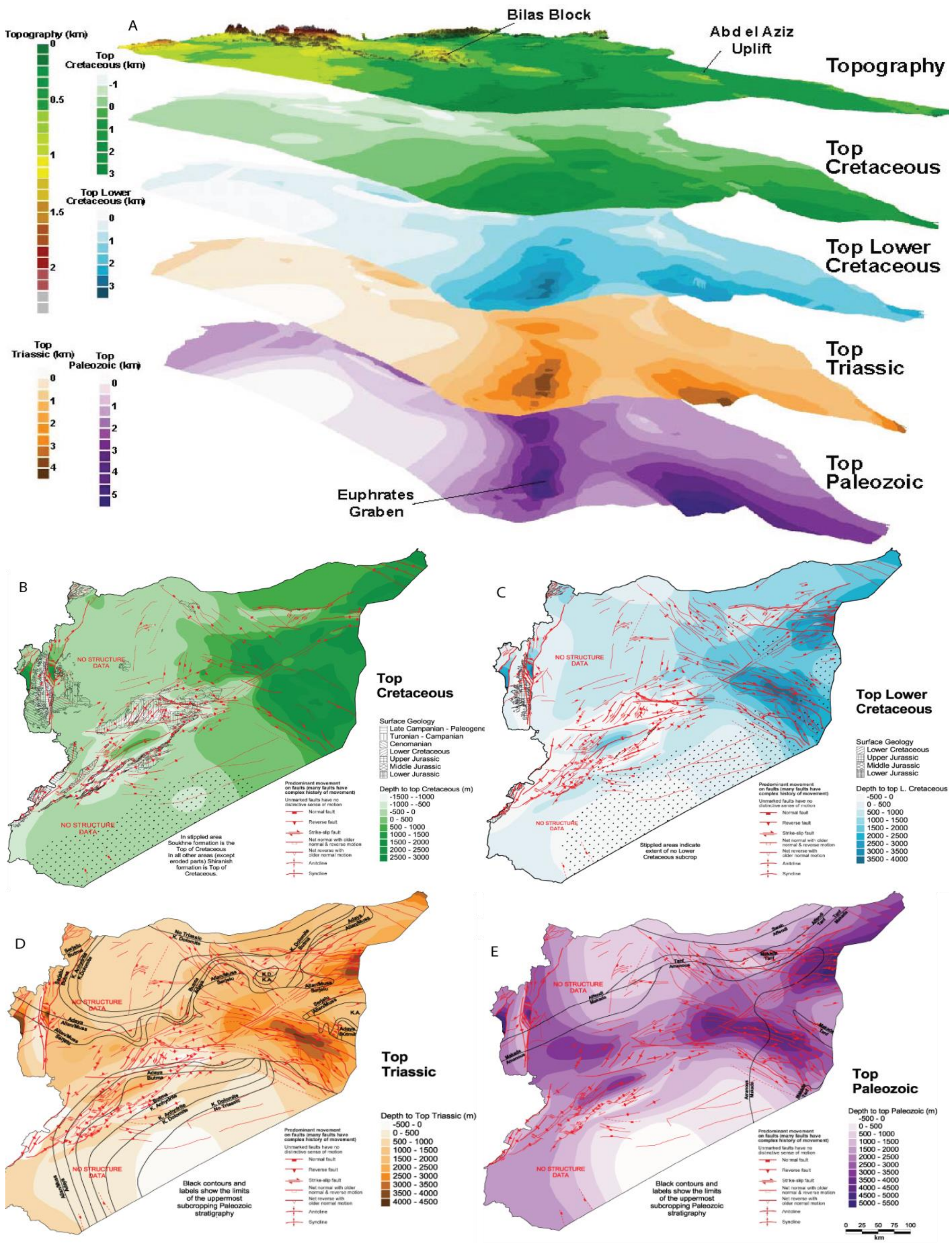

Figure 6.75: *A) Four structural map* derived from seismic *and B-E) interpretations representing top layer of subsurface rock formations (Brew et al., 2001).*

6.9. Applications in seismic method

6.9.1. The strengths and weakness of seismic reflection data

Strengths

- Good area coverage.
- Able to image major depositional units.
- Able to identify potential source, reservoir, and seal units.
- Provides a stratigraphic framework within which other data can be understood.

Weakness

- Limited vertical and lateral resolution: can't resolve "small" features.
- Stratigraphic interpretation is limited by the quality of the seismic data/imaging.
- Seismic responses are non-unique –e.g., low amplitude could be a massive sand or a thick shale.
- In new areas, we often have to 'jump' correlate from adjacent outcrops or basins.
- Post-depositional erosion and/or structuring can hamper stratigraphic correlations and paleo-depositional reconstructions.
- Typically we can't "see" hydrocarbons.

6.9.2. Hydrocarbon exploration

Title: Hydrocarbon leakage interpreted on seismic data (Løseth et al., 2009).

Purpose: To focus on the leakage processes taking place above a hydrocarbon-filled trap and how leakage is expressed on seismic data. A variety of seismic anomalies related to hydrocarbon leakage are interpreted and illustrated.

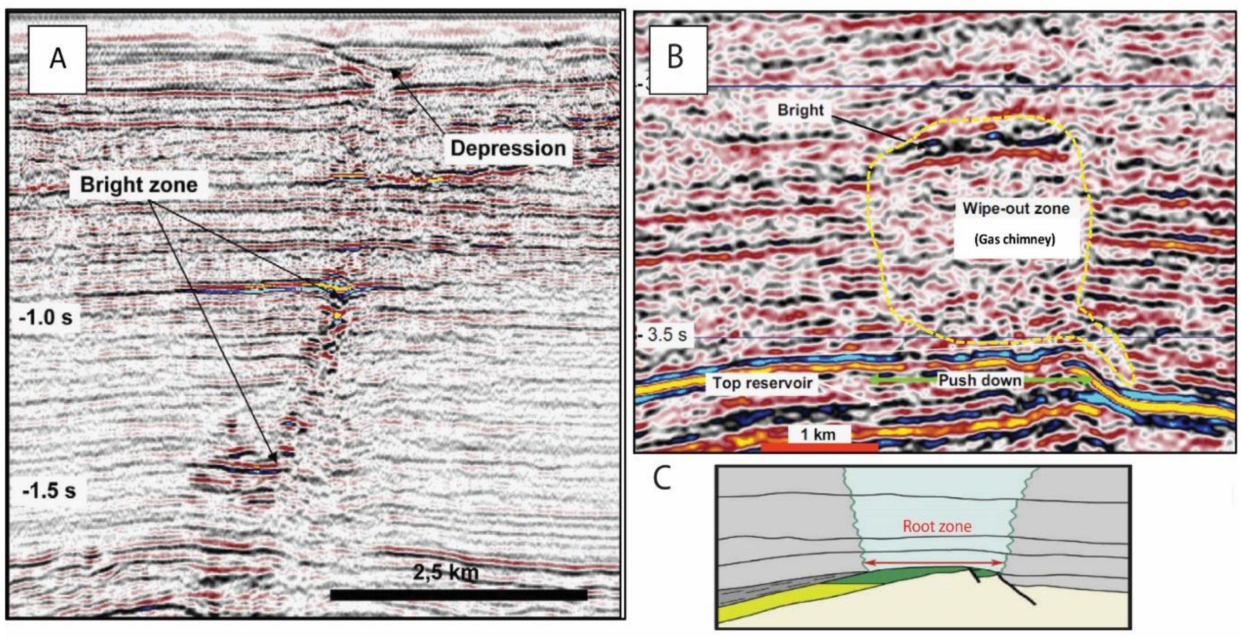

Figure 6.76: *A) Several stacked high amplitude anomalies define a vertical bright zone in the hanging-wall along a fault that can be interpreted as upward gas leakage. B) The vertical zone above a highly pressured fault block where the primary signal is wiped-out is interpreted as a gas chimney with high amplitude anomaly observed at the top of the chimney. Hydrofracturing of the cap rock is believed to have caused vertical hydrocarbon leakage.* C) Schematic illustrations of roots of leakage zones (modified from Løseth et al., 2009).

6.9.3. Hydrogeological application

Title: Seismic refraction methodology for groundwater level determination: "Water seismic index" (Grelle and Guadagno, 2009).

Purpose: To provide a procedure in order to identify groundwater levels by seismic refraction profiles in order to have a perfect overlay of the tomography 2D grids. Assuming the velocity of shear waves increases much less than the velocity of compressional waves in a saturated soil. The WSI, related to the local variations of the P and S wave velocities, has to be high at groundwater level.

Summary: this study highlighted that the WSI is susceptible to lithology and the unsaturated–saturated transition layer.

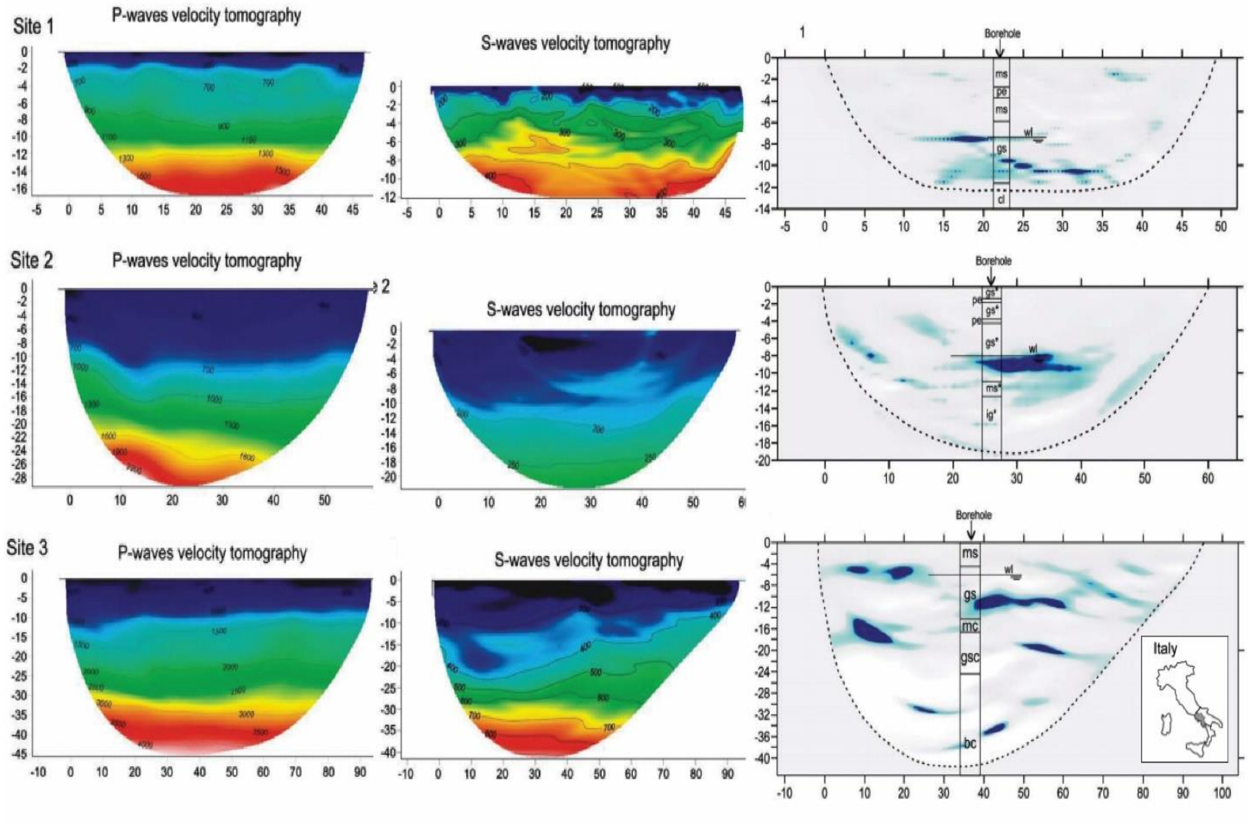

Figure 6.77: *Tomography2D models of P and SH waves velocity and WSI two-dimensional models showing correlation to ground water levels (modified from Grelle and Guadagno, 2009).*

6.9.4. Geohazard application

Title: Geophysical investigations of large landslides in the Carnic Region of southern Austria Mauritsch et al., 2000).

Purpose: To construct velocity model in order to determine the validity of one of the discussed movement models. In-situ velocity measurements were used to identify different lithologies beneath surficial talus deposits.

Summary: They were able to prove that the determination of downslope movement, estimated thickness, relief of the bedrock surface, the presence of water saturated areas, and the internal composition of sliding masses were possible by applying refraction seismic survey.

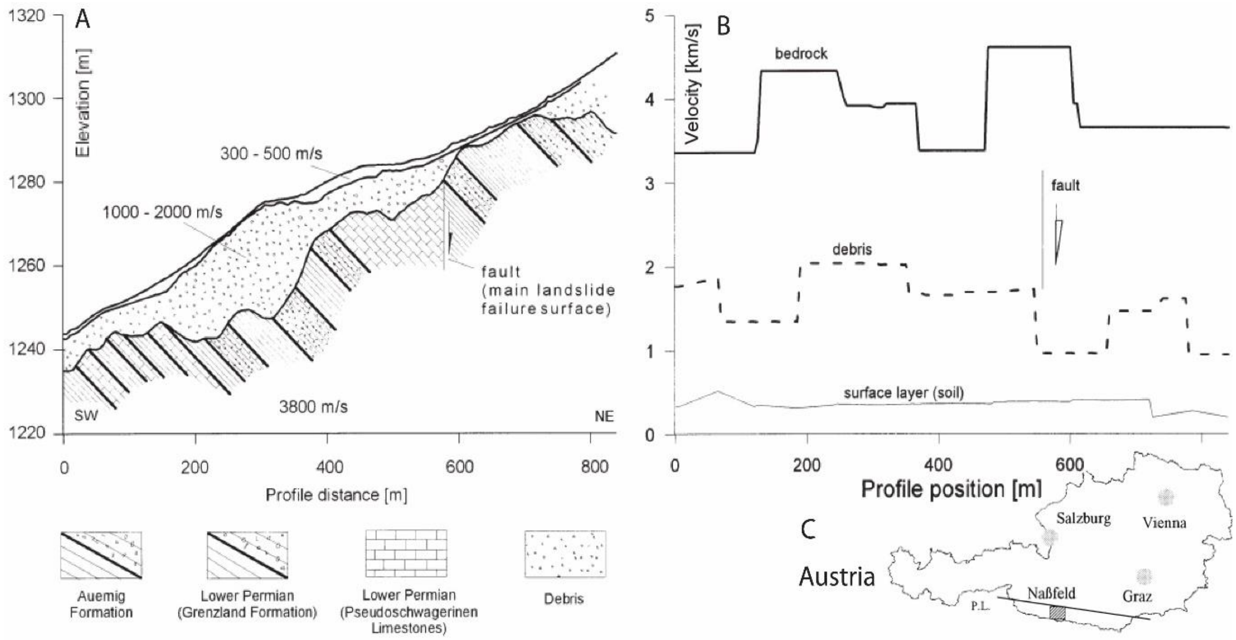

Figure 6.78: *A) Cross-section showing results of refraction seismic surveys. B) Lateral velocity variation along slope profile. C) The study area located in southern Austria (modified from Mauritsch et al., 2000).*

6.10. Exercises in seismic method chapter

1) A seismic wave is incident normally on a reflector with a reflection coefficient R of 0.01. What proportion of the incident energy is transmitted?
2) What is subsurface structure responsible for the given travel-time curves?

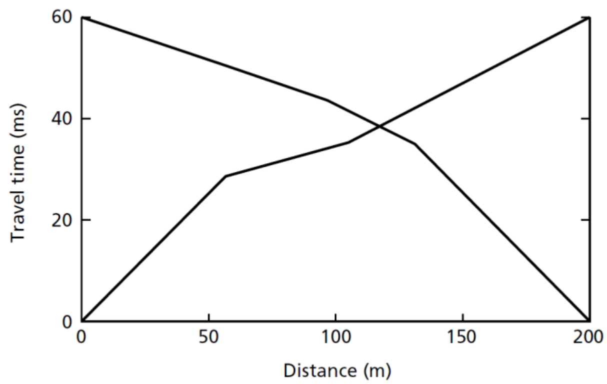

Figure 6.79: *Time–distance curves obtained in the forward and reverse directions along refraction profile (Kearey et al., 2002).*

3) What is the crossover distance for direct and critically refracted rays in the case of a horizontal interface at a depth of 200m separating a top layer of velocity 3.0kms-1 from a lower layer of velocity 5.0kms-1?
4) Construct the X^2-t^2plot and use its method to calculate velocity of the given reflector.

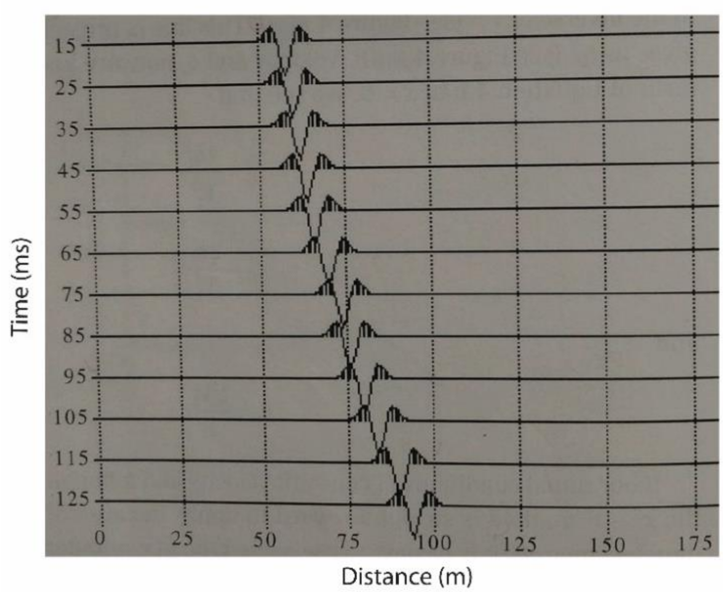

Figure 6.80: *An example of seismogram of reflection data (Burger et al., 1992).*

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เอกสารประกอบการสอนภาคปฏิบัติ

ว.ธณ. 205482 ธรณีฟิสิกส์

GEOL482 GEOPHYSICS

โดย

อาจารย์ ดร. นิติ มั่นเข็มทอง

ภาควิชาธรณีวิทยา

คณะวิทยาศาสตร์

มหาวิทยาลัยเชียงใหม่

ธันวาคม 2559

คำนำ

เอกสารประกอบการสอนฉบับนี้ ได้จัดทำขึ้นโดยการรวบรวมและดัดแปลงจากเอกสารมอบหมายงานแก่นักศึกษาประกอบการเก็บข้อมูลธรณีฟิสิกส์ภาคสนาม เพื่อใช้ประกอบการสอนภาคปฏิบัติในกระบวนวิชา ว.ธณ. 205482 ธรณีฟิสิกส์ (GEOL482: Geophysics) ในภาคส่วนของการศึกษาด้านความถ่วง แม่เหล็ก และคลื่นไหวสะเทือน ซึ่งเป็นกระบวนวิชาเลือกสำหรับนักศึกษาสาขาธรณีวิทยา ชั้นปีที่ 4 สาขาวิชาวิชาธรณีวิทยา และเป็นกระบวนวิชาบังคับสำหรับนักศึกษาสาขาฟิสิกส์ โปรแกรมร่วมธรณีวิทยา ชั้นปีที่ 4 หรือชั้นปีที่ 3 ภาควิชาฟิสิกส์และวัสดุศาสตร์ คณะวิทยาศาสตร์ มหาวิทยาลัยเชียงใหม่ ผู้จัดทำหวังเป็นอย่างยิ่งว่า เอกสารประกอบการสอนฉบับนี้ จะเป็นประโยชน์ต่อนักศึกษาเพื่อประกอบการเรียนและการทำงานวิจัยค้นคว้าอิสระต่อไป

อาจารย์ ดร. นิติ มั่นเข็มทอง

ธันวาคม 2559

ภาควิชาธรณีวิทยา **คณะวิทยาศาสตร์**

ว.ธณ. 482 (205482) ธรณีฟิสิกส์ **4(3-3-6)**

เงื่อนไขที่ต้องผ่านก่อน : ว.ธณ. 324 (205324) ธรณีวิทยาโครงสร้าง

คำอธิบายลักษณะกระบวนวิชา

ธรณีฟิสิกส์และการสำรวจธรณีฟิสิกส์ด้วยวิธีต่างๆ การสำรวจด้านไฟฟ้า: การสำรวจด้านความต้านทานไฟฟ้าจำเพาะ การสำรวจด้านการเหนี่ยวนำโพลาไรซ์ และ การสำรวจด้านศักย์ไฟฟ้าธรรมชาติ การสำรวจด้านแม่เหล็กไฟฟ้า: การสำรวจด้านม่เหล็กไฟฟ้าเหนี่ยวนำและการสำรวจด้านเรดาร์ทะลุพื้นดิน การสำรวจด้านศักย์: การสำรวจด้านความโน้มถ่วงและการสำรวจด้านแม่เหล็ก และการสำรวจด้านคลื่นไหวสะเทือน

วัตถุประสงค์กระบวนวิชา : นักศึกษาสามารถ

1. อธิบายหลักการสำรวจทางธรณีฟิสิกส์
2. วิเคราะห์ข้อมูลทางธรณีฟิสิกส์
3. ประยุกต์ใช้ความรู้ทางธรณีฟิสิกส์เพื่อการสำรวจแหล่งทรัพยากรทางธรณีวิทยาและการศึกษาด้านสิ่งแวดล้อม
4. เก็บข้อมูลภาคสนามด้วยเครื่องมือธรณีฟิสิกส์

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กระบวนวิชานี้ได้ผ่านความเห็นชอบจากที่ประชุมคณะกรรมการบริหารประจำคณะวิทยาศาสตร์ ในคราวประชุมครั้งที่ ..../............... เมื่อวันที่ .... กำหนดเปิดสอนตั้งแต่ภาคการศึกษาที่ 1 ปีการศึกษา 2554 เป็นต้นไป

(ลงนาม) ....

(รองศาสตราจารย์ ดร. สัมพันธ์ สิงหราชวราพันธ์)

คณบดีคณะวิทยาศาสตร์

วันที่........เดือน........................พ.ศ...............

Contents

Laboratory 5: Gravity Prospecting and Implication

Geophysics 205482

*Name*______________________________ *Student ID*__________

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Gravity data interpretation along the western boundary of the Chiang Mai Basin

2 weeks project

Introduction

The changes in gravity from place to place are small: gravity at the Earth's surface is about 9.81 ms^{-2}, but the local variations are a tiny fraction of this; often we are measuring differences of 10^{-6} ms^{-2}. If you can measure how g changes from place to place, you can learn something about how the density and geometry of the rocks below observations varies and understand how geology constrains gravity variations.

Objectives

1) To measure gravity data across the western boundary of the Chiang Mai Basin.
2) To understand the standard gravity reduction method.
3) To interpret subsurface geology along the western boundary of the Chiang Mai Basin based on based on gravity data.

Study area: Along main and local roads trending WNW-ESE (e.g. Huay Kaew Rd. and Suthep Roads), Muang Chiang Ma district, Chiang Mai Province (Fig. 1).

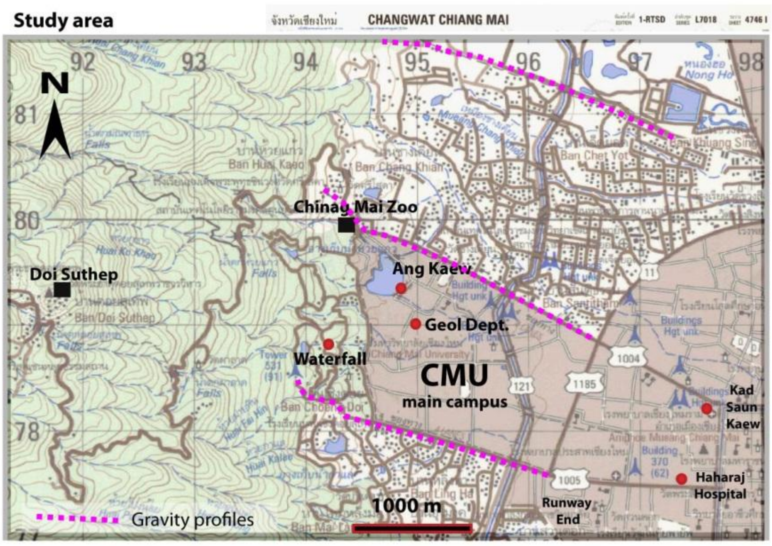

Figure 1. The study area of the gravity survey.

Acquisition and processing assignments

1) Work as a group of 4-5 students
2) Collect the refraction data. Instructors have already located a survey line for all working teams with designed configurations. Each team has two shot points; forward and reverse short with the different locations of short points.
3) Describe topographic terrain and geology along the surveyed line.
4) Use a given spreadsheet profile for gravity reduction process.
5) Use the Google Earth Pro software to location all gravity stations.
6) Use the Surfer software to grid corrected gravity and display simple gravity anomaly
7) Use the Field Geophysics Software Suite to construct a simple gravity model (Fig. 2)
8)

Figure 2. Field Geophysics Software Suite for data modeling.

Interpretation assignments

1) Review previous literature such as geologic and hazard maps and publications, groundwater well information.

2) Integrate all surface geological data constraints to gravity data interpretation to finalize your geological interpretation with structural controls.

Implication assignments

1) Discuss relation between your geological interpretations, and local tectonic setting of the Chiang Mai Basin and Doi Suthep-Inthanon Range.

2) Locate the possible small-sized dam in the locality and design the dam size (e.g. depth of foundation and dam height). Please consider the earthquake and flood hazard possibilities.

Presentation assignments

Prepare 10-15 slides for 10-minute presentations of your team works. No limit of presentation style and creativity.

Suggested readings and GS-NAS map source

Mankhemthong, N., Takaew, P., Morley, K.C., and Rhodes, B., Structure of Baan pong basin, hang dong district, Chiang Mai province, the joint gravity interpretation and geologic field observation, based on joint gravity interpretation and geologic field observations. International Conference and Exhibition, Barcelona, Spain, 3-6 April 2016: pp. 119-119. https://doi.org/10.1190/ice2016-6336262.1

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Laboratory 6: Magnetic Prospecting and Implication

Geophysics 205482

*Name*_____________________________ *Student ID*__________

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Magnetic data interpretation of shallow or surface magnetic sources

2 weeks project

Introduction

A magnetic survey method used as a baseline survey method and the depth of the basement. We also applied the survey method in engineering and archeology aspects to investigate buried/exposed foundations with metals embedded or ancient monuments beneath the subsurface. See Fig 1 for an example target.

Figure 1. An example of the magnetic survey target.

Objectives

1) To measure magnetic data in the city area.
2) To understand magnetic reduction processes and the best practice to minimize unneeded noise signals.
3) To interpret the magnetic anomaly due to the known size magnetic sources.

Study area: Rugby football field and surrounding green areas, the main CMU Campus (Fig. 2).

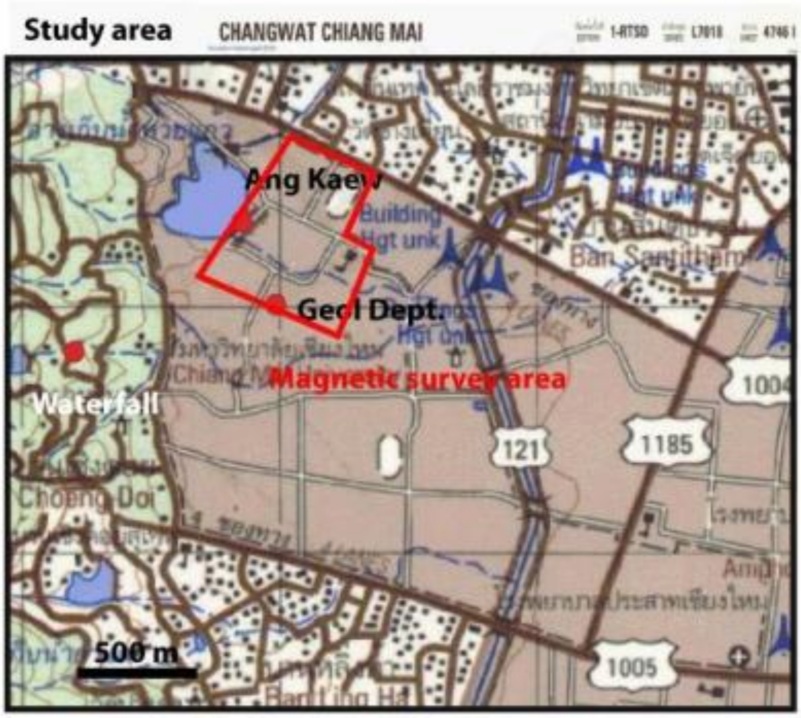

Figure 2. The study area of the magnetic survey.

Acquisition and processing assignments

1) Work as a group of 4-5 students.
2) Select one construction that could be a major magnetic anomaly source, which is a noise problem in the geological propose survey.
3) Collect 30 stations as a random grid survey with a suitable station spacing depending on the size of the anomaly source using the Magnetometer and Topcon G-1000 Differential GPS receiver.
4) Map topographic terrain and man-made constructions in a surveyed area.
5) Use a given spreadsheet profile for magnetic reduction process.
6) Use the Google Earth Pro software to location all magnetic stations.
7) Use the Surfer software to grid corrected magnetic and display corrected magnetic anomaly
8) Use the Field Geophysics Software Suite to construct the 2-D magnetic model.

Interpretation assignments

1) Review available information such as underground pipeline, drainage, and power line maps and other buried foundations.
2) Integrate all surface data/noise constraints to magnetic data interpretation to give a final interpretation with noise signal explanation.

Implication assignment

Discuss the interpretation to engineering factors you the best practice of magnetic surface in the futures.

Presentation assignments

Prepare 10-15 slides for 10-minute presentations of your team works. No limit of presentation style and creativity.

Suggested readings and GS-NAS map sources

Mankhemthong, N., and Linchongsuwan, N., 2015, Geophysical survey for subsurface mapping of ancient kiln locations, On-Tai Subdistrict, San Kamphaeng District, Chiang Mai Province. In Proceedings of 5th GEOINDO 2015 International Conference, 24-25 November 201

นิติ มั่นเข็มทอง, 2561, เอกสารประกอบการสอน ว.ธณ. 205482 ธรณีฟิสิกส์ (GEOL482 GEOPHYSICS), ภาควิชาธรณีวิทยา คณะวิทยาศาสตร์ มหาวิทยาลัยเชียงใหม่

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Laboratory 7: Seismic Refraction Prospecting and Implication

Geophysics 205482

*Name*______________________________ *Student ID*__________

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Seismic Refraction data interpretation of western boundary of the Chiang Mai Basin

2 weeks project

Introduction

A refraction survey uses refracted (or head) waves to deduce velocities of the layered-earth model. So-called first arrival information is used for the analysis. In theory, direct wave and head wave will arrive and/or met each other at some time then head wave will arrive first at receivers because head wave through the compacted rock/soil layers, but direct wave travel through less dense rock/soil and higher porosity within air .as well. This is a simple principal to find geometry of the top layered material.

Objective

1) To measure seismic refraction data across the western boundary of the Chiang Mai Basin
2) To understand the seismic refraction method for the 2-layer model problem.
3) To calculate the thickness of sediment covers close to the western margin of the Chiang Mai Basin
4) To interpret subsurface geology along the western boundary of the Chiang Mai Basin based on seismic data.

Study area: A trail heading to a waterfall located in the backyard of the Faculty of Mass Communication, the main CMU Campus (Fig. 1).

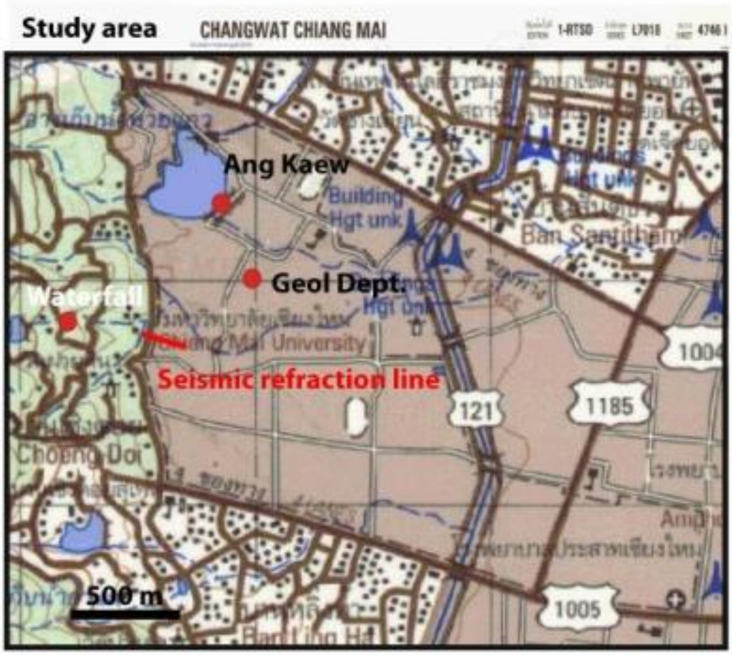

Figure 1. The study area of seismic refraction survey.

Acquisition and processing assignments

1) Work as a group of 4-5 students.
2) Collect refraction data. Note that instructors have already located a survey line for all working teams with designed configurations of 72 geophone channels and a spacing shot of 2 meters.
3) Each team is assigned to make two shot points; forward and reverse short with the different locations.
4) Describe topographic terrain and geology along the surveyed line.
5) Use a given spreadsheet profile for seismic refraction processing method and calculate the thickness of sediment covers.
6) Use the Google Earth Pro software to location all gravity stations.
7) Use the Field Geophysics Software Suite to construct the 2-layer model.

Interpretation assignments

1) Review previous literature such as geologic and hazard maps and publications, shallow geophysical information.
2) Integrate all surface geological data constraints to seismic data interpretation to finalize your geological interpretation with structural controls.

Implication assignments

1) Compare the interpretation to the gravity model we have done a few weeks ago. Which method does give the better data resolution?

2) Discuss the final interpretation of the local tectonic setting of the Chiang Mai Basin and Doi Suthep-Inthanon Range. Are they support each other?

Presentation assignments

Prepare 10-15 slides for 10-minute presentations of your team works. No limit of presentation style and creativity.

Suggested readings and GS-NAS map sources

Morley, C.K., 2009. Geometry and evolution of low-angle normal faults (LANF) within a Cenozoic high-angle rift system, Thailand: Implications for sedimentology and the mechanisms of LANF development. Tectonics 28, doi:10.1029/2007TC002202.

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Laboratory 8: Seismic reflection interpretation implication

Geophysics 205482

*Name*_____________________________ *Student ID*__________

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Seismic reflection data interpretation and petroleum hunting

2 weeks project

Introduction: find the oil - an interpretation

This is an exercise to locate any structural features on the seismic data that could act as trap for hydrocarbons. Working geophysicists called interpreters do this on seismic data all the time.

Reading assignments

A seismic profile runs from South to North with a borehole at CMP 740, labelled well tie. This links the horizons on the seismic profile to the real lithology. Using the results of the well tie, we can identify the following horizons and rock units. **Read an attached handout for details.**

Working assignments

1) Locate the well tie position on the top of the seismic section and draw a pencil line directly under that point right to the bottom of the profile.
2) Mark the TWT of the horizons in the well tie using a different colored pencil for each one. The very black events are peaks, the greyer ones troughs and zero crossings are white.
3) Start with the Purbeck Sandstone and use a pencil to follow the grey colored trough across the page (it may be easier to use a lead pencil to mark your progress). The horizon may not be continuous all the way across the profile. Try to think about possible structures to explain any jumps or breaks in the horizon, (faults for example). Remember if you think you see a break that could be a fault, you should see a similar break in the horizons above and below. When you are confident you can see where the horizon goes, color it with your chosen color.
4) Do the same for the Kimmeridge Clay, but this time follow the black colored peak? Finally, interpret the Corallian Limestone, the zero crossing is the changeover point from a peak to a trough (positive to negative).
5) Next, mark any faults on the profile with a lead pencil. Try to describe it, e.g. is it a normal or a reverse fault, which is the downthrown side (north or south) and how big is the throw (in ms)?

6) Finally when you have traced all the horizons and marked any faults, look at the whole picture and answer the following questions.

   i. Explain the geologic structure, stratigraphy, and related tectonics.

   ii. Describe all seismic profile parameters.

   iii. Show locations of all petroleum systems on the interpreted profile.

      - Reservoir rock
      - Caprock
      - Migration path
      - Source rocks
      - Fault trap

   iv. Where is/are an accumulation(s) of oil or gas? Draw the highest potential wells!

References and data sources

http://www.myoilandgascareer.com/uploaded-files/curriculum-resources-england/find-the-oil-an-interpretation-exercise.pdf

นิติ มั่นเข็มทอง, 2561, เอกสารประกอบการสอน ว.ธณ. 205482 ธรณีฟิสิกส์ (GEOL482 GEOPHYSICS), ภาควิชาธรณีวิทยา คณะวิทยาศาสตร์ มหาวิทยาลัยเชียงใหม่

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