جامعة جنوب الوادي　　　　قسم الجيولجيا

مقرر
الجيوفيزياء التطبيقية

الفرقة الثالثة شعبة جيوفيزياء

أستاذ المقرر

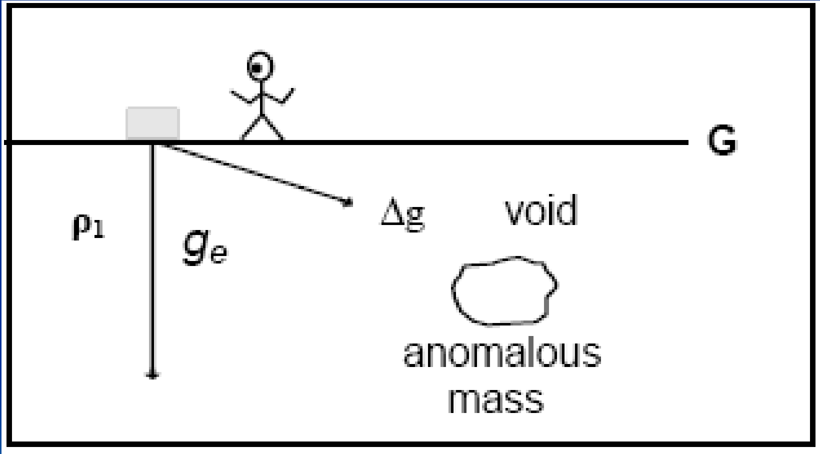
أ.د\ عبدالهادي محمد العقربي

قسم الجيولوجيا - كلية العلوم

العام الجامعي

2023-2022

Gravity Method



General Overview

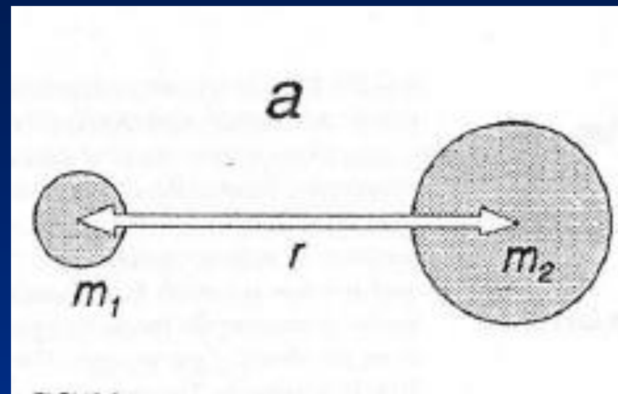
- Gravimeters are designed to measure the vertical component of the vector sum of the earth's background gravitational field (g_E) and superposed secondary fields (Δg) created by localized, causative bodies of anomalous density.

- Generally, the secondary field is isolated and interpreted with a view to elucidating the nature of the secondary causative body

Applications of Gravity Method

Hydrocarbon exploration
Regional geological studies
Isostatic compensation determination
Exploration for, and mass estimation of, mineral deposits
Detection of sub-surface cavities (micro-gravity)
Location of buried rock-valleys
Determination of glacier thickness
Tidal oscillations
Archaeogeophysics (micro-gravity); e.g. location of tombs
Shape of the earth (geodesy)
Military (especially for missile trajectories)
Monitoring volcanoes

THEORY – EARTH'S GRAVITY FIELD



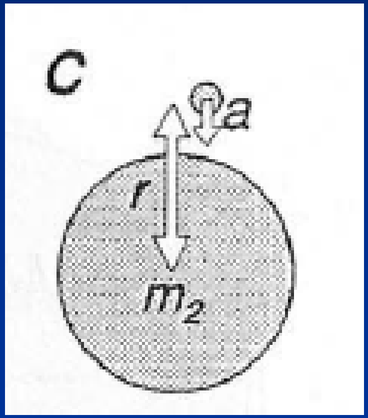
Gravity is the attraction on one body due to the mass of another body. The force of one body acting on another is given by *Newton's Law of Gravitation* (Fig. 8.2a):

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

where:

F = force of attraction between the two objects (N)
G = Universal Gravitational Constant (6.67×10^{-11} Nm^2/kg^2)
m_1, m_2 = mass of the two objects (kg)
r = distance between the centers of mass of the objects (m):

THEORY – EARTH'S GRAVITY FIELD



The force (F) exerted on the object with mass m_1 by the body with mass m_2, is given by *Newton's Second Law of Motion* (Fig. 8.2b):

$$F = m_1 a$$

where:

a = acceleration of object of mass m_1 due to the gravitational attraction of the object with mass m_2 (m/s^2).

Solving for the acceleration, then combining the two equations (Fig. 8.2c):

$$a = \frac{F}{m_1} = \frac{1}{m_1}\frac{Gm_1m_2}{r^2}$$

$$a = \frac{Gm_2}{r^2}$$

For Earth's gravity field (Fig. 8.3a), let:

$a = g$ = gravitational acceleration observed on or above Earth's surface;
$m_2 = M$ = mass of the Earth;
$r = R$ = distance from the observation point to Earth's center of mass;

so that:

$$\boxed{g = \frac{GM}{R^2}}$$

THEORY – EARTH'S GRAVITY FIELD

$$\text{Force} = \text{gravitational constant} \times \frac{\text{mass of Earth } (M) \times \text{mass } (m)}{(\text{distance between masses})^2}$$

$$F = \frac{G \times M \times m}{R^2} \qquad \text{(equation (1))}$$

where the gravitational constant $(G) = 6.67 \times 10^{-11}\ \mathrm{N\,m^2\,kg^{-2}}$

$$\text{Force} = \text{mass } (m) \times \text{acceleration } (g)$$

$$F = m \times g \qquad \text{(equation (2))}$$

$$F = \frac{G \times M \times m}{R^2} = m \times g; \quad \text{thus } g = \frac{G \times M}{R^2} \quad \text{(equation (3))}.$$

UNITS OF "g"

SI unit for g: m/s^2 – though you will rarely see this!

$1\ cm/s^2$ = 1 Gal (for Galileo) = $0.01\ m/s^2$

milliGal or mGal = 10^{-3} Gal – typical unit for field studies

Our text book uses the "gravity unit" (g.u.)

1 g.u. = 0.1 mGal

Normal value of g at the surface of the Earth:

$g_E = 9.8\ m/s^2 = 980\ cm/s^2$ = 980 Gal = 980,000 mGal = 9800 g.u.

VARIATIONS OF "g"

The shape of the Earth is a consequence of the balance between gravitational and centrifugal accelerations causing a slight flattening to form an oblate spheroid. Mathematically it is convenient to refer to the Earth's shape as being an *ellipse of rotation* (Figure 2.1).

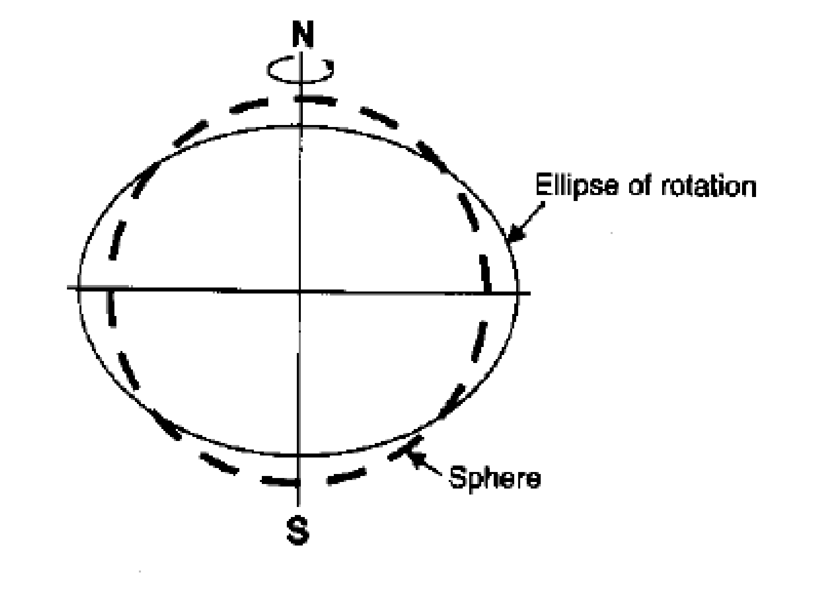


Figure 2.1 Exaggerated difference between a sphere and an ellipse of rotation (spheroid)

VARIATIONS OF "g"

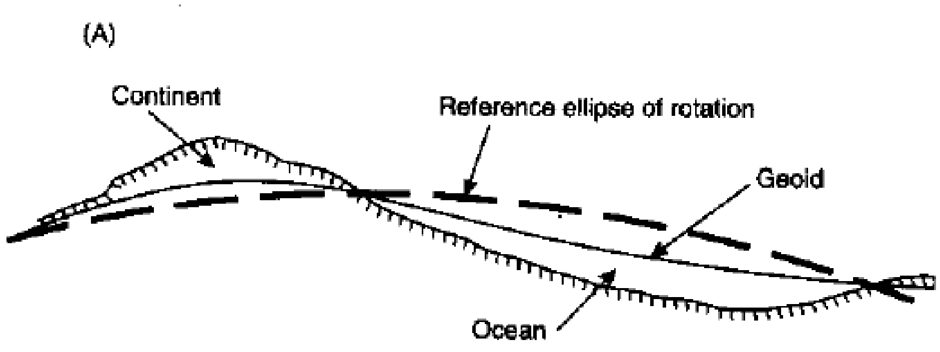
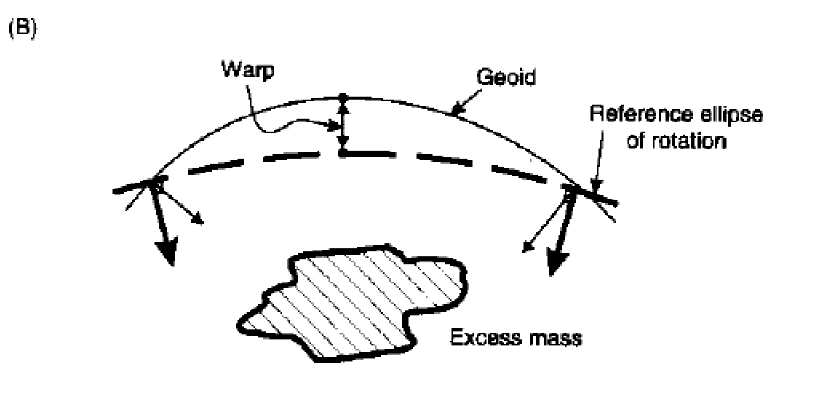


Figure 2.2 Warping of the geoid: (A) continental-scale effects, and (B) localised effects due to a sub-surface excess mass

to gravity everywhere. The geoid represents a surface over which the gravitational field has equal value and is called an *equipotential surface*. The irregular distribution of mass, especially near the Earth's surface, warps the geoid so that it is not identical to the ellipse of rotation (Figure 2.2). Long-wavelength anomalies, which can be

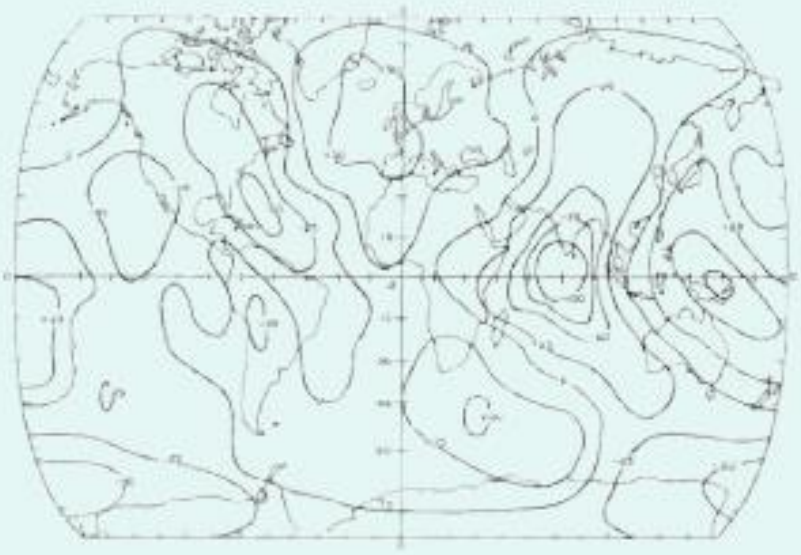


The geoid height (= "geoid anomaly") varies because gravity varies from place to place as a result of variable mass & density distribution

The Geoid

The spheroid and geoid

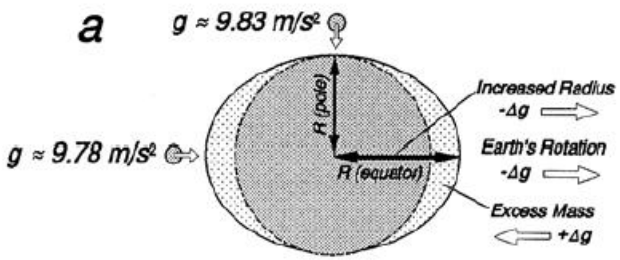
Global-scale geoid anomalies plotted as map



16

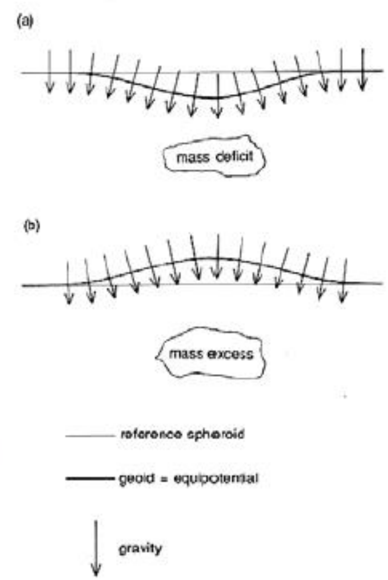
VARIATIONS OF "g"

Variations in g



Large scale variations: global or regions

Figure 5.8. (a) A trough in the geoid, or negative geoid height anomaly, occurs over a region of mass deficit (such as a depression in the seabed). A negative free-air gravity anomaly also occurs over such a mass deficit. (b) A bulge in the geoid, or positive geoid height anomaly, occurs over regions of excess mass (such an elevated region of the seabed). A positive free-air gravity anomaly also occurs over such a mass excess.



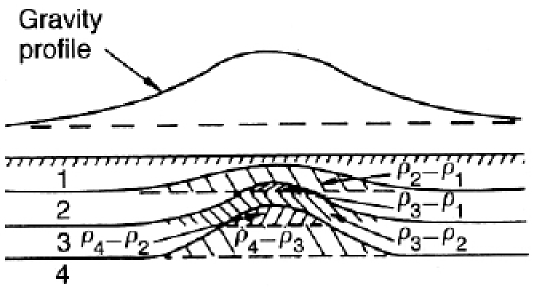
Smaller scale variations: local

This is what we want to make use of

Mass = Density x Volume

Rock Density

Lateral variations in rock density result in gravity anomalies that can be measured at the surface



Gravity surveying is sensitive to rock density. In the following typical values for various types or rocks and minerals are given.

However, laboratory measurements of rock/sediment/mineral density are in error relative to the true *in situ* densities. Rock samples are typically altered either by drying (loss of moisture) or by release of the *in situ* stress.

Sedimentary rocks:

There are at least 7 factors affecting density of sedimentary material. The values given in the brackets are the average change in density possible due to this physical factor:

Composition (35%), cementation (10%), age (25%), depth of burial (25%), tectonic stresses (10%), porosity and pore-fluid type (10%).

Densities of Common Rocks

- Rocks have a range of densities* (*your book gives more)
  - But in general, rock density does not widely vary
  - What does this say about typical gravity anomaly sizes?

Type	Rock	Density
Unconsolidated	Sand	1400-1650 kg/m^3
Sedimentary	Salt	2100-2600
	Limestone	2000-2700
	Shale	2000-2700
Igneous	Granite	2500-2800
	Basalt	2700-3000
Metamorphic	Quartzite	2600-2700
	Gneiss	2600-3000
Ore	Galena	7400-7600
	Pyrite	4900-5200
	Magnetite	4900-5300

Densities of Common Rocks

- Which has higher / lower density?
  - Surface / deep rocks
  - Weathered / unweathered rocks
  - Rocks in the phreatic / vadose zone
  - Fractured / unfractured rocks
- Which rock types make good targets for gravity surveying?
  - Which don't
  - Why

Type	Rock	Density
Unconsolidated	Sand	1400-1650 kg/m^3
Sedimentary	Salt	2100-2600
	Limestone	2000-2700
	Shale	2000-2700
Igneous	Granite	2500-2800
	Basalt	2700-3000
Metamorphic	Quartzite	2600-2700
	Gneiss	2600-3000
Ore	Galena	7400-7600
	Pyrite	4900-5200
	Magnetite	4900-5300

Rock Density

Igneous and metamorphic rocks:

Igneous and metamorphic rocks tend to be denser than sedimentary rocks, but there is considerable overlap.

Density typically increases with silica content, so basic igneous rocks are denser than acid ones. Similarly, plutonic rocks tend to be denser than the volcanic equivalent.

Crystal size	Silica content		
	Acid	Intermediate	Basic
Fine-grained (volcanic)	Rhyolite 2.35 – 2.70 g/cm^3	Andesite 2.40 – 2.80 g/cm^3	Basalt 2.70 – 3.30 g/cm^3
Coarse-grained (plutonic)	Granite 2.50 – 2.81 g/cm^3	Syenite 2.60 – 2.95 g/cm^3	Gabbro 2.70 – 3.50 g/cm^3

Density of Minerals

Material type	Density range (Mg/m^3)	Approximate average density (Mg/m^3)
Metallic minerals		
Oxides, carbonate		
Manganite	4.2–4.4	4.32
Chromite	4.2–4.6	4.36
Magnetite	4.9–5.2	5.12
Haematite	4.9–5.3	5.18
Cuprite	5.7–6.15	5.92
Cassiterite	6.8–7.1	6.92
Wolframite	7.1–7.5	7.32
Uraninite	8.0–9.97	9.17
Copper	n.d.	8.7
Silver	n.d.	10.5
Gold	15.6–19.4	17.0
Sulphides		
Malachite	3.9–4.03	4.0
Stannite	4.3–4.52	4.4
Pyrrhotite	4.5–4.8	4.65
Molybdenite	4.4–4.8	4.7
Pyrite	4.9–5.2	5.0
Cobaltite	5.8–6.3	6.1
Galena	7.4–7.6	7.5
Cinnabar	8.0–8.2	8.1
Non-metallic minerals		
Gypsum	2.2–2.6	2.35
Bauxite	2.3–2.55	2.45
Kaolinite	2.2–2.63	2.53
Baryte	4.3–4.7	4.47
Miscellaneous materials		
Snow	0.05–0.88	n.d.
Petroleum	0.6–0.9	n.d.
Lignite	1.1–1.25	1.19
Anthracite	1.34–1.8	1.50

Rock Density

Factors influencing rock density

Unconsolidated sediments – composition, porosity, saturation

Sedimentary rocks – composition, age and depth of burial (compaction), cementation, porosity, pore fluid

Igneous rocks – composition (esp. silica content), crystal size, fracturing (i.e. porosity)

Metamorphic rocks – composition (esp. silica content), metamorphic grade, fracturing (i.e. porosity)

Porosity and **pore fluid content** are probably the most important factors affecting density in the shallow sub-surface

Measuring Gravity

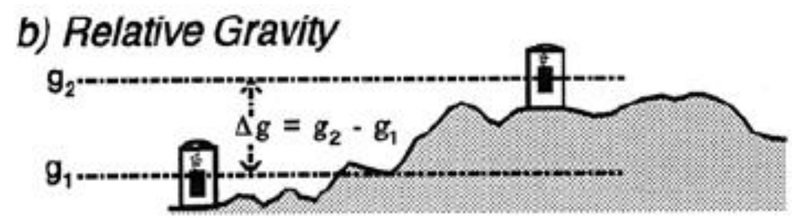
Gravity Measurements

absolute gravimeters

- pendulums
- falling masses

relative gravimeters

- stable gravimeters
- unstable gravimeters



A gravimeter is an instrument used in gravimetry for measuring the local gravitational field of the Earth. A gravimeter is a type of accelerometer, specialized for measuring the constant downward acceleration of gravity, which varies by about 0.5% over the surface of the Earth.

Gravity Measurements

Absolute Gravimeters

a- Pendulum

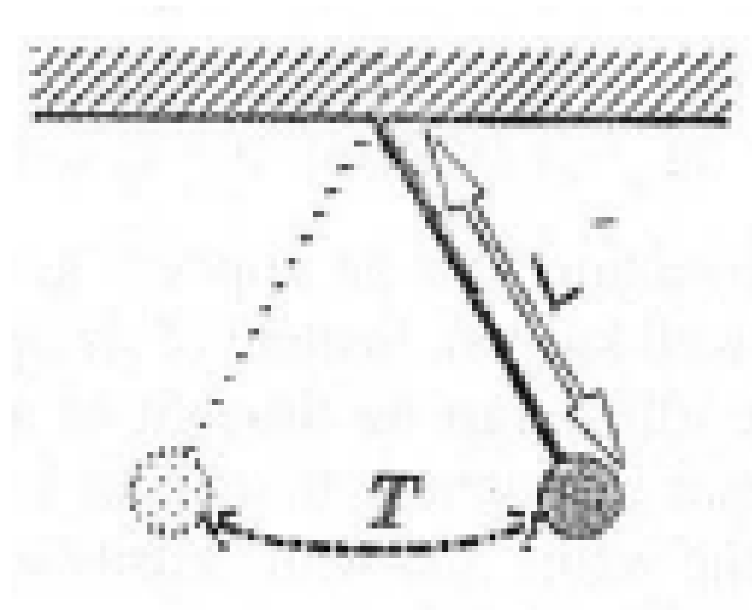
Absolute measurements: pendulums

First done by Pierre Bouguer in 1749

$$g = \frac{4\pi^2 L}{T^2}$$

L = pendulum length

$$\frac{T_2^2}{T_1^1} = \frac{g_1}{g_2}$$



Gravity = constant × pendulum length/period^2 $\quad g = 4\pi^2 L/T^2$

$$\frac{(\text{Period}_1)^2}{(\text{Period}_2)^2} = \frac{\text{gravity}_2}{\text{gravity}_1} \qquad \frac{T_2^2}{T_1^2} = \frac{g_2}{g_1}$$

Gravity Measurements

Absolute Gravimeters

b- Falling body

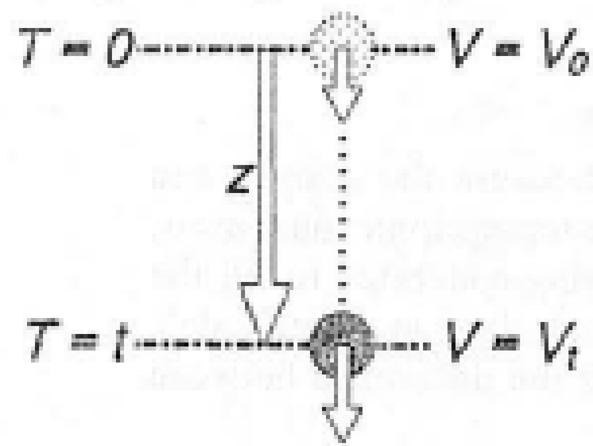
$$z = v_0 t + \tfrac{1}{2} g t^2$$

where:

z = distance the object falls
t = time to fall the distance z
v_0 = initial velocity of the object
g = absolute gravity.

The absolute gravity is thus:

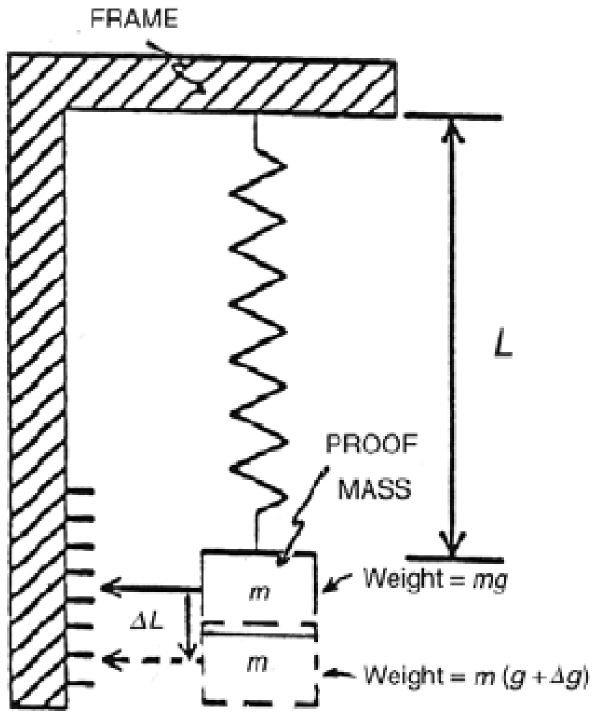
$$\boxed{g = 2(z - v_0 t)/t^2}$$



Gravity Measurements

Relative Gravimeters

a- Stable Gravimeter



change in g ➔ change in spring length

Hooke's Law $\Delta F = -k\ \Delta L$

and $\Delta g = -k\ \Delta L/m$

if $\Delta g/g = 10^{-6}$

then $\Delta L/L = 10^{-6}$

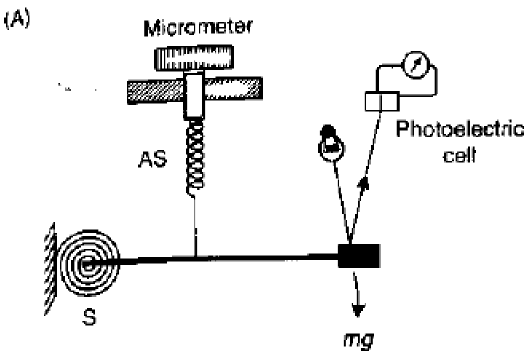
This requires high optical, mechanical or electronic magnification

Work on the principle of a force balancing the force of gravity. Example: the Gulf gravimeter

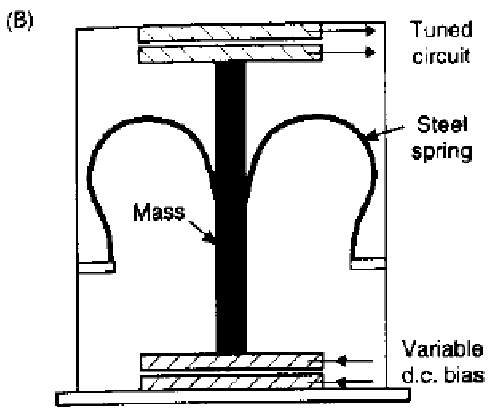
Gravity Measurements

Relative Gravimeters

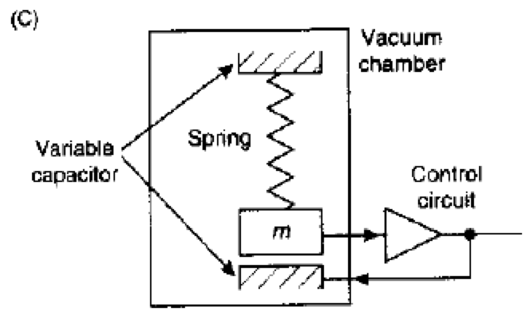
a- Stable Gravimeter



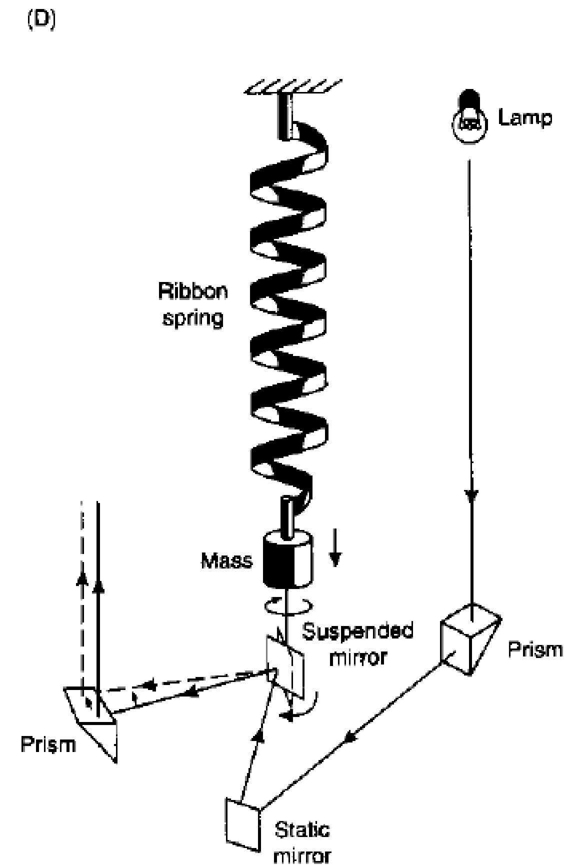
Askania



Boliden



Scintrex (CG-3)



Gulf

Gravity Measurements

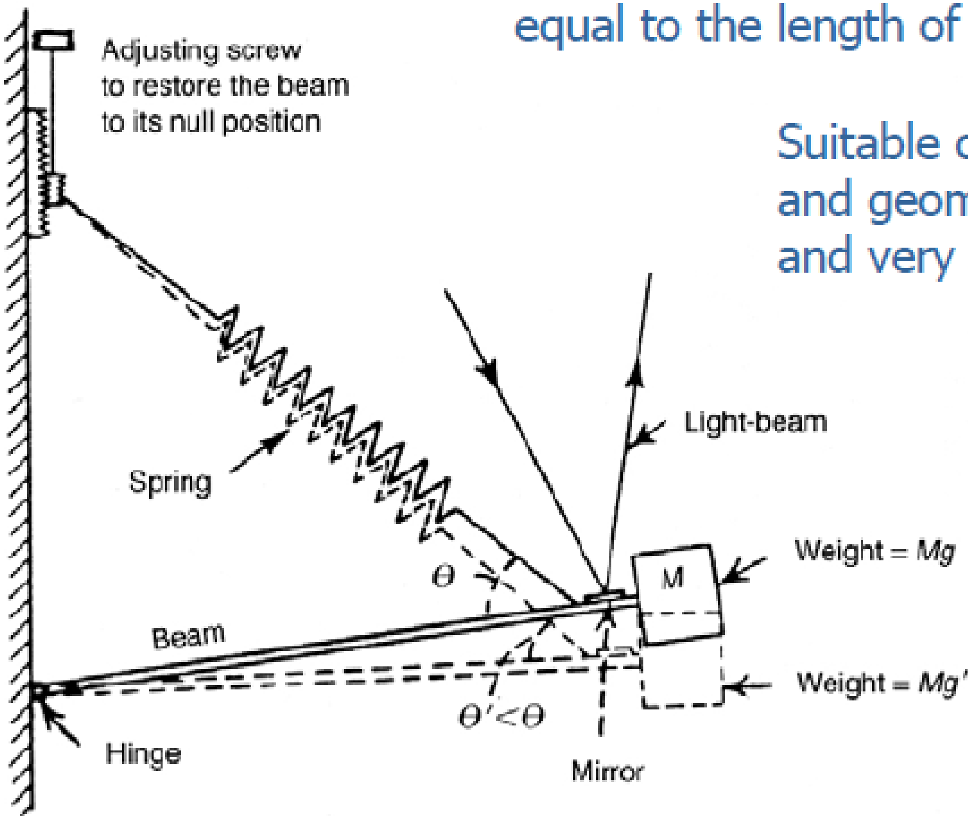
Relative Gravimeters

b- Unstable Gravimeter

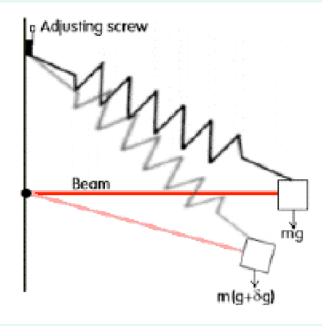
Applies and additional negative restoring force to amplify changes in g

Uses a zero length spring: the restoring force is equal to the length of the spring

Suitable choice of mass, spring constant and geometry makes the system unstable and very sensitive to changes in g



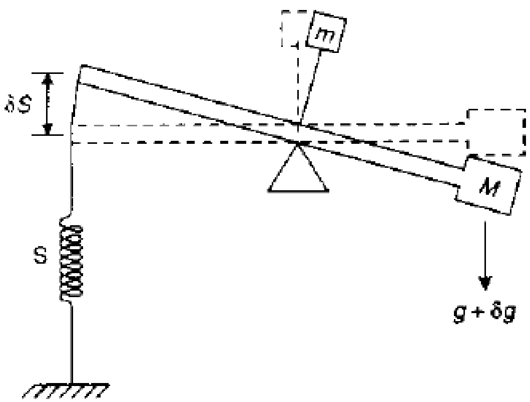
- Cunning mechanical devices
- increases in g cause extension of spring
- extension magnified by mechanical geometry
- Examples: the Wordon and the LaCoste-Romberg gravimeters



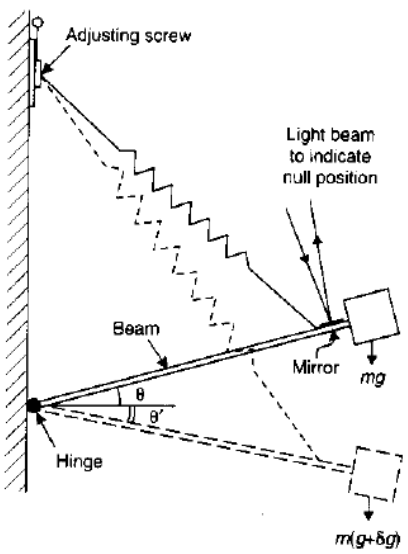
Gravity Measurements

Relative Gravimeters

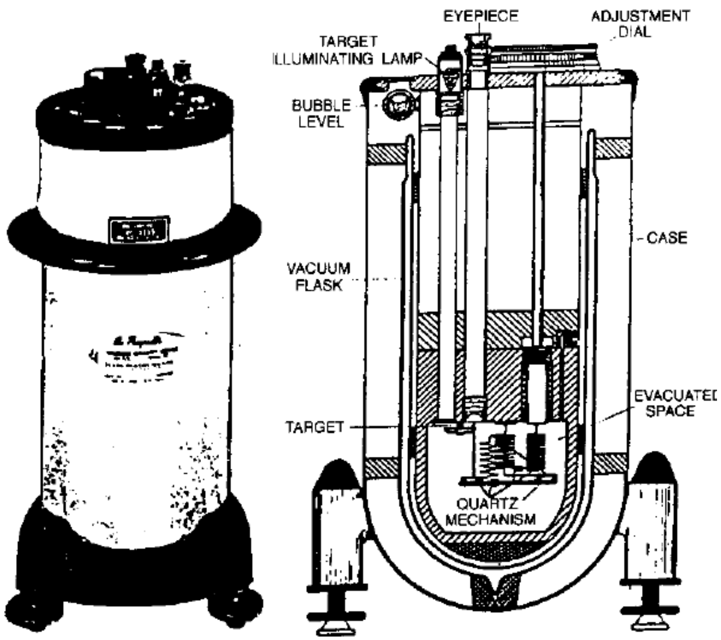
b- Unstable Gravimeter



Thyssen



LaCoste Romberg



Worden

Gravity surveys on land

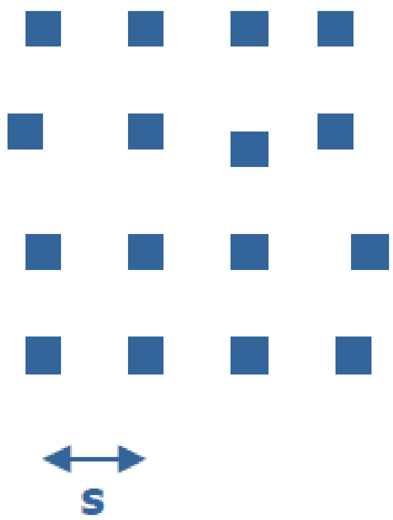
1. Include station where absolute *g* is known?
2. Station spacing must fit anomaly scale.
3. Heights of all stations must be known or measured to ~ 10 cm.
4. Latitudes must be known to 50 m.
5. Topography affects the measurements - locate stations where little topography.
6. Access - keep stations to existing roads or waterways if there are no roads.
7. Design survey well. Computer processing cannot compensate for poor experiment design.

Gravity surveys on land

Method

1. Measure base station,
2. Measure more stations,
3. Remeasure base station approximately every two hours.
4. Record in log book:
   - time of measurement
   - reading
   - terrain

Survey design



Survey design considerations

- **Uniform grid – for easier interpretation**
- **Station spacing: s < h**

  h is the depth of the body of interest
- **Avoid steep tomographic gradients**
- **Absolute and relative station locations are needed** …how accurate?

Typical station spacing

Regional geologic studies: km to 10s of km

Local structure/Engineering/Environmental: 10s to 100s m

Near surface e.g. archeology: few meters

Gravity Corrections/ Reductions

Gravity Corrections

- Take the simple example of a gravity measurement made at two different elevations
  - Because the distance to the center of Earth is farther for higher elevations, the gravity must be less.
  - There are actually several types of corrections that must be applied to raw gravity data in order for anomalies to be identified.

1. Drift
2. Latitude correction
3. Eötvös correction
4. Topographic correction
   a. Free-air correction
   b. Bouguer correction
   c. Terrain correction
5. *Regional / Residual anomaly *(not always necessary)*

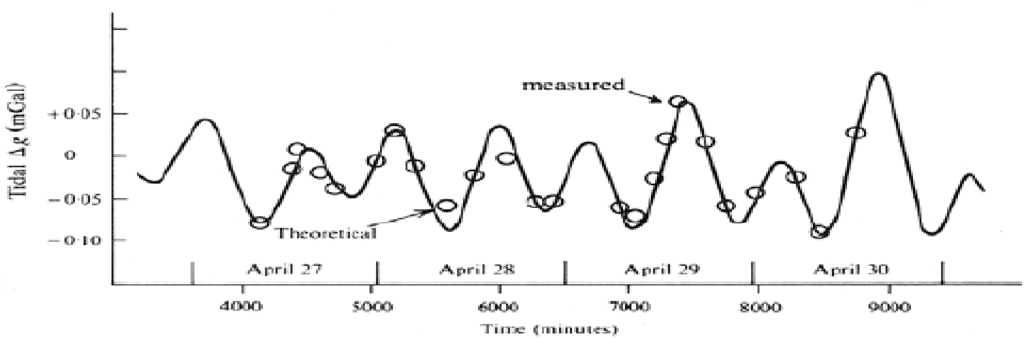
Drift

- **Drift** is a change in readings that would occur even if the device was not moved throughout the day.
  - The spring inside the gravimeter may slowly creep or stretch
  - Diurnal variations in tides
- Drift is corrected by periodically returning to a base station to get the temporal variation.
  - The drift is then subtracted from the rest of the data.

The reading of a gravimeters at a point changes with time!

Causes

- **Instrument drift:** due to environmental changes (P,T) and spring creep
- **Earth tides:** relative rotations of the earth, moon and sun



Drift Example

- Because most land-based surveys can only collect one data point at a time, temporal drift variations must be corrected
  - Base station readings are used to determine the temporal variations
  - The base station readings are normalized and then subtracted from the data to correct for drift.

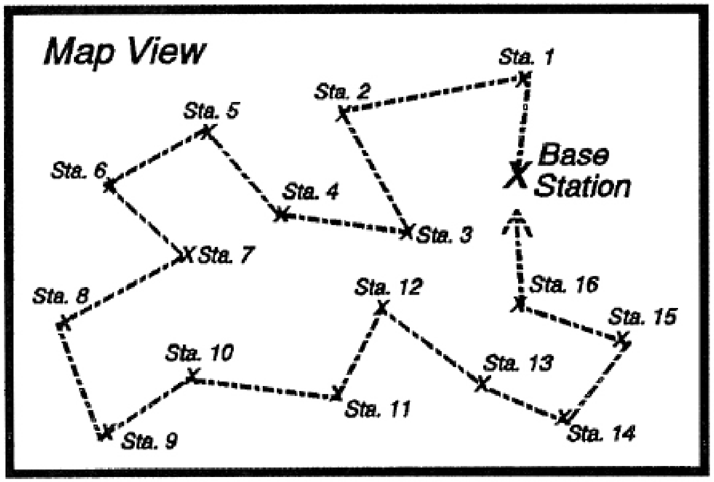
Correcting for drift

1. **Return to base station periodically**
2. **Assume drift is linear**
3. **Correct measurements in loop**

How often?

Depends on requires accuracy

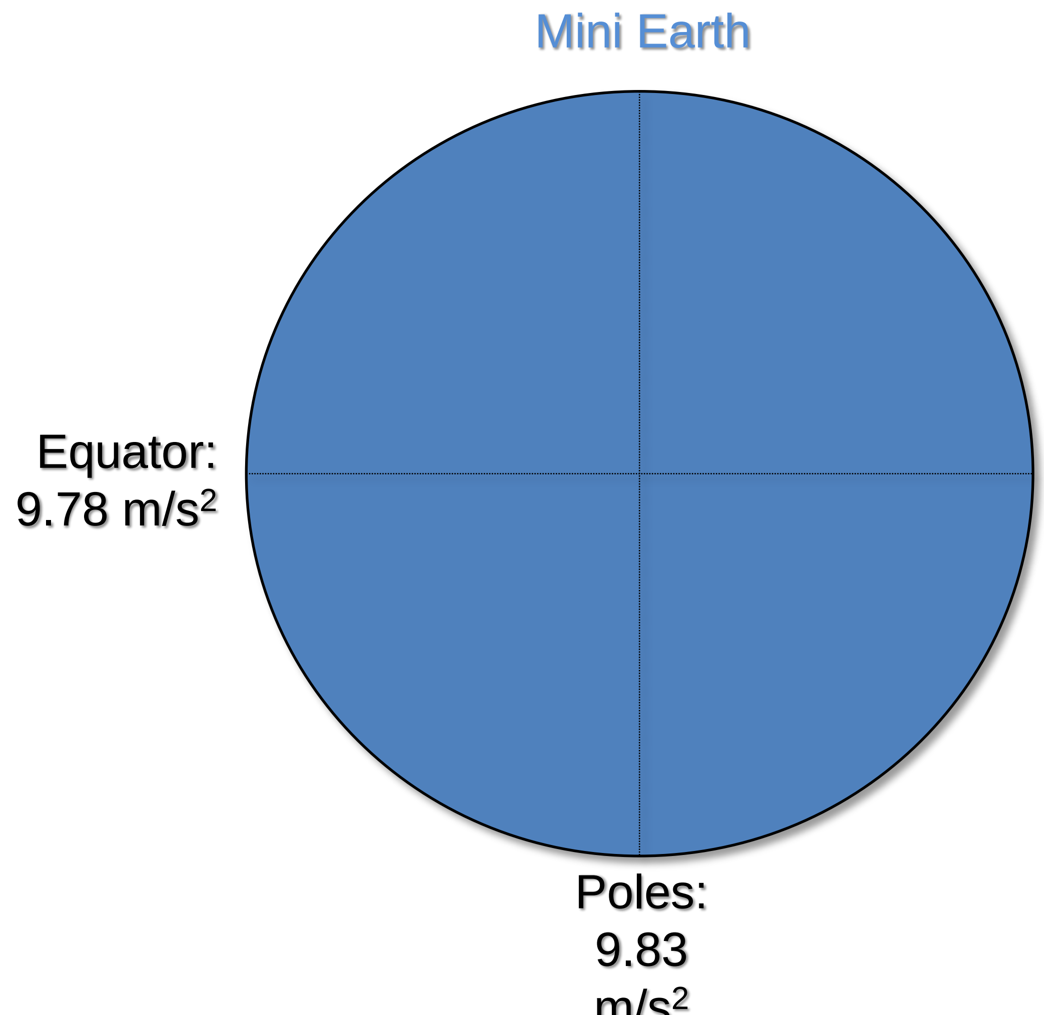
- max tidal rate: 0.05 mGal/hr
- instrument drift usually less



Latitudinal Gravity Variations

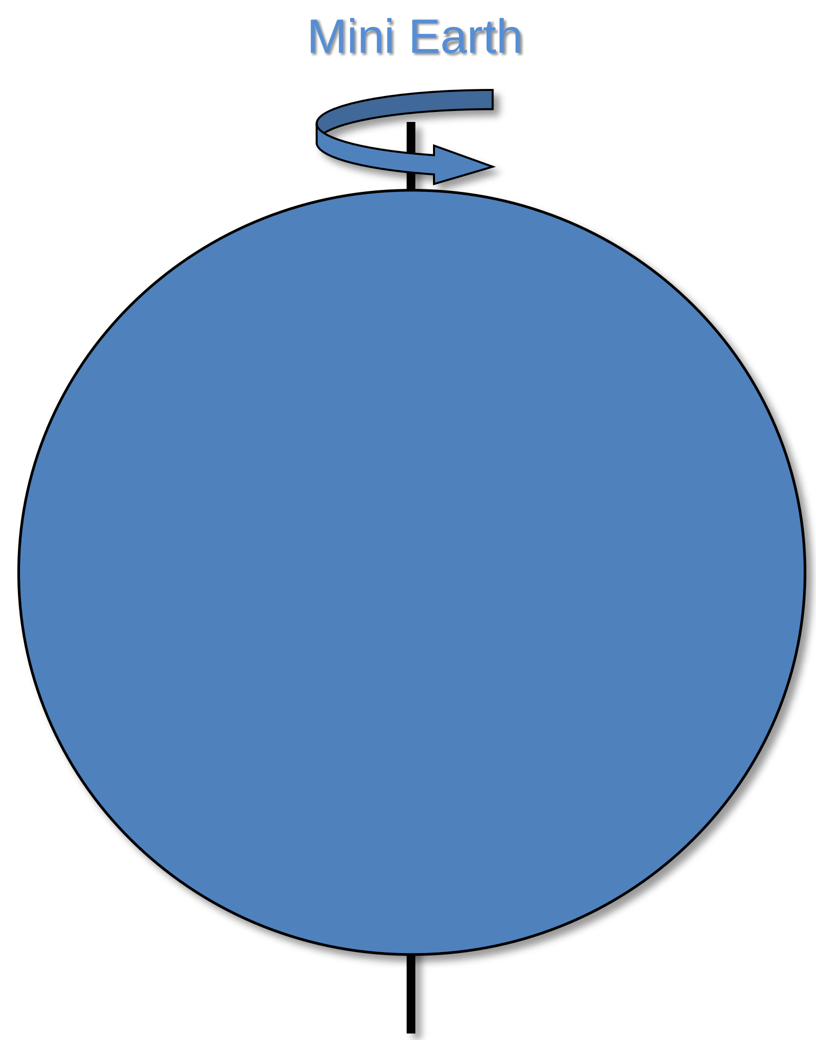
- Even if all rocks were the same and there was no topography gravity would still vary with latitude
  - Least at Equator
  - Most at Poles
  - Total variation ~0.5%

Why?



Latitudinal Gravity Variations: Why?

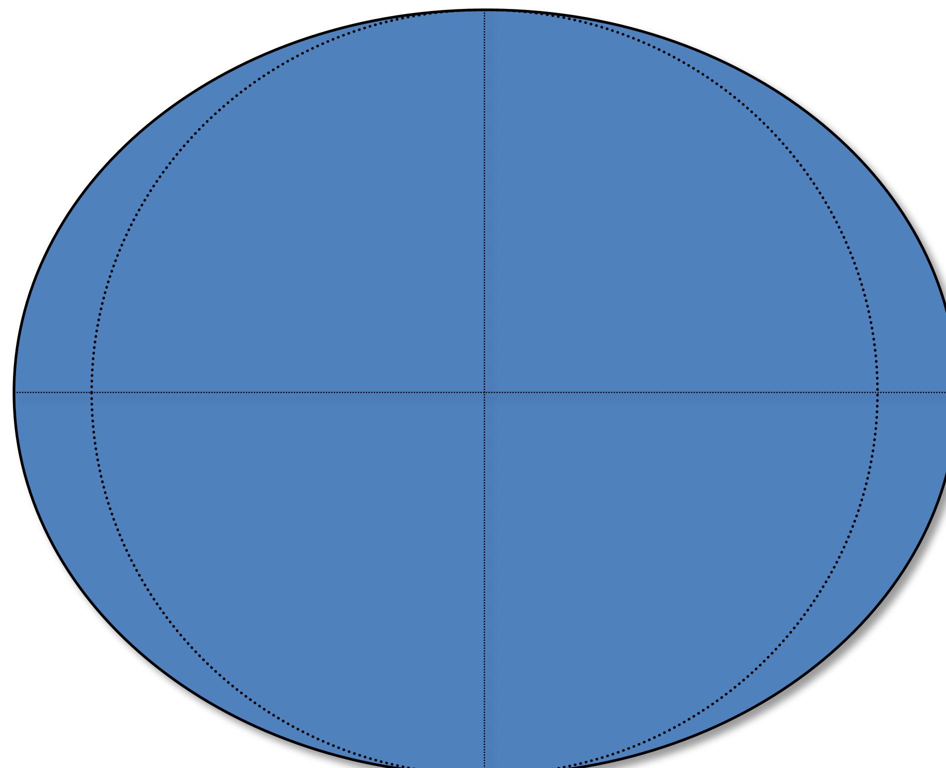
- The Earth rotates on its axis…
  - Creates centrifugal force which depends on:
    - Distance from axis
    - Rate of rotation
  - The linear velocity of a person at the equator is much faster than someone at the poles.
  - Centrifugal force causes Earth to deform
    - Fattest at equator
    - Pinched in at poles



Latitudinal Gravity Variations: Why?

- Because the Earth rotates on its axis…
  - The radius of the Earth is greatest at the equator, least at the poles
    - Gravity depends on distance to center of Earth

$$g = G\frac{M_E}{{R_E}^2}$$

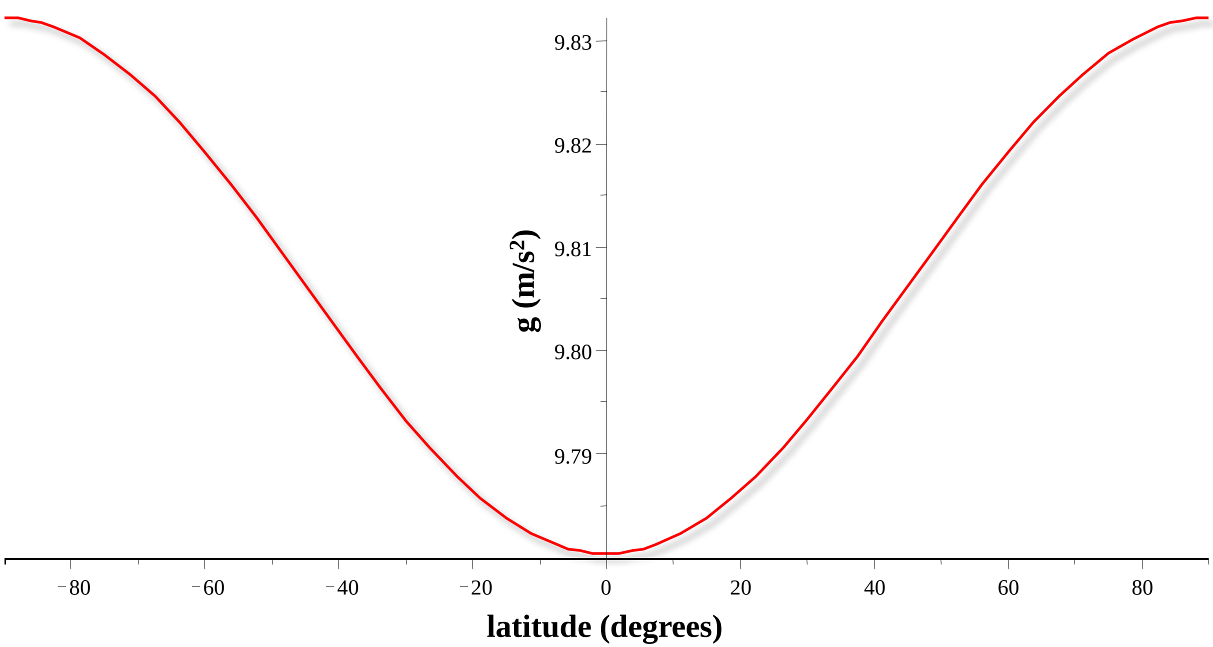


Latitude Correction

- Because gravity varies with latitude:
  - A survey covering a large north/south distance will need to be corrected for latitudinal changes in g.
  - This correction is performed using the International Gravity Formula

$$g_\lambda = 978031.8\left(1 + 0.0053024\sin^2\lambda - 0.0000059\sin^2 2\lambda\right)\text{mGal}$$

  - The correction ends up ~0.8 mGal/km, and given that a good gravimeter can detect a 0.01 mGal change, a N/S movement of only 12 m can be detected.



Eötvös Correction

- If gravity measurements are made on a moving object (car, airplane, ship) a centrifugal acceleration is induced and the gravity measurements must be corrected.
- E.g. because the Earth rotates to the east (counterclockwise when looking down from the north pole):
  - Your weight is reduced due to centrifugal force (~0.34% on the equator due to a rotation of 465 m/s)
  - If you are traveling eastward:
    - measured gravity is less because your motion adds with Earth's rotation
  - If you are traveling westward:
    - You cancel out some of Earth's rotation and measured gravity is more

$$\delta g_{\text{Eötvös}} = 4.040 v \sin\alpha \cos\lambda + 0.001211 v^2 \text{ mGal}$$

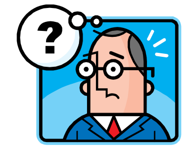
- v = speed in km/hr
- λ = latitude
- α = direction of travel (azimuth)
- This is a huge correction
- ~2.5 mGal per km/hr!
- The main limiting factor in aerial gravity surveys is accurate determination of the airplane velocity

Topographic Corrections

- So far we have assumed that we were taking gravity measurements at the same elevation.
- When gravity measurements are taken at different elevations, up to three further corrections are needed.
  - We also need a way to deal with water!

Free-Air Correction

- Imagine taking a measurement at A (base station) then floating up to evevation B in a balloon (i.e. in the free air!)
  - You just moved farther from the center of Earth, so gravity must decrease!
  - This turns out to be ~0.3086 mGal/meter of elevation change
    - So a gravimeter will respond to changes in elevation of a few cm!

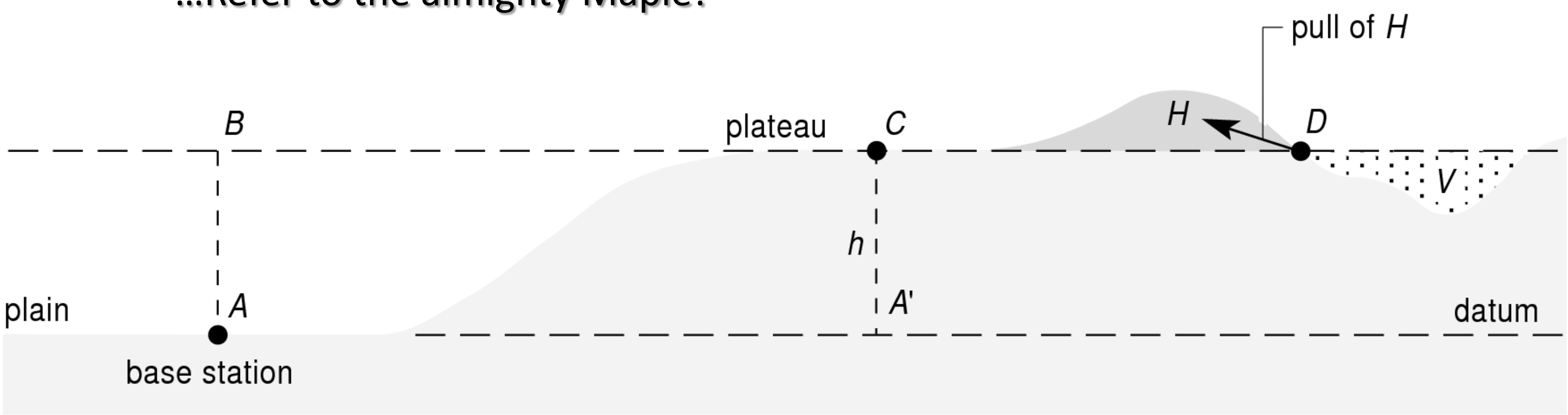


Where did this come from??

  - To find the rate of change in g with elevation, take the derivative of g

$$g = G\frac{M_E}{{R_E}^2} \quad \Rightarrow \quad \frac{dg}{dR} = -2G\frac{M_E}{R^3}$$

  - If the derivative of g (i.e. the rate in change in g w.r.t. R) has $1/R^3$, then how is your book correct when it says that the free-air correction is linear?
  - …Refer to the almighty Maple!



The Free-Air Correction

- The Free-air correction:

This is negative too, so the negatives cancel

$$\delta g_{Free-air} = -h\frac{dg}{dR} \approx 0.3086\,h$$

- When *h* is positive (mountain range):

$$\delta g_{Free-air} > 0$$

- When *h* is negative (Death Valley):

$$\delta g_{Free-air} < 0$$

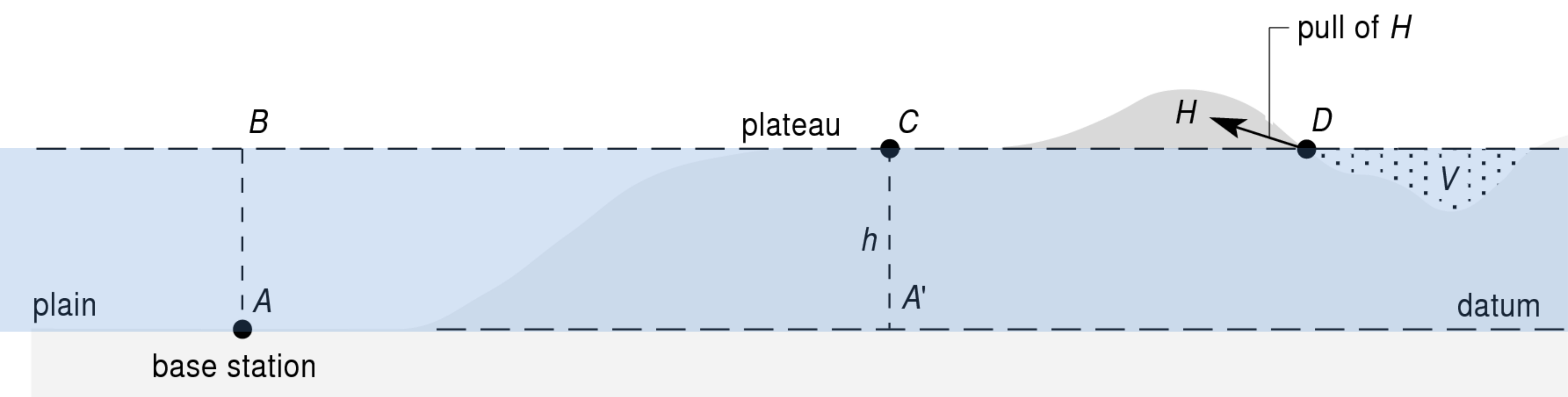
- But what do we do over the ocean?
  - We do not correct using FAC, we use the Bouguer correction

The Bouguer Correction

- Now imagine that you float to point C.
  - You feel less gravity than at A' due to elevation (Free-air correction)
  - You feel more gravity than B due to the mass of rock beneath you (Bouguer correction)
  - If you are on a wide and relatively flat plateau, the extra gravity can be approximated by an infinite sheet/slab

$$\delta g_{Bouguer} = 2\pi G \rho h \left(\frac{m}{s^2} \right)$$

$$\delta g_{Bouguer} = 0.04192 \rho h \, (mGal)$$



The Bouguer Correction

- Since we do not treat negative elevations (i.e. marine surveys) using the Free-air correction
  - We use the Bouguer Correction and apply a negative density to reflect the "missing" mass

LAND: Same as before.
It is common to use:

$$\rho = 2.67\, g/cm^3$$

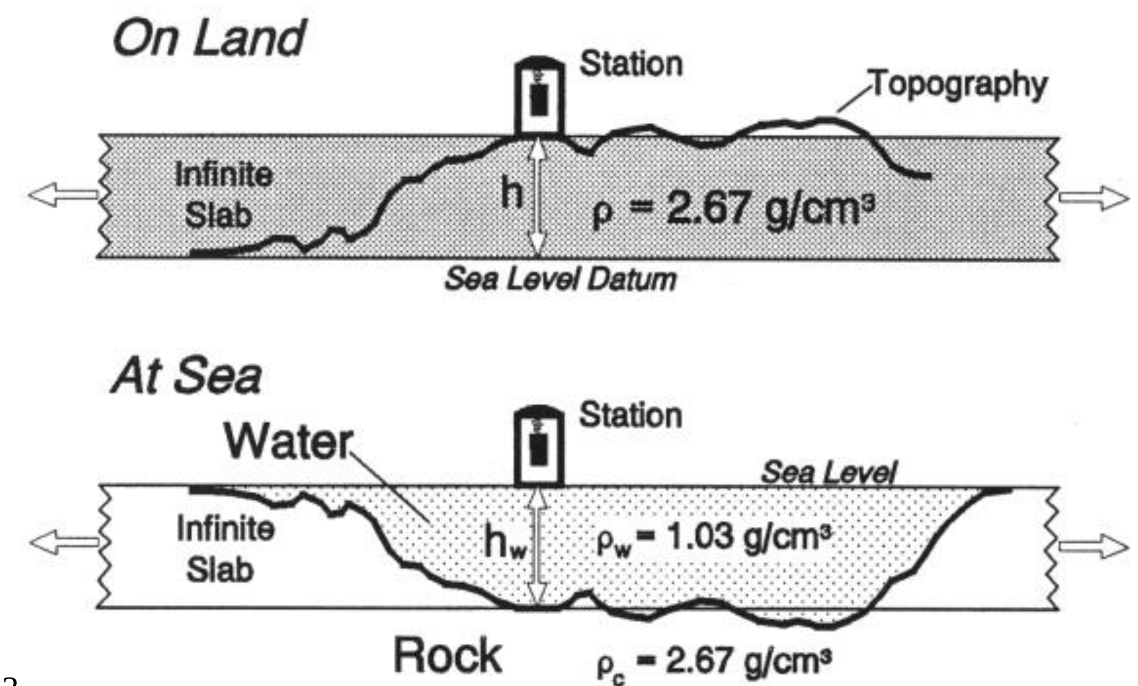
$$\delta g_{Bouguer} = 0.04192\, \rho\, h$$

$$\delta g_{Bouguer} = 0.112\, h\ (mGal)$$

SEA: Bouguer is negative because the water acts like a mass deficit:

$$\rho = (1.03 - 2.67)\, g/cm^3$$

$$\delta g_{Bouguer} = -0.0688\, h_w\ (mGal)$$



$$\delta g_B = \beta(\rho_r - \rho_w) h_w \text{ (g.u.)}.$$

where ρ_r and ρ_w are the densities of rock and sea water respectively, and h_w is the water depth in metres.

The Combined Elevation Correction

- Because the free-air and Bouguer corrections both depend on h (elevation)…
  - we can combine them into a single “combined elevation correction

$$\delta g_{Bouguer} = 0.04192\,\rho\,h\,(mGal)$$

$$\delta g_{Free-air} = 0.3086\,h\,(mGal)$$

$$\delta g_{elevation} = \delta g_{Free-air} - \delta g_{Bouguer}$$

$$\delta g_{elevation} = h(0.3086 - 0.04192\rho)\,(mGal)$$

All together: adding or subtracting the gravity corrections

It is very important to keep physical track of the sign of the corrections; if you do not, you will get the wrong answer. Remember, we are correcting the measured gravity data to remove unwanted effects.

The free-air effect is added if you are above sea-level and is subtracted if you are below sea-level.

The Bouguer effect is subtracted if you are above sea-level (+h) and added if you are below sea-level (-h).

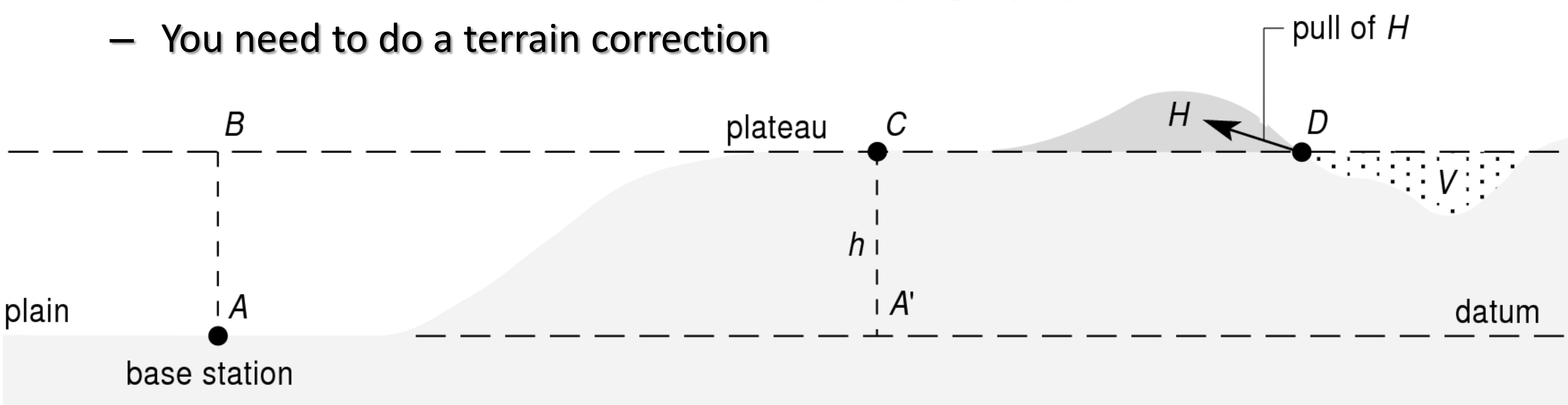
Total Bouguer correction : Bouguer = observed – latitude +/- free-air +/- Bouguer

Total correct to Free-air: Free-air = observed – latitude +/- free-air

The sign of the free-air and Bouguer correction depends on whether the measurements was made above or below ones datum.

Terrain Correction

- The Bouguer correction assumes an infinite slab
  - Reasonable at C, but not at D
- The pull of the mountain, H, would have the same effect as the valley, V
  - Both would reduce g due to the vertical component of pull
- The terrain correction aims to correct for this and depends on:
  - Shape and density of topography
  - Mostly only the nearby features matter (g α $1/R^2$)
- No simple way to do this, so computers are used in conjunction with digital elevation data and knowledge of local rock density
- Rule of thumb: If < 200 m from steep topography
  - You need to do a terrain correction



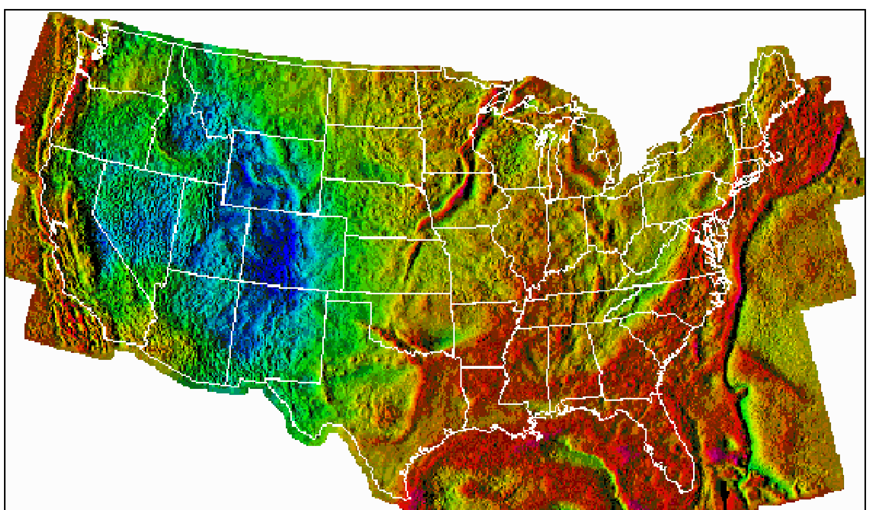
Finally…The Bouguer Anomaly

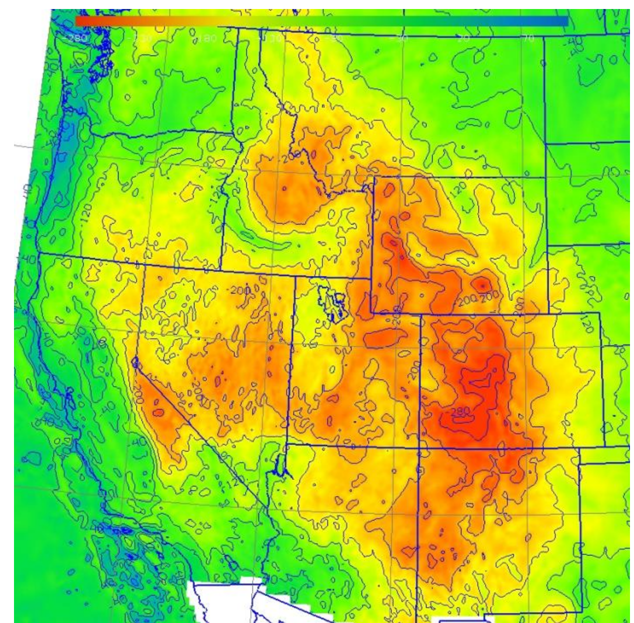
- Once all of the previously mentioned corrections have been made

$$\delta g_{Bouguer\ anomaly} = \delta g_{measured} - \delta g_{latitude} + \delta g_{free-air} - \delta g_{bouguer} + \delta g_{terrain} + \delta g_{Eötvös}$$

  – The result is called the **Bouguer anomaly**
    - Not to be confused with the Bouguer correction, which is different.
    - If the terrain correction is omitted, the result is the **“simple Bouguer anomaly”**

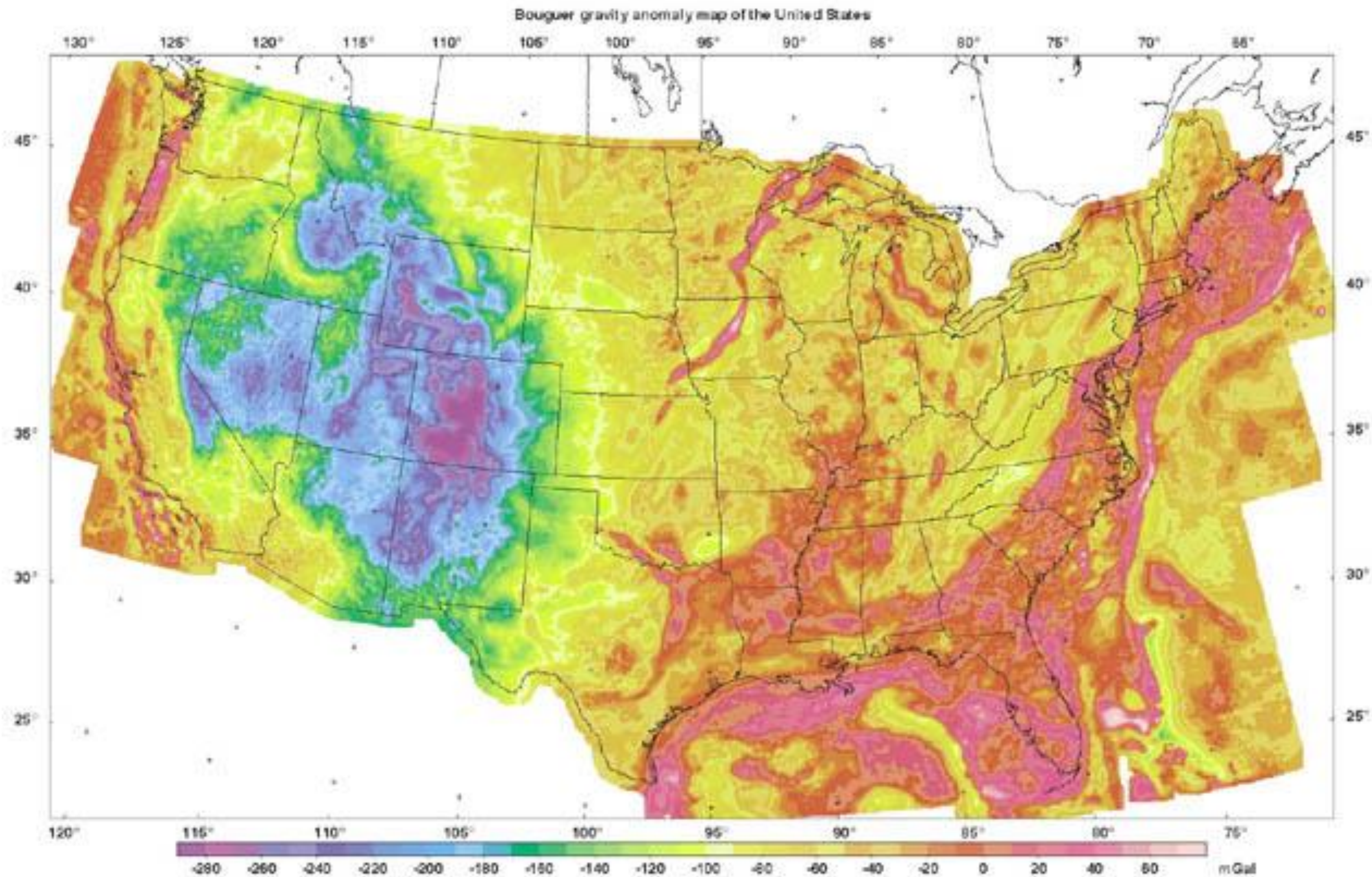
  – The purpose of the Bouguer anomaly is to give the anomaly due to the density variations below the datum, without the effects of topography and latitude





Bouguer Anomaly For The U.S.

- From USGS data

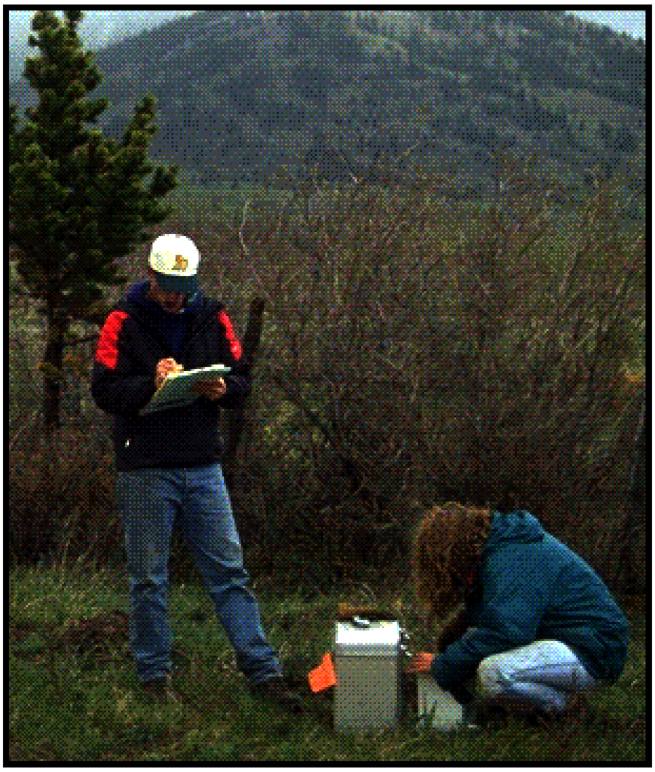


Geophysical Surveys: Active Versus Passive

Introduction:

- Active and Passive Geophysical Methods
- Advantages and Disadvantages of Each Method

Geophysical surveys can be classified into one of two types; *Active* and *Passive*. Passive geophysical surveys are ones that incorporate measurements of naturally occurring fields or properties of the earth. We have already considered passive geophysical surveys in our discussions of gravity and magnetic surveys. In these two cases, the naturally occurring fields are the gravitational and magnetic fields. We simply measure spatial variations in these fields and attempt to infer something about the subsurface geology from these measurements. The fields and properties that we are measuring in this class of experiments exist regardless of our geophysical survey. Examples of other earth properties that could be passively measured include radiometric decay products, certain electrical fields, and certain electromagnetic fields



In conducting active geophysical surveys, on the other hand, a signal is injected into the earth and we then measure how the earth responds to this signal. These signals could take a variety of forms such as displacement, an electrical current, or an active radiometric source. The final two survey methods considered in this short course,

DC resistivity and seismic refraction, are examples of active geophysical experiments.

Advantages and Disadvantages of Active and Passive Experiments

Shown below is a table listing some of the advantages and disadvantages to each of these types of surveys. In reading these, please note that the terms passive and active cover a wide range of geophysical survey methods. Thus, the listed advantages and disadvantages are by necessity generalizations that might not apply to any given specific survey.

Active		Passive	
Advantage	**Disadvantage**	**Advantage**	**Disadvantage**
Better control of noise sources through control of injected signal.	**Because both sources and receivers are under the surveyor's control, he must supply both. Therefore, field equipment tends to be more complex.**	**Surveyor need only record a naturally occurring field, therefore, he need supply only a sensor and a data recorder.**	**Less control of noise because source of the signal is out of the hands of the surveyor.**
Because propagating fields are generally measured, active experiments usually provide better depth control over source of anomalous signal.	**Field operations and logistics are generally more complex and time consuming than passive experiments.**	**Field operations are generally very time efficient. Thus, passive experiments can be run over wider areas in a more cost-effective manner.**	**Because passive fields are generally the result of integrating anomalous geological contributions overwide areas, identification of the source of an anomalous observation can be difficult.**
Many different	**Many different**	**One or two well-**	**One or two well-**

source/receiver configurations can be used allowing for a wide variety of survey designs. This allows survey designers great flexibility in customizing surveys for particular problems.	**source/receiver configurations can be used allowing for a wide variety of survey designs. The increase in the number of field options inevitably leads to greater survey design costs and potentially leads to increased probability of field mishaps.**	**established field procedures are generally used. Contractors can provide these surveys on short notice with relatively easily quantifiable results.**	**established field procedures are generally used. This limits the amount of customization that can be done for specific problems.**
Once set up, active experiments are capable of producing vast quantities of data that can be used to interpret subtle details of the earth's subsurface.	**The large quantity of data obtained in many active experiments can become overwhelming to process and interpret.**	**Interpretation of the limited set of observations can be accomplished with modest computational requirements quickly and efficiently.**	**The data sets collected in passive experiments are smaller than those collected in active experiments and usually do not allow for as detailed an interpretation.**

Electrical Methods Overview

Bridging our subdivision of geophysical techniques into passive and active methods are the electrical and electromagnetic methods. Taken as a whole, the electrical and electromagnetic methods represent the largest class of all geophysical methods; some passively monitor natural signals while others employ active sources. In addition to their great variety, this group of geophysical techniques represents some of the oldest means of exploring the Earth's interior. For example the *SP*(Self Potential) method described below dates back to the 1830s when it was used in Cornwall, England by Robert Fox to find extensions of known copper deposits. Natural electrical currents in the Earth, referred to as *telluric* currents were first identified by Peter Barlow (pictured) in 1847. The *EM* method was developed in the 1920s for the exploration of base-metal deposits.

Electrical methods employ a variety of measurements of the effects of electrical current flow within the Earth. The phenomena that can be measured include current flow, electrical potential (voltages), and electromagnetic fields. A summary of the better-known electrical methods is given below. In this set of notes we will consider only one of these methods, the DC resistivity method, in greater detail.

DC Resistivity - This is an active method that employs measurements of electrical potential associated with subsurface electrical current flow generated by a DC, or slowly varying (very low frequency current) AC, source. Factors that affect the measured potential, and thus can be mapped using this method, include the presence and quality of pore fluids and clays. Our discussions will focus solely on this method.

Induced Polarization (IP) - This is an active method that is commonly done in conjunction with DC Resistivity. It employs measurements of the transient (short-term) variations in potential as the current is initially applied or removed from the ground, or alternatively the variation in the response as the AC frequency is changed. It has been observed that when a current is applied to the ground, the ground behaves much like a capacitor, storing some of the applied current as a charge that is dissipated upon removal of the current. In this process, both capacitative and electrochemical effects are responsible.

IP is commonly used to detect concentrations of clay, and electrically conductive metallic mineral grains.

Electromagnetic (EM) - This is an active method that employs measurements of a time-varying magnetic field generated by induction through current flow within the earth. In this technique, a time-varying magnetic field is generated at the surface of the earth that produces a time-varying electrical current in the earth through induction. A receiver is deployed that compares the magnetic field produced by the current-flow in the earth to that generated at the source. EM is used for locating conductive base-metal deposits, for locating buried pipes and cables, for the detection of unexploded ordnance, and for near-surface geophysical mapping.

Self Potential (SP) - This is a passive method that employs measurements of naturally occurring electrical potentials commonly associated with shallow electrical conductors, such as sulfide ore bodies. Measurable electrical potentials have also been observed in association with groundwater flow and certain biologic processes. The only equipment needed for conducting an SP survey is a high-impedance voltmeter and some means of making good electrical contact to the ground.

Magnetotelluric (MT) - This is a passive method that employs measurements of naturally occurring electrical currents, telluric currents, generated by magnetic induction from electrical currents in the ionosphere. This method can be used to determine electrical properties of materials at relatively great depths (down to and including the mantle) inside the Earth. In this technique, a time variation in electrical potential is measured at a base station and at survey stations. Differences in the recorded signal are used to estimate subsurface distribution of electrical resistivity

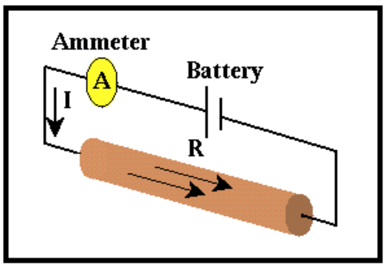
Resistivity Basics

Current Flow and Ohm's Law

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

$$\boxed{V = IR}$$

Ohm found that the current, I, was proportional to the voltage, V, for a broad class of materials that we now refer to as *ohmic* materials. The constant of proportionality is called the *resistance* of the material and has the units of voltage (volts) over current (amperes), $(\Delta V/I)$ or *ohm*.



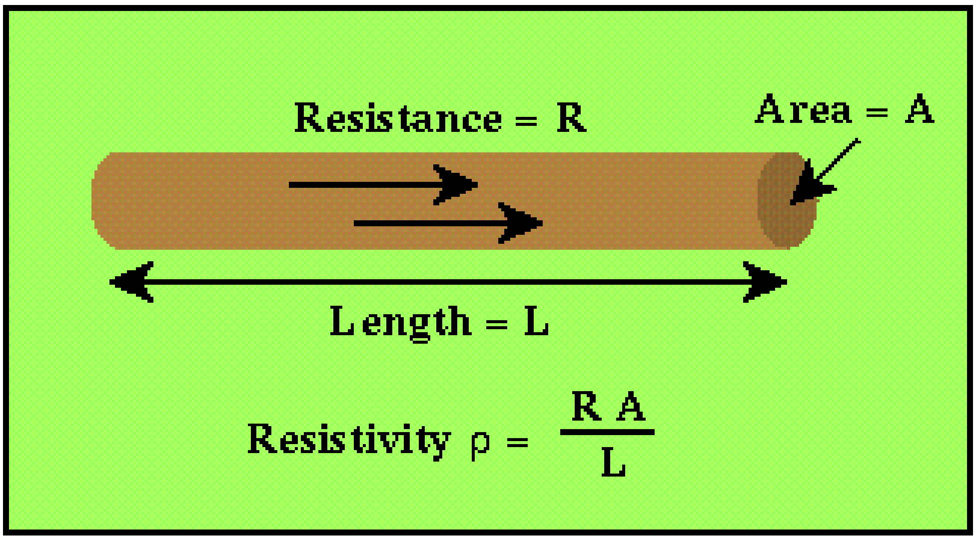
In principle, it is relatively simple to measure the resistance of a strand of wire. Connect a battery to the wire of known voltage and then measure the current flowing through the wire. The voltage divided by the current yields the resistance of the wire. In essence, this is how your multimeter measures resistance. In making this measurement, however, we must ask two crucial questions.

- Is the measured resistance related to some fundamental property of the material from which the wire is made?
- How can we apply this relatively simple experiment to determining electrical properties of earth materials?

<u>It's Resistivity, not Resistance</u>

The problem with using resistance as a measurement is that it depends not only on the material out of which the wire is made, but also the geometry of the wire. If we were to increase the length of wire, for example, the measured resistance would increase. Also, if we were to decrease the diameter of the wire, the measured resistance would increase. We want to define a property that describes a material's ability to transmit electrical current that is independent of the geometrical factors.

The quantity that is used is called *resistivity* and is usually indicated by the Greek symbol ρ(read *rho**,**).



In the case of the wire, resistivity is defined as the resistance in the wire, multiplied by the cross-sectional area of the wire, divided by the length of the wire. The units associated with resistivity are thus **ohm.m** (ohm - meters).

Resistivity is a fundamental parameter of the material making up the wire that describes how easily the wire can transmit an electrical current. High values of resistivity imply that the material making up the wire is very resistant to the flow of electricity. Low values of resistivity imply that the material making up the wire transmits electrical current very easily.

Resistivity of Earth Materials:

Although some native metals and graphite conduct electricity, most rock-forming minerals are electrical insulators. Measured resistivities in Earth materials are primarily controlled by the movement of charged ions in pore fluids. Although water itself is not a good conductor of electricity, ground water generally contains dissolved compounds that greatly enhance its ability to conduct electricity. Hence, porosity and fluid saturation tend to dominate electrical resistivity measurements. In addition to pores, fractures within crystalline rock can lead to low resistivities if they are filled with fluids.

The resistivities of various earth materials are shown below.

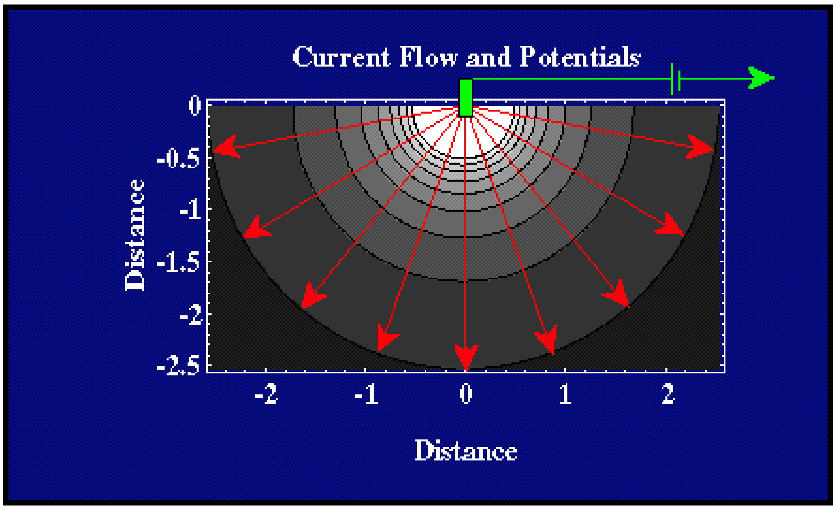
Material	Resistivity (Ohm.meter)
Air	~ ?
Pyrite	2.9 x 10^-5 - 1.5
Galena	3 x 10^-5 - 3 x 10^2
Sphalerite	1.5 - 1 x 10^7
Quartz	4 x 10^10 - 2 x 10^14
Calcite	2 x 10^12
Rock Salt	30 - 1 x 10^13
Mica	9 x 10^12 - 1 x 10^14
Ground Water	0.5 - 300
Sea Water	0.2

Diabase	**20 - 5 x 10^7**
Limestones	**50 - 1 x 10^7**
Sandstones	**1 - 6.4 x 10^8**
Shales	**20 - 2 x 10^3**
Gabbro	**1 x 10^3 - 1 x 10^6**
Basalt	**10 - 1.3 x 10^7**
Dolomite	**3.5 x 10^2 - 5 x 10^3**

Like susceptibilities, there is a large range of resistivities, not only between varying rocks and minerals, but also within rocks of the same type. This range of resistivities, as described above, is primarily a function of fluid content. Thus, a common target for electrical surveys is the identification of fluid saturated zones. For example, resistivity methods are commonly used in engineering and environmental studies for the identification of water table.

Current Densities and Equipotentials

To describe the nature of electrical current flow in media occupying a volume, we must move beyond the simple notions of current and voltage gained from our experience with wires, resistors, and batteries. In the Earth, or any three-dimensional body, electrical current is not constrained to flow along a single path as it does in a wire. Consider as an example the situation shown below. A battery is connected to the earth by wires and electrodes at two remote points (that is the electrical connections to the earth are very distant from one another). The Earth, not being a perfect insulator, conducts the electrical current imported by the battery. At this stage, lets assume the resistivity of the earth is uniform throughout the Earth. How does the current flow through the Earth?



In this example, current flows (the red lines) out from the electrode (the green square) radially along straight lines (the second electrode is far to the right of the figure as it is drawn). The resistivity of the medium imposes a voltage drop as we move away from the electrode. If we could take a voltmeter and measure the voltage drop between a point very far from the current electrode to various places in (on) the medium near the electrode, we would find that the voltage drops would be constant along circular lines centered at the electrode. (That is, one of the leads to the voltmeter would make contact with the ground at a distance very far from the electrode, while the other is then moved throughout the medium). These lines are referred to as *equipotentials* (think equal voltage). In three dimensions, they form hemispheres centered on the electrodes. Several equipotential lines are shown in black with the voltage drop associated by each line shown in gray scale. The darker the gray scale, the smaller the potential drop between this location and a location very far from the current electrode.

$$\boxed{V = \frac{\rho I}{2\pi r}}$$

Voltage differences between any two points in the medium can be computed by simply subtracting the potentials at the two points. Thus, if the two points lie on a hemisphere centered at the current electrode, there will be no voltage difference recorded, because these two locations lie along an equipotential surface. That is, if you were to take your voltmeter and connect to two points within the earth that were on the same equipotential surface, you would read a voltage difference of zero. When compared to the potential near the electrode, voltage differences will increase away from the electrode. This should make sense, what you are measuring with your voltmeter is proportional to the current passing through the media times the resistance of the media as given by Ohm's law. As you move away from the electrode, your current is traveling through more of the media. The resistance increases as the path increases, hence, the voltage increases.

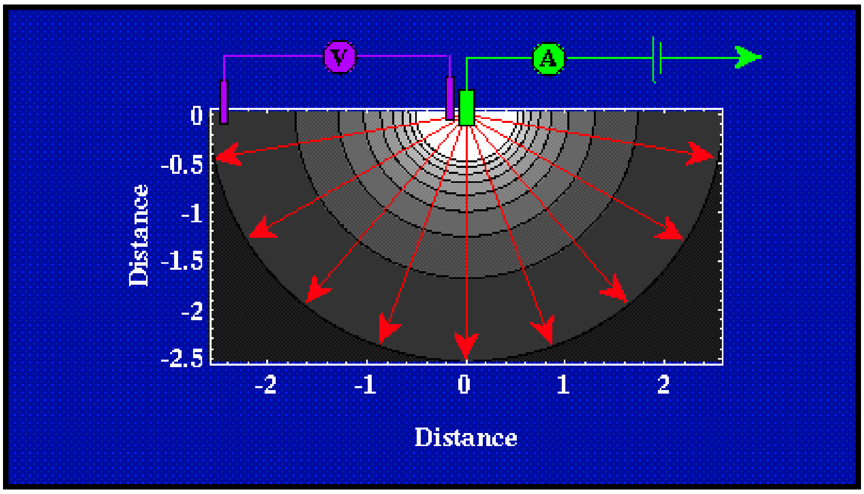
At any point in the medium, the *current density* is defined as the amount of current passing through a unit area of an equipotential surface. Thus, close to the electrode, all of the current is passing through a very small volume. The current crossing any equipotential surface normalized by the area of the surface will thus be high. Far away from the electrode, this same current occupies a much larger volume of the medium. The total current (which is the same regardless of where the surface is in the volume) crossing any equipotential surface, normalized by the area of the surface (which is now large), will be small.

Current density has the units of Amperes per meter squared.

A First Estimate of Resistivity

The voltage change from a single current electrode to some point in the half space representing the earth is given by the expression to the right. In this expression, *V* is voltage, *I* is current, *rho* is resistivity, and *r* is the distance between the current electrode and the point the voltage is measured. Notice that this expression is nothing more than Ohm's law with the resistance, *R* equal to *rho* over *2 ? r*.

If the Earth had a constant resistivity (it doesn't) we could estimate this resistivity by performing the following experiment. Attach to the wire connecting the battery with one of the current electrodes an ammeter to measure the amount of current going into the earth. Place one electrode connected to a voltmeter next to the current electrode and place the other at some distance, *r*, away from the electrode and measure the voltage difference between the two locations. Using the expression given above, compute the resistivity, *rho*.



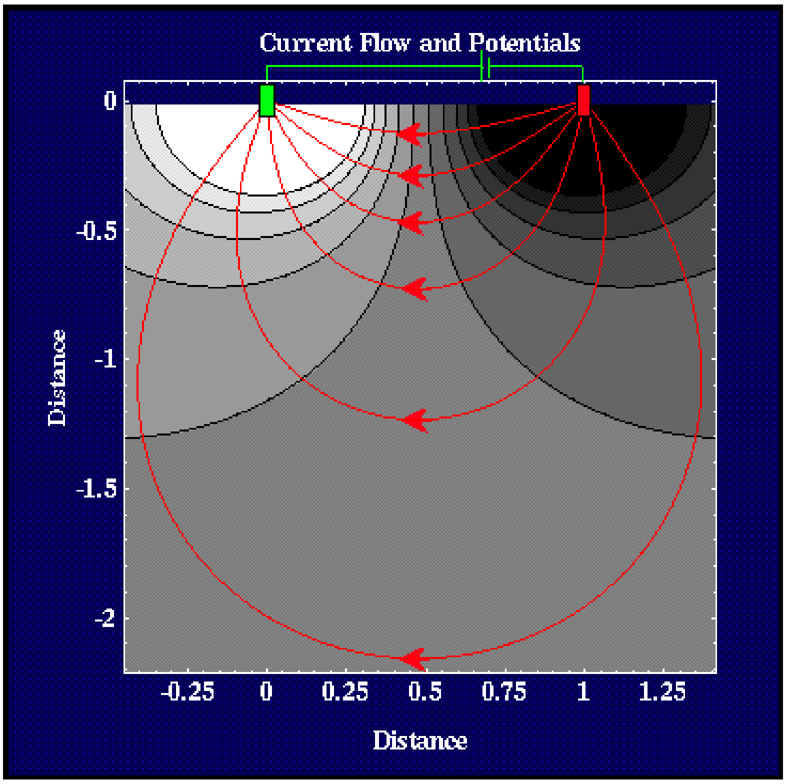
In practice, this experiment could be difficult to implement because the two current electrodes must be placed a great (usually 10 times the distance over which the voltage is being measured) distance from one another. So, why not simply decrease the distance between the two voltage electrodes so the distance between the two current electrodes remains at a practical distance? The problem is that the closer the two voltage electrodes are to each other, the smaller the voltage difference that must be measured. Thus, there is a practical limit to how close the two voltage electrodes can be and thereby how close the two current electrodes can be. More importantly, there is no need to place the second electrode at "infinity" if we develop a method to take its effect into account.

As another note, one may ask why don't we simply attach the voltmeter to the wire in which the current is flowing and measure the voltage drop between the two current electrodes. This could be done. In practice, however, it is impossible to obtain information about the Earth, because what you measure is more a function of the *contact resistance* between the earth and the current electrodes than of the resistivity of the Earth. The contact resistance is the resistance that is encountered to current flow because the electrode does not make perfect electrical contact with the earth. Contact resistances can be quite large, on the order of kilo-ohms (10^4 ohms), although good field practice can reduce these to the order of an ohm, if necessary. However, the *resistance* between the electrodes will be made up of the contact resistance at each electrode, and the resistance of the Earth - which is effectively zero (the "wire" had a very large cross-sectional area!). So, the voltage measured will be dominated by the voltage drop over the electrode/earth contacts, even if it is small.

If a large (infinite) impedance voltmeter is used to make the voltage measurement across two *separate* voltage electrodes, however, very little current actually flows through the voltage electrodes and contact resistance is unimportant to the measurement.

Current Flow From Two Closely Spaced Electrodes

In practice, we will need to place the two current electrodes close to each other. In doing so, however, the current distribution and equipotentials produced within a homogeneous earth become more complicated than those shown previously.



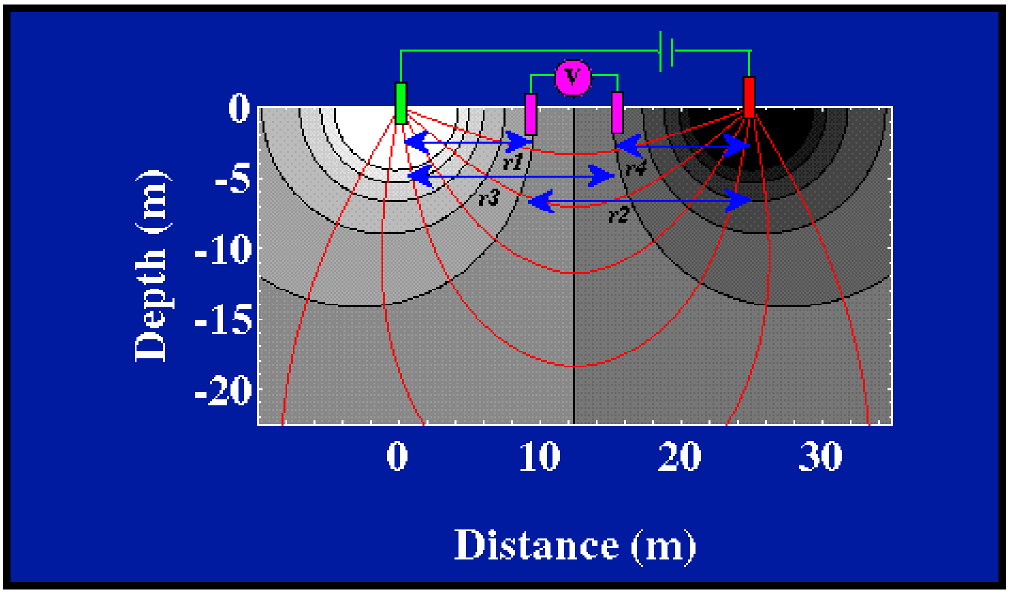
Instead of the current flowing radially out from the current electrodes, it now flows along curved paths connecting the two current electrodes. Six current paths are shown (red lines). Between the surface of the earth and any current path we can compute the total proportion of current encompassed. The table below shows this proportion for the six paths shown above. Current paths are labeled 1 through 6 starting with the topmost path and proceeding to the bottom-most path.

Current Path	% of Total Current
1	17
2	32
3	43
4	49
5	51
6	57

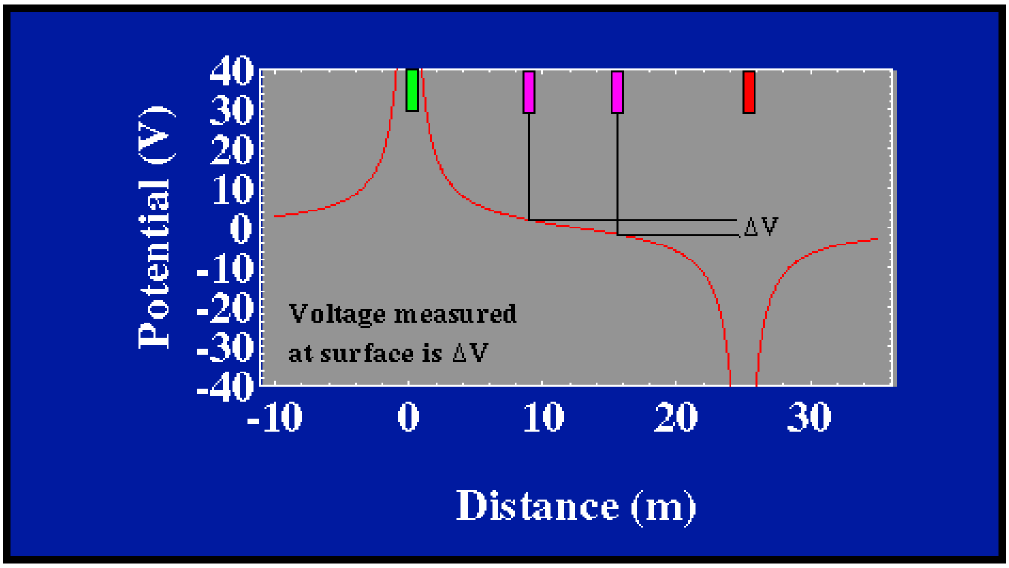
From these calculations and the graph of the current flow shown above, notice that almost 50% of the current placed into the ground flows through rock at depths shallower or equal to the current electrode spacing.

A Practical way of Measuring Resistivity

Using an experimental configuration where the two current electrodes are placed relatively close to one another as described previously and using two potential electrodes place between the two current electrodes, we can now estimate the resistivity of our homogeneous earth. The configuration of the four electrodes for this experiment is shown below. Let the distances between the four electrodes be given by *r1, r2, r3,* and *r4* as shown in the figure.



The <u>potential</u> computed along the surface of the earth is shown in the graph below. The voltage we would observe with our voltmeter is the <u>difference in potential</u> at the two voltage electrodes, l. The horizontal positions of the four electrodes, two current (red and green), and the two potential (purple) are indicated along the top of the figure.



Notice, that in this configuration, the voltage recorded by the voltmeter (l) is relatively small. That is, the difference in the potential at the locations of the two potential electrodes is small. We could increase the size of the voltage recorded by the voltmeter by moving the two potential electrodes outward, closer to the two current electrodes. For a variety of reasons, some related to the reduction of <u>noise</u> and some related to maximizing the depth over which our measurements are sensitive, we will typically not move the potential and current electrodes close together. Thus, a very sensitive voltmeter must be used. In addition to having a large impedance, voltmeters need to be able to record voltage differences down to mV (10^-3 volts). If the potential electrodes were moved closer to the two current electrodes, larger voltages would be recorded. For a variety of reasons, however, we will typically not do this in the field.

Knowing the locations of the four electrodes, and by measuring the amount of current input into the ground, *i*, and the voltage difference between the two potential electrodes, l, we can compute the resistivity of the medium, *rho-a*, using the following equation.

$$\rho_a = \frac{2\pi\Delta V}{i}\left[\frac{1}{\left(\frac{1}{r_1} - \frac{1}{r_2} - \frac{1}{r_3} + \frac{1}{r_4}\right)}\right]$$

In this particular case, regardless of the location of the four electrodes, *rho-a* will be exactly equal to the resistivity of the medium. The resistivity computed using the equation given above is referred to as the *apparent resistivity*. We call it the apparent resistivity for the following reason. We can always compute *rho-a*, we only need to know the locations of the electrodes and measure the current and voltage. If, however, the Earth does not have a constant resistivity (that is if the resistivity varies with depth or horizontally), the resistivity computed by the above equation will not represent the true resistivity of the Earth. Thus, we refer to it as an apparent resistivity.

As a final caveat, as written above, the difference between the apparent and the true resistivity of the medium is not a function of any noise that might be associated with the measurements we are attempting to record. The difference rather comes from the fact that our measurement, in some sense, averages the true resistivities of some region of the earth, yielding an apparent resistivity that may not represent the true resistivity at some specific point within the earth.

Resistivity Surveys and Geology

1- **Sources of Noise**

2- Depth of Current Penetration Versus Current Electrode Spacing

3- Current Flow in Layered Media

4-Variation in Apparent Resistivity: Layered Versus Homogeneous Media

5- Current Flow in Layered Media Versus Electrode Spacing

6- A Second Example of Current Flow in Layered Media

Sources of Noise

Even given the simple experiment outline on the previous page, there are a number of sources of noise that can affect our measurements of voltage and current from which we will compute apparent resistivities.

1-Electrode Polarization - A metallic electrode, like a copper or steel rod, in contact with an electrolyte, ground water, other than a saturated solution of one of its own salts will generate a measurable contact potential. In applications such as SP, these contact potentials can be larger than the natural potential that you are trying to record. Even for the DC methods described here these potentials can be a significant fraction of the total potential measured.

For DC work, there are two possible solutions.

1. a-Use ***nonpolarizing*** *electrodes*. These are electrodes that contain a metallic conducting rod in contact with a saturated solution of its own salt. Copper and copper sulfate solution are commonly used. The rod and solution are placed in a porous ceramic container that allows the saturated solution to slowly leak out and make contact with the ground. Because these solutions are rather environmentally non-friendly, and because the method described below is easy to employ, these so-called *porous pot* electrodes are rarely used in DC work. They are, however, commonly used in SP and IP surveys.

(1-b): A simple method to avoid the influence of these contact potentials is to periodically reverse the current flow in the current electrodes or use a slowly varying, a few cycles per second, AC current. As the current reverses, the polarizations at each electrode break down and begin to reverse. By measuring over several cycles, robust current and voltage measurements can be made with negligible polarization effects.

2--Telluric Currents -As described previously, naturally existing currents flow within the earth. These currents are referred to as telluric currents. The existence of these currents can generate a measurable voltage across the potential electrodes even when no current is flowing through the current electrodes.

Solution of (2):By periodically reversing the current from the current electrodes, or by employing a slowly varying AC current, the effects of telluric currents on the measured voltage can be cancelled.

3-Presence of Nearby Conductors -Electrical surveys can not be performed around conductors that make contact with the ground. For example, the presence of buried pipes or chain-linked fences will act as current sinks. Because of their low resistivity, current will preferentially flow along these structures rather than flowing through the earth. The presence of these nearby conductors essentially act as electrical shorts in the system.

Solution of (3) : choose your VES's locations far away from the conductors connecting the earth.

4-Low Resistivity at the Near Surface -Just as nearby conductors can act as current sinks that short out an electrical resistivity experiment, if the very near surface has a low resistivity, it is difficult to get current to flow more deeply within the earth. Thus a highly conductive* near-surface layer such as a perched water table can prevent current from flowing more deeply within the earth.

Solution (4):Do as we did with near surface conductors previously mentioned.

5-Near-Electrode Geology and Topography - Any variations in geology, or water content localized around an electrode, which produces near-surface variations in resistivity can greatly influence resistivity measurements. In addition, rugged topography will act to concentrate current flow in valleys and disperse current flow on hills.

Solution(5) you can select another sites of smooth topography for your Ves's

6-Current Induction in Measurement Cables - Current flowing through the cables connecting the current source to the current electrodes can produce an induced current in the cables connecting the voltmeter to the voltage electrodes, thereby generating a spurious voltage reading.

Solution (6): This source of noise can be minimized by keeping the current cables physically away from, a meter or two, the voltage cables.

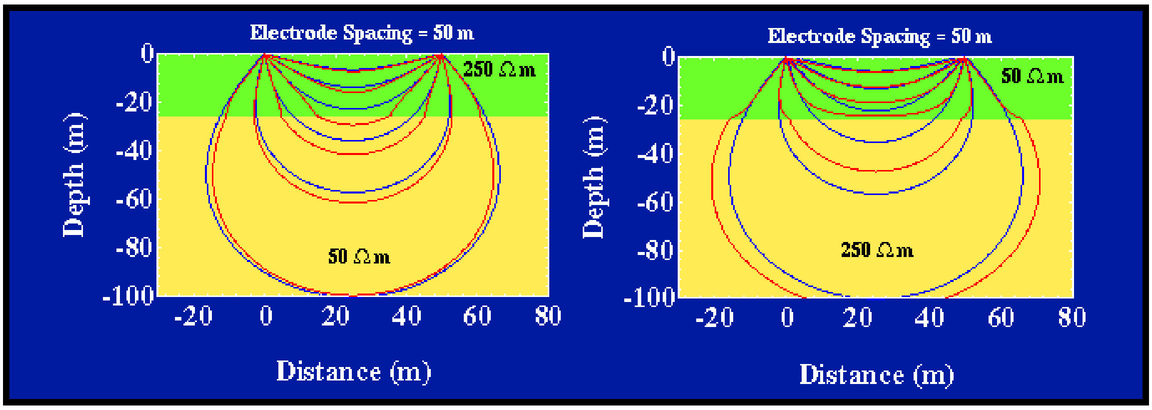
Notes

**Conductivity* is the inverse of resistivity. Highly *conductive* media transmit electrical current with great ease, thus they have a low *resistivity*. Mathematically, conductivity is the reciprocal of resistivity and is measured in the units of 1 over Ohm meters. 1/Ohm is referred to as a *siemen* (S) (sometimes, as "mho"). Thus, the units of conductivity are siemens per meter.

Current Flow in Layered Media

How does the presence of depth variations in resistivity affect the flow of electrical current? In the previous examples, we assumed that the Earth has a constant resistivity. Obviously this isn't true or else we wouldn't be trying to map the variation in resistivity throughout the earth. Although resistivity could conceivably vary in depth and in horizontal position, we will initially only consider variations in depth. In addition, we will assume that these depth variations in resistivity can be quantized into a series of discrete layers, each with a constant resistivity. Thus, initially we will not consider variations in resistivity in the horizontal direction or continuous variations in depth*.

Shown below are current-flow paths (red) from two current electrodes in two simple two-layer models. The model to the left contains a high-resistivity layer (250 ohm.m) overlying a lower resistivity layer (50 ohm.m). This model is characteristic of the resistivity profile that would be found in a region where unsaturated alluvium overlies water saturated alluvium. The model to the right contains a low-resistivity layer (50 ohm.m) overlying a higher resistivity layer (250 ohm.m). This model is characteristic of a perched aquifer. For comparison, we've also shown the paths current would have flowed along if the Earth had a constant resistivity (blue) equal to that of the top layer. These paths are identical to those described previously.



Notice that the current flow in the layered media deviates from that observed in homogeneous media. In particular, notice that in layered media the current flow lines are distorted in such a way that current preferentially seems to be attracted to the lower-resistivity portion of the layered media. In the model on the left, current appears to be pulled downward into the 50 ohm.m layer. In the model on the right, current appears to be bent upward, trying to remain within the lower resistivity layer at the top of the model. This shouldn't be surprising. What we are observing is the current's preference toward flowing through the path of least resistance. For the model on the left, that path is through the deep layer. For the model on the right, that path is through the shallow layer.

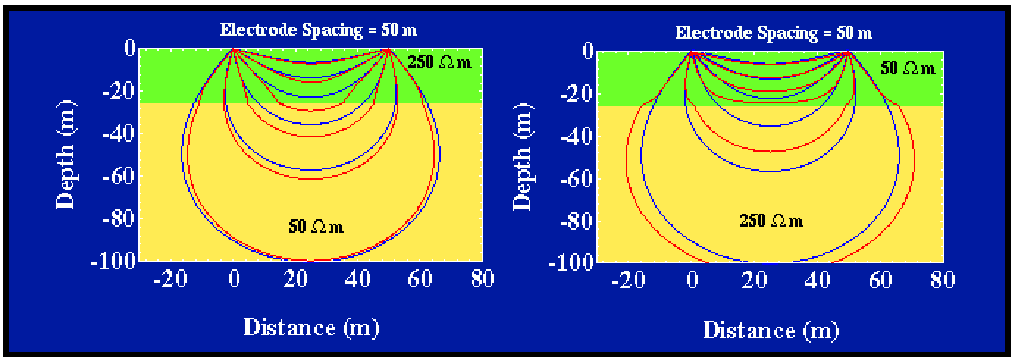
Note

*For all practical purposes this layered model does allow for continuous variations in resistivity with depth because we have made no constraints on the number of layers allowed in the model or their thicknesses. Thus, a smoothly varying resistivity depth profile could be approximated by a large number of very thin, constant resistivity layers.

Variation in Apparent Resistivity: Layered Versus Homogeneous Media

An important consequence of the deviation in current flow in layered media is how it can affect our measurements of apparent resistivity. Imagine that we configured an electrical experiment over these two models by measuring the potential difference between two places on the surface of the earth between the two current electrodes and compute the apparent resistivity. In these examples we will assume that the potential electrodes are between the two current electrodes and have a fixed separation that remains constant throughout the experiment. This is the same geometry for a four electrode experiment, two current and two potential, that was described previously.

Because current is preferentially being pulled into the lower layer for the model on the left, the current density between the two current electrodes near the surface of the Earth (where we are measuring electrical potential) will be smaller than that which would be observed if the Earth were homogeneous. By the same token for the model on the right, the current density would be higher than that observed in a homogeneous earth because the current is being preferentially channeled through the near-surface layer.

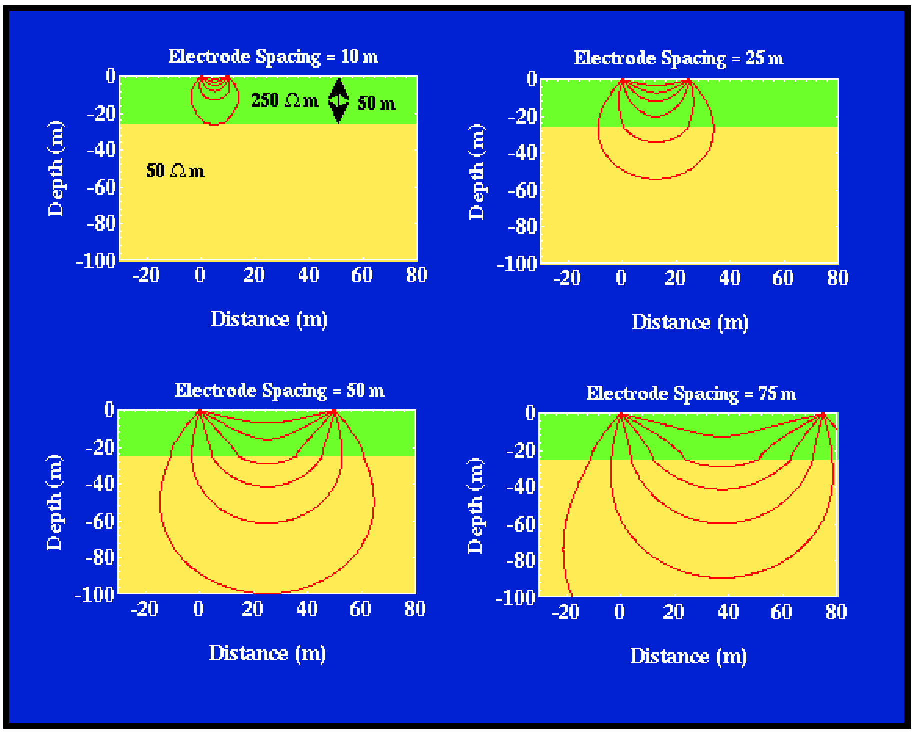


Recall, that our expression for the computation of apparent resistivity, shown below, is a function of electrode spacing r (which is the same for the two situations shown above), current i (assume that we are putting the same current in the ground for each model), and potential difference $\mathcal{l}$ (voltage) between the two potential electrodes. It can be shown that the potential difference, *l*, is proportional to the current density around the potential electrodes. Thus, for the case shown on the left the potential difference will be smaller than would have been observed in a homogeneous Earth because the current density is smaller than that which would have been observed in a homogeneous earth. Therefore, the measured apparent resistivity will be decreased. Conversely, for the case shown of the right the potential difference will be larger than that observed in a homogeneous earth and the measured apparent resistivity will likewise be larger.

Current Flow in Layered Media Versus Current Electrode Spacing

Imagine that we conduct a series of four electrode experiments, each centered about the same point. Let's assume that the potential electrodes remain centered between the current electrodes and that their separation is held fixed. Initially, the current electrodes are placed close together and we measure current and voltage from which we compute apparent resistivity. Then we perform the same experiment, but we systematically increase the current electrode spacing while holding the potential electrode spacing fixed. What will happen?

Consider the earth model shown below, a high resistivity layer over

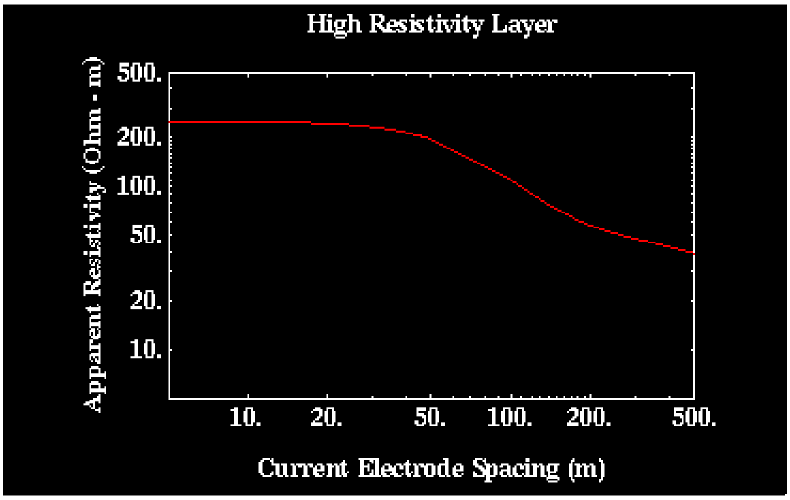


a lower resistivity layer.

When the current electrodes are closely spaced, in the region surrounding the potential electrode positions (between the two current electrodes), most of the current flows through the upper layer along paths that are close to those that they would have flown along if the model were homogeneous. That is, in this electrode configuration, current flow is not sufficiently perturbed near the potential electrodes for us to be able to distinguish between this layered model and a homogeneous Earth model with a resistivity equal to the resistivity of the top layer. Thus, the computed apparent resistivity will be close to the resistivity of the upper layer, 250 ohm.m

Now, we increase the current electrode spacing and repeat the same experiment. At larger current electrode spacings, the current flow near the potential electrodes is significantly altered by the presence of the subsurface boundary. In this case, current is preferentially drawn downward into the lower resistivity layer, decreasing the current density between the two current electrodes where we will measure the voltage with our two potential electrodes. This decrease in current density will cause our computed value of apparent resistivity do decrease from 250 ohm.m

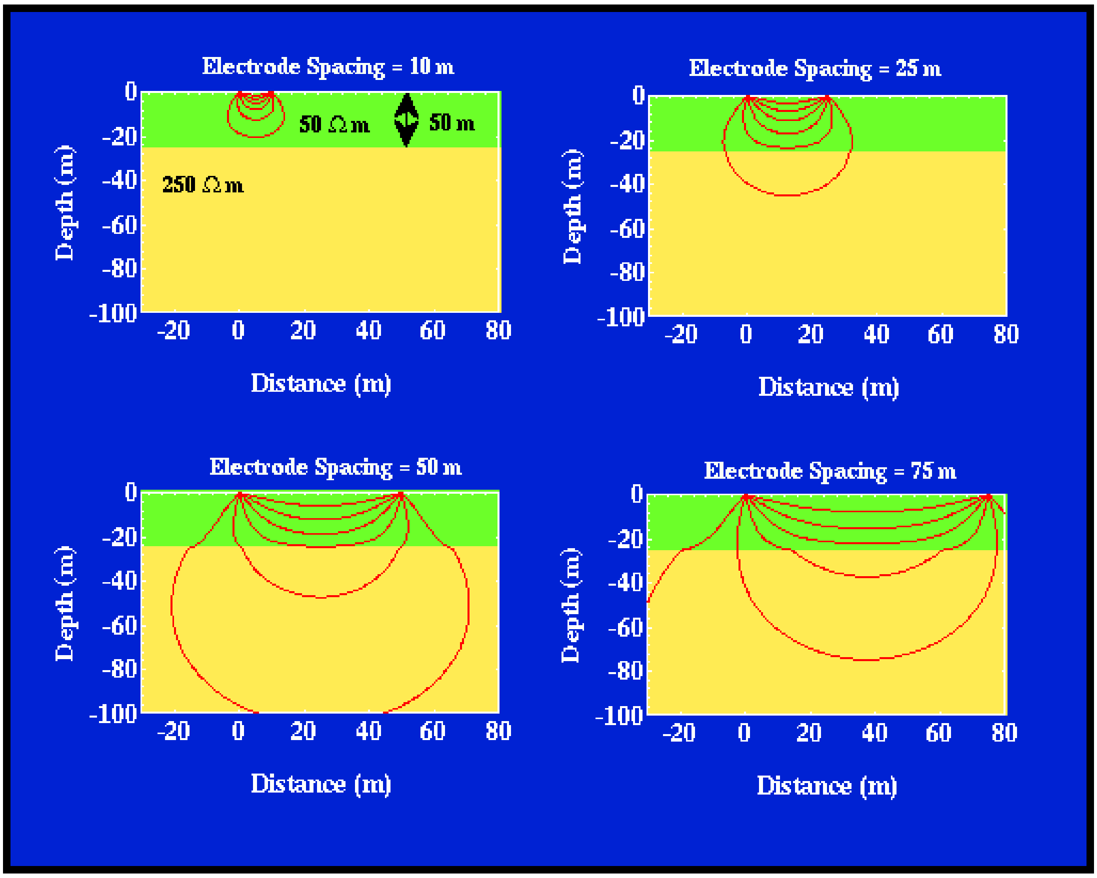
At very large current electrode spacings, underneath our potential electrodes, the pattern of current flow is again similar to that which we would observe in a homogeneous Earth model. In this case, however, the media has a resistivity of 50 ohm.m, not 250 ohm.m Thus, if we were to compute and plot apparent resistivity for a variety of current electrode spacings holding the potential electrodes fixed we would generate a plot similar to that shown below.



As is common for curves of this type, notice that this plot is a *Log-Log* plot. Instead of plotting apparent resistivity versus current electrode spacing, we have plotted the Log (base 10) of the apparent resistivity versus the Log (base 10) of the current electrode spacing. This is done because, in practice, we will find that both the apparent resistivities and the current electrode spacings will vary over two to three orders of magnitude (e.g., spacings can commonly increase from 0.25 m to 250 m). Using Log-Log plots provides us with a means of compressing the relevant information into a single graph. In the example shown above, notice that the apparent resistivity does not approach the resistivity of the lower layer until the electrode spacing approaches 500 m! Thus, large electrode spacings are required to see deep structure. A good rule of thumb is that you will need current electrode spacings on the order of 10 times the depth to which you would like to see.

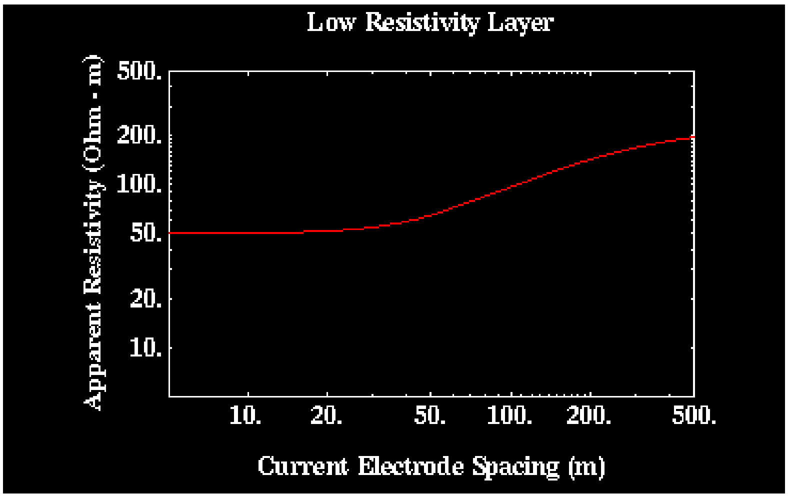
A Second Example of Current Flow in Layered Media

As another example of current flow in layered media and how apparent resistivity can vary with varying electrode spacing*, consider the earth model shown below. In this case, a low resistivity layer overlies a higher resistivity halfspace.



Initially with the current electrodes closely spaced, most of the current confined to the upper layer along paths that are very close to those that they would have assumed if the model were homogeneous. The computed apparent resistivity is very close to the resistivity of the upper layer, 50 ohm.m

At larger current electrode spacings, more current flows to greater depths. Between the two current electrodes, where are potential electrodes are located, the current flow lines become significantly distorted by the presence of the higher-resistivity layer located at depth. Therefore, around the potential electrodes the current density is larger than we would have observed in a homogeneous Earth. This relative increase in current density will cause our computed value of apparent resistivity to increase from 50 ohm.m. At very large current electrode spacings, current flow round our potential electrodes again approximate that we would observed in a homogeneous Earth. In this case, however, because most of the current is flowing through the lower media in the vicinity of the potential electrodes, the computed resistivity will be close to 250 ohm.m Thus, as current electrode spacing is increased the apparent resistivity will increase, eventually approaching 250 ohm.m A plot of apparent resistivity versus current electrode spacing is shown below.



Because current would prefer to flow within the first layer, notice that the apparent resistivity approaches the resistivity of the halfspace more slowly (i.e., with greater electrode spacings) than was found in the previous case.

*Although we have not explicitly said this, the relevant spacing in the phrase *electrode spacing* is not the spacing between the current electrodes or the spacing between the potential electrodes, but rather the spacing between the current and the potential electrodes. Thus, even if our current electrode spacing is large (so that most or our current is flowing through the lower medium), if our potential electrodes are close to the current electrodes we will compute apparent resistivities that are more like the resistivity of the upper layer than of the lower layer. In the previous example as well as in this example, we have explicitly assumed that the positions of the potential electrodes remain fixed throughout the experiment so that the distance between the potential and the current electrodes increases as the distance between the current electrodes increases. As the distance between current and potential electrodes increases, the depth over which we average resistivity structure in computing an apparent resistivity also increases.

Resistivity Equipment and Field Procedures

- Equipment
- Survey Types Overview: Soundings and Profiles
- Soundings: Wenner and Schlumberger
- Electrode Spacings and Apparent Resistivity Plots
- Advantages and Disadvantages of Each Survey Type
- Profiles

DC Resistivity Equipment

Compared to the equipment required for gravity surveying and magnetic surveying, that required for DC resistivity surveying . In fact, it is rather consisting of nothing more than a source of electrical current, an ammeter, a voltmeter, some cable, and electrodes. Given the nature of the that we are making, however, there are some considerations that must be given the equipment used to perform the measurements.

> *Current Source* - A source of DC current is required. In general, batteries are not capable of producing the DC currents required, so that if a pure DC source is used, it has to be produced by a portable electric generator. If, as is commonly done to eliminate the effects of electrode potentials and telluric currents a slowly varying AC current is used, portable, battery driven sources can be employed for DC resistivity surveys commonly used in engineering and environmental applications.

Ammeter - A simple ammeter (a device for measuring electrical current) can be used. The only constraint is that the meter be capable of measuring amperage from a few milliamps to about 0.5 amps. Many of the modern instruments are regulated such that the user determines the amperage input into the ground and the instrument attempts to deliver it. If the instrument can not deliver the specified amperage, usually because the electrode contact resistance is too high, the instrument warns the user.

Voltmeter - A simple voltmeter can also be used. To avoid problems with contact potential, a voltmeter with a very high impedance, above 500,000 Ohms, should be used. The voltmeter must also be capable of measuring voltages from a few millivolts to a few volts.

Electrodes - To avoid problems associated with electrode potentials, sophisticated electrodes known as porous pots can be used. But, because spurious electrode potentials can be mitigated through the use of a slowly varying AC source, these electrodes are not commonly used for DC resistivity measurements. If the conditions in the survey are extremely dry, and contact between the electrode and the ground can not be maintained, one might consider using porous pots.

For DC resistivity surveys, the most commonly used electrodes are nothing more than aluminum, copper, or steel rods about two feet in length. These rods are driven into the ground, and connected with cables to the current source or the voltmeter. Under dry conditions, contact between the rod and the ground can be enhanced by wetting the ground surrounding the electrode.

- *Cables* - To connect the electrodes to the various electrical components cables must be employed. These cables are typically nothing more than insulated wires with stranded, copper-cored conductors. Although long cable lengths may need to be employed, given the high resistivity of the ground, resistance in the cables is typically negligible. A more significant problem is current induction in the cables used to make the voltage measurement from the current flowing in the cables going to the current electrodes. This source of noise is easily avoidable by simply keeping the voltage cables at a distance (a few feet) from the current cables. For easy of deployment, cables are usually stored on reels.

Survey Types Overview: Soundings and Profiles

Thus far we have begun to see how geologically relevant structure can affect electrical current flow and measurements of voltage at the Earth's surface. We've described how depth variations in resistivity can be detected by increasing current electrode spacing by estimating apparent resistivities for various current electrode spacings. We have not, however, described the specific field procedures used in resistivity surveying.

Before describing these procedures, there is an important point to note about the geological structures considered thus far. Notice that the resistivity method represents the first method which can detect depth variations in a geologically relevant parameter. For example, if we conducted gravity or magnetic surveys at top structures that varied in density or magnetic susceptibility *only* in depth, we would observe no spatial variation in the Earth's gravity or magnetic fields. Thus, these methods are insensitive to changes in density and magnetic susceptibility that occur *solely* in depth.

***Resistivity Soundings* - As we've already shown,**

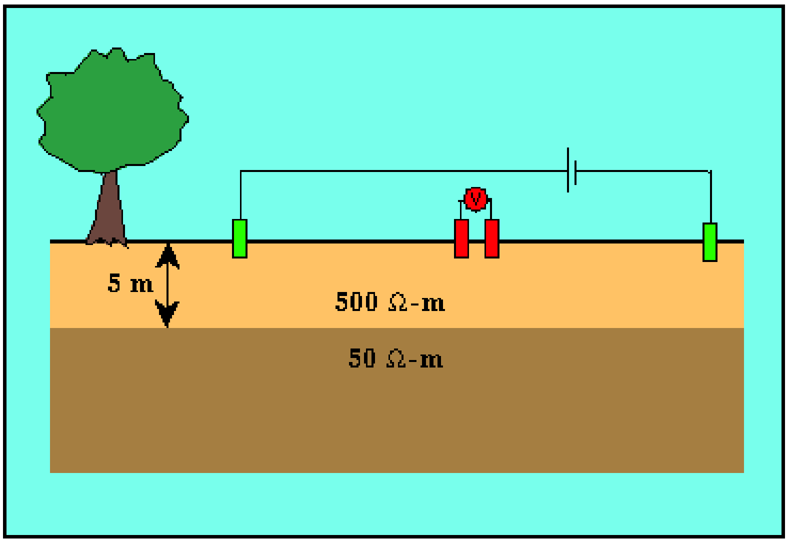
the resistivity method can detect variations in resistivity that occur solely in depth. In fact, this method is most commonly applied to look for variations in resistivity with depth. Surveys that are designed to determine resistivity variations with depth above some fixed surface location are referred to as *resistivity soundings*. In principle, the two-electrode experiments described previously are examples of soundings. In these experiments, electrode spacing is varied for each measurement. The center of the electrode array, where the electrical potential is measured, however, remains fixed. An example of a problem for which one might employ resistivity soundings is the determination of the depth to water table.

Resistivity Profiles - Like the gravity and magnetic methods, resistivity surveys can also be employed to detect lateral variations in resistivity. Unlike soundings, profiles employ fixed electrode spacings The center of the electrode spread is moved for each reading. These experiments thus provide estimates of the spatial variation in resistivity at some fixed electrode spacing. Surveys that are designed to locate lateral variations in resistivity are referred to as *resistivity profiles*. An example of a problem for which one might employ resistivity profiles is the location of a vertical fault.

Electrode Spacings and Apparent Resistivity Plots

You may have noticed on the previously shown plots of apparent resistivity that the data were plotted on *log-log* plots rather than the more traditional linear-linear plots. You should also notice that the electrode distances shown on these plots are evenly spaced in *log distance* rather than being evenly spaced in linear distance. Why have we chosen to acquire and display the data in this fashion?

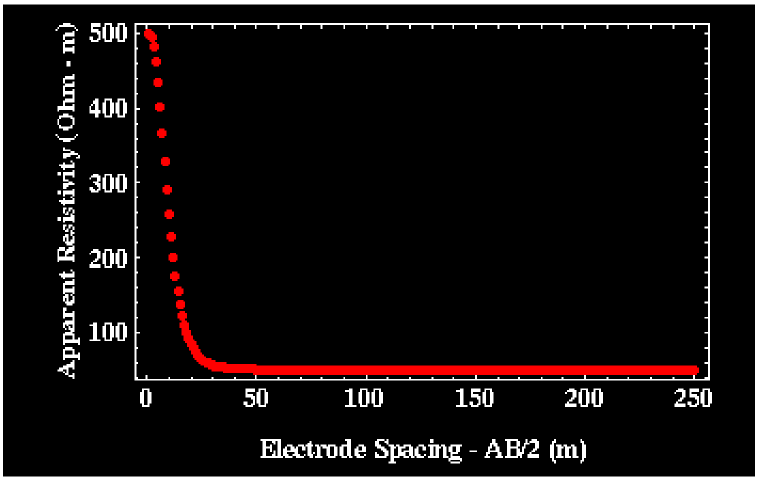
Consider performing a Schlumberger sounding over the geological model shown below.



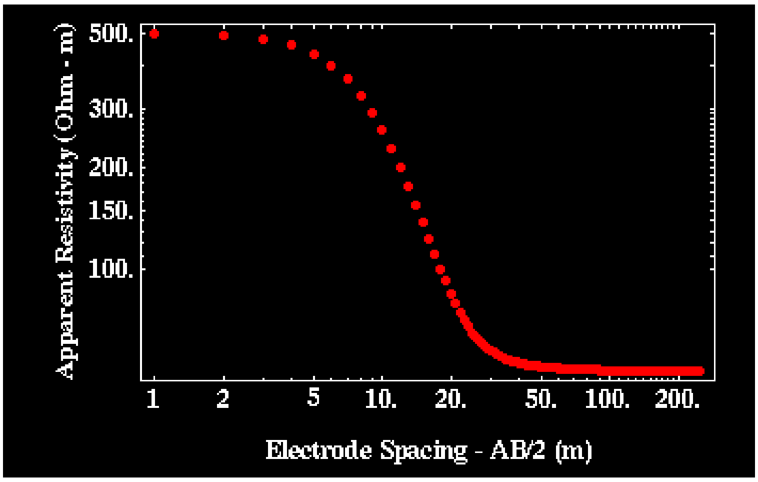
Let's do our Schlumberger sounding by varying current electrode spacing, *AB/2*, from 1 to 250 meters at 1 meter increments. Shown below is a plot of the resulting apparent resistivity versus electrode spacing.

We know that for small electrode spacings the apparent resistivity should approximate the resistivity of the layer. As the electrode spacing increases, the apparent resistivity should approach the resistivity of the halfspace.

These are the features that are shown in the plot. They are not, however, emphasized in this plot.

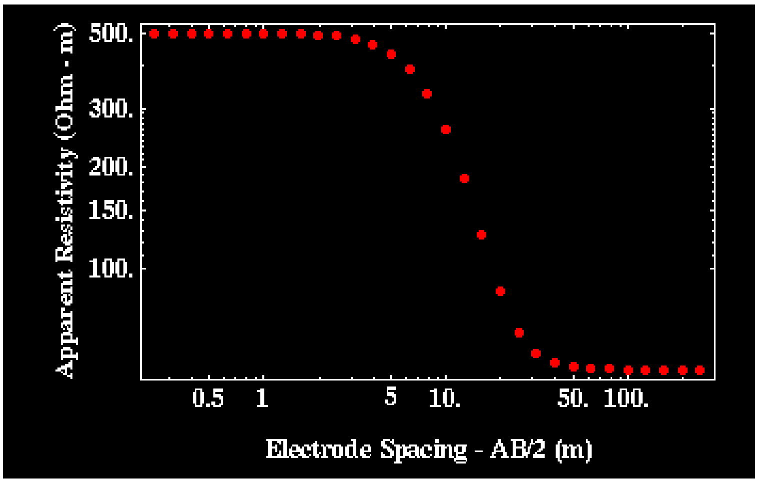


Most of the interesting features of this apparent resistivity curve occur at electrode spacings smaller than 50 meters. When looking at this apparent resistivity curve, because the plot includes so much data at electrode spacings larger than 50 meters, it de-emphasizes the important data at the smaller electrode spacings. One way to help bring out the information content at both the smaller and longer electrode spacings is to plot the same data on a log scale rather than a linear scale. A log-log plot the same data is shown below. Notice how the smaller electrode spacings now occupy more of the plot, thus making it easier to extract important information about how the apparent resistivity varies with electrode spacing.



Although this plot is better, there is still one problem related to how the data were acquired. Notice that there are only a few readings made at the small electrode spacings that are approximately equal to 500 ohm.m while there are many at the larger electrode spacings that are approximately equal to 50 ohm.m We would like more readings at the smaller electrode spacings so that we can be assured that the apparent resistivities plotted are representative of the near-surface resistivity. This could be done at the cost of taking fewer readings at the larger electrode spacings. By reallocating the electrode spacings that we will use, we will not only acquire observations that are more relevant, but we could also speed up our field acquisition by eliminating those readings that do not contain new information.

For electrical soundings, electrode spacings commonly are chosen so that they are evenly spaced in log distance rather than being evenly spaced in distance to address the problem described above. Shown below is a plot of log apparent resistivity versus log electrode spacing, where the distance interval is now chosen to be evenly spaced in log distance rather than distance. Now there are approximately as many samples showing apparent resistivities of 500 ohm.m as there are of 50 ohm.m In addition, the transition between these two extremes is also well sampled. Up to here

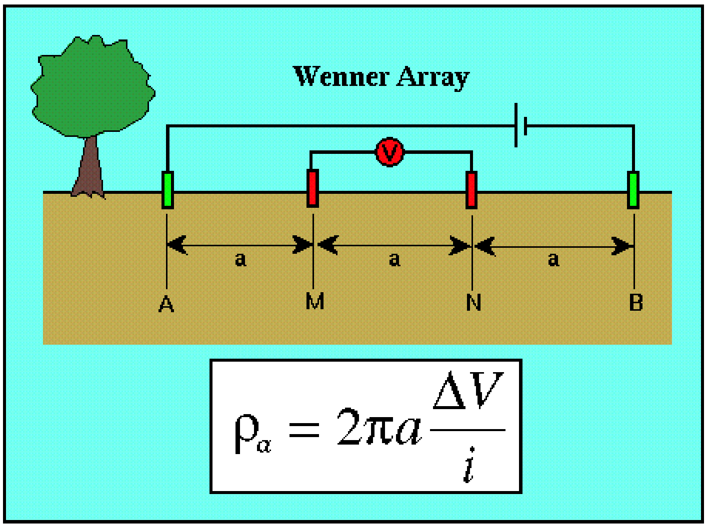


In the example shown above, we acquired the data so that there are 9 soundings for every decade (power of 10) in distance beginning at 0.25 meters. Thus, soundings were taken at 0.25, 0.5, 0.75, 1.0, 1.25, 1.5, 1.75, 2.0, 2.25, 5.0, 7.5, 10.0, 12.5, 15.0, 17.5, 20.0, 22.5, 50.0, 75.0, 100.0, 125.0, 150.0, 175.0, 200.0, 250.0 meter current electrode separations, *AB/2*.

This example clearly shows that using a log-distance scheme to acquire electrical data provides information at the densities required over all distance ranges. For most surveys, acquiring 9 readings per decade of distance is not necessary. The most common electrode spacing used is one that employs 6 soundings for every decade in distance. For this example, using six points per decade would yield electrode spacings of 0.25, 3.67, 5.39, 7.91, 1.16, 1.70, 2.5, 3.67, 5.39, 7.91, 11.6, 17.0, 25.0, 36.7, 53.9, 79.1, 116.0, 170.0, 250.0.

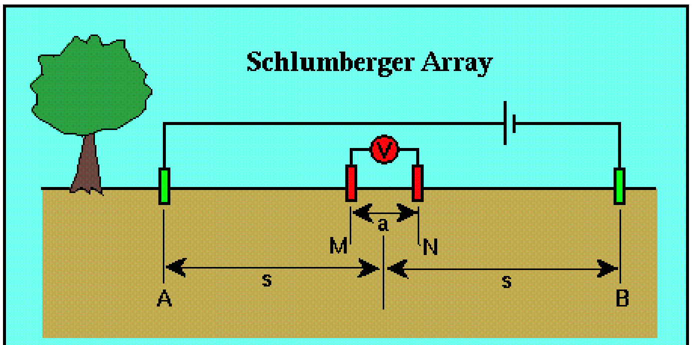
Resistivity Soundings

When doing resistivity sounding surveys, one of two survey types is most commonly used. For both of these surveys, electrodes are distributed along a line, centered about a midpoint that is considered the location of the sounding. The simplest in terms of the geometry of electrode placement is referred to as a *Wenner* survey. The most time-effective in terms of field work is referred to as a *Schlumberger* survey. For a Wenner survey, the two current electrodes (green) and the two potential electrodes (red) are placed in line with each other, equidistant from one another, and centered on some location as shown below.



The apparent resistivity computed from measurements of voltage, *l*, and current, *i*, is given by the relatively simple equation shown above. This equation is nothing more than the apparent resistivity expression shown previously with the electrode distances fixed to *a*. To generate a plot of apparent resistivity versus electrode spacing, from which we could interpret the resistivity variation with depth, we would have to compute apparent resistivity for a variety of electrode spacings, *a*. That is, after making a measurement we would have to move all four electrodes to new positions.

For a Schlumberger survey, the two current electrodes (green) and the two potential electrodes (red) are still placed in line with other, centered on some location, but the potential and current electrodes are not placed equidistant from one another.



$$\rho_a = \frac{\pi(s^2 - a^2/4)}{a}\frac{\Delta V}{i}$$

The current electrodes are at equal distances from the center of the sounding, *s*. The potential electrodes are also at equal distances from the center of the sounding, but this distance, *a/2*, is much less than the distance *s*. Most of the interpretation software available assumes that the potential electrode spacing is negligible compared to the current electrode spacing. In practice, this is usually interpreted as meaning that *a* must be less than *2s/5*.

In principle, this implies that we could set *a* to be less than *2s/5* for the smallest value of *s* we will use in the survey and never move the potential electrodes again. In practice, however, as the current electrodes are moved outward, the potential difference between the two potential electrodes gets smaller. Eventually this difference becomes smaller than our voltmeter is capable of reading, and we will need to increase *a* to increase the potential difference we are attempting to measure.

Advantages and Disadvantages of Wenner and Schlumberger Arrays

The following table lists some of the strengths and weaknesses of Schlumberger and Wenner sounding methods.

Schlumberger		Wenner	
Advantage	**Disadvantage**	**Advantage**	**Disadvantage**
Need to move the two potential electrodes only for most readings. This can significantly decrease the time required to acquire a sounding.			**All four electrodes, two current and two potential must be moved to acquire each reading.**
	Because the potential electrode spacing is small compared to the current electrode spacing, for large current electrode spacings very sensitive	**Potential electrode spacing increases as current electrode spacing increases. Less sensitive voltmeters are**	

	voltmeters are required.	**required.**	
Because the potential electrodes remain in fixed location, the effects of near-surface lateral variations in resistivity are reduced.			**Because all electrodes are moved for each reading, this method can be more susceptible to near-surface, lateral, variations in resistivity. These near-surface lateral variations could potentially be misinterpreted in terms of depth variations in resistivity.**
	In general, interpretations based on DC soundings will be limited to simple, horizontally layered structures.		**In general, interpretations based on DC soundings will be limited to simple, horizontally layered structures**

Seismic Methods

Introduction.

Seismic methods are the most commonly conducted geophysical surveys for engineering investigations. Seismic refraction provides engineers and geologists with the most basic of geologic data via simple procedures with common equipment.

Any mechanical vibration is initiated by a source and travels to the location where the vibration is noted. These vibrations are seismic waves. The vibration is merely a change in the stress state due to a disturbance. The vibration emanates in all directions that support displacement. The vibration readily passes from one medium to another and from solids to liquids or gasses and in reverse. A vacuum cannot support mechanical vibratory waves, while electromagnetic waves can be transmitted through a vacuum. The direction of travel is called the ray, ray vector, or ray path. Since a source produces motion in all directions the locus of first disturbances will form a spherical shell or wave front in a uniform material. There are two major classes of seismic waves: body waves, which pass through the volume of a material; and, surface waves, that exist only near a boundary.

Body waves. These are the fastest traveling of all seismic waves and are called compressional or pressure or primary wave (P-wave). The particle motion of P-waves is extension (dilation) and compression along the propagating direction. P-waves travel through all media that support seismic waves; air waves or noise in gasses, including the atmosphere. Compressional waves in fluids, e.g., water and air, are commonly referred to as acoustic waves.

The second wave type is the secondary or transverse or shear wave (S-wave). S-waves travel slightly slower than P-waves in solids. S-waves have particle motion perpendicular to the propagating direction, like the obvious movement of a rope as a displacement speeds along its length. These transverse waves can only transit material that has shear strength. S-waves therefore do not exist in liquids and gasses, as these media have no shear strength.

S-waves may be produced by a traction source or by conversion of P-waves at boundaries. The dominant particle displacement is vertical for SV-waves traveling in a horizontal plane. Dominant particle displacements are horizontal for SH-waves traveling in the vertical plane. SH-waves are often generated for S-wave refraction evaluations of engineering sites.

Elastic body waves passing through homogeneous, isotropic media have well-defined equations of motion. Most geophysical texts, including Grant and West (1965), include displacement potential and wave equations. Utilizing these equations, computations for the wave speed may be uniquely determined. Field surveys can readily obtain wave velocities, V_P and V_S; velocities are in units of length per time, usually meters/second (m/s). A homogeneous, isotropic medium's engineering properties of Young's or elastic modulus (E) and shear modulus (G) and either density (p_b) <u>or</u> Poisson's ratio (ν) can be determined if V_P and V_S are known. The units of these measures are: moduli in pressure, usually pascals (Pa); density in mass per volume, grams/cubic meter (g/m^3 = 10-6 mg/m^3); and Poisson's ratio, ν, is dimensionless. Manipulation of equations from Grant and West (1965) yields

$$v = \left[\frac{\left(V_p / V_s \right)^2 - 2}{\left\{ 2 \left[\left(V_p / V_s \right)^2 - 1 \right] \right\}} \right],$$

(1)

$$E = \frac{p_b V_p^2 (1-2v)(1+v)}{(1-v)},$$

(2)

$$G = \frac{E}{[2(1+v)]},$$

(3)

$$p_b = \frac{G}{V_s^2},$$

(4)

Note that these are not independent equa tions. Knowing two velocities uniquely determines only two unknowns of p_b, ν, or E. Shear modulus is dependent on two other values. Poisson's ratio, ν, varies from 0.0 to a value less than 0.5 from Equations 1 and 2. For units at the surface, p_b , the density, can be determined from samples or for the subsurface from bore hole samples or downhole logging. Estimates may be assumed for ν by material type. Usually the possible range of p_b is approximated, and ν is estimated. Equations 1 through 4 may be compared to the approximate values with some judgment applied. Table 1, below, provides some typical values selected from Hempen and Hatheway (1992) for V_P; Das (1994) for dryp_b of soils; Blake (1975) for $p_{b,dry}$ of rock; and Prakash (1981) for ν. Other estimates of p_b are contained in the section on gravity methods. Blake (1975) offers laboratory values of all these parameters, but field values will vary considerably from the lab estimates.

Surface waves. **Two recognized vubrations, which exist only at "surfaces" or interfaces, are Love and Rayleigh waves. Traveling along a surface, these waves attenuate rapidly with distance from the surface. Surface waves travel slower than body waves. Love waves travel along the surfaces of layered media, and are most often faster than Rayleigh waves. Love waves have particle displacement similar to SH-waves. A point in the path of a Rayleigh wave moves back, down, forward, and up repetitively in an ellipse like ocean waves.**

Surface waves are produced by surface impacts, explosions, and waveform changes at boundaries. Love and Rayleigh waves are also portions of the surface wave train in earthquakes. These surface waves may carry greater energy content than body waves. These wave types arrive last, following the body waves, but can produce larger

displacements in surface structures. Therefore, surface waves may cause more damage from earthquake vibrations.

Table 1. Typical/representative field values of V_p,P_b and ν for various materials.

Material	V_P (m/s)	$\rho_{b,dry}$ (mg/m³)	ν
Air	330		
Damp loam	300-750		
Dry sand	450-900	1.6-2.0	0.3-0.35
Clay	900-1,800	1.3-1.8	-0.5
Fresh, shallow water	1,430-1,490	1.0	
Saturated, loose sand	1,500		
Basal/lodgement till	1,700-2,300	2.3	
Rock			0.15-0.25
Weathered igneous sand metamorphic rock	450-3,700		
Weathered sedimentary Rock	600-3,000		
Shale	800-3,700		
Sandstone	2,200-4,000	1.9-2.7	
Metamorphic rock	2,400-6,000		
Unweathered basalt	2,600-4,300	2.2-3.0	
Dolostone and limestone	4,300-6,700	2.5-3.0	
Unweathered granite	4,800-6,700	2.6-3.1	
Steel	6,000		

Wave theory. **A seismic disturbance moves away from a source location; the locus of points defining the expanding disturbance is termed the wavefront. Any point on a wavefront acts as a new source and causes displacements in surrounding positions. The vector normal to the wavefront is the ray path through that point, and is the direction of propagation. Upon striking a boundary between differing material properties, wave energy is transmitted, reflected, and converted. The properties of the two media and the angle at which the incident ray path strikes will determine the amount of energy reflected off the surface, refracted into the adjoining material, lost as heat, and changed to other wave types.**

An S-wave in rock approaching a boundary of a lake will have an S-wave reflection, a P-wave reflection, and a likely P-wave refraction into the lake water (depending on the properties and incident angle). Since the rock-water boundary will displace, energy will pass into the lake, but the water cannot support an S-wave. The reflected S-wave departs from the boundary at the same angle normal to the boundary as the arriving S-wave struck. In the case of a P-wave incident on a boundary between two rock types (of differing elastic properties), there may be little conversion to S-waves. Snell's Law provides the angles of reflection and refraction for both the P- and S-waves. [Zoeppritz's equations provide the energy conversion for the body wave forms.] In the rock on the source side (No. 1), the velocities are V_{P1} and V_{S1}; the second rock material (No. 2) has properties of V_{P2} and V_{S2}. Then for the incident P-wave ($P_1 i$), Snell's Law provides the angles of reflections in rock No. 1 and refraction in rock No. 2 as:

$$\frac{\sin \alpha_{P1i}}{V_{P1i}} = \frac{\sin \alpha_{P1}}{V_{P1}} = \frac{\sin \alpha_{S1}}{V_{S1}} = \frac{\sin \alpha_{P2}}{V_{P2}} = \frac{\sin \alpha_{S2}}{V_{S2}}, \qquad (5)$$

The second and third terms of equation 45 are reflections within material No. 1; the fourth and fifth terms are refractions into medium No. 2. Note that none of the angles can exceed 90 degrees, since none of the sin terms can be over 1.0, and $\alpha_{P1i} = \alpha_{P1}$.

Two important considerations develop from understanding equation 5. First is the concept of critical refraction. If rock No. 1 has a lower velocity than rock No. 2 or $V_{P1} < V_{P2}$, then from Equation 5, sin $\alpha_{P2} >$ sin α_{P1i} and the refracted $\alpha_{P2} > \alpha_{P1i}$, the incident angle. Yet sin α_{P2} cannot exceed 1.00. The critical incident angle causes the refraction to occur right along the boundary at 90° from the normal to the surface. The critical angle is that particular incident angle such that sine α_{P2} = 1.0 and α_{P2} = 90 deg, or $\alpha(_{P1i})_{cr}$ = sin-1(V_{P1}/V_{P2}). Secondly, any incident angle > $\alpha(_{P1i})_{cr}$ from the normal will cause total reflection back into the source-

side material, since $\sin \alpha_{P2}$ Ý 1.0. For the latter case, all the P-wave energy will be retained in medium No. 1.

Other wave phenomena occur in the subsurface. Diffractions develop at the end of sharp boundaries. Scattering occurs due to inhomogeneities within the medium. As individual objects shrink in size, their effect on scatter is reduced. Objects with mean dimensions smaller than one- fourth of the wavelength will have little effect on the wave. Losses of energy or attenuation occur with distance of wave passage. Higher frequency waves lose energy more rapidly than waves of lower frequencies, in general.

The wave travels outward from the source in all directions supporting displacements. Energy dissipation is a function of the distance traveled, as the wave propagates away from the source. At boundaries, the disturbance passes into other media. If a wave can pass from a particular point A to another point B, Fermat's principle states that the ray path taken is the one requiring the minimum amount of time. In crossing boundaries of media with different properties, the path will not be the shortest distance (a straight line) due to refraction. The actual ray path will have the shortest travel time. Since every point on a wavefront is a new source, azimuths other than that of the fastest arrival will follow paths to other locations for the ever-expanding wave.

Data Acquisition

Digital electronics have continued to allow the production of better seismic equipment. Newer equipment is hardier, more productive, and able to store greater amounts of data. The choice of seismograph, sensors(geophones), storage medium, and source of the seismic wave depend on the survey being undertaken. The sophistication of the survey, in part, governs the choice of the equipment and the field crew size necessary to obtain the measurements. Costs rise as more elaborate equipment is used. However, there are efficiencies to be gained in proper choice of source, number of geophone emplacements for each line, crew size, channel

capacity of the seismograph, and requirements of the field in terrain type and cultural noise.

Sources. **The seismic source may be a hammer striking the ground or an aluminum plate or weighted plank, drop weights of varying sizes, rifle shot, a harmonic oscillator, waterborne mechanisms, or explosives. The energy disturbance for seismic work is most often called the "shot," an archaic term from petroleum seismic exploration. Reference to the "shot" does not necessarily mean an explosive or rifle source was used. The type of survey dictates some source parameters. Smaller mass, higher frequency sources are preferable. Higher frequencies give shorter wavelengths and more precision in choosing arrivals and estimating depths. Yet, sufficient energy needs to be transmitted to obtain a strong return at the end of the survey line. The type of source for a particular survey is usually known prior to going into the field. A geophysical contractor normally should be given latitude in selecting or changing the source necessary for the task. The client should not hesitate in placing limits on the contractor's indiscriminate use of some sources. In residential or industrial areas, perhaps the maximum explosive ge should be limited. The depth of drilling shot holes for explosives or rifle shots may need to be limited; contractors should be cautious not to exceed requirements of permits, utility easements, and contract agreements.**

Geophones. **The sensor receiving seismic energy is the geophone (hydrophone in waterborne surveys) or phone. These sensors are either accelerometers or velocity transducers, and convert ground movement into a voltage. Typically, the amplification of the ground is many orders of magnitude, but accomplished on a relative basis. The absolute value of particle acceleration cannot be determined, unless the geophones are calibrated.**

Most geophones have vertical, single-axis response to receive the incoming waveform from beneath the surface. Some geophones have horizontal-axis response for S-wave or surface wave assessments. Triaxial phones, capable of

measuring absolute response, are used in specialized surveys. Geophones are chosen for their frequency band response.

The line, spread, or string of phones may contain one to scores of sensors depending on the type of survey. The individual channel of recording normally will have a single phone. Multiple phones per channel may aid in reducing wind noise or air blast or in amplifying deep reflections.

Seismographs. **The equipment that records input geophone voltages in a timed sequence is the seismograph. Current practice uses seismographs that store the channels' signals as digital data at discrete time. Earlier seismographs would record directly to paper or photographic film. Stacking, inputting, and processing the vast volumes of data and archiving the information for the client virtually require digital seismographs. The seismograph system may be an elaborate amalgam of equipment to trigger or sense the source, digitize geophone signals, store multichannel data, and provide some level of processing display. Sophisticated seismograph equipment is not normally required for engineering and environmental surveys. One major exception is the equipment for sub-bottom surveys or nondestructive testing of pavements.**

Data processing of seismic information can be as simple as tabular equations for seismic refraction. Processing is normally the most substantial matter the geophysicists will resolve, except for the interpretation.

A portion of the seismic energy striking an interface between two differing materials will be reflected from the interface. The ratio of the reflected energy to incident energy is called the reflection coefficient. The reflection coefficient is defined in terms of the densities and seismic velocities of the two materials as:

$$R = \frac{(p_{b2}V_2 - p_{b1}V_1)}{(p_{b2}V_2 + p_{b1}V_1)}, \quad (6)$$

where

***R*= reflection coefficient,**

p_{b1}, p_{b2}= densities of the first and second layers, respectively,

V_1, V_2= seismic velocities of the first and second layers, respectively.

Modern reflection methods can ordinarily detect isolated interfaces whose reflection coefficients are as small as 0.02.

Seismic Reflection Methods

The physical process of reflection is illustrated in Figure 1, where the raypaths through successive layers are shown. There are commonly several layers beneath the earth's surface that contribute reflections to a single seismogram. The unique advantage of seismic reflection data is that it permits mapping of many horizon or layers with each shot.. At later times in the record, more noise is present in the record making the reflections difficult to extract from the unprocessed data.

Figure 2 indicates the paths of arrivals that would be recorded on a multichannel seismograph. Note that the subsurface coverage is exactly one-half of the surface distance across the geophone spread. The subsurface sampling interval is one-half of the distance between geophones on the surface. Another important feature of modern reflection-data acquisition is illustrated by figure 3. If multiple shots, S1 and S2, are recorded by multiple receivers, R1 and R2, and the geometry is as shown in the figure, the reflection point for both raypaths is the same. However, the ray paths are not the same length, thus the reflection will occur at different times on the two traces. This time delay, whose magnitude is indicative of the

subsurface velocities, is called normal-moveout. With an appropriate time shift, called the normal-moveout correction, the two traces (S1 to R2 and S2 to R1) can be summed, greatly enhancing the reflected energy and canceling spurious noise.

This method is called the common reflection point, common midpoint, or common depth point (CDP) method. If all receiver locations are used as shot points, the multiplicity of data on one subsurface point (called CDP fold) is equal to one-half of the number of recording channels. Thus, a 24-channel seismograph will record 12-fold data if a shot corresponding to every receiver position is shot into a full spread. Thus, for 12-fold data, every subsurface point will have 12 separate traces added, after appropriate time shifting, to represent that point.

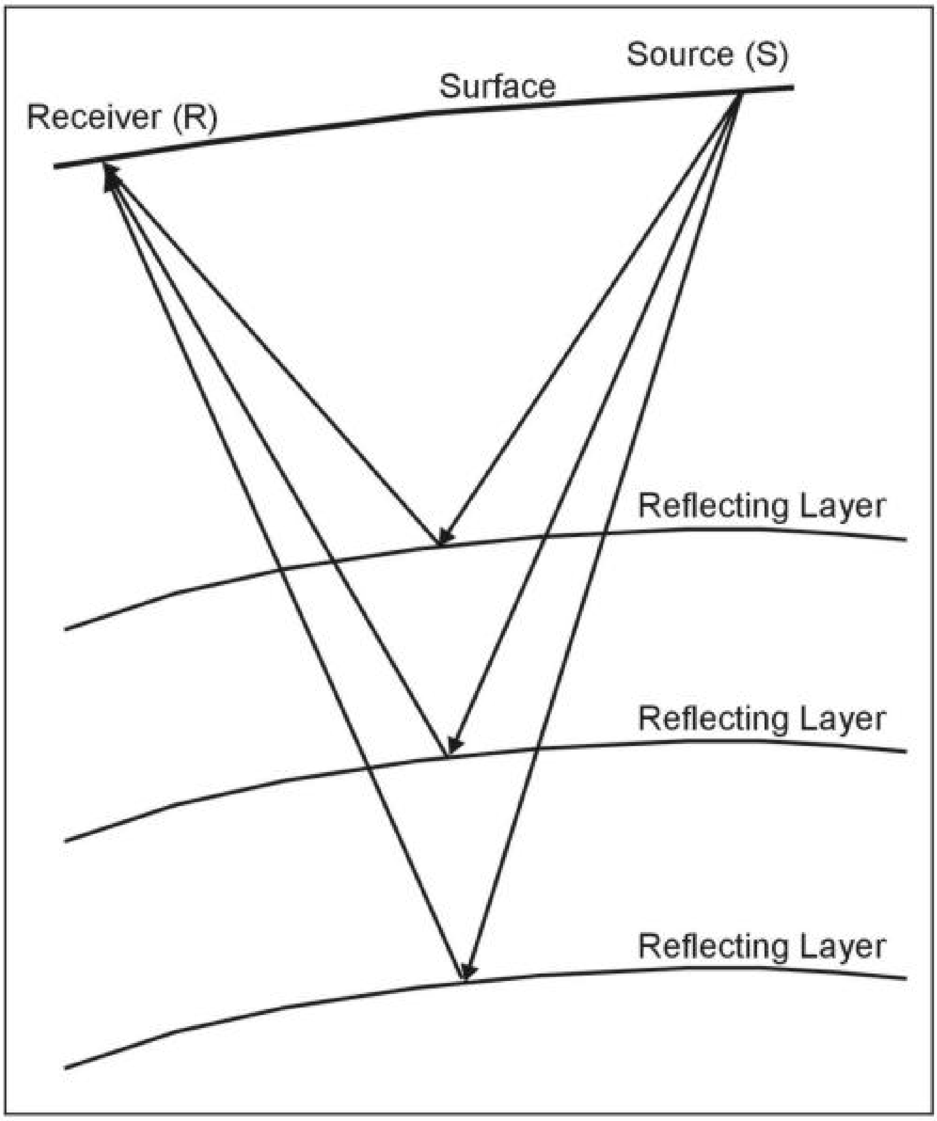


Figure 1. Schematic of the seismic reflection method.

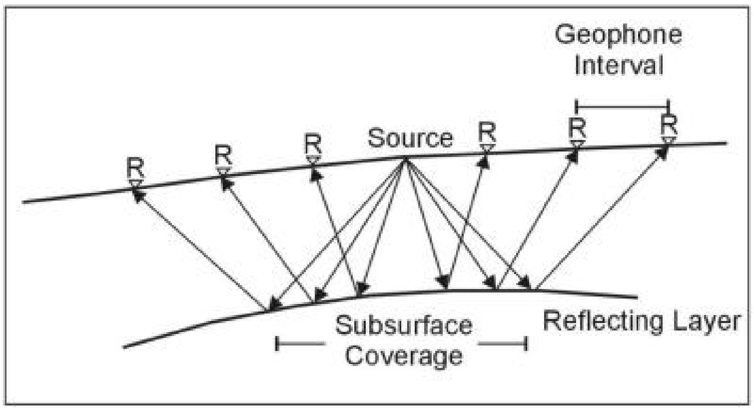


Figure 2. Multichannel recordings for seismic reflection.

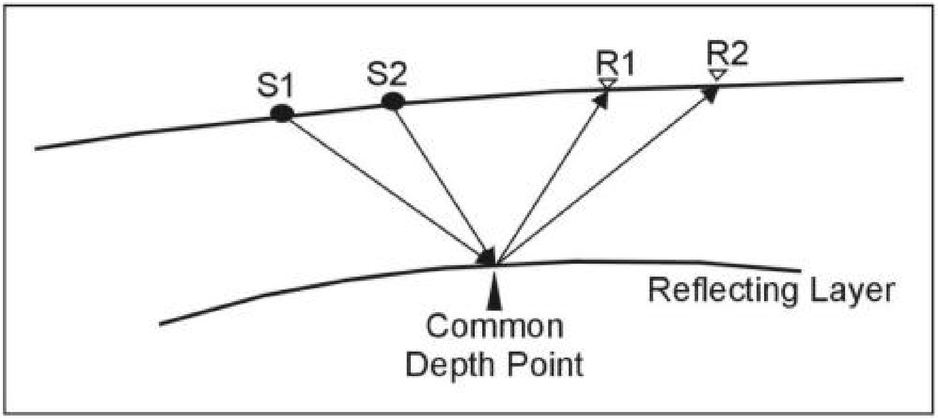


Figure 3. Illustration of common depth point (often called common mid point).

Arrivals on a seismic reflection record can be seen in figure 4. The receivers are arranged to one side of a shot, which is 15 m from the first geophone. Various arrivals are identified on figure 4. Note that the gain is increased down the trace to maintain the signals at about the same size by a process known as automatic gain control (AGC). One side of the traces is shaded to enhance the continuity between traces.

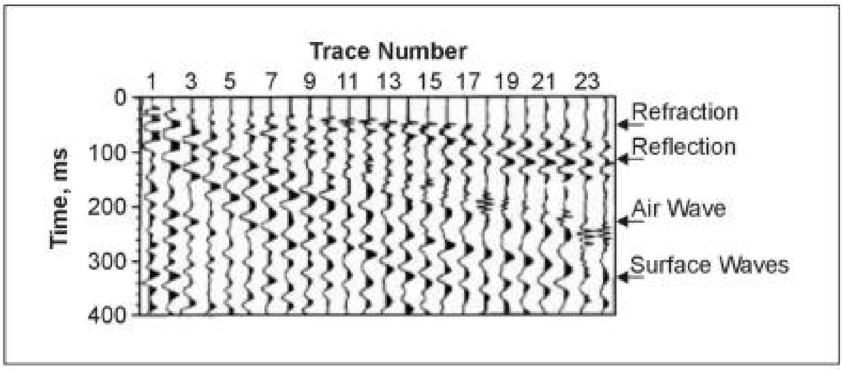


Figure 4. Simple seismic reflection record.

The ultimate product of a seismic reflection survey is a corrected cross section of the earth with reflection events in their true subsurface positions. This section does not present every detail of the acquisition and processing of shallow seismic reflection data. Thus, the difference between deep petroleum-oriented reflection and shallow reflection work suitable for engineering and environmental applications will be stressed.

Cost and frequency bandwidth are the principal differences between the two applications of seismic reflection. One measure of the nominal frequency content of a pulse is the inverse of the time between successive peaks. In the shallow subsurface, the exploration objectives are often at depths of 15 to 45 m. At 450 m/s, a wave with 10 ms peak-to-peak (nominal frequency of 100 Hz) is 45 m long. To detect (much less differentiate between) shallow, closely spaced layers, pulses with nominal frequencies at or above 200 Hz may be required. A value of 1,500 m/s is used as a representative velocity corresponding to saturated, unconsolidated materials because, without saturated sediments, both attenuation and lateral variability make reflection generally difficult.

Common-Offset Seismic Reflection Method

A technique for obtaining one-fold reflection data is called the common-offset method or common-offset gather (COG). It is instructive to review the method, but it has fallen into disuse because of the decreased cost of CDP surveys and the difficulty of quantitative interpretation in most cases. Figure 5 illustrates time-distance curves for the seismic waves that can be recorded. In the optimum offset distance range, the reflected and refracted arrivals will be isolated in time. Note that no quantitative scales are shown as the distances or velocities, and wave modes are distinct at each site. Thus, testing is necessary to establish the existence and location of the optimum offset window. Figure 6 illustrates the COG method. After the optimum offset distance is selected, the source and receiver are moved across the surface. Note that the subsurface coverage is one-fold, and there is no provision for noise cancellation. Figure 7 is a set of data presented as common offset data. The offset between geophone and shot is 14 m. Note that the acoustic wave (visible as an arrival near 40 ms) is attenuated (the shot was buried for this record). Note the prominent reflection near 225 ms that splits into two arrivals near line distance 610 m. Such qualitative changes are the usual interpretative result of a common offset survey. No depth scale is furnished.

Data Acquisition.

A shallow seismic reflection crew consists of three to five persons. The equipment used allows two to three times the number of active receivers to be distributed along the line. A switch (called a roll-along switch) allows the seismograph operator to select the particular set of geophones required for a particular shot from a much larger set of geophones that have been previously laid out. The operator can then switch the active array down the line as the position of the shot progresses. Often the time for a repeat cycle of the source and the archiving time of the seismograph are the determining factors in the production rates. With enough equipment, one or two persons can be continually moving equipment forward on the line while a shooter and an observer are sequencing through the available equipment.

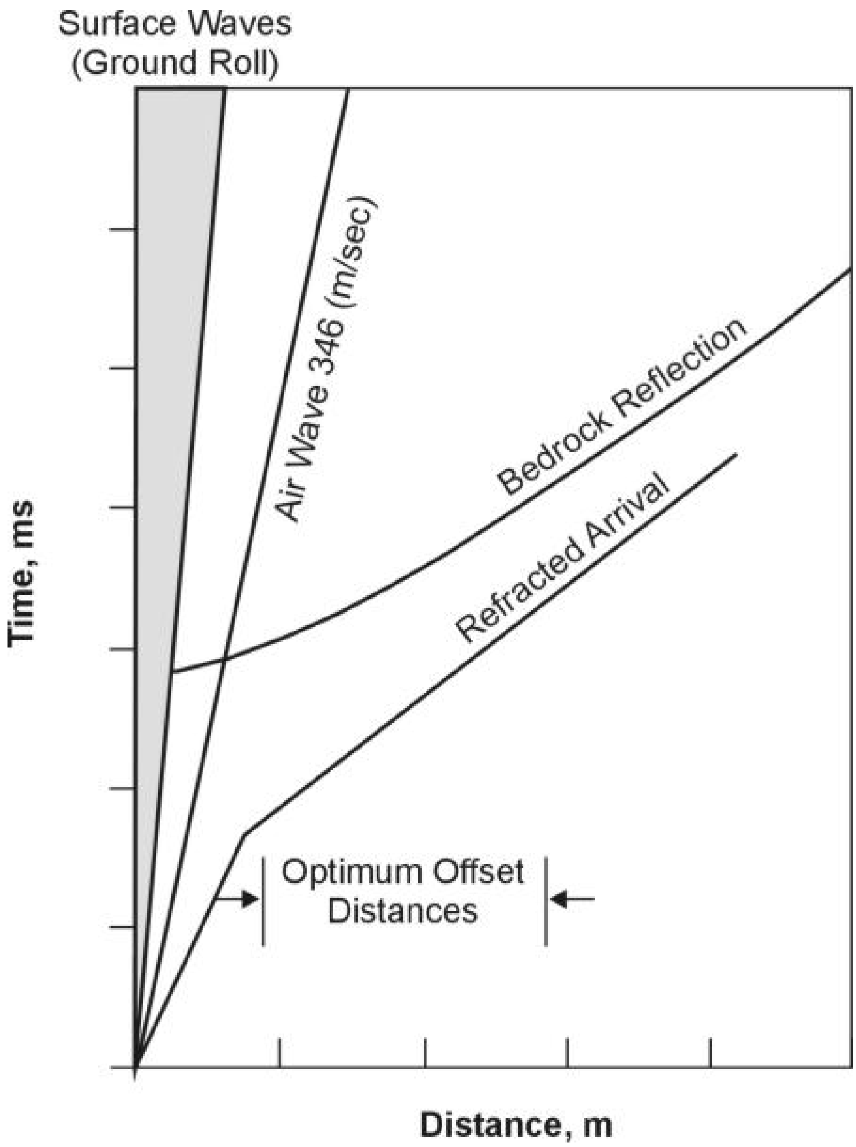


Figure 5. Optimum offset distance determination for the common offset method.

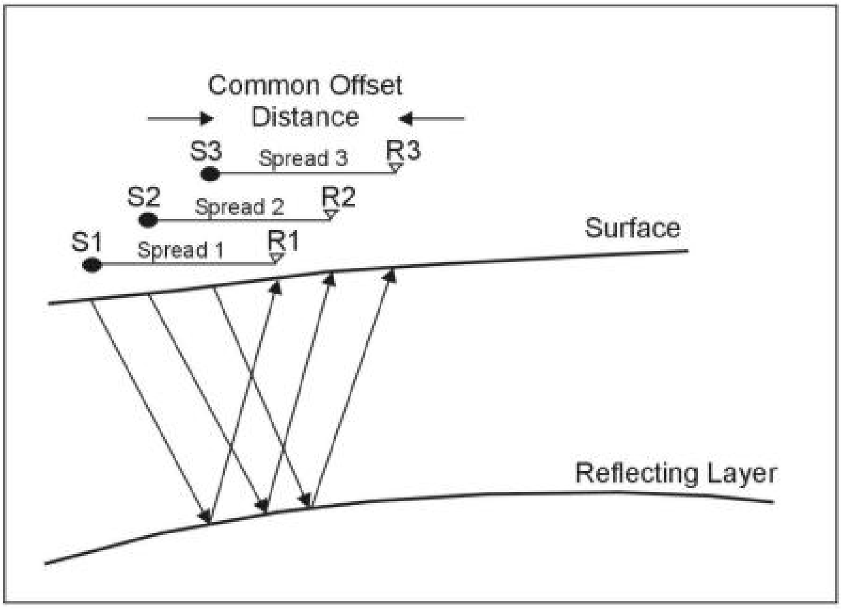


Figure 6. Common offset method schematic.

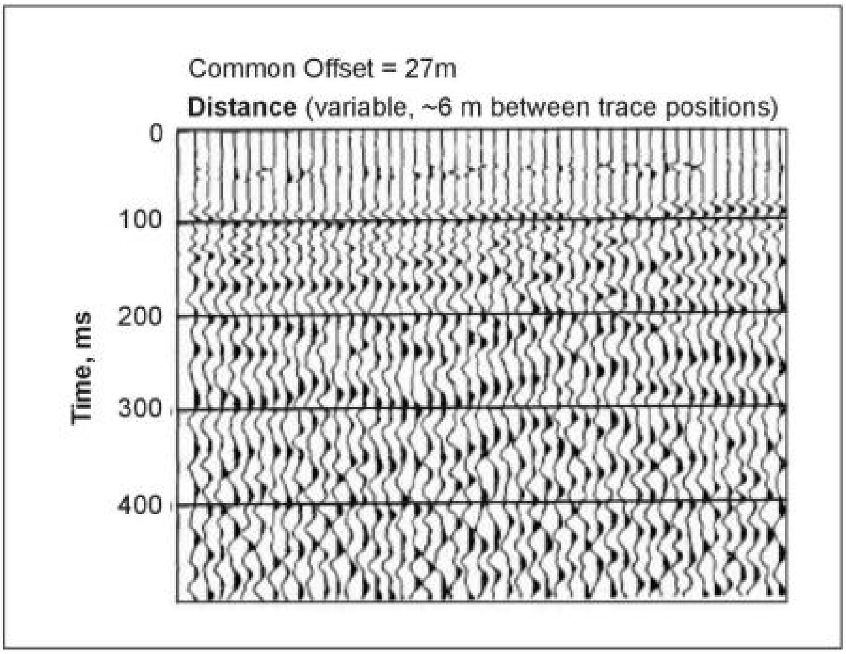


Figure 7. Sample common offset record.

If the requirements for relative and absolute surveying are taken care of at a separate time, excellent production rates, in terms of number of shot points per day, can be achieved. Rates of 1/min or 400 to 500 normal shots can be recorded in a field day. Note that the spacing of these shot points may be only 0.6 to 1.2 m, so the linear progress may be only

about 300 m of line for very shallow surveys. Also, note that the amount of data acquired is enormous. A 24-channel record sampled every 1/8 ms that is 200 ms long consists of nearly 60,000 32-bit numbers, or upwards of 240 KB/record. Three hundred records may represent more than 75 MB of data for 1 day of shooting.

Field data acquisition parameters are highly site specific. Up to a full day of testing with a knowledgeable consultant experienced in shallow seismic work may be required. The objective of these tests is identifiable, demonstrable reflections on the raw records. If arrivals consistent with reflections from the zone of interest cannot be seen, the chances that processing will recover useful data are slim. One useful testing technique is the walkaway noise test. A closely spaced set of receivers is set out with a geophone interval equal to 1% or 2% of the depth of interest - often as little as 30 or 60 cm for engineering applications. By firing shots at different distances from this spread, a well-sampled long-offset spread can be generated. Variables can include geophone arrays, shot patterns, high and low-cut filters, and AGC windows, among others.

Because one objective is to preserve frequency content, table 1 is offered as a comparison between petroleum-oriented and engineering-oriented data acquisition. The remarks column indicates the reason for the differences.

Table 1. Seismic reflection use differences by methodology.

	Petroleum	**Engineering**	**Remarks**
Explosive seismic source	**10-25 kg or more in a distributed pattern in deep holes**	**20 to 50 g, single shot**	**To increase frequency content**
Mechanical seismic source	**1-7 vibrators 5-15,000 kg peak force 10-100 Hz sweep**	**Hammer and Plates, guns**	**Cost, increased frequency**

Geophones	Arrays of 12-48 phones; 25-40 Hz fundamental frequency; 3-20 m spacing	Single or 3-5 geophones 50-100 Hz fundamental frequency; 1-3 m spacing	To preserve frequency content
Recorders	Instantaneous floating point, 48-1,000 channels	Instantaneous floating point, 24-96 channels	Cost
Passband analog filters	10-110 Hz	100-500 Hz	To increase frequency content
Sample interval	1-2 ms	1/4-1/8 ms	Higher frequencies

Data Processing

Processing is typically done by professionals using special purpose computers. These techniques are expensive but technically robust and excellent results can be achieved. A complete discussion of all the processing variables is well beyond the scope of this manual. However a close association of the geophysicist, the processor and the consumer is absolutely essential if the results are to be useful. Well logs, known depths, results from ancillary methods, and the expected results should be furnished to the processor. At least one iteration of the results should be used to ensure that the final outcome is successful.

One important conclusion of the processing is a true depth section. The production of depth sections requires conversion of the times of the reflections to depths by derivation of a velocity profile. Well logs and check shots are often necessary to confirm the accuracy of this conversion.

Advantages and Limitations

It is possible to obtain seismic reflections from very shallow depths, perhaps as shallow as 3 to 5 m.

1. **Variations in field techniques are required depending on depth.**

2. Containment of the air-blast is essential in shallow reflection work.

3. Success is greatly increased if shots and phones are near or in the saturated zone.

4. Severe low-cut filters and arrays of a small number (1-5) of geophones are required.

5. Generally, reflections should be visible on the field records after all recording parameters are optimized.

6. Data processing should be guided by the appearance of the field records and extreme care should be used not to stack refractions or other unwanted artifacts as reflections.

Subbottom Profiling

A variant of seismic reflection used at the surface of water bodies is subbottom profiling or imaging. The advantage of this technique is the ability to tow the seismic source on a sled or catamaran and to tow the line of hydrophones. This procedure makes rapid, continuous reflection soundings of the units below the bottom of the water body, in other words, the subbottom. This method and significant processing requirements have been recently developed by Ballard, et al., (1993) of the U.S. Army Engineer Waterways Experiment Station (WES). The equipment, acquisition, and processing system reduce the need for over-water boring programs. The developed WES imaging procedure resolves material type, density, and thickness (Ballard, et al., 1993).

Basic Concepts

The acoustic impedance method may be used to determine parameters of the soft aqueous materials. The acoustic impedance ***z for a unit is the product of its p_b and V_P. The reflection coefficient R from a particular horizon is***

$$R = \left(\frac{E_{ref}}{E_{inc}} \right)^{1/2} = \left[\frac{(z_i - z_j)}{(z_i + z_{j)}} \right], \tag{7}$$

where

E_{refl} = energy reflected at the i-j unit boundary

E_{inc} = incident energy at the i-j unit boundary

z_i = acoustic impedance of the i (lower) material

z_j = acoustic impedance of the j (upper) medium

At the highest boundary, the water-bottom interface, *z j* ,water is known to be 1.5 * 10 9 g/(m $_2$ s). Since the *E ref* l , 1-2 can be determined, and *E inc* , 1-2 and *z* $_1$,water are known, *z i* , $_2$ may be determined. *V P* ,$_2$ may be assessed from the depth of the 2-3 boundary, and thus ? *i* , $_2$ may be resolved. The material properties of lower units can be found in succession from the reflections of deeper layers.

Data Acquisition

A variety of different strength sources are available for waterborne use. By increasing strength, these sources are: pingers, boomers, sparkers, and airguns. Although there is some strength overlap among these sources, in general, as energy increases, the dominant period of the wave increases. For the larger source strength, therefore, the ability to resolve detail is impaired as period and wavelength become larger. The resolving accuracy of the system may change by more than an order of magnitude from <0.2 m for a pinger to >1.0 m for an airgun.

The conflicting impact of energy sources is the energy available for penetration and deeper reflections. The greater energy content and broad spectrum of the boomer allow significantly greater depth returns. Some near-bottom sediments contain organic material that readily absorbs energy. Higher energy sources may allow penetration of these materials.

Data collection is enormous with a towed subbottoming system. Graphic displays print real-time reflector returns to the hydrophone set. Recording systems retrieve the data for later processing. The field recorders graph time of source firing versus time of arrival returns. Figure 8 provides the field print for Oakland Harbor (Ballard, McGee, and Whalin, 1992).

Data Processing

Office processing of the field data determines the subbottoming properties empirically. The empiricisms are reduced when more sampling (boring) data are available to assess unit ρ and loss parameters for modeling. The processing imposes the Global Positioning System (GPS) locations upon the time of firing records to approximately locate the individual shot along the towed boat path. The seismic evaluation resolves the layer ***V_P and unit depths. From the firing surface locations and unit depths, the field graphs are correlated to tow path distance versus reflector depths. Figure 257 shows cross sections of the Gulfport Ship Channel, Mississippi. These are fence diagrams of depth and material types once all parallel and crossing surveys are resolved.***

Advantages and Limitations

The subbottoming technique can be applied to a large variety of water bodies. Saltwater harbors, shipping channels, and river waterways were the original objective for the technique. The method is now used on locks, dams, reservoirs, and engineering projects such as the location of pipelines.

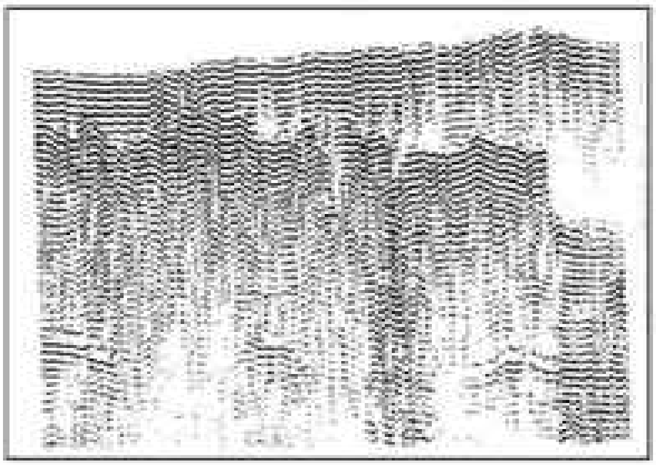


Figure 8. Reflected subbottoming signal amplitude cross section, 3.5 kHz in Oakland Harbor, California. (Ballard, McGee and Whalin, 1992)

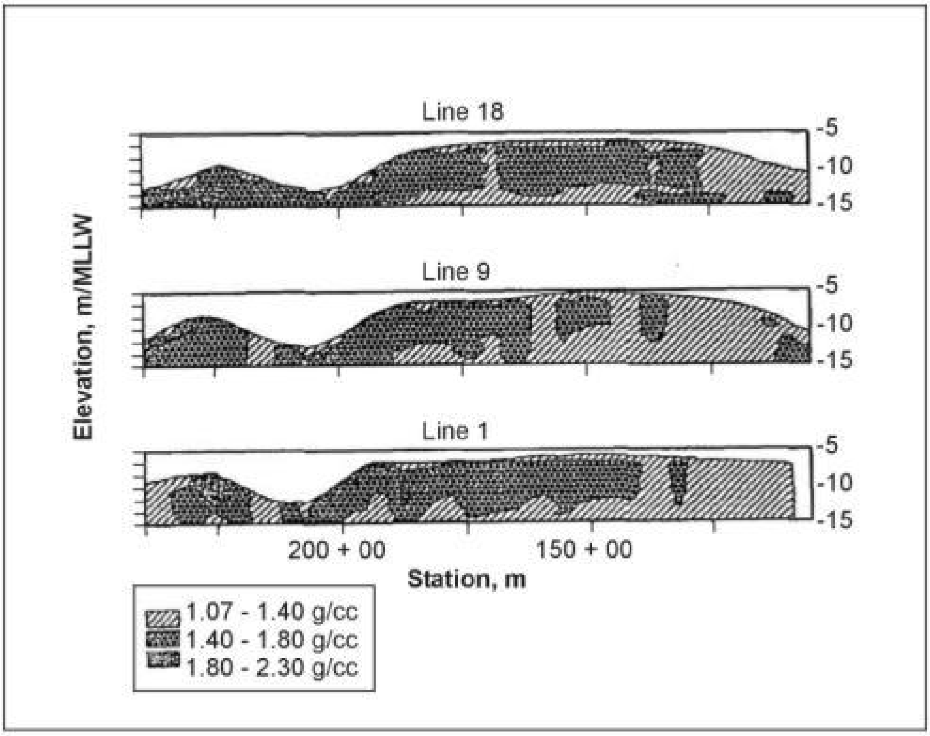


Figure 9. Density cross section in Gulfport ship canal, Mississippi. (Ballard. McGee and Whalin 1992)

Fathometer Surveys

Basic Concepts

Fathometers are also called Echo Sounders and are similar to reflection seismic profilers in that they also employ an acoustic source and receiver placed immediately beneath the surface of the water. However, fathometers differ from reflection seismic profilers in that they use higher frequency acoustic source pulses varying from less than 10 kHz up to about 200 kHz. Some of this energy transmitted by the source is reflected from the sediments at the water bottom, and the reflections are recorded by the receiver and stored digitally.

Data Acquisition

Fathometers determine water depth by repeatedly transmitting seismic energy through the water column and recording the arrival time of the reflected energy from the water bottom. The time required for the seismic signal to travel from its source to a reflector and back is known as the two-way travel time, and it is measured in milliseconds (ms) (equal to 1 x 10^{-3} seconds). The Fathometer calculates the depth to a water bottom by dividing the two-way travel time by two and multiplying the result by the velocity of sound through water:

$$D = V \tfrac{t}{2}, \tag{8}$$

where

D= depth to the water bottom (m),

V= velocity of sound through water (m/s), and

t= two-way travel time (s).

Fathometer surveys are conducted while traveling at a moderate speed in a boat. Typically, the transducer is mounted on the side of the boat and placed in the water. Data recording is essentially automatic with a t recorder plotting providing a hard copy of the data or a computer screen may be used for the display. The data may also be stored on magnetic tape for further processing and plotting. As with the CSP method, GPS can be used to position the data.

Data Processing

The depth value is printed as a continuous graphic profile and/or displayed as a numeric value by the Fathometer. Fathometers are calibrated by adjusting the value of V, which may vary slightly depending on water type. Most Fathometers use a narrow-bandwidth 200 kHz seismic signal. These Fathometers provide accurate depth data, but little or no information about the subbottom. Fathometers that use a lower frequency signal, such as 20 kHz, can detect reflected energy from subbottom interfaces, such as the bottom of an infilled scour hole.

Fathometer systems come with black and white t recording systems and in color systems. Colors are often assigned according to the different amplitudes of the reflected signals. Fathometers color step sizes as low as four dB are now available, allowing quite small changes in the reflected signals to be observed. An event marker button is often available allowing vertical line marks to be placed on the records when specific locations are selected by the operator. Sometimes the data can be downloaded to a computer allowing digital processing to be done along with data plots.

A color Fathometer can be calibrated to measure and display in color the amplitude of the reflected signal, which, in constant water depths, can be related to characteristics of the bottom material.

Data Interpretation

Traces from adjacent source/receiver locations are plotted side-by-side to form an essentially continuous time-depth profile of the stream bottom. Estimated seismic interval velocities can be used to transform the time-depth profile into a depth profile. However, water velocities are a function of suspended sediment load, and can vary appreciably.

The data are interpreted by viewing the plotted data. The response of specific objects may be used if these were noted on the records using a button marker.

Figure 10 shows Fathometer data recorded with a 200 kHz transducer. Note that only the water depth is observed in these data. Because of the high frequency, little energy is transmitted into the bottom sediments, and thus no reflections are observed from within the sediments.

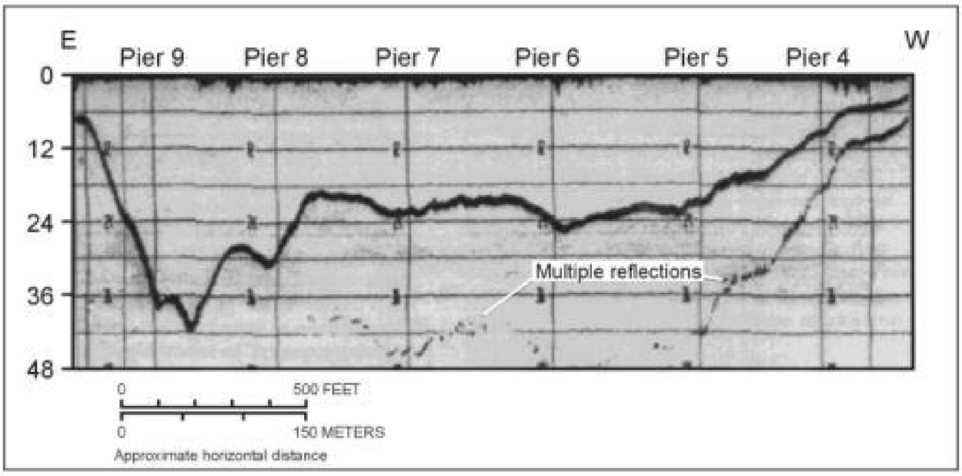


Figure 10. Fathometer data recorded with 200 kHz transducer. (Placzek, et al. 1995, USGS Report 95-4009)

Figure 11 presents fathometer data using a 3.5 kHz transducer. Because of the lower frequency, some of the energy is transmitted into the sediments and reflections are seen.

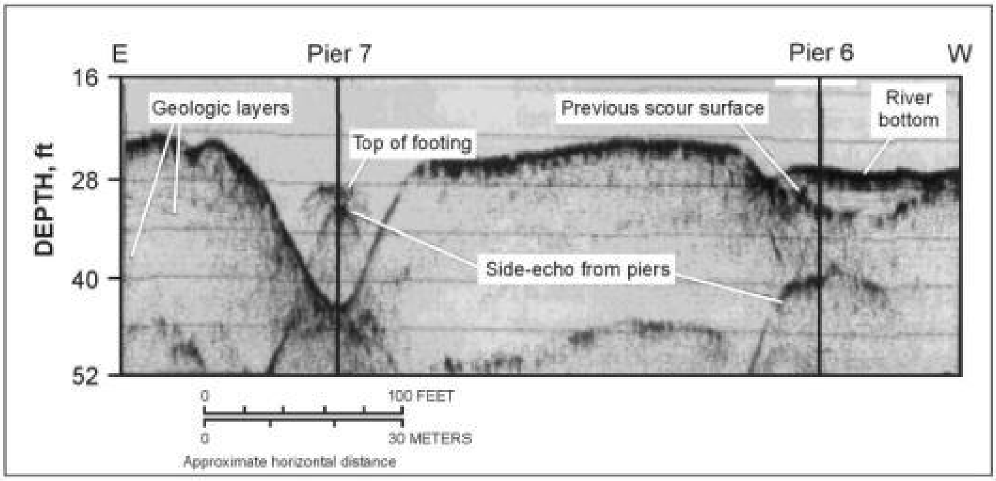


Figure 11. Fathometer data recorded using a 3.5 kHz transducer. (Placzek, et al. 1995, USGS Report 95-4009)

Advantages and Limitations

The main advantages of Fathometers (in continuous mode) are as follows:

1. The tool can provide an accurate depth-structure model of the water bottom (if acoustic velocities are known).

2. Post acquisition processing (migration) can be applied.

The main disadvantages of Fathometers (in continuous mode) are as follows:

1. The source and receiver need to be submerged. Profiles cannot be extended across emerged sand bars or onto the shore.

2. The equipment is relatively expensive (hardware and software).

3. Data may be contaminated by noise (multiple reflections, and echoes from the shoreline, water bottom, and/or piers).

4. Post acquisition processing (migration) may be required in areas where significant structural relief is present.

Fathometers are also employed in a spot survey mode. In this type of survey, sounding data (single reflection traces) are acquired at irregularly (or uniformly) spaced intervals (typically on the order of meters) at the water surface. The first high-amplitude reflected event is usually interpreted to be the water bottom reflection. Note, that spot data usually cannot be accurately migrated because of aliasing problems.

Surface Wave Methods

Introduction

A wide variety of seismic waves propagate along the surface of the earth. They are called surface waves because their amplitude decreases exponentially with increasing depth. The Rayleigh wave is important in engineering studies because of its simplicity and because of the close relationship of its velocity to the shear-wave velocity for earth materials. As most earth materials have Poisson's ratios in the range of 0.25 to 0.48, the approximation of Rayleigh wave velocities as shear-wave velocities causes less than a 10% error. Rayleigh wave studies for engineering purposes have most often been made in the past by direct observation of the Rayleigh wave velocities. One method consists of excitation of a monochromatic wave train and the direct observation of the travel time of this wave train between two points. As the frequency is known, the wavelength is determined by dividing the velocity by the frequency.

The assumption that the depth of investigation is equal to one-half of the wavelength can be used to generate a velocity profile with depth. This last assumption is somewhat supported by surface wave theory, but more modern and comprehensive methods are available for inversion of Rayleigh-wave observations. Similar data can be obtained from impulsive sources if the recording is made at sufficient distance such that the surface wave train has separated into its separate frequency components.

Spectral Analysis of Surface Waves (SASW)

The promise, both theoretical and observational, of surface wave methods has resulted in significant applications of technology to their exploitation. The problem is twofold:

To determine, as a function of frequency, the velocity of surface waves traveling along the surface (this curve, often presented as wavelength versus phase velocity, is called a dispersion curve).

From the dispersion curve, determine an earth structure that would exhibit such dispersion. This inversion, which is ordinarily done by forward modeling, has been automated with varying degrees of success.

Basic Concept

Spectral analysis, via the Fourier transform, can convert any time-domain function into its constituent frequencies. Cross-spectral analysis yields two valuable outputs from the simultaneous spectral analysis of two time functions. One output is the phase difference between the two time functions as a function of frequency. This phase difference spectrum can be converted to a time difference (as a function of frequency) by use of the relationship:

$$\Delta t(f) = \frac{\Phi(f)}{2\pi f}, \tag{1}$$

where

Δt(f)= frequency-dependent time difference,

Φ(f) = cross-spectral phase at frequency *f,*

***f*= frequency to which the time difference applies.**

If the two time functions analyzed are the seismic signals recorded at two geophones a distance *d* apart, then the velocity, as a function of frequency, is given by:

$$V(f) = \frac{d}{t(f)}, \tag{2}$$

where

***d* = distance between geophones,**

t(f)= term determined from the cross-spectral phase.

If the wavelength (λ) is required, it is given by:

$$\lambda(f) = \frac{V(f)}{f}, \tag{3}$$

As these mathematical operations are carried at for a variety of frequencies, an extensive dispersion curve is generated. The second output of the cross-spectral analysis that is useful in this work is the coherence function. This output measures the similarity of the two inputs as a function of frequency. Normalized to lie between 0 and 1, a coherency of greater than 0.9 is often required for effective phase difference estimates. Once the dispersion curve is in hand, the calculation-intensive inversion process can proceed. Although the assumption given above of depth equal to one-

half the wavelength may be adequate if relatively few data are available, the direct calculation of a sample dispersion curve from a layered model is necessary to account for the abundance of data that can be recorded by a modern seismic system. Whether or not the inversion is automated, the requirements for a good geophysical inversion should be followed, and more observations than parameters should be selected.

Calculation methods for the inversion are beyond the scope of this manual. The model used is a set of flat-lying layers made up of thicknesses and shear-wave velocities. More layers are typically used than are suspected to be present, and one useful iteration is to consolidate the model layers into a geologically consistent model and repeat the inversion for the velocities only.

The advantages of this method are:

a) **High frequencies (1-300 Hz) can be used, resulting in definition of very thin layers.**

b) **The refraction requirement of increasing velocities with depth is not present; thus, velocities that decrease with depth are detectable.**

By using both of these advantages, this method has been used to investigate pavement substrate strength. An example of typical data obtained by an SASW experiment is shown in figure 1. The scatter of these data is smaller than typical SASW data. Models obtained by two different inversion schemes are shown in figure 2 along with some crosshole data for comparison. Note that the agreement is excellent above 20 m of depth.

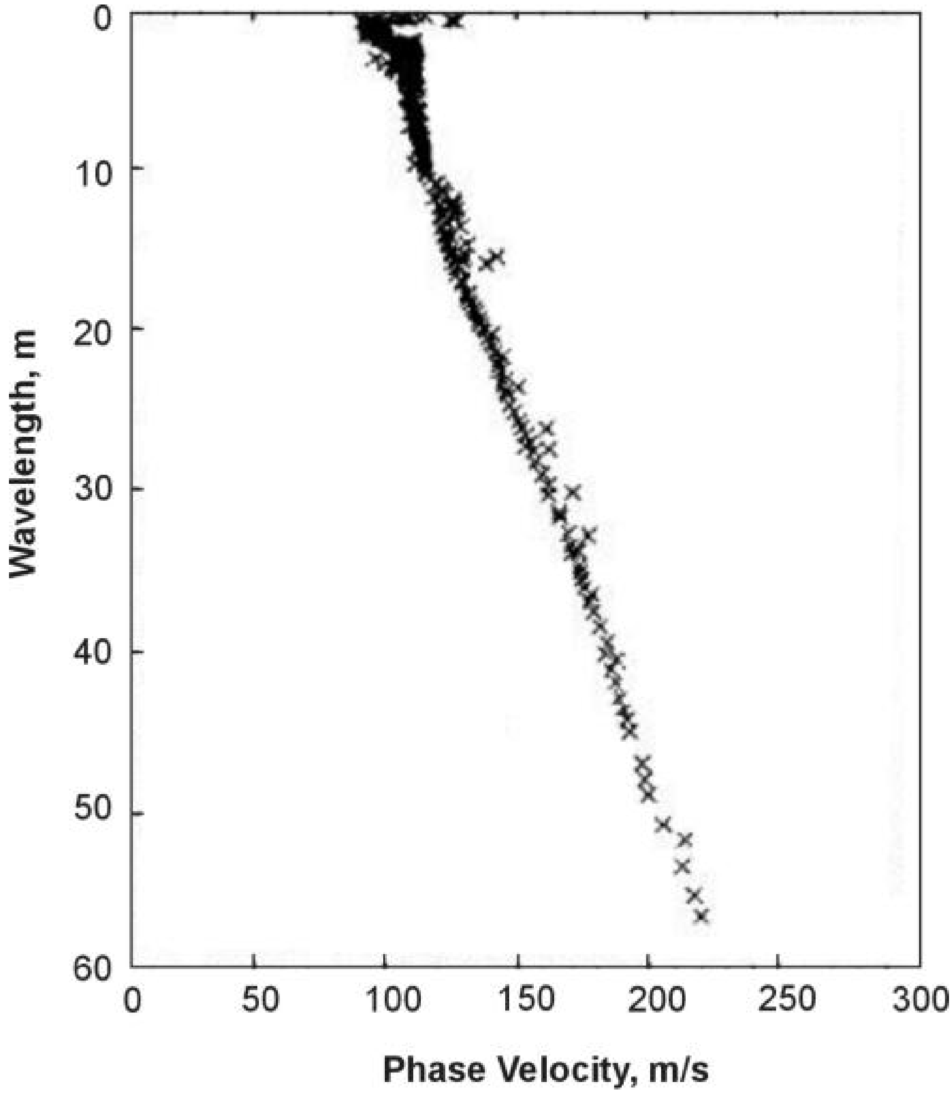


Figure 1. Typical Spectral Analysis of Surface Waves data.

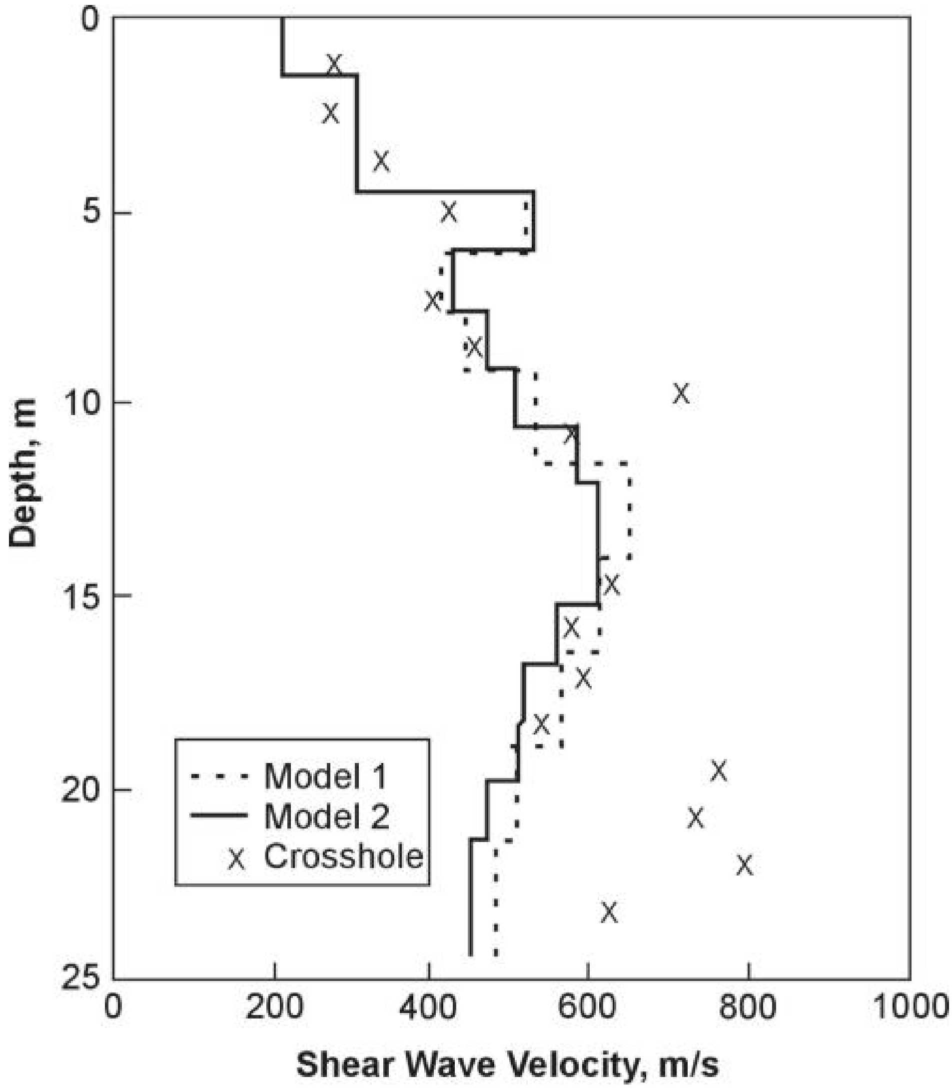


Figure 2. Inversion results of typical Spectral Analysis of Surface Waves data.

Data Acquisition

Most crews are equipped with a two- or four-channel spectrum analyzer, which provides the cross-spectral phase and coherence information. The degree of automation of the subsequent processing varies widely from laborious manual entry of the phase velocities into an analysis program to automated acquisition and preliminary processing. The inversion process similarly can be based on forward modeling with lots of human interaction or true inversion by computer after some manual smoothing.

A typical SASW crew consists of two persons, one to operate and coordinate the source and one to monitor the quality of the results. Typical field procedures are to place two (or four) geophones or accelerometers close together and to turn on the source. The source may be any mechanical source of high-frequency energy; moving bulldozers, dirt whackers, hammer blows, and vibrators have been used. Some discretion is advised as the source must operate for long periods of time, and the physics of what is happening are important. Rayleigh waves have predominantly vertical motion; thus, a source whose impedance is matched to the soil and whose energy is concentrated in the direction and frequency band of interest will be more successful.

Phase velocities are determined for waves with wavelength from 0.5 to 3 times the distance between the geophones. Then the phones are moved apart, usually increasing the separation by a factor of two. Thus, overlapping data are acquired, and the validity of the process is checked. This process continues until the wavelength being measured is equal to the required depth of investigation. Then the apparatus is moved to the next station where a sounding is required. After processing, a vertical profile of the shear-wave velocities is produced.

Advantages and Limitations

1. **The assumption of plane layers from the source to the recording point may not be accurate.**

2. **Higher modes of the Rayleigh wave may be recorded. The usual processing assumption is that the fundamental mode has been measured.**

3. **Spreading the geophones across a lateral inhomogeneity will produce complications beyond the scope of the method.**

4. **Very high frequencies may be difficult to generate and record.**

Common-Offset Rayleigh Wave Method

This method is also called Common-Offset Surface Waves. The method is quite effective at mapping inhomogeneities in the near surface, although it is not a frequently used method. The field techniques are easy to apply, and compared to other seismic methods, it is a rapid technique.

This method uses Rayleigh waves to detect fracture zones and associated voids. Rayleigh waves, also known as surface waves, have a particle motion that is counterclockwise with respect to the direction of travel. Figure 3 illustrates the particle motion for Rayleigh waves traveling in the positive X direction. In addition, the particle displacement is greatest at the ground surface, near the Rayleigh wave source, and decreases with depth. Three shot points are shown, labeled A, B, and C. The particle motion and displacement are shown for five depths under each shot point. For shot B, over the void, no Rayleigh waves are transmitted through the water/air filled void. This affects the measured Rayleigh wave recorded by the geophones over the void. Four parameters are usually observed. The first is an increase in the travel time of the Rayleigh wave as the fracture zone above the void is crossed. The second parameter is a decrease in the amplitude of the Rayleigh wave. The third parameter is reverberations (sometimes called ringing) as the void is crossed. The fourth parameter is a shift in the peak frequency toward lower frequencies. This is caused by trapped waves, similar to a tube wave in a borehole. The effective depth of penetration is approximately one-third to one-half of the wavelength of the Rayleigh wave.

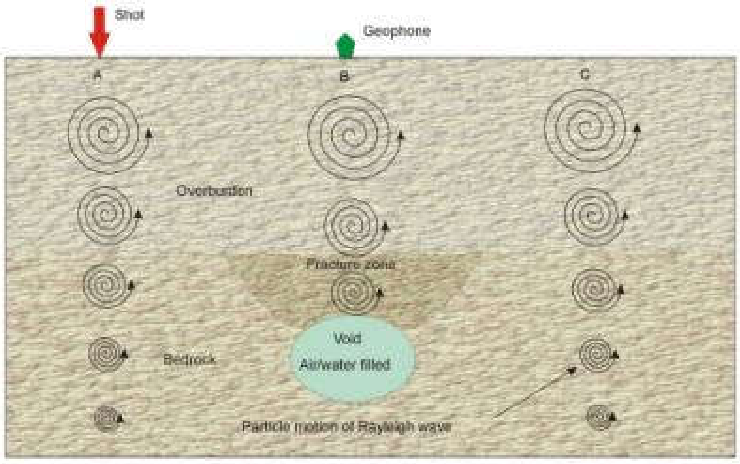


Figure 3. Rayleigh wave particle motion and displacement.

Rayleigh waves are created by any impact source. For shallow investigations, a hammer is all that is needed. Data are recorded using one geophone and one shot point. The distance between the shot and geophone depends on the depth of investigation and is usually about twice the expected target depth. Data are recorded at regular intervals across the traverse while maintaining the same shot geophone separation. The interval between stations depends on the expected size of the void/fracture zone and the desired resolution. Generally, in order to clearly see the void/fracture zone, it is desirable to have several stations that cross the area of interest. Figure 3 presents data from a common offset Rayleigh wave survey over a void/fracture zone in an alluvial basin. The geophone traces are drawn horizontally with the vertical axis being distance (shot stations).

The data may be filtered to highlight the Rayleigh wave frequencies and is then plotted as shown on figure 4. Because the peak frequency of the seismic signal decreases over a void/fracture zone, a spectral analysis of the traces can assist in the interpretation of the data by highlighting the traces with lower frequencies. The data shown in figure 4 illustrate many of the features expected over a void/fracture zone. The travel time to the first arrival of the Rayleigh wave is greater across the void/fracture and is wider than the actual fractured zone. The amplitudes of the

Rayleigh waves decrease as the zone is crossed. In addition, the wavelength of the signals over the fracture/void is longer than those over unfractured rock, showing that the high frequencies have been attenuated. Since the records are not long enough, the ringing effect is not presented in these data.

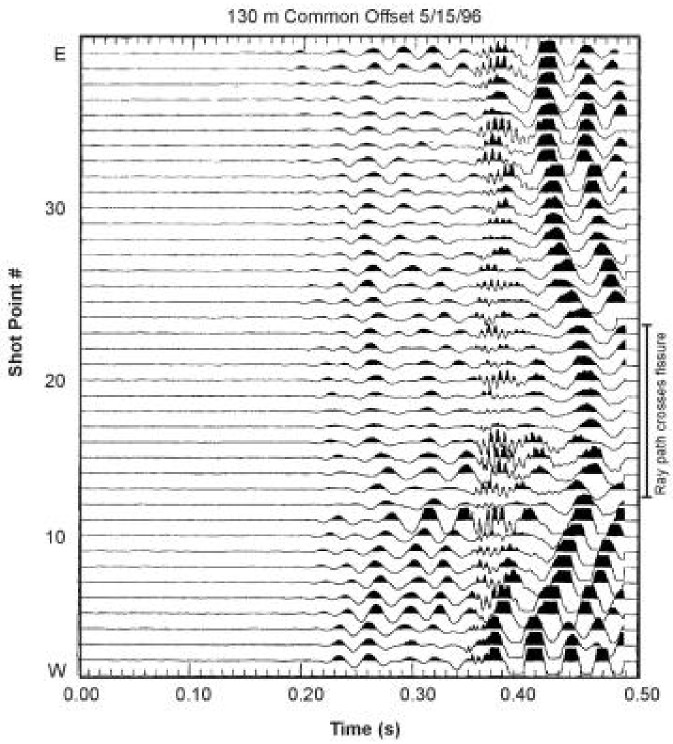


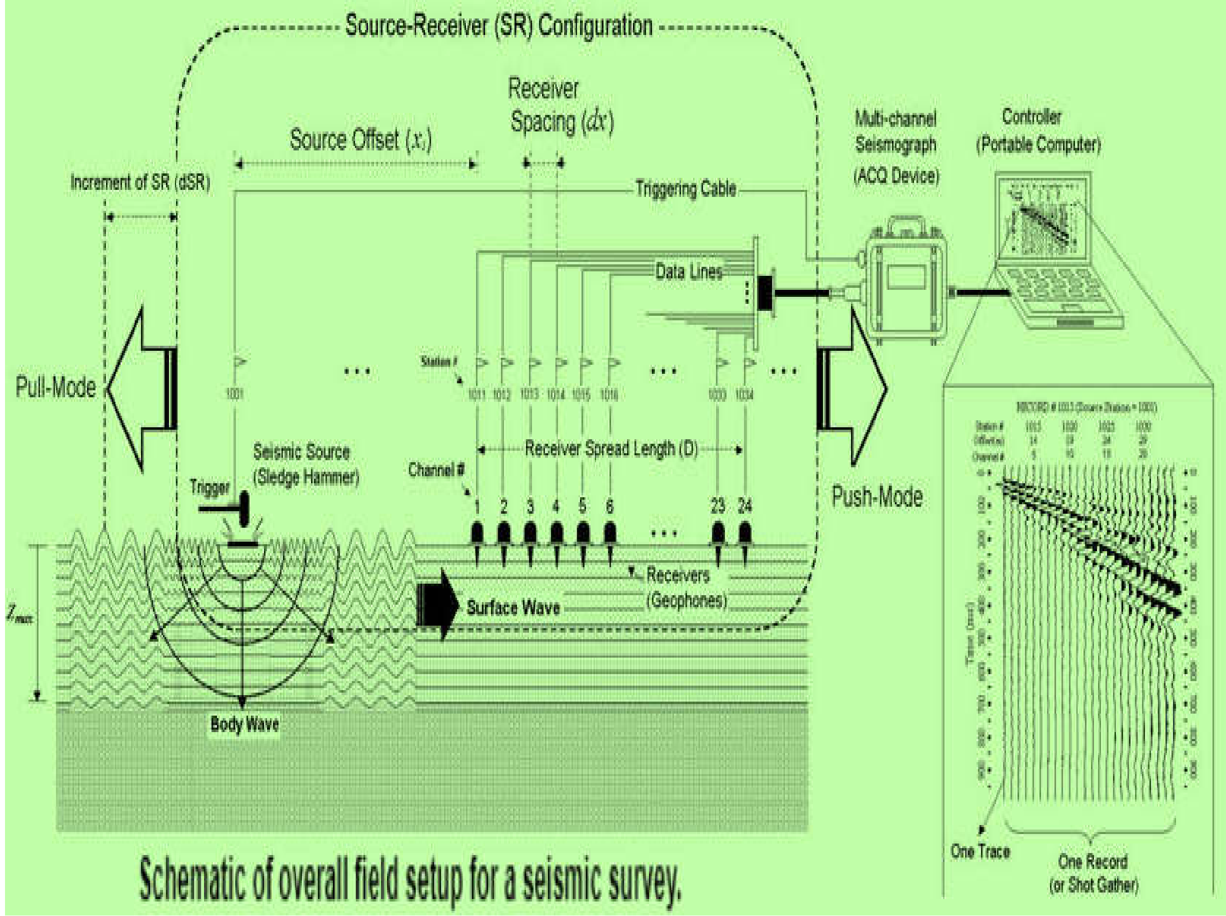
Figure 4. Data from a Rayleigh wave survey over a void/fracture zone.

Rayleigh waves are influenced by the shear strength of the rocks as well as fractures and voids, and changes in shear strength can cause anomalies similar to those obtained over these features. The depth of penetration and target resolution are influenced by the wavelengths generated by the seismic source. Since longer wavelengths, which have lower resolution, are needed to investigate to greater depths, fractures and voids at depth need to be increasingly larger in order to be observed. However, this method is faster and less costly than most other seismic methods.

Seismic

Pertaining to waves of elastic energy, such as that transmitted by P-waves and S-waves, in the frequency range of approximately 1 to 100 Hz. Seismic energy is studied by scientists to interpret the composition, fluid content, extent and geometry of rocks in the subsurface.

seismic data acquisition



The generation and recording of seismic data. Acquisition involves many different receiver configurations, including laying geophones or seismometers on the surface of the Earth or seafloor, towing hydrophones behind a marine seismic vessel, suspending hydrophones vertically in the sea or placing geophones in a wellbore (as in a vertical seismic profile) to record the seismic signal.

A source, such as a vibrator unit, dynamite shot, or an air gun, generates acoustic or elastic vibrations that travel into the Earth, pass through strata with different seismic responses and filtering effects, and return to the surface to be recorded as seismic data.

Optimal acquisition varies according to local conditions and involves employing the appropriate source (both type and intensity), optimal configuration of receivers, and orientation of receiver lines with respect to geological features. This ensures that the highest signal-to-noise ratio can be recorded, resolution is appropriate, and extraneous effects such as air waves, ground roll, multiples and diffractions can be minimized or distinguished, and removed through processing.

2D Seismic

- **2-D seismic is recorded using straight lines of receivers crossing the surface of the earth. A 2-D seismic survey works well for imaging major structures.**
- **2-D surveying is still popular because data gathering and analysis of 2D seismic information is much quicker and cheaper than 3D or 4D.**
- **2-D data requires much less permitting, surveying, and processing time than even small 3D surveys. Large 3D seismic shoots may take one to two years to acquire, and three to four months to process the information.**

3D Seismic

- **The late 1970s saw the development of the 3D seismic survey in which the data imaged was not just a vertical cross-section but an entire volume of earth.**
- **One of the most obvious differences between 2D and 3D seismic is that 3D imaging provides information continuously through the subsurface whereas 2D seismic reveals only strips of information.**
- **Nevertheless, 3D seismic may not be cost-effective in many onshore provinces, especially in the early stages of exploration.**
- **In onshore 3D seismic, many lines of receivers are used and recorded across the earth's surface. The area of receivers recorded is known as a "patch".**
- **In offshore, the main difference between 2D and 3D seismic is that 2D seismic is acquired using a single listening cable towed behind the seismic vessel, whereas 3D seismic is acquired using six parallel listening cables, and the cables can be up to six kilometers long.**
- **3D operations are considerably more elaborate than 2D and the daily cost of the crew is substantially increased.**

- **However, 3D seismic data collection improves exploration performance by allowing for:**
- **fewer dry holes**
- **more optimized well locations**
- **guidance for horizontal drilling projects**
- **more complete evaluation of mineral rights**
- **and a better understanding of the nature of the prospects**

4D Seismic

- **4D, or time-lapse, seismic is the process of using 3D seismic data acquired at different times, over the same area.**
- **It is used to assess changes in a producing hydrocarbon reservoir over time (the fourth dimension). Changes may be observed in fluid location and saturation, pressure and temperature.**
- **To maximize the value of a 4D seismic project, exploration and production assets are carefully screened because of the expense to acquire and analyze the data.**

Seismic data gather

A gather i.e. “A display of seismic traces that share an acquisition parameter ” can be of different types:

- Shotpoint gather
- Common source-receiver offset gather (COS)
- Common midpoint gather

Shotpoint gather

Shotpoint

One of a number of locations or stations at the surface of the Earth at which a seismic source is activated.

Seismic source

A device that provides energy for acquisition of seismic data, such as an air gun, explosive charge or vibrator.

Requirements for seismic sources:

- Produce enough energy in wide enough frequency band
- Energy focused for specific wave type (P or S)
- Repeatable source waveform
- Safe, efficient, environmentally acceptable

Seismic Sources (Land)

- Sledgehammer
- Weight drop

Air shooting

A method of seismic acquisition using charges detonated in the air or on poles above the ground as the source. Air shooting is also called the Poulter method after American geophysicist Thomas Poulter.

Explosive seismic data

Surface seismic data acquired using an explosive energy source, such as dynamite.

dynamite

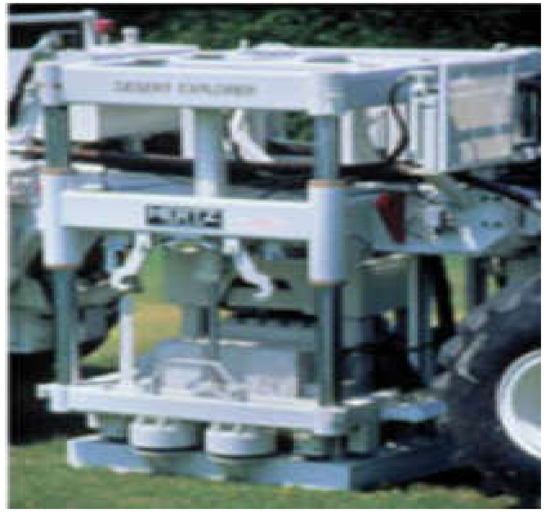
A type of explosive used as a source for seismic energy during data acquisition. Originally, dynamite referred specifically to a nitroglycerin-based explosive formulated in 1866 by Alfred Bernhard Nobel (1833 to 1896), the Swedish inventor who endowed the Nobel prizes. The term is incorrectly used to mean any explosive rather than the original formulation.

shot depth

- The location of an explosive seismic source below the surface.
- Before acquisition of surface seismic data onshore using explosive sources such as dynamite, holes are drilled at shotpoints and dynamite is placed in the holes. The shotholes can be more than 50 m [164 ft] deep, although depths of 6 to 30 m [20 to 98 ft] are most common and depth is selected according to local conditions.
- With other "surface" sources, such as vibrators and shots from air shooting, the shots occur at the Earth's surface.

Vibratory seismic data

- Seismic data whose energy source is a truck-mounted device called a vibrator that uses a vibrating plate to generate waves of seismic energy; also known as Vibroseis data (Vibroseis is a mark of Conoco).
- The frequency and duration of the energy can be controlled and varied according to the terrain and type of seismic data desired.
- The vibrator typically emits a linear "sweep" of at least seven seconds, beginning with high frequencies and decreasing with time ("downsweeping") or going from low to high frequency ("upsweeping").
- The frequency can also be changed in a nonlinear manner, such that certain frequencies are emitted longer than others. The resulting source wavelet is not impulsive.
- Vibrators are employed in land acquisition in areas where explosive sources cannot be used, and more than one vibrator can be used simultaneously to improve data quality.

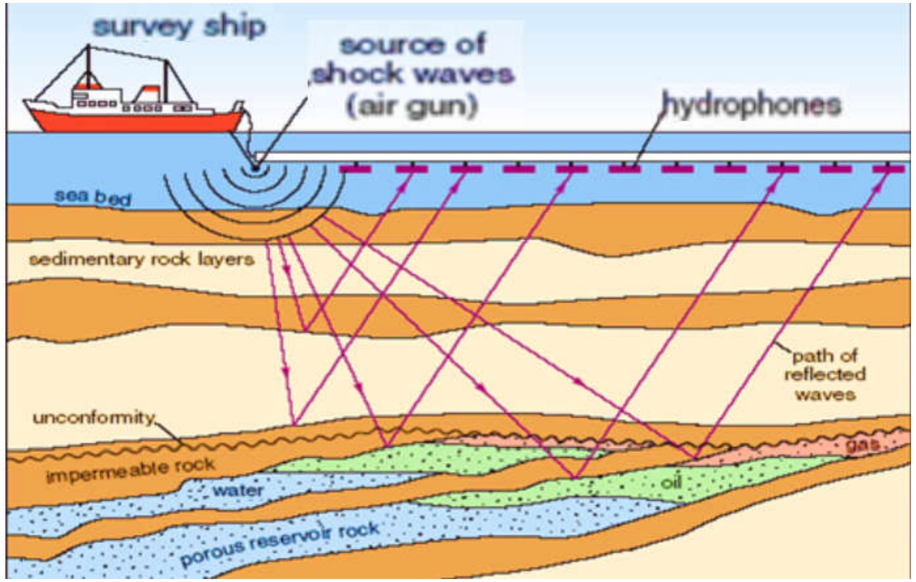


impulsive seismic data

Seismic data whose energy source is impulsive and of short duration, as with an air gun, rather than vibratory, as with a vibrator.

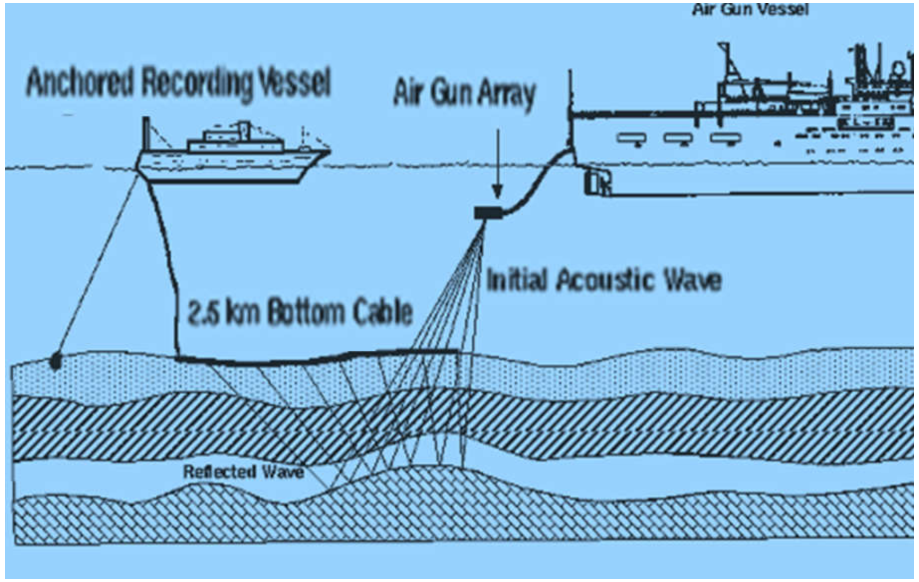
Marine Seismic Methods

- A variety of seismic sources are available for marine applications, including :
- water guns (20-1500 Hz), Air Gun (100-1500 Hz),Sparkers (50-4000 Hz), Boomers (300-3000 Hz), and Chirp Systems (500 Hz-12 kHz, 2-7 kHz, 4-24 kHz, 3.5 kHz, and 200 kHz).
- The greatest resolution of near surface structure is generally obtained from the higher frequency sources such as the Chirp systems, while the lower frequency tend to better characterize structure at depth.



water gun

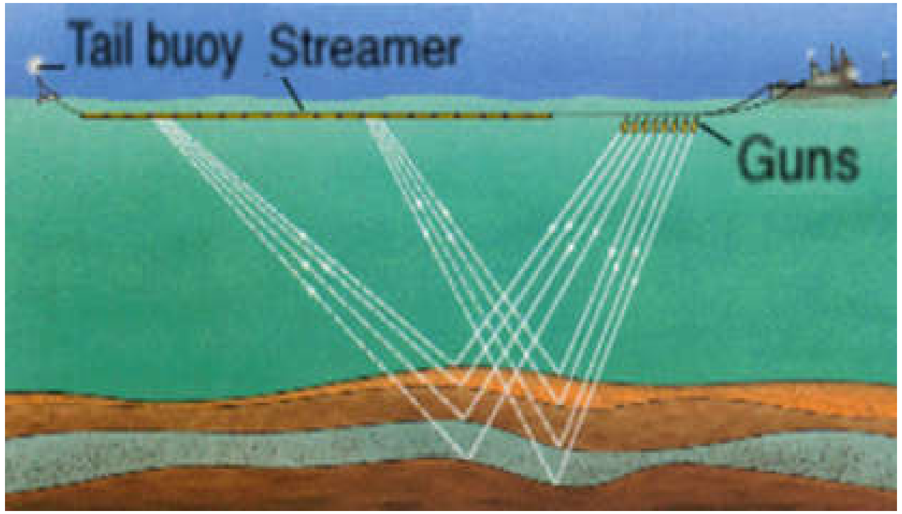
A source of energy for acquisition of marine seismic data that shoots water from a chamber in the tool into a larger body of water, creating cavitation. The cavity is a vacuum and implodes without creating secondary bubbles. This provides a short time Signature and higher resolution than an air-gun source.



Ocean-bottom cable

- Typically, an assembly of vertically oriented geophones and hydrophones connected by electrical wires and deployed on the seafloor to record and relay data to a seismic recording vessel.
- Such systems were originally introduced to enable surveying in areas of obstructions (such as production platforms) or shallow water inaccessible to ships towing seismic streamers (floating cables).
- Recent developments provide four component (4C) seabed systems to record shear wave (S-wave) as well as P-wave energy.

Marine Seismic Methods



- Marine seismic vessels are typically about 75 m [246 ft] long and travel about 5 knots [9.3 km/hr or 5.75 statute miles/hr] while towing arrays of air guns and streamers containing hydrophones a few meters below the surface of the water.
- The tail buoy helps the crew locate the end of the streamers.
- The air guns are activated periodically, such as every 25 m (about 10 seconds), and the resulting sound wave travels into the Earth, is reflected back by the underlying rock layers to hydrophones on the streamer and then relayed to the recording vessel.

hydrophone

A device designed for use in detecting seismic energy in the form of pressure changes under water during marine seismic acquisition. Hydrophones are combined to form streamers that are towed by seismic vessels or deployed in a borehole. Geophones, unlike hydrophones, detect motion rather than pressure.

streamer

- A surface marine cable, usually a buoyant assembly of electrical wires that connects hydrophones and relays seismic data to the recording seismic vessel.
- Multistreamer vessels tow more than one streamer cable to increase the amount of data acquired in one pass.

depth controller

A device used in acquisition of marine seismic data that keeps streamers at a certain depth in the water.

Seismic sensors

Detector

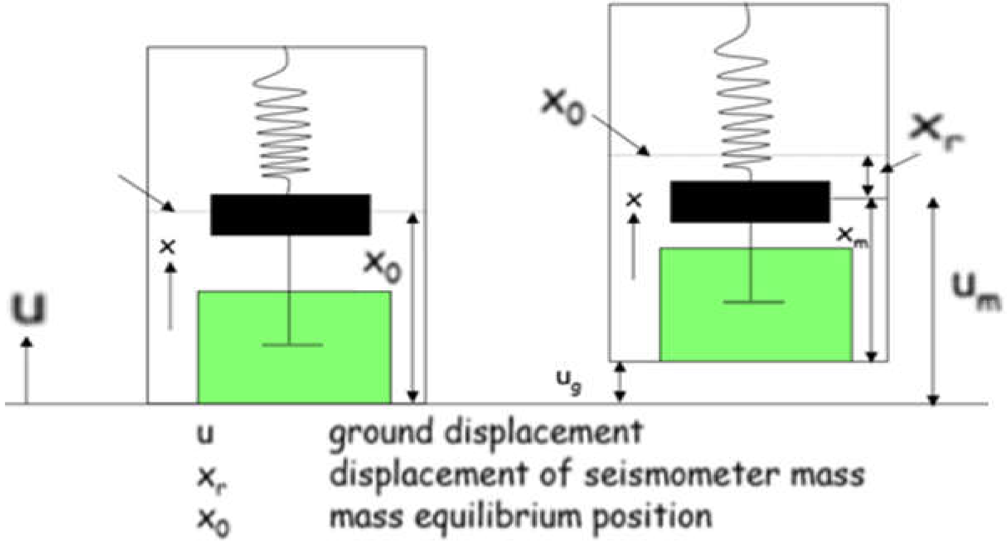
A sensor or receiver, such as a geophone or hydrophone

A device used in surface seismic acquisition, both onshore and on the seabed offshore, that detects ground velocity produced by seismic waves and transforms the motion into electrical impulses.
Geophones detect motion in only one direction.
Conventional seismic surveys on land use one geophone per receiver location to detect motion in the vertical direction.
Three mutually orthogonal geophones are typically used in combination to collect 3C seismic data.

Jug: Archaic slang for a geophone.

Seismometer

A device that records seismic energy in the form of ground motion and transforms it to an electrical impulse.



Seismogram

Traces recorded from a single shot point. Numerous seismograms are displayed together in a single seismic section.

ground roll

- A type of coherent noise generated by a surface wave, typically a low-velocity, low-frequency, high-amplitude Rayleigh wave.
- Ground roll can obscure signal and degrade overall data quality, but can be alleviated through careful selection of source and geophone arrays, filters and stacking parameters.

water-bottom roll

- The marine equivalent of ground roll.
- Water-bottom roll consists of a pseudo-Rayleigh wave traveling along the interface of the water and the seafloor.
- As the use of seabed receiver systems increases, noise from water-bottom roll has become more of a concern.

offset data

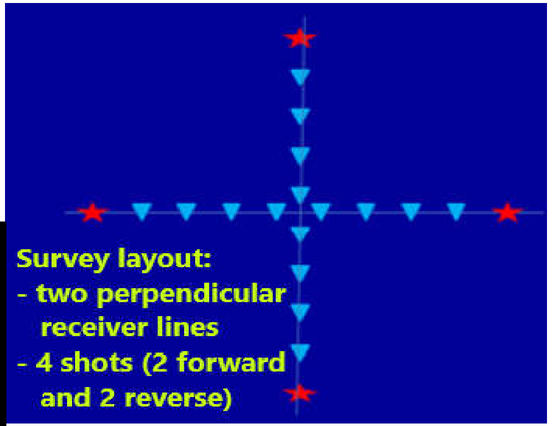
- The horizontal displacement between points on either side of the measuring traverse
- High-quality offset data are coveted by competent field measurements planners to optimize data designs. When lacking offset data, the field planner must be more conservative in designing traversers of measurements.

offset VSP

Abbreviation for offset vertical seismic profile, a type of vertical seismic profile in which the source is located at an offset from the drilling rig during acquisition. This allows imaging to some distance away from the wellbore.

perpendicular offset

In seismic surveys, perpendicular or normal offset is the component of the distance between the source and geophones at a right angle to the spread.



zero-offset data

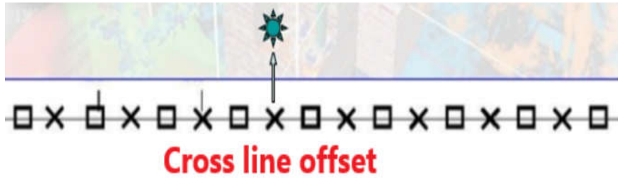
Seismic data acquired with no horizontal distance between the source and receiver.

check-shot survey

- A type of borehole seismic data designed to measure the seismic travel time from the surface to a known depth.
- P-wave velocity of the formations encountered in a wellbore can be measured directly by lowering a geophone to each formation of interest, sending out a source of energy from the surface of the Earth, and recording the resultant signal.
- The data can then be correlated to surface seismic data by correcting the sonic log and generating a synthetic seismogram to confirm or modify seismic interpretations.
- It differs from a vertical seismic profile in the number and density of receiver depths recorded; geophone positions may be widely and irregularly located in the wellbore, whereas a vertical seismic profile usually has numerous geophones positioned at closely and regularly spaced intervals in the wellbore.

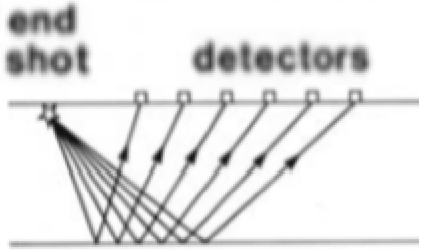
geophone interval

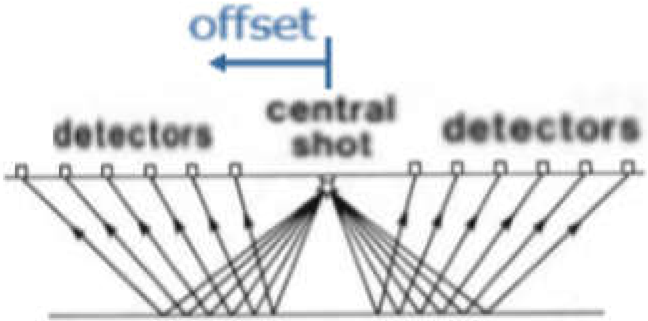
The distance between geophones or the centers of groups of geophones.

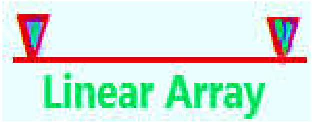


spread

The geometrical pattern of groups of geophones relative to the Seismic source. Common spread geometries include in-line offset, L-spread, split-spread and T-spread.

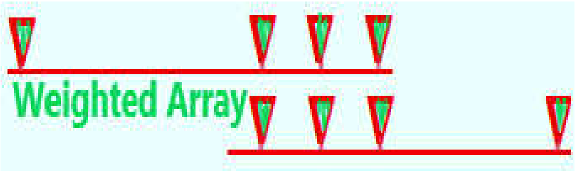


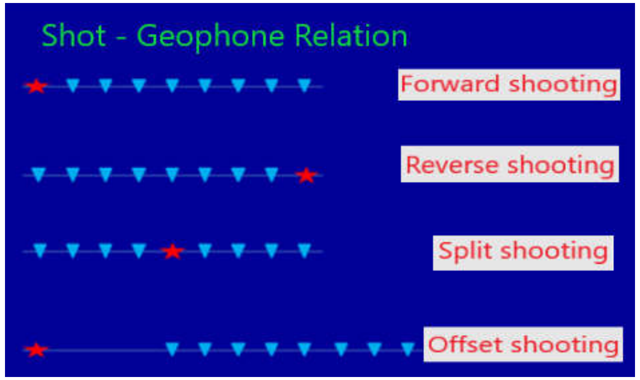




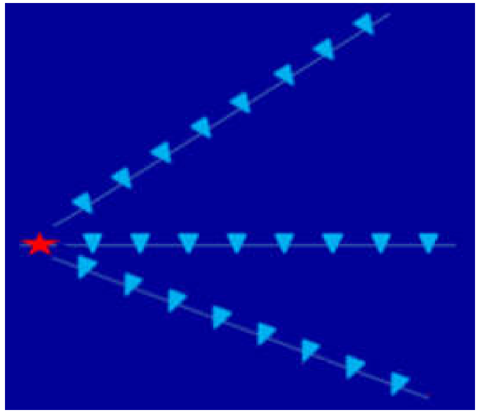
array

Generally, a geometrical configuration of transducers (sources or receivers) used to generate or record a physical field, such as an acoustic or electromagnetic wavefield or the Earth's gravity field.





radial array

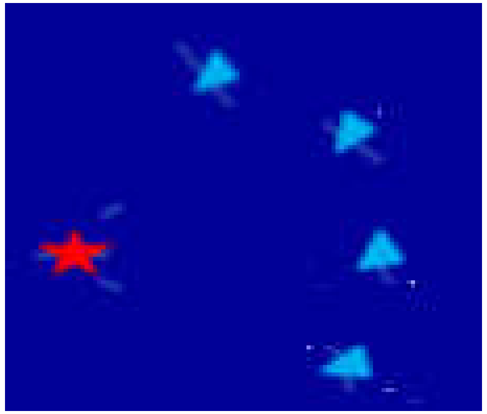


An array of sources or receivers radiating outward from a central point, usually a borehole.

radial refraction

A surveying technique used to identify local, high-velocity features such as salt domes, also called [fan shooting].

fan shooting



A technique for acquiring seismic refraction data around local, high-velocity features such as salt domes by using a fan or arc-shaped geophone array around a central shotpoint. The data from the fan-shaped array are calibrated against a control profile acquired some distance from the anomalous feature.

footprint

The area covered by an array of towed streamers in marine seismic acquisition.

signal

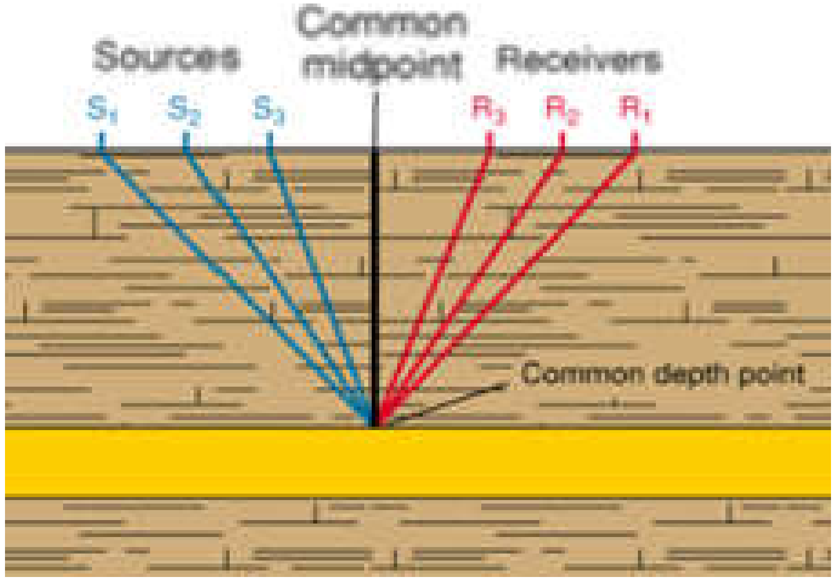
The portion of the seismic wave that contains desirable information. Noise is the undesirable information that typically accompanies the signal and can, to some extent, be filtered out of the data.

signal-to-noise ratio

The ratio of desirable to undesirable (or total) energy. The signal-to-noise ratio can be expressed mathematically as S/N or S/(S+N), although S/N is more commonly used. The signal-to-noise ratio is difficult to quantify accurately because it is difficult to completely separate signal from noise. It also depends on how noise is defined.

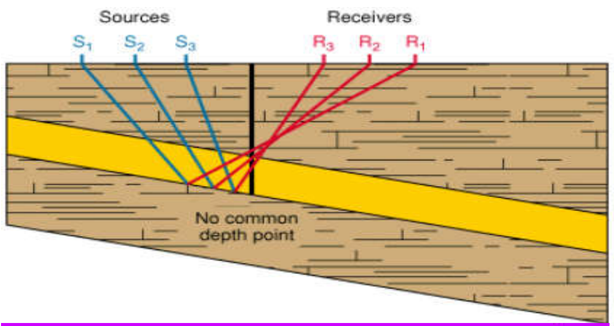
Common depth point (common reflection point)

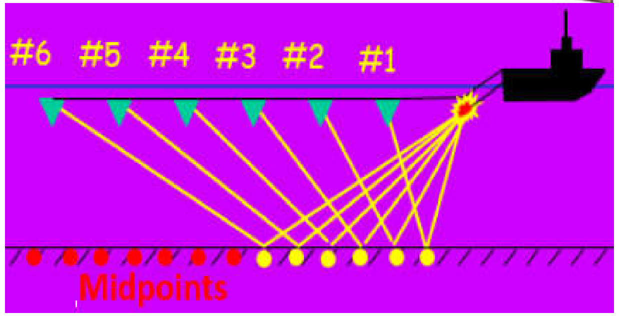
- In multichannel seismic acquisition where beds do not dip, the common reflection point at depth on a reflector, or the halfway point when a wave travels from a source to a reflector to a receiver.
- In the case of flat layers, the common depth point is vertically below the common midpoint.
- In the case of dipping beds, there is no common depth point shared by multiple sources and receivers, so dip moveout processing is necessary to reduce smearing, or inappropriate mixing, of the data.

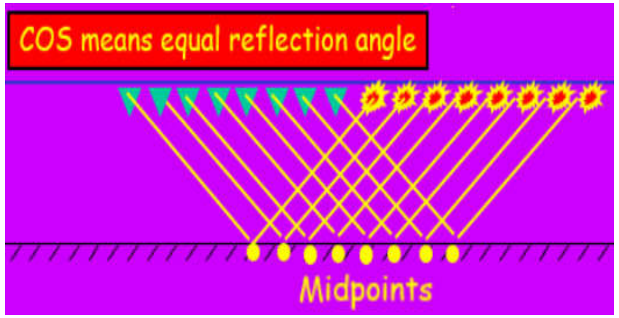


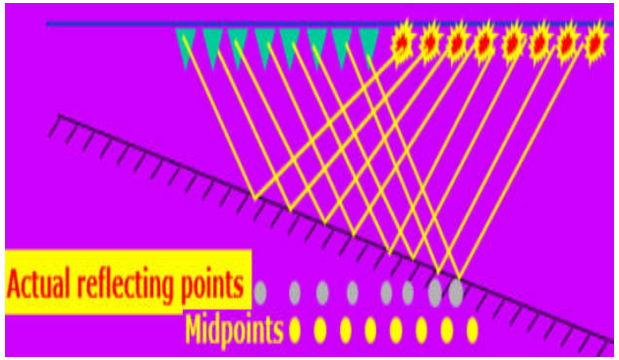
common midpoint

- Method of seismic reflection surveying and processing that exploits the redundancy of multiple fold to enhance data quality by reducing noise.
- In multichannel seismic acquisition, the point on the surface halfway between the source and receiver that is shared by numerous source-receiver pairs.
- During acquisition, an energy source is supplied to a number of shot points simultaneously. Once data have been recorded, the energy source is moved farther down the line of acquisition, but enough overlap is left that some of the reflection points are re-recorded with a different source-to-receiver offset.
- Multiple shotpoints that share a source-receiver midpoint are stacked and enhances the quality of seismic data.
- The number of times that a common midpoint is recorded is the fold of the data.
- The common midpoint is vertically above the common depth point, or common reflection point. Common midpoint is not the same as common depth point, but the terms are often incorrectly used as synonyms.



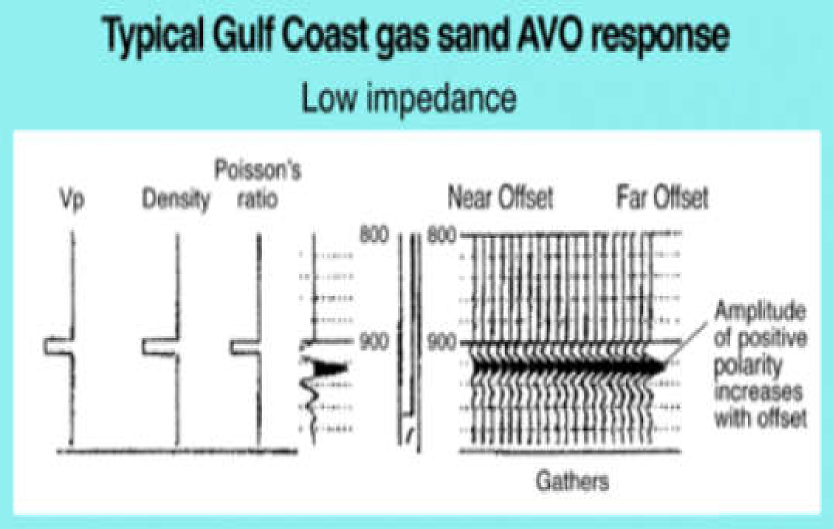


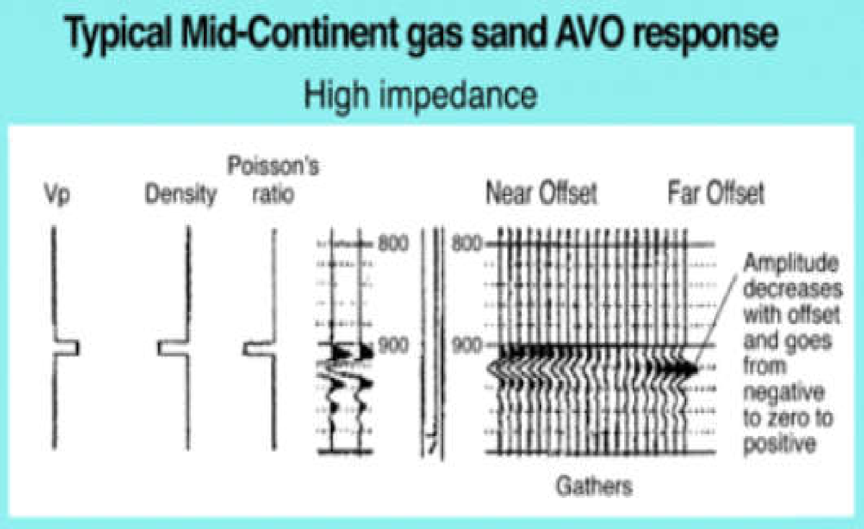




Amplitude variation with offset

Variation in seismic reflection amplitude with change in distance between shotpoints and receiver that indicates differences in lithology and fluid content in rocks above and below the reflector. AVO analysis is a technique by which geophysicists attempt to determine thickness, porosity, density, velocity, lithology and fluid content of rocks. Successful AVO analysis requires special processing of seismic data and seismic modeling to determine rock properties with a known fluid content. With that knowledge, it is possible to model other types of fluid content. A gas-filled sandstone might show increasing amplitude with offset, whereas a coal might show decreasing amplitude with offset. A limitation of AVO analysis using only P-energy is its failure to yield a unique solution, so AVO results are prone to misinterpretation. One common misinterpretation is the failure to distinguish a gas-filled reservoir from a reservoir having only partial gas saturation ("fizz water"). However, AVO analysis using source-generated or mode-converted shear wave energy allows differentiation of degrees of gas saturation. AVO analysis is more successful in young, poorly consolidated rocks, such as those in the Gulf of Mexico, than in older, well-cemented sediments.





seismic line (seismic section)

- A display of seismic data along a line, such as a 2D seismic profile
- or a profile extracted from a volume of 3D seismic data.
- A seismic section consists of numerous traces with location given along the x-axis and two-way traveltime or depth along the y-axis.
- The section is called a time section

depth section

A display of seismic data with a scale of units of depth rather than time along the vertical axis. Careful migration and depth conversion are essential for creating depth sections.

trace

The seismic data recorded for one channel. A trace is a recording of the Earth's response to seismic energy passing from the source, through subsurface layers, and back to the receiver.

zero crossing

The null point of a seismic trace. At zero deflection, the phase of a periodic signal is zero or pi.

one-way time

The time measured from a check-shot survey or vertical seismic profile (VSP), which is the time energy takes to travel from an energy source at the surface of the Earth to a receiver at a depth of interest.

Bias

An adjustment of the relative positive and negative excursions of reflections during seismic processing by bulk shifting the null point, or baseline, of the data to emphasize peaks at the expense of troughs or vice versa. Some authors describe bias as a systematic distortion of seismic data to achieve greater continuity.

baseline

A line joining base stations whose transmissions are synchronized during surveying.
A reference line, such as a “shale baseline," a line representing the typical value of a given measurement for a shale on a well log, or the zero-amplitude line of a seismic trace.

Aliasing

The distortion of frequency introduced by inadequately sampling a signal, which results in ambiguity between signal and noise.
Aliasing can be avoided by sampling at least twice the highest frequency of the waveform or by filtering frequencies above the Nyquist frequency, the highest frequency that can be defined accurately by that sampling interval.

alias filter

A filter, or a set of limits used to :

- **eliminate unwanted portions of the spectra of the seismic data,**
- **remove frequencies that might cause aliasing during the process of sampling an analog signal during acquisition or when the sample rate of digital data is being decreased during seismic processing.**

root-mean-square velocity

- The value of the square root of the sum of the squares of the velocity values divided by the number of values, symbolized by v_{rms}.
- The root-mean-square velocity is that of a wave through subsurface layers of different interval velocity along a specific raypath, and is typically several percent higher than the average velocity.
- The stacking velocity and the root-mean-square velocity approach equality when source-reciver offset approaches zero and layers are horizontal and isotropic.

interval velocity

The velocity, typically P-wave velocity, of a specific layer or layers of rock, symbolized by v_{int} and commonly calculated from acoustic logs or from the change in stacking velocity between seismic events on a common midpoint gather.

average velocity

In geophysics, the depth divided by the travel time of a wave to that depth. 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.

Dix formula

An equation used to calculate the interval velocity within a series of flat, parallel layers, named for American geophysist C. Hewitt Dix (1905 to 1984). Sheriff (1991) cautions that the equation is misused in situations that do not match Dix's assumptions. The equation is as follows:

$V_{\text{int}} = [(t_2\ V_{\text{RMS2}}{}^2 - t_1\ V_{\text{RMS1}}{}^2) / (t_2 - t_1)]^{1/2}$,

Where: V_{int} = interval velocity t_1 = traveltime to the first reflector t_2 = traveltime to the second reflector

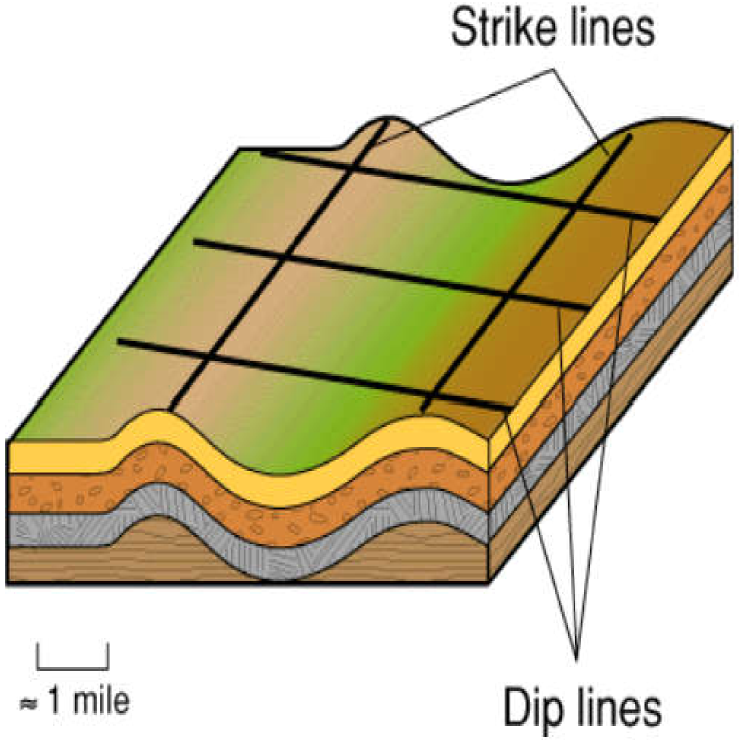
V_{RMS1} = root-mean-square velocity to the first reflector V_{RMS2} = root-mean-square velocity to the second reflector.

stacking velocity

The distance-time relationship determined from analysis of normal moveout (NMO) measurements from common depth point gathers of seismic data. The stacking velocity is used to correct the arrival times of events in the traces for their varying offsets prior to summing, or stacking, the traces to improve the signal-to-noise ratio of the data.

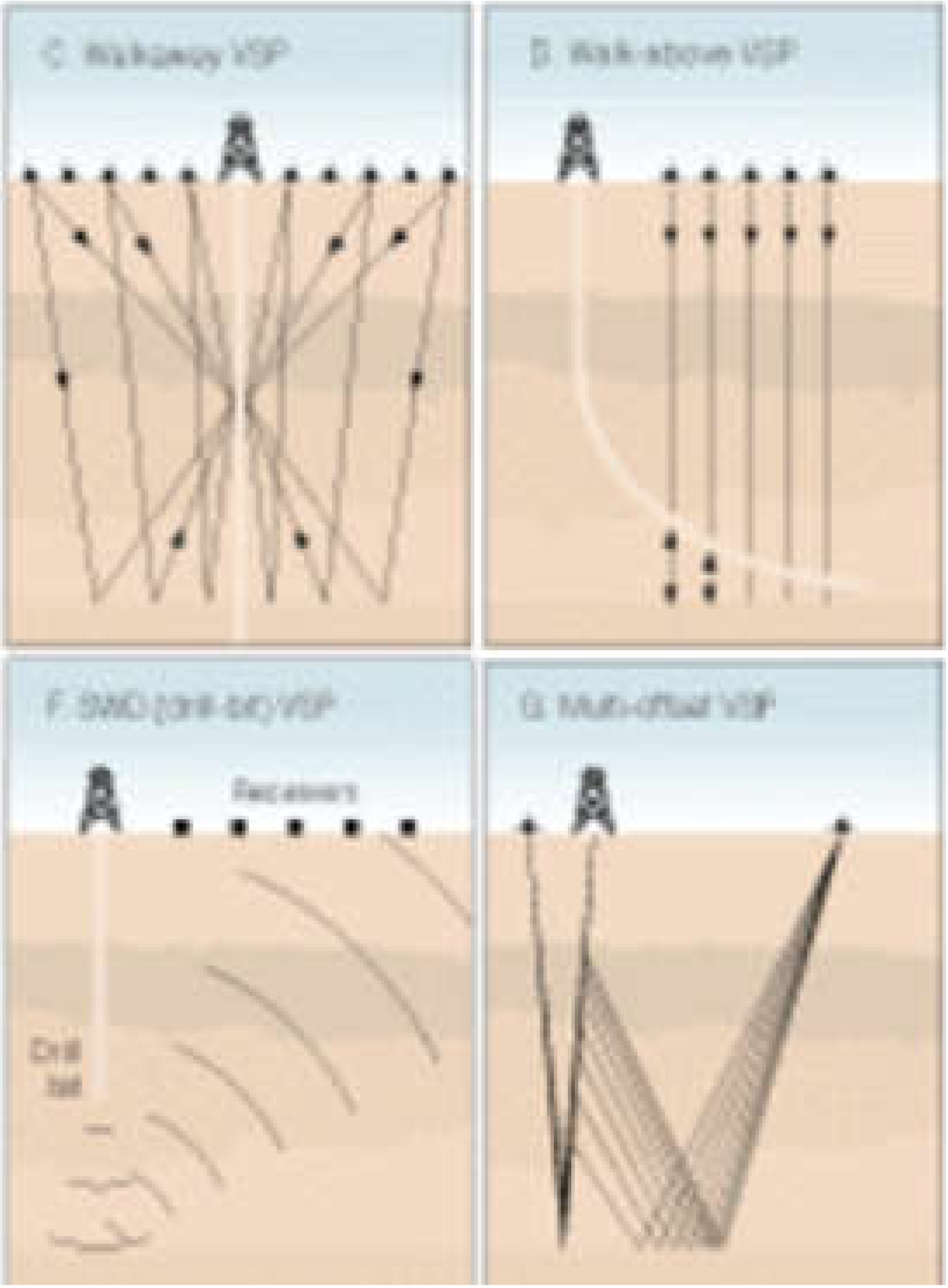
Two-dimensional survey

Seismic data or a group of seismic lines acquired individually such that there typically are significant gaps (commonly 1 km or more) between adjacent lines. A 2D survey typically contains numerous lines acquired orthogonally of geological structures (such as faults and folds) with a minimum of lines acquired parallel to geological structures to allow line-to-line tying of the seismic data and interpretation and mapping of structures.



vertical seismic profile

A class of borehole seismic measurements used for correlation with surface seismic data, for obtaining images of higher resolution than surface seismic images and for looking ahead of the drill bit; also called a VSP. Purely defined, VSP refers to measurements made in a vertical wellbore using geophones inside the wellbore and a source at the surface near the well. In the more general context, VSPs vary in the well configuration, the number and location of sources and geophones, and how they are deployed. Most VSPs use a surface seismic source, which is commonly a vibrator on land and an air gun in offshore or marine environments. VSPs include the zero-offset VSP, offset VSP, walkaway VSP, walk-above VSP, salt-proximity VSP, shear-wave VSP, and drill-noise or seismic-while-drilling VSP. A VSP is a much more detailed survey than a check-shot survey because the geophones are more closely spaced, typically on the order of 25 m [82 ft], whereas a check-shot survey might include measurements of intervals hundreds of meters apart. Also, a VSP uses the reflected energy contained in the recorded trace at each receiver position as well as the first direct path from source to receiver. The check-shot survey uses only the direct path traveltime. In addition to tying well data to seismic data, the vertical seismic profile also enables converting seismic data to zero-phase data and distinguishing primary reflections from multiples.



Multicomponent seismic data

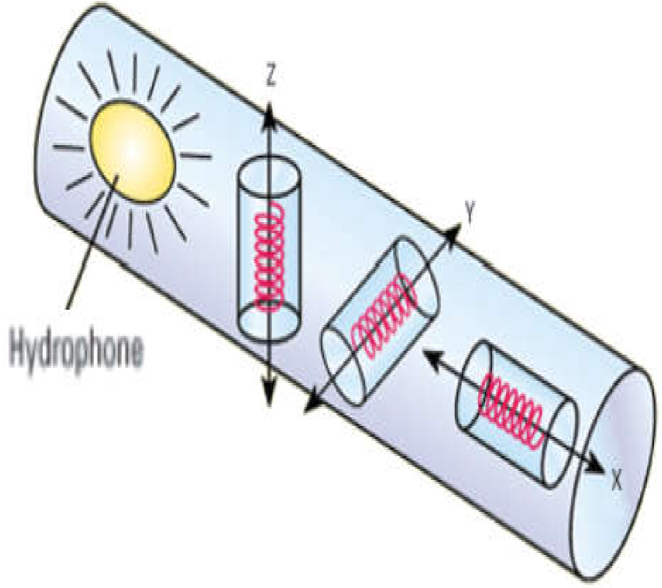
Seismic data acquired in a land, marine, or borehole environment by using more than one geophone or accelerometer.

Three-component seismic data

A type of multicomponent seismic data acquired in a land, marine, or borehole environment by using three orthogonally oriented geophones or accelerometers. 3C is particularly appropriate when the addition of a hydrophone (the basis for 4C seismic data) adds no value to the measurement, as for example, on land. This technique allows determination of both the type of wave and its direction of propagation.

Four-component seismic data

Four-component (4C) borehole or marine seismic data are typically acquired using three orthogonally-oriented geophones and a hydrophone within an ocean-bottom sensor (deployed in node-type systems as well as cables). Provided the system is in contact with the seabed or the borehole wall, the addition of geophones allows measurement of shear (S) waves, whereas the hydrophone measures compressional (P) waves.



Reflection surveying

- Sensitive to impedance contrasts
- Use near-normal incidence i.e. P-waves
- Target scale:

  10’s m: Ground water, engineering and environmental studies

  km’s: Oil exploration

  10’s km: Crustal structure

Deep Seismic Reflection and Refraction Profiling

Multi-channel recording, along a measurement line, of seismic waves, artificially generated using large energy sources, after these have travelled deep through the earth’s crust (and upper mantle).

Deep-reflection profiling is done using vibrators (on land) or air guns (in water) at near-vertical distances (8–12 km) to image the structure of the crust and upper mantle. Wide-angle reflection/refraction profiling uses large explosions and recording distances (200–300 km), primarily to obtain velocity information down to upper mantle.

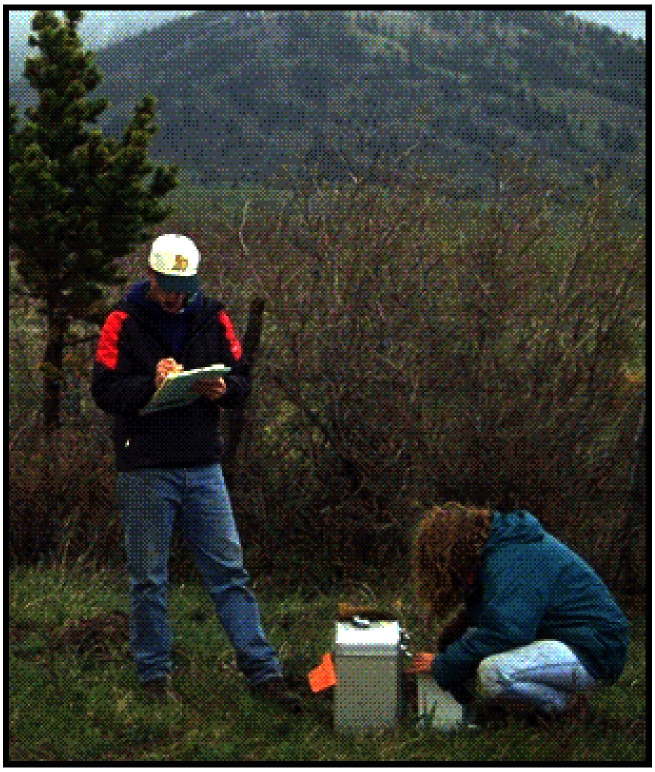
Geophysical Surveys: Active Versus Passive

Introduction:

- Active and Passive Geophysical Methods
- Advantages and Disadvantages of Each Method

Geophysical surveys can be classified into one of two types; *Active* and *Passive*. Passive geophysical surveys are ones that incorporate measurements of naturally occurring fields or properties of the earth. We have already considered passive geophysical surveys in our discussions of gravity and magnetic surveys. In these two cases, the naturally occurring fields are the gravitational and magnetic fields. We simply measure spatial variations in these fields and attempt to infer something about the subsurface geology from these measurements. The fields and properties that we are measuring in this class of experiments exist regardless of our geophysical survey. Examples of other earth properties that could be passively measured include radiometric decay products, certain electrical fields, and certain electromagnetic fields

-
-



In conducting active geophysical surveys, on the other hand, a signal is injected into the earth and we then measure how the earth responds to this signal. These signals could take a variety of forms such as displacement, an electrical current, or an active radiometric source. The final two survey methods considered in this short course,

DC resistivity and seismic refraction, are examples of active geophysical experiments.

Advantages and Disadvantages of Active and Passive Experiments

Shown below is a table listing some of the advantages and disadvantages to each of these types of surveys. In reading these, please note that the terms passive and active cover a wide range of geophysical survey methods. Thus, the listed advantages and disadvantages are by necessity generalizations that might not apply to any given specific survey.

Active		Passive	
Advantage	**Disadvantage**	**Advantage**	**Disadvantage**
Better control of noise sources through control of injected signal.	**Because both sources and receivers are under the surveyor's control, he must supply both. Therefore, field equipment tends to be more complex.**	**Surveyor need only record a naturally occurring field, therefore, he need supply only a sensor and a data recorder.**	**Less control of noise because source of the signal is out of the hands of the surveyor.**
Because propagating fields are generally measured, active experiments usually provide better depth control over source of anomalous signal.	**Field operations and logistics are generally more complex and time consuming than passive experiments.**	**Field operations are generally very time efficient. Thus, passive experiments can be run over wider areas in a more cost-effective manner.**	**Because passive fields are generally the result of integrating anomalous geological contributions overwide areas, identification of the source of an anomalous observation can be difficult.**
Many different	**Many different**	**One or two well-**	**One or two well-**

source/receiver configurations can be used allowing for a wide variety of survey designs. This allows survey designers great flexibility in customizing surveys for particular problems.	**source/receiver configurations can be used allowing for a wide variety of survey designs. The increase in the number of field options inevitably leads to greater survey design costs and potentially leads to increased probability of field mishaps.**	**established field procedures are generally used. Contractors can provide these surveys on short notice with relatively easily quantifiable results.**	**established field procedures are generally used. This limits the amount of customization that can be done for specific problems.**
Once set up, active experiments are capable of producing vast quantities of data that can be used to interpret subtle details of the earth's subsurface.	**The large quantity of data obtained in many active experiments can become overwhelming to process and interpret.**	**Interpretation of the limited set of observations can be accomplished with modest computational requirements quickly and efficiently.**	**The data sets collected in passive experiments are smaller than those collected in active experiments and usually do not allow for as detailed an interpretation.**

Electrical Methods Overview

Bridging our subdivision of geophysical techniques into passive and active methods are the electrical and electromagnetic methods. Taken as a whole, the electrical and electromagnetic methods represent the largest class of all geophysical methods; some passively monitor natural signals while others employ active sources. In addition to their great variety, this group of geophysical techniques represents some of the oldest means of exploring the Earth's interior. For example the *SP*(Self Potential) method described below dates back to the 1830s when it was used in Cornwall, England by Robert Fox to find extensions of known copper deposits. Natural electrical currents in the Earth, referred to as *telluric* currents were first identified by Peter Barlow (pictured) in 1847. The *EM* method was developed in the 1920s for the exploration of base-metal deposits.

Electrical methods employ a variety of measurements of the effects of electrical current flow within the Earth. The phenomena that can be measured include current flow, electrical potential (voltages), and electromagnetic fields. A summary of the better-known electrical methods is given below. In this set of notes we will consider only one of these methods, the DC resistivity method, in greater detail.

DC Resistivity - This is an active method that employs measurements of electrical potential associated with subsurface electrical current flow generated by a DC, or slowly varying (very low frequency current) AC, source. Factors that affect the measured potential, and thus can be mapped using this method, include the presence and quality of pore fluids and clays. Our discussions will focus solely on this method.

Induced Polarization (IP) - This is an active method that is commonly done in conjunction with DC Resistivity. It employs measurements of the transient (short-term) variations in potential as the current is initially applied or removed from the ground, or alternatively the variation in the response as the AC frequency is changed. It has been observed that when a current is applied to the ground, the ground behaves much like a capacitor, storing some of the applied current as a charge that is dissipated upon removal of the current. In this process, both capacitative and electrochemical effects are responsible.

IP is commonly used to detect concentrations of clay, and electrically conductive metallic mineral grains.

Electromagnetic (EM) - This is an active method that employs measurements of a time-varying magnetic field generated by induction through current flow within the earth. In this technique, a time-varying magnetic field is generated at the surface of the earth that produces a time-varying electrical current in the earth through induction. A receiver is deployed that compares the magnetic field produced by the current-flow in the earth to that generated at the source. EM is used for locating conductive base-metal deposits, for locating buried pipes and cables, for the detection of unexploded ordnance, and for near-surface geophysical mapping.

Self Potential (SP) - This is a passive method that employs measurements of naturally occurring electrical potentials commonly associated with shallow electrical conductors, such as sulfide ore bodies. Measurable electrical potentials have also been observed in association with groundwater flow and certain biologic processes. The only equipment needed for conducting an SP survey is a high-impedance voltmeter and some means of making good electrical contact to the ground.

Magnetotelluric (MT) - This is a passive method that employs measurements of naturally occurring electrical currents, telluric currents, generated by magnetic induction from electrical currents in the ionosphere. This method can be used to determine electrical properties of materials at relatively great depths (down to and including the mantle) inside the Earth. In this technique, a time variation in electrical potential is measured at a base station and at survey stations. Differences in the recorded signal are used to estimate subsurface distribution of electrical resistivity

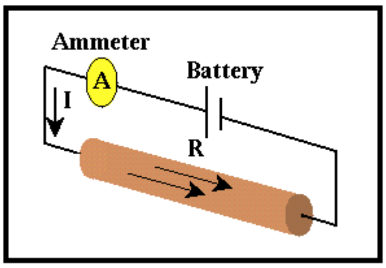
Resistivity Basics

Current Flow and Ohm's Law

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

$$\boxed{V = IR}$$

Ohm found that the current, I, was proportional to the voltage, V, for a broad class of materials that we now refer to as *ohmic* materials. The constant of proportionality is called the *resistance* of the material and has the units of voltage (volts) over current (amperes), *(ΔV/I)* or *ohm*.



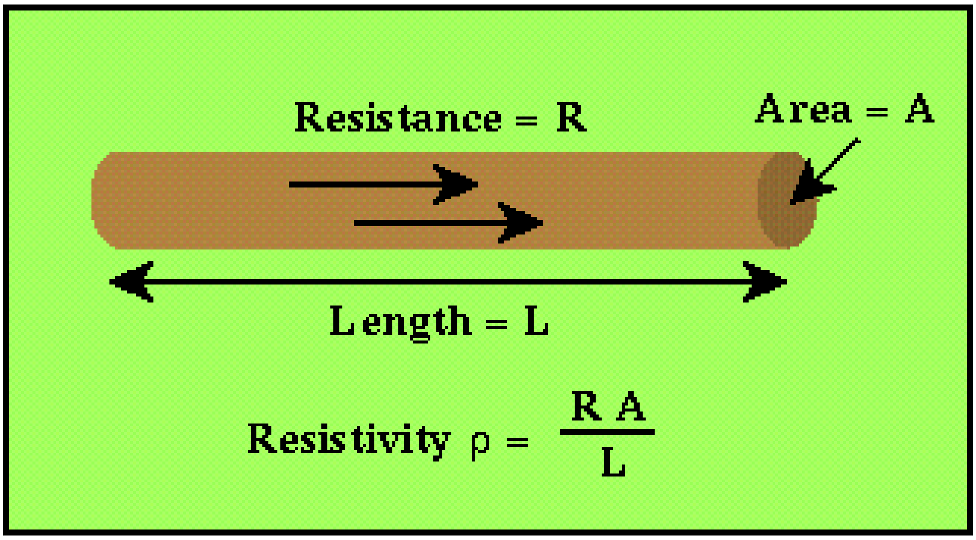
In principle, it is relatively simple to measure the resistance of a strand of wire. Connect a battery to the wire of known voltage and then measure the current flowing through the wire. The voltage divided by the current yields the resistance of the wire. In essence, this is how your multimeter measures resistance. In making this measurement, however, we must ask two crucial questions.

- Is the measured resistance related to some fundamental property of the material from which the wire is made?
- How can we apply this relatively simple experiment to determining electrical properties of earth materials?

It's Resistivity, not Resistance

The problem with using resistance as a measurement is that it depends not only on the material out of which the wire is made, but also the geometry of the wire. If we were to increase the length of wire, for example, the measured resistance would increase. Also, if we were to decrease the diameter of the wire, the measured resistance would increase. We want to define a property that describes a material's ability to transmit electrical current that is independent of the geometrical factors.

The quantity that is used is called *resistivity* and is usually indicated by the Greek symbol ρ(read *rho**,**).



In the case of the wire, resistivity is defined as the resistance in the wire, multiplied by the cross-sectional area of the wire, divided by the length of the wire. The units associated with resistivity are thus **ohm.m** (ohm - meters).

Resistivity is a fundamental parameter of the material making up the wire that describes how easily the wire can transmit an electrical current. High values of resistivity imply that the material making up the wire is very resistant to the flow of electricity. Low values of resistivity imply that the material making up the wire transmits electrical current very easily.

Resistivity of Earth Materials:

Although some native metals and graphite conduct electricity, most rock-forming minerals are electrical insulators. Measured resistivities in Earth materials are primarily controlled by the movement of charged ions in pore fluids. Although water itself is not a good conductor of electricity, ground water generally contains dissolved compounds that greatly enhance its ability to conduct electricity. Hence, porosity and fluid saturation tend to dominate electrical resistivity measurements. In addition to pores, fractures within crystalline rock can lead to low resistivities if they are filled with fluids.

The resistivities of various earth materials are shown below.

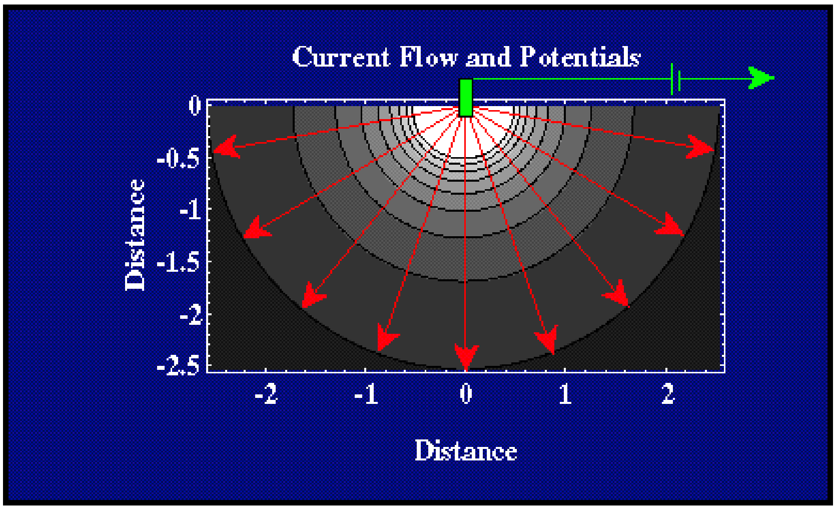
Material	Resistivity (Ohm.meter)
Air	**~ ?**
Pyrite	**2.9 x 10^-5 - 1.5**
Galena	**3 x 10^-5 - 3 x 10^2**
Sphalerite	**1.5 - 1 x 10^7**
Quartz	**4 x 10^10 - 2 x 10^14**
Calcite	**2 x 10^12**
Rock Salt	**30 - 1 x 10^13**
Mica	**9 x 10^12 - 1 x 10^14**
Ground Water	**0.5 - 300**
Sea Water	**0.2**

Diabase	**20 - 5 x 10^7**
Limestones	**50 - 1 x 10^7**
Sandstones	**1 - 6.4 x 10^8**
Shales	**20 - 2 x 10^3**
Gabbro	**1 x 10^3 - 1 x 10^6**
Basalt	**10 - 1.3 x 10^7**
Dolomite	**3.5 x 10^2 - 5 x 10^3**

Like susceptibilities, there is a large range of resistivities, not only between varying rocks and minerals, but also within rocks of the same type. This range of resistivities, as described above, is primarily a function of fluid content. Thus, a common target for electrical surveys is the identification of fluid saturated zones. For example, resistivity methods are commonly used in engineering and environmental studies for the identification of water table.

Current Densities and Equipotentials

To describe the nature of electrical current flow in media occupying a volume, we must move beyond the simple notions of current and voltage gained from our experience with wires, resistors, and batteries. In the Earth, or any three-dimensional body, electrical current is not constrained to flow along a single path as it does in a wire. Consider as an example the situation shown below. A battery is connected to the earth by wires and electrodes at two remote points (that is the electrical connections to the earth are very distant from one another). The Earth, not being a perfect insulator, conducts the electrical current imported by the battery. At this stage, lets assume the resistivity of the earth is uniform throughout the Earth. How does the current flow through the Earth?



In this example, current flows (the red lines) out from the electrode (the green square) radially along straight lines (the second electrode is far to the right of the figure as it is drawn). The resistivity of the medium imposes a voltage drop as we move away from the electrode. If we could take a voltmeter and measure the voltage drop between a point very far from the current electrode to various places in (on) the medium near the electrode, we would find that the voltage drops would be constant along circular lines centered at the electrode. (That is, one of the leads to the voltmeter would make contact with the ground at a distance very far from the electrode, while the other is then moved throughout the medium). These lines are referred to as *equipotentials* (think equal voltage). In three dimensions, they form hemispheres centered on the electrodes. Several equipotential lines are shown in black with the voltage drop associated by each line shown in gray scale. The darker the gray scale, the smaller the potential drop between this location and a location very far from the current electrode.

$$\boxed{V = \frac{\rho I}{2\pi r}}$$

Voltage differences between any two points in the medium can be computed by simply subtracting the potentials at the two points. Thus, if the two points lie on a hemisphere centered at the current electrode, there will be no voltage difference recorded, because these two locations lie along an equipotential surface. That is, if you were to take your voltmeter and connect to two points within the earth that were on the same equipotential surface, you would read a voltage difference of zero. When compared to the potential near the electrode, voltage differences will increase away from the electrode. This should make sense, what you are measuring with your voltmeter is proportional to the current passing through the media times the resistance of the media as given by Ohm's law. As you move away from the electrode, your current is traveling through more of the media. The resistance increases as the path increases, hence, the voltage increases.

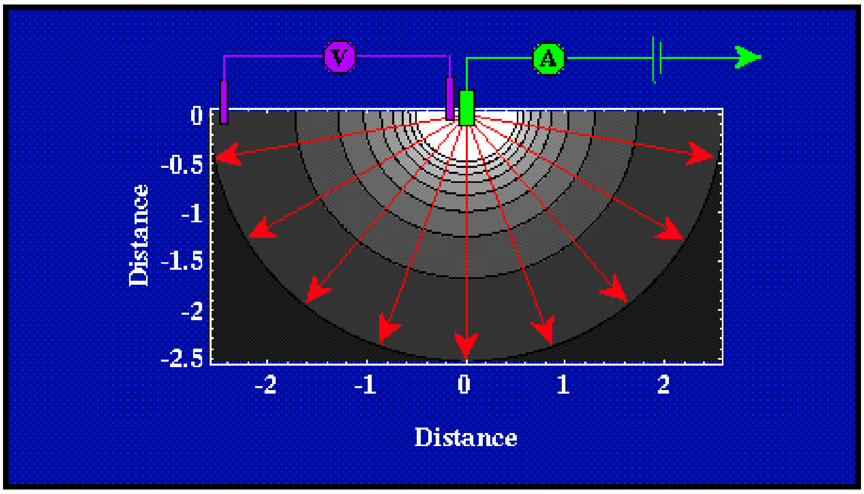
At any point in the medium, the *current density* is defined as the amount of current passing through a unit area of an equipotential surface. Thus, close to the electrode, all of the current is passing through a very small volume. The current crossing any equipotential surface normalized by the area of the surface will thus be high. Far away from the electrode, this same current occupies a much larger volume of the medium. The total current (which is the same regardless of where the surface is in the volume) crossing any equipotential surface, normalized by the area of the surface (which is now large), will be small.

Current density has the units of Amperes per meter squared.

A First Estimate of Resistivity

The voltage change from a single current electrode to some point in the half space representing the earth is given by the expression to the right. In this expression, *V* is voltage, *I* is current, *rho* is resistivity, and *r* is the distance between the current electrode and the point the voltage is measured. Notice that this expression is nothing more than Ohm's law with the resistance, *R* equal to *rho* over *2 ? r*.

If the Earth had a constant resistivity (it doesn't) we could estimate this resistivity by performing the following experiment. Attach to the wire connecting the battery with one of the current electrodes an ammeter to measure the amount of current going into the earth. Place one electrode connected to a voltmeter next to the current electrode and place the other at some distance, *r*, away from the electrode and measure the voltage difference between the two locations. Using the expression given above, compute the resistivity, *rho*.



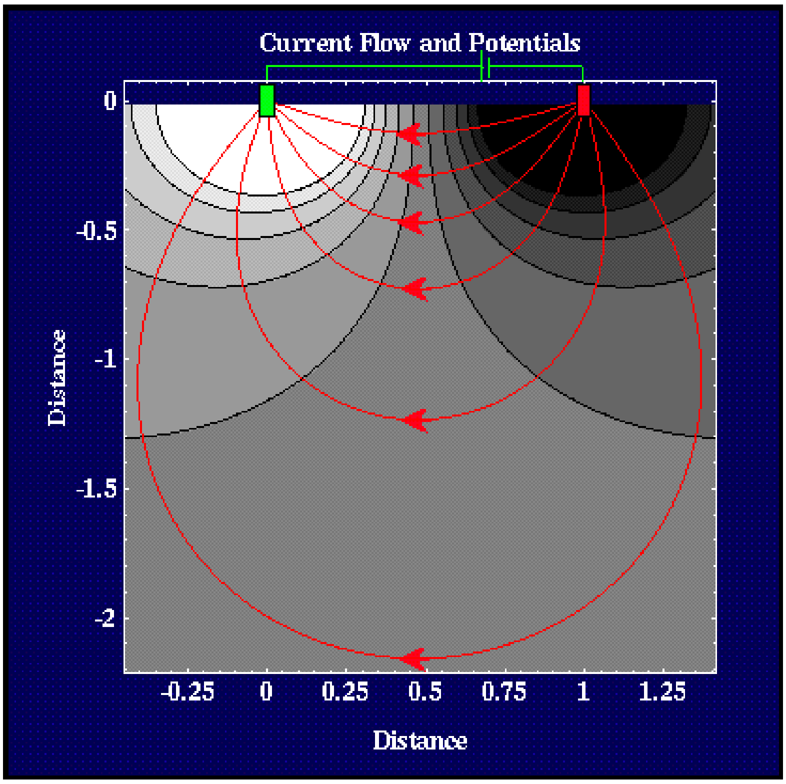
In practice, this experiment could be difficult to implement because the two current electrodes must be placed a great (usually 10 times the distance over which the voltage is being measured) distance from one another. So, why not simply decrease the distance between the two voltage electrodes so the distance between the two current electrodes remains at a practical distance? The problem is that the closer the two voltage electrodes are to each other, the smaller the voltage difference that must be measured. Thus, there is a practical limit to how close the two voltage electrodes can be and thereby how close the two current electrodes can be. More importantly, there is no need to place the second electrode at "infinity" if we develop a method to take its effect into account.

As another note, one may ask why don't we simply attach the voltmeter to the wire in which the current is flowing and measure the voltage drop between the two current electrodes. This could be done. In practice, however, it is impossible to obtain information about the Earth, because what you measure is more a function of the *contact resistance* between the earth and the current electrodes than of the resistivity of the Earth. The contact resistance is the resistance that is encountered to current flow because the electrode does not make perfect electrical contact with the earth. Contact resistances can be quite large, on the order of kilo-ohms (10^4 ohms), although good field practice can reduce these to the order of an ohm, if necessary. However, the *resistance* between the electrodes will be made up of the contact resistance at each electrode, and the resistance of the Earth - which is effectively zero (the "wire" had a very large cross-sectional area!). So, the voltage measured will be dominated by the voltage drop over the electrode/earth contacts, even if it is small.

If a large (infinite) impedance voltmeter is used to make the voltage measurement across two *separate* voltage electrodes, however, very little current actually flows through the voltage electrodes and contact resistance is unimportant to the measurement.

Current Flow From Two Closely Spaced Electrodes

In practice, we will need to place the two current electrodes close to each other. In doing so, however, the current distribution and equipotentials produced within a homogeneous earth become more complicated than those shown previously.



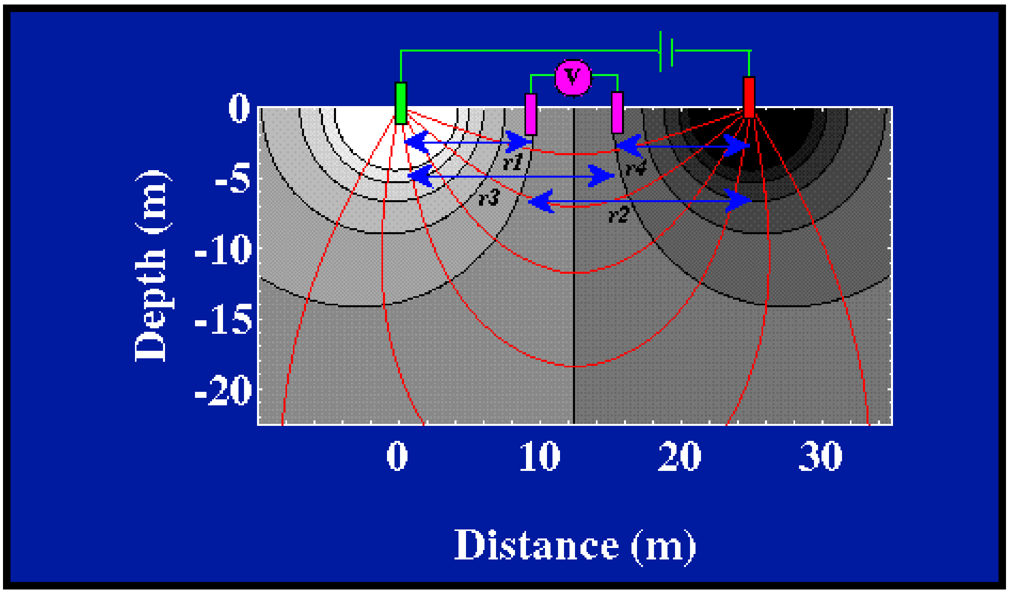
Instead of the current flowing radially out from the current electrodes, it now flows along curved paths connecting the two current electrodes. Six current paths are shown (red lines). Between the surface of the earth and any current path we can compute the total proportion of current encompassed. The table below shows this proportion for the six paths shown above. Current paths are labeled 1 through 6 starting with the topmost path and proceeding to the bottom-most path.

Current Path	% of Total Current
1	17
2	32
3	43
4	49
5	51
6	57

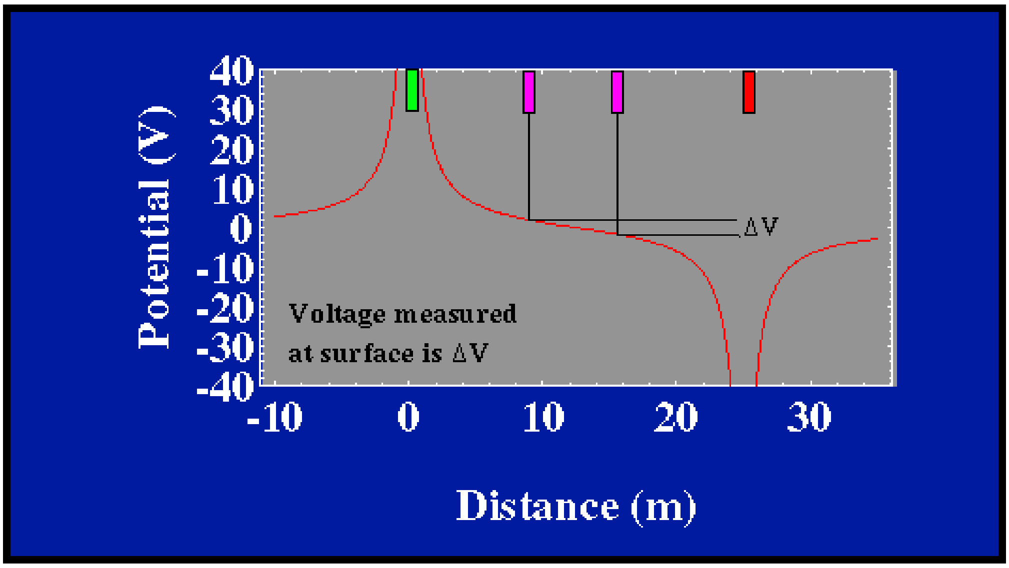
From these calculations and the graph of the current flow shown above, notice that almost 50% of the current placed into the ground flows through rock at depths shallower or equal to the current electrode spacing.

A Practical way of Measuring Resistivity

Using an experimental configuration where the two current electrodes are placed relatively close to one another as described previously and using two potential electrodes place between the two current electrodes, we can now estimate the resistivity of our homogeneous earth. The configuration of the four electrodes for this experiment is shown below. Let the distances between the four electrodes be given by *r1, r2, r3,* and *r4* as shown in the figure.



The <u>potential</u> computed along the surface of the earth is shown in the graph below. The voltage we would observe with our voltmeter is the <u>difference in potential</u> at the two voltage electrodes, l. The horizontal positions of the four electrodes, two current (red and green), and the two potential (purple) are indicated along the top of the figure.



Notice, that in this configuration, the voltage recorded by the voltmeter (l) is relatively small. That is, the difference in the potential at the locations of the two potential electrodes is small. We could increase the size of the voltage recorded by the voltmeter by moving the two potential electrodes outward, closer to the two current electrodes. For a variety of reasons, some related to the reduction of <u>noise</u> and some related to maximizing the depth over which our measurements are sensitive, we will typically not move the potential and current electrodes close together. Thus, a very sensitive voltmeter must be used. In addition to having a large impedance, voltmeters need to be able to record voltage differences down to mV (10^-3 volts). If the potential electrodes were moved closer to the two current electrodes, larger voltages would be recorded. For a variety of reasons, however, we will typically not do this in the field.

Knowing the locations of the four electrodes, and by measuring the amount of current input into the ground, *i*, and the voltage difference between the two potential electrodes, *l*, we can compute the resistivity of the medium, *rho-a*, using the following equation.

$$\rho_a = \frac{2\pi\Delta V}{i}\left[\frac{1}{\left(\frac{1}{r_1}-\frac{1}{r_2}-\frac{1}{r_3}+\frac{1}{r_4}\right)}\right]$$

In this particular case, regardless of the location of the four electrodes, *rho-a* will be exactly equal to the resistivity of the medium. The resistivity computed using the equation given above is referred to as the *apparent resistivity*. We call it the apparent resistivity for the following reason. We can always compute *rho-a*, we only need to know the locations of the electrodes and measure the current and voltage. If, however, the Earth does not have a constant resistivity (that is if the resistivity varies with depth or horizontally), the resistivity computed by the above equation will not represent the true resistivity of the Earth. Thus, we refer to it as an apparent resistivity.

As a final caveat, as written above, the difference between the apparent and the true resistivity of the medium is not a function of any noise that might be associated with the measurements we are attempting to record. The difference rather comes from the fact that our measurement, in some sense, averages the true resistivities of some region of the earth, yielding an apparent resistivity that may not represent the true resistivity at some specific point within the earth.

Resistivity Surveys and Geology

1- **Sources of Noise**

2- Depth of Current Penetration Versus Current Electrode Spacing

3- Current Flow in Layered Media

4-Variation in Apparent Resistivity: Layered Versus Homogeneous Media

5- Current Flow in Layered Media Versus Electrode Spacing

6- A Second Example of Current Flow in Layered Media

Sources of Noise

Even given the simple experiment outline on the previous page, there are a number of sources of noise that can affect our measurements of voltage and current from which we will compute apparent resistivities.

1-Electrode Polarization - A metallic electrode, like a copper or steel rod, in contact with an electrolyte, ground water, other than a saturated solution of one of its own salts will generate a measurable contact potential. In applications such as SP, these contact potentials can be larger than the natural potential that you are trying to record. Even for the DC methods described here these potentials can be a significant fraction of the total potential measured.

For DC work, there are two possible solutions.

1. a-Use ***nonpolarizing*** *electrodes*. These are electrodes that contain a metallic conducting rod in contact with a saturated solution of its own salt. Copper and copper sulfate solution are commonly used. The rod and solution are placed in a porous ceramic container that allows the saturated solution to slowly leak out and make contact with the ground. Because these solutions are rather environmentally non-friendly, and because the method described below is easy to employ, these so-called *porous pot* electrodes are rarely used in DC work. They are, however, commonly used in SP and IP surveys.

(1-b): A simple method to avoid the influence of these contact potentials is to periodically reverse the current flow in the current electrodes or use a slowly varying, a few cycles per second, AC current. As the current reverses, the polarizations at each electrode break down and begin to reverse. By measuring over several cycles, robust current and voltage measurements can be made with negligible polarization effects.

2--Telluric Currents -As described previously, naturally existing currents flow within the earth. These currents are referred to as telluric currents. The existence of these currents can generate a measurable voltage across the potential electrodes even when no current is flowing through the current electrodes.

Solution of (2):By periodically reversing the current from the current electrodes, or by employing a slowly varying AC current, the effects of telluric currents on the measured voltage can be cancelled.

3-Presence of Nearby Conductors -Electrical surveys can not be performed around conductors that make contact with the ground. For example, the presence of buried pipes or chain-linked fences will act as current sinks. Because of their low resistivity, current will preferentially flow along these structures rather than flowing through the earth. The presence of these nearby conductors essentially act as electrical shorts in the system.

Solution of (3) : choose your VES's locations far away from the conductors connecting the earth.

4-Low Resistivity at the Near Surface -Just as nearby conductors can act as current sinks that short out an electrical resistivity experiment, if the very near surface has a low resistivity, it is difficult to get current to flow more deeply within the earth. Thus a highly conductive* near-surface layer such as a perched water table can prevent current from flowing more deeply within the earth.

Solution (4):Do as we did with near surface conductors previously mentioned.

5-Near-Electrode Geology and Topography - Any variations in geology, or water content localized around an electrode, which produces near-surface variations in resistivity can greatly influence resistivity measurements. In addition, rugged topography will act to concentrate current flow in valleys and disperse current flow on hills.

Solution(5) you can select another sites of smooth topography for your Ves's

6-Current Induction in Measurement Cables - Current flowing through the cables connecting the current source to the current electrodes can produce an induced current in the cables connecting the voltmeter to the voltage electrodes, thereby generating a spurious voltage reading.

Solution (6): This source of noise can be minimized by keeping the current cables physically away from, a meter or two, the voltage cables.

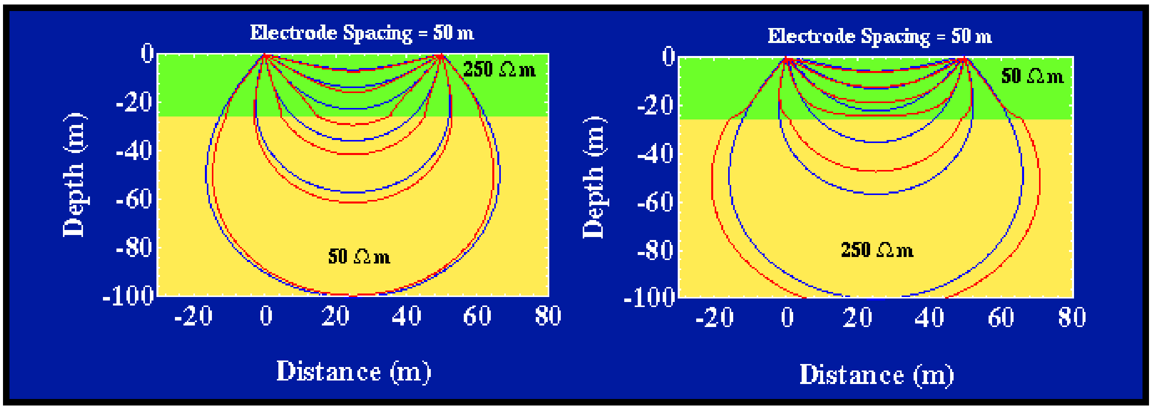
Notes

**Conductivity* is the inverse of resistivity. Highly *conductive* media transmit electrical current with great ease, thus they have a low *resistivity*. Mathematically, conductivity is the reciprocal of resistivity and is measured in the units of 1 over Ohm meters. 1/Ohm is referred to as a *siemen* (S) (sometimes, as "mho"). Thus, the units of conductivity are siemens per meter.

Current Flow in Layered Media

How does the presence of depth variations in resistivity affect the flow of electrical current? In the previous examples, we assumed that the Earth has a constant resistivity. Obviously this isn't true or else we wouldn't be trying to map the variation in resistivity throughout the earth. Although resistivity could conceivably vary in depth and in horizontal position, we will initially only consider variations in depth. In addition, we will assume that these depth variations in resistivity can be quantized into a series of discrete layers, each with a constant resistivity. Thus, initially we will not consider variations in resistivity in the horizontal direction or continuous variations in depth*.

Shown below are current-flow paths (red) from two current electrodes in two simple two-layer models. The model to the left contains a high-resistivity layer (250 ohm.m) overlying a lower resistivity layer (50 ohm.m). This model is characteristic of the resistivity profile that would be found in a region where unsaturated alluvium overlies water saturated alluvium. The model to the right contains a low-resistivity layer (50 ohm.m) overlying a higher resistivity layer (250 ohm.m). This model is characteristic of a perched aquifer. For comparison, we've also shown the paths current would have flowed along if the Earth had a constant resistivity (blue) equal to that of the top layer. These paths are identical to those described previously.



Notice that the current flow in the layered media deviates from that observed in homogeneous media. In particular, notice that in layered media the current flow lines are distorted in such a way that current preferentially seems to be attracted to the lower-resistivity portion of the layered media. In the model on the left, current appears to be pulled downward into the 50 ohm.m layer. In the model on the right, current appears to be bent upward, trying to remain within the lower resistivity layer at the top of the model. This shouldn't be surprising. What we are observing is the current's preference toward flowing through the path of least resistance. For the model on the left, that path is through the deep layer. For the model on the right, that path is through the shallow layer.

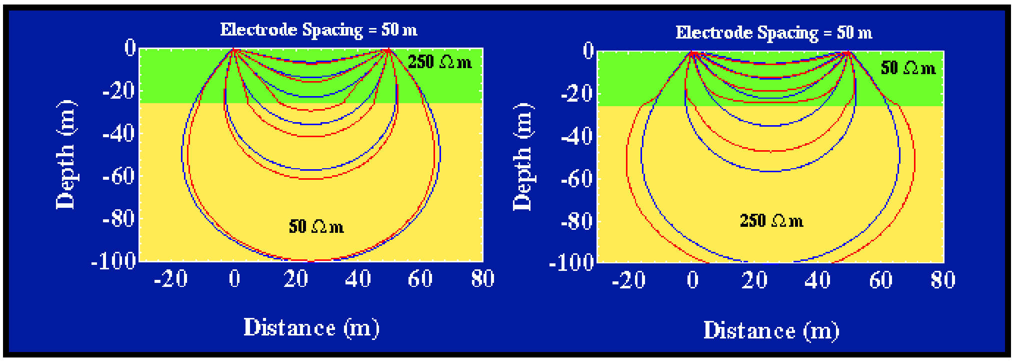
Note

*For all practical purposes this layered model does allow for continuous variations in resistivity with depth because we have made no constraints on the number of layers allowed in the model or their thicknesses. Thus, a smoothly varying resistivity depth profile could be approximated by a large number of very thin, constant resistivity layers.

Variation in Apparent Resistivity: Layered Versus Homogeneous Media

An important consequence of the deviation in current flow in layered media is how it can affect our measurements of apparent resistivity. Imagine that we configured an electrical experiment over these two models by measuring the potential difference between two places on the surface of the earth between the two current electrodes and compute the apparent resistivity. In these examples we will assume that the potential electrodes are between the two current electrodes and have a fixed separation that remains constant throughout the experiment. This is the same geometry for a four electrode experiment, two current and two potential, that was described previously.

Because current is preferentially being pulled into the lower layer for the model on the left, the current density between the two current electrodes near the surface of the Earth (where we are measuring electrical potential) will be smaller than that which would be observed if the Earth were homogeneous. By the same token for the model on the right, the current density would be higher than that observed in a homogeneous earth because the current is being preferentially channeled through the near-surface layer.

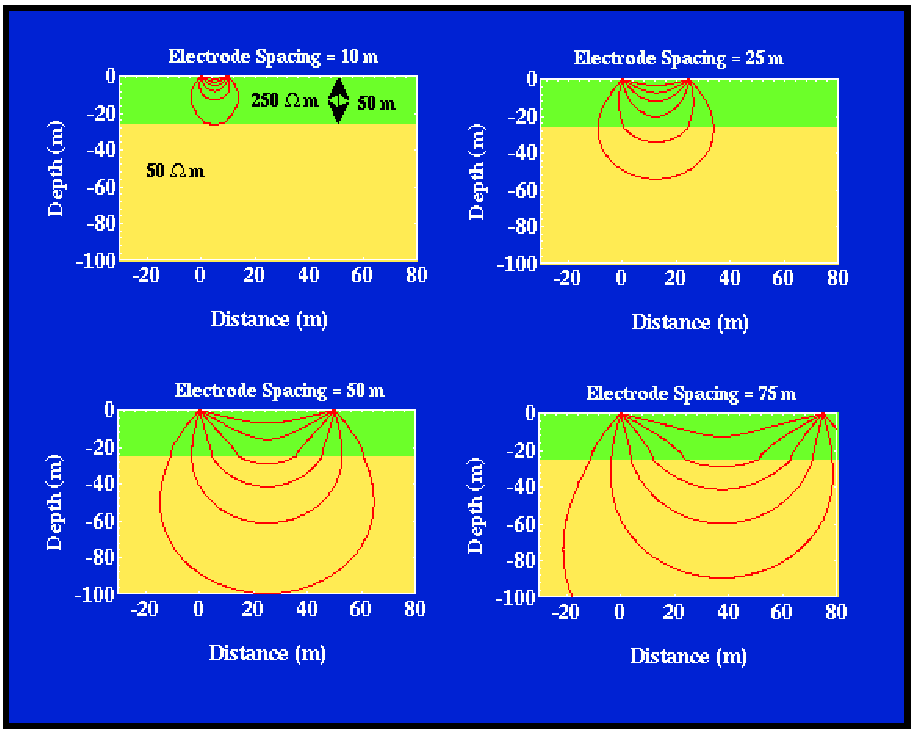


Recall, that our expression for the computation of apparent resistivity, shown below, is a function of electrode spacing r (which is the same for the two situations shown above), current i (assume that we are putting the same current in the ground for each model), and potential difference $\mathcal{l}$ (voltage) between the two potential electrodes. It can be shown that the potential difference, *l*, is proportional to the current density around the potential electrodes. Thus, for the case shown on the left the potential difference will be smaller than would have been observed in a homogeneous Earth because the current density is smaller than that which would have been observed in a homogeneous earth. Therefore, the measured apparent resistivity will be decreased. Conversely, for the case shown of the right the potential difference will be larger than that observed in a homogeneous earth and the measured apparent resistivity will likewise be larger.

Current Flow in Layered Media Versus Current Electrode Spacing

Imagine that we conduct a series of four electrode experiments, each centered about the same point. Let's assume that the potential electrodes remain centered between the current electrodes and that their separation is held fixed. Initially, the current electrodes are placed close together and we measure current and voltage from which we compute apparent resistivity. Then we perform the same experiment, but we systematically increase the current electrode spacing while holding the potential electrode spacing fixed. What will happen?

Consider the earth model shown below, a high resistivity layer over

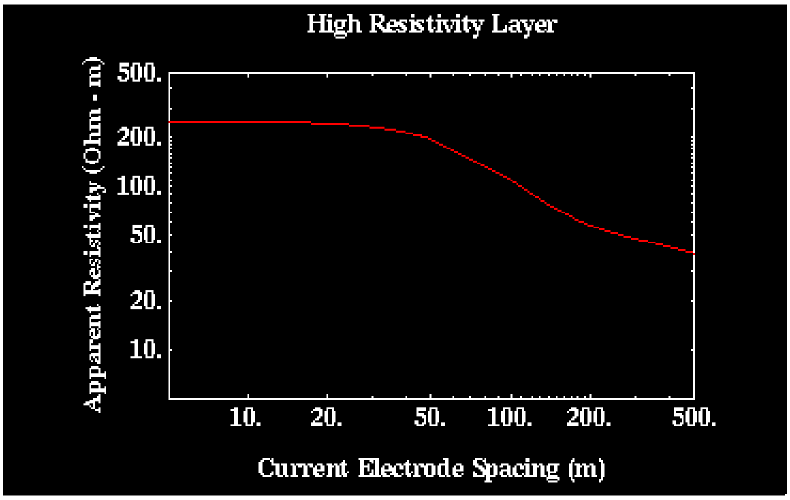


a lower resistivity layer.

When the current electrodes are closely spaced, in the region surrounding the potential electrode positions (between the two current electrodes), most of the current flows through the upper layer along paths that are close to those that they would have flown along if the model were homogeneous. That is, in this electrode configuration, current flow is not sufficiently perturbed near the potential electrodes for us to be able to distinguish between this layered model and a homogeneous Earth model with a resistivity equal to the resistivity of the top layer. Thus, the computed apparent resistivity will be close to the resistivity of the upper layer, 250 ohm.m

Now, we increase the current electrode spacing and repeat the same experiment. At larger current electrode spacings, the current flow near the potential electrodes is significantly altered by the presence of the subsurface boundary. In this case, current is preferentially drawn downward into the lower resistivity layer, decreasing the current density between the two current electrodes where we will measure the voltage with our two potential electrodes. This decrease in current density will cause our computed value of apparent resistivity do decrease from 250 ohm.m

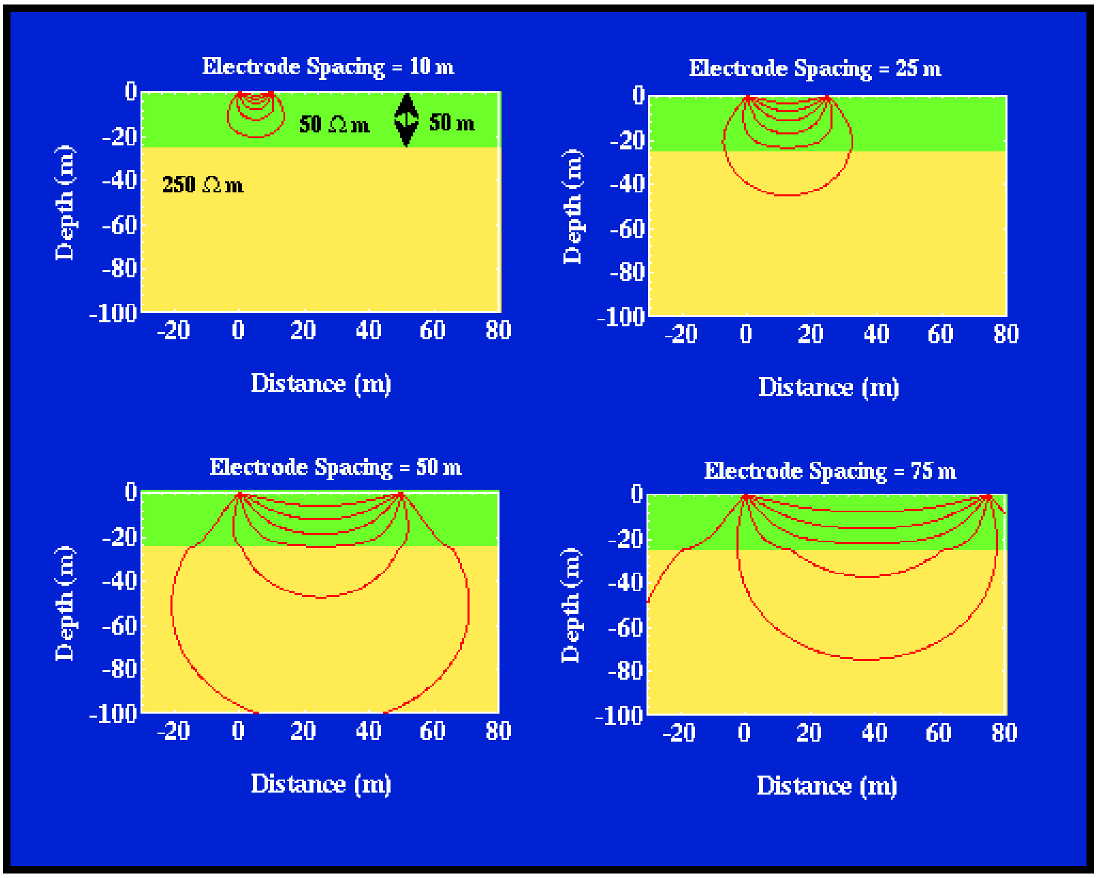
At very large current electrode spacings, underneath our potential electrodes, the pattern of current flow is again similar to that which we would observe in a homogeneous Earth model. In this case, however, the media has a resistivity of 50 ohm.m, not 250 ohm.m Thus, if we were to compute and plot apparent resistivity for a variety of current electrode spacings holding the potential electrodes fixed we would generate a plot similar to that shown below.



As is common for curves of this type, notice that this plot is a *Log-Log* plot. Instead of plotting apparent resistivity versus current electrode spacing, we have plotted the Log (base 10) of the apparent resistivity versus the Log (base 10) of the current electrode spacing. This is done because, in practice, we will find that both the apparent resistivities and the current electrode spacings will vary over two to three orders of magnitude (e.g., spacings can commonly increase from 0.25 m to 250 m). Using Log-Log plots provides us with a means of compressing the relevant information into a single graph. In the example shown above, notice that the apparent resistivity does not approach the resistivity of the lower layer until the electrode spacing approaches 500 m! Thus, large electrode spacings are required to see deep structure. A good rule of thumb is that you will need current electrode spacings on the order of 10 times the depth to which you would like to see.

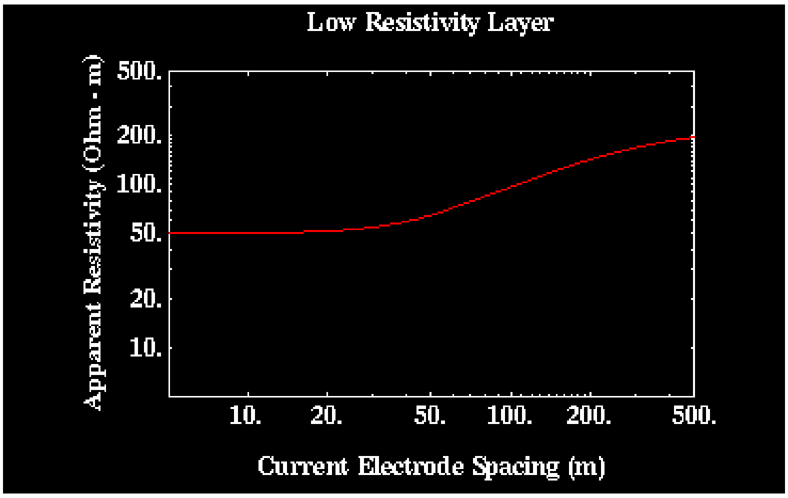
A Second Example of Current Flow in Layered Media

As another example of current flow in layered media and how apparent resistivity can vary with varying electrode spacing*, consider the earth model shown below. In this case, a low resistivity layer overlies a higher resistivity halfspace.



Initially with the current electrodes closely spaced, most of the current confined to the upper layer along paths that are very close to those that they would have assumed if the model were homogeneous. The computed apparent resistivity is very close to the resistivity of the upper layer, 50 ohm.m

At larger current electrode spacings, more current flows to greater depths. Between the two current electrodes, where are potential electrodes are located, the current flow lines become significantly distorted by the presence of the higher-resistivity layer located at depth. Therefore, around the potential electrodes the current density is larger than we would have observed in a homogeneous Earth. This relative increase in current density will cause our computed value of apparent resistivity to increase from 50 ohm.m. At very large current electrode spacings, current flow round our potential electrodes again approximate that we would observed in a homogeneous Earth. In this case, however, because most of the current is flowing through the lower media in the vicinity of the potential electrodes, the computed resistivity will be close to 250 ohm.m Thus, as current electrode spacing is increased the apparent resistivity will increase, eventually approaching 250 ohm.m A plot of apparent resistivity versus current electrode spacing is shown below.



Because current would prefer to flow within the first layer, notice that the apparent resistivity approaches the resistivity of the halfspace more slowly (i.e., with greater electrode spacings) than was found in the previous case.

*Although we have not explicitly said this, the relevant spacing in the phrase *electrode spacing* is not the spacing between the current electrodes or the spacing between the potential electrodes, but rather the spacing between the current and the potential electrodes. Thus, even if our current electrode spacing is large (so that most or our current is flowing through the lower medium), if our potential electrodes are close to the current electrodes we will compute apparent resistivities that are more like the resistivity of the upper layer than of the lower layer. In the previous example as well as in this example, we have explicitly assumed that the positions of the potential electrodes remain fixed throughout the experiment so that the distance between the potential and the current electrodes increases as the distance between the current electrodes increases. As the distance between current and potential electrodes increases, the depth over which we average resistivity structure in computing an apparent resistivity also increases.

Resistivity Equipment and Field Procedures

DC Resistivity Equipment

Compared to the equipment required for gravity surveying and magnetic surveying, that required for DC resistivity surveying . In fact, it is rather consisting of nothing more than a source of electrical current, an ammeter, a voltmeter, some cable, and electrodes. Given the nature of the that we are making, however, there are some considerations that must be given the equipment used to perform the measurements.

> *Current Source* - A source of DC current is required. In general, batteries are not capable of producing the DC currents required, so that if a pure DC source is used, it has to be produced by a portable electric generator. If, as is commonly done to eliminate the effects of electrode potentials and telluric currents a slowly varying AC current is used, portable, battery driven sources can be employed for DC resistivity surveys commonly used in engineering and environmental applications.

Ammeter - A simple ammeter (a device for measuring electrical current) can be used. The only constraint is that the meter be capable of measuring amperage from a few milliamps to about 0.5 amps. Many of the modern instruments are regulated such that the user determines the amperage input into the ground and the instrument attempts to deliver it. If the instrument can not deliver the specified amperage, usually because the electrode contact resistance is too high, the instrument warns the user.

Voltmeter - A simple voltmeter can also be used. To avoid problems with contact potential, a voltmeter with a very high impedance, above 500,000 Ohms, should be used. The voltmeter must also be capable of measuring voltages from a few millivolts to a few volts.

Electrodes - To avoid problems associated with electrode potentials, sophisticated electrodes known as porous pots can be used. But, because spurious electrode potentials can be mitigated through the use of a slowly varying AC source, these electrodes are not commonly used for DC resistivity measurements. If the conditions in the survey are extremely dry, and contact between the electrode and the ground can not be maintained, one might consider using porous pots.

For DC resistivity surveys, the most commonly used electrodes are nothing more than aluminum, copper, or steel rods about two feet in length. These rods are driven into the ground, and connected with cables to the current source or the voltmeter. Under dry conditions, contact between the rod and the ground can be enhanced by wetting the ground surrounding the electrode.

- *Cables* - To connect the electrodes to the various electrical components cables must be employed. These cables are typically nothing more than insulated wires with stranded, copper-cored conductors. Although long cable lengths may need to be employed, given the high resistivity of the ground, resistance in the cables is typically negligible. A more significant problem is current induction in the cables used to make the voltage measurement from the current flowing in the cables going to the current electrodes. This source of noise is easily avoidable by simply keeping the voltage cables at a distance (a few feet) from the current cables. For easy of deployment, cables are usually stored on reels.

Survey Types Overview: Soundings and Profiles

Thus far we have begun to see how geologically relevant structure can affect electrical current flow and measurements of voltage at the Earth's surface. We've described how depth variations in resistivity can be detected by increasing current electrode spacing by estimating apparent resistivities for various current electrode spacings. We have not, however, described the specific field procedures used in resistivity surveying.

Before describing these procedures, there is an important point to note about the geological structures considered thus far. Notice that the resistivity method represents the first method which can detect depth variations in a geologically relevant parameter. For example, if we conducted gravity or magnetic surveys at top structures that varied in density or magnetic susceptibility *only* in depth, we would observe no spatial variation in the Earth's gravity or magnetic fields. Thus, these methods are insensitive to changes in density and magnetic susceptibility that occur *solely* in depth.

***Resistivity Soundings* - As we've already shown**,

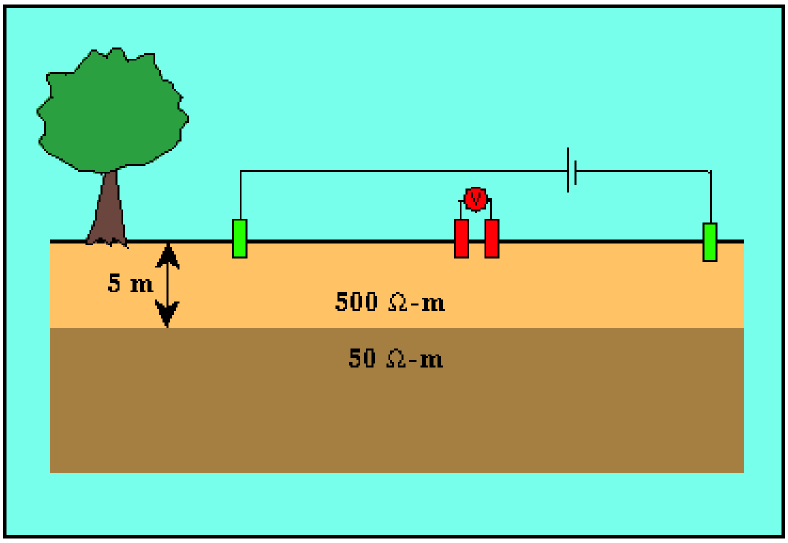
the resistivity method can detect variations in resistivity that occur solely in depth. In fact, this method is most commonly applied to look for variations in resistivity with depth. Surveys that are designed to determine resistivity variations with depth above some fixed surface location are referred to as *resistivity soundings*. In principle, the two-electrode experiments described previously are examples of soundings. In these experiments, electrode spacing is varied for each measurement. The center of the electrode array, where the electrical potential is measured, however, remains fixed. An example of a problem for which one might employ resistivity soundings is the determination of the depth to water table.

Resistivity Profiles - Like the gravity and magnetic methods, resistivity surveys can also be employed to detect lateral variations in resistivity. Unlike soundings, profiles employ fixed electrode spacings The center of the electrode spread is moved for each reading. These experiments thus provide estimates of the spatial variation in resistivity at some fixed electrode spacing. Surveys that are designed to locate lateral variations in resistivity are referred to as *resistivity profiles*. An example of a problem for which one might employ resistivity profiles is the location of a vertical fault.

Electrode Spacings and Apparent Resistivity Plots

You may have noticed on the previously shown plots of apparent resistivity that the data were plotted on *log-log* plots rather than the more traditional linear-linear plots. You should also notice that the electrode distances shown on these plots are evenly spaced in *log distance* rather than being evenly spaced in linear distance. Why have we chosen to acquire and display the data in this fashion?

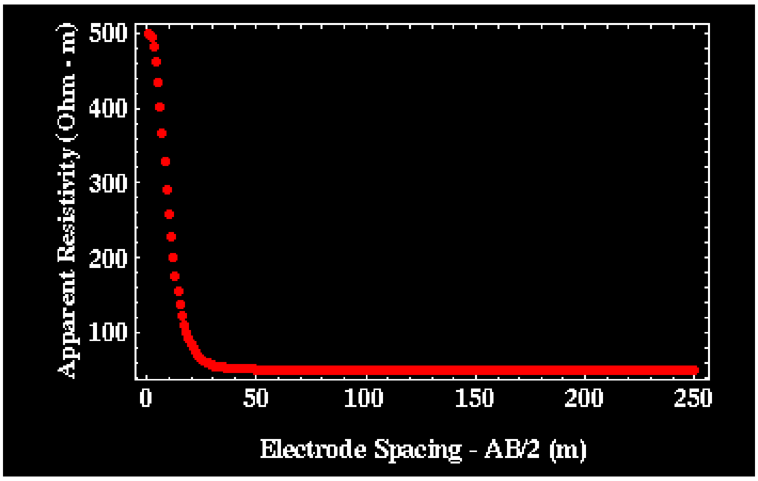
Consider performing a Schlumberger sounding over the geological model shown below.



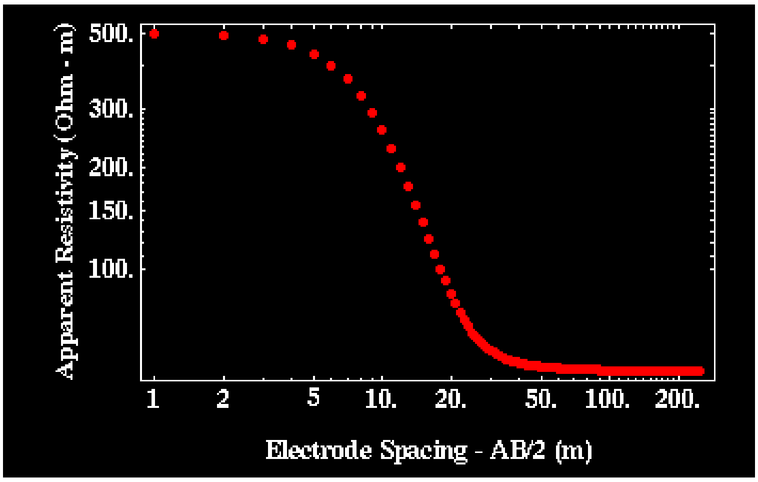
Let's do our Schlumberger sounding by varying current electrode spacing, *AB/2*, from 1 to 250 meters at 1 meter increments. Shown below is a plot of the resulting apparent resistivity versus electrode spacing.

We know that for small electrode spacings the apparent resistivity should approximate the resistivity of the layer. As the electrode spacing increases, the apparent resistivity should approach the resistivity of the halfspace.

These are the features that are shown in the plot. They are not, however, emphasized in this plot.

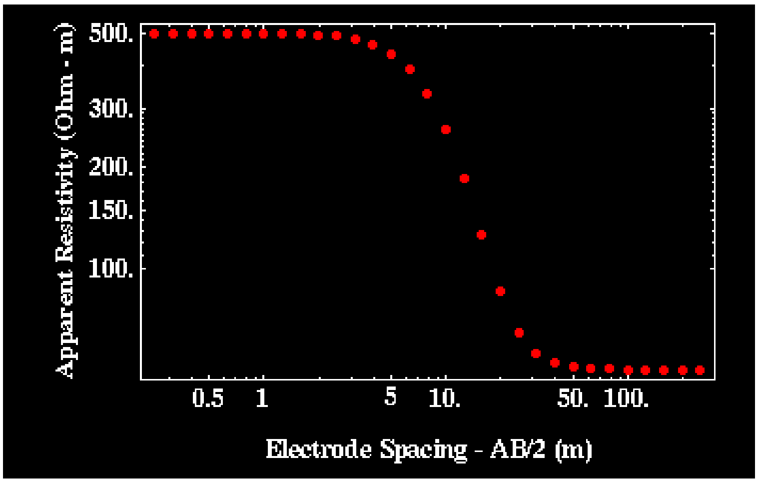


Most of the interesting features of this apparent resistivity curve occur at electrode spacings smaller than 50 meters. When looking at this apparent resistivity curve, because the plot includes so much data at electrode spacings larger than 50 meters, it de-emphasizes the important data at the smaller electrode spacings. One way to help bring out the information content at both the smaller and longer electrode spacings is to plot the same data on a log scale rather than a linear scale. A log-log plot the same data is shown below. Notice how the smaller electrode spacings now occupy more of the plot, thus making it easier to extract important information about how the apparent resistivity varies with electrode spacing.



Although this plot is better, there is still one problem related to how the data were acquired. Notice that there are only a few readings made at the small electrode spacings that are approximately equal to 500 ohm.m while there are many at the larger electrode spacings that are approximately equal to 50 ohm.m We would like more readings at the smaller electrode spacings so that we can be assured that the apparent resistivities plotted are representative of the near-surface resistivity. This could be done at the cost of taking fewer readings at the larger electrode spacings. By reallocating the electrode spacings that we will use, we will not only acquire observations that are more relevant, but we could also speed up our field acquisition by eliminating those readings that do not contain new information.

For electrical soundings, electrode spacings commonly are chosen so that they are evenly spaced in log distance rather than being evenly spaced in distance to address the problem described above. Shown below is a plot of log apparent resistivity versus log electrode spacing, where the distance interval is now chosen to be evenly spaced in log distance rather than distance. Now there are approximately as many samples showing apparent resistivities of 500 ohm.m as there are of 50 ohm.m In addition, the transition between these two extremes is also well sampled. Up to here

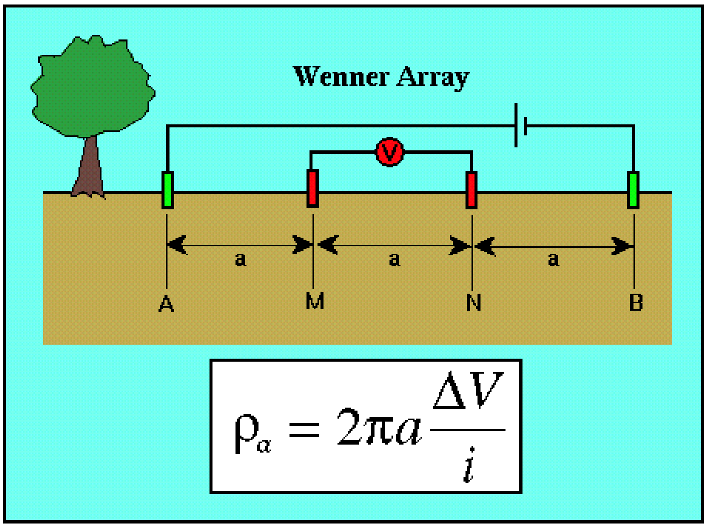


In the example shown above, we acquired the data so that there are 9 soundings for every decade (power of 10) in distance beginning at 0.25 meters. Thus, soundings were taken at 0.25, 0.5, 0.75, 1.0, 1.25, 1.5, 1.75, 2.0, 2.25, 5.0, 7.5, 10.0, 12.5, 15.0, 17.5, 20.0, 22.5, 50.0, 75.0, 100.0, 125.0, 150.0, 175.0, 200.0, 250.0 meter current electrode separations, *AB/2*.

This example clearly shows that using a log-distance scheme to acquire electrical data provides information at the densities required over all distance ranges. For most surveys, acquiring 9 readings per decade of distance is not necessary. The most common electrode spacing used is one that employs 6 soundings for every decade in distance. For this example, using six points per decade would yield electrode spacings of 0.25, 3.67, 5.39, 7.91, 1.16, 1.70, 2.5, 3.67, 5.39, 7.91, 11.6, 17.0, 25.0, 36.7, 53.9, 79.1, 116.0, 170.0, 250.0.

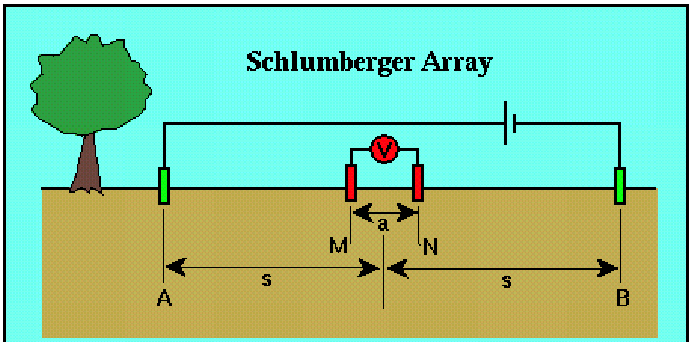
Resistivity Soundings

When doing resistivity sounding surveys, one of two survey types is most commonly used. For both of these surveys, electrodes are distributed along a line, centered about a midpoint that is considered the location of the sounding. The simplest in terms of the geometry of electrode placement is referred to as a *Wenner* survey. The most time-effective in terms of field work is referred to as a *Schlumberger* survey. For a Wenner survey, the two current electrodes (green) and the two potential electrodes (red) are placed in line with each other, equidistant from one another, and centered on some location as shown below.



The apparent resistivity computed from measurements of voltage, ℓ, and current, i, is given by the relatively simple equation shown above. This equation is nothing more than the apparent resistivity expression shown previously with the electrode distances fixed to a. To generate a plot of apparent resistivity versus electrode spacing, from which we could interpret the resistivity variation with depth, we would have to compute apparent resistivity for a variety of electrode spacings, a. That is, after making a measurement we would have to move all four electrodes to new positions.

For a Schlumberger survey, the two current electrodes (green) and the two potential electrodes (red) are still placed in line with other, centered on some location, but the potential and current electrodes are not placed equidistant from one another.



$$\rho_a = \frac{\pi(s^2 - a^2/4)}{a}\frac{\Delta V}{i}$$

The current electrodes are at equal distances from the center of the sounding, s. The potential electrodes are also at equal distances from the center of the sounding, but this distance, $a/2$, is much less than the distance s. Most of the interpretation software available assumes that the potential electrode spacing is negligible compared to the current electrode spacing. In practice, this is usually interpreted as meaning that a must be less than $2s/5$.

In principle, this implies that we could set a to be less than $2s/5$ for the smallest value of s we will use in the survey and never move the potential electrodes again. In practice, however, as the current electrodes are moved outward, the potential difference between the two potential electrodes gets smaller. Eventually this difference becomes smaller than our voltmeter is capable of reading, and we will need to increase *a* to increase the potential difference we are attempting to measure.

Advantages and Disadvantages of Wenner and Schlumberger Arrays

The following table lists some of the strengths and weaknesses of Schlumberger and Wenner sounding methods.

Schlumberger		Wenner	
Advantage	**Disadvantage**	**Advantage**	**Disadvantage**
Need to move the two potential electrodes only for most readings. This can significantly decrease the time required to acquire a sounding.			**All four electrodes, two current and two potential must be moved to acquire each reading.**
	Because the potential electrode spacing is small compared to the current electrode spacing, for large current electrode spacings very sensitive	**Potential electrode spacing increases as current electrode spacing increases. Less sensitive voltmeters are**	

	voltmeters are required.	**required.**	
Because the potential electrodes remain in fixed location, the effects of near-surface lateral variations in resistivity are reduced.			**Because all electrodes are moved for each reading, this method can be more susceptible to near-surface, lateral, variations in resistivity. These near-surface lateral variations could potentially be misinterpreted in terms of depth variations in resistivity.**
	In general, interpretations based on DC soundings will be limited to simple, horizontally layered structures.		**In general, interpretations based on DC soundings will be limited to simple, horizontally layered structures**