Earthquake Seismology

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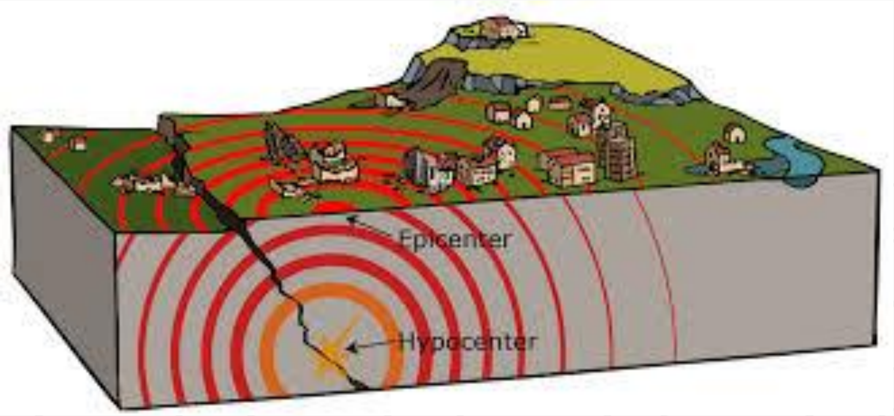
Introduction

- Definitions
- Type of seismic waves
- Seismic waves and seismogram components
- Elastic Rebound theory
- Earthquake distributions and plate tectonics



What is Seismology?

Seismology is the study of the generation, propagation and recording of elastic waves in the Earth (and other celestial bodies) and of the sources that produce them.

What is earthquake?

An earthquake is the result of a sudden release of energy in the earth’s crust that creates seismic waves

Seismology

- ***Earthquake:*** an event of ground shaking usually caused by the rupturing of a fault within the Earth.

- Who studies Earthquakes?
  - Seismologists
    - waves
  - Geophysicists
    - Mechanics
    - Geodesy
  - Geologists
    - Structures
    - Paleoseismology

Causes of earthquake

The primary cause of an earthquake is faults that occur in the earth's crust in response to stress

The fault is a break of fracture that represents by a plane separates two block of rocks (hang and foot walls)

The movement along the fault plane may occur rapidly in the form of **earthquake** or slowly in the form of **creep**.

Major causes of earthquakes

Tectonic causes

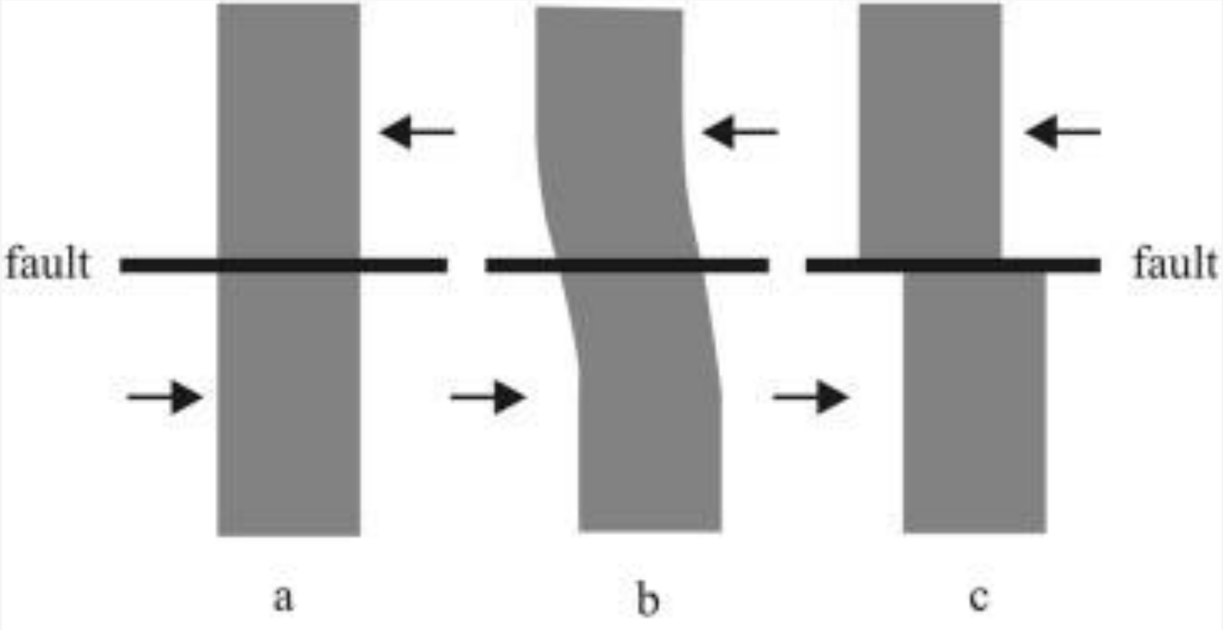
Volcanic causes

Surface causes

Elastic rebound theory

The elastic rebound theory is the theory used by geologists to explain the mechanism of earthquake occurrences (how do earthquakes occur). The theory explains the mechanism of how energy is stored in rocks:

- Rocks bend until the strength of the rock is exceeded
- Rupture occurs and the rocks quickly rebound to an undeformed shape
- Energy is released in the form of seismic waves that radiate outward from the hypocentre

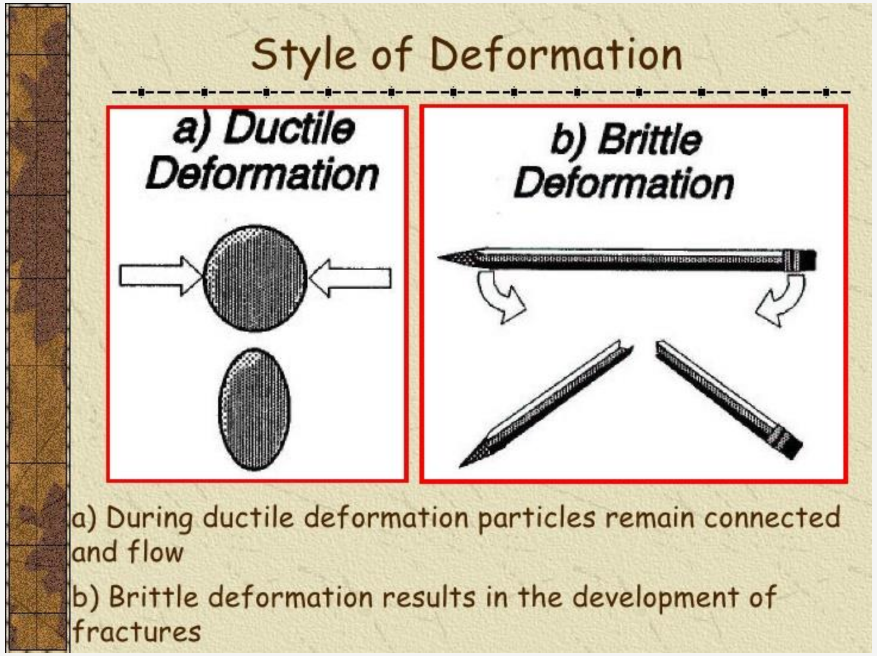


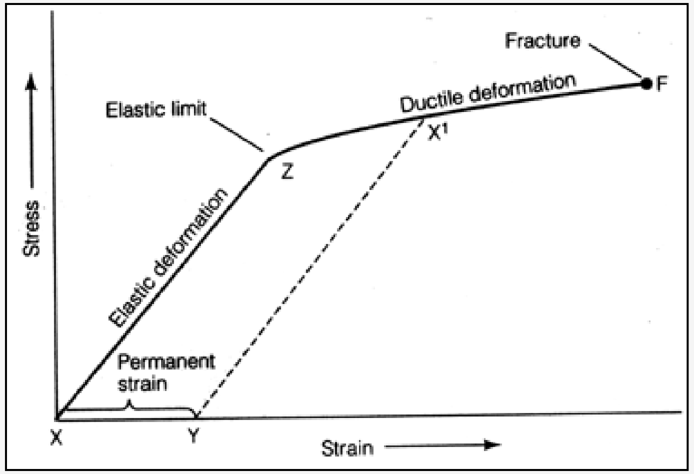
a–No slip potential accumulation

b– Interseismic slip potential accumulation (fault locking)

c– Coseismic slip potential release

Style of deformation





During ductile deformation particles remain connected and flow

Brittle deformation results in the development of fractures

What is Brittle?

Brittle is the opposite of ductile.

Brittle describes a material property of rock.

A brittle rock fractures when it is forced to change its shape and deform only a small amount.

Brittle can also describe the way rock deforms under pressure.

Fracturing, including earthquake faulting, is a brittle form or deformation.

Example, lithosphere

What is ductility?

Ductility is toughness, the ability to deform permanently without breaking.

A ductile material can stretch, compress, or distort inelastically.

Ductile rock, such as rock heated to a high temperature in the interior of Earth slowly flows rather than breaks; does not suddenly crack under load as brittle rock does.

Example; athenosphere

Definitions should be memorized before starting our course

Seismology	A branch of science focused on the study of earthquakes, seismic activity, and seismic waves
Earthquake	A sudden rupture of Earth’s lithosphere caused by the release of stress accumulated along geologic faults or by volcanic activity, and released seismic waves.
Seismic waves	Waves of energy transfer by the way of particle motions caused either by earthquakes, by massive man-made explosions or volcanos. Seismic waves travel through and on top of the surface of Earth causing the shaking and vibrations on the ground. Earthquake waves can travel hundreds of kilometers causing shaking to be felt a long way away from the origin.
Focus or hypocentre	The focus or hypocentre of an earthquake is where the earthquake originated from, usually underground on the fault zone (subsurface).
Epicentre	The epicentre of an earthquake is the point on the surface of Earth directly above the ~~epicentre~~ hypocentre
Seismometer	An instrument that detects the seismic waves released from earthquakes and/or other ground movements such as explosions.
Seismograph	The device takes the readings produced by a Seismometer and produces a Seismogram
Seismogram	A graph demonstrates seismic waves that looks like a squiggly line

Definitions should be memorized before starting our course

Foreshocks	Foreshocks are earthquakes of smaller sizes occur in the same area as a larger earthquake (mainshock) that follows. Not all earthquakes have foreshocks or aftershocks.
Aftershocks	Aftershocks are smaller earthquakes that may occur after the mainshock in the same area. They are caused by the area readjusting to the fault movement, and some may be the result of continuing movement along the same fault zone
Earthquake swarm	Sometimes a series of similar sized earthquakes, called an earthquake swarm, happens over months without being followed by a significantly larger earthquake.
Magnitude	Magnitude is a scale used to describe the earthquake size. There are a number of magnitude scale, including the Richter Scale. In Australia, seismologists prefer the use of the moment magnitude scale, which calculates the magnitude of an earthquake based on physical properties such as the area of movement (slip) along the fault plane.
Intensity	Intensity is a scale measures how much damage was done during an earthquake. It is based on what people in the area felt by seismic waves.
Fault plane	A fault is a weak point within a tectonic plate where pressure from beneath the surface can break through and causing shaking in an earthquake.
Lithosphere	All natural earthquakes take place in the lithosphere. The lithosphere refers to the portion of up to 100 kilometers from the surface of the earth.

Summary

Seismology is the study of generation, propagation and recording of seismic waves in the Earth

An earthquake is the result of a sudden release of energy in the earth's crust that creates seismic waves

The movement along the fault plane may occur rapidly in the form of **earthquake (seismic deformation, brittle deformation, elastic materials)** or slowly in the form of **creep (aseismic deformation, ductile deformation, inelastic materials)**.

A brittle rock fractures when it is forced to change its shape and deform only a small amount.

Ductility is toughness, the ability to deform permanently without breaking

Brittle deformation occurs when elastic materials are forced

Ductile deformation occurs when anelatic (plastic) materials are forced

An example of brittle deformation: Earthquakes are associated with brittle deformation in lithosphere

An example of ductile deformation: Creeping is associated with ductile deformation in asthenosphere

Seismic Waves

- Stress and strain
  - Elastic constants
  - Body waves and surface waves
  - Seismic wave velocities
  - Attenuation
- Seismic rays
  - Snell's law
  - Reflection, transmission
  - Refraction, Diffraction
- Seismic sources
- Seismic recording systems

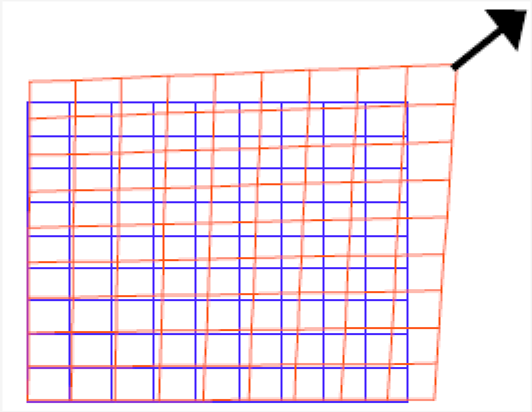
Questions?

- Why do we observe waves in the Earth?
- What kind of waves are there?
- How fast do they propagate?
- What determines their speed?
- Do seismic waves vary with rock type?
- Are seismic waves attenuated?
- What waves do we use in seismic exploration?
- How are seismic waves generated (on land, at sea)?
- With what instruments can we observe seismic waves?

Stress & Strain

The earth's crust deforms like an elastic body when the deformation (strain) is small

In other words, if the force that causes the deformation is stopped the rock will go back to its original form.



The change in shape or volume (i.e., the deformation) is called strain, the forces that cause this strain are called stresses.

Quantifying Deformation: Stress & Strain

A simplistic view…

Stress = force/area

So both force and area of contact are important

Stress [=] Pascals [=] $Kg \cdot m^{-1} \cdot s^{-2}$

Types: tension, compression, shear

$\text{Strain} = \Delta l / l_0$

Measures change in size/shape (i.e. deformation)

Dimensionless (i.e. it is a percent).

Some may say, e.g., 2.5 µstrains or 3.1 nstrains

Types: extension/dilatation, contraction, shear

Most scientists agree that stress causes strain

Chicken and egg argument

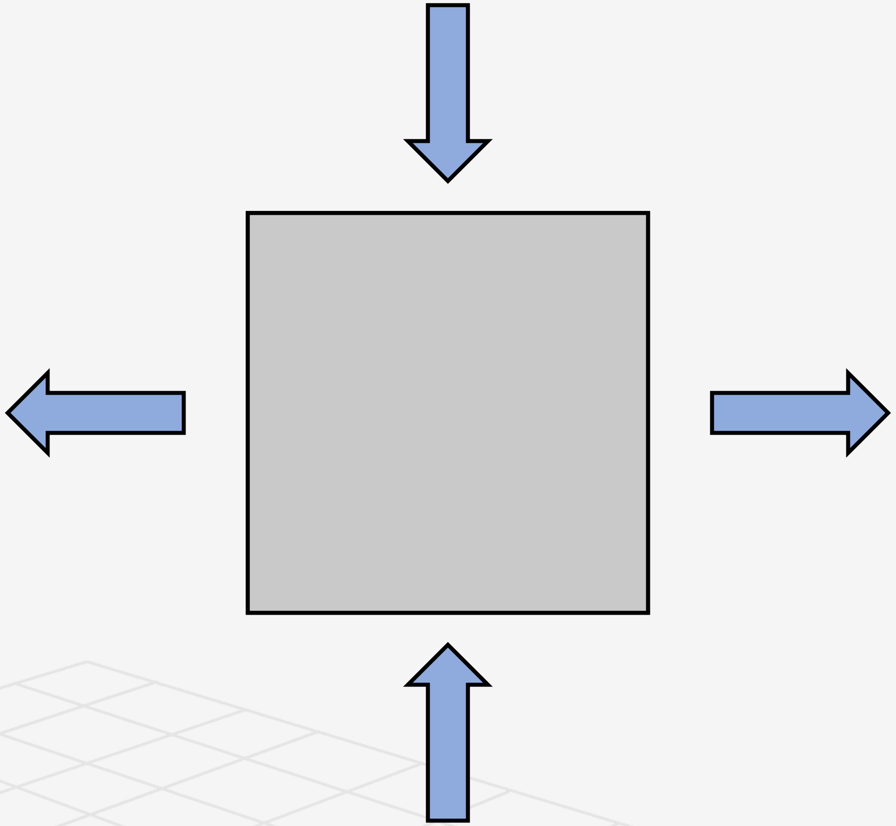
Normal Stress / Normal Strain

A normal stress acts perpendicular to the applied surface.

A normal strain results from a normal stress

Deforms a square into a rectangle

- Angles between sides remain unchanged



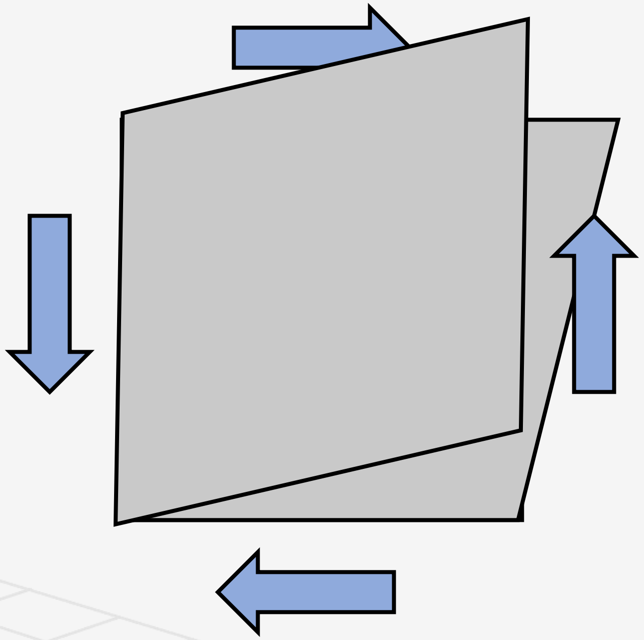
Shear Stress / Normal Strain

A shear stress acts parallel to the applied surface.

A normal strain results from a normal stress

Deforms a square into a lozenge.

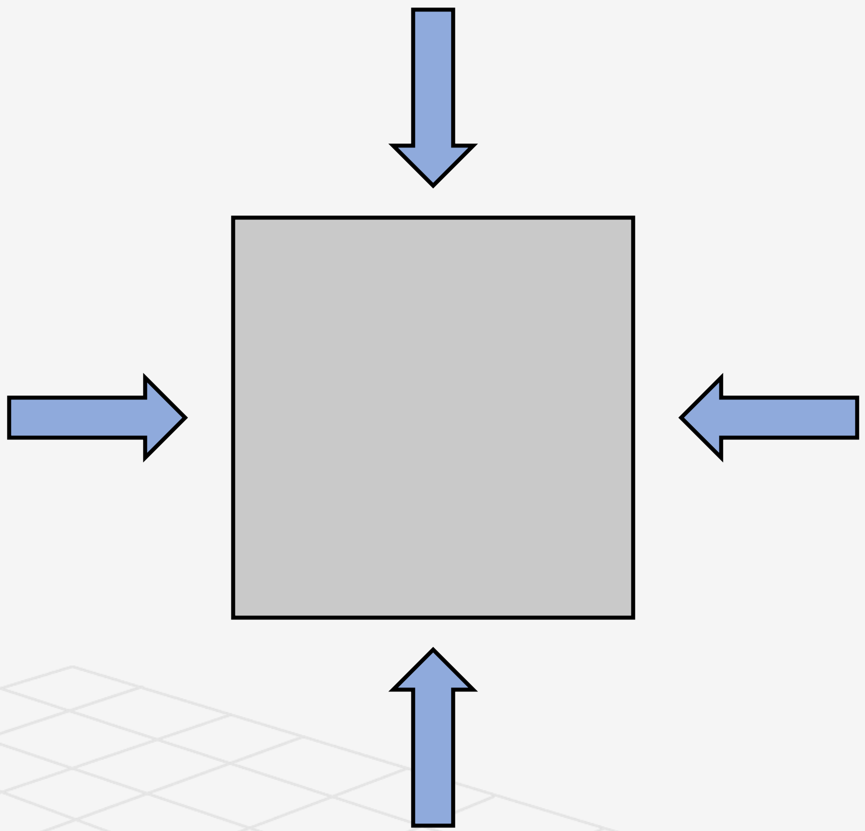
- Angles between sides change.



Pressure

Pressure is a special state of stress where the stresses are equal in all directions

- i.e. all normal stresses
- Zero shear stress
- Occurs in fluids (liquids/gasses)
  - E.g. air pressure
- Causes only volumetric changes



Elastic Constants

The elastic constants link the stresses and the strains (think of the spring constants in the 1D problem)

Stress = Elastic constants * strain

F = D * s   Hooke's Law

Elastic constants

The elastic constants describe how a material deforms when it's stressed. Unfortunately there are many different approaches. The most important are (compare with previous slide):

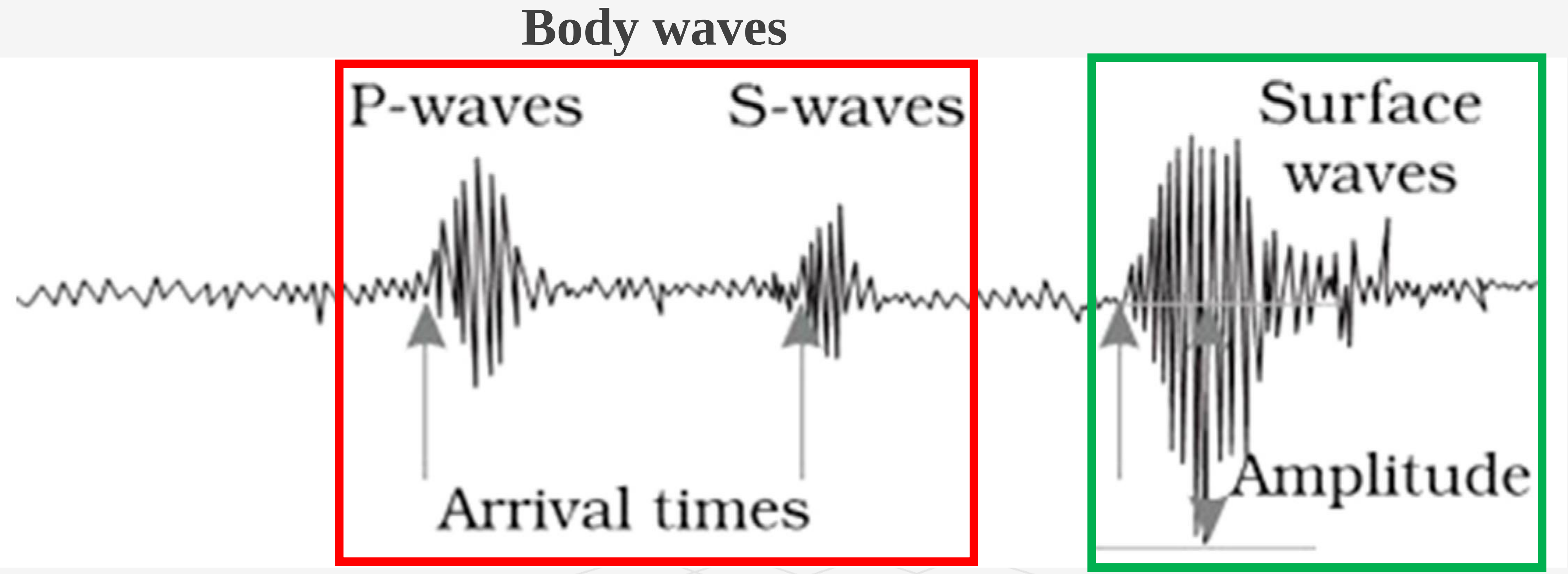
Young's modulus $E = \dfrac{\text{longitudinal stress } F/A}{\text{longitudinal strain } \Delta l/l}$

Bulk modulus $K = \dfrac{\text{volume stress } P}{\text{volume strain } \Delta V/V}$

Shear modulus $\mu = \dfrac{\text{shear stress}}{\text{shear strain } (\tan \varphi)}$

Others: Lame' parameters, Poisson's ratio, etc.

Type of seismic waves



How seismic movement is produced earthquake and seismic waves

The motion we fell on the surface of the earth during an earthquake comes from energy released deep with the earth. This energy is transmitted to the surface by seismic waves. The study of earthquake and seismic energy is called seismology.

Earthquake occur when rocks deep suddenly break under pressure of slip along a fault. The point of release is known as the focus (hypocentre) of the earthquake. The energy released by the earthquake radiates from the focus as body waves.

Body waves

There are two types of body waves.

The **first type** is called a **P wave** or **primary wave**, the fastest wave, that moves between four to seven kilometers per second depending on the elastic constants of the rock it's moving through. The P wave is a compression type wave. Rock materials in its direction of propagation compress then expand as the wave passes. The P wave is similar to a wave travelling through a spring. The coils compress and expand in the direction of the wave is travelling.

The **second type** of body wave is called a **S wave** or **secondary wave**. It travels about two to five kilometers per second through rock about half the speed of a P wave. The S wave is a transfer shear wave. Rock materials in its path move up and down or side to side perpendicular to the direction of the waves travel. The S wave is similar to the wave traveling along a piece of rope. The wave moves along the rope by moving a section of the rope up and then down.

Surface Waves

Surface waves radiate outward from the epicentre (the point on the surface above the focus). Surface waves are slower than body waves. They traveling at two to three kilometers per second. The can change the surface of the earth as well as damage resident buildings and other structures. **These waves are more destructive and damaging**.

There are two types of surface waves. The **first type** is the **Love waves** that cause side-to-side motion perpendicular to the direction of wave travel. The Love wave can cause damage by breaking roads and pipes. The **second type** of surface wave is called a **Raleigh wave**. The Raleigh waves move the surface of the earth up forward down and back in a circle. They can cause damage by knocking buildings off their foundations.

In most earthquake, combinations of love and Raleigh waves caused the most destruction because the ground shakes up and down and side-to-side at the same time.

How Surface waves generate?

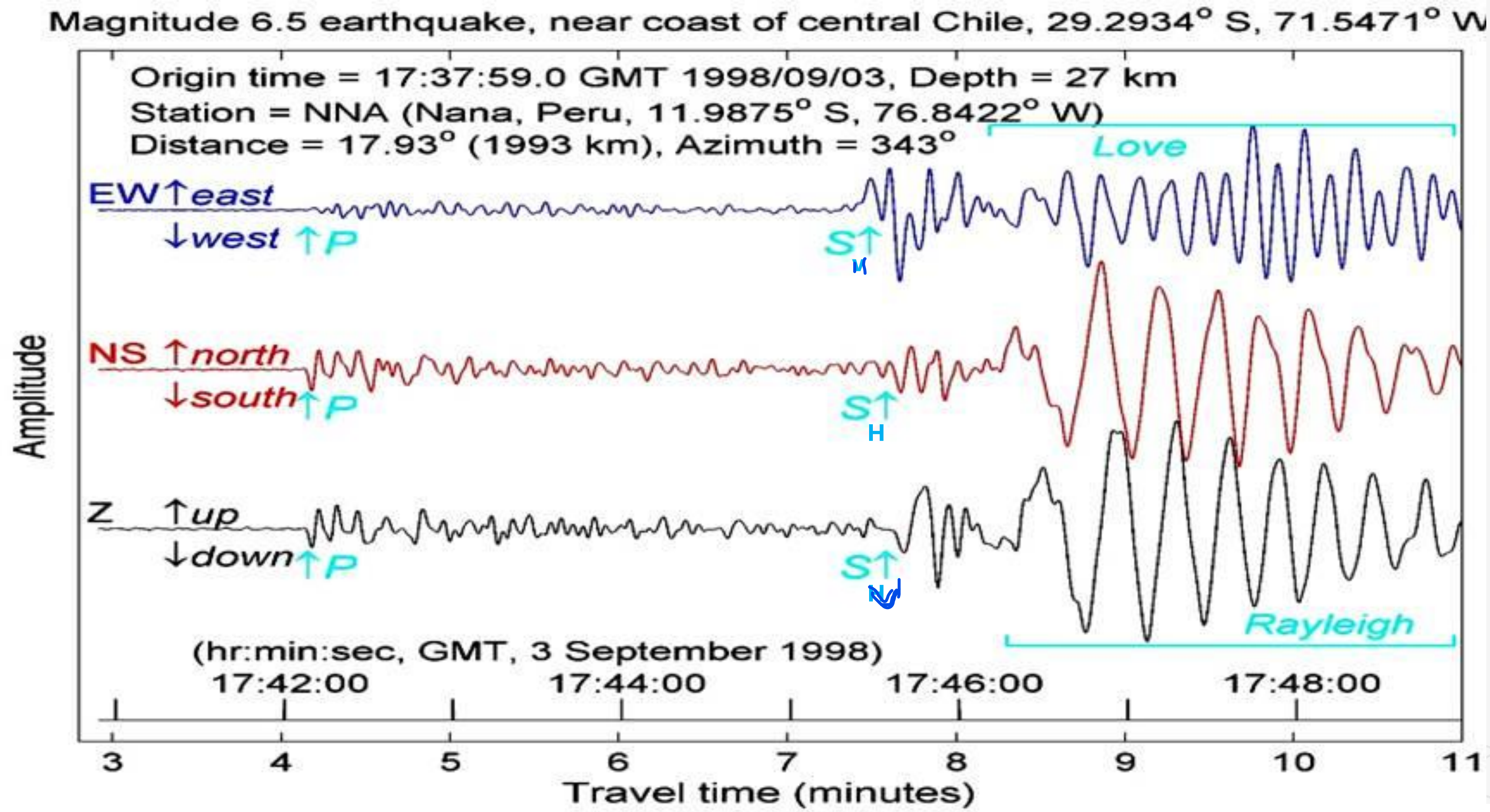
How to distinguish between Raleigh and Love waves

Using seismograms, you can distinguish between the Raleigh waves and Love waves where Raleigh waves recorded on the vertical component of seismograms while the Love waves recorded on the horizontal components of seismograms (see the previous slide).

Vertical component (Z) ➔ P + Sv + Raleigh waves

Horizontal components (EW and/or NS) ➔ P + SH + Love waves

Seismic waves and seismogram components



Questions

- At a seismic station the first waves to arrive are _________________.
- At a seismic station the second waves to arrive are __________________.
- At a seismic station the last waves to arrive are ______________________.
- How would you distinguish between Love waves and Rayleigh waves if you were given an earthquake record from a three-component seismic station?
- Which type of seismic waves can be detected by a vertical component seismometer?
- Which type of seismic waves can be detected by a horizontal component seismometer?
- How seismic movement is produced earthquake and seismic waves?
- What is the elastic rebound theory?

Seismic wave velocities
P-waves

Material	V_p (km/s)
Unconsolidated material	
Sand (dry)	0.2-1.0
Sand (wet)	1.5-2.0
Sediments	
Sandstones	2.0-6.0
Limestones	2.0-6.0
Igneous rocks	
Granite	5.5-6.0
Gabbro	6.5-8.5
Pore fluids	
Air	0.3
Water	1.4-1.5
Oil	1.3-1.4
Other material	
Steel	6.1
Concrete	3.6

Seismic wave velocities
shear waves

The relation between P-waves velocities and shear wave velocities is often described by the v_P/v_s ratio or Poisson's ratio.

A commonly used assumption for crustal rocks is:

$v_P/v_s = \mathrm{sqrt}(3) \approx 1.7$

This corresponds to the Poisson ratio σ

$\sigma = 0.25$

With the relation:

$$\frac{v_p}{v_s} = \left[\frac{2(1-\sigma)}{(1-2\sigma)}\right]^{1/2}$$

Fluids or gas in rocks strongly influence the v_P/v_s ratio that is one of the most important diagnostics in seismic exploration!

Seismic velocities and density
Porosity

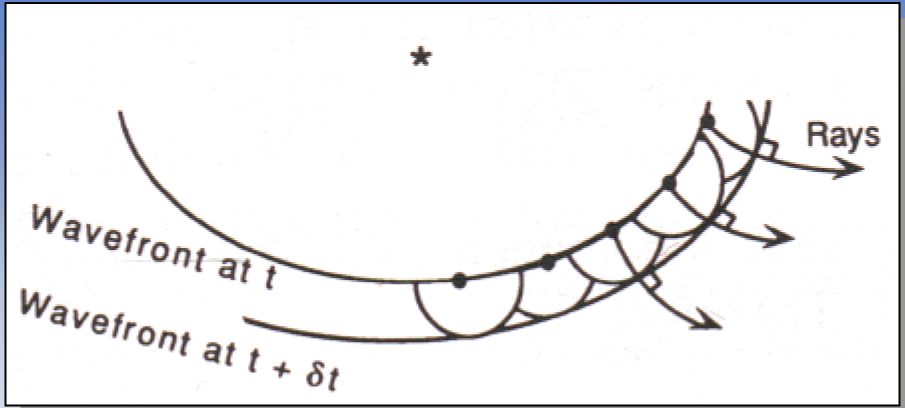
We want to quantify the effect of porosity Φ on the seismic wave velocity and density. With ρ_b the bulk density, ρ_f the pore fluid density and ρ_m the rock matrix density:

$$\rho_b = \rho_f \Phi + (1-\Phi)\rho_m$$

... and a corresponding relation exists for P-velocity

$$\frac{1}{v_b} = \frac{\Phi}{v_f} + \frac{(1-\Phi)}{v_m}$$

Seismic rays



Huygens principle states that each point on the wavefront serves as a secondary source. The tangent surface of the expanding waves gives the wavefront at later times. Rays are the trajectories perpendicular to the wavefronts.

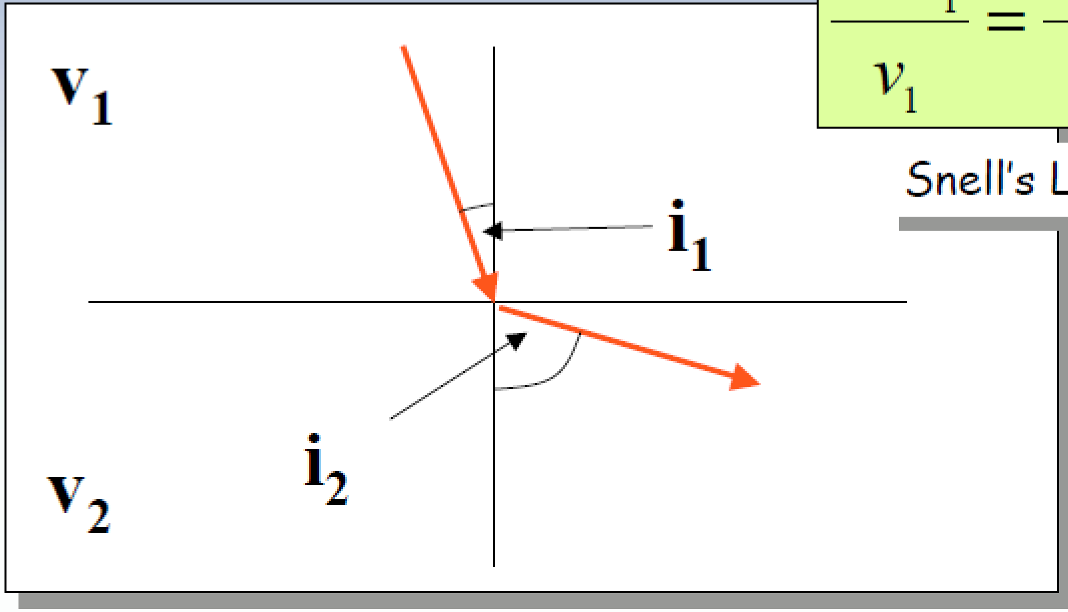
Fermat's principle and Snell's law
ray transmission

Fermat's principle governs the geometry of the ray path. The ray will follow *a minimum-time* path. From Fermat's principle follows directly Snell's Law

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

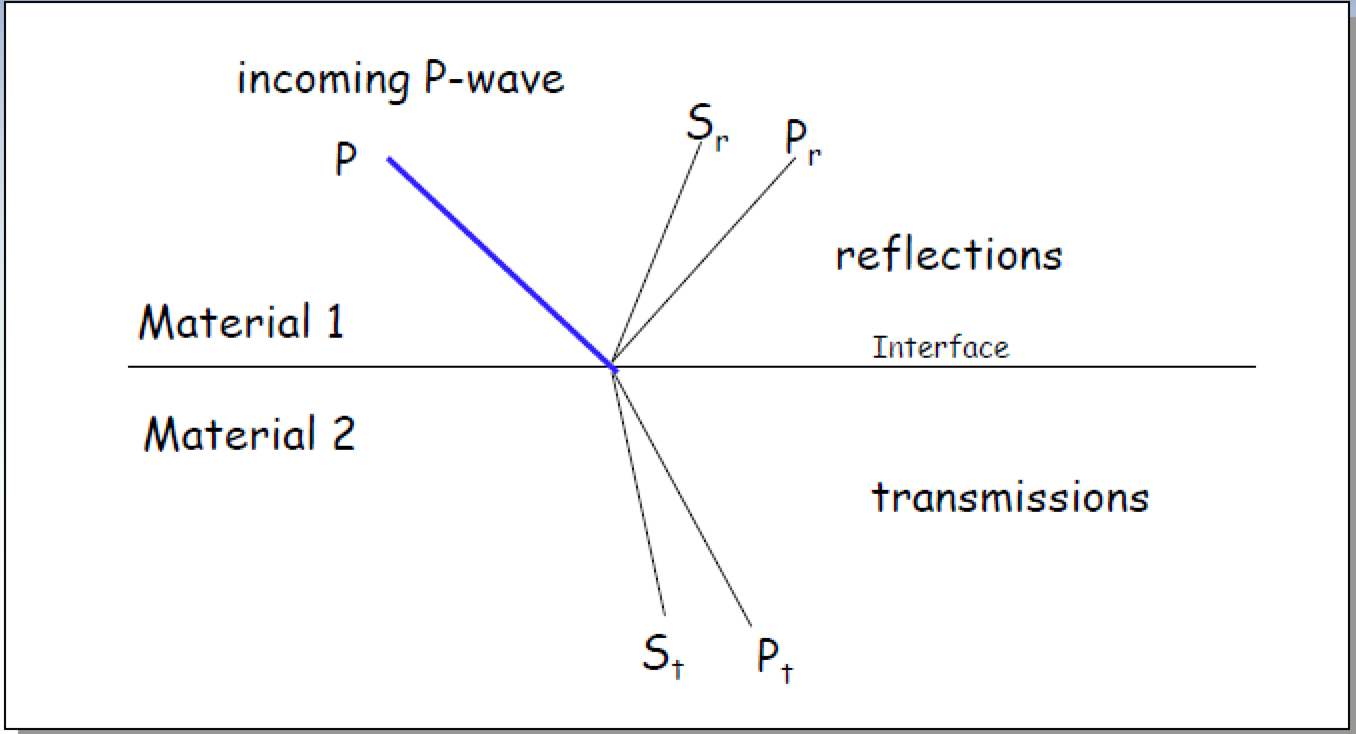
Snell's Law

$V_2 > V_1$

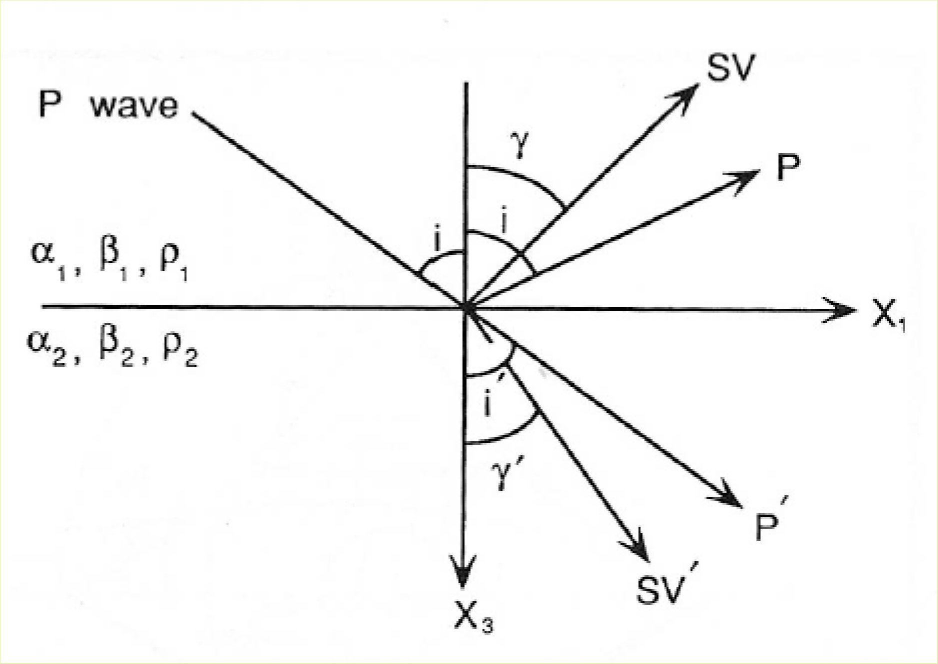


Reflection and transmission at boundaries
oblique incidence - conversion

P waves can be converted to S waves and vice versa. This creates a quite complex behavior of wave amplitudes and wave forms at interfaces. This behavior can be used to constrain the properties of the material interface.



Mode conversions and partitioning of seismic energy at a boundary



- S waves can be decomposed to SV (the motion within the vertical plane) and SH components (the motion perpendicular to the vertical plane)

- P can be transformed to SV and vice versa

Boundary condition

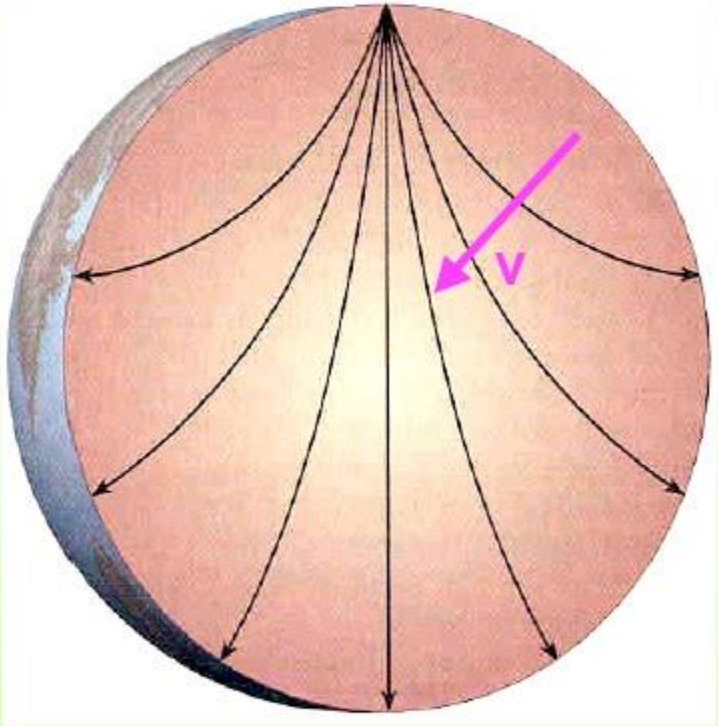
- Wave propagation:
  Stresses and displacements must be transmitted across the interface

- Free surface: all the stresses must be zero; no restrictions on displacements
- Solid-solid interface: all the stresses and displacements must be continuous
- Solid-liquid interface: tangential stresses must vanish; normal displacements and normal stresses must be continuous

Why do seismic waves travel as a curving path not as a straight path through the Earth ?

Seismic rays

- The nature of seismic waves
  - wave paths are “bent” when going deeper in Earth
  - P and S-wave seismic velocities generally increase with depth



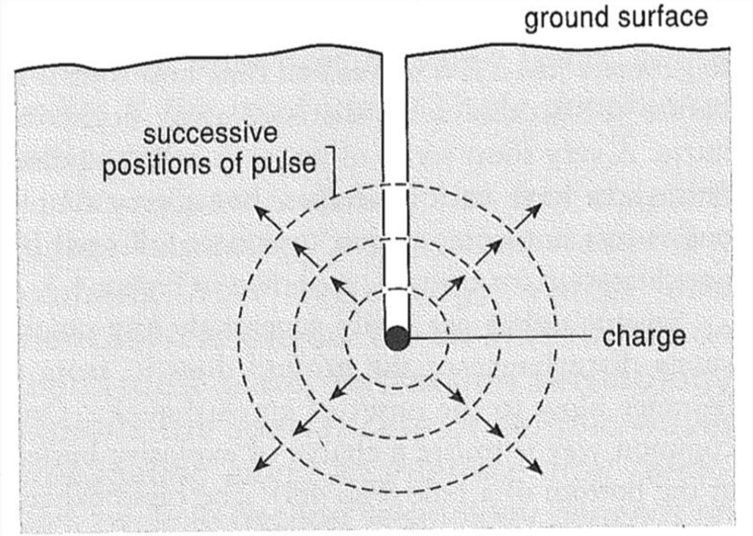
Seismic Rays

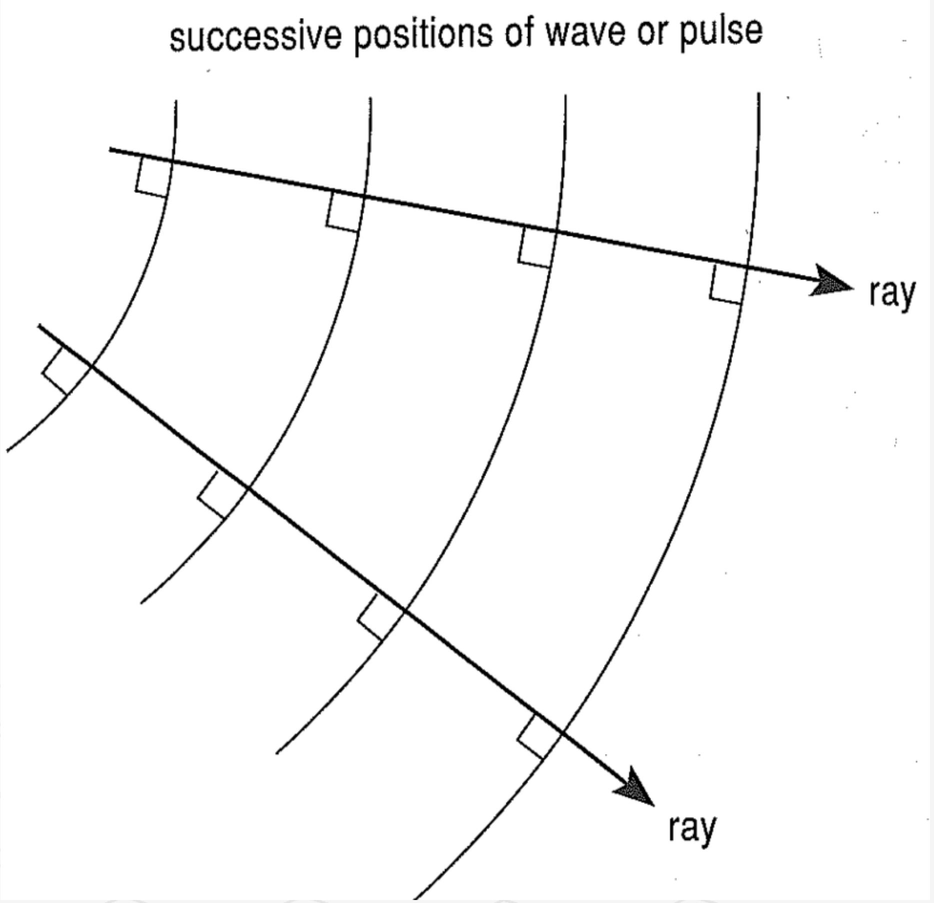
- Discuss in class

Seismic Rays

Ray: The path of a tiny portion of the wave front

- Perpendicular to wave front
- Easier to understand
- Most of seismology involves rays, although we understand that wave fronts are what is really occurring.





Do Seismic Velocities Vary with Depth?

To test this question:

Look at arrival times from stations at varying distances from each other

If velocity is uniform with depth…

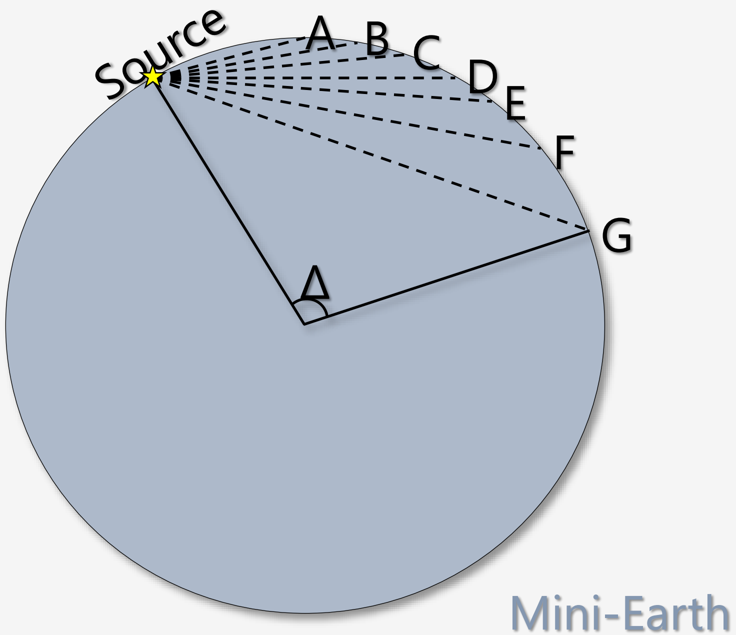
Distance will be linearly related to time; waves will be traveling in straight lines

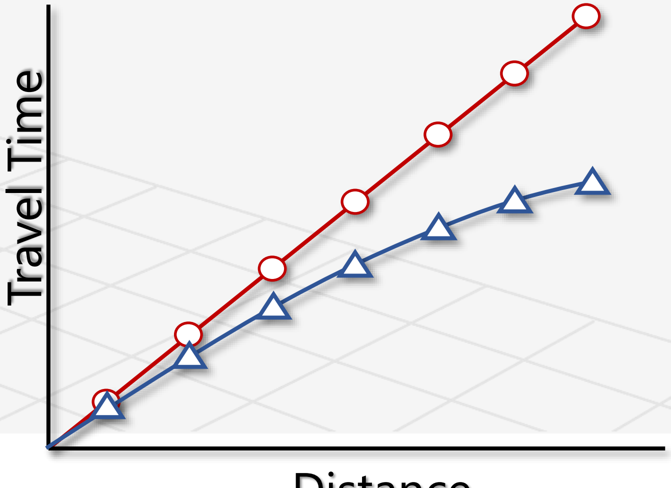
Global measurements show…

Travel times are not linearly related to distance

Velocity is not uniform with depth; waves do not travel in straight lines

Velocity must increase with depth





Finding Ray Paths :: Refraction

Just like light, seismic rays refract, or bend, when they encounter a medium of different seismic velocity

Refraction is quantified by Snell's Law

An orangutan spear fishing

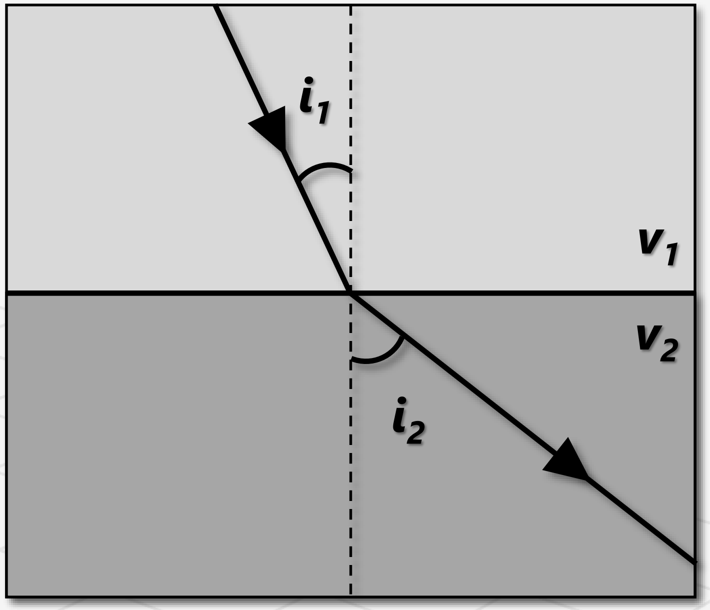
I = angle of incidence

Measured normal to interface

v = seismic velocity of material

At some value of *i*, i_{crit}, the wave is completely reflected

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



$v_1 < v_2$

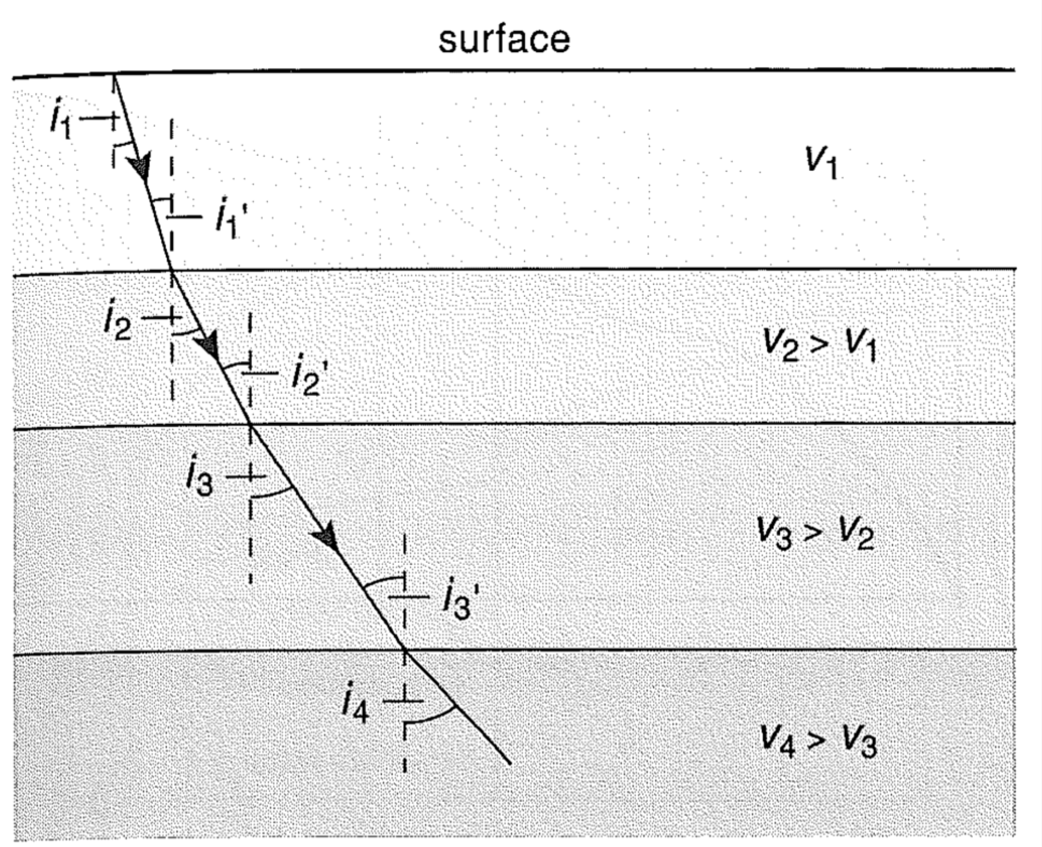
Refraction Through Multiple Layers

Snell's law can be applied to multiple layers

Snell's law also applies to reflected rays

we'll cover this later

$$\frac{\sin i_1}{v_1} = \frac{\sin i_2}{v_2} = \frac{\sin i_3}{v_3}$$



Refraction Through Curved Layers

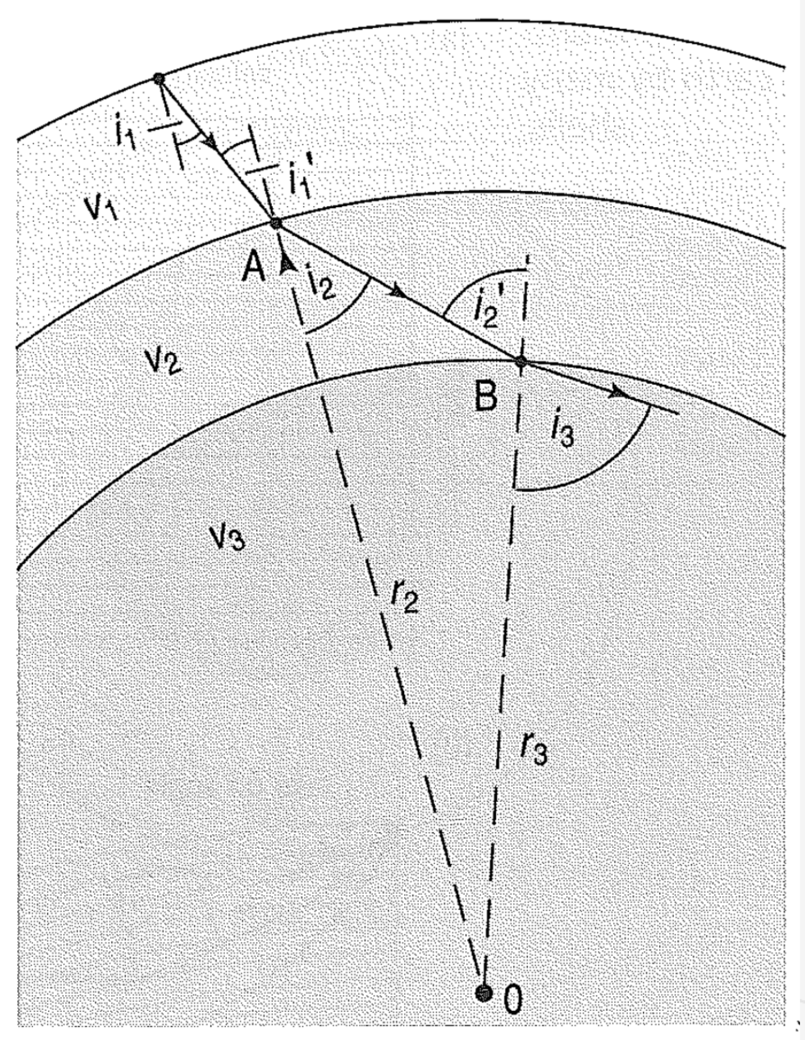
When dealing with large epicentral distances

Must account for the Earth being spherical

Snell's Law can be derived for spherical layers

$$\frac{r_1 \sin i_1}{v_1} = \frac{r_2 \sin i_2}{v_2} = \frac{r_3 \sin i_3}{v_3} = p$$

- P = is the "ray parameter"
  - Has the same value along the entire path of any given ray assuming:
    - v, i, and r are measured at the same place



Reflection, Conversion, and Snell's Law

Snell's Law applies to reflected rays

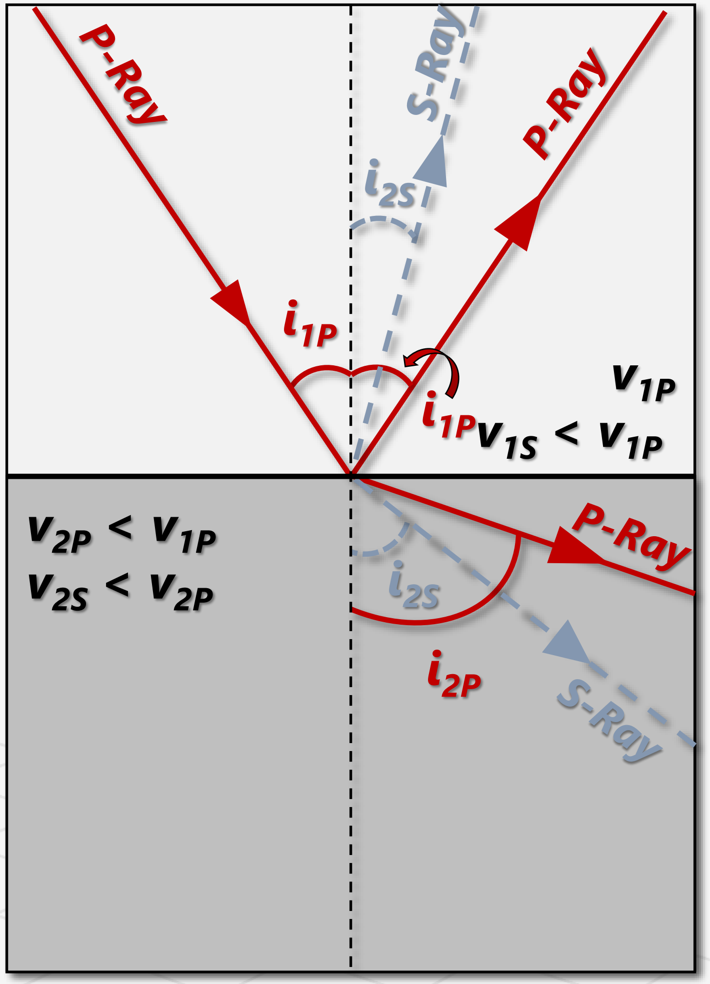
Conversion: when a ray meets an interface, new rays are typically created

- A P-wave can generate reflected P-waves, and S-waves as well as refracted P-waves and S-waves
  - Same is true for S-waves

Snell's law also applies to converted rays that are:

Reflected $$\frac{\sin i_{1P}}{v_{1P}} = \frac{\sin i_{1S}}{v_{1S}}$$

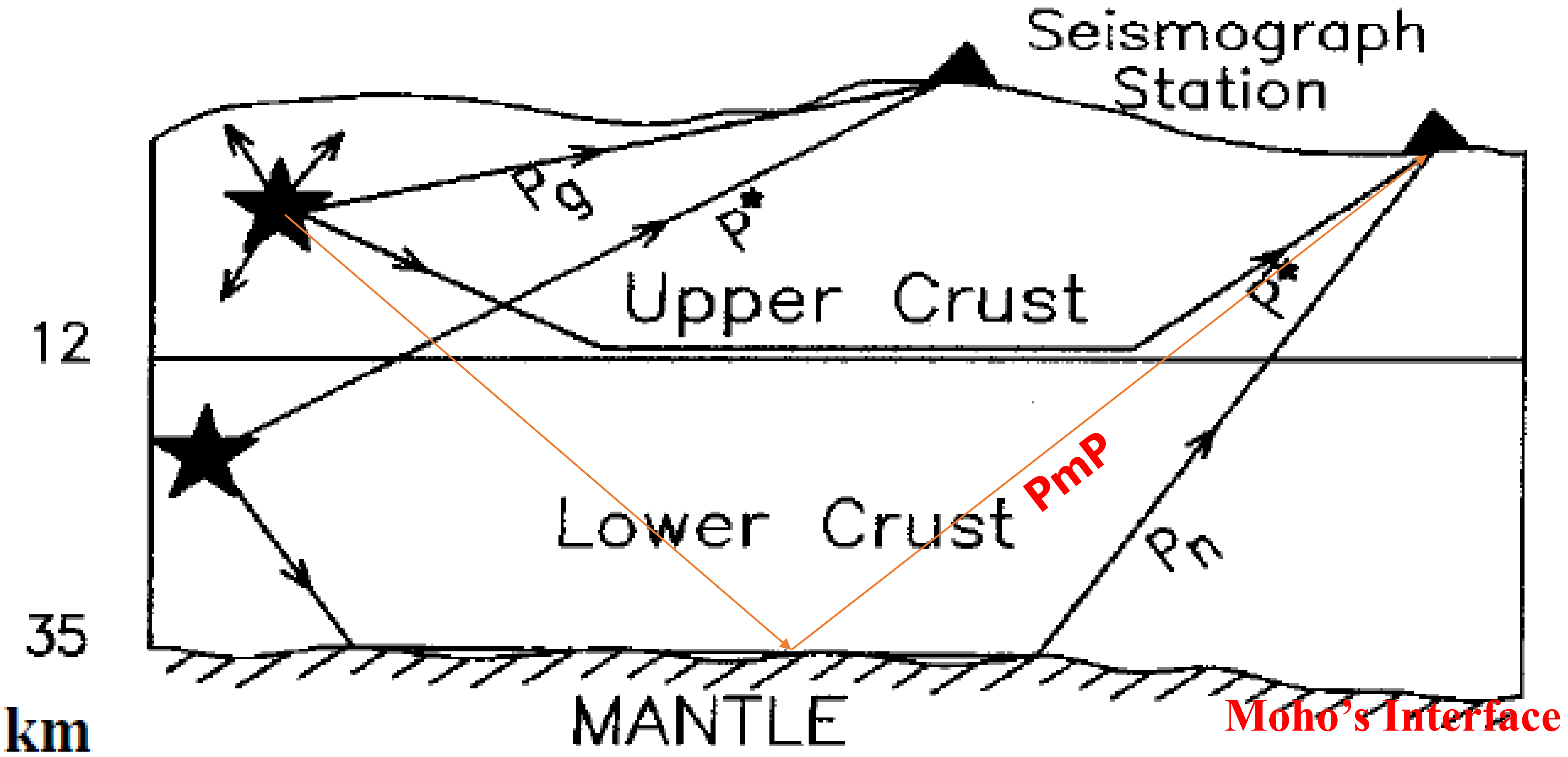
Refracted $$\frac{\sin i_{1P}}{v_{1P}} = \frac{\sin i_{2S}}{v_{2S}}$$



Wave Phases

- Local phases
  - Distance < 1000 km → local earthquakes
- Tele phases
  - Distances > 1000 km → Distant earthquakes

Local phases



Wave Phases

On the global scale, waves converted from reflection or refraction with the major layers of the Earth are called **phases**.

PS

begins as P-wave

reflected off of the surface

converted to S-wave

S_cP

Begins as S-wave

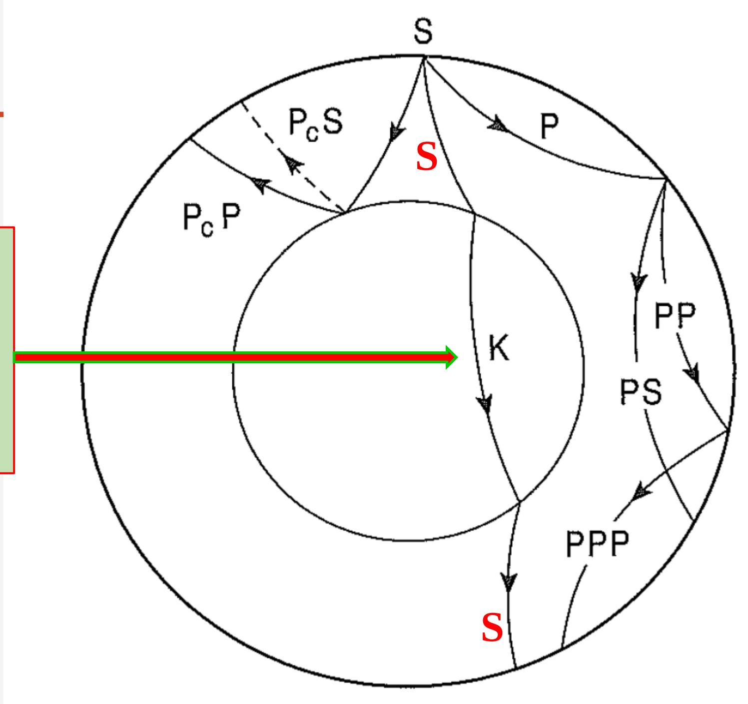
reflected off outer core

Converted to P-wave

i

Reflected off the outer core/inner core boundary

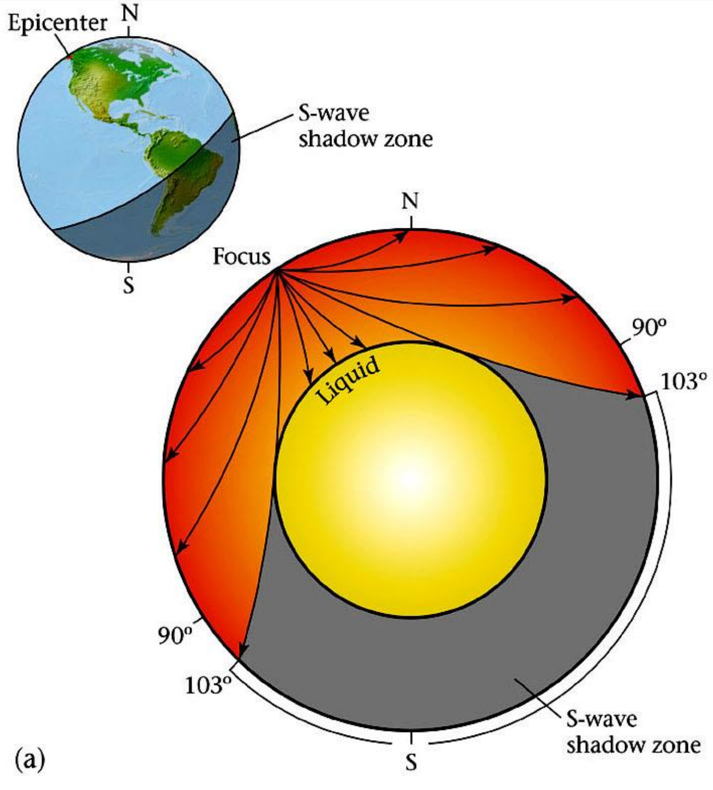
How is the SKS phase traveling through the outer core although the S-waves do not travel through fluids?



- K
  - P-wave in the outer core
- I
  - P-Wave in the inner core
- J
  - S-Wave in the inner core
- Test: What is SKJKP?
- Why is PSKIKP not possible?

The S-wave Shadow Zone

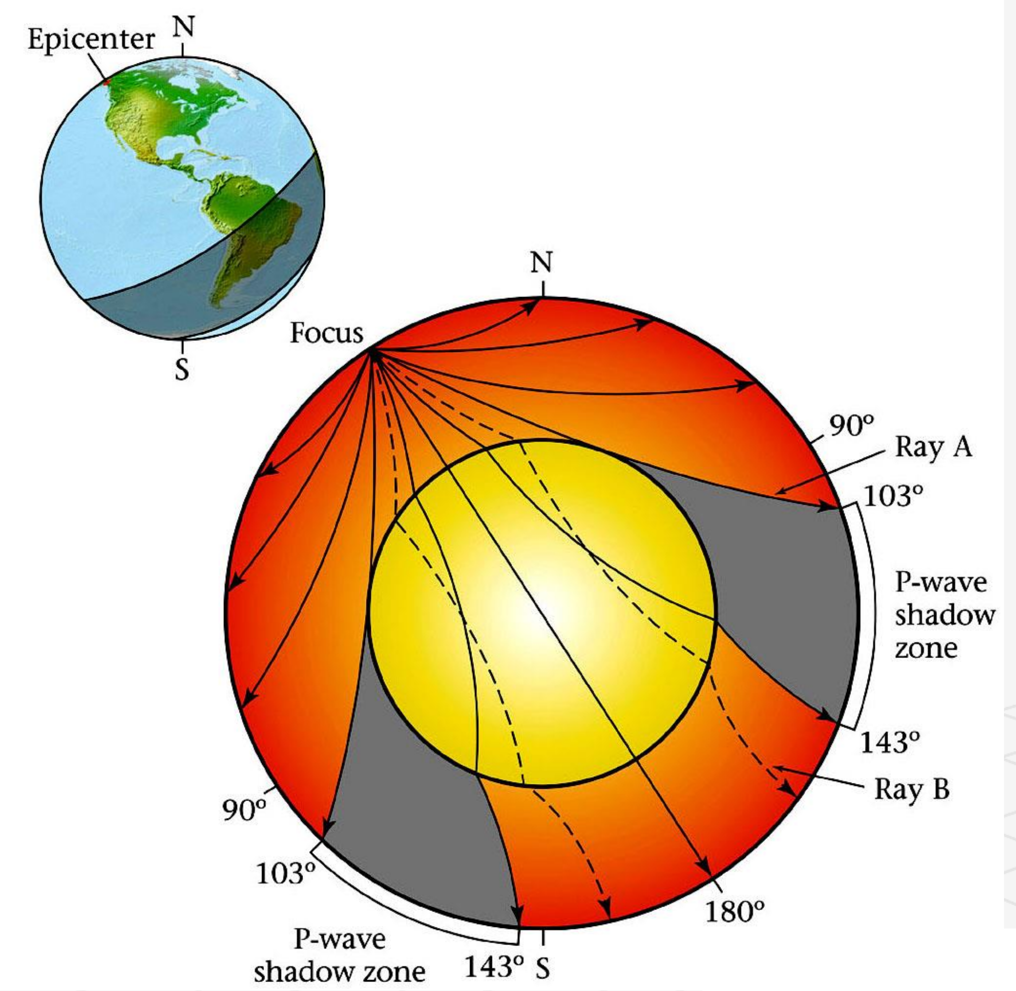
- If velocity gradually increased with depth
  - waves would be recorded at all stations globally
- Seismic waves are not recorded at all seismic stations world wide
  - Called "shadow zones"



- Earth's core is made of an iron alloy based on meteorites
- The S-wave shadow zone is a direct consequence of the liquid outer core.
- Seismologists have since discovered that P-waves reflect off of a discontinuity within the core suggesting a two-layer core.

S-waves only reach stations that are within epicentral angles of < 103° of the epicenter.

The P-Wave Shadow Zone



P-waves also do not reach all stations globally

P-wave shadow zone is between 103° and 143°

- This can be easily explained by a large seismic velocity discontinuity at depth
- The core-mantle boundary

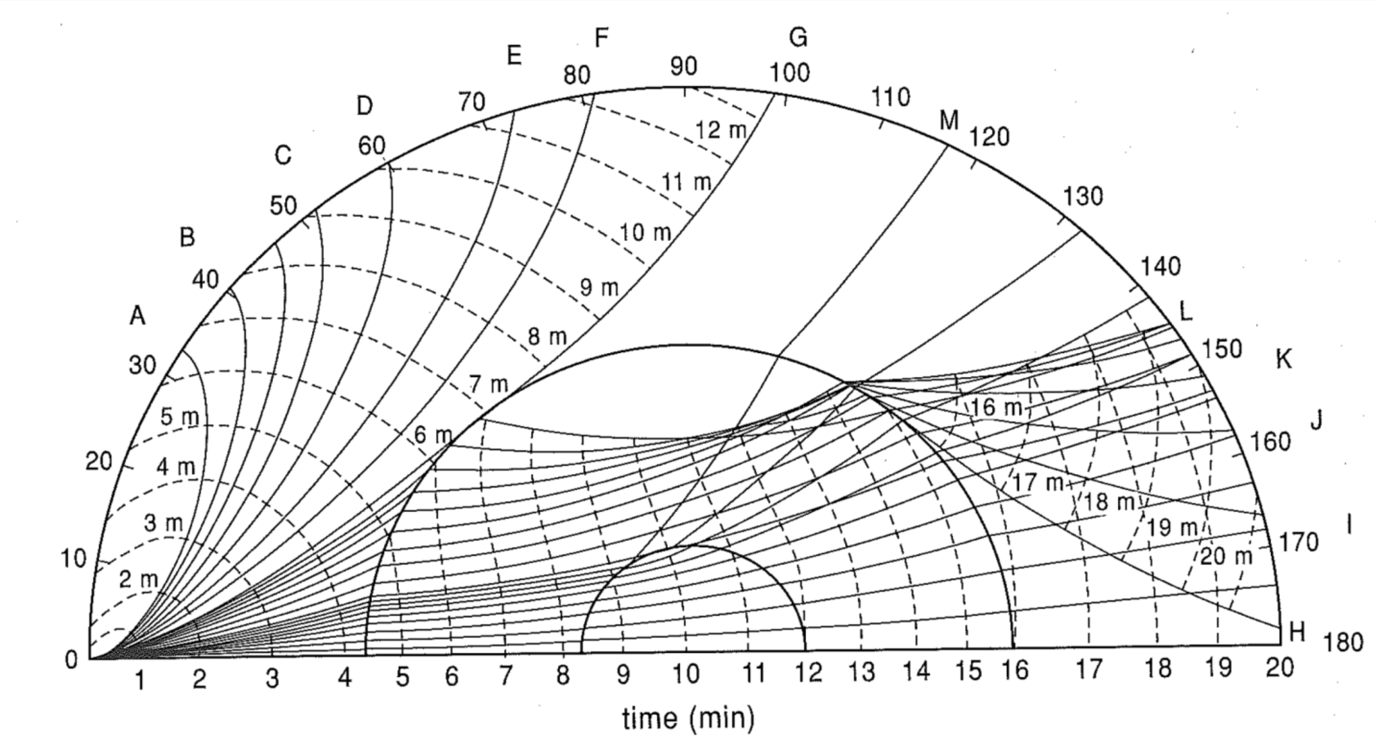
- The shadow zone means that seismic velocities in the core must be slower on average than the lower mantle.

How Long Does it take?

Seismic waves can travel through Earth in a matter of minutes!

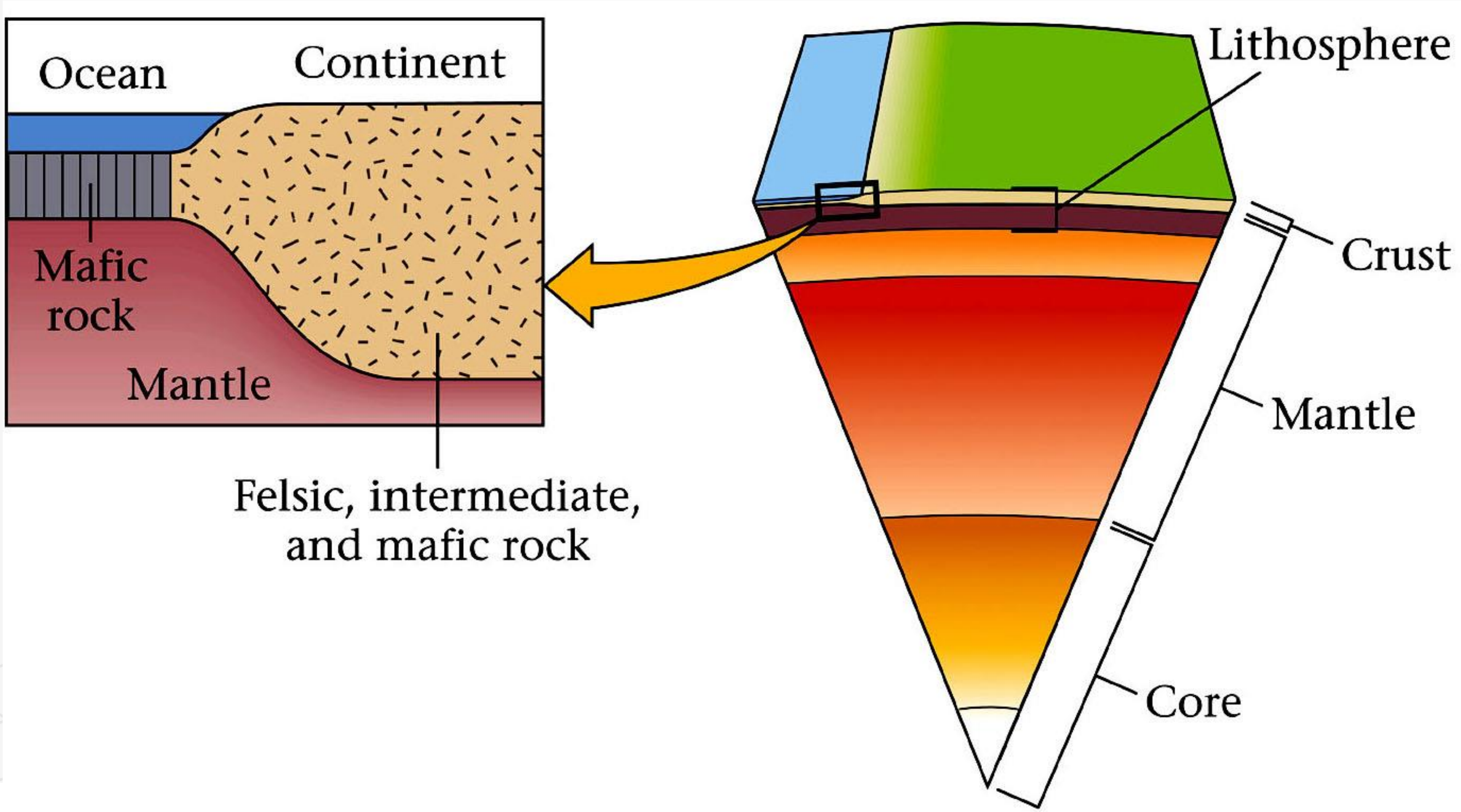
Teleseismic: Rays that arrive at > 18° away from their source.

- Spend little time in the crust
- Useful for investigating the deep Earth



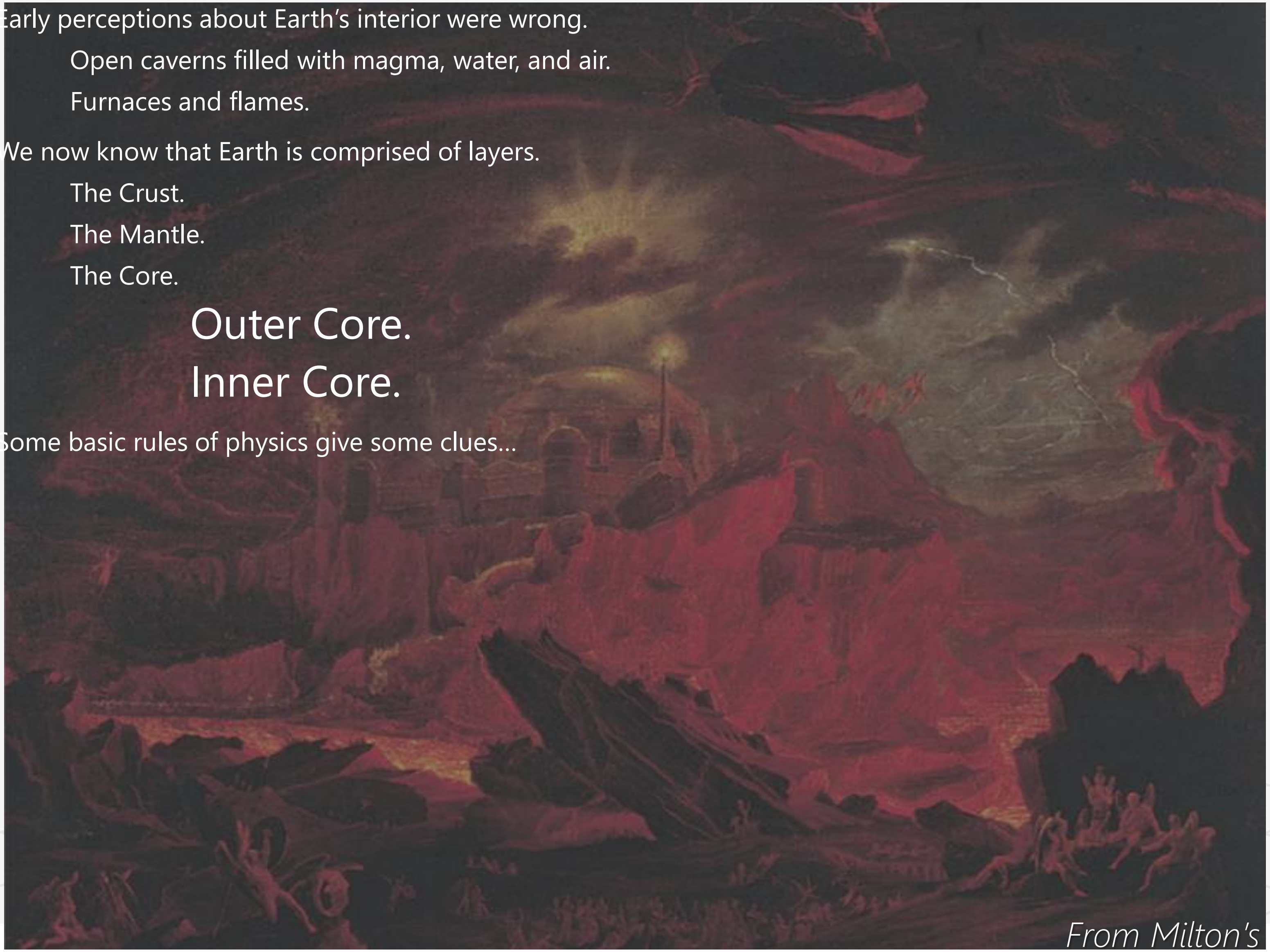
Using Earthquakes to See Inside the Earth

Using what we have learned (and will learn now) about seismic waves, we can now look at how the various layers of the Earth were discovered and some of their properties.



A Layered Earth

- We live on the thin outer skin of Earth.
- Early perceptions about Earth's interior were wrong.
  - Open caverns filled with magma, water, and air.
  - Furnaces and flames.
- We now know that Earth is comprised of layers.
  - The Crust.
  - The Mantle.
  - The Core.
    - Outer Core.
    - Inner Core.
- Some basic rules of physics give some clues…



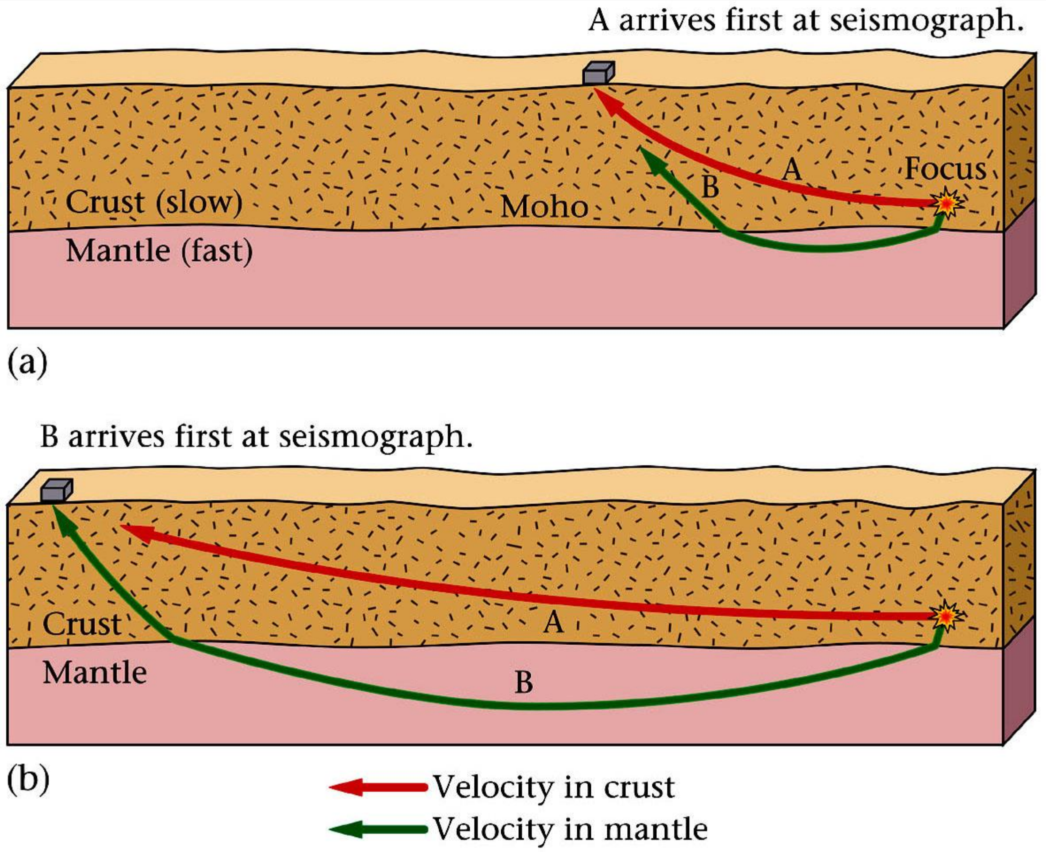
From Milton's "Paradise Lost"

Refraction, Velocity and Arrival Times = Moho Discovery

Seismic waves travel faster through the mantle than the crust

In 1909, Andrija Mohorovičić discovered that waves first arriving at seismic stations within 200 km of an epicenter had an average velocity of 6 km/s

- Stations > 200 km away
  - average wave speed 8 km/s
- To explain this
  - nearby stations received waves that only went through the crust
  - far away stations received waves that travel through the mantle.
- The crust mantle boundary is now called the ***Moho***, in honor of this discovery



Velocity of P-Waves at Depth

Mantle rock = Peridotite

- Ultramafic rock, mostly olivine

In general, seismic velocity increases with depth.

In oceanic crust

- low-velocity zone at ~100-200 km depth.

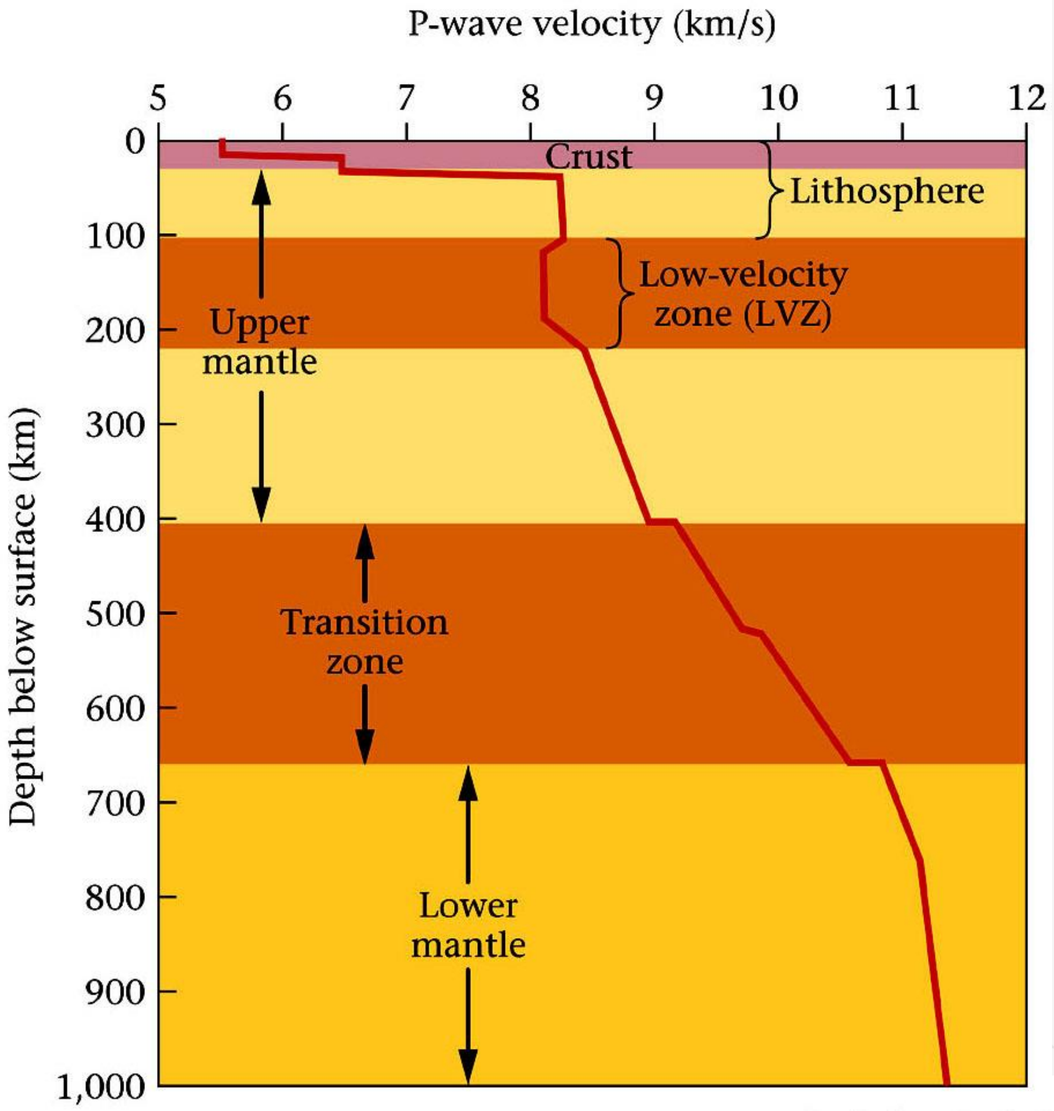
At this depth (pressure) and temperature

- peridotite partially melts < 2%

This zone permits the movement of oceanic plates.

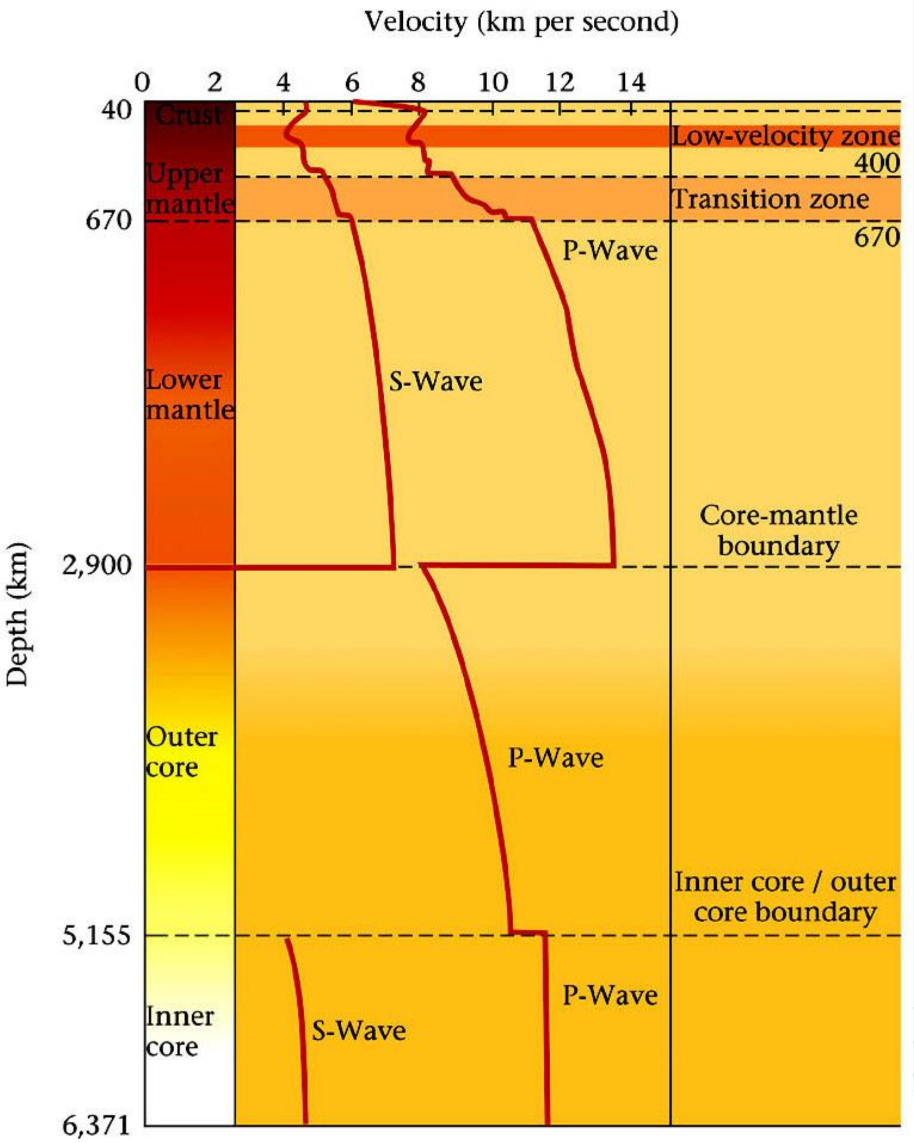
Below the LVZ

- velocities increase until the core mantle boundary

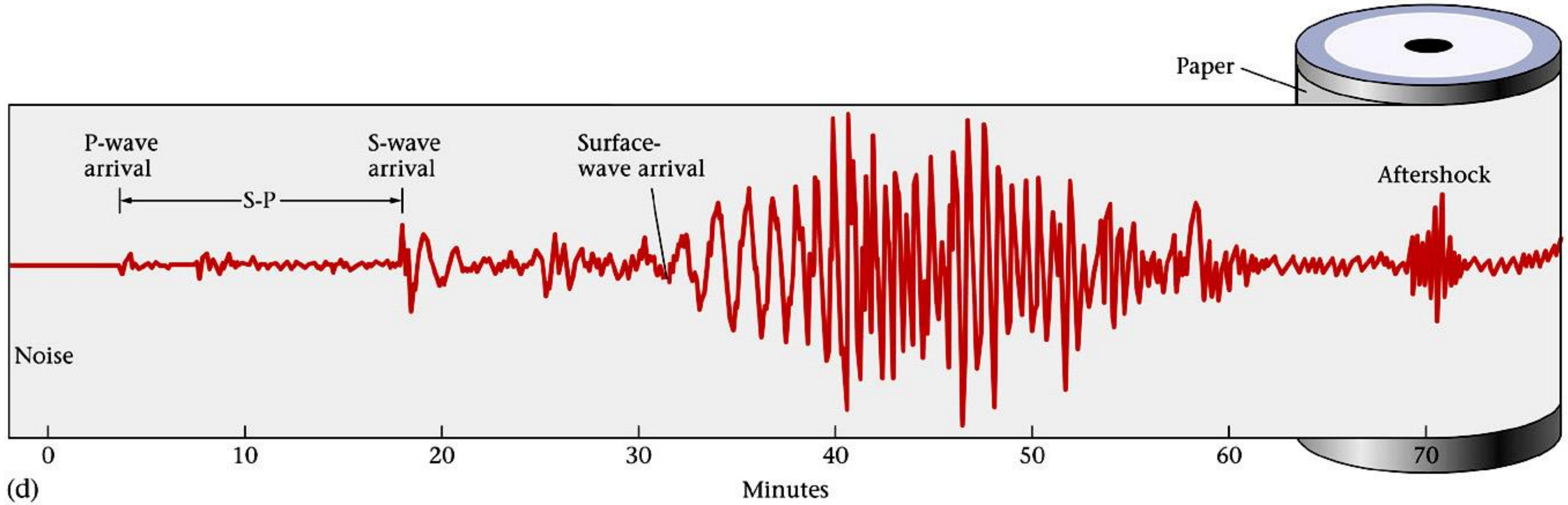


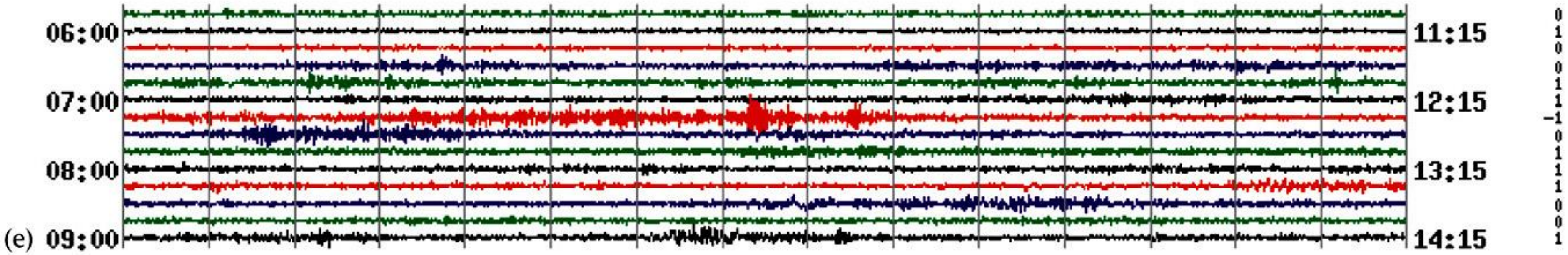
Seismic Velocity vs. Depth for the Whole Earth

Over the past hundred years, seismologists have created a pretty robust picture of seismic velocities at various depths within the Earth.



Attenuation





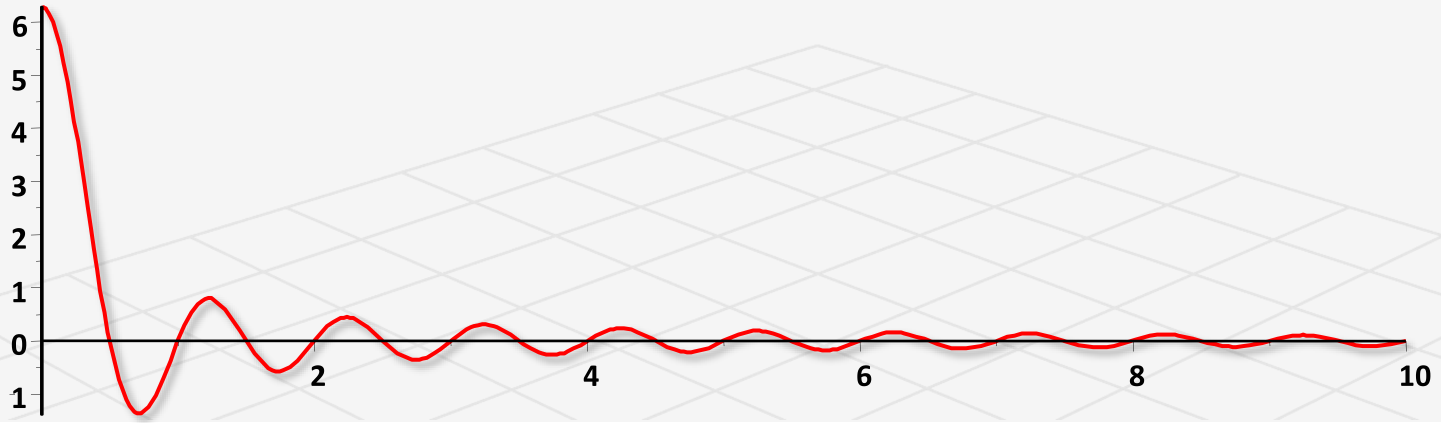
Attenuation

Wave amplitudes generally decrease away from their source

- Energy is spread over larger volume as the wave front expands

Attenuation: the gradual loss of energy (amplitude) of a wave as it travels through a medium

- Causes of attenuation in seismic waves:
  - Encountering liquids or partial melts
    - E.g. the low velocity zone
  - Encountering unconsolidated (or non-elastic) material
    - E.g. Sand



What are the factors described the attenuation mechanism of seismic waves? discuss each factor in details?

Attenuation

Propagating seismic waves loose energy due to

- geometrical spreading

  e.g. the energy of spherical wavefront emanating from a point source is distributed over a spherical surface of ever increasing size

- intrinsic attenuation

  elastic wave propagation consists of a permanent exchange between potential (displacement) and kinetic (velocity) energy. This process is not completely reversible. There is energy loss due to shear heating at grain boundaries, mineral dislocations etc.

- scattering attenuation

  whenever there are material changes the energy of a wavefield is scattered in different phases. Depending on the material properties this will lead to amplitude decay and dispersive effects.

Geometrical Spreading

Decay of wavefront amplitude/energy for spherical waves

- Energy

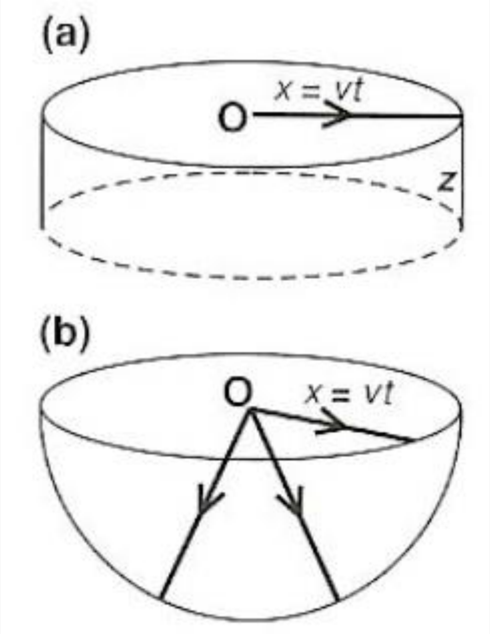
  Decay is proportional to $1/r^2$

- Amplitude

  Decay is proportional to $1/r$

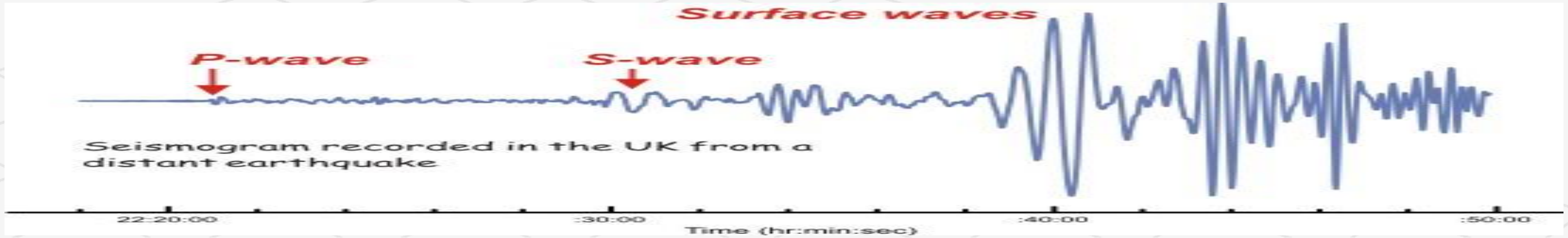
Amplitude of Body waves & Surface waves

Surface waves have much larger amplitudes than Body waves because the energy is spread over a circle ($2\pi x$) rather than a half sphere ($2\pi x^2$)



Why do surface waves show large amplitudes on seismograms rather than the body wave amplitudes?

- Surface wave energy decays with $1/x$, surface wave amplitude decays as $1/\sqrt{x}$
- Body wave energy decays with $1/x^2$, Body wave amplitude decays as $1/x$



Attenuation Q

Attenuation of seismic wave s if usually desribed by the attenuation factor Q. Q is the energy loss per cycle. Q is usually different for P and S waves. Why?

Rock Type	Q_p	Q_S
Shale	30	10
Sandstone	58	31
Granite	250	70-250
Peridotite	650	280
Midmantle	360	200
Lowermantle	1200	520
Outer Core	8000	0

Seismic anisotropy

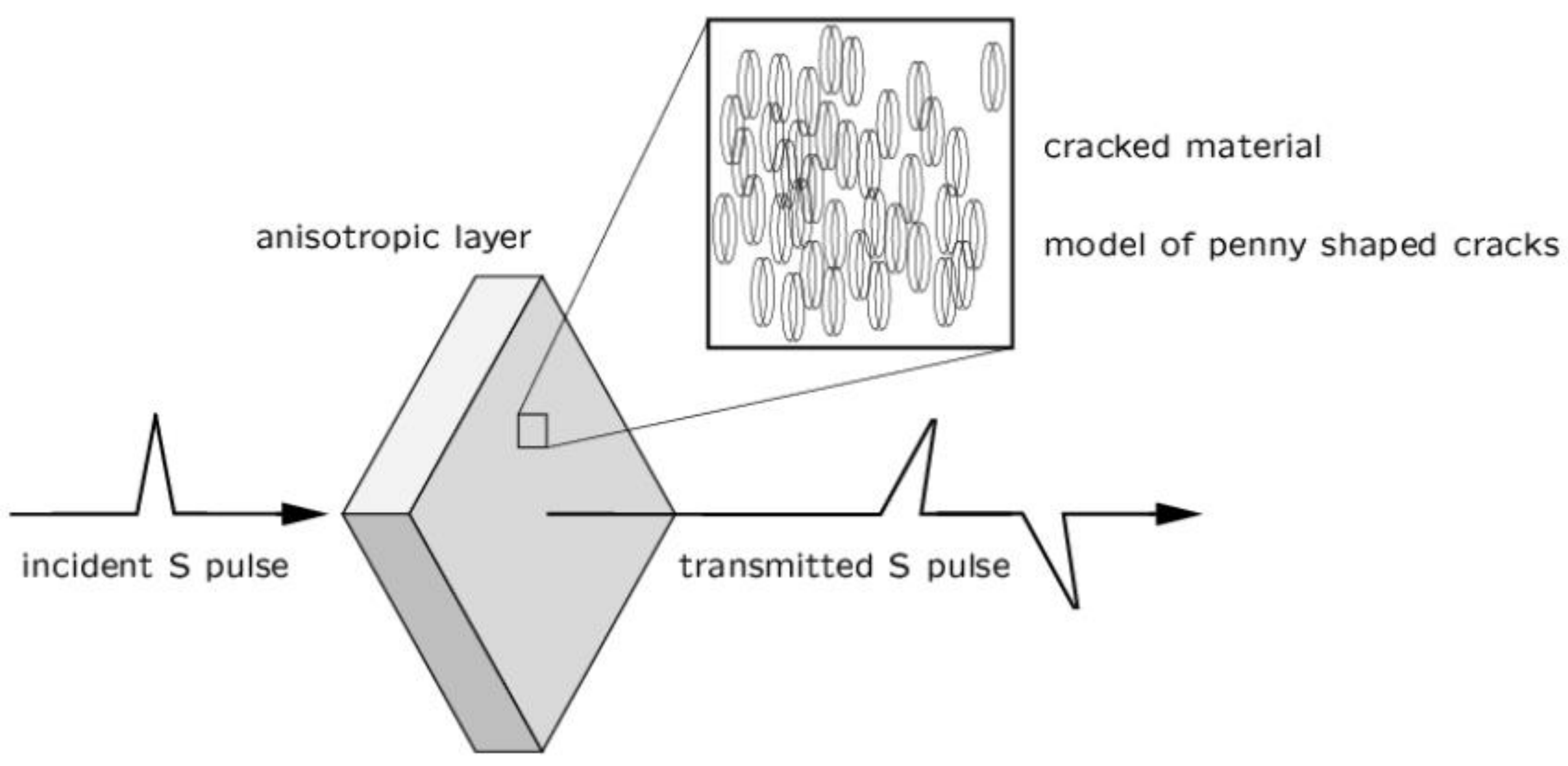
What is seismic anisotropy?

$$\boxed{\sigma_{ij} = c_{ijkl}\varepsilon_{kl}}$$

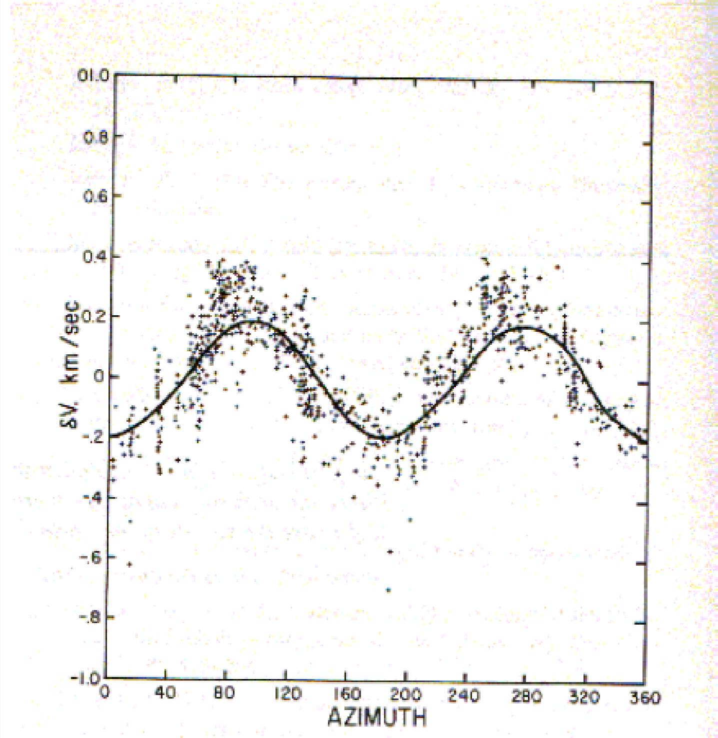
Seismic wave propagation in anisotropic media is quite different from isotropic media:

- There are in general 21 independent elastic constants (instead of 2 in the isotropic case)
- there is shear wave splitting (analogous to optical birefringence)
- waves travel at different speeds depending in the direction of propagation
- The polarization of compressional and shear waves may not be perpendicular or parallel to the wavefront, resp.

Shear wave splitting



Anisotropy



Azimuthal variation of velocities in the upper mantle observed under the pacific ocean.

What are possible causes for this anisotropy?

- Aligned crystals
- Flow processes

Azimuthal anisotropy of Pn waves in the Pacific upper mantle. Deviations are from the mean velocity of 8.159 km/s. Data points from seismic-refraction results of Morris and others (1969). The curve is the velocity measured in the laboratory for samples from the Bay of Islands ophiolite (Christensen and Salisbury, 1979).

Summary

Wave propagation in the Earth due to the elastic properties of its materials

For seismic exploration the most important wave types are P and S waves

Waves are reflected and transmitted by internal interfaces

Seismic wave velocities are important discriminates for rock types and change in lithology

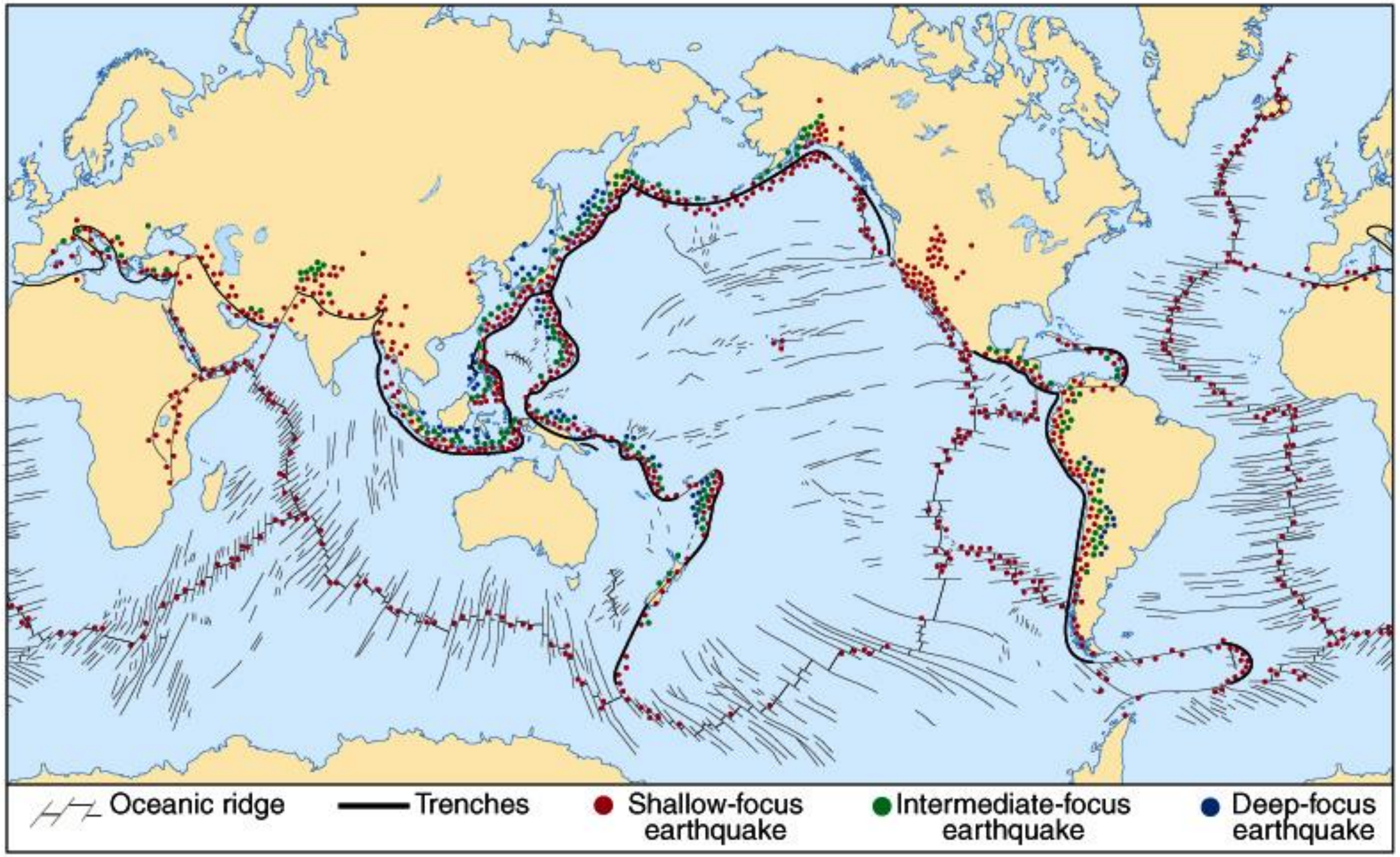
Wave velocities are affected by density, rock type, porosity, pore space content

Seismic waves loose energy through geometrical spreading, intrinsic attenuation and scattering attenuation

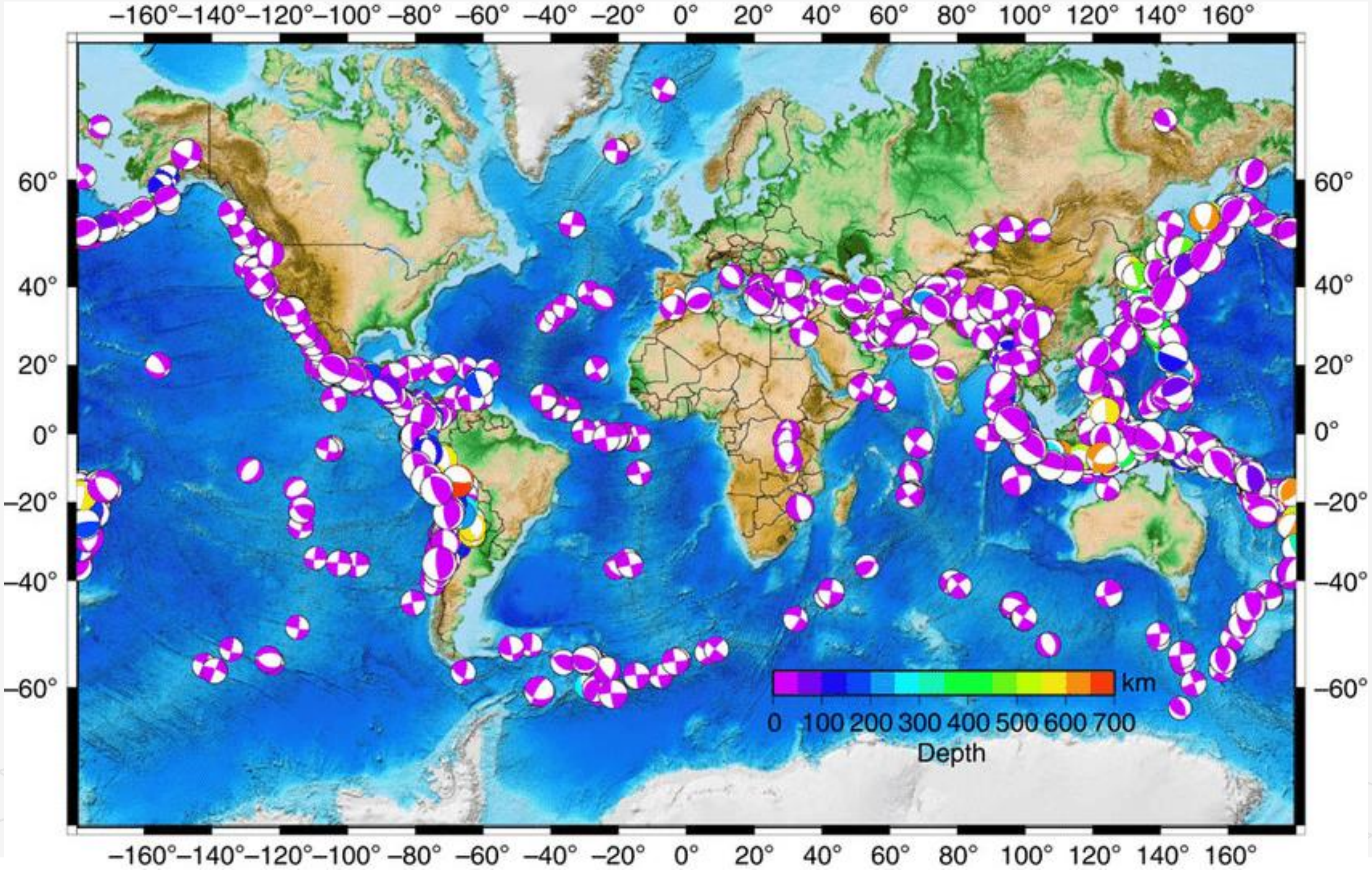
Earthquake and plate tectonics

- Where do earthquakes occur
- Plate boundaries
- Type of faults
- Relation between type of faults and plate boundaries

Earthquake distributions and plate tectonics



Earthquake distributions and plate tectonics

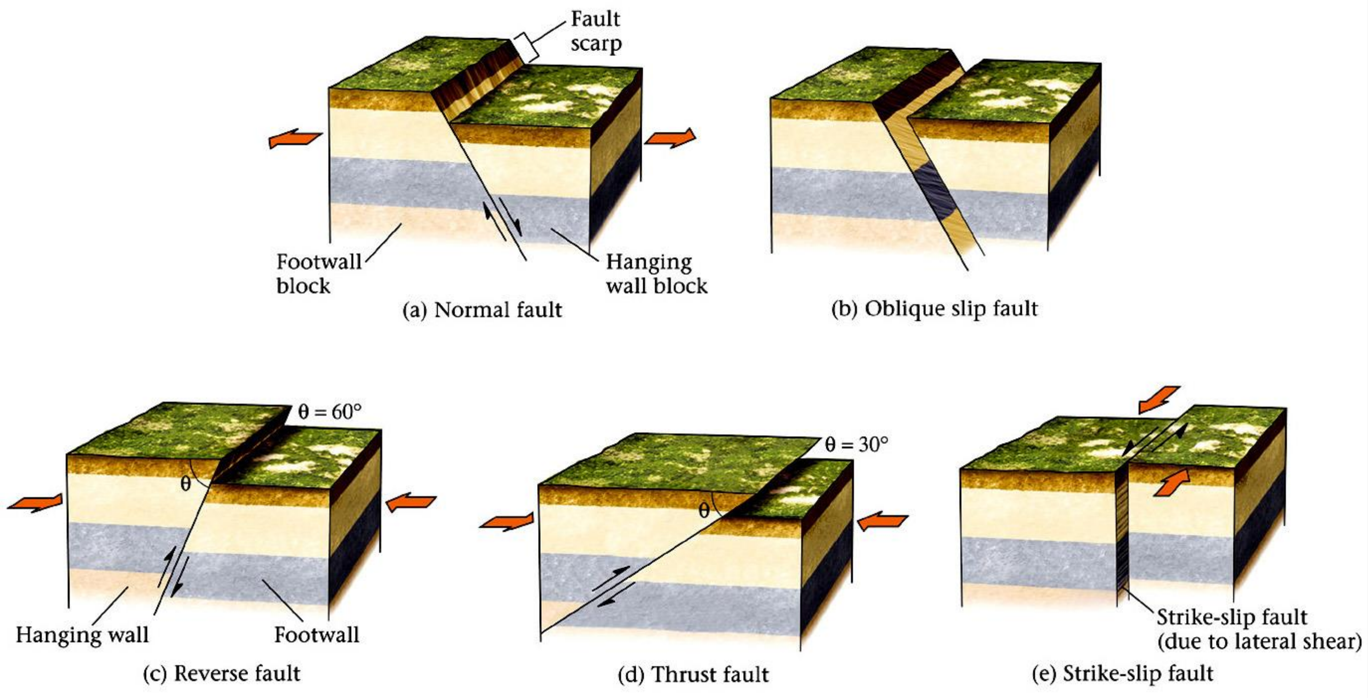


Types of Faults

In general, faults come in three different types: ***Normal, Reverse, and Strike-Slip***

Shallow angle (< 30°) reverse faults are called ***thrust faults***

Faults that have a mix of slip styles are called ***oblique slip faults***



Why are there different types of faults?

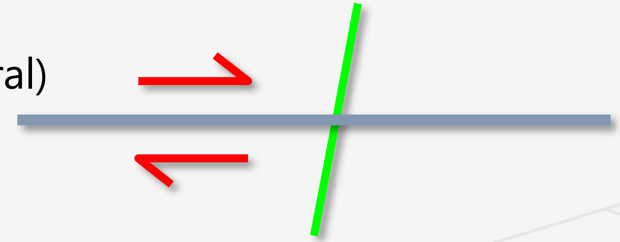
Normal Faults: from stretching of or extending rock; points on opposite sides of a fault are father apart after an earthquake

Reverse Faults: from contracting or squishing rock; points on opposite sides of the fault are closer together after an earthquake

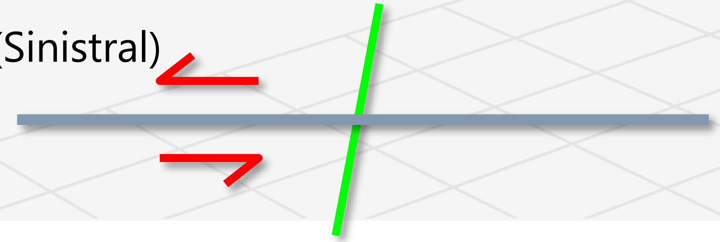
Strike-Slip: can form in either areas of stretching or squishing, material slides laterally past each side of the fault.

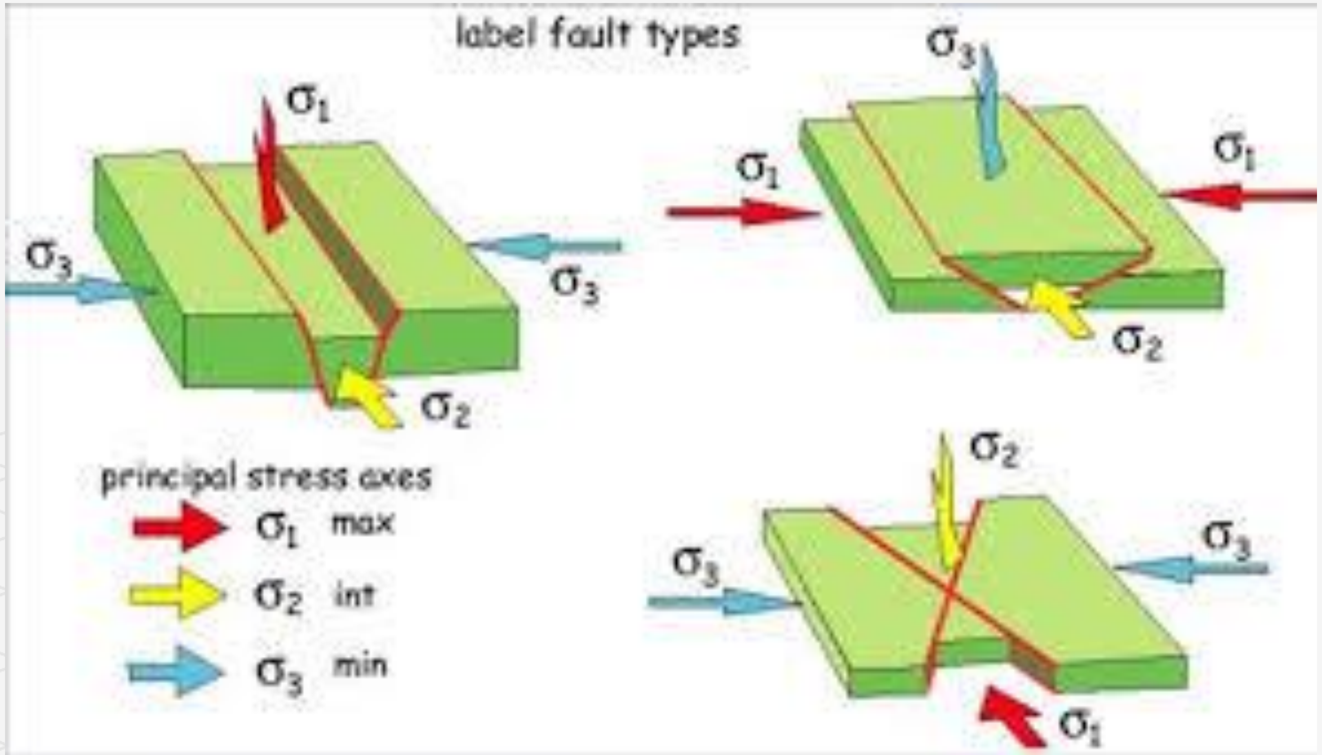
Described by sense of motion:

Right-lateral (Dextral)



Left-lateral (Sinistral)





How are earthquakes recording?

To answer this question, we need to take about the followings:

Earthquake seismic sources

How are earthquakes recorded?

Quantitative methods that can be used to measure an earthquake

Seismic sources

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

Typical sources are:

- Explosive sources (shallow boreholes)
- Vibroseis®
- Airguns (marine surveys)

How are earthquakes recorded?

An important tool to study the Earth's interior is the seismograph.

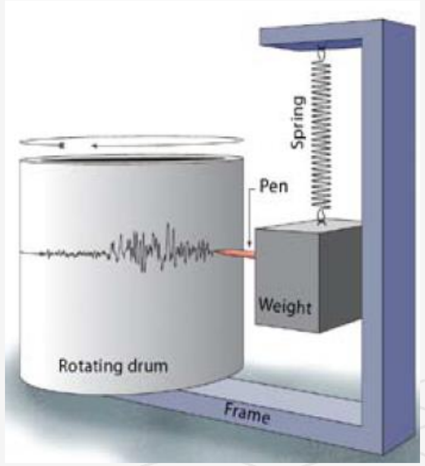
The seismograph is an instrument that records ground motion, or seismic waves, generated by earthquakes.

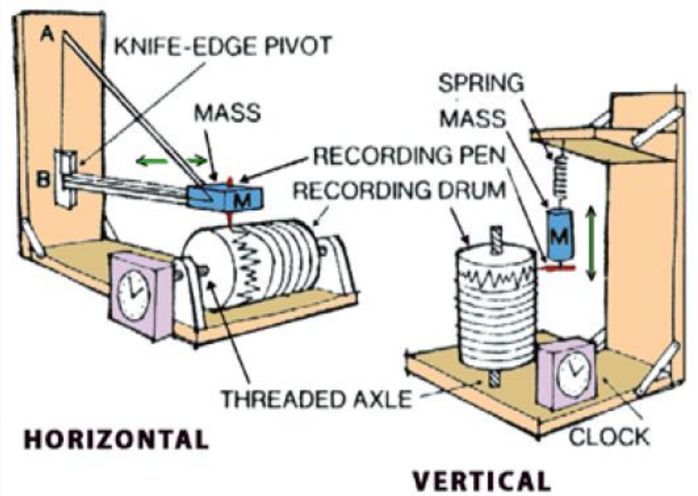
Seismographs can be installed permanently or temporarily.

Seismographs

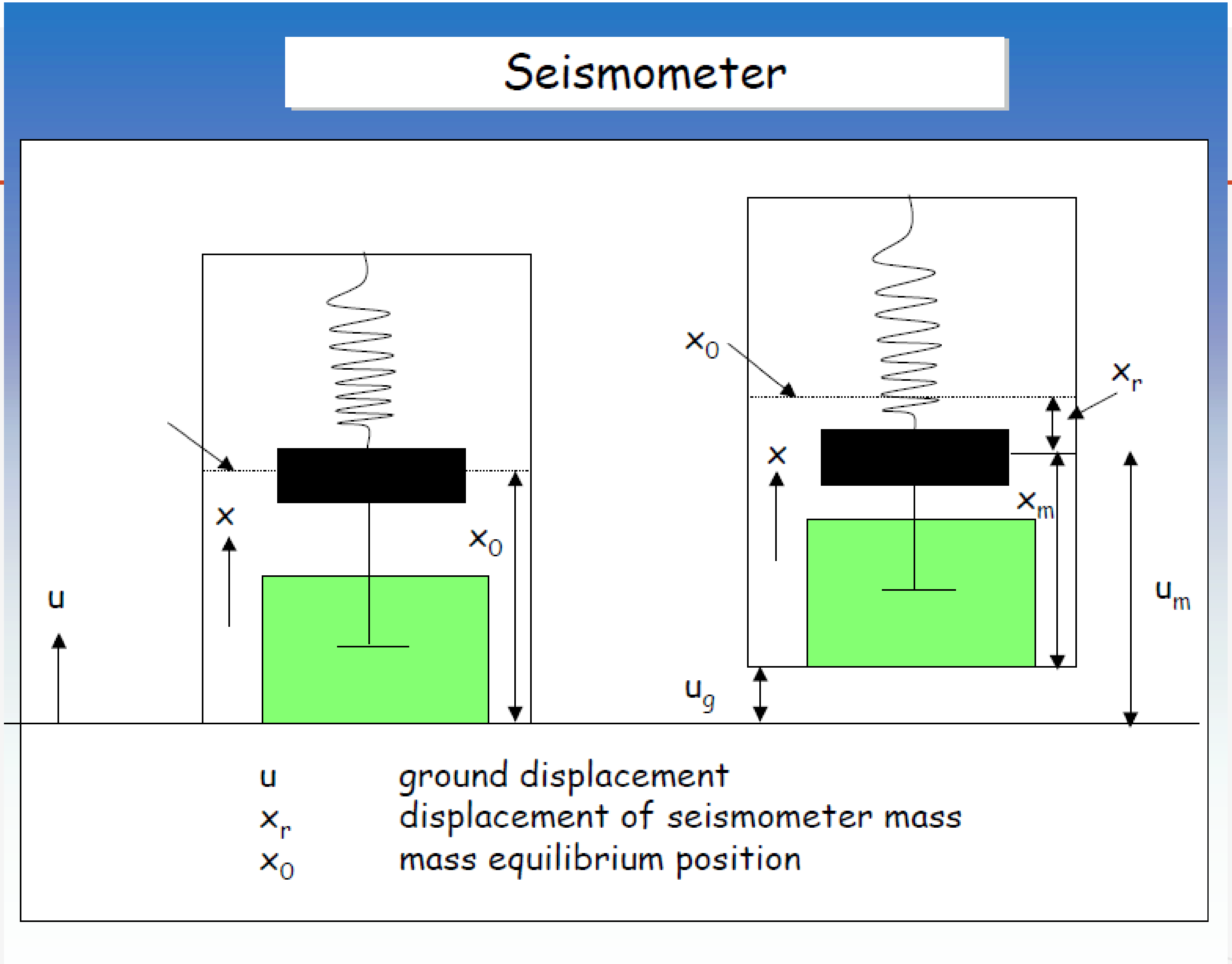
https://www.youtube.com/watch?v=DX5VXGmdnAg

https://www.youtube.com/watch?v=pmf4TXroRJM





Seismograph is an instrument that records vibrations of the Earth, especially earthquakes. Seismograph generally refers to the seismometer plus a recording device as a single unit.



Seismometers measure the components of ground motions. The seismometer is the ground-motion detector part of the seismograph system.

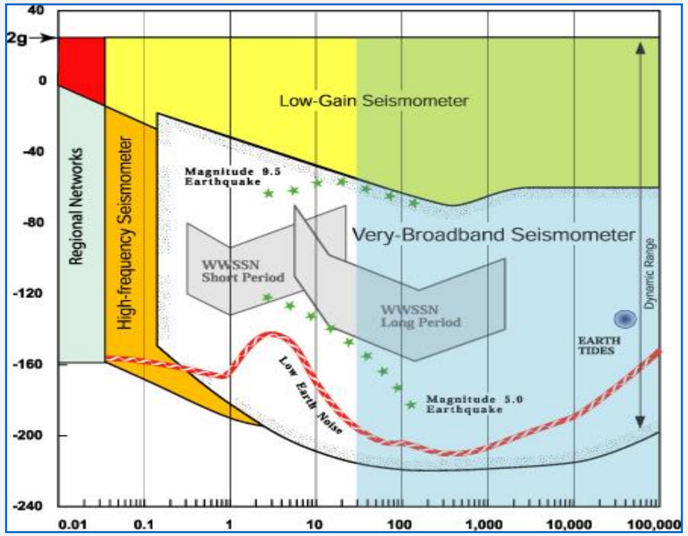
Why more than one kind of seismometer is used

Seismic ground motions generate from different size earthquakes depend on frequency and distance.

At low frequencies ≤ .01 Hz, ground motions range 1 micron to several centimetres for earthquakes of different magnitudes at recorded by distant seismic stations.

For a small micro-earthquake of magnitude 2 recorded at local distances ≤ 30 km, ground motions are as small as 10^{-9} meters.

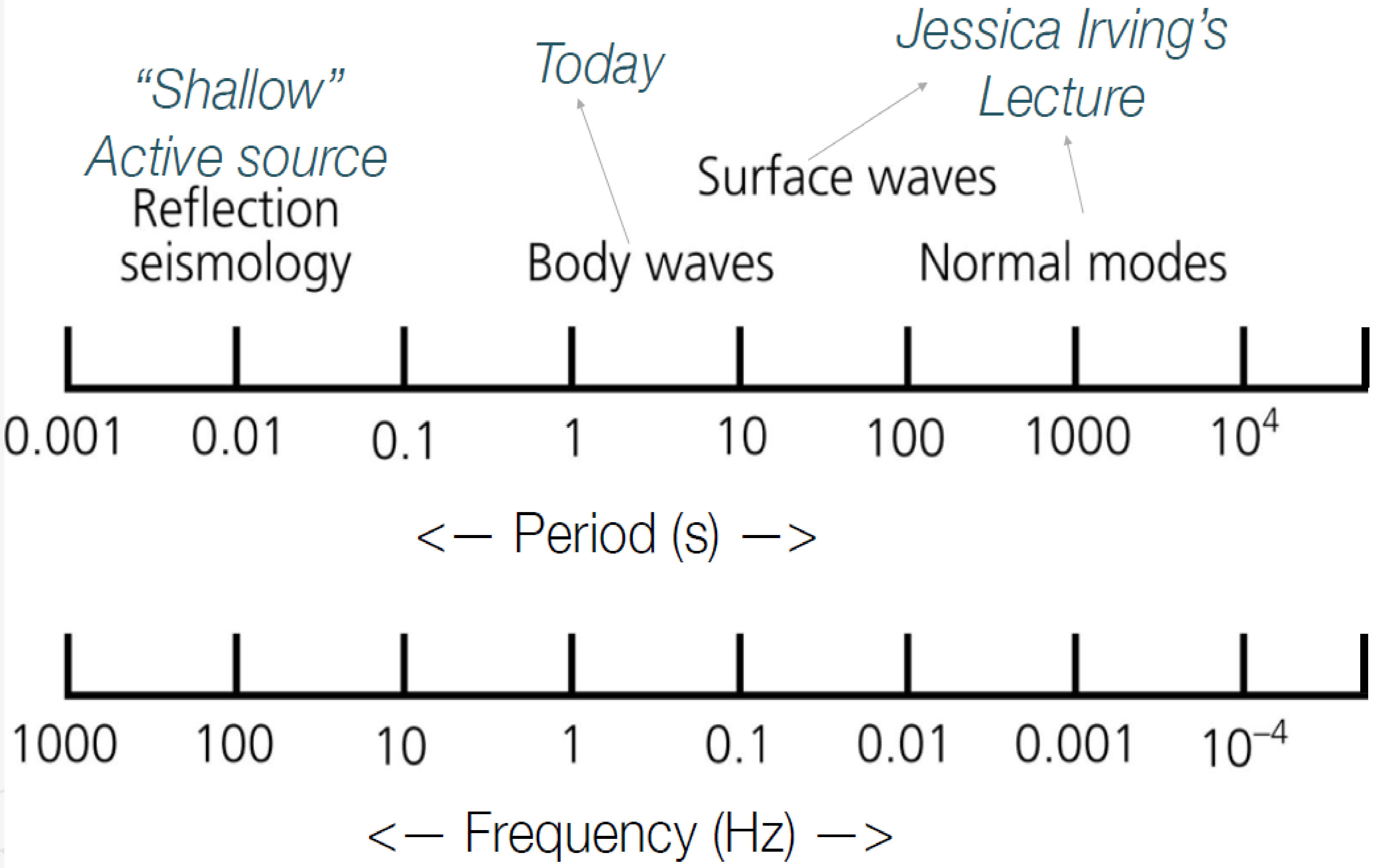
EQUIVALENT EARTH PEAK ACCELERATION (20 LOG M/SEC 2)



PERIOD (SECONDS)

- No single instrument can record all these frequencies and amplitudes
- Thus, many different types of seismometers are needed to study a great range of frequency of the seismic signals

Wave types across frequencies



Types of seismometers

Long-period seismometer

Short-period seismometer

Broadband seismometer

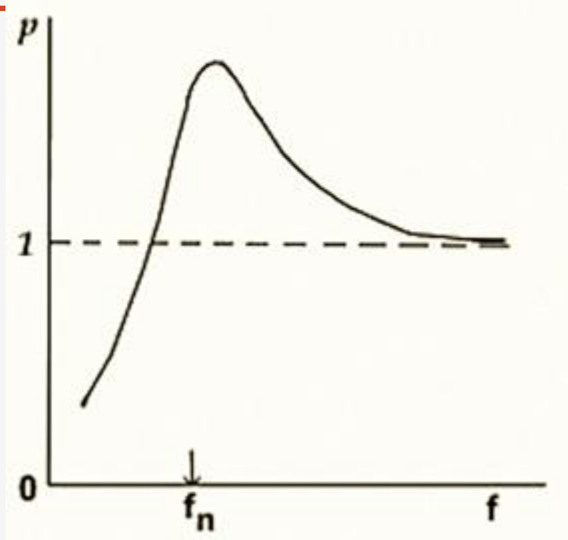
Long-Period seismometer

The natural period of a seismometer is an important factor in determining what is actually records

The Long-period seismometer is an instrument that designed to have long natural period [$\frac{2\pi}{\omega_o}$] and correspondingly low natural frequency ($w_o << w$)

The phase shift between the seismometer and the ground motion is zero

The long-period seismometer is usually designed to record seismic signals with low frequency band rages from 0.01 to 0.1 Hz (10 – 100 seconds)



Amplitude-frequency response curve of a seismometer, *fn is a natural frequency*

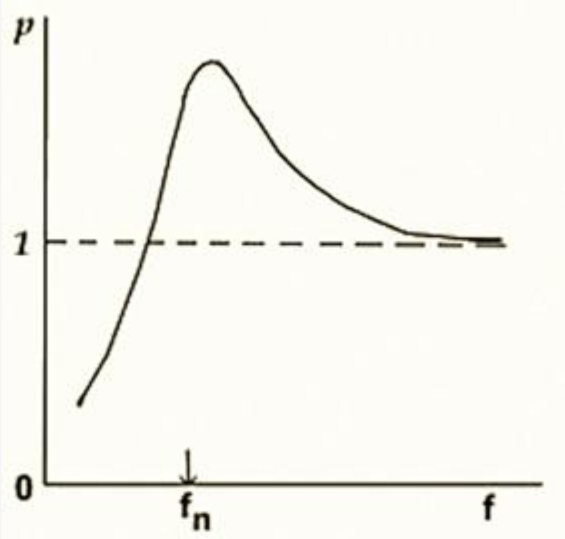
Sate the characteristics of long-period seismometer?

Short-period seismometer

The short-period seismometer is an instrument that designed to have a very short natural period [$\frac{2\pi}{\omega_o}$] and correspondingly high natural frequency ($w_o >> w$)

The phase shift between the seismometer and the ground motion is small

The short-period seismometer is usually designed to record seismic signals with intermediate frequency band rages from **0.01** to 1 Hz (1 – 10 seconds)



Amplitude-frequency response curve of a seismometer, *fn is a natural frequency*. Pendulums have a special property of swing at a characteristic frequency, *fn (or period Tn)*, ***called the natural frequency or natural period***.

Sate the characteristics of short-period seismometer?

Broadband sensors are good for studying:

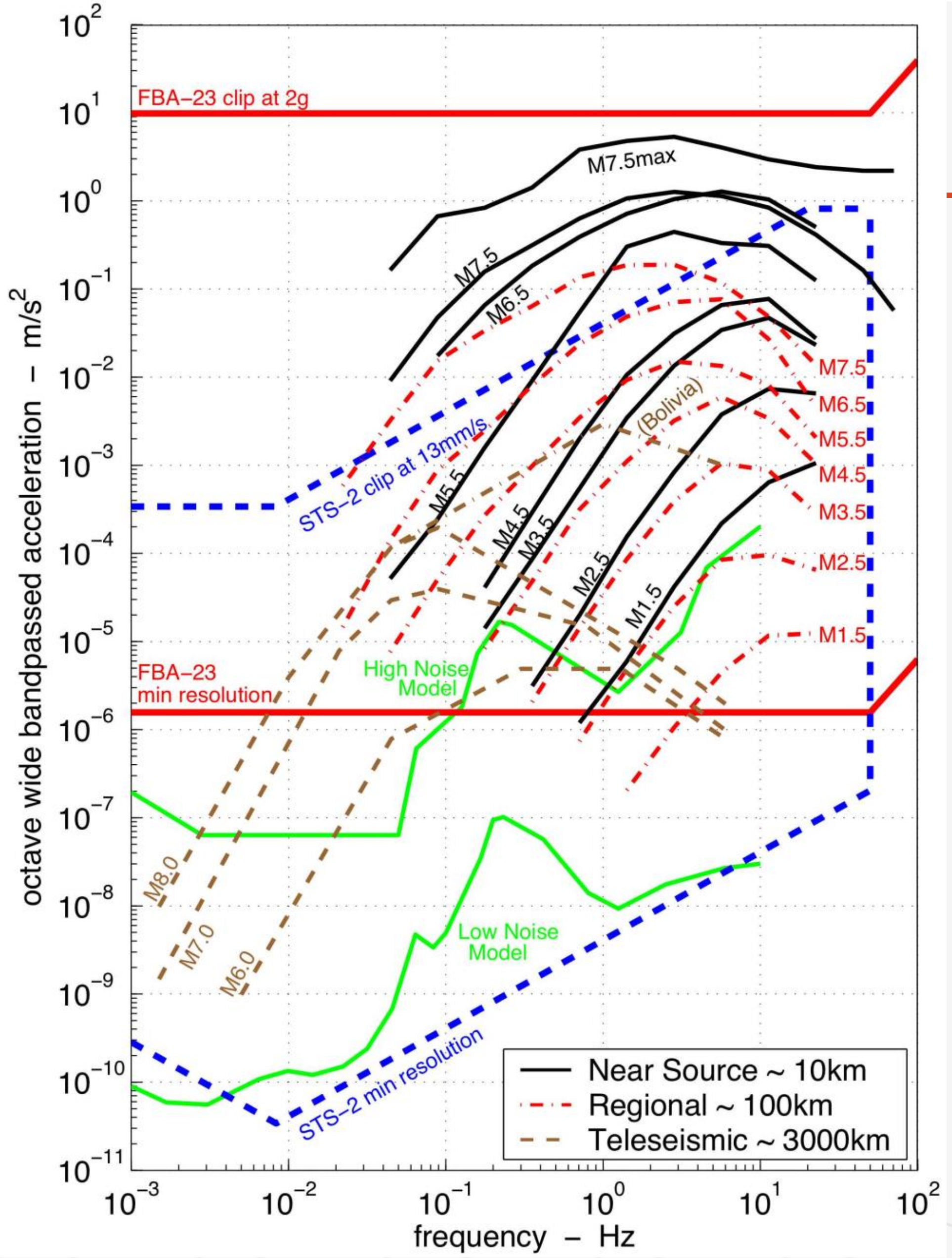
Large – Great events at teleseismic distances

Regional events up to M6.5

Local events up to M5.3

The dynamic range extends from ground noise to large amplitudes (120 dB) and periods range from high-frequency to very long-period signals

(0.001 Hz to more than 50 Hz)



Dynamic range

What is the precision of the sampling of our physical signal in amplitude? It can be measured by estimating the dynamic range.

Dynamic range: the ratio between largest (A_{max}) and weakest (A_{min}) measurable amplitude. The unit is Decibel (dB) and is defined as the ratio of two power values (and power is proportional to amplitude square)

In terms of amplitudes

Dynamic range = $20 \log_{10}(A_{max}/A_{min})$ dB

Example: with 1024 units of amplitude ($A_{min}=1$, $A_{max}=1024$)

$$20 \log_{10}(1024/1) \text{ dB} \approx 60 \text{ dB}$$

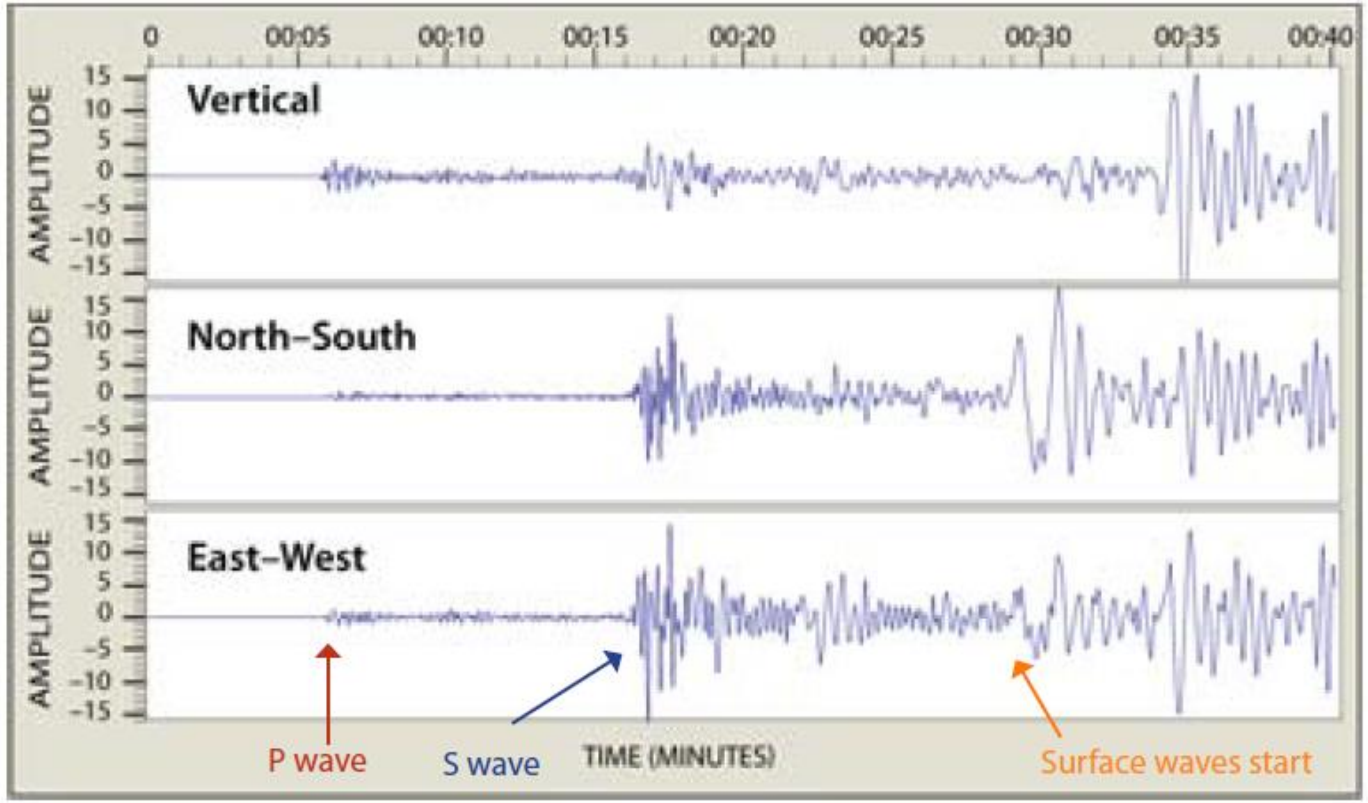
Dynamic range

dynamic range of different digitisers

bits	dynamic range ($2^{\#bits-1}$)	DR_{dB} (#bits-1).6	Orders of magnitude
8	256/2	42	2
12	4,096/2	66	3
16	65,536/2	90	4.5
20	1,048,576/2	114	6
24	16,777,216/2	138	7

Seismograms

Seismogram is the real-time record of earthquake ground motion recorded by a seismograph. Seismograms are the records (paper copy or computer image) used to calculate the location and magnitude of an earthquake.



https://www.iris.edu/hq/inclass/search#type=1

Seismograms

The response of a seismometer is commonly proportionally to the ground velocity because its depend on the relative velocity of the magnet and coil.

However, the ground motion can be recorded as displacement or velocity or acceleration. This depends on the equation of motion.

The seismometer characterizes by **very rapid earth motion** (oscillations), the seismometer is designed to **measure displacement**.

For very **slow earth motion**, the seismometer is designed to **measure acceleration**.

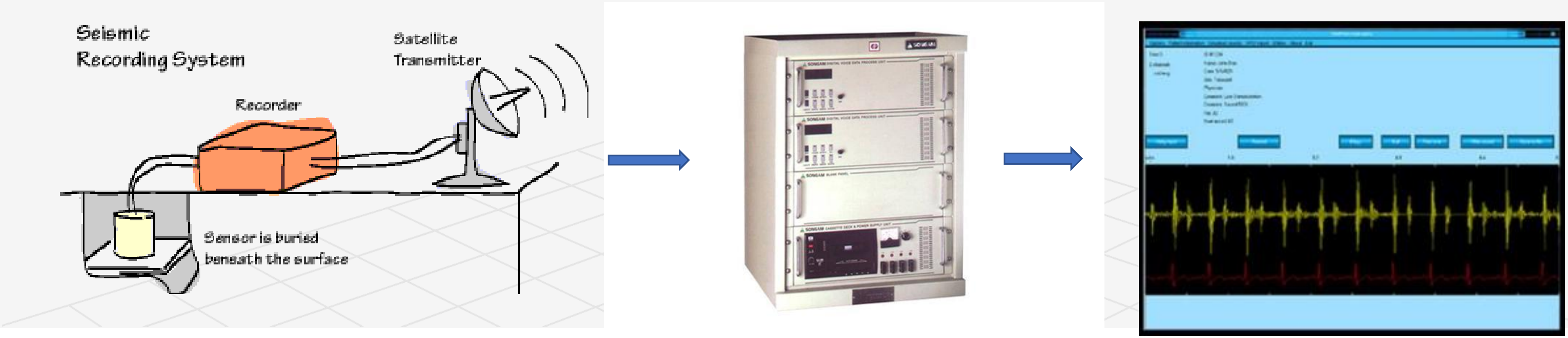
Magnification

The oscillations of earth motion due to an earthquake is very small and should be magnified by seismometer, digitizer, and recording system.

The equation uses to magnify the amplitude of ground motions is defined as

Magnified amplitude of the ground motion equals true ground motion multiplied by magnification factor

Analog and digital recording systems



Earthquake size

Earthquake Seismology

- How can we determine the origin time of an earthquake?

- How can we determine the epicentre and hypocentre of earthquakes?

- How can we measure the size of an earthquake (Richter scale)?

- What describes the damage of an earthquake (seismic intensity, Mercalli scale)?

Earthquake size

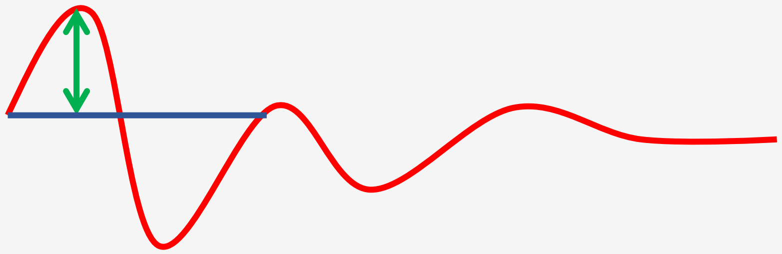
- **Earthquake “size” described by two measurements:**
  - **Intensity:**
    - **The intensity is defined as a measure of the degree of earthquake shaking at a given location based on the amount of damage.**
  - **Magnitude:**
    - **It is defined as estimates the amount of energy released at the source of the earthquake.**

Earthquake magnitudes

Magnitude scales

Due to different types of seismometers and different types of seismic waves can be used to estimate the magnitude of an earthquake, there are many magnitude scales based on the type of instruments and the type of seismic waves used in calculations.

Magnitude scales thus have the general form:



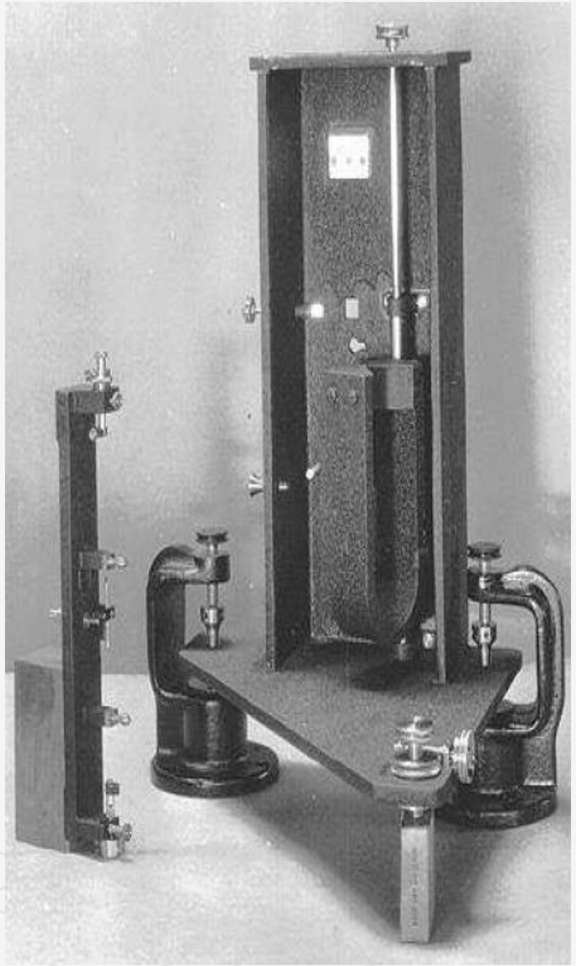
$$M = log\left[\frac{A}{T}\right] + F(h, \Delta) + C$$

where A is the amplitude of the signal, T is its dominant period, F is a correction for the variation of amplitude with the earthquake's depth h and Δangular distance from the seismometer, and C is a regional scaling factor.

Since a logarithmic scale is used to calculate earthquake magnitude, magnitudes can be negative for very small displacements.

The first magnitude scale “Richter magnitude scale”

- The picture on the left shows a short period Wood-Anderson torsion seismometer.
- The instrument as a whole is sensitive to horizontal motions.
- The magnitude scale devised by Richter is now referred to as the local magnitude M_L.



Richter magnitude scale

- The Richter magnitude scale is a logarithmic scale with commonly reported magnitudes varying from -0.1 to 9.
- *Each unit increase in* magnitude on the Richter scale corresponds to an increase in seismic wave amplitude of *10 times*.
- Example: Magnitude 5 is 10 times greater in Richter magnitude than Magnitude 4 and 100 times greater than Magnitude 3. The difference in wave amplitude between two earthquakes of different sizes can be measured by $10^{\Delta m}$, where Δm is the difference between two magnitudes.
- *Each unit increase in* magnitude on the Richter scale corresponds to an increase in seismic wave energy of approximately 30 *times*
- Example" Magnitude 7 releases approximately 30 times the strain energy as Magnitude 6 and 900 times the energy of Magnitude 5. The difference in strain energy can be measured as $32^{\Delta m}$, where Δm is the difference between two magnitudes.
- The Richter magnitude scale is determined by the amplitudes of S- waves at a distance of 100 km from the center of the earthquake (hypocenter).

Magnitude versus ground motion and energy

The relationship between earthquake size and the ground motion and energy release. For example, an earthquake of magnitude 7.0 produces 10 times more ground motion and releases 32 times more energy than an earthquake of magnitude 6.0.

Magnitude differences	Seismic wave amplitudes	Seismic wave Energy
1.0	10.0	32
0.5	3.2	5.5
0.3	2.0	3
0.1	1.3	1.4

Richter magnitude scale

Due to heterogeneity of the geological setting from place to other, each region requires its own calibration curve, this is because the attenuation of seismic waves with distance can be very different for different geological provinces.

The magnitude of the largest arrival (often the S-wave) is measured and corrected for the distance between source and receiver, given by the P- and S-wave differential arrival times. The scale for Southern California is defined by:

$$ML = logA + 2.76log\Delta - 2.48$$

Although the Wood-Anderson seismograph is now rarely used for recording the seismic waves, local magnitudes are sometimes still reported due to many buildings have resonant frequencies near 1 Hz, which is close to that of a Wood-Anderson seismograph. Therefore, ML is often a good indicator of the potential for structural damage.

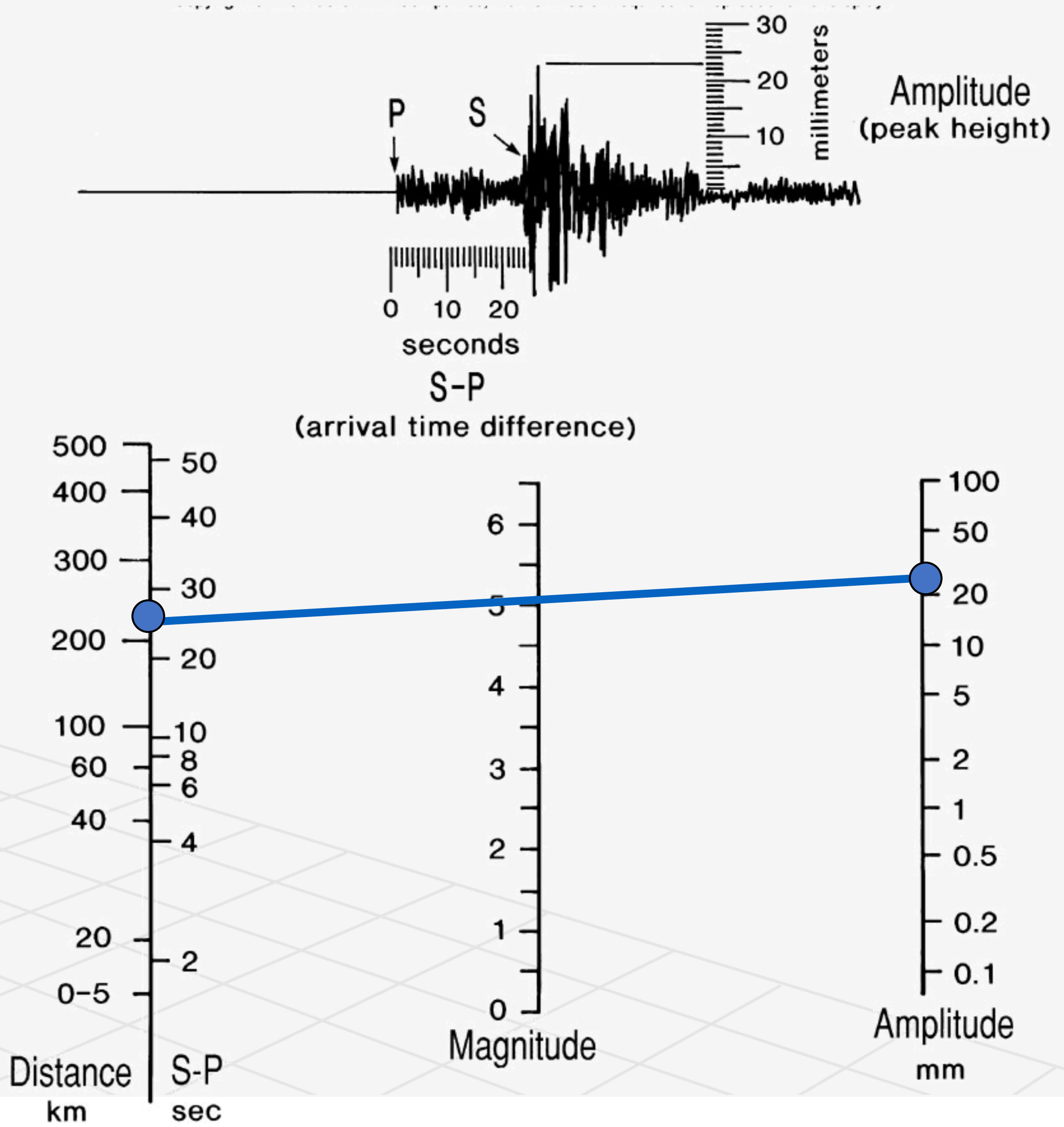
How big was it? The Richter Scale

What is the Richter magnitude of this EQ?

S — P = 26 sec

Amplitude = 23mm

Magnitude = 5



Other magnitude scales

A number of different global and local magnitude scales have been produced since the original formulation of Richter. Some of these scales will be discussed.

Body wave magnitude scale (m_b):

This scale is measured from the early portion of the body waves, usually that associated with the P-wave, and is defined as:

$$m_b = \log\left[\frac{A}{T}\right] + Q(h, \Delta)$$

In this case, A is the ground motion amplitude in microns after the effects of the seismometer are removed, T is the wave period in seconds, and Q is an empirical term that is a function of angular distance and focal depth. Q can be derived as a global average or for a specific region.

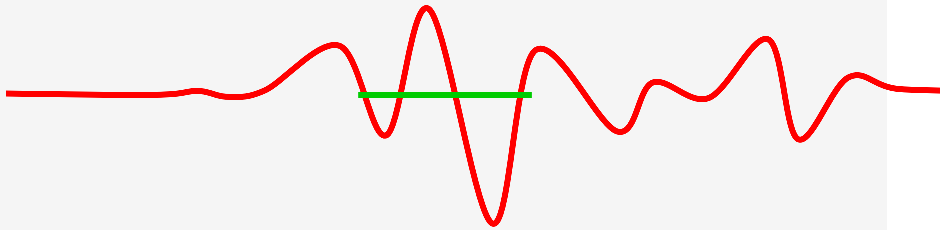
Surface wave Magnitude

- The Richter local magnitude does not distinguish between different types of waves. Other magnitude scales that base the magnitude on the amplitude of a particular wave have been developed.

- At large epicentral distances, body waves have usually been attenuated and scattered sufficiently that the resulting motion is dominated by surface waves.

- The surface wave magnitude is a worldwide magnitude scale based on the amplitude of Rayleigh waves with a period of about 20 sec. The surface wave magnitude is obtained from

$$M_s = \log A + 1.66 \log D + 2.0$$



- where A is the maximum ground displacement in micrometers and D is the epicentral distance of the seismometer measured in degrees (360^o) corresponding to the circumference of the earth).

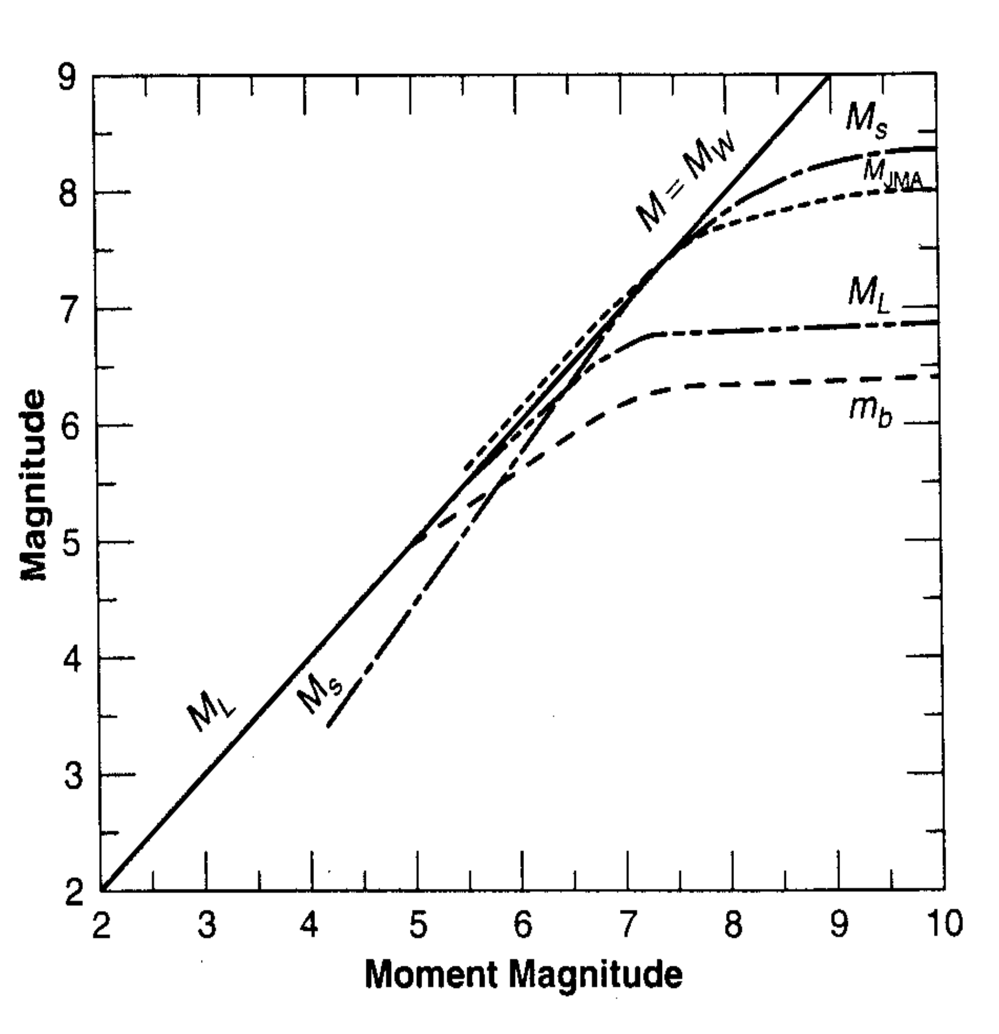
Surface wave Magnitude

- Note that the surface wave magnitude is based on the maximum ground displacement amplitude at priod within a range of 18 to 22 seconds (rather than the maximum trace amplitude of a particular seismograph); therefore, it can be determined from any type of seismograph.

- The surface wave magnitude is most commonly used to describe the size of shallow focal depth (less than about 70 km)), and distant (farther than about 1000 km) for moderate to large size earthquakes.

Saturation of magnitude scales

- The Richter, body wave, and surface wave magnitude scales suffered from what we call saturation.
- Saturation of the instrumental scales is indicated by their flattening at larger magnitudes
- Moment magnitude and energy magnitude scales overcome this problem.



Moment magnitude

- Moment magnitude is based on the total elastic energy released by the fault rupture and is related to the seismic moment Mo defined by

$$M_o = \mu DA$$

where μ = Modulus of rigidity of the rock (dyne/cm2) A = Area of rupture surface of the fault (cm2) D = Average fault displacement (cm) Moment Magnitude is Defined by Hank and Kanamori (1979) as

$$M_w = \frac{2}{3} log M_o - 10.7$$

- Different magnitude scales also yield different values, and body and surface wave magnitudes do not correctly reflect the size of large earthquakes.

Energy Magnitude Me

- Energy magnitude is related to the total energy released as seismic waves.
- This is important for seismic hazard from ground shaking and landsliding.
- Done by integrating radiated energy flux in velocity-squared seismograms over the duration of the rupture.

How to distinguish destructive earthquakes using Me scale

- The Me scale can be used to distinguish between two earthquakes which one has released larger energy than the other

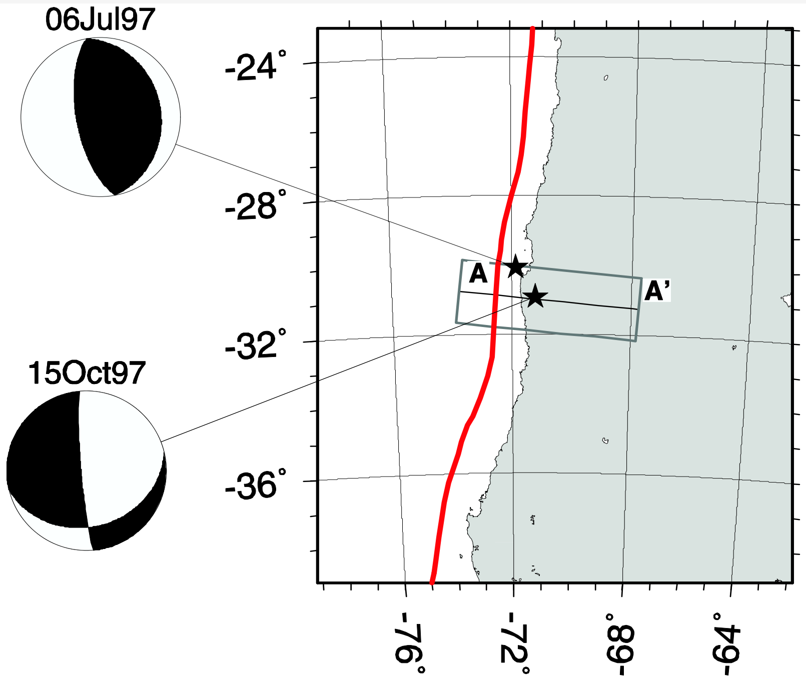
- Example: two earthquakes in Chile had the same M_w but different M_e

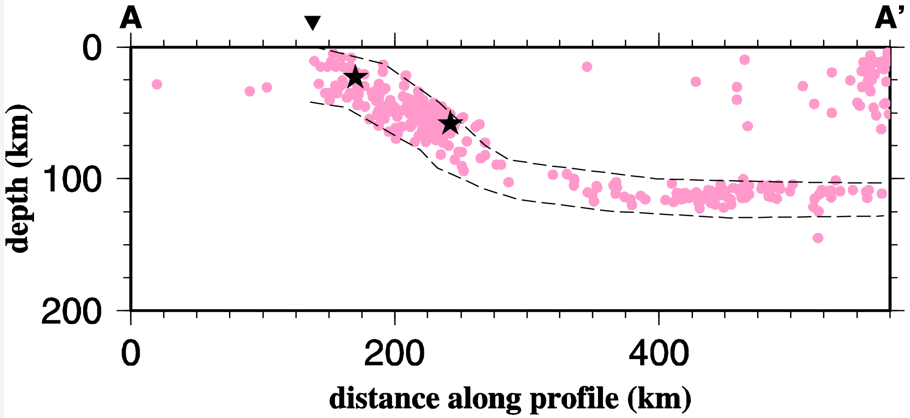
  - **Earthquake 1**: 6 July 1997, *Me* 6.1, *Mw* 6.9

    No fatalities, no houses destroyed.

  - **Earthquake 2**: 15 October 1997, *Me* 7.6 *Mw* 7.1

    7 people killed, more than 300 people injured. 5,000 houses destroyed. Landslides and rockslides in the epicentral region.





Magnitudes and Energy of Earthquakes

Annual Numbers of EQs

Frequency of Occurrence of Earthquakes

Descriptor	Magnitude	Average Annually
Great	8 and higher	1 [1]
Major	7 - 7.9	17 [2]
Strong	6 - 6.9	134 [2]
Moderate	5 - 5.9	1319 [2]
Light	4 - 4.9	13,000 (estimated)
Minor	3 - 3.9	130,000 (estimated)
Very Minor	2 - 2.9	1,300,000 (estimated)

[1] Based on observations since 1900.
[2] Based on observations since 1990.

Magnitude vs. Ground Motion and Energy

Magnitude Change	Ground Motion Change (Displacement)	Energy Change
1.0	10.0 times	about 32 times
0.5	3.2 times	about 5.5 times
0.3	2.0 times	about 3 times
0.1	1.3 times	about 1.4 times

This table shows that a magnitude 7.2 earthquake produces 10 times more ground motion than a magnitude 6.2 earthquake, but it releases about 32 times more energy. The energy release best indicates the destructive power of an earthquake.

MOST of the energy is released by around 20 magnitude 7 and larger earthquakes every year.

Summary

As magnitude increases by 1.0, ground motion changes by a factor of 10. The change in energy with magnitude is even more dramatic. A change of magnitude by 1.0 corresponds to a change in energy released by a factor of 32!
Notice the dramatic change in number of earthquakes of different sizes. Small earthquakes are MUCH more frequent than large earthquakes. However, because the energy changes by a factor of 32 with an increase of 1.0 in magnitude, large earthquakes account for most of the energy released in earthquakes. There are about 20 earthquakes of magnitude 7.0 or greater each year and these release 80% of all seismic energy.

Earthquake Energy—To rephrase, for each unit of magnitude the amplitude of the waves increases by a factor of 10, but the duration also increases, so the energy released increases by a factor of 32!

RESOURCE: See file “How Often do Earthquakes Occur?” on the IRIS web site (URL below). If you visit this web site, you should also examine the other informative “one-pagers” that have been developed by the IRIS Education and Outreach staff.

Intensity

Modified Mercalli Intensity Scale (MMI)

- Do not confuse MMI with magnitude
- An intensity scale is the intensity of the ground shaken intensity as determined by human feeling and by the effects ground motion on structures and living things
- MMI is graded based on intensity:
  - Ranges from I to XII (from imperceptible to catastrophic destruction earthquakes)
  - The scale that measures the effect of an earthquake on human aspects
  - Establish based on visible damage and human feelings

Intensity Scale (MMI)

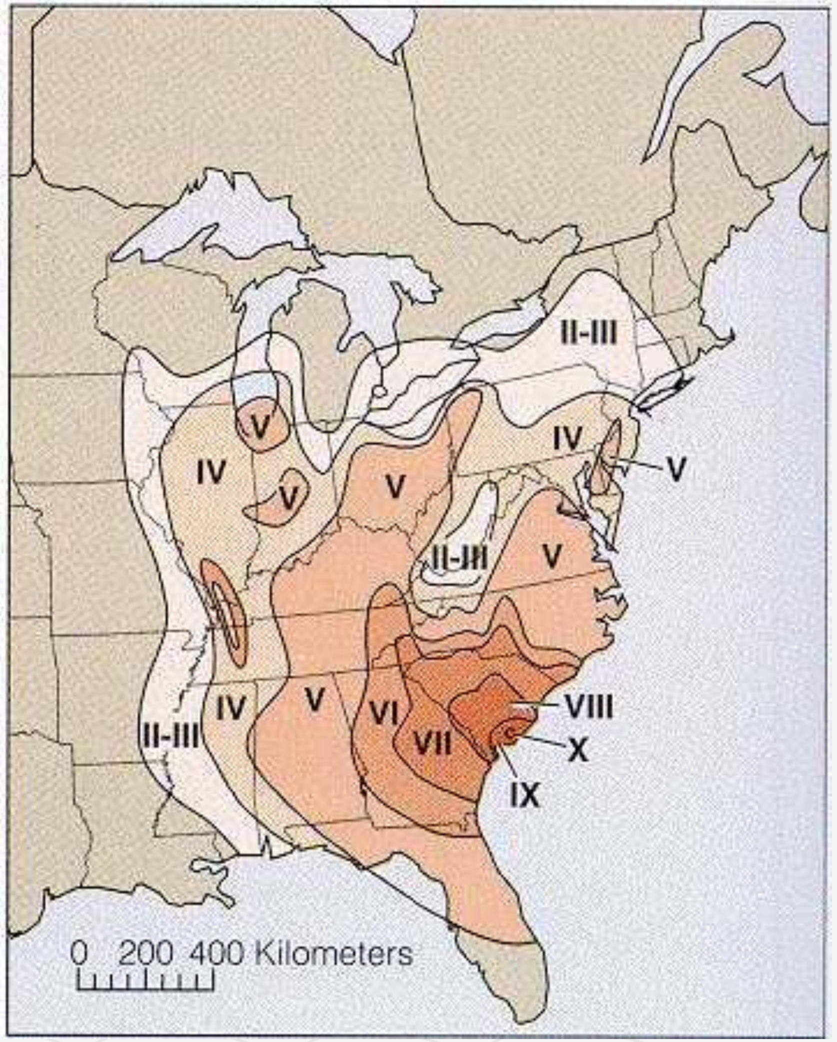
Modified Mercalli Intensity Scale

I	Not felt
II	Felt only by persons at rest
III–IV	Felt by persons indoors only
V–VI	Felt by all; some damage to plaster, chimneys
VII	People run outdoors, damage to poorly built structures
VIII	Well-built structures slightly damaged; poorly built structures suffer major damage
IX	Buildings shifted off foundations
X	Some well-built structures destroyed
XI	Few masonry structures remain standing; bridges destroyed
XII	Damage total; waves seen on ground; objects thrown into air

Isoseismal map

- Ideal isoseismal pattern shows a bell shape

- Isoseismal pattern depends on:
    - Condition at epicentre
    - The route of seismic wave from focus to the observation point
    - Geological conditions



Example of Intensity maps for
1886 Charleston, USA, earthquake.

Earthquake Location

Terms related to seismology

Epicenter distance

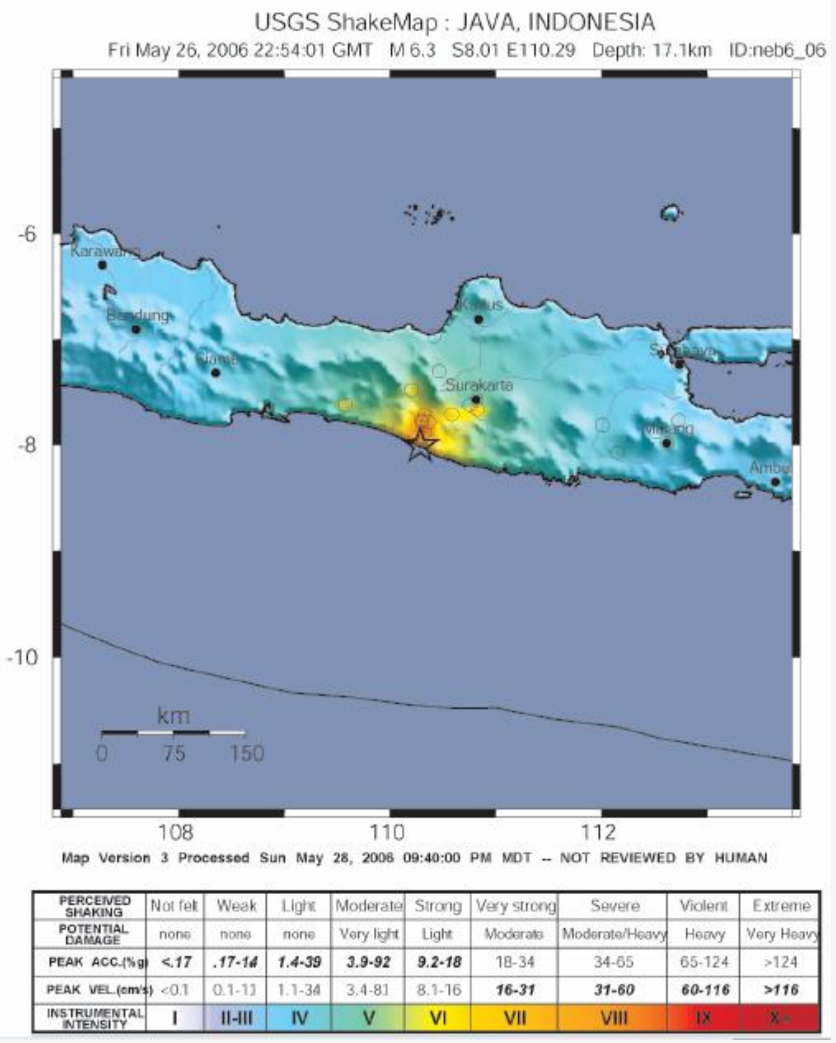
Hypocenter distance

Station azimuth

Take-off angle

Earthquake Location

- The basic principles
  - S-P location (manual)
  - location by inversion
  - single station location
  - depth assessment
  - velocity models

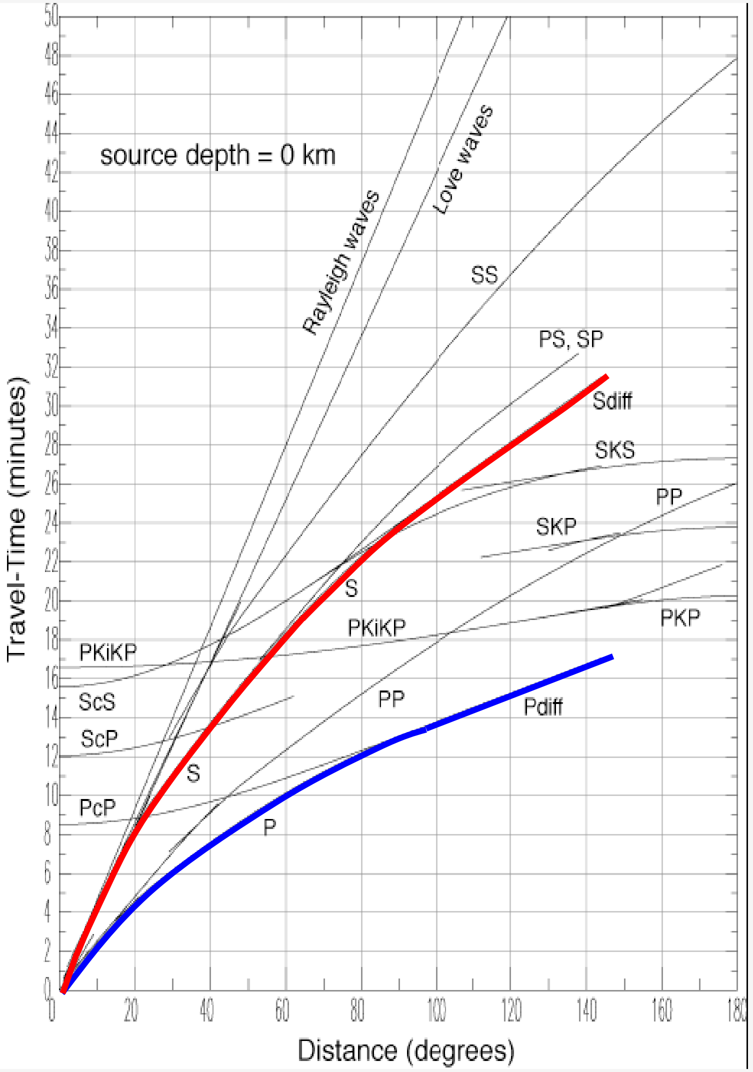


S-P time

- Time between P and S arrivals increases with distance from the focus.
  - A single trace can therefore give the origin time and distance (but not azimuth)

$$T_S - T_P = D\left(\frac{1}{V_S} - \frac{1}{V_P}\right)$$

approximates to $D = 8\left(T_S - T_P\right)$



How to calculate the distance from earthquake source to the recorded seismic stations

This can be calculated using the difference between the arrival times of P and S waves

Travel times for P and S waves are given by:

$Tp=D/vp$ & $Ts=D/vs$

Arrival times of P and S waves are given by

$tp=Tp+To$ & $ts=Ts+To$

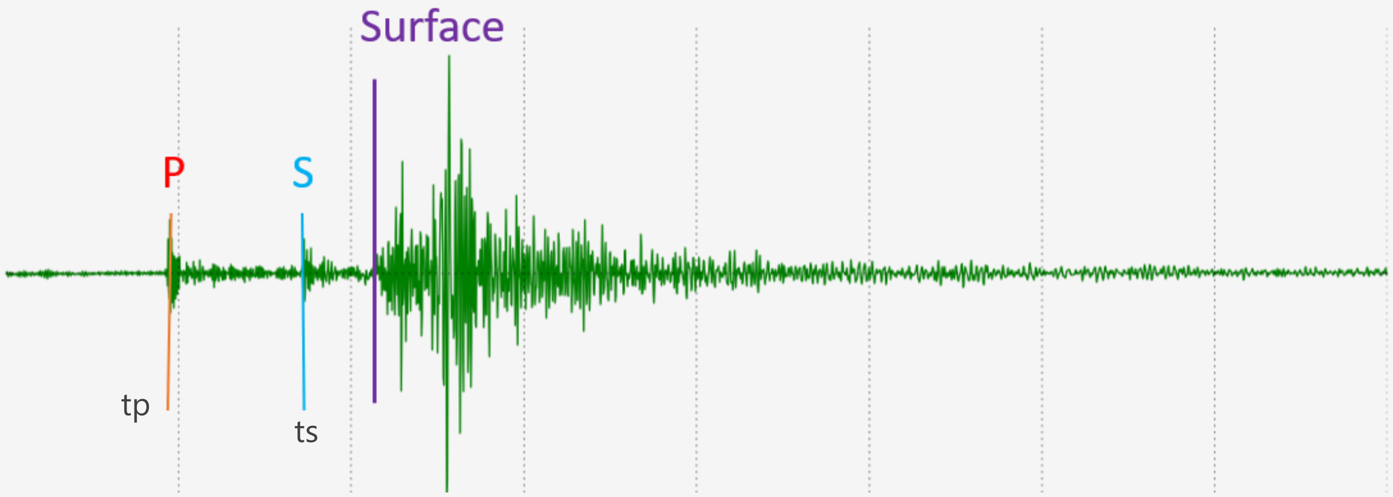
Then;

$tp=D/vp+To$ & $ts=D/vs+To$

Then,

$ts-tp=D/vs-D/vp$

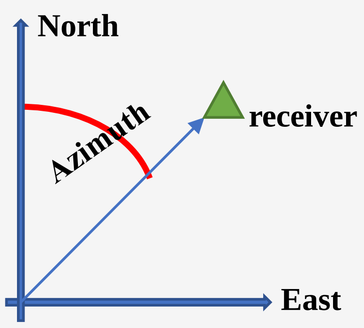
$D= (ts-tp)vsvp/(vp-vs)$



where tp and ts are the arrival times of P and S waves, Tp and Ts are the travel times of P and S waves, To is the origin time of earthquake, and D is the **hypocenter distance**.

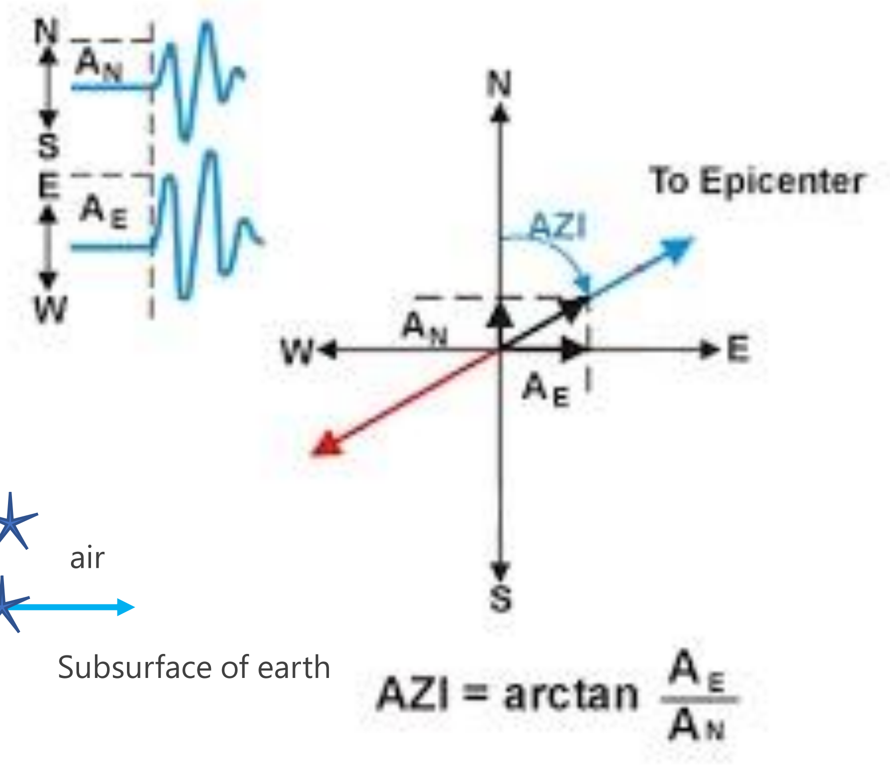
How to calculate the azimuth from the recorded seismic stations to the earthquake source

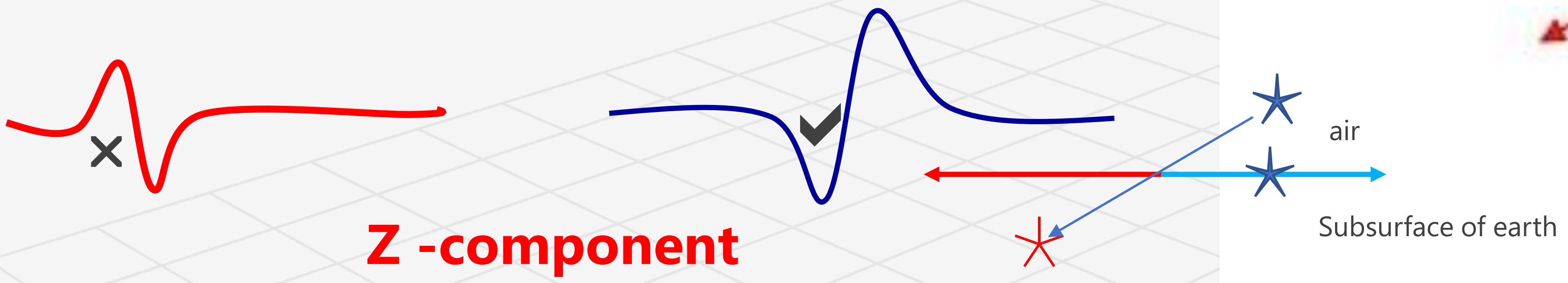
- In seismology we call the direction a receiver is from a source the azimuth:



- The azimuth is always measured clockwise from North and varies between 0º and 360º.

- This can be measured using the P - wave amplitudes and polarities from two horizontal components at a given seismic station.





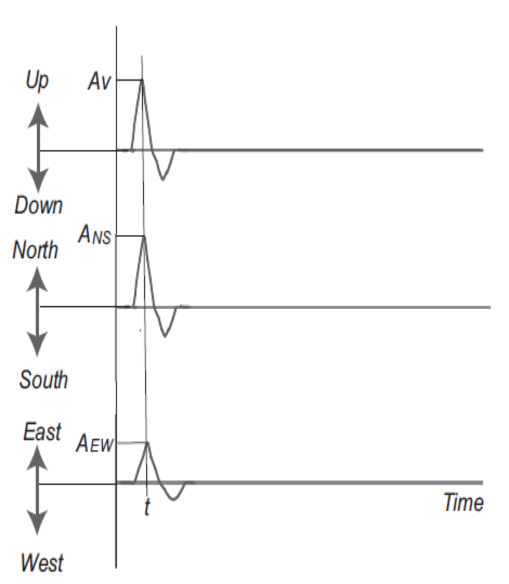
Examine the true location of the epicenter based on the polarity of the P-wave first arrival

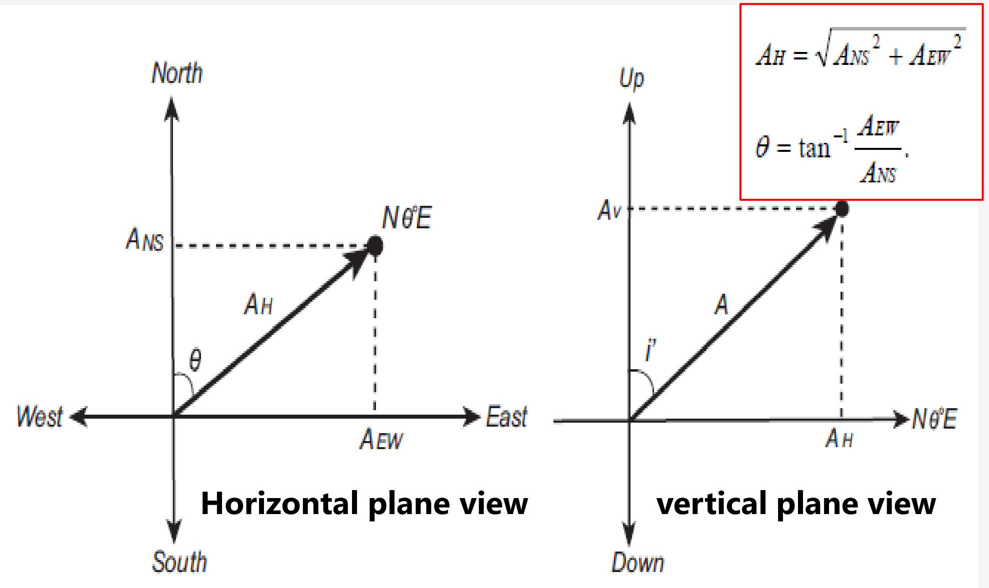
The location of epicenter may change from the north-east direction to the south-west direction base on the polarity of P-wave recorded by the vertical component. If the polarity of the first arrival of P-wave recorded on the vertical component is down, then the epicenter location will be fixed to the north-east direction. Otherwise, if the polarity of the first arrival of P-wave recorded on the vertical component, the epicenter location will change to be located to the south-west direction.

Example

Procedure

To obtain the particle motion of an observed seismic wave, suppose amplitudes of vertical and two horizontal (North-South and East-West) component seismograms are A_V, A_{NS}, and A_{EW}, respectively, at a certain time.

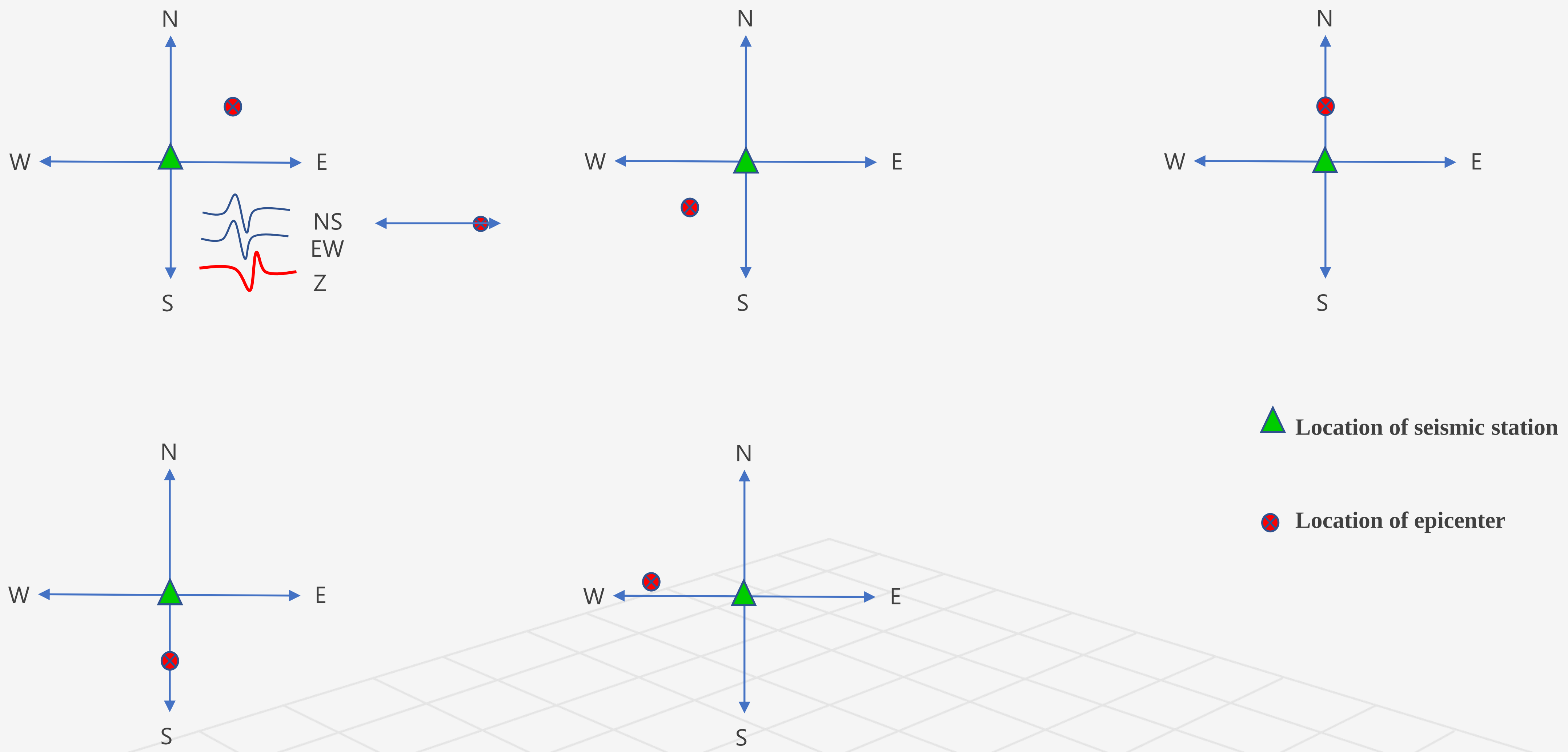




Since this direction is parallel to the direction of the *P*-wave particle motion, the direction is Nθ°E when the polarity of the *P* wave is down. Otherwise, it is N(θ+180°)E. The reason is that the *P* wave travels in underground, not through air.

Different examples should be proposed by students

Examples: predict the particle motions of P-wave recorded at the seismic station:

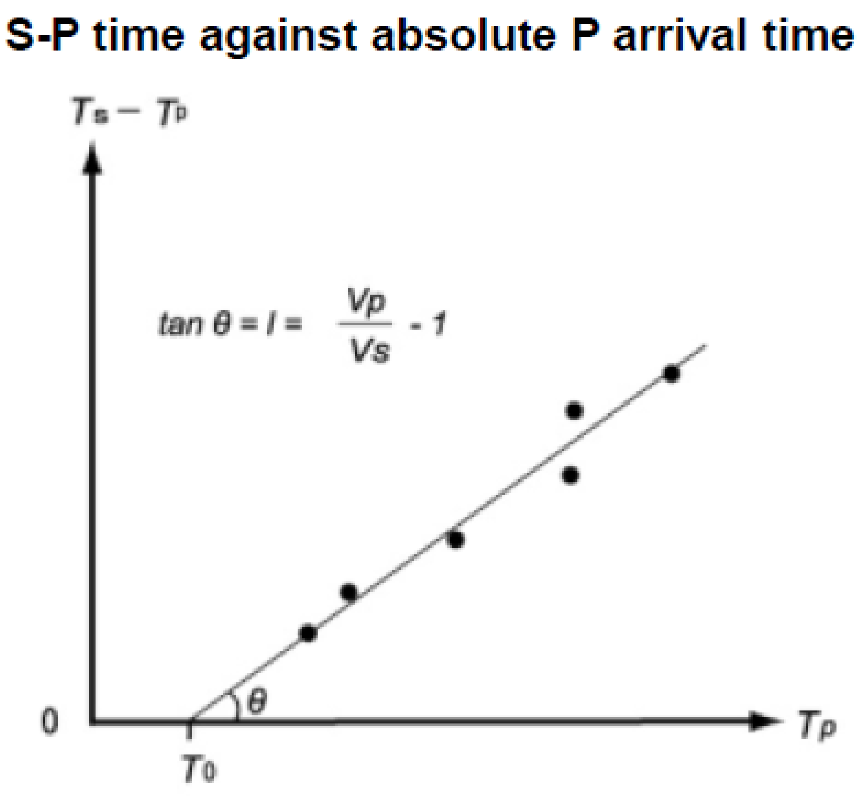


Origin time and Poisson's ratio

- The S-P time is plotted against the absolute P arrival time. As S-P time goes to zero at the source, by back-projecting the best fit line through the picks we retrieve the origin time of the event. The slope gives Vp/Vs-1. Gives the origin time (where S-P time = 0).

- Determines Vp/Vs (assuming it's constant and the P and S phases are the same type – e.g. Pn and Sn, or Pg and Sg).

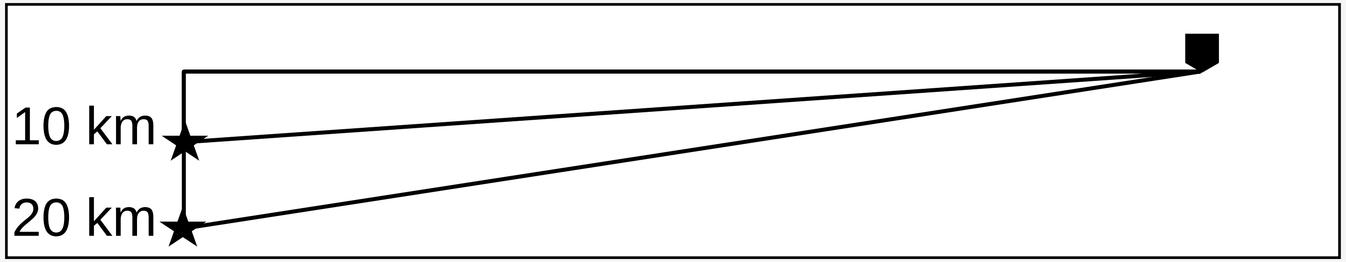
- Indicates pick errors



$$poission\ ratio\ is\ given\ by\ \frac{1}{2}\left[1-\frac{1}{(V_P/V_S)-1}\right]$$

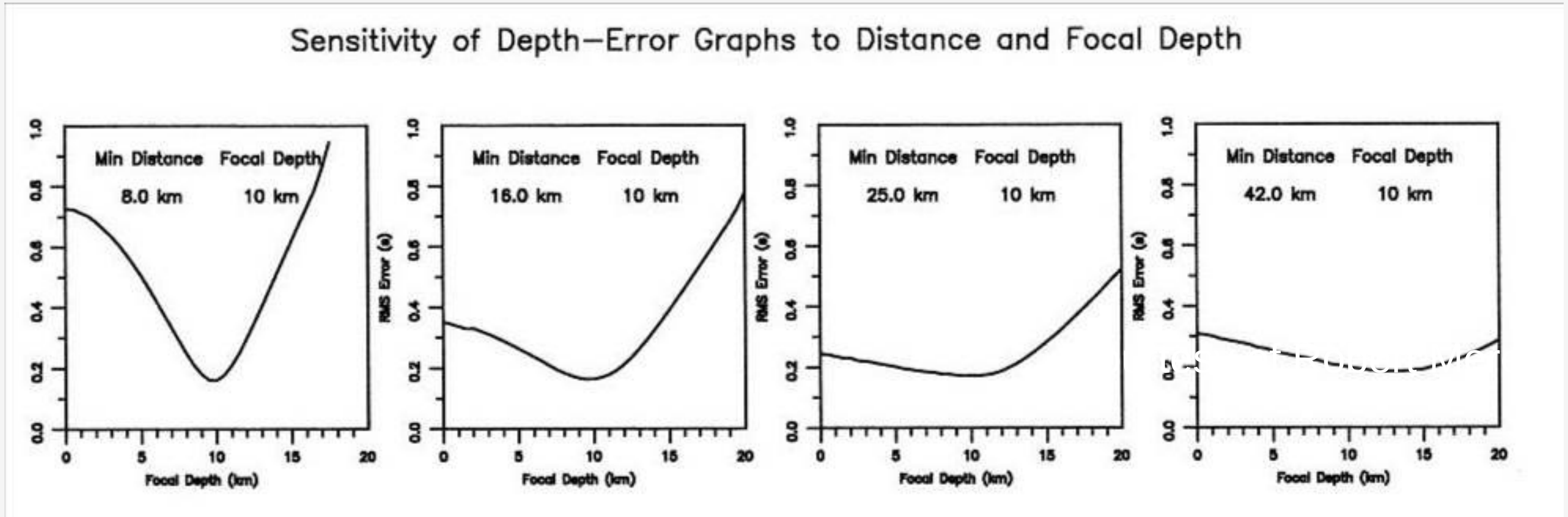
Depth estimation

- The distance between the station and the event is likely to be many kilometers. Therefore a small variation in focal depth (e.g. 5 km) will have little effect on the distance between the event and the station.
- Therefore the S-P time and P arrival time are insensitive to focal depth



Knowing the depth of the event important for all applications of earthquake seismology, and especially for tsunami warnings. The tsunami is generated by disturbances in the sea floor, so if the event to too deep to rupture (or significantly deform) the seabed then there is a greatly reduced tsunami risk – there is still a risk of a tsunami generated by a landslide triggered by the ground shaking, but the risk is smaller than for a significant seabed rupture. Therefore it is important to know if the epicenter is 5 km or 50 km depth. Unfortunately the geometry of the stations is likely to make the depth very uncertain if we only have S-P or P arrival times. The station spacing is likely to be several tens of kilometers at best (and for the Indian Ocean at the moment it is more like hundreds of kilometers in most regions). For example, the backbone coverage of the US Advance National Seismic System has a station spacing of roughly 280 km (although the coverage is far greater in high hazard areas, like California, where there is about a 20 km spacing on average). With large station intervals the distance between the quake and the station is very insensitive to the depth of the event.

Depth assessment

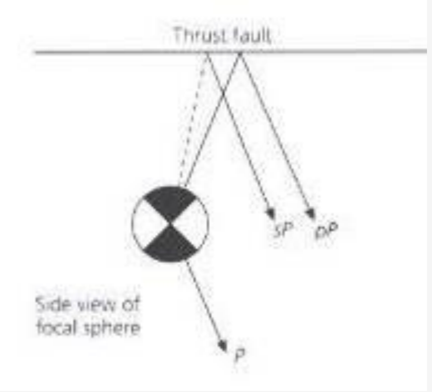


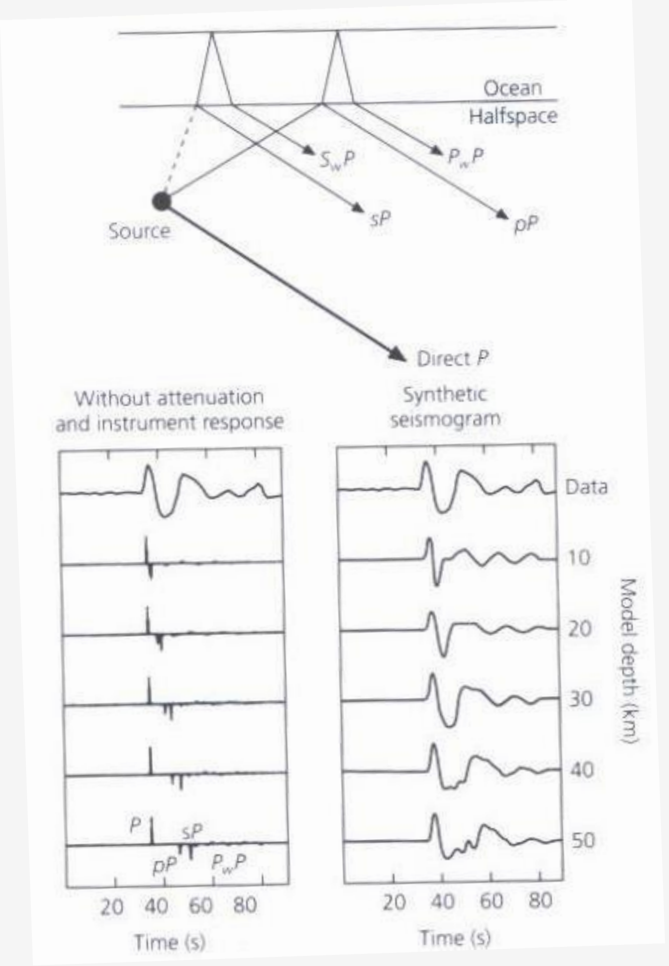
- Synthetic tests of variation in depth resolution - used in designing the network.
- As the distance for the quake to the nearest station increases the network becomes insensitive to the depth of the event (which was 10km for this test data).

- These are synthetic tests to illustrate how sensitivity to earthquake depth decreases with distance to the nearest station. When the closest station is less than 10 km from the event the location is very sensitive to the depth of the event, as shown by the high error for solution with different depths to the 10 km of the simulated earthquake. With the nearest station at 25 km, the depth is much more poorly resolved and if the closest station is more then 40 km from the event then there is almost no depth resolution.

Depth – pP and sP

- The phases that reflect from the Earth surface near the course (pP and sP) can be used to get a more accurate depth estimate.
- However, because energy is radiated from the source in all directions, for distant events it may be possible to identify a phase that travels up from the source and reflects from the surface of the Earth. If the energy travels as a P wave from the source this arrival is called pP; but energy that leaves the source as an S wave can also be converted to a P when reflected, this arrival is called sP. Almost the only difference in travel path between pP and P (or sP and P) is the path from the focus to the earth surface and back. Therefore, the difference in travel time for P and pP (or P and sP) is controlled by the depth of the source. So this can be used to get a much more accurate depth estimate for events when there isn't a very local seismic station.

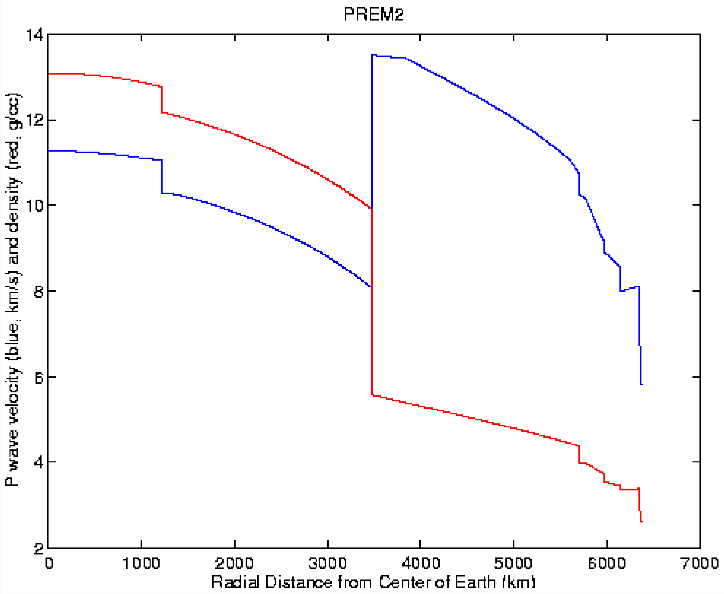


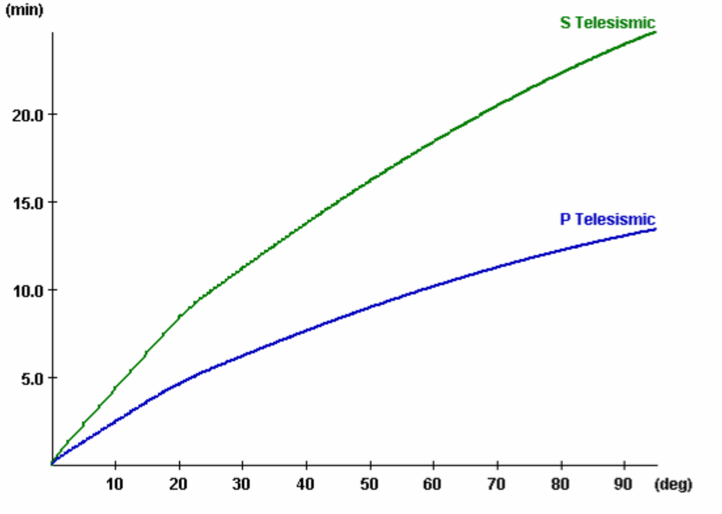


Stein and Wysession "An Introduction to Seismology, Earthquakes, and Earth Structure"

Velocity models

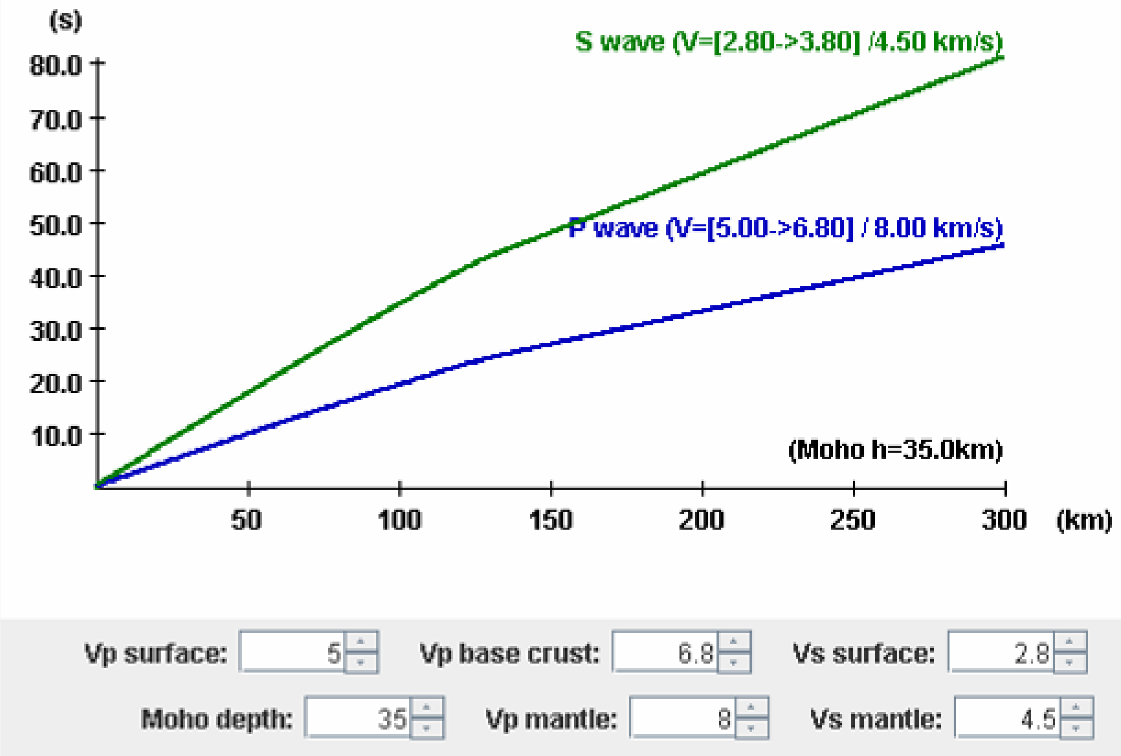
- For distant events may use a 1-D reference model (e.g. PREM) and station corrections. All locations rely on a good velocity model to convert time to distance, so now we'll look a little more a velocity models. Generally we use a simple 1D velocity model as this allows relatively easy conversion of time to distance, saving computational time. Generally this seems to give adequate locations (with small RMS misfits). For distant events we use a standard reference earth model (e.g. PREM, AK135, ISAP91) which includes the velocity structure of the deep earth. In this slide we see the velocity structure on the left and the distance-time plot this generates on the right.





Local velocity model

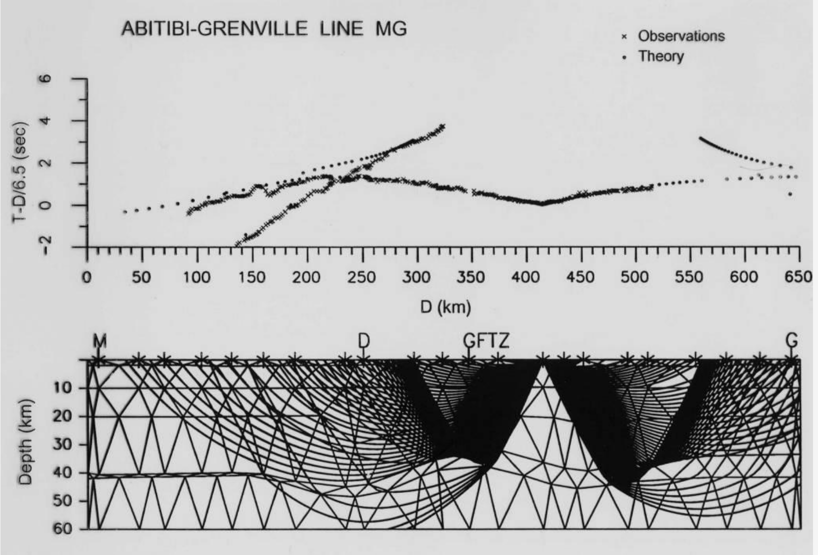
- For local earthquakes need a model that represents the (1D) structure of the local crust.
- For local earthquakes we usually use a local 1D model of the crust and upper mantle, often with a simple linear gradient and a velocity step at the Moho. The example on this slide is for a Moho depth of 35 km (a little thinner than the global average), a crustal velocity gradient from 5 km/s-6.8 km/s for P waves and an upper mantle P wave velocity of 8. This model would be adjusted for different areas, for example, changing the crustal thickness or allowing for slow sediments at the surface.

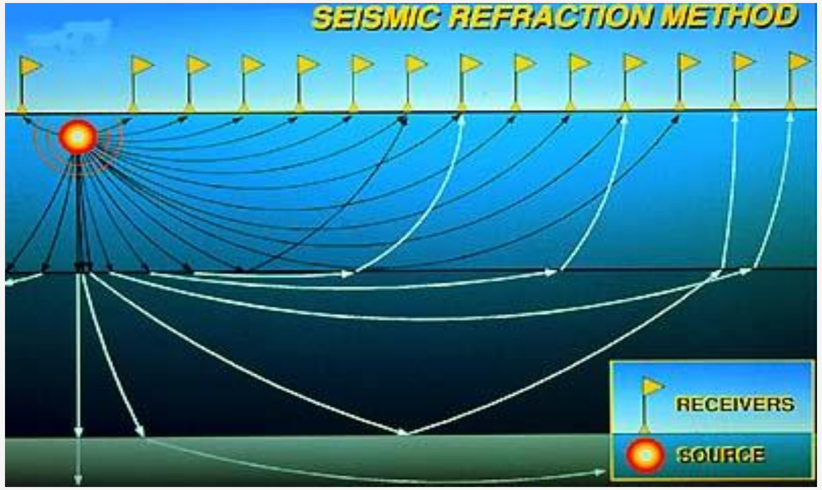


SeisGram2K

Determining the local velocity model

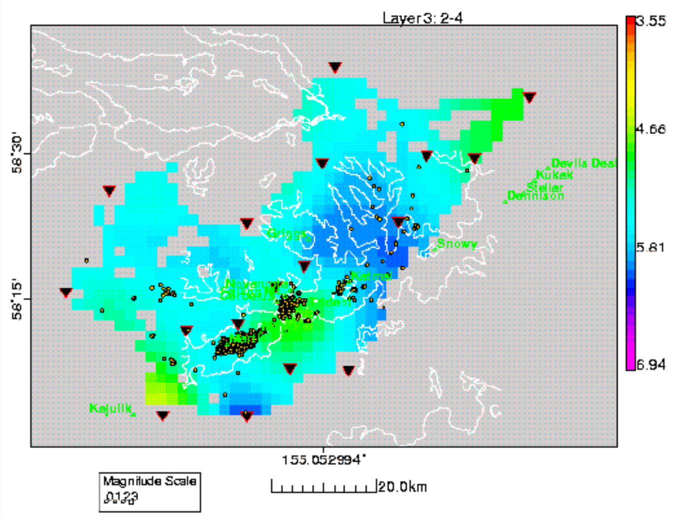
- Refraction data the best for Moho depth and velocity structure of the crust.
- To determine the local velocity structure we often use controlled source (also called active source) techniques. These are experiments done using controlled sources (explosions or vibrators on land, air gun shots at sea). They use a very dense coverage of either sources or temporary stations and optimize the distribution to look at the structure of the crust. The best method for determining seismic velocity of the crust and upper mantle and the depth of the Moho is the refraction method. However, these experiments are relatively expensive to run and until recently have mostly be done for academic research and so are not always available. Normal incidence reflection methods can also provide information on the velocity of the crust and depth/structure of the Moho, but do not get this information as precisely. But reflection surveying has been done by oil companies over much more extensive areas than refraction surveying (although often for only the upper crust).

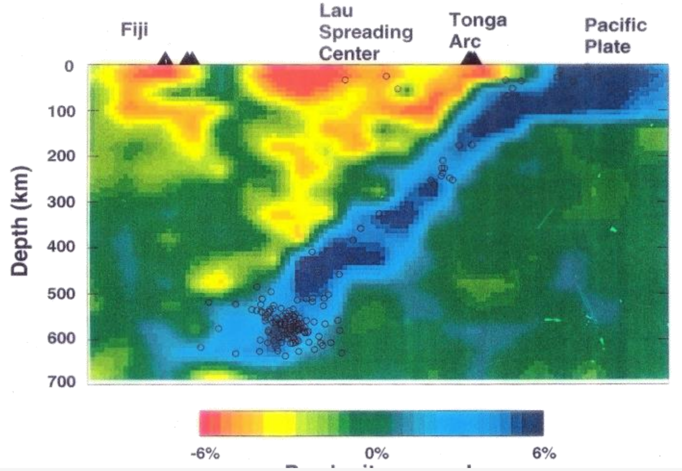




Tomography

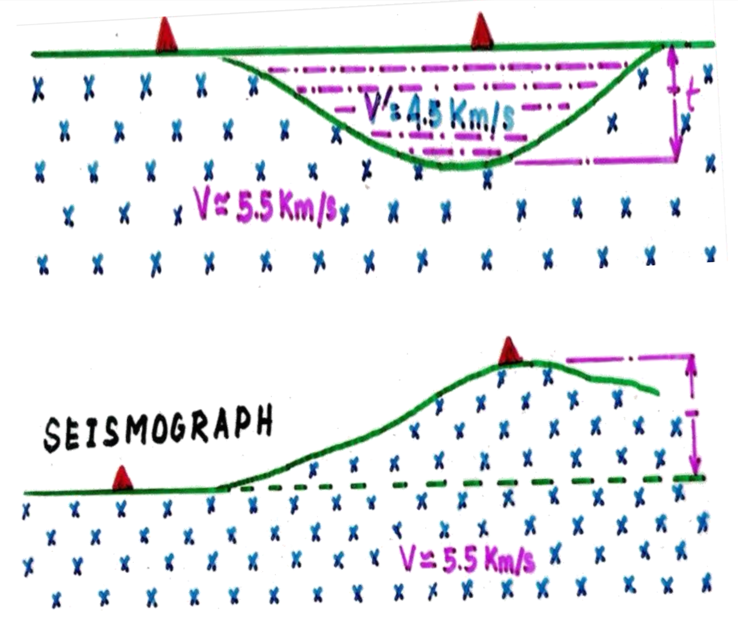
- Local tomography from local earthquakes can give crust and upper mantle velocities
- Regional/Global tomography from global events gives mantle velocity structure.
- seismic tomography is the method of mapping out the velocity structure in very well sampled area by looking at the variation in travel time of different paths through the area. Local tomography from local earthquakes or controlled sources can give detailed maps of the crustal velocity, for example the top image is the velocity (at 2-4 km below the surface) around a volcano in Alaska, resolved using the earthquakes from magma movement (black dots) recorded at a number of local stations (black triangles). The lower image is an example of regional seismic tomography done using regional or teleseismic earthquakes. Here we see a very clear variation in seismic velocity associated with the subduction Pacific Plate and melting in the mantle wedge above the plate. We can see that such images can provide very detailed information on the velocity variation, but require a large number of sources and stations and considerable computation, so high resolution local models are not widely available around the globe.

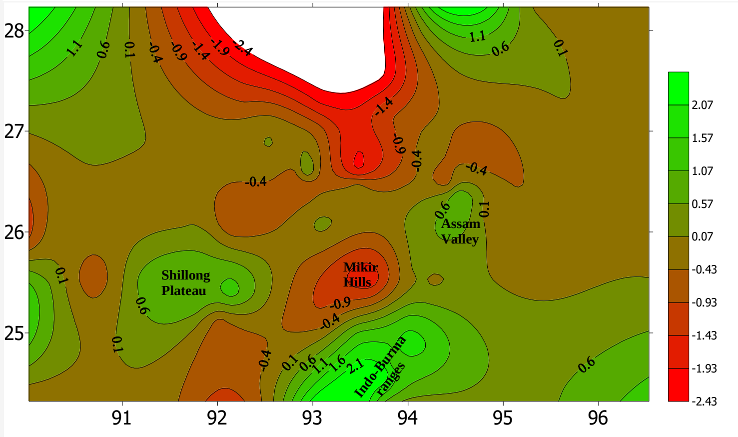




Station Corrections

- Station corrections allow for local structure and differences from the 1D model.
- The many refraction, reflection and tomographic studies around the globe show that the velocity structure of the crust is not a simple 1D model, but has considerable local variation. However, as mentioned earlier, a 1D model is far easier to use in the location calculation. One way of reducing the errors introduced by using a 1D model is to add station corrections. These are effectively a static shift applied to the arrival times at a recording station to account for local differences to the 1D model. For example a station located far above sea level will record delayed arrivals with respect to the 1D model, due to the extra distance the energy is traveling. Or a station on in a sedimentary basin will also have a delay due to the low velocity beneath the seismometer. However, it must be noted that station correction is a constant correction applied to all arrivals (from all distances and directions) and so can only correct for the structure very close to the station, not for deeper structure that will affect only arrivals from a certain distance and azimuth.

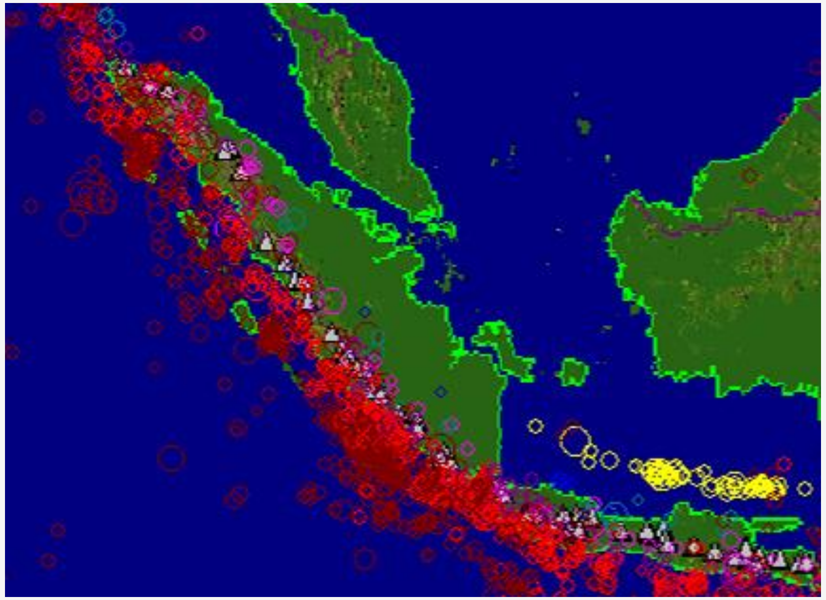


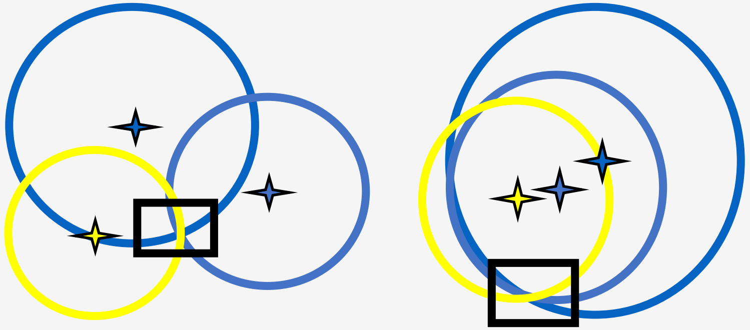


(Bhattacharya et al., 2005)

Location in subduction zones

- Poor station distribution
- Locations for tsunami warning centers often face a problem with station distribution. The smallest uncertainties are for events surrounded by stations. When all stations are at a similar azimuth then the location is much more poorly constrained. For shallow subduction zone earthquakes, this is often a problem as the seismometers are normally only deployed on land (only the JMA has ocean bottom seismometers for tsunami warnings) and the geological processes mean the land is almost all on the back arc side of the trench. Therefore, the stations are often all on one side of the earthquake.





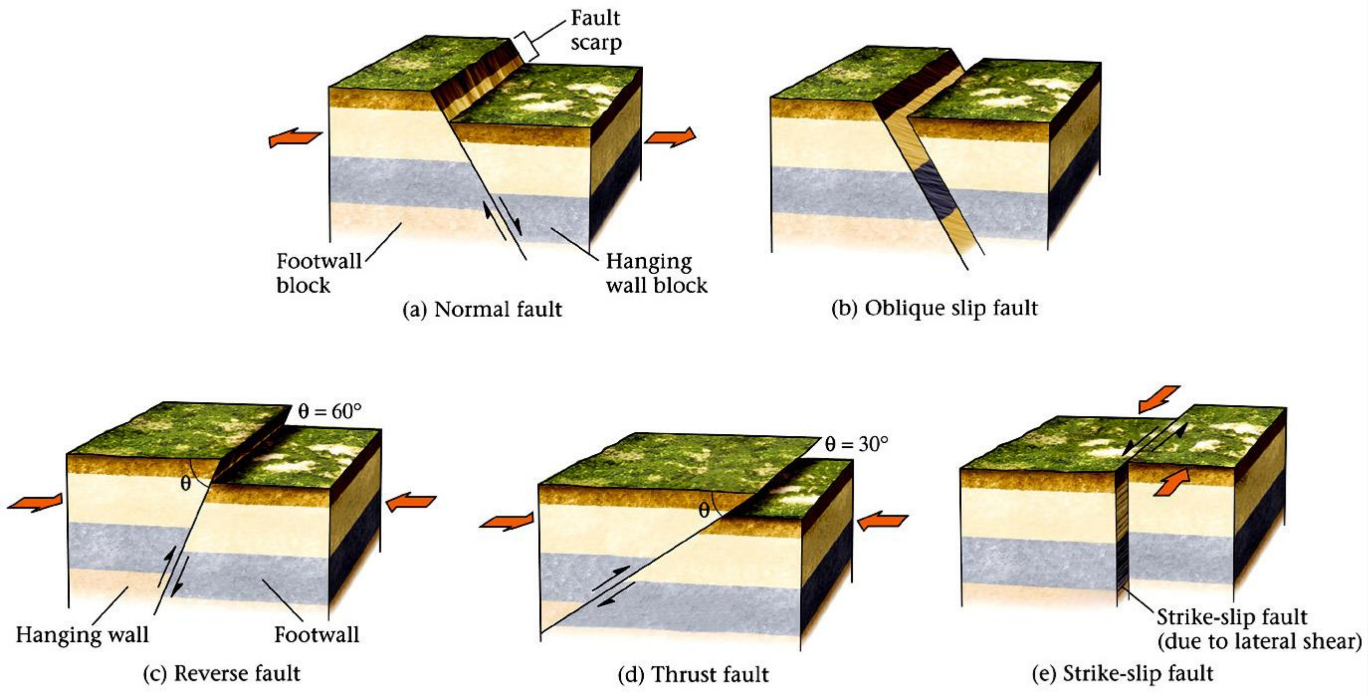
Focal mechanism

Types of Faults

In general, faults come in three different types: ***Normal, Reverse, and Strike-Slip***

Shallow angle (< 30°) reverse faults are called ***thrust faults***

Faults that have a mix of slip styles are called ***oblique slip faults***



(a) Normal fault
(b) Oblique slip fault
(c) Reverse fault
(d) Thrust fault
(e) Strike-slip fault

Why are there different types of faults?

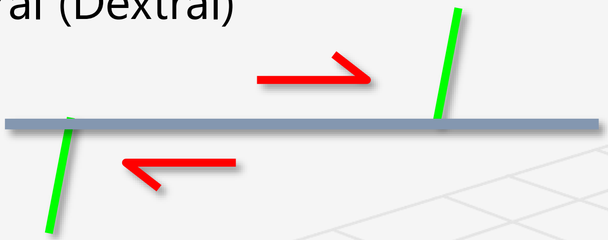
Normal Faults: from stretching of or extending rock; points on opposite sides of a fault are father apart after an earthquake

Reverse Faults: from contracting or squishing rock; points on opposite sides of the fault are closer together after an earthquake

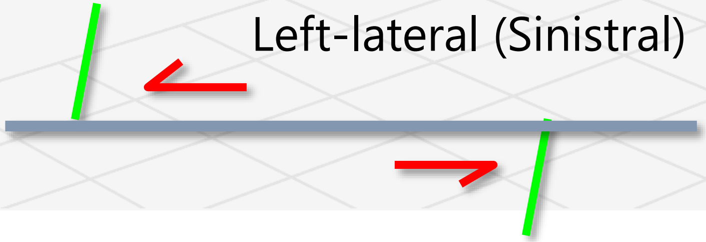
Strike-Slip: can form in either areas of stretching or squishing, material slides laterally past each side of the fault.

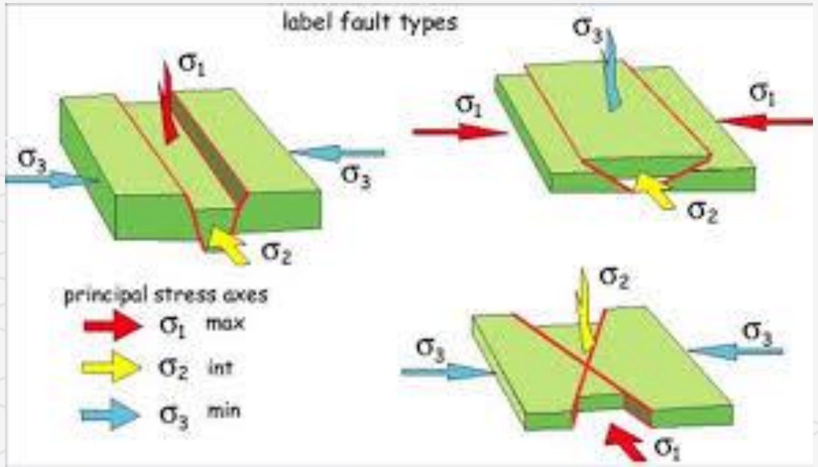
Described by sense of motion:

Right-lateral (Dextral)



Left-lateral (Sinistral)





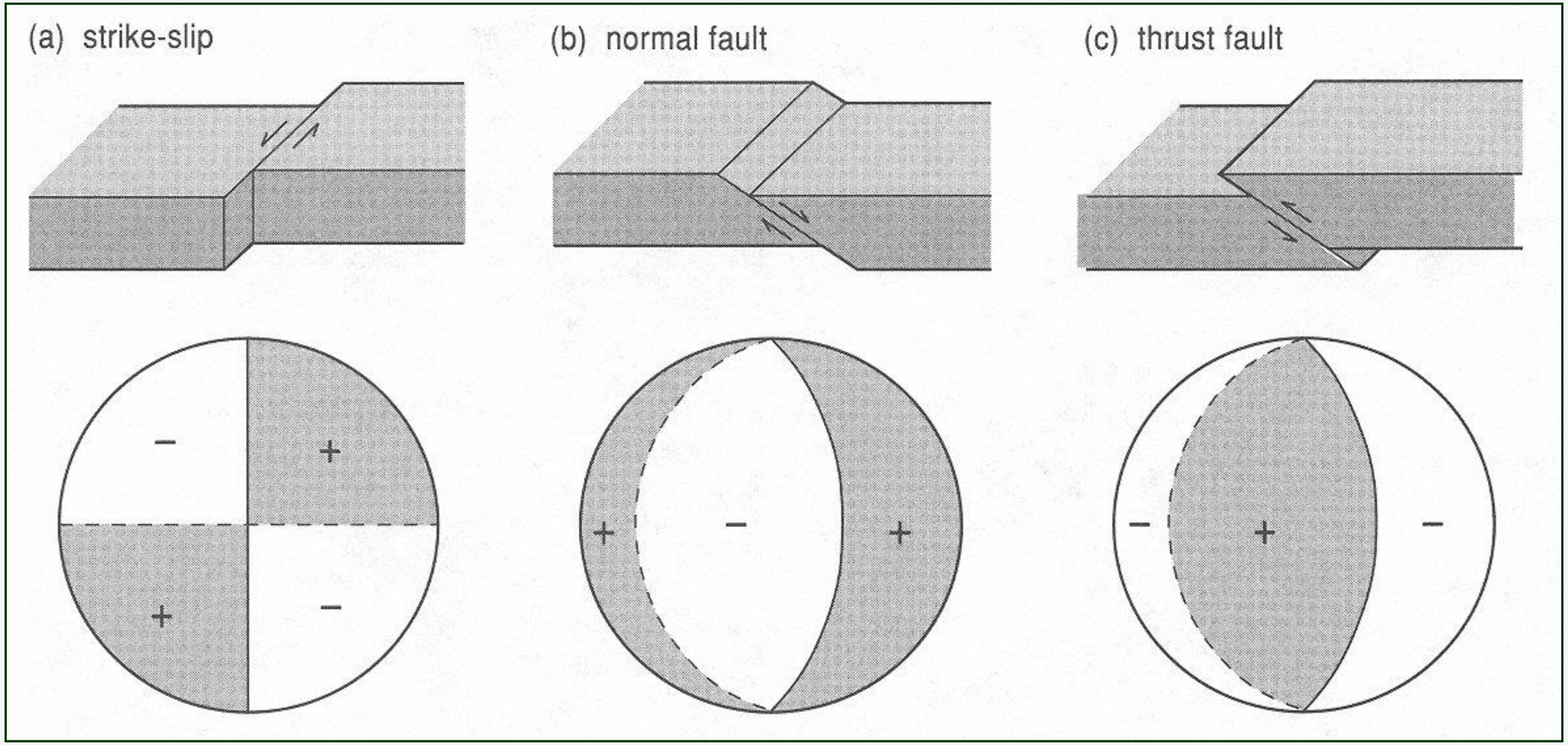
Focal Mechanism Solutions

Also called “beachball diagrams” “fault plane solutions”

Tell us the geometry and mechanism of the fault in a simple diagram

Generally from the moment tensor (which is more general), but originally calculated using first motions – done here to illustrate the concepts

Examples



Seismic "Beach Balls"

Project the pattern of initial motions intersecting an imaginary sphere around the source (the "focal sphere") onto a flat surface.

We use the radiation patterns of P-waves to construct a graphical representation of earthquake faulting geometry (two planes intersecting one another at right angles)

The symbols are called "Focal Mechanisms" or "Beach Balls", and they contain information on the fault orientation and the direction of slip.

Focal Mechanism

When mapping the focal sphere to a circle (beachball) two things happen:

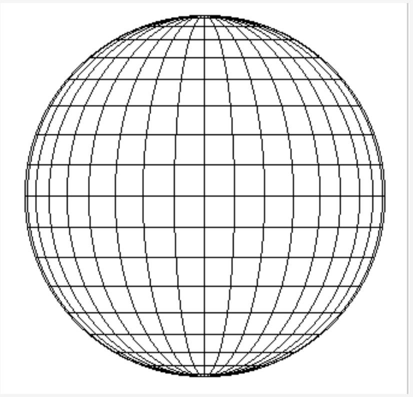
Lines (vectors) become points

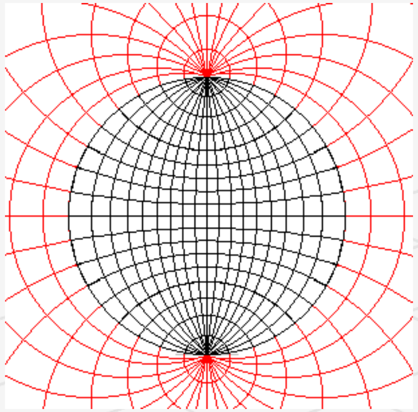
Planes become curved lines

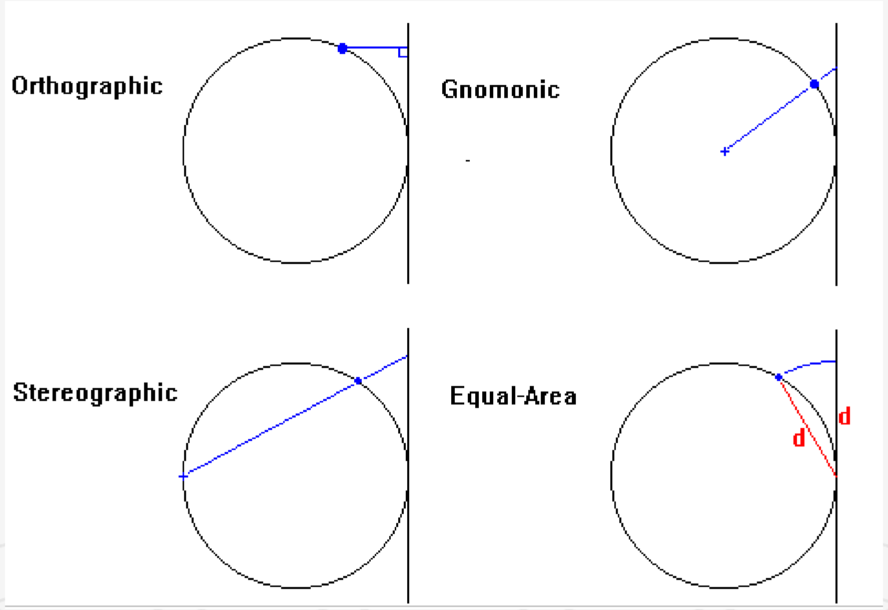
Two steps to understanding

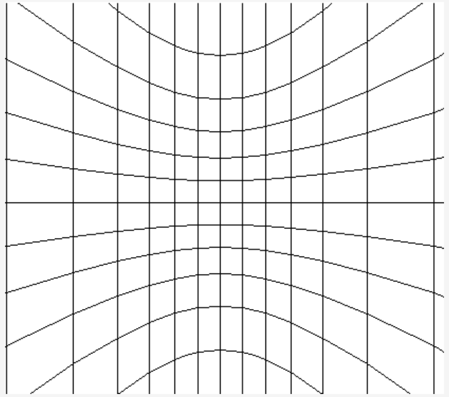
1) The stereographic projection

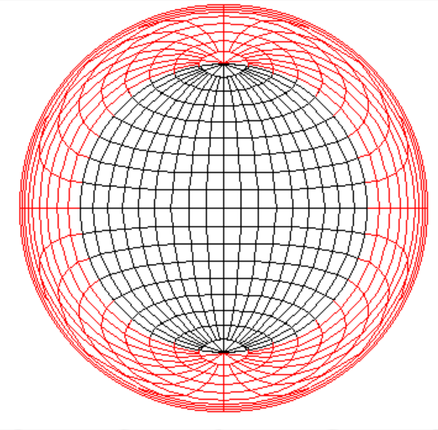
2) The geometry of first motions and how this is used to define fault motion.







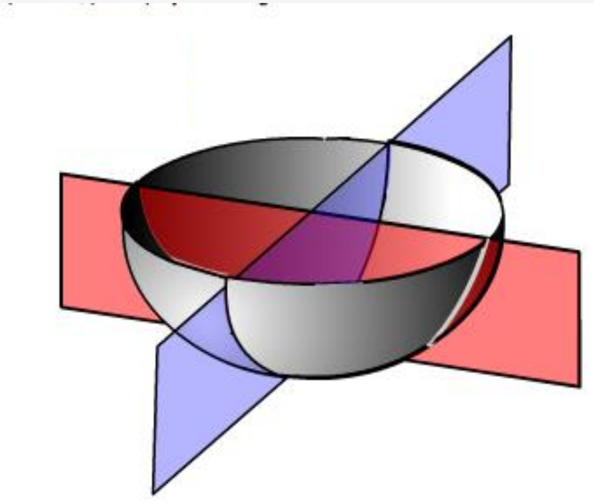


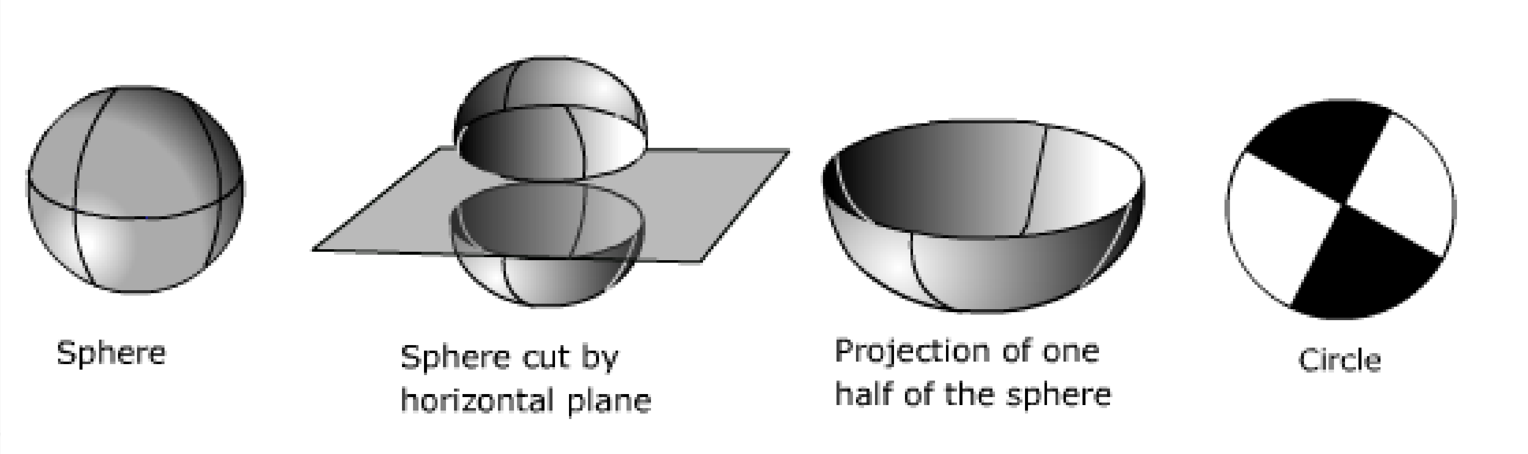


Stereographic projection

A method of projecting half a sphere onto a circle.

e.g. planes cutting vertically through the sphere plot as straight lines



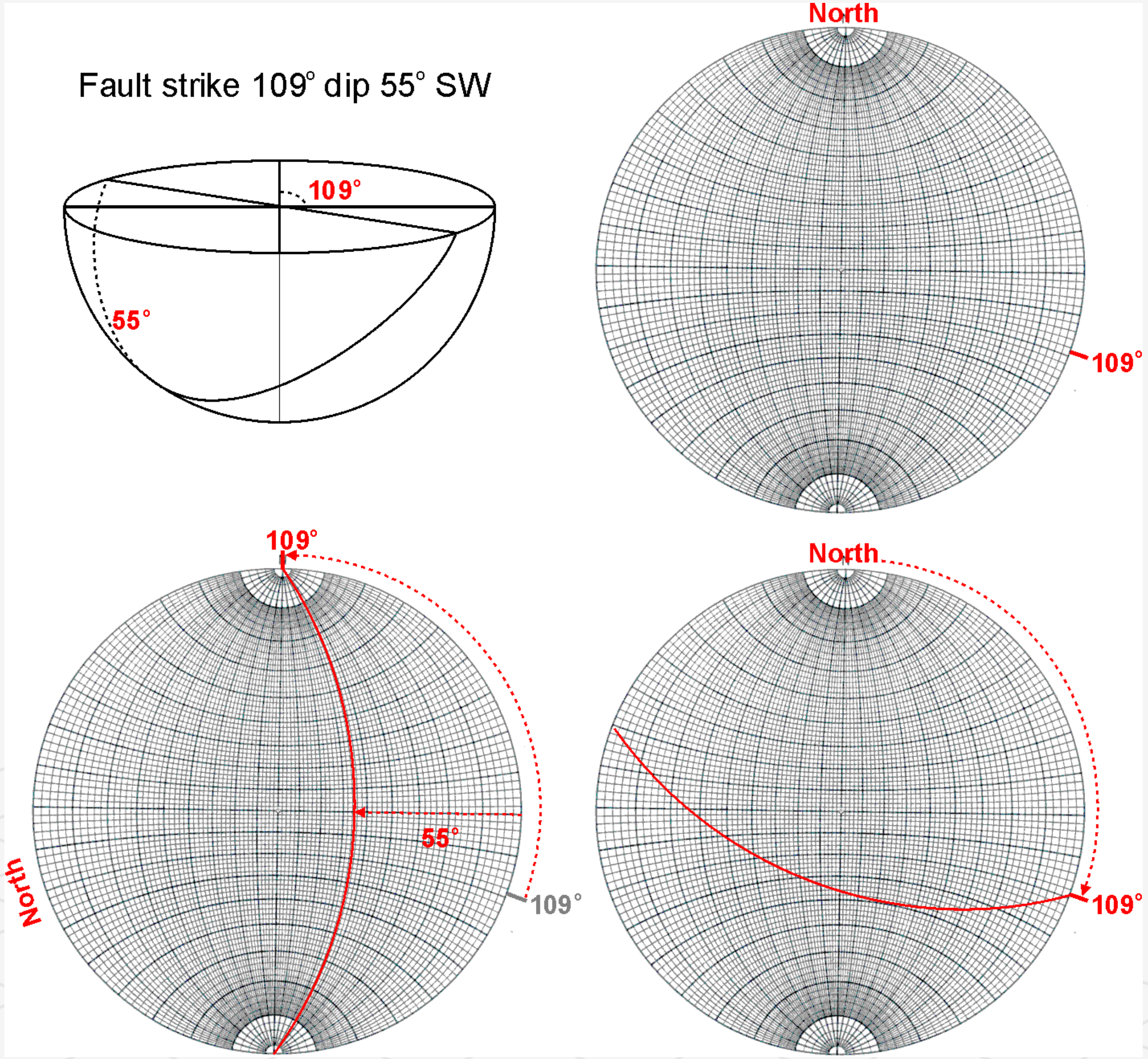


Images from http://www.learninggeoscience.net/free/00071/index.html

Stereonets

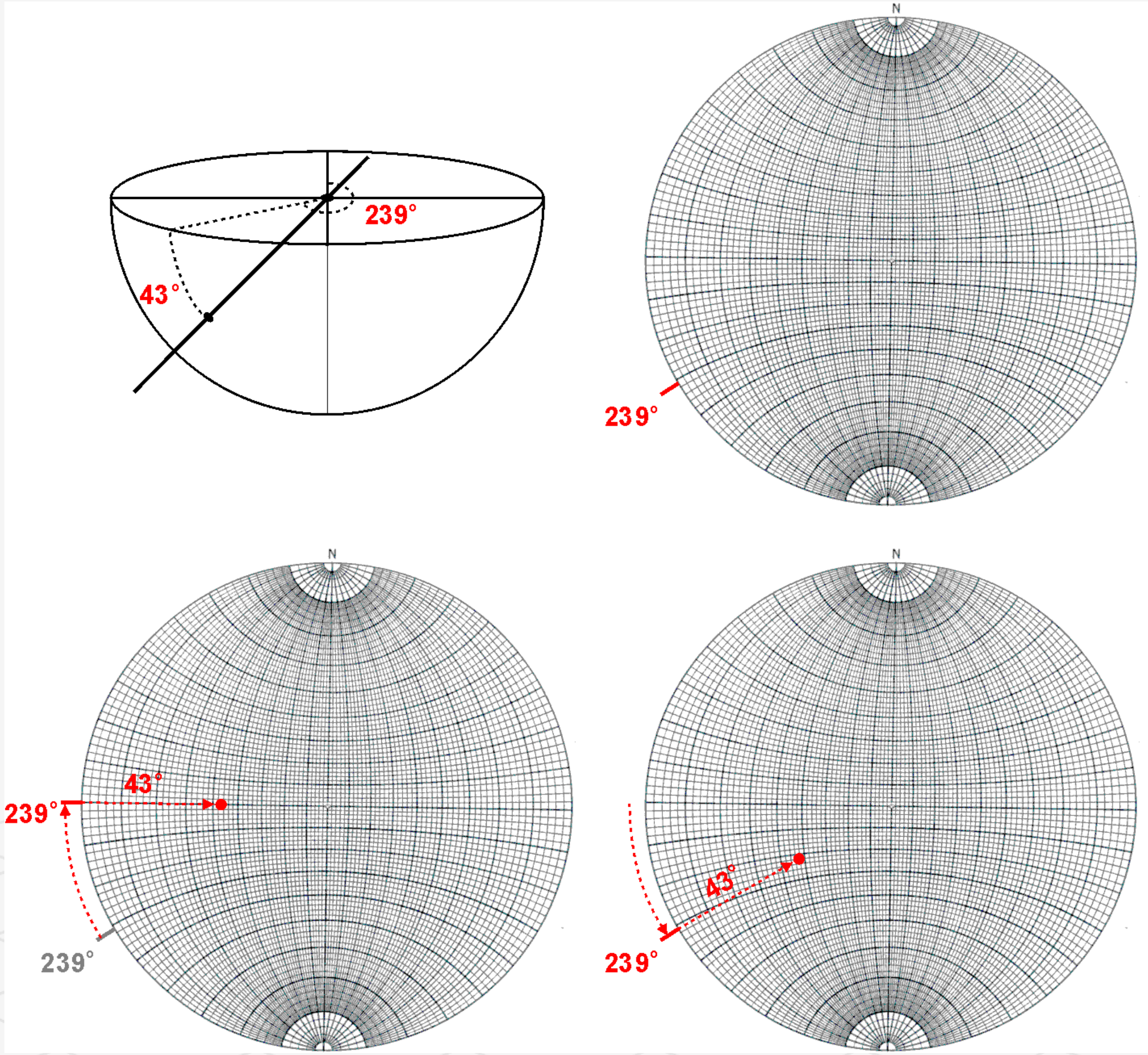
A template called a stereonet is used to plot data.

Example – plotting planes (e.g. faults)



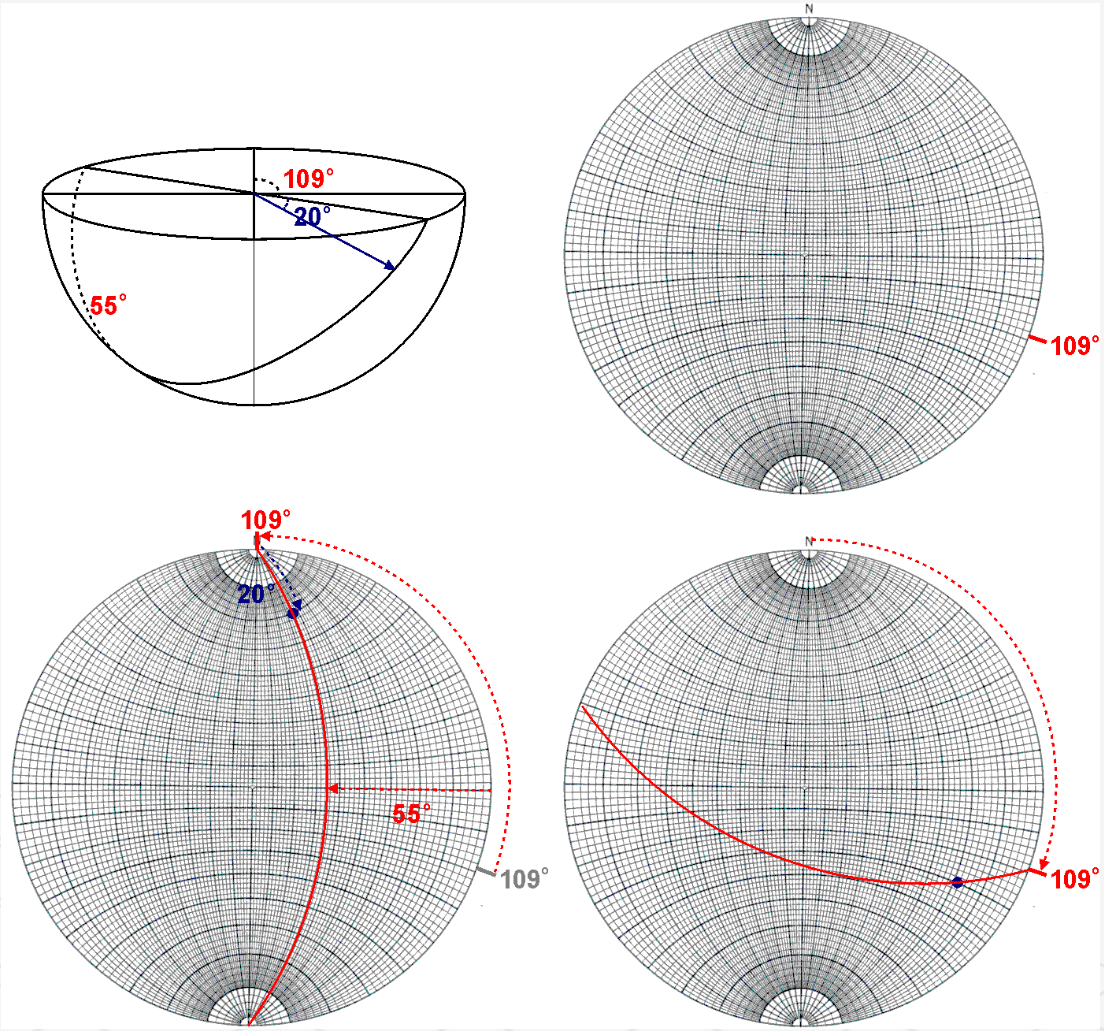
Stereonets

Example – plotting lines (e.g. ray paths)

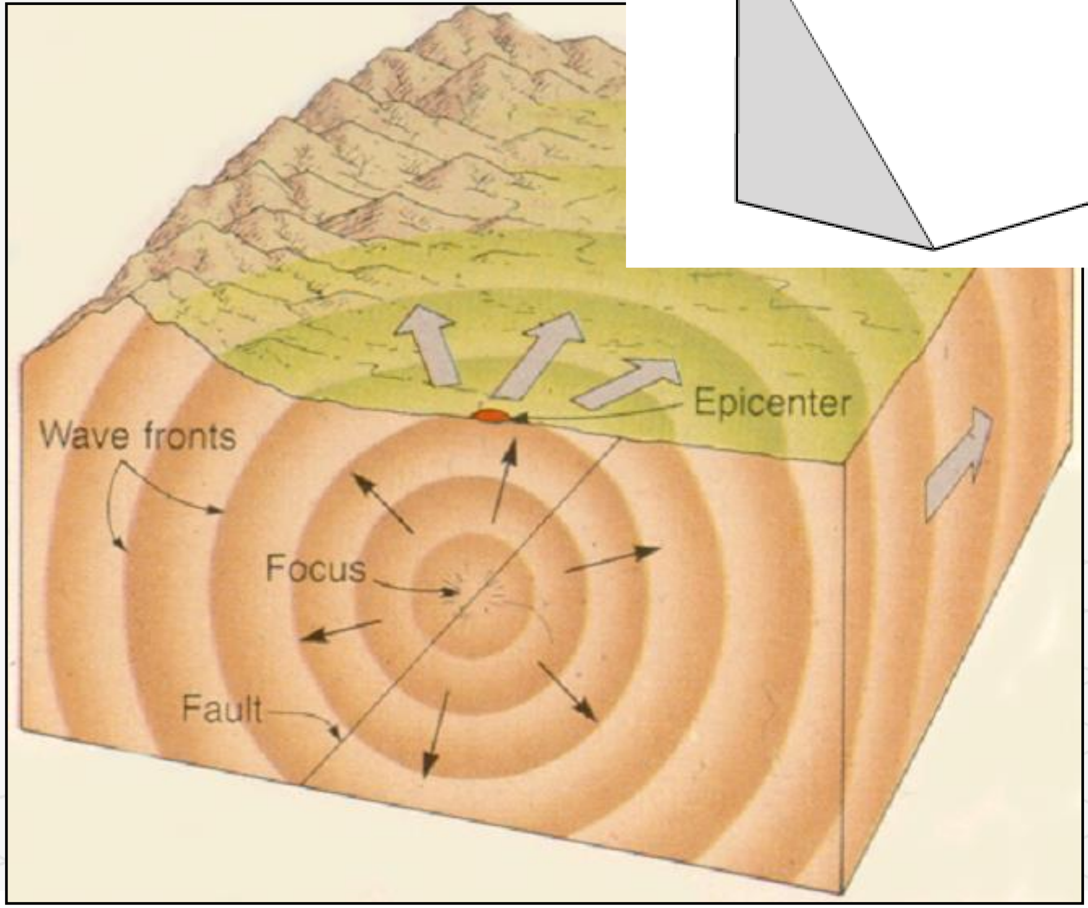


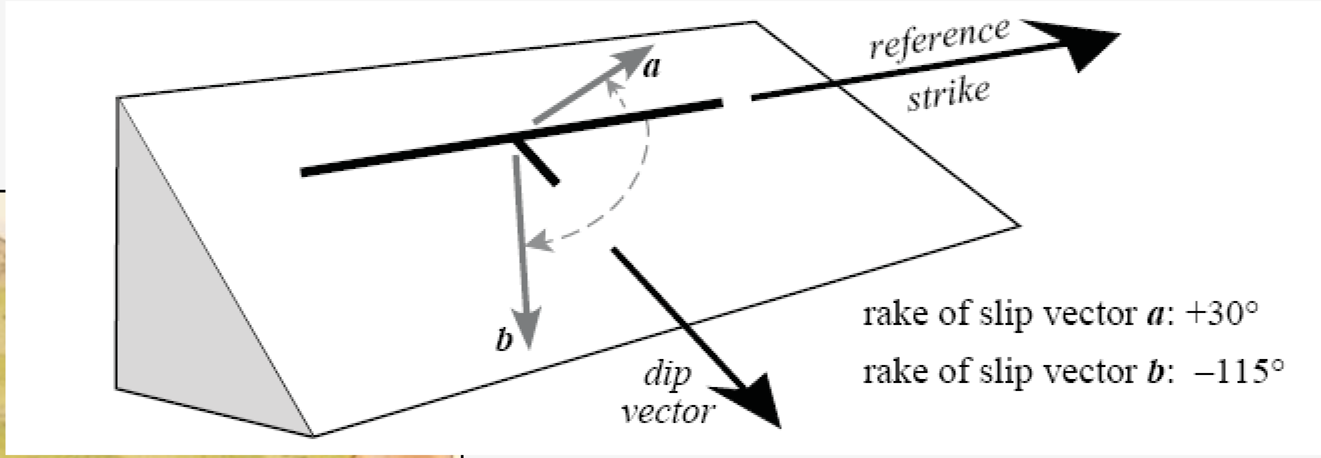
Stereonets

Example – pitch (or rake) of a line on a plane (e.g. the slip direction on a fault)

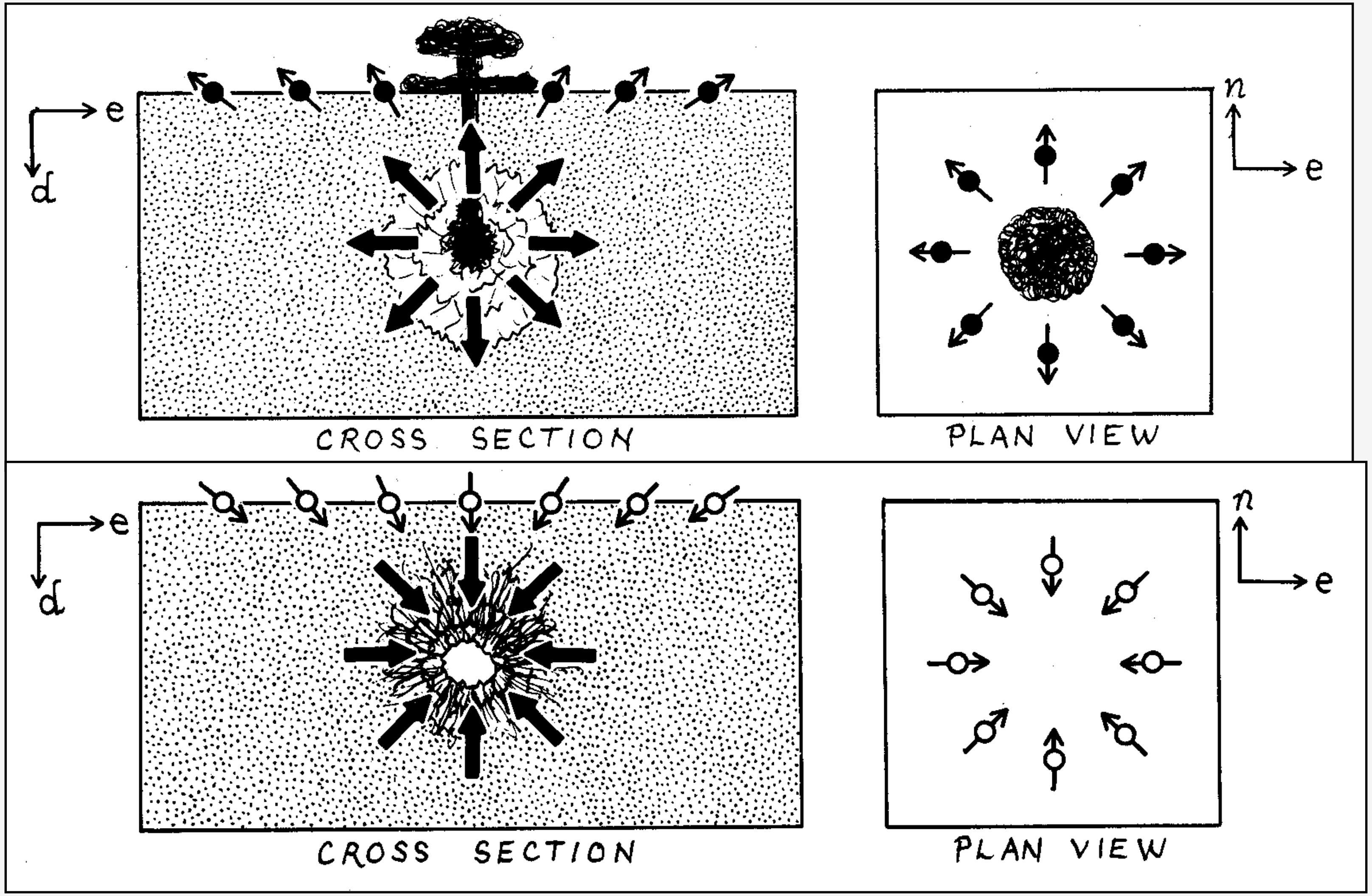


Refresher on terminology



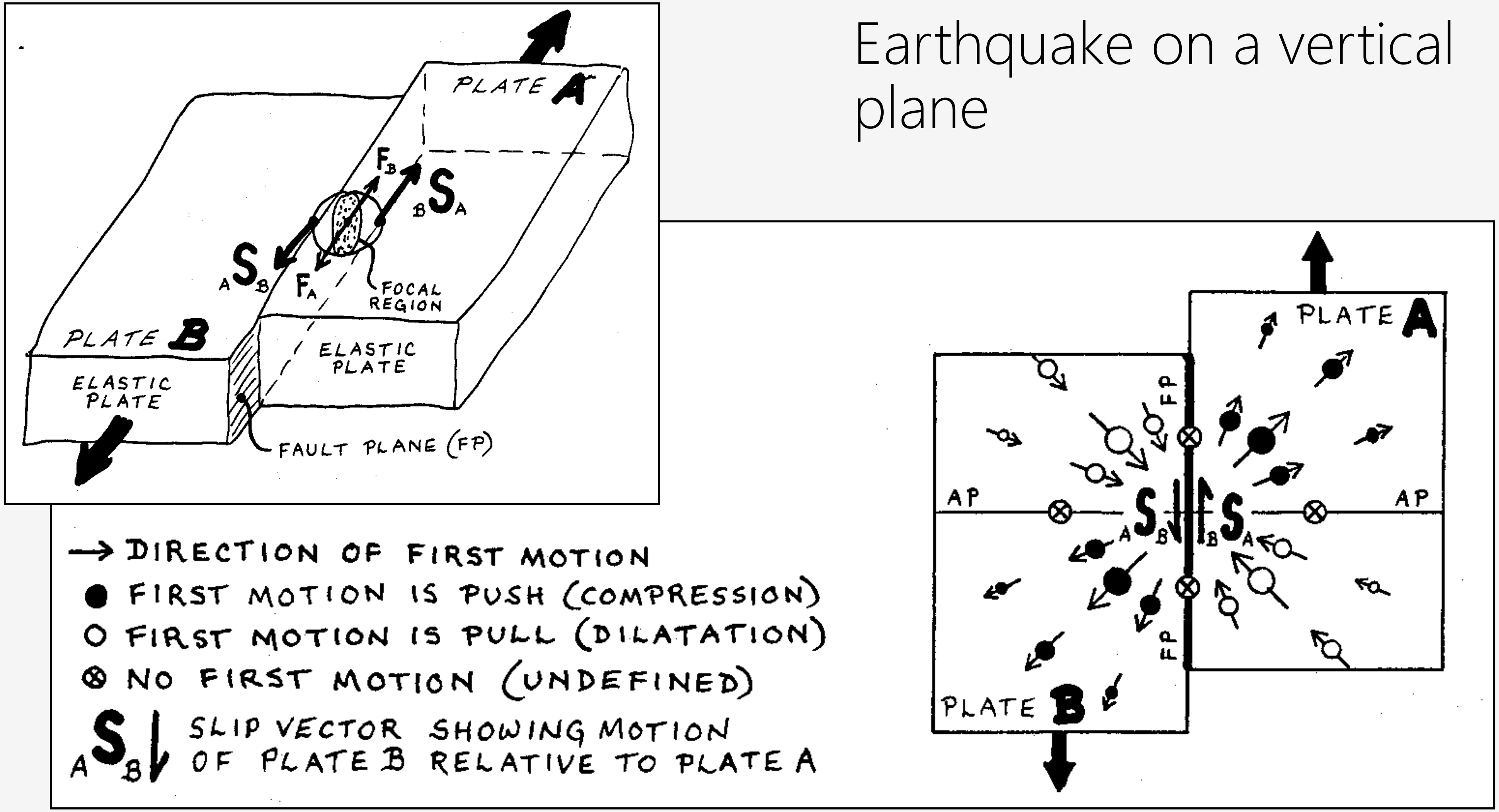


Energy and Polarity of "First Motions"



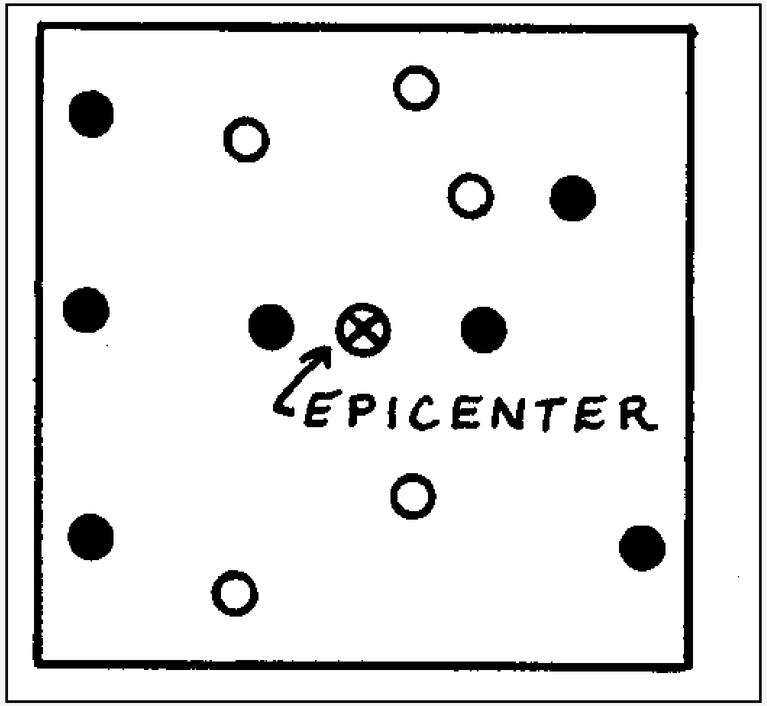
courtesy of Ian Hill, University of Leicester, UK

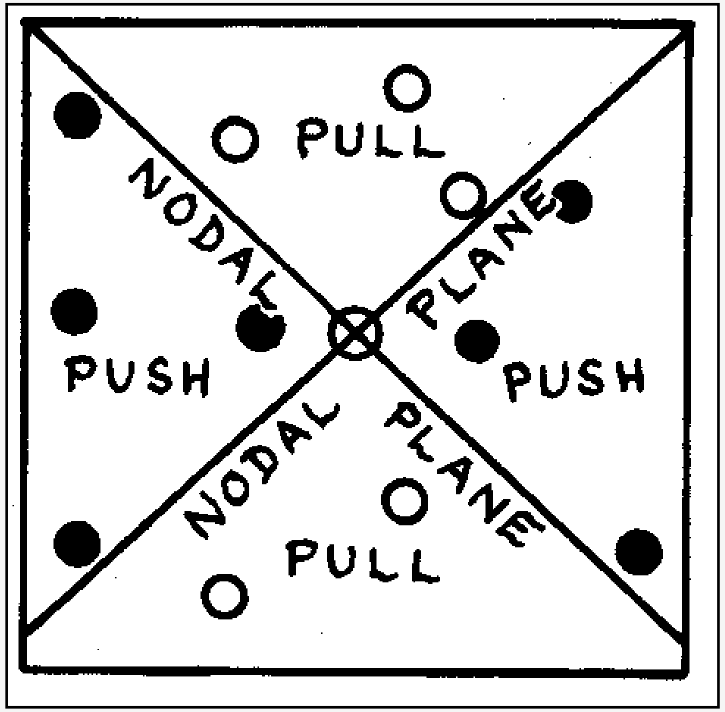
Earthquake on a vertical plane



courtesy of Ian Hill, University of Leicester, UK

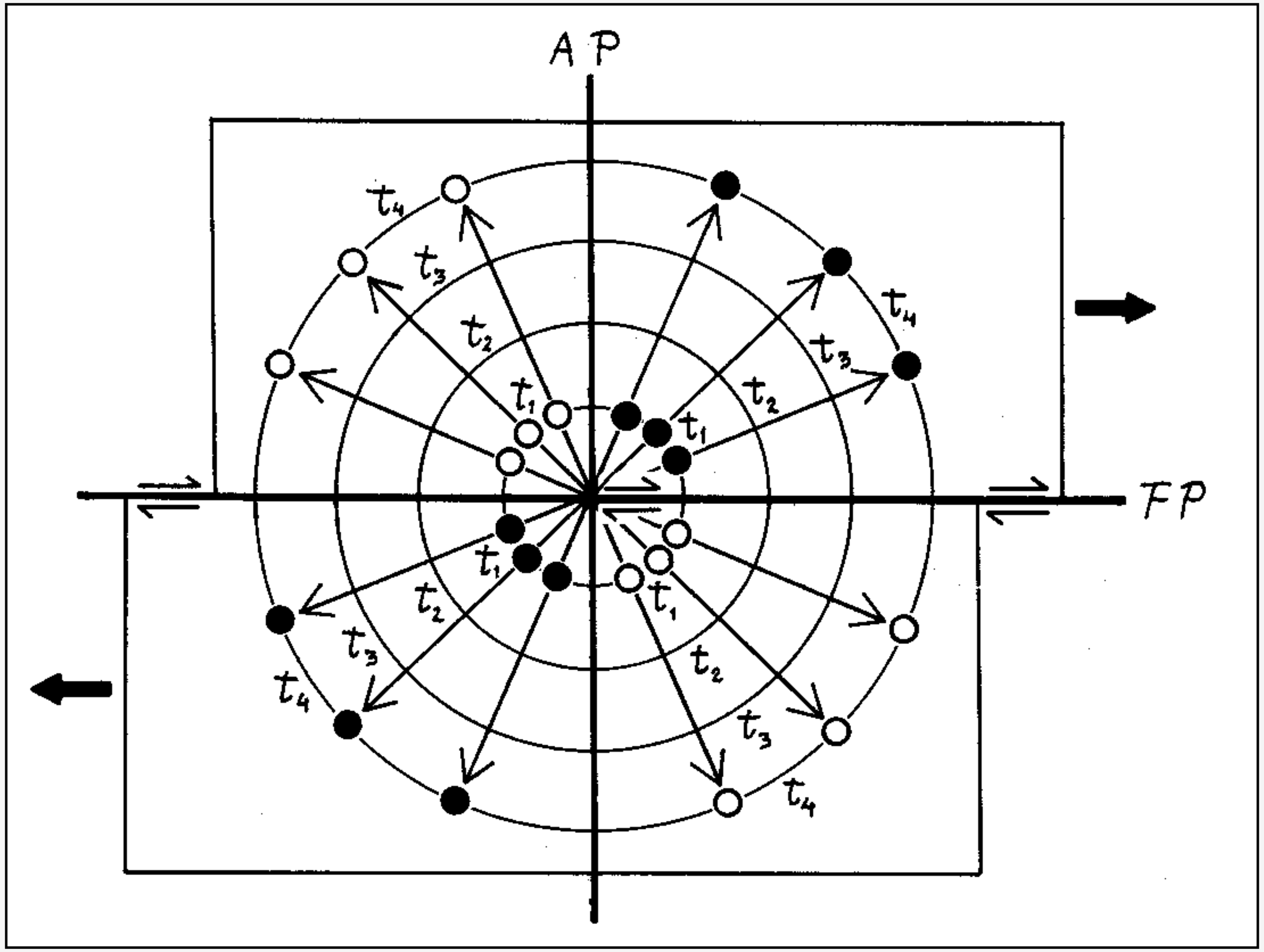
Determination of nodal planes



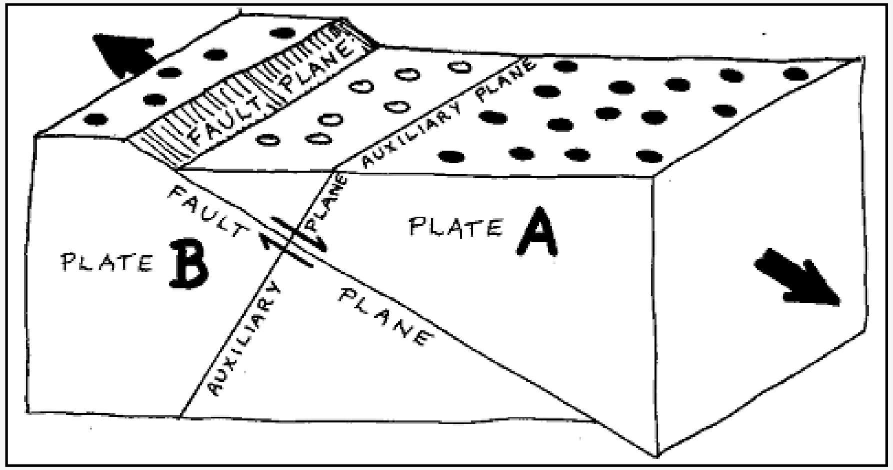


courtesy of Ian Hill, University of Leicester, UK

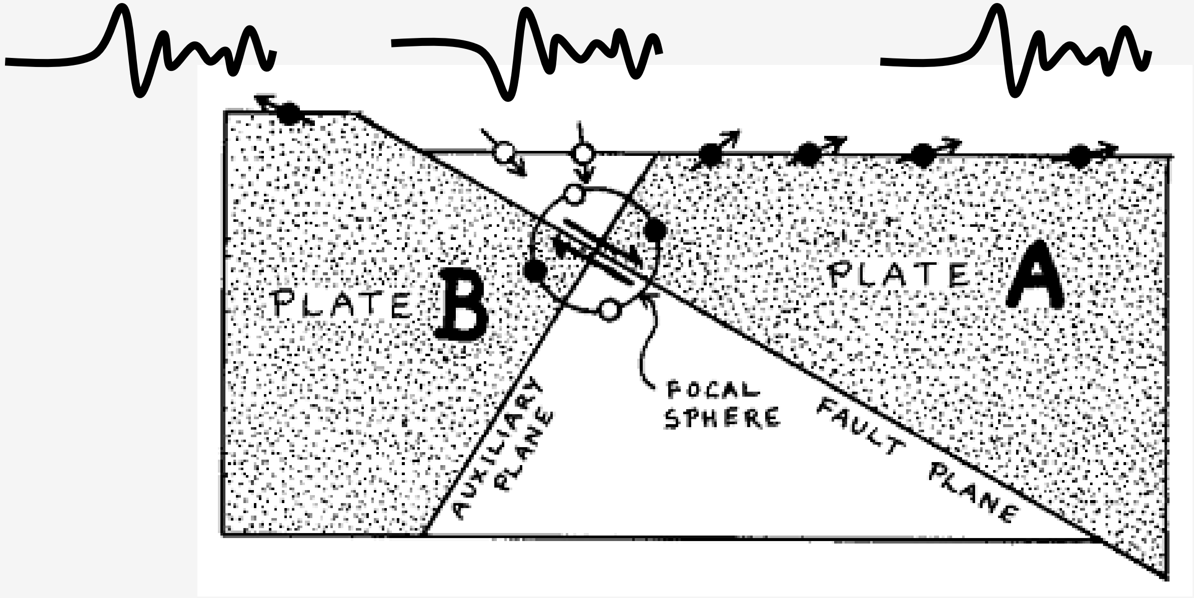
Spreading of the seismic wave



courtesy of Ian Hill, University of Leicester, UK



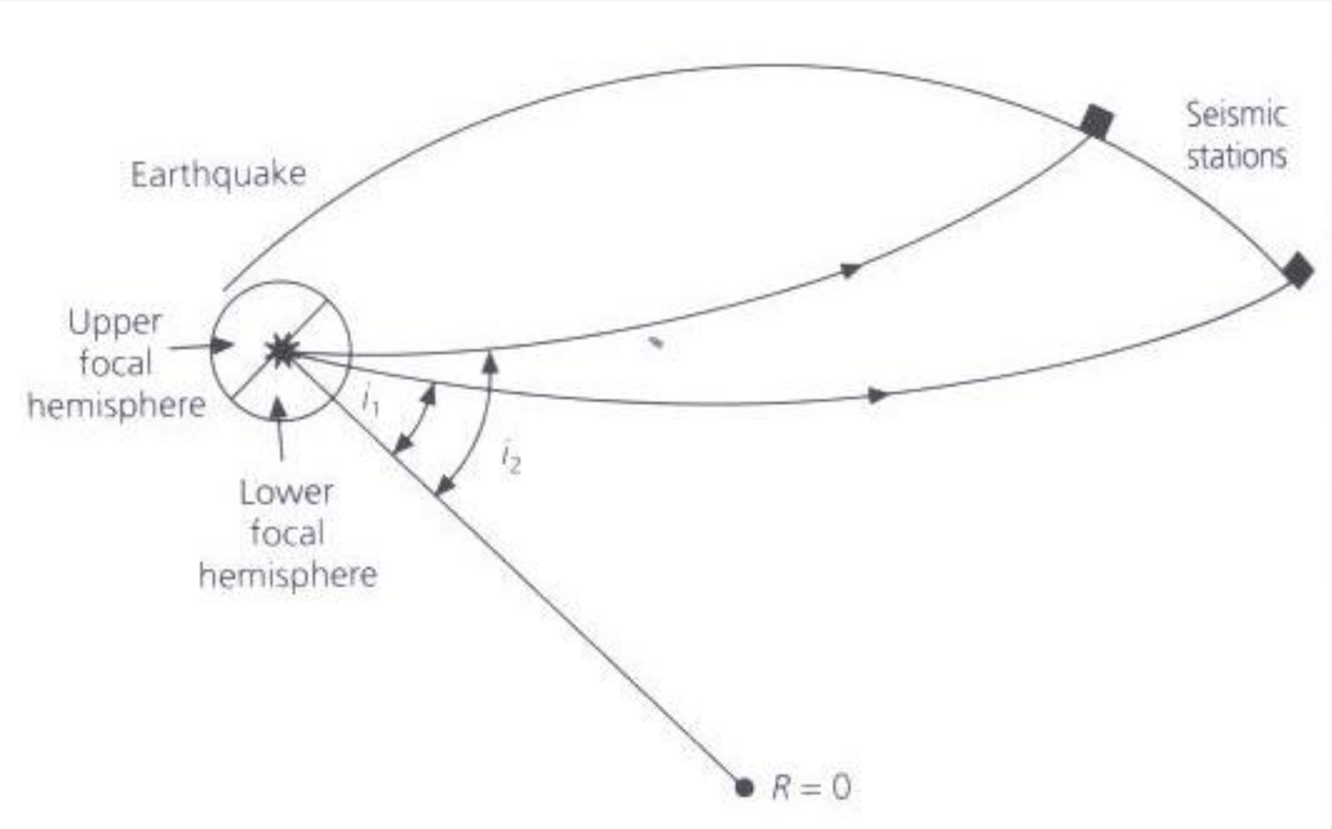
Data on the surface, interpreted in 3D



courtesy of Ian Hill,

Take-off angle

The angle (from vertical) that the ray leaves the earthquake = take-off angle



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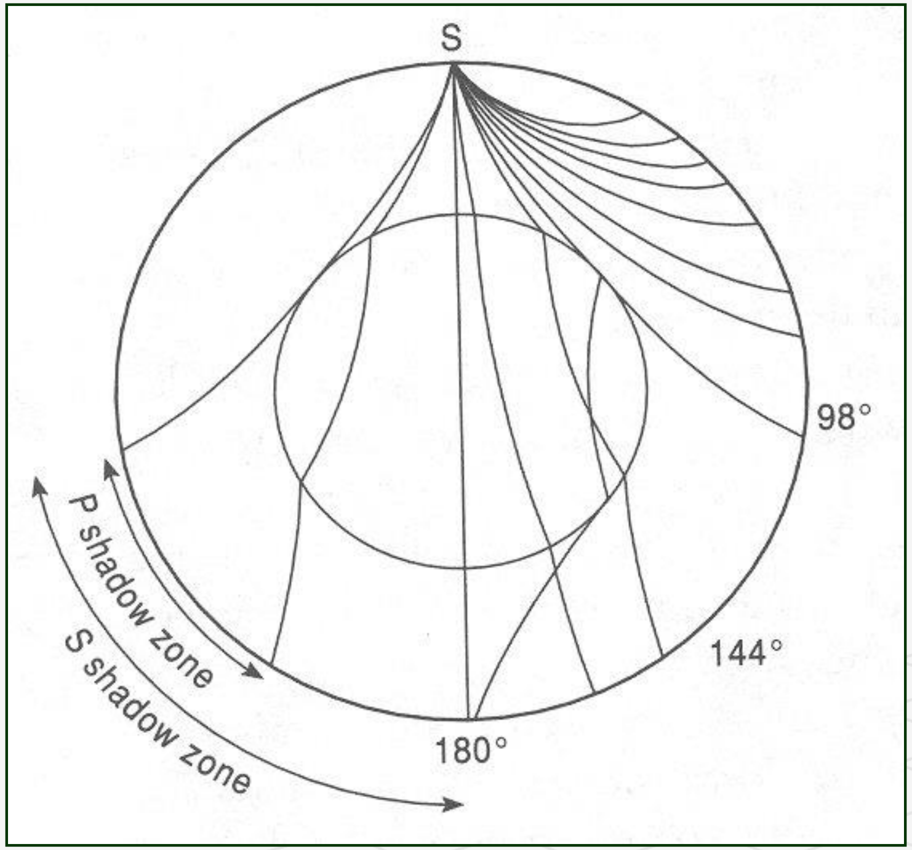
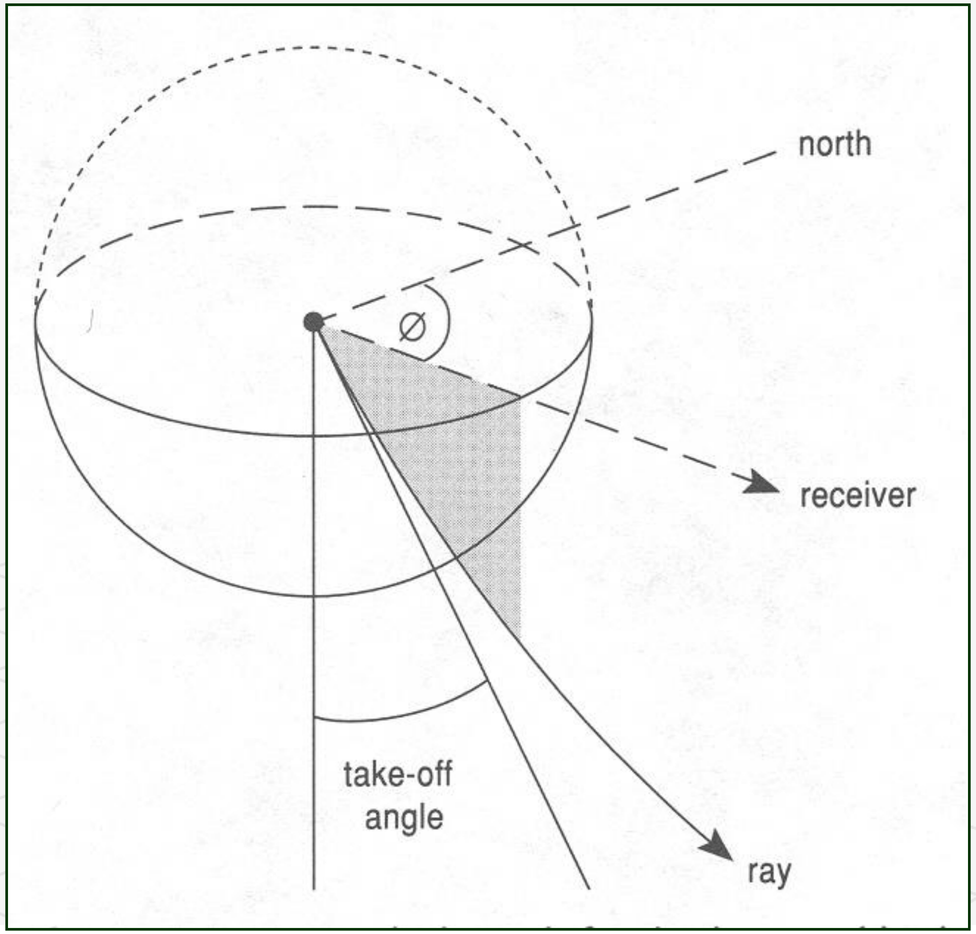
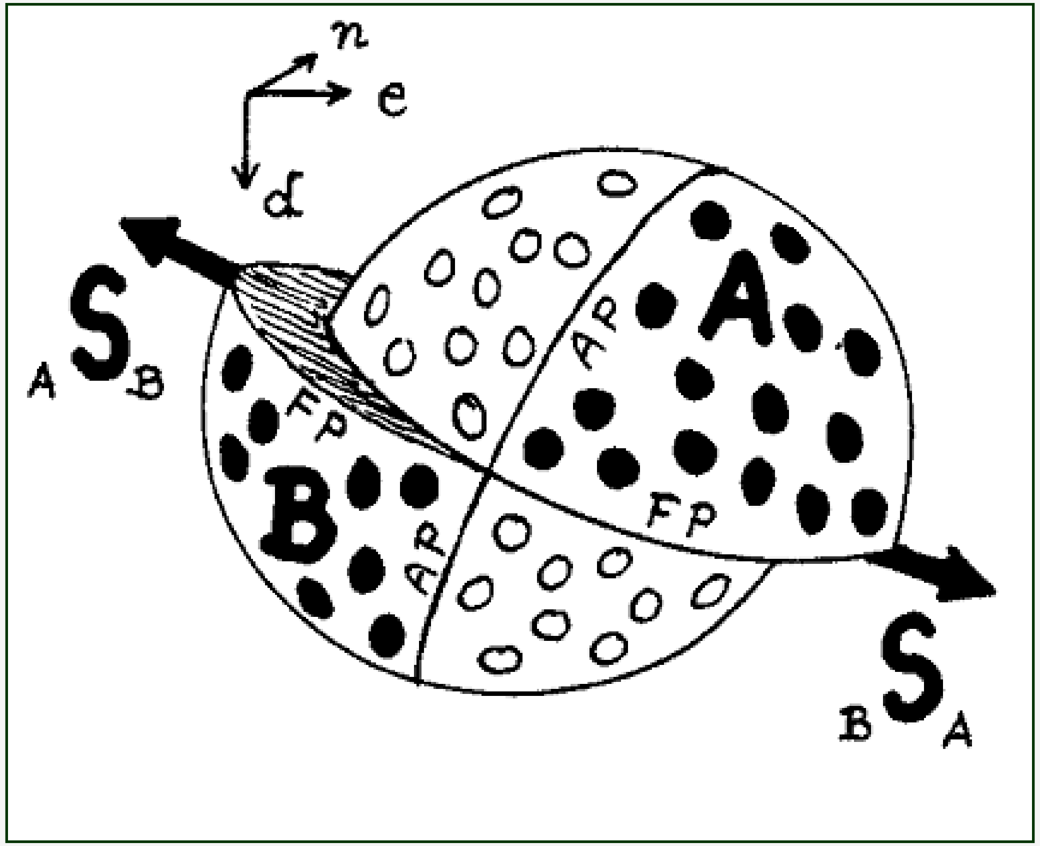
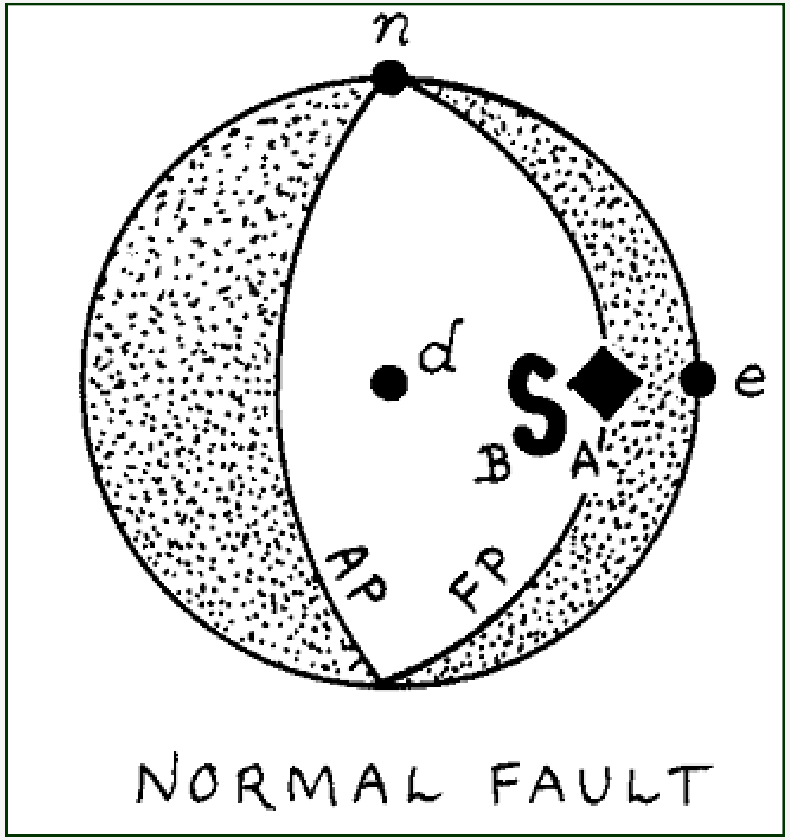
Azimuth (ϕ) and take-off angle





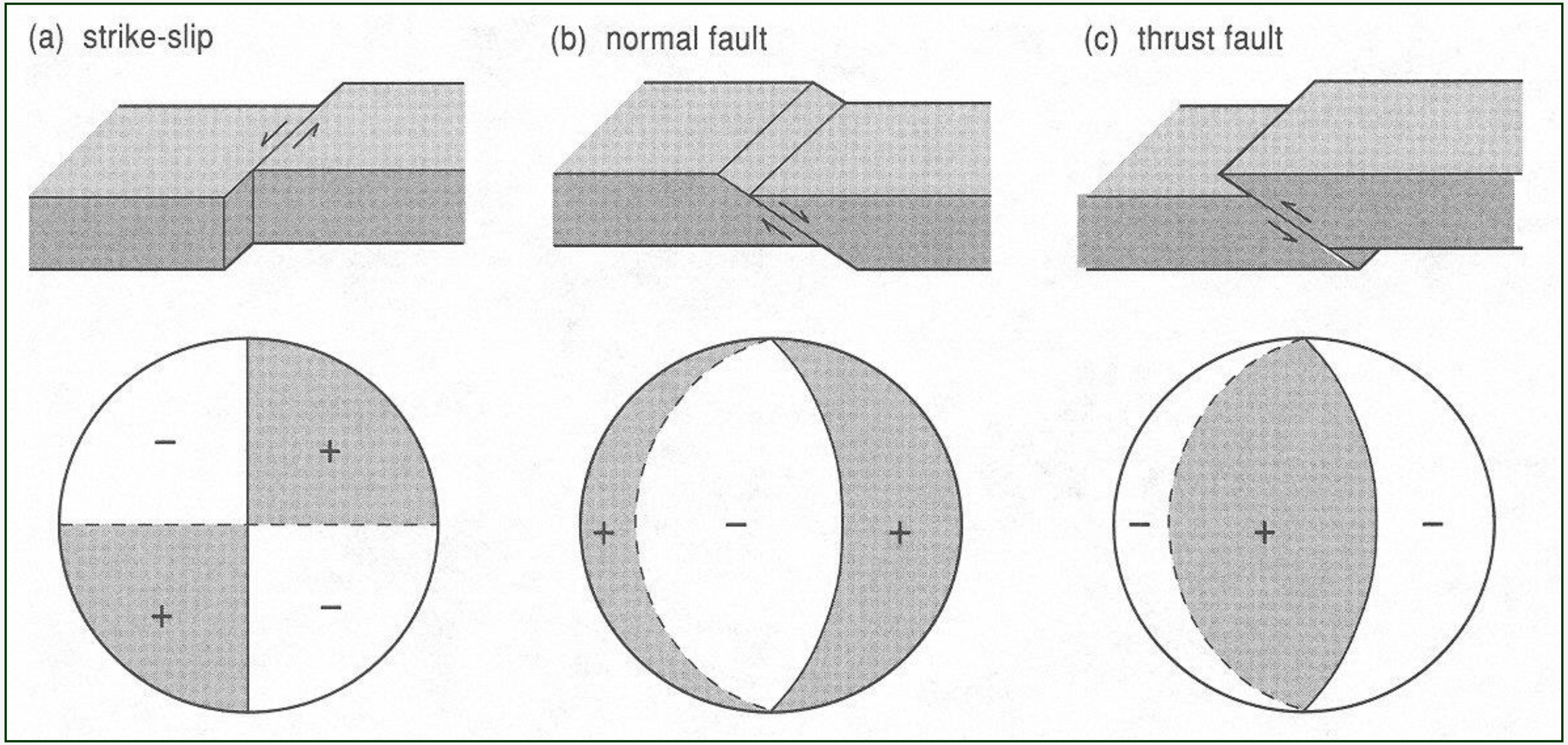
With a lot of recordings we can reconstruct faults with any orientations





courtesy of Ian Hill, University of Leicester, UK

Fault types and "Beach Ball" plots

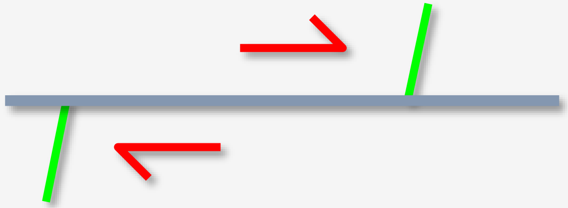


courtesy of Ian Hill, University of Leicester, UK

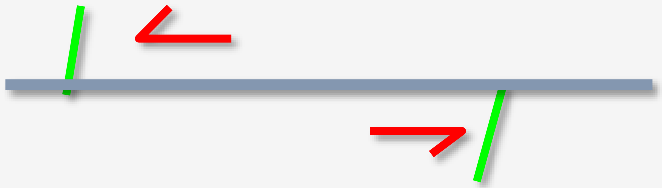
Fault types and the distribution of stress axes (P & T)

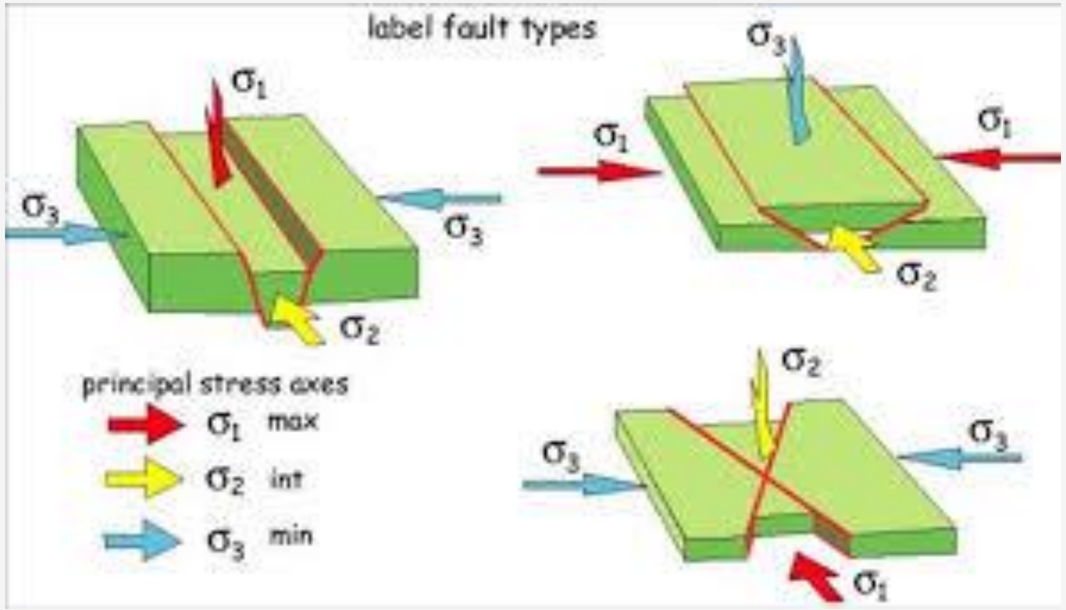
- **Normal Faults: from stretching of or extending rock; points on opposite sides of a fault are father apart after an earthquake**
- **Reverse Faults: from contracting or squishing rock; points on opposite sides of the fault are closer together after an earthquake**
- **Strike-Slip: can form in either areas of stretching or squishing, material slides laterally past each side of the fault.**

Described by sense of motion:
Right-lateral (Dextral)

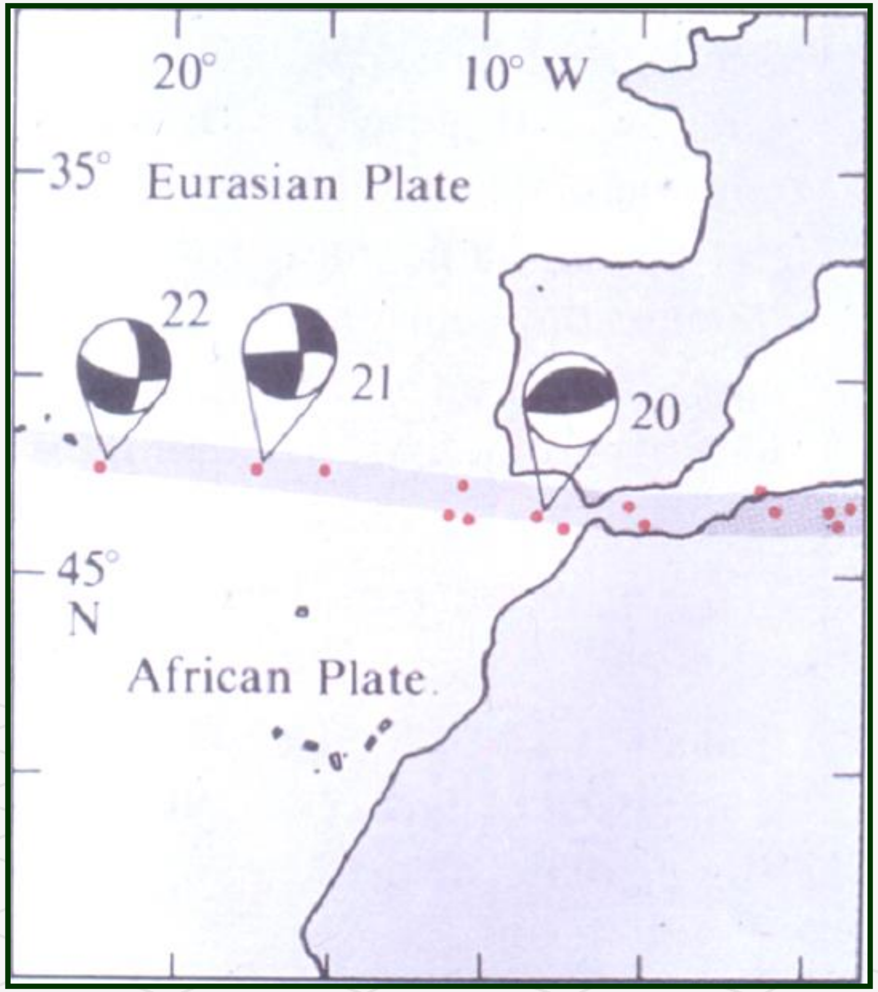


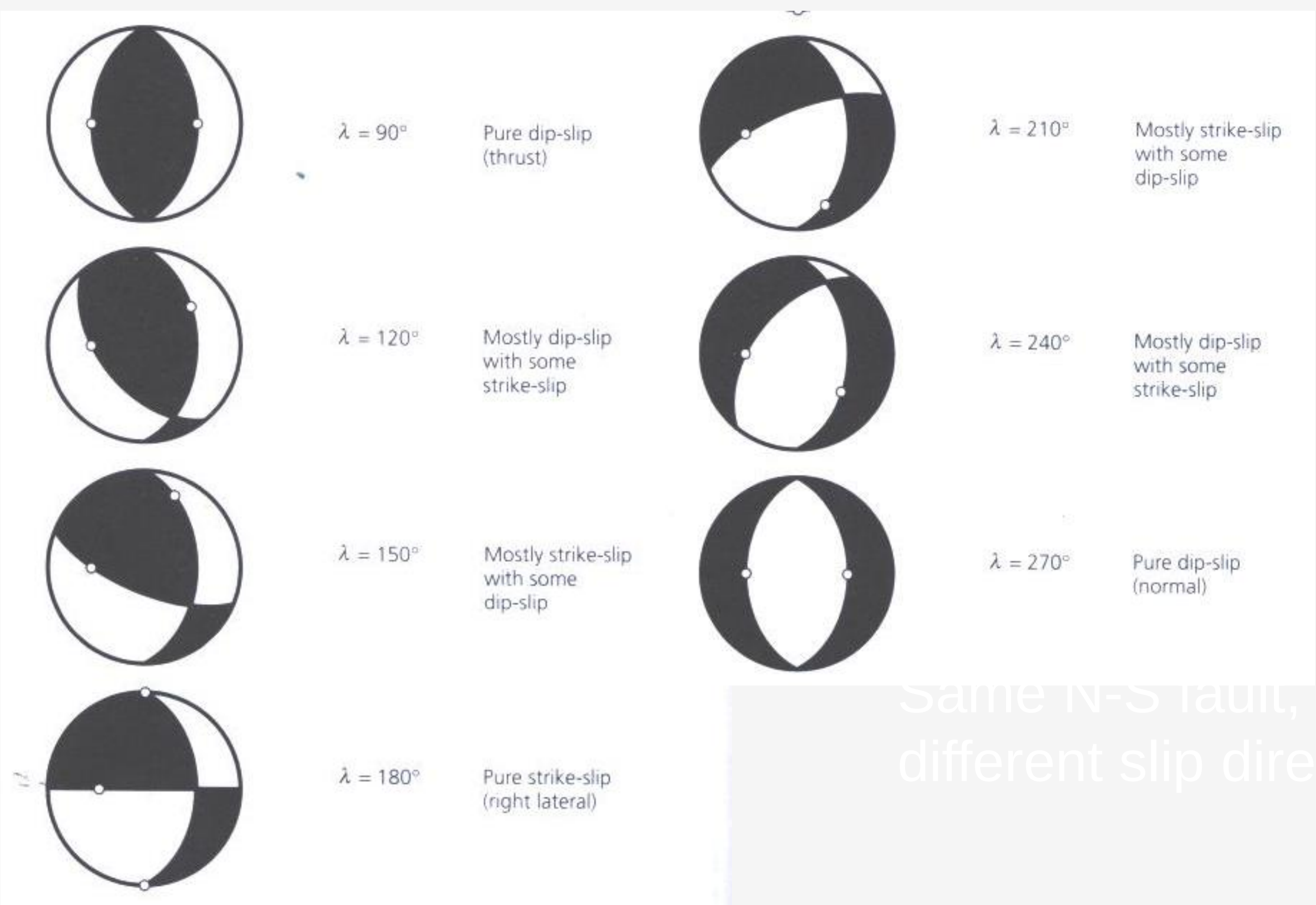
Left-lateral (Sinistral)





Example Focal mechanism diagrams on the Azores-Gibraltar fracture zone





Same N-S fault,
different slip direction

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