SEISMOLOGY I

Laurea Magistralis in Physics of the Earth and of the Environment

Introduction

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What is seismology?

Seismology (from Greek "Seismòs") is the science studying **earthquakes**:

"PHYSICAL" SEISMOLOGY

Study of the propagation of **seismic waves**

Study of the properties of the **Earth's interior**

Study of the physics of **seismic sources**

OBSERVATIONAL SEISMOLOGY

Recording earthquakes (microseismology)

Cataloguing earthquakes

Observing earthquake effects (macroseismology)

ENGINEERING SEISMOLOGY

Estimation of **seismic hazard** and risk

Definition of the Seismic input

EXPLORATIONAL SEISMOLOGY (Applied seismic methods)...

History – milestones

In Europe, seismological research was sparked by two devastating earthquakes in the 18th century:

1755 earthquake in Lisbon, Portugal
32000 killed

1783 earthquake sequence in Calabria, Italy
30000 killed

Experimental seismology	Theoretical seismology
1846 Mallet	1831 Poisson, waves in infinite media
1880 Milne (first real seismograph)	1849 Stokes, P and S waves as
1889 First teleseismic recording	dilatation and shear waves
(Potsdam)	1885 Rayleigh, waves in half space,
1884 Intensity scale (Rossi-Forel)	surface waves

History – milestones (cont'd)

1900	Oldham: identification of P, S, and surface waves
1901	Wiechert: first geophysical institute in Göttingen, Germany. Development of seismometers
1903	Foundation of International Seismological Association
1906	San Francisco earthquake: 1000 killed. Galitzin seismograph
1909	Mohorovicic discontinuity (MOHO)
1911	Theory of Love waves Seismological Society of America

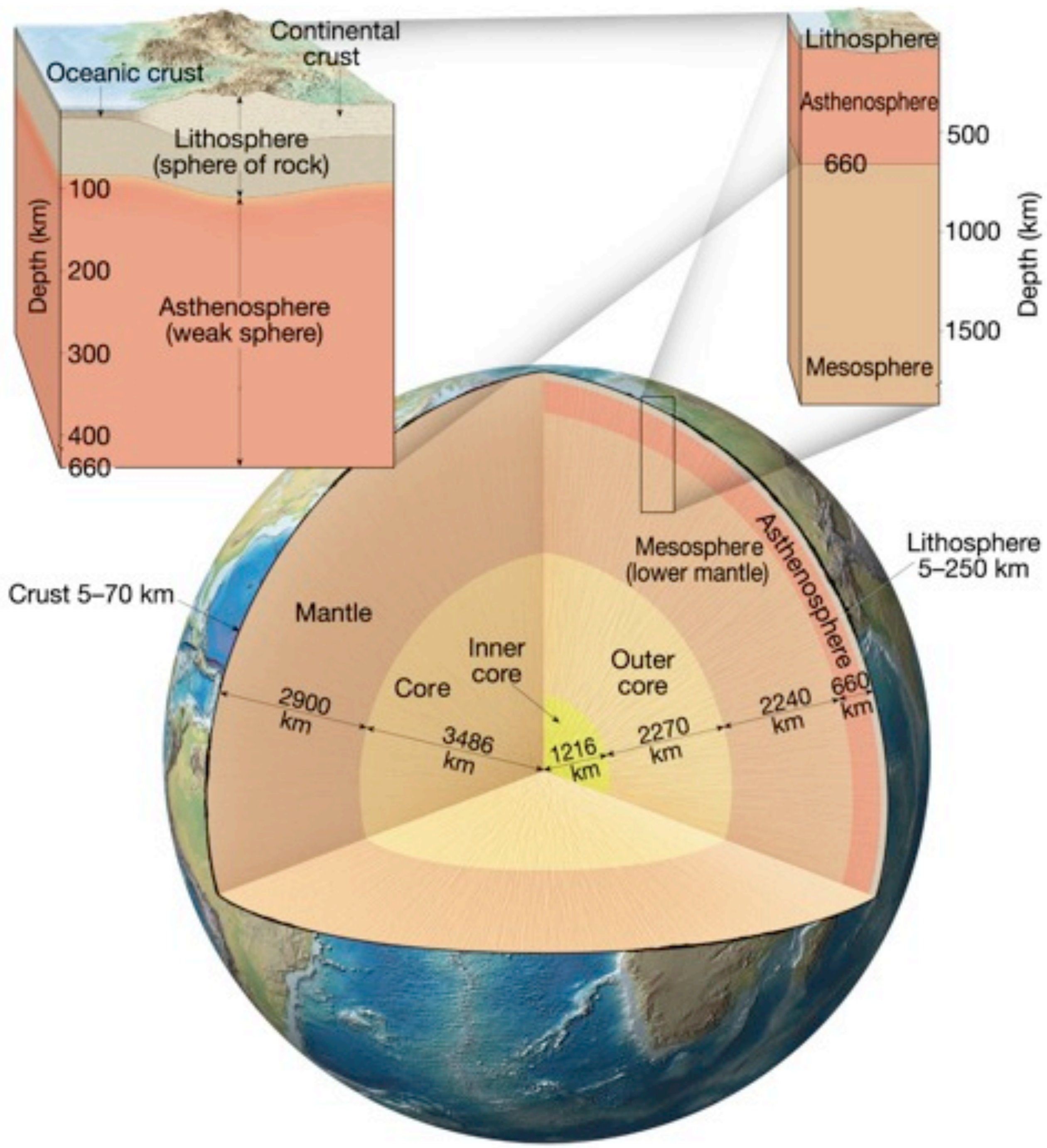
History – milestones (cont'd)

1913	Determination of radius of Earth's core by Beno Gutenberg (Göttingen)
1923	Tokyo earthquake („Great Japanese Quake") 250000 killed, Foundation of Earthquake Research Institute (ERI)
1903	Foundation of International Seismological Association
1931 1932	Benioff Seismometer Strain seismometer
1935 1936	Richter magnitude Discovery of the Earth's inner core by Inge Lehmann (**1888-1993**)
1940	Sir Harold Jeffreys, Cambridge Traveltime tables. Bullen – density model

History – milestones (after 1950)

1960	Observation of Earth's free oscillations after the 1960 Chile earthquake
1963	Limited Test Ban Treaty, World Wide Standard Seismograph Network (WWSSN)
Late 60s	The concept of plate tectonics is recognized
1981	Preliminary Reference Earth Model (PREM)
Mid 80s	First 3-D tomographic images of mantle heterogeneity
1997	Rotation of the Earth's inner core?

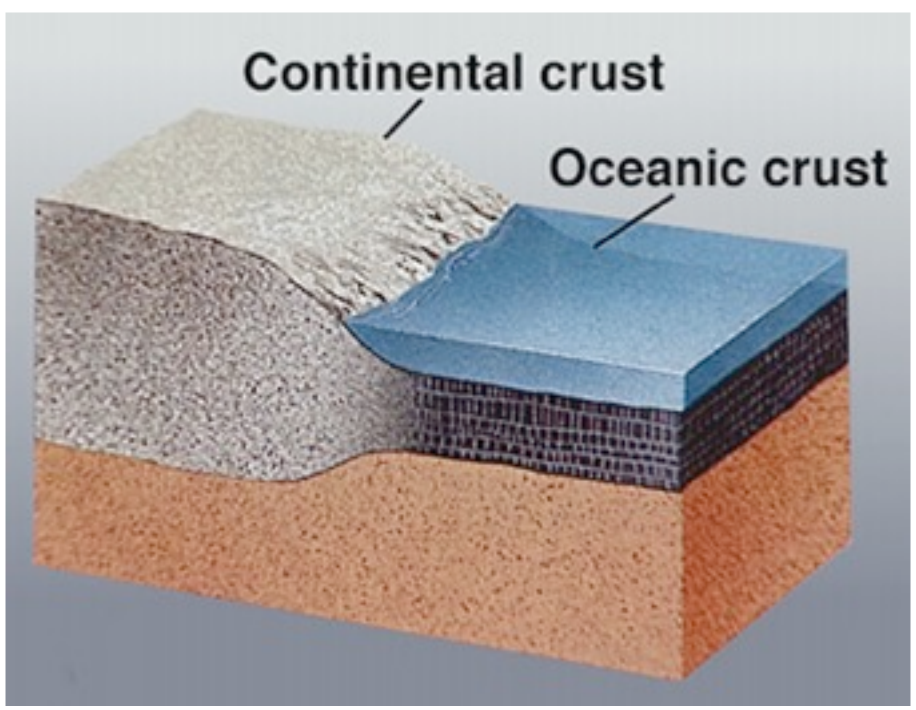
To see how earthquakes really occur, we first need to learn about constitution of the Earth!



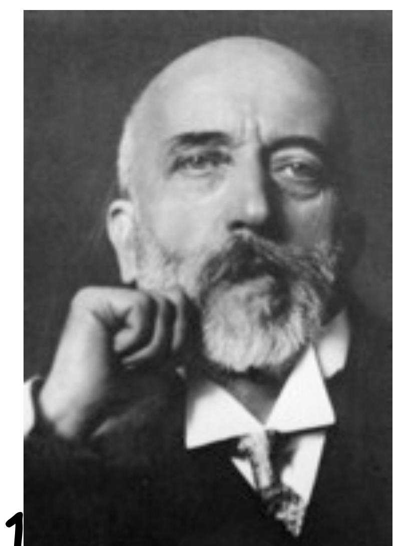
- Crust
- Mantle
- Core

The Shallowest Layer: the Crust

- The crust is the most heterogeneous layer in the Earth
- The crust is on average 33 km thick for continents and 10 km thick beneath oceans; however it varies from just a few km to over 70 km globally.

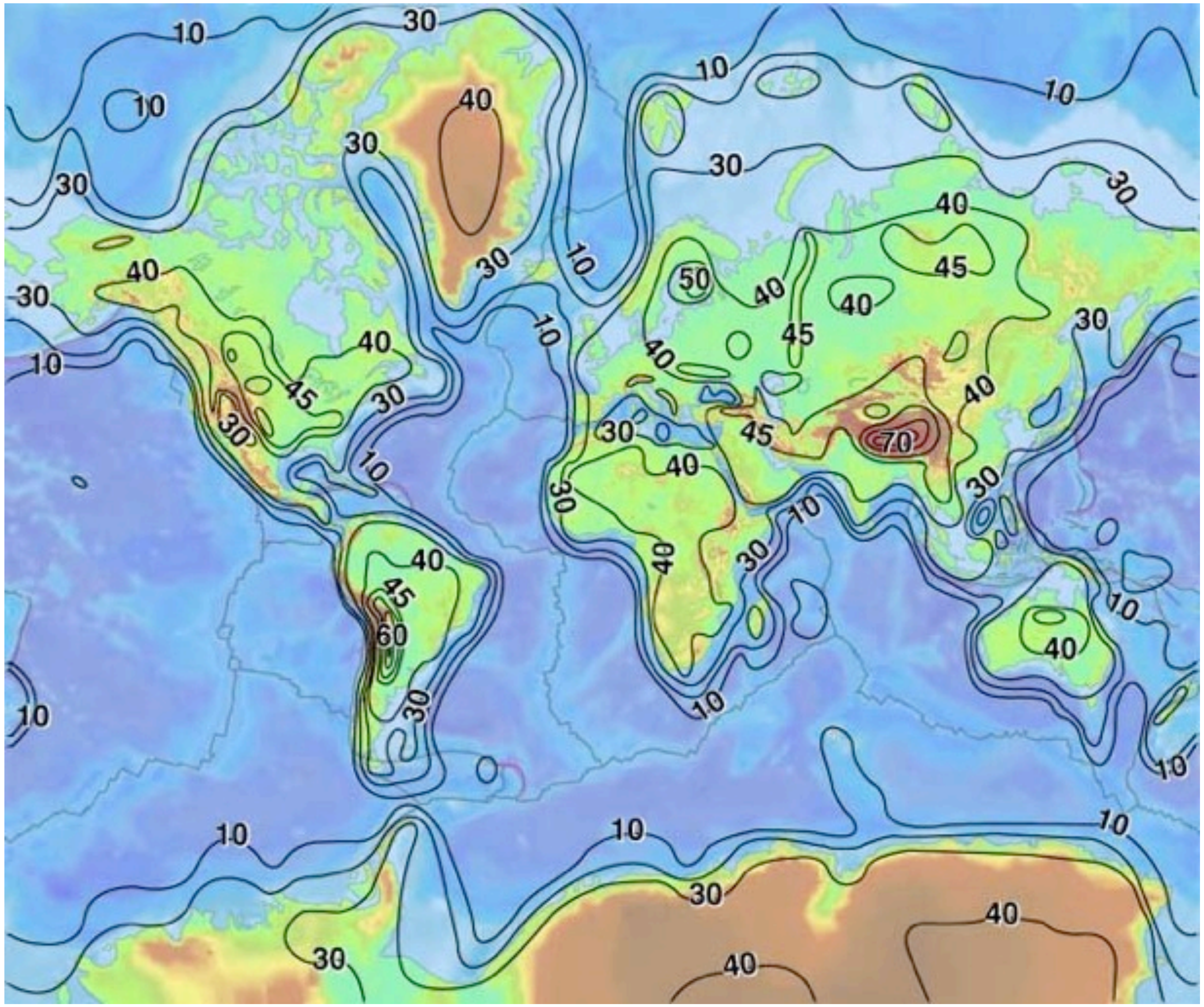


- The boundary between the crust and the mantle is mostly chemical. The crust and mantle have different compositions.
- This boundary is referred to as the Mohorovičić discontinuity or "Moho".



- It was discovered in 1910 by the Croatian seismologist Andrija Mohorovičić.

Crustal thickness

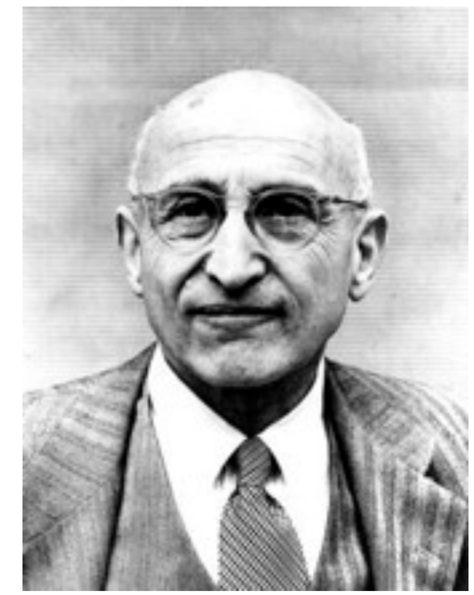


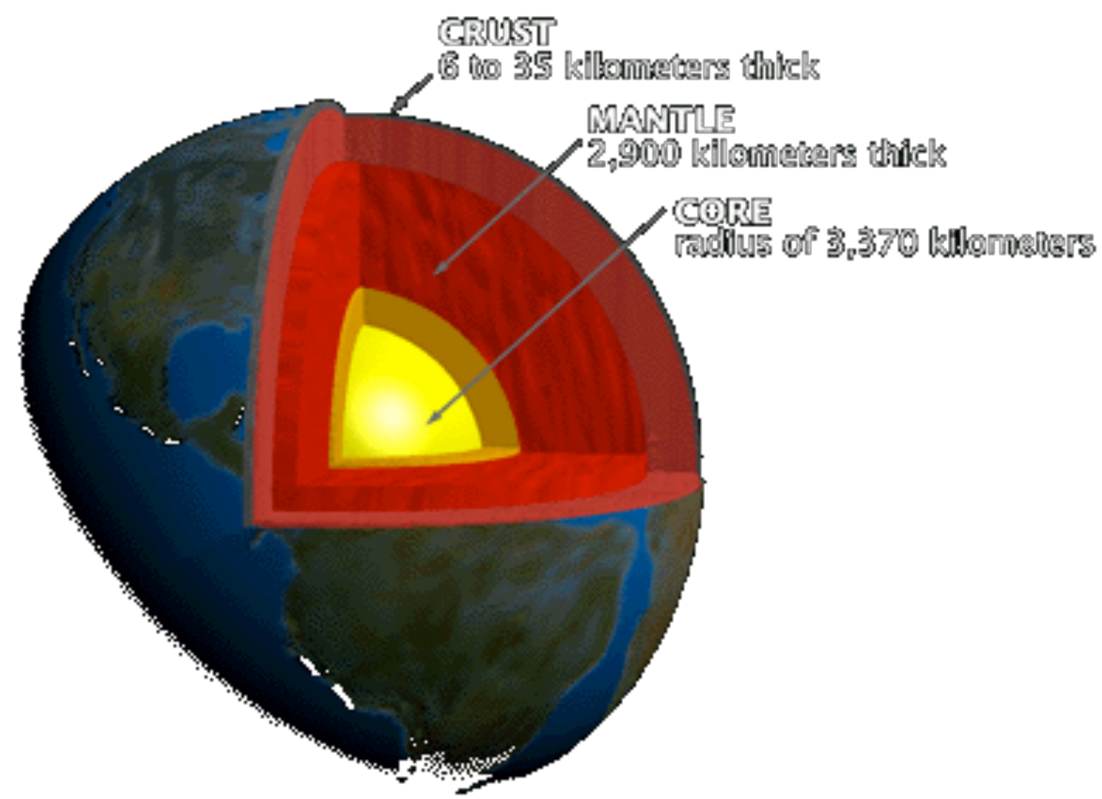
http://quake.wr.usgs.gov/research/structure/CrustalStructure/index.html

Middle Earth: The Mantle

- Earth's mantle exists from the bottom of the crust to a depth of 2891 km (radius of 3480 km) – **Gutenberg discontinuity**
- It is further subdivided into:
  - The uppermost mantle (crust to 400 km depth)
  - The transition zone (400 – 700 km depth)
  - The mid-mantle (700 to ~2650 km depth)
  - The lowermost mantle (~2650 – 2891 km depth)
- The uppermost mantle is composed dominantly of olivine; lesser components include pyroxene, enstatite, and garnet

Beno Gutenberg



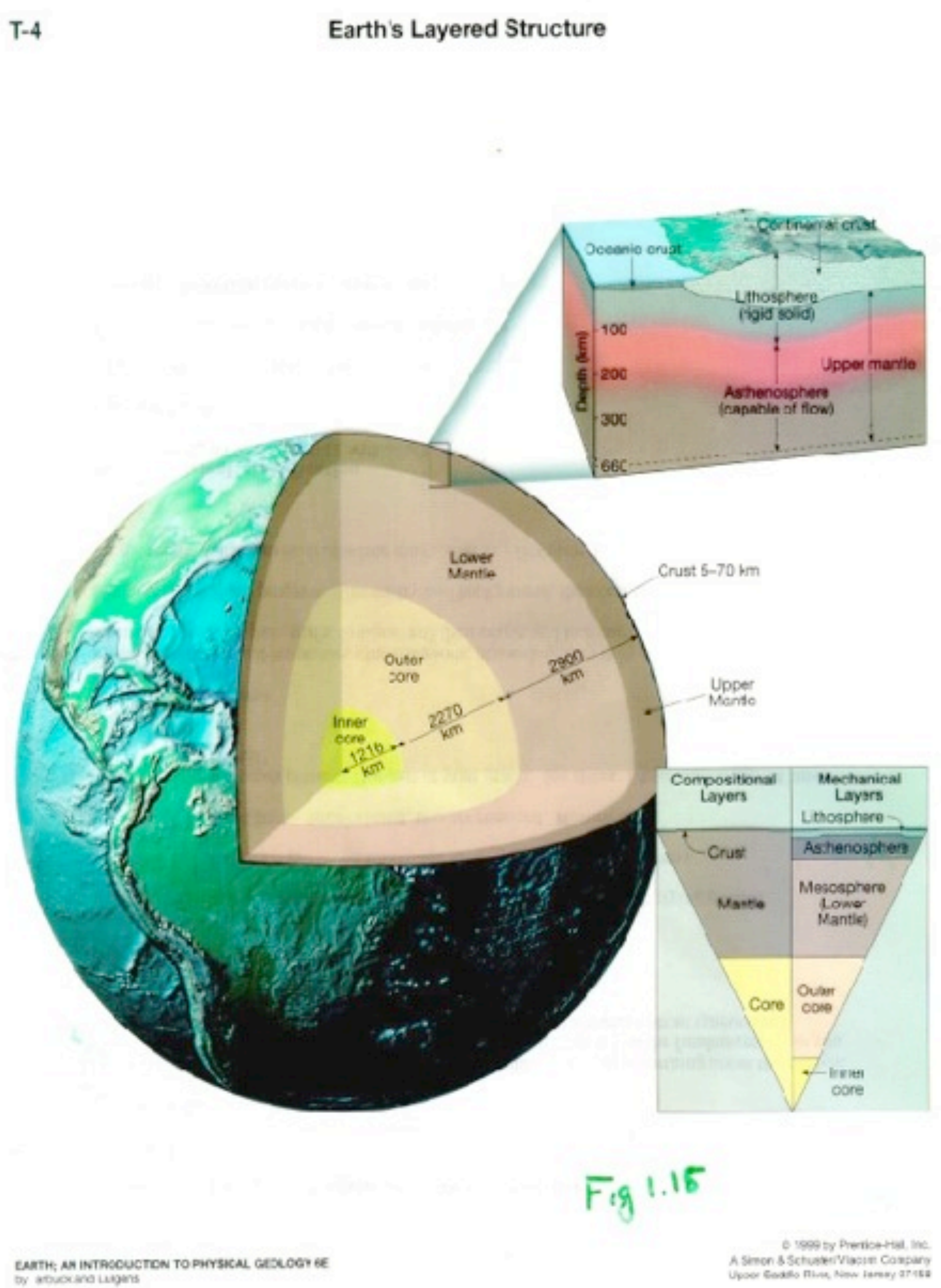


Earth's Core

- The boundary between the liquid outer core and the solid inner core occurs at a radius of about 1220 km – **Lehman discontinuity**, after Inge Lehman from Denmark.
- The boundary between the mantle and outer core is sharp. The outer core is the most homogeneous part of the Earth
- The change in density across the core-mantle boundary is greater than that at the Earth's surface!
- The outer core is mostly an alloy of iron and nickel in liquid form.

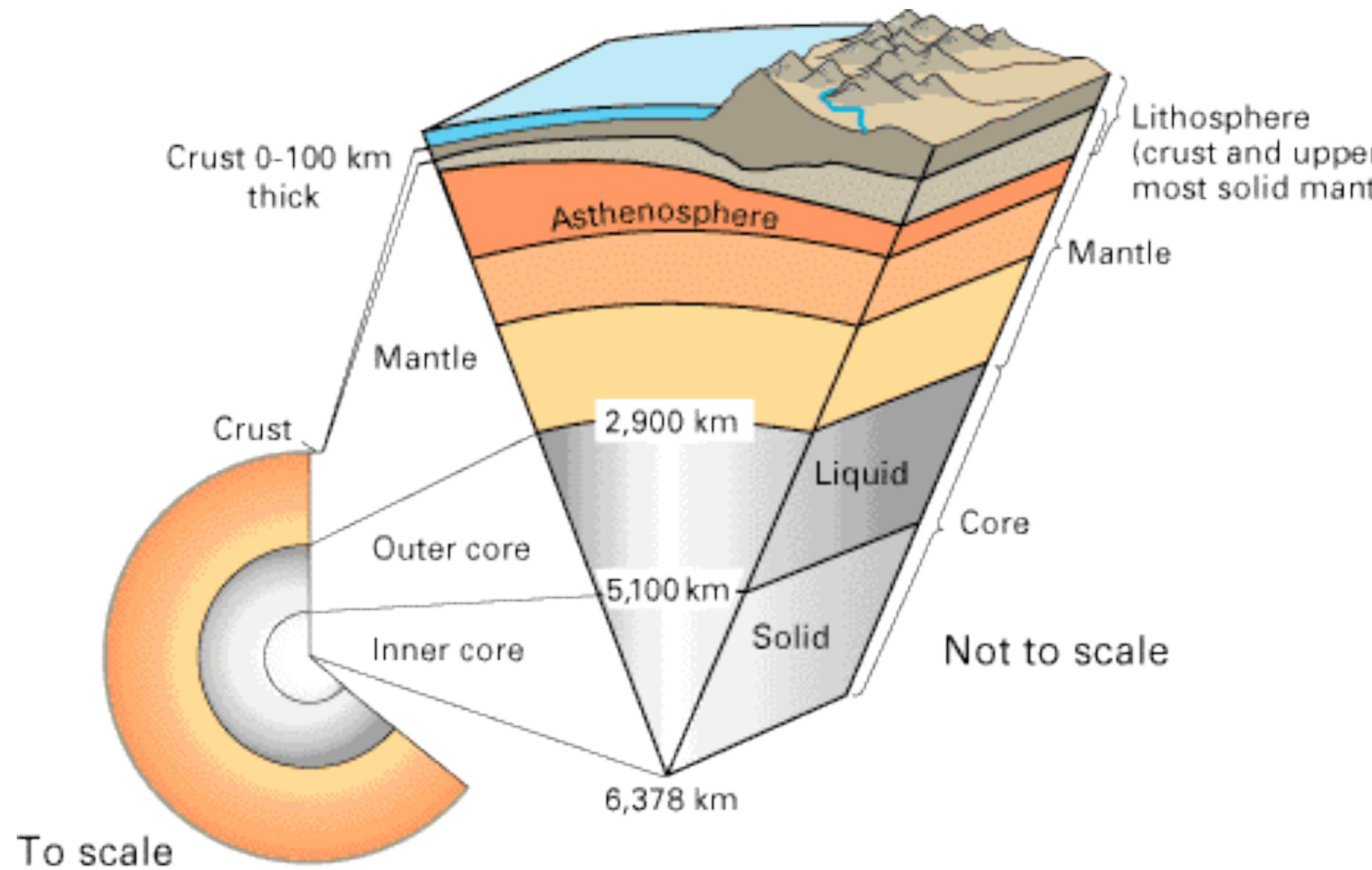
- The viscosity of the outer core is similar to that of water, it flows kilometers per year and creates the Earth's magnetic field.
- Owing to the great pressure inside the Earth the Earth's core is actually freezing as the Earth gradually cools.
- As the core freezes latent heat is released; this heat causes the outer core to convect and so generates a magnetic field.

Mechanical Layers:



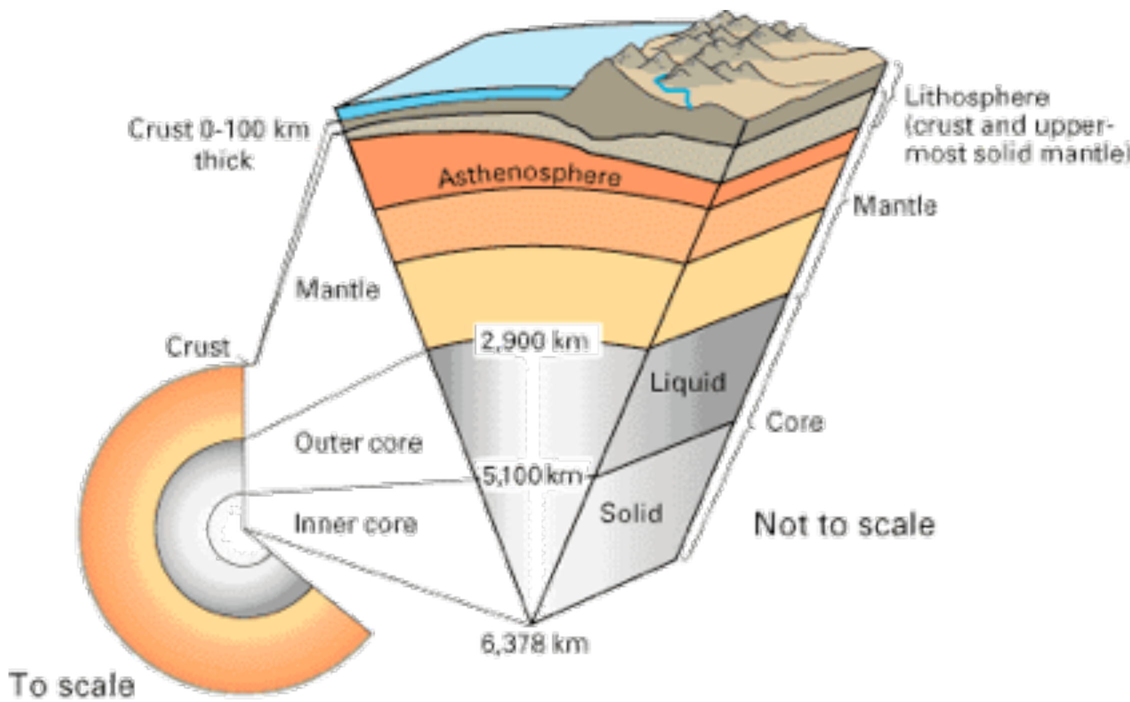
- Lithosphere
- Asthenosphere
- Mesosphere

Lithosphere



- The lithosphere is the uppermost 50-100 km of the Earth.
- There is not a strict boundary between the lithosphere and the asthenosphere as there is between the crust and mantle.
- It consists of both crust and upper parts of mantle.
- It behaves rigidly, like a solid, over very long time periods.

Asthenosphere



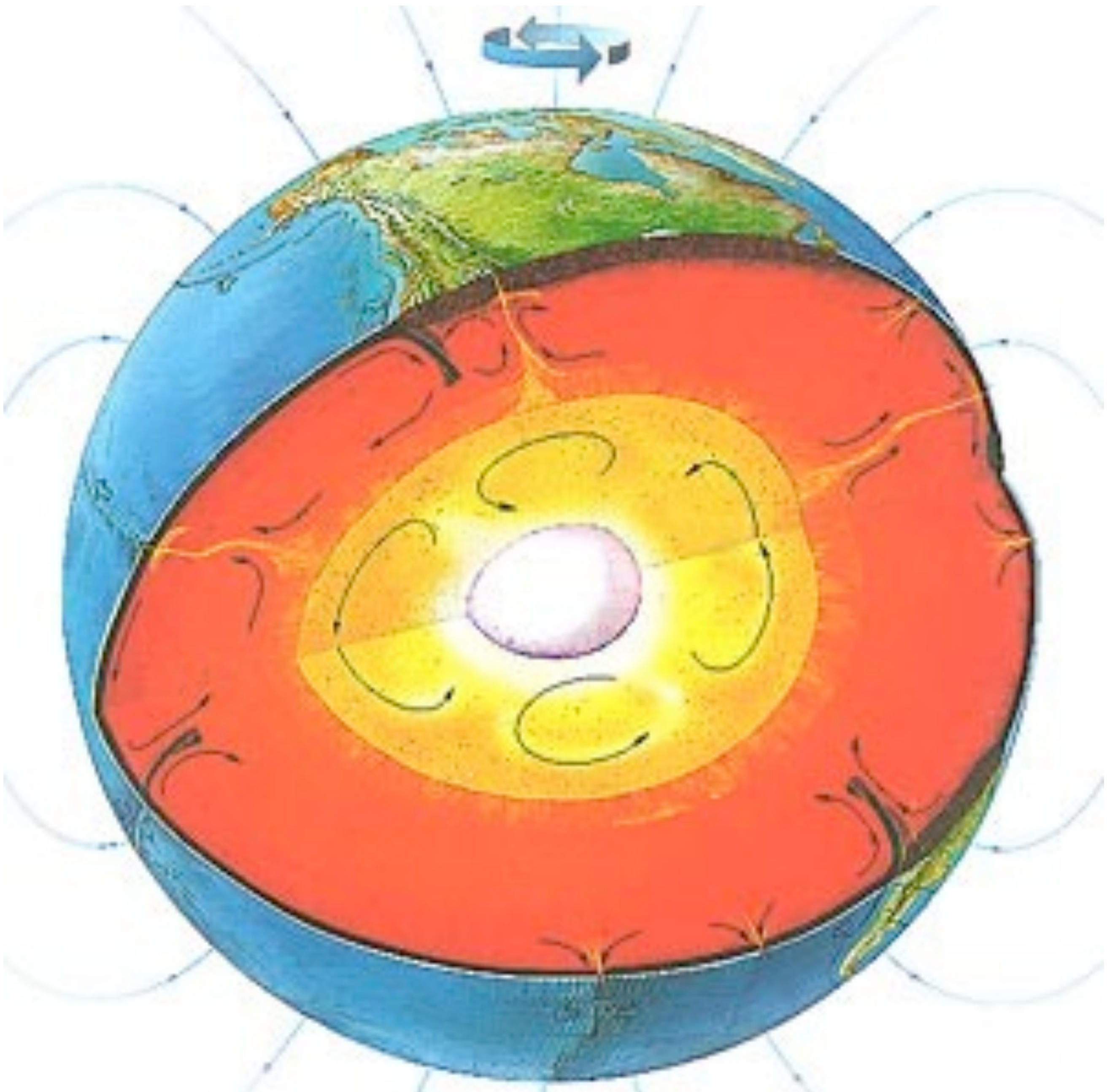
- The asthenosphere exists between depths of 100-200 km.
- It is the weakest part of the mantle.
- It is a solid over short time scales, but behaves like a fluid over millions of years.
- The asthenosphere decouples the lithosphere (tectonic plates) from the rest of the mantle.

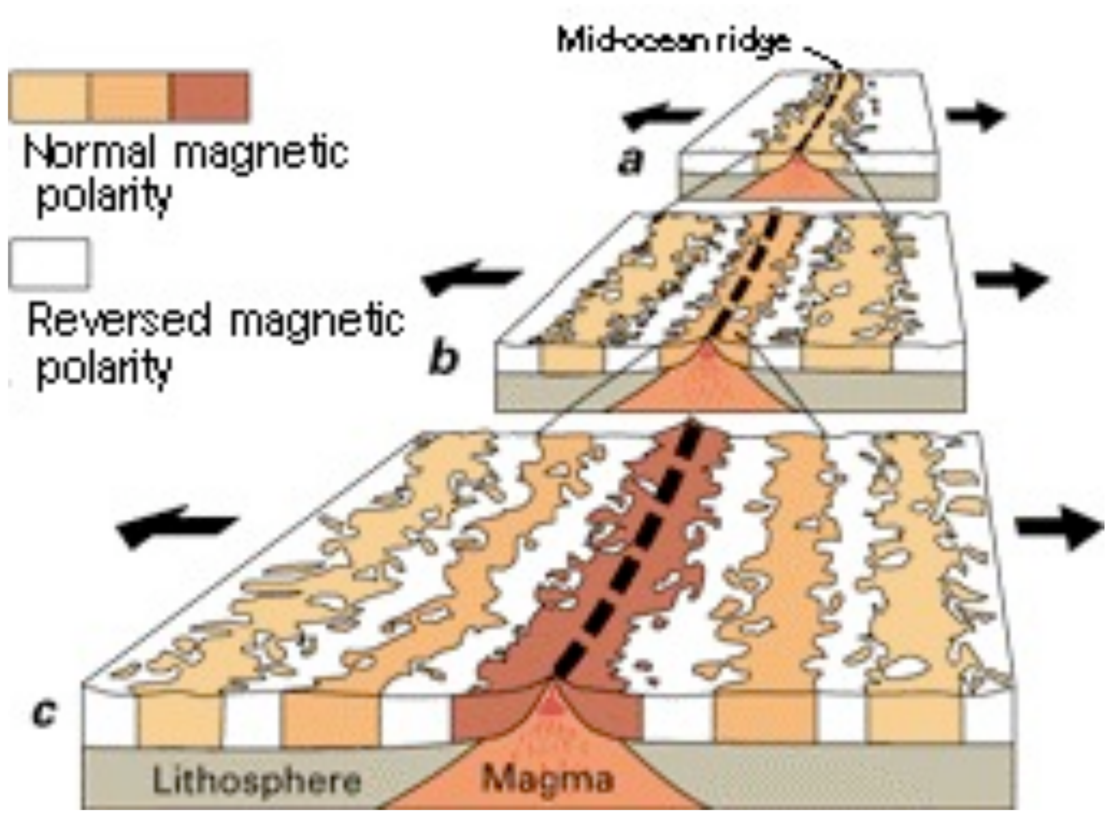
Tectonic Forces & Convection

- The interior of the Earth is dynamic – it cools down and thus provides energy for convective currents in the outer core and in the astenosphere.

- Additional energy comes from radioactive decay...

Convection in the astenosphere enables tectonic processes – **PLATE TECTONICS**





The proof of plate tectonics came from the magnetization of the seafloor as a function of distance from the ridge axes.

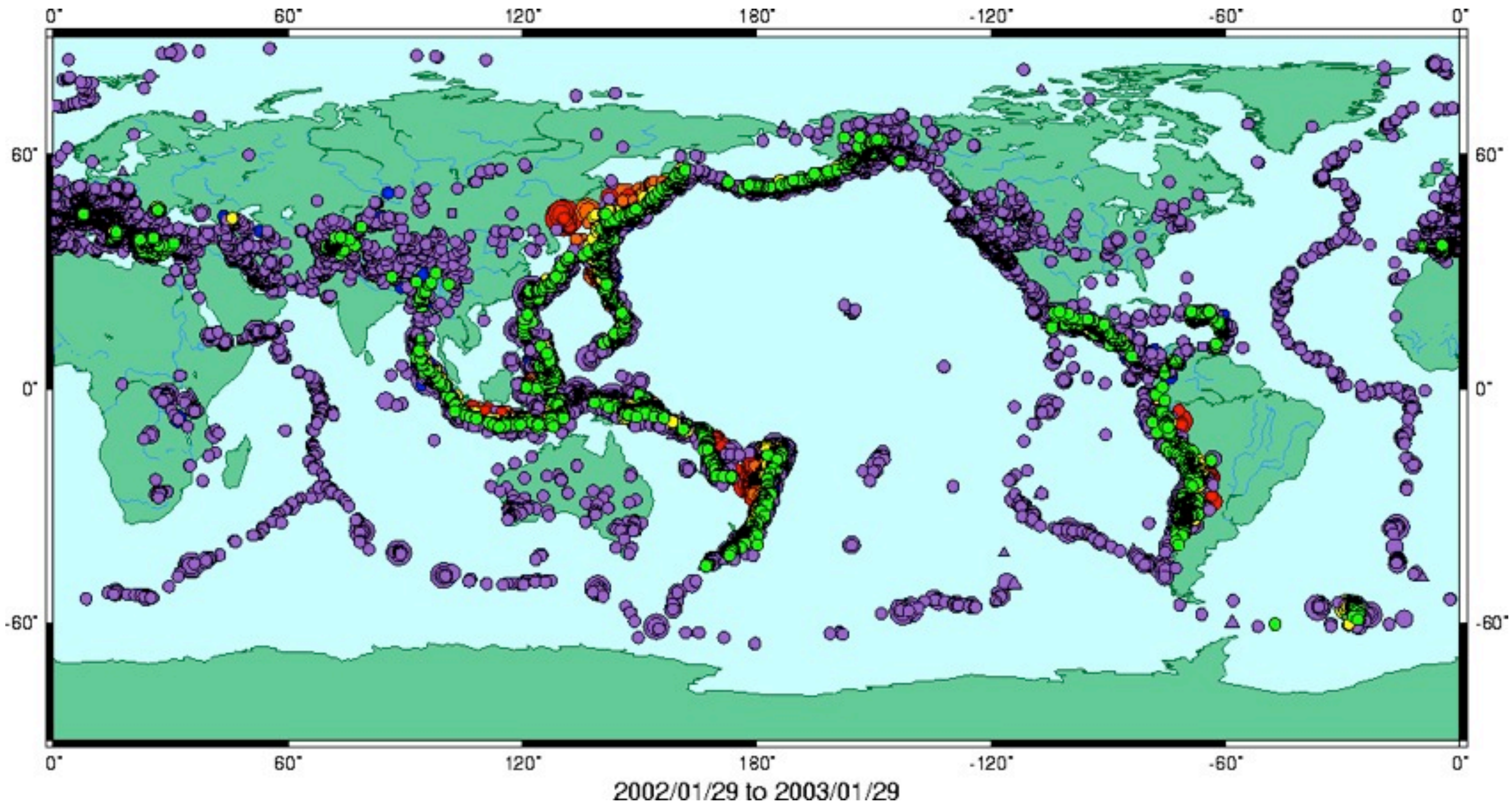
Plate tectonics

PLATE TECTONICS theory is very young (1960-ies)

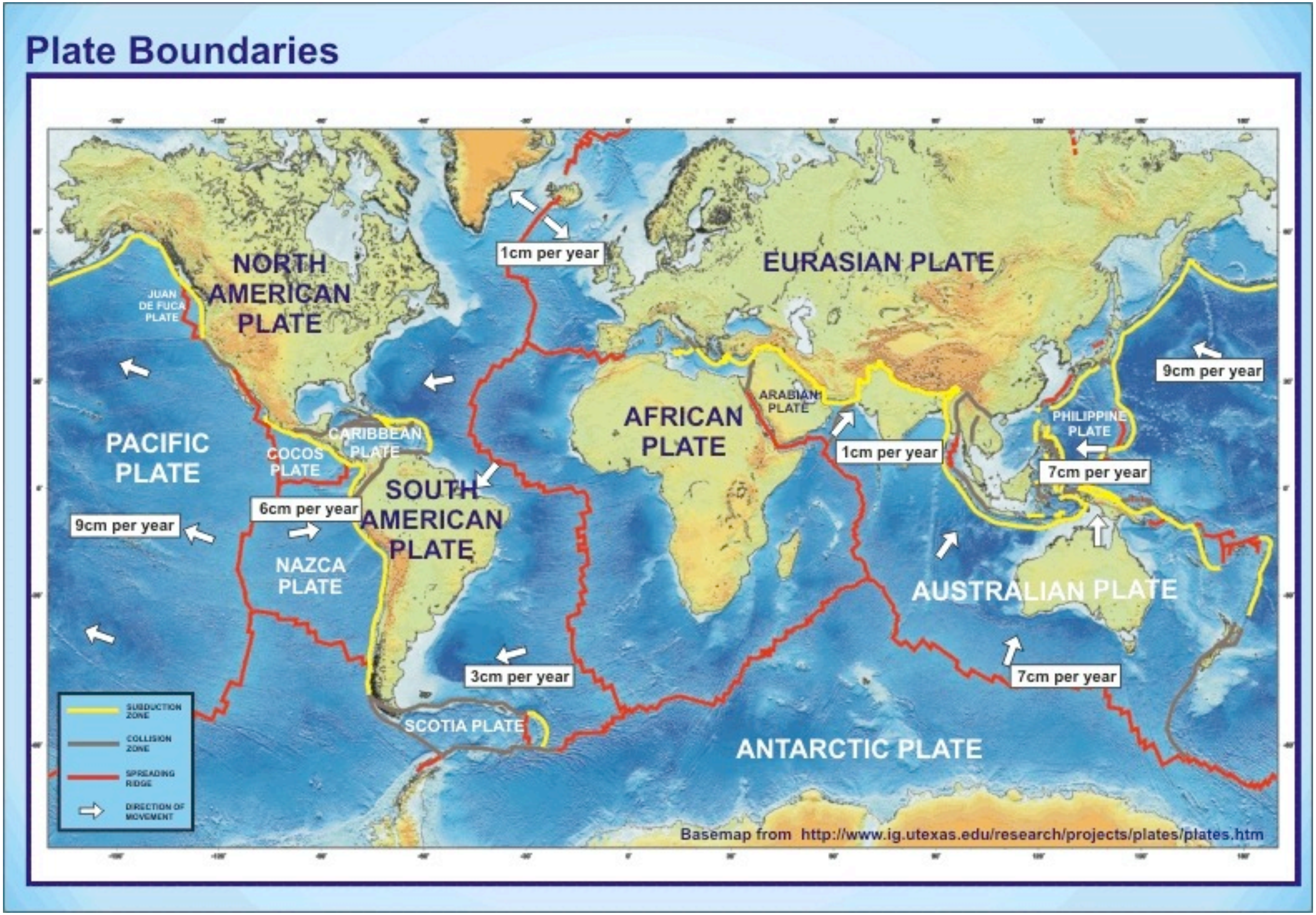
It provides answers to the most fundamental questions in seismology:

- Why earthquakes occur?
- Why are earthquake epicenters not uniformly distributed around the globe?
- At what depths are their foci?

One year of seismicity

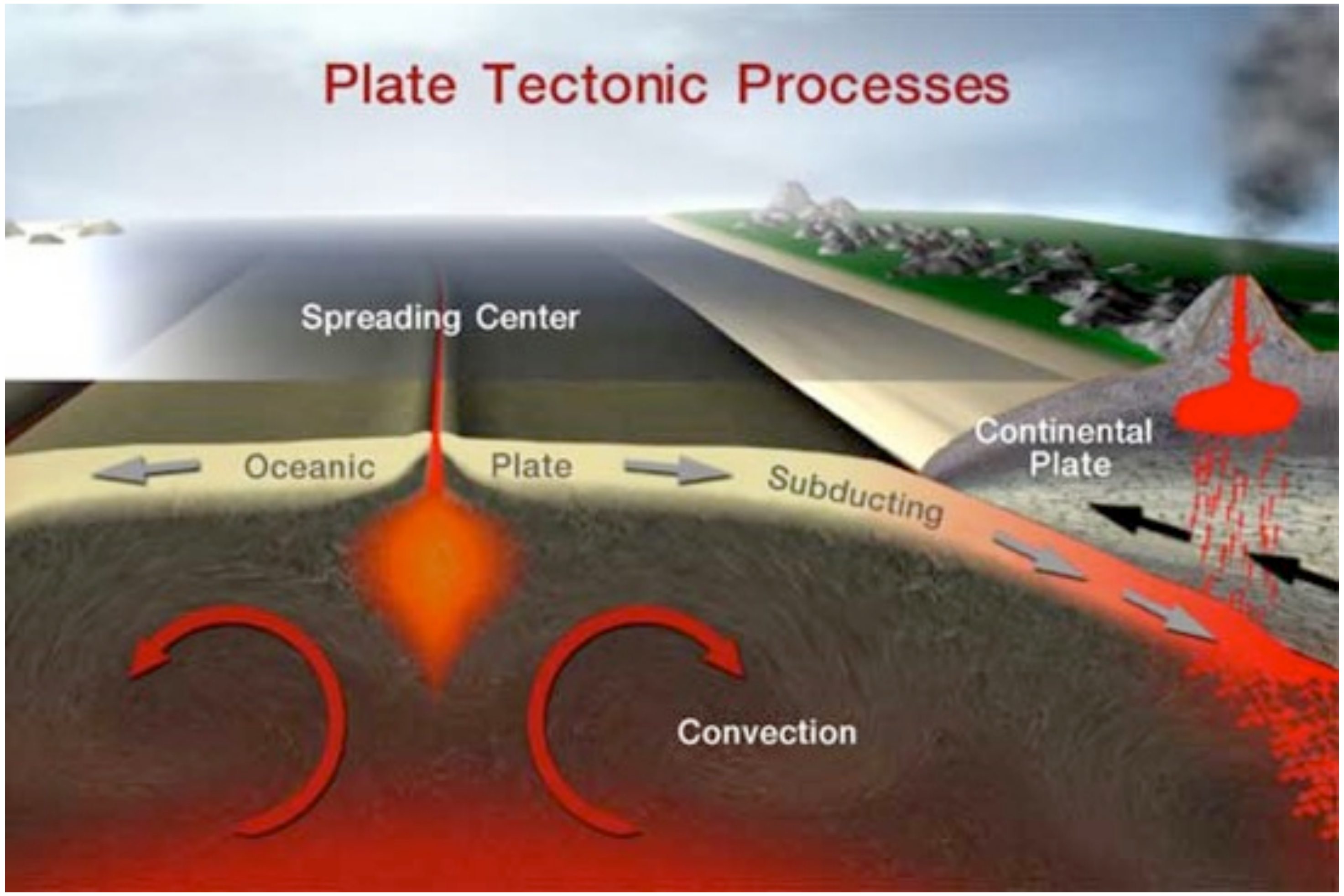


Major tectonic plates



Tectonic plates

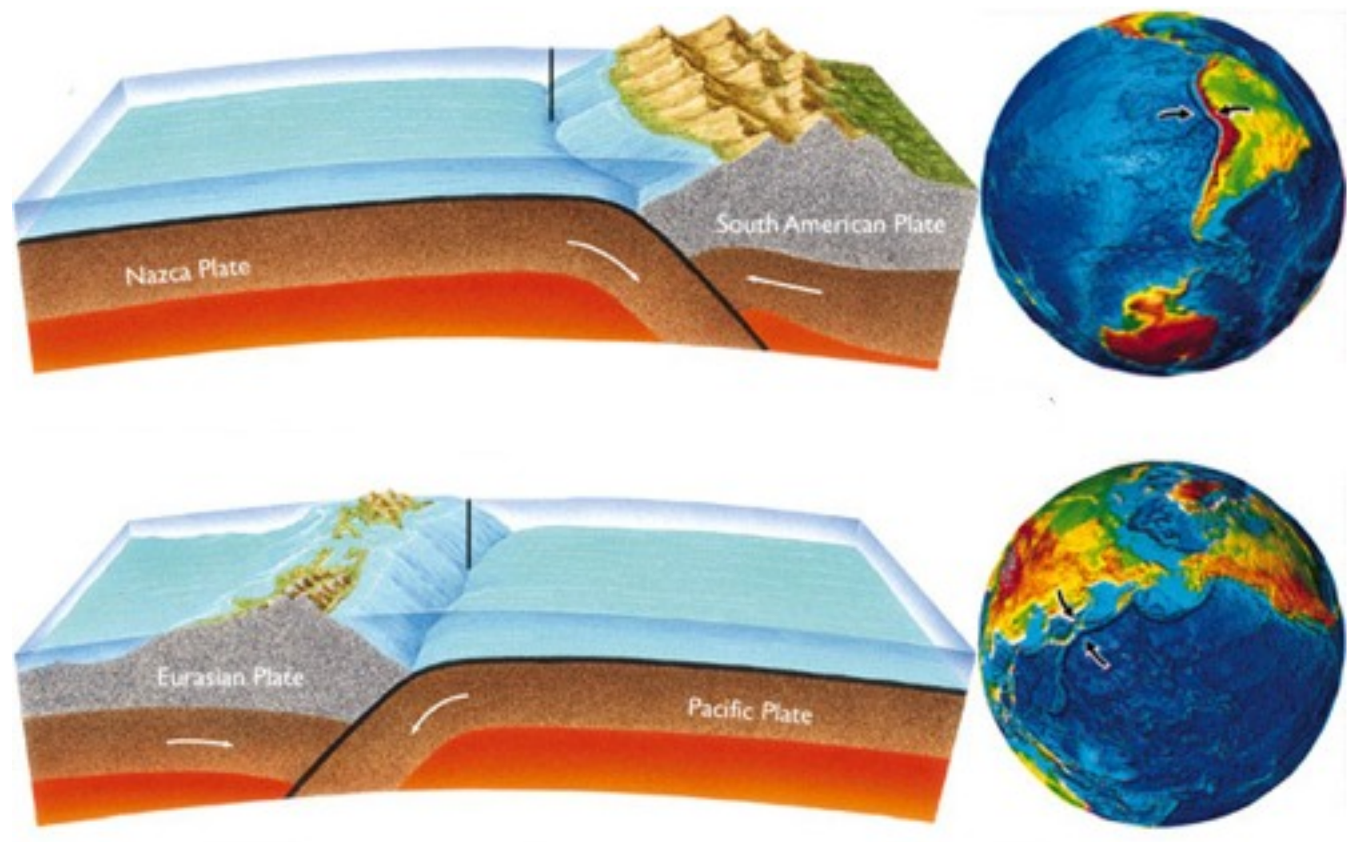
- Tectonic plates are large parts of lithosphere ‘floating’ on the asthenosphere



- Convective currents move them around with velocities of several cm/year.
- The plates interact with one another in three basic ways:
- They collide
- They move away from each other
- They slide one past another

Interacting plates

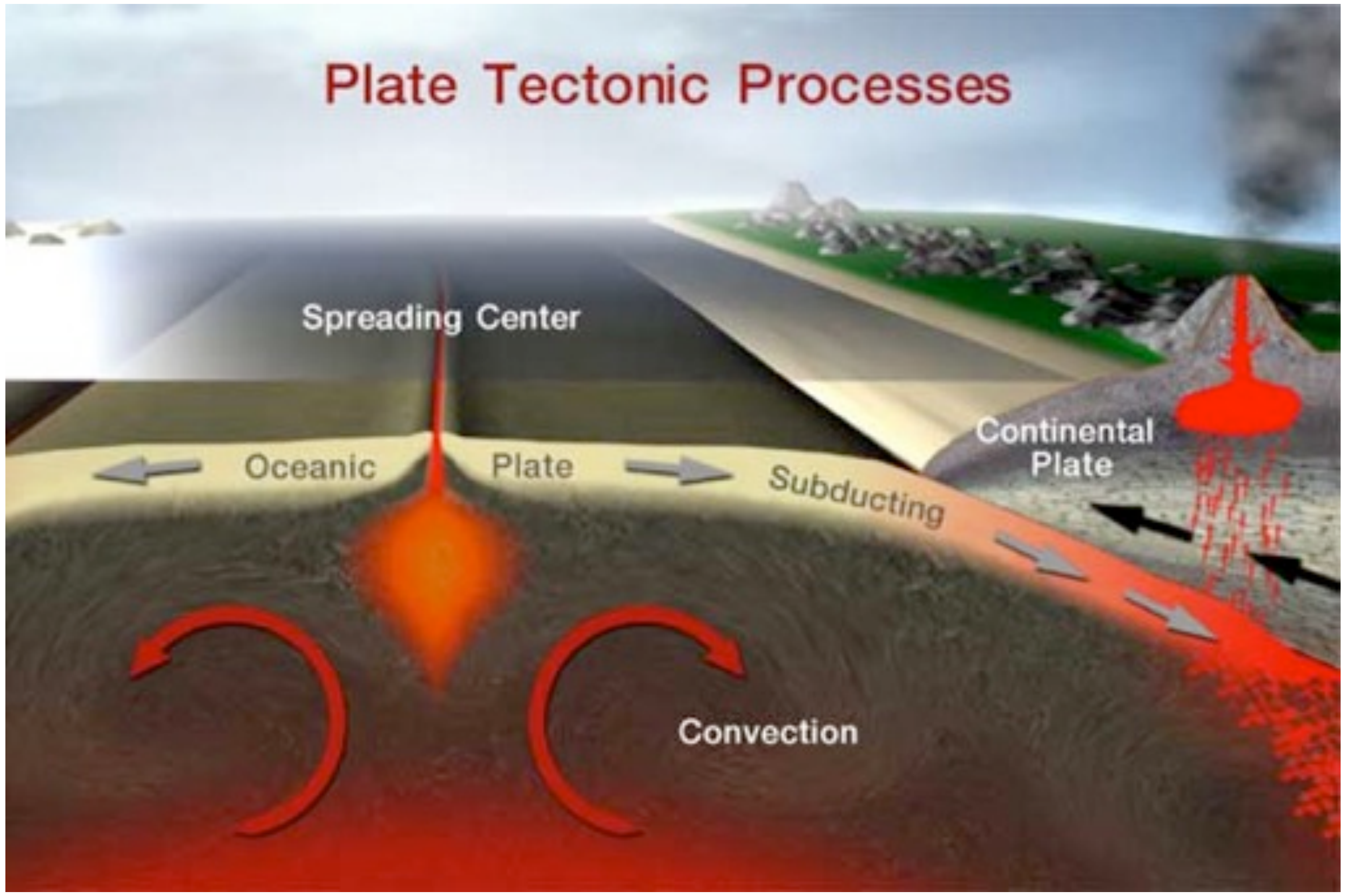
- Collision leads to SUBDUCTION of one plate under another. Mountain ranges may also be formed (Himalayas, Alps...).
- It produces strong and sometimes very deep earthquakes (up to 700 km).
- Volcanoes also occur there.



EXAMPLES: Nazca - South America - Eurasia - Pacific

Interacting plates

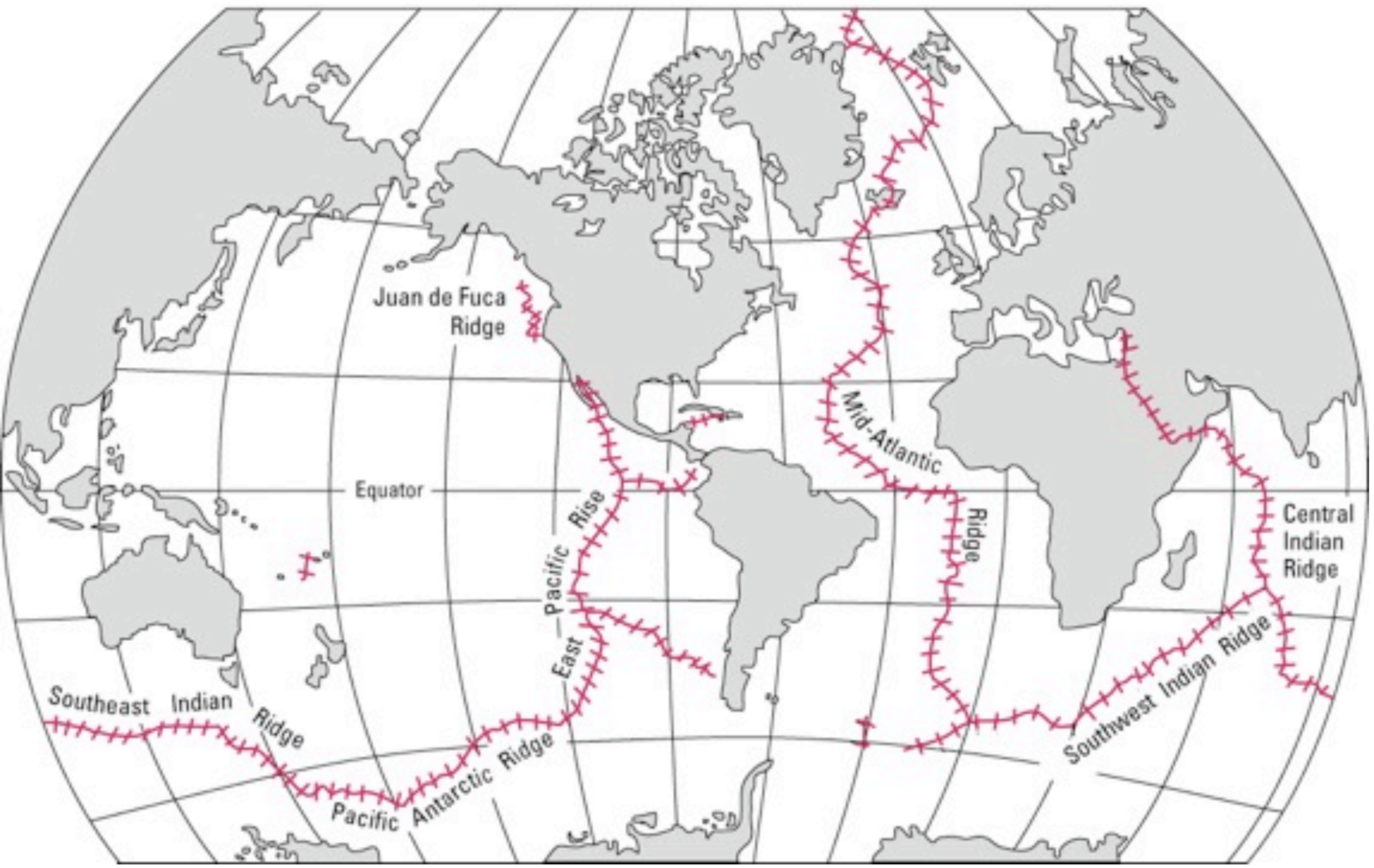
- Plates moving away from each other produce **RIDGES** between them (spreading centres).
- The earthquakes are generally weaker than in the case of subduction.



EXAMPLES: Mid-Atlantic ridge (African - South American plates, Euroasian -North American plates)

Plate Tectonics – Mid-oceanic ridges

Global ridge system



Topography mid-atlantic ridge

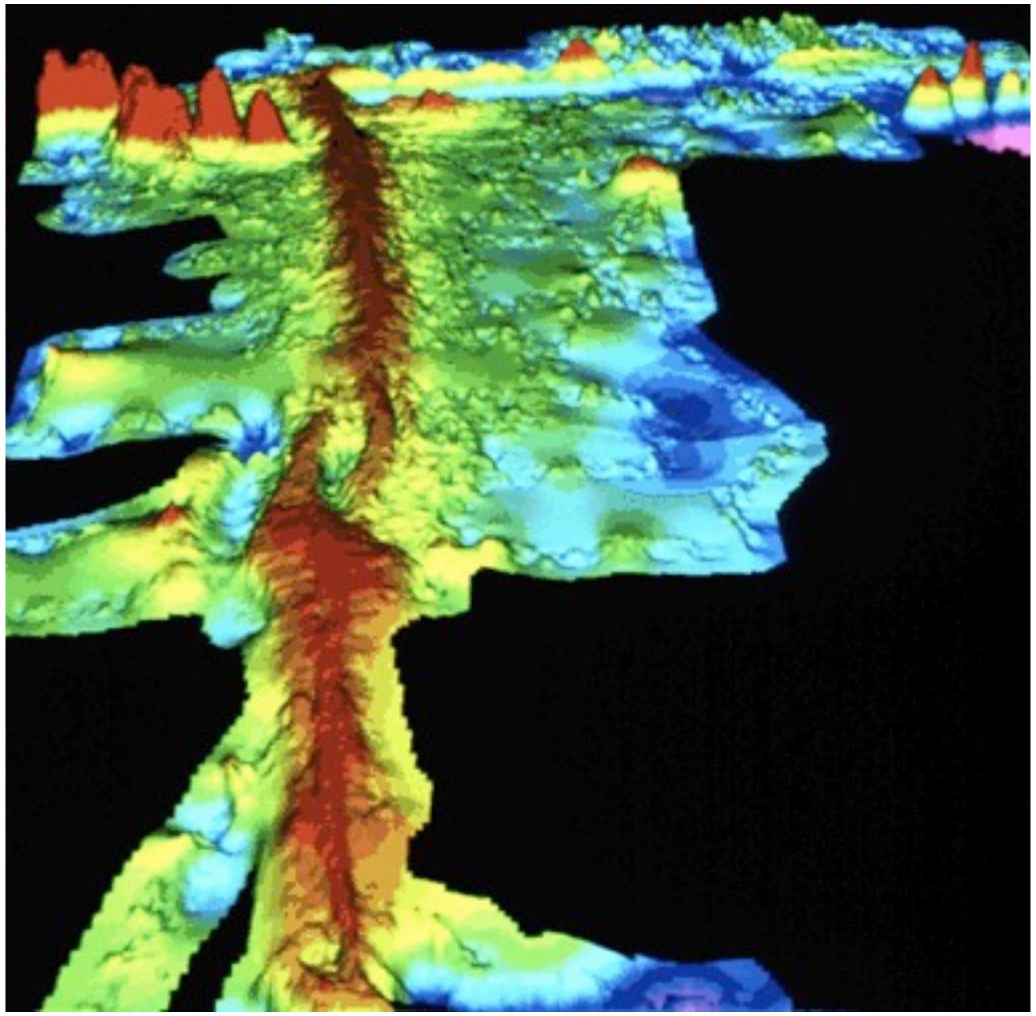
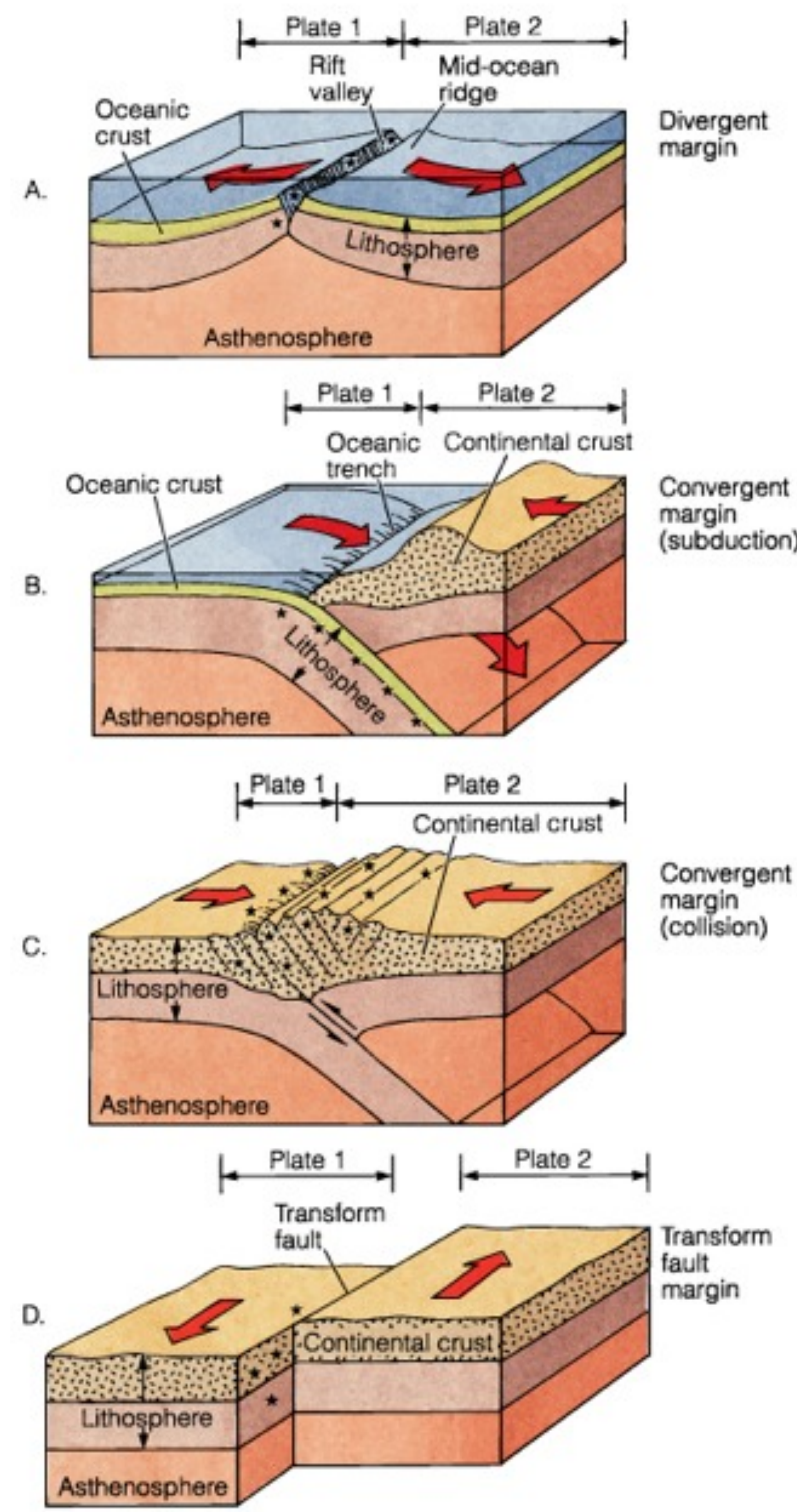


Plate motions are up to 15cm per year

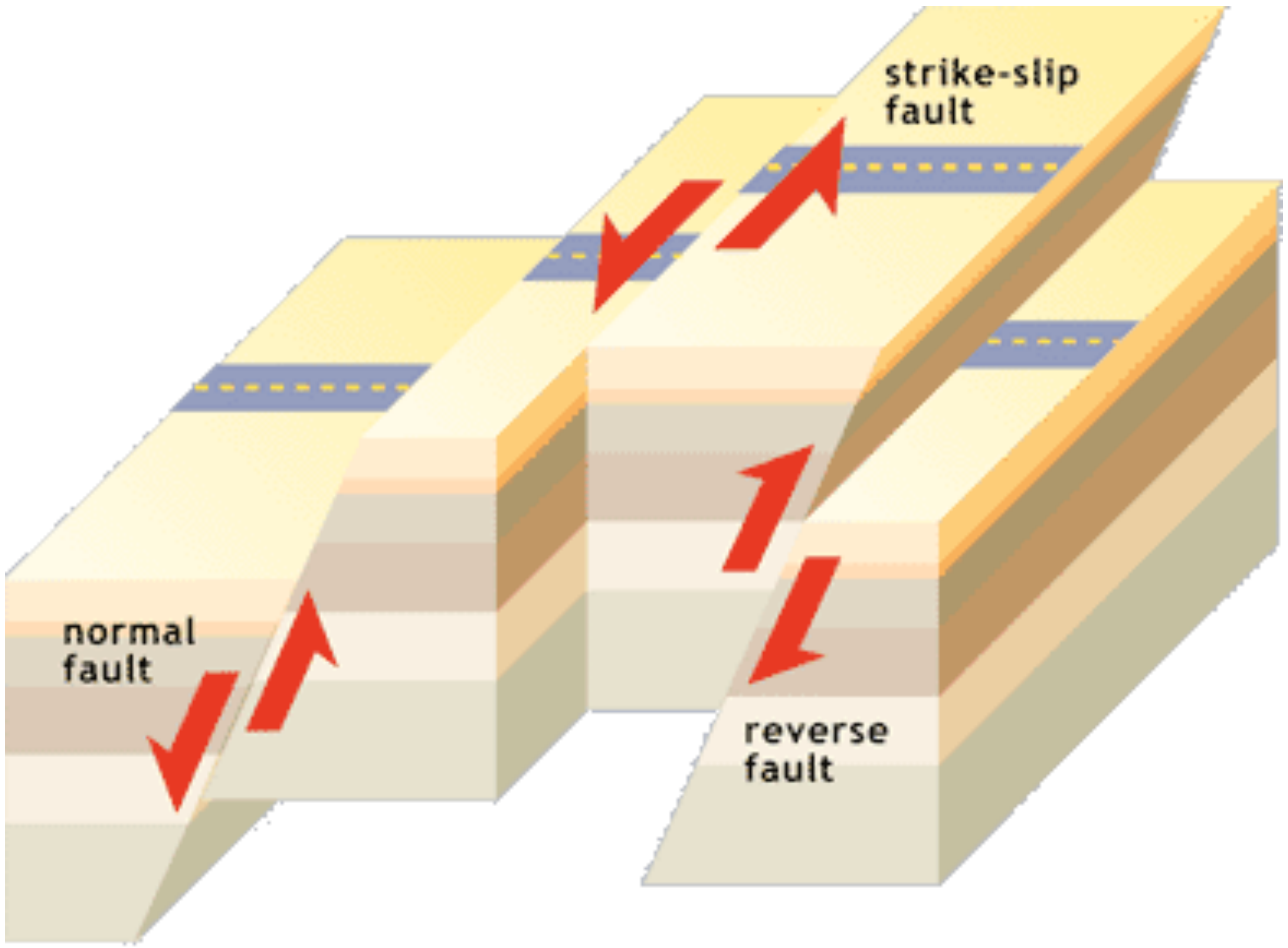
Interacting plates

- Plates moving past each other do so along the **TRANSFORM FAULTS.**
- The earthquakes may be very strong.

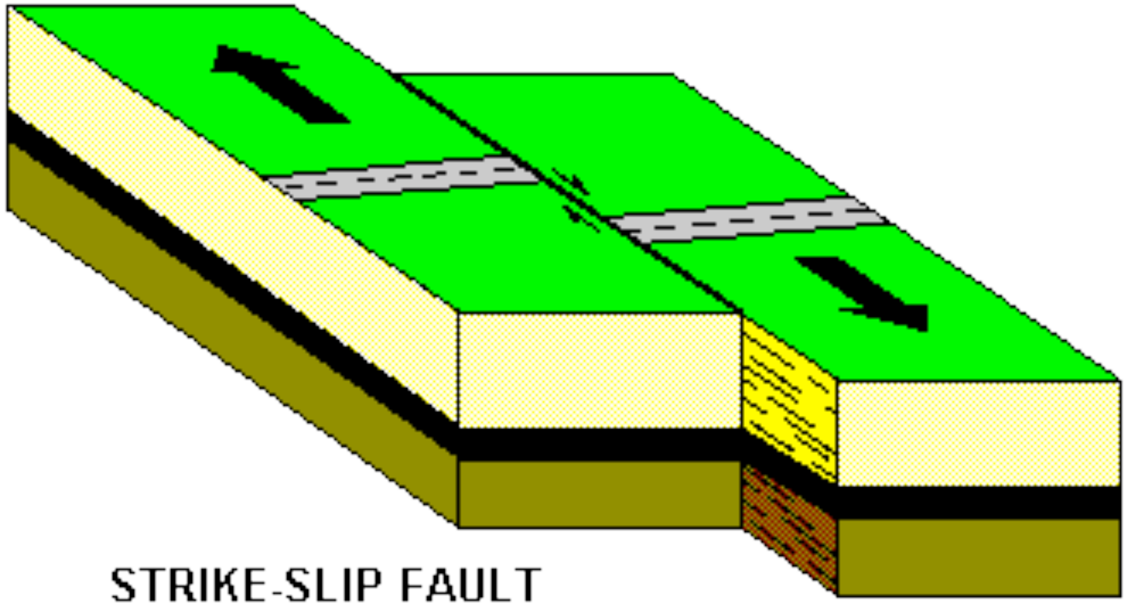
EXAMPLES: San Andreas Fault (Pacific – North American plate)



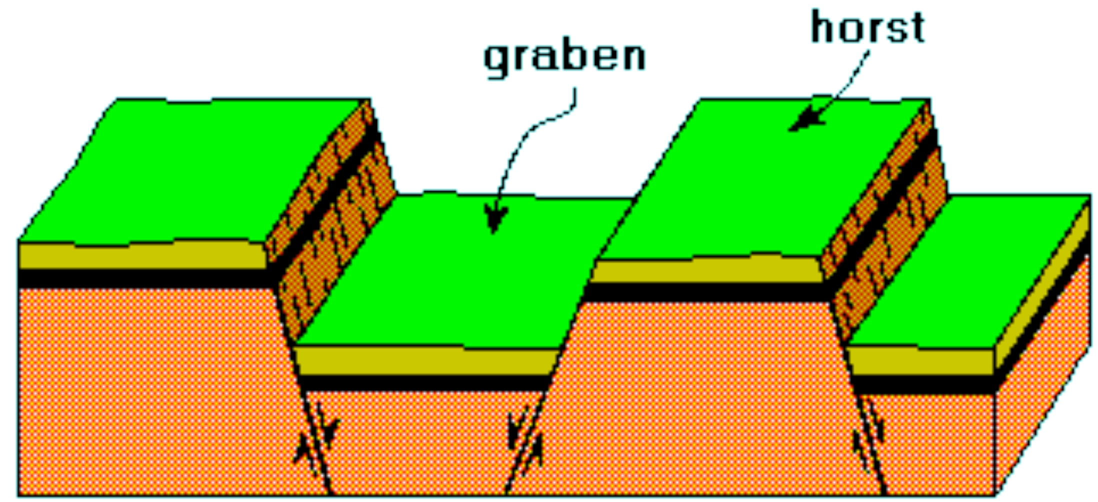
- Earthquakes occur at **FAULTS**.
- Fault is a weak zone separating two geological blocks.
- Tectonic forces cause the blocks to move relative one to another.



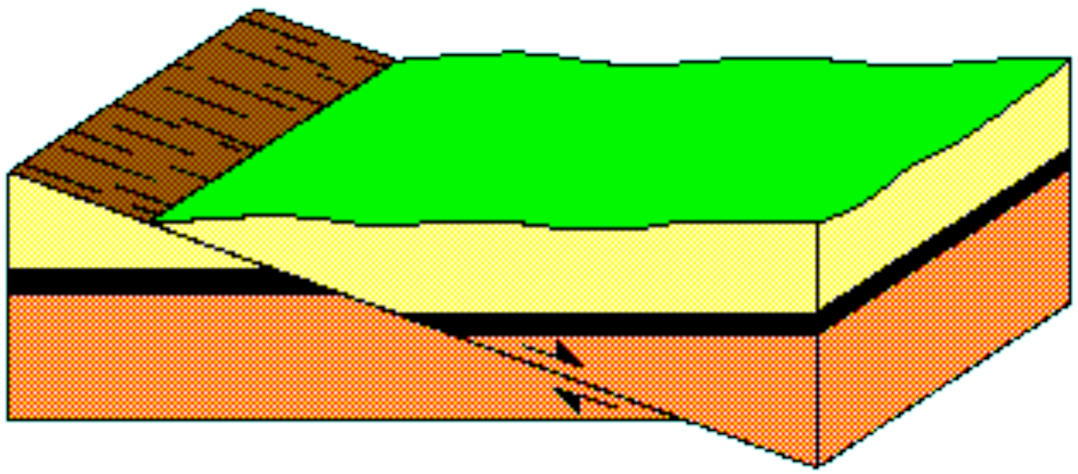
Earthquake sources



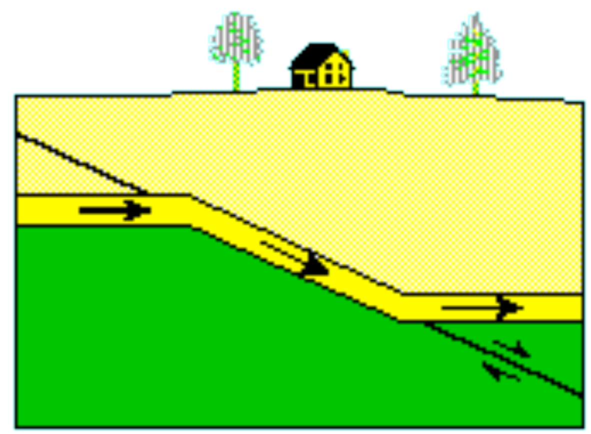
STRIKE-SLIP FAULT



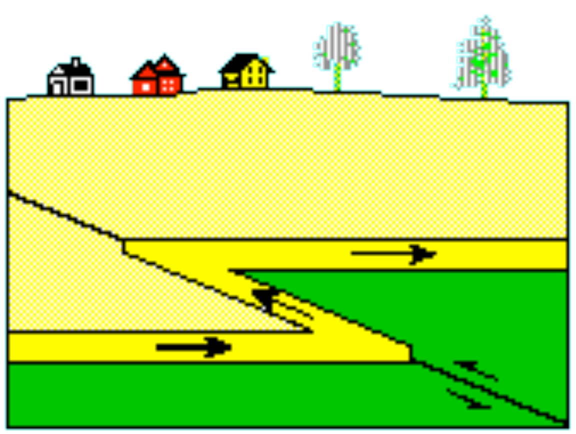
HORST AND GRABEN



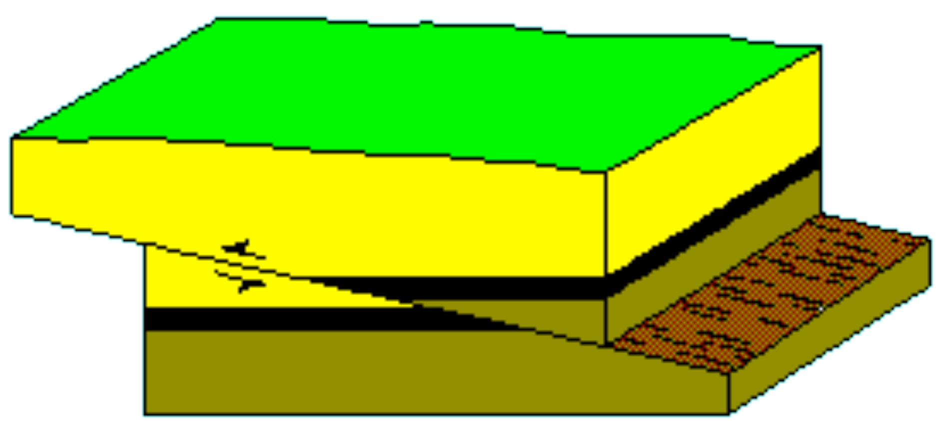
DETACHMENT FAULT



NORMAL

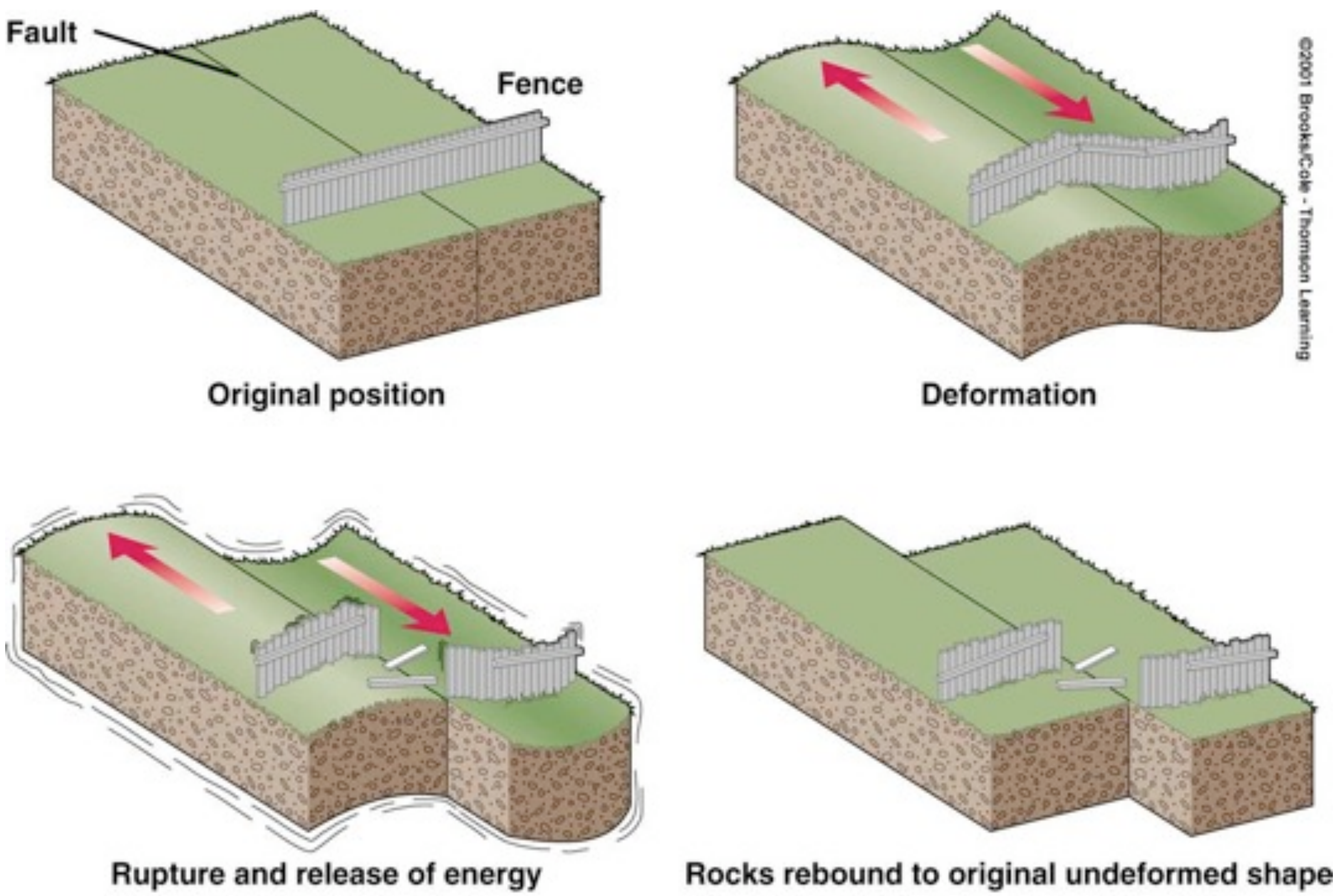


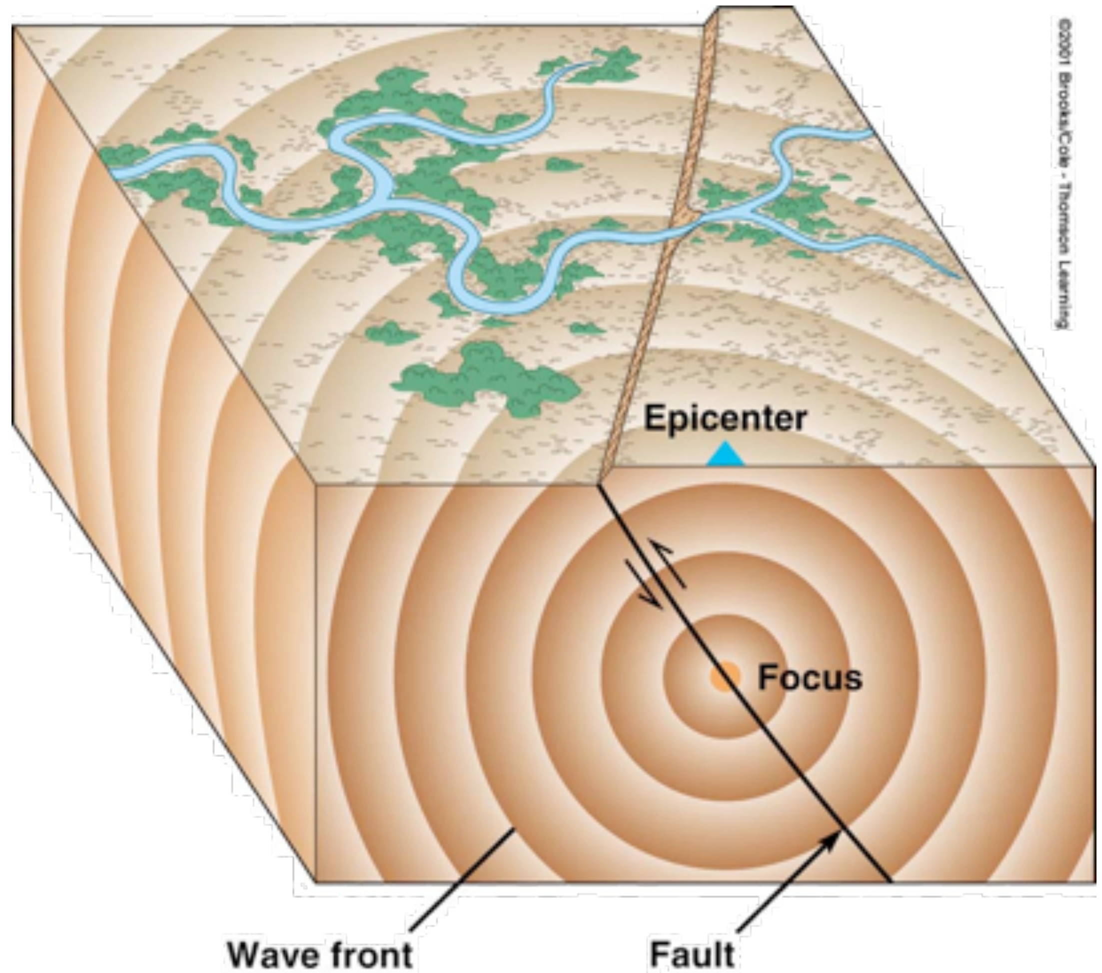
REVERSE



THRUST FAULT

Elastic rebound theory





- Because of friction, the blocks do not slide, but are deformed.
- When the stresses within rocks exceed friction, rupture occurs.
- Elastic energy, stored in the system, is released after rupture in waves that radiate outward from the fault.

Elastic waves – Body waves

Longitudinal waves:

- They are faster than transversal waves and thus arrive first.
- The particles oscillate in the direction of spreading of the wave.
- Compressional waves
- P-waves

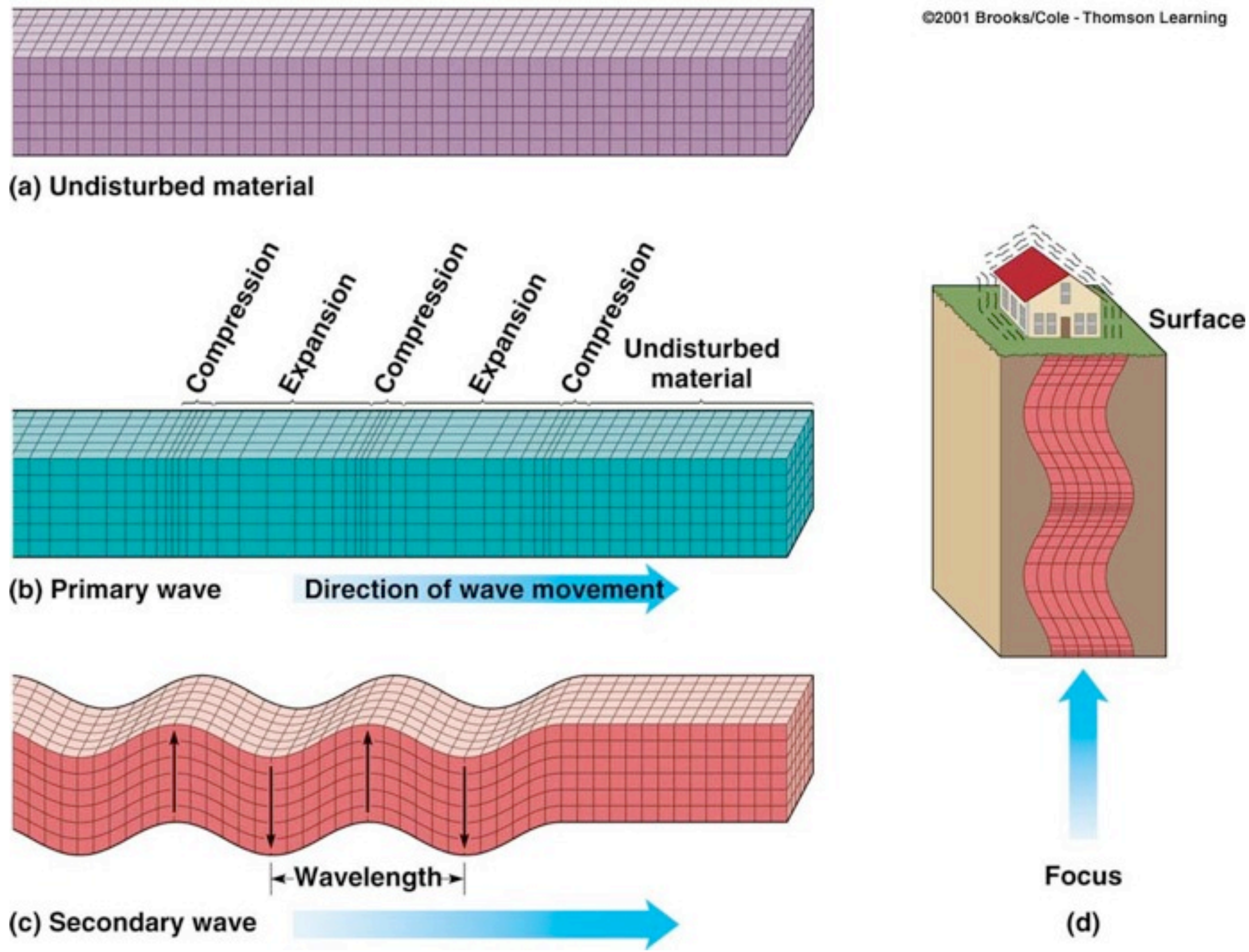
Transversal waves:

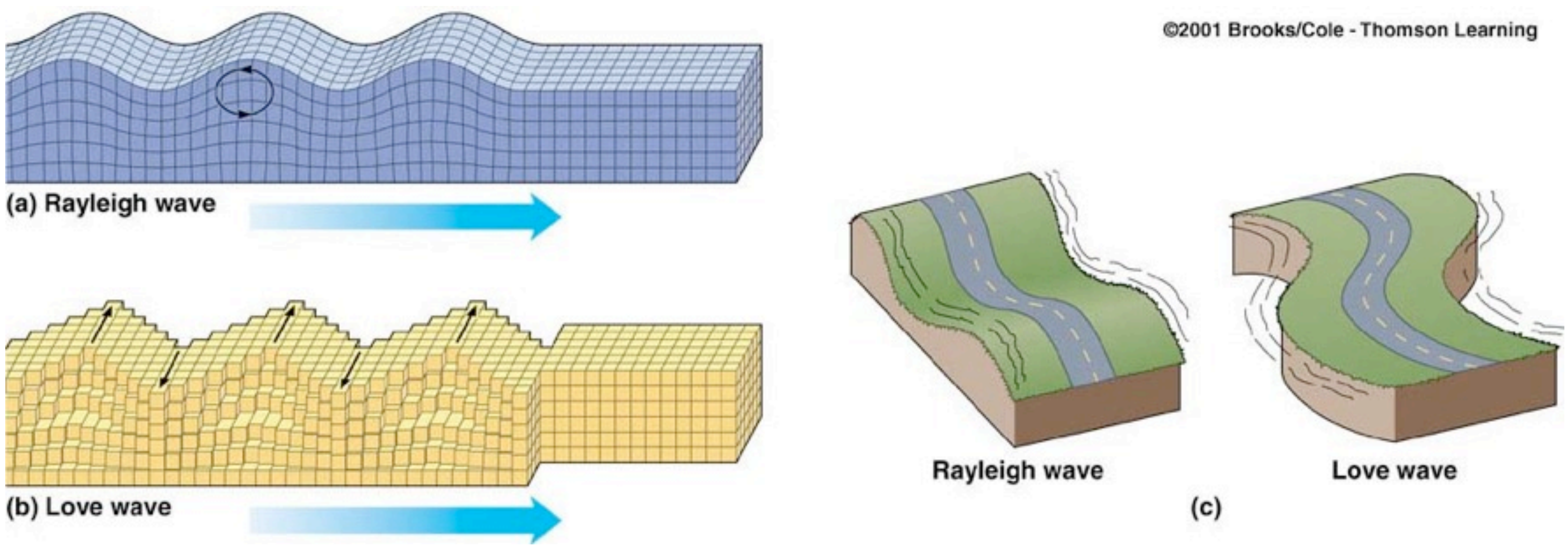
- The particles oscillate in the direction perpendicular to the spreading direction.
- Shear waves – they do not propagate through fluids (e.g. through the outer core).
- S-waves

Elastic waves – Body waves

P-waves:

S-waves:





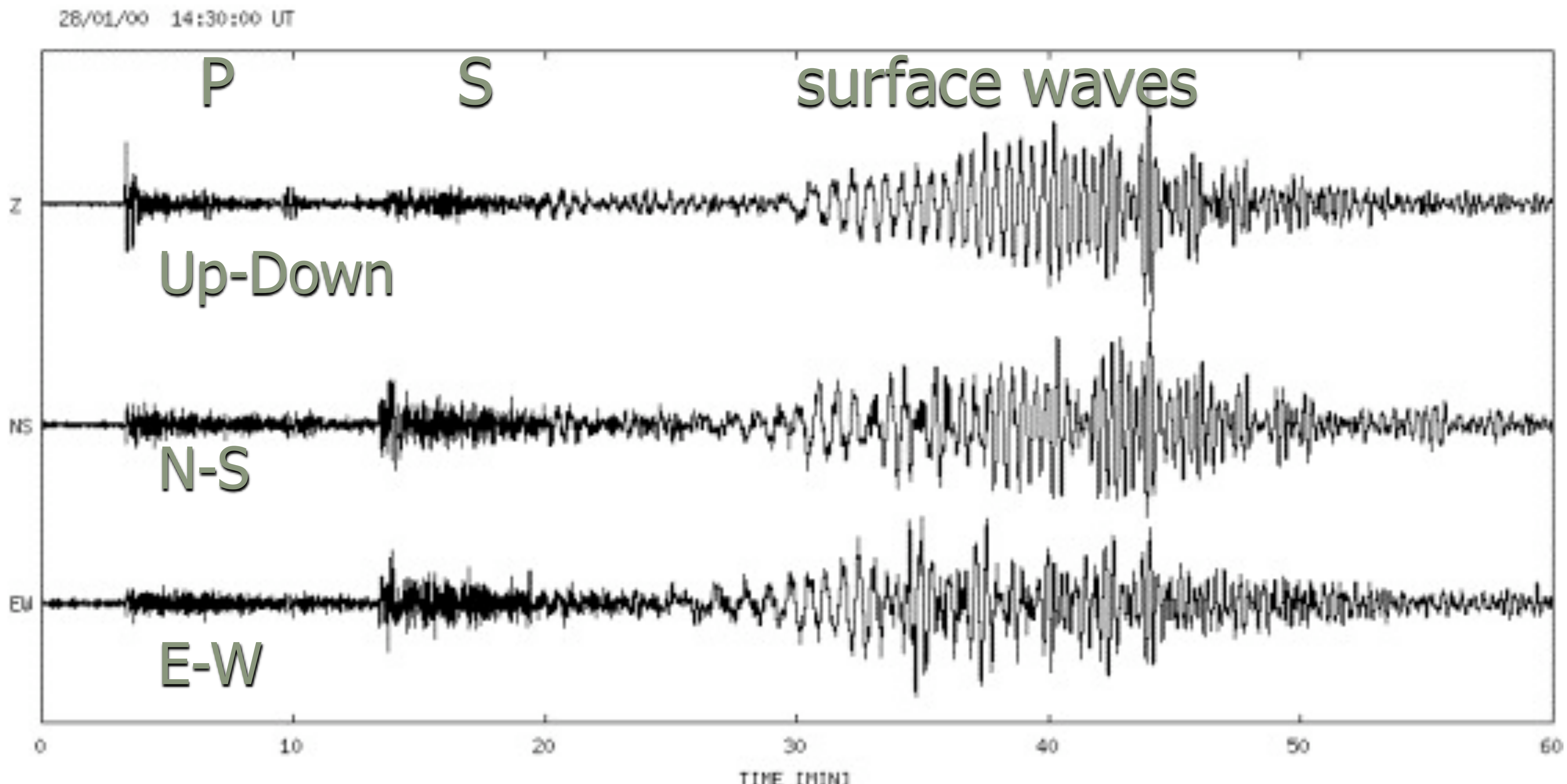
Surface waves: Rayleigh and Love waves

- Their amplitude diminishes with the depth.
- They have large amplitudes and are **slower** than body waves.
- These are dispersive waves (large periods are faster).

Earthquake in Japan

Station in Germany

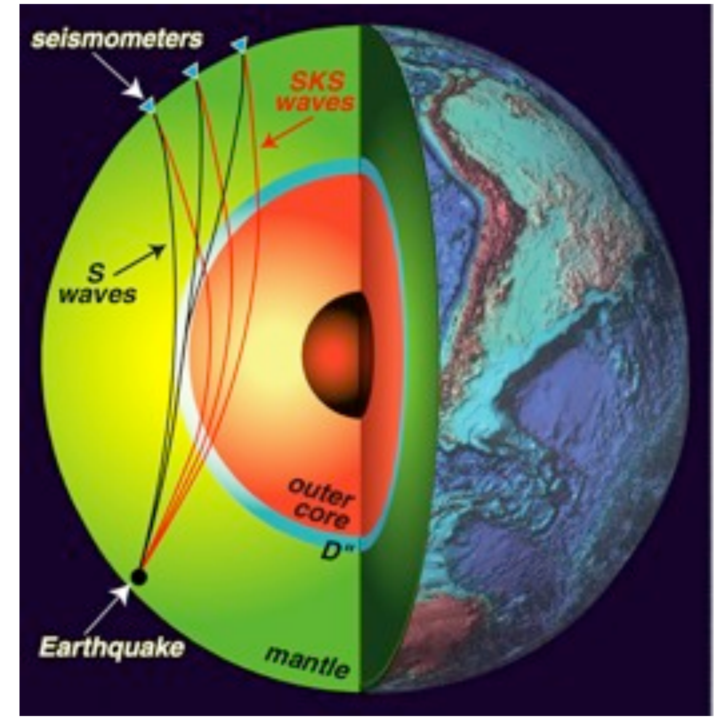
Magnitude 6.5

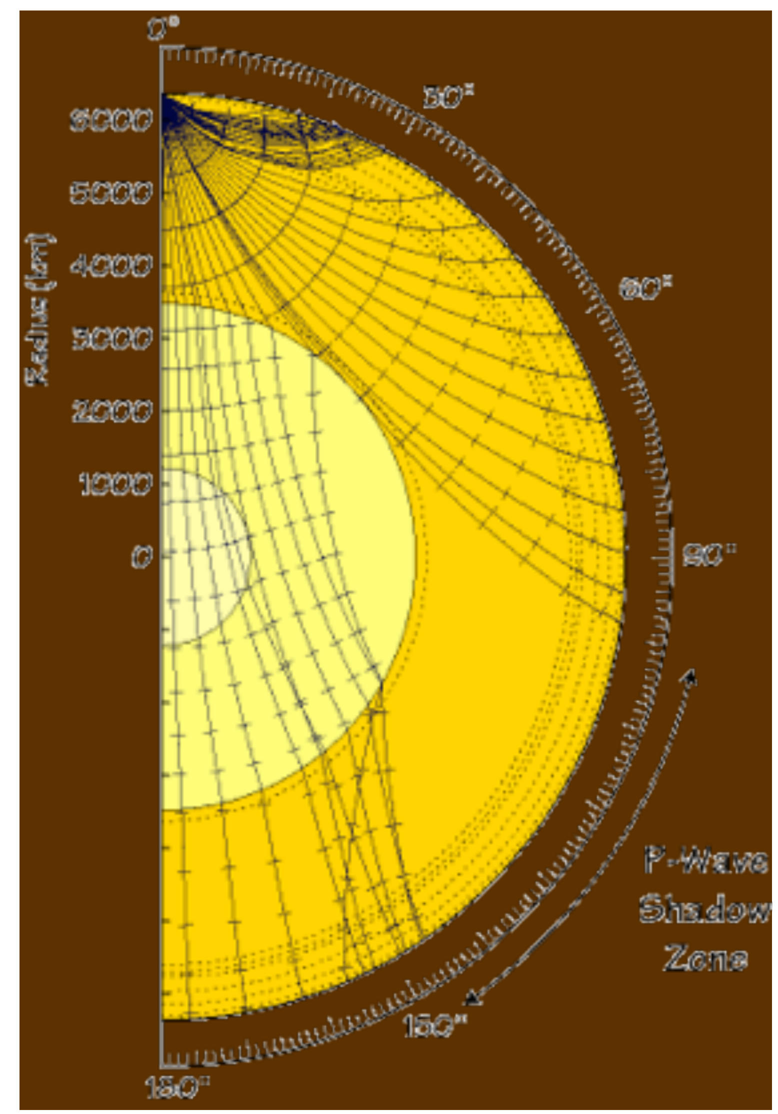


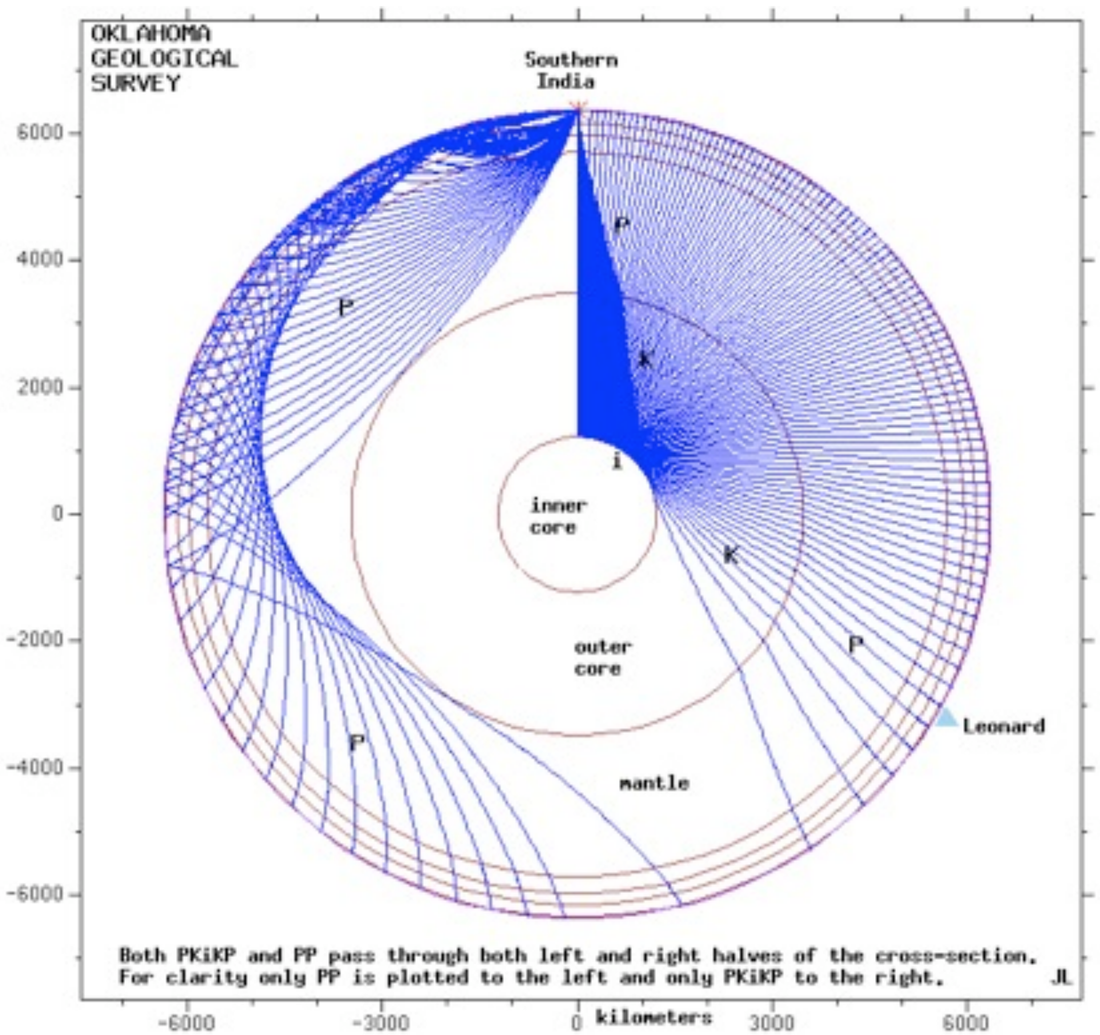
'Physical' Seismology

- Our knowledge about the structure of the Earth deeper than several km was gained almost exclusively using seismological methods.

- Seismologists use seismic rays to look into the interior of the Earth in the same way doctors use X-rays.







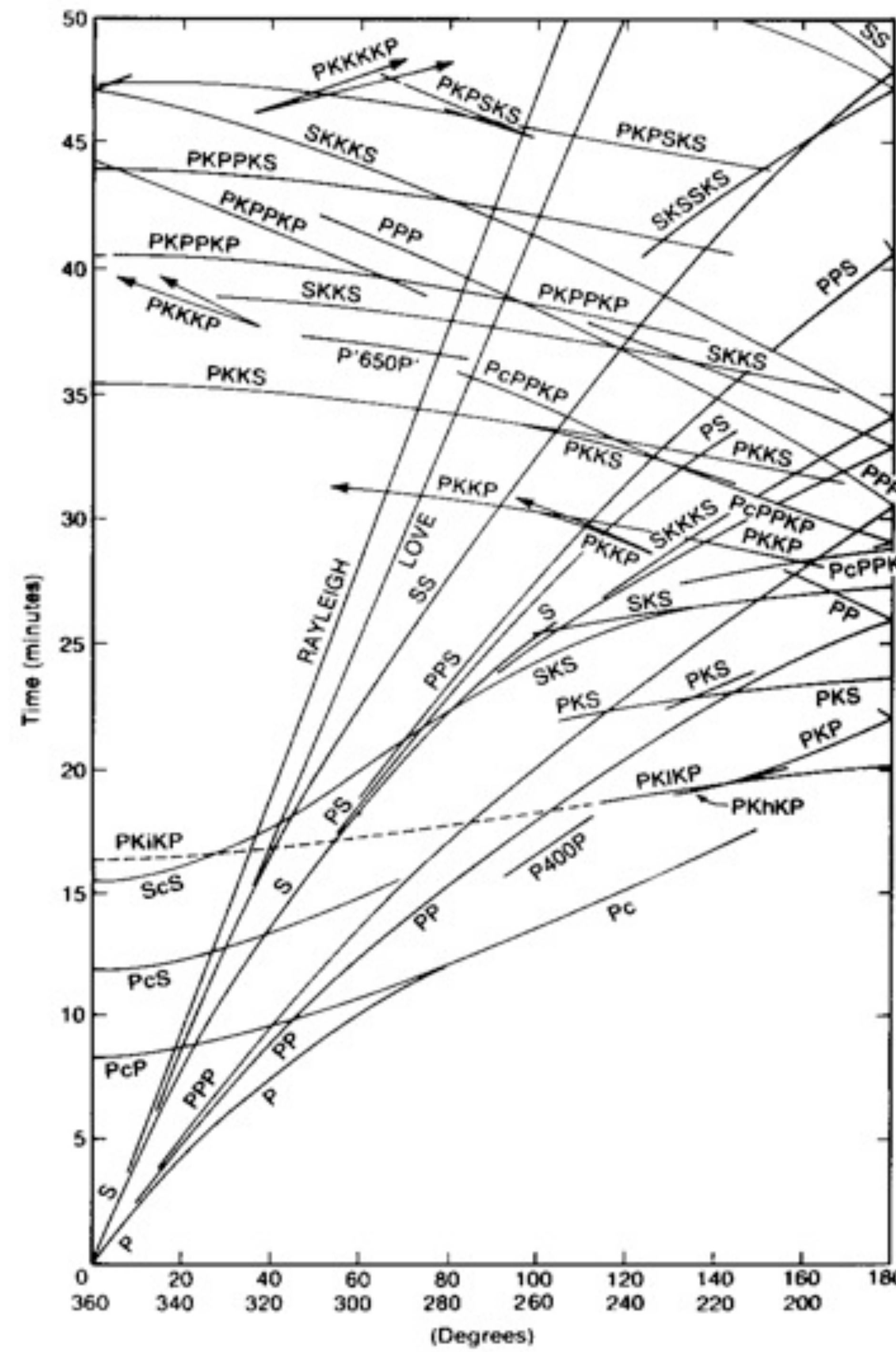


Figure 1.1.2a Travel-time curves for earthquake phases (Bolt, 1976).

Seismic waves get reflected, refracted and converted on many discontinuities within the earth thus forming numerous seismic phases. The rays also bend because the velocity of elsastic waves changes with depth.

Forward problem

- Given the distribution of velocity, density and attenuation coefficient with depth, and positions of all discontinuities, calculate travel times and amplitudes of some seismic phase (e.g. pP or SKS).

This is relatively easy and always gives unique solution.

Inverse problem

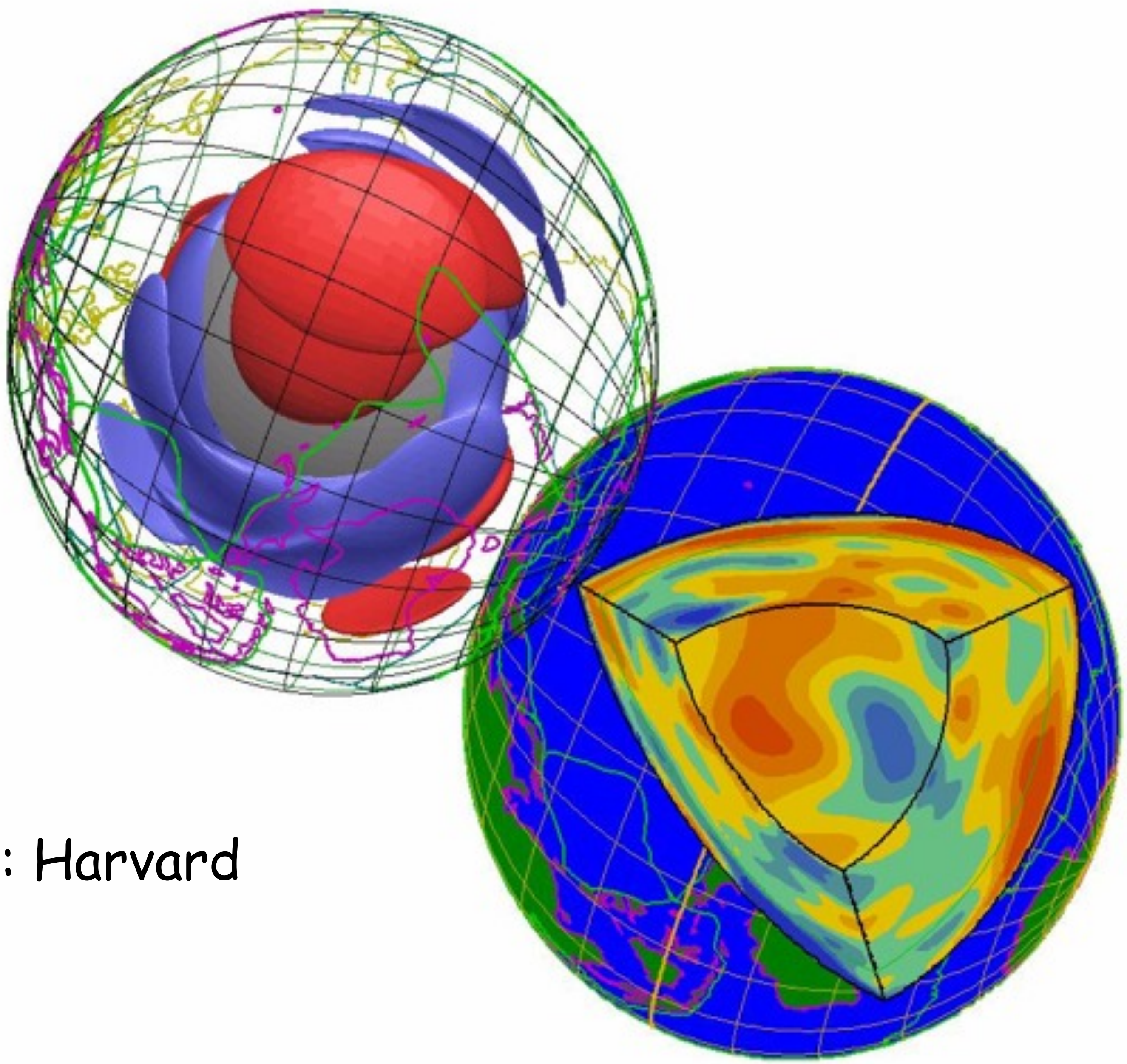
- Given the arrival times and amplitudes of several seismic phases on a number of stations, compute distribution of velocity, density and attenuation coefficient with depth, and positions of all discontinuities.

This is very difficult and often does not give a unique solution. Instead, a range of solutions is offered, each with its own probability of being correct. The solution is better the more data we have.

Tomography

- Seismic tomography gives us 3-D or 2-D images of shallow and deep structures in the Earth. They may be obtanied using earthquake data, or explosions (controlled source seismology). These methods are also widely used in explorational geophysics in prospecting for oil and ore deposits.

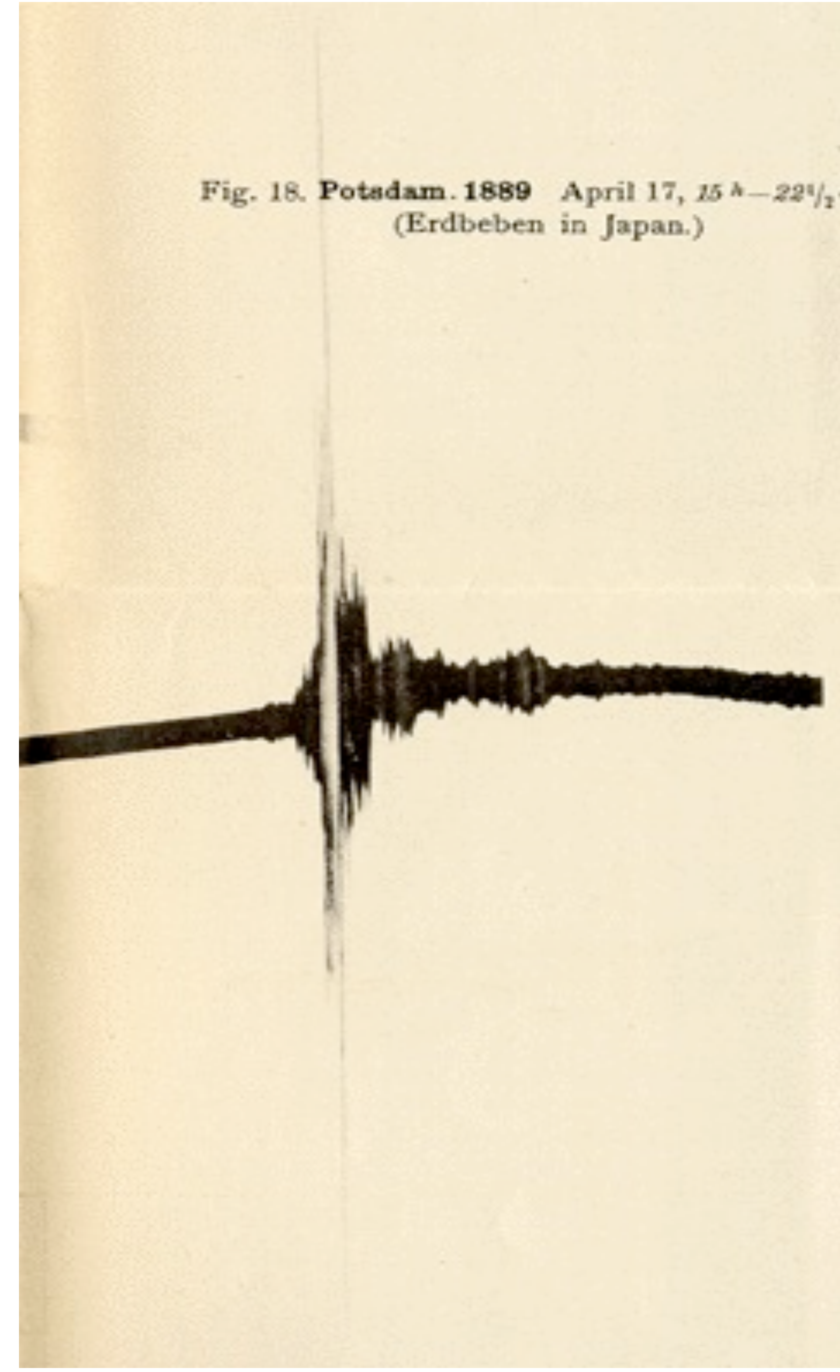
Maybe the most important goal in global seismology today is to determine the Earth's global 3-D structure with high resolution-



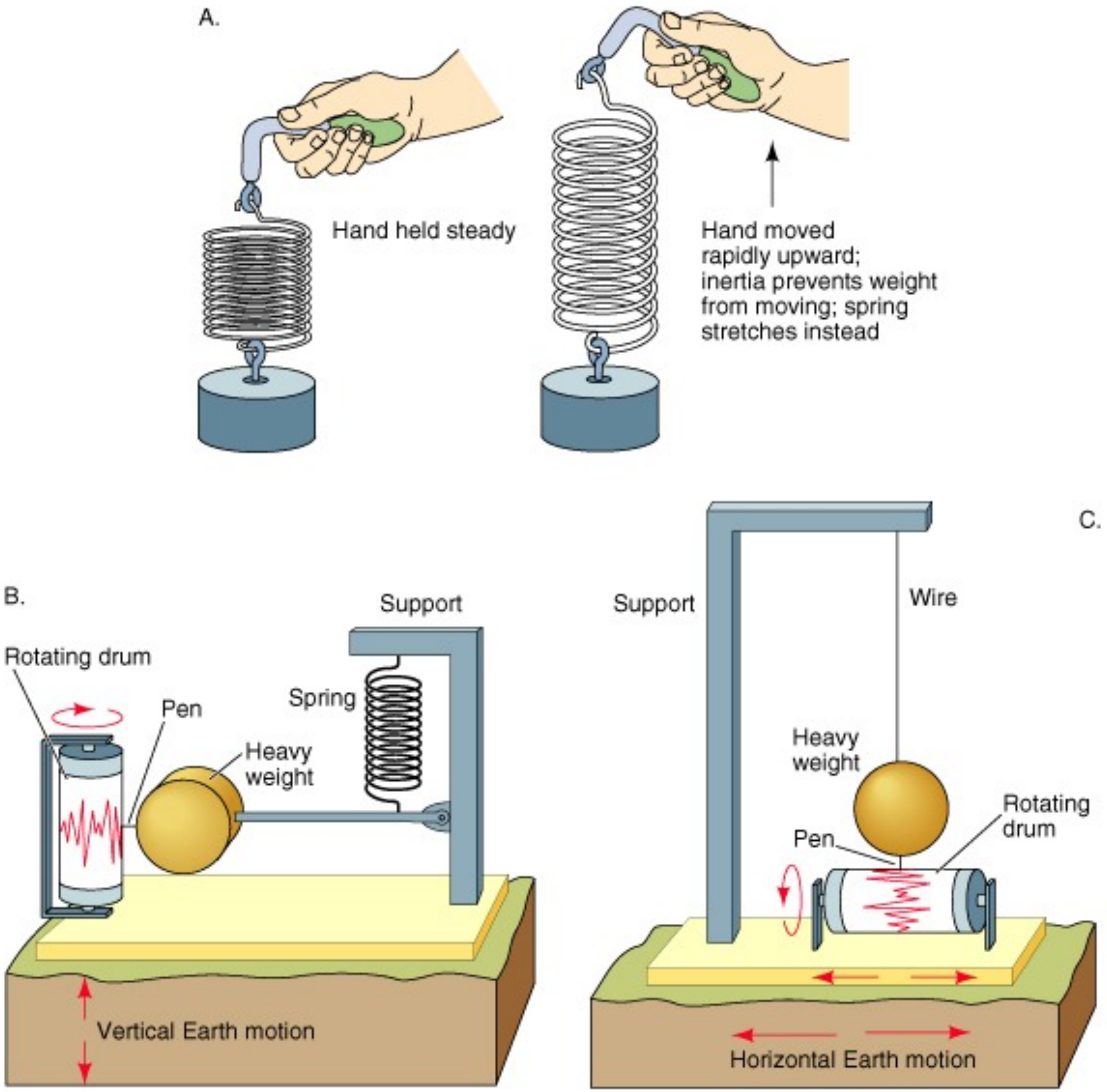
Source: Harvard

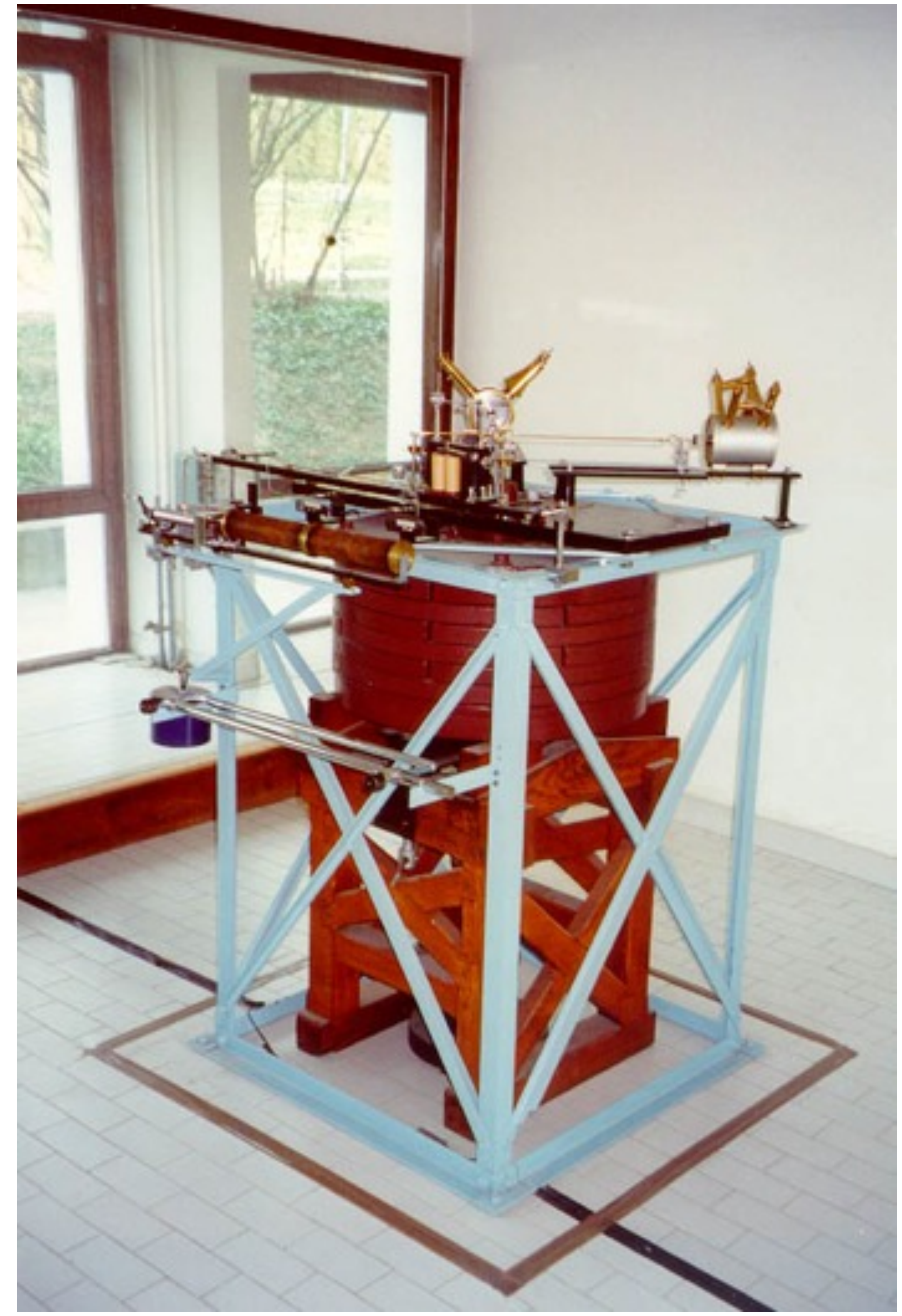
Seismographs

- Seismographs are devices that record ground motion during earthquakes.
- The first seismographs were constructed at the very end of the 19th century in Italy and Germany.



Seismographs





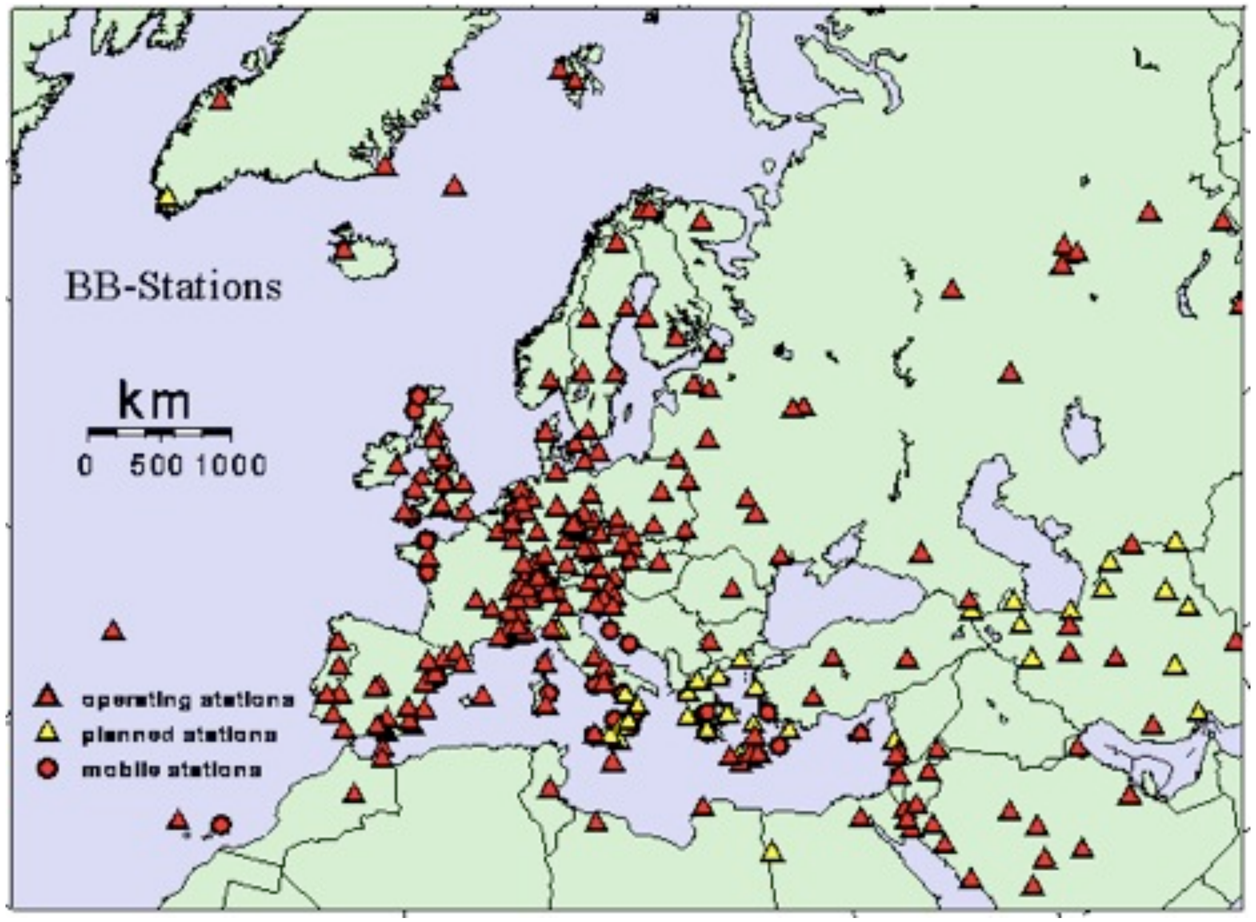
Horizontal 1000 kg Wiechert seismograph in Zagreb (built in 1909)

Seismographs

- Modern digital broadband seismographs are capable of recording almost the whole seismological spectrum (50 Hz – 0.002 Hz).
- Their resolution of 24 bits (high dynamic range) allows for precise recording of small quakes, as well as unsaturated registration of the largest ones.

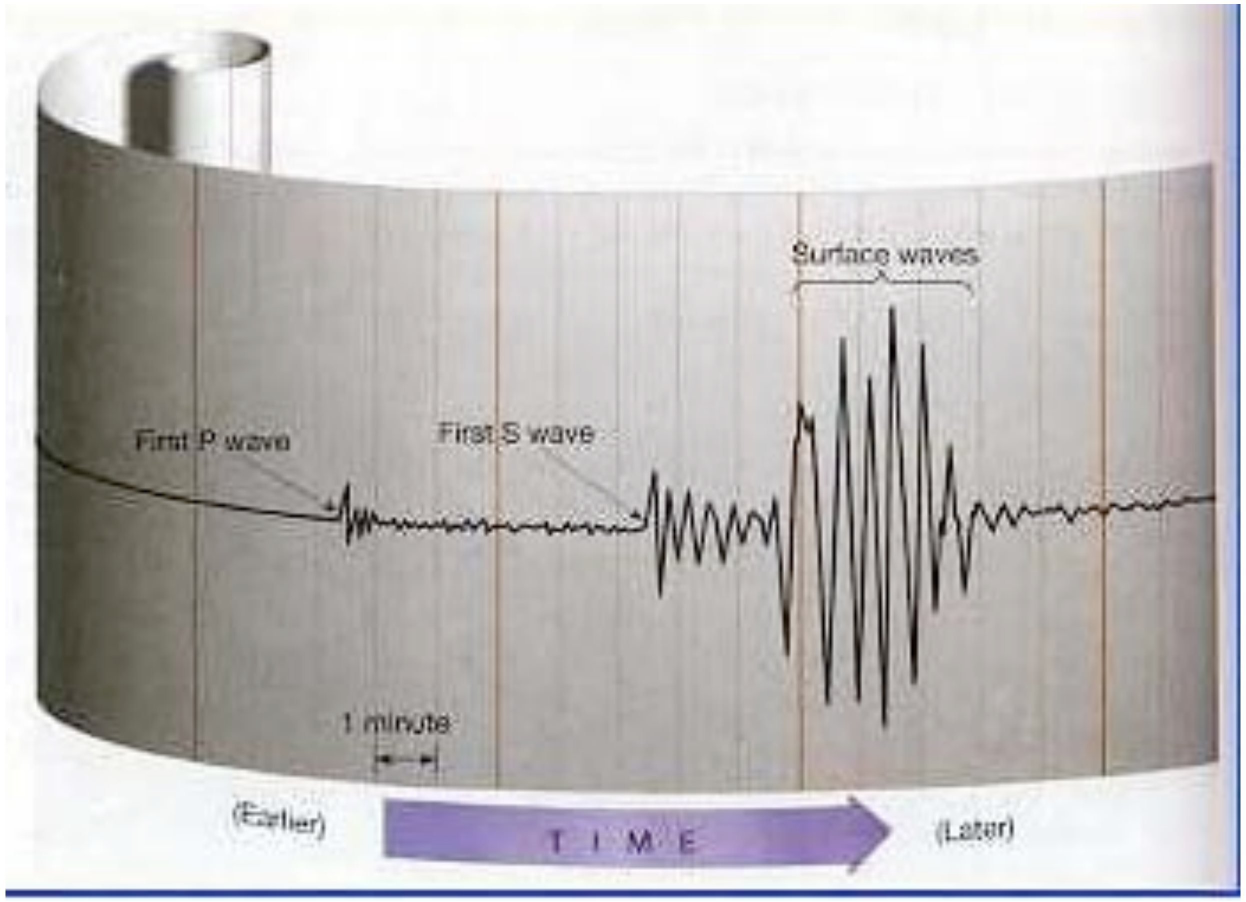
Observational Seismology

- We are now equipped to start recording and locating earthquakes. For that we need a seismic network of as many stations as possible.
- Minimal number of stations needed to locate the position of an earthquake epicentre is three.



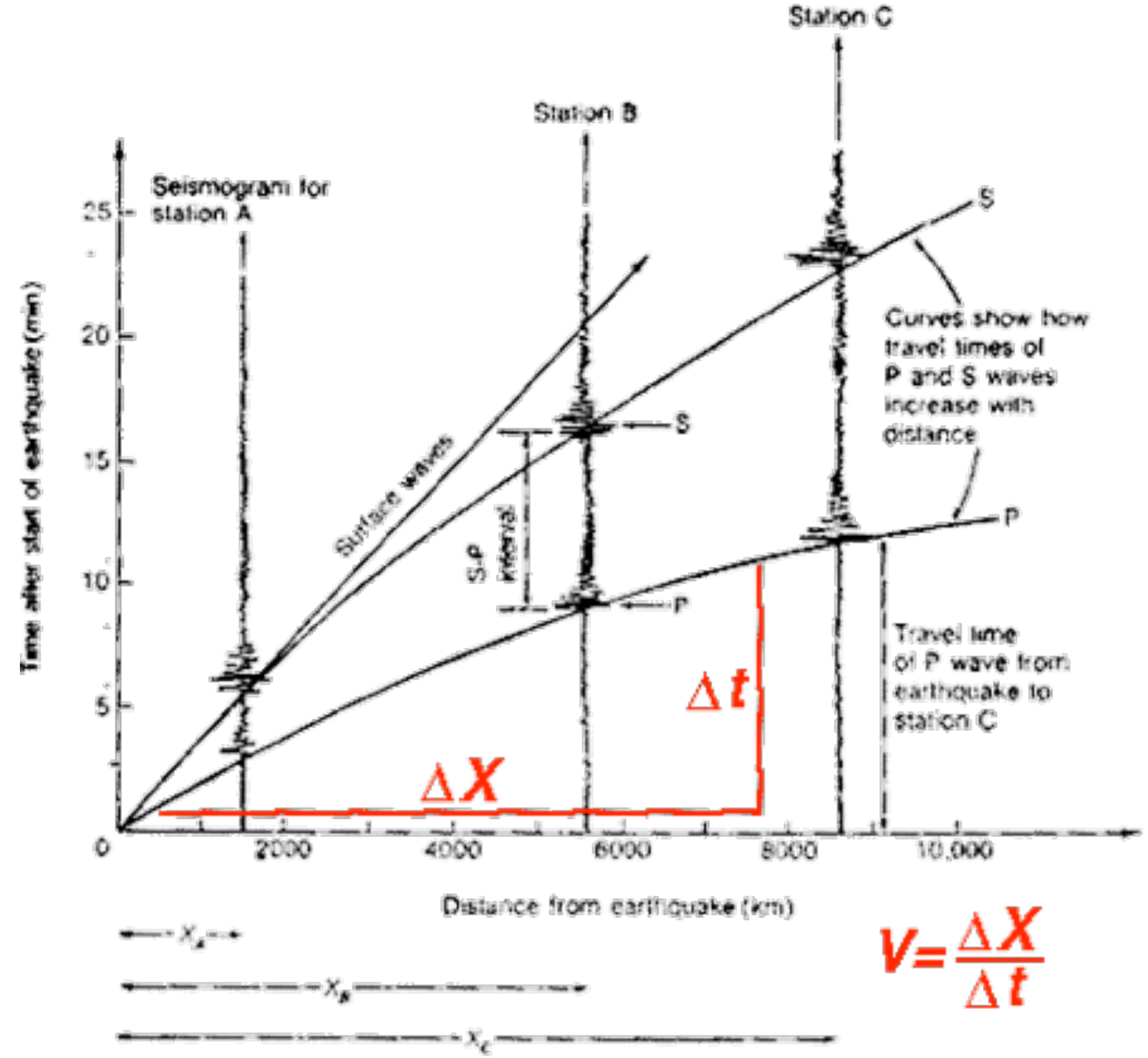
Broad-band seismological stations in Europe

- To locate an earthquake we need precise readings of the times when P- and S-waves arrive at a number of seismic stations.
- Accurate absolute timing (with a precission of 0.01 s) is essential in seismology!



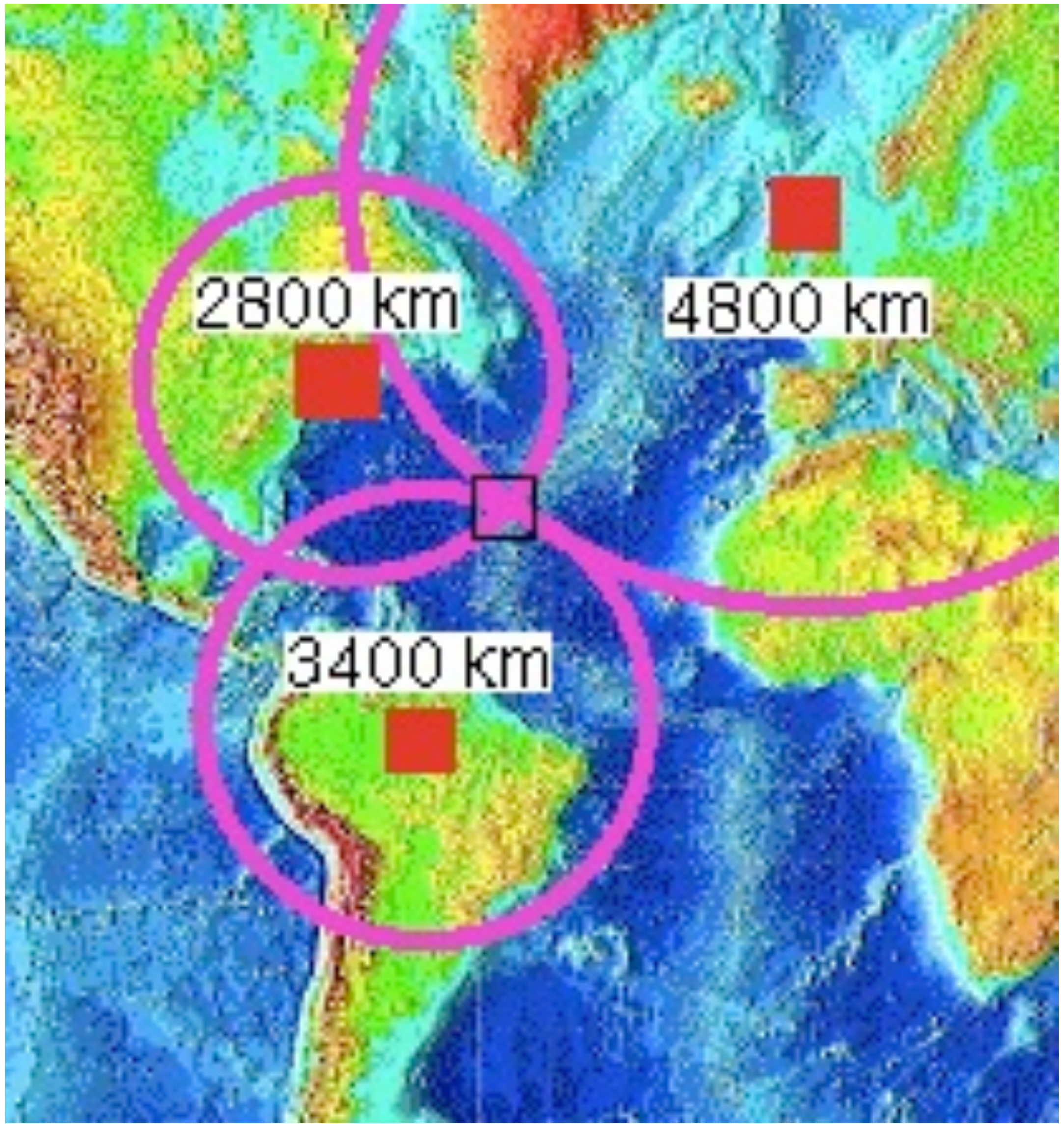
Locating Earthquakes

- Knowing the difference in arrival times of the two waves, and knowing their velocity, we may calculate the distance of the epicentre.
- This is done using the travel-time curves which show how long does it take for P- and S-waves to reach some epicentral distance.



Locating Earthquakes

- After we know the distance of epicentre from at least three stations we may find the epicentre like this
- There are more sophisticated methods of locating positions of earthquake foci. This is a classic example of an inverse problem.

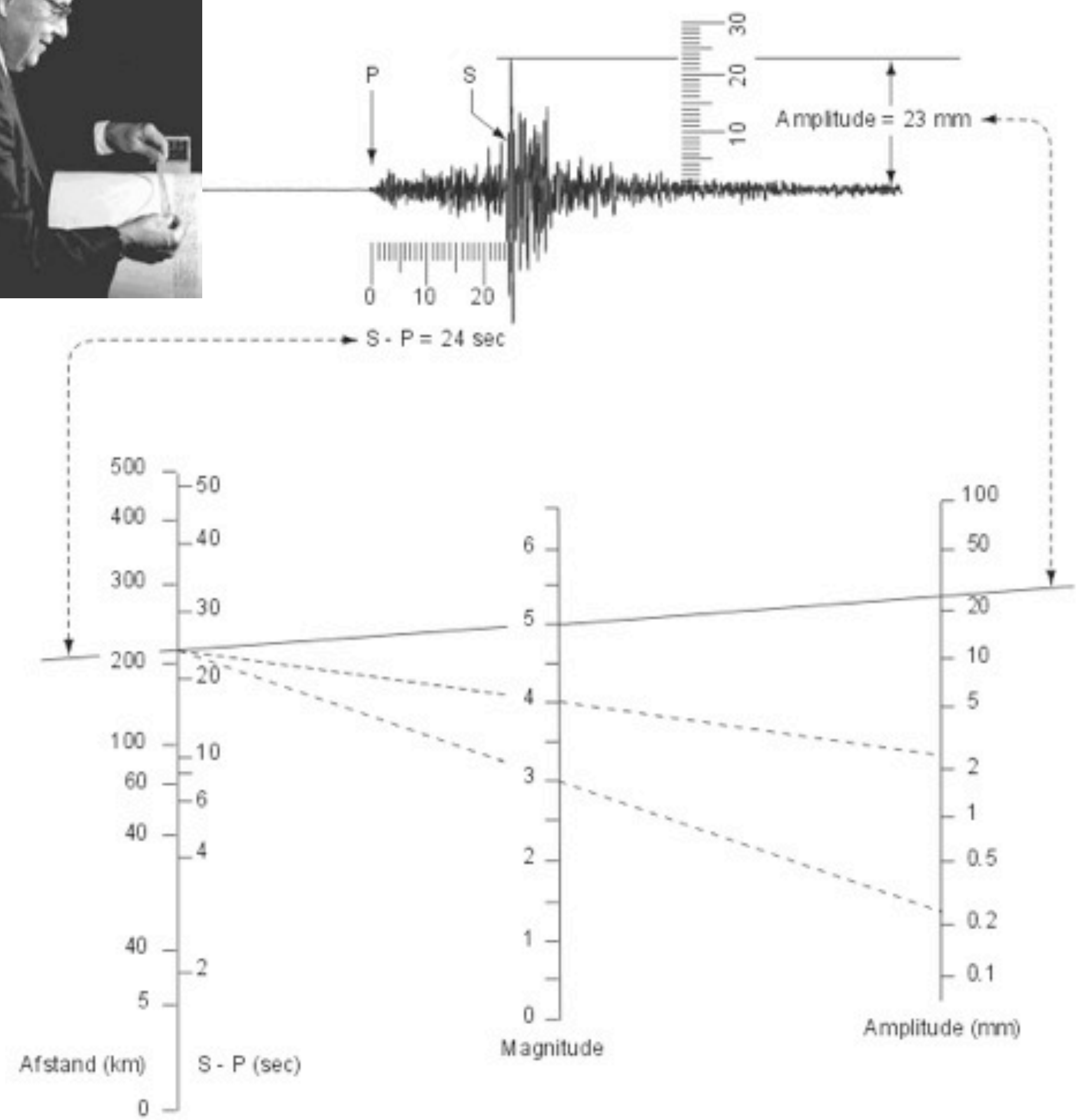


Magnitude determination

Besides the position of the epicentre and the depth of focus, the earthquake magnitude is another defining element of each earthquake.

Magnitude (defined by Charles Richter in 1935) is proportional to the amount of energy released from the focus.

Magnitude is calculated from the amplitudes of ground motion as measured from the seismograms. You also need to know the epicentral distance to take attenuation into account.



Magnitude determination

Formula:

$M = \log(A) + c_1 \log(D) + c_2$

where A is amplitude of ground motion, D is epicentral distance, and c_1, c_2 are constants.

There are many types of magnitude in seismological practice, depending which waves are used to measure the amplitude: M_L, m_b, M_c, M_s, M_w, ...

Increase of 1 magnitude unit means ~32 times more released seismic energy!

Some statistics

Magnitude	Effects	Number per year
less than 2	Not felt by humans. Recorded by instruments only.	Numerous
2	Felt only by the most sensitive. Suspended objects swing	>1 000 000
3	Felt by some people. Vibration like a passing heavy vehicle	100 000
4	Felt by most people. Hanging objects swing. Dishes and windows rattle and may break	12 000
5	Felt by all; people frightened. Chimneys topple; furniture moves	1 400
6	Panic. Buildings may suffer substantial damage	160
7-8	Widespread panic. Few buildings remain standing. Large landslides; fissures in ground	20
8-9	Complete devastation. Ground waves	~2

http://earthquake.usgs.gov/eqcenter/top10.php

Macroseismology

- MACROSEISMOLOGY deals with effects of earthquakes on humans, animals, objects and surroundings.
- The data are collected by field trips into the shaken area, and/or by questionaires sent there.
- The effects are then expressed as earthquake INTENSITY at each of the studied places.
- Intensity is graded according to **macroseismic scales** – Mercalli-Cancani-Sieberg (**MCS**), Medvedev-Sponheuer-Karnik (**MSK**), Modified Mercalli (**MM**), European Macroseismic Scale (**EMS**).
- This is a subjective method.

Engineering Seismology

- Earthquakes are the only natural disasters that are mostly harmless to humans! The only danger comes from buildings designed not to withstand the largest possible earthquakes in the area.
- **Engineering seismology** provides civil engineers parameters they need in order to construct seismically safe and sound structures.
- **Engineering seismology** is a bridge between seismology and earthquake engineering.

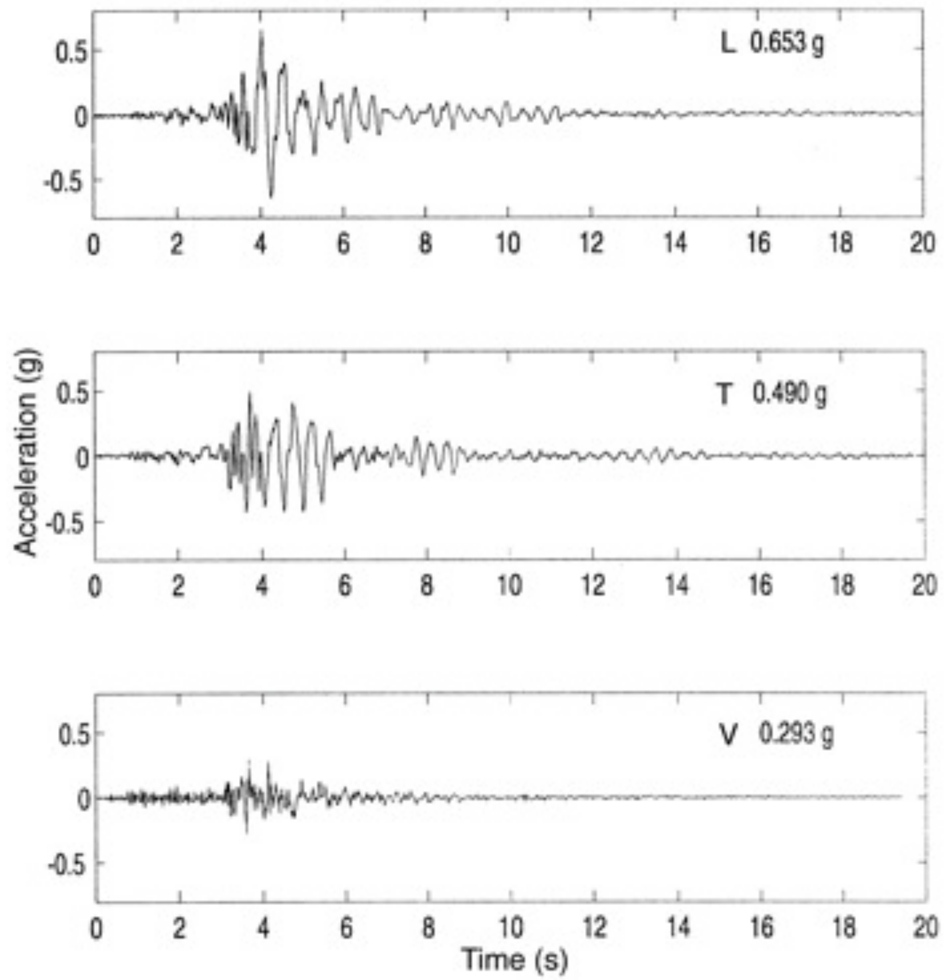
Izmit, Turkey, 1999

Engineering Seismology

Most common input parameters are:

- maximal expected horizontal ground acceleration (PGA)
- maximal expected horizontal ground velocity (PGV)
- maximal expected horizontal ground displacement (PGD)
- (response) spectral acceleration (SA)
- maximal expected intensity (Imax)
- duration of significant shaking
- dominant period of shaking.

Engineering seismologists mostly use records of ground acceleration obtained by strong-motion accelerographs.



Accelerogram of the Ston-Slano (Croatia, M = 6.0, 1996) event

Engineering Seismology

In order to estimate the parameters, seismologists need:

- Complete earthquake catalogues that extend well into the past,
- Information on the soil structure and properties at the construction site, as well as on the path between epicentre and the site,
- Records of strong earthquakes and small events from near-by epicentral regions,
- Results of geological surveys ...

Seismology – Schematically

Seismic Source

Ruptures, crack propagation, physics of earthquakes, magnitude, faulting, seismic creep, radiation pattern, Earthquake precursors, aftershocks, fault planes, etc.

Seismometer

Filtering, (de)convolution, three components, spectrum, broadband, strong-motion, tilt, long-period, amplification, etc.

Propagation Effects

heterogeneities, scattering, attenuation, anisotropy, rays, body waves, surface waves, free oscillations, reflections, refractions, trapped waves, geometrical spreading, etc.