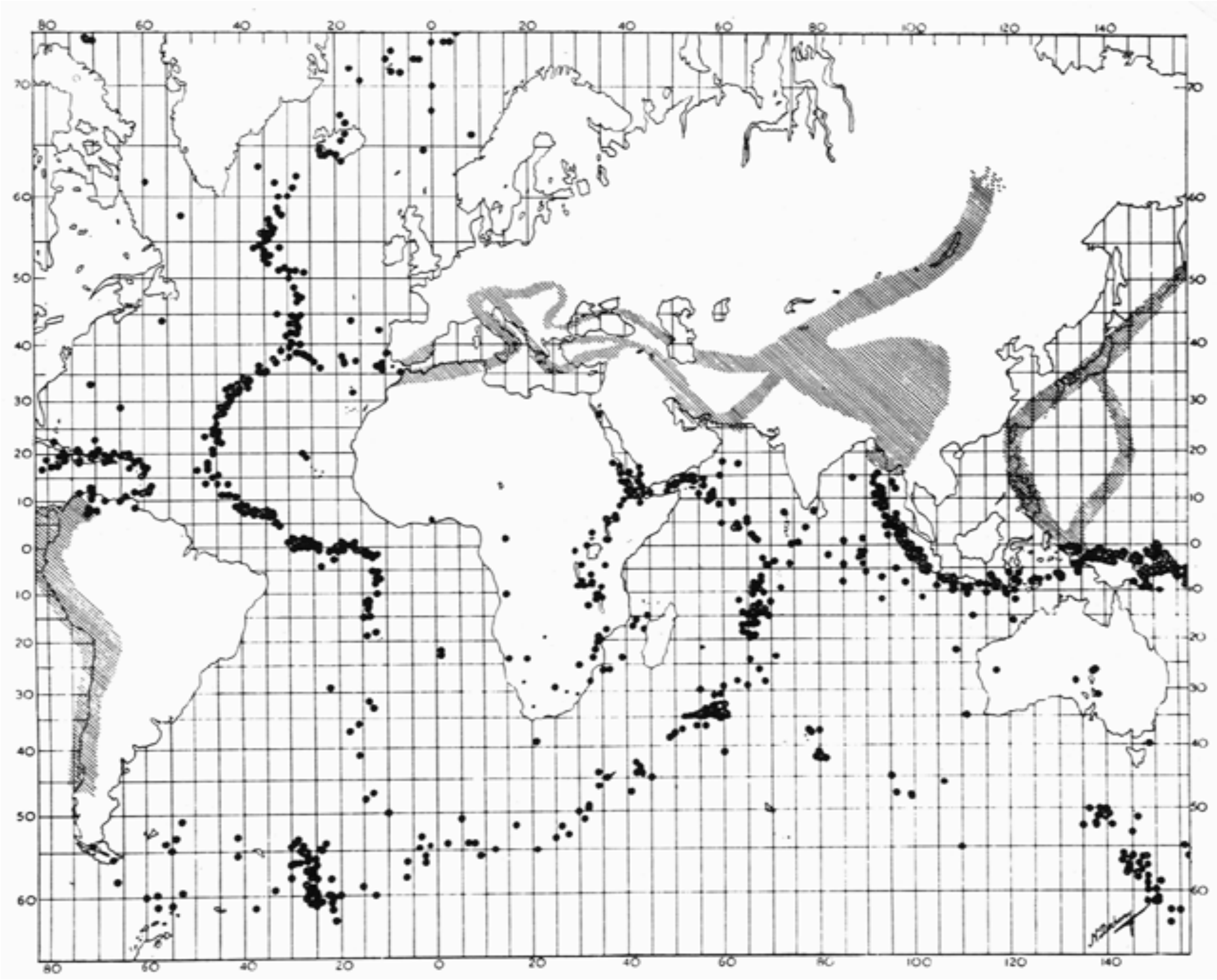
Lecture 4: Earthquakes

Read Chapters 2.1, 2.10.1-3 in KK&V

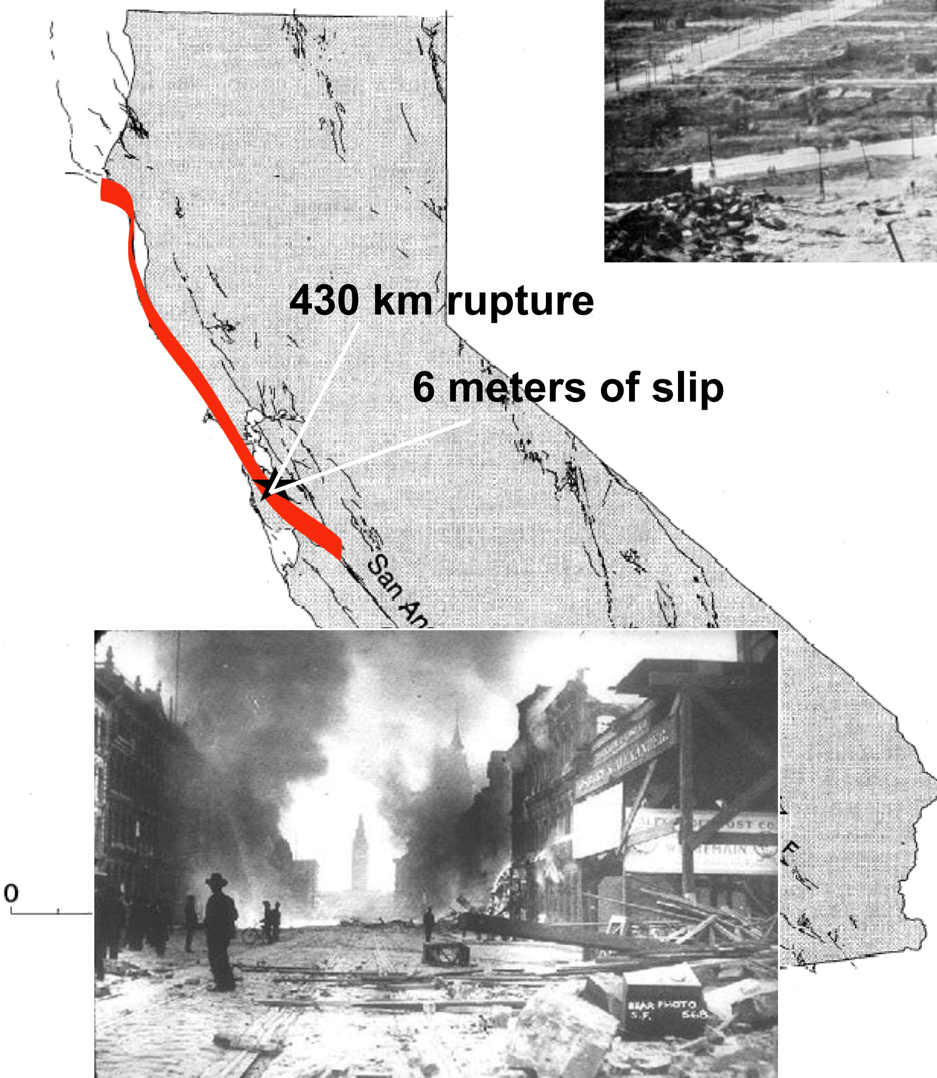
Homework 1! (due next Thursday)

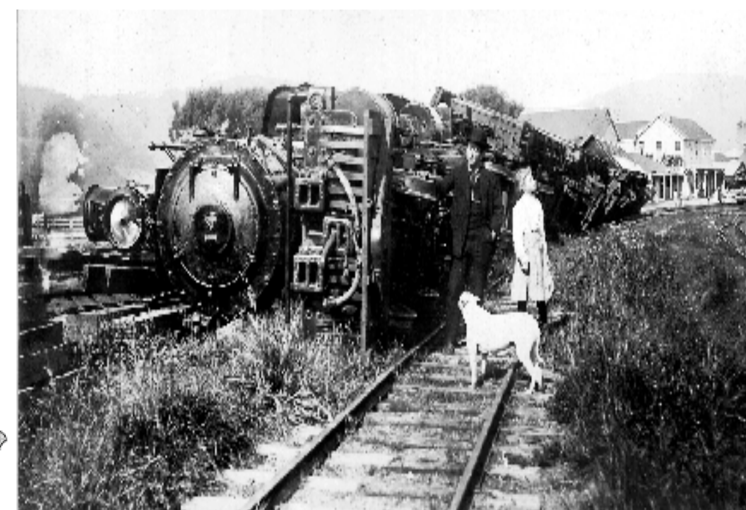
Global distribution of earthquakes (circa 1960)



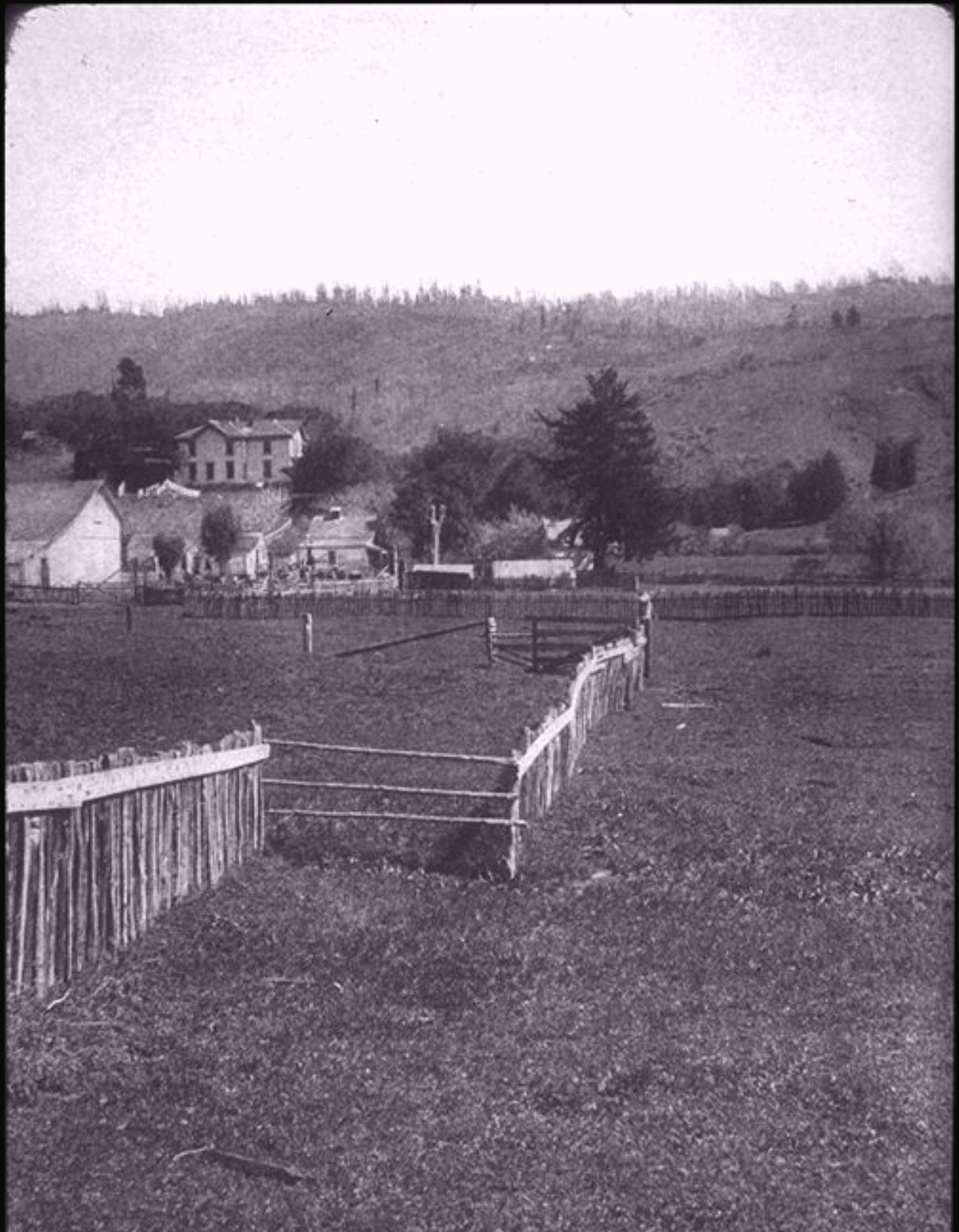
San Francisco Earthquake

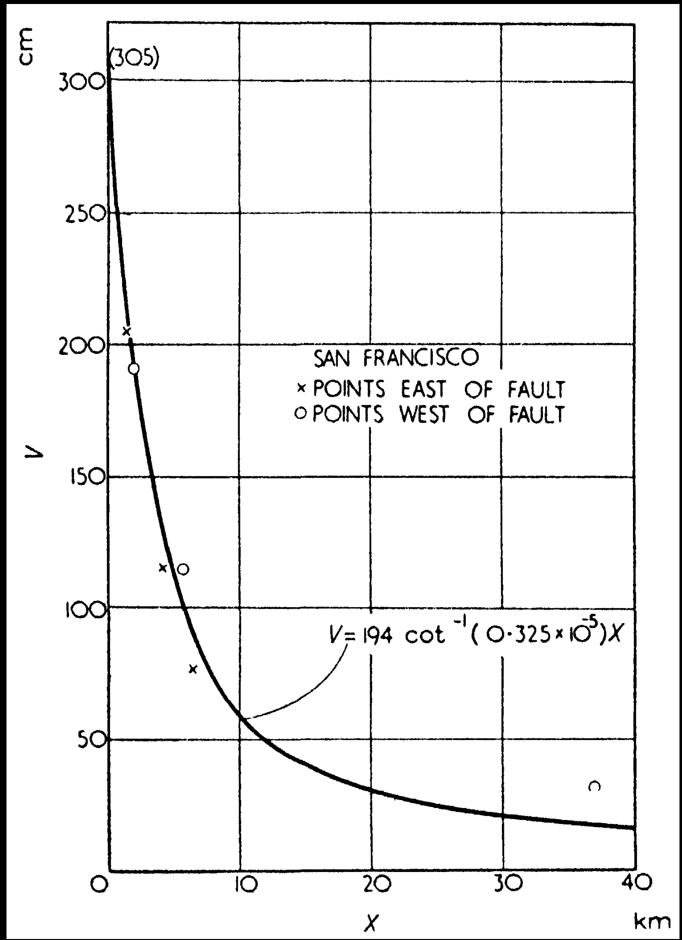
April 18, 1906 5:12 am
Magnitude 8.3





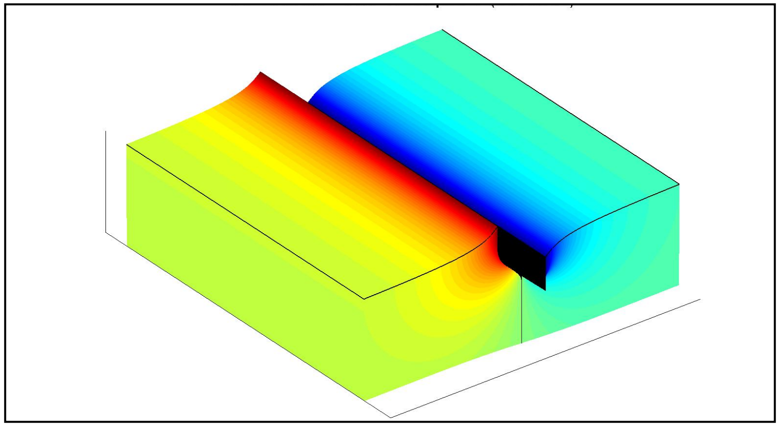
Reid, 1910; Byerly & DeNoyer, 1958; Knopoff, 1958; Chinnery, 1964





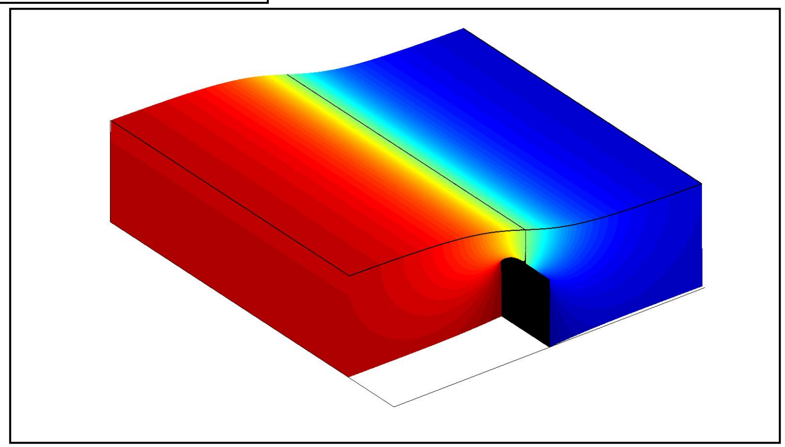
cm
(305)
300
250
200
150
100
50
0
10
20
30
40
V
X
km
SAN FRANCISCO
× POINTS EAST OF FAULT
○ POINTS WEST OF FAULT
V= 194 cot⁻¹(0·325×10⁻⁵)X

The Earthquake Cycle

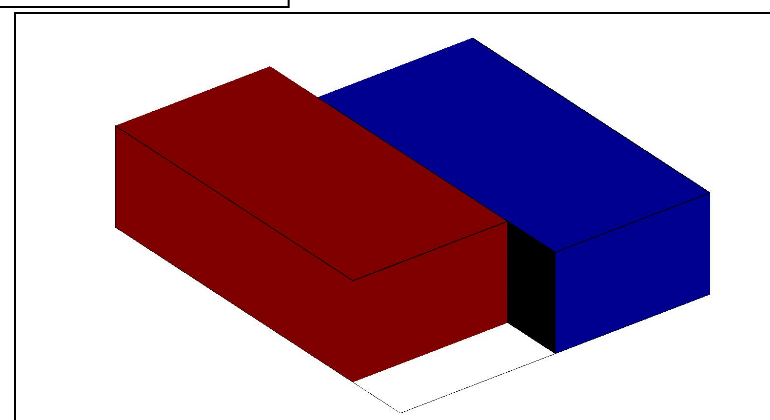


Coseismic
(seconds)

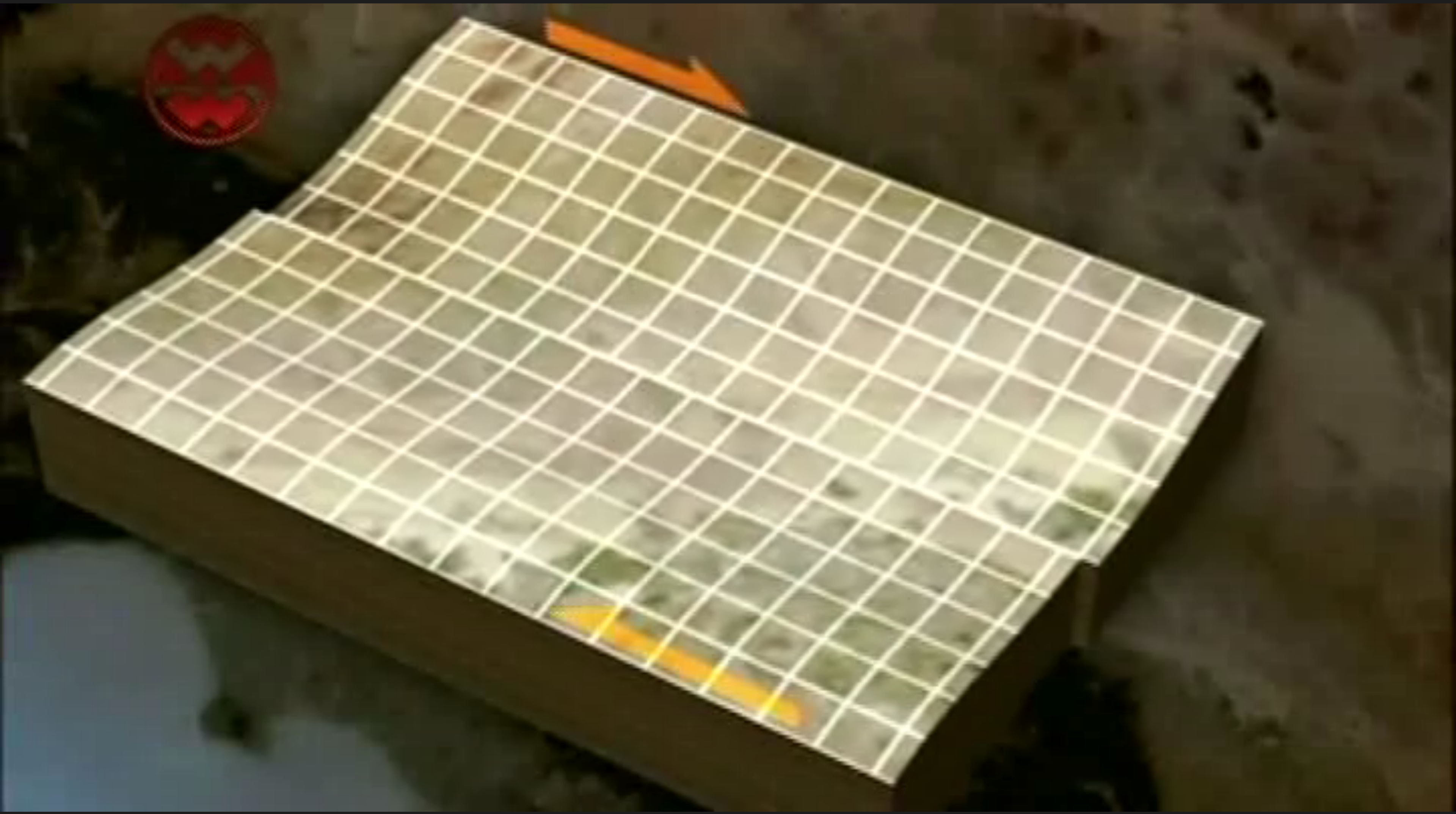
Interseismic
(10^2-10^3 yrs)



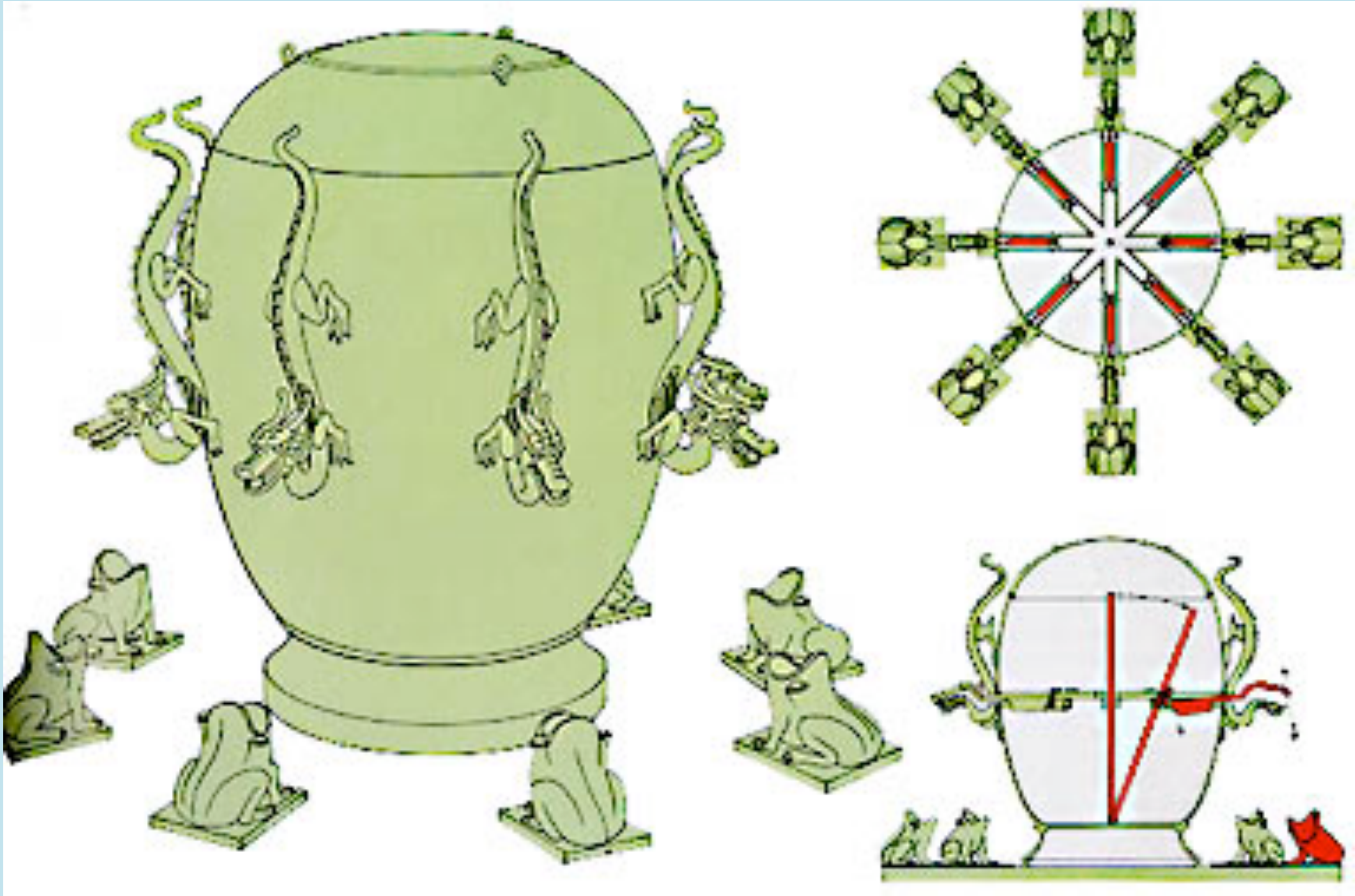
Geologic - Multiple Earthquakes
(10^3-10^6 yrs)



The Earthquake Cycle



Measuring earthquakes: the old way (China, 132 AD)



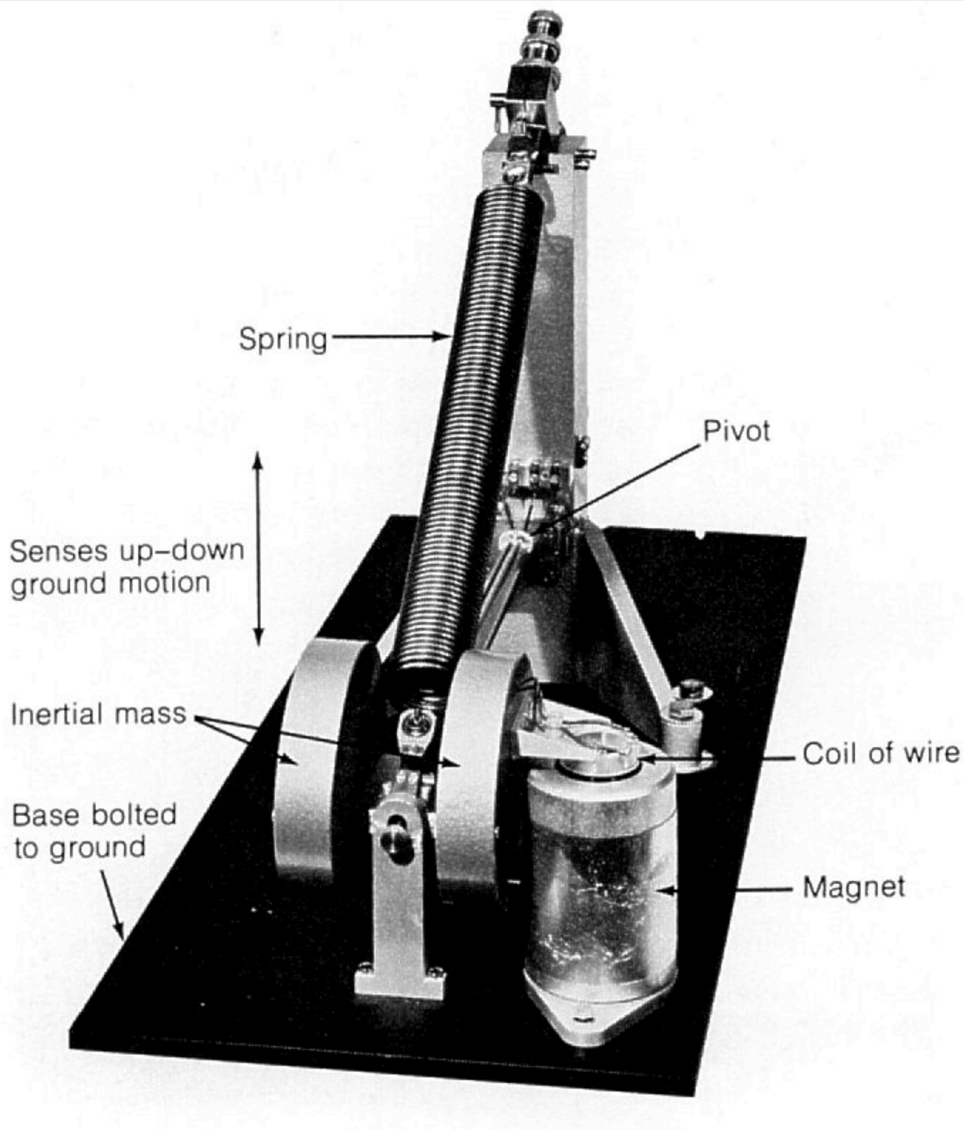
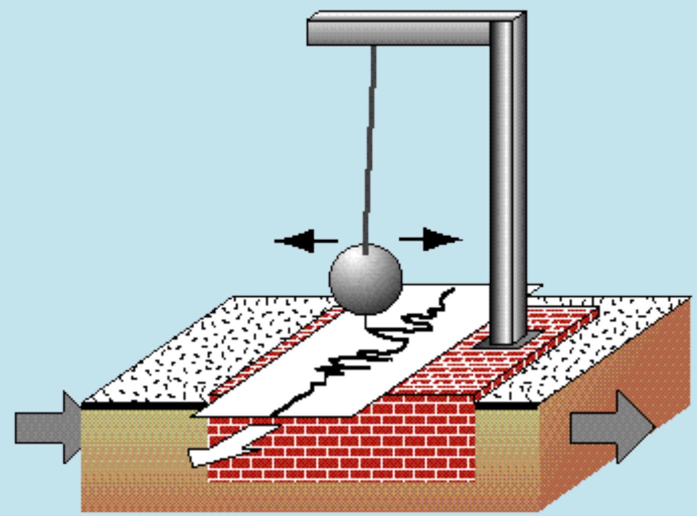
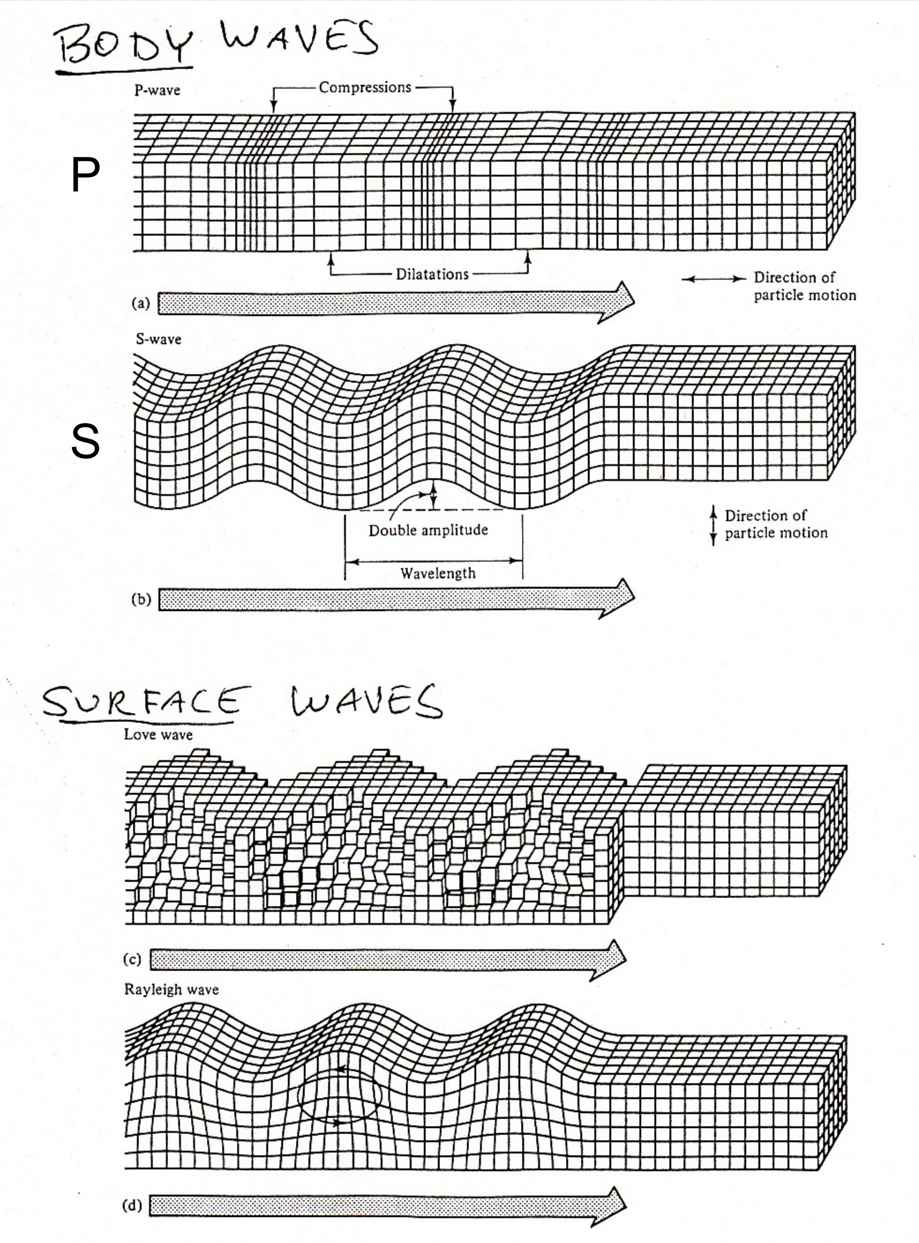


Figure 18-3

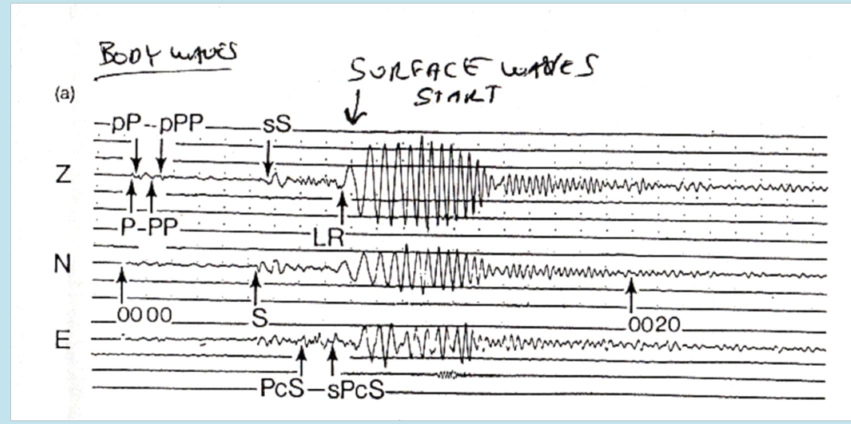
The Press-Ewing seismograph, an example of a modern instrument. Airtight cover, electronics, and recording system not shown.

Fundamentally, a seismograph is a simple pendulum. When the ground shakes, the base and frame of the instrument move with it, but inertia keeps the pendulum bob in place. It will then appear to move, relative to the shaking ground.





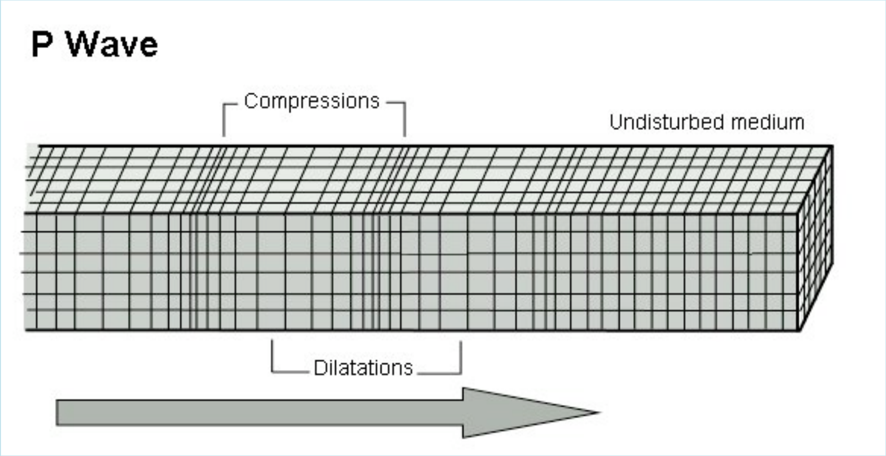
Two basic types of earthquake waves: body waves and surface waves:

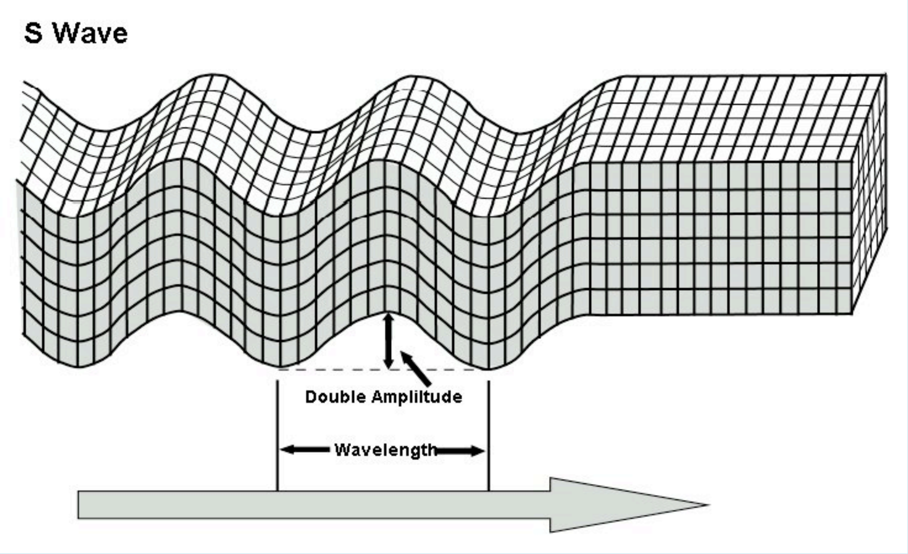


Body waves (P and S waves) travel through the deep interior of the Earth while surface waves do not.

Body waves arrive before surface waves (shorter path), but surface waves have larger amplitude

Body waves



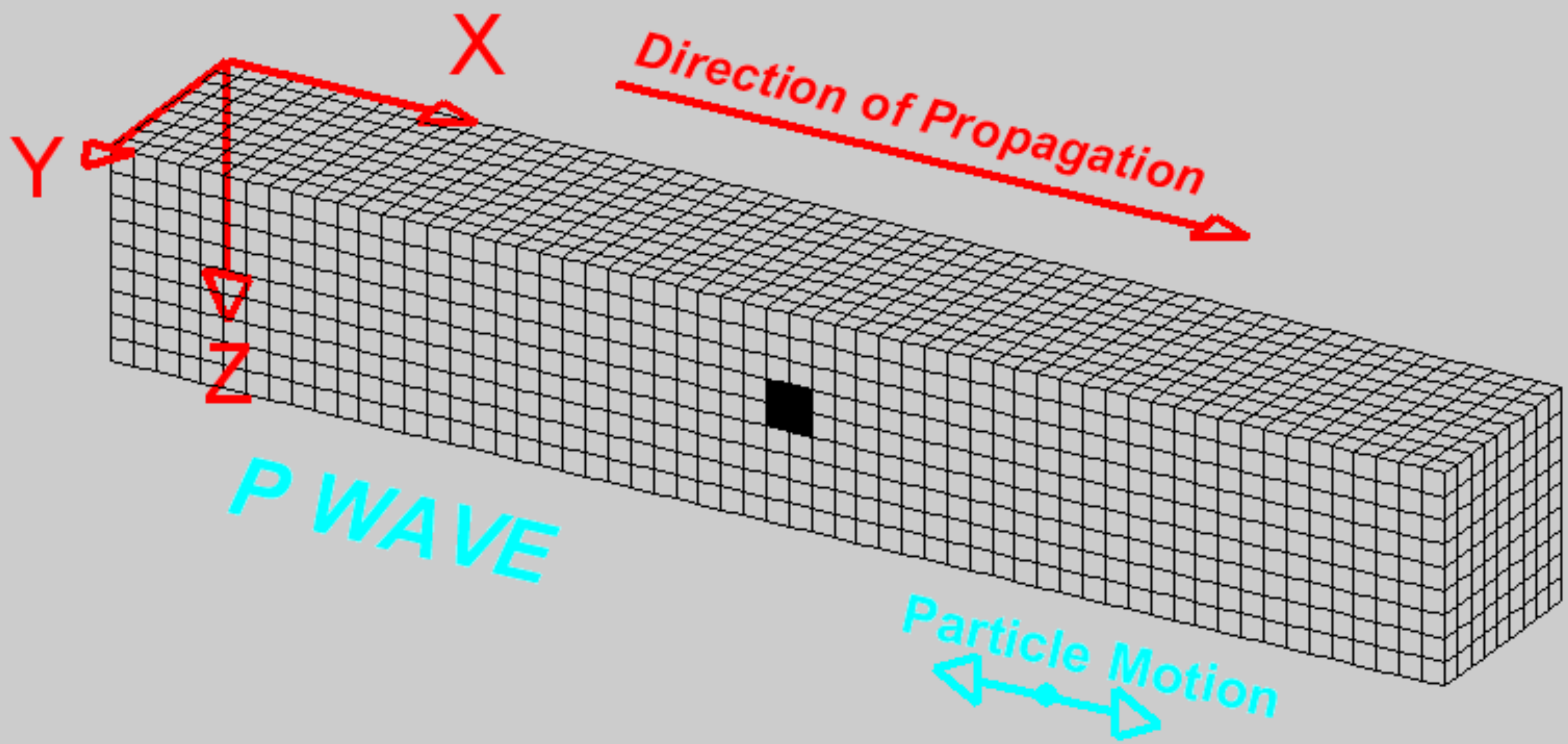


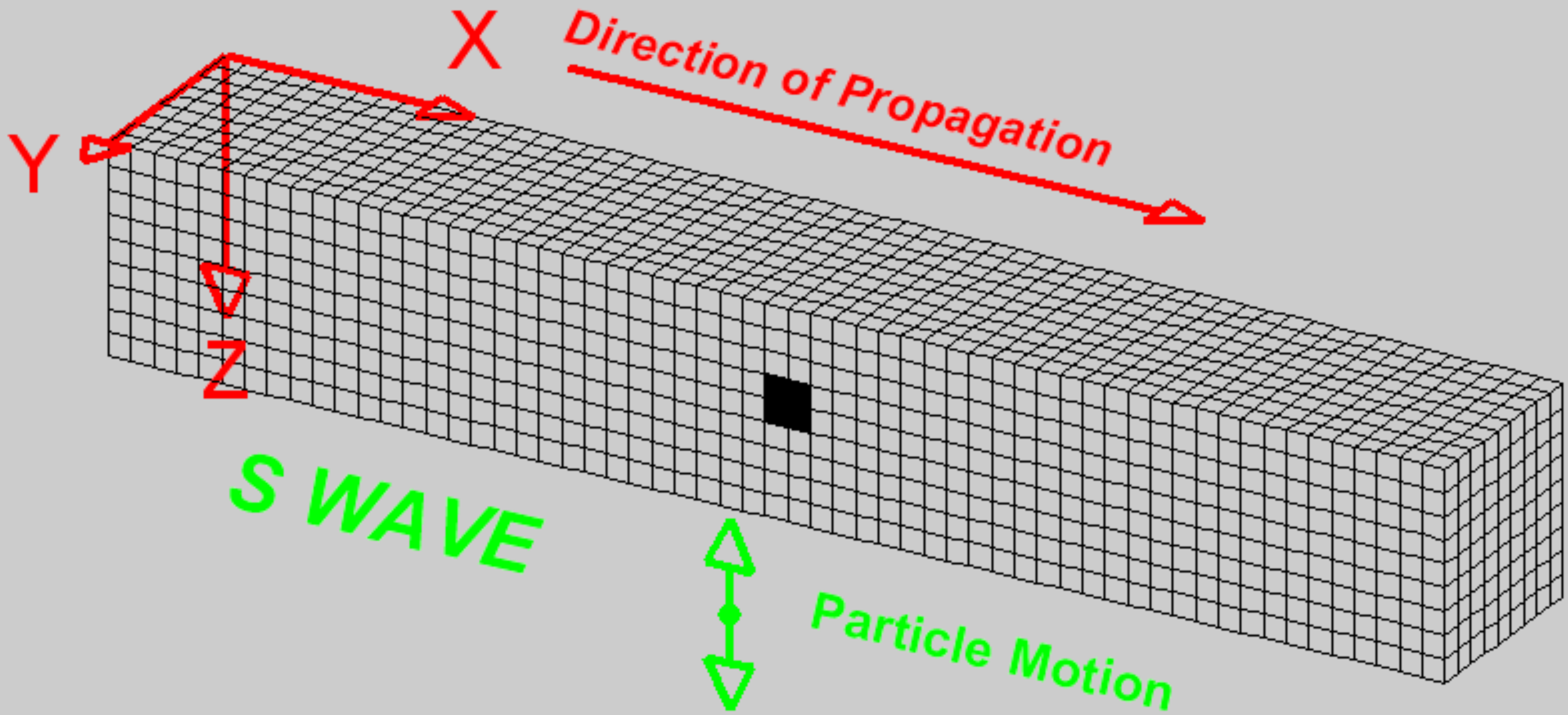
P waves (primary or pressure or push-pull). Involves compression or dilatation (rarefaction) of the material as it passes. Analogous to sound waves in the air.

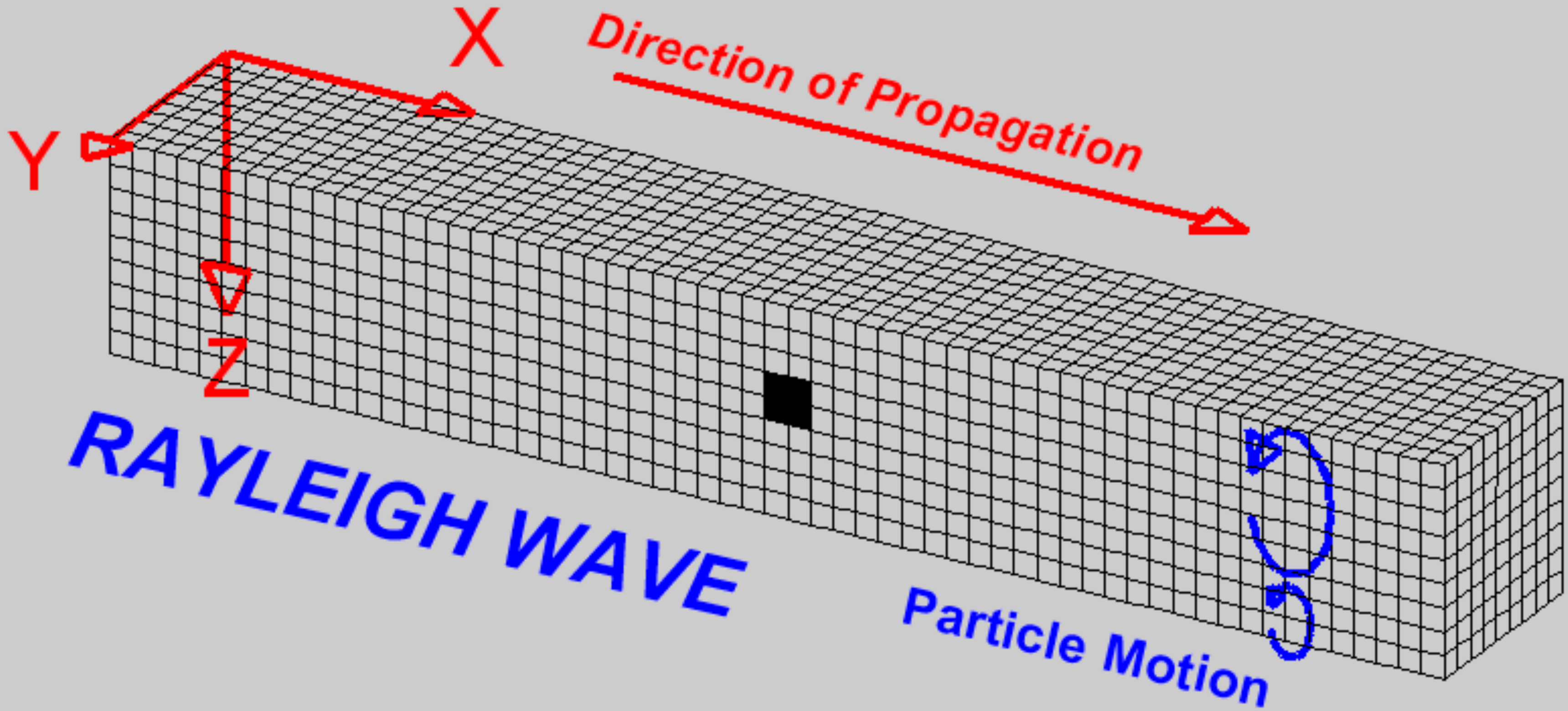
S waves (secondary or shear or shake). Involves shearing and rotation of the material but no change in volume.

S waves cannot propagate through a liquid (e.g. outer core)

Let's do a demonstration…

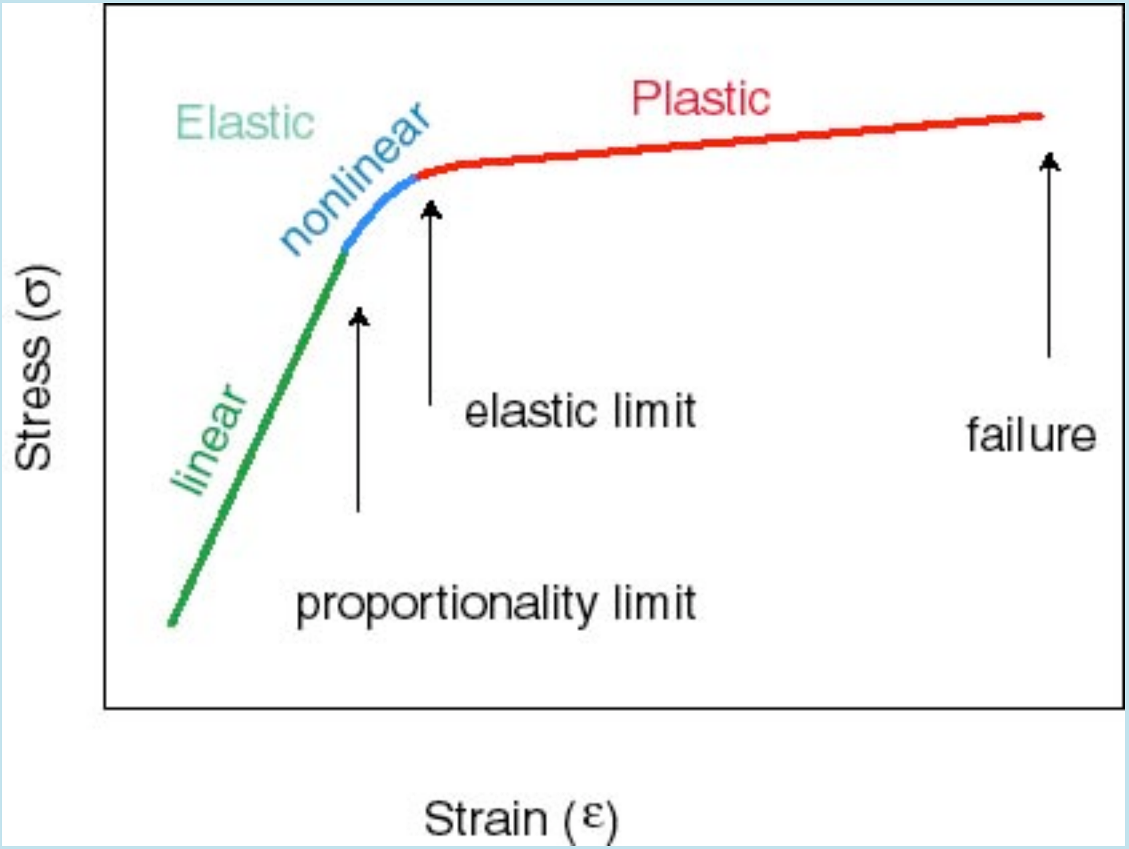

X
Y
Z
Direction of Propagation
P WAVE
Particle Motion


X
Y
Z
Direction of Propagation
S WAVE
Particle Motion



X
Direction of Propagation
Y
Z
RAYLEIGH WAVE
Particle Motion

Stress-strain relationship



Dealing with elastic behavior of materials

Stress is proportional to strain and there is no permanent deformation

$$V_P = \sqrt{\frac{K + \frac{4}{3}\mu}{\rho}}$$

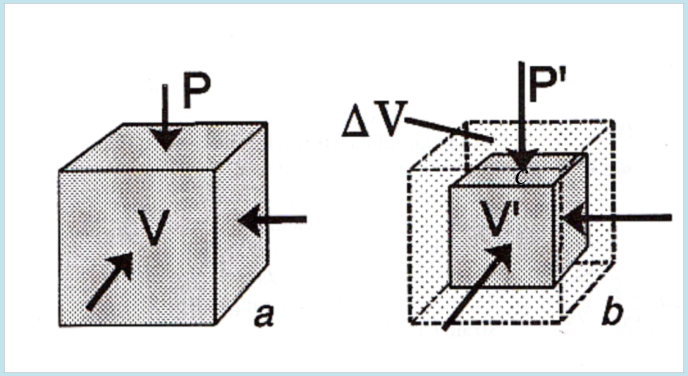
$$V_S = \sqrt{\frac{\mu}{\rho}}$$

ρ – density

Velocities are a function of bulk modulus, shear modulus and density

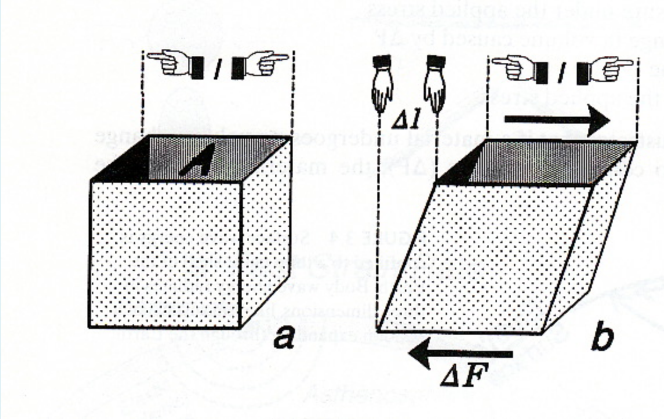
S waves involve no change in volume; Vs is only a function of μ, which is zero in a liquid

Bulk modulus K
incompressibility
(change in volume)



$$k = \frac{\text{stress}}{\text{strain}} = \frac{\Delta P}{\Delta V/V}$$

Shear modulus μ
rigidity



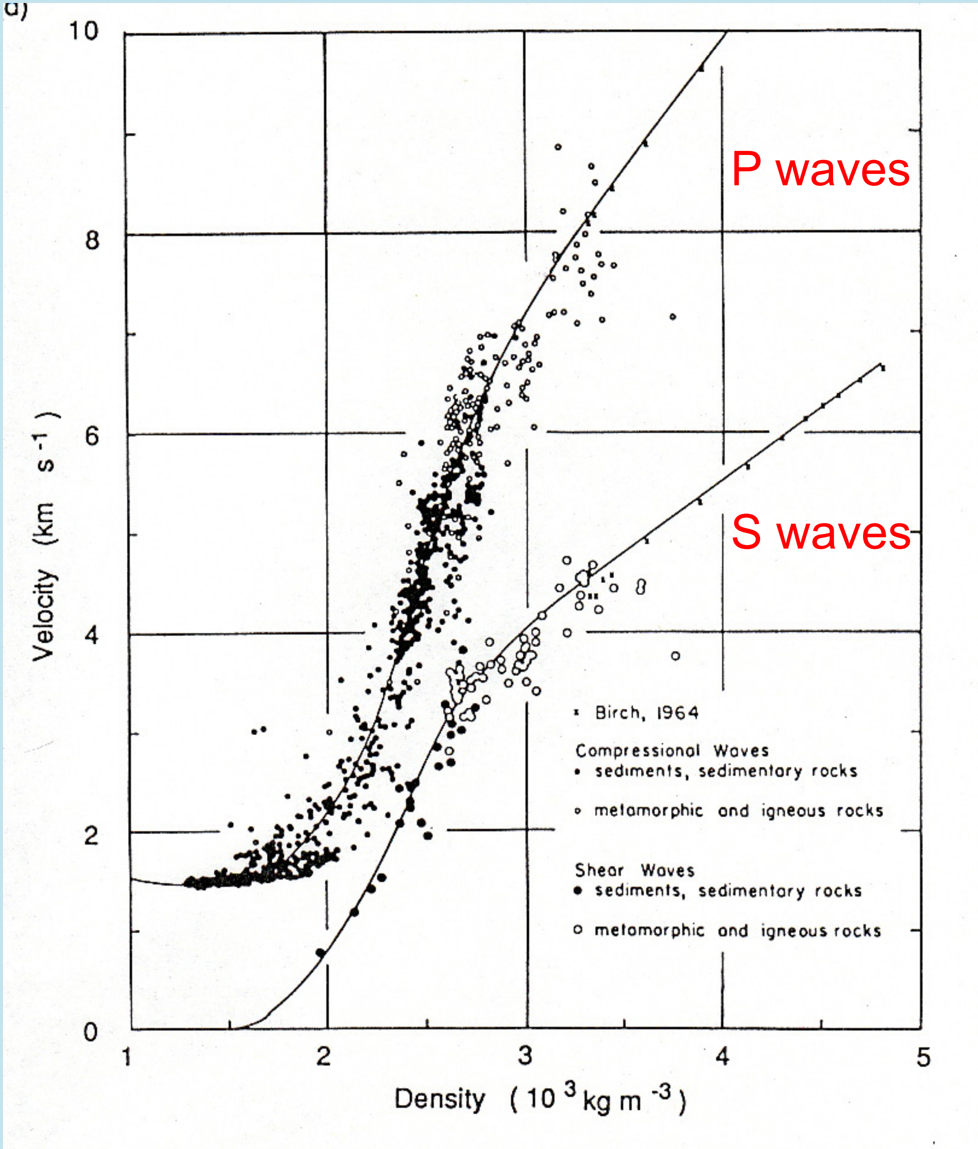
$$\mu = \frac{\text{stress}}{\text{strain}} = \frac{\Delta F/A}{\Delta l/l}$$

“Nafe-Drake” curves: empirical relationship between density and velocity

Denser rocks have higher velocity

$V = a\rho + b$ Birch’s law

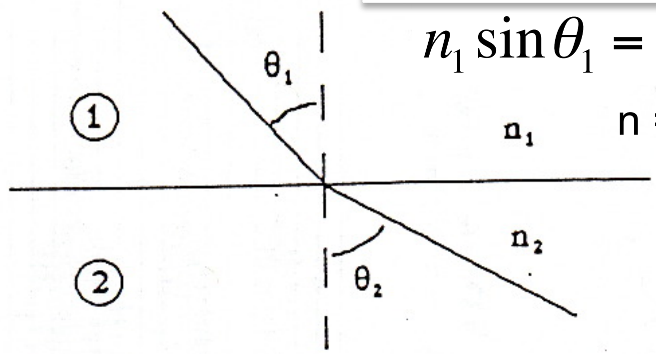
where *a* and *b* are constants determined empirically



Snell's law

$$n_1 \sin\theta_1 = n_2 \sin\theta_2$$

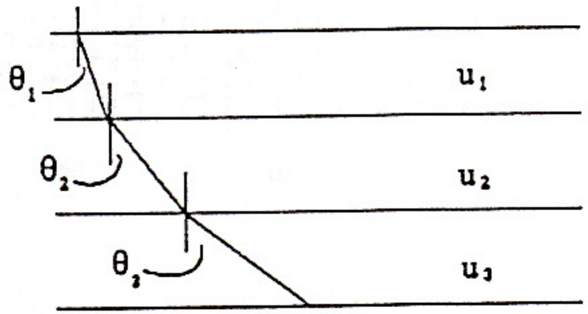
n = 1/v

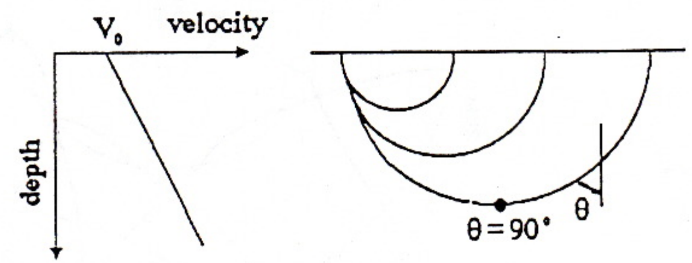


Body waves follow the laws of refraction …

multiple layers

u = 1/v

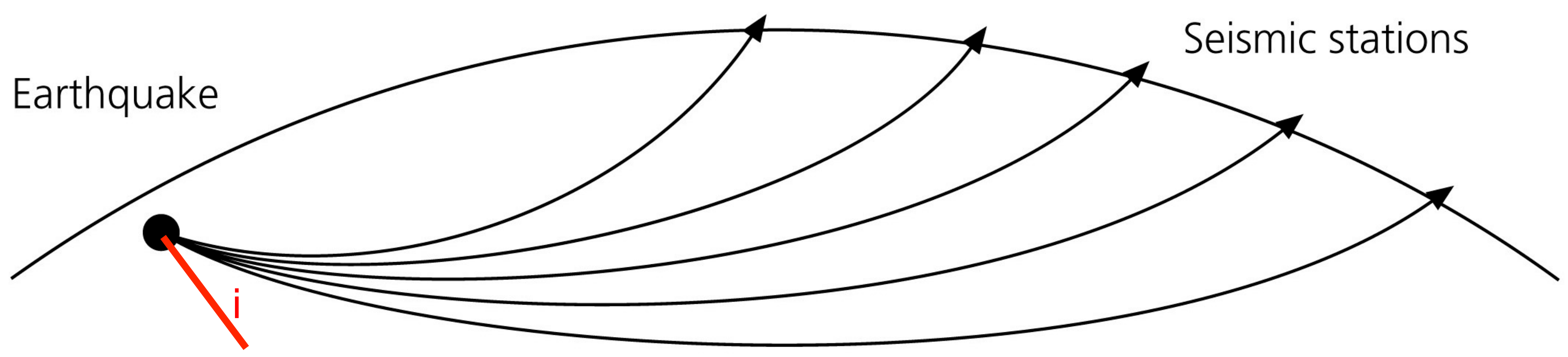




…. and in the Earth follow curved trajectories

SEISMIC RAY PATHS BEND AS VELOCITY INCREASES WITH DEPTH

Figure 1.1-2: Schematic ray paths for an increase in seismic velocity with depth.

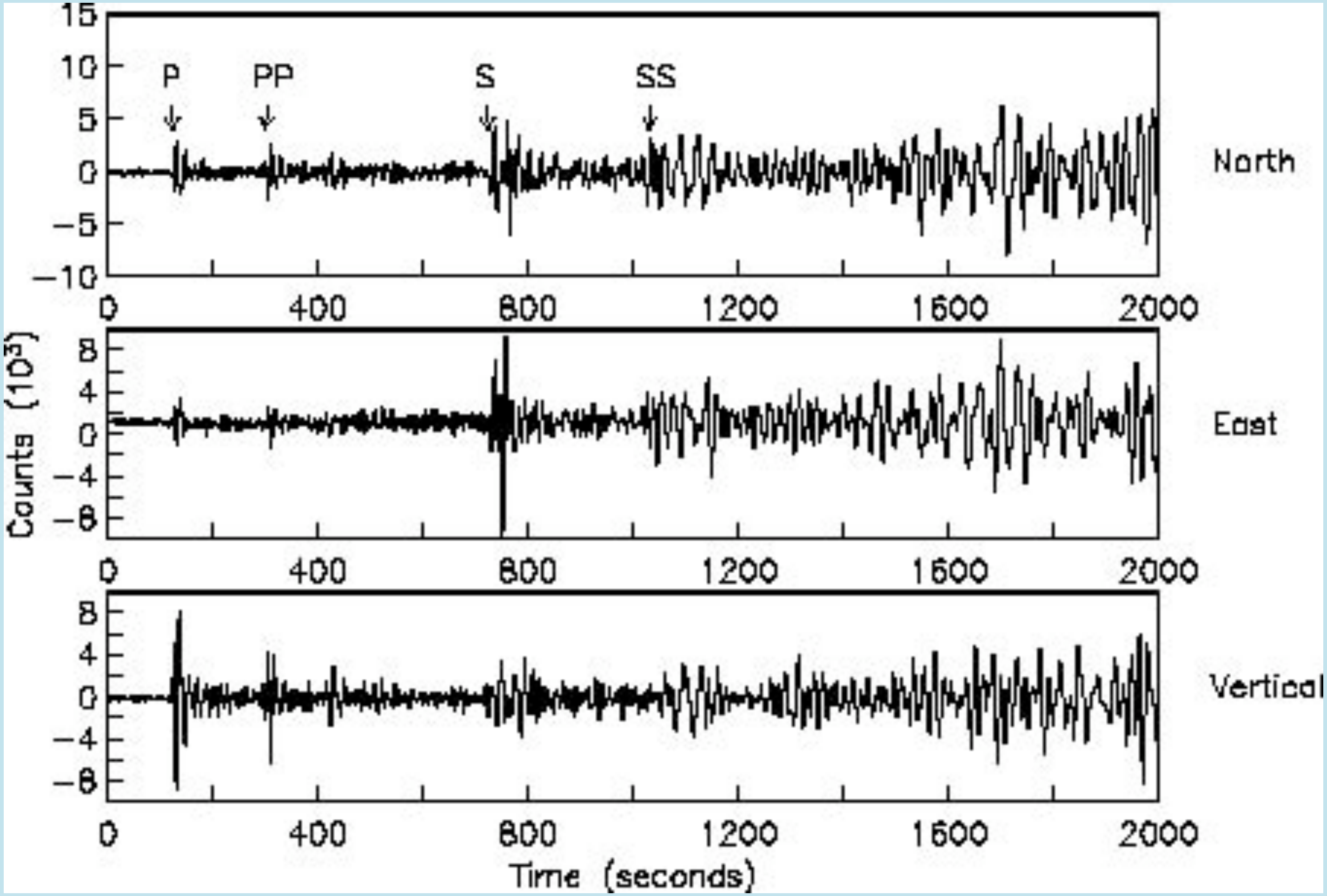


Can define ray paths in which the earthquake wave has a constant value of p

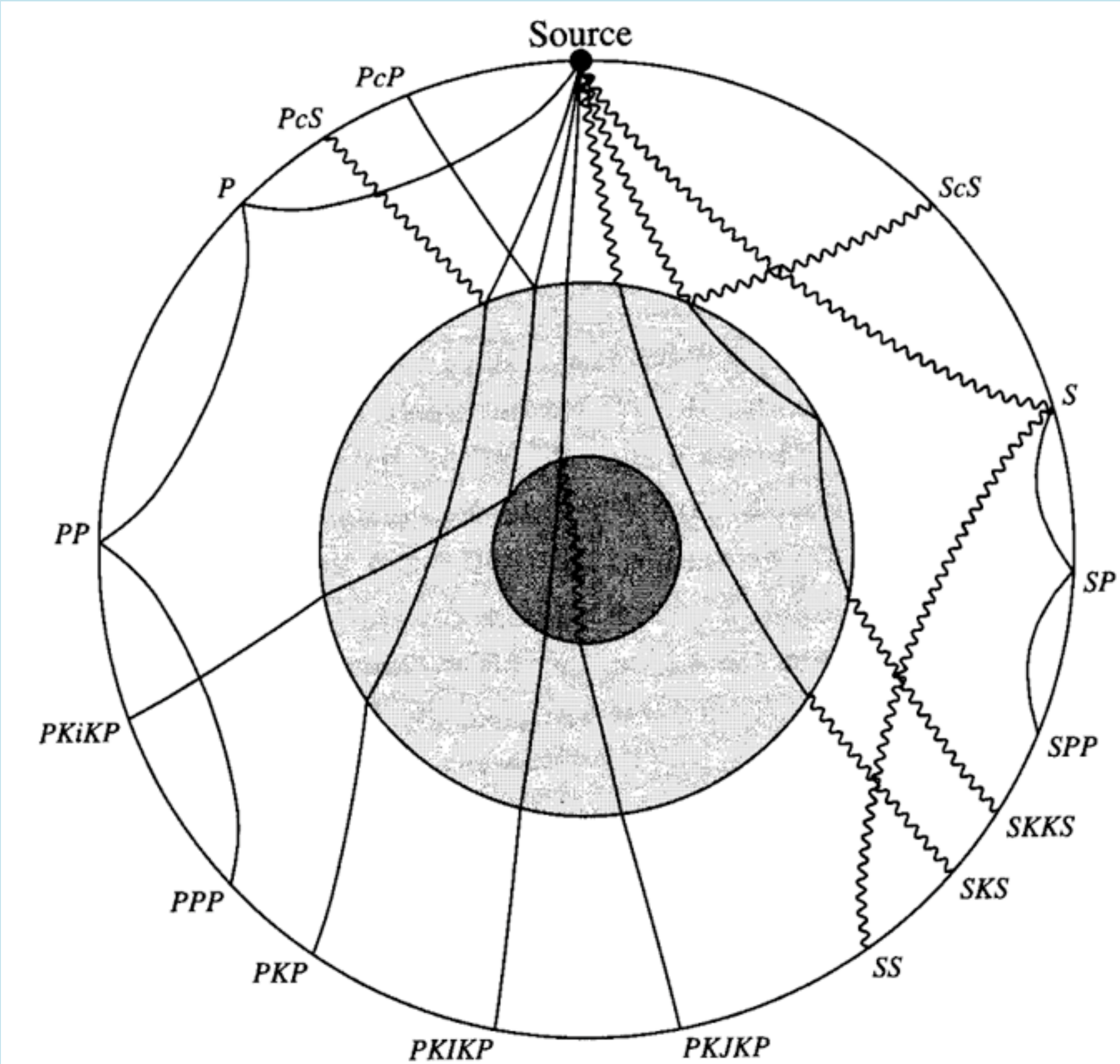
p is called the ray parameter

$$p = \frac{r \sin i}{\mathrm{v}}$$

Snell's law for spherical earth with velocity v at radius r

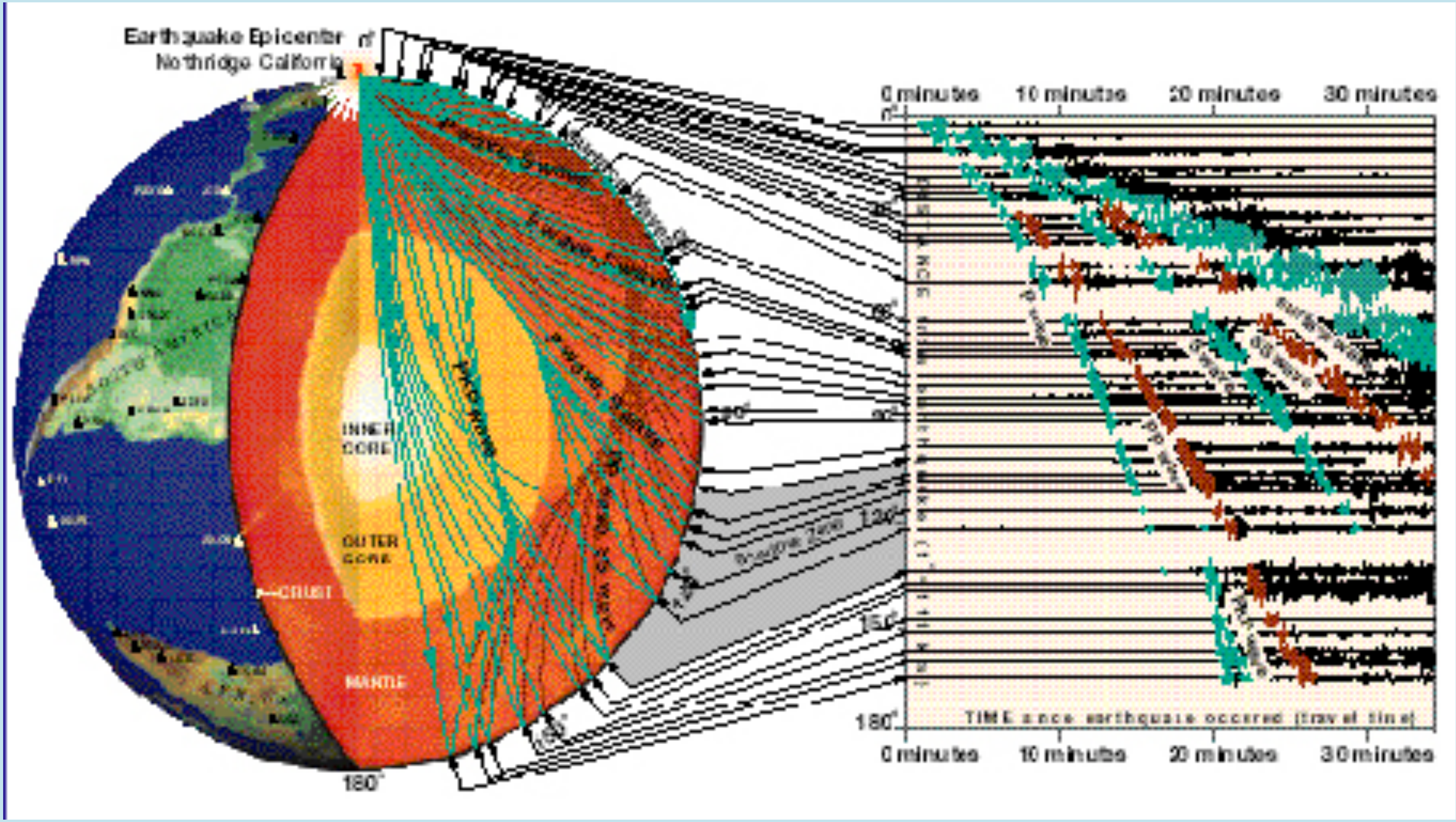


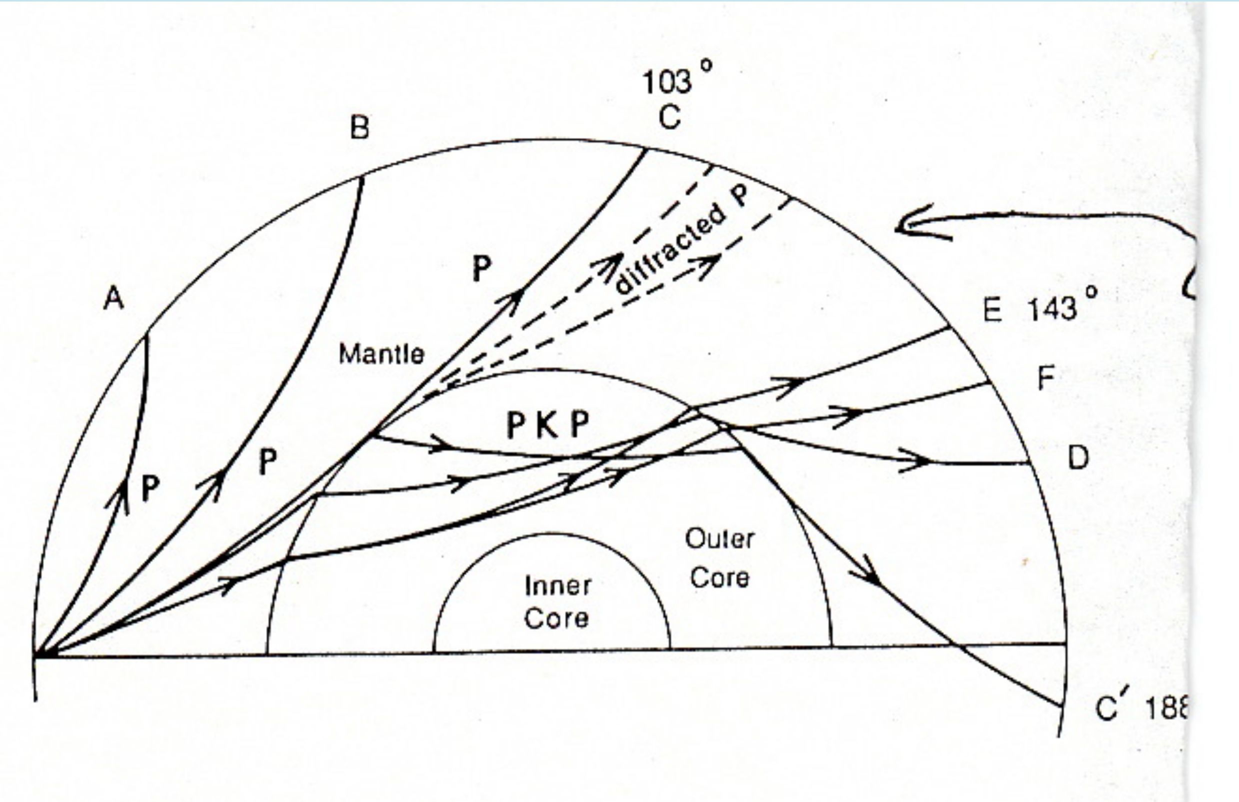
The seismogram measured at the receiver is analyzed and the arrival times of the P, S, and various other waves are measured.



P: P-wave in the mantle,
K: P-wave in the outer core,
I: P-wave in the inner core,
S: S-wave in the mantle,
J: S-wave in the inner core,
c: reflection off the CMB,
i: reflection off the ICB.

Travel times





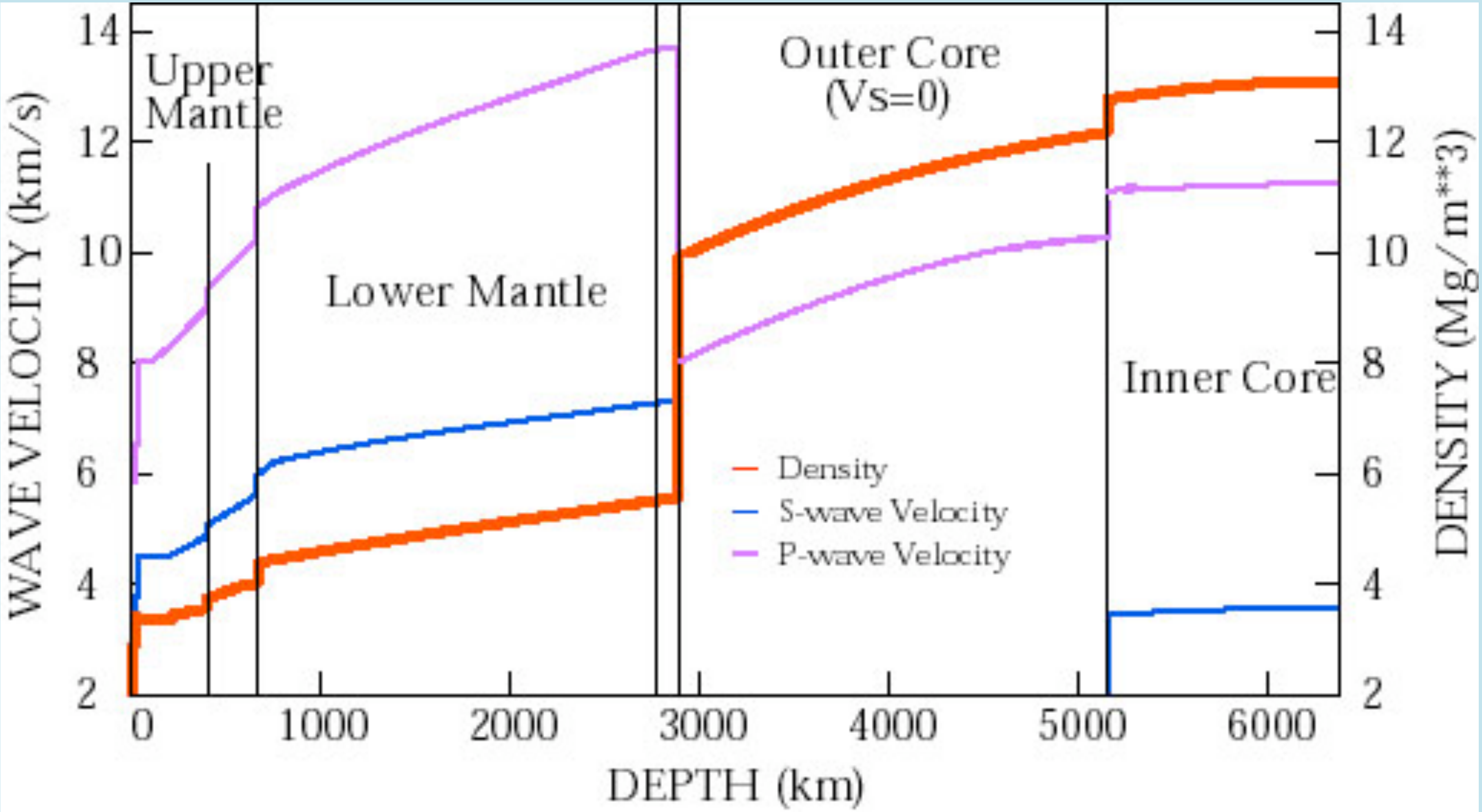
Note shadow zone due to refraction of P-wave energy at core-mantle boundary

Indicates a sharp decrease in velocity of P waves in liquid outer core

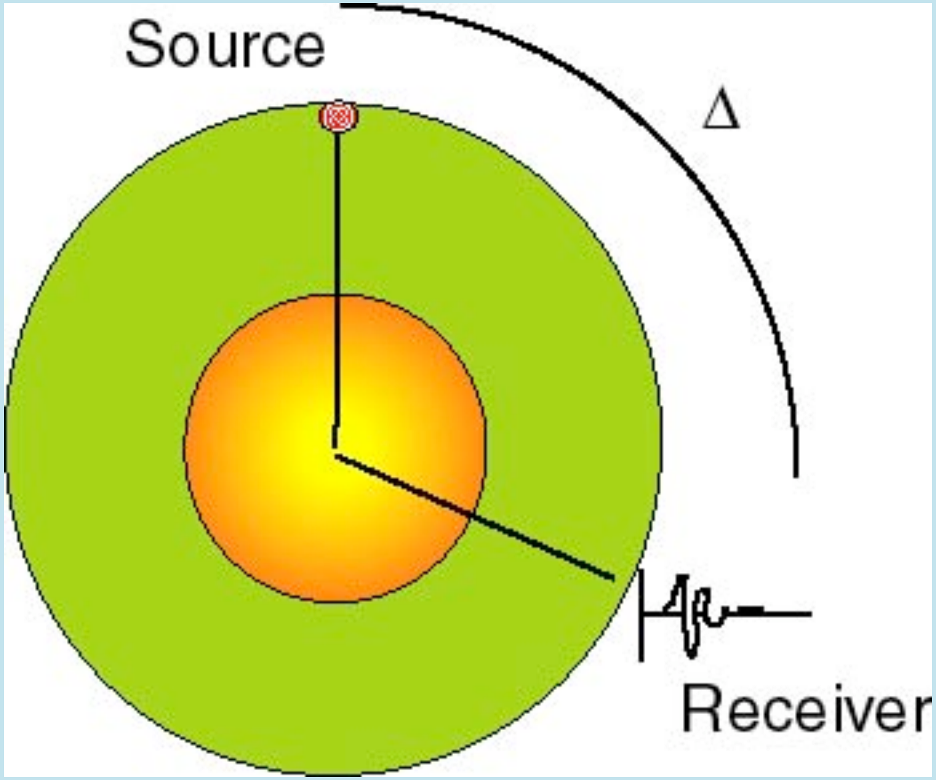
1-D Model

Outer core:
- density goes up (Fe),
- P wave vel. drops;
- No S waves = liquid

Inner core:
- S waves = solid

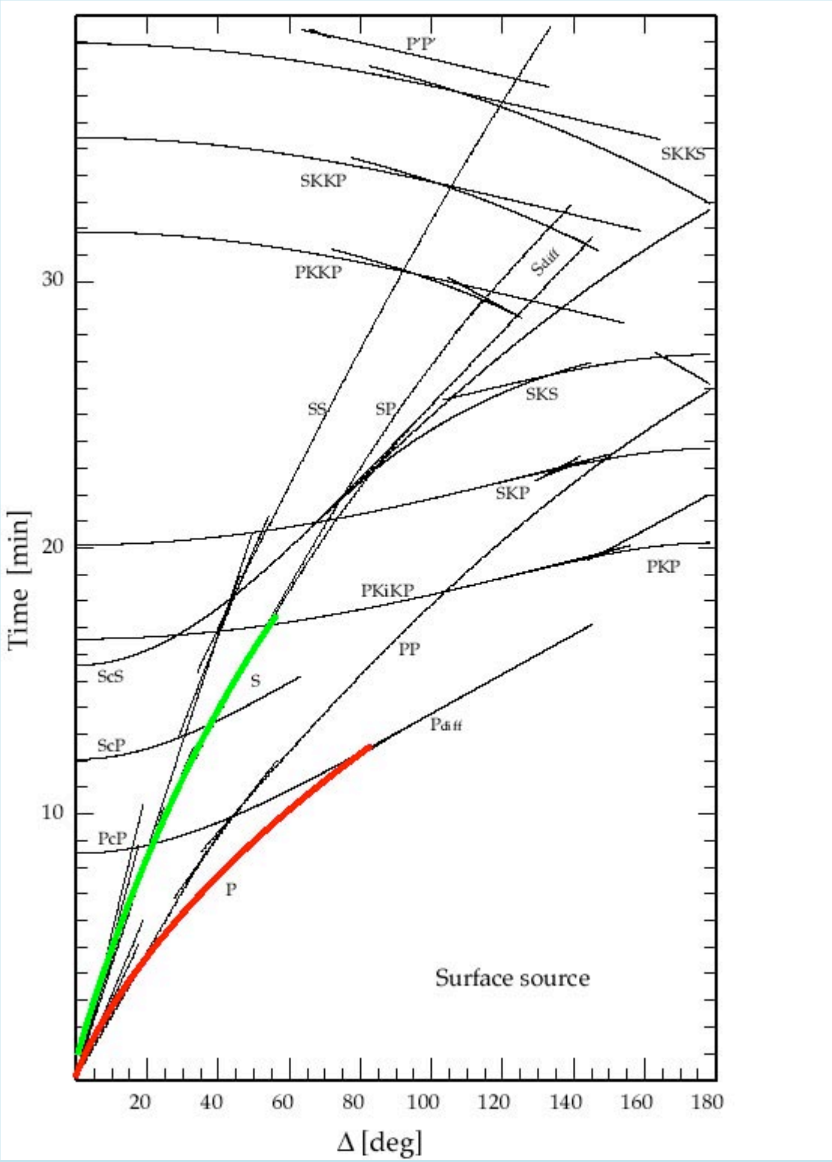


Travel time plots:
Delta



International Association of Seismology and Physics of the Earth's Interior

IASPEI



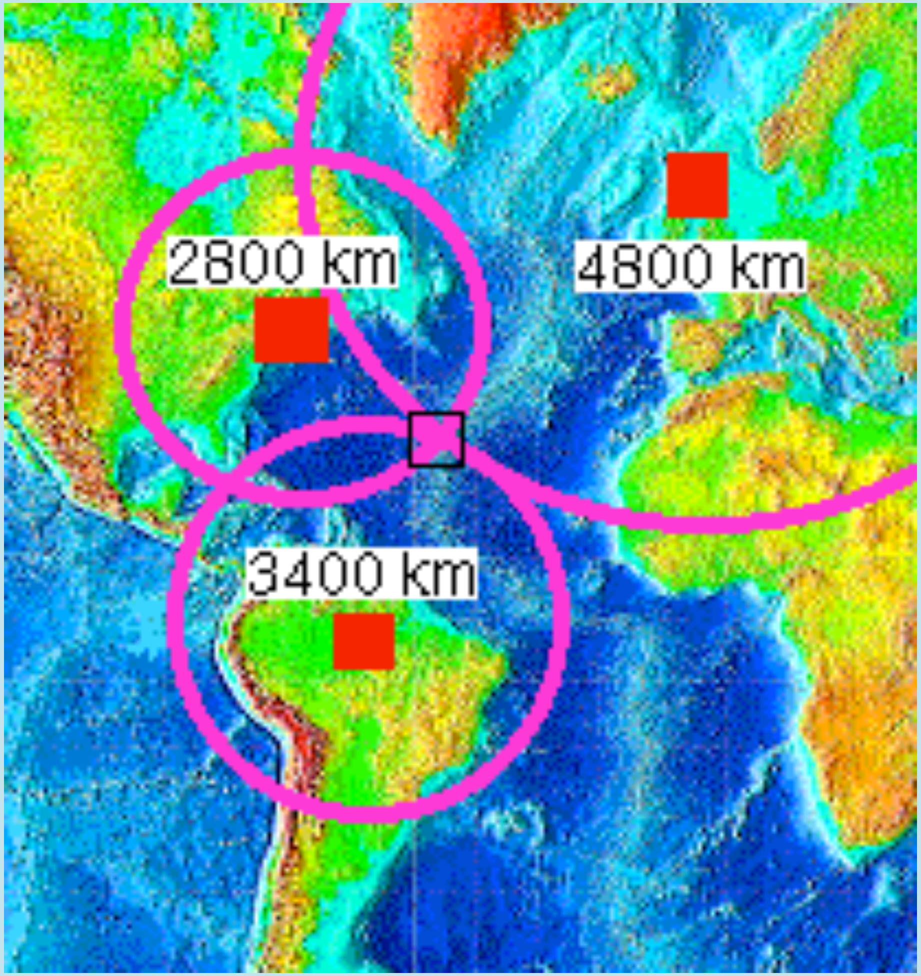
Standard travel time curves for earthquakes with a focus at the surface; accurate to within a few seconds

Structure of the Earth …

Use this information to locate earthquakes even if don't know when earthquake occurred

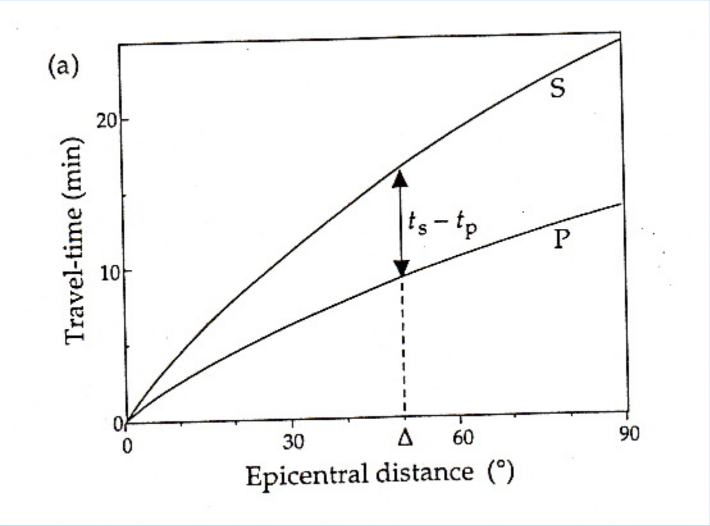
How?
Note P and S

Earthquake locations

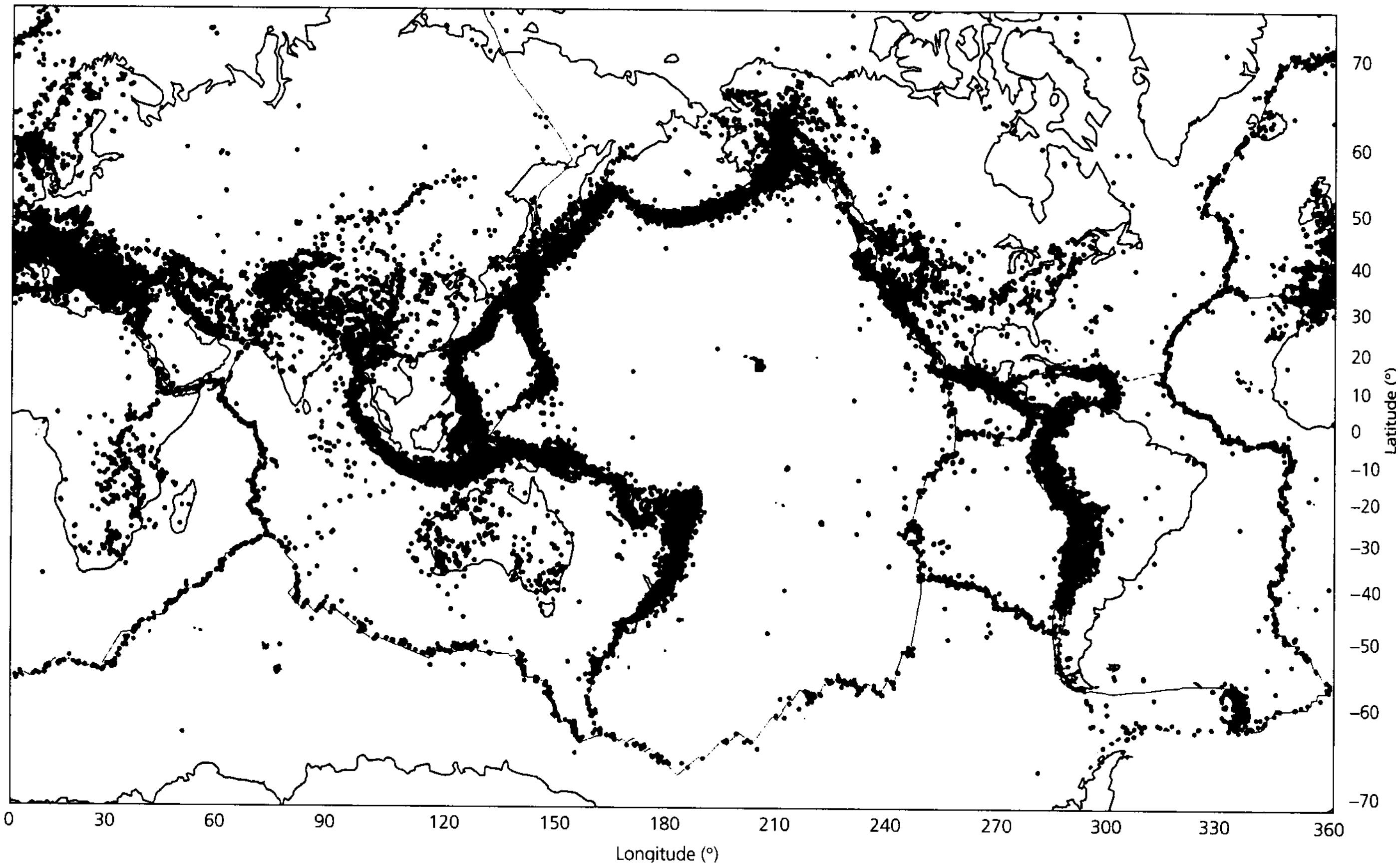


Triangulate

The difference in time between the P and S waves can be used to determine the distance to the epicenter

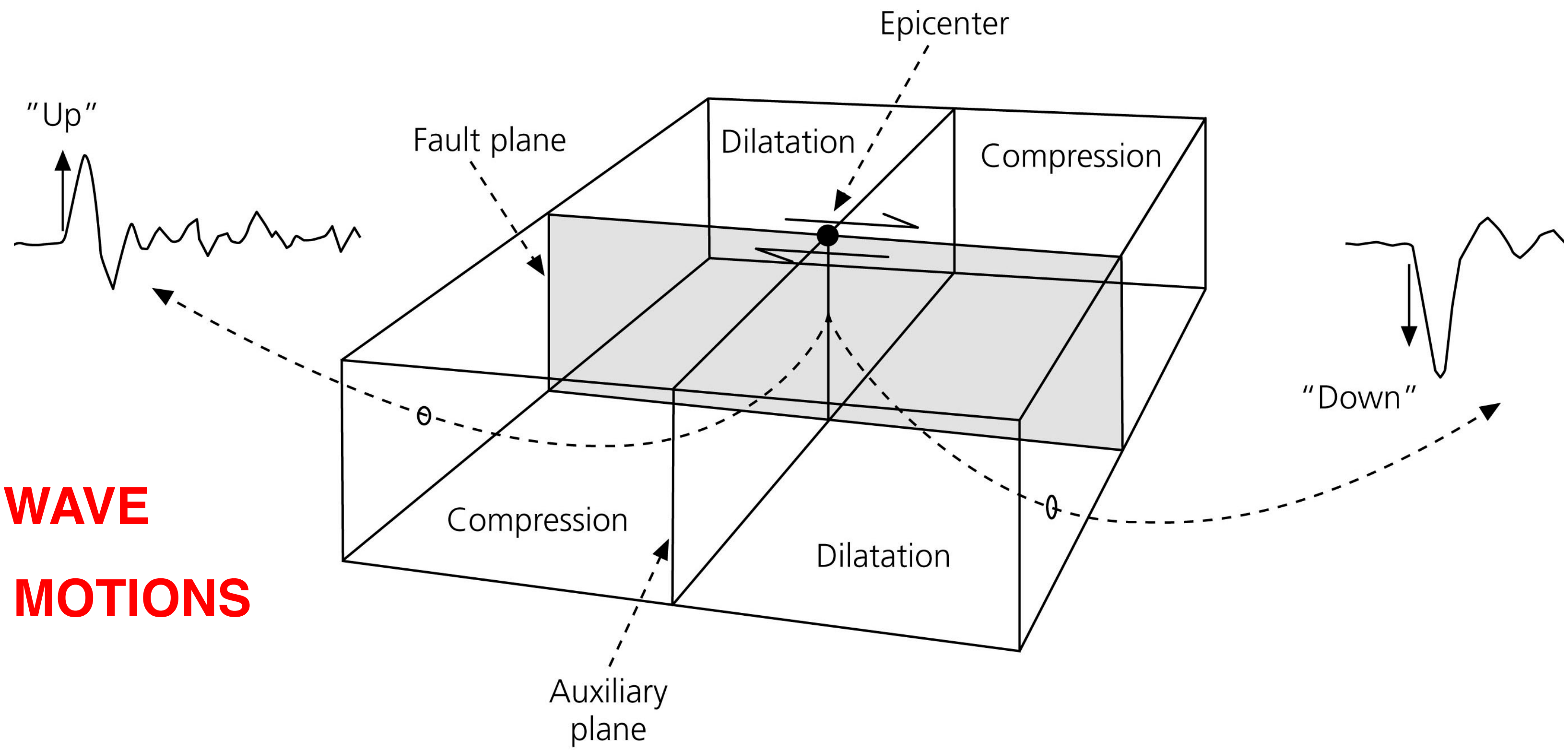


…much sharper picture of plate boundaries!



P WAVE

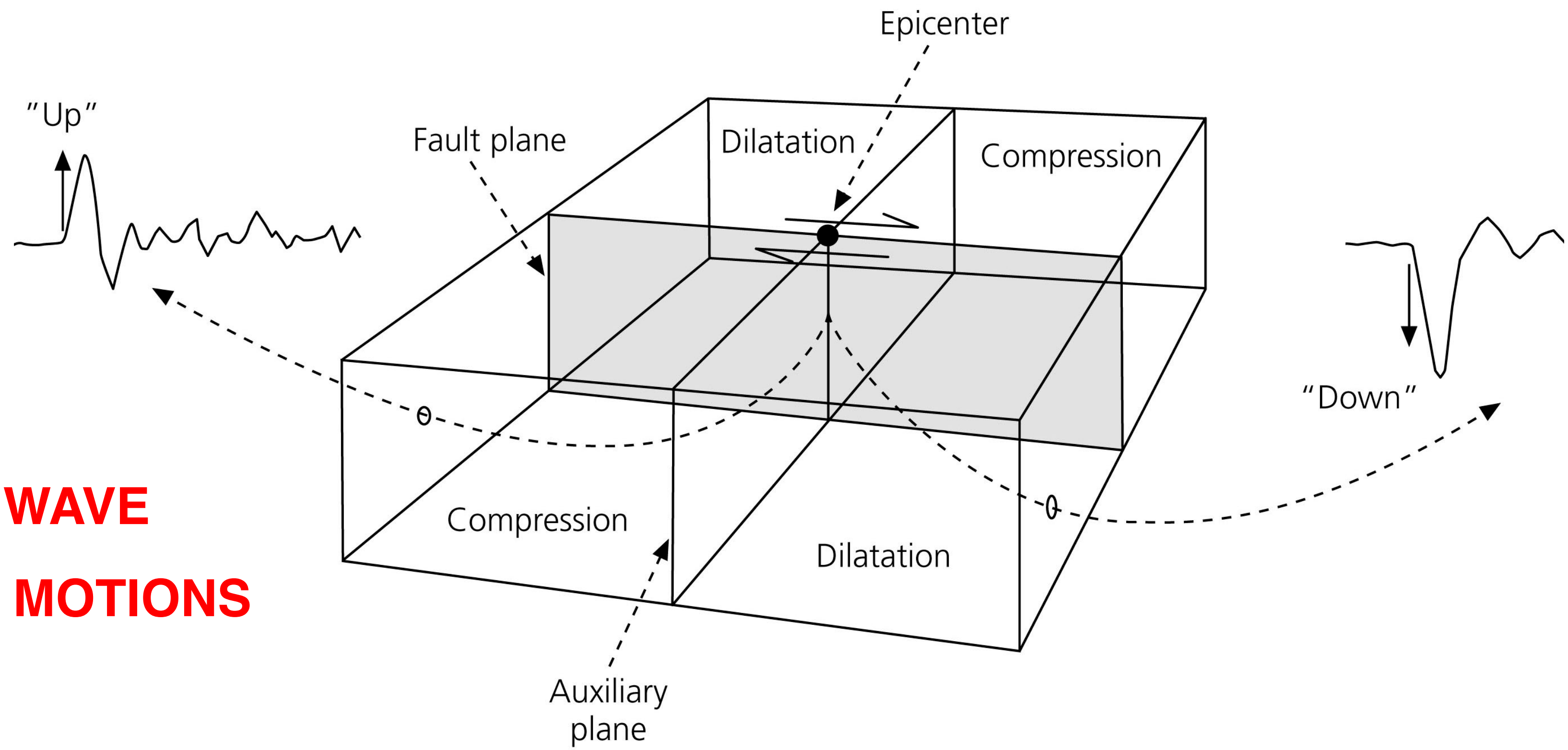
FIRST MOTIONS



Polarity of first P-wave arrival varies between seismic stations in different directions.

First motion is compression for stations located such that material near the fault moves ``toward'' the station, or dilatation, where motion is ``away from'' the station.

When a P wave arrives at a seismometer from below, a vertical component seismogram records up or down first motion, corresponding to either compression or dilatation.



P WAVE FIRST MOTIONS

First motions define four quadrants; two compressional and two dilatational.

Quadrants separated by nodal planes: the fault plane and auxiliary plane perpendicular to it.

From the nodal planes fault geometry is known.

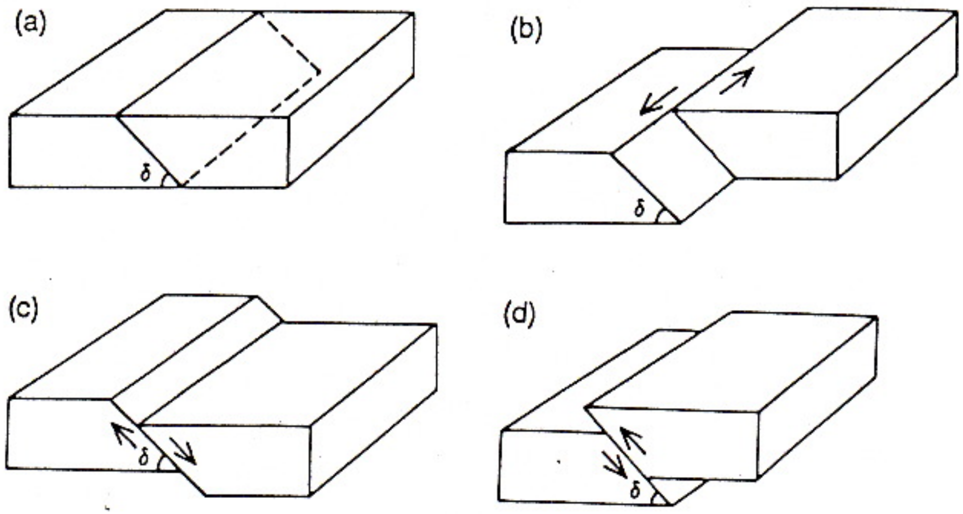
Because motions from slip on the actual fault plane and from slip on the auxiliary plane would be the same, first motions alone cannot resolve which is the actual fault plane.

Fundamental types of faulting

Transcurrent fault
Strike-slip fault

Lateral fault

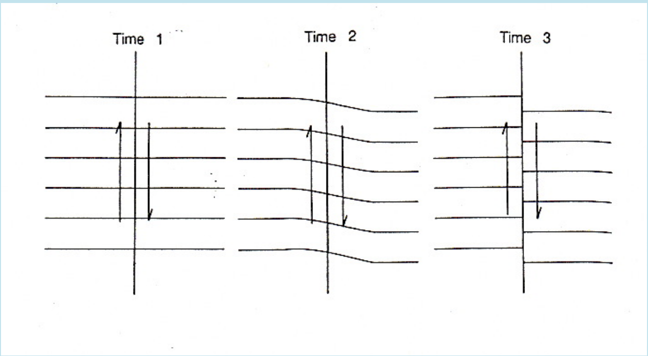
δ = dip



Dip-slip fault

Normal fault
Tensional fault

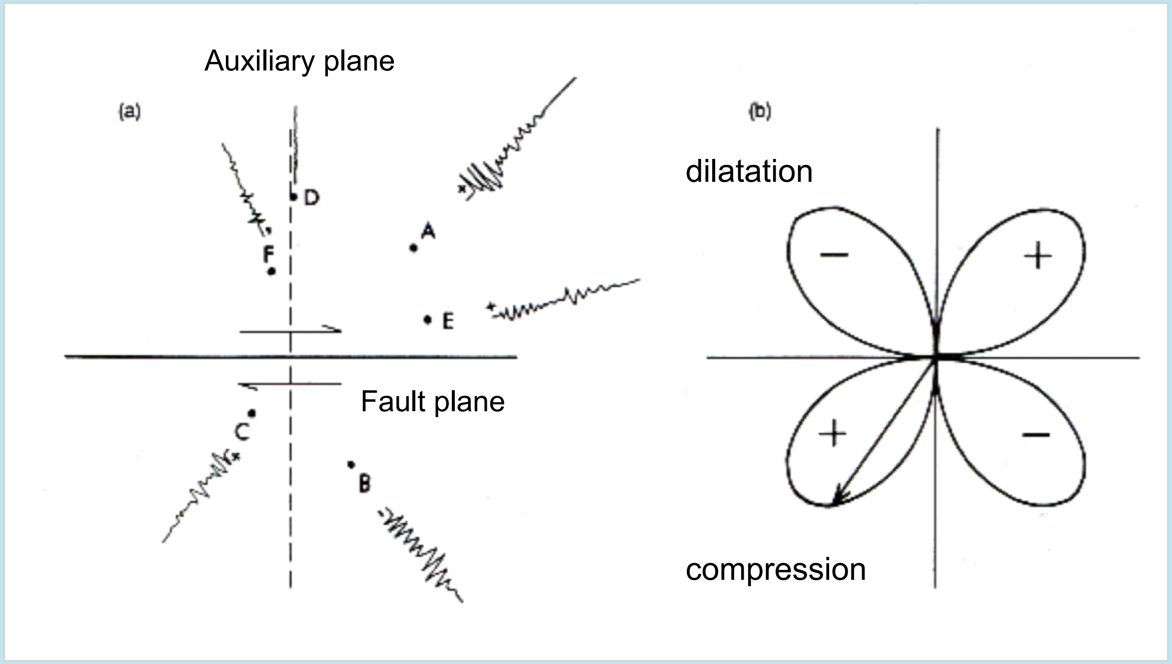
Thrust fault
Reverse fault



Elastic rebound model of earthquakes

1: Unstrained
2: Accumulation of strain near fault
3: Release of strain

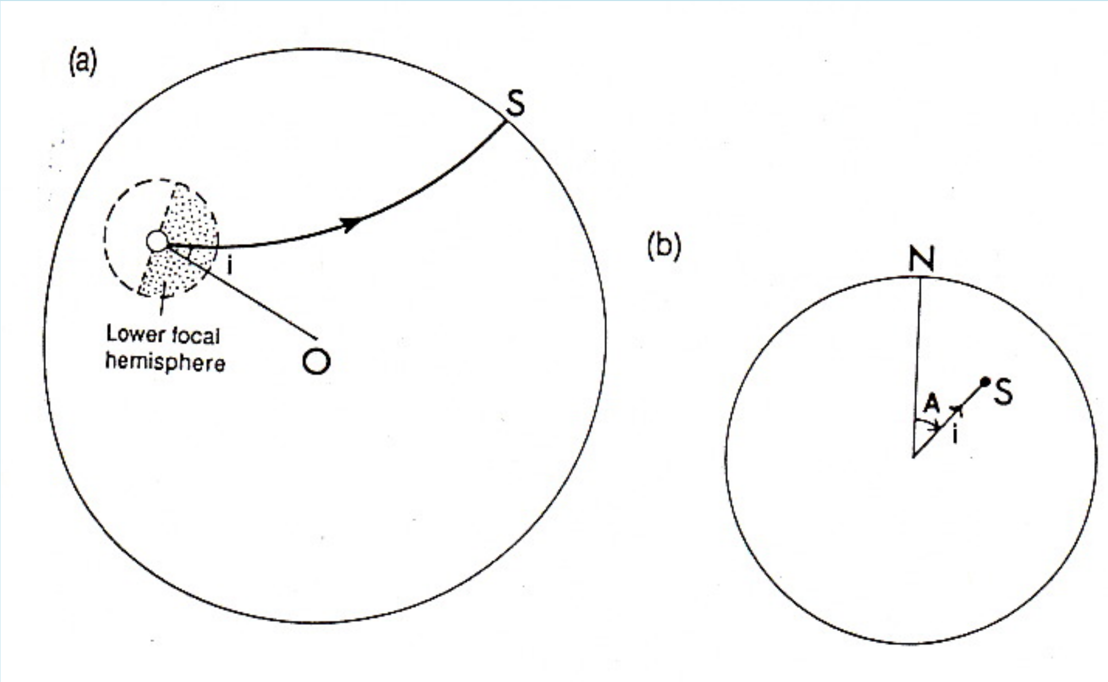
Plane view of an earthquake



a) First motion polarity changes from quadrant to quadrant

b) Amplitude also varies: zero on nodal planes

Focal sphere



Imaginary small sphere centered on the earthquake focus

Plot where the ray path intersects the lower hemisphere

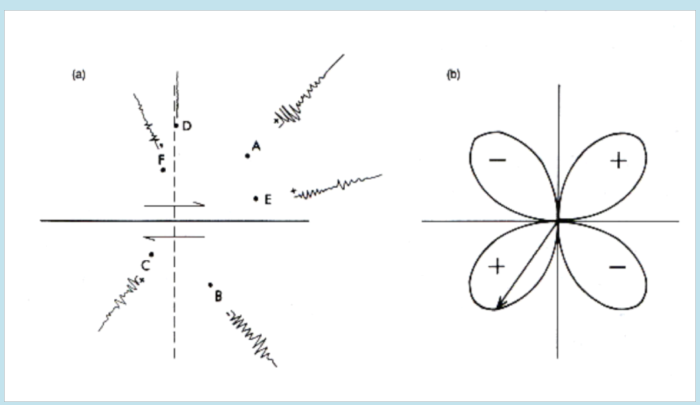
i = angle of incidence

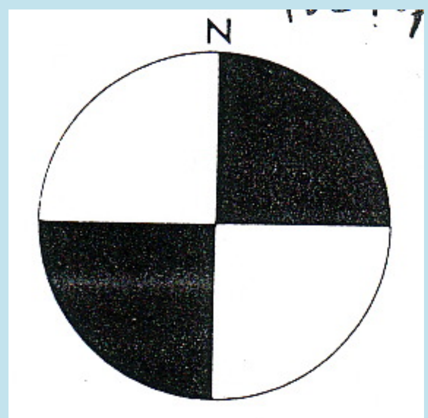
A = azimuth from North

Plot first motion:

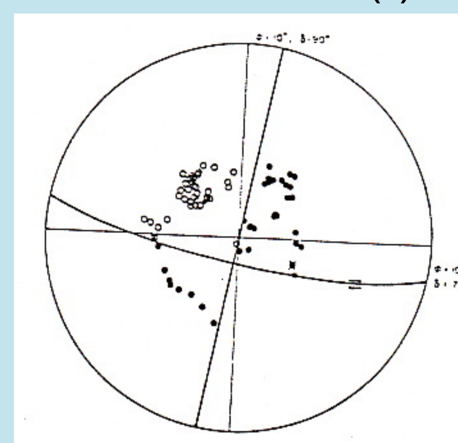
black = compression (+)

white = dilatation (-)

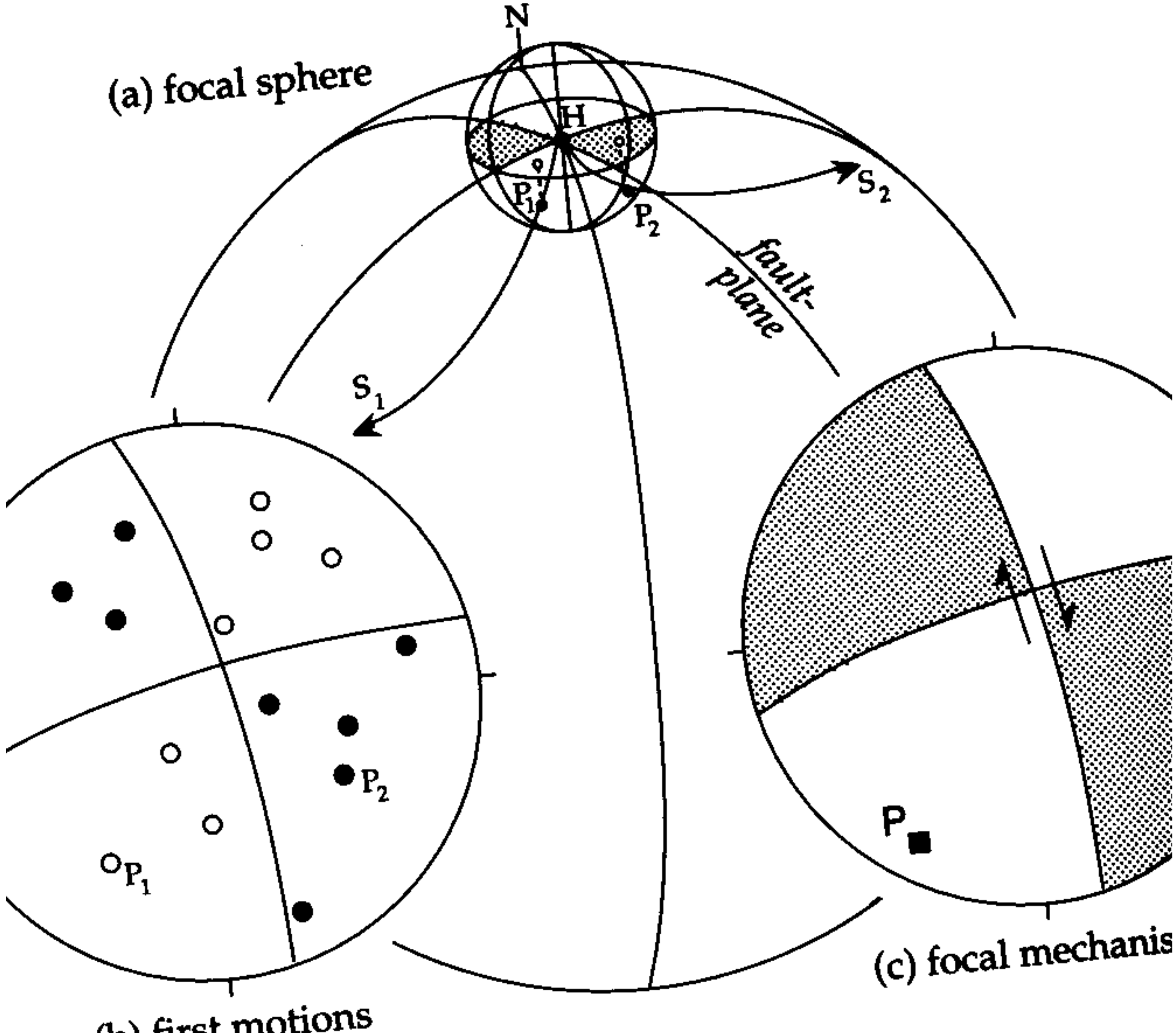




Theoretical solution for strike slip fault

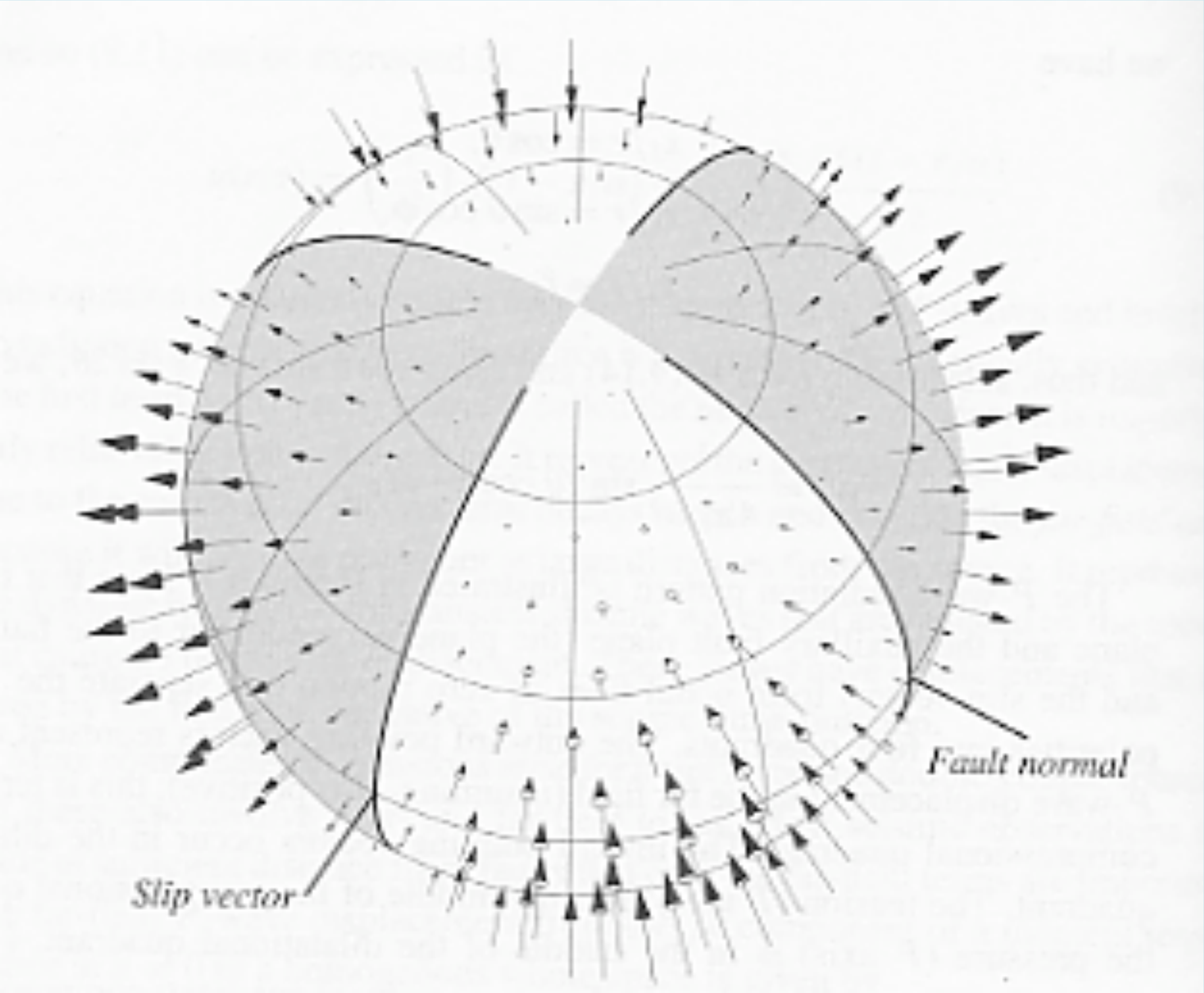


More typical solution

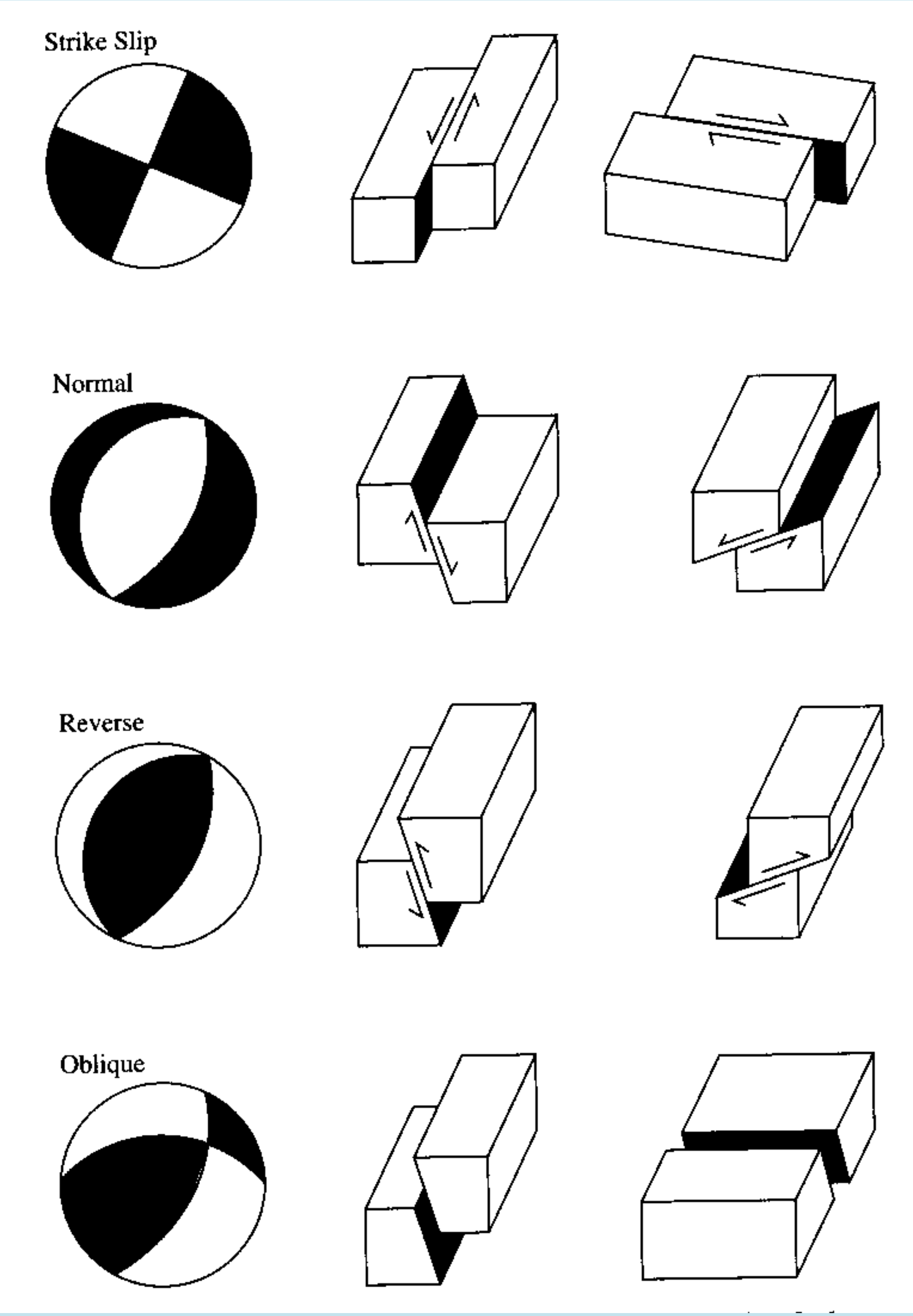


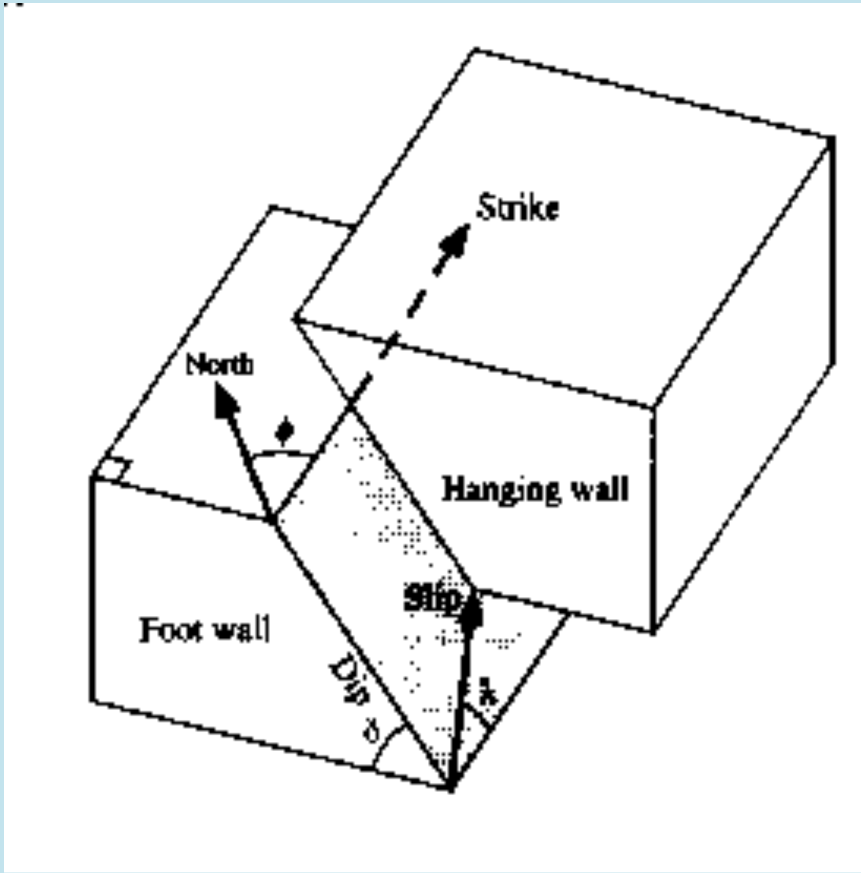
N
(a) focal sphere
H
P1
P2
S2
fault-plane
S1
P2
P1
(b) first motions
P
(c) focal mechanis

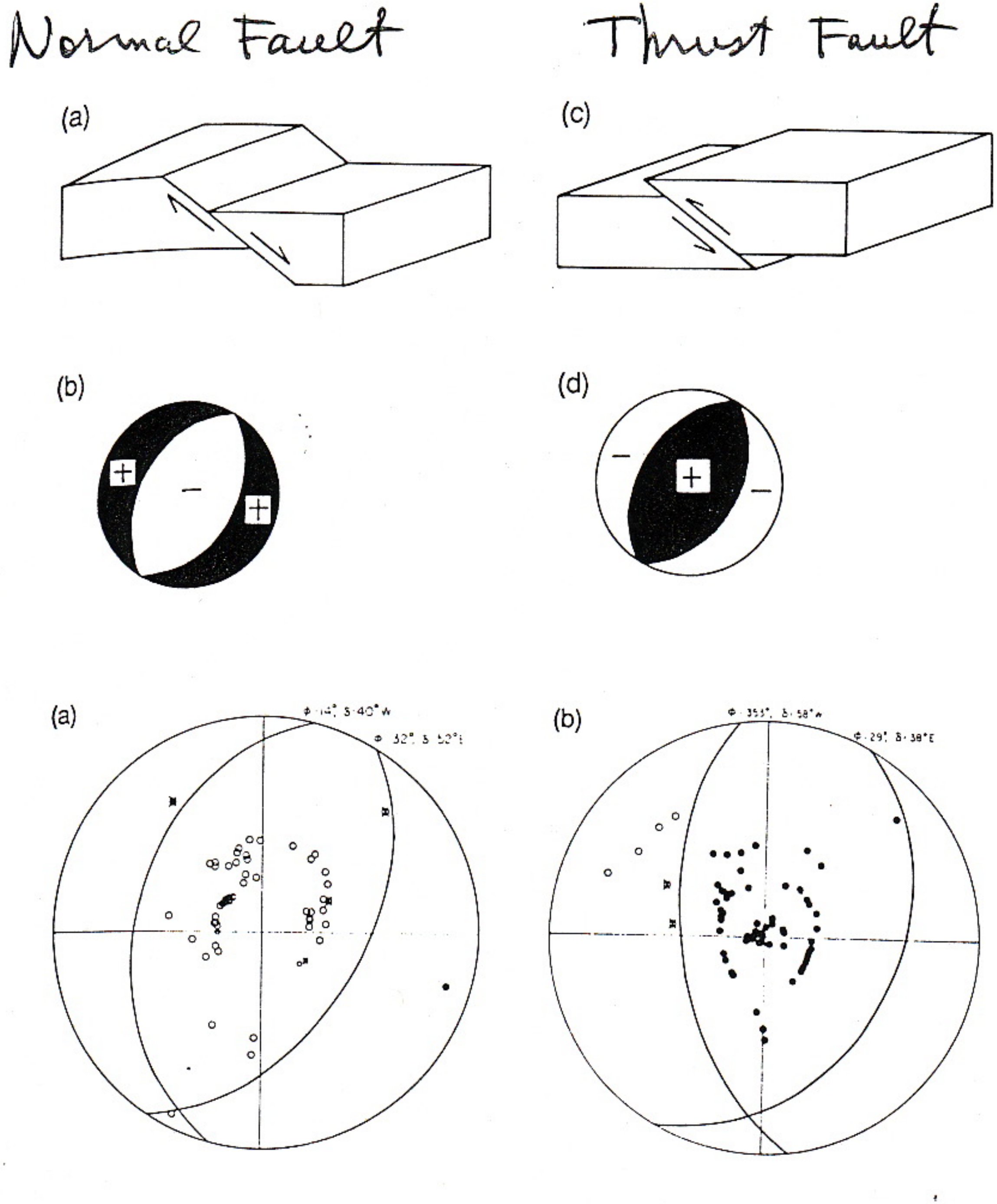
Focal mechanism



Focal mechanisms: “beach balls”







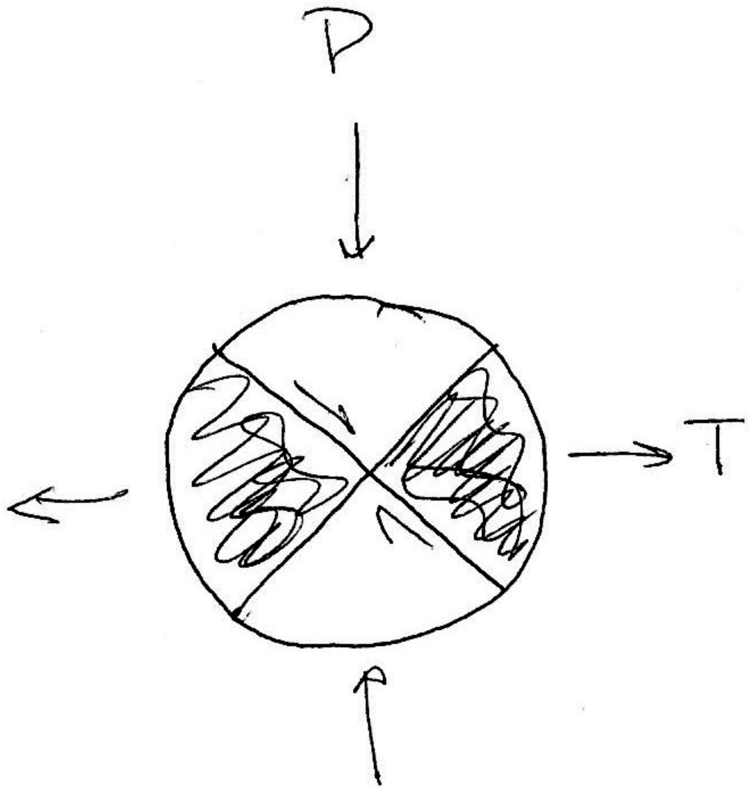
Focal mechanisms for normal and thrust faults

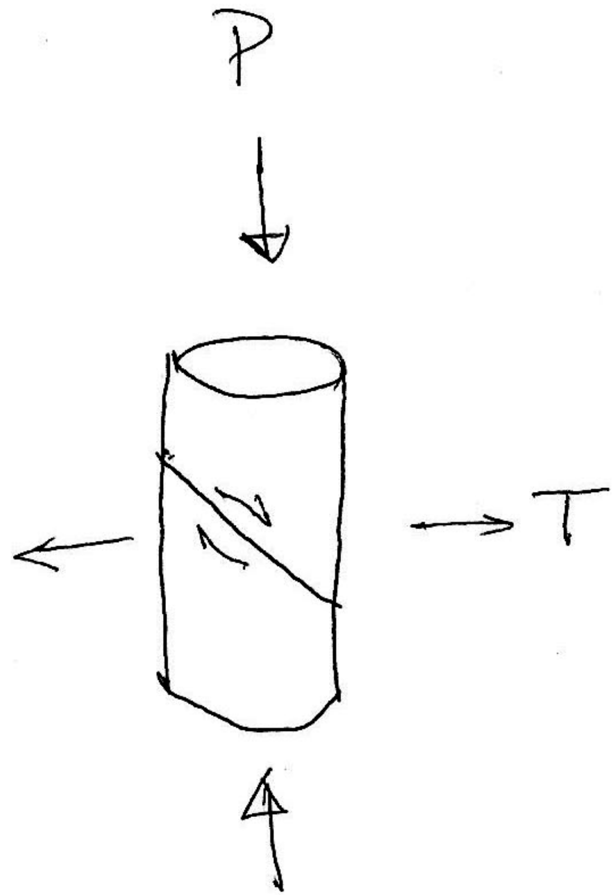
Easy way to remember …

Fundamental ambiguity: can not distinguish between fault plane and auxiliary plane

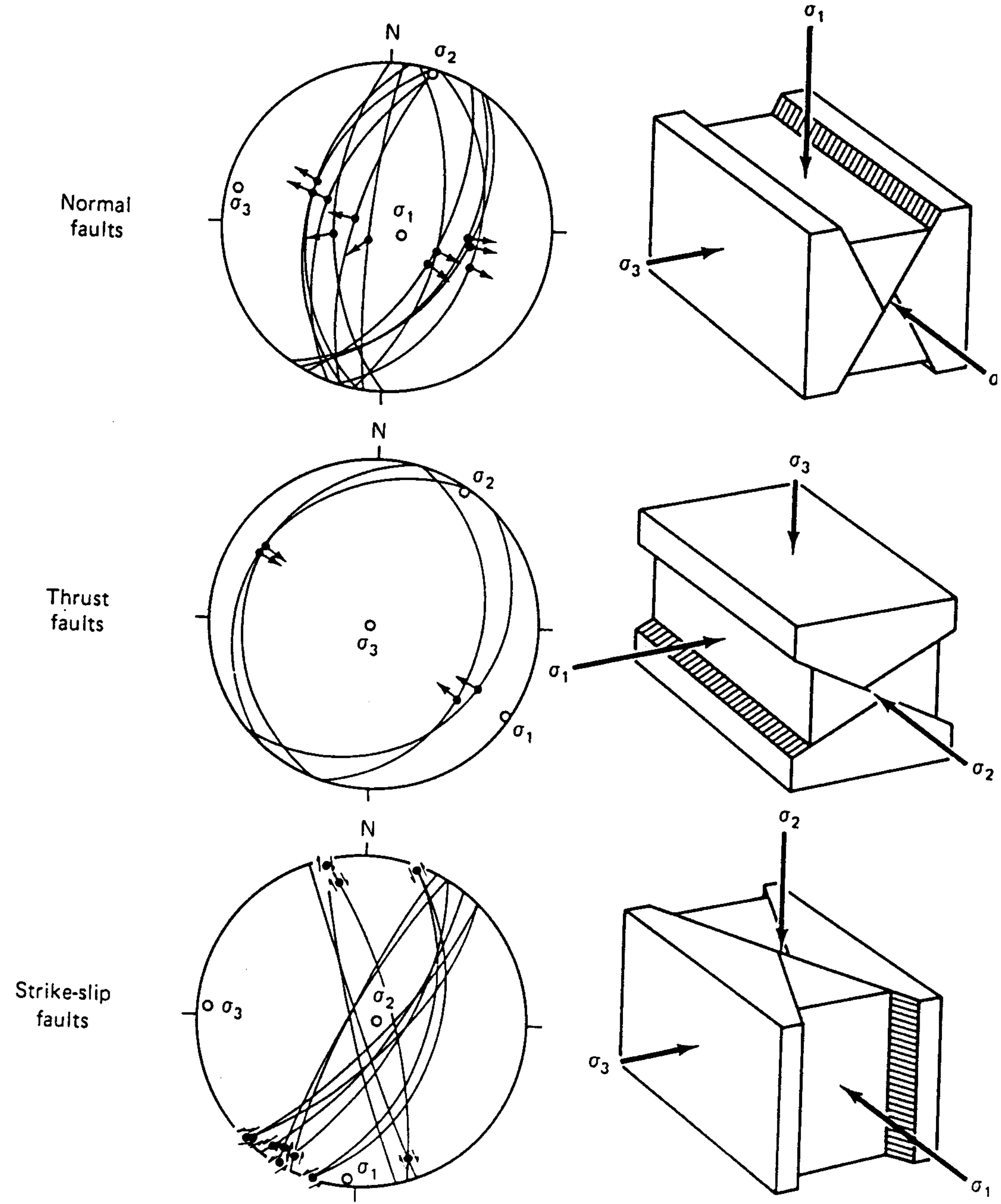
Infer fault plane based on other tectonic information

Principal Stress Axes

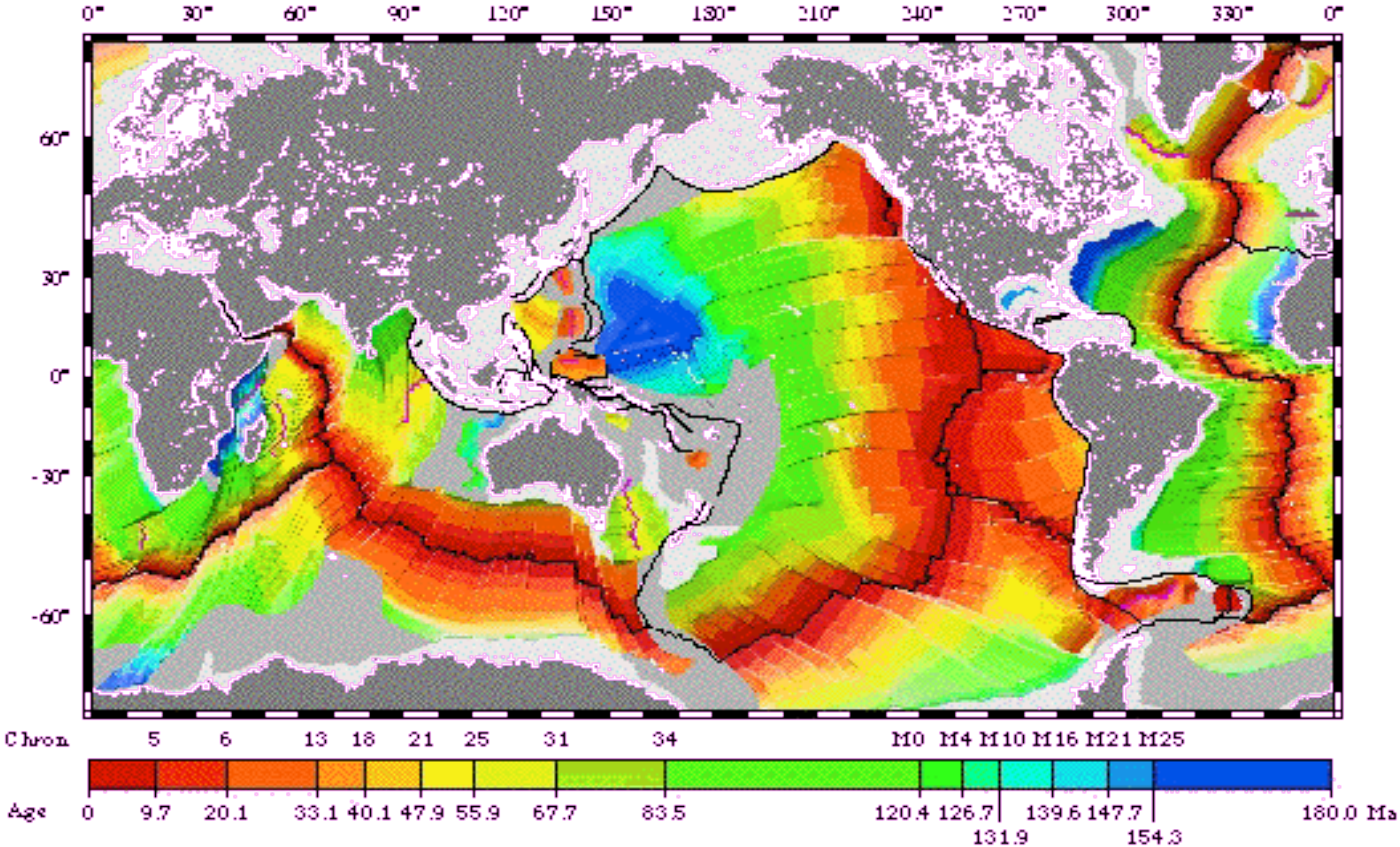




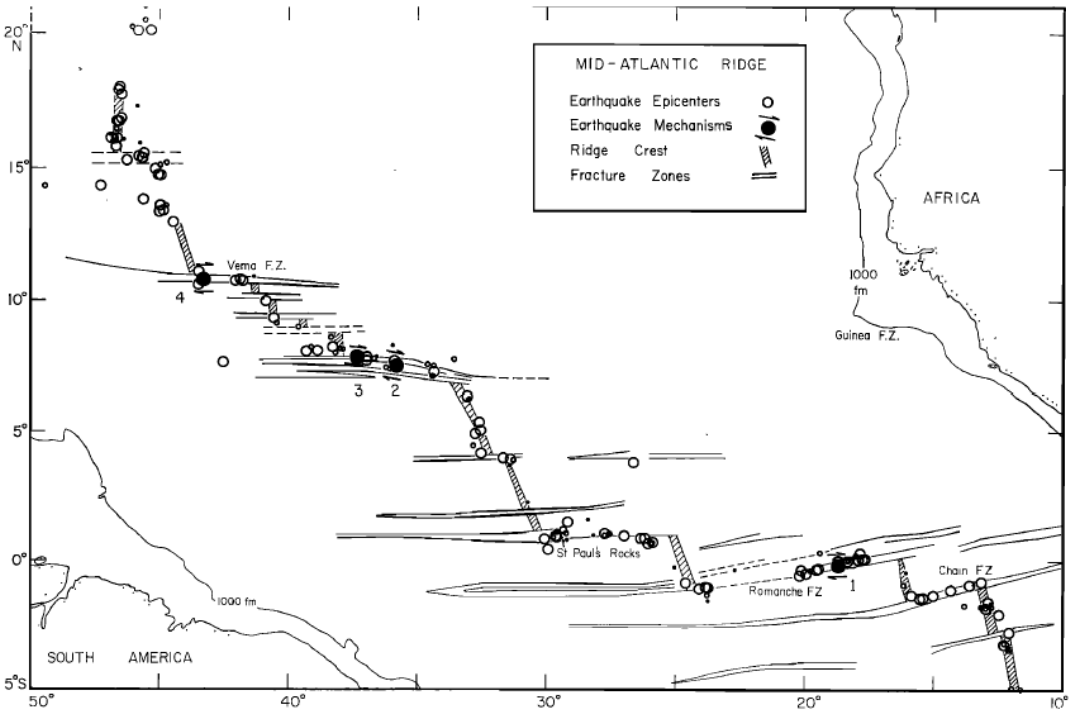
Anderson's fault theory:

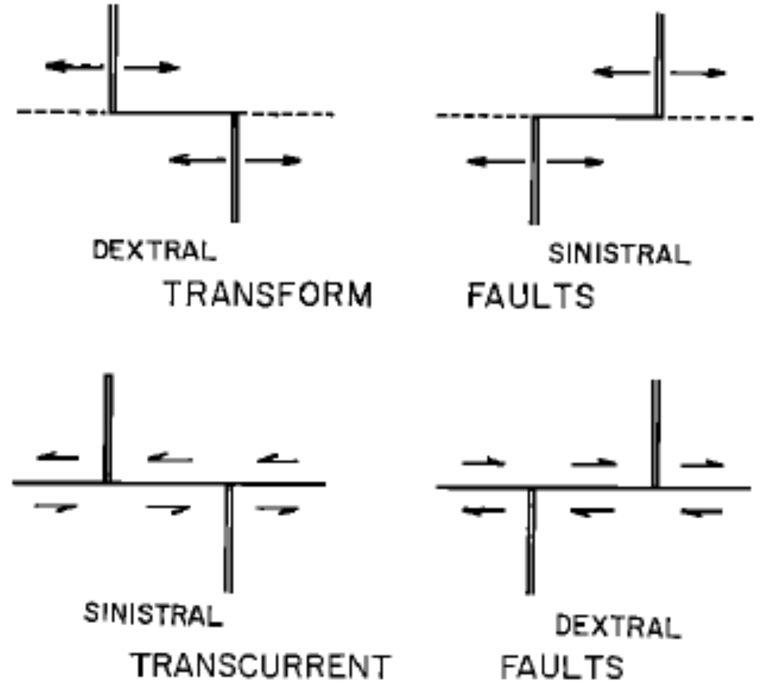


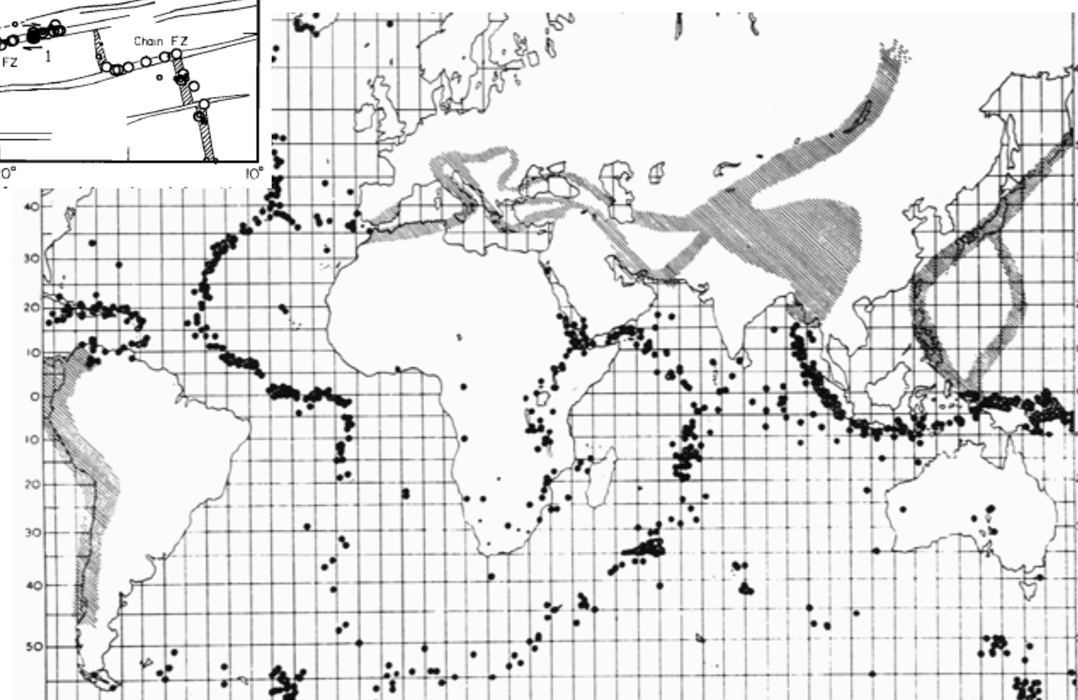
Oceanic transforms: seismic test of plate tectonics



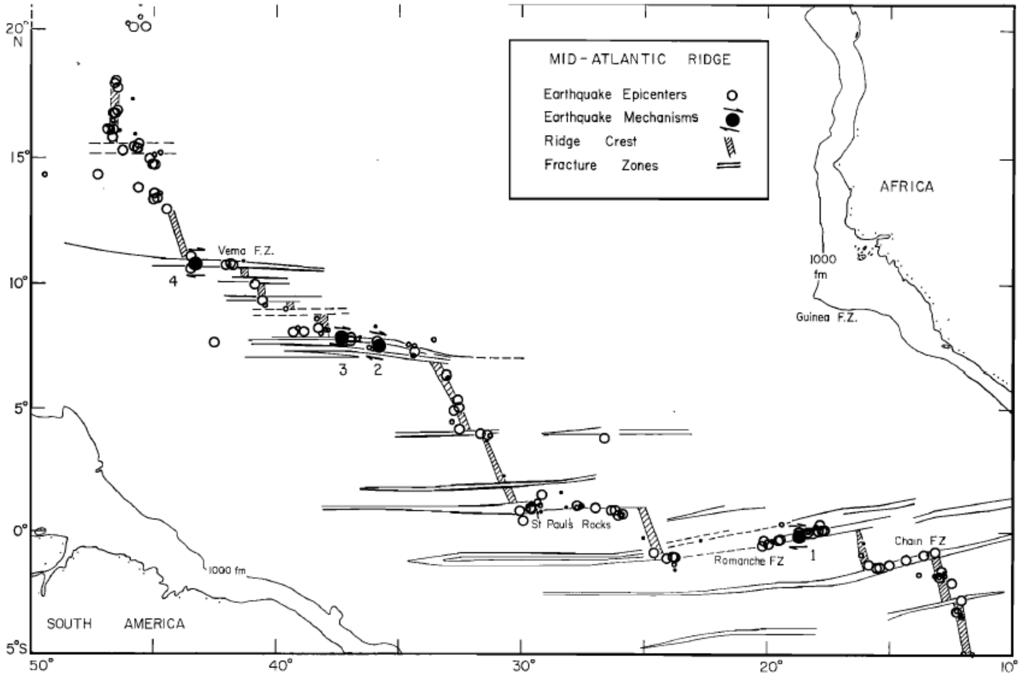
Transform Faults

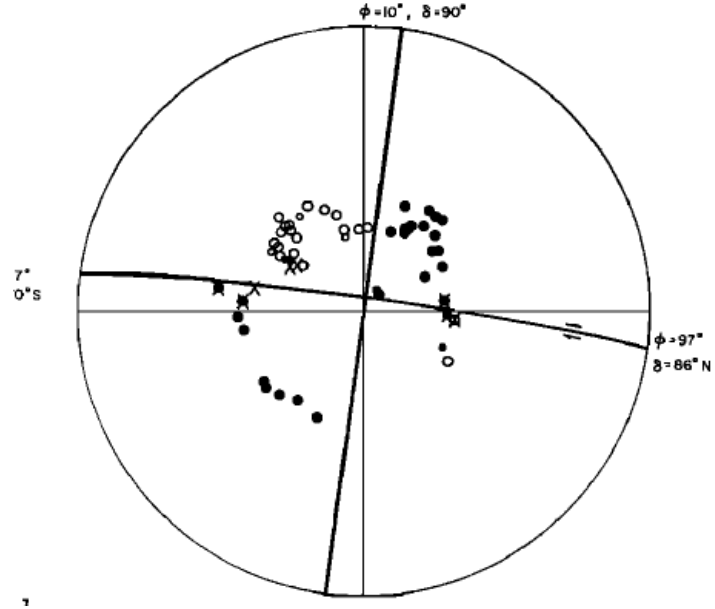




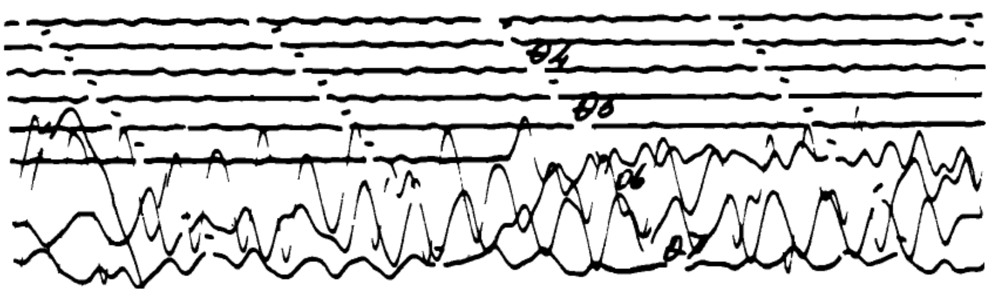


J. Tuzo Wilson (1965)

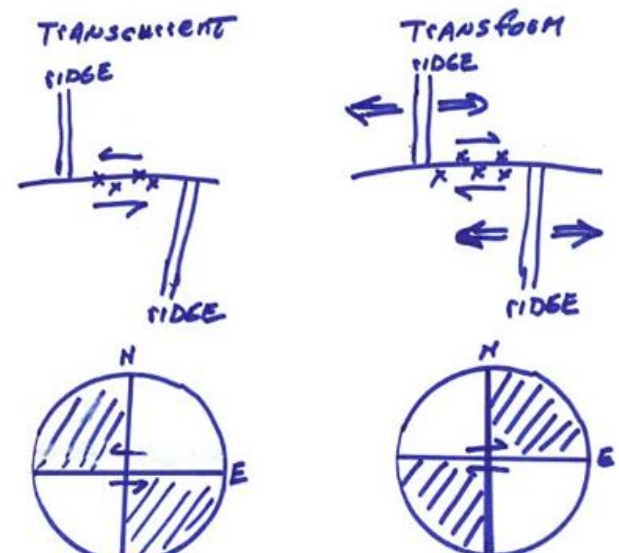




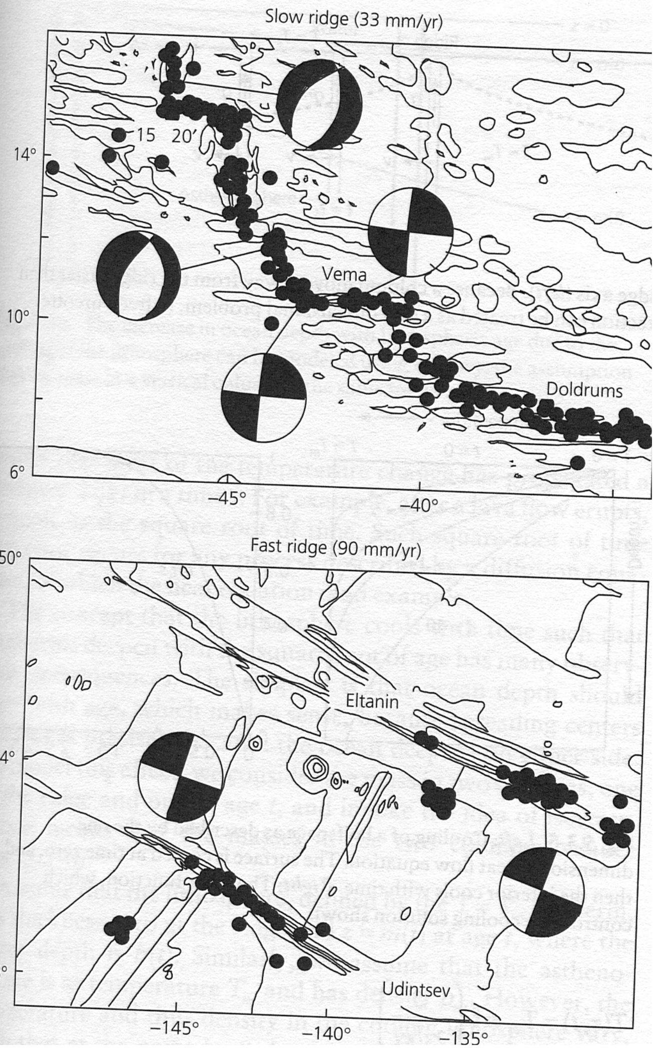
Focal mechanisms; fault plane solutions



The seismologists verify Wilson's transform fault hypothesis;
World wide seismic network

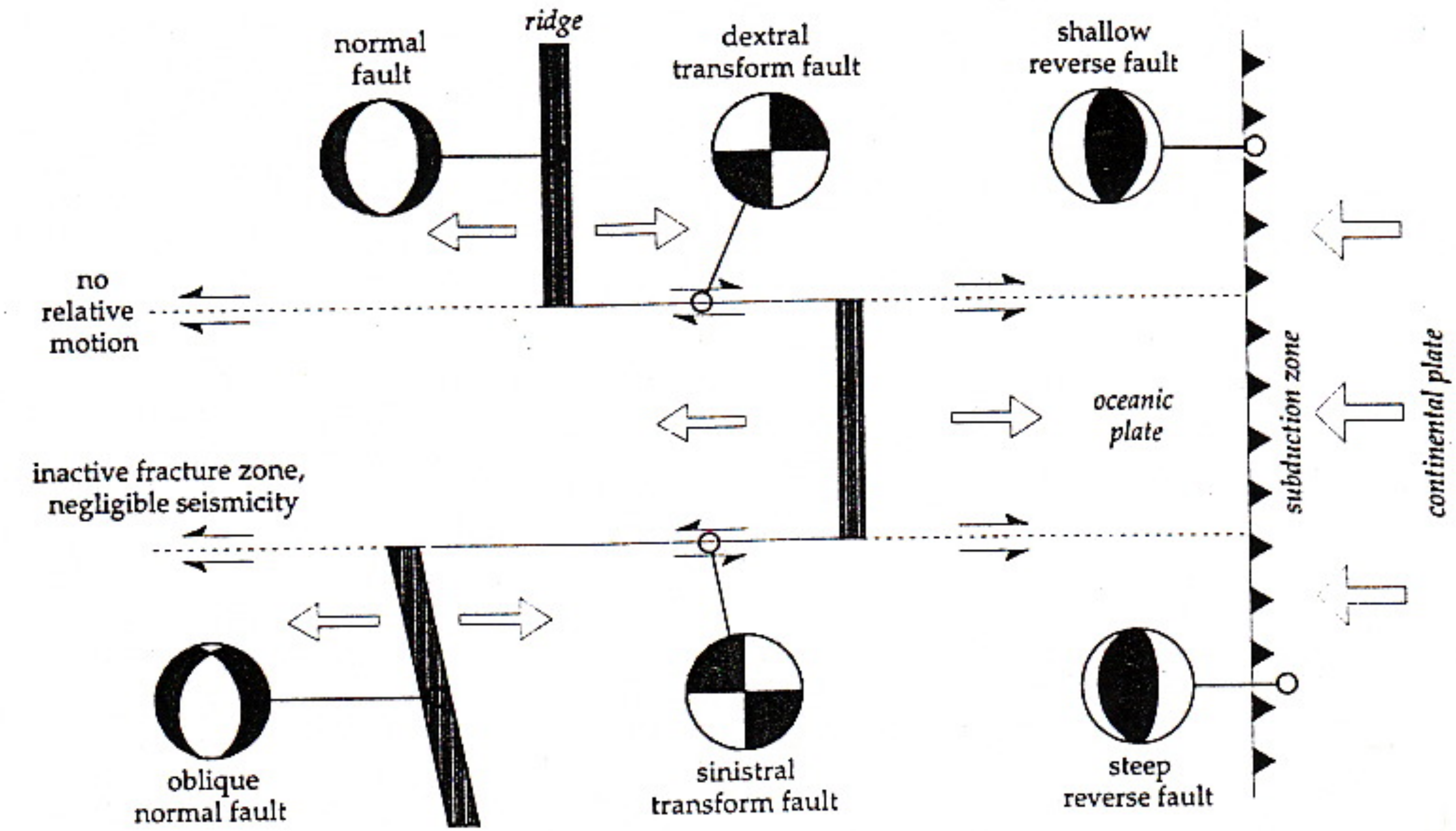


Lynn Sykes (1967)

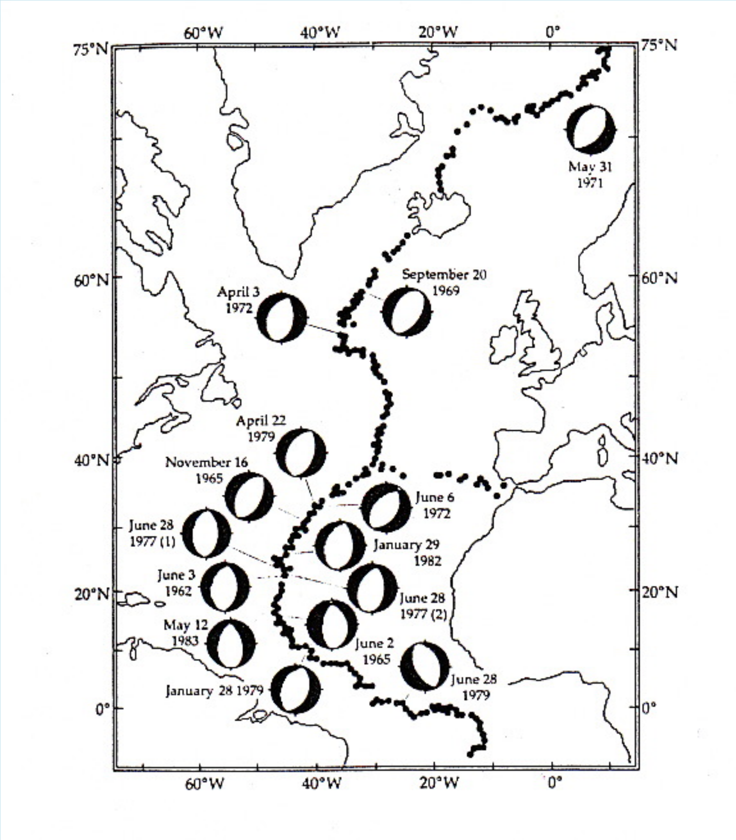


Slow ridge (33 mm/yr)
15
20′
14°
10°
6°
Vema
Doldrums
−45°
−40°
Fast ridge (90 mm/yr)
−50°
−54°
−58°
Eltanin
Udintsev
−145°
−140°
−135°

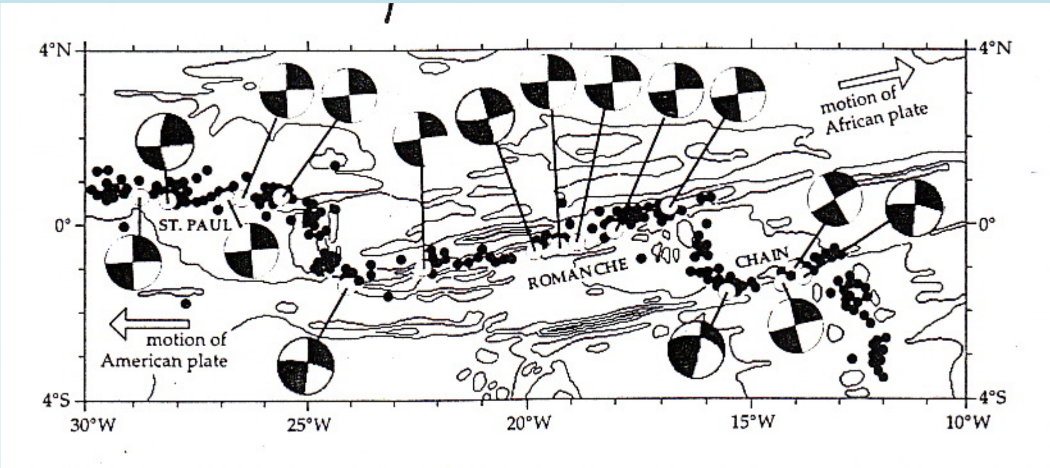
Examples of hypothetical fault plane solutions



Mid-Atlantic ridge



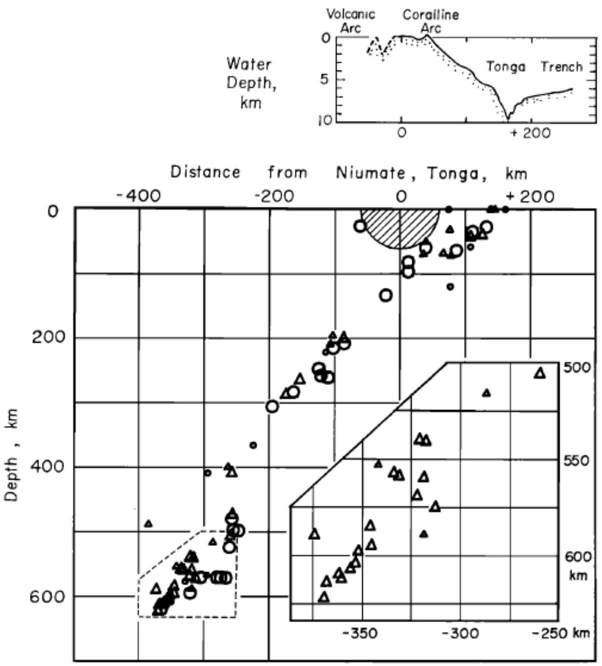
Romanche transform fault



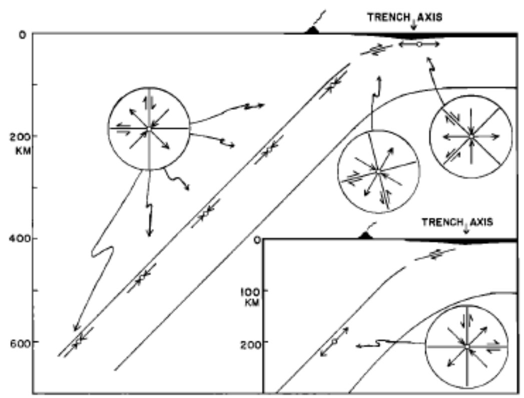
Strike slip motion

Normal motion

Imaging the subducting slab



ISACKS, OLIVER, AND SYKES



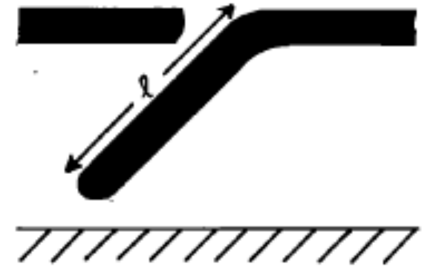
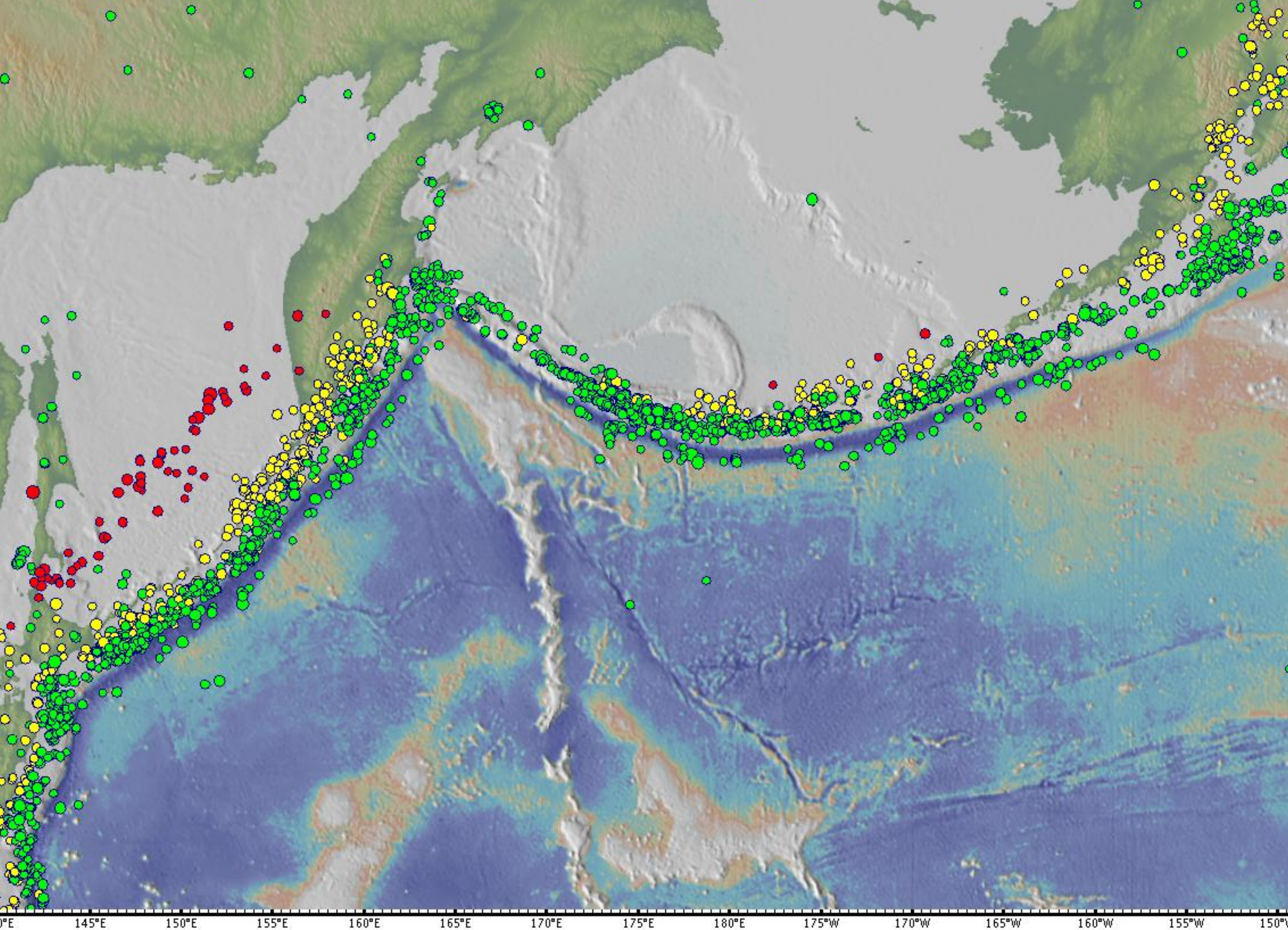


Fig. 14*a*. Length *l* is a measure of the amount of underthrusting during the most recent period of sea-floor spreading.

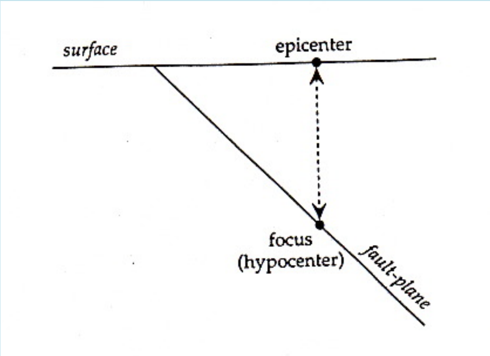
Fig. 14*b*. Lithosphere is deformed along its lower edge as it encounters a more resistant layer (the mesosphere).

Fig. 14*c*. Length of seismic zone is the product of rate of underthrusting and time constant for assimilation of slab by upper mantle.

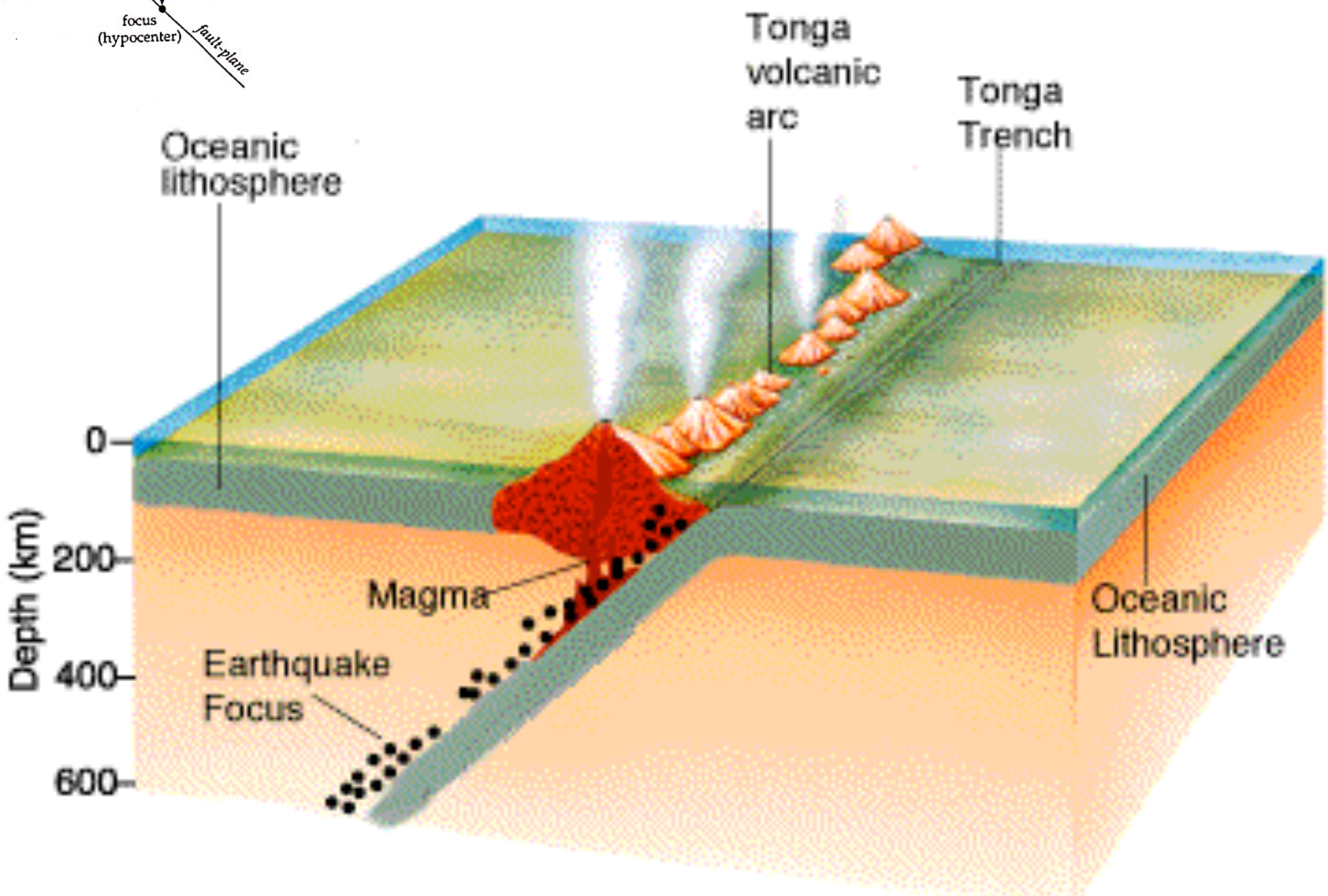
Isacks, Oliver and Sykes (1968)

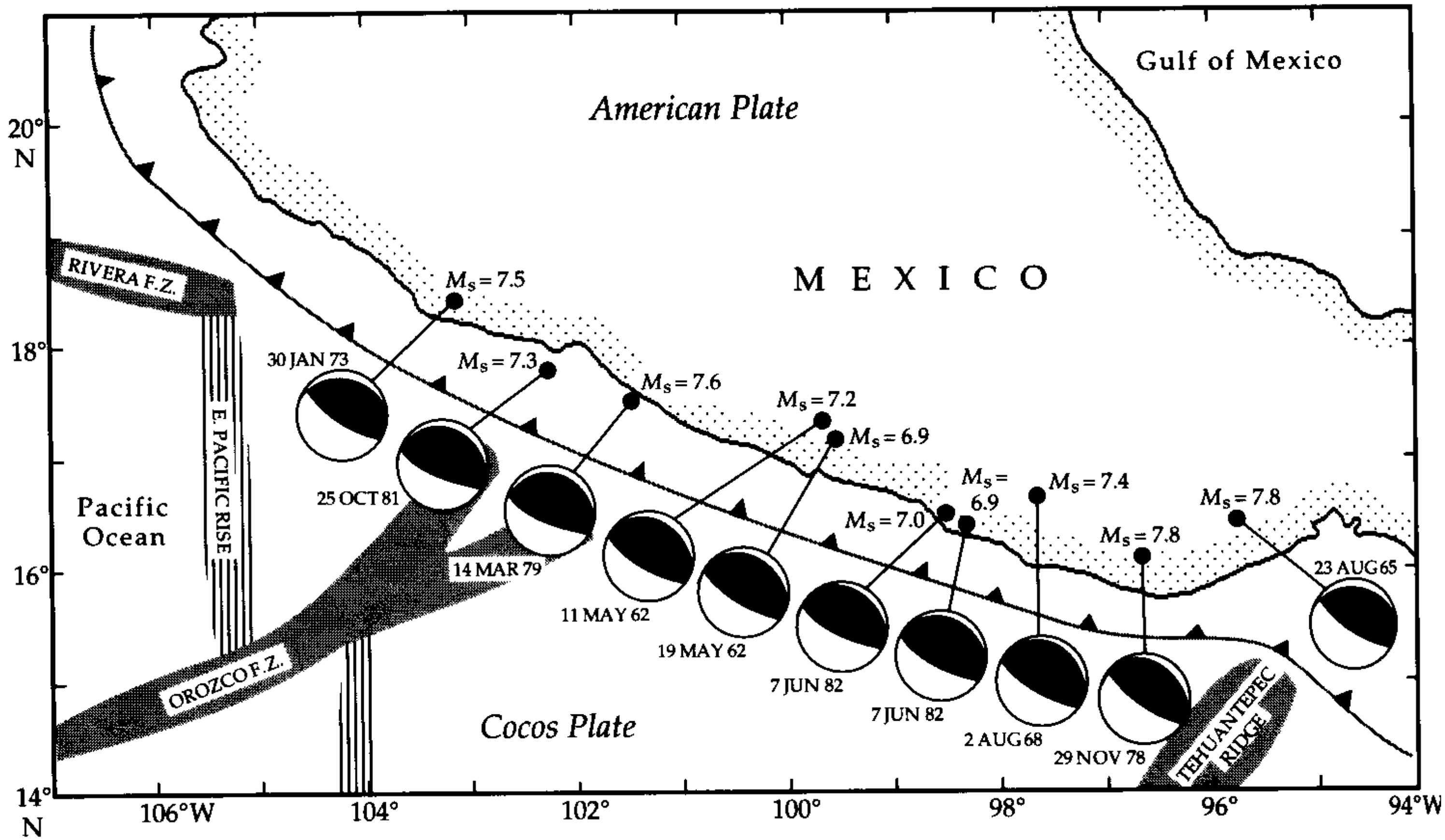

145°E
150°E
155°E
160°E
165°E
170°E
175°E
180°E
175°W
170°W
165°W
160°W
155°W

Benioff Zone



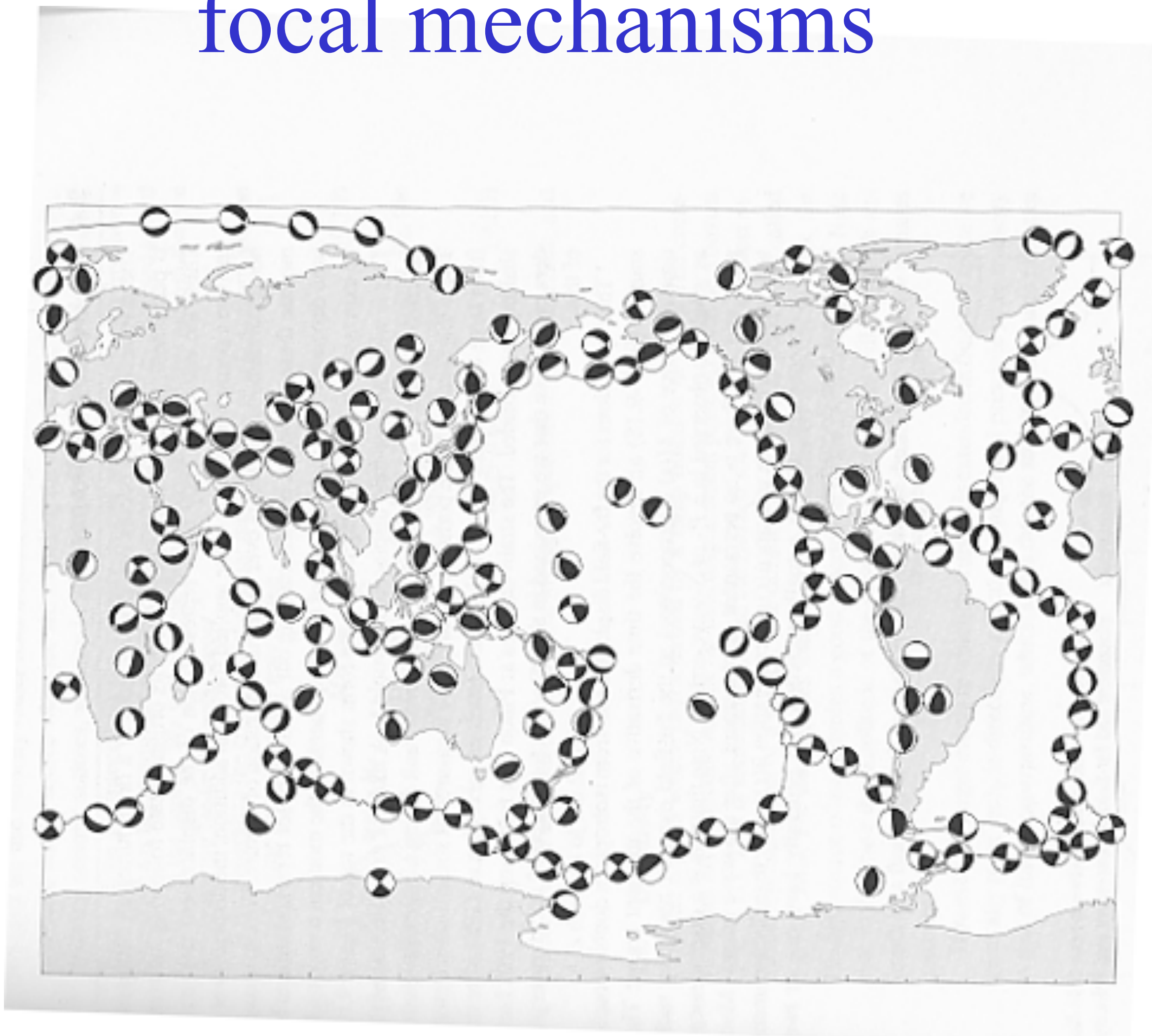
Note difference between epicenter and hypocenter (focus)

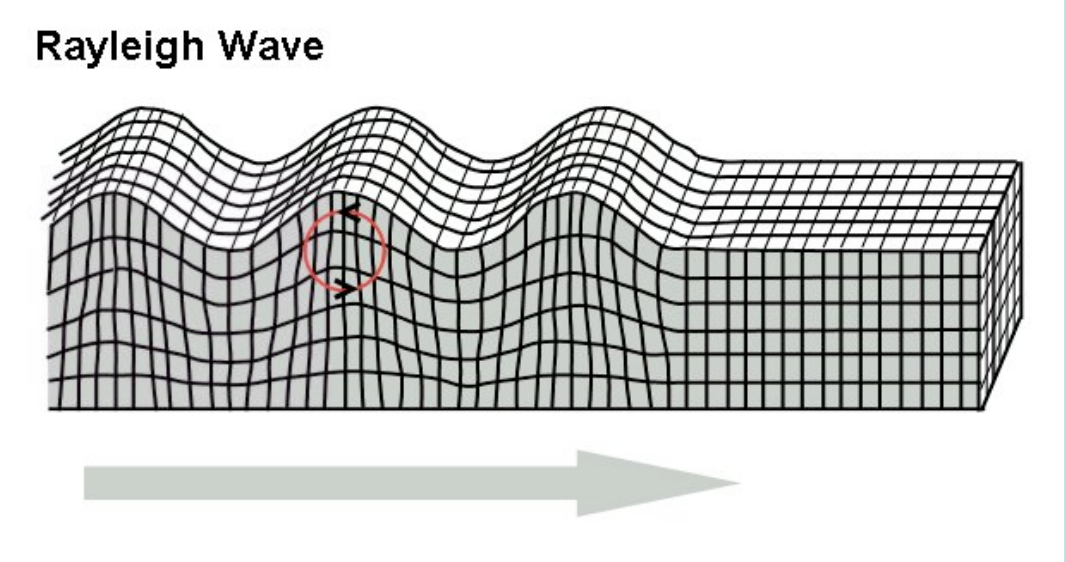


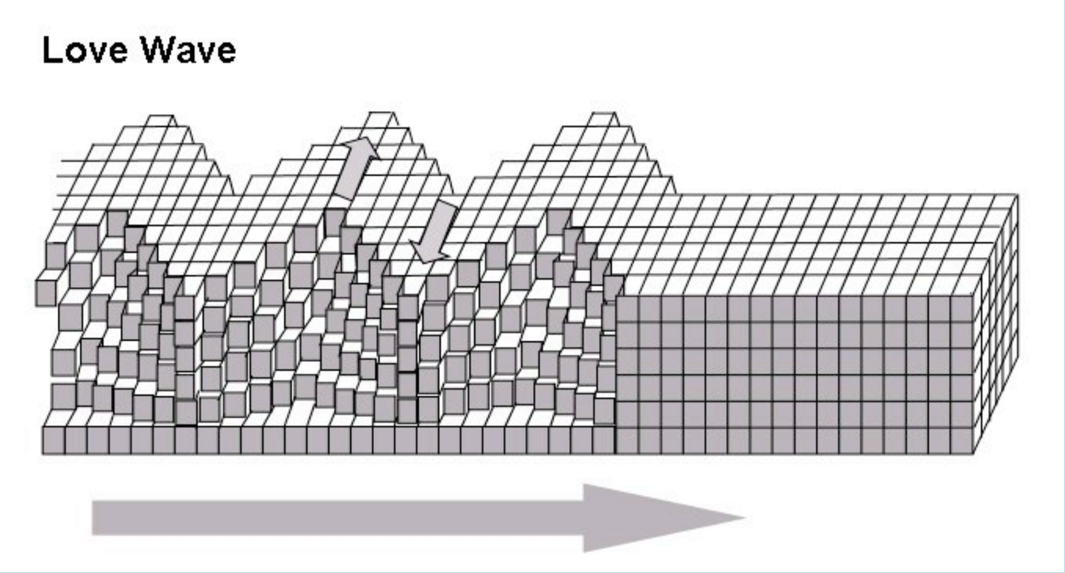


Gulf of Mexico
American Plate
MEXICO
Pacific Ocean
Cocos Plate
RIVERA F.Z.
E. PACIFIC RISE
OROZCO F.Z.
TEHUANTEPEC RIDGE
M_s = 7.5
30 JAN 73
M_s = 7.3
25 OCT 81
M_s = 7.6
14 MAR 79
M_s = 7.2
11 MAY 62
M_s = 6.9
19 MAY 62
M_s = 7.0
7 JUN 82
M_s = 6.9
7 JUN 82
M_s = 7.4
2 AUG 68
M_s = 7.8
29 NOV 78
M_s = 7.8
23 AUG 65
20° N
18°
16°
14° N
106°W
104°
102°
100°
98°
96°
94°W

Global distribution of the earthquake focal mechanisms





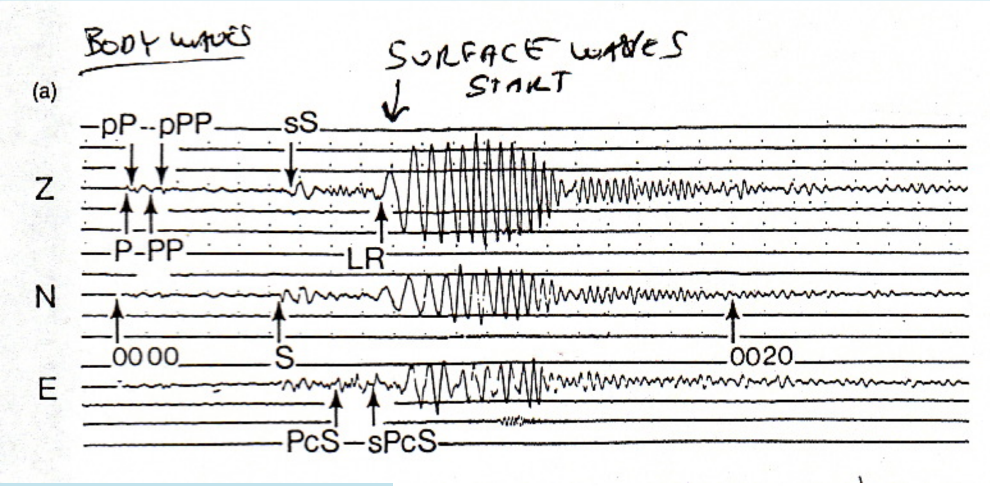


Surface waves

Amplitude decreases with depth
"Samples" to depth = wavelength/3
Use to determine gross crustal shear wave structure

Rayleigh waves: "ground roll";
Motion at the surface is a retrograde vertical ellipse;
LR or R (L stands for long, R for Rayleigh)

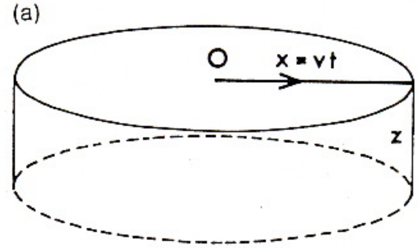
Love waves: horizontally polarized S waves;
Motion is transverse and horizontal;
LQ or Q (querwellen: German for "transverse wave")



Amplitudes of surface waves fall off slower and are larger than body waves

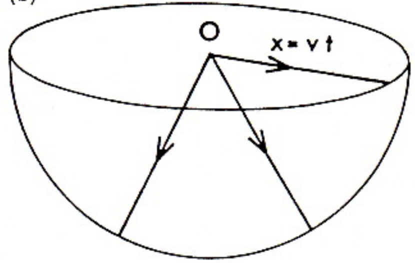
Comparison of Amplitudes of Surface (A) & Body waves (B)

(a)



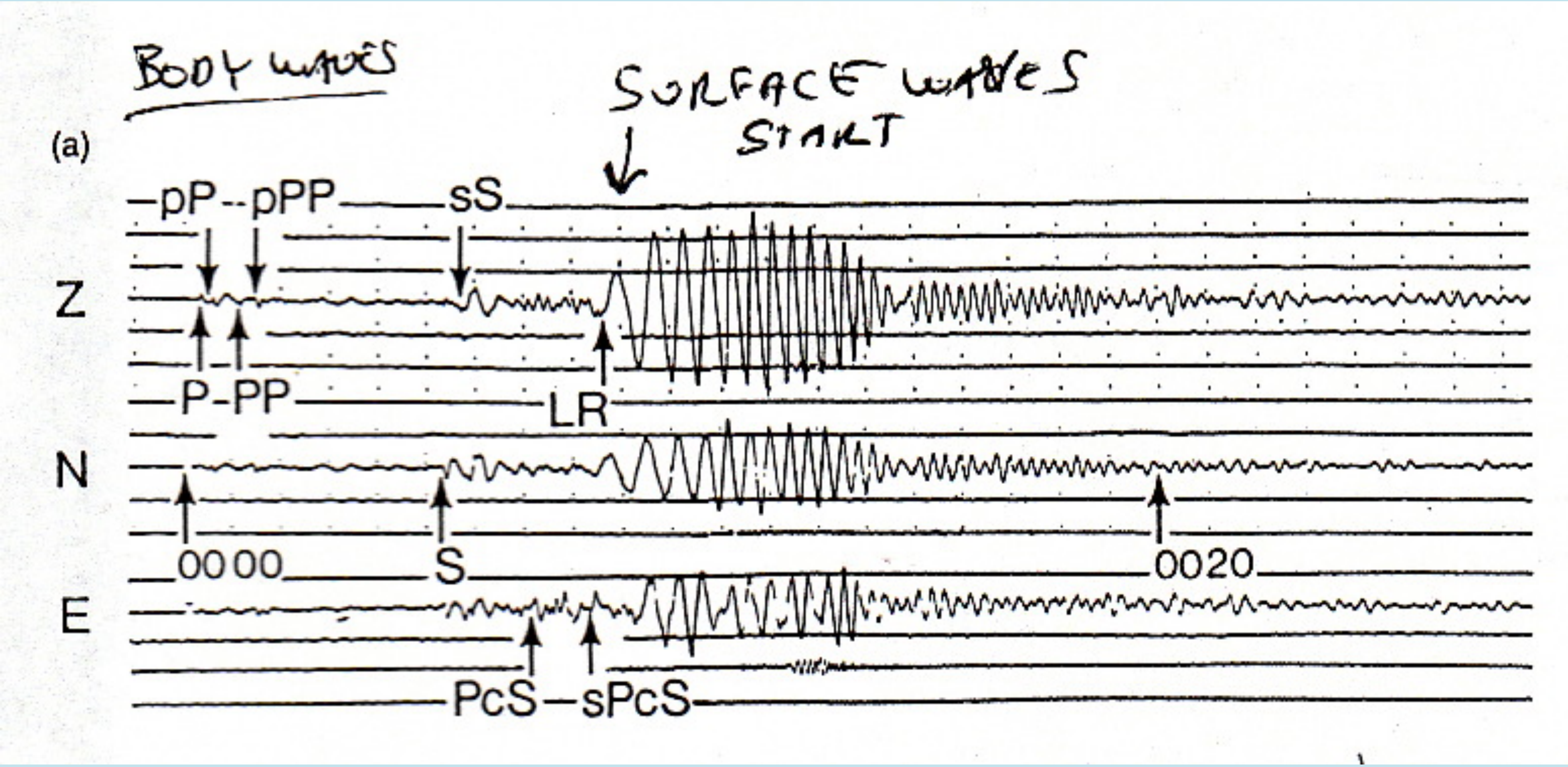
Amplitude at $x \propto \frac{1}{\sqrt{x}}$

(b)

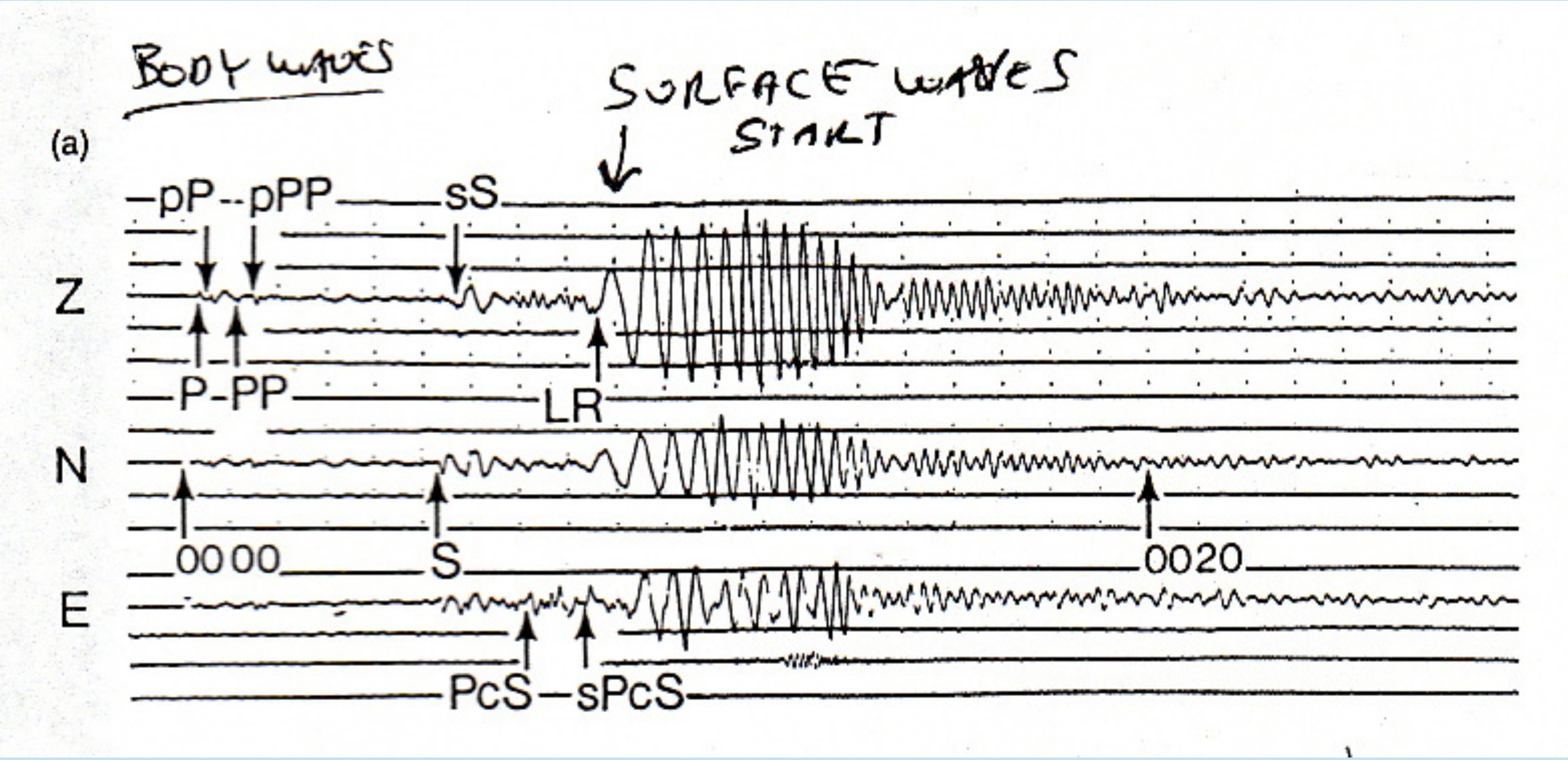


Amplitude at $x \propto \frac{1}{x}$

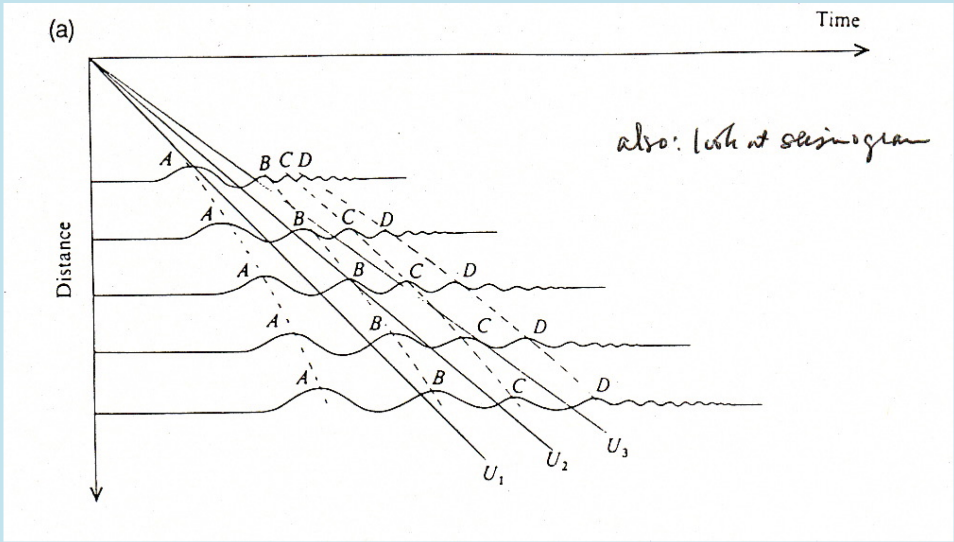
What else is noticeable about surface waves?

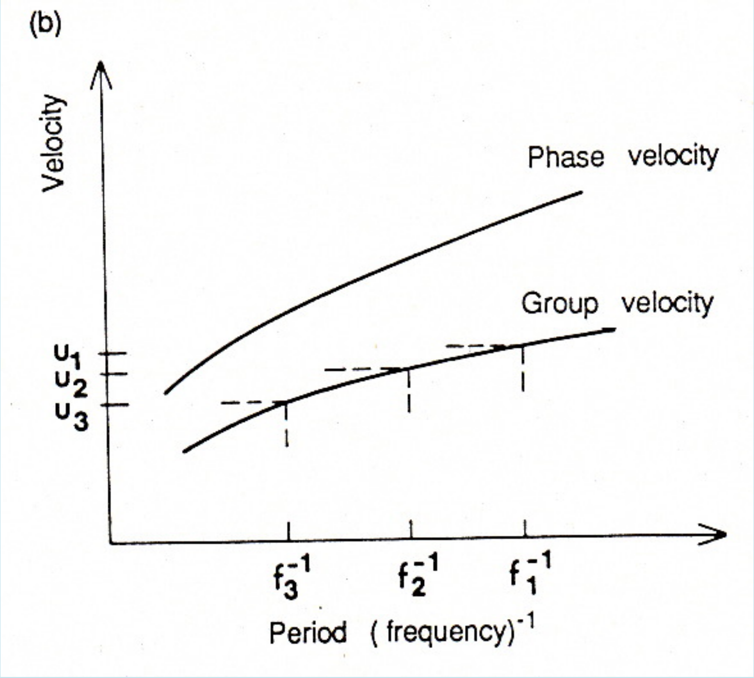


Surface wave dispersion:
note that longer period surface waves arrive first



Surface wave dispersion





Longer period portion of signal moves faster (arrives first)

As signal moves away from source, length of wave train increases

Phase velocity (v) = velocity of a particular crest or trough (dashed lines)

Group velocity (u) = velocity of a constant frequency portion of the signal;

group velocity decreases with increasing frequency

Similar to storm waves at sea:
Long period waves precede a storm;
Wavelengths get shorter as a storm approaches

Classic study: Use surface waves that cross different age oceanic crust to map thickening of lithosphere as it ages

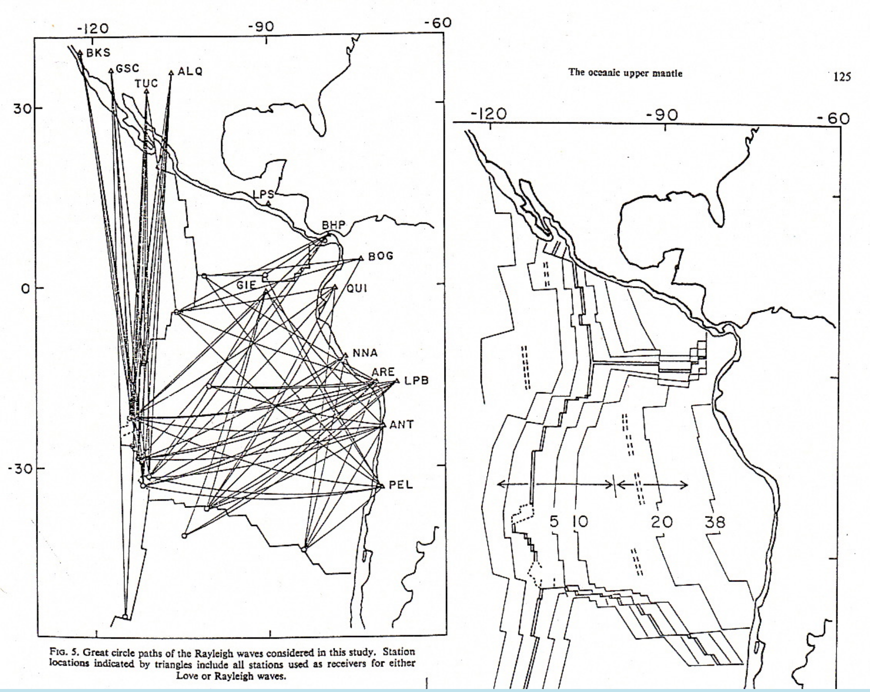
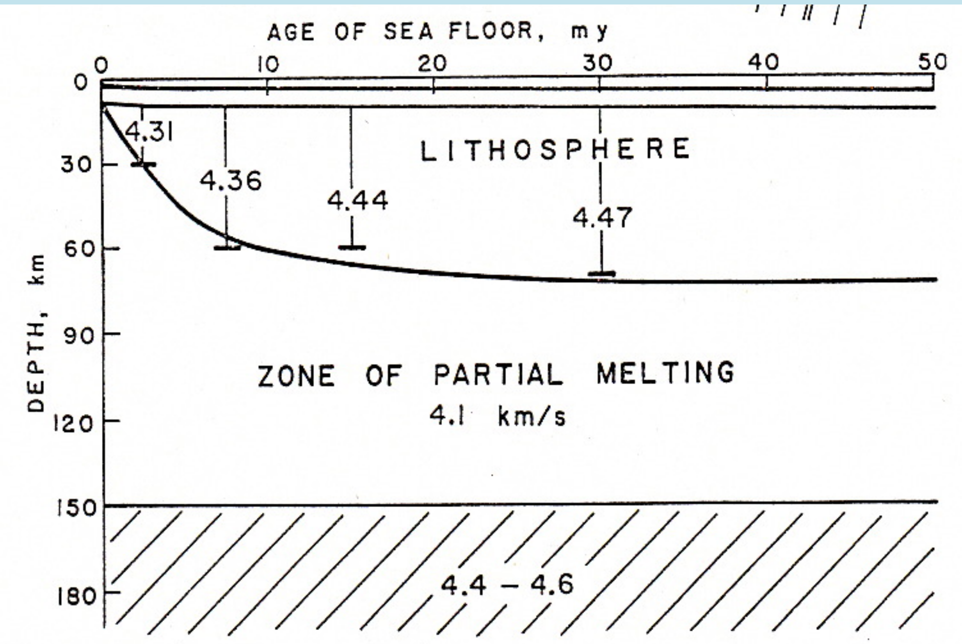
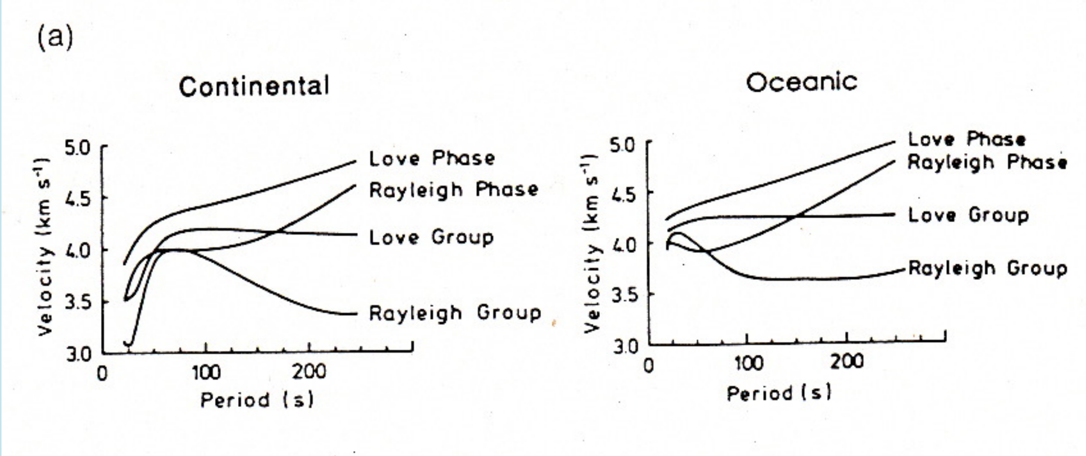


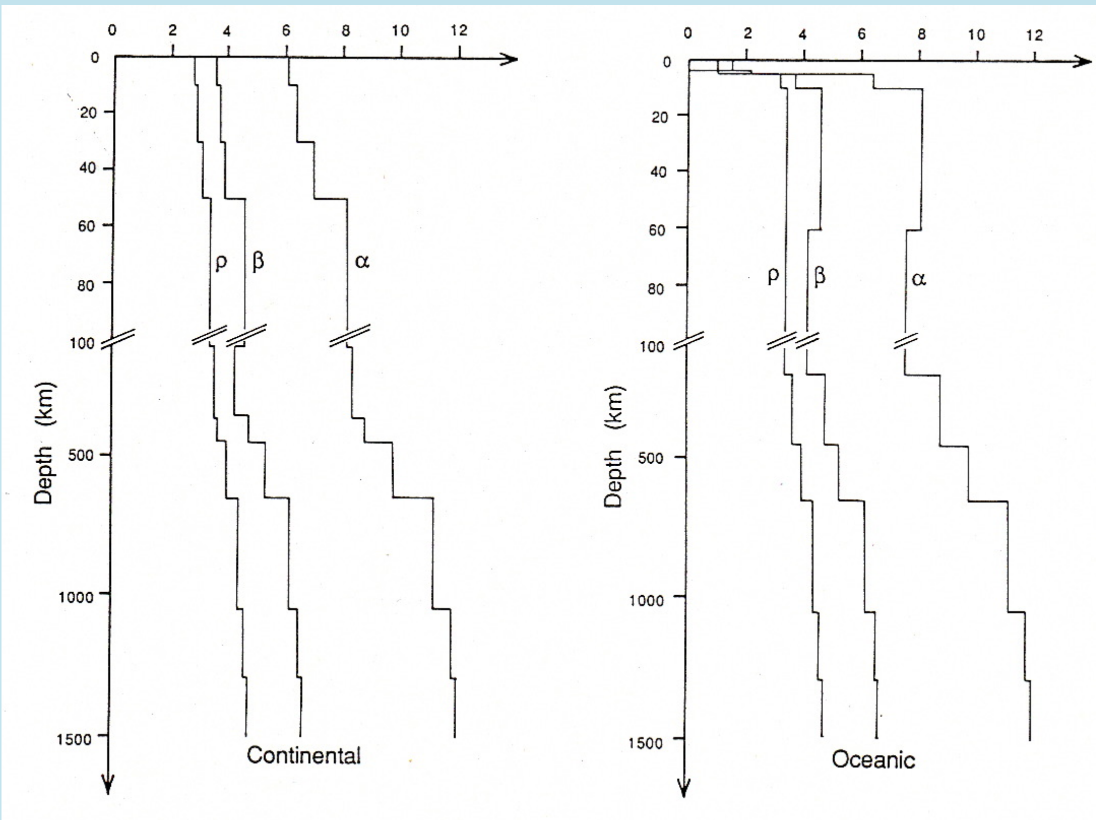
Fig. 5. Great circle paths of the Rayleigh waves considered in this study. Station locations indicated by triangles include all stations used as receivers for either Love or Rayleigh waves.

The pattern of surface wave dispersion is related to the shear wave velocity and thickness of the lithosphere





Standard surface wave dispersion curves for continental and oceanic paths



Upper mantle structure based on standard dispersion curves