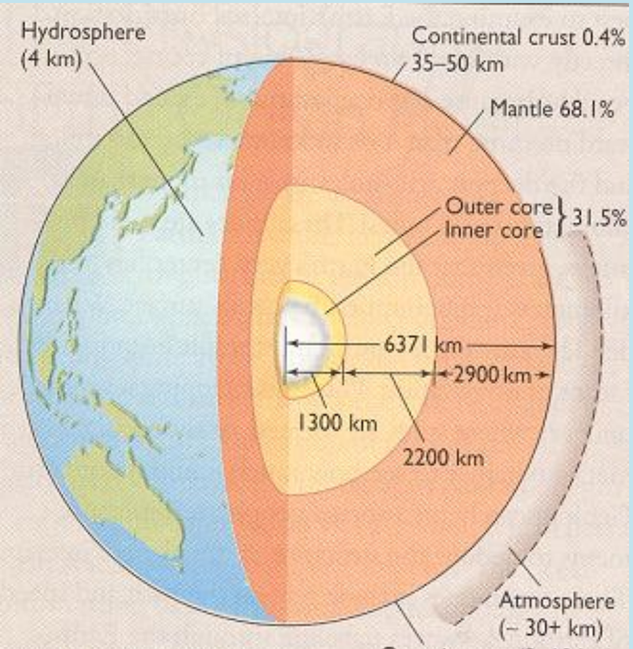
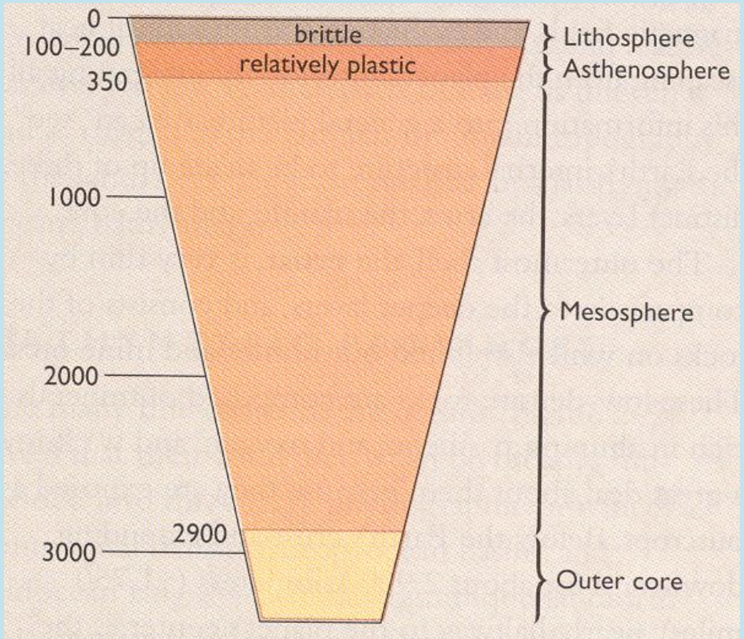
Global Tectonics

Lecture 7 – Earth's Interior

Read Chapters 2.3, 2.4, 2.8, 2.9, 2.12 of KK&V



(a) THE EARTH'S STRUCTURE



(b) LAYERED, INTERNAL STRUCTURE OF THE EARTH

How do we know what's inside?

Geodesy: shape and size of the Earth (R ~6370 km)

Gravity: mass of the Earth ($6\ 10^{24}$ kg)

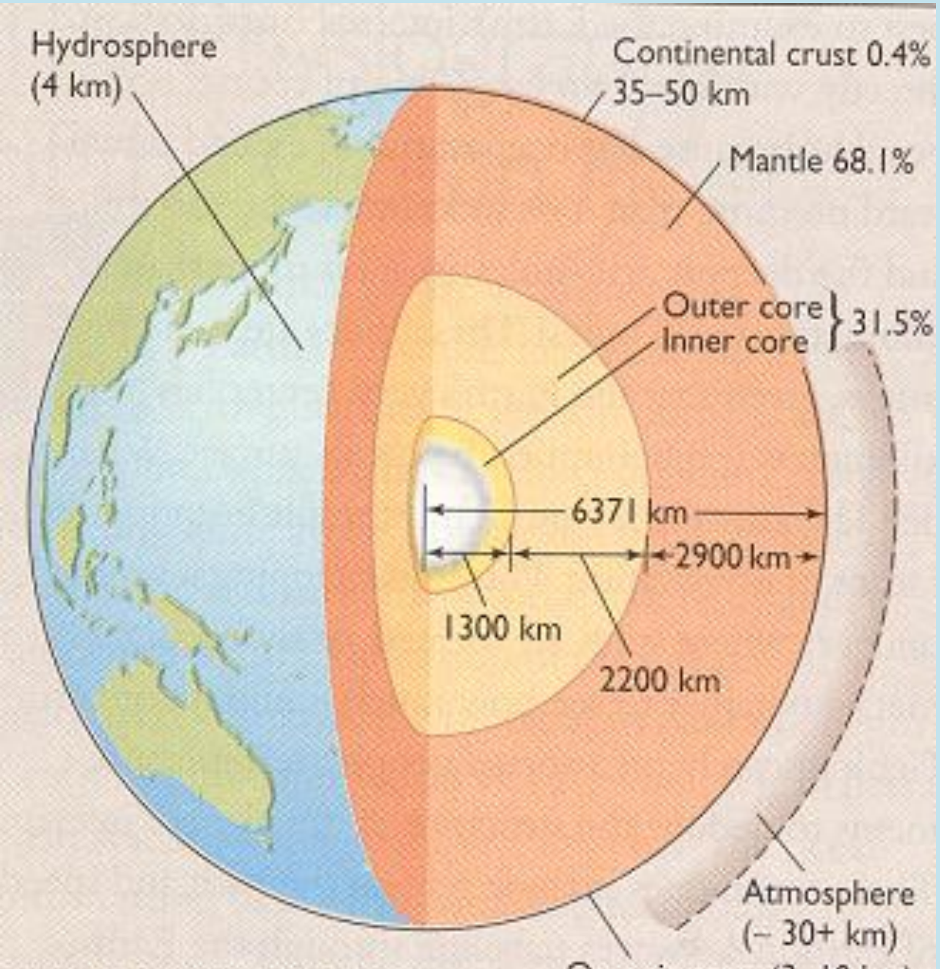
-> average density is $5.5\ 10^{3}$ kg/m^{3} (compare to density of rocks we find on the surface)

Magnetism: geodynamo

Seismology: seismic velocities and discontinuities

Geochemistry: the bulk composition of the Earth is close to the composition of meteorites

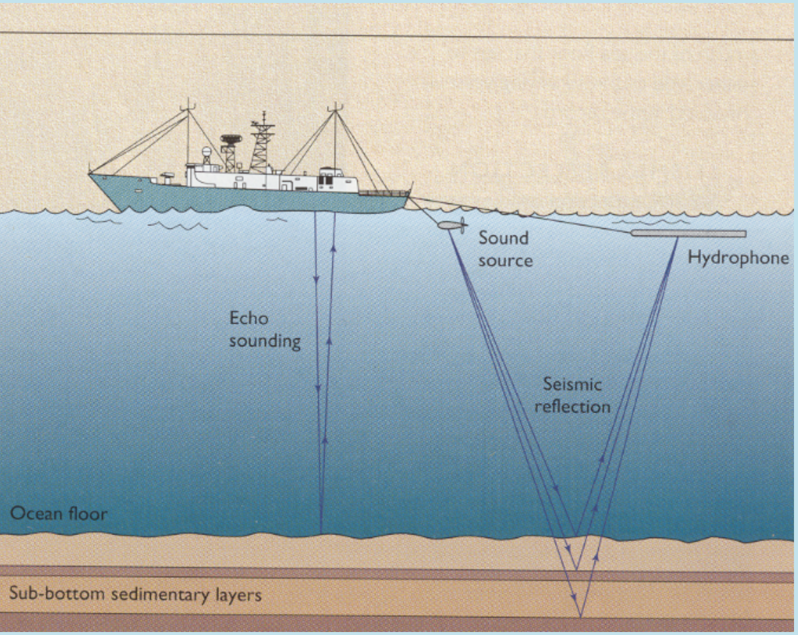
-> accretion from the solar nebula



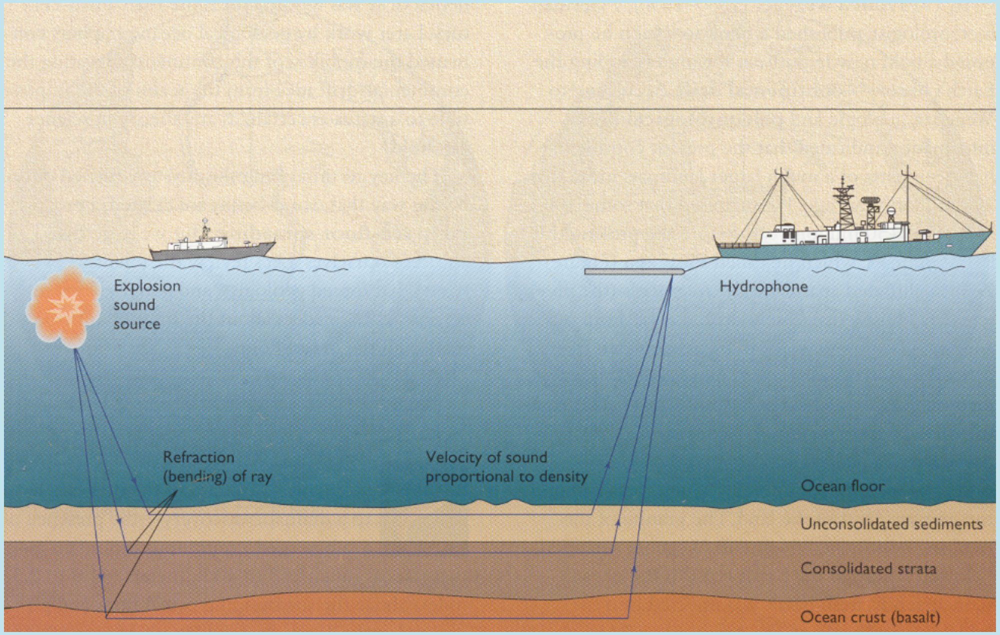
(a) THE EARTH'S STRUCTURE

Mapping the ocean crustal layers

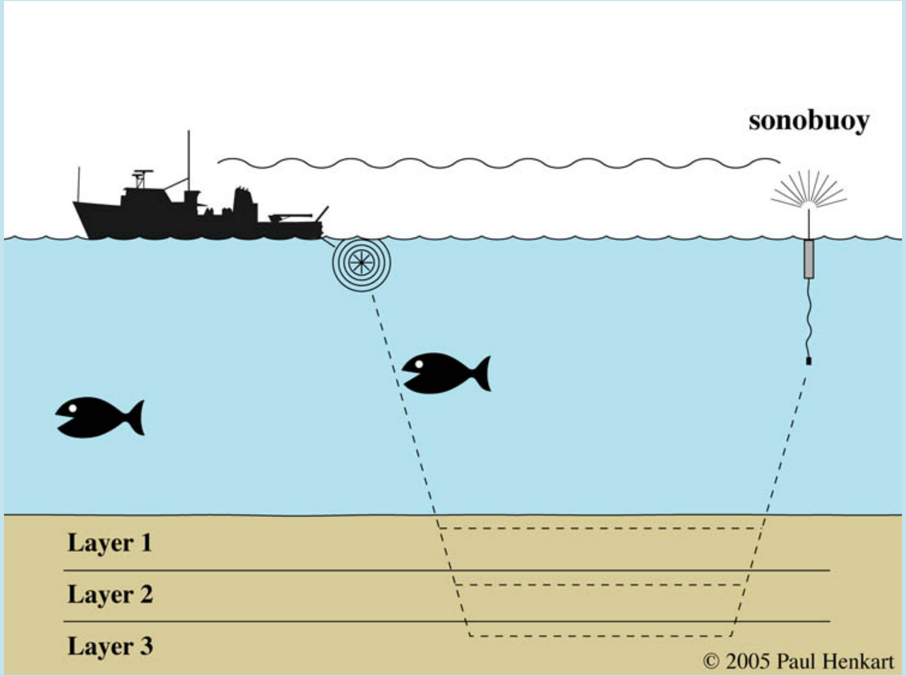
Seismic reflection



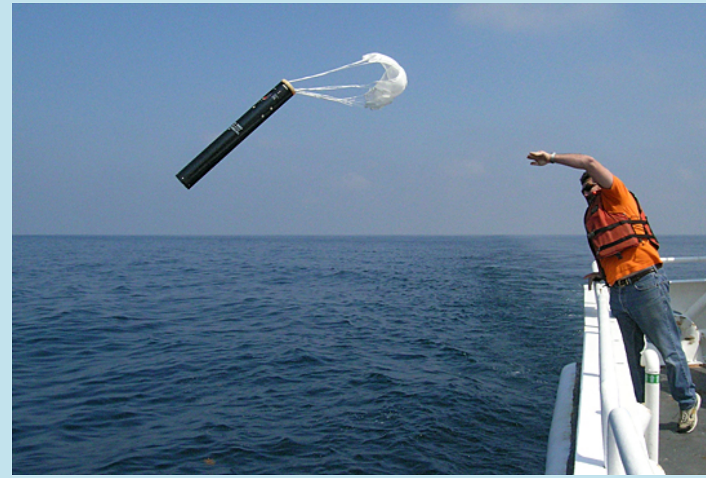
Seismic refraction



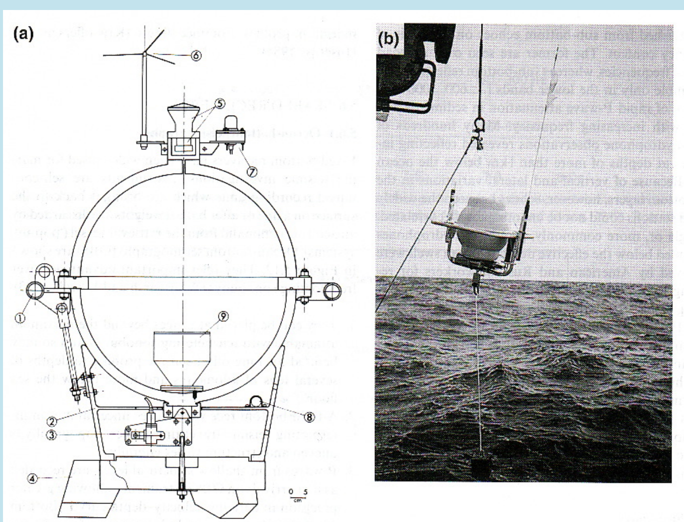
Seismic refraction



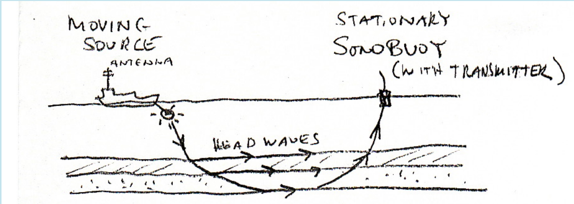
Sonobouys (3-4 hours)

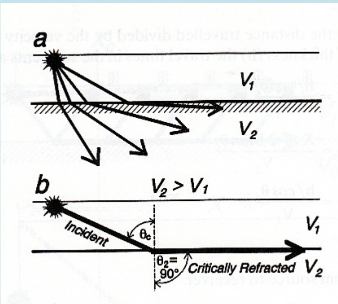


OBS: ocean bottom seismometer (months)



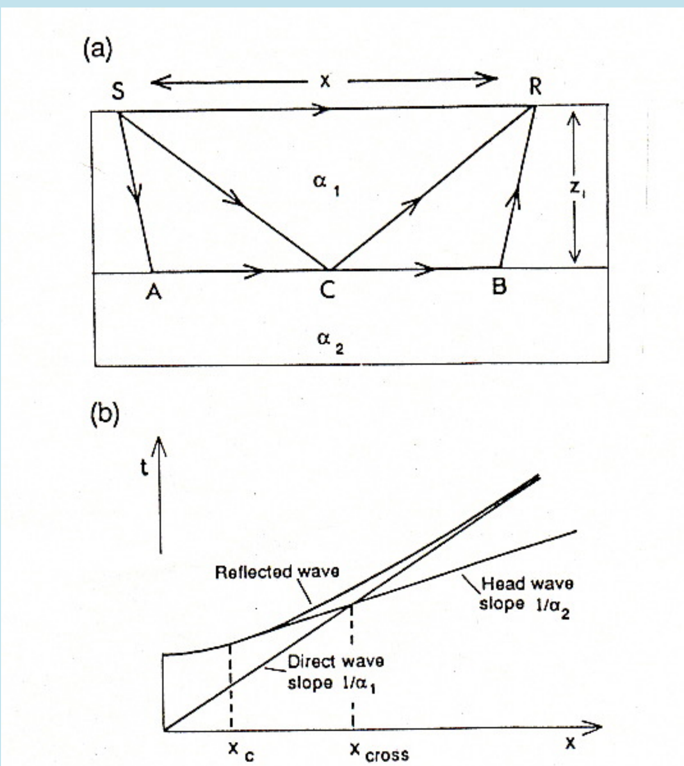
P3 Orion
Tail boom has
magnetometer

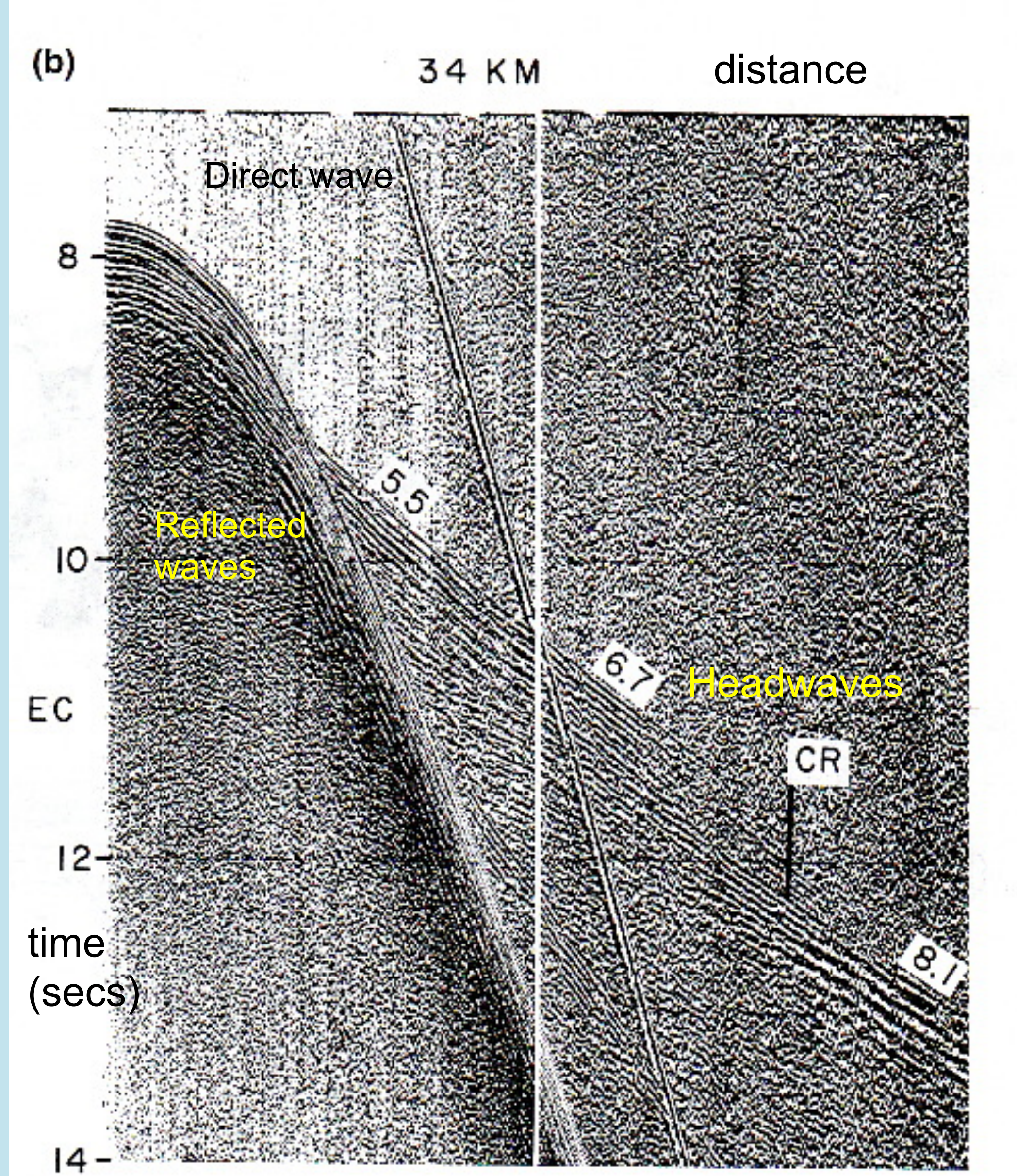




Seismic waves whose angle of refraction is 90° are known as headwaves

Headwaves travel at velocity of underlying layer; tangent at Xc

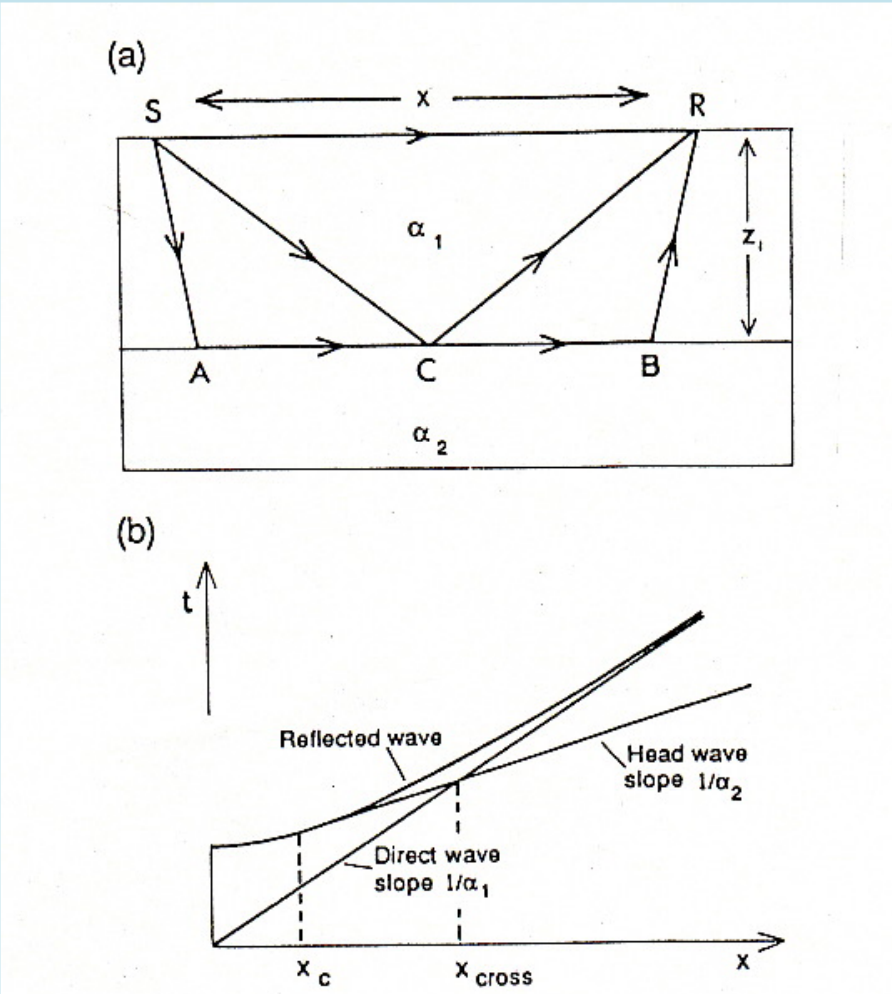




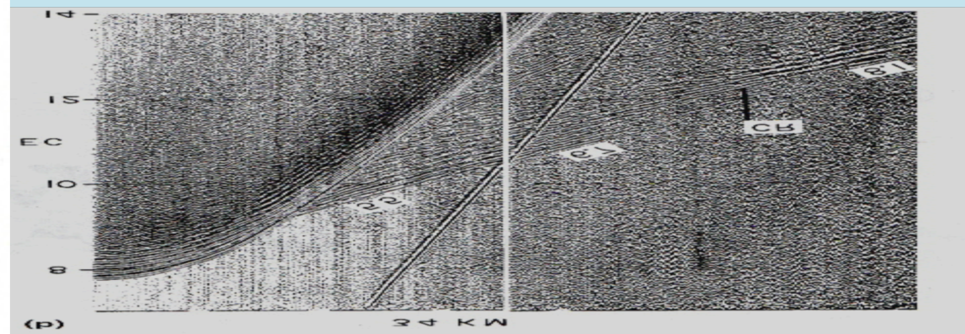
Upside down version of (b) left

Velocity = $\Delta x/\Delta t$

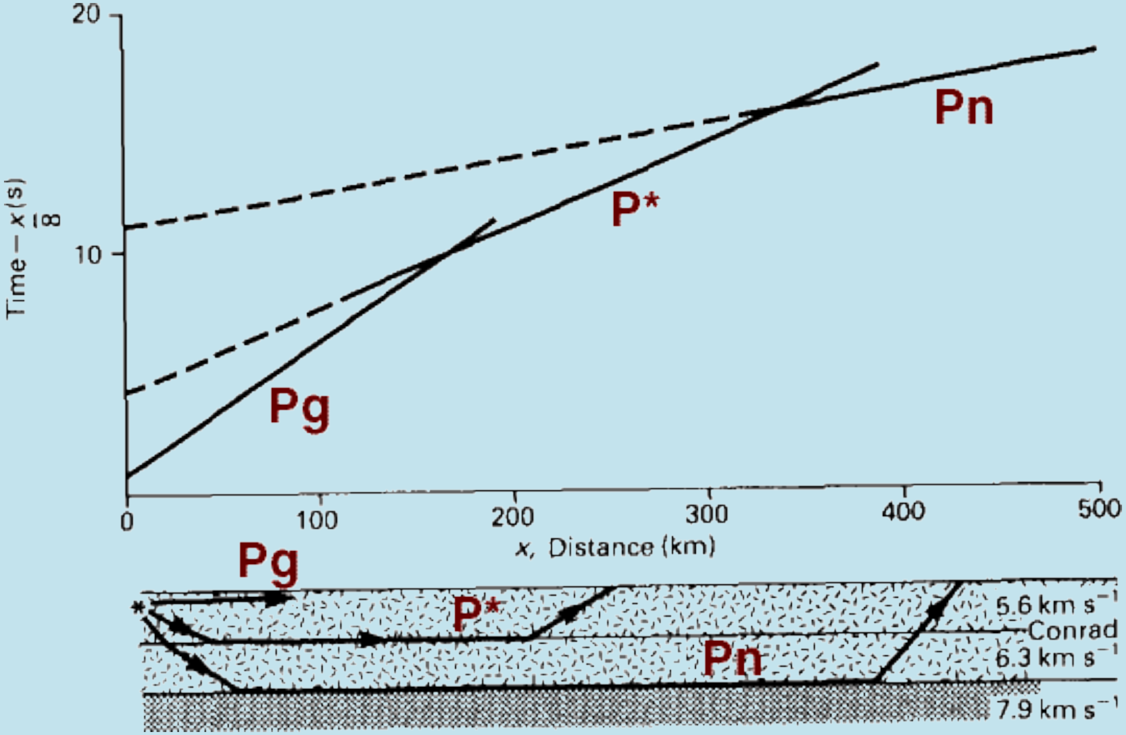
Classic velocities:
layer 3 = 6.7 km/sec
mantle = 8.1 km/sec



Inverted (and compressed) sonobuoy record



Y axis starts at 7 secs. Direct wave cut off at bottom



Continental crust has a lower velocity than oceanic layer 3

Typical lower continental crust is 6.3 km/sec

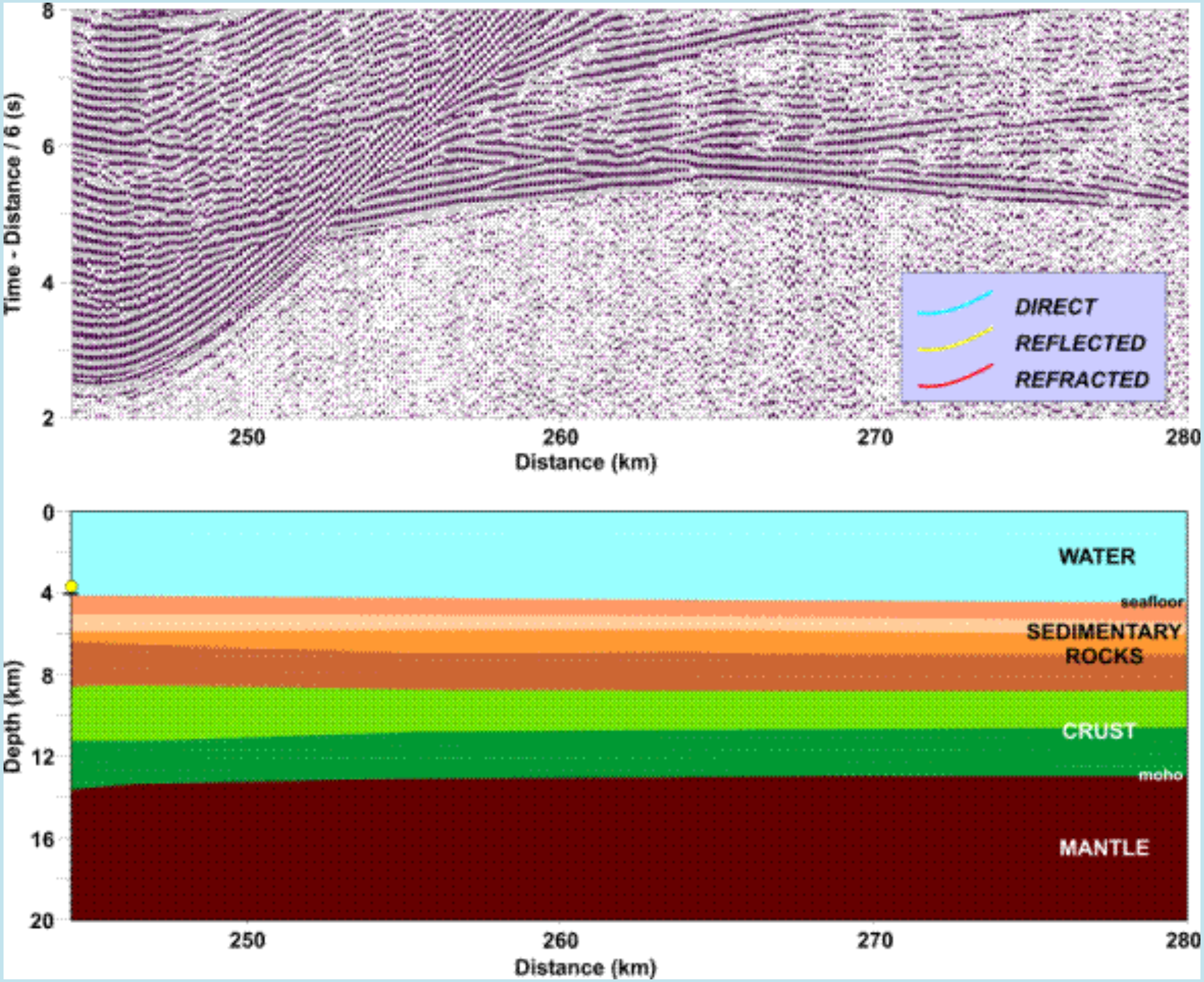
(oceanic layer 3 is 6.7 km/sec)

Another way of presenting results (KK&V)

"Reduced" time-distance plots; Y axis shows time – x(dis)/8

A refracted head wave with a velocity of 8 kms/sec will be horizontal

Here Pg is direct wave,
P* is mid-crust discontinuity, Pn is Moho

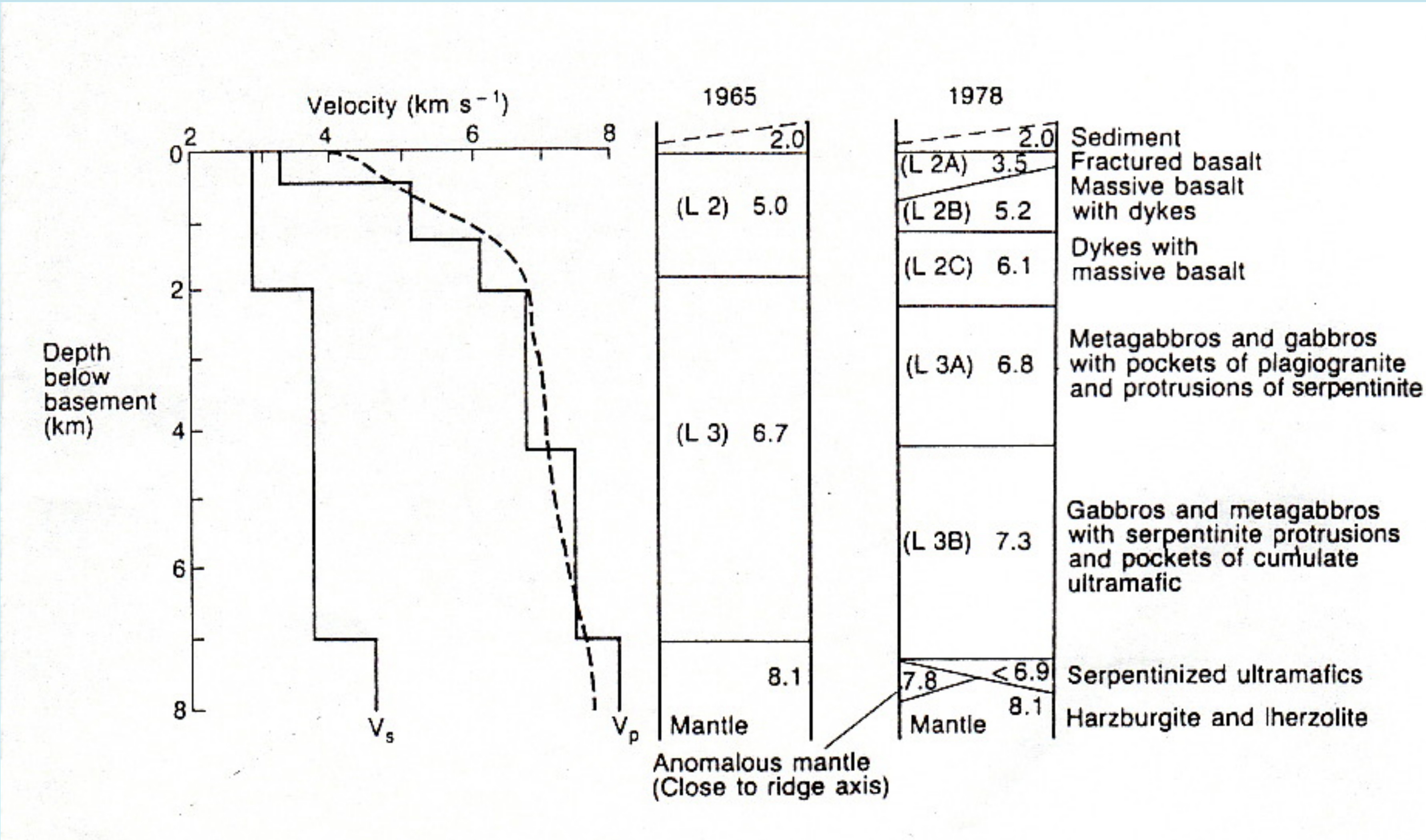


Another example of reduced time-distance plot for oceanic crust showing three strong refraction arrivals

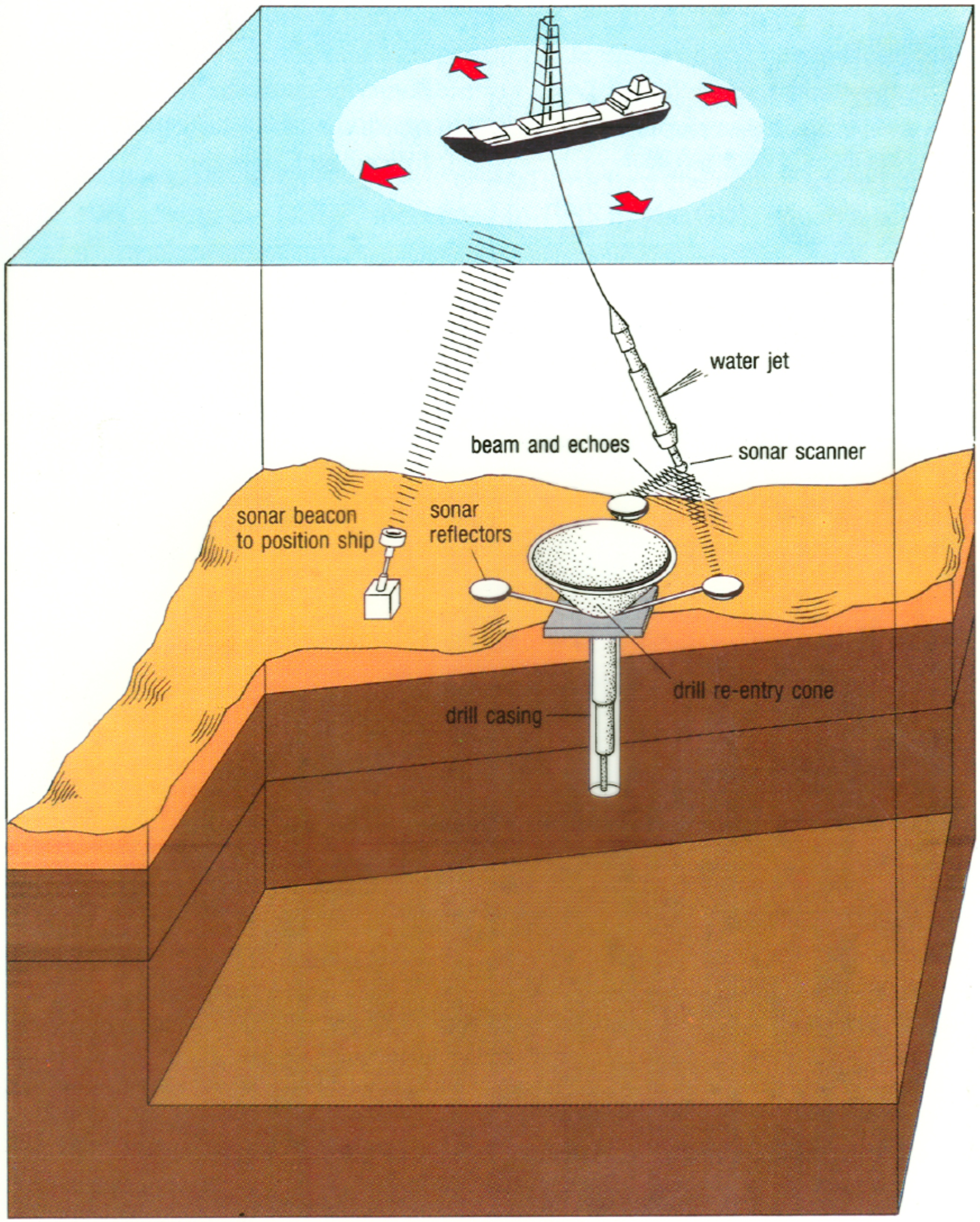
Mantle head wave has a velocity > 8 kms/sec

Structure of oceanic crust

Seismic refraction gives you layer thickness and velocity



KK&V



Drilling from the top of the oceanic crust has barely penetrated layer 2 and reached the top of layer 3. (about 1.5 km)

And took multiple legs, several months …

However, in places where layer 2 is tectonically missing, drilling has penetrated about 1.5 km of layer 3 (gabbro)

55th Anniversary of Project Mohole: 1961 – 2016

“Project Mohole represented the earth sciences' answer to the space program. If successful, this highly ambitious exploration of the intraterrestrial frontier would provide invaluable information on the earth's age, makeup, and internal processes.”

Goal was to drill to the crust/ mantle boundary (Moho) in the ocean where crust is thinnest; a step too far …

Ophiolites

KK&V 2.19

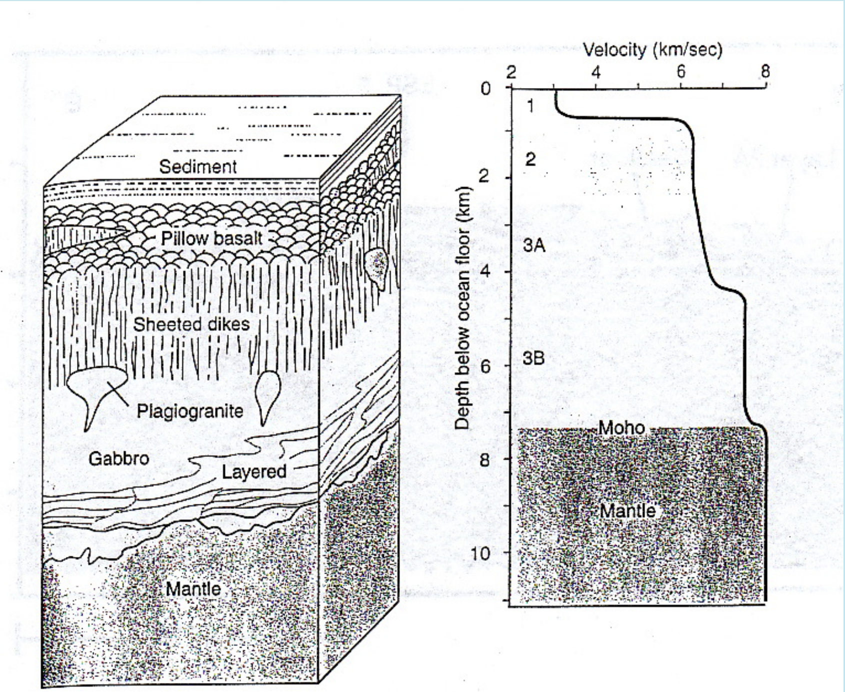
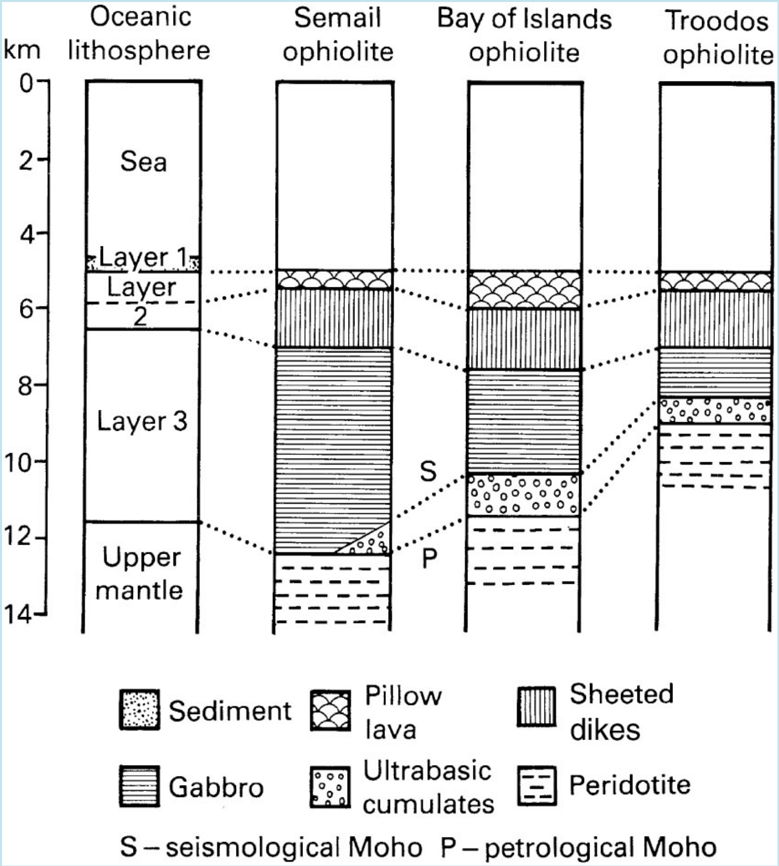
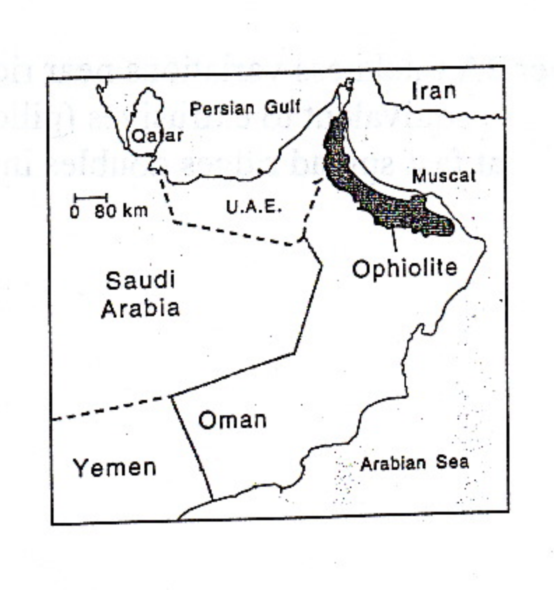


FIGURE 19.12 The major rock units in an ophiolite sequence

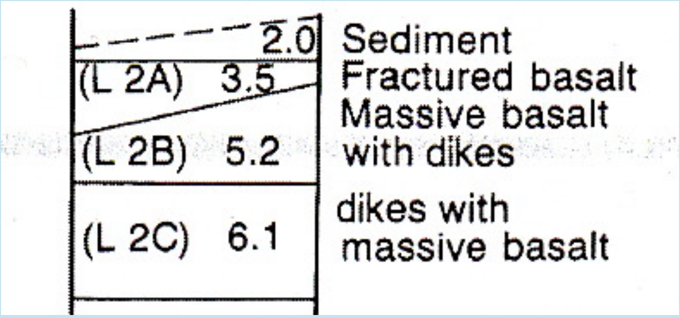




Semail ophiolite
In Oman

Classic in So Calif is the
Point Sal ophiolite (Jurrasic age)
in Vandenburg AFB
- e.g. outcrop of dikes (layer 2b)

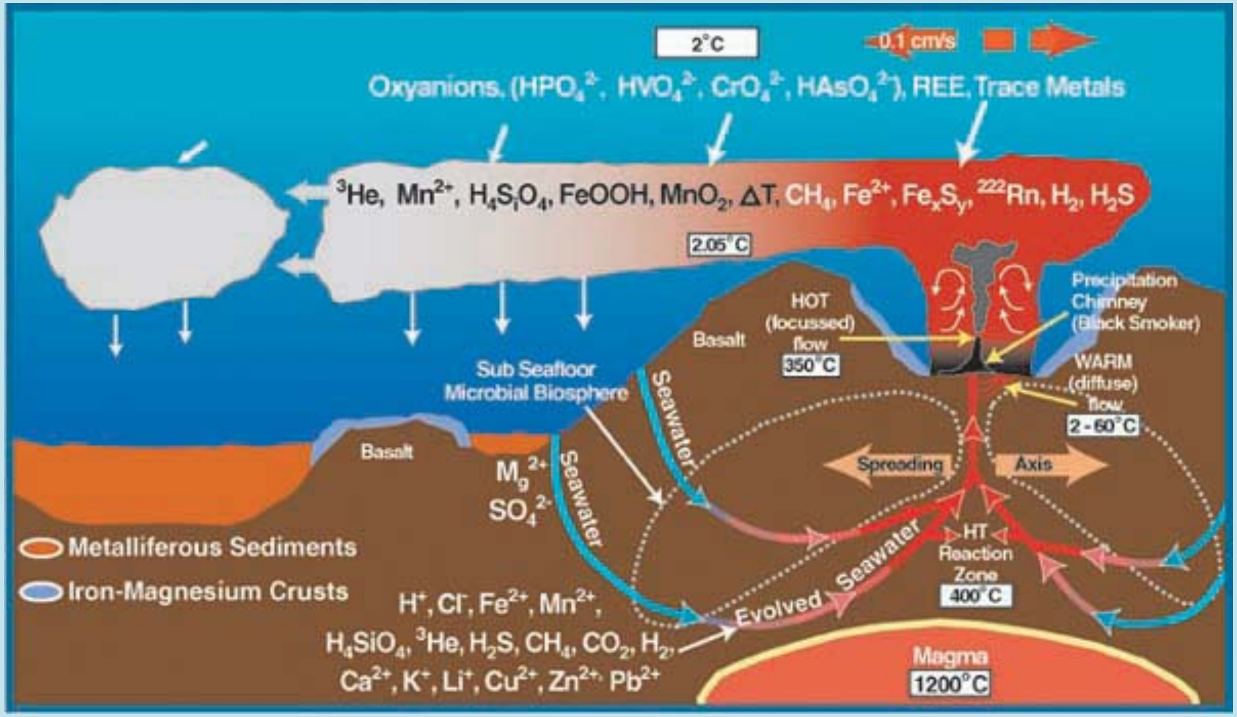
CSUN



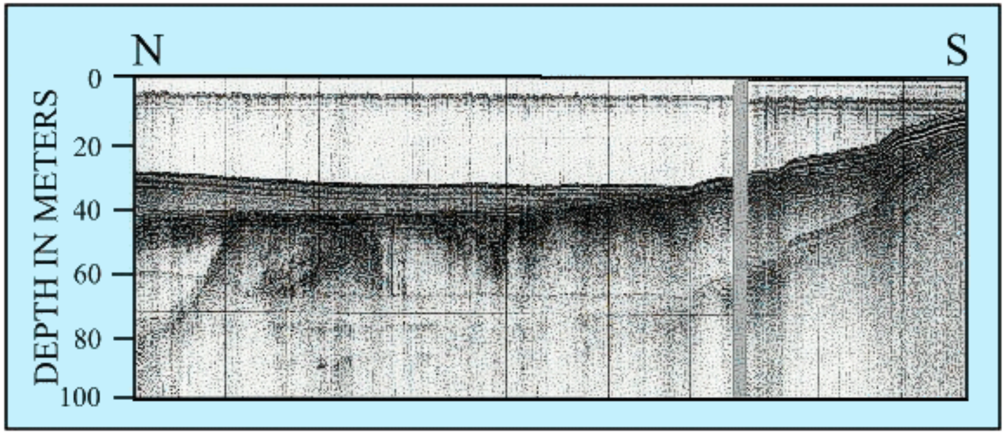
Velocities in the upper seismic layers (2a and 2b), increase over the first few million years. This is generally attributed to seawater circulation in the upper 3 km of the crust leading to metamorphism of the extrusive rocks and the gradual filling of fissures and cracks.

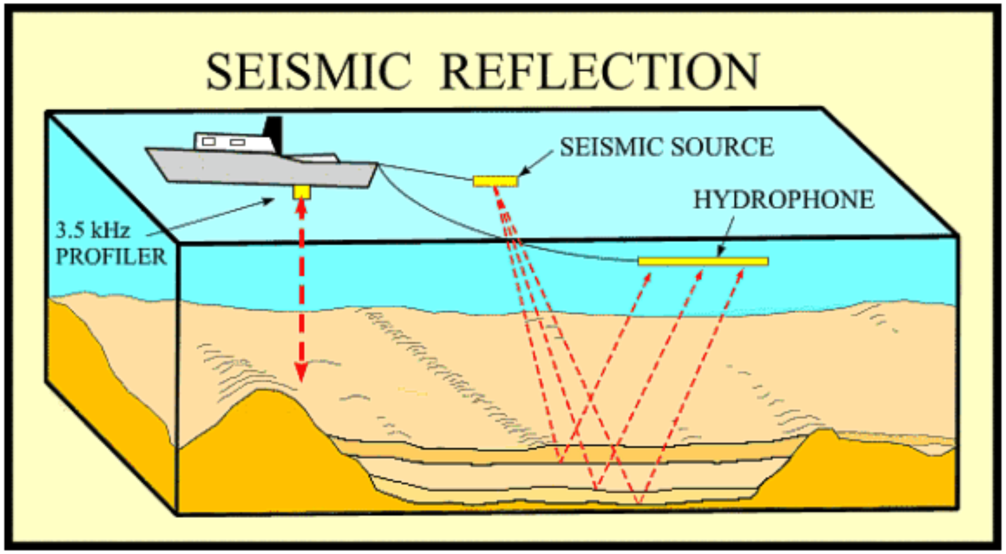
Draw down of seawater through fissures on flanks

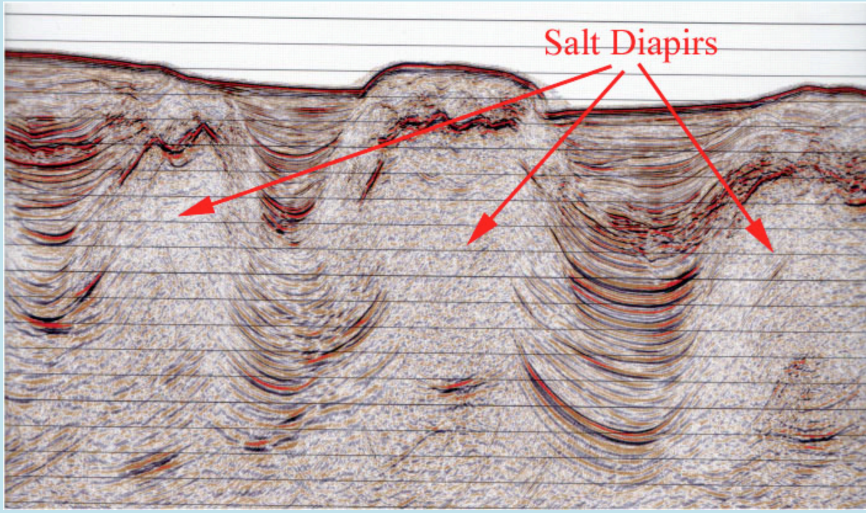
Outflow of seawater through black smokers



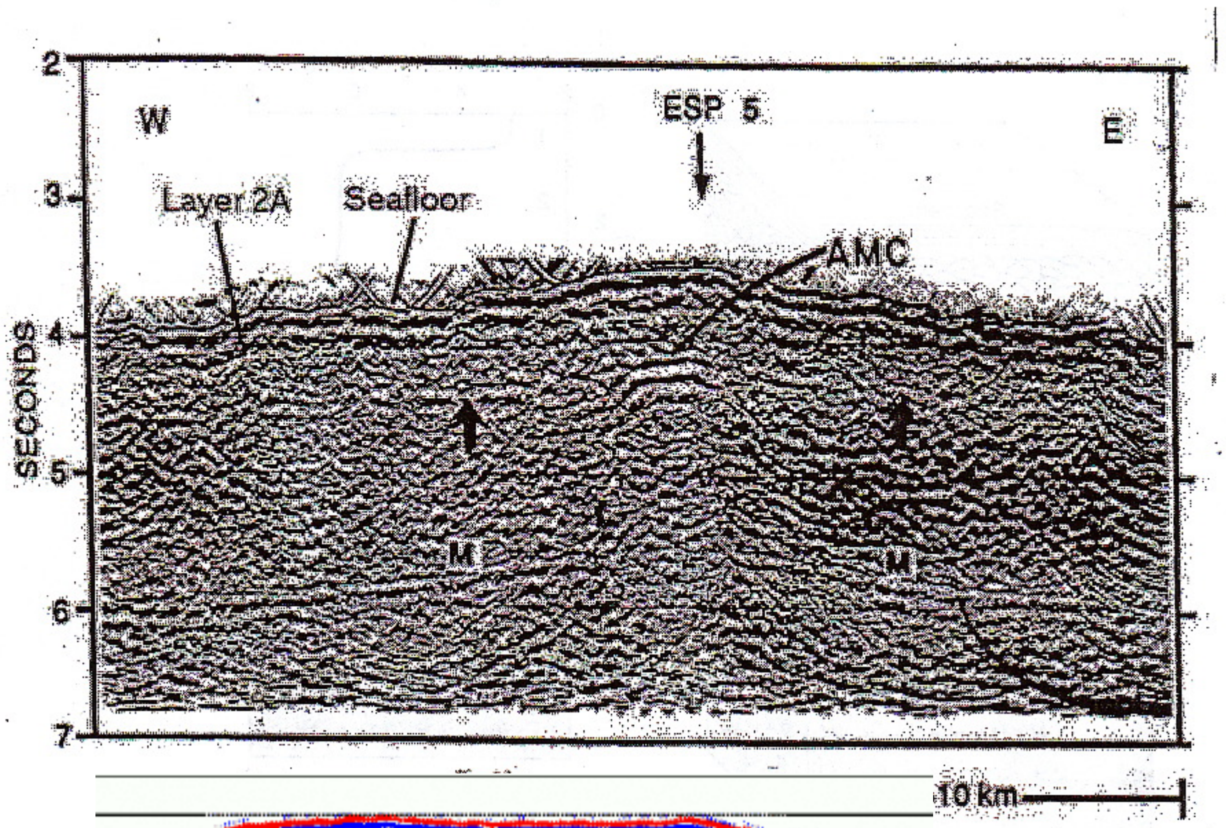
Seismic reflection: vertical (or near vertical) reflections
Mainly used to explore sedimentary basins

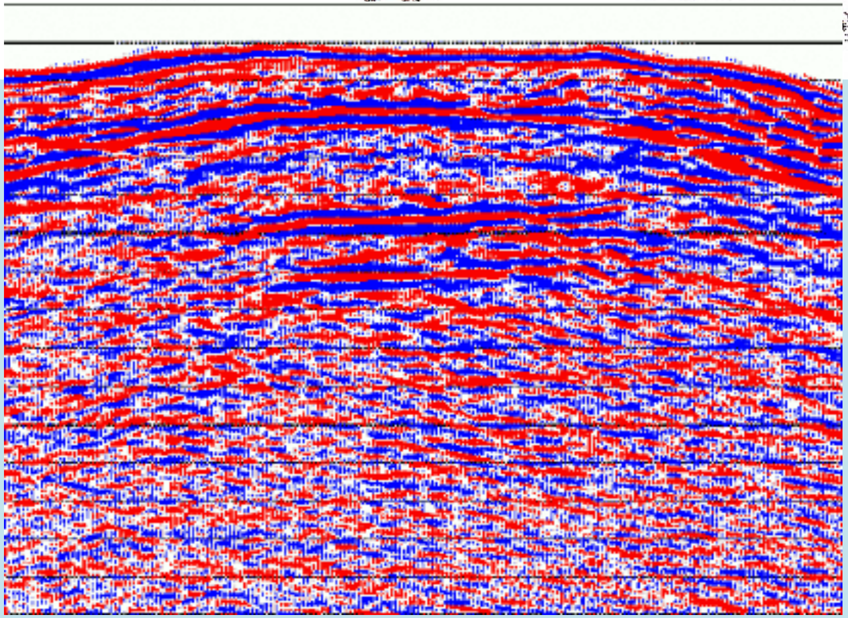






Can image very complex structures

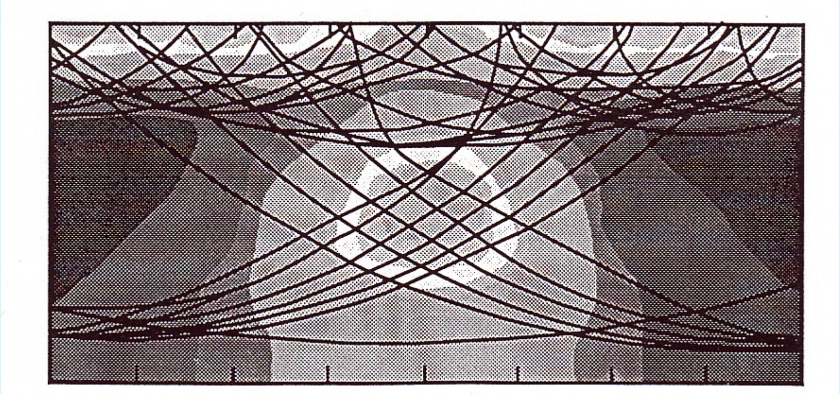




Application of seismic reflection to ridge axis:

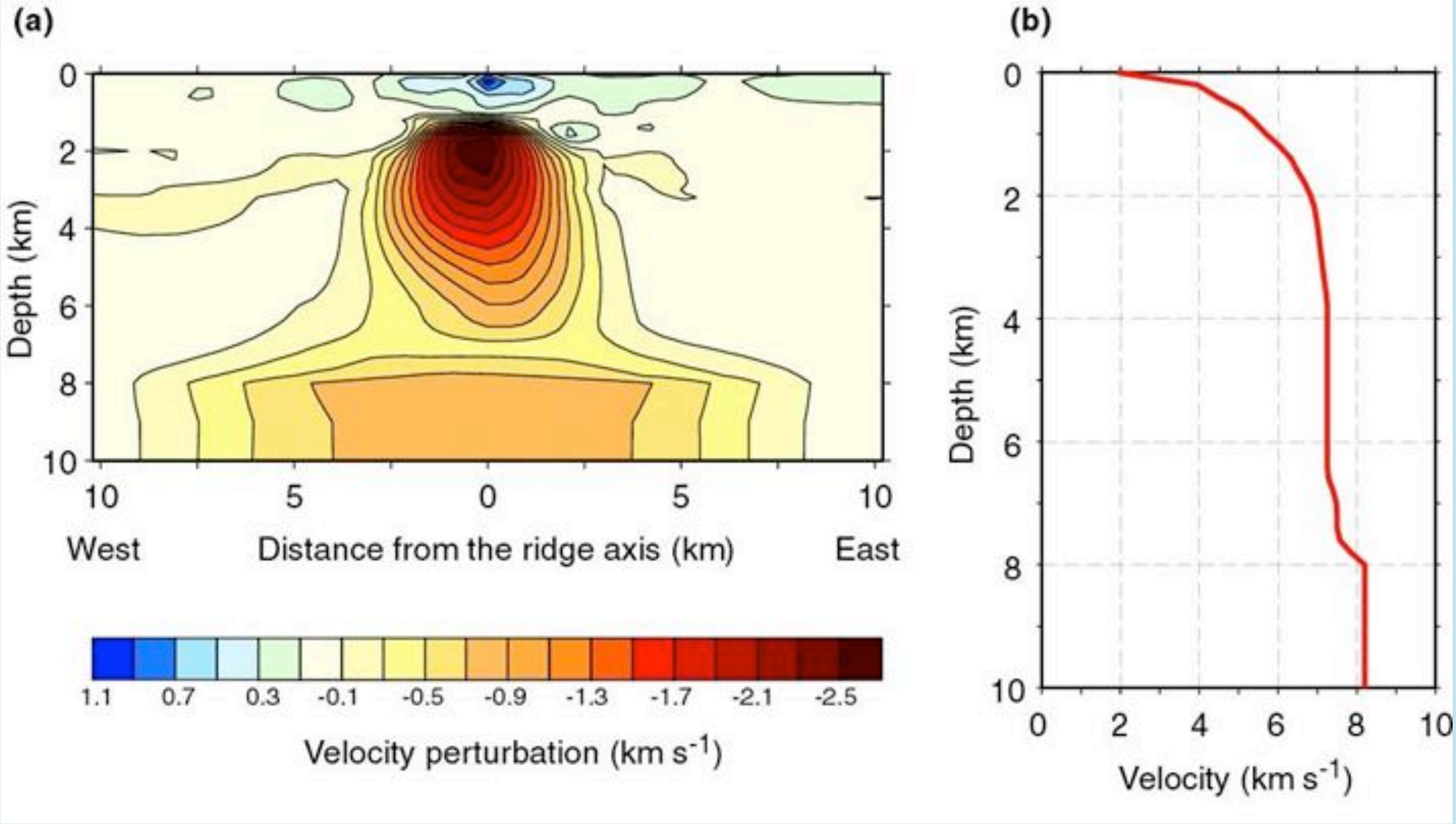
Can image

1) boundary between layers 2a and 2b
2) top of mantle (Moho)
3) top of axial magma chamber (reversed polarity reflection from liquid magma lens)

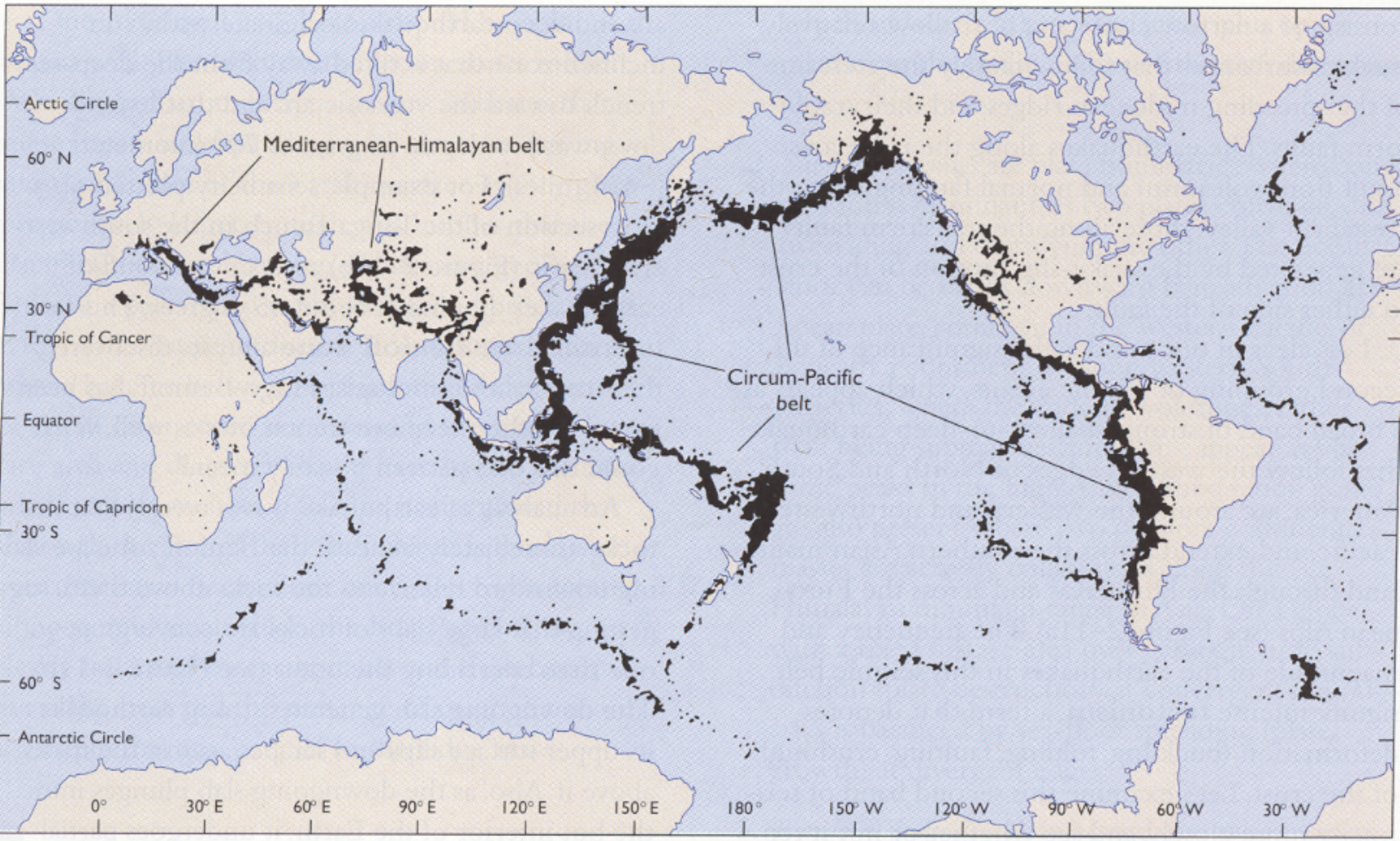


Seismic tomography:

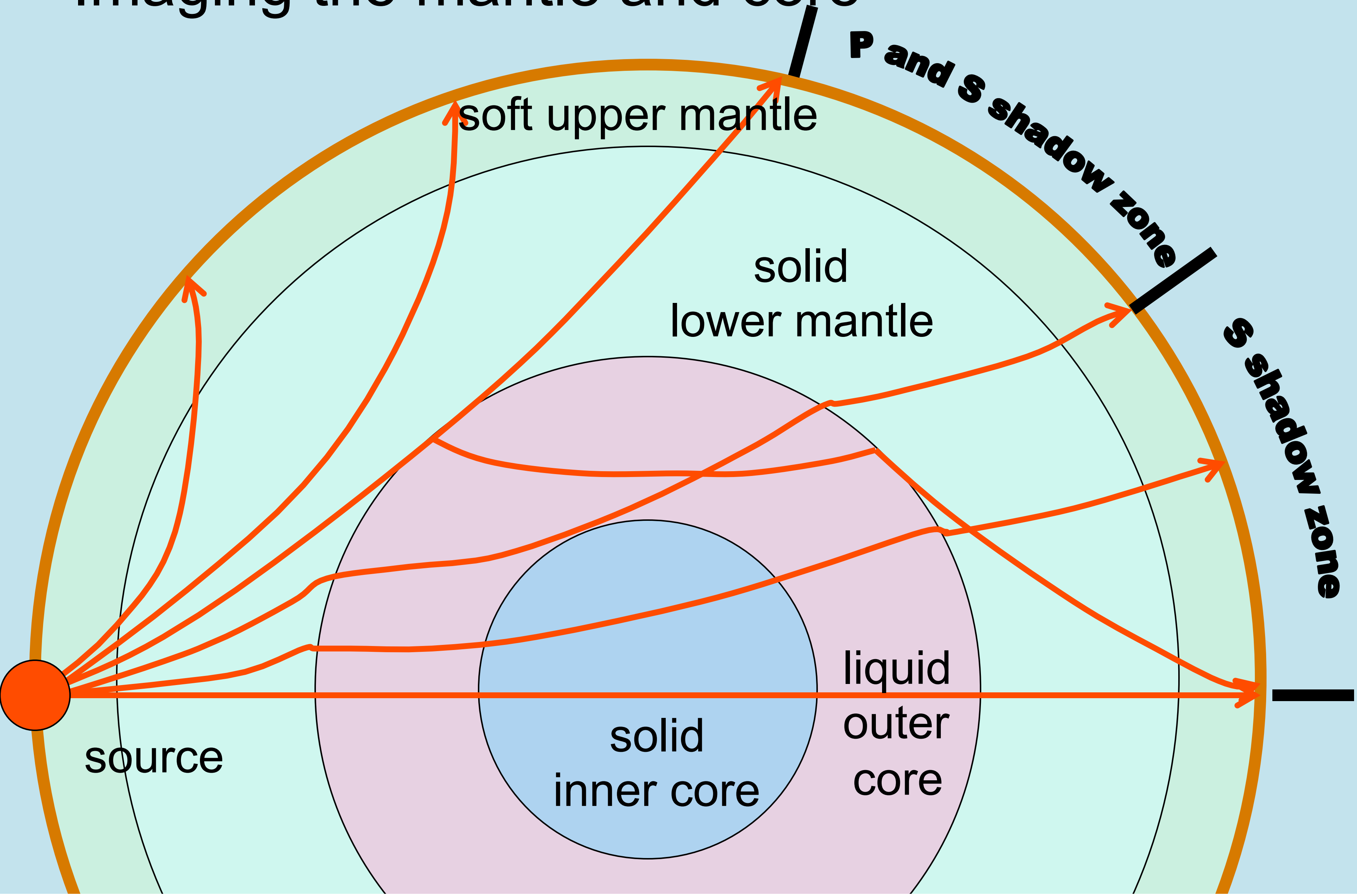
many sources, many receivers = detailed velocity models of ridge axis



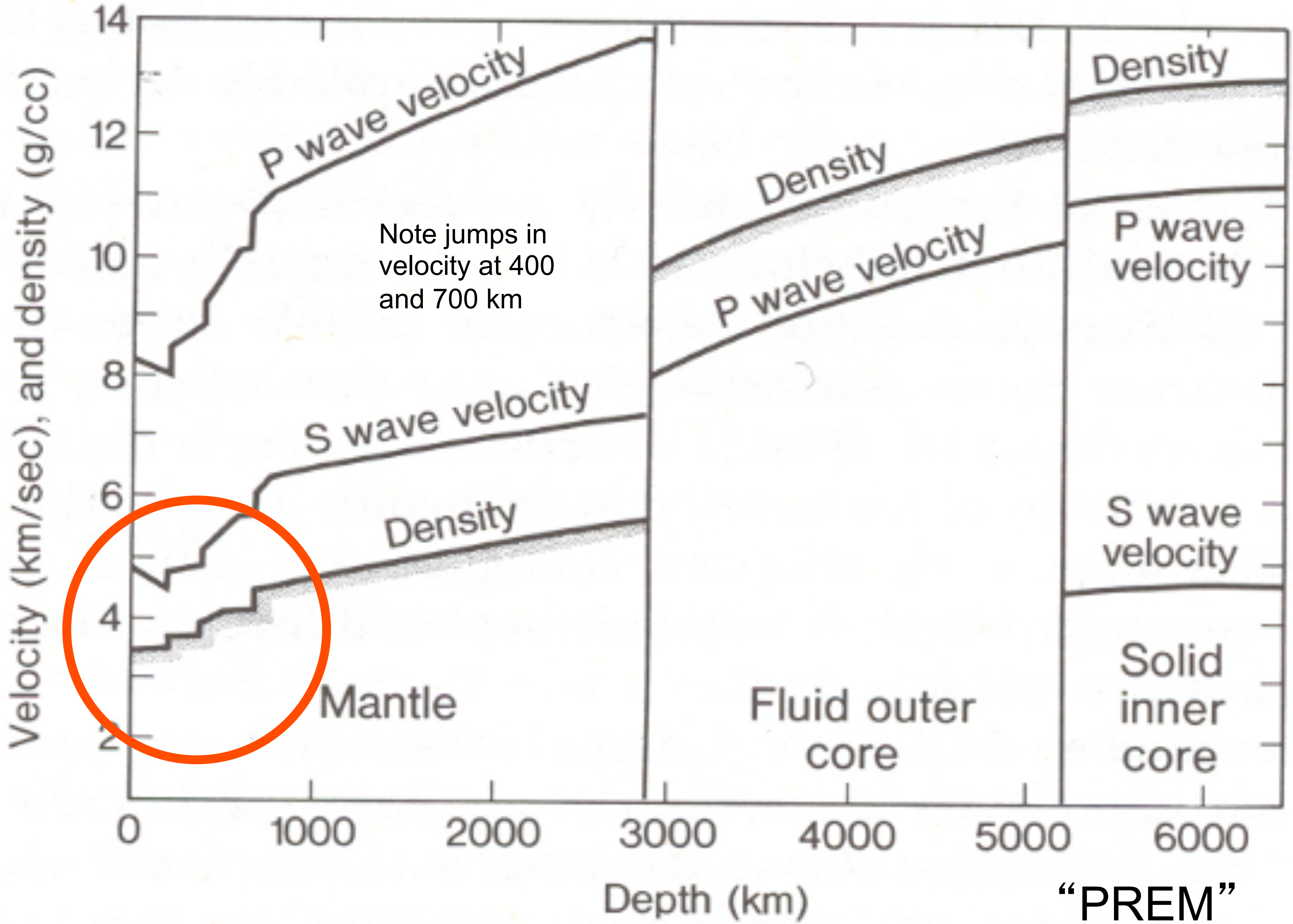
imaging earth's interior with earthquakes



Imaging the mantle and core

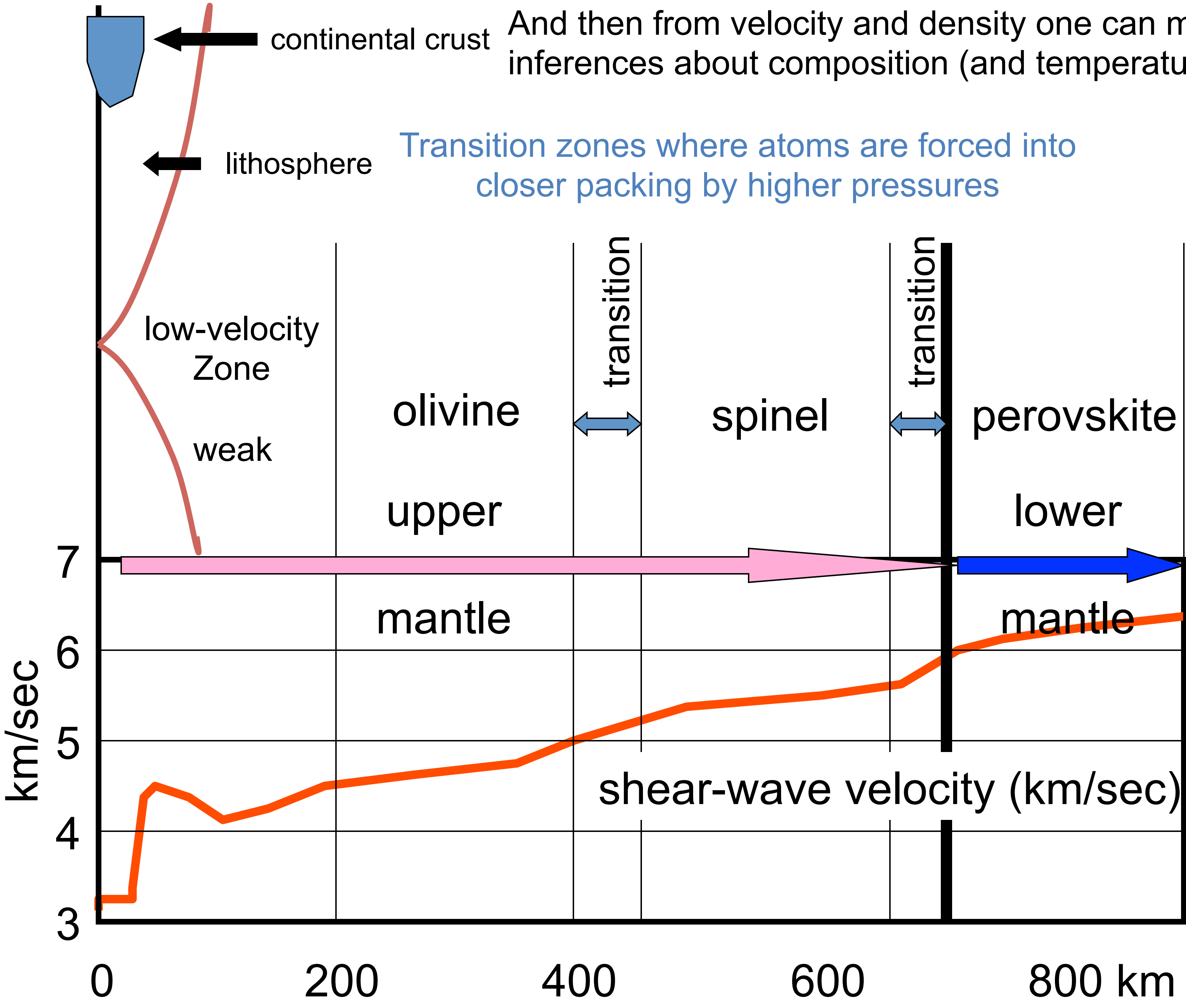


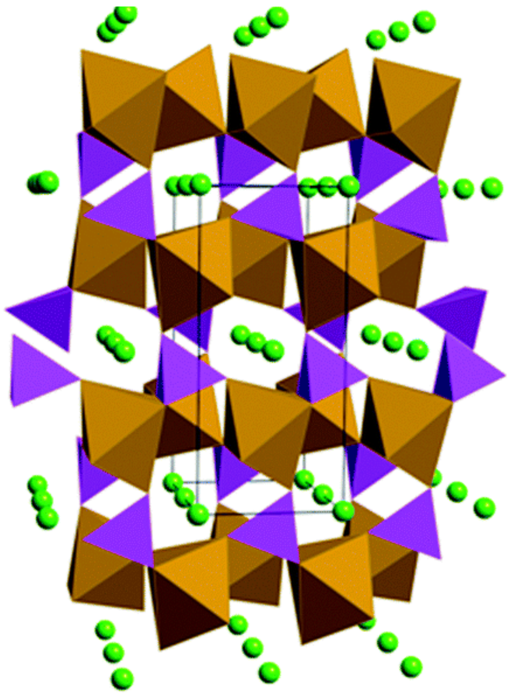
For example a velocity structure and from it density



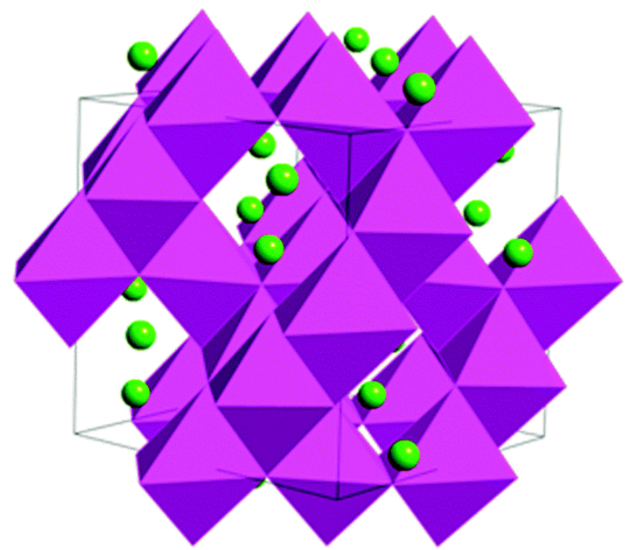
"PREM"

And then from velocity and density one can make inferences about composition (and temperature)



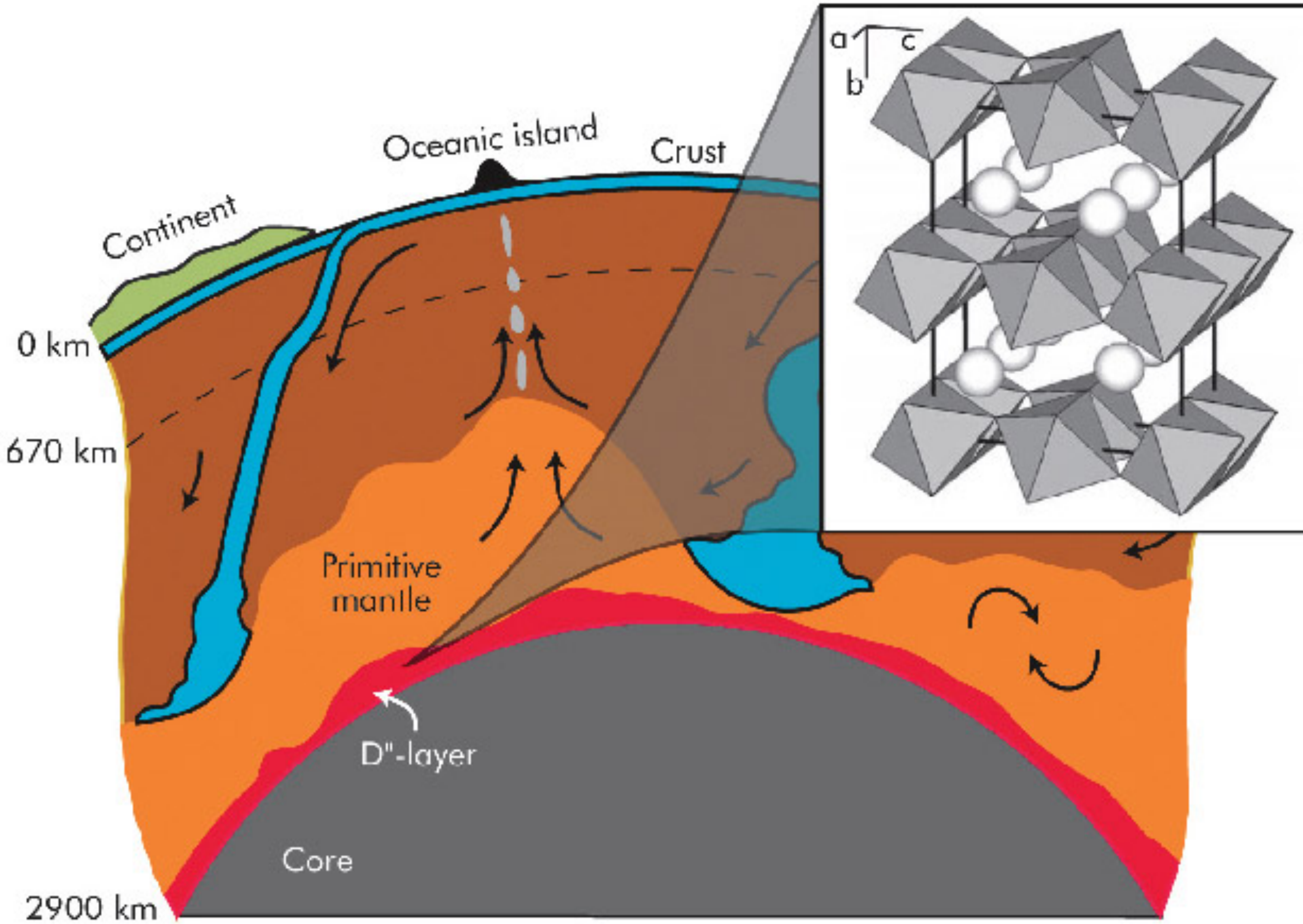


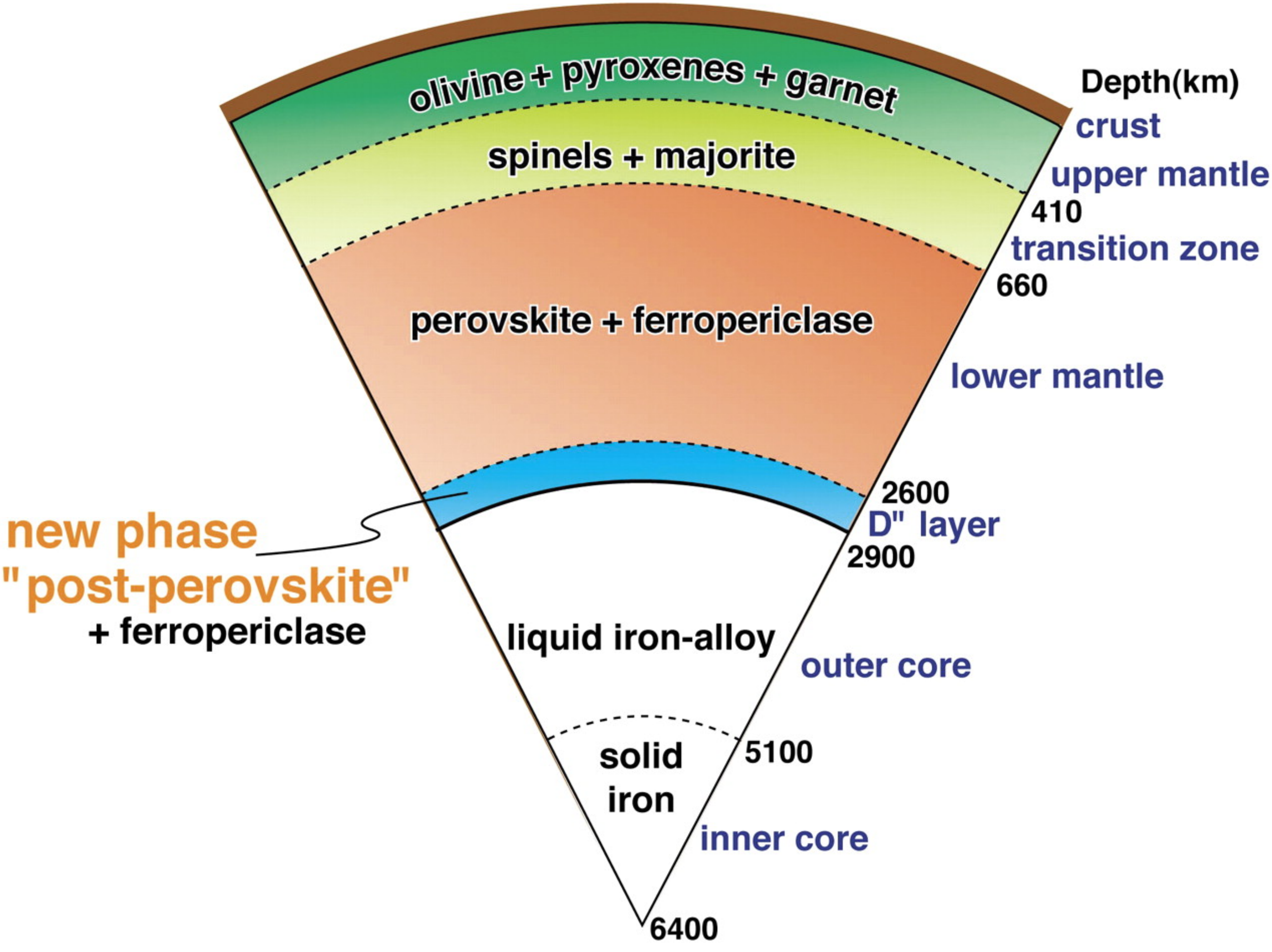
olivine



spinel

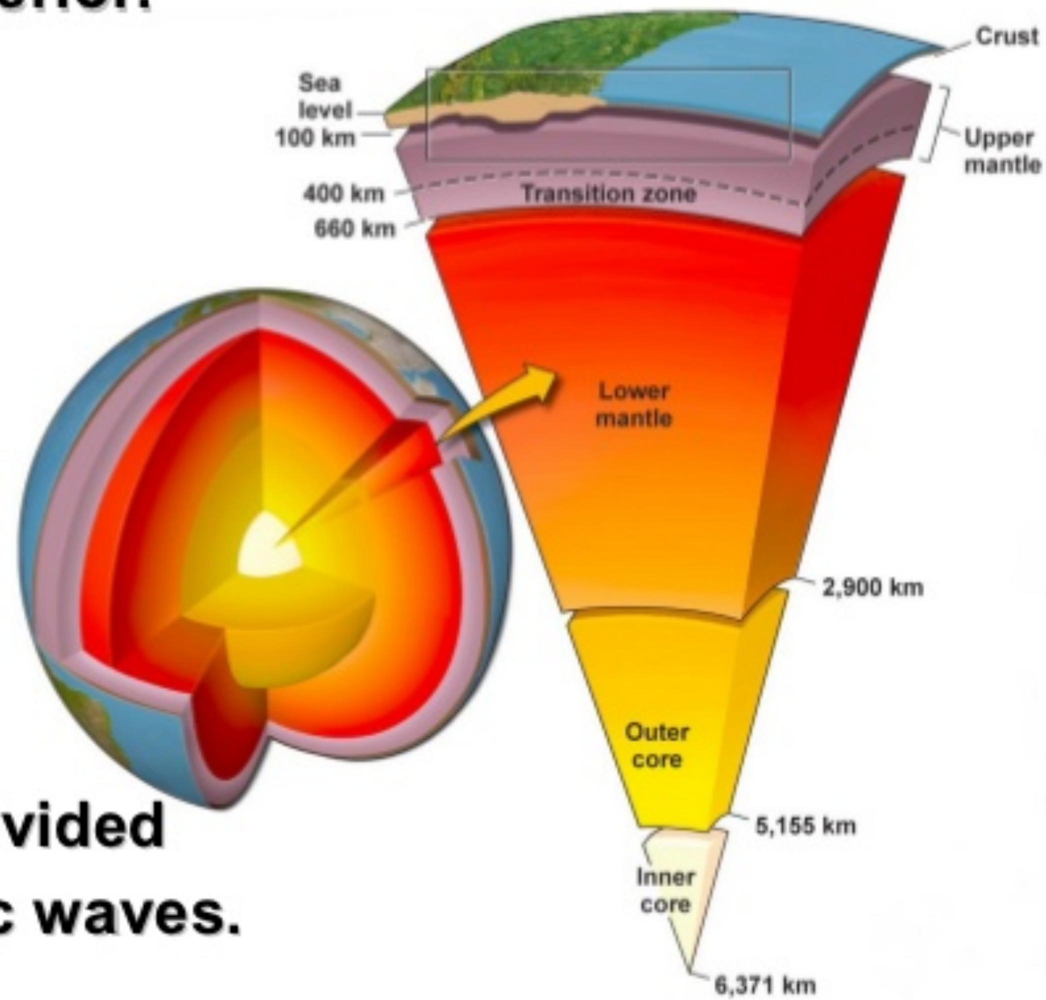
… more recently: post-perovskite transition at the CMB




olivine + pyroxenes + garnet
spinels + majorite
perovskite + ferropericlase
new phase
"post-perovskite"
+ ferropericlase
liquid iron-alloy
solid
iron
Depth(km)
crust
upper mantle
410
transition zone
660
lower mantle
2600
D" layer
2900
outer core
5100
inner core
6400

Summarizing Earth's Layers

- Earth has a layered interior.
  - Crust
    - Continental
    - Oceanic
  - Mantle
    - Upper
    - Transitional
    - Lower
  - Core
    - Outer—liquid
    - Inner—solid
- These layers are subdivided on the basis of seismic waves.



Rheological layering

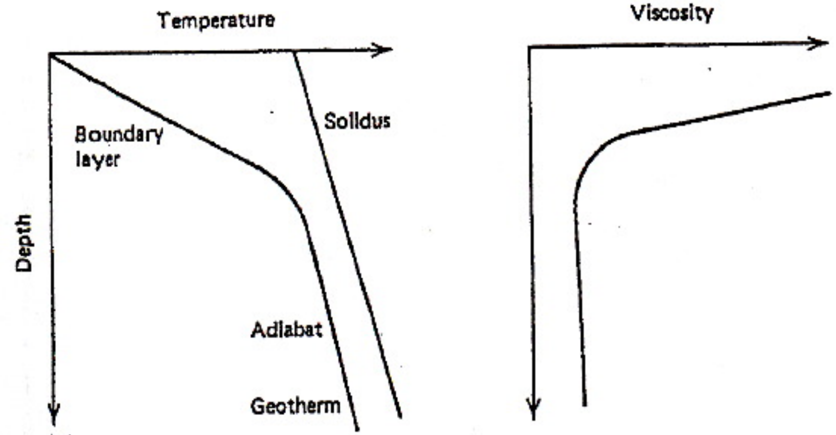
The strong dependence of rheology on temperature, together with the thermal structure of the earth, leads to a pronounced rheological layering

Lithosphere (Greek lithos = rock)
rigid plate of finite thickness, typically on the order of 100 km) that can support both horizontal and vertical stress
the “plate” of plate tectonics;
the crust is embedded in the lithosphere
conductively cooled “lid” over convecting mantle

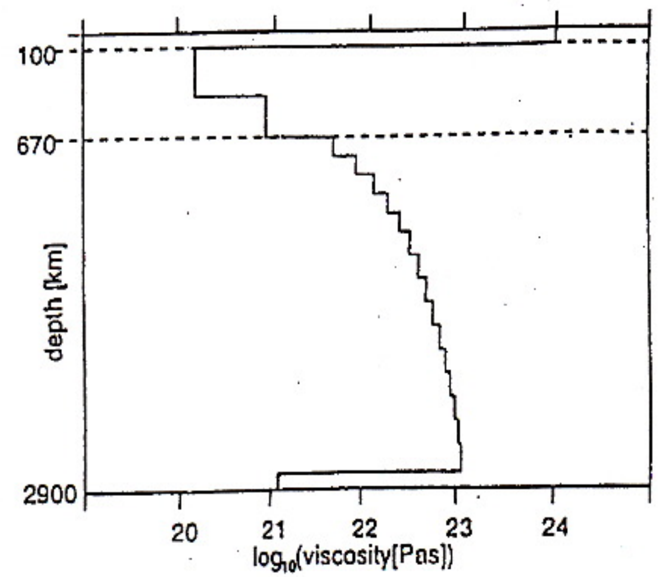
Asthenosphere (Greek asthenia = weak)
Solid but weak, deformable (fluid on long timescales) upper mantle that extends for some 200 -300 km below the lithosphere; seismic low velocity zone (LVZ) - few % melt

Mesosphere
Solid mantle between the asthenosphere and CMB which is less deformable than astheosphere, but with viscosities low enough to allow convection over long time scales

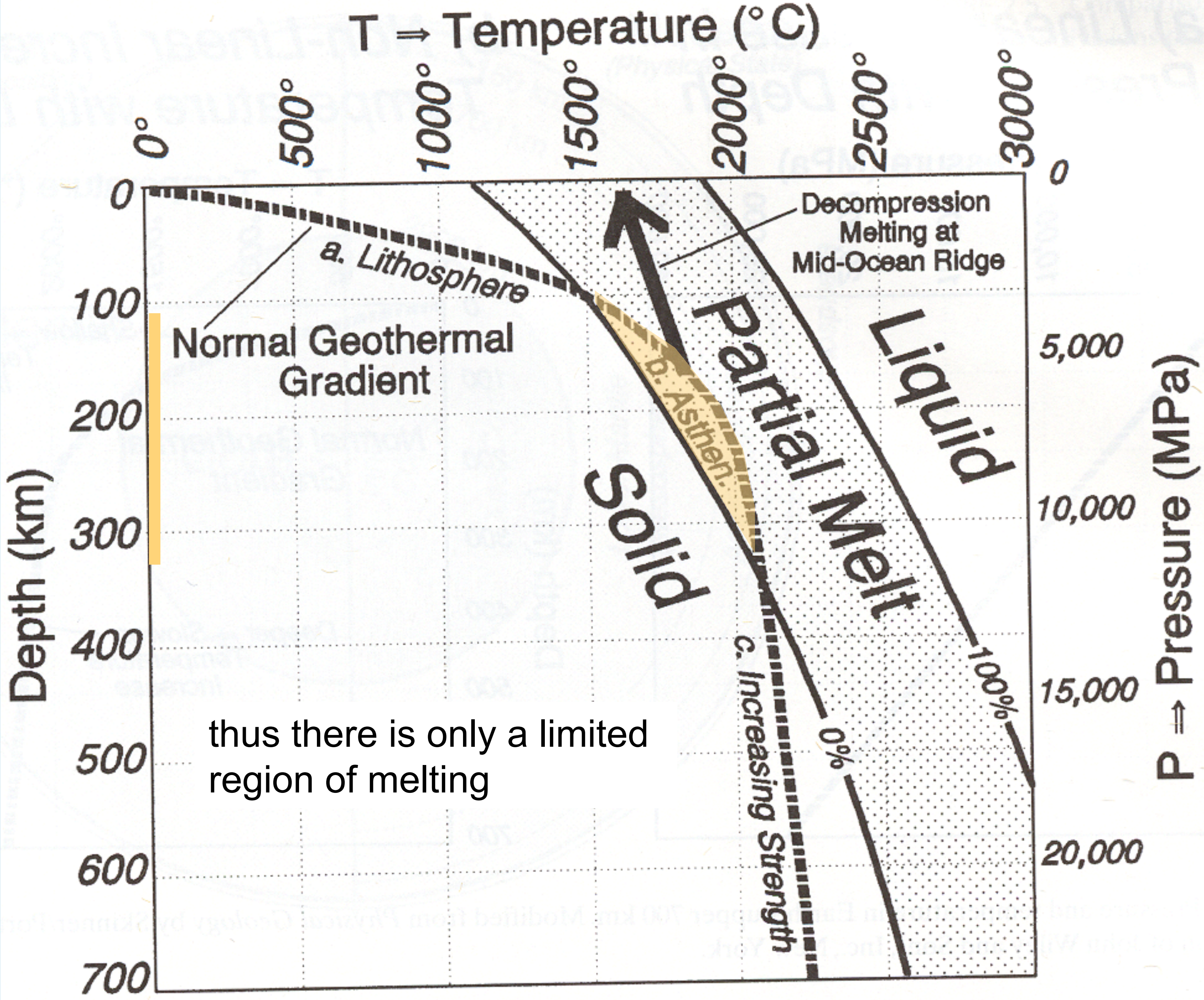
Outer core (2885 – 5145 km) – liquid
Inner core (> 5145 km) - solid



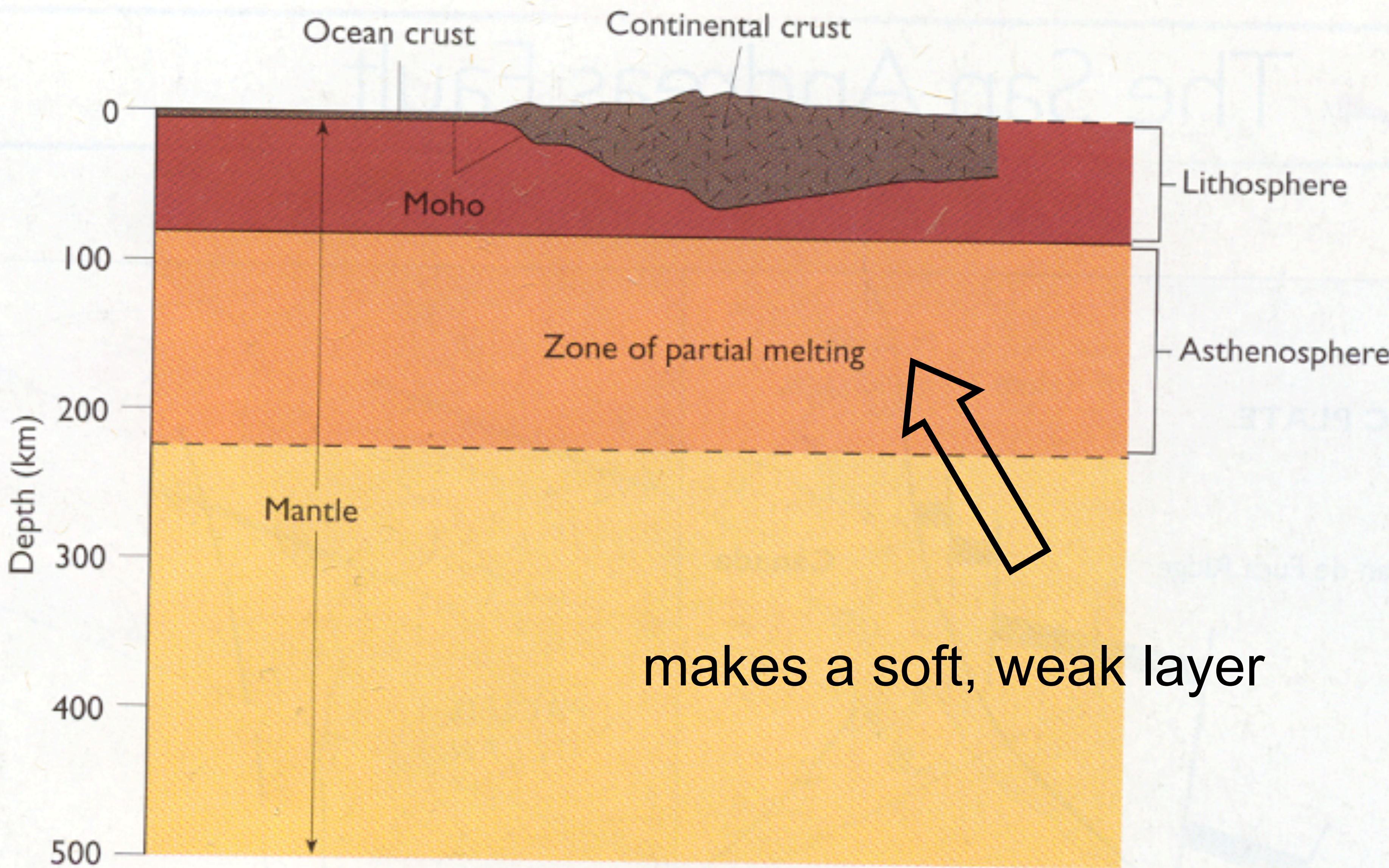
*from Turcotte and Schubert, 2002, Geodynamics



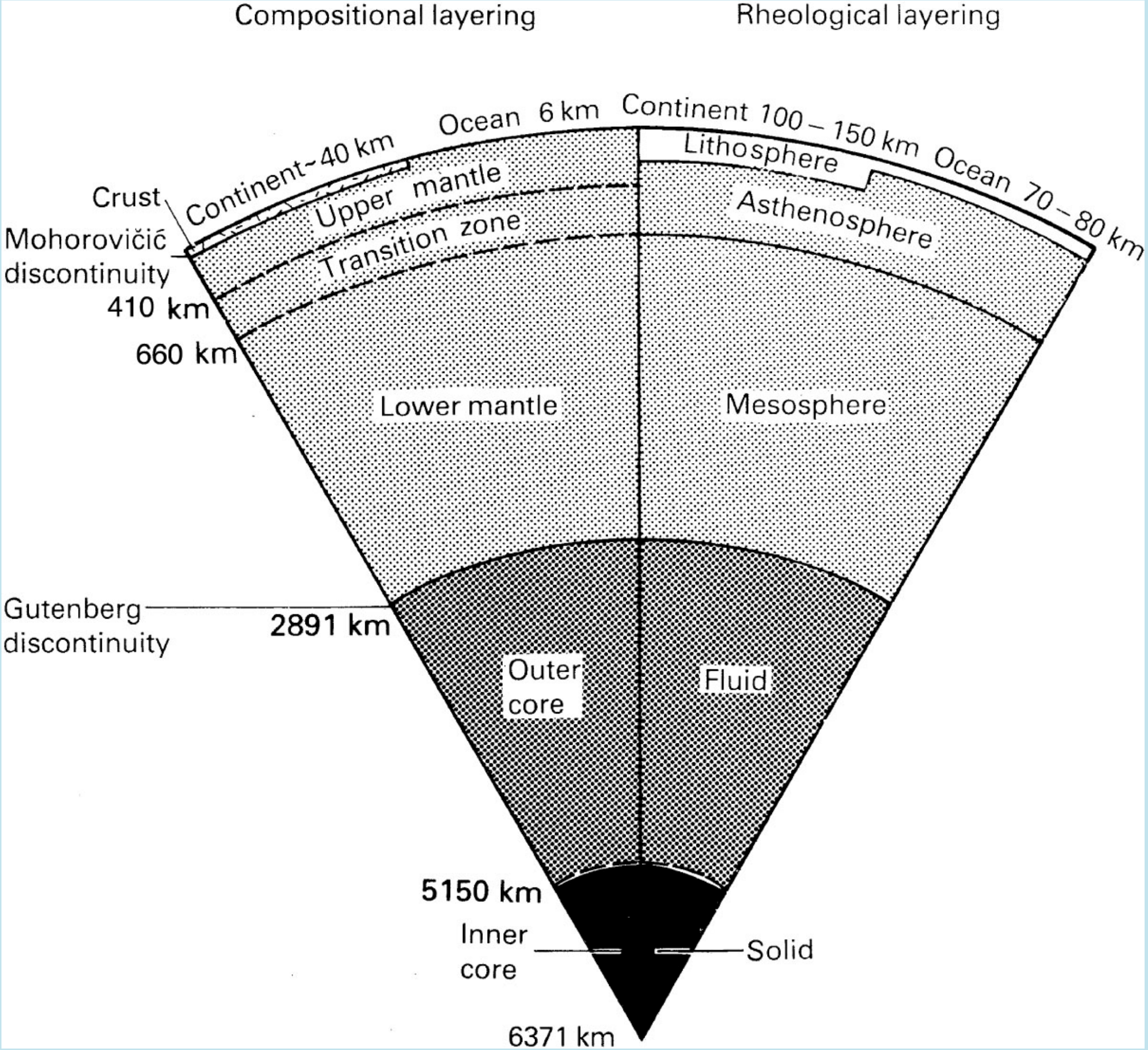
Steinberger and O’Connell (2000)



thus there is only a limited region of melting

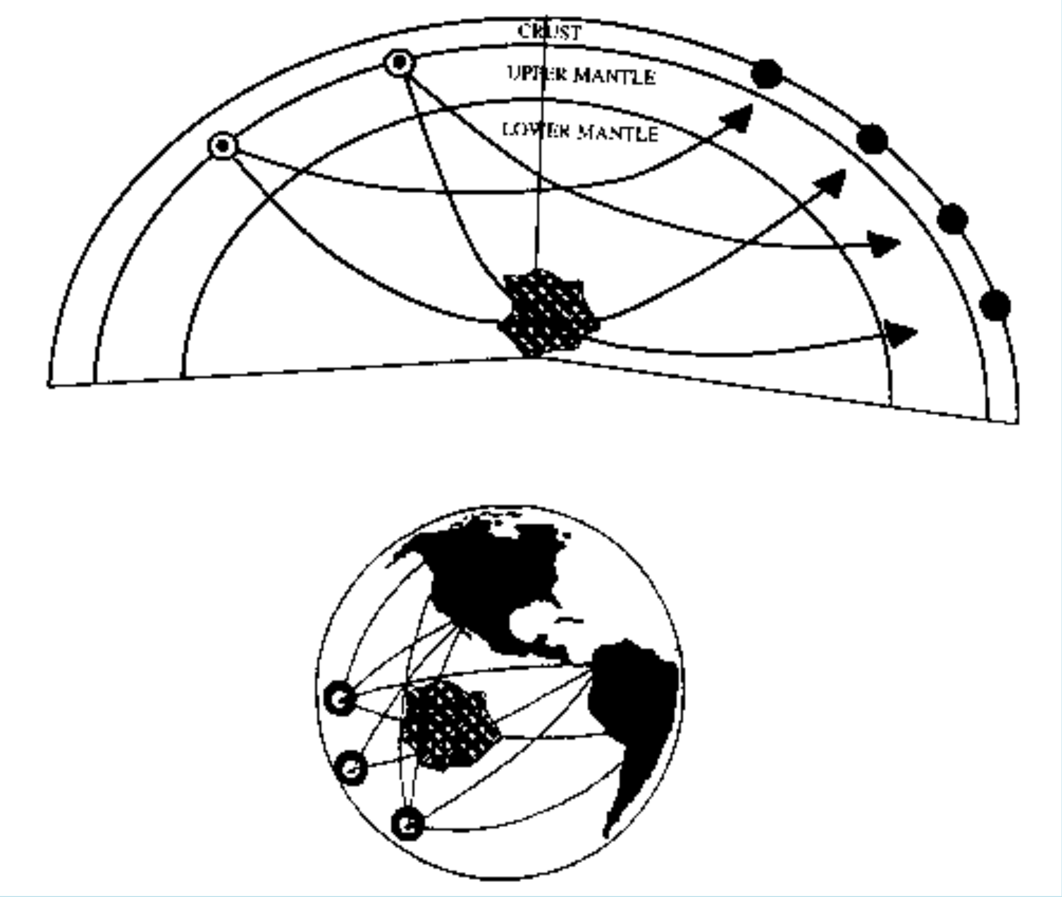


(b) THE LITHOSPHERE AND THE ATHENOSPHERE



Compositional versus rheological layering of the earth

Lateral (3-D) variations in deep earth structure



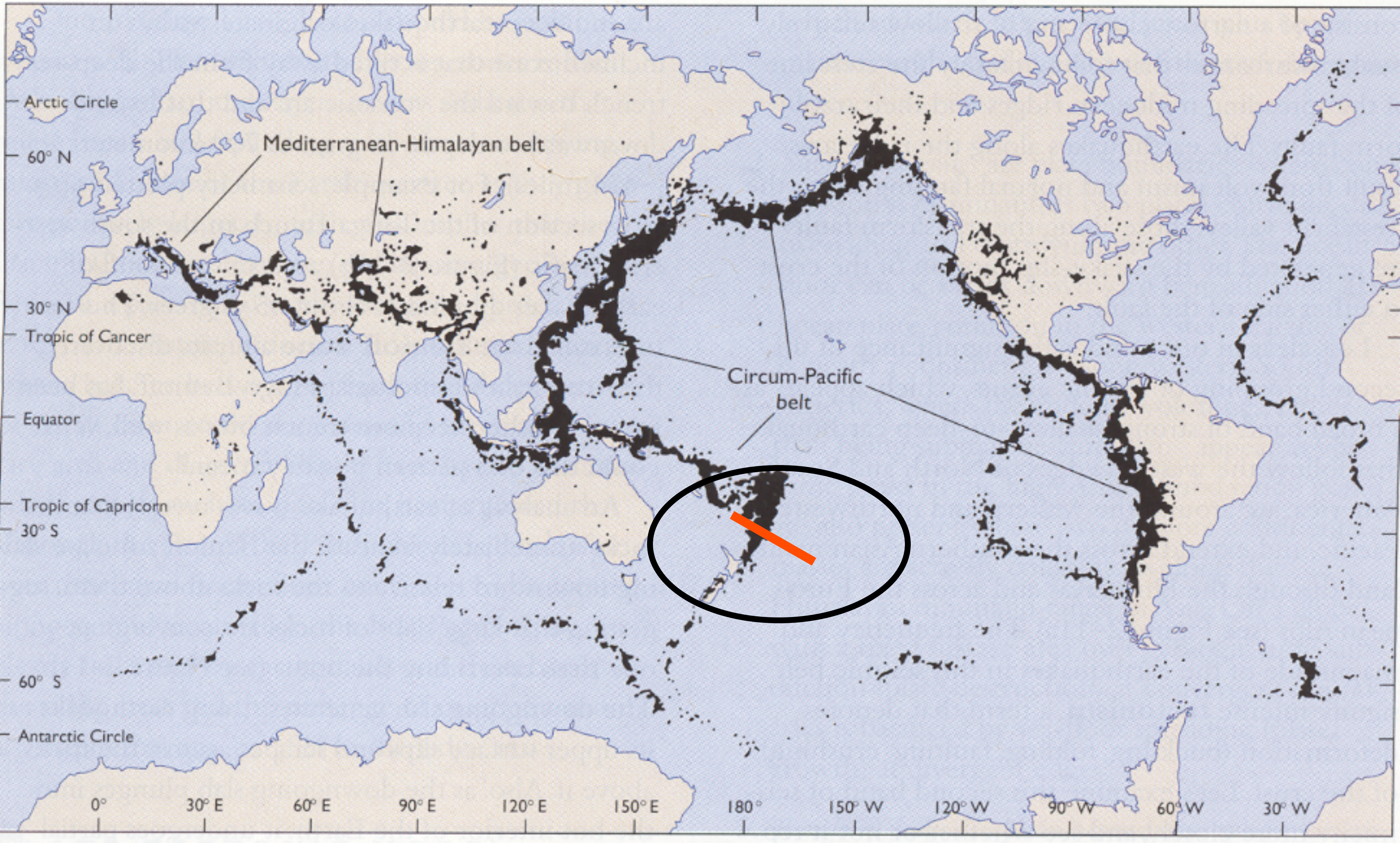
Seismic tomography:

many ray paths through the earth measure small anomalies in travel time

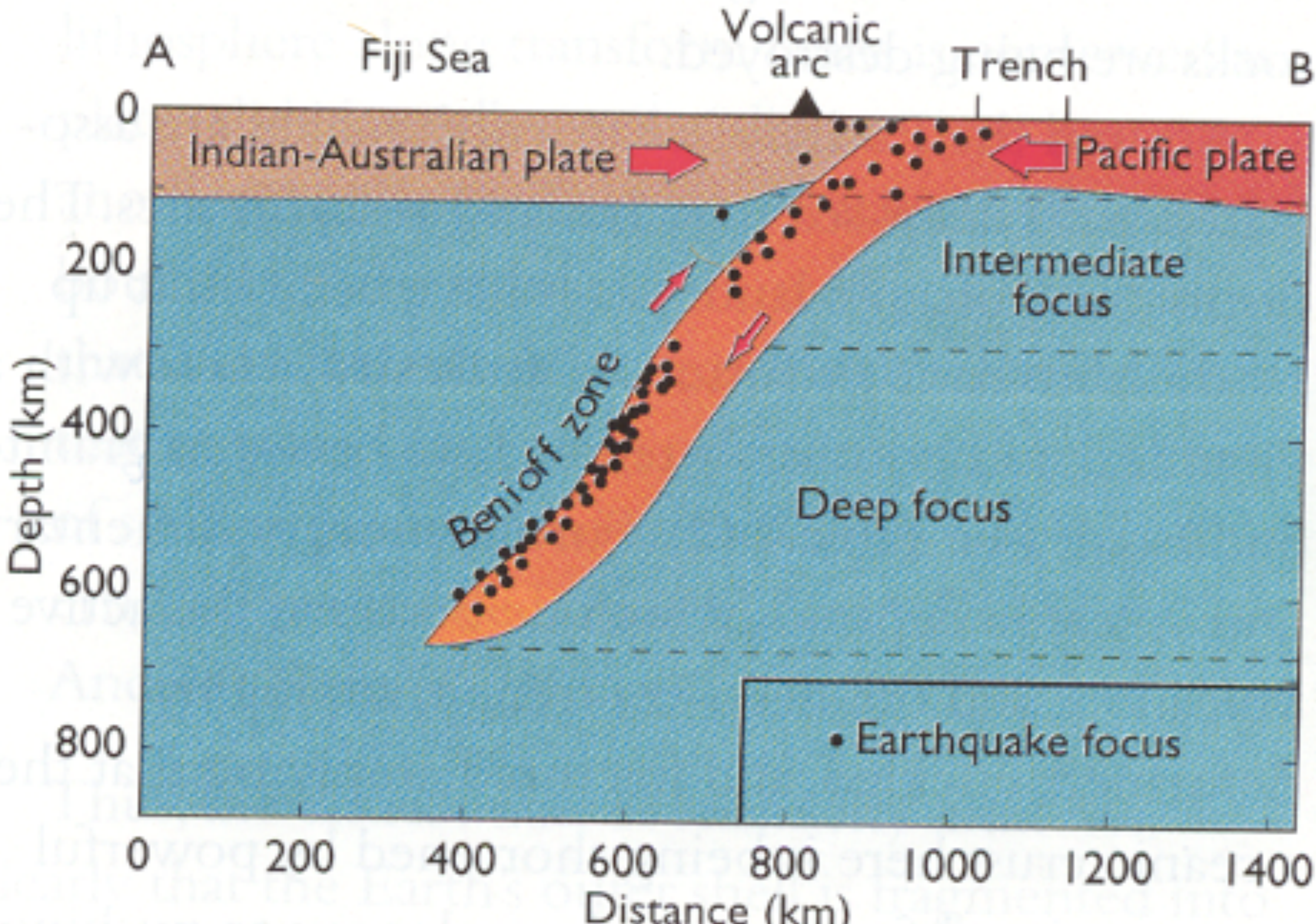
locate region of anomalous material
early arrivals = fast = blue (cold)
late arrivals = slow = red (hot)

See KK&V section 2.1.8

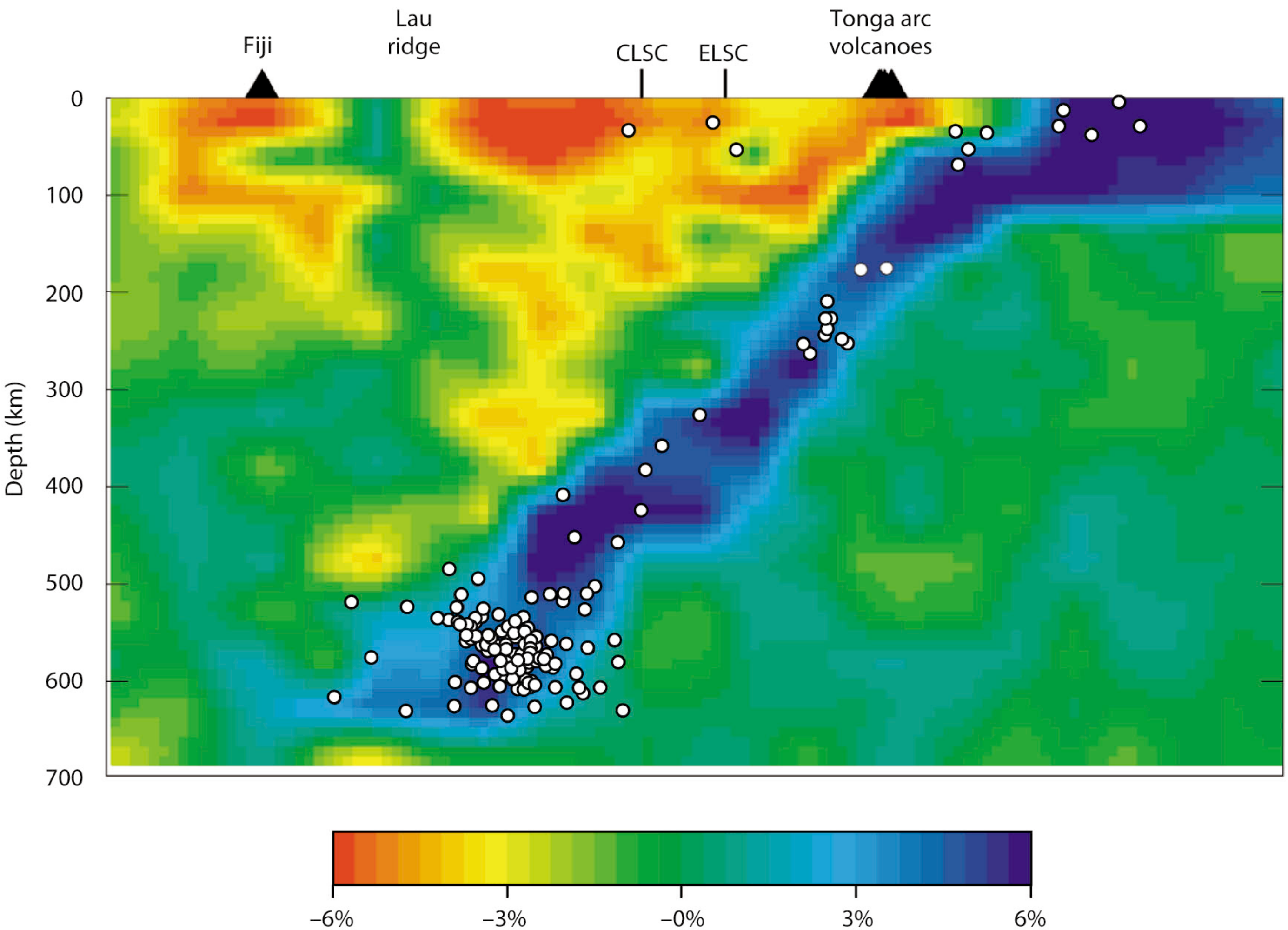
Example: Tonga Trench



by the geometry of earthquake focal enters



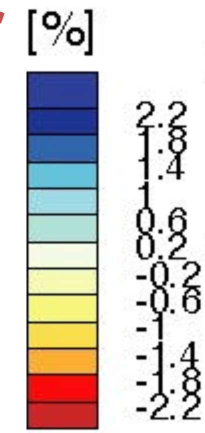
Cross section of Lau Basin-Tonga Trench from seismic tomography

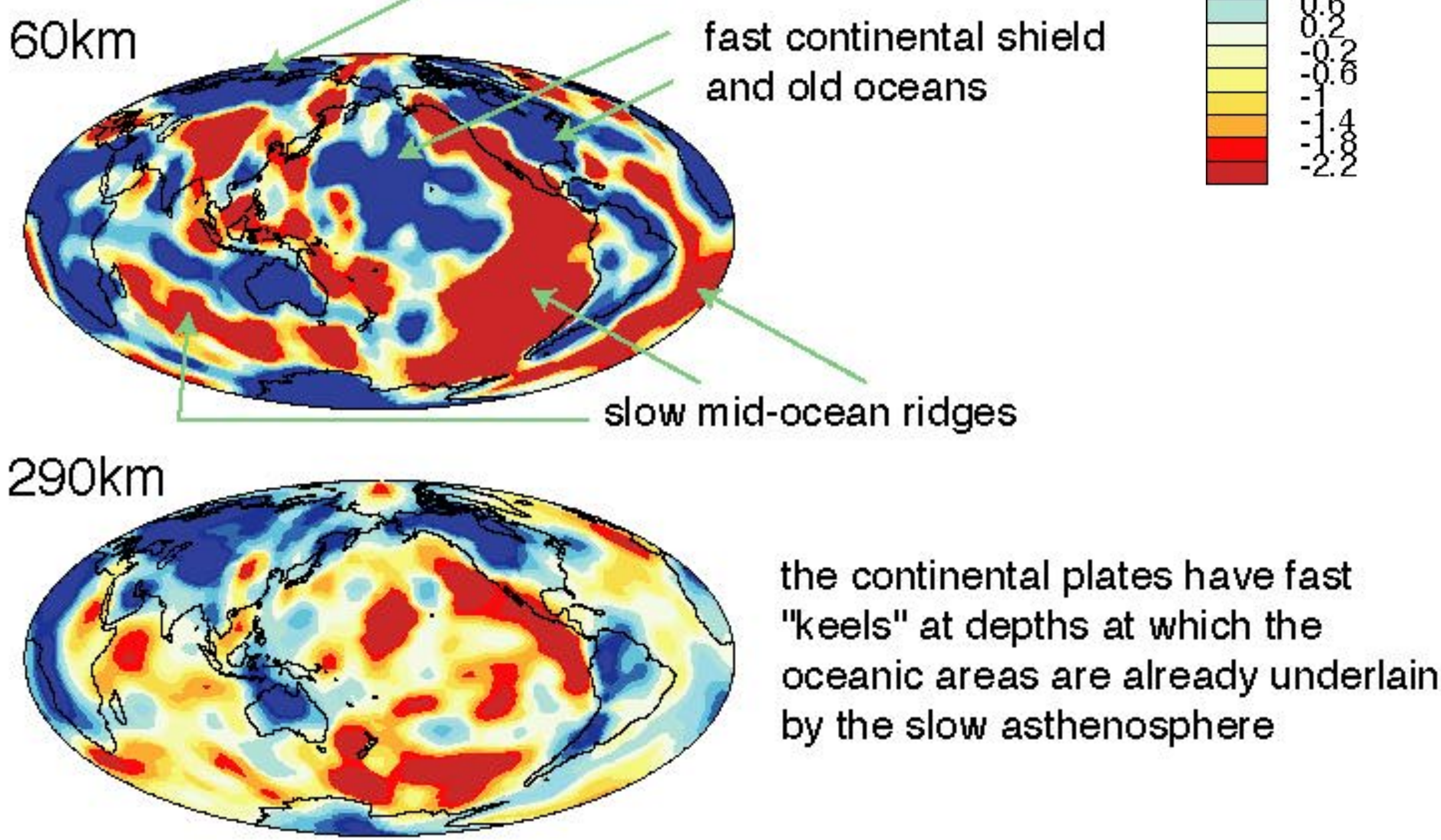


seismic tomography permits a view of the earth's deep interior

blue = faster = colder = stronger

red = slower = hotter = weaker



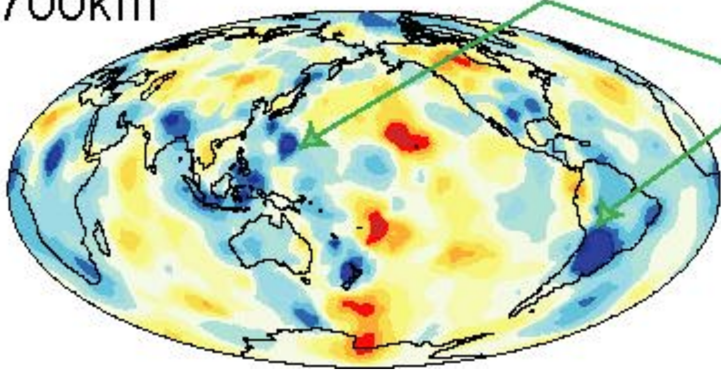


blue = faster = colder = stronger

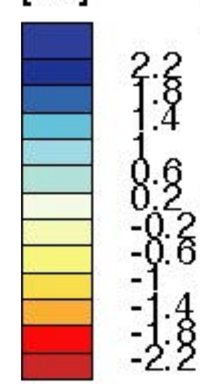
red = slower = hotter = weaker

to the base of upper mantle

700km

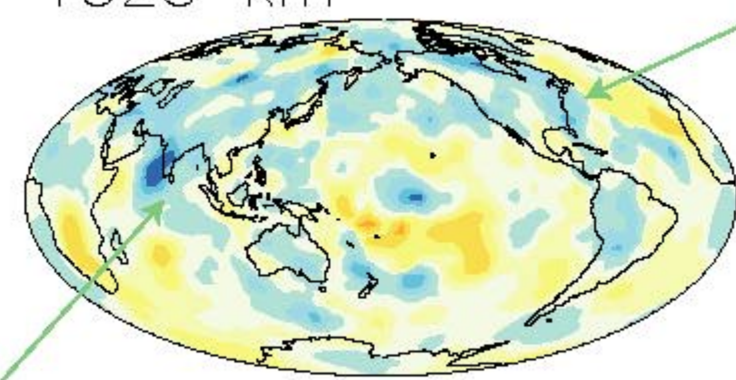


The "cold" subducting slabs show up as seismically fast areas. They pass the 670km discontinuity between upper and lower mantle and penetrate well into the lower mantle.

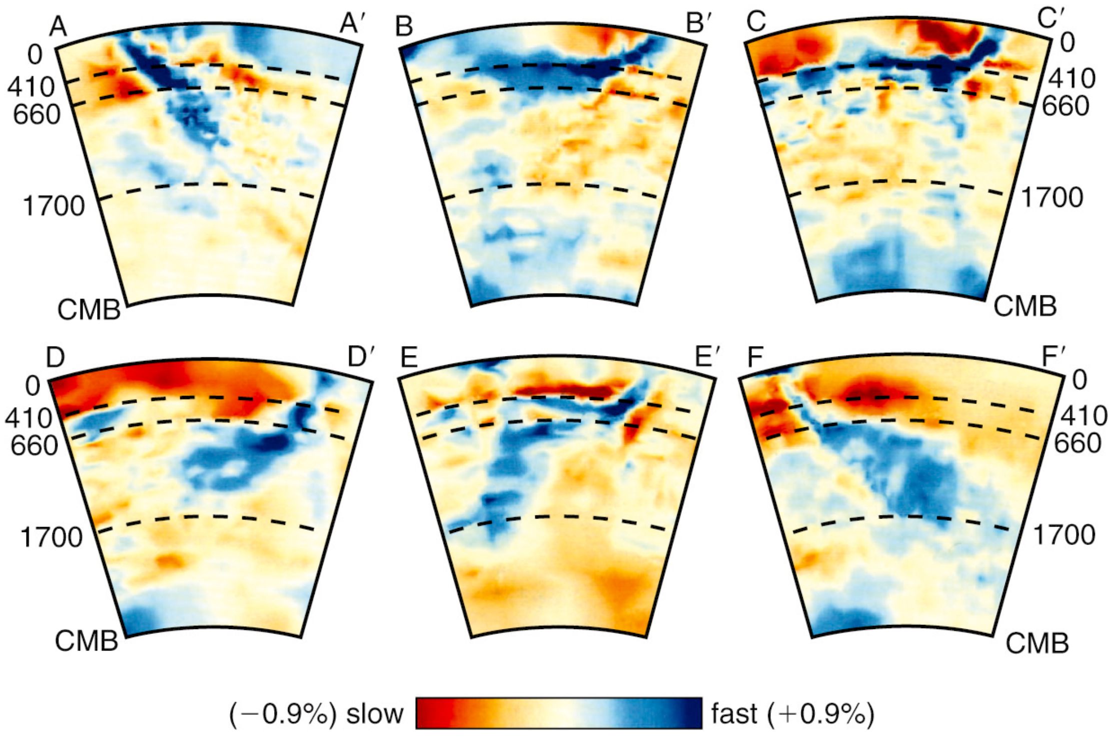


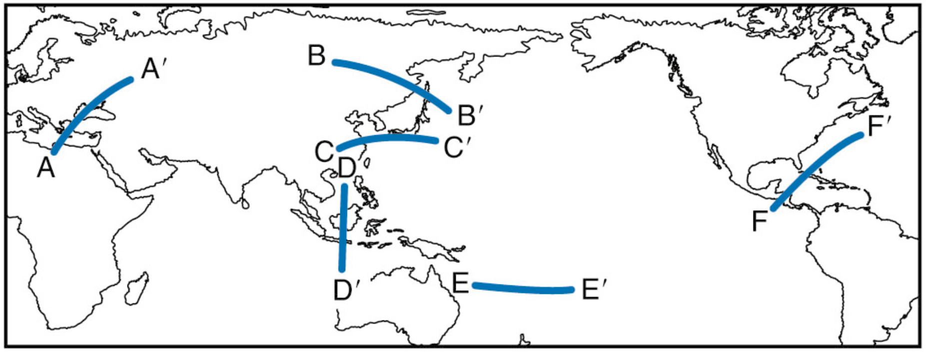
and even way down in the lower mantle

1525 km



Some of the "cold" subducting slabs can be traced well into the lower mantle. E.g. old Farallon and Tethian subducting slabs.





KK&V plate 9.2

Cross sections through the mantle from seismic tomography reveal subducted slabs penetrating into lower mantle

Addresses (resolves) issue of single or two layers of convection within mantle

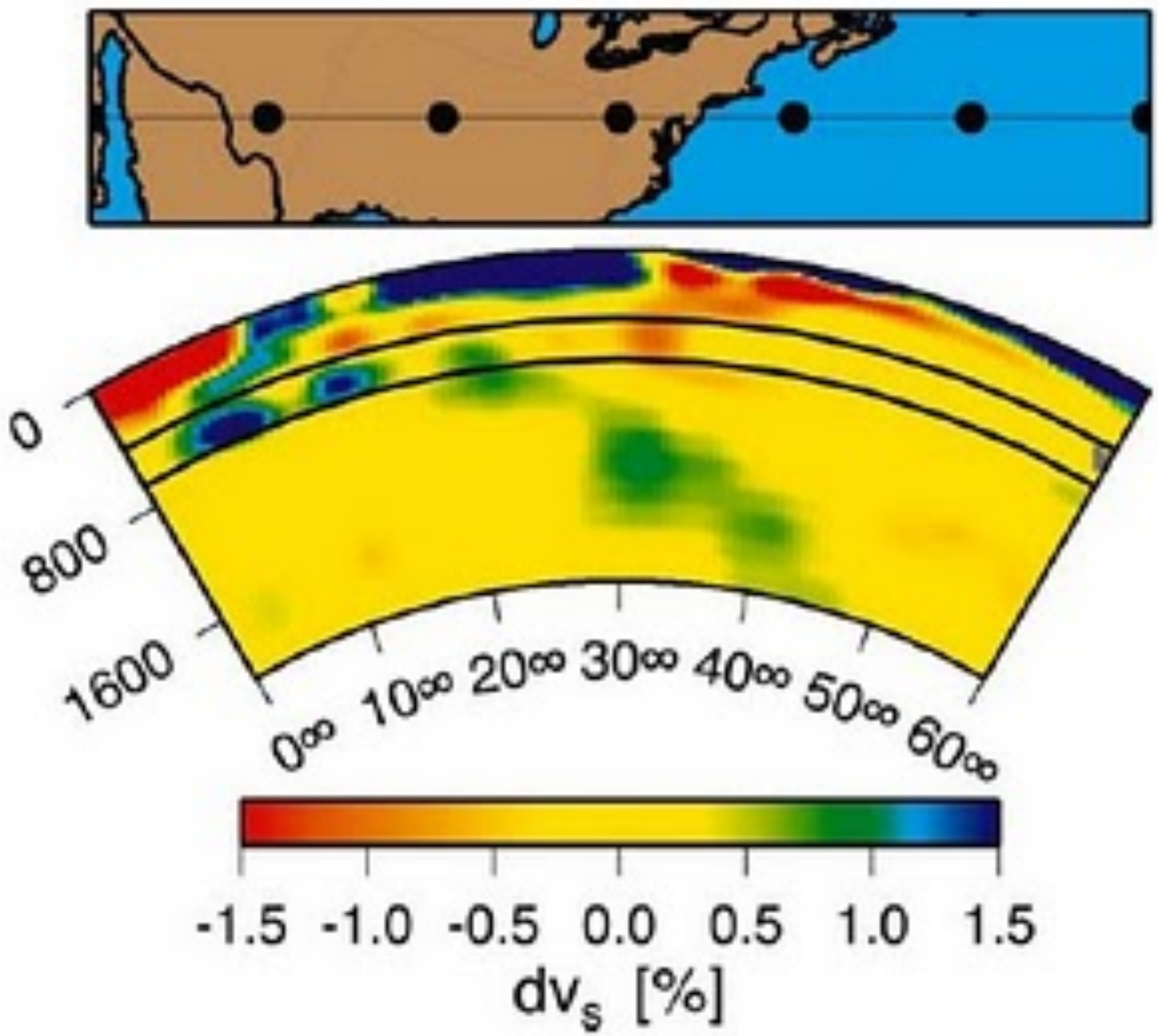


Fig. 3. Combined cross section through the tomographic

Can image Farallon slab (cold) beneath North America

Now are trying to reconstruct slab back to 50 Ma using superimposed plate motion models

Can model effect of deep slab on vertical motion in overlying continents

“dynamic” topography

May explain ups and downs along eastern seaboard formerly attributed to sea level change

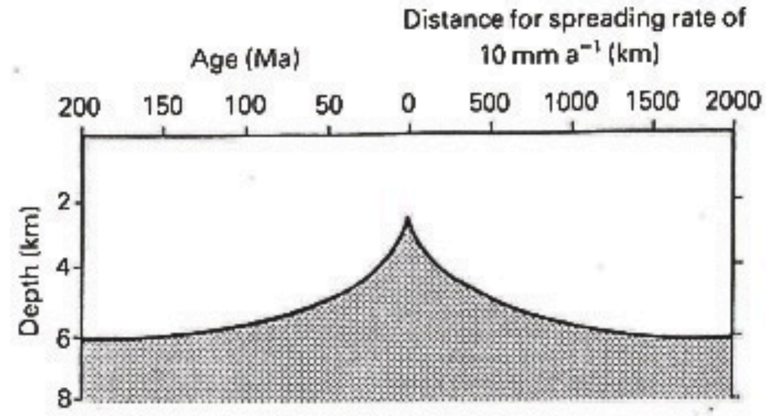
Wannamaker et al. (2003)

Age – depth relationship

“Square root of age” cooling

Read chapter 6.4 in KK&V

Empirical depth/age relationship

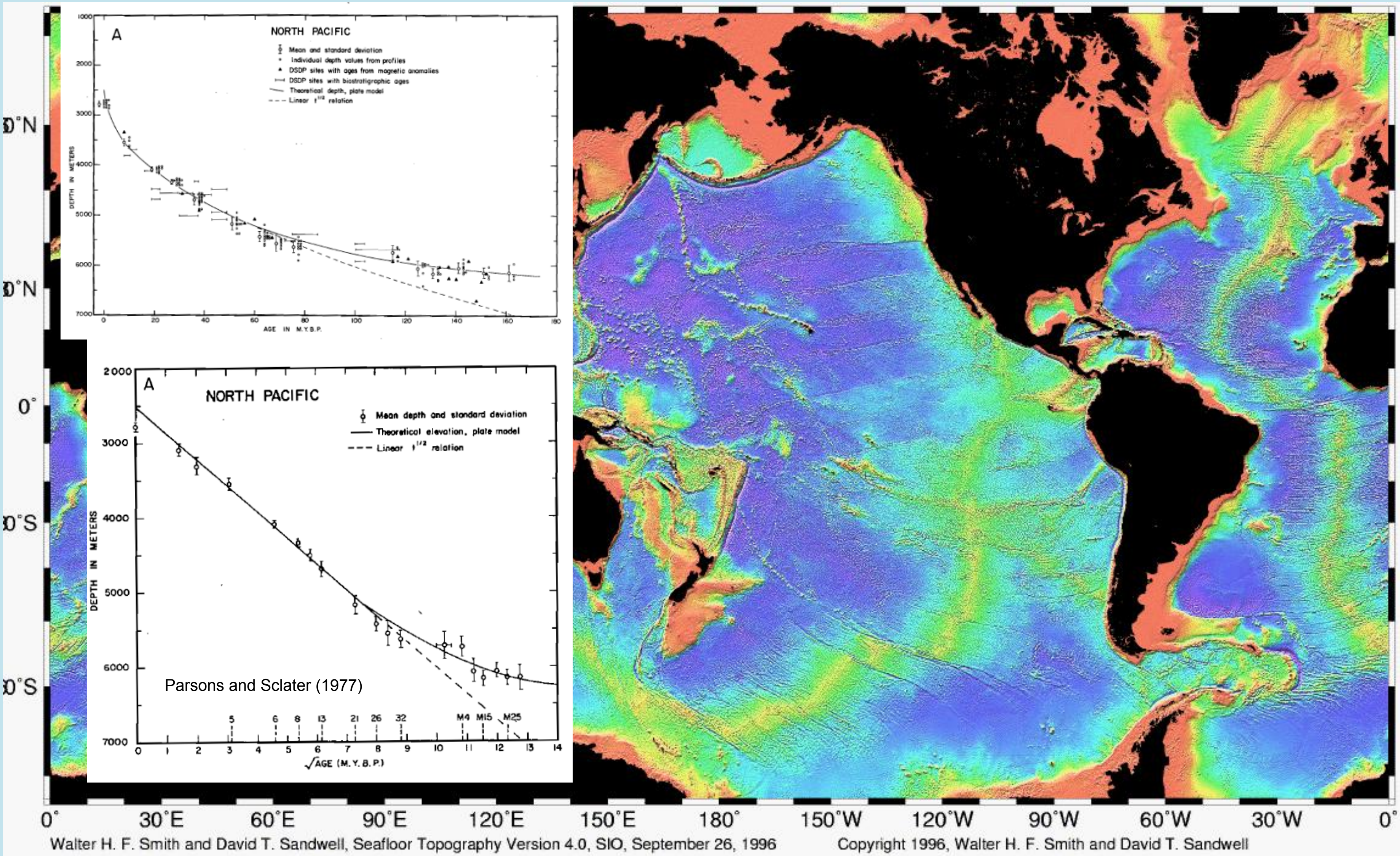


$D = 2500 + 350 \sqrt{T}$ where T is age in Ma

Valid out to about 80 Ma,
then depth subsides more slowly

Controls long term sea level changes
Most notably – high sea levels in the late Cretaceous

Depths in North Pacific follow square root of age out to roughly 80 Ma



Conductive cooling and the thermal lithosphere

$q = -K\, dT/dx$ Fourier's Equation

where q = heat flux (W/m^2)
dT/dx = temperature gradient (K/m)
K = thermal conductivity (W/mK)

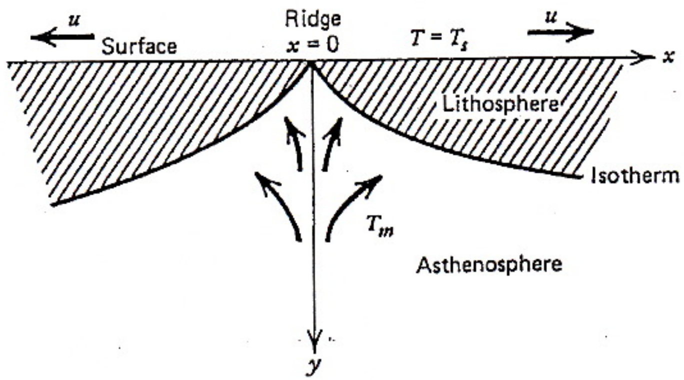


Figure 4-22 Schematic of the cooling oceanic lithosphere.

Turcotte and Schubert (2002)

Unlike the mantle, which cools by convection, the lithosphere cools by conduction

Why does the lithosphere thicken as the sq root of time?

Start with Fourier's equation

Heat flux is proportional to temperature gradient

Consider the relationship between time and depth of cooling:

Take two blocks, one twice as tall as the other (h and 2h);
The blocks have the same initial T (T_o); so the “2h” block has twice as much heat as the “h” block;
Cool through the top by ΔT;

The temp gradients in the blocks are $\Delta T/h$ and $\Delta T/2h$;
So “h” block loses heat twice as fast as “2h” block;
But “h” block has only half the heat of the “2h” block;
so the “h” block cools in ¼ the time

So, (using Z instead of h)

t is proportional to Z^2

By adding a constant of proportionality we write:

$$t \approx Z^2/k$$

where k is the thermal diffusivity, given by

$$k = K/\rho C_p$$

Where K is the thermal conductivity in units of W/mK
ρ is the density in kg/m^3
C_p is the specific heat capacity in J/kgK

For the lithosphere, typical values of K = 3 W/mK, ρ = 3300 kg/m^3 and C_p = 1000 J/kgK

give $k = 10^{-6}\ m^2/s$

(Thermal diffusivity is the ease with which a material undergoes a change in temperature)

The relationship between thermal diffusivity, layer thickness and time is a remarkably powerful one.

For example:

How long will it take a 1 m thick dike to cool?

$$t \approx Z^2/k = 1^2/10^{-6} = 10^6 \text{ sec} = 10 \text{ days}$$

The relationship can also be used to determine how deep the lithosphere will conductively cool. In time, heat flow causes temperature changes to occur over a length scale Z

$$Z \approx \sqrt{kt}$$

So, for 100 m.y. old oceanic lithosphere cooled by the overlying ocean we would expect a lithospheric thickness of

$$Z = \sqrt{10^{-6} \times 10^{8} \times (\pi \times 10^{7})} = \sqrt{\pi \times 10^{9}} = 56 km$$

A more accurate thickness of the lithosphere

Since lithosphere cools asymptotically,
the "thickness" of the "thermal boundary layer" is arbitrary.

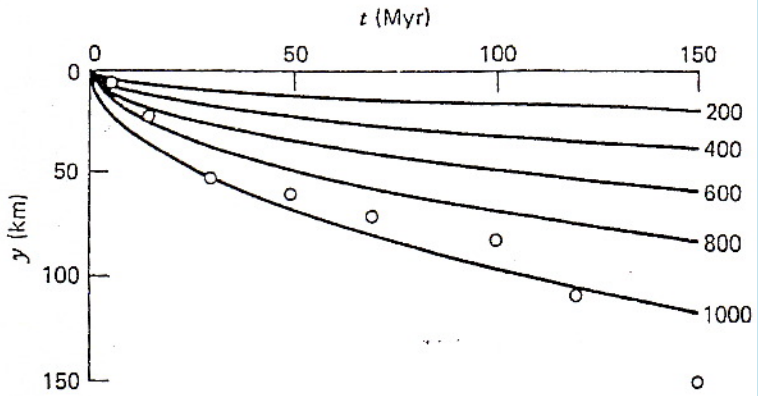
We take the thickness to be the depth at which layer has cooled by 90%

$$Z_{90\%} = 2.32 \times \sqrt{kt} = 2.32 \times \sqrt{k\frac{x}{u}}$$

Where u is spreading rate

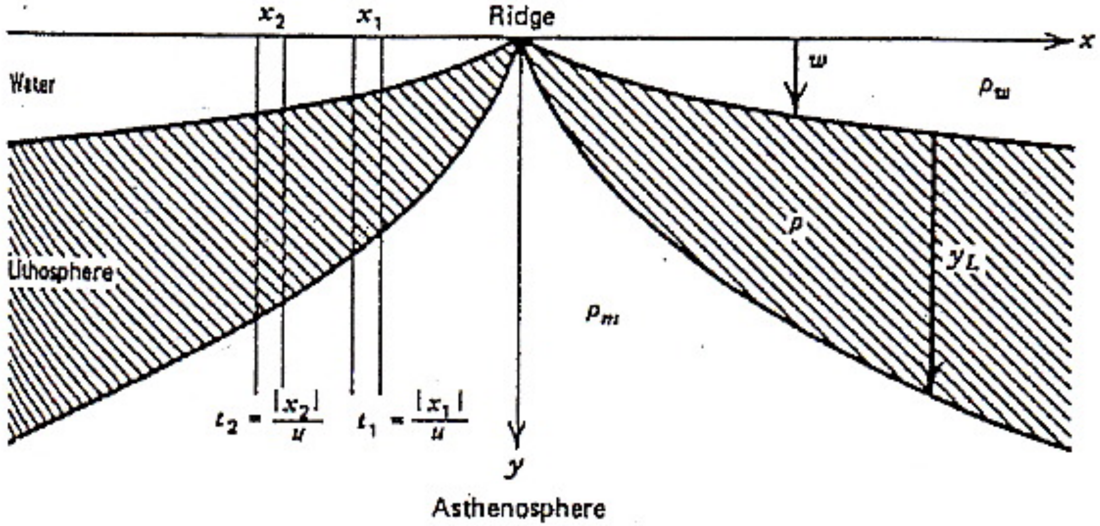
Factor of 2.32 comes from the full solution of the problem

Since typical mantle is 1300°C,
we define the thickness to be the depth to the 1170° isotherm



Above: Theoretical isotherms

Circles are data points from Rayleigh wave dispersion studies



*from Turcotte and Schubert, 2002, Geodynamics

So, why does the seafloor also follow a square root of age curve?

We can use the principle of isostasy ("beneath the depth of compensation the pressures generated by all overlying materials are everywhere equal") to determine how the depth of the seafloor varies as a function of time.

For the case of the oceanic lithosphere, the mass, M, in any column of unit cross sectional area is:

$$M = w\rho_w + Z\rho_L + (H_c - w - Z)\rho_m$$

where w is the increase in the depth of the ocean (relative to the mid-ocean ridge), Z is the thickness of the lithosphere, H_c is the depth of compensation, ρ_w is the density of the oceans (1030 kg/m^3), ρ_L is the density of the lithosphere (~3350 kg/m^3), and ρ_m is the density of the asthenosphere (3300 kg/m^3). The density of the lithosphere is higher because it is colder (thermal contraction causes density to increase).

Using our relationship for the thermal boundary thickness of the lithosphere, we can write this as

$$M = w\rho_w + 2.32\sqrt{kt}\rho_L + (H_c - w - Z)\rho_m$$

For a column of material at the ridge crest, $w = 0$ and Z is near zero.

$$M = H_c\rho_m$$

Equating these two equations gives

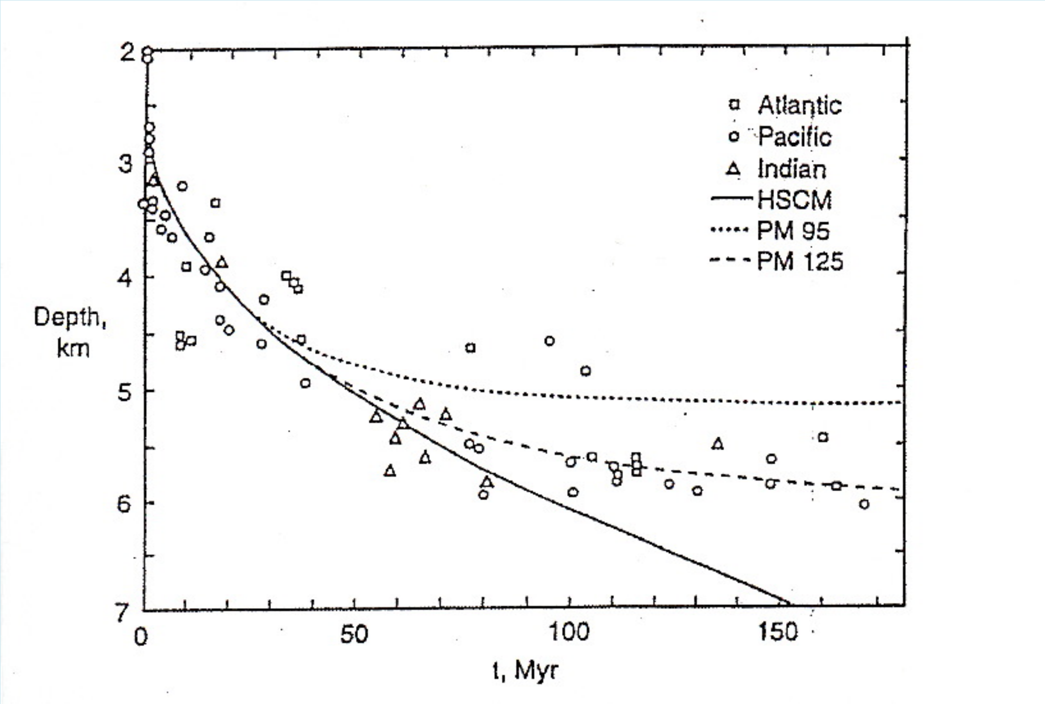
$$H_c\rho_m = w\rho_w + 2.32\sqrt{kt}\rho_L + (H_c - w - Z)\rho_m$$

Solving for w gives

$$w = 2.32\sqrt{kt}\frac{(\rho_L - \rho_m)}{(\rho_L - \rho_w)}$$

Thus, the depth of the seafloor increases as the square root of age.

Depth-age relationship



HSCM = half space cooling model
= simple model

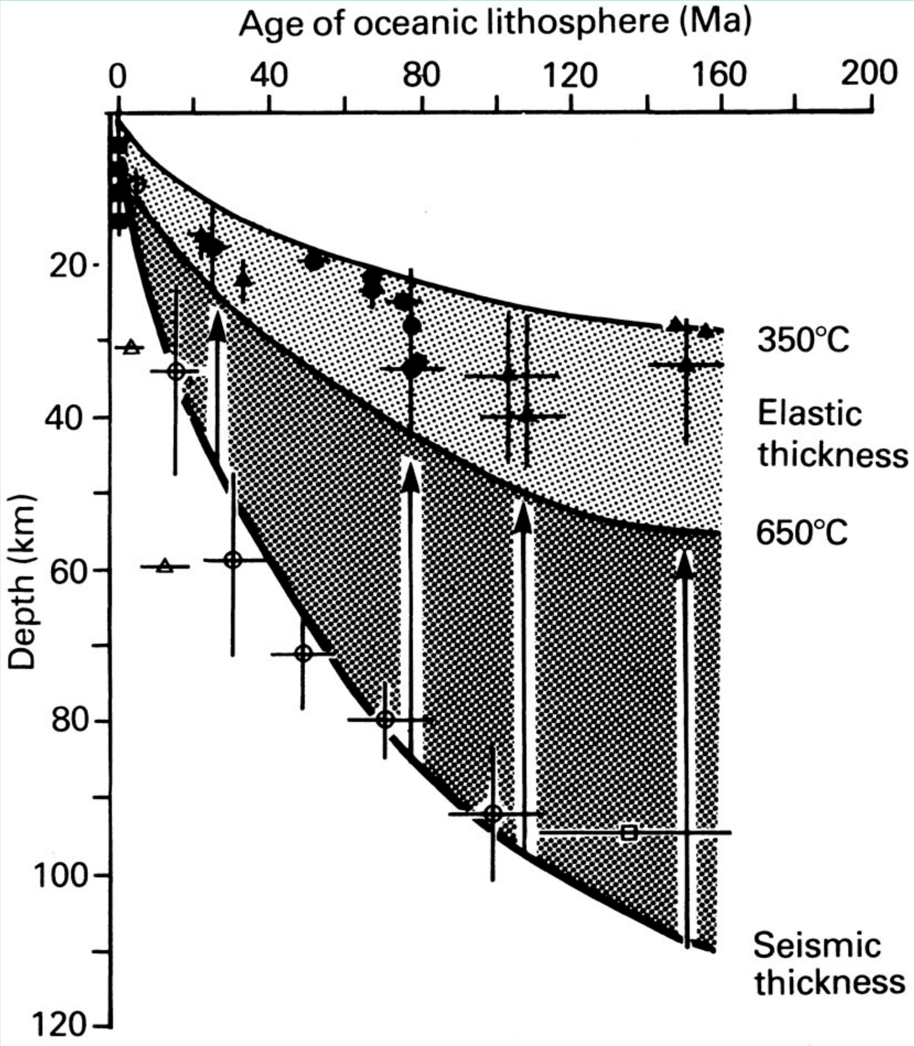
Observations of seafloor depth generally follow a √t curve for the first 80 m.y. or so.

After that they tend to deepen much more slowly than predicted from a simple “half-space” model.

This difference can be attributed to additional heat being added to the base of the lithosphere

E.g., may reflect secondary mantle convection or shear heating.

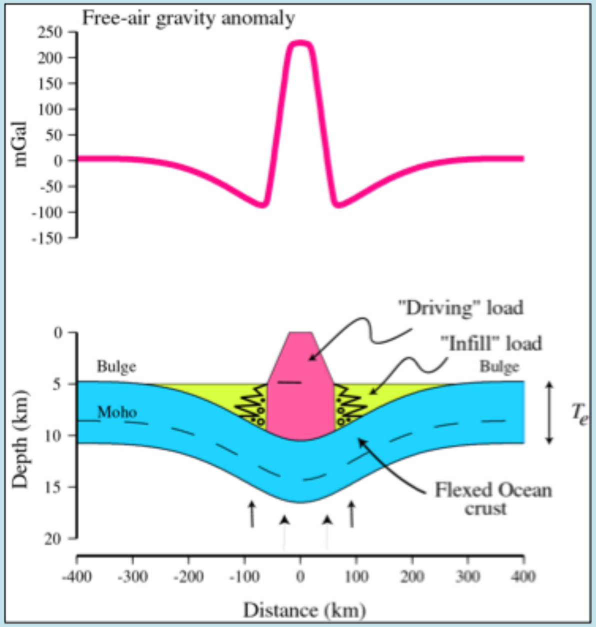
Controversial …



Short term versus long term thickness of lithosphere

Short term (seismic) = .85 Tm

Long term (elastic/flexure) = .55 Tm

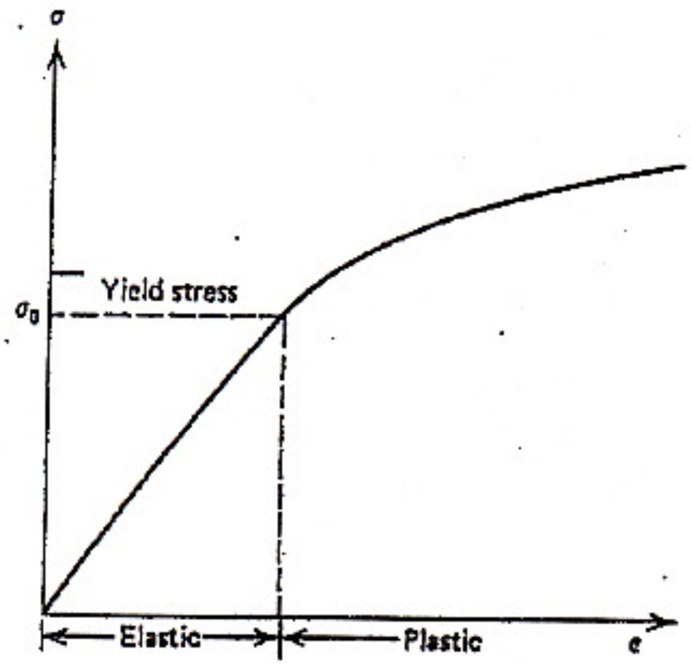


Rheology

From Greek rheo = flow

Elastic-perfectly plastic rheology

Deformation of a solid exhibiting an elastic–plastic transformation.



*from Turcotte and Schubert, 2002, Geodynamics

viscosity proportional to $\exp\left(\frac{E_a + pV_a}{RT}\right)$

Ea = activation energy/mole

Va = activation volume/mole

Geologic materials exhibit brittle (elastic) behavior at surface conditions but deeper these materials deform in a ductile (viscous) fashion

P,T and strain rates are important in determining brittle-ductile transition

Modes of ductile deformation:

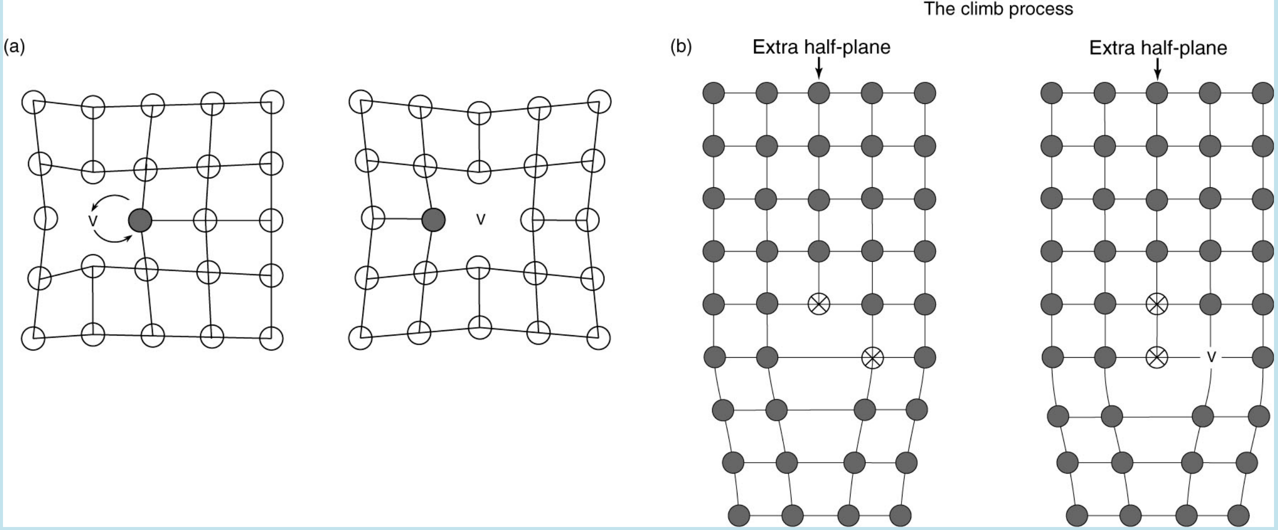
1) Diffusion creep – diffusion of atoms/vacancies (low stress)

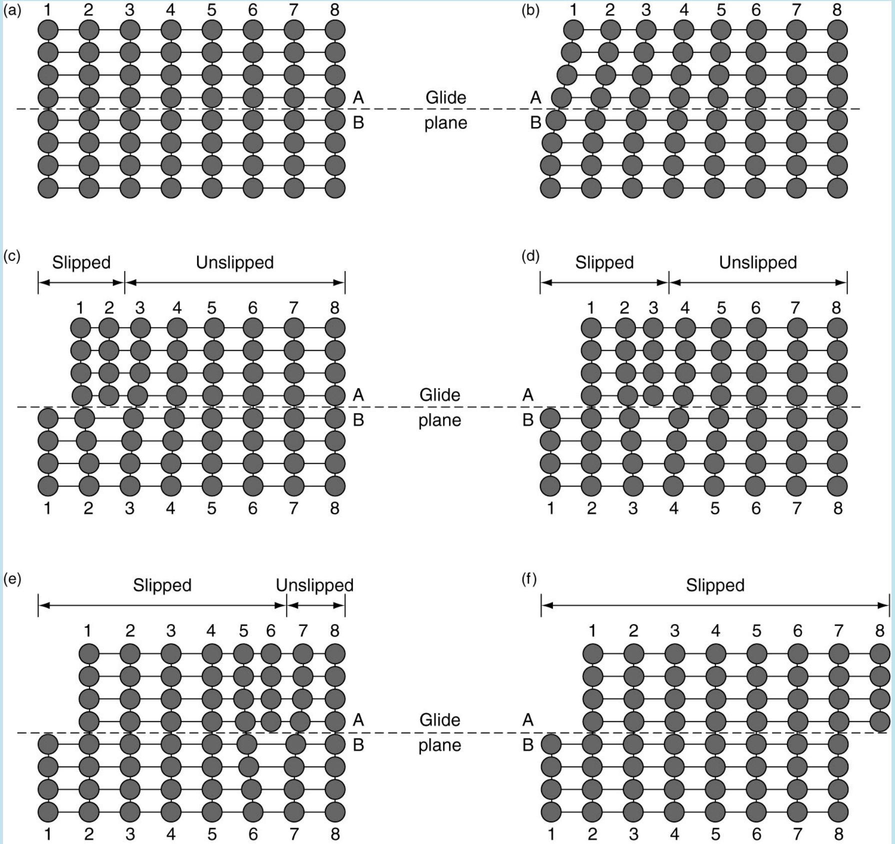
2) Dislocation creep - migrations of imperfections in crystal lattice (high stress)

Both modes have viscosity that depend exponentially on pressure and inverse of temperature

Two methods of ductile flow in the mantle

Can distinguish based on presence or absence of seismic anisotropy





(Above) Diffusion creep

KK&V

When T > .85 Tm

Migration of individual atoms and vacancies in a stress gradient

Occurs in asthenosphere and lower mantle

Leads to isotropic seismic velocities

(left) Dislocation creep (or power law creep)

When T > .55 Tm

Deformation takes place by dislocation glide

Controls convective flow in upper mantle

Leads to anisotropic seismic velocities

Have a gliding motion of large numbers of defects in the crystal lattices

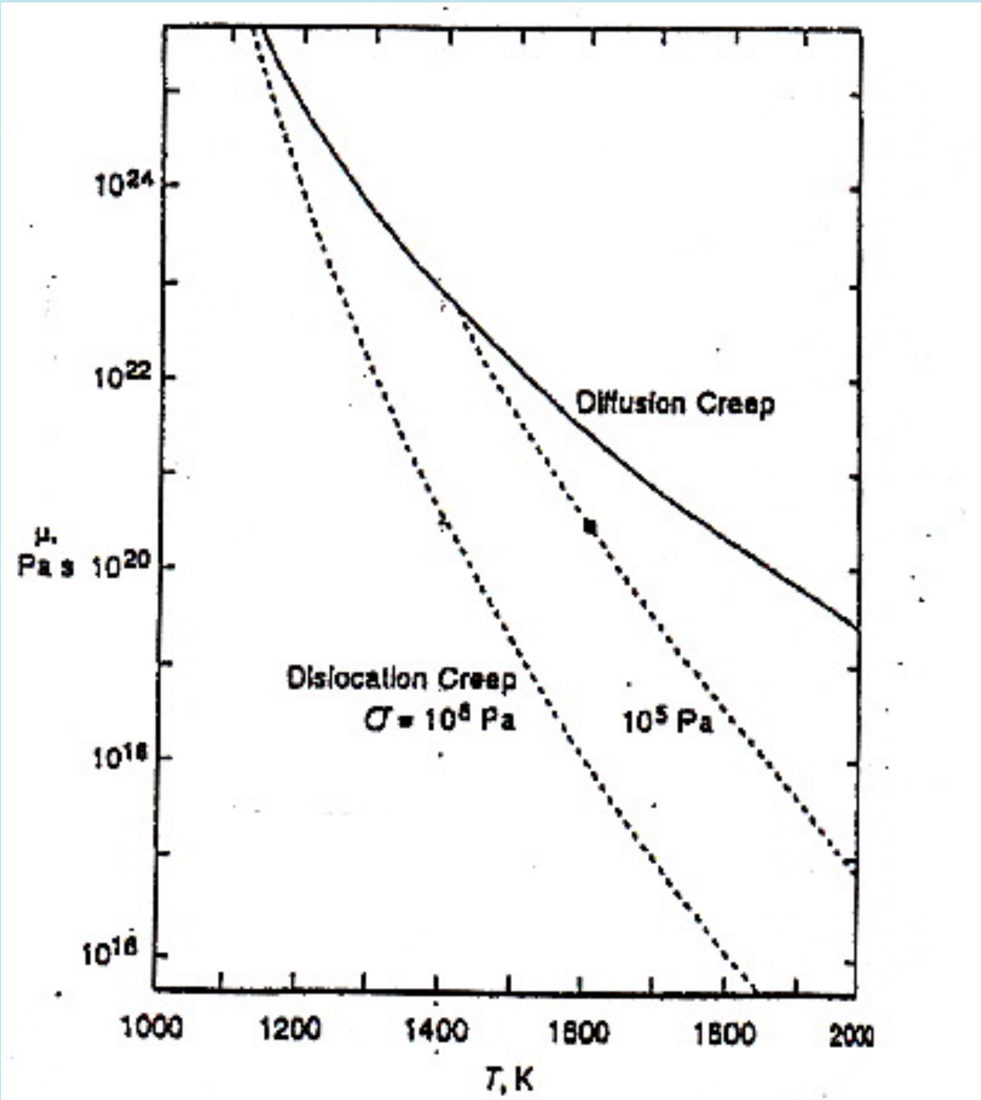


Illustration of the strong dependence of the viscosity (μ) of a dry upper mantle on temperature for several stress levels.

For diffusion creep (low differential stress) - μ is not dependent on stress

For dislocation creep (dashed lines) μ is dependent on stress

Turcotte and Schubert (2002)