COMPARATIVE PLANETOLOGY

Interiors & Surfaces

Note: Calendar modifications (on Website)

This Week: Planetary Interiors (heat flow & Basics, EOS, etc. & "Observations" - ~~other probes~~ seismic, - other probes) — Read Chapter 8 to p. 211

SILLY PUTTY

Our S.S. insides — 211-223, 230-231

Next Week Planetary Surfaces — Read Chap. 9

- Basics: Rock types, densities, & geochem — 237-246, briefly
- Impact Cratering — 246-264
- Volcanism & Tectonics — Ch. 10, 271-284
- Weathering/Erosion — 287-293

Our SS outsides — 294-313

So...—What do we know about planetary interiors?

- what do we want to know?

$T(r)$, $\rho(r)$, comp(r), state(r)

↳ similarities & diff's between planets

↳ formation, history, and future

22-141 50 SHEETS
22-142 100 SHEETS
22-144 200 SHEETS
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· Where to start?
- Study the ⊕
- " " Moon
- Have theory, apply to planets

⊕ → surface composition, morphology
↳ peeks inside via volcanos
↳ seismology
} also, may the Moon

Even more basic; general
- MASS - via moons
- RADIUS & SHAPE
→ mean density
- surface composition
- samples
- spectroscopy / spectrophotometry

Earth: 5.52 g cm^{-3}
Moon: 3.34 g cm^{-3}
Venus: 5.25 g cm^{-3}
Mercury: 5.42 g cm^{-3}
Mars: 3.94 g cm^{-3}
Vesta: ~2.9 g cm^{-3}
Ganymede: 1.93 g cm^{-3}
Titan: 1.88 g cm^{-3}

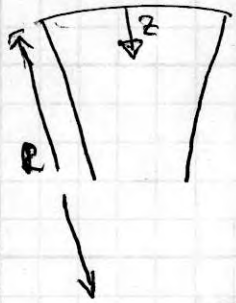
Mass Rank
① ⑤ ② ④ ③ ⑧ ⑥ ⑦

* Typical surface rock has density of ~ 2.8 g/cm^3

- What does bulk ranking say?
Generally, big M → big ρ
- * says that Earth is highly differentiated (much denser inside); Moon, Mars less so

Planetary Interior Physics / Chemistry

Remember Hydrostatic Equil. for ⊙? Still relevant for massive-enough planets!



$$\frac{dP}{dz} = \rho \frac{GM}{(R-z)^2} = \rho g$$

So as you go deeper ($z\uparrow$) $P\uparrow$

But what about ρ?

Equation of State: Gas: $P = \frac{\rho}{\mu} N_A k T$ (mmw)

① Planets (Rock, ice, high P liquid) → more complex

② need to know $T(z)$ to find $\rho(z)$ given $P(z)$

First, like in case of ⊙: • source(s) of internal heat
- resid. of accretion
- radioactive decay

balanced by • Heat radiated away
- surface radiation
- reduced by abs. of ⊙ light

* expect $T(z)\uparrow$ w/ (z)

③ i.e. State changes of rock:

1- melting

2- high P → solidification @ high T?

3- Solid state phase changes

↳ different phases, different compressibility

i.e. diamond ↔ graphite

* DATA FROM THEORY + EXPERIMENT
BUT TOUGH TO DO AT BOTH HIGH T & P, SUSTAINED

October 6
HW reminder

That said, what you talkin' 'bout, Willis?

Use HSE to estimate P_c

$$\int_0^{P_c} dP = -\frac{4}{3}\pi G \bar{\rho}^2 \int_0^R r\,dr$$

$$-P_c = -\frac{4}{3}\pi G \bar{\rho}^2 R^2$$

$\oplus$ $\bar{\rho} = 5.500\ g\,cm^{-3}$

$R = 6.33 \times 10^8\ cm$

$P_c \approx \frac{3}{2} \times 10^{12}\ dyne/cm^2$

$1\ bar = 10^6\ dyne/cm^2$

$\rightarrow$ $P_c = 1.5$ Mbar

Moon: $\rightarrow$ 47 k bar

Via Geophysics, Geochem. can make a phase diagram for planetary components (ie. rocks of various types, ices, iron, etc) at given pressure (general P(ρ) depends only slightly on T); get

- FIGURE 8.2 - from text

FIND

- Terrestrial Planets
  - Silicates + Iron (Merc., Venus, $\oplus$)
  - Silicates, less Fe (Moon, Mars)
  - Ices (Ur, Nep)
    - $\rightarrow$ liq (Sa, Jup)
  - Gas (Sun)

Another Key Issue: Differentiation

Cooling planet: Xtalizing minerals sink or float in liquid remaining.

lithophiles: elements happy w/ loop silicates; Na, Ca $\rightarrow$ Granite

siderophiles Fe, Ni

Bulk Properties (cont'd)

Plasticity: still 'solid' but long-term flow to relieve stress

↳ heavy minerals in bulk can settle down through mantle

SUMMING UP

HSE + DIFFERENTIATION

↳ Jeuser core made of "naturally" Jeuser constituents

OTHER PROBES OF INTERIOR STRUCTURE

- **ROTATION**

Total angular momentum $J = I\Omega$

$$I = \int_0^{\pi}\int_0^{R} r^2 \, dm \sin^2\phi = \frac{\pi}{2}\int_0^{R} 4\pi r^2 \rho(r)\, dr$$

ρ = constant $I = \frac{4}{5}\pi R^5 \bar{\rho} = \frac{3}{5} R^2 \cdot \frac{4}{3}\pi R^3 \bar{\rho}$

$= \frac{3}{5} MR^2$

Uniform shell $I = 2/3\ MR^2$

Uniform sphere $I = \frac{2}{5} MR^2$

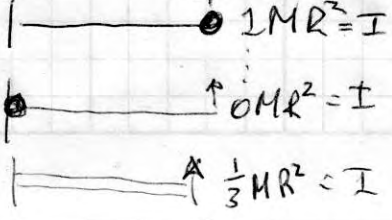
Core $\rho = \bar{\rho} \times 2$, $R = R/2$ $I = 0.347\ MR^2$

Moon $I = 0.40 MR^2$

⊕ $\sim 0.33 MR^2$

- Measure via precessional "wobble"
- also by $F_g(r)$ by passing spacecraft

Limiting Cases:



$I = \int_0^M r^2 \, dm$

$\int r^2 dr = \frac{r^3}{3}$

$= \frac{1}{3} MR$

- HW Due Fri
- Rever spotted Friday at 5

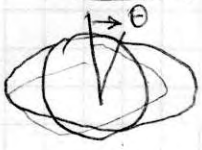
- **Oblateness** $\epsilon = \frac{a-b}{a}$ (b: polar diameter, a: equatorial diameter)

$\epsilon(M, R, \Omega, \text{comp})$ ← rock, mantle/core fraction, etc.

Earth: $\epsilon = 0.0034$ } same Ω
Mars: $\epsilon = 0.0006$ }
Jupiter: $\epsilon = 0.0650$ fast rotator
Venus $\epsilon = 0$ slow rotator

Py 198

"J2"



$$\Phi_g = -\frac{GM}{r}\left(1 - J_2\left(\frac{R_e}{r}\right)^2 P_2(\cos\theta)\right)$$

$P_2 \propto \frac{3}{2}\cos^2\theta$

J_2 calculated, given

so a rotating planet will have an excess gravitational effect towards equator, ~~can measure by~~

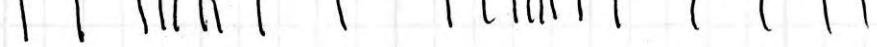
- Rotation rate
- Plasticity of interior

- Measure by anomalous orbital motion
  - ↳ orbits assume pointlike center of mass i.e. central potential
  - But J_2 → non-central potential
- further departures → non axisymmetric
  - ↳ "Mass cons"

• SEISMOLOGY

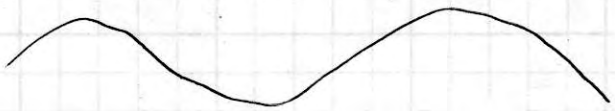
use ~~sound~~ seismic waves to probe density + state vs. depth & position

P (pressure) waves:



- compression waves; i.e. sound waves
- longitudinal (motion in direction of propagation)

S (shear) waves



- transverse (⊥ to propagation)

- Speed at given point depends on:
  - density
  - 2 elastic "constants"
  - wave type
- Generally — P faster than S
  - no S waves in liquid

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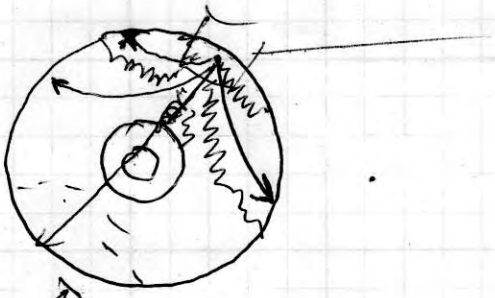
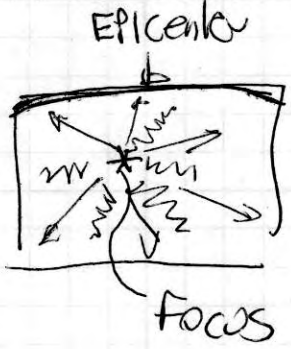
Phenomenology



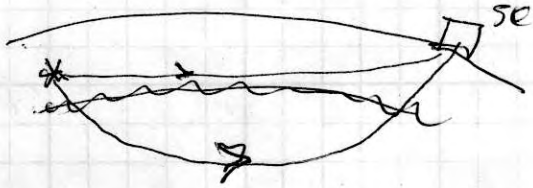


- ↑ swave "shadow zone"
- increased P-S delay further away
- refraction of P, S by changing density

$$t_{s,p} = \int \frac{dr}{V_{s,p}(r)} \quad \text{where } V_{s,p}(r) = V_{s,p}(\rho, \epsilon)$$

- Observe $t_{s,p}$ over ⊕ surface
- guess $\rho(r), \epsilon(r)$ to calculate $V(r)$
- integrate model, compare w/data
- correct guesses, repeat

ie. discovery of crust/mantle boundary (text):



- 2 sets of P, S
- 1st set, fast arrival, saw high ρ
- 2nd set, slow, low ρ crust

(Hartmann Teanspi's)

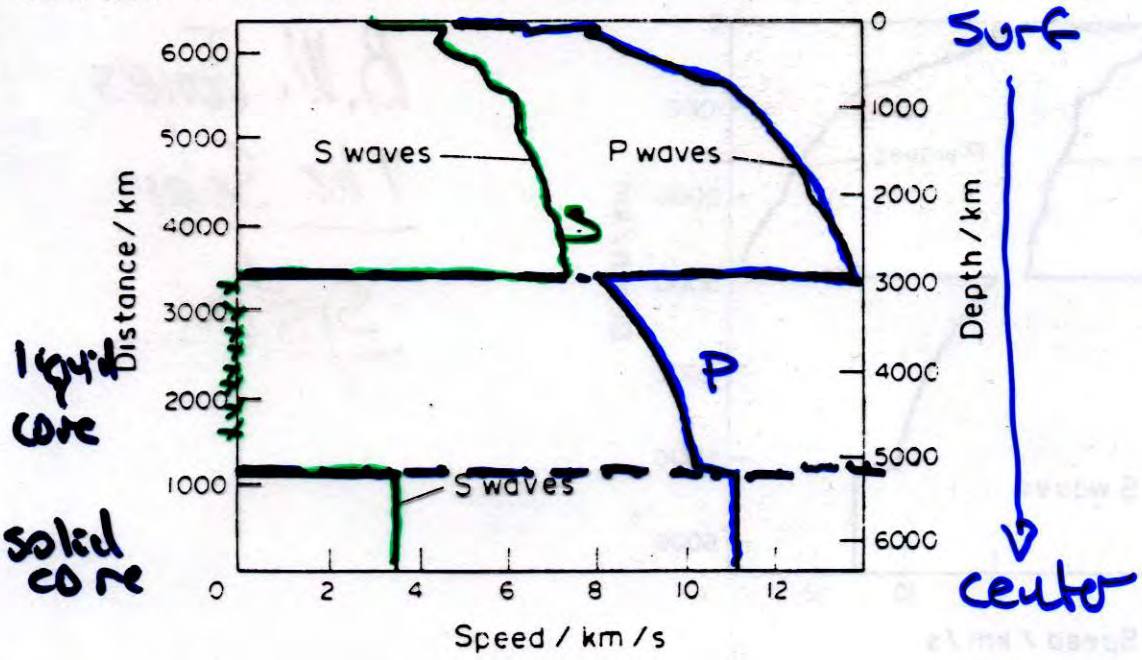
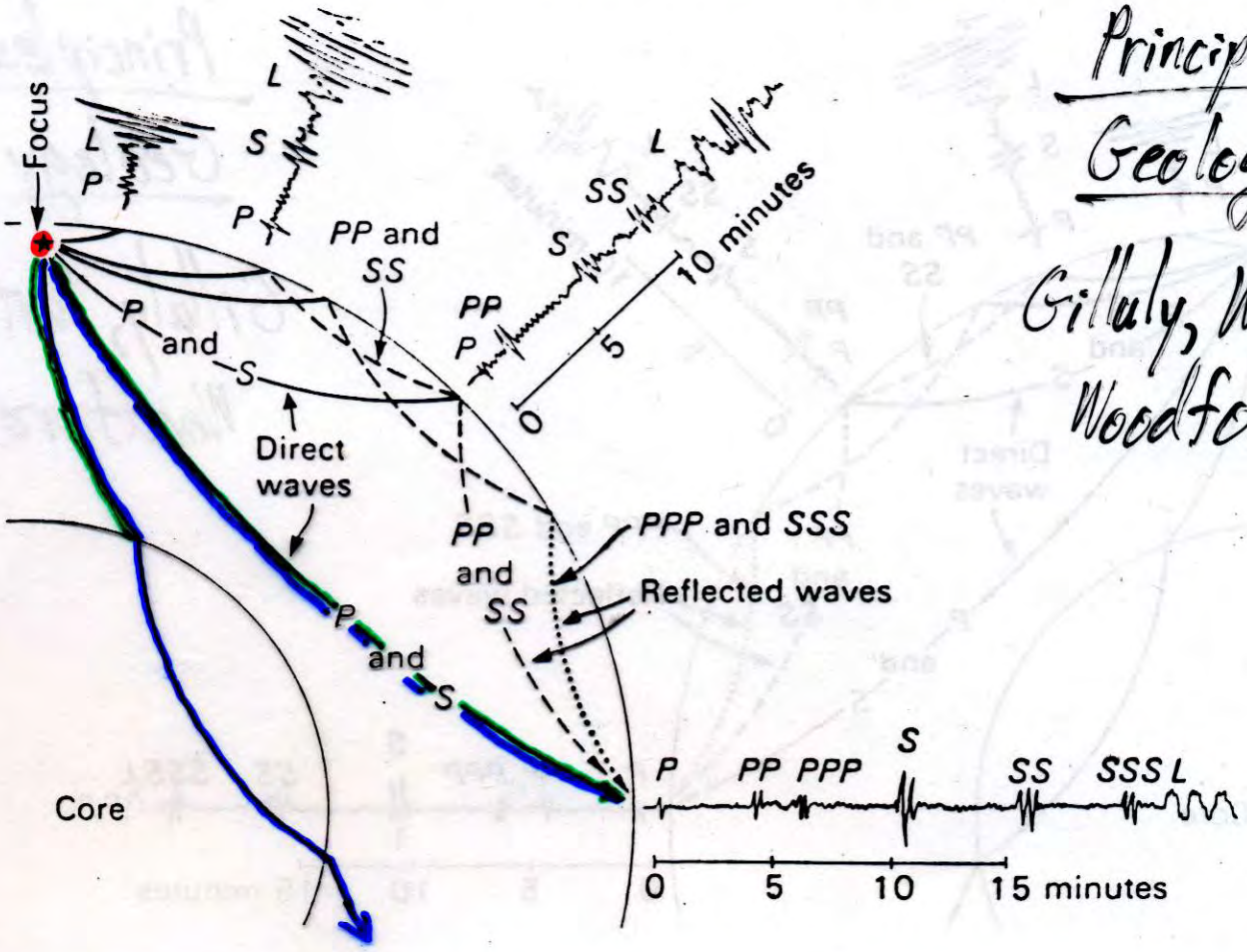


FIGURE 2.7
P-wave and S-wave speeds in the Earth.



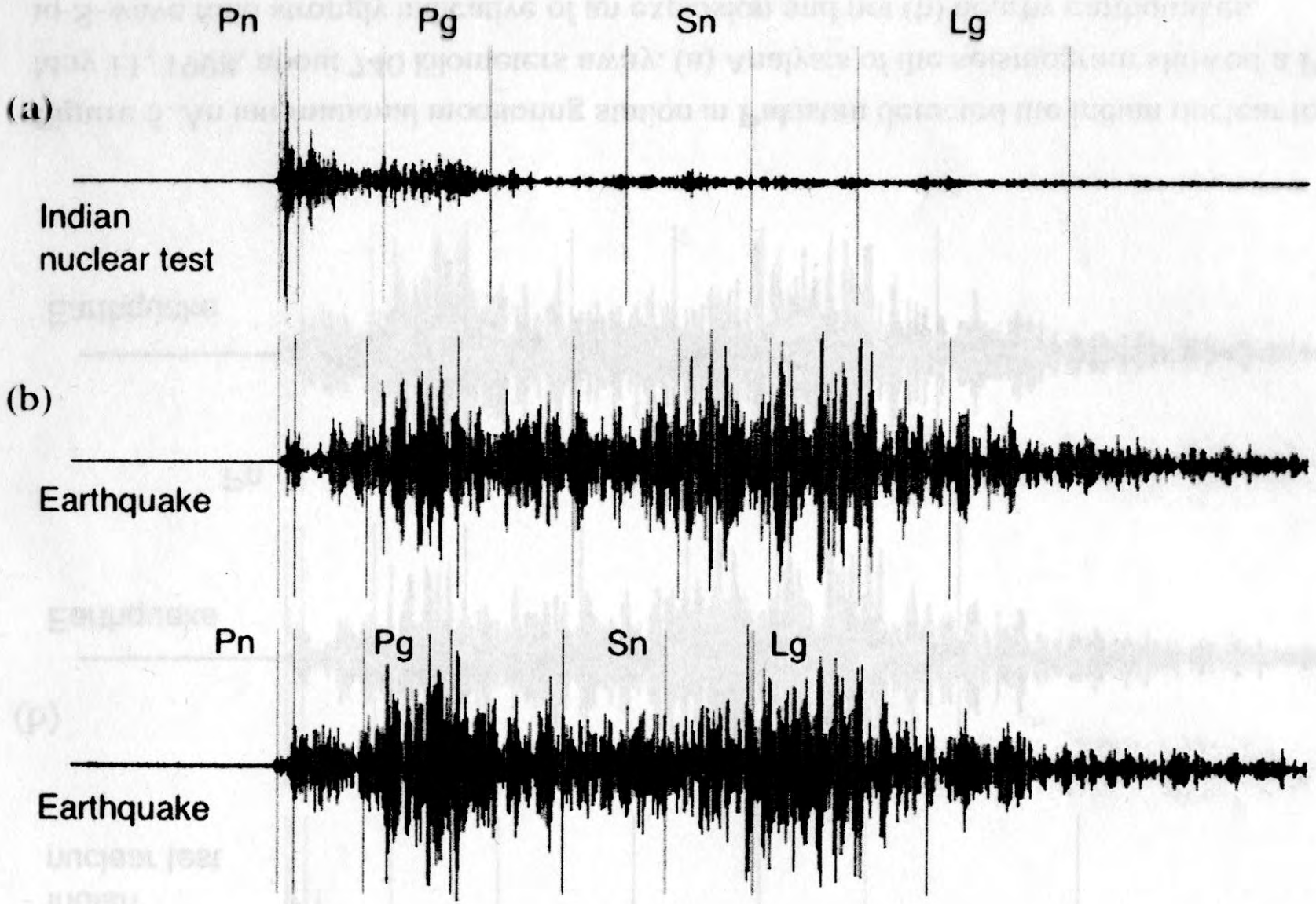
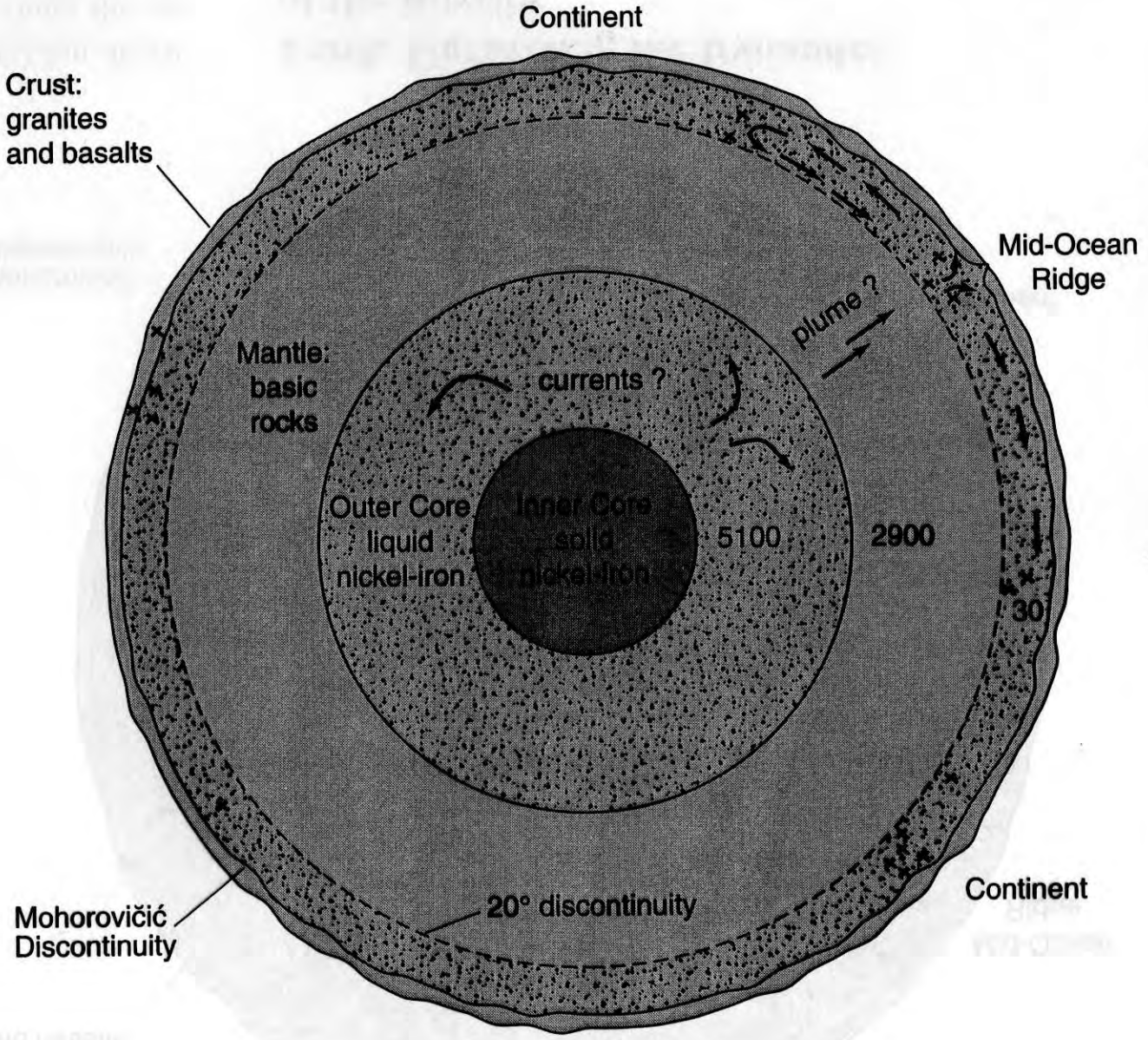


Figure 5. An international monitoring station in Pakistan detected the Indian nuclear test of May 11, 1998, about 740 kilometers away. (a) Analysis of the seismogram showed a P-wave-to-S-wave ratio strongly indicative of an explosion and not (b) nearby earthquakes.

re 8-13. Structure of Earth's
or as revealed by seismology, with
positions inferred from additional
nce. Earthquake positions are
ated by Xs. Stippling indicates
en material in the outer core and
ally molten material in the astheno-
re. Numbers indicate depth in
meters.



n the olivine to the spinel phase about 400 km down. ecognized by the 1960s, the heat flow through the top ne mantle under continents is substantially less than heat flow out of the mantle under the seas; hence,

Earth: Unraveling the Dynamics of the Interior

For centuries, naturalists pondered features of Earth's sur-

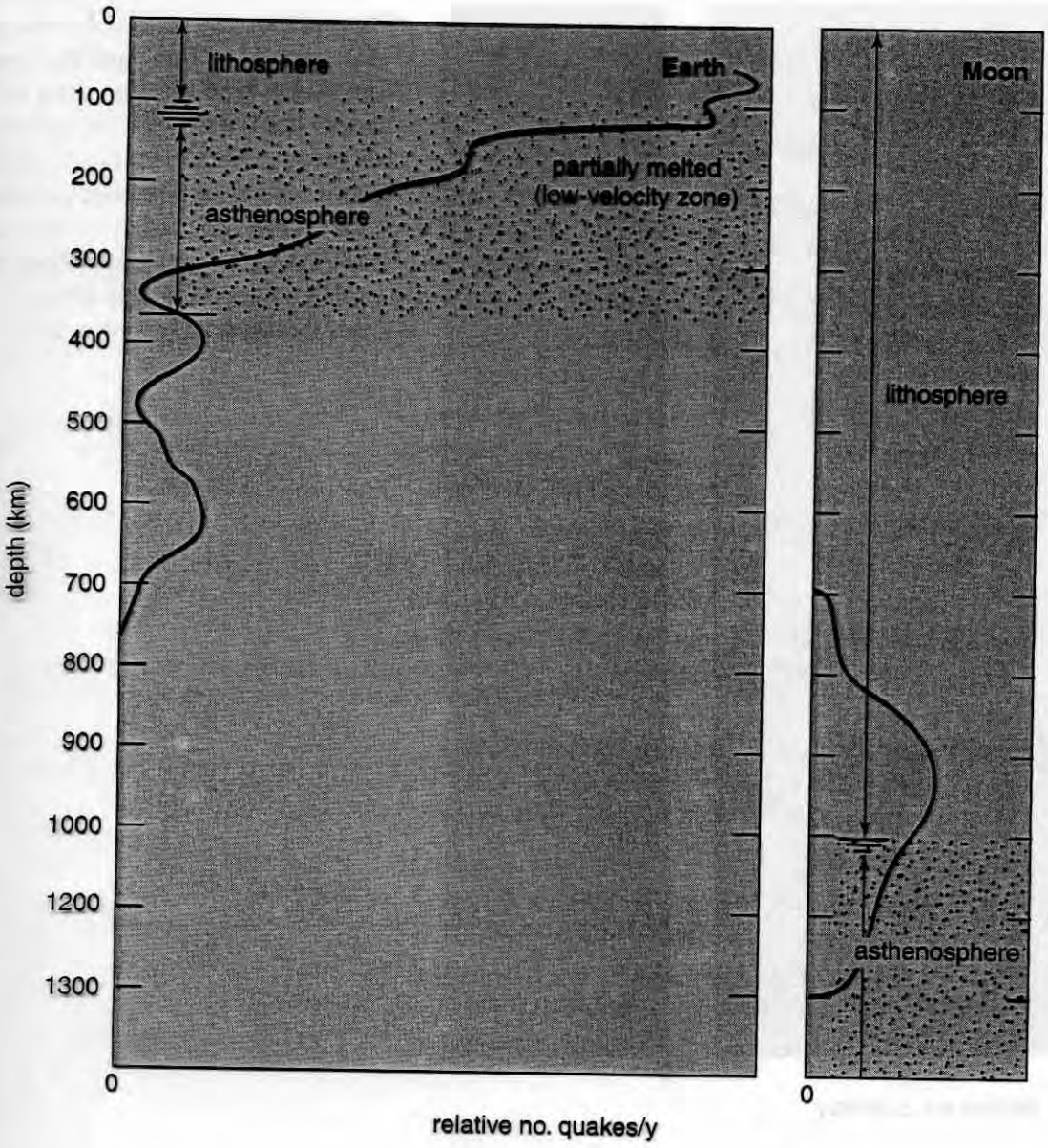


Figure 8-9. Depth distributions of "earthquakes" on Earth and the moon. Earthquakes tend to occur along the bottom edge of the solid lithosphere and the top of the more plastic asthenosphere, where currents build up stresses in the rocks. Moonquakes number near 3,000/y but are much less energetic than larger earthquakes, so that the two scales are not directly comparable.

in the moon, which is seismically much quieter than Earth. During the first year of lunar observations, lunar quakes rated only 1 to 2 on the well-known Richter scale, contrasted with ratings of 5 to 8 for major terrestrial quakes. The total seismic energy released annually in the moon is estimated at about 10^7 J (Latham, 1971), contrasted with 5×10^{17} J in Earth.

As shown in Figure 8-9, the quakes inside the moon occur in a band around 800 to 1000 km in depth, a level that is interpreted as the base of the lunar lithosphere. Various geophysical measures, such as seismic data and [illegible]

Thermal Histories of Planets

Definition of the Problem

Most of this chapter so far has dealt with ways of determining the **statics of planets**—that is, descriptions of their interiors in their present-day states. But planets are evolving systems. We must therefore also come to grips

~~Other~~ important seismic results (4)

① - CRUST - MANTLE BDRY

② - DISTINCT CORE - MANTLE BDRY
- slow region deep → low speed waves in Fe core

③ - slow zone in UPPER mantle (~250 km)
- Partially molten mantle → convecting Silly Putty, fluid driven, tectonics ↑ asthenosphere

④ - S-wave shadow zone
- liquid Fe, outer core

LUNAR Seismology

- Apollo Seismographs + Saturn V 3rd stage impacts
  - much thicker crust
  - deeper natural quakes
  - possible deep asthenosphere

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CURT (I was in Baltimore)
-+ examples of planetary interiors

Finally

Magnetic Fields

- generated by moving a magnet conductor
- DYNAMO
  - planetary fields generated by circulating conducting fluids
    - ⊕ molten outer Fe core
- long-term changes
  - field reversals via circulation changes
  - field decrease as core cools/solidifies
- residual fields after freezing are small

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mG = 0.001 G

Sun	2 G
Mercury	2.2 mG ← why Big?
Venus	< 0.3 mG
Earth	~~0.31 mG~~ 305 mG
Moon	0.1 mG
Mars	0.4 mG
Jupiter	4.2 G
Saturn, Uj	~~10~~ 200 mG

EUROPA 2.4 mG !

] why big?

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