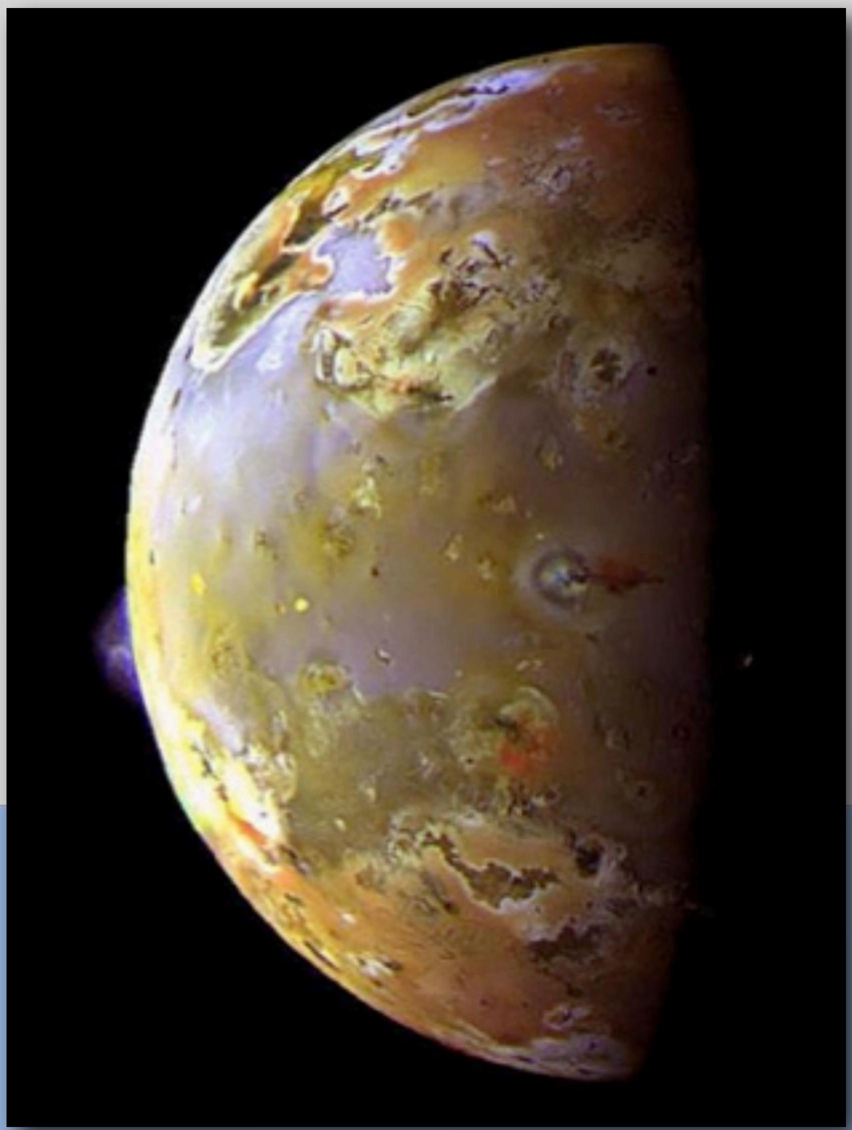
Planetary interiors

Planetary Sciences Chapters 6 & 5

Daphne Stam



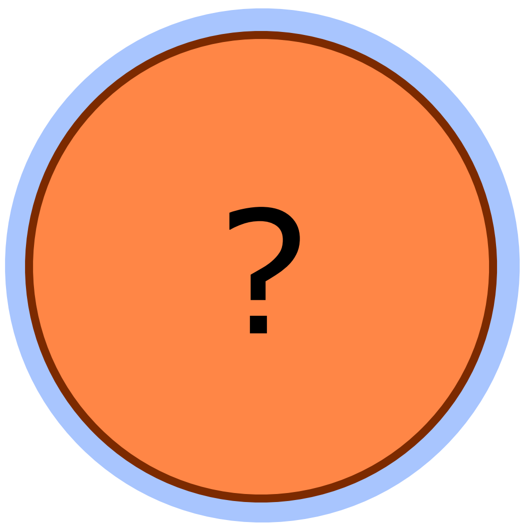
SRON
Netherlands Institute for Space Research

Outline

- Studying the interior structures of planets
- Magnetic fields
- Hydrostatic equilibrium
- A planet’s gravity field
- Interiors of the gaseous planets
- Interiors of the terrestrial planets
- Planet quakes
- Volcanoes

Studying the interior structures of planets

What can we say about the interior structure of a planet?



Of Earth and the moon there is seismic data, which gives information on the transport of waves through the interior, hence on the deepest density variations.

For another planet or moon we should use its:

- average density (mass/volume)
- rotational period and geometric oblateness
- gravity field
- magnetic field
- total energy output
- atmospheric and surface characteristics

Measuring a planet's size*

- A bodies' diameter equals its angular size times its distance from the observer. In the Solar System, distances can be derived from the orbits.

  From Earth, angular sizes are difficult to estimate due to the limited spatial resolution of observations!

- The diameter of a body can be derived when the body occults a star.

  This doesn't happen too often!

- For solid bodies, radar echoes can be used.

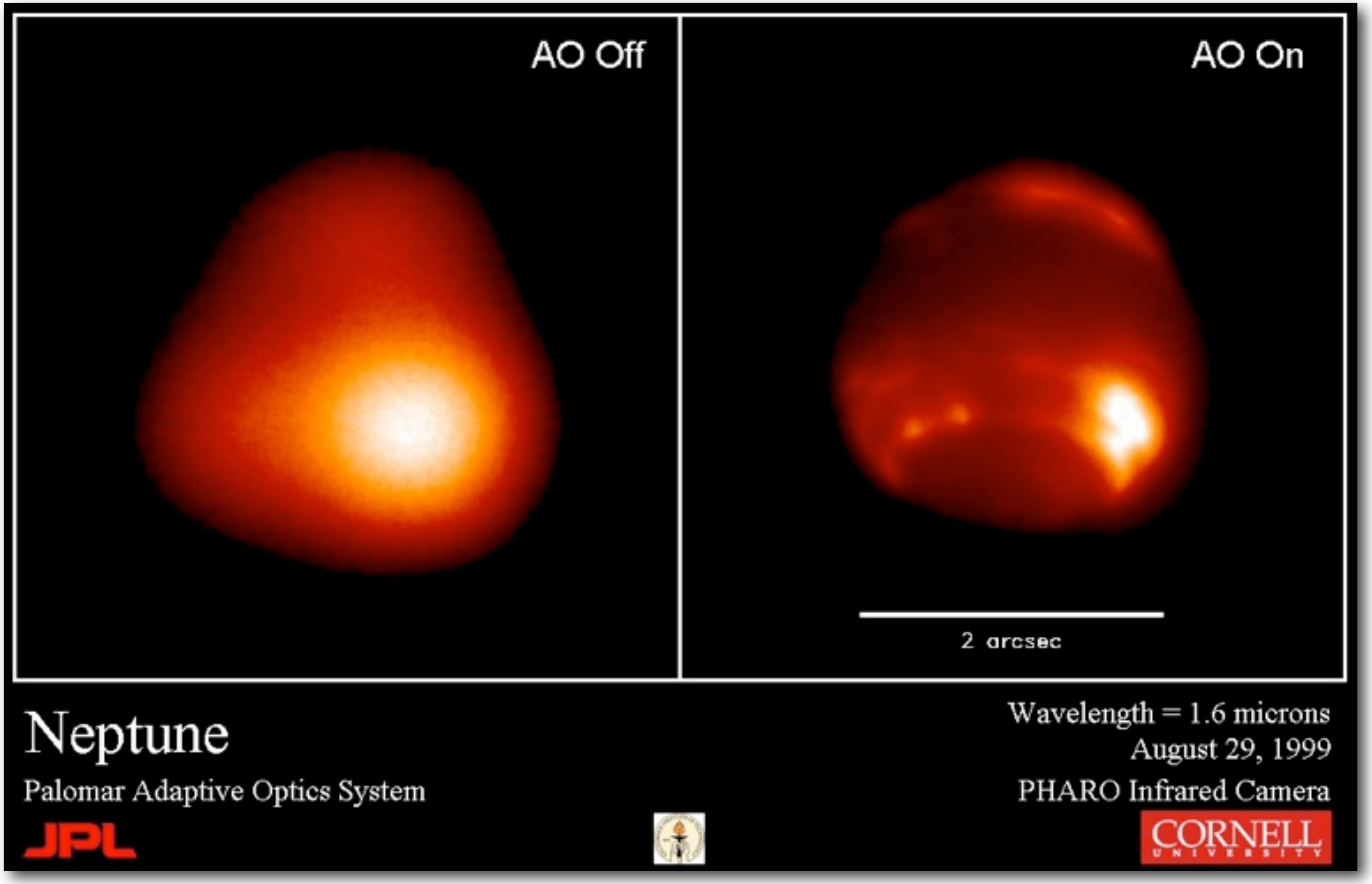
  The signal drops off as $1/r^4$, with r the distance to the body!

- From triangulation with e.g. a lander-orbiter combination.

  Not an option for gas planets!

* By measuring the equatorial and polar radii, a body's shape can also be determined

An example of planet observations from the Earth:



Near-IR observations with a 5-m telescope with and without Adaptive Optics

Measuring a planet's mass

- The orbital periods of moons can be used to solve for the mass.

  You obtain the sum of the masses of the planet and the moons!

- The gravity of each planet perturbs the orbits of the other planets.

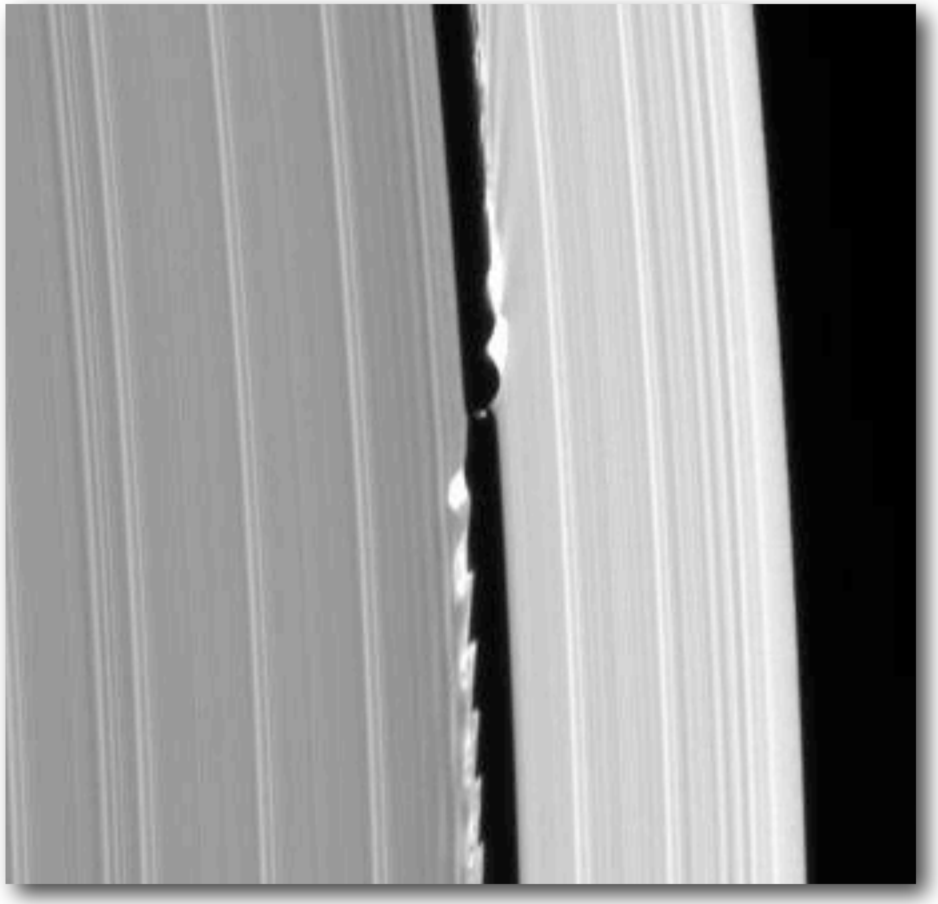
  Neptune was discovered thanks to its perturbation on Uranus' orbit!

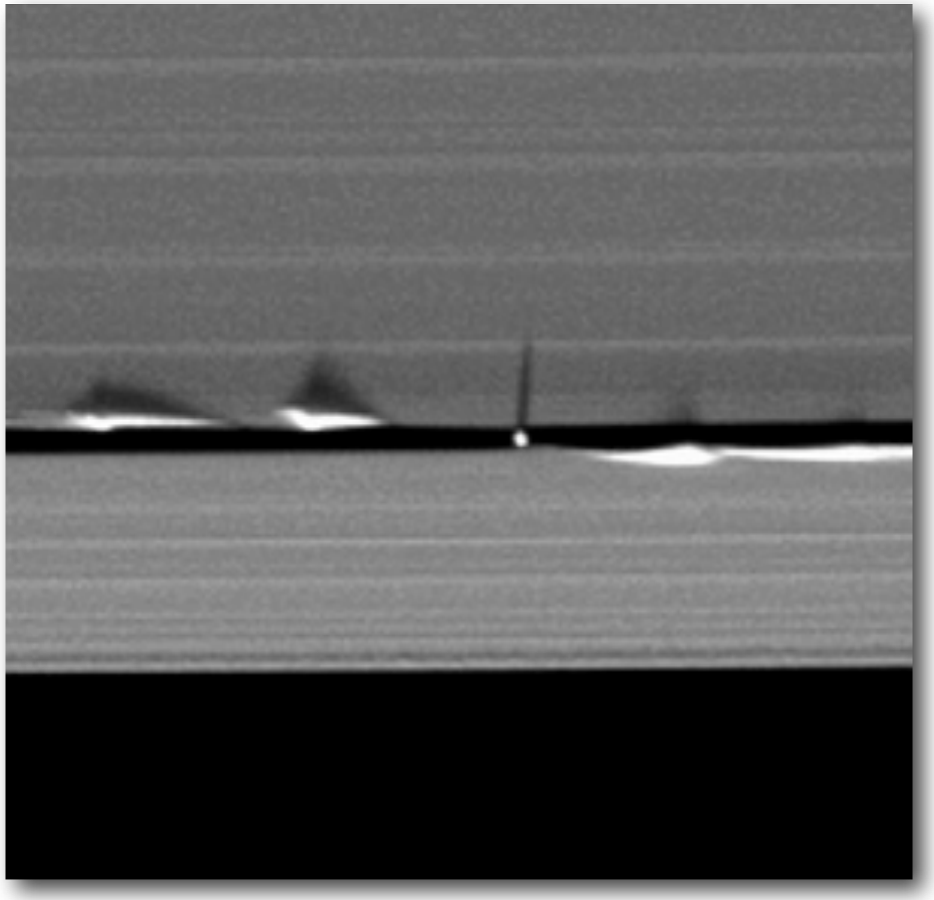
- The orbits of spacecraft are perturbed when they orbit a planet or when they have a fly-by.

  With the Voyager fly-by, the mass of Neptune was revised, and left-over anomalies in Uranus' orbit could finally be explained! (Standish, 1993)

- Planetary rings can show density waves that are caused by moons, and that can give information about their masses.

An example of moons leaving waves in Saturn’s rings:

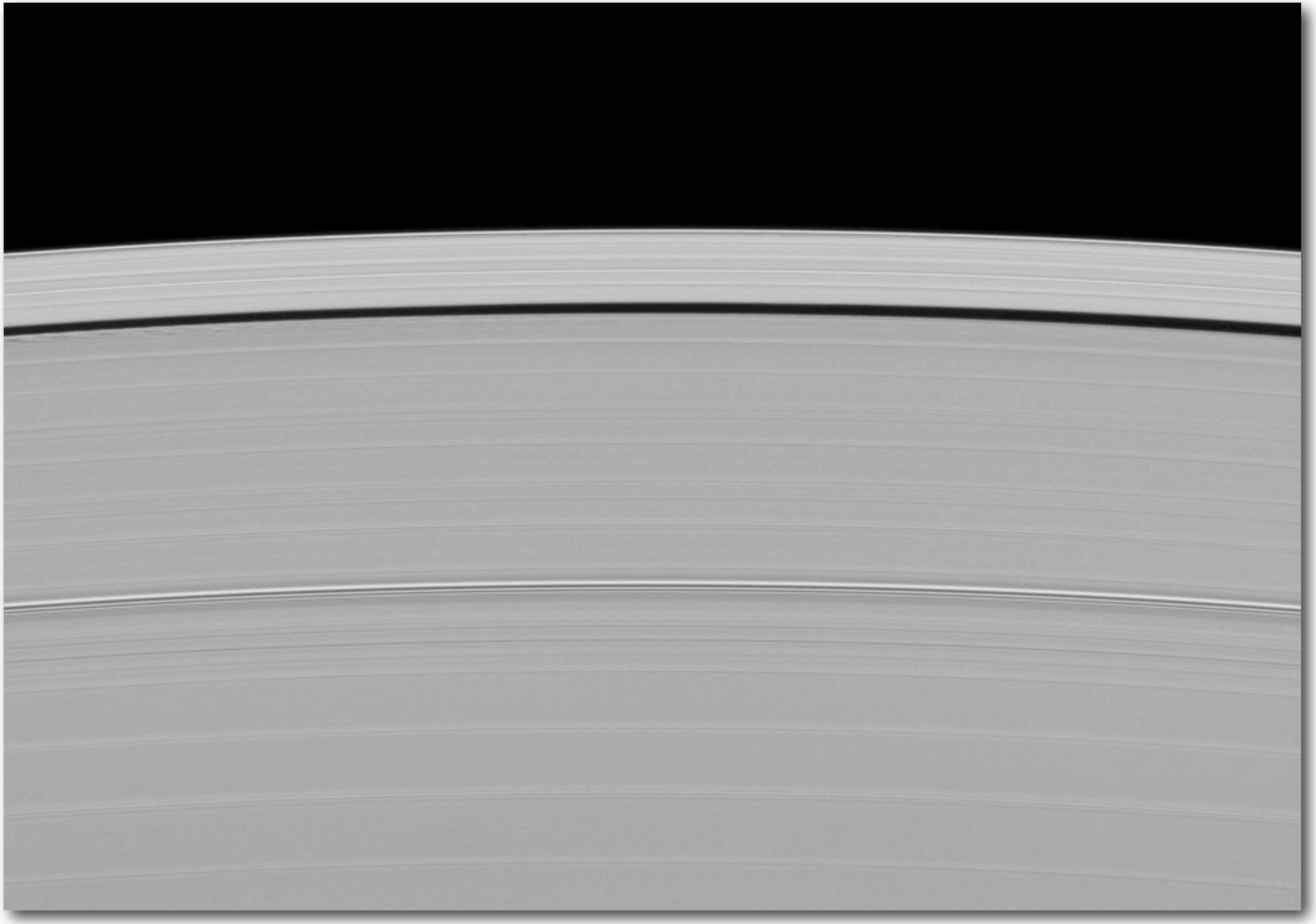




The tiny (8 km wide) moon Daphnis causes waves in Saturn’s A ring, seen from above, casting shadows during the recent (August 2009) ring-plane crossing (Credit: NASA/JPL/SSI). The waves rise ~ 1 km up from the ring surface, while the rings themselves are ~ 10 m thick!

The moon’s mass is about 7×10^{13} kg (Weiss, Porco & Tiscareno, 2009).

A movie of the waves:



Daphnis causing waves in Saturn's A ring, casting shadows during the recent (August 2009) ring-plane crossing (Credit: NASA/JPL/SSI).

Information in a planet's average density

The average density of a planet gives an indication of its composition:

- For a small object, a density $\rho \leq 1$ g/cm^3 implies an icy and/or porous object

- For a large object, a density $\rho \leq 1$ g/cm^3 implies it consists primarily of H_2 and/or He

- A density $\rho \sim 3$ g/cm^3 suggests a rocky object

- Larger densities indicate the presence of heavy elements, in particular iron ($\rho \sim 7$-8 g/cm^3), one of the most abundant heavy elements

Characteristics of Solar System planets

Planet	$R_{equator}$ (in km)	Oblateness*	Density (g cm^{-3})	Central P (Mbar)	Central T (K)
Mercury	2440	-	5.43	~ 0.4	~ 2000
Venus	6042	-	5.20	~ 3	~ 5000
Earth	6378	0.0034	5.52	3.6	6000
Moon	1738	0.0012	3.34	~0.045	~ 1800
Mars	3390	0.0065	3.93	~ 0.4	~ 2000
Jupiter	71,492*	0.0649	1.33	~ 80	~ 20,000
Saturn	60,268*	0.0980	0.69	~ 50	~ 10,000
Uranus	25,559*	0.0229	1.32	~ 20	~ 7000
Neptune	24,766*	0.0171	1.64	~ 20	~ 7000

Table 6.1 from *Planetary Sciences* (oblatenesses from Tables 1.2 and 1.3)

* A bodies' oblateness or flattening ε is defined as $\varepsilon = (R_{equator} - R_{pole})/R_{equator}$

* This radius is measured at a pressure of 1 bar

Magnetic fields

Most Solar System planets have global magnetic fields, that are generated by an internal dynamo. Requirements:

- a rotating body

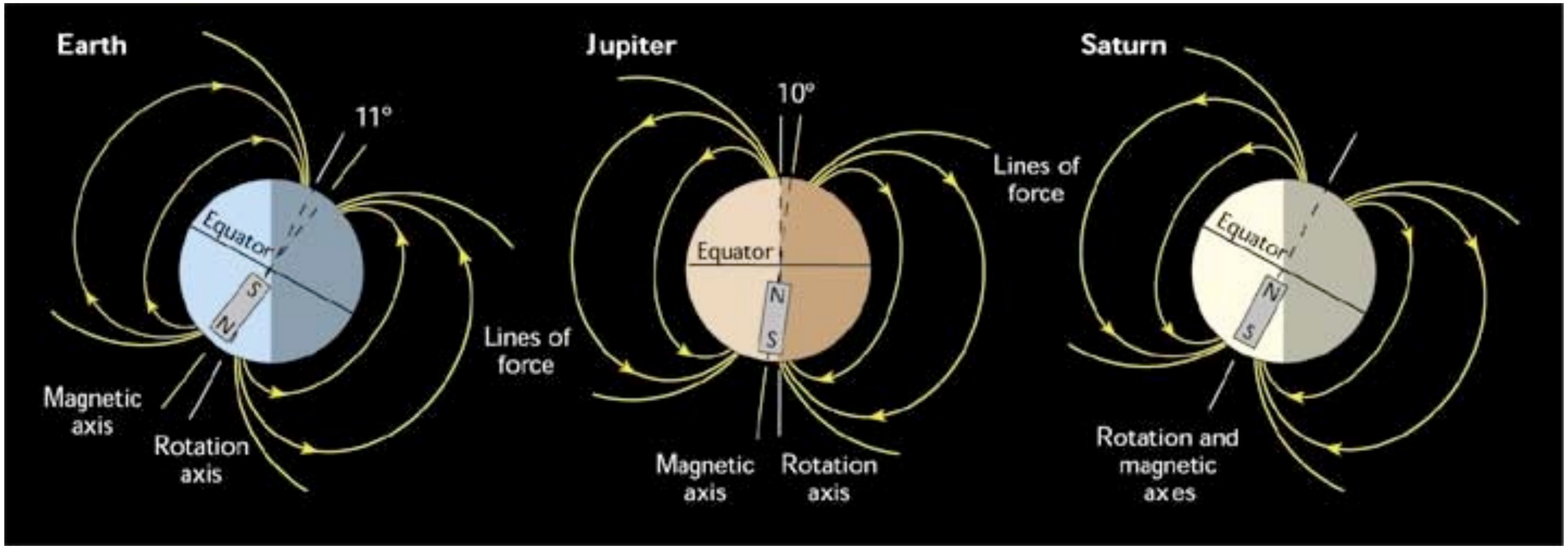
  the necessary rotation rate is unknown (Venus might rotate too slowly)

- containing a fluid, electrically conducting region

  molten rock/iron, metallic hydrogen, mixtures of “ionic” ices, ...

- within which convective motion occurs

  this requires a temperature gradient across the fluid, conducting region

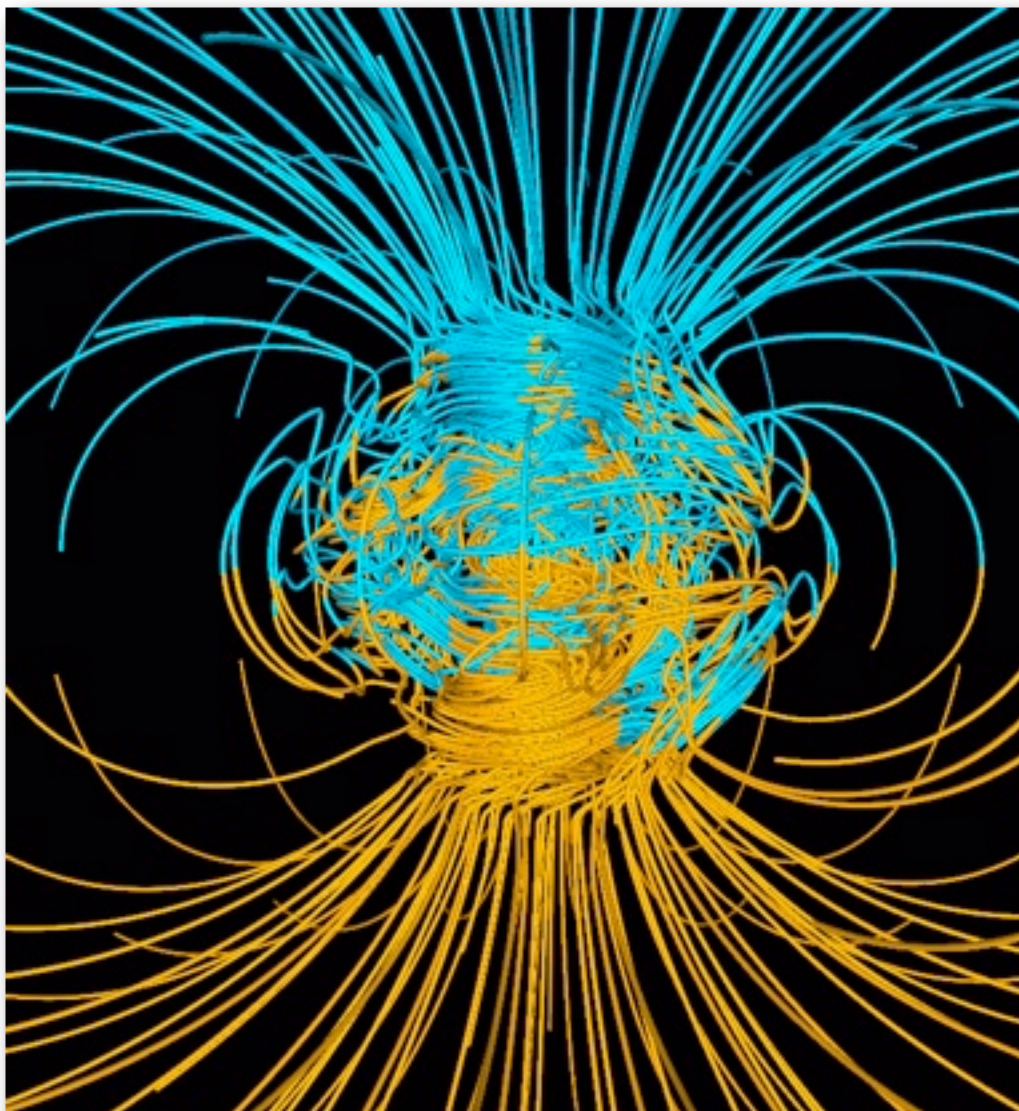


The simplest magnetic fields have dipole configurations

Credit: The New Solar System [1999]

Magnetic field: the geodynamo

In close-up, the Earth's magnetic field is very complicated, with chaotic internal behaviour, on the outside it deviates from a dipole, too:



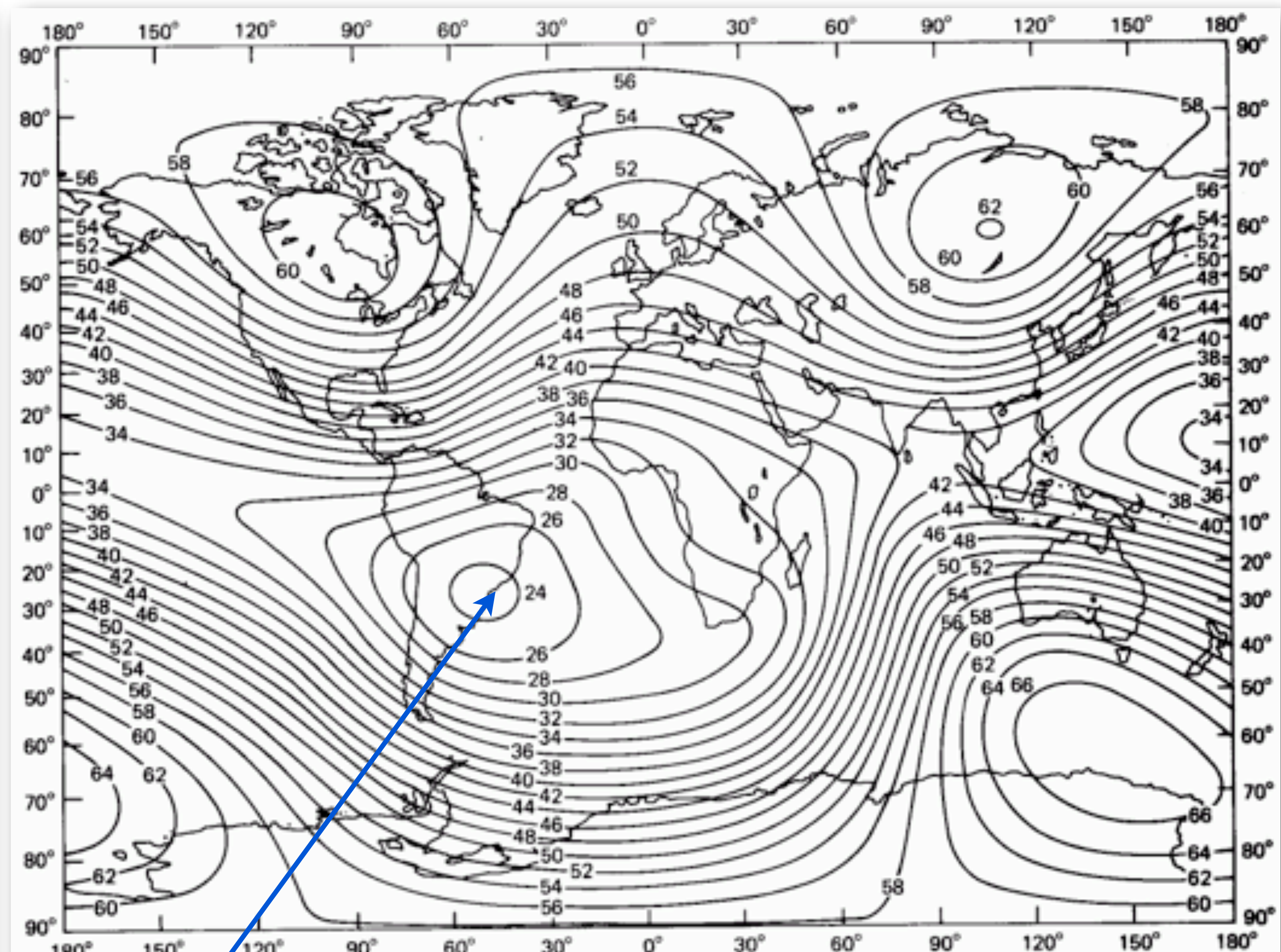


Figure 1-18 Present-day magnetic field of the earth. (a) Magnitude, μT.

Credit: Blakely (IGRF 1990)

South Atlantic Anomaly

Magnetic field: the geodynamo

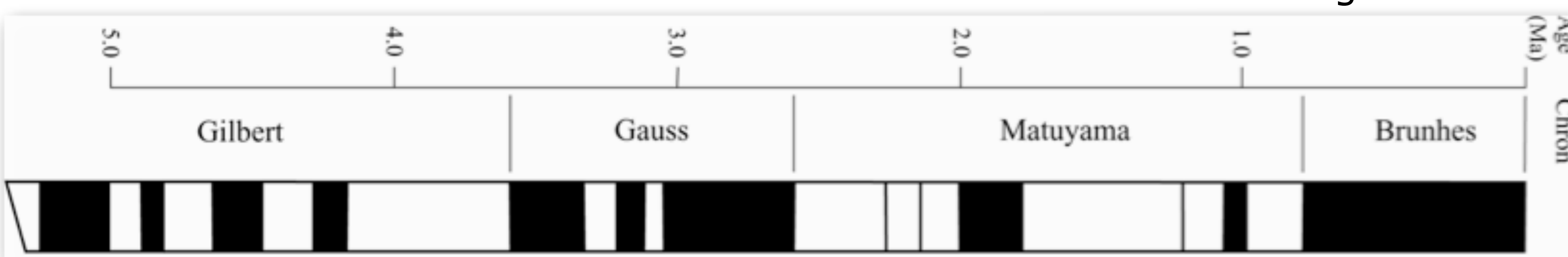
The polarity of the Earth’s magnetic field changes in time. Reversals occur ~ every 300.000 years (last one took place > 700.000 years ago):

Ancient magnetic fields can be studied from ancient lava flows with ‘frozen in’ field lines.

It is unknown how reversals happen:

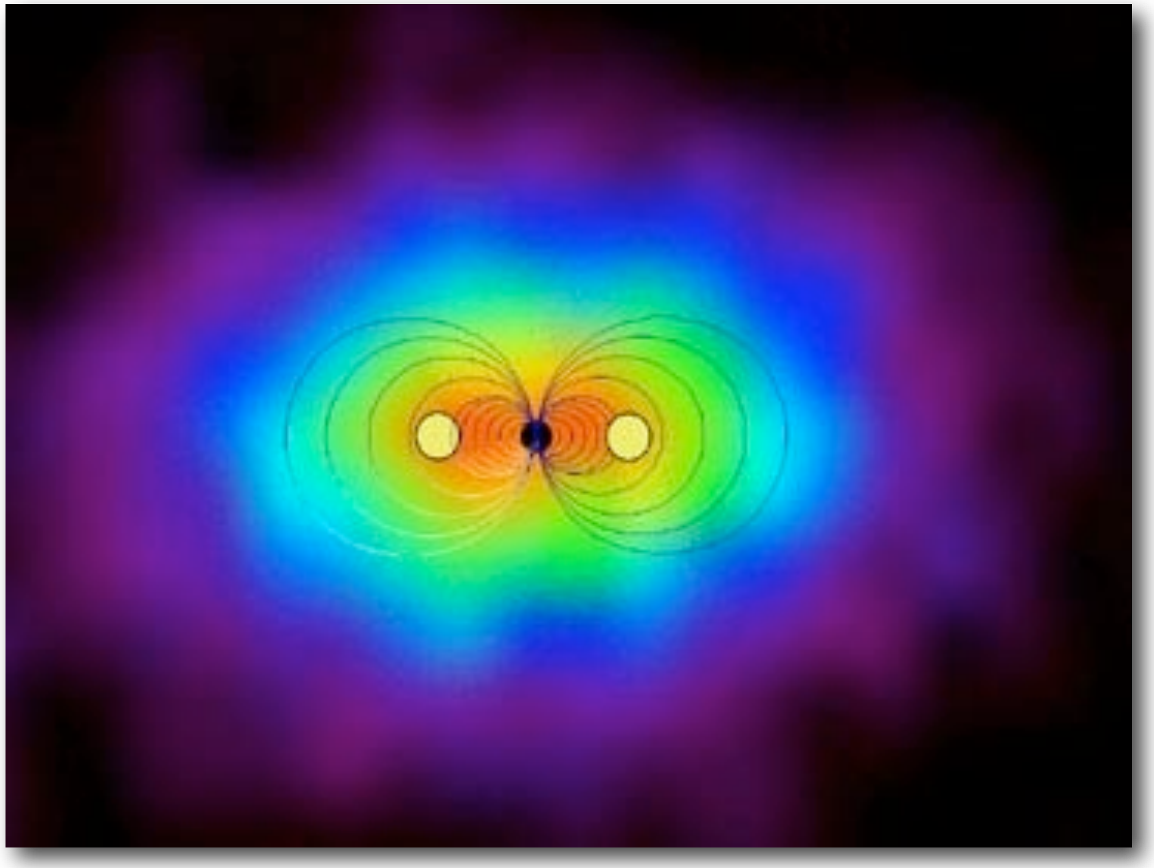
- A gradual (~1000 years) shift of the poles
- A magnetic field vanishes, starts up again reversed

The polarity of the Earth’s field changes in time:

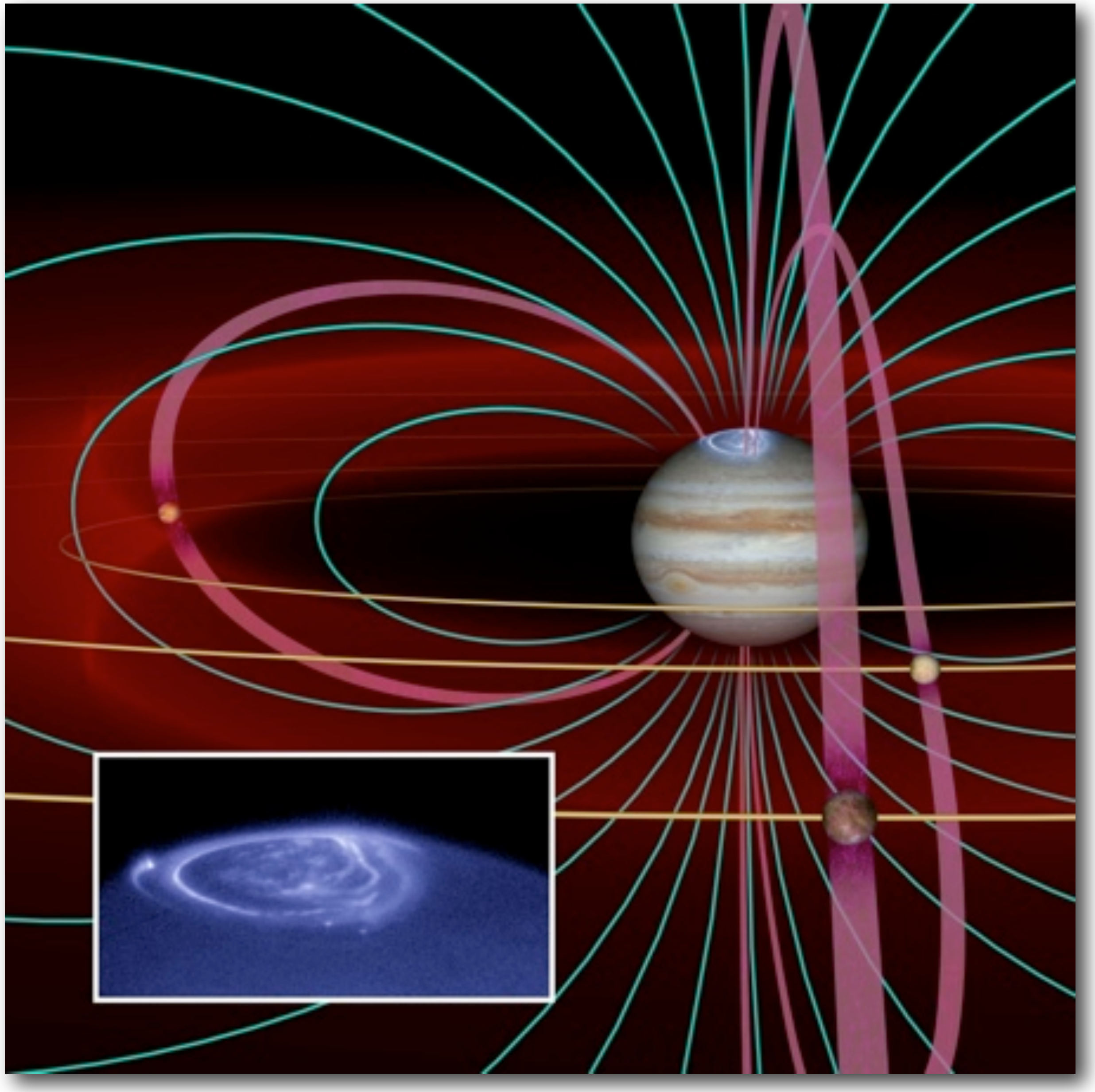


The majestic magnetic field of Jupiter

Jupiter's magnetosphere has a radius of 50-100 R_J. The Galilean satellites orbit the planet within the intense radiation field in the inner parts of the magnetosphere.



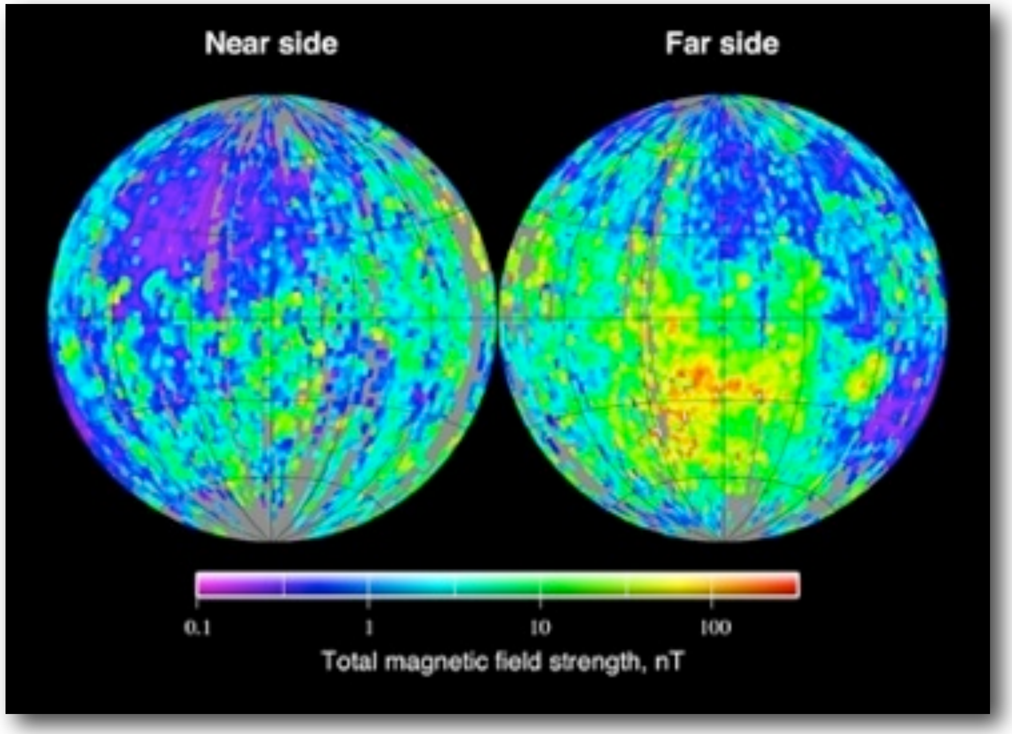
Black dot: Jupiter; Yellow dots: Io's orbit.
Credit: Cassini mission NASA/ESA



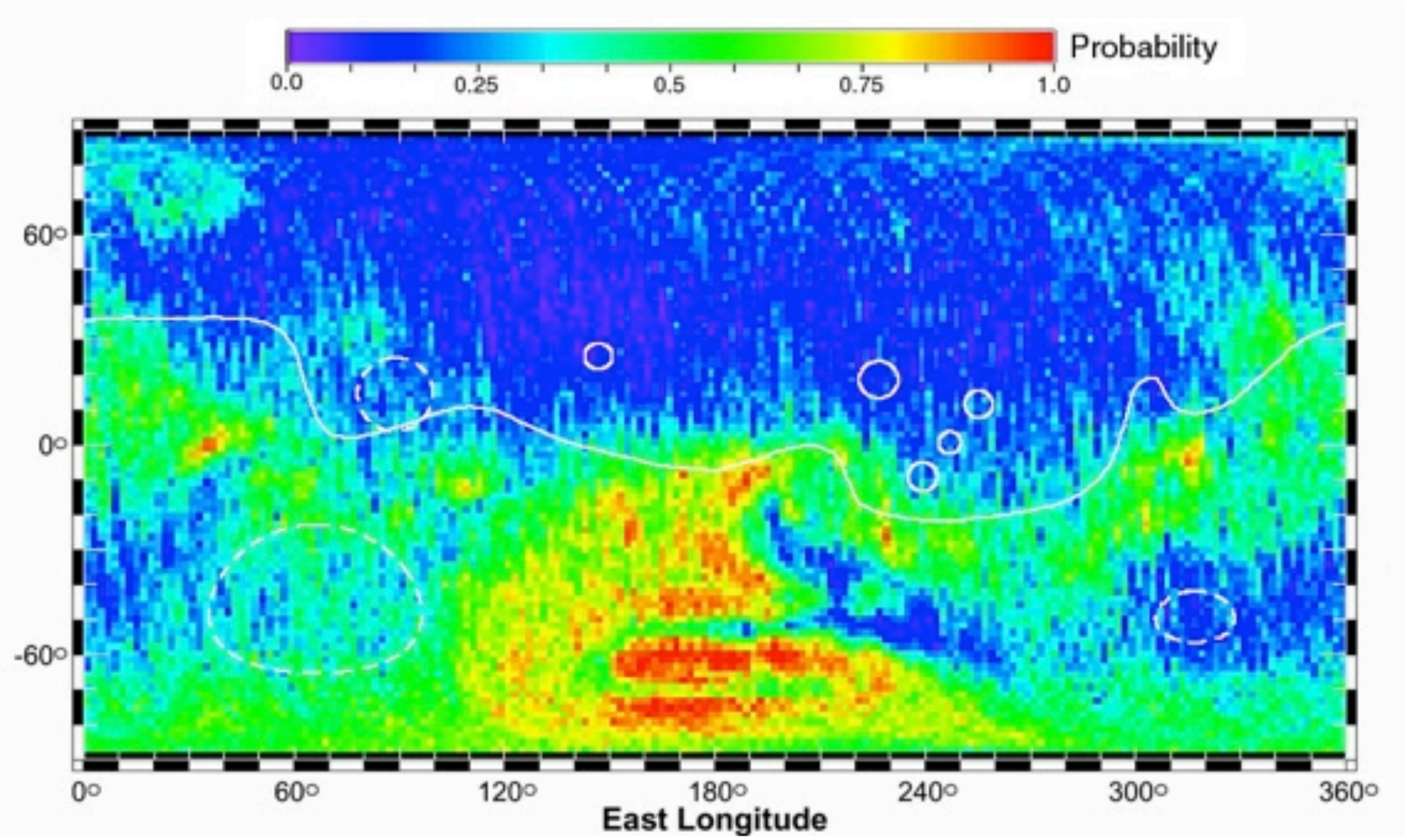
Credit: John Spencer (drawing), HST (image)

Remnant magnetic fields

Weak magnetic fields can be due to the presence of an internal dynamo in the past, or it can be frozen in from the era of planet formation. Processes such as impact cratering can significantly change these fields.

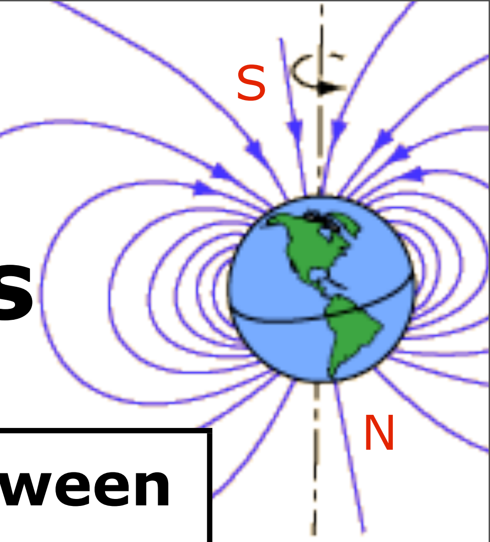


The magnetic field of the moon.
Credit: Lunar Prospector (NASA)



The Martian magnetic field (Credit: MGS/NASA)

Planetary magnetic field characteristics



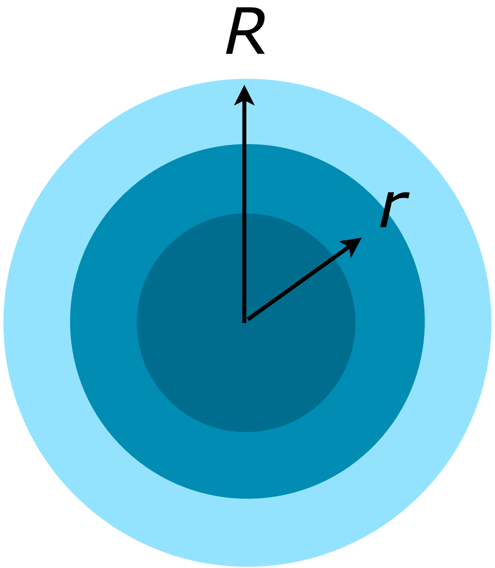
Planet	Dipole moment (Earth=1)	Polarity (same as Earth)	Angle between axes
Mercury	0.0007	yes	14°
Venus	< 0.0004	-	-
Earth	1	yes	10.8°
Moon	~ 0	-	-
Mars	< 0.0002	-	-
Jupiter	20,000	no	9.6°
Saturn	600	no	< 1°
Uranus	50	no	59°*
Neptune	25	no	47°*

* Offset from the centre of the planet

Hydrostatic equilibrium

To first order, the internal structure of a spherical body is determined by *hydrostatic equilibrium*, the balance between gravity and pressure:

$$P(r) = - \int_r^R g(r') \, \rho(r') \, dr'$$



With P the pressure, r the distance to the centre, R the radius, g the acceleration of gravity, and ρ the density.

Assuming a constant density throughout the planet, the pressure in the centre of the planet equals:

$$P_c = \frac{2\pi \, G\rho^2 R^2}{3} = \frac{3 \, GM^2}{8 \, \pi R^4}$$

This is a good estimate for small, homogeneous bodies like the moon

The Equation of State

A general description of the pressure-density can be written as:

$$P = K\,\rho^{(n+1)/n}$$

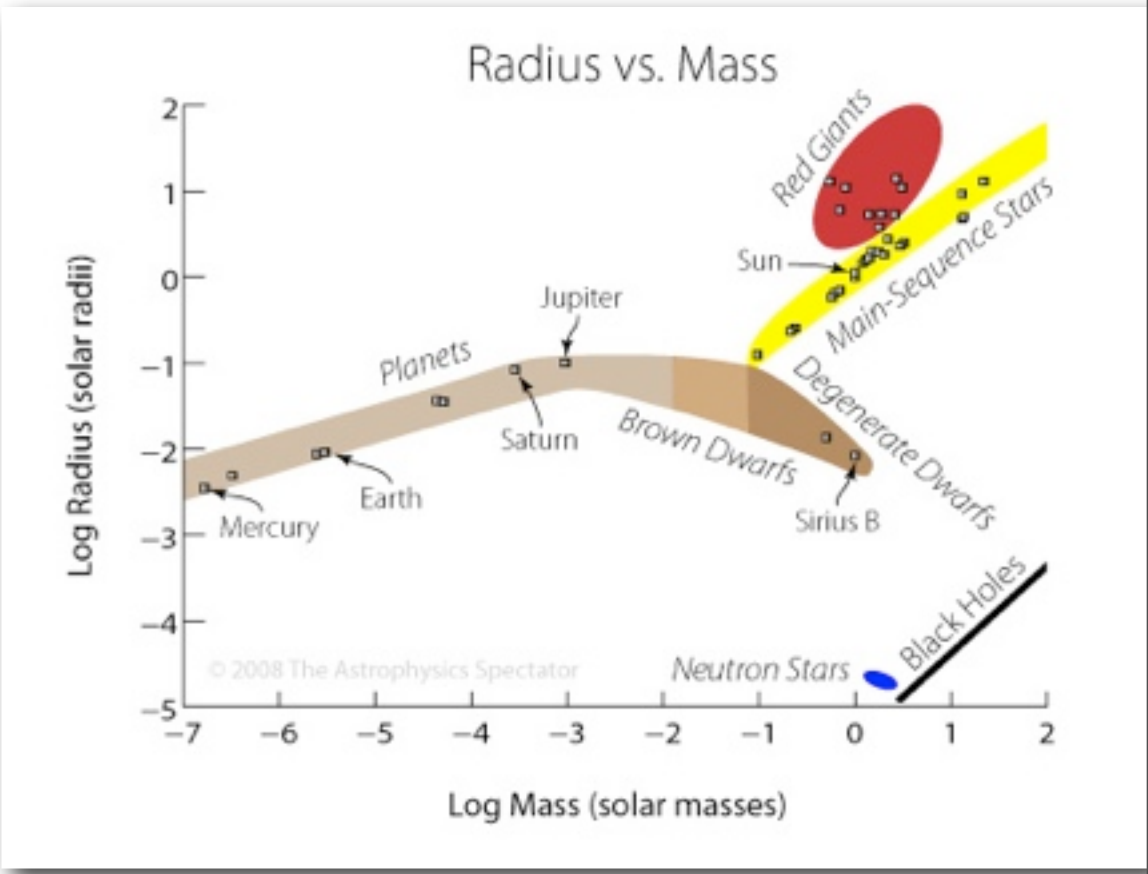
where K is the polytropic constant and n the polytropic index.

At very low pressures, $n \approx \infty$
At very high pressures, n=3/2, and $P \propto \rho^{5/3}$

For an incompressible planet: $M \propto R^3$

When the internal P increases, material will become compressed, and adding mass will increase R slower and slower.

When the internal P gets very high, material will become degenerate, and adding mass will decrease R: $M \propto R^{-3}$



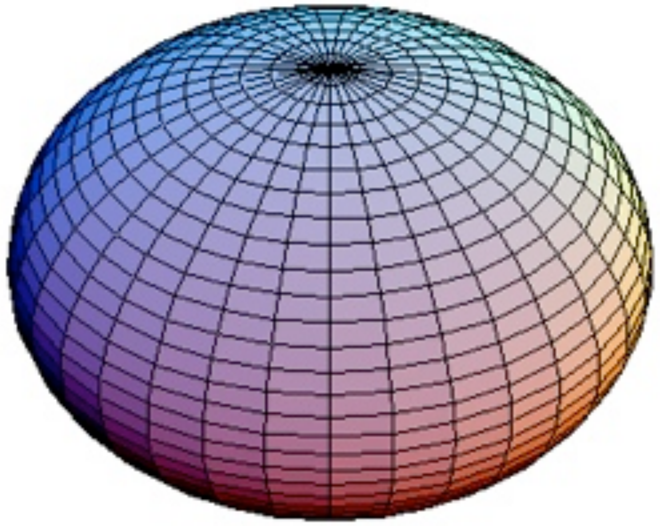
A planet's gravity field

Internally, a planet is usually not homogeneous, but differentiated. Information about its internal structure can be derived from its gravity field.

The gravitational potential at a distance r of a non-rotating (spherical) fluid-like* body with mass M in hydrostatic equilibrium is:

$$\phi_g(r) = -\frac{GM}{r}$$

More general, the gravity potential of an axisymmetric body, with the origin at its centre of mass, is given by:



$$\phi_g(r,\theta,\varphi) = -\frac{GM}{r}\left[1 - \sum_{n=2}^{\infty}\left(\frac{R_e}{r}\right)^n J_n\, P_n\,(\cos\theta)\right]$$

With θ the colatitude, φ the azimuthal angle, R_e the equatorial radius of the planet, J_n the gravitational moments and P_n Legendre polynomials.

* fluid-like: deformable over geological time-scales (i.e. > milions of years)

Legendre polynomials

The Legendre polynomials are given by: $P_n(x) = \dfrac{1}{2^n\, n!} \dfrac{d^n}{dx^n} (x^2 - 1)^n$

The first few Legendre polynomials are:

$P_0(x) = 1$

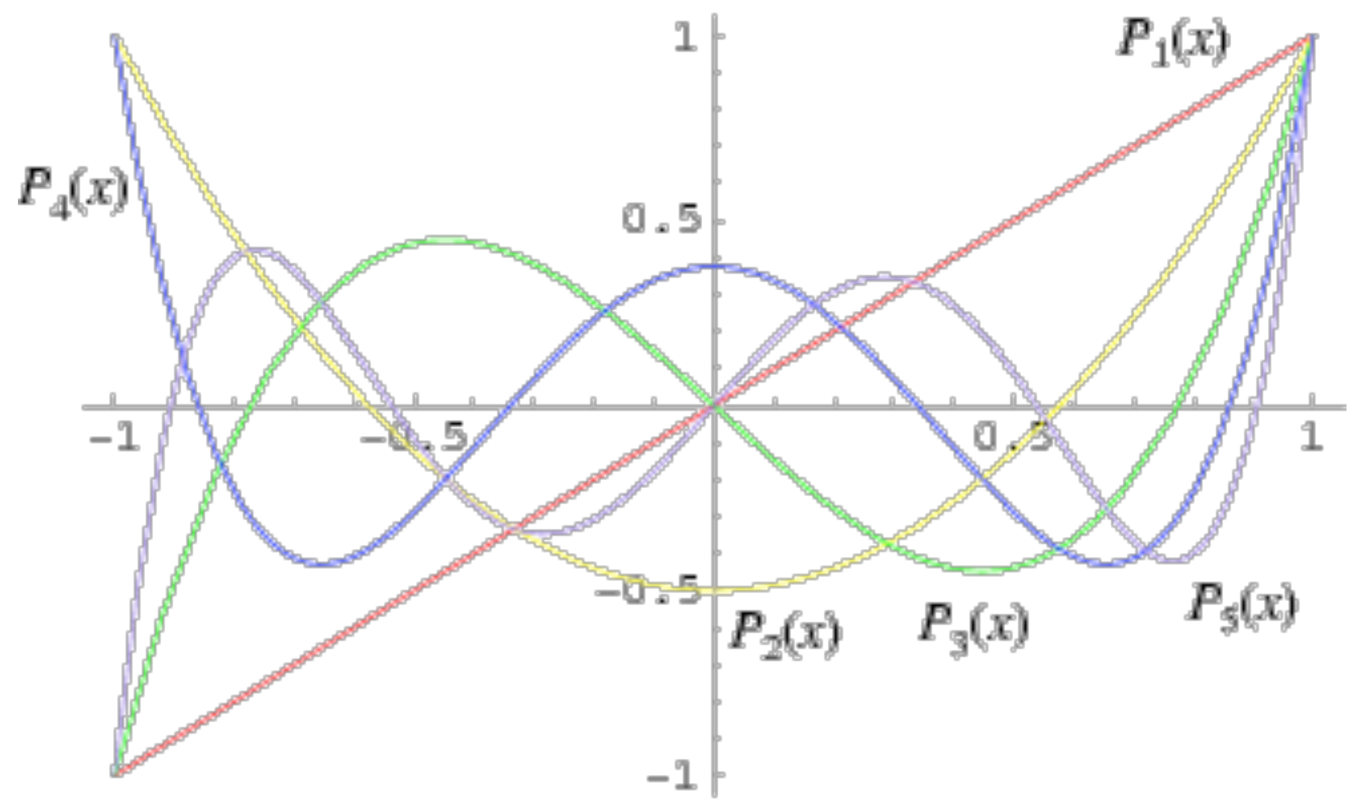
$P_1(x) = x$

$P_2(x) = (3x^2 - 1)/2$

$P_3(x) = (5x^3 - 3x)/2$

$P_4(x) = (35x^3 - 30x^2 + 3)/8$

$P_5(x) = (63x^5 - 70x^3 + 15x)/8$



http://mathworld.wolfram.com/LegendrePolynomial.html

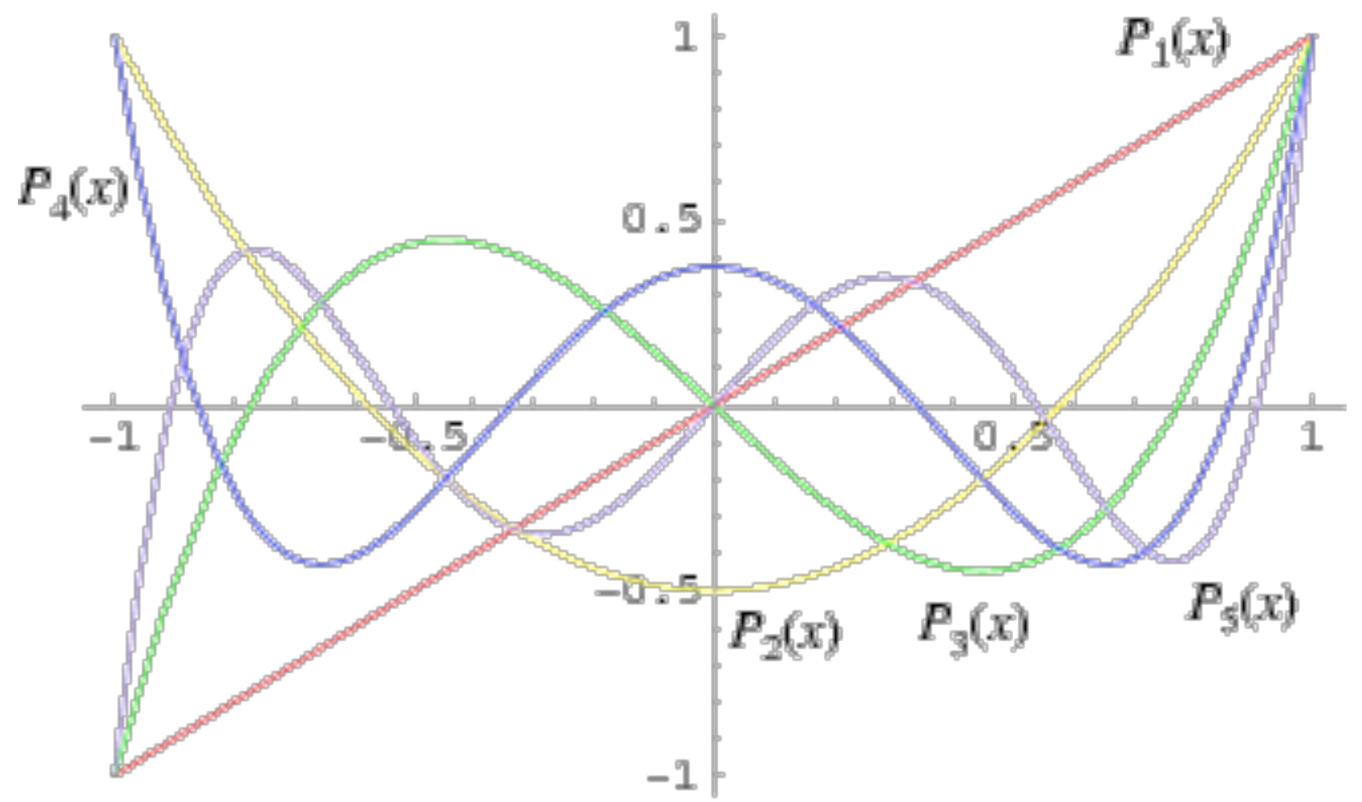
A planet's gravity field

$$\phi_g(r,\theta,\varphi) = -\frac{GM}{r}\left[1 - \sum_{n=2}^{\infty}\left(\frac{R_e}{r}\right)^n J_n\, P_n\,(\cos\theta)\right]$$

- For a non-rotating fluid body in hydrostatic equilibrium, J_n=0
- Rotating fluid bodies in hydrostatic equilibrium have J_n=0 for odd n

This is a good approximation for the gaseous planets!

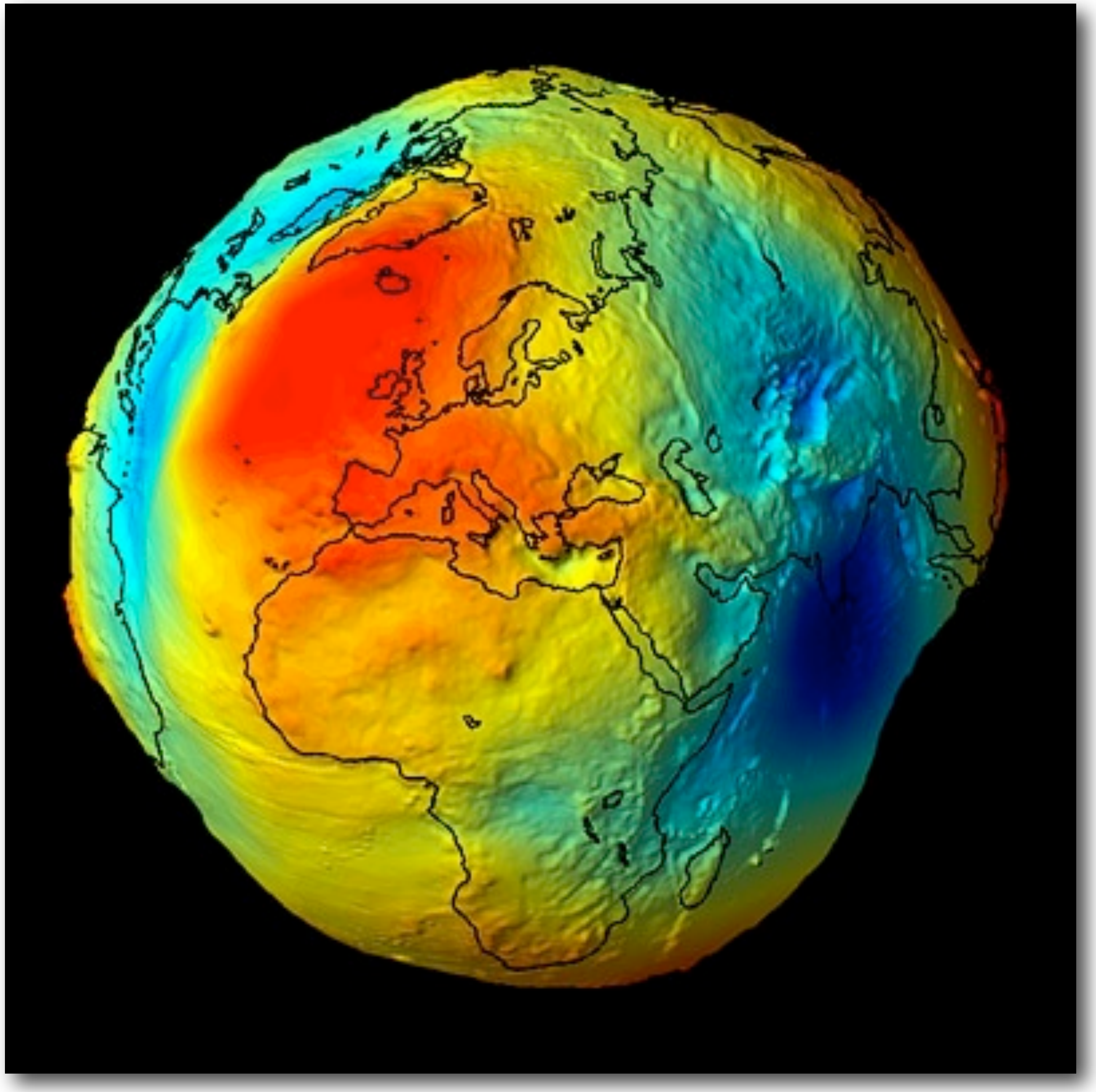
For terrestrial planets, non-zero odd moments and non-axisymmetric figures have been measured!



http://mathworld.wolfram.com/LegendrePolynomial.html

The Earth’s gravity field

The precise shape of the Earth’s gravity field is currently being measured by ESA’s GOCE (Gravity field and steady-state Ocean Circulation Explorer), that was launched March 17th, 2009:



The Earth’s gravity field (left), GOCE (above) (Credit: ESA)

Gravitational moments

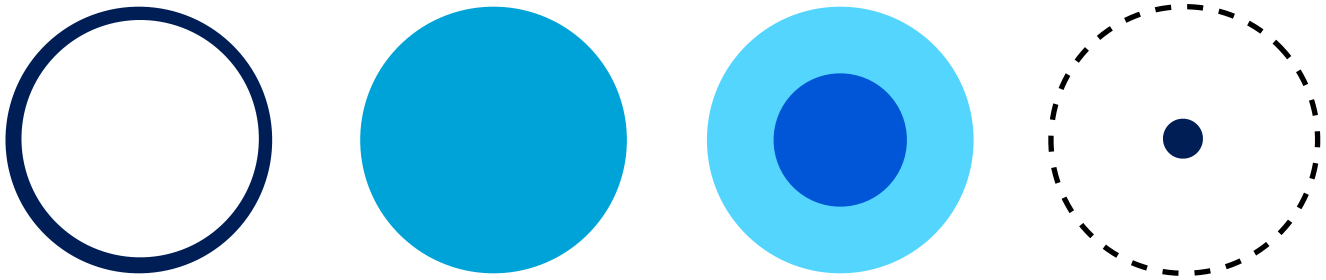
The shape of a planet depends on its internal structure, its plasticity, and its rotation rate. A planet's gravitational moments hold information on its structure:

$$J_2 = \frac{1}{2}\, q_r \equiv \frac{\omega^2_{rot}\, R^3}{GM}$$

where ω^2_{rot} is the planet's spin angular velocity

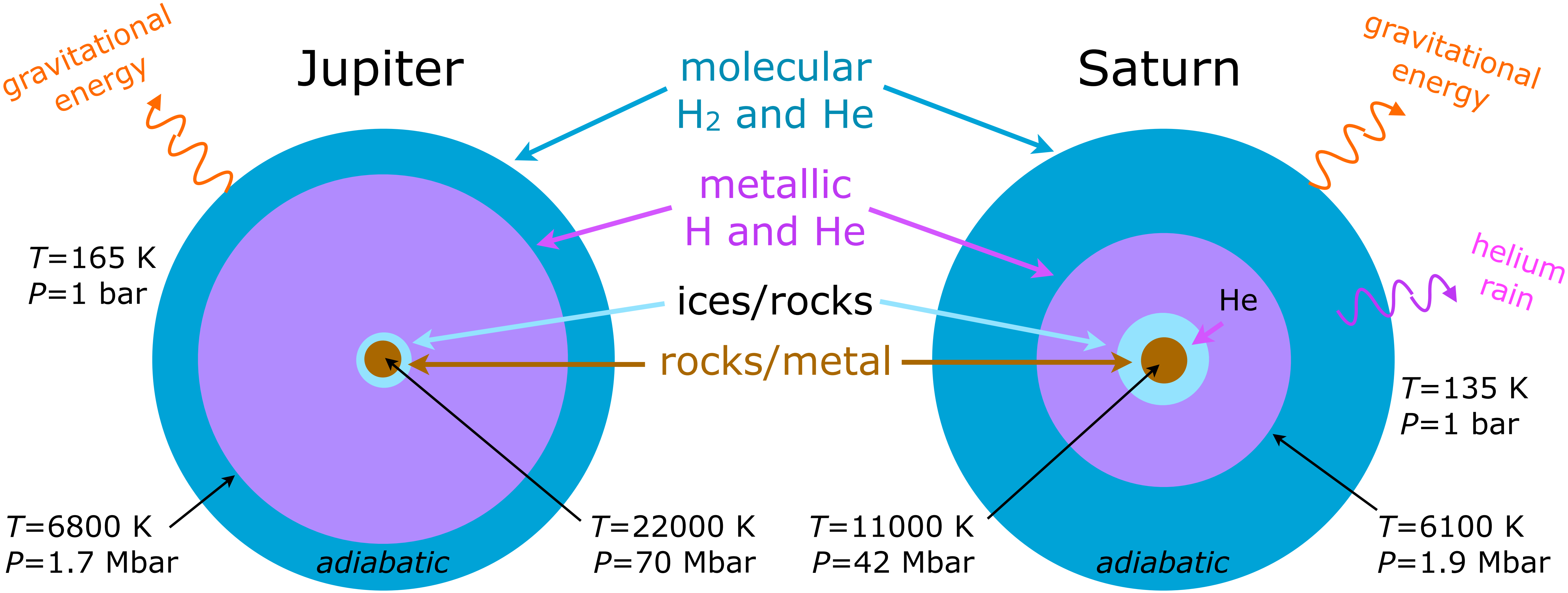
For a rapidly rotating planet in hydrostatic equilibrium, its moment of inertia I can be approximated by:

$$\frac{I}{M\,R^2} \approx \frac{\frac{3}{2} J_2}{J_2 + \frac{1}{3} q_r}$$



$$\frac{I}{M\,R^2} = \quad 0.6667 \qquad 0.4 \qquad < 0.4 \qquad 0.0$$

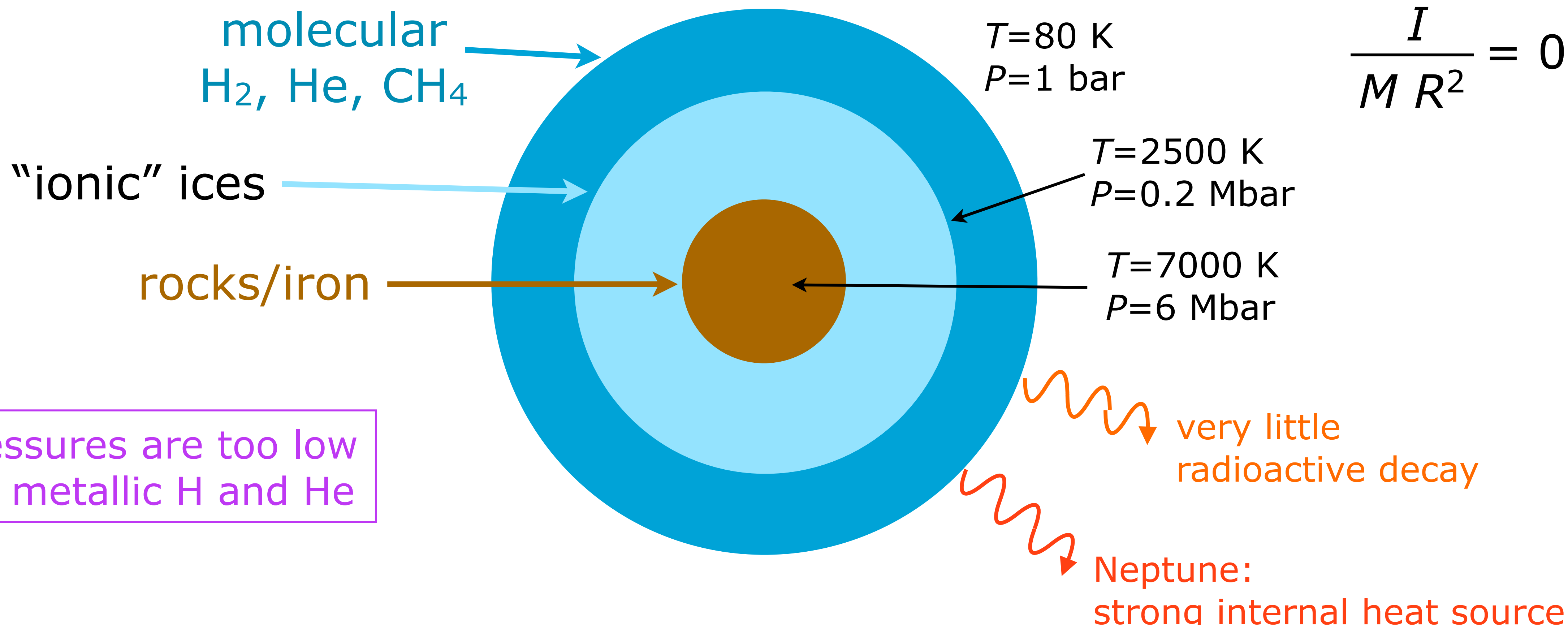
The internal structures of Jupiter & Saturn



- Both planets have ~15-30 $M_\oplus$ high-Z (>2) material in their cores and surrounding envelopes (I/MR^2= 0.25 for Jupiter and 0.21 for Saturn)
- The cores probably consist of iron and rocks from the accretion of planetesimals and gravitational settling later on.
- The core mantle probably contains a relatively large amount of ices of H_2O, NH_3, CH_4, and S-bearing materials.

The internal structures of Uranus & Neptune

Uranus and Neptune models are less well constrained than those of Jupiter and Saturn because there are more solutions possible!



$$\frac{I}{M R^2} = 0.23$$

Pressures are too low for metallic H and He

- Both planets have ~1 $M_\oplus$ high-Z (>2) core.
- Both planets have the same absolute amount of high-Z material as Jupiter & Saturn: they are significantly enhanced.
- Neptune is 3% smaller than Uranus, with a 15% larger mass, hence a 24% larger density.

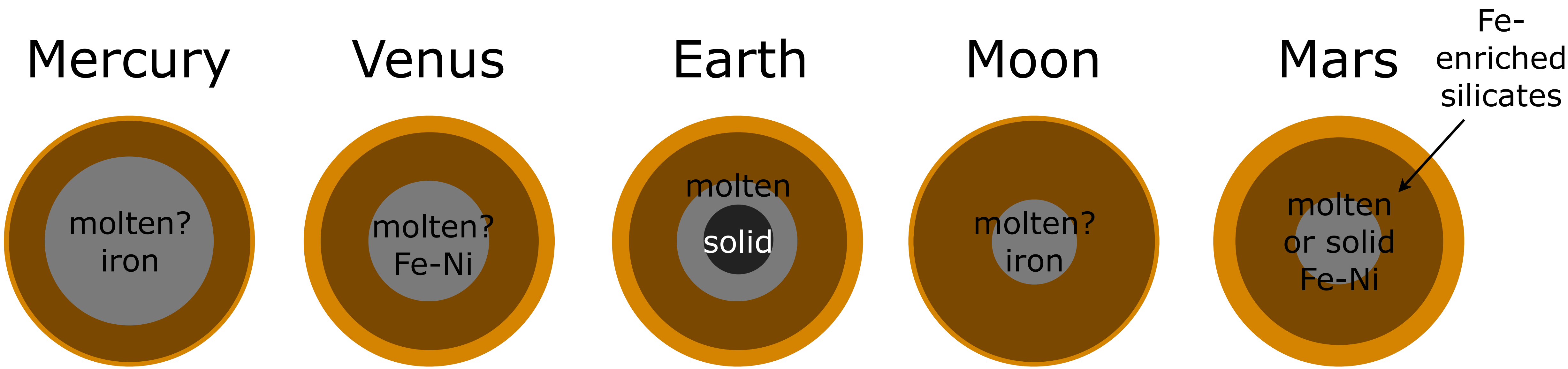
The internal structure of the Galilean moons

The Galilean moons orbit close to Jupiter, and are subject to strong tidal forces. All moons have magnetic fields, intrinsic and/or induced by Jupiter.

The Galilean moons as depicted in NASA/ESA's Europa Jupiter System Mission (EJSM) proposal.

- Io: 3.53 g/cm^3, I/MR^2=0.37; its mass is centrally concentrated: iron core + liquid rocky mantle?
- Europa: 3.02 g/cm^3, I/MR^2=0.35; centrally concentrated mass: liquid (?) iron core + rocky mantle + (liquid and) ~150 km solid H_2O
- Ganymede: 1.94 g/cm^3, I/MR^2=0.31; *heavily* centrally concentrated mass: liquid (?) iron core + rocky mantle + (~150 km liquid and) solid H_2O
- Callisto: 1.85 g/cm^3, I/MR^2=0.35; quite homogeneous composition: icy/rocky mantle+(liquid +) solid H_2O

The internal structure of the terrestrial planets



$\frac{I}{M R^2} =$	0.33	0.33	0.33	0.39	0.37

lithosphere: lighter rocks

mantle: heavier silicates

core: iron

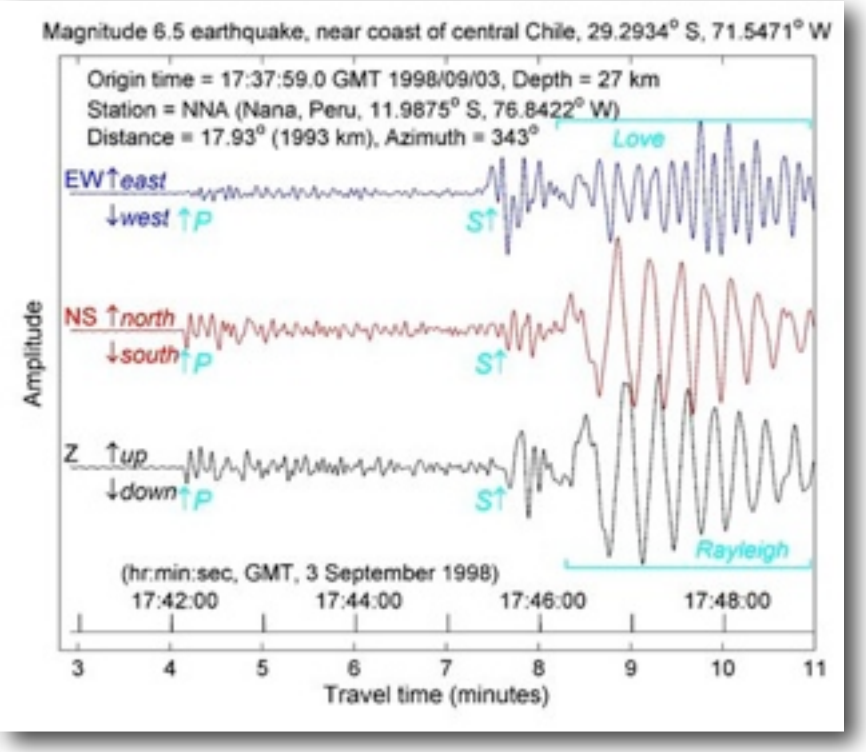
- Mercury might have lost most of its outer, rocky mantle
- Venus has no magnetic field: no convection in its mantle/core?
- Earth has a solid inner core, and a liquid outer core
- The moon has a relatively small iron core, with magnetic field
- Mars has no magnetic field: no liquid core?

Probing the internal structure of the Earth

The internal structure of the Earth is well known thanks to *seismology*, the study of the passage of elastic waves through the planet.

These waves are induced by earthquakes, meteoritic impacts, volcanic or man-made explosions. The waves are detected by seismometers, instruments that measure the motion of the ground on which they are located.

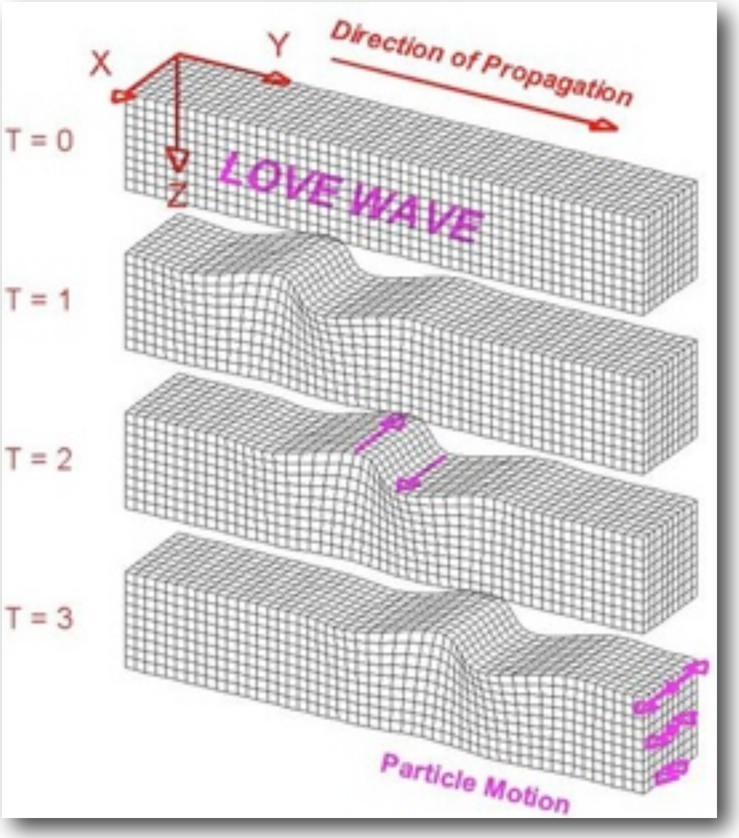
Ancient Chinese seismometer (invented in 132 AD); the exact mechanism has been lost ...

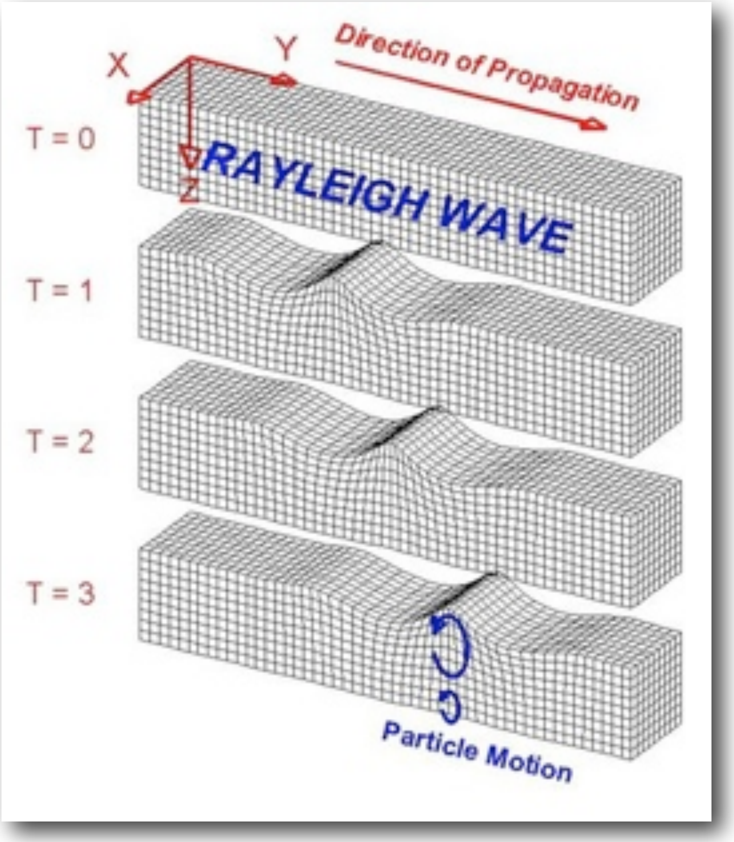


Seismograph of an earthquake in Chile, recorded in Peru. Horizontal and vertical waves of different types are registered. Timing differences between the waves are due to different travel speeds and paths (from www.iris.edu)

On Earth, there are body and surface waves:

Surface waves:





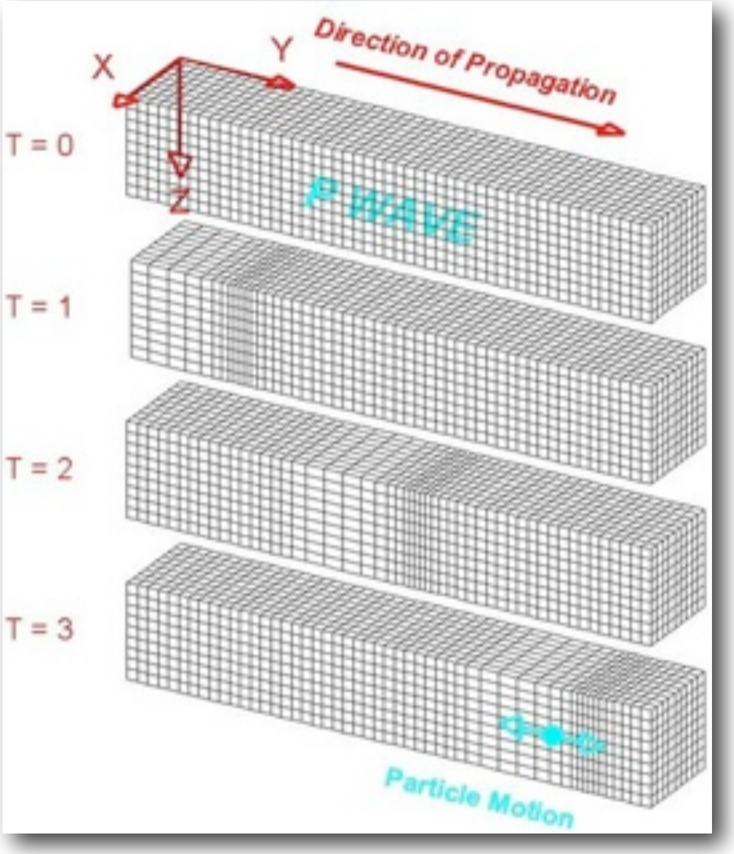
Surface waves are confined to the near-surface layers on Earth. They have a larger amplitude and longer duration than the body waves, and a smaller velocity.

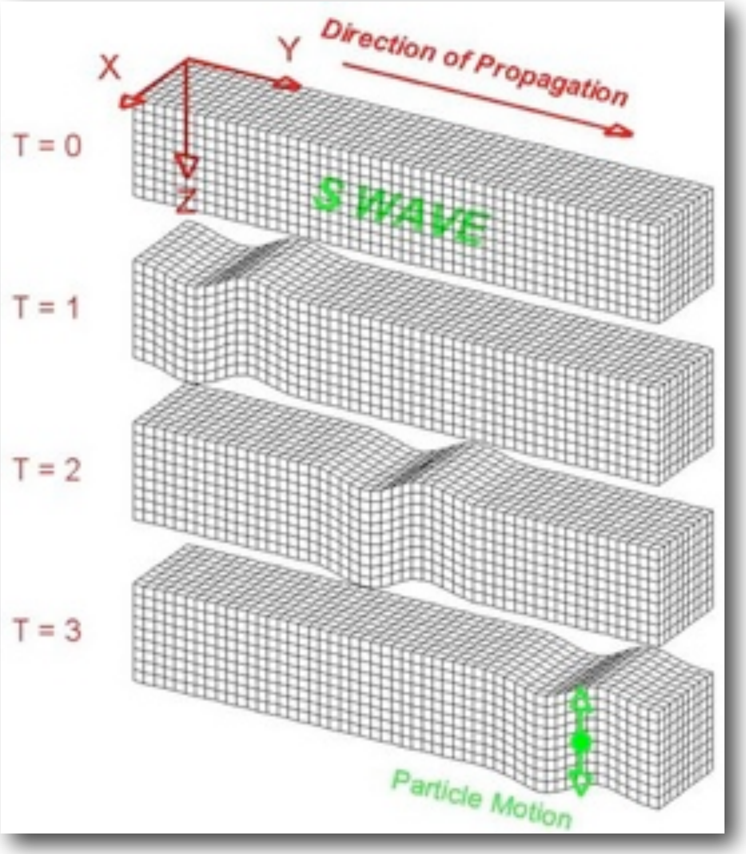
Love waves: transverse motion, entirely horizontal

Rayleigh waves: similar to waves on water; their amplitude decreases with depth.

On Earth, there are body and surface waves:

Body waves:



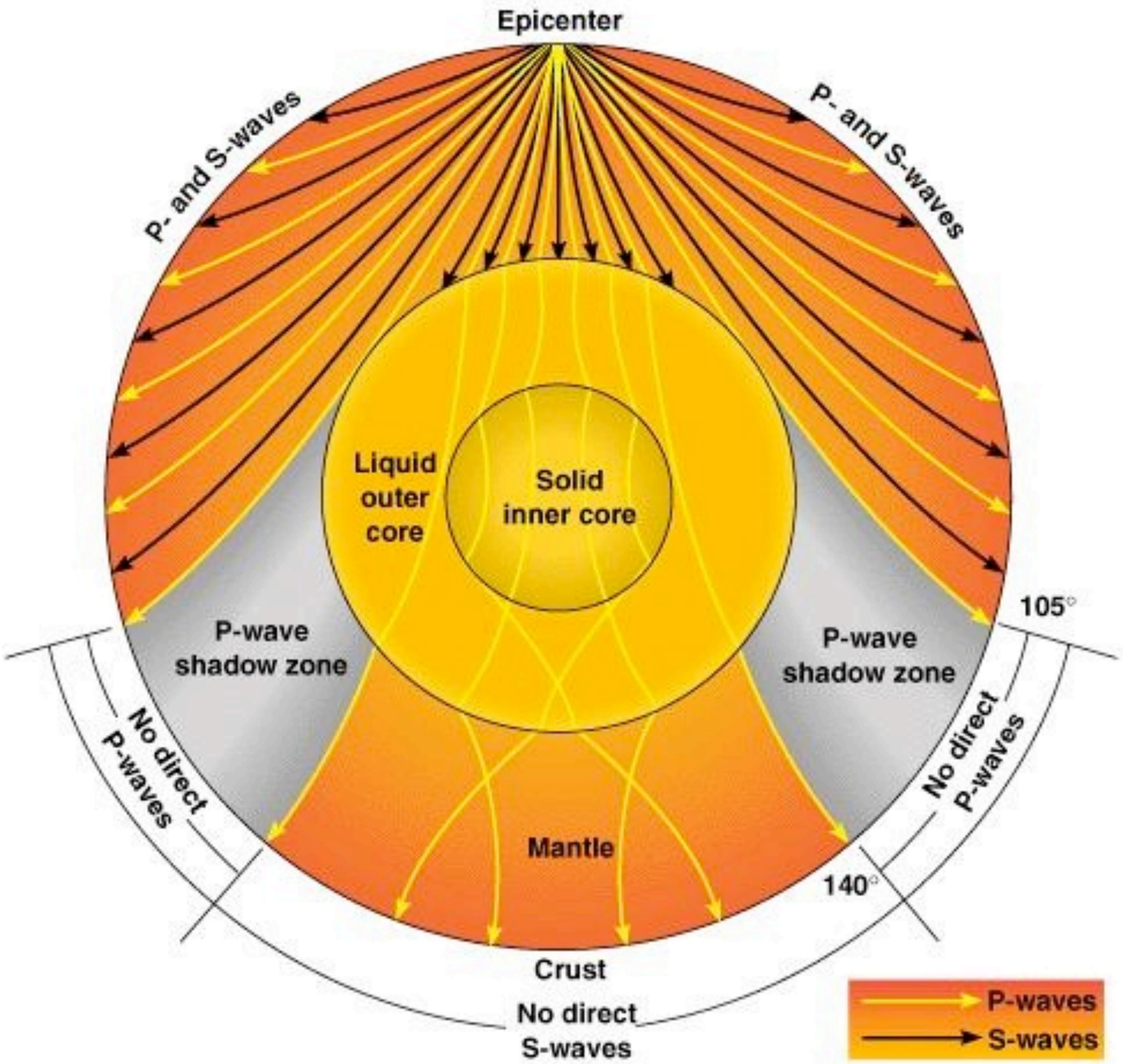


P-waves: Primary, Push, or Pressure waves. The fastest waves, they travel through solids and liquids.

S-waves: Secondary, Shake, or Shear waves. They cannot travel through liquids.

Body waves propagating through the Earth

The velocities of the body waves depend on the density, compressibility, and the rigidity of the material they travel through: velocities increase with density or depth. The waves behave according to Snell's law!



- Waves can also reflect from the crust and/or the core: SS-waves, or PP-waves, ...
- When an S-waves is incident on the outer core, part of it is reflected and part transmitted as P-wave
- A reflected P-wave is called a PCP-wave
- A refracted P-wave is a PKP-wave

From: The Earth through Time, Levin

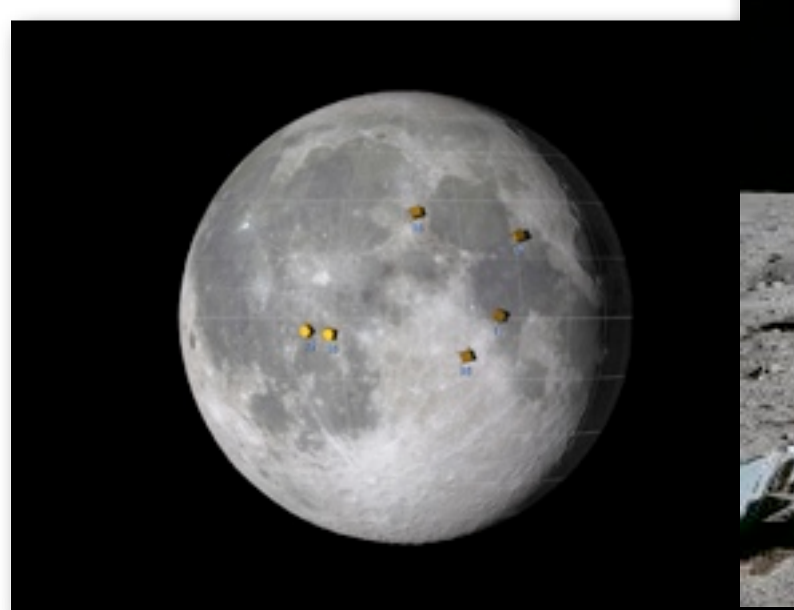
Moon quakes

From 1969 to 1972, astronauts placed seismometers at their landing sites around the moon. Data was radioed back to Earth until 1977.

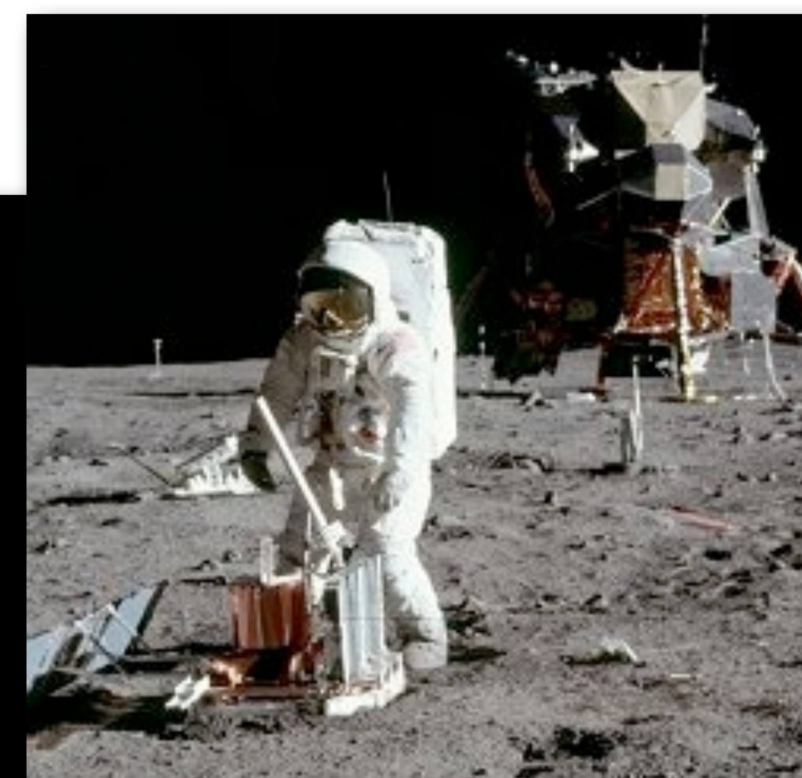
Four types of quakes have been detected:

- Deep quakes (~700 km below the surface) probably caused by tides
- Vibrations of meteorite impacts
- Thermal quakes caused by the expanding crust when heated up
- Shallow quakes (~ 20-30 km below the surface), of unknown origin

→ 28 counted, up to 5.5 on the Richter scale!



The Apollo landing sites



Buzz Aldrin placing a seismometer in the Sea of Tranquility (Credit: NASA)

Mars quakes

NASA's Viking I and II missions (landing in 1976) carried seismometers, only one of them worked. During several years, no quake was detected ...

View from one of the Viking landers (Credit: NASA)

Avalanche seen by the High Resolution Imaging Science Experiment (HiRISE) on NASA's Mars Reconnaissance Orbiter (MRO)

The structure of the Earth: details

The upper mantle consists of:

- the brittle crust, with a thickness of ~6 (oceans) to ~50 km (continents)
- the cool, 'elastic' lithosphere
- the hot, viscous, liquid asthenosphere

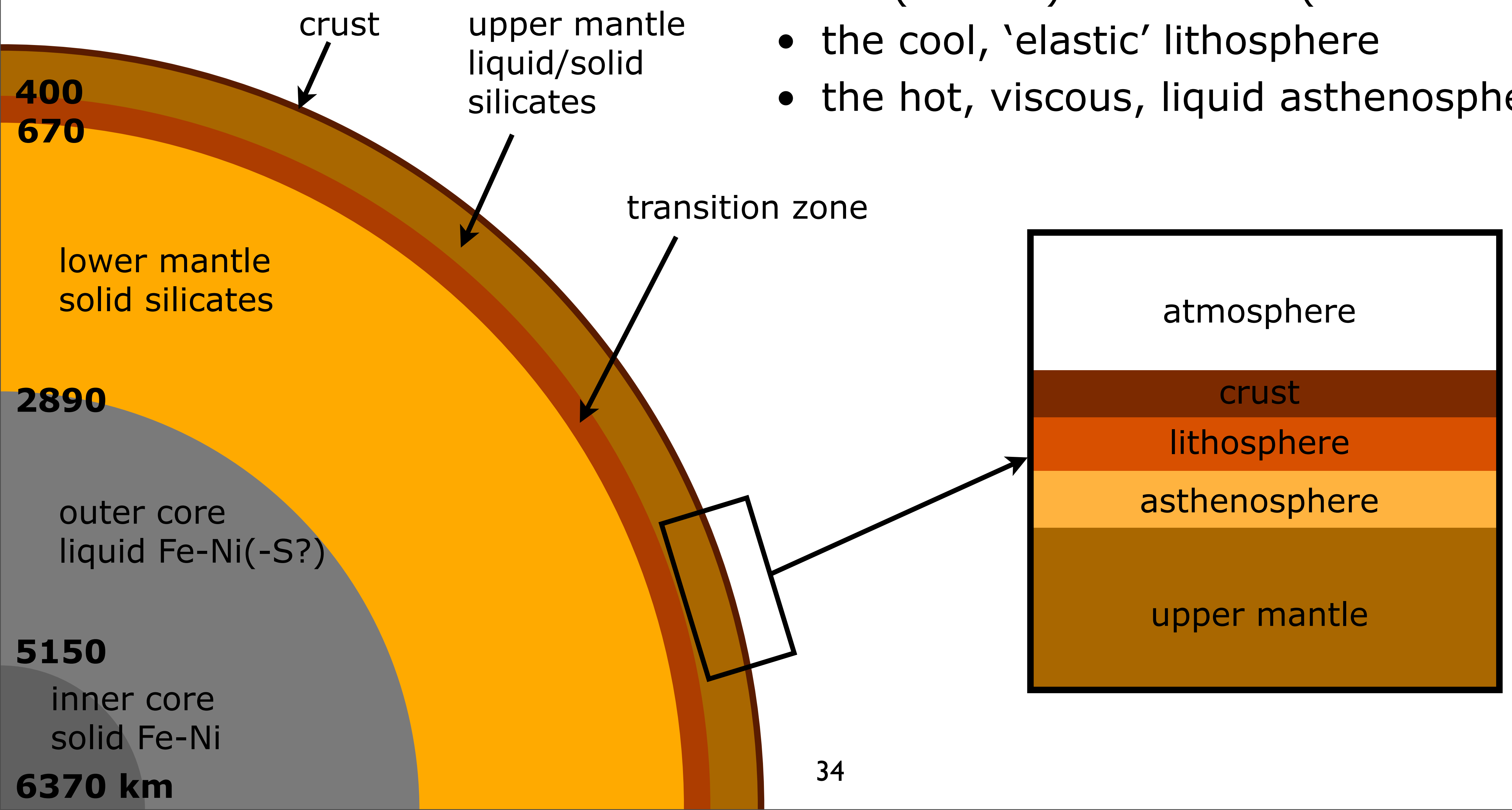
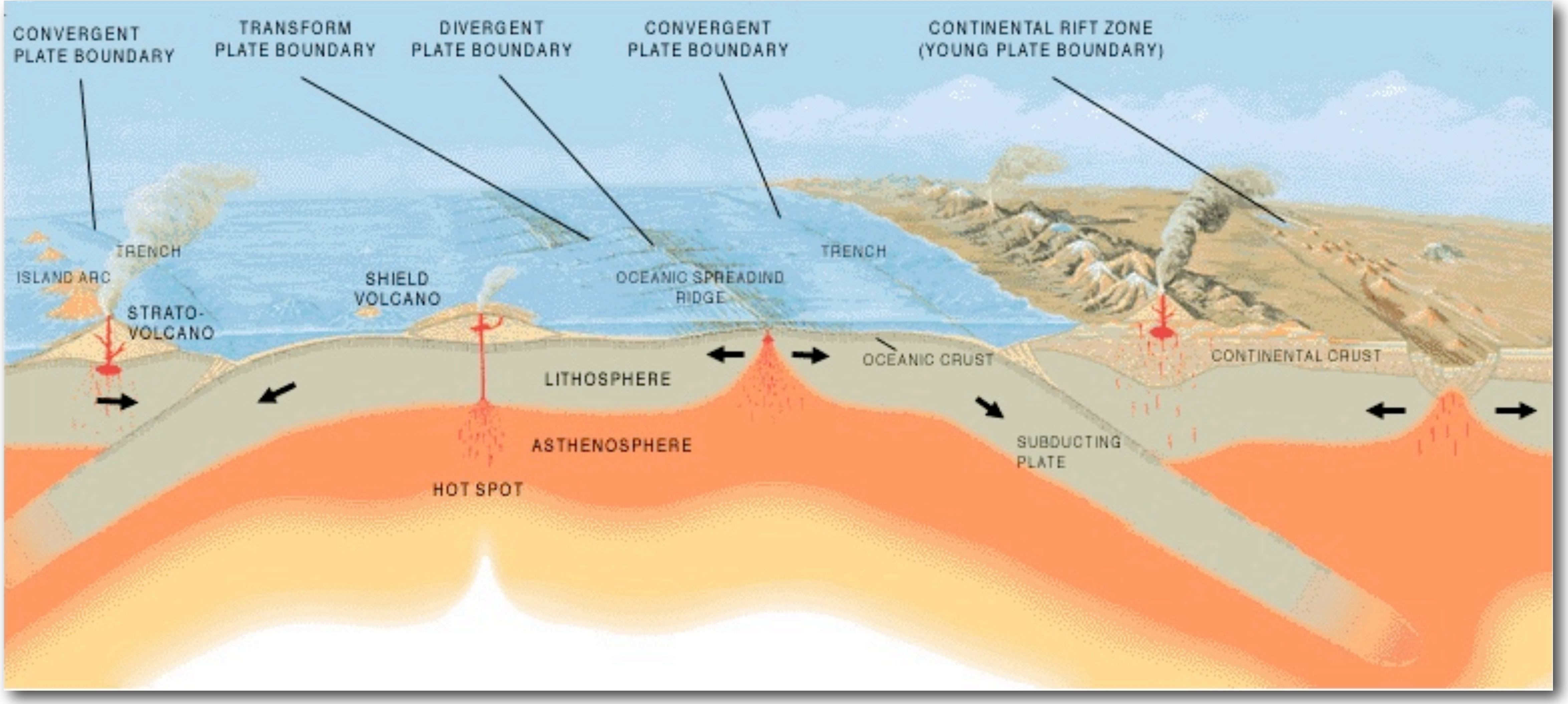


Plate tectonics in a nutshell

Principle of plate tectonics on Earth:

- the lithosphere grows in ridge zones
- oceanic crust is denser than continental crust and subducts upon collision
- where continental plates collide, one subducts and mountains rise up
- recycling of oceanic crust in ~10^8 years



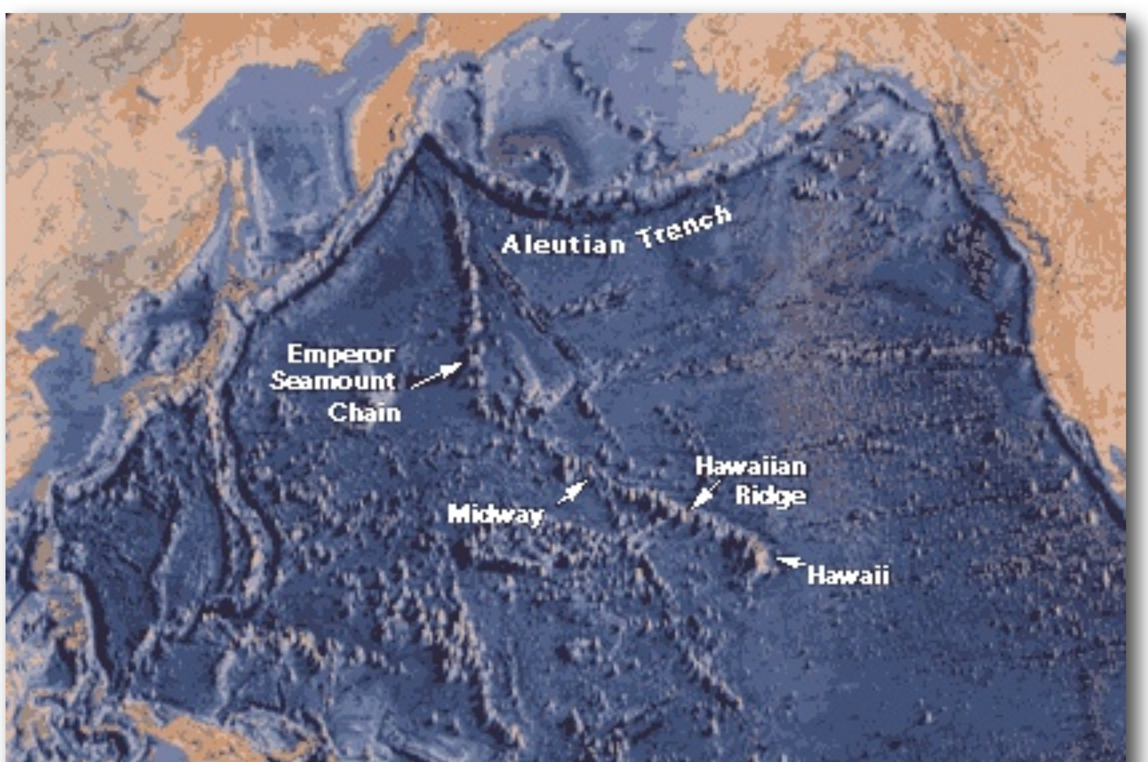
Credit: http://pubs.usgs.gov/gip/dynamic/dynamic.html

Volcanoes on Earth

Many planets and several satellites show signs of past volcanism, while a few bodies, like the Earth (> 1000 active volcanoes) and Io, are still active.

Some types of volcanoes:

- fissure volcanoes (Icelandic)
- shield volcanoes (Hawaiian)
- eruptive volcanoes (Pelean and Plinian)



‘Hot-spot’ volcanic chain (70 Myears old): currently on the eastern coast of Hawaii: Loihi Seamount (altitude: 3 km)

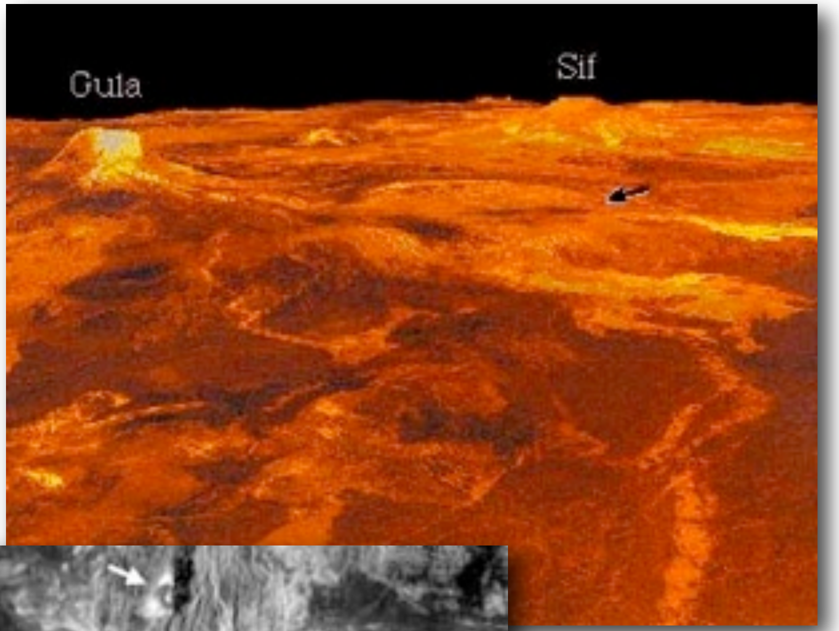
Ultra-Plinian eruption of Mount Pinatubo (1991): it injected ashes and aerosol up into the stratosphere

Volcanoes on Venus

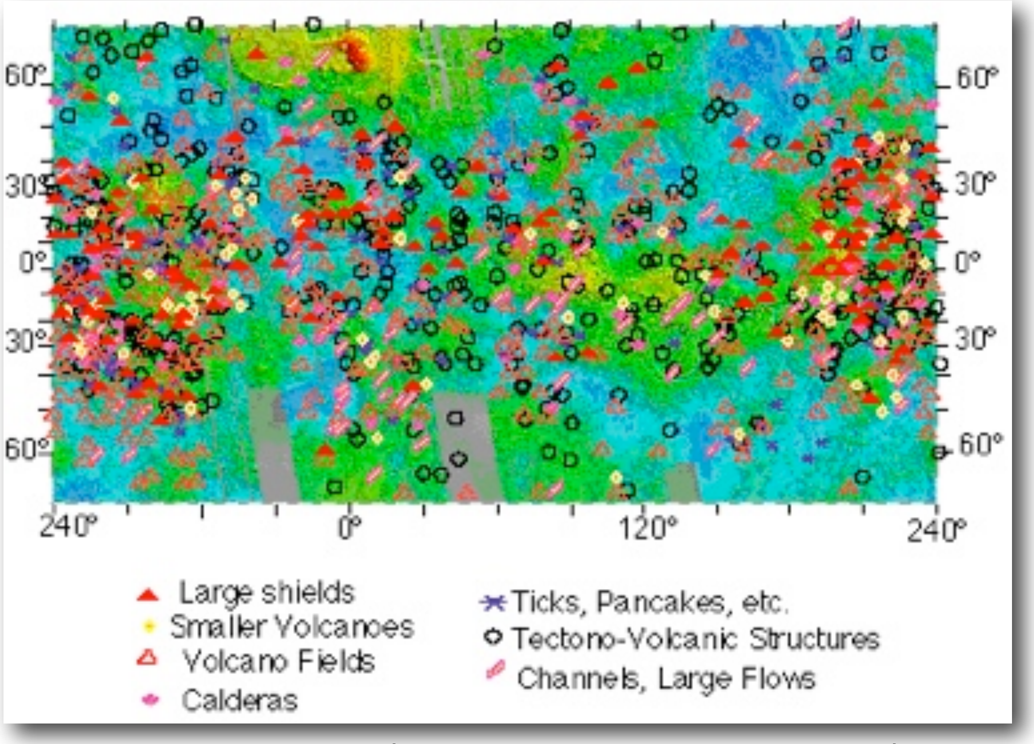
Venus has no plate tectonics; the surface is very young (< few 10^8 years): internal heat builds up and is released through catastrophic resurfacing or episodic plate tectonics?

There are ~1500 volcanoes, however, no evidence for explosive eruptions:

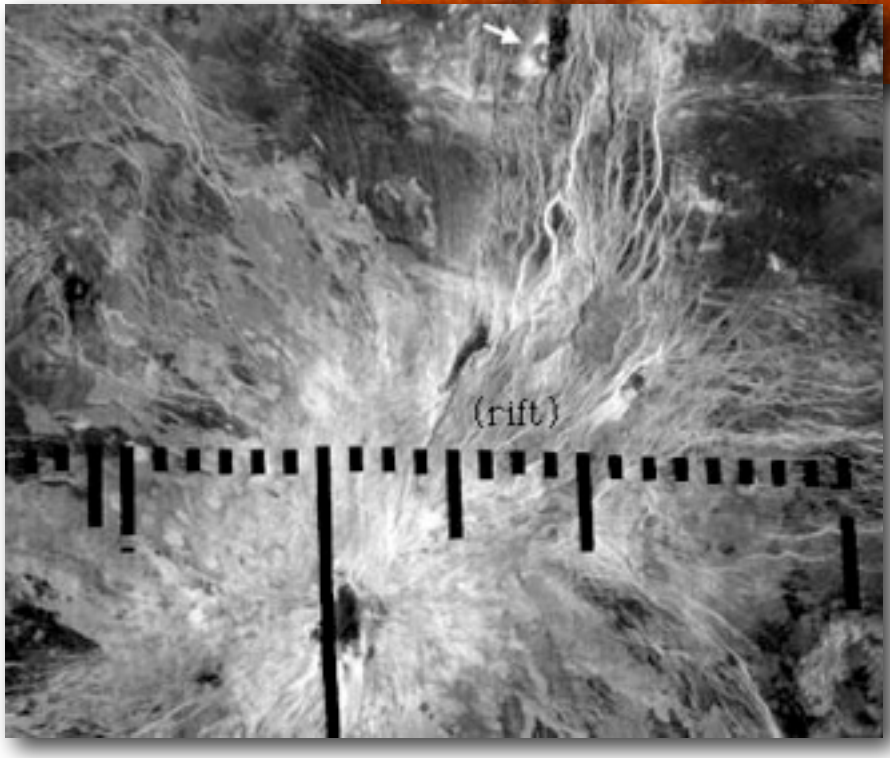
- High atmospheric pressure
- Hardly any water vapour



Magellan data



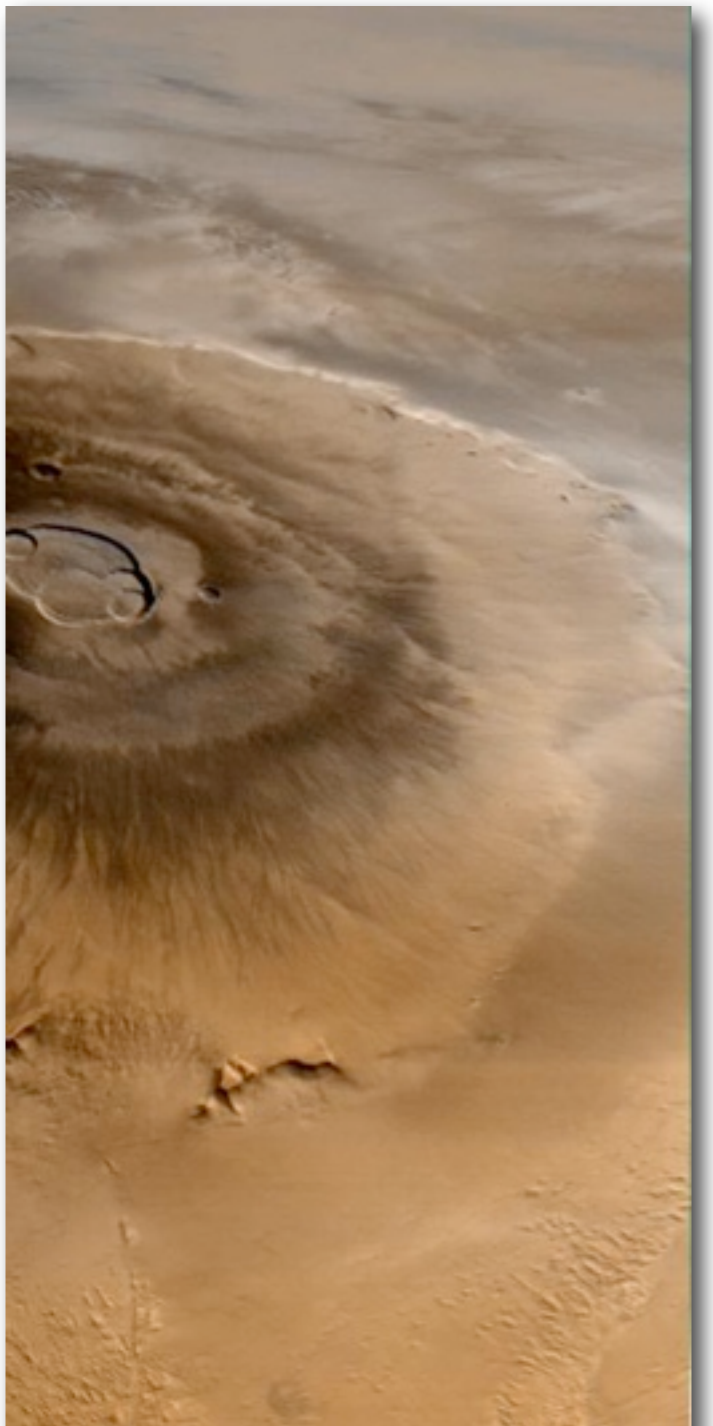
Magellan data (vertical scale exaggerated)



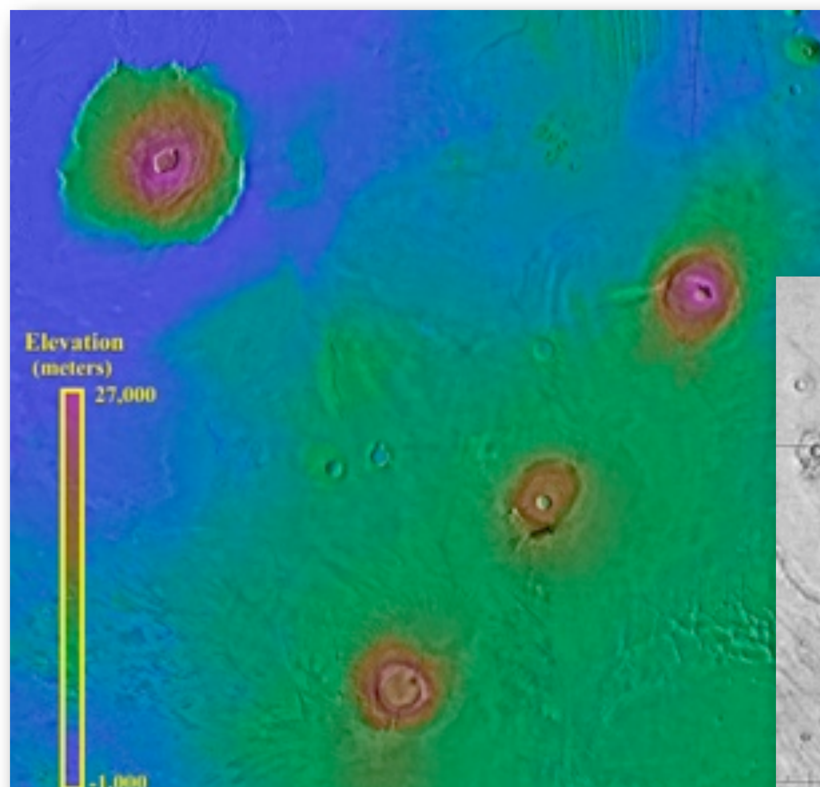
The largest volcano: Theia Mons (4 km high, caldera ~ 50 km wide)

Volcanoes on Mars

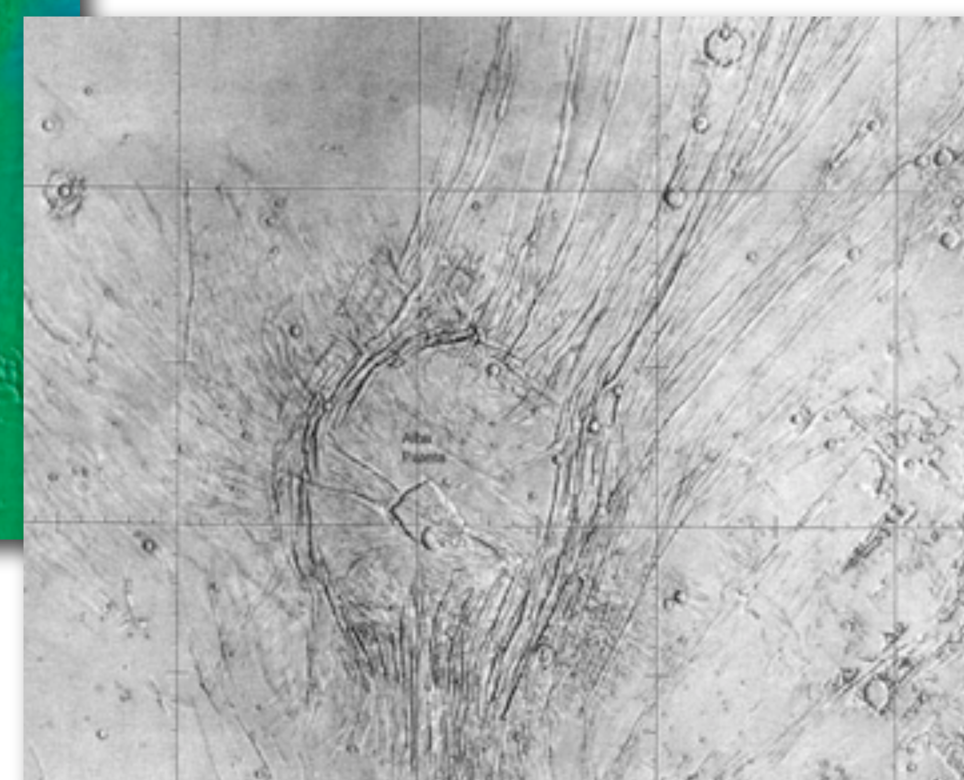
Mars is a one-plate planet, with a thin crust and huge, inactive volcanoes.



Olympus Mons (25 km high, 550 km wide)
Last eruption ~ 200 million years ago (???)



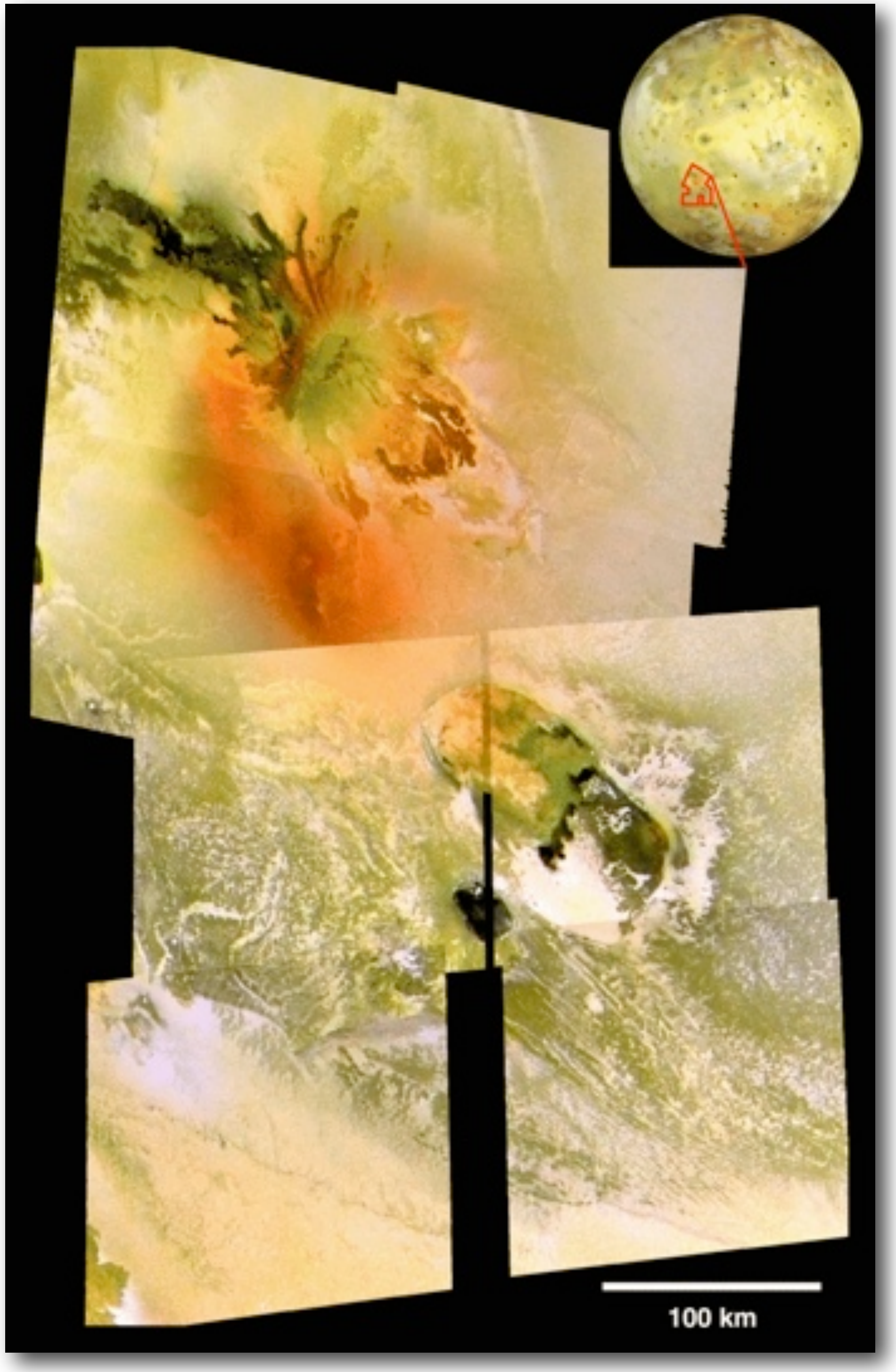
Tharsis region: Arsia Mons (20 km), Pavonis Mons (18 km), Ascraeus Mons (26 km)



Alba Patera (1500 km across)

Volcanoes on Io (Jupiter)

Io is internally heated by Jupiter's tidal forces, resulting in volcanic eruptions, driven by evaporation of SO_2:

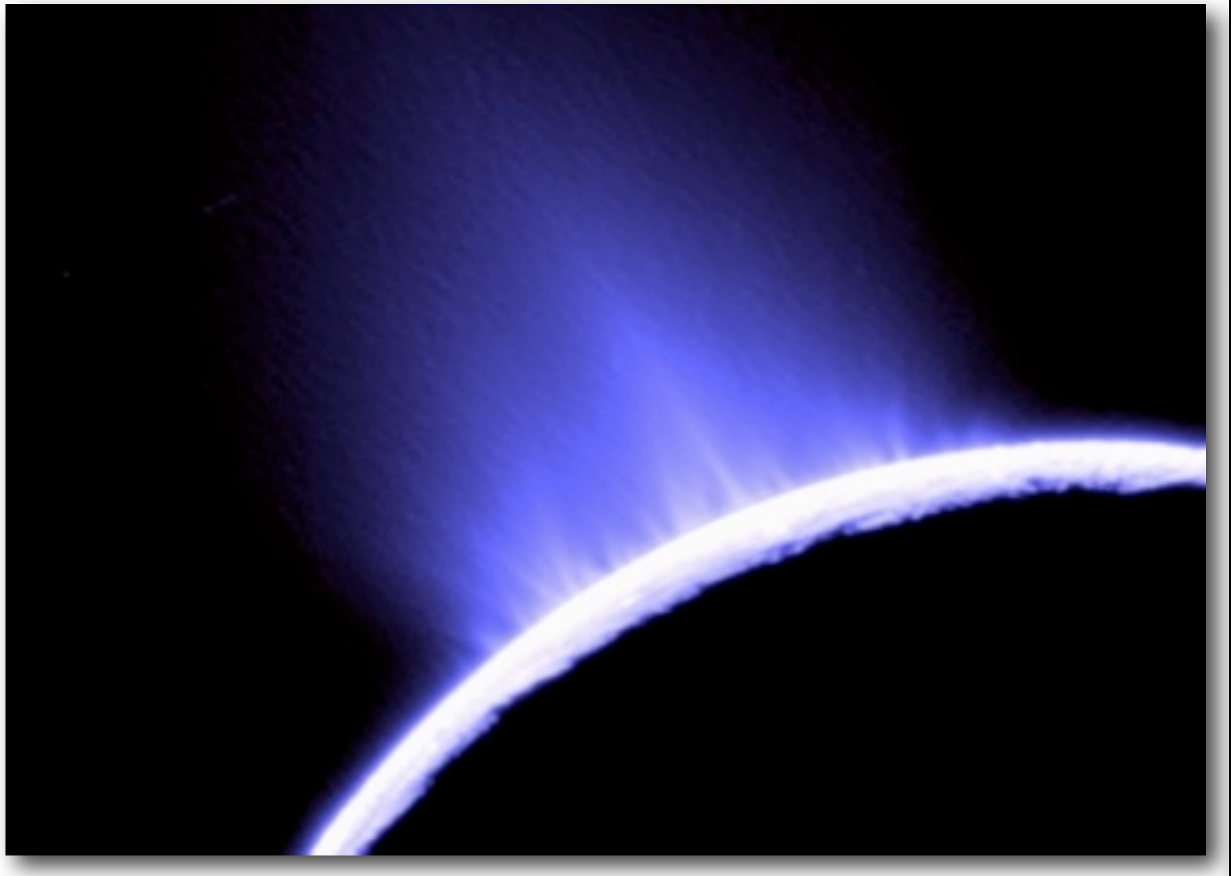


Volcanic regions on Io (Credit: Galileo/NASA)

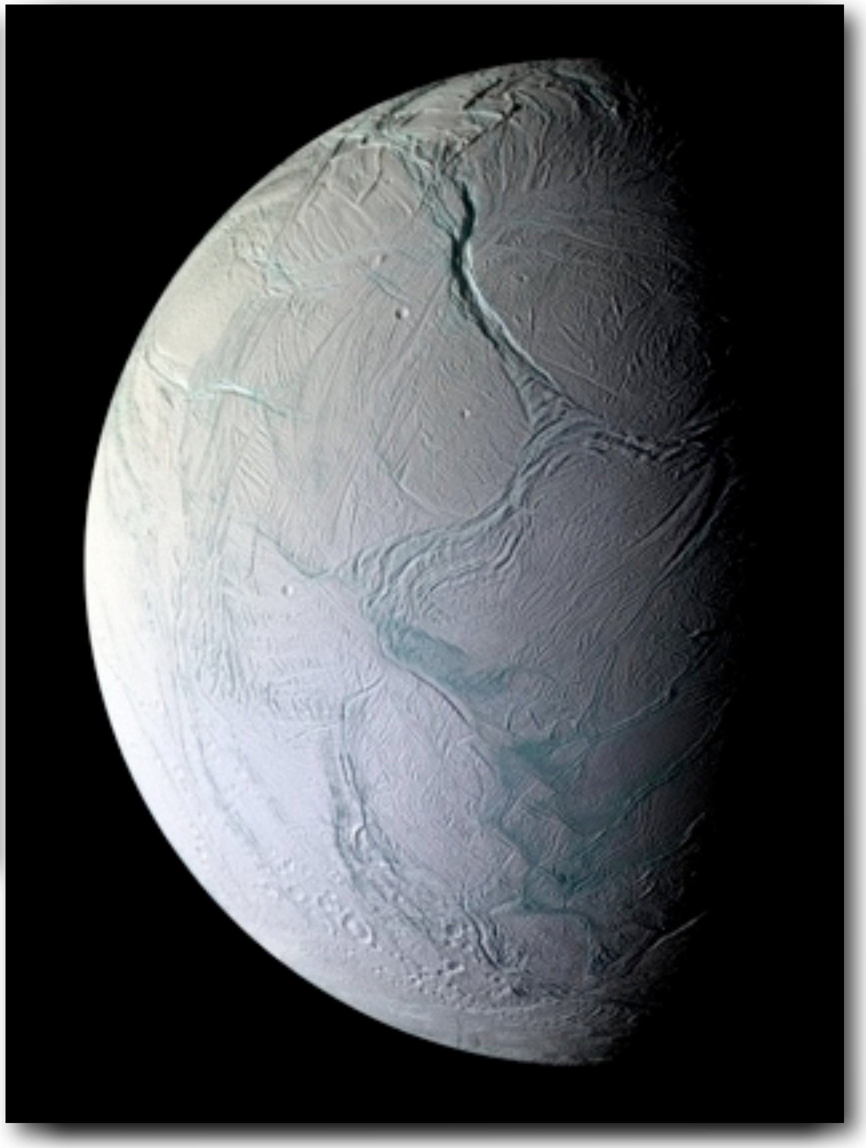
Credit: Voyager/NASA

Cryovolcanoes on Enceladus (Saturn)

Enceladus has an outer water ice layer, with liquid water below. Cryovolcanism occurs through cracks (the "Tiger stripes"?) in the ice layer:



Water ice particles erupt from Enceladus' surface, seen in forward scattered light (Credit:Cassini/NASA)



Credit: Cassini/NASA