Terrestrial Planets

Week 12

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Earth and Planetary Sciences
FD Adams 131C

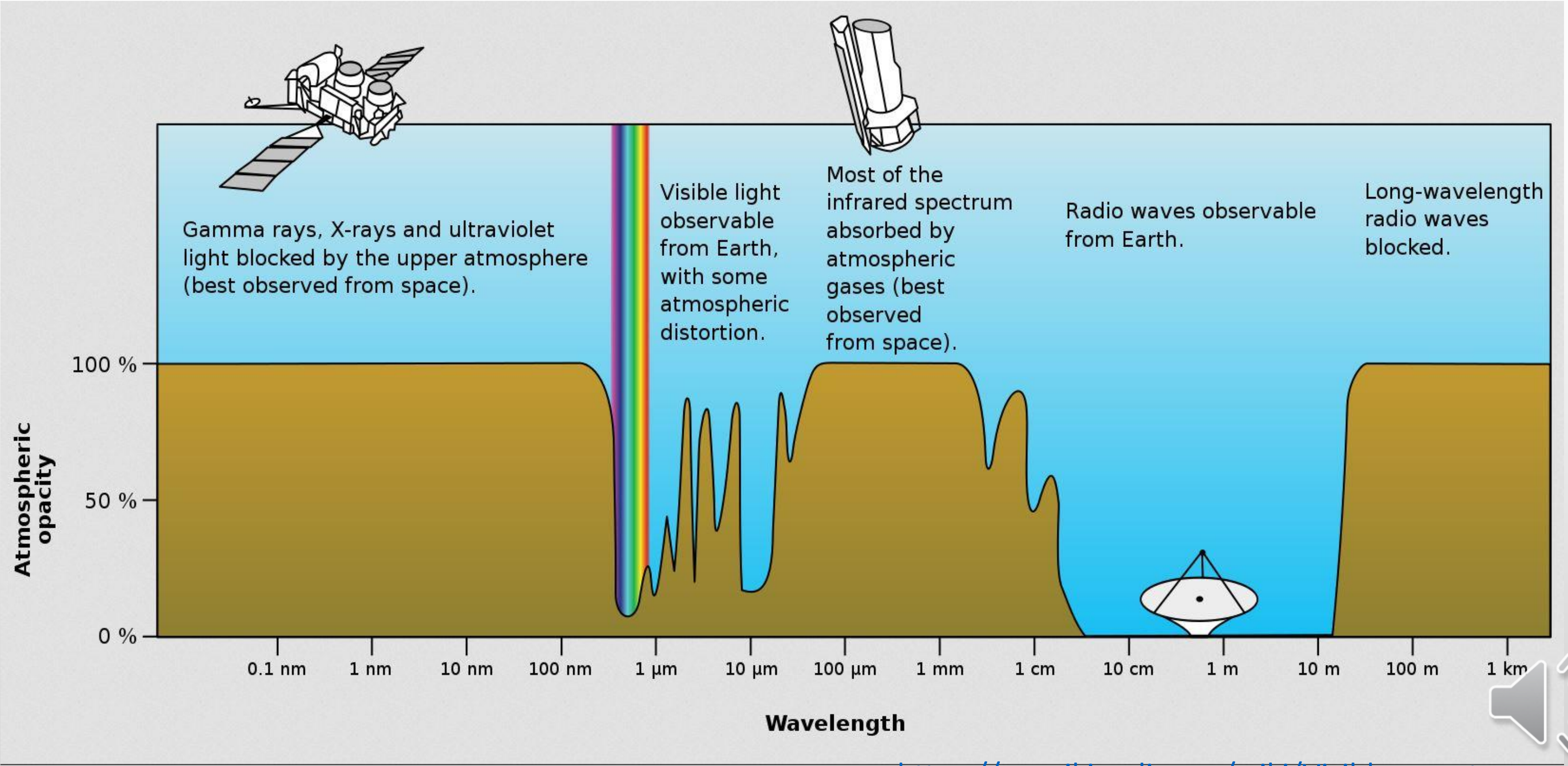
Geophysical Mapping

While direct imaging of planetary surfaces can tell us a great deal about the processes at play on a planet or moon, we have other tools to tell us even more.

- Spectroscopic imaging
- Magnetics
- Gravity variations
- Seismic imaging
- Crater count density

Spectroscopy

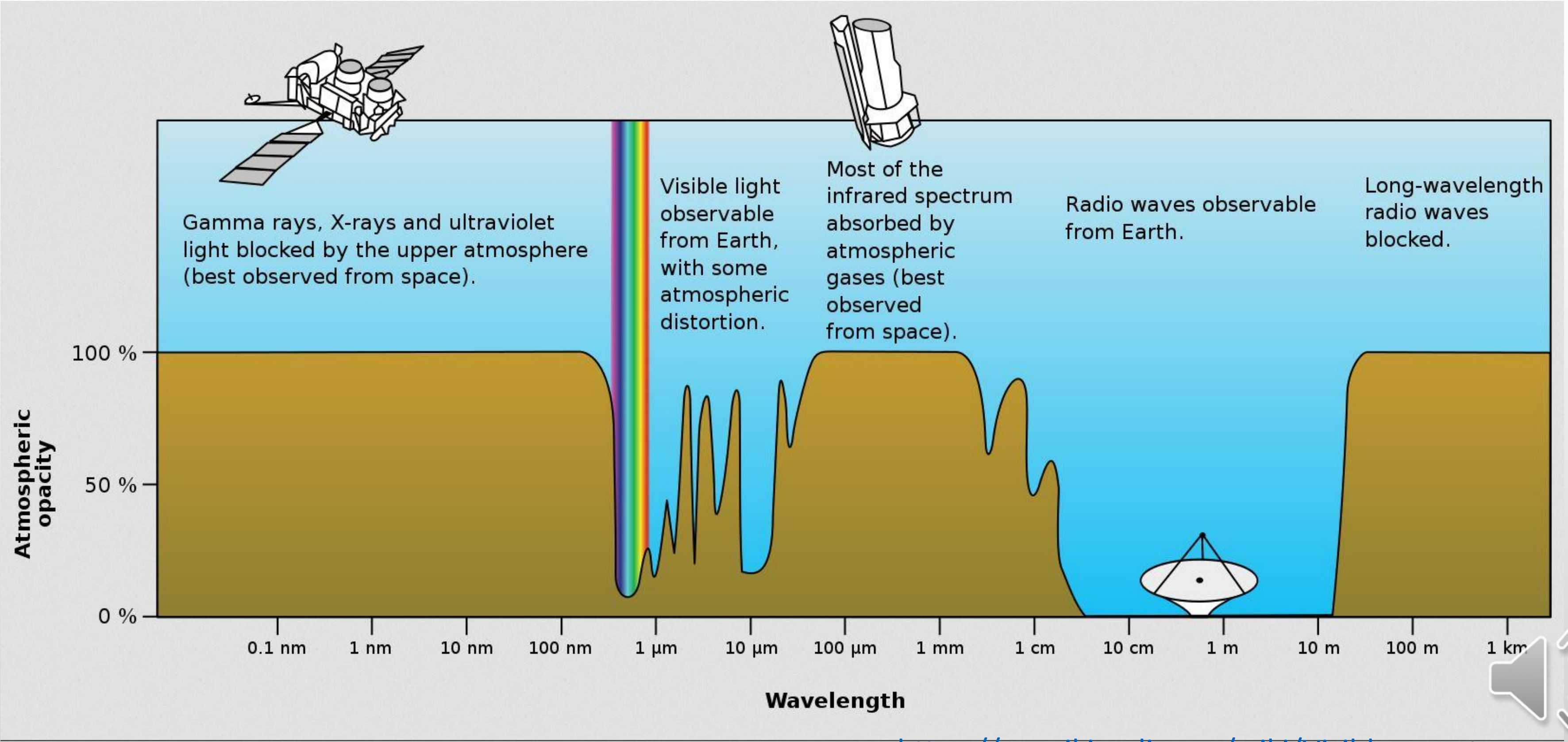
When we look at rocks or minerals, we see clear differences in their colours. Reflected light visible to our eye, spans only a very narrow band of the electromagnetic spectrum. Our atmosphere is opaque to many "colours".



https://en.wikipedia.org/wiki/Visible_spectrum

Spectroscopy -- Atmospheres

The atmospheres of planets are also opaque according to their densities and their compositions. Still, we can determine the gases of their atmospheres according to the wavelengths of absorption of light.



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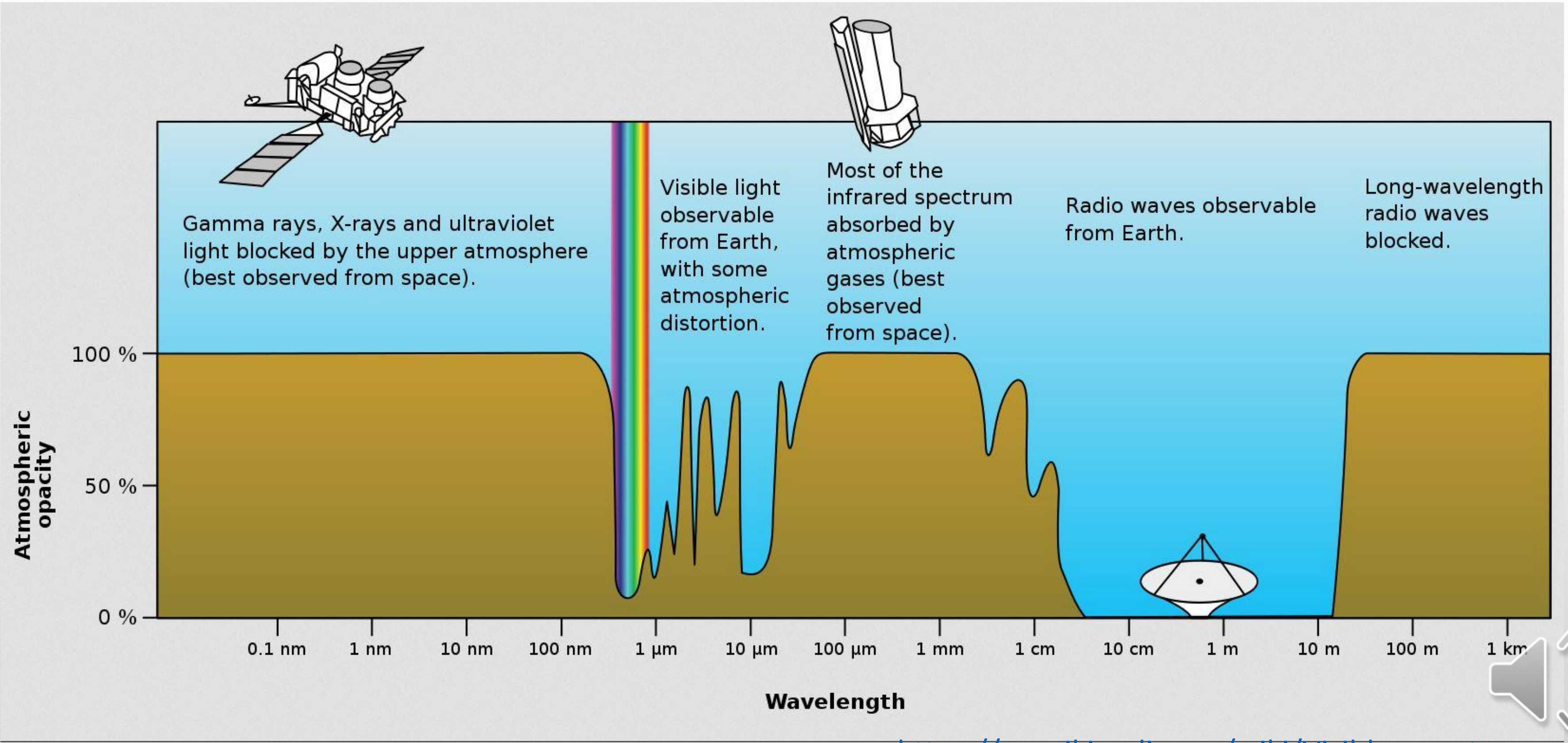
Spectroscopy -- Surfaces

For planets and moons with atmospheres thin enough to be penetrated by visible, infrared and microwave radiation, we can measure the wavelength dependent reflectance of rocks and minerals on the surface.



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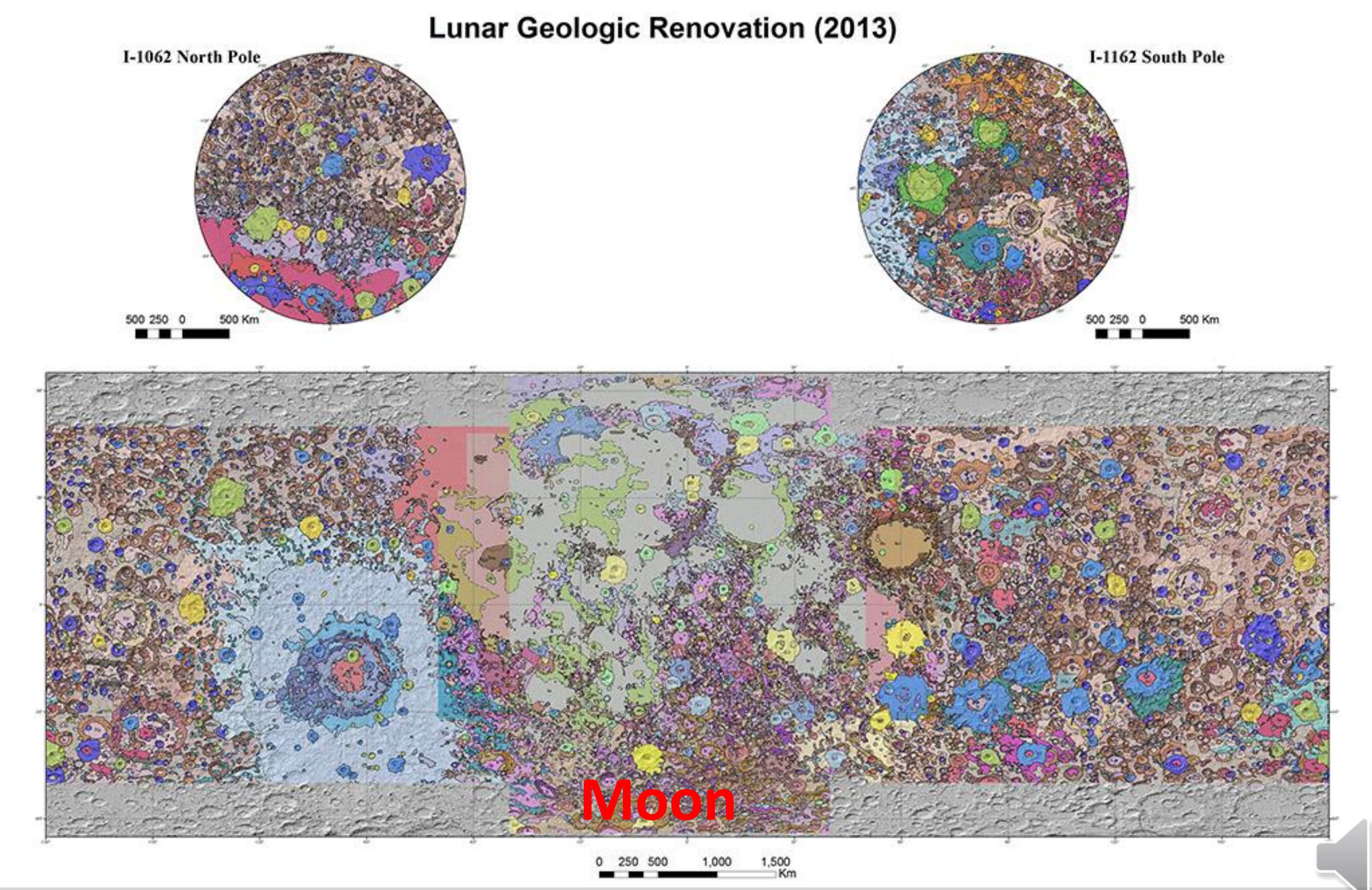
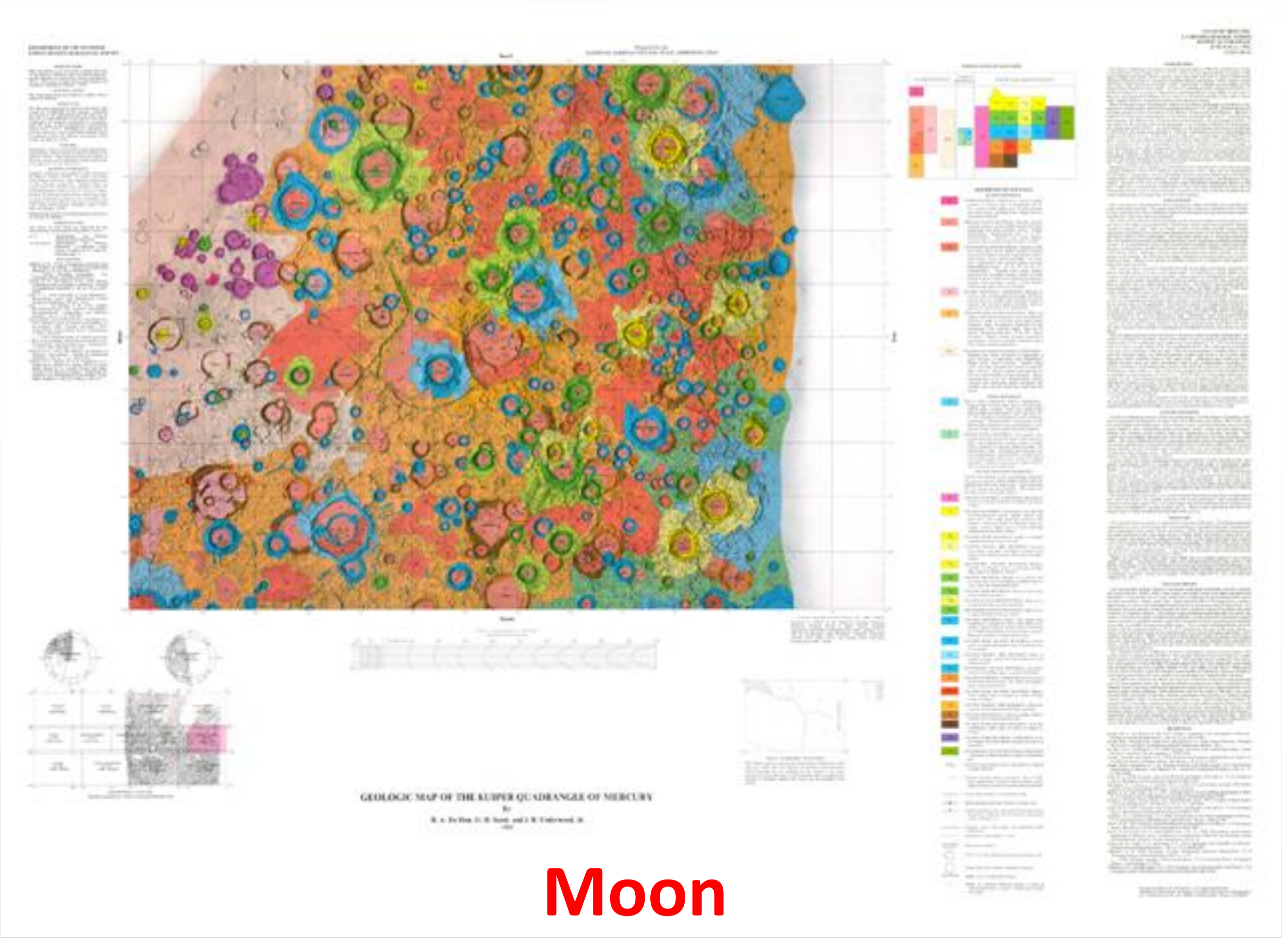
Geology by spectroscopy

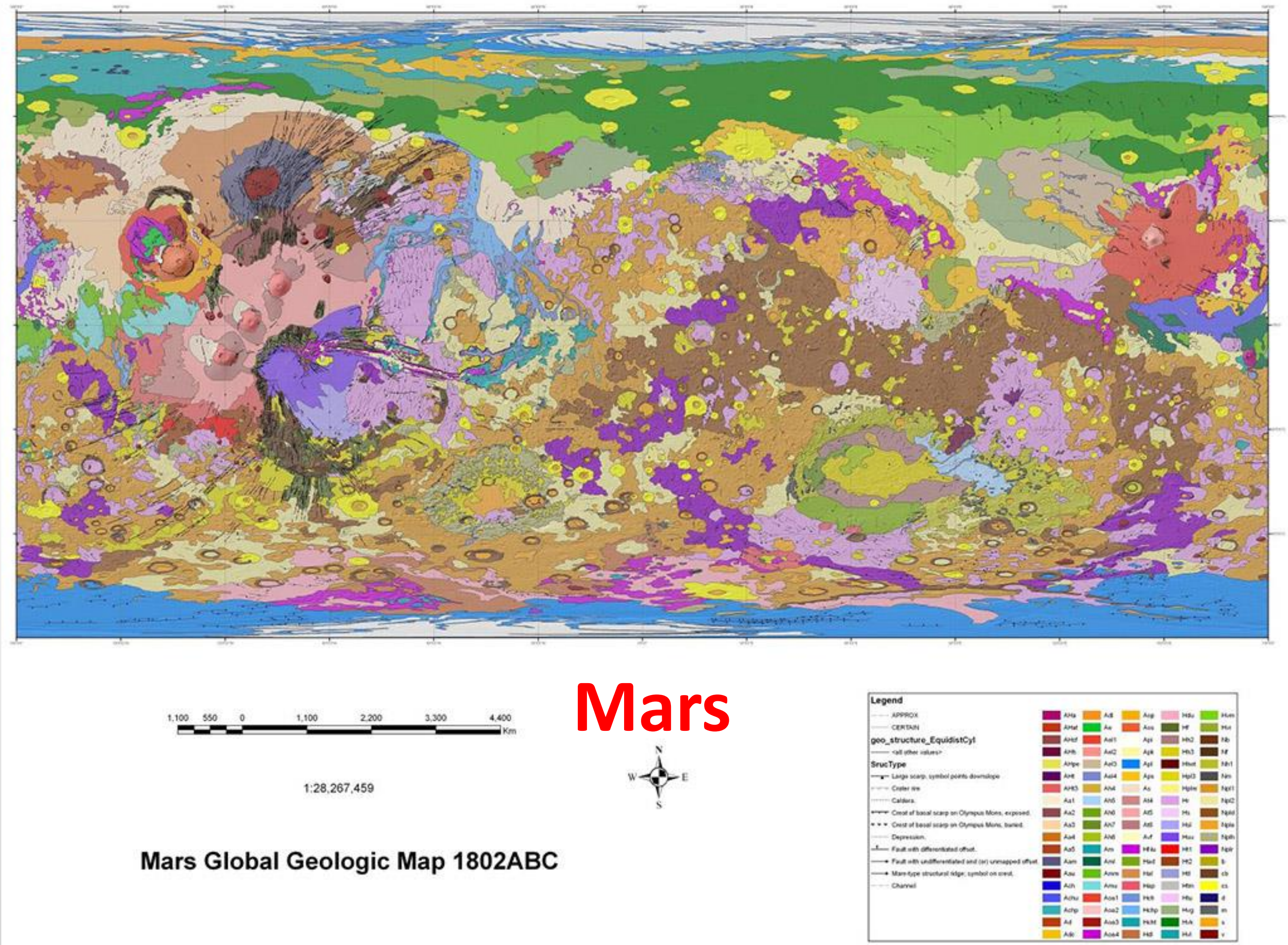


Geology by spectroscopy



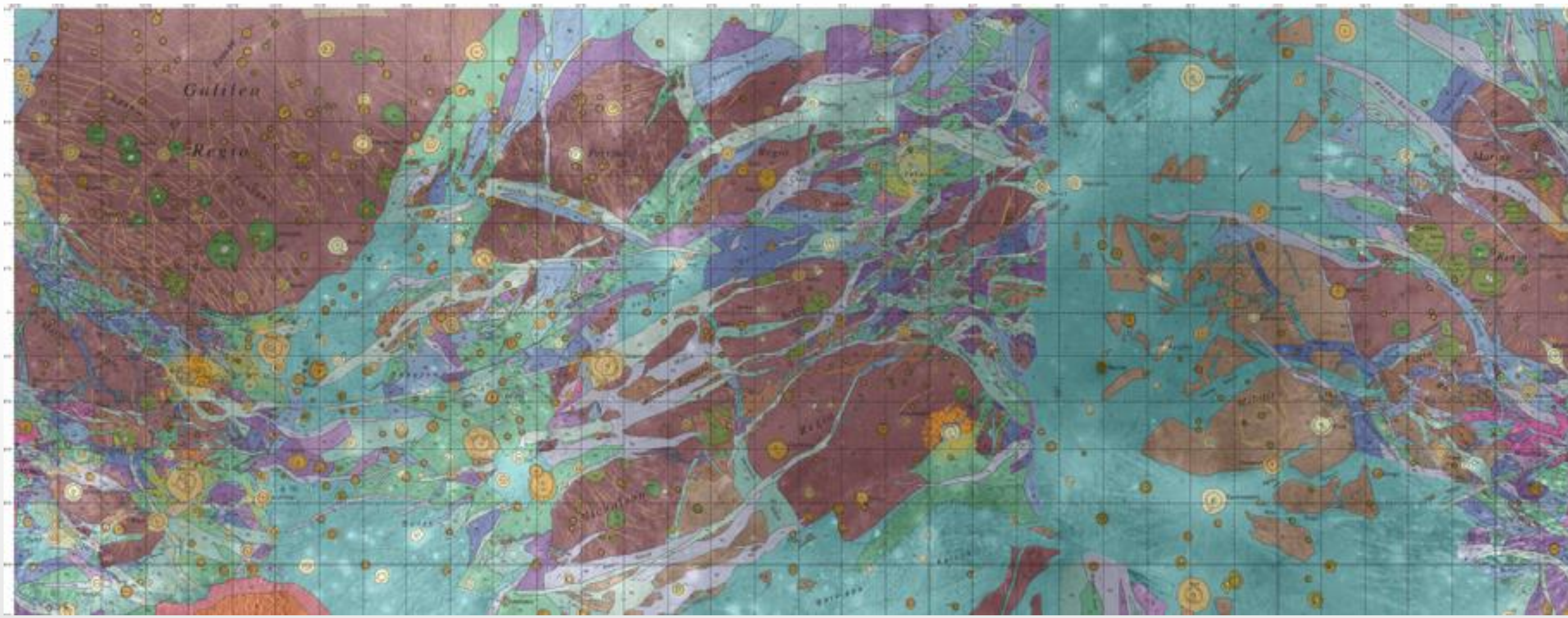
Moon

Geology by spectroscopy



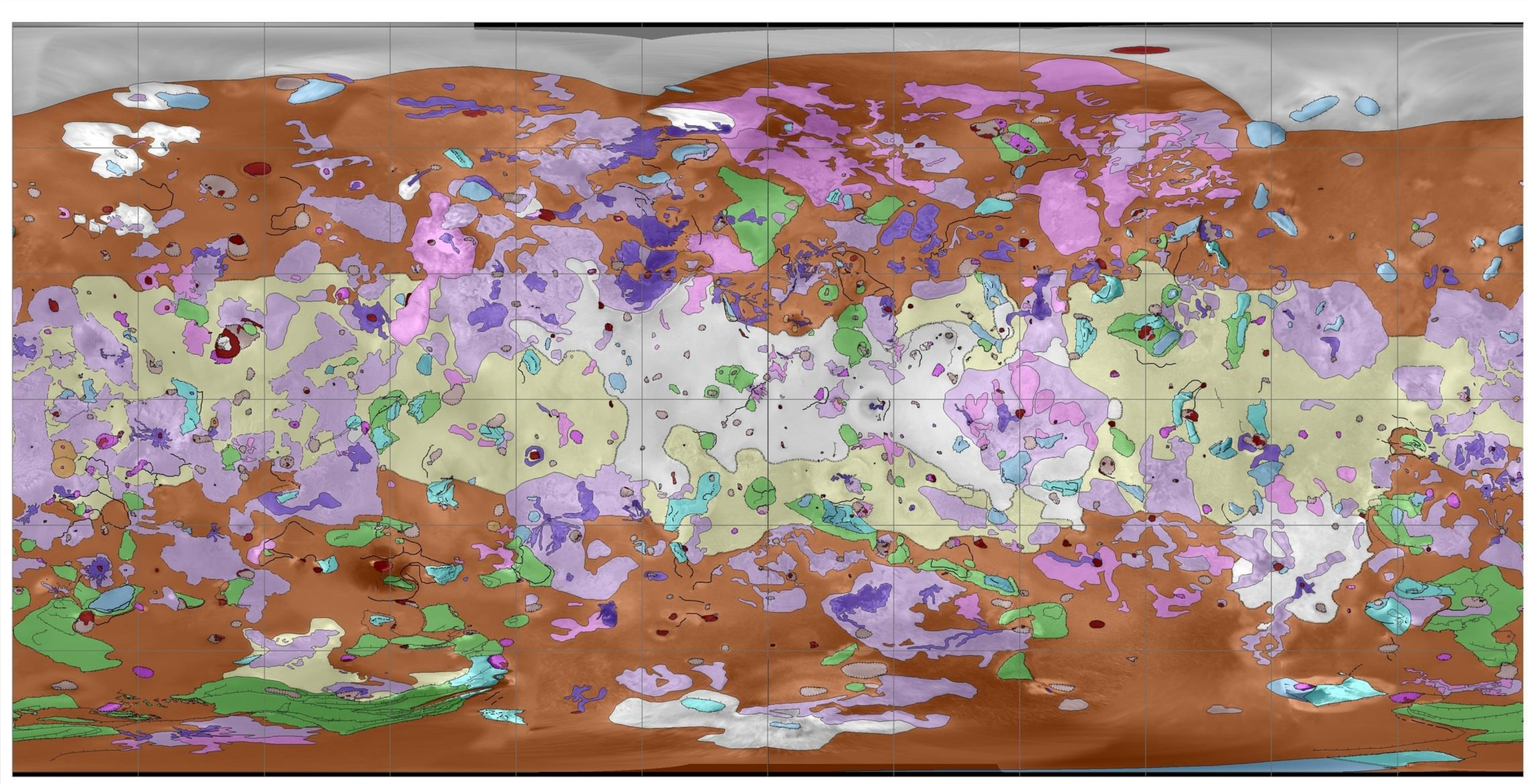
Mars

Geology by spectroscopy



Ganymede

Geology by spectroscopy



Io

Magnetics

All "first pass" probes of other planets and moons have carried magnetometers to measure possible global magnetic fields. By mid-1970s, we knew that the ***Moon, Venus and Mars have no significant global magnetic fields while Mercury, like Earth did***.
Internally generated global magnetic fields are a clear indication that a ***geodynamo*** action in which circulating conductive materials generate a self-sustaining field.

In 1974, Mariner 10's close fly-by of Mercury detected a dipolar field with a dipole moment about 1% that of Earth.

Magnetics

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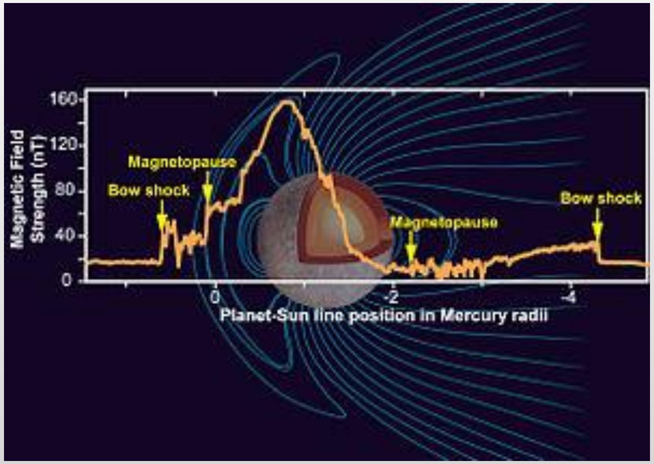
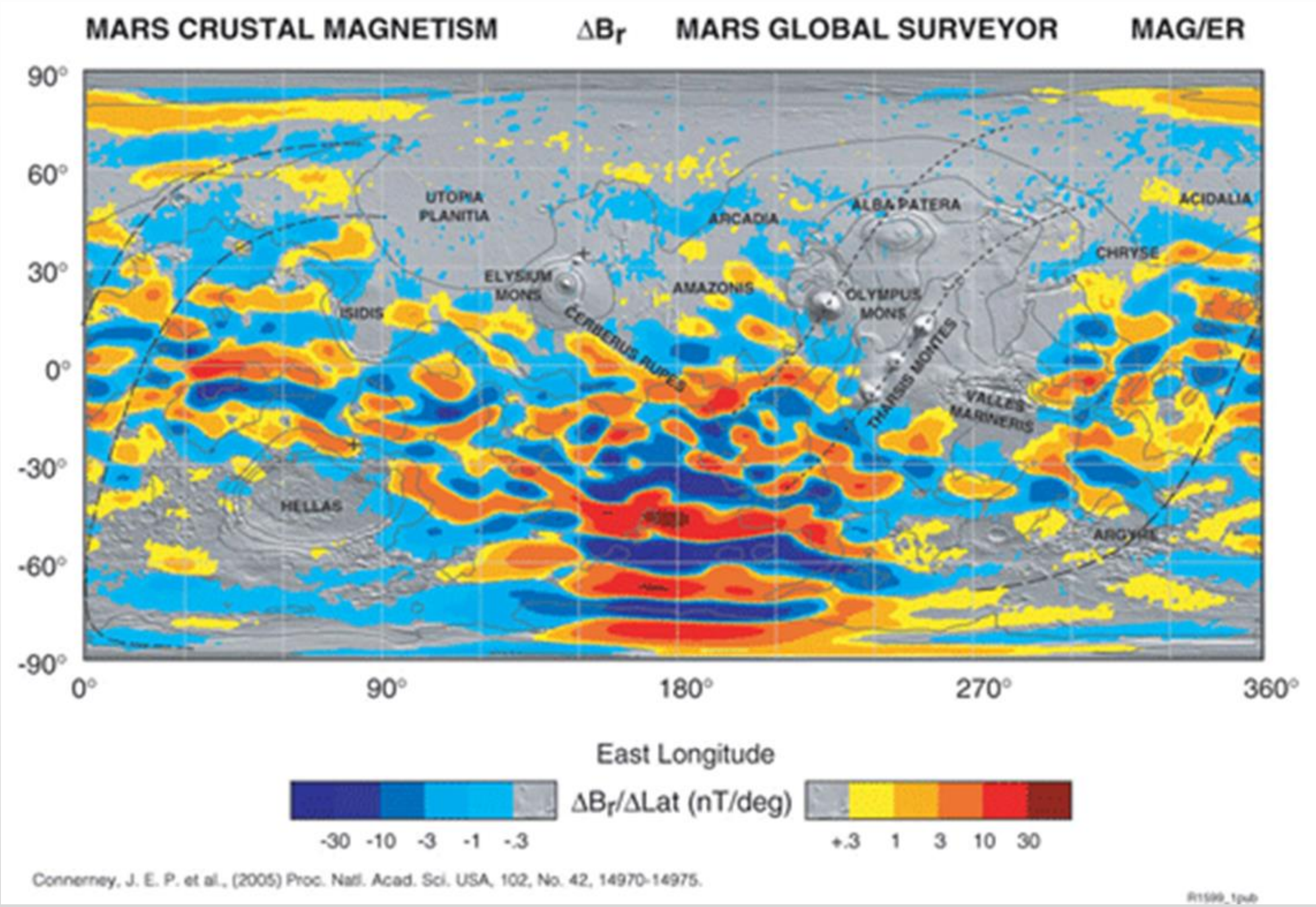
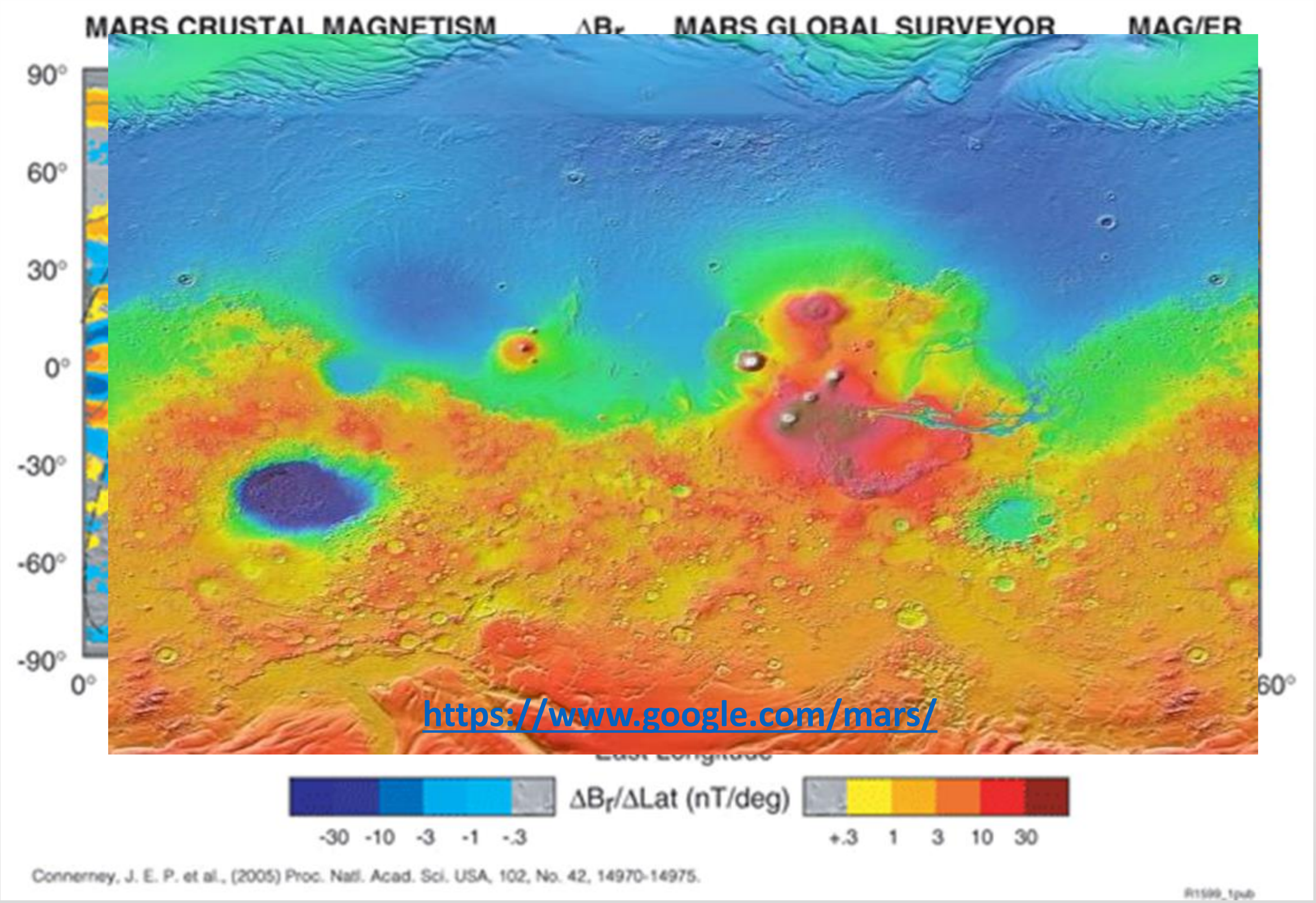


Magnetics



https://www.nasa.gov/mission_pages/mars/news/mgs_plates.html

Magnetics



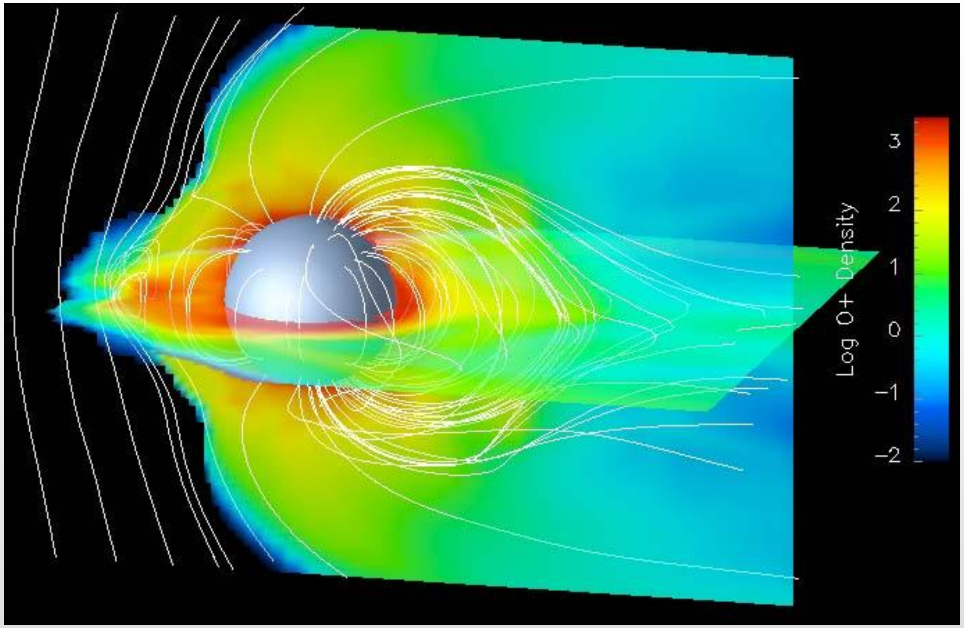
https://www.google.com/mars/

https://www.nasa.gov/mission_pages/mars/news/mgs_plates.html

Magnetics

On other bodies?

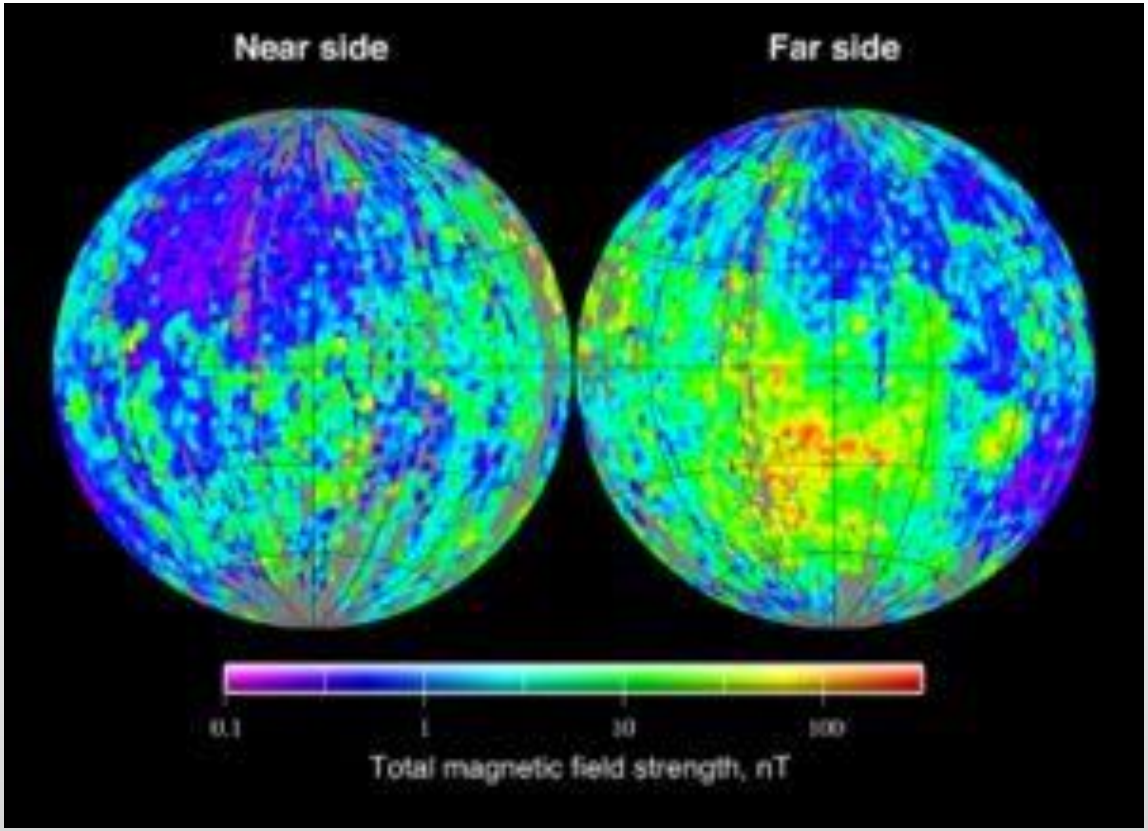
Ganymede does have an active geodynamo and ***Io*** may have one.



Carol Paty, Georgia Institute of Technology

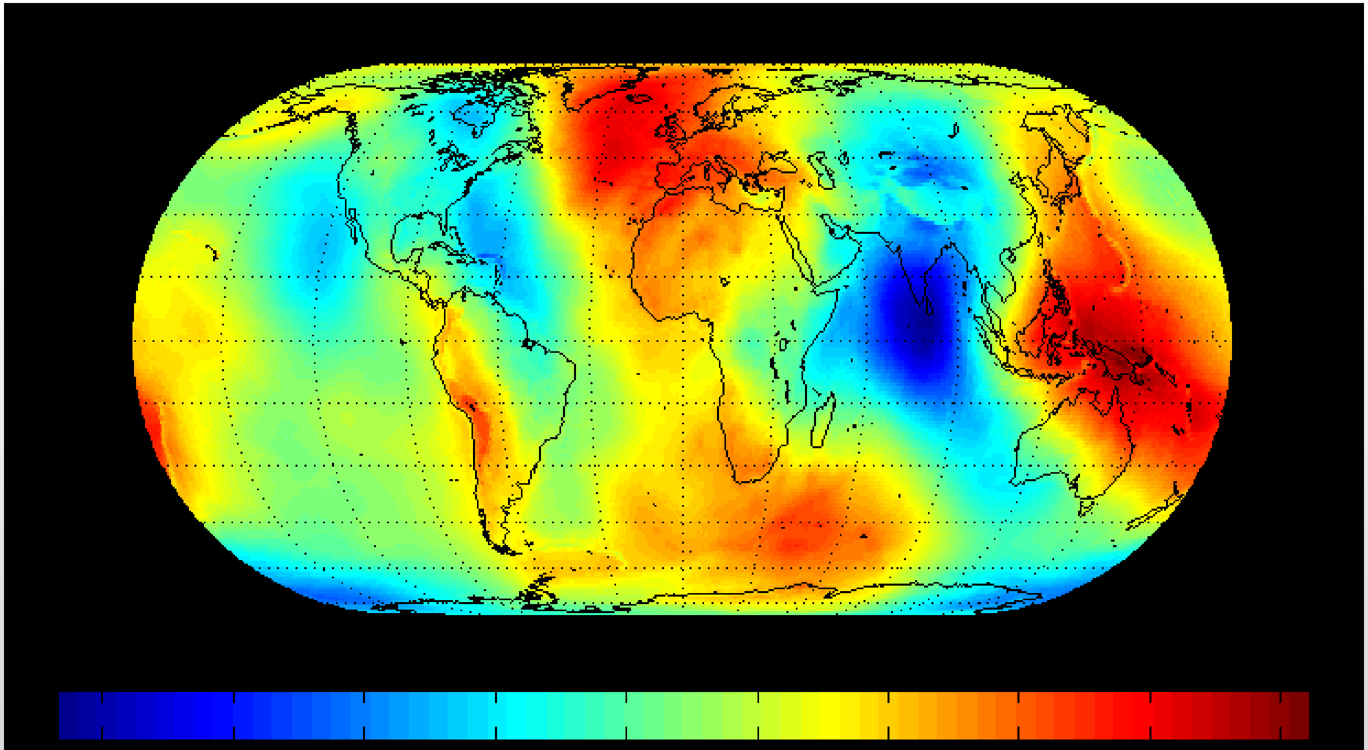
Magnetics

Crustal rocks on the Moon show weak remanence probably resulting from the original “Big Whack” or due to an ancient geodynamo.



Gravity

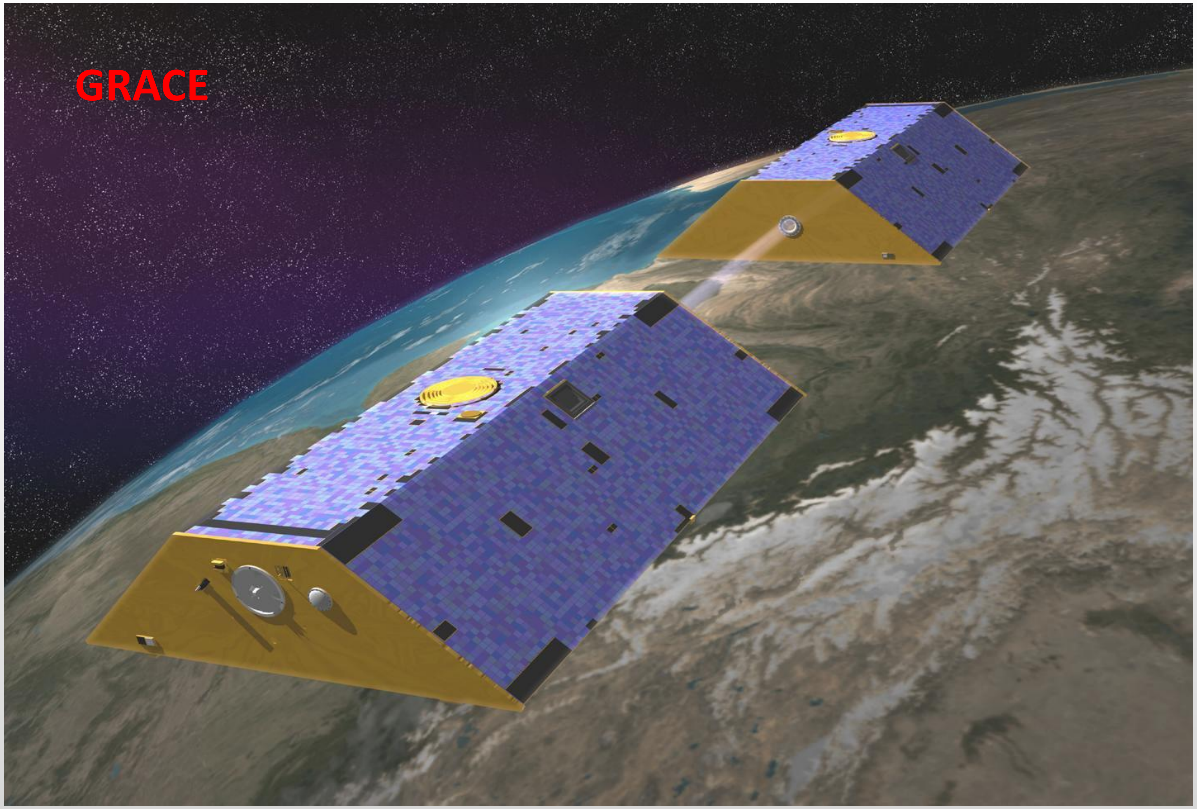
Satellites in close orbit around planets and moons can map the detailed gravity variations that indicates anomalous mass distributions. The Earth's "geoid" maps the gravitational equipotential surface deviations… approximately "sea level"



https://en.wikipedia.org/wiki/EGM96

Grace and Grail

GRACE (for Earth) and **GRAIL** (for Moon) were paired satellites that measure the gradient of the gravity geopotential or geoid. The gradient is much more sensitive to near-surface density variations.



https://en.wikipedia.org/wiki/Gravity_Recovery_and_Climate_Experiment

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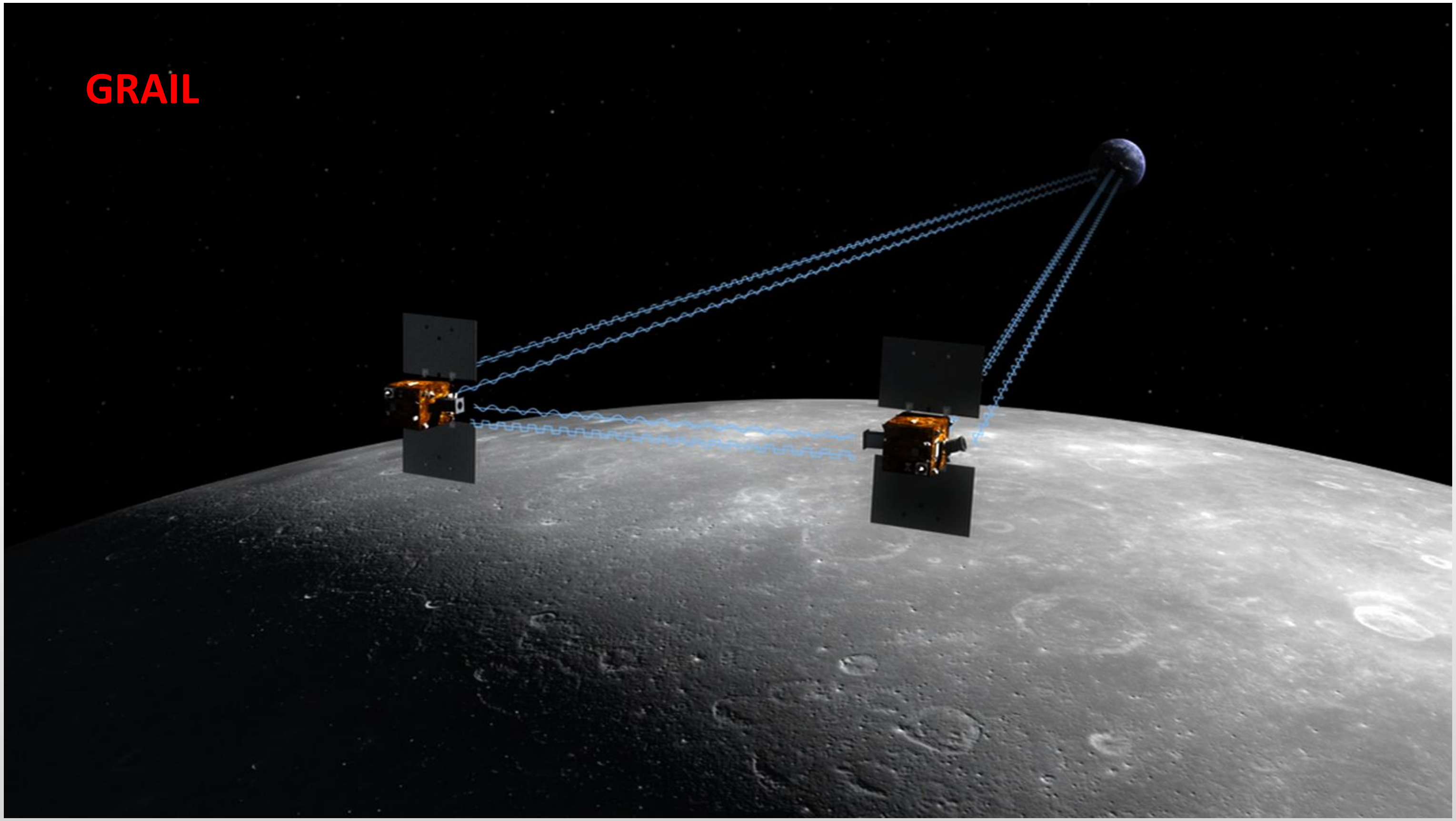

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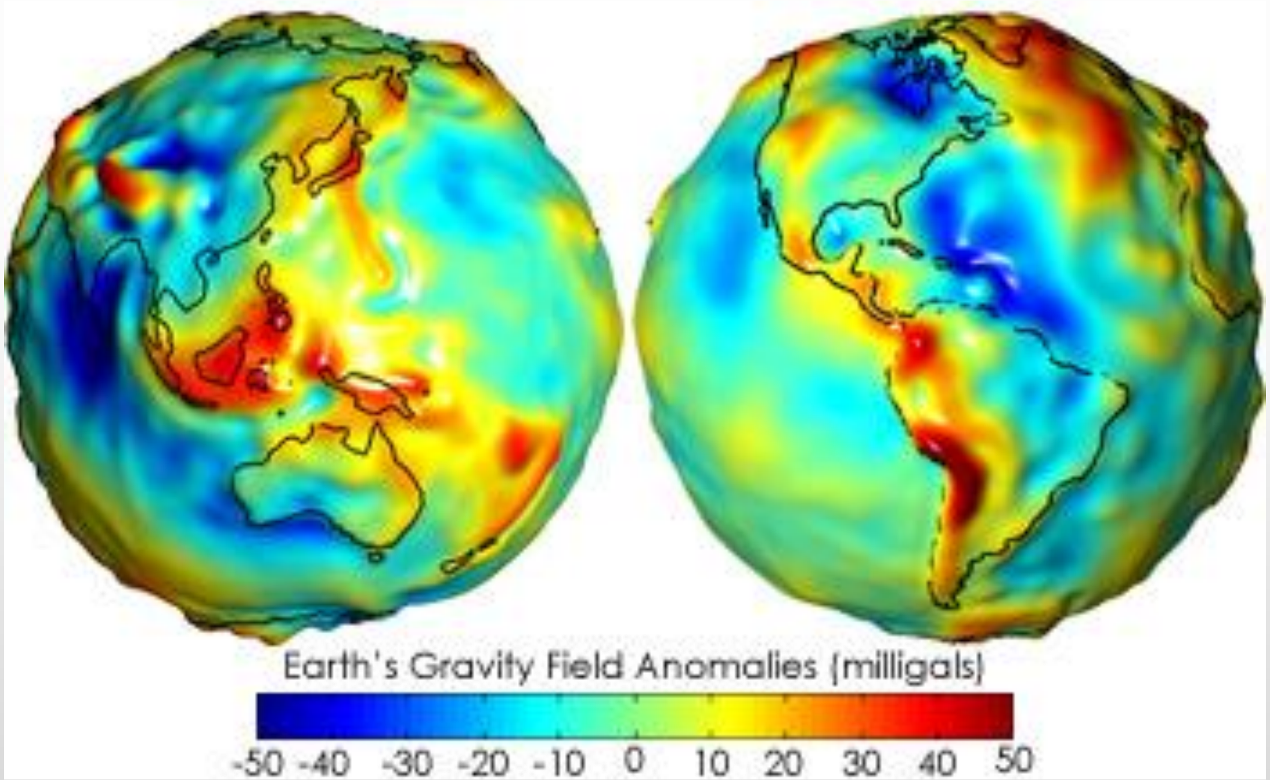
Gravity

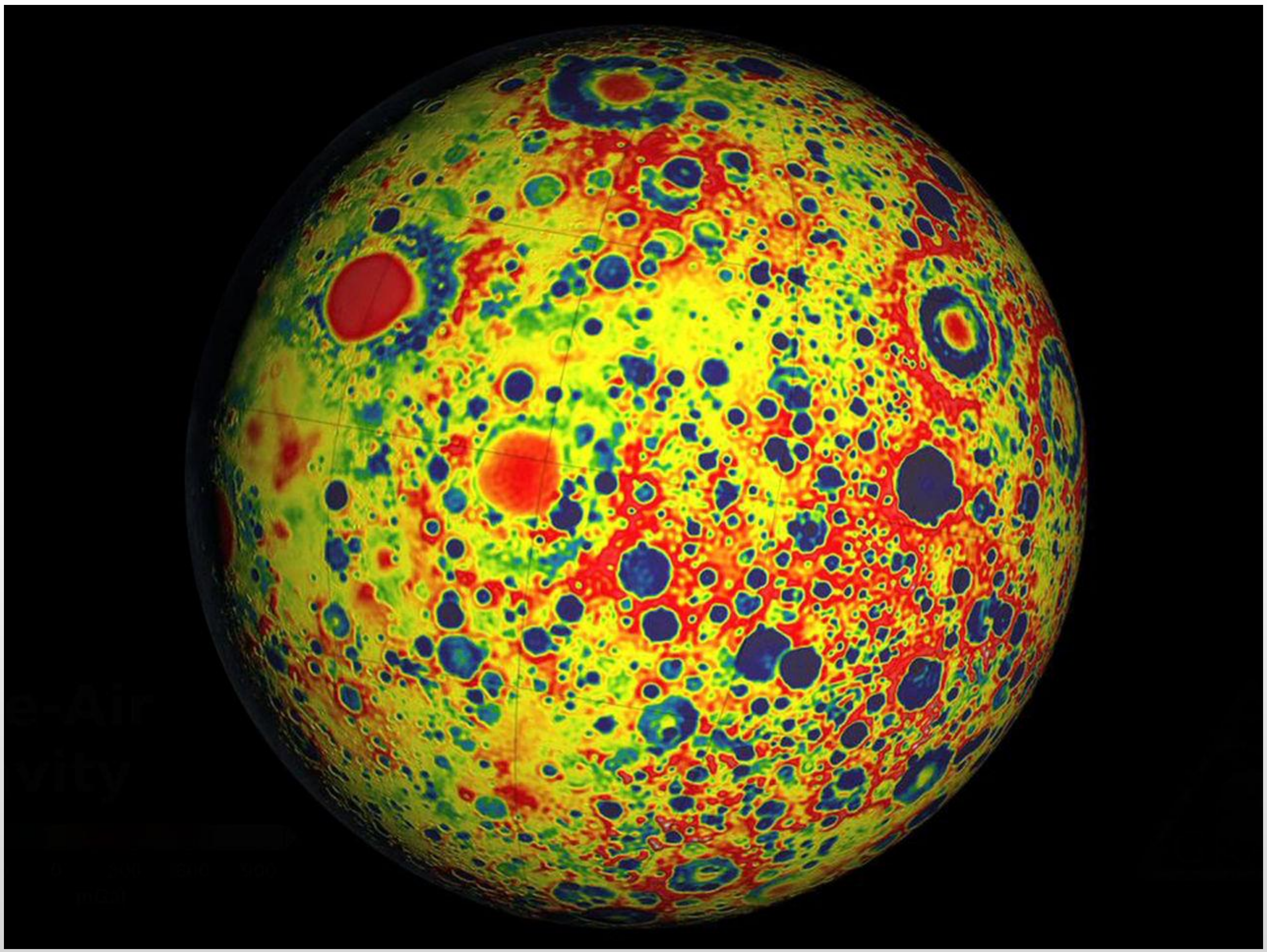
The "gradient" of the geoid determines the gravitational acceleration on the surface. Variations in the acceleration are sensitive to crustal and shallow variations in density. Field anomalies from data by GRACE.



https://en.wikipedia.org/wiki/Gravity_of_Earth

Gravity

Lunar gravity from the GRAIL satellite mission



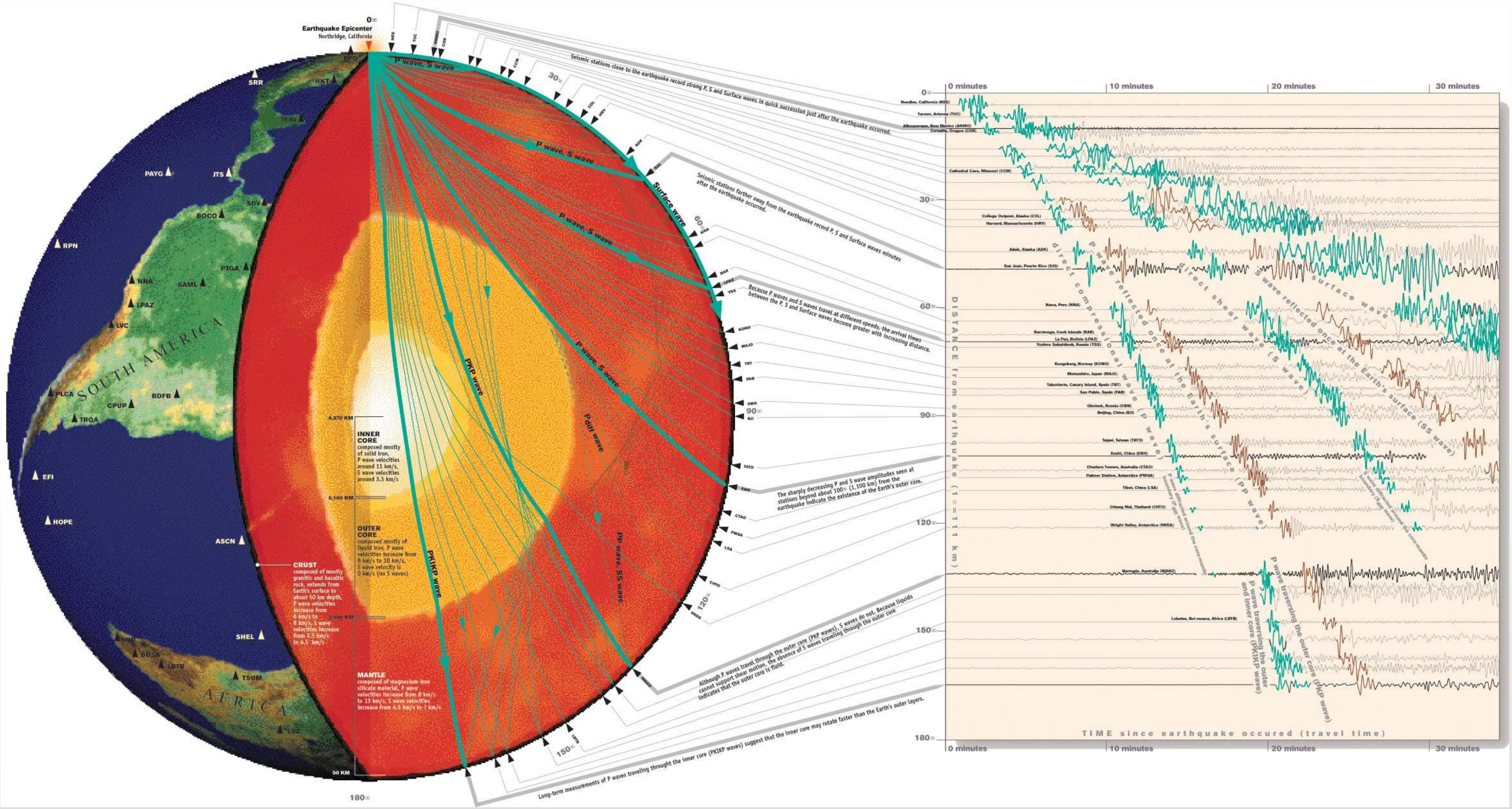
https://en.wikipedia.org/wiki/File:GRAIL%27s_gravity_map_of_the_moon.jpg

Seismology – internal structure

On Earth, earthquakes provide us with seismic waves that penetrate throughout the body of the Earth. We can "deconvolve" the wavefields to determine variations in the internal elastic structure of our planet.

If we were to have seismic instruments on other planets and moons that produced "quakes" like Earth, we could map their interiors too.

Seismology – internal structure



https://www.iris.edu/hq/inclass/lesson/imaging_earths_interior_with_seismic_waves

On Earthquakes

<u>Focus</u>: The point at which an earthquake starts to fracture is called the focus.

<u>Epicentre</u>: The place on the surface directly above the focus is called the earthquake's epicentre.

<u>Fault plane</u>: The surface area over which the fracture occurs is called the fault plane.

<u>Slip</u>: The distance of motion of the fracture.

<u>Seismic waves</u>: By measuring the field of seismic waves that issue from an earthquake, seismologists can determine the plane of faulting, the amount and direction of slip along that fault plane and the amount of energy released in the event and so understand the mechanism and tectonic stresses involved in precipitating an earthquake.

Seismic Wave Types

Earthquakes generate several different wave fields.

- ***P-waves*** (sound waves, compressional waves): they travel through the 3-D volume of the Earth with local speed:

$$\boldsymbol{\alpha} = \sqrt{\frac{\boldsymbol{k} + \frac{2\boldsymbol{\mu}}{3}}{\boldsymbol{\rho}}}$$

- ***S-waves*** (shear waves – only in elastic solids): They travel through the volume of Earth with local speed:

$$\boldsymbol{\beta} = \sqrt{\frac{\boldsymbol{\mu}}{\boldsymbol{\rho}}}$$

k: bulk incompressibility; $\boldsymbol{\mu}$: elastic rigidity; $\boldsymbol{\rho}$: material density

Seismic Waves

- Body waves—pass through Earth's interior.
  - P-waves (primary or compressional waves).
    - Waves travel by compressing and expanding material.
    - Material moves back and forth parallel to wave direction.
    - P-waves are the fastest.
    - They travel through solids, liquids, and gases.

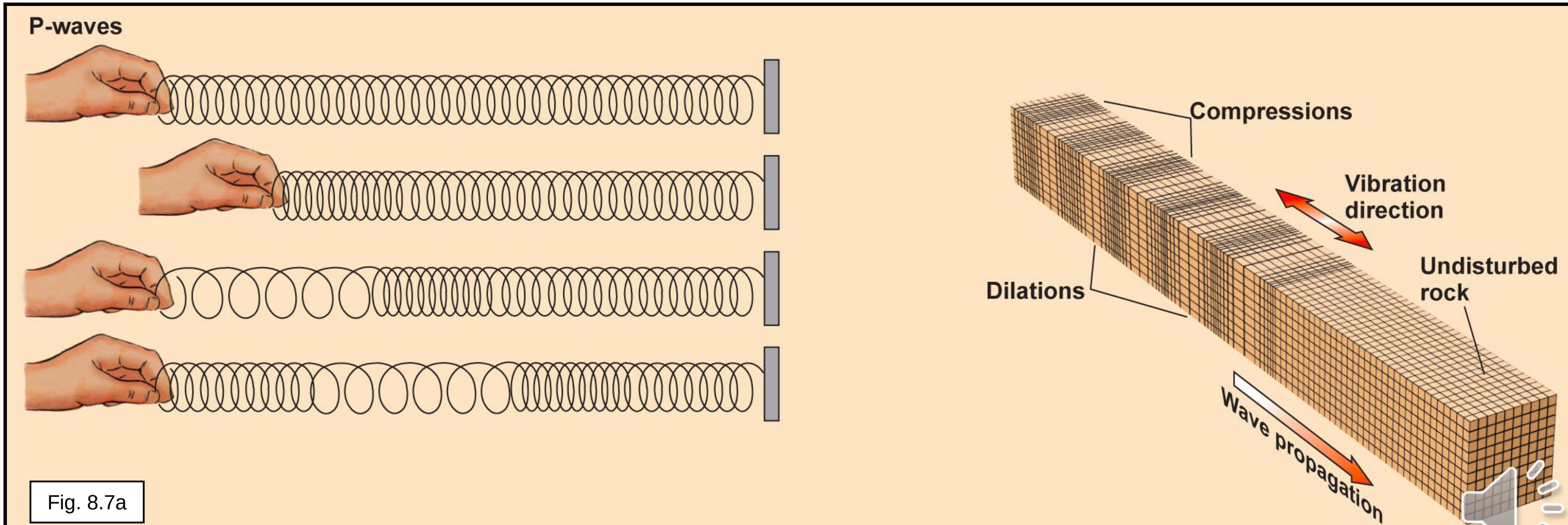


Fig. 8.7a

Seismic Waves

- **Body waves—pass through Earth's interior.**
  - **S-waves (secondary or shear waves).**
    - **Waves travel by moving material back and forth.**
    - **Material moves perpendicular to wave travel direction.**
    - **S-waves are slower than P-waves.**
    - **They travel only through solids, never liquids or gases.**

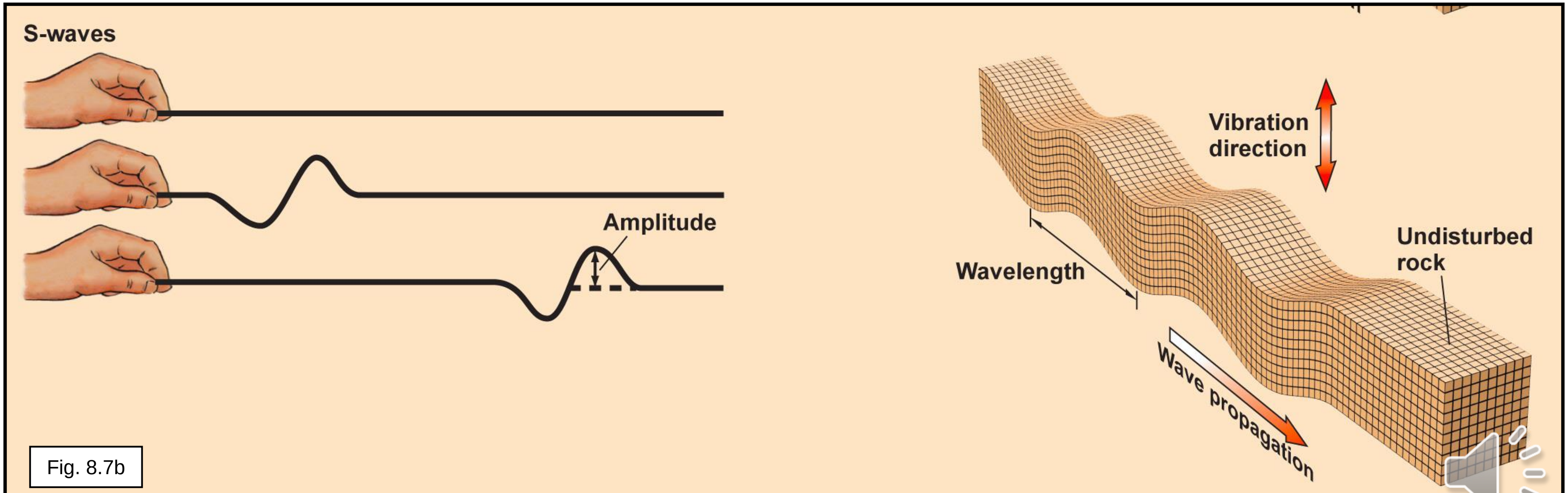


Fig. 8.7b

Surface waves

Surface waves are composed of P-waves and S-waves that are confined to travelling as a boundary wave across the surface.

- ***Rayleigh waves*** (rolling waves producing up-down surface motions): These are composed of elements of both P-waves and S-wave. They travel with a speed that is approximately $\boldsymbol{0.9\,\beta}$.

- ***Love waves*** (waves with lateral motions confined to the surface): These are S-waves that are trapped or guided by the elastic layering of the near surface layers. Their speeds depend strongly on their wavelengths or periods.

Seismic Waves

- **Surface waves—travel along Earth's exterior.**
- **Surface waves are the slowest and most destructive.**
  - **LQ-waves (Love waves)**
    - **S-waves that intersect the land surface.**
    - **Move the ground back and forth like a writhing snake.**

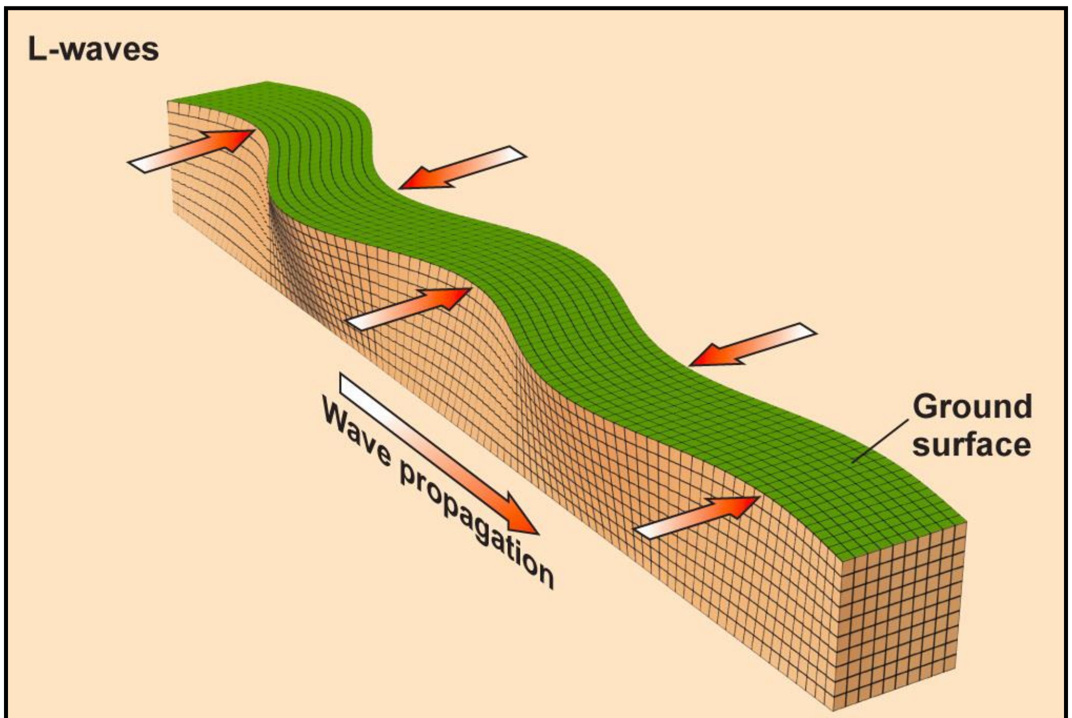


Fig. 8.7c

Seismic Waves

- **Surface waves—travel along Earth's exterior.**
  - **LR-waves (Rayleigh waves)**
    - **P-waves that intersect the land surface.**
    - **Cause the ground to ripple up and down like water.**

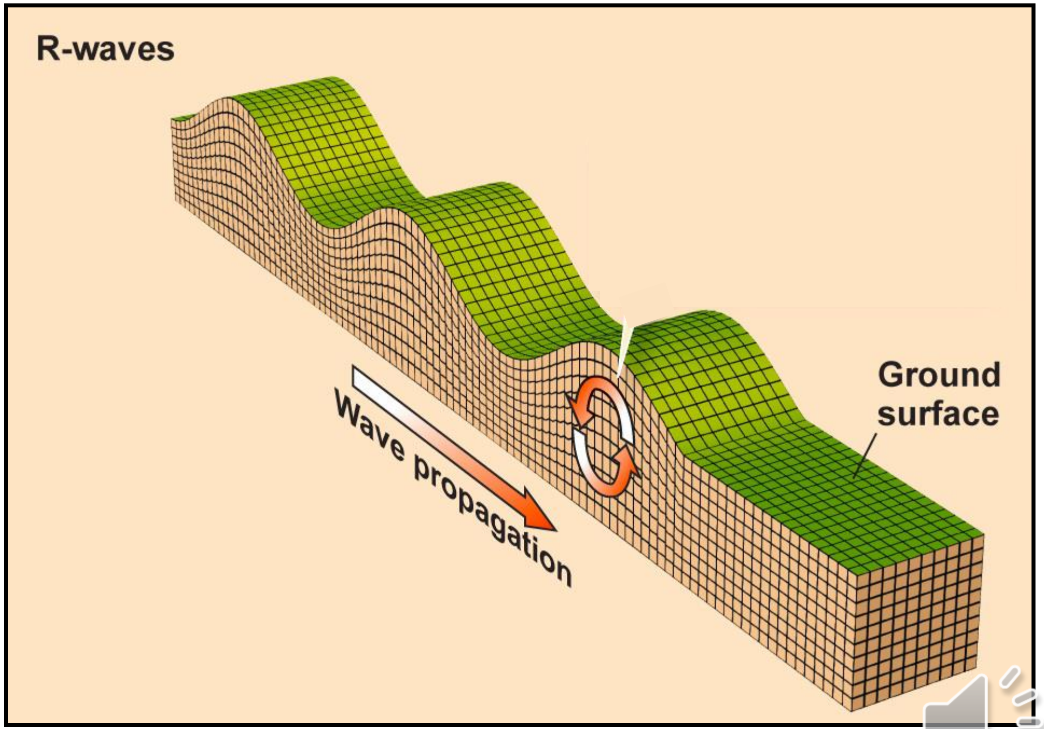


Fig. 8.7c

Magnitude scales

Earthquakes scale over an enormous range of sizes, from small local crackings to the great megathrust events. Magnitude scales describe the "size", the source-energy release of an earthquake.

Seismic moment: A simple model of an earthquake in which a slip $\boldsymbol{s}$ occurs over a fault plane surface $\boldsymbol{A}$ under frictional resistance $\boldsymbol{\mu}$ determines the seismic moment:

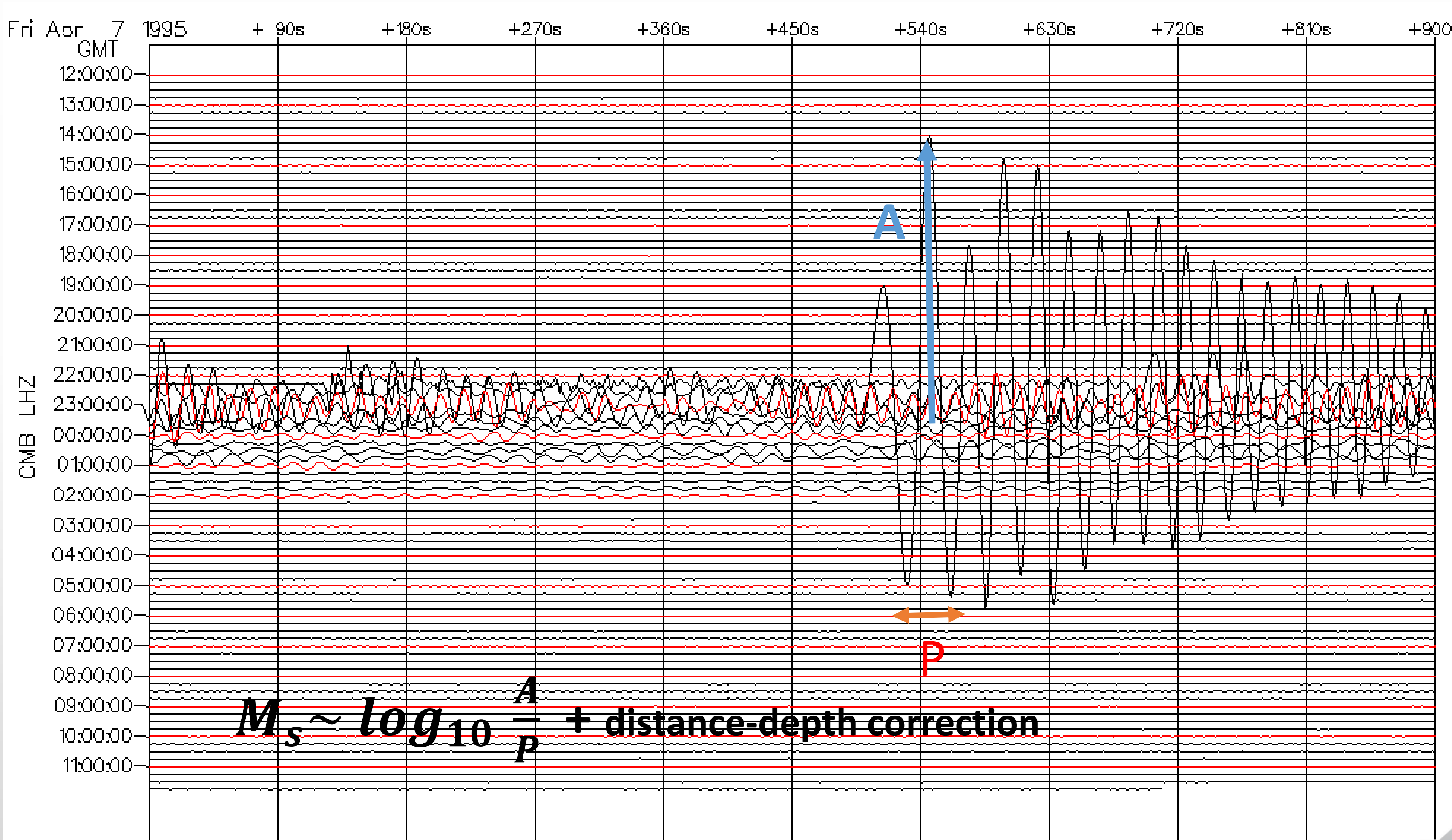
$$\boldsymbol{M_0 = \mu \cdot s \cdot A}$$

Moment magnitude scale: If we calculate $\boldsymbol{M_0}$ using **SI units** (i.e., in newton-metre), we determine the moment magnitude based on the "***number***" result of that calculation:

$$\boldsymbol{M_w = \frac{2}{3} \log_{10} M_0} - 6.1$$

You might notice that each increasing step in $\boldsymbol{M_w}$ by 1 unit represents a factor of 32 increase in $\boldsymbol{M_0}$ and energy release.

Older magnitude scales



$$M_s \sim log_{10} \frac{A}{P} + \text{distance-depth correction}$$

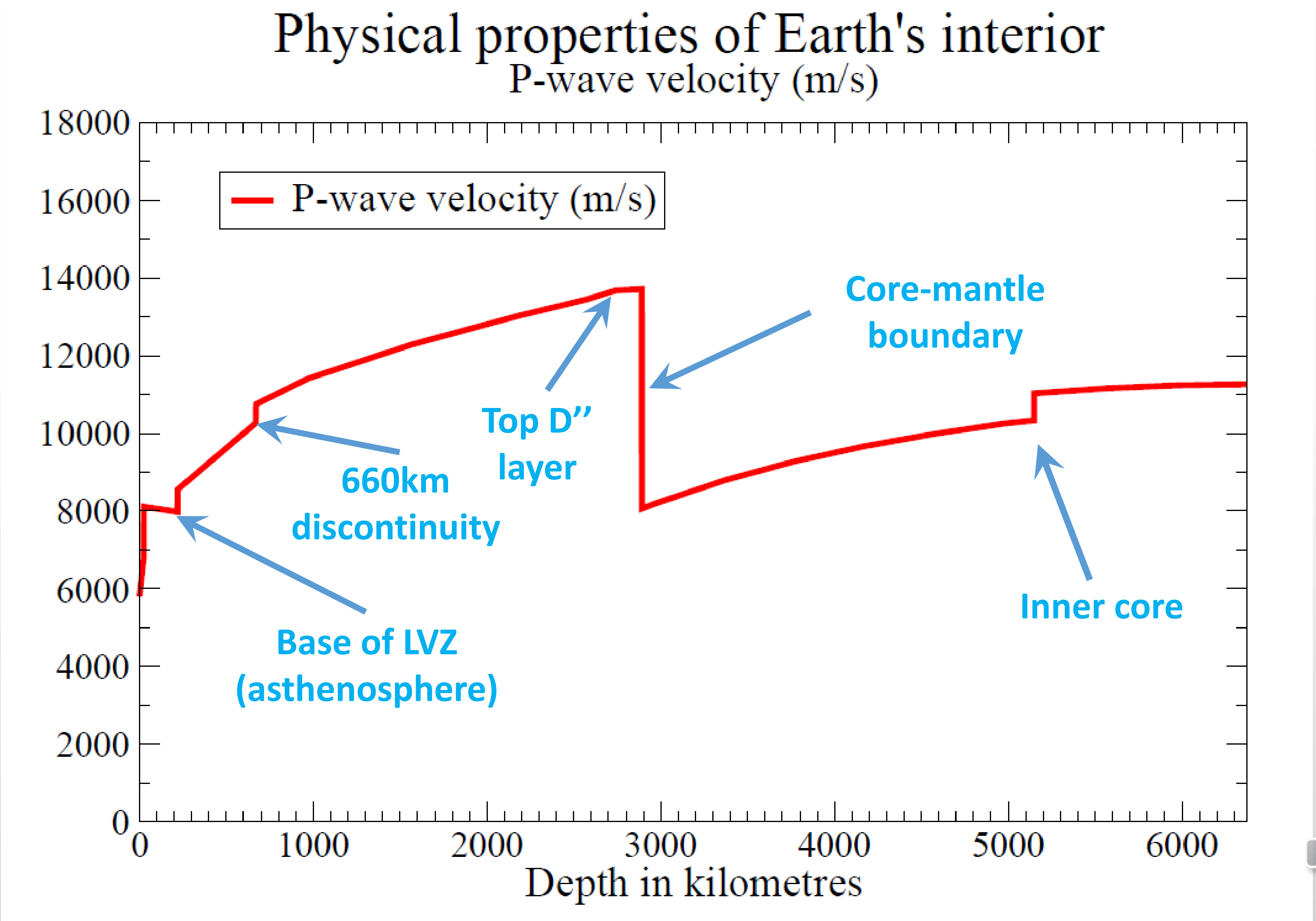
Classical seismogram: Tonga April 7, 1995

Earth Structure by Seismology

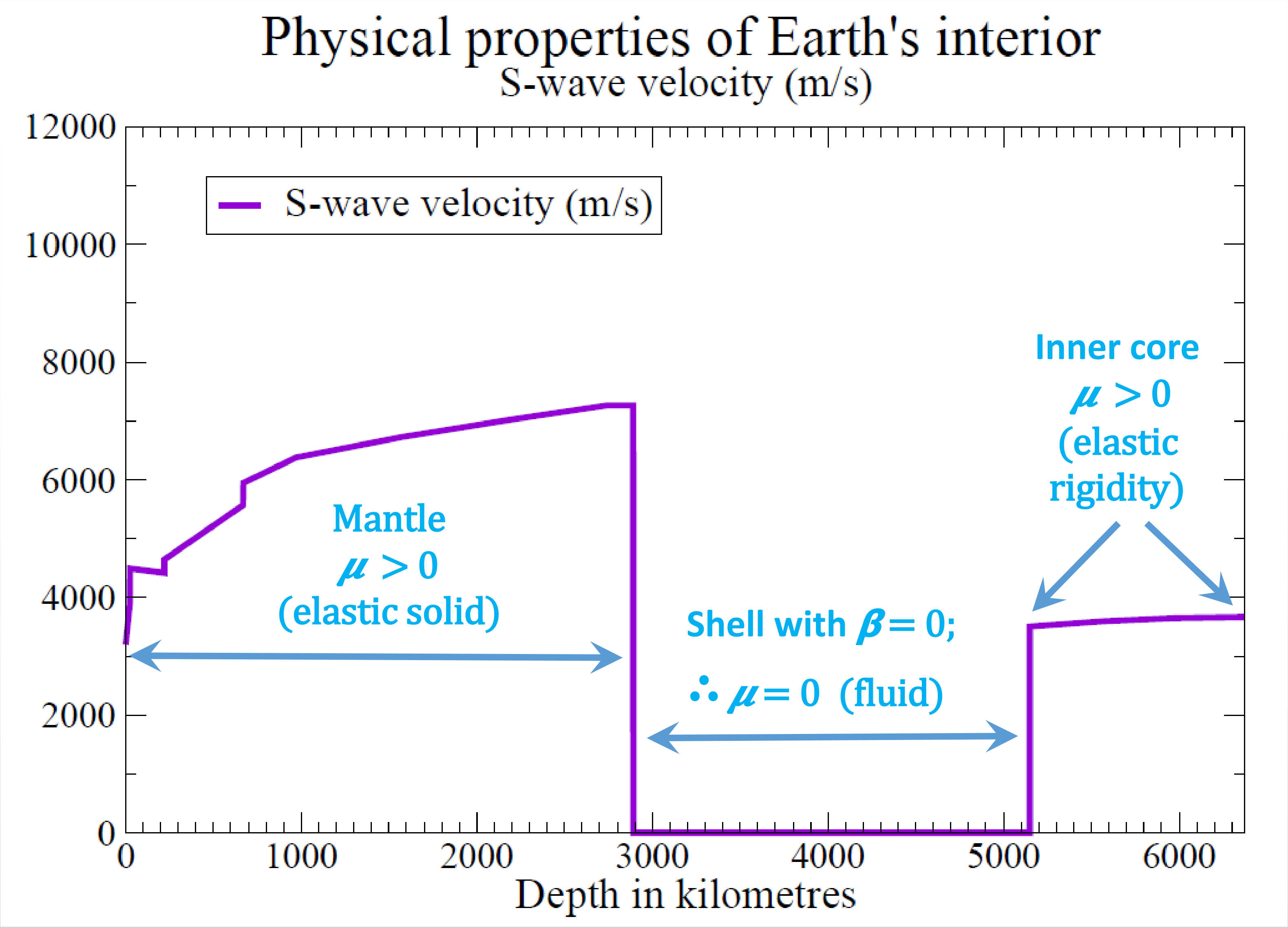
With the virtually millions of pairing of seismic observation sites and earthquake sources we had, by the early 1970s, obtained a detailed view of the seismic velocities within the Earth. Constrained by our knowledge of the Earth's moment of inertia, inversion of this data obtains the raw mechanical properties.

Geochemists and mineral physicists accord these the mechanical properties with mantle mineralogy.

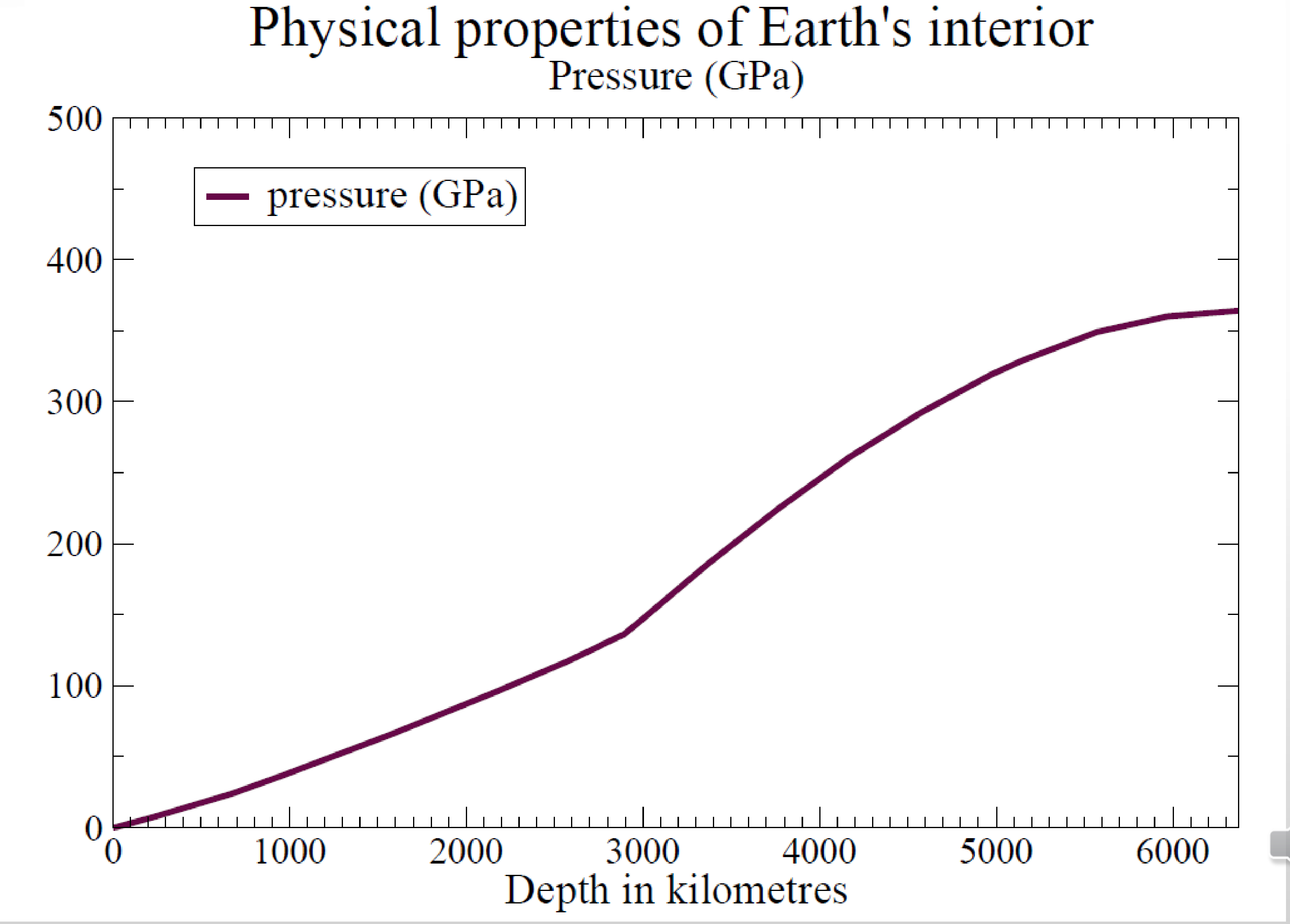
Earth Structure by Seismology



Earth Structure by Seismology



Earth Structure by Seismology

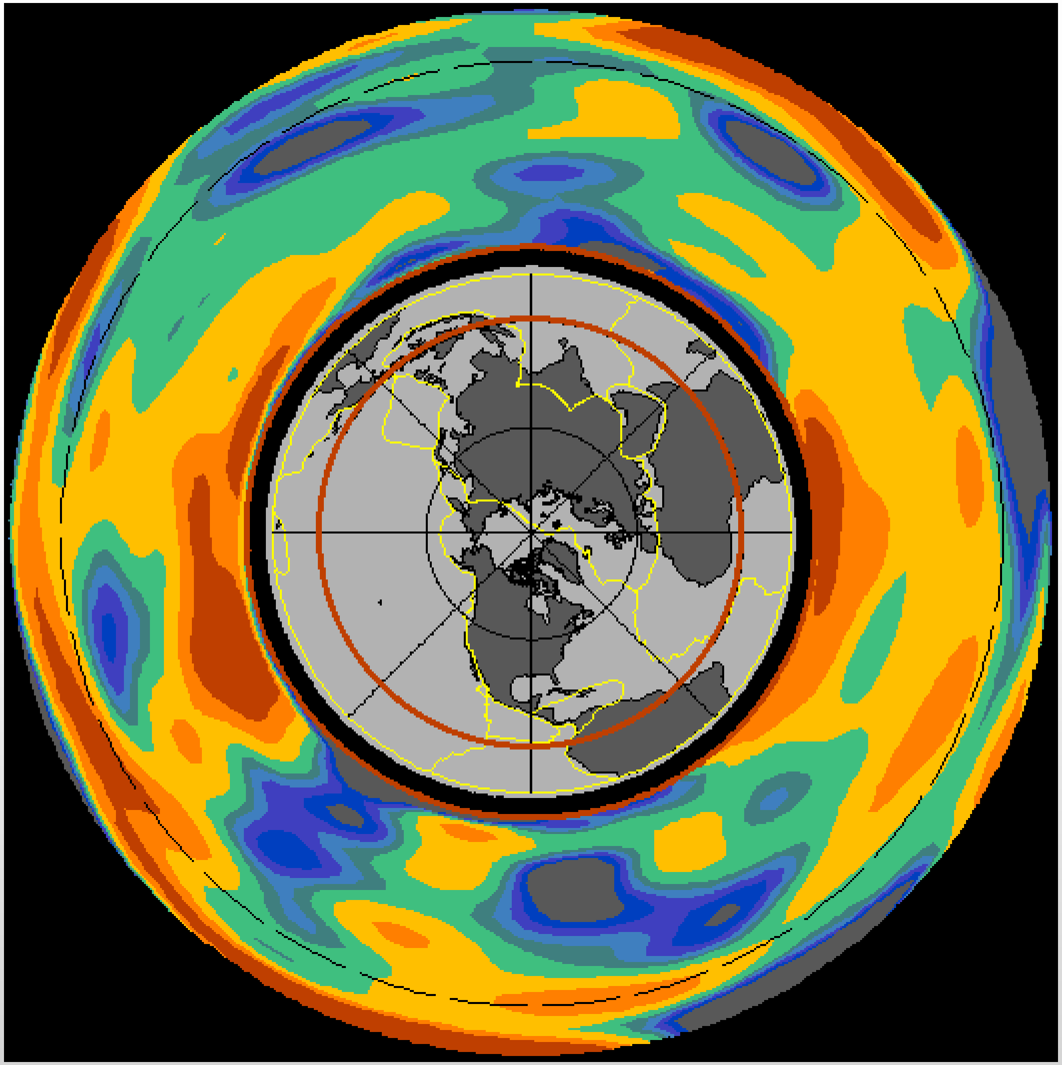


Seismic Tomography

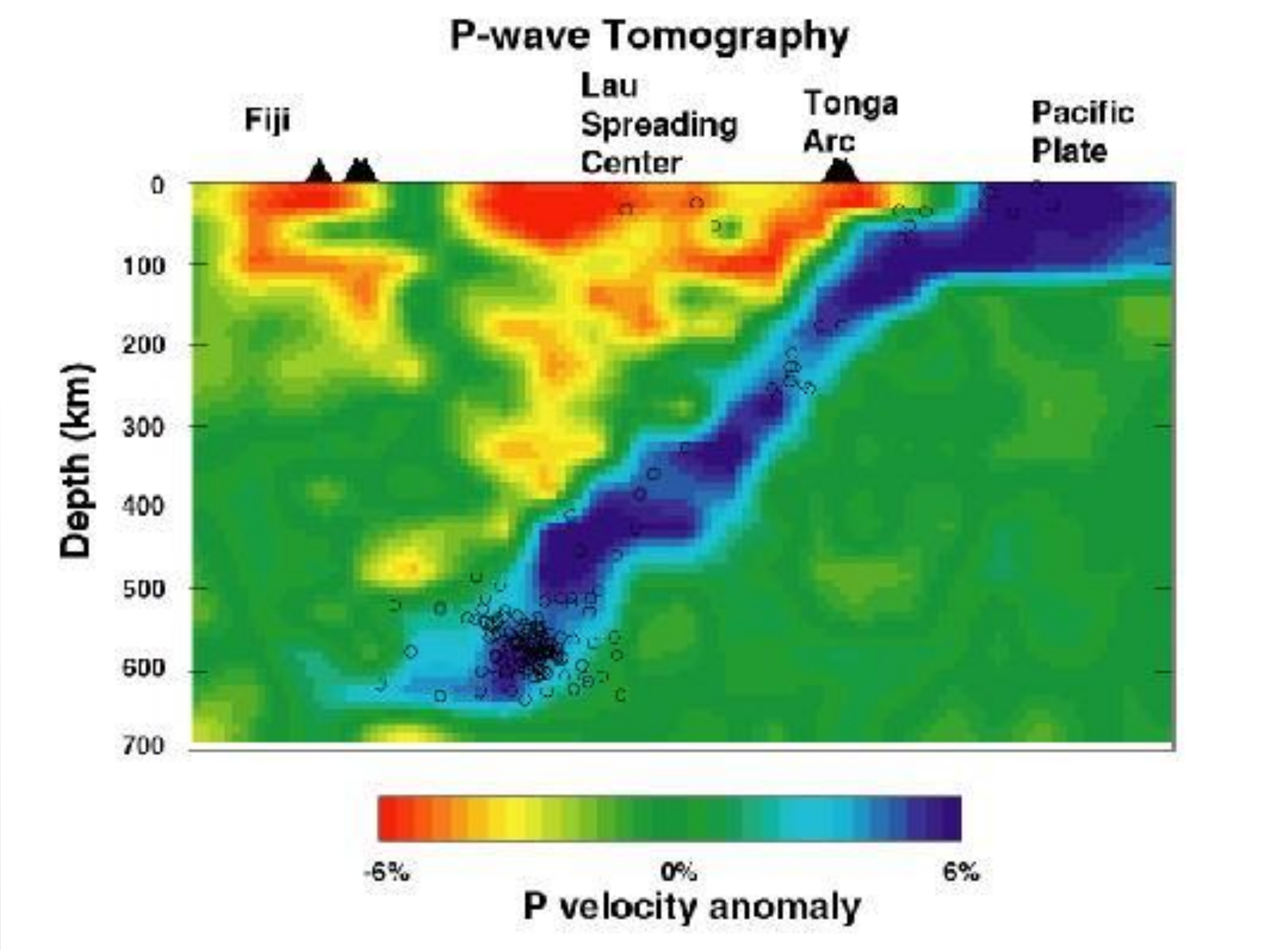
One can perform the equivalent of a CAT-Scan on the Earth using seismic waves. In fact, seismic tomography is an older science than is medical tomography. By looking to the localized anomalies in the seismic wave velocities according to places within the Earth, we can actually track the history of descending tectonic plates and upwellings beneath spreading ridges. As the science evolves and as ever more data are obtained, we are producing a detailed picture of Earth's interior.

Tomographic anomalies in seismic velocities are normally referenced to the **PREM** (Preliminary Reference Earth Model), that from which the previous velocity-depth profiles were taken. Small variations in the temperature of flowing mantle materials produces slightly **negative velocity anomaly if warm** than the depth average and **slightly positive anomaly if cool**.

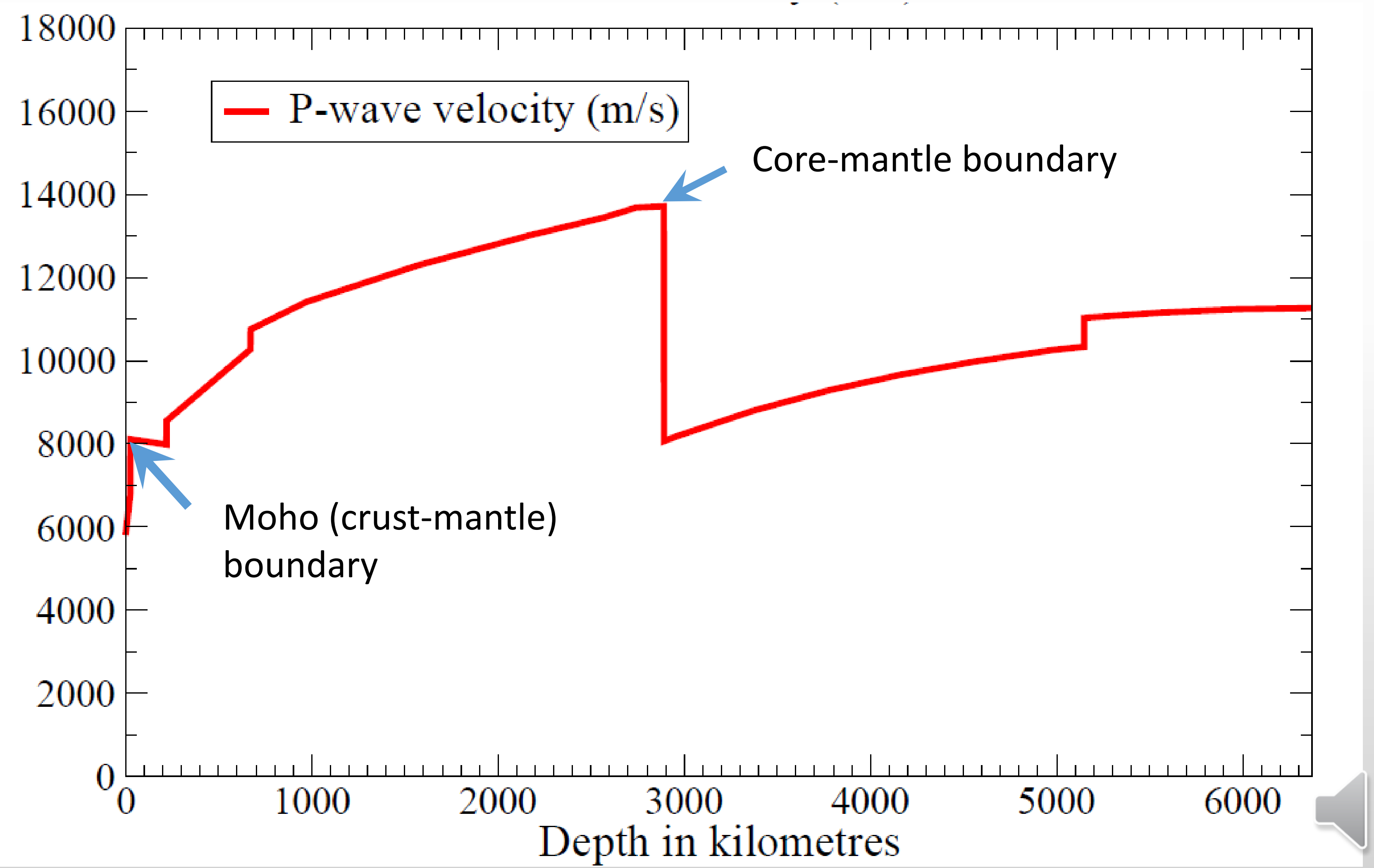
Seismic Tomography



Seismic Tomography



Geology/Mineralogy by Seismology



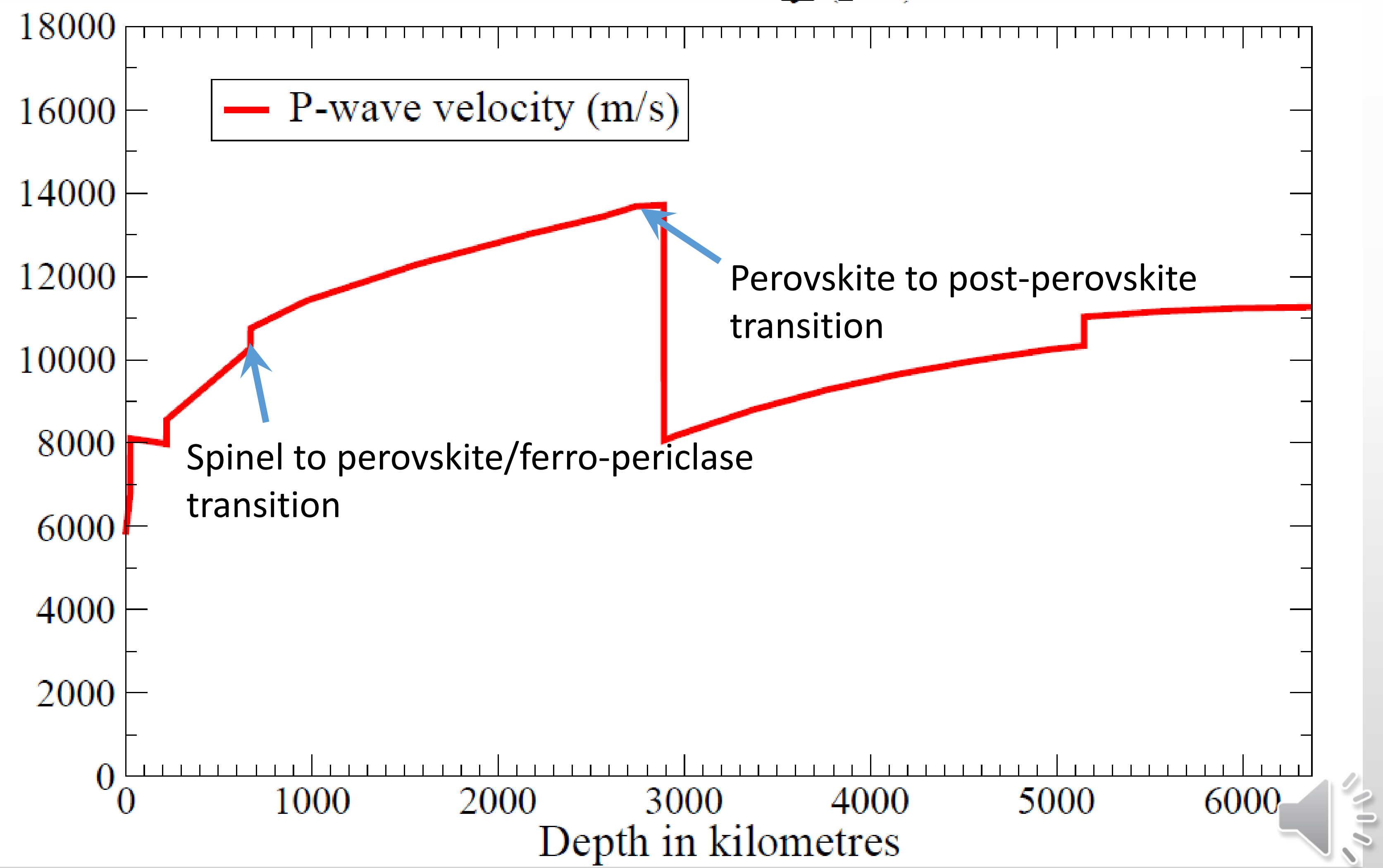
Geology/Mineralogy by Seismology

– **Crust**: This outer skin, typically about 33km thick, shows a relatively low velocity for seismic sound waves, the (P-waves). Its base is characterized by a sharp increase in the P-wave velocity to about 8km/s.

– **Mantle**: Most of the volume (≈ 87%) of the Earth comprises the silicate, rocky mantle. The mantle has a layered structure that affects seismic wave velocities. The structure is caused by phase changes of the mantle minerals as they are compressed at depth into higher density forms.

The most distinct boundaries are at ≈ **440km** where olivine, **$[Mg, Fe]_2SiO_4$**, becomes compressed into the spinel structure and then at ≈ **660km** where the spinel becomes further compressed into a mix of **perovskite**, **$[Mg, Fe]SiO_3$**, and f**erro-periclase**, (or **magnesiowustite**) **[Fe,Mg]O**. The lower mantle is thought to maintain this mix to the core-mantle boundary at 2970km depth. At, perhaps 2770km (the **D'' discontinuity**) the pressure of **1.25GPa** allows for the perovskite mineral form to compress into the **post-perovskite** structure. If this is the cause of the elastic discontinuity, **D''**, observed by seismic experiments, the temperature there is established at about 2500K.

Geology/Mineralogy by Seismology



Geology/Mineralogy by Seismology

– **<u>Outer core</u>**: The outer core is essentially a liquid mix of iron, **Fe**, and nickel, **Ni**, with some alloying lighter elements, probably, sulfur, **S**, and possibly oxygen, **O**, or silicon, **Si**, and even carbon, **C**.

-- **<u>Inner core</u>**: At the centre of this liquid core is the frozen, crystalline probably almost pure iron (though perhaps with some **Ni** and C**o** and even traces of potassium, **K**) inner core. Its radius is 1222km.

The outer core and inner core contain about 35% of the Earth's total mass in less than 1/8 of its volume.

Geology/Mineralogy by Seismology – other planets

– **Moon:** In 1969, the **Apollo 11** mission installed seismograph systems on the Moon. Moonquakes revealed that there is a small core of the Moon; the density of the Moon suggests that it is like a mix of **Fe** and **FeS**.

-- **Mars**: In 1976, the **Viking 2 Lander** installed a seismograph on Mars. It operated for months and saw nothing that could be securely attributed to a ***tectonic marsquake***. We have no information about Mars' deep interior from seismology…. Yet! We did learn that it has a thin crust, 14-18km thick.

In May, 2018, the **InSight** lander was launched toward Mars. It installed another seismic station on the surface in December, 2018. Over the following few months, it is recording seismic surface motions that may well reveal detail about the planet's deep interior structure.

Age determination of Planetary Surfaces

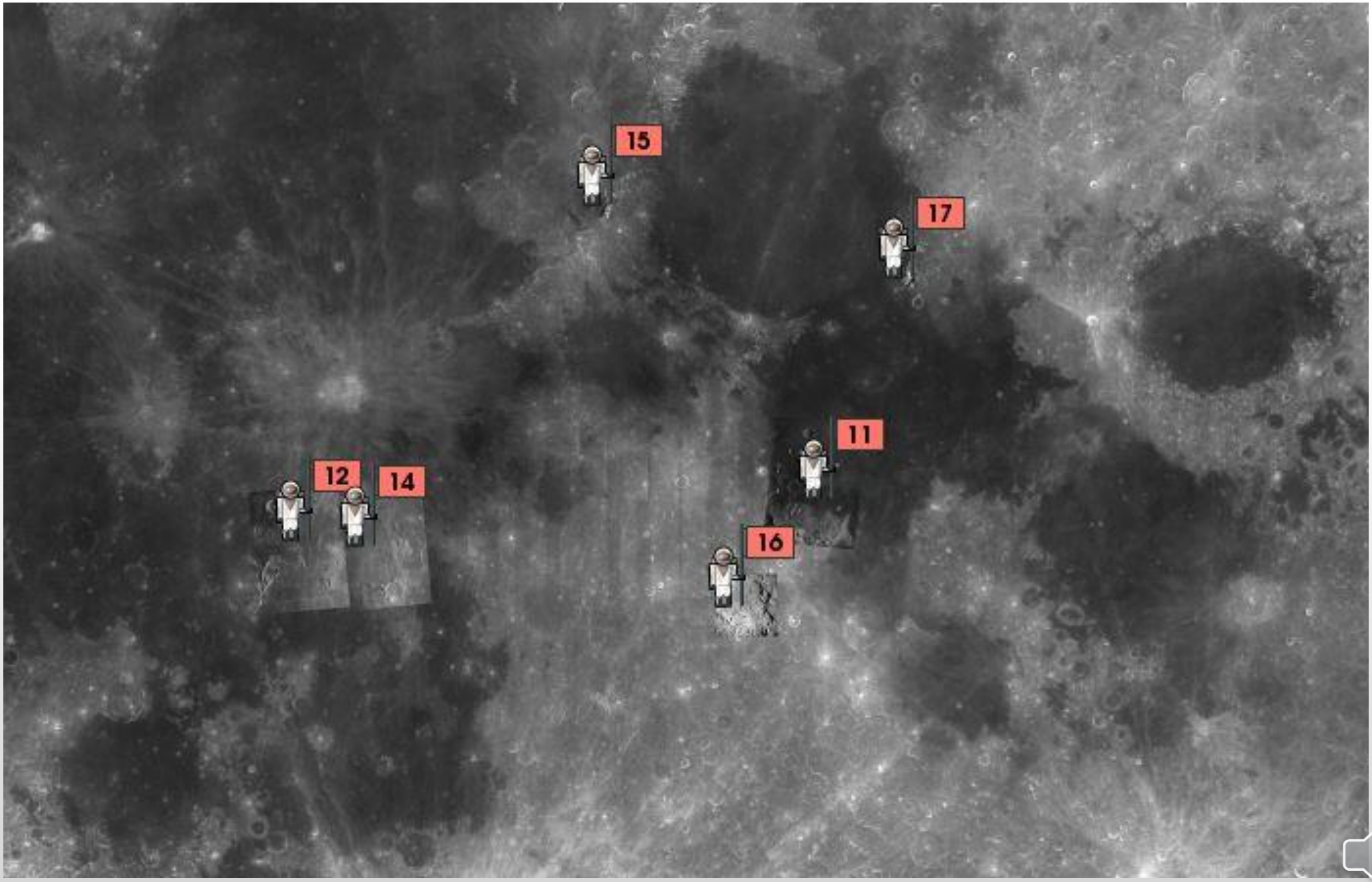
The final topic for the course

Except for the Moon , we have no rocks or soils collected from other bodies in the Solar System that we can use in the age-determination of their surfaces.

The 1969-72 Apollo mission to the Moon returned over 300kg of rock and soil from various locations on the Moon. These have been extensively studied during these past 45+ years.

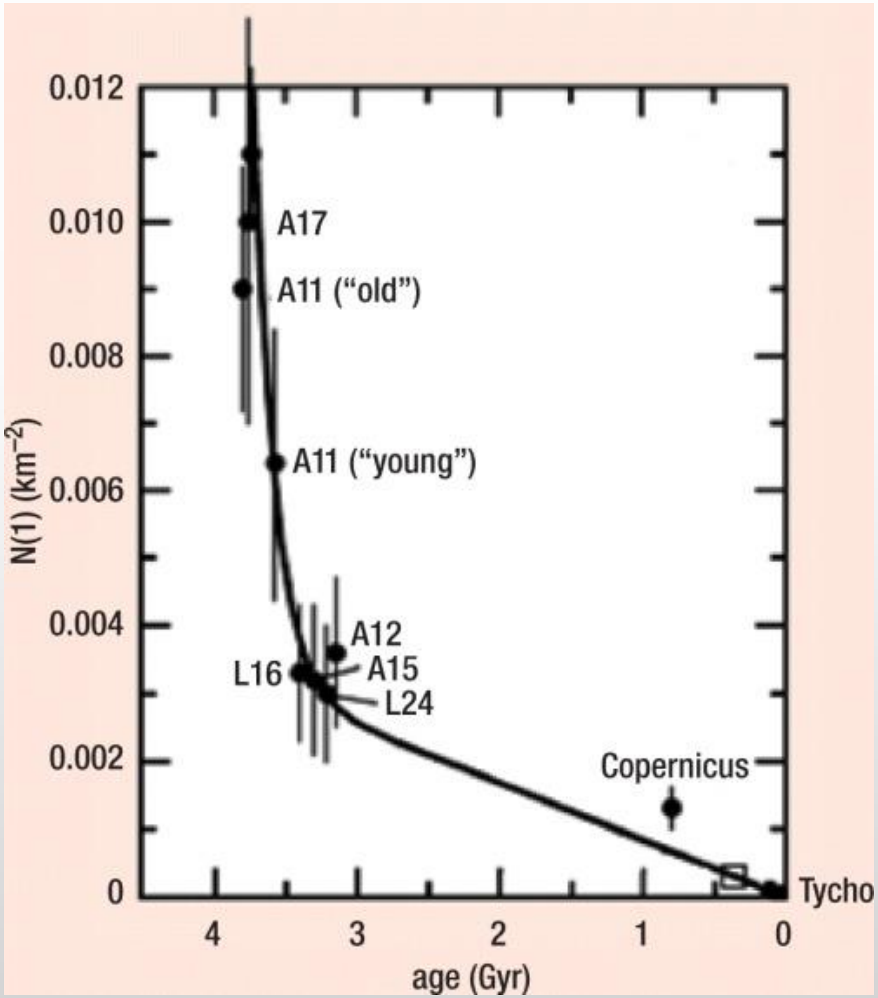
On major use of these rocks was to determine the “age” of the surface at various lunar locations. This age determination was then compared to the surface blemishes due to impactors in the region of collection. It allowed for the calibration of a “crater-density clock”.

The Crater-density Clock



https://www.google.com/moon/

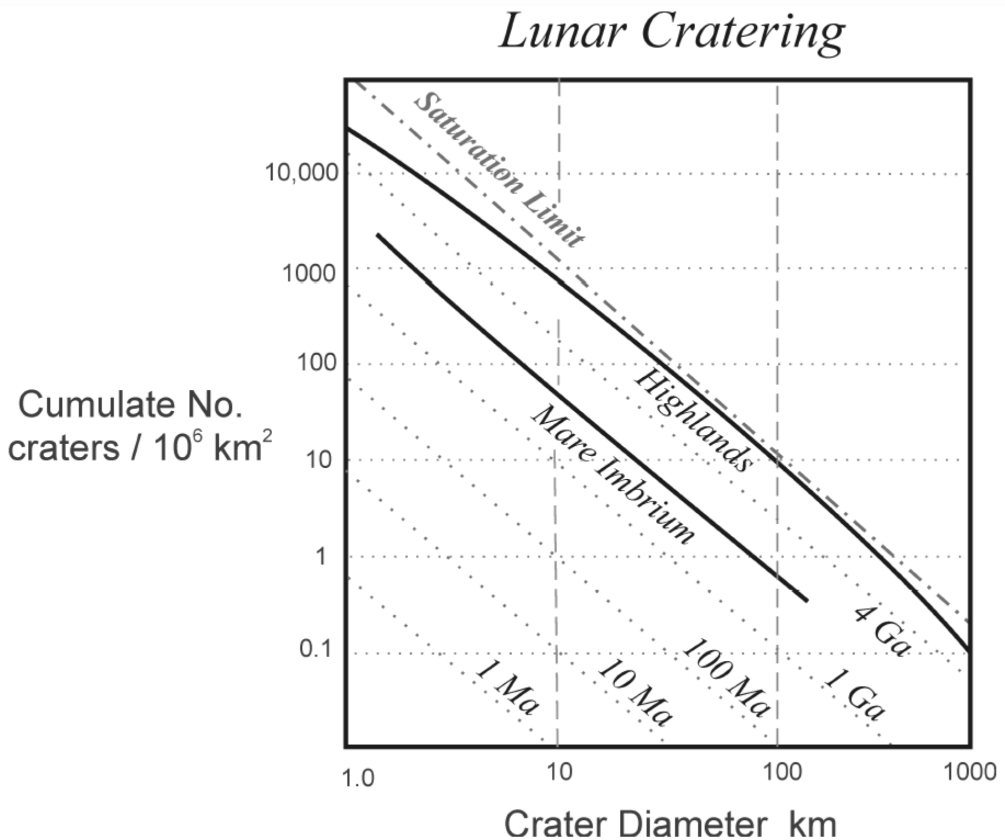
Calibration: Crater-density clock



This model (the line) compares the number of craters/km^2 to the rock ages obtained by the Apollo Missions and the rock/soil return by the robotic Luna 16 and 24 missions by the USSR.

Diagram by [Mineralogical Society of America]

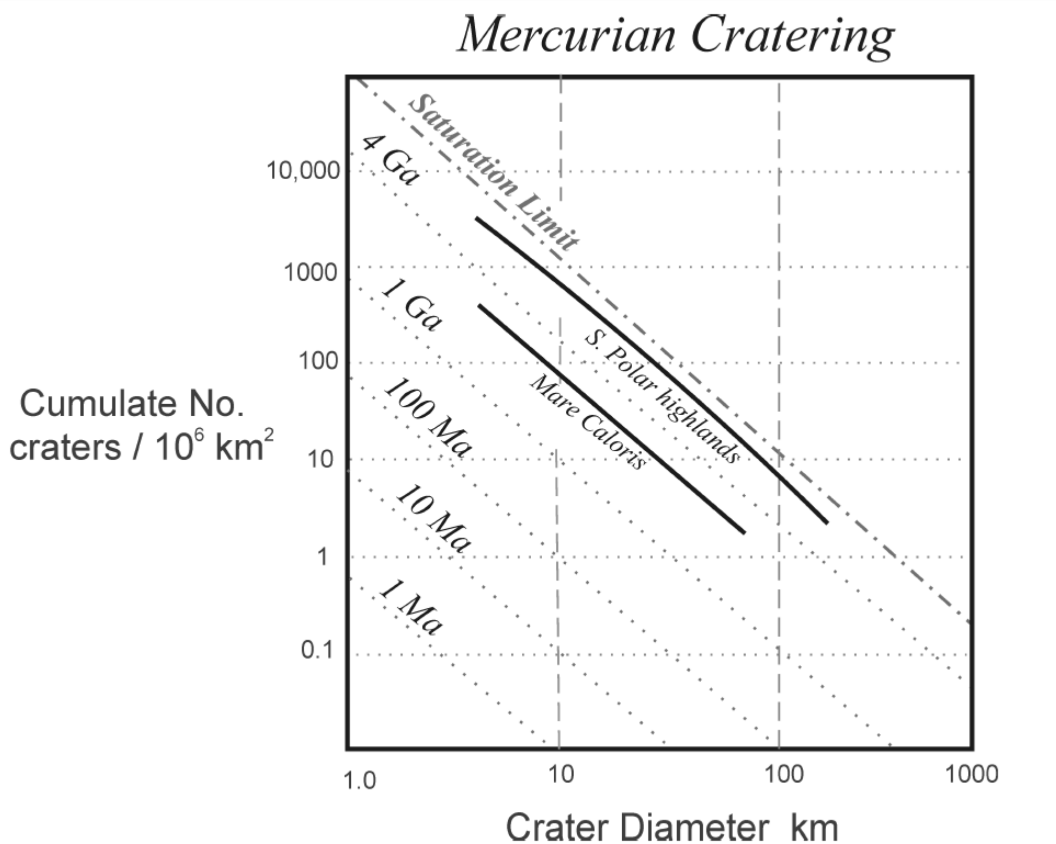
Crater Density Age: Moon



Lunar cratering and age of the Moon's surface features
(after D. Francis, Earth & Planetary Sciences, McGill)

The highlands of the Moon show crater-density measured ages at the "**saturation limit**" dating back to the formation of the Moon. The **Mare** basins are younger, formed by subsequent impacts.

Crater Density Age: Mercury

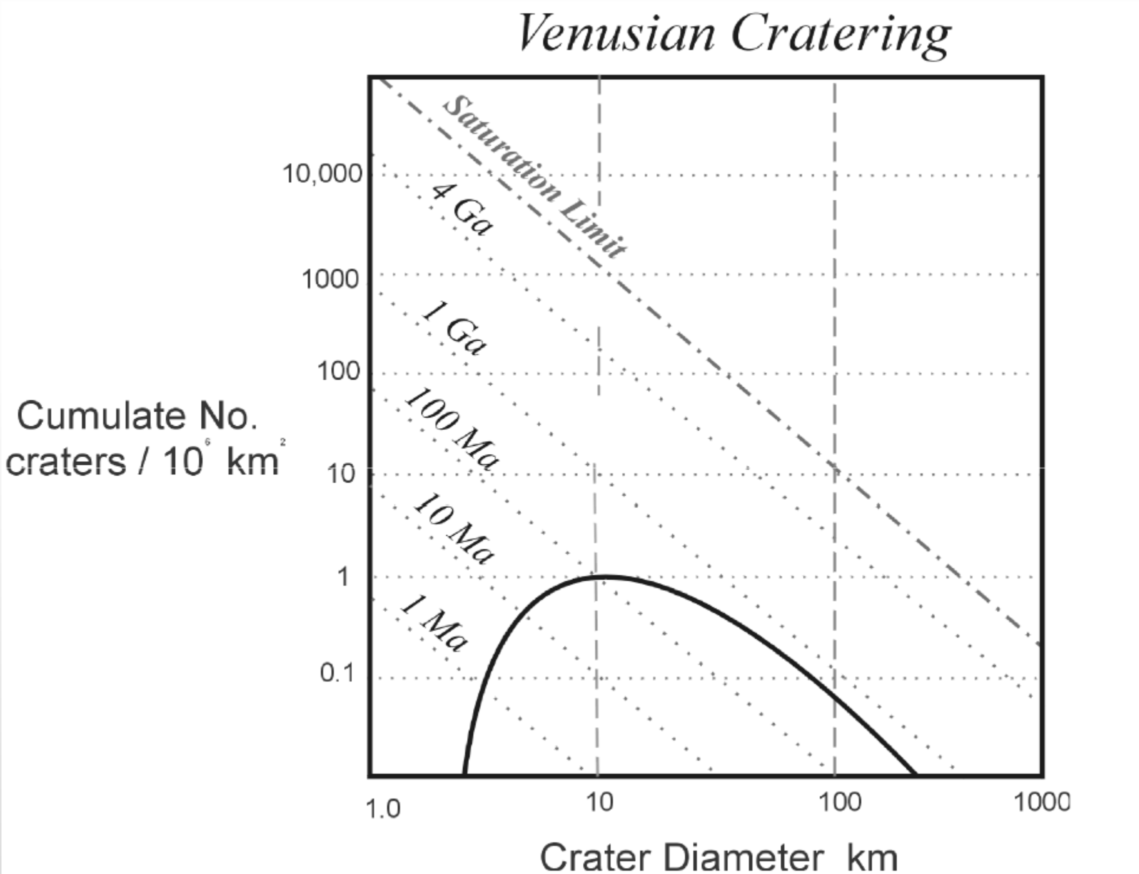


The age of Mercury's surface

(after D. Francis, Earth & Planetary Sciences, McGill)

The highlands of Mercury are old; like the Moon the **Caloris basin** is younger suggesting that it was formed by an impact only about 3 billion years ago.

Crater Density Age: Venus

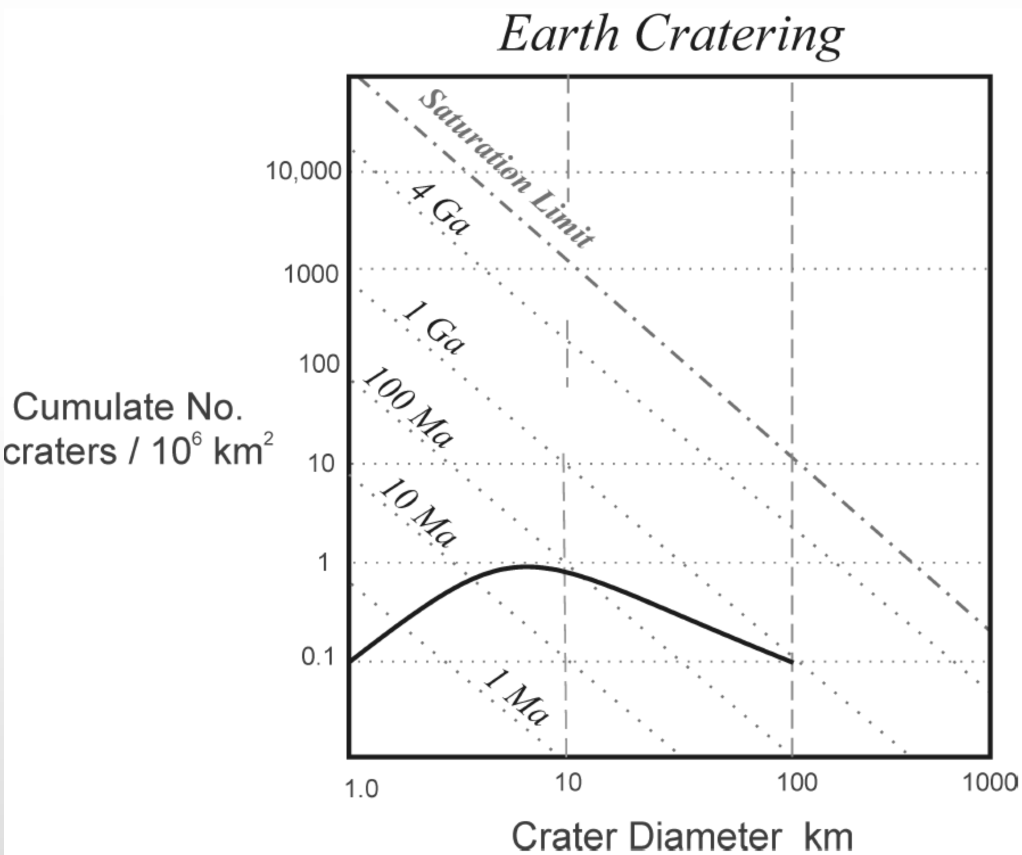


The age of Venus' surface

(after D. Francis, Earth & Planetary Sciences, McGill)

Venus' surface is young. The planet appears to have been completely resurfaced between 400My and 700My ago.
There are few small craters as the thick atmosphere of Venus so slows the small asteroids that they don't form significant craters on impact if they reach the surface.

Crater Density Age: Earth



The *apparent* age of Earth's' surface

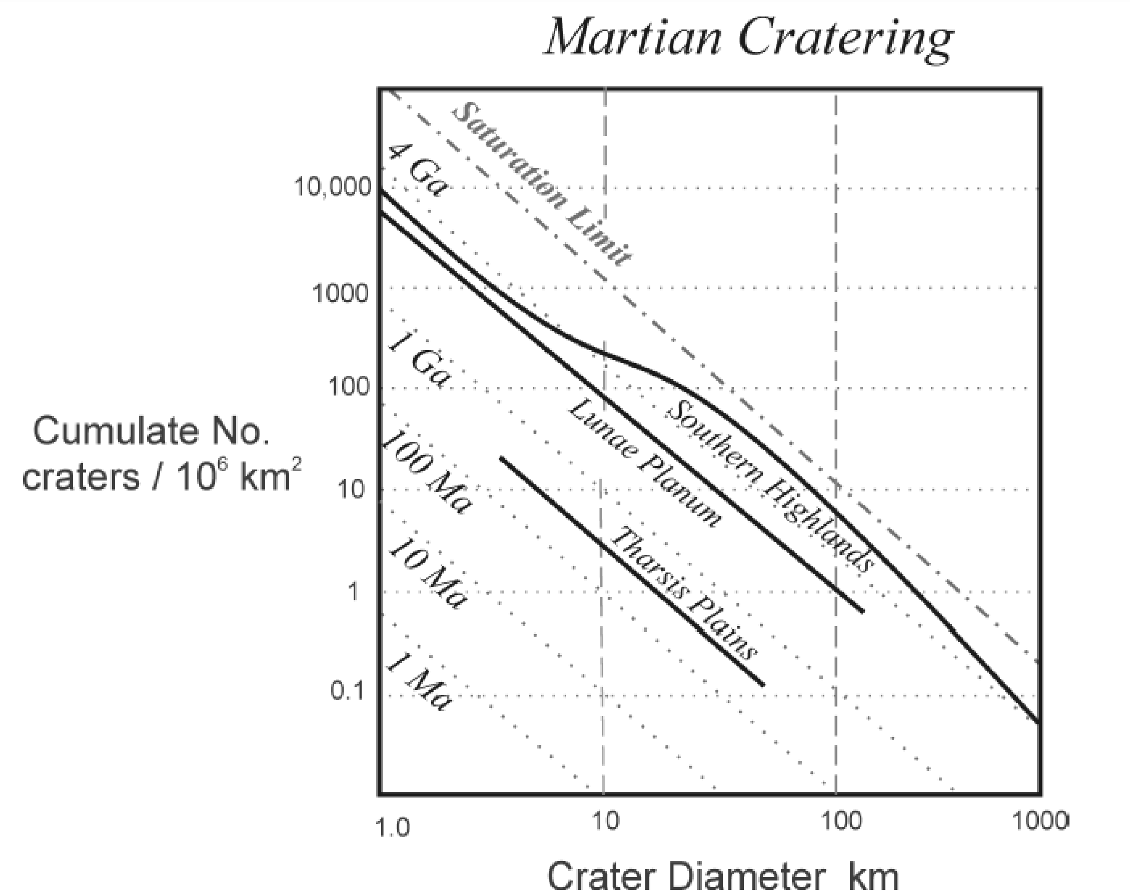
(after D. Francis, Earth & Planetary Sciences, McGill)

Only about 10% of Earth's surface is older than about 1 billion years and all of this is continental. Erosion has erased the history of cratering.

Small craters are few because our atmosphere slows small asteroids that, then, don't produce significant surface craters.

70% of Earth is ocean covered and not cratered.

Crater Density Age: Mars

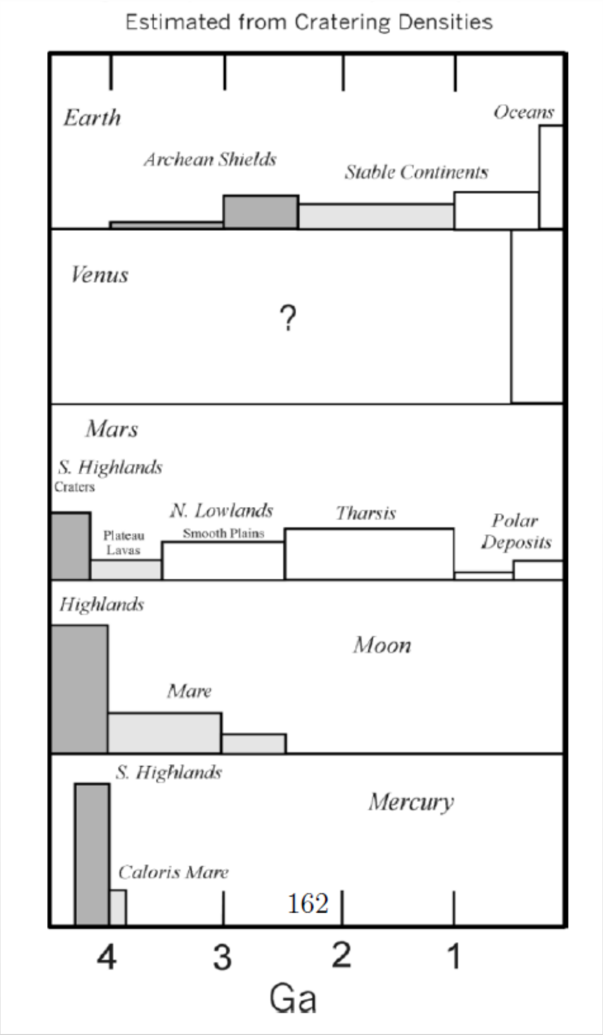


The cratering age of Mars' surface

(after D. Francis, Earth & Planetary Sciences, McGill)

Mars shows very old surface in the **Southern Highlands**. For small craters, though, their lower density suggests either a once-thick atmosphere or perhaps even a large areal ocean. The **Tharsis Plains** near the 4 large volcanoes is younger, about 1 billion years old or less, indicating that the volcanic erruptions were relatively recent.

Age of Planetary Surfaces



Earth shows a distribution of ages. Ocean basins (~70% of the surface ares) are all younger than 200 My.

Venus' entire surface is relatively young, about 400My to 700My old.

Mars shows surfaces with a range of ages spanning the whole time since formation of the Solar System.

The Moon's surface dates to the original formation of the Moon except for the Mare impact basins.

Mercury's surface is mostly very old though, seemingly, not dating to the age of the Solar System.

We, now, end our story … I thank you for your attention during these 12 weeks

The examination is scheduled during the April exam period. It will be online. I shall make myself available via Zoom to answer any questions you might have about the materials of the course or, in fact, any other questions you may have during the late afternoon of Sunday, April 15 (after 3:00PM). I shall remain available via email at olivia.jensen@mcgill.ca .

I wish you all the best in your exams in this and your other courses and success throughout your time at McGill.