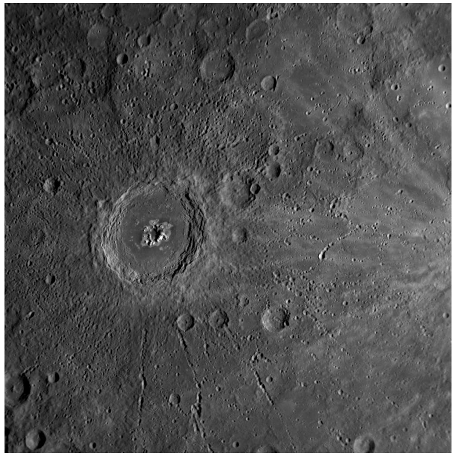
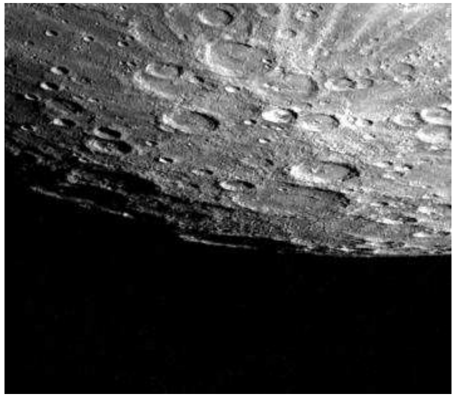
Planets: Surfaces and Interiors

Major landforms: Craters



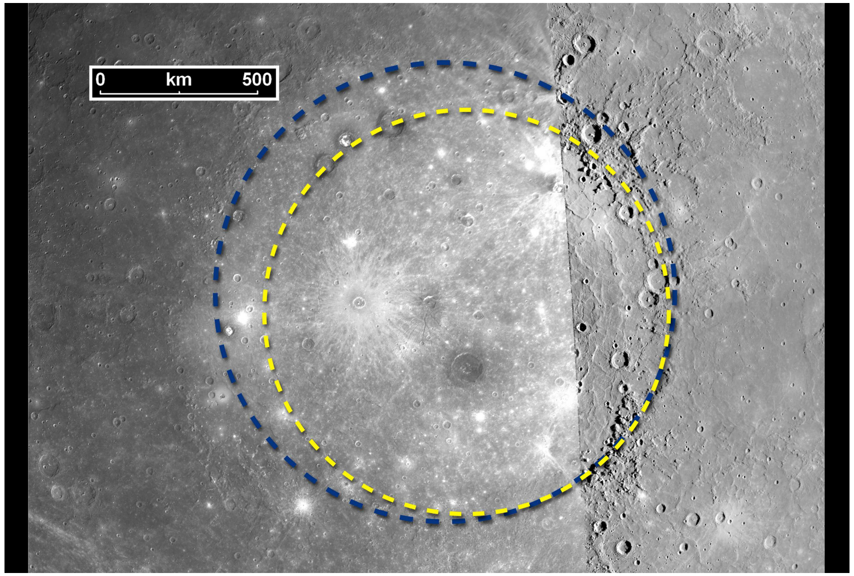
NASA/MESSENGER

Terraced craters, with central mountains.



S-Pole; NASA/JPL (Mariner)

50 km diam craters with rays (remains from impact)



NASA/MESSENGER

Mercury: Caloris Basin (1300 km diameter)

close to sub-solar point at perihelion $\Longrightarrow$ hot! ($T > 400^\circ$C on day, $T \sim -170^\circ$C during night)

Basin: result of *large* impact event

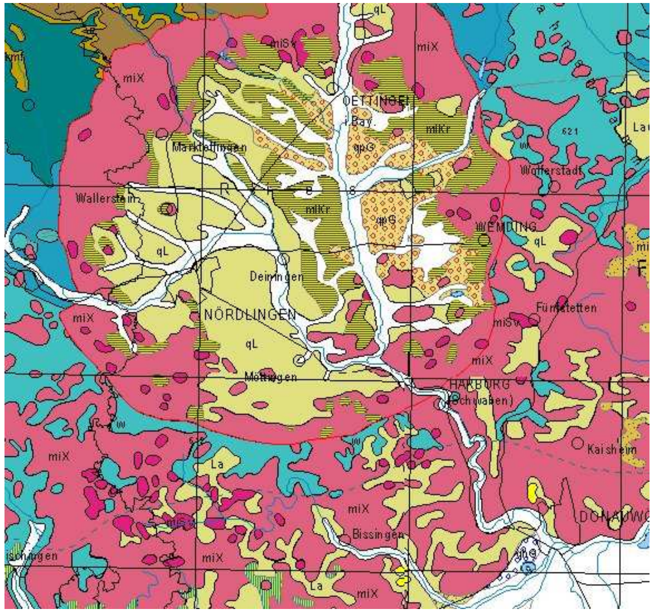
V. L. Sharpton

Earth: Wolf Creek Crater, Australia

Currently 172 confirmed impact structures on Earth

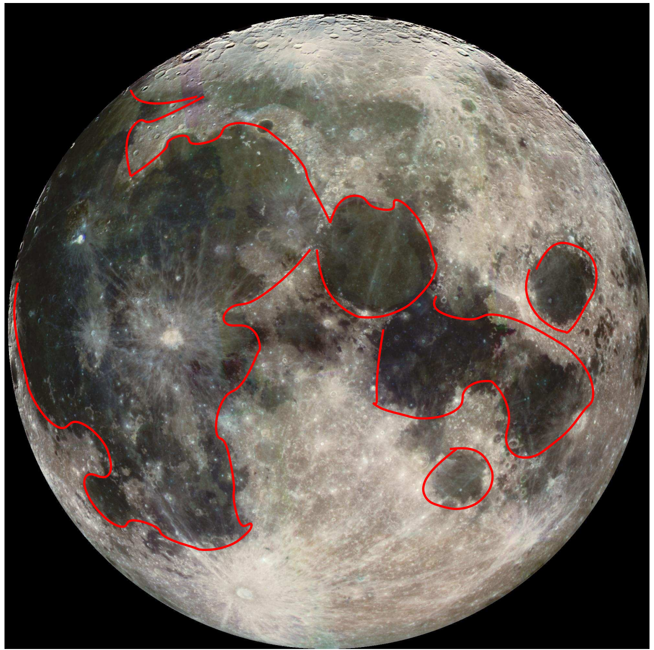
J. Wilms

Earth's Moon

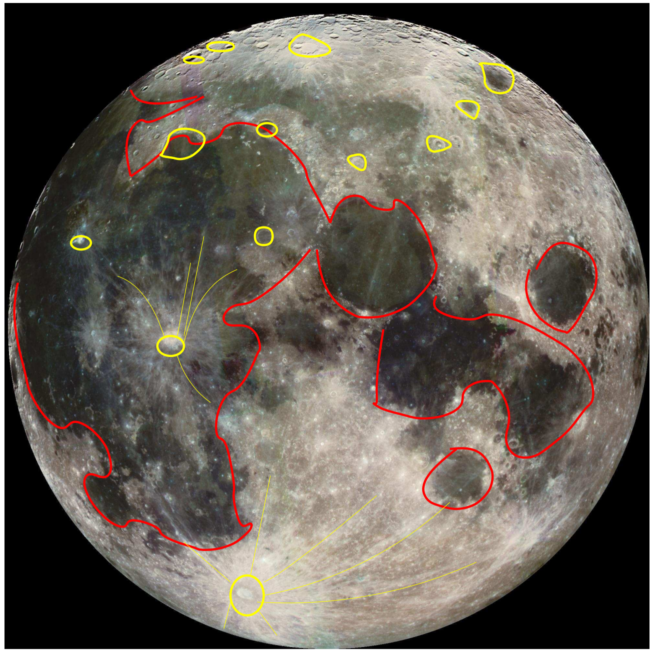


Earth: Nördlinger Ries

Impact crater identified by Shoemaker in 1960s



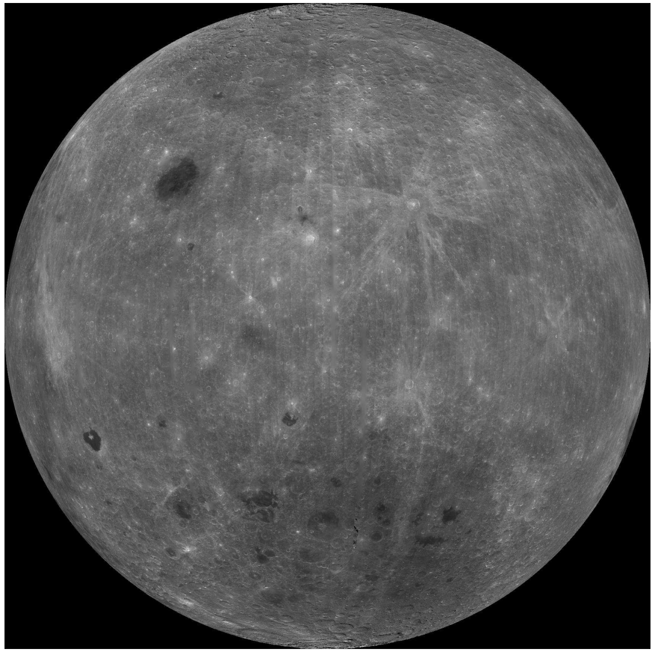
Earth's Moon : surface dominated by mariae (large, dark lava basins)



Earth’s Moon : surface dominated by **mariae** (large, dark lava basins) and **craters** (only most prominent shown).

Moon: Apollo 16, 1972 Apr, Descartes Highlands

Moon: Crater Copernicus

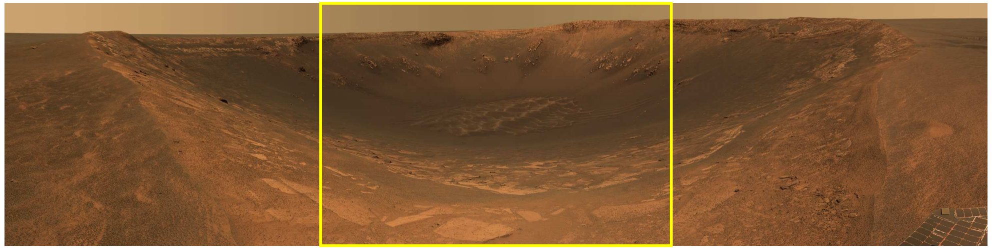


Far side of the Moon:

- few maria
- stronger relief (16 km roughness, compared to 5–6 km on near side)
- Aitkin basin near South pole: 12 km deep

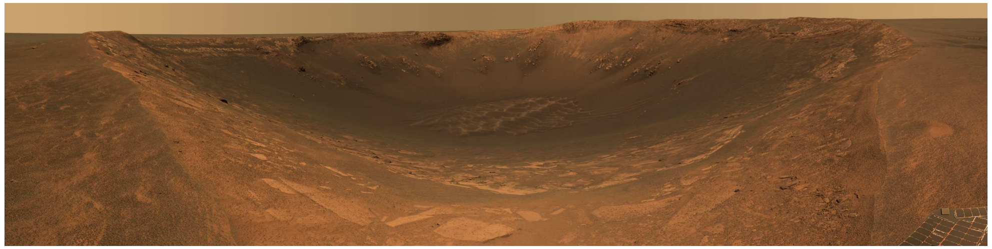
NASA/Clementine

Mars: Surface panorama, Exploration Rover “Opportunity” looks back to lander (2004 Feb 09)



NASA/JPL/Cornell

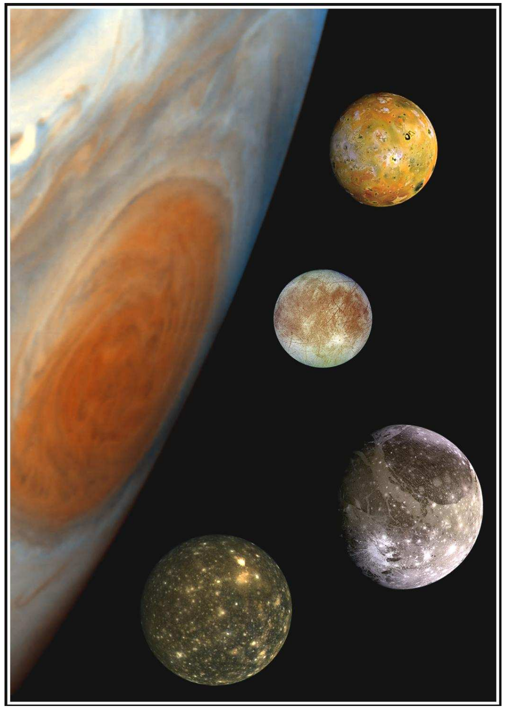
Mars: Crater Endurance



NASA/JPL/Cornell

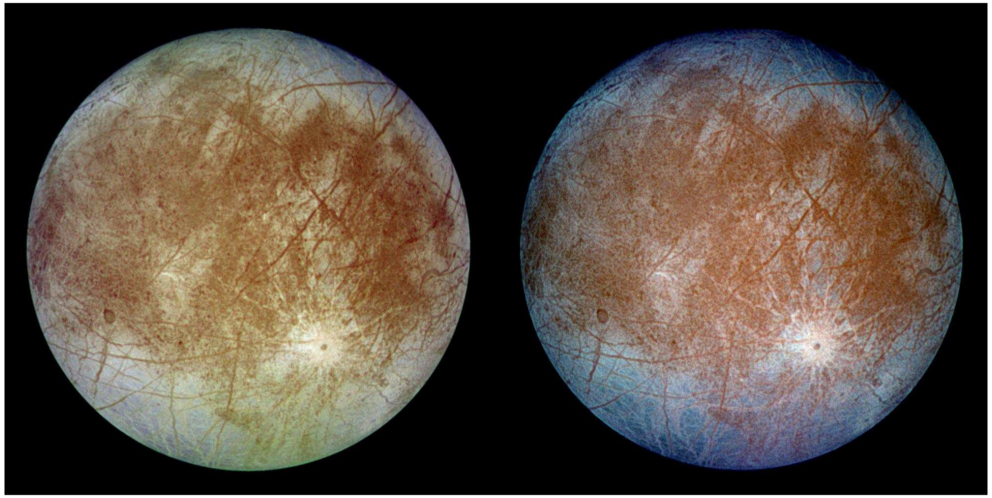
NASA/JPL/Cornell

Mars: Crater Endurance



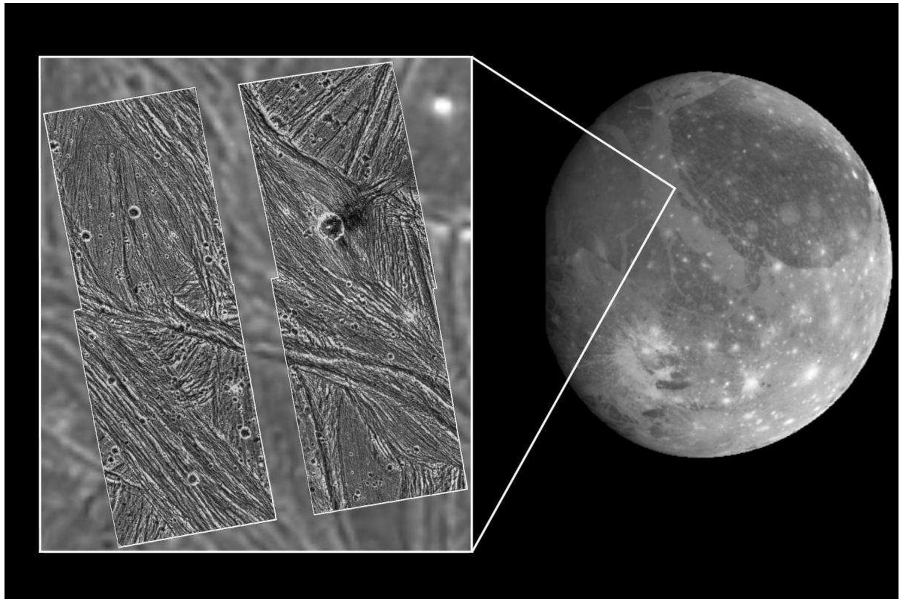
Montage of Jupiter and Galilean Moons: top to bottom: Io, Europa, Ganymede and Callisto.

(N.B.: All Galilean moons tidally locked to Jupiter – always same side is facing Jupiter)



NASA Galileo / DLR, 1996 September 7

Europa – icy surface with ridges (colors: different kinds of ice)
Radius: 1565 km (∼ Earth Moon)
possibility of water ocean below surface



NASA Galileo / DLR, inset: 120×110 km

Ganymede – icy surface, ice hills and valleys, craters
Radius: 2634 km (∼ Mercury!)

Callisto: “pock faced”,
mainly impact craters.
white: ice
dark: ice-poor material

Radius: 2406 km (similar
to Mercury!)

Impact Craters, I

Physics of impact cratering:

Kinetic energy:

$$E = \frac{1}{2}mv^2 = \frac{1}{2}\cdot\frac{4}{3}\pi r^3\rho v^2 = \frac{\pi d^3\rho v^2}{12}$$

Important numbers:

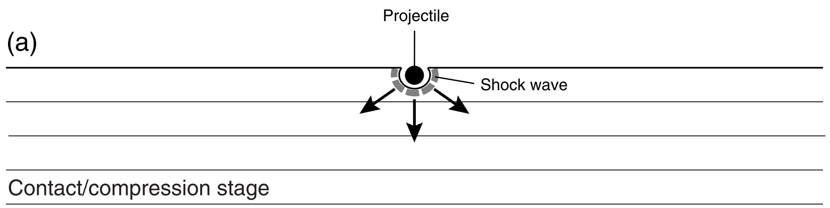
- Velocity of impact: several times orbital speed of planet
- Impacting body: rock or Fe, several meters to kilometers in size

Example:

E.g., $v = 10\,\mathrm{km\,s^{-1}}$, $d = 25\,\mathrm{m}$, $\rho = 7900\,\mathrm{kg\,m^{-3}}$
$\Longrightarrow E = 3\times 10^{15}\,\mathrm{J}$ ($\sim$1 Megaton of TNT)

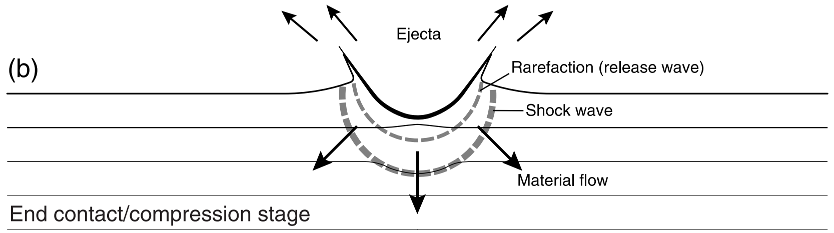
1 Megaton TNT is typical strength of US nuclear bombs [B-83 bomb]

Impact Craters, II



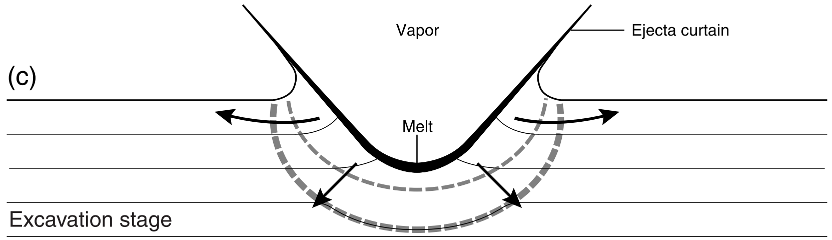
French, 1998, LPI Cont. 954

Impact Craters, III



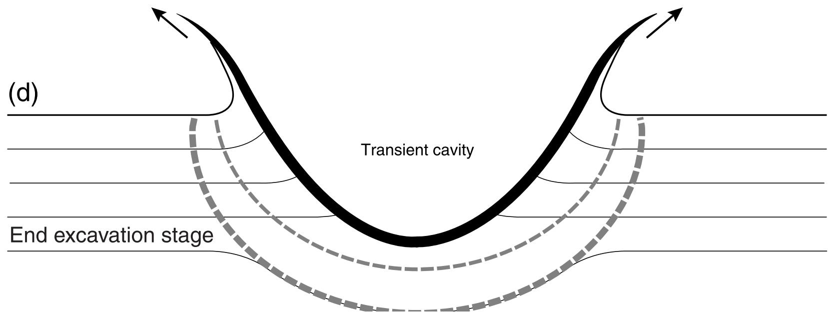
French, 1998, LPI Cont. 954

Impact Craters, IV



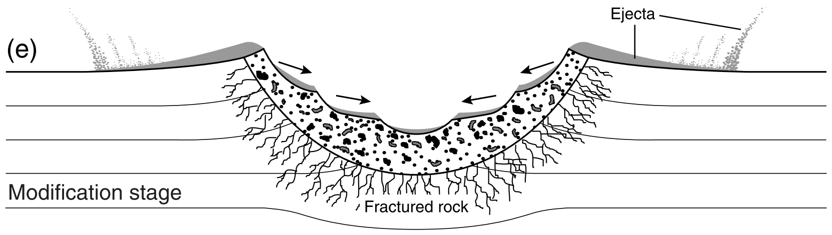
French, 1998, LPI Cont. 954

Impact Craters, V



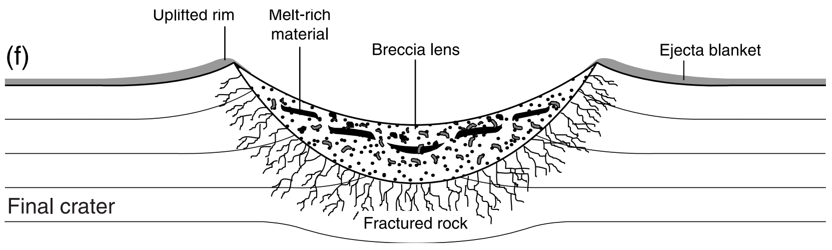
French, 1998, LPI Cont. 954

Impact Craters, VI

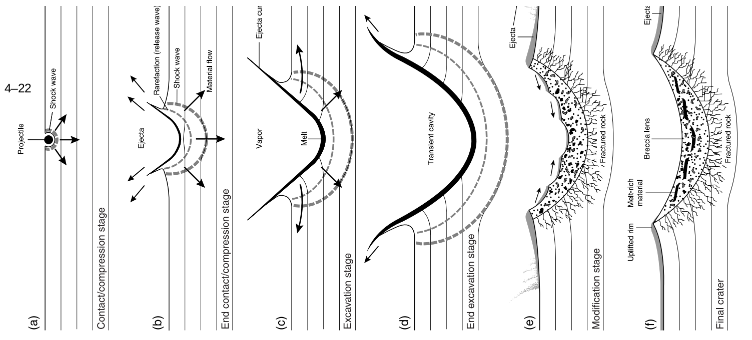


French, 1998, LPI Cont. 954

Impact Craters, VII

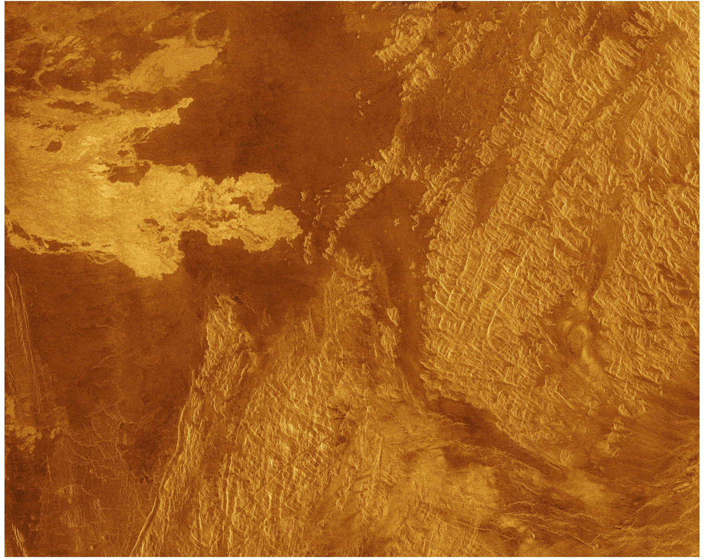


French, 1998, LPI Cont. 954



French, 1998, Traces of Catastrophe, LPI Contribution 954, p. 21

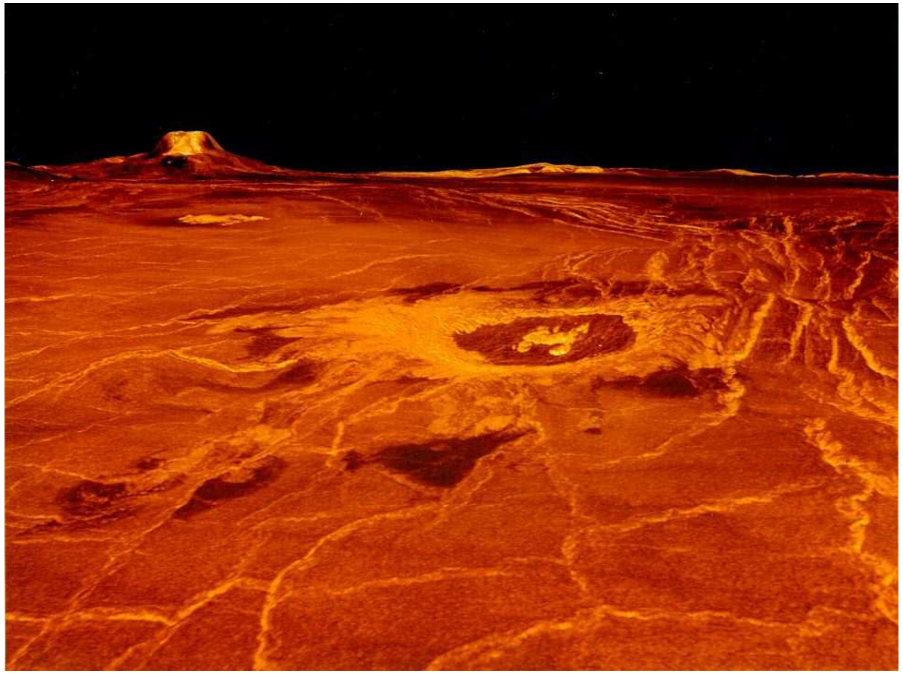
Venus



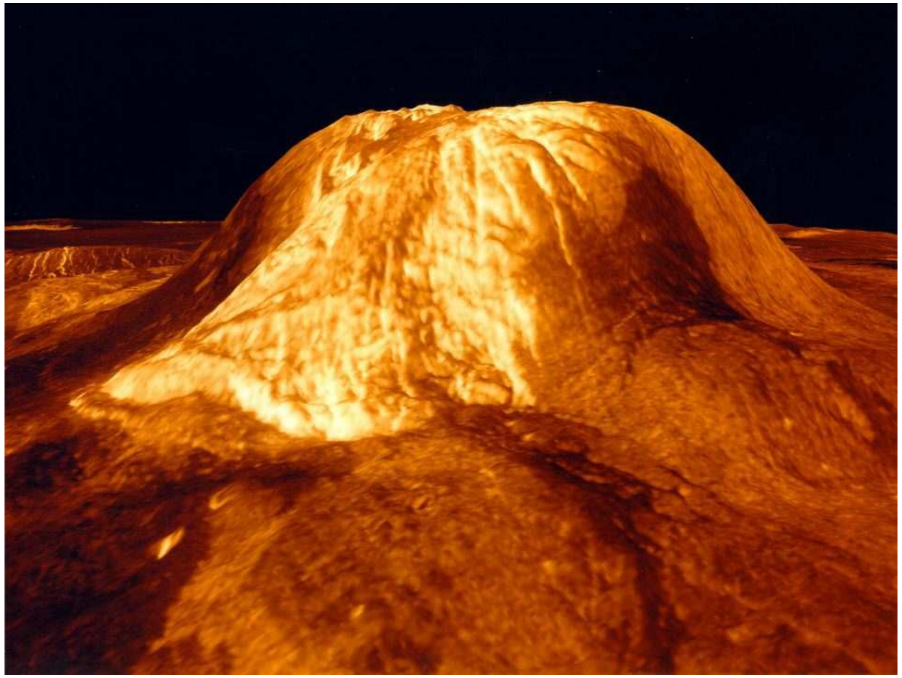
NASA, Magellan

$440 \times 350\,\mathrm{km}^2$ area in Eistla Regio, shows basic stratigraphy (sequence of geologic events): right half: old highlands, fractured structure ($\sim$15% of surface), left part: lowlands, younger area, origin in former volcanism?

Craters (note: strong erosion $\Longrightarrow$ fewer craters overall)



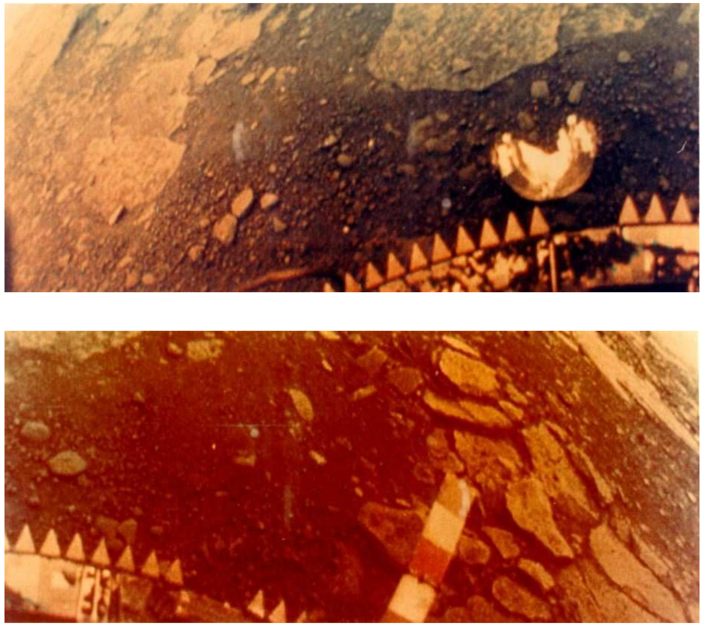
Eistla Regio; heights exagerrated by factor 22.5



Gula Mons; heights exagerrated by factor 22.5

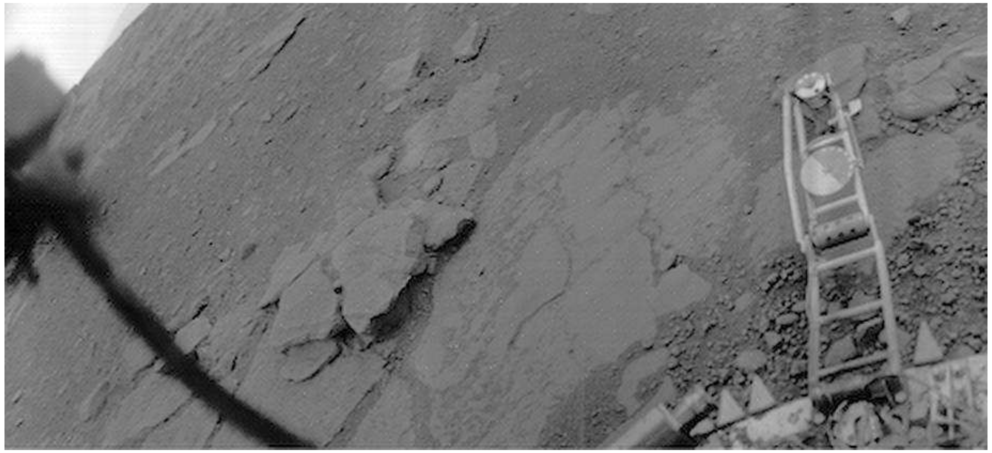
Gula Mons; real heights

Venus surface images:



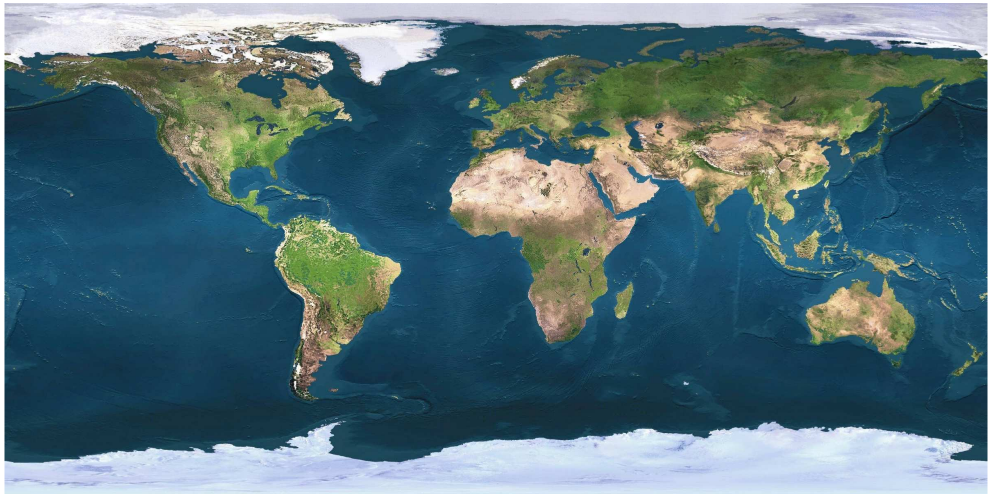
Venera 13 (3 March 1982): images from color TV camera

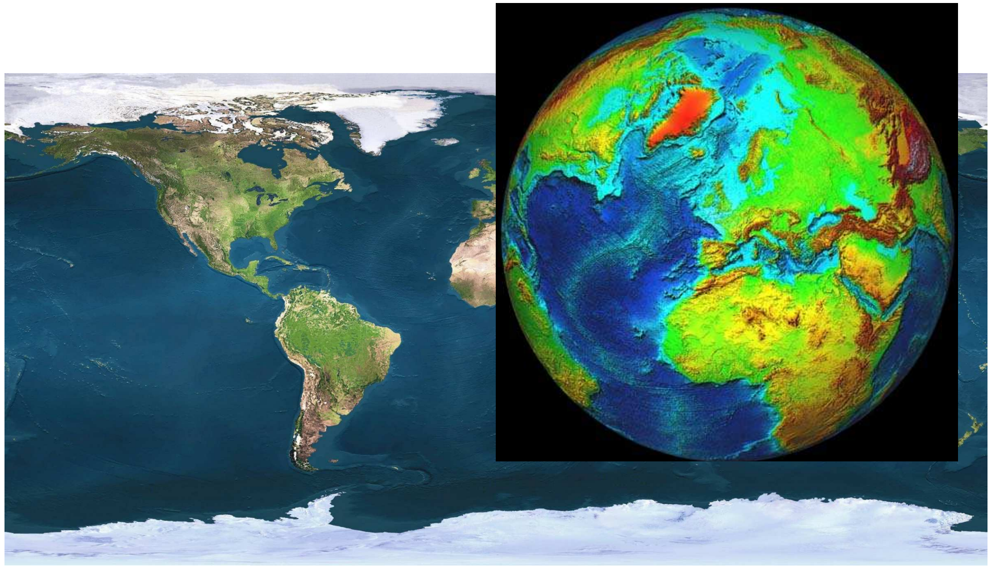
Venera 14 (5 May 1982)



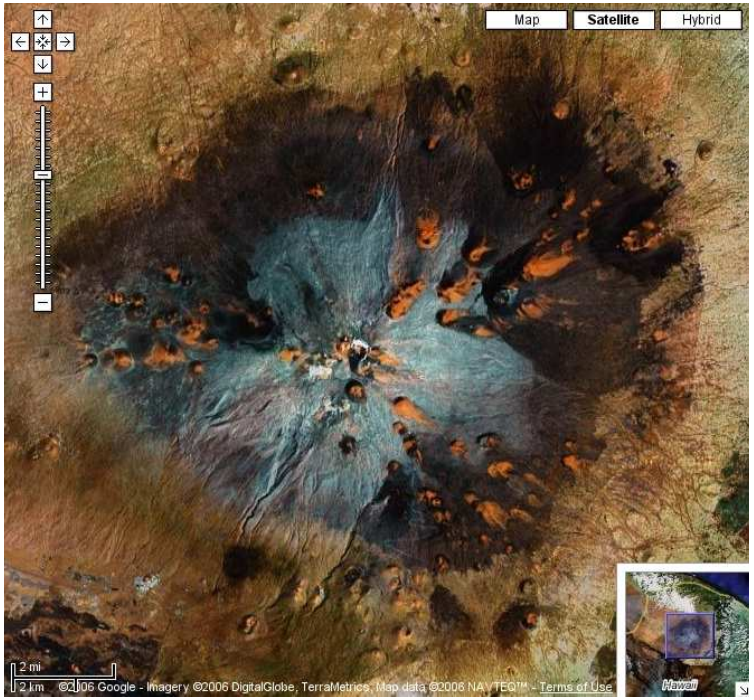
courtesy D.P. Mitchell

Venera 13 (3 March 1982): reanalysed image without camera distortion

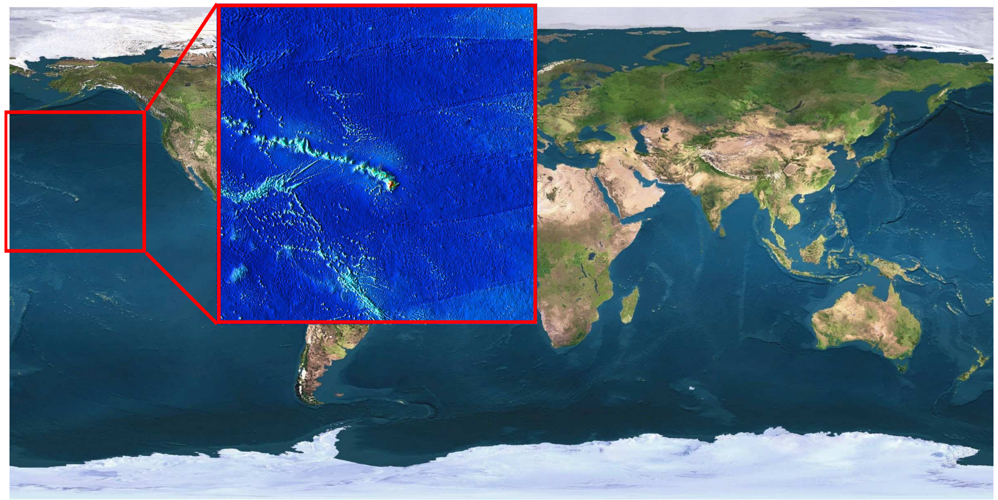




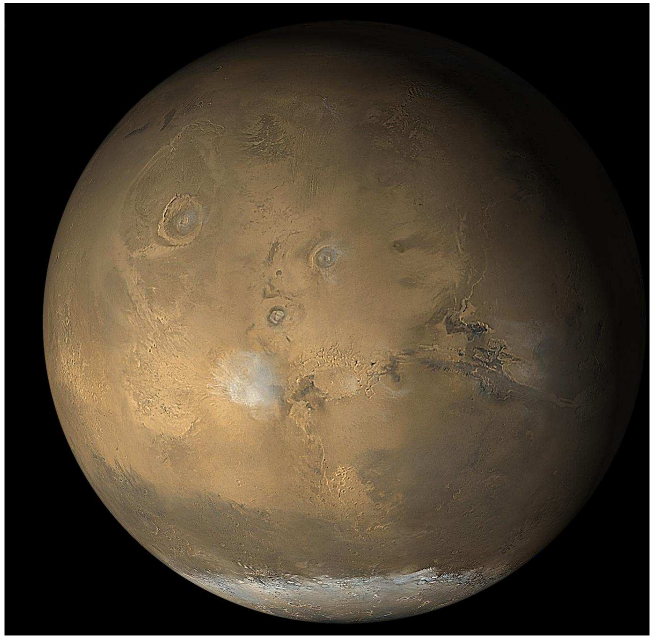
Evidence for plate tectonics (few craters!)



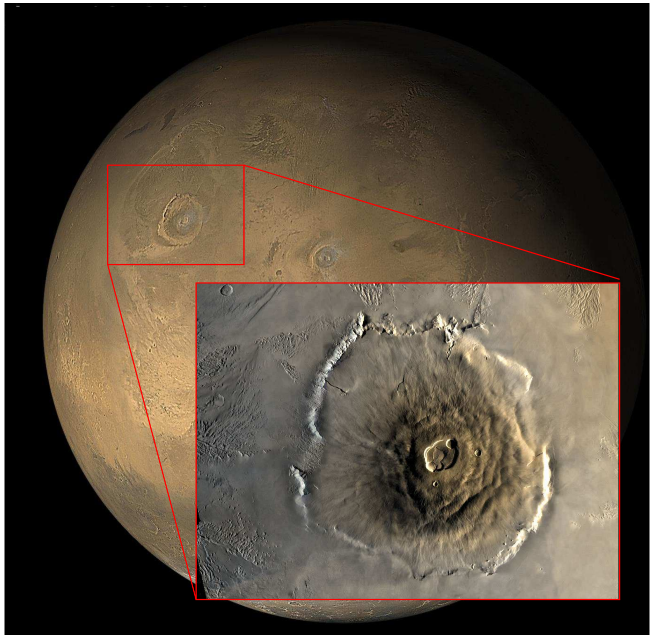
© Google



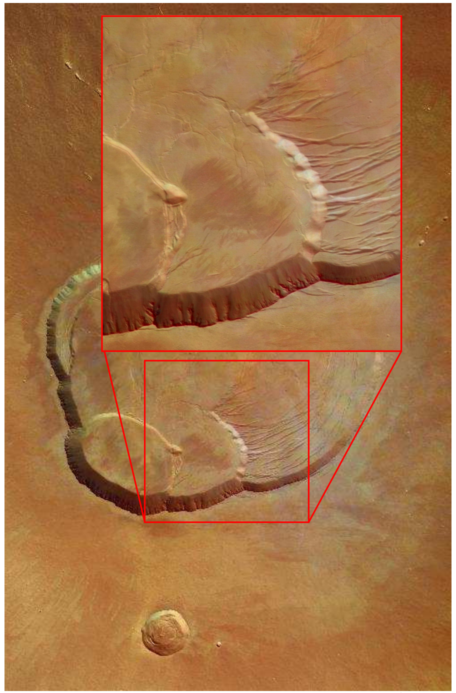
Evidence for plate tectonics (few craters!) , volcanism,. . .



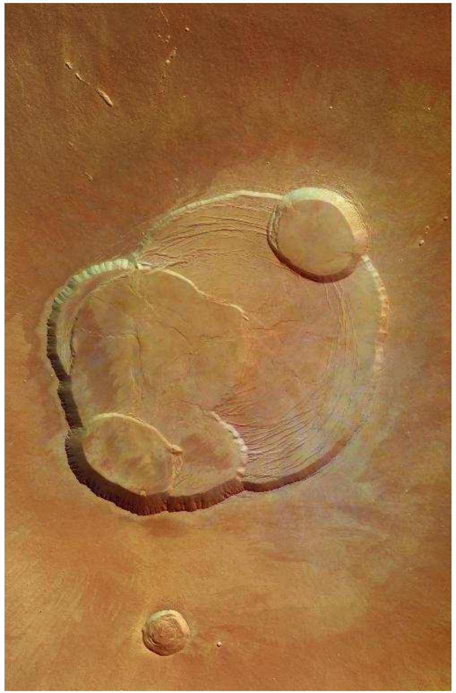
Mars: Tharsis vulcanos: Large shield vulcanos, now extinct
$\Longrightarrow$ no plate tectonics $\Longrightarrow$ Mars interior is colder than Earth.



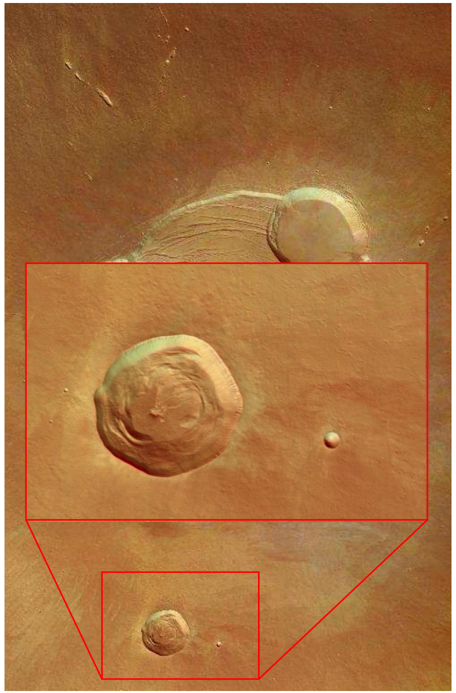
Olympus Mons: highest volcano in solar system
(25 km above surrounding plain; but slope only 2° to 5°).



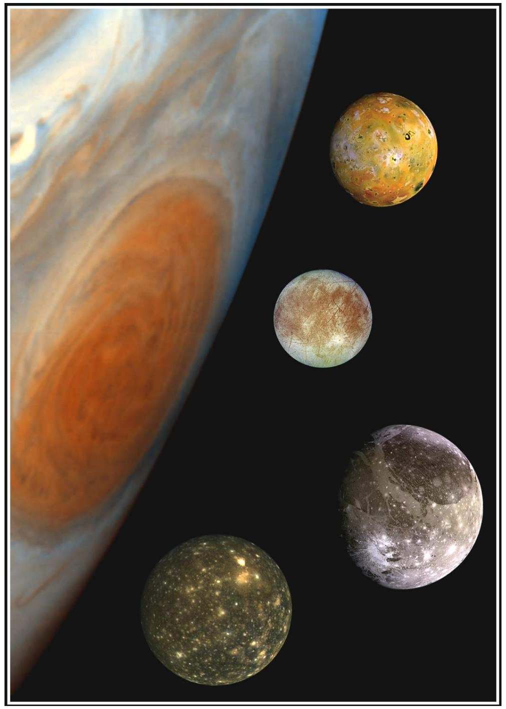
ESA/Mars Express, HRSC, 11.02.2004



ESA/Mars Express, HRSC, 11.02.2004

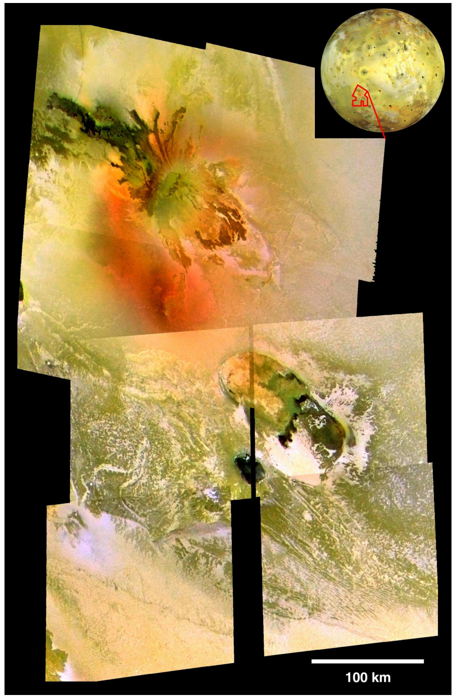


ESA/Mars Express, HRSC, 11.02.2004



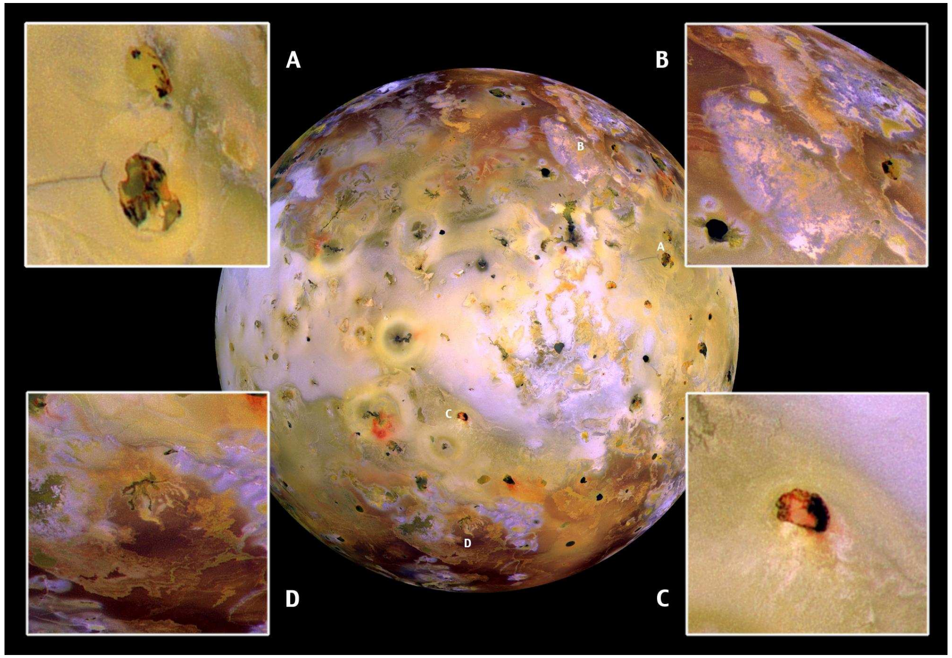
Montage of Jupiter and Galilean Moons: top to bottom: Io, Europa, Ganymede and Callisto.

(N.B.: All Galilean moons tidally locked to Jupiter – always same side is facing Jupiter)

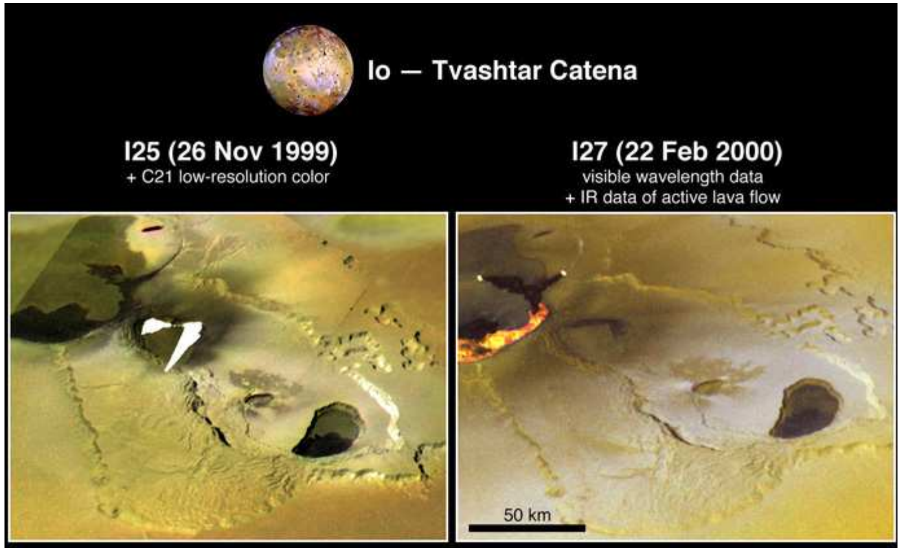


Active vulcanoes on Io (interior heated by tidal forces from Jupiter), color due to large contents of sulphur and sulphur oxides in lava.

Height of vulcanoes: 6 km or higher

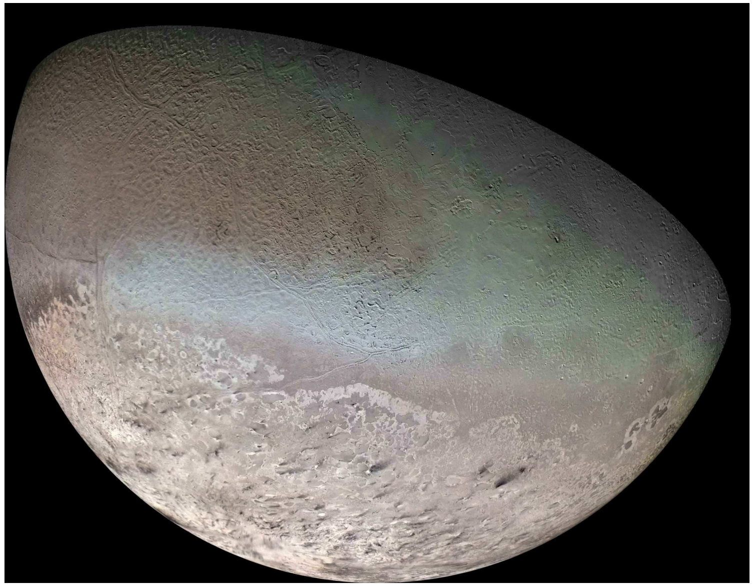


Jupiter's moon Io – the vulcano moon (Diam. 1821 km [Earth moon: 1738 km])



curtains of lava fountains [white: overexposed]
NASA Galileo, 1999 Nov 26

High temperature volcanism (2000 K; hotter than on Earth [1700 K]!)



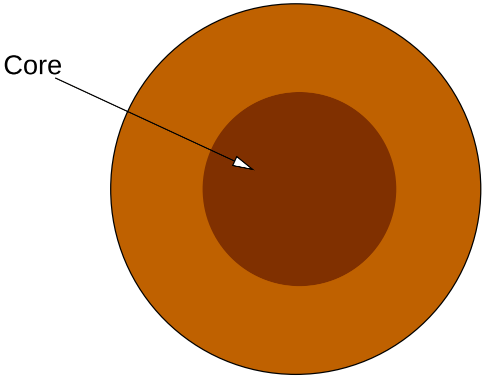
NASA/Voyager 2/Calvin J. Hamilton

Neptune's Moon Triton:

ice cap of frozen methane (freezing point 90 K) and frozen nitrogen (freezing point 60 K).

Few impact craters $\Longrightarrow$ young surface $\Longrightarrow$ volcanism (dark spots: nitrogen geysers with $T \sim 70\,\mathrm{K}$)

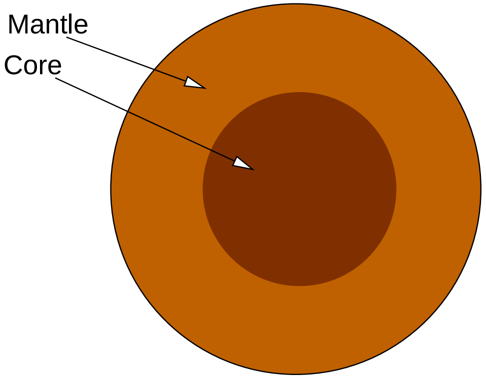
Interiors: Terrestrial Planets



Structure of terrestrial planets:

- Core: high-density material (Fe)

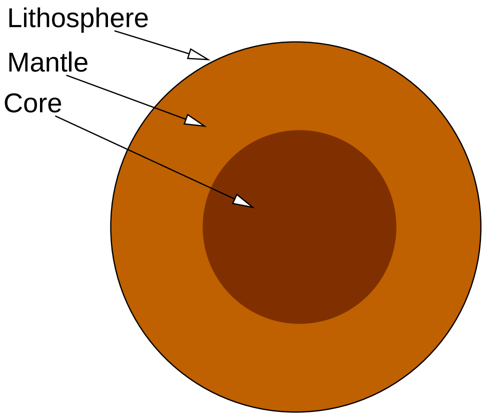
Interiors: Terrestrial Planets



Structure of terrestrial planets:

- Core: high-density material (Fe)
- Mantle: plastic materials, hot (e.g., Earth: molten rocks)

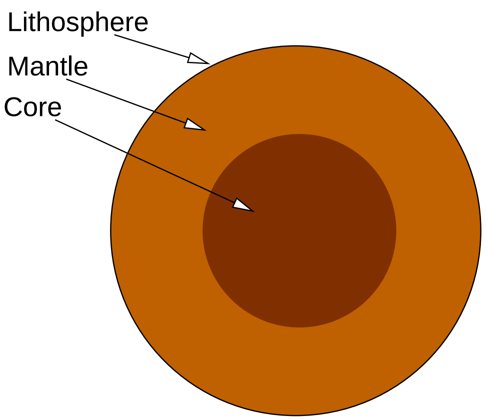
Interiors: Terrestrial Planets



Structure of terrestrial planets:

- Core: high-density material (Fe)
- Mantle: plastic materials, hot (e.g., Earth: molten rocks)
- Lithosphere: rigid material, e.g., Silicates

Interiors: Terrestrial Planets

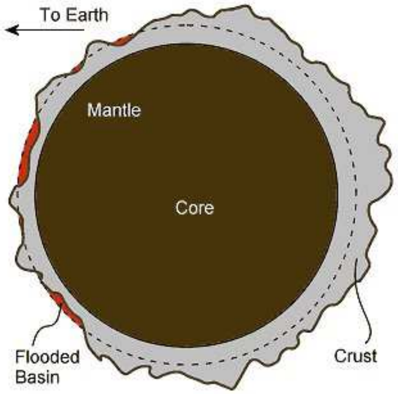


Structure of terrestrial planets:

- Core: high-density material (Fe)
- Mantle: plastic materials, hot (e.g., Earth: molten rocks)
- Lithosphere: rigid material, e.g., Silicates

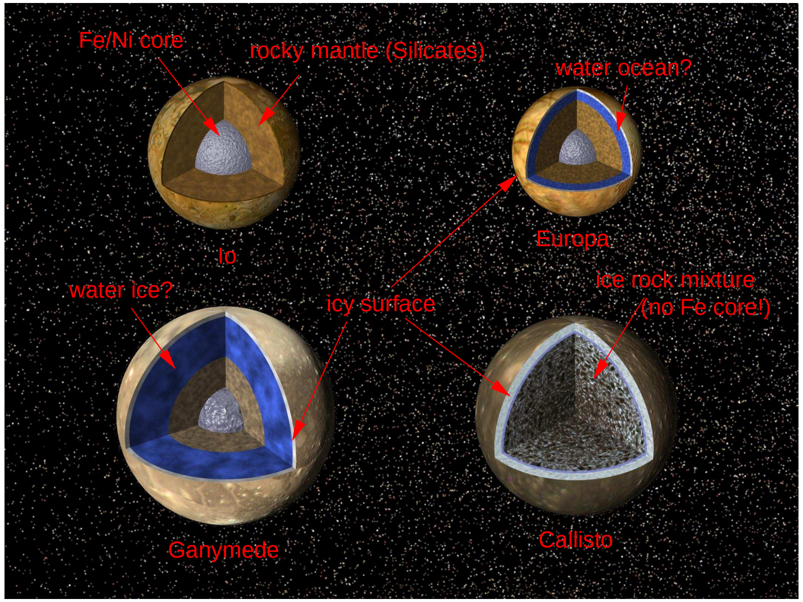
Knowledge of structure important for, e.g.,

- origin of magnetic fields (thought to be caused by molten core $\Longrightarrow$ currents $\Longrightarrow$ B-field ("dynamo"). Details unknown).
- atmospheric composition (molten mantle $\Longrightarrow$ volcanism $\Longrightarrow$ CO_2, CH_4,. . .)

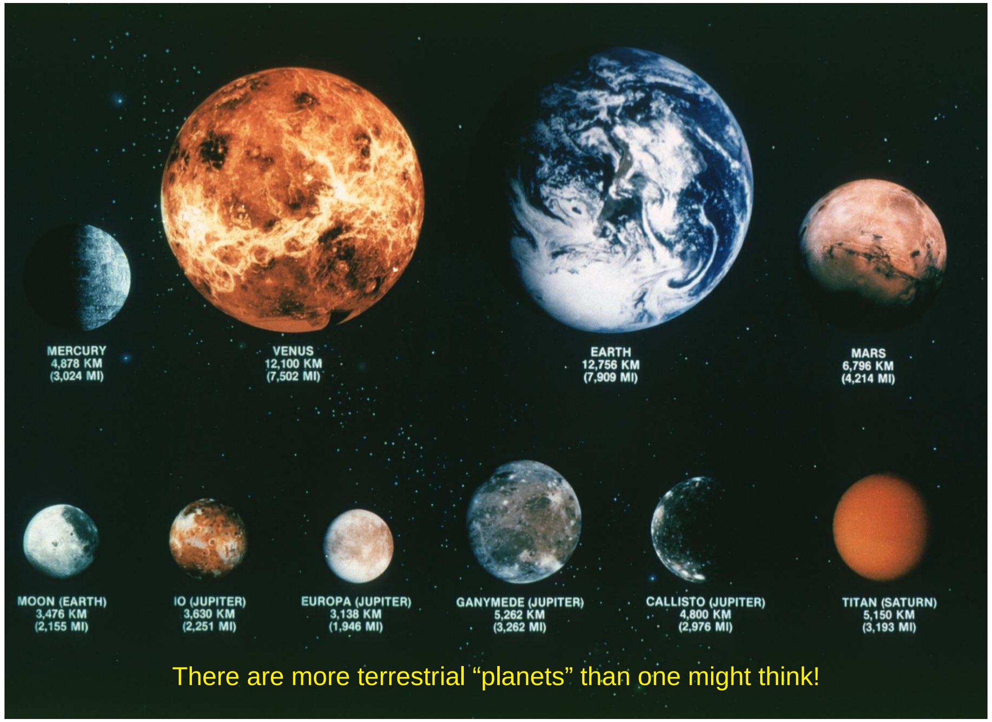


Structure of the Moon:

- gravimetric measurements
- seismometry from Apollo 12, 14, 15 & 16: mild moonquakes from 800–1000 km depth; generated by tidal forces?
- center of mass off-set by 2 km from center of sphere
- crust much thicker on the far side (100 km) than on the near side (60 km)
- iron core must be small ($<$400 km) if present (no magnetic field detected)



Structure of Jupiter's Galilean Moons similar to terrestrial planets
(but some also have very thick ice layer on top)



There are more terrestrial "planets" than one might think!

Structure: Gas Giants

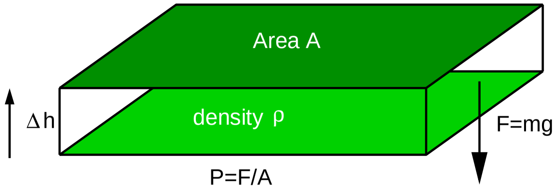
In general, gas giants have very different properties from terrestrial planets:

- average density low, e.g.,
  - Jupiter: $\langle\rho\rangle \sim 1.3\,\mathrm{g\,cm^{-3}}$
  - Saturn: $\langle\rho\rangle \sim 0.7\,\mathrm{g\,cm^{-3}}$

  (compare to terrestrial planets: $\langle\rho\rangle \sim 5.5\,\mathrm{g\,cm^{-3}}$; water has $\rho = 1\,\mathrm{g\,cm^{-3}}$).
- elemental composition similar to stars (by mass):
  - 75% H
  - 24% He
  - 1% rest (“metals”)

$\Longrightarrow$ expect fundamentally different internal structure!

Hydrostatic Equilibrium



Gas giants: structure defined by equilibrium between gravitation (downwards) and gas pressure (upwards): hydrostatic equilibrium.

Force excerted by material with density ρ sitting on top of an area A:

$$F = mg = \rho V g = Ah\rho g \quad \text{where} \quad g = \frac{GM(r)}{r^2} \quad \Longrightarrow \quad P = \frac{F}{A} = \rho h g \qquad (4.1)$$

where $M(r)$: mass contained within radius r.

Change in pressure between two points separated in radius by Δr:

$$\Delta P = \rho(r)g(r)(h+\Delta r) - \rho(r)g(r)h \quad \xrightarrow{\lim_{\Delta r\to 0}} \quad \frac{\mathrm{d}P}{\mathrm{d}r} = -\rho(r)\frac{GM(r)}{r^2} \qquad (4.2)$$

This is the equation of hydrostatic equilibrium.

Structure: Gas Giants, I

Structure of a gas giant from equation of hydrostatic equilibrium:

$$\frac{\mathrm{d}P}{\mathrm{d}r} = -\rho(r)\frac{GM(r)}{r^2} \qquad (4.2)$$

To solve, need $\rho(r)$, $M(r)$ $\Longrightarrow$ complicated, but doable if properties of material are known.

To guesstimate the central pressure, one can show for a planet of radius R (see writeup):

$$P_{\text{central}} = \frac{2\pi}{3}G\langle\rho\rangle^2 R^2 \qquad (4.3)$$

Plug in numbers for Jupiter: $R = 70000\,\mathrm{km}$, $\langle\rho\rangle = 1.3\,\mathrm{g\,cm^{-3}}$, get $P_{\text{central}} = 1.2 \times 10^{12}\,\mathrm{Pa}$ (10× Earth).

At this pressure: existence of metallic hydrogen (i.e., electrons can move freely around).

More detailed computations: metallic hydrogen from 14000–45000 km away from center

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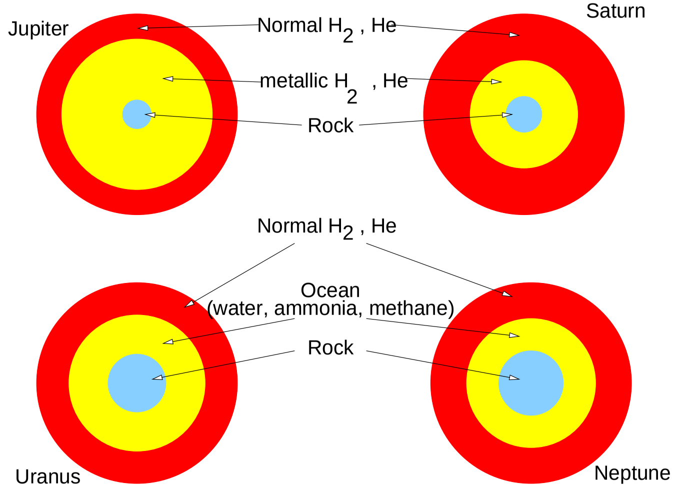
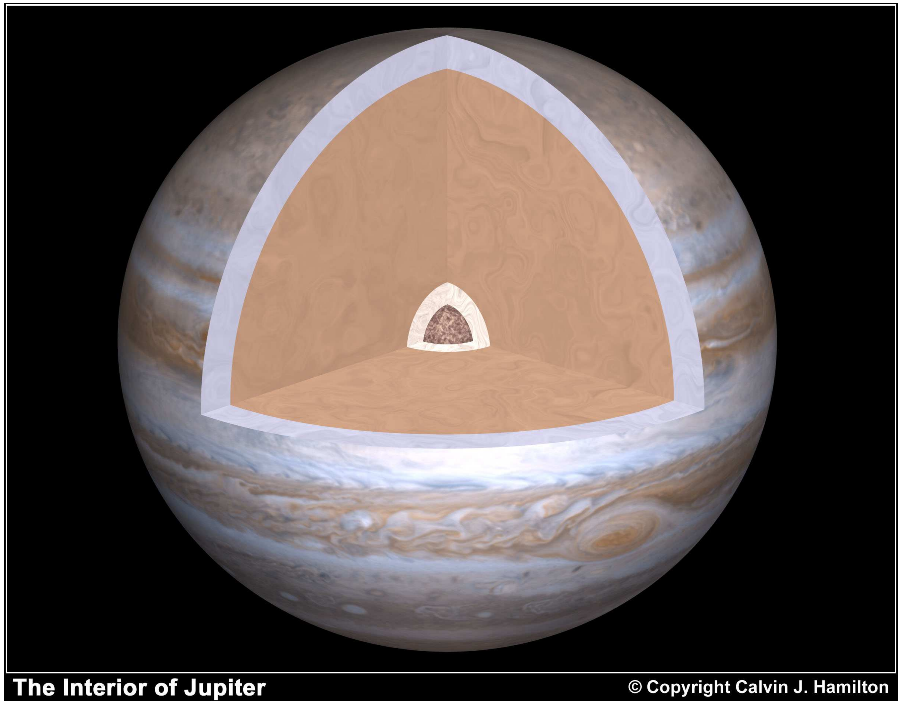
Structure: Gas Giants, II



Note: relative sizes of planets not to scale! Also rotational flattening not taken into account.



The Interior of Jupiter

© Copyright Calvin J. Hamilton

To obtain information on the pressure structure of any gravitationally supported static body we can use the *concept of hydrostatic equilibrium*, which we will encounter also later when looking at the structure of atmospheres and at the end of Intro I, when we will be looking at the structure of stars. As we have seen before, the radial pressure gradient is given by

$$\frac{\mathrm{d}P}{\mathrm{d}r} = -\rho(r)g(r) \tag{4.4}$$

here, r is now the radial distance from the planetary centre. In contrast to atmospheres, the acceleration g depends on the position, $g = g(r)$. It is easy to show that

$$g(r) = \frac{GM(r)}{r^2} \tag{4.5}$$

where $M(r)$ is the mass of the planet contained within a radius r:

$$M(r) = \int_0^r 4\pi\rho(r)r^2\mathrm{d}r \tag{4.6}$$

(interpretation: integrate over onion shells of thickness dr and density $\rho(r)$; the mass in each of these shells is $4\pi\rho(r)\mathrm{d}r$, summing over all onion shells gives the above answer).

To solve the equation of the hydrostatic equilibrium one needs to know the equation of state, i.e., the pressure as a function of the parameters of the material. Unfortunately, this equation of state is generally much more complicated than for gases (where $P = nkT$) and often only roughly known. One can estimate, however, the order of magnitude for the pressure within a planet. In order to do so, we assume that the density is the same throughout the planet, and that it equals the planet's average density $\rho(r) = \langle\rho\rangle =$ const.. This is o.k. to an order of magnitude. Under this assumption,

$$M(r) = (4/3)\pi r^3\langle\rho\rangle \tag{4.7}$$

such that the equation of hydrostatic equilibrium reads

$$\frac{\mathrm{d}P}{\mathrm{d}r} = -\langle\rho\rangle^2 G(4/3)\pi r \tag{4.8}$$

Differential equations looking like this are called separable. They can be solved “separation of variables”, as we already did when computing the structure of an isothermal atmosphere.

First integrate both sides of the equation from $r = 0$ to the surface of the planet at $r = R$:

$$\int_0^R \frac{\mathrm{d}P}{\mathrm{d}r}\,\mathrm{d}r = -\int_0^R \langle\rho\rangle^2 G(4/3)\pi r\,\mathrm{d}r \tag{4.9}$$

To integrate the left hand side of the equation, substitute $r \longrightarrow P(r)$ where $P(r)$ is an unknown function (the pressure as a function of radius r). Luckily enough, we only need to know its values at $r = 0$ and $r = R$ (the “boundary conditions”). By definition of the surface of the planet, the pressure at $r = R$ will be $P(R) = 0$ to very good

accuracy, while the pressure at $r = 0$ is the (unknown) central pressure, $P(0) = P_\mathrm{c}$. Therefore

$$\int_0^R \frac{\mathrm{d}P}{\mathrm{d}r}\,\mathrm{d}r = P(R) - P(0) = -P(0) =: -P_\mathrm{c} \tag{4.10}$$

The right hand side of the equation is easily found as well:

$$-\int_0^R \langle\rho\rangle^2 G(4/3)\pi r\,\mathrm{d}r = -\langle\rho\rangle^2(4\pi/3)G\int_0^R r\,\mathrm{d}r = -\langle\rho\rangle^2(4\pi/3)GR^2/2 = -\frac{2\pi}{3}\langle\rho\rangle^2 R^2 \tag{4.11}$$

such that

$$P_\mathrm{c} = \frac{2\pi}{3}\langle\rho\rangle^2 R^2 \tag{4.12}$$

As a rule of thumb, this formula gives central pressures that are correct to better than a factor of 10 compared to the detailed theory.