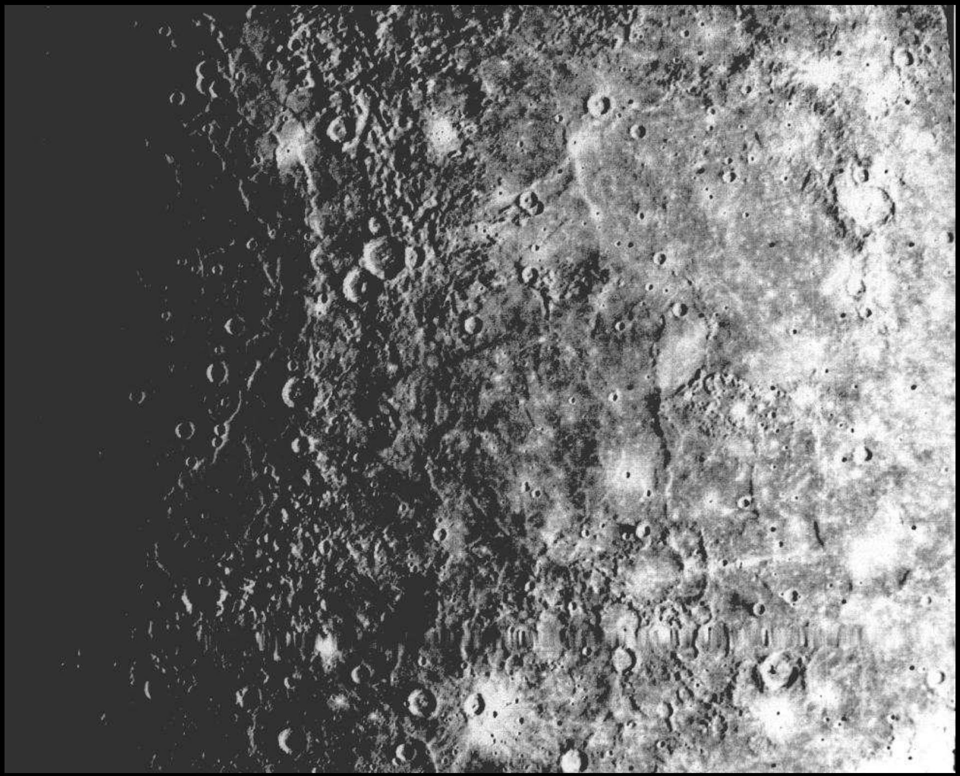
Mercury:



Caloris Basin (1300 km diameter)

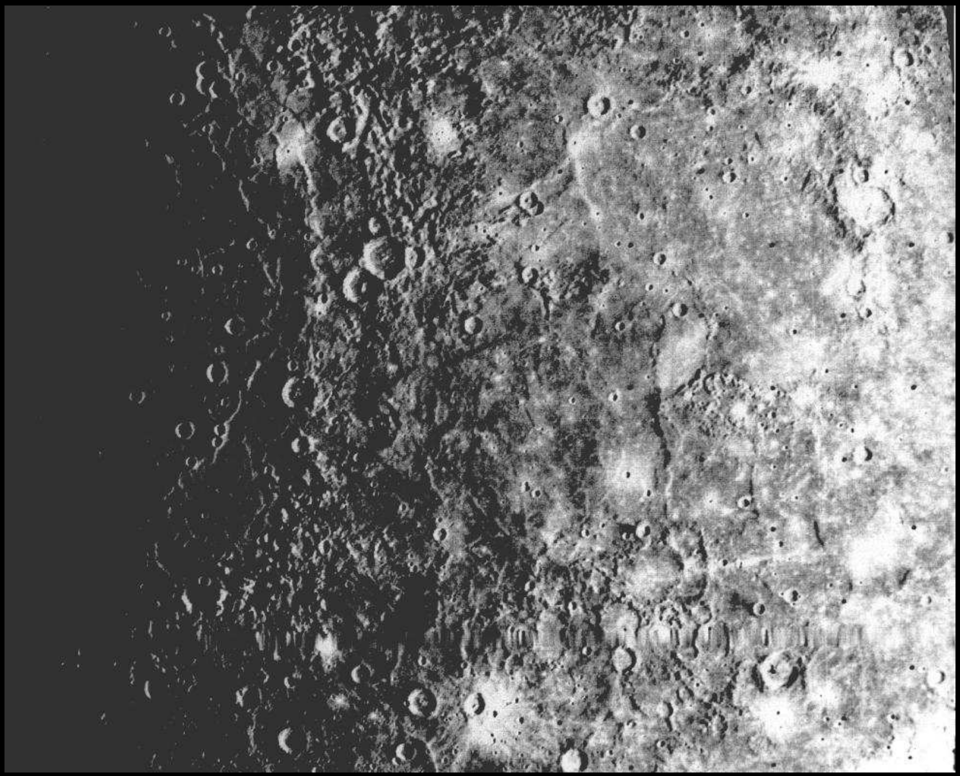
close to sub-solar point at perihelion

$\Longrightarrow$ hot ($T > 400^\circ$C on day,

$T \sim -170^\circ$C during night)

result of *large* impact event

Mercury:

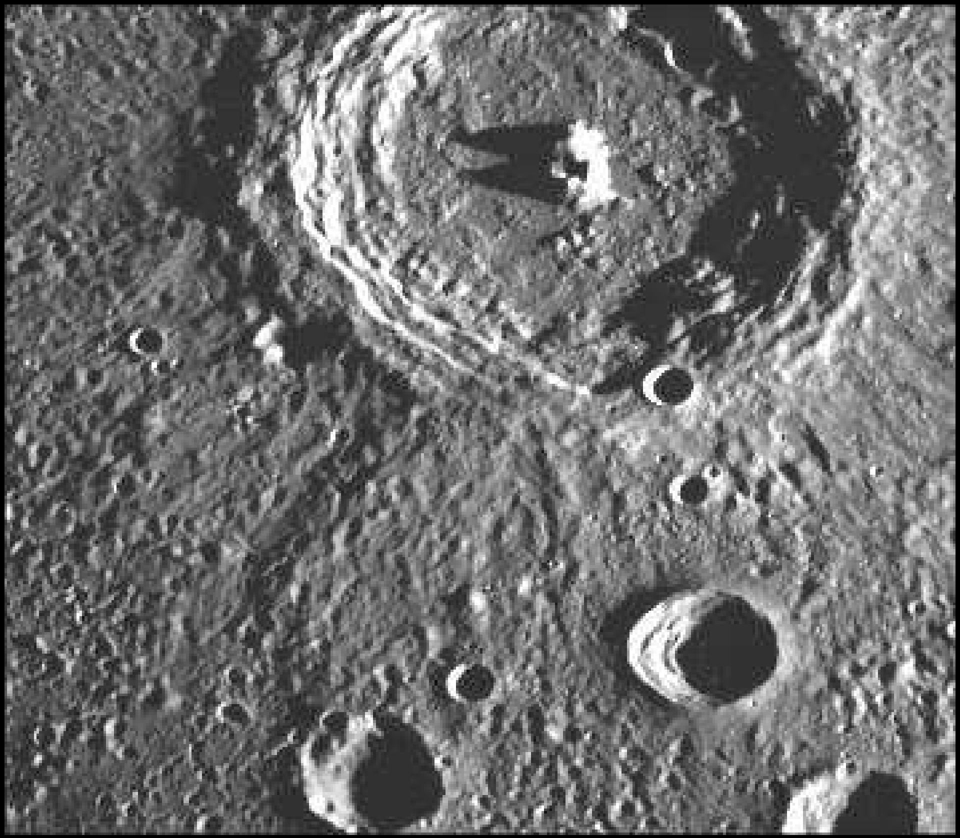


Caloris Basin (1300 km diameter)
close to sub-solar point at perihelion
$\Longrightarrow$ hot ($T > 400^\circ$C on day,
$T \sim -170^\circ$C during night)
result of *large* impact event

Robinson, NWU / NASA

Hilly/lineated terrain antipodal to Caloris (120 km across)
$\Longrightarrow$ effect of shock from Caloris impact.

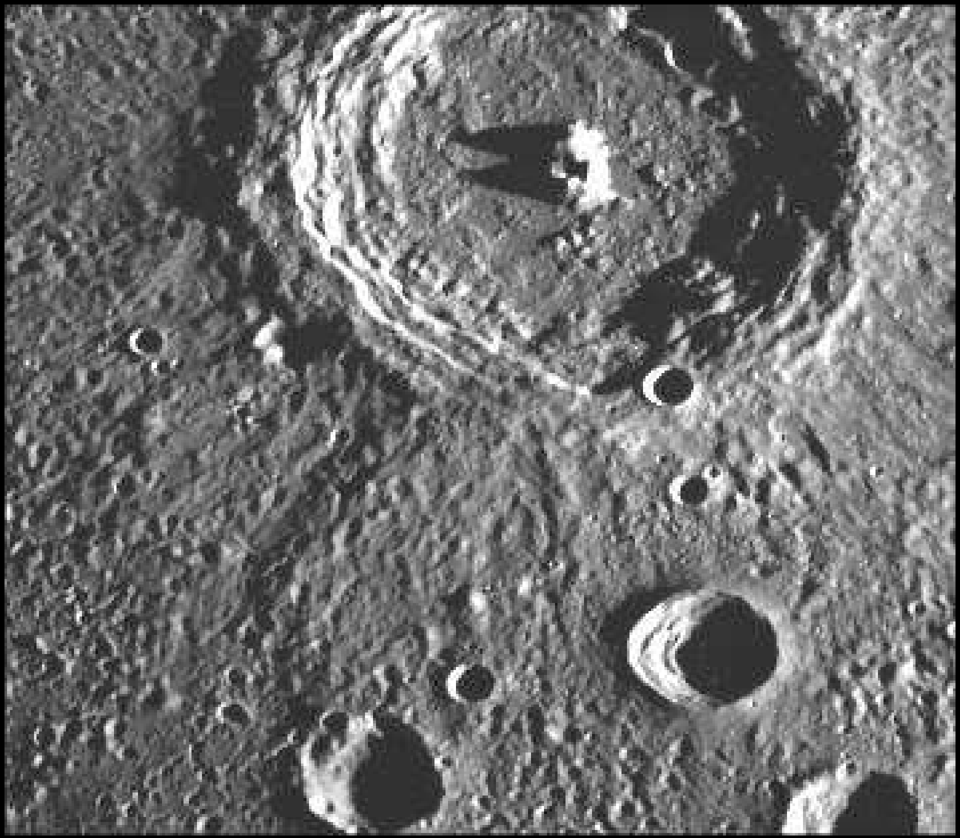
Major landforms: Craters



NASA/JPL

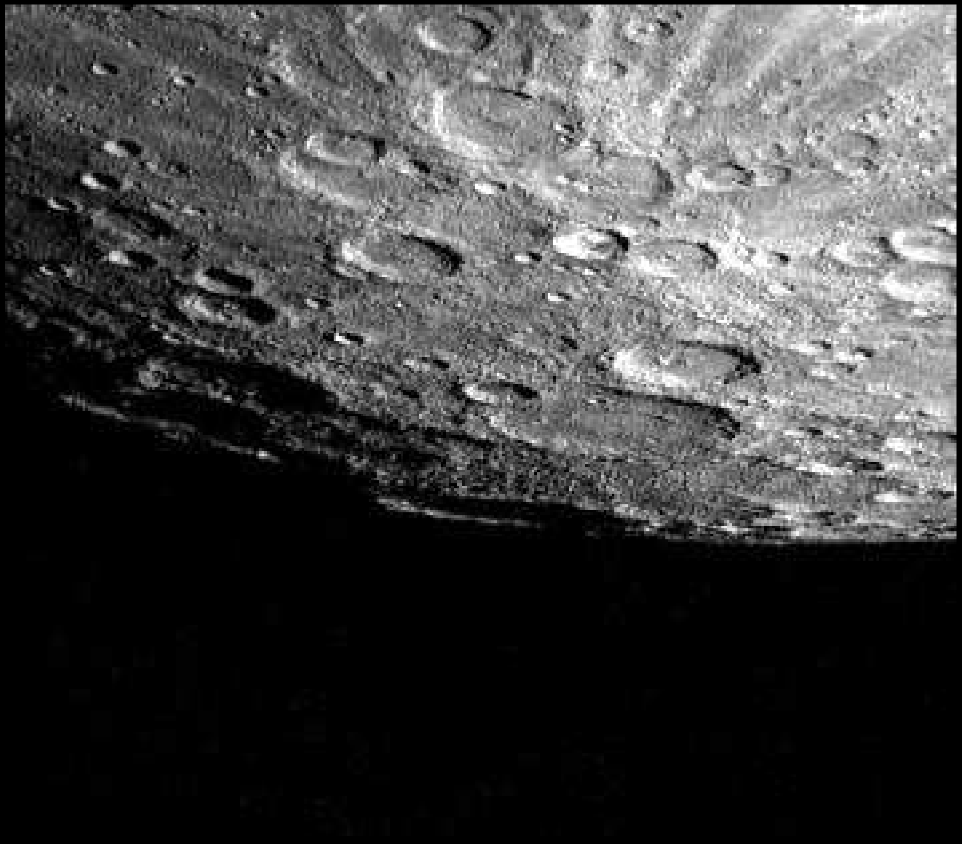
Terraced craters, with central mountains.

Major landforms: Craters



NASA/JPL

Terraced craters, with central mountains.



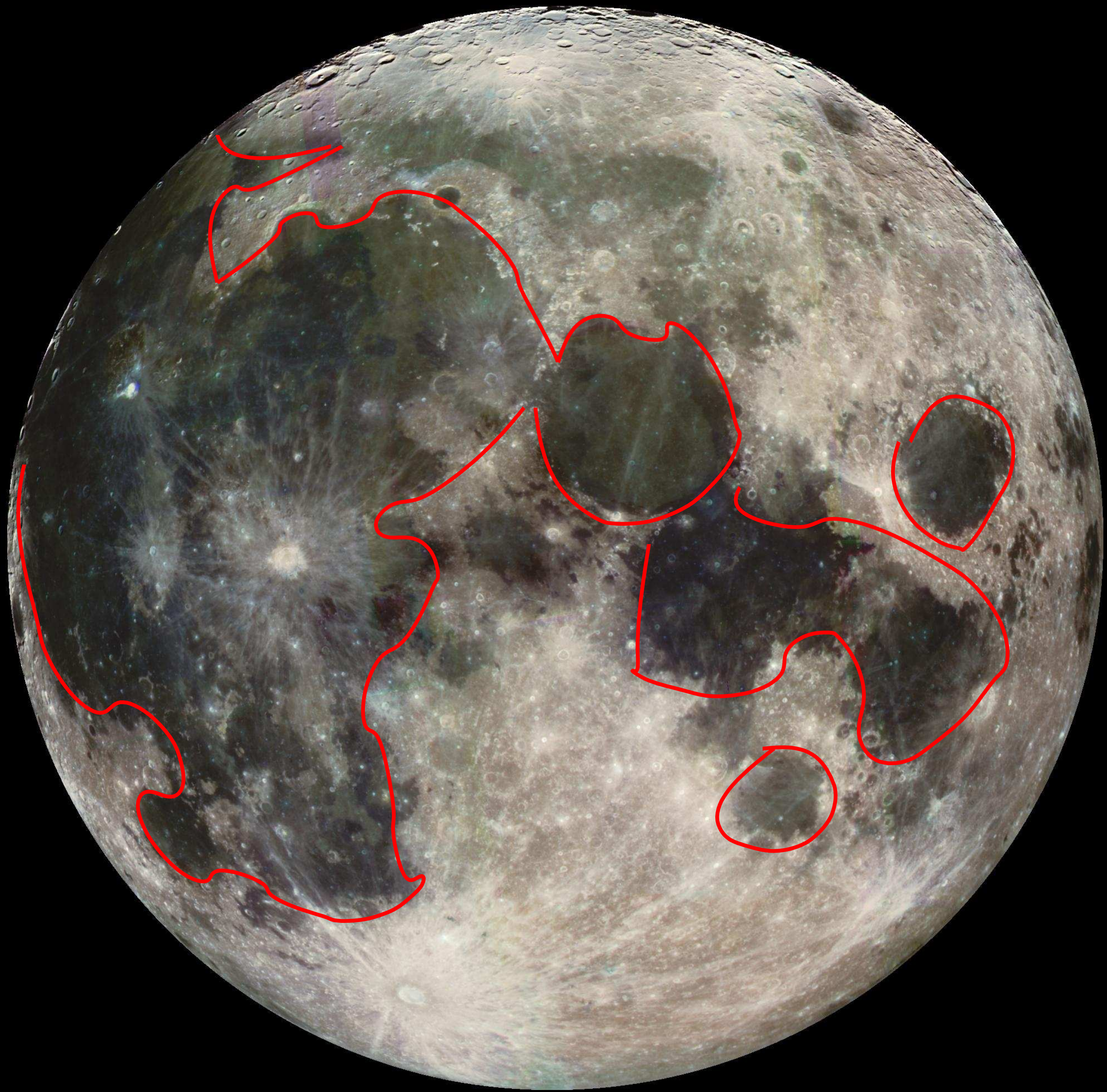
S-Pole; NASA/JPL

50 km diam craters with rays (remains from impact)

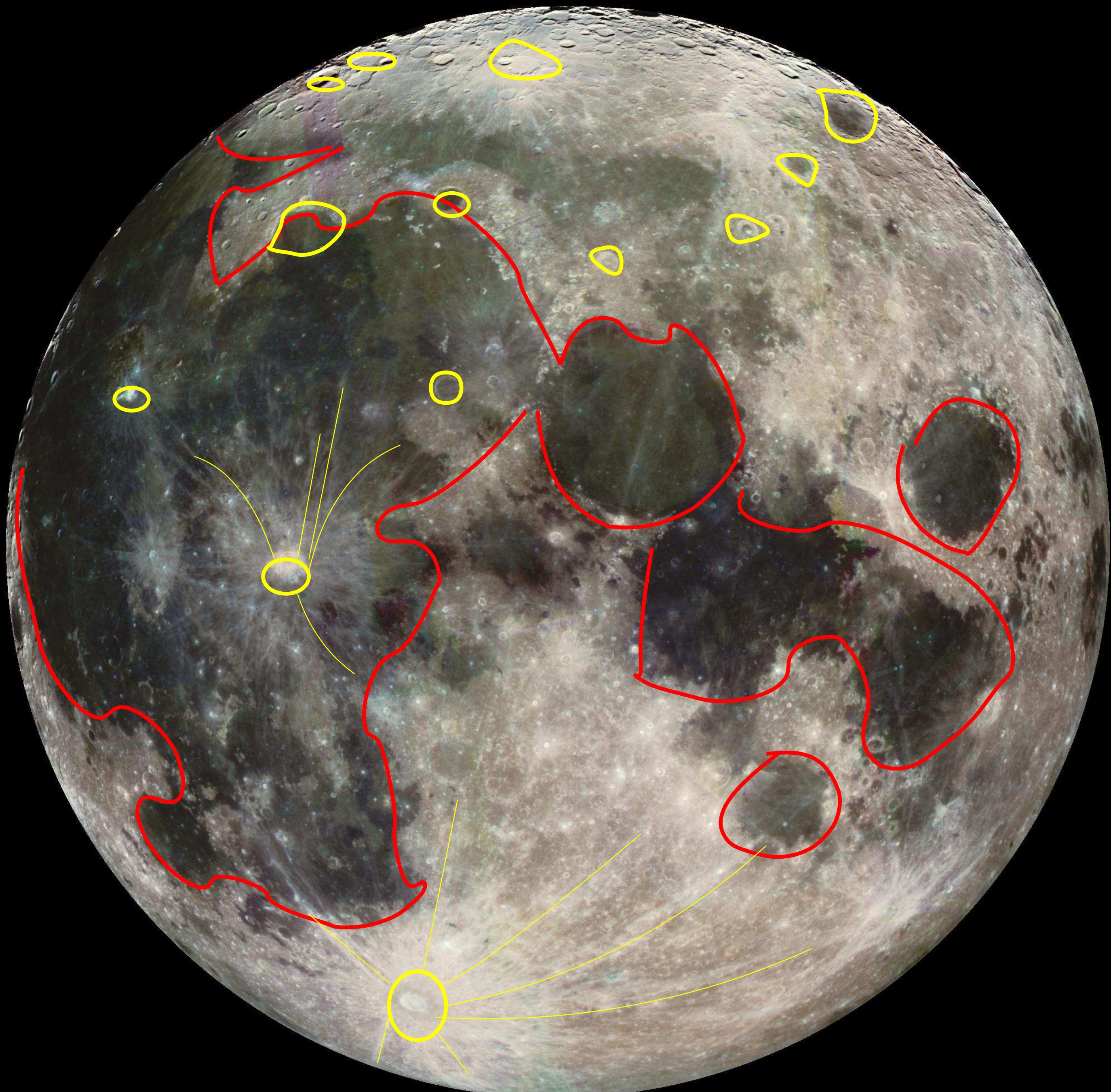
Earth: Wolf Creek Crater, Australia

Currently 172 confirmed impact structures on Earth

Earth’s Moon



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: Crater Copernicus

Moon: Apollo 16, 1972 Apr, Descartes Highlands

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

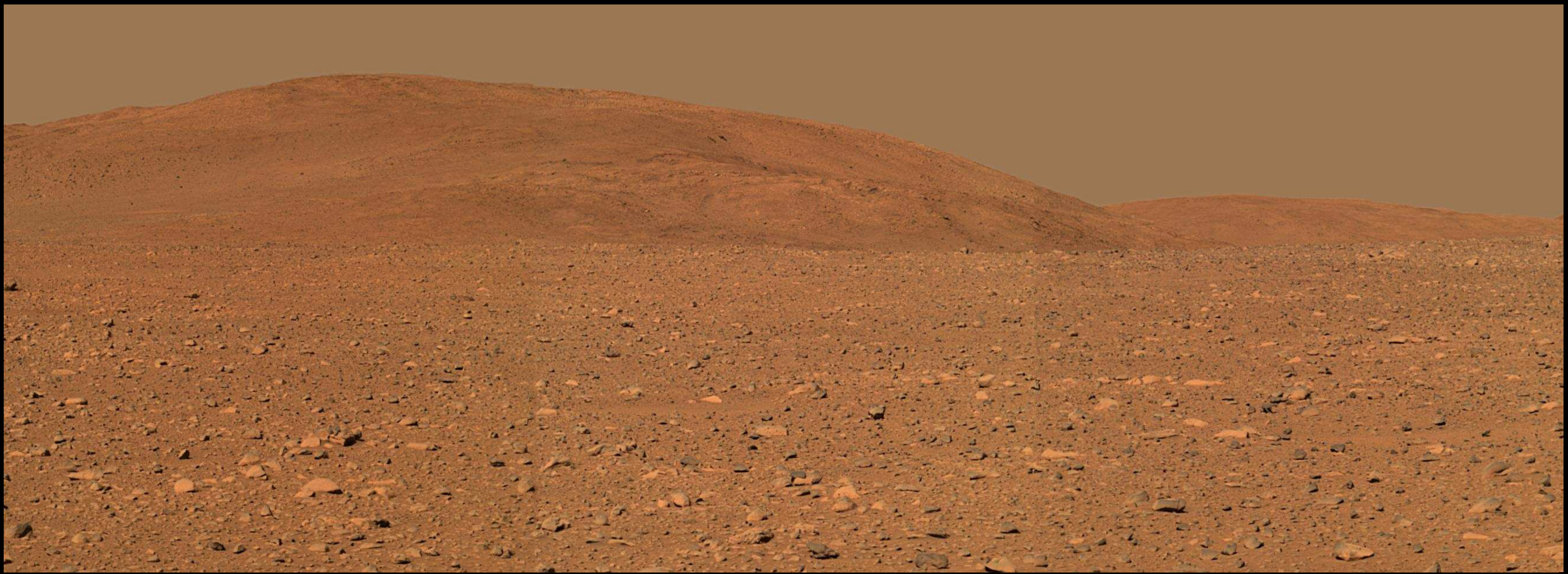
NASA/JPL/Cornell

NASA/JPL/Cornell

Mars: Crater Endurance

NASA/JPL/Cornell

Mars: Crater Endurance

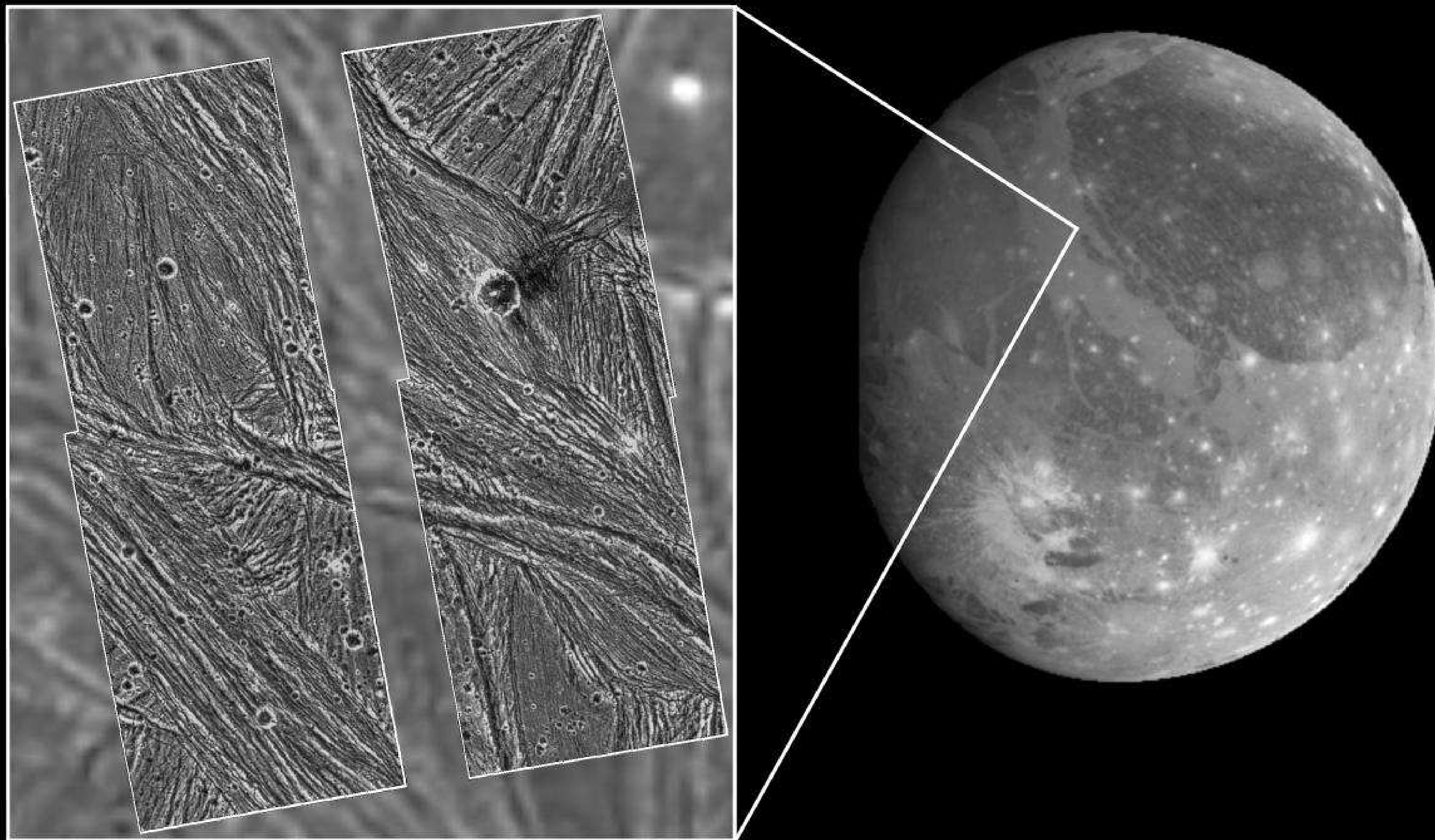


Mars: “Spirit” rolls towards Columbia Hills
(2004 June)

Mars: “Spirit” looks from Columbia Hills towards Gusev crater
(2004 Aug)

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

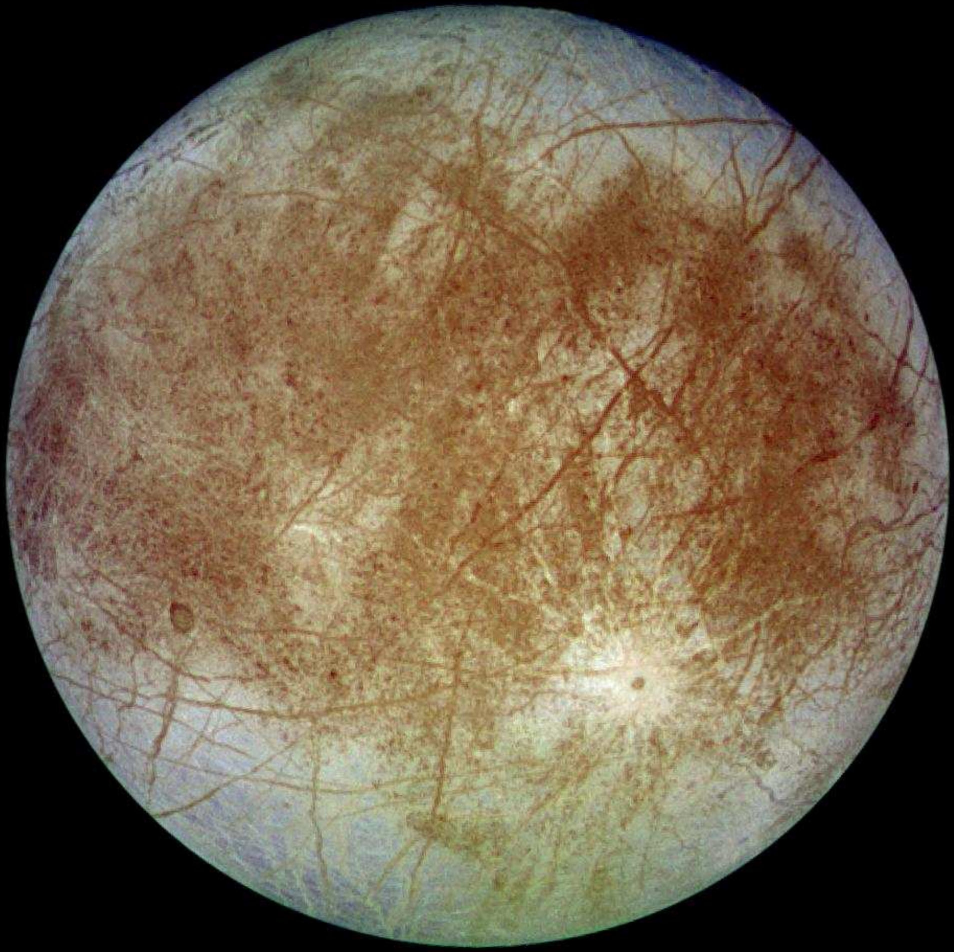
(N.B.: All Galilean moons tidally locked to Jupiter – always same side is facing Jupiter)

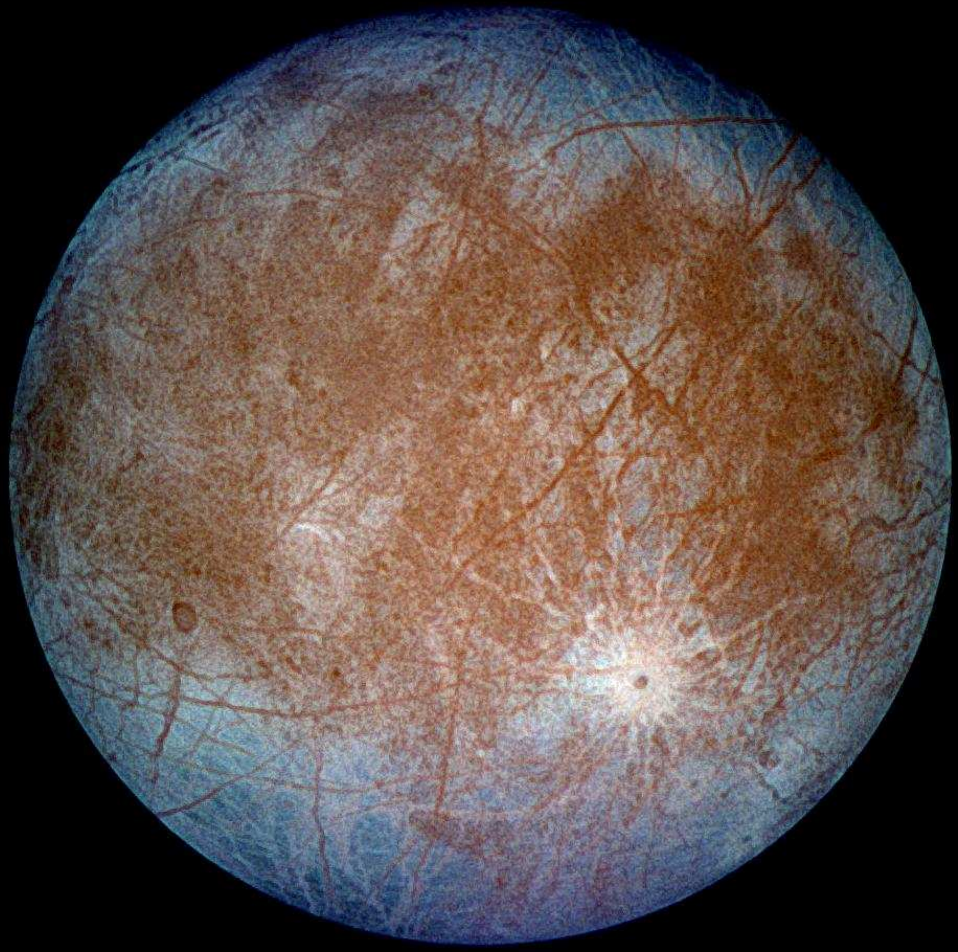


NASA Galileo / DLR, inset: 120×110 km

Ganymede – icy surface, ice hills and valleys, craters

Radius: 2634 km (∼ Mercury!)





NASA Galileo / DLR, 1996 September 7

Europa – icy surface with ridges (colors: different kinds of ice)

Radius: 1565 km ($\sim$ Earth Moon)

possibility of water ocean below surface

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

Radius: 2406 km (similar to Mercury!)

Impact Craters

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

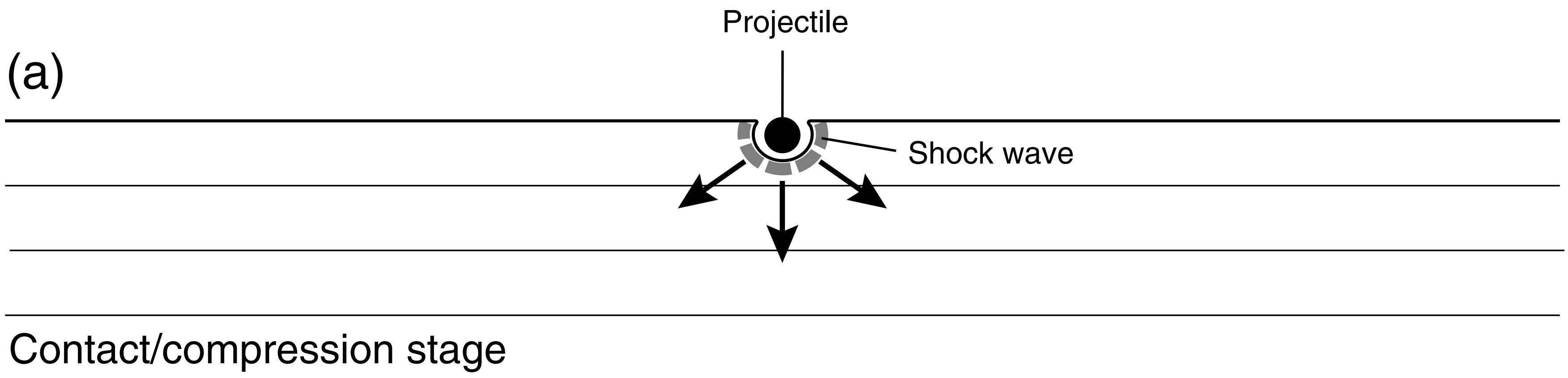
Example:

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)

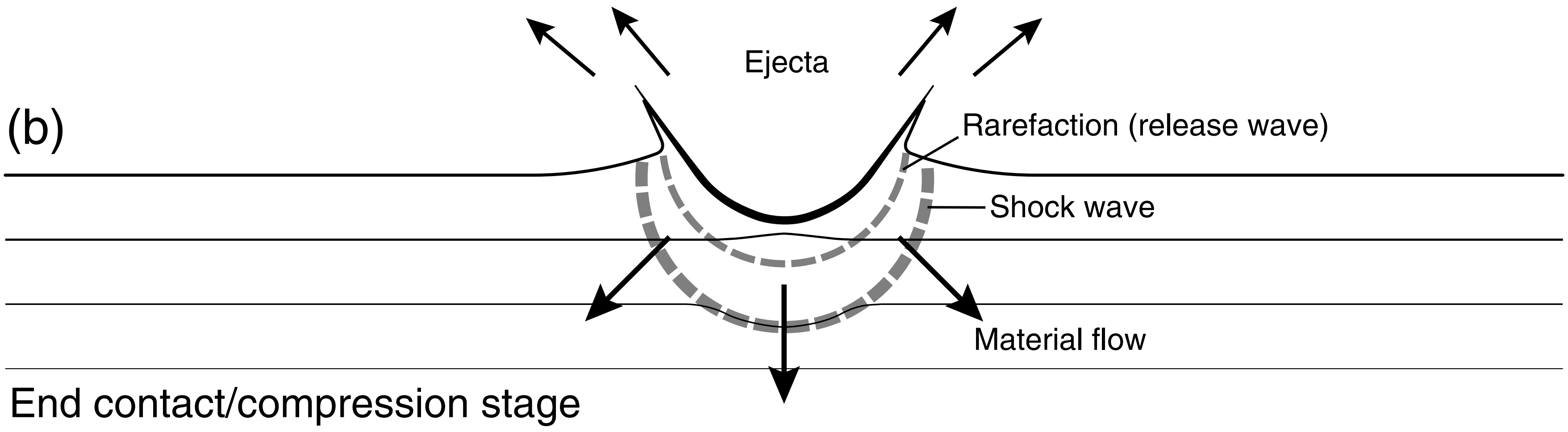
1 Megaton TNT is typical strength of US nuclear bombs [B-83 bomb]

Impact Craters



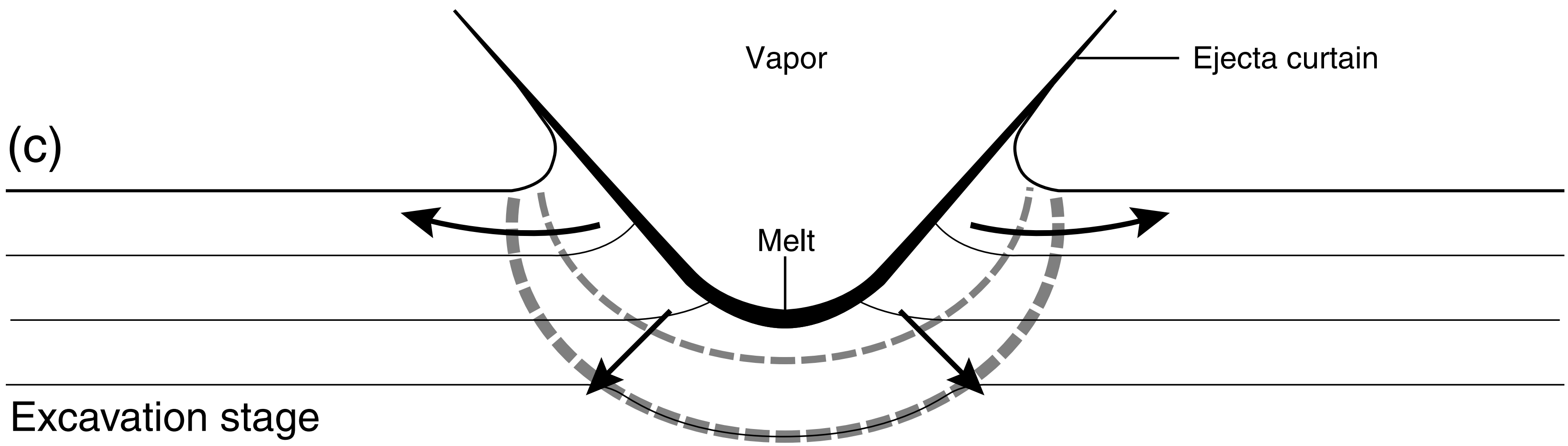
French, 1998, LPI Cont. 954

Impact Craters



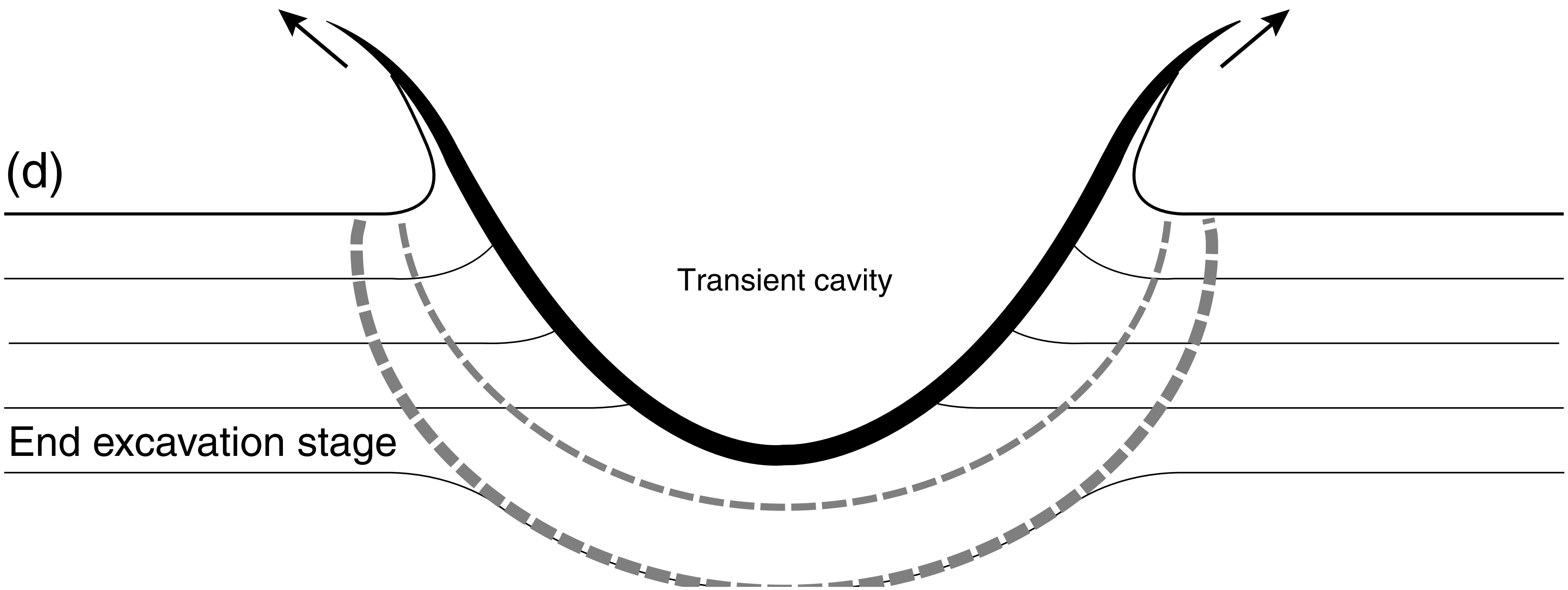
French, 1998, LPI Cont. 954

Impact Craters



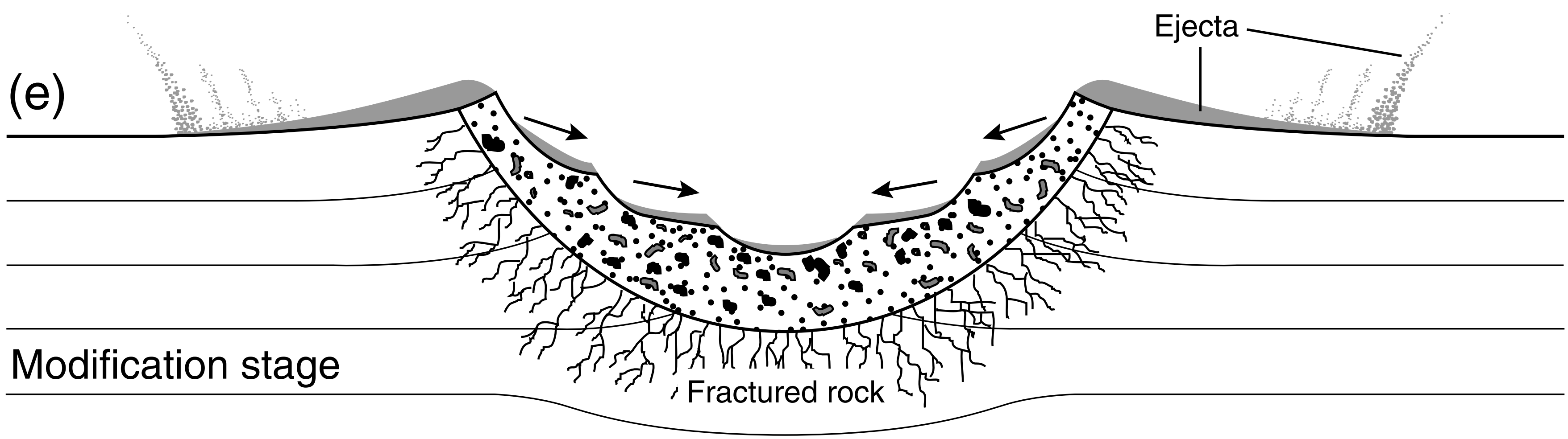
French, 1998, LPI Cont. 954

Impact Craters



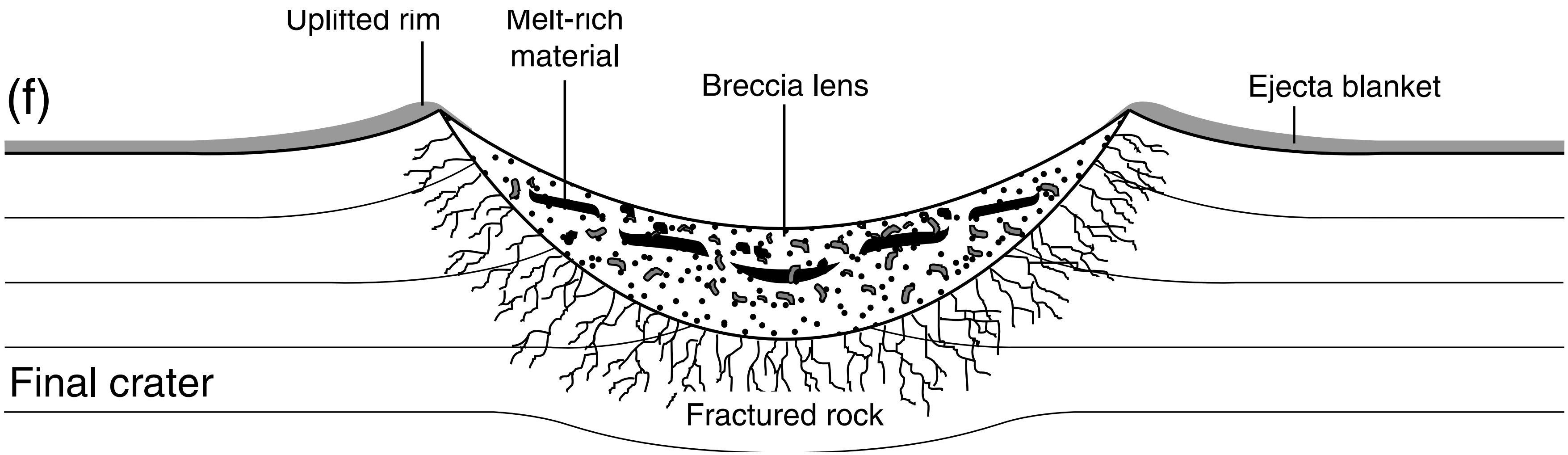
French, 1998, LPI Cont. 954

Impact Craters



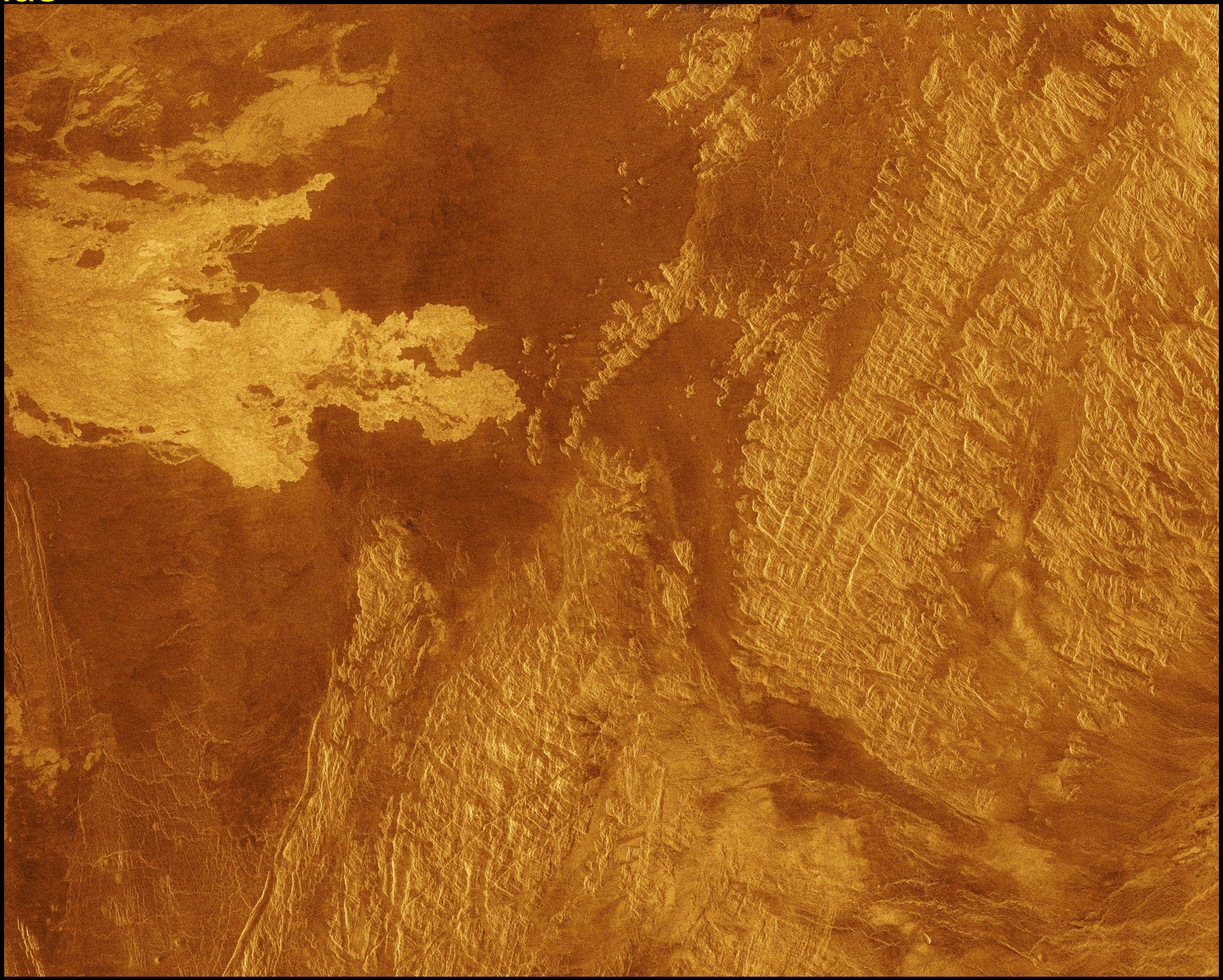
French, 1998, LPI Cont. 954

Impact Craters



French, 1998, LPI Cont. 954

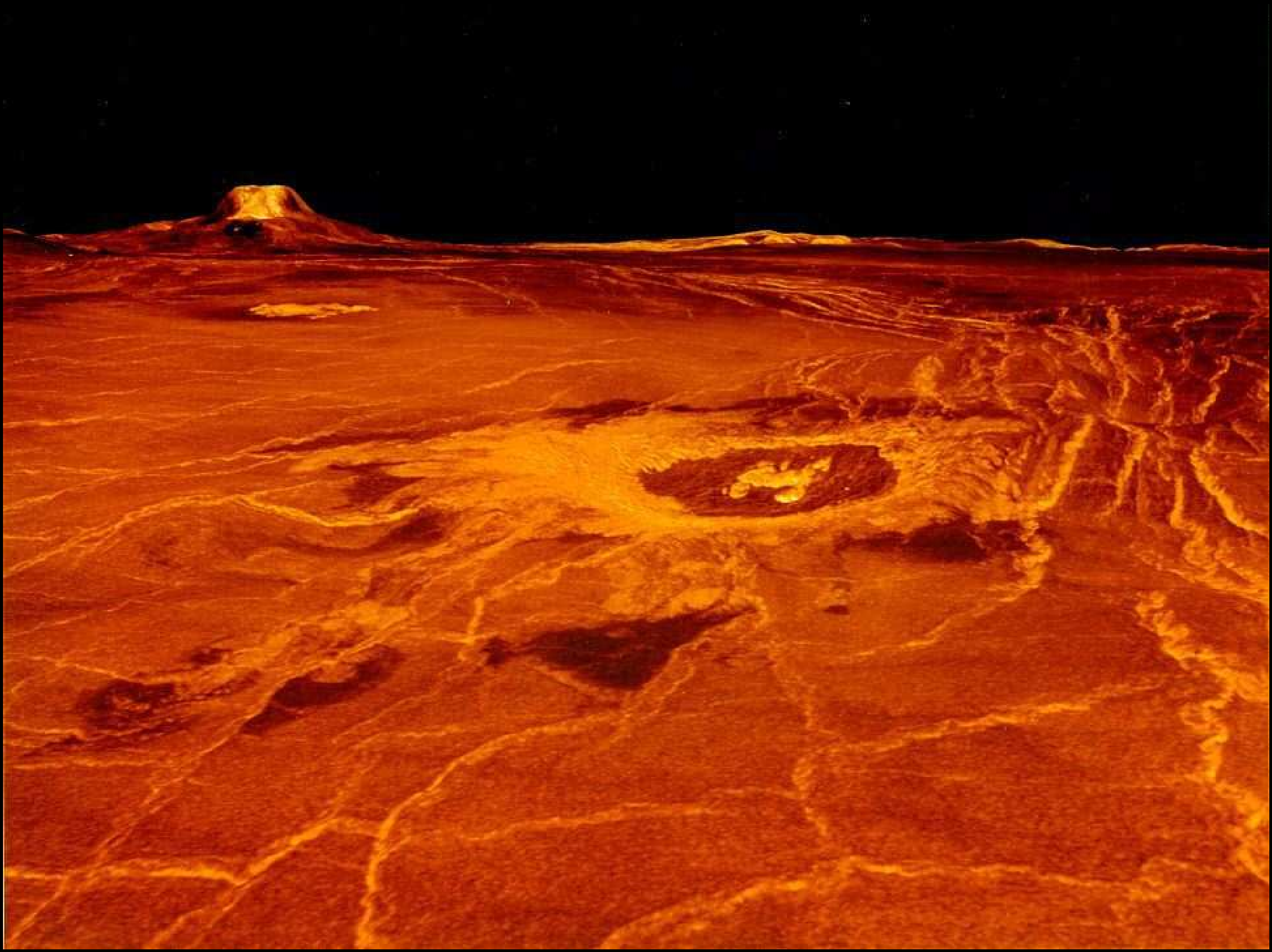
Venus



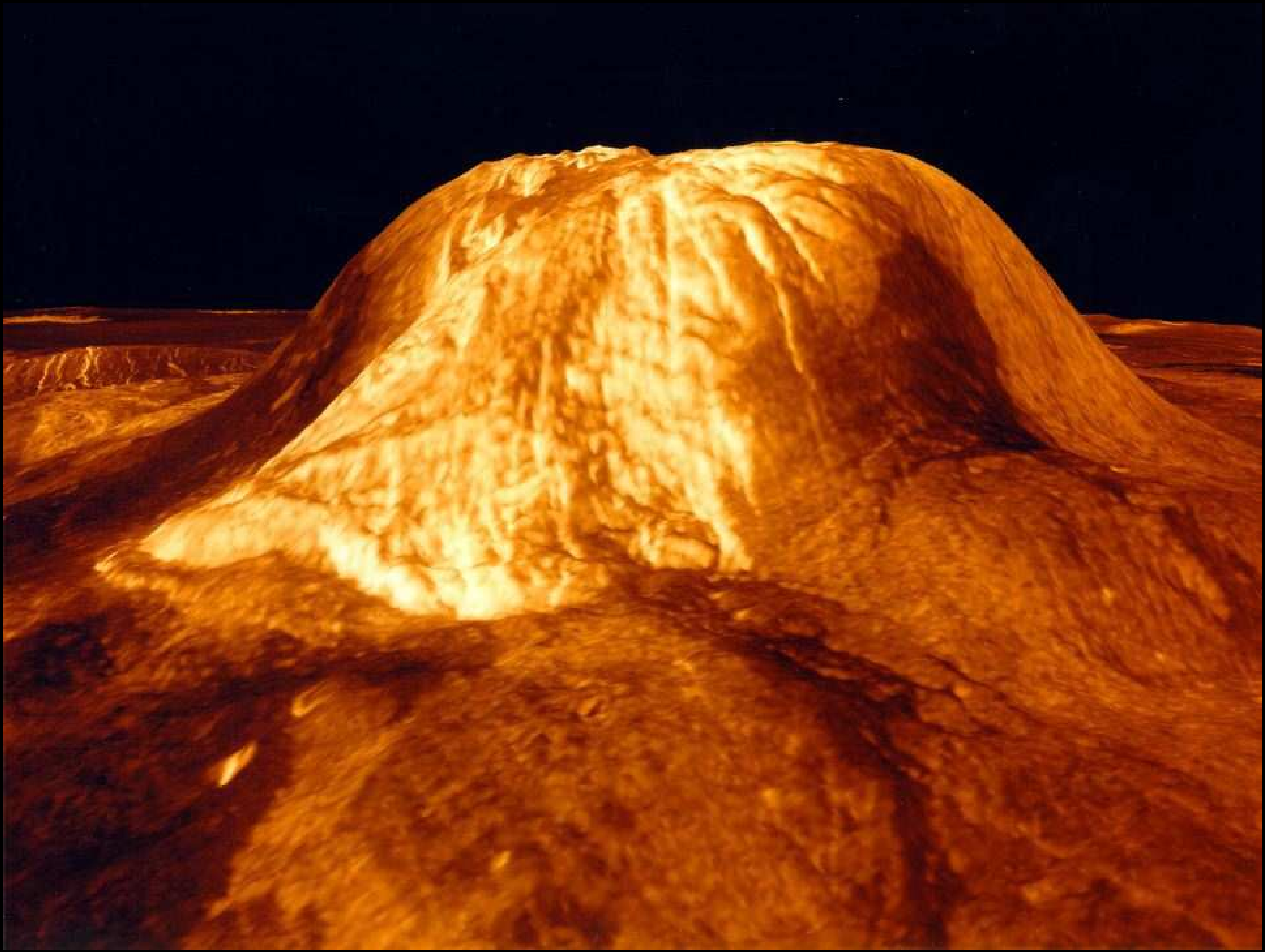
NASA, Magellan

$440 \times 350\,km^2$ 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)



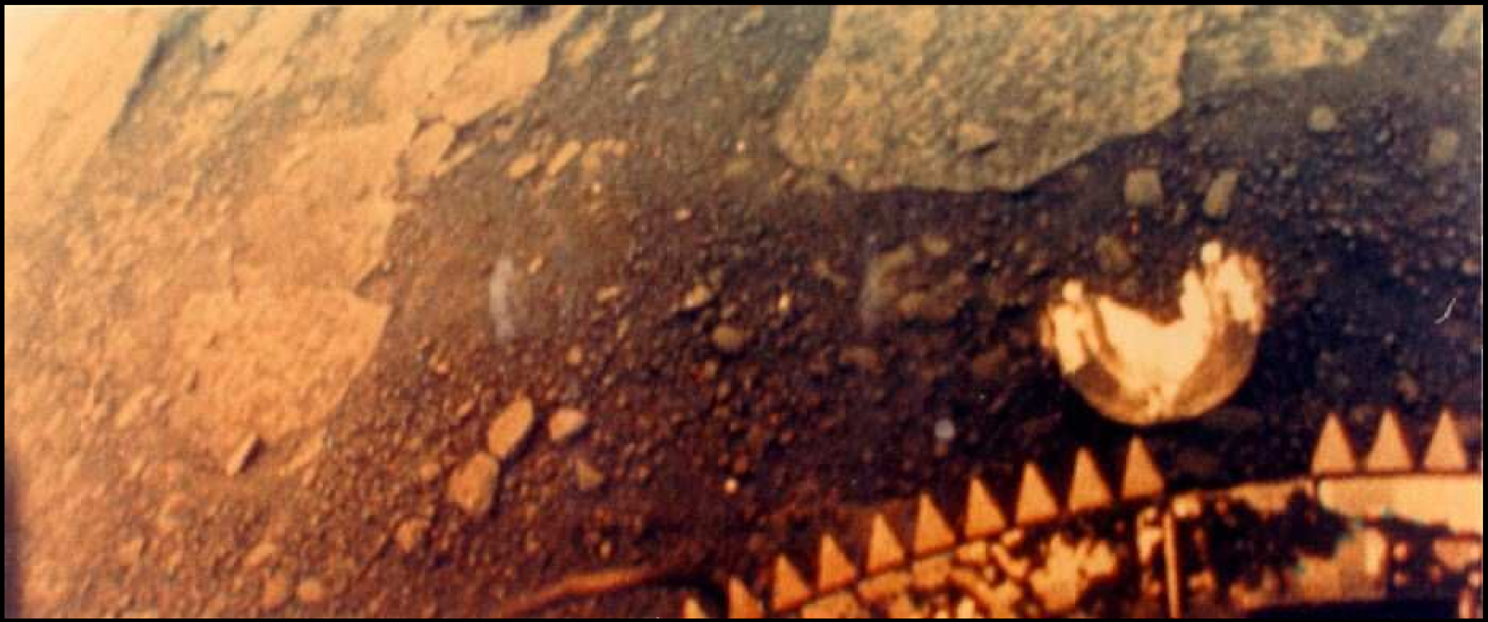
Eistla Regio; heights exagerrated by factor 22.5

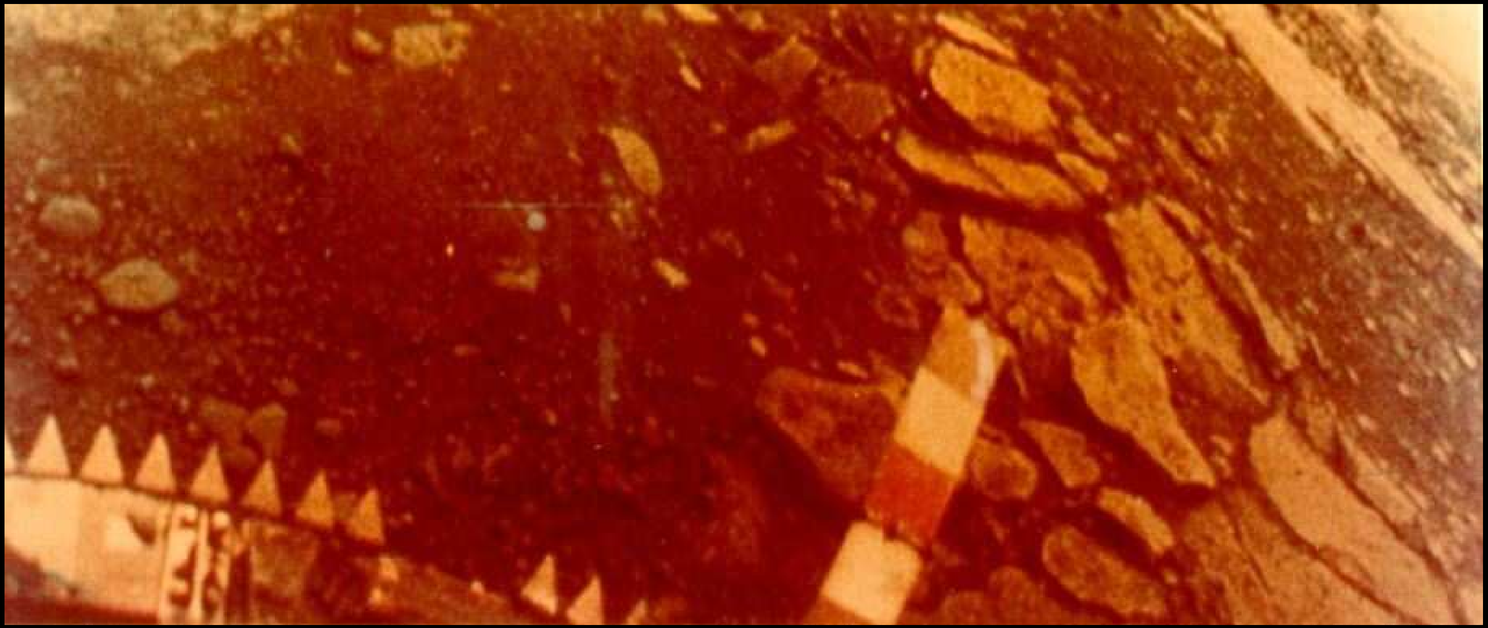


Gula Mons; heights exagerrated by factor 22.5

Gula Mons; real heights

Venus surface images:



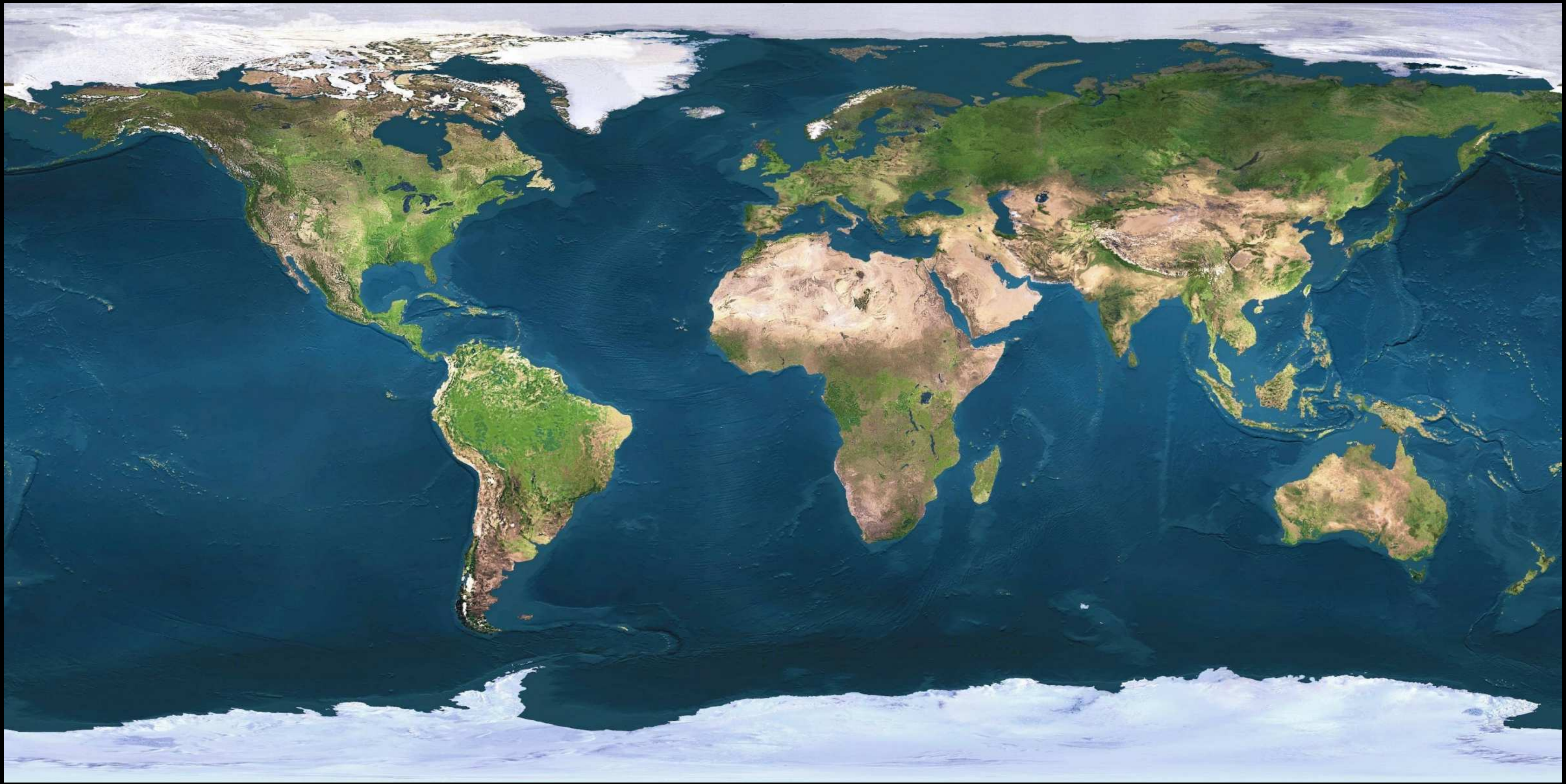


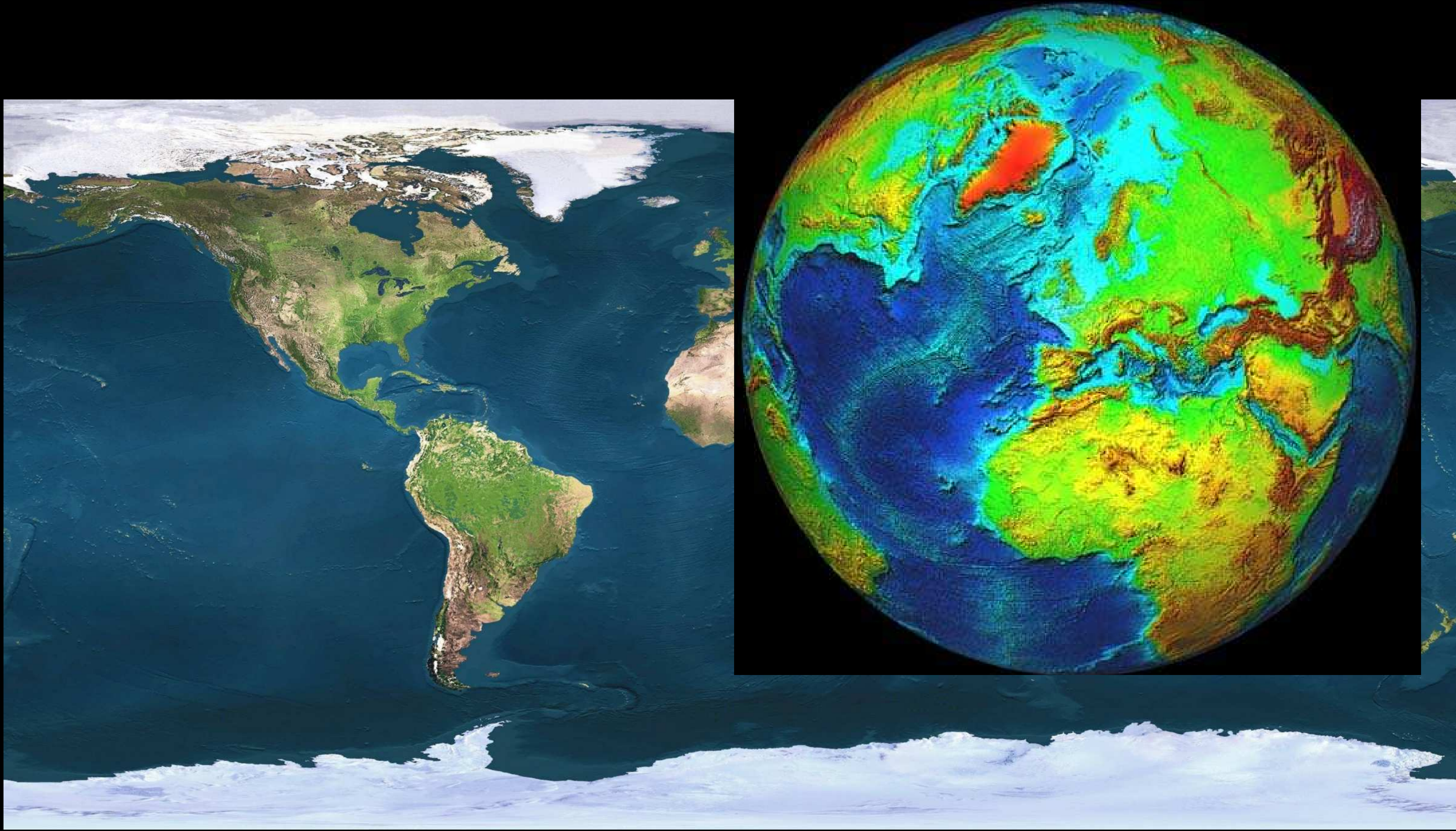
Venera 13 (3 March 1982): images from color TV camera

courtesy D.P. Mitchell

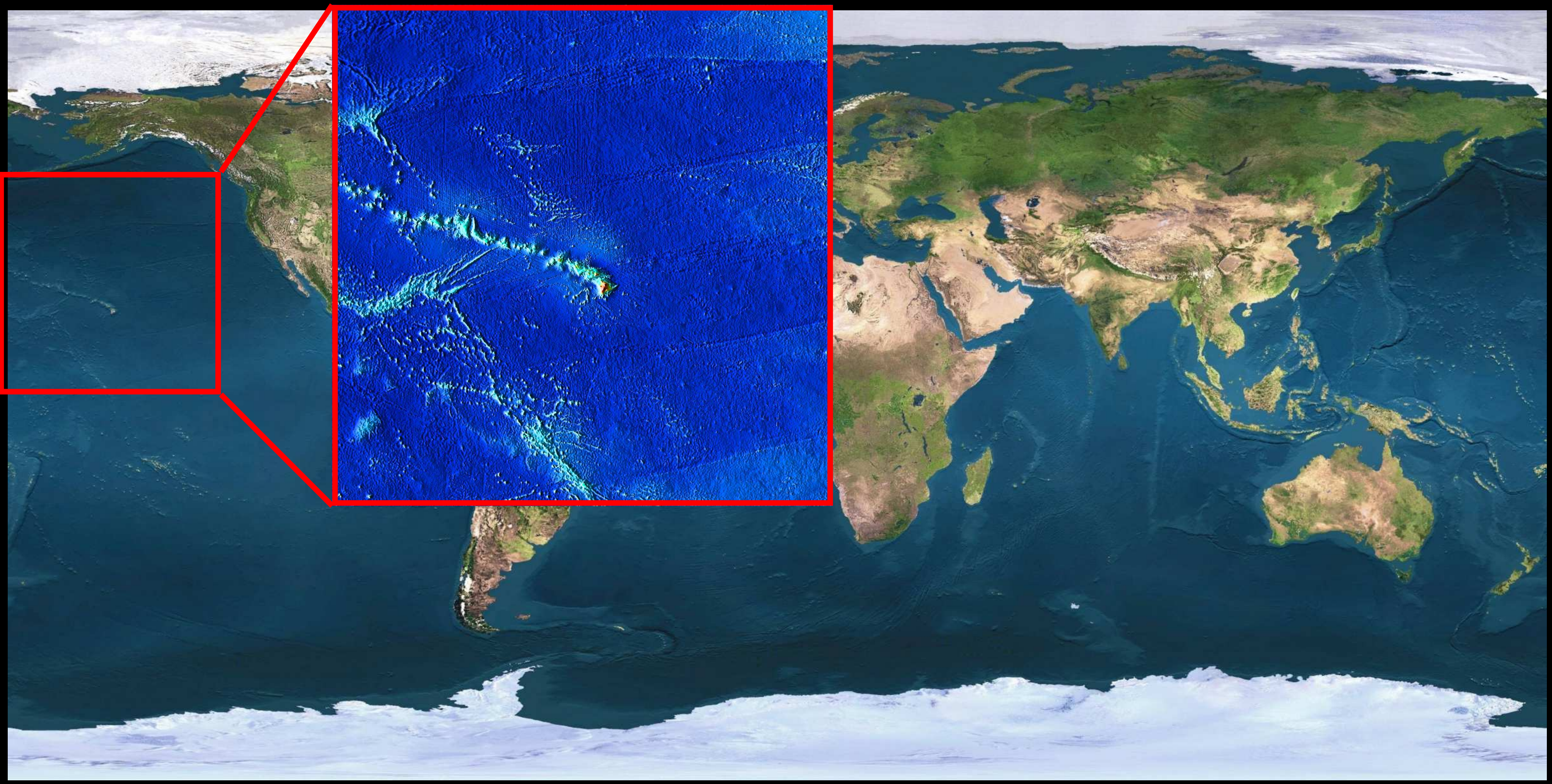
Venera 13 (3 March 1982): reanalysed image without camera distortion

Venera 14 (5 May 1982)



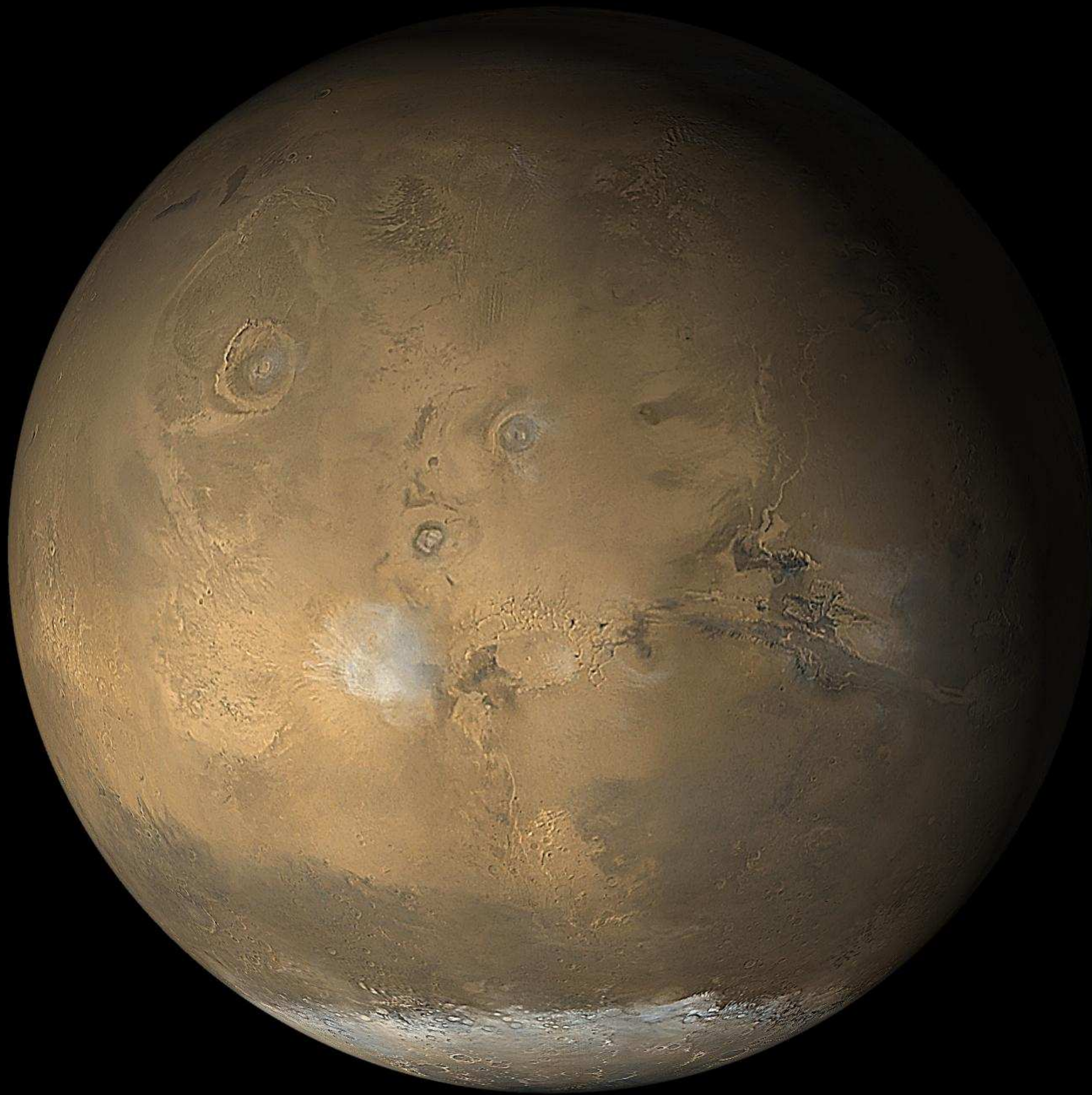


Evidence for plate tectonics (few craters!)



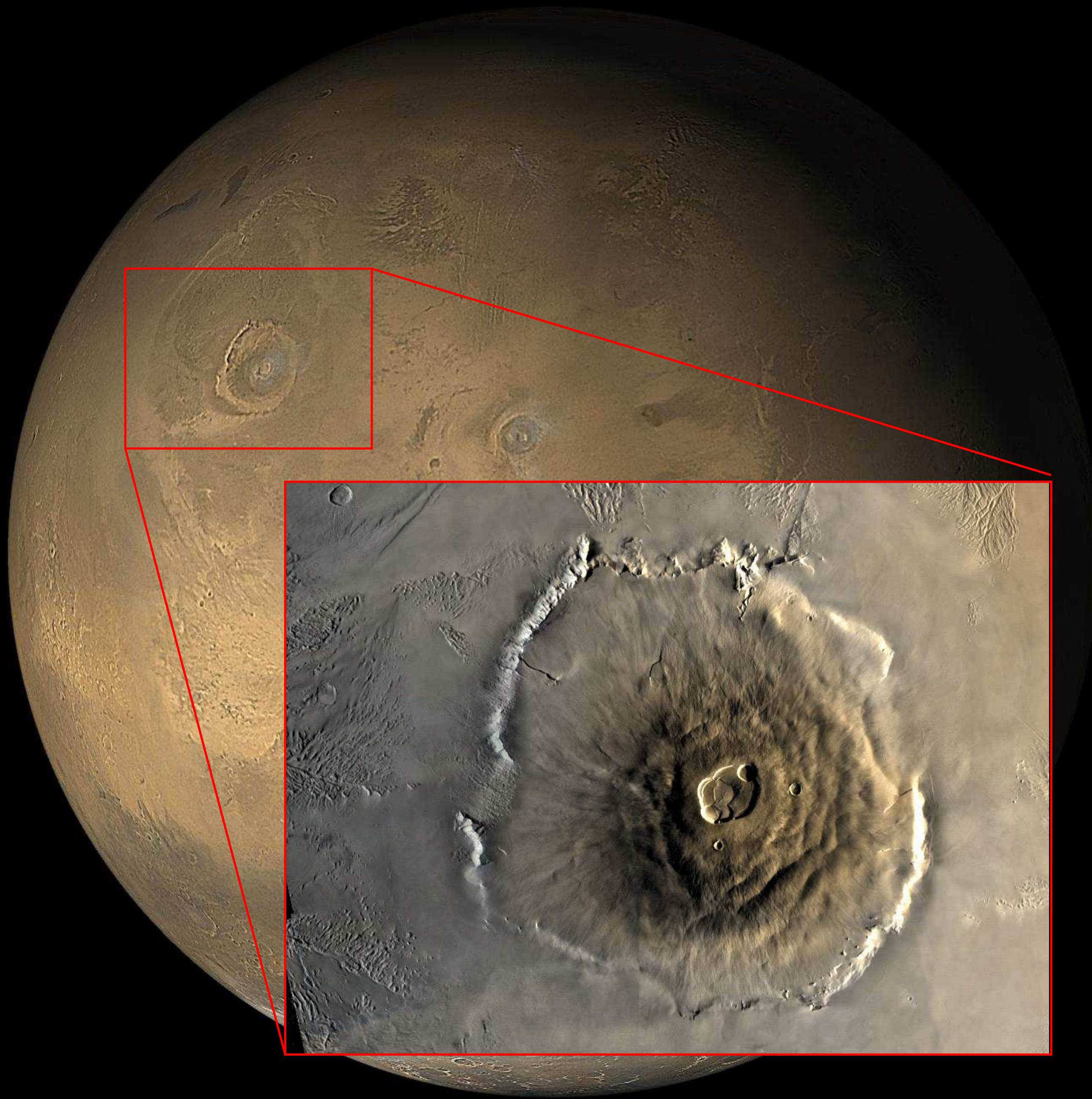
Evidence for plate tectonics (few craters!) , volcanism,. . .


Map
Satellite
Hybrid
2 mi
2 km
©2006 Google - Imagery ©2006 DigitalGlobe, TerraMetrics, Map data ©2006 NAVTEQ™ - Terms of Use
Hawaii



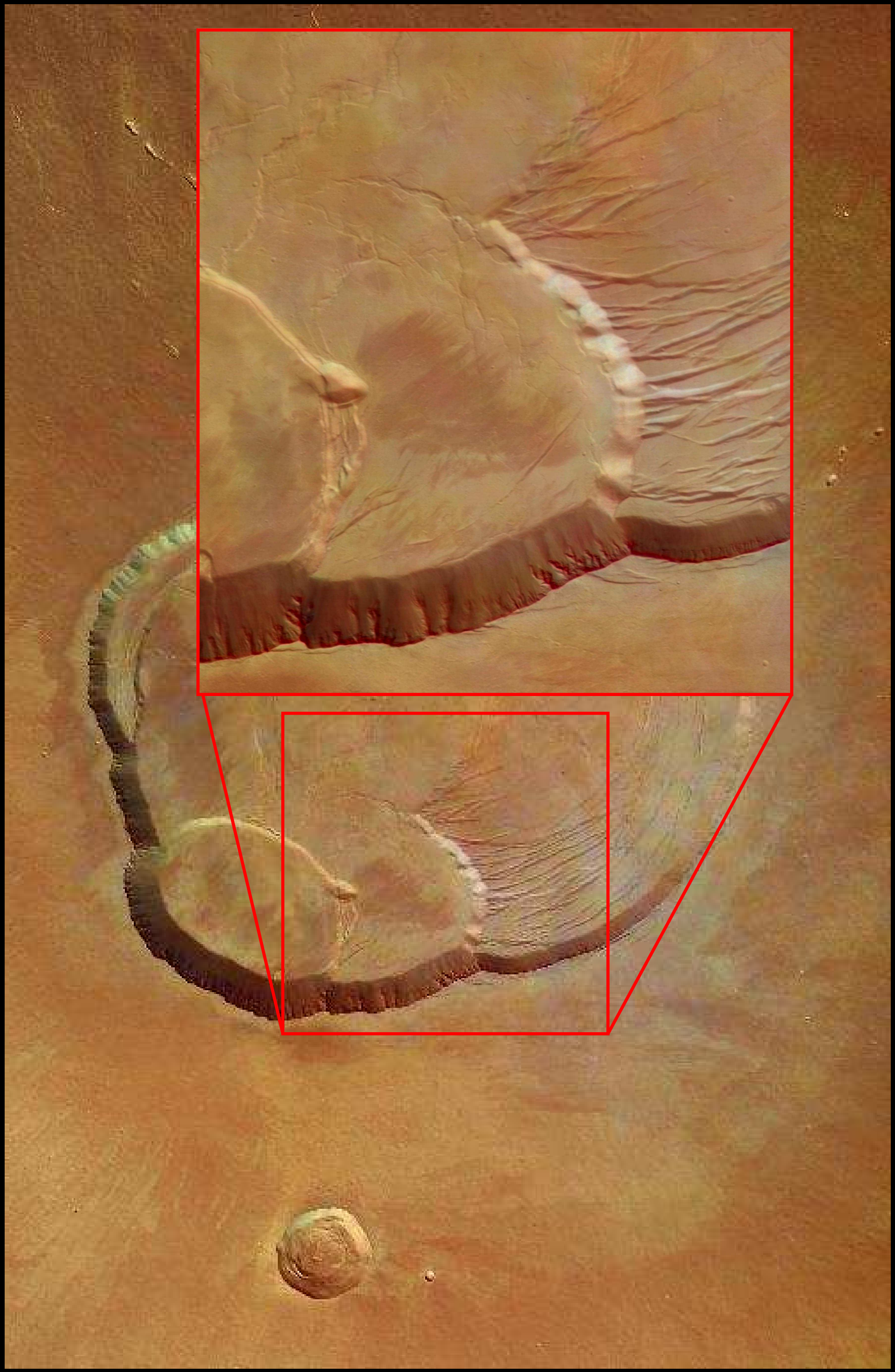
Mars: Tharsis vulcanos: Large shield vulcanos, now extinct

$\Longrightarrow$ no plate tectonics $\Longrightarrow$ Mars interior is colder than Earth.



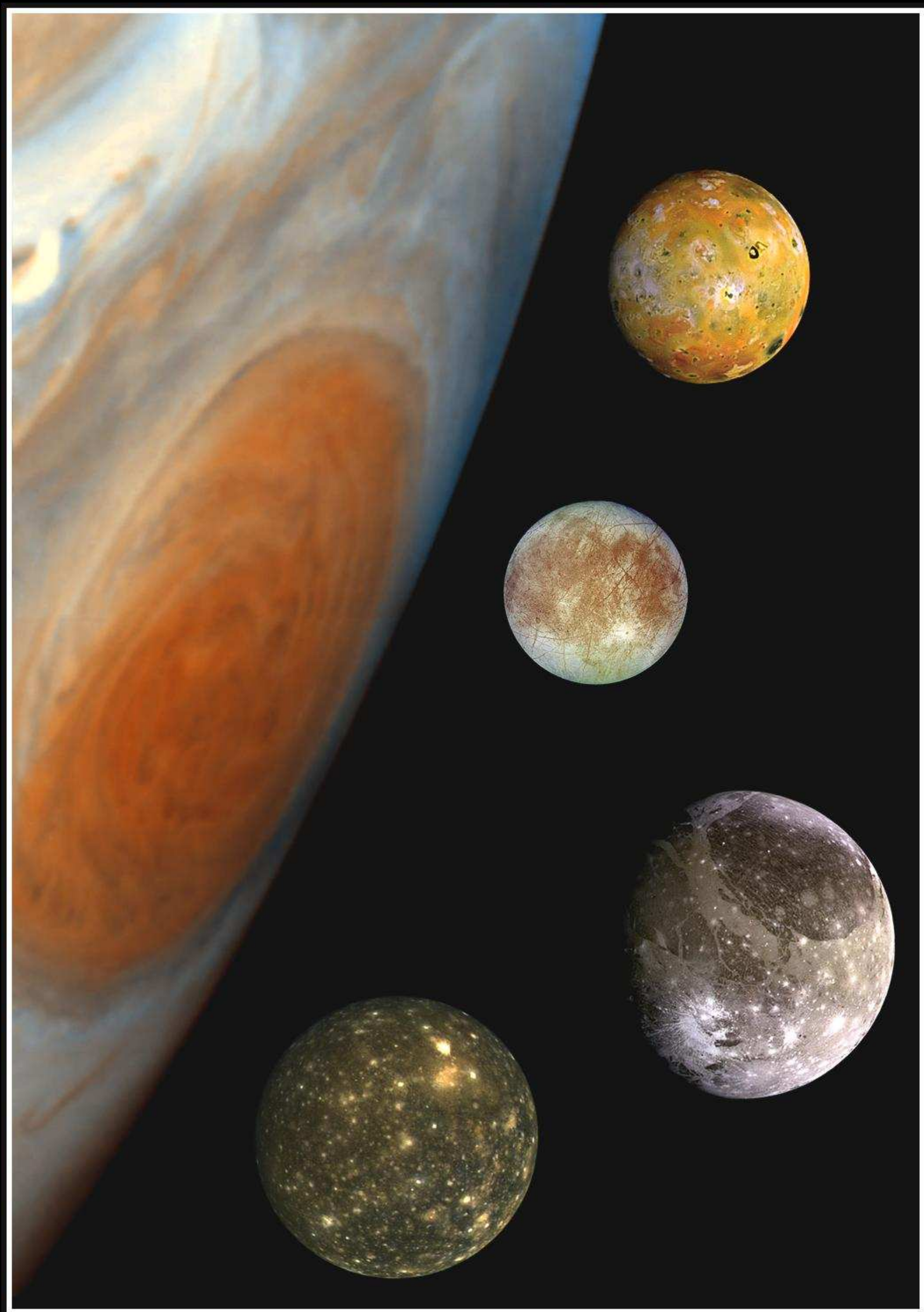
Olympus Mons: highest volcano in solar system
(25 km above surrounding plain; but slope only 2° to 5°).

ESA/Mars Express, HRSC, 11.02.2004



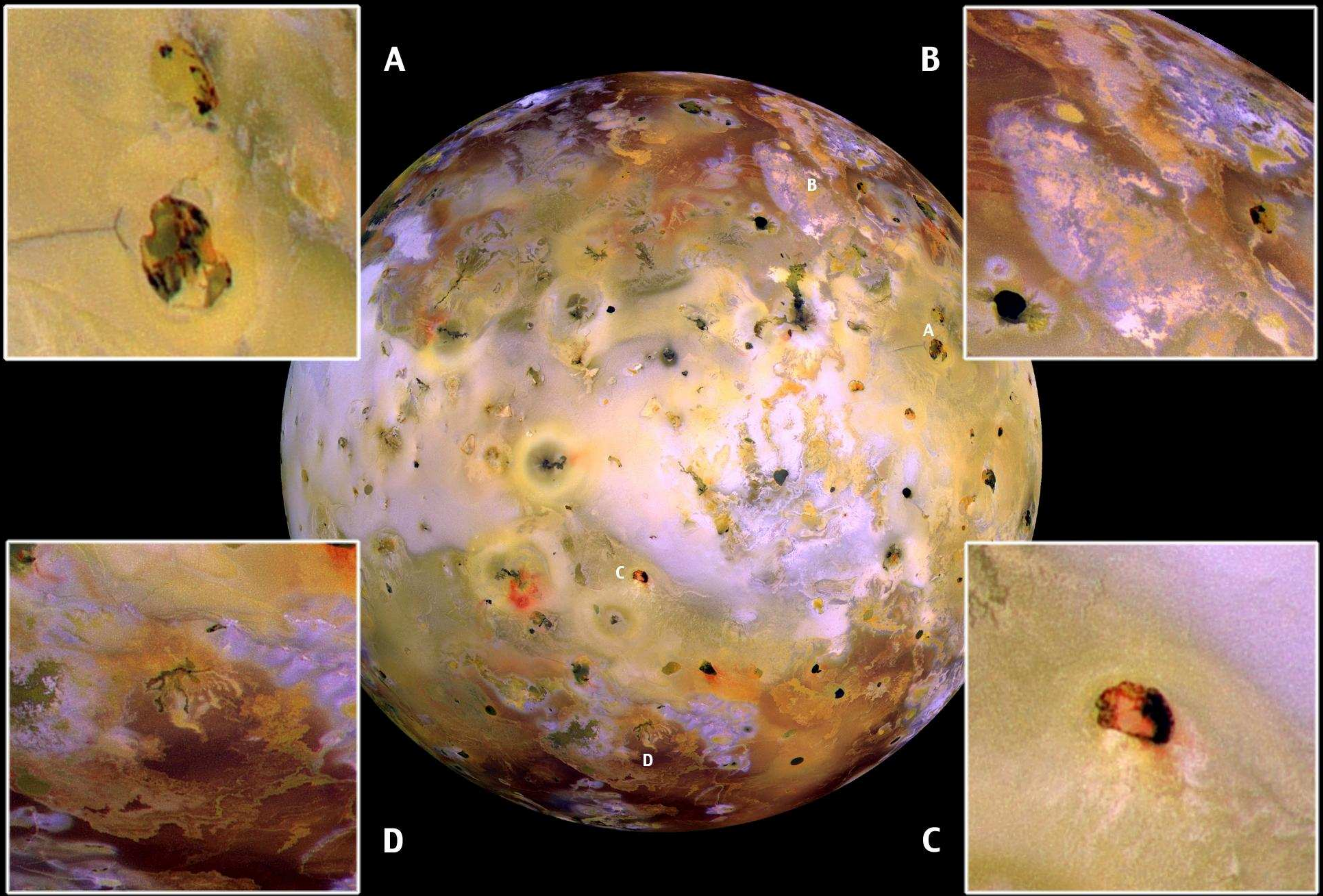
ESA/Mars Express, HRSC, 11.02.2004

ESA/Mars Express, HRSC, 11.02.2004

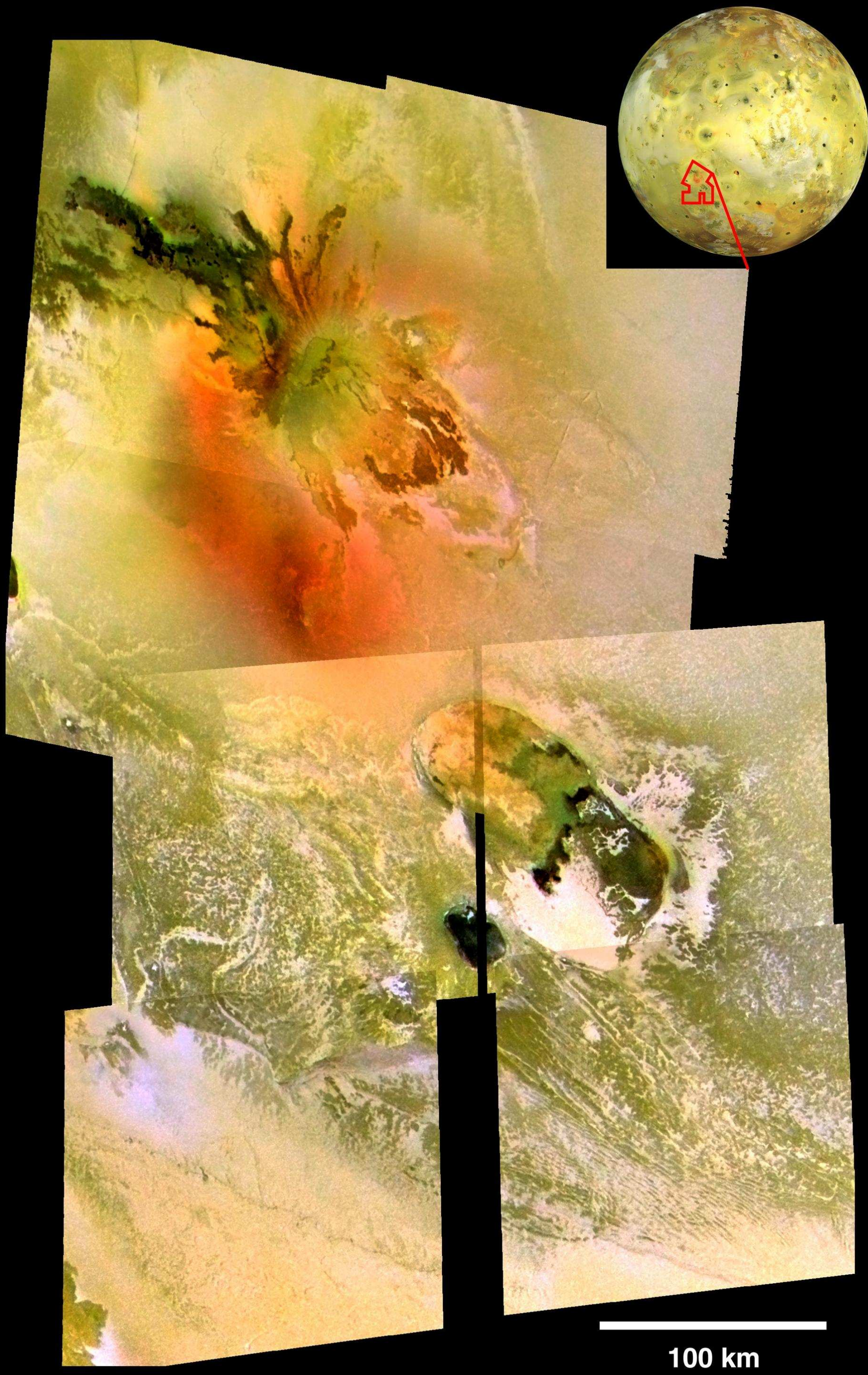


Montage of Jupiter and Galilean Moons: top to bottom: Io, Europa, Ganymede and Callisto.

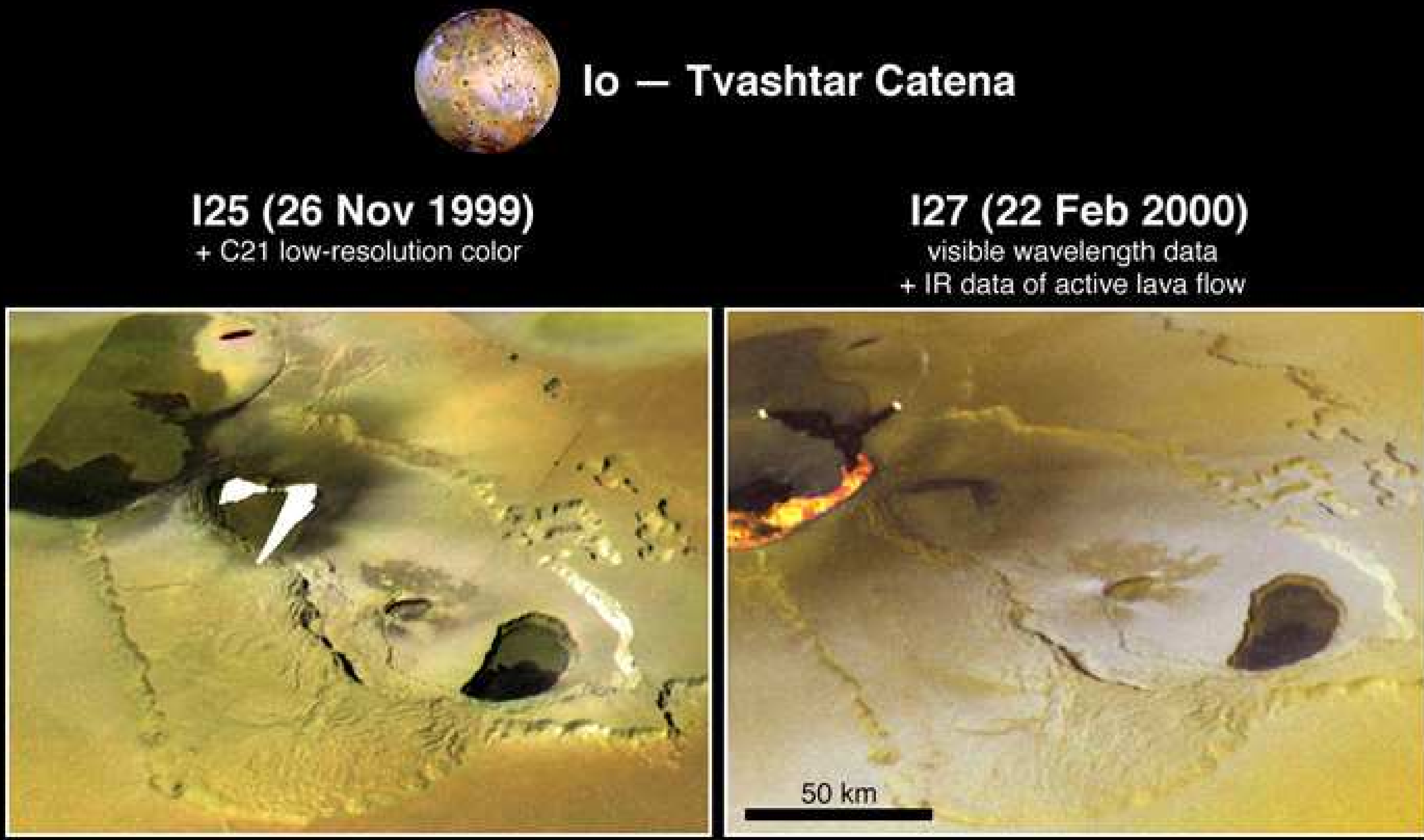
(N.B.: All Galilean moons tidally locked to Jupiter – always same side is facing Jupiter)



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

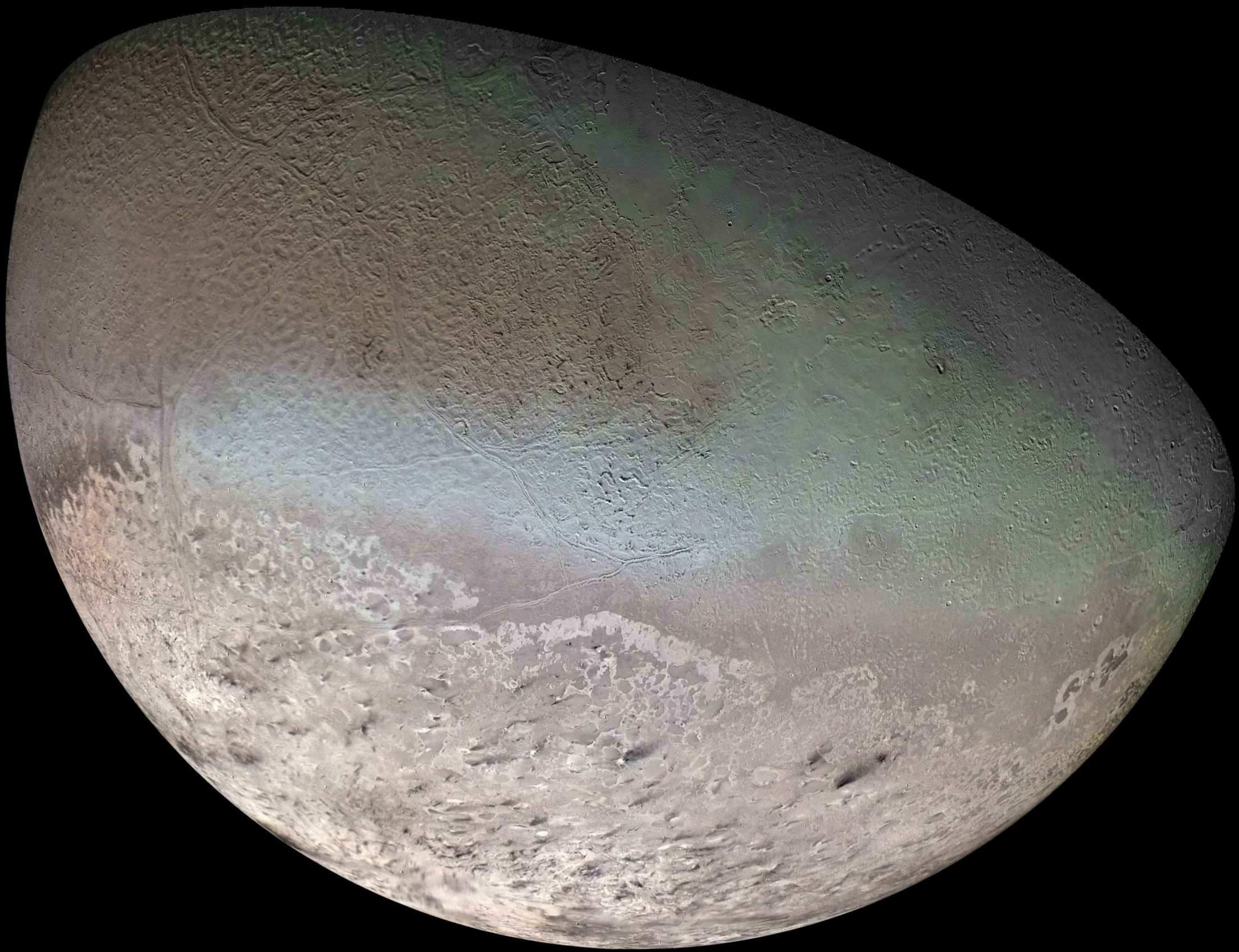


Active vulcanoes on Io
(interior heated by tidal forces from Jupiter), color due to large contents of sulphur and sulphur oxides in lava.
Height of vulcanoes: 6 km or higher



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

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



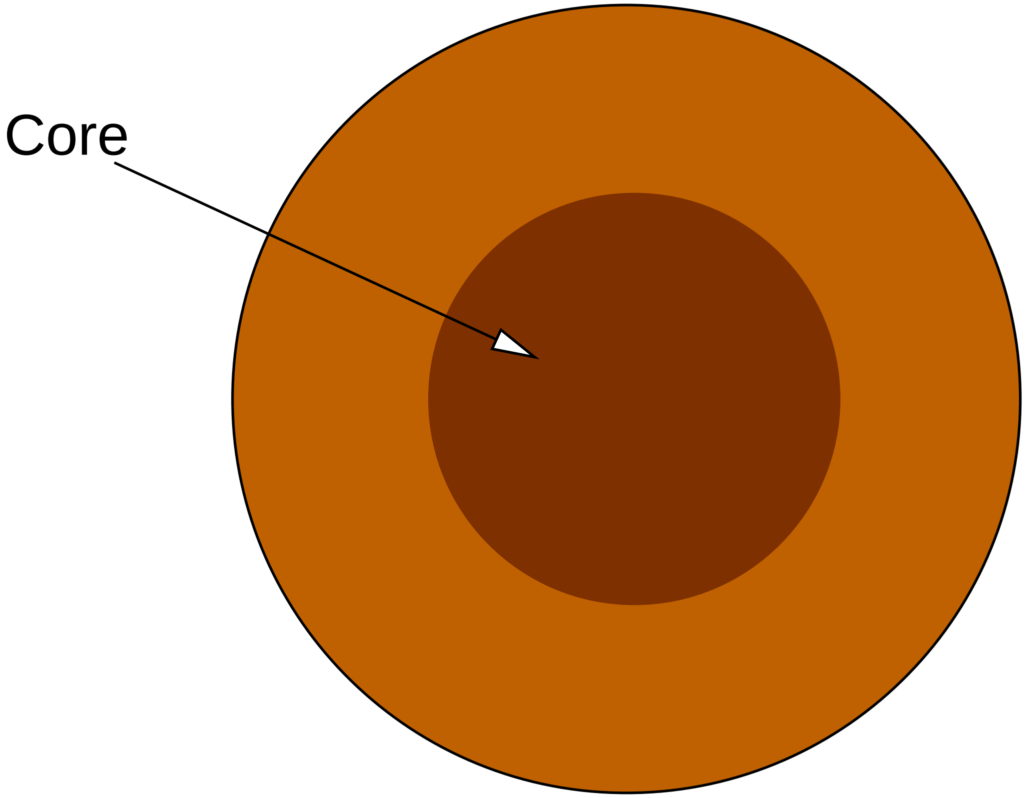
NASA/Voyager 2/Calvin J. Hamilton

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 $T \sim 70\,\mathrm{K}$)

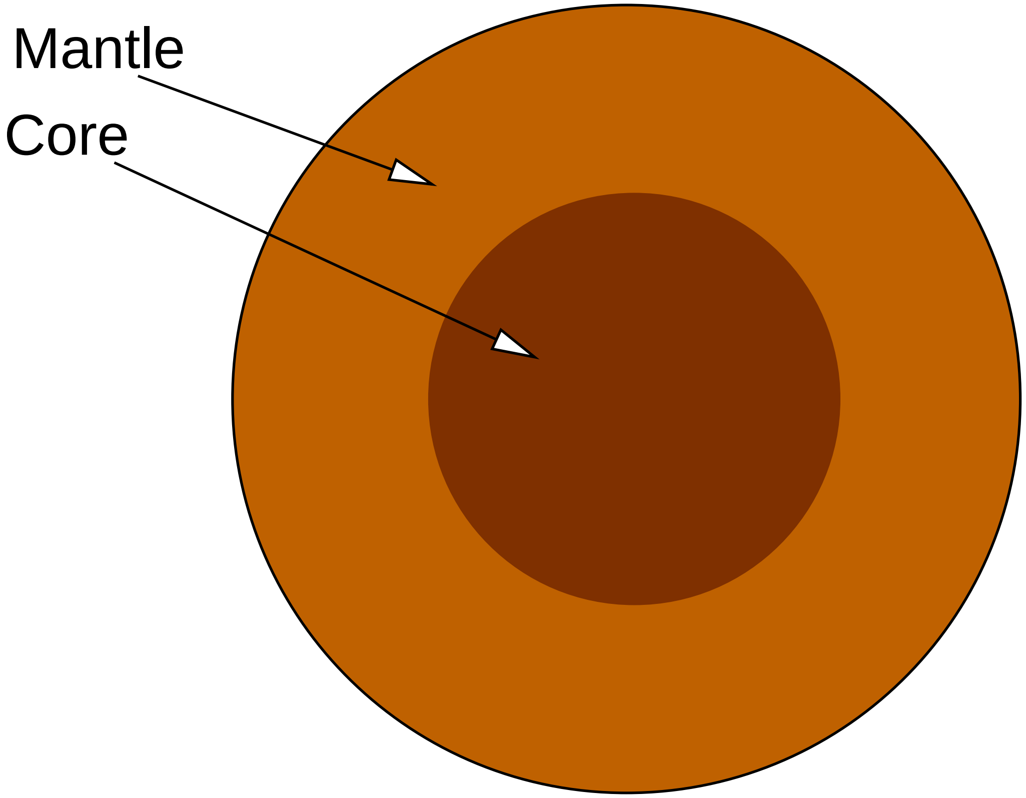
Interiors: Terrestrial Planets, I



Structure of terrestrial planets:

- Core: high-density material (Fe)

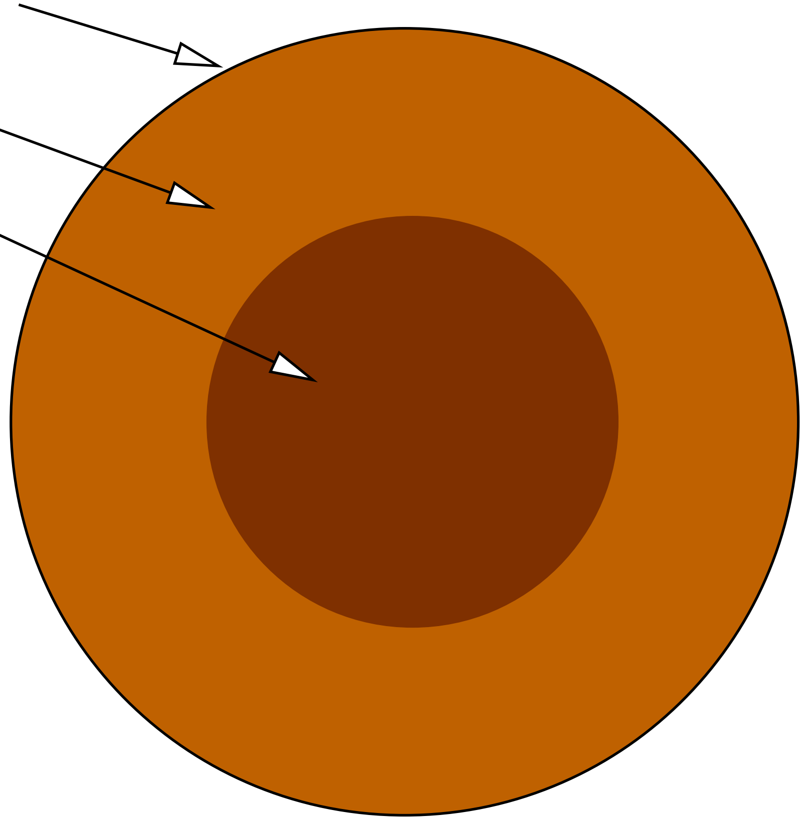
Interiors: Terrestrial Planets, II



Structure of terrestrial planets:

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

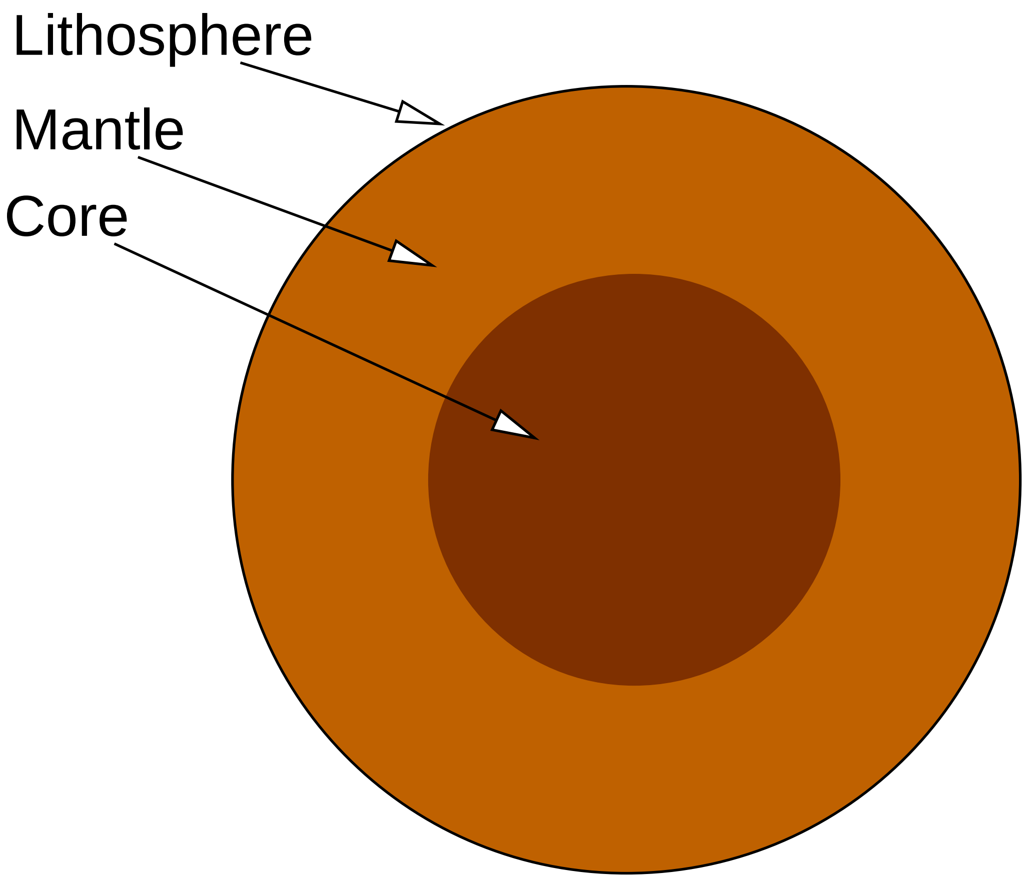
Interiors: Terrestrial Planets, III



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, IV

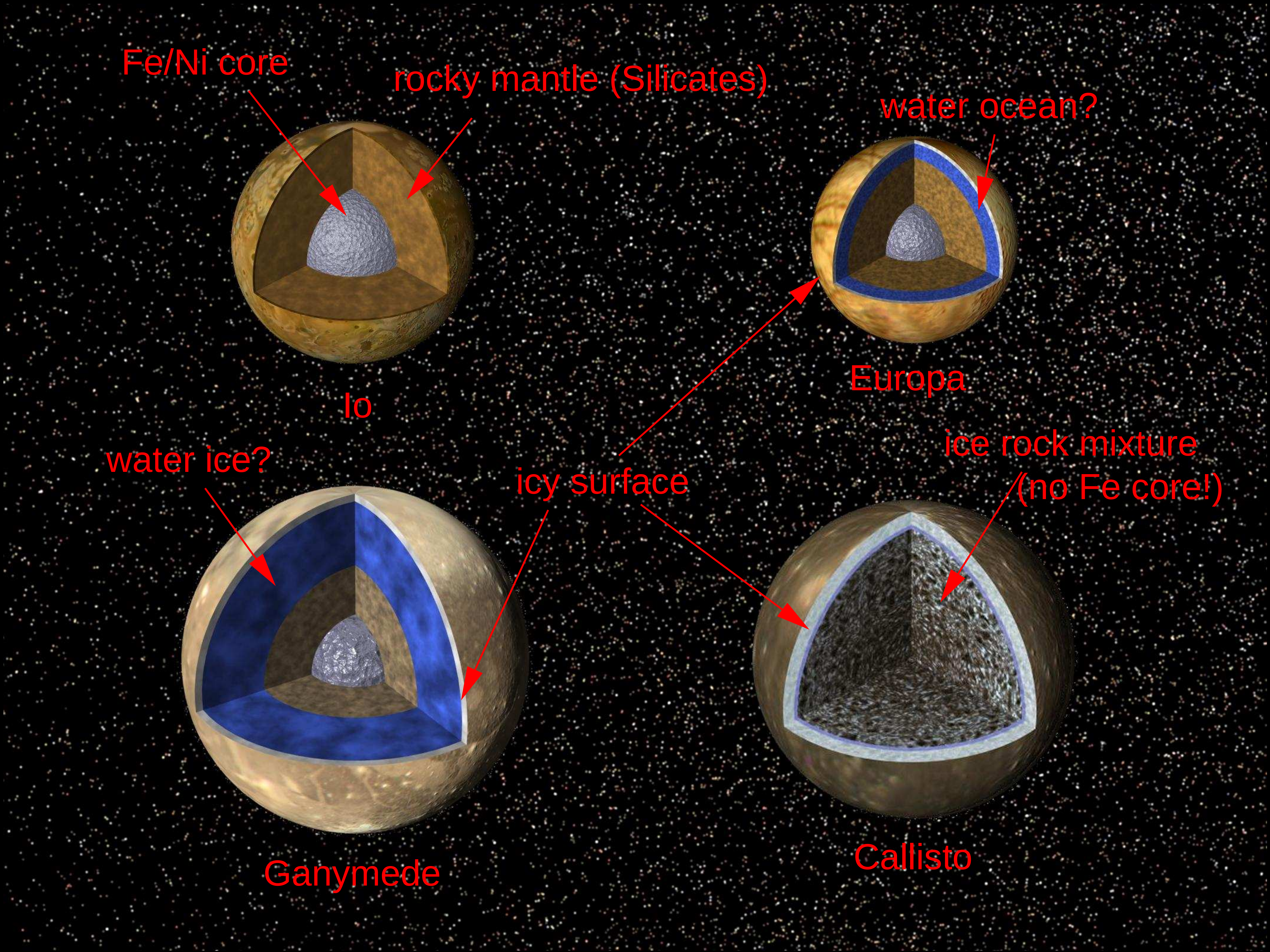


Structure of terrestrial planets:

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

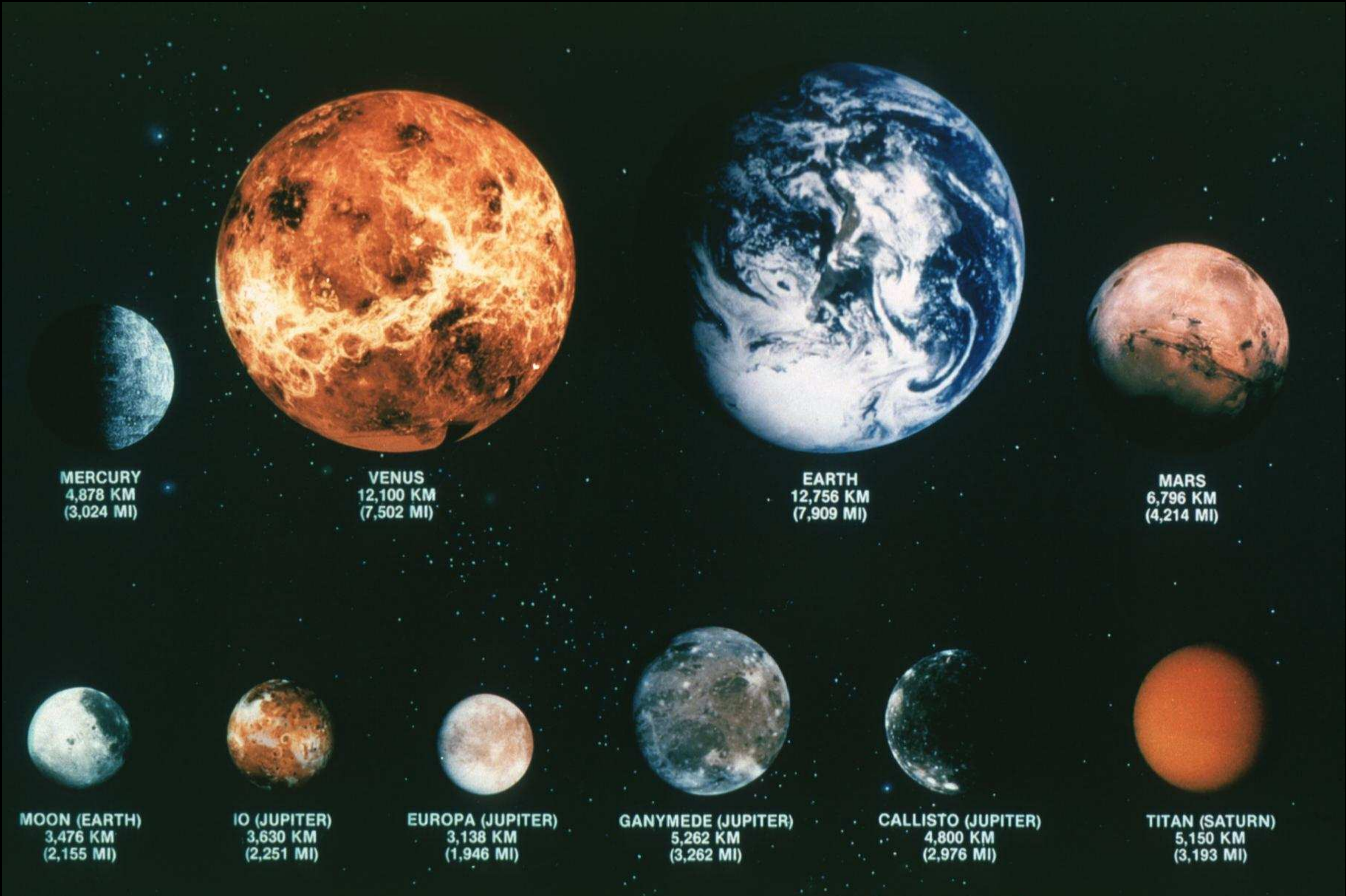
Knowledge of structure important for, e.g.,

- 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 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!

Structure: Gas Giants

Structure of a gas giant from equation of hydrostatic equilibrium:

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

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

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

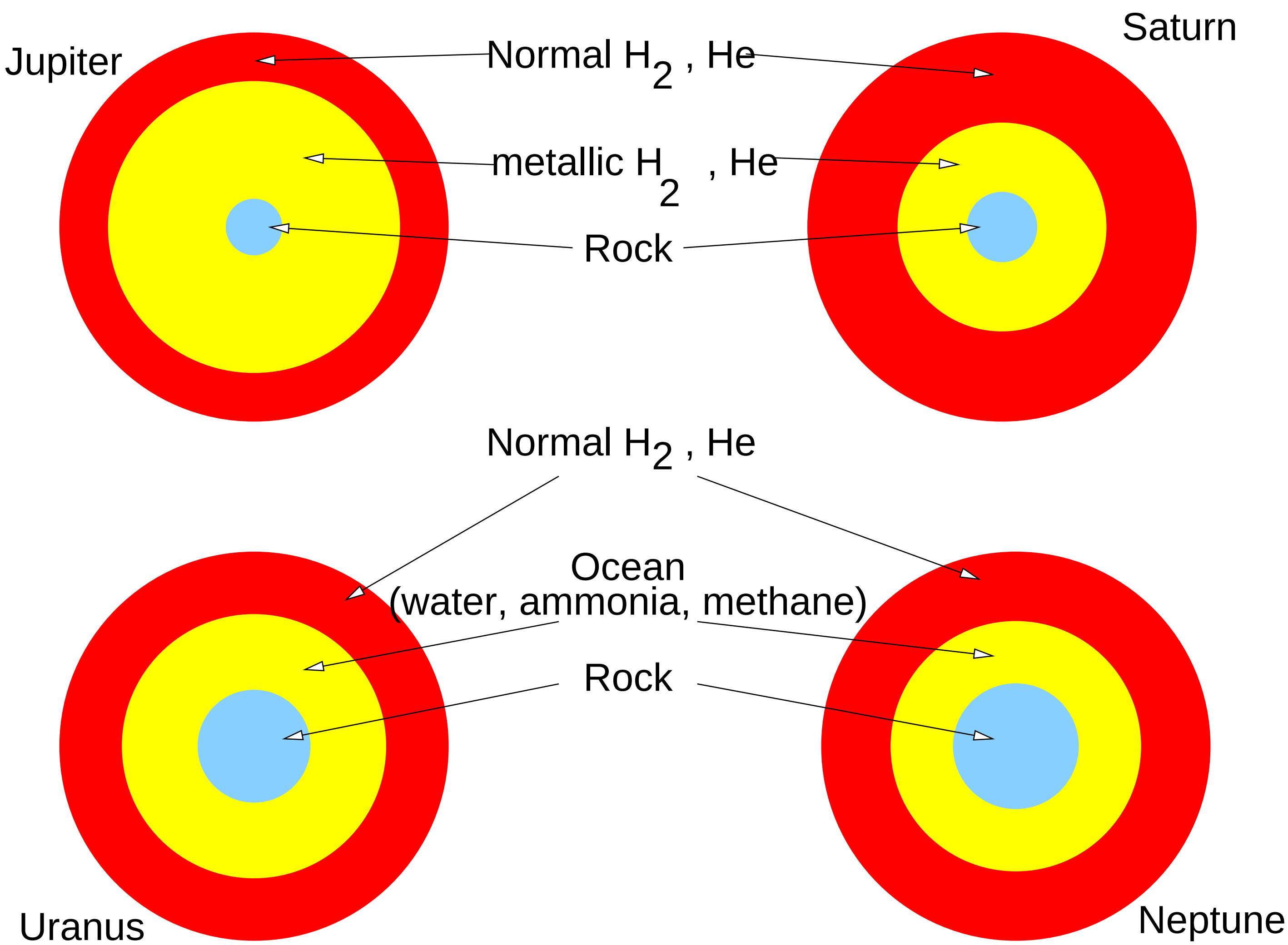
$$P_{\text{central}} = \frac{2\pi}{3} G \langle\rho\rangle^2 R^2$$

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

Structure: Gas Giants



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

To obtain information on the pressure structure of any gravitationally supported static body we can use the *concept of hydrostatic equilibrium*, which we already used for estimating the structure of atmospheres,

$$\frac{dP}{dr} = -\rho(r)g(r)$$

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

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

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

$$M(r) = \int_0^r 4\pi\rho(r)r^2 dr$$

(interpretation: integrate over onion shells of thickness dr and density $\rho(r)$; the mass in each of these shells is $4\pi\rho(r)dr$, 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. Unfortunately, this equation of state is generally much more complicated than for gases 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,

$$M(r) = (4/3)\pi r^3 \langle\rho\rangle$$

such that the equation of hydrostatic equilibrium reads

$$\frac{dP}{dr} = -\langle\rho\rangle^2 G(4/3)\pi r$$

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 $r = R$:

$$\int_0^R \frac{dP}{dr}\, dr = -\int_0^R \langle\rho\rangle^2 G(4/3)\pi r\, dr$$

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 defi nition 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, $P(0) = P_c$. Therefore

$$\int_0^R \frac{dP}{dr}\, dr = P(R) - P(0) = -P(0) =: -P_c$$

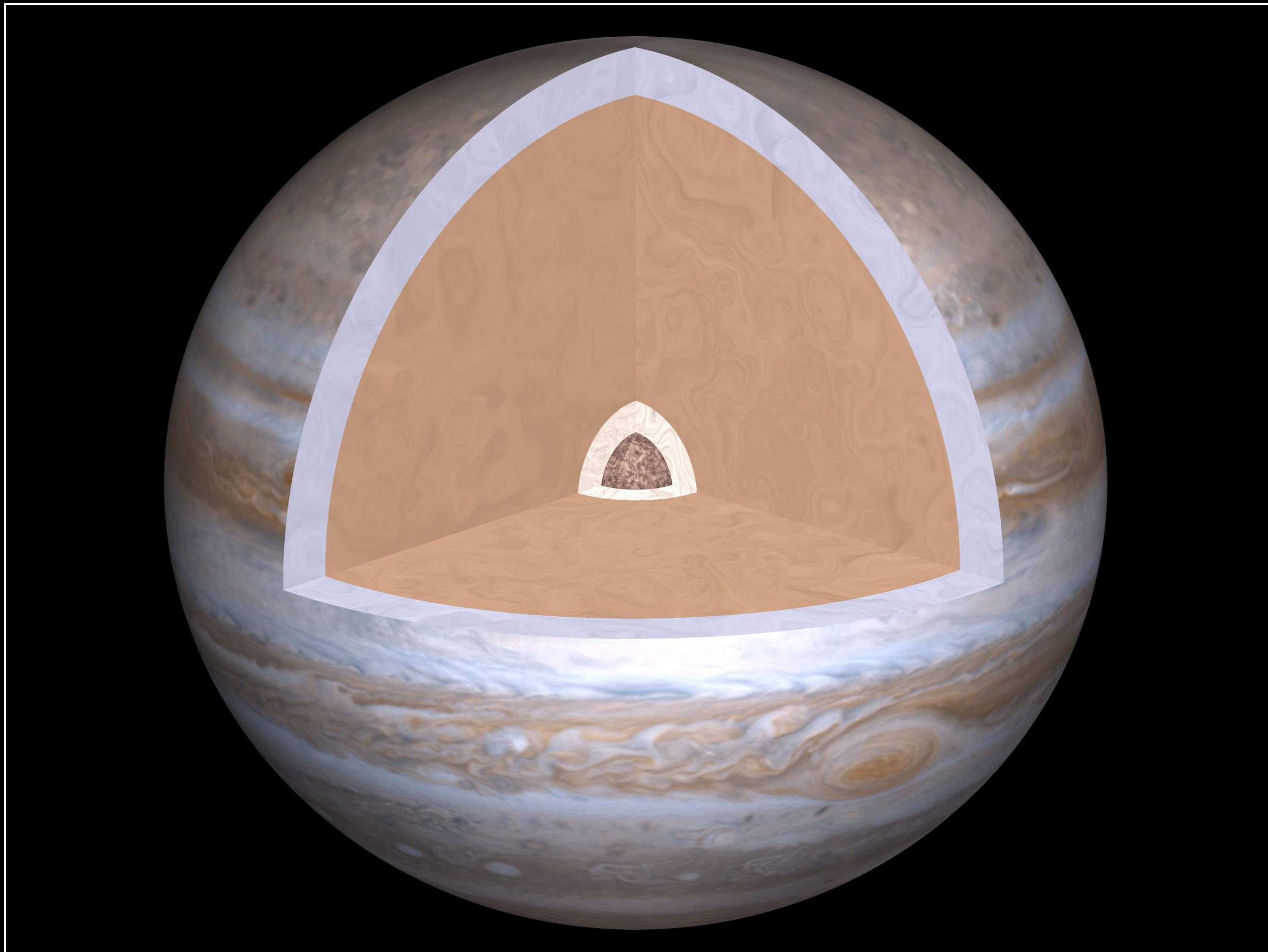
The right hand side of the equation is easily found as well:

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

such that

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

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.



The Interior of Jupiter

Small Solar System Bodies: Asteroids, Comets, and Transneptunians