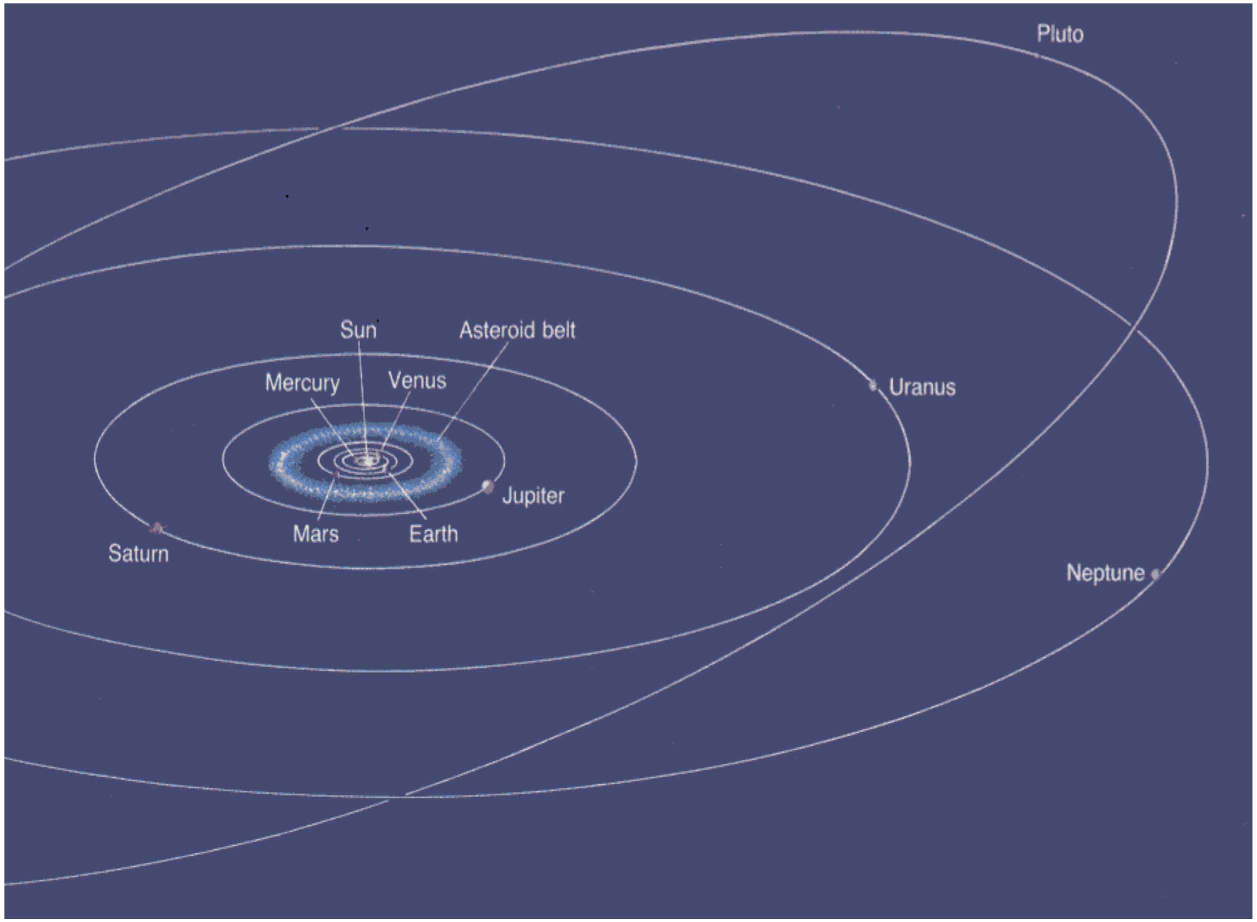
From last week:

In my (astronomical) view there are three necessary (but not sufficient?) requirements for Life in the Universe

- A diversity of atomic species to "enable" complexity.
- Environments that are
    - reasonably warm so that chemical processes occur and
    - reasonably stable over long periods of time for complexity to develop through replication and selection
- Sources of low entropy energy and sinks for high entropy waste energy, i.e. heat flow down a temperature gradient (i.e. an absence of thermal equilibrium)

Chapter 2: Our Solar System

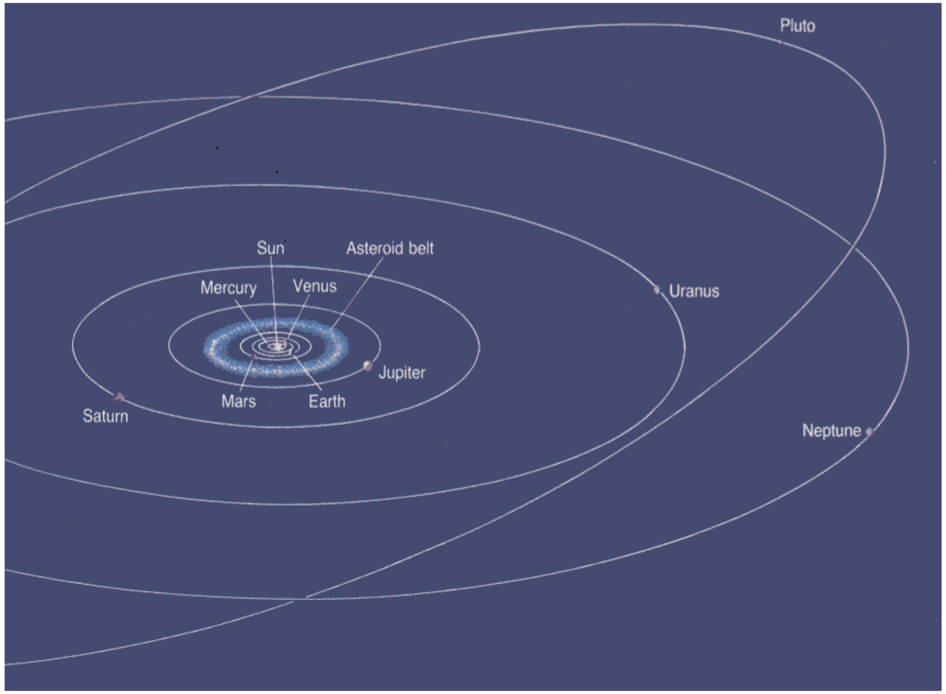
1. The formation of planetary systems, based on our own



99.9% of mass is in the Sun, so the gravitational field is simple $\propto r^{-2}$

Kepler's three laws of planetary motion:

- Orbits are ellipses with Sun at one focus (Kepler 1)
- Relation between period T and semi-major axis A : $T^2 = (4\pi^2/GM_\odot)\, A^3$ (Kepler 2)
- Constant "areal velocity" $\underline{\mathbf{v}} \times \underline{\mathbf{r}}$ (= conservation of angular momentum) (Kepler 3)



Object	ε
Mercury	0.206
Venus	0.007
Earth	0.017
Mars	0.093
Jupiter	0.048
Saturn	0.054
Uranus	0.047
Neptune	0.009
Pluto	0.249
Sedna	0.855
Halley's Comet	0.967
Comet Hale-Bopp	0.995

- In our Solar System, the eccentricities of the planets are very low $\varepsilon < 0.1$ (except for Mercury and Pluto), i.e. the planets have almost circular orbits

$$r_{\mathrm{peri}}=(1-\varepsilon)a \qquad r_{\mathrm{ap}}=(1+\varepsilon)a$$

  So, the variation in solar heating is modest, and close encounters between planets do not happen

- Aligned angular momentum vectors of Sun's spin, all planetary orbits (except Pluto), and (almost all) satellite orbits and most planetary spins

Rotating disks are the natural result of the conservation of angular momentum, since circular orbits have the highest L for a given energy. Easy to show for Keplerian orbits around central mass that:

$$E = -\frac{GMm}{2a}$$

$$L^2 = GMm^2 \frac{b^2}{a}$$

$$= 2m|E|b^2$$

L is maximised for a given E, or E is minimized for a given L, when $b = a$ (given that $b \leq a$)

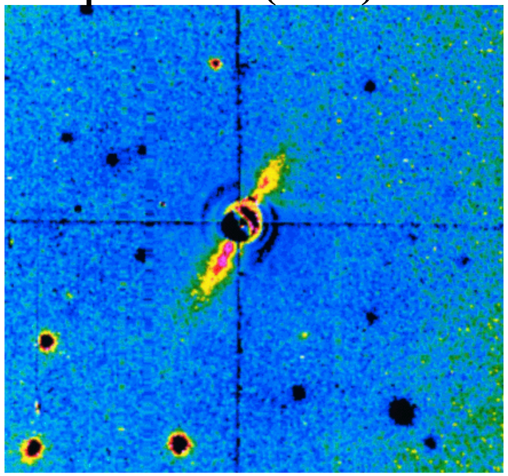
Gas that is losing kinetic energy but conserving angular momentum will naturally settle into a flattened spinning disk with circular orbits.

Circular orbits minimize collisions between gas particles. Collisions → heating → radiative cooling. This is how bulk kinetic energy of the gas can be lost from the gas.

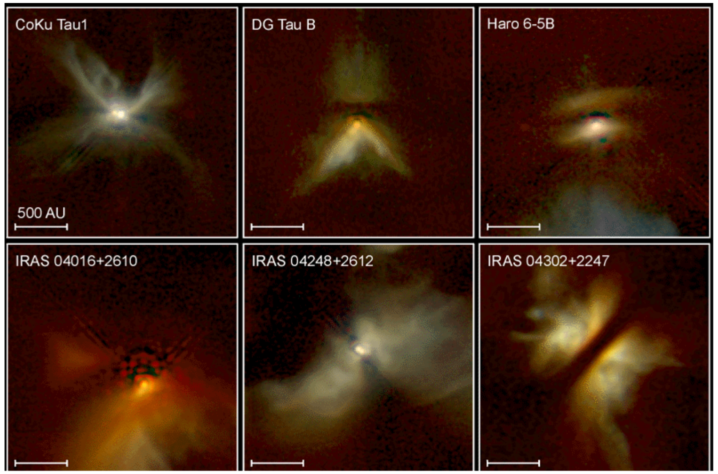
→ Formation of the Solar System out of a flattened rotating gas cloud that also produced the Sun at the center – the **"Solar Nebula" accretion disk.** Idea dates from ~1700's (Kant, Laplace). But no idea if they were common!

Similar disks have now been seen ubiquitously associated with forming and recently formed stars (seen in all of reflected star-light, dust obscuration and thermal emission)

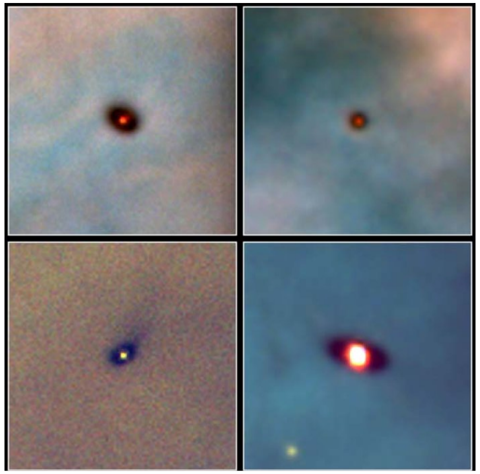
βPic disk (1984)

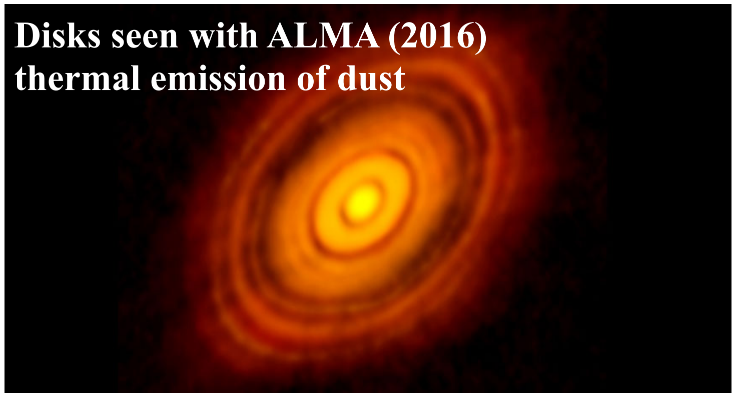


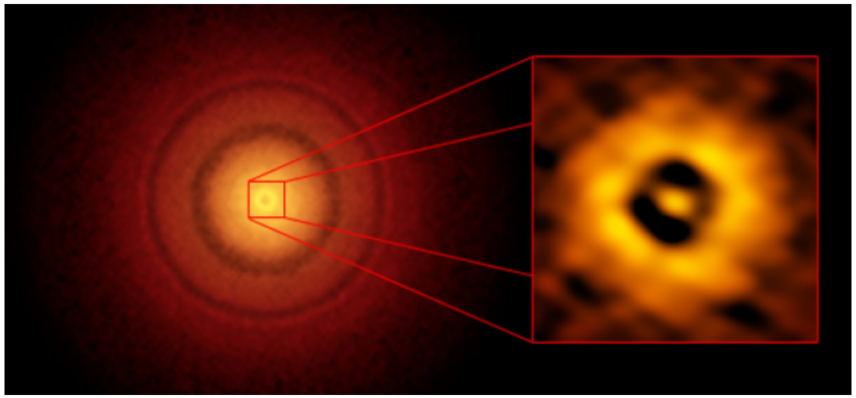
Disks seen edge on around young stars (1995)



Protoplanetary disks shadowing in Orion (1995)




Disks seen with ALMA (2016)
thermal emission of dust



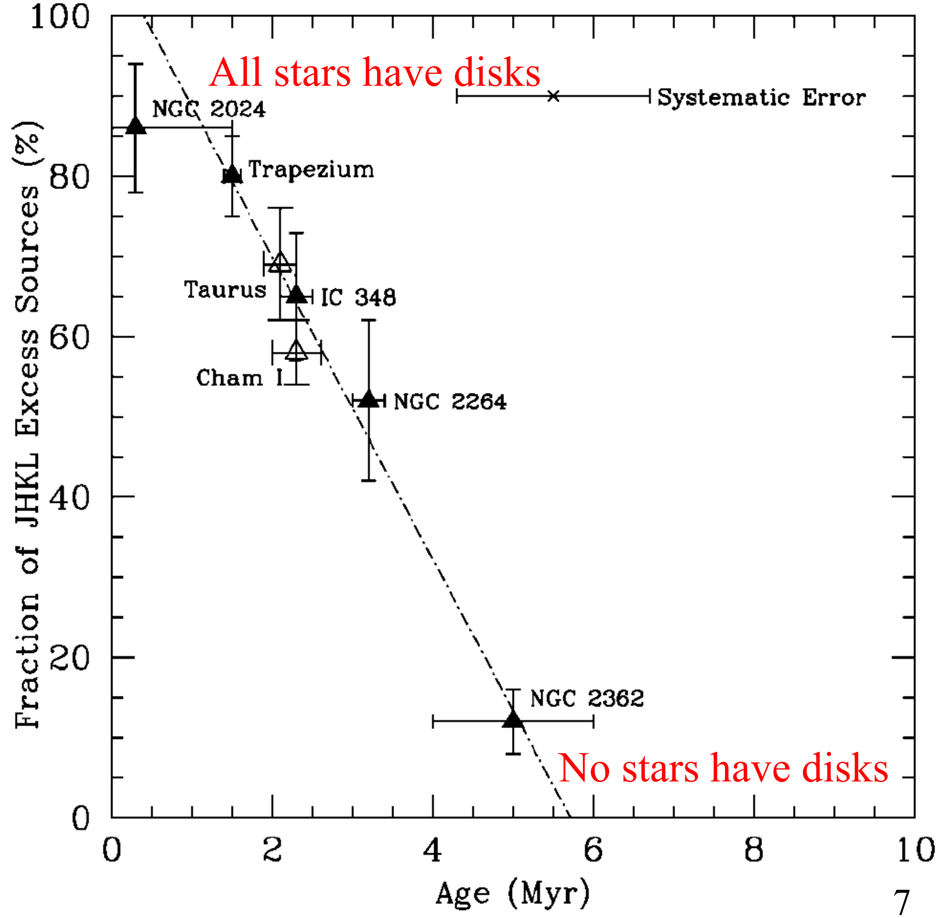
What is the timescale for planetary formation in disks?

The lifetime of the pre-planetary dust+gas disks around young stars seems to be short: they disappear after a few million years.

The "JHKL Excess" is due to thermal emission from hot dust in the disks, which radiates in excess of the star at 1-4 μm.

The fraction of stars showing excess steadily decreases with the age of star clusters (a group of stars formed at the same time)

Haisch+2001



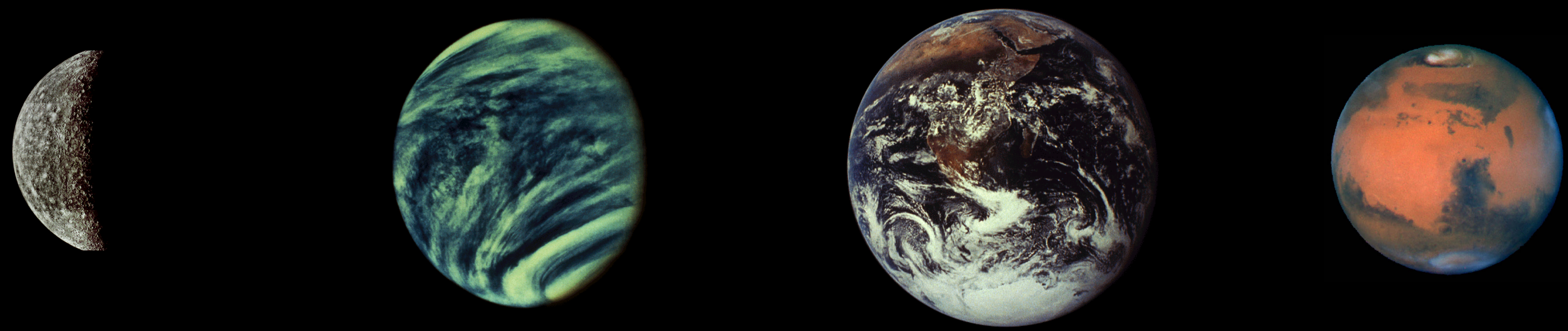
Angular momentum in the Solar System in spins and orbits

$$L_{spin} = \frac{2\pi I}{T_{spin}} \sim \frac{2\pi\ 0.1mr^2}{T_{spin}}; \qquad L_{orb} = \frac{2\pi mR^2}{T_{orb}}$$

- Almost all of the angular momentum in Solar System is in the orbits of the planets (especially Jupiter) and <u>not</u> in the spin of the Sun.
- The specific angular momentum of planetary orbits is ~10^5 larger than for the Sun's spin, and quite similar to that of a presumed progenitor gas cloud.

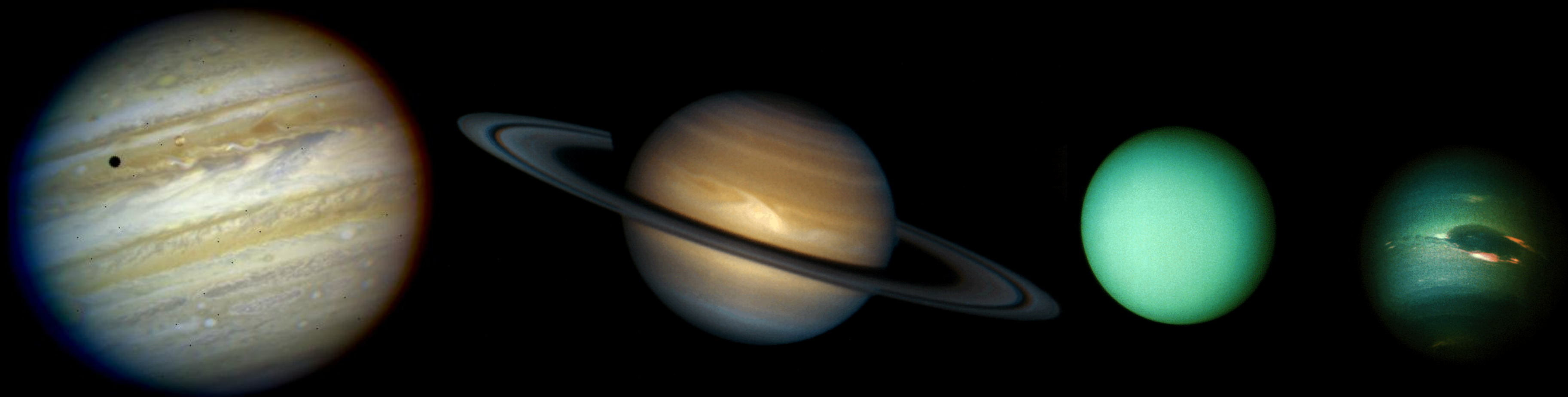
	m (kg)	r (m)	T (s)	L(kg m²/s)	L/m (m²/s)
Sun spin	2.10^{30}	6.10^{8}	$2.6.10^{6}$	$1.7.10^{41}$	8.10^{10}
Earth orbit	6.10^{24}	$1.5.10^{11}$	$3.1.10^{7}$	$2.7.10^{40}$	$4.5.10^{15}$
Jupiter orbit	2.10^{27}	$7.8.10^{11}$	$3.7.10^{8}$	$2.1.10^{43}$	$1.0.10^{16}$
Initial gas cloud	2.10^{30}	$9.5.10^{15}$ Light year	3.10^{15} 10^8 years	10^{47}	5.10^{16}

Conclusion: The material in the Sun must have lost almost all of it's angular momentum during its formation. Not so for planet formation.



Properties of the planets:

- Inner “terrestrial” planets
- Outer “Jovian” or “Gas Giant” planets



Bulk differences between terrestrial and Giant planets

	Terrestrial planets	Gas Giants
Basic form	Rocky	Primarily gas
Orbital distance (R_{earth} = AU)	0.39-1.52	5.2-30.1
"Surface" temperature (K)	200-750	75-170
Mass (M_{Earth})	0.055-1.0	14.5-320
Radius (r_{earth})	0.38-1.0	3.9-11.2
Mean density (gm cm^{-3})	3.95-5.52	0.7-1.64
Rotation period	24h - 243d	9.8h - 19.2h

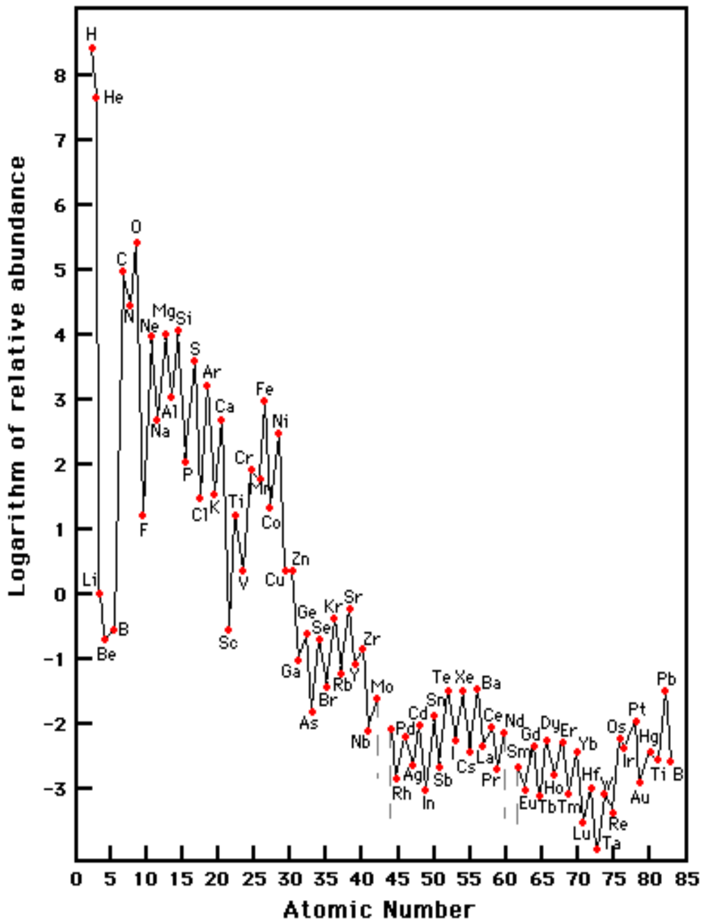
Density differences due to differences in bulk composition

Object	Distance(AU)	Density (g/cm^3)	Bulk composition, dominant species
Mercury	0.4	5.4	iron, nickel, silicates
Venus	0.7	5.4	silicates, iron, nickel
Earth	1.0	5.5	silicates, iron, nickel
Moon	1.0	3.3	silicates
Mars	1.4	3.9	silicates, iron, sulfur
Jupiter**	5.2	1.3	H, He
Io		3.4	silicates
Europa		3.0	silicates, water, ice
Ganymede		1.9	water ice, silicates
Callisto		1.8	water ice, silicates
Saturn	9.6	0.7	H, He
Titan		1.8	water ice, silicates
Uranus	19.2	1.2	ices, H, He
Neptune	30.1	1.6	ices, H, He
Triton		2.1	silicates, ices
Pluto	39.4	2.1	silicates, ices

Increasingly volatile substances

** Note the similar gradient within the mini-system of e.g. Jupiter's major moons

Remember: also big differences with abundances in the Universe as a whole



Relative abundance of the elements in the Universe. Abundances are scaled so that silicon (Si) = 10000. From Mason (1968).

	Sun**	Earth	Human
H	70.537	0.000	9.271
He	27.505	0.000	0.000
O	0.967	29.293	63.222
C	0.307	0.000	19.149
Ne	0.171	0.000	0.000
N	0.109	0.000	5.106
Mg	0.074	12.303	0.000
Si	0.065	14.354	0.000
Fe	0.130	34.859	0.000
S	0.099	3.750	1.264
Ar	0.009	0.000	0.000
Al	0.006	1.088	0.000
Ca	0.006	1.084	1.398
Na	0.004	0.556	0.000
Ni	0.009	2.376	0.000
Cr	0.001	0.248	0.000
P	0.001	0.091	0.612

Fractional abundances by <u>mass</u>

**The Sun is typical of other stars, and gas in the Galaxy, and indeed of the Universe as a whole

The formation of the central Sun must have involved the loss of angular momentum (transport of angular momentum outwards and of mass inwards):

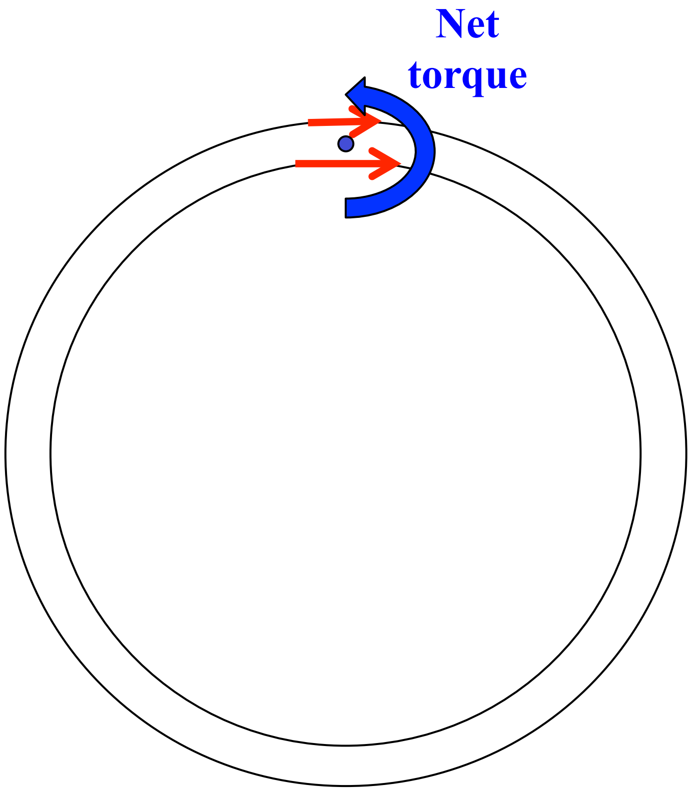
→ Accretion disk physics

The formation of the planets must have involved some process(es) that were *chemical* specific – i.e. not simply gravity (or most other astrophysical processes).

→ How to grow dust grains from μm size (ubiquitous in interstellar clouds) to 10^{4+} km size of planets?

Angular momentum transport in accretion disks

A disk of material orbiting around a (dominant) central object has differential rotation $\omega \propto r^{-3/2}$ or $v \propto r^{-1/2}$



Torques acting on material in the disk transfer angular momentum from fast rotating inner parts to slower rotating outer parts of disks: Torques arise from the differential orbital velocities, via:

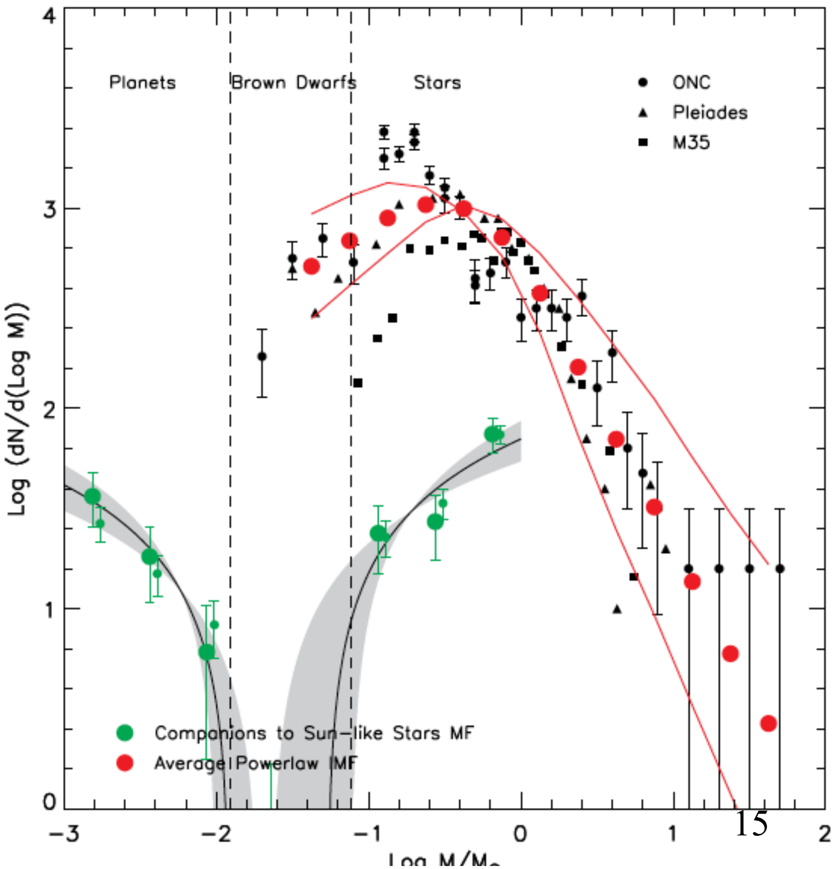
1. Magnetic fields anchored to ionized material.
2. Density inhomogeneities sheared to spiral waves, producing gravitational torques
3. Friction due to convective (vertical) motions in disk

Net effect is for mass to be transported inwards and angular momentum outwards. *This is an accretion disk.* Accretion disks are often encountered in astrophysics (e.g. accretion onto a black hole in an AGN)

Comment: Are we forming planets, stars or brown dwarfs**?

** brown dwarf = "failed star"

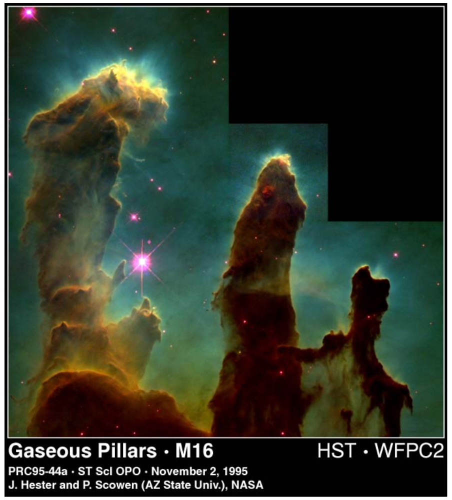
- More than half of all stars in the solar neighbourhood are in binary systems.
- "Stars" less than 0.08 $M_\odot$ (= 80 $M_{Jupiter}$) never ignite H fusion and never become a star → "brown dwarf". They cool and fade
- What is the difference between making a brown dwarf and a planet?
- There appears to be a dearth of "companions" formed with 0.01 to 0.1 $M_\odot$. This likely reflects different formation processes.
  above: bulk gravitational instability
  below: growth of dust grains, i.e. "planet formation"

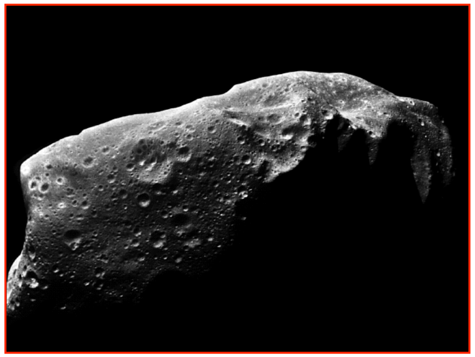


μm dust grains are ubiquitous in gas in the galaxy. Typical $m_{dust}/m_{gas} \sim 1\%$ by mass. A significant fraction of elements above H and He are in dust for $T_{gas} < 1000$ K.

Three phases during the formation of the bodies in the Solar System ($\times 10^4$!)

- Initial growth of dust grains (μm to cm size)
- Formation of "planetessimals" (cm to km size)
- Growth of planetessimals to make (small number) of large planets (km to 10^4 km)





growth

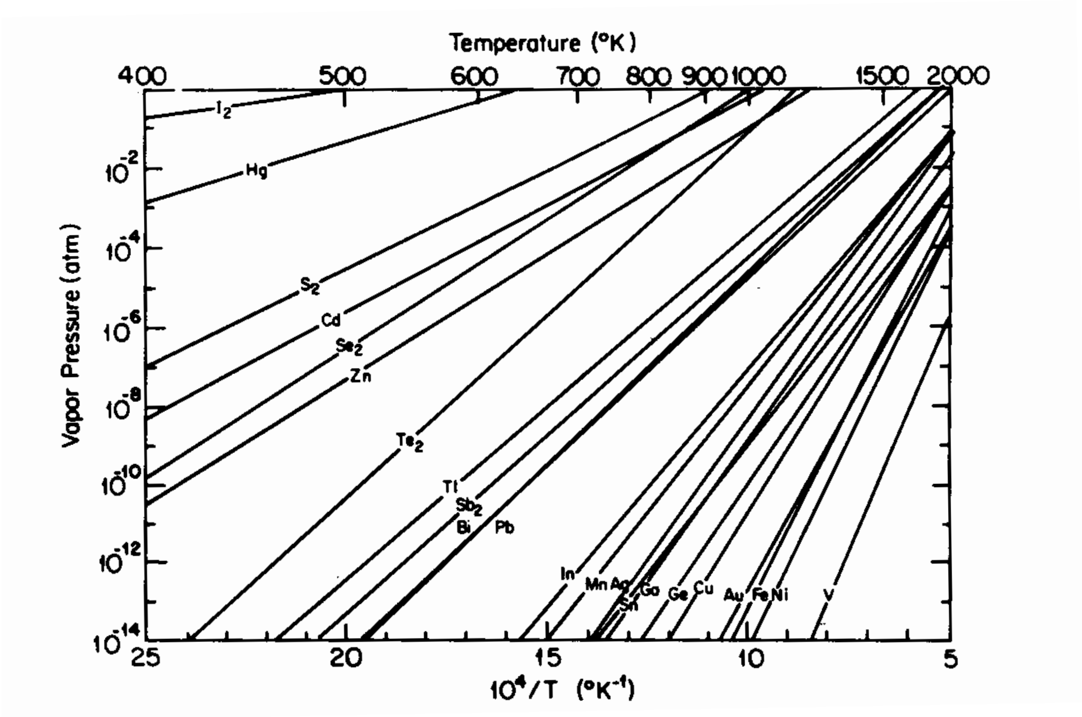
Step 1: Condensation and other non-gravitational effects

Different atomic/molecular species will condense out of the gaseous phase at different temperatures.

Antoine's Law

Vapor pressure P above a surface drops exponentially at temperatures below $T \sim \lambda/R$

$$\boxed{\ln P = \frac{-\lambda}{RT} + c}$$



P_T = 1 atm		$P_T = 6.6 \times 10^{-3}$ atm	
Compound or Element	T (°K)	Compound or Element	T (°K)
$MgAl_2O_4$	2050	$CaTiO_3$	1740
$CaTiO_3$	2010	$MgAl_2O_4$	1680
Al_2SiO_5	1920	Al_2SiO_5	1650
Ca_2SiO_4	1900	Fe	1620
$CaAl_2Si_2O_8$	1900	$CaAl_2Si_2O_8$	1620
$CaSiO_3$	1860	Ca_2SiO_4	1600
Fe	1790	$CaSiO_3$	1580
$CaMgSi_2O_6$	1770	$CaMgSi_2O_6$	1560
$KAlSi_3O_8$	1720	$KAlSi_3O_8$	1470
Ni	1690	$MgSiO_3$	1470
$MgSiO_3$	1670*	SiO_2	1450
SiO_2	1650	Ni	1440
Mg_2SiO_4	1620*	Mg_2SiO_4	1420
$NaAlSi_3O_8$	1550	$NaAlSi_3O_8$	1320
$MnSiO_3$	1410	$MnSiO_3$	1240
Na_2SiO_3	1350	MnS	1160
K_2SiO_3	1320	Na_2SiO_3	1160
MnS	1300	K_2SiO_3	1120
Cu	1260	Cu	1090
Ge	1150	Ge	970
Au	1100	Au	920
Ga	1015	Ga	880
Sn	940	Zn_2SiO_4	820
Zn_2SiO_4	930	Sn	806
Ag	880	Ag	788
ZnS	790	ZnS	730
FeS	680	Fes	680
Pb	655	Pb	570
CdS	625	CdS	570
Bi	620	$PbCl_2$	535
$PbCl_2$	570	Bi	530
Tl	540	Tl	475
In	400	Fe_3O_4	400
Fe_3O_4	400	In	360
H_2O	260	H_2O	210
Hg	196	Hg	181

The “Condensation Sequence” in the young Solar System

Temperature	Condensate	Where
1500	Metal Oxides	Mercury
1300	Iron and Nickel	
1200	Silicates	
1000	Feldspars	Venus
700	Trolite	Earth, Mars
“snow line”		
175	H_2O ice	Jupiter, Saturn
150	NH_3 ice	Uranus, Neptune
120	CH_4 ice	

Condensation is a chemical process, not a gravitational one, and it leads to changes in the chemical composition relative to the surrounding gas

Pressure effects in the disk will enhance grain growth

- Gas in an accretion disk feels (at a low level) a radial force from the pressure gradient in the disk, as well as from the dominant gravity
- Effect of the pressure relative to gravity is much smaller for large and/or dense grains relative to gas atoms/molecules

$$\frac{F_G}{F_P} \propto \frac{m}{r^2} \propto \rho r$$

Dust grains therefore migrate towards the center (i.e. gas orbits at slightly below Keplerian speed because of non-gravity force, less felt by dust, exerting drag force on dust).

Dust grains also sink towards the plane of the disk since they experience less vertical pressure in the disk.

Both effects enhance grain-grain collisions and sticking together of grains

Step 2: Gravitational effects

Both condensation and non-gravitational accretion (i.e. collisions between leading to sticking) will be surface effects (and therefore $\propto r^2$).

Purely gravitational effects will depend on the mass ($\propto r^3$) and will become more important as r increases.

How to determine this gravitational growth? *(Don't worry too much for non-physicists)*

Consider stability of spinning disk of material of surface density Σ that has a small density perturbation in the form of a wave-like disturbance:

$$\boxed{\Delta = \frac{\delta\Sigma}{\Sigma} = \Delta_0 \exp\left(i(\mathbf{k}.\mathbf{r} - \omega t)\right)}$$

real $\omega \rightarrow$ sound-wave oscillations

imaginary $\omega \rightarrow$ exponential collapse

It turns out that the dispersion relation for ω, depends on the wave number $k = 2\pi/\lambda$, the angular rotation speed Ω, the surface density Σ, and the sound speed c_s.

$$\boxed{\Delta = \Delta_0 \exp\left(i(\mathbf{k.r} - \omega t)\right)}$$

$$\omega^2 = k^2 c_s^2 + \Omega^2 - 2\pi G \Sigma k$$

Oscillation or collapse depends on the sign of the RHS.
Collapse requires imaginary ω, i.e. negative ω^2.

(1) When the sound speed c_s is <u>negligible</u>, (i.e. there is insignificant "pressure") then collapse will occur on all scales up to some maximum size λ given by the surface density and rotation rate, producing objects upto mass $M_{\max}$

$$2\pi G \Sigma k \geq \Omega^2$$

$$\lambda \leq \lambda_{\max} \sim 4\pi^2 G \Sigma \Omega^{-2}$$

$$M_{\max} = 16\pi^4 G^2 \Sigma^3 \Omega^{-4}$$

Does it work? In the proto-Solar System, we expect $\lambda_{\max} \sim 10^4$ km from expected Σ (and Ω). This is about right for producing collapsed objects of about 10km size (planetessimals).

Aside: note that $M_{\max}$ expected to vary as Σ^3.

(2) Note that if the sound speed is <u>not</u> negligible, then the analysis reverts to the classic Jeans analysis: small scale fluctuations do <u>not</u> grow on interesting length scales: i.e. high c_s sound speed *stabilizes* the disk

$$2\pi G\Sigma k \geq k^2 c_s^2$$

$$\lambda \geq \frac{c_s^2}{G\Sigma} \sim 10^{18} m \sim 10^7 AU$$

The threshold c_s is given by

$$k^2 c_s^2 \sim \Omega^2$$

... and we get collapse on scales above

$$\lambda \sim c_s \frac{2\pi}{\Omega} \sim c_s T_{rot}$$

Makes sense: istance pressure wave travels in rotation period

i.e. a gas composed of slow moving massive particles will be more gravitationally unstable than a disk of lower mass particles with higher speeds

Sound speed is given by the mass of the particles since collisions lead to equipartition of energy (e.g. 10^{-7} ms^{-1} for milligram masses).

$$c_s \sim v = \left(\frac{3kT}{m}\right)^{1/2}$$

Bottom line: Growth of grains through condensation and non-gravitational sticking increases the mass, <u>reducing</u> the sound speed c_s, and allowing the material in the disk to become gravitationally unstable to produce 10 km-sized bodies

Do the timescales work?

Growth time for collapse is given by $\tau \sim \omega^{-1}$

This is about 10^6 years for $\lambda \sim 10^4$ km and $c_s \sim 10^{-7}$ ms^{-1}

Again, this is just about OK given the observed constraints on the lifetime of disks discussed earlier.

Step 3: Clearing the Nebula

- Assembly of 10 km planetessimals into planets through collisions
- Removal of remaining planetessimals and removal of gas (ejection through close encounters with planets and solar pressure respectively)

The end result is likely to be rather stochastic and unpredictable

Note:

- large collisions at late epochs between "proto-planets" are likely
- Transport of volatile rich objects from beyond "snow-line" in outer Solar System into inner Solar System is possible.

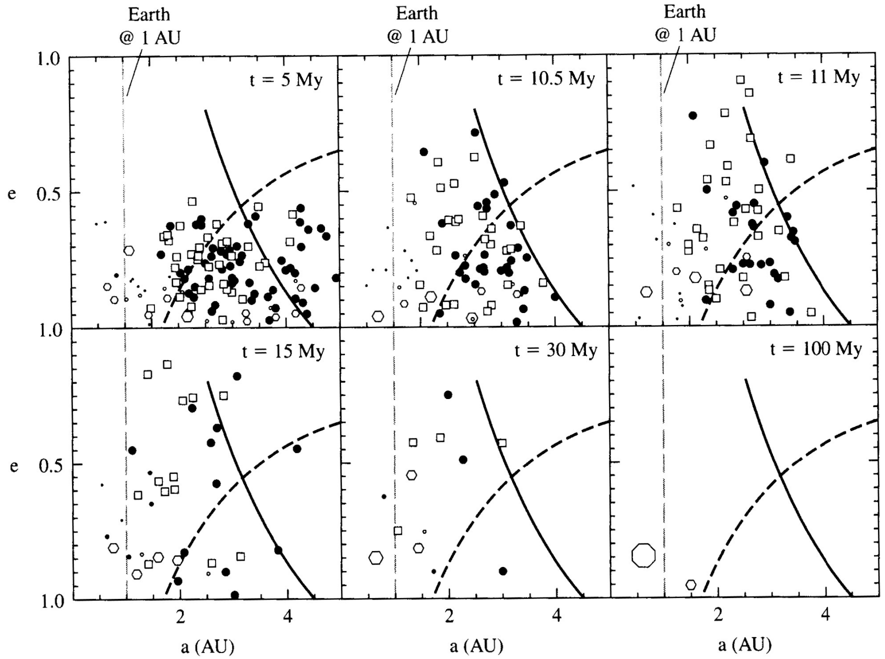
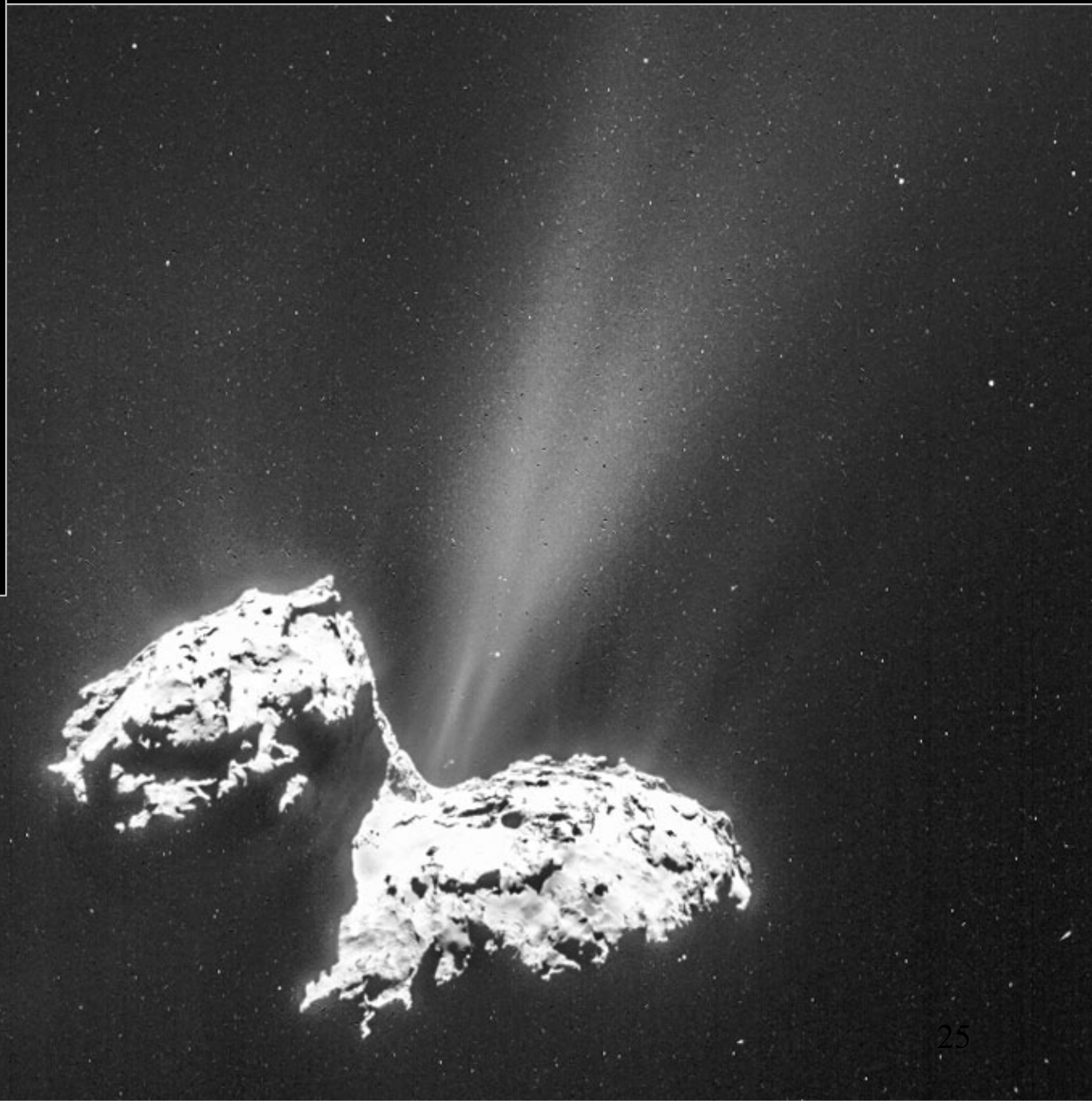


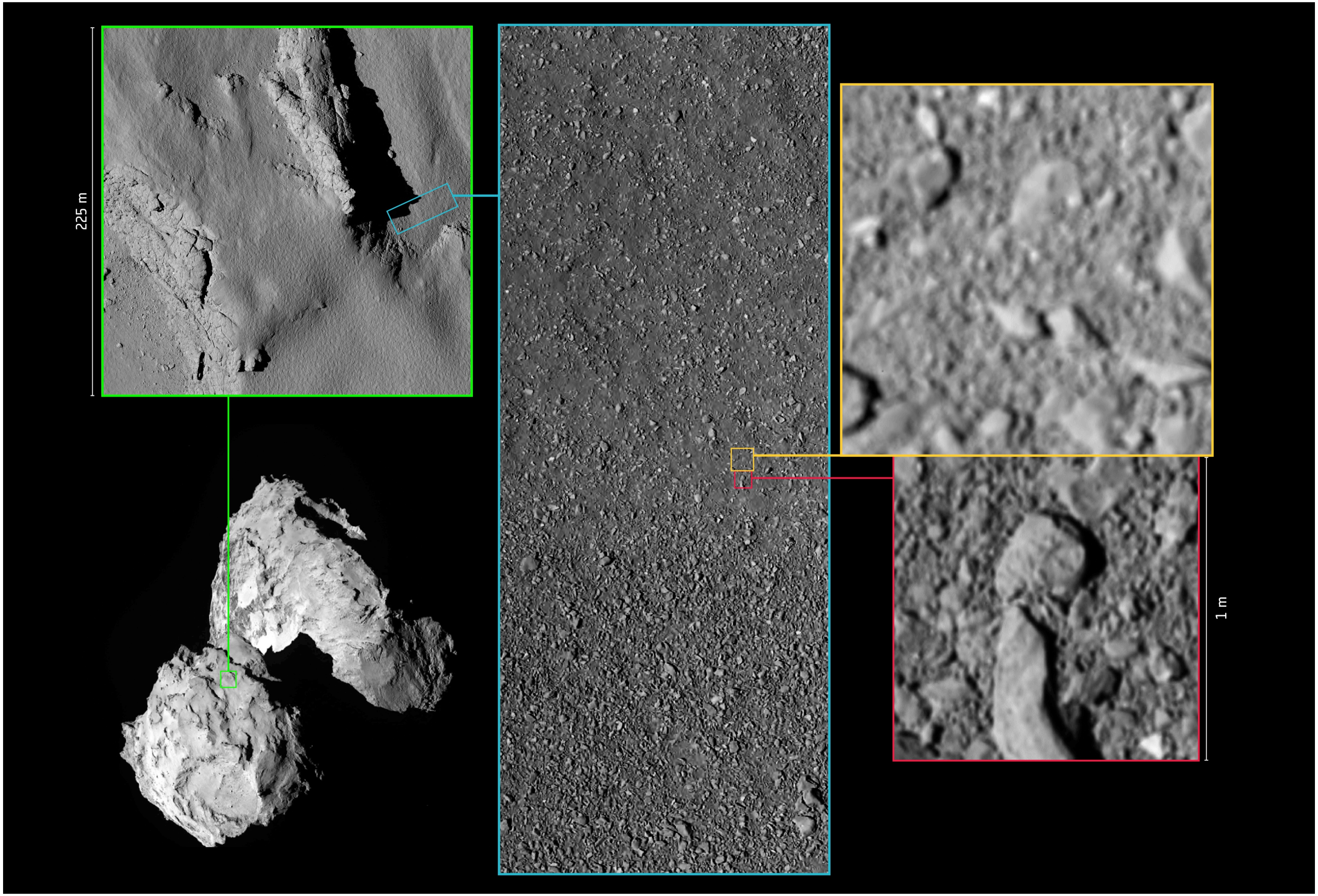
FIGURE 6.6 Simulation of the assembly of the terrestrial planets from lunar- to Mars-sized fragments. Each panel shows the distribution in orbital semimajor axis and eccentricity (Chapter 1), and the series of panels are a progression in time (labeled in millions of years). Planetesimals grow by collision, and their increasing mass is schematically shown by the size of the circles. The dotted line illustrates the realm (to the right of the line) where Jupiter's gravity is so large that rapid ejection of planetesimals occurs. To simplify the calculation, Jupiter's influence is "inserted" into the picture beginning in the second panel. The location of water-bearing planetesimals is shown as being at 2.5 AU and beyond. At the end of this simulation two terrestrial planets are formed.

Comets and asteroids are surviving planetessimals still in a relatively pristine state – being volatile-rich and less volatile respectively

Asteroid Vesta



Comet 67P

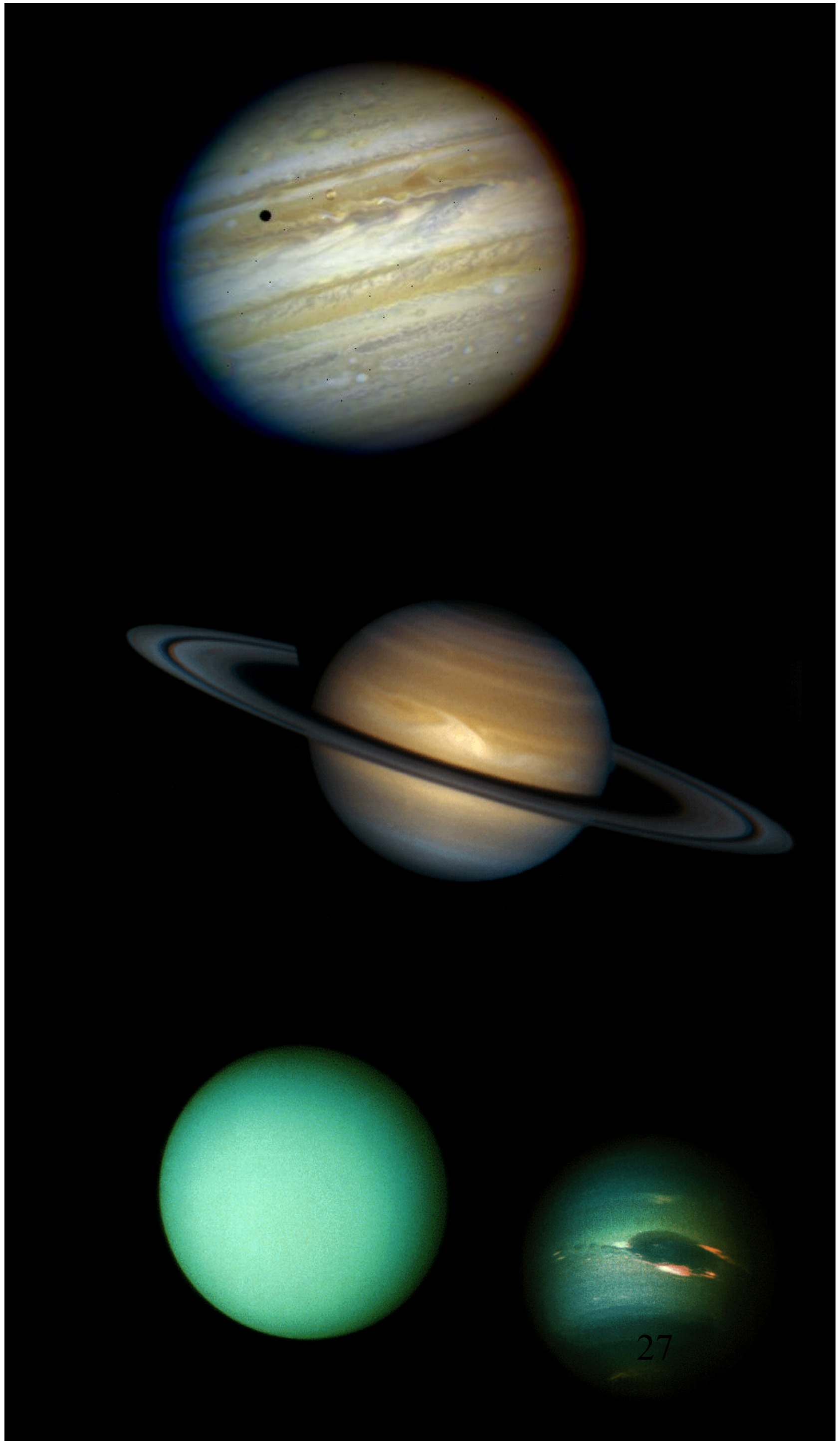

225 m
1 m

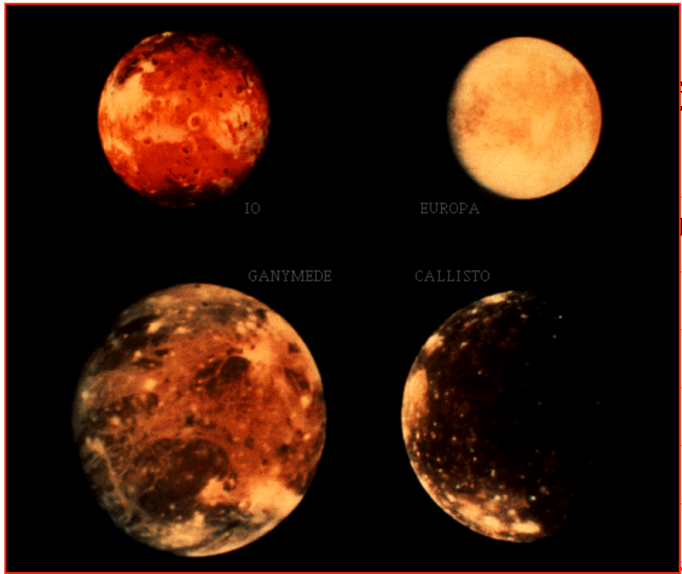
Outer Planets

Giant planets were able to gravitationally attract substantial amounts of H and He gas from the Nebula. This requires an initial solid core of $>5\ M_{earth}$.

Major moons of Jovian planets formed *in situ* out of spinning disk of gas around then (Solar System formation in miniature) and also display gradients of composition etc. But some smaller moons were likely late captures.

Note: it is not clear that Jupiter and Saturn do actually possess a rocky core (esp. Saturn). This plus concerns about timescale has led to alternative scenario of coherent collapse of gas cloud (but what about non-solar abundance?)





eflect differences in bulk composition

	)	Density (g/cm^3)	Bulk composition
		5.4	iron, nickel, silicates
		5.4	silicates, iron, nickel
		5.5	silicates, iron, nickel
		3.3	silicates
		3.9	silicates, iron, sulfur
Jupiter	5.2	1.3	H, He
Io		3.4	silicates
Europa		3.0	silicates, water, ice
Ganymede		1.9	water ice, silicates
Callisto		1.8	water ice, silicates
Saturn	9.6	0.7	H, He
Titan		1.8	water ice, silicates
Uranus	19.2	1.2	ices, H, He
Neptune	30.1	1.6	ices, H, He
Triton		2.1	silicates, ices
Pluto	39.4	2.1	silicates, ices

Formation of the Earth's Moon (with unusually large mass ratio 1:83)

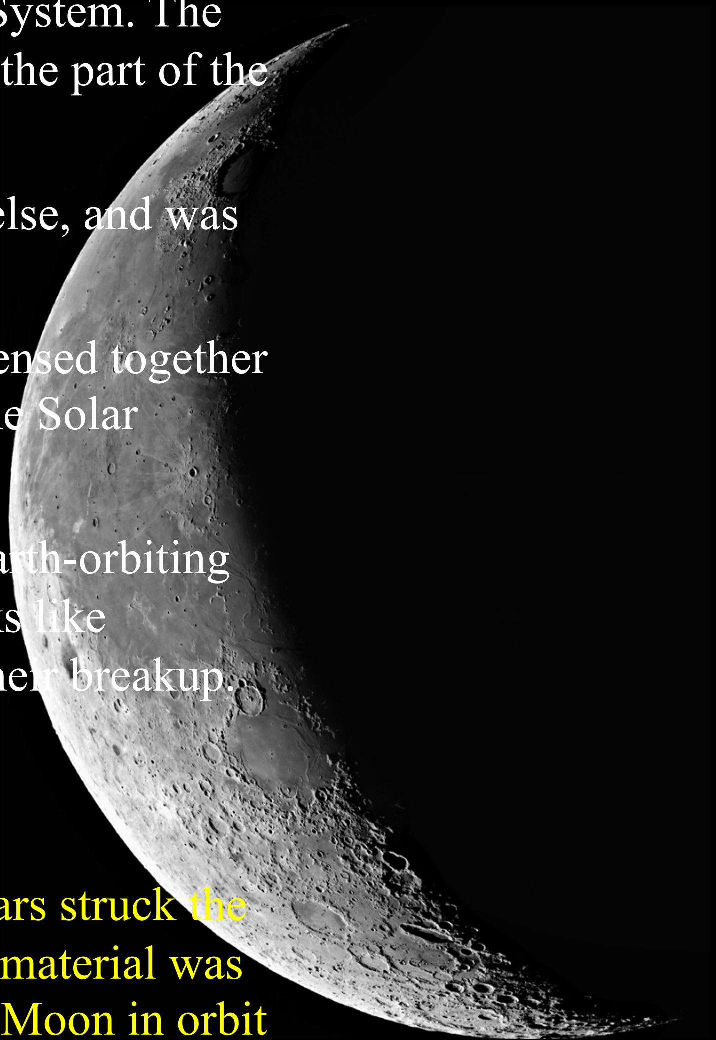
Historical ideas:

The Fission Theory: The Moon was once part of the Earth and somehow separated from the Earth early in the history of the Solar System. The present Pacific Ocean basin was the most popular site for the part of the Earth from which the Moon came.

The Capture Theory: The Moon was formed somewhere else, and was later captured by the gravitational field of the Earth.

The Condensation Theory: The Moon and the Earth condensed together as a binary system from the original nebula that formed the Solar System.

The Colliding Planetesimals Theory: The interaction of earth-orbiting and Sun-orbiting planetesimals (very large chunks of rocks like asteroids) early in the history of the Solar System led to their breakup. The Moon condensed from this debris.



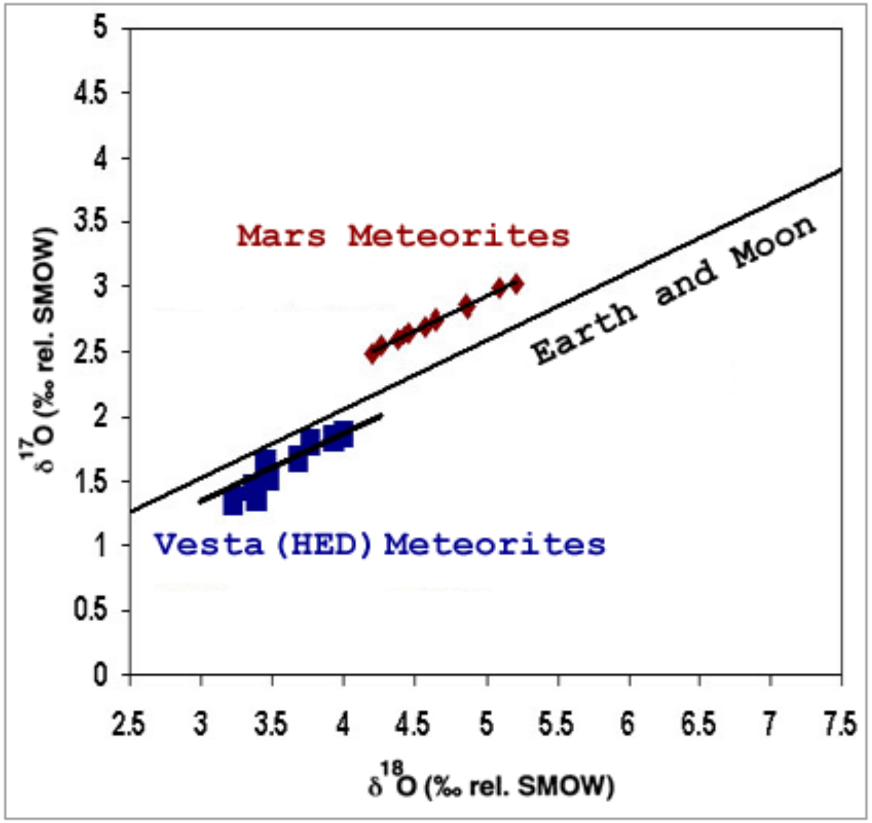
Now almost universally accepted:

The Ejected Ring Theory: A "planetesimal" the size of Mars struck the earth, ejecting large volumes of matter. A disk of orbiting material was formed, and this matter eventually condensed to form the Moon in orbit around the Earth.

Why?

- There is an imperfectly understood but clearly defined gradient in $^{16}O/^{17}O/^{18}O$ within the Solar System, Moon has identical Oxygen isotopic ratios to Earth (c.f. Mars and Vesta).
- But, the chemical composition is different from the Earth as a whole, being more similar to just the Earth's crust without the Fe and Ni that is in Earth's core

Currently accepted idea: impact with Mars-sized body that formed at the same distance from the Sun (perhaps in L4/L5 point??)



PSRD graphic

Age of solid bodies: Radioactive dating of rock solidification

Radioactive decay producing "daughter" ${}^{i}D$ from "parent" ${}^{k}P$ (and no other source)

$$\frac{d\,{}^{i}D}{dt} = -\frac{d\,{}^{k}P}{dt} = \lambda\ {}^{k}P$$

Surviving ${}^{k}P$:

$${}^{k}P = {}^{k}P_0\, e^{-\lambda t}$$

Number of ${}^{i}D$ produced from radioactivity

$${}^{i}D_r = {}^{k}P_0\,(1 - e^{-\lambda t}) = {}^{k}P(e^{\lambda t} - 1)$$

Total number of ${}^{i}D$ present:

$${}^{i}D = {}^{i}D_0 + {}^{i}D_r = {}^{i}D_0 + {}^{k}P\left(e^{\lambda t} - 1\right)$$

Divide by another (stable and not-produced) isotope of D. If the P/D ratio varies within the rock due to initial chemical inhomogeneities, then the slope of the line gives the age (in terms of decay constant λ)

$$\boxed{\frac{{}^{i}D}{{}^{j}D} = \frac{{}^{i}D_0}{{}^{j}D} + \frac{{}^{k}P}{{}^{j}D}\left(e^{\lambda t} - 1\right)}$$

Half-lives:	$^{87}Rb \rightarrow {}^{87}Sr$	4.99×10^{10} yrs
$\tau = 0.693\lambda^{-1}$	$^{232}Th \rightarrow {}^{208}Pb$	1.39×10^{10} yrs
	$^{238}U \rightarrow {}^{206}Pb$	4.50×10^{9} yrs
	$^{235}U \rightarrow {}^{207}Pb$	7.13×10^{8} yrs
	$^{147}Sm \rightarrow {}^{143}Nd$	10.6×10^{10} yrs

Note: the method relies on

(a) The *presence* of chemical inhomogeneities within the sample;

(b) The *absence* of any initial isotopic inhomogeneities;

(c) atoms remaining in place $\rightarrow$ it works only after solidification of rock, i.e. the "age" of the rock is the time since solidification

2 year old rock (Hawaiian lava) and 4.5 billion year old rock (meteorite)

Meteorites: $4.55 \pm 0.01 \times 10^{9}$ yr

Moon (Apollo 17) 4.4×10^{9} yr

Consistent with estimate of age of Sun, i.e. well less than the solar lifetime $\sim 1.0 \times 10^{10}$ yr, (and the Moon forming about 100 million years after the Earth).

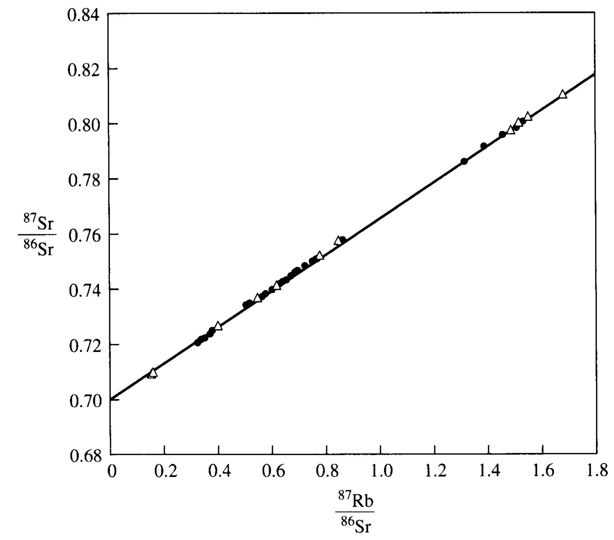
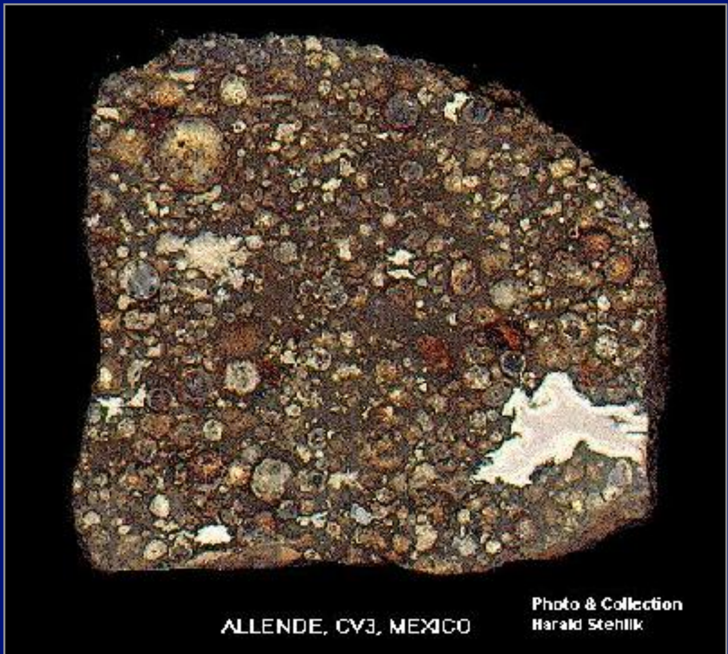


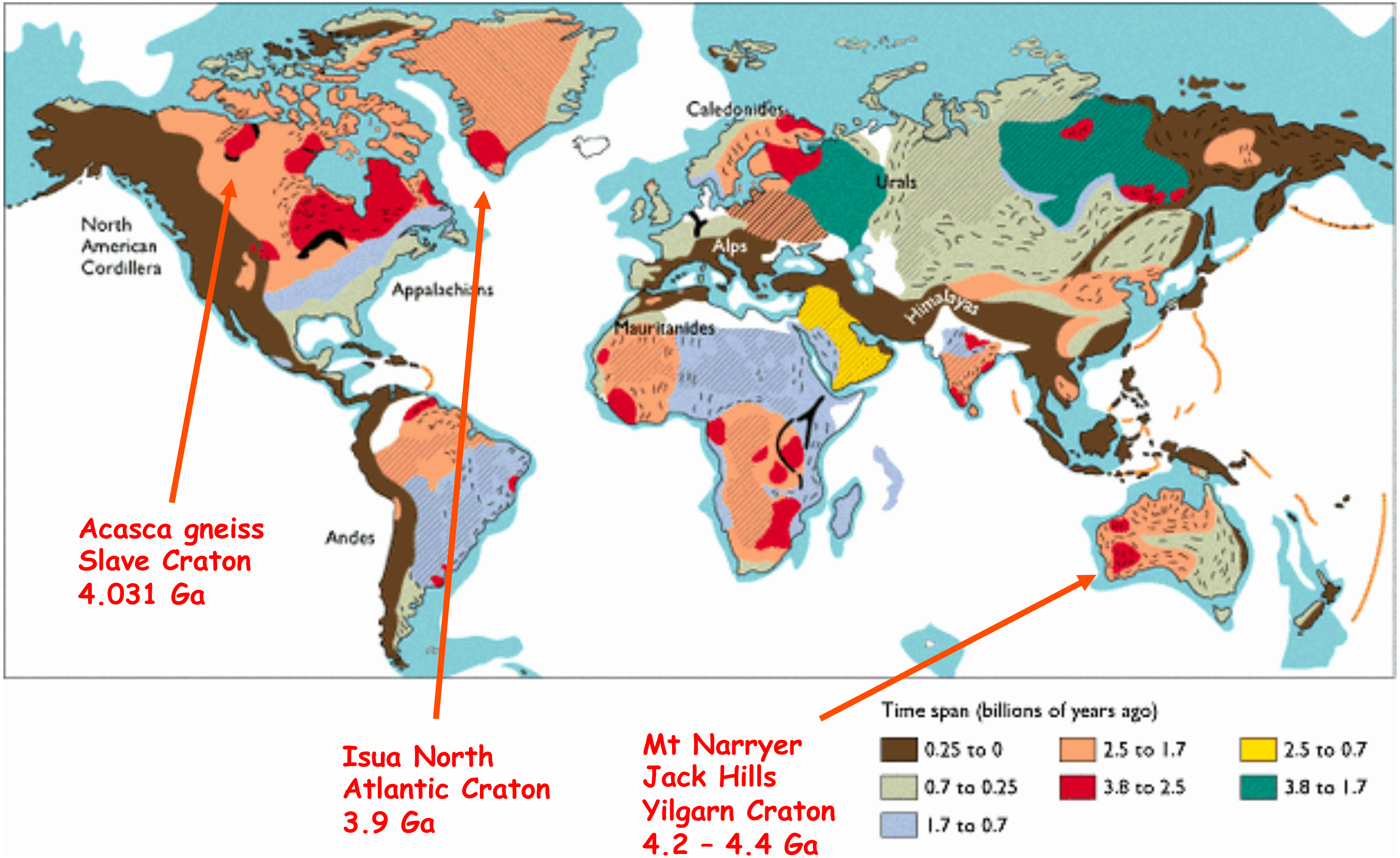
FIGURE 6.9 Rubidium–strontium diagram for a number of different chondritic samples of various types (labeled by different symbols), forming a beautiful straight line with an age of 4.55 ± 0.01 billion years.



Ages of meteorites

Meteorite Name	*Material Dated*	*Method*	*Age (billions of years)*
Allende	whole rock	Ar-Ar	4.52 +/- 0.02
	whole rock	Ar-Ar	4.53 +/- 0.02
	whole rock	Ar-Ar	4.48 +/- 0.02
	whole rock	Ar-Ar	4.55 +/- 0.03
	whole rock	Ar-Ar	4.55 +/- 0.03
	whole rock	Ar-Ar	4.57 +/- 0.03
	whole rock	Ar-Ar	4.50 +/- 0.02
	whole rock	Ar-Ar	4.56 +/- 0.05
Guarena	whole rock	Ar-Ar	4.44 +/- 0.06
	13 isochron	Rb-Sr	4.46 +/- 0.08
Shaw	whole rock	Ar-Ar	4.43 +/- 0.06
	whole rock	Ar-Ar	4.40 +/- 0.06
	whole rock	Ar-Ar	4.29 +/- 0.06
Olivenza	18 isochron	Rb-Sr	4.53 +/- 0.16
	whole rock	Ar-Ar	4.49 +/- 0.06
St. Severin	4 isochron	Sm-Nd	4.55 +/- 0.33
	10 isochron	Rb-Sr	4.51 +/- 0.15
	whole rock	Ar-Ar	4.43 +/- 0.04
	whole rock	Ar-Ar	4.38 +/- 0.04
	whole rock	Ar-Ar	4.42 +/- 0.04
Indarch	9 isochron	Rb-Sr	4.46 +/- 0.08
	12 isochron	Rb-Sr	4.39 +/- 0.04
Juvinas	5 isochron	Sm-Nd	4.56 +/- 0.08
	5 isochron	Rb-Sr	4.50 +/- 0.07
Moama	3 isochron	Sm-Nd	4.46 +/- 0.03
	4 isochron	Sm-Nd	4.52 +/- 0.05
Y-75011	9 isochron	Rb-Sr	4.50 +/- 0.05
	7 isochron	Sm-Nd	4.52 +/- 0.16
	5 isochron	Rb-Sr	4.46 +/- 0.06
	4 isochron	Sm-Nd	4.52 +/- 0.33
Angra dos Reis	7 isochron	Sm-Nd	4.55 +/- 0.04
	3 isochron	Sm-Nd	4.56 +/- 0.04
Mundrabrilla	silicates	Ar-Ar	4.57 +/- 0.06
	olivine	Ar-Ar	4.54 +/- 0.04
	plagioclase	Ar-Ar	4.50 +/- 0.04
Weekeroo Station	4 isochron	Rb-Sr	4.39 +/- 0.07
	silicates	Ar-Ar	4.54 +/- 0.03

Ages of oldest rocks at surface of the Earth



Continental rock on Earth is much older than oceanic basalt. On continents, the oldest rocks only rarely exposed on the surface

Chronology of formation of Earth and inner Solar System

Age	When	What
0 (Sun formed)	4.55 Gyr before present	First solids formed
5 million years		Gas and dust ejected from young Solar System
30 million years		Earth melts and differentiates
100 million years		Large impact formed the Moon
500 million years		Cratering declines sharply, Solar System more or less as it is today
Probably about 700 million years	Probably about 3.8 Gyr before present	First evidence for Life on Earth??

Aside on (quite different) radioactive carbon dating for biological material

$$^{14}C \rightarrow {}^{14}N + e^{-} \qquad\qquad \tau_{0.5} = 5370 \text{ yr}$$

The ratio of $^{14}C/^{12}C$ in the atmosphere is maintained at an equilibrium value (1.5×10^{-12}) by production of new ^{14}C in upper atmosphere from cosmic ray impacts with ^{12}C

Living things continually exchange carbon with atmosphere. Dead things do not.

Once a living thing dies the ratio of $^{14}C/^{12}C$ declines due to the decay of ^{14}C.

$$\frac{^{14}C}{^{12}C} = \left.\frac{^{14}C}{^{12}C}\right|_{atmos} e^{-\lambda t}$$

2. Geological evolution of terrestrial bodies: sources of heat

Clearing the nebula phase → impacts of planetessimals

Kinetic energy of an impact:

$$\boxed{KE = \frac{1}{2}mv^2}$$

Which velocity is relevant?

e.g. for impact with Earth

$$v_{esc} = \sqrt{\frac{2GM_{Earth}}{r}} \sim 11 \text{ kms}^{-1}$$

Escape speed is speed of something dropped from infinity

$$v_{orb} = \sqrt{\frac{GM_{sun}}{R}} \sim 29 \text{ kms}^{-1}$$

Orbital speeds in Solar System (but these are not randomly oriented, so impact velocity will be less)

Impact energy per unit mass of impactor:

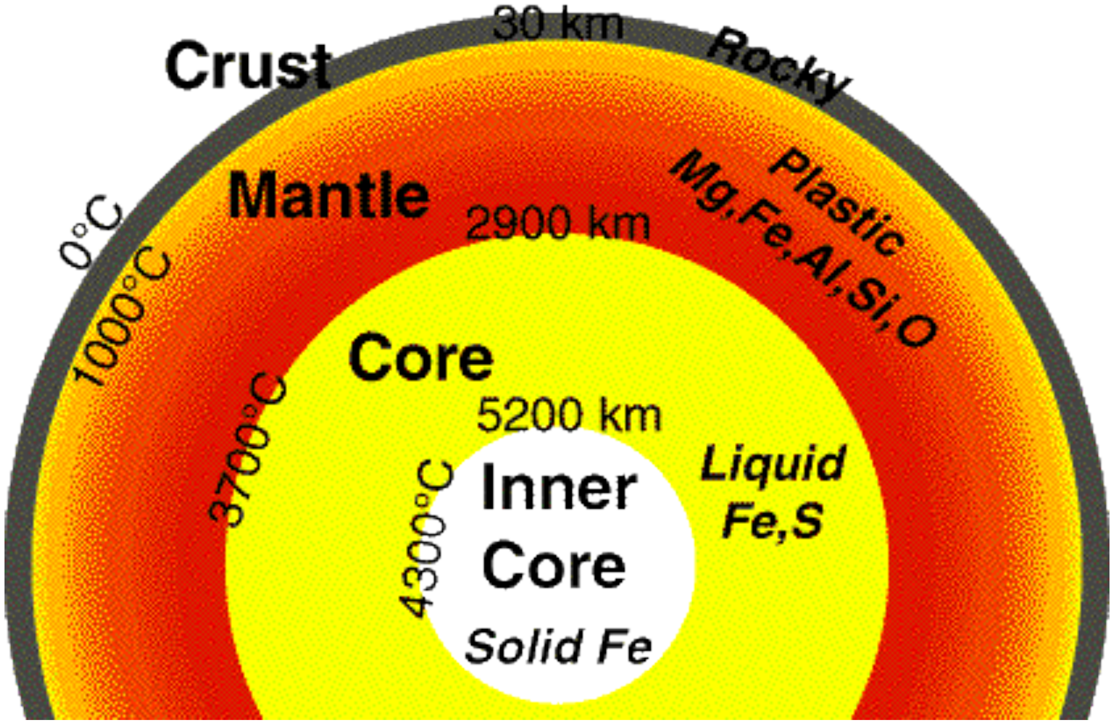
$$\boxed{\frac{1}{2}v_{esc}^2 \sim 10^8 \text{ Jkg}^{-1}}$$

Compare this with the specific heat capacity (10^3 Jkg^{-1} K^{-1}) and the latent heat of fusion (4×10^5 Jkg^{-1}) of Basalt rock

Conclusion: An impacting planetessimal can melt about <u>10^2 times</u> its own mass

Melting leads to:

- Differentiation (dense substances sink to center, e.g. Fe, Ni)
- Outgassing of any volatile substances from interior



Cooling of hot planets

Heat losses ∝ area ∝ r^2

Heat production from gravity ∝ -PE ∝ mass^2/r ∝ r^5

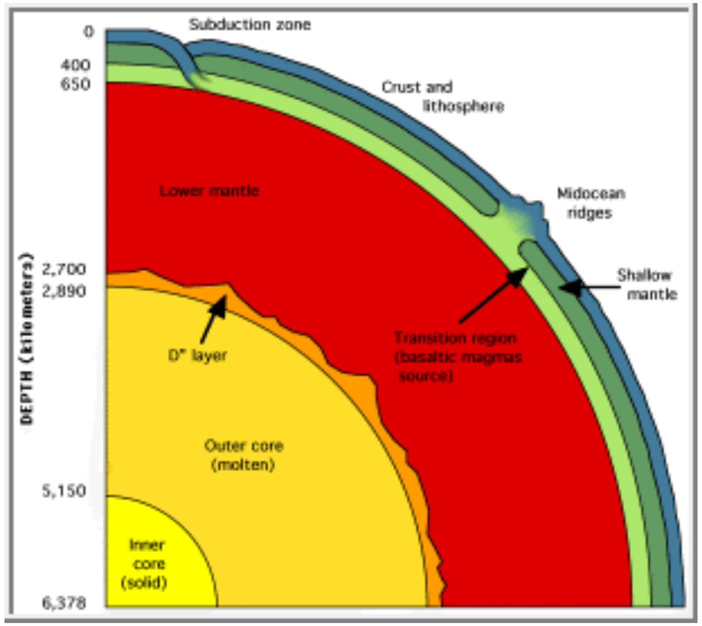
→ Small objects will cool quicker (e.g. Moon, Mars, Mercury) leading to early termination of geological activity(c.f. Venus and Earth)

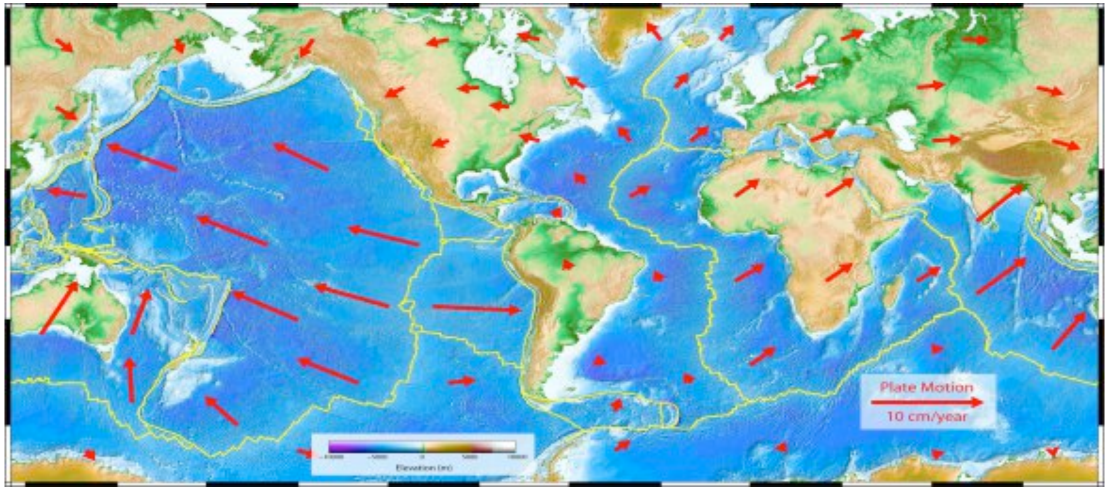
→ These objects show old (heavily cratered) surfaces because they solidified rapidly and have had little surface reprocessing by geological activity

Planetary interiors can also be kept hot by any of the following additional heat sources

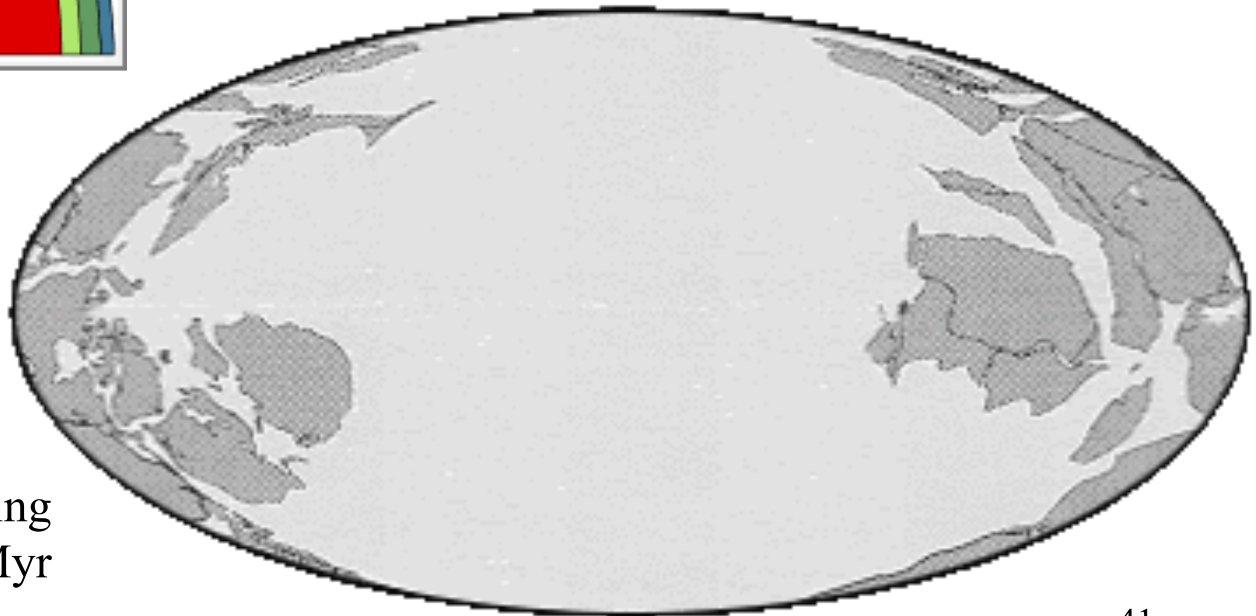
- Decay of radioactive nuclides within interior
- Tidal heating (compression) effects (e.g. Io)
- Gravitational Kelvin-Helmholz contraction (gas giants)

Non-rigid interiors: geological activity on Earth





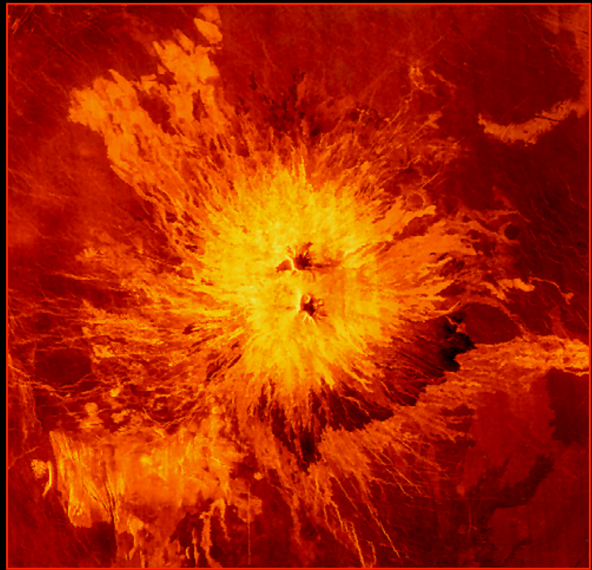
Average age of <u>oceanic</u> crust (< 100 Myr), much less than average age of <u>continental</u> crust (~ 2Gyr)

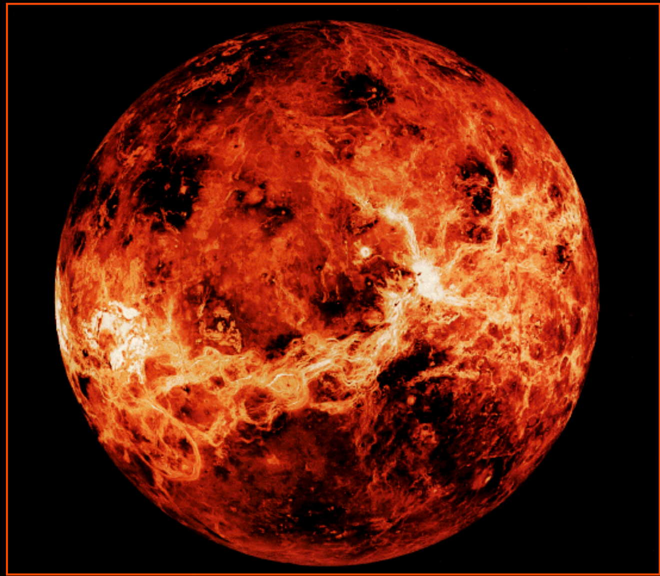


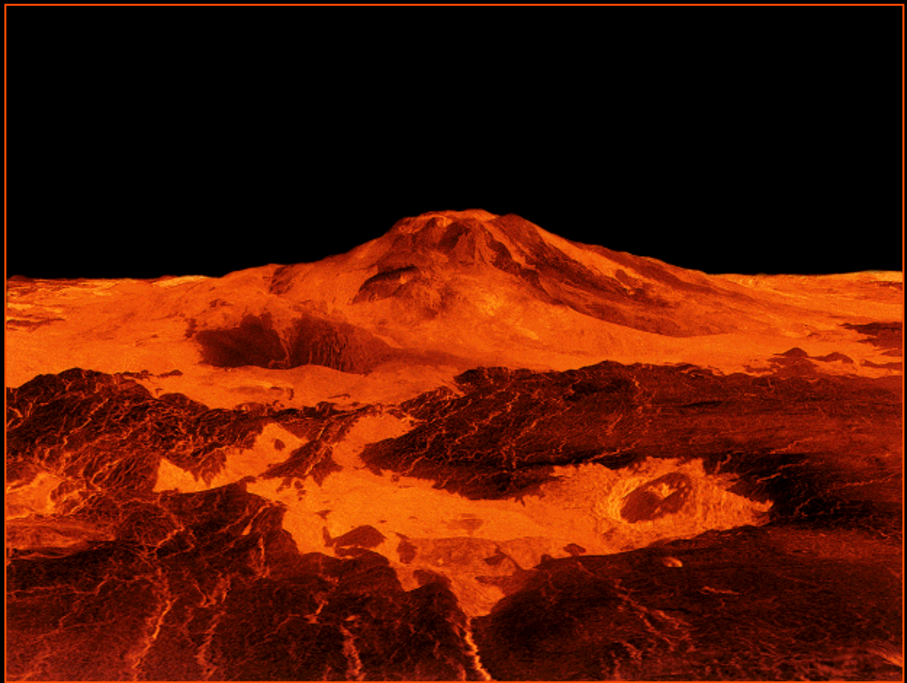
Continental motions during the last 750 Myr

Venus

- Ongoing activity (young surface 300-600 Myr)
- No tectonic plates
- Difference with Earth reflects water content?

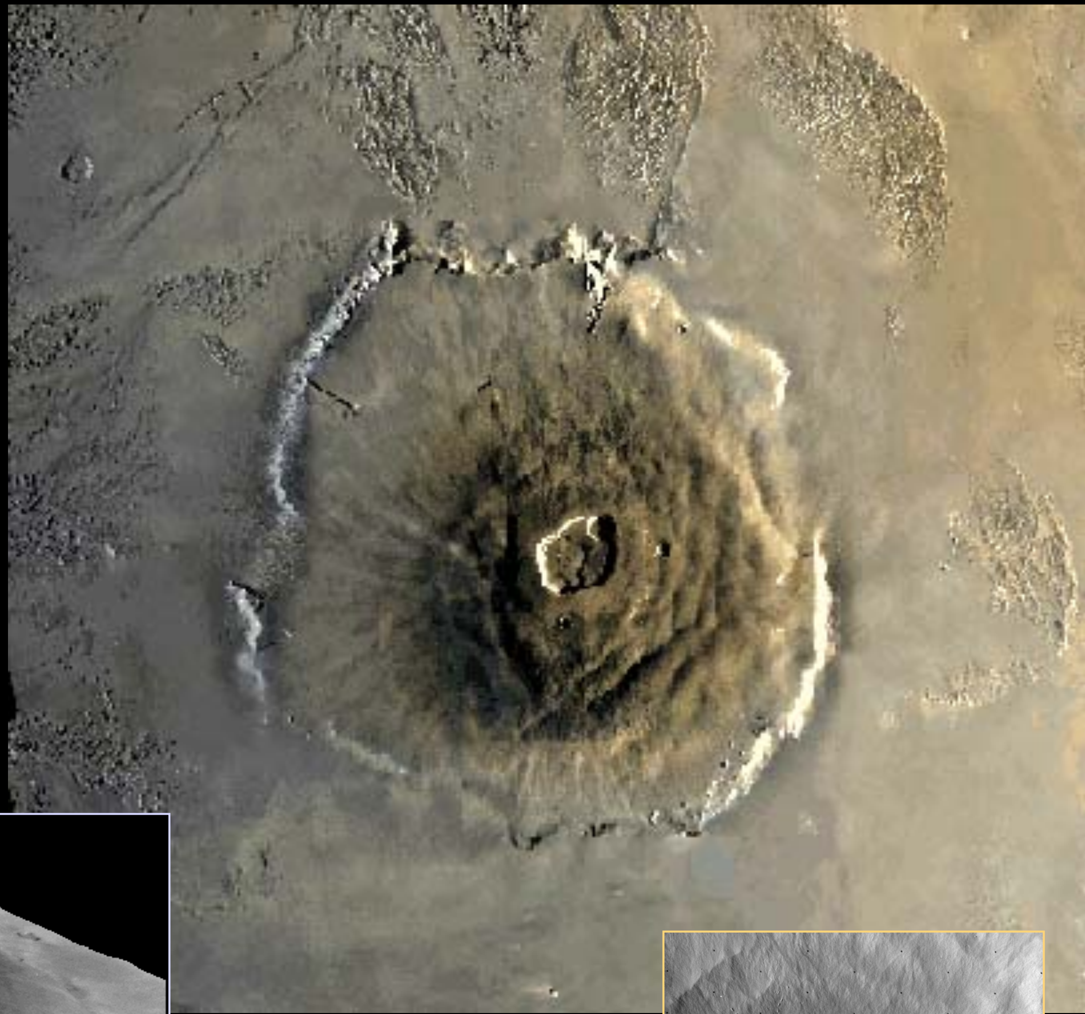


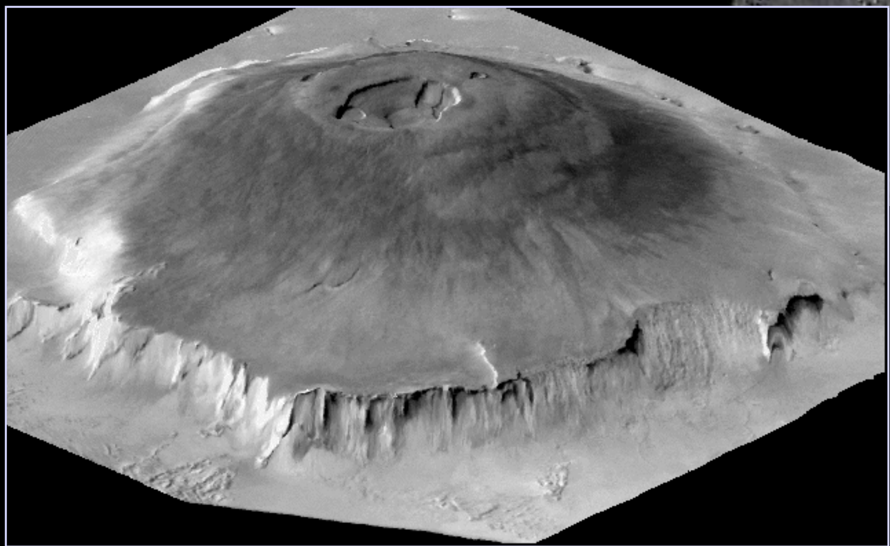


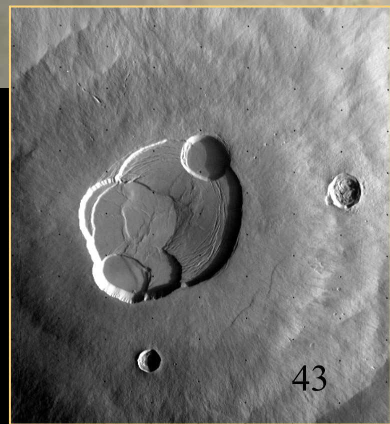


Mars does have (huge) volcanoes, but:

- No plate tectonics
- No ongoing volcanic activity (last major episode 500 Myr ago)







Tidal forces on objects

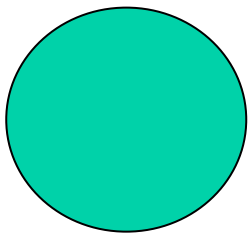
Consider a small object orbiting in the gravitational field of a larger one

$$\boxed{g = -\frac{GM}{R^2}} \quad \boxed{\Delta a = -\frac{d}{dR}\left(\frac{GM}{R^2}\right)\Delta R = 2\frac{GM}{R^3}\Delta R}$$

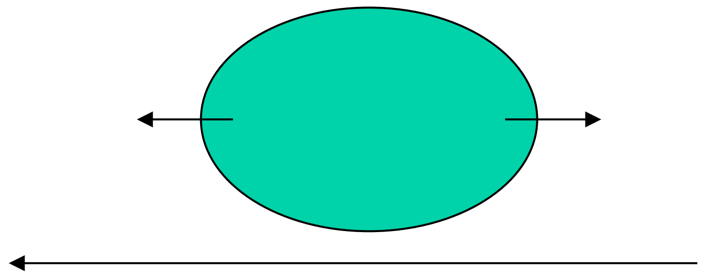
Differential acceleration

Distance to primary object

= size of secondary object



Uniform g



Varying g across an object produces local tidal forces, which can distorting object (e.g. ocean tides on Earth)

Tidal effects have several interesting roles in our story ...

Tidal forces on objects

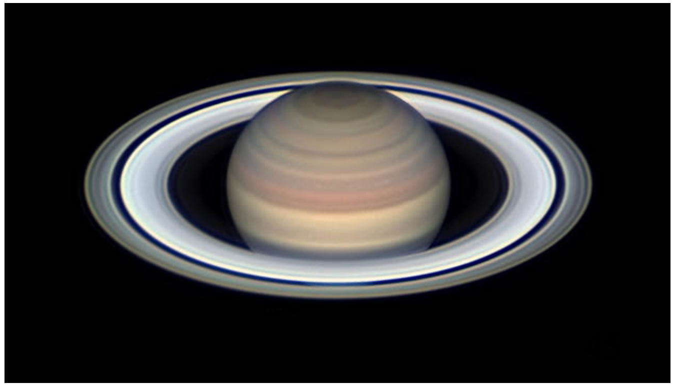
$$\boxed{\Delta a = 2\frac{GM}{R^3}r}$$

(1) *Roche limit*: When the tidal forces exceed the satellite's own gravity holding it together, disruption of satellite of radius *r* orbiting at distance *R*

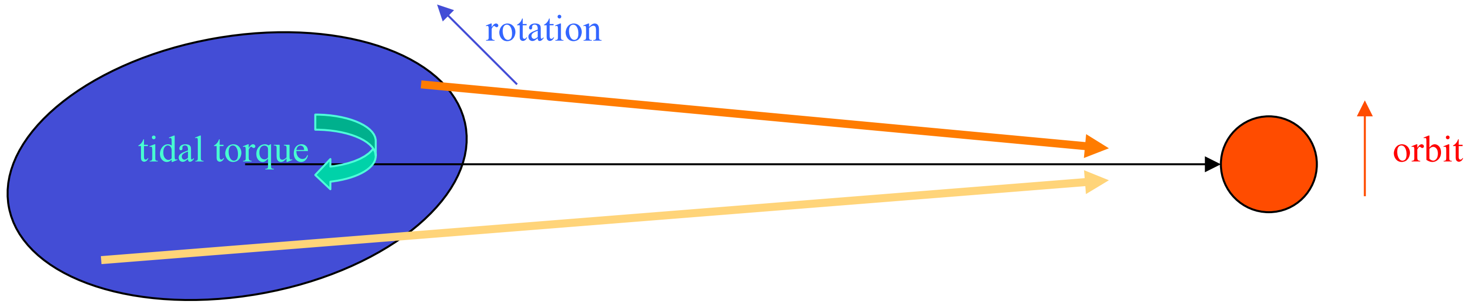
$$\boxed{2\frac{GM_{planet}}{R^3}r > \frac{GM_{moon}}{r^2}}$$

$$\boxed{R < 2^{1/3}\left(\frac{\bar{\rho}_{planet}}{\bar{\rho}_{moon}}\right)^{1/3} r_{planet}}$$

So, if $\rho_{planet} \sim \rho_{satellite}$, the satellite is disrupted when the orbital radius is comparable to the primary radius (e.g. Saturn's rings)



(2) *Tidal locking*: Rotation at $\omega_{rot} \neq \omega_{orbit}$ produces misalignment of tidal bulge due to friction (whether solid or liquid body) and thus torques



These torques act to make $\omega_{rot} = \omega_{orbit}$

- Moon's spin already synchronised with orbit (keeps one face towards us)
- Earth's spin is slowing (our "day" is lengthening by 2ms per century)

Tidal torques transfer angular momentum from the spins to the orbit

- Radius (and period) of the Moon's orbit is <u>increasing</u> as the angular momentum increases

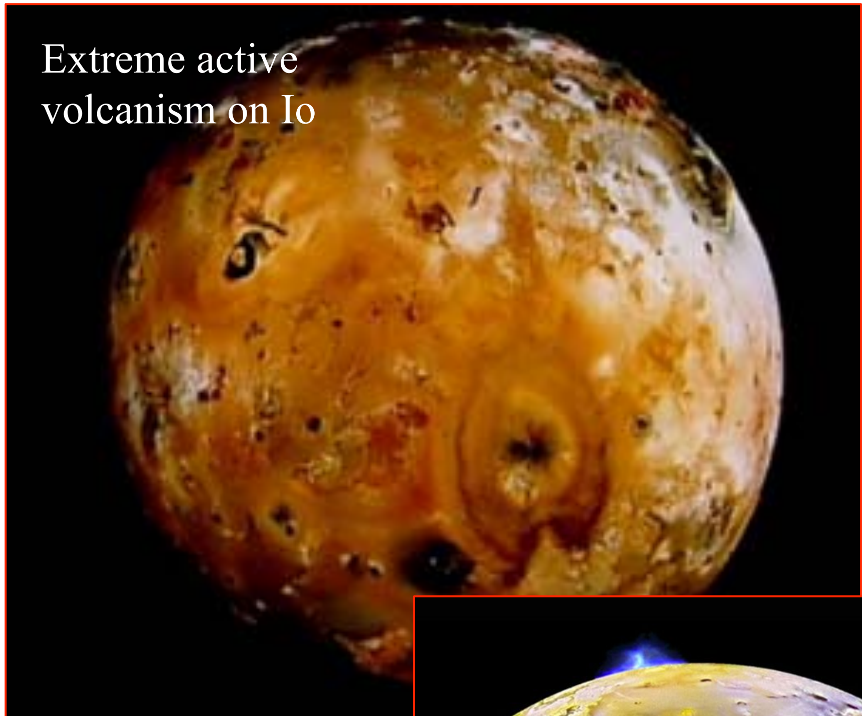
Implications for Life?

- Tidal pools on shores of ocean as ideal containers for chemistry?
- Tidal locking of planets with stars will produce extreme temperature variations across surface?

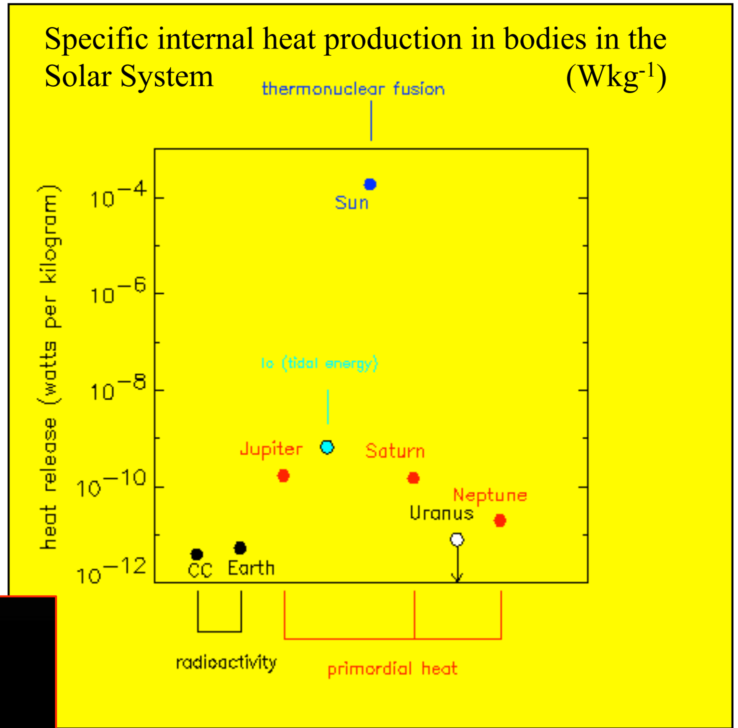
(3) Internal heating: the friction associated with repeated tidal deformations, due to non-synchronous rotation and/or eccentric orbit (note R^3 dependence), produces <u>internal heating</u>

$$\Delta a = 2\frac{GM}{R^3}r$$

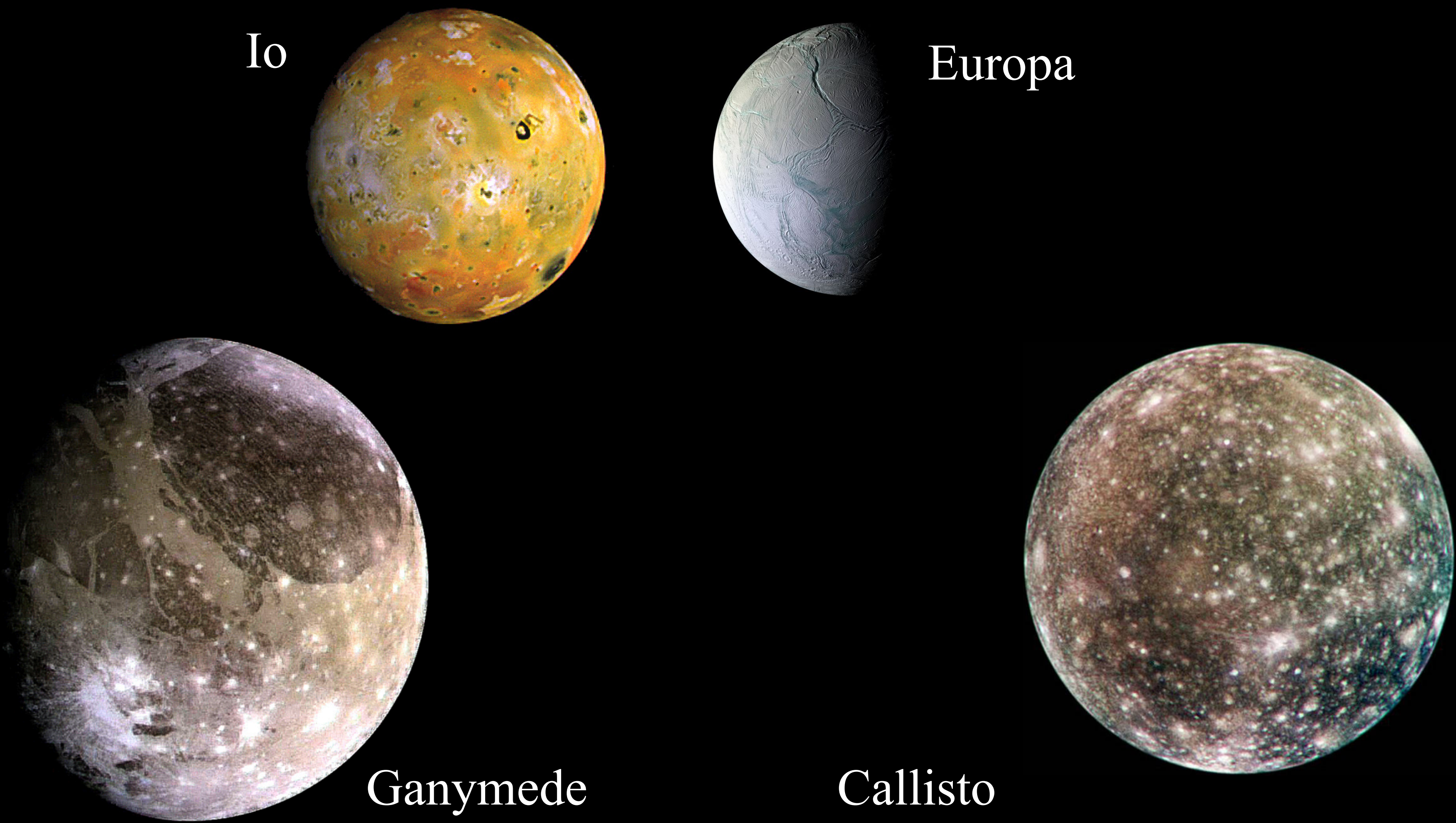
e.g. Io, innermost satellite of Jupiter, whose surface elevation changes by up to 100m during its 41 hour orbit!



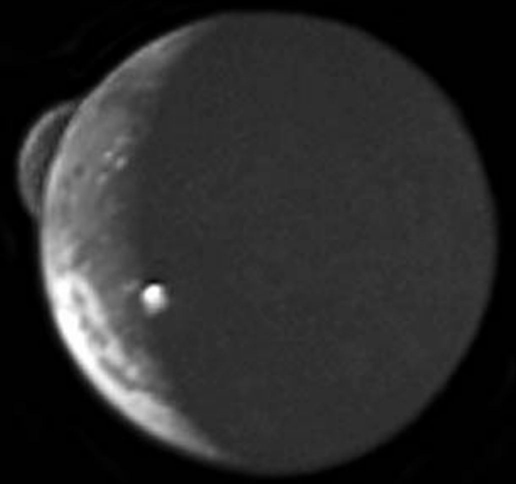
Extreme active volcanism on Io

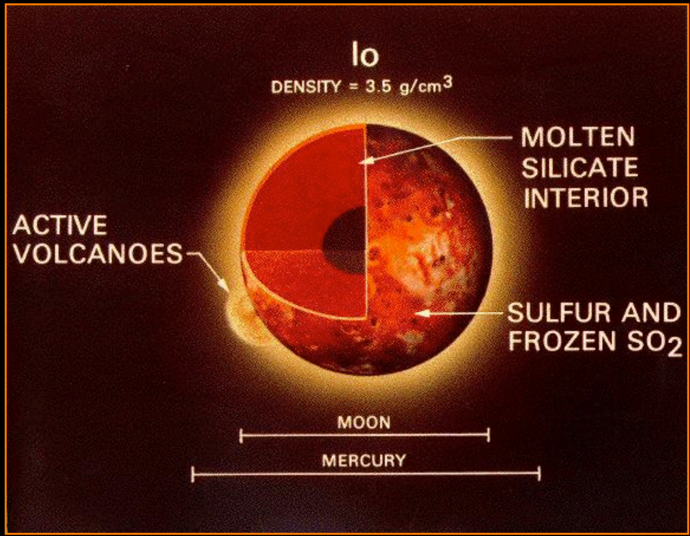


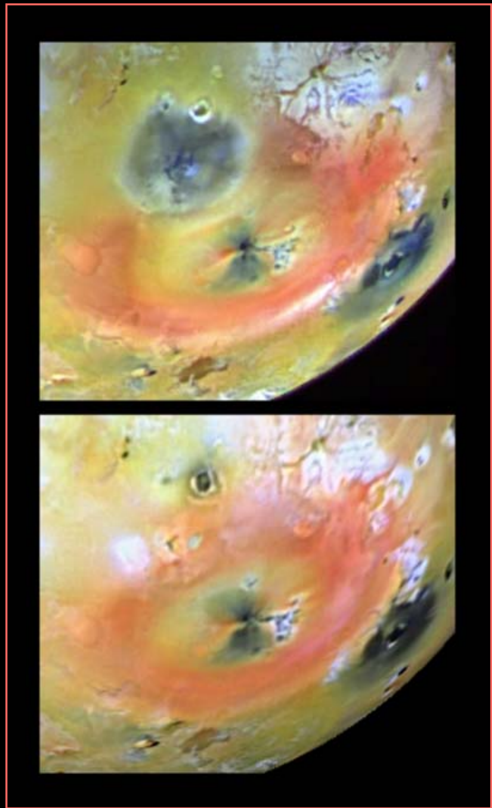
Moon	Period (days)	Diam (km)	Density (g/cc)	Eccentricity
Io	1.769	3642	3.529	0.0041
Europa	3.551	3120	3.018	0.0101
Ganymede	7.155	5268	1.936	0.0015
Callisto	16.689	4800	1.851	0.007



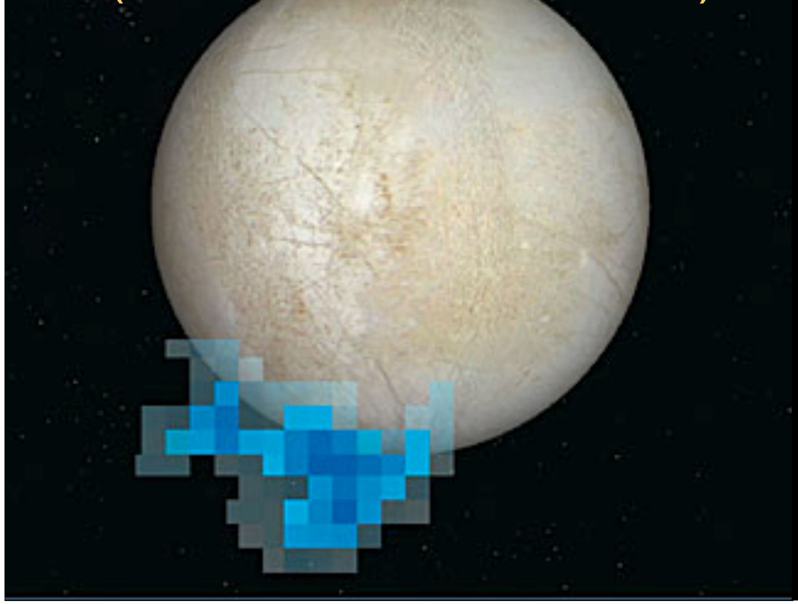
Volcanoes on Io

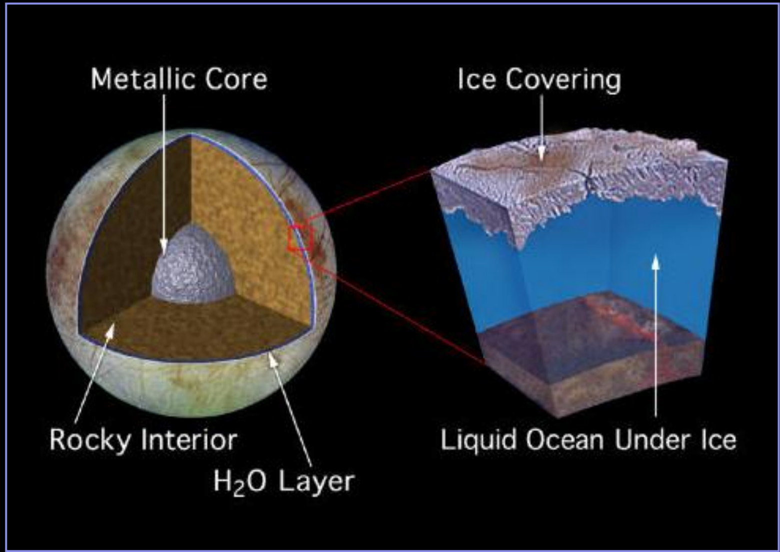






Subsurface oceans and H_2O Geysers on Europa (also Saturn's Enceladus)





3. Origin and evolution of planetary and satellite atmospheres

Sources of planetary and satellite atmospheres

First, note that the atmosphere on Earth is a very small fraction (0.02%) of the total mass (even including the water oceans as "atmosphere")

- Solid material 6.0×10^{24} kg
- Water ocean 1.4×10^{21} kg
- Gas atmosphere 5×10^{18} kg

Possible sources of the atmospheres

- Capture of gases from original Solar Nebula (e.g. H, He): only relevant for the massive Outer Planets
- Outgassing of volatile substances from the interior during molten phase
- *Most favoured:* Subsequent impacts by volatile rich planetessimals (perturbed from outer solar system) during the "clearing of the Nebula" phase. NB. It is "easy" to get "Water Worlds", that are completely covered by water (On Earth: note that the average depth of the ocean is 3.6 km, the highest mountain above sea level = 8.8 km!)

Loss of planetary and satellite atmospheres

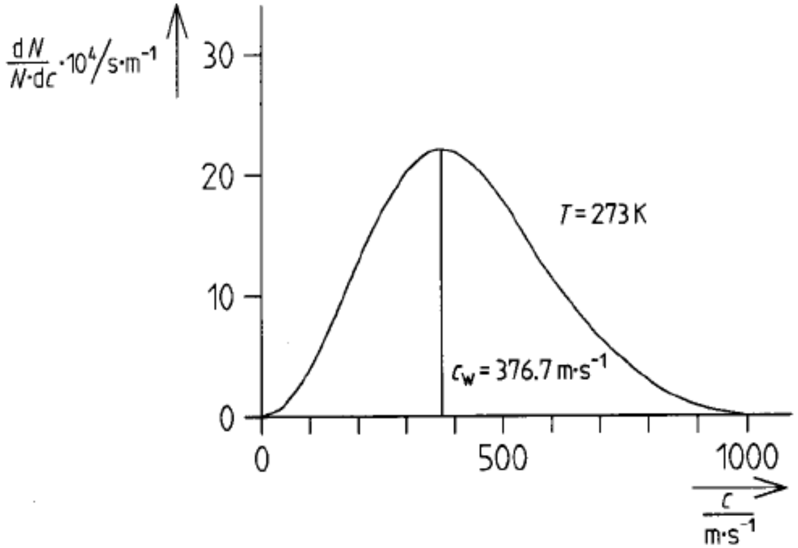
- Thermal leakage to space
- (Stripping due to large impacts)

At some altitude (= the "exosphere"), collisions between gas particles become negligible, and they move on <u>ballistic</u> trajectories determined only by gravity, and those with $v > v_{\rm esc}$ can leave the planet.

Repartition of thermal energy at lower altitudes means the atmosphere can continually "leak" through this high velocity tail, even if $v_{\rm rms} << v_{\rm esc}$

Good rule of thumb in Solar System: loss of atmosphere results if

$$v_{rms} > \frac{1}{6} v_{esc}$$



Maxwellian velocity distribution:

$$\boxed{n_v dv = n\left(\frac{m}{2\pi kT}\right)^{3/2} \exp\left\{\frac{-mv^2}{2kT}\right\} 4\pi v^2 \, dv}$$

$$v_{rms} > \frac{1}{6} v_{esc}$$

v_{rms} for different gas species

$$v_{rms} = \sqrt{\frac{3kT}{m_{gas}}}$$

v_{esc} for each planet

$$v_{esc} = \sqrt{\frac{2GM_{planet}}{R_{planet}}} \longrightarrow$$

$$T_{esc} \geq \frac{1}{54} \frac{GM_{planet} m_{gas}}{kR_{planet}}$$

Gives the critical temperature for each species on each planet

If the actual atmospheric T for a planet is > T_{esc} for a given atomic/molecular species, that species escapes

e.g. for molecular N_2:	T_{esc}(Earth) ~ 3900K:	N_2 stays (T_{exo} ~ 1000K)
	T_{esc}(Moon) ~ 180K:	N_2 lost
	T_{esc}(Mars) ~ 700K:	marginal

Note: Photodissociation by solar ultraviolet radiation of volatile species like CH_4, NH_3, (H_20) produces the very light Hydrogen (either as H or H_2) which is almost always quickly lost.

Comparing the very different atmospheres of three terrestrial planets: How can we understand these differences?

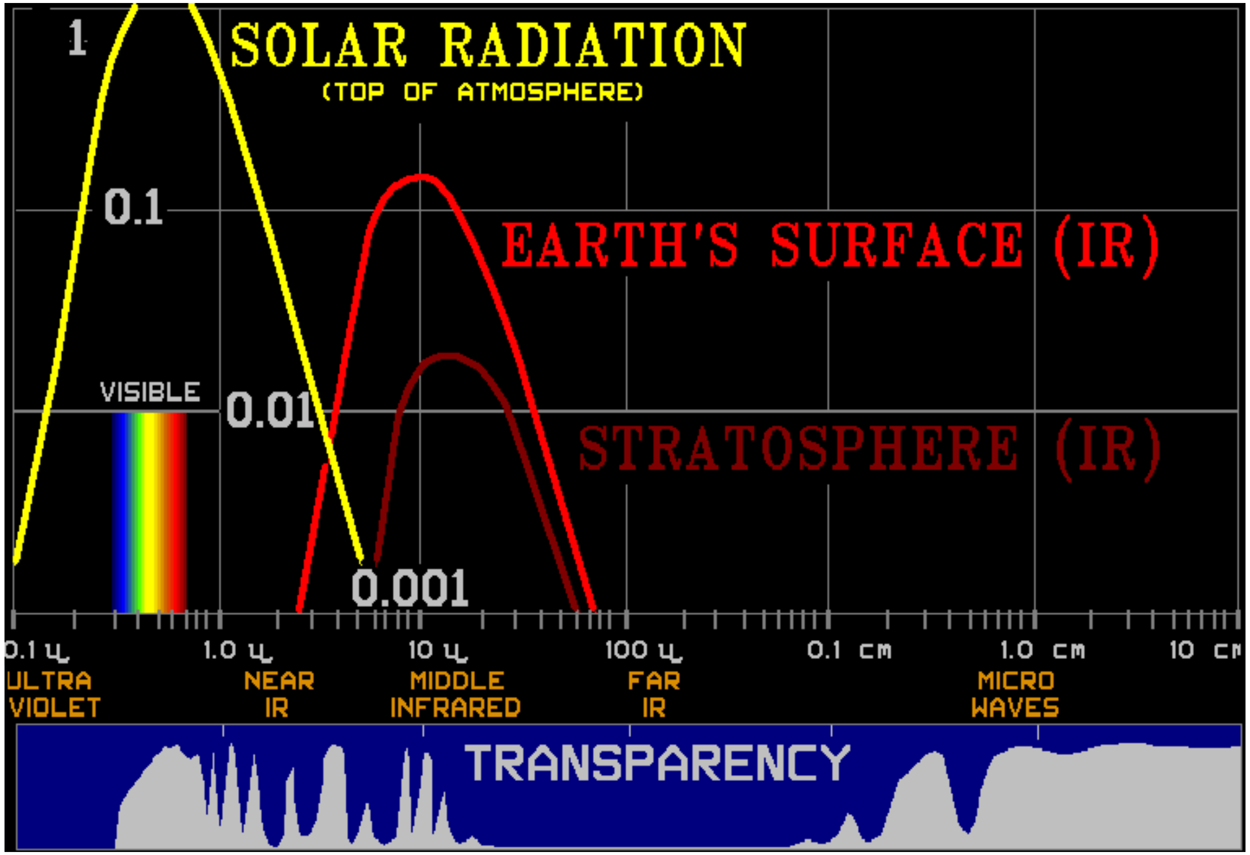
	mass 10^{24} kg	radius km	v_{esc} kms^{-1}	Solar dist. A.U.	Density gm/cm^3	Atm. pressure (bar)	Surface Temp. K	Primary atmos. compon.
Venus	4.9	6000	10.4	0.72	5.24	90	750	$C0_2$ (96%) N_2 (3.5%)
Earth	6.0	6400	11.2	1.00	5.52	1	300	N_2 (78%) O_2 (21%) H_20 (~1%) Ar (1%) (+300 bar in liquid H_20)
Mars	0.6	3400	4.8	1.52	3.9	< 0.01	180-290	$C0_2$ (95%) N_2 (3%) (+ten times more CO_2 in polar ice caps)

Greenhouse effects in atmospheres:

A planet receives solar radiation at visible wavelengths ($T_{eff} \sim 5800K$) and itself radiates at infrared wavelengths ($T_{eff} \sim 300K$).

What happens if the atmosphere is transparent at one wavelength but not at the other?

e.g. Earth's atmosphere



Equilibrium temperature in the absence of any atmosphere:

The Sun has temperature, T_S, radius R_S, and lies at a distance D

The planet has radius r, and reflects a fraction a of the incoming light (the "albedo")

In principle (without an atmosphere) this will set up an equilibrium temperature T_E

$$4\pi r^2 \sigma T_E^4 = (1-a)\ 4\pi R_S^2 \sigma T_S^4 \frac{2\pi r^2}{4\pi D^2}$$

$$T_E = T_S (1-a)^{1/4} \sqrt{R_S / 2D}$$

Note: T_E is <u>independent</u> the radius r of the planet (because both absorption and re-emission depend on surface area)

Now add a partially transparent/opaque atmosphere

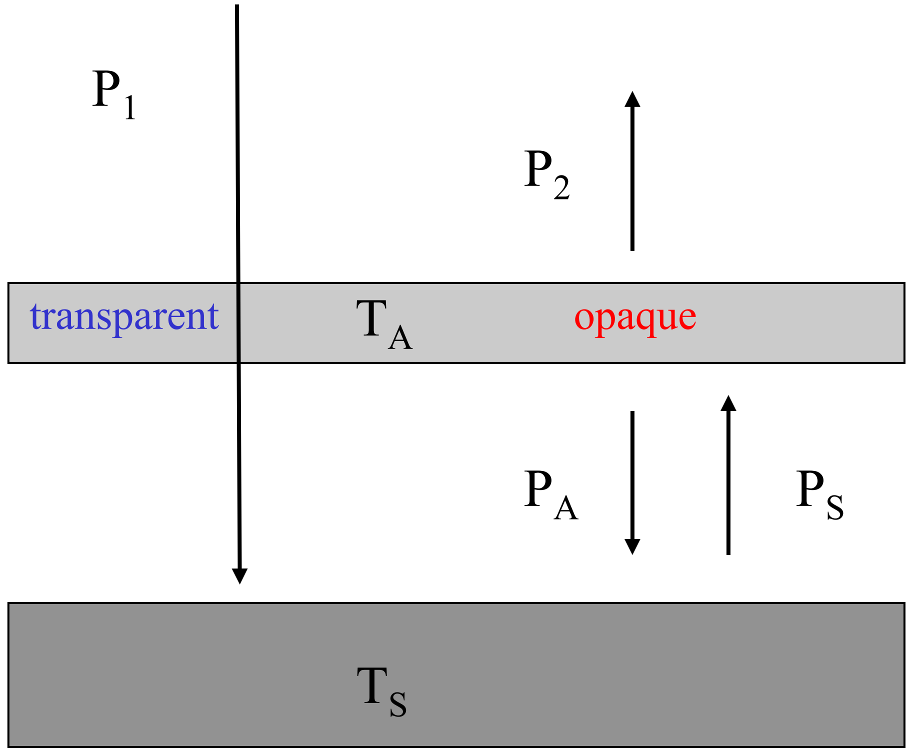
Set up some simultaneous equations involving energy flows

$$P_1 = P_2 = P_A = 0.5\sigma T_A^4$$

$$P_S = \sigma T_S^4 = P_1 + P_A = 2P_1$$

$$\sigma T_E^4 = P_1$$ *(equilibrium temperature in absence of an atmosphere)*

$$\boxed{T_S = \sqrt[4]{2} T_E}$$ About a 20% effect for this simple case



Note for more complex situations bigger effects:

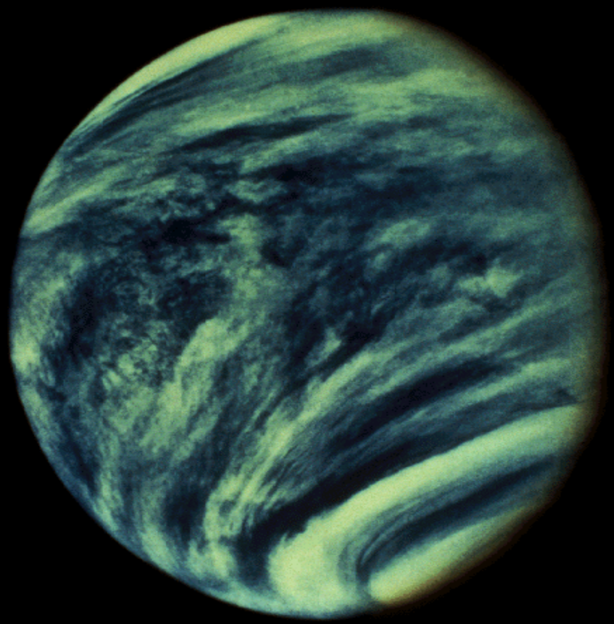
(a) For incomplete opacity in atmosphere

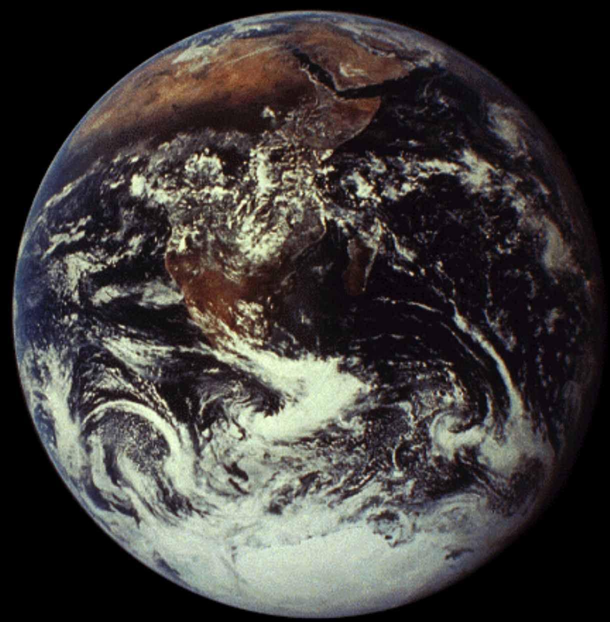
$$\boxed{P_S = \sigma T_S^4 = P_1 + P_A = (2 - f)\, P_1}$$

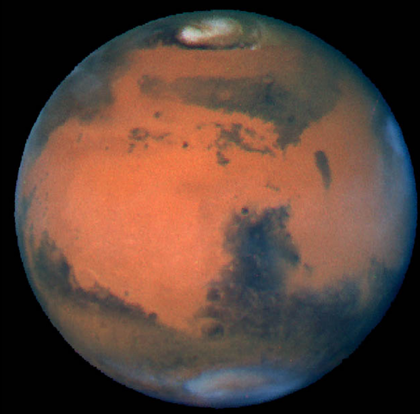
(b) For multiple n (or continuous τ) layers in the atmosphere

$$\boxed{T_S = \sqrt[4]{(1+n)}\ T_E = \sqrt[4]{(1+\tau)}T_E}$$

Greenhouse effects on inner planets







Venus

$T \sim 2T_e$

Earth

$T \sim 1.1T_e$

Mars

$T \sim 1.02T_e$

Now let's go back and look at the early evolution of our three planets:

- All with have an initial atmosphere rich in volatiles brought in by "comets"– H_20, NH_3 and CH_4

  (Note: O,C and N are cosmically the most abundant elements – ignoring He and H – with O:C:N ratio ~ 10:3:1

- Photo-dissociation of these molecules and the loss of the H will convert the initial H-rich reducing atmosphere to a H-poor atmosphere of CO, CO_2, N_2 and H_20 and small amounts of free O_2 Note: the amount of H loss is indicated by the D/H ratio since D ($={}^2H$) is less easily lost than H because it has twice the mass.

- The outer planets remain rich in H, He and hydrogenated gases

Evolution of the Earth (1)

- This is dominated by the fact that H_20 condensed out and formed Earth's oceans (from a very early time).
- CO_2 is highly soluble in water: So rain (produced by the "water cycle") "scrubbed" the CO_2 from atmosphere → producing solution of H_2CO_3 (carbonic acid) in water.
- Reactions with metal ions in oceans → e.g. $CaCO_3$ (rock) (marine life helps but is not essential, most $CaCO_3$ rock is not biological)
- Very small quantities of CO_2 and H_20 that remained in the Earth's atmosphere produce a modest greenhouse effect (boost of +35K)

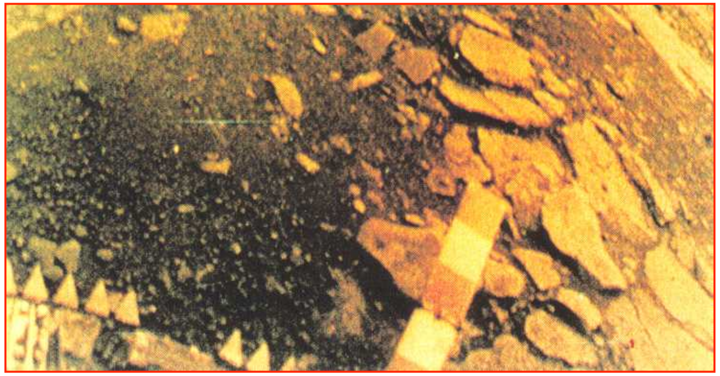
Note that the CO_2 currently "locked" in near-surface rocks would be sufficient to make 70 bar atmosphere of CO_2, i.e. similar to that seen on Venus

Evolution of the Earth (2)

- Subsequent biological activity (photosynthesis) produced free O_2 out of CO_2.
- This initially oxidized the CO $\rightarrow$ CO_2, plus oxidized Fe, S in surface rocks.
- After these were saturated, continued production of O_2 raised the atmospheric O_2 level to 20% (about 2.5 billion years ago) and maintained it there ever since.

The O_2 abundance in the atmosphere is far from equilibrium and is a strong signature of Life that could in principle be detected from far away. Especially, the <u>simultaneous</u> presence of O_2 and trace CH_4 is a strong biosignature that one or other species is being pumped by a source.

What happened to Venus?



H_20 did not condense and did not scrub out the CO_2. Why? The $T_e \sim 330$K, not so hot!

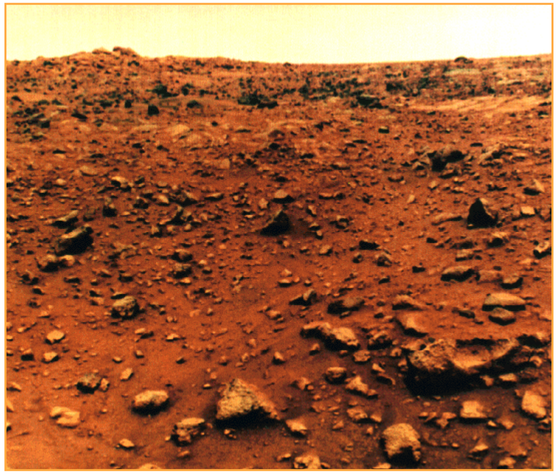
The initial greenhouse effect was strong enough to vaporise water in a runaway effect:

higher T → more evaporation → more greenhouse effect → higher T → more evaporation etc.

The water in the atmosphere was then almost entirely destroyed through photo-dissociation, leaving CO_2 to dominate the atmosphere: [D/H is 100 times higher on Venus than Earth, suggesting > 99.9% of the H was lost from Venus.

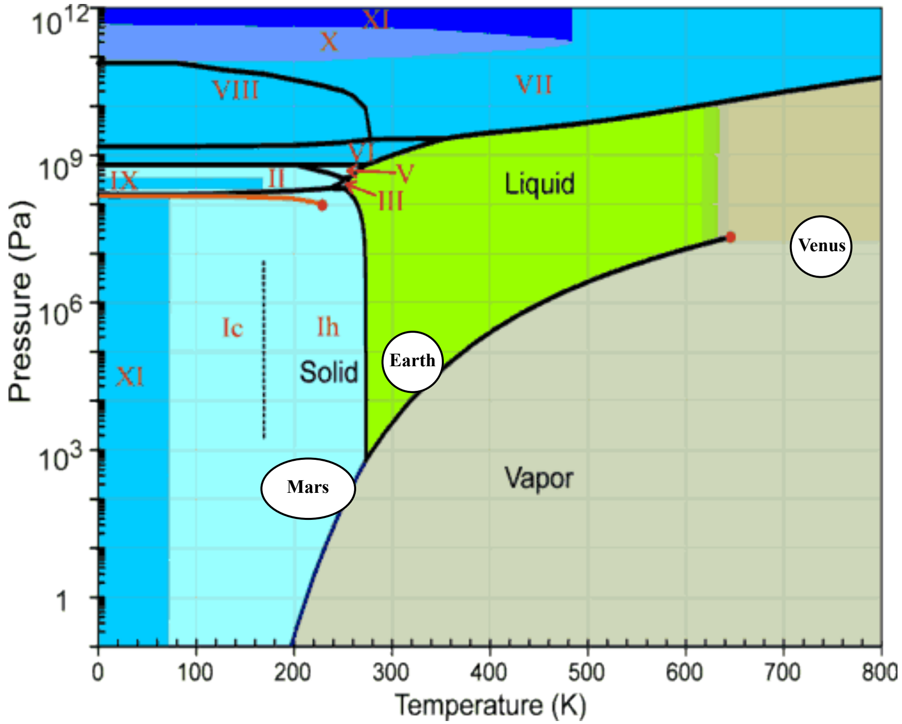
As noted above, the current CO_2 content of Venus' atmosphere is comparable to that in Earth's $CaCO_3$ rocks, while the atmospheric N_2 content is similar on both planets.

What happened to Mars?

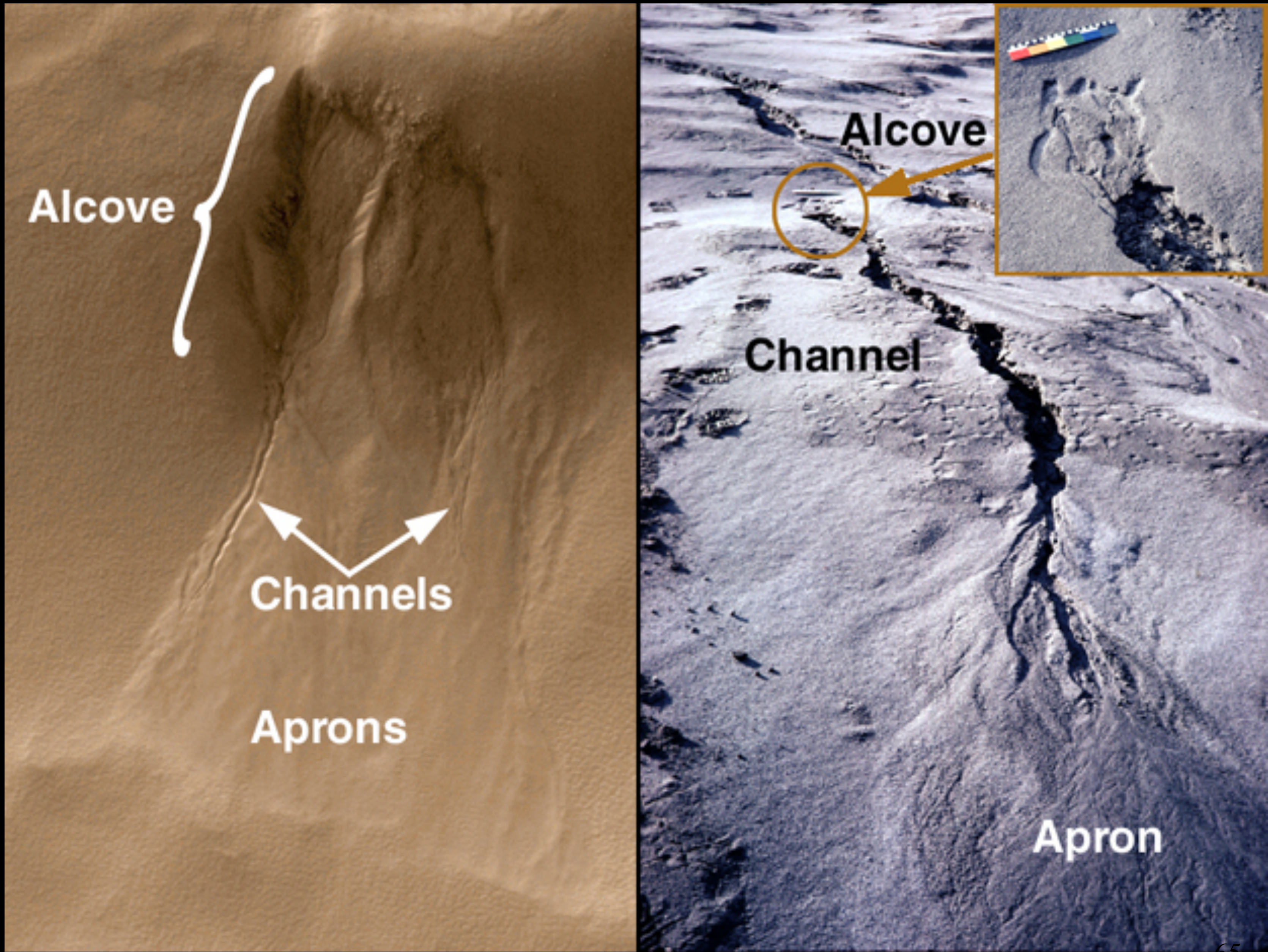


Retention of the atmosphere is marginal. Impacts thought to have stripped away much of the atmosphere (e.g. Ar ratio→ suggests a factor of 100?).

The weakness of the greenhouse effect (+5K) leads to condensation of CO_2 and H_20 on polar ice caps (removing ~ 90% of available CO_2 → runaway "freezeout", reducing greenhouse effect)



The present atmosphere is unable to support liquid water (temperature is below the triple point), yet there is surface evidence for "flows"?


Alcove
Channels
Aprons
Alcove
Channel
Apron

So let’s return to our comparison:

	Apparent atmospheric compositions			
	Pressure	CO_2	H_2O	N_2
Venus	**90**	**0.96**	**-**	**0.04**
Earth	**1**	**-**	**0.01**	**0.78**** **+0.21 O_2**
Mars	**0.01**	**0.95**	**-**	**0.03**
	After taking into account oceans and rocks, and dividing by the mass of the planet			
Venus	**…**	9.6×10^{-5}	$>2\times10^{-5}$	2×10^{-6}
Earth	**…**	16×10^{-5}	2.8×10^{-4}	2×10^{-6}
Mars	**…**	$>3.5\times10^{-8}$	$>5\times10^{-6}$	4×10^{-8}

Note how these three rather similar planets had three very different histories reflecting rather small differences in their initial mass and distance from the Sun

Key ideas: Formation and evolution of the Solar System

- A proto-planetary disk of material (gas+dust) is a <u>natural consequence</u> of the star-formation process. *We would expect planetary systems to be common (and they are!)*

- Planet formation involves chemical differentiation because of the <u>condensation process</u> on grain surfaces. *2% atomic diversity is concentrated up to to ~ 100%*

- The difficult step of growth is from cm to km sized bodies. Stochastic evolution of later stages building planets. *Don't be surprised by a diversity of planetary systems?*

- The heating of massive bodies during impacts in late growth leads to chemical differentiation within the object.

- Potential importance of other sources of heat (e.g. tides) on small bodies. *Especially important are tidal heat sources, e.g. sub-surface water ocean on Europa and Enceladus*

- Atmospheres of rocky planets come from impacts of volatile rich planetessimals during the last "clearing of the Nebula" phase.
- *Provided favourable conditions for Life on Earth...*

- Variety of subsequent evolution due to "amplification" of small initial differences.

  ... but the conditions for Life may not be met everywhere.