Lecture #1 — ~~[illegible]~~ Water on the Earth and in the Solar System

Part II of Press & Siever Chs. 11-17

Part III of Judson & Richardson (and this course) devoted to surface processes, in particular the hydrological cycle, and sediment and solute transport by rivers and glaciers, erosion and sculpting of landforms on ⊕'s surface, i.e., geomorphology.

Water has a dominant role in what makes ⊕ geologically distinctive among planets in the solar system

- important for weathering and ~~[illegible]~~ erosion, presence of sedimentary rocks
- lowers T_{melt} of magmas — explosive volcanism — probably required for differentiation of continental crust (other planets don't have continents)
- important role in rock deformation — presence of water lowers rock strength — might be required for plate tectonics
- necessary for (~~terrestrial~~) life

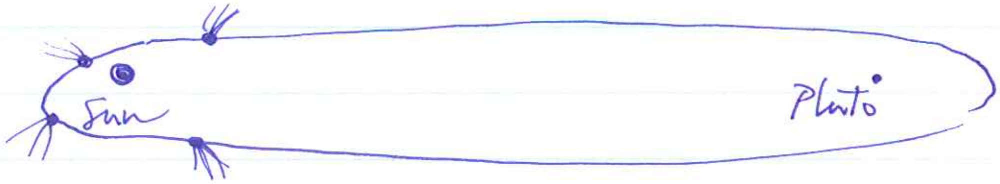
Limited range of temperatures for liquid water — at $P = 1$ atm,

°C (273 K) freezing point
100°C (373 K) boiling point

To be an ⊕-like planet, surface temperature must be in this range

Water in the Solar System

- **Comets** made of water ice & dust (dirty snowballs)
  Come from out beyond Pluto



May be source of water in early Solar System — or water on ⊕ may have been in hydrous minerals in chondritic [illegible] meteorites — thought to be many more comets in early Solar System.

What about other planets:

- **Mercury** — dry — no evidence of volatiles whatsoever

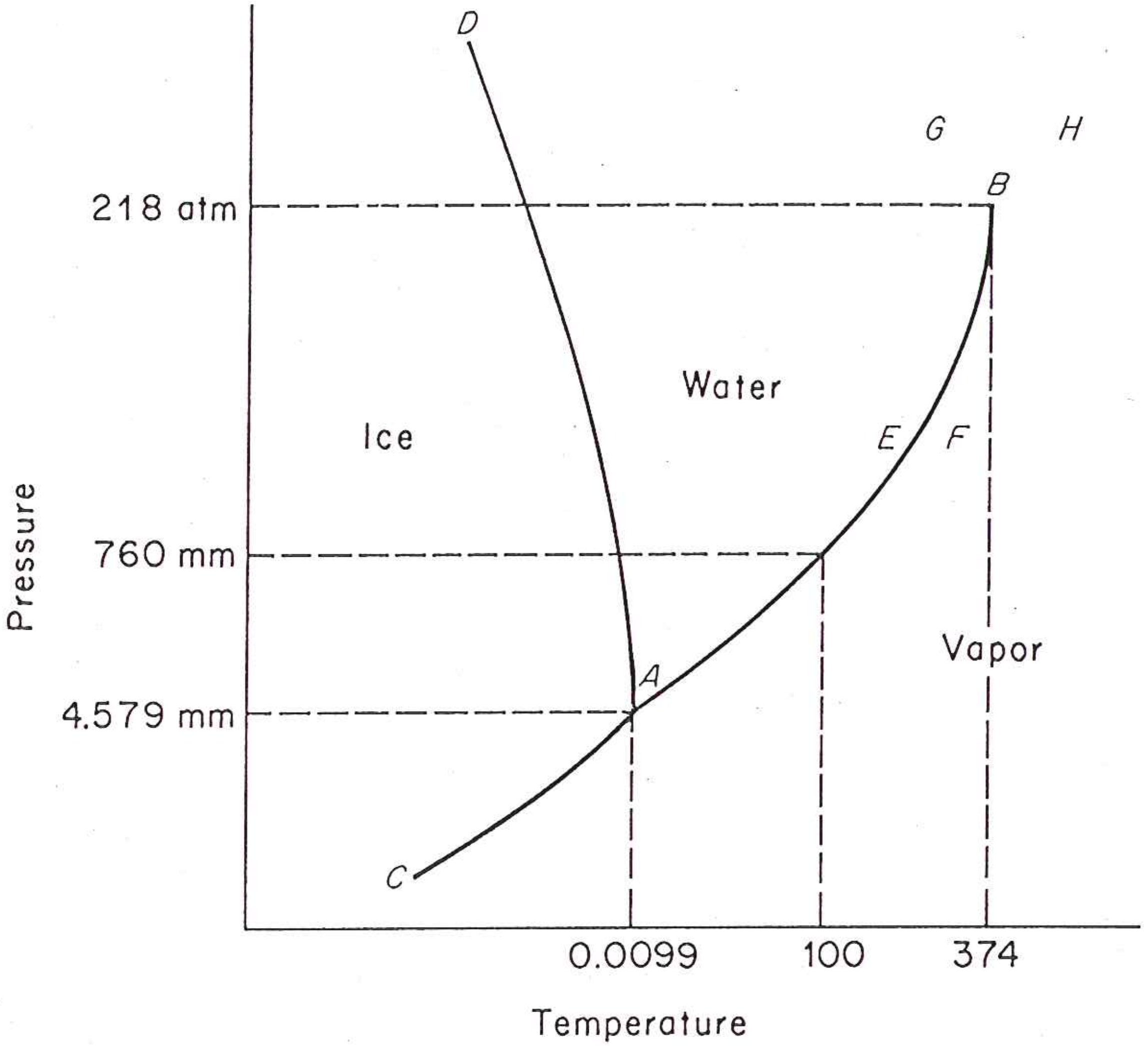


FIGURE 147
Schematic *PT* diagram, showing the relations between ice, water, and vapor.

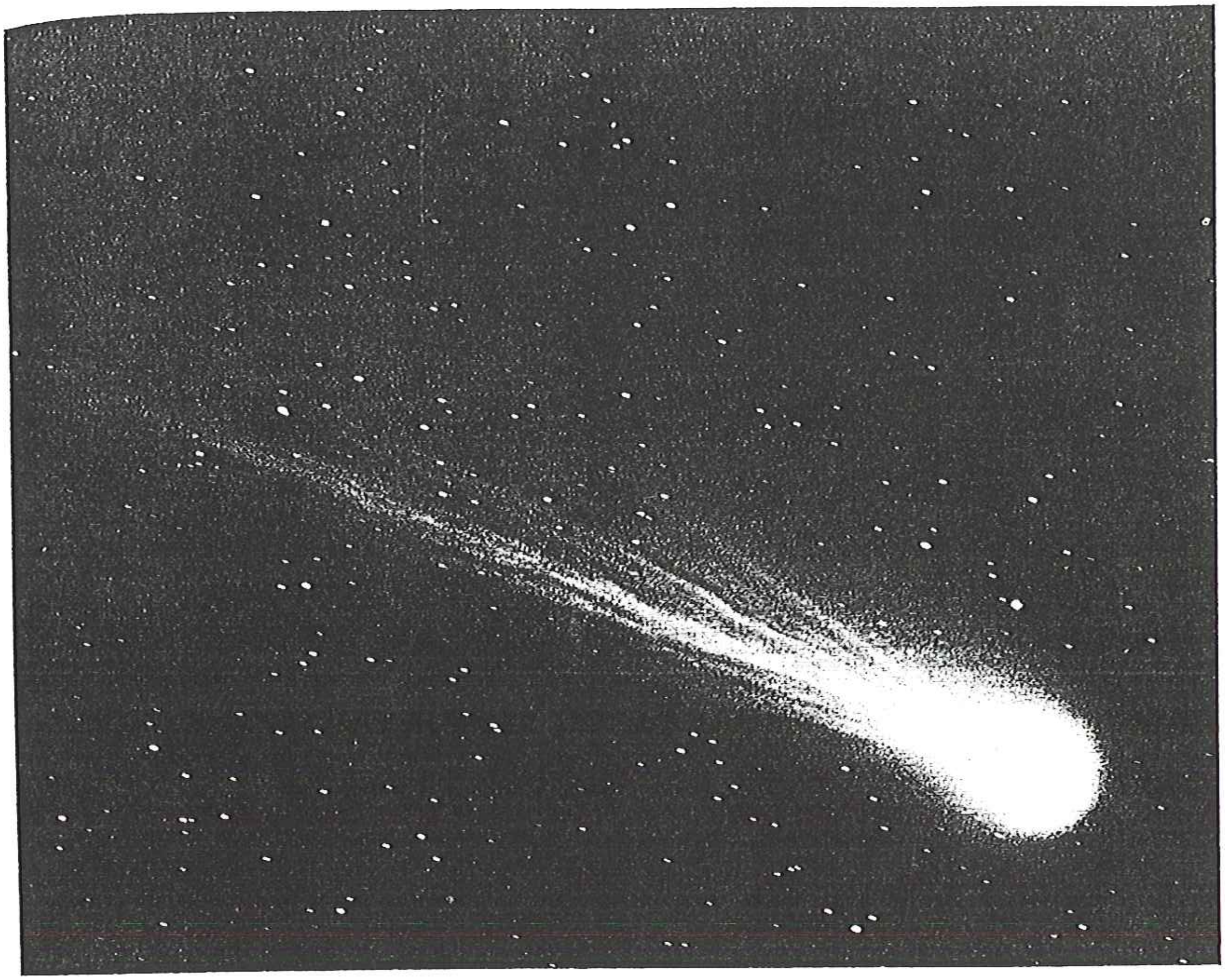


Figure 7.2. Comet Kohoutek, January 1974. (Courtesy Hale Observatories, Pasadena, Calif.)

Table 7.4. Chemical composition of comets

Comas	Tails
Nonmetals	
H, OH, O, S, S_2, H_2O, H_2CO, $(H_2CO)_n$, C, C_2, C_3, CH, CN, CO, CS, N_2, NH, NH_2, NH_3, HCN, CH_3CN	CO^+, CO_2^+, H_2O^+, OH^+, CH^+, CN^+, N_2^+, C^+
Metals	
Na, K, Al, Mg, Si, Ca, Ti, V, Cr, Mn, Fe, Co, Ni, Cu	Ca^+

Table 7.5. Planets: mean surface temperatures and atmospheres

Planet	Surface temperature (K)	Gas Pressure (bars)	Gas	Volume percent
Mercury	440	$2 \cdot 10^{-15}$	He	42
			N	42
			O	15
			H	1
Venus	730	90	CO_2	96.4
			N_2	3.4
			SO_2	0.015
			H_2O	0.010
			Ar	0.007
Earth	288	1.013	N_2	78.804
			O_2	20.946
			Ar	0.934
			H_2O	variable
Moon	257	$2 \cdot 10^{-14}$	Ne	40
			Ar	40
			He	20
Mars	218	0.007	CO_2	95.7
			N_2	2.7
			Ar	1.6
Jupiter	165	$\gg 100$	H_2	82
			He	17
			Other	1
Saturn	140	$\gg 100$	H_2	82
			He	17
			Other	1
Uranus	57	$\gg 100$	H_2	85
			He	15
Neptune	57	$\gg 100$	H_2	85
			He	15
Pluto	42	10^{-5}	N_2	100

96% CO_2, 3.5% N_2

100 × ⊕

- Venus — dense atmosphere — 100 bars — mostly CO_2 — essentially no H_2O ← no erosion on Venus surface — very old geological structures are preserved for millions – billions of years, unlike on ⊕

- Earth — The Water Planet, oceans, rivers, lakes, water vapor in atmosphere, polar ice sheets, etc.

has been suggested that Mars N hemisphere once an ocean

- Moon — devoid of atmosphere — dry

600 Pa = $\frac{1}{167}$ × ⊕

- Mars — tenuous atmosphere — mostly CO_2 — but imagers have provided evidence of erosional features ⟹ water must have been present in the past
Example — dendritic stream channels
Mars does have CO_2 – rich polar caps — may be some water ice there too — also may be a deep water-ice permafrost — or it may be all gone — we're not sure.

hang MOLA map of Mars

Also — why does the ⊕ have water & the moon not?

How can Mars have once had water and not have any now?
Answer — water in the (upper) atmosphere can be dissociated by UV radiation

$$2H_2O + UV \rightarrow 2H_2 + O_2$$

FIGURE 14.46 Water-formed channels on Mars. *(NASA)*

Part of Nanedi Vallis, a channel thought to have been cut by flowing water early in Mars' history. The image shows an area 9.8 km by 18.5 km (Malin Space Science Systems/NASA MGS).

Mars

The light H_2 can escape by literally ~~boiling off~~ evaporating off the planet

Governed by the <u>Jeans escape law</u>
(Sir James Jeans, The Dynamical Theory of Gases, 1904)

The mean <u>square</u> velocity of a molecule in a gas depends only upon the (absolute) temperature T. For a diatomic gas such as H_2 or O_2

since 5 degrees of freedom

$$\frac{1}{2} m \langle v^2 \rangle = \frac{5}{2} kT$$

where m = mass of molecule
$T = T(°C) + 273$ ← measured in K
k = Boltzmann's constant
$= 1.38 \cdot 10^{-23}$ J/K

Multiply both sides by Avogadro's number

$$N = 6.022 \cdot 10^{23} \frac{\text{molecules}}{\text{mole}}$$

$$\frac{1}{2} \underbrace{mN}_{\frac{\text{kg}}{\text{molecule}} \times \frac{\text{molecules}}{\text{mole}}} \langle v^2 \rangle = \frac{5}{2} \underbrace{kN}_{\text{this is the gas constant}} T$$

$= M$ (atomic wt in $\frac{\text{kg}}{\text{mole}}$)

this is the <u>gas constant</u>

$$R = kN = 8.31 \frac{J}{K \text{ mole}}$$

$$\langle v^2 \rangle = \frac{5RT}{M}$$ or

$$\boxed{\langle v^2 \rangle^{1/2} = \sqrt{\frac{5RT}{M}}}$$ — maybe call this v_{rms}

For H_2 on the ⊕ ($T_{observed} = 290\,K$)

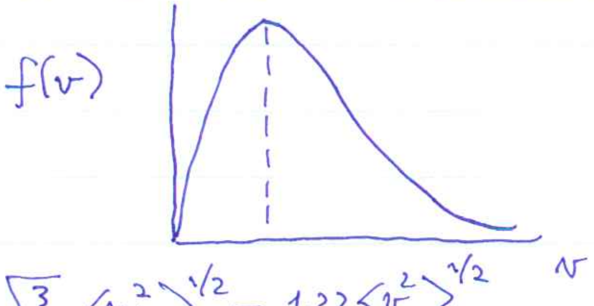
$$M = 2\,\frac{gm}{mole} = 2 \cdot 10^{-3}\,\frac{kg}{mole}$$

$$\boxed{\langle v^2 \rangle^{1/2} = 2.5\ km/sec}$$

This is just the rms velocity. There are molecules moving both faster and slower. The distribution of molecular velocities in an ideal gas is given by

$$f(v) = 4\pi v^2 \left(\frac{m}{2\pi kT}\right)^{3/2} e^{-\frac{1}{2}mv^2/kT}$$

the Maxwell-Boltzmann distribution

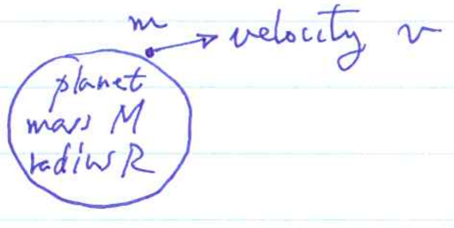


$$\int_0^\infty f(v)\,dv = 1$$

$$\sqrt{\frac{3}{2}}\,\langle v^2 \rangle^{1/2} = 1.22\langle v^2 \rangle^{1/2}$$

location of maximum = 3 km/s for H_2

We need to compare the thermal velocity with that required to <u>escape</u> from the planet. To deduce this we consider the energy of a particle of mass m in the gravitational field of a planet of mass M:

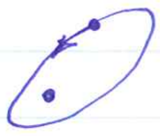


$$\text{Energy} = \underbrace{\tfrac{1}{2}mv^2}_{\text{kinetic}} - \underbrace{\frac{GMm}{r}}_{\text{gravitational potential}} = \text{constant, since energy is } \underline{\text{conserved}}$$

Note: the gravitational (binding) energy is <u>negative</u>.

example: cannonball fired from mtn top
(see figure 7.5 Emiliani)

There are three cases:



$E < 0$: elliptical orbit

$E > 0$: hyperbolic orbit

$E = 0$: parabolic orbit
(particle can escape)

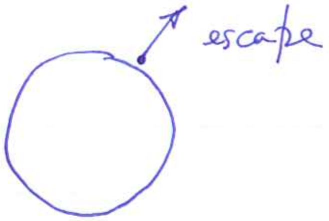
So to find the (minimum) escape velocity v_e we set

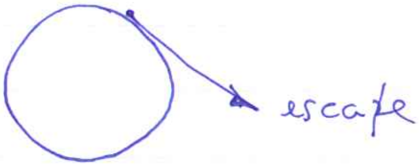
$$\frac{1}{2}mv_e^2 - \frac{GMm}{R} = 0$$

← for a particle at surface

$$\boxed{v_e = \sqrt{\frac{2GM}{R}}}$$

Note — it does not matter which <u>direction</u> the particle is moving (unless it is down)



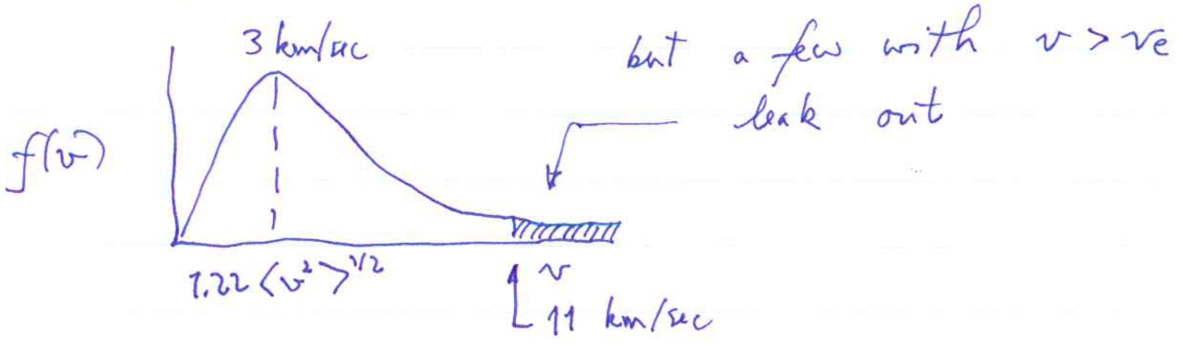


$$\boxed{\text{For the } \oplus : \quad v_e = 11.1 \text{ km/sec}}$$

↑ Needed to launch spacecraft

For the $\oplus$: $\dfrac{\langle v^2\rangle^{1/2}}{v_e} = \dfrac{2.5}{11.1} = 0.22$

Most H_2 molecules are unable to escape.



Can do the same calculation for the other terrestrial planets.

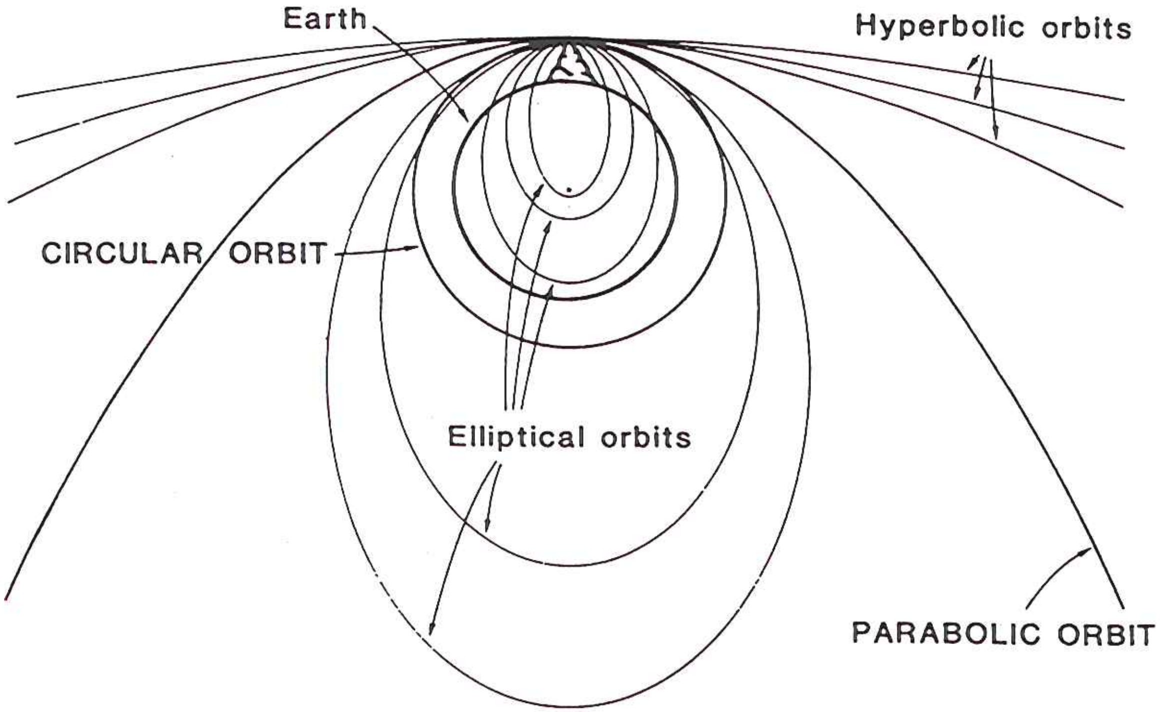


Figure 7.5. A cannonball shot horizontally from a mountaintop will reach farther the greater its initial speed, following segments of elliptical orbits as it falls back to Earth. If the cannonball is shot at a speed of 7.91 km/s, it will circle the Earth and return to the cannon. At speeds between 7.91 and 11.18 km/s, the ball will follow elliptical orbits of increasing ellipticity. At a speed of 11.18 km/s the ball will follow a parabolic orbit and escape. At speeds greater than 11.18 km/s the ball will follow hyperbolic orbits increasingly more open. (Emiliani, 1988, p. 119, Fig. 8.10.)

still has 0.01% H_2O in atmosphere → Venus

Planet	v_e (km/s)	T (K)	$\langle v^2 \rangle^{1/2}_{H_2}$ (km/s)	$\frac{\langle v^2 \rangle^{1/2}}{v_e}$
Mercury	4.3	700	3.7	0.88
Venus	10.3	740	3.8	0.37
Earth	11.1	290	2.4	0.22
Moon	2.4	380 ← due to ~~higher albedo~~ lower albedo	2.8	1.2
Mars	5.1	240	2.2	0.43

This is the reason for the absence of any H_2O on the small [illegible] bodies, Mercury & [illegible] the Moon

Rule of thumb – if $\langle v^2 \rangle^{1/2} > \frac{1}{3} v_e$ then essentially all the H_2O will be lost by ~~evaporation of H_2~~ photodissociation $2H_2O + UV \rightarrow 2H_2 + O_2$ and evaporation of H_2 in 4.5 b.y.

for terrestrial (inside Mars orbit) planets

The ⊕ owes its H_2O to its distance from the Sun ([illegible] this governs its temperature) and its size.

If it were Mars-sized ($v_e = 5.1$ km/s) it would have little or no water. Venus is ⊕-sized ($v_e = 10.3$ km/s) but is much hotter (because closer to Sun & because of runaway greenhouse effect); hence, it too is [illegible] (essentially) dry.

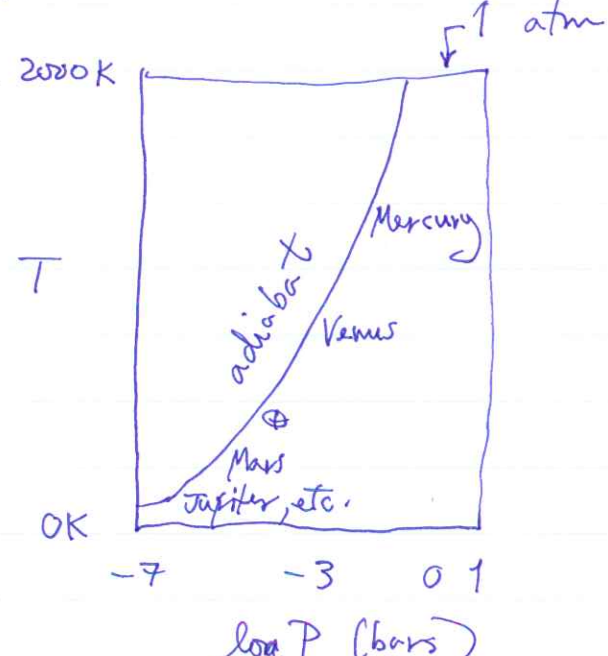
Planet	v_e (km/s)	T (K)	$\langle v^2 \rangle_{H_2}^{1/2}$ (km/s)	$\frac{\langle v^2 \rangle^{1/2}}{v_e}$
Mercury	4.3	700	3.7	0.88
Venus	10.3	740	3.8	0.37
Earth	11.1	290	2.4	0.22
Moon	2.4	380	2.8	1.2
Mars	5.1	240	2.2	0.43

This is not the whole story because we have only examined the loss mechanisms and not how the H_2O got there in the first place.

Condensation of planitesimals from the solar nebula (cloud of gas & dust surrounding the proto-sun)

~~Suppose we have such a disk shaped cloud, very hot, [illegible] above 2000 K. As it cools it will condense~~

The P–T conditions within the solar nebula 4.5 b.y. ago have been modelled



Other curves in Fig. 14.4 show the P–T conditions at which certain mineral phases begin to condense

Perovskite $CaTiO_3$
most refractory
Also Al_2O_3 corundum

found in chondritic meteorites →

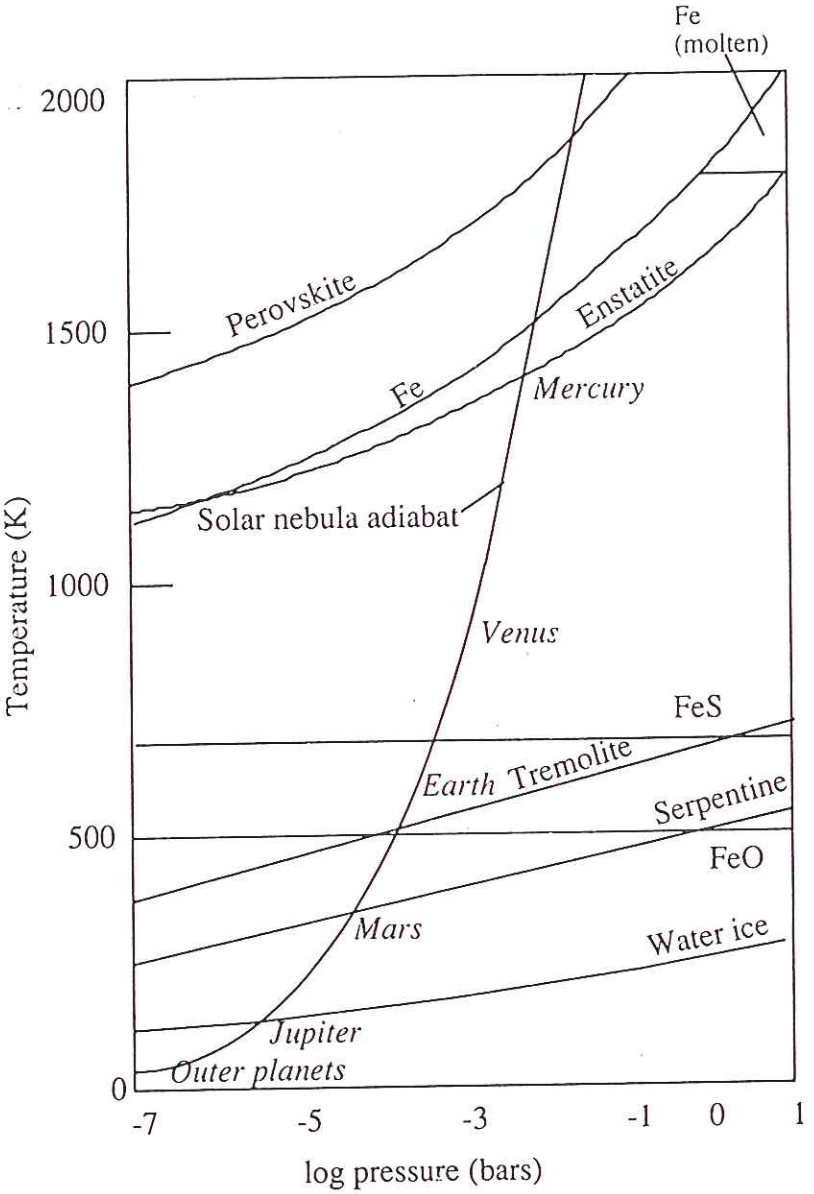


FIGURE 14.4 The equilibrium condensation sequence in a solar nebula. The lines mark the P–T conditions where various representative minerals begin to condense or are formed by a reaction with a preexisting condensate and the gas. An adiabat for the solar nebula is shown, with suggested condensation ranges of each planet, Mercury through Jupiter and the outer planets.

from Consolmagno

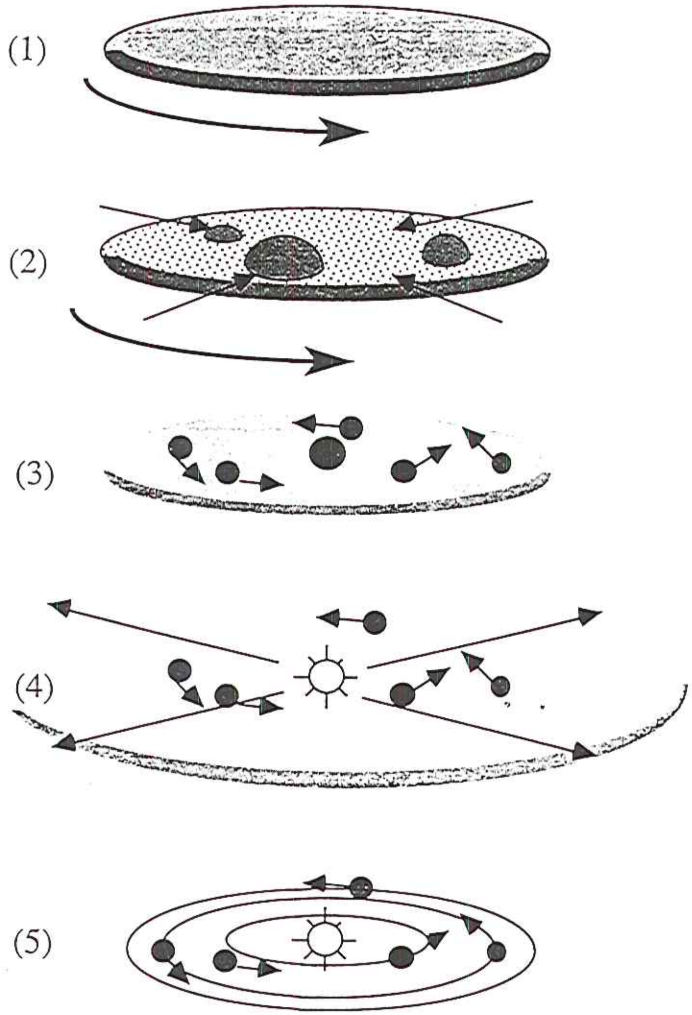


FIGURE 14.3 Five stages in the evolution of the solar nebula. (1) Starting as a disk-shaped cloud of gas and dust. . . (2) the cloud collapsed into fragments. . . (3) that began to orbit about the largest fragment, the proto-Sun. (4) As the Sun ignited, the energy from its solar wind and the pressure of its light combined to blow away the rest of the gas. . . (5) leaving behind rocky planetesimals, which would eventually collide and form into planets.

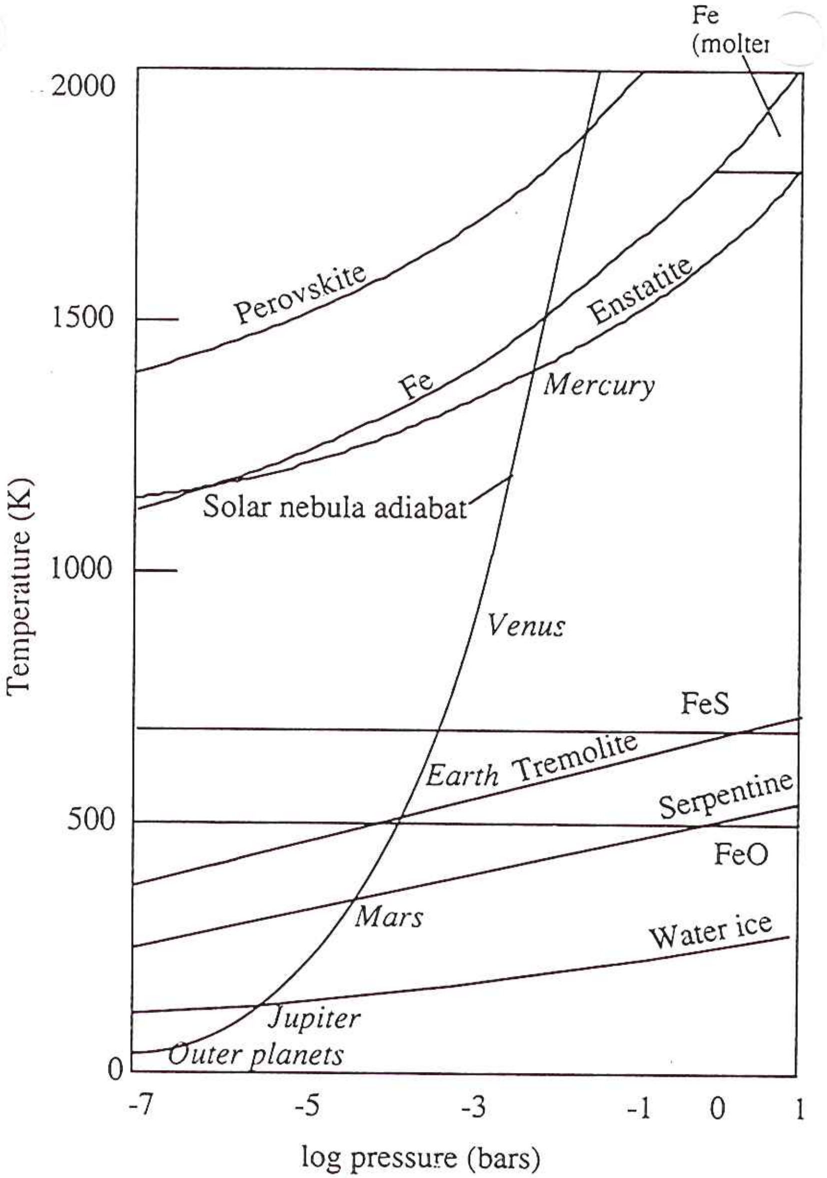


FIGURE 14.4 The equilibrium condensation sequence in a solar nebula. The lines mark the P–T conditions where various representative minerals begin to condense or are formed by a reaction with a preexisting condensate and the gas. An adiabat for the solar nebula is shown, with suggested condensation ranges of each planet, Mercury through Jupiter and the outer planets.

Fe condenses at 1000 – 1500K
FeS and FeO at 500 – 700 K

~~First H_2O in hydrous mineral phases such as~~

First silicates at ~ 1300K — enstatite $MgSiO_3$

First H_2O in hydrous mineral phases such as tremolite $Ca_2 Mg_5 Si_8 O_{22} (OH)_2$ — a Ca-rich pyroxene.

Amount of H_2O that can be incorporated into a planitesimal at this P, T is limited by availability of Ca.

First hydrated Fe, Mg silicates at ~ 400 K serpentine (asbestos) $Mg_3 Si_2 O_5 (OH)_4$. This can be the source of a lot of water on ⊕ but not on Venus. This another reason for Venus – ⊕ difference.

Not known for certain how much of ⊕'s H_2O from this source (hydrated Fe, Mg silicates in planitesimals that accreted to form ⊕, later released during magma ocean phase) versus arrival as comets 4.5 by ago.

At the low temperatures prevalent in the outer solar system ($T < 200\ K$) water ice can condense directly from the solar nebula.

This is the source of comets. Also important for the Galilean moons of Jupiter.

Moon	density (kg/m^3)	v_e ($\frac{km}{s}$)	$\langle v^2\rangle^{1/2}_{H_2}/v_e$
Io	3500	2.6	0.58
Europa	3000	2.0	0.75
Ganymede	1900	2.7	0.56
Callisto	1800	2.4	0.62

$\Uparrow$ all roughly moon-sized

$\Uparrow$ assuming $T = T_{Jupiter} = 120\ K$

$\langle v^2\rangle^{1/2}_{H_2} = 1.5\ \frac{km}{sec}$
$T = 120 K$

If the Galilean moons had been formed with the same H_2O content as the $\oplus$'s moon, they would have lost it all by evaporation now. But there is abundant evidence they are H_2O rich

Europa – water-ice crust 100 - 200 km thick
visual evidence of ice rafting, few craters
focus of search for life

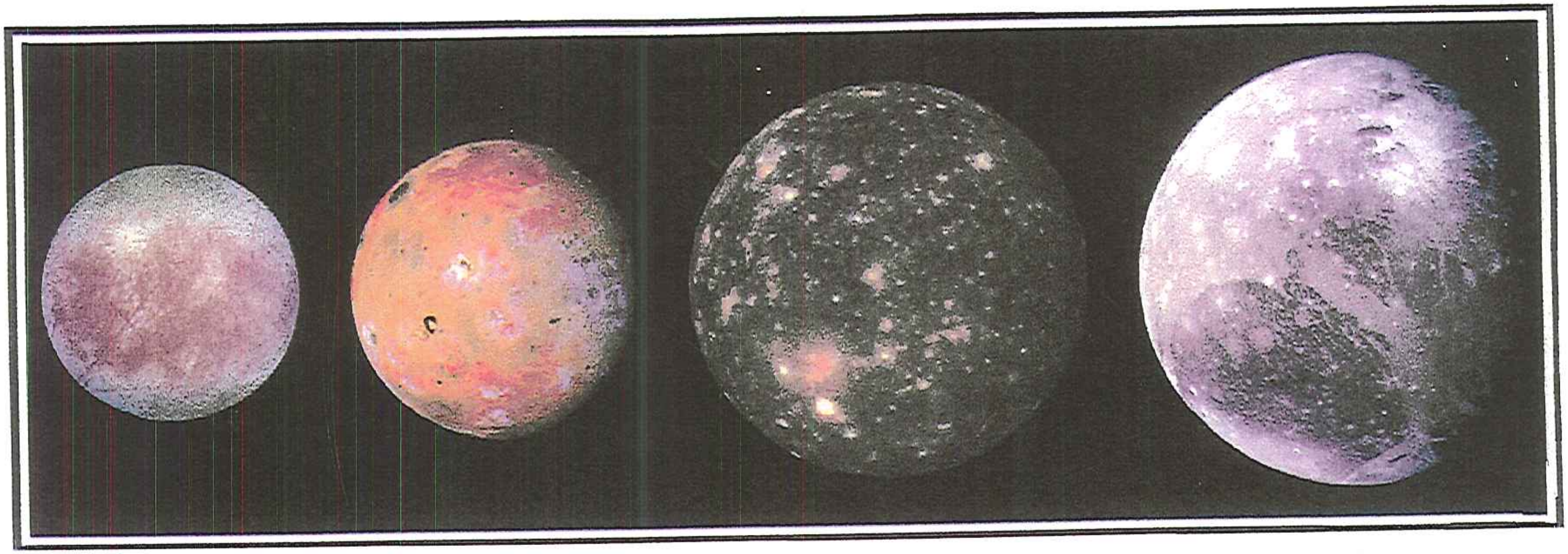
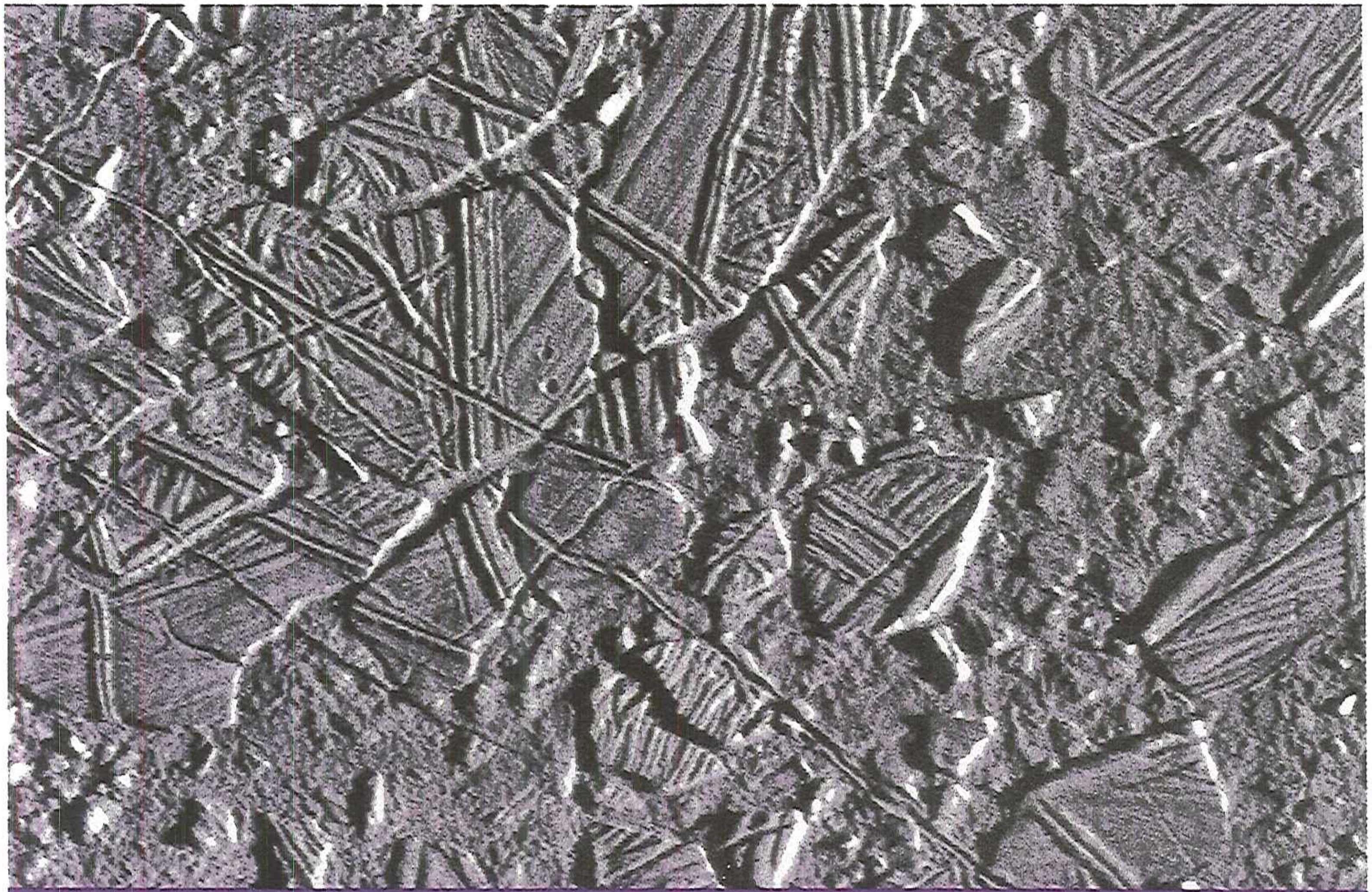


Plate I: The four Galilean moons
From left: Europa, Io, Callisto, and Ganymede



The ridged and faulted surface of Europa. This moon of Jupiter appears to have a thin skin of water ice that is a kilometre or so thick, overlying a widespread ocean of liquid water. Fracturing and relative movement of fragments of the ice seems to have happened repeatedly. The frame width is 42 km (NASA/JPL P48526).

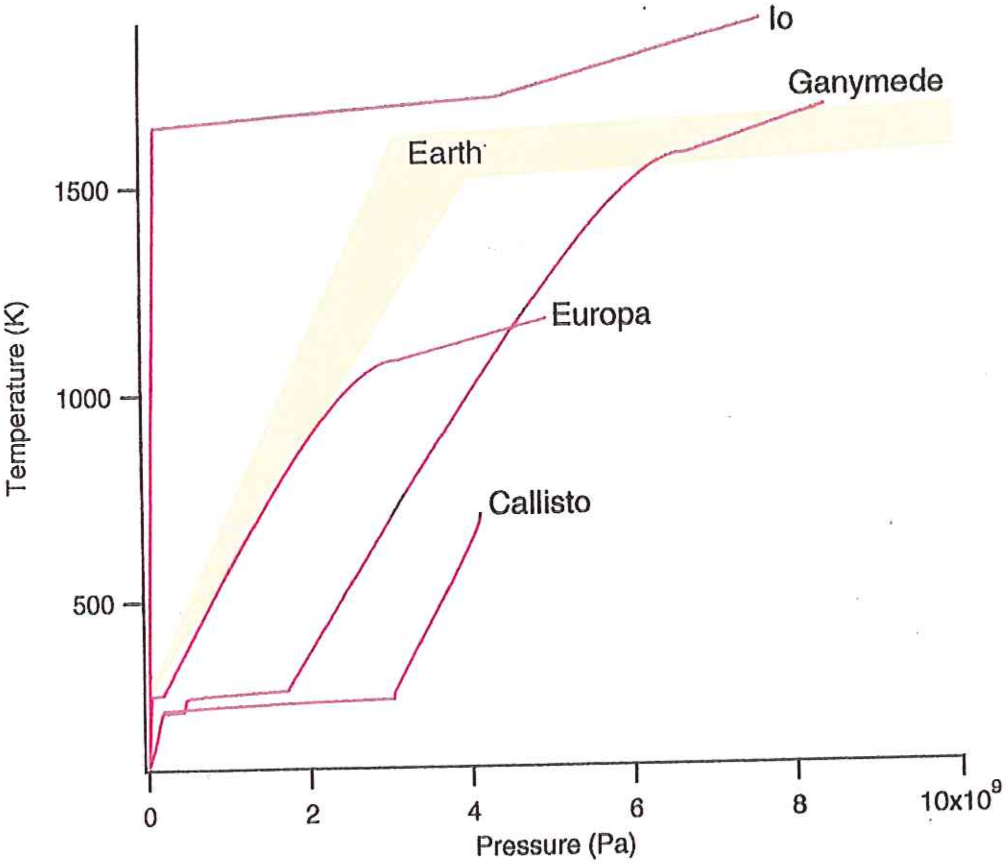


Figure 5.23: Thermal profiles of the moons of Jupiter and the Earth

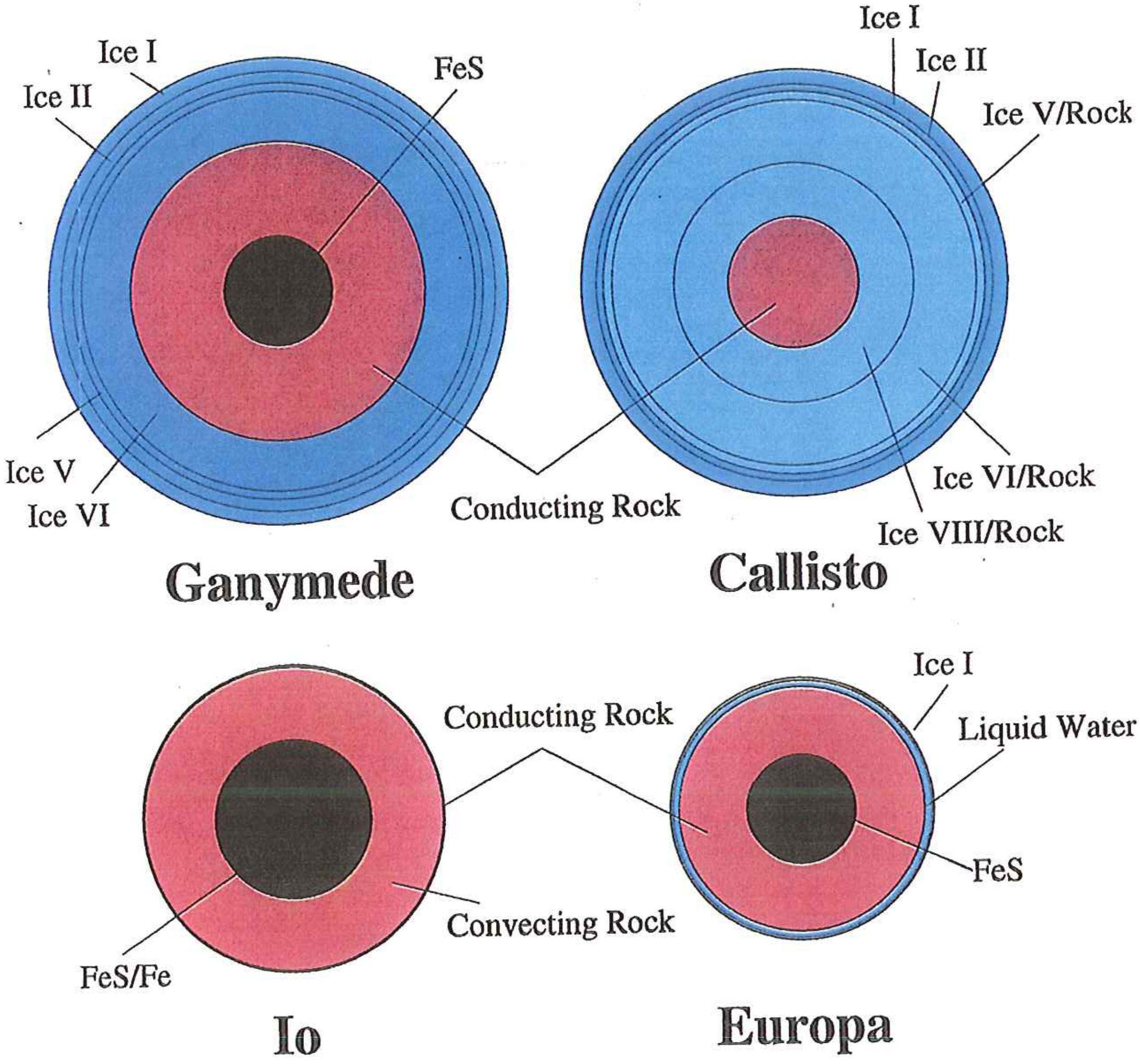
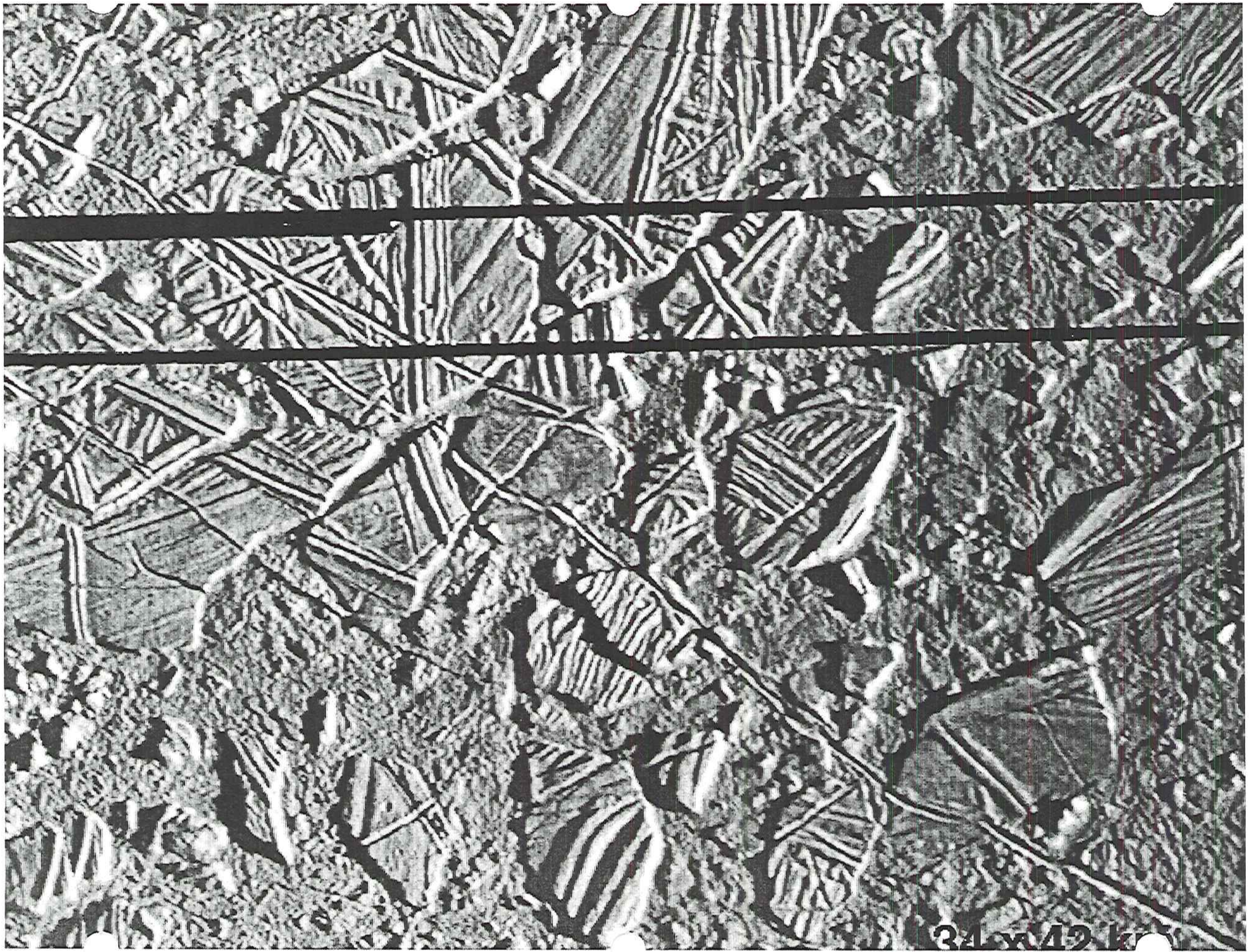


Figure 5.24: Schematics of the final models presented in figure 23 and table 5; all radii are to scale. Geotherms are taken from [117].

Table I: Fundamental Data on the Galilean Moons of Jupiter				
	Io	Europa	Ganymede	Callisto
Mass (MG) (km^3/s^2)	5959.91	3202.86	9886.6	7181
Radius(km)	1821.3	1565'	2634[3]	2403
Density (kg/m^3)	3529.13	3018	1936	1852
Moment (I/MR^2) +/-	0.378 0.007	0.330 0.014	0.3117 0.0026	0.359 0.005
Period (d)	1.769	3.551	7.155	16.69
v_e (km/s)	2.6	2.0	2.7	2.4
$\frac{\langle v^2 \rangle_{H_2}^{1/2}}{v_e}$	0.58	0.75	0.56	0.62

1 assuming $T = T_{Jupiter} = 120$ K



Europa ↑ Scale

rifted ice
on Europa

Ganymede & Callisto — densities ≈ 2× ice
heavily cratered icy crusts — ice-rock
aggregates below this

Io an anomaly — strong tidal heating gives rise to active silicic volcanism — very hot — all H_2O has been driven off.

We turn next to the question — what are the reservoirs of H_2O on $\oplus$:

Tables from Berner & Berner and Turekian — Similar table in Judson & Richardson, appendix.

All the compilations are a little different — some things hard to estimate or determine.

Let us make our own table, from first principles — learn how to estimate

Biggest reservoir by far is oceans.
Need to know ocean bathymetry

Mean ocean depth 3.9 km
Oceans cover 70% of $\oplus$ surface

$$3.9 \times 0.7 \times 4\pi \times (6371)^2 = 1400 \times 10^6 \text{ km}^3$$

from Berner & Berner — a similar compilation in Judson & Richardson Appendix

TABLE 1.1 Inventory of Water at the Earth's Surface

Reservoir	Volume 10^6 km^3 (10^{18} kg)	Percent of Total
Oceans	1400.	95.96
Mixed layer	50.	
Thermocline	460.	
Abyssal	890.	
Ice caps and glaciers	43.4	2.97
Groundwater	15.3	1.05
Lakes	0.125	0.009
Rivers	0.0017	0.0001
Soil Moisture	0.065	0.0045
Atmosphere total[a]	0.0155	0.001
Terrestrial	0.0045	
Oceanic	0.0110	
Biosphere	0.002	0.0001
Approximate total	1459.	

Sources: NRC 1986; Berner and Berner 1987.

[a] As liquid volume equivalent of water vapor.

TABLE 3–2 Mass of the Present Hydrosphere

	Total Mass (units of 10^{20} g)	Percentage of Total Hydrosphere
Oceans	13,700	80.0
Pore waters in sediments	3,300	18.8
Ice	200	1.2
Rivers, lakes	0.3	0.002
Atmosphere	0.13	0.0008
Total hydrosphere	17,200	100.0

Turekian

(NB: start drawing onion diagram here)

If spread equally over ⊕ surface, depth = 2.7 km

Ice caps & glaciers — need to know ice thickness

Antarctica is 85% of cryosphere

Roughly — 3 km thick × (area of Greenland & Antarctica) ≈ $40 \cdot 10^6$ km^3

Press & Siever
numbers with the
agree 2
Berner

Turekian estimate too low [illegible]

Only 80 meters (0.08 km) thick if spread out over whole ⊕ — 3% of hydrosphere

Lakes & rivers even smaller — only [illegible] 0.01% of hydrosphere — 0.30 meter thick layer

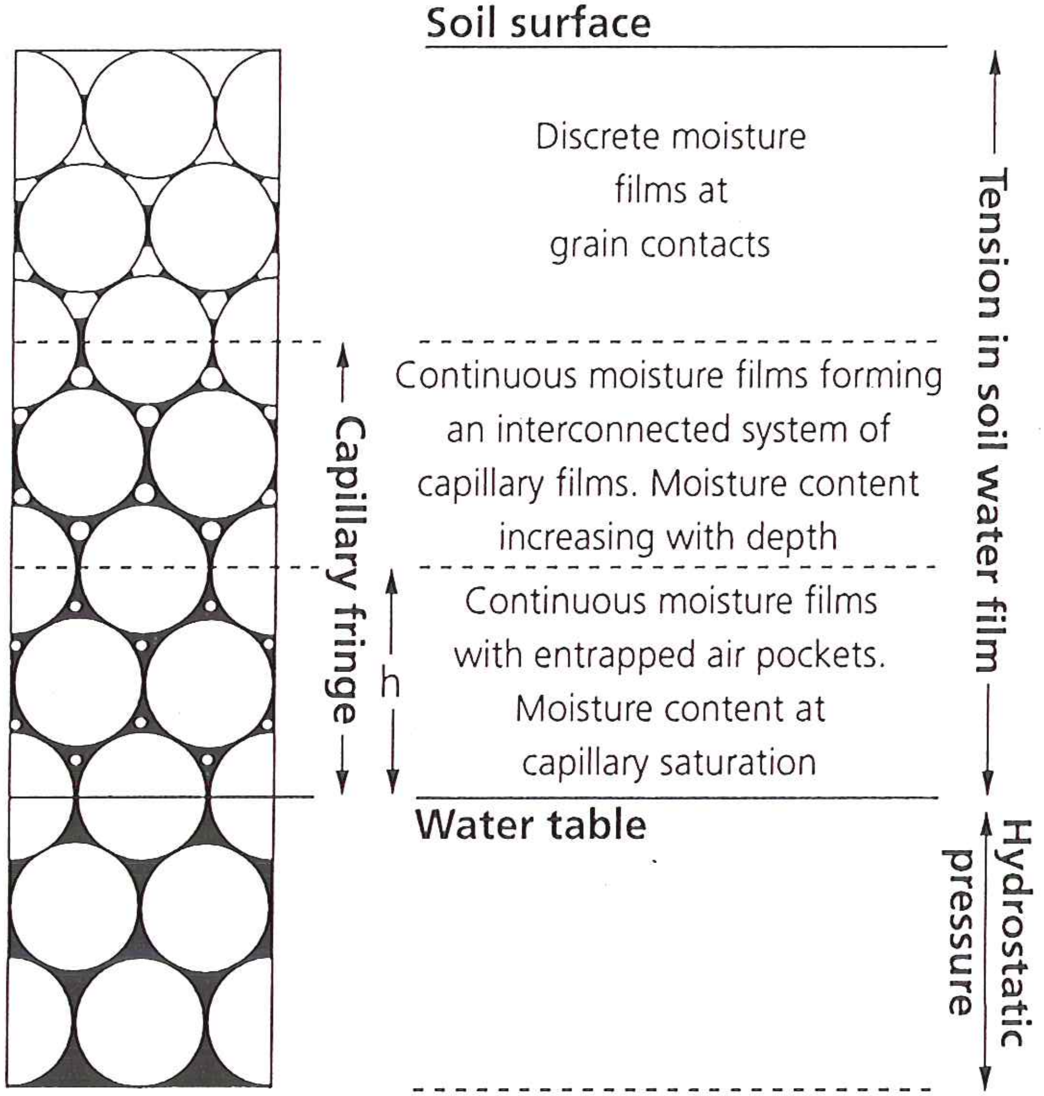
Ground water — two Yale estimates very different:

also Press & Siever

B&B: $15 \cdot 10^6$ km^3 = 1% = 0.03 km
Karl: $330 \cdot 10^6$ km^3 = 20% = 0.6 km

Depends on what you are estimating — pumpable fresh ground water or saline pore water, which may not be recoverable.

Turekian estimate more reasonable on following grounds



h=height of equivalent rise in a capillary tube

Fig. 3.1 The zones of moisture in regolith, showing the extent of the capillary fringe. After Carson 1969 [12].

$$\text{Porosity} = \frac{\text{pore volume}}{\text{total volume}}$$ measured in %

Table 4-4. Porosity and Bulk Density, Dry and Saturated, of Sedimentary Rocks*
(G. Edward Manger)

Formation	Age	Locality	Depth of sample, feet	Number of samples	Porosity, per cent			Sat'd. bulk density (average) g per cm^{-3}	Ref.	Porosity method: T (Total) A (Apparent)
					Aver.	Min.	Max.			
			SANDSTONE							
"Mount Simon" ss.	Cambrian	W. Va., Wood County	13,005–13,065	9	.7	.2	2.5	2.70	30	A
Southern "Potsdam" ss.	Cambrian	Wis.	quarry	14	11.4	4.8	28.3	2.41	5	A
Northern "Potsdam" ss.	Cambrian	Wis.	quarry	16	19.4	10.4	22.6	2.32	5	A
St. Peter Ss.	Ordovician	Ark., Ozark Plateau	outcrop	12	8.8	3.6	14.1	2.50	4	T
Bradford ss.	Devonian	Pa.	≈600–≈2300	297	15.0	6.0	23.3	2.40	7	T
					(assumed grain density, 2.65 g cm^{-3})					
Chemung Formation (ss's.)	Devonian	Pa.	≈1700–≈2300	49	9.5	4.5	22.2	2.51	20, 21	T
Berea Ss.	Mississippian	Ohio, W. Va.	0–2160	18	14.1	4.7	19.5	2.39	20, 21 3, 36	T, A
Atoka Formation (and other ss's.)	Pennsylvanian	Ark.:								
		Ozark Plat.	outcrop	17	12.3	4.7	19.8	2.44	4	T
		Ark. Valley	outcrop	35	7.8	0	20.6	2.51	4	T
		Ouachita Mts.	outcrop	25	5.1	0	10.4	2.56	4	T
Bartlesville sand	Pennsylvanian	Okla.	1570–2680	26	18.3	7.6	32.0	2.40	20	T
Bunter Ss.	Triassic	Gt. Britain	outcrop	18	20.4	5.8	30.8	2.29	22	A
Keuper Ss.	Triassic	Gt. Britain	outcrop	16	22.6	16.5	28.6	2.25	22	A
Woodbine sand	Cretaceous	Tex.	2436–3701	10	24.7	19.0	32.0	2.25	28	A
Sandstones and siltstones	Cretaceous	Montana, eastern	outcrop	22	33.7	22.6	38.3	2.17	34	T
Sandstones	Cretaceous	Wyo.	0–3187	38	19.7	8.8	27.0	2.32	20, 21	T
Sandstones	Miocene	Switzerland	quarry	15	18.7	13.3	22.1	2.37	11	T
					(dips 7° or less)					
			LIMESTONE, DOLOMITE, CHALK, AND MARBLE							
Ellenburger Group (1s. and dol.)	Ordovician	Tex., Llano County	outcrop	57	3.0	.1	12.6	2.75	10	T
Beekmantown Group (dol.)	Ordovician	W. Va., Wood County	10,531–11,945	56	.4	.1	1.1	2.80	30	A
Black River Ls.	Ordovician	Ontario	quarry	11	.46	.07	1.67	2.72	26	A
Niagara Dolomite	Silurian	Wis.	quarry	14	2.9	.5	6.7	2.77	5	A
Marl	Carboniferous	Russia	subsurface	19	8.2	..	..	2.63	24	T
Oolites	Jurassic	Gt. Britain	outcrop	5	14.6	5.5	24.0	2.44	22	A
Limestones	Jurassic	Switzerland	quarry	114	3.6	.4	25.6	2.66	11	T
					(dips 8° or more)					
Glen Rose Ls.	Cretaceous	Tex.	20.5–30.5	10	16.8	16.0	18.8	2.37	28	A
Chalk	Cretaceous	Gt. Britain	outcrop	3	28.8	17.6	42.8	2.23	22	A
Limestone	Cretaceous	Switzerland	quarry	29	4.3	.4	18.3	2.65	11	T
					(dips 10° or more)					
Green River Fm. (marlstone)	Eocene	Colo.	mine	11	2.9	.2	12.0	2.26	36	A
			SHALE, CLAYSTONE, AND SLATE							
Shale	Pennsylvanian	Okla.	1000	..	17	..	..	2.42	1	graph
			3000	..	7	..	..	2.59	1	graph
			5000	..	4	..	..	2.66	1	graph
Shales	Cretaceous	Wyo., Mont.	outcrop	9	29.5	23.8	37.6	2.17	31	T
Shale, nearly horizontal and undisturbed	Oligocene and Miocene	Venezuela	≈600	6	33.5	31.3	35.8	2.06	15	T
			≈2500	9	25.4	22.9	28.9	2.25	15	T
			≈3500	9	21.1	17.8	25.6	2.35	15	T
			≈6100	3	9.6	9.1	10.6	2.52	15	T
			≈7850	2	10.4	10.3	10.4	..	15	T
			SAND, CLAY, GRAVEL, ALLUVIUM, AND SOILS							
Cape May Fm. (sd.)	Pleistocene	N.J.	mostly pits	12	38.9	30.8	45.3	1.93	34	T
Loess soil	Quaternary	Idaho	surface	3	61.2	53.2	69.4	1.61	34	T
Fine sand	Quaternary	Calif.	sea-floor sediments,	54	46.2	..	..	1.93	12	†
Very fine sand			0–1 inch below the	15	47.7	..	..	1.92	12	†
Sand-silt-clay			depositional surface	3	74.7	..	..	1.44	12	†
Mud	Quaternary	Hudson River	submerged crate	..	88.2	..	..	..	19	†

TABLE B.6 Average Composition of the Crust[a]

	AV. IGNEOUS ROCK, %	AV. SHALE, %	AV. SANDSTONE, %	AV. LIMESTONE, %	WEIGHTED-AV. CRUST,[b]%
SiO_2	59.12	58.11	78.31	5.19	59.07
TiO_2	1.05	0.65	0.25	0.06	1.03
Al_2O_3	15.34	15.40	4.76	0.81	15.22
Fe_2O_3	3.08	4.02	1.08 }	0.54	{ 3.10
FeO	3.80	2.45	0.30 }		{ 3.71
MgO	3.49	2.44	1.16	7.89	3.45
CaO	5.08	3.10	5.50	42.57	5.10
Na_2O	3.84	1.30	0.45	0.05	3.71
K_2O	3.13	3.24	1.32	0.33	3.11
H_2O	1.15	4.99	1.63	0.77	1.30
CO_2	0.10	2.63	5.04	41.54	0.35
ZrO_2	0.04	—	—	—	0.04
P_2O_5	0.30	0.17	0.08	0.04	0.30
Cl	0.05	—	Tr[c]	0.02	0.05
F	0.03	—	—	—	0.03
SO_3	—	0.65	0.07	0.05	—
S	0.05	—	—	0.09	0.06
$(Ce, Y)_2O_3$	0.02	—	—	—	0.02
Cr_2O_3	0.06	—	—	—	0.05
V_2O_3	0.03	—	—	—	0.03
MnO	0.12	Tr[c]	Tr[c]	0.05	0.11
NiO	0.03	—	—	—	0.03
BaO	0.05	0.05	0.05	0.00	0.05
SrO	0.02	0.00	0.00	0.00	0.02
Li_2O	0.01	Tr[c]	Tr[c]	Tr[c]	0.01
Cu	0.01	—	—	—	0.01
C	0.00	0.80	—	—	0.04
Total	100.00	100.00	100.00	100.00	100.00

---

[a] After F. W. Clarke and H. S. Washington, "The Composition of the Earth's Crust," *U.S. Geol. Surv. Prof. Paper* 127, p. 32, 1924.

[b] Weighted average: igneous rock, 95%; shale, 4%; sandstone, 0.75%; limestone, 0.25%.

[c] Trace.

Shale is most important sedimentary rock (95% of total)

Average shale porosity 20%

Say 3 km of sedimentary pore spaces filled (on both continents & seafloor)

$$0.2 \times 3 = 0.6 \text{ km of } H_2O$$

Deeper pore water (in crystalline rocks)

Say porosity = 5% for 3 km ≤ depth ≤ 9 kms

$$\Rightarrow \text{another } (0.05)(6) = 0.3 \text{ km of } H_2O$$

This is H_2O just held in the pore spaces.

Also H_2O in hydrous mineral phases (clays, mica, amphibole, etc.)

Average wt. % of H_2O in a (dry) rock is 1.3% (Judson & Richardson)

Average crustal thickness

$$\underbrace{35 \text{ km } (0.3)}_{\text{continents}} + \underbrace{10 \text{ km } (0.7)}_{\text{oceans}} = 18 \text{ km}$$

1.3 wt% = 0.013 × 2.7 (specific gravity)

= 3.5% by volume

0.035 × 18 = another 0.6 km of bound water

Finally, there is bound water in mantle rocks — though estimates of how much are very uncertain.

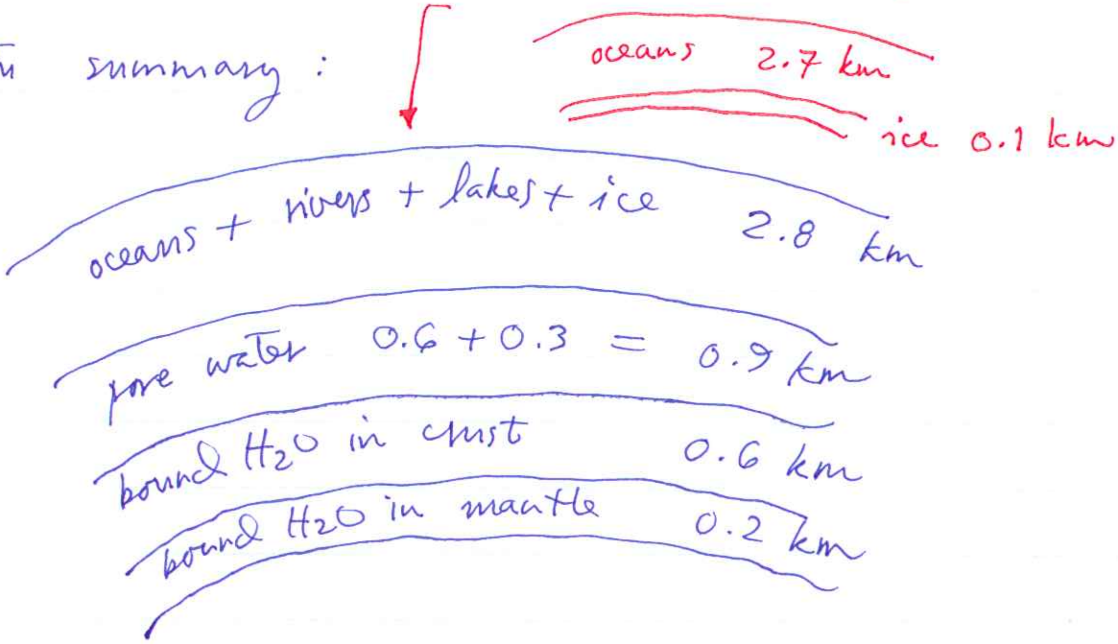
Recall, basalts are 1 wt% = 0.1 × 2.8
= 3 vol% H_2O
If they form by, say, 1% partial melting of mantle peridotite, then it must be (0.03)(0.01) = 0.03% H_2O

If this H_2O extends to a depth of 600 km below the Moho, i.e., to transition zone

(0.0003)(600) = 0.2 km of mantle H_2O

Nature of stable H_2O-bearing high-pressure minerals is a research topic of great current interest — e.g., Prof. Duffy in this Dept.

In summary :


could break top up into:

By this tally the total reservoir is
$4.5 \text{ km} = 2300 \times 10^6 \text{ km}^3$

The oceans are only 60% of the total

This is a reminder that there might be water stored in the crust of Mars.