

The volatile component of rocky planets (hydrospheres and atmospheres)

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G. Vladilo

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The Earth's Hydrosphere

The Earth is the only planet of the Solar System with surface liquid oceans

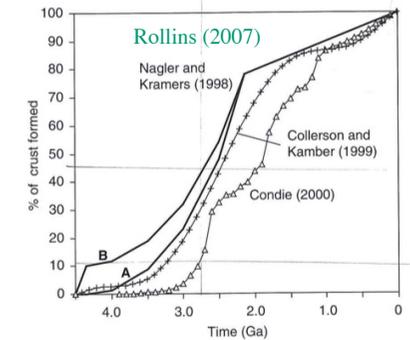
Current ocean coverage: ~70%

Typical ocean depth: 3 km

Current continent coverage: ~30%

The fraction of continents has increased over geological time due to the emergence of continents as a result of tectonic activity

Water is present in three different phases: liquid, vapour and ice



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Hydrospheres of rocky planets

- **Mercury**
 - No hydrosphere
 - Not clear if it has ever accreted water given its vicinity to the Sun
- **Venus**
 - Currently does not have a hydrosphere
 - Water was possibly present in the early stages of its history

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Earth water reservoirs

Rollins (2007)

TABLE 5.2 Estimates of the fraction of water and mass of water in the different mantle reservoirs. Also shown are estimates of water balance in the modern mantle.

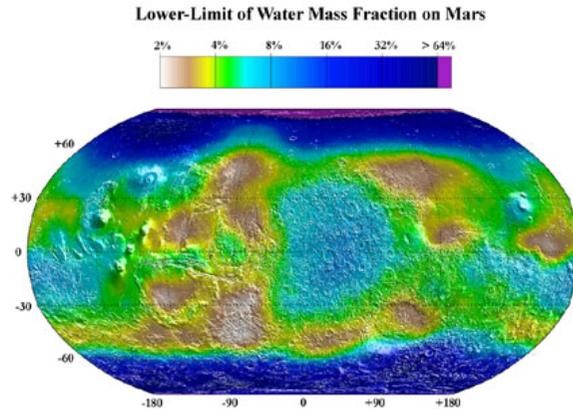
Reservoir	Fraction of water %	Mass of water	Reference
Water budget			
Surface water	100	1.4×10^{21} kg	Lecuyer et al. (1998)
Sedimentary rocks	0.11	0.23×10^{21} kg	Lecuyer et al. (1998)
Upper mantle	0.02–0.03	$0.2–0.3 \times 10^{21}$ kg	Data from Saal et al. (2002)
Mantle transition zone			
Lower mantle	0.5–1.0	$2.6–5.2 \times 10^{21}$ kg	Ohtani (2005)
Total Mantle	0.15–0.2	$3.4–4.5 \times 10^{21}$ kg	Ohtani (2005)
Water balance			
Total water returned to mantle (subduction)	1.83×10^{12} kg/yr		Jarrard (2003)
Total outgassing (arc + ridge)	2.0×10^{11} kg/yr		Peacock (1990)

Water in the outer reservoirs (1.4×10^{21} kg) make $\sim 2 \times 10^{-4}$ of Earth mass
The mass fraction of water has an optimal value that avoids the extremes of an ocean planet and a dry planet

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Mars water content

Water-equivalent hydrogen content of sub-surface water-bearing soils derived from the *Mars Odyssey* neutron spectrometer



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Rocky planet atmospheres

- Atmospheres have a strong influence on the climate and habitability
- Processes that shape the atmospheres of rocky planet:
 - gravitational capture of gas from the protoplanetary disk
important for massive planets during the epoch of planet formation
 - emission of gas from the surface
evaporation, sublimation, loss from the interior, and de-absorption
 - loss of the atmosphere to space

Primary atmosphere:

generated at the epoch of planetary formation

Secondary atmosphere:

created after the loss of the primary atmosphere

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Mars volatile reservoirs

Water (H₂O) Reservoir	Equivalent Global Ocean Depth
Atmosphere	10 ⁻⁵ m
Polar caps and layered terrains	5–30 m
Ice, adsorbed water, and/or hydrated salts stored in the regolith	0.1–100 m
Deep aquifers	Unknown
Carbon Dioxide (CO₂) Reservoir	Equivalent Surface Pressure
Atmosphere	~6 mbar
Carbonate in weathered dust	~200 mbar per 100 m global average layer of weathered dust
Adsorbed in regolith	<200 mbar
Carbonate sedimentary rock	~0 (at surface)
Sulfur Dioxide (SO₂) Reservoir	Equivalent Global Layer Depth
Atmosphere	0
Sulfate in weathered dust	~8 m per 100 m global average layer of weathered dust
Sulfate sedimentary rock reservoirs	Extensive, but not yet quantifiable

Radar evidence of subglacial liquid water on Mars (Orosei 2018)

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Vertical structure of planetary atmospheres

In a planet with a solid surface the atmospheric pressure, p , decreases with increasing height from the surface, z

If the atmosphere is in *hydrostatic equilibrium* the pressure decreases with an exponential law

$$p(z) = p_s \exp(-z/H)$$

where H is the scale height of the atmosphere.

Let us now express H in terms of atmospheric parameters

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Atmospheric scale height

- Derivation of H

- If the atmosphere is much thinner than the planet radius, the surface gravity acceleration, g , can be considered constant
- From the condition of hydrostatic equilibrium we obtain a relationship between the pressure, p , and the mean density, ρ

$$dp = -g \rho dz$$

In an *ideal* case of uniform temperature, T , from the law of perfect gases, $p = \rho kT/\mu$, we obtain

$$p(z) = p_s \exp(-z/H)$$

$$H = kT/\mu g$$

where $p_s = p(0)$ is the surface pressure and μ the *mean molecular weight*, which depends on the atmospheric chemical composition

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Atmospheres of rocky planets

Allen (2000)

Planet	Surface pressure p [bar]	Scale height H [km]	Surface gravitational acceleration g [m/s ²]
Mercury	3.7
Venus	90	15	8.9
Earth	1	8	9.8
Mars	~0.007	11	3.7

Huge differences in chemical composition, p and H make hard to predict the properties of rocky exoplanet atmospheres

Atmospheric scale height

- In the real case the atmospheric temperature is not uniform:

$$T = T(z)$$

and the relationship between surface pressure and height becomes

$$p(z) = p_s \exp(-z/H_z)$$

$$H_z = k T_h(z) / \mu g$$

where $T_h(z)$ is the *harmonic mean* of the temperature between the surface and the geometric height z

Since the temperature is in absolute scale (kelvin degrees), the percent variations of $T_h(z)$ can be relatively small over a significant part of the atmosphere

Therefore, to first approximation, the pressure follows an exponential decay even considering the vertical temperature gradient

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Chemical composition of rocky planet atmospheres

Gas	Venus [2]	Earth [2]	Mars [2]
N ₂	→ 0.035	→ 0.78084	0.027 ←
O ₂		→ 0.20948	1.3 × 10 ⁻³
CO ₂	→ 0.965	3.33 × 10 ⁻⁴	0.953 ←
CO	3 × 10 ⁻⁷	2 × 10 ⁻⁷	2.7 × 10 ⁻³
CH ₄		2.0 × 10 ⁻⁶	
NH ₃		4 × 10 ⁻⁹	
H ₂ O	2 × 10 ⁻⁵	~ 10 ⁻⁶	3 × 10 ⁻⁴
H ₂		5 × 10 ⁻⁷	
He	1.2 × 10 ⁻⁵	5.24 × 10 ⁻⁶	
Ar	7 × 10 ⁻⁵	9.34 × 10 ⁻³	0.016
Ne	1 × 10 ⁻⁵	1.818 × 10 ⁻⁵	2.5 × 10 ⁻⁶
Kr	3 × 10 ⁻⁹	1.14 × 10 ⁻⁶	3 × 10 ⁻⁷
Xe		8.7 × 10 ⁻⁸	8 × 10 ⁻⁸
H ₂ S	3 × 10 ⁻⁶	2 × 10 ⁻⁸	
SO ₂	1.5 × 10 ⁻⁴	1 × 10 ⁻⁹	

Allen (2000)

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Physical regimes of the atmosphere at low and high z

- **Mixing and diffusion**

Bakulin , Kononovic & Moroz 1984

- Low z

- *convection* and *turbulence* dominate
- the gas is well mixed
 - the chemical composition and mean molecular weight do not change significantly with height

- High z

- *diffusion* dominates because this process scales inversely with the density (concentration by number), which decreases with z
 - the mean molecular weight decreases with increasing z
 - H and He become the main constituents

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Physical regimes of the atmosphere at low and high z

- **Optical depth in the infrared band, τ_{IR}**

Bakulin , Kononovic & Moroz 1984

- Low z

$$\tau_{\text{IR}} \gg 1$$

- in thick atmospheres the lowest layers, with highest density, are optically thick to thermal radiation
 - thermal radiation is mostly due to surface heating by stellar radiation
 - also geothermal heating could be present

- High z

$$\tau_{\text{IR}} \sim 0$$

- the highest atmospheric layers are transparent to thermal radiation; the thermal radiation from underlying layers can escape to space

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Physical regimes of the atmosphere at low and high z

- **Mean free path of molecules, ℓ**

Bertotti & Farinella 1990

- Low z

$$\ell < H$$

- molecules can attain *local thermal equilibrium* through collisions
- the atmosphere is confined and can be described as a fluid

- High z

$$\ell > H$$

- there is no gravitational confinement of the gas
- the fastest molecules can escape to the interplanetary medium
- interplanetary molecules can be captured

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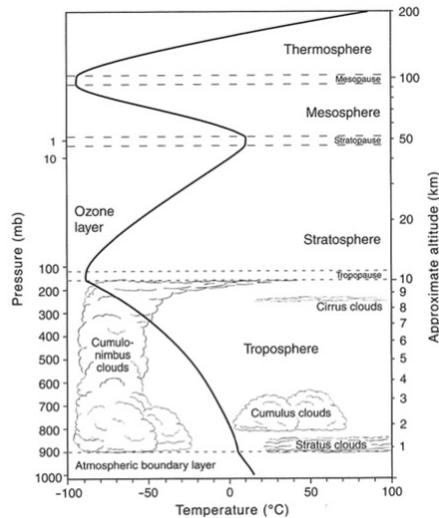
Vertical structure of the atmosphere

- The transitions between the above mentioned different regimes may take place at different heights, depending on the type of process under consideration
 - mixing/diffusion, molecular mean free path and infrared optical depth
- The lowest atmospheric layer where $\tau_{\text{IR}} > 1$ is called *troposphere*
 - It is characterized by a negative gradient of the temperature

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Vertical temperature distribution in Earth's atmosphere

- **Negative gradient in the troposphere**
 - The solar radiation penetrates the atmosphere and heats the surface
 - The lower atmospheric layers are opaque to thermal radiation emitted by the surface
 - The infrared flux is directed upwards, creating the observed negative gradient
- **Complex temperature profile in the highest layers**
 - The pressure and temperature must gradually merge with the values of the interplanetary medium



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Containment of the atmosphere

- The escape velocity from the planet plays a key role in the containment of the atmosphere

$$v_{\text{esc}} \propto (M/R)^{1/2}$$

M, R : planet mass and radius

- **Condition for the containment of the atmosphere**

$$\langle v_{\text{thermal}} \rangle \ll v_{\text{esc}}$$

The mean thermal velocity of atmospheric atoms or molecules must be significantly lower than the escape velocity

$$\langle v_{\text{thermal}} \rangle \propto (T/\mu)^{1/2}$$

Where μ is the atomic or molecular mass

For a given temperature, atoms and molecules with low μ will be lost more easily

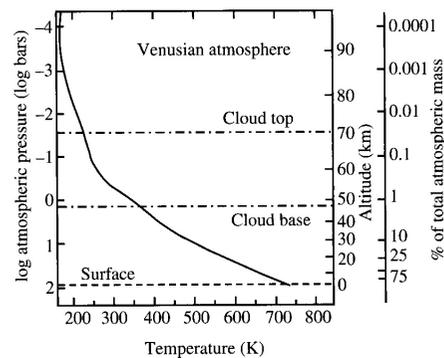
Planet	Equatorial escape velocity (km s ⁻¹)
Mercury	4.3
Venus	10.4
Earth	11.2
Mars	5.0
Jupiter	59.5
Saturn	35.5
Uranus	21.3
Neptun	23.7
Moon	2.4

Allen (2000)

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Vertical temperature distribution in Venus atmosphere

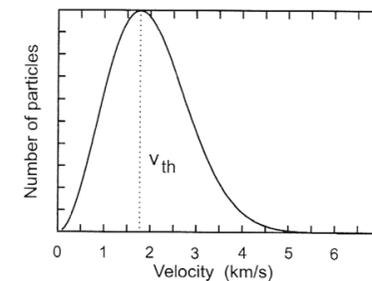
- Negative temperature gradient in the lower layers, as in the case of the Earth
- At the surface, p and T are significantly higher than in the case of the Earth
- At a height of ~50-60 km, p and T are roughly similar to the values found at the surface of the Earth



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Loss of atmospheric layers

- **“Jean escape” mechanism**
 - Loss of atmospheric molecules with velocities in the high-velocity tail of the Maxwellian distribution
 - Takes place in the external layers, where the mean free path becomes larger than the atmospheric scale height, i.e., $\ell > H$
 - Figure: Maxwellian distribution for H₂ molecules at $T=390\text{K}$



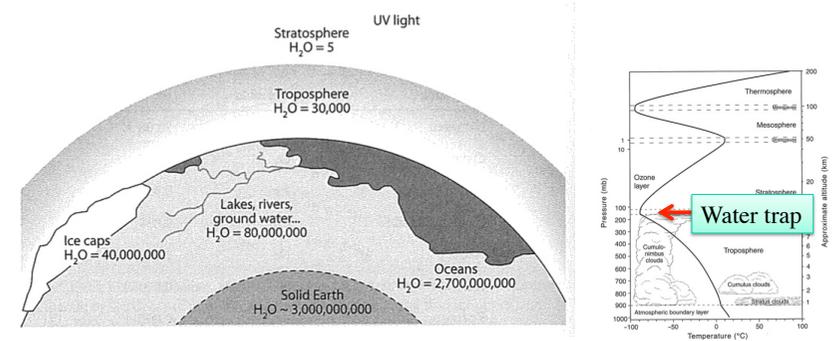
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Water vapour in the atmosphere

- Water vapour plays a key role in planetary atmospheres
 - It is a very effective greenhouse gas
 - Transformation between liquid and vapour phases provides a way of storing energy in the form of “latent heat”
- The atmospheric water vapour content depends on
 - presence of surface reservoirs of liquid water
 - existence of other sources of water vapour from surface or underground layers
 - mechanisms that prevent the loss of water vapour from the planet
 - atmospheric temperature

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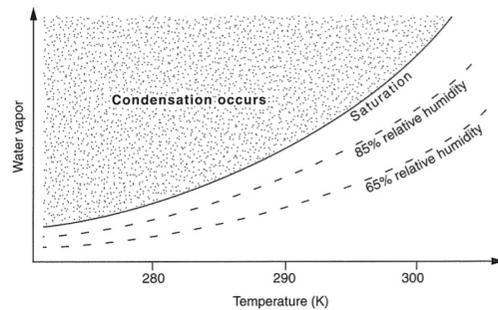
The water trap on Earth



Thanks to the negative temperature gradient in the troposphere, all water precipitates in the higher tropospheric layers, and none migrates to the upper atmosphere where it could be photodissociated and lost to space

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Water vapour and temperature



Clausius-Clapeyron equation (simplified form)

$$p_{sat}(T) = p_{sat}(T_0)e^{-\frac{L}{R_A}\left(\frac{1}{T} - \frac{1}{T_0}\right)}$$

R : gas constant; L : latent heat constant

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