Planetary atmospheres 2: their formation and evolution

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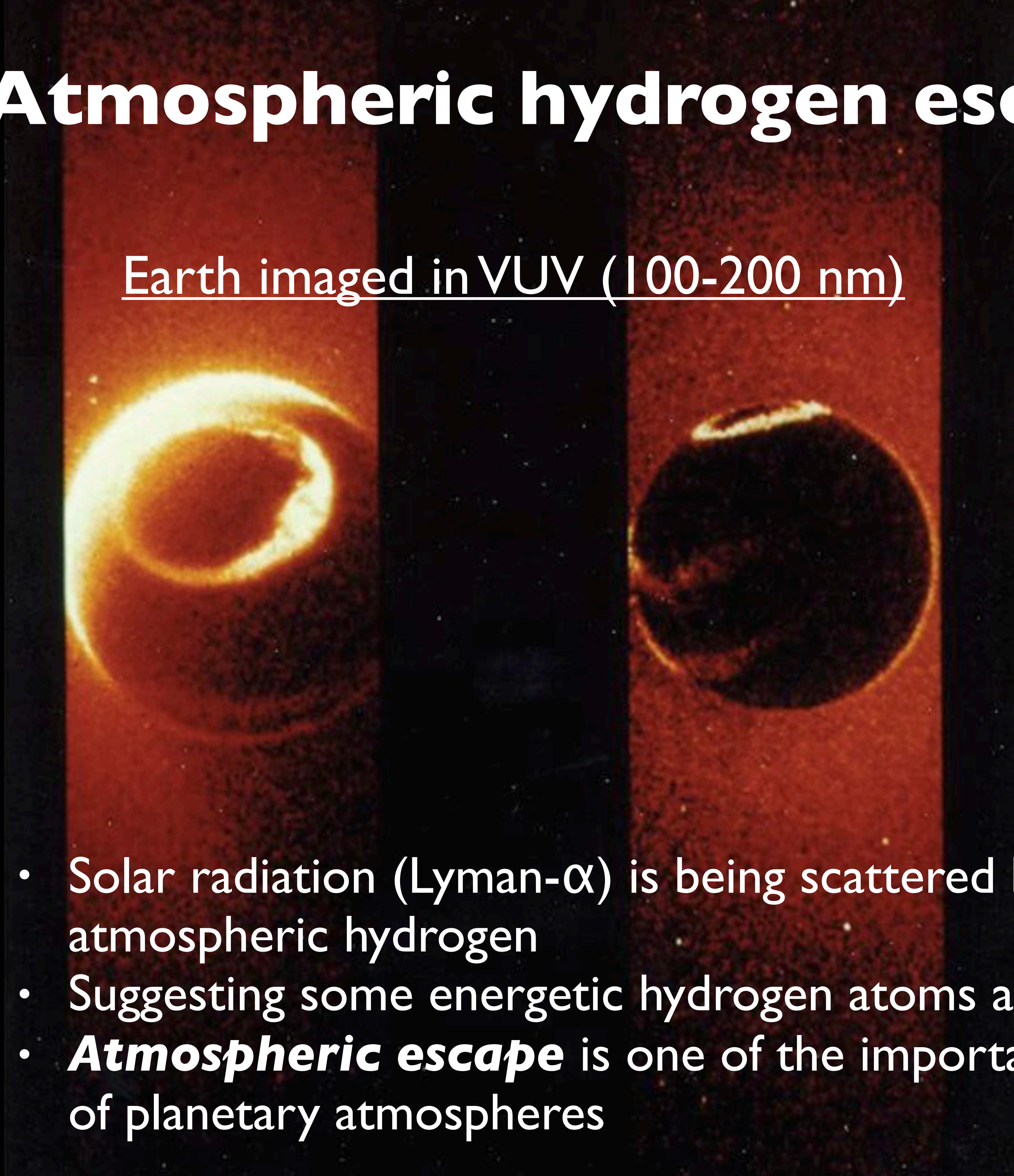
What we learn

In the 1st half of Planetary Atmosphere Lecture,
We learned that planetary climate is tied to the mass and composition of a planet's atmosphere.

To understand how climate has changed over time,
we must consider how the atmospheric mass and composition has evolved.

In this 2nd half,
I introduce the driving forces of atmospheric formation and evolution, and discuss the history of the atmospheres of terrestrial planets in our solar system

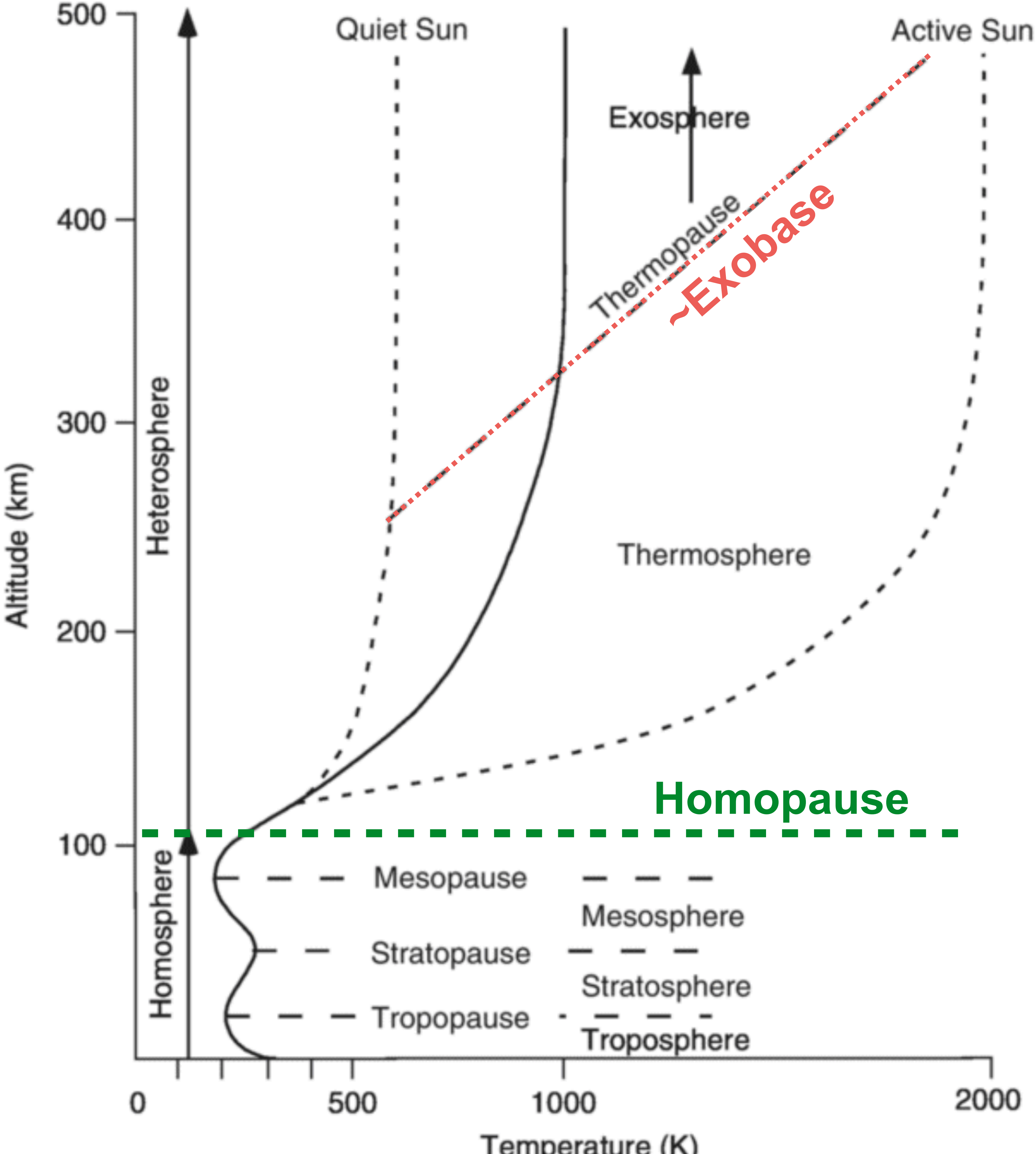
Atmospheric hydrogen escaping to space



<u>Earth imaged in VUV (100-200 nm)</u>

Rairden et al. (1986)

- Solar radiation (Lyman-α) is being scattered by Earth's extended atmospheric hydrogen
- Suggesting some energetic hydrogen atoms are escaping to space
- ***Atmospheric escape*** is one of the important drivers of the evolution of planetary atmospheres



Catling & Kasting (2017)

In an atmosphere, pressure and density decrease exponentially as a function of height (z),

$$p(z) = p_0 \exp\left(-\frac{z}{H}\right)$$

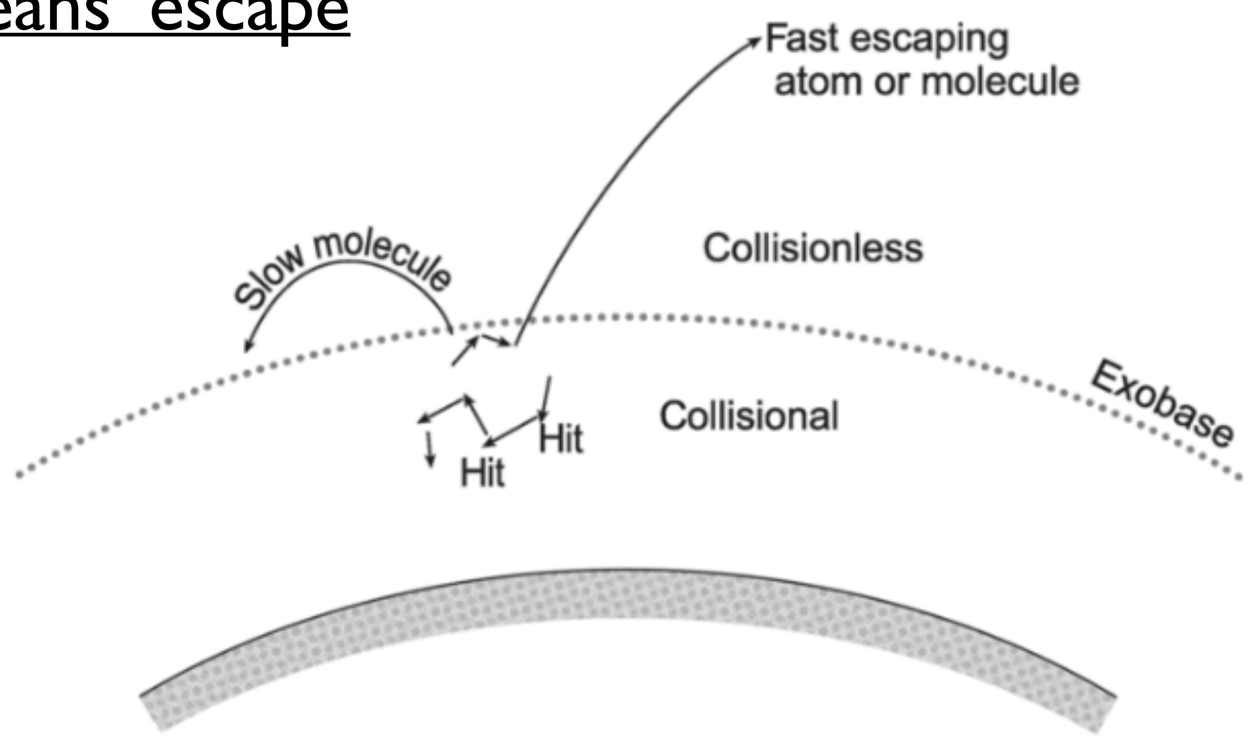
Where H is ***the scale height***, defined by,

$$H \equiv \frac{kT}{\overline{m}g}$$

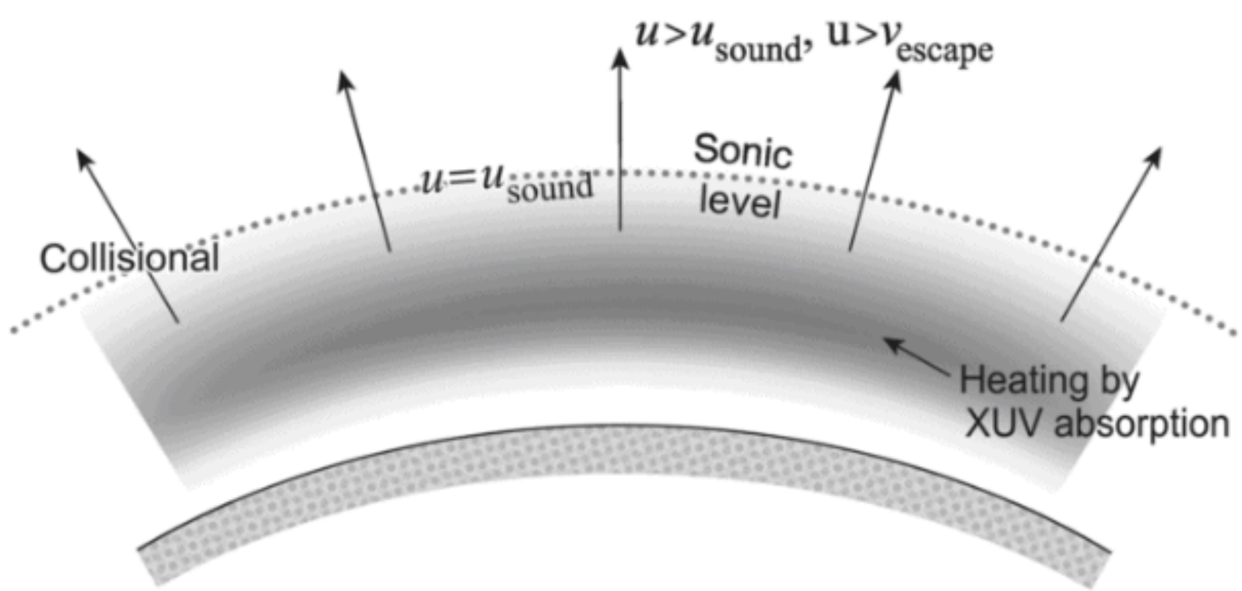
As you go to the upper atmosphere, the composition become heterogeneous (***homopause***). The atmospheric gas eventually becomes colisionless (***exobase***), and can escape to space

Thermal escape

Jeans' escape

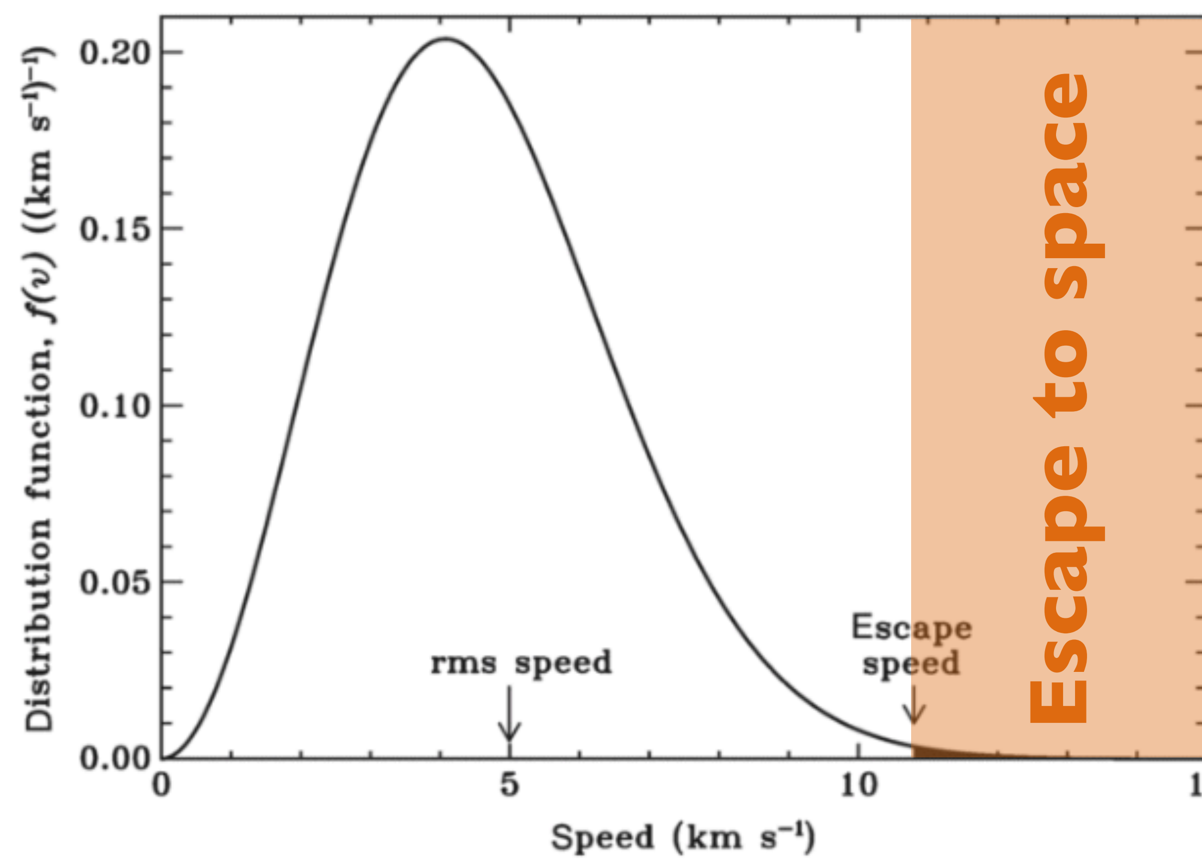


Hydrodynamic escape



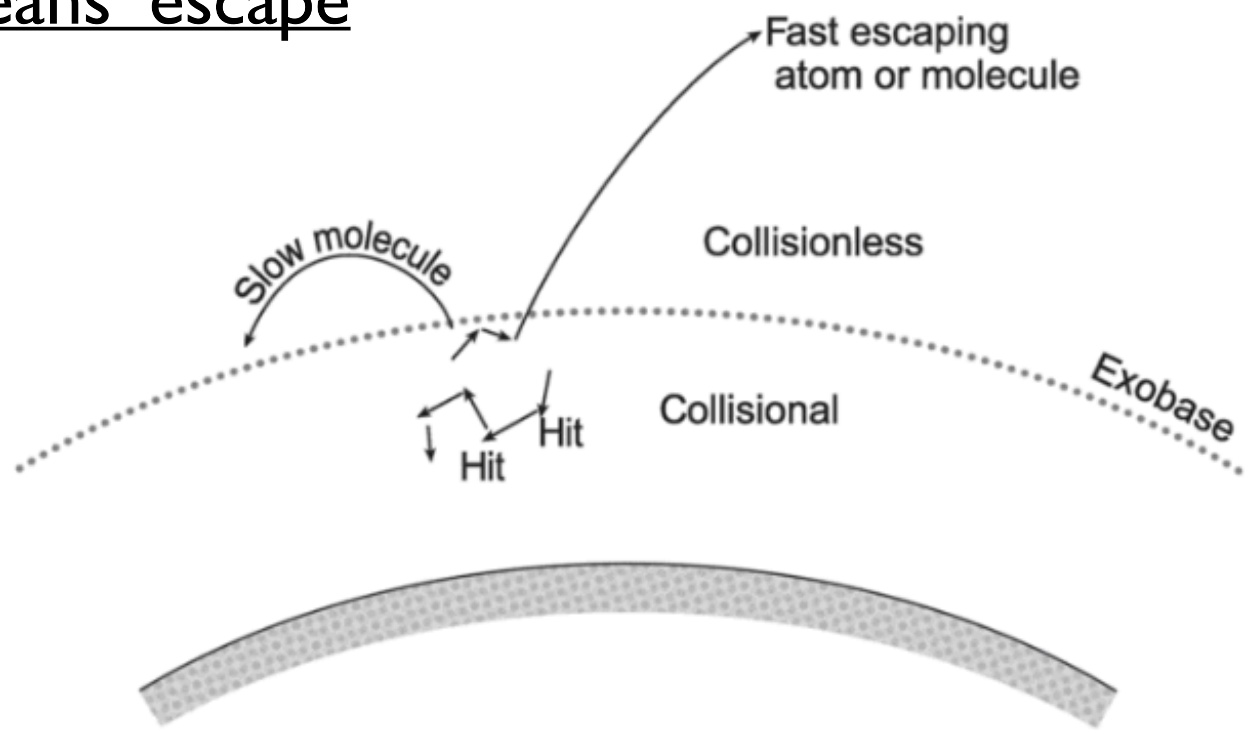
Catling & Kasting (2017)

Maxwell-Bolzmann speed distribution func.

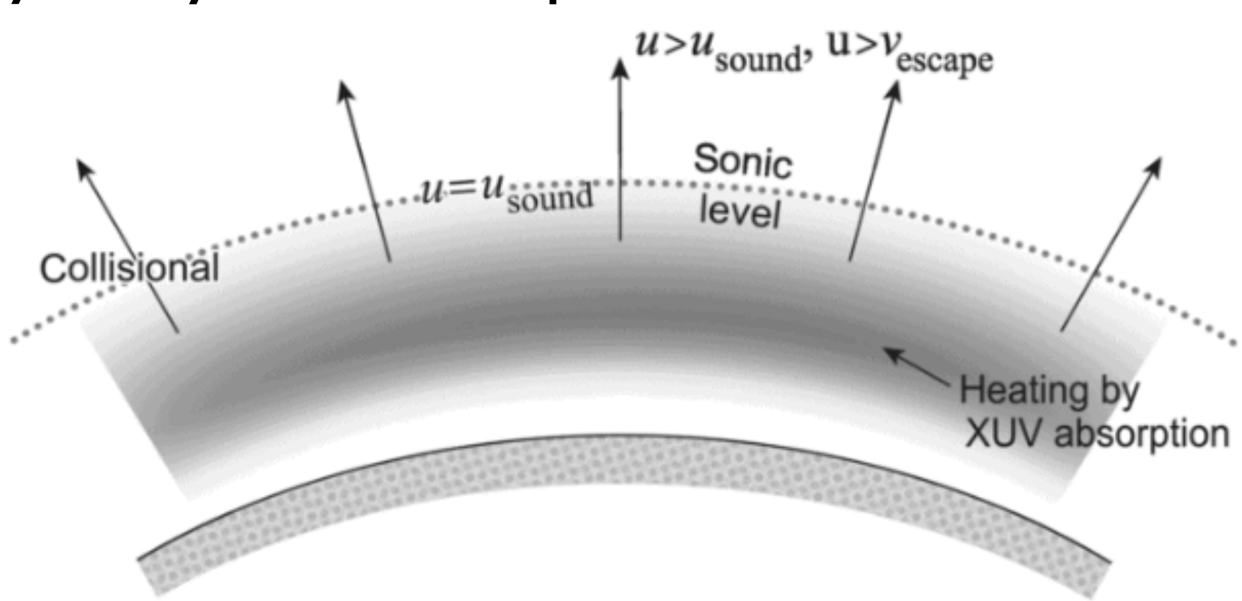


Thermal escape

Jeans' escape



Hydrodynamic escape



Catling & Kasting (2017)

Suppose hydrogen atoms at exobase.
The root-mean-square speed is given by,

$$\frac{1}{2}mv^2 = \frac{3}{2}kT$$

m: atomic weight 1.67×10^{-27} kg
v: RMS velocity
k: Bolzmann constant 1.38×10^{-23} J K^{-1}
T: temperature at exobase 1000 K

$$\longrightarrow \; v \simeq 5 \text{ km s}^{-1}$$

On the other hand, the escape speed is,

$$v_{\text{esc}} = 11.2 \text{ km s}^{-1}$$

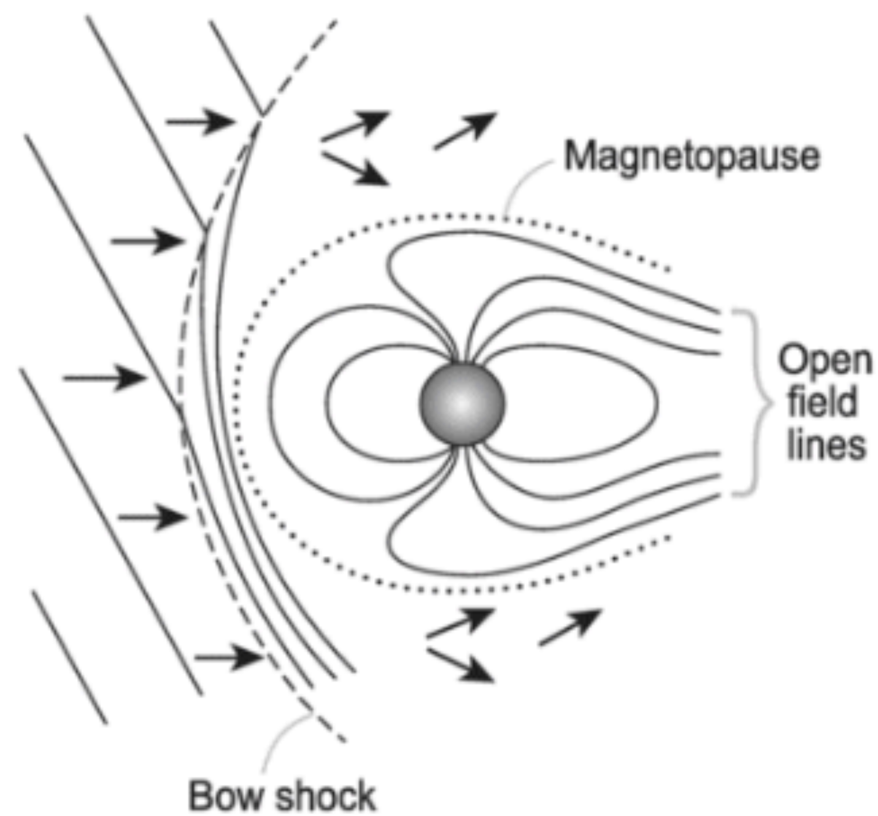
In Earth's present atmosphere, thermal escape rate of hydrogen can be approximated by Jeans' escape

Nonthermal escape of Earth's hydrogen

In the upper atmosphere, there exist hot (fast) H^+ ions ionized by solar UV

- **The polar wind:**
  Hot H^+ ions can escape from the magnetic poles

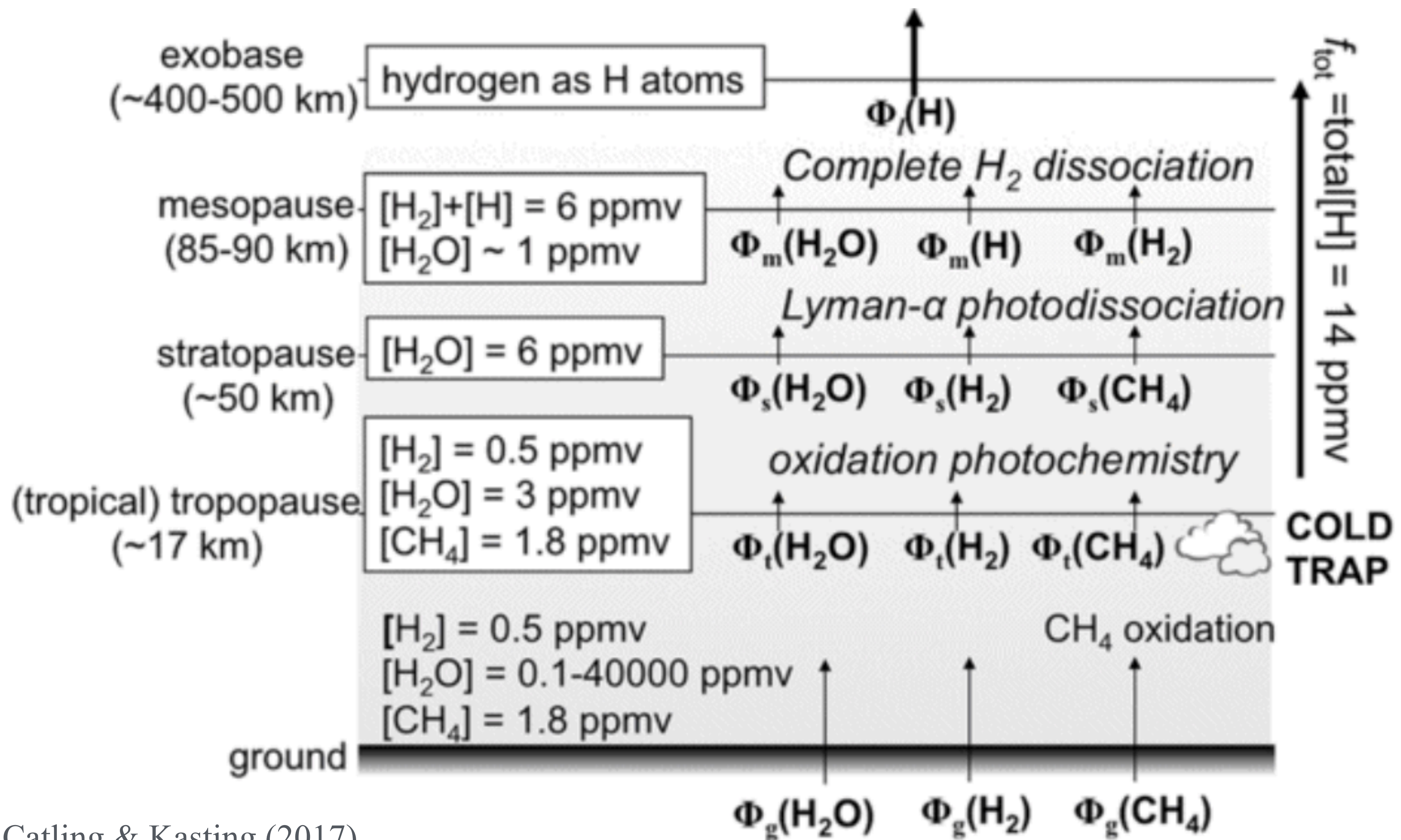
- **Charge exchange (H-H^+):**
  Hot H^+ ions exchange charge with H atoms, producing fast-moving H atoms which can escape to space



On present-day Earth, their contributions on H escape are,
60-90% for charge exchange, 10-40% for Jeans' escape, and 10-15% for the polar wind

"Diffusion-limited escape" concept

Because the processes to remove hydrogen from Earth's exobase are efficient, the total escape flux is controlled by the diffusion from the homopause (~100 km)



Catling & Kasting (2017)

“Diffusion-limited escape” concept

The rate of hydrogen supply from the homopause to the exobase by diffusion is given by,

$$\Phi_{\mathrm{dif},i} \simeq \frac{b_i f_i}{H_\mathrm{a}} \simeq 2.5 \times 10^{17} f_\mathrm{T}(\mathrm{H})\ \mathrm{m}^{-2}\mathrm{s}^{-1}$$

H_a : scale height of atmosphere

b_i : binary diffusion coefficient of i

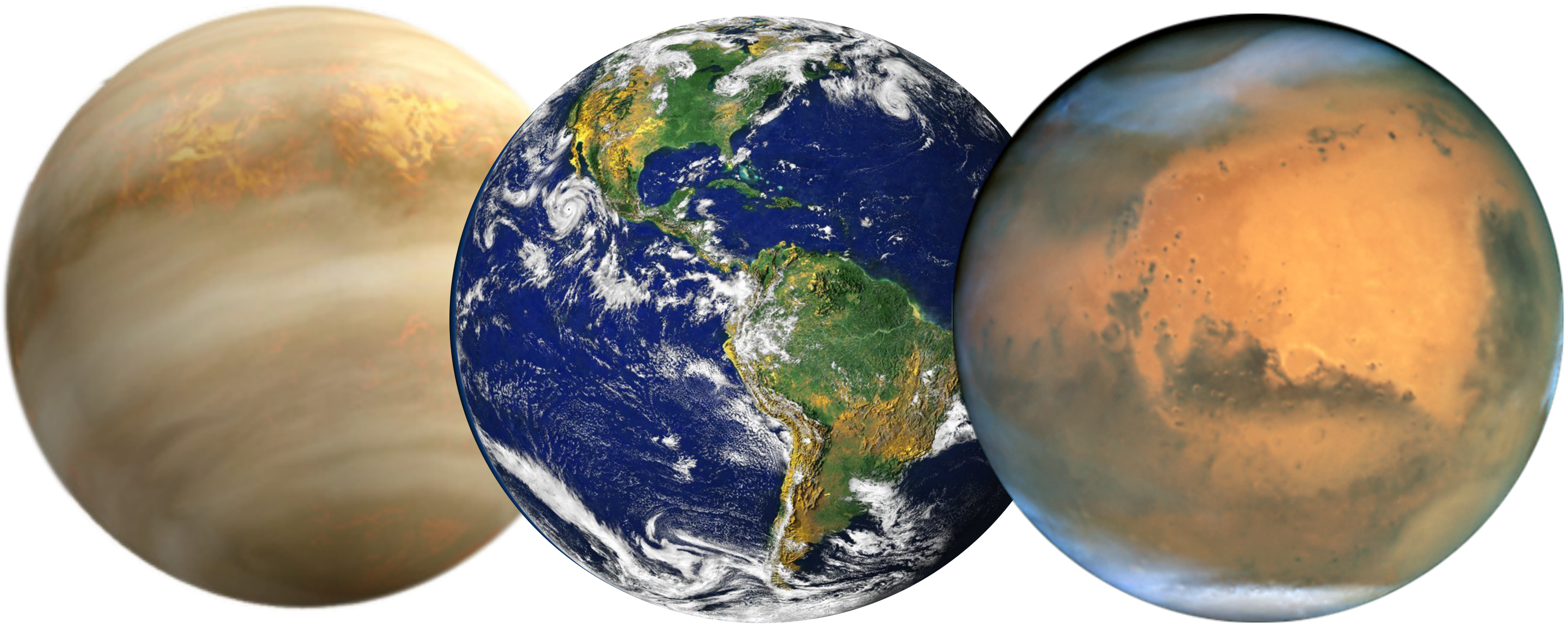
f_i : volume mixing ratio of i

$f_\mathrm{T}(\mathrm{H})$: total hydrogen mixing ratio

$$f_\mathrm{T}(\mathrm{H}) = f_\mathrm{H} + 2f_{\mathrm{H_2}} + 2f_{\mathrm{H_2O}} + 4f_{\mathrm{CH_4}} + \ldots$$

H escape flux from present Earth is small, because H_2O abundance at the tropopause is kept small (condensation and precipitation), which is called ***cold trap***

Surface environments of terrestrial planets

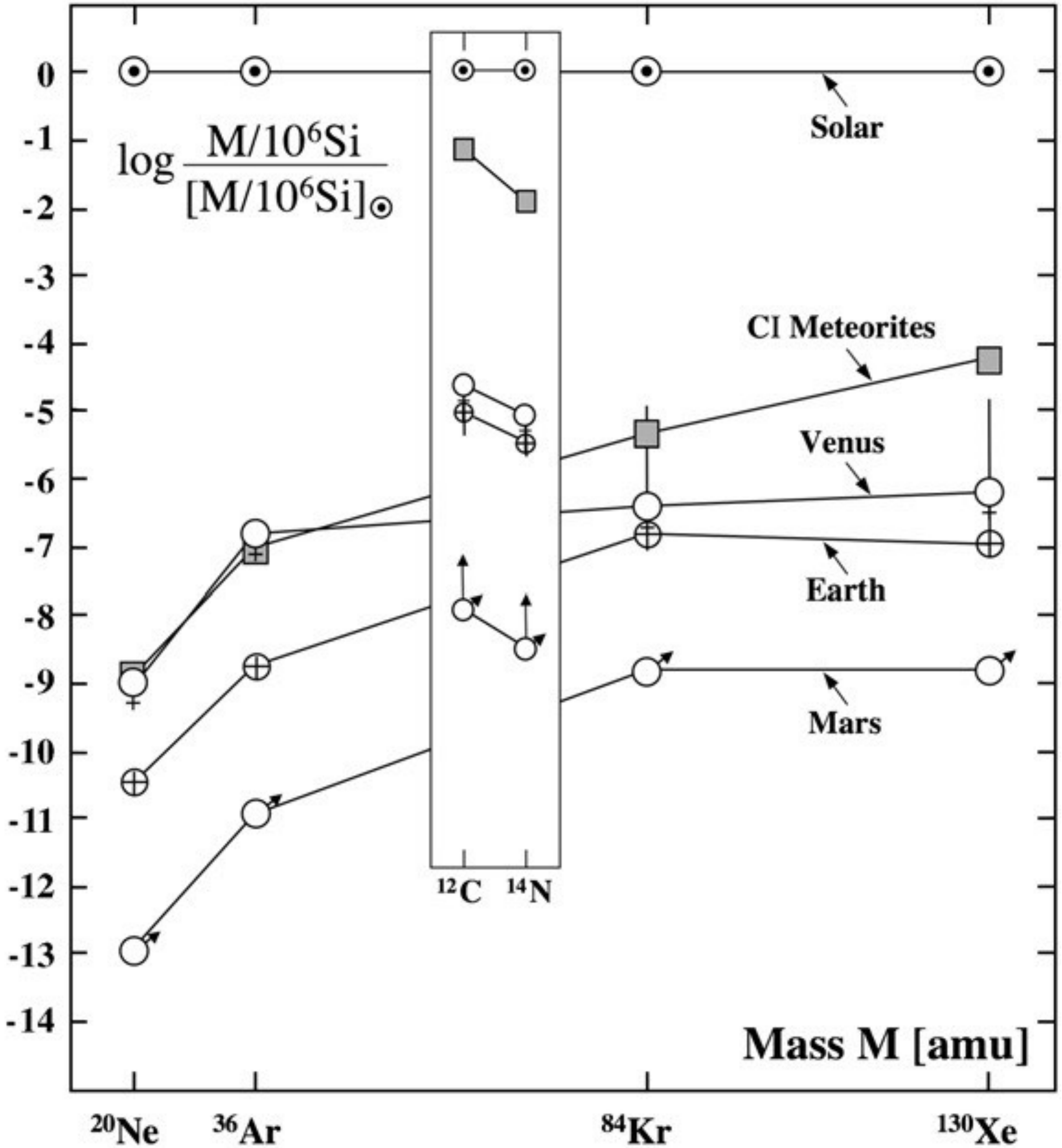


Orbital radius	0.7 AU	1 AU	1.5 AU
Surface pressure	90 bar	1 bar	0.006 bar
Major atm. composition	CO_2	N_2, O_2	CO_2
Surface temperature	735 K	288 K	210 K
Global equivalent depth H_2O	30 mm	2700 m	>20 m
Major H_2O reservoir	water vapor	seawater	Polar ice (+ subsurface ice)

Planets formed by impacts in protoplanetary disk

- **Two sources of atmospheres:**
  protoplanetary disk gas (primary atmospheres),
  volatiles delivered by impactors (secondary atmospheres)
- For impactors whose velocities exceed 5 km s^{-1}, which occur for ~1/10 Earth mass planets, an impact is followed by the release of volatiles (impact degassing) →
  Atmospheres should have formed at the same time with the volatile delivery during the planet formation
  (Matsui & Abe 1986; ABe & Matsui 1988)

Origin of atmospheres

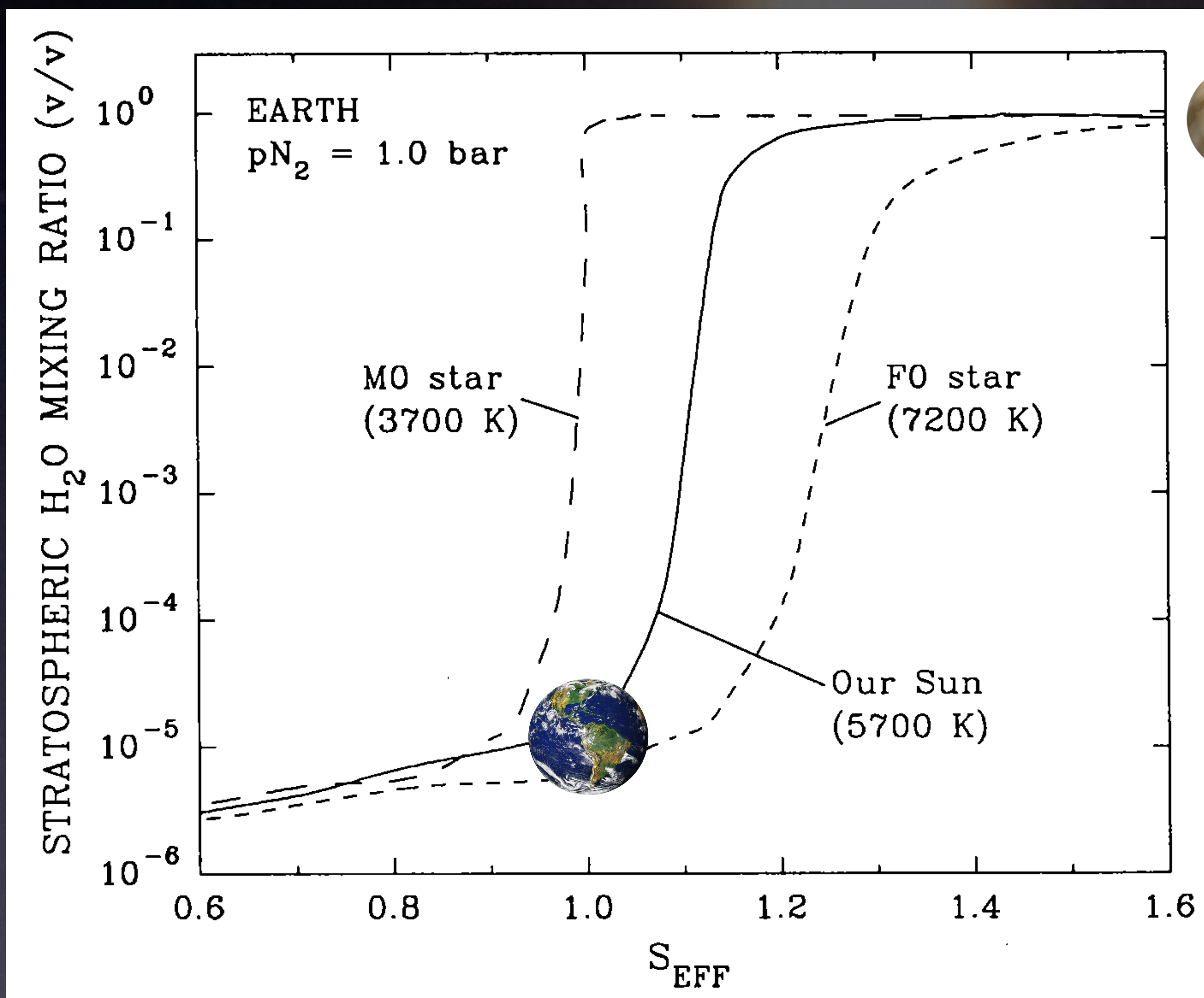


Pepin (2006)

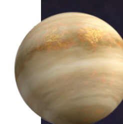
- **Depletion of noble gases (atmophiles) tell us the secondary origin**, presumably sourced by asteroids and comets
- **Hydrodynamic escape is likely to be responsible for the loss of the captured disk gas (primary atmospheres)** (Pepin 1991)
- Note: some noble gases might be remnants of the primary atmosphere (Dauphas, 2003; Genda & Abe 2005)

Fate of Venus and Earth

<u>H_2O mixing ratio in upper atmosphere</u>



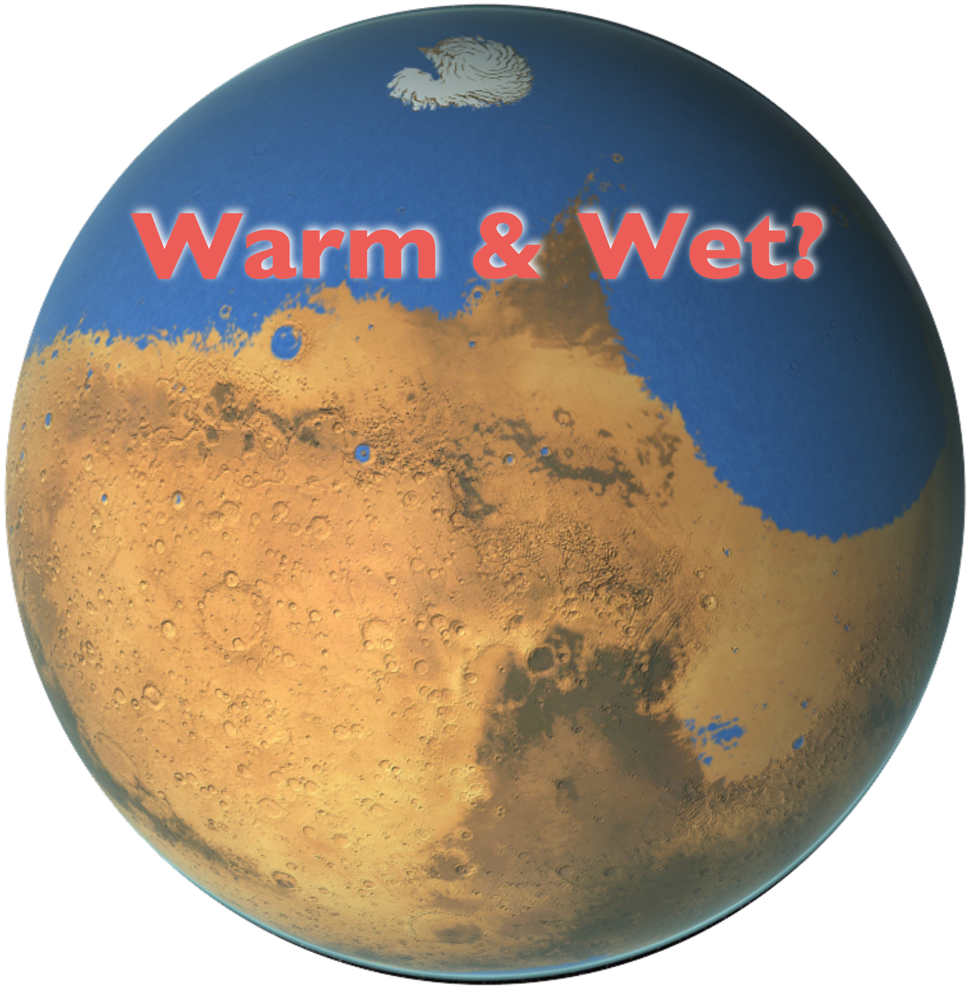
Kasting et al. (1993)

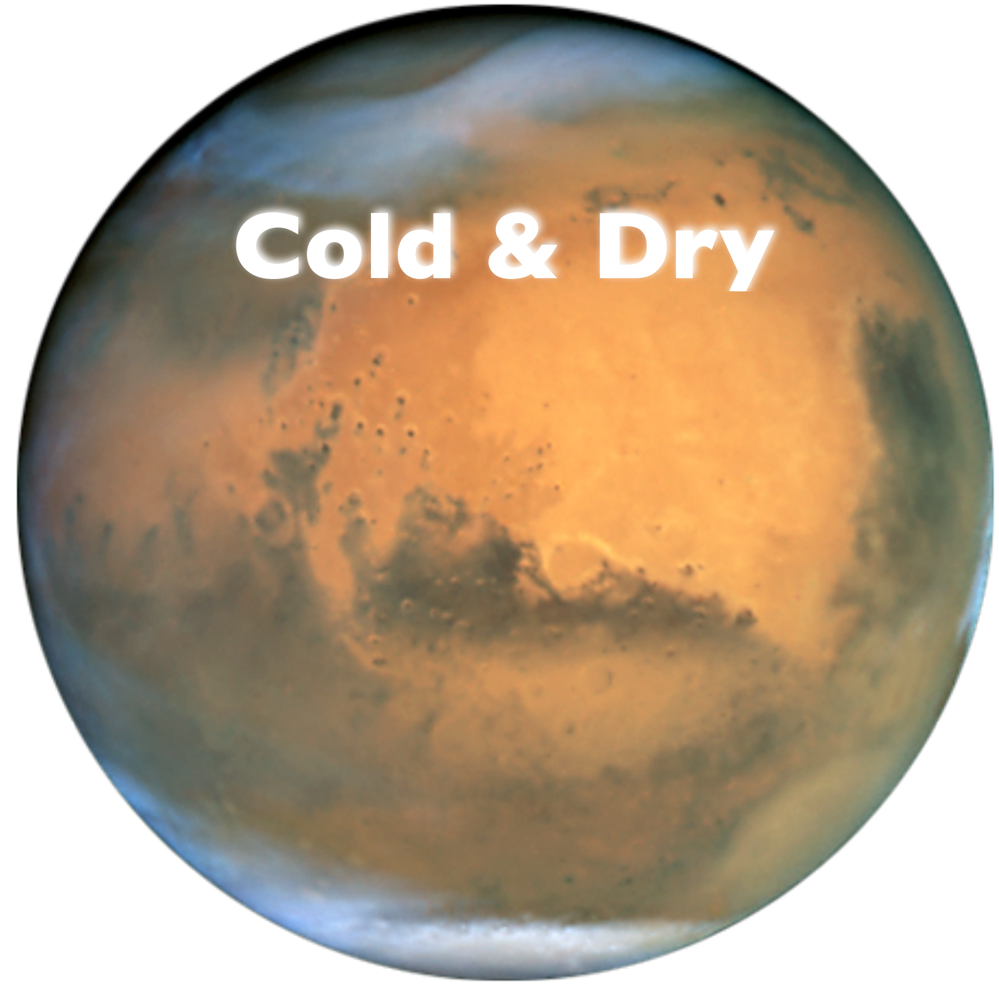


- Venus is located near the inner edge of the habitable zone
- While cold trap at the tropopause kept H_2O mixing ratio in Earth's upper atmosphere lower, the mechanism did not work for early Venus
- **The diffusion-limit on H escape sustained Earth's water, but water on Venus was lost almost completely**

Climate change on Mars

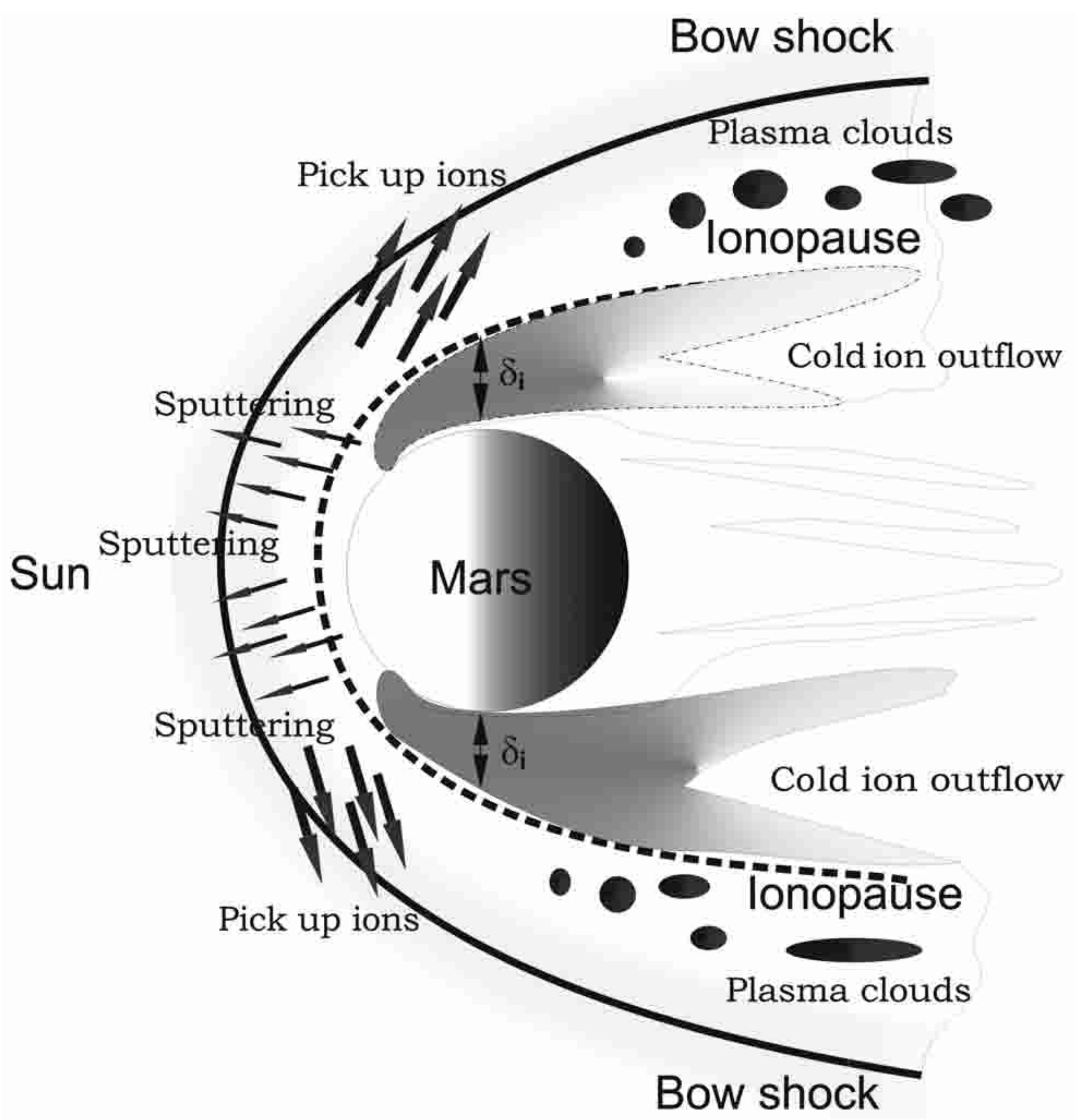
~4 billion years ago | **Present-day**





- Mars lost its surface water and atmosphere which sustained warm climate
- What caused the loss?
  - The loss of magnetic dynamo caused nonthermal escape (e.g., Jakosky et al. 2017) ?
  - The low gravity allowed significant thermal escape (e.g., Tian et al. 2009) ?

Nonthermal escape from non-magnetized planets



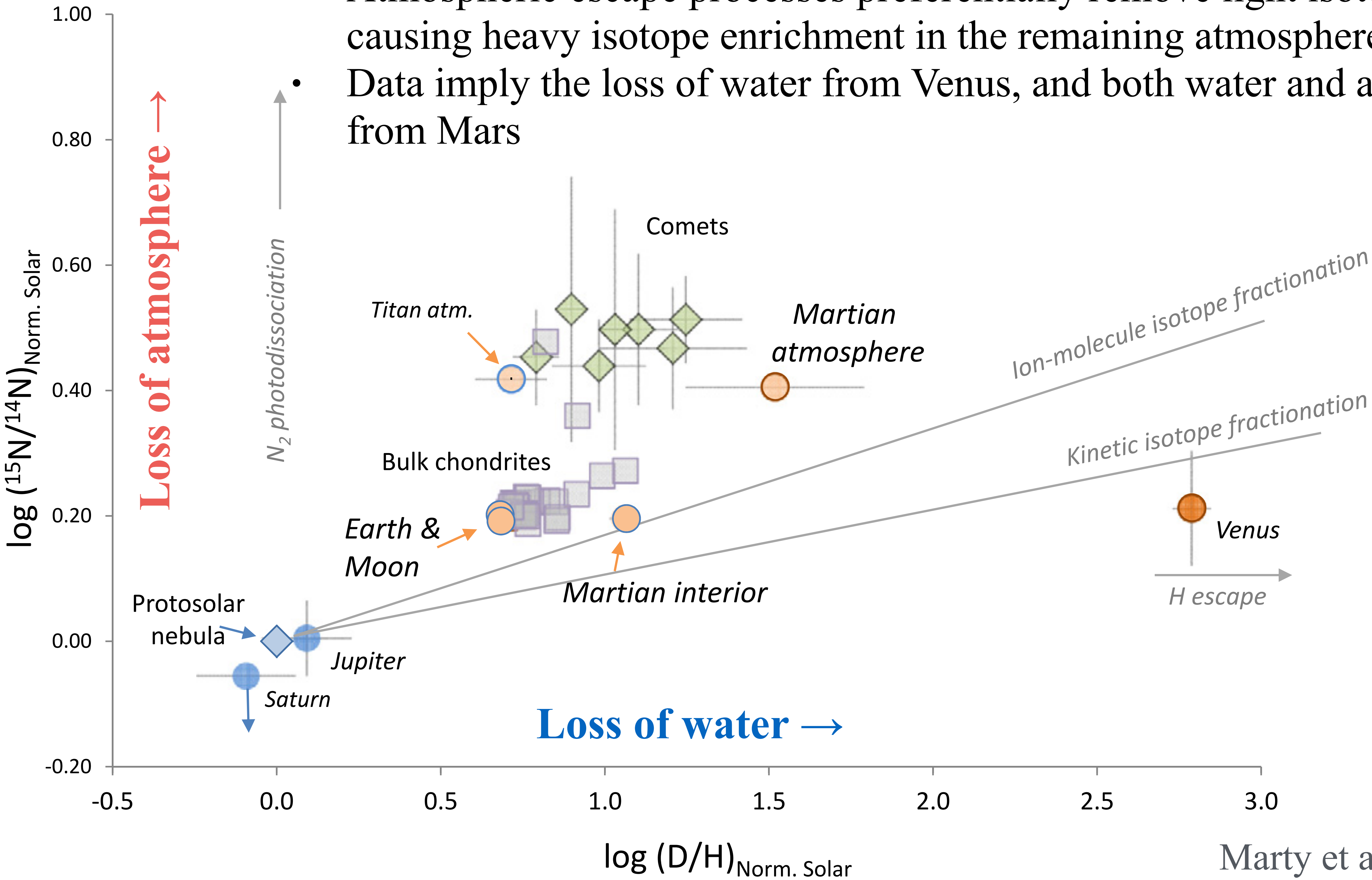
Terada et al. (2009)

In the case of non-magnetized planets, the solar wind directly interacts the upper atmosphere, and causes escape

- **Ion pickup:**
  Ions in the upper atmosphere can be removed by the interaction with the solar-wind magnetic field
- **Sputtering:**
  Picked up ions re-enters the atmosphere and impacts neutral molecules/atoms
- **Bulk removal (cold ion flow):**
  Ionized atmosphere interacts with the solar-wind plasma and is removed as bulk

Isotope geochemistry

- Atmospheric escape processes preferentially remove light isotopes, causing heavy isotope enrichment in the remaining atmosphere
- Data imply the loss of water from Venus, and both water and atmosphere from Mars



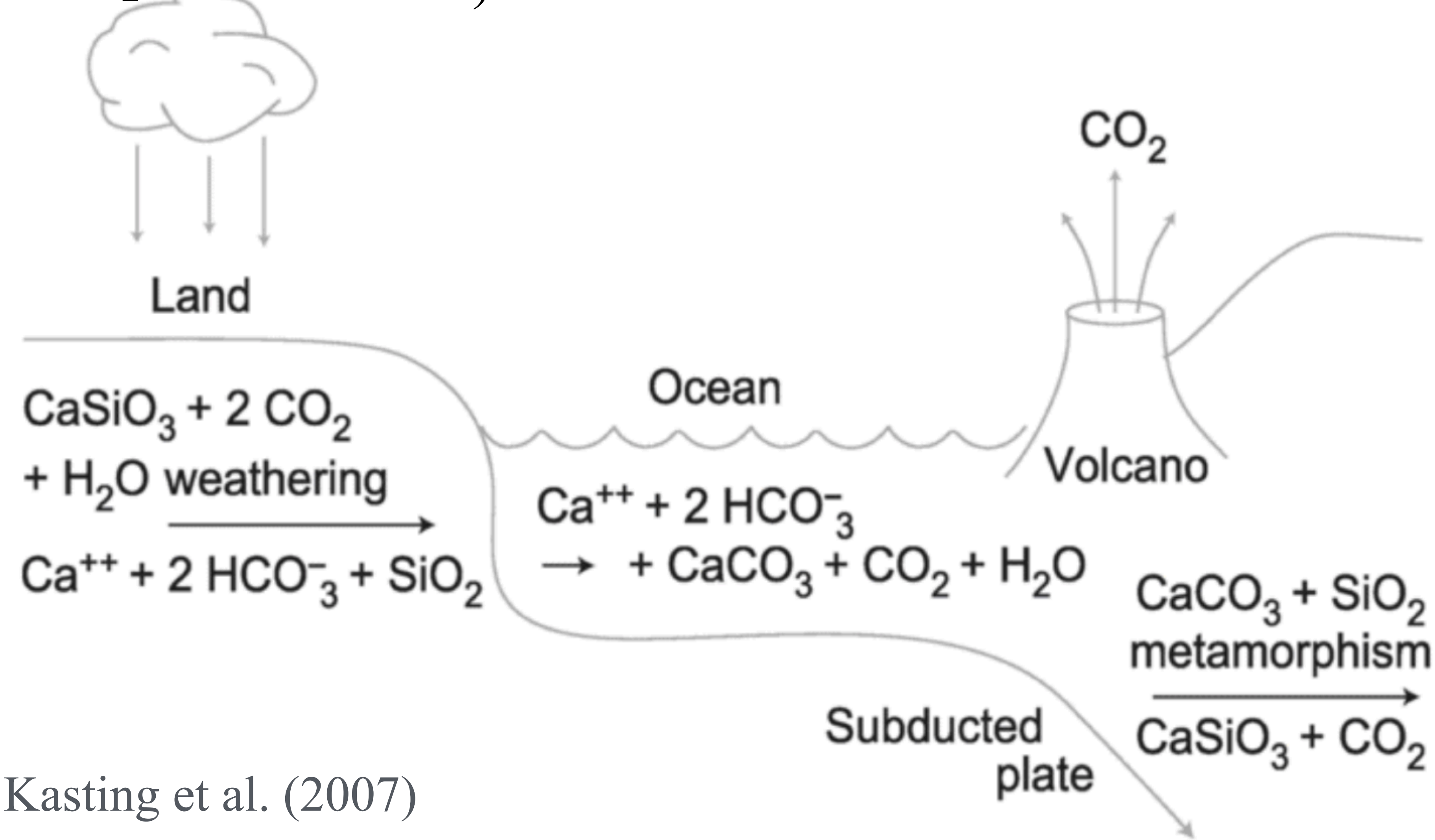
Marty et al. (2016)

Faint young Sun paradox

- ~4 billion years ago, the Sun's output would be only 70% as intense as it is during the modern epoch
- Assuming the same volume & composition of Earth's atmosphere as present-day, early Earth would have been too cold to sustain its oceans
- However, geologic evidence suggest that Earth has sustained oceans from that period → need more greenhouse gas!

Carbon-Silicate cycle

- Presence of oceans and plate tectonics have removed a large amount of atmospheric CO_2 on early Earth by carbonate ($CaCO_3$) formation, which explains the difference in Venus and Earth
- ***The carbon-silicate cycle*** has a negative feedback, it may compensate for the faint young Sun (low T → less weathering → less carbonate formation → high atmospheric CO_2 accumulation)

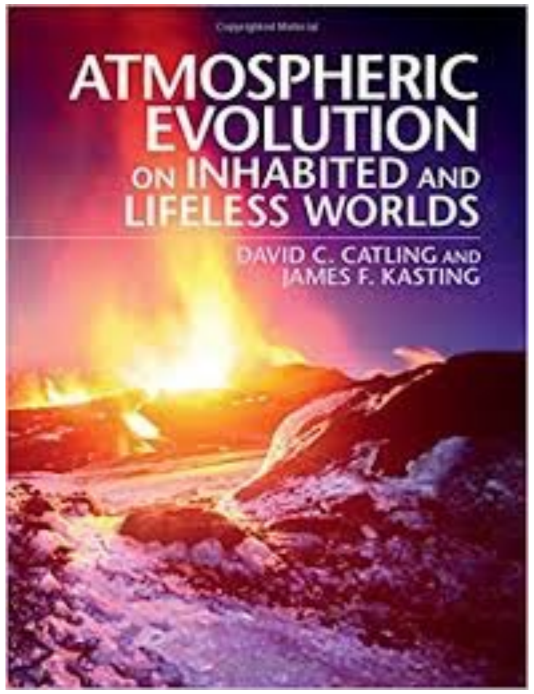


Catling & Kasting et al. (2007)

Summary

- Atmospheric escape is an important driver of long-term climate change
  - Thermal escape: Jeans' escape, hydrodynamic escape
  - Nonthermal escape: charge exchange, polar wind, ion pickup, sputtering, …
- Captured protoplanetary-disk gas (primary atmospheres) are likely to have been lost by hydrodynamic escape
- Present atmospheres of Venus, Earth, and Mars originated from volatiles delivered by their building blocks (secondary atmospheres)
- Because impact degassing is efficient, the atmospheres should have formed at the same time with the volatile delivery during the planet formation
- Cold trap and diffusion limit on hydrogen escape have kept Earth's water, whereas Venus lost its water because of the difference in their distances from the Sun
- Mars has lost its water and atmosphere because of its absence of global magnetic field and/or its small gravity
- Carbon-Silicate cycle has worked to stabilize Earth's climate

Reference



For further reading,
Catling, D. C. and Kasting. J. F.
Atmospheric evolution on inhabited and lifeless worlds.
Cambridge University Press, 2017.

The slides have been upload to
https://members.elsi.jp/~hiro.kurokawa/lecture_files/ELSI_Winter_School_2018_Atmospheres2_Kurokawa.pdf

You may be able to find by searching "ELSI Hiroyuki Kurokawa" on the internet

Exercises

1. Given the estimate of the diffusion-limited escape rate of hydrogen and the total H_2O mixing ratio at the homopause of present Earth (H_2: 0.5 ppmv, H_2O: 3 ppmv, CH_4: 1.8 ppmv), calculate the fraction (wt. %) of seawater that can be lost in 4 billion years

$$\Phi_{\mathrm{dif},i} \simeq 2.5 \times 10^{17} f_{\mathrm{T}}(\mathrm{H})\ \mathrm{m}^{-2}\mathrm{s}^{-1}$$

$f_{\mathrm{T}}(\mathrm{H})$: total hydrogen mixing ratio

$$f_{\mathrm{T}}(\mathrm{H}) = f_{\mathrm{H}} + 2 f_{\mathrm{H_2}} + 2 f_{\mathrm{H_2O}} + 4 f_{\mathrm{CH_4}} + \ldots$$

You may use these values: Mass of a hydrogen atom 1.7×10^{-27} kg,
Earth radius = 6.4×10^{6} m, Mass of seawater = 1.4×10^{21} kg

2. Methane (CH_4) has been proposed as a possible greenhouse-effect gas to compensate the faint young Sun paradox. Assuming f_{CH4} = 1000 ppmv, calculate the fraction of seawater that can be lost during the Archean (4 Ga to 2.5 Ga)

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A. 0.24 wt.%

2. Methane (CH_4) has been proposed as a possible greenhouse-effect gas to compensate the faint young Sun paradox. Assuming f_{CH4} = 1000 ppmv, calculate the fraction of seawater that can be lost during the Archean (4 Ga to 2.5 Ga)

A. 26 wt. %