VI Astrobion – September 2026

Core Formation and Magma Oceans

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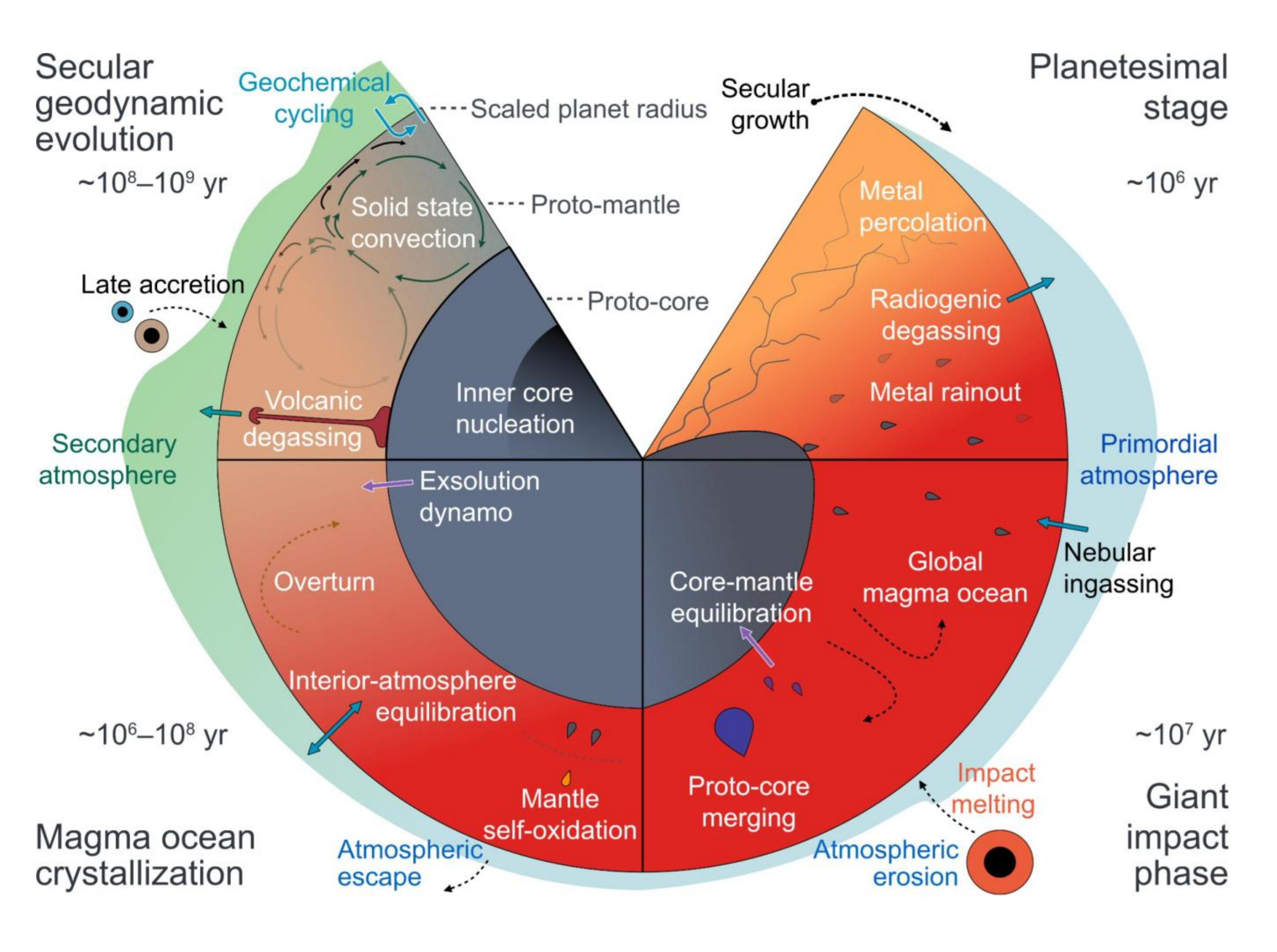
Atmosphere-Interior Evolution



Lichtenberg et al. (2022) Protostars and Planets VII

Partition Coefficients

$M(metal) + O_2(g) = MO_2(silicate)$

$$D_i = \frac{C_i^{metal}}{C_i^{silicate}}$$

xFe(metal) + MO_x(silicate) = M (metal) + xFeO(silicate)

$$K_D = \frac{a_{M,metal}\,a_{FeO,sil}^{x}}{\boxed{a_{MO_x,sil}}\,a_{Fe,metal}^{x}}$$

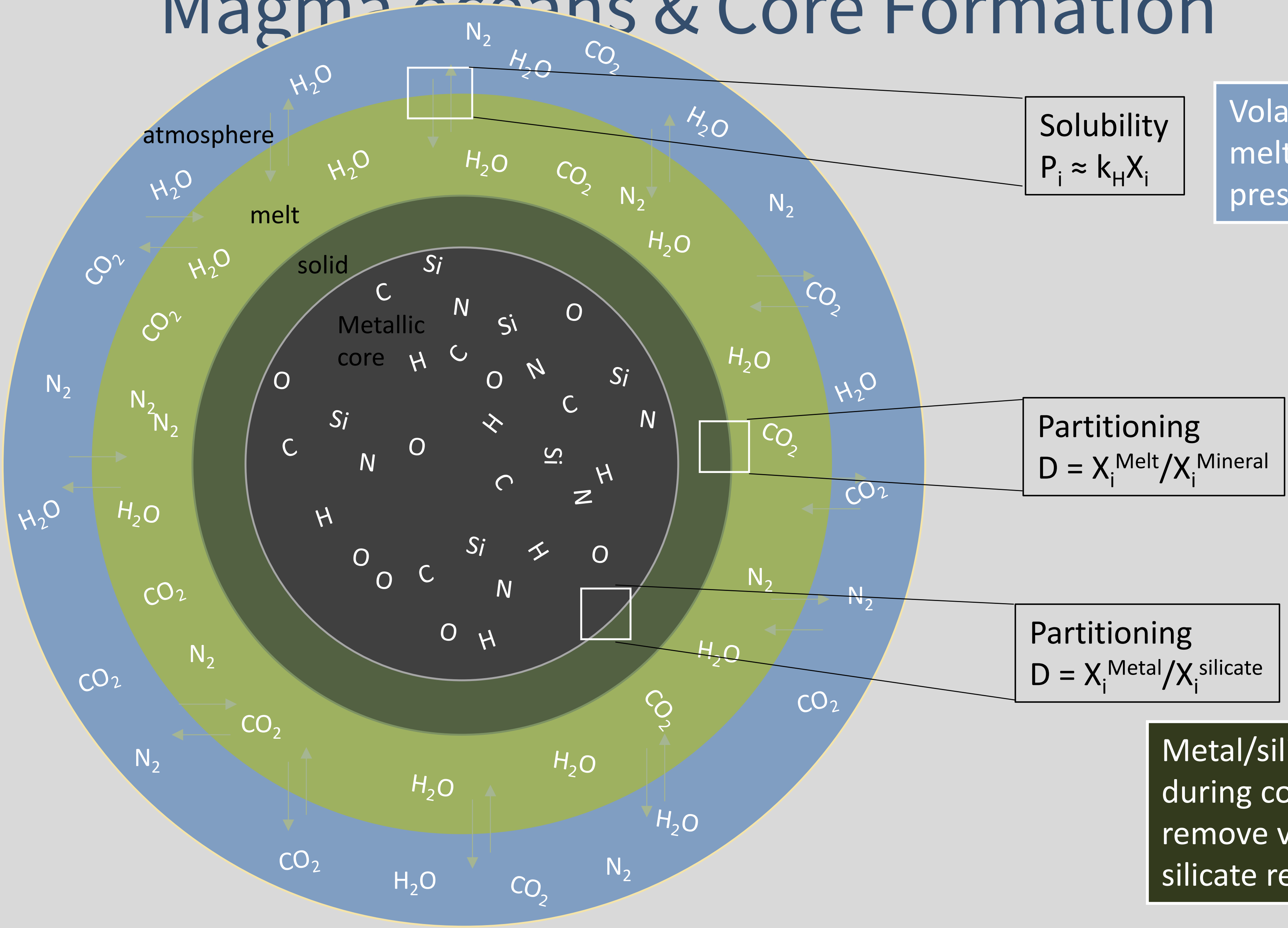
$$a_{MO_x,sil} = \gamma_{MO_x,sil} X_{MO_x,sil}$$

Experimentally determined

Dependent on temperature, pressure and fO_2

$$\ln D_{\mathrm{M}}^{\mathrm{met-sil}} = a \ln f\mathrm{O}_2 + b/T + cP/T + g$$

Magma oceans & Core Formation



Volatile solubilities in melt control atmospheric pressure at surface

Solid/melt partitioning sets up the initial volatile abundances in the mantle

Metal/silicate partitioning during core formation may remove volatiles from the silicate reservoir.

Core Formation

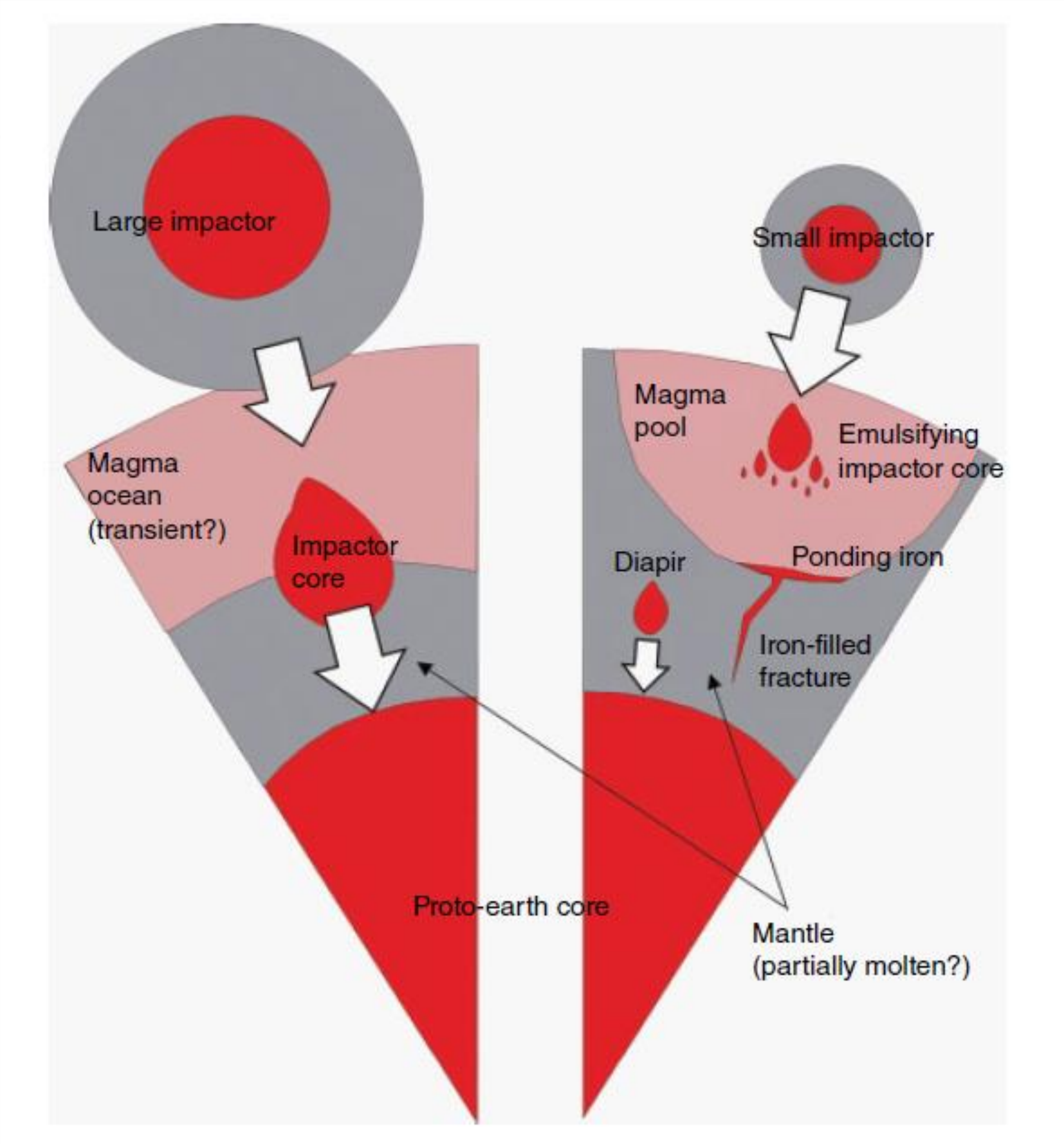
Core Formation

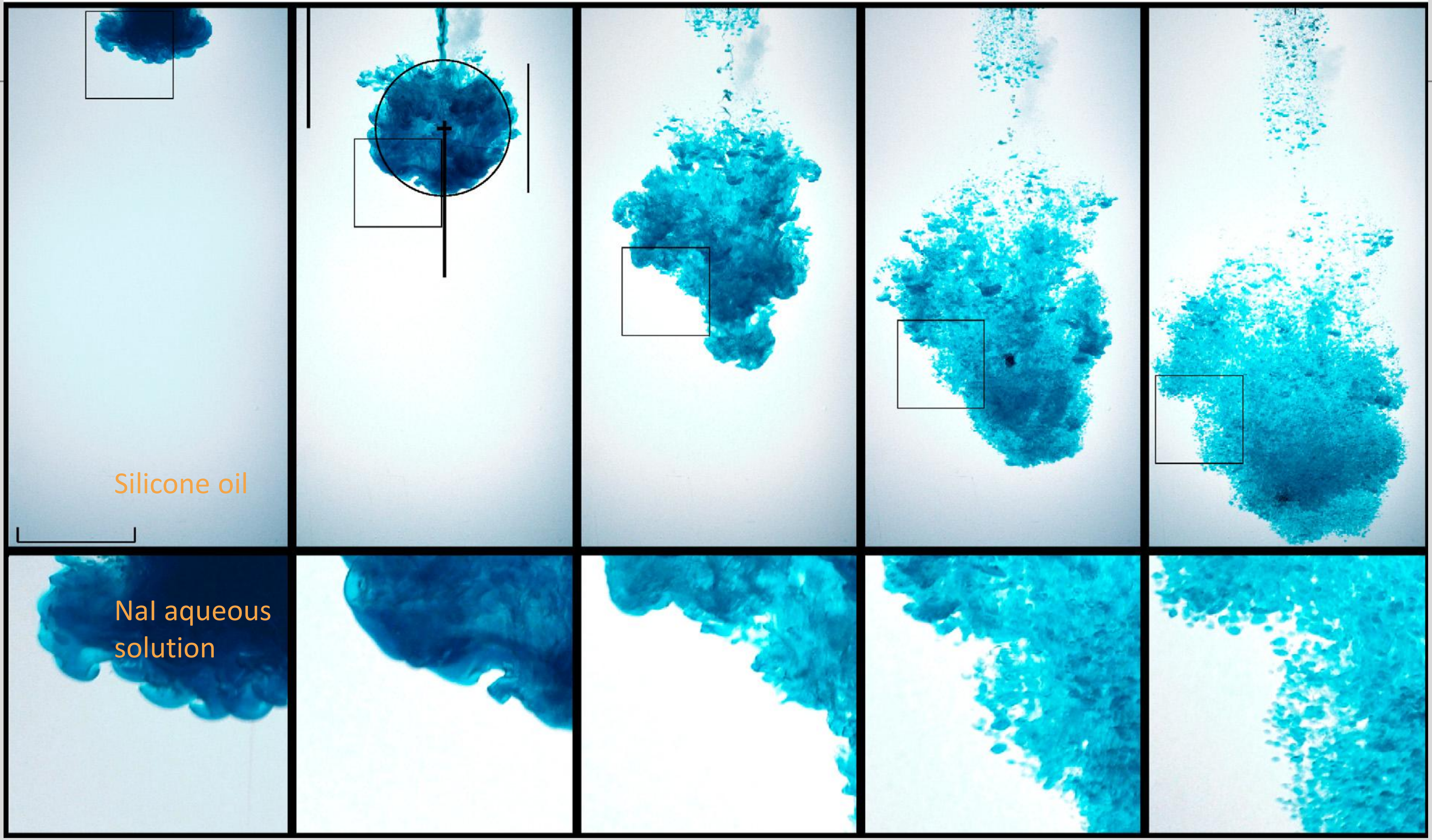
Impactors larger than planetesimals are likely to already be differentiated.

This has implications for how iron from the impactor will interact with the mantle of the target.

Smaller impactor cores may break up into small droplets (“emulsify”) and mix with the magma ocean long enough to react.

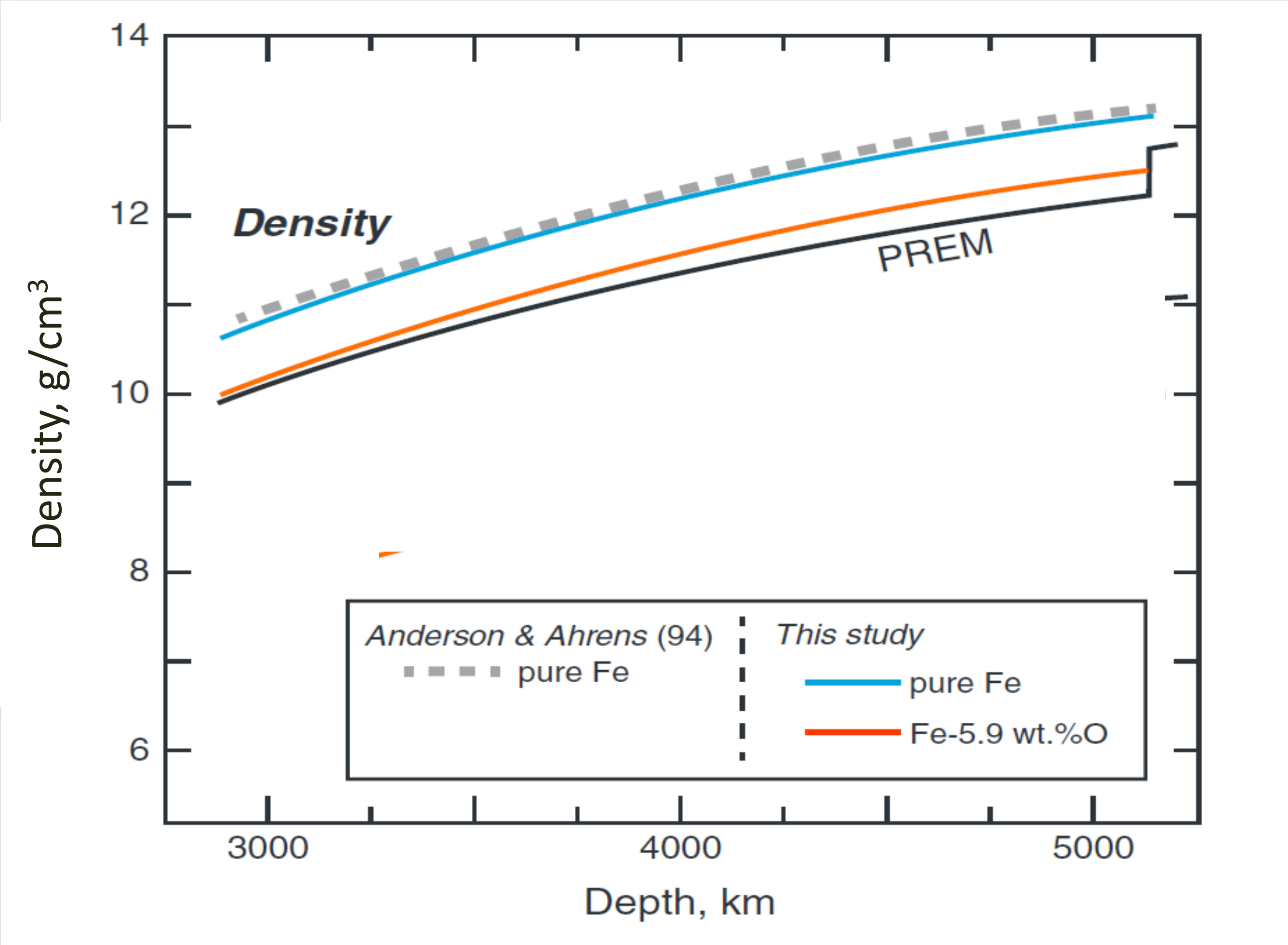
Larger impactor cores do not break up, and will sink directly to the target core without substantial reaction with the mantle.





Deguen et al. 2014

Earth's core composition

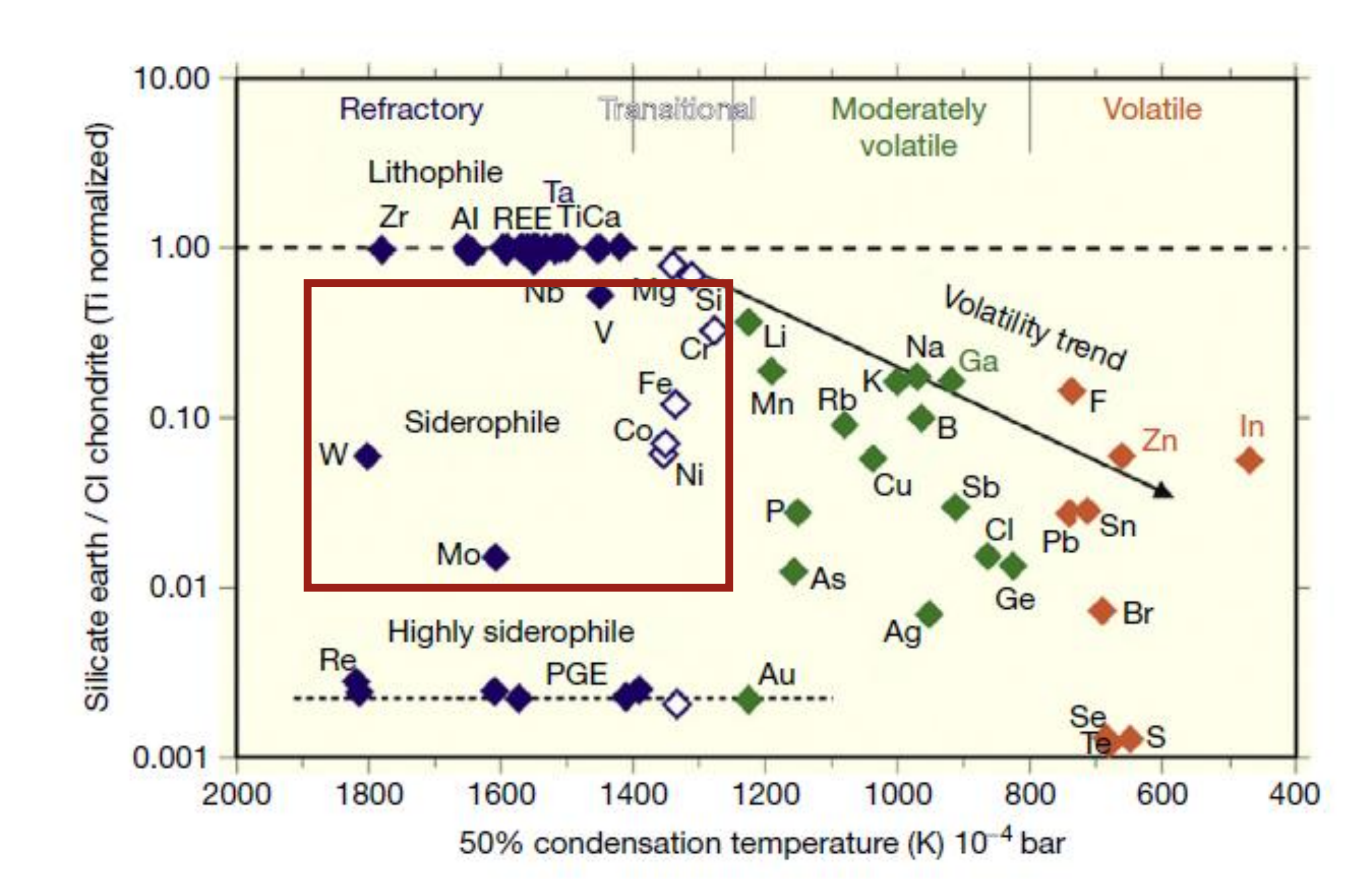


Komabayashi (2015)

Earth's core is less dense than pure Fe. The core must contain up to 10 wt% of a lighter element (H, C, O, S, Si) to account for the deficit.

Observations of iron meteorites indicate that the cores also contain significant amounts of Ni and Co.

Depletion of Siderophile elements

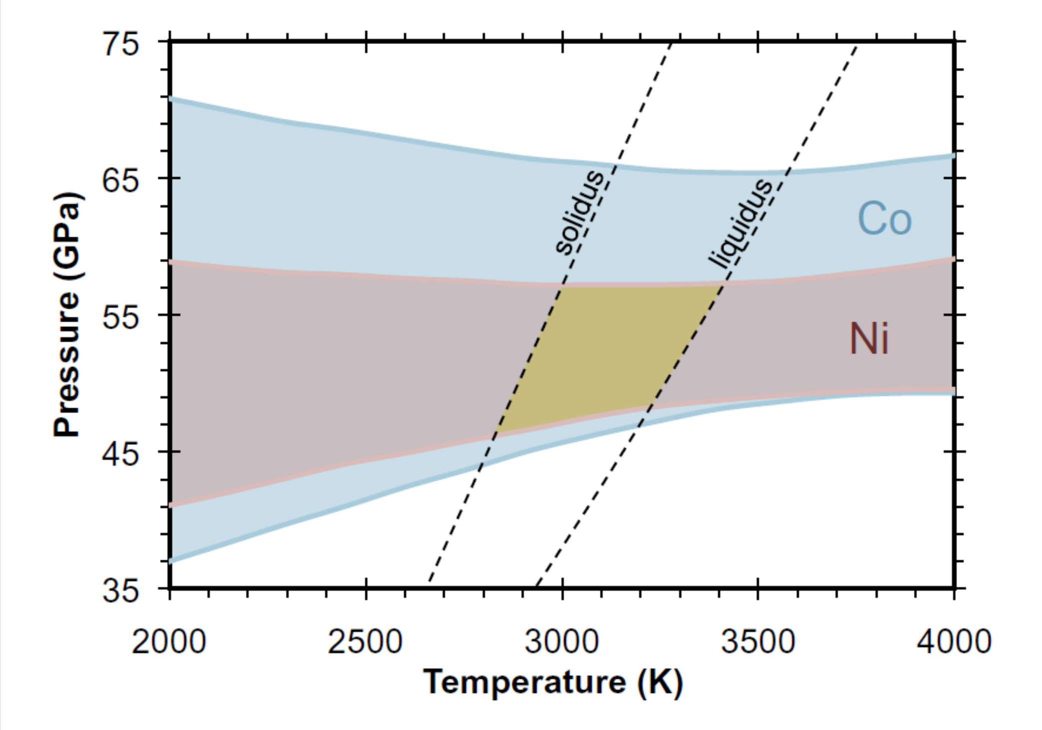


Refractory siderophile elements are depleted in the Earth's mantle.

These elements chemically exist in metallic or sulfidic phases.

Assuming an overall chondritic complement of these elements in the Bulk Earth allows an estimate of their concentration in the core.

Core formation conditions determined from trace elements



Fischer et al. (2015)

Ni and Co become more siderophile at higher pressure and temperature.

For single stage core formation model, a range of "average equilibration" conditions can be found that match the Ni & Co abundances in the mantle.*

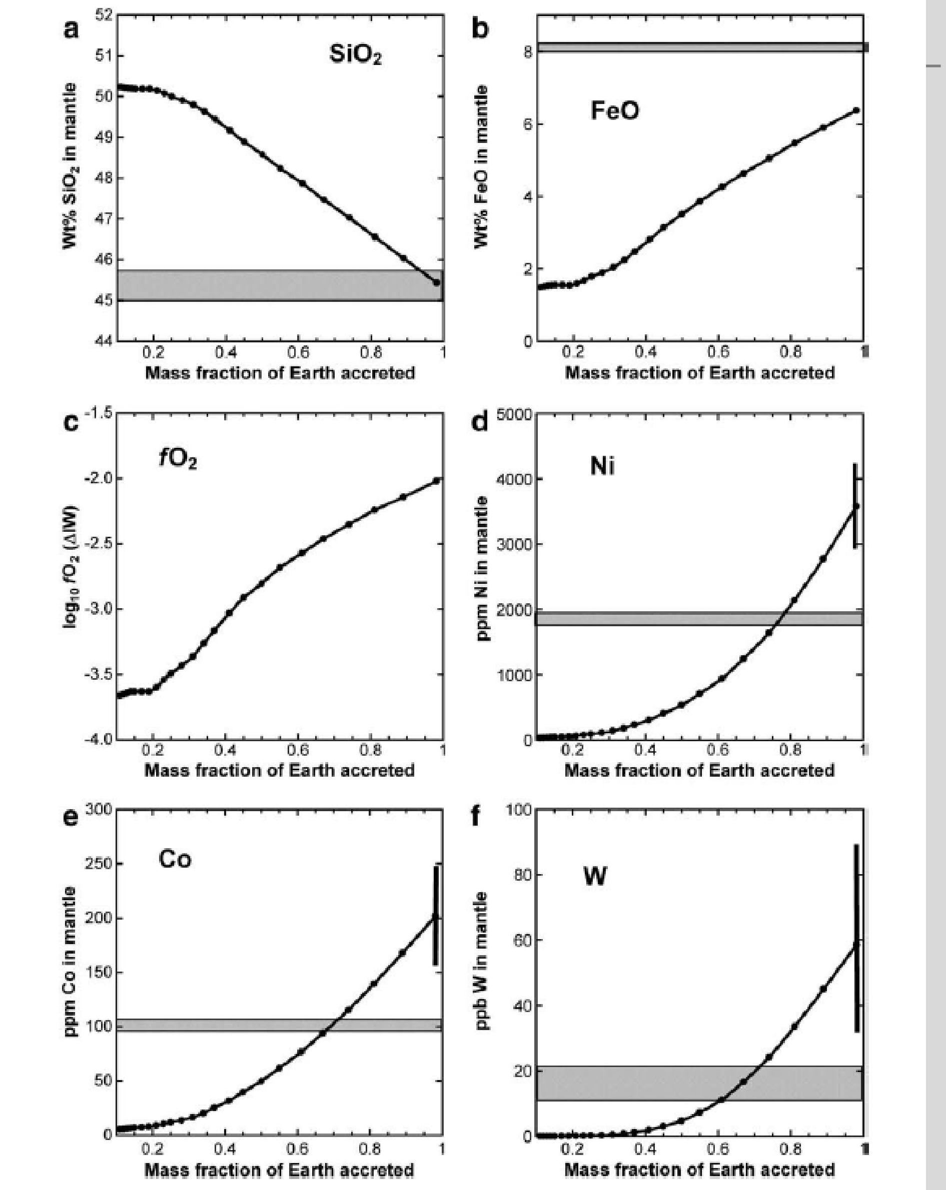
*assuming chondritic bulk abundances

Homogeneous accretion model

This model tracks the evolution of the Earth's composition as it grows from carbonaceous chondrite material.

This model assumes perfect equilibration of impactor cores with target mantle, and homogeneous composition of impactors.

This model cannot match most of the observed trace element abundances in the mantle.



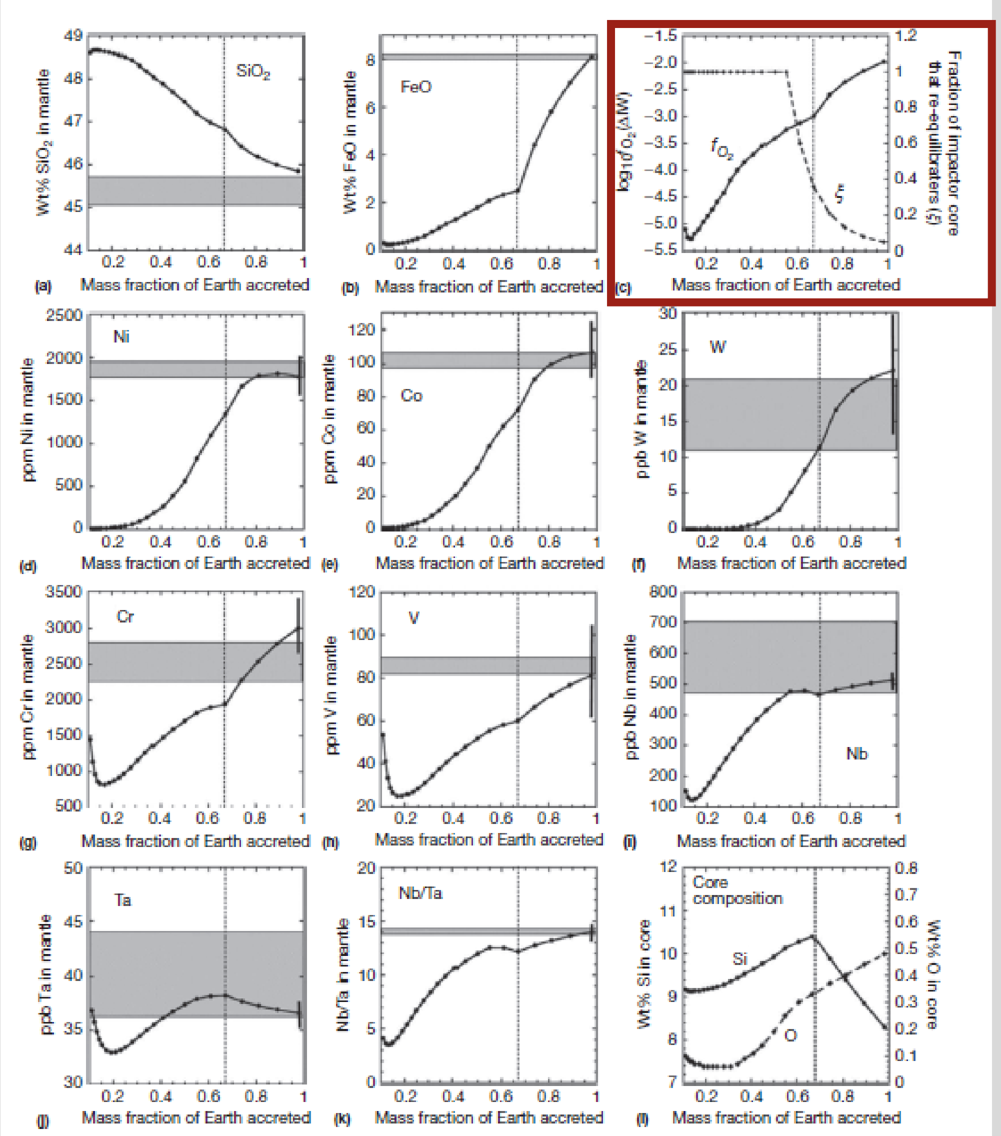
Rubie et al. (2011)

Heterogeneous accretion model

This model tracks the evolution of the Earth's composition as it grows.

This model assumes imperfect equilibration of impactor cores with target mantle. The composition of early impactors is very reduced (all Fe in metal), and later material is more oxidized (0.6 Fe in metal).

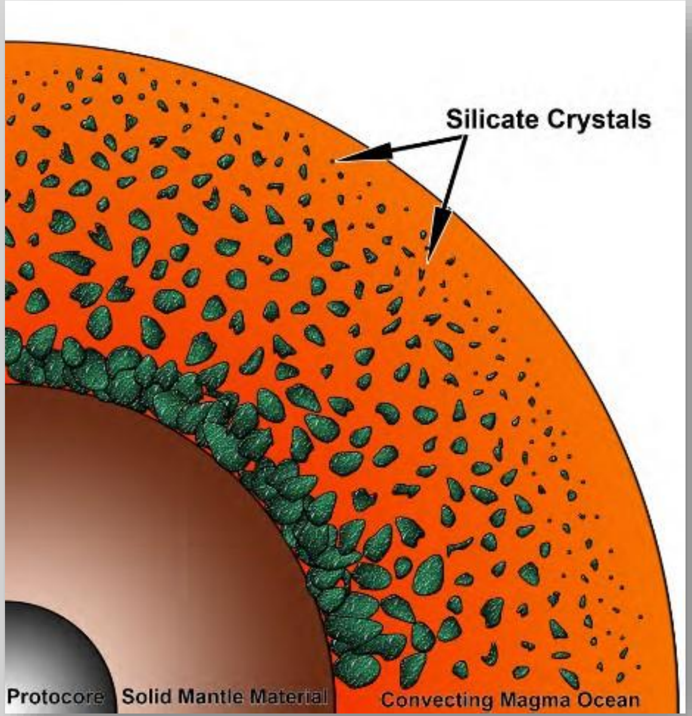
This model provides a much better match for many trace elements, plus the SiO_2 and FeO of the Earth's mantle.



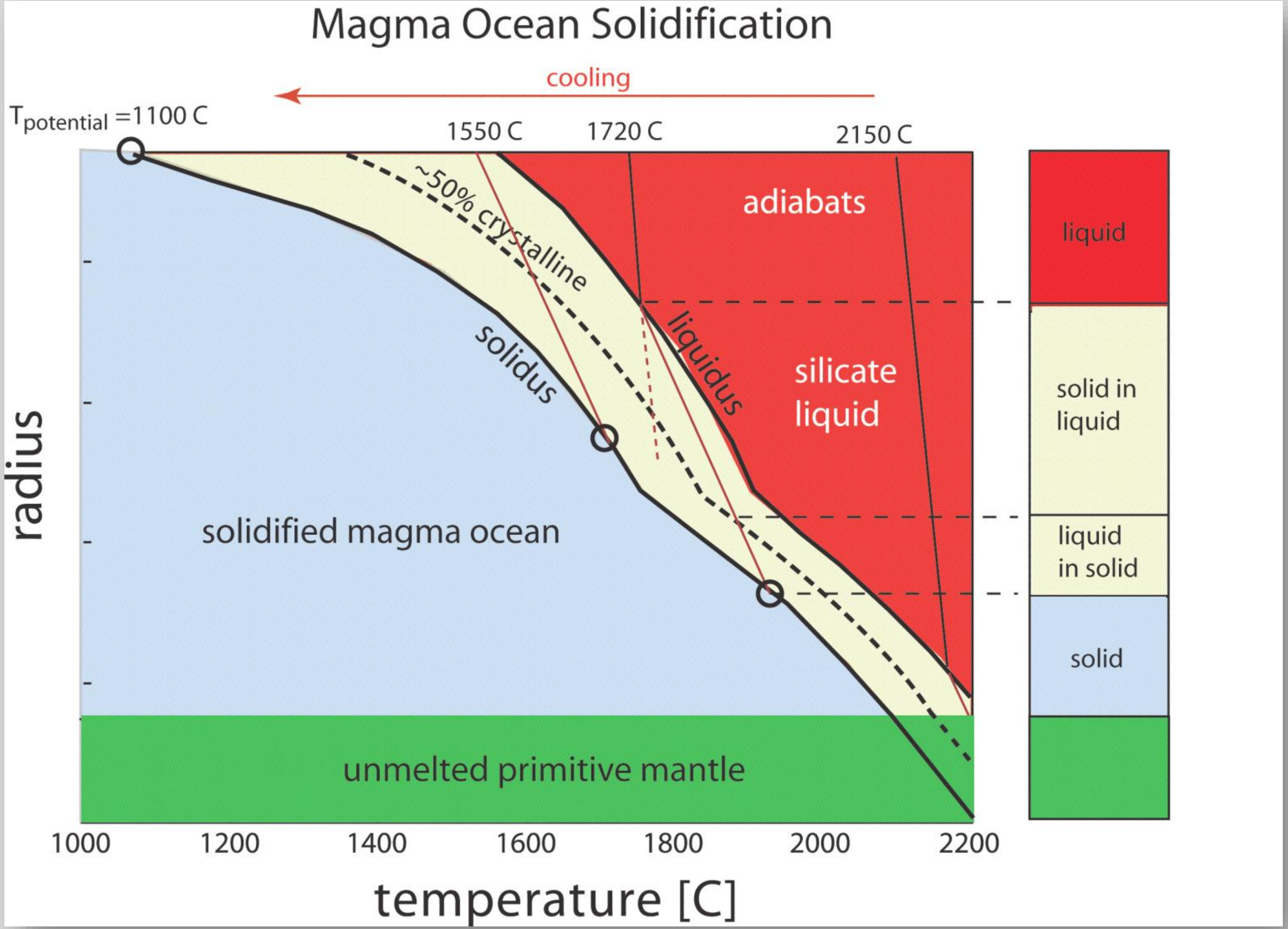
Rubie et al. (2011)

Magma Oceans

Magma Oceans

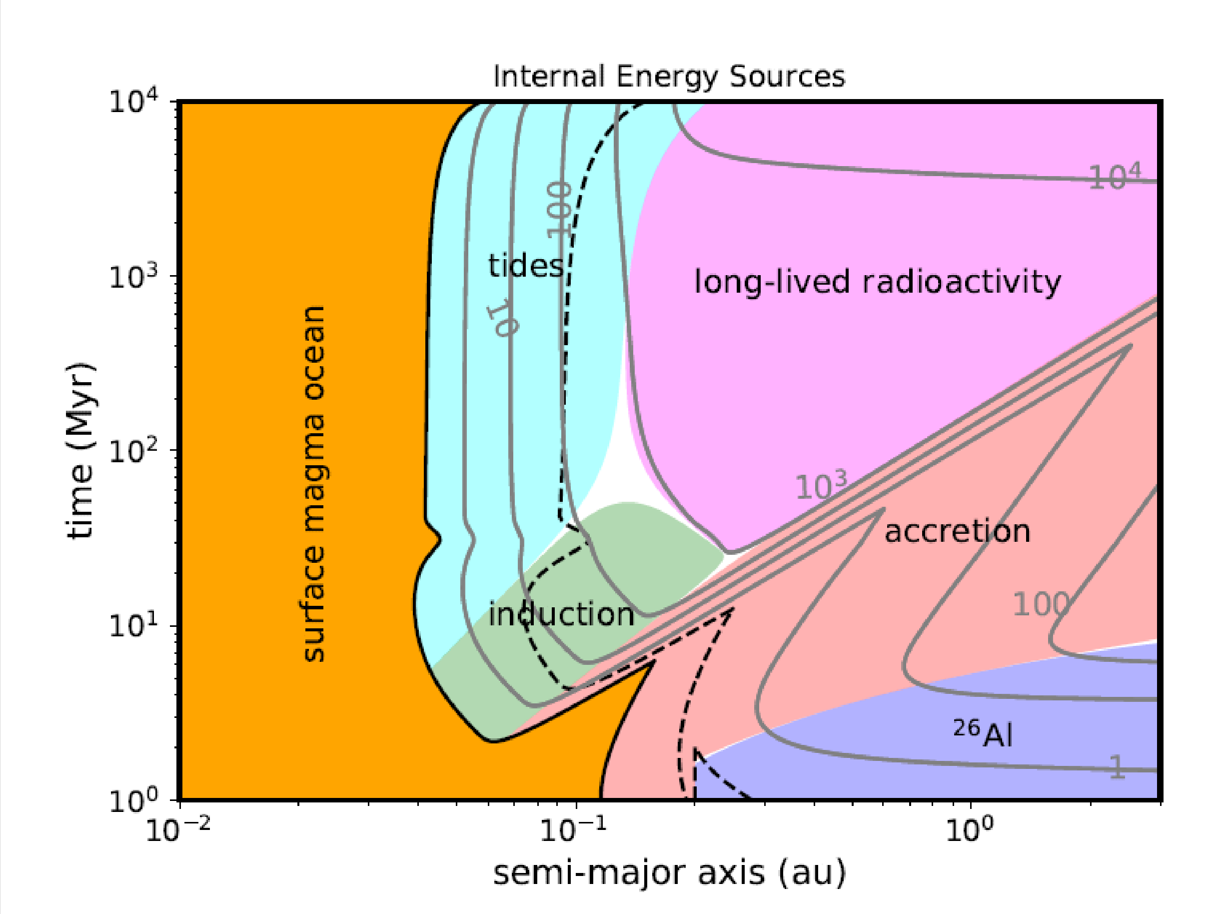


Mass & Hansen (2015)



Elkins-Tanton et al. (2003)

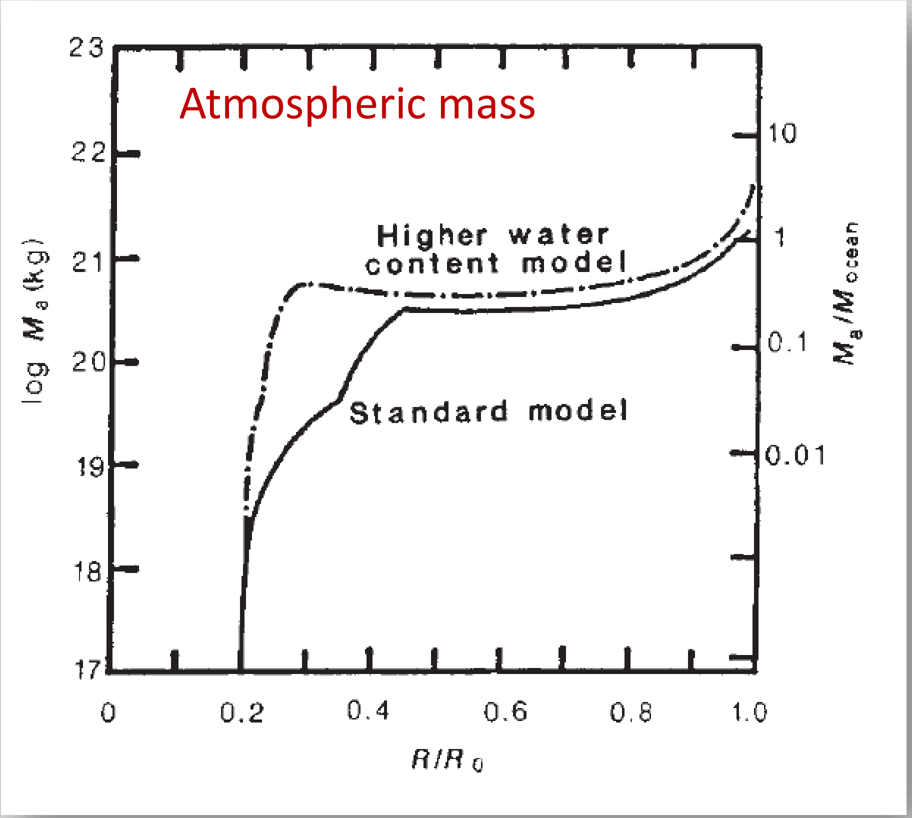
Where do we find magma oceans?



Chao et al. (2021)

- Young planets
- Extremely hot, bare planets
- Hot runaway-greenhouse planets

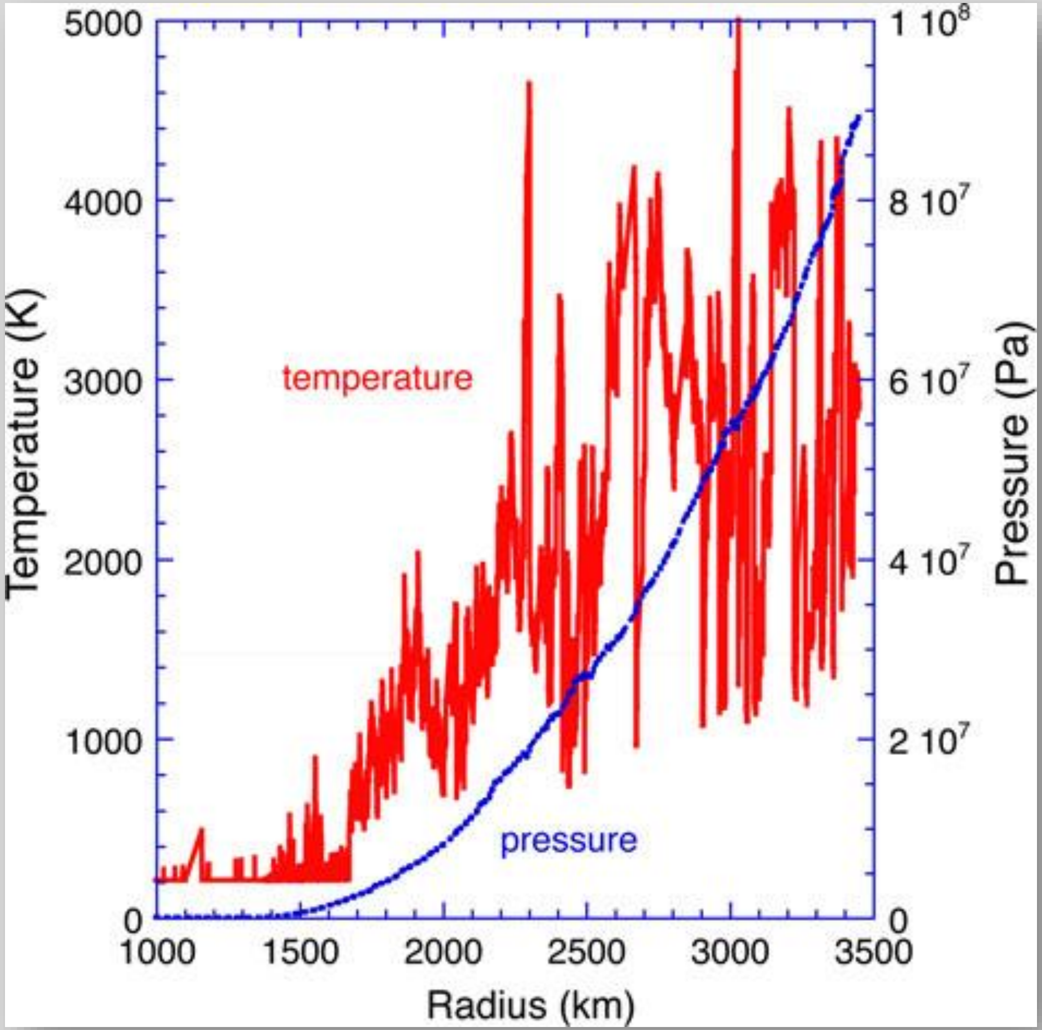
Heat source: Accretionary Impacts



Matsui & Abe (1986) Nature

The addition of an atmosphere to the growing planet limits the cooling rate and allows it to heat up more.

Atmospheric Blanketing Effect



Abe (2011)

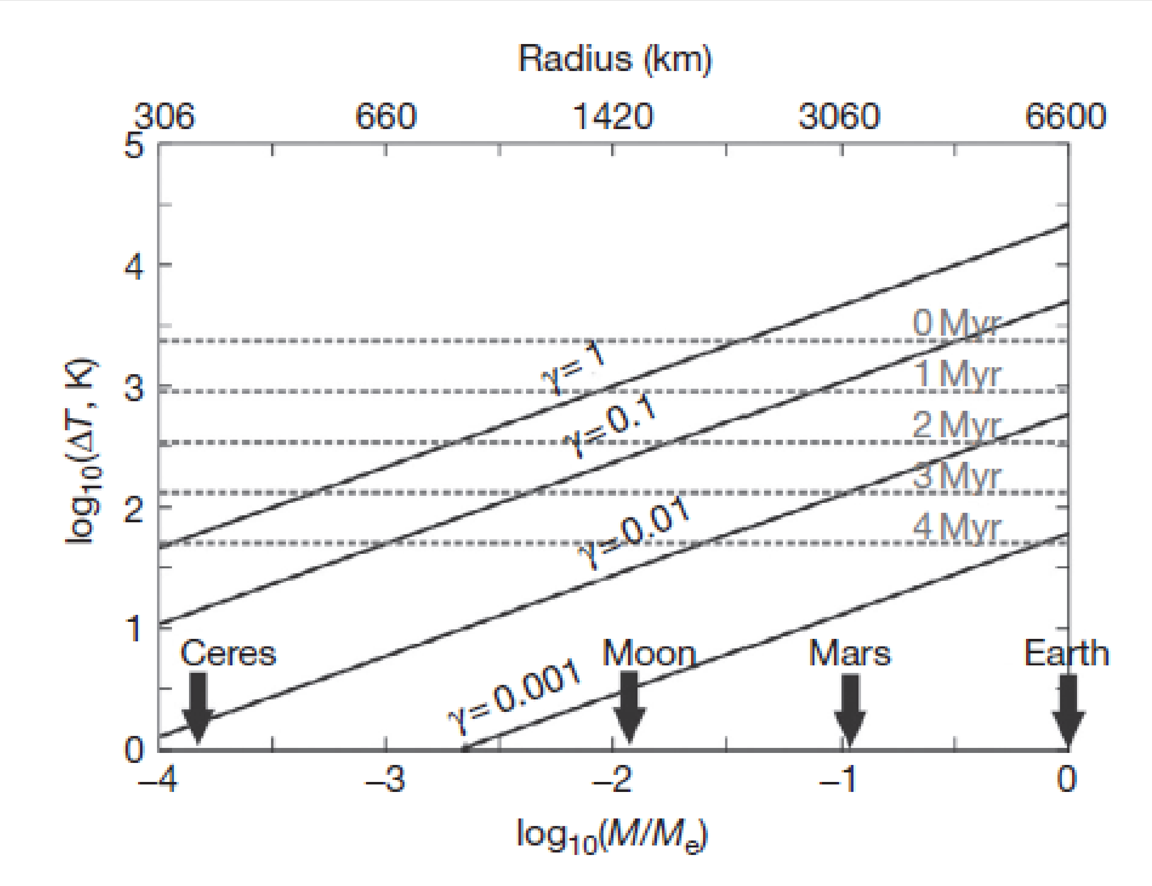
Giant impact

$$\Delta T \approx 6000K \left(\frac{\gamma}{0.1}\right)\left(\frac{M}{M_e}\right)^{2/3}$$

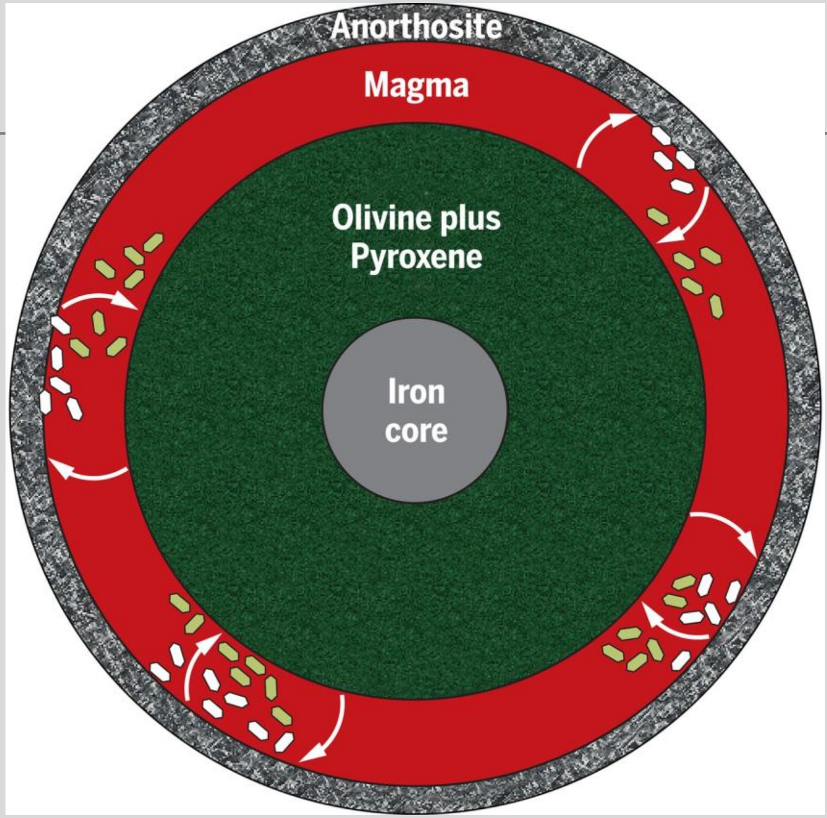
Globally averaged temperature change for a giant impact, with impactor mass ratio γ

Figure compares heating from ^{26}Al (dashed lines) to giant impact heating.

Widespread melting happens for ΔT > 1000 K.



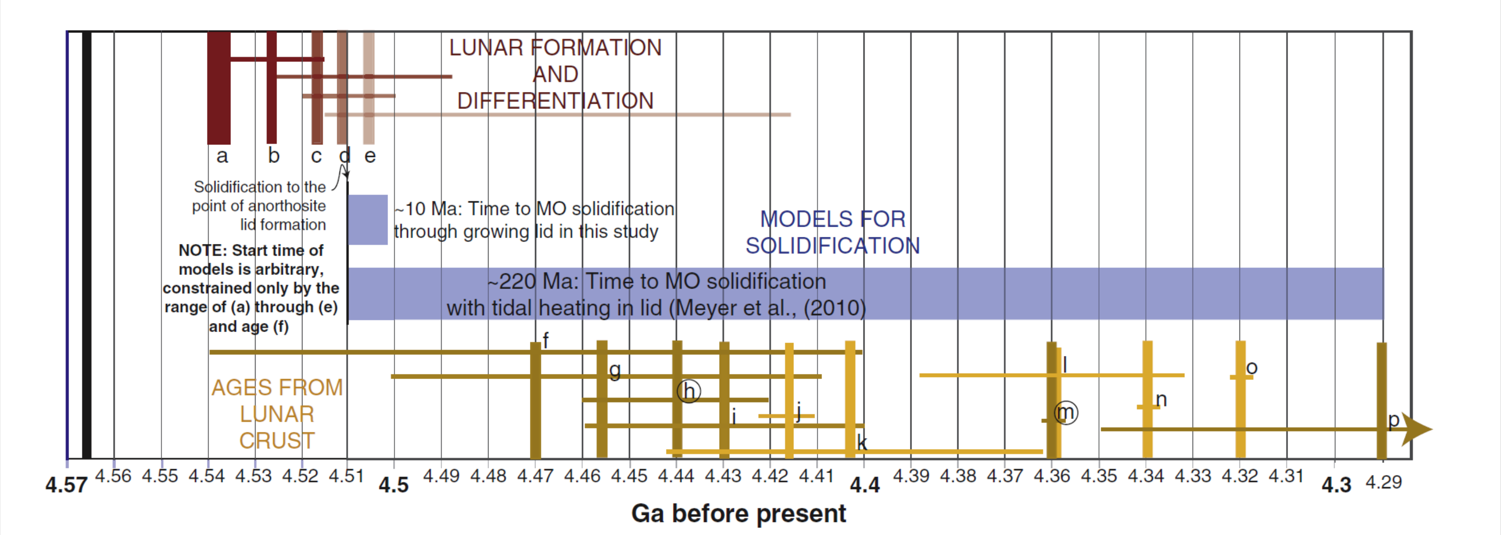
Rubie et al. (2014) Treatise on Geochemistry



Strongest geochemical evidence of a magma ocean in the Solar System is for the Moon.

The dominant crustal material is Anorthosite, which is thought to crystallize within a fractionally solidifying magma ocean and float to the surface due to its low density compared with the silicate melt.

KREEP: a component of some lunar basalts that formed through extensive fractionation = last dregs of the lunar MO

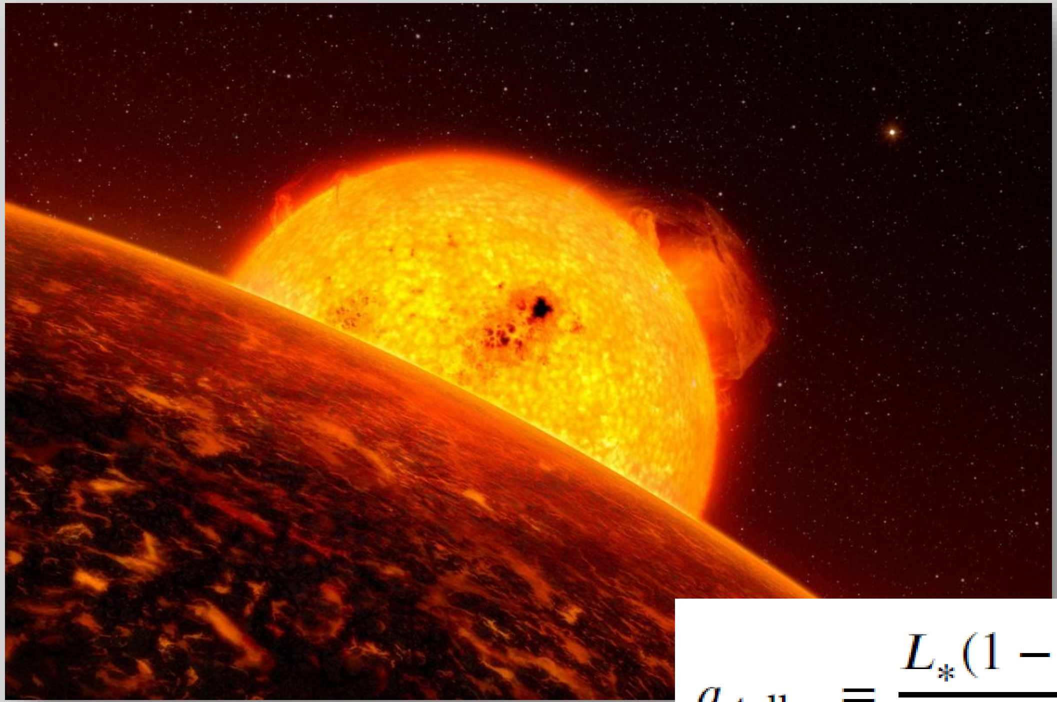


Elkins-Tanton et al. (2011)

Ages of lunar crustal minerals are consistent with very early formation shortly after planet formation.

Controversy over timing of lunar magma ocean solidification = too short!

Ultra-short period planets

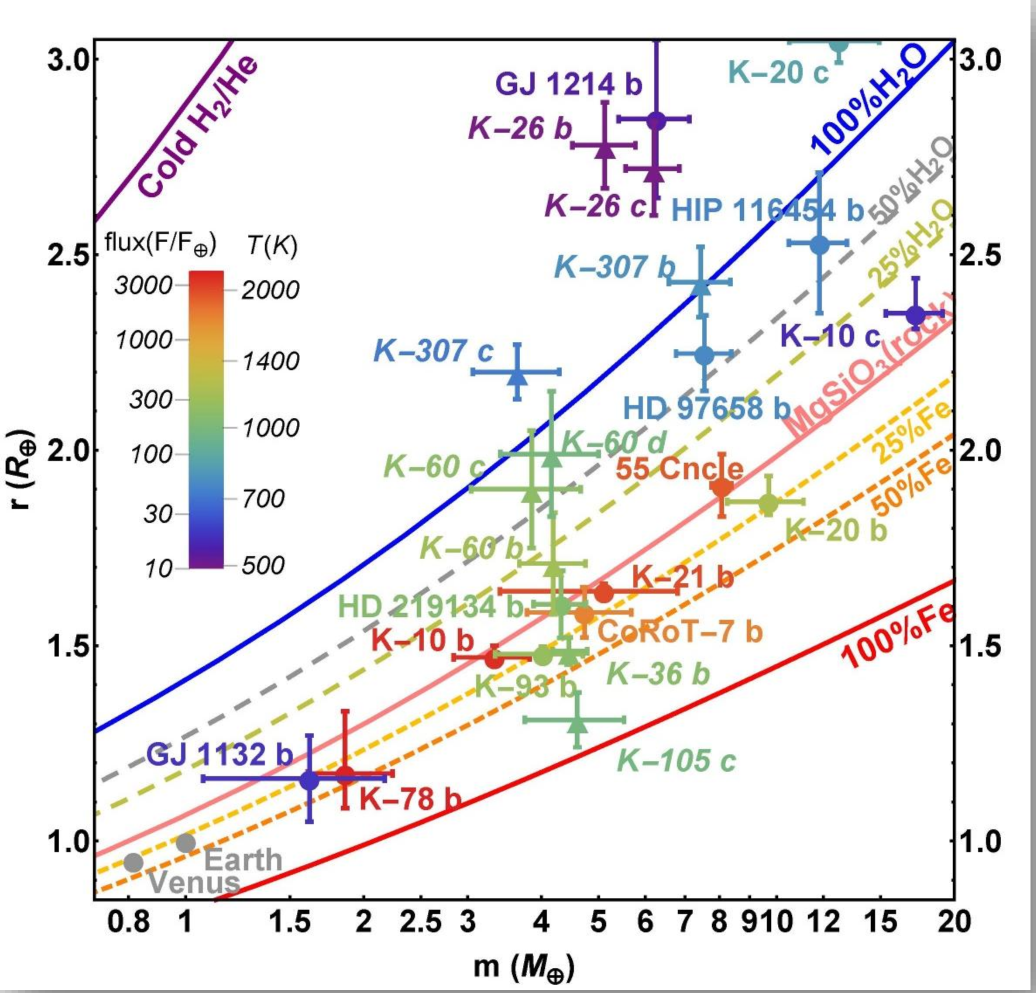


$$q_{\mathrm{stellar}} = \frac{L_*(1 - A)}{16\pi a^2}$$

Artist's impression of CoRoT-7b

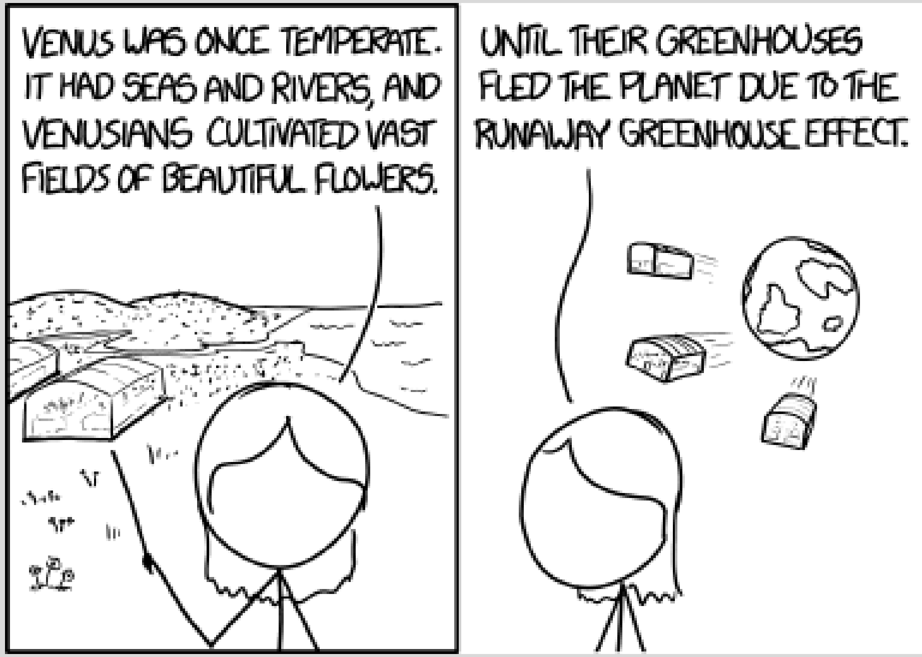
Stellar radiation alone is enough to melt silicate surfaces for some short-period exoplanets.

Exoplanet Mass-Radius Diagram

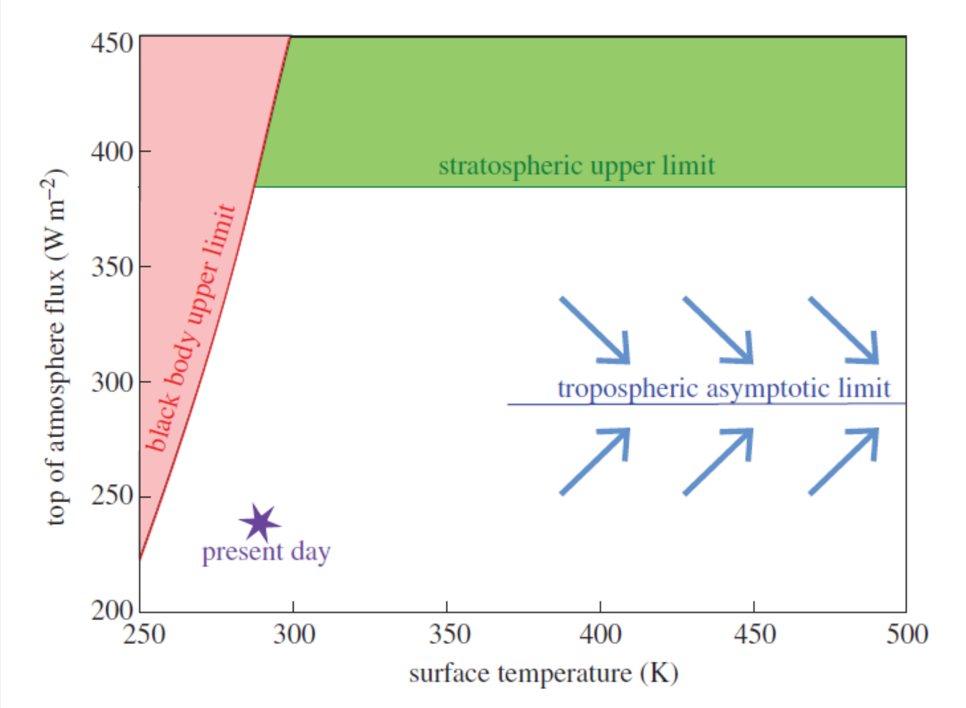


Lopez-Morales et al. (2016)

Runaway Greenhouse Limit



The **runaway greenhouse limit** describes how increasing the concentration of water vapor in a planetary atmosphere imposes a limit on the outgoing longwave radiation that the planet can emit.

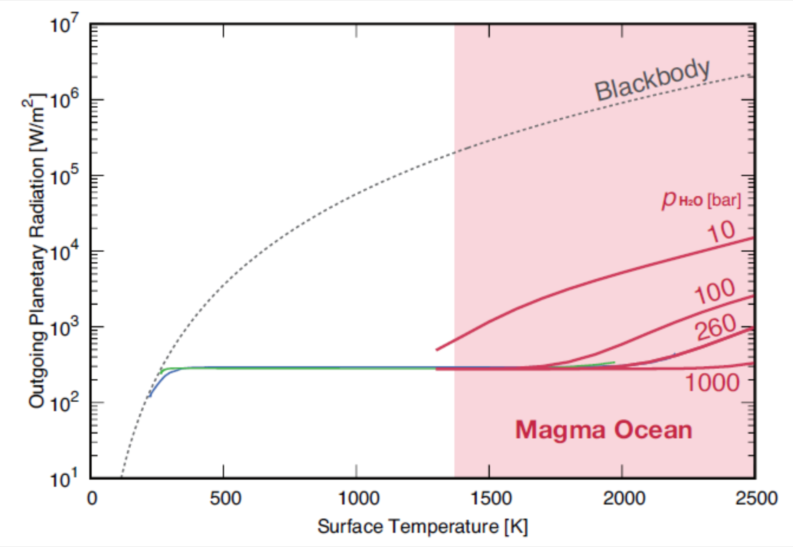


Increasing temperature will increase water vapor concentration in the atmosphere, which will cause a further increase in surface temperature.

Goldblatt & Watson (2012) Phil Trans Roy Soc A

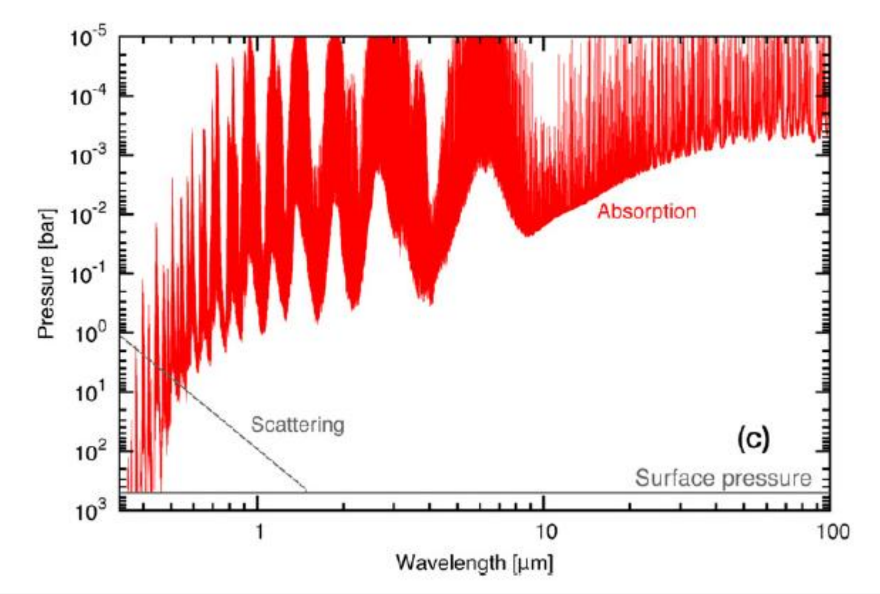
Runaway Greenhouse Limit

Ikoma et al. (2018)



If the total energy deposited into the bottom of a planet's atmosphere (e.g. sum of geothermal + solar radiation) exceeds the runaway greenhouse limit, the surface temperature will increase until the outgoing radiation (OLR) can balance it.

OLR will only begin to increase once the surface temperature is so high that significant radiation can escape from the windows between H_2O absorption bands in the IR. This can lead to surface temperatures in excess of 1500 K.



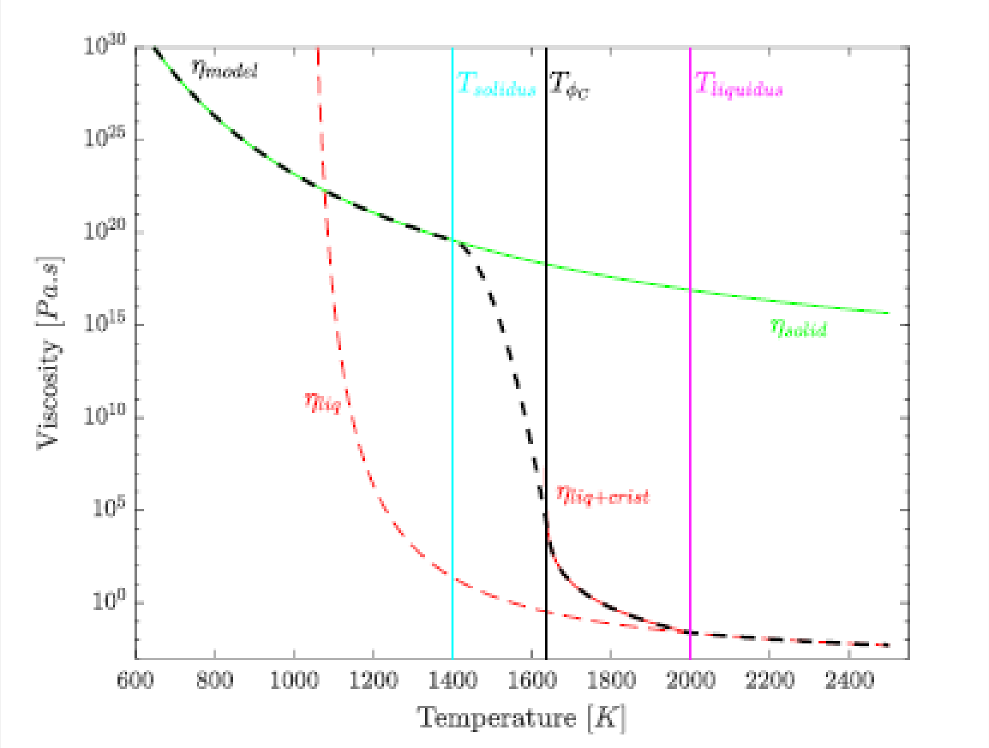
Hamano et al. (2015)

Viscosity

Viscosities of silicate melts are extremely low (~0.1 Pa s), approaching that of water.

However, viscosity will depend strongly on the crystal fraction.

Above a critical crystal fraction (~60%), viscosity jumps to near solid-like values.

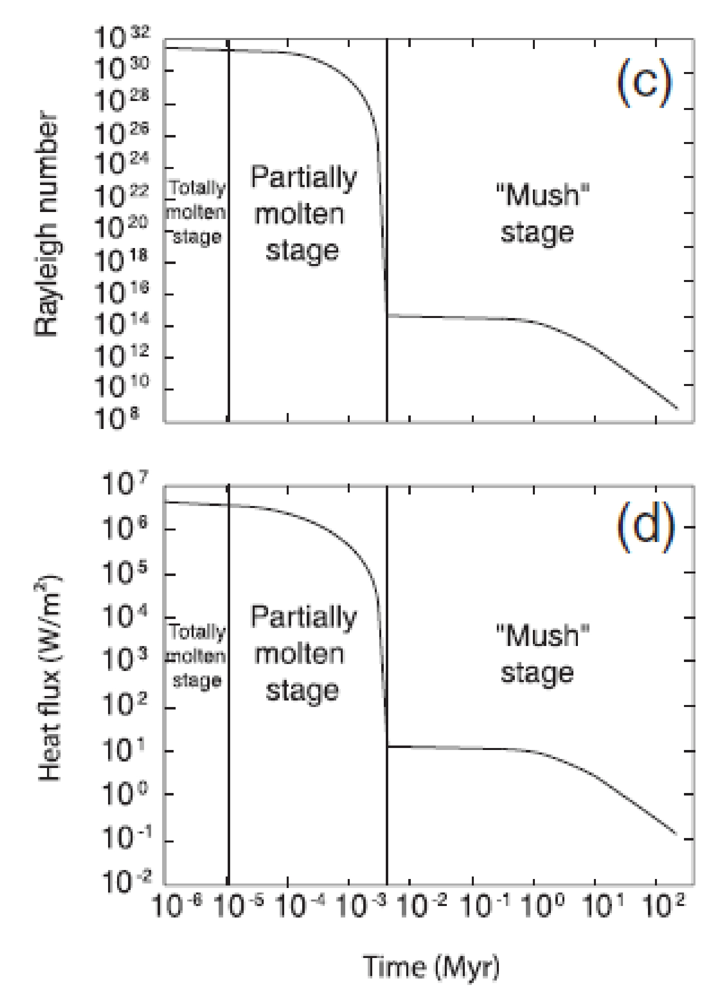


Salvador et al. (2017)

viscosity

$$\eta = \frac{\eta_l}{\left(1 - \frac{1-\phi}{1-\phi_c}\right)^{2.5}}$$

Magma Ocean Models



LeBrun et al. (2013)

Rayleigh number

$$Ra = \frac{\alpha g (T_m - T_s) l^3}{\nu \kappa}$$

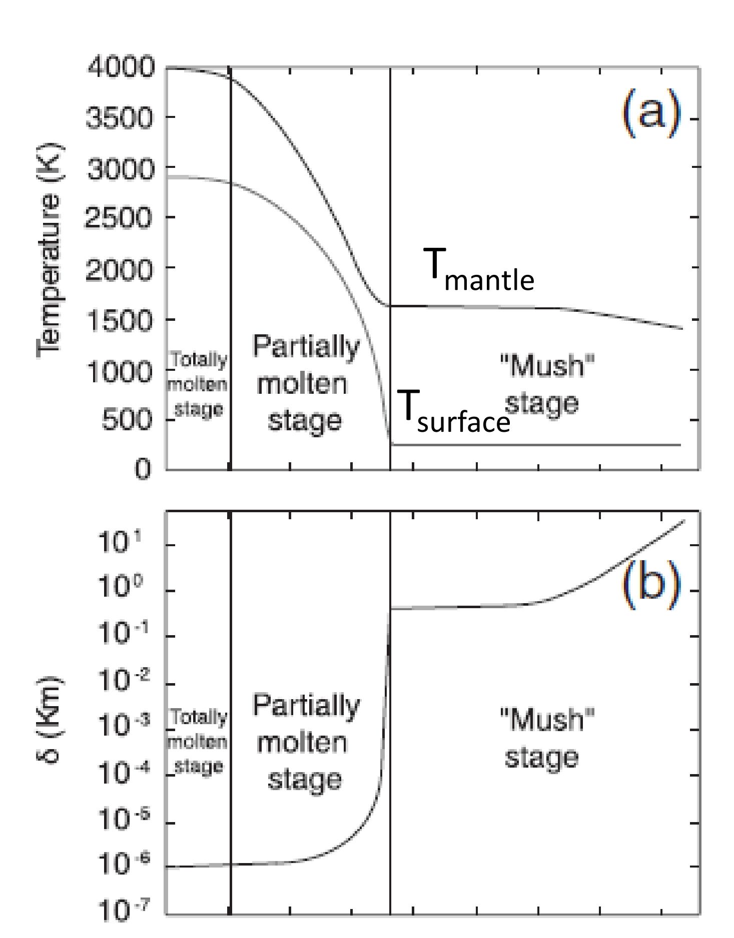
viscosity

$$\eta = \frac{\eta_l}{\left(1 - \frac{1-\phi}{1-\phi_c}\right)^{2.5}}$$

Heat flux

$$q_m = \frac{k(T_m - T_s)}{l} \left(\frac{Ra}{Ra_{cr}}\right)^{\beta}$$

Magma Ocean Models



LeBrun et al. (2013)

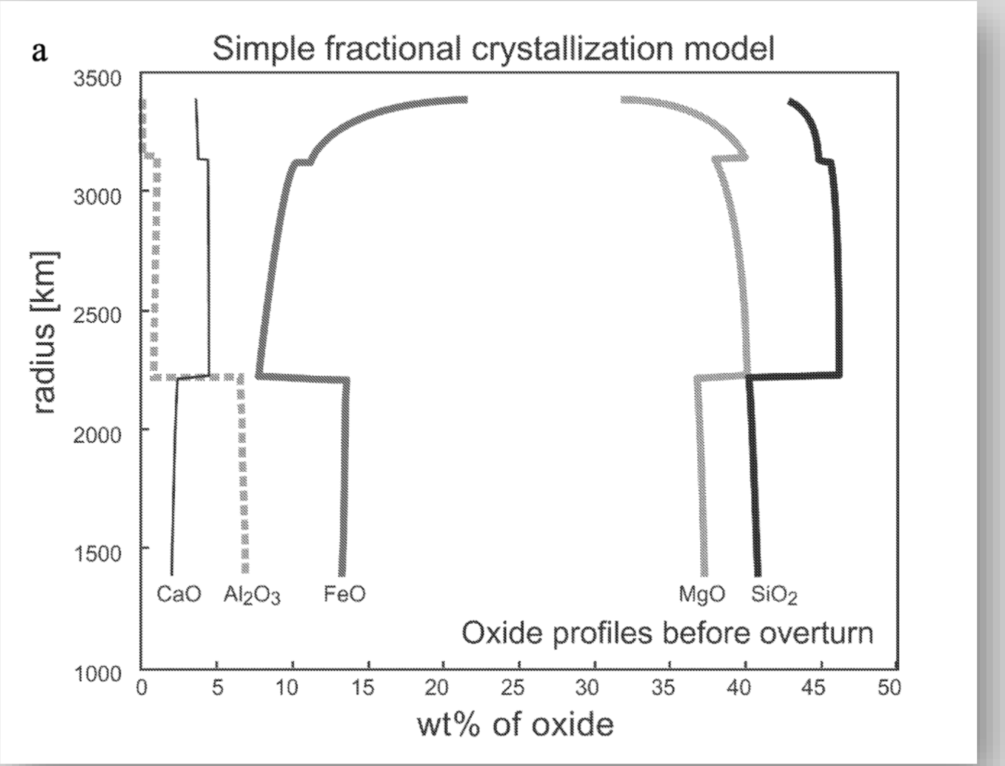
Mantle temperature

$$\rho c \frac{\partial \langle T \rangle}{\partial t} = -A_p q_m + Q + \Delta H_f A_s \rho \frac{dr_s}{dt}$$

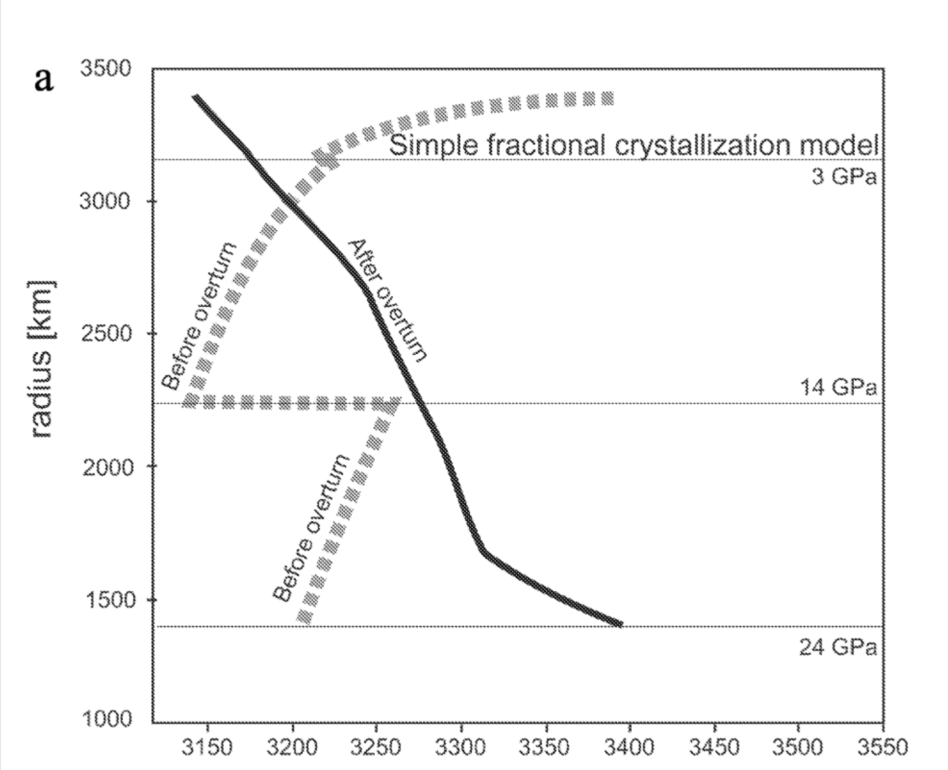
Thermal boundary layer

$$\delta = \frac{k\,(T_m - T_s)}{q_m}$$

Crystallization of magma ocean



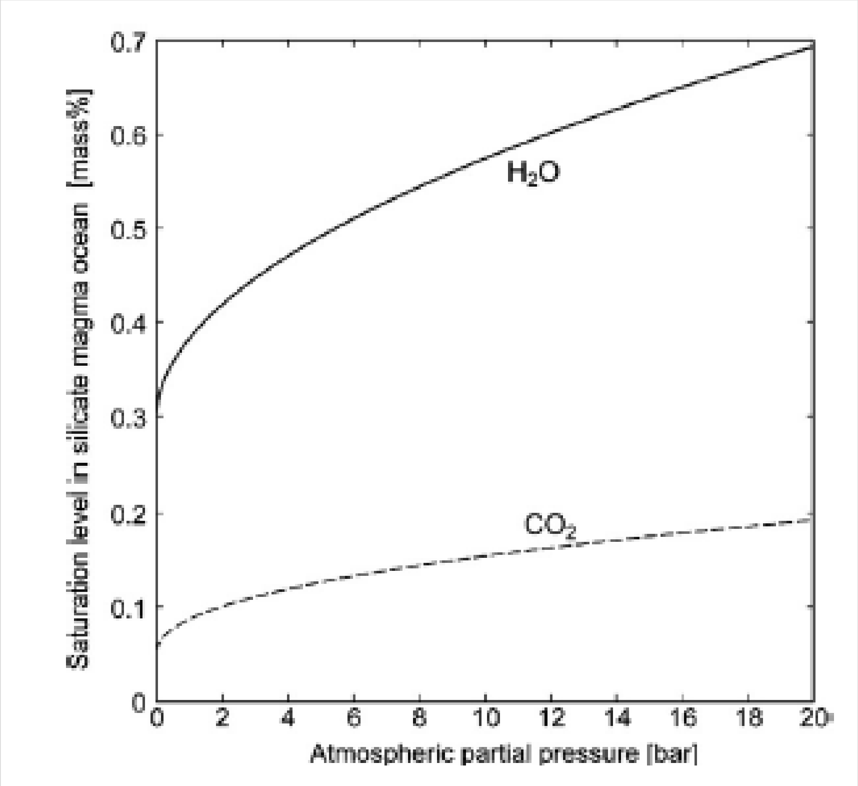
Elkins-Tanton et al. (2003)



Fractional crystallization occurs when the solid crystals settle out of the magma ocean and the liquid evolves in composition.

Fractional crystallization enriches the melt in FeO, which leads to higher densities of crystals forming at top of the mantle. This would produce a density inversion.

Volatile solubility in melts

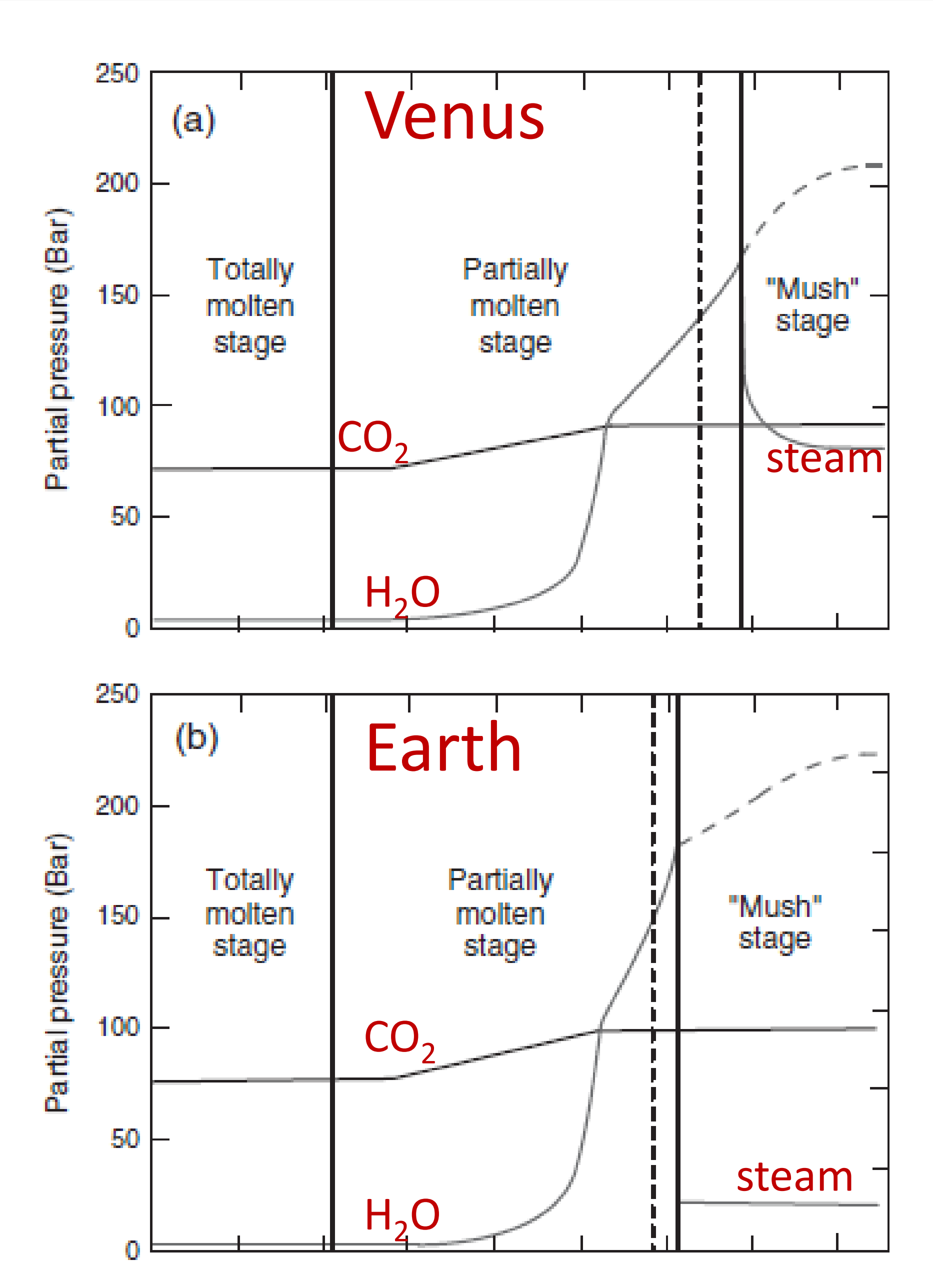


Elkins-Tanton (2008)

Volatiles are soluble in the silicate melt, depending on silicate composition.

Water is more soluble than CO_2. This will govern their relative abundances in the magma ocean atmosphere during solidification.

Volatile evolution



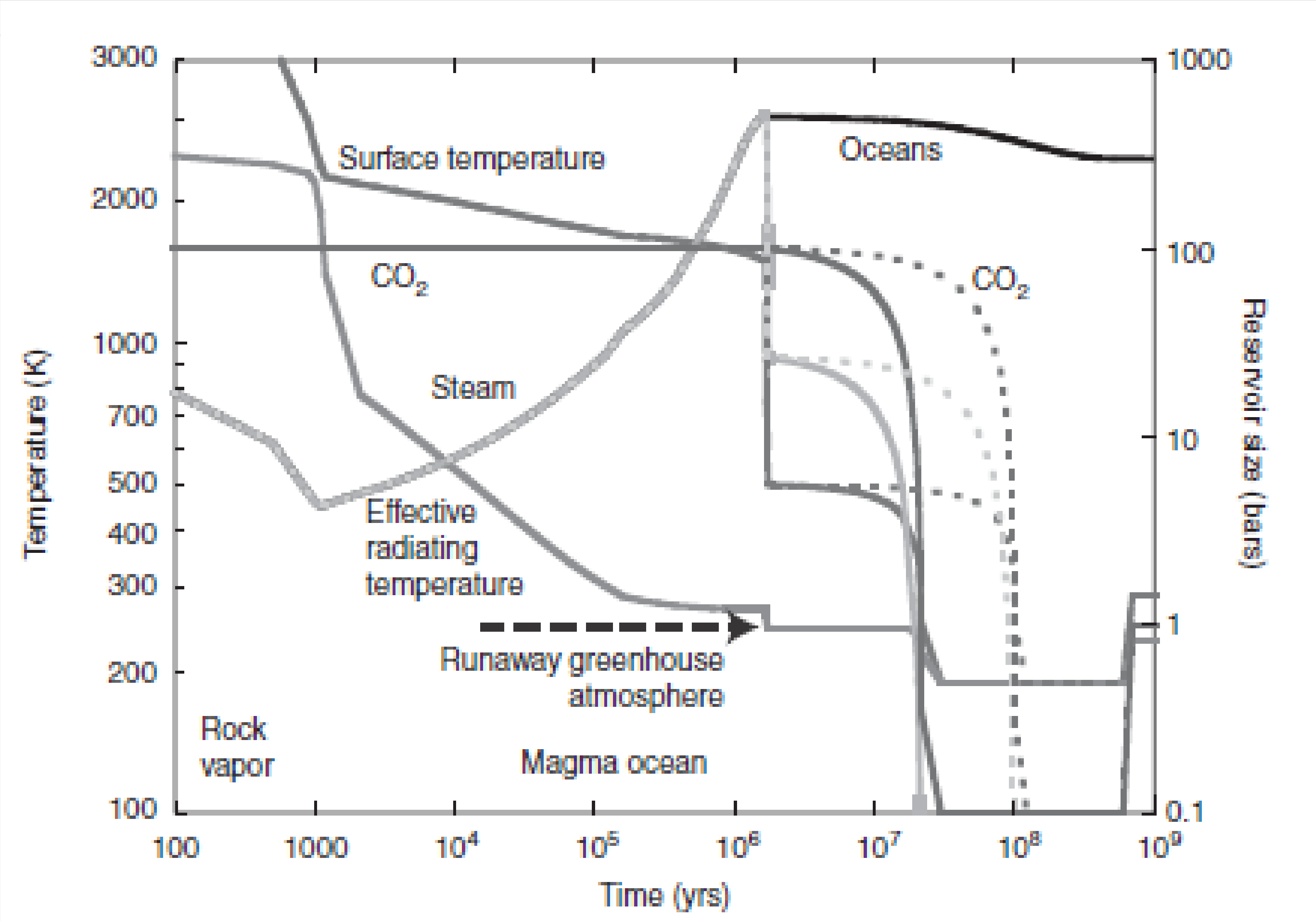
LeBrun et al. (2013)

Because CO_2 is less soluble in the silicate melt, it will be found dominantly in the atmosphere, and its evolution is only minimally affected by solidification.

Water is initially minimally present in the atmosphere until saturation is reached in the magma ocean. Late outgassing is substantial.

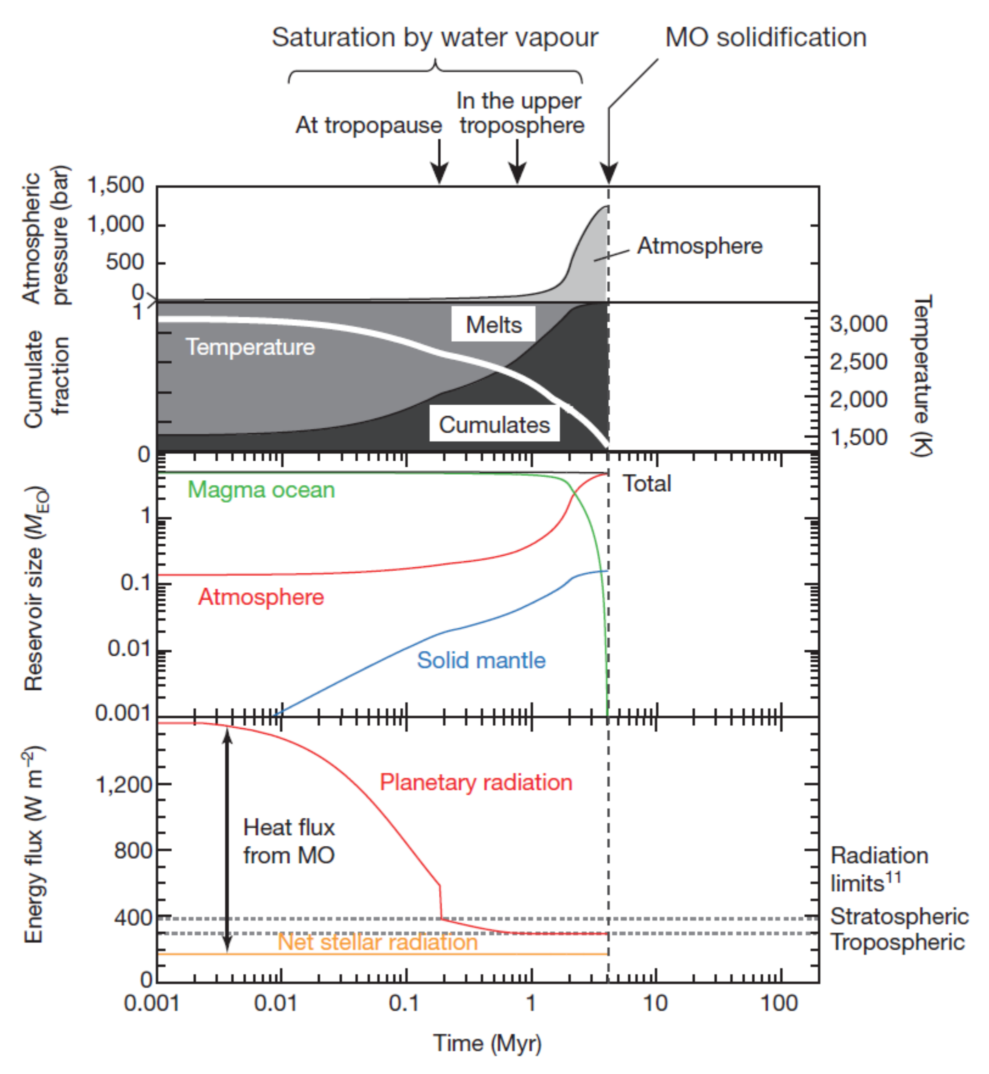
Condensation produces early clouds and then oceans.

Stages of the giant impact atmosphere on Earth



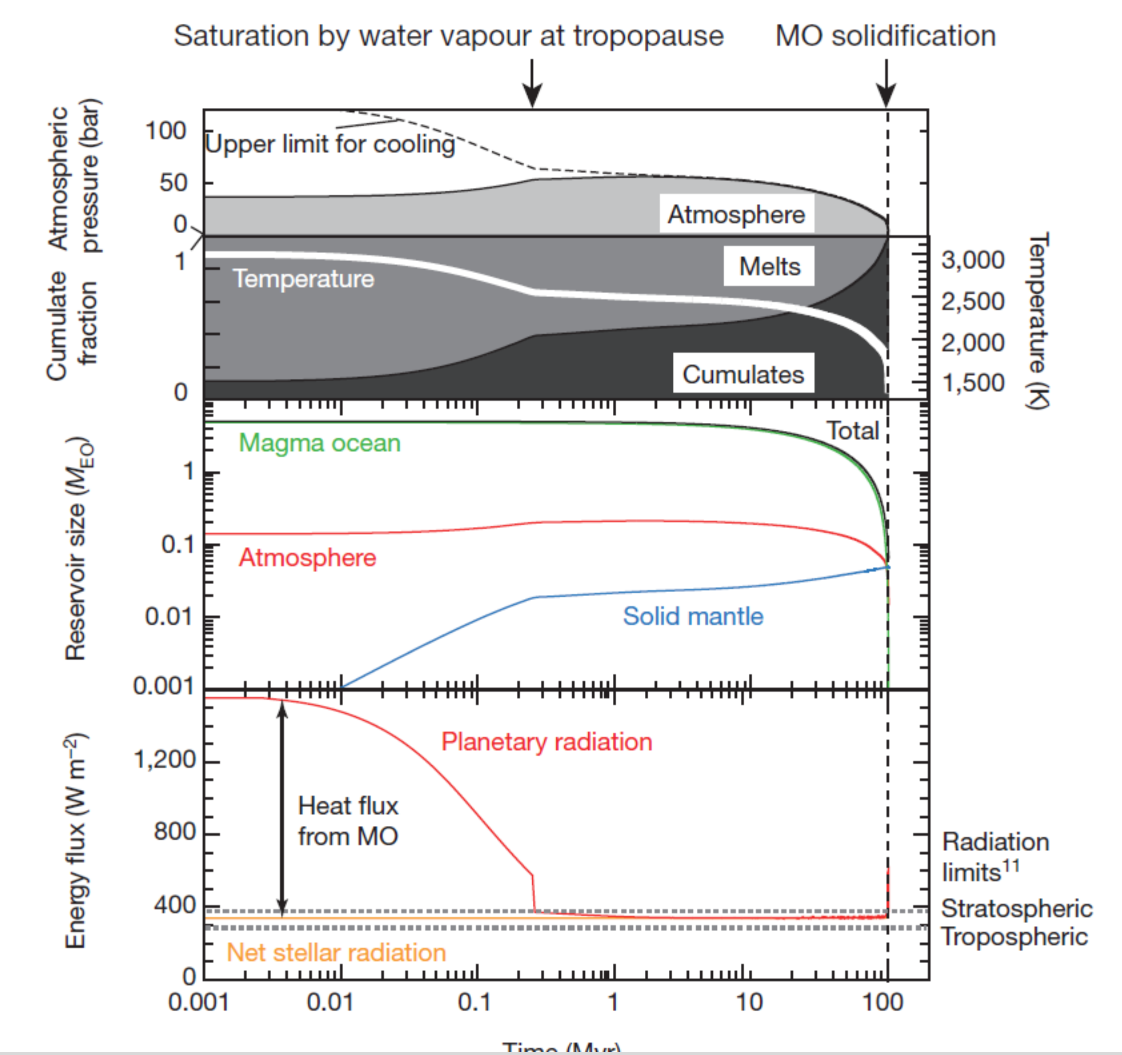
Based on Zahnle, Schaefer, Fegley (2010)

Earth-like planets have oceans.

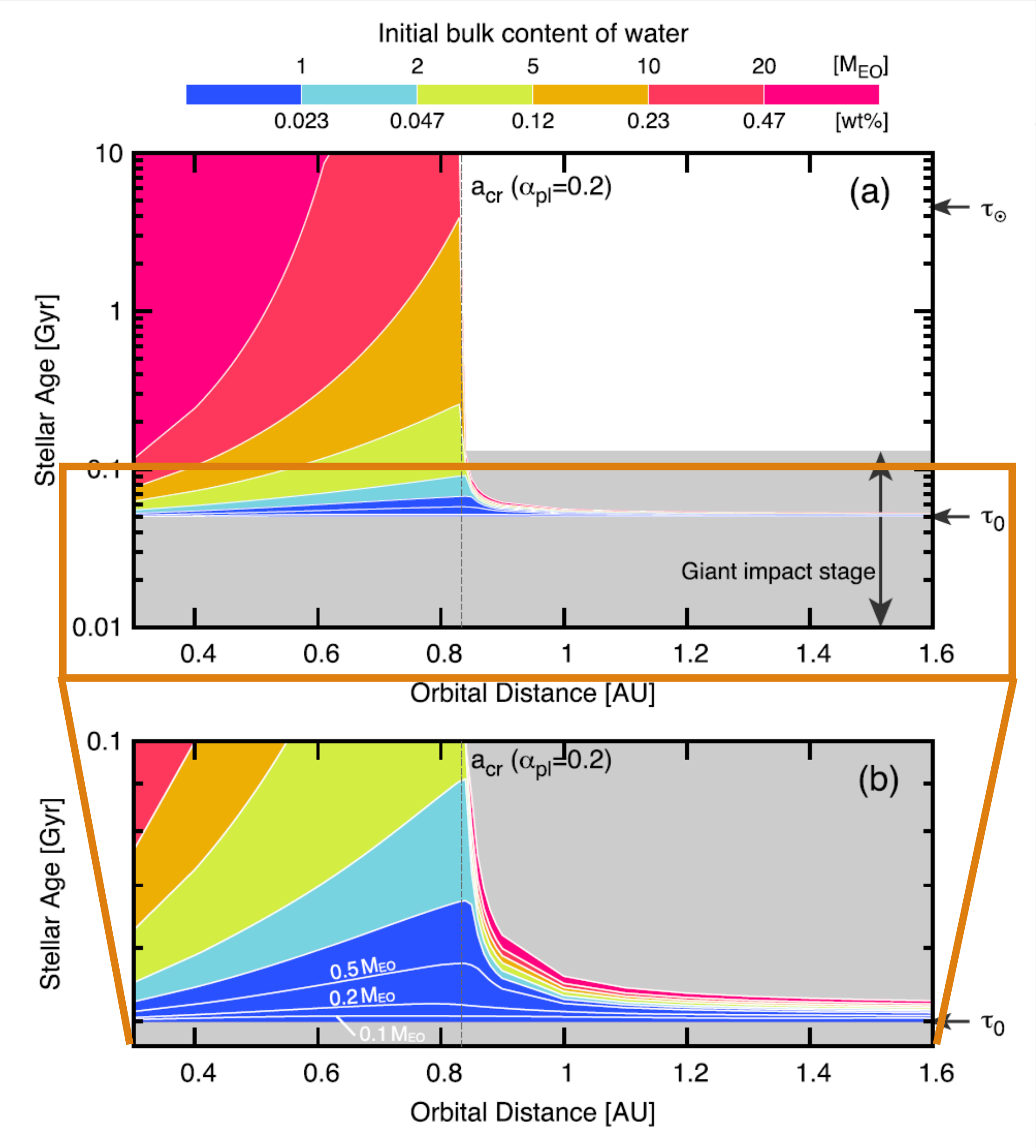


Hamano et al. (2013) Nature

Venus-like Planets don't.



Total Hydrogen loss (= water loss) happens before planet cools enough that water would have condensed



Hamano et al. (2015)

Magma ocean occurrence as a function of stellar age and orbital distance for a G dwarf

Colors show how much water the planet must have for a magma ocean to survive to these ages.

Oxygen Fugacity

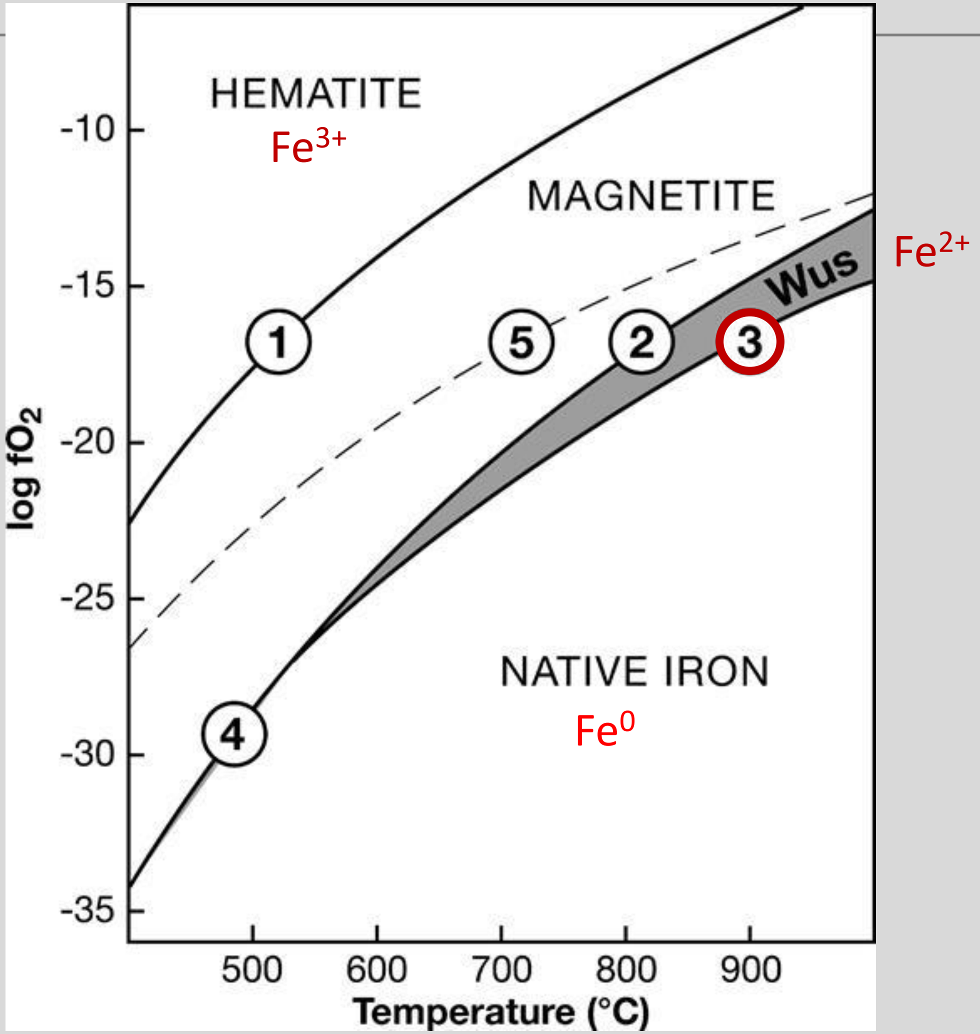
Iron – Wüstite (IW) buffer

$2(1\text{-}x)\ Fe + O_2(g) = 2\ Fe_{1\text{-}x}O$ (wüstite)

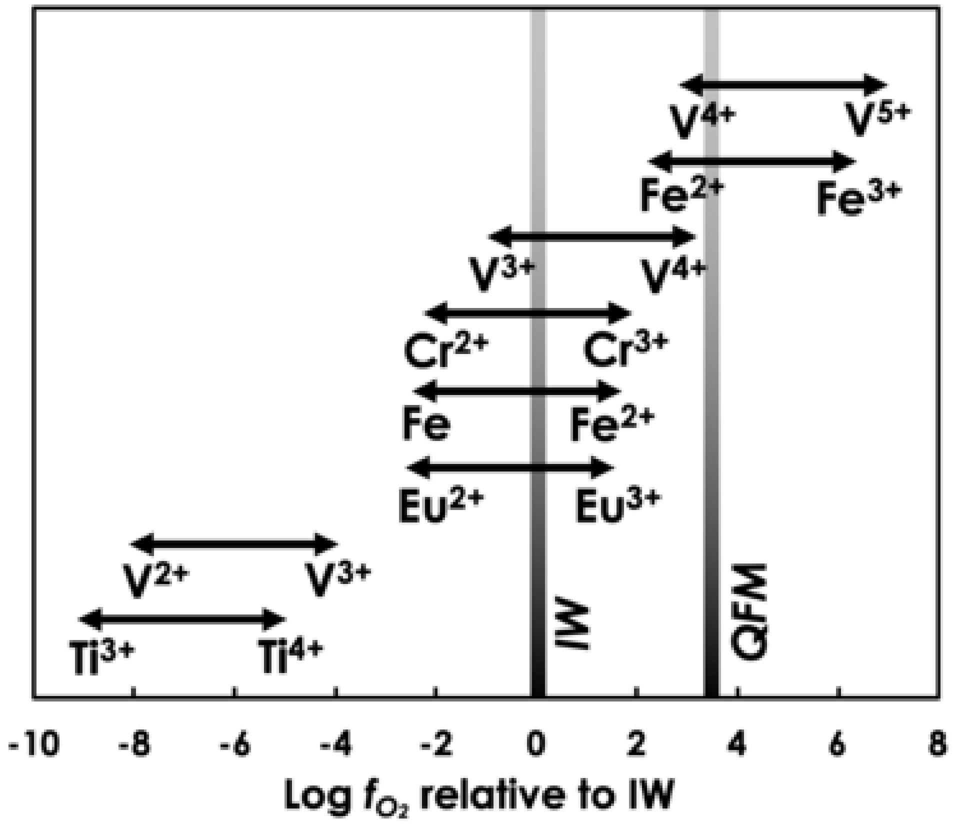
Gas reactions

$H_2(g) + ½\ O_2(g) = H_2O(g)$ $\Rightarrow$ $$\frac{P_{H_2O}}{P_{H_2}} \propto f_{O_2}^{1/2}$$

$CO(g) + ½\ O_2(g) = CO_2(g)$ $\Rightarrow$ $$\frac{P_{CO_2}}{P_{CO}} \propto f_{O_2}^{1/2}$$



Seifert et al. (2010) Eu. J. Min.

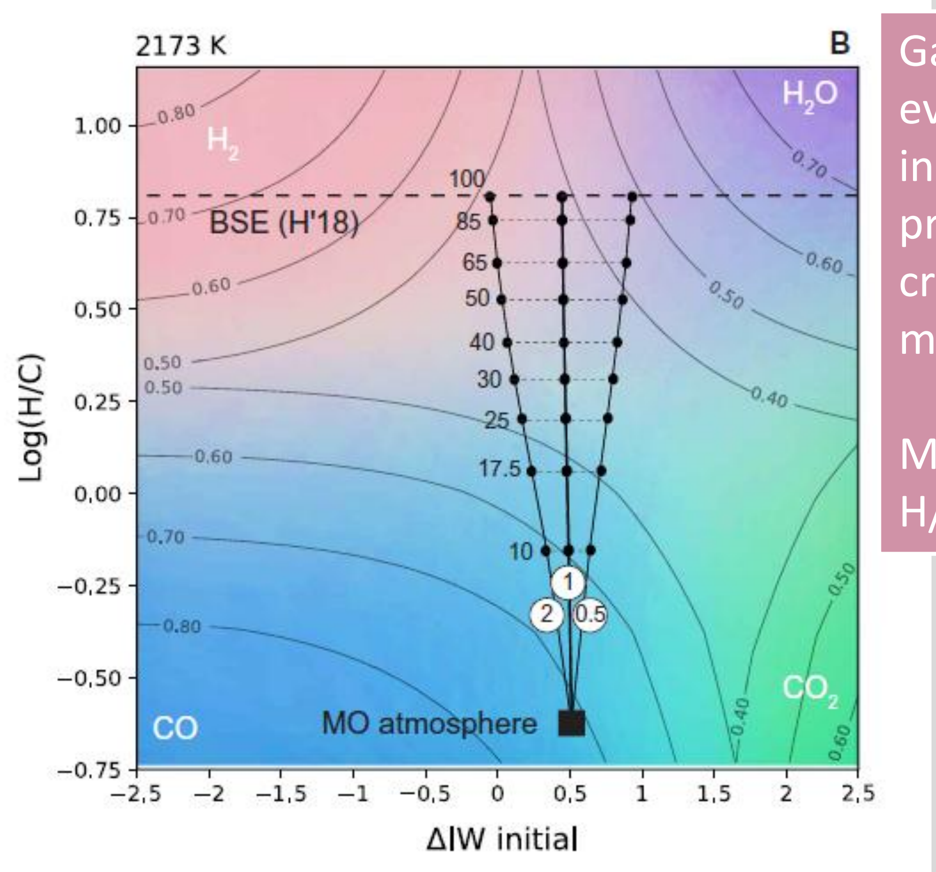


This figure shows how the **oxidation state** of different transition elements depends on oxygen fugacity.

Iron is the most abundant of these elements, so it has the biggest effect on mantle redox.

The oxygen fugacity (f_{O2}) in equilibrium with mantle rocks depends on a) how much Fe^{3+} is present, and b) which minerals the Fe^{3+} is found in.

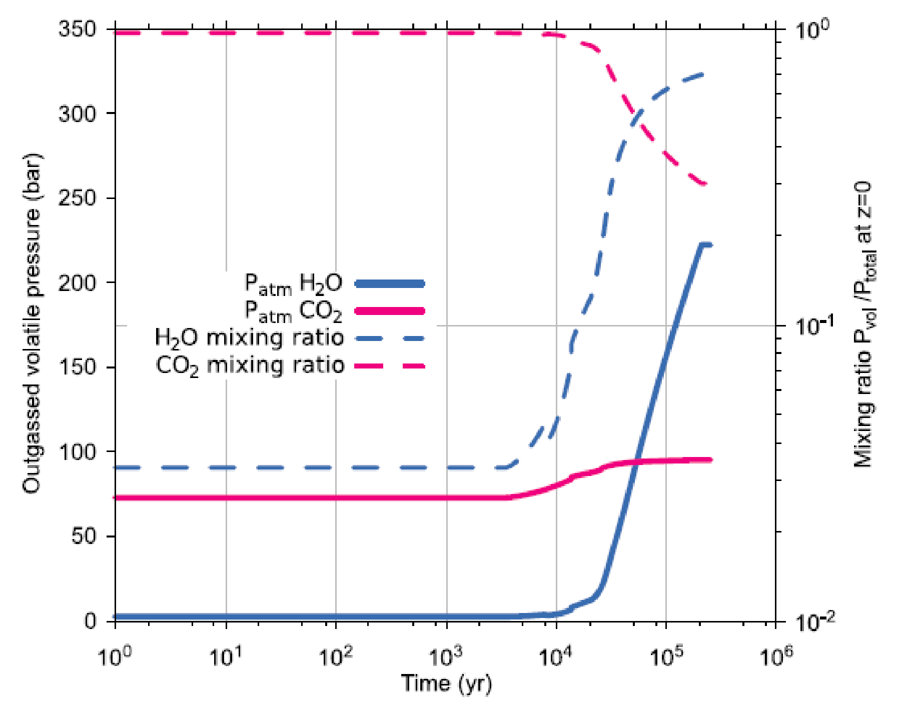
Magma ocean atmospheres



Sossi et al. (2020)

Gas compositions will evolve due to differences in volatile solubilities and progressive crystallization of the mantle.

More outgassing = higher H/C ratios in gas

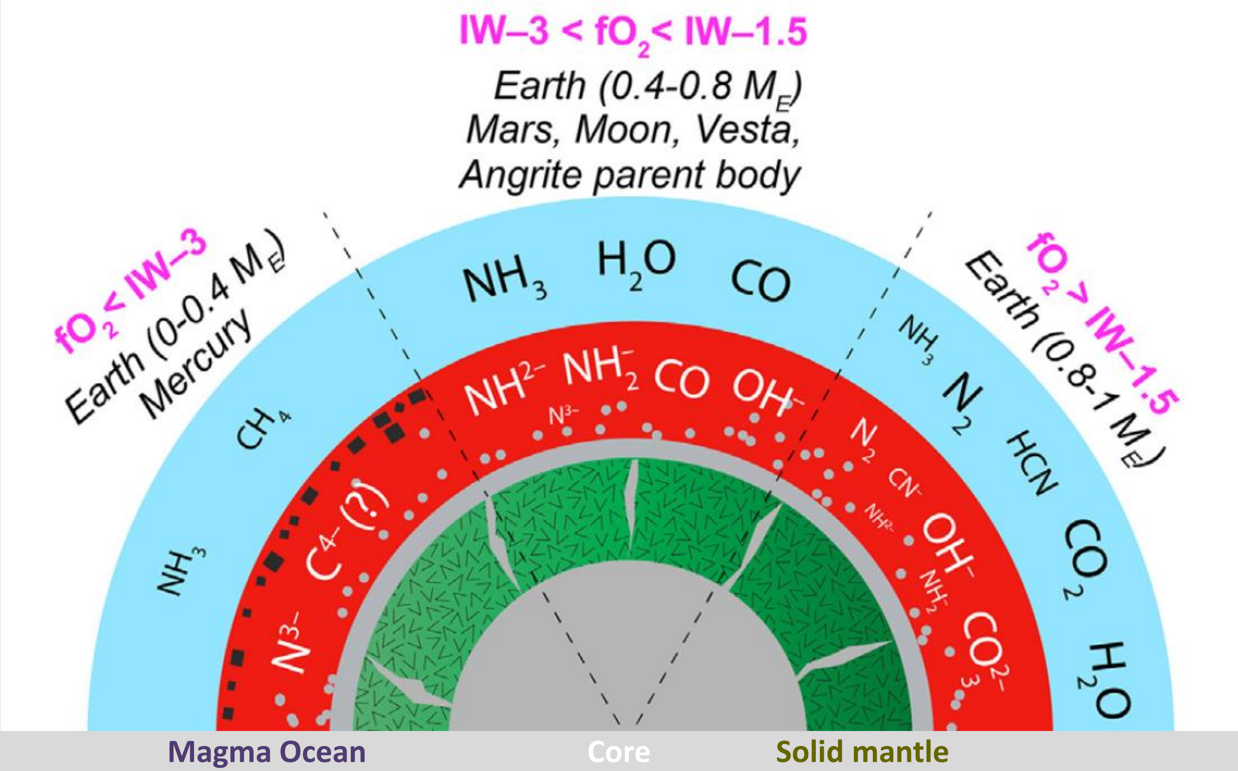


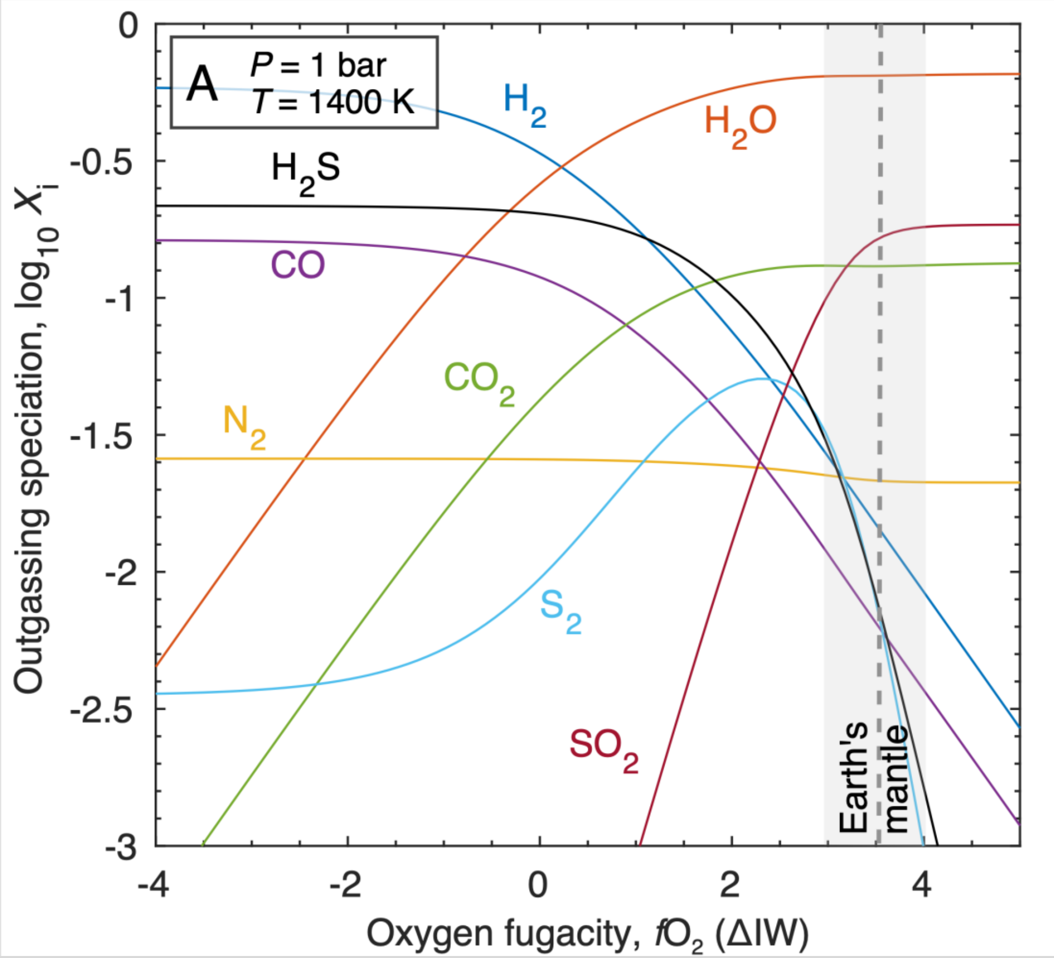
Nikolaou et al. (2019)

CO_2 is less soluble, therefore more abundant in early gas, H_2O outgasses later.

More outgassing = higher H/C ratios in gas

Mantle fO_2 controls the outgassed atmosphere composition

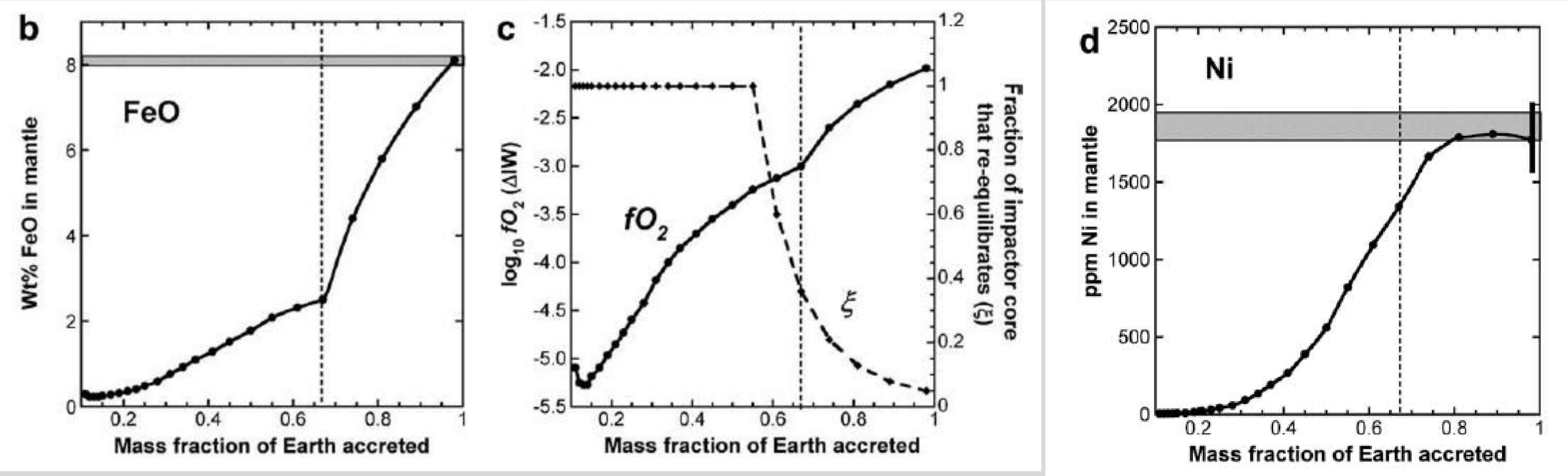




Grewal et al. 2020

Lichtenberg, Schaefer, Nakajima, Fischer, *PPVII*

Core formation models require fO_2 evolution for Earth

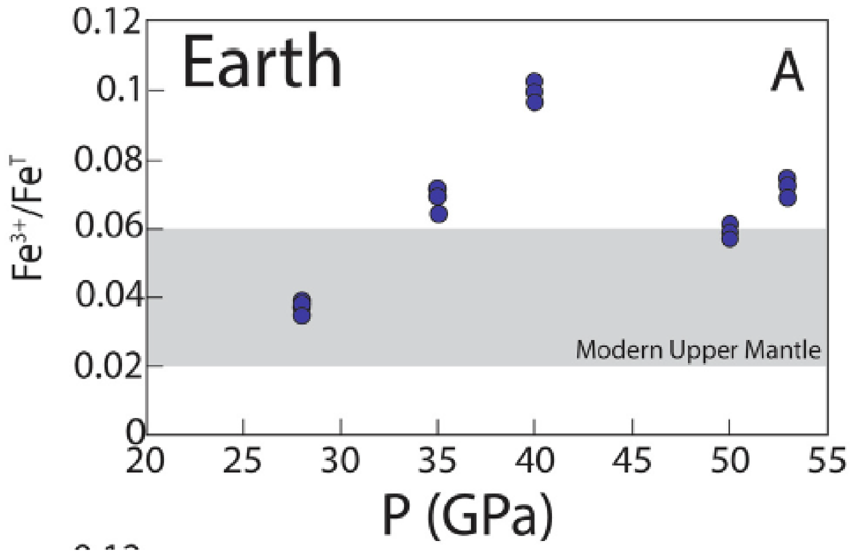


To match the trace element concentrations of siderophile elements in the mantle, core formation models typically require the FeO content and bulk fO_2 to increase during the formation of the Earth.

Rubie et al. 2011, 2015, Badro et al. 2015 and many others

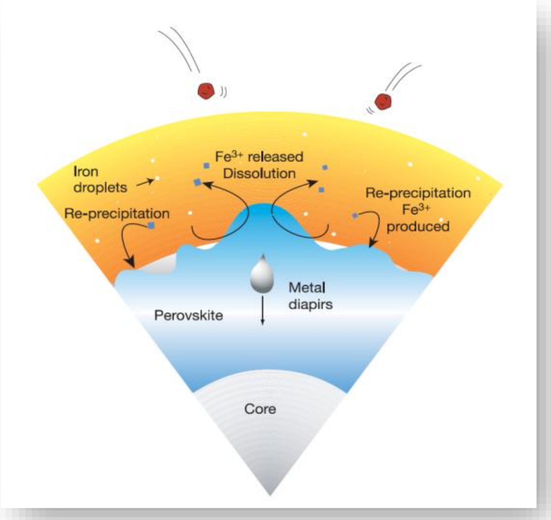
Rocky planets are born with non-zero Fe^{3+}

Fe^{3+} can be produced by mantle self-oxidation of iron
at high pressure by the disproportionation reaction:

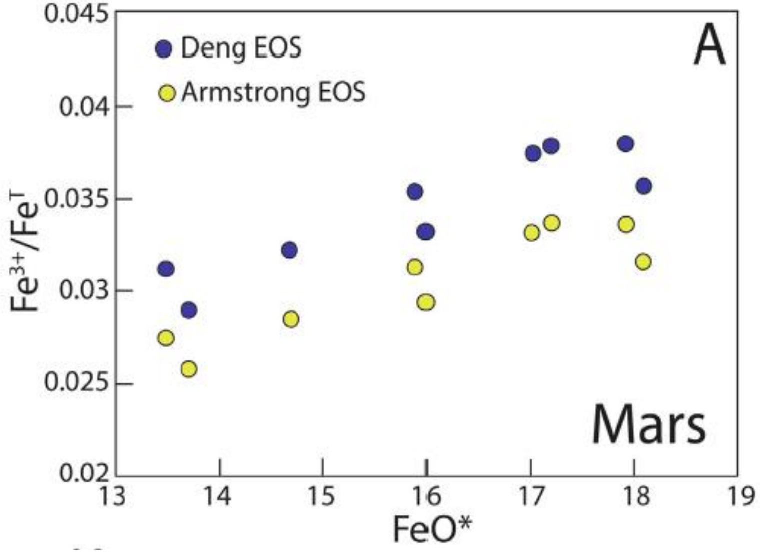


$$3\ Fe^{2+}O = 2\ Fe^{3+}O_{1.5} + Fe^{0}$$

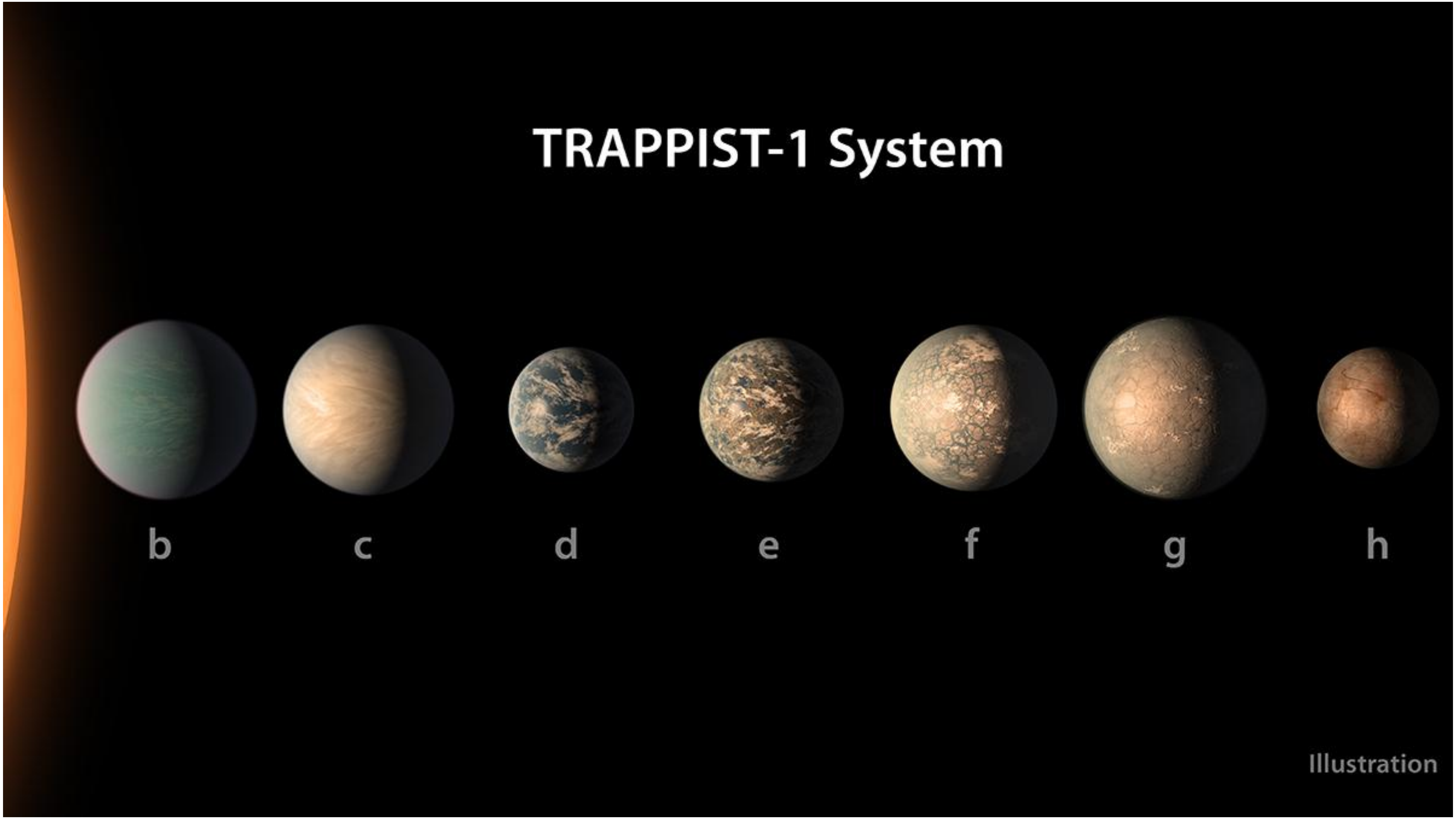
Silicate melt	Silicate melt	Liquid metal



Wood et al. (2006)

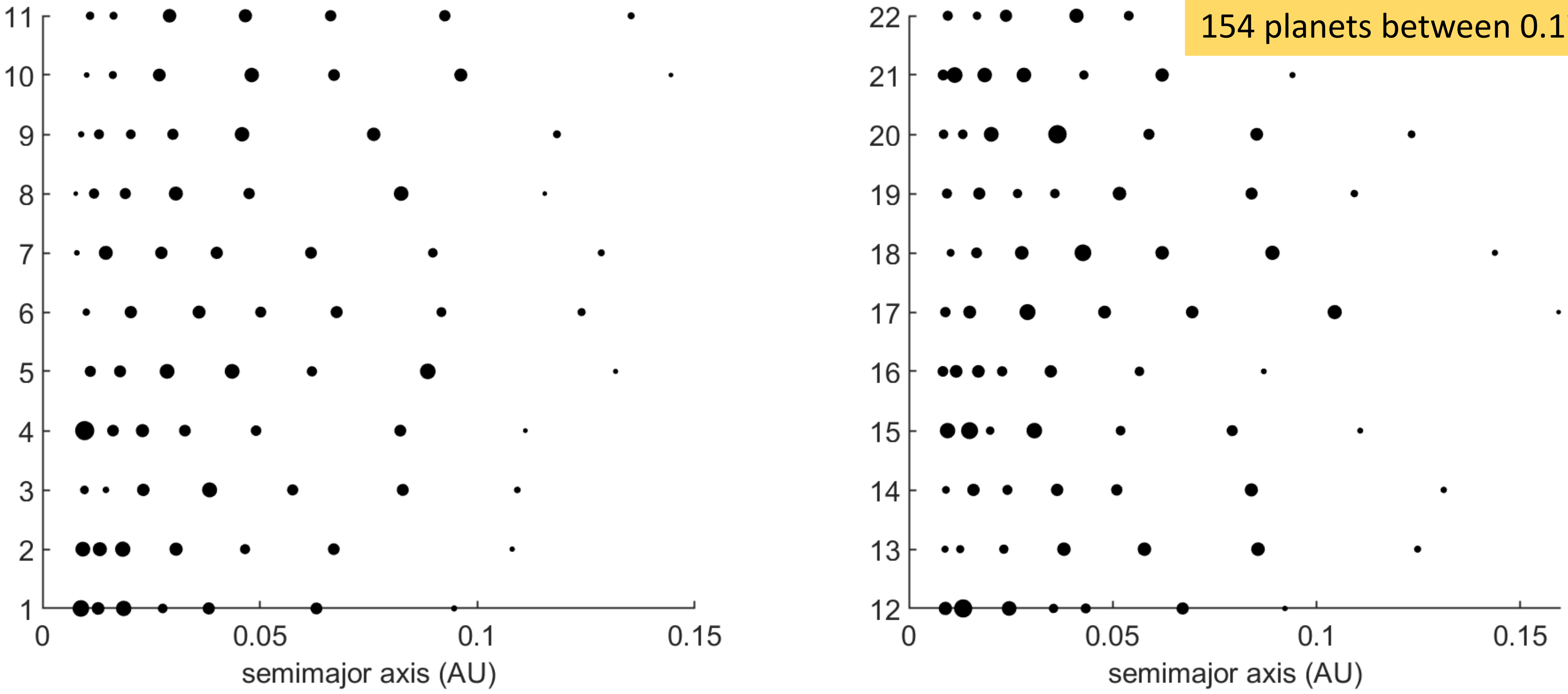


Hirschmann (2022)


TRAPPIST-1 System
b
c
d
e
f
g
h
Illustration

Local Formation of the TRAPPIST-1 system

Clement et al. (2024)
22 simulations
154 planets between 0.17 – 2.7 M_{Earth}



Metal-silicate equilibration model

$$2Fe^{0}_{metal} + SiO_{2,\,silicate} = Si_{metal} + 2Fe^{2+}O_{silicate}$$

$$Fe^{2+}O_{silicate} = Fe^{0}_{metal} + O_{metal}$$

$$3\ Fe^{2+}O = 2\ Fe^{3+}O_{1.5} + Fe^{0}$$

Silicate melt | Silicate melt | Liquid metal

This is called the **iron disproportionation reaction**. It is only energetically-favorable at pressures above ~25-30 GPa, so it is likely to happen during high pressure metal-silicate equilibration.

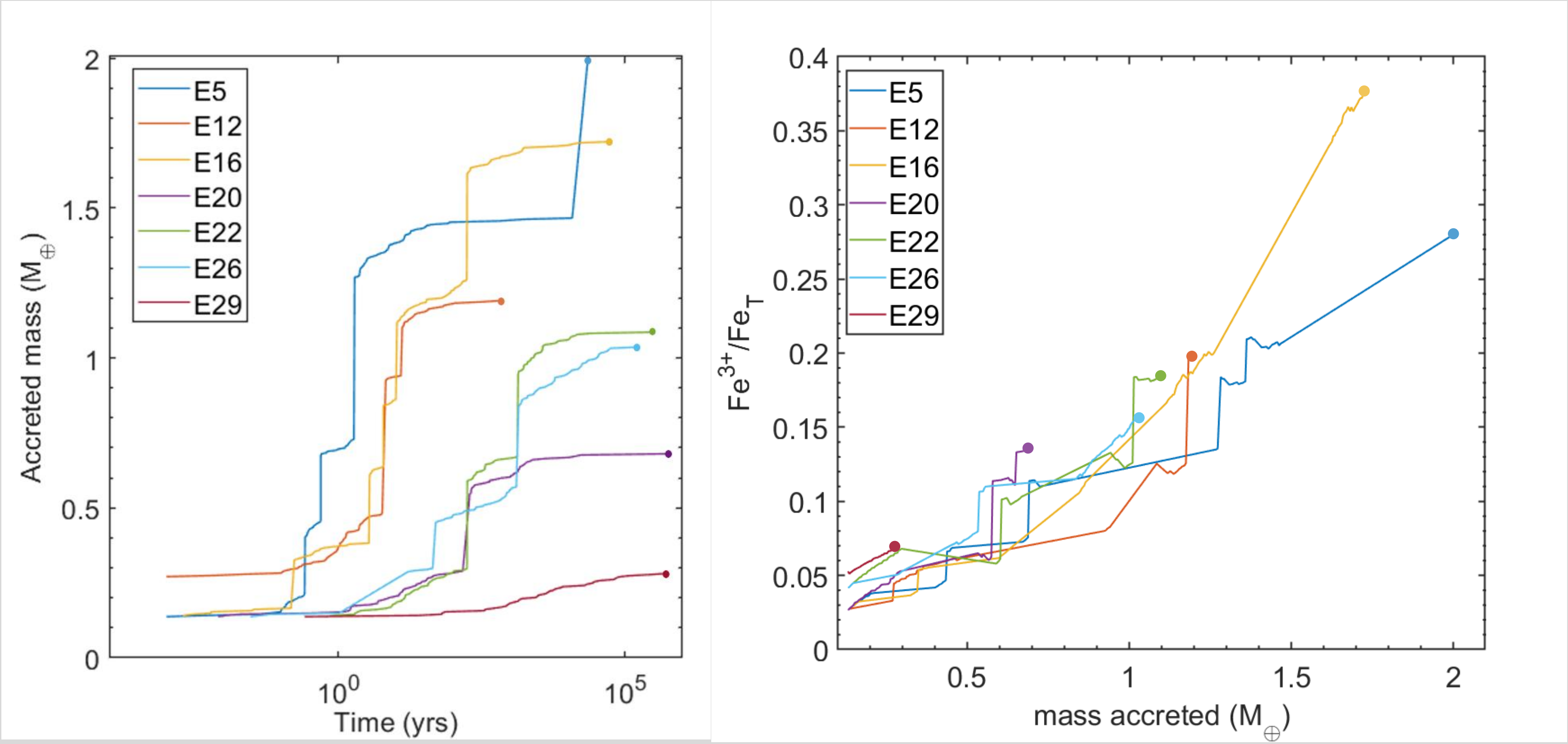
$$K_d = \frac{(a_{Fe}^{metal})^2\, a_{SiO2}^{sil}}{(a_{FeO}^{sil})^2\, a_{Si}^{metal}}$$

$$K_d = \frac{a_{Fe}^{metal}\, a_{O}^{metal}}{a_{Fe}^{mw}}$$

$$K_d = \frac{(a_{FeO_{1.5}}^{silicate})^2\, a_{Fe}^{metal}}{(a_{FeO}^{silicate})^3}$$

These are the three main equations that we solve together. The K_d's are given by experimental constraints, we solve for the concentrations of the different phases.

Mantle Fe^{3+}/Fe_T jumps after impacts



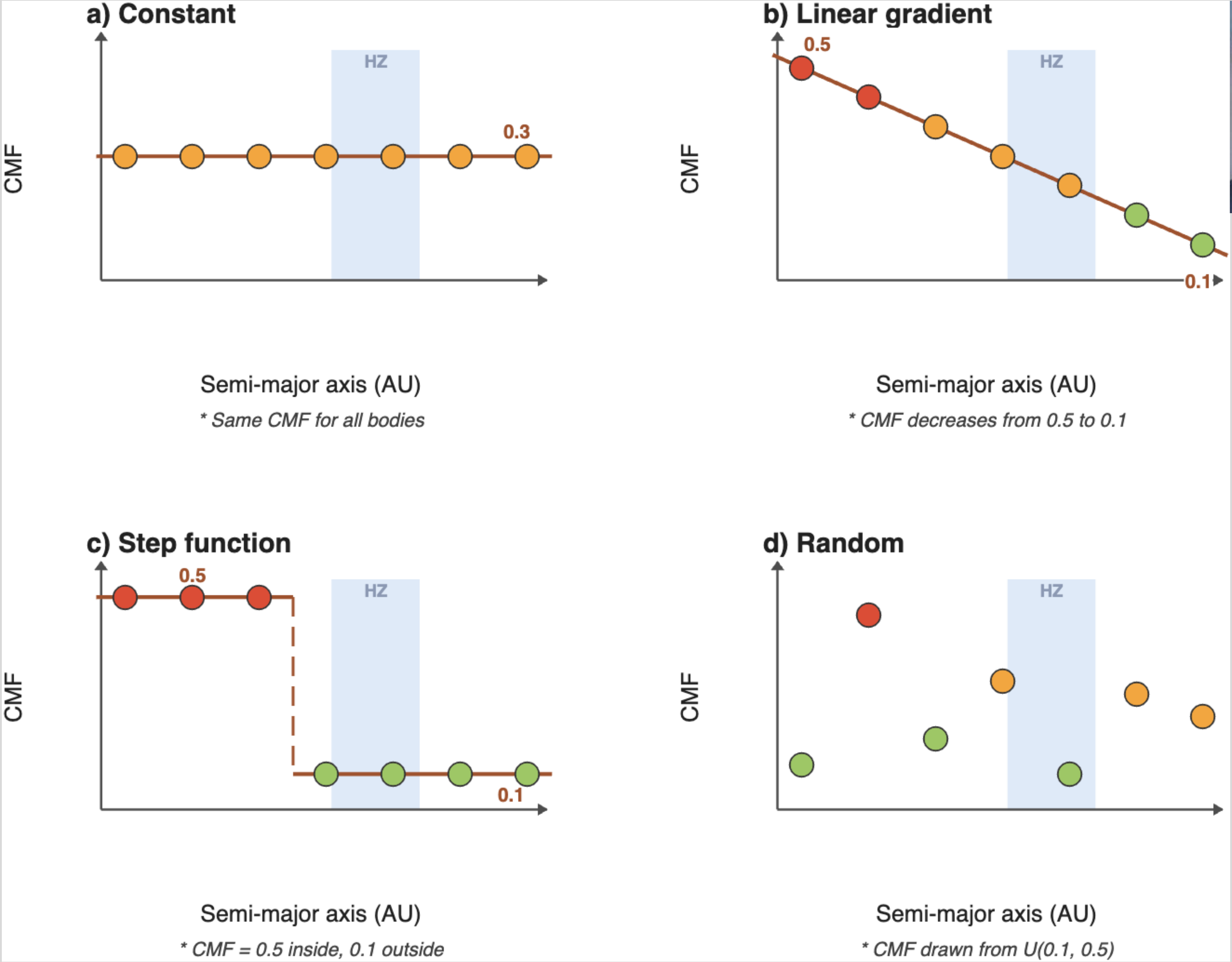
Using coupled N-body simulations with metal-silicate equilibration models can allow us to start computing mantle oxygen fugacity during accretion.

Mantle Fe^{3+} content responds primarily to changes in the pressure of metal-silicate equilibration.

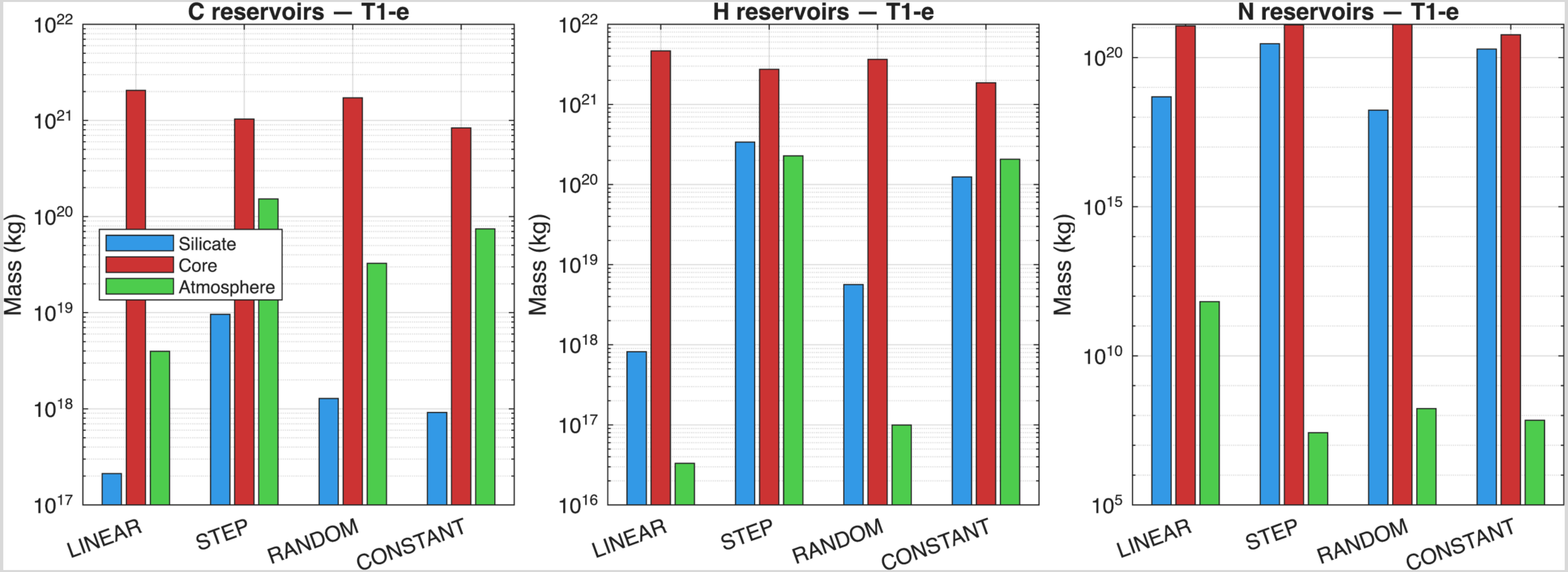
We used the same set of impact simulations but applied different initial compositions to the planetesimals.

We varied the core-mass fractions (CMF) in different ways, to see if patterns in the planetesimal population are preserved in the final planet population.

Carbon and Nitrogen are more soluble in the metal alloy under very reducing (low oxygen fugacity) conditions.



Volatile Partitioning



The initial CMF distribution in the planetesimal population can lead to vary different distributions of the volatile elements between core, mantle and atmosphere, even with identical impact histories.

Takeaways

- Planetesimals are likely differentiated before accreting onto larger planets. Cores of smaller impactors may emulsify when descending through the silicate melt
  - Trace element partitioning between mantle and core is used to quantify the conditions (e.g. pressure, temperature) of core formation
- Magma oceans solidify from the bottom up and control the first atmosphere composition by volatile solubilty.
  - Outgassed atmospheres change with the oxygen fugacity of the magma ocean.