Predictions of the Atmospheric Compositions of GJ 1132b

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GJ 1132b ?

- Earth-like exoplanet
- GJ1132: M3.5 dwarf(0.181 M_{solar})
- Discovered by transit method
- located only 39 light years away

Credit: Dana Berry, from NASA website

Parameter	Values
Present-day stellar luminosity L ($L_\odot$)	0.00438
Orbital distance a (au)	0.0153
Planetary mass M_p ($M_\oplus$)	1.62
Planetary radius r_p ($r_\oplus$)	1.16
Core mass fraction M_c (M_p)	0.262
Core radius r_c (r_p)	0.54
Surface gravity g_0 (m s^{-2})	11.8
Planetary albedo A	0.75

Closer than habitable zone

Transit method

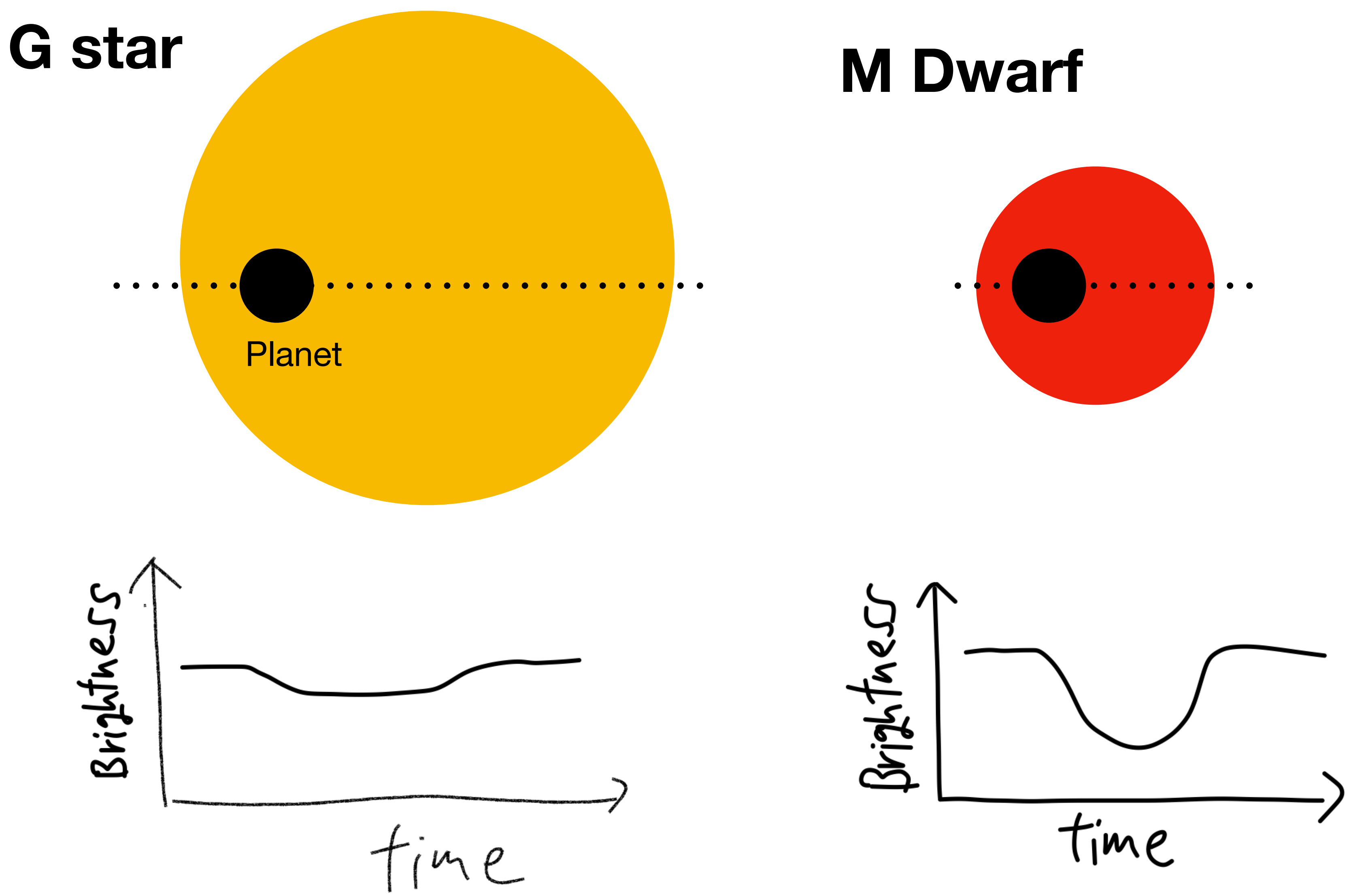


Fig. Transition method

M dwarf is suitable for transit method!

Exoplanets around M dwarf

M dwarf

- Habitable zone is close to the star
  → suitable for observation

Habitability

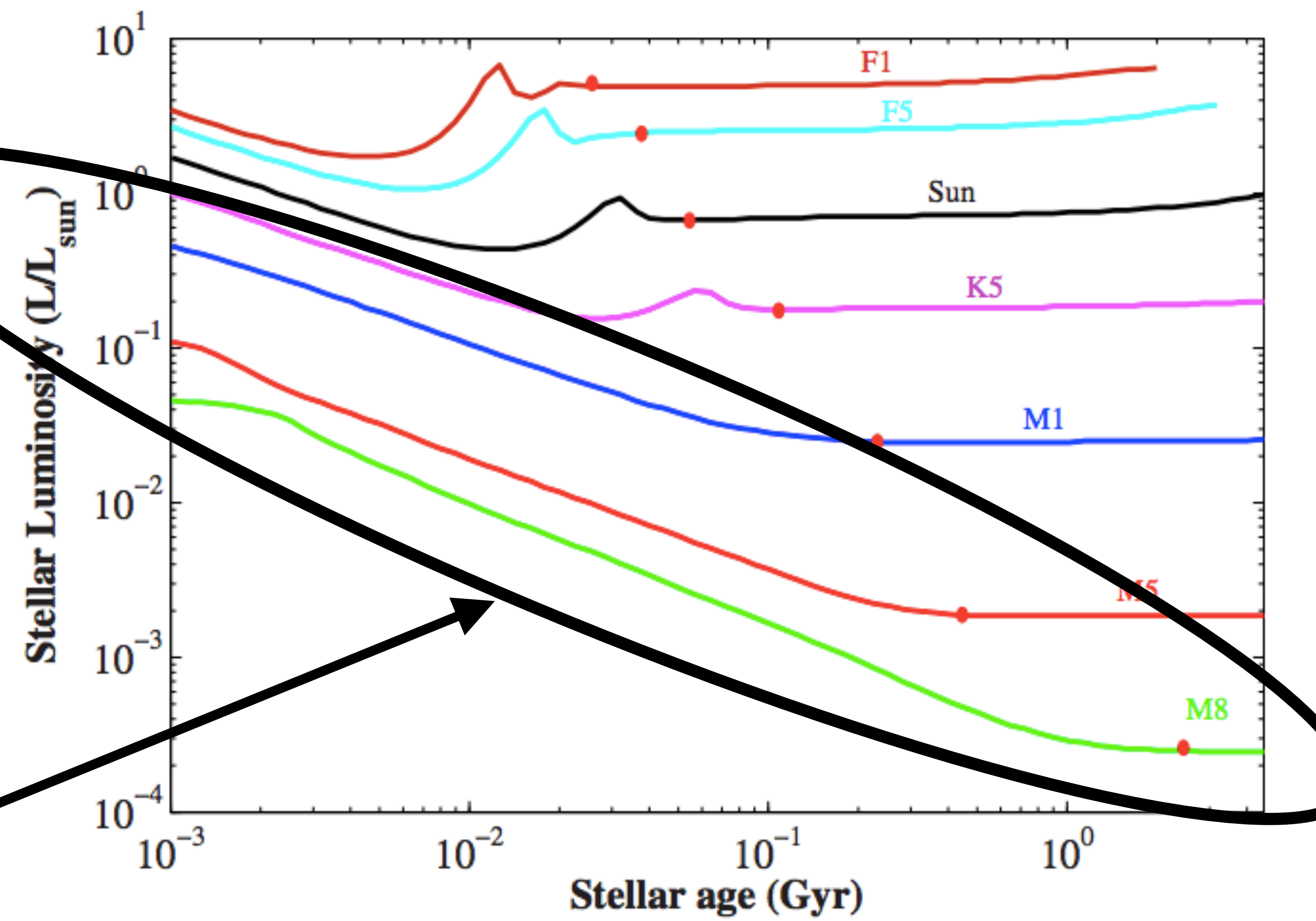


Fig. Evolution of stellar luminosity (Ramirez et al. 2014)

Strong stellar flux is expected

XUV = X-ray + EUV

We need to know the early evolution of exoplanets around M dwarf

Early stage of Planet history

Accretion of Planetesimals → **Magma ocean** → **Magma ocean solidification** → **Crust is formed**

Questions

H$_2$O would remain?

Abiotic O$_2$ would build up?

Essential components

1. Stellar XUV flux
2. initial H$_2$O inventory
3. initial FeO content

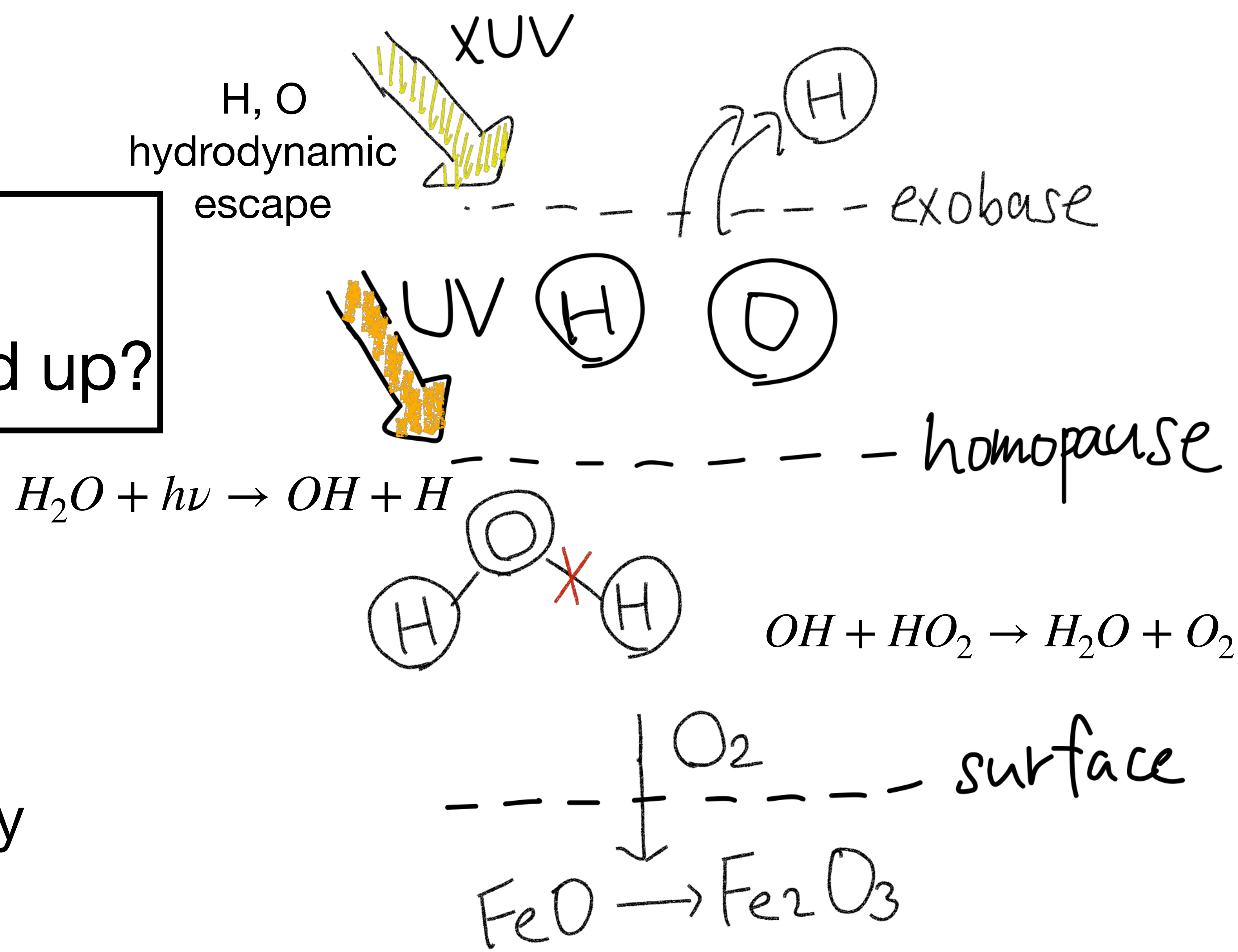


Fig. Magma ocean model illustration

Purpose of this study

Previous Study

Several studies of atmospheric loss of H_2O-rich rocky planets orbiting M dwarf have done but without fully interior model.
(Luger & Barnes 2015; Tian & Ida 2015; Bolmont et al. 2016)

↓

Purpose of this study

To study atmospheric loss of H_2O and O_2 build up during and after magma ocean with a fully atmosphere-interior coupled model about GJ 1132b and predict its present atmosphere

このような研究の最終目標

- 水が保持できるか？
- 生命由来でない酸素がいつ、どのくらいの期間、どれくらいの量が大気に残るか？

$\longrightarrow$ **JWST, 大型地上望遠鏡によるフォローアップ**

$\longrightarrow$ **ハビタブル惑星の制約**

太陽系内の惑星に対して

- 恒星に近い系外惑星を観測することでマグマオーシャンモデルのテストができる

ex) 金星における水と酸素の散逸

Model Description

Model outline

q_m: mantle heat flux

Q_r: heat from radioactive decay

ΔH_f: heat of fusion of silicates

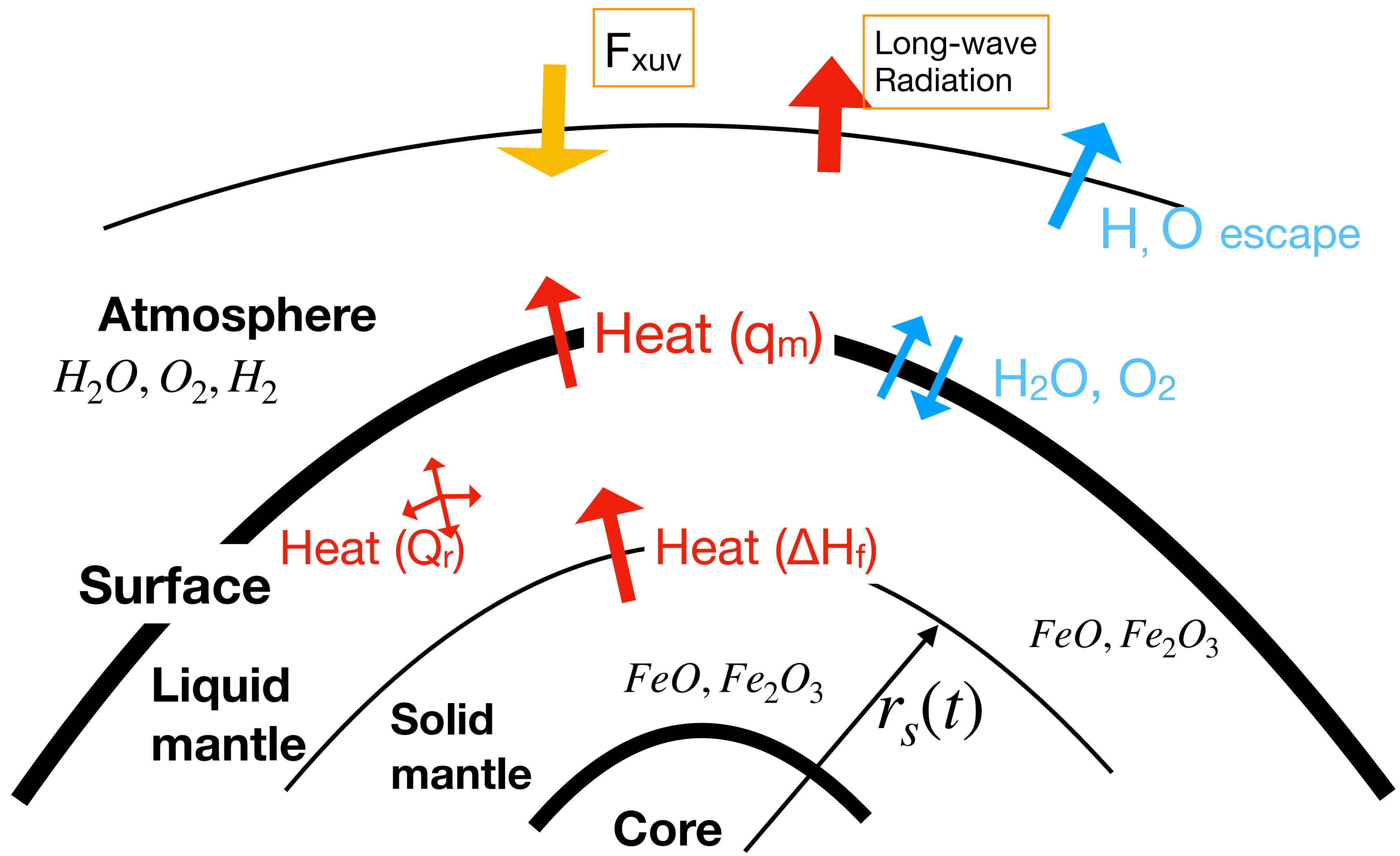


Fig. atmosphere-interior coupled model

Model components

- Heat flux balance
- Mantle solidification
- Exchange rate of H, O species at boundaries
- Escape flux of H and O
- Time-dependent stellar XUV flux

Thermal parametarization:

Lebrun et al. 2013,
Elkins-Tanton 2008,
Hamano et al. 2013

Atmospheric escape model

- XUV-driven hydrodynamic escape of H (Zahnle et al. 1990)

$$\phi = -\frac{\epsilon F_{XUV}}{4V_{pot}} \qquad V_{pot} = -\frac{GM_p}{r_p}$$

- The number flux Φ_2 of O is given by (Hunten et al. 1987)

$$\Phi_2 = \begin{cases} \Phi_1 \frac{X_2}{X_1} \frac{\mu_c - \mu_2}{\mu_c - \mu_1} & (\mu_2 < \mu_c) \\ 0 & (\mu_2 > \mu_c) \end{cases} \quad , \mu_c = \mu_1 + \frac{k_b T \Phi_1}{b_{12} g X_1 m_p}$$ (Cross-over mass flux)

After the abundances of O_2 exceeds that of H_2O

- Diffusion-limited escape of H and no more O escape

XUV assumption

The present-day XUV flux from GJ 1132 has not yet been measured

→ GJ 1214 as a proxy(Parke Loyde et al. 2016)

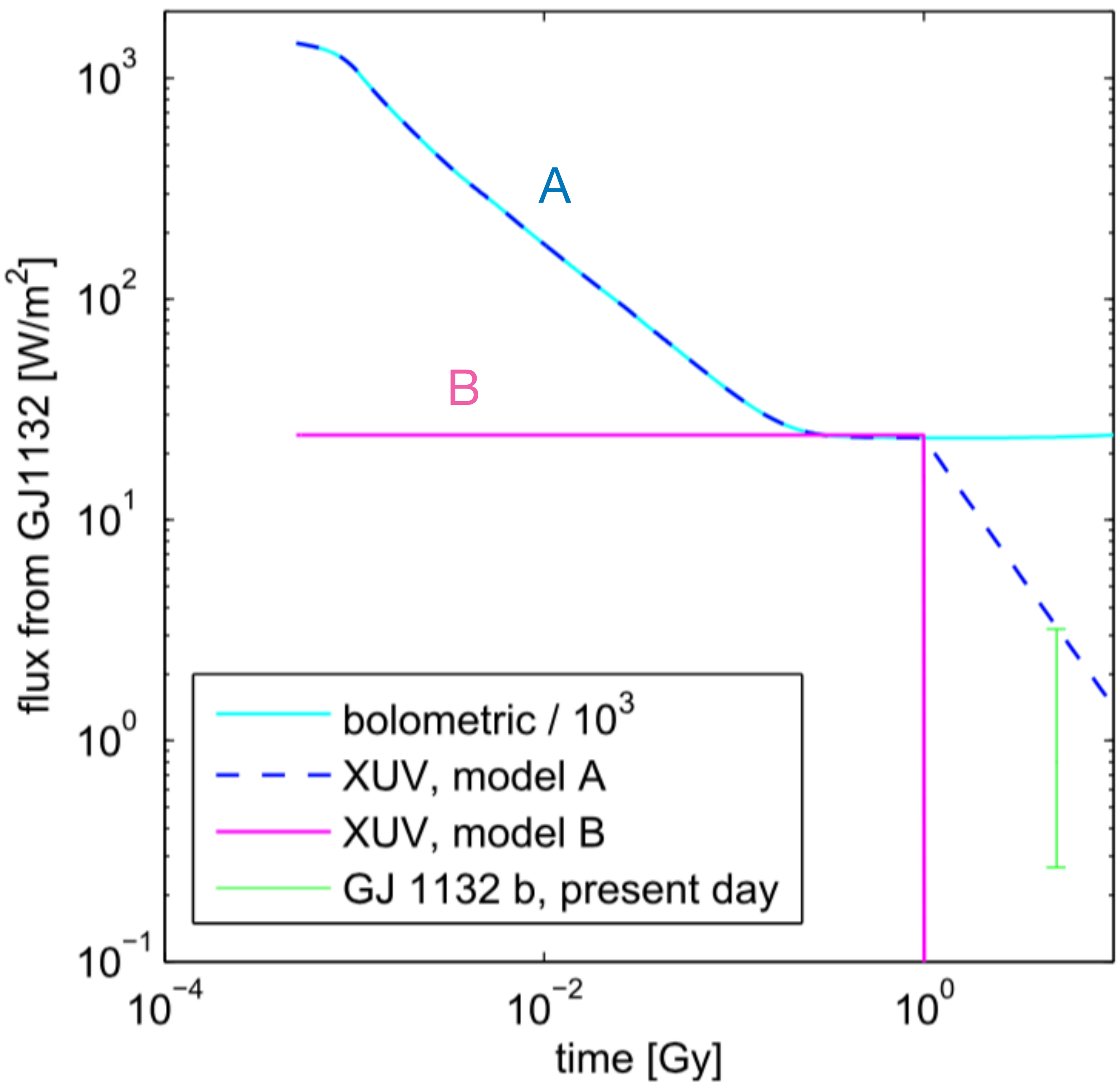


Fig. Scaled bolometric and XUV flux from GJ1132 at GJ 1132's orbit as a function of time

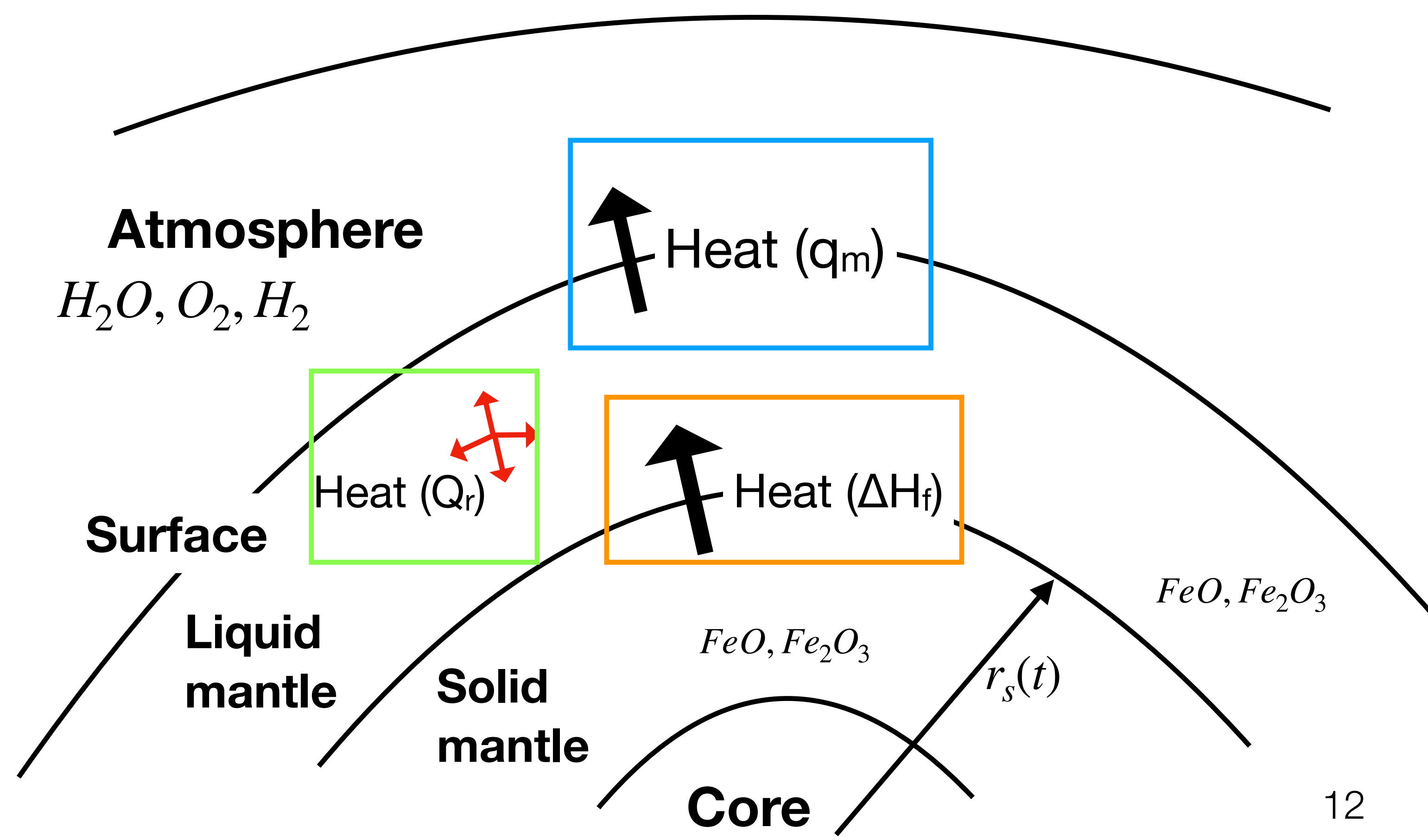
Thermal model①

- Equation of Heat balance in the magma ocean (Schaefer & Sasselov. 2015)

$$\frac{4}{3}\pi\rho_m C_p(r_p^3 - r_s^3)\frac{dT_p}{dt} = \boxed{4\pi r_s^2 \Delta H_f \rho_m \frac{dr_s}{dt}} - \boxed{4\pi r_p^2 q_m} + \boxed{\frac{4}{3}\pi\rho_m Q_r(r_p^3 - r_s^3)}$$

Heat change in the mantle | Fusion of silicates | Mantle heat flux | Radioactive decay

T_p: Potential temperature which governs the degree of melting and convection in the mantle
T_s: Surface temperature which is governed by heat balance
r_s : Radius of solidification



Thermal model②

- **Radius of solidification (Hirschmann. 2000)**

$$T_{\mathrm{p}}\left[1+\frac{\alpha g}{c_{\mathrm{p}}}(r_{\mathrm{p}}-r_{\mathrm{s}})\right]=ag\rho_{\mathrm{m}}(r_{\mathrm{p}}-r_{\mathrm{s}})+b$$

$$\frac{dr_{\mathrm{s}}}{dt}=\frac{c_{\mathrm{p}}(b\alpha-a\rho_{\mathrm{m}}c_{\mathrm{p}})}{g(a\rho_{\mathrm{m}}c_{\mathrm{p}}-\alpha T_{\mathrm{p}})^2}\frac{dT_{\mathrm{p}}}{dt}.$$

Solidification of the magma ocean proceeds from the bottom up

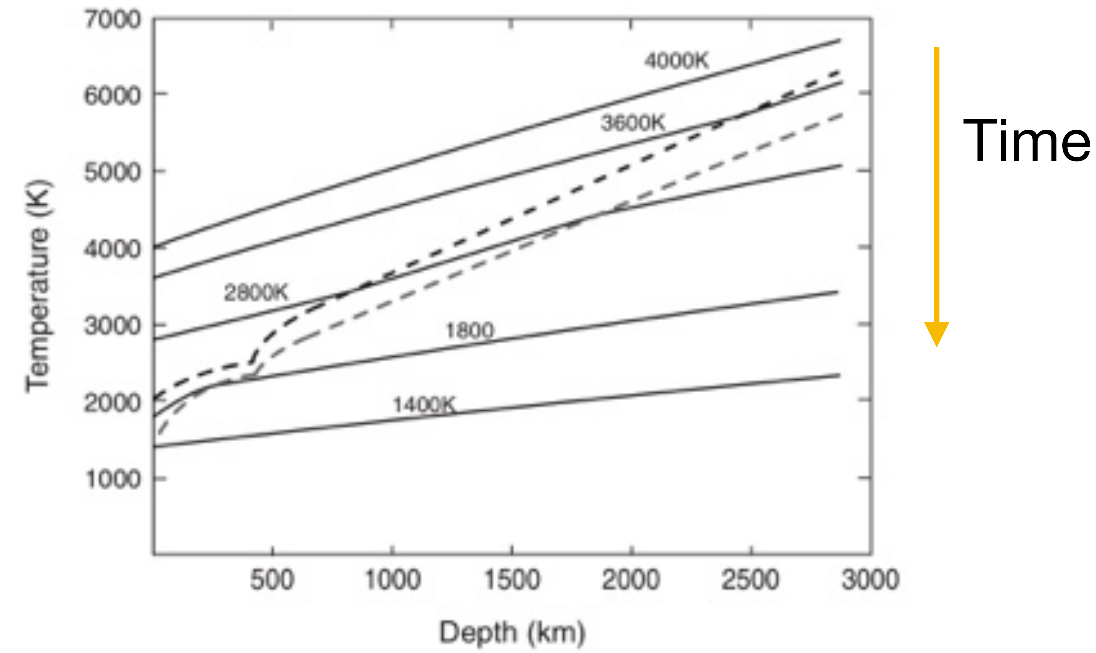


Fig. Adiabatic temperature profile, Liquidus: black dashed line, Solidus: grey dashed line (Lebrun et al. 2013)

- **Calculation of the surface Temperature**

$$\left(c_{p,\mathrm{H_2O}}M_{\mathrm{atm}}+c_{\mathrm{p,m}}\frac{4}{3}\pi\rho_{\mathrm{m}}(r_{\mathrm{p}}^3-\delta^3)\right)\frac{dT_{\mathrm{surf}}}{dt}=4\pi r_{\mathrm{p}}^2(q_{\mathrm{m}}-\mathfrak{F})$$

Heat flux from the atmosphere from LBL model

Volatile model

- H2O pressure is set by its solubility in the magma ocean (Papale. 1997)

$$p\,(Pa) = \left(\frac{F_{\mathrm{H_2O}}}{3.44\,\times\,10^{-8}}\right)^{1/0.74}$$

F_{H2O}: fraction of water in the liquid silicate melt

- H2O exchange between boundaries with 2 differential equaitons

$$\begin{aligned} M_{\mathrm{H_2O}}^{\mathrm{mo},t} &= M_{\mathrm{H_2O}}^{\mathrm{crystal}} + M_{\mathrm{H_2O}}^{\mathrm{liq}} + M_{\mathrm{H_2O}}^{\mathrm{atm}} \\ &= k_{\mathrm{H_2O}} F_{\mathrm{H_2O}} M^{\mathrm{crystal}} + F_{\mathrm{H_2O}} M^{\mathrm{liq}} \\ &\quad + \frac{4\pi R_{\mathrm{p}}^2}{g}\left(\frac{F_{\mathrm{H_2O}}}{3.44\,\times\,10^{-8}}\right)^{1/0.74} \end{aligned}$$

We calculated O exchange between boundaries in the same way for H2O

- Outgassing is parameterized similar to (Sandu & Kiefer 2012)

$$r_{outgass} = 4\pi r_p^2 \rho_m F_{melt}^{avg} f_{melt}^{avg} u_c x_d$$

F_{melt}^{avg} : volume-averaged mass fraction of water in the melt
f_{melt}^{avg} : volume-averaged mass fraction of the mantle
u_c : the mantle convection velocity
x_d :degassing efficiency

Initial condition

- Mantle FeO Abundance: 0.1~20 wt%
- Initial Water abundances : 0.1~1000 EO (Earth Ocean)
- $\dfrac{Fe^{3+}}{Fe_{total}} \approx 0$
- Comparison of the effect of degassing coefficient X_d = 0 and X_d =1

Both Model A and B were run for a total integration of 5 Gyr

Result

Result: Magma Ocean Solidification time

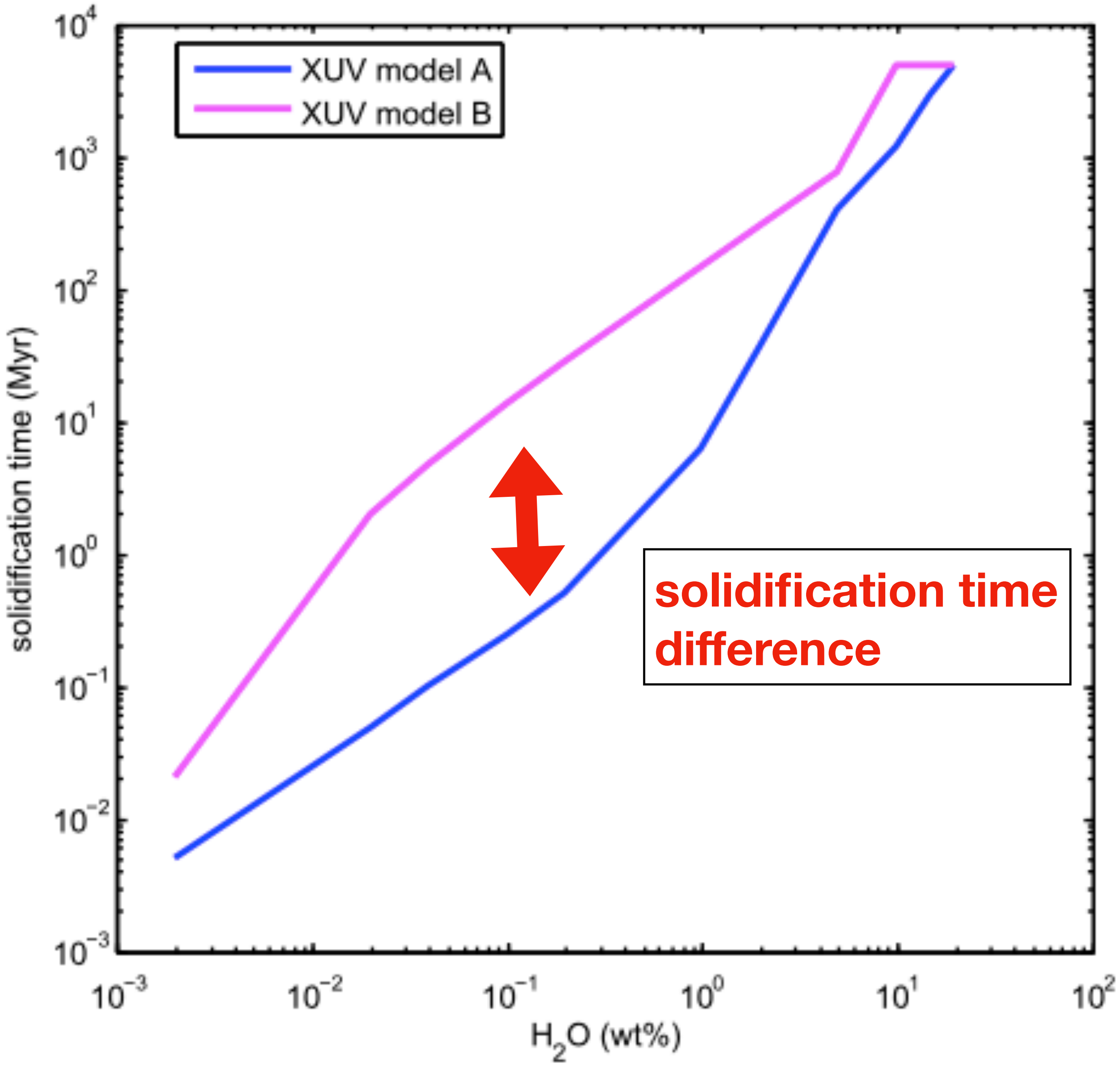
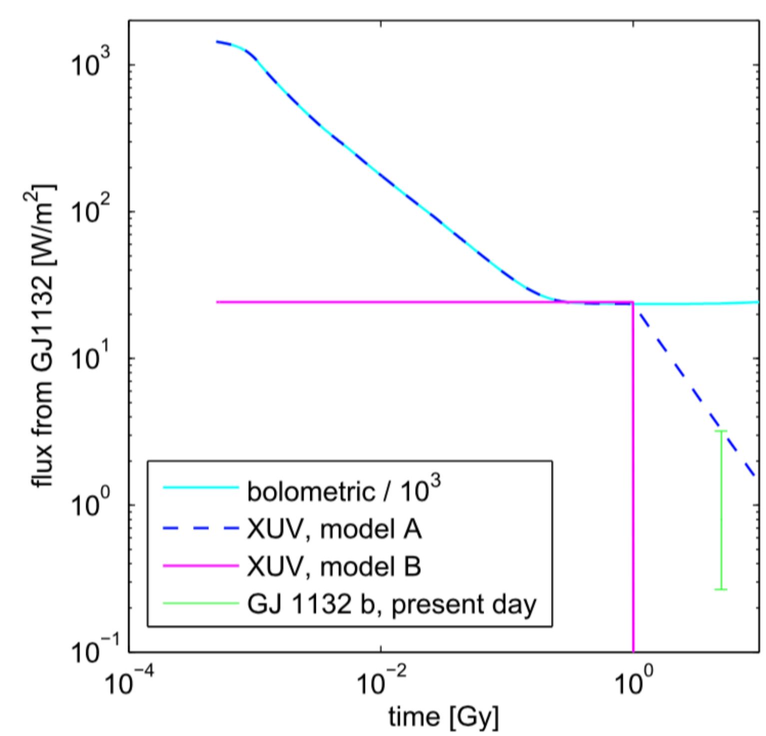
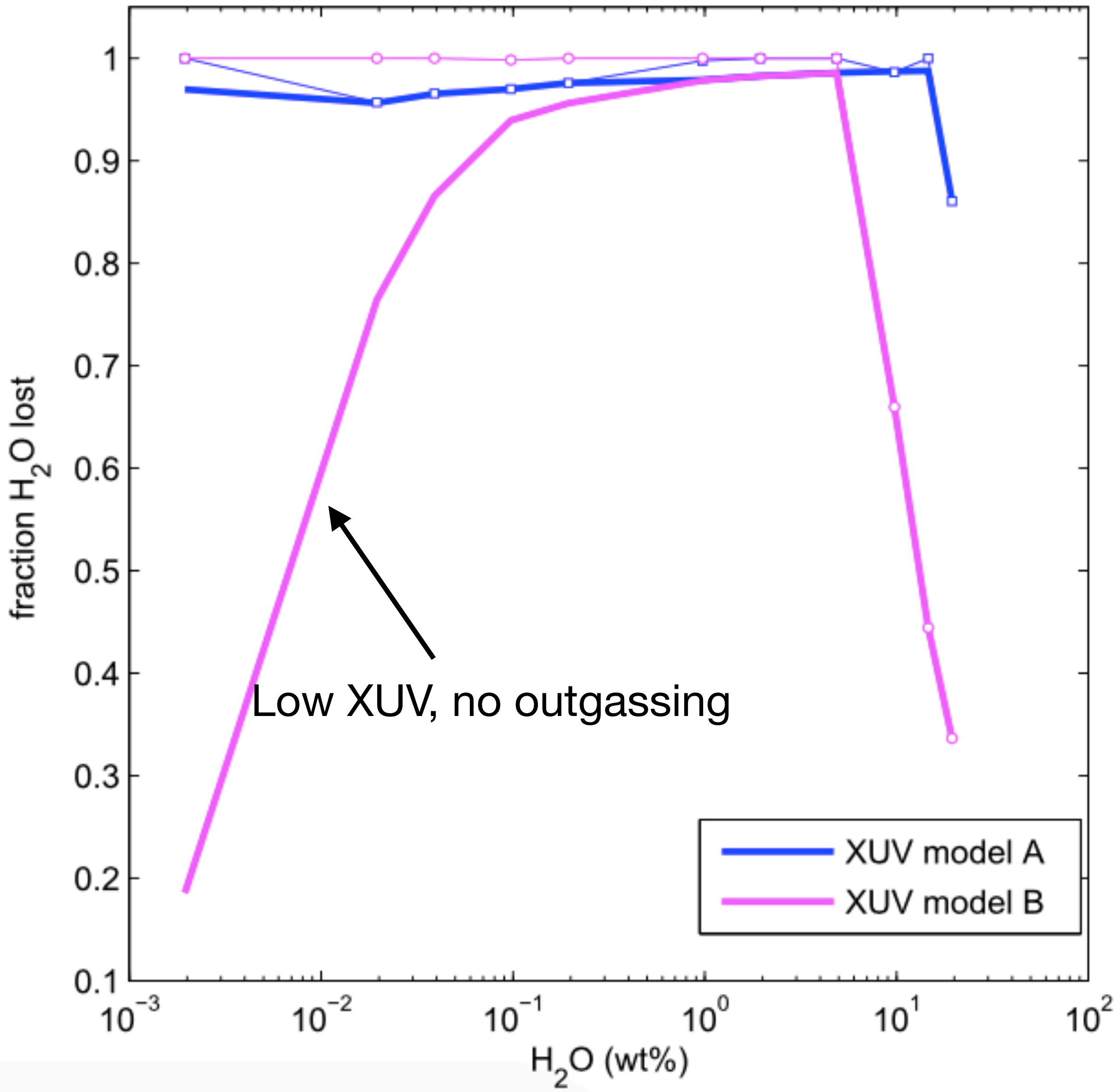


Fig. Solidification time for different initial water abundances



Stronger XUV of Model A enhances escape of H2O

Result: Water loss



Thick lines: $X_d = 0$ (**no outgassing**)
Thin lines : $X_d = 1$(**efficient outgassing**)

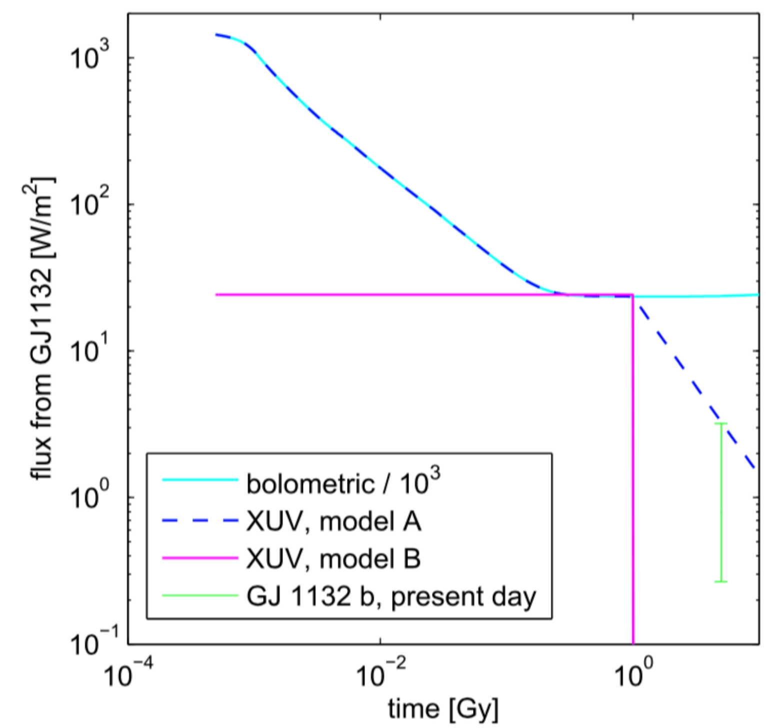


Fig. total water loss for different initial water abundances

Water remaining conditions

- High initial water abundances
- Low XUV and inefficient outgassing

Result: Atmospheric Oxygen buildup

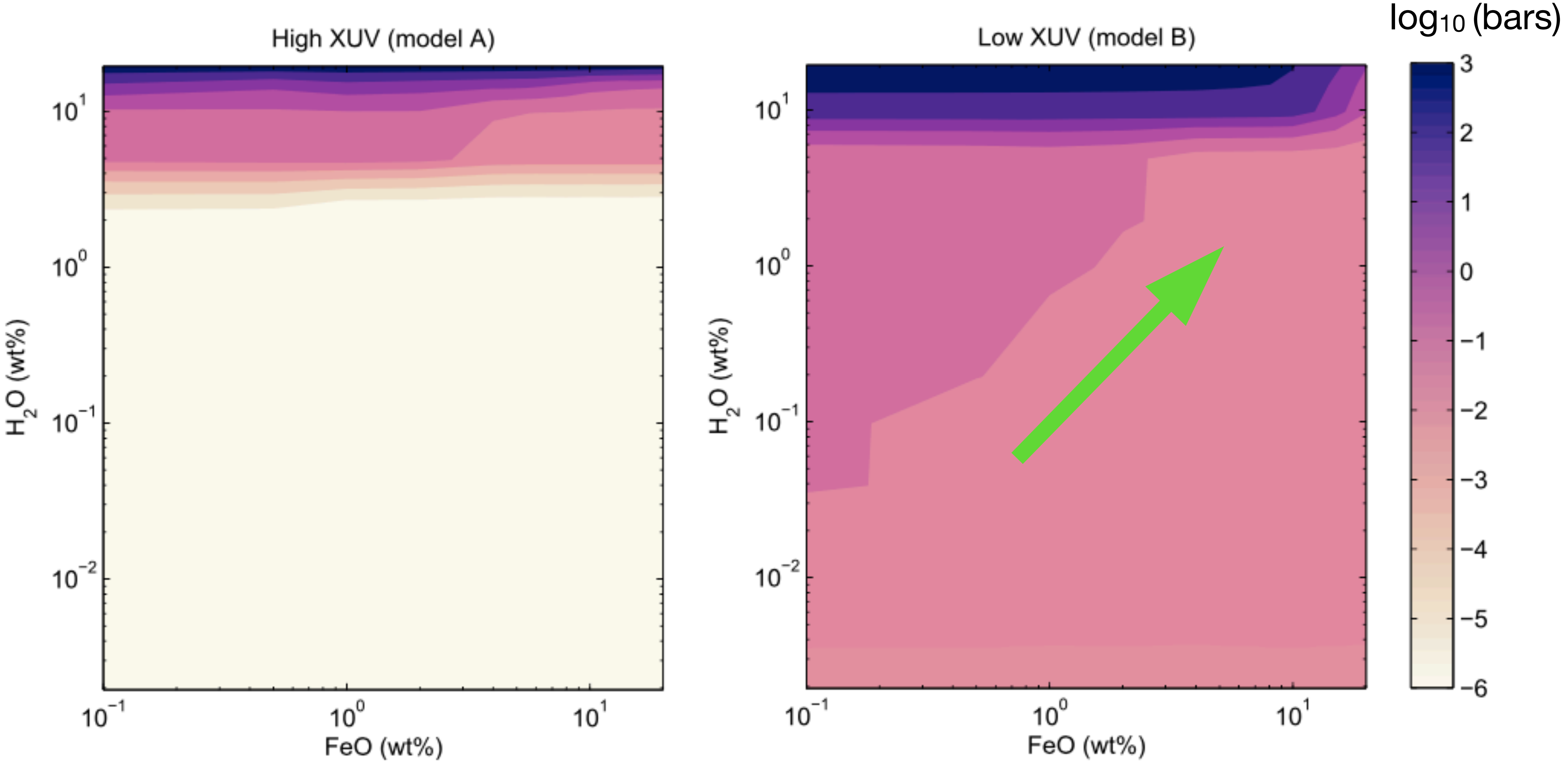


Fig. Final O_2 pressure as a function of initial H_2O and F_eO contents, $X_d = 0$

- For both XUV models, O_2 > H_2O vapor for nearly all of our parameter space

The atmosphere of GJ 1132b may be tenuous and dominated by O_2. If abundant water vapor is observed, both a low flux and high initial water is expected.

Result: Change of FeO to $FeO_{1.5}$ in the mantle

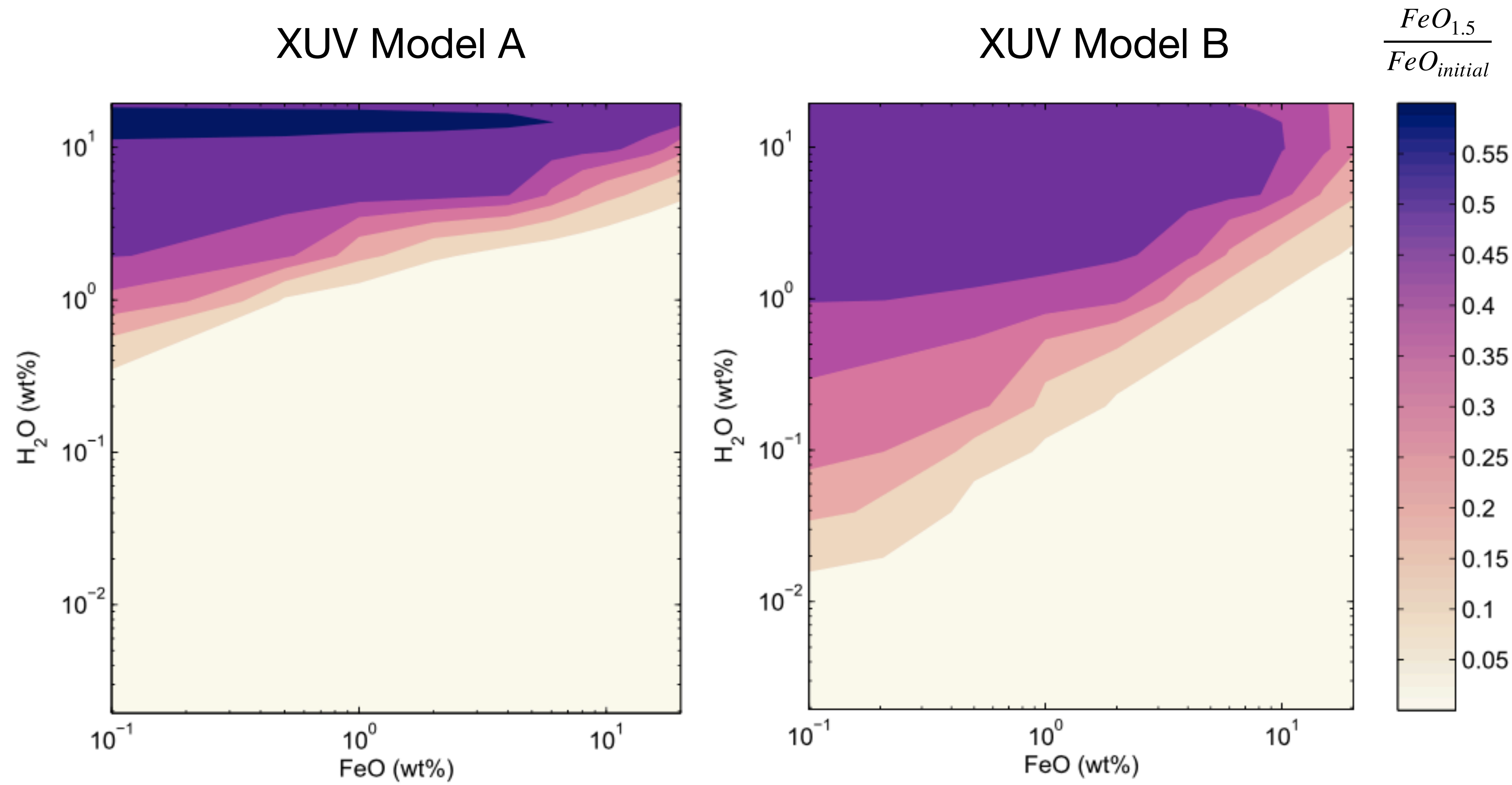


Fig. Mantle averaged ratio of $FeO_{1.5}$ to initial FeO as a function of initial H_2O and FeO, x_d =0

- At most, 10% of O2 is absorbed in magma ocean for model B

$$FeO + \frac{1}{4}O_2 \leftrightarrow FeO_{1.5}$$

In the case of Low XUV (model B), initial FeO abundances have stronger effect on O2

Discussion

Discussion: Effect of CO_2

- Greenhouse effect of CO_2

→ **Longer magma ocean lifetimes & Additional loss of both water vapor and O_2**

- CO_2 is soluble in metal alloy.

↓

It would be possible that later outgassing of CO_2 during post-magma ocean

→ **We need more a detailed model to evaluate**

Discussion: Initial H_2 envelope

It is possible that GJ 1132b began with an envelope dominated by H_2 gas

→ Interaction of an H2 and FeO may be expected

- Reduction of FeO → Fe metal through a reaction below

$$H_2(g) + FeO \leftrightarrow H_2O(g) + Fe$$

$$\longleftarrow$$

The forward reaction is thermodynamically unfavorable on the present and early Earth (Fukai 1984, Kuramoto & Matsui 1996)

Initial H2 envelope may prolong the magma ocean lifetimes and reduce the loss of water and O_2

Conclusion (Predictions of GJ 1132b)

- Predictions of the atmosphere

Tenuous atmosphere with a few bars of O_2
and little or no steam for most of starting conditions

- Requirement for water remaining

More than ~5wt% of initial water would be required to retain a substantial steam envelope

- Effect of interaction of magma ocean with a H_2O-rich atmosphere

Magma ocean absorbs at most 10% of the O_2
produced while 90% is lost to space

Implications of special cases

- O_2 abundances > 500bars
  & Significant steam >500 bars

Implicating

Presence of an earlier H_2-rich envelope

Low XUV flux over the system's lifetime
initial water abundances > 250EO

- The presence of a steam atmosphere

Implicating

Continued existence of a magma ocean at the surface

It was reported that transmission spectrum is best fit to the atmosphere with H2O volume mixing ratios of 1~10%
(Southworth et al. 2017)

Appendix

Appendix XUV assumption

NUV-FUV: Hubble COS and STIS (France et al. 2016)
EUV: model-dependent scaling from Lyα (Linsky et al. 2014)
X-ray: from a plasma model matched to an earlier XMM detection of flare from GJ1214
(Lalitha et al. 2014)

XUV flux(1-120nm) = 3×10^{-5} bolometric flux

XUV flux(120-130 nm) = 3×10^{-5} bolometric flux contributed by Lyα

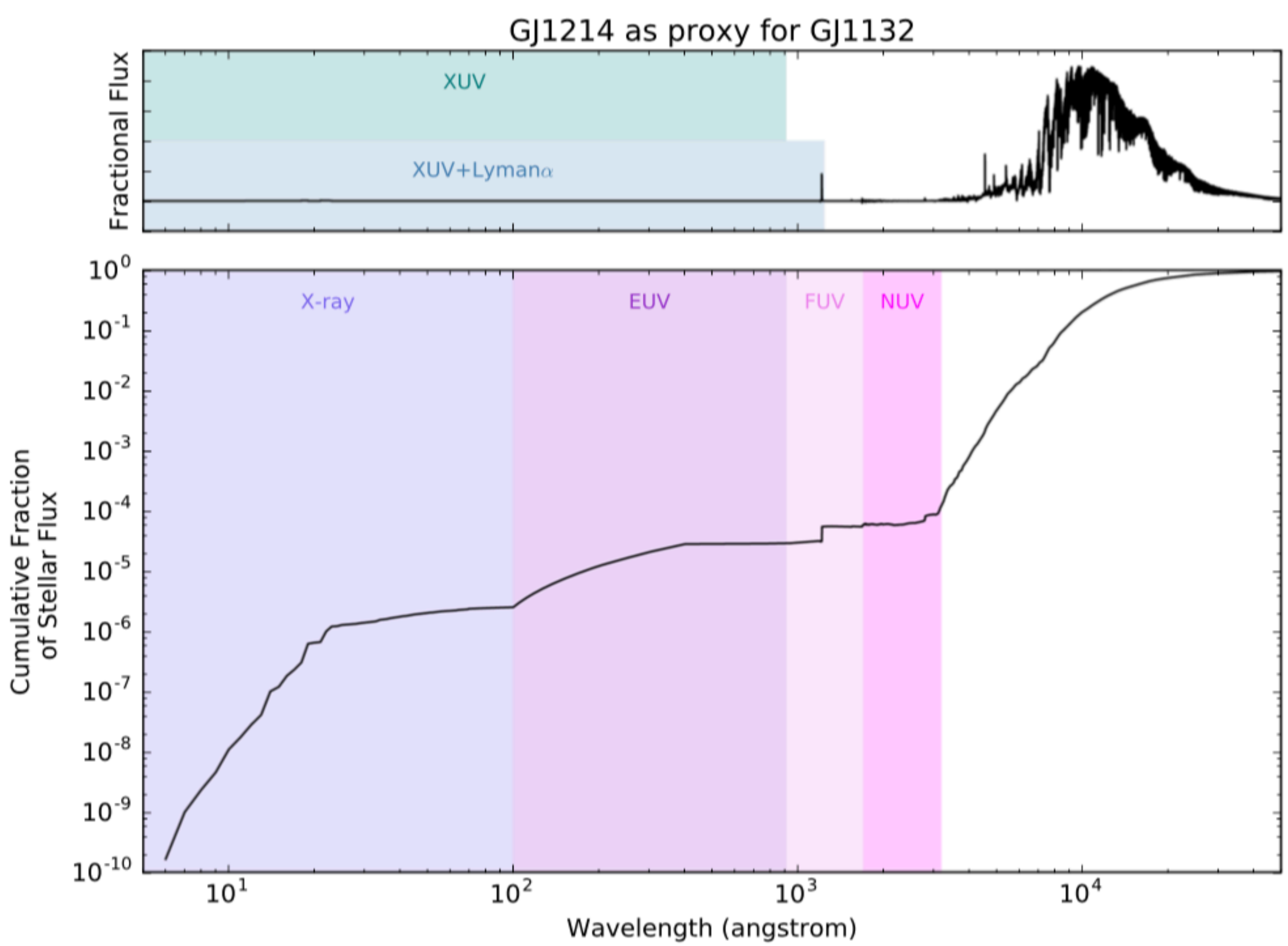


Fig. XUV flux assumption

Time evolution

$L_{XUV} = 10^{-3} L_{bol}$ Saturation

Model A:
XUV flux is 10^{-3} time-evolving stellar luminosity and after 1 Gyr, declining as a power of law with ε=0.3

Model B:
XUV flux is 10^{-3} present-day stellar luminosity and 0 after 1Gyr, with ε=0.15

Appendix

Thermal model

- Mantle heat flux is parameterized by the mantle Rayleigh number

$$q_m = \frac{k(T_p - T_{surf})}{l}(\frac{Ra}{Ra_{cr}})^{\beta}$$

$$Ra = \frac{\alpha g(T_p - T_{surf})l^3}{\kappa \nu}$$

k: thermal conductivity
α: thermal expansion coefficient
κ: thermal conductivity
ν: kinematic viscosity

- Melt fraction ψ (Lebrun et al. 2013)

$$\psi = \frac{T_p - T_{solidus}}{T_{liquidus} - T_{solidus}}$$

The viscosity depends on the melt fraction ψ.
Below ψ_c ~0.4, viscosity becomes solid-like.

Solid-like viscosity η

$$\eta = \eta_0 \exp(-E_a/(RT))$$

Appendix

Line by Line model

To calculate out-going-radiation

- Monochromatic equation

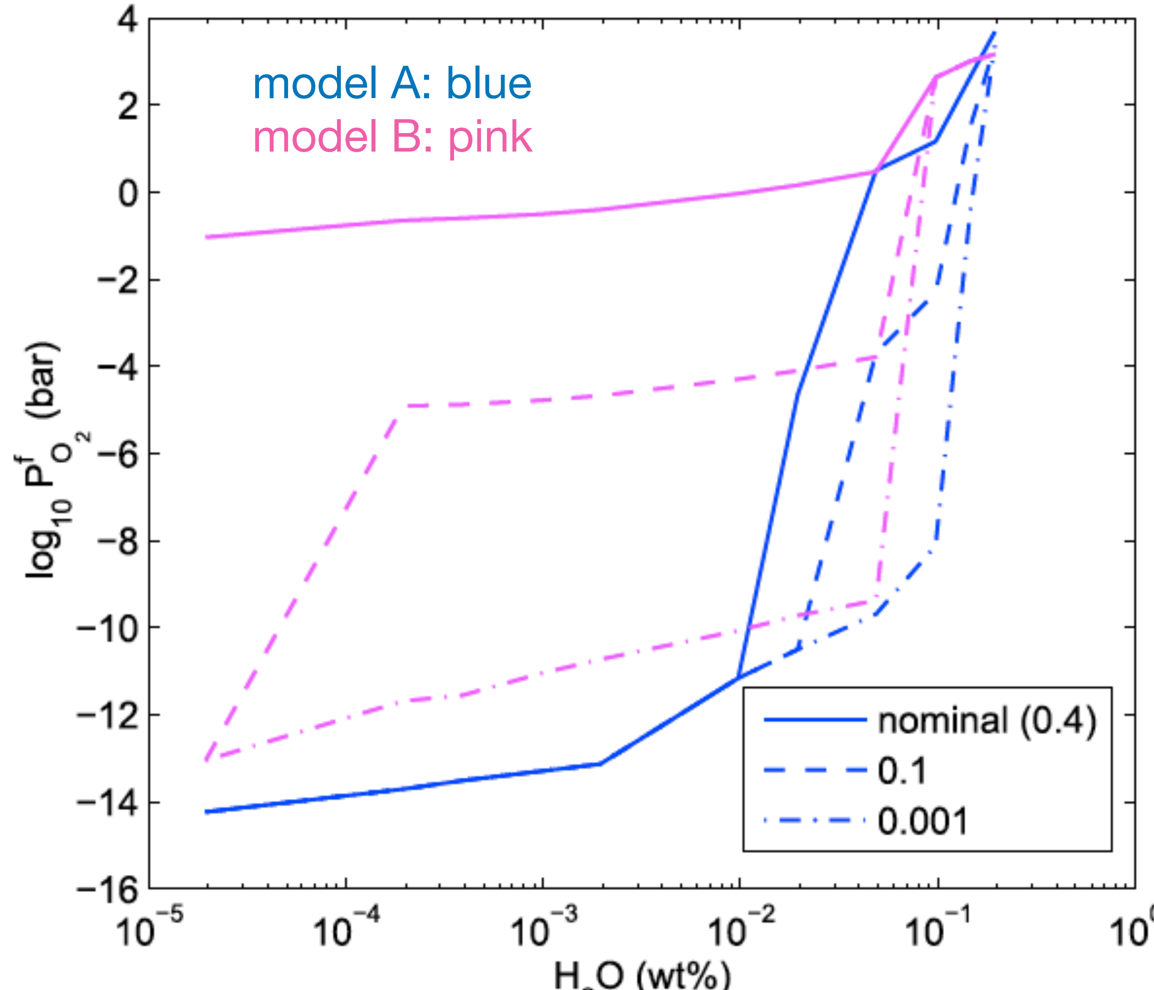
$$F_+(\bar{\tau}_\infty) = \pi B_\nu(T_{surf})e^{-\bar{\tau}_\infty} + \pi \int_0^{\bar{\tau}_\infty} B_\nu(\bar{\tau})e^{\bar{\tau}-\bar{\tau}_\infty}\mathrm{d}\bar{\tau}$$

Conditions

- 30 layer up to a minimum atmospheric pressure of 1 Pa
- Spectral calculations; 1 cm^{-1} ~ five times Wien peak number of the Planck function at the given surface Temperature
- sensitivity; 5000 points in wavenumber
- Temperature assumption; a dry adiabat from the surface to the tropopause, after which a stratospheric temperature = skin temperature of GJ 1132b (Albedo = 0.75)
- MT-CKD parametrization for far-wing absorption of strong H2O line

Appendix

Discussion: Transition point of Escape model



FeO: 8 wt%
X_d : 1

Fig. Sensitivity of the final O2 pressure to the transition point of 3 different X_H

Energy-limited hydrodynamic Escape where $P_{O^2} < P_{H_2O}$

$P_{O^2} = P_{H_2O}$:Transition point

Diffusion-limited Escape: H diffuse through O background with no more O escape

Reducing the transition abundances → more tenuous O2