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2 **Supplementary Information for**

3 **Narrow range of early habitable Venus scenarios permitted by modelling of oxygen loss and**
4 **radiogenic argon degassing**

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8 **This PDF file includes:**

- 9 Supplementary text
- 10 Figs. S1 to S10
- 11 Tables S1 to S2
- 12 SI References

Supporting Information Text

Oxygen loss model

Crustal production upper limit. For an Earth-like internal heating rate and a 500 K temperature drop between the melting region and the surface the maximum crustal production rate is $\sim 175 \text{ km}^3 \text{ yr}^{-1}$ (1). Volcanism at this rate over 4 Gyr would produce $\sim 1500 \text{ km}^3$ of crust (not accounting for changing heating rate over time or mantle cooling), implying substantial crustal recycling, and eruption of 80% of Venus' mantle. For a melt fraction of 15%, at the upper end of terrestrial Mid Ocean Ridge melting, this crustal production rate would imply melt processing of Venus' mantle volume 5 times. Crustal recycling may aid Venus in retaining its water (2). Parameterized stagnant lid convection models constrained by ^{40}Ar measurements suggest total crustal production of a global layer $< 65 \text{ km}$, and melt processing of up to 1.2 Venus mantle volumes (3). However, 2D simulations (1) of Venus' thermal evolution including episodic lid convection are able to produce up to 1.2 mantle volumes of crust over Venus' lifetime (implying crustal recycling).

Oxidation of lava flows. Experimental work on oxidising basaltic lava flows (4–6) suggests that exposed olivine, pyroxene, and basaltic glasses should oxidise under present-day Venus conditions within weeks to months, and pyroxene on timescales of weeks to decades, or even faster. However, these timescales apply only to materials exposed directly to the atmosphere. The formation of a surface coating of oxidized minerals may prevent thick oxidation rinds from forming in Venus basalts (4). To model oxidation of Venusian lava flows, we assume that basaltic lava flows on Venus have a typical thickness of $\sim 1 \text{ m}$ (7), resembling terrestrial lavas. The depth of oxidation is controlled by the diffusivity of cations and oxygen in basalt (8). These diffusivities are much higher in melts than in solid basalt (8–10). Eruption rates on present day Venus are estimated to be of order $0.1 - 0.2 \text{ km}^3 \text{ yr}^{-1}$. Taking an average lava flow thickness of 1 m , this corresponds to 1 m of lava being erupted every $\sim 4 \text{ Myr}$. Even under present-day Venus surface temperatures, lavas are expected to chill crusts 4 m thick in 2 months (7). Using the approximation:

$$\Delta z \approx \sqrt{D\Delta t} \quad [1]$$

thermal diffusivity in the lava flow is $10^{-2} \text{ cm}^2 \text{ s}^{-1}$, 4 orders of magnitude greater than the diffusivity of oxygen in basaltic melts ($D \approx 10^{-6} \text{ cm}^2 \text{ s}^{-1}$) (9, 10), suggesting that oxidation of lava before solidification is limited to $< 1 \text{ mm}$. However, slower oxidation of solidified lava flows may still proceed, slowing down as the lava flow cools. For a lava flow exposed to the surface at 700 K for 40,000 years (1 timestep), the diffusive length scale is $\sim 0.1 \text{ cm}$. Also taking into account a) faster diffusion while the lava flow is still hot, and b) surface undulations and cooling cracks that increase the surface area in contact with the atmosphere.

Runaway Greenhouse Surface Temperature. We find surface temperature at each timestep, given the incoming solar flux relative to Earth's insolation (S_{eff}) and partial pressure of H_2O in the atmosphere in bars ($p_{\text{H}_2\text{O}}$), using a curve fit to existing numerical results (11):

$$\log_{10}(T_{surf}(x, y, z)) = c_1 + c_2x + c_3y + c_4z + c_5x^2 + c_6xy + c_8z^2 + c_9y^3 + c_{10}z^3$$

where:

$$x = (\log_{10}(P_{\text{H}_2\text{O}}) - k_1) / k_2$$

$$y = (\log_{10}(g) - k_3) / k_4$$

$$z = (\log_{10}(S_{eff}) - k_5) / k_6$$

Coefficient values for c_{1-10} and k_{1-6} are given in Table S1.

Drag of heavy species during H escape. If H escape to space is sufficiently rapid, the H can drag heavier species (such as O) along with it (12). To determine whether this process is active at a given timestep in our model, we calculate the critical mass (m_c) of the heaviest molecule escaping H can drag along with it in a background atmosphere of mass m_{atm} (12):

$$m_c = \frac{X_H + q^2 X_{atm}}{X_H + q X_{atm}} m_H + \frac{kT \dot{M}_{H,ref}}{m_H b g (X_H + q X_{atm})} \quad [2]$$

where $q = m_{atm} / m_H$. In our model runs, we only see $m_c \geq m_O$ at the very beginning of models with $> 700 \text{ m GEL}$ H_2O with hypothetical habitable eras ending at 4 Ga and $f_{volc} \geq 0.9$, i.e. H escape is not limited by diffusion through the background CO_2 atmosphere. Under these conditions there is more than an order of magnitude more O in the background atmosphere than CO_2 , and we approximate the atmosphere as a mixture of H and O only to find the resulting H and O escape rates (12):

$$\dot{M}_{O,esc} = \frac{\eta q}{1 + \eta q} \dot{M}_{H,ref} \quad [3]$$

$$\dot{M}_{H,esc} = \frac{1}{1 + \eta q} \dot{M}_{H,ref} \quad [4]$$

where:

$$\eta = \frac{Y - 1}{Y X_H / X_O + q} \quad [5]$$

$$Y = \frac{kT \dot{M}_{H,ref}}{(q - 1) b g X_H m_H^2} \quad [6]$$

Explosive Volcanism. Although high atmospheric pressure disfavors explosive volcanism on Venus today, dissolved magmatic volatiles drive explosive volcanism at pressures greater than Venus' atmospheric pressure in subaqueous environments, such as MORs (13, 14). Several authors report observations of evidence for explosive volcanism on Venus, invoking volatile concentrations of order 1-3 wt% CO₂, SO₂ and H₂O (15, 16). However, CO₂ as the primary magmatic volatile phase is unlikely to drive explosive volcanism on present day Venus because concentrations above 6.5 wt% would be required (17). The conditions required for explosive volcanism are sensitive not only to atmospheric pressure but also to melt ascent rates, magma temperatures, melt viscosity, and vent geometry (17, 18). We choose to implement the simplest possible threshold for the effusive-explosive transition, allowing explosive volcanism to occur when the volume fraction of gas (F_{gas}) exceeds ~75% (19, 20). We find the volume of gas per m³ melt delivered to the surface (V_{gas}) using the ideal gas law:

$$V_{gas} = \frac{n_{degas} R T_{melt}}{P_{atm}} \quad [7]$$

where n_{degas} is the number of moles of degassed CO₂ and H₂O. The corresponding volume fraction of gas is:

$$F_{gas} = \frac{V_{gas}}{V_{gas} + 1} \quad [8]$$

For $F_{gas} \geq 0.75$, the mass of O₂ removed from the atmosphere in a given timestep ($\dot{M}_{O_2,ox}$) is:

$$\dot{M}_{O_2,ox} = M_{O_2,atm} - m_{O_2} \frac{0.25 \rho_{FeO} f_{ex} z_{erupt}(t)}{m_{FeO}} \quad [9]$$

Radiogenic Ar Degassing Model

The initial concentration of ⁴⁰K in Venus' mantle is given by (3, 21):

$$[^{40}\text{K}(0)] = r_{(^{40}\text{K}/\text{K})} r_{(\text{K}/\text{U})} c_{(\text{U},M)} \exp(\lambda_{tot} t) \quad [10]$$

where $r_{(\text{K}/\text{U})}$ is the ratio of K to U for Venus' crust as measured by spacecraft (21), $r_{(^{40}\text{K}/\text{K})} = 1.165 \times 10^{-4}$ is the present ⁴⁰K/K ratio (21), $c_{(\text{U},M)}$ is the primordial mantle U concentration based on values calculated for the Earth (SI Appendix: Table S2) (3), where t is present time after Venus' formation (4.5 Gyr), and λ_{tot} is the decay constant for ⁴⁰K:

$$\lambda_{tot} = \lambda_{Ar} + \lambda_{Ca} \quad [11]$$

Where $\lambda_{Ar} = 0.0581 \text{ Gyr}^{-1}$ and $\lambda_{Ca} = 0.4962 \text{ Gyr}^{-1}$ are the decay constants for decay of ⁴⁰K to ⁴⁰Ar and ⁴⁰Ca respectively (21). Our models begin at the end of a hypothetical habitable era. We calculate the starting mantle ⁴⁰K concentration:

$$[^{40}\text{K}(t_0)] = [^{40}\text{K}(0)] \exp(-\lambda_{tot}(4.5 - t_0)) \quad [12]$$

starting mantle ⁴⁰Ar concentration:

$$[^{40}\text{Ar}_M(t_0)] = (1 - f_{pre}) [^{40}\text{K}(0)] (1 - \exp(-\lambda_{tot}(4.5 - t_0))) \quad [13]$$

and starting atmospheric ⁴⁰Ar abundance:

$$^{40}\text{Ar}(t_0)_a = f_{pre} V_M \rho_M [^{40}\text{K}(0)] (1 - \exp(-\lambda_{tot}(4.5 - t_0))) \quad [14]$$

where f_{pre} is the fraction of radiogenic ⁴⁰Ar degassed prior to the end of the habitable era, $\rho_M = 3550 \text{ kg m}^{-3}$ is the density of the mantle (22), and V_M is the mantle volume, taking Venus' core radius to be 3110 km (23).

Next, we evolve the mantle and atmosphere ⁴⁰K and ⁴⁰Ar inventories forwards in time (SI Appendix: Fig S4) until the present day based on the parameters in each of our O₂ loss models, calculating the crustal production rate using Equations ??-??. We assume that crustal thickness does not evolve over time, and instead consider a range of fixed thicknesses.

The concentration of ⁴⁰K in Venus' mantle at each new timestep ($[^{40}\text{K}]_{M,i+1}$) is given by:

$$\begin{aligned}
[^{40}\text{K}]_{M,i+1} &= [^{40}\text{K}]_{M,i} - [^{40}\text{K}]_{M,i} \exp(-\Delta t \lambda_K) \\
&\quad - \frac{V_{new,i} \rho_c}{V_M \rho_M} [^{40}\text{K}]_{m,i} \\
&\quad + \frac{V_{rec,i} \rho_c}{V_M \rho_M} \int_{D_c+z_{rec}}^{D_c} [^{40}\text{K}(z)]_{c,i} dz
\end{aligned} \tag{15}$$

where $\rho_c = 2800 \text{ kg m}^{-3}$ is the density of the crust, and D_c is the crustal thickness. $V_{new,i} = z_{new} A_{Venus}$ is the volume of new crust produced in one timestep.

V_{rec} is the volume of crust recycled back into the mantle at a given timestep, which is given by the difference between the total crustal thickness at timestep i and the fixed maximum crustal thickness, D_c . We track the concentration of ^{40}K throughout the crust with depth in 1 m intervals to calculate the amount of ^{40}K recycled back into Venus' mantle over time. We also consider a case in which recycled crust is not remixed into the mantle, and instead sinks into a separate mantle reservoir, in which case we set the final term of Equation 15 to 0. This depletes the melting mantle of ^{40}K over time, and permits a wider range of crustal production histories for Venus that can still fall within or below the measured range for Venus' atmospheric ^{40}Ar abundance.

New crust is produced by partial melting of the mantle. This sets the initial ^{40}K concentration of the crust. We consider melt fractions of $\Phi = 0.025$ and 0.05 (SI Appendix: Table S2) (3). During partial melting, ^{40}K and ^{40}Ar preferentially partition into the melt, enriching the melt in these species. The concentration of ^{40}K in new crust (which is also the concentration of ^{40}K in mantle melts) formed through fractional melting of the mantle is:

$$[K]_{c,i} = [K]_{m,i} = [K]_{M,i} \frac{1}{\phi} [1 - (1 - \phi)^{1/D_K}] \tag{16}$$

where D_K is the partition coefficient of K in the melt, which is very small (24), so we make the approximation $(1 - \phi)^{1/D} \approx 0$ (3, 21). In our model, we assume that ^{40}Ar in the melt is released instantaneously to the atmosphere (3). The mass of ^{40}Ar released to the atmosphere at each timestep by partial melting and crustal production is:

$$^{40}\text{Ar}_{M \rightarrow a,i} = V_{new,i} \rho_c [\text{Ar}]_M \frac{1}{\phi} [1 - (1 - \phi)^{1/D_{\text{Ar}}}] \tag{17}$$

where D_{Ar} is the partition coefficient of Ar in the melt, approximated in the same way as for ^{40}K (3, 21). The concentration of ^{40}K in the crust at each timestep is given by:

$$[^{40}\text{K}(z)]_{c,i} = [^{40}\text{K}]_{c,i} - [^{40}\text{K}]_{c,i} \exp(-\Delta t \lambda_K) \tag{18}$$

The corresponding release of ^{40}Ar to the atmosphere by ^{40}K decay in the crust and rapid diffusion of ^{40}Ar to the surface (3) is given by:

$$^{40}\text{Ar}_{c \rightarrow a,i} = \rho_c A_{Venus} \frac{\lambda_{\text{Ar}}}{\lambda_K} \exp(-\Delta t \lambda_K) \int_{D_c}^0 [^{40}\text{K}(z)]_{c,i+1} dz \tag{19}$$

Therefore, the total ^{40}Ar in the atmosphere at each timestep is:

$$^{40}\text{Ar}_{a,i+1} = ^{40}\text{Ar}_{a,i} + ^{40}\text{Ar}_{c \rightarrow a} + ^{40}\text{Ar}_{M \rightarrow a,i} \tag{20}$$

Habitable Era Water Recycling Model

We assume Venus produced enough crust to degas 80 bar of CO_2 prior to the end of the habitable era to maximize the potential for extensive recycling of hydrated crust into the mantle by increasing the rate of crustal burial. Heat transport in a stagnant lid regime with downwards advection of crust caused by continuous resurfacing by lava flows with diffusion of heat and internal radiogenic heat production, can be expressed as (25, 26):

$$\alpha \frac{\partial^2 T}{\partial z^2} = v \frac{\partial T}{\partial z} + \frac{H}{\rho c_p} \tag{21}$$

where $\alpha = \frac{k}{\rho c_p}$ is the diffusion coefficient, k is conductivity, ρ is density, and c_p is specific heat capacity, and H is internal heat production by decay of radioactive elements. The steady state solution is (26):

$$T(z) = \frac{z_{lith}}{v} \frac{H}{\rho c_p} + c_1 \frac{\alpha}{v} \exp \frac{v z_{lith}}{\alpha} + c_2 \tag{22}$$

where c_1 and c_2 are constants:

$$c_1 = \frac{v}{\alpha} \frac{\left(\Delta T - \frac{\rho c_p H z_{lith}}{v} \right)}{\exp \left(\frac{v z_{lith}}{\alpha} \right) - 1} \tag{23}$$

$$c_2 = T_s - c_1 \frac{\alpha}{v} \quad [24]$$

where $\Delta T = T(z_{lith}) - T(0)$.

The advection velocity v depends on the timing of the end of the habitable era (t_0 from our O₂ loss model), the proportion of Venus' modern CO₂ degassed prior to the end of the habitable era ($1 - f_{volc}$), and melt CO₂ concentration. We assume that the eruption rate (and therefore v) is constant over the period 4.4 Ga to t_0 , such that:

$$v = \frac{1}{3.15 \times 10^{16} (4.4 - t_0)} \frac{(1 - f_{volc})}{c_{CO_2} \rho_{melt}} M_{CO_2, f} \quad [25]$$

where the factor of 3.15×10^{16} converts from Gyrs to seconds.

We compare our calculated geotherms to the phase boundaries for serpentinite dehydration and eclogite formation (27, 28) (SI Appendix, Fig S5) to find combinations of lithosphere thickness and volumetric heat production rate that permit eclogite to form before serpentinite dehydration occurs. When the temperature-pressure profile crosses into the serpentinite dehydration field first, we assume that water returns to the surface or shallower depths in the crust, as suggested by the upward migration of water released by serpentinization during subduction on Earth. Additionally, we omit all geotherms where temperatures exceed 1573 K at depths shallower than z_{lith} , as the base of the lithosphere is defined by this temperature in our model. We find that eclogite forms before dehydration occurs when v and z_{lith} are large, and H is small (SI Appendix, Fig S5).

Eclogite formation before dehydration provides a mechanism for recycling of H₂O into the mantle. We consider the implications of removal of water to the mantle on the evolution of the total surface water inventory of a hypothetical habitable era.

We initiate our models with an initial surface water inventory h_0 of up to 4000 m GEL, at the beginning of a hypothetical habitable era. At each timestep, we calculate the water removed from the surface inventory, and the water added by volcanic degassing. We assume that any hydrated crust is removed due to burial by new lava being erupted at the surface, so the crust moves downward at the advection velocity v (Equation 25). We use the same volcanic degassing procedure as for our initial O₂ loss models (Main Text Methods: C. Degassing), and the same range of melt H₂O concentrations (Main Text Table 1). However, we fix atmospheric pressure to <10 bar, based on climate models (29), rather than tracking accumulation of atmospheric CO₂ as climate models for early Venus indicate that low atmospheric pressures are required to sustain surface liquid water (30). The total GEL of water present at each timestep i is then:

$$\begin{aligned} h_i &= h_{i-1} \\ &- \frac{1}{\rho_w} \Delta t f_{hyd} A_h f_{ext} v \rho_c \\ &+ \frac{1}{\rho_w} \Delta t f_{ext} v \rho_c (c_{H_2O} - s_{H_2O}) \end{aligned} \quad [26]$$

where $\rho_w = 1000 \text{ kg m}^{-3}$ is the density of water, $\rho_c = 2800 \text{ kg m}^{-3}$ is the density of the crust, f_{hyd} is the mass fraction of water in hydrated crust (up to 3 wt% (31)), A_h is the fraction of Venus' surface area submerged by the existing surface water inventory, and s_{H_2O} is the solubility of water in the crust-producing melt as calculated using VolcGasses. To obtain A_h , we use modern Venus hypsometry (32) to calculate the global equivalent layer of water that can be stored beneath any given elevation on Venus, and the cumulative area of Venus' surface beneath each elevation. We then interpolate to find the cumulative area of Venus' surface submerged at each timestep in our model.

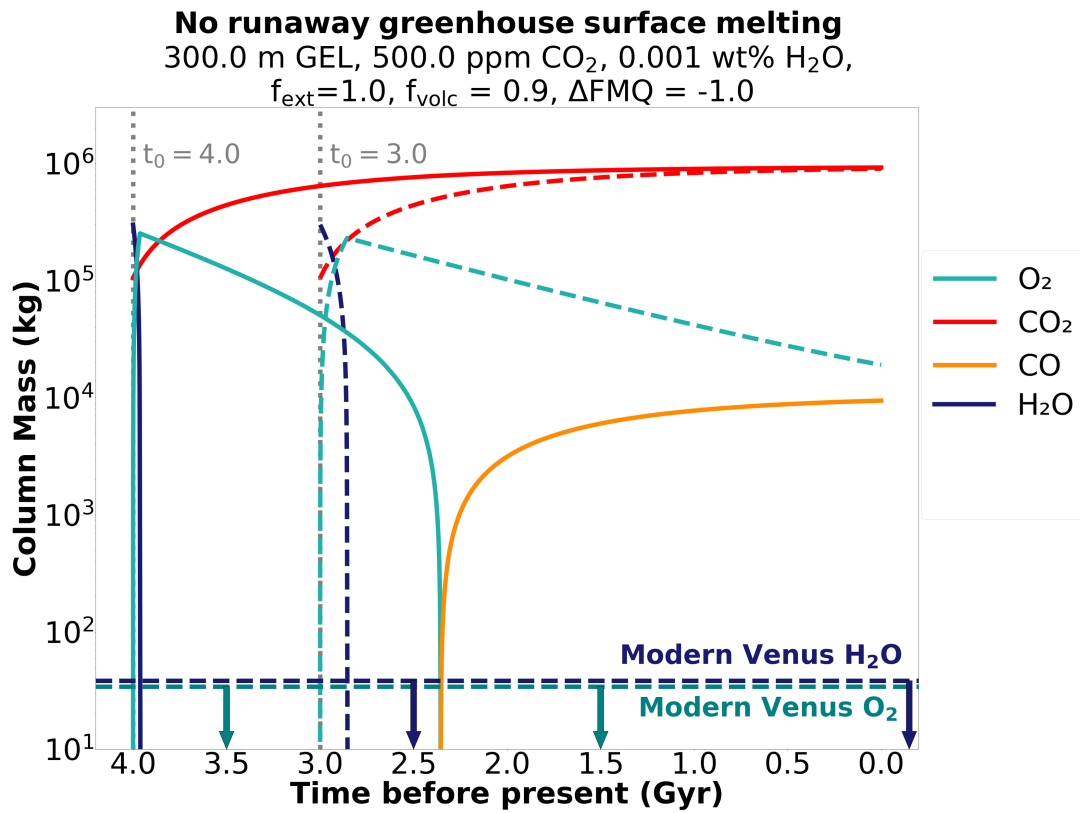


Fig. S1. Timing of the end of a hypothetical habitable era is an important parameter for determining whether the habitable era can be reconciled with modern Venus' low atmospheric O₂. Time evolution of column masses of H₂O, H₂, O₂, CO₂, CO, and CH₄ for two post habitable era evolutions, each with a 300 m GEL water inventory, average melt H₂O 0.001 wt%, average melt CO₂ 500 ppm, extrusive volcanism fraction $f_{ext} = 1.0$, fraction of Venus' modern CO₂ volcanically degassed since the end of a hypothetical habitable era $f_{volc} = 1.0$, and melt oxygen fugacity $\log f_{O_2} = FMQ-1$ in our baseline model (no runaway greenhouse surface melting). The solid line shows that a short-lived habitable era ending at 4.0 Ga can be reconciled with Venus' modern atmospheric O₂ (solid lines), whereas a habitable era ending at 3.0 Ga cannot (dashed line).

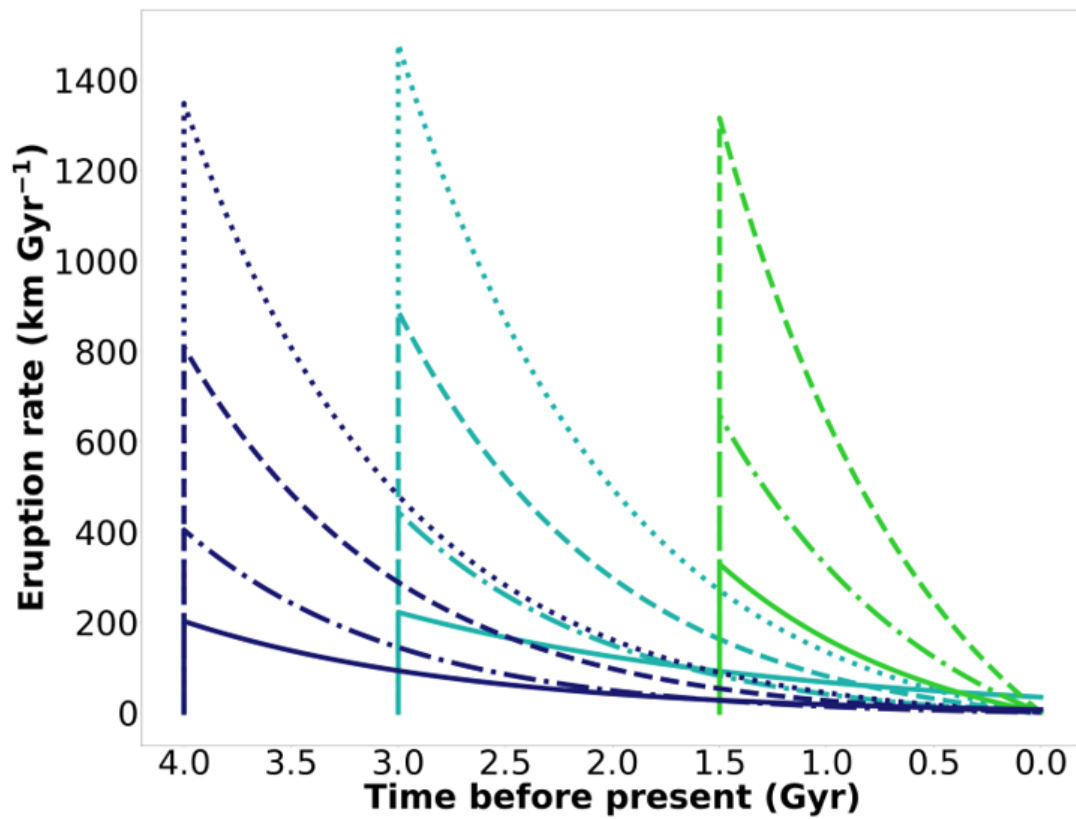


Fig. S2. Magmatism rates (including both intrusive and extrusive volcanism) as a function of time in our model for end habitable era times $t_0 = 4.0, 3.0$, and 1.5 Ga (dark blue, blue, and green) and average melt $\text{CO}_2 = 300, 500, 1000$, and 2000 ppm (solid, dashed, dashed-dotted, dotted) with $f_{volc} = 1.0$. There is no eruption rate line for $t_0 = 1.5$ Ga and $\text{CO}_2 = 300$ ppm because this would lead to a total erupted volume requiring processing of more than $1.2 \times$ Venus' mantle volume in 1.5 Ga in order to degas the full 90 bar of CO_2 in Venus' modern atmosphere. The eruption rates shown in these curves will be lower for smaller values of f_{volc} as less of the total CO_2 in Venus' modern atmosphere is derived from volcanic degassing in these cases.

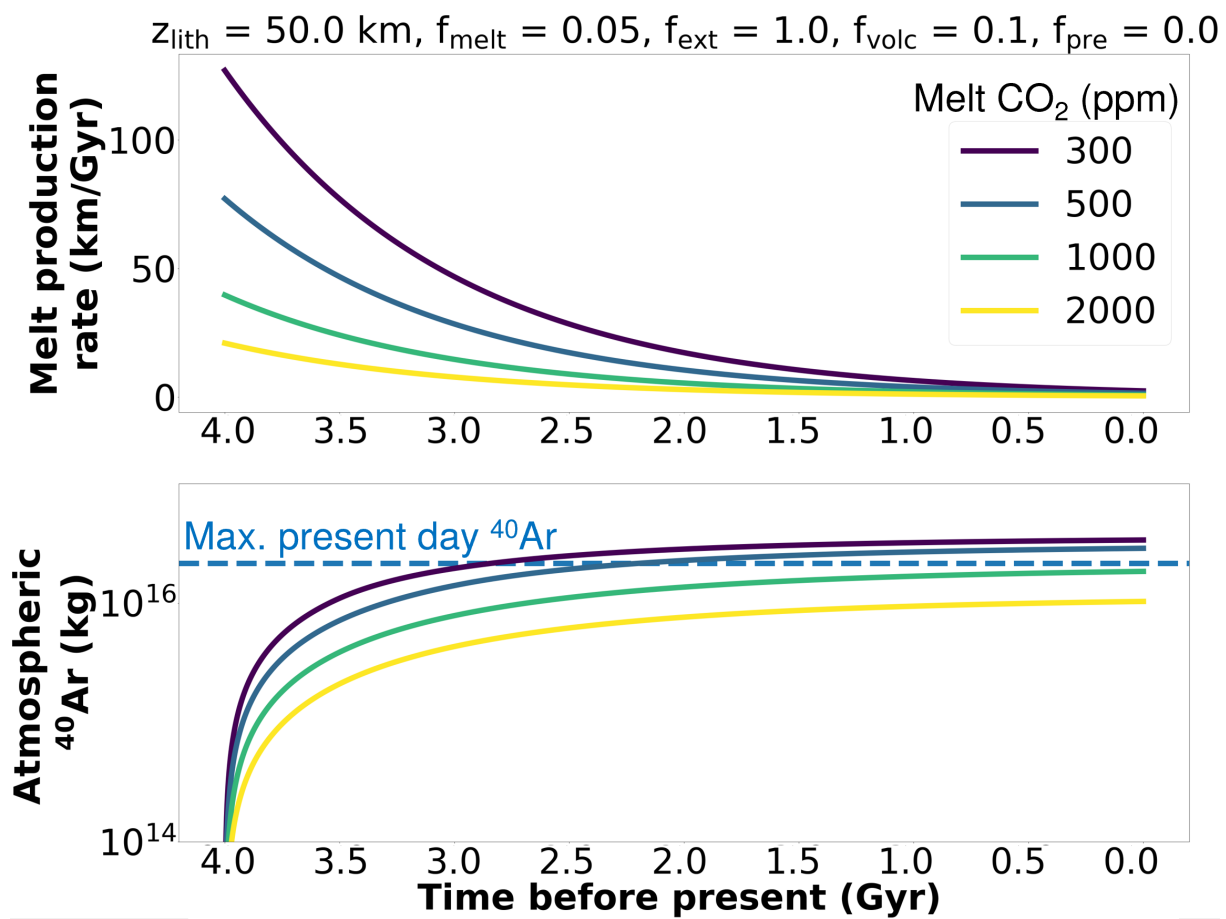


Fig. S3. Example atmospheric ^{40}Ar accumulation for a range of crustal production rates, set by melt CO_2 concentrations (as in Fig S2 and Equations 13-16 in the Main Text).

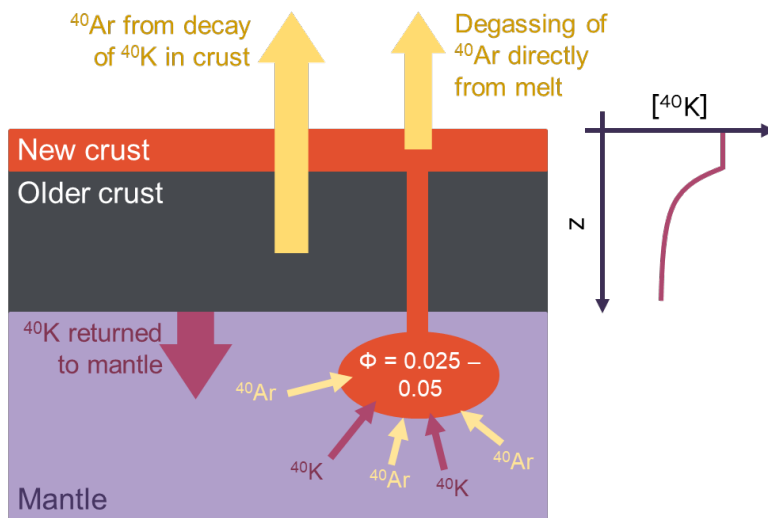


Fig. S4. Schematic illustration of our ^{40}Ar degassing model based on: (3, 21).

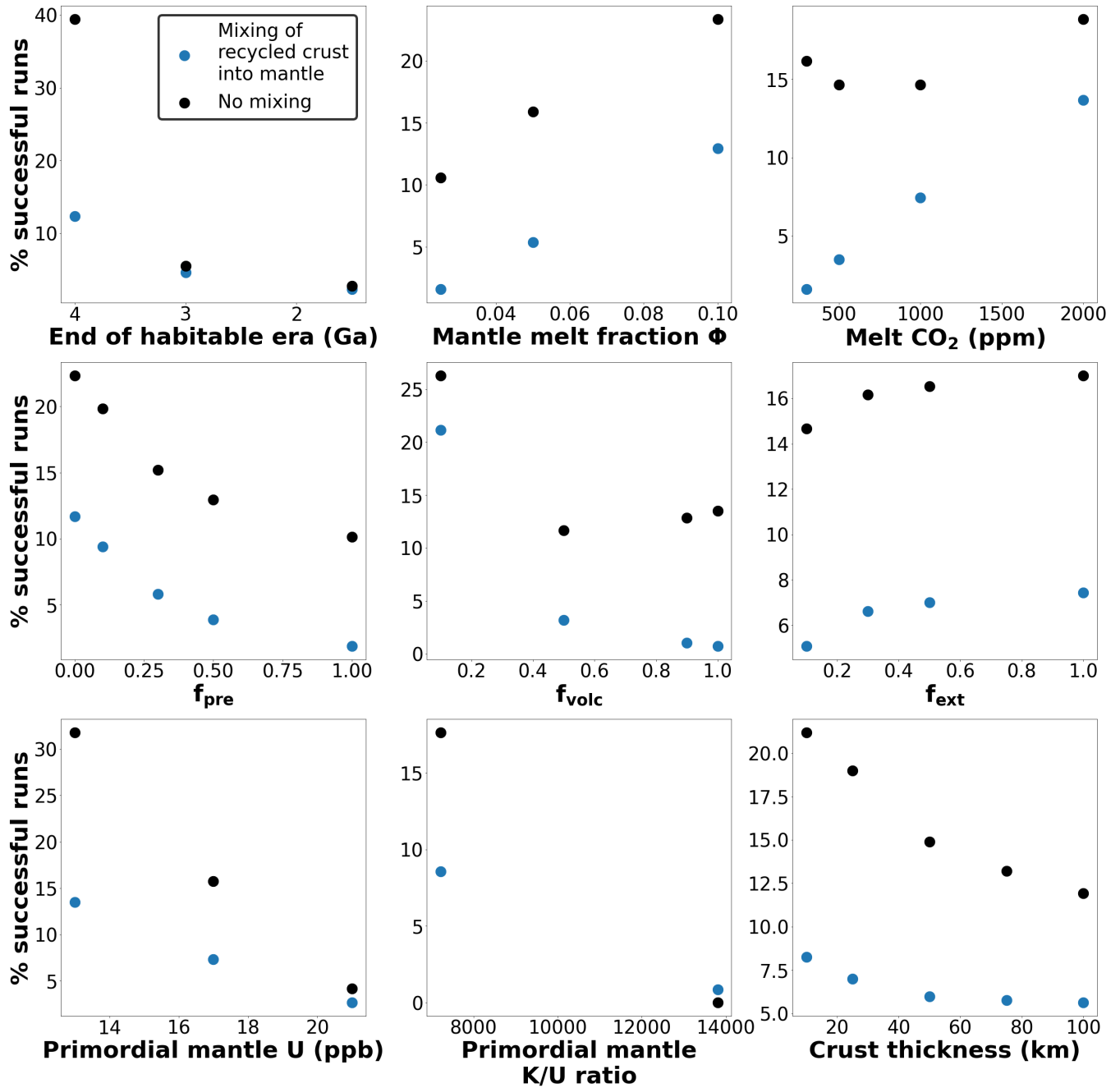


Fig. S5. Summary of ^{40}Ar degassing model results shown as the percentage of ^{40}Ar evolution models (e.g. Fig S3) that contribute less than $1.61 \pm 0.54 \times 10^{16} \text{ kg } ^{40}\text{Ar}$ (21) to Venus' atmosphere over the course of Venus' post-habitable era evolution. The most favorable parameters for successful Ar degassing model runs are anticorrelated with parameters that reconcile a hypothetical early habitable era with Venus' modern atmospheric O_2 . ^{40}Ar degassing models where ^{40}K crust is not re-mixed into the mantle after recycling at the base of the crust permit a wider range of parameter space to be compatible with modern ^{40}Ar in Venus' atmosphere. Blue circles correspond to results of 57,000 models where material recycled from the base of the crust is mixed uniformly throughout the mantle. Black circles correspond to results of 57,000 models where recycled crust becomes part of a separate mantle reservoir and does not alter the ^{40}K concentration of the mantle from which new crust is derived. f_{pre} is the fraction of mantle ^{40}Ar produced from the formation of Venus until the end of the hypothetical habitable era that is assumed to have already degassed and remained in the atmosphere. Melt CO_2 concentration, fraction of Venus' present day atmosphere contributed by post-habitable era volcanic degassing (f_{volc}), and extrusive volcanism fraction (f_{ext}) have the same values as in the O_2 loss models. Table S2 lists all model variables and their values.

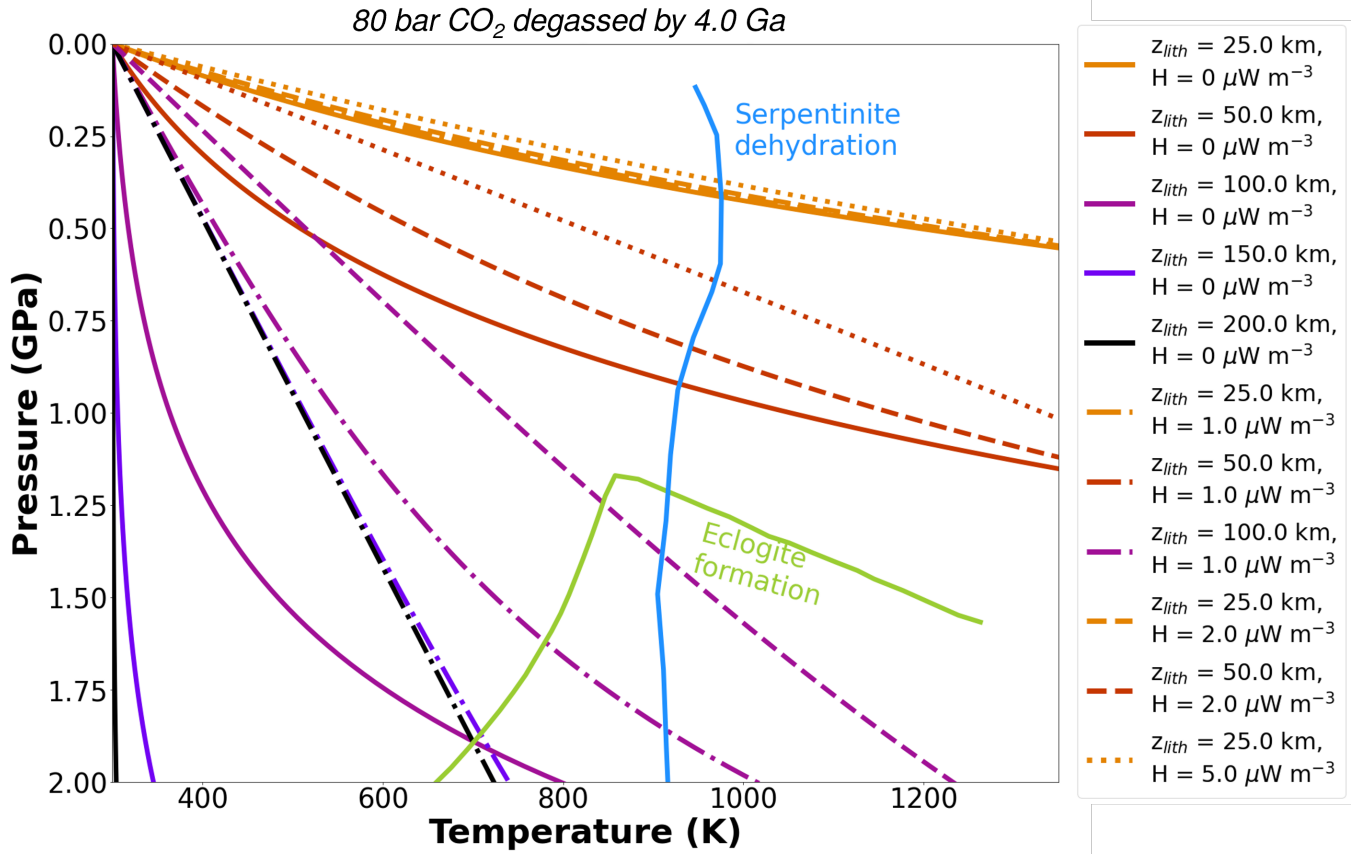


Fig. S6. Steady-state pressure-temperature profiles for Venus lithosphere with a constant advection velocity of $v = 400 \text{ km Gyr}^{-1}$ (to degas a total of 90 bar CO₂ over the course of a 0.5 Gyr hypothetical habitable era on Venus). Colors indicate different lithospheric thicknesses ranging from $z_{lith} = 25\text{--}200$ km, and solid, dash-dotted, dashed, and dotted lines indicate crustal internal heat production rates from $H = 0\text{--}5 \mu W m^{-3}$. The blue and green lines show phase boundaries corresponding to serpentinite dehydration (27) and eclogite formation (28). We assume that all pressure-temperature profiles that cross into the eclogite formation field before the serpentinite dehydration field are able to recycle hydrated lithosphere into the mantle through entrainment of dense eclogite into the convecting mantle (1).

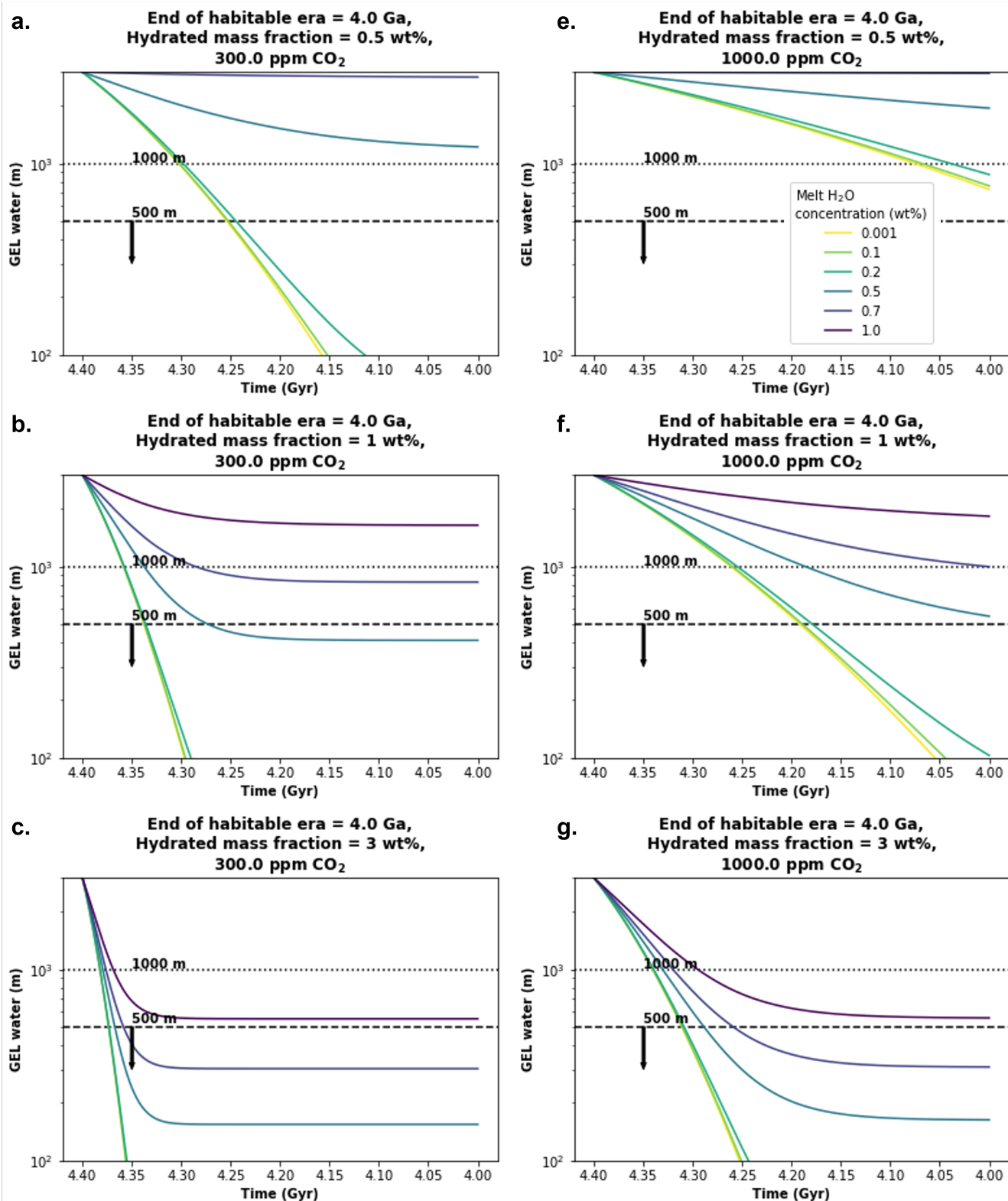


Fig. S7. Example evolutions of habitable era water inventory in km GEL for a habitable era starting with 4000 m water on the surface, with a constant atmospheric CO₂ pressure of 1 bar, for a range of H₂O mass fractions in hydrated crust (a & d. 0.5 wt%; b & e. 1 wt%; c & f. 3 wt%) and melt CO₂ concentrations (a,b, & c. 300 ppm; d, e & f. 1000 ppm). Dashed lines indicate upper limits on end-habitable-era surface water inventory from O₂ loss and ⁴⁰Ar degassing models (500 m—black dashed, 1000 m—black dotted for very reduced melt oxygen fugacities). The final size of the surface liquid water inventory must fall below these lines by the end of the habitable era at 4.0 Ga to be consistent with Venus' modern atmospheric composition.

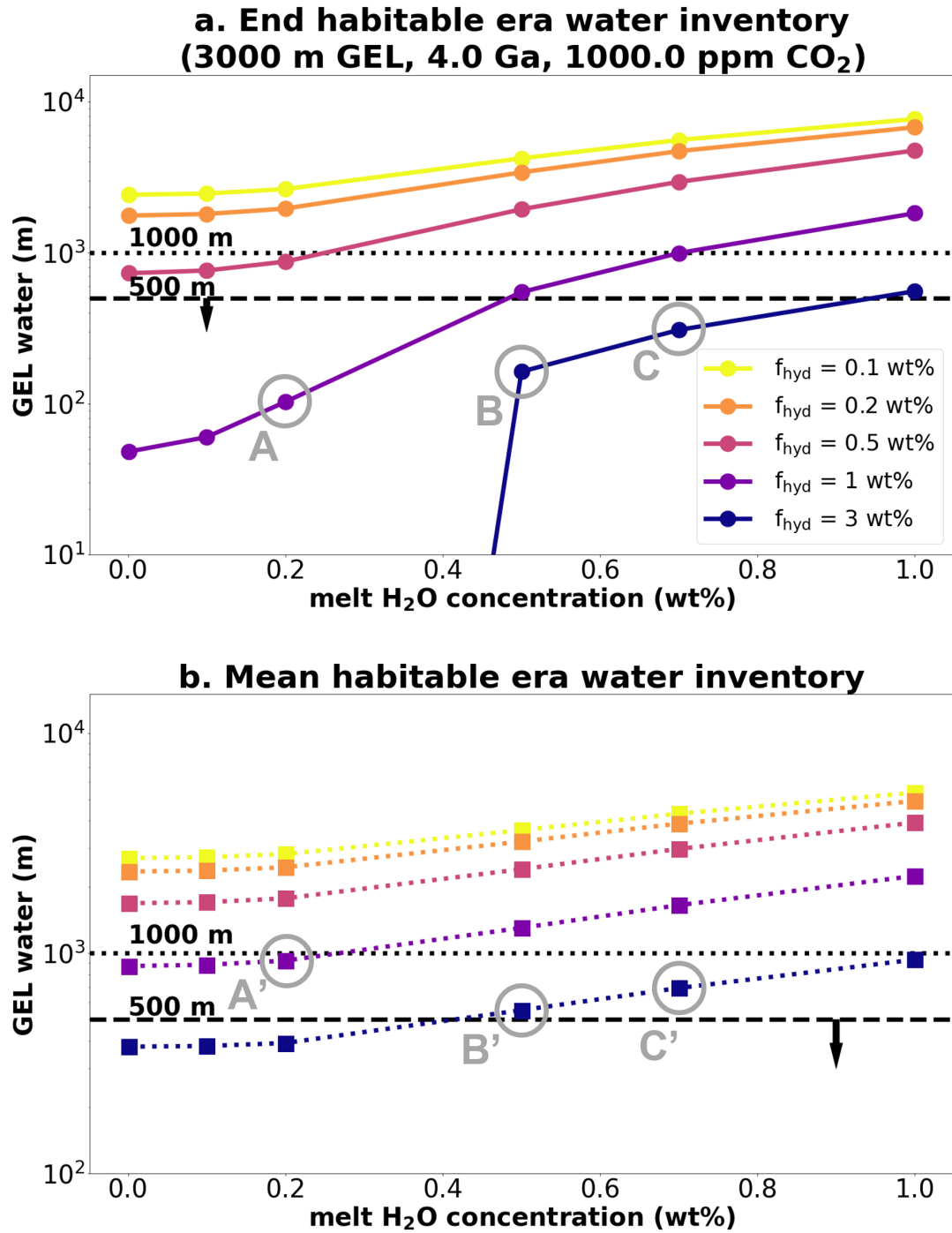


Fig. S8. Although the final surface water inventory at the end of the habitable era must be less than 500 m GEL (or 1000 m GEL for a very reducing melt oxygen fugacity), recycling of water into the mantle during the habitable era can lead to an average surface water inventory larger than these upper limits. Example runs that meet these conditions are shown by the circled points A (Panel a.) and A' (Panel b.), B (Panel a.) and B' (Panel b.), C (Panel a.) and C' (Panel b.). a. Final size of the habitable era water inventory after evolution of the surface water inventory in km GEL. Horizontal dashed line indicates upper limit on end-habitable-era surface water inventory from O₂ loss and ⁴⁰Ar models. The final size of the surface liquid water inventory must fall below these lines to be consistent with Venus' modern atmospheric composition based on our O₂ loss model. b. Time-averaged size of the habitable era water inventory during the same habitable era. Horizontal black line is the same as for a., but the average water inventory does not need to fall beneath these lines to be consistent with oxygen and ⁴⁰Ar in Venus' modern atmosphere.

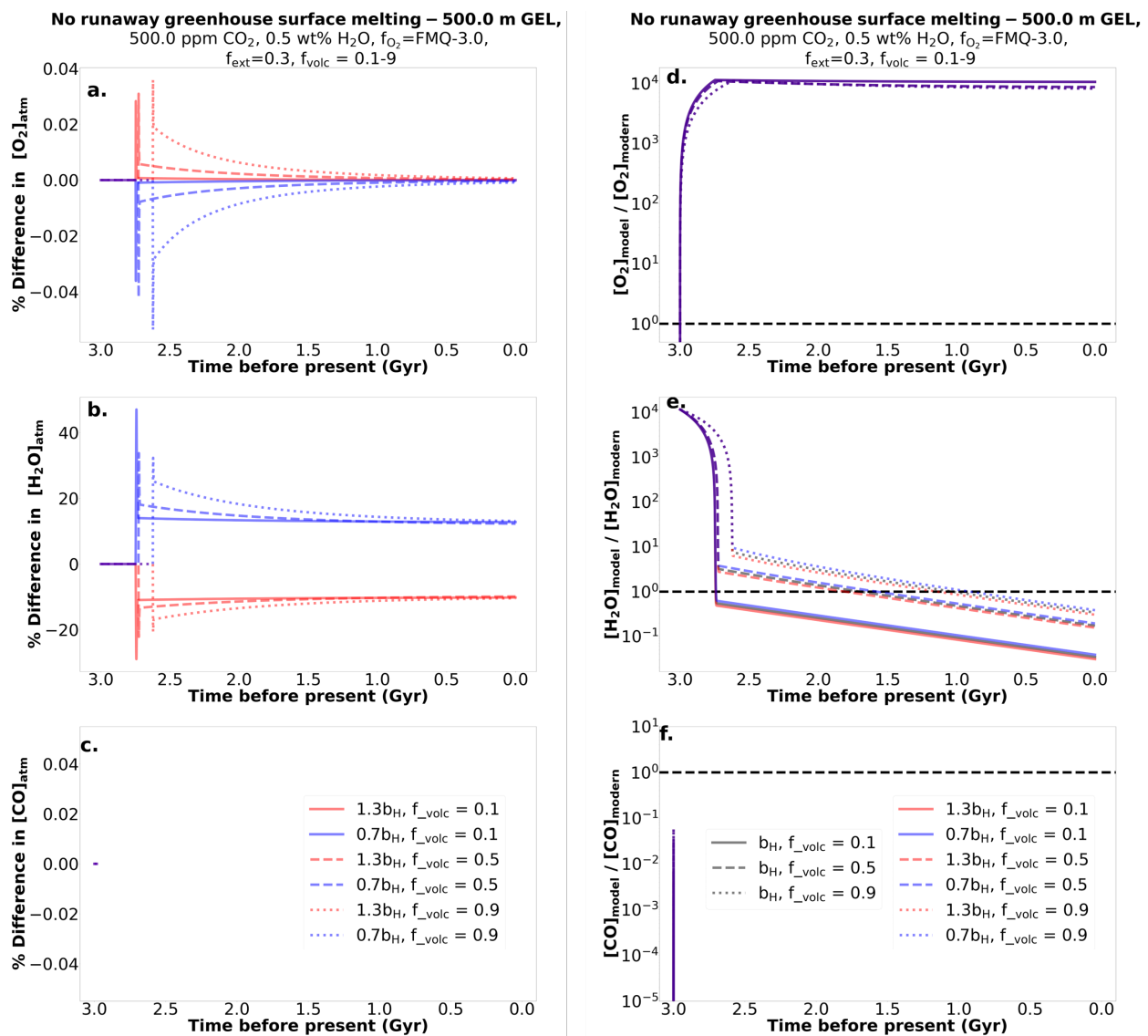


Fig. S9. Example sensitivity tests with values of binary diffusion coefficient of hydrogen through the background atmosphere (b_H) 30% lower ($0.7b_H$, blue) and 30% higher ($1.3b_H$, red) than our baseline model (black, shown in d, e, and f). Colors are transparent and curves appear purple where $0.7b_H$, baseline b_H , and $1.3b_H$ results overlap. We set 500m GEL end habitable era water inventory, 500 ppm melt CO₂, 0.5 wt% melt H₂O, $f_{ext} = 0.3$, $\log f_{O_2} = \text{FMQ-3}$, and a $f_{volc} =$ ranging from 0.1 to 0.9 (indicated by solid, dotted, and dashed lines). a-c show the percent difference in the atmospheric concentration of O₂, H₂O, and CO relative to the baseline b_H used in our O₂ loss model. CO curves are absent where the atmospheric CO concentration in the baseline model is 0. Initial, short-lived spikes in CO concentration are due to CO degassing prior to O₂ accumulation due to H escape to space. d-f show the atmospheric concentration of each species in the model divided by the maximum atmospheric concentration as constrained by modern Venus measurements. The horizontal dashed black line indicates where atmospheric concentration exactly matches the relevant modern constraint—curves that end below this line are deemed “successful” in our model. $\pm 30\%$ variations in b_H do not affect whether a given model run is successful for any parameter combination.

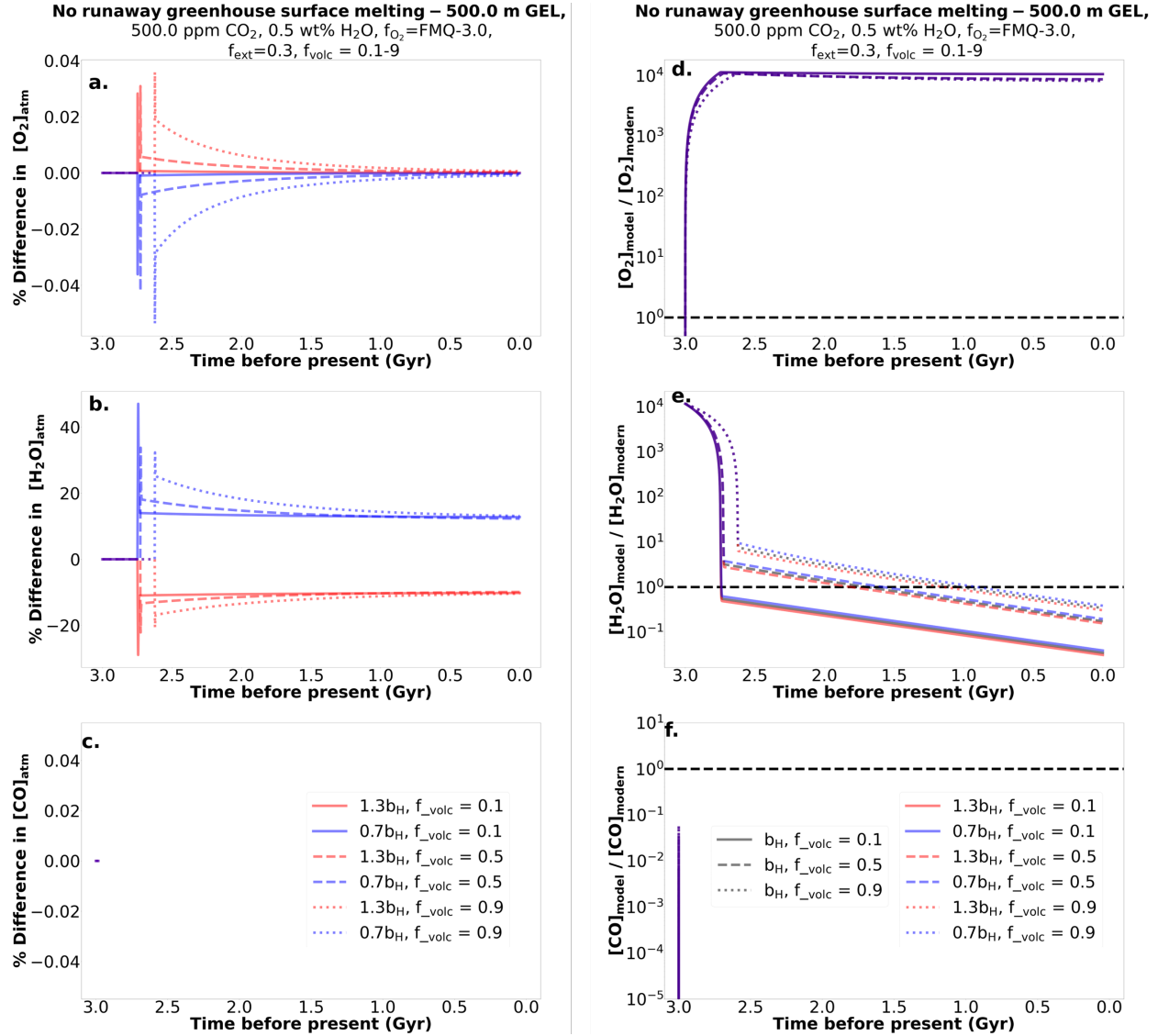


Fig. S10. Example sensitivity tests with values of $T_{atm} = 170$ K and 215 K compared with our baseline model that uses $T_{atm} = 215$ K (black, shown in d, e, and f). Colors are transparent and curves appear purple where 170 K, baseline, and 215 K results overlap. We set 500m GEL end habitable era water inventory, 500 ppm melt CO₂, 0.5 wt% melt H₂O, $f_{ext} = 0.3$, $\log f_{O_2} = \text{FMQ}-3$, and a f_{volc} = ranging from 0.1 to 1.0 (indicated by solid, dotted, dashed lines, and dashed-dotted lines). a-c show the percent difference in the atmospheric concentration of O₂, H₂O, and CO relative to the baseline b_H used in our O₂ loss model. CO curves are absent where the atmospheric CO concentration in the baseline model is 0. Initial, short-lived spikes in CO concentration are due to CO degassing prior to O₂ accumulation due to H escape to space. d-f show the atmospheric concentration of each species in the model divided by the maximum atmospheric concentration as constrained by modern Venus measurements. The horizontal dashed black line indicates where atmospheric concentration exactly matches the relevant modern constraint—curves that end below this line are deemed “successful” in our model. These variations in T_{atm} do not affect whether a given model run is successful for any parameter combination.

Table S1. Coefficient values used in calculation of surface temperature as a function of effective stellar luminosity and partial pressure of water in the atmosphere (11).

Coefficient	Value	Coefficient	Value
c_1	3.401	k_1	2.668
c_2	1.501×10^{-1}	k_2	1.019
c_3	-3.401×10^{-2}	k_3	1.099
c_4	4.702×10^{-2}	k_4	0.4683
c_5	-4.911×10^{-3}	k_5	0.7664
c_6	8.519×10^{-1}	k_6	0.4224
c_7	-1.467×10^{-2}		
c_8	-7.091×10^{-3}		
c_9	-7.627×10^{-3}		
c_{10}	8.348×10^{-3}		

Table S2. Parameter values used in ^{40}Ar degassing model.

Symbol	Description	Values	Units
t_0	End of habitable era (as in Table 1)	4, 3, 1.5	Ga
z_{crust}	Crust thickness	5, 10, 20, 30, 50	km
ϕ	Mantle melt fraction for crust production	0.025, 0.05	
c_{CO_2}	Melt CO_2 concentration (as in Table 1)	300, 500, 1000, 2000	ppm
f_{pre}	Fraction of radiogenic ^{40}Ar degassed before end of habitable era	0, 0.1, 0.25, 0.5, 1	
f_{volc}	Fraction of present day atmospheric CO_2 degassed since end of habitable era (as in Table 1)	0.1, 0.5, 0.9, 1	
f_{ext}	Extrusive volcanism fraction (as in Table 1)	0.1, 0.3, 0.5, 1	
$c_{U,M}$	Primordial mantle U concentration	13, 17, 21 (3, 21)	ppb
$r_{K/U}$	K/U ratio	7200 (21), 13800 (33)	

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