Venus: A Thick Basal Magma Ocean May Exist Today

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Key Points:

- Extensive melting of the deep mantle during Earth's accretion and differentiation was proposed to solve geochemical and geodynamic puzzles
- High temperatures and slow mantle cooling relative to Earth naturally extend the predicted lifetime of a basal magma ocean in Venus
- A basal magma ocean in Venus would sequester incompatible elements such as potassium and could have sustained a dynamo until recently

Abstract

Basal magma oceans develop in Earth and Venus after accretion as their mantles solidify from the middle outwards. Fractional crystallization of the basal mantle is buffered by the core and radiogenic and latent heat in the magma ocean. Previous studies showed that Earth's basal magma ocean would have solidified after two or three billion years. Venus has a relatively hot interior that cools slowly in the absence of plate tectonics, which reduces heat flow through the solid mantle. Consequentially, the basal magma ocean could remain as thick as ~200–400 km today. Vigorous convection of liquid silicates could power a global magnetic field until recently while a core-hosted dynamo is suppressed. The basal mantle ocean may be a hidden reservoir of potassium and other incompatible elements. A high tidal Love number could reveal a basal magma ocean and would definitively establish that the core is at least partially liquid.

Plain Language Summary

Venus is Earth's nearest neighbor but arguably the least-studied planet in the inner solar system. Although there are no direct constraints on its deep structure, the mantle of Venus is assumedly solid by analogy to Earth's current condition. However, recent models of Earth focus on the prospect that a thick layer of melt called a "basal magma ocean" persisted in the lowermost mantle for billions of years. This layer cools orders-of-magnitude more slowly than a magma ocean near the surface because the solid mantle acts as a ~1500-mile-thick blanket. Moreover, the solid mantle itself remains hot for longer in Venus compared to Earth because its surface is scorched and desiccated. This study argues that the lifetime of the basal magma ocean in Venus extends to the present. Detecting a thick, molten layer with future spacecraft missions would support the hypothesis that Venus and Earth formed under similarly energetic conditions.

1 Introduction

Magma oceans were ubiquitous during the formation of rocky planets. Giant impacts, radiogenic heating, and core formation delivered enough heat to melt entire mantles (e.g., Canup, 2012; Ćuk & Stewart, 2012; Elkins-Tanton, 2012; Nakajima & Stevenson, 2015). Crystallization of the mantle proceeded from the middle outwards because melt is gravitationally stable near both the surface and core/mantle boundary (CMB) (e.g., Labrosse et al., 2007; Stixrude & Karki, 2005). In particular, bridgmanite crystals are neutrally buoyant at mid-mantle depths where pressures are ~50 GPa (e.g., Caracas et al., 2019; Mosenfelder et al., 2007). The surficial magma ocean would have solidified within ~10 or ~100 Myr if it cools directly to space or is blanketed by a thick steam atmosphere, respectively (e.g., Hamano et al., 2013). In contrast, the basal magma ocean (BMO) may survive for billions of years because cooling through the solid mantle is orders-of-magnitude less efficient.

Earth's putative BMO has received increasing scrutiny because liquid silicates are a reservoir of incompatible elements. Labrosse et al. (2007) proposed that a long-lived BMO could explain differences in ^{142}Nd/^{144}Nd ratios between terrestrial rocks and chondritic meteorites and also solve the so-called "missing heat source" problem (e.g., Korenaga, 2008). Primordial iron-rich melt (e.g., Zhang et al., 2016) and compositional anomalies (e.g., Li et al., 2017) in the deep mantle are possibly the last residua of a BMO that has all but finished solidifying.

Models that feature a terrestrial BMO predict suppressed cooling of the core and a delayed start for the geodynamo. While the basal mantle is liquid and thus low viscosity, the thermal contrast across the CMB is negligible (e.g., Ulvrová et al., 2012). In other words, the

BMO and uppermost core should have the same temperature and must cool in tandem. Latent and radiogenic heat in the BMO buffers its cooling rate, so the core cannot cool rapidly enough to drive thermal convection until the BMO has started solidifying. Assuming that the BMO was well-mixed, Labrosse et al. (2007) suggested that convection in Earth's core would not initiate until ~3.4–4 Gyr ago. Compositional stratification in the BMO could lead to non-continuous dynamo history with a brief burst of activity before an extended pause (Laneuville et al., 2017). For now, these predictions stand untested because whether a dynamo existed in the Hadean and/or Eoarchean is a mystery (e.g., Tang et al., 2019; Weiss et al., 2015).

Why would Earth but not Venus have a BMO? These two planets have nearly identical sizes and bulk densities that are thought to reflect similar bulk compositions (e.g., Smrekar et al., 2018). Venus likely suffered energetic impacts during accretion although they produced no moon (e.g., Gillmann et al., 2016; Jacobson et al., 2017). Regardless, gravitational and radiogenic heating alone were probably sufficient to melt the primordial mantle (e.g., Elkins-Tanton, 2012; Ikoma et al., 2018). Habitable conditions may have continued until approximately one billion years ago on Venus (e.g., Way et al., 2016). Alternatively, the proximity of Venus to the Sun may have delayed solidification of the surficial magma ocean (e.g., Hamano et al., 2013) and desiccated the atmosphere and surface (e.g., Gillmann et al., 2009). In the absence of colder temperatures and oceans, Venus entered a geodynamic regime that is less efficient at cooling the mantle than plate tectonics. Models indicate that the heat flow from the solid mantle to the surface in Venus is roughly half Earth's modern value, i.e., ~20 TW versus 44 TW (e.g., Driscoll & Bercovici, 2013, 2014; Gillmann & Tackley, 2014; Weller & Kiefer, 2019). Ultimately, a BMO should have formed inside Venus and would solidify at a slower rate over time.

This study argues that the basal mantle of Venus plausibly remains molten today. Section 2 adapts earlier models of Earth's BMO (Labrosse et al., 2007) and Venus' core (O'Rourke et al., 2018). The BMO may host a dynamo (e.g., Blanc et al., 2019; Ziegler & Stegman, 2013) if liquid silicates become semi-metallic at extreme pressures and temperatures (Holmström et al., 2018; Scipioni et al., 2017; Soubiran & Militzer, 2018). Section 3 presents nominal models for Earth and Venus. Sensitivity analyses reveal the range of initial thicknesses and temperatures that are compatible with extant constraints. Section 4 discusses the implications of a BMO for Venus' geochemistry (e.g., Kaula, 1999), magnetic history (e.g., O'Rourke et al., 2018, 2019), and tidal response (e.g., Dumoulin et al., 2017). Ultimately, the prospect that a major feature such as a BMO could await detection highlights the pressing need to explore Earth's near twin.

2 Numerical Methods

Figure 1 illustrates the imagined internal structure of Venus. Broadly speaking, the compositional layering of the deep interior resembles that of Earth roughly two billion years in the past. No inner core has yet nucleated within Venus—Earth's core only began freezing from the inside out within the past billion years (e.g., Labrosse, 2015; Nimmo, 2015; O'Rourke et al., 2017). Fractional crystallization of the BMO enriched the lower mantle in iron. Convection in the solid mantle of Venus may have organized iron-rich material into thermochemical piles (e.g., Labrosse et al., 2007; Li et al., 2017), while the BMO itself remains thick enough to constitute a global layer with limited topography. However, structural similarities give way to dynamical differences because modern Venus cools slowly compared to middle-aged Earth. Crucially, the core and BMO of Venus are entirely stagnant or convecting too sluggishly to drive a dynamo.

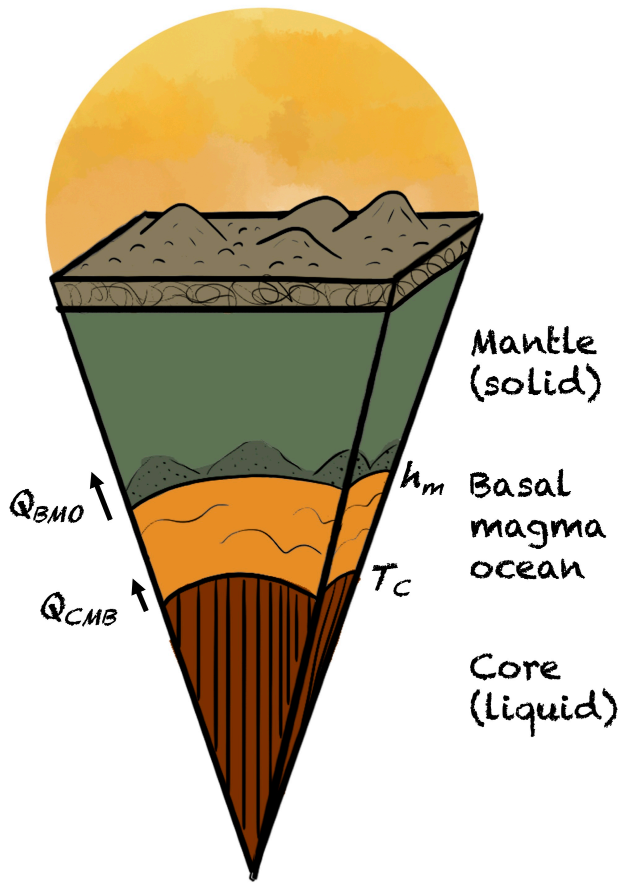

Figure 1. Cartoon of the internal structure of Venus. Four key parameters control the thermal evolution and any dynamo: the heat flow from the basal magma ocean to the solid mantle (Q_{BMO}), the heat flow across the core/mantle boundary (Q_{CMB}), the temperature at the core/mantle boundary (T_C), and the height of the basal magma ocean measured from the core/mantle boundary (h_M). Illustration by JoAnna Wendel.

2.1. Thermal histories

Parameterizations of energy sources and sinks are used to track the thermochemical evolution of the BMO and core. The Supporting Information describes how the numerical methods were adapted from Labrosse et al. (2007) and O'Rourke et al. (2018). To summarize, the heat budget of the BMO is

$$Q_{BMO} = Q_{SM} + Q_{RM} + Q_{LM} + Q_{CMB}, \quad (1)$$

where the heat flow across the solid/liquid interface in the basal mantle (Q_{BMO}) is imposed as a boundary condition. The first three terms on the right side represent secular cooling (Q_{SM}), radiogenic heating (Q_{RM}), and latent heat (Q_{LM}) in the BMO (e.g., Labrosse et al., 2007; Ziegler & Stegman, 2013). The heat flow across the CMB (Q_{CMB}) comprises additional terms that describe the energy budget of the core (e.g., Labrosse, 2015; O'Rourke et al., 2018):

$$Q_{CMB} = Q_{SC} + Q_{RC} + Q_{PC} + Q_{GC} + Q_{LC} + Q_{IC}, \quad (2)$$

which include secular cooling of the outer core (Q_{SC}), radiogenic heating (Q_{RC}), and precipitation of light species such as magnesium oxide near the CMB (Q_{PC}). After nucleation, the inner core contributes gravitational energy from the exclusion of light elements (Q_{GC}), latent heat (Q_{LC}), and secular cooling under the assumption of efficient thermal conduction (Q_{IC}). Terms representing the heat of reaction and pressure changes from thermal contraction are considered negligible (e.g., Blanc et al., 2019). Exothermic reactions between the core and BMO such as oxygen partitioning (e.g., Pozzo et al., 2019) are also ignored, but could also slow the solidification of the BMO. Large amounts of heat could be transported laterally within the BMO and core and/or conducted upwards. However, the energy budgets only include heat that crosses the CMB or the upper boundary of the BMO because heats that are generated and lost within a single layer sum to zero. Labrosse et al. (2007) assumed that Q_{CMB} was proportional to one specific heat for the core, which was held constant. In reality, the effective specific heat of the core should decrease over time with radiogenic heating but increase once the inner core nucleates. This study avoids that simplification in order to delineate the inner core and dynamo.

The key to generating thermal histories is realizing that nearly all of the heat sources are directly proportional to the cooling rate of the core. Radiogenic heating is the exception but easy to calculate as an absolute value given the initial abundances of heat-producing elements. For all other terms in Eq. 1,

$$Q_i = \tilde{Q}_i \left(\frac{dT_C}{dt}\right), \tag{3}$$

where $\tilde{Q}_i$ depends only on the thermodynamic properties of the BMO and core (i.e., physical constants from Tables S1 and S2) and T_C is the temperature at the CMB. Secular cooling in the BMO is parameterized using a specific heat that is invariant with depth. Latent heat is computed with an idealized phase diagram for a well-mixed BMO (Labrosse et al., 2007). Energetic terms for the core are derived by integrating fourth-order polynomials, which describe the radial density and temperature profiles in the core, over the volume(s) of the outer and/or inner cores (e.g., Labrosse, 2015; Nimmo, 2015; O'Rourke et al., 2018). Combining Eq. 1 and 2,

$$\frac{dT_C}{dt} = \frac{Q_{BMO} - Q_{RM} - Q_{RC}}{\tilde{Q}_{SM} + \tilde{Q}_{LM} + \tilde{Q}_{SC} + \tilde{Q}_{PC} + \tilde{Q}_{GC} + \tilde{Q}_{LC} + \tilde{Q}_{IC}} \,. \tag{4}$$

This equation places the boundary and initial conditions in the numerator and the structural parameters in the denominator. As discussed in the Supporting Information, the thickness of the BMO and the radius of the inner core are also directly proportional to dT_C/dt. Given initial values for the size of the BMO and T_C, the forward Euler method generates a thermal history consisting of all the time-dependent quantities listed in Table S3. Timesteps of ~1 Myr suffice because halving the timestep yielded no discernable change in the model results.

2.2. Prospects for dynamo action

Thermal histories reveal when and where a dynamo could exist. The basic criterion for a dynamo is that a rapidly rotating, electrically conductive fluid convects with sufficient vigor. Although the rotation of Venus is "slow" relative to Earth, it is "rapid" in the context of dynamo physics because the Coriolis force would strongly affect fluid flow in its core and/or BMO as expressed by small Rossby numbers ($\sim 10^{-5} \ll 1$) at the equator (e.g., Stevenson, 2003). Simply put, the rotation rate of Venus is not to blame for the lack of a strong magnetic field. The dynamo criterion thus reduces to convection in the BMO and/or core. If the heat flow across the upper boundary does not exceed that conducted upwards along an adiabatic gradient, then thermal conduction—the bane of dynamos—transports heat without fluid motion. Chemical sources of buoyancy such as radiogenic heating, precipitation of light elements, and growth of an inner core decrease the minimum heat flow required to support a dynamo.

There are two general methods for estimating the power of a dynamo. First, scaling laws relate the flux of compositional and/or thermal buoyancy to the velocities of convective flows (e.g., Christensen, 2010). Higher velocities translate to stronger magnetic fields. Second, entropy budgets reveal whether non-zero amounts of dissipation are available for the dynamo (e.g., Labrosse, 2015). Classical thermodynamics explicitly compares the entropy sink associated with thermal conductivity to the entropy production from the myriad heat sources. To enable straightforward comparisons to previous studies, velocity scalings are applied to the BMO (e.g., Labrosse et al., 2007; Ziegler & Stegman, 2013) while the entropy budget is calculated for the core (e.g., O'Rourke et al., 2018). In contrast, Driscoll & Bercovici (2014) used velocity scalings for the core while Blanc et al. (2019) recently formulated the entropy budget for the BMO. Either method yields similar results. Ultimately, no dynamo exists when the heat flow is sub-critical.

Velocity scalings quantify convective vigor within the BMO. The magnetic Reynolds number is Rm = $\mu_0 v_M h_M \sigma_M$. A dynamo may exist if Rm > $O(10)$, where an exact cutoff of 40 is commonly used (e.g., Stevenson, 2003). Here μ_0 is the permittivity of free space and h_M is the thickness of the BMO (e.g., Ziegler & Stegman, 2013). The electrical conductivity (σ_M) is assumed to equal 2×10^4 S/M (Holmström et al., 2018; Scipioni et al., 2017; Soubiran & Militzer, 2018). Nominal values for the convective velocity (v_M) are taken from the scaling based on the Coriolis-Inertial-Archimedean (CIA) force balance:

$$v_{CIA} = \left(\frac{Q_{BMO}}{\rho_M H_T}\right)^{\frac{2}{5}} \left(\frac{h_M}{\Omega}\right)^{\frac{1}{5}}, \tag{5}$$

where ρ_M is the density of the BMO, H_T is its thermal scale height, and Ω is the planetary rotation rate. Scalings based on mixing length theory and the Magnetic-Archimedean-Coriolis (MAC) force balance are also considered (Supporting Information). If the Rm-criterion is satisfied, then the magnetic field at the equatorial surface is calculated as

$$B_S = \frac{1}{7}\left(2\epsilon f_{ohm} \mu_0 \rho_M v_{CIA}^2\right)^{\frac{1}{2}} \left(\frac{r_B}{r_P}\right)^3, \tag{6}$$

where ε is a constant prefactor, f_{ohm} is the fraction of available power that is converted to ohmic dissipation as magnetic energy, r_B is the radial distance from the planetary center to the upper boundary of the BMO, and r_P is the planetary radius (Christensen, 2010). These scalings predict Earth-like surface strengths of ~30 μT for flow velocities of ~1 cm/s in the BMO.

Entropy budgets determine whether the core is convective. The total dissipation available for a dynamo is the sum of various sources and sinks:

$$\Phi = \frac{T_{DC}[T_L(r_I) - T_C]}{T_L(r_I)T_C}(Q_{LC} + Q_{IC}) + \frac{T_{DC}}{T_C}(Q_{GC}) + \frac{T_{DC} - T_C}{T_C}(Q_{RC}) + \frac{T_{DC}(T_{SC} - T_C)}{T_{SC}T_C}(Q_{SC}) + \frac{T_{DC}}{T_C}(Q_{PC}) - T_{DC}E_K. \tag{7}$$

Here T_{DC} is the average temperature in the outer core, $T_L(r_I)$ is the liquidus temperature of the core at the inner core boundary, and T_{SC} is an effective temperature association with dissipation from secular cooling. The entropy sink (E_K) is associated with and directly proportional to thermal conductivity. The thermal conductivity of the core near the CMB is uncertain within a broad range: ~33–226 W/m/K (e.g., Konôpková et al., 2016; Ohta et al., 2016). In contrast, the thermal conductivity of perovskite at similar conditions is unambiguously lower at ~10–20 W/m/K (e.g., Ohta et al., 2012). Gravitational energies associated with chemical buoyancy (Q_{GC} and Q_{PC}) are efficient sources of dissipation that are not penalized by "Carnot-like" efficiency terms (e.g., Labrosse, 2015; O'Rourke et al., 2018). The total dissipation is translated into a true dipole moment (TDM) using a scaling law that considers the relative amounts of dissipation generated at the CMB and inner core boundaries (Aubert et al., 2009). This scaling is only approximate because how the dynamo changes once the inner core nucleates remains uncertain (e.g., Landeau et al., 2017). In any case, the surface magnetic field at the equator follows as

$$B_S = \frac{\mu_0}{4\pi}\left(\frac{\mathrm{TDM}}{r_P^3}\right). \tag{8}$$

These formulations assume that flows in the core and BMO are independent. Future studies should investigate possible magnetohydrodynamic couplings between flows in both regions.

3 Results

3.1. Models for Earth

Earth is a benchmark for models of Venus. Augmenting Labrosse et al. (2007) with a detailed description of the core was the first step in this study. The nominal model was initialized with the thickness and temperature of the BMO equaling 750 km and 5250 K, respectively. The BMO started with 20 TW of radiogenic heating, which is ~14% of the total heating expected for bulk silicate Earth (e.g., Lay et al., 2008). The core contained 50 ppm of potassium (e.g., Hirose et al., 2013) and precipitated light elements at a rate of ~10^{19} kg/K or 5×10^{-6} K^{-1} normalized to the mass of the core (e.g., Badro et al., 2018; O'Rourke & Stevenson, 2016). Finally, Q_{BMO} decreased linearly from 55 TW at the start to 15 TW at present, which approximates the cooling history obtained using boundary layer models (e.g., Blanc et al., 2019; Labrosse et al., 2007; Ziegler & Stegman, 2013) and dynamical simulations (e.g., Nakagawa & Tackley, 2010, 2015).

Figure S1 shows that this model reproduces major features of Earth's history. First, the globally averaged thickness of the BMO is only ~1 km at present (Figure S1c), which is small enough that solid-state mantle convection would concentrate the BMO into pockets of primordial melt consistent with seismology (Labrosse et al., 2007). Second, the predicted radius of the inner core today is 1206 km, close to the correct value of 1220 km (Figure S1d). Third, the inner core nucleated ~500 Myr ago (Figure S1e) but a dynamo persisted at all times (Figure S1f). If the thermal conductivity of the core is relatively low at ~40 W/m/K, then a core dynamo always operated. Higher thermal conductivity could suppress a core dynamo at early times, but the BMO would still host a dynamo with roughly the same strength for the first ~0.5–1.5 Gyr (Figure S1f).

Figure S2 reports a sensitivity test for the terrestrial models. Initial values of h_M and Q_{BMO} were varied from ~600 to 1500 km and ~35 to 60 TW, respectively. Decreasing the heat flow to the solid mantle is equivalent to decreasing the amount of radiogenic heating and/or the latent heat of solidification in the BMO, so those other parameters were not separately permuted. Key outputs were the present-day thickness of the BMO (Figure S2a) and the lifetime of the dynamo in the BMO (Figure S2b). Neither key output was very sensitive to the initial temperature of the BMO. Likewise, the thermal history of the BMO does not depend on the particulars of the evolution of the core. For example, changing the amount of radiogenic heating or the rate of elemental precipitation only adjusts the proportions in which the available heat flow (Q_{CMB}) is distributed between the various sources. Models are invalidated if $h_M \geq 10$ km today because seismology has revealed no global melt layer in the basal mantle. Ultimately, Earth's BMO could have started as thick as ~1000 km, in which case a dynamo in the BMO may have survived for ~2 Gyr. Assuming that Earth and Venus accreted in an equivalently energetic environment, expecting that the BMO in Venus began with a similar size seems logical.

3.2. Models for Venus

A nominal model for Venus was obtained through two modifications to the terrestrial benchmark. First, various structural parameters were adjusted to slightly lower internal pressures (Table S2), e.g., ~125 versus 130 GPa at the CMB for Venus and Earth, respectively. Second, the heat flow to the solid mantle was halved at all times, i.e., Q_{BMO} decreased linearly from 27 to 7 TW over 4.5 Gyr to match the cooling history in dynamical simulations (e.g., Gillmann & Tackley, 2014; O'Rourke et al., 2018). All other model parameters—the initial temperature and thickness of the BMO and the potassium content of the core—were held constant. These treatments are faithful to the hypothesis that Venus and Earth began as twin planets and then diverged because their surficial magma oceans solidified on different timescales.

Figure 2 elucidates the dramatic consequences of slow mantle cooling for the structure and dynamics of the deeper interior. Radiogenic and latent heat dominate the energy budget of the BMO (Fig. 2a), so the core and BMO only cool by ~216 K over 4.5 Gyr (Fig. 2b). The thickness of the BMO decreases by ~494 km to ~256 km (Fig. 2c), which is deep enough to constitute a global layer today. Driving a dynamo in the core with thermal convection alone would require $Q_{CMB} > 4.8$ TW compared to its maximum and modern values of ~2.2 and 1.6 TW, respectively. Dissipation in the core remains negative at all times (Fig. 2d) even though the thermal conductivity was set to the lowest plausible value (40 W/m/K). If there were no BMO, then $Q_{CMB} \sim Q_{BMO}$ and the core would have powered a dynamo at all times. The BMO may have hosted a dynamo until recently, although the predicted lifetime depends on the choice of velocity scaling law (Fig. 2e). Mixing length theory suggests that a strong dynamo exists today, which is incorrect (Phillips & Russell, 1987). In contrast, the MAC scaling indicates that the dynamo died ~2.8 Gyr ago. The favored, CIA scaling suggests that Venus had a magnetic field with surface strengths of ~10–30 μT until ~0.9 Gyr ago, less than the estimated ages of many surface units.

Figure 3 shares a sensitivity test for these models of Venus. Initial values of h_M and Q_{BMO} were varied again (Fig. 3a and 3c), alongside a range of starting values for T_C (Fig. 3b and 3d). Models were invalidated if the CIA scaling predicted that the BMO would host a dynamo today because Rm > 40. The derived upper limits on the initial thickness of the BMO are similar for both Earth and Venus although the criterion for Earth was based on a different requirement: complete solidification of the BMO. For Venus, setting $h_M \leq 860$ km initially would yield acceptable models using the nominal cooling history, which return $h_M \leq 370$ km at present day. The upper limit for h_M increases to ~1250 km but decreases to ~630 km if Q_{BMO} is raised or lowered initially to 40 or 20 TW, respectively. As for Earth, these results are mostly insensitive to the absolute temperature of the BMO. Changing T_C by ~1500 K is equivalent to adjusting Q_{BMO} by ~3 TW only, so the a priori uncertainty on T_C is relatively unimportant albeit very large.

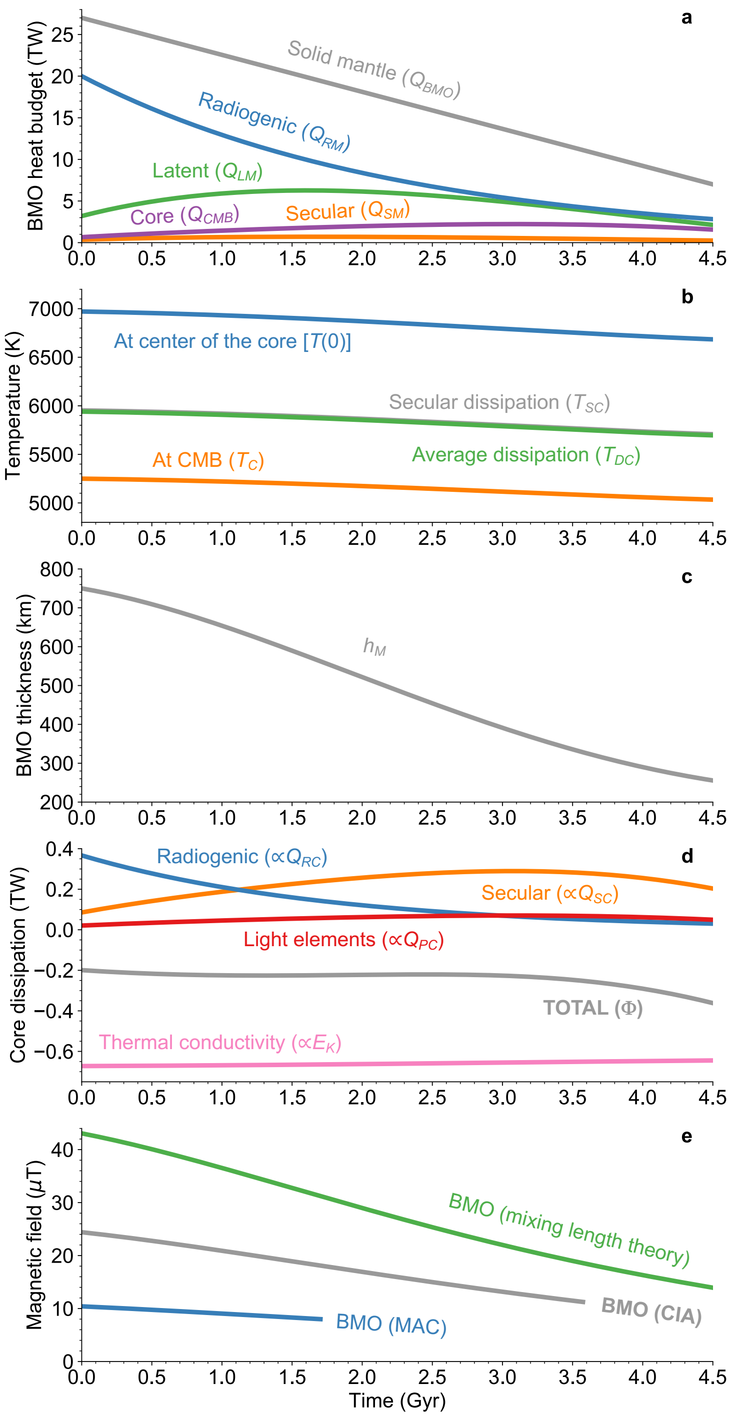

Figure 2. Nominal model for Venus. A long-lived basal magma ocean is the natural outcome of the conventional cooling history for the solid mantle, which is compatible with the observed lack of a strong magnetic field today. (a) Heat budget of the basal magma ocean. (b) Temperatures at the core/mantle boundary and deeper in the core. (c) Thickness of the basal magma ocean. (d) Dissipation budget for the core including all non-zero terms from Equation 7, which is always negative in total. (e) Estimated strength of the magnetic field at the surface based on three velocity scalings for the basal magma ocean. The CIA scaling reproduces the lack of a dynamo today but further motivates a search for crustal remanent magnetization.

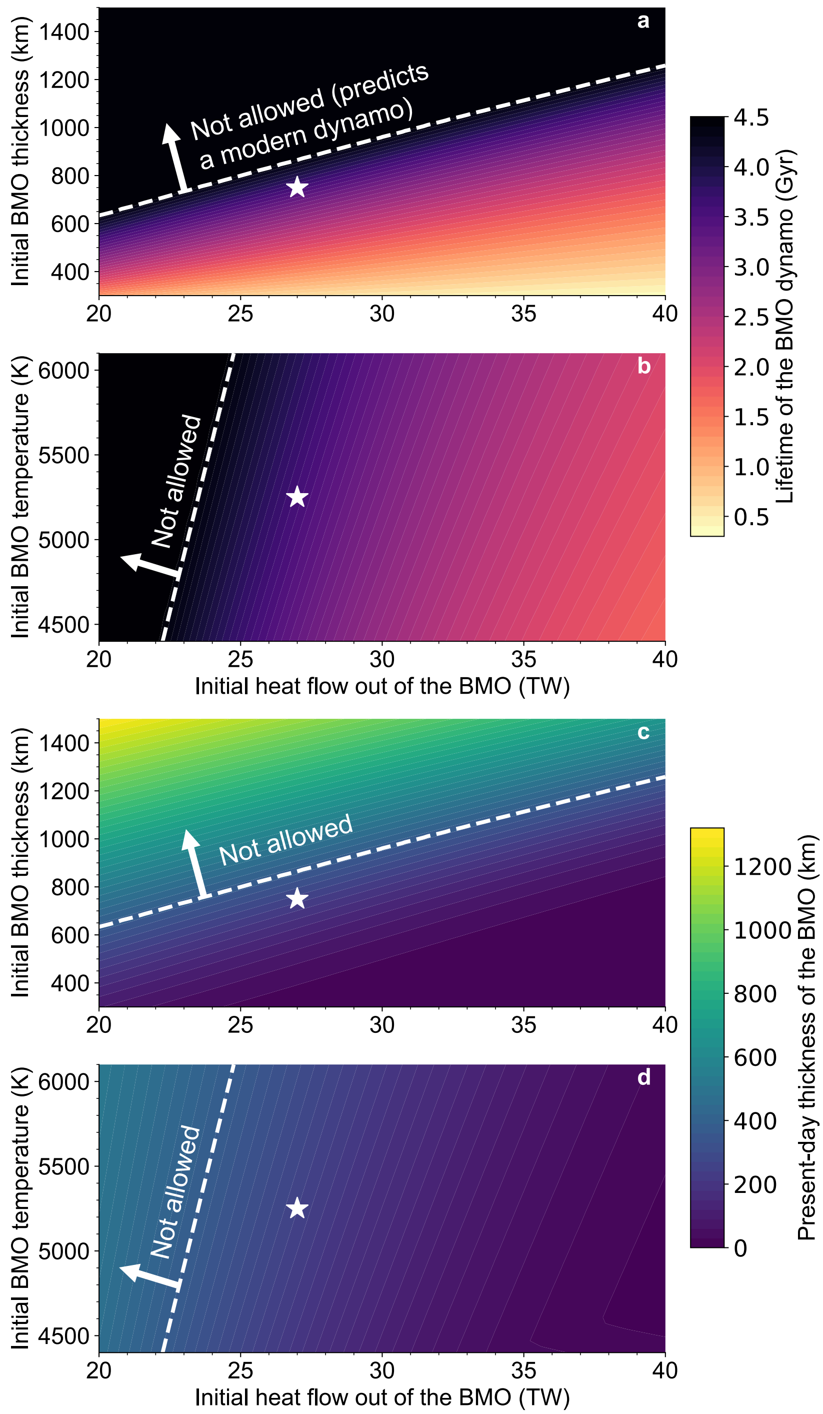

Figure 3. Sensitivity analysis for models of Venus. The maximum thickness of the basal magma ocean at present day is ~400 km. Models with larger melt layers conflict with observational evidence against a strong intrinsic magnetic field. Arrows point towards invalid initial conditions on one side of the dashed white lines. Lifetimes of the dynamo in the basal magma ocean are estimated with the Coriolis-Inertial-Archimedean velocity scaling as functions of the initial heat flow to the solid mantle and the initial (a) temperature and (b) thickness of the basal magma ocean. Present-day thicknesses of the basal magma ocean are reported versus its initial (c) temperature and (d) thickness. White stars represent the nominal model.

4 Discussion

Future investigations should build on the simplifications that this study employed. In particular, solidification timescales were estimated with an idealized phase diagram (e.g., Labrosse et al., 2007) but actually depend on whether compositional layering develops (e.g., Laneuville et al., 2017) and on the partition coefficient for FeO between the BMO and the solid mantle (e.g., Blanc et al., 2019). Fully dynamical simulations are required to generate self-consistent cooling histories and track the fate of iron-rich residua in the solid mantle. Despite these limitations, the likelihood that a BMO persisted within Venus has myriad implications. For example, incompatible elements from the lowermost ~650–1250 km of the mantle (e.g., ~11–25% of the mantle's total volume), including potassium and associated decay products such as argon-40 (e.g., Kaula, 1999; Namiki & Solomon, 1998; O'Rourke & Korenaga, 2015; Xie & Tackley, 2004), could remain hidden in a reservoir that surface volcanism and degassing would not sample. Beyond geochemistry, two other issues deserve particular attention.

4.1. Tidal response of a basal magma ocean

Tidal deformation of Venus constrains the structure of the deep interior. Assuming an elastic response from the mantle to solar tides, Yoder (1995) predicted that a Love number k_2 above 0.23 would signal that the core is at least partially liquid today. Doppler tracking of the NASA Magellan and Pioneer Venus Orbiter missions then determined that $k_2 = 0.295 \pm 0.066$ (Konopliv & Yoder, 1996). However, Dumoulin et al. (2017) recently showed that realistic viscoelasticity of the mantle strongly increases k_2 relative to predictions from elastic models, so envisioned missions would need to constrain $k_2 > 0.27$ to verify that the core is liquid. A BMO would decouple the solid mantle from the core and, in principle, raise k_2 even if the underlying core were completely solid. In reality, the solidus of the core is far lower than that of the basal mantle—no realistic thermal history features a BMO and a solid core (O'Rourke et al., 2018). Overall, measuring a high k_2 would thus prove that the core remains partially liquid.

4.2. Magnetic history of Venus

Venus could have sustained an Earth-strength magnetic field until recently. O'Rourke et al. (2018) assumed that the mantle had fully solidified and found that the core would power a dynamo for billions of years. High thermal conductivity for the core (e.g., >100 W/m/K) was invoked because simulations using lower conductivity over-predicted the lifetime of the dynamo. In this study, a dynamo exists instead in the BMO but survives for similar timescales. Thermal conductivity of the core is no longer a critical uncertainty because the cooling rate of the core is sub-adiabatic regardless. Crustal remanent magnetism is a potentially observable consequence of an early dynamo in either the core or BMO (O'Rourke et al., 2019). Any detection of crustal magnetization would indicate that Venus and Earth accreted under similarly energetic conditions. Alternatively, Venus could have accreted under less energetic conditions where any BMO was short-lived and chemical stratification precluded convection in the core (Jacobson et al., 2017).

5 Conclusions

Earth's early evolution featured a basal magma ocean that took several billion years to solidify. Until now these models have not been extended to Venus. Slow mantle cooling in the absence of plate tectonics on Venus is a common feature of dynamical simulations. The natural consequence is an extended lifetime for the basal magma ocean. Roughly speaking, halving the cooling rate doubles the solidification timescale. Models indicate that the lowermost ~200–400 km of the mantle of Venus plausibly remains molten today. Seismology would enable the direct detection of a thick melt layer, which should also yield a high tidal Love number that is degenerate with a partially liquid core. The basal magma ocean is a hidden reservoir of incompatible elements that the solid mantle will not ingest for billions of years. Vigorous fluid motions in the basal magma ocean can drive a dynamo until recent times (<1 Gyr ago), but latent and radiogenic heat keeps the cooling rate of the core below the adiabatic limit for a dynamo driven by thermal convection. The basal magma ocean gives, and the basal magma ocean has taken away.

Acknowledgments

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