A Subsurface Magma Ocean on Io: Exploring the Steady State of Partially Molten Planetary Bodies

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Abstract

Intense tidal heating within Io produces active volcanism on the surface, and its internal structure has long been a subject of debate. A recent reanalysis of the Galileo magnetometer data suggested the presence of a high-melt-fraction layer with >50 km thickness in the subsurface region of Io. Whether this layer is a “magmatic sponge” with interconnected solid or a rheologically liquid “magma ocean” would alter the distribution of tidal heating and would also influence the interpretation of various observations. To this end, we explore the steady state of a magmatic sponge and estimate the amount of internal heating necessary to sustain such a layer with a high degree of melting. Our results show that the rate of tidal dissipation within Io is insufficient to sustain a partial-melt layer of \( \phi > 0.2 \) for a wide range of parameters, suggesting that such a layer would swiftly separate into two phases. Unless melt and/or solid viscosities are at the higher end of the estimated range, a magmatic sponge would be unstable, and thus a high-melt-fraction layer suggested in Khurana et al. is likely to be a subsurface magma ocean.

1. Introduction

Io, the innermost among four Galilean satellites, is tidally heated by forced eccentricity, exhibiting the most active surface volcanism in our solar system. The absence of impact craters, the presence of substantial topography, and globally distributed volcanoes are best explained by the dominance of melt transport through a mostly rigid lid (O’Reilly & Davies 1981; Moore 2003). The extent of partial melting in the interior, however, remains elusive (e.g., Bierson & Nimmo 2016; Spencer et al. 2020). The internal structure characterizes the mode of tidal dissipation, including its magnitude, phase, and distribution, and thus is crucial for delineating the energy budget of a tidally heated body.

The reanalysis of the magnetometer data collected by the Galileo mission has shown a strongly induced magnetic signature within Io, and such an induction feature suggests the presence of a global high-melt-fraction (>20%) layer thicker than 50 km in a subsurface region (Khurana et al. 2011). This has raised the question of whether this layer is a rheologically liquid “magma ocean” or a melt-rich “magmatic sponge” with interconnected silicate solid (Figures 1(a) and (b)). The existence of a magma ocean would decouple the lithosphere from the interior, and in the original model for tidal heating proposed by Peale et al. (1979), the tidal heating is then enhanced by the far larger strain (larger Love number) of the overlying shell. A magma ocean would likely not be internally very dissipative because of its low viscosity, while a magmatic sponge might be highly dissipative. We do not seek to resolve the dissipation issue here, but a magma ocean is likely to alter the internal distribution of tidal heating and may erase the correlation between the depth of tidal dissipation and surface volcanism pattern. A concentration of volcanoes at low latitudes has been interpreted as a concentration of tidal dissipation in the shallow mantle (e.g., Tackley et al. 2001; Tyler et al. 2015), but a magma ocean could buffer the localization of magma upwelling, and the spatial distribution of eruptions could rather be a result of heterogeneity in lithospheric weakness.

In this paper, we build a 1D melt percolation model and examine the stability of a partially molten layer. A partial melt under large-scale deformation is known to be unstable and accumulates in veins (Stevenson 1989), so melt migration in Io may not necessarily be characterized by percolation. Veins could eventually be interconnected, and melt may efficiently be drained to the surface. The growth of the instability, however, is yet to be quantified due to its nonlinearity, and how efficiently partial melt would escape to the surface remains unclear. Instead, we assume percolative flow and show that, even in this conservative approach, a mixture of interconnected melt and solid would swiftly separate into two phases. Our results indicate that a melt-rich layer beneath the Io’s subsurface is thus more likely to be a magma ocean rather than a magmatic sponge. The observations of libration amplitude (Van Hoolst et al. 2020) or the Love number could potentially resolve the two scenarios. The latter will be measured by the Juno extended mission, perhaps to sufficient accuracy to delineate between a magma ocean and a magma sponge, and we aim to make a theoretical prediction before the planned measurements of the Love number. This paper is organized as follows. In Section 2, we describe our model setup, focusing on the boundary conditions of the model. Model results with a range of parameter values are presented in Section 3, and the presence of a subsurface magma ocean is suggested in the last part of this paper.

2. Methods

Melt percolation within a tidally heated mantle is considered in this study to examine the stability of a partially molten layer. Instead of solving for a time evolution, we explore steady-state structures and investigate the relation between the degree of
tidal dissipation within the mantle and the resulting extent of partial melting. The goal is to estimate the amount of tidal heating necessary to maintain a “magmatic sponge” to show that tidal dissipation within Io is insufficient to sustain interconnected solid containing a high degree of partial melting. For simplicity, a spherically symmetric structure is assumed, and 1D radial profiles of percolation velocity and melt fraction are solved for various degrees of tidal dissipation.

Given the uncertainty in the state of Io, this study aims to model heat transport with a minimum number of parameters. The parameters characterizing the mantle system are limited to solid viscosity, $\eta_s$, the ratio of melt viscosity to permeability, $\eta_m/k_m$, and melt–solid density difference, $\Delta \rho$ (300 kg m$^{-3}$ in this study; e.g., Stevenson & Scott 1991; Spiegelman 1993a), and we explore how the profile of melt fraction changes with different rates of tidal heating, $\psi$. In addition, we specify the liquid overpressure at the top boundary, $\tau_0$. A reasonable value for the liquid overpressure is discussed in detail in Section 2.2, but fortunately, the choice of overpressure does not affect our key results. We note that the top boundary does not necessarily represent the crust–mantle boundary. The crust and the partial-melt region are likely to be separated by a transition zone of several kilometers in thickness, which is characterized by transient behaviors and may not be described by percolation or eruption. We discuss the nature of such a transition zone in Sections 2.2 and 4.2. It is noted that partial melt is compositionally different from the bulk mantle, and thus the near-surface solid layer, which is consisted of a solidified partial melt, would also differ in composition from the mantle and can be defined as a crust.

The model presented here can be applied to any body size, but with an application to Io in mind, a planet with 1820 km radius and a partially molten mantle of ~200–700 km is considered in this study. We first summarize the governing equations and then discuss boundary conditions.

### 2.1. Governing Equations

Governing equations for melt percolation within a solid–melt mixture consist of the conservation of mass,

$$\frac{\partial \phi}{\partial t} = -\nabla \cdot (\phi \nabla \psi) + \Gamma,$$

the conservation of momentum (Stevenson & Scott 1991),

$$(1 - \phi) \Delta \rho g = \frac{\eta_l}{k_m} \phi (v_l - v_s)$$

$$+ \frac{d}{d\zeta} \left[ (1 - \phi) \left( \zeta + \frac{\eta_s}{3} \right) \frac{dv_s}{d\zeta} \right],$$

and the conservation of energy (Katz 2008),

$$\rho c_p \frac{\partial T}{\partial t} + \nabla \cdot ((1 - \phi) v_s \cdot \rho c_p T)$$

$$+ \nabla \cdot (\phi (v_l - v_s) (\rho c_p T + L)) = k \nabla^2 T + \psi.$$  

The parameter $\phi$ denotes volumetric melt fraction, $T$ is temperature, and $\Gamma$ is a change in melt fraction due to tidal heating. The constant $g$ indicates gravitational acceleration (1.5 m s$^{-2}$), $\rho$ is the mantle density (3300 kg m$^{-3}$), $c_p$ is specific heat capacity (1000 J kg$^{-1}$ K$^{-1}$), $L$ is the latent heat of silicates (4 × 10$^7$ J kg$^{-1}$), and $k$ is thermal conductivity (3.3 W m$^{-1}$ K$^{-1}$). We set effective bulk viscosity, $\zeta$, to $\eta_s/\phi$ based on microscopic models (Sleep 1988; Hewitt & Fowler 2008).

By considering a horizontally uniform steady-state system, a frame-invariant variable $u \equiv \phi (v_l - v_s)$ can be defined (Scott & Stevenson 1984), which becomes the same as the solid velocity under a barycentric coordinate. Using the separation flux $u$,
Equations (1)–(3) are simplified to

\[(1 - \phi)\frac{\Delta \rho g}{\rho L} + \frac{d}{dz}\left(1 - \phi\right)\left(\zeta + \frac{4}{3}\eta_l\right)\frac{du}{dz} = \frac{\eta_l u}{k_m}, \quad (4)\]

\[\rho L \nabla \cdot ((1 - \phi)u) = k \nabla^2 T + \psi, \quad (5)\]

and the nondimensionalized version of the two equations are solved here (Appendix). The latter equation indicates that heat loss by the advection of latent heat mostly balances tidal heating. Thermal conduction is mostly negligible in the partial-melt layer except in the near-surface region where the overlying cold crust affects the thermal structure (Section 4.1).

### 2.2. Boundary Conditions

Boundary conditions of the momentum equation (Equation (2)) pose a challenge in constraining the profile and advection of melt within the partially molten layer. Whereas the bottom boundary condition is straightforward \((\phi = 0, u = 0)\), the top boundary involves interaction with the overlying layer. Percolated melt that reached the top of the partial-melt layer has to be delivered to the planet’s surface to release heat through eruption (O’Reilly & Davies 1981).

The top of the partially molten region is often considered to be a crust–mantle boundary, and the transition from a partial-melt regime to a solidified crust has often been thought of as a “boundary condition” on the former. This, however, is a questionable approach since the two regimes are so different: the partial-melt layer is dominated by viscous flow with pervasive yet microscopically distributed melt, whereas the crust is elastic–plastic with viscoelasticity in its lowermost regions, which is capable of temporal and spatial fluctuations that might include crack formation. In reality, the “boundary” is rather likely to be a transition zone of several kilometers in thickness (Section 4.2), and neither partial melt nor standard crack formation recipes may apply. The perplexing issue of correct petrology remains unsolved as well.

While governing equations in the transition zone remain unresolved, energy balance requires that a large fraction of melt that originated in the partial-melt region solidifies within the transition zone to heat the cold subsiding crust (see Section 4.1 for details). The upwelling melt flux is thus expected to decrease while going through the transition zone, whereas in most of the partially molten regions, the melt flux increases toward the surface to release tidal heating (Figures 2–4). This implies that \(du/dz\) would be 0 in the vicinity of the boundary between the transition zone and the partial-melt layer. Here, the transition zone is defined as a region where the standard percolation recipe does not apply, but it is unclear under what conditions percolative flow becomes inapplicable.

Instead of determining the exact value of \(du/dz\) at the boundary, we assume that the pressure difference between the liquid and solid phases is close to 0. The liquid overpressure is related to the compaction rate as

\[\Delta P = P_{\text{liq}} - P_{\text{sol}} = \zeta \nabla \cdot \mathbf{v}_e = -\zeta \frac{du}{dz} = \tau_0, \quad (6)\]

where the \(z\)-direction is defined positive toward the surface. The top boundary of the partial-melt layer is set to have a small liquid overpressure of \(\tau_0 \sim 1\) MPa, which is typical stress needed to initiate fracture (Sleep 1988). We, however, stress that standard crack formation may not apply, and \(\tau_0\) can potentially take any value around 0 MPa. Values between \(-1\) and \(3\) MPa are tested, but we find that changing \(\tau_0\) by a few MPa has little influence on the overall results. The melt fraction at the top boundary, \(\phi_{\text{top}}\), is set so that the liquid overpressure within the partial-melt layer does not exceed \(\tau_0\).

### 2.3. Plausible Range of Rheological Parameters

Given the uncertainty in the Ionian interior, the expected range of \(\eta_l, \eta_h,\) and \(k_m\) could span a few orders of magnitude.
Solid viscosity $\eta_s$ is known to follow (Hirth & Kohlstedt 1996; Mei & Kohlstedt 2000; Jain et al. 2019)

$$\eta_s = \eta_0 \left( \frac{c_w}{c_w} \right) d^2 \exp \left( \frac{E}{R \left( \frac{1}{T} - \frac{1}{T_0} \right)} \right),$$  

(7)

where $c_w$ shows volatile content, $d$ is typical grain size, $E$ is the activation energy, $R$ is the universal gas constant, and the subscript 0 denotes a reference value. Diffusion creep, which is likely to be the dominant deformation mechanism under a near-solidus temperature, is assumed here, and Equation (7) shows that solid viscosity decreases with a higher volatile content, a higher temperature, and a smaller grain size. Sulfur-rich volcanism indicates that the interior of Io contains more volatile than the terrestrial mantle (e.g., McGrath et al. 2000), and basaltic lava composition implies that the mantle temperature is similar to Earth. Modeling of the eruption temperature also suggests a similar mantle temperature (Keszthelyi et al. 2007), but a lava temperature higher than $\sim$1800 K has been observed on some occasions (Davies et al. 2001; de Kleer et al. 2014), possibly pointing to a hotter interior than Earth. Therefore, the volatile content and interior temperature of Io suggest that the solid matrix of Io is less viscous than the terrestrial mantle. Grain size within Io remains uncertain because efficient grain growth may occur under a low-stress environment, and it is unlikely that the grain size becomes larger by 1 order of magnitude than in the Earth’s interior ($\sim$1 cm). We thus estimate that solid viscosity would not be higher than 100 times the terrestrial value. Rock mechanics and the possibility of small-scale convection beneath oceanic lithosphere predict a value of $10^{19}$ Pa s for Earth (Davaille & Jaupart 1994; Hirth & Kohlstedt 1996; Dumoulin et al. 1999), so we estimate that $\eta_s$ is smaller than $10^{21}$ Pa s on Io.

Melt viscosity follows a similar trend, and $\eta_l$ is smaller under a higher temperature and volatile content. The SiO$_2$ content also affects melt viscosity, with ultramafic magma having a significantly smaller viscosity than rhyolitic. Silicate lava on Io is best explained by basaltic or ultramafic compositions (Mcewen et al. 1998), so we take the viscosity of basaltic melt as a reference value: $\sim 100$ Pa s under $\sim 1300$ K (Giordano et al. 2008). With an abundant amount of sulfur in Io’s magma and a potentially higher temperature, $\eta_l$ is expected to be smaller than the terrestrial value; thus, $\eta_l < 100$ Pa s in Io’s mantle. We note that this maximum value is subject to uncertainty because Io’s interior and melt rheology could be controversial. Sulfur may be mainly stored in a segregated surface layer, and volatile-depleted melt in the interior may be highly viscous. Melt viscosity also increases when small crystals are abundantly suspended in the percolating melt phase. Under the same $\phi$, however, suspended crystals result in a more porous matrix and a higher permeability, so it could cancel out the effect of viscosity increase ($\eta_l/k_m$ in Equation (4)). It is thus not straightforward to determine the upper bound of viscosity, and we may need to consider $\eta_l > 100$ Pa s as a higher-end value. The lowest possible values are not constrained here because, as shown later, a layer with a high melt fraction becomes increasingly unstable under lower melt or solid viscosities. The conclusion of our study thus remains unchanged even for small values of rheological parameters.

Permeability $k_m$ is also a function of grain size $d$, and dimension analysis suggests that $k_m$ is proportional to $d^2$. For a permeability model, a scaling relation $k_m = k_0 \phi^2$ is adopted based on a simple tube model. This is likely to be valid under a high melt fraction of around 0.1–0.2, which is the interest in this work. A permeability model for small porosity is subject to debate, which is applicable to studies of midocean ridges, for example. Smaller values of the permeability prefactor $k_0$ and cubic dependence on melt fraction in the literature correspond to a low melt fraction where the effects of nonuniform tubule
radii and surface tension are pronounced. Such an effect, however, is irrelevant to the melt fraction of interest here. The coefficient $k_0$ is thus approximated by $d^2/32$, and assuming a grain size of 1 mm to 1 cm, $k_0$ is in the range of $3 \times 10^{-8}$–$3 \times 10^{-6}$. Large grain size may increase permeability, but it has the same effect as lowering $\eta_s$. A high-melt-fraction layer would become increasingly unstable, and it would not change the conclusion of this study.

3. Results

We first consider a partially molten layer with a thickness of 400 km, corresponding to the upper half of the Ionian mantle, and different degrees of tidal heating are applied to this layer. For simplicity, tidal heating is uniformly distributed throughout the partial-melt layer. The volumetric heating rate may differ by a factor of $\sim 4$ with depth (Bierson & Nimmo 2016), but the profile of melt fraction does not change significantly as long as
the total amount of tidal heating remains the same (see Section 3.3).

Figure 2 shows a typical mantle structure under $\eta_l/k_0 = 10/2 \times 10^{-5}$ and $\eta_s = 10^{20}$ Pa s. As expected, the degree of melting increases as tidal heating intensifies, and a larger separation flux is observed to allow for efficient heat loss. Whereas melt and solid viscosities increase toward the surface, melt fraction exhibits an oscillatory behavior with a peak near the top of the partial-melt layer (Hewitt & Fowler 2008). This peak arises due to a pressure difference imposed at the top boundary and can be understood from the net mass flux:

$$\nabla \cdot ((1 - \phi) u) \simeq -u \frac{d\phi}{dz} + (1 - \phi) \frac{du}{dz} \geq 0.$$

(8)

Tidal dissipation continuously melts the partially molten layer, so the net mass flux has to be positive at all depths to maintain a steady state. At the same time, cooling by the subsiding crust cools the uppermost region to create small or negative solid overpressure (Section 2.2), resulting in a negative value of the factor $d\phi/dz$. The melt fraction rapidly increases with depth in the subsurface region, and its value becomes higher by a factor of 2–4 compared to the average melt fraction.

The thickness of a high-melt-fraction layer (>20%), however, only exceeds 50 km when tidal dissipation within the partially molten layer is larger than 550 TW (Figure 2(c)). This is significantly larger than the estimated rate of tidal dissipation for Io, where astrometry ($93 \pm 19$ TW; Lainey et al. 2009) and ground-based observations of global thermal emission (105 ± 12 TW; Veeder et al. 1994) both suggest a value of ~100 TW. The latter may not reflect high-latitude heat flow and long-wavelength emission (Stevenson & McNamara 1988), and with the contribution from those sources, the global heat flow would increase to 122 ± 40 TW (Veeder et al. 2004) or potentially to a larger value. Yet, the magnitudes of heat loss and tidal dissipation are expected to be similar considering that Io has maintained thermal equilibrium in a long term, and the rate of tidal dissipation is likely to fall into a range of 100–200 TW. For parameters used in Figure 2, tidal dissipation in Io is therefore insufficient to sustain a high-melt-fraction layer, and a “magmatic sponge” is likely to swiftly separate into two phases. In the following sections, we explore whether such a shortage of tidal heating is ubiquitous among a plausible range of values.

3.1. Dependency on Rheological Parameters

As discussed in Section 2.3, both melt and solid viscosities could vary by a few orders of magnitude with temperature and volatile content, so we explore the relationship between the porosity field and the rheological parameters under a steady state. It is noted that the velocity field is mostly unaffected by a change in the rheological parameters. The transport of latent heat dominates heat loss from the partial-melt layer, and thus under the same rate of tidal dissipation, the melt flux at the top boundary ($=(1 - \phi) u$) becomes similar regardless of the values of $\eta_l$ and $\eta_s$ (Figures 3(a), 4(a)).

When melt is less viscous, the overall degree of melting decreases because melt can escape easily under a smaller porosity (Figure 3(b)). The resistance of solid matrix to compaction is insignificant except in the top boundary region, and the velocity field is determined so that viscous drag mostly balances buoyancy (Figure 3(c)). The velocity profile is similar regardless of $\eta_l$, so the melt fraction decreases with a smaller $\eta_l$ to produce the same level of viscous drag. Under a tidal dissipation rate of 100 TW, a high-melt-fraction layer consistent with the result of Khurana et al. (2011) appears only when $\eta_l > 100$ Pa s.

Solid matrix with a higher bulk viscosity also results in a larger degree of melting (Figure 4(b)). A larger bulk viscosity produces a greater pressure difference between the solid and melt phases (Equation (6), Figure 4(c)), and thus the resistance of solid matrix to compaction becomes significant at a wider depth range (Figure 4(d)). In such a region characterized by compaction resistance, the profile of the melt fraction exhibits oscillation. The wavelength of porosity oscillation is longer under a more viscous solid matrix (Figure 4(c)), which is related to the compaction length (Spiegelman 1993b):

$$\delta_0 = \sqrt{\frac{k_0 \phi_0^2 \cdot \eta_l / \phi_0}{\eta_s}}.$$

(9)

Under a similar velocity profile, the peak melt fraction increases with a longer wavelength, so a higher melt fraction is stable under a more viscous matrix. For parameters used in Figure 4, a 50 km thick layer with $\phi > 20$% appears when $\eta_l$ is higher than $10^{21}$ Pa s. The compaction length reaches ~140 km in such a case, but the actual wavelength is longer than the compaction length under a larger amplitude, so the resistance to compaction plays a dominant role in governing the entire partial-melt layer (Figure 4(d)).

We repeated the calculation for a range of values, and Figure 5 shows that a high-melt-fraction layer with 50 km thickness is stable only under a higher end of parameter ranges. With an estimated rate of tidal dissipation (~100 TW), a high-melt-fraction layer of $\phi > 0.2$ can be sustained only with $\eta_l \geq 100$ Pa s (Figures 5(a)-(c)). Because the contribution from low-temperature heat sources may be underestimated in the observed heat flux (e.g., Stevenson & McNamara 1988; Veeder et al. 2012), we also tested a larger value of 160 TW, but the stability field of the high-melt-fraction layer did not change much (Figure 5(d)). It is noted that some amount of tidal dissipation should be taking place in the overlying crust or in the underlying solid layer (e.g., Peale et al. 1979; Bierson & Nimmo 2016), so only a fraction of the total tidal heating can be used to sustain a magmatic sponge. If $\eta_l < 10^{20}$ and $\eta_s < 10$ Pa s, the rate of tidal dissipation needs to be higher than 500 TW to maintain a layer of $\phi > 0.2$ (Figure 5(b)), and thus a magmatic sponge is unlikely to be stable inside the present-day Io.

3.2. Scaling Law

Although the details of the porosity profile depend on the imposed boundary condition, the characteristic melt fraction within the partial-melt layer can be deduced from a simple scaling that follows Equation (4). In the limit where the viscosity of the solid matrix is negligible, separation flux $u$ asymptotes to

$$u = \frac{k_m}{\eta_l} (1 - \phi) \Delta \rho g,$$

(10)

which is valid under $\eta_s \ll 10^{20}$ Pa s in Figure 4. A typical melt fraction in the upper region of the partial-melt layer, $\phi_0$, can
then be approximated by
\[
\phi_0^2 (1 - \phi_0) \simeq \frac{\eta_l \mu_0}{k_0 \Delta \rho g},
\]  
(11)

using the flux leaving the partial-melt layer from the top boundary, \(u_0\). The surface volcanic flux is estimated as \(\sim 1.5 \text{ cm yr}^{-1}\) from the global heat flux (Section 4.1), but considering the amount of melt solidifying before the eruption, the amount of melt leaving the partial-melt layer, \((1 - \phi_0)u_0\), is likely to be \(\sim 7 \text{ cm yr}^{-1}\) (Section 4.1). Figure 6(a) shows that the melt fraction estimated from the scaling law (\(\phi_0\) in Equation (11)) roughly matches the maximum porosity obtained by solving Equations (4) and (5).

Similarly, in the limit where the bulk viscosity of the solid matrix dominates the system, the following scaling can be obtained by integrating Equation (4):
\[
u \simeq \frac{\Delta \rho g (D/2)^2}{\eta_s/\phi}.
\]
(12)

where \(D\) is the thickness of the partial-melt layer. This scaling relates \(u\) to the average melt fraction over \(D\), \(\phi\), and the average melt fraction \(\bar{\phi}\) can be approximated by
\[
\bar{\phi} \simeq \frac{4u_0 \eta_s}{\Delta \rho g D^2},
\]
(13)

which explains the porosity profile calculated in Figure 4 for high solid viscosity (Figure 6(b)). Therefore, even without solving for differential equations, scaling relations indicate that a layer with a high melt fraction is incompatible with an
observed volcanic flux on Io unless melt or solid viscosity is sufficiently high.

3.3. Dependency on Model Setting

The most unconstrained parameter used in this study may be the pressure difference imposed at the top boundary ($\tau_0$). This value, however, has little influence on the overall results. Increasing the value of liquid overpressure by a factor of 3 increases the maximum melt fraction only by a few percent (Figure 7). On the other hand, when a solid overpressure is set at the top boundary, the maximum value of melt fraction decreases, and a high-melt-fraction layer becomes less stable. Therefore, the conclusion that a magmatic sponge is unstable in Io remains valid regardless of $\tau_0$. Liquid overpressure larger than 15 MPa has been suggested to create a high-melt-fraction layer.
layer beneath the crust (Spencer et al. 2020), but the value of $\tau_0$ is unlikely to deviate largely from 0 MPa (Section 2.2). Also, partial melts can evolve under much lower melt–solid pressure differences (e.g., Sleep 1988; Stevenson 1989), so such a large overpressure may not be sustained in the near-surface region.

Although we assumed that tidal heating is uniformly distributed within the partial-melt layer, the volumetric heating rate may be enhanced at certain depths (Tyler et al. 2015; Bierson & Nimmo 2016). To discuss how a nonuniform distribution may affect the results, we solved for the porosity profile when tidal dissipation is concentrated in a narrower depth range. Figure 8 shows a case when 160 TW of tidal heating is applied to a 200 km thick partially molten layer, which is half of the 400 km assumed in Figures 2–6. With the same total amount of dissipation, the degree of melting increases when heating is concentrated, and such a trend can be understood from the scaling law (Equations (11) and (13)): under the same $\omega_0$, $\phi$ increases with a smaller $D$. The maximum melt fraction, however, does not exceed 0.2 even when 160 TW of dissipation is applied to a 200 km thick layer, and a high-melt-fraction layer compatible with Khurana et al. (2011) requires even more concentrated heating: 160 TW of melting is happening in the solid lower mantle (Bierson & Nimmo 2016), such a condition is unlikely to be met. For Io, a magmatic sponge is thus expected to be unstable under the current dissipation rate regardless of its distribution within the interior.

4. Discussion

4.1. The Fate of Melt Leaving the High-melt-fraction Layer

A fraction of magma leaving the high-melt-fraction layer solidifies before reaching the surface (Spencer et al. 2020), and the degree of solidification is discussed in this section. Regardless of the nature of a high-melt-fraction layer, whether it is a magmatic sponge or a magma ocean, heat transport within the crust would predominantly be achieved through heat pipes (O’Reilly & Davies 1981). Therefore, the energy balance of the near-surface region (crust + transition zone) can be expressed as

$$-k\frac{dT}{dz}\bigg|_{z=z_s} + \rho c_p \Delta T v_s = -k\frac{dT}{dz}\bigg|_{z=z_b} + \rho L \Delta v,$$

(14)

where $\Delta T$ is the temperature difference between the erupting magma and the surface and $v_s(>0)$ is the subsiding velocity of the crust, which should be the same as the surface magma flux (Figure 1(c)). Subscript $s$ denotes the surface, and $b$ describes the base of the near-surface layer, which is equivalent to the top of the partial-melt layer described in Figure 2–4. Conductive heat flux from the partially molten layer is more than 2 orders of magnitude smaller than the latent heat flux, and its magnitude decreases further toward the surface as the temperature gradient asymptotes to zero (O’Reilly & Davies 1981). The amount of melt solidifying within this region, $\Delta v \equiv v_b - v_s$, can thus be estimated as

$$\Delta v = \frac{c_p \Delta T}{L} \approx 3.5 v_s.$$

(15)

With a temperature difference of $\Delta T = 1400$ K, this equation suggests that nearly 3.5 times the surface magma flux solidifies before magma erupts to the surface (Spencer et al. 2020). The latent heat of solidifying magma is used to heat the cold subsiding crust.

It is noted that conductive heat flux at the surface could still be significant in some regions, even though the temperature...
detailed understanding of how melt migrates and solidifies within the crust and the boundary layer, but this rough scaling suggests that the boundary layer is considerably thinner than the radius of Io. At a steady state, the total tidal heating should balance the surface heat flux:

\[ \int_{V_m} \psi dV \approx 4\pi r_i^2 \cdot \rho (L + c_p \Delta T) v_i, \]

where \( V_m \) is the volume of partial-melt layer experiencing tidal deformation. Assuming a total dissipation rate of \( 122 \pm 40 \text{ TW} \) (Veeder et al. 2004), Equation (16) suggests that the globally averaged surface volcanic flux is \( v_i \approx 1.5 \pm 0.5 \text{ cm yr}^{-1} \), and the melt flux leaving the partial-melt layer would be \( v_b \approx 6.8 \pm 2.2 \text{ cm yr}^{-1} \) (Equation (15)).

### 4.2. Characterization of the Subsurface Structure

The solidification of upwelling melt (\( \Delta v \) in Section 4.1) is considered to be happening in a “transition” zone, which connects the crust and the partial-melt layer. The top boundary of the partial-melt layer considered in Section 3 corresponds to the base of this transition zone. The transition zone is cooled from above by the cold subsiding crust, while upwelling melt delivers heat from the deeper region. The heat balance of the boundary layer is thus characterized as

\[ \rho L \nabla \cdot ((1 - \phi) u) \approx k \nabla^2 T (<0), \]

which describes the conversion of latent heat into conductive heat. The thickness of this boundary layer, \( D \), can be estimated using scaling analysis (\( \rho L \Delta u \sim k \Delta T/D \)). The upwelling flux that solidifies before eruption is denoted as \( \Delta U \), and \( \Delta U \) of 5 cm yr\(^{-1} \) is expected for Io from Equation (15). The scaling suggests that \( \Delta U/D \approx 0.5 \text{ K m}^{-1} \), so the layer thickness should only be \( \sim 1 \text{ km} \) when the temperature change across this boundary layer is 500 K. The exact value of \( D \) requires a detailed understanding of how melt migrates and solidifies within the crust and the boundary layer, but this rough scaling shows that the boundary layer is considerably thinner than the radius of Io (Spencer et al. 2020).

The temperature is expected to rapidly drop in a narrow depth range, and with melt fraction quickly decreasing as well, the standard recipe of percolation would cease to apply in the transition zone. How a percolative flow induces cracking and turns into a channelized flow remains elusive and is left for future work. For now, our calculation is based on the fact that the value of liquid overpressure would be around \( \tau_0 \approx 0 \text{ MPa} \) at the top of the partial-melt layer, considering that melt flux (\( u \)) starts to decrease toward the surface in the transition zone (Section 2.2). This approach allows us to solve for the thermal state of the crust and partial-melt layer separately, and calculations in Section 3 would not depend on the details of the transition zone.

### 4.3. The Interior of a Partially Molten Body

Figure 5(a) shows that the amount of tidal heating required to sustain a partial-melt layer increases with the average melt fraction, whereas the rate of tidal dissipation is known to peak at a low melt fraction (\( \phi < 0.05 - 0.1 \); Ross & Schubert 1986; Zahnle et al. 2015). A planetary body with a partially molten interior is thus likely to maintain an equilibrium state with an average melt fraction of \( \phi \sim 0.1 \). When \( \phi \sim 0.1 \), heat loss by percolation is expected to be larger than the degree of tidal dissipation, and the average melt fraction would decrease with time. An excess amount of melt would either escape to the surface or separate from the solid matrix to create a subsurface magma ocean. A rheological model is needed to solve for the detailed structure, but the typical melt fraction of a partially molten layer experiencing extreme tidal heating is likely to fall into the range of \( \phi \sim 0.05 - 0.1 \).

### 5. Conclusions

In this study, we calculated the amount of internal heating necessary to sustain a magmatic sponge: a high degree of melting with interconnected solid. Our results suggest that the presence of such a partially molten layer requires internal heating much larger than the rate of tidal dissipation expected within Io. A high-melt-fraction layer could be stable when the rheological parameter is at the higher end of the estimated range, but such a case is less likely given that Io contains an abundant amount of sulfur in the interior. Because the amount of internal heating is insufficient to maintain a high degree of melting, a magmatic sponge would separate into two phases, creating a subsurface magma ocean. We note that the magma ocean may not necessarily be pure liquid but could contain some amount of crystals. Crystals start to disaggregate once the melt fraction reaches \( \phi \geq 0.4 \), and a partially molten material would behave rheologically as liquid as long as \( \phi \geq 0.4 \) (e.g., Abe 1993; Solomatov & Stevenson 1993). A crystal-rich magma ocean in Io has been supported by petrologic arguments (Keszthelyi et al. 1999).

Governing equations of such a subsurface magma ocean would differ from Equations (4) and (5). Some tidal dissipation may occur in this fluid layer (Tyler et al. 2015), but even without such heating, sufficient heating could be provided from the underlying partial-melt layer (Figure 1(b)). A magma ocean may simply be acting as a layer that transports heat from the partial-melt layer to the crust. Compositions of such a magma ocean and the partial-melt layer are likely to be different, but knowledge of petrology would be required to delineate its detail.

A strong induction signal from Io’s interior has recently been questioned (Blöcker et al. 2018), but if a melt-rich layer is indeed present in the subsurface of Io, such a layer should exist as a magma ocean. Its existence may be confirmed if a Love number around \( \sim 0.5 \) is measured by Juno flybys. If the measurement otherwise suggests a larger value of around \( \sim 0.1 \), a strong induction signal may simply be a result of a plasma interaction in Io’s atmosphere (Blöcker et al. 2018). In such a case, the melt fraction is likely to be smaller than \( \phi < 0.2 \) in the Io’s interior, and a high-melt-fraction layer would be absent within Io.

This work was motivated by NASA’s Juno mission. Y.M. was supported by Stanback Postdoctoral Fellowship from...
Caltech Center for Comparative Planetary Evolution. The authors also thank two anonymous reviewers, whose comments were helpful to improve the clarity of the manuscript.

Appendix

Method Details: Nondimensionalized Equations

Using scales \(z = \delta_0 z^*, u = u_0 u^*, \) and \(\phi = \phi_0 \phi^*, \) Equation (4) is nondimensionalized as follows:

\[
\frac{k_0 \Delta \rho g}{\eta_0 u_0} (1 - \phi_0 \phi^*) + \frac{k_0 \delta_0}{\eta_0 \delta_0^2} \frac{d}{dz^*} \left( \frac{1 - \phi_0 \phi^*}{\phi^*} \frac{du^*}{dz^*} \right) = \frac{u^*}{\phi^*},
\]

(A1)

where \(\zeta_0 = \eta_0 / \phi_0 \) is a reference bulk viscosity at \(\phi = \phi_0.\) By taking \(\delta_0\) as the compaction length (Equation (8)) and \(u_0\) as \((k_0 \Delta \rho g)/\eta_0,\) the equation for momentum conservation can be simplified to (Scott & Stevenson 1984)

\[
(1 - \phi_0 \phi^*) + \frac{d}{dz^*} \left( \frac{1 - \phi_0 \phi^*}{\phi^*} \frac{du^*}{dz^*} \right) = \frac{u^*}{\phi^*}.
\]

(A2)

The conservation of energy is also nondimensionalized as

\[
\nabla^* \cdot \left( (1 - \phi_0 \phi^*) u^* \right) = \frac{\kappa}{\delta_0 u_0} \frac{c_p T_0}{L} \nabla^* T + \frac{\delta_0 \psi}{u_0 \rho L},
\]

(A3)

and these two equations are solved in our model.

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References

Abe, Y. 1993, Litho, 30, 223
Bierson, C. J., & Nimmo, F. 2016, JGRE, 121, 2211
Bloch, A., Saur, J., Roth, L., & Strobel, D. F. 2018, JGRA, 123, 9286

Davaille, A., & Jaupart, C. 1994, JGR, 99, 19853
Davies, A. G., Keszthelyi, L. P., Williams, D. A., et al. 2001, JGR, 106, 33079
Davies, A. G., Wilson, L., Matson, D., et al. 2006, Icar, 184, 460
de Kleer, K., Pater, I. D., Gerard, A., & Adámek, M. 2014, Icar, 242, 352
Dumoulin, C., Doin, M. P., & Fleitout, L. 1999, JGR, 104, 12759
Giordano, D., Russell, J. K., & Dingwell, D. B. 2008, E&PSL, 271, 123
Hewitt, I. J., & Fowler, A. C. 2008, RSPSA, 464, 2467
Hirth, G., & Kohlstedt, D. L. 1996, E&PSL, 144, 93
Jain, C., Korenaga, J., & Karato, S. i. 2019, JGRB, 124, 310
Katz, R. F. 2008, IPE, 49, 2099
Keszthelyi, L., Jaeger, W., Milazzo, M., et al. 2007, Icar, 192, 491
Keszthelyi, L., McEwen, A. S., & Taylor, G. J. 1999, Icar, 141, 415
Khurana, K. K., Jia, X., Kivelson, M. G., et al. 2011, Sci, 332, 1186
Lainey, V., Arlot, J. E., Karatekin, O., & Van Hoolst, T. 2009, Natur, 459, 957
McEwen, A. S., Keszthelyi, L., Spencer, J. R., & Schubert, G. 1998, Sci, 281, 87
McGrath, M. A., Belton, M. J., Spencer, J. R., & Sartoretti, P. 2000, Icar, 146, 476
Mei, S., & Kohlstedt, D. L. 2000, JGR, 105, 21457
Moore, W. B. 2003, JGRE, 108, E85096
O’Reilly, T. C., & Davies, G. F. 1981, GeoRL, 8, 313
Peale, S. J., Cassen, P. M., & Reynolds, R. T. 1979, Sci, 203, 892
Ross, M., & Schubert, G. 1986, JGR, 91, 447
Scott, D. R., & Stevenson, D. J. 1984, GeoRL, 11, 1161
Sleep, N. H. 1988, JGR, 93, 10255
Solomatov, V. S., & Stevenson, D. J. 1993, JGR, 98, 5375
Spencer, D. C., Katz, R. F., & Hewitt, I. J. 2020, JGRE, 125, e06443
Spiegelman, M. 1993a, JFM, 17, 247
Spiegelman, M. 1993b, JFM, 427, 39
Stevenson, D. J. 1989, GeoRL, 16, 1067
Stevenson, D. J., & Mcnamara, S. C. 1988, GeoRL, 15, 1455
Stevenson, D. J., & Scott, D. R. 1991, AnRFM, 23, 305
Tackley, P. J., Schubert, G., Glatzmaier, G. A., et al. 2001, Icar, 149, 79
Tyler, R. H., Henning, W. G., & Hamilton, C. W. 2015, ApJS, 218, 22
Van Hoolst, T., Baland, R. M., Trinh, A., Yseboodt, M., & Nimmo, F. 2020, JGRE, 125, e06473
Veer, G. J., Davies, A. G., Matson, D. L., et al. 2012, Icar, 219, 701
Veer, G. J., Matson, D. L., Johnson, T. V., Davies, A. G., & Blaney, D. L. 2004, Icar, 169, 264
Veer, G. J., Matson, L., Johnson, V., Blaney, D. L., & Goguen, J. D. 1994, JGR, 99, 17095
Zahnle, K. J., Lupu, D., Dobrovolskis, A., & Sleep, N. H. 2015, E&PSL, 427, 74