The Internal Structure of Io

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Gravitational interactions with Jupiter drive Io's internal heating. Image of Io passing in front of Jupiter, acquired by the Voyager 1 spacecraft in 1979.

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Io is a differentiated body with a silicate crust and mantle, and an iron-rich core. However, its internal structure, especially that of its mantle, differs from that of other terrestrial bodies, as a result of the intense heat supply by tidal dissipation. The amount and distribution of melt in Io's interior strongly depend on the composition, as well as the heat and mass transport mechanisms operating at depth and in the near-surface. This article discusses melting processes and the mechanisms of magma segregation inside Io, informed by Earth-based observations and spacecraft measurements, as well as thermo-chemical and thermo-physical modeling.

KEYWORDS: interior structure; partial melt; melt segregation; magma ocean; tidal dissipation; asthenosphere

INTRODUCTION

Active volcanism on Io is due to tidal dissipation and has led to considerable interest in understanding its internal structure. In particular, the nature and depth of an internal melt layer has motivated numerous investigations. Like Earth, Io has a layered structure consisting of an iron-rich core, a silicate mantle, and a crust, inferred from its gravitational field and rotational state. The low-degree gravity harmonics of Io, determined from radio-tracking the *Galileo* spacecraft, indicate that Io is in hydrostatic equilibrium. Under this assumption and given the planetary equatorial and polar radii, mass, and rotation rate, a dimensionless moment of inertia (MoI) factor with a value of 0.378 was derived (Schubert et al. 2004). The difference in the MoI from a value of 0.4 for a sphere with a homogeneous density indicates that Io has a core that is denser than its mantle. Based on this observation, core radius estimates range from 650 to 950 km, with core compositions ranging from dense pure iron to low-density iron–sulfur, respectively, and with mantle density being consistent with an olivine-dominated mineralogy (e.g., Sohl et al. 2002; see FIG. 1). Io's core is most likely completely liquid, which can be inferred from the strong melt production in the mantle resulting from tidal dissipation, the higher melting temperatures of silicate compared with iron, and considering that the core is unlikely to be significantly cooler than the mantle.

How the interior is further structured is widely debated (e.g., Tylor et al. 2015; Spencer et al. 2020b), with the major issues being the distribution and amount of melt in the mantle, the possible stratification of the mantle, the crustal thickness, and how the internal structure has evolved with time. Depending on the distribution of tidal heat and the strength of tidal dissipation, there will likely be a consistent distribution of melt. The accumulation of melt in turn affects the mechanical properties and thus the tidal response of the mantle. Convective heat transport and melt extraction control the rate of cooling and determine whether and where melt can accumulate in the interior. Indirect observations, such as the distribution of the surface heat flux and volcanic activity (e.g., Kirchoff et al. 2011; Hamilton et al. 2013), the eruption temperature (e.g., Davies et al. 2001), and the melt distribution in the mantle derived from magnetic induction data (Khurana et al. 2011)—although current results are controversial (Blöcker et al. 2018)—in combination with thermo-chemical and thermo-physical modeling (e.g., Spencer et al. 2020a; Kervazo et al. 2022), can provide valuable information about Io's interior.

CONSTRAINTS ON MELT DISTRIBUTION IN IO FROM OBSERVATIONS OF HEAT FLUX

Estimates of Io's global mean surface heat flux range from 1.5 to 4.0 $W \cdot m^{-2}$ (e.g., Moore et al. 2007), with more recent observations supporting a value of 2.24 ± 0.45 $W \cdot m^{-2}$ (Lainey et al. 2009). These values correspond to an estimated global heat output between 65 and 125 TW. Io's thermal emission, combined with observations of its high-temperature eruptions, implies that the moon includes silicate melt at depth (Keszthelyi et al. 2007).

Major expressions of volcanism on Io include paterae, which are interpreted to be volcano-tectonic depressions that are analogous to caldera on Earth, and hotspots, which correspond to high-temperature anomalies, associated primarily with active explosive eruptions, lava lakes, and lava flows. Globally, there are over 250 hotspots on Io and their temperature and spatial distribution provide information about Io's interior (Davies and Vorburger 2022 this issue). Keszthelyi et al. (2007) estimated that the upper mantle should have a melt fraction of about 20%–30% if there is widespread ultramafic volcanism. A high melt fraction within Io is consistent with models describing the release of interior heat via melt extraction (e.g., Bierson and Nimmo 2016) and with *Galileo* magnetic induction measurements (Khurana et al. 2011). However, the total amount of heat produced by tidal dissipation within Io depends on the distribution of heat and melt, which in turn

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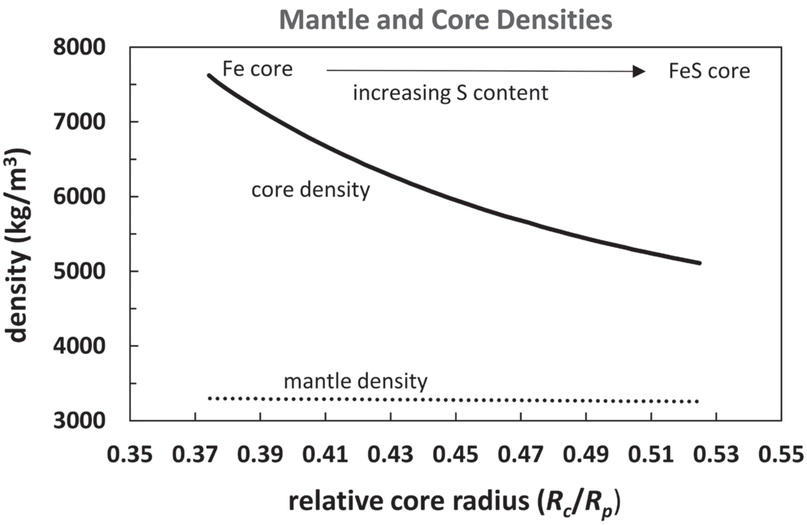

FIGURE 1 Mantle and core densities of Io as a function of relative core radius. The models satisfy the mass and moment-of-inertia constraints imposed by the *Galileo* radio tracking data and are consistent with an olivine silicate mantle mineralogy and Fe–FeS core. A pure FeS core contains about 37% S, and a eutectic Fe–FeS core contains about 25% S. The relative core radius is the ratio between the core radius R_c and the satellite radius R_p. ADAPTED FROM SOHL ET AL. (2002).

affect the temperature-dependent mechanical properties. There are also complex feedbacks involving the mechanical properties, processes of tidal dissipation, melt generation, melt migration, and lithospheric loading and subsidence, as well as the states of stress within the crust, which can initiate deep faults and facilitate the ascent of magma to the surface (Keszthelyi et al. 2022 this issue).

Two endmember models are discussed to describe the internal structure of Io: either the entire mantle is rigid and has relatively little melt (e.g., in an asthenosphere), or it contains a high-melt fraction (i.e., fluid) layer, possibly decoupling the lithosphere from the deeper interior (FIG. 2). This distinction is based on the concept of a rheologically critical melt fraction (RCMF), which is associated with a sharp transition from solid to liquid behavior and typically occurs at melt fractions between 25% and 40% (e.g., Kervazo et al. 2022). At low melt fractions, the material is best described as a solid matrix with fluid pores. Its deformation is dominated by solid-state rheology. At large melt fractions (or low crystal fractions), the material loses shear strength and tends to behave like a liquid. In the case of a solid structure, solid dissipation predominates as a heating mechanism, while the presence of a fluid layer means that fluid dissipation is possible.

INSIGHTS FROM OBSERVATIONS AND MODELING

Heat Distribution and Partial Melting

The surface heat flow pattern and distribution of volcanoes can be used to understand which of these models best represents Io's interior. This information is obtained by analyzing data collected by *Galileo*'s near infrared mapping spectrometer (NIMS) and photopolarimeter–radiometer (PPR) and infrared radiometric data of various wavelengths from large ground-based telescopes. A global distribution of volcanism on Io has been revealed, with volcanic centers on Io broadly distributed but showing a concentration at low latitudes (e.g., Kirchoff et al. 2011; Hamilton et al. 2013). In particular, a spherical-harmonic analysis identified statistically significant clustering at degrees 2 and 6 (Kirchoff et al. 2011). In addition, results using a distance-based cluster approach show a near-equatorial concentration of volcanism with a 30°–60° eastward shift of the volcanic centers relative to sub-Jovian or anti-Jovian points (Hamilton et al. 2013). However, because of the observational conditions of the *Voyager* and *Galileo* missions, only limited polar coverage is available.

The distribution of volcanoes can then be used to test models for Io's interior structure, because the relative strength of tidal heating in an asthenosphere or magma ocean, and deep mantle, strongly influences the expected heat flow patterns. This relationship assumes that the radially integrated dissipated energy inside the planet is perceived at the surface as volcanic heat transported through the lithosphere. Typically, the mechanical properties of the lithosphere that can redirect the upward flow are not considered in convection and/or melt migration models. In solid mantle models, where solid dissipation dominates, a further distinction is made concerning whether an asthenosphere is present or the mantle rheology (i.e., the flow and deformation of mantle materials) is relatively homogeneous. In the former, dissipation takes place in the asthenosphere, while in the latter, dissipation is essentially located in the deep mantle. In representative viscoelastic models (e.g., Beuthe 2013), the radially integrated tidal heat production yields enhanced polar heat flow in the deep-mantle case with absolute minima occurring at the sub-Jovian (0° N, 0° W) and anti-Jovian points (0° N, 180° W). In the asthenospheric case, enhanced heat flow occurs at lower latitudes with primary maxima at sub-Jovian and anti-Jovian longitudes (0° W and 180° W), secondary maxima at 270° W and 90° W, and minima at the poles.

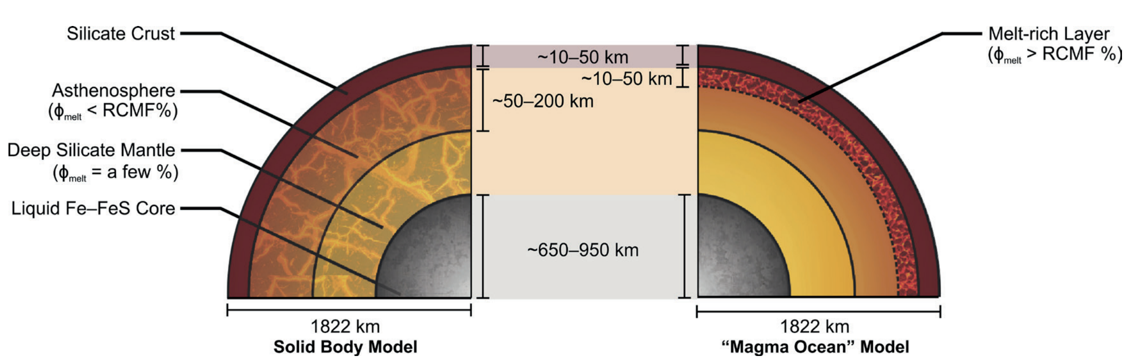

FIGURE 2 Schematic representations of Io's interior structure for two endmember models of the interior structure assuming a solid mantle with partial melting (**LEFT**) or a melt-rich layer (magma ocean) (**RIGHT**). Here, Φ_{melt} is the melt fraction, and RCMF is the rheological critical melt fraction. Values in kilometers reflect present-day estimates of layer thicknesses. The layers are not drawn to scale.

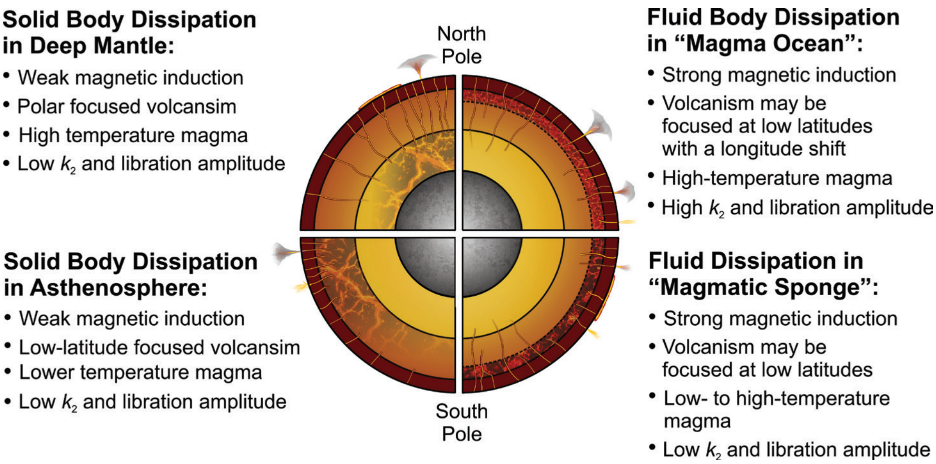

FIGURE 3 Schematic representations of Io’s interior structure. Layer thicknesses are arbitrary. Deep (TOP) and shallow mantle (BOTTOM) tidal dissipation scenarios are considered. A solid interior (LEFT) is compared with scenarios where dissipation processes occur through either a magma ocean or a globally extended layer with a high melt fraction (i.e., a magmatic sponge; RIGHT). AFTER DE KLEER ET AL. (2019).

The observed surface heat flux with enhanced values at lower latitudes generally supports asthenospheric-dominated tidal heating models (FIG. 3, BOTTOM LEFT). For the concentration of heat in the low-latitude region, the degree of melting in the asthenosphere must not be too small and the thickness of this layer must be sufficiently thick. For a melting degree of 20%, the asthenosphere must be thicker than 200 km; and for a higher melting degree of 25%, it must be thicker than 100 km (Bierson and Nimmo 2016). Melt accumulation in a layer at the top of the mantle beneath the lithosphere is generally expected as a result of decompression melting. This is also supported by work on melt transport using a one-dimensional, two-phase flow approach (Spencer et al. 2020a; FIG. 4), showing the formation of a layer with a high degree of melting beneath the crust, especially when a large excess fluid pressure is required to inject dikes into the crust. The overpressure may be caused by subsidence of the continuously forming crust. Whether the melt fraction in the layer is below or above the RCMF is controversial. Assuming a balance between tidal heat production and heat loss resulting from the rapid rise of magma, the degree of melting is below (or slightly above) the RCMF, and in the range of 5%–30% (e.g., Moore et al. 2007; Bierson and Nimmo 2016; Spencer et al. 2020a; Steinke et al. 2020). However, the effectiveness of solid-body tidal dissipation breaks down when the degree of melting exceeds this rheological critical melt fraction, and a higher degree of melting requires a different heating mechanism, such as tidal dissipation in a fluid generated by excited waves (Tyler et al. 2015; Hay et al. 2020).

A longitudinal offset between observed concentrations of volcanism and predicted heat flows is difficult to explain solely by combining the two endmember solid-body dissipation models (i.e., a mixture of heating in the asthenosphere and in the deep mantle). Modifying the proportion of tidal heading between these two zones does not affect the overall pattern, only the locations of the primary (and secondary) maxima based on an increase in heat flux at the poles with increasing contribution from the deep mantle (e.g., Beuthe 2013). Similarly, thermal convection in Io’s mantle does not fundamentally change this pattern in the heat flux distribution and occurrence of volcanism (e.g., Steinke et al. 2020). This is a consequence of the different time scales associated with these processes—melt migration acts on a much shorter time scale than solid-state convection. When solid-state convection becomes more vigorous (i.e., the Rayleigh number, which describes the strength of convection, becomes larger with decreasing viscosity as a result of increasing temperatures and higher degrees of melting), horizontal flows smooth out the lateral temperature variations and thus the lateral variations in the melt content. Interestingly, the heat production maxima along the equator are shifted by 25°–30° relative to the sub-Jovian and anti-Jovian points when bulk dissipation (i.e., dissipation resulting from volumetric changes) is also considered, but these models also show additional maxima that are not observed (Kervazo et al. 2022).

The apparent eastward offset of the observed volcanism patterns relative to the tidal axis could also be explained by fluid dissipation in a magma ocean. This has been demonstrated in models that examine the tidal response of a fluid layer decoupled from the lithosphere and underlying mantle. Excited by eccentricity-driven motions, a shift in surface heat flux at low latitude can be observed (Tyler et al. 2015). Other fluid tidal models, driven by interactions with Jupiter and Europa, predict no offset in surface heat flux with longitude (Hay et al. 2020). Furthermore, most models have very low surface heat flux at high latitudes and are difficult to reconcile with the occurrence of volcanoes in this region. This may require an additional component of solid tidal heating in the deep mantle. Interestingly, the *Galileo* PPR data show anomalously warm nighttime temperatures at the south pole, about 15 K higher than expected (Rathbun et al. 2004), and further support a contribution by deep-mantle heating in addition to dissipation in a shallow layer. Alternatively, the elevated temperatures at the poles may be caused by a surface unit with high thermal inertia.

An important caveat of the abovementioned fluid dissipation models is that their melt configurations assume a globally extensive region with a high degree of interconnected melt (FIG. 3, RIGHT), which suggests that volcanic eruptions can occur anywhere. Properties of the lithosphere (e.g., local zones of weakness) may play the most important role in the spatial distribution of volcanism, rather than the location of the most heat dissipation in the interior, contrary to the original assumption. However, tidal heating, melt production, eruptions, and subsidence can lead to feedbacks that may promote ongoing volcanism by creating zones of weakness in the lithosphere where melt production is the most prodigious.

Probing Io’s Interior using Electrical Conductivity from Magnetic Induction Measurements

A high degree of melting below the lithosphere is also suggested by magnetometer data collected by the *Galileo* spacecraft (Khurana et al. 2011). Jupiter’s time-varying magnetic field induces a magnetic field in the conducting layer, the strength of which depends on the electrical conductivity of the rock (FIG. 5). Electrical studies in the

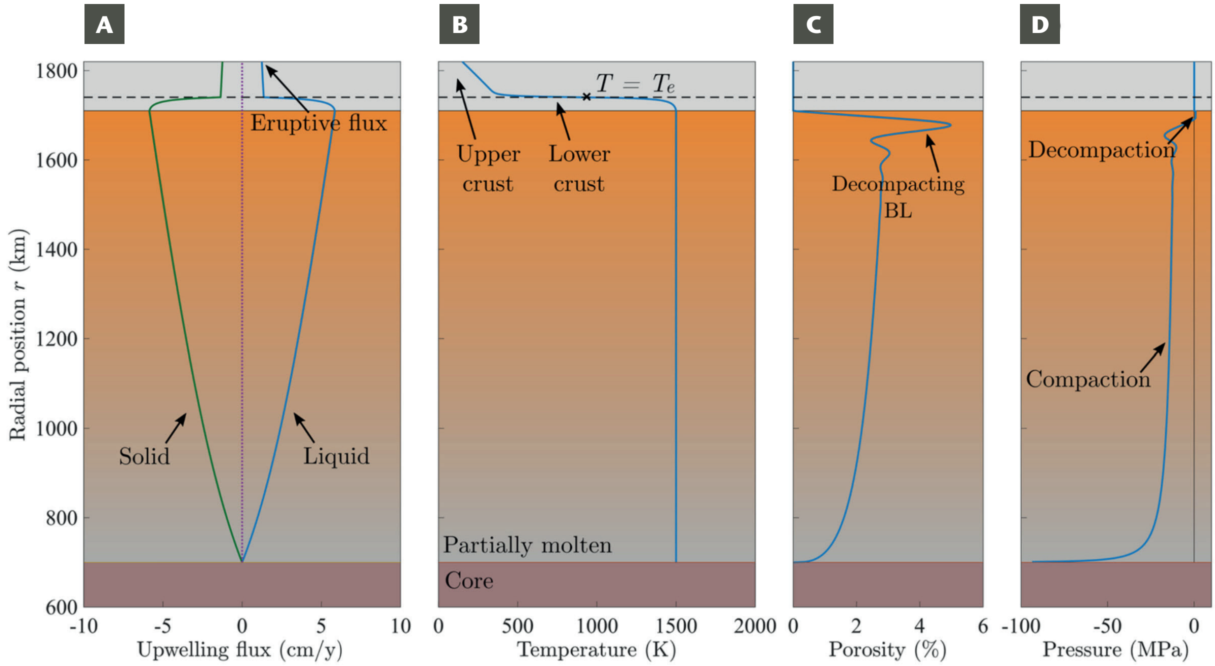

Figure 4 Representative solution of a two-phase model of Io's interior in steady state with tidal heating. (**A**) Upwelling magma is replaced by downwelling solid. (**B**) The lower crust (temperature $T > T_e$ (elastic limit temperature)) is heated by magmatic intrusions (emplacement), but the upper crust ($T < T_e$) is cold as a result of the downwelling cold surface material. (**C**) Melt fractions are low throughout the mantle but increase in a thin boundary layer on the order of 50 km thick. (**D**) Compaction occurs throughout most of the mantle because the liquid is subjected to low pressure. However, beneath the crust, the pressure $P \cong P_c = 0.8$ MPa, where P_c is the critical compaction pressure that characterizes the strength of the downwelling crust, and the downwelling solid is "decompacted" by the liquid pressure. Higher P_c values cause the material downwelling from the crust to decompact more rapidly, leading to the accumulation of even larger amounts of melt. The elastic thickness is 80 km and the eruption rate is 1.1 cm/y, with 99.5% of the heat transported through the surface being volcanic. Figure from Spencer et al. (2020a).

laboratory have shown that rock conductivity depends significantly on temperature, melt fraction, and composition (e.g., Pommier and Garnero 2014 and references therein). The electromagnetic sounding technique uses the electrical response of Io's interior to derive the temperature and estimate the extent of melting at depth.

The *Galileo* magnetic field data from four different close Io encounters show that the induced field is global, dipolar, and almost out of phase with the inducing field. A completely solid mantle provides an inadequate fit to these observations (Khurana et al. 2011) (Fig. 5). Instead, the field data require a strongly electrically conductive layer, such as a global subsurface magma ocean. The best fit is obtained with a melt layer that starts at ~50 km below the surface, has a thickness of >50 km, and contains an interconnected melt fraction of ≥20%. The data, however, cannot distinguish whether this layer is a magma ocean behaving like a continuous fluid decoupling the shell from the interior (Fig. 3, top right), or whether it is a melt-rich "magmatic sponge" with an interconnected silicate melt in an interconnected silicate solid matrix (Fig. 3, bottom right). The reason for this ambiguity is that the induction from Io's interior at the synodic rotation period of Jupiter (13 h), which has a skin depth of <50 km (for ≥20% melt fraction), is unable to fully penetrate the melt layer and, therefore, the signal response saturates to its maximum value. Additionally, strong magnetic induction is possible even with a melting degree lower than 20% if the conductivity is increased by dissolved volatiles (e.g., Pommier et al. 2008). Silicate melts rich in sulfur-containing volatiles could play an important role in Io. However, further laboratory studies are needed to understand the effect of sulfur on the electrical properties of rocks and melts. Note that the requirement of a strong electrically conducting layer to explain the *Galileo* data has been questioned. Some of the measured perturbations could also result from plasma interactions with a highly asymmetric atmosphere (Blöcker et al. 2018), and very high melt fractions may also be at odds with the phase of the auroral spot oscillations observed with the *Hubble Space Telescope* (Roth et al. 2017). Ŝebek et al. (2019) revisited Io's interaction with the Jovian plasma and concluded that a strong dipolar induced field is indeed required to explain the observed data. Nevertheless, further induction measurements from future missions are desirable, especially at other longer wave periods such as the orbital period of Io (42.4 h), which is able to penetrate deeper into the melt layer and does not create a saturated induction response. Multiple-frequency soundings at three or more frequencies would also allow unique determinations of the thickness of the overlying crustal layer, the thickness of the melt layer, and the conductivity of the melt layer (de Kleer et al. 2019).

Lava and Magma Temperatures

Another piece of evidence about the thermo-rheological state of Io's interior comes from the temperature of the erupted magma. The eruption temperature provides information about the composition of the magma as well as its temperature and the degree of melting. If basaltic magma is erupting, then we would expect a maximum temperature of about 1475 K, the liquidus temperature of basalt. Similarly, if ultramafic lavas are erupting, we would expect a maximum temperature of about 1900 K. Estimates of Io's eruption temperatures have been derived from data collected using the *Galileo* Solid-State Imaging (SSI) experiment in combination with NIMS measurements and lava-cooling models. Such cooling models require an accurate understanding of the eruption style, and hence the dynamics of the lava. Eruption temperatures of ~1400 K and even extreme temperatures of 1900 K were determined

(Davies et al. 2001). However, these very high temperatures may have been overestimated because of the simplified models of the SSI scan platform motion and do not consider the effects of viscous dissipation during magma ascent. Accounting for these issues lowers the maximum temperature estimates to ~1613 K (Keszthelyi et al. 2007). High eruption temperatures are indicative of a basaltic or even ultramafic magma composition (e.g., Davies et al. 2001) and a high degree of melting of the silicate mantle. Ultramafic compositions could be consistent with a magma ocean. Associated fluid-body dissipation could potentially sustain these high melt temperatures.

Alternatively, high eruption temperatures can be explained by melt-bearing regions made of a more refractory material than a typical mantle rock (e.g., peridotite) with a high melting temperature. In this case, a high degree of melting is not necessarily required. This scenario was obtained using melt transport models that assume homogeneous solid-body tidal dissipation and track melt composition as well as mantle depletion (Spencer et al. 2020b). These models also imply that Io's mantle may be stratified in composition and that magma forms in both the upper and lower mantle. The segregation of magma then leads to the formation of a lower refractory mantle, while the upper mantle and crust consist of more fusible material. The formation of this mantle structure, in equilibrium with the tidal heat dissipation, occurs on time scales ranging from tens of millions to hundreds of millions of years. The high-temperature eruptions observed on Io can arise when melt from the lower mantle can reach the surface rapidly without reaching thermal equilibration (e.g., through magma plumes that form deep in the mantle, possibly at the core–mantle boundary and rise adiabatically toward the crust). It is unclear whether this chemical stratification remains stable because thermal and chemical convection, which has been neglected in these models, leads to the mixing of the two mantle reservoirs. However, even if mixing occurs, there may still be localized areas in the lower mantle that are refractory and, thus, strongly depleted in crustal components.

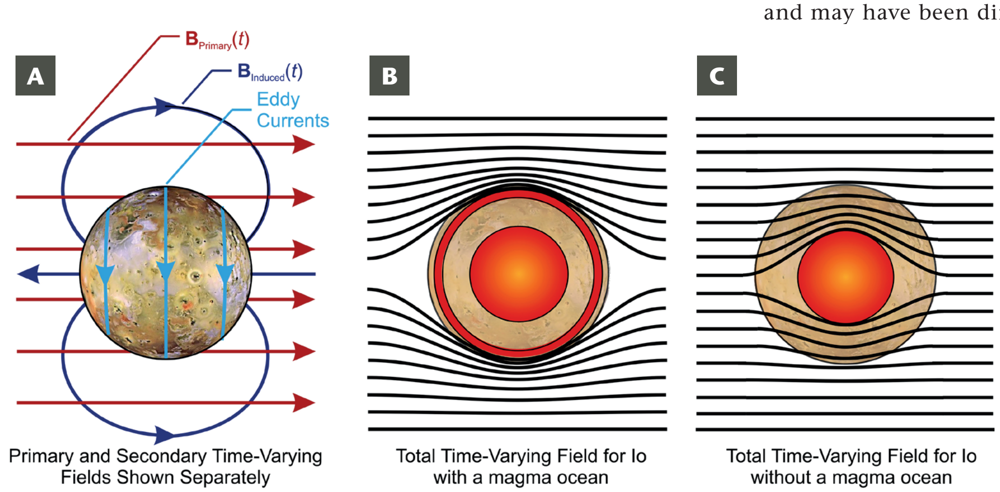

Figure 5 (**A**) In response to a time-varying primary field (red lines with red arrows), eddy currents flow on the surface of the conductor (cyan curves) creating a dipolar induced field (blue lines) that shields the conductor's interior from the primary field. (**B**) The time-varying primary and induced fields summed together (black curves) avoid the interior of the conductor, in this case, the magma ocean. (**C**) The expected external perturbation field from a conducting core would be much weaker than that expected from a magma ocean because the induction signal from a core has to travel out to a greater distance.

Eruption Rates, Surface Topography, and Io's Interior

The melt that rises from the mantle to the surface and forms Io's crust provides information about its internal processes. Io's crust is continuously resurfaced by strong mafic/ultramafic volcanism at rates between 0.1 and 1 cm/y. The minimum value is based on the paucity of impact craters observed on Io's surface (Blaney et al. 1995), whereas the maximum estimate is derived from the observed heat flux. To explain the observed heat flow, Moore et al. (2007) suggested that Io must be covered everywhere with an average of about 1 cm of lava per year. Because of these high crustal formation rates and associated high temperatures, the thickness of the crust, unlike other differentiated planets or moons, is limited by its melting temperature and is continuously recycled. Therefore, not only the melt production rate in the mantle, but also the temperature distribution in the crust and feedbacks between subsidence and the stress state within the lithosphere, play an important role in determining the crustal thickness. The crustal temperature distribution can be indirectly inferred from the observed topography, which shows high and steep mountains with heights of up to 17 km. To support Io's topography, the thermal lithosphere, which for the most part consists of a rigid layer, must be at least 12 km thick, with maximum estimates as high as 100 km, assuming a subsidence rate of 1 cm/y (Jaeger et al. 2003). Within Io, the base of the crust may coincide with the base of the lithosphere if its thickness is controlled by melting at the top of the asthenosphere. However, intrusions into the base of the crust could shift the elastic limit temperature to shallower depths (Fig. 4). Magma segregation models using two-phase flow show that the steady-state crustal thickness would be too large as a result of effective cooling if all tidal heat were removed by extrusive volcanism, but the thermal input of heat from freezing magmatic intrusions into the lower crust could sufficiently raise the temperature of the cold, subsiding crust to keep the lithospheric thickness within the observed limits (Spencer et al. 2020a).

EVOLUTION OF IO'S INTERIOR STRUCTURE

The internal structure of Io probably changed over time and may have been different during Io's past than from what it is today. During the early evolution of Io, the first broad-scale differentiation likely took place in the form of core formation. The rapid separation of dense iron-rich material and less dense silicate, typical of terrestrial planets, indicates that a large part of the interior had to be molten—the necessary energy supply came from accretion, especially with large impacts at the end of the planet's formation. Whether impacts on Io or tidal dissipation helped to sufficiently melt the interior, or core formation took longer because of the missing energy required to melt the interior, is unknown. If Io has a very sulfur-rich core, as is often suggested, core formation would be facilitated by the low-melting temperature of FeS, compared with that of Fe, and melt percolation (Sohl et al. 2002).

The heating rates and associated energy required for melting and differentiation evolve with Io's orbit because they depend on both Io's distance from Jupiter and the eccentricity of Io's orbit. Thus, the coupling of the thermal evolution and resonant orbital interactions of Io, Europa, and Ganymede plays an important role in how the strength of the tidal dissipation, and subsequently the temperature distribution and internal structure, has evolved. For example, it is not clear at present whether the tidal and melting processes described above are currently in a state of equilibrium or whether tidal dissipation rates and volcanic activity vary episodically. It is possible that Io is currently migrating toward Jupiter (Lainey et al. 2009) as a result of tidal heating removing energy from the orbit. When Io's eccentricity is damped, tidal heating decreases and the interior cools and is even less dissipative. Io will then move outward again into resonance with Europa, increasing its eccentricity and tidal strength. Coupled models of thermal and orbital evolution suggest oscillation periods on the order of 100 My (Hussmann and Spohn 2004) with strong implications for volcanic activity and, thus, the internal structure of Io. At times when tidal dissipation does not provide significant heat and melting, the degree of melting in the mantle will be strongly reduced and any chemical layering of the mantle may be mixed more efficiently by thermal convection. However, it is not necessarily expected that volcanic activity will cease completely because of the continuing decay of radioactive elements and the reduced, but still significant, tidal heating (Renaud and Henning 2018).

FUTURE RESEARCH DIRECTIONS

The primary means of testing models of Io's interior come from examinations of its surface expressions of volcanism (e.g., the distribution of volcanoes, patterns of thermal emission, magma temperature, composition). These observations may also be supported by improved measurements of tidal deformation from spacecraft tracking, altimetry, and high-resolution imaging to determine the diurnal tidal Love numbers k_2, h_2, and l_2. These parameters quantify the gravitational potential and the surface radial and horizontal displacements associated with tidal motions, respectively. Improved constraints on these parameters can fill a critical knowledge gap and therefore improve our understanding of Io's interior structure (Kervazo et al. 2022). Finally, future magnetic multi-frequency induction measurements, which are sensitive to the location and thickness of the high-conductivity melt layer, can provide important information on the melt fraction and composition of the lower mantle and should be a priority for future spacecraft observations of Io.

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