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Key Points:

- We use Bayesian inference to constrain Io's interior with gravity and volcanic observations
- Io has a 200 ± 100 km low-viscosity ($10^{13 \pm 0.9}$ Pa s) asthenosphere
- The inferred low viscosity indicates Io has a partially molten layer with a melt fraction of $>30\%$

Supporting Information:

Supporting Information may be found in the online version of this article.

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Inferring Io's Internal Properties by Combining Gravity Measurements and Volcanic Activity

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Abstract With its extreme geological activity, Io represents an archetype for tidally heated exoplanets/moons and provides insight into early stages of terrestrial planet evolution. Decades of ground-based observations and multiple space missions have refined our understanding of Io, yet it remains debated where inside Io the tidal heating takes place. Here, we synthesize several independent data sets in a Bayesian approach to provide an updated view of Io's interior. Tidal response data constrain the bulk mechanical properties of the silicate envelope but do not resolve the radial profile. We propose adding constraints based on the observed distribution of volcanic activity, thereby introducing sensitivity to the radial viscosity profile. The combined observations are consistent with the majority of the dissipation occurring in a low-viscosity asthenosphere approximately 200 km thick, with a maximum viscosity of $\sim 10^{14}$ Pa s, which is also valid under a more conservative estimate of spatially uniform heat flux. This low viscosity implies either enhanced viscous dissipation under Io's conditions due to poorly understood rheological laws, an asthenosphere near the disaggregation threshold, corresponding to a melt fraction of $>30\%$, or a combination of both. Together, these results provide a coherent interior structure and thermal state that can be used in future Io studies.

Plain Language Summary Io, one of Jupiter's moons, exhibits extreme volcanic activity driven by tidal heating, a process where the gravitational pull of Jupiter continuously squeezes and flexes Io's interior, generating enormous amounts of heat. Despite decades of observations from telescopes and spacecraft, it is still unknown where inside Io most of this heat is generated. In this study, we combined several different types of measurements, such as data on how Io deforms and maps of where volcanoes are located on its surface, and analyzed them together using a statistical method. Our results suggest that most of the heat is produced in a roughly 200 km thick layer, underneath Io's outer crust. This layer appears to be exceptionally soft and weak, which may indicate either that rock behaves differently under Io's extreme conditions, or that this layer contains a very high proportion of molten rock (more than 30%). Together, these findings offer a consistent picture of Io's internal structure and heat budget, and provide a foundation for future research.

1. Introduction

Besides the Moon, one could argue that Io is the most observed solar system moon. Since Voyager captured the first glimpse of extraterrestrial volcanism more than 40 years ago (Morabito et al., 1979), five probes have visited the Jovian moon (Anderson et al., 2001; Lopes et al., 2004; Morabito et al., 1979; Mura et al., 2020; Park, Jacobson, et al., 2025; Porco et al., 2003; Tsang et al., 2014), and extensive Earth-based observational campaigns have been conducted (de Kleer & de Pater, 2016b; de Kleer, McEwen, et al., 2019; de Kleer, de Pater, et al., 2019; Veeder et al., 1994). These observations have revealed that Io's volcanism is driven by friction within Io's interior, a consequence of its continuous deformation due to the strong tidal forces that Io experiences (Peale et al., 1979; Segatz et al., 1988). However, where in Io's interior this heat is generated, is still debated (Aygün & Čadež, 2024; Davies et al., 2015; Davies, Perry, Williams, Veeder, & Nelson, 2024; Hamilton et al., 2013; Pettine et al., 2024; Steinke, van Slidregt, et al., 2020; Tackley et al., 2001; Tyler et al., 2015).

Observations of Io's static gravitational field indicate that Io must be differentiated into a large metallic core—between 0.35 and 0.5 times Io's radius (Moore et al., 2007)—overlaid by a rocky envelope. Joint analysis of Galileo and Juno data revealed tidally-driven changes in the moon's gravity field (Park, Jacobson, et al., 2025). The tidally driven changes are smaller than those expected for an Io with a fully molten shallow magma ocean (Bierson & Nimmo, 2016; Kervazo et al., 2022; Park, Jacobson, et al., 2025). Besides precluding a shallow, global magma ocean, the measurement of $\Re\{k_2\}$ constrains the mean shear modulus of Io. Changes in Io's orbit inferred from astrometric observations starting at the end of the 19th century have allowed the determination of the amount

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of tidal dissipation within Io, which is directly linked to the tidal response phase-lag ($\Im\{k_2\}$) (Lainey et al., 2009). The average tidal dissipation of $\sim 2.4 \text{ W m}^{-2}$, bounds Io's mean viscosity and melt content (depending on the rheological model used), but cannot resolve the radial viscosity and melt profile (Bierson & Nimmo, 2016; Kervazo et al., 2022).

The heat flux inferred from astrometry is generally consistent with independent, Earth-based infrared observations (de Kleer, McEwen, et al., 2019; de Kleer, de Pater, et al., 2019; Lainey et al., 2009; Veeder et al., 1994). The Earth-based observations, together with in situ ones, provide a picture of the geographical distribution of Io's volcanoes (e.g., Davies et al., 2015; Hamilton et al., 2013; Steinke, van Sliedregt, et al., 2020). While interpreting these observations is challenging (see Section 2), they are believed to provide a proxy for heat dissipation, assuming that it is similar to the radially integrated tidal dissipation pattern (i.e., negligible lateral heat transport). This would make the surface heat flux intimately linked to the radial rheology profile (i.e., viscosity and shear modulus) (Beuthe, 2013; Segatz et al., 1988). For example, a low-viscosity asthenosphere, and thus greater heating, leads to a higher heat flux at the equator than in the polar regions. Similarly, a rheologically homogeneous mantle leads to more heating in the deep mantle and a higher heat flux at the poles (Segatz et al., 1988). Models containing only asthenospheric or deep mantle heating have been referred to as the two end-member cases (e.g., de Kleer, McEwen, et al., 2019; de Kleer, de Pater, et al., 2019), with actual conditions likely in between. Following this idea, several authors have linked Io's enhanced equatorial heat flux to argue in favor of a low-viscosity asthenosphere (Davies et al., 2015; Davies, Perry, Williams, Veeder, & Nelson, 2024; Hamilton et al., 2013). Nevertheless, this claim has not directly been connected to interior property values and has recently been challenged (e.g., Pettine et al., 2024; Sowell & Stevenson, 2026).

Despite years of observations and many analyses of Io's interior (e.g., Bierson & Nimmo, 2016; Kervazo et al., 2022; Lopes et al., 2023; Park, Jacobson, et al., 2025; Segatz et al., 1988), there has not been a full inversion of Io's interior properties using both gravity and volcanic activity observations. Park, Jacobson, et al. (2025) performed an inversion to show the impact of the measured $\Re\{k_2\}$, but they did not include the possibility for an asthenosphere, nor any measure of Io's surface heat flux.

Here, we investigate whether combining gravitational and heat flux observations provides evidence for a low-viscosity asthenosphere, and if so, what its characteristics are. We exploit the complementary sensitivities of geophysical observables within a Bayesian framework to infer the most likely spherically symmetric interior of Io. In our inversion, we include layer thicknesses, densities, and rheologies of a four-layered interior model (core, mantle, asthenosphere, crust). This allows us to identify the most likely layer in which Io's heat dissipation occurs and provide the most likely interior property values.

We start by describing the existing observational constraints to be used in this work—that is, gravitational and volcanic observations (Section 2). Afterward, we describe how these observations are integrated into a Bayesian framework (Section 3) to infer the most likely density and rheology structure (Section 4). Finally, we discuss the implications for our understanding of Io's interior and other rocky worlds (Section 5).

2. Observations

2.1. Static and Temporal Gravity Measurements

The gravitational potential of a spherical body, centered at the center of mass, is given as

$$U(r, \theta, \phi) = \frac{Gm}{r} \left[1 + \sum_{l=2}^{\infty} \sum_{m=0}^l \left(\frac{R}{r} \right)^l (C_{lm} \cos(m\phi) + S_{lm} \sin(m\phi)) P_{lm}(\cos \theta) \right]. \quad (1)$$

Here, G is the gravitational constant, m the mass of the moon, $R = 1822.7 \text{ km}$ the reference radius of the moon, r the radial distance from the center of the moon, θ the colatitude, ϕ the longitude, P_{lm} the associated Legendre polynomial of degree l and order m , and the gravitational coefficients C_{lm} and S_{lm} , with $J_2 = -C_{20}$. The gravitational coefficients can be split into a static and a dynamic component. The static part relates to the long-term, radial density structure of Io, and radiometric data from Galileo and Juno allowed the retrieval of the static degree-2 components J_2 and C_{22} without the assumption of hydrostatic equilibrium (Anderson et al., 2001; Park, Jacobson, et al., 2025).

Table 1
Observables Used in the MCMC, We Have Rescaled J_2 , C_{22} , and $\Re\{k_2\}$ to Match Our Reference Radius of $R = 1822.7$ km

Parameter	Value
J_2	$1848.1 \pm 1.5 \times 10^{-6a}$
C_{22}	$553.6 \pm 0.3 \times 10^{-6a}$
$\Re\{k_2\}$	0.127 ± 0.048^a
$\Im\{k_2\}$	-0.015 ± 0.003^b
$F_{\text{ratio}, 30^\circ}$	0.4 ± 0.11
$F_{\text{ratio}, 45^\circ}$	0.69 ± 0.097
$F_{\text{ratio}, 60^\circ}$	0.741 ± 0.074

^aPark, Jacobson, et al. (2025). ^bLainey et al. (2009).

The dynamic component of C_{lm} and S_{lm} changes periodically due to the effect of diurnal Jovian tides, and gravitational perturbation due to the tidal response is described by k_2 . This is a complex parameter where the real component, $\Re\{k_2\}$, is associated with the amplitude of tidal deformation. The imaginary component, $\Im\{k_2\}$, is related to the phase-lag between the tidal force and the body's response, and is a measure of tidal heating. The amount of tidal dissipation is coupled to the orbital evolution of Io (e.g., Peale et al., 1979; Yoder, 1979) and so $\Im\{k_2\}$ could be estimated using astrometric observations before the flybys of Juno (Lainey et al., 2009).

The values used in this work are listed in Table 1. While Park, Jacobson, et al. (2025) also found a value for $\Im\{k_2\}$ ($\Im\{k_2\} = -0.0109 \pm 0.0054$) they report lower accuracy compared to the findings of Lainey et al. (2009). Because $\Im\{k_2\}$ and $\Re\{k_2\}$ are determined independently (Park, Jacobson, et al., 2025), we still choose to use the value found in Lainey et al. (2009) for

the rest of this study. We use the equal volume radius as our reference radius (de Pater et al., 2021), which differs from the one used by Park, Jacobson, et al. (2025). Accordingly, we have rescaled J_2 , C_{22} , and $\Re\{k_2\}$.

2.2. Volcanic Activity

The decade-spanning observations of Io have led to near-global maps of Io's volcanic activity (Davies et al., 2015; Davies, Perry, Williams, Veeder, & Nelson, 2024; Hamilton et al., 2013; Ross et al., 1990). Besides the fact that the volcanic activity is not evenly distributed, two trends are visible on these maps. First, the volcanic heat flow seems to cluster at low latitudes (between $\pm 60^\circ$ N) (Davies et al., 2015; Davies, Perry, Williams, Veeder, & Nelson, 2024; Pettine et al., 2024). Interestingly, the number of hotspots per area does not show a latitudinal dependency and is roughly uniform (Davies, Perry, Williams, & Nelson, 2024). Second, both the volcanic heat flow and the number of volcanic features peak at roughly 30–60 degrees to the east of the sub-Jovian and anti-Jovian points (Cantrall et al., 2018; Davies, Perry, Williams, Veeder, & Nelson, 2024; Davies et al., 2015; de Kleer & de Pater, 2016b; Hamilton et al., 2013; Pettine et al., 2024; Steinke, van Sliedregt, et al., 2020; Veeder et al., 2012, 2015; Veenstra et al., 2025). We use the volcanic heat flux data from Davies, Perry, Williams, Veeder, and Nelson (2024), as it is the only data set that incorporates observations from all spacecraft and ground-based telescopes and spans the longest time. We cannot account for the reportedly higher heat flux on Io's North pole (Davies, Perry, Williams, & Nelson, 2024; Davies, Perry, Williams, Veeder, & Nelson, 2024) because our spherically symmetric interior model cannot produce a tidal heating pattern with a polar dichotomy (Beuthe, 2013). Instead, we average the flux over both polar caps and are thus fitting the background model onto which lateral variations might exist.

While volcanic heat flow observations are believed to be the most direct connection to Io's heat flux distribution that we have, linking them to tidal heating models is challenging. The difficulty stems from three main problems: volcanic heat flux only represents a fraction of Io's total heat output, volcanic events are discrete elements in time and space, and most widely used tidal heating models assume a spherically symmetric interior. First, volcanic heat flow only explains about 60 TW of the observed ~ 100 TW thermal emissions of Io (Lainey et al., 2009; Veeder et al., 1994). How this “missing” heat flux is distributed is unknown, but it is likely related to near-surface intrusive activity, such as emplacement of magma in the crust (Spencer et al., 2020). This would mean that the distribution of the missing heat flux would follow the observed distribution of volcanic heat flow. The only existing constraint on how Io's missing heat flux could be distributed is a PPR measurement by Galileo (Rathbun et al., 2004). Rathbun et al. (2004) report on nighttime polar temperatures that are higher than expected. However, due to missing daytime measurements of the same locations, they could not determine whether the higher temperatures were due to a higher endogenic heat flow at the poles or a locally larger heat capacity. In this study, we will initially assume that the missing heat flux is evenly distributed across Io's surface, and so does not distort the observed difference between low and high latitudes. Later, we will revisit this assumption to investigate the effect of a latitudinally uniform heat flux distribution, which implies a larger share of the missing heat flux originating from the poles.

The second challenge is that most volcanic eruptions are discrete events in time and/or location, and can thus be hard to link to models that assume long-time averages. Io's observed volcanic activity spans a broad range of

volcanic expressions, from eruptions that produce gaseous plumes reaching hundreds of kilometers above the surface, to forms of effusive volcanism, such as lava flows and lakes (Davies & Veeder, 2023; Milazzo et al., 2005). These different volcanic expressions also differ in their characteristic timescales and degree of spatial stability. Active lava lakes, hosted within paterae, are among the more spatially stable volcanic features on Io: as caldera-like depressions, paterae mark sites of volcanic activity on geological timescales of up to ~1 Myr (Hamilton et al., 2013). Yet spatial stability does not imply a stable heat output. This challenge of capturing a stable heat output is illustrated by Dazhbog Patera, a large patera, whose emitted power appears to have declined from ~4 TW measured by Galileo to ~1.3 TW observed by Juno (Davies, 2026; Davies, Perry, Williams, Veeder, & Nelson, 2024; Perry et al., 2025). Such a change is large enough that its variability alone perturbs the global heat flux distribution. Loki Patera, the largest thermal source on Io, is a singular outlier within this class, so it is routinely treated separately in global heat flow analyses (Davies & Veeder, 2023; de Kleer et al., 2017; Matson et al., 2006; Mura et al., 2025). In contrast, flow fields and certain plume-forming eruptions are inherently transient and spatially variable: they may shift location between observation epochs or cease entirely, complicating any attempt to assign them a representative long-term flux (Davies et al., 2010; de Kleer & de Pater, 2016a; Perry et al., 2025). Both the temporal and spatial variations make it hard to distill a long-term average heat flux map, which is necessary to compare to tidal dissipation models because they assume steady-state conditions (Peale et al., 1979; Rovira-Navarro et al., 2024; Segatz et al., 1988). Furthermore, observations of Io are not spatially uniform, so observational biases (such as the lack of observations of Io's poles) may impact our understanding of the spatial distribution of volcanic activity.

The third challenge is that most interior models assume spherical symmetry and thus cannot account for any small-scale variations or hemispherical dichotomies in heat flux. Modeling these phenomena requires a model capable of computing 3D tides (Berne et al., 2024; Lau et al., 2015; Qin et al., 2016; Rovira-Navarro et al., 2024; Veenstra et al., 2025), but this is computationally expensive to run, prohibiting exhaustive inversions. Therefore, we attempt to filter out these small-scale variations and the longitudinal asymmetry by focusing on the long-wavelength signals in the observations. We only use latitudinal flux variations, as that distribution is the most sensitive to radial variations in rheology (Beuthe, 2013; Segatz et al., 1988; Tyler et al., 2015). The latitudinal heat flux distribution can be represented by the ratio of polar and equatorial heat flux (Davies, Perry, Williams, & Nelson, 2024; Davies, Perry, Williams, Veeder, & Nelson, 2024; Kervazo et al., 2022). We define the polar heat flux as the average flux over both polar caps. To compute it, we divide the total volcanic heat flow of both polar caps by their area, given as twice $A_{\text{cap}} = 2\pi R^2(1 - \cos(\theta_{\text{cap}}))$, where R is the radius of Io, and θ_{cap} is the colatitude of the cap. The equatorial flux we define as the average flux over the remaining surface area (total area of Io minus the area of both polar caps). Because the total heat flux can not be derived from the volcanic heat flux, we use the ratio of polar-to-equatorial heat flux instead of the total fluxes. The boundary between equatorial and polar heat flux is somewhat arbitrary. Traditionally, a co-latitude of 30° has been used to separate equatorial and polar heat flux (e.g., Davies, Perry, Williams, Veeder, & Nelson, 2024). This co-latitude is where end-member tidal dissipation models exhibit their largest differences. Still, this boundary is not sharp, and no good argument can be made for 30° over, say, 32° (Tosi et al., 2025). Instead, we use three cap angles, at 30°, 45°, and 60°, in the MCMC. Doing so, we are not only fitting our model to a particular polar-to-equatorial heat flux but also to latitudinal variations.

Figure 1 shows the polar to equatorial heat flux ratio for various cap angles, with the effect of the cap angle clearly visible. Particularly powerful hotspots, such as Dazhbog Patera at a co-latitude of 35°, can strongly influence the ratio. We interpret the variance in the black line as an expression of the partial stochastic nature of volcanic activity. To capture this spatial variation, we use a sliding window to average the ratio and use the standard deviation in this averaging range as an uncertainty in our inversion. We use a sliding window of 15°, based on the distance for which suspected synchronized eruptions at different locations have occurred, thus seemingly originating from the same source. Examples are: the apparent simultaneous eruptions at Rarog and Heno Patera, separated by ~500 km (or ~15 degrees) (de Pater et al., 2014); the simultaneous eruption of three paterae within the Bosphorus Regio, separated by a distance of almost 1,000 km (Perry et al., 2025); and the recent outburst observed by Juno in the southern hemisphere, at Illyrikon, that showed synchronized eruptions over a spatial range of roughly 500 km (Mura et al., 2026). Changing this averaging range does not significantly change the results.

To get some estimate for the uncertainty in the ratio due to temporal variability, we do the above for data sets containing either the low or high power output estimates for Dazhbog Patera, 1.3 and 4 TW, respectively

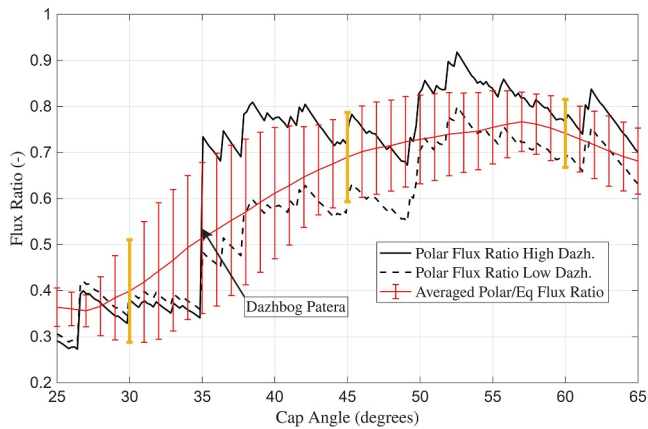


Figure 1. Polar-to-equatorial heat flux ratio as a function of the cap angle used to define the polar cap, plotted for both the high and low power output estimates for Dazhbog Patera (solid and dotted black lines, respectively). The heat flux data is taken from Davies, Perry, Williams, Veeder, and Nelson (2024). The big jump at 35° is due to Dazhbog Patera. The averaged polar-to-equatorial heat flux in a sliding 15° window is plotted in red. It is the average of the high- and low-power estimates for Dazhbog Patera. The error bars are both a measure of the variance with cap angle within the sliding window and the variations in time of Dazhbog Patera's output. Cap angles 30°, 45°, and 60° are indicated in orange and are our used observables.

(Davies, 2026; Davies, Perry, Williams, Veeder, & Nelson, 2024). While other volcanic features also change output with time (Davies, Perry, Williams, Veeder, & Nelson, 2024; de Kleer & de Pater, 2016a), the large output and observed variability make it the dominant factor. The change in output of Dazhbog is also comparable to the total change in volcanic thermal output between the global volcanic heat flow map in 2015 (Davies et al., 2015) and 2024 (Davies, Perry, Williams, & Nelson, 2024). Although that change was mainly due to the addition of new features, it shows that a variation of ~3 TW in total volcanic heat flux over time is reasonable. We average between the high and low power output estimates for Dazhbog Patera (red line in Figure 1), thus assuming that the data with a high estimate for Dazhbog and the one with a low estimate are equally likely. The difference between the solid and dotted black lines is used as an additional uncertainty and added to the uncertainty from the spatial variability σ_{sp} . The spatial uncertainty thus combines data from both the low and high estimates as

$$\sigma_{sp} = \sqrt{\frac{\sigma_{sp, high}^2 + \sigma_{sp, low}^2}{2}}, \quad (2)$$

and the total uncertainty, including the difference in average ratio, as

$$\sigma_{tot} = \sqrt{\sigma_{sp}^2 + \left(\frac{\mu_{low} - \mu_{high}}{2}\right)^2}, \quad (3)$$

with μ_{low} and μ_{high} , the average ratio at a given cap angle for the low and high estimates respectively. The three heat flux ratios and their uncertainties can be found in Table 1.

3. Methods

We combine the observations using Bayesian inference implemented via Markov-Chain Monte Carlo (MCMC) sampling. To constrain Io's interior properties using the observations above we need a forward model that can map interior parameters to the observations (Section 3.1.1) and a statistical framework for the inversion (Section 3.2).

3.1. Forward Model

3.1.1. Interior Model

We consider four-layer models of Io, containing a liquid core, mantle, asthenosphere, and lithosphere. Each layer has constant properties. We assume that density increases with depth and set the asthenosphere to the same density as the mantle because the observations do not allow us to estimate them separately (Moore et al., 2007). We don't vary Io's total mass or radius, and so the density of the lithosphere is derived from the remaining mass and thickness. We assume the core is liquid, and all layers are incompressible, as Hay et al. (2025) showed that bulk dissipation is likely negligible on Io.

The viscoelastic response of all layers, except the core, is modeled assuming Andrade rheology (Bierson, 2024; Renaud & Henning, 2018; Walterová et al., 2023). Specifically, we use the ζ model of Andrade rheology, as presented in Bierson (2024), in favor of the β model used by Park, Jacobson, et al. (2025). Because the β parameter depends on material properties such as temperature, grain size, and pressure, similar to the viscosity, varying β and η separately can be unrealistic (Bierson, 2024). Setting β to a high value while keeping η also high can lead to an overestimation of the amount of dissipation. To keep the direct scaling with viscosity, the ζ model was developed (Castillo-Rogez et al., 2011; Efroimsky, 2012). Andrade rheology introduces two additional parameters: α , which has the units of time and is related to the duration of the transient response (Castillo-Rogez et al., 2011), and the aforementioned ζ . We use a single ζ value for the mantle, asthenosphere, and lithosphere, which we vary between 0.01 and 100 (Bierson, 2024; Walterová et al., 2023). The α parameter in the Andrade rheology is better constrained by experiments, with values typically between 0.2 and 0.4 (Bierson, 2024). We

Table 2

Model Parameters and Their Ranges and Distributions, as Used in the Inversions

Parameter	Units	Range	Distribution	Reference
Core outer radius r_c	km	650–1,100	Uniform	
Mantle thickness h_m	km	250–1,200	Uniform	
Asthenosphere thickness h_a	km	40–500	Uniform	
Lithosphere thickness h_l	km	10–100	Uniform	
Core density ρ_c	kg m ^{−3}	5,000–8,000	Uniform	
Mantle density ρ_m	kg m ^{−3}	2700–3,600	Uniform	
Lithosphere density ρ_l	kg m ^{−3}	2700–3,600	Uniform	
Mantle shear modulus μ_m	Pa	10^8 – 7×10^{10}	Log-uniform	
Asthenosphere shear modulus μ_a	Pa	10^8 – 7×10^{10}	Log-uniform	
Lithosphere shear modulus μ_l	Pa	10^{10} – 7×10^{10}	Log-uniform	
Mantle viscosity η_m	Pa s	10^{11} – 10^{24}	Log-uniform	
Asthenosphere viscosity η_a	Pa s	10^{11} – 10^{24}	Log-uniform	
Lithosphere viscosity η_l	Pa s	10^{19} – 10^{24}	Log-uniform	
Andrade parameter ζ	–	0.01–100	Log-uniform	
Non-hydrostatic contribution J_2	–	$-4\sigma_{20}^{\text{nhyd}}$ – $-4\sigma_{20}^{\text{nhyd}}$	Normal	
Non-hydrostatic contribution C_{22}	–	$-4\sigma_{22}^{\text{nhyd}}$ – $-4\sigma_{22}^{\text{nhyd}}$	Normal	
Orbital period T	hr	42.456	Fixed	
eccentricity e	–	0.004031	Fixed	
Reference radius R	km	1822.7	Fixed	de Pater et al. (2021)
Bulk modulus κ	Pa	2×10^{13}	Fixed	
Gravitational parameter GM	m ³ s ^{−2}	5.9599×10^{12}	Fixed	Park, Jacobson, et al. (2025)
Andrade parameter α	–	0.21	Fixed	Bierson (2024)
Non-hydrostatic contribution σ_{20}	–	$\sqrt{5} \cdot 8 \cdot 10^{-6}$	Fixed	
Non-hydrostatic contribution σ_{22}	–	$\sqrt{5/12} \cdot 8 \cdot 10^{-6}$	Fixed	
Correlation between C_{20} and C_{22}	–	−0.00256203491215322	Fixed	Park, Jacobson, et al. (2025)
Correlation between $\Re\{k_2\}$ and C_{20}	–	−0.029120556410328	Fixed	Park, Jacobson, et al. (2025)
Correlation between $\Re\{k_2\}$ and C_{22}	–	0.521951972001789	Fixed	Park, Jacobson, et al. (2025)

found during initial testing that α had only minimal impact on our results because Io must be close to the Maxwell peak to match its dissipation (Bierson, 2024). Therefore, we kept α fixed at 0.21 (Bierson, 2024) during our inversions to reduce the number of free parameters. The ranges for all free parameters are given in Table 2.

We consider core densities between those of a fully iron core ($\sim 8,000$ kg m^{−3}) and an iron-iron sulfide eutectic ($\sim 5,000$ kg m^{−3}) (Moore et al., 2007). The mantle, asthenosphere, and lithosphere densities vary between 2,700 and 3,600 kg m^{−3}, a broad range of values for rock under low pressures. We limit the maximum size of the asthenosphere to be 500 km, a conservatively large value that is mainly based on how the heating patterns change with asthenosphere size, as described by Tyler et al. (2015). The minimum size of the asthenosphere is 40 km, slightly smaller than the minimum value suggested from magnetic induction measurements (>50 km) (Khurana et al., 2011). We limit the lithosphere thickness to 100 km, much larger than current estimates (e.g., Steinke, Hu, et al., 2020; Spencer et al., 2021), which find values around 50 km because magmatic emplacement in the crust likely limits the formation of a thick lithosphere (Spencer et al., 2020). Because the allowed range of viscosities and shear moduli is the same for the mantle and asthenosphere, there is no inherent bias toward the formation of a low-viscosity asthenosphere. The shear modulus ranges are broader than typically assumed to both show the strong effect that $\Re\{k_2\}$ has on them, and because we anticipate that there is significant melt within Io, which lowers the shear modulus (Hay et al., 2025; Kervazo et al., 2022).

3.1.2. Static Gravity

We compute the two measured degree-2 static gravity coefficients starting from a spherically symmetric layer structure and density distribution under hydrostatic equilibrium conditions. However, Io is not necessarily in perfect hydrostatic equilibrium, and so the observed coefficients can deviate from these theoretical values. Thus, as is typically done, we split the computed coefficients into a hydrostatic and non-hydrostatic contribution (e.g., Iess et al., 2014),

$$J_2 = J_2^{\text{nhyd}} + J_2^{\text{hyd}}, \quad (4)$$

$$C_{22} = C_{22}^{\text{nhyd}} + C_{22}^{\text{hyd}}. \quad (5)$$

The hydrostatic contributions, denoted by hyd, are determined by computing the figure of equilibrium for a layered body (e.g., Murray & Dermott, 2000) to second-order accuracy (Beuthe et al., 2016; Zharkov & Gudkova, 2010). Although Gyalay and Nimmo (2024) already showed that the second-order contribution is small for Io, it is of comparable size to the uncertainty in J_2 and C_{22} , which is why we still include it. The expressions for the hydrostatic contributions are

$$J_2^{\text{hyd}} = \frac{5}{6} \left(1 + \frac{16}{21} h_2^f q \right) k_2^f q, \quad (6)$$

$$C_{22}^{\text{hyd}} = \frac{1}{4} \left(1 + \frac{64}{21} h_2^f q \right) k_2^f q, \quad (7)$$

with $q = \omega^2 R^3 / GM$ the nondimensional rotational parameter, and k_2^f and $h_2^f = 1 + k_2^f$ the fluid Love numbers. The fluid Love number k_2^f is computed using a matrix-propagator method, as described in Padovan et al. (2018). The non-hydrostatic contributions are free parameters in the MCMC inversion, and are subject to rotational stability conditions (Park, Jacobson, et al., 2025)

$$\begin{aligned} J_2^{\text{nhyd}} &> 2C_{22}^{\text{nhyd}}, \\ C_{22}^{\text{nhyd}} &> 0. \end{aligned}$$

3.1.3. Tidal Response and Dissipation Pattern

The degree-2 Love number k_2 is a function of the radial structure as well as the rheological properties of the body: the shear modulus and viscosity of every layer. This makes them a very complementary set of observations when determining the interior properties. The Love numbers are computed by solving the mass, momentum, and Poisson's equation for a self-gravitating viscoelastic body. We use the spherically symmetric version of *LOV3D* (Park, Berne, et al., 2025; Rovira-Navarro et al., 2024, 2025) to solve these equations semi-analytically in the Fourier domain, allowing for the efficient computation needed to do the ~ 60 million computations in a reasonable time. The flux ratios are computed from the radially integrated volumetric heating rate.

3.2. Setup of the MCMC

To explore the range of interior properties that match the observed geophysical parameters, we invert them using a MCMC sampler. We assume that the prior distribution of the interior properties is uniform and sample them using a Differential Evolution (DE) algorithm (Nelson et al., 2014), as implemented in the *emcee* package (Foreman-Mackey et al., 2013). Because of the relatively large number of interior parameters (14) and the strong correlation that exists between some of them, using DE sampling can speed up convergence and more efficiently explore the parameter space compared to the standard Metropolis-Hastings algorithm (Nelson et al., 2014).

We run 60 chains, each taking 1,000,000 steps after an initial 80,000 steps burn-in. To ensure convergence, we make sure the average autocorrelation time over all 14 parameters is less than $N/100$, with N the number of steps. By keeping the acceptance ratio between 0.2 and 0.3, we ensure sufficient exploration of the parameter space. We tune the γ parameter, which influences the stepsize in parameter space, in the DE algorithm to accomplish the

desired acceptance ratio (Nelson et al., 2014). The other algorithm parameters we keep at their default settings (Foreman-Mackey et al., 2013). Besides the 80,000 burn-in steps, we also remove the first 100,000 steps from the main run to reduce any potential effect from starting conditions.

To find the most likely interior property ranges, the computed geophysical parameters are compared against the observed values using a log-likelihood function, defined as

$$\log P_x = -\frac{1}{2}(X - X_{\text{obs}})^T \Sigma^{-1} (X - X_{\text{obs}}), \quad (8)$$

with X the computed geophysical parameters, X_{obs} the observations, and Σ the covariance matrix of the observations. Only J_2 , C_{22} , and $\Re\{k_2\}$ are assumed to be correlated, with the correlations taken from Park, Jacobson, et al. (2025) and given in Table 2. We add a fitness check to limit the size of the non-hydrostatic contribution to J_2 and C_{22} , as will be explained in the next section. To avoid putting too much emphasis on the heat flux observations, we have added a hyperparameter ν to scale down the three heat flux observables relative to the gravity observations. The total log-likelihood function is

$$\log P = \log P_{\text{grav}} + \nu \log P_{\text{flux}} + \log P_{\text{nhyd}}, \quad (9)$$

where $\log P_{\text{grav}}$ is the log-likelihood function for the geophysical observables [C_{20} , C_{22} , $\Re\{k_2\}$, $\Im\{k_2\}$], $\log P_{\text{flux}}$ is the log-likelihood function for the heat flux observations [$F_{\text{ratio}, 30^\circ}$, $F_{\text{ratio}, 45^\circ}$, $F_{\text{ratio}, 60^\circ}$], and $\log P_{\text{nhyd}}$ the log-likelihood function for the non-hydrostatic contribution. We treat ν as a variable and vary it between 0.1 and 0.25, such that the posterior distribution functions of the observables are Gaussian-shaped and centered around the observation value.

4. Results and Discussion

We use the method described above to infer Io's interior properties. To highlight the importance of the heat flux ratio constraint, we consider an inversion with and without the constraint to match this observation. Both inversions converged to the desired autocorrelation time with acceptance ratios between 0.2 and 0.3. The fits of both inversions to the observables are good (see Figures S2 and S4 in Supporting Information S1); the fits and the full inversions, with and without the constraint to match the polar-to-equatorial heat flux ratio, are shown in Supporting Information S1 (Figures S1 and S3). Apart from these, additional inversions were performed to test the sensitivity of the results to our assumptions. More specifically, the value of the heat flux ratio (Section 4.3) and the number of layers used in our interior model (Section 4.4).

4.1. Density Profile

We first discuss the posterior PDFs of the core thickness and density and the non-hydrostatic contributions. The first is a validation of our model, since the gravity coefficients estimated by Park, Jacobson, et al. (2025) do not differ much from previous estimates (Moore et al., 2007). The second showcases the limited but still significant effect of using the second-order figure of equilibrium equations.

Figure 2 shows the core thickness and density, and non-hydrostatic contributions obtained in both inversions. They are comparable to the results in Park, Jacobson, et al. (2025). As expected, the distributions of density and thickness are unaffected by the introduction of the polar heat flux ratio as a constraint. Given the large degeneracy in the solution, the large range in core radius is to be expected (Moore et al., 2007), see subplot c in Figure 2. In fact, the core density posterior PDFs are roughly uniform over the full range of the prior distribution, from 5,000 to 8,000 kg/m³. Without seismic measurements or a better understanding of Io's chemical composition, the core radius and density will remain unresolved. The mantle and asthenosphere density PDFs, see Figures S1 and S3 in Supporting Information S1, are in the expected range of around 3,300 kg/m³ (Moore et al., 2007). The lithosphere thickness, see Figures S1 and S3 in Supporting Information S1, is unconstrained due to the small density difference between the lithosphere and the mantle/asthenosphere. Measurements of Io's libration amplitude, combined with the measurement of $\Re\{k_2\}$, could provide an additional constraint on the thickness and rigidity of the lithosphere (Van Hoolst et al., 2020).

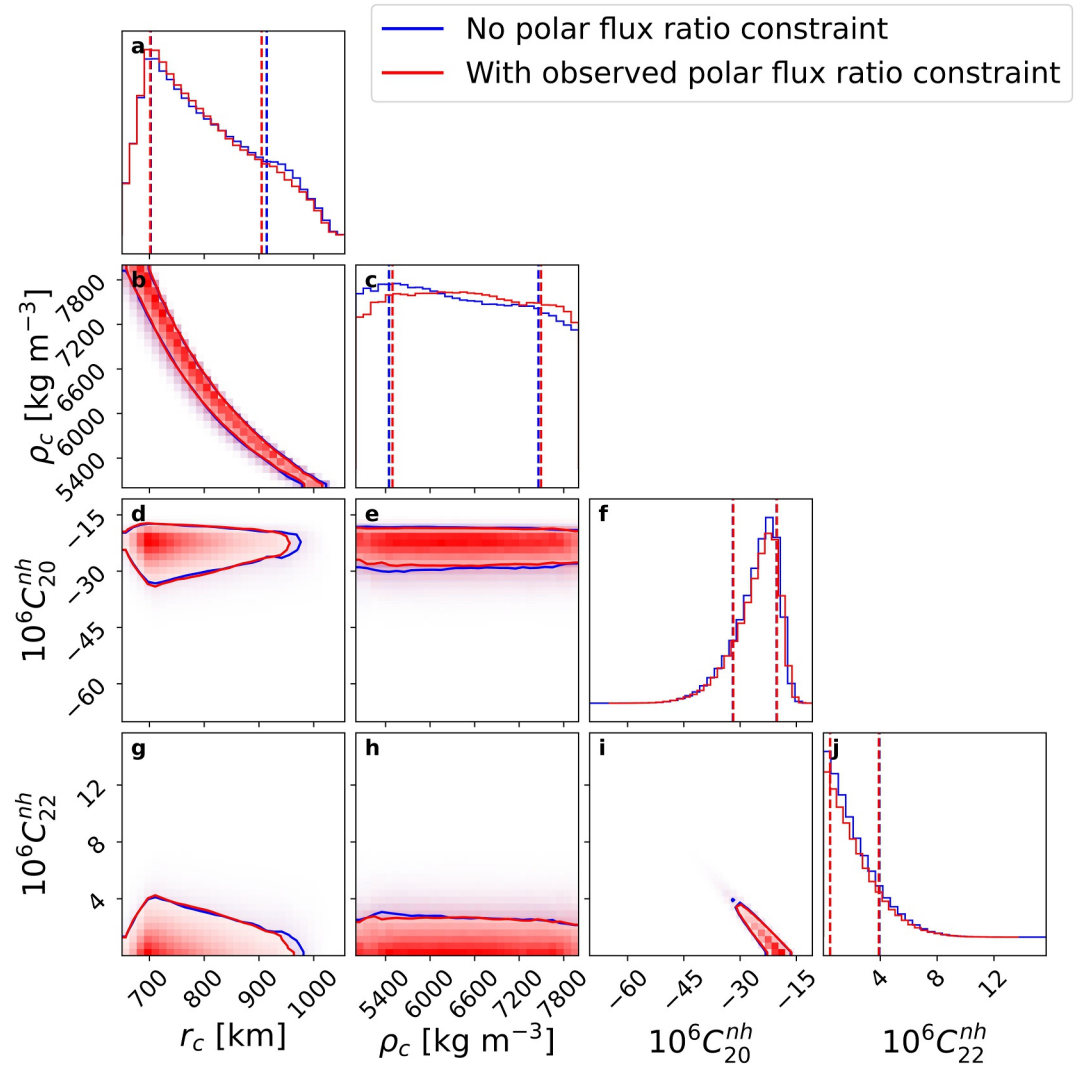


Figure 2. The posterior PDFs of the core radius r_c (subplot a), core density ρ_c (subplot c), and 10^6 times the non-hydrostatic contributions to J_2 and C_{22} (subplots f and j respectively). The 2-dimensional PDFs, indicating the correlations between two parameters, are given in the off-diagonal subplots. The cases with and without the constraint to match the observed polar-to-equatorial flux ratio are plotted on top of each other, in red and blue, respectively. The dotted lines in the posterior PDFs and the solid contours in the 2D PDFs contain 68.3% of the cases.

Our use of a second-order figure of equilibrium results in significantly larger non-hydrostatic contributions compared to Park, Jacobson, et al. (2025). Our J_2^{nh} and C_{22}^{nh} ranges are $24.56^{+7.20}_{-4.35} \times 10^{-6}$ and $1.70^{+2.18}_{-1.24} \times 10^{-6}$, while Park, Jacobson, et al. (2025) report $4.2^{+2.9}_{-4} \times 10^{-6}$ and $0.6^{+1.1}_{-0.5} \times 10^{-6}$. This indicates that the reported error on the static gravity coefficients is indeed low enough that second-order methods are relevant since the non-hydrostatic contributions are larger than the error (see Table 1). Without measurement of higher-order gravity coefficients, or more accurate measurements of Io's shape, these non-hydrostatic contributions cannot be attributed to any specific surface features (Gyalay & Nimmo, 2024).

4.2. Rheology Profile

We now investigate where the dissipation takes place, and for that, we look mainly at asthenosphere and mantle viscosity. In Figure 3, we show the PDFs of these rheological parameters as well as the mantle shear modulus and the asthenosphere thickness. We have not plotted the ζ parameter because it is unconstrained due to degeneracy between viscosity and the ζ parameter since they have similar effects on the amount of tidal dissipation in a layer

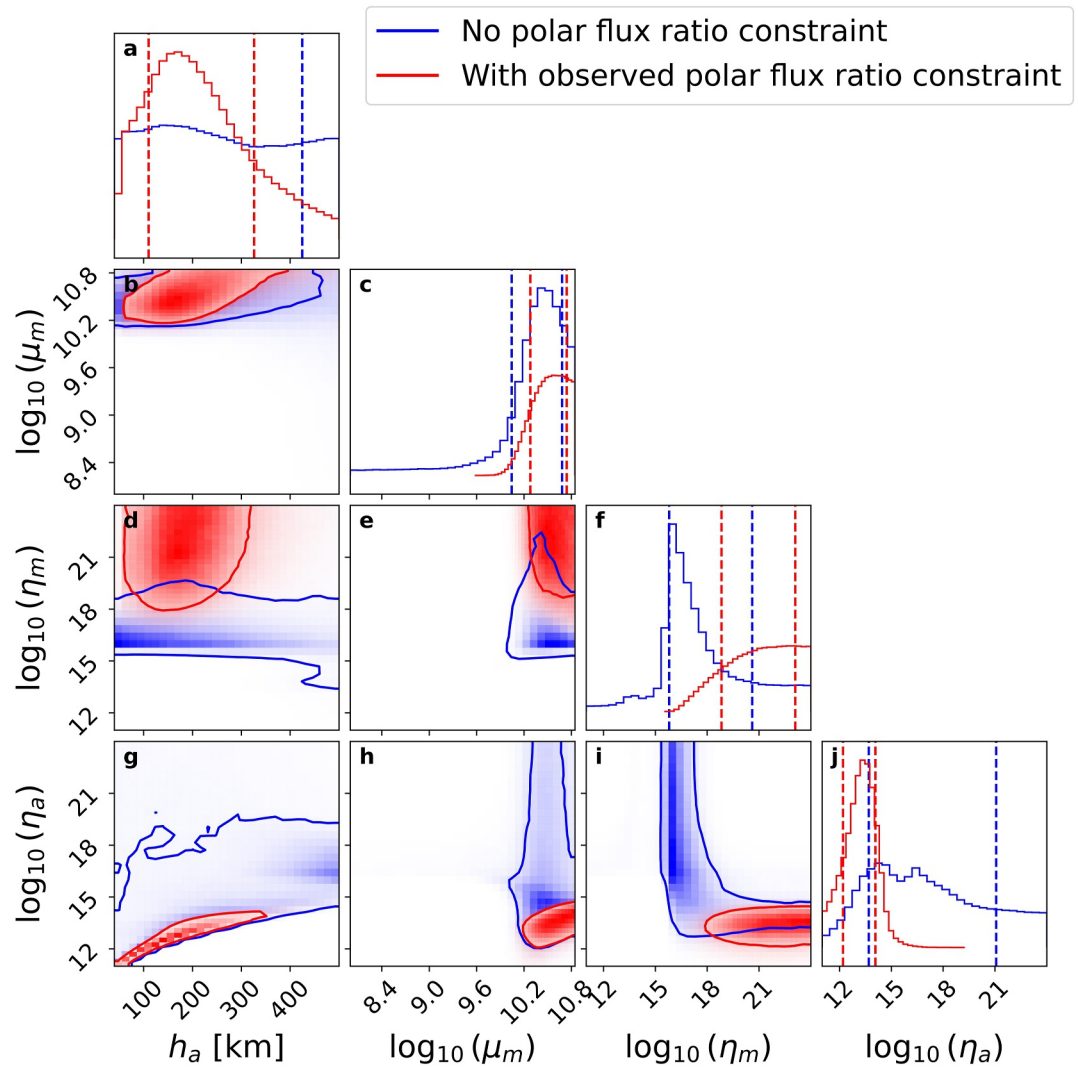


Figure 3. The posterior PDFs of the asthenosphere thickness h_a (subplot a), $\log_{10}$ of the shear modulus of the mantle μ_m (subplot c), the $\log_{10}$ of the viscosity of the mantle η_m (subplot f), and the $\log_{10}$ of the viscosity of the asthenosphere η_a (subplot j). The 2-dimensional PDFs, indicating the correlations between two parameters, are given in the off-diagonal subplots. The cases with and without the constraint to match the observed polar flux ratio as one of the observables are plotted on top of each other, in red and blue, respectively. The dotted lines in the histograms and the contours in the 2D PDFs contain 68.3% of the cases.

(see Figures S1 and S3 in Supporting Information S1) (Bierson, 2024; Renaud & Henning, 2018; Walterová et al., 2023).

The inclusion of the polar flux ratio as a constraint significantly changes the PDF of the viscosity in the mantle and the asthenosphere, as shown in subplots f, i, and j in Figure 3. The inversion without the polar flux ratio as a constraint has a slight preference for an Io with a relatively low-viscosity mantle, while the asthenosphere viscosity is relatively unconstrained; 68% of all cases fall within $\eta_a \approx 10^{14}$ Pa s and $\eta_a \approx 10^{21}$ Pa s (subplot j). This corresponds to an Io that is dominated by mantle heating (see subplot a in Figure 4), which is one of the two familiar end-member models (de Kleer, McEwen, et al., 2019; de Kleer, de Pater, et al., 2019; Segatz et al., 1988). The addition of the heat flux constraint drastically changes the inferred radial viscosity profile. With the added constraint, a low-viscosity mantle is unlikely, leaving only the possibility of a low-viscosity and slightly thicker asthenosphere; in 68% of all cases, the asthenosphere viscosity falls between $\eta_a \approx 10^{12}$ and $\eta_a \approx 10^{14}$ (subplot j). This also means that most heating occurs in the asthenosphere (see subplot d of Figure 4). Thus, the ratios we observe (Figure 1) favor an Io dominated by asthenospheric heating and thus a low-viscosity asthenosphere.

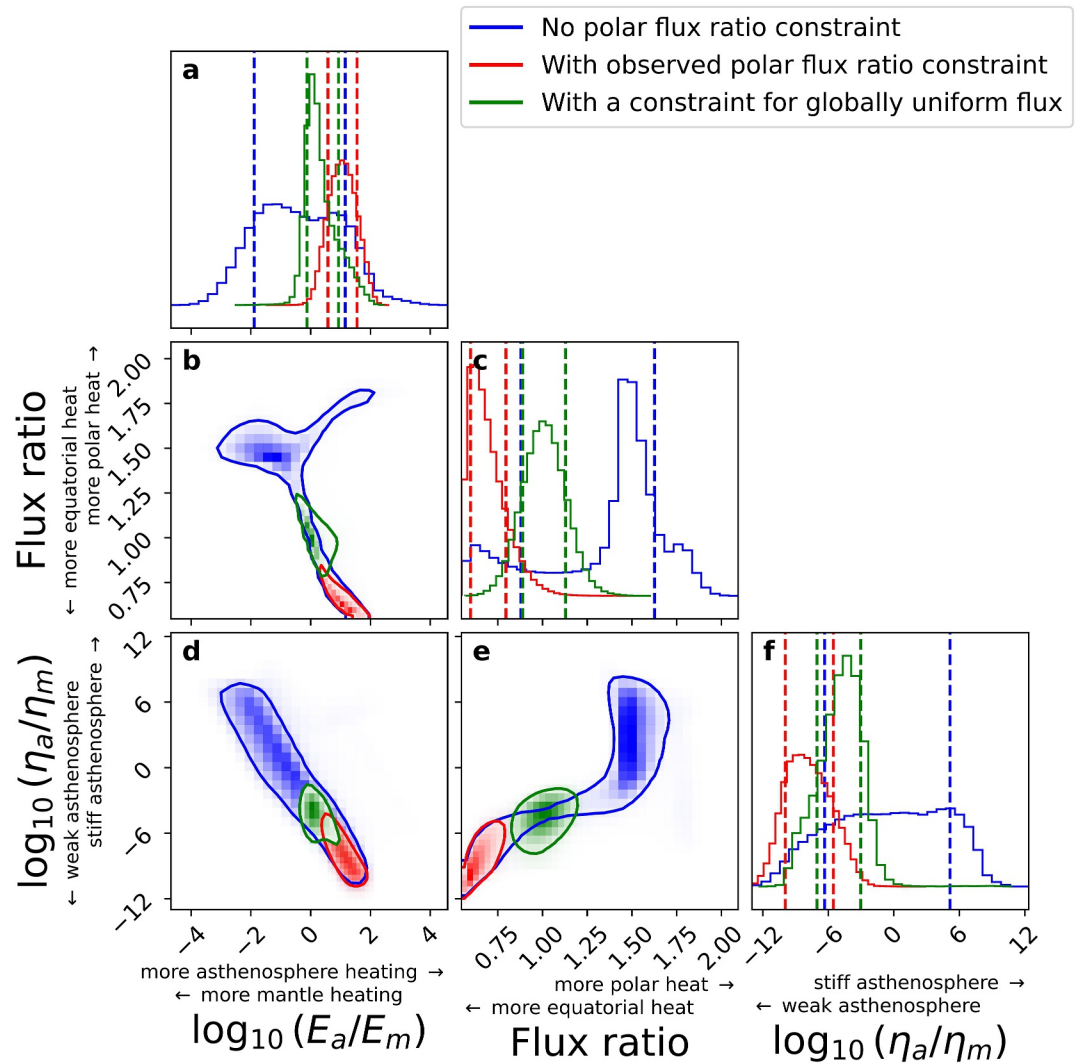


Figure 4. The posterior PDFs of the $\log_{10}$ of the ratio between energy dissipation in the asthenosphere (E_a) and mantle (E_m) (subplot a), the polar-to-equatorial flux ratio at a cap angle of 30° (subplot c), and the $\log_{10}$ of the ratio between asthenosphere and mantle viscosity (subplot f). The 2-dimensional PDFs, indicating the correlations between two parameters, are given in the off-diagonal subplots. The three cases, without the constraint to match the observed polar flux ratio (blue), with the constraint to match the observed ratio (red), and with a constraint to match a globally uniform flux ratio (green), are plotted on top of each other. The dotted lines in the histograms and the contours in the 2D PDFs contain 68.3% of the cases.

We note that the viscosity values we find are significantly lower than those found in Bierson and Nimmo (2016) and Park, Jacobson, et al. (2025). Although we expected this given the ζ model (see discussion above about the two formulations of the Andrade model), the values we find require further analysis, and we will return to this in Section 4.6.

The $\Re\{k_2\}$ observable limits the range of possible mantle and asthenosphere shear moduli to those typical for olivine (Turcotte & Schubert, 2002). This agrees with Park, Jacobson, et al. (2025), who showed that $\Re\{k_2\}$ precludes the possibility of a fully molten magma ocean, which, in our model, would amount to a very low shear modulus μ_a in the asthenosphere. Other than excluding the existence of a magma ocean, the measurement of $\Re\{k_2\}$ does not directly constrain any of the other end-member interior models, and adding the polar flux ratio as a constraint does not significantly change the PDF for μ_m .

Finally, there is also a trade-off between asthenosphere thickness and viscosity for $\Im\{k_2\}$ (Segatz et al., 1988). However, this effect is only strong if most of the heat dissipates in the asthenosphere. Without the polar-to-equatorial heat flux ratio as a constraint, this connection does not strongly influence the layer thickness, as

shown in Figure 3a. Introducing the polar heat flux ratio reduces the likelihood of a thick asthenosphere and shrinks the range of likely thicknesses.

4.3. Sensitivity to the Polar-Equatorial Heat Flux Ratios

The previous results demonstrate how much the polar-equatorial flux ratios can constrain Io's radial viscosity profile. However, despite our efforts to include an estimate for the uncertainty for this flux ratio, the true uncertainty can be larger because it is still only based on half of Io's heat output (Davies, Perry, Williams, Veeder, & Nelson, 2024). In the previous results, we assumed that the other half of Io's heat flux is distributed such that it does not alter the observed volcanic heat flux ratios. Here, we investigate the sensitivity of our inversion to the exact flux ratio values and examine what happens under a uniform heat flux distribution.

Figure 4 shows the sensitivity of the radial viscosity and heating profiles to the heat flux ratio constraint. It contains the relations between the polar-to-equatorial heat flux ratio at a polar cap of 30°, the $\log_{10}$ of the ratio of asthenosphere and mantle viscosity, and the $\log_{10}$ of the ratio of the amount of heating in asthenosphere and mantle. For the inversion without the flux ratio constraint (the blue data in Figure 4), the flux ratio is a model output, and it thus shows the full relation between the flux ratio and the viscosity and dissipation profiles allowed by the observed static gravity and tidal response. As expected, it contains both end-member tidal heating patterns, represented in subplot e as the two ends of the blue area. With the observed heat flux ratio as a constraint (the red data in Figure 4), the data indeed falls in the asthenospheric heating lobe of subplot e. Interestingly, the apparent hysteresis in subplot b, with asthenospheric heating dominated models producing both low and high flux ratios, comes from a population of models with very thick asthenospheres (>400 km), which is also visible in Figure 3g.

In between the two end-member models in Figure 4 lie interior models that produce heat flux ratios of ~ 1 . This heat flux ratio aligns with the density distribution of volcanic features, which also appears to be latitudinally constant (Davies, Perry, Williams, Veeder, & Nelson, 2024). Setting the heat flux constraints to 1 implies that the other half of Io's heat flux, about 40 TW, is distributed such that it smooths the distribution of observed volcanic heat flux. This is a conservative assumption because a uniform heat flux contradicts the observed nighttime temperatures by the PPR instrument (Rathbun et al., 2004). If, for example, all the missing heat flux escapes Io as thermal radiation due to conduction, the resulting surface temperatures would be too high. Rathbun et al. (2004) estimated that a maximum of $\sim 1 \text{ W m}^2$ additional endogenic heat flux could exist, whereas this uniform flux ratio would put that number between 2 and 1.3 W m^2 depending on the latitude, with the higher estimate for the higher latitudes. Other means of heat output, such as “stealth volcanism” (de Pater et al., 2020; Johnson et al., 1995) could still allow for a uniform heat flux, but there is no evidence of large activity at the poles for this phenomenon.

Still, Pettine et al. (2024), using only JIRAM data, found a stronger correlation between the observed volcanic heat flux and a uniform flux distribution than with any of the end-member heating patterns. Thus, to illustrate the impact of a uniform heat flux on Io's interior, we also perform an inversion where we set the polar-to-equatorial heat flux ratio to 1 ± 0.1 for all three cap angles (green data in Figure 4). The uncertainty of ± 0.1 is based on the uncertainty in the observed $F_{\text{ratio}, 45^\circ}$ and $F_{\text{ratio}, 60^\circ}$, but changing it does not affect the results. The results show that up to heat flux ratios of ~ 1 , a majority of the heating will still be in the asthenosphere. Subplot e also shows that even for a globally uniform flux, the asthenosphere viscosity should be orders of magnitude smaller than the viscosity in the mantle (the one-sigma interval is $\eta_a = 10^{12.21} - 10^{14.07} \text{ Pa s}$).

Thus, we find that, even with uniform heating, a significant amount of heating (>50%) occurs in the asthenosphere, and it always has a lower viscosity than the mantle. However, the new constraint does alter the ratio of asthenosphere to mantle viscosity, especially favoring lower viscosities in the mantle (see Table 3). All posterior distributions and the fits to the geophysical parameters are shown in Figures S5 and S6 in Supporting Information S1, and the interior parameter value ranges are given in Table 3.

4.4. Sensitivity to the Number of Layers

Our assumption of a four-layer model is the simplest model still able to produce the two known end-member tidal dissipation states (e.g., de Kleer, McEwen, et al., 2019; de Kleer, de Pater, et al., 2019). However, the low viscosities in the asthenosphere that we find are hard to explain, even with a significant presence of melt (see also Section 4.6). Introducing another layer might allow spreading the amount of dissipation over the asthenosphere and the upper mantle, increasing the viscosity in the asthenosphere. To check this, we ran an inversion with a

Table 3

Median Model Parameter Values and Their One-Sigma Range, As Well As the Parameters From the Best-Fitting Model, Given for Both the MCMC Inversions: With the Observed Polar-to-Equatorial Heat Flux Ratio and With the Assumption of Uniform Flux as Constraints

Parameter	Units	Observed flux ratio		Uniform flux ratio	
		Median and range	Best fit	Median and range	Best fit
r_c	km	$779.72^{+124.45}_{-78.03}$	721.51	$784.68^{+127.72}_{-82.05}$	719.65
h_m	km	$745.86^{+131.19}_{-140.39}$	866.63	$695.45^{+155.71}_{-155.36}$	629.72
h_a	km	$200.58^{+124.68}_{-89.98}$	173.98	$249.50^{+154.46}_{-131.78}$	374.73
h_l	km	$73.70^{+18.94}_{-29.50}$	60.58	$75.07^{+18.00}_{-28.77}$	98.59
ρ_c	kg m ⁻³	$6478.04^{+1012.29}_{-980.32}$	7147.4	$6421.47^{+1039.03}_{-974.80}$	6772.4
ρ_m	kg m ⁻³	$3299.81^{+63.68}_{-44.97}$	3,316.5	$3300.86^{+63.38}_{-46.36}$	3,390.4
ρ_l	kg m ⁻³	$2957.941^{+184.99}_{-176.73}$	2967.77	$2957.08^{+184.51}_{-175.70}$	2882.04
$\log_{10}(\mu_m)$	Pa	$10.53^{+0.21}_{-0.24}$	10.39	$10.50^{+0.22}_{-0.23}$	10.77
$\log_{10}(\mu_a)$	Pa	$9.45^{+0.90}_{-0.87}$	9.022	$9.77^{+0.67}_{-0.92}$	9.375
$\log_{10}(\mu_l)$	Pa	$10.42^{+0.29}_{-0.29}$	10.08	$10.44^{+0.28}_{-0.30}$	10.16
$\log_{10}(\eta_m)$	Pa s	$21.14^{+1.95}_{-2.30}$	22.78	$18.52^{+2.59}_{-1.57}$	16.99
$\log_{10}(\eta_a)$	Pa s	$13.25^{+0.82}_{-1.04}$	12.95	$14.11^{+0.71}_{-1.18}$	14.67
$\log_{10}(\eta_l)$	Pa s	$21.54^{+1.68}_{-1.71}$	21.18	$21.47^{+1.72}_{-1.70}$	21.70
$\log_{10}(\zeta)$	—	$0.17^{+1.27}_{-1.42}$	−0.618	$−0.42^{+1.49}_{-1.13}$	−0.518
$J_2^{\text{hyd}} (\times 10^6)$	—	$24.56^{+7.20}_{-4.35}$	19.02	$24.60^{+7.23}_{-4.32}$	19.15
$C_{22}^{\text{hyd}} (\times 10^6)$	—	$1.70^{+2.18}_{-1.24}$	0.0333	$1.71^{+2.20}_{-1.24}$	0.0465

Note. These values are generated using $n = 0.21$ and are only valid for that value.

model containing an upper mantle layer. In addition, we force the viscosity in the lower mantle, the upper mantle, and the asthenosphere to be larger than 10^{15} Pa s.

With the addition of an extra layer, the inversion does not produce a better fit to the geophysical parameters (see Figure S8 in Supporting Information S1). The higher minimum viscosity also causes the posterior distributions of the parameters to deviate from a Gaussian distribution. Instead, they are concentrated at the edges of their prior ranges, signaling that the inversion prefers viscosities that are lower than the 10^{15} Pa s limit (see Figure S7 in Supporting Information S1). While this does not statistically exclude a five-layer model with higher viscosities, it does show that a more complicated interior model is not necessary to explain the polar-to-equatorial heat flux ratio.

4.5. Preferred Interior Parameters

To allow our interior parameter values to be used in future studies, we include here both the values obtained in the inversion using the observed flux ratio and those obtained using the globally uniform flux estimate as constraints. The median values and their one-sigma interval (16% and 84%) for each interior parameter that we find in our inversions are given in Table 3. We note that taking all the median values does not lead to a good-fitting model because not all the posterior PDFs are purely Gaussian. Furthermore, freely varying parameters within the given range can violate constraints such as total mass, and so should be done with caution. Therefore, we also give the parameter values for the best-fitting model (with the highest log-likelihood function) in Table 3. However, given how, for some parameters, the 68% confidence interval still covers orders of magnitude, we stress that these best-fitting values are not the ground-truth, and should only be used as a starting point in future studies. By the nature of the MCMC inversion, it is also possible that the actual values lie outside our one-sigma interval, especially as new observational constraints are added, as we have seen here with the introduction of the polar heat flux ratio as a constraint.

With the addition of the polar-to-equatorial flux ratio and the measurement of $\Re\{k_2\}$, the parameter space has shrunk significantly compared to previous studies (e.g., de Kleer, McEwen, et al., 2019; de Kleer, de Pater,

et al., 2019; Segatz et al., 1988). The results now exclude all interior model parameters that lead to only deep-mantle heating. Furthermore, parameter value ranges for asthenospheric heating models have also been shrunk due to the strong constraint on the mantle shear modulus μ_m and asthenosphere shear modulus μ_a from $\Re\{k_2\}$. Using Andrade rheology has also increased the viscosity values we find compared to for example, Segatz et al. (1988); Tyler et al. (2015). The median values and their one-sigma interval (16% and 84%) for each interior parameter that we find in our inversions are given in Table 3. In general, the found ranges show agreement with previous studies looking at tidal dissipation in Io (Bierson & Nimmo, 2016; Kervazo et al., 2022; Segatz et al., 1988).

4.6. Implications of the Low Viscosity Asthenosphere

The posterior distribution of the asthenosphere viscosity peaks around 10^{13} Pa s, with most values within the range 10^{12} – 10^{14} Pa s. This is a very low viscosity compared to typical values found in rocky bodies (Broquet et al., 2025; Karato & Wu, 1993). We discuss three possible reasons for these low viscosities: the rheological model, the rheological effect of melt on rock, and the assumed creep mechanism.

While currently the ζ model is the most used rheological model for tidal heating computations (e.g., Běhounková et al., 2021; Kervazo et al., 2021; Renaud & Henning, 2018), there are still large unknowns. Most experimental work (e.g., Faul & Jackson, 2015; Jackson et al., 2004; Yamauchi & Takei, 2024) is on seismic attenuation in the upper mantle of the Earth, where conditions are different from Io and from tidal heating applications in general (Bierson, 2024). Furthermore, there has been no experimental work directly on the ζ model, making it solely based on secondary analysis of laboratory data (Bierson, 2024; Castillo-Rogez et al., 2011; Walterová et al., 2023). The lack of dedicated experiments is a factor in the large uncertainty in the value for ζ , which has a large influence on the amount of dissipation at a given viscosity (Walterová et al., 2023). However, we found that even the lowest bound value of 10^{-2} that is often used (Bierson, 2024; Walterová et al., 2023) is not enough to lift the viscosity to above $\sim 10^{15}$ Pa s.

The presence of melt in the asthenosphere is an obvious cause of a low-viscosity asthenosphere, given its large effect on viscosity (Scott & Kohlstedt, 2006) and the obvious prevalence of lava on the surface. Using three different experimental relations between viscosity and melt fraction, Figure 5 shows the average melt fraction that is needed to explain the low viscosity (Costa et al., 2009; Scott & Kohlstedt, 2006). These relations are normalized with respect to the material's viscosity at the solidus temperature, η_{sol} , that is, the temperature at which the material begins to melt. The blue, solid line in Figure 5 is taken from Costa et al. (2009) and shows that the relation between viscosity and melt fraction has two distinct domains. For melt fractions below ~ 0.3 the relation follows a power-law relation, and above a melt fraction of ~ 0.3 the behavior changes to more fluid-like, and thus the viscosity drops significantly. The power-law relation can be characterized as $\eta(\phi) = \eta_{sol}e^{-B\phi}$, with B a proportionality constant which depends on the creep mechanism. For diffusion creep, B is 21, while for dislocation creep it is 32 (Scott & Kohlstedt, 2006). Dislocation creep thus results in lower effective viscosities (Karato & Wu, 1993).

To match the best-fitting viscosity assuming diffusion creep, which is almost always the assumption for Io, and using a solidus viscosity of 10^{19} (Karato & Wu, 1993), there would need to be a melt fraction of $> 30\%$. However, at such high melt fractions, viscous dissipation likely breaks down (Hay et al., 2025). At slightly lower melt fractions of 20%, Darcy dissipation might contribute to the tidal dissipation, reducing the need for lower viscosities, but this seems unlikely at realistic grain sizes and melt viscosities (Hay et al., 2025). Besides, studies that modeled magma advection in Io have shown that such melt fractions are hard to reach with Io's tidal dissipation of ~ 100 TW (Miyazaki & Stevenson, 2022; Spencer et al., 2020). Miyazaki and Stevenson (2022) show that melt fractions above 20% are only possible if all dissipation happens within a thin region and if the melt has a very high viscosity ($\eta_l > 100$ Pa s). Thus, melt is likely not enough to explain the effective viscosities we find, which is consistent with other studies that found large melt fractions to match Io's dissipation (Bierson & Nimmo, 2016; Kervazo et al., 2022).

One mechanism that could lead to lower effective viscosities is the possibility that Io is dominated by dislocation creep. This would lead to a stronger decrease in viscosity due to melting (Scott & Kohlstedt, 2006), as well as a lower solidus viscosity (Karato & Wu, 1993). For dislocation creep to occur there must be significant stresses. However, tidal stresses, which are on the order of 10–100 kPa, are likely not high enough to cause a significant viscosity drop due to dislocation creep. New research into transient dislocation creep behavior might change this

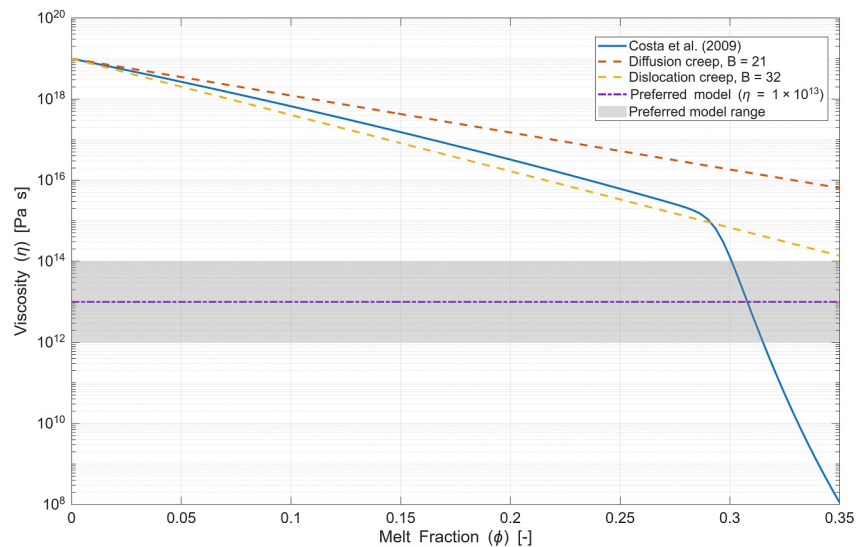


Figure 5. Viscosity values as a function of melt fraction in different regimes. Below melt fractions of ~ 0.3 , the relation between viscosity and melt fraction follows an exponential decay depending on the creep mechanism, present in both the blue line and the two dashed lines. Above this critical melt fraction, the solid structure begins to break down, with an even stronger melt dependency. The blue line captures both regimes, and is based on experiments with crystal-bearing melts (Costa et al., 2009). The two dashed lines are based on deformation experiments of melt-bearing, solid rock in two creep mechanism regimes (Scott & Kohlstedt, 2006). The purple dashed-dotted line and the gray area are the preferred viscosity and its uncertainty range we find in our inversion.

perspective (Breithaupt et al., 2023), especially given the periodicity of tidal stresses and their relatively high frequency (Hein et al., 2025), but this will require further investigation.

The three possible explanations listed above for the low-viscosity asthenosphere are not mutually exclusive and are also by no means exhaustive. Given that low viscosities and/or large melt fractions have been found before in tidal dissipation studies (Bierson & Nimmo, 2016; Kervazo et al., 2022), a more comprehensive study into possible explanations seems prudent. In particular, additional experimental work on anelastic deformation under Io-like conditions is needed. For example, while the tidal stresses might not be high enough to trigger dislocation creep, the physics of repeated stresses in the same direction could lead to different, unknown behavior, and it has been shown that pre-stressed material shows higher seismic attenuation (Farla et al., 2012). However, until these experiments have been done, much of this will remain speculative.

5. Conclusion

In this study, we incorporated both gravity and volcanic activity observations to obtain the most likely spherically symmetric interior model of Io, and our inversions point toward a partially molten Io with predominantly tidal heating in the asthenosphere. We used both static and tidal gravity measurements, including the recently measured real part of Io's tidal Love number, $\Re\{k_2\}$. As a measure of Io's heat flux distribution, we used the polar-to-equatorial heat flux ratio, and show that metric is highly sensitive to radial rheology changes. Because of their synergy, our approach of combining geophysical measurements in a Bayesian framework significantly constrains the possible values of interior parameters. The inversions indicate that Io must have a sizable ($h_a = 200.58^{+124.68}_{-89.98}$ km) low-viscosity asthenosphere beneath the crust, even with conservative heat flux ratio estimates. We also present updated values for interior models that can be used in future simulations of Io. However, to explain the low viscosity asthenosphere requires high melt fractions of $>30\%$, which may be challenging to produce unless other mechanisms are at play, such as dislocation creep or heightened dissipation at the tidal frequency from as-yet poorly understood transient rheological laws. This highlights that more experimental work on the rheological behavior of melt-bearing silicates, especially under Io-like conditions, is needed.

Besides gravity and volcanic activity observations, there are several other observables that could help tighten the constraints on Io's interior. As we showed in this work, heat flux measurements are essential to constrain the radial

dissipation profiles, and thus Io's interior properties, but their use is currently limited by the unresolved background heat flux. Ongoing efforts using the Juno Microwave Radiometer (Bolton et al., 2024; Brown et al., 2026) and Earth-based observations with ALMA or VLA might provide the first estimates of this quantity, while a dedicated Io mission, such as the Io Volcano Observer (Hamilton et al., 2025), could deliver more precise, higher-resolution global maps. Such a mission would also improve Io's static and time-dependent gravity field measurements. A better determined k_2 would place tighter constraints on the interior and, depending on the accuracy, might even detect the lateral structure expected to arise from the feedback between melt generation and tidal heating (Veenstra et al., 2025).

Topography offers a further route to the interior through isostasy. Degree-2 topography variations can be converted into lithosphere thickness variations, which in turn can be linked to tidal dissipation (O'Reilly & Davies, 1981; Spencer et al., 2021). This route is presently blocked by the accuracy of Io's global shape, as the current uncertainty exceeds the expected topographic signal and would need to be at least halved, from 0.3 to 0.15 km, before it could be used directly (Gyalay & Nimmo, 2024). Additionally, measurements of Io's libration amplitude could provide independent information on the thickness and rigidity of the lithosphere (Van Hoolst et al., 2020). Finally, Io's magnetic environment offers another window into its interior (Khurana et al., 2011), although the interpretation of Galileo's magnetometer data remains debated (Blöcker et al., 2018; Roth et al., 2017) and we therefore did not incorporate it here. Any of these additional constraints may, of course, shift the inferred interior properties outside the one-sigma ranges reported in this work.

In this work, we focused on the radial structure of Io. However, it is likely that both the mantle and lithosphere exhibit lateral variations (Davies, Perry, Williams, Veeder, & Nelson, 2024; de Kleer, McEwen, et al., 2019; de Kleer, de Pater, et al., 2019; Steinke, Hu, et al., 2020; Veenstra et al., 2025). Variations in melt fraction can also be amplified in a feedback mechanism, producing significantly altered tidal heating patterns (Veenstra et al., 2025). Furthermore, we have ignored the effect of melt advection on the relation between volcanic activity and tidal heating (Spencer et al., 2021). While we do not model these more complex behaviors, our results for the latitudinal distribution seem to be valid as a baseline structure, on top of which lateral variations exist. Therefore, our best-fitting interior structure parameters can serve as the basis for future studies investigating these more complex mechanisms.

Conflict of Interest

The authors declare no conflicts of interest relevant to this study.

Availability Statement

The software used in this article, which can be used to reproduce the results, LOV3D, is available in a repository at Rovira Navarro et al. (2025). The data and code are also published on Zenodo (<https://doi.org/10.5281/zenodo.19854018>) (Veenstra et al., 2026).

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