

New Constraints on the Distribution of Silicate Melt and the Depth of Tidal Heating in Io's Mantle

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Introduction: Io's prodigious active volcanism is maintained by tidal dissipation within its partially molten interior. There are three fundamentally different models for tidal dissipation inside Io. Two of these models treat Io as a solid visco-elastic body, using either a Maxwell [1, 2] or the more complex Andrade [3] rheology law. The third model treats Io as a magma ocean whose dynamics are akin to ocean tides on Earth [4].

We focus here on solid body models, which are usually classified into two end-member models, involving either a shallow, weak asthenosphere or deep tidal heating in a body with similar visco-elastic properties at all mantle depths. It has sometimes been suggested that Io is approximately 2/3 shallow heating and 1/3 deep heating [5, 6], with more recent studies favoring a heating distribution closer to 1/2 deep and 1/2 shallow [7]. It is important to recognize, however, that because these end member tidal heating distributions are derived using different mechanical layerings inside Io, the resulting tidal heating distributions can not be simply averaged. Rather, one must derive physical models for Io's visco-elastic structure that are consistent with available geophysical constraints, which is the focus of the work discussed here.

Based on our models and comparison with a variety of geophysical constraints, we conclude that Io's mantle must be a magma mush that is close to its rheologically critical melt fraction to a depth of many hundreds of kilometers. This model resembles the deep heating model of past studies. Our results are inconsistent with a thin, low strength asthenosphere.

Models: We use the numerical tidal dissipation model TiRADE (Tidal Response And Dissipation of Energy, [8]) to calculate tidal dissipation of Io. TiRADE solves for tidal deformation using a propagator matrix method, which requires that the rheological properties be strictly a function of radius and cannot vary laterally. We model Io as a Maxwell visco-elastic body, with a liquid core and three layers of varying strength in the silicate mantle. The lithosphere is assumed to be 20 km thick, with a viscosity of 10^{25} Pa-s and a shear modulus of 25 GPa.

The rheology of the upper mantle depends on the melt concentration. Peridotite has a modest decrease in viscosity at low melt fraction followed by a sharp decrease in viscosity in the range 25-30% partial melt

[9, 10]. This sharp drop in viscosity over a narrow melt fraction is termed the rheologically critical melt fraction (RCMF). In this melt fraction range, the peridotite solid + melt system is transitioning from interconnected solid grains coated with a small amount of melt to disconnected grains that are fully coated by melt. It is this transition in the physical state that leads to the rapid transition in viscosity at the RCMF. The shear modulus also decreases with temperature and melt fraction. This is constrained by experiment at low melt fraction [11]. At high melt fraction, the shear modulus goes to zero, and we therefore assume that this occurs at the same RCMF range that governs the viscosity. The lower mantle in our model is assumed to be at or just below the peridotite solidus, with a viscosity of 10^{20} Pa-s.

Results: The key model parameters in this work are the rheology (viscosity and shear modulus) of the low strength upper mantle and the depth of the transition between upper and lower mantle. Although the weak layer is often referred to as the asthenosphere in other studies, this nomenclature sometimes carries the implicit understanding that the asthenosphere is relatively thin, only a small fraction of the overall mantle thickness. Because the low strength layer can be quite thick in our models, we refer to this layer as the upper mantle and refer to the underlying, stronger layer as the lower mantle.

A variety of geophysical observations can be used to constrain the values of these model parameters. Magnetic induction observations of Io by the Galileo spacecraft require the presence of a shallow melt layer that is at least 50 km thick with at least 20% partial melt [12]. However, the magnetic induction observations do not provide upper bounds on either the melt fraction or the thickness of the melt-rich layer.

Observations of the heat flux out of Io provide important additional constraints. Io is estimated to have a total global heat flow of about 10^{14} W [13, 14]. This can be used to estimate the rheological properties of Io's upper mantle. Figure 1 shows the total tidally dissipated power as a function of upper mantle viscosity for upper mantles with thicknesses of 50 km (blue line with diamonds), 200 km (black line with squares), and 600 km (red line with triangles). As the upper mantle viscosity decreases, tidal dissipation increases and thus the total heat flow increases. However, when the system exceeds the

RCMF, the crash in both viscosity and shear modulus causes a sharp drop in tidal heating [15]. Io's observed heat flux can be explained by any of these mantle thickness models. When the upper mantle is thin (blue line), a very low viscosity is needed to explain the heat flux, while for thicker layers (red and black lines), the necessary dissipation can be achieved at higher viscosities. Although the required upper mantle viscosities differ by a factor of 100, all of these models reach their peak tidal dissipation between 26% melt fraction (9.4×10^{13} Pa-s) and 28% melt fraction (6.2×10^8 Pa-s), in the rheologically critical melt fraction range of 25-30% for peridotite [10]. Moreover, it is clear that even if Io's overall heat flux was mis-estimated by a factor of 2 or 3, only a small change in melt fraction would be required to meet the revised heat flux.

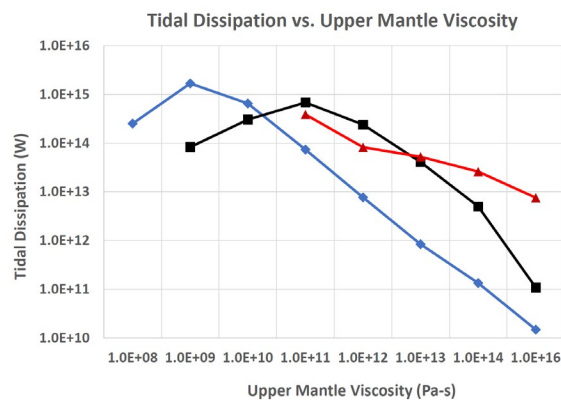


Figure 1: Tidal dissipation as a function of the viscosity of the upper mantle layer. Blue line with diamonds is for a 50 km thick upper mantle weak layer, black line with squares is for a 200 km thick upper mantle layer, and red line with triangles is for a 600 km thick upper mantle layer.

It is well known that the thickness of the upper mantle weak layer has a strong effect on the spatial distribution of the tidal heating. Models with a shallow asthenosphere have maximum heating at the equator and mid-latitudes and zero tidal heating at the poles. Models with mostly deep heating have maximum dissipation at the poles but vary only by a factor of ~ 2 from pole to equator [2, 5, 6]. Measurements of volcanic heat flux from the Galileo Near Infrared Mapping Spectrometer are larger at the equator than the poles, but the distribution is much flatter with latitude than for a model with a thin asthenosphere [7, 13, 16]. In our models, this distribution is best explained by a deep upper mantle layer, extending to at least 600 km below the surface and possibly close to the core mantle boundary. A limitation of the

NIMS data is that it only measures heat flux where active volcanism is occurring. Mountains on Io are distributed at all latitudes, but the long-wavelength component of the distribution of mountains on Io peaks at equatorial to mid-latitudes [17]. If the high mountain topography, which sometimes exceeds 10 km elevation [18], requires support from a thicker than average lithosphere, this observation favors the presence of regions of low heat flux near the equator. This again supports a predominantly deep heating model. The magma ocean tidal heating model [4] also has a strong equator to pole gradient in the predicted tidal heating. Although we have not formally modeled ocean tides, the spatial patterns of the observed heat flux and mountain distribution likely also requires a strong deep tidal heating component even if there is a shallow magma ocean.

A final observational constraint comes from the degree 2 gravitational tidal Love number, k_2 , which was determined from Doppler tracking of the Galileo spacecraft [19]. An initial search of the model parameter space shows that the observed value of k_2 is inconsistent with a shallow, weak asthenosphere and is best explained by a melt concentration in the rheologically critical range that extends for many hundreds of kilometers into the mantle. This conclusion is consistent with that reached using the spatial distributions of heat flow and mountains on Io. This model with a very thick, low strength upper mantle resembles the deep heating model of [1, 2]. On the other hand, our results are strongly inconsistent with a thin, shallow asthenosphere on Io.

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