Internal structure of Mars: viscosity constraints from short period tides and loads

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Introduction: We present a suite of radially stratified visco-elastic models for the internal structure of Mars. One of the more challenging constraints for such models is to simultaneously support significant long wavelength topography and accommodate the observed rate of secular evolution of the orbit of Phobos. The latter indicates that, at sub-diurnal periods, Mars is more dissipative than Earth's mantle. Our favored (though non-unique) explanation is that there is a relatively shallow partial melt zone in the mantle of Mars.

Background: In the absence of seismic data, internal structure models of Mars depend heavily upon remotely accessible observations, such as gravity, topography, and response to periodic forces. The response of Mars to annual solar tidal forcing is sensed via the resulting gravitational perturbation to spacecraft orbits, and is characterized by a degree two tidal Love number [1,2]. The secular orbital acceleration of Phobos depends upon both the amplitude and phase lag of the tide it raises on Mars, with the latter characterized by a tidal Q [3,4]. The gravitational effect of seasonal mass loads, associated with volatile transport into and out of the polar regions, has already been measured [2,5] and is potentially measureable with sufficient accuracy to distinguish the response of the "solid" body to that load.

For simplicity, we assume a spherically symmetric and incompressible model. At each level within Mars layer we thus need to specify density, rigidity, and viscosity. The density structure is only directly constained by two observations, the mean density [6] and polar moment of inertia [7]. These two constaints, and the stipulation of monotonic variations, defines an envelope within which the density must lie [8], but does not provide more definitive results without additional geochemical constraints [9,10,11].

Plausible solid materials within Mars have a relatively narrow range of rigidity values, but might have viscosity values spanning many orders of magnitude, depending upon composition, temperature, and pressure [12,13,14]. We are chiefly interested in the inverse problem of finding relatively simple models which satisfy the relevant geodynamical constraints.

Method: For a layered elastic model, we use the formulation of [15] to construct a solution to the appropriate differential equations (elastic constitutive relation, Poisson's equation, and momentum conservation). In consideration of surface loading on Maxwell viscoelastic bodies, a Laplace transform on the time dependence is often used [16,17]. However, since the forcing of interest here is periodic, we use a Fourier transform. When elastic and viscous elements are coupled in series, the resulting complex modulus, at forcing frequecy ω is

$$\tilde{\mu}[\omega] = \mu\left(\frac{I\,\omega\,\tau}{1 + I\,\omega\,\tau}\right)$$

where $I = \sqrt{-1}$ and the Maxwell relaxation time τ is the viscosity η divided by the rigidity μ

The Fourier transformed Love numbers for a layered Maxwell model have the form

$$k[\omega] = k_e + \sum_j \frac{I\, r_j}{\omega - I\, p_j}$$

where p_j and r_j are poles and residues, respectively. The poles are the same for both tide and load responses, but the residues are different. In our model, we adjusted the Maxwell times in the core and mantle to match the tidal forcing at annual and Phobos orbital periods. The immediate response is dominated by the first term k_e which is the purely elastic response, and the long term, or fluid response is given by the limit $\omega \rightarrow 0$

$$k_f = k[0] = k_e - \sum_j \frac{r_j}{p_j}$$

That is obviously also the value that would result from computing the elastic Love number for a body in which all of the viscous layers have their rigidities set to zero. In our case, only the outer elastic layer retains long term strength.

Preliminary Results: The simplest case we evaluate is a 3-layer model (elastic lithosphere, viscoleastic mantle and core). For this model to reproduce the degree 2 Love number estimates and the secular acceleration of Phobos, we require a mantle relaxation time of ~10^7sec, and a core relaxation time of ~10^5 sec. Using a simplified representation of the annual polar ice loads, we calculate the response of this model at harmonic degrees 2 and 3 (Figure 1). We find that the response to annual forcing is substantially smaller in magnitude than expected, which is surprising given the extremely low mantle viscosity implied by the mantle Maxwell relaxation time.

One interpretation of the relatively large size of the degree 2 Love number is that it implies existence of a substantial fluid core [1]. We are currently testing models that employ a fluid core and additional layers within the mantle. It is possible that a low-viscosity layer (e.g., a layer of partial melt) could provide the energy dissipation required to satisfy the Phobos orbital constraints without invoking an unrealistically low viscosity throughout the mantle.

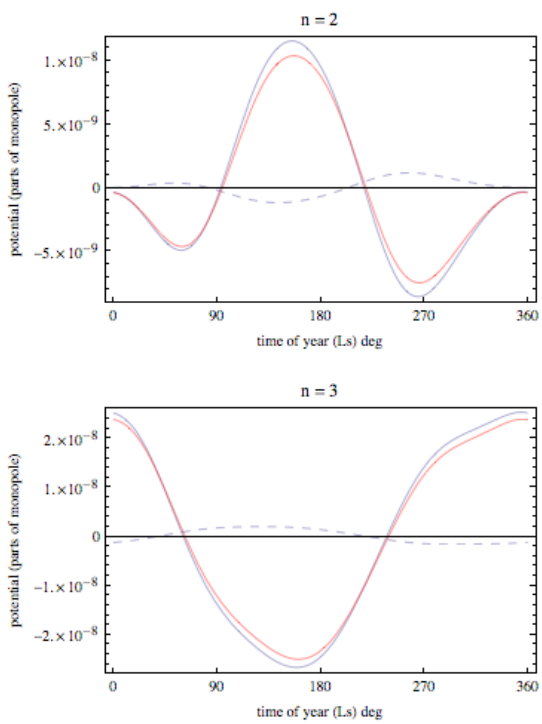

Figure 1. Response of 3-layer Maxwell viscoelastic model at degrees 2 (top) and 3 (bottom) to forcing arising from annual changes in polar ice cap mass load. In this model, elastic lithosphere is 150 km thick. Solid blue curve: potential due to load; dashed curve: potential due to response of Mars model; solid red curve: total observed potential (load + model response).

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