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Introduction: Planetary bodies typically exhibit a phase lag between an applied stress and a response, leading to dissipation (heating). This phase lag ϕ can be measured at frequencies f from ~1 Hz (seismic attenuation[1]) to ~1nHz (Earth tides [2]), and is described by the dissipation factor Q, where one possible definition is Q^{-1}=tan ϕ [3]. Large values of Q denote small dissipation. Phase lags can also be measured in experimentally-deformed samples [3-5], over a smaller frequency range.

Phase lags and dissipation are central to the evolution of planetary satellites, because they control the rate of rotational despinning [6], orbital circularization [7] and tidal heating [8]. Unfortunately, there is currently a disconnect between orbital treatments of Q and the experimental and observational constraints.

Satellite evolution is typically modelled either assuming that Q is constant [9], or that the satellite behaves as a Maxwell viscoelastic body [10], which in simple cases gives Q~f. (Note that tidal *heating*, as opposed to Q, exhibits a more complex frequency dependence [11]). On the other hand, both experimental [3-4] and observationally-constrained [1-2] values suggest that Q~f^{α}, where α~0.2-0.4. This frequency-dependence can give rise to satellite behaviour differing from conventional results [12].

Part of the disconnect arises from the fact that very different mechanisms may be responsible for dissipation in experimental samples (e.g. dislocation glide [13]) compared with planets (e.g. inertial waves [14]). It is also likely that dissipation in a complex body like a satellite is not well-represented by a simple phase-lag description.

Dissipation: Measurement of phase lags in silicate samples using a torsion apparatus is now relatively straightforward [3-4], but much less work has been done on icy materials [cf. 5]. An advantage of the torsion apparatus is that the periods (~10^5 s) and strain rates (~10^{-10} s^{-1}) appropriate to icy satellites are achievable. Despite the potential pitfalls of relating experimentally-determined Q values to real satellites, **measurement of Q and its frequency-dependence for icy materials would represent a major step forward**. It may also be important to look at the behaviour of partially-molten systems, which tend to exhibit different responses [4].

A factor which complicates ice rheology is its tendency to undergo transient creep [15]. This behaviour may be especially important at the relatively low strains and short timescales characteristic of tidal deformation [Melosh, pers. comm.]. One way of representing this behaviour is with an Andrade model [3], where the strain ε is given by

$$\varepsilon = \sigma\left(\frac{1}{G} + \frac{t}{\eta_{ss}} + \beta t^n\right)$$

Here σ is the stress, G the shear modulus, t is time, η_{ss} is the steady-state viscosity and β and n are empirical constants. The three terms on the RHS are, respectively, the elastic, viscous and transient response.

For this model, it may be shown that the corresponding dissipation function Q is given by

$$Q^{-1} = \frac{\frac{f^{n-1}}{\eta} + \beta \sin\frac{n\pi}{2}\Gamma(n+1)}{\frac{f^n}{G} + \beta\cos\frac{n\pi}{2}\Gamma(n+1)}$$

where here f is the angular frequency and Γ is a gamma function [3]. If the transient contribution is negligible (β~0) then Q~f, the usual Maxwell response. In the general case, at high frequencies (elastic limit) Q^{-1}~f^{-n} and at low frequencies (viscous) Q^{-1}~f^{n-1}.

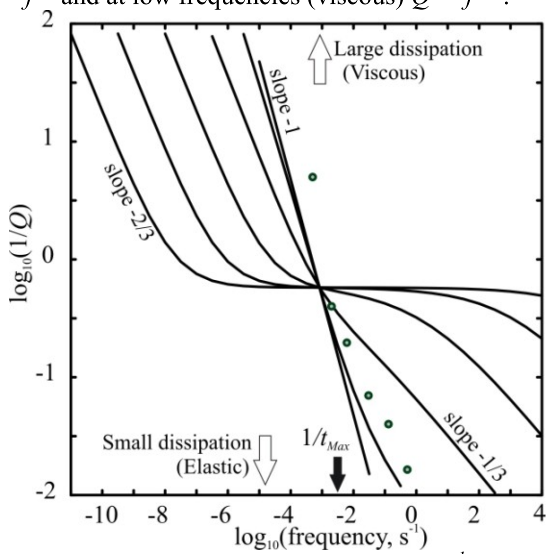

Figure 1. Frequency dependence of Q^{-1} for Andrade rheology. Shear modulus and viscosity are 3 GPa and 10^{12} Pa s, respectively, t_{Max} is the Maxwell time. Andrade model assumes n=1/3 and variable β. For small β, the behaviour is viscoelastic (Q^{-1}~f^{-1}), while at larger β, Q^{-1} shows a shallower frequency-dependence and becomes constant over an intermdiate frequency range. Dots are data for an ice-$MgSO_4$ compression creep test, from [5].

Figure 1 shows how Q changes as a function of the frequency and the strength of the transient component, and also plots the experimentally-determined values of [5]. The Andrade model is attractive because it has some theoretical justification [13], and can also reproduce the observed frequency-dependent behaviour of icy and silicate materials [3,5]. It may therefore become necessary to develop a theory for dissipation in multi-layered Andrade-rheology bodies, as has already been done for viscoelastic bodies [10].

Friction: Although most satellite models assume that dissipation is volumetric and viscoelastic, other mechanisms may also play a role. One such mechanism is frictional heating due to fault motion [16,17]. We will assume that the tidal stresses σ are capable of causing motion on a fault to a depth d, where $d=\sigma/\rho gF$, ρ is the surface density, g is gravity and F is the friction coefficient. In this case, the total heating H is given by

$$H = n^7 e^3 h_2^3 \frac{L_{tot} w \mu'^2 c^3}{2\rho g F G_g^3 \rho_b^3} \qquad (1)$$

where n is the mean motion, e the eccentricity, h_2 the Love number, L_{tot} and w the total fault length and mean separation, μ' an effective modulus, c a constant, G_g the gravitational constant and ρ_b the bulk density. This heating is different from the standard tidal dissipation, which depends on Q^{-1}, e^2 and n^5, and can lead to differing satellite evolution [17]. Heating increases as F decreases because faults can be active at greater depths if friction is lower.

Assuming global tectonic activity (that is, $L_{tot}w$ ~ satellite surface area), the expected mean frictional heat flux on Europa is 2 mWm^{-2} for F=0.3. For Enceladus, the expected value is 5 mWm^{-2} (h_2/0.1) for the same F. This value is larger mainly because of the lower gravity, and for h_2=0.1 represents a total heat flow of 4 GW, comparable to that observed at the South Pole [18]. This heat production is shallow, and does not significantly alter the ice shell temperature structure at depth.

An important factor in equation (1) is the friction coefficient F, which also controls the depth to which brittle faulting is likely to occur. At low strain rates ice demonstrates a constant friction coefficient similar to the behaviour of silicate materials [19]. However, at higher strain rates, more complicated behaviour results [20]. **It may therefore be of interest to more fully characterize the frictional behaviour of ice under the range of conditions appropriate to icy satellites**. It is also important to understand the role of melt in such systems. Melt can reduce the friction coefficient [e.g. 21], potentially giving rise to periodic behaviour in which frictional heating drops dramatically as soon as melt is produced.

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