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Tidal dissipation in the oceans of icy satellites

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ABSTRACT

Dissipation of tidal energy is an important mechanism for the evolution of outer Solar System satellites, several of which are likely to contain subsurface oceans. We extend previous theoretical treatments for ocean tidal dissipation by taking into account the effects of ocean loading, self-attraction, and deformation of the solid regions. These effects modify both the forcing potential and the ocean thicknesses for which energy dissipation is resonantly enhanced, potentially resulting in orders of magnitude changes in the dissipated energy flux. Assuming a Cassini state obliquity, Enceladus' dissipated energy flux due to the obliquity tide is smaller than the observed value by many orders of magnitude. On the other hand, the dissipated energy flux due to the resonant response to the eccentricity tide can be large enough to explain Enceladus' observed heat flow.

1. Introduction

Several lines of evidence suggest the presence of subsurface oceans in icy satellites of the outer Solar System. On Europa, the detection of an induced magnetic field has been explained as due to a layer of liquid salty water a few kilometers beneath the icy surface (Kivelson et al., 2000; Zimmer et al., 2000; Hand and Chyba, 2007). Similar to Europa, Ganymede and Callisto also exhibit an induced magnetic field, indicating the presence of a liquid salty water layer beneath their surfaces. However, the conducting layer must be a few hundreds of kilometer beneath the surface (Zimmer et al., 2000; Kivelson et al., 2002). On Titan, the detection of a Schumann-like resonance has been explained as due to a subsurface, conductive, liquid layer (Béghin et al., 2010). Similarly, estimates of the mean moment of inertia (Bills and Nimmo, 2011; Baland et al., 2011) and the degree-2 tidal Love number (Iess et al., 2012) suggest the presence of a subsurface liquid layer. On Enceladus, the composition of particles ejected from fissures across the south pole (Hansen et al., 2011; Postberg et al., 2011) and the observed gravity field (Iess et al., 2014) support models with a subsurface liquid water reservoir.

Dissipation of tidal energy is an important mechanism for the long-term evolution of icy satellites because it affects their thermal, rotational, and orbital states. By analogy with Earth, one might expect that ocean dissipation would dominate the energy loss. However, the majority of previous studies that have investigated tidal dissipation in outer Solar System satellites assume that dissipation occurs only in their solid regions (e.g., Poirier et al., 1983; Segatz et al., 1988; Ojakangas and Stevenson, 1989; Ross and Schubert, 1989; Sohl et al., 1995), despite evidence for the presence of subsurface oceans. The fewer studies considering ocean dissipation ignore ocean loading, self-attraction, and deformation of the solid regions (Sears, 1995; Tyler, 2008, 2009, 2011; Chen and Nimmo, 2011; Chen et al., 2013). Therefore, these studies are only applicable to idealized bodies with infinite rigidity and without self-gravity. As a notable example, Tyler (2011) concluded that resonant tides driven by forces associated with obliquity and eccentricity provided a plausible source for Enceladus notable global heat flux. However, Tyler (2009) suggested that the obliquity of Enceladus (which has not been directly measured) could be very small, and this was confirmed by Chen and Nimmo (2011) on the basis of dynamical considerations. Chen and Nimmo (2011) have argued that the obliquity of Enceladus should be less than $\sim 0.002^{\circ}$ and they thus discounted the possibility of a non-negligible contribution to heating from obliquity tides.

We extend previous theoretical treatments for ocean tidal dissipation by taking into account the effects of ocean loading, self-attraction, and deformation of the solid regions. The rest of this paper is organized as follows. Section 2 extends previous theoretical treatments by taking into account the effects described above. Section 3 presents the difference between cases where these effects are considered against cases where they are not. In Section 4, we reassess arguments related to tidal heating on Enceladus using the extended theoretical treatment.

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2. Theory

2.1. Equilibrium tide theory

The equilibrium tide theory assumes that the surface of the ocean coincides at all times with a gravitational equipotential. Sagan and Dermott (1982) used this theory to consider tidal dissipation on a surface ocean on Titan. The equilibrium tide theory ignores the dynamics of the ocean movement. That is, it assumes that the ocean responds instantaneously to the imposed tidal potential. However, it is useful to consider this theory because it provides insight into the effects of ocean loading, self-attraction, and deformation of the solid regions caused by both the external tidal potential and ocean loading.

The equilibrium tide on a satellite with infinite rigidity is

$$\eta^{eq} = U^T/g, \tag{1}$$

where U^T is the imposed tidal potential, η^{eq} is an equipotential ocean surface and g is the surface gravity. If self-attraction and finite rigidity (i.e., deformation of the solid regions) are taken into account, the equilibrium tide is given by

$$\eta^{eq} = (1 + k_2 - h_2)U^T/g, \tag{2}$$

where k_2 and h_2 are the degree-2 tidal and tidal displacement Love numbers. The ocean bottom is lifted by $h_2 U^T/g$, and the additional gravitational potential arising solely from this mass redistribution is $k_2 U^T$. Therefore, $1 + k_2$ is a factor allowing for the self-attraction of the solid regions, and the response by $h_2 U^T/g$ takes this self-attraction into account. Eq. (2) is an expression for the amount an ocean surface covering the satellite is lifted relative to the ocean bottom. Because we are considering a surface ocean in this paper, the Love numbers must be evaluated at the bottom of the ocean. If ocean loading, and self-attraction and deformation of the solid regions in response to ocean loading are also taken into account, the equilibrium tide for a satellite with finite rigidity can be written as (Agnew and Farrell, 1978)

$$\eta^{eq} = \gamma_{eq} \frac{U^T}{g}, \tag{3}$$

where the equilibrium tide amplification factor is given by

$$\gamma_{eq} \equiv \frac{1 + k_2 - h_2}{1 - \frac{3\rho_o}{5\bar{\rho}}(1 + k_2' - h_2')}. \tag{4}$$

In Eq. (4), ρ_o is the ocean density, $\bar{\rho}$ is the mean density of the solid regions beneath the ocean, and k_2' and h_2' are degree-2 load and load displacement Love numbers evaluated at the bottom of the ocean.

The tidal and load Love numbers depend on the interior structure and rheology of the satellite. These dimensionless numbers can be calculated using the classical propagator matrix method (e.g. Sabadini and Vermeersen, 2004). In this method, the Love numbers are computed by solving mass and momentum conservation equations, and Poisson's equation at each layer. The propagator matrix method can be used to calculate the Love numbers for any spherically symmetric interior structure, including radial variations in density and rigidity.

Assuming a uniform interior with density ρ and rigidity μ, the degree-ℓ Love numbers are given by (e.g., Munk and MacDonald, 1960)

$$\{k_\ell, h_\ell, k_\ell', h_\ell'\} = \frac{1}{1+\bar{\mu}}\left\{\frac{3}{2(\ell-1)}, \frac{2\ell+1}{2(\ell-1)}, -1, -\frac{2\ell+1}{3}\right\}, \tag{5}$$

where

$$\bar{\mu} \equiv \frac{2\ell^2 + 4\ell + 3}{\ell}\frac{\mu}{\rho g R} \tag{6}$$

is a dimensionless effective rigidity. Table 2 shows the Love numbers for icy satellites assuming a uniform interior with the parameters in Table 1.

The denominator in Eq. (4) is a factor allowing for ocean loading, and self-attraction and deformation of the solid regions in response to ocean loading. This factor is always smaller than unity if $\rho_o < \bar{\rho}$ as observed in icy satellites; therefore, the combination of these effects increases the equilibrium tide. In the limit of a vanishing ocean, there is no ocean loading and $\gamma_{eq} = 1 + k_2 - h_2$. Independently of the interior structure, if the Love numbers are significantly smaller than unity, which is likely for typical rock and ice rigidities, γ_{eq} (Eq. (4)) is well approximated by

$$\gamma_{eq}' \equiv \left(1 - \frac{3\rho_o}{5\bar{\rho}}\right)^{-1}. \tag{7}$$

Table 2 compares γ_{eq} and γ_{eq}' for icy satellites assuming a uniform interior with the parameters in Table 1.

2.2. Laplace tidal equations

The Laplace tidal equations describe mass and momentum conservation taking into account the dynamics of ocean movement (Lamb, 1932; Longuet-Higgins, 1968). These equations describe tides in an incompressible ocean in the shallow water limit. That is, the ocean thickness $h \ll R$, where R is the radius of the satellite.

Table 1
Satellite parameters taken from Chen et al. (2013). Columns correspond to mean radius (R), mean density ($\bar{\rho}$), surface gravity (g), rigidity (μ), rotation rate (Ω), orbital eccentricity (e), and obliquity (θ_0). The rigidities are chosen so as to have values between typical of ice and rock rigidities using the formula $\mu = 4\,\text{GPa}\left(\frac{\bar{\rho}}{950\ \text{kg m}^{-3}}\right)$.

	R (km)	$\bar{\rho}$ (kg m^{-3})	g (m s^{-3})	μ (GPa)	Ω (10^{-5} rad s^{-1})	e	$-\theta_0$ (°)
Europa	1565.0	2989	1.31	12.6	2.05	0.0094	0.053
Ganymede	2631.2	1942	1.43	8.18	1.02	0.0013	0.033
Callisto	2410.3	1834	1.24	7.72	0.436	0.0074	0.24
Mimas	198.2	1150	0.064	4.84	7.72	0.0196	0.041
Encelauds	252.1	1610	0.11	6.78	5.31	0.0044	0.0027
Tethys	531.0	985	0.15	4.15	3.85	0.0001	0.039
Dione	561.4	1478	0.23	6.22	2.66	0.0022	0.002
Rhea	763.5	1237	0.26	5.21	1.61	0.0002	0.030
Titan	2574.7	1882	1.35	7.92	0.456	0.0288	0.32
Miranda	235.8	1200	0.079	5.05	5.15	0.0013	0.021
Ariel	578.9	1665	0.27	7.01	2.89	0.0012	0.0005
Umbriel	584.7	1399	0.23	5.89	1.76	0.0039	0.0026
Titania	788.9	1714	0.38	7.22	0.835	0.0011	0.014
Oberon	761.4	1630	0.35	6.86	0.540	0.0014	0.075
Triton	1353.4	2060	0.78	8.67	1.24	0.0000	0.35