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Ocean tides heat Enceladus

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[1] With a young, highly reflective surface and vigorous geological activity, Saturn's tiny moon Enceladus has been one of the most mysterious planetary bodies in the solar system. Recent observations from the Cassini spacecraft show vast plumes of vapor erupting from geysers near the south pole, and inferred heat fluxes of about 100 mW/m^2 for the same region have helped consolidate the essential enigma of Enceladus: there must be a relatively large and unidentified source of heat in the tiny moon. Here we present a case for heating from dissipation by tidal flow in an ice-covered ocean. We show that if the spin axis of Enceladus is tilted with respect to its orbital plane by at least 0.05 degree then strong tidal flow will be generated with enough dissipative heating to explain the observed heat flux. In an alternative case of a shallow (10 km or less) ocean, comparable flow velocities and heating may be obtained by eccentricity tidal forces. **Citation:** Tyler, R. H. (2009), Ocean tides heat Enceladus, *Geophys. Res. Lett.*, 36, L15205, doi:10.1029/2009GL038300.

1. Introduction

[2] Saturn's moon Enceladus is one of only three solid planetary bodies in the solar system with an internal heat source large enough to have been detected by remote sensing [Spencer *et al.*, 2006; Porco, 2006]. The other two—Earth, and Jupiter's moon Io—are much larger than Enceladus. The ratio of surface area to volume decreases with the size of the moon and so these larger moons present less of a mystery because there is relatively more volume for radiogenic and tidal-stress heat sources, and relatively less surface area through which heat is lost. But with a diameter of only 500 km, the volume of Enceladus is so small that a radiogenic source for the observed heat flux is ruled out. The possibility of extraordinary tidal stresses in the solid moon, such as is the case for Io, has been closely considered but also appear insufficient [Meyer and Wisdom, 2007; Tobie *et al.*, 2008; Roberts and Nimmo, 2008]. Indeed, the usual calculations of tidal heat depend on the orbital eccentricity of the moon, and the eccentricity of Enceladus (0.0047) is relatively small. The neighboring moon Mimas, for example, has a larger eccentricity (0.020) and for similar composition should receive about eleven times as much tidal heating as Enceladus [Meyer and Wisdom, 2007]. Mimas, however, is cool and geologically dead. An ocean layer under the ice of Enceladus has long been suspected as the source of Saturn's E-ring, and with the observations of Cassini an ocean on Enceladus seems

now to be very likely [Kargel, 2006; Collins and Goodman, 2007]. In this case the ice shell would become uncoupled from the solid moon. Several proposals have been put forth to describe how tidal stresses on an uncoupled ice shell can, under restrictive assumptions, lead to elevated heat fluxes similar to those observed [Nimmo *et al.*, 2007], but a consensus has not been reached and the heat source of Enceladus remains a mystery. An explanation must provide enhanced heat fluxes over the broad region south of about -55° latitude, an average of about 100 W/m^2 south of -65° , and peak local heat fluxes should associate with the four prominent troughs near the south pole dubbed the “tiger stripes”.

[3] Here we propose that the primary source of heat in Enceladus can be provided by dissipation of strong ocean tidal flow if either or both of the following conditions are met: a) Enceladus' unmeasured obliquity angle is 0.05 degree or greater; b) the ocean depth is shallower than about 10 km. When we consider that it has long been known that tidal dissipation on Earth occurs primarily in the ocean, and not in the solid earth or ice sheets, it may seem surprising that ocean tidal dissipation has not received much attention for the outer moons. This neglect is likely because previous models suggested meager tidal flow speeds for which dissipation would indeed be small, and because a modern understanding of ocean dissipation was not applied. With the assembly of several new considerations, a new view of strong tidal flow on these moons emerges. We shall first address these considerations because they are important in understanding the results we will present.

[4] First, tidal forces on the outer moons come in at least two different types. The one considered in all previous work involving tidal-heating effects is due to the eccentricity of the moon's orbit around the planet. A second tidal force is due to the moon's obliquity (the axial tilt of the moon's spin axis relative to its orbital plane). The reason for previous focus on the eccentricity tidal forces, and the consequent neglect of obliquity, has surely been because the eccentricity tidal forces are typically much larger. Let us make the point that when considering the response of a dynamical system to forces, a comparison based merely on the amplitudes of the forces is inadequate. Indeed, these two forces are not even in the same symmetry group, and as we shall see the response of the dynamical system (the ocean) is quite different in each case. Technically, the important difference is that the obliquity tidal forces include a component with a degree-two, order-one spherical-harmonic spatial form propagating westward around the moon with a diurnal frequency; while the eccentricity tidal forces do not have such a component.

[5] Second, it has probably appeared that the ocean response to tidal forcing should be a quasi-static one because the fastest waves the ocean has to respond with, the so-called “shallow-water waves,” are fast enough to circle the moon within the relatively long tidal periods.

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This quasi-static assumption is implicit in a previous calculation of the ocean tidal response of Europa [Moore and Schubert, 2000], but it is shown by Tyler [2008] that the correct obliquity tidal flow response is not obtained under this assumption. Notably, tidal forces can induce ocean flows that are tangentially non-divergent. With such flow there is no vertical motion and therefore gravity waves such as the fast shallow-water waves cannot help adjust these flows to the ever-changing tidal forces. In this case, it is other waves such as the Rossby-Haurwitz wave that become important.

[6] Third, until rather recently it was thought that almost all (~99 percent) of the tidal dissipation in Earth's ocean occurred on the continental shelves and shallow seas. And because oceans on the outer moons are usually considered to be global and deep, it would seem that the case of Earth, with its tidal dissipation predominantly in the ocean rather than the solid parts, would be a poor analogy for the case of the outer moons. Recent evidence from satellite as well as in-situ observations have indicated that this earlier view is incorrect, and that in fact about a third of the total tidal dissipation on Earth occurs in the deep ocean [Egbert and Ray, 2000; Jayne and Laurent, 2001]. This was not expected from the traditional model of dissipation based only on frictional boundary-layer theory for which the dissipation D is described by

$$D = \rho C_D u^3 \quad (1)$$

(where ρ is the water density, C_D is the drag coefficient, and u is the amplitude of the ocean flow velocity). Although the deep ocean carries a major fraction of the tidal kinetic energy, the flow velocities are small and therefore the dissipation, varying with the cube of the velocity, was expected to be especially small. Now it is appreciated that about 25–30% of the total tidal dissipation occurs in the deep ocean. In the modern view, significant tidal dissipation is generated by deep-ocean tidal flow as it crosses rough bathymetry (sea floor topography) and transfers energy to internal waves that eventually lose this energy to smaller-scale dissipative processes (heat).

[7] Fourth, we shall present ocean tidal-flow results that can reach amplitudes much larger than any previously discussed for these moons, and we will claim that the high amplitudes we find are due to a resonant excitation of Rossby-Haurwitz waves by the typically-disregarded obliquity tidal forces. But Rossby waves and tidal forces, both have been studied in comprehensive detail—at least in the application to Earth's atmosphere and ocean. If such an amplitude-raising resonance were to exist, would it not already be famously described in terrestrial applications? No; although Earth has a large obliquity angle, and this generates forcing of the right spatial pattern for the resonance, the frequency is wrong. On synchronously rotating moons the orbit and spin have the same frequency, but such is not the case for Earth. Hence, the conditions for the resonance we shall describe have been known for over a century [Hough, 1898; Longuet-Higgins, 1968] but there are not familiar examples of this resonance on Earth.

[8] The method for estimating the heat generated by the tidal flow (see auxiliary material for details) involves essentially three steps: 1) Calculate the tidal forces at

Enceladus; 2) Calculate the ocean flow response due to these tidal forces; 3) Calculate the dissipation (heating rate) due to this ocean flow.¹ Because the dissipation increases with the square or the cube of the flow velocity, much of the argument for ocean tidal heat amounts to identifying a scenario for strong tidal flow. This paper primarily addresses the scenario of non-zero (i.e. >0.05 degree) obliquity. We do not know the obliquity angle of Enceladus, but theoretical work for Europa [Bills, 2005] indicates a minimum of 0.1° is needed to satisfy orbital requirements. Such estimates tend to scale with the inclination which is small for Enceladus. Therefore, the obliquity of Enceladus could be quite small (B. Bills, personal communication, 2008), of the order 0.01 degree or even less. The recent observation of 0.3° for Titan (inferred from Cassini observations [Lorenz *et al.*, 2008]) is, however, about four times larger than such scaling referenced from the European value would suggest. Clearly, we do not know the obliquity of Enceladus. One could make a consistent argument, however, that the obliquity tides we present were indeed strong in the past but have now driven the obliquity angle to zero. As the dominant obliquity tidal heat decreases, the ocean thickness h decreases due to freezing. The dominant tidal flow is then due to eccentricity, but this flow increases with $1/h$ (see auxiliary material) and must therefore become significant for some small h . Interestingly, the h value for this transition is larger (≈ 10 km) for Enceladus than it is for the other moons with suspected oceans. While bigger moons like Europa have not likely spun down to transition from deep-ocean obliquity tides to shallow-ocean eccentricity tides, there is this possibility for Enceladus and so we include in places discussion of a second scenario describing the zero-obliquity, shallow-ocean case.

2. Results

[9] The obliquity tidal forces are described by the gravitational potential, which we calculate to be

$$\Phi = \frac{3}{2} \Omega^2 a^2 \theta_o \sin \theta \cos \theta (\cos(\phi - \Omega t) + \cos(\phi + \Omega t)), \quad (2)$$

where $\Omega = 5.3 \times 10^{-5} \text{ s}^{-1}$ is the rotation rate, $a = 252 \text{ km}$ is the radius, θ_o , θ , and ϕ are the obliquity, colatitude, and longitude (all in radians), respectively, and t (s) is time. More specifically, it is only the second of the cosine terms in (2) that is expected to deliver the resonant response and we focus on just this component. (Physically, the first term can not excite this Rossby mode because it travels only westward.)

[10] The response of the ocean to this forcing is calculated using the Laplace tidal equations [Hough, 1898; Longuet-Higgins, 1968]. This particular application is a rare case where a closed-form analytical solution to these equations can be found. The solution for the flow can be given compactly in terms of a stream function

$$\psi = \frac{3}{2} \Omega a^2 \theta_o \sin(\theta) \cos(\phi + \Omega t) \quad (3)$$

¹Auxiliary materials are available in the HTML. doi:10.1029/2009GL038300.

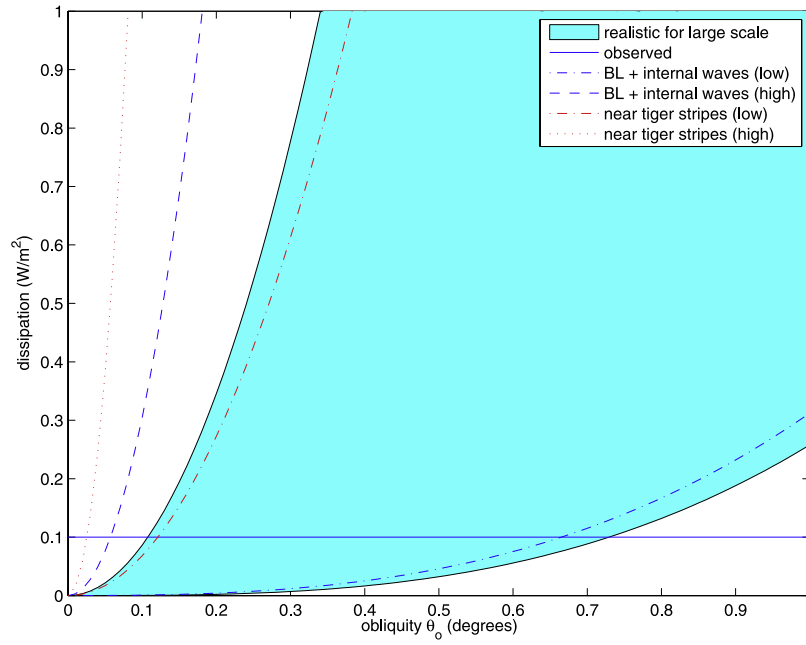


Figure 1. Ocean tidal heating (dissipation) as a function of Enceladus's axial tilt (obliquity angle). Large-scale average rates must fall in the shaded region to satisfy minimum boundary-layer dissipation as well as assure that the dissipation does not overdamp the tidal oscillations. The curves can reach beyond this region when they represent only local conditions. As in the legend, these curves represent the observed heat flux at the south pole of Enceladus, low/high estimates for total boundary-layer + internal-wave dissipation, and low/high estimates for internal-wave dissipation under the “tiger-stripe” features.

where the ocean flow velocities \mathbf{u} are obtained as $\mathbf{u} = \nabla \times (\psi \hat{\mathbf{r}})$. This solution describes rotation of the ocean about an equatorial axis that propagates westward around the moon once per day. In geophysical fluid dynamics, this solution is recognized as the lowest mode of the class of Rossby-Haurwitz waves. Remarkably, the obliquity is the only parameter in this flow estimate that is not accurately known; this can be contrasted with similar calculations for the eccentricity tides for which the solutions depend strongly on the ocean depth assumed. The reason for this difference is that in this solution, the flow is tangentially non-divergent. As another consequence, all energy of the ocean response is carried in the flow kinetic energy and there is no potential energy associated with vertical displacements of the ocean fluid. The remainder of this paper shall involve only two products from the flow solution: the time/longitude average of the squared flow velocity amplitude

$$u^2 = u^2(\theta) = u_{pole}^2 \frac{1}{2} (1 + \cos^2 \theta), \quad (4)$$

where

$$u_{pole} = \frac{3}{2} \Omega a \theta_o. \quad (5)$$

is the flow amplitude at the pole; and a similar average for u^3 which we obtain through numerical integration. Taking $\theta_o = 0.1^\circ$, for example, the flow amplitude u_{pole} is 3.5 cm/s which is remarkably strong. By comparison, the eccentricity tides are of the order of 1 mm/s for depth $h = 50$ –100 km,

but they increase as $1/h$ (see auxiliary material) such that at a depth of 4.3 km the eccentricity tides are as strong as the obliquity tides assuming $\theta_o = 0.1^\circ$. This is the second, zero-obliquity, shallow-ocean scenario and can be included in the dissipation results of Figure 1 by simply replacing the axis label θ_o with $2.3 \times 10^{-3} (m)/h (m)$. One should note though that the eccentricity tides are strongest at the equator, rather than at the poles.

[11] We shall now estimate the dissipation associated with this tidal flow. We shall focus on the polar region where (for the first scenario) the flow is strongest, and then add discussion of the latitude dependence. First, from the boundary-layer approach, and assuming for simplicity that the ice shell remains fixed with respect to the Enceladus frame, and therefore provides a second boundary layer in addition to that at the seafloor, we substitute (5) for u in (1), assuming the values $\rho = 10^3 \text{ kg/m}^3$, and $C_D = 0.003$ [Munk, 1997], and produce an estimate of dissipation, which we call D_{pole} , as a function of obliquity angle θ_o . (The value for C_D we use here is close to the value commonly used in terrestrial as well as icy-satellite [Sagan and Dermott, 1982; Sears, 1995] applications, but the value is empirical and shows a range of values that depend on local conditions [Wadhams, 2000].) This curve, described by the lower boundary of the shaded region in Figure 1, we regard as a minimum estimate because it does not account for transfer of energy to internal waves as described above. The upper bound of the shaded region in Figure 1 describes a requirement for consistency with the model we have used to produce the flow. Above this boundary, we expect that dissipation begins to affect the flow and we would need to

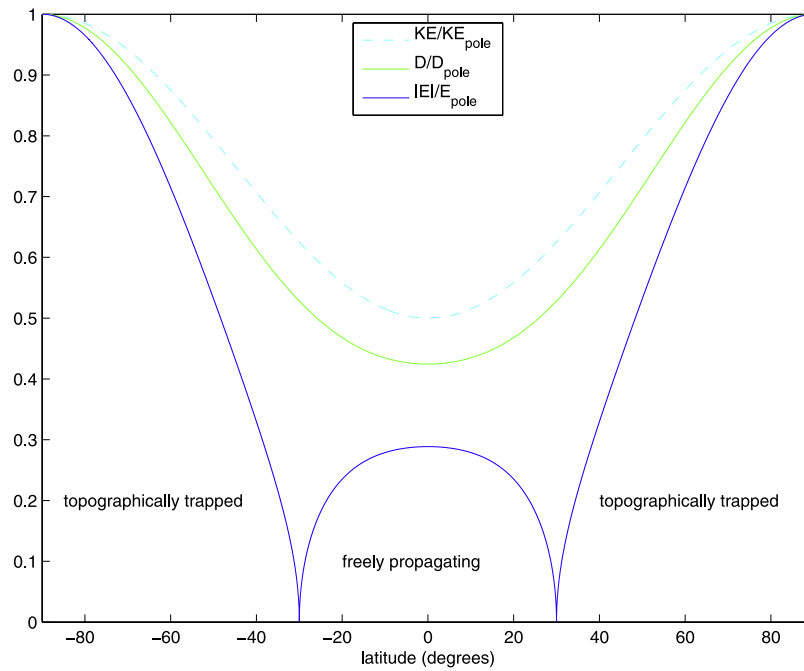


Figure 2. Latitude dependence of ocean flow kinetic energy (KE), boundary-layer dissipation (D), and internal-wave dissipation (E), each normalized with respect to their value at the poles. Beyond the diurnal-frequency “turning latitudes” ($\pm 30^\circ$) internal waves become trapped to topography. All curves show highest values near the poles.

include damping terms in the Laplace tidal equations to account for this. But the resonance is also lost with such strong damping, so we expect this curve defines a maximum dissipation rate. The criterion we used to produce this upper boundary was that the tidal oscillations decay in no less than 3 tidal cycles if forcing were to stop. Hence, the shaded region satisfies the requirement $Q \geq 6\pi$, where Q is the so-called “quality factor”. (This can be compared with a Q of about 15 for Earth’s dominant diurnal tide, where the presence of continents leads to major damping and a low Q .) So long as a global ocean on Enceladus exists, we expect this shaded region to contain a realistic description of dissipation. Note, however, that the upper bound pertains only to large-scale average dissipation; higher localized values are allowed. Parts of this diagram would indicate extremely large heating rates, and an important consideration in deciding the realism would be to consider the impact of this energy extraction on the moon’s orbital evolution. Such considerations (beyond the scope of the work here) would probably not preclude time-dependent or episodically extreme tidal heating, but they would indicate limitations on the timescales over which such strong tidal heating could be sustained.

[12] We see from the lower boundary of the shaded region in Figure 1 that for obliquity angles greater than about 0.7 degrees, even this minimum estimate shows ocean tidal heating in excess of the 100 mW/m² observed. In light of the modern understanding of how dissipation occurs in Earth’s deep ocean, as described above, we should expect that this is an underestimate. Surely an ocean on Enceladus will also include stratification and topography and therefore enhanced dissipation through transfer of tidal-flow energy to internal waves. We estimate this dissipation, which we shall call E , using the parameterization described by Jayne

and Laurent [2001] for diurnal frequency, and using our solution (4), to give:

$$E(\theta) = E_{pole} \left(\frac{1 + \cos^2 \theta}{2} \right) \left(\frac{1 - 4 \cos^2 \theta}{3} \right)^{1/2} \quad (6)$$

where the dissipation amplitude at the poles

$$E_{pole} = \frac{\sqrt{3}}{2} \rho \kappa \tilde{h}^2 N u_{pole}^2 \quad (7)$$

depends on the wavenumber and height anomaly of the topography (κ , \tilde{h}), and the stratification-dependent buoyancy frequency N . Specifying the parameters in (6) is obviously guess-work, but let us consider a wide range $N = 8.2 \times 10^{-5} - 1.8 \times 10^{-3} \text{ (s}^{-1}\text{)}$ that includes conditions for Earth as well as assumptions used previously for Europa, and $\kappa \tilde{h}^2 = 6.3 - 160 \text{ (m)}$. We calculate low and high estimates for E_{pole} and add these to D_{pole} to obtain estimates for the total dissipation, labeled “BL + internal waves” in Figure 1. These estimates are chosen to encompass possible conditions of seafloor topography. We can also consider flow interaction with the topography of the bottom of the ice sheet, and attempt a more specific example focusing on the local region under the tiger stripes. We make the rather rough assumption that this topography has the same height ($\approx 500 \text{ m}$) and length scales (2 km) as that observed on the surface of these regions [Porco, 2006]. In this case it is only the unknown stratification that accounts for the difference in the low and high estimates for tiger-stripe dissipation plotted in Figure 1.

[13] Finally, let us relate our results to the observation that high heat fluxes were observed only near the south

pole, below about -55° latitude. In Figure 2 we plot the latitude dependence of our results. While the flow kinetic energy (KE) and the boundary-layer dissipation (D) vary with latitude only through dependence on the latitude-dependent quantities u^2 and u^3 , respectively, the internal-wave dissipation E has an additional latitude-dependent factor given by the second term in brackets in (6) which describes the latitude dependence of the internal wave propagation. Although our results show there is indeed larger dissipation near the poles, this may not be as important as it first seems because the distribution of the topography will likely be more important in controlling where dissipation occurs. It is useful to compare with the second scenario for which very little of the dissipation should appear in the polar regions because the eccentricity tides are very weak there (auxiliary material). Neither of our ocean tidal dissipation scenarios would alone account for the lack of a corresponding heat flux at the north pole, but note that this will be common to any tidal dissipation scheme that assumes symmetric composition parameters because the astronomical tidal potential lacks the necessary degree 1 spherical harmonic to produce such an asymmetry. Moreover, with a liquid ocean capable of forming convective circulation to transport heat, there is no longer the expectation that observed heat fluxes reflect local heat sources.

[14] In summary, global ocean tides appear capable of providing large ocean heat, and this would explain many of the curious features of Enceladus. This requires either that there be at least some small fraction of a degree of obliquity angle, or a shallow ocean. Of the moons Europa, Ganymede, Callisto, Enceladus, and Titan, Enceladus allows for the deepest ocean where the transition from obliquity to eccentricity tidal flow dominance occurs. It also has the smallest size. Hence, if one seeks a candidate where strong obliquity tides have diminished the obliquity angle to zero and the ocean is now heated by eccentricity tides in a shallow ocean, then Enceladus may be it.

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