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Title: Methane hydrate creates the thick oceans in Mimas and Enceladus

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Abstract

The difference between the inactive surface of Mimas and the active surface of Enceladus is puzzling. We investigate the conditions under which the both have a thick subsurface ocean and the thermal lithosphere of Mimas is thicker than that of Enceladus by using a one-dimensional simulation of thermal evolution. We adopt the initial core temperature, initial methane concentration, and tidal heating rate as free parameters in the calculation. The initial methane concentration and tidal heating rate greatly affect the current ocean thickness, although the initial core temperature does not affect the thickness. Methane hydrate forms in a subsurface ocean if the initial methane concentration is not 0. The methane hydrate layer plays an insulative role in an icy shell. When the initial methane concentration is 1000 $\mathrm{mol\,m^{-3}}$, ~3 GW is needed to achieve more than 50 km of the subsurface ocean on Mimas and ~10 GW is needed to achieve more than 25 km of the subsurface ocean on Enceladus. These values are smaller than those needed for when the initial methane concentration is 0 $\mathrm{mol\,m^{-3}}$. The existence of the methane hydrate layer promotes the survival of the subsurface ocean because it insulates internal heat. In addition, it is found that the surface heat flux is depressed if

the methane hydrate layer exists, which is consistent with the unrelaxed craters in Mimas. Methane hydrate may explain the thick oceans in Mimas and Enceladus and the inactive shell of Mimas.

Keywords

Mimas; Enceladus; Subsurface ocean; Methane hydrate

Main Text

Introduction

Cassini has revealed that Saturnian moons have various appearances, although their surfaces are mainly covered with the same substance – H_2O ice. This fact raises many questions. One of the issues is the contrast between Enceladus and Mimas.

Enceladus has a high albedo surface. In the south polar region of Enceladus, an active plume that includes ice particles and gaseous species is emitted (Porco et al. (2006)). A composite infrared spectrometer (CIRS) aboard Cassini found high heat flow around the region, and its power is estimated to be ~15 GW (Howett et al. (2011)). Gravity, physical libration, and topography indicate that Enceladus has a global ocean beneath its icy shell (Iess et al. 2014; Thomas et al. 2016; Čadek et al. 2016). According to these

data, the interior consists of a low-density core, thick ocean, and thin icy shell. The plume is found to include nanosized silica particles, which indicates ongoing hydrothermal activity at the seafloor (Hsu et al. (2015)). These observations show that Enceladus is a hot and active body despite its small size.

In contrast to Enceladus, Mimas is dark. Many craters cover its surface. Although absolute surface ages for Mimas and Enceladus are vague due to uncertainty of a cratering rate, their crater densities indicate that Mimas surface is much older than most of Enceladus surface (Kirchoff and Schenk (2009)). Additionally, there are few fractures derived from geophysical activities. The largest crater, Herschell, is not overly relaxed. Hence, Mimas is thought to be inactive. However, strong libration is observed, which indicates a thick subsurface ocean or lumpy rocky core (Tajeddine et al. (2014)). The phase lag of the libration supports the existence of a subsurface ocean. (Noyelles (2017)). The suggestion of a subsurface ocean is confusing since craters on Mimas should be more relaxed if the inside of Mimas is hot enough to have a global ocean (Neveu and Rhoden (2017)). Hence, it is still unclear whether Mimas possesses a subsurface ocean.

The physical and orbital properties of these two moons are summarized in Table 1. Enceladus and Mimas are similar in size and components, and they possibly have the oceans, but the appearances are quite different. The difference is thought to reflect the difference in their interior structures and thermal histories. To constrain the conditions to achieve the current internal structures and surfaces of Enceladus and Mimas, it is essential to know the history of these moons. Therefore, we investigate the long-term thermal evolution of each moon under various conditions.

The major heat sources needed to sustain global oceans are long-lived radioactive decay in rocky cores and tidal heating in the solid parts of the moons. Although other heating sources may contribute, their heating rates are much lower than the aforementioned major sources (Malamud and Prialnik 2013; Hand et al. 2011; Nimmo et al. 2007). The total heat derived from radioactive decay depends on the rocky core size, and those of Mimas and Enceladus are too small to melt ice shells and sustain their subsurface oceans. Therefore, tidal heating is more important to maintain these oceans. When a moon is in a synchronous state with its primary, the tidal heating rate can be calculated. The rate is given by

$$\dot{E} = -\frac{21}{2}\mathrm{Im}(k_2)\frac{(nR_s)^5}{G}e^2 \#(1)$$

where $\mathrm{Im}(k_2)$ is the imaginary part of the Love number, G is the gravitational constant, n is the mean motion, R_s is the radius of the moon, and e is the orbital eccentricity (Segatz et al. (1988)). The heating rate is proportional to the square of eccentricity. As the orbital eccentricity of Mimas is higher than that of Enceladus, Mimas may possess a subsurface ocean because of tidal heat. Generally, evaluating the Love number is difficult because the value depends on how a moon deforms. On the other hand, if two moons are in a mean motion resonance (MMR), the equilibrium tidal heating rate can be calculated by considering the conservation of mass and angular momentum (Meyer and Wisdom (2007)). The heating rate is described by

$$\dot{E} = \frac{n_0 T_0}{\sqrt{1-e_0^2}} + \frac{n_1 T_1}{\sqrt{1-e_1^2}} - \frac{T_0 + T_1}{L_0 + L_1}\left(\frac{GMm_0}{a_0} + \frac{GMm_1}{a_1}\right) \#(2)$$

$$T = \frac{3}{2}\frac{Gm^2 R_p^5 k_{2p}}{a^6 Q_p} \#(3)$$

where T is the torque caused by the primary, L is the angular momentum, M is the mass of the primary, m is the mass of the moon, and a is the semimajor axis. Subscripts 0 and 1 represent the inner moon and outer moon, respectively. Note that this equation does not rely on the interior structure of the moon but on that of the primary.

The current Q_{p} is suggested to be ~2000 from an astrometric study (Lainey et al. (2012)). At present Enceladus and Dione are in a 2:1 eccentricity-type MMR. Assuming Q_{p} is now 2000, the equilibrium tidal heating rate for Enceladus is ~10 GW. This tidal heating rate is insufficient for the survival of a thick ocean (Kamata (2018)). Mimas is not currently in any eccentricity resonance, and it is difficult to determine its equilibrium tidal heating rate.

Dissolved ingredients in an ocean, such as ammonia, make it easier for the ocean to survive because they depress the freezing point (Choukroun and Grasset (2010)). However, high concentrations of ammonia (~15 wt%) are needed to achieve thick oceans (Kamata (2018)), and this is unlikely in Enceladus and Mimas since such amounts of ammonia were not found in the Enceladus plume (Waite et al. (2017)). Alternatively, we consider the existence of a methane hydrate layer beneath a pure ice layer in an icy shell. Methane hydrate is a crystalline compound where methane gas is included in a cage made of a hydrogen bond network of water (Sloan Jr and Koh (2007)). Its appearance and some physical properties are similar to those of ice. However, the thermal conductivity of methane hydrate is approximately four times

lower than that of ice (Waite et al. (2009)), and its viscosity is one order of magnitude higher than that of ice (Durham et al. (2003)). These two characteristics may contribute to the thick subsurface ocean, as heat conduction and convection are depressed by the low thermal conductivity and the high viscosity.

Additionally, this insulating layer lowers the temperature of the icy shell, so the icy shell becomes more rigid than the shell which is composed of only ice. This rigid shell has an advantage in the maintenance of the unrelaxed craters of Mimas. The assumption of a methane hydrate insulating layer succeeded in achieving both the maintenance of the topography of Sputnik Planitia and the existence of an inner ocean on Pluto (Kamata et al. (2019)). We assume that methane hydrate may similarly exist in Enceladus and Mimas. The plume composition of Enceladus observed by Cassini includes not only water but also some hydrophobic gases such as methane. With these plume compositions, gas hydrates can form under the pressure and temperature of the subsurface ocean of Enceladus (Bouquet et al. (2015)). If there are the methane hydrate layers exist in Mimas and Enceladus, subsurface oceans of Mimas and Enceladus could survive with small amount of tidal heating rate.

In this paper, we investigate the conditions under which both Enceladus and Mimas have a thick inner ocean and where Mimas is less active than Enceladus, which means that Mimas has thicker thermal lithosphere than Enceladus, by using a one-dimensional simulation of thermal evolution under various conditions. In section 2, we describe the interior structure models we used, the details of the thermal evolution model, and the parameter settings. In section 3, we show the results. Changes in the thickness of the shell and ocean with time and the dependence of the final ocean and lithosphere thicknesses on the parameters are shown. In section 4, we discuss the relevance between our results and other observations and simulations. In section 5, we offer conclusions.

Methods

Interior structure model

We consider Enceladus and Mimas to be 4-layered spheres of rock - ocean - methane hydrate - ice. This layered structure is assumed to be achieved initially by early differentiation caused by the energies of decay of short-lived radioactive elements and accretion energy (Schubert et al. (2007)). The interior structure models are shown in Figure 1, and the model parameters are summarized in Table 2. The initial thickness of

the ice layer is set to 10 km, and those of the other layers are calculated from the bulk density and the densities of each layer. These values are compatible with the values estimated from observation data of Enceladus and Mimas (e.g., Tajeddine et al. (2014)). In the ocean, we assume that only methane has dissolved. We do not consider the presence of ammonia because the amount of ammonia observed by Cassini (~1 %) would not notably affect the ice freezing point (Choukroun and Grasset 2010; Vu et al. 2014). It is assumed that the composition formula of methane hydrate is $CH_4 \cdot 6H_2O$, and its density is 920 $\mathrm{kg\,m^{-3}}$.

Thermal evolution

Heat transfer

To calculate the evolution of heat transport with time, including conduction and convection, the modified mixing length theory, the one-dimensional calculation scheme developed by (Kamata (2018)), is used. The governing equation of heat transfer is given by

$$\rho C_p \frac{\mathrm{d}T}{\mathrm{d}t} = -\frac{1}{r^2}\frac{\mathrm{d}}{\mathrm{d}r}\left(r^2 F_{\mathrm{cond}} + r^2 F_{\mathrm{conv}}\right) + Q$$

where ρ is the density, C_p is the specific heat, T is the temperature, r is the radius,

F_{cond} is the conductive heat flux, F_{conv} is the convective heat flux, and Q is the volumetric heating rate. The thermal conductivity and specific heat of each layer depend on temperature (Waite et al. 2007; Hobbs 2010; Choukroun and Grasset 2010). F_{cond} and F_{conv} are given by

$$F_{\mathrm{cond}} = -k\frac{\mathrm{d}T}{\mathrm{d}r}$$
$$F_{\mathrm{conv}} = -k_v\left\{\frac{\mathrm{d}T}{\mathrm{d}r} - \left(\frac{\mathrm{d}T}{\mathrm{d}r}\right)_s\right\}$$

where k is the thermal conductivity, k_v is the effective thermal conductivity, and $(\mathrm{d}T/\mathrm{d}r)_s$ is the adiabatic temperature gradient, that is $-\alpha g T/C_p$. α is the thermal expansivity. k_v is given by

$$k_v = \begin{cases} -\dfrac{\alpha C_p \rho^2 g l^4}{18\eta}\left\{\dfrac{\mathrm{d}T}{\mathrm{d}r} - \left(\dfrac{\mathrm{d}T}{\mathrm{d}r}\right)_s\right\}, & \left(\dfrac{\mathrm{d}T}{\mathrm{d}r} \le \left(\dfrac{\mathrm{d}T}{\mathrm{d}r}\right)_s\right) \\ 0, & \left(\dfrac{\mathrm{d}T}{\mathrm{d}r} > \left(\dfrac{\mathrm{d}T}{\mathrm{d}r}\right)_s\right) \end{cases}$$

where g is the gravitational acceleration, η is the viscosity, and l is the mixing length. l depends on the computational grid temperature, shell thickness, and other parameters (Kamata (2018)). The viscosities of ice and methane hydrate are as follows:

$$\eta(T) = \eta_{\mathrm{ref}}\exp\left(\frac{E_a}{R_g T_{\mathrm{ref}}}\left(\frac{T_{\mathrm{ref}}}{T} - 1\right)\right)$$

where η_{ref} is the reference viscosity and corresponds to the viscosity at T_{ref}, T_{ref} is the reference temperature 273 K, E_a is the activation energy, and R_g is the gas

constant (e.g., Goldsby and Kohlstedt (2001)). We assume that the reference viscosity of ice is 10^{14} Pa s and that of methane hydrate is 2.0×10^{15} Pa s (Kamata et al. 2019; Durham et al. 2003). These reference viscosities have uncertainties since these values depend on the porosity and grain sizes of the icy shell, and they are unknown. Solid convection in the rocky core is not considered, i.e., F_{conv} is 0 in the rocky core. We assume that heat received from the rocky core transfers to the bottom of the icy shell immediately in the ocean, and the temperature remains constant at the freezing point of the shell bottom. The adiabatic gradient in the ocean is ignored since the gradient is small. Heat sources are the decay of long-lived radioactive elements included in the rocky core and tidal heat in solid parts. The rocky core is assumed to be carbonaceous chondrites. The amounts of long-lived radioactive elements are summarized in Table 3 (Lodders (2003)). The details of the tidal heating rate are described in a later section.

Shell growth

Icy shell growth is described by the energy balance at the shell/ocean boundary. If the temperature at the shell bottom exceeds the melting point, differences between

incoming heat flux and outgoing heat flux will melt the icy shell. The change in the shell thickness is given by

$$\rho_{\mathrm{ice}} L_{\mathrm{eff}} \frac{\mathrm{d}D}{\mathrm{d}t} = F_{\mathrm{out}} - F_{\mathrm{in}}$$

where D is the icy shell thickness, L_{eff} is the effective latent heat, F_{out} is the heat flux outgoing from the shell bottom, and F_{in} is the heat flux incoming into the shell bottom. L_{eff} is the sum of the latent heat of the shell and the heat used for warming the ocean to the melting point at the shell bottom. This value is given by

$$L_{\mathrm{eff}} = L - \frac{\rho_{\mathrm{oce}} g_{bot} C_{p,\mathrm{oce}} V_{\mathrm{oce}}}{4\pi R_{\mathrm{bot}}^2} \frac{\mathrm{d}T_{\mathrm{m}}}{\mathrm{d}P_{\mathrm{bot}}}$$

where L is the latent heat, V_{oce} is the volume of the ocean, and R_{bot} is the radius at the bottom of the layer. $C_{p,\mathrm{oce}}$ is the specific heat of pure water, which depends on temperature (Choukroun and Grasset (2010)). The melting point of ice $T_{\mathrm{m,ice}}$ depends on pressure, and is given by

$$T_{\mathrm{m,ice}}(P) = 273.1 - c_1 P - c_2 P^2$$

where P is the pressure, $c_1 = 7.95 \times 10^{-8}\ \mathrm{K\,Pa^{-1}}$, and $c_2 = 9.6 \times 10^{-17}\ \mathrm{K\,Pa^{-2}}$ (Leliwa-Kopystyński et al. (2002)). The melting point of methane hydrate is given by

$$T_{\mathrm{m,hyd}} = \frac{b}{\ln(\frac{P}{1000}) - a}$$

where $a = 38.980$ and $b = -8533.80\ \mathrm{K}$ (Sloan Jr and Koh (2007)). The growth of the methane hydrate layer is slightly different from that of the ice layer. We adopt a method for methane hydrate layer growth similar to that of (Kamata et al. (2019)) and summarize it below. The methane hydrate layer grows thicker if the outgoing heat flux is higher than the incoming heat flux. The temperature of the ocean remains constant at that time. If the incoming heat flux is higher than the outgoing heat flux, the ocean temperature increases until it reaches the melting point. After the ocean temperature exceeds the melting point of methane hydrate, the hydrate layer starts to dissociate. We change the method slightly to introduce methane concentrations into the ocean. In our calculation, methane hydrate can form when methane is supersaturated in the ocean and the melting point of methane hydrate is lower than that of ice. The methane solubility at the ocean/shell boundary, which is at methane gas - water - methane hydrate equilibrium, can be determined from the temperature or the pressure (Duan and Mao (2006)). We note a fitting curve that reproduces the methane solubility in table 14 in (Duan and Mao (2006)), which is given by

$$m_{\mathrm{CH_4,satu,bou}} = \exp\left(A\left(\frac{1}{T_\mathrm{m}}\right)^3 + B\left(\frac{1}{T_\mathrm{m}}\right)^2 + C\left(\frac{1}{T_\mathrm{m}}\right) + D\right)$$

where $m_{\mathrm{CH_4,satu,bou}}$ is the methane solubility at the ocean/shell boundary and its unit is $\mathrm{mol\,m^{-3}}$, $A = 9.1589 \times 10^9\ \mathrm{K^3\,mol\,m^{-3}}$, $B = -9.7261 \times 10^7\ \mathrm{K^2\,mol\,m^{-3}}$, $C = 3.3963 \times 10^5\ \mathrm{K\,mol\,m^{-3}}$, and $D = -3.8513 \times 10^2\ \mathrm{mol\,m^{-3}}$. The methane concentration in the ocean decreases with methane hydrate formation. When methane hydrate dissociates, methane is dissolved into the ocean and is mixed immediately. Once the methane concentration falls below the soluble concentration by methane hydrate formation, the ice layer starts to form. Methane in the ocean will condense by ice layer formation until saturated. When the methane is supersaturated in the ocean, methane hydrate can form again. Consequently, the methane hydrate layer and ice layer form alternately, and the methane concentration in the ocean remains near the soluble concentration. This layered region is considered as a mixed layer of methane hydrate and ice for simplicity in our calculation. The mixture properties are a linear combination of methane hydrate and ice, which is given by

$$A_{\mathrm{mix}} = \phi_{\mathrm{hyd}} A_{\mathrm{hyd}} + (1 - \phi_{\mathrm{hyd}}) A_{\mathrm{ice}}$$

where A represents a material property and ϕ_{hyd} is the volume fraction of methane hydrate to ice in a calculated cell.

Calculation

Our code can reproduce the results for Kamata et al. (2019), Kamata (2018), and Kimura and Kamata (2020). The icy shell and rocky core are divided into 100 thin layers. The time step is chosen to satisfy the Courant-Friedrichs-Lewy (CFL) condition. The calculation is finished when the sum of the time steps exceeds 4.5 Gyr.

Initial conditions and parameter settings

Initial structure and temperature

Our calculation starts after differentiation, which finishes at approximately 100 Myr after CAI formation (Schubert et al. (2007)). Differentiation was caused by heat derived from short-lived radioactive species such as ^{26}Al and accretion energy. During differentiation of Enceladus, its core temperature reached ~ 1273 K (Schubert et al. (2007)). Mimas is also differentiated, but its core temperature is not heated as much (Neveu and Rhoden (2017)). We believe the icy shell must have been melted initially, and the liquid layer survives for the existence of the ocean because tidal dissipation does not effectively occur once the ocean completely freezes (Roberts and Nimmo (2008)). The highest temperature that the rocky core experienced during differentiation

depends on the amount of ^{26}Al in the core when the moon accreted. However, determining when the moons formed and how much ^{26}Al is in the core is complicated. Here, we consider the initial isothermal core temperature to be a free parameter in a range between 273 K and 1273 K. The nominal values are 1273 K for Enceladus and 573 K for Mimas. The thickness of the initial icy shell is set to 10 km and the temperature has a linear gradient from the surface to the ocean. The initial shell does not include any methane hydrate. The surface temperature is set to a constant value described in Table 1. The initial temperature of the ocean is set to 273 K. The differences in the initial structures except for completely frozen state does not considerably affect the final structures (Kimura and Kamata (2020)). If a subsurface ocean is initially completely frozen, an icy shell never melts because tidal heat which is a major heat source on Mimas and Enceladus is much depressed (Roberts and Nimmo (2008)).

Tidal heating rate

In previous studies, tidal heating rates on Saturnian moons were estimated (e.g., Meyer and Wisdom (2007)). However, determining the history of tidal heating rates is difficult

because the heating rates strongly depend on the interior structure and orbital properties, and their evolution is quite complicated. Therefore, we adopt an assumption similar to Kamata (2018), where the constant tidal heating rate is added at the bottom of the icy shell. In our calculation, we consider the average tidal heating rate as a free parameter in a range between 0 and 30 GW for simplicity. The tidal heat is assumed to be generated in solid parts of the moons, namely, the rocky core and icy shell. We do not consider dissipation in the ocean because its value is small enough to ignore (Chen et al. (2014)). In a rocky core, how the amount of tidal dissipation depends on the properties of the rocky core. If the core is rigid, dissipation does not effectively occur due to its high viscosity (Roberts and Nimmo (2008)). Porous rocky cores such as gravel dissipate well, and high heat flow is generated (Choblet et al. (2017)). In an icy shell, the dissipation depends on its viscosity. Tidal heat is produced most effectively when the shell is near the melting point (Tobie et al. (2003)); i.e., tidal heat is generated most at the shell bottom. The average tidal heat in our calculation represents the sum of the tidal heat generated in the icy shell and the rocky core. Since we are focused on the growth of the icy shell and the subsurface ocean, whether dissipation occurs at the rocky core or the

bottom of the icy shell is not important; the total tidal heat added to the icy shell is important. Thus, F_{in} is described by

$$4\pi R_{\mathrm{bot}}^2 F_{\mathrm{in}} = 4\pi R_{\mathrm{c}}^2 F_{\mathrm{c}} + Q_t$$

where R_{c} is the radius of the rocky core, F_{c} is the heat flux at the surface of the rocky core, and Q_t is the average tidal heating rate. Since tidal heating becomes ineffective once the ocean freezes completely (Roberts and Nimmo (2008)), the tidal heating rate is set to 0 if there is no ocean.

Initial methane concentration

The rocky core needs to be heated to approximately 300 °C so that methane is formed by the serpentinization and reduction of carbon dioxide (McCollom (2016)). However, heat only from long-lived radioactive decay cannot sufficiently warm the core due to the small-sized cores of Enceladus and Mimas. Therefore, we assume that methane is produced during early differentiation by the heat derived from decay of short-lived radioactive elements and/or that methane existed initially in the moons. The amount of methane produced by the serpentinization of the rocky core depends on the initial amount of olivine in the core and the degree of hydration of the core. Comets

considered to be parts of icy moons contain $\sim$1 % CH_4 relative to water (Mumma and Charnley (2011)). This value is approximately equivalent to $\sim 600\ \mathrm{mol\,m^{-3}}$ methane. As the amount of methane in an initial subsurface ocean depends on a complicated condition at the accretion of the moons and it is unknown, this amount is designated a free parameter in a range of 0-1000 $\mathrm{mol\,m^{-3}}$. In most of the concentration range, methane is supersaturated in the subsurface ocean because methane must be supersaturated in the subsurface ocean so that methane hydrate forms. The previous study which considered methane hydrate layer in Pluto similarly assumed that methane is supersaturated in the ocean (Kamata et al. (2019)). Supplementary Figure 7 in Kamata et al. (2019) shows that $\sim$1 % CH_4 allows >10 km methane hydrate layer to form in the subsurface ocean of Pluto. For another source of methane, ongoing methane production may be considered. If tidal dissipation occurs in the rocky core, methane could possibly be produced by tidal heat. Additionally, it is known that methanogen can survive under the conditions present in the ocean of Enceladus (Taubner et al. (2018)), and methane may be produced by them now. To simplify our calculation, we do not consider such ongoing methane production.

Results and Discussion

Results

Change in shell thickness with time

Figure 2 shows one of the results of the icy shell thickness change with time for Mimas. If a methane hydrate layer exists beneath the pure ice layer, the final ocean is thicker than that in the case without methane hydrate layer under the same tidal heating rate and initial core temperature. This is because the viscosity of the methane hydrate layer is higher than that of ice, and the convection is depressed as shown in the formula of k_v. The methane hydrate layer prevents the icy shell from convecting and insulates the heat from the bottom of the shell or deeper region. As a consequence of the insulation of the methane hydrate layer, the ice layer over the methane hydrate layer becomes cooler, which leads to a higher viscosity of ice. Hence, the icy shell convection in the case with methane hydrate is depressed and F_{conv} is nearly 0 in all areas of the shell.

The current shell thickness is achieved in the early stage of evolution. This is because the energy of radioactive decay is much lower than tidal heating due to the small rocky core, and the incoming tidal heat into the icy shell immediately becomes closely

balanced with the outgoing heat. Therefore, the thermal evolution is mainly controlled by tidal heating.

The core temperature evolution was also investigated, and one of the results for Mimas is shown in Figure 3. The temperature rapidly decreases to near the ocean temperature. Tidal heating does not affect the core temperature increasing in our calculation. Therefore, this result implies that the decay of long-lived radioactive elements does not warm the core effectively. The effect of the differences in the initial core temperature on the final shell thickness of Mimas is shown in the next subsection. Similar tendencies are also found in the results for Enceladus.

The dependency of final shell thickness on initial core temperature and initial methane concentration

Figure 4 shows the different growths of the icy shell with time among different initial core temperatures for Mimas. The shell thicknesses for each evolution attain a similar value in approximately 100 Myr, and the final shell thicknesses of all initial core temperatures achieve the same value. It is found that the final shell thickness does not depend on the initial core temperature, in other words, the heat stored initially in the

rocky core is much lower than the heat added into the icy shell, namely the tidal heat and the heat generated by radioactive decay. In our calculation, the typical initial temperatures are 573 K for Mimas and 1273 K for Enceladus. These temperatures are close to the core temperatures indicated by the previous studies in which the thermal evolution considering the heat from the decay of short-lived radioactive elements was calculated (Schubert et al. 2007; Neveu and Rhoden 2017).

Figure 5 shows the differences in final structures under different initial methane concentrations in the ocean and 4 GW for the tidal heating rate for Mimas. As the initial methane concentration decreases, the final ocean thickness decreases. If the initial methane concentration is less than 1000 $\mathrm{mol\,m^{-3}}$, the mixed layer of ice and methane hydrate is laid beneath the pure methane hydrate layer. The properties of the mixed layer are similar to those of ice, and therefore, the mixed layer cannot insulate as much incoming heat as the methane hydrate layer. Under the condition that the initial concentration is less than 50 $\mathrm{mol\,m^{-3}}$, the final ocean thickness is almost equal to that of the case without methane hydrate. This is because the initial subsurface ocean is not supersaturated by methane, and the pure methane hydrate layer cannot form. Under the

condition that the initial concentration is more than 1000 $\mathrm{mol\,m^{-3}}$, the final ocean thicknesses under any methane concentration become the same value since the ocean is supersaturated during growth, and the mixed layer does not appear. In our calculation, the typical values are 1000 $\mathrm{mol\,m^{-3}}$ for the case with the methane hydrate layer and 0 $\mathrm{mol\,m^{-3}}$ for the case without the methane hydrate layer.

Differences in final shell thickness and lithosphere among different tidal heating rates

Figure 6 shows the final shell and ocean thickness and the thermal lithosphere of Enceladus and Mimas under different tidal heating rates. The thermal lithosphere denotes the nonconvective area in the shell where F_{conv} is 0. Comparing Mimas with Enceladus, the subsurface ocean is more likely to survive in Mimas than in Enceladus under the same tidal heating rates. This is because the heating rates per unit mass for Mimas are higher than those for Enceladus due to its smaller radius: the final shell of Mimas is thinner than that of Enceladus if the tidal heating rates of Mimas and Enceladus are equivalent. In the regions between 4 GW and 7.5 GW for Figure 6(a) and between 15 GW and 25 GW for Figure 6(c), the final shell thickness is a plateau, where

the methane hydrate layer insulates the shell from the inner heat and the convective area in the shell is thin; in other words, the thermal lithosphere is thick. If the methane hydrate layer exists in an icy shell, a thermal lithosphere is thicker than that of the icy shell without the methane hydrate layer. This is attributed to methane hydrate insulation. The methane hydrate layer cannot form in the regions more than the tidal heating rates because the melting point of methane hydrate is higher than that of ice due to low pressure. The final thicknesses for the case with methane hydrate are equivalent to those for the case without methane hydrate.

Surface heat flux of Mimas

Figure 7 shows the relevance between the current surface heat flux for Mimas and the final shell thicknesses. In the region where the final shell thickness is more than 40 km in the figure, the surface heat flux of the case with methane hydrate is depressed more than that of the case without methane hydrate due to methane hydrate insulation. In the other region, the current surface heat flux for the case with methane hydrate is equal to that for the case without methane hydrate since there is no methane hydrate layer due to the high tidal heating rate for the case with methane hydrate. The low surface heat flux

on Mimas may contribute to the maintenance of surface topography on Mimas. Details are discussed in the discussion section.

Summary of results

In Figure 8, the results in Figure 6 are summarized, and the ranges in which a thick subsurface ocean can exist in both moons and where the current thermal lithosphere of Mimas is thicker than that of Enceladus are shown. For the thick ocean criteria, thicknesses greater than 50 km for Mimas and greater than 25 km for Enceladus are assumed since the ocean thicknesses of both moons are not completely known. The area that is colored red and surrounded by the dashed line in the figure is *the prefer region* where the thick subsurface ocean exists and the thermal lithosphere of Mimas is thicker than that of Enceladus. To achieve a thick ocean in these bodies, Mimas needs ~3 GW and Enceladus needs ~10 GW of tidal heating rate in the case of methane hydrate, while Mimas needs ~5.5 GW and Enceladus needs ~14 GW in the case without methane hydrate. *The prefer region* for the case with methane hydrate is larger than that for the case without methane hydrate, which means that the current states of Mimas and Enceladus are achieved more easily if a methane hydrate layer exists beneath the ice

layer.

Discussion

The dependence of tidal dissipation factor Q_{p}

The thermal evolution of the moons is closely associated with their orbital evolution (Nimmo et al. (2018)). This is because the tidal heating rate, which controls the thermal evolution of Mimas and Enceladus, depends not only on their internal structures but also on their orbits. For moons with MMR, the equilibrium tidal heating rate can be estimated as shown in equations (2) and (3). The tidal dissipation factor of the primary Q_{p} in the equations influences the tidal heating rate in the moons and the orbital evolution of the moons. An astrometric study estimated that the Q_{p} of Saturn is ~ 2000 (Lainey et al. (2012)). If $Q_{\mathrm{p}} \sim 2000$ was constant, Mimas and Enceladus were possibly formed more recently (e.g., Nimmo et al. (2018)). As shown in Figure 2, it is found that the current interior of Mimas was achieved in a few Myr from the initial state, and the interior structure can be achieved even though Mimas formed a few Myr ago. The same finding applies to Enceladus, whose size is similar to that of Mimas. Note, however, that we do not consider differentiation, failure, and the relaxation of

craters in the calculations.

Fuller et al. (2016) proposed that Q_{p} does not have to be constant and that the value used to be higher; this was called the resonance locking hypothesis. Under this hypothesis, Mimas and Enceladus could have formed ~4.6 Gyr ago and Mimas went through 3:2 Mimas-Enceladus MMR 0.34 Gyr ago when Q_{p} was ~4800, and the equilibrium tidal heating rate for Mimas, following equations (2) and (3), was ~3.6 GW (Tian and Nimmo (2019)). The tidal heating rate can thicken the subsurface ocean in Mimas if the methane hydrate layer exists. Now, Enceladus is in the Enceladus-Dione MMR and the equilibrium tidal heating rate is ~10 GW. The tidal heating rate can thicken the subsurface ocean in Enceladus if the methane hydrate layer exists. Observational data are used to estimate the current shell thicknesses of both bodies. The estimated shell thickness of Enceladus ranges between 25 km and 60 km, which means that the subsurface ocean is at least 10 km (e.g., Iess et al. (2014)). Figure 8 shows that the 10 GW tidal heating for Enceladus rate can achieve the subsurface ocean thickness in both cases – with methane hydrate and without methane hydrate. Libration data are used to estimate the shell thickness of Mimas: the thickness ranges between 24 km and

31 km (Tajeddine et al. (2014)), which corresponds to the ocean thickness between 68.3 km and 75.3 km assuming that the core radius is 98.9 km. The 3.6 GW for the tidal heating rate of Mimas is insufficient to maintain the ocean thickness between 68.3 km and 75.3 km. This may be explained by an effect not considered in the previous study. The estimation of shell thickness by libration depends on the viscosity of the shell. If the methane hydrate layer exists beneath the ice layer, the viscosity of the shell increases due to the insulation of methane hydrate, which may lead to the estimated shell becoming thicker. In addition, dissolved gas, which can be encapsulated in gas hydrate cages, may affect the final ocean thickness. The stable pressure for mixed hydrates of methane and other gaseous species is different from that for pure methane hydrate, which changes the ice/gas hydrate boundary depth and the final ocean thickness. Details of the dissolved gaseous species in the ocean are discussed in a later subsection.

Relevance to geophysical features

Mimas has few tectonic features on its surface (Rhoden et al. (2017)). This is a large constraint for the thermal evolution of Mimas. Rhoden et al. (2017) estimated the tidal

stress on the ice shell of Mimas when the subsurface ocean existed. The tidal stress is over 800 kPa when the viscosity of the shell of the lower parts is 10^{12} Pa s, which leads to the shell failure. In order not to crack, the shell must be cool and exhibit high viscosity. The existence of a methane hydrate layer may offer an advantage in that the icy shell remains cool due to the insulation of methane hydrate. Thermal stress derived from phase change may also cause shell failure. To avoid thermal stress, changes in ocean thickness must be much slower. As shown in Figure 2, the changes in ocean thickness with time under a constant tidal heating rate is little except for the initial 100 Myr. Hence, the effects of change in the tidal heating rate on the shell thickness is more significant. As shown in Figure 6(a) and 6(c), the final shell thickness changes little in the regions where methane hydrate exists (i.e., 4-7.5 GW for Mimas and 15-25 GW for Enceladus), which means that the change in shell thickness is insensitive to the change in tidal heating rate under the condition that a methane hydrate layer exists. Methane hydrate layer suppresses the changes in the shell thickness. Note, however, that the tidal heating rate is closely associated with orbital evolution; the details must be investigated by another study.

In Tian and Nimmo (2019), the surface heat flux on Mimas is estimated to be less than 10 mW/m^2 from the condition of the largest crater Herschell, which is not notably relaxed. As shown in Figure 7, the heat flux is depressed sufficiently if the methane hydrate layer exists. As mentioned above, the visual appearance of Mimas may be explained by assuming a methane hydrate layer beneath the icy shell.

Craters on Enceladus are highly relaxed, and the surface is thought to experience high heat flux. The high heat flow is supposed to be derived from the regolith layer near the surface, which originates from the plume and E-ring (Bland et al. (2012)). The regolith layer has high porosity, and relatively low thermal conductivity, which plays a role in insulation like the methane hydrate layer. The regolith layer creates a high-temperature gradient near the surface. The high-temperature gradient results in viscous relaxation of the craters on Enceladus surface. Therefore, the presence of relaxed craters may be independent of whether the methane hydrate layer exists, and it is difficult to discuss any relevance between the geophysical features on the Enceladus surface and methane hydrate existence. In contrast to the methane hydrate layer, however, too thick regolith layer causes convection, and leads to thin the final ocean thickness (Kamata et al.

(2019)). Therefore, the methane hydrate layer is superior to the regolith layer to achieve a thick ocean in Enceladus. Although it is difficult to consider a local geophysical feature because our calculation has only one dimension, it may be unlikely that methane hydrate or any other hydrate would form at the south polar terrain since the shell thickness is thinner there than at any other region of Enceladus (Čadek et al. (2016)), and the melting points of gas hydrates are lower than that of ice. Hence, the viscosity of a shell at the south polar terrain that does not include methane hydrate may be lower than that at the other region.

Dissolved gaseous species

Hydrate cages can incorporate not only methane but also other gases dissolved in the ocean. Components observed in the Enceladus plume would form mixed hydrates under pressure and temperature in the ocean (Bouquet et al. (2015)). The differences between methane hydrates and mixed hydrates do not greatly affect our results except for melting point and density. Because of the differences in melting points, the final ocean thickness may change because the depth at which gas hydrates instead of ice appear may change. The difference in density affects whether the gas hydrate layer can exist

beneath the ice layer. The density of mixed hydrates depends on the structure of the hydrates and the involved gaseous species. If gas hydrates form from the Enceladus plume composition, two types of hydrates can exist: structure I mixed hydrates, with a density of 1040 $\mathrm{kg\,m^{-3}}$, and structure II mixed hydrates, with a density of 970 $\mathrm{kg\,m^{-3}}$ (Bouquet et al. (2015)). Structure II mixed hydrates can ascend in an ocean, but whether structure I mixed hydrates ascend in the ocean depends on the density of the ocean. The ascended hydrates can exist beneath the icy shell, while the sunken hydrates may be dissolved by heat from the rocky core. CO_2 is important in the dissolved gaseous species since CO_2 is heavier than CH_4 and CO_2 coexists with CH_4. The ratio of CO_2 to CH_4 in the ocean depends on the temperature near the core. CO_2 dominates in the ocean at over 450 °C (McCollom and Seewald (2007)). Temperature increase in the rocky core caused by tidal dissipation is not considered in our calculation. As shown in Figure 3, the core temperature is insufficient to produce much CO_2, and CH_4 dominates in the ocean. However, the temperature of the core is influenced by how much tidal dissipation occurs in the core, which is beyond our focus. We assume that the initial methane concentration would be 1000 $\mathrm{mol\,m^{-3}}$ for most of the cases with methane

hydrate in our calculation. This value is much higher than the methane solubility of the water and the amount produced by the serpentinization of the core. Hence, our hypothesis requires methane ice and/or methane hydrate accretion when moons form.

Conclusion

We investigate the conditions under which a thick subsurface ocean could be achieved in Mimas and Enceladus, and Mimas is found to have a thicker thermal lithosphere than that of Enceladus in a one-dimensional thermal evolution simulation. In our calculation, the current shell thicknesses of Mimas and Enceladus are achieved within almost 100 Myr since both moons are small. If a methane hydrate layer exists beneath the ice layer, the final ocean thickness is thicker than that in the case without methane hydrate because the methane hydrate layer plays a role in shell insulation. If there is a methane hydrate layer, ~3 GW is needed to sustain more than 50 km of subsurface ocean on Mimas, and ~10 GW is needed for more than 25 km ocean on Enceladus. Additionally, the thermal lithosphere of Mimas is thicker than that of Enceladus if the tidal heating rate for Mimas is less than ~20 GW. These values are consistent with the tidal heating rate evaluated by the orbital conditions of Mimas and Enceladus. The surface heat flux

is depressed if there is a methane hydrate layer, which is consistent with the unrelaxed surface of Mimas. It is assumed that a methane hydrate layer explains the thick subsurface oceans of both moons and the inactive shell of Mimas. Although we need a more detailed investigation of the tidal heating rate since it depends on an orbital evolution that we do not consider in this study, methane hydrate may be a key compound for understanding the histories of icy moons.

Declarations

Ethics approval and consent to participate

Not applicable

Consent for publication

Not applicable

List of abbreviations

CIRS: Composite infrared spectrometer; MMR: Mean motion resonance.

Availability of data and materials

The datasets used and/or analyzed during the current study are available

from the corresponding author on reasonable request.

Competing interests

No conflict of interest

Funding

This study was supported by KAKENHI from the Japan Society for the Promotion of Science (Grant No. JP17K05635 and JP17H06456).

Authors' contributions

RN designed and performed the numerical simulations. JK helped to develop the code and to interpret the results. AT and SS supervised this work. All authors read and approved the manuscript.

Acknowledgements

RN thank the economic support of Osaka University honors program for Graduate Schools in Science.

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Figures

Figure 1; Interior structure models of Enceladus and Mimas; These are snapshots of the interior structure of Enceladus and Mimas in a case where methane hydrate exists. From top to bottom, each layer represents ice, methane hydrate, ocean, and a rocky core.

Figure 2; Shell thickness evolution of Mimas with time; The changes in shell thickness of Mimas under 4 GW for tidal heating, 573 K for initial core temperature, and different initial methane concentrations are shown. (a) is the case of 1000 $\mathrm{mol\,m^{-3}}$ for initial methane concentration and (b) is the case of 0 $\mathrm{mol\,m^{-3}}$ (in other words, only ice exists). Each line indicates the boundary of each layer. The colour contour in both figures represents the ratio of F_{conv} to the total heat flux (=F_{conv} +F_{cond}). The coloured region shows the area in which the convection occurs.

Figure 3; Core temperature evolution of Mimas with time; The change in the rocky core temperature of Mimas under 4 GW for tidal heating rate, 573 K for initial core temperature, and 1000 $\mathrm{mol\,m^{-3}}$ of the initial methane concentration is shown.

Figure 4; The differences in shell thickness evolution among different initial core temperatures for Mimas; The icy shell growths of Mimas for an initial 100 Myr under

different initial core temperatures are shown. The added tidal heating rate is 4 GW, and the initial methane concentration is 1000 $\mathrm{mol\,m^{-3}}$. The bottom black line represents the ocean/core boundary. The top line represents the surface. Other lines represent ocean/shell boundaries under different initial core temperatures.

Figure 5; The differences in shell thickness evolution among different initial methane concentrations for Mimas; The final structures of Mimas under different initial methane concentrations are shown. The initial core temperature is 573 K, and the tidal heating rate is 4 GW.

Figure 6; The dependency of final shell thickness and thermal lithosphere on tidal heating rate; The final thickness of the shell and ocean and thermal lithosphere under different tidal heating rates are shown. The initial core temperature for Mimas is 573 K and that for Enceladus is 1273 K. (a) is the case with the methane hydrate layer for Mimas, that is, 1000 $\mathrm{mol\,m^{-3}}$ initial methane concentration, (b) is the case without the methane hydrate layer for Mimas, that is, 0 $\mathrm{mol\,m^{-3}}$ initial methane concentration, (c) is the case with the methane hydrate layer for Enceladus, and (d) is the case without the methane hydrate layer for Enceladus. The vertical axis represents the sum of the shell

and ocean. The horizontal axis represents the added tidal heating rate. In the area painted red in the figure, convection occurs; in other words, F_{conv} is not 0. The thermal lithosphere means that convection does not occur in the region, which is the white region of the shell in the figure.

Figure 7; Surface heat flux on Mimas; The current surface heat flux of Mimas is shown under an initial core temperature of 573 K. The solid line represents the case with methane hydrate, that is, the initial methane concentration is 1000 $\mathrm{mol\,m^{-3}}$. The dashed line represents the case without methane hydrate, that is, the initial methane concentration is 1000 $\mathrm{mol\,m^{-3}}$.

Figure 8; Summary of the results; The final ocean thicknesses for Mimas and Enceladus under different tidal heating rates are summarized. (a) is the case without the methane hydrate layer, that is, the initial methane hydrate is 0 $\mathrm{mol\,m^{-3}}$, and (b) is the case with the methane hydrate layer, that is, the initial methane hydrate is 1000 $\mathrm{mol\,m^{-3}}$. The vertical axis shows the tidal heating rate for Enceladus, and the horizontal axis shows that for Mimas. Each tidal heating rate can correspond to the final ocean thickness which is described at the opposite axis. The dashed lines correspond to 50 km of the

final ocean thickness for Mimas and to 25 km for Enceladus. The area surrounded by a dashed line means that a thick ocean exists in each moon. The grey area represents that at least one satellite – Mimas or Enceladus – does not have an ocean. The red area represents that the final lithosphere of Mimas is thicker than that of Enceladus. The blue area represents that the lithosphere of Mimas is thinner than that of Enceladus, although both moons have the inner ocean.

Preparing Tables

Table 1 The thermal and physical properties of Mimas and Enceladus.

Parameter	Symbol	Units	Mimas	Enceladus
Mean surface radius	R	km	198.2	252.0
Bulk density	ρ	$\mathrm{kg\,m^{-3}}$	1149	1611
Surface temperature	T_{surf}	K	76	59
Semimajor axis	a	km	185.539	237.948
Orbital eccentricity	e		0.0196	0.0047

Table 2 Parameters used in numerical simulation.

Parameter	Symbol	Units	Value
Density of ice	ρ_{ice}	$\mathrm{kg\,m^{-3}}$	920
Density of methane hydrate	ρ_{hyd}	$\mathrm{kg\,m^{-3}}$	920
Density of ocean	ρ_{oce}	$\mathrm{kg\,m^{-3}}$	1000

Density of rocky core	ρ_{core}	$\mathrm{kg\,m^{-3}}$	2450
Core radius of Mimas	$r_{\mathrm{c,mim}}$	km	98.9
Core radius of Enceladus	$r_{\mathrm{c,enc}}$	km	190.7
Specific heat of rocky core	$C_{\mathrm{p,core}}$	$\mathrm{J\,kg^{-1}\,K^{-1}}$	1000
Thermal conductivity of rocky core	k_{core}	$\mathrm{W\,m^{-1}\,K^{-1}}$	3.0
Activation energy of ice	$E_{\mathrm{a,ice}}$	$\mathrm{J\,mol^{-1}}$	60000
Activation energy of methane hydrate	$E_{\mathrm{a,hyd}}$	$\mathrm{J\,mol^{-1}}$	90000
Thermal expansivity of ice	α_{ice}	$\mathrm{K^{-1}}$	10^{-4}
Thermal expansivity of methane hydrate	α_{hyd}	$\mathrm{K^{-1}}$	10^{-4}
Latent heat of ice	L_{ice}	$\mathrm{J\,kg^{-1}}$	333000
Latent heat of methane hydrate	L_{ice}	$\mathrm{J\,kg^{-1}}$	437000

Table 3 The concentration and properties of radioactive elements.

	Concentration (ppb)	Decay energy (W/kg)	Half life time (Myr)
^{238}U	19.9	94.65×10^{-6}	4468
^{235}U	5.4	568.7×10^{-6}	703.81
^{232}Th	38.7	26.38×10^{-6}	14030
^{40}K	738	29.17×10^{-6}	1277