Methane hydrates achieve both thick oceans and thick lithospheres in Enceladus and Mimas

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Title page:

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16 **Abstract**

17 The difference between the inactive surface of Mimas and the active surface of
18 Enceladus is puzzling. We investigate the conditions under which both have a thick
19 subsurface ocean and the thermal lithosphere of Mimas is thicker than that of Enceladus
20 by using a one-dimensional simulation of thermal evolution. We adopt the initial core
21 temperature, initial methane concentration, and tidal heating rate as free parameters in
22 the calculation. The initial methane concentration and tidal heating rate greatly affect
23 the current ocean thickness, although the initial core temperature does not affect the
24 thickness. Methane hydrate forms at the base of the icy shell if the initial methane
25 concentration is not 0. The methane hydrate layer plays an insulative role in an icy shell.
26 When the initial methane concentration is 1000 mol m^{-3} , $\sim 2 \text{ GW}$ is needed to achieve
27 more than 50 km of the subsurface ocean on Mimas and $\sim 7.5 \text{ GW}$ is needed to achieve
28 more than 25 km of the subsurface ocean on Enceladus. These values are smaller than
29 those needed when the initial methane concentration is 0 mol m^{-3} . The existence of the
30 methane hydrate layer promotes the survival of the subsurface ocean because it
31 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

1. Introduction

Cassini has revealed that Saturnian moons have various appearances, although their surfaces are mainly covered with the same substance – H₂O 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; Čadež et al. 2016). According to these

48 data, the interior consists of a low-density core, thick ocean, and thin icy shell. The
49 plume is found to include nanosized silica particles and hydrogen, which indicates
50 ongoing hydrothermal activity at the seafloor (Waite et al., (2017); Hsu et al. (2015)).
51 These observations show that Enceladus is a hot and active body despite its small size.
52 In contrast to Enceladus, Mimas is dark. Many craters cover its surface. Although
53 absolute surface ages for Mimas and Enceladus are vague due to uncertainty of a
54 cratering rate, their crater densities indicate that Mimas surface is much older than most
55 of Enceladus surface (Kirchoff and Schenk (2009)). Additionally, there are few
56 fractures derived from geophysical activities. The largest crater, Herschell, is not overly
57 relaxed. Hence, Mimas is thought to be inactive. However, strong libration is observed,
58 which indicates a thick subsurface ocean or lumpy rocky core (Tajeddine et al. (2014)).
59 The phase lag of the libration supports the existence of a subsurface ocean. (Noyelles
60 (2017)). The suggestion of a subsurface ocean is confusing since craters on Mimas
61 should be more relaxed if the inside of Mimas is hot enough to have a global ocean
62 (Neveu and Rhoden (2017)). Rhoden et al. (2017) found that the tidal stresses evaluated
63 for the model of Mimas that included a deep ocean are hard to reconcile with Mimas'

64 lack of tectonic activity. Hence, it is still unclear whether Mimas possesses a subsurface
65 ocean.

66 The physical and orbital properties of these two moons are summarized in Table 1.

67 Enceladus is larger in size and has a higher bulk density than Mimas. These differences
68 in properties are thought to produce differences in the thermal evolution of Mimas and
69 Enceladus and cause the contrast between Mimas and Enceladus. To constrain the
70 conditions to achieve the current internal structures and surfaces of Enceladus and
71 Mimas, it is essential to know the history of these moons. Therefore, we investigate the
72 long-term thermal evolution of each moon under various conditions.

73 The major heat sources needed to sustain global oceans are decay of long-lived
74 radioactive elements in rocky cores and tidal heating in the solid parts of the moons.

75 Although other heating sources may contribute, their heating rates are much lower than
76 the aforementioned major sources (Malamud and Prialnik 2013; Hand et al. 2011;
77 Nimmo et al. 2007). The total heat derived from radioactive decay depends on the rocky
78 core size, and those of Mimas and Enceladus are too small to melt ice shells and sustain
79 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

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

where $\text{Im}(k_2)$ 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. Since Mimas' orbital eccentricity is larger than Enceladus' orbital eccentricity, depending on the value of $\text{Im}(k_2)$, Mimas may have experienced enough tidal heating to sustain the ocean. If $\text{Im}(k_2)$ is small, it will not experience tidal dissipation that would pull down the eccentricity. 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

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

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

96 where T is the torque caused by the primary, L is the angular momentum, M is the
97 mass of the primary, m is the mass of the moon, Q is the tidal dissipation factor, and
98 a is the semimajor axis. Subscripts 0 and 1 represent the inner moon and outer moon,
99 respectively. Subscript p represents the primary. Note that this equation does not rely on
100 the interior structure of the moon but on that of the primary (subscript p). The current
101 Q_p is suggested to be ~ 2000 from an astrometric study (Lainey et al. (2012)). At
102 present Enceladus and Dione are in a 2:1 eccentricity-type MMR. Assuming Q_p is now
103 2000, the equilibrium tidal heating rate for Enceladus is ~ 10 GW. This tidal heating
104 rate is insufficient for the survival of a thick ocean (Kamata (2018)). Mimas is not
105 currently in any eccentricity resonance, but it is likely to have experienced eccentric
106 resonance in the past (e.g., Meyer and Wisdom (2008)). The most recent eccentric
107 resonance that Mimas may have experienced is 3:2 with Enceladus. Mimas may have
108 experienced this resonance 0.34 Gyr ago, at which time Q_p was 4800 and the
109 equilibrium tidal heating rate was 3.6 GW (Tian and Nimmo, (2019)).
110 Dissolved ingredients in an ocean, such as ammonia, make it easier for the ocean to
111 survive because they depress the freezing point (Choukroun and Grasset (2010)).

112 However, high concentrations of ammonia (~ 15 wt%) and tidal heating of 15 GW are
113 needed to achieve an ocean of ~ 25 km consistent with observations (Kamata (2018)),
114 and this is unlikely in Enceladus and Mimas since such amounts of ammonia were not
115 found in the Enceladus plume (Waite et al. (2017)).

116 Alternatively, we consider the existence of a methane hydrate layer beneath a pure ice
117 layer in an icy shell. Methane hydrate is a crystalline compound where methane gas is
118 included in a cage made of a hydrogen bond network of water (Sloan Jr and Koh
119 (2007)). Its appearance and some physical properties are similar to those of ice.

120 However, the thermal conductivity of methane hydrate is approximately four times
121 lower than that of ice at 263 K (Waite et al. (2009)), and its viscosity is one order of
122 magnitude higher than that of ice (Durham et al. (2003)). These two characteristics may
123 contribute to the thick subsurface ocean, as heat conduction and convection are
124 depressed by the low thermal conductivity and the high viscosity.

125 Additionally, this insulating layer lowers the temperature of the icy shell, so the icy
126 shell becomes more rigid than the shell which is composed of only ice. This rigid shell
127 has an advantage in the maintenance of the unrelaxed craters of Mimas. The assumption

128 of a methane hydrate insulating layer succeeded in achieving both the maintenance
129 of the thick subsurface ocean and the ice shell thickness contrasts of Pluto
130 for a long time (Kamata et al. (2019)). In addition, Kalousová and Sotin (2020)
131 revealed that surface clathrate layer could prevent Titan's ocean freezing. We assume
132 that methane hydrate may similarly exist in Enceladus and Mimas. The plume
133 composition of Enceladus observed by Cassini includes not only water but also some
134 hydrophobic gases such as methane. With these plume compositions, gas hydrates can
135 form under the pressure and temperature of the subsurface ocean of Enceladus (Bouquet
136 et al. (2015)). If there are the methane hydrate layers in Mimas and Enceladus,
137 subsurface oceans of Mimas and Enceladus could survive with small amount of tidal
138 heating rate.

139 In this paper, we investigate the conditions under which both Enceladus and Mimas
140 have a thick inner ocean and where Mimas is less active than Enceladus, which means
141 that Mimas has thicker thermal lithosphere than Enceladus, by using a one-dimensional
142 simulation of thermal evolution under various conditions. In section 2, we describe the
143 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.

2. Methods

2.1 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 size of the rocky cores was calculated using the ice shell thickness and density currently estimated from observations. The ice thickness of Enceladus is assumed to be 20 km, the ice shell thickness of Mimas is assumed to be 30 km, and the density of the rocky cores is assumed to be 2450 kg m^{-3} . Other values are given in Table 3. Then, we obtained the size of the Enceladus rocky core is 190.7 km, and the size of the Mimas rocky core is

160 98.9 km. These values are consistent with the values estimated from observation data of
 161 Enceladus and Mimas (e.g., Tajeddine et al. (2014)). In the ocean, we assume that only
 162 methane has dissolved. We do not consider the presence of ammonia because the
 163 amount of ammonia observed by Cassini (~1 %) would not notably affect the ice
 164 freezing point (Choukroun and Grasset 2010; Vu et al. 2014). It is assumed that the
 165 composition formula of methane hydrate is $\text{CH}_4 \cdot 6\text{H}_2\text{O}$, and its density is 920 kg m^{-3} .

166 **2.2 Thermal evolution**

167 **2.2.1 Heat transfer**

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

$$172 \quad \rho C_p \frac{dT}{dt} = -\frac{1}{r^2} \frac{d}{dr} (r^2 F_{\text{cond}} + r^2 F_{\text{conv}}) + Q \quad (4)$$

173 where ρ is the density, C_p is the specific heat, T is the temperature, r is the radius,
 174 F_{cond} is the conductive heat flux, F_{conv} is the convective heat flux, and Q is the
 175 volumetric heating rate. The thermal conductivity and specific heat of each layer are

176 given in Table 2. F_{cond} and F_{conv} are given by

$$\begin{aligned}
 177 \quad F_{\text{cond}} &= -k \frac{dT}{dr} \\
 F_{\text{conv}} &= -k_v \left\{ \frac{dT}{dr} - \left(\frac{dT}{dr} \right)_s \right\}
 \end{aligned} \tag{5}$$

178 where k is the thermal conductivity, k_v is the effective thermal conductivity, and

179 $(dT/dr)_s$ is the adiabatic temperature gradient, that is $-\alpha g T / C_p$. α is the thermal

180 expansivity. k_v is given by

$$181 \quad k_v = \begin{cases} -\frac{\alpha C_p \rho^2 g l^4}{18\eta} \left\{ \frac{dT}{dr} - \left(\frac{dT}{dr} \right)_s \right\}, & \left(\frac{dT}{dr} \leq \left(\frac{dT}{dr} \right)_s \right) \\ 0, & \left(\frac{dT}{dr} > \left(\frac{dT}{dr} \right)_s \right) \end{cases} \tag{6}$$

182 where g is the gravitational acceleration, η is the viscosity, and l is the mixing

183 length. l depends on the computational grid temperature, shell thickness, and other

184 parameters (Kamata (2018)). Gravitational acceleration is determined from the equation

185 of hydrostatic equilibrium. The viscosities of ice and methane hydrate are as follows:

$$186 \quad \eta(T) = \eta_{\text{ref}} \exp \left(\frac{E_a}{R_g T_{\text{ref}}} \left(\frac{T_{\text{ref}}}{T} - 1 \right) \right) \tag{7}$$

187 where η_{ref} is the reference viscosity and corresponds to the viscosity at T_{ref} , T_{ref} is

188 the melting point of ice or the dissociation point of methane hydrate, E_a is the

189 activation energy, and R_g is the gas constant (e.g., Goldsby and Kohlstedt (2001)). We

190 assume that the reference viscosity of ice is 10^{14} Pa s and that of methane hydrate is

191 2.0×10^{15} Pa s (Kamata et al. 2019; Durham et al. 2003). These reference viscosities
192 have uncertainties since these values depend on the porosity and grain sizes of the icy
193 shell, and they are unknown. Solid convection in the rocky core is not considered, i.e.,
194 F_{conv} is 0 in the rocky core. We assume that heat received from the rocky core transfers
195 to the bottom of the icy shell immediately in the ocean, and the temperature remains
196 constant at the freezing point of the shell bottom. The adiabatic gradient in the ocean is
197 ignored since the gradient is small. Heat sources are the decay of long-lived radioactive
198 elements included in the rocky core and tidal heat in solid parts. The rocky core is
199 assumed to be carbonaceous chondrites. The amounts of long-lived radioactive elements
200 are summarized in Table 3 (Lodders (2003)). The details of the tidal heating rate are
201 described in a later section.

202 **2.2.2 Shell growth**

203 Icy shell growth is described by the energy balance at the shell/ocean boundary. If the
204 temperature at the shell bottom exceeds the melting point, differences between
205 incoming heat flux and outgoing heat flux will melt the icy shell. The change in the
206 shell thickness is given by

$$\rho_{ice} L_{eff} \frac{dD}{dt} = F_{out} - F_{in} \quad (8)$$

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

$$L_{eff} = L - \frac{\rho_{oce} g_{bot} C_{p,oce} V_{oce}}{4\pi R_{bot}^2} \frac{dT_m}{dP_{bot}} \quad (9)$$

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. $C_{p,oce}$ is the specific heat of pure water, which depends on temperature (Choukroun and Grasset (2010)), and is given by

$$C_{p,oce} = 4190 + 9 \exp(-0.11(T_{oce} - 281.6)) \quad (10)$$

where T_{oce} is the ocean temperature. The melting point of ice $T_{m,ice}$ depends on pressure, and is given by

$$T_{m,ice}(P) = 273.1 - c_1 P - c_2 P^2 \quad (11)$$

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

$$T_{m,hyd} = \frac{b}{\ln\left(\frac{P}{1000}\right) - a} \quad (12)$$

where $a = 38.980$ and $b = -8533.80$ 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 dissociation point. After the ocean temperature exceeds the dissociation 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 dissociation 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

$$m_{\text{CH}_4, \text{satu}, \text{bou}} = \exp \left(A \left(\frac{1}{T_{\text{oce}}} \right)^3 + B \left(\frac{1}{T_{\text{oce}}} \right)^2 + C \left(\frac{1}{T_{\text{oce}}} \right) + D \right) \quad (13)$$

where T_{oce} is the ocean temperature, $m_{\text{CH}_4, \text{satu}, \text{bou}}$ is the methane solubility at the ocean/shell boundary and its unit is mol m^{-3} , $A = 9.1589 \times 10^9 \text{ K}^3 \text{ mol m}^{-3}$, $B = -9.7261 \times 10^7 \text{ K}^2 \text{ mol m}^{-3}$, $C = 3.3963 \times 10^5 \text{ K mol m}^{-3}$, and $D = -3.8513 \times 10^2 \text{ mol m}^{-3}$. When methane hydrate dissociates or forms, the methane concentration in the ocean changes. Let the methane concentration in the ocean be $C_{\text{CH}_4, \text{oce}}$ and the change in methane concentration be $\Delta C_{\text{CH}_4, \text{oce}}$. $\Delta C_{\text{CH}_4, \text{oce}}$ is given by

$$\Delta C_{\text{CH}_4, \text{oce}} = \frac{C_{\text{CH}_4, \text{oce}} V_{\text{oce}} - C_{\text{CH}_4, \text{MH}} \Delta V_{\text{MH}}}{V_{\text{oce}} - \Delta V_{\text{oce}}} - C_{\text{CH}_4, \text{oce}} \quad (14)$$

where V_{oce} is the ocean volume, ΔV_{oce} is the change in the ocean volume, ΔV_{MH} is the change in the methane hydrate layer volume, and $C_{\text{CH}_4, \text{MH}}$ is the methane concentration included in methane hydrate. Since the composition formula of methane hydrate is $\text{CH}_4 \cdot 6\text{H}_2\text{O}$ and its density is 920 kg m^{-3} , $C_{\text{CH}_4, \text{MH}}$ is 7419 mol m^{-3} .

Once the methane concentration falls below the soluble concentration by methane hydrate formation, the dissociation points of methane hydrate decreases (Choukroun et al., 2010). Even if ocean is methane unsaturated, if the dissociation point of methane hydrate is higher than the melting point of ice, methane hydrate layer will grow.

255 However, when the methane in the ocean is less than 0.1%, the dissociation point of
 256 methane hydrate is about 40K lower than that of the case where ocean is supersaturated
 257 with methane. When methane hydrate is in equilibrium with water and methane gas, the
 258 methane solubility is about 0.11% under the conditions of 275.15 K and 3.2 MPa (Duan
 259 and Mao (2006)). Therefore, the growth of methane hydrate alone in the methane
 260 unsaturated ocean is ignored in this study. Instead, we introduced a mix layer in which
 261 methane hydrate and ice grow simultaneously when the ocean becomes methane
 262 unsaturated, and the mix layer begins to form when the ocean becomes methane
 263 unsaturated and the ocean temperature becomes equal to the melting point of ice at that
 264 depth. The properties of the mixture are a linear combination of methane hydrate and
 265 ice, with the exception of viscosity., which is given by

$$266 \quad A_{\text{mix}} = \phi_{\text{hyd}} A_{\text{hyd}} + (1 - \phi_{\text{hyd}}) A_{\text{ice}} \quad (15)$$

267 where A represents a material property and ϕ_{hyd} is the volume fraction of methane
 268 hydrate to ice in a calculated cell. The viscosity of the mixture is given by

$$269 \quad \eta_{\text{mix}} = \eta_{\text{hyd}}^{\phi_{\text{hyd}}} \eta_{\text{ice}}^{(1-\phi_{\text{hyd}})}. \quad (16)$$

270 **2.2.3 Calculation**

271 Our code can reproduce the results for Kamata et al. (2019), Kamata (2018), and
272 Kimura and Kamata (2020). The icy shell is divided into 100 thin layers. The maximum
273 thickness of the layer is ~ 1 km when there is no ocean in Mimas. The rocky core is also
274 divided into 100 thin layers. The time step is chosen to satisfy the Courant-Friedrichs-
275 Lewy (CFL) condition. The calculation is finished when the sum of the time steps
276 exceeds 4.5 Gyr.

277 **2.3 Initial conditions and parameter settings**

278 **2.3.1 Initial structure and temperature**

279 Our calculation starts after differentiation, which finishes within approximately 100
280 Myr after CAI formation (Schubert et al. (2007)). Differentiation was caused by heat
281 derived from short-lived radioactive species such as ^{26}Al and accretion energy. During
282 differentiation of Enceladus, its core temperature reached ~ 1273 K (Schubert et al.
283 (2007)). Mimas is also differentiated, but its core temperature is not heated as much
284 (Neveu and Rhoden (2017)). We believe the icy shell must have been melted initially,
285 and the liquid layer survives for the existence of the ocean because tidal dissipation
286 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. Even at the size of Pluto, the initial internal structure has little effect on the final internal structure (Kimura and Kamata (2020)).

2.3.2 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,

303 and their evolution is quite complicated. Tidal heating which depends on the viscosity
304 structure inside the icy shell, has been proposed (Tobie et al., (2003)), but the modified
305 mixing length theory cannot handle the heat generated inside the icy shell. Therefore,
306 we adopt an assumption similar to Kamata (2018), where the constant tidal heating rate
307 is added at the bottom of the icy shell. In our calculation, we consider the average tidal
308 heating rate as a free parameter in a range between 0 and 30 GW for simplicity. The
309 tidal heat is assumed to be generated in solid parts of the moons, namely, the rocky core
310 and icy shell. We do not consider dissipation in the ocean because its value is small
311 enough to ignore (Chen et al. (2014)). In a rocky core, the amount of tidal dissipation
312 depends on the properties of the rocky core. If the core is rigid, dissipation does not
313 effectively occur due to its high viscosity (Roberts and Nimmo (2008)). Porous rocky
314 cores such as gravel dissipate well, and high heat flow is generated (Choblet et al.
315 (2017)). In an icy shell, the dissipation depends on its viscosity. Tidal heat is produced
316 most effectively when the shell is near the melting point (Tobie et al. (2003)); i.e., tidal
317 heat is generated most at the shell bottom. Souček et al. (2019), which investigated the
318 effect of variations in ice shell thickness on tidal deformation in Enceladus, found that

319 tidal heating inside the ice shell is less than 1 GW, except near the south pole.
320 Therefore, it is not completely clear where and how much tidal heating is occurring in
321 the solid parts of Enceladus and Mimas. We assume that all of the tidal heating is used
322 at the base of the ice shell to melt the shell and increase ocean temperature. The average
323 tidal heat in our calculation represents the sum of the tidal heat generated in the icy shell
324 and the rocky core. Thus, F_{in} is described by

$$325 \qquad 4\pi R_{\text{bot}}^2 F_{\text{in}} = 4\pi R_c^2 F_c + Q_t \qquad (17)$$

326 where R_c is the radius of the rocky core, F_c is the heat flux at the surface of the rocky
327 core, and Q_t is the average tidal heating rate. Since tidal heating becomes ineffective
328 once the ocean freezes completely (Roberts and Nimmo (2008)), the tidal heating rate is
329 set to 0 if there is no ocean. In our model, we assume that all tidal heating is used to
330 melt the icy shell and increase the ocean temperature. Numerical experiment for the
331 convection of the ice shell suggests that the tidal heating occurs most at the bottom of
332 the icy shell (Tobie et al., (2003)), and recent numerical model for the tidal deformation
333 of Enceladus' icy shell indicates that the tidal dissipation in the icy shell is small
334 (Souček et al. (2019)). Nevertheless, tidal heating warms the lower part of the icy shell

and affects convection in the icy shell (Běhouňková et al., 2013). Therefore, in the case of a well-convected icy shell, our assumption is an oversimplification. If tidal heating were also used to increase the temperature of the icy shell, then convection in the icy shell would be promoted. As a result, the icy shell would be cooled more efficiently and the final ocean thickness would be thinner than the thickness determined in this study.

2.3.3 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 $\sim 600\text{ mol m}^{-3}$ 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 mol m⁻³. When the ocean temperature is 273.15 K and the pressure is 2.6 MPa, the solubility of methane is ~ 60 mol m⁻³. 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 ~1 % CH₄ 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.

3 Results

3.1 Change in shell thickness with time

Figure 2 shows one of the results of the icy shell thickness change with time for Mimas.

367 If a methane hydrate layer exists beneath the pure ice layer, the final ocean is thicker
368 than that in the case without methane hydrate layer under the same tidal heating rate and
369 initial core temperature. This is because the viscosity of the methane hydrate layer is
370 higher than that of ice, and the convection is depressed as shown in the formula of k_v .
371 The methane hydrate layer prevents the icy shell from convecting and insulates the heat
372 from the bottom of the shell or deeper region. As a consequence of the insulation of the
373 methane hydrate layer, the ice layer over the methane hydrate layer becomes cooler,
374 which leads to a higher viscosity of ice. Hence, the icy shell convection in the case with
375 methane hydrate is depressed and F_{conv} is nearly 0 in all areas of the shell.
376 The current shell thickness is achieved in the early stage of evolution. This is because
377 the energy of radioactive decay is much lower than tidal heating due to the small rocky
378 core. The thicknesses of the icy shells are controlled by both the radiogenic heating the
379 tidal heating. Because of the small size of both Enceladus and Mimas, the tidal heat at
380 the base of the icy shell immediately reaches a near equilibrium with the outgoing heat.
381 In fact, tidal heat is not constant with time, as it strongly depends on orbital evolution
382 and structural evolution of the interior. Therefore, the evolution of the internal structure

383 shown in Fig. 2 does not account for variations in tidal heating. However, as mentioned
384 above, the internal structure almost reaches equilibrium in 100 Myr, so the final internal
385 structure shown in Figure 2 can be achieved if the average tidal heating over the last
386 100 Myr is close to constant.

387 The core temperature evolution of Mimas was also investigated. However, the core
388 temperature rapidly decreases to near the ocean temperature without tidal heating. In
389 addition, the effect of the differences in the initial core temperature on the final shell
390 thickness is investigated. The shell thicknesses for each evolution which starts from
391 different initial core temperature attain a similar value in approximately 100 Myr, and
392 the final shell thicknesses of all initial core temperatures achieve the same value. It is
393 found that the final shell thickness does not depend on the initial core temperature, in
394 other words, the heat stored initially in the rocky core is much lower than the heat added
395 into the icy shell, namely the tidal heat and the heat generated by the decay of the
396 radioactive elements. Similar tendencies are also found in the results for Enceladus.
397 Therefore, in our calculation, the typical initial temperatures are chosen 573 K for
398 Mimas and 1273 K for Enceladus. These temperatures are consistent with 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).

3.2 The dependency of final shell thickness on initial methane concentration

Figure 3 shows the differences in final structures under different initial methane concentrations in the ocean and 3 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 500 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. 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 500 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 mol m^{-3} for the case with the methane hydrate

415 layer and 0 mol m^{-3} for the case without the methane hydrate layer.

416 **3.3 Differences in final shell thickness and lithosphere among different tidal** 417 **heating rates**

418 Figure 4 shows the final shell and ocean thickness and the thermal lithosphere of
419 Enceladus and Mimas under different tidal heating rates. The thermal lithosphere
420 denotes the nonconvective area in the shell where F_{conv} is 0. If the total dissipated
421 power is the same for Enceladus and Mimas, then the heating rate per unit mass for
422 Mimas is higher because it is smaller. In the regions between 2.0 GW and 5.0 GW for
423 Figure 4(a) and between 7.5 GW and 20 GW for Figure 4(c), the final shell thickness is
424 a plateau, where the methane hydrate layer insulates the shell from the inner heat and
425 the convective area in the shell is thin; in other words, the thermal lithosphere is thick.
426 If the methane hydrate layer exists in an icy shell, a thermal lithosphere is thicker than
427 that of the icy shell without the methane hydrate layer. This is attributed to methane
428 hydrate insulation. If the tidal heating rate for Enceladus is more than 20 GW and the
429 tidal heating rate for Mimas is more than 7.5 GW, no methane hydrate layers appear
430 inside the icy shell. This means that there is no region where the dissociation point of

methane hydrate is higher than the melting point of ice because the entire icy shell is at low pressure.

3.4 Surface heat flux of Mimas

Figure 5 shows the current surface heat flux versus the final shell thickness of Mimas.

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. When the thickness of the icy shell is smaller than 30 km, the surface heat fluxes are equal for the case with and without in the ocean. This is because, as mentioned in the previous subsection, if the icy shell is too thin, methane hydrate layers do not appear. The low surface heat flux on Mimas may contribute to the maintenance of surface topography on Mimas. Details are discussed in the discussion section.

3.5 Summary of results

In Figure 6, the results from Figure 4 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, we assumed that the criteria are about half of the H₂O layer (including the icy shell and the ocean). The area that is colored red and surrounded by the dashed line in the figure is *the preferred 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 ~2.0 GW and Enceladus needs ~7.5 GW of tidal heating rate in the case of methane hydrate, while Mimas needs ~5.0 GW and Enceladus needs ~15 GW in the case without methane hydrate. *The preferred 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.

4. Discussion

4.1 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

463 evolution of Mimas and Enceladus, depends not only on their internal structures but
464 also on their orbits. For moons with MMR, the equilibrium tidal heating rate can be
465 estimated as shown in equations (2) and (3). The tidal dissipation factor of the primary
466 Q_p in the equations influences the tidal heating rate in the moons and the orbital
467 evolution of the moons. An astrometric study estimated that the Q_p of Saturn is $\sim$
468 2000 (Lainey et al. (2012)). If $Q_p \sim 2000$ was constant, Mimas and Enceladus were
469 possibly formed more recently (e.g., Charnoz et al. (2011)). As shown in Figure 2, it is
470 found that the current interior of Mimas was achieved in 100 Myr from the initial state,
471 and the interior structure can be achieved even though Mimas formed 100 Myr ago. The
472 same finding applies to Enceladus whose size is slightly larger than that of Mimas.
473 Note, however, that we do not consider differentiation, failure, and the relaxation of
474 craters in the calculations.

475 Fuller et al. (2016) proposed that Q_p does not have to be constant and that the value
476 used to be higher; this was called the resonance locking hypothesis. Under this
477 hypothesis, Mimas and Enceladus could have formed ~ 4.6 Gyr ago and Mimas went
478 through 3:2 Mimas-Enceladus MMR 0.34 Gyr ago when Q_p was ~ 4800 , and the

479 equilibrium tidal heating rate for Mimas, following equations (2) and (3), was ~ 3.6
480 GW (Tian and Nimmo (2019)). The tidal heating rate can thicken the subsurface ocean
481 in Mimas if the methane hydrate layer exists. Now, Enceladus is in the Enceladus-Dione
482 MMR and the equilibrium tidal heating rate is ~ 10 GW. The tidal heating rate can
483 thicken the subsurface ocean in Enceladus if the methane hydrate layer exists.
484 Assuming constant tidal heating, we found that Enceladus and Mimas reach near
485 equilibrium within about 100 Myr. If tidal heating of 10 GW for Enceladus and 3.6 GW
486 for Mimas had been added for about last 100 Myr, a thick subsurface ocean would be
487 realized in both satellites if methane hydrate is present. Observational data are used to
488 estimate the current shell thicknesses of both bodies. The estimated shell thickness of
489 Enceladus ranges between 14 km and 50 km, which means that the subsurface ocean is
490 at least 10 km (e.g., Iess et al. (2014)). Figure 6 shows that 10 GW of tidal heating in
491 Enceladus can be achieved with ice alone for 10 km of ocean, but methane hydrate is
492 needed to achieve thicker ocean of 25 km or more. Libration data are used to estimate
493 the shell thickness of Mimas: the thickness ranges between 24 km and 31 km
494 (Tajeddine et al. (2014)), which corresponds to the ocean thickness between 68.3 km

495 and 75.3 km assuming that the core radius is 98.9 km. A tidal heating rate of 3.6 GW
496 for Mimas is sufficient to maintain 50 km of ocean in the presence of methane hydrate
497 layer, but not sufficient to maintain the ocean thickness between 68.3 km and 75.3 km.
498 This may be explained by the presence of dissolved gases other than methane in the
499 ocean. Dissolved gas, which can be encapsulated in gas hydrate cages, may affect the
500 final ocean thickness. The stable pressure for mixed hydrates of methane and other
501 gaseous species is different from that for pure methane hydrate, which changes the
502 ice/gas hydrate boundary depth and the final ocean thickness. Details of the dissolved
503 gaseous species in the ocean are discussed in a later subsection.

504 **4.2 Relevance to geophysical features**

505 Mimas has few tectonic features on its surface (Rhoden et al. (2017)). This is a large
506 constraint for the thermal evolution of Mimas. Rhoden et al. (2017) estimated the tidal
507 stress on the icy shell of Mimas when the subsurface ocean existed. The tidal stress is
508 over 1300 kPa when the viscosity of the shell of the lower parts is 10^{12} Pa s, which
509 leads to the shell failure. In order to reduce the tidal stress, the viscosity of the icy shell
510 must be high, i.e. the temperature of the icy shell must be low. 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, there must be no ocean or the 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 4(a) and 4(c), the final shell thickness changes little in the regions where methane hydrate exists (i.e., 2.0-5.0 GW for Mimas and 7.5-20 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 5, the heat flux is depressed sufficiently if the methane

hydrate layer exists. The surface heat flux is mainly determined by the total heat retained in the interior. As shown in Figure 4, if the same tidal heating rate is applied to the icy bodies with and without methane in the ocean, the final ocean thickness is thicker in the case with methane in the ocean. Therefore, when methane hydrate layers are present in the icy shell, both a thin ice shell and low surface heat flux can be achieved. 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

543 hydrate existence. Although it is difficult to consider a local geophysical feature
544 because our calculation has only one dimension, it may be unlikely that methane
545 hydrate or any other hydrate would form at the south polar terrain since the shell
546 thickness is thinner there than at any other region of Enceladus (Čadek et al. (2016)),
547 and the dissociation points of gas hydrates are lower than that of ice. Hence, the
548 viscosity of a shell at the south polar terrain that does not include methane hydrate may
549 be lower than that at the other region.

550 **4.3 Dissolved gaseous species**

551 Hydrate cages can incorporate not only methane but also other gases dissolved in the
552 ocean. Components observed in the Enceladus plume would form mixed hydrates under
553 pressure and temperature in the ocean (Bouquet et al. (2015)). Pure methane hydrate
554 and mixed hydrate have different properties such as dissociation point and density.
555 Because of the differences in dissociation points, the final ocean thickness may change
556 because the depth at which gas hydrates appear may change. The difference in density
557 affects whether the gas hydrate layer can exist beneath the ice layer. The density of
558 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 kg m^{-3} , and structure II mixed hydrates, with a density of 970 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)). If the temperature of the rocky core is high, carbon dioxide dominate in the ocean and gas hydrates may not form. 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 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

575 methane hydrate accretion when moons form.

576 The thermal evolution of icy bodies when such multiple gases are dissolved in the ocean
577 is a future work.

578 **5. Conclusion**

579 We investigate the conditions under which a thick subsurface ocean could be achieved
580 in Mimas and Enceladus, and Mimas is found to have a thicker thermal lithosphere than
581 that of Enceladus in a one-dimensional thermal evolution simulation. In our calculation,
582 the current shell thicknesses of Mimas and Enceladus are achieved within almost 100
583 Myr since both moons are small. If a methane hydrate layer exists beneath the ice layer,
584 the final ocean thickness is thicker than that in the case without methane hydrate
585 because the methane hydrate layer plays a role in shell insulation. If there is a methane
586 hydrate layer, ~ 2 GW is needed to sustain more than 50 km of subsurface ocean on
587 Mimas, and ~ 7.5 GW is needed for more than 25 km ocean on Enceladus. These
588 values are consistent with the tidal heating rate evaluated by the orbital conditions of
589 Mimas and Enceladus. Additionally, the thermal lithosphere of Mimas is thicker than
590 that of Enceladus if the tidal heating rate for Mimas is less than ~ 20 GW. 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

607 from the corresponding author on reasonable request.

608 **Competing interests**

609 No conflict of interest

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613 **Authors' contributions**

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

617

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760 **Figures**

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

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

771 Figure 3; The differences in shell thickness evolution among different initial methane
772 concentrations for Mimas; The final structures of Mimas under different initial methane
773 concentrations are shown. The initial core temperature is 573 K, and the tidal heating rate
774 is 3 GW.

775 Figure 4; The dependency of final shell thickness and thermal lithosphere on tidal heating

776 rate; The final thickness of the shell and ocean and thermal lithosphere under different
777 tidal heating rates are shown. The initial core temperature for Mimas is 573 K and that
778 for Enceladus is 1273 K. (a) is the case with the methane hydrate layer for Mimas, that
779 is, 1000 mol m^{-3} initial methane concentration, (b) is the case without the methane
780 hydrate layer for Mimas, that is, 0 mol m^{-3} initial methane concentration, (c) is the case
781 with the methane hydrate layer for Enceladus, and (d) is the case without the methane
782 hydrate layer for Enceladus. The vertical axis represents the sum of the shell and ocean.
783 The horizontal axis represents the added tidal heating rate. In the area painted red in the
784 figure, convection occurs; in other words, F_{conv} is not 0. The thermal lithosphere means
785 that convection does not occur in the region, which is the white region of the shell in the
786 figure.

787 Figure 5; Surface heat flux on Mimas; The current surface heat flux of Mimas is shown
788 under an initial core temperature of 573 K. The solid line represents the case with methane
789 hydrate, that is, the initial methane concentration is 1000 mol m^{-3} . The dashed line
790 represents the case without methane hydrate, that is, the initial methane concentration is
791 1000 mol m^{-3} .

Figure 6; 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 concentration is 0 mol m^{-3} , and (b) is the case with the methane hydrate layer, that is, the initial methane concentration is 1000 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	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

808 Table 2 Parameters used in numerical simulation.

Parameter	Symbol	Units	Value
Density of ice	ρ_{ice}	kg m^{-3}	920
Density of methane hydrate	ρ_{hyd}	kg m^{-3}	920
Density of ocean	ρ_{oce}	kg m^{-3}	1000
Density of rocky core	ρ_{core}	kg m^{-3}	2450
Core radius of Mimas	$r_{\text{c,mim}}$	km	98.9
Core radius of Enceladus	$r_{\text{c,enc}}$	km	190.7
Specific heat of rocky core	$C_{\text{p,core}}$	$\text{J kg}^{-1} \text{K}^{-1}$	1000
Specific heat of ice	$C_{\text{p,ice}}$	$\text{J kg}^{-1} \text{K}^{-1}$	2100
Specific heat of methane hydrate	$C_{\text{p,hyd}}$	$\text{J kg}^{-1} \text{K}^{-1}$	2100
Thermal conductivity of rocky core	k_{core}	$\text{W m}^{-1} \text{K}^{-1}$	3.0
Thermal conductivity of ice	k_{ice}	$\text{W m}^{-1} \text{K}^{-1}$	2.6
Thermal conductivity of methane hydrate	k_{hyd}	$\text{W m}^{-1} \text{K}^{-1}$	0.6
Activation energy of ice	$E_{\text{a,ice}}$	J mol^{-1}	60000
Activation energy of methane hydrate	$E_{\text{a,hyd}}$	J mol^{-1}	90000
Reference viscosity of ice	$\eta_{\text{ref,ice}}$	Pa s	1.0×10^{14}
Reference viscosity of methane hydrate	$\eta_{\text{ref,hyd}}$	Pa s	2.0×10^{15}
Thermal expansivity of ice	α_{ice}	K^{-1}	10^{-4}

Thermal expansivity of methane hydrate	α_{hyd}	K^{-1}	10^{-4}
Latent heat of ice	L_{ice}	J kg^{-1}	333000
Latent heat of methane hydrate	L_{hyd}	J kg^{-1}	437000

809 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

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 3 GW for tidal heating, 573 K for initial core temperature, and different initial methane concentrations are shown. (a) is the case of 1000 mol m^{-3} for initial methane concentration and (b) is the case of 0 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 $F_{\text{(conv)}}$ to the total heat flux ($=F_{\text{(conv)}}+F_{\text{(cond)}}$). The coloured region shows the area in which the convection occurs.

Figure 3

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 3 GW.

Figure 4

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 mol m^{-3} initial methane concentration, (b) is the case without the methane hydrate layer for Mimas, that is, 0 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 5

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 mol m^{-3} . The dashed line represents the case without methane hydrate, that is, the initial methane concentration is 1000 mol m^{-3} .

Figure 6

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 concentration is 0 mol m^{-3} , and (b) is the case with the methane hydrate layer, that is, the initial methane concentration is 1000 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.

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