Title: Pluto's ocean is capped and insulated by gas hydrates

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Many icy solar system bodies possess subsurface oceans. At Pluto, Sputnik Planitia's location near the equator suggests the presence of a subsurface ocean and a locally thinned ice shell. To maintain an ocean, Pluto needs to retain heat inside. On the other hand, to maintain large variations in ice shell thickness, Pluto's ice shell needs to be cold. Achieving such an interior structure is problematic. Here we show that the presence of a thin layer of clathrate hydrates (gas hydrates) at the base of the ice shell can explain both the long-term survival of the ocean and the maintenance of shell thickness contrasts. Clathrate hydrates act as a thermal insulator, preventing the ocean from complete freezing while keeping the ice shell cold and immobile. The most likely clathrate guest gas is methane either contained in precursor bodies and/or produced by cracking of organic materials in the hot rocky core. Nitrogen molecules initially contained and/or produced later in the core would likely not be trapped as clathrate hydrates, instead supplying the nitrogen-rich surface and atmosphere. The formation of a thin clathrate hydrate layer capping a subsurface ocean may be an important generic mechanism maintaining long-lived subsurface oceans in relatively large but minimally-heated icy satellites and Kuiper Belt Objects.

Liquid water oceans are thought to exist inside icy satellites of gas giants such as Europa and Enceladus and the icy dwarf planet Pluto[1]. Understanding the survival of subsurface oceans is of fundamental importance not only to planetary science but also to astrobiology. One indication of a subsurface ocean on Pluto is Sputnik Planitia, a ~1000-km-wide basin. It is a topographical low and is located near the equator, indicating that it is a positive gravity anomaly. To make this basin a positive gravity anomaly, a subsurface ocean beneath a locally thinned ice shell (by ~90 km) is inferred[2].

The presence of an ocean suggests a “warm” (i.e., not completely frozen) Pluto. However, unlike icy satellites, tides do not play an important role in heating the dwarf planet[3], and radiogenic heating is insufficient to avoid complete freezing unless the ice shell is highly viscous (>~10^{16} Pa s (ref. 4)). Heat may be retained inside Pluto owing to a thick surface thermal insulating layer resulting from high porosity and a high concentration of nitrogen ice[5]. However, if the ice shell is warm, its viscosity should be low[6]. Substantial viscous flow of ice would then occur, eliminating any local thickness contrasts[7]. Consequently, a locally thinned shell suggests a “cold” Pluto (<~200 K at the base of the ice shell[2]). One way to reconcile these contradictory requirements is to depress the freezing point of water by contaminating the ocean with anti-freeze molecules, such as ammonia[2]. However, to avoid substantial viscous relaxation of the ice shell, the ammonia concentration needs to be ~30 wt% (Supplementary Fig. 1). Such a high concentration is hard to justify; the ammonia concentration in comets, whose chemical composition should be representative of Kuiper Belt Objects (KBOs) including Pluto[8], are mostly less than 1 % with respect to water[9]. In addition, an ammonia concentration >20 wt% leads to an ocean density <1000 kg/m^3 (Supplementary Fig. 2, ref. 10), which makes it difficult for Sputnik Planitia to be a positive gravity anomaly[11]. Another possibility is that the viscosity of the ice shell is high due to the presence of high-strength material, such as salts[12] or silicate particles[13]. However, in a multiphase system, the weaker phase (i.e., water ice) dominates the flow[12,13]. Consequently, inhibiting substantial viscous relaxation requires the stronger phase (i.e., salts) to be the dominant component. Such an ice shell would have a high density (e.g., ~1.3 g/cm^3 for a mixture of 50% H_2O ice and 50% epsomite). While the density of the ocean might also be high due to the presence of salts in the ocean[11], their concentrations would be largely controlled by the formation of clay minerals and evaporates in the rocky core, keeping the ocean salinity < 4–5 mol/L or < 10 mol% relative to H_2O (ref. 14) and, thus, the ocean density to be < 1.3 g/cm^3. In order for the ice shell to float on the ocean and for Sputnik Planitia to be a positive gravity anomaly, the ice shell has a density much lower than the ocean (at least by ~0.2 g/cm^3 (ref. 11)), and its major constituent needs to be water ice.

Instead, we propose that the ice shell has a thin layer of clathrate hydrates at its base (Fig. 1). Clathrate hydrates, or gas hydrates, are solids in which water molecules create cages trapping gas molecules[15]. Because the formation temperatures of clathrate hydrates are higher than the melting point of pure water ice[15] and the subsurface ocean is sufficiently pressurized by the overlying ice shell, a freezing ocean would form clathrate hydrates rather than water ice if dissolved gas concentrations are sufficiently high[16]. The thermal conductivity of clathrate hydrates is about a factor of 5–10 smaller than that of water ice[17] and the viscosity of clathrate hydrates is about an order of magnitude higher than that of water ice[18]. Because of these physical properties, a cap of clathrate hydrates would act as a highly viscous thermal insulator between a subsurface ocean and an ice shell. The temperature difference across this layer allows the presence of a subsurface ocean with a temperature near the melting point of pure water ice while the overlying ice shell maintains a much lower temperature at the same time. Its high viscosity would also maintain shell thickness contrasts for a long time.

Global thermal evolution of Pluto:

We first evaluate the effect of clathrate hydrates on the thermal evolution of Pluto using the physical properties of methane hydrates ($CH_4 \cdot nH_2O$ where n ~ 6 (refs.

15,17)) (Methods). In the absence of clathrate hydrates, a subsurface ocean is expected to freeze completely because thermal convection in the ice shell removes heat effectively from the deep interior[4] (Fig. 2a, b). To maintain a thick subsurface ocean, the ice shell needs to be conductive, requiring a reference viscosity about one or two orders of magnitude higher than 10^{14} Pa s, which is a typical value for terrestrial ice sheets[19]. This may not be impossible but would require an extremely large ice grain size (a few cm)[20]. Interestingly, a thicker surface insulating layer leads to a shorter ocean lifetime, because such a layer enhances thermal convection in the ice shell beneath (Supplementary Fig. 3).

In contrast, if clathrate hydrates form, a thick subsurface ocean can be maintained for billions of years even if the reference viscosity of water ice is assumed to be 10^{14} Pa s (Fig. 2c). This is because heat from the ocean cannot be removed efficiently through the conductive clathrate hydrate layer. As the clathrate layer thickens the overlying water ice layer cools, leading to a higher viscosity. Thus, thermal convection in the ice shell becomes less vigorous, further reducing the freezing rate of the ocean (Fig. 2d). Consequently, the clathrate hydrate layer remains thin, while reducing the freezing rate of the ocean. If the clathrate hydrate layer grows from the beginning, its current thickness can reach ~30 km (Supplementary Fig. 4a, b). If its formation is delayed, its current thickness would be reduced (Supplementary Fig. 4c, d). Nevertheless, the efficiency of heat removal decreases significantly and the ocean thickness remains approximately constant following the formation of a clathrate hydrate layer.

As the ocean freezes and the ice shell thickens, the radius and surface area of Pluto increase, leading to the formation of normal faults on the surface[4]. Pluto's surface is covered by many such faults[21], and their pattern indeed supports global expansion of Pluto[22]. A future detailed image analysis estimating the change of Pluto's radius would provide a constraint on the clathrate hydrate layer thickness and the start timing and/or the duration of clathrate hydrate formation, which would inhibit freezing and faulting.

Viscous relaxation of the ice shell:

We next examine the effect of clathrate hydrates on the timescale of viscous relaxation of the ice shell using the physical properties of methane hydrates[15,17,18] (Methods). The origin of the local interior structure beneath Sputnik Planitia (a thin ice shell above a thick ocean) is considered to be associated with the large impact that formed the Sputnik Planitia basin[2,11]. Although the age of this impact is unknown, the eroded rim of the basin suggests that it is likely to be billions of years[23]. Consequently, the timescale of viscous relaxation should be at least one billion years.

If a clathrate hydrate layer does not exist, shell thickness contrasts can be maintained for only a few million years unless an extremely viscous ice shell and/or an extremely cold ocean is assumed (Fig. 3, Supplementary Fig. 5). In contrast, if a clathrate hydrate layer of a few to 10 km in thickness exists at the base of the ice shell, the timescale of ice shell relaxation is increased even by 3 to 4 orders of magnitudes and can exceed one billion years even if we assume typical pure-water properties both for the water ice layer and the ocean (Fig. 3). Such a large increase in the timescale results from the combined effect of a low temperature in the water ice layer and the high viscosity of clathrate hydrates (Supplementary Fig. 6). Thus, under the presence of a thin clathrate hydrate layer, an extremely ammonia- or salt-rich Pluto is no longer necessary to explain a long viscous relaxation timescale for the ice shell.

Implications for geochemical evolution:

Various gas species can form clathrate hydrates, though some species (e.g., CH_4) occupy clathrate hydrates more readily than others (e.g., N_2) (refs. 15,16). This behavior may explain the unique volatile composition observed on Pluto. Comets contain ~1 % of CH_4 and a few % of CO with respect to water[9], and even these low concentrations are sufficient to form a clathrate hydrate layer several tens of km in thickness (Methods, Supplementary Fig. 7). When Pluto formed, volatiles trapped in precursor bodies would have been partitioned between the atmosphere, ice shell, and the subsurface ocean. Gases may also have been initially trapped as clathrate hydrates near the surface[24], which may also have happened at Titan[25]. At an early stage, primordial CH_4 and CO dissolved in the ocean would likely form mixed clathrate hydrates at depth. Although precursor bodies of Pluto may be rich in CO_2 (ref. 9), CO_2 may not be a major guest molecule of clathrate hydrates above the ocean because its high density indicates that CO_2-rich hydrates would not float on the ocean unless the ocean is highly salty (Supplementary Table 1). CO_2 clathrate hydrates at the seafloor could have acted as a thermostat to prevent heat transfer from the core to the ocean. Primordial CO_2, however, may have been converted into CH_4 through hydrothermal reactions within early Pluto under the presence of Fe–Ni metals[26]. Because CH_4 and CO predominantly occupy clathrate hydrates, the components that degassed into the surface-atmosphere system would be rich in other species, such as N_2 (refs. 8,27,28). Trapping of CO in deep clathrate hydrates and degassing of N_2 may explain the low CO/N_2 on the surface of Pluto[29]. Further constraints on likely incorporation of cometary species, particularly CO, into clathrate hydrates are desirable, either via experiments or a detailed statistical thermodynamic approach[15].

As clathrate hydrate formation continues, concentrations of dissolved gases decrease if gases are not supplied to the ocean, eventually leading to the formation of pure water ice instead of clathrate hydrates at the interface between the ocean and ice shell. Thus, to form clathrate hydrates continuously in the ocean, secondary gases need to be continuously supplied to the ocean in the later stages. One plausible mechanism to supply gases is thermal cracking of organic materials in the rocky core, which would mainly produce CH_4 (refs. 30,31). Organic materials are abundant in cometary solids (~45 wt% of dust grains of 67P/Churyumov-Gerasimenko[32]). Thermogenic CH_4 can be produced where temperatures exceed ~150°C (ref. 33), and such a condition can be achieved in a large portion of Pluto's core for most of its history[8] (Supplementary Fig. 8). A high-temperature origin of CH_4 would leave a trace in its isotopic composition[34]. N_2 can also be produced via pyrolysis of organic matter when temperatures exceed ~350°C (ref. 35), though CH_4 would be preferentially trapped in clathrate hydrates[27]. Within a hot and porous core, high-temperature water-rock reactions would also occur. These can produce various gas species depending on many factors, including the redox state of the reactions, but the main gas species would be H_2 for chondritic rocks[36]. However, H_2 hydrates do not form unless H_2 is dominant in gases[37]. If organic materials within the rocky core contact with the hydrothermal fluids, a large quantity of C-bearing gas species, such as CH_4, also would be included in the fluids through hydrothermal decomposition[33]. Thus, under the presence of abundant CH_4, H_2 would be degassed to the surface and rapidly lost to the space because of its small mass. In contrast, heavier N_2 becomes a major volatile at Pluto's surface[38,39]. Gas production processes in a hot rocky core are unlikely to occur in small icy bodies. This may be why CH_4 and N_2 are found only on large KBOs such as Pluto and Eris but not on small KBOs such as Charon[38,40]. Furthermore, icy bodies possessing subsurface oceans may have low CO/N_2 and/or CH_4/N_2 ratios on their surface.

The current presence of subsurface oceans in outer solar system bodies is often explained by a high concentration of ammonia[1,41], though it is only rarely detected[9,38,42]. In contrast, a thin clathrate hydrate layer is equally effective at maintaining subsurface oceans and preventing motion of the ice shell, while requiring much lower concentrations of secondary species (e.g., CH_4) whose presence is commonly inferred. Such layers provide a likely explanation for minimally-heated but ocean-bearing worlds.

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Figure captions:

Figure 1: Schematic diagram of the interior structure of Pluto. The ice shell has a thin clathrate hydrate layer at its base. Temperature changes substantially across this layer immediately above the subsurface ocean, leading to a conductive shell rather than a convective shell. Nitrogen-rich ice on the surface is the bright surface of Sputnik Planitia.

Figure 2: Time evolution of the interior thermal profile above the rocky core. Here

the reference viscosity of water ice is 10^{14} Pa s, the initial ice shell thickness is 100 km, and the surface insulating layer thickness is 5 km. The green solid, dashed, and dotted curves indicate the surface of Pluto, the boundary between the ice shell and the ocean, and the boundary between the water ice layer and the clathrate hydrate layer, respectively. **a**, The temperature profile for the case without clathrate hydrate formation. The subsurface ocean becomes thin rapidly and freezes completely at ~3.8 Gyr. **b**, The ratio of the convective heat flux to the total heat flux (the sum of the convective and conductive heat fluxes) for the case of **a**. The lower part of the ice shell where temperature is nearly constant is highly convective. **c**, The temperature profile for the case with clathrate hydrate formation. Clathrate hydrate formation starts from the beginning (0 yr). The subsurface ocean remains thick, and the water ice layer is cold throughout. **d**, The ratio of the convective heat flux to the total heat flux for the case of **c**. Convection does not occur in the ice shell.

Figure 3: Timescale of viscous relaxation of the ice shell. Results for different layer thicknesses are shown. The reference viscosity of water ice is 10^{14} Pa s, and freezing-point depression due to impurities in the ocean is not considered. The presence of a clathrate hydrate layer ~5–10 km in thickness leads to a timescale of viscous relaxation longer than 10^9 yr.

Methods:

Thermal evolution:

To calculate the time evolution of the radial temperature profile of Pluto, we used the code developed by ref. 43. We modified the code to incorporate the effects of a clathrate layer and those of a surface thermal insulating layer.

We assume a 3-layer Pluto, consisting of an ice shell (solid), a subsurface ocean (liquid), and a rocky core (solid). The time evolution of temperature in the solid parts is obtained by solving the equation:

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

where ρ is density, C_p is specific heat, T is temperature, t is time, r is radial distance from the center, F_{cond} is conductive heat flux, F_{conv} is convective heat flux, and Q is volumetric heating rate. F_{cond} is given by the product of thermal conductivity and the local thermal gradient while F_{conv} is estimated using a modified mixing length theory[43] (see below).

The core is assumed to be purely conductive (F_{conv} = 0 W/m^2) and to be uniformly heated from within due to the decay of long-lived radioactive elements. The heating rate for carbonaceous chondrites[4,43] is used. The density, specific heat, and thermal conductivity of the core are 3000 kg/m^3, 1000 J/kg/K, and 3 W/m/K, respectively. The radius of the core is ≈910 km, which is calculated from eq. (2) of ref. 5 assuming the mean density of Pluto[44] (1854 kg/m^3), the present-day radius of Pluto[44] (1188 km), and the density of ices (including clathrate hydrates; 920 kg/m^3).

The subsurface ocean is assumed to be an inviscid fluid, and the ocean temperature is assumed to be uniform. For the case without clathrate hydrate formation, the ocean temperature is calculated from the pressure-dependent melting point of pure water ice[45]. For the case with clathrate hydrate formation, it is between the melting point of pure water ice and the dissociation temperature of methane hydrate[15]. No heat production in the ocean is assumed.

The ice shell is divided into 3 layers: a surface insulating layer (i.e., a top low-conductivity layer), a water ice layer, and a basal clathrate hydrate layer. The thickness of the top layer is assumed to be constant and to be 5 km unless otherwise noted. The thin top and basal layers are assumed to be purely conductive, and the thick intermediate layer is assumed to be convective or conductive depending on the viscosity. The thermal conductivities from the top to the bottom layers are 1 W/m/K (ref. 5), 0.4685 + 488.49/T W/m/K where T is temperature in Kelvin (ref. 46), and 0.6 W/m/K (ref. 17), respectively. The temperature-dependent specific heat of water ice[47] and that of methane hydrate[48] are used. No heat production (Q = 0 W/m^3) in the ice shell is assumed. F_{conv} for the intermediate part is given by

$$F_{\mathrm{conv}} = \begin{cases} -\frac{\alpha C_p \rho^2 g l^4}{18\eta}\left\{\frac{dT}{dr} - \left(\frac{dT}{dr}\right)_s\right\} \ (\text{for } \frac{dT}{dr} \leq \left(\frac{dT}{dr}\right)_s) \\ 0 \ \ (\text{for } \frac{dT}{dr} > \left(\frac{dT}{dr}\right)_s) \end{cases} \tag{2}$$

where $\alpha = 10^{-4}$ /K is thermal expansivity of ice, g is gravitational acceleration, l is the mixing length, η is viscosity of ice, and $(dT/dr)_s$ is the adiabatic thermal gradient. dT/dr is calculated by local thermal gradient, while $(dT/dr)_s$ is given by $-\alpha g T/C_p$. l is chosen so that it reproduces a scaling law between the Rayleigh number and the Nusselt number based on 3D numerical calculations[43,49]. l is updated at each time step since it depends on the thickness of the layer, rheological parameters, and the temperature difference across the layer[43]. η is given by

$$\eta = \eta_{\mathrm{ref}} \exp\left(\frac{E_a}{R_g T_{\mathrm{ref}}}\left(\frac{T_{\mathrm{ref}}}{T} - 1\right)\right) \tag{3}$$

where η_{ref} is reference viscosity of ice, E_a = 60 kJ/mol is the activation energy, R_g is the gas constant, T_{ref} = 273 K is the reference temperature[4,43,50]. The nominal value of η_{ref} is 10^{14} Pa s (ref. 19). Note that we assume a Newtonian rheology. This assumption is appropriate under typical conditions for terrestrial ice sheets (i.e., grain size ~ 1 mm, stress ~ 10^{-3} MPa, temperature near the melting point) (ref. 51). Nevertheless, Pluto's ice shell may exhibit non-Newtonian behavior (due to a large grain size, for example). The effect of non-Newtonian flow can be imitated by a Newtonian fluid with a smaller activation energy[52,53]. Calculation results using different activation energies are shown in Supplementary Fig. 9. We find that a smaller activation energy leads to a faster freezing of a subsurface ocean beneath a convective ice shell. Consequently, our model calculations using a Newtonian rheology provide the longest ocean lifetime for cases without a clathrate hydrate layer. On the other hand, for cases with a clathrate hydrate layer, different activation energies lead to nearly the same result because the ice shell is conductive. Thus, our conclusion does not change even if Pluto's ice shell exhibits non-Newtonian behavior. It is noted that results shown in Supplementary Fig. 9 assume the reference viscosity of water ice of 10^{14} Pa s. Although different creep mechanisms may lead to different reference viscosities, their quantification is left for another study.

The initial thickness of the ice shell is assumed to be 100 km. Initial temperature in the ice shell linearly increases with depth from 40 K at the surface to the pressure-dependent melting point of water ice[45] at the base of the ice shell. The initial temperature of the ocean and the rocky core is assumed to be uniform (i.e., the melting point of water ice). Different initial conditions do not affect the long-term evolution[4,43]. The exception is an initially completely frozen case; ref. 4 reported that a subsurface ocean does not appear if an initially completely frozen Pluto and $\eta_{\mathrm{ref}} \leq 10^{15}$ Pa s are assumed. However, the conditions required for the formation of a subsurface ocean based on their results are too strict because a freezing-point depression due to pressure is not incorporated in their calculations. A detailed investigation of such conditions is beyond the scope of this study and is left for another study.

The temporal change in the thickness of the ice shell is calculated from the difference between the outgoing and incoming heat fluxes at the base of the ice shell. Note that this difference in heat flux is used not only to change the thickness of the ice shell but also to change the temperature of the ocean. For the case without clathrate hydrate formation, the effect of the ocean temperature change caused by a change in the ice shell thickness is incorporated by using an effective latent heat[43]. For the cases with clathrate hydrate formation, the change in the thickness of the ice shell is interpreted as that of the clathrate hydrate layer; we assume the growth and dissociation rates of clathrate hydrates are simply controlled by heat transfer[15]. If the outgoing heat flux is higher than the incoming heat flux, the clathrate hydrate layer becomes thicker, keeping the ocean temperature constant. If the outgoing heat flux is lower than the incoming heat flux, the ocean temperature increases, keeping the layer constant until the ocean temperature reaches the pressure-dependent dissociation temperature of methane hydrate[15]. If the ocean temperature reaches the dissociation temperature, the clathrate hydrate layer becomes thinner. As the clathrate hydrate layer becomes thinner and the pressure at the top of the ocean decreases, the ocean temperature decreases because of the pressure dependence of dissociation temperature. Similar to the case without clathrate hydrate formation, the effect of pressure dependence of dissociation temperature is included by using an effective latent heat. Latent heats of water ice and methane hydrate are 333 kJ/kg and 437 kJ/kg (ref. 54), respectively. Note that the radius of Pluto changes as the thicknesses of the ice shell and the ocean (the density of 1000 kg/m^3 is assumed for the latter) change in order to conserve the total mass of Pluto. The initial radius of Pluto is determined so that the final radius becomes the present-day value through trial and error.

The surface temperature is fixed to 40 K, and the thermal gradient at the center is fixed to 0 K/m. The temperatures at the base of the ice shell and the top of the rocky core are the same and are given by temperature of the ocean if a subsurface ocean exists. If the ocean is completely solidified, the temperature at the boundary between the ice shell and the rocky core is obtained by equating the heat flux at the top of the rocky core to that at the base of the ice shell. Temperatures and heat fluxes at the boundaries within the ice shell are assumed to be continuous. For the case with clathrate hydrate formation, the start time of clathrate hydrate formation of 0 yr and no pre-existing clathrate hydrate layer are assumed unless otherwise noted. We calculate the thermal evolution for 4.6 Gyr for each calculation.

Viscous relaxation:

To calculate the timescale of viscous relaxation for the ice shell of Pluto, we followed the procedure adopted by ref. 7. We assume that the thinned portion of the ice shell has a bowl-shaped topography at the base of the ice shell. More specifically, the cross section can be described using a quadratic function, and the height and radius of the bowl are assumed to be 80 km and 500 km, respectively (Supplementary Fig. 5c). These values are chosen assuming a nearly (Airy) isostatically compensated basin before the loading of nitrogen-rich ice[2,11]. The shape of the basal topography is then expressed as a superposition of zonal components of spherical harmonic functions. In this study, we consider spherical harmonic degrees from 1 to 20. The time evolution of the amplitudes (coefficients) of each spherical harmonic for 10^{10} yr is obtained using the numerical code calculating spheroidal viscoelastic deformation of a planetary body developed by ref. 55 (see below). The time evolution of the basal topography can be calculated by superposing the spherical harmonics with time-dependent amplitudes. The timescale of viscous relaxation is defined as the time when the volume of the bowl becomes $1/e$ of the initial condition where e is Napier's constant.

The governing equations are the linearized equation of momentum conservation given by

$$\nabla_j \cdot (\sigma_{ij} - P\delta_{ji}) + \rho\nabla_i\phi = 0, \tag{4}$$

the Poisson's equation for the gravitational field given by

$$\nabla^2\phi = -4\pi G\rho, \tag{5}$$

and the constitutive equation for a Maxwell medium given by

$$\frac{d\sigma_{ji}}{dt} + \frac{\mu}{\eta}\left(\sigma_{ji} - \frac{\sigma_{kk}}{3}\delta_{ji}\right) = \left(\kappa - \frac{2\mu}{3}\right)\frac{de_{kk}}{dt}\delta_{ji} + 2\mu\frac{de_{ji}}{dt}, \tag{6}$$

where ∇_i is a spatial differentiation in direction of i (= x, y, z), σ_{ji} is stress tensor, e_{ji} is strain tensor, P is hydrostatic pressure, δ is the Kronecker delta, ϕ is gravitational potential, G is the gravitational constant, ρ is density, μ is shear modulus, η is viscosity, κ is bulk modulus. Application of spherical harmonic expansion to the governing equations leads to a six-component, time-dependent, inhomogeneous first-order ordinary differential equation system. The major assumptions are a spherically-symmetric steady-state interior structure, small deformation amplitudes, and a linear viscoelasticity[55]. These assumptions are valid to estimate the timescale of viscous relaxation, though more detailed numerical calculations would be necessary for precisely estimating the shape of the ice shell.

Following the thermal evolution calculations, we use a 3-layer Pluto model, though a steady-state thermal profile is adopted because what we calculate is the timescale of viscous relaxation under a given interior structure. We assume that the ice shell consists of Maxwell viscoelastic material, that the subsurface ocean is an inviscid liquid, and that the rocky core consists of purely elastic material. The radius of the core is determined in the same manner as that done in thermal evolution calculations. Different ice shell thicknesses, clathrate hydrate layer thicknesses, and surface insulating layer thicknesses are considered.

The densities of the ice shell, the ocean, and the core are 920 kg/m^3, 1000 kg/m^3, and 3000 kg/m^3, respectively, which are used in the thermal evolution calculations. The shear moduli of the ice shell, the ocean, and the core are 3.3 GPa, 0 GPa, and 10 GPa, respectively. We adopt an incompressible ($\kappa \rightarrow \infty$ and $e_{kk} \rightarrow 0$) limit because of the small size of Pluto. The viscosity in the ice shell is calculated from the temperature profile adopting the same rheological model. The surface temperature is fixed to 40 K. The temperature at the base of the ice shell is assumed to be the pressure-dependent melting point of pure water ice[56] unless otherwise noted. Assuming a steady-state conductive profile with given boundary temperatures, we calculate the temperature profile in the ice shell analytically. The thermal conductivity profile is the same as that used in thermal evolution calculations. The reference viscosities of water ice and clathrate hydrates are 10^{14} Pa s (ref. 19) and 2×10^{15} Pa s (ref. 18), respectively, unless otherwise noted. The activation energy of water ice and clathrate hydrates are 60 kJ/mol (ref. 50) and 90 kJ/mol (ref. 18), respectively, unless otherwise noted. The upper limit of the viscosity is 10^{30} Pa s, though the choice of this value does not affect the timescale of viscous relaxation of topography at the base of the ice shell.

Mass balance:

We calculate the amount of methane with respect to water in the ice shell and subsurface ocean. The thicknesses of these layers are determined in the same manner as in viscous relaxation calculations; the thickness of the ice shell is a free parameter while that of the ocean is calculated so that the mean density becomes the observed value. The ice shell is divided into an outer pure water ice layer and a deeper methane hydrate layer. We assume methane hydrate (Structure I) of full cage occupancy (i.e., $CH_4 \cdot 5.75\ H_2O$) (ref. 15). The ocean is assumed to consist of water and methane only. We use the pressure at the top of the ocean and the melting point of pure water ice[56] to calculate the (pressure- and temperature-dependent) solubility of methane in the ocean[57].

Data availability:

The data that support the plots within this paper and other findings of this study are available from the corresponding author on reasonable request.

Code availability:

Codes for the thermal evolution and viscous relaxation calculations are available upon reasonable request from S.K.

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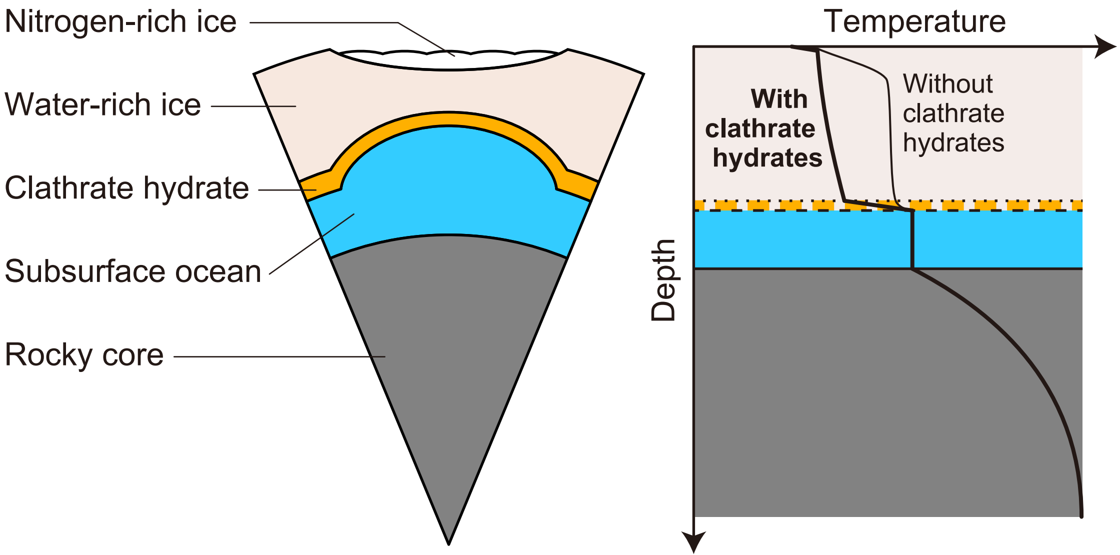
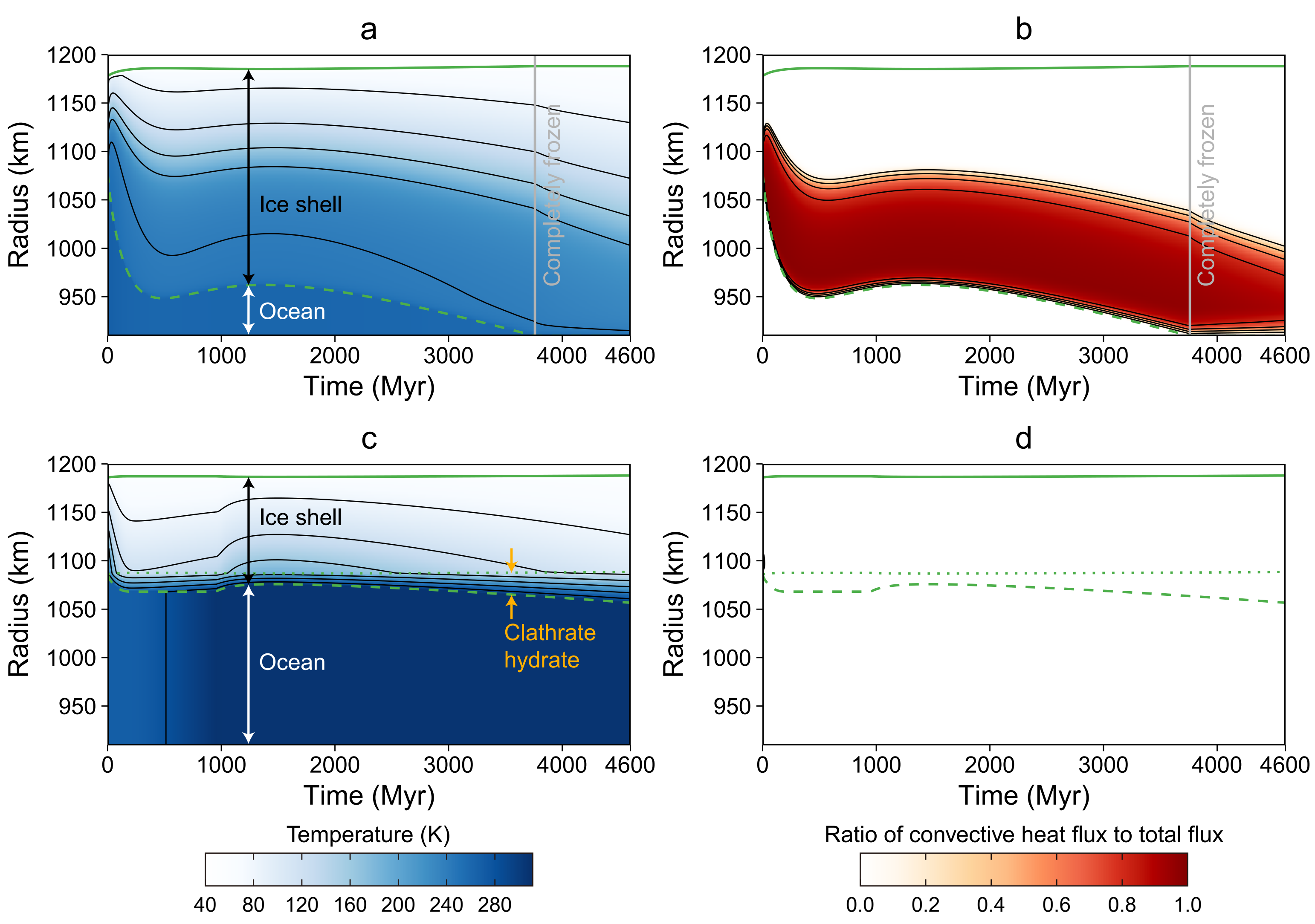
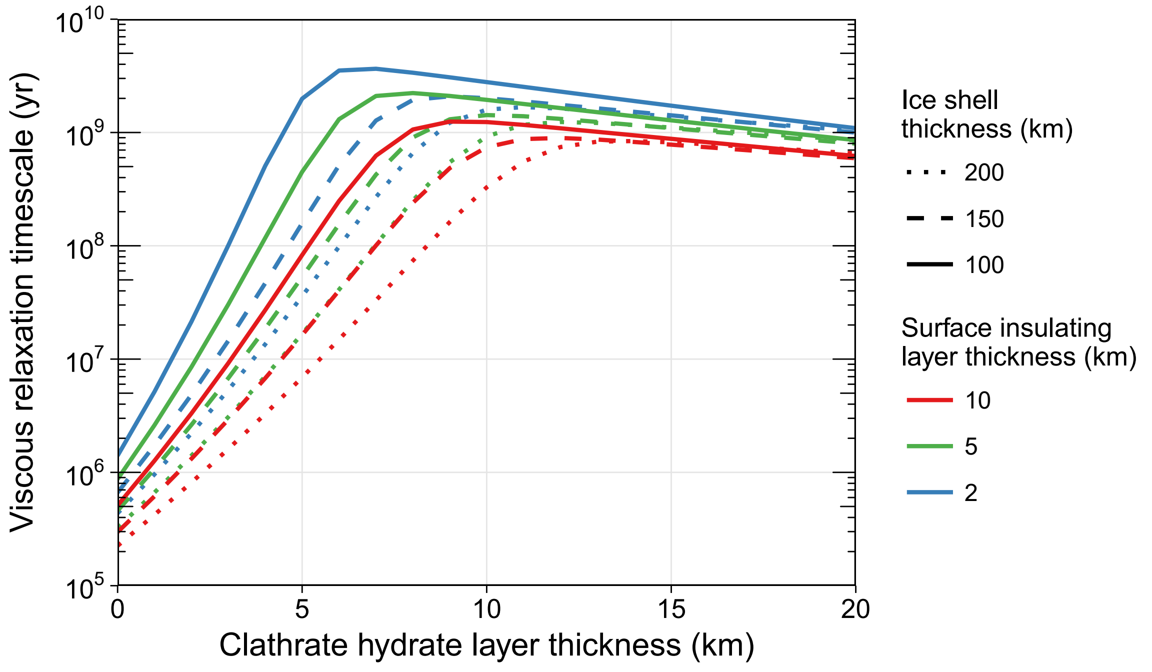

Nitrogen-rich ice
Water-rich ice
Clathrate hydrate
Subsurface ocean
Rocky core
Temperature
Depth
With clathrate hydrates
Without clathrate hydrates



a
b
c
d
Radius (km)
Time (Myr)
Ice shell
Ocean
Completely frozen
Clathrate
hydrate
Temperature (K)
40 80 120 160 200 240 280
Ratio of convective heat flux to total flux
0.0 0.2 0.4 0.6 0.8 1.0



Viscous relaxation timescale (yr)
10^{10}
10^9
10^8
10^7
10^6
10^5
0
5
10
15
20
Clathrate hydrate layer thickness (km)
Ice shell thickness (km)
200
150
100
Surface insulating layer thickness (km)
10
5
2