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Oceans in the icy Galilean satellites of Jupiter?

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Abstract

Equilibrium models of heat transfer by heat conduction and thermal convection show that the three satellites of Jupiter—Europa, Ganymede, and Callisto—may have internal oceans underneath ice shells tens of kilometers to more than a hundred kilometers thick. A wide range of rheology and heat transfer parameter values and present-day heat production rates have been considered. The rheology was cast in terms of a reference viscosity ν_0 calculated at the melting temperature and the rate of change A of viscosity with inverse homologous temperature. The temperature dependence of the thermal conductivity k of ice I has been taken into account by calculating the average conductivity along the temperature profile. Heating rates are based on a chondritic radiogenic heating rate of 4.5 pW kg^{-1} but have been varied around this value over a wide range. The phase diagrams of H_2O (ice I) and H_2O + 5 wt% NH_3 ice have been considered. The ice I models are worst-case scenarios for the existence of a subsurface liquid water ocean because ice I has the highest possible melting temperature and the highest thermal conductivity of candidate ices and the assumption of equilibrium ignores the contribution to ice shell heating from deep interior cooling. In the context of ice I models, we find that Europa is the satellite most likely to have a subsurface liquid ocean. Even with radiogenic heating alone the ocean is tens of kilometers thick in the nominal model. If tidal heating is invoked, the ocean will be much thicker and the ice shell will be a few tens of kilometers thick. Ganymede and Callisto have frozen their oceans in the nominal ice I models, but since these models represent the worst-case scenario, it is conceivable that these satellites also have oceans at the present time. The most important factor working against the existence of subsurface oceans is contamination of the outer ice shell by rock. Rock increases the density and the pressure gradient and shifts the triple point of ice I to shallower depths where the temperature is likely to be lower then the triple point temperature. According to present knowledge of ice phase diagrams, ammonia produces one of the largest reductions of the melting temperature. If we assume a bulk concentration of 5 wt% ammonia we find that all the satellites have substantial oceans. For a model of Europa heated only by radiogenic decay, the ice shell will be a few tens of kilometers thinner than in the ice I case. The underlying rock mantle will limit the depth of the ocean to 80–100 km. For Ganymede and Callisto, the ice I shell on top of the H_2O–NH_3 ocean will be around 60- to 80-km thick and the oceans may be 200- to 350-km deep. Previous models have suggested that efficient convection in the ice will freeze any existing ocean. The present conclusions are different mainly because they are based on a parameterization of convective heat transport in fluids with strongly temperature dependent viscosity rather than a parameterization derived from constant-viscosity convection models. The present parameterization introduces a conductive stagnant lid at the expense of the thickness of the convecting sublayer, if the latter exists at all. The stagnant lid causes the temperature in the sublayer to be warmer than in a comparable constant-viscosity convecting layer. We have further modified the parameterization to account for the strong increase in homologous temperature, and therefore decrease in viscosity, with depth along an adiabat. This modification causes even thicker stagnant lids and further elevated temperatures in the well-mixed sublayer. It is the stagnant lid and the comparatively large temperature in the sublayer that frustrates ocean freezing.

Keywords: Satellites of Jupiter; Ices; Europa; Ganymede; Callisto

Introduction

Observations of electromagnetic induction signatures at Europa and Callisto have been interpreted as indicative of sublithosphere liquid water oceans in these satellites (Khurana et al., 1998; Neubauer, 1998; Kivelson et al., 1999, 2000; Zimmer et al., 2000). The magnetic field data gathered during several close fly-bys of Ganymede have also been proposed to be in part due to induction in an ocean in this satellite as well (Kivelson et al., 2002). If these interpretations are correct, then the magnetic field data will

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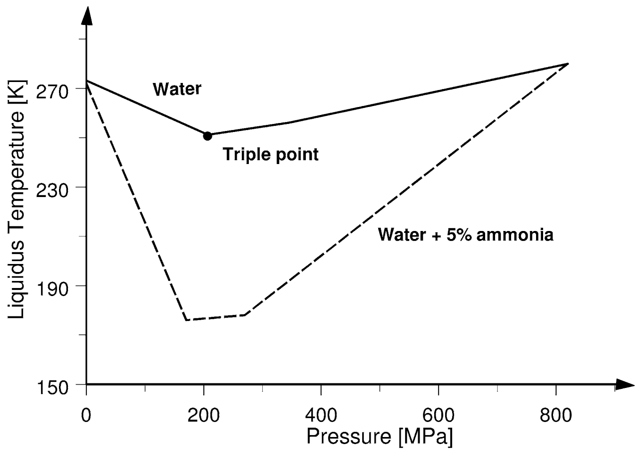

Fig. 1. Simplified phase diagram of water ice and of water ice + 5 wt% ammonia (Grasset and Sotin, 1996) as used in the present study.

allow estimates of the thickness of the ice layers covering the oceans. For Europa, the strength of the induced signal suggests an ocean underneath a thin ice shell a few tens of kilometers thick. Hussmann et al. (2002) have shown that tidal heating in such an ice shell can keep a subsurface europan ocean from freezing. For Ganymede and Callisto the oceans are at greater depths of 100 km or more. For Europa, the existence of an ocean underneath tens of kilometers of ice is supported by geological evidence (Pappalardo et al., 1999) and some authors favor even thinner ice layers (e.g., Greenberg et al., 1998; Kattenhorn, 2002). Similar geologic evidence is not available for Ganymede and Callisto. At least for Callisto, the inferred existence of an ocean is surprising since this satellite is incompletely differentiated (Anderson et al., 2001; Sohl et al., 2002) and shows little sign of past endogenic activity (as witnessed by its old surface).

Subsurface oceans on satellites are possible because of the anomalous melting behavior of ice I for which the melting temperature decreases with pressure until it joins the ice I/ice II transition and the ice III melting curve in a triple point at a pressure of 207 MPa and a temperature of 251.15 K (Chizov, 1993; Fig. 1). This triple point pressure translates into different depths for the three satellites because of their differing masses and, possibly, ice shell densities. For Europa, assuming an ice shell density of 1000 kg m^{-3}, a depth of 160 km is obtained, about as deep as the thickness of the water layer, which is believed to be around 150 km (Anderson et al., 1998; Sohl et al., 2002). The depth of the minimum melting temperature is about 145 km in Ganymede if an ice shell density of 1000 kg m^{-3} is assumed. This density is reasonable since Ganymede is most likely differentiated (Anderson et al., 1996). For Callisto, the ice shell density could be about 1600 kg m^{-3} if the satellite is undifferentiated or has a mostly undifferentiated, cold and stiff outer shell. The value is calculated from the density of ice I, a concentration of rock of about 50 wt%, and a rock density of 3500 kg m^{-3}. With the latter ice shell density, the depth to the triple point pressure is about 104 km. If the outer shell in Callisto were pure ice, the depth to the triple point would be about 166 km.

It is possible that the melting point is even further depressed if the ice in the satellites is not pure H_2O but contains other components such as ammonia, methane, and/or salts (e.g., Kargel, 1992). The phase diagram of the water–ammonia system is well studied (e.g., Hogenboom et al., 1997; Sotin et al., 1998) and can serve as a model. Ammonia hydrates are predicted condensates in the satellites of Jupiter (e.g., Lewis, 1971) and the concept of internal layers of ammonia–water liquid is well developed in the literature (see, e.g., Kargel, 1998; Sotin et al., 1998 for reviews). The water–ammonia liquidus temperature depends on pressure and on the concentration of ammonia in the water. The liquidus surface in a concentration–pressure–temperature phase diagram contains the points 5 wt% NH_3, $P = 0.1$ MPa, and $T = 266.92$ K; the low-pressure peritectic at 32.1 wt% NH_3, 0.1 MPa, and 176 K, and the high-pressure eutectic at 29 wt% NH_3, 170 MPa, and 176 K. The solidus up to 170 MPa is an isothermal surface at 176 K to a good approximation.

The evolution of an ice shell and ocean will depend not only on the bulk concentration of ammonia but also, to some extent, on the initial conditions. If the shell grew on top of an ocean with some initial concentration of ammonia and began from zero thickness, then the shell will be pure water ice. The composition of the ocean will be determined by the mass of the water ice removed from the ocean and the constancy of the ammonia mass. The ice shell bottom temperature will be the liquidus temperature and will be determined by the composition of the ocean and the pressure at the bottom of the lid. As the shell thickens, the shell bottom temperature will describe a path on the liquidus surface. Grasset and Sotin (1996) have suggested a possible path on the liquidus surface starting at 5 wt% NH_3, $P = 0.1$ MPa, and $T = 266.92$ K and continuing to the high-pressure eutectic at 29 wt% NH_3, 170 MPa, and 176 K (compare Fig. 1). The pressure of 170 MPa is equivalent to a depth of about 130 km on Europa, 120 km on Ganymede, and 135 km on Callisto.

If the satellite were initially at subsolidus temperatures and was warmed to reach the ammonia hydrate solidus, a partial melt would form at a temperature of 176 K and mostly independent of pressure. The composition of the liquid, under these circumstances, will be around 30 wt% ammonia, decreasing slightly with pressure, because the composition along the eutectic/peritectic is almost constant. The solid phase in the partial melt will be H_2O ice. Since the ice is less dense than the fluid, the ice will tend to float on top of the forming ocean. The rate at which the ice will separate and float will depend on the rheology and the concentration of the melt in the solid. As the temperature increases, water will dilute the melt. The increasing concentration of the melt will further the formation of an ocean by facilitating the gravitational separation of ice from the melt. The composition of the ocean once formed and the temperature at the base of the ice shell will be determined by the pressure at the base of the ice shell, the bulk concentration of ammonia, and the liquidus surface.

Since we are interested in equilibrium configurations, we will neglect any rates of separation of the ice from the liquid. Rather, we assume that ice and melt have separated and that the composition of the ocean and the temperature at the bottom of the ice shell are given by the Grasset and Sotin (1996) liquidus.

For Europa, it is possible to attribute the existence of an ocean to tidal heating. This is not possible for Ganymede and Callisto; tidal heating is negligible at their present orbital distances and eccentricities. On these satellites, oceans must be due to radiogenic heating or to heat buried at depth and released through satellite cooling. It has been speculated (e.g., Malhotra, 1991; Showman and Malhotra, 1997; Showman et al., 1997) that Ganymede went through a phase of intense tidal heating perhaps as little as 1 Gyr ago when it passed into the present resonance through a 3:2 resonance. During such a transition, the eccentricity would have increased by as much as a factor of 10, which would have increased the heating rate by approximately a factor of 2 above the radiogenic heating rate (Spohn and Breuer, 1998).

In this paper, we will investigate the possibility of present oceans in the Galilean satellites as a consequence of an assumed heat flow from below the ocean. That heat flow may be due to radiogenic heating, tidal heating, and/or satellite cooling, but the model is general and does not *a priori* attribute the heat flow to a particular mechanism.

The model

The model is shown in Fig. 2. An ice layer of thickness D lies on top of an ocean of thickness D_o. The basal temperature of that ice layer is T_b. The thickness of the ocean is determined by equating T_b with both branches of the ice liquidus that define the dip in the ice melting temperature profile with pressure (depth) as shown in Fig. 2. The heat flow from the deeper interior is q. The existence of an ocean is dependent on the rate of heat transfer through the top ice layer of thickness D. The rate of heat transfer depends on the temperature difference $(T_b - T_s)$ across the layer regardless of whether heat transfer is by convection or entirely by conduction. Since the surface temperature T_s is constant at about 100 K, this translates into a dependence on the basal temperature of the layer. In equilibrium, the basal temperature T_b is determined by the heat flow from below.

We approximate the first branch of the ice liquidus between zero pressure and 207 MPa by the linear relation

$$T_m(P) = T_{m0} + \frac{dT_m}{dP} P, \tag{1}$$

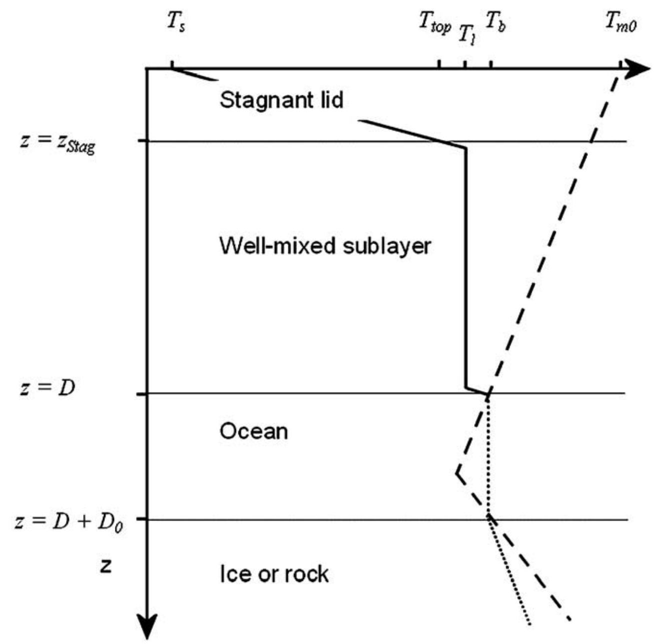

Fig. 2. Sketch of the model. The top layer is the stagnant lid. Below the stagnant lid is the well-mixed sublayer through which heat is transferred by convection. The temperature profile in the ice shell is shown with solid lines. The melting temperature is shown with dashed lines and the temperature profile in the ocean and the ice layer below the ocean is shown with dotted lines. Note that on the top and bottom of the well-mixed layer there are thermal boundary layers with steep temperature gradients.

where T_m is the liquidus temperature, P is pressure, and T_{m0} is T_m at zero pressure. Values of parameters are summarized in Table 1. Heat transfer through the ice shell can be either by conduction or by thermal convection. With heat conduction alone, the heat flow through the ice layer is

$$q_c = k\frac{(T_b - T_s)}{D}, \tag{2}$$

where k is the thermal conductivity, assumed constant for simplicity in this study.

With $T_b = T_m(P)$ and $q_c = Q/4\pi(R_P - D)^2$, where Q is the heat flow from below and R_P is the satellite radius, the shell thickness D from Eq. (2) is

$$D \approx \frac{T_{m0} - T_s}{\dfrac{Q}{4\pi R_P^2 k} - \dfrac{dT_m}{dz}}, \tag{3}$$

where we have replaced $(R_P - D)^2$ with R_P^2 for simplicity, z is depth,

$$\frac{dT_m}{dz} = \frac{dT_m}{dP}\rho_i g, \tag{4}$$

ρ_i is the ice shell density, and g is the acceleration of gravity.

Table 1
Parameter values used for the standard models

Parameter	Symbol	Moon		
		Europa	Ganymede	Callisto
Ice I_h melting temperature at $P = 0$	T_{m0} [K]	273.16		
Ice I_h melting temperature at $P = 207$ MPa	T_{m207} [K]	251.15		
Slope of ice I_h melting temperature	$\frac{dT_m}{dP}[10^{-7}KPa^{-1}]$	−1.063		
Ice I + 5% NH_3 melting temperature at $P = 0$	T_{m0} [K]	271.6		
Ice I + 5% NH_3 melting temperature at $P = 70$ MPa	T_{m207} [K]	176.0		
Slope of Ice I + 5% NH_3 melting temperature	$\frac{dT_m}{dP}[10^{-7}KPa^{-1}]$	−5.647		
Density of ice shell	ρ_l [10^3 kg m^{-3}]	1.0	1.0	1.6 (1.0)
Satellite surface gravity	g[m s^{-2}]	1.30	1.58	1.25
Chondritic heat production rate	$Q[10^{11}$W]	1.74	3.36	2.42
Surface heat flow in equilibrium with chondritic heating rate	$\frac{Q}{4\pi R_p^2}[mWm^{-2}]$	5.62	3.87	3.35
Surface temperature	T_s[K]	105.	120.	130.
Average ice I_h thermal conductivity (between 100 and 273 K)	k[Wm^{-1} K^{-1}]	3.3		
Average ice I_h thermal conductivity (between 100 and 177 K)	k[Wm^{-1} K^{-1}]	4.3		
Average thermal diffusivity	$\kappa[10^{-6}$ m^2 s^{-1}]	3.7		
Activation parameter for solid state creep	A	24.		
Melting point viscosity	v_0[m^2 s^{-1}]	10^{11}		
Thermal expansion coefficient	$\alpha[10^{-4}$ K^{-1}]	1.6		
Nu–Ra exponent	β	0.25		
Nu–Ra prefactor	a	0.25		

If the temperature difference across the ice layer is large enough and the viscosity of the ice is suitably small, the ice layer will be unstable to thermal convection. In that case the temperature-dependent viscosity of the ice becomes important. We assume that

$$\nu = \nu_0 \exp A\left(\frac{T_m}{T} - 1\right), \tag{5}$$

where A is a constant and ν_0 is the viscosity at T_m. The dependence of the viscosity on the homologous temperature T/T_m is supported by experimental data (Durham et al., 1997) and accounts for both the pressure and temperature dependence of the viscosity. Experimental and theoretical studies of convection in fluids with strongly temperature dependent viscosity (e.g., Davaille and Jaupart, 1993; Solomatov, 1995; Grasset and Parmentier, 1998) have shown that the convecting layer consists of a stagnant near-surface lid (see Fig. 2), in which temperature increases rapidly and viscosity decreases steeply with depth, and a well-mixed sublayer below, through which viscosity decreases by approximately one order of magnitude with depth. The depth of transition between the stagnant lid and the well-mixed sublayer is z_{stag}. The bottom temperature of the stagnant lid and the top temperature T_{top} of the convecting well-mixed layer are then

$$T_{top} = T_m\,(z = z_{stag})\left(\frac{A}{A + \ln 10}\right). \tag{6}$$

The temperature of the well-mixed sublayer is

$$T_1 = \frac{T_{top} + T_m(D)}{2}. \tag{7}$$

The viscosity of the well-mixed sublayer is simply given by

$$\nu\left(\frac{T_{top} + T_m(D)}{2}\right) \equiv \nu_{conv} = \nu_0\sqrt{10}. \tag{8}$$

The convective heat flow through the well-mixed sublayer can be calculated from

$$q_{conv} = k\frac{(T_m(D) - T_{top})}{d_{conv}} \cdot Nu, \tag{9}$$

where d_{conv} is the thickness of the well-mixed sublayer and Nu is its Nusselt number, given by

$$Nu = aRa^{\beta}. \tag{10}$$

The parameter a and β are constants; a takes values between 1 and 10 and β takes values between 0.2 and 1/3. The Rayleigh number, Ra, is defined as

$$Ra \equiv \frac{\alpha g\,(T_m(D) - T_{top})d^3_{conv}}{\kappa\nu_{conv}}, \tag{11}$$

where α is the thermal expansivity and κ is the thermal diffusivity. In thermal equilibrium, the (conductive) heat flow through the stagnant lid equals q_{conv}. With (10), Eq. (9) becomes

$$q_{\mathrm{conv}} = a\left(\frac{\alpha g}{\kappa \nu_{\mathrm{conv}}}\right)^{\beta} k(T_{\mathrm{m}}(D) - T_{\mathrm{top}})^{1+\beta} d_{\mathrm{conv}}^{3\beta}$$
$$= k\left(\frac{T_{\mathrm{m}}(D) - T_{\mathrm{top}}}{d_{\mathrm{conv}}}\right)\left(\frac{T_{\mathrm{m}}(D) - T_{\mathrm{top}}}{T_{\mathrm{m0}}}\frac{d_{\mathrm{conv}}^3}{l_0^3}\right)^{\beta} Nu_0, \tag{12}$$

where Nu_0 is given by

$$Nu_0 \equiv a \cdot Ra^{\beta}{}_0 \equiv a \cdot \left(\frac{\alpha g T_{\mathrm{m0}} l_0^3}{\kappa \nu_{\mathrm{conv}}}\right)^{\beta}. \tag{13}$$

The thickness of the stagnant lid z_{stag} is then given by

$$z_{\mathrm{stag}} \approx \frac{T_{\mathrm{m0}}\dfrac{A}{(A+\ln 10)} - T_{\mathrm{s}}}{\dfrac{Q}{4\pi R_{\mathrm{p}}^2 k} - \dfrac{dT_{\mathrm{m}}}{dz}\dfrac{A}{(A+\ln 10)}} \tag{14}$$

and the thickness of the entire ice layer is obtained by solving

$$\left(\frac{\Theta}{Nu_0}\right) \approx \left(1 + \frac{1}{T_{\mathrm{m0}}}\frac{dT_{\mathrm{m}}}{dz} D - \frac{T_{\mathrm{top}}}{T_{\mathrm{m0}}}\right)^{1+\beta} \times \left(\frac{D - z_{\mathrm{stag}}}{l_0}\right)^{3\beta-1} \tag{15}$$

for D, where we have again replaced $(R_{\mathrm{p}} - D)^2$ with R_{p}^2 and where

$$\Theta \equiv \frac{Q l_0}{4\pi R_{\mathrm{p}}^2 k T_{\mathrm{m0}}}. \tag{16}$$

The transition between the conductive and the convective regimes occurs according to whichever heat flow is larger.

Parameter values

The most important model parameters are the thermal conductivity of the ice, k, the convective heat transfer parameters a and β, the rheology parameters, and the heat flow from below. The thermal conductivity of ice is known to be an inverse function of temperature. For compact, hexagonal crystalline ice I_{h} the empirical relation is (Klinger, 1973; Ross and Kargel, 1998)

$$k(T) = \frac{567}{T}, \tag{17}$$

where temperature T is in kelvins. In the temperature range of interest, between 100 and 300 K, this translates into a variation of k between 2 and 6 W m^{-1} K^{-1}. Noncrystalline and porous ices have significantly smaller thermal conductivities of less than 1 W m^{-1} K^{-1} (e.g., Ross and Kargel, 1998) and less than 10^{-1} W m^{-1} K^{-1} (Seiferlin et al., 1996), respectively. Depending on the thickness of the regolith, a few meters or tens of meters perhaps, and the heating rate, the temperature difference through the regolith can be up to some tens of kelvins and may, therefore, be relevant. Amorphous ice has been observed at the surface, but it is not likely to play a major role in the interior, at least in satellites that have experienced some level of endogenic or exogenic activity. Ammonia reduces the thermal conductivity of ice with increasing concentration by up to a factor of 3–4 at the peritectic composition of 32.1 wt% (Lorenz and Shandera, 2001).

Contamination of the ice by rock, gas bubbles, and dislocations may also alter the thermal conductivity. Studies of a sample of a carbonaceous chondrite (Allende C3) gave a thermal conductivity of about 2 to 3 W m^{-1} K^{-1} (Horai and Susaki, 1989). Thermal conductivities of rock in the temperature range of interest may be even an order of magnitude larger, but if the concentration of rock is well below 60 vol% this will be irrelevant. If we neglect the very small values of k of 10^{-1} W m^{-1} K^{-1} that should be applicable only to the ice regolith layer, then the thermal conductivity is uncertain by less than a factor of 5. The dimensionless heat flow parameter Θ will then be uncertain by the same factor of 5 while Nu_0 will be uncertain by a factor of 5^{β} or by about a factor of 1.5.

The values of the convective heat transfer parameters a and β in the Nu–Ra relation (10) depend on the boundary conditions, the planform of convection, the rheology of the convecting fluid, the Rayleigh number, and the Prandtl number $Pr = \nu/\kappa$. For infinite Prandtl number convection, as is applicable to solid state convection in ice and rock shells of planets and satellites, and for strongly temperature dependent viscosity and convection in the stagnant lid regime, it is believed that convection in the well-mixed sublayer below the lid can be parameterized by using the constant-viscosity parameterization (Schubert et al., 2001). Most of the data available to constrain the prefactor a and the exponent β have been determined for convection in plane layers where β is usually around 0.3 and a is between 0.1 and 0.3 (Schubert et al., 2001). Deschamps and Sotin (2000) have calculated numerical models of convection in fluids with strongly temperature dependent viscosity in the stagnant lid regime. Their model is two dimensional in plane layers and the fluid is heated from below. Deschamps and Sotin (2001) have used their results to derive a parameterization of convective heat transport and to model thermal convection in the outer shells of large icy satellites. Their numerical results and their parameterization are consistent with the general principles of stagnant lid convection. Their parameterization differs from the one used here but can be recast in terms of the present parameters. Using their data we find a value of 0.263 for β and a value of 0.79 for the prefactor a.

Ratcliff et al. (1996) numerically calculated steady-state convection in spherical shells with varying viscosity variation. The ratio between the inner and outer radii of the shell was 0.55 and boundaries were assumed to be impermeable, isothermal, and stress free. They found the heat transfer parameters to depend on the viscosity ratio across the fluid and on the planform of the convection. For constant viscosity and tetrahedral convection they found β equal to about 0.25. For cubic convection they obtained β equal to 0.26. Tetrahedral convection was found to be the dominant mode under these circumstances. These values are consistent with those of Bercovici et al. (1989), who obtained β equal to 0.26 and 0.28, for tetrahedral and cubic convection, respectively. For the prefactor a, Ratcliff et al. (1996) found a value of 0.39 for tetrahedral convection. The value of a decreases with increasing viscosity ratio to become 0.29 at the largest viscosity variation of 10^3 considered.

In this paper we take β equal to about 0.25 as a standard value but vary this value over a broad range. For a, we consider values between 0.1 and the Deschamps and Sotin (2001) value of 0.79.

The rheology of ice I and H_2O–NH_3 ice has been the subject of a number of studies (e.g., Goodman et al., 1981; Kirby et al., 1985; Kargel et al., 1991; Durham et al., 1992, 1993, 1997, 1998; Goldsby and Kohlstedt, 1997, 2001). The rheology of water ice is grain size, strain rate, and temperature dependent, but only weakly pressure dependent in the pressure range of interest. The pressure and temperature dependence of the rheology can be represented by the dependence of the rheology on the homologous temperature as in Eq. (5). The applicability of this equation has been demonstrated experimentally by Durham et al. (1997). They show that viscosity will decrease with increasing pressure approximately in parallel with the decrease of the melting temperature with increasing pressure, for pressures below the triple point pressure. The activation parameter A is usually taken to be between 18 and 36, which corresponds to an activation energy between about 40 and 80 kJ mole^{-1}. We consider that range of A in our calculations. As for ν_0, a generous range of values for ice I is between 10^9 and 10^{14} m^2 s^{-1}. Ammonia has been found to reduce the viscosity substantially, between 2 and 4 orders of magnitude (Durham et al., 1993). However, the solid ice shell that forms on top of an H_2O–NH_3 ocean will be composed of ice I up to the peritectic temperature and concentration.

The heat flow from below is taken to be primarily due to heat produced by radioactive decay and cooling. In addition, there may be heat released by the growth of an inner core in Ganymede. It is very likely that Ganymede has a substantial, self-generated magnetic field (Schubert et al., 1996). Tidal heating in the rock core or the ice shell may be applicable for Europa. Although we model convection in the ice shell as heated from below it is likely that the results of our calculations can also be applied to the ice shell of Europa if it is heated by tidal dissipation from within.

The concentration of radiogenic elements in the Galilean satellites is basically unknown. In modeling radiogenic heating of planets and satellites of unknown composition, the chondritic composition is often used as a reference. Using compositions listed in Lodders and Fegley (1998), and correcting for the nonrock components, we find a typical value of about 4 pW kg^{-1}. CV and CO type chondrites have higher values of specific heat production, 4.5 and 5 pW kg^{-1}, respectively. Heat production in the Earth and Moon has been estimated to be approximately 4 pW kg^{-1}. In the present paper, we use 4.5 $\pm$ 0.5 pW kg^{-1} as a reference value. This value translates into 1.74×10^{11} W for Europa, 3.36×10^{11} W for Ganymede, and 2.42×10^{11} W for Callisto. These values have been used as reference values in Figs. 3 and 6.

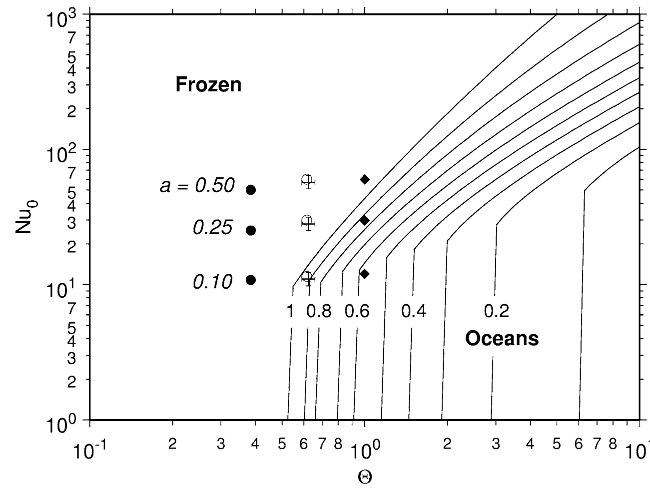

Fig. 3. Results of the model calculations for H_2O ice for $\beta = 0.25$. Shown are lines of constant lid thickness divided by the depth to the triple point in the Nu_0 vs Θ-plane. The ice shells are completely frozen for fractional lid thicknesses of 1. Models with three different values of the convective heat transfer parameter a are marked in the figure. Diamonds refer to models of Europa, crosses to models of Ganymede, open circles to models of a (partially) differentiated Callisto, and solid circles to models of an undifferentiated Callisto. The kinks in the isolines indicate the transition from conductive heat transfer in the ice shell for which the shell thickness is independent of Nu_0 to convective heat transfer. Note that Nu_0 is a scaling Nusselt number and is not the Nusselt number of the convecting sublayer.

Results

The results of the calculations are shown in Figs. 3–7. Figs. 3 to 5 show results for H_2O ice while Fig. 6 shows results for NH_3–H_2O ice. Plotted in Fig. 3 are lines of constant ice shell thickness scaled by the depth l_0 to the triple point in a plane defined by the scaled heat production rate Θ and the scaling Nusselt number Nu_0. To be exact, the lines of constant ice shell thickness are strictly valid only for Ganymede. However, the error introduced by applying Fig. 3 to Europa and Callisto is small, only a few percent in shell thickness values. The diamonds in Fig. 3 refer to standard models of Europa, the crosses to models of Ganymede, the open circles to (partially) differentiated models of Callisto, and the solid circles to undifferentiated models of Callisto, respectively, with the following parameter values: chondritic Q, $k = 3.3$ W m^{-1} K^{-1}, $A = 24$, $\nu_0 = 10^{11}$ m^2 s^{-1}, and $\beta = 0.25$. The standard models differ mostly in Θ among the satellites. Since the same specific heat production rate of 4.5 pW kg^{-1} has been used, a larger value of Θ reflects a larger concentration of silicates. This is why Europa models plot around $\Theta = 1$, while Ganymede and the (partially) differentiated Callisto models have smaller, albeit similar, values of Θ. The value of Θ is also proportional to the depth to the triple point pressure l_0 and is therefore inversely proportional to the pressure gradient. Since the pressure gradient in an undifferentiated Callisto is steeper than in a differentiated Callisto, Θ is still smaller for the former model.

The lines of constant ice shell thickness are independent of Nu_0 for values of Nu_0 smaller than 10 for $D/l_0 = 1$ and smaller than about 50 for $D/l_0 = 0.1$. Heat is transferred by conduction when the ice shell thickness is independent of Nu_0, and the ice shell is then stable against thermal convection. For larger values of Nu_0, the ice shell is unstable to convection.

The undifferentiated models of Callisto are all frozen for the parameter values chosen in Fig. 3. For (partially) differentiated models of this satellite and for Ganymede and Europa the presence of an ocean and its thickness and the thickness of the ice shell will, with all other parameters fixed, depend on the heat transfer parameter a. For values of a larger than about 0.1, Ganymede and Callisto are predicted to be frozen; for Europa a value of a larger than about 0.4 is required.

Fig. 4 shows the results of variations of β and of the *Nu–Ra* prefactor a for Europa and Ganymede. The Ganymede model is also applicable to a (partially) differentiated model of Callisto. Other parameters have their standard value; in particular, Θ is 0.998 for Europa and 0.623 for Ganymede as in Fig. 3. Increasing β at constant values of a will tend to freeze the ocean. The same is true if a is increased at constant values of β. The smallest ice shell thickness for Europa is 0.6 because conduction with the standard thermal conductivity value will result in an ice shell thickness of 0.58. For Ganymede the latter value is 0.8.

Fig. 5 shows the results of variations of ν_0 and A, again for Europa and Ganymede. As in Fig. 4, the Ganymede model is also applicable to a (partially) differentiated Callisto. Parameter values other than A and ν_0 are standard values. Ice shell thicknesses are inversely proportional to both A and ν_0.

Fig. 6 shows the results for NH_3–H_2O ice and $\beta = 0.25$. The lines of constant ice shell thickness vary little with Nu_0 for values of Θ below about 1. All models have oceans in this case, with Europa having the thinnest ice shell and undifferentiated Callisto having the thickest shell. Completely frozen satellites would require a value of Θ of about 0.2.

Fig. 7a and 7b show the thickness of the oceans as a function of the thickness of the ice shell for models of Ganymede and Callisto. For Europa it is likely that the depth to the bottom of the ice + water layer and the top of the rock shell coincides nearly with the depth to the triple point. Thus the thickness of the ocean for Europa is simply calculated by subtracting the ice shell thickness from the depth to the rock shell. To calculate the thickness of the water ocean for Ganymede and Callisto we have used the water liquidus up to a pressure of 400 MPa in the parameterization of Chizhov (1993). The kink in the thickness curves in Fig. 7a indicates the transition from ice III to ice V. The thickness of the water + ammonia ocean will depend on the path of the system in the three-dimensional phase diagram. For simplicity, we have linearly interpolated between the melting temperature of ice I + 5 wt% NH_3 at $P = 0$ MPa (271.6 K) and the peritectic temperature of 176 K at $P = 170$ MPa, along the peritectic to 270 Pa and 178 K, and between the latter temperature and the melting temperature of ice VI + 5 wt% NH_3 at $P = 820$ MPa, thereby following Grasset and Sotin (1996) for a possible path in the phase diagram. The kinks in the thickness curves here in-

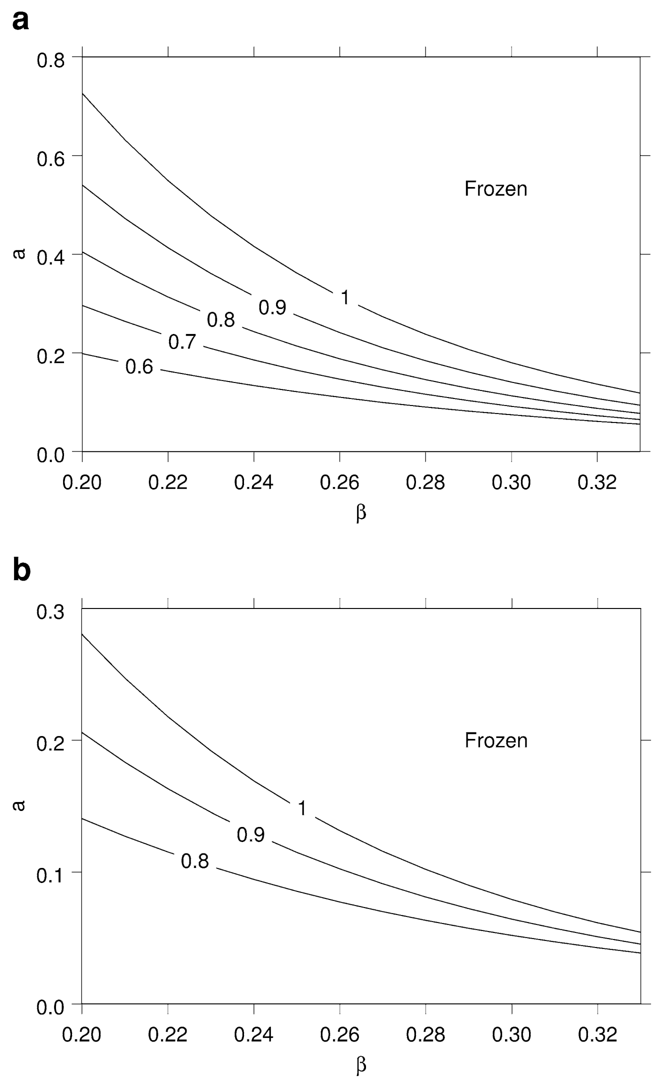

Fig. 4. Results of variations of α and β for (a) Europa and (b) Ganymede. Other parameter values are as listed in Table 1. The Ganymede model is also applicable to a (partially) differentiated model of Callisto.

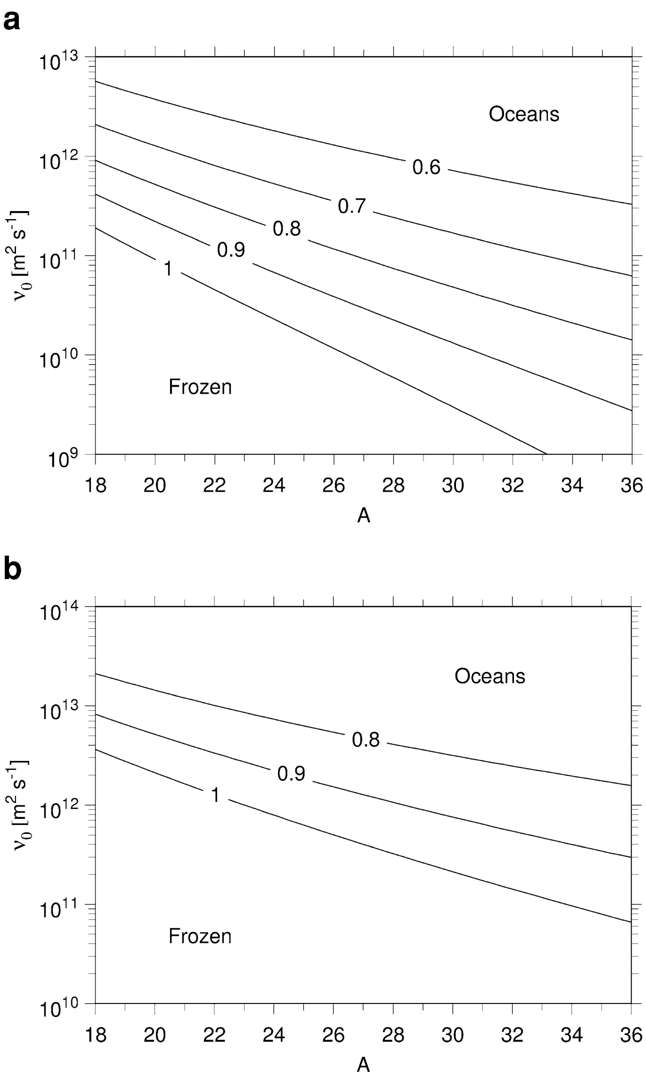

Fig. 5. Results of variations of ν_0 and A for (a) Europa and (b) Ganymede. Other parameter values are as listed in Table 1. The Ganymede model is also applicable to a (partially) differentiated model of Callisto.

dicate that the peritectic has been reached at the high-pressure flank of the melting curve.

For the nominal models, H_2O oceans are 60- to 70-km thick on Europa if $\nu_0 = 10$, depending on the depth to the bottom of the shell being 150 or 160 km, respectively, and 35- to 45-km thick if $\nu_0 = 30$. With the former value of ν_0, the oceans are about 50-km thick on Ganymede and on the (partially) differentiated Callisto model. H_2O–NH_3 oceans are 85- to 95-km thick on Europa, about 230-km thick on Ganymede, almost 200-km thick for a Callisto with an undifferentiated ice shell, and almost 300-km thick for a (partially) differentiated Callisto model.

Discussion and conclusions

In the following we will defend a main conclusion that we draw from our model calculations: While an ocean can hardly be doubted for Europa, oceans are possible for Ganymede and a (partially) differentiated Callisto because we find that equilibrium states plot close to the transition line between oceans and complete freezing even for pure H_2O. Pure H_2O models are extreme because they have the highest melting temperature. Equilibrium models are extreme because satellite cooling should contribute to the energy balance and increase the ice shell bottom heat flow above the values derived from radiogenic heating alone. We also argue that the most important factor acting against oceans, aside from uncertainty in convection heat transfer efficiency, is rock left in the outer ice shell during interior differentiation or accumulated after differentiation. Rock increases the pressure gradient and moves the triple point of the phase diagram to shallower and colder depths.

Our results are clear-cut for water + ammonia. Oceans are predicted for all three satellites and the conclusion is robust against parameter variations. The water ice shells on top of the H_2O– NH_3 oceans are similar in thickness (70–90 km) and the shells, because of the steep gradient in homologous temperature and its associated rapid increase in viscosity with proximity to the surface, have thick stagnant lids or are entirely nonconvective. A variation in rheological and heat transfer parameter values may result in more efficient convection in the ice shell but, most likely, the increased heat transfer will not suffice to freeze the ocean. Since the lines of constant ice shell thickness in Fig. 6 are almost independent of Nu_0 an ocean is almost impossible to avoid. The thicknesses of these oceans are substantial. For Europa, the ocean thickness is limited by the thickness of the water shell (about 150 km), leaving only about 80 km for the ocean. This limitation does not apply to Ganymede and Callisto, whose bulk densities and moment of inertia factors indicate a thick water shell (e.g., Anderson et al., 2001; Sohl et al., 2002). The thickness of the H_2O–NH_3 oceans may well be 200 to 350 km. For Callisto this would allow a structure of an ice shell with or without rock in it overlying

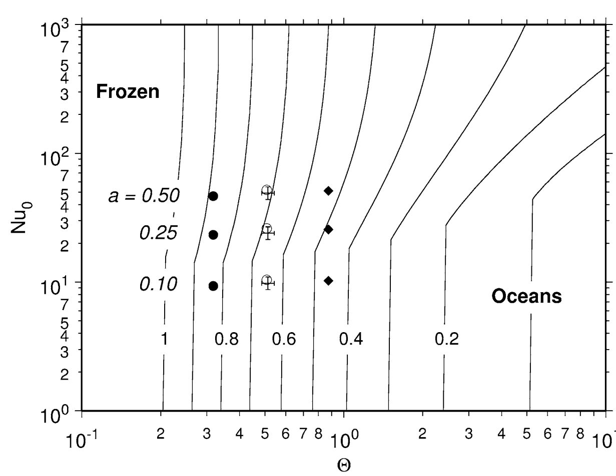

Fig. 6. Results of the model calculations for H_2O–NH_3 ice for $\beta = 0.25$. Shown are lines of constant lid thickness divided by the depth to the triple point in the Nu_0 vs Θ-plane. For further explanation see Fig. 3.

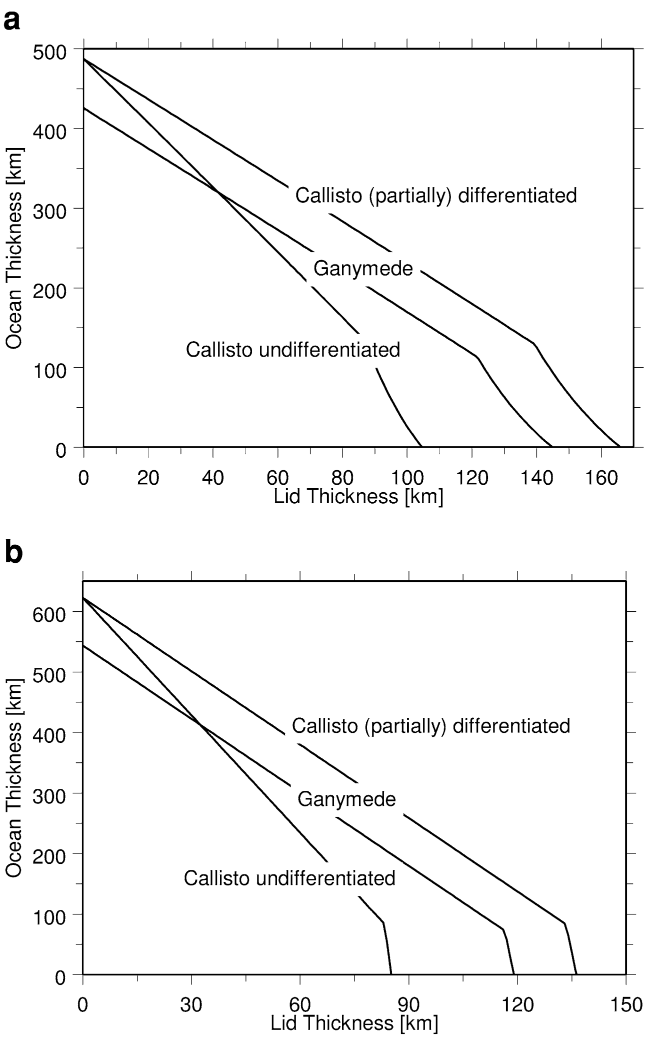

Fig. 7. Ocean thickness as a function of ice shell thickness for H_2O ice (a) and H_2O–NH_3 ice (b). The thickness values in panel a have been calculated from the known phase diagram of ice. The values in panel b have been calculated by linearly interpolating between the melting temperature of ice I + 5 wt% NH_3 at $P = 0$ MPa and the peritectic temperature at $P = 170$ MPa, along the peritectic up to a pressure of 270 MPa, and between the peritectic temperature and the melting temperature of ice VI + 5 wt% NH_3 at $P = 820$ MPa.

an ocean that could extend to close to the surface of the ice–rock core. The ice shells would be thinner and the oceans even thicker if the bulk concentration of ammonia were greater. Unfortunately, we cannot constrain the concentration of ammonia in the Galilean satellites.

The case for oceans in Ganymede and Callisto is less clear if the ice is pure or if the melting point depression is small. In these cases the differences among the satellites will matter as well as choices of uncertain parameter values. For Europa, however, an ocean is very likely even under these circumstances. An ocean is more likely for smaller pressure gradients ρg in a satellite and for greater heat flows per unit area. This is why Europa, which has the greatest heat flow per unit area and the smallest pressure gradient (see Table 1) of the satellites, is the most likely to have an ocean. In the parameter range considered by us, Europa may have an ocean even if the water ice were pure. This conclusion will be further strengthened if tidal heating is considered. Hussmann et al. (2002) have calculated tidal heating rates in Europa's ice shell and have searched for equilibrium between the tidal heating and the heat transfer rates, both of which are functions of the ice shell thickness and the ice rheology. For values of the rheology parameters similar to ours, they find that the equilibrium value of the tidal heating rate when expressed as a surface heat flow is about three times the heat flow due to radiogenic heating and the ice shell thickness is 20–40 km. This much thinner ice shell for Europa is probably the reason why there is considerably more geologic evidence for an ocean on Europa than there is for oceans on Ganymede and Callisto.

Our results are also clear-cut and robust for an undifferentiated H_2O-ice Callisto. An ocean is unlikely, in that case, mostly because the rapid increase in pressure caused by the assumed density of 1600 kg m^{-3} moves the triple point to a shallow depth of around 100 km. An ice shell of this thickness will be capable of removing the heat due to radiogenic decay in the deeper interior by conduction. Of course, it is unlikely that Callisto is undifferentiated because its moment of inertia factor is 0.355 (Anderson et al., 2001), smaller than the value of 0.38 for a homogeneous Callisto with density increases through ice phase transformations (McKinnon, 1997). Still, the value of the "observed" moment of inertia factor has been calculated under the assumption that Callisto is in hydrostatic equilibrium, which may not be true. In any case, it is possible, if not likely, that the outer shell of Callisto carries some rock if the satellite differentiated gradually by the slow inward motion of the rock component. Since the outer shell is cold and rigid, the separation of rock and ice will be very slow there. The effect on the depth to the triple point and the presence of an ocean will be the same as in the case of a completely undifferentiated satellite. However, our assumption of a more or less primordial concentration of rock in the outer shell may be too extreme. Any reduction in the rock concentration will make an ocean more likely.

The cases of a (partially) differentiated Callisto and of Ganymede, for which the density in the ice shells is only around 1000 kg m^{-3}, represent the other extreme. While an ocean is marginal for our nominal parameter values in this case, there are ways of increasing the likelihood of a more substantial ocean even if ammonia is not present. Other components, e.g., salts such as $MgSO_4$ and $NaSO_4$ (Kargel, 1998), could reduce the melting temperature. Of these and other possible candidates, only $MgSO_4$–H_2O has been studied experimentally in sufficient detail at elevated pressures, but at low pressures all the major salt-forming systems are well understood. The effect of $MgSO_4$ on the melting temperature of ice seems to be much smaller than that of ammonia, resulting in a reduction of only several kelvins (Hogenboom et al., 1995), but chloride salts are known to have a more potent effect on the melting point than $MgSO_4$.

Other parameters that may increase the likelihood of an ocean are the thermal conductivity and the heating rate. The nominal value of the thermal conductivity in our study may be an overestimate if there are substantial ice regoliths on the satellites. The temperature increase through a regolith a few tens of meters thick could be a few tens of kelvins, as discussed earlier. This increase relative to the values considered in our model will act similar to a reduction of the melting temperature by approximately the same value, or similar to the effect of salt on the melting temperature. The assumed chondritic heat flow from below could, on the other hand, represent an underestimate of the actual heat flow from the deep interior of the satellites. Thermal history calculations for terrestrial planets and satellites (see Schubert et al. (1986) for a review of the evolution of satellites) have shown that cooling will contribute substantially to the heat flow from the interior. The ratio between the surface heat flow and the rate of radiogenic heating (if expressed as a heat flow) could be a factor of 2 for the Earth and the Moon. The value for the Moon, a body closer in size to the Galilean satellites than the Earth, has been estimated from the numerical results of Konrad and Spohn (1997) and Spohn et al. (2001). If this is applicable to the icy Galilean satellites then there is no problem explaining oceans on Ganymede and a (partially) differentiated Callisto, as our calculations suggest. Other uncertainties concern the rheology and the convective heat transfer parameters. For instance, if the rheology of the ice is stiffer than considered here either through a larger activation parameter or through a large melting point viscosity then the likelihood of oceans will increase. Ruiz (2001) has recently considered stress-dependent rheologies for Callisto's (differentiated) ice shell. He finds that the shell is stable against convection for both grain boundary sliding with a quadratic dependence of strain rate on stress and dislocation creep with a dependence on stress to the power of 4. His results should at least qualitatively be applicable for Ganymede as well. Applying Fig. 3 under the assumption that the ice shell is stable against convection gives Nu_0 smaller than 10. An ocean is then predicted for both Ganymede and the (partially) differentiated Callisto underneath an ice shell of a little more than 100-km thickness.

Uncertainties in parameter values that work against oceans can also be listed. For instance, if we overestimated the specific chondritic heating rate of 4.5 pW kg^{-1} and the rheology parameters then a correction will tend to make an ocean less likely. Moreover, we have used values of β and a of 0.25 each. If β should be closer to 0.3, as suggested by boundary layer theory and by some numerical calculations, then a must be close to 0.1 to prevent freezing. The Deschamps and Sotin (2001) parameterization of convective heat transfer has $\beta = 0.26$ as in our models, but their value of a would be as large as 0.79. Their model applied to Ganymede and Callisto predicts frozen ice shells. However, for Europa the ice shell is just frozen. This is due to a larger value of chondritic heat production used by these authors and underlines the uncertainty in some of the model parameters.

The fact that Ganymede's deep interior is differentiated and Callisto's most likely is not does not affect our results in any major way. It is possible that the outer ice shell of Ganymede has also kept some rock during its differentiation that will, as previously discussed, work against an ocean. On the other hand, Ganymede may have experienced tidal heating in its past and may still be losing some of the heat deposited during this phase.

Previous discussions, as, e.g., reviewed by Schubert et al. (1986), have predicted that the ice shells of the Galilean satellites would be frozen solid because subsolidus convection would easily remove the heat generated by radioactive decay in the interior. These conclusions were based on models of convective heat transfer derived from a parameterization of constant-viscosity convection. The parameterization of heat transfer by convection in fluids with strongly temperature dependent viscosity developed in recent years does not predict much less total heat transfer. The difference is mostly in the distribution of the heat in the interior. The variable-viscosity parameterization removes heat mostly by thickening a stagnant lid on top of the convecting deeper interior. Thermal history calculations for the Moon using two- and three-dimensional convection models with the viscosity depending on the laterally averaged temperature profile show the same characteristics (Spohn et al., 2001). The constant-viscosity parameterization removes heat mostly from the deep interior. It is the lack of cooling of the deep interior that effectively results in temperatures above the melting temperature in our model. Because the stagnant lid is so important, care must be taken when calculating its thickness. We have allowed the melting temperature (liquidus temperature for NH_3–H_2O) and the homologous temperature to be functions of depth in Eqs. (6) to (15), where the thickness of the stagnant lid and the ice shell are calculated. In this respect our model differs from the previous models of Grasset et al. (2000) and Deschamps and Sotin (2001). Accounting for the decrease of the melting (liquidus) temperature with depth results in smaller ice shell thicknesses D compared with models in which the variation of the melting (liquidus) temperature in the convecting layer is neglected.

Opposite conclusions are reached about the existence of oceans in the Galilean satellites depending on whether thick stagnant lids form on top of convecting ice shells as predicted by the previously referenced experimental and theoretical studies of convection in fluids with strongly temperature dependent viscosity. We have assumed the latter in this paper, but the applicability of these results to the Galilean satellites is not certain. Because of plate tectonics, stagnant-lid convection is not relevant to the Earth. If heat transfer processes like ice shell delamination or water or soft ice "magmatism" occur on the Galilean satellites, thick stagnant lids may not form. Such processes are perhaps not likely for Ganymede and Callisto because of their thick ice shells, but they could be relevant for Europa. On the basis of heat balance considerations, however, Europa is the satellite for which an ocean is most likely.

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