Equilibrium Convection on a Tidally Heated and Stressed Icy Shell of Europa for a Composite Water Ice Rheology

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Abstract Water ice I rheology is a key factor for understanding the thermal and mechanical state of the outer shell of the icy satellites. Ice flow involves several deformation mechanisms (both Newtonian and non-Newtonian), which contribute to different extents depending on the temperature, grain size, and applied stress. In this work I analyze tidally heated and stressed equilibrium convection in the ice shell of Europa by considering a composite viscosity law which includes diffusion creep, basal slip, grain boundary sliding and dislocation creep, and. The calculations take into account the effect of tidal stresses on ice flow and use grain sizes between 0.1 and 100 mm. An Arrhenius-type relation (useful for parameterized convective models) is found then by fitting the calculated viscosity between 170 and 273 K to an exponential regression, which can be expressed in terms of pre-exponential constant and effective activation energy. I obtain convective heat flows between $\sim$40 and $\sim$60 mW m^{-2}, values lower than those usually deduced ($\sim$100 mW m^{-2}) from geological indicators of lithospheric thermal state, probably indicating heterogeneous tidal heating. On the other hand, for grain sizes larger than $\sim$0.3 mm the thicknesses of the ice shell and convective sublayer are $\sim$20–30 km and $\sim$5–20 km respectively, values in good agreement with the available information for Europa. So, some fundamental geophysical characteristics of the ice shell of Europa could be arising from the properties of the composite water ice rheology.

Keywords Europa · Satellites of Jupiter · Thermal convection · Ice rheology · Heat flow · Ice shell thickness

1 Introduction

Ice rheology is a central issue for research the thermal and mechanical properties, as well as the evolution, of the icy satellites. Concretely, water ice I viscosity is decisive for the possible existence and character of convection in the outer layer of these satellites, and hence for the possible maintenance of internal oceans (e.g., Spohn and Schubert 2002; Barr and Pappalardo 2005; Freeman et al. 2006; McKinnon 2006).

Ice flow is a complex phenomenon involving several deformation mechanisms (e.g., Duval et al. 1983; Weertman 1983; Budd and Jacka 1989; Durham and Stern 2001; Goldsby and Kohlstedt 2001), both Newtonian (such as volume and grain boundary diffusion creep) and non-Newtonian (dislocation creep, grain boundary sliding and basal slip), which contribute to different extents depending on the temperature, grain size, and applied stress. By taking together these diverse mechanism, Goldsby and Kohlstedt (2001) have proposed a composite flow law, which should be capable of describe the flow of water ice for a wide variety of situations.

Convection in the ice shell of Europa has been investigated by numerous works, which include parameterized (e.g., Pappalardo et al. 1998; McKinnon 1999; Hussmann et al. 2002; Nimmo and Manga 2002; Ruiz and Tejero 2003; López et al. 2003; Ruiz et al. 2007) and numerical (e.g., Sotin et al. 2002; Tobie et al. 2003; Showman and Han 2004, 2005; Mitri and Showman 2005) treatments. These works, as a whole, cover a wide variety of rheological parameters, creep mechanisms, and geodynamic settings and models.

The only previous work which has analyzed convection on Europa for a composite water ice rheology was performed by Moore (2006). This author used the parameters proposed for low temperature diffusion creep, grain boundary sliding, basal slip and dislocation creep, but did not take into account the effect of tidal stresses on non-Newtonian flow mechanism. However, tidal stresses are dominant for Europa (McKinnon 1999), and therefore they should have an important influence on non-Newtonian mechanisms, and hence on the total (composite) viscosity, of the water ice in Europa. In this work I analyze thermal equilibrium convection in the ice shell of Europa by considering a composite viscosity law for water ice I, but considering the effect of tidal stresses on the viscosity of non-Newtonian creep mechanisms (see McKinnon 1999).

2 Convective Model

Convection in the outer shell of icy satellites operates in the stagnant lid regime (e.g., McKinnon 1998; Freeman et al. 2006), in which a cold and essentially immobile lid develops above the actively convective sublayer. Although the viscosity contrast across the entire ice shell can be very large, the viscosity variation within the convective sublayer is typically of one order of magnitude [see Grasset and Parmentier (1998) and references therein]. Grasset and Parmentier (1998) have shown that convective parameterization laws derived for constant viscosity convection are applicable if the boundary conditions are properly defined. In fact, this procedure has been previously used for icy satellites (e.g., Hussmann et al. 2002, 2006; Ruiz and Tejero 2003; Multhaup and Spohn 2007; Ruiz et al. 2007). This is useful for Europa, because tidal heating is strongly temperature-depedent (Ojakangas and Stevenson 1989), and it is largely restricted to the warmest ice. In these conditions, tidal heating is negligible in the stagnant lid, which can be treated separately (Hussmann et al. 2002; Ruiz and Tejero 2003). Otherwise, parameterizations for internally heated stagnant lid convection considering the same heating rate in both the stagnant lid and the convective sublayer (which could be more useful for the radioactively heated mantles of the terrestrial planets) are not a good analogous for the case of the ice shell of Europa, where tidal strain rates in the stagnant lid would be strongly overestimated.

Thus, here I consider a steady-state convective layer heated from within, which satisfies the relations found by Grasset and Parmentier (1998). So, the relation between the dimensionless temperature ratio and the Rayleigh number is given by

$$\frac{k(T_i - T_t)}{Hb_c^2} = a\mathrm{Ra}_{\mathrm{H}}^{\beta}, \tag{1}$$

where k is the thermal conductivity, T_i is the temperature of the well-mixed convective interior, T_t is the temperature of the top of the convective layer, H is the volumetric heating rate, b_c is the thickness of the actively convective layer, a and β are constants (determined from numerical investigation as 2.383 and -0.227 respectively; Grasset and Parmentier 1998), and $\mathrm{Ra_H}$ is the Rayleigh number defined for an internally heated layer,

$$\mathrm{Ra_H} = \frac{\alpha\rho g H b_c^5}{k\kappa\eta_i}, \tag{2}$$

where, in turn, α is the thermal expansion coefficient, ρ is the density, g is the gravity (1.31 m s^{-2} for Europa), κ is the heat diffusion coefficient, and η_i is the effective viscosity calculated for $T = T_i$. Several terms of Eqs. 1 and 2 are functions of temperature: $k = k_0 T^{-1}$, $\alpha = \alpha_0 T$, and $\kappa = \kappa_0 T^{-2}$, where the constants are $k_0 = 621$ W m^{-1} (Petrenko and Whitworth 1999), $\alpha_0 = 6.24 \times 10^{-7}$ K^{-2} and $\kappa_0 = 9.1875 \times 10^{-2}$ m^2 K^2 s^{-2} (Kirk and Stevenson 1987); these functions are calculated for $T = T_i$, since most of the convective layer is nearly isothermal. Also, ice density varies slightly with temperature and pressure (e.g., Lupo and Lewis 1979), yet adopting a constant value does not alter the results significantly; here this value is taken as 930 kg m^{-3}. The convective heat flow can be obtained by combining Eqs. 1 and 2

$$F_c = Hb_c = \left[\frac{k(T_i - T_t)H^{1+4\beta}}{a}\left(\frac{k\kappa\eta_i}{\alpha\rho g}\right)\right]^{1/(2+5\beta)}. \tag{3}$$

In order to use isoviscous convection equations for the convective sublayer, I use (Grasset and Parmentier 1998)

$$T_t = T_i - 2.23\frac{RT_i^2}{Q}. \tag{4}$$

for adequately define the temperature at the top of the convective layer,

Equations (1–3) work if the convective layer is only heated from within, and so a lower boundary layer does not exist, but the Europa's ice shell must also be heated from below by tidal and radioactive heating in the rocky core [the radiogenic contribution to the surface heat flow is ~6–8 mW m^{-2} (Cassen et al. 1982; Spohn and Schubert 2002)]. However, the method here described can be used as an approximation, since the surface heat flow of Europa must be mostly generated in the warm ice of the convective sublayer (Ruiz 2005). So, here I take $T_i = T_b$, where T_b is the temperature at the shell base, which in turn is given by the water ice melting point, as there are solid evidences for an internal ocean on Europa (e.g., Kivelson et al. 2000), which depends on pressure P as (Chizhov 1993)

$$T_b = 273.16\left(1 - \frac{P(\mathrm{MPa})}{395.2}\right)^{1/9}, \tag{5}$$

the pressure at the ice shell base is given by $\rho g b$, where b is the total shell thickness. The stagnant lid, which is thermally conductive and heated from below, contributes to the total ice shell thickness. For a temperature-dependent thermal conductivity the thickness of the stagnant lid is

$$b_{sl} = \frac{k_0}{F_c}\ln\left(\frac{T_t}{T_s}\right), \tag{6}$$

where T_s is the surface temperature, taken as 100 K, a value considered as representative of the mean temperature at Europa's surface (e.g., Ojakangas and Stevenson 1989).

Finally, to calculate tidal heating rates, I assume that under tidal stresses ice behave like a viscoelastic (Maxwell) material: thus, the tidal volumetric dissipation rate can be calculated according to (Ojakangas and Stevenson 1989)

$$H = \frac{2\eta\dot{\varepsilon}^2\mu^2}{\mu^2 + \omega^2\eta^2}, \tag{7}$$

where $\dot{\varepsilon}$ is the strain rate, here taken as 2×10^{-10} s^{-1}, value considered as representative for the average strain rate on the icy shell of Europa (Ojakangas and Stevenson 1989; McKinnon 1999), $\mu = 4 \times 10^9$ Pa is the ice rigidity, and ω is the frequency of the forcing, which can be equated with Europa's mean motion, 2.05×10^{-5} rad s^{-1}. I take $H = H_i$, which somewhat overestimates tidal dissipation within the upper boundary layer (where temperatures are lower than T_i).
temperatures are higher than T_i)
total tidal heating. I thus consider our tidal dissipation rate calculations to be representative for the convective layer.

I calculate ice shell structure and heat flow by simultaneously solving Eqs. 3–7. The total ice shell thickness is taken as $b = b_{sl} + b_c$ in Eq. 5.

3 Composite Water Ice Viscosity

The composite flow law for pure water ice I proposed by Goldsby and Kohlstedt (2001) is

$$\dot{\varepsilon} = \dot{\varepsilon}_{diff} + \left(\dot{\varepsilon}_{basal}^{-1} + \dot{\varepsilon}_{gbs}^{-1}\right)^{-1} + \dot{\varepsilon}_{disl}, \tag{8}$$

where $\dot{\varepsilon}$ is the total strain rate, and $\dot{\varepsilon}_{diff}$, $\dot{\varepsilon}_{basal}$, $\dot{\varepsilon}_{gbs}$ and $\dot{\varepsilon}_{disl}$ are the strain rates due, respectively, to diffusion creep, basal slip, grain boundary sliding and dislocation creep. The flow law for each individual mechanism is given by a relation of the form

$$\dot{\varepsilon} = \frac{A\sigma^n}{d^p}\exp\left(-\frac{Q}{nRT}\right), \tag{9}$$

where A is the pre-exponential coefficient, σ is the applied stress, n is the stress exponent, d is the grain size, p is the grain size exponent, Q is the activation energy, $R = 8.31447$ J mol^{-1} K^{-1} is the gas constant, and T is the absolute temperature. For basal slip, grain boundary sliding and dislocation creep, A is a constant depending on the creep mechanism; for diffusion creep, $A_{diff} = 14V_m\, D_v/(RT)$, volume and D_v is the pre-exponential volume diffusion coefficient. Diffusion creep has not been experimentally observed in water ice, although Goldsby and Kohlstedt (2001) proposed a theoretical flow law for this deformation mechanism based on the values of the constants involved.

Based upon Eq. 8 a composite viscosity law for water ice can be written as (Freeman et al. 2006)

$$\eta = \frac{1}{\eta_{diff}^{-1} + (\eta_{basal} + \eta_{gbs})^{-1} + \eta_{disl}^{-1}}, \tag{10}$$

where the subindexes have the same meaning as in Eq. 8. For diffusion creep, the viscosity is given by

$$\eta = \frac{d^2}{2A_{diff}} \exp\left(\frac{Q}{RT}\right), \tag{11}$$

For non-Newtonian mechanisms the viscosity is given by $\eta = \sigma/(3\dot{\varepsilon})$, where the factor 3 in the denominator is related to the axisymetric and divergent nature of the flow (e.g., Durham and Stern 2001). In the ice shell of Europa convective stresses are significantly less than fluctuating tidal stresses. McKinnon (1999) has argued that in these conditions an average effective viscosity (which can be treated as Newtonian), can be calculated for non-Newtonian mechanisms from

$$\eta_{\text{eff}} = \eta\, n^{-1/2} = \sigma_{tidal}/\left(3\dot{\varepsilon} n^{1/2}\right), \tag{12}$$

where σ_{tidal} is tidal strain.

In order to find an Arrhenius-type relation, which is useful for the parameterized convective model described in Sect. 2, I solve Eq. 10 between 170 and 273 K [the temperature range in the experiments performed by Goldsby and Kohlstedt (2001)], for creep parameters shown in Table 1, and then a exponential regression is found. Calculation of tidal heat dissipation rates in Sect. 2 assumes a Maxwell behavior for water ice. The strain rate for a Maxwell material is the sum of a viscous and an elastic term. The contribution of the viscous term decreases for lowering temperatures, whereas the elastic term increases. Here I assume that the ice behave totally viscous for $T = 273$ K. Thus, Eqs. 8 and 10 are solved for tidal stresses satisfying $\dot{\varepsilon} = 2 \times 10^{-10} \text{s}^{-1}$ for $T = 273$ K. As diffusion creep and grain boundary sliding are sensitive to the grain size, a range between 0.1 and 100 mm is used for this parameter. Based on polar glacial ice observations, a grain size smaller than 0.1 mm is unlikely (McKinnon 1999), at least if there are no impurities limiting crystal growth. On the other hand, ice grains in equilibrium with dynamic recrystallization due to convective stresses could approach sizes of 50–100 mm (Barr and McKinnon 2006, 2007;

Table 1 Flow laws parameters for water ice (Goldsby and Kohlstedt 2001)

	A ($\text{MPa}^{-n}\ \text{m}^p\ \text{s}^{-1}$)	n	p	Q (kJ mol^{-1})
Diffusion creep[a]	$3.02 \times 10^{-8}/T$	1	2	59.4
Basal slip	5.5×10^{7}	2.4	0	60
Grain boundary sliding	3.9×10^{-3}	1.8	1.4	49
Dislocation creep B	4.0×10^{5}	4	0	60

[a] Flow law for diffusion creep as adjusted by Barr and Pappalardo (2005)

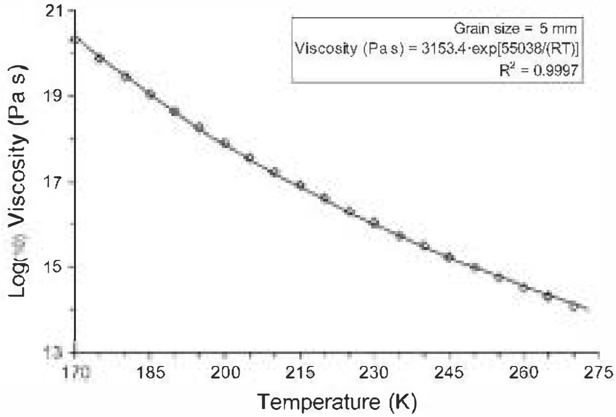

Fig. 1 Example of Arrhenius equation obtained from an exponential regression for a grain size of 5 mm. *Open diamonds* shown viscosities obtained from (5) between 170 and 273 K, drawn each 5° intervals in order to improve clarity. *Thick solid line* shows viscosities obtained from the Arrhenius equation

Tobie et al. 2006), although tidal stresses could further limit crystal growth (maybe to grain sizes close to ~4 mm; Barr and McKinnon 2006). Grain size values must be taken as averages, since this parameter could be very heterogeneous both vertically and horizontally (Tobie et al. 2006; Barr and McKinnon 2007).

The obtained exponential regression can be expressed in terms of a pre-exponential constant B_0, and an effective activation energy Q_{eff}, such as

$$\eta = B_0 \exp\left(\frac{Q_{eff}}{RT}\right). \tag{13}$$

Figure 1 shown an example for $d = 5$ mm. The R^2 coefficient for the so-obtained exponential regressions is above 0.99 in all the cases explored. Stresses giving $\dot{\varepsilon} = 2 \times 10^{-10}\,\mathrm{s}^{-1}$ for $T = 273$ K are in general between ~10^4 and ~10^5 Pa, in agreement with theoretical calculations of tidal stresses in the ice shell of Europa (e.g., Greenberg et al. 1998; Harada and Kurita 2006).

Figure 2 shows the value of B_0 and Q_{eff} as functions of grain size. The curves for B_0 and Q_{eff} have, respectively, a maximum and minimum for grain size close to 1 mm, due to the dominant role of grain boundary sliding for this grain size. The relative importance of diffusion creep and dislocation creep increases for grain sizes lesser and greater, respectively, than ~1 mm. The implications of the properties of the composite viscosity law for convection in the Europa's ice shell are analyzed in Sect. 4.

4 Results and Discussion

Figure 3 shows that the convective heat flow varies between ~60 and ~40 mW m^{-2} for grain sizes from 0.1 to 100 mm. There is a transition from higher to lower heat flows, which roughly coincides with the maximum and the minimum obtained, respectively, for B_0 for Q_{eff}. These heat flows are lower than the values of ~100 mW m^{-2} usually deduced from diverse geological features used as temperature-in-depth indicators (Ruiz and Tejero 1999, 2000; Pappalardo et al. 1999; Ruiz 2005; Dombard and McKinnon 2006;

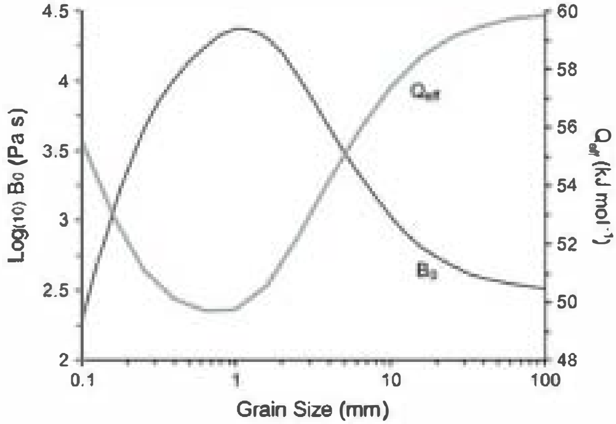

Fig. 2 Values of the pre-exponential coefficient and effective activation energy for the composite viscosity law for water ice, shown as a function of grain size

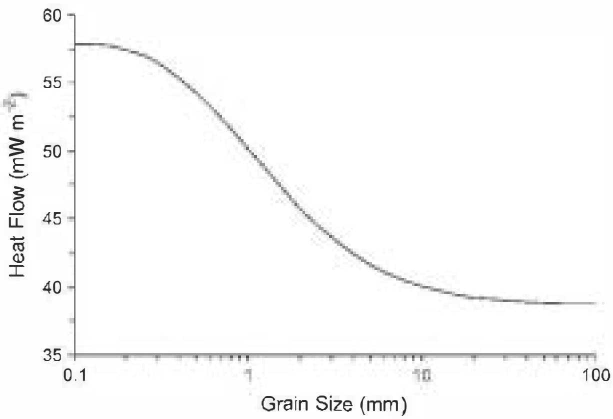

Fig. 3 Convective heat flow shows as a function of grain size

Lichtenberg et al. 2006). Previously, some works found high convective heat flows for Europa, but only for individual grain size-dependent rheologies and grain sizes lower than 1 mm (Nimmo and Manga 2002, Ruiz and Tejero 2003; Ruiz et al. 2007). An explanation of this discrepancy may be that high heat flows correspond to local settings with enhanced tidal dissipation. Indeed, coupled enhanced tidal heating and deformation has been previously invoked to explain both formation of double-ridges and lenticulae on Europa (e.g., Sotin et al. 2002; Nimmo and Gaidos 2002), and high heat flows (e.g., Pappalardo et al. 1999), and it has been recently shown that tidal dissipation could be much higher, even until 20-times more, in thermal plumes than in sourronding ice (Mitri and Showman 2008; Han and Showman 2010). Moreover, heat flows of 25–35 mW m^{-2} have been proposed from the effective elastic thickness of the lithosphere supporting a plateau near the Cilix impact crater (Nimmo et al. 2003; Ruiz 2005), which suggest an heterogeneous surface heat flow on Europa. Alternatively, parameterized convective models may not be adequate to describe heat transfer in a heterogeneously tidally heated icy shell.

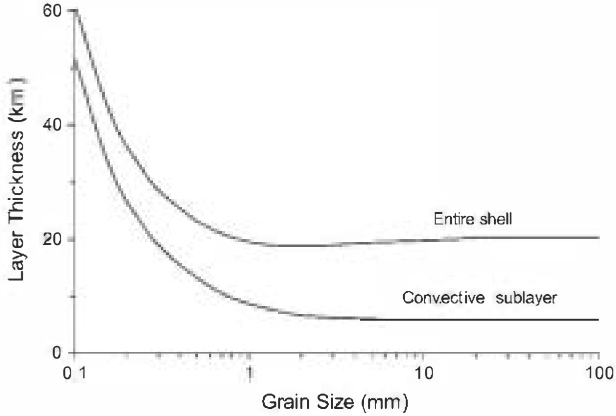

Fig. 4 Convective sublayer and total ice shell thickness as functions of grain size

Figure 4 shows the thicknesses of the actively convective sublayer and the total ice shell (given as the sum of stagnant lid and convective layer thicknesses), as functions of grain size. Both thicknesses first strongly decrease, but remain roughly constants for grain sizes larger than a few millimeters. The nearly horizontal curves for larger grain sizes are due to the dominance of dislocation creep, for which $p = 0$, for these grain sizes. The ice shell is ~20–30 km thick for grain sizes larger than ~0.3 mm, which is fairly consistent with the minimum thickness of ~19–25 km inferred from impact craters (Schenk 2002), and with the ~20 km of depth until an internal ocean suggested by the magnetic evidence (Schilling et al. 2004).

The origin of the areas of microchaos and features known as lenticulae (which include domes, pits and dark spots) could be related to thermal diapirs raising from a convective layer (e.g., Pappalardo et al. 1998; Nimmo and Manga 2002; Sotin et al. 2002). In this case, the convective layer would be ~7–18 km thick (Spaun et al. 2004), roughly corresponding to the half of the observed lenticulae spacing (as proposed by Pappalardo et al. 1998). But a lower boundary layer does not exists for internally heated convection, and hence there are no hot plumes rising (e.g., Turcotte and Schubert 2002). However, a certain amount of heat enters the ice shell from below, and tidal heating, which is strongly temperature-dependent, could therefore intensifier rising plumes (e.g., McKinnon 1999). Figure 4 shows a convective layer of ~6–20 km for grain sizes larger than ~0.6 mm, which is in good agreement with the proposed from the lenticulae spacing.

The analysis of Moore (2006) for a composite rheology, but not considering the effects of tidal stresses on ice flow, found a shell thickness strongly dependent on the grain size. This is a consequence of the domination of ice flow by diffusion creep and grain boundary sliding when only convective stresses are taking into account in the calculations. However a significant role for dislocation creep arises when tidal stresses are incorporated, as expected from the high value of the stress exponent in Eq. 9 for this flow mechanism. This produces nearly constant values for the convective layer and total shell thicknesses for grain sizes larger than ~0.3 mm. Moreover, this increased role for dislocation creep would explain that heat flows calculated for the composed rheology are lower than those obtained for grain boundary sliding in lower grain sizes (Ruiz and Tejero 2003; Ruiz et al. 2007).

5 Conclusions

The results presented in this works suggest that some fundamental geophysical characteristics of the ice shell of Europa may be a natural consequence of the properties of a composite water ice rheology coupled with tidal stresses. Indeed, the use of a composite water ice rheology, alongside the effect of tidal stresses on ice flow, for calculate equilibrium convection on Europa gives convective layer and total ice shell thicknesses consistent, for a wide range of grain size values, with the current knowledge about these parameters. Further application of composite rheologies to Europa and other icy worlds, as well as to different geodynamic situations, could therefore be useful for the research of the evolution and present state of the outer layers of these bodies.

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