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# Note

# Non-Newtonian topographic relaxation on Europa

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#### Abstract

Models of topographic support on Europa by lateral shell thickness variations have previously assumed a Newtonian ice viscosity. Here I show that using a more realistic stress-dependent viscosity gives relaxation times which can be significantly different. Topography of wavelength 100 km cannot be supported by lateral shell thickness variations for  $\sim$  50 Myr, unless the shell thickness is < 10 km or the ice grain size > 10 mm. Shorter wavelength topography would require even thinner shells, but may be supported elastically. Global-scale variations in shell thickness, however, can be supported for geological timescales if the shell thickness is O(10 km).

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## 1. Introduction

An important result of the Galileo mission was the discovery of significant topography on Europa, associated with features such as ridges, bands, chaos regions and plateaus (e.g., Schenk and McKinnon, 2001). The topography is typically 100-1000 m in amplitude, over wavelengths of a few km (ridges) to tens of km (bands, plateaus). The mean surface age of 30-80 Myr (Zahnle et al., 2003) provides an upper bound on the likely topography age. Some topography could be younger, but only if it can be created without significantly modifying pre-existing surface features (an example of such topography is the folds observed by Prockter and Pappalardo (2000)). Although short-wavelength topography can be supported by the rigidity of the ice shell, such support is not effective at longer wavelengths (see Section 5). One potential mechanism which can support long-wavelength loads is lateral variations in shell thickness (Airy isostasy). Because Europa's ice shell is underlain by an ocean (Kivelson et al., 2000), topography of O(100 m) implies variations in shell thickness of O(1 km). The shell thickness of Europa is unknown; estimates vary from  $\approx 2$  to  $\approx 20$  km (e.g., Hoppa et al., 1999; Schenk, 2002). Lateral variations in shell thickness cause pressure gradients which drive flow of ductile ice near the base of the shell (Stevenson, 2000). This flow reduces the shell thickness variations over time. The flow rate is determined by the viscosity structure near the base of the shell.

Stevenson (2000), O'Brien et al. (2002), and Buck et al. (2002) have examined the decay in lateral thickness contrasts for Europa's ice shell. One disadvantage of these models is that they assume an ice viscosity which is independent of stress (i.e., Newtonian), whereas ice is in fact a moderately non-Newtonian material (Goldsby and Kohlstedt, 2001). The effect of nonNewtonian behavior on ductile flow in silicate crusts has been addressed by various authors (e.g., Bird, 1991; Nimmo and Stevenson, 2001). The purpose of this note is to model the decay of topography for realistic, non-Newtonian ice viscosities.

## 2. Theory

For most geological materials, including ice, the relationship between deviatoric stress  $\sigma$  and strain rate  $\dot{\epsilon}$  may be written (e.g., Goldsby and Kohlstedt, 2001)

$$\dot{\epsilon} = A' \sigma^n \exp(-Q/RT),\tag{1}$$

where  $A' = Ad^{-p}$ , *d* is the grain size, *Q* is the activation energy, *R* is the universal gas constant, *T* is the absolute temperature and *A*, *n*, and *p* are rheological constants. The effective viscosity of the material  $\eta_e$  is given by  $\sigma \dot{\epsilon}^{-1}$ .

For ice, the thermal conductivity varies as c/T where c = 567 W m<sup>-1</sup> (Klinger, 1980). Consider an ice shell of thickness  $t_c$  in which heat transfer occurs by conduction only, and there is no internal heat generation. The temperature structure is given by

$$T(z) = T_b \left(\frac{T_s}{T_b}\right)^{z/t_c},\tag{2}$$

where  $T_b$  and  $T_s$  are the bottom and top temperatures, respectively, and z is the height above the base of the shell.

Combining Eqs. (1) and (2) and assuming constant stress, the variation in effective viscosity with position in the ice shell may be linearized as follows:

$$\eta(z) = \eta_0 \exp(z/\delta),\tag{3}$$

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where  $\eta_0$  is the effective viscosity at the base of the ice shell and

$$\delta = \frac{RT_b t_c}{Q \ln(T_b/T_s)}.$$
(4)

The rate of lateral flow, and hence the rate of change in shell thickness variations, is governed by the viscosity structure. Nimmo and Stevenson (2001) showed that for a non-Newtonian material the rate of change of shell thickness  $t_c$  is given by:

$$\frac{dt_c}{dt} = -A'BE_b \frac{d}{dx} \left| \Delta \rho g \frac{\partial t_c}{\partial x} \right|^{n-1} \Delta \rho g \frac{\partial t_c}{\partial x}.$$
(5)

Here *B* is related to the vertically integrated rate of lateral flow and increases with increasing  $\delta$ ,  $E_b = \exp(-Q/RT_b)$ ,  $\Delta\rho$  is the density contrast between the shell and the underlying material and *g* is the acceleration due to gravity. For a Newtonian material, n = 1 and Eq. (5) may be solved analytically. For a sinusoidal disturbance of wavelength  $\lambda$  the time  $\tau$  for the crustal thickness variation to relax to 1/e of its initial value is given by

$$\tau = \frac{\eta_0}{\Delta \rho g \delta^3 k^2},\tag{6}$$

where  $k = 2\pi/\lambda$  is the wavenumber. Hence, short wavelength topography is removed more rapidly than longer wavelength topography.

For the general case  $(n \neq 1)$  the rate of decay depends on the amplitude of the disturbance. Nimmo and Stevenson (2001) used a Cartesian finitedifference version of Eq. (5) to calculate the variation in crustal thickness with time. The same model was used in this work, with the exception that  $\delta$  was calculated using Eq. (4) and parameters appropriate to ice were used (see below). The numerical solution was checked in two ways. Firstly, for integer *n*, analytical expressions for *B* may be derived and it was verified that the numerical approach gave the same values. Secondly, for n = 1 the relaxation timescale is given by Eq. (6); the numerical code gave a timescale which differed by < 1%.

# 3. Parameters

Table 1 gives the properties adopted for this work. The base temperature of the ice  $T_b$  is lower than the melting temperature of pure water to account for the effect of impurities and the effect of pressure. The effect of this parameter on the results is discussed below. The surface temperature of Europa is based on the results of Ojakangas and Stevenson (1989); the results below are more sensitive to  $T_b$  than  $T_s$ .

The stress driving lateral flow will scale as  $\sigma \sim \Delta \rho g \Delta t_c$  where  $\Delta t_c$  is the ice shell thickness variation. Assuming  $\Delta t_c \sim 1$  km, the stress is O(0.1 MPa), significantly larger than diurnal tidal stresses. Goldsby and Kohlstedt (2001) give rheological parameters for ice in a variety of deformation regimes. At the likely stresses and temperatures, the strain rate will be set by whichever is the slower of: basal-slip accommodated grainboundary sliding (GBS) (rheology 1); and GBS-accommodated basal-slip

Table 1Rheologies are from Goldsby and Kohlstedt (2001)

(rheology 2). The reason for the slower rheology controlling the deformation rate is that both mechanisms operate in parallel; this issue is discussed in more detail by Goldsby and Kohlstedt (2001). Which rheology will dominate depends on the grain size of the ice. While the grain size of ice in Europa's shell is unknown, typical terrestrial sea-ice has grain sizes in the range 1–10 mm (Budd and Jacka, 1989). Assuming similar grain sizes for Europa, rheology 2 will dominate.

# 4. Results

Figure 1 plots the relaxation time  $\tau$  calculated using the numerical solution to Eq. (5) for different rheologies as a function of shell thickness  $t_c$  and wavelength  $\lambda$ . The initial variation in shell thickness was 1 km.

As expected from Eq. (6), the timescale decreases with increasing shell thickness and increases with increasing wavelength. For rheology 2 and a 10 mm grain size, lateral shell thickness contrasts will be significantly reduced after 30–80 Myr even at 100 km wavelength, unless  $t_c < 10$  km (Fig. 1b). At shorter wavelengths, relaxation is orders of magnitude more rapid (Fig. 1a). Figure 2 shows the increase in relaxation time with increasing wavelength or decreasing shell thickness. In general, rheology 1 produces more rapid relaxation than rheology 2, as expected.

There are significant differences between the realistic rheologies and those of two Newtonian examples. The Newtonian curves show a gradient of -3, in agreement with Eq. (6), while the slopes of the non-Newtonian curves are steeper. More importantly, the Newtonian rheologies tend to underestimate the relaxation time, especially at long wavelengths. When  $\lambda = 1000$  km, Newtonian relaxation timescales are 1–2 orders of magnitude faster than for rheology 1. Thus, relaxation calculations which assume a Newtonian viscosity (Stevenson, 2000; O'Brien et al., 2002; Buck et al., 2002) may under-estimate the true relaxation timescale, depending on the wavelength of topography being investigated.

### 5. Discussion and conclusions

The results of Fig. 1 strongly suggest that topography with a wavelength of 100 km cannot be supported by Airy isostasy for timescales comparable to the surface age of Europa unless  $t_c < 10$  km or d > 10 mm. Shorter wavelength loads would require even lower shell thicknesses. Conversely, global shell thickness variations, such as those due to spatial variations in tidal heating (Ojakangas and Stevenson, 1989), can survive for O(100 Myr) if  $t_c < 10$ –20 km, depending on grain size (Figs. 1c and 2). The non-Newtonian analysis carried out here thus broadly agrees with the conclusions of Stevenson (2000).

It is important to examine the uncertainties in the above conclusions. The relaxation timescale increases with larger grain sizes or lower temperatures. Increasing the timescale by an order of magnitude requires  $T_b$  to drop to 235 K or  $T_s$  to 40 K. This value of  $T_b$  is unlikely unless Europa's ocean is extremely salt-rich (Kargel et al., 2000), while the value of  $T_s$  is probably colder than polar temperatures on Europa (Ojakangas and Stevenson, 1989).

Variable	Value	Units	Variable	Value	Units
$T_s$	105	К	$T_b$	260	К
g	1.3	$\mathrm{ms}^{-2}$	$\Delta  ho$	100	$\mathrm{kg}\mathrm{m}^{-3}$
	Rheology 1			Rheology 2	
Α	$5.5 \times 10^{7}$	$MPa^{-2.4} s^{-1}$	A	$3.9 \times 10^{-3}$	$MPa^{-1.8} m^{1.4} s^{-1}$
Q	60	kJ mol <sup>−1</sup>	Q	49	$kJ mol^{-1}$
n	2.4	_	n	1.8	_
p	0	-	р	1.4	-

For rheology 2, it is assumed that T < 258 K.



Fig. 1. Relaxation time as a function of shell thickness and rheology. Initial shell thickness variation is sinusoidal with an amplitude of 1 km and wavelength  $\lambda$ . Evolution of shell thickness is calculated using Eq. (5) and numerical method of Nimmo and Stevenson (2001). Rheologies 1 and 2 are defined in Table 1; rheology 2 uses variable grainsizes (*d*). Newtonian rheologies have n = 1, Q = 60 kJ mol<sup>-1</sup> and reference viscosities  $\eta_0$  as noted in the figure; relaxation time is calculated using Eq. (6). Note that a conductive temperature profile is assumed; for thick shells, convection may be occurring. For likely conditions on Europa, rheology 2 is the most appropriate. (a)  $\lambda = 10$  km. Shaded region is the mean surface age of Europa of 30–80 Myr (Zahnle et al., 2003). (b)  $\lambda = 100$  km. (c)  $\lambda = 100$  km.

Rheology 2 of ice changes above 258 K (Goldsby and Kohlstedt, 2001). The activation energy increases, reducing the effective channel thickness in which flow occurs (Eq. (4)), and increasing the flow timescale by about two orders of magnitude. However, most of the flow is likely to take place at temperatures colder than 258 K, suggesting that this particular effect is unlikely to apply.

There are many examples of short-wavelength (O(10 km)) topography on Europa, such as bands and ridges (e.g., Prockter et al., 2002). Since Airy isostasy does not appear to be an effective means of supporting such topography, other mechanisms must be occurring. The cold, near-surface ice on Europa probably has appreciable rigidity (e.g., Nimmo et al., 2003a), which can support short-wavelength loads. Another possibility is topography produced by lateral variations in density (e.g., Nimmo et al., 2003b). Although



Fig. 2. Relaxation time as a function of wavelength. Times calculated as for Fig. 1, using rheology 2 with a grainsize of 1 mm and variable shell thickness ( $t_c$ ).

these density variations will also cause lateral pressure gradients, if they are restricted to relatively shallow depths, the ice will not be warm enough to flow appreciably and the topography may be maintained indefinitely.

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