NON-NEWTONIAN ICE RHEOLOGY AND THE RETENTION OF CRATERS ON GANYMEDE

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Recent laboratory data (Durham *et al.*, 1983; Kirby *et al.*, 1985) reveals that H_2O ice has a strongly non- Newtonian (power-law) rheology at temperatures appropriate to the near-surface regions of the Galilean satellites *i.e.* that

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

where $\dot{\epsilon}$ is strain-rate, τ is effective stress and $A = 2.51 \times 10^{-26} Pa^{-3.7} s^{-1}$, $Q = 27\ kJ\ mole^{-1}$ and $n = 3.7$ for $158\ K < T < 195\ K$. Numerical calculations incorporating this rheological law indicate that crater relaxation in a non-Newtonian rheology has significantly different characteristics to viscous relaxation in a Newtonian medium (Thomas and Schubert, 1986). These arise principally by virtue of the fact that viscosity η varies as $\eta \sim \tau^{1-n}$. Thus high stresses are associated with low viscosities, which permit relatively rapid relaxation flow. However, once the initial stress is reduced, viscosity increases and relaxation flow becomes markedly slower. Consequences of this phenomenon observed for numerical simulations of crater relaxation include the formation of a high-stress, low-viscosity region immediately below the center of the crater; a relatively rapid relaxation rate intially, followed by a marked slowing of the relaxation flow as stresses decrease (and viscosities increase) and, concomitantly, much greater differences between relaxation timescales for small *vs.* large craters than would be expected for a Newtonian medium.

Calculations carried out for craters of varying sizes in a medium with constant temperature $T = 173\ K$ yield values for the crater relaxation time t_e (defined as the time required for the crater depth to become $1/e$ of its original value) that appear to be too small to account for the observed retention of craters on Ganymede and the other icy satellites. Typical values were $t_e \approx 10^2 y$ for craters with a diameter of $D = 300\ km$ and $t_e \approx 10^5 y$ for $D = 50\ km$. Although the rate of crater relaxation at times greater than t_e is much slower than the initial rate, as mentioned above, as many as 10^6 e-folding times are available to reduce the remaining topography. Such a calculation is thus seriously in conflict with the observed crater population of the surfaces of the icy satellites.

In an attempt to reconcile this conflict, we consider possible explanations for the much slower relaxation rate of craters on the icy satellites. It is possible that an admixture of silicates in the surface ice regions of the icy satellites may raise the viscosity to some extent. However, even a silicate volume fraction of the order of 0.6, implausible for the icy satellites, whose surfaces seem to be predominantely ice, only increases the viscosity (and thus the relaxation time) by at most two orders of magnitude (Friedson and Stevenson, 1983).

A rheological transition to a Newtonian law (where $n = 1$) will not occur unless the the Newtonian rheology permits a greater strain-rate for a given stress than the non-Newtonian rheology. One cannot thus increase viscosity by appealing to a rheological transition mechanism. In fact, the critical stress τ_c below which a

Newtonian rheology is dominant, given currently accepted values for Newtonian ice viscosity laws ($Q = 57.7\ kJ\ mole^{-1}$) and ($A = 9.06 \times 10^{-15} Pa^{-1}s^{-1}$) (Reynolds and Cassen, 1979) is of the order of 10 Pa, many orders of magnitude less than the stresses associated with even small craters.

It may be argued that the grain size of the ice in the upper layers of the icy satellites is unknown, and this may render inapplicable the experimental results of Kirby *et al.* (1986). While the creep of polycrytalline ice is sensitive to the grain size at temperatures close to the melting point (Baker, 1978; Goodman *et al.*, 1981), it is probably of less importance at low temperatures (S.H. Kirby, personal communication, 1986). In any event, this effect alone is not likely to be responsible for the orders of magnitude increase required in viscosity.

A solution to the problem can be achieved by accounting for temperature variations in the surface regions of the icy satellites, and utilizing more recent experimental data (Kirby *et al.*, 1986). This data, the product of an over 200% increase in experimental runs at low temperature, yields $n = 4.8$, $A = 2.10 \times 10^{-33} Pa^{-4.8} s^{-1}$ and $Q = 29\ kJ\ mole^{-1}$. The significantly higher value of n is observed to enhance the non-Newtonian phenomena associated with crater relaxation and mentioned above.

To incorporate the effects of a temperature gradient, we take the case of Ganymede ($T_{surface} = 120K$) and assume a temperature gradient $dT/dz = 1\ K\ km^{-1}$, consistent, for the present day Ganymede, with global thermal models (Schubert *et al.*, 1981). These models predict the onset of convective motion at a depth where $T \approx 0.6\ T_m = 164K$. The adiabatic temperature gradient for the icy satellites is small, and thus we assume that the convecting region is isothermal, with $T = T_m$. Thus the maximum temperature determining the viscosity for crater relaxation is 164 K, although higher temperatures may be appropriate to the early history of Ganymede. We further assume that the dominant viscosity is that determined by the temperature T_a at a depth $z = D/2$ (which lies in the center of the relaxation flow region (Thomas, 1986)) and a typical stress τ_a determined from numerical analysis of craterform stresses (Thomas and Schubert, 1986). We can thus construct estimates of relaxation times for various craters as follows:

$D(km)$	$T_a(K)$	$\tau_a(MPa)$	$\eta(\tau)(Pas)$	$t_e(y)$
300	164	1.00	5.76×10^{19}	5.10×10^{4}
250	164	0.83	1.17×10^{20}	1.24×10^{5}
200	164	0.67	2.64×10^{20}	3.50×10^{5}
150	164	0.50	8.02×10^{20}	1.42×10^{6}
100	164	0.33	3.89×10^{21}	1.03×10^{7}
75	157.5	0.25	2.69×10^{22}	9.52×10^{7}
50	145	0.17	∞	∞
25	132.5	0.08	∞	∞
10	125	0.03	∞	∞

Kirby *et al.* (1986) did not observe viscous behavior for $T < 158\ K$, and thus viscous relaxation is not expected for the craters of smallest sizes. Relaxation times are calculated for the $D = 75\ km$ crater, a borderline case for which viscous relaxation may occur.

These results are in good qualitative agreement with observations of the surface of Ganymede: they account for the paucity of well-preserved large ($D > 100\ km$) craters without requiring any *ad hoc* mechanisms such as impact melt-lubrication (Croft, 1983) or acoustic fluidization (Melosh, 1979) while allowing for the long-term retention of smaller craters. Further research will concentrate on the quantitative comparison of calculated relaxation times to the observed population of craters on Ganymede.

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