GEOPHYSICAL RESEARCH LETTERS, VOL. 106, NO. A6, PAGES 1?? , MAY 2002

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Persistence of Thin Ice Regions in Europa's Ice Crust

Abstract. Extensive data from planetary spacecraft and celestial mechanics models support the existence of a subsurface ocean on Europa $\sim$ 100km thick, maintained by a tidal heat flux. These models are consistent with a thin, thermally conductive H_2O ice crust ($\sim$ 5-20 km thick) which would be subject to breaches by impact events and thermal plumes from the tidally heated core. We apply a two dimensional thermal model to the analysis of the refreezing of a hole in the ice crust following a breach event. Our model incorporates heat produced by tidal heating of Europa in two ways: a basal heat flux from Europa's silicate and iron core, together with volumetric heating of the ice shell. We compare our refreezing timescales to those obtained from a model where viscous flow in the base of the ice crust fills the hole. We find that catastrophic breaches in Europa's core may produce regions of relatively thin ice persisting up to $\sim$1 My.

1. Introduction

The presence of a subsurface ocean on Europa is strongly implied from a wide range of imaging data, including the sparsity of impact craters; low-surface relief with lateral separation of crustal plates [*Lucchita and Soderblom*, 1982; *Greeley et al.*, 1998]; models for the formation of flexi by periodic tidal stresses [*Hoppa et al.*, 1999]; and crater morphologies [*Moore et al.*, 1998; *Turtle et al.*, 1999]. The young age (10 My) of Europa's surface, determined from impact crater densities, argues for ongoing geological activity [*Zahnle et al.*, 1998]. Galileo magnetometer data also supports the presence of an ocean via the detection of a possible induced magnetic field in a saline liquid water layer [*Khurana*, 1998].

Tidal heating models for Europa are also consistent with a stable ocean sustained by tidal heating in both Europa's core and the ice crust [*Cassen et al.*, 1979; *Cassen et al.*, 1980; *Squyres*, 1983; *Ross and Schubert*, 1987; *Stevenson and Ojakangas*, 1989]. Tidal model estimates for global ice crust thickness generally lie in the range 10-30 km, consistent with several models of Europan tectonic expression from Galileo images (summarized in [*Pappalardo et al.*, 1999]).

Such a thin icy crust would experience occasional melt through events as a result of both large impacts [*Turtle et al.*, 1999] and the penetration of large scale ocean plumes through the bottom of the ice crust [*Thomson and Delaney*, 2001]. The chaos regions (e.g. Conamara Chaos) are possibly the remnants of such catastrophic events, with "ice rafts" preserving temporarily floating blocks that were frozen in place in the refreezing ice layer [*Carr et al.*, 1998; *Greenberg et al.*, 1999]. Assuming the ice rafts were floating in liquid at the time of their formation (consistent with the record of significant lateral motion preserved in the surface), buoyancy considerations require that the blocks were < 3 km thick when they were formed [*Williams and Greeley*, 1998]. In the model of *Thomson and Delaney*, [2001]. the ocean is assumed to be weakly stratified thermally, so that buoyant plumes can rise from regions of active volcanism on the core to impinge on the ice crust. Such plumes would be narrowly confined by the Coriolis forces arising from Europa's short (3.55 day) rotation period. Using order of magnitude estimates for heat flows in an ice crust, Thomson and Delany estimate that plume heating could result in a melt through of the crust after $\sim 10^3$ y for plumes with heat fluxes $\approx 10^{11}$ W, about 1% of Europa's global heat flux. A more detailed melt-through model that incorporates volumetric heating in the crust [*O'Brien et al.*, 2002] yields broadly similar results.

In this paper, we present the application of a numerical model for the refreezing of a cylindrical hole in Europa's ice crust following a melt-through event (for example, after the hot material from the ascending plume has dissipated). Our model solves the Fourier heat equation within the crust by finite differences in two dimensional cylindrical coordinates (r, z): we have $T = T(r, z, t)$ with

$$c(T)\rho(T)\frac{\partial T}{\partial t} = \nabla \cdot \Big(k(T)\nabla T\Big) + q(T)$$

where $c(T)$, $\rho(T)$ and $k(T)$ are specific heat, density and thermal conductivity, respectively (temperature dependencies given in [*Stevenson and Ojakangas*, 1989]). The term

Paper number 2001JA900021.
0094-8276/02/2001JA900021$5.00

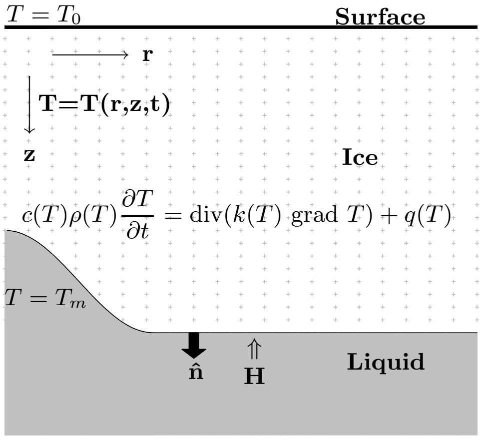

Figure 1. Diagram of numerical model. Velocity $\mathbf{v}$ of ice/ocean interface governed by the equation $L_h\mathbf{v} = k(T_m)\nabla T|_{\text{interface}} - H\hat{\mathbf{n}}$.

$q(T)$ represents a temperature dependent volumetric heating due to tidal flexure of the icy shell. We assume the hole's axis of symmetry to lie at $r = 0$.

The thickness of the model ice crust varies in response to the heat sources applied, maintaining constant temperature boundary conditions, T_0 at the surface (assumed to be 100 K) and T_m (assumed to be 273 K) at the ocean/ice interface. The effect of a surface solid state greenhouse effect on Europa [*Urquhart and Jakosky*, 1982], which we modeled by increasing T_0 to 110 K, produced a negligible change in both equilibrium ice thicknesses and freezing rates. In addition to the volumetric heat source $q(T)$, there is a basal heat source H arising from heat passing through the subsurface ocean from Europa's core. The velocity $\mathbf{v}$ of the front between the ice and the ocean [*Untersteiner and Maykut*, 1971] is governed by the equation

$$L_h\mathbf{v} = k(T_m)\nabla T - H\hat{\mathbf{n}}$$

where $\hat{\mathbf{n}}$ is the unit normal at the interface, L_h is the latent heat required for phase transition, and the thermal gradient ∇T in the ice is evaluated at the interface (Figure 1). When $k(T_m)||\nabla T|| > H$, a new computational cell is added to the ice sheet with its temperature held at T_m for as many timesteps as is needed for the net heat flow into the cell to supply the latent heat required for the phase transition [*Reynolds et al.*, 1966]. This process can be reversed for melting of a cell. Convective heat transfer during the initial stages of refreezing is likely to be unimportant as the ice layer is thin so we do not incorporate convection in our model.

A variety of estimates for H are available in the literature. Early models [*Squyres et al.*, 1983] estimated $H = 24\mathrm{mWm}^{-2}$, a combination of $8\mathrm{mWm}^{-2}$ radiogenic heating and $16\mathrm{mWm}^{-2}$ from tidal dissipation in the core. However, if the core is partially melted, enhanced tidal dissipation could produce much higher heat flows. Models incorporating partially melted cores have estimated much higher basal heat flows: $H = 150\mathrm{mWm}^{-2}$ [*Sjogren and Yoder*, 1996] or $290\mathrm{mWm}^{-2}$ [*Geissler*, 2001]. Thermal plumes from Europa's ocean floor would also produce significant local enhancements in H [*Thomson and Delaney*, 2001; *O'Brien et al.*, 2002].

Early models [*Cassen et al.*, 1982; *Squyres et al.*, 1983] assumed constant values for the volumetric heating rate q within the ice crust. More recent viscoelastic models for ice [*Stevenson and Ojakangas*, 1989] assume a temperature-dependent $q(T)$ that concentrates heating in the lowest and warmest part of the shell.

Assuming a Maxwell viscoelastic rheology for the ice crust, the strain rate $\dot{\varepsilon}_{ij}$ is related to stress σ_{ij} and its time derivative $\dot{\sigma}_{ij}$ by

$$2\dot{\varepsilon}_{ij} = \frac{\sigma_{ij}}{\eta} + \frac{\dot{\sigma}_{ij}}{\mu}$$

where the temperature-dependent viscosity η is given by

$$\eta(T) = \eta_0 \exp\left[l\left(\frac{T_m}{T} - 1\right)\right],$$

where μ is the modulus of rigidity and l is the activation energy constant. Alternative rheological choices, such as a non-Newtonian viscous flow law given by the Glen rheology, do not appear to influence the outcome of thermal models significantly [*Stevenson and Ojakangas*, 1989]. For an icy shell with this Maxwell rheology, the volumetric heating

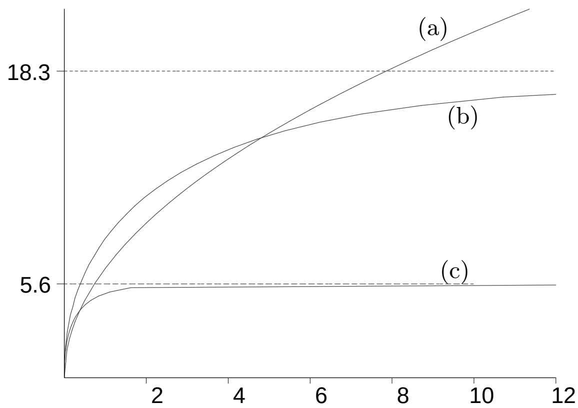

Figure 2. Ice thickness (km) vs. time (My) for (a) $H = 0$, $q = 0$ (Stefan); (b) $H = 24$ mW m^{-2}, Maxwell q; (c) $H = 100$ mW m^{-2}, Maxwell q for the one-dimensional freezing case.

Table 1. Persistence timescales for "warm core" and "hot core" models. Timescales for refreezing (to a thickness of $0.5L_{eq}$) and viscous relaxation are compared.

Model		**Thickening Time (My)**		L_{eq}
H (mW m^{-2})	r_0 (km)	Refreezing	Relaxation	(km)
24	5	0.7	0.003	18.3
24	10	0.8	0.01	
24	50	1.0	0.3	
24	100	1.1	1.1	
100	5	0.1	0.07	5.6
100	10	0.2	0.3	
100	50	0.3	7.0	

rate is given by

$$q(T) = \overline{\sigma_{ij}\dot{\varepsilon}_{ij}} = \frac{2\mu\dot{\varepsilon}_{ij}^2}{\omega}\left[\frac{\omega\tau_M}{1+(\omega\tau_M)^2}\right]$$

where $\tau_M = \eta(T)/\mu$ is the Maxwell Time. For Europa, $\omega = 2\pi/3.55$ days. In this work we assume two values for H: $24\mathrm{mWm}^{-2}$ and $100\mathrm{mWm}^{-2}$, the latter appropriate to a partially melted silicate core with enhanced tidal dissipation. In this work, we will refer to these two models as the "warm core" and "hot core" models, respectively. The equilibrium ice crust thickness L_{eq} is primarily determined by the value of H.

For a Maxwell viscoelastic ice crust, the equilibrium thicknesses are $L_{eq} = 18.3$ km (for a "warm core") and 5.6 km (for a "hot core"). The effect of q is to reduce the freezing rate and to establish a smaller L_{eq} (by approximately 25%). We consider only the case of a conductive ice shell, appropriate to the values of H assumed. For somewhat smaller values of H, the ice crust of Europa may be partly convecting, which would permit more efficient heat transfer and substantially thicker ($\approx$30 km) ice shells. For these ice shells, it has been suggested that viscous flow of the lower boundary will be sufficient to prevent catastrophic melt through events and maintain a generally uniform thickness of the ice [*Stevenson*, 2000]. However, a combination of buoyancy forces and enhanced tidal heating within an upwelling hot plume may overcome this inward viscous flow [*Sotin et al.*, 2002]. In any case, for the thin, conductive ice layers considered in this work ($<$ 20 km), viscous flow will in general be strongly constrained, and catastrophic melting and refreezing will allow substantial time-dependent variations in the thickness of the ice [*O'Brien et al.*, 2002]. The impact of viscous flow on the persistence of thin ice following a melt through event is discussed in the next section.

2. Refreezing Timescales

Galileo spacecraft imaging has revealed the presence of widespread chaotic terrain, comprising 9% of the most highly-resolved surface area [*Riley et al.*, 2000]. The terrain appears to occur over a wide range of length scales, and so here we consider the refreezing of a range of holes in Europa's ice crust with a wide range of initial radii r_0: 5, 10 and 50 km. A schematic view of the numerical model is shown in Figure 1.

Results of this model for a variety of one-dimensional cases (effectively equivalent to $r_0 = \infty$) are shown in Figure 2. Excellent agreement for a range of analytic solutions (e.g. Stefan's law, where $H = q = 0$) and numerical solutions to the heat diffusion equation with non-zero H and q were found. One-dimensional models for refreezing are applicable to large-scale melt through regions ($r_0 \gtrsim 100$ km) [*Thomas et al.*, 2000]. However, for smaller holes, horizontal freezing from the side of the hole will play an important role in refreezing the crust.

The results of our model are summarized in Table 1. Since we are interested in the persistence of relatively thin ice, we estimate the time required to refreeze the crust to 50% of its original equilibrium thickness L_{eq}. For a "hot core," refreezing to 50% of L_{eq} occurs in timescales of $\sim$ 0.2 My. For a "warm core," refreezing takes somewhat longer, typically $\sim$ 0.8 My.

It should be noted that our timescale estimates for refreezing tend to be underestimates, as we assume that no enhancement of heat flow due to the melt through event remains.

We now compare our refreezing timescales to those estimated for relaxation of the hole via flow in a low-viscosity channel at the base of the ice crust [*Stevenson*, 2000]. This model assumes an exponential viscosity gradient arising from a linear temperature gradient within the ice crust. However, a self-consistent solution to the Fourier heat equation for the model discussed above results in an exponential temperature gradient [*Chyba et al.*, 1998]. This reduces the thickness of the channel (typically to the bottom 3% of the layer as a whole) and thus increases the viscous relaxation timescales of those previoulsy estimated [*Stevenson*, 2000]. We find that for a "warm core" model, the viscous flow rate is greater than the refreezing rate for $r_0 < 100$ km (timescales $\sim$ 1 My). For a "hot core" model, the viscous flow rate is greater than the refreezing rate for $r_0 < 10$ km (timescales $\sim$ 0.2 My). For models with even greater core heat flows [*Sjogren and Yoder*, 1996; *Geissler*, 2001], the ice crust is so thin that viscous relaxation is likely to be unimportant for chaos regions of any size.

3. Conclusions

If the chaos regions are the remnants of catastrophic melt through events, reasonable estimates for basal and shell heating predict that relatively thin regions of the ice crust, particularly in the largest chaos regions, will persist for up to $\sim$ 1My, which is 10% of the observed surface age.

Assuming the "hot core" model, chaos regions with linear dimensions greater than 20 km are likely to have thickened to steady state via refreezing, and 200 km assuming the "warm core" model. These linear dimensions should be compared to $\sim$ 1300 km, which is that of the largest contiguous chaos region [*Riley et al.*, 2000]. In either case, it is likely that the ice will remain relatively thin (50% of the surrounding crust thickness) for periods exceeding 0.1 My and up to 1 My for the largest chaos regions.

The detection of the ice/ocean interface by orbital radar sounding is therefore most likely in these areas. This prediction could be tested by future spacecraft missions, for example by radar sounding [*Chyba et al.*, 1998].

4. Acknowledgements

We wish to acknowledge the contributions of our undergraduate student collaborators: Bradley Barth, Derrick Whitelaw, Alex Johnson, Jon Mitchell, Alanna Klucas. This work was supported by the UW-Eau Claire Offices of Research and Sponsored Programs and Professional Development. **Chris's funding at SETI**.

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(Received June 1, 2002; revised July 1, 2002; accepted I hope!.)