Europa: Tidal heating of upwelling thermal plumes and the origin of lenticulae and chaos melting

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[1] Tidal heating models are linked to thermal convection models for ice having strongly temperature dependent viscosity. In the range of ice viscosity inferred from laboratory experiments, tidal forces will heat up rising diapirs on Europa. Partial melt produced in the rising diapirs is predicted to create disruption of near-surface materials and formation of lenticulae and chaos, even if the average ice layer thickness overlying an ocean is larger than 20 km. *INDEX TERMS:* 6218 Planetology: Solar System Objects: Jovian satellites; 8121 Tectonophysics: Dynamics, convection currents and mantle plumes; 8130 Evolution of the Earth: Heat generation and transport; 8147 Evolution of the Earth: Planetary interiors (5430, 5724)

1. Introduction

[2] High-resolution images of surface features on Europa from the Galileo Solid State Imaging (SSI) experiment [*Carr et al.*, 1998; *Greeley et al.*, 1998; *Pappalardo et al.*, 1998; *Spaun et al.*, 1998, 1999; *Greenberg et al.*, 1998; *Riley et al.*, 2000] have revealed the presence of two types of features, chaos and lenticulae, whose characteristics suggest that melting and disruption of the surface have occurred (Figure 1), even though the present temperatures on Europa average about 100 K. This conundrum has resulted in the development of three hypotheses to account for these features and other characteristics of the Europan crust and lithosphere: 1) a thin-shelled model, in which the outer solid layer is less than a few km thick. The surface is constantly in contact with an ocean below due to tidal cracking and disruption, and small variations in the thermal gradient cause chaos and lenticulae [*Carr et al.*, 1998; *Greenberg et al.*, 1998, 1999, 2001]; 2) a thick-shelled model, in which the outer solid layer is of the order of 20–30 km thick and the lenticulae and chaos are the result of diapirism, which delivers warmer material from depth to near the surface, causing disruption and partial melting [*Pappalardo et al.*, 1998; *Rathbun et al.*, 1998; *McKinnon*, 1999; *Pappalardo and Head*, 2001], and 3) a seafloor plume model in which tidal energy focused in the silicate interior at the base of an ocean causes thermal plumes to rise through the ocean and melt through an outer rigid surface layer 2–5 km thick [*Thomson and Delaney*, 2001]. A major difficulty with all three models is how to obtain sufficient heat at shallow depths in the Europan crust and lithosphere to produce the melting apparently required by the observed surface modification and disruption [e.g., *Collins et al.*, 2000].

[3] We report here on a mechanism that can produce melting temperatures and disruption in water ice in the shallow crust of Europa. We show that tidal energy can be preferentially focused in rising plumes for viscosities in agreement with laboratory experiments. When the plume cores reach the base of the outer cold brittle layer, they spread laterally, causing shallow melting, disruption, and formation of terrain similar to lenticulae and chaos. We show that this mechanism can readily explain the major characteristics of chaos and lenticulae even if the average outer solid ice layer thickness is larger than 20 km.

2. Model for the Rise of Diapirs and the Influence of Tidal Heating

[4] Compared to previous description of plumes within Europa [*Rathbun et al.*, 1998], the present study describes plumes arising in a fully developed convective layer. In addition, these 2D numerical thermal convection models [*Deschamps and Sotin*, 2000] take into account the fact that viscosity (η) is strongly temperature dependent by using the following relationship:

$$\eta = \eta(T_{1/2}).\exp\left(-avis.\left(\frac{T - T_{1/2}}{\Delta T}\right)\right) \tag{1}$$

where T is temperature, $T_{1/2}$ is the horizontally averaged temperature at mid-depth, ΔT is the temperature variation across the layer, and *avis* is a coefficient that describes the amount of viscosity variation across the fluid layer. There are two main input parameters of the 2D numerical models: the parameter *avis* and the Rayleigh number ($Ra_{1/2}$) defined as:

$$Ra_{1/2} = \frac{\alpha\rho g \Delta T b^3}{\kappa\eta_{1/2}} \tag{2}$$

where α is thermal expansion, ρ is density, g is gravitational acceleration, b is ice crust thickness, κ is thermal diffusivity, and $\eta_{1/2}$ is viscosity at median temperature. In all models, the temperature at the base of the crust is the melting temperature of H_2O ice (Tm). An example of a steady-state temperature field in the case $Ra_{1/2} = 8 \times 10^3$ and $avis = 13.8$ (viscosity variation of 10^6 between the lower and upper surfaces) is given in Figure 2. The Rayleigh number, calculated with the viscosity at a temperature T_c equal to the mean temperature of the convective fluid, is around 2×10^6. Note the thick conductive lid on top of the convective fluid. For a given set ($Ra_{1/2}$, *avis*), the characteristics of the ice crust are determined as follows. The relationship between the temperature difference across the thermal boundary layer and the viscous temperature scale [*Deschamps*

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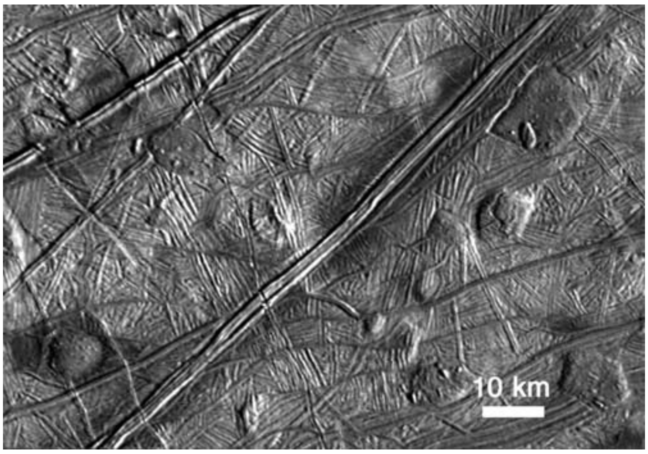

Figure 1. Galileo image of lenticulae near the Conamara Chaos region (~10°N, 273°W).

and Sotin, 2001] is first used. Because the viscous law of ice is better described by an Arrhenius type of law (η = A.exp[Q/RT]), this equation gives a relationship between melting temperature (T_m), core temperature (T_c), *avis*, and activation energy (Q):

$$T_m - T_s = 1.43\frac{T_m - T_s}{avis} = 1.43\frac{RT_c^2}{Q} \tag{3}$$

where T_s is surface temperature. Values of Q range between 40 and 60 kJ/mole [*Durham and Stern*, 2001; *Goldsby and Kohlstedt*, 2001]. The viscosity is supposed to be Newtonian. This is not exactly the case since recent work [*Goldsby and Kohlstedt*, 2001] suggests a superplastic flow with a stress exponent equal to 0.8 for viscosity (1.8 for strain-rate). Although such a small value will slightly change the temperature difference across the thermal boundary layer, it appears insufficient to alter the conclusions relevant to the effect of tidal heating and the formation of partial melt in the uprising plume. The Arrhenius and exponential laws are very close to each other in the domain where there is convection (Figure 2b). The two curves become different in the conductive lid where there is no advection and therefore no influence on the thermal convection model. Using the temperature of the convective fluid and the activation energy from equation (3), the viscosity of the convective fluid can be calculated if the viscosity at a given temperature (we choose $T = T_m$) is provided. Using data in *Goldsby and Kohlstedt* [2001], we find that viscosity at the melting point would be between 10^{13} and 10^{14} Pa.s. The viscosity profile used in the calculation of tidal heating is the one given by the Arrhenius law. The thickness of the crust is then deduced from equation (2). Iterations are required because melting temperature depends on pressure and therefore on the crust thickness. In the example shown in Figure 2, we choose the viscosity of ice at the melting point to be equal to 10^{14} Pa.s, a value which provides the maximum amount of tidal heating (see below). The crust thickness is therefore equal to 35 km.

[5] The thickness and viscosity profile of the ice crust are input parameters of the model computing the deformation of the satellite as it orbits Jupiter. The ice crust overlies a deep ocean (the thickness of the H_2O layer is ~160 km). The silicate core is homogeneous and its radius is 1415 km. The density of the silicate core is determined by the value of the moment of inertia [*Anderson et al.*, 1998]. The spatial distribution of tidal heating is calculated using a perturbation approach: the viscous response of the planet is considered as a first order perturbation of the elastic response to the tidal potential [*Takeushi and Saito*, 1972]. In the case of a Maxwell body, the perturbation ($\delta\mu$) of the elastic shear modulus (μ) is defined as:

$$\delta\mu = \frac{\eta\omega\mu^2}{\mu^2 + \omega^2\eta^2} \tag{4}$$

where ω is circular frequency of straining which can be identified with the satellite's orbital mean motion (e.g., *Segatz et al.* [1988]). The volumetric tidal dissipation rate is the product of an elastic term, which depends on both elastic and orbital parameters, and the viscoelastic term given by equation (4). The tidal dissipated power depends very strongly on viscosity. A maximum value is obtained

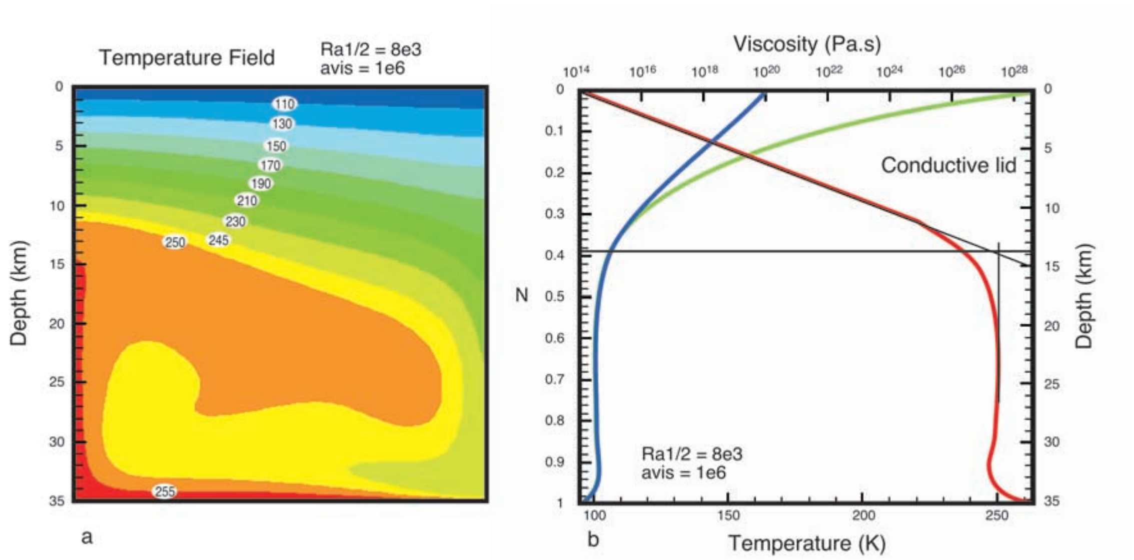

Figure 2. Temperature field (2a) and horizontally averaged temperature profile (2b-red curve) for a model with Rayleigh number ($Ra_{1/2}$) equal to 8×10^3 and viscosity ratio equal to 10^6. The viscosity along the temperature profile is also plotted for an Arrhenius law (green curve) and for the law given by equation (1) (blue curve).

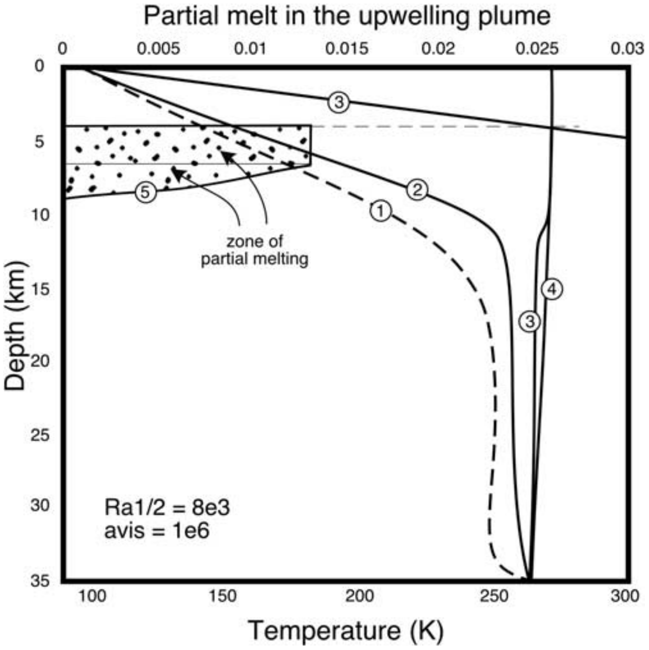

Figure 3. Temperature profiles: curve 1 is the horizontally averaged temperature (red curve in Figure 2a); curve 2 is the temperature in and above the plume core without taking into account tidal heating; and, curve 3 is the same as curve 2 but taking into account tidal heating and it matches the melting temperature (curve 4) when the plume velocity decreases and partial melt starts to be created. Curve 5 denotes the amount of partial melt in the upwelling plume and the location of partial melt is emphasized with dots.

for a viscosity equal to 1.5×10^{14} Pa.s. Using the viscosity field of the 2D thermal convection models, one can derive the tidal heating field.

3. Calculation of Temperature Variations in the Plume

[6] Tidal heating heats up the upwelling plume because the plume moves rapidly relative to loss of heat by lateral conduction. In Figure 3, the mean temperature profile as well as the temperature in the center of the plume are shown. In that case, the upwelling velocity is so high that the temperature does not exceed the melting temperature until the plume slows down as it arrives beneath the conductive lid (around 10 km). There, tidal heating melts the ice and one can compute the amount of ice transformed into water by dividing tidal heating by latent heat. In the models we have examined, 1 to 10% of the ice could be melt in the upwelling warm ice. This process can lead to the formation of a water-rich reservoir (melt lens) at the top of the plume (Figures 3 and 4). The increase of heat flux induced by tidal heating increases the slope of the conductive temperature profile just above the plume by a factor 2 or 3. The conductive temperature profile intercepts the melting curve at a depth (Figures 3 and 4) which can represent the depth where partial melt can still be present (4 km in the present example). Models with smaller values of the Rayleigh number (smaller crustal thickness) predict limited partial melting in the plume tail (1%) and more partial melt just beneath the conductive lid. Continued pumping of warm buoyant plume material into the plume head should laterally enlarge the zone of melting and disruption. The presence of partial melt modifies the viscosity and therefore the amount of tidal heating and focuses the stream lines. Our next step in the modeling is to develop a thermal convection model which includes at each time step the tidal heating field calculated by the method described above [*Tobie et al.*, 2001] and which investigates the effect of partial melting on tectonic activity.

4. Implications for Shallow Melting

[7] On the basis of our analysis, we believe that the incorporation of tidal heating into models for the ascent of diapirs from the base of a thick ice layer can produce melting temperatures in broad areas near the surface of Europa and account for many of the features observed in lenticulae and chaos. Consider a single diapir rising in a water-ice dominated crust (Figure 4). As the rising diapir nears the surface, it is sufficiently buoyant to cause upbowing and fracturing of the overlying lineated plains surface to cause domes (e.g., Figure 4a). With tidal heating (Figure 4b), as the melting isotherm is reached in the subsurface (Figure 3) and the plume head spreads laterally (Figure 2), a zone of partial melting and basal heating exists just below the surface. This has several potential consequences. First, the overlying brittle layer becomes heated, and depending on the initial thickness of the layer and the peak heat flux, can undergo partial melting and disaggregation to produce micro-chaos. Brines and other lower melting temperature impurities [*Kargel et al.*, 2000] will enhance this effect, as will other processes such as brine mobilization [e.g. *Head and Pappalardo*, 1999]. Secondly, the decrease in volume accompanying melting will result in subsidence of the overlying material, creating further surface disruption, and potentially leading to the low topography such as that associated with pits and topographic moats (Figure 4b). Thirdly, the negative buoyancy of the resulting melt will mean that any melt will tend to migrate downward, enhancing subsidence in the surface layer and further removing support from the overlying brittle layer to produce disruption and chaos formation. Fourthly, the thermal perturbation associated with the enhanced near-surface temperatures could lead to conditions that would enhance surface albedo change and potential discoloration through processes of surface frost sublimation, brine mobilization and exposure, and concentration of melt impurities and residues [e.g., *Spencer*, 1987; *Pappalardo et al.*, 1999]. Lastly, continued tidal deformation may overpressurize accumulations of partial melt and propagate dikes and sills upward from the melt zone [*Wilson et al.*, 1997; *Collins et al.*, 2000].

[8] The present work assumes pure H_2O ice. The presence of brines and impurities such as those envisaged by *Kargel et al.* [2000] may affect the value of parameters such as that of viscosity or melting temperature. However, their influence is minor compared to that of first order parameters such as temperature dependent viscosity and tidal heating.

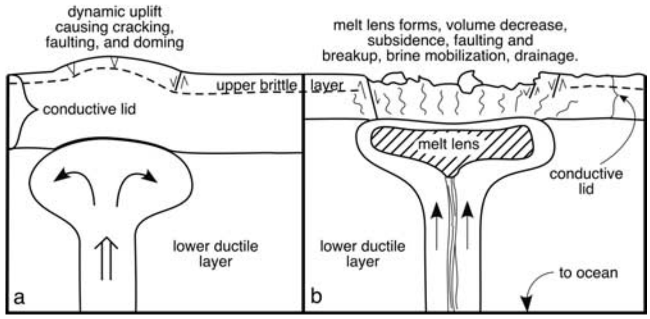

Figure 4. Interpretation of the effects of tidal heating on a plume. Without tidal heating (4a), plume rises buoyantly, impinges on overlying conductive lid, and causes doming. Plume core is tidally heated (4b) and melt lens builds up in the upper part and leads to the collapse and modification seen in chaos and lenticulae.

[9] Despite a 100 K surface temperature, Europa is an environment of major exobiological interest because of the probable presence of a subsurface ocean, and thin-shelled models [e.g., *Greenberg et al.*, 2001; *Thomson and Delaney*, 2001] are attractive because they offer the hope of the ocean and potential access for cryobots. We have shown, however, that even with the thick-shelled model (ice layers in excess of 20 km thickness) [e.g., *Pappalardo et al.*, 1998], exchange between an ocean crystallizing at the base of the ice crust and the surface is ensured by subsolidus convective upwelling of deep material and its partial melting near the surface.

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