Coupled convection and tidal dissipation in Europa's ice shell

Lijie Han a, *, Adam P. Showman b

a Planetary Science Institute, 1700 E. Fort Lowell, Suite 106, Tucson, AZ 85719, USA
b Department of Planetary Sciences, Lunar and Planetary Laboratory, University of Arizona, Tucson, AZ 85721, USA

A B S T R A C T

We performed 2D numerical simulations of oscillatory tidal flexing to study the interrelationship between tidal dissipation (calculated using the Maxwell model) and a heterogeneous temperature structure in Europa's ice shell. Our 2D simulations show that, if the temperature is spatially uniform, the tidal dissipation rate peaks when the Maxwell time is close to the tidal period, consistent with previous studies. The tidal dissipation rate in a convective plume encased in a different background temperature depends on both the plume and background temperature. At a fixed background temperature, the dissipation increases strongly with plume temperature at low temperatures, peaks, and then decreases with temperature near the melting point when a melting-temperature viscosity of $10^{13}$ Pa s is used; however, the peak occurs at significantly higher temperature in this heterogeneous case than in a homogeneous medium for equivalent rheology. For constant plume temperature, the dissipation rate in a plume increases as the surrounding temperature increases; plumes that are warmer than their surroundings can exhibit enhanced heating not only relative to their surroundings but relative to the Maxwell-model prediction for a homogeneous medium at the plume temperature. These results have important implications for thermal feedbacks in Europa's ice shell.

To self-consistently determine how convection interacts with tidal heating that is correctly calculated from the time-evolving heterogeneous temperature field, we coupled viscoelastic simulations of oscillatory tidal flexing (using Tekton) to long-term simulations of the convective evolution (using ComMan). Our simulations show that the tidal dissipation rate resulting from heterogeneous temperature can have a strong impact on thermal convection in Europa's ice shell. Temperatures within upwelling plumes are greatly enhanced and can reach the melting temperature under plausible tidal-flexing amplitude for Europa. A pre-existing fracture zone (at least 6 km deep) promotes the concentration of tidal dissipation (up to ~20 times more than that in the surroundings), leading to lithospheric thinning. This supports the idea that spatially variable tidal dissipation could lead locally to high temperatures, partial melting, and play an important role in the formation of ridges, chaos, or other features.

© 2010 Elsevier Inc. All rights reserved.

1. Introduction

Galileo data show that Europa's surface consists primarily of subdued plains, ridges, bands, chaos, pits, domes, and craters (Greeley et al., 1998, 2000, 2004; Senske et al., 1998; Head et al., 1999; Prockter et al., 1999; Figueredo and Greeley, 2000, 2004). Ridges on Europa's surface can extend several hundreds of kilometers with widths of a few kilometers (Smith et al., 1979; Lucchitta and Soderblom, 1982) and heights up to a few hundred meters (Greeley et al., 2004; Malin and Pieri, 1986). Chaos regions vary from a few kilometers to hundreds of kilometers across (Greenberg et al., 1999; Riley et al., 2000) and up to 250 m in height (Schenk and Pappalardo, 2004). Galileo images indicate that Europa's surface has numerous small (~3–30 km-diameter) landforms, including uplifts, pits, and irregular lobate features (Pappalardo et al., 1998; Greeley et al., 1998; Spaun et al., 1999; Spaun, 2002; Riley et al., 2000; Greenberg et al., 2003). The uplifts (domes) and pits can reach 100–300 m or more topographically (Schenk and McKinnon, 2001; Schenk and Pappalardo, 2004; Greenberg et al., 2003), although debate exists about the size distribution of the features.

Two end-member models have been suggested for the thickness of the ice shell and the formation mechanism of the surface features. Greenberg et al. (1998, 1999, 2002, 2003) and O'Brien et al. (2002) preferred a thin-shell model (~few kilometers thickness). In this scenario, non-synchronous and tidal stress dominate the fracturing on Europa's surface, and diurnal stress makes the existing fractures open and close daily to form ridges and, eventually, bands. Chaos results from local melting of the ice shell due to the local concentration of (tidal) heating at the bottom of a thin-ice...
shell, while incomplete or aborted melt-through events would produce pits (Greenberg et al., 2003). An alternate formation scenario proposes that Europa has a thick ice shell (>10 km). Pits, domes, and chaos resulted from solid-state convection in the ice shell, perhaps aided by partial melting from tidal dissipation (Pappalardo et al., 1988; Rathbun et al., 1998; Head and Pappalardo, 1999; Collins et al., 2000; Spaun et al., 1998; Spaun, 2005). In this scenario, domes would occur when ascending diapirs of buoyant ice impinge against the lithosphere, while pits would form when cold, dense plumes detach from the lithosphere and descend into the interior of the ice shell. Surface disruption (leading to chaos) would naturally result if the convective stresses exceeded the lithospheric yield strength.

In either scenario, heterogeneous tidal heating could play an important role in generating surface features. Heterogeneous, temperature-dependent heating has been suggested to promote the formation of Europa’s ridges (Nimmo and Gaidos, 2002), pits and uplifts (Sotin et al., 2002), and chaos (Collins et al., 2000). In these models, modest temperature increases that occur locally can soften the ice, enhance the tidal dissipation, and potentially lead to a runaway heating that enhances horizontal temperature differences, allows localized partial melting, and promotes surface disruption.

Several authors have made attempts to test the hypothesis that shear heating could cause Europa’s ridges under plausible assumptions about the tidal-flexing amplitude and other parameters. Nimmo and Gaidos (2002) demonstrated that frictional heating can indeed localize along tidally sheared faults, and using a 2D diffusional model, they estimated that temperature increases along the fault could reach ~60 K. Simple buoyancy arguments suggest that these temperature increases can lead to flexural uplift of up to ~100 m, which is encouragingly close to typical ridge heights on Europa. However, their shear-heating calculation did not consider advection of the ice and so could not rigorously determine how the ice deforms in response to the shear heating. Moreover, they calculated the shear heating under the assumption that the shear velocity along the fault zone is constant rather than oscillatory, as would occur with tidal flexing. This forces the full strain rate in the brittle layer to occur along the fault rather than partitioning between fault slip and distributed elastic deformation (Han and Showman, 2008) and changes the temperature dependence of the heating in the viscous layer. Taking a different strategy, Han and Showman (2008) modeled the convective response to an imposed shear heating, and they showed that the shear heating can locally thin the stagnant lid, potentially trigger linear diapirism, and generate ridge-like topographic features ~100 m tall. However, their shear heating was imposed by hand rather than self-consistently determined from the heterogeneous temperature field. Finally, based on finite-element calculations of tidal “walking,” Preblich et al. (2007) performed order-of-magnitude estimates suggesting that only 3–5 K of heating would result from shear heating. However, they did not calculate the dissipation rigorously, nor did they model the long-term thermal evolution, so their estimate does not account for the positive feedback where modest initial temperature increases induce softening that strengthens the localization of strain below the fault, leading to greater heating rates as the runaway progresses.

Many authors have performed convection simulations to study the thermal evolution of Europa’s ice shell (Sotin et al., 2002; Tobie et al., 2003; Barr et al., 2004; Barr and Pappalardo, 2005; Showman and Han, 2004, 2005; Han and Showman, 2005; see Barr and Showman (2009) for a review). So far, all these studies have included a tidal heating that is either constant or depends only on local temperature, following the predictions of an isothermal Maxwell model with zero spatial dimensions (Sotin et al., 2002; Tobie et al., 2003; Showman and Han, 2004, 2005; Han and Showman, 2005). However, this is not self-consistent: in reality, tidal dissipation depends not only on local temperature, but also on surrounding temperature (Mitri and Showman, 2008). Mitri and Showman (2008) constructed an idealized 2D analytical model to show that a convective plume encased in a background of different temperature can experience strongly temperature-dependent tidal dissipation. However, because of their analytic approach, their calculation was limited to Newtonian rheology; moreover, they did not couple their heating calculation to a convection calculation. To fully understand the impact of tidal dissipation on convection and tectonics on Europa, it is important to solve the fully coupled problem of thermal evolution and oscillatory tidal flexing.

Here, we perform several related calculations to study tidal heating and its impact on convection in Europa’s ice shell. In Section 2, we present 2D viscoelastic numerical simulations of the tidal oscillation process in isolation, to determine how a heterogeneous temperature structure—such as an isolated warm or cold convective plume—affects the spatially variable tidal heating. This is similar in approach to Mitri and Showman (2008) but allows us to explore a range of creep-flow mechanisms, both Newtonian and non-Newtonian. In Section 3, we self-consistently couple these viscoelastic tidal–heating calculations to simulations of the convective evolution, which reveal how the feedback between convection and tidal dissipation affects the state of the ice shell. In Section 4, we focus on the shear-heating problem by including a localized weak zone to represent a pre-existing fracture, and determine how the coupled convection and tidal dissipation responds to such a weak zone. Section 5 concludes and discusses geological implications. The emphasis at this stage is exploratory, with the primary goal of identifying the effects that coupling has in idealized models. The model setups are thus intentionally kept simple.

2. Tidal dissipation with heterogeneous temperature structure

We use the 2D finite-element code Tekton (Melosh and Raefsky, 1980) to solve for the oscillatory tidal stress, strain, and tidal heating rate in the presence of temperature heterogeneities in Europa’s ice shell. Following Mitri and Showman (2008), we adopt a 2D model with a circular region of temperature $T_{\text{plume}}$ surrounded by background ice of a different temperature $T_{\text{base}}$, representing a horizontal cross section through an isolated, vertically oriented convective plume. See Fig. 1 for the model setup. We adopt viscoelastic rheology, which is important because the tidal period is close to the Maxwell time. The Young’s modulus is $10^{10}$ Pa and Poisson ratio is 0.25 (Gammon et al., 1983).

Please cite this article in press as: Han, L., Showman, A.P. Coupled convection and tidal dissipation in Europa’s ice shell. Icarus (2010), doi:10.1016/j.icarus.2009.12.028
The viscous rheology of ice can be described using a power-law relationship (Durham et al., 1997; Goldsby and Kohlstedt, 2001)

\[
\dot{\varepsilon} = A \frac{\sigma^n}{\sigma_0^n} \exp \left( -\frac{Q}{RT} \right)
\]

(1)

where \( \dot{\varepsilon} \) is strain rate, \( A \) is a material constant, \( \sigma \) is stress, \( n \) is stress exponent, \( d \) is grain size, \( p \) is grain-size exponent, \( Q \) is activation energy, \( R \) is the gas constant, and \( T \) is temperature. We explore three different creep-flow mechanisms: diffusion, grain boundary sliding (GBS), and dislocation creep. See Table 1 for the rheological parameters. While the grain sizes are unknown, estimates have suggested 0.1–1 mm (Kirk and Stevenson, 1987; Barr and McKinnon, 2007). To bracket this range, we explore values spanning 0.1–1 mm. For the present study, we neglect pre-melting, which could lead to viscosity drop and greater strain rates near the melting temperature than are captured in Eq. (1) (e.g., Dash et al., 1995; Duval, 1977; De La Chapelle et al., 1999; Tobie et al., 2003).

In the numerical models, the calculation domain is a square with dimensions \( L \) of 20 km \( \times \) 20 km and a resolution of 100 \( \times \) 100 finite elements (200 m grid spacing). To simulate the tidal-flexing process, we specify a displacement boundary condition wherein the position of the sidewalls at \( x = L/2 \) and \( -L/2 \) are displaced sinusoidally in time as \( \zeta_0 \sin \omega t \) and \( -\zeta_0 \sin \omega t \), respectively, where \( \zeta_0 \) is 0.125 m and \( \omega = 2.07 \times 10^{-5} \) s\(^{-1} \), corresponding to Europa’s orbital period of 3.5 Earth days. The amplitude is chosen so that the domain-averaged peak-to-peak tidal-strain amplitude is \( 2\zeta_0/L = 1.25 \times 10^{-5} \), which is within the range of global values of \( \epsilon_0 \sim 1 \times 10^{-5} \) calculated for Europa’s expected internal structure (i.e., an ice shell, internal ocean, and silicate core) and current orbital eccentricity (Ojakangas and Stevenson, 1989; Ross and Schubert, 1987; Moore and Schubert, 2000). Note, however, that depending on the temperature structure, the local value of tidal-flexing strain within the domain can deviate from the spatial-mean value \( 2\zeta_0/L \); this is an important aspect, since strain localization is a major mechanism for generating spatially concentrated tidal heating. The horizontal boundaries at \( y = \pm L/2 \) are free (deformable) surfaces, along which the stress-tensor components \( \sigma_{xy} \) and \( \sigma_{yy} \) are taken to be zero. The calculations adopt plane strain.

![Image](340x120 to 357x327)

Globally tidal potential varies with latitude and longitude (Kaula, 1964). However, our models only consider a regional geometry, less than \( 10^{-4} \) of the entire spherical shell in area. The variations in the mean tidal-flexing strain that occur across distances comparable to Europa’s radius (e.g., between the equator and pole) can therefore be neglected on the scale of our domain (20 km-width). Likewise, Europa’s tidal strain may vary in radius (Kaula, 1964; Tobie et al., 2005). But our models only consider the top 20 km of the ice layer, representing about 1% of Europa in radius. Thus, we neglect the radial tidal strain variation within this layer.

We start the simulations with zero stress, and run the simulation for 5–10 tidal cycles until the initial transients die out and a temporally periodic solution is achieved. Each tidal cycle is resolved with 85 steps. We then calculate the tidal dissipation rate at each cell of the 2D domain by integrating stress times strain rate over the last tidal cycle, yielding a heating rate per volume:

\[
q(x,y) = \frac{1}{\Delta T} \int \sigma_{ij}(x,y,t) \dot{\varepsilon}_{ij}(x,y,t) \, dt
\]

(2)

where \( q \) is the tidal heating rate, \( \sigma_{ij} \) is stress tensor, \( \dot{\varepsilon}_{ij} \) is strain-rate tensor, \( t \) is time, and \( \Delta T \) is the tidal cycle period. Subscripts correspond to the \( x \) and \( y \) coordinate axes; repeated indices imply summation.

**Fig. 2** displays the tidal dissipation rates versus temperature from models with a homogeneous (spatially independent) temperature distribution. Each point corresponds to a distinct Tekton simulation; the curves connect families of simulations with the same rheology but different temperatures. Under diffusion-creep rheology with an ice grain size of 0.5 mm (corresponding to melting-temperature viscosity at \( 10^{14} \) Pa s), the tidal dissipation rate peaks at a temperature \( \sim 270 \) K. If the ice grain size decreases to 0.15 mm (corresponding to melting-temperature viscosity of \( 10^{13} \) Pa s), the tidal dissipation rate peaks at temperature of \( \sim 250 \) K. This is consistent with previous estimates of tidal dissipation of uniform temperature structure with diffusion-creep rheology (Mitri and Showman, 2008). Under grain boundary sliding (GBS) with an ice grain size of 0.1 mm, our results show that the tidal dissipation rate is comparable to that obtained from models using diffusion rheology with a melting-temperature viscosity of \( 10^{14} \) Pa s. When the ice grain size increases to 1 mm, GBS models show that tidal dissipation is very close to diffusion models with a melting-temperature viscosity of \( 10^{13} \) Pa s. The tidal dissipation rate is about two orders of magnitude smaller if dislocation creep is implemented. For the temperatures, stresses, and grain sizes that are likely in a convecting Europa’s ice shell, deformation by diffusion and grain boundary sliding (GBS) are most relevant (Goldsby and Kohlstedt, 2001; Durham and Stern, 2001; McKinnon, 1998, 1999).

Note that, although previous studies have explored the tidal dissipation rate under the assumption of Newtonian rheology (which in the case of sinuosoidal forcing leads to a sinusoidal response that is straightforward to treat analytically), this is the first study to rigorously quantify the tidal dissipation rate for non-Newtonian rheology.

**Fig. 3** shows that the tidal dissipation rate in a convective plume depends strongly on plume temperature when the background ice...
temperature remains fixed and diffusion creep is implemented. Tidal dissipation rates depend on both local and background temperature. **Diffusion-creep** rheology with a melting-temperature viscosity of $10^{13}$ Pa s is implemented in the simulations. Fig. 3 considers a heterogeneous ice shell containing a plume embedded in an environment of different temperature as in Mitri and Showman (2008). The plume is 1 km in radius within a domain of 20 km in length and width (see Fig. 1). Each point corresponds to a simulation with a convective plume at one temperature and background ice at a different temperature, and the curves connect simulations that hold the background temperature constant (at 200, 250, or 260 K) yet consider different plume temperatures. For comparison, Fig. 3 also includes simulations with homogeneous temperature in the calculation domain (plus signs). The tidal dissipation rate in a plume with lower temperature than the background is smaller than the tidal dissipation rate calculated from a homogeneous temperature. Tidal dissipation rate in a plume with higher temperature (than the background) is greatly enhanced compared with the results from a homogeneous temperature. Furthermore, tidal dissipation rate in a convective plume at one particular temperature decreases as the surrounding ice temperature increases.

In our simulations, the strong dependence of the tidal dissipation rate on the temperature of the plume and its surroundings is qualitatively consistent with Mitri and Showman’s (2008) results and supports the idea that runaway heating can potentially occur in convective plumes for appropriate parameters, as originally suggested by Sotin et al. (2002). Nevertheless, our results exhibit some modest differences with Mitri and Showman in terms of dissipation magnitude and the temperature at which the dissipation peaks. These differences may be caused by several factors: First, our models implemented displacement boundary conditions, while Mitri and Showman (2008) used stress boundary conditions. Second, an infinite domain was used in Mitri and Showman’s (2008) model, while our numerical simulations implement a finite domain. Third, several physical parameters used in the models are slightly different (for example, the Poisson ratio is 0.314 in Mitri and Showman (2008), and it is 0.25 in our models).

### 3. Coupled thermal convection with tidal dissipation

Next, we perform fully coupled numerical simulations of thermal convection and tidal heating that we self-consistently calculate from the time-evolving temperature structure. We again adopt 2D cartesian (rectangular) geometry, with the dimensions in this case representing horizontal position $x$ and height $z$. Cartesian geometry is appropriate for regional studies of Europa’s ice shell because Europa’s ice-shell thickness is much smaller than its radius. We neglect inertia and adopt the Boussinesq approximation. We use the finite-element code ComMan (King et al., 1990) to solve the dimensionless equations of momentum, continuity, and energy, respectively given by

Please cite this article in press as: Han, L., Showman, A.P. Coupled convection and tidal dissipation in Europa’s ice shell. Icarus (2010). doi:10.1016/j.icarus.2009.12.028
\[
\frac{\partial \sigma_{ij}}{\partial x_j} + Ra \frac{\partial \theta}{\partial x_i} = 0
\]  
(3)

\[
\frac{\partial u_i}{\partial x_i} = 0
\]  
(4)

\[
\frac{\partial \theta}{\partial t} + u_i \frac{\partial \theta}{\partial x_i} = \frac{\partial^2 \theta}{\partial x_i^2} + q
\]  
(5)

where \( \sigma_{ij} \) is the dimensionless stress tensor, \( u_i \) is dimensionless velocity, \( \theta \) is dimensionless temperature, \( q \) is dimensionless heating rate, \( k_i \) is the vertical unit vector, \( t \) is dimensionless time, \( x_i \) and \( x_j \) are the dimensionless spatial coordinates, and \( i \) and \( j \) are the coordinate indices. Repeated spatial indices imply summation. The Rayleigh number \( Ra \) is given by

\[
Ra = \frac{g \rho x A T_0^3}{\kappa \eta_0}
\]  
(6)

where \( g \) is gravity, \( \rho \) is density, \( x \) is thermal expansivity, \( A T \) is the temperature drop between the top and bottom boundaries, \( d \) is the depth of the system, \( \kappa \) is the thermal diffusivity, and \( \eta_0 \) is the dynamic viscosity at the melting temperature. For reference, the model parameters are presented in Table 2.

The purpose of the numerical simulations here is to study the impact of tidal heating resulting from heterogeneous temperature on thermal convection in Europa's ice shell. We run two sets of simulations. For one set, we adopt a tidal dissipation rate depending on local temperature only, based on the zero-dimensional isothermal Maxwell model prediction as follows (Sotin et al., 2002; Tobie et al., 2003; Showman and Han, 2004, 2005).

\[
q = \frac{\epsilon_0 (\omega^2)^2}{2 \left(1 + \omega T^2 \right)}
\]  
(7)

where \( \omega \) is Europa's tidal-flexing frequency, \( \mu = 4 \times 10^7 \) Pa is the rigidity of ice, and \( \epsilon_0 \) is the amplitude of tidal-flexing strain (assumed constant throughout the domain). While this is not fully self-consistent, this approach has been adopted by most previous authors and thus this set of simulations serves as a reference.

For another set of simulations, we study the tidal dissipation rate and its impact on thermal convection in fully coupled models of thermal evolution and tidal heating. This set of simulations includes two components: convection (which is solved using ConMan and involves timescale of millions of years) and oscillatory tidal flexing (which is solved using the finite-element code Tekton and involves timescale of days to weeks).

We adopt a 2D approach (with dimensions corresponding to horizontal distance, \( x \), and height, \( z \)). This is equivalent to assuming no variation in properties along the coordinate \( y \) perpendicular to the domain. Because Europa ridges are linear structures with minimal variation along strike, this approach is ideally suited to investigate shear heating as a ridge-formation mechanism. Performing coupled 3D simulations is a long-term goal but is computationally extremely expensive; the 2D simulations described here represent an important first step.

The coupled calculation proceeds as follows. (1) For a given temperature structure \( T(x,z) \) throughout the ice shell, we perform viscoelastic calculations assuming oscillatory tidal flexing to determine the stress/strain response throughout a tidal cycle. We then determine the tidal dissipation \( q(x,z) \) for that thermal structure by integrating stress and strain rate over a cycle (Eq. (2)). (2) Including this dissipation as a heat source in Eq. (5), we then integrate forward the thermal-evolution equations, accounting for both thermal diffusion and advection/convection in the ice shell, until the temperature has changed by a modest amount. (3) This new thermal structure \( T(x,z) \) is then used to recalculate the tidal dissipation, hence repeating step (1), and the cycle repeats. The viscoelastic tidal-heating calculation can thus be viewed as a self-consistent means for specifying the time and spatially varying tidal-dissipation source term in the thermal evolution calculation. We carry the calculation forward until the system reaches a statistically steady state.

Because the tidal-oscillation model needs to run \( \sim 5 \)–10 tidal cycles (each cycle being solved with 85 timesteps) it is very computationally expensive to calculate the dissipation rate at every thermal-convection timestep in the coupled model. Moreover, the change in thermal structure (and the resulting change in tidal dissipation rate) per convective timestep is small. We ran test cases to calculate the tidal dissipation rate every 20, 50, 100, or 200 timesteps in the thermal convection model, and the thermal structure is identical in each case. Therefore, in practice, we generally run the tidal-oscillation model to calculate the dissipation rate every 100–200 timesteps in the thermal convection model.

We perform (both Tekton and ConMan) simulations with ice-shell thicknesses of 15 or 20 km, consistent with ice-shell thickness estimated from evolution, crater morphology, and crater size-distribution studies (Hussmann et al., 2002; Ojakangas and Stevenson, 1989; Turtle and Ivanov, 2002; Schenk, 2002). We maintain constant ice-shell thickness in any given simulation. The simulations are performed with aspect ratio (ratio of width to depth) of 1 or 3. The aspect ratio has little effect on the qualitative results. The resolution is typically 100 elements in the vertical direction and 100 or 300 elements in the horizontal direction depending on the geometric ratio.

As we have described in Section 2, diffusion creep and grain boundary sliding (GBS) creep (the two creep mechanisms most relevant for Europa) have qualitatively the same effects on tidal dissipation rates under the same heterogeneous temperature structure. Because the current simulations are exploratory attempts to understand the coupling between convection and tidal dissipation, for simplicity, here we only consider diffusion creep (i.e. Newtonian, temperature-dependent viscosity) appropriate for ice (Sotin et al., 2002; Tobie et al., 2003; Showman and Han, 2004, 2005; Han and Showman, 2005, 2008).

\[
\eta(T) = \eta_0 \exp \left( A \frac{T_m}{T} - 1 \right)
\]  
(8)

where \( T \) is temperature, \( T_m \) is melting temperature (here assumed constant at 270 K) and \( \eta_0 \) is the viscosity at the melting temperature. We fix \( \eta_0 \) at \( 10^{13} \) Pa s or \( 10^{14} \) Pa s, appropriate to ice with grain sizes of \( \sim 0.1 \) mm or \( \sim 0.5 \) mm, respectively (Goldsby and Kohlstedt, 2001). The value of \( A \) is assumed constant at 26, corresponding to activation energy of 60 kJ mol\(^{-1}\), which implies a viscosity contrast of \( 10^{20} \). We neglect the influence of partial melting on viscosity in Eq. (8). Modest increase in melt fraction greatly increase the percolation rate (Stevenson and Scott, 1991), which tends to stabilize the melt fraction at small values of \( 0.001 \)–0.01 for conditions relevant to Europa (Showman et al., 2004; Gaidos and Nimmo, 2004).

### Table 2

<table>
<thead>
<tr>
<th>Name</th>
<th>Symbol</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Acceleration of gravity</td>
<td>( g )</td>
<td>( 1.3 ) m s(^{-1})</td>
</tr>
<tr>
<td>Density</td>
<td>( \rho )</td>
<td>( 917 ) kg m(^{-3})</td>
</tr>
<tr>
<td>Thermal expansivity</td>
<td>( x )</td>
<td>( 6.5 \times 10^{-4} ) K(^{-1})</td>
</tr>
<tr>
<td>Thermal diffusivity</td>
<td>( \kappa )</td>
<td>( 1 \times 10^{-6} ) m(^2) s(^{-1})</td>
</tr>
<tr>
<td>Specific heat</td>
<td>( c_p )</td>
<td>( 2000 ) J kg(^{-1}) K(^{-1})</td>
</tr>
<tr>
<td>Surface temperature</td>
<td>( T_s )</td>
<td>95 K</td>
</tr>
<tr>
<td>Bottom temperature</td>
<td>( T_b )</td>
<td>270 K</td>
</tr>
<tr>
<td>Melting-temperature viscosity</td>
<td>( \eta_b )</td>
<td>( 10^{13} ) or ( 10^{14} ) Pa s</td>
</tr>
<tr>
<td>Thickness of ice shell</td>
<td>( d )</td>
<td>( 15 ) or 20 km</td>
</tr>
</tbody>
</table>
Therefore we assume the effect of partial melting on viscosity can be neglected. Nevertheless, pre-melting may cause additional decrease in viscosity near the melting temperature (Dash et al., 1995; Duval, 1977; De La Chapelle et al., 1999) beyond that represented by Eq. (8) and should be considered in future work.

In the thermal convection models, we fix the bottom boundary to a temperature of 270 K, the melting temperature. The top surface is maintained at 95 K, 35% of the basal temperature. The temperature boundary condition is periodic on the sides. The temperature was initialized as a linear function of depth (i.e., a conductive profile), with a weak perturbation near the bottom to break the lateral symmetry. The velocity boundary conditions are periodic on the sides and free-slip rigid walls on the top and bottom.

In the Tekton calculation, we implement a displacement boundary condition where the sidewall positions at \( x = L/2 \) and \(-L/2\) vary in time as \( \zeta_0 \sin \omega t \) and \(-\zeta_0 \sin \omega t\), respectively, with a displacement amplitude \( \zeta_0 \) specified to give a spatial-mean peak-to-peak strain amplitude \( 2\zeta_0/L \) (where \( L \) is the width of the domain) of \( 1.25 \times 10^{-5} \). The oscillation period is 3.5 Earth days. This boundary condition forces the domain to experience a temporally periodic cycle of extension/contraction. On Europa, the stress field generally rotates in time such that a tidal cycle involves both extension/contraction and shear (e.g., Preblich et al., 2007). This could be taken into account by including in the displacement boundary condition a component of oscillation perpendicular to the domain, but we would not expect this phenomenon to qualitatively alter the dissipation patterns, and we leave an exploration of it for future work. The top/bottom boundaries are open (deformable) surfaces where \( \sigma_{xx} \) and \( \sigma_{yy} \) are taken to be zero. In comparison, in our reference simulations where we determine tidal heating using Eq. (7), we adopt \( \zeta_0 = 1.25 \times 10^{-5} \) or \( 2.1 \times 10^{-5} \). The strain magnitude of \( 2.1 \times 10^{-5} \) is chosen so that the peak values of tidal dissipation rates from the Tekton solution and zero-dimensional estimate are identical.

Fig. 5 shows the temperature distribution from a fully coupled simulation with a mean strain amplitude of \( 1.25 \times 10^{-5} \) and a melting-temperature viscosity of \( 10^{13} \) Pa s. The domain has a thickness of 15 km, with an aspect ratio (width versus depth) of 3. A thick thermal conductive layer is formed at the top of the domain, and the temperature reaches 260–270 K throughout the bottom two-thirds of the domain (top panel in Fig. 5). To examine the temperature field in more detail, we replot the temperature distribution using a modified scale (bottom panel in Fig. 5) that highlights the structure in the convection region. Two strong upwellings plumes and several downwellings dominate the lower part of the domain, with temperature reaching melting temperature (270 K) in these plumes.

Fig. 6 shows the temperature distribution from a convection simulation using the exact same parameter setup as the model shown in Fig. 5, but determining the tidal dissipation rate using Eq. (7) rather than Tekton. The model implements a strain magnitude of \( 1.25 \times 10^{-5} \). Results show that temperature within the top one-third of the domain is similar to that in Fig. 5, forming a thermal conductive layer. However, the temperature in the bottom half of the domain is about 5 K lower, and the melting temperature is not reached anywhere in the simulation. On the other hand, Fig. 7 shows the temperature distribution from a simulation similar to the model in Fig. 6 (implementing tidal dissipation rate using Eq. (7)), but with strain magnitude of \( 2.1 \times 10^{-5} \). The temperature is 2–3 K lower than the model using the fully coupled convection and tidal dissipation with a strain magnitude of \( 1.25 \times 10^{-5} \) (Fig. 5). The melting temperature is not reached in this simulation.

To understand what causes the temperature differences in different models, we plot out the tidal dissipation rates for a given heterogeneous temperature structure using different dissipation calculation methods (2D tidal oscillation simulation using Tekton versus simple estimation by Eq. (7)) (Fig. 8). Fig. 8a displays the tidal dissipation rate associated with the temperature structure shown in Fig. 5 using a 2D Tekton simulation based on a mean
Fig. 7. Temperature distribution from a model of thermal convection with tidal heating rate calculated using Eq. (7). Tidal flexing amplitude of \(2.1 \times 10^{-5}\) and a tidal period of 3.5 days are used. The thickness of the ice shell is 15 km, and the Rayleigh number is \(1.81 \times 10^7\). Top: temperature range of 90–270 K. Bottom: temperature range of 260–270 K.

Fig. 8. Tidal dissipation rates for a given heterogeneous temperature structure (shown in Fig. 5) using different dissipation calculation methods. (a) Tidal dissipation rates using Tekton simulation based on a tidal-flexing amplitude of \(2.1 \times 10^{-5}\). (b) Tidal dissipation rates calculated using Eq. (7) with a strain magnitude of \(2.1 \times 10^{-5}\). (c) Tidal dissipation rates calculated using Eq. (7) with a strain magnitude of \(1.25 \times 10^{-5}\).

Fig. 9. Temperature distribution from a fully coupled ConMan/Tekton model of thermal convection and oscillatory tidal flexing (see Fig. 4). Tidal flexing amplitude of \(1.25 \times 10^{-5}\) and tidal period of 3.5 days are used. The thickness of the ice shell is 15 km, and the Rayleigh number is \(1.81 \times 10^7\). Top: temperature range of 90–270 K. Bottom: temperature range of 260–270 K.

strain amplitude of \(1.25 \times 10^{-5}\). Tidal dissipation rates under the same temperature structure using Eq. (7) with different strain magnitudes of \(2.1 \times 10^{-5}\) and \(1.25 \times 10^{-5}\) are shown in Fig. 8b and c, respectively. The tidal dissipation rate in Fig. 8a considers the effects from both local and surrounding temperature. On the other hand, the tidal dissipation rates in Fig. 8b and c depend only on local temperature. We see that tidal dissipation rate peaks at the location where the temperature is about 250 K, because the Maxwell times in those locations are close to the tidal oscillation period. However, the peak tidal dissipation rate and average tidal dissipation rate in the upwellings are larger (by a factor of 2–5) in Fig. 8a compared with those rates in Fig. 8b and c. This explains why the mean temperature is higher in the fully coupled tidal dissipation and convection simulation (Fig. 5).

The patterns of thermal convection and tidal dissipation change dramatically if the melting-temperature viscosity is altered. Fig. 9 shows the temperature distribution from a fully coupled simulation, with the same set up as that in Fig. 5, but with a melting-temperature viscosity of \(10^{14}\) Pa s. A thick thermal conductive layer is formed at the top of the domain, and several strong upwellings and downwellings dominate the lower part of the domain, with temperature reaching melting (270 K) in these upwellings. Fig. 10 displays the tidal dissipation rate for a given heterogeneous temperature (shown in Fig. 9) using different dissipation calculation methods (2D tidal oscillation simulation using Tekton in Fig. 10a versus simple estimation by Eq. (7) in Fig. 10b and c). The tidal dissipation rate within the upwellings (265–270 K) is greatly enhanced due to the fact the Maxwell time in these areas are very close to the tidal oscillation period.

What overall implication do these results have? An examination of Figs. 8 and 10 reveals that the horizontal variation of tidal dissipation in the fully coupled case (top panels) exceeds that calculated from Eq. (7) with identical temperature structure (bottom two panels). In Fig. 8, the peak dissipation occurs at the base of the stagnant lid, but the fully coupled case (Fig. 8a) shows that the dissipation in this layer is significantly greater over ascending
plumes than in other regions; in contrast, the dissipation at the base of the stagnant lid as estimated by Eq. (7) is more horizontally uniform (Fig. 8b and c). Likewise, in Fig. 10, the ratio of dissipation in convective upwellings and downwellings is greater in the fully coupled case than that estimated by Eq. (7). These results would suggest that the fully coupled case might be able to develop greater spatial heterogeneity than the non-coupled counterparts, perhaps promoting surface disruption. Examination of the temperature fields (Figs. 5–7, 9 and 10) suggests that the fully coupled cases more readily reach the melting temperature within local regions such as plumes. This is consistent with our horizontal 2D models from Figs. 2 to 3, which show that, for a given plume temperature and spatially-averaged tidal-flexing amplitude, a warm plume experiences greater dissipation when the dissipation is calculated self-consistently than when calculated via Eq. (7). Thus, our results support the idea that tidal heating could be sufficiently heterogeneous to allow (for example) localized partial melting, increasing the likelihood of chaos formation or other surface disruption. Nevertheless, additional work will be needed to fully quantify the implications of self-consistent coupling for surface tectonics as well as its effect on metrics such as Nusselt number–Rayleigh number relationships.

4. Coupled convection and tidal dissipation in the presence of a weak zone

Next, we explore the effect of a weak zone—representing a pre-existing fracture—on the coupled convection and tidal dissipation. The motivation here is to better understand the extent to which localized heating, modulated by a fracture, could play a role in the formation of Europan ridges. A weak zone can lead to lithospheric thinning and regulate thermal convection patterns within the ice shell (Tobie et al., 2005; Han and Showman, 2008). Fracture zones in Europa’s ice shell may weaken the material and cause concentration of tidal stress and/or strain, which may result in strongly localized tidal dissipation (Nimmo and Gaidos, 2002).

To study the impacts of fracture zones on tidal heating and convection, we impose a narrow (few kilometer wide) weak zone extending vertically downward from the surface to the base of the lithosphere. The weak zone is implemented by simply using weaker material (one tenth of the viscosity calculated from Eq. (8)) compared with the surroundings. This approach has been widely used to study weak zones (i.e. ridges or subduction zones) in the Earth mantle convection community (e.g., Zhong and Gurnis, 1994; Zhong et al., 1998). For simplicity, we did not consider frictional heating of the fault movement in our simulations.

We perform simulations with ice shell thicknesses 20 km, with an aspect ratio (ratio of width to depth) of 1. The weak zone is located at the center of the domain horizontally, with a width of 2 km wide, extending from the surface to the depth of 6–10 km. The resolution is 100 elements in the vertical direction and 100 elements in the horizontal direction in the domain.

Fig. 11 shows the results from a fully coupled convection and tidal dissipation model with a weak zone 2 km wide, extending from the surface to a depth of 10 km. The melting-temperature viscosity is 10^{13} Pa s in this case. The pre-existing fracture zone promotes the concentration of tidal dissipation. The tidal dissipation rate is greatly enhanced at temperature of about 250 K within the weak zone (see Fig. 11a). The upwelling plume reaches much shallower depth in the weak zone location. This supports the idea that enhanced heating along pre-existing fractures may be important for the formation of tectonic features (e.g., ridges). Simulations varying the depth of weak zone show that the weak zone has to be at least 6 km deep (i.e. fracture zone extending from the surface to a depth of 6 km) for it to have a visible influence on the temperature field. The viscosity near the surface is extremely large and tidal dissipation rate is very small due to the low local temperature.

For comparison, we ran a model implementing the same parameters as those in Fig. 11, but with tidal dissipation calculated by Eq. (7). The tidal dissipation rate and temperature structure are shown in Fig. 12. The tidal dissipation rate peaks over the top of the upwelling plume (the location where temperature is 250 K), due to the fact tidal dissipation depends on local temperature only. The high tidal dissipation rate (top panel in Fig. 12) is slightly elevated to a shallower depth within the weak zone because of the viscosity decrease there. The upwelling plume penetrates to a shallower depth compared with the model without weak zone. However, the tidal dissipation rate is not greatly enhanced in the weak zone as that estimated by Eq. (8).

5. Conclusions and discussions

We presented three calculations of the influence of tidal heating on Europa’s ice shell, emphasizing the possible role that spatially heterogeneous temperatures play in affecting the tidal dissipation and its feedback with convection. First, we used the 2D viscoelastic code Tekton to explore the tidal dissipation that occurs in the presence of an isolated cold or warm plume surrounded by ice of differing temperature. The pre-existing fracture zone enhances dissipation when the dissipation is calculated self-consistently, whereas the non-coupled case yields lower dissipation rates. This suggests that the fully coupled case may be able to develop greater spatial heterogeneity than the non-coupled counterparts, perhaps promoting surface disruption. Second, we explored homogeneous temperature structures to examine the influence of spatially varying tidal dissipation. We ran simulations with a range of temperature distributions, including uniform and spatially-averaged tidal-flexing amplitude, to understand the effect of temperature variations on tidal dissipation.

Next, we explored the effect of a weak zone—representing a pre-existing fracture—on the coupled convection and tidal dissipation. Our results showed that localized heating near fractures may be important for the formation of tectonic features on Europa, such as ridges. A weak zone can lead to lithospheric thinning and regulate internal circulation patterns within the ice shell. This underscores the significance of incorporating fracture zones into models of tidal heating and convection on Europa. Overall, our study highlights the importance of understanding the interplay between tidal dissipation and fracture zones in shaping the surface of Europa.

Please cite this article in press as: Han, L., Showman, A.P. Coupled convection and tidal dissipation in Europa’s ice shell. Icarus (2010), doi:10.1016/j.icarus.2009.12.028
Maxwell time is close to tidal period, consistent with previous estimates of tidal dissipation (Kirk and Stevenson, 1987; Tobie et al., 2003; Mitri and Showman, 2008). Under grain boundary sliding (GBS) rheology with an ice grain size of 0.1 mm or 1 mm, our results show that the tidal dissipation rate is comparable to that obtained from models using diffusion rheology with a melting-temperature viscosity of $10^{13}$ Pa s, or $10^{14}$ Pa s, respectively. The tidal dissipation rate in Europa’s ice shell is about two orders-of-magnitude smaller if dislocation creep rheology is implemented in our model.

A heterogeneous temperature structure has a strong impact on the tidal dissipation rate. The tidal dissipation rate depends not only on local temperature, but also on surrounding temperature. Our simulations show that tidal dissipation rate in a convective plume encased in a different background temperature depends on both plume and background temperature, consistent with the results of Mitri and Showman (2008). For a given plume temperature, the dissipation rate in the plume decreases as the background temperature is increased. The dissipation rate is smaller than the homogeneous prediction (Eq. (7)) when the plume is cold relative its surroundings. On the other hand, tidal dissipation rate in a plume of higher temperature is significantly enhanced compared with the results from the homogeneous prediction.

Second, we performed numerical simulations of coupled convection and oscillatory tidal flexing to study the interrelationship between tidal dissipation and the thermal evolution of Europa’s ice shell. The tidal dissipation rate based on a simple zero-dimensional isothermal Maxwell model (see Eq. (7)) exhibits qualitative differences from the self-consistently calculated dissipation rate; the former underestimates the horizontal variation of tidal dissipation, and such spatial variations could play an important role in the geophysics of the ice shell.

Nevertheless, the temperature dependences indicate that for small grain size (corresponding to $\eta_0$ less than $10^{13}$ Pa s), the peak dissipation in a convecting ice shell will occur in cold plumes and at the base of the stagnant lid (where temperature is about 250 K). In contrast, for large grain size (corresponding $\eta_0$ larger than $10^{14}$ Pa s), the peak dissipation will occur in warm plumes and near the base of the ice shell. These differences have implications for
geophysical feedbacks depending on the grain size (Barr and Showman, 2009).

Third, our simulations demonstrate that pre-existing fracture zones or weak zones (at least ~6 km deep) promote the concentration of tidal dissipation. The tidal dissipation rate is increased five times or more at the weak zone (Fig. 10a). Upwelling plumes reach shallower depth at the location of the weak zone, leading to enhanced lithospheric thinning. Melting temperature is reached within certain locations of the upwelling plume. This supports the idea that enhanced heating along pre-existing fractures may be important for the formation of tectonic features (e.g., ridges).

As we noted earlier, the purpose of this study is to test the influence of heterogeneous temperature structures on the tidal-heating patterns, and explore the effects that such dissipation has on the convective structure in fully coupled models. Our study here is not intended to describe the detailed convection patterns in Europa's ice shell, for which more detailed models will be required in the future.

In our coupled thermal convection and tidal oscillation process models, we implemented 2D cartesian geometry by assuming no variation in temperature and tidal oscillation perpendicu- lar to the plane of the simulation. This is the sufficient for idealized tests, especially of the efficacy of the localization of tidal heating in gener- ating European ridges. However, 3D geometry is needed to distin- guish cylindrical plume structures from linear structures such as ridges. Furthermore, 3D models will allow a better representation of the cyclical tidal stresses, which involve tension, compression, and shear that may have any orientation of pre-existing fractures.

In the future, fully coupled convection and tidal oscillation process models should be studied in models with 3D geometry.

Our coupled models neglect the impact of melting on viscosity and effects of latent heating of melting. Melting causes viscosity to drop, and allowing larger strain rate (Dash et al., 1995; Duval, 1977; De La Chapelle et al., 1997), which in turn influences the ti- dal dissipation rate. Latent heating of melting buffers the temper- ature increase within the Europa's ice shell. Future studies should include these effects.

Although our coupled convection/tidal-oscillation models assumed a diffusion-creep rheology, the non-Newtonian grain boundary sliding (GBS) rheology may play an important role and should be explored in future work. Other areas for future investiga- tion include allowing the grain sizes to evolve, with growth rates depending on temperature and strain-rate conditions (cf. Barr and McKinnon, 2007), and incorporating plastic or damage rheo- logy into the simulations to allow for the effects of brittle deformation.

Acknowledgments

This work was supported by Grant NNX06AC41G from the NASA Planetary Geology and Geophysics Research Program to LH. APS was supported by P&G Grant NNX07AR27G.

References


