The impact of a weak south pole on thermal convection in Enceladus’ ice shell

Lijie Han a,⇑, Gabriel Tobie b, Adam P. Showman c

a Planetary Science Institute, 1700 E Fort Lowell, Suite 106, Tucson, AZ 85719, United States
b Laboratoire de Planétologie et Géodynamique, CNRS, UMR 6112, Université de Nantes, Nantes 44322, France
c Department of Planetary Sciences, Lunar and Planetary Laboratory, University of Arizona, Tucson, AZ 85721, United States

1. Introduction

Saturn’s moon Enceladus exhibits a diversity of heavily tectonized terrains, including a complex assortment of ridges, grooves, graben, and rifts that cover a substantial fraction of the surface (Porco et al., 2006; Kargel and Pozio, 1996; Squyres et al., 1983). Among the most striking regions is the South Polar Terrain (SPT), which contains the tiger stripes (Porco et al., 2006)—geologically young, ~130 km-long fractures exhibiting strongly elevated surface temperature. The average heat flux at Enceladus’ SPT reaches 100–200 mW m−2 (Spencer et al., 2006; Howett et al., 2011). Original estimates placed the heat flow of the SPT at 4–7 GW (Spencer et al., 2006), but this number has been recently revised to 16 ± 3 GW using Cassini data that extends to longer infrared wavelengths and hence can sense emission from lower temperatures (Howett et al., 2011). These heat flows greatly exceed the heat flow of 0.3 GW predicted by radiogenic heat (Porco et al., 2006; Schubert et al., 2007). Tidal dissipation, most likely from the Dione and Enceladus 2:1 resonance, has been suggested as a possible heat source (Ross and Schubert, 1989; Spencer et al., 2006; Meyer and Wisdom, 2008). Plumes of water vapor, ice, methane, carbon dioxide, nitrogen, ammonia, and other minor species emanate from the tiger stripes (Porco et al., 2006; Hansen et al., 2006; Brown et al., 2006; Waite et al., 2006, 2009; Spitale and Porco, 2007; Matson et al., 2007; Schneider et al., 2009; Kieffer et al., 2009; Postberg et al., 2009, 2011). The plumes are probably generated by evaporation or sublimation of liquid water, ice, or clathrates (Spencer et al., 2006; Kieffer et al., 2006; Postberg et al., 2011), powered by tidal dissipation along the tiger stripes or within the ice shell (Nimmo et al., 2007; Hurford et al., 2007, 2009; Smith-Konter and Pappalardo, 2008; Tobie et al., 2008).

Scientists have been puzzled by the tectonic contrast between Enceladus’ southern hemisphere, which is young and geologically active, and its northern hemisphere, which is heavily cratered and relatively ancient. What caused the formation of the tiger stripes and their erupting plumes? Why are the temperature and heat flux in the SPT so highly elevated?

Several authors have suggested that Enceladus’ tectonics may result, directly or indirectly, from convection in its interior (Nimmo and Pappalardo, 2006; Barr and McKinnon, 2007; Mitri and Showman, 2008; Roberts and Nimmo, 2008a,b; Bekoukova et al., 2010). Nimmo and Pappalardo (2006) suggested that a low-density diapir...
within the ice shell or putative silicate core could have caused a satellite reorientation, hence explaining the polar location of the active SPT. Large tectonic stress (over 10 MPa) could be generated by this reorientation, which could produce strong tectonic deformation (Nimmo and Pappalardo, 2006). Thermal convection can occur in Enceladus’ ice shell if the ice grain size is less than 0.1–0.3 mm (Barr and McKinnon, 2007; Mitri and Showman, 2008). Mitri and Showman (2008) described scenarios in which the shell could repeatedly switch between conductive and convective states. If an ocean exists, this would induce changes in the ice-shell thickness, thereby causing satellite volume changes, large stresses, and tectonic disruption. However, these studies could not show whether a large-scale and long-lived diapir underneath the SPT could be produced and maintained, which is a crucial requirement for reorientation to occur (cf. Nimmo and Pappalardo, 2006). Several groups have attempted to understand convection, high heat flux, and geological activity at Enceladus’ SPT. Grott et al. (2007) suggested that degree-one convection appears only if Enceladus’ core radius is less than 100 km and Enceladus is not fully differentiated. Stagnant-lid convection without concentrated tidal heating in the near surface cannot explain the high heat flux at Enceladus’ SPT, because the thick stagnant lid limits the near-surface thermal gradient (Barr and McKinnon, 2007; Barr, 2008; Roberts and Nimmo, 2008a; Mitri and Showman, 2008; Bekounkova et al., 2010). For grain sizes of 0.1–0.3 mm, the estimated heat flux would be ~10 mW/m² (Barr and McKinnon, 2007), over 10 times smaller than the observed SPT heat flux (Spencer et al., 2006). Stegman et al. (2009) suggested that compositional convection may become a dominant process if enough ammonia is present in the ice shell, leading to polar wander and other tectonics on Enceladus. Previous work has shed light on the mechanisms for maintaining a high heat flux at the SPT. By enforcing a very small viscosity contrast (10²–10¹⁵) between the bottom and surface of the ice shell to represent brittle failure on the surface, Barr (2008) argued that convection at Enceladus’ south pole occurs in the “mobile lid” regime and that a heat flux comparable to the observed ~100 mW/m² can be produced. Stegman et al. (2009) suggested an even weaker lithosphere with a viscosity contrast of 10⁶–10¹⁵ in their thermo-compositional models. O’Neill and Nimmo (2010) suggested that the current activity could be explained if the convection is episodic, with short, ~10–Myr spikes of high heat flux interspersed between ~0.1 and 1 Gyr-long periods of lower heat flux. Heterogeneous tidal dissipation, especially strongly enhanced tidal dissipation in the south pole region, may also play an important role in driving tectonic deformation on Enceladus. Roberts and Nimmo (2008b) showed that localized shear heating enhances the heat flux in Enceladus’ south polar region, although their imposed tidal heating is much larger than the maximum steady-state value estimated by Meyer and Wisdom (2007, 2008). The existence of a south polar sea (Collins and Goodman, 2007; Schubert et al., 2007) could help to promote tectonism in the overlying ice shell. Tobie et al. (2008) implemented a viscoelastic model to simulate the response of Enceladus to tidal oscillation. They showed that a subsurface liquid layer is required for strong enhancement of tidal dissipation and heat flux at the SPT.

However, previous three-dimensional spherical models of thermal convection in Enceladus’ ice shell exhibit convective patterns that are highly symmetrical about the equatorial plane and therefore fail to explain the hemispheric dichotomy in tectonics, heat flux, and ages (Roberts and Nimmo, 2008a; Bekounkova et al., 2010). Here we present three-dimensional spherical numerical simulations to explain how a hemispheric dichotomy of thermal convection, surface heat flux, and tectonics can arise in Enceladus’ ice shell. In particular, we demonstrate that an imposed mechanical weakening of the ice shell underlying the SPT can lead to an enormous hemispheric dichotomy in heat flux—similar to that observed—with important implications for tectonics. This provides a possible explanation for how the SPT can exhibit such a high heat flux despite presumably much smaller heat fluxes elsewhere on the satellite. In Section 2, we describe the model. In Section 3, we present numerical simulations to show how a regional-scale weakening affects the behavior, and we compare these results to models without such regional weakening. In Section 4, we conclude and discuss the implications.

2. Models and methods

2.1. Models and parameters

We study the problem of thermal convection with basal heating and tidal heating in global, three-dimensional (3D) spherical geometry with parameters appropriate to Enceladus’ ice shell. We neglect inertia and adopt the Boussinesq approximation. The governing dimensionless momentum, continuity, and thermal-energy equations are respectively given by

\[ \frac{\partial D_T}{\partial t} + u \cdot \frac{\partial D_T}{\partial x_i} = \frac{\partial^2 T}{\partial x_i^2} + q^i, \]  

where \( q^i \) is the stress tensor, \( u \) is velocity, \( T \) is temperature, \( q \) is internal tidal heating rate, \( k \) is the vertical unit vector, \( t \) is time, \( x_i \) and \( x_j \) are the spatial coordinates, and \( i \) and \( j \) are the coordinate indices. All the variables are dimensionless. Repeated spatial indices imply summation.

The Rayleigh number \( Ra \) is given by

\[ Ra = \frac{g \rho \alpha \Delta T D^3}{\kappa \eta_0}, \]

where \( g \) is gravity, \( \rho \) is density, \( \alpha \) is thermal expansivity, \( \Delta T \) is the temperature drop between the bottom and top boundaries, \( D \) is the depth of the system, \( \kappa \) is the thermal diffusivity, and \( \eta_0 \) is the reference viscosity at the melting temperature. The model parameters are presented in Table 1.

We use the finite-element code CitcomS (Zhong et al., 2000) to solve the problem in global, 3D spherical geometry. Rather than using a longitude–latitude grid, the model divides the sphere into 12 diamond-shaped “caps” of approximately equal area, each of which is further subdivided into a three-dimensional, non-orthogonal grid (see Zhong et al. (2000) for further details). Our models implement a resolution of 49 × 49 × 49 finite elements within each cap, corresponding to a grid size of approximately 1.5° in longitude.

<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>0.114 m s⁻¹</td>
</tr>
<tr>
<td>Density</td>
<td>( \rho )</td>
<td>930 kg m⁻³</td>
</tr>
<tr>
<td>Thermal expansivity</td>
<td>( \alpha )</td>
<td>1.5 × 10⁻⁵ K⁻¹</td>
</tr>
<tr>
<td>Thermal diffusivity</td>
<td>( \kappa )</td>
<td>1.5 × 10⁻⁴ m² s⁻¹</td>
</tr>
<tr>
<td>Specific heat</td>
<td>( c_p )</td>
<td>2150 J kg⁻¹ K⁻¹</td>
</tr>
<tr>
<td>Surface temperature</td>
<td>( T_s )</td>
<td>73 K</td>
</tr>
<tr>
<td>Bottom temperature</td>
<td>( T_b )</td>
<td>273 K</td>
</tr>
<tr>
<td>Angular frequency</td>
<td>( \omega )</td>
<td>2 × 10⁻⁷ s⁻¹</td>
</tr>
<tr>
<td>Young’s modulus</td>
<td>( E )</td>
<td>10¹² Pa</td>
</tr>
<tr>
<td>Activation energy</td>
<td>( Q )</td>
<td>60,000 J mol⁻¹</td>
</tr>
<tr>
<td>Ice grain size</td>
<td>( d )</td>
<td>0.1–2 mm</td>
</tr>
<tr>
<td>Thickness of ice shell</td>
<td>( D )</td>
<td>70 or 100 km</td>
</tr>
</tbody>
</table>
and latitude per finite element. Although the grid is non-orthogonal and requires careful treatment at the boundaries between the caps, it has the major advantage that it lacks the pole problems associated with longitude–latitude grids.

Because it is uncertain whether a subsurface ocean exists in Enceladus’ ice shell, we run different simulations with and without a subsurface ocean. In the models with a subsurface ocean, we implement a free-slip boundary condition at the bottom. Temperature is fixed on the bottom of the ice shell with a temperature of 273 K. In the models without a subsurface ocean, we use a no-slip velocity boundary condition at the bottom. If the rocky core is cold, the temperature at the bottom of the ice shell can be as low as 200–220 K (Schubert et al., 2007); therefore, in these models, we again fix the lower-boundary temperature and explore values ranging from 200 to 260 K. In both types of models, a free-slip velocity boundary condition and a temperature of 73 K is imposed at the surface. The initial temperature decreases linearly from the inner to the outer surface, with an initial disturbance of amplitude 20 K to start the convection. For simplicity, we do not consider cases with a localized (non-global) subsurface ocean.

While the ice grain sizes are unknown for Enceladus, estimates have suggested 0.1–1 mm within icy satellite interiors (Kirk and Stevenson, 1987; Barr and McKinnon, 2007). We vary the ice-grain size from 0.1 to 2 mm in our simulations to bracket the plausible range. Depending on the exact regime of stress and grain size, the expected flow regime is (Newtonian) diffusion creep or mildly non-Newtonian grain-size-sensitive creep (such as grain-boundary sliding creep) (Durham and Stern, 2001; Goldsby and Kohlstedt, 1997, 2001). Here, we adopt a Newtonian, temperature-dependent viscosity as follows:

\[
\eta(T) = \eta_\text{ref} \min \left\{ \Delta \eta_\text{ref} \exp \left[ \frac{A(1 - 1)}{T} \right] \right\},
\]

where \( \eta_\text{ref} \) is the reference viscosity at the melting temperature, varying from \( 5 \times 10^2 \) Pa s to \( 2 \times 10^5 \) Pa s (corresponding to ice grain size of 0.1–2 mm), \( A \) is a rheological parameter with a value of 26 (corresponding to an activation energy of 60 kJ mol\(^{-1}\)), and \( T \) is the nondimensional temperature (actual temperature divided by 273 K). \( \Delta \eta \) is the dimensionless viscosity contrast cutoff, as discussed further below.

For the present study, we neglect pre-melting, which could lead to a viscosity drop and greater strain rates near the melting temperature than are captured in Eq. (5) (e.g., Dash et al., 1995; Duval, 1977; De La Chapelle et al., 1999; Tobie et al., 2003). For simplicity, we also neglect the possible dynamic growth of ice grains in Enceladus’ ice shell (Barr and McKinnon, 2007; Tobie et al., 2006).

### 2.2. Weak south pole

The Tiger stripes are the predominant tectonic features within Enceladus’ SPT. The plumes, high surface temperature and heat flux along the Tiger Stripes indicate that the SPT is undergoing active geological processes. Enceladus’ South Polar Terrain exhibits significant tectonic disruption, suggesting mechanically weak behavior. Enceladus’ SPT has experienced a complex tectonic history including folding (Barr and Preuss, 2010) and fracturing (Patthoff and Kattenhorn, 2011). The folds may have formed under conditions similar to those producing lava flow on Earth (Barr and Preuss, 2010). Producing the observed surface disruption in response to the (small) convective and tidal stresses requires that the ice at the SPT is weak; for example, in their models attempting to explain SPT tectonics, Barr (2008) invoked an effective viscosity contrast between the surface and the deeper, warmer ice of \( 10^{-2} - 10^{2.5} \) and Stegman et al. (2009) adopted an even lower viscosity contrast of \( 10^{-8} - 10^{-15} \). Barr (2008) discussed the importance of brittle deformation in Enceladus’ South Polar Terrain and used two-dimensional numerical simulations and scaling laws to determine the impact on heat flux and geological age. Unfractured ice at Enceladus’ 70-K surface temperature has a viscosity at least \( 10^{10} \) times that at the melting temperature. The high viscosity near the surface would preclude the surface layers from participating in the convection, thus leading to stagnant-lid convection (Solomatov, 1995; Moresi and Solomatov, 1995). The use of viscosity contrasts \( 10^{-2} - 10^{2.5} \) in Barr’s (2008) simulations was intended as a simple means of allowing near-surface deformation, under which condition, a mobile-lid convection regime occurs. Such an approach has previously been successfully used to investigate the effect of convection on Europa tectonics (Showman and Han, 2004).

Brittle/plastic rheology has been widely implemented in convection models of Earth and is making an entrance into the icy-satellite literature. One set of models adopts strain-rate or strain softening rheologies and attempts to self-consistently generate brittle or semi-brittle behavior from the simulations (e.g., Bercovici, 1993, 2003; Moresi and Solomatov, 1998; Tackley, 2000a,b; Showman and Han, 2005). Perhaps the simplest such model is plastic rheology, which allows deformation only for deviatoric stresses exceeding a specified yield stress \( \sigma_Y \) (and because increases in strain rate are envisioned to be accommodated by increased slip on fractures with minimal stress increase, the deviatoric stress in plastic rheology cannot exceed the yield stress). For stresses less than \( \sigma_Y \), the temperature-dependent viscosity dominates and an immobile stagnant lid would form at the surface, but for stresses exceeding \( \sigma_Y \), plastic deformation would occur, and the convection would be forced away from the stagnant-lid regime. Showman and Han (2005) applied this approach to icy satellites; they showed for Europa-like conditions that mobile-lid convection can occur when the yield stress is a fraction of a bar or smaller, leading to significant disruption of the surface, which in some cases can be highly episodic. O’Neill and Nimmo (2010) explored a similar brittle or semi-brittle model for Enceladus. The advantage of these models are that they do not presuppose any particular spatial configuration for the weakening. However, the solutions for these models are very unstable due to the non-linear effects of the plasticity (Moresi and Solomatov, 1998).

Another class of models explicitly imposes weak zones (such as faults or subduction zones) within the system and investigates how these specified weak zones interact with the fluid flow (Zhong et al., 1998; Zhong and Gurnis, 1994; Han and Gurnis, 1999, and others) but avoids the question of how the weak zones were generated to begin with. In this type of model, a fault or weak zone is typically parameterized by introducing a smaller viscosity contrast cutoff in the desired region. Given the uncertainties about how Enceladus’ SPT was formed, we implement a weak south pole region by reducing the viscosity-contrast cutoff poleward of either 45° or 60° latitude. This allows us to determine how mechanical weakening influences the convective patterns and resulting heat flux. At the same time, a strong viscosity contrast cutoff (larger than \( 10^6 \)) is implemented everywhere northward of either 45° or 60° latitude.

We present two set of models here. For the first set of models, we do not include a weak south pole. A uniform viscosity contrast cutoff of \( 10^{2} - 10^{10} \) due to temperature variation was implemented. This set of models is comparable to previous 3D spherical numerical models of thermal convection in Enceladus’ ice shell (Roberts and Nimmo, 2008a; Bekoukova et al., 2010). For the second set of models, we include a weak south pole by implementing a viscosity cutoff of \( 10^{-2} - 10^{3} \). This set of models allows us to investigate whether a weak south pole influences thermal convection, tectonics, and the heat flux on Enceladus.
2.3. Tidal dissipation

The viscoelastic tidal deformation is computed in the frequency domain following the procedure described in Tobie et al. (2005). A Maxwell compressible rheology, characterized by the elastic shear modulus \( \mu \), the elastic bulk modulus \( K_e \) and the Newtonian viscosity \( \eta \), is assumed. In the frequency domain, the stress–strain constitutive relationship can be written in the form of a complex Hooke-like law:

\[
\sigma_y = 2\mu(\omega)\hat{\varepsilon}_y + \left[ K - \frac{2}{3} \mu(\omega) \right] \hat{\varepsilon}_y, \tag{6}
\]

where

\[
\hat{\mu}(\omega) = \frac{\mu_0 \omega^2 \eta^2}{\mu_0^2 + \omega^2 \eta^2} + i \left[ \frac{\mu_0 \omega^2 \eta^2}{\mu_0^2 + \omega^2 \eta^2} \right] \tag{7}
\]

and \( \hat{\sigma}_y \) and \( \hat{\varepsilon}_y \) the Fourier transform of the stress and strain tensor components, and \( \hat{\mu}(\omega) \) the complex shear modulus in the frequency domain.

Using the correspondence principle established by Biot (1954), the viscoelastic solutions can be determined by solving the equivalent elastic spherical oscillation problem initially developed by Al'terman et al. (1959) and Takeushi and Saito (1972) in the frequency domain and by imposing the tidal potential as the source of excitation for the spherical oscillation. In the frequency domain, the tidal potential at the surface of a satellite in spin–orbit 1:1 resonance (e.g., Moore and Schubert, 2000; Tobie et al., 2005) is:

\[
\Phi_{\text{tid}}(R, \omega) = R_0^2 \omega^2 e^{-\frac{3}{2}} \left( \frac{2}{P_j^2(\cos \theta) - \frac{1}{2} P_j^2(\cos \theta) \cos 2\phi} \right) \left[ \frac{1}{2} P_j^2(\cos \theta) \sin 2\phi \right], \tag{8}
\]

where \( R_0 \) is the surface radius, \( \omega \) is the orbital angular frequency, \( e \) is the orbital eccentricity, \( P_j(\cos \theta) \) and \( P_j^2(\cos \theta) \) are the associated Legendre polynomials, \( \theta \) and \( \phi \) are latitude and longitude coordinates.

The present formulation is valid only for radially layered internal models, and lateral viscosity variations associated with thermal convection cannot be explicitly included, contrary to the method employed in Tobie et al. (2008) and Bekounkova et al. (2010). However, the effect of lateral viscosity variations on the tidal dissipation field can be included by considering their local effect on the specific dissipation function, \( Q^{-1} \). Assuming that dissipation is only associated with shear motions, the specific dissipation function can be estimated locally from the ratio between the imaginary part of the complex shear modulus and its modulus:

\[
Q^{-1} = \frac{\Im(\hat{\mu}(\omega))}{|\hat{\mu}(\omega)|} = \frac{\mu}{\eta \omega} \times \left[ 1 + \frac{\mu_0^2 \omega^2}{\mu^2} \right]^{-1/2}. \tag{9}
\]

The local volumetric tidal dissipation rate, \( \eta_{\text{tid}} \), can then be obtained from the 3D viscosity field by multiplying the strain energy function (corresponding to the product of the stress and strain tensor components) and the specific dissipation function, \( Q^{-1} \):

\[
\eta_{\text{tid}}(r, \theta, \phi) = \frac{\mu}{\eta \omega} \times \left[ 1 + \frac{\mu_0^2 \omega^2}{\mu^2} \right]^{1/2} \times \frac{|\omega \bar{\sigma}_y(r, \theta, \phi) \times \hat{\varepsilon}_y(r, \theta, \phi)|}{2}. \tag{10}
\]

As \( \eta \omega > \mu \), i.e., \( \eta > 6 \times 10^{13} \text{ Pa s} \), \( |\omega \bar{\sigma}_y(r, \theta, \phi) \times \hat{\varepsilon}_y(r, \theta, \phi)| \) is constant and does not vary significantly with the local viscosity, \( Q^{-1} \) and hence \( \eta_{\text{tid}} \) are inversely proportional to \( \eta \). Hence, Eq. (10) can then be reduced to:

\[
\eta_{\text{tid}}(r, \theta, \phi) = \frac{\mu}{\eta \omega |\Gamma(r, \theta, \phi)|} \times \frac{|\omega \bar{\sigma}_y(r, \theta, \phi) \times \hat{\varepsilon}_y(r, \theta, \phi)|}{2}. \tag{11}
\]

Neglecting the lateral variations of viscosity when computing \( |\omega \bar{\sigma}_y(r, \theta, \phi) \times \hat{\varepsilon}_y(r, \theta, \phi)| \) remain a valid approximation.

For \( \eta \omega < \mu \), the situation is more complex. \( |\omega \bar{\sigma}_y(r, \theta, \phi) \times \hat{\varepsilon}_y(r, \theta, \phi)| \) increases as a function of decreasing viscosity, while \( Q^{-1} \) decreases. Only a full 3D formulation of the viscoelastic problem can correctly assess these two terms (Bekounkova et al., 2010). As demonstrated by Bekounkova et al. (2010) and Han and Showman (2010), the stress and strain fields, and therefore the strain energy function, can be significantly affected around regions with small viscosities (\( \eta \leq 6 \times 10^{13} \text{ Pa s} \)). The formulation based on radial functions (Tobie et al., 2005) starts to diverge from the exact 3D solution.

For simplicity, we adopt here the simplified Eq. (11) to compute the dissipation rate. This approximation is justified in most of the simulations where \( \eta > 6 \times 10^{13} \text{ Pa s} \). It is only problematic for

### Table 2

<table>
<thead>
<tr>
<th>Layer</th>
<th>b (km)</th>
<th>( \rho ) (kg m(^{-3}))</th>
<th>( \mu_e ) (GPa)</th>
<th>( K_e ) (GPa)</th>
<th>( \eta ) (Pa s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rocky core</td>
<td>149–160</td>
<td>3625–4370</td>
<td>58–70</td>
<td>155–187</td>
<td>10(^{23})</td>
</tr>
<tr>
<td>Internal ocean</td>
<td>1–20</td>
<td>1000</td>
<td>0</td>
<td>2.25</td>
<td>0</td>
</tr>
<tr>
<td>Ice shell</td>
<td>70–100</td>
<td>920</td>
<td>3.7</td>
<td>10</td>
<td>10 Eq. (5)</td>
</tr>
</tbody>
</table>

Fig. 1. Temperature structure from a full 3D spherical model showing thermal convection in Enceladus' ice shell with a Rayleigh number of 6.5 \times 10^5. The simulation implements a temperature-dependent viscosity contrast of 10^10. To depict the 3D temperature structure, the top panel displays the shape of an isothermal surface of 216 K. The bottom panel displays the nondimensional temperature distribution (i.e., temperature divided by 273 K) along a radial cross-section passing through the poles. Multiple plumes develop at the bottom of the ice shell beneath a thick stagnant-lid. Convection show no north–south asymmetry.
simulations with the highest Rayleigh number ($Ra = 2 \times 10^8$, corresponding to $\eta_0 = 6 \times 10^{12} \text{ Pa s}$). As we will show, however, our results are insensitive to the details of the tidal-heating formulation, because tidal heating constitutes only a small part of the heat budget for the models considered here, especially at high Rayleigh number. Tidal heating rate in the ice shell is very small in the cases without a subsurface ocean, so we neglect tidal dissipation in our thermal convection simulations for this case.

To compute the viscoelastic response to tidal forcing in the cases with a subsurface ocean, the interior is assumed to be differentiated into a rocky core, an internal ocean and an ice shell. Table 2 summarizes the parameters of the different internal layers used to compute the solutions. The internal ocean has a thickness of 5 km or 30 km. A model with 48 layers of constant viscosity in ice shell to simulate the temperature-dependent viscosity profile along the radius direction in the simulations.

3. Results

3.1. No mechanically weak south pole

When the viscosity contrast cutoff is the same everywhere, the convection in our models does not exhibit a hemispheric dichotomy. Fig. 1 shows the temperature structure from a model with a Rayleigh number of $6.5 \times 10^7$, an ice shell thickness of 100 km, ice grain size of 0.3 mm ($\eta_0 = 3.2 \times 10^{13} \text{ Pa s}$), and a viscosity contrast cutoff of $10^{10}$. As expected, stagnant-lid convection occurs in the ice shell. Multiple plumes develop in the lower layer, but the overall convective pattern is similar in the northern and southern hemispheres. Fig. 2a and b shows the tidal dissipation distribution at the bottom and 33-km depth of the model. The tidal dissipation rate in the bottom of the ice shell is about $10^{-7} \text{ W m}^{-3}$ and the dissipation rate at a depth of 33 km is several orders of magnitude less ($\sim 10^{-15} \text{ W m}^{-3}$). The dissipation rate is so small as to have no real impact on the surface heat flux. The surface heat flux is very low in the simulation, about 5–10 mW m$^{-2}$. The areally integrated heat transport in the SPT is about 0.5 GW, much less than the observationally inferred values of $\sim 10$ GW or more (Spencer et al., 2006; Howett et al., 2011). These results are consistent with the previous 3D spherical numerical simulation results (Roberts and Nimmo, 2008a; Bekounkova et al., 2010).

When the viscosity contrast is sufficiently small, the convection can occur in a mobile or sluggish-lid regime but still lacks a hemispheric dichotomy. Fig. 3 shows temperature from a model with a viscosity contrast cutoff of $10^4$ and, as before, a Rayleigh number of $6.5 \times 10^7$. Two big plumes develop in the convection shell, with one plume at each pole, respectively. Four smaller plumes appear in the equatorial region. The convection patterns are very different from the multiple-plume stagnant-lid convection patterns in Fig. 1.

![Image of temperature distribution](image_url)

Fig. 2. Spatial distribution of the tidal dissipation rate for four different cases once they have reached an equilibrated state. Each row shows a different model; for each row, the left and right panels show $\log_{10}$ of the tidal dissipation rate ($\text{W m}^{-3}$) at the bottom and a depth of 33 km, respectively. All four models use an ice-shell thickness of 100 km and adopt an orbital eccentricity of 0.0045 (the current value) in the tidal dissipation calculation. (a and b) No southern hemisphere weak zone; $Ra = 6.5 \times 10^7$ and the viscosity contrast cutoff is $10^{10}$ everywhere. (c and d) No southern hemisphere weak zone; $Ra = 6.5 \times 10^7$ and the viscosity contrast cutoff is $10^6$ everywhere. (e and f): $Ra = 5 \times 10^6$; southern hemisphere weak zone (southward of 60°S latitude) imposes a viscosity contrast cutoff of $10^3$ with a cutoff of $10^6$ elsewhere. (g and h) $Ra = 2.3 \times 10^7$; southern hemisphere weak zone (southward of 60°S latitude) imposes a viscosity contrast cutoff of $10^3$ with a cutoff of $10^6$ elsewhere.
Still, with a globally homogeneous lithospheric strength, the convection patterns do not show a hemispheric dichotomy and fail to explain the locally enhanced tectonism and heat flux at the SPT. Again, the tidal dissipation rate is too small to have a strong impact on the surface heat flux (see Fig. 2c and d).

Simulations without a subsurface ocean also produce symmetrical convection patterns in Enceladus’ ice shell. We integrated models with a no-slip velocity boundary condition and a fixed temperature of 250 K or 260 K at the bottom of the ice shell. For a viscosity contrast of $10^{10}$, the convection exhibits a pattern similar to that shown in Fig. 1, with multiple plumes underlying a thick stagnant lid. For a viscosity contrast of $10^{4}$, two big plumes appear in the ice shell, as shown in Fig. 3.

The above simulations show that the viscosity contrast cutoff can strongly impact the thermal-convection patterns. However, the thermal convection is generally symmetrical across the equatorial plane. No north/south dichotomy in convection develops.

### 3.2. Impact of mechanically weak south pole

We now demonstrate that including a mechanically weak SPT dramatically alters the convection patterns in Enceladus’ ice shell and naturally leads to a strong hemispheric dichotomy in the surface heat flux. Fig. 4 shows results from a model with a maximum viscosity contrast of $10^{2}$ south of 60°S latitude but $10^{6}$ everywhere else. The Rayleigh number is $5 \times 10^{6}$ and ice-grain size is 1 mm ($\eta_{0} = 4.2 \times 10^{14}$ Pa s). From Fig. 4, we can see that a single robust upwelling plume develops in the south polar region, while no convection occurs in the northern hemisphere because of the low Rayleigh number. The tidal dissipation rate exhibits a strong dichotomy in this model (Fig. 2e and f), however, the dissipation rate is too small to have a significant effect on the surface heat flux.

**Fig. 3.** Temperature structure from a full 3D spherical model showing thermal convection in Enceladus’ ice shell with a Rayleigh number of $6.5 \times 10^{6}$. The simulation implements a temperature-dependent viscosity contrast of $10^{4}$. The top panel displays the shape of the 216-K isotherm. The bottom panel displays the nondimensional temperature distribution along a radial cross-section passing through the poles. Two large plumes are evident, one in the southern hemisphere and the other in the northern hemisphere. No dichotomy develops between the northern and southern hemispheres.

**Fig. 4.** Temperature profile from a full 3D spherical model showing thermal convection in Enceladus’ ice shell with a Rayleigh number of $5 \times 10^{6}$. The simulation implements temperature-dependent viscosity contrast of $10^{6}$, except a viscosity contrast of $10^{2}$ southward of 60°S latitude. The top panel displays the shape of the 216-K isothermal surface. The bottom panel displays the nondimensional temperature distribution along a radial cross-section passing through the poles. A convecting plume develops at the south pole, but no convection occurs outside the SPT.
Changing the Rayleigh number alters the details of the convection, but a strong hemispheric dichotomy in convection persists as long as a weak SPT is included. Fig. 5 shows the results from a model with a Rayleigh number of $2.3 \times 10^7$ and an ice-grain size of $0.4 \text{ mm}\left(\eta_0 = 9 \times 10^{13} \text{ Pa s}\right)$. A maximum viscosity contrast of $10^6$ was adopted except, as before, southward of 60°S latitude, where the viscosity contrast was $10^7$. In this case, a single upwelling plume ascends toward the surface in the south polar region. Convection occurs in the northern hemisphere under a thick stagnant lid, with upwellings and downwellings at the bottom of the ice shell. At even higher Rayleigh numbers ($Ra = 2.1 \times 10^8$, corresponding to an ice-grain size of 0.1 mm, $\eta_0 = 6 \times 10^{12} \text{ Pa s}$, not shown), the SPT exhibits chaotic, mobile-lid convection, with stagnant-lid convection occurring elsewhere.

In either case with a lower Rayleigh number (Fig. 4) or higher Rayleigh number (Fig. 5), the thermal convection pattern exhibits a strong dichotomy between the northern and southern hemispheres. We emphasize that our simulations contain no imposed thermal boundary asymmetries nor asymmetries in the imposed tidal-heating formulation; rather, the dichotomy in convection develops solely from the imposed dichotomy in lithospheric strength.

The mechanically weak south pole also causes a strong dichotomy of surface heat flux in our models. Fig. 6 displays the heat flux from a model with a Rayleigh number of $2 \times 10^6$, corresponding to an ice grain size of 0.1 mm. The surface heat flux can be strongly elevated in the southern hemisphere (Fig. 6), while the average heat flux in the northern hemisphere remains low at about $10 \text{ mW m}^{-2}$. Early in our simulations, the heat flux within the SPT peaks at values in some cases exceeding $200 \text{ mW m}^{-2}$ (Fig. 6b), implying a total integrated power of 10 GW across the SPT, comparable to the observed power (Spencer et al., 2006; Howett et al., 2011). The surface heat flux decreases modestly as time goes on, reaching equilibrium values in our models as high as 60–200 mW m$^{-2}$ (Fig. 6e), equivalent to an integrated power across the SPT of ~4 GW. This is close to the value inferred by Spencer et al. (2006) but remains a factor of 2–3 less than that inferred by Howett et al. (2011). Fig. 6c and f show the radially integrated tidal heating, illustrating that tidal heating is only a small fraction of the total heat budget.

Fig. 7 summarizes how the global and SPT heat fluxes vary with Rayleigh number and lithospheric strength. When no southern-hemisphere weak zone is used, the global heat flow is 2.7 GW, while the heat flow in the SPT is only 0.2 GW (Fig. 7a). This value is several dozen times less than observational estimates (Spencer et al., 2006; Howett et al., 2011). By implementing a weak south pole, however, the heat flow in SPT increases to 1 GW or more, depending on the Rayleigh number and viscosity contrast in the SPT. At a constant Rayleigh number, smaller viscosity cutoff values within the SPT lead to greater SPT heat fluxes; specifically, when the SPT viscosity contrast is $10^2$, the heat flux is ~50–70% greater than when the SPT viscosity contrast is $10^3$. This can be seen by comparing Fig. 7b and c, where the steady-state SPT heat flux increases from ~1 to 1.5 GW as the SPT viscosity contrast decreases from $10^3$ to $10^2$ at a constant Rayleigh number of $6.5 \times 10^7$, or by comparing Fig. 7e and f, where the steady-state SPT heat flux increases from ~2.2 to 3.8 GW as the SPT viscosity contrast decreases from $10^3$ to $10^2$ at a constant Rayleigh number of $2 \times 10^6$. Likewise, Fig. 7 shows that, for a constant viscosity cutoff in the SPT, the SPT heat flux increases with Rayleigh number. For example, with SPT viscosity contrast cutoff of $10^2$, the SPT surface heat flux increases by a factor of ~2.5 (from ~1.5 to 3.8 GW) as the Rayleigh number is increased from $6.5 \times 10^7$ to $2 \times 10^8$.

In all the models presented so far, the weak zone was implemented southward of 60°S latitude. To test the effects of the SPT’s size on convection and tectonics, we ran some simulations with a wider weak zone extending from 45°S latitude to the south pole. Qualitatively, the results remain unchanged in the two cases, although the case with the wider weak zone initiates convection at a slightly lower Rayleigh number (i.e., with a slightly larger ice-grain size). Nevertheless, because the region of intense tectonism and high heat flux on Enceladus is confined primarily southward of 60°S, the simulations with the smaller weak zone (southward of 60°S latitude) may represent Enceladus more accurately.

The thickness of Enceladus’ ice shell is uncertain and could range from ~40 km (Olgin et al., 2011) to a maximum of ~100 km. The minimum ice shell thickness for convection to occur is ~40 km (Barr and McKinnon, 2007). To investigate the effect of the ice-shell thickness, we ran a few simulations with an ice shell thickness of 70 km. Decreasing the shell thickness from 100 to 70 km decreases the Rayleigh number by a factor of three if other parameters remain the same, implying that the convective vigor and heat flux decrease correspondingly.

Thermal evolution models have shown that if the rocky core remains sufficiently cold, then no subsurface ocean exists in
Enceladus’ ice shell, with temperatures at the bottom of ice shell as low as 200–220 K (Schubert et al., 2007). We ran a series of simulations considering a cold ice shell without a subsurface ocean. A no-slip velocity bottom boundary condition, a weak south pole with a viscosity contrast of $10^1/C_0^3$, and a fixed bottom temperature of 200 K, 220 K, 250 K, or 260 K were used in different simulations. As in models with a subsurface ocean, stagnant-lid convection or no convection at all occurs in all regions in the northern hemisphere. Thermal convection can occur in Enceladus’ south pole if the ice grain size is 0.5 mm and bottom temperature is higher than 250 K. If the bottom temperature is 220 K, thermal convection can occur in SPT only if the ice grain size is as small as 0.1 mm. However, if the bottom of the ice shell is colder than 200 K, no convection can occur in Enceladus’ SPT even with a small ice grain size of 0.1 mm. In summary, convection can occur in Enceladus’ SPT if the bottom of the ice shell is warmer than 220 K and the ice-grain size is sufficiently small.

Our simulations also provide an explanation for the strong age dichotomy between Enceladus’ young southern hemisphere and relatively ancient northern hemisphere. With a weak south pole, the surface velocities are greatly elevated over values for stagnant-lid convection with no weak SPT. The weaker the south-polar region (i.e., the smaller the SPT viscosity cutoff), the larger the SPT surface velocities. For a simulation with a Rayleigh number of $6.5 \times 10^5$ (corresponding to an ice-grain size of 0.3 mm) and an SPT viscosity contrast of $10^2$, the average horizontal velocity is 1 mm yr$^{-1}$ in the SPT. By decreasing the viscosity contrast cutoff in the SPT to $10^1$ and using the same Rayleigh number, the average horizontal velocity achieves values of 27 mm yr$^{-1}$. With a horizontal velocity of 10 mm yr$^{-1}$, the time for surface materials to circulate from the south pole to 60°S latitude is about 10 Myr. The intense strains accompanying this deformation should destroy the pre-existing surface and reset its apparent age. Thus, our models exhibit a surface age in the northern hemisphere exceeding $10^9$ years but a surface age in the SPT of $\sim 10^7$ years depending on parameters. (These estimates, obtained from direct numerical simulation, are consistent with Barr’s (2008) estimates of surface velocity and SPT age from scaling analysis using a viscosity contrast cutoff in the SPT of $10^2$–$10^3$.) Broadly, these predictions are consistent with the young age of Enceladus’ SPT as inferred from crater analysis (Porco et al., 2006) and, simultaneously, the ancient ages of $\sim 1$ Ga or more inferred for the heavily cratered northern terrains.

4. Conclusion and discussion

The striking hemispheric dichotomy of tectonism and heat flux on Enceladus—with intense, geologically young tectonic deformation and enhanced heat transport at high southern latitudes but relatively ancient terrains elsewhere—remains a major puzzle in icy-satellite science. We have shown that, in the absence any hemispheric asymmetries in heating or material strength, the convection (if any) does not exhibit a hemispheric dichotomy. However, we demonstrated that, in the presence of an imposed weakened lithosphere over the south pole, the convection in the ice shell naturally develops a strong hemispheric dichotomy: ascending convective plume(s) underlie the SPT and reach the near-surface regions, while stagnant-lid convection or no convection at all occurs in the northern hemisphere. This behavior naturally leads to a strong dichotomy in heat flux: in our models, the local peak heat flux at the SPT becomes enhanced by a factor of $\sim 10$–20 over that...
in surrounding regions and can reach \(\sim 200 \text{ mW m}^{-2}\), implying a heat transport integrated over the SPT of up to \(\sim 5 \text{ GW}\). The surface strains in our models imply a resurfacing age of \(\sim 10^6 - 10^7\) years at the SPT but \(> 10^9\) years elsewhere, consistent with the observed surface ages. We emphasize that our models do not include any north–south asymmetry in the imposed thermal boundary conditions or tidal-heating formulation; rather, the dichotomy of heat flux and tectonics in our models result solely from the imposed dichotomy of lithospheric strength. Therefore, our results help to naturally explain the observed hemispheric dichotomy of tectonics and heat flux as long as a mechanism for weakening the south-polar regions exists.

A subsurface ocean under Enceladus’ south pole region has been suggested by several authors (Collins and Goodman, 2007; Schubert et al., 2007; Tobie et al., 2008). The existence of a subsurface ocean can help explain the high heat flux in the SPT (Tobie et al., 2008). However, our simulations show that the tidal heating rate in the ice shell underlying the SPT is only \(\sim 0.5 \text{ GW}\) at the current orbital eccentricity, which is much lower than the typical SPT surface heat flux of several GW in our models. The heat lost due to thermal convection is so huge that it would result in a very rapid crystallization of the ocean if there is one. The subsurface ocean freezes within a few million years if the heat flux at the SPT is \(100 \text{ mW m}^{-2}\). This is consistent with previous findings that tidal heating in Enceladus is not strong enough to sustain a long-term ocean (Roberts and Nimmo, 2008a).

The latest heat flow estimates in Enceladus’ SPT from Cassini thermal data yield values over 10 GW (Howett et al., 2011), significantly larger than the original estimates of 4–7 GW (Spencer et al., 2006). While our models can easily explain an SPT heat transport of 1 GW (see Fig. 7), and even reach 3–5 GW in some cases for a short time period, none of our models produce SPT heat transports of 10 GW in steady state. It could be that the system is episodic, allowing occasional spikes in heat transport, which might allow the current high heat flux to be explained as an extreme event in a system with a lower time-mean flux. Such episodicity could occur if the eccentricity, hence tidal heating rate, are time variable. Alternately, they might occur if the convection itself is episodic, as suggested for example by O’Neill and Nimmo (2010). In this light, it is interesting that our models, initialized from a conductive initial temperature profile, all generate a sharp spike in SPT heat flux before settling into the steady state (Fig. 7), hinting that such spikes might generally accompany switches between conductive and convective regimes in an episodic system. Still, detailed modeling of episodic convection would be needed to test this hypothesis further.

Fig. 7. Surface heat flow versus time in our Enceladus models. Horizontal axis represents non-dimensional time. Vertical axis represents heat flow. Lines with green triangles represent global surface heat flow, and lines with red circles represent surface heat flow in the SPT. Different panels display results from models with different Rayleigh number or mechanical strength of Enceladus’ south pole terrain. (a) \(Ra = 6.5 \times 10^7\), viscosity cutoff is \(10^{10}\) in all the regions. (b) \(Ra = 6.5 \times 10^7\), viscosity cutoff is \(10^9\) in SPT. (c) \(Ra = 6.5 \times 10^7\), viscosity cutoff is \(10^8\) in SPT. (d) \(Ra = 5 \times 10^7\), viscosity cutoff is \(10^9\) in SPT. (e) \(Ra = 2 \times 10^8\), viscosity cutoff is \(10^8\) in SPT. (f) \(Ra = 2 \times 10^8\), viscosity cutoff is \(10^7\) in SPT. For all cases that have a weak SPT (all cases except (a)), the SPT is implemented southward of 60°S latitude, and the viscosity contrast is \(10^6\) outside the SPT.
Our models did not include the possibility of enhanced shear heating along the tiger stripes. Concentrated tidal heating can appear along the fractures and would strongly affect the surface tectonics. In some cases, runaway heating along faults is possible (Nimmo and Gaidos, 2002; Han and Showman, 2008, 2010). Roberts and Nimmo (2008b) have suggested that enhanced tidal heating just below the surface can help to explain the high heat flux at the SPT. In total, the observed SPT heat flux may be the sum of a broadly distributed heat flux component—like that modeled here—and localized surface heating along the tiger stripes. Future convection models may need to consider enhanced shear heating along the tiger stripes in addition to the convective heat transport in the subsurface.

Our models did not consider the compositional convection that may result from the density contrast between ice and other chemical species. With enough ammonia, compositional convection could become a dominant mechanism in Enceladus’ ice shell (Stegman et al., 2009). However, the composition of the ice shells of icy satellites are poorly constrained, and the viscosity as a function of temperature, stress, and grain size of ice–salt and ice–ammonia mixtures is less well understood than the viscosity of pure ice. Perhaps for this reason, only two convection modeling papers in the icy-satellite literature (Han and Showman, 2005; Stegman et al., 2009) consider thermo-compositional convection, and in those cases, ad hoc assumptions about the composition were required; the remainder of the convection literature has assumed pure water–ice interior structure and rheology. In the specific case of Enceladus, even though constituents other than water ice (including ammonia) were detected by Cassini, these contaminants may still exert a minor influence on the ice behavior. Cassini INMS data indicate that there is not more than 1% ammonia in the vapor plume (Waite et al., 2009). Salts have also been detected in the ejected icy particles, but their concentration is also of the order of 1% (Postberg et al., 2011). Surface infrared spectra (Brown et al., 2006) indicate also that the ice at Enceladus’ surface is one of the purest among the icy satellites. These concentrations of non-ice constituents are probably not sufficient to generate compositional convection. They can however exert an effect on thermal convection by controlling the viscosity of ice (notably by limiting the grain growth). Such an effect is indirectly included when considering different reference viscosity values as we have done. Because of these issues, we have only considered thermal convection here, in line with previous Enceladus modeling studies (Barr and McKinney, 2007; Barr, 2008; Roberts and Nimmo, 2008a,b; Mitri and Showman, 2008; O’Neill and Nimmo, 2010; Bekoukova et al., 2010).

The intense tectonism observed in the SPT provides a strong motivation for our use of a weakened lithosphere at high southern latitudes, and the fact that our models can explain the strong surface deformation, young surface ages, and high heat flux at those latitudes as a result of such weakening is encouraging. Nevertheless, this raises a chicken-and-egg question of what caused the weakening in those regions. One possibility is that the lithospheric strength and convective/heat-flux behavior are coupled and mutually self-generating: the numerous fractures associated with the intense tectonism may keep the ice underlying the SPT weak, thereby allowing continued tectonism and a high heat flux in that region. In this scenario, the weakening would develop naturally as a result of the intense tectonism—and vice versa. It should be possible to test this hypothesis by including in the convection models a strain– or strain-rate-weakening rheology to represent distributed brittle deformation. The development of a south polar sea (Collins and Goodman, 2007; Schubert et al., 2007) may aid this feedback, because it would locally increase the tidal heating and tidal flexing amplitudes, promoting the weakened conditions at high southern latitudes. Alternatively, a weakened region could be exogenous, resulting for example from an impact of a large body with Enceladus in the distant past. In a similar vein, an impact has been suggested as the cause of the degree-one tectonic pattern on the Martian surface (Marinova et al., 2008; Nimmo et al., 2008). Future impact modeling will be required to evaluate this possibility for Enceladus. In either case, the active region could have developed away from the pole; the resulting thermal structure would then naturally tend to rotate to the pole (Nimmo and Pappalardo, 2006), explaining its current polar location and producing additional stresses that would encourage the tectonic deformation.

Acknowledgments

This work was supported by Grant NNX10AB82G from the NASA Outer Planets Research Program to L.H. A.P.S. was supported by PG&G Grant NN07AR27GC.

References


Brown, R.H. et al., 2006; Composition and physical properties of Enceladus’ surface. Science 311, 1425–1428.


Han, L., Gurnis, M., 1999. How valid are dynamic models of subduction and convection when plate motions are prescribed? Phys. Earth Planet. Inter. 110, 235–246.


