The Dichotomy of Thermal Convection in Enceladus’ Ice Shell

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Abstract

Enceladus exhibits a strong hemispheric dichotomy of tectonism and heat flux, with geologically young, heavily tectonized terrains and a high heat flux in the south polar terrains (SPT) and relatively ancient terrains with presumably lower heat fluxes over the rest of the satellite. To understand the conditions that can give rise to this dichotomy, we present three-dimensional numerical models of convection in Enceladus’ ice shell including tidal heating. Our thermal boundary conditions exhibit no north-south asymmetries, but because the tectonism at the SPT may weaken the ice there, we implement a mechanically weak lithosphere within the SPT. The weakening is parameterized by adopting a reduced viscosity contrast within the SPT. Without such a weak zone, convection (if any) resides in stagnant-lid mode and exhibits no hemispheric dichotomy. In the presence of such a SPT weak zone, however, we find vigorous convection in the ice underneath the SPT, with convective plumes rising close to the surface. In contrast, only stagnant lid convection, or no convection at all, occurs elsewhere over the satellite. Away from the SPT, the heat flux in our models is small (5–10 mW m\(^{-2}\)) and the surface strains are small enough to imply surface ages \(> 10^9\) years. Within the SPT,
however, our models yield heat fluxes of $\sim 70-200 \text{ mW m}^{-2}$, implying heat flows integrated across the SPT of up to 5 GW, similar to that inferred from Cassini thermal observations. The surface strains in our models are high enough near the south pole to cause intense tectonism and imply surface ages of $\sim 10^6-10^7$ years, consistent with age estimates of the SPT. Thus, our models explain the high heat flows, intense tectonism, and young surface ages in the SPT and small heat flows and greater surface ages inferred elsewhere over the satellite.

*Keywords:* Enceladus; Satellites; Saturn; ices.
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). Cassini thermal infrared observations show that the surface temperature is strongly elevated along the tiger stripes, which are geologically young, ∼130 km-long fractures exhibiting elevated temperatures. 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. 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 (Yoder 1979; Ross and Schubert 1989; Spencer et al. 2006; Meyer and Wisdom 2008). Plumes of water vapor, ice, methane, carbon dioxide, nitrogen, ammonia, and possibly $^{40}$Ar emanate from the tiger stripes (Hansen et al. 2006; Brown et al. 2006, Spitalle and Porco 2007, Waite et al. 2006). The plumes are probably generated by evaporation or sublimation of liquid water, ice, or clathrates (Spencer et al. 2006; Kieffer et al. 2006), powered by tidal dissipation along the tiger stripes or within the ice shell beneath (Smith-Konter et al. 2007; Hurford et al. 2007; Nimmo et al. 2007; 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 hemi-
sphere, 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 Ence-
ladus’ SPT, because the thick stagnant lid limits the near-surface thermal gradient (Barr and McKinnon 2007; Barr 2008; Roberts and Nimmo 2008 a; Mitri and Showman 2008). For grain sizes of 0.1–0.3 mm, the estimated heat flux would be \( \sim 10 \text{ mW/m}^2 \) (Barr and McKinnon 2007), over ten times smaller than the observed SPT heat flux (Spencer et al. 2006).

Previous work has shed light on the mechanisms for maintaining a high heat flux at the SPT. By enforcing a very small viscosity contrast \( (10^2 - 10^{3.5}) \) 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 \( \sim 100 \text{ mW/m}^2 \) can be produced. O’Neill and Nimmo (2010) suggested that the current activity could be explained if the convection is episodic, with short, \( \sim 10\)-Myr spikes of high heat flux interspersed between \( \sim 0.1-1 \) 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 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 modeling of thermal convection in Enceladus’ ice shell has not been able to explain the dichotomy between the northern hemisphere and
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 a 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 natural 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 to the 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 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

\[ \frac{\partial \sigma_{ij}}{\partial x_j} + Ra \theta k_i = 0 \]  \hspace{1cm} (1)

\[ \frac{\partial u_i}{\partial x_i} = 0 \]  \hspace{1cm} (2)
\[
\frac{\partial \theta}{\partial t} + u_i \frac{\partial \theta}{\partial x_i} = \frac{\partial^2 \theta}{\partial x_i^2} + q'
\]  

(3)

where \(\sigma_{ij}\) is the stress tensor, \(u_i\) is velocity, \(\theta\) is temperature, \(q'\) is internal tidal heating rate, \(k_i\) 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

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

(4)

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 \(49 \times 49 \times 49\) finite elements within each cap, corresponding to a grid size of approximately \(1.5^\circ\) in longitude 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. Velocity boundary conditions on the inner and outer radii are free-slip. Temperature is fixed on the inner and outer surfaces, with an inner boundary temperature of 273 K, and an outer boundary temperature of 73
The initial temperature decreases linearly from the inner to the outer surface, with an initial disturbance of amplitude 20 K to start the convection.

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). Ice grain sizes in terrestrial glaciers have been estimated at 1-10 mm (Budd and Jacka 1989; Thorsteinsson et al. 1997; De La Chapelle et al. 1998). 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:

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

where \( \eta_0 \) is the reference viscosity at the melting temperature, varying from \( 5 \times 10^{12} \) Pa s to \( 2 \times 10^{15} \) Pa s (corresponding to ice grain size of 0.1 to 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).

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 ejected plumes, high surface temperature and heat flux along the Tiger Stripes suggest that the SPT is undergoing active geological processes. Enceladus’ south polar terrain exhibits significant tectonic disruption, suggesting mechanically weak behavior.

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 $\lesssim 10^2 - 10^{3.5}$ in Barr (2008)’s 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 surface features on Europa (Showman and Han 2004).

Brittle/plastic rheology has been widely implemented in convection models of Earth or icy satellites. 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, 2000b; 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 easily 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, these models are infamous for the instability of the solutions due to the non-linear effects of the plasticity.

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 polar region by reducing the viscosity-contrast cutoff poleward of either 45° or 60°S 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°S latitude (see Figure 1).

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 $10^4 - 10^{10}$ due to temperature variation was implemented. This set of models are comparable to previous 3D spherical numerical models of thermal convection in Enceladus’ ice shell (Roberts and Nimmo 2008a; Bekounkova et al. 2010). For the second set of models, we include a weak south pole by implementing a viscosity cutoff of $10^2 - 10^3$ (see Figure 1). 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 $\mu_E$, 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:

$$\tilde{\sigma}_{ij} = 2\tilde{\mu}(\omega)\tilde{\epsilon}_{ij} + \left[K - \frac{2}{3}\tilde{\mu}(\omega)\right]\tilde{\epsilon}_{kk}\delta_{ij},$$

where

$$\tilde{\mu}(\omega) = \frac{\mu_E\omega^2\eta^2}{\mu_E^2 + \omega^2\eta^2} + i\frac{\mu_E^2\omega\eta}{\mu_E^2 + \omega^2\eta^2},$$

with $\tilde{\sigma}_{ij}$ and $\tilde{\epsilon}_{ij}$ the Fourier transform of the stress and strain tensor components, and $\tilde{\mu}(\omega)$ 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 spheroidal oscillation problem initially developed by Alterman et al. (1959) and Takeuchi and
Saito (1972) in the frequency domain and by imposing the tidal potential as the source of excitation for the spheroidal 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:

$$\tilde{\Phi}_{\text{tide}}(R_s, \omega) = R_s^2 \omega^2 e \left[ -\frac{3}{2} \left( P^0_2(\cos \theta) - \frac{1}{2} P^2_2(\cos \theta) \cos 2\phi \right) - iP^2_2(\cos \theta) \sin 2\phi \right],$$  

(8)

where $R_s$ is the surface radius, $\omega$ is the orbital angular frequency, $e$ is the orbital eccentricity, $P^0_2(\cos \theta)$ and $P^2_2(\cos \theta)$ are the associated Legendre polynomials. 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 Behounkova 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:

$$Q^{-1} \approx \frac{\Im(\tilde{\mu}(\omega))}{|\mu(\omega)|} = \frac{1}{\eta \omega}. \quad (9)$$

The local volumetric tidal dissipation rate, $h_{\text{tide}}$, 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, $Q^{-1}$:

$$h_{\text{tide}}(r, \theta, \phi) = \frac{\left| \tilde{\sigma}_{ij}(r, \theta, \phi) \times \tilde{\epsilon}_{ij}^*(r, \theta, \phi) \right|}{2\omega \eta(T(r, \theta, \phi))}. \quad (10)$$

This formulation remains a reasonable simplification as long as the viscosity
variations do not significantly modify the stress and strain fields. To check this assumption, we have computed the radial solutions by using different viscosity profiles obtained by the thermal convection code and we showed that the radial solution in the ice layer only faintly varies as long as a global decoupling layer is considered between the rocky core and the outer layer. Behounkova et al. (2010) and Han and Showman (2010) have, however, demonstrated that the stress and strain fields can be significantly affected around hot upwellings due to large and sharp variations in viscosity. In the present formulation, such an effect cannot be described. Therefore, the dissipation rate predicted in hot upwellings here are slightly smaller than in a more complete solution.

To compute the viscoelastic response to tidal forcing, 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. 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.

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. Figure 2 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, 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. Figure 3 (a,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 $10^{-7}$ W m$^{-3}$ and the dissipation rate at a depth of 33 km is several orders of magnitude less ($\sim 10^{-15}$ 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. Figure 4 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 Figure 2. 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 3c and Fig 3d).

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 con-
vvection 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. Figure 5 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. From Figure 5, 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. Tidal dissipation rate shows strong dichotomy in this model (Fig 3e and Fig 3f), however, the dissipation rate is too small to have real effects on surface heat flux.

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. Figure 6 shows the results from a model with a Rayleigh number of $2.3 \times 10^7$ and an ice-grain size of 0.4 mm. A maximum viscosity contrast of $10^6$ was adopted except, as before, southward of 60°S latitude, where the viscosity contrast was $10^3$. 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, not shown), the SPT exhibits chaotic, mobile-lid convection, with stagnant-lid convection occurring elsewhere.

In either case with a lower Rayleigh number (Figure 5) or higher Rayleigh number (Figure 6), 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. Figure 7 displays heat flux from a model with a Rayleigh number of $2 \times 10^8$, corresponding to an ice grain size of 0.1 mm. The surface heat flux can be strongly elevated in the southern hemisphere (Figure 7), 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. 7b), 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 $60-200 \text{ mW m}^{-2}$ (Fig 7e), equivalent to an integrated power across the SPT of $\sim 4$ 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 7c and Fig 7f show the radially integrated tidal heating, illustrating that tidal heating is only a small fraction of the total heat budget.

Figure 8 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 8a). 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 $\sim$50–70% greater than when the SPT viscosity contrast is $10^3$. This can be seen by comparing Fig. 8b and 8c, where the steady-state SPT heat flux increases from $\sim$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. 8e and Fig. 8f, where the steady-state SPT heat flux increases from $\sim$2.2 to 3.8 GW as the SPT viscosity contrast decreases from $10^3$ to $10^2$ at a constant Rayleigh number of $2 \times 10^8$. Likewise, Figure 8 shows that, for a constant viscosity cutoff in the SPT, the SPT heat flux increases with Rayleigh number. For example, with an SPT viscosity contrast cutoff of $10^2$, the SPT surface heat flux increases by a factor of $\sim$2.5 (from $\sim$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 $60^\circ$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 $45^\circ$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 $60^\circ$S, the simulations with the smaller weak zone (southward of $60^\circ$S latitude) may represent Enceladus more accurately.

The thickness of Enceladus’ ice shell is uncertain and could range from $\sim$40 km (Olgin et al. 2011) to a maximum of $\sim$100 km. The minimum ice shell thickness for convection to occur is $\sim$40 km (Barr and Mckinnon 2007). To in-
vestigate 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.

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 $6.5 \times 10^7$ (corresponding to an ice-grain size of 0.3 mm) and an SPT viscosity contrast of $10^3$, the average horizontal velocity is 1 mm yr$^{-1}$ in the SPT. By decreasing the viscosity contrast cutoff in the SPT to $10^2$ and using the same Rayleigh number, the average surface velocities reach 5 mm yr$^{-1}$ in the SPT. At the same time, the horizontal surface velocities in the SPT increase if the Rayleigh number increases. For a simulation with a Rayleigh number of $2 \times 10^8$ (corresponding to ice grain size of 0.1 mm) and a viscosity contrast of $10^3$ in the SPT, the average horizontal velocity reaches 10 mm yr$^{-1}$. By decreasing the viscosity contrast cutoff to $10^2$ in the SPT 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 $\sim1$ 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 tidal heating or material strength, the convection (if any) does not exhibit a hemispheric dichotomy. However, we demonstrated that, in the presence of a 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 heat flux at the SPT becomes enhanced by a factor of $\sim10$–20 over that in surrounding regions and can reach $\sim200\,\text{mW m}^{-2}$, implying a heat transport integrated over the SPT of up to $\sim5$ GW. The surface strains in our models imply a resurfacing age of $\sim10^6$-$10^7$ 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 tectonism in our models result solely from the imposed dichotomy of lithospheric strength. Therefore, our results help to naturally explain the observed hemispheric dichotomy of tectonism and heat flux as long as a mechanism for weakening the south-polar regions exists.

The latest heat flow estimates in Enceladus’ SPT from Cassini thermal data yield values of 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 Figure 8), and even reach 3–5 GW in some cases, 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 (Figure 8), 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.

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 tidal dissipation
heating rate in the ice shell underlying the SPT is only $\sim 0.5$ GW, which is much lower than the simulated value of 3.5 GW for heat dissipation at SPT in convective ice shell. The heat lost due to thermal convection is so huge that would result in a very rapid crystallization of the ocean if there is one. Tidal heating in Enceladus is not strong enough to sustain a long-term ocean (Roberts and Nimmo 2008a).

Our models did not include the possibility of enhanced shear heating along the tiger stripes. Concentrated tidal heating can appear along 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 explain 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.

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 convec-
tion models a strain- or strain-rate-weakening rheology to represent distributed
brittle deformation. The development of a south polar sea (Collins and Good-
man 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 pos-
sibility 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.

5. Acknowledgments

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<th>Name</th>
<th>Symbol</th>
<th>Value</th>
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<tr>
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<td>Density</td>
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<td>Thermal expansivity</td>
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<td>Thermal diffusivity</td>
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<td>Thickness of ice shell</td>
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Table 1: Model Parameters
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<th>Layer</th>
<th>(b) (km)</th>
<th>(\rho) (kg.m(^{-3}))</th>
<th>(\mu_E) (GPa)</th>
<th>(K_E) (GPa)</th>
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<td>3625-4370</td>
<td>58-70</td>
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<td>10</td>
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Table 2: Physical parameters for Enceladus’ interior. The density of the rocky core is calculated from the density of the ocean and of the ice shell in order to verify the mean density of Enceladus.
FIGURE CAPTIONS

Figure 1. Schematic of how we implement a weak South Polar Terrain (SPT). In most models, we assumed that the ice is weak in the region from 60°S to 90°S latitude (i.e., southward of 60°S latitude) to evaluate how mechanical weakening influences the convective patterns and resulting heat flux. A small viscosity contrast cutoff, similar to that in mantle convection simulations of Earth to study the behavior of faults and subductions, is implemented in this region, denoted schematically in light blue. Northward of that region (dark blue) the viscosity contrast cutoff is larger than $10^6$ to enforce stagnant-lid convection there. In some models, the weakened zone extends to 45°S rather than 60°S latitude.

Figure 2. Temperature structure from a full 3D spherical model showing thermal convection in Enceladus’ ice shell with a Rayleigh number of $6.5 \times 10^7$. 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.

Figure 3. 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 ($W m^{-3}$) 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–b): No southern hemisphere
weak zone; $Ra = 6.5 \times 10^7$ and the viscosity contrast cutoff is $10^{10}$ everywhere. (c–d): No southern hemisphere weak zone; $Ra = 6.5 \times 10^7$ and the viscosity contrast cutoff is $10^4$ everywhere. (e–f): $Ra = 5 \times 10^6$; southern hemisphere weak zone (southward of 60°S latitude) imposes a viscosity contrast cutoff of $10^2$ with a cutoff of $10^6$ elsewhere. (g–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.

Figure 4. Temperature structure from a full 3D spherical model showing thermal convection in Enceladus’ ice shell with a Rayleigh number of $6.5 \times 10^7$. 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.

Figure 5. 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 $10^3$ 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.

Figure 6. Temperature structure from a full 3D spherical model showing ther-
mal convection in Enceladus’ ice shell with a Rayleigh number of $2.35 \times 10^7$. The
simulation implements temperature-dependent viscosity contrast of $10^6$, except
a viscosity contrast of $10^3$ southward of $60^\circ$S latitude. The top panel displays
the shape of the 216-K isothermal surface. The bottom panel displays the nondi-
mensional temperature distribution along a radial cross-section passing through
the poles. An active ascending plume reaches near the surface at the SPT, while
stagnant-lid convection occurs elsewhere.

**Figure 7.** Heat flux from a model with a Rayleigh number of $2 \times 10^8$, corre-
ponding to an ice grain size of 0.1 mm. The simulation implements temperature-
dependent viscosity contrast of $10^6$, except southward of $60^\circ$S latitude, where the
viscosity contrast is $10^2$. Left panels (a, b, c) show fluxes at a nondimensional
time of 0.0002, while right panels (d, e, f) show fluxes once the model has equi-
librated (nondimensional time of 0.01). Top row shows heat flux through the
bottom boundary. Middle row shows the surface heat flux, demonstrating the en-
hancement of heat flux at the SPT with values exceeding 200 mW m$^{-2}$. Bottom
row shows the radially integrated tidal heating, illustrating that tidal heating is
only a small fraction of the total heat budget.

**Figure 8.** 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 mod-
els 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 $10^3$ in SPT. (c). $Ra = 6.5 \times 10^7$, viscos-

ity cutoff is $10^2$ in SPT. (d). $Ra = 5 \times 10^6$, viscosity cutoff is $10^2$ in SPT. (e). $Ra = 2 \times 10^8$, viscosity cutoff of $10^3$ in SPT. (f). $Ra = 2 \times 10^8$, viscosity cutoff is $10^2$ in SPT. For all cases that have a weak SPT (all cases except (a)), the SPT is implemented southward of $60^\circ$S latitude, and the viscosity contrast is $10^6$ outside the SPT.
Figure 1:
Figure 2:
Figure 4:
Figure 5:
Figure 7:
Figure 8: