Mobile lid convection beneath Enceladus’ south polar terrain

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Enceladus’ south polar region has a large heat flux, 55–110 mW m$^{-2}$, that is spatially associated with cryovolcanic and tectonic activity. Tidal dissipation and vigorous convection in the underlying ice shell are possible sources of heat, however, prior predictions of the heat flux carried by stagnant lid convection range from $F_{\text{conv}} \sim 15$ to 30 mW m$^{-2}$, too low to explain the observed heat flux. The high heat flux and increased cryovolcanic and tectonic activity suggest that near-surface ice in the region has become rheologically and mechanically weakened enough to permit convective plumes to reach close to the surface. If the yield strength of Enceladus’ lithosphere is less than $\sim 1–10$ kPa, convection may instead occur in the “mobile lid” regime, which is characterized by large heat fluxes and large horizontal velocities in the near-surface ice. I show that model ice shells with effective surface viscosities between $10^{16}$ and $10^{17}$ Pa s and basal viscosities between $10^{13}$ and $10^{15}$ Pa s have convective heat fluxes comparable to that observed by Composite Infrared Spectrometer. If this style of convection is occurring, the south polar terrain should be spreading horizontally with $v \sim 1–10$ mm a$^{-1}$ (where $a$ is years) and should be resurfaced in $\sim 0.1–10$ Ma. On the basis of Cassini imaging data, the south polar terrain is $\sim 0.5$ Ma old, consistent with the mobile lid hypothesis. Maxwell viscoelastic tidal dissipation in such ice shells is not capable of generating enough heat to balance convective heat transport. However, tidal heat may also be generated in the near-surface along faults as suggested by Nimmo et al. (2007) and/or viscous dissipation within the ice shell may occur by other processes not accounted for by the canonical Maxwell dissipation model.


1. Introduction

[2] Observations of Enceladus by the Cassini spacecraft indicate that this tiny satellite is geologically active, with plumes of water vapor, dust, and other materials erupting from a region centered near its south pole dubbed the “south polar terrain” (SPT) [Porco et al., 2006; Spencer et al., 2006]. Enceladus’ plumes are spatially associated with a region of increased heat flux, with a total power output of $5.8 \pm 1.9$ GW spread over a region $\sim 70,000$ km$^2$ [Porco et al., 2006; Spencer et al., 2006]. The total power output corresponds to a heat flux of $F = 55$ to 110 mW m$^{-2}$. The Cassini Composite Infrared Spectrometer (CIRS) instrument measured brightness temperatures $\sim 80–90$ K near the centers of the tiger stripes, four quasi parallel features $\sim 500$ km long, centered near the south pole. The active region at the south pole is bounded by cycloidal arcs, with wedge-shaped regions of intense folding at their cusps [Helfenstein et al., 2006]. The SPT has very few craters, and none larger than 1 km, suggesting a young surface age $\sim 0.5$ Ma [Porco et al., 2006]. On Enceladus, the CIRS data provide a unique opportunity to observe the heat flux from the surface of an icy satellite at the time of active resurfacing. The heat flux estimates can be used to argue for or against certain types of convective behavior, can clarify the relationship between convection and resurfacing on Enceladus, and may provide insight into convective-driven resurfacing on other icy satellites.

[3] The activity in the SPT is likely driven by tidal dissipation within Enceladus. The high brightness temperatures along the tiger stripes led Nimmo et al. [2007] to suggest that shear heating due to cyclical strike-slip deformation along fault zones within the stripes could be a dominant source of heat generation in the SPT. Another possibility is that Enceladus’ ice shell is heated from within by tidal dissipation and is vigorously convecting [Barr and McKinnon, 2007; Roberts and Nimmo, 2008]. In such a scenario, the tiger stripes could represent passive spreading centers, or shear heating within the centers of the tiger stripes [Nimmo et al., 2007] and underlying solid state convection could combine to create the large regional heat flux. If convective motions could reach close to the surface of Enceladus, convective heat transport and associated...
lithospheric spreading may provide a natural explanation for the region’s high heat flux [Barr and McKinnon, 2007] and the convergent morphology at the SPT margins. Here, I explore the hypothesis that the SPT is a region on Enceladus where the surface ice has become locally rheologically and mechanically weakened permitting “mobile lid” convection.

[4] In a convecting planetary mantle, the heat flux and the potential for convective-driven lithospheric deformation are controlled in part by the ratio between the viscosity of the material (ice or rock) at the planet’s cold surface (η0) and the viscosity of the mantle at its warm base (η1), $\Delta \eta = \eta_0/\eta_1$. A basic requirement for mobile lid convection is $\Delta \eta < \exp (4(n + 1))$ [Solomatov, 1995, 2004], where $n \sim 1–4$ is the rheological stress exponent, which implies mobile lid behavior can occur for $\Delta \eta \lesssim 10^4$ for $n = 1$. The predicted $\Delta \eta$ for the outer ice I shells of outer solar system satellites, however, is so large that convection in the outer ice I shells of icy satellites is expected to be confined beneath a thick lid of immobile, cold, and highly viscous material that does not participate in convection, called a “stagnant lid.” The presence of the stagnant lid inhibits advective heat transport by preventing convective plumes from reaching close to the surface of the satellite and also inhibits convective-driven resurfacing because thermal buoyancy forces are too small to deform the overlying high-viscosity lithosphere.

[5] Stagnant lid convection, however, does not seem to be compatible with the appearance of the surfaces of terrestrial planets like Venus and Earth (see, e.g., Schubert et al. [2001] for general discussion), nor with the surfaces of tidally flexed icy satellites such as Europa and Enceladus, all of which display a rich variety of surface features suggestive of convective-driven resurfacing [e.g., Helfenstein et al., 2006; Pappalardo et al., 1998; Prockter et al., 2002]. On terrestrial planets, microscale and macroscale defects (cracks and faults) accommodate deformation and facilitate the creation of zones of weakness in the Earth’s crust, which limits the effective viscosity of the lithosphere and permits mobile lid behavior [Kohlstedt et al., 1995; Tackley, 2000; Bercovici, 2003; Solomatov, 2004]. As illustrated in Figure 1, mobile lid convection is characterized by high heat fluxes, substantial horizontal velocities of surface materials, and elongated convection cells, behaviors that are reminiscent of plate tectonics [Schubert et al., 2001]. Europa’s ice shelf can exhibit mobile lid behavior if the effective viscosity of its near-surface ice is limited by the finite yield strength of the lithosphere. Mobile lid convection is possible on Europa if its lithosphere has a yield strength less than 0.02 MPa [Showman and Han, 2005]. The upper limit on convective heat transport and the amount of tidal heating in a stagnant lid ice shell on Enceladus is a factor of $\sim 3–4$ smaller than the heat flux observed by CIRS [Roberts and Nimmo, 2008] (see also section 2).

[6] To determine whether mobile lid convection can occur on Enceladus, and is consistent with the observed heat flux and morphology of the SPT, I perform numerical simulations of mobile lid convection to constrain the relationship between rheological properties of ice, physical properties of the ice shell, and convective behavior. For activity to persist over geologically long timescales, heat transported by convection must be balanced by tidal heat generated within Enceladus’ floating ice shell. A standard Maxwell viscoelastic tidal dissipation model is used to compare tidal dissipation and convective heat transport in a mobile lid ice shell on Enceladus as a function of the shell’s viscosity, physical, and mechanical properties. For mobile lid convection to occur, the ice shell on Enceladus must have a low effective yield strength, and the Rayleigh number of the ice shell must exceed a critical value. I evaluate these quantities using estimates of the yield stress of ice based on terrestrial analogs and studies of lithospheric deformation on Europa. If mobile lid convection is occurring within the SPT, regional lithospheric extension and/or spreading will resurface the region on a timescale controlled by the rheological properties of the near-surface ice. Regional spreading/extension must be balanced by lithospheric loss at the margins of the terrain. These geological consequences provide a means by which the mobile lid hypothesis can be tested using Cassini data.

2. A Limit to Convective Heat Transport in the Stagnant Lid Regime

[7] Enceladus is likely fully differentiated at present, with a rocky core of radius $\sim 150–160$ km [Schubert et al., 2007; Barr and McKinnon, 2007] (depending on the density of its rocky component), and an outer ice I shell between 80 to 100 km thick. It is not known whether Enceladus has a global subsurface ocean, however, the activity in the south polar region would seem to require tidal strains of the magnitude available only to an ice shell decoupled from the rocky core [Nimmo et al., 2007; Roberts and Nimmo, 2008]. A floating outer ice I shell on Enceladus would be heated largely from within by tidal dissipation, which may occur as a result of viscous dissipation due to daily tidal deformation [e.g., Ojakangas and Stevenson, 1989] and/or strike-slip motion along shallow faults [Nimmo et al., 2007]. Tidal dissipation within Enceladus’ rocky core is expected to be extremely small [Roberts and Nimmo, 2008], and radiogenic heating currently provides only $O(1)$ mW m$^{-2}$ to the global heat flux [Schubert et al., 2007]. Therefore, the processes that create and transport the observed heat flux likely occur entirely within Enceladus’ ice shell.

[8] At the low stresses associated with thermal buoyancy on a small satellite like Enceladus, deformation in solid water ice is likely accommodated by diffusion creep [Moore, 2006; Barr and McKinnon, 2007], which occurs by volume diffusion creep. The effective viscosity from volume diffusion is strongly dependent on temperature (T) [Goodman et al., 1981; Goldsby and Kohlstedt, 2001]

$$\eta = \frac{R_GT^2}{14V_mD_{os}} \exp \left( \frac{Q_o}{R_GT} \right)$$

where $R_G$ is the gas constant, $V_m = 1.95 \times 10^{-5}$ m$^3$ mol$^{-1}$ is the molar volume, $D_{os} = 9.10 \times 10^{-4}$ m$^2$ s$^{-1}$ is an estimate of the diffusion coefficient [Goldsby and Kohlstedt, 2001], and $Q_o = 59.4$ kJ mol$^{-1}$ is the activation energy, and $d$ is the ice grain size. On the basis of equation (1), the expected ratio between the viscosity at $T = T_b = 273$ K at the base of Enceladus’ ice shell and $T = T_i = 70$ K is $\Delta \eta \sim 10^{32}$, which suggests that convection in Enceladus’ outer ice I shell would be confined beneath a thick stagnant lid. Equation (1), however, does not adequately describe how extremely cold ice responds to an applied stress. Like any planetary
Figure 1. Comparison between convective temperature, flow, and viscosity fields for an ice shell with $Ra_1 = 10^7$, appropriate for Enceladus’ ice shell with a basal ice viscosity $\eta_1 = 7 \times 10^{13}$ Pa s for (left) mobile lid convection with $\Delta \eta = 10^{2.5}$ and (middle) stagnant lid convection with $\Delta \eta = 10^{6}$. (a) In mobile lid convection, convective plumes rise close to the surface of the ice shell, permitting a large convective heat flux. In this case, $F_{\text{conv}} = 110$ mW m$^{-2}$ ($Nu = 14.35$), compared to (b) the large $\Delta \eta$ case, where plumes are trapped beneath a thick stagnant lid and $F_{\text{conv}} = 15$ mW m$^{-2}$ ($Nu = 1.97$). (c) The horizontally averaged temperature in the mobile lid case (black) is much lower than the stagnant lid case (gray). Dots on both lines indicate the locations of the characteristic temperatures in the convective interior $T_i$. Contours of stream function for the (d) mobile lid and (e) stagnant lid cases (f) and horizontally averaged velocity. Surface velocities are $\sim 170$ mma$^{-1}$ (where a is years) in the stagnant lid case (black) whereas $v \sim 0$ at the surface for $\Delta \eta = 10^{6}$ (gray). Viscosity in the ice shell for the (g) mobile lid and (h) stagnant lid cases (i) and horizontally averaged viscosity as a function of depth for both cases. Dots on the viscosity profiles show the value of the characteristic viscosity in the convecting interior $\eta_i$ and labels illustrate the effective surface viscosity $\eta_0$ and basal viscosity $\eta_1$. Although the viscosities in the convecting interior are comparable in both regimes, the strongly temperature-dependent viscosity in the stagnant lid case dictates high surface viscosities $\eta_0 \sim 10^{22}$ Pa s.
is related to the Rayleigh number of the ice shell evaluated at the surface.

\[ D = \frac{\Delta T_b - T_s}{\Delta T} \]

where \( D \) is the thickness of the shell, \( \Delta T_b = T_b - T_s \) is the temperature difference between the surface and the basal temperature, \( T_b \) is the basal temperature, and \( T_s \) is the surface temperature. In equation (3), \( \Delta T \) is the temperature difference between the surface and the basal temperature, \( T_b \) is the basal temperature, and \( T_s \) is the surface temperature.

\[ F_{\text{conv}} = \frac{k \Delta T}{D} \eta \]

where \( k = 2.27 \text{ W m}^{-1} \text{ K}^{-1} \) is the thermal conductivity of water ice, \( \eta \) is the Nusselt number, which expresses the relative efficiency of convective heat transport over conduction alone (i.e., in a conductive ice shell, \( \eta = 1 \)). In a vigorously convecting, internally heated fluid with a strongly temperature-dependent viscosity, \( \eta \) is related to the Rayleigh number \( (Ra) \), which expresses the vigor of convection [Solomatov and Moresi, 2000]

\[ \eta = 0.53 \theta^{-4/3} R_a^{1/3} \]

where \( \theta \approx Q_i^\frac{g\Delta T}{R_a G i T_b^3} \) and \( T_i \) is the temperature in the well-mixed convecting ice shell. In equation (3), \( R_a \) is the Rayleigh number of the ice shell evaluated at the viscosity in the convecting interior (see Figure 1), \( T_i \), and \( \Delta T \) is the temperature difference between the surface and the basal temperature.

\[ R_a = \frac{\rho g \alpha \Delta T D^3}{\kappa \eta i} \]

where \( \rho = 920 \text{ kg m}^{-3}, g = 0.13 \text{ m s}^{-2} \) is the local acceleration of gravity within the ice shell (\( g = 0.11 \text{ m s}^{-2} \) at the surface, and increases to 0.15 m s\(^{-2}\) at the base of the ice shell [Barr and McKinnon, 2007]), \( \alpha = 1.56 \times 10^{-4} \text{ (150 K)} = 1.7 \times 10^{-4} \text{ K}^{-1} \) is the coefficient of thermal expansion, and \( \kappa = 1.47 \times 10^{-6} \text{ (250 K)} = 1.23 \times 10^{-6} \text{ m}^2 \text{ s}^{-1} \) is the thermal diffusivity [Kirk and Stevenson, 1987].

[11] Evaluating \( \eta \) in equation (2) using equations (3) and (4) gives an expression for \( F_{\text{conv}} \) that depends only on \( T_i \) and descriptive properties of ice

\[ F_{\text{conv}} = 0.53 \left( \frac{Q_i}{R_a G i T_b^3} \right)^{-4/3} \left( \frac{\rho g \alpha D^3}{\kappa \eta (T_i)} \right)^{1/3} \]

Note that for vigorous convection in the stagnant lid regime, the convective heat flux does not depend on the thickness of the ice shell, \( D \). Therefore, the upper limit on heat transfer is controlled simply by the rheology of the ice (through the activation energy and \( \eta_i \)). The convective heat flux is maximized when \( T_i \approx T_s \), a state that is also required for long-term thermodynamic stability of Enceladus’ ocean [McKinnon and Barr, 2008]. A lower limit on \( \eta_i \) can be obtained by evaluating the volume diffusion rheology at \( T = T_b \) and \( d = 0.1 \text{ mm} \), a plausible lower limit for ice grain size in natural systems [McKinnon, 2006], which gives \( \eta = 6 \times 10^{13} \text{ Pa s} \). Using these parameters, the maximum heat flux from tidally or basally heated convection is \( F_{\text{conv}} \approx 20 \text{ mW m}^{-2} \), a factor of \( \sim 3-4 \) lower than the heat flux observed at the SPT. Relaxing the constraint on ice grain size to permit \( \eta = 10^{15} \text{ Pa s} \) permits only a modest increase in \( F_{\text{conv}} \) to 33 mW m\(^{-2}\), still less than the observed SPT value. Roberts and Nimmo [2008] have shown that tidal dissipation in an ice shell with the nominal volume diffusion rheology produces a heat flux from tides \( F_{\text{tidal}} < F_{\text{conv}} \) and \( F_{\text{tidal}} \) much less than the regional heat flux observed by CIRS. This strongly suggests that processes other than a simple balance between convective and tidal heating in the stagnant lid regime are at work in the SPT.

3. Heat Transport in the Mobile Lid Regime

[12] To determine whether mobile lid convection can produce heat flows comparable to those observed in the SPT, it is necessary to perform numerical simulations of this style of convection to obtain a relationship between the Rayleigh number of the ice shell and the Nusselt number. I simulate Newtonian basally heated convection with a \( \Delta \eta \) low enough for mobile lid behavior (the simplest mobile lid system) because a description of its behavior and implications for heat transport on icy satellites is lacking. I first describe the theoretical relationships between the rheology of the fluid, the Rayleigh number, and the convective heat flux in the stagnant and mobile lid regimes on the basis of the work of Solomatov [1995]. These theoretical relationships and numerical results are used to determine a relationship between convective heat flux, rheological properties of ice, and physical properties of Enceladus’ ice shell. The scaling developed here represents a first step toward understanding mobile lid convection on icy satellites and can inform future studies of more complex mobile lid systems.

3.1. Background

[13] As its name implies, the mobile lid regime of convective behavior is characterized by large horizontal surface velocities of the convecting fluid, but also by large convective heat fluxes and relatively elongated convective cells (see Schubert et al. [2001] for general discussion). In
stagnant lid convection, near-surface material is completely immobile and convective plumes are trapped beneath the high-viscosity lid, which imposes a limit on convective heat transfer. The stagnant lid also suppresses the dynamic topography created by convective upwellings: the modest thermal buoyancy stresses associated with convection beneath the lid are not able to create uplift in the overlying cold and viscous surface material. In basally heated stagnant lid convection, the natural wavelength of convective upwellings and downwellings is comparable to, or smaller than, the thickness of the convecting fluid layer. By contrast, convective plumes reach close to the surface in mobile lid convection, permitting high heat fluxes; the conductive boundary layer at the surface of the planet is extremely thin. In mobile lid convection, thermal buoyancy stresses are able to drive deformation in the near-surface material. The natural wavelength of convective upwellings in basally heated mobile lid convection can be very large, between 1 to 4 times the thickness of the fluid layer [Ratcliffe et al., 1997; Kameyama and Ogawa, 2000].

[14] Mobile lid convection has been most thoroughly characterized in the regime of parameter space appropriate for terrestrial planets because some of the behaviors in this regime are reminiscent of plate tectonics. Such studies are focused on determining why Earth is seemingly the only planet with plate tectonics at present; studying the criteria required for mobile lid convection and the motion of near-surface material clarifies the coupling between the behavior of the Earth’s lithosphere and underlying mantle convection [see, e.g., Moresi and Solomatov, 1998; Solomatov, 2004]. Recently, mobile lid convection has been studied as a possible explanation for the variety of features on Europa’s surface inferred to result from tidally driven convection in its ice shell. Showman and Han [2005] induced mobile lid behavior in Europa’s ice shell by limiting the viscosity of its near surface ice to 

$$\eta_{\text{eff}} \sim \sigma_y \varepsilon^2 \varepsilon^2$$

where \(\sigma_y\) is the yield stress of the ice, and \(\varepsilon\) is the second invariant of the strain rate tensor (a similar procedure has been used to model mobile lid behavior on Venus and the Earth [Moresi and Solomatov, 1998; Solomatov, 2004]). Showman and Han [2005] characterized the topography and overall fluid dynamical behavior of the shell as a function of yield stress. They find that periodic mobilization and foundering of Europa’s near-surface icy crust can occur for \(\sigma_y \approx 0.01–0.06 \text{ MPa}\), and continuous recycling of surface material (true mobile lid behavior) can occur for \(\sigma_y \approx 0.02 \text{ MPa}\).

[15] Although the works described above shed light upon the behavior of mobile lid convection, they are almost exclusively focused on the formation of and deformation within the lid, and do not provide much insight into the relationship between convective heat flux and physical and rheological properties of the planetary mantle. Here, I use numerical simulations of convection with a low effective \(\Delta \eta\) to constrain the relationship between Ra and Nu in the mobile lid regime, and to determine the conditions under which the heat flux from mobile lid convection is comparable to that observed by CIRS.

3.2. Scaling of Mobile Lid Convection

[16] In a convecting fluid layer heated from beneath, heat must be conducted across boundary layers at its warm base and cold upper surface. The thickness of each boundary layer is inversely proportional to the vigor of convection: when convection is extremely vigorous (for large \(Ra\)), plumes can transport heat close to the surface of the fluid, so the boundary layer is thin. In a fluid with a temperature-dependent viscosity, the thickness of each boundary layer is also proportional to \(\theta = \ln(\Delta \eta)\), with large \(\theta\) implying a thick conductive boundary layer (or stagnant lid) at the cold upper surface (see Figure 1).

[17] If the upper boundary layer has thickness \(\delta_0\) and the bottom boundary layer \(\delta_i\), the dimensionless heat flux, \(Nu = F_{\text{conv}}(k\Delta T/D)^{-1}\) is related to the boundary layer widths as [Solomatov, 1995]

$$Nu \sim \frac{D}{(b_0 + b_1)}.$$  \hfill (6)

In the stagnant lid regime, \(\delta_0 \sim D\theta^{1/3}Ra_i^{-1/3}\), and the bottom thermal boundary layer is thinner by a factor of \(1/\theta\): \(\delta_i \sim \delta_0/\theta\) [Solomatov, 1995]. In stagnant lid convection, \(\theta \gg 1\), which implies \(\delta_0 \gg \delta_i\) in equation (6), so \(Nu \sim D/\delta_0\). This gives rise to equation (3), where \(Nu \sim \theta^{-4/3}Ra_i^{1/3}\).

[18] Similar to the stagnant lid regime, the Ra-Nu relationship in the mobile lid regime has the form [Olson and Corcos, 1980; Moresi and Solomatov, 1998]

$$Nu \sim Ra_i^{1/3}.$$  \hfill (7)

On Enceladus, the ice viscosity in the well-mixed convecting interior of the ice shell, \(\eta_s\), and Rayleigh number, \(Ra_i\), are not known a priori, so a scaling relationship involving these parameters is not useful for making predictions about its heat flux. A similar uncertainty regarding the properties of Venus’ mantle was faced by Moresi and Solomatov [1998], who suggest a scaling of form

$$Nu = aRa_0^{1/3} \exp(\theta/b),$$  \hfill (8)

where \(a\) and \(b\) are empirical constants,

$$Ra_0 = \frac{\rho g \alpha \Delta T D^3}{\eta_0},$$  \hfill (9)

and \(\eta_0\) is the effective viscosity at the satellite’s surface. If \(\eta_i\) is evaluated using the experimental result from constant viscosity convection, \(\eta_i \approx \eta(\Delta T/2) \approx \eta_0 \exp(\theta/2)\) [e.g., Booker, 1976], this gives \(b = 6\).

[19] To constrain \(a\) and \(b\) to obtain a relationship between the convective heat flux and descriptive properties of a Newtonian ice shell, I performed numerical simulations of convection in basally heated ice shells with \(10^5 < \Delta \eta < 10^{12}\). For Newtonian ice (where \(n = 1\), mobile lid convection occurs for \(\Delta \eta < 10^6\) or \(\theta < 8\). Figure 2 (top) summarizes the locations of the simulations in \(Ra_1-\Delta \eta\) space. Simulations were performed using the finite element convection model CITCOM [Moresi and Solomatov, 1995], in a 2-D Cartesian domain with 64 × 64 elements. For ease of comparison between my scaling and those of previous studies, and to easily relate the behavior of the convection to the behavior of the thermal boundary layers, a temperature-dependent viscosity of form

$$\eta = \eta_0 \exp(-\gamma T)$$  \hfill (10)
is used, where $\gamma = \theta / \Delta T$. With this type of temperature dependence, $Ra_0$ is related to the more commonly used $Ra_1$ as $Ra_0 = Ra_1 \exp(-\theta)$. Free slip boundary conditions were used on the vertical surfaces of the domain, and constant temperature boundary conditions were used on the surface and base of the ice shell. The constant temperature boundary condition at the base of the ice shell ensures that it is forced to be in thermodynamic steady state with an underlying ocean at $T = T_m \approx 273$ K [McKinnon and Barr, 2008]. The surface temperature $T_s = 70$ K used here is appropriate for the south polar region in the absence of internal heating [Spencer et al., 2006].

Table 1 summarizes the values of $Nu$ obtained in the convection simulations. The majority of the simulations have $10^6 < Ra_1 < 10^7$, close to expected values for Enceladus’ ice shell with melting point viscosities for ice, $10^{13}$ Pa s $< \eta_1 < 10^{16}$ Pa s. Simulations were run until $Nu$ and the r.m.s. velocity were constant as a function of time; none of the solutions were time dependent, although, time-dependent behavior is expected for larger $Ra_1$ [Moresi and Solomatov, 1998] and in internally heated cases [Solomatov, 2004].

In the mobile lid regime, if the numerical domain width is comparable to the natural wavelength of convective downwellings, free slip boundary conditions on the vertical edges of the domain can promote the formation of artificially elongated cells [Kameyama and Ogawa, 2000]. As a result, obtaining extremely accurate values of $Nu$ requires running simulations in long aspect ratio domains. Table 2 summarizes $Nu$ values for $Ra_1 = 10^6$ and $\Delta \eta$ as a function of aspect ratio for $1 \times 1$ through $8 \times 1$ domains. The values of $Nu$ are roughly the same for aspect ratios up to $4 \times 1$, because in each of these cases, the heat is transported by a single elongated cell with downwellings at the edge of the

Table 1. Simulation Results for Newtonian Convection in the Mobile Lid Regime

<table>
<thead>
<tr>
<th>$\Delta \eta$</th>
<th>$Nu$</th>
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<th>$v_{rms}$</th>
<th>$max(v_{rms})$</th>
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$^a$Nondimensional value.

Figure 2. (top) Convective regime diagram for a Newtonian fluid [cf. Solomatov, 1995]. Typically, simulations of convection on icy satellites are performed in the stagnant lid regime (II) where large $\Delta \eta$ values lead to the development of a rigid lid of immobile ice near the surface of a convecting shell. Boxes represent locations in $Ra_1$ and $\Delta \eta$ parameter spaces of simulations used to constrain the coefficients in the scaling relationship between $Ra$ and $Nu$ in the mobile lid convective regime (II). Filled boxes represent locations in parameter space of simulations shown in Figure 1. (middle) $Nu$ values as a function of $Ra_0$. (bottom) Comparison between $Nu$ values obtained in numerical simulations and predicted by the scaling relationship (equation (8)) with $a = 0.32$ and $b = 19$. 

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domain. The value of \( Nu \) decreases by 2% in an 8 \times 1 domain because two full convection cells can be maintained. For a 50 km thick ice shell on Enceladus, an 8 \times 1 aspect ratio domain would encompass 25% of the circumference of the satellite: a situation that is better simulated in a fully 3-D spherical geometry. Here, I simulate convection in a 1 \times 1 domain because the measured SPT heat flux is uncertain to within a factor of \( \sim 2 \), so it is not necessary to know \( Nu \) to better than \( \sim 10\% \) accuracy. If mobile lid convection appears to be consistent with the geology of the SPT, more sophisticated geometries should be used in future work to provide tighter constraints on heat flux.

[22] A fit to the \( Ra - Nu \) data gives

\[
Nu = 0.32 Ra_0^{1/3} \exp(\theta/19),
\]

similar to the relationship obtained by Moresi and Solomatov [1998], whose simulations were performed for higher \( Ra \) than used here. The scaling is accurate to within 0.1% for low \( Ra \), high \( \Delta \eta \), and low \( Nu \), and the accuracy decreases as \( Nu \) increases. This likely occurs because \( \eta_i \) is weakly dependent on \( Ra_0 \) [Moresi and Solomatov, 1998]; inclusion of this effect would cause the power on \( Ra_0 \) in equation (8) to change by \( \sim 10\% \).

### 3.3. Convective Heat Flux

[23] The convective heat flux is related to \( Ra_0 \) and \( \Delta \eta \) as

\[
F_{\text{conv}} = \frac{k\Delta T}{D} Nu = \frac{k\Delta T}{D} a Ra_0^{1/3} \exp(\theta/b),
\]

where \( a = 0.32 \) and \( b = 19 \) (see section 3.2). Evaluating \( Ra_0 \) in terms of the physical properties of the ice shell and the thermal properties of water ice,

\[
F_{\text{conv}} = 0.32 \left( \frac{\rho g \Delta T^4 k_0^3}{k} \right)^{1/3} \frac{\exp(\theta/19)}{\eta_0^{1/3}},
\]

where \( \eta_0 \) is the effective surface viscosity of the ice shell. Similar to the stagnant lid regime, the convective heat flux is independent of the ice shell thickness, but depends on the boundary temperatures (through \( \Delta T \)) and is critically dependent upon the ice rheology, through \( \eta_0 \) and \( \theta \). Figure 3 (right) illustrates that only a narrow range of surface viscosities, \( 10^{10} \text{ Pa s} < \eta_0 < 10^{17} \text{ Pa s} \) gives convective heat fluxes comparable to those observed by CIRS.

[24] The free parameters in \( F_{\text{conv}} \) are the basal ice viscosity, \( \eta_1 \), and \( \theta \), which appears explicitly in the exponential term, and implicitly in \( \eta_0 = \eta_1 \Delta \eta = \eta_1 \exp(\theta) \). The heat flux can be expressed in terms of \( \eta_1 \) as

\[
F = 0.32 \left( \frac{\rho g \Delta T^4 k_0^3}{k} \right)^{1/3} \frac{\exp(\theta/19)}{\eta_1^{1/3} \exp(\theta/3)}.
\]

Figure 3 (left) illustrates how the convective heat flux varies as a function of \( \Delta \eta \) and \( \eta_1 \). For the highest value of \( \Delta \eta = 10^{3.25} \), a basal viscosity between \( \eta_1 = 10^{13} \) and \( 10^{16} \text{ Pa s} \), similar to values expected from the volume diffusion rheology with grain sizes \( d \sim 0.1–1 \text{ mm} \), gives \( F_{\text{conv}} \) comparable to the regional heat flux observed in the SPT. For smaller \( \Delta \eta \), convection becomes more vigorous, so larger basal viscosities can give \( F_{\text{conv}} \) comparable to the values observed.

### 4. Tidal Heat Generation

[25] To remain thermodynamically stable, the amount of heat generated within Enceladus’ ice shell from tidal dissi-

---

**Table 2. Dependence of \( Nu \) on Domain Aspect Ratio for \( Ra_1 = 10^6 \) and \( \Delta \eta = 10^{2.5} \)**

<table>
<thead>
<tr>
<th>Aspect Ratio</th>
<th>( Nu )</th>
<th>( \eta_{\text{mm}} )</th>
<th>Cells</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 \times 1</td>
<td>6.30</td>
<td>140</td>
<td>1</td>
</tr>
<tr>
<td>2 \times 1</td>
<td>6.25</td>
<td>132</td>
<td>1</td>
</tr>
<tr>
<td>4 \times 1</td>
<td>6.25</td>
<td>132</td>
<td>1</td>
</tr>
<tr>
<td>8 \times 1</td>
<td>6.04</td>
<td>149</td>
<td>2</td>
</tr>
</tbody>
</table>

*Nondimensional value.*
Table 3. Physical and Rheological Properties of Enceladus’ Interior

<table>
<thead>
<tr>
<th>Layer</th>
<th>Parameter</th>
<th>Symbol</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rocky Core</td>
<td>Radius</td>
<td>$R_c$</td>
<td>161 km</td>
</tr>
<tr>
<td>Rocky Core</td>
<td>Density</td>
<td>$\rho_c$</td>
<td>3500 kg m$^{-3}$</td>
</tr>
<tr>
<td>Rocky Core</td>
<td>Young’s modulus</td>
<td>$E_c$</td>
<td>$2.5 \times 10^7$ Pa</td>
</tr>
<tr>
<td>Rocky Core</td>
<td>Poisson’s ratio</td>
<td>$\nu_c$</td>
<td>0.25</td>
</tr>
<tr>
<td>Ocean</td>
<td>Thickness</td>
<td>$D_{ocean}$</td>
<td>Variable</td>
</tr>
<tr>
<td>Ocean</td>
<td>Density</td>
<td>$\rho_o$</td>
<td>1000 kg m$^{-3}$</td>
</tr>
<tr>
<td>Ocean</td>
<td>P-wave velocity</td>
<td>$V_p, o$</td>
<td>1.45 km s$^{-1}$</td>
</tr>
<tr>
<td>Ice Shell</td>
<td>Thickness</td>
<td>$D$</td>
<td>Variable</td>
</tr>
<tr>
<td>Ice Shell</td>
<td>Density</td>
<td>$\rho$</td>
<td>920 kg m$^{-3}$</td>
</tr>
<tr>
<td>Ice Shell</td>
<td>Young’s modulus</td>
<td>$E_i$</td>
<td>$8 \times 10^7$ Pa</td>
</tr>
<tr>
<td>Ice Shell</td>
<td>Poisson’s ratio</td>
<td>$\nu_i$</td>
<td>0.325</td>
</tr>
<tr>
<td>Ice Shell</td>
<td>Mean ice viscosity</td>
<td>$\eta$</td>
<td>$10^{14} - 10^{22}$ Pa s</td>
</tr>
</tbody>
</table>

Tidal dissipation must approximately equal the amount of heat transported by convection. If the convective heat flux exceeds the amount of tidal heating, the interior of the satellite will cool, causing the underlying ocean to freeze [Moore, 2006; Roberts and Nimmo, 2008]. In addition to controlling the convective heat flux, the rheology of Enceladus’ shell controls the height of the daily tidal deformation experienced by the satellite, by affecting its degree-2 Love numbers [Mullen, 2006; Nimmo et al., 2007]. Previous work shows that the volumetric tidal dissipation rates in an ice shell in the stagnant lid regime are significantly smaller than the convective heat flux [Roberts and Nimmo, 2008]. If the ice shell had a rheology governed purely by volume diffusion, the presence of the high-viscosity stagnant lid would suppress tidal deformation and limit the amount of tidal dissipation [Moore, 2006; Roberts and Nimmo, 2008]. However, this balance has not been explored in the small $\Delta \eta$ regime. Here, tidal dissipation within the shell is modeled as viscous dissipation in a Maxwell viscoelastic solid, and related to the tidal potential from Saturn using the $h_2$ Love number of a model Enceladus with a floating ice shell. The surface heat flux from tidal dissipation is compared to the convective heat flux and the observed heat flux from CIRS.

[26] The heating rate per unit mass within the ice shell due to viscous dissipation driven by diurnal tidal deformation is [Ojakangas and Stevenson, 1989; Showman and Han, 2004]

$$H = \frac{\dot{\varepsilon}^2 \eta}{2 \rho \left(1 + \frac{\omega^2 h_2^2}{\mu^2}ight)},$$  (15)

where

$$\dot{\varepsilon} \sim 3 \times 10^{-9} (h_2) s^{-1}$$  (16)

is the tidal strain rate [Nimmo et al., 2007], $h_2$ is the Love number relating diurnal tidal deformation to the tidal force on Enceladus from Saturn, $\omega = 2\pi/T_E = 5.308 \times 10^{-5}$ s$^{-1}$ is Enceladus’ orbital frequency, $T_E = 1.37$ days is Enceladus’ orbital period, and $\mu = 3.5 \times 10^9$ Pa is the shear modulus of ice. Similar to section 2, here, I make the approximation that tidal dissipation is uniformly distributed within an ice shell that is thin compared to the radius of the satellite, with an average heating rate given by equation (15) with $\eta$ evaluated at the viscosity in the well-mixed convective interior of the ice shell. In this case, the surface heat flux from tidal dissipation is

$$F_{t {\text{ tidal}}} = \rho HD = \frac{\dot{\varepsilon}^2 \eta D}{2\left(1 + \left(\frac{\omega^2 h_2^2}{\mu^2}\right)\right)}.$$  (17)

[27] The strain rate in Enceladus’ ice shell is calculated using $h_2$ for a satellite consisting of an elastic rocky core, a subsurface liquid water ocean, and a viscoelastic ice shell (J. Wahr et al., Modeling stresses on satellites due to non-synchronous rotation and orbital eccentricity using gravitational potential theory, submitted to Icarus, 2008). In the simulations of mobile lid convection, the viscosity in the ice shell is $\eta_i \approx \eta_0 \exp(\theta/2)$ (see, e.g., Figure 1 (bottom)). Because the ice viscosity is approximately constant, Love numbers of the satellite are determined using a single layer of ice of variable thickness with a range of $\eta$ values appropriate for the well-mixed interiors of ice shells convection in the mobile lid regime.

[28] Values of $h_2$ for the model satellite due to the applied tidal potential from Saturn are calculated using the correspondence principle [Sabadini et al., 1982], using the SatStress software package (J. Wahr et al., submitted manuscript, 2008). This approach has been used to study tidal heating in both rocky and icy satellites including Io, Europa, and Enceladus [Segatz et al., 1988; Ross and Schubert, 1989; Moore and Schubert, 2000; Wahr et al., 2006; Moore, 2006; Nimmo et al., 2007]. Mechanical and rheological parameters used for each layer of the satellite are summarized in Table 3. Using the same mechanical and rheological properties of ice, and the same physical structure for the interior of Enceladus, my values of $h_2$ match those reported by Nimmo et al. [2007] to 10%.

[29] Figure 4 (top) illustrates how $h_2$ varies as a function of ice shell viscosity and thickness. For an ice shell 30 km thick, $h_2$ decreases from 0.18 to 0.03 as the viscosity in the ice shell is increased from $10^{13}$ to $10^{16}$ Pa s. The maximum $h_2 \sim 0.65$ occurs for the thinnest ice shell, $D = 8$ km and the lowest viscosity, $\eta = 10^{13}$ Pa s. Figure 4 (bottom) illustrates how the global mean diurnal tidal strain rate $\dot{\varepsilon} \approx 3 \times 10^{-9}$ $h_2$ s$^{-1}$ varies as a function of ice shell thickness and viscosity. Strain rates vary from $\dot{\varepsilon} \sim 3 \times 10^{-10}$ to $3 \times 10^{-11}$ s$^{-1}$ and are maximized in thin, low-viscosity ice shells.

[30] Figure 5 summarizes a comparison between $F_{t {\text{ tidal}}}$ and $F_{\text{conv}}$ a function of the surface viscosity $\eta_0$. For a broad range of plausible ice shell viscosities, the tidal heat generation falls short of the convective heat flux, in most cases, by a factor of $\sim 10$ or more. The amount of tidal dissipation calculated here represents an upper limit on heat generation in a mobile lid shell. Because mobile lid behavior of the ice shell is confined within Enceladus’ SPT and is not occurring everywhere on Enceladus, the remainder of its ice shell may have a high effective near-surface viscosity. However, the tidal deformation model described above assumes a radially symmetric viscosity structure and thus, that Enceladus’ ice shell has a low near-surface viscosity everywhere. The values of $h_2$ calculated using the globally low viscosity assumed here are therefore upper limits. In a more realistic model, tidal deformation in the SPT might be limited by the high viscosity of the remainder of Enceladus’ lithosphere and may be much lower than.
For convection to occur, the Rayleigh number in the ice shell must exceed a critical value. The critical Rayleigh number for convection in a basally heated fluid layer convecting in the moderate viscosity contrast regime has been constrained by Stengel et al. [1982] using linear stability analysis and laboratory experiments. Unlike in the stagnant lid regime, there is no clear relationship between the critical Rayleigh number and $\Delta \eta$ for moderate $\Delta \eta$. Table 4 summarizes the values of critical Rayleigh number for convection in a Newtonian fluid as a function of $\Delta \eta$ obtained by Stengel et al. [1982]. For mobile lid convection to occur, $Ra_1 > Ra_{cr,1}$. Evaluating $Ra_1$ in terms of the properties of the ice shell,

$$Ra_1 = \frac{\rho g a \Delta T D^3}{\kappa \eta \Delta},$$

and solving for the critical ice shell thickness where convection is possible,

$$D > \left( \frac{\eta_b Ra_{cr,1} \kappa}{\rho g a \Delta T \Delta \eta} \right)^{1/3}.$$ 

For $\Delta \eta = 10^3$, in an ice shell with a heat flux comparable to that measured by CIRS, $\eta_b \approx 10^{16}$ Pa s, which implies that convection can occur if $D > 5$ km. The critical Rayleigh number criterion is easily achieved even in a thin ice shell on Enceladus in the mobile lid regime. However, if convection itself (perhaps coupled with tides) is required to damage the near-surface ice and induce mobile lid behavior, convection may have to be initially triggered in an ice shell with a high effective viscosity contrast. In this case, the basal viscosity of the shell must be relatively low, $\sim 10^{13} - 10^{14}$ Pa s, and the ice shell relatively thick, $D \approx 100$ km, for convection to start [Barr and McKinnon, 2007].

So far, I have used the criterion that mobile lid behavior can occur in Enceladus’ ice shell if $\Delta \eta < 10^3$. The most straightforward method of lowering the effective viscosity contrast in the ice shell is to suppose that the effective viscosity of the cold near-surface ice is limited by the finite yield stress of ice. Recently, Solomatov [2004] has developed a general criterion for the critical yield strength for lid mobilization on the basis of the rheology of the convecting layer. For convection to occur in the mobile lid regime, stresses built up in the lithosphere have to be comparable to the yield strength of the lithosphere [Solomatov, 2004]

$$\sigma_{yield} \leq 13 \frac{\rho g a}{\Delta T} \left( \frac{RT_h}{\rho \sigma_h} \right)^2 l_h,$$

where $T_h \sim 273$ K is the characteristic temperature beneath the upper thermal boundary layer, and $l_h \sim D$ is the horizontal length scale over which stress accumulates in the lithosphere [Solomatov, 2004], which is comparable to the width of convective upwellings. For parameters appropriate to convection in a 30 km thick ice shell on Enceladus, where deformation is accommodated by volume diffusion, $\sigma_{yield} \sim 3 \times 10^{-3}$ MPa.

Field characterization of fracture in the Ross Ice Shelf suggest that the yield stress of ice is $\sim 0.1$ MPa.

Figure 4. (top) Contours of $\log_{10}(h_2)$, the magnitude of degree 2 Love number $h$, as a function of ice shell thickness and viscosity. The amount of diurnal deformation on Enceladus that gives rise to tidal dissipation in the ice shell is proportional to $h_2$. (bottom) Contours of $\log_{10}(\dot{\varepsilon})$ (in units of s$^{-1}$), the global, time-averaged diurnal tidal strain rate in a floating ice shell on Enceladus, as a function of ice shell thickness and viscosity. Diurnal deformation and strain rates are large in thin ice shells and in shells with high viscosity.

estimates in Figure 5. It is also possible that tidal deformation could become concentrated within a low-viscosity zone at the south pole [Moore, 2001]. Further work to determine how tidal strain is partitioned in an ice shell with a heterogeneous viscosity structure is required to properly evaluate the balance between tidal heating and convection.

5. Requirements for Mobile Lid Behavior

The presence of solid state convection in Enceladus’ ice shell is not a foregone conclusion [Barr and McKinnon, 2007]. For convection to occur, the Rayleigh number in the ice shell must exceed a critical value. The critical Rayleigh number for convection in a basally heated fluid layer convecting in the moderate viscosity contrast regime has been constrained by Stengel et al. [1982] using linear stability analysis and laboratory experiments. Unlike in the stagnant lid regime, there is no clear relationship between the critical Rayleigh number and $\Delta \eta$ for moderate $\Delta \eta$. Table 4 summarizes the values of critical Rayleigh number for convection in a Newtonian fluid as a function of $\Delta \eta$ obtained by Stengel et al. [1982]. For mobile lid convection to occur, $Ra_1 > Ra_{cr,1}$. Evaluating $Ra_1$ in terms of the properties of the ice shell,

$$Ra_1 = \frac{\rho g a \Delta T D^3}{\kappa \eta \Delta},$$

and solving for the critical ice shell thickness where convection is possible,

$$D > \left( \frac{\eta_b Ra_{cr,1} \kappa}{\rho g a \Delta T \Delta \eta} \right)^{1/3}.$$ 

For $\Delta \eta = 10^3$, in an ice shell with a heat flux comparable to that measured by CIRS, $\eta_b \approx 10^{16}$ Pa s, which implies that convection can occur if $D > 5$ km. The critical Rayleigh number criterion is easily achieved even in a thin ice shell on Enceladus in the mobile lid regime. However, if convection itself (perhaps coupled with tides) is required to damage the near-surface ice and induce mobile lid behavior, convection may have to be initially triggered in an ice shell with a high effective viscosity contrast. In this case, the basal viscosity of the shell must be relatively low, $\sim 10^{13} - 10^{14}$ Pa s, and the ice shell relatively thick, $D \approx 100$ km, for convection to start [Barr and McKinnon, 2007].

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$$\sigma_{yield} \leq 13 \frac{\rho g a}{\Delta T} \left( \frac{RT_h}{\rho \sigma_h} \right)^2 l_h,$$

where $T_h \sim 273$ K is the characteristic temperature beneath the upper thermal boundary layer, and $l_h \sim D$ is the horizontal length scale over which stress accumulates in the lithosphere [Solomatov, 2004], which is comparable to the width of convective upwellings. For parameters appropriate to convection in a 30 km thick ice shell on Enceladus, where deformation is accommodated by volume diffusion, $\sigma_{yield} \sim 3 \times 10^{-3}$ MPa.

Field characterization of fracture in the Ross Ice Shelf suggest that the yield stress of ice is $\sim 0.1$ MPa.
[Kehle, 1964; Showman and Han, 2005], much larger than the critical yield stress for mobile lid convection. Terrestrial field estimates for $\sigma_Y$ may not be directly applicable to the surfaces of icy satellites, because measurements are usually acquired at ambient temperatures $\sim 240$–273 K. The microphysical processes responsible for fracture in macroscopic terrestrial ice bodies (e.g., grain boundary sliding and dislocation pileups [Frost, 2001]) are thermally driven [Sinha, 1991], and are not likely to be responsible for fracture in ice with temperatures relevant to Enceladus’ lithosphere, $T \sim 70$–180 K.

[34] Estimates of the yield strength of Europa’s lithosphere, which may be more similar in terms of temperature and microtexture to the lithosphere of Enceladus than terrestrial ice, range from $\sim 0.01$–0.1 MPa. The lower bound on yield stress arises from the observation that cycloidal ridges on Europa propagate in response to daily tidal stresses, which have magnitudes $\sim 0.01$ MPa assuming a floating ice shell [Hoppa et al., 1999]. It is possible, however, that stresses twice as high are required to initiate cycloids on Europa [Hurford et al., 2007b], and this larger initiation stress may be more relevant to lithospheric fracturing than the lower propagation stress. If Europa’s bands form in a style similar to terrestrial midocean ridge spreading [Prockter et al., 2002], a lithospheric yield strength of 0.4 to 2 MPa is consistent with the width of fault blocks flanking the band’s central trough [Stempel et al., 2005].

[35] It has been suggested that plume activity in the SPT is driven in part by opening and closing of portions of the tiger stripes because of the diurnal tidal stresses [Hurford et al., 2007a]. If that were the case, diurnal stresses on Enceladus, similar to Europa, may be able to drive at least superficial lithospheric cracking. Using the values of $h_2$

### Table 4. Critical Rayleigh Number for Mobile Lid Convection

<table>
<thead>
<tr>
<th>$\Delta \eta$</th>
<th>$Ra_{1,cr}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>$10^{2.5}$</td>
<td>$1.01 \times 10^4$</td>
</tr>
<tr>
<td>$10^3$</td>
<td>$2.12 \times 10^4$</td>
</tr>
<tr>
<td>$10^{3.25}$</td>
<td>$4.70 \times 10^4$</td>
</tr>
<tr>
<td>$10^4$</td>
<td>$6.41 \times 10^4$</td>
</tr>
</tbody>
</table>
Love number calculated in section 4, this suggests yield stresses for Enceladus’ lithosphere of less than \( \sim 500 (h^2) \) kPa, or \( \sim 5–50 \) kPa for \( h \sim 0.01 \) to 0.1, depending on the properties of the near-surface ice.

Because Enceladus is a small body with relatively low gravity, and its convection is driven by small temperature fluctuations, the yield stress required for mobile lid convection is extremely low. The required yield stress is seemingly a factor of \( \sim 10 \) lower than estimates of the yield stress derived from models of Europa’s surface geology, and a factor of \( \sim 1000 \) lower than yield strength estimates from terrestrial ice sheets. It should be noted that a similar situation occurs on Earth, where the yield stress of rock measured in laboratory contexts is a factor of \( \sim 10 \) higher than the critical yield stress for the initiation of subduction [Solomatov, 2004]. On Earth, one can appeal to chemical variations such as the presence of water [Solomatov, 2004] and/or semibrittle behavior of the lithosphere due to microcracking and faulting [Kohlstedt et al., 1995; Tackley, 2000; Bercovici, 2003] to mobilize the lid. On Enceladus, the presence of the tiger stripes suggest that tidal stresses alone, or acting in concert with convective stresses, may be sufficient to fracture the near-surface ice. Because tidal strain and strain rates are maximized on Enceladus’ poles [Ojakangas and Stevenson, 1989], the likelihood of tidally induced lithospheric fracture is highest at this location. Improved knowledge of the friction and failure of cold ice and the interaction between convective and tidal stresses in icy satellite lithospheres can shed light upon the likelihood of mobile lid convection on icy bodies.

6. Geological Consequences

As the name implies, convection in the mobile lid regime leads to substantial horizontal velocities of near-surface ice. If mobile lid convection is responsible for lithospheric recycling and surface deformation at the SPT, the SPT should have a younger surface than the rest of Enceladus, and the age of the surface should be related to the horizontal velocity of the surface ice. The horizontal velocities are related to the convective heat flux through \( Ra_0 \) and \( \eta_0 \). Rapid extension or spreading in the SPT must be balanced by material loss at its flanks, so understanding the processes by which extension/spreading in the SPT is accommodated by the regional geology is a key to evaluating the plausibility of mobile lid convection. Here I describe how these geological consequences provide a means by which the mobile lid convection hypothesis can be tested using Cassini data.

6.1. Relationship Between SPT Age and Surface Viscosity

The characteristic convective velocities in the upper boundary layer, \( v_0 \), are inversely proportional to the boundary layer thickness, \( \delta_0 \) [Solomatov, 1995]

\[
v_0 \sim \left( \frac{D}{\delta_0} \right)^2 \kappa \frac{\kappa}{D} \sim Ra_0^{2/3} \frac{\kappa}{D}.
\]

Note that nondimensional velocities such as those reported in Tables 1 and 2 may be expressed in meters per second by multiplying by \( \kappa/D \), where \( D \) is the ice shell thickness. Because \( \delta_0 \sim Ra_0^{-1/3} \), the near-surface velocity scales with the surface Rayleigh number as \( v_0 \propto Ra_0^{-1/3} \). Nondimensional values of the maximum horizontal velocity of surface ice, \( \max(|v_{x,sf}|) \), obtained in each of the mobile lid simulations are summarized in Table 1. The maximum horizontal velocity scales as

\[
\max(|v_{x,sf}|) = 0.08Ra_0^{0.8} \frac{\kappa}{D}.
\]

Figure 6 illustrates the dependence of \( \max(|v_{x,sf}|) \) on \( Ra_0 \) and shows the fit to the values obtained in the numerical
simulations. The dependence of surface velocity on $Ra_0$ implies that surface velocities for the ice shell will depend on $\eta_0$ and the shell’s thermal and physical parameters,

$$\max \left( |v_{x, sf}| \right) = 0.08 \frac{k}{D} \left( \frac{\rho \alpha \Delta T D^3}{k} \right)^{0.8} \frac{1}{\eta_0^{0.8}}. \quad (23)$$

Unlike the convective heat flux, the surface velocity is dependent on the ice shell thickness $D$; even if the fit had yielded an exact $Ra_0^{2/3}$ dependence, the surface velocity would still depend on $D$. Evaluating the velocity using thermal and physical parameters appropriate for Enceladus’ ice shell,

$$\max \left( |v_{x, sf}| \right) = 25 \text{ mm a}^{-1} \left( \frac{10^{17} \text{ Pa s}}{\eta_0} \right)^{0.8} \left( \frac{D}{30 \text{ km}} \right)^{1.4}. \quad (24)$$

comparable to spreading rates on midocean ridges on Earth [e.g., Turcotte and Schubert, 1982] and comparable to estimates of the spreading rates on band-type features on Europa [Stempel et al., 2005].

For a circular SPT with an area 70,000 km² [Porco et al., 2006], the equivalent radius of the region is $R_{\text{spt}} \sim 150$ km. If mobile lid convection is the driving force for resurfacing, the age of the surface is related to the shell thickness and viscosity as

$$\tau_{\text{spt}} \sim \frac{R_{\text{spt}}}{\max \left( |v_{x, sf}| \right)} \sim 6 \text{ Ma} \left( \frac{\eta_0}{10^{17} \text{ Pa s}} \right)^{0.8} \left( \frac{30 \text{ km}}{D} \right)^{1.4}. \quad (25)$$

Figure 6 (right) illustrates how the predicted age of the SPT varies as a function of $\eta_0$ and $D$. For the range of surface viscosities that give $F_{\text{conv}}$ close to the heat flux observed by CIRS, the predicted age of the SPT is between 0.2 and 7 Ma, depending on the ice shell thickness. This is consistent with the 0.5 Ma age inferred from the lack of craters with diameter greater than 1 km [Porco et al., 2006], but more detailed studies are needed to place tighter constraints on the cratering statistics. Ages significantly longer than 7 Ma require convection that is too sluggish to give the observed heat flux; conversely, ages much less than 0.2 Ma imply $F_{\text{conv}} > F$.

### 6.2. Lithospheric Loss at Cusps

Regional extension or spreading at the SPT driven by mobile lid convection must be balanced by the loss of material at the flanks of the terrain. On Enceladus, the SPT seems to be bounded by a series of arcuate features, which join at cusps. Folding at the cusps suggest that they may be sites of convergent tectonic activity [Helfenstein et al., 2006] and are a natural location to accommodate the regional extension or spreading predicted to occur in mobile lid convection.

Unfortunately, the processes by which icy satellites accommodate compressional deformation and lose surface material are not well understood. On terrestrial planets, crustal loss occurs by subduction. Although it is tempting to appeal to subduction or a subduction-like process on Enceladus, it is not clear how local density differences of the magnitude required for lithospheric fracture and foun-
dering might be created. One way of generating negative buoyancy within an ice shell is to generate local melting, but whether the melt simply drains through the ice shell or is retained to create local negative buoyancy depends on the poorly understood details of how melt drains within satellite ice shells.

On terrestrial planets, crustal loss can also occur beneath continents where the thickened, cold, lithosphere can become gravitationally unstable and sink into the mantle [e.g., Molnar et al., 1998]. Similar processes have been suggested to balance the seemingly ubiquitous lithospheric extension on Europa. For example, Prockter and Pappalardo [2000] have identified a set of long-wavelength, low-amplitude folds on the surface of Europa which could represent locations of local thickening of the lithosphere and crustal loss. On Enceladus, the folded terrain within the cusps could represent sites of locally thickened lithosphere and cold, negatively buoyant ice dripping into the underlying warm convecting ice shell. Characterization of the long-wavelength topography within the SPT and at its margins may be able to shed light on the processes by which regional extension is accommodated.

### 7. Discussion

The discovery of the warm cryovolcanically and possibly tectonically active south polar region of Enceladus provides a unique opportunity to understand the relationship between tidal dissipation, heat transport, and resurfacing on icy satellites. The high regional heat flux at the SPT, $F = 55$ to $110$ mW m$^{-2}$, is a factor of $\sim 3-4$ higher than predicted by prior studies of stagnant lid convection on Enceladus [Barr and McKinnon, 2007; Roberts and Nimmo, 2008]. If Enceladus’ surface ice has a viscosity as high as values predicted by the theoretically constrained volume diffusion rheology, convective heat transport is limited to $\sim 15-30$ mW m$^{-2}$, and the amount of tidal dissipation within the ice shell falls short of the observed heat flux, as well [Roberts and Nimmo, 2008].

Here, I have proposed that Enceladus’ south polar region is a location where the near-surface ice is relatively weak, which permits convective fluid motions to reach closer to the surface and drive lithospheric deformation. The effects of a weak brittle lithosphere are mimicked in the framework of a purely viscous convection model by limiting the effective viscosity of the near-surface ice. If the ratio between the effective surface viscosity and basal viscosity of the ice shell is less than $10^4$, Enceladus’ SPT convects in the so-called “mobile lid” regime of convective behavior, which is characterized by large heat fluxes.

Heat fluxes comparable to those observed by CIRS are possible if the effective viscosity of the surface ice is $\eta_0 \sim 10^{10}$ to $10^{11}$ Pa s. Such low effective viscosities near the surface of the ice shell permit the near-surface ice to be mobilized by the underlying convection, which may drive rapid resurfacing in the SPT. If mobile lid convection is occurring, the age of the SPT should be $\tau_{\text{spt}} \sim O(0.1-10)$ Ma. A paucity of craters larger than 1 km on the south polar terrain suggests that its surface is less than 0.5 Ma old [Porco et al., 2006], which is consistent with mobile lid behavior. If mobile lid convection is occurring in the SPT, the lithospheric spreading associated with convective
upwellings must be balanced by loss of surface ice at the margins of the region. This would be consistent with the presence of convergent tectonics at the margins of the SPT [Helfenstein et al., 2006]. Folded terrain observed at the cusps between cycloidal arcs that bound the SPT may represent locations of thickened lithosphere where cold near surface ice pushed into the cusps drips down into the underlying convecting ice shell in the style of sublithospheric instabilities beneath terrestrial continents.

[46] The amount of tidal heating generated in a mobile lid ice shell falls short of the amount of heat transported by mobile lid convection. At face value, this suggests that mobile lid convection might be a transient phenomenon because the efficient convective heat transport in the mobile lid regime would cause Enceladus’ ocean to freeze [Roberts and Nimmo, 2008]. The model used to calculate tidal dissipation assumes that Enceladus has a radially symmetric viscosity structure. A more realistic model would consider the SPT to be a low-viscosity region embedded in an ice shell with an otherwise high viscosity. If such a structure allowed tidal deformation to become concentrated within the SPT, it may be able to generate higher local heating rates than calculated here [cf. Moore, 2001] and perhaps permit an equilibrium between tidal dissipation and convection. Another possibility is that Newtonian volume diffusion does not accommodate diurnal tidal strain in Enceladus’ ice shell. If another processes, such as grain boundary sliding, accommodated tidal deformation, the effective viscosity of the ice shell over tidal timescales would be different from the effective viscosity over convective timescales [McKinnon, 1999]. Very little is known about the microphysical processes responsible for attenuation in water ice that is being cyclically flexed at frequencies appropriate to tidal deformation on Enceladus. Such information is critically needed to help clarify the relationship between tidal flexing and the activity at the SPT.

[47] It has been previously suggested that shear heating due to slip motion along fault zones within the centers of the tiger stripes provide a substantial amount of heat to the total observed heat flux at the SPT [Nimmo et al., 2007]. Such a heat source is compatible with the presence of mobile lid convection, and indeed, the two processes acting in concert may be able to provide a total heat flux comparable to that observed by CIRS. Observations of the heat flux in regions of the SPT away from the tiger stripes may shed light upon how tidal dissipation in the region is partitioned between shear heating and viscous dissipation in the underlying ice shell. In addition, although a heat source representative of ridges was not included as part of the convection simulations here, the interaction between ridge heating and mobile lid convection may be a promising avenue of future work.

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