A Common Origin for Ridge-and-Trough Terrain on Icy Satellites by Sluggish Lid Convection

Amy C. Barr\textsuperscript{a}, Noah P. Hammond\textsuperscript{a}

\textsuperscript{a}Department of Geological Sciences, Brown University, Providence, RI 02912

Abstract

Ridge and trough terrain is a common landform on icy satellites of the outer solar system. Examples include the grooved terrain on Ganymede, gray bands on Europa, coronae on Uranus’s moon Miranda, and ridges and troughs in the northern plains of Saturn’s small, but active, moon Enceladus. Regardless of setting, the heat flow and strain rates associated with the formation of each of these terrains are similar: heat flows of order tens to a hundred milliwatts per meter squared, and deformation rates of order $10^{-16}$ to $10^{-12}$ s$^{-1}$. Barr (2008) and Hammond and Barr (2014a) have previously shown that the conditions associated with the formation of ridge and trough terrain on Ganymede and the south polar terrain on Enceladus are consistent with solid-state ice shell convection in a shell with a weak surface. Here, we show that sluggish lid convection can simultaneously create the heat flow and deformation appropriate for the formation of ridge and trough terrains on a number of satellites. This conclusion holds regardless of the thickness of the satellites’ ice shells. For convection to deform their surfaces, the ice shells must have yield stresses similar in magnitude to the daily tidal stresses exerted by the gravitational pull from their parent planets. This suggests that tidal and convective stresses must act together to deform the surface.
and that the spatial pattern of tidal cracking on the surfaces of the moons controls the locations of ridge and trough terrain.

**Keywords:**

1. Introduction

Each of our solar system’s outer planets, Jupiter, Saturn, Uranus, and Neptune, harbors a family of regular satellites ranging from ~200 to 2600 km in radius. Most of these moons have mean densities $1000 \text{ kg m}^{-3} < \rho < 2000 \text{ kg m}^{-3}$, in between the values inferred for ice and silicate rock ($\rho_i = 920 \text{ kg m}^{-3}$ and $\rho_r = 3000 \text{ kg m}^{-3}$), suggesting mixed ice+rock compositions. Many of the satellites show signs of endogenic resurfacing, some in the form of extensional tectonics, which may occur as organized systems of sub-parallel ridges and troughs, so-called “ridge and trough terrain,” or as systems of either large canyons (e.g., Ithaca Chasma on Tethys), or smaller faults (e.g., “wispy” terrain on Dione). Radiogenic heating supplies a modest heat flow in the interiors of most moons, with surface heat fluxes of order one to tens of mW m$^{-2}$ (Spohn and Schubert, 2003; Hussmann et al., 2006; Schubert et al., 2007). Thus, much of the endogenic resurfacing is thought to have occurred early in the satellites’ histories, when additional sources of heat from accretion, short-lived radioisotopes, and/or differentiation were available to drive resurfacing.

Organized groups of sub-parallel ridges and troughs, which we refer to as "ridge-and-trough terrain," occurs in a number of settings on various satellites. Examples include Ganymede’s grooved terrain (see Pappalardo et al., 2004 for discussion), Europa’s bands (Prockter et al., 2002; Stempel et al., 2007).
Miranda’s coronae (Pappalardo et al., 1997; Hammond and Barr, 2014b), and swaths of ridges and troughs in the northern plains of Enceladus (Bland et al., 2007). Each of these terrains are characterized by sub-parallel ridges and troughs with kilometer-scale spacing (see Table 1). The fault spacing implies a shallow brittle/ductile depth and thus, a high thermal gradient at the time of formation (e.g., Nimmo et al. 2002). Some of the terrains, such as Europa’s bands, subdued grooves on Ganymede, and Elsinore corona on Miranda, are bounded by a sharp groove, and seem to represent sites of emplacement of fresh material from the subsurface (e.g., Pappalardo and Sullivan 1996; Sullivan et al. 1998; Head et al. 2002). Other terrains, such as Arden corona on Miranda and ridge systems on Enceladus appear to be sites of extension without complete lithospheric separation (Pappalardo et al., 1997; Bland et al., 2007).

Although other satellites show evidence of tectonics, closely spaced sub-parallel ridges and troughs seem to occur primarily on satellites that have experienced tidal flexing, which provides a large heat source and a possible means of lithospheric weakening. When the orbital periods of satellites within the same system are integer multiples, gravitational interactions between the bodies increase their orbital eccentricities and/or inclinations (Malhotra and Dermott, 1990; Showman and Malhotra, 1997; Meyer and Wisdom, 2008), leading to the periodic raising and lowering of a tidal bulge. If the satellite has a subsurface liquid water ocean, which appears to be common (Zimmer et al., 2000; Kivelson et al., 2002; Hussmann et al., 2006), most of the tidal energy is deposited in the outer ice I shell of the satellite, which floats atop the liquid ocean and is free to deform (e.g., Ojakangas and Stevenson...
Through the process of tidal heating, spin energy from the parent planet is converted to mechanical energy via tidal flexing of the satellites’ interiors, where it is ultimately dissipated as heat. Tidally flexed icy satellites can experience heat flows of tens to a hundred mW m$^{-2}$ during their time in resonance (e.g., Cassen et al. 1979, 1980; Malhotra and Dermott 1990; Meyer and Wisdom 2008; Hammond and Barr 2014b). The cyclical deformation of the satellites’ outermost ice shells may also weaken the near-surface ice, facilitating deformation.

Solid-state convection has long been suggested to play a role in driving deformation on the icy satellites (e.g., Parmentier et al. 1982; Shoemaker et al. 1982; Pappalardo et al. 1998). However, the relationship between convection and resurfacing remains unclear (see, e.g., Barr and Showman 2009 for discussion). Because the viscosity of water ice depends strongly on temperature (Goldsby and Kohlstedt, 2001), fluid motions in the ice shells of the outer planet satellites are thought to be confined to a relatively thin layer at the base of the shell, beneath a thick stagnant lid of cold ice that is too stiff to participate in convection. In the stagnant lid regime, convective stresses are too small to deform the near-surface ice, preventing resurfacing and limiting the convective heat flow to $\sim 30$ mW m$^{-2}$ (Barr, 2008).

Prior work has shown that if the near-surface ice in the shells of Enceladus (Barr, 2008) or Ganymede (Hammond and Barr, 2014a) is extremely weak, convective plumes can reach close to the surface, supplying heat fluxes up to $\sim 200$ mW m$^{-2}$ and $\sim$ mm/yr deformation rates; this style of convection is known as “sluggish lid” because the near-surface “lid” of cold ice is dragged along by the underlying convective flow (Schubert et al., 2001). Here, we use
numerical simulations of ice shell convection to show that similar rheological parameters can give rise to the heat flows and deformation rates inferred for the formation of several examples of extensional ridge and trough terrain. We determine scaling relationships for the heat flow and strain rate in extensional regions on the satellites and show that these quantities are independent of the thickness of the ice shell. The yield stresses required for sluggish lid behavior are similar in magnitude to the daily tidal stresses on each satellite. Tides may weaken the near-surface ice and/or provide lines of pre-existing weakness that can be exploited by convective buoyancy stresses to make swaths of extensional ridges and troughs.

2. Observations

2.1. Inferred Formation Conditions

There are several methods by which the effective heat flow during deformation can be estimated from characterization of topography on the surface of a planet. One common approach is to derive an estimate of the thickness of the brittle/elastic layer at the surface of the satellite based on measurements of the dominant spacing between ridges and troughs. The topography measurements can be performed on digital elevation models, or on images (where brightness is taken as a proxy for topography, e.g., Patel et al. 1999). At the base of the brittle layer, the behavior of the ice is assumed to change from elastic to viscous. Using assumptions about the rheology of ice and the strain rate of deformation, lithospheric deformation models can yield estimates of the dominant wavelength of deformation and its relationship to the the temperature and depth of the brittle/ductile transition.
(e.g., Dombard and McKinnon 2001; Bland and Showman 2007; Bland et al. 2010). The temperature and depth of the brittle/ductile transition depth can be used to estimate the thermal gradient and thus, the heat flow.

A second common approach is to estimate the effective elastic thickness of the ice shell by looking for flexural uplift near deformed terrains (e.g., Nimmo et al. 2002). The deformed terrain is assumed to represent a load emplaced on the lithosphere, which drives flexural warping. The wavelength of the flexural deformation is proportional to the thickness of the elastic portion of the lithosphere. An assumption of strain rate and ice rheology can yield an estimate of the temperature at which the ice behavior transitions from elastic to viscous. Similar to the method based on fault spacing described above, the temperature at the base of the elastic layer and the elastic layer thickness permit an estimate of heat flow.

Table I summarizes geological and geophysical constraints on deformation wavelength, heat flux, and strain rate derived from photogeology of several examples of extensional ridge and trough terrain. Here, we mostly use information determined from measurements of fault spacing. The main sources of error in both the fault spacing and flexural methods are the assumptions about ice rheology and the strain rate of deformation. Although the rheology of the ice in the upper few kilometers can be constrained by laboratory measurements (Goldsby and Kohlstedt, 2001), constraints on strain rates are much looser. Minimum and maximum strain rates are often estimated based on the ages of the terrain (e.g., a feature estimated to be $\tau = 1$ Gyr old could have formed with a strain rate as low as $\dot{\varepsilon} \sim 1/\tau \sim 3 \times 10^{-17}$ s$^{-1}$). Uncertainties in thermal gradient are typically a factor of $\sim 2$, and strain
rates may be constrained only to within several orders of magnitude. These uncertainties are reflected in the values listed in Table 1.

2.2. Morphology

Roughly two thirds of the surface of Jupiter’s ice/rock moon Ganymede (with radius $R = 2631$ km and mean density $\rho = 1940$ kg m$^{-3}$), is covered with bright, relatively young “grooved terrain,” swaths of sub-parallel ridges and troughs, often within a sharp bounding groove (see Pappalardo et al., 2004 for discussion). Groove lanes are $\sim$ 10 to 100 km wide (Collins et al., 2000) and contain groups of sub-parallel ridges and troughs spaced by $\sim$ 1-2 km, superimposed on broad pinches and swells $\sim$8 km apart (Patel et al., 1999). The double wavelength structure of grooved terrain is most readily created by extensional necking at strain rates $\sim 10^{-16}$ to $10^{-13}$ s$^{-1}$ in an ice shell with a very high thermal gradient, $\sim$ 5 to 30 K km$^{-1}$ (Bland and Showman, 2007; Bland et al., 2010). Flexural studies indicate heat flows at the time of deformation between 80 to 200 mW m$^{-2}$ (Nimmo et al., 2002). This heat flow is far in excess of that implied by radiogenic heating (Dombard and McKinnon, 2001), suggesting that the terrain formed during Ganymede’s passage through an orbital resonance, which may have also triggered global ice/rock separation and the formation of an ocean (Showman et al., 1997). Some groove lanes, so-called “subdued grooves,” show signs of extensive strike-slip motion and may be sites of emplacement of fresh material from beneath (Head et al., 2002).

Features called “bands” on Jupiter’s moon Europa ($R = 1569$ km; $\rho = 3040$ kg m$^{-3}$) share morphological similarities with the subdued grooves on Ganymede (Head et al., 2002) and with terrestrial mid-ocean ridges (Prockter et al., 2002).
Bands are lanes of sub-parallel ridges and troughs roughly 6 to 25 km wide, characterized by a central trough, a hummocky zone, and sets of imbricate fault blocks (Prockter et al., 2002) with ~0.5 km spacing (Stempel et al., 2005). Similar to subdued grooves on Ganymede, bands appear to be sites of complete lithospheric separation and emplacement of fresh material from below (Schenk and McKinnon, 1989; Pappalardo and Sullivan, 1996; Sullivan et al., 1998). Stempel et al. (2005) applied a simple model of mid ocean ridge spreading to show that the characteristic fault block spacing in bands can form in an ice shell deforming with strain rates $\sim 10^{-15}$ to $10^{-12}$ s$^{-1}$ with a brittle/ductile transition temperature at $T_{BDT} \sim 150$ to 190 K, at a depth of 2 to ~ 10 km. For a nominal thermal conductivity $k = 3.52$ W m$^{-1}$ K$^{-1}$ (see Table 2), this corresponds to a heat flow of 15 to 150 mW m$^{-2}$.

The surface of Uranus’s icy moon Miranda ($R = 236$ km; $\rho = 1200$ kg m$^{-3}$) is dominated by three zones of intense deformation, dubbed “coronae.” Coronae are 200–300 km in diameter, polygonal to ovoidal in shape and have concentric outer belts of sub-parallel linea (Smith et al., 1986). The outer belts of each corona have distinct morphologies. Arden corona has concentric ridges and troughs with $\approx 5$ km spacing and 2 km or relief. Inverness corona has ridges and troughs whose spacing increases with increasing distance from the corona center (Pappalardo et al., 1997). Elsinore corona has relatively widely spaced ridges and troughs with subdued topography (Schenk, 1991). The outer belts are interpreted as normal faults and cryovolcanic materials (Greenberg et al., 1991; Schenk, 1991; Pappalardo et al., 1997) and each corona is consistent with forming under concentric tensional stresses that ra-
iated from the feature’s center (Collins et al., 2010). Topography along the flanks of Arden corona suggest the surface may be supported by flexure and that the elastic thickness during corona formation was likely \( \approx 2 \) km, suggesting a thermal gradient of \( 8 \rightarrow 20 \) K km\(^{-1}\) (Pappalardo et al., 1997). With a thermal conductivity of 4.2 W m\(^{-1}\) K\(^{-1}\), this implies a heat flow of 34 to 84 mW m\(^{-2}\). Such a large thermal gradient could be generated by energy dissipation in Miranda’s interior during an orbital resonance with neighboring satellite Umbriel (Tittemore and Wisdom, 1989). Convection driven by tidal heating during resonance can match the distribution of surface deformation and the thermal gradient implied by flexure.

Saturn’s small moon Enceladus has two types of ridge and trough terrain. Its south polar terrain, a \( \sim 70,000 \) km\(^2\) quasi-circular region, is a site of intense tectonic deformation and high regional heat flow, with a power output measured by the Cassini CIRS instrument at 3-7 GW (Spencer et al., 2006). Much of the thermal emission is being emitted from four sub-parallel linear features dubbed “tiger stripes,” which are also sites of eruptions of water ice particles with small amounts of silicate and salt (e.g., Porco et al., 2006; Waite et al., 2006; Postberg et al., 2009). Regions in between the tiger stripes are characterized by sub-parallel folds with a spacing of \( 1.1 \pm 0.4 \) km and a funicular texture, reminiscent of ropy pahoehoe (Barr and Preuss, 2010). A simple folding model shows that strain rates \( 10^{-14} \rightarrow 10^{-12} \) s\(^{-1}\) can recreate the observed fold spacing (Barr and Preuss, 2010). Away from the south polar terrain, Enceladus has several large systems of extensional ridge and trough terrain with a dominant spacing between faults \( \sim 3\)-4 km (Bland et al., 2007). Creating such short-wavelength features requires an
effective elastic thickness of only 0.4 to 1.4 km, implying a local heat flow
\( F \sim 110 \) to \( 220 \) mW m\(^{-2}\) at the time of deformation (Bland et al., 2007),
similar to the heat flow currently estimated for the south polar terrain.

3. Methods

The high heat flow and intense deformation of the south polar terrain of
Enceladus led many to suggest that the terrain represented a region where the
near-surface ice was weak, permitting convective plumes to reach close to the
surface (Barr, 2008; Roberts and Nimmo, 2008; O’Neill and Nimmo, 2010;
Han et al., 2012). Here, we use numerical simulations of ice shell convection
to determine whether the heat flow and deformation rates arising from this
style of convection is consistent with those inferred for the formation of ridge
and trough terrains on each of the icy satellites.

3.1. Model

We simulate convection using the finite element model CITCOM (Moresi and Solomatov,
1995). The vigor of convection is expressed by the Rayleigh number,

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

where \( \rho \) is the density of ice, \( g \) is the acceleration of gravity, \( \alpha = 1.56 \times 10^{-4}(T_b/250 \) K\) \( K^{-1} \) (Kirk and Stevenson, 1987) is the coefficient of thermal
expansion, \( \Delta T \) is the difference in temperature between the surface \( (T_s) \) and
basal \( (T_b) \) ice, \( D \) is the thickness of the ice shell, \( \kappa = 1.47 \times 10^{-6}(250 \) K/\( T_b) ^2 \) m s\(^{-2}\) is the thermal diffusivity (Kirk and Stevenson, 1987), and \( \eta_b \) is the
basal viscosity. Table 2 summarizes values of these parameters used for each
satellite in the study. We vary the ice shell thickness between \( \sim 10 \) and 100
km, consistent with ice shell thickness estimates for each of the satellites. We use a temperature dependent Newtonian rheology for ice (cf., Solomatov 1995),

$$\eta(T) = \eta_0 \exp(-\gamma T)$$  \hspace{1cm} (2)

where $$\gamma = \theta / \Delta T$$ and $$\theta = \ln(\Delta \eta)$$, where $$\Delta \eta = \eta_0 / \eta_b$$ is the ratio between the viscosity at the surface of the ice shell and its base. We use a value of $$\eta_b = 3 \times 10^{14}$$ Pa s, close to the melting point viscosity of ice for a nominal grain size $$d = 0.1$$ mm (Barr and McKinnon 2007; Barr 2008).

The effect of a finite yield stress in rock and ice is commonly modeled in purely viscous convection models such as CITCOM by setting $$\max(\eta) \approx \sigma_Y / \dot{\varepsilon}_{II}$$, where $$\sigma_Y$$ is the yield stress of the convecting material (described by Byerlee’s law; Beeman et al. 1988), and $$\dot{\varepsilon}_{II}$$ is the second invariant of the strain rate tensor (e.g., Trompert and Hansen 1998; Moresi and Solomatov 1998; Showman and Han 2005; O’Neill and Nimmo 2010). Here, we use a simpler approach, where we simply limit the viscosity of ice to a constant value, between $$10^2$$ to $$10^3$$ times larger than the basal viscosity (Barr 2008; Hammond and Barr 2014a,b). In a Newtonian fluid, sluggish lid behavior occurs for $$\Delta \eta \lesssim 10^4$$ to $$10^5$$ (Solomatov 1995). We favor a simpler approach because it limits the number of free parameters in our calculations, allowing us to focus on the relationship between the behavior of the near-surface ice and its rheology.

Figure 1 illustrates the values of $$Rab$$ and $$\Delta \eta$$ explored in our study. Our simulations are performed in an 8×1 two-dimensional Cartesian domain with 1024×128 elements, with periodic boundary conditions (Hammond and Barr 2014a). The surface and base of the ice shell are held at constant tempera-
tures $T_s$ and $T_b$, implying that the ice shell is heated purely from its base. Although we imagine that ridge and trough terrain is formed in ice shells undergoing active tidal flexing and heating, the details of how the mechanical energy of tidal flexing is converted to heat in the satellites interiors are not well understood (see e.g., Barr and Showman 2009 for discussion). Here, we focus on basally heated ice shells with the knowledge that the details of internal heating are not likely to change the behavior of the upper thermal boundary layer (Solomatov, 2004) where the ridge and trough terrain will be formed.

3.2. Outputs

Simulations are run until the dimensionless heat flow, or Nusselt number ($Nu$) has reached a statistical steady state, so that

$$Nu_{rms} = \frac{\int_0^t (Nu(t))dt}{\int_0^t dt}$$

(3)

has converged to the $10^{-5}$ level (Solomatov and Moresi, 2000). We measure the strain rate, $\dot{\varepsilon}_x = dv_{x,sf}/dx$, where $v_{x,sf}$ is the $x$–velocity at the surface. The surface experiences extension where $\dot{\varepsilon}_x > 0$. In these regions, we record the average heat flow, $Nu_{ext}$, and the strain rate, $\dot{\varepsilon}_{ext}$. Successful simulations are those in which both of these quantities matches the values in Table 1.

4. Results

4.1. Heat Flow

The heat flow across the convecting ice shell is related to the Nusselt number,

$$F = \frac{k\Delta T}{D}Nu,$$

(4)
where \( k = 651/T \) W m\(^{-1}\) K\(^{-1}\) is the temperature-dependent thermal conductivity of ice, which we evaluate at \( T = T_s + \frac{1}{2} \Delta T \) (see Table 2). For Newtonian convection, the relationship between the Rayleigh number and Nusselt number, \( Nu \sim Ra^{1/3} \) (see Schubert et al. 2001 for discussion). Moresi and Solomatov (1998) propose that the relationship for sluggish lid convection has the form, \( Nu = aRa_0^b \exp(\theta/c) \), where \( b \equiv 1/3 \). We find a good fit between our \( Nu \) data and this function (see Appendix A). However, here we are interested primarily in the value of \( Nu \) in extensional zones, \( Nu_{\text{ext}} \), for which \( a = 0.48 \pm 0.09 \), \( b = 0.31 \pm 0.009 \), and \( c = 14.21 \pm 2.7 \) (see Appendix A for discussion). In general, \( Nu_{\text{ext}}/Nu \sim 2 \). The heat flow in extensional zones,

\[
F_{\text{ext}} = 0.48 \frac{k\Delta T}{D} Ra_0^{0.31} \exp\left(\frac{\theta}{14.21}\right),
\]  

is essentially independent of \( D \) because \( b \sim 1/3 \) and \( Ra \propto D^3 \). If we impose \( b = 1/3 \) and evaluate the Rayleigh number, we can obtain a rough estimate of \( F_{\text{ext}} \),

\[
F_{\text{ext}} \approx 0.48 \left(\frac{\rho g \alpha T^4 k^3}{\kappa}\right)^{1/3} \exp(\theta/14.21) \frac{\eta_b^{1/3} \exp(\theta/3)}{\eta_{\text{b}}^{1/3} \exp(\theta/3)},
\]  

where, for ease of application to the icy satellites, we have evaluated the viscosity at the base of the ice shell, \( \eta_b \). Any weak dependence of \( F_{\text{ext}} \) on \( D \) arising because \( b \neq 1/3 \) is due to limitations of our numerical model (resolution and domain width) and should not be construed as physically meaningful.

Figure 2 illustrates the range of \( \Delta \eta \) values for which \( F_{\text{ext}} \) matches geological constraints for the formation of Europa’s bands, Miranda’s coronae, Enceladus’ South Polar Terrain, the fold belts in the northern plains
of Enceladus, and Ganymede’s grooved terrain. We find that ice shells with $10^{2.5} < \Delta \eta < 10^{4.25}$ can create heat flows high enough to create each of these features. Ice shells with $10^{3.25} < \Delta \eta < 10^{3.5}$ are consistent with the creation of each of the four sets of features we consider. This range of $\Delta \eta$ is also consistent with the heat flow and morphology of convective upwellings required to create the global tetrahedral distribution of Miranda’s coronae (Hammond and Barr, 2014b).

4.2. Deformation Rates

In isoviscous convection, the velocity of near-surface material, $v_{x,sf} \propto Ra_0^{2/3}$ (Solomatov, 1995). Thus, we expect the near-surface strain rate $\dot{\varepsilon}_x$, to be related to the surface Rayleigh number,

$$\dot{\varepsilon}_{\text{ext}} = a_e Ra_0^{b_e} \frac{\kappa}{D^2},$$

(7)

where $\kappa/D^2$ is used to convert from non-dimensional strain rate output by CITCOM to strain rate in s$^{-1}$. Fits to our data give $a_e = 0.42 \pm 0.12$ and $b_e = 0.71 \pm 0.03$. Note that similar to the heat flow, $\dot{\varepsilon}_{\text{ext}}$ should be independent of the ice shell thickness, $D$, if $b_e = 2/3$. Again a weak dependence on $D$ arises from limitations in our numerical model due to resolution and box width, but this should not be interpreted as a means of constraining $D$ from terrain morphology.

Figure 3 illustrates how the strain rates in extensional zones predicted by equation (7) vary as a function of $\Delta \eta$ and the thickness of the ice shells on Europa, Ganymede, and Enceladus. Strain rates associated with the formation of Europa’s bands are readily obtained for a wide range of $\Delta \eta$. The low strain rates associated with grooved terrain formation imply $\Delta \eta \gtrsim 10^{3.25}$. 

14
Strain rates predicted for the Enceladus SPT can be created only in very weak shells with $\Delta\eta < 10^{3.75}$, but strain rates associated with the formation of ridges and troughs in the northern plains are created for a wide range of $\Delta\eta$.

5. Discussion

Extensional ridge-and-trough terrain is a common landform on tidally flexed icy satellites. Examples include the grooved terrain on Ganymede, bands on Europa, Miranda’s coronae, and ridges and troughs in the northern plains of Enceladus. On each of these satellites, the spacing between ridges and troughs is of order a kilometer, up to ten kilometers. Each of these terrains is inferred to form in an ice shell with a high thermal gradient, and thus, heat flow. Strain rates between about $10^{-16}$ and $10^{-12}$ s$^{-1}$ (see Table 1) are consistent with the spacing between ridges and troughs observed. Our numerical simulations of convection in an ice shell with a weak surface show that the heat flow and strain rate associated with ridge and trough terrain can be created by sluggish lid convection. This conclusion holds on each satellite regardless of the thickness of the ice shell because the heat flow and strain rate are independent of $D$.

We also find that a single (albeit narrow) set of rheological parameters, $10^{3.25} < \Delta\eta < 10^{3.75}$ can give rise to conditions appropriate for the formation of all of the terrains. By limiting the value of $\Delta\eta$ to less than $10^4$ to $10^5$, we are mimicking the effect of weak near-surface ice. A crude estimate of the ice strength in our model, $\sigma_y \sim \eta_0 \dot{\varepsilon}_{ext}$, implies that the near-surface ice must have a yield strength $\sim 1$ to 300 kPa for sluggish lid behavior. This
yield stress is many orders of magnitude lower than that inferred for the yield stress of ice based on terrestrial field studies (Kehle, 1964) and laboratory experiments (Beeman et al., 1988). However, the formation of the cycloidal cracks on Europa (Hurford et al., 2007b), the timing and duration of plume eruptions on Enceladus (Hurford et al., 2007a), and the putative eruptions on Europa (Roth et al., 2014), suggest that the modest daily tidal stresses exerted on these bodies are capable of cracking near-surface ice. These tidal stresses are of similar magnitude to $\sigma_y$ (Hurford et al., 2007b,a; Wahr et al., 2009).

This suggests to us that the stresses from tides and the stresses from solid-state convection are both required to create ridge and trough terrain. The cyclical tidal flexing of the ice shells has been suggested to produce a significant amount of heat in the shells via tidal dissipation (e.g., Ojakangas and Stevenson, 1989). However, we suggest that the cyclical flexing is also modifying the structure of the near-surface ice, at the macro- and possibly micro-scale.

At macro-scales, pre-existing lines of weakness driven by tidal forces may provide weak locations in the ice shell where thermal buoyancy stresses from underlying convection can pull the surface apart. This may explain why only tidally flexed icy satellites have global occurrences of extensional ridge and trough terrain, whereas satellites of similar size and composition that have not experienced tidal flexing have little endogenic activity. Classic examples of this apparent paradox include Ganymede (tidally flexed) and Callisto (no tidal flexing and no endogenic resurfacing); and Enceladus (tidally flexed) and Mimas (no tidal flexing and no endogenic resurfacing). If convection can only drive deformation along lines of weakness, this
might explain, for example, the global degree-2 pattern of grooved terrain on Ganymede (Patterson et al., 2010). Convection may also create ridge and trough terrain by pulling apart pre-existing tidal cracks in the satellites’ ice shells. On Europa, reconstruction of pre-existing features across bands suggests that these features represent locations of the emplacement of new material (Pappalardo and Sullivan, 1996; Sullivan et al., 1998). When the regions of fresh material are removed, the remaining landform resembles a simple ubiquitous europa crack in the ice shell (see, e.g., Figure 2 of Pappalardo and Sullivan 1996).

At micro-scales, the cyclical working of the ice shell at low temperatures (where kinetic processes such as grain growth and annealing are slow compared to the timescale of tidal flexing), may introduce cracks and defects in the ice that severely weakens its structure, decreasing its yield stress. Similar processes, along with the presence of pore fluids, may be responsible for decreasing the yield stress of the crust on the Earth, permitting plate tectonics (see, e.g., Kohlstedt et al. 1995 for discussion). Another possibility is that tidal flexing may introduce heat in the near-surface ice, providing thermal softening (Roberts and Nimmo, 2008) which allows for intense deformation driven by convection. In either case, laboratory experiments are needed to clarify the response of ice to cyclical flexing at frequencies and temperature conditions appropriate for the ice mantles of the outer planet satellites.

Here, we have focused on the role of convection in driving, principally, extensional deformation on the icy satellites. Definitive morphological evidence for compression on icy satellites is rare (e.g., Prockter and Pappalardo 2006). However, recent work by Bland and McKinnon (2012) suggests that
low-amplitude compressional folds can form at strain rates and thermal gradients similar to conditions arising in the compressional zones in our simulations (Hammond and Barr, 2014a). Thus, it may be possible that the seemingly ubiquitous extension on icy satellites may be accommodated by folds that are difficult to detect in existing images. Further images from spacecraft, e.g., the forthcoming ESA JUICE mission, could shed light on the relationship between extension and compression on icy bodies.

Acknowledgements

This work was supported by NASA OPR NNX12AL22G and NESSF NNX13AN99H.
| Satellite | Feature                      | $\lambda$ (km) | $w$ (km) | $F$ (mW m$^{-2}$) | $\dot{\varepsilon}$ (s$^{-1}$) |
|-----------|------------------------------|----------------|----------|------------------|-------------------------------|
| Enceladus | South Polar Terrain$^a$      | 1.1 ±0.4       | 55–110   | $10^{-14}$ – $10^{-12}$ |
| Europa    | Bands$^{b,c}$                | 0.5            | 6–25     | $10^{-15}$ – $10^{-12}$ |
| Enceladus | Northern Plains$^d$           | 3–4            | 30–50    | $10^{-15}$ – $10^{-12}$ |
| Ganymede  | Grooved Terrain$^{e,f,g,h}$  | 2; 8           | 10–100   | $10^{-16}$ – $10^{-13}$ |
| Miranda   | Coronae$^i$                  | 5–10           | 34 – 84  | –                |

Table 1: Wavelength of deformation ($\lambda$), width of deformed zones ($w$), inferred heat flow ($F$), and strain rates ($\dot{\varepsilon}$) estimates for ridge-and-trough terrains on various icy satellites.

$^a$Barr and Preuss (2010), $^b$Prockter et al. (2002), $^c$Stempel et al. (2005), $^d$Bland et al. (2007), $^e$Nimmo et al. (2002), $^f$Patel et al. (1999), $^g$Dombard and McKinnon (2001), $^h$Bland and Showman (2007), $^i$Pappalardo et al. (1997).
| Parameter                  | Symbol | Europa | Ganymede | Enceladus | Miranda |
|----------------------------|--------|--------|----------|-----------|---------|
| Surface Temperature        | $T_s$ (K) | 110    | 130      | 80        | 60      |
| Basal Temperature          | $T_b$ (K) | 260    | 260      | 273       | 250     |
| Ice Shell Density          | $\rho$ (kg m$^{-3}$) | 920    | 920      | 920       | 920     |
| Gravity                    | $g$ (m s$^{-2}$) | 1.3    | 1.42     | 0.11      | 0.079   |
| Ice Thermal Expansion      | $\alpha$ (K$^{-1}$) | $1.62 \times 10^{-4}$ | $1.62 \times 10^{-4}$ | $1.70 \times 10^{-4}$ | $1.56 \times 10^{-4}$ |
| Thermal Diffusivity        | $\kappa$ (m s$^{-2}$) | $1.36 \times 10^{-6}$ | $1.36 \times 10^{-6}$ | $1.23 \times 10^{-6}$ | $1.47 \times 10^{-6}$ |
| Basal Viscosity            | $\eta$ (Pa s) | $3 \times 10^{14}$ | $3 \times 10^{14}$ | $3 \times 10^{14}$ | $3 \times 10^{14}$ |
| Thermal Conductivity       | $k$ (W m$^{-1}$ K$^{-1}$) | 3.52   | 3.34     | 3.69      | 4.2     |

Table 2: Thermal and physical properties of satellite ice shells.
Figure 1: Locations of our simulations in $Ra$-$\Delta \eta$ space. Lines show the boundaries between the constant-viscosity convection regime (I), transitional (or sluggish lid) regime (II), the stagnant lid regime (III), and no convection. The boundary between the sluggish and stagnant lid regimes (dashed line) is thought to lie near $\Delta \eta \sim 4(n + 1) \sim 3000$ for a Newtonian fluid ($n = 1$), but its exact location is not precisely defined (Solomatov, 1995).
Figure 2: Heat flow in regions of extension $F_{ext}$ as a function of $\Delta \eta$ for Europa (black), Ganymede (gray), Enceladus (blue/red), and Miranda (green). Dotted lines indicate values of $\Delta \eta$ where heat flows are outside the range inferred from geological characterization. Solid lines indicate values of $\Delta \eta$ where heat flows are consistent with the formation of each of the class of surface features we consider: Europa’s bands, Ganymede’s grooved terrain, Miranda’s coronae; and the south polar terrain and fold systems in the northern plains of Enceladus.
Figure 3: Strain rate in extensional regions ($\dot{\varepsilon}_{\text{ext}}$) as a function of viscosity contrast ($\Delta\eta$) for Europa (left), Ganymede (middle), and Enceladus (right). Strain rates depend weakly on ice shell thickness ($D$) and are reported for a range of values: $D = 10$ km (dotted), $D = 30$ km (light gray), $D = 50$ km (dark gray), and $D = 100$ km (black). Across all $\Delta\eta$ values in our study, strain rates in extensional regions match those for Europa's bands. Strain rates associated with grooved terrain formation are achieved for $\Delta\eta \gtrsim 10^{3.25}$. On Enceladus, strain rates in the south polar terrain imply $\Delta\eta < 10^{3.75}$ at that location, but the north polar extensional features may form for any $\Delta\eta$ in our study.
Appendix A. Scaling of Sluggish Lid Convection

A.1 Heat Flow

Solomatov (1995) and Solomatov and Moresi (2000) propose that the Nusselt number is related to the thickness of the upper and lower thermal boundary layers,

$$\text{Nu} \sim \frac{D}{\delta_0 + \delta_1},$$  \hspace{1cm} (A.1)

where $\delta_0$ is the thickness of the thermal boundary layer at the cold surface and $\delta_1$ is the thickness of the boundary layer at the warm base of the convecting layer.

For Newtonian fluids, the dependence on Rayleigh number, $\text{Nu} \propto Ra^{1/3}$, holds in both the stagnant lid and constant-viscosity regimes (Solomatov, 1995; Solomatov and Moresi, 2000). In stagnant lid convection, the value of $\text{Nu}$ is largely controlled by the thickness of the upper thermal boundary layer, which is too cold and stiff to permit advective heat transport. The thick stagnant lid serves as a “bottleneck” to heat transport because heat must be transferred across the lid by conduction. The upper thermal boundary layer has a thickness $\delta_0 \sim \theta^{4/3}Ra_i^{1/3}$ (Solomatov, 1995), where $Ra_i$ is the Rayleigh number (equation 1) where the temperature-dependent viscosity evaluated in the roughly isothermal convecting interior (in between the boundary layers), $\eta = \eta(T_i)$ (Solomatov and Moresi, 2000). The bottom thermal boundary layer thickness, $\delta_1 \sim \delta_0/\theta$. Because $\theta \gtrsim 8$ in the stagnant lid regime, $\delta_0$ overwhelmingly controls the value of $\text{Nu}$. This gives rise to the $Ra - \text{Nu}$ relationship, $\text{Nu} \sim \theta^{-4/3}Ra_i^{1/3}$ (Solomatov and Moresi, 2000). In icy satellite settings, $T_i$ is not known \textit{a priori}, so it is common, and usually not a bad assumption, to set $T_i \sim T_b$, which allows for a rough estimate of $\text{Nu}$ (see, e.g.,
In constant-viscosity convection, $\delta_0 = \delta_1$ and $T_i$ is equal to the average between the surface and basal temperatures.

Sluggish lid convection can be viewed as a transitional regime between these two end-members. Olson and Corcos (1980) suggest that in this regime, $Nu \propto Ra_i^{1/3}$. However, since $\eta_i$ is very poorly constrained for planets other than Earth, Moresi and Solomatov (1998) propose a scaling relationship of the form,

$$Nu = a Ra_i^b \exp(\theta/c),$$

in their work on mantle convection for Venus. This is also an advantageous approach for icy bodies (Barr, 2008). Here, $Ra_i$ is the Rayleigh number evaluated using $\eta = \eta(T_s)$, which can be calculated from the basal Rayleigh number, $Ra_0 = Ra_i / \exp(\theta)$. We expect $b \approx 1/3$ but consider it a free parameter. We note that Barr (2008) obtained values of $a$ and $c$ by fitting $Nu$ data (with $b \equiv 1/3$). However, these fits were constrained by relatively few simulations, all which were performed entirely in $1 \times 1$ computational domains.

The left panel of Figure A.4 shows how the values of $Nu$ obtained in our simulations vary as a function of $Ra_0$. We find that the slope of the $Ra - Nu$ relationship remains constant over several orders of magnitude in $Ra$ space. We note that we slightly underestimate $Nu$ at high $Ra$, which is likely due to our limited resolution. The $Ra - Nu$ relationship close to the critical Rayleigh number has a different dependence, likely $Ra^{1/2}$, which would be expected for extremely low-amplitude convection (Solomatov and Barr, 2007). However, a multivariate least squares fit on all of the $Nu$ values in our data set gives $a = 0.23 \pm 0.03$, $b = 0.32 \pm 0.007$, and $c = 9.8 \pm 0.12$. The right panel of Figure
A.4 shows how equation (A.2) compares to the data. The slight discrepancy between model and data at high $Nu$ is due to our numerical resolution.

Here, we are particularly interested in the relationship between physical properties of the ice shell and the heat flow in extensional regions, $Nu_{ext}$. Figure A.5 summarizes the values of $Nu_{ext}$ obtained in our study, and their relationship with $Ra_0$. One might expect that the relationship between this quantity and the Rayleigh number might be similar in form. We find that $Nu_{ext}$ can be described by,

$$Nu_{ext} = (0.48 \pm 0.09) Ra^{0.31 \pm 0.009} \exp\left(\frac{\theta}{14.21 \pm 2.7}\right).$$  \hspace{1cm} (A.3)

which is extremely similar to the scaling for $Nu$, with roughly a factor of $\sim 2$ difference between the two.

A.2 Strain Rate

In constant viscosity convection, the velocities in the upper thermal boundary layer, $v_0 \propto \delta_0$, giving rise to a relationship between $v_0$, $Ra_0$, and $\delta_0$ of form (Solomatov, 1995),

$$v_0 \propto Ra_0^{2/3} \frac{\kappa}{D}.$$  \hspace{1cm} (A.4)

where $\kappa/D$ is used to convert from non-dimensional velocities to velocities in meters per second. Barr (2008) find that the maximum $x$–velocity at the surface, $\max(|v_{x,sf}|) \sim a_v Ra_0^{b_v} (\kappa/D)$, where $a_v = 0.08$ and $b_v = 0.8$. Here, we are interested in the relationship between strain rates in extensional zones and $Ra_0$, which we can expect to follow

$$\dot{\varepsilon}_{ext} \sim \frac{v_0}{D} \propto Ra_0^{2/3} \frac{\kappa}{D^2}.$$  \hspace{1cm} (A.5)
This predicts that surface strain rates should be independent of the thickness of the ice shell, $D$. We fit our data for $\dot{\varepsilon}_{\text{ext}}$ to an equation of form,

$$\dot{\varepsilon}_{\text{ext}} = a_e R a_0^{b_e} \frac{\kappa}{D^2}, \quad (A.6)$$

where $a_e = 0.42 \pm 0.12$ and $b_e = 0.71 \pm 0.03$. Figure A.6 illustrates the values of $\dot{\varepsilon}_{\text{ext}}$ obtained in our study, and a comparison between these values and equation (A.6). For the same reasons we have errors in $Nu$ at low and high $Ra_0$ we find evidence of systematic error in the $\dot{\varepsilon}_{\text{ext}}$ data at low $Ra_0$ and high $Ra_0$. 

27
Figure A.4: Figure A1. (left) Dimensionless heat flow, $Nu$, from our simulations as a function of surface Rayleigh number, $Ra_0$. (right) Comparison between the values of $Nu$ from our simulations (“data”) with the scaling relationship (equation A.2 “model”).
Figure A.5: Figure A2. (left) Dimensionless heat flow in extensional zones, $N_{u_{ext}}$, from our simulations as a function of surface Rayleigh number, $Ra_0$. (right) Comparison between the values of $N_{u_{ext}}$ from our simulations with the scaling relationship (equation A.3).
Figure A.6: Figure A3. (left) Dimensionless values of strain rate in extensional zones, $\dot{\varepsilon}_{\text{ext}}$, from our simulations as a function of surface Rayleigh number, $Ra_0$. (right) Comparison between the values of $\dot{\varepsilon}_{\text{ext}}$ from our simulations ("data") with the scaling relationship (equation A.6; "model").
References

Barr, A. C., 2008. Mobile lid convection beneath Enceladus’ south polar terrain. Journal of Geophysical Research 113, E07009.

Barr, A. C., McKinnon, W. B., 2007. Convection in ice I shells and mantles with self-consistent grain size. J. Geophys. Res. 112.

Barr, A. C., Preuss, L. J., 2010. On the origin of south polar folds on Enceladus. Icarus 208, 499–503.

Barr, A. C., Showman, A. P., 2009. Heat Transfer in Europa’s Icy Shell. In: Europa, Robert T. Pappalardo, William B. McKinnon, Krishan K. Khurana, eds. University of Arizona Press, Tucson, pp. 405–430.

Beeman, M., Durham, W. B., Kirby, S. H., 1988. Friction of Ice. J. Geophys. Res. 93, 7625–7633.

Bland, M. T., Beyer, R. A., Showman, A. P., Dec. 2007. Unstable extension of Enceladus’ lithosphere. Icarus 192, 92–105.

Bland, M. T., McKinnon, W. B., 2012. Forming Europa’s folds: Strain requirements for the production of large-amplitude deformation. Icarus 221 (2), 694–709.

Bland, M. T., McKinnon, W. B., Showman, A. P., 2010. The effects of strain localization on the formation of Ganymede’s grooved terrain. Icarus 210 (1), 396–410.
Bland, M. T., Showman, A. P., 2007. The formation of Ganymede’s grooved terrain: Numerical modeling of extensional necking instabilities. Icarus 189, 439–456.

Cassen, P., Peale, S. J., Reynolds, R. T., Nov. 1980. Tidal dissipation in Europa - A correction. Geophys. Res. Lett. 7, 987–988.

Cassen, P., Reynolds, R. T., Peale, S. J., Sep. 1979. Is there liquid water on Europa? Geophys. Res. Lett. 6, 731–734.

Collins, G. C., Head, J. W., Pappalardo, R. T., Spaun, N. A., Jan. 2000. Evaluation of models for the formation of chaotic terrain on Europa. J. Geophys. Res. 105, 1709–1716.

Collins, G. C., McKinnon, W. B., Moore, J. M., Nimmo, F., Pappalardo, R. T., Prockter, L. M., Schenk, P. M., 2010. Tectonics of the outer planet satellites. Planetary Tectonics (11), 264.

Dombard, A. J., McKinnon, W. B., 2001. Formation of grooved terrain on Ganymede: Extensional instability mediated by cold, superplastic creep. Icarus 154, 321–336.

Goldsby, D. L., Kohlstedt, D. L., 2001. Superplastic deformation of ice: Experimental observations. J. Geophys. Res. 106, 11017–11030.

Greenberg, R., Croft, S. K., Janes, D. M., Kargel, J. S., Lebofsky, L. A., Lunine, J. I., Marcialis, R. L., Melosh, H. J., Ojakangas, G. W., Strom, R. G., 1991. Miranda. in Uranus, University of Arizona Press, Tucson, AZ, pp. 693–735.
Hammond, N. P., Barr, A. C., 2014a. Formation of Ganymede’s grooved terrain by convection-driven resurfacing. Icarus 227, 206–209.

Hammond, N. P., Barr, A. C., 2014b. Global resurfacing of Uranus’s moon Miranda by convection. Geology, submitted.

Han, L., Tobie, G., Showman, A. P., 2012. The impact of a weak south pole on thermal convection in Enceladus’ ice shell. Icarus 218, 320–330.

Head, J., Pappalardo, R., Collins, G., Belton, M. J. S., Giese, B., Wagner, R., Breneman, H., Spaun, N., Nixon, B., Neukum, G., Moore, J., Dec. 2002. Evidence for Europa-like tectonic resurfacing styles on Ganymede. Geophys. Res. Lett. 29, 2151.

Hurford, T. A., Helfenstein, P., Hoppa, G. V., Greenberg, R., Bills, B. G., 2007a. Eruptions arising from tidally controlled periodic openings of rifts on Enceladus. Nature 447, 292–294.

Hurford, T. A., Sarid, A. R., Greenberg, R., 2007b. Cycloidal cracks on Europa: Improved modeling and non-synchronous rotation implications. Icarus 186, 218–233.

Hussmann, H., Sohl, F., Spohn, T., 2006. Subsurface oceans and deep interiors of medium-sized outer planet satellites and large trans-neptunian objects. Icarus 185, 258–273.

Kehle, R. O., 1964. Deformation of the Ross Ice Shelf. Geol. Soc. Amer. Bull 75, 259–286.
Kirk, R. L., Stevenson, D. J., Jan. 1987. Thermal evolution of a differentiated Ganymede and implications for surface features. Icarus 69, 91–134.

Kivelson, M. G., Khurana, K. K., Volwerk, M., 2002. The permanent and inductive magnetic moments of Ganymede. Icarus 157, 507–522.

Kohlstedt, D. L., Evans, B., Mackwell, S. J., 1995. Strength of the lithosphere: Constraints imposed by laboratory experiments. J. Geophys. Res. 100, 17,587–17,602.

Malhotra, R., Dermott, S. F., 1990. The role of secondary resonances in the orbital history of Miranda. Icarus 85, 444–480.

McKinnon, W. B., 2006. On convection in ice I shells of outer solar system bodies, with specific application to Callisto. Icarus 183, 435–450.

Meyer, J., Wisdom, J., 2008. Tidal Evolution of Mimas, Enceladus, and Dione. Icarus 193, 213–223.

Moresi, L.-N., Solomatov, V. S., 1995. Numerical investigation of 2D convection with extremely large viscosity variations. Physics of Fluids 7, 2154–2162.

Moresi, L.-N., Solomatov, V. S., 1998. Mantle convection with a brittle lithosphere: Thoughts on the global tectonic styles of Earth and Venus. Geophys. J. Int. 133, 669–682.

Nimmo, F., Pappalardo, R. T., Giese, B., 2002. Effective elastic thickness and heat flux estimates on Ganymede. Geophys. Res. Lett. 29, 1158.
Ojakangas, G. W., Stevenson, D. J., 1989. Thermal state of an ice shell on Europa. Icarus 81, 220–241.

Olson, P., Corcos, G. M., 1980. A boundary layer model for mantle convection with surface plates. Geophys. J. Roy. Astr. Soc. 62, 195–219.

O’Neill, C., Nimmo, F., Feb. 2010. The role of episodic overturn in generating the surface geology and heat flow on Enceladus. Nat. Geosci. 3, 88–91.

Pappalardo, R. T., Collins, G. C., Head III, J. W., Helfenstein, P., McCord, T., Moore, J. M., Prockter, L. M., Schenk, P. M., Spencer, J. R., 2004. Geology of Ganymede. In: Jupiter: The Planet, Satellites & Magnetosphere. Cambridge University Press, New York, pp. 363–396.

Pappalardo, R. T., Head, J. W., Greeley, R., Sullivan, R. J., Pilcher, C., Schubert, G., Moore, W. B., Carr, M. H., Moore, J. M., Belton, M. J. S., 1998. Geological evidence for solid-state convection in Europa’s ice shell. Nature 391, 365–368.

Pappalardo, R. T., Reynolds, S. J., Greeley, R., 1997. Extensional tilt blocks on Miranda: Evidence for an upwelling origin of Arden Corona. J. Geophys. Res. 102, 13,369–13,379.

Pappalardo, R. T., Sullivan, R. J., 1996. Evidence for separation across a gray band on Europa. Icarus 123, 557–567.

Parmentier, E. M., Squyres, S. W., Head, J. W., Allison, M. L., 1982. The tectonics of Ganymede. Nature 295, 290–293.
Patel, J. G., Pappalardo, R. T., Head, J. W., Collins, G. C., Hiesinger, H., Sun, J., Oct. 1999. Topographic wavelengths of Ganymede groove lanes from Fourier analysis of Galileo images. J. Geophys. Res. 104, 24057–24074.

Patterson, G., Collins, G. C., Head, J. W., Pappalardo, R. T., Prockter, L. M., Lucchitta, B. K., Kay, J. P., 2010. Global geological mapping of Ganymede. Icarus 207 (2), 845–867.

Porco, C. C., Helfenstein, P., Thomas, P. C., Ingersoll, A. P., Wisdom, J., West, R., Neukum, G., Denk, T., Wagner, R., Roatsch, T., Kieffer, S., Turtle, E., McEwen, A., Johnson, T. V., Rathbun, J., Veverka, J., Wilson, D., Perry, J., Spitale, J., Brahic, A., Burns, J. A., DelGenio, A. D., Dones, L., Murray, C. D., Squyres, S., Mar. 2006. Cassini observes the active south pole of Enceladus. Science 311, 1393–1401.

Postberg, F., Kempf, S., Schmidt, J., Brilliantov, N., Beinsen, A., Abel, B., Buck, U., Srama, R., 2009. Sodium salts in E-ring ice grains from an ocean below the surface of Enceladus. Nature 459 (7250), 1098–1101.

Prockter, L. M., Head, J. W., Pappalardo, R. T., Sullivan, R. J., Clifton, A. E., Giese, B., Wagner, R., Neukum, G., 2002. Morphology of Europen bands at high resolution: A mid-ocean ridge-type rift mechanism. J. Geophys. Res. 107, 5028.

Prockter, L. M., Pappalardo, R. T., Aug. 2000. Folds on Europa: Implications for crustal cycling and accommodation of extension. Science 289, 941–944.
Roberts, J. H., Nimmo, F., 2008. Near-surface heating on Enceladus and the south polar thermal anomaly. Geophys. Res. Lett. 35, L09201.

Roth, L., Saur, J., Retherford, K. D., Strobel, D. F., Feldman, P. D., McGrath, M. A., Nimmo, F., 2014. Transient water vapor at Europa’s south pole. Science 343 (6167), 171–174.

Schenk, P. M., 1991. Fluid volcanism on Miranda and Ariel: Flow morphology and composition. Journal of Geophysical Research: Solid Earth 96 (B2), 1887–1906.

Schenk, P. M., McKinnon, W. B., 1989. Fault offsets and lateral crustal movement on Europa: Evidence for a mobile ice shell. Icarus 79 (1), 75–100.

Schubert, G., Anderson, J. D., Travis, B. J., Palguta, J., 2007. Enceladus: Present internal structure and differentiation by early and long-term radiogenic heating. Icarus 188, 345–355.

Schubert, G., Turcotte, D. L., Olson, P., 2001. Mantle Convection in the Earth and Planets. Cambridge University Press, New York.

Shoemaker, E. M., Lucchitta, B. K., Wilhelms, D. E., Plescia, J. B., Squyres, S. W., 1982. The geology of Ganymede. In: Satellites of Jupiter. pp. 435–520.

Showman, A. P., Han, L., 2005. Effects of plasticity on convection in an ice shell: Implications for Europa. Icarus 177, 425–437.
Showman, A. P., Malhotra, R., 1997. Tidal evolution into the Laplace resonance and the resurfacing of Ganymede. Icarus 127, 93–111.

Showman, A. P., Stevenson, D. J., Malhotra, R., 1997. Coupled orbital and thermal evolution of Ganymede. Icarus 129, 367–383.

Smith, B. A., Soderblom, L., Beebe, R., Bliss, D., Boyce, J., Brahic, A., Briggs, G., Brown, R., Collins, S., Cook, A., et al., 1986. Voyager 2 in the Uranian system: Imaging science results. Science 233, 43–64.

Solomatov, V. S., 1995. Scaling of temperature- and stress-dependent viscosity convection. Physics of Fluids 7, 266–274.

Solomatov, V. S., 2004. Initiation of subduction by small-scale convection. J. Geophys. Res. 109.

Solomatov, V. S., Barr, A. C., 2007. Onset of convection in fluids with strongly temperature-dependent, power-law viscosity 2. Dependence on the initial perturbation. Phys. Earth. Planet. Interiors 165, 1–13.

Solomatov, V. S., Moresi, L.-N., 2000. Scaling of time-dependent stagnant lid convection: Application to small-scale convection on Earth and other terrestrial planets. J. Geophys. Res. 105, 21795–21818.

Spencer, J. R., Pearl, J. C., Segura, M., Flasar, F. M., Mamoutkine, A., Romani, P., Buratti, B. J., Hendrix, A. R., Spilker, L. J., Lopes, R. M. C., Mar. 2006. Cassini encounters Enceladus: Background and the discovery of a south polar hot spot. Science 311, 1401–1405.
Spohn, T., Schubert, G., Feb. 2003. Oceans in the icy Galilean satellites of Jupiter? Icarus 161, 456–467.

Stempel, M. M., Barr, A. C., Pappalardo, R. T., 2005. Model constraints on the opening rates of bands on Europa. Icarus 177, 297–304.

Sullivan, R., Greeley, R., Homan, K., Klemaszewski, J., Belton, M. J. S., Carr, M. H., Chapman, C. R., Tufts, R., Head, J. W., Pappalardo, R., 1998. Episodic plate separation and fracture infill on the surface of Europa. Nature 391, 371–373.

Tittermore, W. C., Wisdom, J., 1989. Tidal evolution of the uranian satellites: II. An explanation of the anomalously high orbital inclination of Miranda. Icarus 78 (1), 63–89.

Trompert, R., Hansen, U., 1998. Mantle convection simulations with rheologies that generate plate-like behaviour. Nature 395, 686–689.

Wahr, J., Pappalardo, R. T., Barr, A. C., Crawford, Z., Gleeson, D., Stempel, M. M., Mullen, M. E., Collins, G. C., 2009. Modeling stresses on satellites due to nonsynchronous rotation and orbital eccentricity using gravitational potential theory. Icarus 200 (1), 188–206.

Waite, J. H., Combi, M. R., Ip, W.-H., Cravens, T. E., McNutt, R. L., Kasprzak, W., Yelle, R., Luhmann, J., Niemann, H., Gell, D., Magee, B., Fletcher, G., Lunine, J., Tseng, W.-L., Mar. 2006. Cassini Ion and Neutral Mass Spectrometer: Enceladus plume composition and structure. Science 311, 1419–1422.
Zimmer, C., Khurana, K. K., Kivelson, M. G., 2000. Subsurface oceans on Europa and Callisto: Constraints from Galileo magnetometer observations. Icarus 147, 329–347.