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Degree-one convection and the origin of Enceladus' dichotomy

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Abstract

Recently, the Cassini spacecraft has detected ongoing geologic activity near the south pole of Saturn's moon Enceladus. In contrast, the satellite's north-polar region is heavily cratered and appears to have been geologically inactive for a long time. We propose that this hemispheric dichotomy is caused by interior dynamics with degree-one convection driving the south-polar activity. We investigate a number of core sizes and internal heating rates for which degree-one convection occurs. The numerical simulations imply that a core radius of less than 100 ± 20 km and an energy input at a rate of 3.0 to 5.5 GW would be required for degree-one convection to prevail. This is within the range of the observed thermal power release near Enceladus' south pole. Provided that Enceladus is not fully differentiated, degree-one convection is found to be a viable mechanism to explain Enceladus' hemispheric dichotomy.

Keywords: Saturn, satellites; Enceladus; Geophysics; Interiors; Thermal histories

1. Introduction

Enceladus is among the few Solar System bodies for which ongoing geologic activity has been reported (Porco et al., 2006). The activity is concentrated near Enceladus' south pole, a recently resurfaced region dissected by extensional graben structures which exhibit anomalously high surface temperatures (Spencer et al., 2006) and are thought to be the source of geysers feeding Saturn's E-ring (Spahn et al., 2006). At a latitude of 55° the active region is enclosed by a concentric fold margin, indicative of compressional tectonic activity. This suggests that the activity is driven by a buoyant mantle plume, which is internally heated by tidal dissipation. Reorientation of the spin axis by true polar wander could have then caused the active region to migrate towards the pole (Nimmo and Pappalardo, 2006).

Voyager 2 images show that in contrast to the active south pole, Enceladus' north-polar region is heavily cratered and appears to have been geologically inactive for Gyr (Smith et al., 1982). The cause of this asymmetric appearance remains unknown, but an endogenic origin seems likely. This dichotomy is reminiscent of other hemispheric dichotomies observed on Solar System bodies, e.g. the lunar near-side–far-side dichotomy and the martian crustal dichotomy. Degree-one mantle convection has been suspected to cause these phenomena (Zhong et al., 2000; Zhong and Zuber, 2001; Roberts and Zhong, 2006), and it is conceivable that low-degree mantle convection is also responsible for the development of the hemispheric dichotomy of Enceladus.

One of the factors controlling the morphology of the convective flow pattern is the specific viscosity structure of the mantle. In particular, the presence of a low-viscosity zone near the surface and mechanical decoupling of the core from the mantle promote the formation of large-scale convection cells (Zhong and Zuber, 2001; Roberts and Zhong, 2006; Yoshida and Kageyama, 2006). Furthermore, the size of the core (Zhong et al., 2000) and the amount of internal heating (McNamara and Zhong, 2005; Stemmer et al., 2006) affect the wavelength of the flow. Czechowski and Leliwa-Kopystyński (2005) have shown that degree-one convection can develop in homogeneous mid-sized icy satellites and we will here investigate the range of core sizes and heating rates that allow for the development of a degree-one pattern in Enceladus.

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With a radius of 252 km Enceladus is one of the smallest mid-sized icy satellites, comparable in size to Mimas, its inner neighbor, and the uranian satellite Miranda. Enceladus has the highest rock-content of these three, and—apart from Titan—the largest rock-mass fraction of all the saturnian satellites. Pressures in Enceladus can reach about 12 MPa at the base of the ice shell for a fully differentiated model, and a central pressure of about 23 MPa for a homogeneous model (Hussmann et al., 2006b). These pressures are insufficient for structural phase transitions to occur within the ice, and we therefore assume that the interior of Enceladus consists of ice-I and rock (including metal), only.

It is likely that the relatively large rock content of Enceladus is linked to its geological activity. Unfortunately, the state of internal differentiation is still not well-constrained by suitable observations. Thomas et al. (2006) have concluded that Enceladus' ellipsoidal shape acquired from Cassini ISS measurements of satellite limb positions would be consistent with a homogeneous interior if hydrostatic equilibrium with respect to the satellite's present orbit configuration applies. In contrast, the enormous geological activity at Enceladus' south polar terrain strongly suggests that its interior is differentiated (Schubert et al., 2007). Since the shapes of tidally-locked, differentiated bodies would be less oblate than of those with homogeneous interiors spinning at the same rate (e.g., Castillo-Rogez, 2006), it can be envisioned that Enceladus is partly differentiated and acquired its ellipsoidal shape closer to Saturn (Sohl et al., 2006), i.e. the assumption of hydrostaticity does not apply. Furthermore, it is also conceivable that, even at present, demixing of ice and silicate inside Enceladus is progressing (Ziethe et al., 2006), similar to an evolution scenario suggested for Callisto (Nagel et al., 2004). Future Cassini measurements of second-degree gravitational field parameters from which the satellite's moment of inertia could be deduced will be particularly useful to distinguish between possible models of Enceladus' interior. In the following, we consider the entire range of core sizes from essentially homogeneous to fully differentiated interior structures.

2. Model description

We investigate Enceladus' interior dynamics using two-dimensional convection models and explore the range of core sizes and the amount of internal heating that lead to the development of large-scale convection cells. For each model, the silicate volume and mass fraction coefficients f_v and f_m are calculated self-consistently from mass-balance constraints. The silicate volume fraction is given by

$$f_v = \frac{\rho - \rho_{\mathrm{ice}}}{\rho_{\mathrm{sil}} - \rho_{\mathrm{ice}}}, \tag{1}$$

where ρ is the calculated mantle density and ρ_{ice} and ρ_{sil} are the ice-I and silicate density, respectively. The rock mass fraction f_m is then calculated from f_v by

$$f_m = \frac{\rho_{\mathrm{sil}} f_v}{\rho}. \tag{2}$$

We model Enceladus' interior as a suspension of rock particles in an icy matrix. This suspension will behave as a viscous continuum as long as the individual rock particles embedded in the ice are large enough to not interfere with the crystal structure, yet small enough such that their settling motion can be neglected with respect to the convective motion (Friedson and Stevenson, 1983). The effective viscosity of the rock–ice mixture is slightly larger than that of pure ice and can be approximated by

$$\eta = \eta_{\mathrm{ice}}\left(1 - \frac{f_v}{0.62}\right)^{-\beta}, \tag{3}$$

where $\beta = 1.55$ (Rudman, 1992).

The temperature dependence of the viscosity of ice is modeled by an Arrhenius-type law of the form

$$\eta_{\mathrm{ice}} = \eta_0 e^{A(T_m/T - 1)}, \tag{4}$$

where η_0 is the melting point viscosity of ice-I, T_m the pressure-dependent ice-I melting temperature, and A is a dimensionless parameter controlling the viscosity variations due to temperature changes. It is connected to the activation energy for the predominant creep mechanism. This viscosity law is supported by experimental data (Durham et al., 1997). Here we will assume T_m to be constant, as the pressure dependence of the melting temperature is very weak for the pressures encountered in Enceladus (e.g., Sotin et al., 1998). For the parameter values adopted here (see Table 1), such a viscosity law is consistent with grain sizes of $\sim$3 mm if diffusion creep is the rate-limiting deformation mechanism (Goldsby and Kohlstedt, 2001). To avoid numerical problems associated with the high-viscosity ratios arising from the low surface temperatures, a viscosity cut-off of 10^{25} Pa s is used.

The material properties of ice-I are known to be functions of temperature and we use the following parameterizations

$$k_{\mathrm{ice}}\ [\mathrm{W\,m^{-1}\,K^{-1}}] = \frac{488.12}{T\ [\mathrm{K}]} + 0.4685, \tag{5}$$

$$\alpha_{\mathrm{ice}}\ [\mathrm{K^{-1}}] = 2.5 \times 10^{-7} \cdot T\ [\mathrm{K}] - 1.25 \times 10^{-5}, \tag{6}$$

$$c_{\mathrm{ice}}\ [\mathrm{kJ\,kg^{-1}\,K^{-1}}] = 7.037 \cdot T\ [\mathrm{K}] + 185, \tag{7}$$

where k_{ice}, α_{ice} and c_{ice} are the ice-I thermal conductivity, thermal expansion coefficient and specific heat, respectively (Hobbs, 1974).

The material properties of the rock-ice mixture are then calculated from the properties of pure water ice and rock, the latter of which are assumed to be constant. The thermal conductivity k and thermal expansion coefficient α of the mixture are given by volume averages and the heat capacity c_p by the mass average of the components

$$k = (1 - f_v) \cdot k_{\mathrm{ice}} + f_v \cdot k_{\mathrm{sil}}, \tag{8}$$

$$\alpha = (1 - f_v) \cdot \alpha_{\mathrm{ice}} + f_v \cdot \alpha_{\mathrm{sil}}, \tag{9}$$

$$c_p = (1 - f_m) \cdot c_{\mathrm{ice}} + f_m \cdot c_{\mathrm{sil}}, \tag{10}$$

where, k_{ice}, k_{sil}, α_{ice}, α_{sil}, c_{ice} and c_{sil} are the ice-I and silicate thermal conductivity, thermal expansion coefficients and specific heat capacities, respectively.