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Note

Origin of a partially differentiated Titan

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ABSTRACT

Accretional temperature profiles for Saturn's large moon Titan are used to determine the conditions needed for accretion to avoid global melting as a function of the timing, duration, and nebular conditions of Titan's accretion. We find that Titan can accrete undifferentiated in a "gas-starved" disk even with modest quantities of ammonia mixed in with its ices. Simulations of impact-induced core formation are used to show that Titan can remain only partially differentiated after an outer Solar System late heavy bombardment capable of melting its outer layers, permitting some of its rock to consolidate into a core.

1. Introduction

Titan's interior state is important to interpretation of radio tracking data (Iess et al., 2010), its spin state (Stiles et al., 2008, 2010; Bills and Nimmo, 2008), and its atmospheric composition (Tobie et al., 2006). Here, we use recent models of satellite formation (Canup and Ward, 2002, 2006; Barr and Canup, 2008) to assess whether Titan can avoid fully differentiating while it accretes. We use geophysical models of impact-induced core formation in icy satellites (Barr and Canup, 2010) to assess whether Titan can remain partially differentiated after an outer Solar System late heavy bombardment (LHB) predicted by recently proposed scenarios for the dynamical evolution of the outer Solar System (Levison et al., 2001; Tsiganis et al., 2005; Gomes et al., 2005).

2. Titan can accrete undifferentiated

Historically, satellite formation models focus on the evolution of a minimum-mass subnebula (MMSN), wherein the masses of the currently observed satellites are combined with gas to create a massive solar-composition disk around the parent planet. MMSN models predict dense, gas-rich protosatellite disks and rapid formation of satellites whose interiors are heated close to, or well above their melting points (Stevenson et al., 1986; Schubert et al., 1986; Squyres et al., 1988; Canup and Ward, 2002; Barr and Canup, 2008). Canup and Ward (2002) instead proposed satellite growth in a "gas-starved" disk supplied by the slow inflow of gas and $\leqslant$1 m-sized rock and ice particles from solar to planetary orbit. In the Canup and Ward model, the jovian and saturnian satellites form during the waning stages of gas accretion by the planets, implying that the observed satellites form in much lower-density disks than the MMSN, and that the satellites accrete in 10^5–10^6 yr, at a rate controlled by the supply of ice/rock solids to the disk from the solar nebula (Canup and Ward, 2002, 2006, 2009; Barr and Canup, 2008).

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2.1. Methods

We calculate accretion temperature profiles for Titan (final mass $M_S = 1.345 \times 10^{26}$ g and radius R_S = 2575 km) by balancing radiation from its surface, heating of accreted material from its initial temperature (assumed to be the disk temperature, T_d) to the satellite surface temperature T, and accretional heating (Squyres et al., 1988; Barr and Canup, 2008),

$$\bar{\rho} C_p (T - T_d)\frac{dr}{dt} = \frac{1}{2}\frac{\dot{M} v_i^2}{4\pi r^2} - \sigma_{SB}\left(T^4 - T_d^4\right), \quad (1)$$

with time-dependent satellite radius r, Stefan–Boltzmann constant σ_{SB}, satellite specific heat C_p, satellite mean density $\bar{\rho}$, and impact velocity v_i. The mass accretion rate is $\dot{M} = \sigma_s \Omega \pi r^2 F_g$ with the gravitational focusing factor $F_g \equiv 1 + (v_{esc}/v_\infty)^2$, surface mass density of solids in the protosatellite disk σ_s, and satellite angular velocity $\Omega = \sqrt{GM_p/a^3}$, where M_p is Saturn's mass and a is Titan's semi-major axis. We set $\tau_{acc}(r) \equiv M/\dot{M} \propto r/F_g$, where M is the time-dependent satellite mass. The impact velocity is $v_i^2 = v_{esc}^2 + v_\infty^2$, where v_∞ is a function of the satellite's time dependent escape velocity ($v_{esc} = (2GM(t)/r(t))^{1/2}$), and we consider $v_\infty = v_{esc}/2$, corresponding to a constant gravitational focusing factor $F_g = 5$, so that the satellite's growth is in the orderly rather than runaway regime (e.g., Barr and Canup, 2008). With this value of F_g, impact velocities during accretion are 1–3 km/s, with 3 km/s sufficient to create a small pool of melt at each impact site, but too low to vaporize Titan's surface (Pierazzo et al., 1997).

After accretion, material at radius r is heated by ΔT_r,

$$\Delta T_r(r,t) = \int_{t_f(r)}^{\infty} m_r q_{26}(t)\,dt = \frac{m_r q_{26}(0)}{C_p \lambda_{26}} \exp(-\lambda_{26} t_f(r)), \quad (2)$$

due to radiogenic heating, where $q_{26}(0) = 1.82 \times 10^{-3}$ erg/g is the ^{26}Al heating rate at the time of CAI formation (Barr and Canup, 2008) assuming initial $^{26}Al/^{27}Al = 5.85 \times 10^{-5}$ (Thrane et al., 2006), $\lambda_{26} = 9.68 \times 10^{-7}$ yr^{-1}, and $t_f(r)$ is the time at which a layer at radius r forms, $t_f(r) = t_{start} + \tau_{acc}(r/R_S)$ (Barr and Canup, 2008), where t_{start} is the time at which the satellite starts accreting relative to CAI formation. By integrating to $t \to \infty$ we are assuming that all ^{26}Al heating is retained in the satellite (Barr and Canup, 2008). The rock mass fraction is $m_r = (\rho_r(\bar{\rho} - \rho_i))/(\bar{\rho}(\rho_r - \rho_i))$. We use a simple interior model that assumes constant ice and rock densities as a function of depth. We consider $\rho_r = 3.0$ g/cm^3, mid-way between the density of Prinn–Fegley rock and CI chondrite, two endmember compositions for the rocky component of Callisto used by Mueller and McKinnon (1988). The ice density, ρ_i = 1.4 g/cm^3, is chosen to represent the compressed densities of the water ice phases V–VII, whose densities at atmospheric pressure range from 1.2 to 1.5 g/cm^3 (Hobbs, 1974), which may exist deep in Titan's interior. With these densities, Titan has m_r = 0.48. The specific heat, $C_p \approx m_r\ C_{p,r} + (1 - m_r)C_{p,i}$ = 1.429×10^7 erg/g K, where the specific heat of rock $C_{p,r} = 7 \times 10^6$ erg/g K and the specific heat of ice $C_{p,i} = 2.1 \times 10^7$ erg/g K.

To determine whether melting occurs, we compare accretional temperature profiles generated using Eqs. (1) and (2) to the pressure-dependent melting point of water ice, T_m. Data from Hobbs (1974) are used to construct $T_m(P)$ as a function of depth in the satellite (Barr and Canup, 2008), and the interior pressure P at radius r_i from the satellite's center is calculated assuming the unmelted satellite has a uniform density, $P(r_i, t) = (2\pi/3)G\bar{\rho}^2(r^2(t) - r_i^2)$, where r is the radius of the growing satellite at time t during accretion. The temperature at which an ice/ammonia mixture becomes completely melted is determined using $T_m(P,X) \approx T_o + AP + BP^2 - CX - DX^2$, where T_o = 273.2 K, $A = -7.95 \times 10^{-9}$ K/Poise, $B = -9.6 \times 10^{-19}$ K/Poise^2, C = 53.8, D = 650, and $0 \leqslant X \leqslant 1$ is the ammonia concentration by mass in the liquid phase (Leliwa-Kopystyński et al., 2002). During accretion when the satellite is heated at its surface by impacts, the temperature increases as a function of radius and is highest at the surface. Thus, melting is most likely to occur first at the ice I–III phase boundary at a depth of ~100 km in Titan, and we do not include the effect of ammonia on the melting points of the high-pressure ice phases (Grasset and Pargamin, 2005). We check for melting by comparing $T(r) + \Delta T_r$ to the pressure-dependent melting curves in the satellite after the accretion of each layer. We then identify conditions necessary to avoid melting during accretion and yield an initially undifferentiated Titan.

2.2. Results

Solution of Eq. (1) in the absence of radiogenic heating (Barr and Canup, 2008),

$$\tau_{acc,min} = \frac{\bar{\rho}r\left[\frac{4\pi}{3}\frac{F_g}{(F_g-1)}\bar{\rho}Gr^2 - C_p(T - T_d)\right]}{3\sigma_{SB}(T^4 - T_d^4)} \quad (3)$$

to give the absolute minimum accretion time scale for formation of an undifferentiated Titan, $\tau_{acc,min}$. For Titan to avoid accretional melting, $T \leqslant 253$ K at a depth of 100 km when it reaches its final radius R_s = 2575 km (i.e., r = 2475 km). This implies $\tau_{acc,min} \geqslant 0.8$ Myr for T_d = 100 K. If the impacts that assemble Titan deposit accretional energy at depth, if Titan accretes in a gas-rich environment, has ammonia in its interior, and/or experiences radiogenic heating, longer accretional time scales are required to avoid large-scale melting during formation. Fig. 1a illustrates how $\tau_{acc,min}$ varies as a function of T_d and ammonia concentration in the absence of radiogenic heating. If the ammonia concentration for Titan ~15%, $\tau_{acc} \geqslant 1.3$ Myr is required to avoid complete ice/ammonia melting for T_d = 100 K.

Fig. 1b–d illustrates constraints on the timing and duration of formation of an undifferentiated Titan as a function of T_d, τ_{acc}, and X. Considering accretional and radiogenic heating, a Titan without ammonia (X = 0) must finish forming no earlier than t_{end} = 4.26 Myr after CAI formation to avoid melting during accretion for T_d = 100 K. For T_d = 70 K, $t_{end} \geqslant 4.1$ Myr to avoid melting. The corresponding accretion time scale for Titan must be $\geqslant$0.8 Myr for $T_d \geqslant 70$ K. If Titan contains 5% ammonia by mass, the lowest temperature for complete melting of its outer icy layers is T_m = 248 K at 209 MPa (compared to T_m = 253 K for pure water ice). In this case, to avoid melting, Titan must finish forming no earlier than 4.33 Myr for T_d = 100 K. If Titan contains 15% ammonia by mass, it must finish forming no earlier than 4.8 Myr after CAI's for T_d = 100 K.

In addition to accreting slowly, an undifferentiated Titan must also be assembled from objects small enough to deposit their energy in a boundary layer close to the surface where it can be removed by radiation between successive overlapping impacts (an implicit assumption of Eq. (1)). Small impactors, ~50–100 m in radius, are likely in the Canup and Ward model (see Appendix B of Barr and Canup, 2008). If impactors are large, and a significant fraction of impact energy is deposited at depth, longer accretion time scales and later accretion are required (Barr and

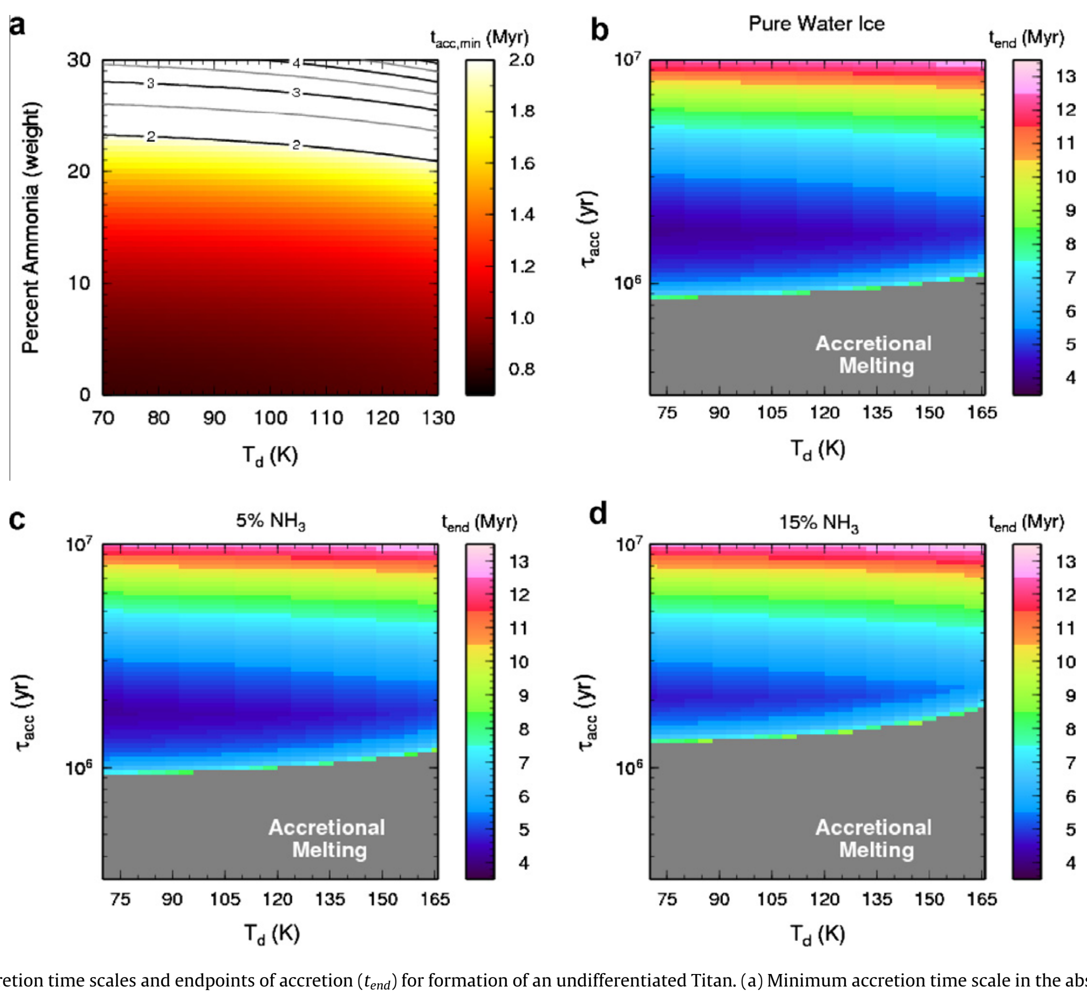

Fig. 1. Critical accretion time scales and endpoints of accretion (t_{end}) for formation of an undifferentiated Titan. (a) Minimum accretion time scale in the absence of radiogenic heating as a function of disk mid-plane temperature (T_d) and mass fraction of ammonia. (b) Earliest endpoint of accretion of an undifferentiated Titan as a function of T_d and τ_{acc}. (c) Same as (b), but for 5% ammonia, which decreases the minimum melting temperature to 248 K at a depth of 100 km. (d) Same as (b), for 15% ammonia, which gives T_m = 230 K at 100 km depth. As ammonia concentration is increased, accretion must occur more slowly and end later in Solar System history for Titan to remain undifferentiated during formation.

Canup, 2008). Titan must also be assembled from objects which are themselves undifferentiated, containing small rock grains rather than large fully formed rock cores in order to form in an undifferentiated state (Barr and Canup, 2008). These conditions are implicit assumptions of Eqs. (1) and (2).

3. Titan remains partially differentiated after the late heavy bombardment

A leading theory is that the lunar late heavy bombardment (LHB) was triggered by the dynamical evolution of the outer Solar System (Levison et al., 2001; Gomes et al., 2005). In the so-called Nice model, gravitational interactions between the four outer planets and a $\sim 35M_\oplus$ disk of icy planetesimals drives the outer planets to migrate to their current orbits (Tsiganis et al., 2005). It has been proposed that ~700 Myr after planet formation, Jupiter and Saturn cross a mutual mean-motion resonance, destabilizing the planetesimal disk (which has been depleted to $M_D \approx 20M_\oplus$ at the time of the LHB), and causing 8×10^{21} g of cometary material, and a comparable mass of asteroids, to impact Earth's Moon over a 10–100 Myr time span (Gomes et al., 2005).

Any model that invokes an outer Solar System source for the LHB will produce an intense bombardment of Titan. During an outer Solar System LHB, Titan would receive an impacting mass ~35 times larger than Earth's Moon (Zahnle et al., 2003), or $M^T_{\rm LHB} = 3 \times 10^{23}$ g due to gravitational focusing by Saturn. The cometary impactors originate in heliocentric orbit and impact Titan with $\langle v_i \rangle$ = 10.5 km/s (Zahnle et al., 2003) and deposit their impact energy at depth in the satellite, in contrast to smaller and low-velocity impactors during its accretion. Such large LHB impacts (see below) would melt the outer layers of Titan, permitting any rock suspended in the ice to sink rapidly to its center, creating a rock core (Tonks et al., 1997; Barr and Canup, 2010). If enough potential energy is released by partial differentiation to melt Titan's ice, runaway differentiation will occur (Barr and Canup, 2010).

3.1. Methods

We use a numerical model of impact-induced core formation in an icy satellite developed by Barr and Canup (2010) to simulate the effect of an outer Solar System LHB on Titan for a range of LHB masses. As per arguments in Section 2, Titan may accrete undifferentiated if it forms slowly and from small impactors. We assume an initially uniform-density ice/rock Titan with rock density ρ_r = 3.0 g/cm^3, ρ_i = 1.4 g/cm^3, and rock fraction ϕ_o = 0.30.

Before the LHB, Titan's interior must convect to avoid melting from long-lived radiogenic species (cf., Friedson and Stevenson, 1983). During the period of time between the endpoint of accretion and the onset of convection, radiogenic heating will build up in its interior, however, it is unlikely that this will cause global melting. Over the first billion years of Solar System history, the decay of ^{40}K → ^{40}Ca is the dominant contributor to radiogenic heating, providing $q(0) = 1.43 \times 10^{-7}$ erg/s per gram of CI chondrite, decaying with a half-life of 1.39 Gyr. The onset of convection in the outer ice shell of Titan may take (Zaranek and Parmentier, 2004) $t_o \approx (500/\kappa)((\rho\alpha\Delta T_i g)/(\eta_o \kappa))^{-2/3}$, with satellite density ρ = 1.88 g/cm^3, coefficient of thermal expansion $\alpha = 10^{-4}$ 1/K, temperature difference driving convection $\Delta T_i \sim 20$ K (Barr and Canup, 2008), gravity g = 130 cm/s^2, ice viscosity η_o, and thermal diffusivity $\kappa \sim 10^{-2}$ cm^2/s. In the outer ice I shell, $\eta_o \sim 10^{15}$ Poise, implying $t_o \sim 2 \times 10^6$ yr. In the first 2 Myr after accretion, we estimate that Titan's interior warms by less than 1 K due to ^{40}K decay. In the deep interior, where the dominant ice phases have higher viscosities, and the local gravity is lower, it can take $t_o \sim 100$ Myr to start convection (Barr and Canup, 2008). During this period, the temperature rise from ^{40}K is ~10 K.

A Monte Carlo approach is used to select random impact angle, position, v_i and impactor radius r_p. We set $\langle v_i \rangle$ = 10.5 km/s (Zahnle et al., 2003). Because an outer Solar System LHB is thought to populate Jupiter's trojan asteroids (Morbidelli et al., 2005), impactor radii are drawn from a double-power law size distribution constrained by that population: for small impactors, $dN_{sm}/dr_p \propto r_p^{-3}$, and for large impactors, $dN_{bg}/dr_p \propto r_p^{-6.5}$ (Jewitt et al., 2000; Morbidelli et al., 2005). The division between large and small impactors occurs at an absolute visual magnitude $V \approx 9$. Using an albedo p_v = 0.056 (Morbidelli et al., 2005), this corresponds to and a transition radius r_0 = 47 km. The most likely impactor size in the distribution is $\langle r_p \rangle$ = 30 km.

Titan's interior is represented by $\sim 10^7$ cubic elements with sides of 15 km that initially contain a uniform rock volume fraction ϕ_0. Each impact induces melting in a roughly spherical region beneath the impact site. We use the scaling relations from Barr and Canup (2010) to determine the radius of the region where the volume fraction of the melt is >50% post-impact. This degree of melting allows the rock particles in this region to sink to the bottom of the melt pool, where they coalesce and rapidly sink to the core of the satellite (e.g., Tonks et al., 1997) within a few thousand years (much less than the duration of the LHB Barr and Canup, 2010). The melted region is described by a sphere of radius $(r_{melt}/r_p) = 4.09\ (v_i/10.5\ {\rm km/s})^{0.6}$ buried at depth $(z_{melt}/r_p) = 2.40\ (v_i/10.5\ {\rm km/s})^{0.47}$ (Barr and Canup, 2010).

The core formation model sums the volume fraction of rock, ϕ, from impact-melted elements to determine the amount of rock added to the core from each impact. That amount of rock is removed from impact-melted elements in the icy mantle and new rock core elements are added to the core's outer edge, replacing ice/rock elements at the core's outer edge in a radially symmetric fashion. The transfer of rock mass from the outer mantle of the satellite to the core generates heat from the liberated gravitational potential energy. If the energy released by the impact-induced ice/rock separation is sufficient to melt the remainder of the satellite's ice, the process becomes energetically self-sustaining and will drive itself to completion (so-called runaway differentiation; Friedson and Stevenson, 1983). A simple estimate is that runaway differentiation of the entire satellite will occur if $\Delta E_{gr} \geqslant (1 - m_r)(1 - (x^3/\phi_0))M_S L$, where $\Delta E_{gr} = E_{gr,f} - E_{gr,i}$ is the difference in gravitational potential energy between the initial uniform-density and final differentiated states (Friedson and Stevenson, 1983), $L = 3 \times 10^9$ erg/g is a representative latent heat for the various water ice phases (Kirk and Stevenson, 1987), and x is the fractional radius of the rocky core ($x \equiv R_{core}/R_S$). We calculate ΔE_{gr} using $E_{gr,f} = \int (Gm/r(m))dm$ based on the model satellite's final heterogeneous density structure, and $E_{gr,i} = -(3/5)GM_S^2/R_S$. The LHB impactors are delivered on a time scale short compared to solid-state heat transport, so these processes do not affect model outcomes and are not included (Barr and Canup, 2010). For each LHB mass we calculate 100 bombardment histories to determine the probability of runaway differentiation.

3.2. Results

Fig. 2 illustrates that an initially undifferentiated Titan has a >99% probability of avoiding runaway differentiation for $M^T_{\rm LHB} \leqslant 4.8 \times 10^{23}$ g. This corresponds to an icy planetesimal disk mass at the time of the LHB, $M_D \leqslant 32M_\oplus$, where $M_D \approx (M^T_{\rm LHB}/1.5 \times 10^{22}\ {\rm g})M_\oplus$. Our prior work shows that Callisto avoids complete differentiation if $M_D \leqslant 21M_\oplus$ (Barr and Canup, 2010). Titan avoids runaway differentiation during an outer Solar System LHB of magnitude and distribution predicted by the Nice model.

The likelihood of avoiding complete differentiation during a Nice-model LHB is only weakly dependent on the assumed ice/rock density and the size distribution of impactors. If the size distribution of LHB impactors changes slope at absolute magnitude V = 10, $\langle r_p \rangle$ = 20 km. Fig. 2 shows that $M^T_{\rm LHB} \gg$ predicted by the Nice model is required to trigger runaway differentiation if the impactors are this small. In our "low-density" model, a lower limit for the density of Titan's ice, ρ_i = 1.2 g/cm^3, is paired with the density of CI chondrite, ρ_r = 2.8 g/cm^3 (Mueller and McKinnon, 1988) to give ϕ_o = 0.43. In our "high-density" model, ρ_i = 1.5 g/cm^3, representative of the compressed densities of the high-pressure ice polymorphs, and a plausible upper limit on the density of Titan's rock, ρ_r = 3.8 g/cm^3. Fig. 2 illustrates that regardless of assumed ice/rock densities, a Nice-model LHB does not fully differentiate Titan. Fig. 3 shows a sample partially differentiated Titan interior resulting from a nominal Nice-model LHB.

4. Discussion

We find that the timing and duration of formation required for an undifferentiated Titan is similar to the timetable required to accrete an undifferentiated Callisto (Barr and Canup, 2008). If satellite formation occurred at Saturn at the same time or later as it did at Jupiter, and over comparable or slower timescales, an undifferen-

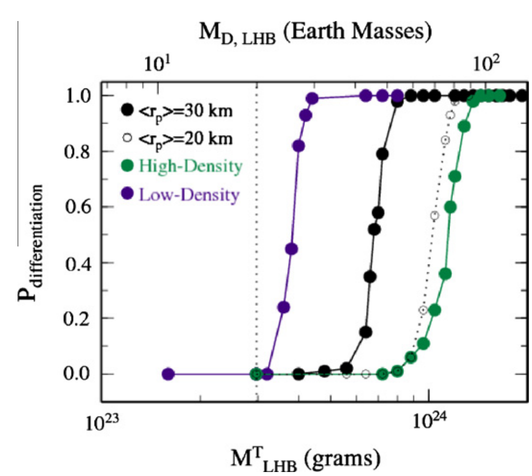

Fig. 2. Probability of triggering impact-induced runaway differentiation in Titan during an outer Solar System late heavy bombardment for our nominal case, $\langle r_p \rangle$ = 30 km, ρ_i = 1.4 g/cm^3, ρ_r = 3.0 g/cm^3 (black dots). The likelihood of triggering runaway differentiation in Titan during the LHB is only weakly dependent on the size distribution of impactors; if $\langle r_p \rangle$ is decreased to 20 km (open circles), the probability of differentiation during a Nice-model LHB is still $\leqslant$1%. The probability of runaway differentiation is also $\leqslant$1% if Titan's ice and rock have low densities (purple) or high densities (green), spanning the range of plausible initial rock fractions between ϕ = 0.17 and ϕ = 0.43.

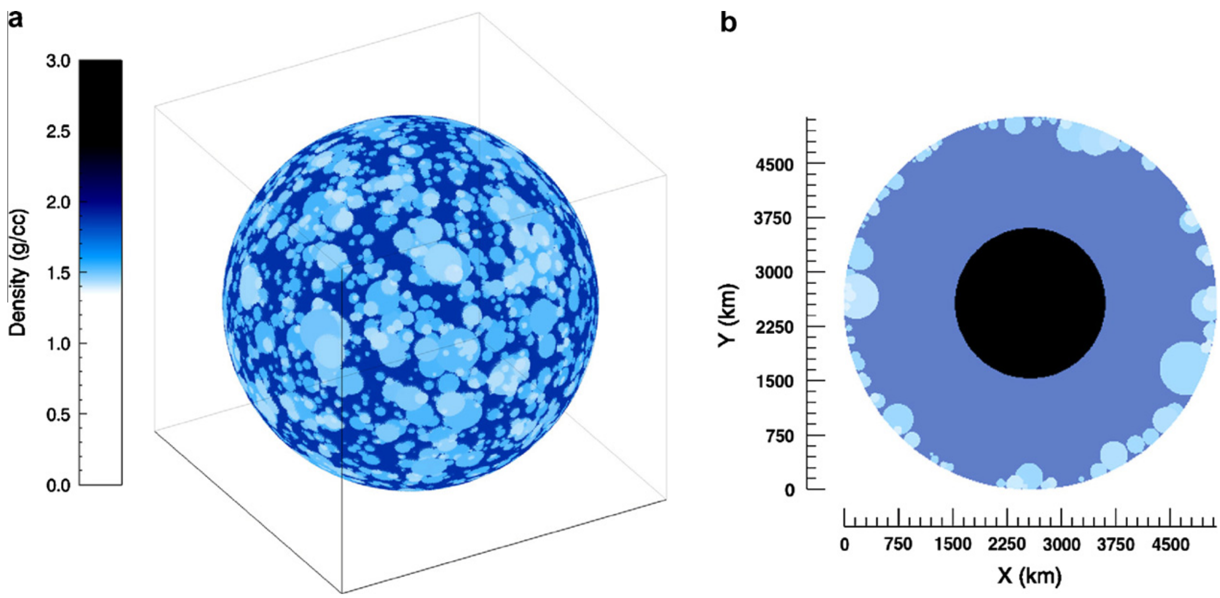

Fig. 3. (a) Post-LHB density (colors) of Titan's surface as a function of latitude and longitude for our nominal ice/rock composition model, $\langle r_p \rangle$ = 30 km, and $M_D = 20M_{\oplus}$. Dark blue indicates $\phi_0 = 0.30$ and $\rho = 1.88$ g/cm^3, black indicates $\rho = 3.0$ g/cm^3, and light blue/white indicates rock-free ice $\rho \sim 1.4$ g/cm^3. (b) Slice through the globe in (a), illustrating Titan's partially differentiated state after the LHB, with a rocky core (black), ice/rock mantle (blue) and ice-rich upper surface (light blue/white).

tiated interior state for Titan post-accretion can occur, although partially differentiated states could also result. Avoiding differentiation during accretion also requires that Titan form without a thick atmosphere (e.g., Stevenson et al., 1986) and from small planetocentric impactors. The formation of a mixed ice/rock Titan with limited accretional melting and vaporization lends support to a deep interior source for atmospheric methane, which could be liberated due to the satellite's later interior evolution (Tobie et al., 2006). Formation of an undifferentiated Titan is precluded in an MMSN because of short accretion time scales (e.g., Stevenson et al., 1986).

An outer Solar System LHB implies that Titan was subjected to an intense bombardment from heliocentric impactors at velocities $v_i \sim 10$ km/s (Zahnle et al., 2003; Barr and Canup, 2010). For the LHB mass predicted by the Nice model, Titan remains only partially differentiated. These conclusions are robust against variations in the density of Titan's rock and ice, and plausible variations in the impactor size distribution. In the context of the Tobie et al. (2006) evolution scenario, LHB-induced formation could have squeezed out clathrate-rich ices from the primordial core, ~700 Myr after its formation, jump-starting the gradual liberation of methane into the atmosphere. Vigorous convection is required to prevent Titan from melting over its long term thermal evolution and remain partially differentiated at present (Friedson and Stevenson, 1983).

If the LHB had a significant outer Solar System source (Levison et al., 2001; Gomes et al., 2005), Titan should have an outer rock-poor layer, an ice/rock mantle, and a rocky core occupying at least 40% of its radius. By calculating the moment of inertia of model titans based on their heterogeneous post-LHB density structures (and neglecting compression of ice and rock at depth), we find that Titan's $C/MR^2 \lesssim 0.38$ if an LHB of the magnitude predicted by the Nice model triggered partial differentiation. By analogy with Callisto, where compression of ice and rock decreases its C/MR^2 by 5% (McKinnon, 1997), we estimate that compression may plausibly decrease the upper limit on Titan's C/MR^2 to 0.36. Our estimates of Titan's C/MR^2 are upper limits only: melting during Titan's formation and/or later thermal evolution can decrease the moment of inertia, plausibly to the estimated value, $C/MR^2 = 0.3419 \pm 0.0005$ (Iess et al., 2010).

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References

Barr, A.C., Canup, R.M., 2008. Constraints on gas giant satellite formation from the interior states of partially differentiated satellites. Icarus 198, 163–177.

Barr, A.C., Canup, R.M., 2010. Origin of the Ganymede–Callisto dichotomy by impacts during the late heavy bombardment. Nat. Geosci. 3, 164–167.

Bills, B.G., Nimmo, F., 2008. Forced obliquity and moments of inertia of Titan. Icarus 196, 293–297.

Canup, R.M., Ward, W.R., 2002. Formation of the Galilean satellites: Conditions of accretion. Astron. J. 124, 3404–3423.

Canup, R.M., Ward, W.R., 2006. A common mass scaling for satellite systems of gaseous planets. Nature 441, 834–839.

Canup, R.M., Ward, W.R., 2009. Origin of Europa and the Galilean satellites. In: Pappalardo, R.T., McKinnon, W.B., Khurana, K.K. (Eds.), Europa. University of Arizona Press, Tucson, AZ, pp. 59–83.

Friedson, A.J., Stevenson, D.J., 1983. Viscosity of rock–ice mixtures and applications to the evolution of icy satellites. Icarus 56, 1–14.

Gomes, R., Levison, H.F., Tsiganis, K., Morbidelli, A., 2005. Origin of the cataclysmic late heavy bombardment period of the terrestrial planets. Nature 435, 466–469.

Grasset, O., Pargamin, J., 2005. The ammonia–water system at high pressures: Implications for the methane of Titan. Planet Space Sci. 53, 371–384.

Hobbs, P.V., 1974. Ice Physics. Clarendon Press, Oxford.

Iess, L., Rappaport, N.J., Jacobson, R.A., Racioppa, P., Stevenson, D.J., Tortora, P., Armstrong, J.W., Asmar, S.W., 2010. Gravity field, shape, and moment of inertia of Titan. Science 327, 1367–1369.

Jewitt, D.C., Trujillo, C.A., Luu, J.X., 2000. Population and size distribution of small jovian trojan asteroids. Astron. J. 120, 1140–1147.

Kirk, R.L., Stevenson, D.J., 1987. Thermal evolution of a differentiated Ganymede and implications for surface features. Icarus 69, 91–134.

Leliwa-Kopystyński, J., Maruyama, M., Nakajima, T., 2002. The water–ammonia phase diagram up to 300 MPa: Application to icy satellites. Icarus 159, 518–528.

Levison, H.F., Dones, L., Chapman, C.R., Stern, S.A., Duncan, M.J., Zahnle, K., 2001. Could the lunar "late heavy bombardment" have been triggered by the formation of Uranus and Neptune? Icarus 151, 286–306.

McKinnon, W.B., 1997. NOTE: Mystery of Callisto: Is it undifferentiated? Icarus 130, 540–543.

Morbidelli, A., Levison, H.F., Tsiganis, K., Gomes, R., 2005. Chaotic capture of Jupiter's trojan asteroids in the early Solar System. Nature 435, 462–465.

Mueller, S., McKinnon, W.B., 1988. Three-layered models of Ganymede and Callisto – Compositions, structures, and aspects of evolution. Icarus 76, 437–464.

Pierazzo, E., Vickery, A.M., Melosh, H.J., 1997. A reevaluation of impact melt production. Icarus 127, 408–423.

Schubert, G., Spohn, T., Reynolds, R.T., 1986. Thermal histories, compositions, and internal structures of the moons of the Solar System. In: Burns, J.A., Matthews, M.S. (Eds.), Satellites. University of Arizona Press, Tucson, AZ, pp. 224–292.

Squyres, S.W., Reynolds, R.T., Summers, A.L., Shung, F., 1988. Accretional heating of the satellites of Saturn and Uranus. J. Geophys. Res. 93, 8779–8794.

Stevenson, D.J., Harris, A.W., Lunine, J.L., 1986. Origins of satellites. In: Burns, J.A., Matthews, M.S. (Eds.), Satellites. University of Arizona Press, Tucson, AZ, pp. 39–88.

Stiles, B.W., and 14 colleagues, 2008. Determining Titan's spin state from Cassini RADAR images. Astron. J. 135, 1669–1680.

Stiles, B.W., and 14 colleagues, 2010. ERRATUM: "Determining Titan's spin state from Cassini RADAR images". Astron. J. 139, 311.

Thrane, K., Bizzarro, M., Baker, J.A., 2006. Brief formation interval for refractory inclusions and uniform distribution of ^{26}Al in the early Solar System. Astrophys. J. 646, L159–L162.

Tobie, G., Lunine, J.I., Sotin, C., 2006. Episodic outgassing as the origin of atmospheric methane. Nature 440, 61–64.
Tonks, W.B., Pierazzo, E., Melosh, H.J., 1997. Impact-induced differentiation in icy bodies. Icarus, unpublished.
Tsiganis, K., Gomes, R., Morbidelli, A., Levison, H.F., 2005. Origin of the orbital architecture of the giant planets of the Solar System. Nature 435, 459–461.
Zahnle, K., Schenk, P.M., Levison, H.F., 2003. Cratering rates in the outer Solar System. Icarus 163, 263–289.
Zaranek, S.E., Parmentier, E.M., 2004. The onset of convection in fluids with strongly temperature-dependent viscosity cooled from above with implications for planetary lithospheres. Earth Planet. Sci. Lett. 224, 371–386.