Titan's Interior Structure and Dynamics After the Cassini-Huygens Mission

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Abstract

The Cassini-Huygens mission that explored the Saturn system during the period 2004-2017, revolutionized our understanding of Titan, the only known moon with a dense atmosphere and the only body, besides Earth, with stable surface liquids. Its predominantly nitrogen atmosphere also contains a few percent of methane that is photolyzed on short geological timescales to form ethane and more complex organic molecules. The presence of a significant amount of methane and ^{40}Ar, the decay product of ^{40}K, argue for exchange processes from the interior to the surface. Here we review the information that constrains Titan's interior structure. Gravity and orbital data suggest that Titan is an ocean world, which implies differentiation into a hydrosphere and a rocky core. The mass and gravity data complemented by equations of state constrain the ocean density and composition as well as the hydrosphere thickness. We present end-member models, review the dynamics of each layer, and discuss the global evolution consistent with the Cassini-Huygens data.

1. Introduction

Titan is the second largest satellite in the solar system after Jupiter's moon Ganymede. It is the only icy moon with a dense atmosphere made of mostly nitrogen (~98 wt%) and methane (<2 wt%) (Bézard, 2014). The methane abundance varies significantly throughout the atmosphere, from 1.5% in the stratosphere to more than 5% in the troposphere (Niemann et al., 2010; Bézard, 2014). The surface conditions, pressure of 0.15 MPa and temperature of 94 K, are such that methane and ethane (produced by photolysis of methane) are liquid and form lakes and seas mostly located at the North pole (Stofan et al., 2007). But methane is irreversibly photolyzed in the upper atmosphere and the present amount would disappear in a few tens of millions of years (e.g. Yung et al., 1984; Griffith et al., 2013), which is short on geological timescales, thus requiring an interior reservoir and a replenishment process. The presence of ^{40}Ar, a daughter element from the decay of ^{40}K, also implies exchange processes between the K-reservoir, likely silicates, and the atmosphere. Finally, the observed ^{15}N/^{14}N isotope ratio suggests that up to 50% of the atmospheric nitrogen could come from the decomposition of insoluble organic matter (IOM) present in Titan's interior (Miller et al., 2019). Understanding the formation of Titan's atmosphere requires knowledge of its interior structure and models of its evolution constrained by observations.

The NASA-ESA-ASI Cassini-Huygens mission has observed the Saturn system during the period 2004-2017 using Titan's gravity assist to explore the diversity of objects and make in situ measurements at different locations. Determining the interior structure of a planet or satellite from orbit is best achieved by measuring the gravity and magnetic fields. On January 14, 2005, the ESA Huygens probe plunged into Titan's atmosphere and provided the first

detailed information on Titan's atmosphere (Niemmann et al. 2005, Fulchignonni et al. 2005) and close-up views of Titan's surface (Tomasko et al. 2005). During 13 years, the NASA Cassini spacecraft performed 127 close Titan flybys including nine dedicated to the determination of the gravity field (Iess et al. 2010, 2012; Durante et al., 2019). One important result has been the determination of the tidal Love number k_2, whose large value implies the presence of a deep ocean (Iess et al., 2012). In addition to being the only moon with a dense atmosphere and the only object in the solar system, besides Earth, with liquids at its surface, Titan is also an ocean world (Fig. 1). The gravity measurements also provided values of the moment of inertia that is an integrated value of the density distribution, suggesting that Titan is differentiated into a hydrosphere and an inner rocky core (Castillo-Rogez & Lunine, 2010). After reviewing the different observations (Section 2), the present paper describes the interior structure models that are consistent with these observations and use the most recent Equations of State (EoS) (Section 3). Section 4 discusses the thermal evolution models and is followed by a discussion of the hydrosphere dynamics and implications for the atmosphere recycling (Section 5).

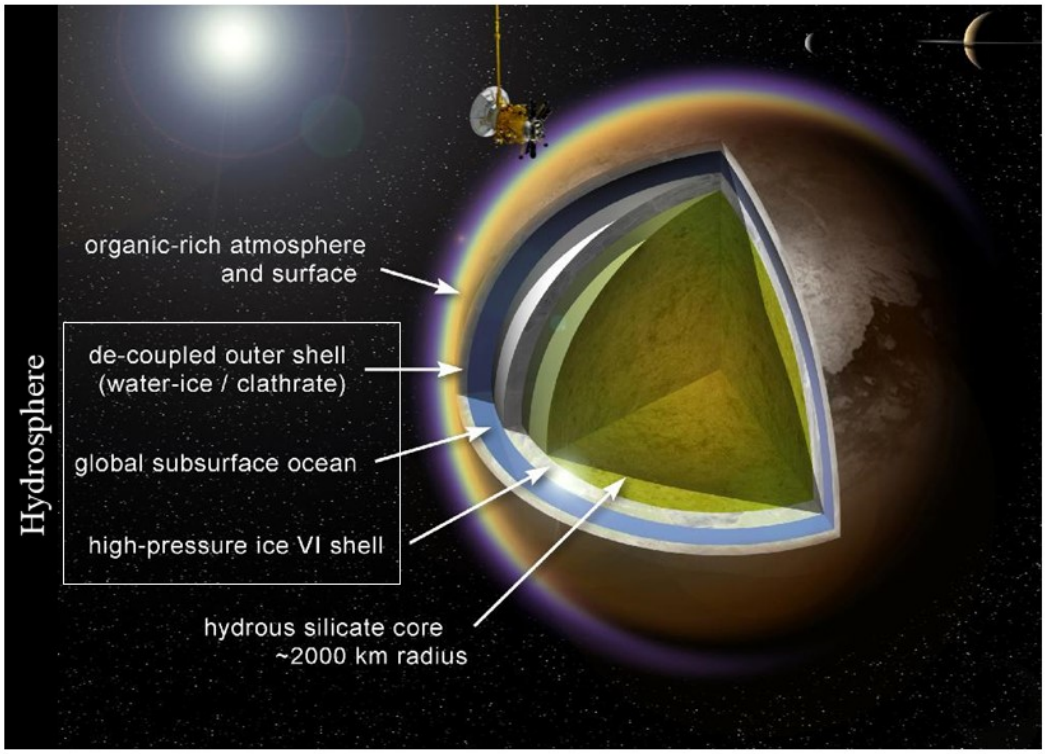

Figure 1: Titan's interior structure adapted from Fortes (2012). The rocky core is likely composed of hydrated silicates to account for the high value of the moment of inertia. It is overlaid by a hydrosphere that includes a deep salty ocean.

2. Observations

Two science questions related to Titan's interior structure that the Cassini-Huygens mission was aimed to answer are: 1) Is there a deep ocean? and 2) How much is Titan differentiated? As proven by previous missions, determining the gravity field, in particular the degree 2 coefficients, provides answers to these questions. Additional constraints come from the shape,

the electro-magnetic field, the surface morphological features and composition, and the orbital and rotational characteristics.

2.1 Gravity coefficients

The gravity field provides information on the mass distribution inside Titan. The gravity coefficients are determined by monitoring the line of sight velocity change of the Cassini spacecraft as it flies by Titan. The closer to the moon, the more precise the determination of the gravity field. However, Titan has an extended atmosphere and the drag would add a large contribution to the spacecraft velocity. An optimum closest approach flyby altitude of 1500 km was a compromise between the atmospheric perturbations and precision. After the first 4 flybys, Iess et al. (2010) were able to retrieve the gravity coefficients up to degree 3. The degree 2 gravity coefficients, J_2 and C_{22} (Table 1) include a constant component and a periodic component that is proportional to the eccentricity and therefore less than a few percent of the constant component. Because Titan is in a spin-orbit resonance, it has an ellipsoidal shape that leads to a high value of the C_{22} coefficient. If Titan is in a hydrostatic equilibrium, the degree 2 coefficients are related by a simple relation: $J_2 = \frac{10}{3} C_{22}$ The retrieved values (Table 1) give a ratio of 3.19 which is 1.7 standard deviation from the theoretical value of 10/3, suggesting a slight departure from hydrostatic equilibrium (Durante et al., 2019) (see Section 3.1 for the implication on the interior structure). With the 9 dedicated flybys, Durante et al. (2019) were able to obtain the gravity field up to degree 5.

Table 1: Parameters for Titan. GM, J_2, C_{22}, k_2 from Durante et al. (2019), shape parameters from Corlies et al. (2017), estimated power from Chen et al. (2014). The obliquity value comes from Stiles et al. (2008).

Semi-major axis to Saturn (km)	$1.2218\ 10^6$
Eccentricity	2.846%
Synchronous sidereal rotational period w (s)	1,377,684 ~ 15.945 Earth days
GM (km^3/s^2)	8978.1383(3)
Mass (10^{22} kg) / density (kg/m3)	13.4522 (3) / 1881.46 (11)
Mean radius (R) in km	2574.765 (18)
Shape triaxial ellipsoid (a,b,c) in km	2575.124 (26), 2574.746 (45), 2574.414 (28)
Degree 2 gravity coefficients: J_2 and C_{22} (x 10^6)	33.1 (6) 10.38 (8)
Reduced Moment of Inertia (MoI) C/MR^2	0.341
Tidal Love number k_2	0.616 (67)
$q=w^2R^3/GM$	$3.96\ 10^{-5}$

Gravity acceleration (m/s^2) at the surface	1.3543
Estimated radiogenic power (GW)	300 – 400
Obliquity	0.3 deg

It was critical to measure the degree 2 tidal Love number because it puts a strong constraint on the presence of an ocean. This was made possible by having dedicated gravity flybys performed when Titan was near its orbit pericenter and apocenter, maximizing gravity changes (Iess et al., 2012). The value of k_2 depends on a number of parameters including the ocean density, the ice shell thickness, and the heat transfer mechanism through the ice shell (Mitri et al., 2014). It also depends on the presence of ocean waves that can be triggered if the ocean is stratified. If the ocean is made of pure water, the value of k_2 is around 0.5 (Mitri et al., 2014). Durante et al. (2019) report a value of 0.616, about 1.7 standard deviation larger than the value of 0.5 for pure water. This large value can be explained by either a higher density due to the presence of salts, waves in a stratified ocean, or a combination of the two.

2.2 Shape and topography

Titan's shape was determined by the radar onboard the Cassini spacecraft (Elachi et al., 2004). Three data sets are available (Corlies et al., 2017). First, the Synthetic Aperture Radar (SAR) topography covers ~5.2% of Titan's surface with an average error in elevation of ~160 m over all 122 profiles. Second, the radar in its altimeter mode, mapped ~1.6% of Titan's surface along 69 profiles. Although altimetry provides a much smaller coverage compared to SAR topography, the vertical error is only ~35 m (Zebker et al., 2009). Finally, 19 digital terrain models (DTMs) were constructed through stereophotogrammetry. They cover an additional ~2.1% of Titan's surface, with a typical error of ~100 m (Kirk et al., 2013). These three data sets provide an incomplete coverage of Titan's surface. However, it was possible to retrieve Titan's shape - i.e. the difference between the surface and a sphere of radius 2575 km (Figure 2) - up to a degree 8. The shape is characterized by polar depressions up to 800 m deep. The topography is the difference between Titan's shape and an equipotential surface which is mainly dominated by the first 2 degrees of the gravity field. Titan is characterized by small topography variations at large wavelengths, from about -600 m for the south polar depression to +400 m for some mid-latitude highlands. The polar dynamical flattening can only explain half of the polar depressions observed in the shape model. If these depressions were not isostatically compensated, the J_2/C_{22} ratio would be much larger than the observed value because J_2 strongly depends on uncompensated polar topography (Gao and Stevenson, 2013) whereas C_{22} does not. Therefore, the depressions must be compensated by either a different thickness of the ice shell between the pole and the equator (Airy model proposed by Nimmo and Bills, 2010) or by differences in density (Pratt model proposed by Choukroun and Sotin, 2012). These two models are discussed in Section 5.3 and linked to different physical processes that may operate at Titan - differences in the heat flux transported through the ocean for the Airy model (Section 5.2) and formation of a dense ethane-clathrate layer at the poles for the Pratt model.

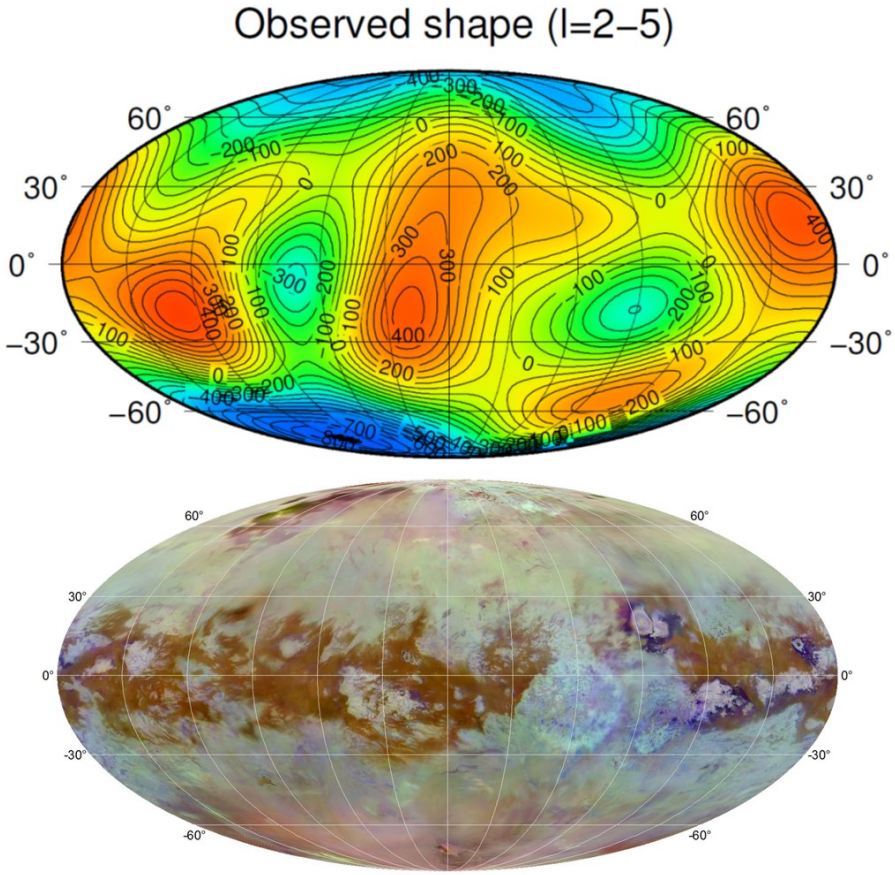

Figure 2: Mollweide projections centered at 180 deg show Titan's shape relative to a 2575 km sphere (top) and VIMS/ISS color map (bottom) showing the different geological regions (Le Mouelic et al., 2019). In the upper figure, contours are in m (Corlies et al; 2017 truncated to degree 5).

The depressions at the North pole are consistent with the presence of hydrocarbon seas that were first observed by the radar instruments (Stofan et al., 2007). A large region named Xanadu centered at (15 S, 120 W) and extending from 60 W to 150 W also correlates with a depression (Fig. 2). Other equatorial topographic variations do not seem to correlate with geological features which are dominated by dune fields (in brown in the VIMS/ISS map, Fig. 2 bottom). As pointed out by Corlies et al. (2017), the absence of dune fields in the Xanadu region is not explained as the sand would be likely transported to low topography. It suggests that Xanadu may be a very recent feature.

2.3 Rotational and orbital characteristics

Titan's rotation can provide additional constraints on the internal structure. Neglecting polar motion, the rotational state of a synchronous satellite can be considered as consisting of two components: the rotation rate variations with respect to the expected synchronous rotation rate, i.e. the non-synchronous rotation (NSR) rate, and the obliquity, corresponding to the angle between the rotation axis and the axis normal to the orbital plane (e.g. Baland et al., 2014). The value of the NSR rate has been updated several times since its first estimation of 0.36°/year in Lorenz et al. (2008) from radar surface images (see also Stiles et al., 2008). Lorenz et al. (2008) have interpreted this high NSR rate as evidence of a subsurface ocean, allowing a decoupling between the deep interior and the ice shell forced by the atmospheric

torque. Subsequent data reprocessing by Stiles et al. (2010) indicated a smaller NSR rate (0.11°/year). Using two additional years of data, Meriggiola & Iess (2012) found a very small NSR rate compatible with synchronous rotation (±0.02°/year). By modeling the gravitational and pressure coupling between the shell and the interior, Van Hoolst et al. (2009, 2013) predicted a maximal NSR rate of about 0.013°/year, smaller than the last available estimation of Meriggiola and Iess (2012), which excludes the possibility to retrieve useful constraints on the interior structure.

The obliquity of Titan is estimated to be about 0.3° (Stiles et al., 2008; Meriggiola & Iess, 2012), which is about three times larger than 0.12° - the value expected for an entirely solid Titan (Bills & Nimmo, 2008). This high value has been interpreted as being indicative of the presence of an internal ocean (Bills & Nimmo, 2008; 2011; Baland et al., 2011; 2014). By computing the obliquity from a Cassini state model for a differentiated interior structure with each layer having an ellipsoidal shape consistent with the measured surface shape and gravity field, Baland et al. (2014) showed that the observed obliquity implies a relatively dense and deep ocean. However, as a wide range of internal models can explain the observed value, no firm conclusion about the interior structure can be drawn from the obliquity alone.

Another important constraint on the interior of Titan and its past evolution is provided by the orbital characteristics. Titan exhibits an orbital eccentricity (3%) several times higher than its jovian cousins Europa, Io, and Ganymede. Contrary to the jovian moons, no orbital resonance in the Saturnian system is able to force Titan's eccentricity and thus the frictional damping due to tides raised by Saturn should lead to progressive circularization of Titan's eccentric orbit. The high eccentricity has been classically interpreted as an evidence of reduced dissipation on Titan implying limited surface liquid bodies (Sagan & Dermott, 1982; Sears, 1995) and no liquid water ocean at depth (Sohl et al., 1995). Using a coupled thermal-orbital model, Tobie et al. (2005a, 2006) showed that the current eccentricity may be compatible with the presence of an internal ocean, but would imply a larger eccentricity in the past and an ocean beneath a relatively thin and weakly dissipative ice shell during most of Titan's evolution. The proximity of small Hyperion suggests that the past eccentricity probably never exceeded 0.1-0.2. However, the 4:3 orbital resonance between Hyperion and Titan remains problematic. Recently, using very accurate radio tracking of the Cassini spacecraft during multiple flybys, Lainey et al. (2020) measured Titan's orbital expansion rate. They showed that Titan is migrating outward much faster than initially anticipated, owing to a strong dissipation inside Saturn. This finding implies that Titan formed much closer to Saturn and has migrated outward to its current position. Consequently, Titan experienced much stronger tidal forcing in the past, which may have major implications for its past internal activity.

2.4 Atmosphere and surface constraints

The atmosphere composition depends on a number of processes including endogenic processes that lead to outgassing. One major enigma of Titan's global dynamics is the need for methane replenishment against its upper atmosphere photolysis - the present amount of atmospheric methane would be photolyzed in a few tens of million years, which is short on geological timescales (Yung et al., 1984). In the absence of a global surface ocean, the methane reservoir must be in the subsurface where it can be destabilized by internal processes (e.g. Tobie et al., 2006; Choukroun et al., 2010) or by impacts (Choukroun and Sotin, 2012; Zahnle et al. 2014). A large amount of ^{40}Ar (mole fraction of 3.39 (± 0.12) × 10^{-5})

was measured in Titan's atmosphere by the gas chromatograph mass spectrometer (GCMS) onboard the Huygens probe (Niemann et al., 2010). This corresponds to a total mass of 4.5 (±0.2) 10^{14} kg. Since ^{40}Ar is a product of the radioactive decay of ^{40}K, its presence in the atmosphere indicates its release from the silicates which represent the most likely initial reservoir of ^{40}K. Depending on assumptions on the K/Si ratio in the silicates that composed Titan's rocky core (i.e. the type of chondritic material), the amount of atmospheric ^{40}Ar represents between 10% and 20% of the potential ^{40}Ar. The remaining ^{40}Ar could still be present in the rocky core or dissolved in the ocean or the hydrocarbon seas (e.g. Tobie et al. 2012). Finally, a recent study by Miller et al. (2019) suggests that up to 50% of Titan's nitrogen comes from the decomposition of insoluble organic matter (IOM) contained in Titan's rocky core when it warms up. The amount of outgassed nitrogen is inferred from dual considerations of the ^{15}N/^{14}N isotope ratio (167.7 ± 0.6) and the ^{36}Ar/N ratio in Titan's atmosphere from Niemann et al. (2010). All the above points demonstrate that exchange processes between the interior and the surface have existed during Titan's geological history although the timings of these outgassing events are not well constrained. Models of Titan's evolution should account for these results.

Another characteristic of Titan is the paucity of impact craters with only about 60 features that rank from certain (12) to nearly certain (25) to probable (25) (Werynski et al., 2019). The density of impact craters is a proxy for determining the age of a planetary surface. The scarcity of impact craters favors a young age for Titan's surface, between 0.2 and 1 Gyr (Neish & Lorenz, 2012). In addition, the presence of impact craters provides information on the thickness of Titan's lithosphere. The size of Titan's largest impact crater Menrva (diameter of 425 km, Figure 2) suggests a lithosphere at least some tens of kilometers thick at the time of formation.

Finally, the surface composition can provide information about a potential resurfacing. Titan's surface is covered with organic aerosol forming in its upper atmosphere by the photolysis of methane and further reactions with nitrogen (e.g. Hörst, 2017). All of Titan's surface should be coated with these aerosols that eventually turn into sand to form the dune fields. However, some areas appear different in false-color infrared images of Titan's surface (Fig. 2, bottom) - the dark blue color is interpreted as regions enriched in water ice (Griffith et al., 2019; Le Mouelic et al., 2019). The origin of these terrains seems to be related to exogenic processes such as impact craters (e.g. Le Mouelic et al., 2008) and erosion (Jaumann et al., 2008; Griffith et al., 2019). Evidence for endogenic processes is limited to a potential cryovolcano in the Sotra Facula regio (Lopes et al., 2013) and tentative interpretation of flow-like features in Tui and Hotei regio (Barnes et al., 2006). Confirmation that these features have a volcanic origin would require higher resolution observations as well as composition information.

2.5 Electro-magnetic measurements

Magnetic field measurements can provide key information on the interior structure of an ocean world as demonstrated by the Galileo mission. The magnetometers on the Galileo spacecraft detected Ganymede's permanent magnetic field best interpreted by the presence of an iron-rich liquid core (Kivelson et al., 1996). They also detected inductive magnetic fields at Ganymede, Europa and Callisto (Kivelson et al., 2002; Khurana et al., 1998) which are best interpreted by the presence of a deep salty ocean. However such measurements at Titan are challenged by three facts (e.g. Saur et al. 2010). First, the presence of the dense atmosphere prevents any flyby closer than 1000 km. Second, the almost perfect alignment of Saturn's

magnetic dipole axis with its spin axis prevents large variations of Saturn's magnetic field along Titan's orbit. Third, Titan's ionosphere adds a large external component that must be removed before retrieving the internal component. The Cassini magnetometer team has not reported any evidence of a permanent magnetic field or an inductive one.

During its descent through Titan's atmosphere in January 2005, the Huygens probe measured the horizontal and vertical components of the electric field with the Permittivity, Wave and Altimetry (PWA) instrument (Béghin et al., 2007). A few Hz wide signal centered at 36 Hz is present almost continuously throughout the descent. A careful analysis of the data (Simoes et al., 2007) suggested that this signal is of natural origin and would have been the second eigenmode of a Schumann-like resonance in Titan's atmospheric cavity. The Schumann resonance on Earth is the propagation of an electric signal within a cavity limited by two conductive layers which are the ionosphere on the top and the surface (mainly the ocean) at the bottom. The frequency of the resonance is linked to the size of the cavity. On Earth, the Schumann resonance is triggered by powerful atmospheric lightning activity. To date, such a source has not been detected on Titan (Fischer & Gurnett, 2011). Another process that could trigger the propagation of the waves is a plasma instability mechanism associated with the corotating Saturnian plasma flow (Béghin et al., 2007, 2012). However, a recent analysis of the signal recorded by the PWA instrument demonstrates a very strong correlation with the history of mechanical vibrations such as parachute release (Lorenz & Le Gall, 2020). It seems likely that the PWA signal is due to vibrations of the boom in the electric field present in Titan's atmosphere and cannot be used to infer the depth of an electrically conductive ocean.

During the course of the Cassini-Huygens mission, four independent observations were discussed as evidence for the presence of a deep ocean: obliquity, degree 2 tidal Love number, nonsynchronous rotation, and Schumann resonance. Only two of them survived a more thorough analysis: obliquity and degree 2 tidal Love number. The degree 2 gravity coefficients suggest that Titan is differentiated although not as much as Ganymede, which is consistent with the lack of a permanent magnetic field generated in an iron-rich liquid core. Titan thus consists of a silicate core overlaid by a hydrosphere that includes a deep ocean (Fig. 1). The next section describes interior structure models that are consistent with EoS for aqueous solutions and silicates.

3. Interior structure

3.1 Modeling the interior structure

Any model of Titan's interior must satisfy the observational constraints presented in Section 2. Its average density of 1881 $kg.m^{-3}$ suggests a large fraction of water ice and other low-density compounds. The interior structure of Titan may consist of up to 5 main layers (from center to surface): a rocky core, a high-pressure (HP) ice VI-V layer, a liquid water ocean, an ice I shell and possibly a chemically distinct icy crust enriched in hydrocarbon elements (Figure 3). The radius and density of these different layers is determined in order to satisfy the average surface radius, total mass and the moment of inertia (MoI) factor (Table 1). Even if those quantities are well constrained, there are no unique solutions of the problem - different combinations of layer densities and thicknesses can reproduce the observed global values.

In each layer, the mass, gravity, pressure and MoI are integrated from the material density which is determined by a given composition/phase using the appropriate Equation of State (EoS, see Section 3.2). The density variations with depth are determined from the pressure

and the temperature, which requires the construction of a temperature profile in each layer. The position of the top and bottom ocean interfaces is determined by the intersection of the water ice melting curve and the temperature profile in the ocean (see Figure 3a). Both the melting curve and the ocean adiabat depend on the ocean composition (e.g. Vance et al., 2018).

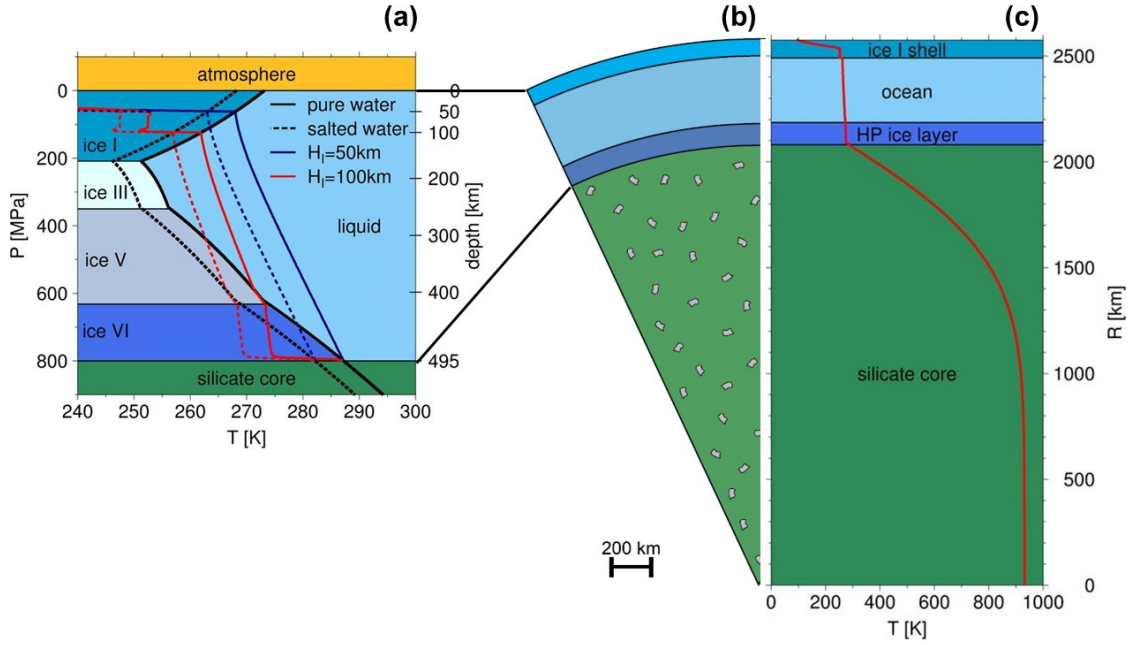

Figure 3: (a) Pressure-temperature diagram for conditions in Titan's hydrosphere. Black lines indicate the melting curve for pure H_2O (full lines, Wagner et al., 1994, 2011) and salty hydrosphere (dashed lines, 5K colder). Blue and red profiles correspond to ice I shell thickness H_I= of 50 and 100km, respectively. The depth on the right vertical axis is approximate and corresponds to pure hydrosphere. Adapted from Kalousová & Sotin (2020a). (b) Interior structure of Titan from Néri et al. (2020) where the orange symbols indicate the presence of insoluble organic matter (IOM). (c) A possible temperature profile in Titan (see text for details).

Due to the physics of the crystallization processes, the outer ice shell and the high-pressure ice layer are believed to be composed of almost pure water ice, for which the density is well known in the pressure range expected in Titan's interior (e.g. Choukroun & Grasset, 2010). The main uncertainties are the thickness of the outer ice shell, the density (composition) of the ocean and the density of the rocky core which can vary significantly depending on the assumed composition. The ocean composition also affects the melting curves. Owing to its strong depression of the crystallization temperature, ammonia has been commonly considered when modeling the evolution of Titan's hydrosphere (Lunine & Stevenson, 1987; Grasset & Sotin, 1996; Tobie et al., 2005a). These models showed that the presence of a few percent of ammonia would help the preservation of a subsurface ocean inside Titan. A few percent of ammonia, however, reduces the density of the aqueous solution compared to pure water (Croft et al., 1988). This appears incompatible with the estimate of the tidal Love number (Iess et al., 2012; Durante et al., 2019), which points to an ocean denser than pure water. According to the analysis of Mitri et al. (2014) (Figure 4b), the average ocean density should be at least 1150 kg.m^{-3} if the ice shell is 50 km thick, and at least 1190 kg.m^{-3} for a 100 km thick convective ice shell. This implies that the ocean should contain a non-water component that increases the density by about 5-10% relative to pure water (Figure 4a). Various compounds have been proposed such as ammonium sulfate (Fortes et al., 2007), sodium

chloride or magnesium sulfate (Vance et al., 2018). Based on the existing data, it is impossible to determine which one is more likely.

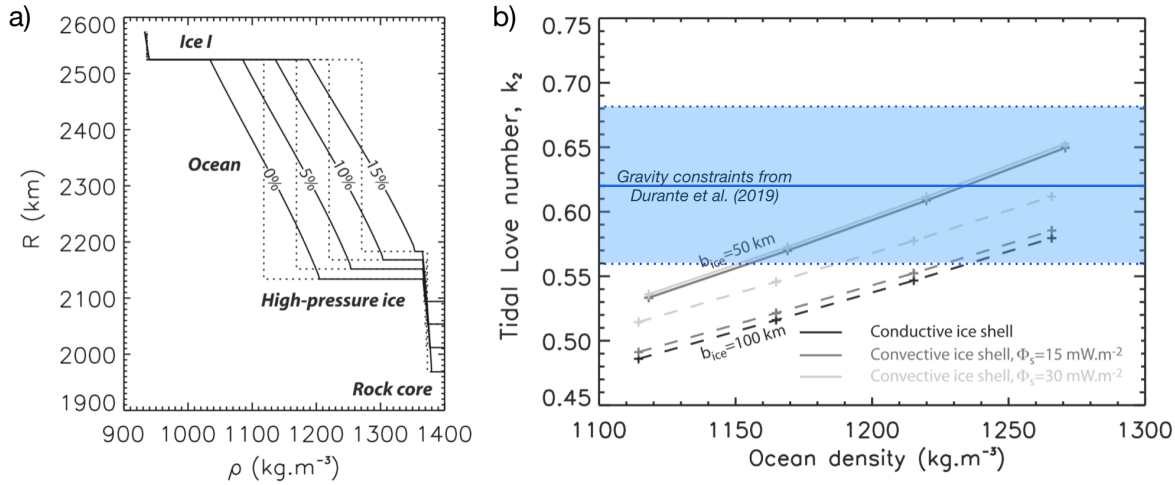

Figure 4: (a) Density profiles in the hydrosphere for different oceanic compositions. (b) Predicted tidal Love number as a function of average ocean density for two different ice shell thicknesses (50 and 100 km) in a conductive or convective state. Adapted from Mitri et al. (2014).

By constraining the hydrosphere structure from the tidal Love number estimate and the phase diagram consideration, it is then possible to determine the size and density of the rocky core from the moment of inertia as explained in Vance et al. (2014) for Ganymede. The MoI factor (C/MR^2) is inferred from the degree 2 gravity coefficients, J_2 and C_{22}, using the Radau-Darwin approximation which assumes that the gravity field is mostly determined by the dynamical distortion of a radially-distributed density structure in response to rotational and tidal forces. The fact that the J_2/C_{22} ratio is not exactly equal to 10/3 (Durante et al., 2019) and the existence of non-negligible higher degrees gravity coefficients indicate a departure from the ideal hydrostatic response. As noted in Section 2.2, the topography variations do not correlate with the higher degree gravity coefficients. A global inversion of gravity and topography is required in order to determine the magnitude of the non-hydrostaticity, which would lead to a slightly smaller value of the MoI. The lower the MoI, the larger the differentiation. It seems however very unlikely that the MoI would be as low as for Ganymede (0.311) where an iron-rich liquid core is present.

Even small variations in the value of the MoI factor can have major implications for the composition of the rocky core. A value of 0.34 implies a rocky core density of the order of 2550-2650 kg.m^{-3} for a core radius ranging between 2100 and 1970 km, whereas the density would range between 2850 and 3100 kg.m^{-3} and the radius from 1950 to 1770 km for a value of 0.33 (Mitri et al., 2014; Néri et al., 2020). The variability in core density and radius for a given value of the MoI is mostly controlled by the ocean density and the outer ice shell thickness (Mitri et al., 2014). As discussed in Section 3.2, these values of density are smaller than values inferred from the EoS of silicates and iron phases that form carbonaceous chondrites which are assumed to be one of the building blocks of silicate rich bodies in the outer solar system. The low density of the core implies either an incomplete differentiation, or a content in iron that would be much less than in carbonaceous chondrites, or the existence of a low density compound mixed with the rock phase. O'Rourke & Stevenson (2014)

performed numerical simulations of double diffusive convection to assess whether Titan could be partially differentiated, i.e. have a mixture of ice and rock in its core. Their study shows that although the differentiation could be delayed, present Titan is not partially differentiated even if ^{40}K, a major component of the radioactive heating, was leached from the core. Similarly, there is no observation or theory that would support an iron content in the building rocks of the Saturnian moons being smaller than in chondrites. The same is true for the Jovian system. If iron concentration is left as a free parameter, it leads to Fe/Si ratios much lower than the chondritic value but no formation model accounts for large variations of the Fe/Si ratio inferred in Jovian satellites (Sohl et al., 2002). A remaining possibility is a low density compound of the rocky core for which Néri et al. (2020) proposed that it may be the insoluble organic matter (IOM) whose mass fraction could range between 15 and 25 wt% for MoI factor between 0.33 and 0.34 (Figure 3b). This proposition is consistent with the finding of Miller et al. (2019) that up to 50% of Titan's atmospheric nitrogen comes from the decomposition of IOM. The presence of IOM in Titan's core would have major consequences for the differentiation and evolution of Titan's interior.

3.2 Equations of State of water and silicates

Determining the internal structure of Titan compatible with the existing observational constraints requires the use of appropriate EoS for water compounds and silicate minerals. For the pressure and temperature ranges expected in Titan's hydrosphere (100-350 K, 1.5 bar - 1 GPa), several ice polymorphs, aqueous solutions of varying compositions, gas hydrates and hydrated salts are predicted. Following the pioneer work of Lunine & Stevenson (1987), most of the internal models considered ammonia-water solutions and NH_3-H_2O phase diagram as representative materials for Titan's hydrosphere. Consequently, the experimental works have been focused on the acquisition of thermodynamic data for the NH_3-H_2O system (Croft et al., 1988; Grasset & Sotin, 1996; Hogenboom et al., 1997; Grasset & Pargamin, 2005; Choukroun & Grasset, 2010). The density constraint provided by Titan's tidal Love number ruled out ammonia as a main constituent of Titan's ocean and suggests rather heavier ionic species such as NaCl, $MgCl_2$, $MgSO_4$, Na_2SO_4, $NH_4(SO_4)_2$, etc.

Quantifying the effect of these ionic species on the thermodynamic properties of aqueous solutions and on the phase diagram of salt-water system is essential to correctly predict the structure and thermal state of the hydrosphere. The Na-Mg-Cl-SO_4-water systems have been extensively studied below 100 MPa (Journaux et al., 2020a), but, with the exception of $MgSO_4$-water system (Vance & Brown, 2013), they remain poorly known at higher pressure that is relevant for Titan's hydrosphere. Sound speed measurements at high pressure (Bollengier et al., 2019) are currently under finalization in order to constrain these different salt-water systems. Following the approach developed for pure water system (Brown, 2018; Bollengier et al., 2019), it is possible to derive a Gibbs energy representation of these systems that allows the derivation of the different thermodynamic quantities (density, thermal expansion, heat capacity, chemical potential) relevant for modeling the adiabatic profile through the ocean and the water/ice phase boundaries. First application using this representation has already been applied to the $MgSO_4$-water system (Vance et al., 2018) and should be generalized in the future to other systems using the SeaFreeze software developed for that purpose (Journaux et al., 2020b).

The density profile in the rocky core can be calculated by using an elementary composition that is consistent with cosmochemical models (Mueller & McKinnon, 1988). Alternatively, Néri

et al. (2020) took the elementary composition of carbonaceous chondrites from Wasson & Kallemeyn (1988). The density of the silicates is computed using the thermodynamic software Perple_X (Connolly, 1990) along a pressure-temperature profile consistent with the conductive heat transfer in a rocky core heated by the decay of the long-lived radioactive elements (Castillo-Rogez & Lunine, 2010, Figure 3c). The density of other components (iron sulfide and IOM) is determined from laboratory measurements.

Table 2: Characteristics of 2 possible models of Titan's interior structure. Model 1 is a nominal pure H_2O model with a thick convective ice shell whereas Model 2 has a salty ocean (denser) with a thinner conductive crust.

		Model 1	Model 2
		convection in the outer ice shell - MoI of 0.341	conduction in the outer ice shell - MoI of 0.330
Ocean depth	km	112	36
Ocean density	kg/m3	1.122	1.219
Ocean thickness	km	250	502
HP ice thickness	km	130	180
Ocean mass	10^{22} kg	1.92	4.04
Rocky core radius	Km	2083	1856
Rocky core density	kg/m3	2565	2975
Hydrosphere mass	10^{22} kg	3.74	5.48
Rocky core mass	10^{22} kg	9.71	7.97

3.3 Comparison between different interior structure models

Although the measurements from the Cassini-Huygens mission coupled with the recent laboratory data on the EoS of the different components provide new constraints on Titan's interior structure, there are still a lot of unknowns such as the thickness of the outer ice shell, the presence and distribution of clathrates therein, the thickness, density and composition of the ocean, the thickness of the high-pressure (HP) ice layer, and the composition of the rocky core. Figure 3a (full red line) shows a nominal model for a pure H_2O ocean and a ~100 km thick ice shell (Kalousová & Sotin, 2020a). Using the H_2O EoS of the different phases, one can compute the thickness of the ocean, the thickness of the HP ice layer, and the radius and density of the rocky core. The very low density of the silicate core (2565 kg m^{-3}, Table 2) compared to that of carbonaceous chondrites at the relevant (P,T) conditions (Figure 3c) requires the addition of about 25% IOM. In this model, the temperature profile remains below the dehydration temperature of the silicates (Fig. 3c), which suggests that a large fraction of potassium was leached during differentiation. Model 2 has a thinner conductive ice shell, a

denser (salty) ocean, and a smaller value of the MoI to account for some potential non-hydrostatic effects (e.g. Durante et al., 2019). This model is also consistent with a value of 0.62 for the tidal Love number (Table 2 and Fig. 4). As a result, the radius of the rocky core is smaller, its density larger but still lower than that of carbonaceous chondrites at these (P,T) conditions, and the thickness of the HP ice layer is larger. If the salt is sodium chloride that was leached from the rock fraction during differentiation, the amount of Chlorine required to get a density of 1.219 is about 50 times as large as the amount available in the silicates assuming a carbonaceous chondrite Cl/Si ration (Wasson and Kallemeyn, 1988). However, the space of possible models is quite large. Further studies are required to investigate the most likely interior model that can satisfy all the information available for Titan.

4. Internal differentiation and evolution

4.1 Accretion and formation of a primitive ocean and atmosphere

During the very early stage of Titan's evolution, the accretional (impact) heating is the predominant heat source and determines the initial structure and post-accretional thermal state of Titan's interior (Figure 5). The impact heating results from the deposition of impactor kinetic energy during the satellite accretion and it thus depends on the impactor velocity as well as on the fraction of kinetic energy that is converted into heat. The impact creates a shock wave that compresses the satellite - below the impact site, the peak pressure is almost uniform in a quasi-spherical region called the isobaric core (e.g. Croft, 1982; Senshu et al., 2002). Energy balance between the impactor kinetic energy and the impact heating leads to a local temperature increase proportional to the square of the growing icy moon radius (e.g. Monteux et al., 2007). As long as the growing moon radius is below 1000 km, the local temperature increase is less than 10-15 K after each impact. Above 2000 km, the increase of temperature is larger than 50 K. During accretion, the global increase of surface temperature depends on how deep the energy is buried, which is controlled by the impactor size, and how frequently the successive impacts occurred.

By modeling the satellite accretion in 3D from a swarm of impactors of various sizes, Monteux et al. (2014) tested the influence of impactor size distribution on the thermal evolution of a growing satellite. They showed that for satellites exceeding 1500-2000 km, surface melting can be avoided only if the satellite accreted relatively slowly (>1 Myr) from small impactors (<1 km) and if the conversion of impact energy into heat is unrealistically inefficient (<10-15%), confirming the earlier estimations of Barr et al. (2010). However, as soon as a small fraction (>10%) of the impactors exceeds 1km, global melting of the outer layer of the growing satellite cannot be avoided. As a result, a ~~the~~ surface ocean forms that is in equilibrium with a massive primitive atmosphere generated by the release of volatiles brought by the icy impactors (Kuramoto & Matsui, 1994; Marounina et al., 2018).

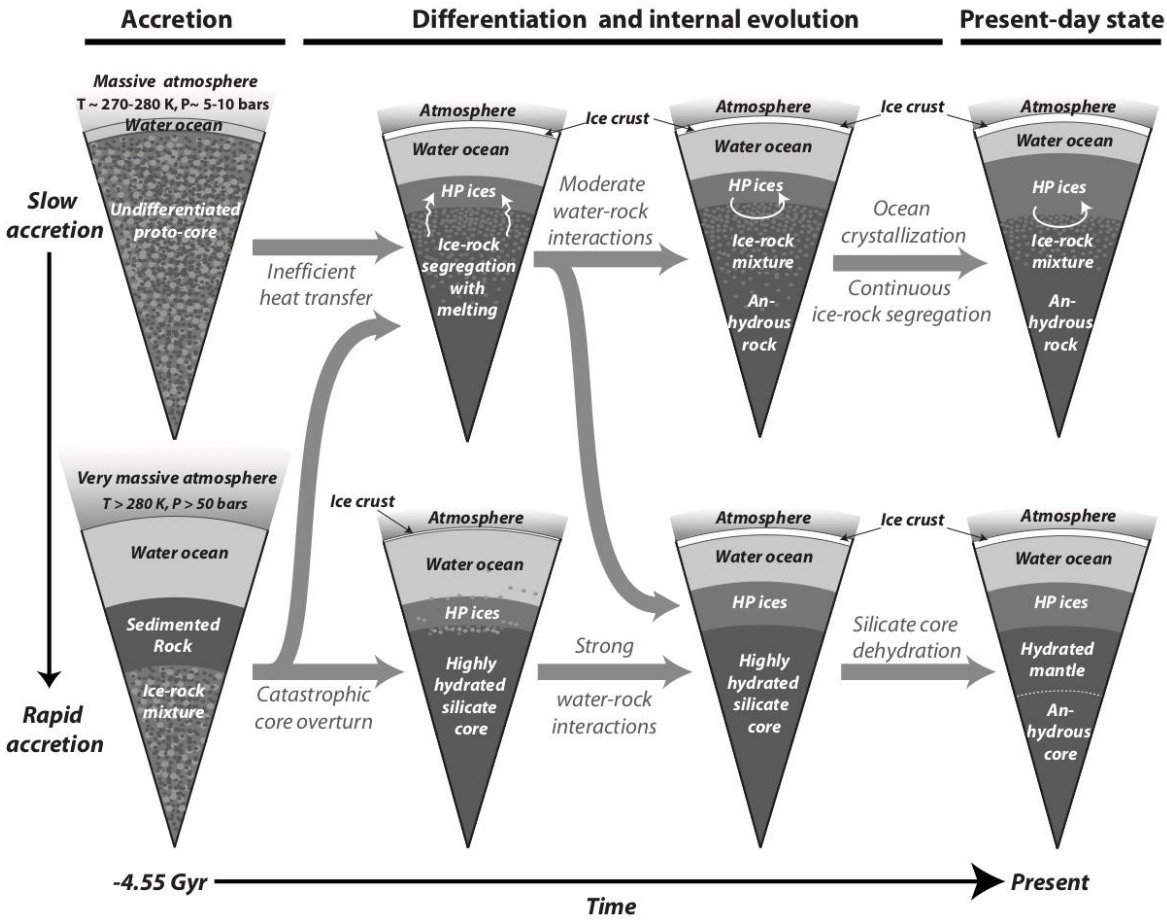

Figure 5: Possible evolution scenarios for the interior of Titan for different initial states. Depending mostly on the efficiency of heat transfer in the interior, different bifurcations in the evolutionary path may have occurred. Adapted from Tobie et al. (2014).

The volume of the surface ocean depends on the efficiency of accretion processes. For a slow accretion with small impactors, melting occurred at the end of accretion, resulting in the formation of a few tens of kilometers thick ocean and an atmosphere of <10 bars. On the other hand, for rapid accretion with large impactors, melting could occur when the satellite radius reached 1000 km (about 40% of its final radius and less than 10% of its final mass), leading to the formation of a very thick water layer in equilibrium with an atmosphere with the surface pressure exceeding 50 bar (Tobie et al., 2012; 2014). Marounina et al. (2018) showed that the atmosphere originating from the devolatilization of icy impactors was mostly composed of CO_2 and CH_4, which contrasts with the present N_2-dominated atmosphere. Melting of infalling icy impactors also released rock particles that sedimented at the base of the ocean. During this post-accretional period, the pressure at the base of the ocean was not large enough to lead to the formation of a high-pressure layer and the water ocean directly interacted with the sedimented rock layer, potentially promoting large-scale water-rock interactions and production of CH_4 and other gas compounds (Glein, 2015).

4.2 Heat budget and water-ice-rock segregation

Saturn likely formed after Jupiter (e.g. Sasaki et al., 2010) so when Titan formed, most of the short-lived radioactive isotopes (mainly ^{26}Al) were already decayed and did not contribute significantly to the early internal heat budget. The main heat source was provided by the decay

of long-lived radiogenic elements, mostly ^{40}K and ^{235}U, during the first billion of years. To estimate the heat power generated by the radiogenic decay of the rock phase in the moons, chondrites can be used. For Titan, carbonaceous chondrites, which are believed to be the dominant types of chondrites beyond Jupiter (Kruijer et al., 2017), are commonly assumed as a good proxy of the rock phase (Fortes, 2012; Tobie et al., 2012; Glein, 2015; Néri et al., 2020). Assuming a composition dominated by CI chondrites, the total radiogenic power just after accretion is estimated to be about 2.5 TW and progressively decay with time to ~300 GW at present (Hussmann et al., 2010).

Heat generated by tidal friction inside Titan due to the gravitational interaction with Saturn may have also significantly varied during Titan's history. Contrary to the Jupiter system where the Laplace resonance forces the moons' eccentricity, the absence of strong orbital resonance forcing Titan's eccentricity suggests that the eccentricity of Titan, which is already large (~3%), was even larger in the past, potentially resulting in tidal heating comparable to Europa's (Tobie et al., 2005b). The recent analysis of Lainey et al. (2020) further indicates that Titan may have formed much closer to Saturn and migrated outward due to resonance locking with Saturn. At present, the global dissipation in Titan is estimated to be <200 GW (Tobie et al., 2005b; Kalousová & Sotin, 2020b), but it may have been much larger in the past, potentially exceeding several TWs. During the very early stage, Titan also experienced strong tidal despinning before reaching the tidally-locked spin-orbit resonance. Although the associated dissipation rate is very large, it lasted for short periods of time, on the order of 100,000 years (Hussmann et al., 2010), resulting in a moderate increase of internal temperature of 25-50 K (Tobie et al., 2014).

A third source of energy resulted from the release of gravitational energy associated with the internal differentiation (Friedson & Stevenson, 1983; Hussmann et al., 2010). An approximative estimate of the temperature increase due viscous heating associated to ice-rock separation can be obtained by considering the difference of gravitational energy between an initially homogeneous interior and a differentiated interior with a full separation of rock and ice phases. This corresponds to a maximum temperature increase of the order of 100-150 K (Tobie et al., 2014). This does not include the heat loss associated to convective transfer, which is particularly important if a large fraction of the interior is molten, as discussed in section 4.1. The total heat budget, including radiogenic heating, tidal viscoelastic heating and viscous heating, determines the timing of the differentiation process. Enhanced tidal dissipation may for instance trigger earlier differentiation process, and was suggested to explain the difference in differentiation state between Callisto and Ganymede (Showman & Malhotra, 1999).

If the ice-rock separation is fast enough (<0.5 Gyr), the dissipation of potential energy may induce runaway melting and thus lead to a catastrophic differentiation, as proposed for Ganymede (Friedson & Stevenson, 1983; Kirk & Stevenson, 1987). This corresponds to the scenario shown on Figure 5 (second row). If the differentiation process is slower and more gradual (>1 Gyr), the convective heat transfer should be able to transport the additional energy and prevent internal melting, as proposed for Callisto (Nagel et al., 2004). More recently, O'Rourke & Stevenson (2014) showed that double-diffusive convection in a mixed ice-rock interior can delay internal melting and ice-rock separation, but cannot prevent it, even if reduced radiogenic power is assumed (see Figure 5, first row). This indicates that it is difficult to prevent full separation between the rock and the hydrosphere. We should keep in mind, however, that the duration of the differentiation process depends on the rheology of ice-rock

mixtures, which is poorly constrained at high pressures. Further experimental and modeling efforts are required to better understand the differentiation processes of large icy moons.

4.3 Water-rock interactions and conditions for the (non)formation of an iron- core

After the separation of rock and ice, water-rock interactions are limited to the ice-rock interface at the base of the high-pressure layer. Even though water at this interface is mostly in the form of a high-pressure ice phase owing to the elevated pressure (>0.8−1 GPa), liquid water may locally exist and potentially circulate through the upper part of the rocky core. Depending on the efficiency of heat transfer through the high-pressure ice layer which is mostly controlled by the poorly-known viscosity of HP ice (Kalousová et al., 2018, see also Section 5), the formation of transiently present liquids at the rock-ice interface may be possible (Kalousová & Sotin, 2020a). These liquids could interact with the underlying rocks before percolating upward to the ocean. Warm liquid water may be also released from the core when the dehydration temperature is reached (Castillo-Rogez & Lunine, 2010). This dehydration event may have occurred relatively late during Titan's evolution (>3−3.5 Gyr after formation), provided that sufficient amounts of radiogenic potassium were leached from the core and that thermal diffusivity in the core is not too low (Castillo-Rogez & Lunine, 2010). In the warm evolution scenarios, intensive water-rock interactions were likely during two main sequences of Titan's evolution - before 1 Gyr during the differentiation and after 3-3.5 Gyr during the core dehydration.

Dehydration processes may not occur if the hydrated silicate core can convect, thus removing the radioactive heat efficiently. Hydrated silicates have a deformation rate that strongly depends on stress and only weakly on temperature (Hilairet et al., 2007). To date, no study has investigated the possibility of convection in the hydrated silicates. If convection does not start in the hydrated silicates, then dehydration may start in the center of the moon (Castillo-Rogez and Lunine, 2010). The dehydration temperature is 350 K lower than the eutectic temperature of the Fe-FeS system (1200 K at Titan's interior pressure, Buono et al., 2011). If the amount of internal heating is large enough, then the temperature may exceed the Fe-FeS system eutectic temperature and an iron core may form, a process that may have happened on Ganymede (Bonnet-Gibet et al., 2020). The lack of intrinsic magnetic field at Titan may indicate that such high temperatures may not have been reached inside Titan.

5. Dynamics and evolution of the hydrosphere

Based on the interior structure models, Titan's hydrosphere is about 500 km thick (see Section 3). It harbors a subsurface water ocean which is sandwiched between the outer ice shell and a deep ice layer possibly composed of several high-pressure (HP) ice phases (ice VI, V, and III, see Fig. 3). We first review recent models describing the dynamics and heat transfer in the particular hydrosphere layers before addressing the coupled evolution of the hydrosphere as a whole.

5.1 Dynamics of the HP ice layer

Titan's HP ice layer is supplied with radiogenic heat flow from the underlying silicate core (Section 4) and its top boundary is determined by the intersection of the ocean adiabat and the melting curve (Section 3). Its dynamics is mainly controlled by the HP ices viscosity which has been measured by several groups (Sotin et al. 1985; Sotin and Poirier, 1987; Durham et al. 1996; 1997; see also Journaux et al., 2020a, for a compilation). The temperature difference between the bottom and top boundary is less than ~30 K (Fig. 3) and thus the ice temperature is always close to the melting curve. This has two implications. First, the viscosity is low enough for thermal convection to occur (Choblet et al., 2017a; Kalousová & Sotin, 2020a). Second, the melting temperature is reached for a broad range of models (Choblet et al., 2017a) and thus a two-phase mixture model is necessary to self consistently treat the melting process and the subsequent melt evolution. Figure 6 shows a snapshot from a two-phase convection simulation tailored to Titan's HP ice layer from Kalousová & Sotin (2020a). The left panel depicts the temperature difference from the melting temperature (note that $\Delta T \leq 20K$) with the dark red denoting the partially molten (temperate) ice where temperature equals the melting temperature ($T=T_m$) and where some water may be present. The right panel shows the corresponding water content (porosity Φ) - note that only a few percent are present and no melt accumulation is observed. Due to the decrease of the melting temperature with the decreasing pressure (i.e. increasing vertical coordinate, see Fig. 3), most of the water is produced in the top layer of temperate ice at the ocean interface from where it is extracted into the ocean. Due to this temperate layer, no cold thermal boundary layer (TBL) is present and the convection is thus characterized by the hot upwelling plumes generated in the hot TBL and a passive downwelling. Some water may be also present in the upwelling plumes and at the silicates interface. Scaling performed by Kalousová & Sotin (2020a) showed that the occurrence or not of bottom melting mainly depends on the incoming heat flux, the ice viscosity and the HP ice layer thickness. For the currently expected interior structure (Section 3), the amount of heat produced in the core (Section 4) and reasonable viscosity values, Kalousová & Sotin (2020a) predict that melt may be present at the interface of Titan's core and HP ice layer. This has profound astrobiology implications since volatiles and organics present in Titan's core may be leached and transported by the liquids into the ocean (Section 4). This makes Titan quite different from Ganymede whose HP ice layer is probably much thicker and melt is not currently predicted at its bottom (Kalousová et al., 2018; Kalousová & Sotin, 2018). Let us note that all the described models were performed assuming a pure H_2O setting. If other compounds such as ammonium sulfate, magnesium sulfate, or sodium chloride are present in the ocean (Section 3), the processes described above may be altered since the melting temperatures as well as the liquid densities would be changed substantially.

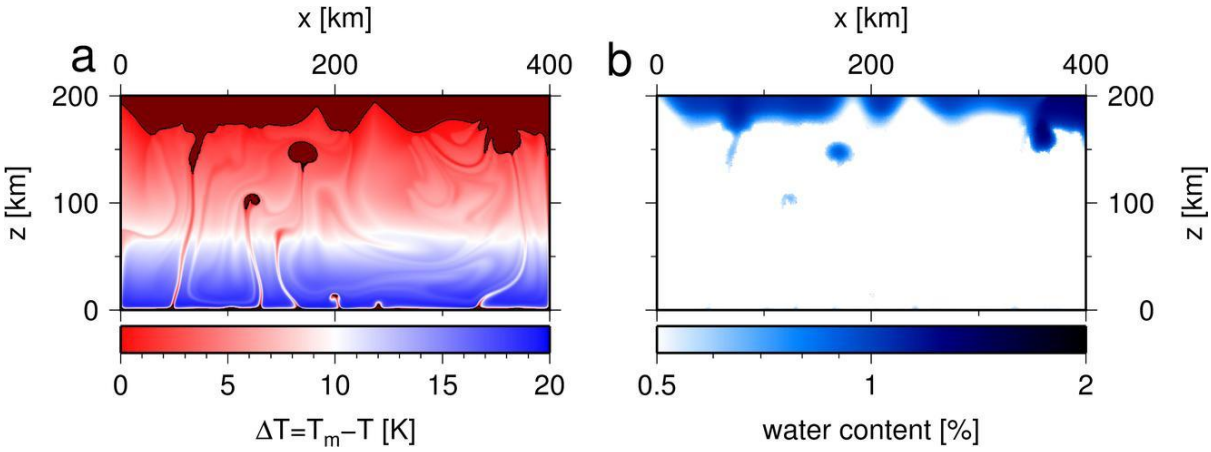

Figure 6: Heat and water transport through Titan's HP ice layer. (a) Temperature difference from the melting temperature ($\Delta T = T_m - T$). Dark red marks the temperate ice ($T = T_m$). (b) Water content (porosity). Adapted from Kalousová & Sotin (2020a).

5.2 Dynamics of the ocean

Flows in Titan's subsurface ocean can be driven by thermo-compositional convection as well as mechanically by tides (for a review, see Soderlund et al., 2020). For the rotating subsurface oceans, the convective heat transfer efficiency and the outgoing heat flux pattern depend strongly on the particular importance of driving buoyancy, Coriolis, and inertia forces when compared to retarding viscous forces (e.g. Gastine et al., 2016; Soderlund et al., 2019; Amit et al., 2020; Figure 7). Amit et al. (2020) described two distinct regimes - (i) equatorial cooling characterized by a dominant top boundary heat flux in the equatorial region which was found for cases when rotation was important, and (ii) polar cooling characterized by dominant top boundary heat flux at polar regions found for cases when rotation was less important (see Figure 7). The expected range of Ekman and Rayleigh numbers relevant to Titan's ocean (blue box in Figure 7) suggests a rather limited effect of rotation on its dynamics, consistent with the polar cooling pattern. These results are in agreement with the findings of Kvorka et al. (2018) who predicted dominant heat fluxes at the poles to explain Titan's long-wavelength topography (see also next section for more details). Using the same spectral code as Amit et al. (2020), Soderlund (2019) found rather different results with the largest polar heat flux (i.e. polar cooling) for cases with a strong influence of rotation and equatorial heat flux peaks (i.e. equatorial cooling) for a more moderate rotational influence. This obvious contradiction between their findings may be the result of different choice of mechanical boundary conditions – while Soderlund (2019) prescribed impenetrable, stress-free boundaries, Amit et al. (2020) used rigid boundaries. Similarly, the effect of thermal boundary conditions should be carefully investigated since in both studies, isothermal boundaries were prescribed. However, given that the ocean is sandwiched between two ice layers (HP ice layer below and ice I shell above) with much slower dynamics, the incoming and outgoing heat flux are likely to be governed by the dynamics of these two solid layers, i.e. a prescribed incoming and outgoing heat flux would be a more appropriate boundary condition then prescribed temperature. A dedicated numerical study investigating these different settings would certainly bring more insight into the ocean dynamics of Titan as well as other ocean worlds.

Titan's eccentric orbit and nonzero obliquity result in time-variable tidal forcing that drives ocean flow. For Europa, the associated tidal dissipation was suggested as an important source

of energy by Tyler (2008) although the assumed value of tidal quality factor remained debatable. To provide an independent constraint on the possible amount of ocean tidal dissipation, Chen et al. (2014) used the shallow-water equations that neglect the radial ocean currents with respect to horizontal currents (making the problem a two-dimensional one) with an Earth-like value of bottom drag coefficient. They also neglected the effect of the overlying ice shell. Their results show that the dissipated power in Titan's ocean due to obliquity tides is on the order of 10 GW, which was later confirmed by Hay & Matsuyama (2017). The corresponding surface heat flux of about 0.1 mW/m^2 is however negligible compared to the tidal heating in solid layers as well as to the radiogenic heating in the core (Chen et al., 2014, see also Section 4). The inclusion of the outer ice shell covering the ocean has a minor effect on both, the eccentricity and obliquity tidal heating in Titan's ocean. Recently, Rovira-Navarro et al. (2019) and Rekier et al. (2019) performed simulations tailored to Europa and Enceladus that took into account the three-dimensional nature of ocean flows. Although the observed pattern of periodic inertial waves differed from the shallow-water results, the tidal dissipation due to these inertial waves was found negligible when compared to the traditional heat sources. Finally, let us note that future studies should also investigate the interaction between convection and the tidally-driven flows (Soderlund et al., 2020).

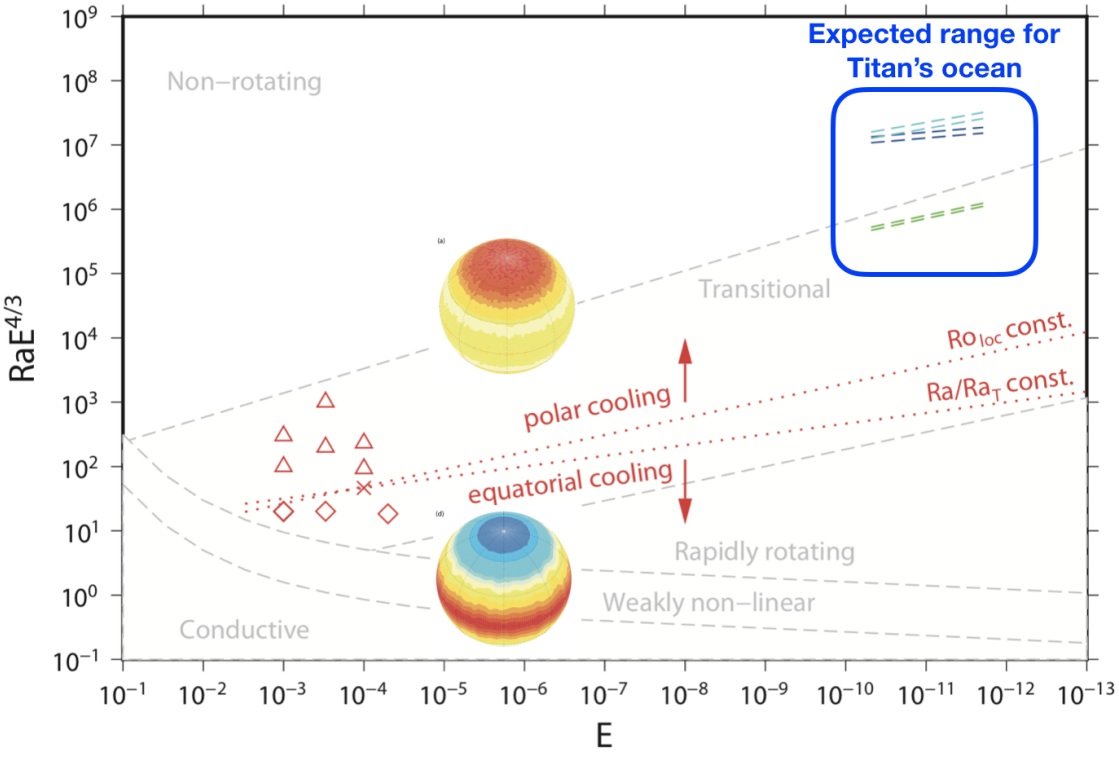

Figure 7: Regime diagram proposed by Gastine et al. (2016) showing the different regimes for convecting rotating flows in a spherical shell depending on Ekman number (E, ratio of viscous and Coriolis forces) and product of Rayleigh number and Ekman number to the 4/3 ~~Ekman and Rayleigh numbers~~ (Ra, ratio of thermal buoyancy to momentum and heat diffusion). The blue box highlights the expected values for Titan's ocean (dashed colored lines, see Amit et al., 2020, for more details). Results by Amit et al. (2020) are denoted by red diamonds (equatorial cooling) and red triangles (polar cooling). The inserted spheres depict the long-term time-averaged heat flux anomaly (i.e. heat flux with respect to the mean value)

across the ocean top boundary. Red/blue denote positive/negative heat flux anomaly, respectively. Modified from Amit et al. (2020).

5.3 Dynamics of the outer ice shell

The thermal state of Titan's outer ice shell is not well constrained. Depending on its thickness and the ice viscosity, heat can be transferred either by conduction (thin shells, large viscosity) or convection (thick shells, small viscosity). The viscosity is mainly determined by the ice grain size (Durham et al., 2001; Goldsby & Kohlstedt, 2001) which may vary from less than 1 mm to more than a few mm and would evolve dynamically (Barr & McKinnon, 2007). Therefore, the critical thickness for the convection to occur depends mainly on the ice grain size, often represented by the viscosity value at the melting temperature, η_0. Lefevre et al. (2014) found that Titan's 100 km thick ice shell is likely to convect unless $\eta_0 \geq 3 \times 10^{16}$ Pa s, roughly corresponding to a 7 mm grain size. If the shell is 50 km thick, the threshold viscosity would be about one order of magnitude smaller, corresponding to a 3 mm grain size. The heat transfer by convection is far more efficient than the conductive cooling and leads to an efficient crystallization of the deep ocean unless another mechanism is acting to delay the freezing.

The presence of ammonia has been suggested (e.g., Grasset & Sotin, 1996) that would decrease the ocean temperature, thus increasing viscosity and making convection less efficient or not occurring at all. However, a significant ammonia concentration would be necessary to substantially decrease the ocean temperature which is not compatible with the densities inferred by Mitri et al. (2014) as well as with the recent work by Leitner & Lunine (2020) who suggest that only up to about 5% of ammonia could be present in Titan's ocean. Another possibility is tidal heating in the ice shell or the core that keeps oceans liquid in bodies such as Europa or Enceladus (Tobie et al., 2005b; Choblet et al., 2017b). However, given the high value of Titan's eccentricity and no resonance to maintain it on geological timescales, it is not likely that significant dissipation is currently occurring in Titan's interior (Tobie et al., 2005b) although more energy could have been dissipated earlier in the evolution (Section 4). Alternatively, a crust made of methane clathrates - that have about ten times smaller thermal conductivity than ice under Titan's surface conditions (Sloan et al., 2007) - has been proposed to insulate Titan's interior and delay the ocean crystallization (e.g. Loveday et al., 2001; Tobie et al., 2006). Recently, Kalousová & Sotin (2020b) revisited this concept and performed simulations of thermal convection in Titan's clathrate capped ice shell. Figure 8 shows the comparison of temperature for simulations without clathrates (panel a) and with a 10km thick clathrate crust (panel c). The middle panel (b) shows the average temperature profiles for these two cases. The inclusion of the clathrate crust results in: (i) a significant reduction of the stagnant lid thickness (h_{sl}, dashed lines) - in this particular setting from 42km (h_c=0km) to 15km (h_c=10km), and (ii) a substantial decrease in the amount of heat extracted from Titan's ocean (q) - for the depicted case by about 40%. As a consequence of the thinner stagnant lid, a warm material is brought closer to Titan's surface which may facilitate material exchange with the surface e.g. due to cryovolcanism (Mitri et al., 2008; Lopes et al., 2013).

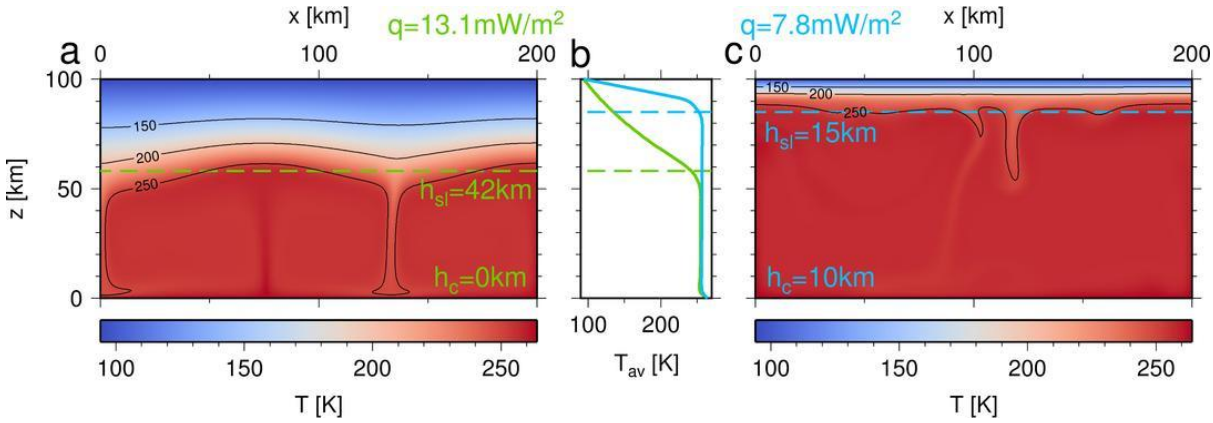

Figure 8: Comparison of temperature fields assuming a clathrate crust 0km (a) and 10km (c) thick. Panel (b) shows the average temperature profiles (full lines). The dashed lines indicate the stagnant lid (h_{sl}). Adapted from Kalousová & Sotin (2020b).

As discussed in Section 2, Titan's long wavelength topography is characterized by polar depressions of about 300m and the negligible correlation of topography with the gravity field anomalies points to a high degree of compensation (Durante et al., 2019). To explain the observed topography, a concept of isostasy has been considered. Nimmo & Bills (2010) assumed that Titan's shell is rigid and conductive and proposed that lateral ice shell thickness variations (corresponding to Airy isostasy) can lead to topographies that fit well in the observations (Figure 9a). The topic was further investigated by Lefevre et al. (2014) and Kvorka et al. (2018) who found that lateral flows driven by the shell thickness variations would tend to remove any ice shell/ocean interface topography unless the ice viscosity at the base of the shell exceeds 10^{16} Pa s, corresponding to either coarse grain (≥ several mm) or very cold ocean (≤240 K). As discussed above, such a cold ocean is not compatible with the high value of k_2 (Section 3). Kvorka et al. (2018) further showed that Titan's topography cannot be reproduced by the pattern of tidal heating in the ice shell as initially proposed by Nimmo & Bills (2010). Their results however indicated that Titan's polar flattening could be explained by lateral variations of ocean heat flux with amplitudes between 0.1 and 1 mW/m^2, warm water upwellings at the poles and cold downwellings close to the equator. Such a pattern would agree with the polar cooling of Amit et al. (2020) (see previous section). Alternatively, Choukroun & Sotin (2012) proposed that the polar depressions could be explained by subsidence due to the presence of ethane clathrate that is about 8% denser than water ice (corresponding to Pratt isostasy, Figure 9b). This model is supported by the observation of ethane rains at the poles (Griffith et al., 2006). The ethane polar caps would be the ethane reservoir that could explain the absence of atmospheric ethane although it is the main product of methane photolysis in the upper atmosphere (e.g. Lavvas et al., 2008). The ethane clathrate would form by substitution of methane by ethane in the methane clathrate crust, a process observed in laboratory experiments (Murshed et al., 2010) or by the reaction of liquid ethane with water ice as observed experimentally by Vu et al. (2020).

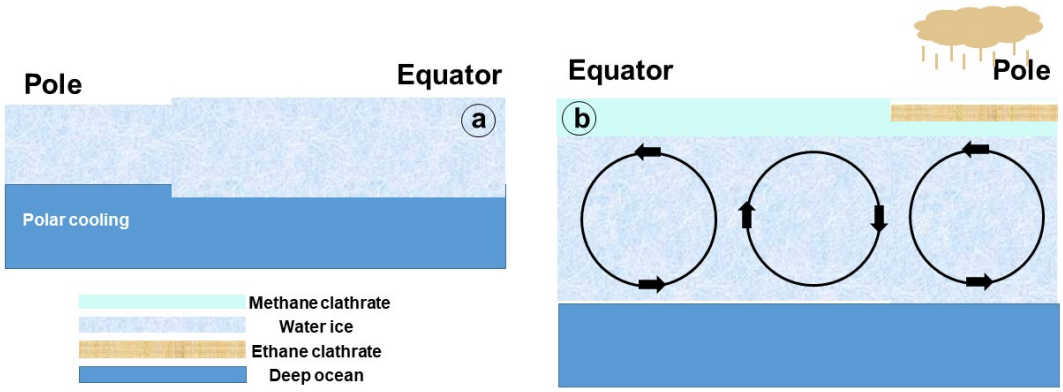

Figure 9: Models of ice I shell that can explain the observed gravity and topography data: (a) Airy model (Nimmo & Bills, 2010), (b) Pratt model (Choukroun & Sotin, 2012).

5.4 Coupled evolution of the hydrosphere and implications for the atmosphere

As stated in Mitchell and Lora (2016), models need to consider the interactions between Titan's atmosphere, surface, and subsurface in order to make further progress in understanding Titan's complex climate system. The greatest mystery is how can Titan retain an abundance of atmospheric methane with only limited surface liquids, while methane is being irreversibly destroyed by photochemistry. A related mystery is how can Titan hide all ethane that is produced in this process (Lavvas et al., 2008). All along Titan's history, the evolution of the hydrosphere has likely influenced the composition of the atmosphere, by controlling the outgassing rate (e.g. Tobie et al., 2006) and possible internal recycling of hydrocarbon compounds (e.g. Choukroun and Sotin, 2012). The hydrosphere may contain large quantities of volatiles, either dissolved in the ocean or stored in the form of gas clathrate hydrates in the icy shell (Tobie et al., 2012; Choukroun et al., 2010, 2013).

The release of these volatile compounds is, however, limited due to their solubility in the internal ocean and the clathration processes within the H_2O layers. Outgassing from the interior may occur through two different mechanisms: (1) destabilization of clathrates stored in the outer ice shell by thermal convective plumes or by impacts, (2) mobilization of dissolved gas molecules contained in the water ocean and subsequent bubble formation if the ocean gets saturated (Tobie et al., 2006; 2009). Both processes are controlled by methane, since it is expected to be the dominant gas in the form of clathrate hydrate in the crust and since it is the only gas compound that may reach the saturation point in the ocean during periods of reduced ice shell thicknesses. The amount of ^{40}Ar detected in the atmosphere provides constraints on the level of methane outgassing through time. Owing to its low solubility in water, argon should be preferentially incorporated in CH_4-rich bubbles when they form at the ocean-ice interface and should then be transported upwards by the percolating CH_4-rich fluids. Following this extraction mechanism, Tobie et al. (2012) estimated that the atmospheric mass of methane has been renewed at least 20-30 times during Titan's evolution in order to explain the atmospheric abundance of ^{40}Ar as detected by Huygens (Niemann et al., 2010).

Following the evolution scenario proposed by Tobie et al. (2006), the outgassing of methane may result in episodic destabilization of the methane clathrate reservoir, accumulated in the crust during the differentiation process. This crust would then thin in response to the release of heat associated with the core evolution (Tobie et al., 2006), resulting in the dissolution of

methane in the ocean and the outgassing of the gas in excess (Tobie et al., 2009). According to this scenario, several kilometers of methane clathrates may still be present in the outer icy shell, limiting the cooling rate of the ocean (Kalousová & Sotin, 2020b). Recent thermal destabilizations of this reservoir may explain the present-day atmospheric methane (Tobie et al., 2006). This evolution model was based on a simplified 1D approach, assuming an ammonia-water ocean and ignoring the coupling between the ocean and the high-pressure ice layer. More sophisticated evolution models taking into account the latest advances in physical and thermodynamic description of the hydrosphere should be developed to better understand the impact of the hydrosphere evolution on the atmosphere. The recent advances made in the description of the large-scale dynamics of the high-pressure ice layer (Section 5.1) and of the ocean (Section 5.2) motivate the need for developing a new generation of models describing the exchange processes between the different layers in a 3D framework and using a consistent thermodynamic approach for phase changes and chemical exchanges between each layer of the hydrosphere.

6. Discussion and conclusions

The Cassini-Huygens mission has provided key information on Titan's interior structure such as the presence of a deep ocean, the near hydrostatic equilibrium of its long wavelength topography, and the low density of its rocky core. Titan is almost the same size and mass as Ganymede, Jupiter's largest moon but the apparent absence of an inner iron-rich core seems to indicate that Titan did not experience as high temperatures in its rocky core as Ganymede. The upcoming JUICE mission managed by the European Space Agency (launch expected in 2022) will orbit Ganymede and will provide a unique set of gravity, topography, rotation, electro-magnetic field, and radar sounder measurements. These data will provide a very good picture of Ganymede's interior structure. The comparison between the two moons may help understand the interior structure and thermal evolution of each of them.

The habitability requirements for a planet or a moon include water, essential elements (CHNOPS), chemical energy, and time (Cockell et al., 2016). The Cassini-Huygens mission demonstrated that Titan has a deep ocean. The thermal evolution included several episodes when water and rocks interacted and most likely left elements in the ocean. Organic material forms in the atmosphere and after falling on the surface, it may be transferred to the deep ocean. As discussed in Section 2.4 (Atmospheric constraints) and in Section 5.1 (HP ice layer dynamics), exchanges have occurred between the rocky core and the atmosphere, implying that transfer of volatiles and other elements coming from the decomposition of IOM in the core have existed between the rocky core and the ocean. Whether the transfer from the surface to the ocean and the transfer from the rocky core to the ocean can provide the chemical disequilibrium and chemical energy that could lead to life is an open issue.

Models describing the geological evolution of Titan must take into account all the information that has been acquired with the Cassini-Huygens data as well as the results of laboratory experiments. Tobie et al. (2006) proposed a model that was consistent with the amount of methane in the atmosphere, the presence of a deep ocean, the large eccentricity, and the mass. New information from Cassini, the recent progress on EoS and other key parameters of the compounds that built Titan, as well as recent advances in the modelling of multi-phase layers point to the development of a new thermal evolution model. Such a model would be useful to define questions that could be answered by the upcoming Dragonfly mission (Turtle et al., 2017). Questions such as the presence or not of a clathrate crust, the thickness of the

ice shell, the tectonic activity of Titan may be answered thanks to the geophysical package onboard that mission (Lorenz et al., 2019). It will shed more light on Titan, the satellite that resembles the Earth more than any other.

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