Note

Ceres – Neither a porous nor salty ball

Julie C. Castillo-Rogez

Jet Propulsion Laboratory, California Institute of Technology, Pasadena, CA 91109, USA

ARTICLE INFO

ABSTRACT

This study explores the geophysical implications of two compositional models recently proposed for Ceres, which assume that the dwarf planet is a homogeneous mixture of chondritic material devoid with free water. In order to reproduce Ceres' density, the rock density has to be offset by the presence of porosity and/or an abundance of hydrated salts resulting from the extensive hydration and oxidation of the chondritic material. Thermal modeling shows that a mixture of hydrated minerals is bound to compact and partly dehydrate as a consequence of long-lived radioisotope decay heat. The resulting interior structure is differentiated in a silicate-rich core and water-rich shell, with little porosity. Hence, this study confirms previous suggestion that Ceres contains a large fraction of free water.

1. Introduction

The nature of water in C-class asteroids is a long-standing question that bears important implications for assessing the astrobiological significance of these objects. The present study addresses the question of the state of water in the dwarf planet Ceres, as the current state-of-the-art opposes two very different views of its interior: the "icy" model assumes that Ceres can be compared to a "warm icy satellite" with an icy shell and rocky core (McCord and Sotin, 2005; Castillo-Rogez and McCord, 2010), while the "rocky model" assumes that water is bounded to minerals in the form of hydrated salts and silicates (Zolotov, 2009 – hereafter referred to as Z09). These two models entail very different framework for formulating science questions that will drive the science planning of the *Dawn* mission, which will reach Ceres in 2015. Comparison between these models, for example in terms of geological setting, is limited by the absence of geophysical modeling for the rocky model. Indeed, the interior structure proposed by Z09 is a static chemical model meant to explain Ceres' current state, but its geodynamical evolution remains to be investigated. This is the focus of the present study.

The recent detections of water ice and organics at the surface of the largest member of the Themis family, 24 Themis (e.g., Rivkin and Emery, 2010) and of 65 Cybele (Licandro et al., 2011) are important clues for this problem. 24 Themis appears to be the remnant core of a larger water-rich asteroid (Castillo-Rogez and Schmidt, 2010), indicating that at least in some of the large, wet asteroids, water must be buried below a thick regolith layer, as suggested by, e.g., Fanale and Salvail (1989). The simplest explanation for the absence of ice at the surface of Ceres is that water is not stable in that region of the Solar System – ~2.7 AU from the Sun versus ~3.1 AU for Themis (e.g., Schörghofer, 2008). Alternatively, the gradient in hydrated mineral abundance across the asteroid belt has been interpreted as increased thermal processing closer to the Sun due to variations in the timing of formation, which determines the amount of short-lived radioisotopes accreted into the objects (Ghosh et al., 2006). It has also been suggested that icy planetesimals were extensively processed prior to accretion, leading to aqueous alteration of the rocky material and water loss (Wilson et al., 1999). The "rocky" model may have been formed from such planetesimals, after ^{26}Al was mostly extinct (Z09). This requires a late formation timeframe with respect to that generally envisioned for Ceres-size asteroids (e.g., Weidenschilling, 2008), but this may be acceptable in absence of definitive constraints on the early Solar System history.

In order to assess the viability of the "rocky" interior scenario, I investigate its evolution over Ceres' lifetime. Especially, I test the assertion made in Z09 that "*After the accretion (after 7–8 Ma after the formation CAIs), limited heat sources (*McCord and Sotin, 2005*) could have not caused mineral dehydration and major density stratification.*" This claim is unsupported by both McCord and Sotin (2005) and Castillo-Rogez and McCord (2010). Both studies agree that Ceres' core could reach a temperature of 400 K in a few hundred Myr after accretion, at which time many hydrated materials and organics start melting and separate from the silicate phase (Kargel, 1991). I tailor a geophysical model using the same conditions as assumed by Z09 in terms of time of formation time and chemistry (Section 2). The results of this investigation are discussed in Section 3.

2. Geophysical evolution

Zolotov (2009) introduced two variants of the same idea, as a function of the degree of oxidation of planetesimal material. One model assumes that CI chondrites are good analogs for Ceres' material and requires at least 10% bulk porosity to account for the observed density (Table 1). CI material is a mixture of serpentines, tochilinite, montmorrillonite, magnetite, gypsum, organics, epsomite, bloedite, and a number of minor species. I choose a content in hydrated salts based after Kargel (1991) who studied in detail the fate of these species in large chondritic bodies. Our model also includes 7 vol.% magnetite (after Nagy et al., 1963), 8 vol.% organics (Pearson et al., 2006), the rest being dominated by antigorite (the serpentine mineral most stable at Ceres' pressures), and montmorillonite (Table 1).

The other model assumes extensive oxidation of carbonaceous chondrite material in planetesimals leading to a significant volume fraction of hydrated salts (Table 2). The latter model, detailed in Table 3 of Z09, is an assemblage of hydrated silicates (37 vol.%), hydrated salts (36 vol.%), oxides (14 vol.%), and organics (6 vol.%). Its mean density is ~2.24 g/cm³. Z09 quotes 1.9 g/cm³, but this does not appear consistent with the content of his Table 3. We refer to these models as the high- and low-grain density endmembers, respectively. Both models assume accretion after short-lived radioisotopes are mostly decayed and bear little geophysical significance. I also assume no heating converted from accretional energy. Thus, the main heat source driving geophysical evolution in this framework comes from long-lived radioisotope decay.

2.1. Initial interior structures

The chemical models introduced in Z09 are representative of the initial states of the geophysical models I aim to develop. The low-density endmember does not

E-mail address: Julie.C.Castillo@jpl.nasa.gov

Table 1

Mineralogical composition of the high-density model considered in this study and physical properties relevant to this study. Minerals representing less than 1 vol.% are not included in the discussion, because they have little affect on the mechanical or thermal behavior of the mixture (Ni-sulfide, dolomite, quartz, Ca-phosphate, hydrohalite, and minor phases). The average density for that mixture is ~2.46 g/cm³, consistently with average carbonaceous chondrite material (Britt et al., 2002). Pyrene is used as a proxy to represent the organic material, following Z09's approach. Thermal data are from Prieto-Ballesteros and Kargel (2005) (epsomite and mirabilite), and Akbulut et al. (2006) (pyrene). The thermal conductivity of magnetite is expressed as a function of temperature T (Molgaard and Smeltzer, 1971). Specific heat data for serpentine, montmorillonite, magnetite, and halite are from Waples and Waples (2004). Full dehydration of epsomite proceeds in several steps, starting at 298 K and completing at 548 K (Van Essen et al., 2009). The most significant step occurs at a temperature of about 353 K. For the sake of simplicity, we have assumed in the modeling a one-step dehydration event at the latter temperature.

Mineral	Volume fraction	Density (g/cm³)	Dehydration (D) or melting (M) temperature (K)	Thermal conductivity (W/m/K)	Specific heat capacity (J/kg/K)
Serpentine	0.45	2.52 (average)	775 (D)	$(0.404 + 0.000246T)^{-1}$	~2000
Montmorillonite	0.15	2.32	1073 (D)	0.75	900
Magnetite	0.07	5.175	N/A	$4.23 - 1.37 \times 10^{-3}T$	586
Epsomite	0.1	1.68	353 (D)	~0.4	1546
Mirabilite	0.07	1.57	305 (D)	~0.6	1825
Pyrene	0.06	1.27	~420 (M)	~0.27	1300
Halite	0.05	~2.07	1074 (M)	5.78–1.5 f(T)	926

Table 2

Same as Table 1 but for the low-density endmember model (see text for details), based after Zolotov (2009), Table 3. The average density for that mixture is 2.24 g/cm³. Gypsum starts dehydrating at 373 K with the production of a hemihydrate, and then the production of anhydrite at 423 K (Brantut et al., 2011). Mirabilite is considered, instead of bloedite, due to the lack of thermophysical data on the latter. Thermal data are from Clauser and Huenges (1995) (gypsum), Molgaard and Smeltzer (1971) (hematite).

Mineral	Volume fraction	Density (g/cm³)	Dehydration (D) or melting (M) temperature (K)	Thermal conductivity (W/m/K)	Specific heat capacity (J/kg/K)
Saponite	0.37	2.32	1073 (D)	0.75	900
Epsomite	0.29	1.68	353 (D)	~0.4	1546
Hematite	0.14	5.3	>1800 (M)	$8.39 - 6.63 \times 10^{-3}T$	620
Mirabilite	0.07	1.57	305 (D)	~0.6	1825
Pyrene	0.06	1.27	~420 (M)	~0.27	1300
Gypsum	0.03	2.36	373–423 (D)	~0.15	1090–1800

require any porosity to explain the observed density, between 2.07 and 2.24 g/cm³, depending on available measurements (Thomas et al., 2005; Carry et al., 2008). On the other hand, the high-density endmember requires at least 10% of silicate in bulk to offset the density of CI-like material, in order to match the observed density. Zolotov (2009) suggested that the preservation of porosity down to Ceres' core, i.e., a pressure of ~150 MPa, is supported by the experimental literature, taking sandstones as an analog. Although I adopt the porosity profile presented in that study, I do not necessarily endorse it, because the compressive strengths of hydrated materials are generally lower than those characteristic of sandstones. The compressive strength of chondritic material appears to be of the order of 10 MPa (Grimm and McSween, 1989), instead of the value of 60 MPa used by Z09.[1] This means that extensive compaction of the interior (porosity ⩽10%) could be reached 20 km below the surface, instead of the ~120 km considered in Z09.

2.2. Thermal properties

Most of the minerals listed in Tables 1 and 2 have a thermal conductivity less than 1 W/m/K, and possibly as low as 0.1 W/m/K. The thermal conductivity of a solid mixture is computed as a linear combination of the minerals conductivities and volume fractions. In the case of the low-density endmember, the thermal conductivity is of the order of 1.4 W/m/K at 200 K and decreases to 1 W/m/K at 700 K, mainly due to the strong dependence of hematite on temperature.

In the case of the high-density endmember, the presence of porosity makes it more difficult to estimate thermal conductivity. Its impact is a function of the void geometry, which determines the number of contacts between grains. The assemblage of various compounds, some of which may act as cement between grains or fill in veins within the material, makes the problem even more difficult to constrain. The thermal conductivity of the solid endmember composition is about 2.4 W/m/K at 200 K. However, if the porosity is significant, it becomes the main factor determining thermal conductivity (e.g., Shoshany et al., 2002). As a reference, a CM chondrite sample with a porosity of ~37% has a bulk thermal conductivity of 0.5 W/m/K, with little dependence on temperature (Opeil et al., 2010). Also, the thermal conductivity of martian regolith analog with a porosity of ~50% is of the order of 0.1 W/m/K at a temperature of 200 K (Zent et al., 2010) similar to Ceres' subsolar temperature (Fanale and Salvail, 1989). I use the latter value in the modeling, keeping in mind that that parameter may be even further reduced in Ceres' outermost material (Jeff Kargel, personal communication).

---

[1] Compressive strength is the maximum stress a material can undergo until failure. It determines compaction as a consequence of brittle deformation (as a difference to creep-driven porosity collapse).

The mean specific heat capacities of the mixtures can be computed as a linear combination of the capacities of each mineral and their mass fractions. Specifics for each models are indicated in Fig. 1.

2.3. Thermal evolution

As a first-order approximation, I compute the thermal evolution of a spherical body subject to long-lived radioactive decay heat (e.g., McCord and Sotin, 2005; Castillo-Rogez et al., 2007): Heat transfer is computed after:

$$\frac{\partial\left(k(T) \cdot \frac{\partial T(r)}{dr}\right)}{\partial r} + \frac{2}{r}\left(k(T)\frac{\partial T(r)}{\partial r}\right) = \rho(r)C_p(T)\left(\frac{dT(r)}{dt}\right) - H(r) \qquad (1)$$

where T is temperature, r is the local radius, k is the thermal conductivity, ρ is the material density, C_p is the specific heat capacity, t is time, and H represents the internal heating. Internal evolution is marked by two major processes: the dehydration and/or melting of most minerals involved in the two endmember models and the compaction of the silicate compounds. By thermogravimetry analysis, Nagy et al. (1963) tracked the loss of 22% of material mass from CI chondrite gradually heated from 373 to 1273 K, with major steps between 373 and 473 K, 773–873 K and 1050–1200 K. These may be attributed to the dehydration of salt hydrates, serpentine (antigorite), and montmorillonite, respectively (Nagy et al., 1963; Tables 1 and 2). Accounting for the melting of these compounds is a major improvement upon the modeling approach previously used by Castillo-Rogez and McCord (2010). Indeed, part of radioisotope decay heating is used as latent heat for these multiple reactions. Hence these events delay the time at which silicate dehydration occurs. The latent heat for gypsum is 98 kJ/kg (Brantut et al., 2011), 932 kJ/kg for epsomite (Van Essen et al., 2009), and 131 kJ/mol for clay. Montmorillonite and serpentine dehydrate at higher temperatures.

The creep of antigorite, the serpentine endmember relevant to Ceres' relatively high pressures, is activated with an energy as low as a few tens kJ/mol (Hilairet et al., 2007). These authors found a relaxation time shorter than ~10⁴ years at a temperature of 473 K, for a differential stress greater than 1 MPa, which is relevant to our study. Measurements of serpentine creep properties are scarce though, and the weakness of antigorite is currently a matter of debate. Still, even if the relaxation times obtained by Hilairet et al. (2007) results were off by several orders of magnitude, silicate relaxation would still be warranted since Ceres' interior was likely to bear temperatures greater than 473 K for most of its history (Fig. 1).

The temperature achieved in the interior model as a function of time and radius is displayed in Fig. 1. For both models, part of Ceres volume starts melting about 200 Myr after formation. Melting and separation of the hydrated salts, pyrene, and water from gypsum occur over a timescale of about 200 Myr determined in part by the large latent heat of epsomite. This event affects about 23 vol.% of the