# The interior structure of Ceres as revealed by surface topography 

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#### Abstract

Ceres, the largest body in the asteroid belt ( 940 km diameter), provides a unique opportunity to study the interior structure of a volatile-rich dwarf planet. Variations in a planetary body's subsurface rheology and density affect the rate of topographic relaxation. Preferential attenuation of long wavelength topography ( $\geq 150 \mathrm{~km}$ ) on Ceres suggests that the viscosity of its crust decreases with increasing depth. We present finite element (FE) geodynamical simulations of Ceres to identify the internal structures and compositions that best reproduce its topography as observed by the NASA Dawn mission. We infer that Ceres has a mechanically strong crust with maximum effective viscosity $\sim 10^{25}$ Pas. Combined with density constraints, this rheology suggests a crustal composition of carbonates or phyllosilicates, water ice, and at least 30 volume percent (vol.\%) low-density, high-strength phases most consistent with salt and/or clathrate hydrates. The inference of these crustal materials supports the past existence of a global ocean, consistent with the observed surface composition. Meanwhile, we infer that the uppermost $\geq 60 \mathrm{~km}$ of the silicate-rich mantle is mechanically weak with viscosity $<10^{21}$ Pas, suggesting the presence of liquid pore fluids in this region and a low temperature history that avoided igneous differentiation due to late accretion or efficient heat loss through hydrothermal processes.


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## 1. Introduction

Despite the unique status of Ceres as a volatile-rich dwarf planet in the asteroid belt, broad uncertainties remain regarding its interior structure. Recent Dawn spacecraft measurements of Ceres' surface topography and gravity field have confirmed groundbased measurements indicating some degree of central condensation (Thomas et al., 2005; Drummond et al., 2014; Park et al., 2016; Ermakov et al., in preparation), as predicted by thermal evolution models (McCord and Sotin, 2005; Castillo-Rogez and McCord, 2010). Even so, the derived density structures permit a range of compositions for the crust and mantle, defined herein as the two principal layers with contrasting densities (see Section 2.1).

[^0]Observations of topographic relief over a range of wavelengths can constrain interior structure and composition. On geologic timescales, topography may decay in amplitude due to viscous relaxation. Importantly, the decay of longer wavelength topography is sensitive to viscosity at greater depths. Taking advantage of this relationship, observed viscous relaxation timescales for topographic relief on Earth have been used to demonstrate the viscosity difference between the upper and lower mantle (Anderson and O'Connell, 1967). Analogously, topographic data have been used to estimate the sub-surface rheological profile of the Moon and Mars, revealing decreasing viscosities at greater depths (Solomon et al., 1982; Zuber et al., 2000).

Stereophotogrammetry (SPG) using the Dawn Framing Camera has been used to construct a shape model of Ceres' surface with 135 m spatial resolution and 10 m vertical resolution $(1 \sigma)$. In contrast to other solar system bodies such as Venus, Mars (Turcotte, 1987), and Vesta (Ermakov et al., 2014), the resulting topographic


Fig. 1. Topographic spectrum evidence for viscous relaxation of topography on Ceres. Observed topographic power spectral density (PSD) for Ceres and Vesta are shown in black and gray, respectively. Blue line represents a power law for Ceres' unrelaxed topographic PSD fitted across medium wavelengths (36.5-134.2 km) extrapolated to spherical harmonic degree 4. The power-law Ceres PSD does not extend to degrees 2 and 3 due to lack of analogous wavelengths on Vesta that support a power-law model. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
power spectral density (PSD) of Ceres does not conform to a single power-law, with wavelengths ( $\lambda$ ) greater than $\sim 150 \mathrm{~km}$ showing lower amplitudes (Fig. 1). This dichotomy between Vesta and Ceres cannot be explained by differing fluxes of large ( $\geq 15 \mathrm{~km}$ diameter) impactors (Marchi et al., 2016). At the same time, the topographic PSDs of some icy satellites such as Enceladus and Iapetus follow a single power-law, suggesting that the crater formation process on volatile-rich bodies cannot by itself explain the loss of long wavelength topography on Ceres (Nimmo et al., 2011). Instead, systematically lower topographic relief at equatorial latitudes on Ceres for $\lambda \geq 120 \mathrm{~km}$ (Ermakov et al., in preparation) strongly suggests that viscous relaxation, a highly temperature-dependent process, is responsible for the reduction of topography.

The observed topographic PSD of modern Ceres therefore represents a balance between the removal of topography due to viscous relaxation and its creation due processes such as impacts. In this work, we quantify the rate of topographic relaxation over a range of wavelengths using a viscoelastoplastic finite element (FE) model written using the deal.II finite element library. We then assume that impact cratering is the dominant mechanism of topography creation on present-day Ceres and estimate the rate of ongoing resurfacing using a Vesta-calibrated, main-belt impactor flux. By combining these rates to compute the evolution of Ceres' topographic PSD since 4.3 Ga , we find that the observed PSD is best reproduced by high surface viscosities that decay rapidly with depth. Meanwhile, a strong, lithified deep interior, if one exists, must be $\geq 100 \mathrm{~km}$ below the surface. We combine these rheological constraints with density estimates from gravity field analysis to infer the interior composition of Ceres and its thermal evolution.

## 2. Model description

### 2.1. Nomenclature for interior layers

Previous analysis has shown that a minimum of two uniform density layers are required to explain Ceres' spherical harmonic degree 2 gravity field and topography (Park et al., 2016; Ermakov et al., in preparation). In this work, we adopt a two-layer density structure and refer to the two regions as the crust and mantle. Due to the $\sim 1000 \mathrm{~kg} \mathrm{~m}^{-3}$ difference between these layers (see Section 2.2), we regard the crust and mantle as compositionally distinct and describe them as volatile-rich and silicate-rich, respectively. In addition, we test for the existence of a rheologically
strong deep region in the mantle and refer to these rheologically weak and strong partitions as the upper and lower mantles, respectively. We note that, although the high strength of the lower mantle may result from a distinct composition rich in anhydrous silicate or metal, the modeling presented here is sensitive to a rheological discontinuity only and does not require such a compositional contrast. We therefore refrain from using the term core, which may be used in the future to describe any compositionally distinct layer in the deep interior.

### 2.2. Topographic relaxation models

To quantify the decay of topography, we perform viscoelastoplastic FE simulations of topographic relaxation on Ceres assuming a range of internal rheological structures. Due to uncertainties of the hydrostatic component of flattening at spherical harmonic degree 2 and the lack of analogously long wavelengths on Vesta, we limit ourselves to spherical harmonic degrees $\geq 4$. Our code, which is available online (Fu and Ermakov, 2016), is written using the deal.II finite element library and based on a previous viscoplastic version used to model topography on Vesta (Bangerth et al., 2007; Fu et al., 2014). We carry out our FE computations on 2D Lagrangian meshes in axisymmetric geometry with between $4.5 \times 10^{3}$ and $1.1 \times 10^{4}$ cells (Fig. 2A). To optimize computation time, we limit our domain to the $x>0, y>0$ quadrant where $x$ is the radius projected to the spin axis and $y$ is height along the spin axis, taking advantage of symmetry across the equator to model all zonal even spherical harmonic degrees. Due to symmetry, we apply zero normal flux boundary conditions on the $x=0$ and $y=0$ edges of the FE domain, which represent the spin axis and the equatorial plane of Ceres, respectively. We specify zero pressure on boundary of the FE mesh that corresponds to the outer surface of Ceres.

All FE model runs except for benchmarking setups assume a two-layer density structure consisting of a 41.0 km crust with $1287 \mathrm{~kg} \mathrm{~m}^{-3}$ density overlying a mantle of 428.8 km radius with $2434 \mathrm{~kg} \mathrm{~m}^{-3}$ density. These densities and thicknesses are based on the analysis of the observed admittance (topography to gravity ratio) between spherical harmonic degrees 3 and 16 (Ermakov et al., in preparation). We calculate the best-fit ellipsoids for the outer surface and the crust-mantle interface at each time step and use an analytical formulation to calculate the internal gravity field (Pohánka, 2011). We then assume a rotation period of 9.0 h to determine the centrifugal acceleration.

Following Keller et al. (2013), we implement viscoelasticity using a modified incompressible Stokes flow formulation. Once the viscosity and shear modulus are defined everywhere in the domain (see below), we precondition the stiffness matrix using lower upper (LU) decomposition and compute pressure at all FE nodes using an iterative conjugate gradient algorithm (Bangerth and Kronbichler, 2008). Displacement at each FE node is then calculated from the pressure solution. To implement plasticity, we evaluate the full stress tensor for each FE cell and find the principal stress components using the Armadillo Linear Algebra Library (Sanderson, 2010). We assume that the continuous bombardment of Ceres' surface results in negligible tensile strength and use Byerlee's rule with rock-like frictional coefficient 0.85 to determine the distribution of brittle failure given the inferred low abundance of water ice (Byerlee, 1978; Bland et al., 2016). After each run of the FE model, we reduce the effective viscosity of FE cells where failure occurred (Fig. 2B) and iterate until we converge on an effective viscosity field that leads to no further significant brittle failure (e.g., King and Hager, 1990; Kaus, 2010). Typically, one to four FE iterations are required to converge on an effective viscosity field. This final flow field solution is then used to advect the mesh according to the specified time step, the duration of which is chosen such


Fig. 2. (A) Initial mesh, (B) initial viscosities, (C) shear stresses ( $\sigma_{1}-\sigma_{3}$ ) and plastic failure locations (red points), and (D) plasticity-adjusted effective viscosities of a typical FE time step. Example shown has viscosity gradient of one logarithmic unit per 10 km and no stiff lower mantle. Logarithmic contours in panels B and D are at $10 \times$ intervals and those in C are at $10^{1 / 2} \times$ intervals. Grey shaded region denotes the higher density mantle. The $x=0$ line is the rotation axis while the $y=0$ line is the equator.
that the largest surface displacement in the first time step is equal to a displacement target parameter set to 175 m .

The viscosity profile is the most critical input for controlling the rate of topographic decay. Broad uncertainties in the composition of Ceres and in the rheology of constituent phases preclude the adoption of an a priori viscosity profile based on an assumed composition. Instead, we test a wide range of plausible viscosity profiles and determine the best-fit viscosity model by comparison to the observed topographic PSD.

We vary three parameters to obtain the initial viscosity profile of our model body: the surface viscosity, the viscosity gradient, and the depth of a rheologically strong deep interior, if present. For the lower bound on the surface viscosity, we adopt a value of $8 \times 10^{19}$ Pas corresponding to deformation of 1 mm pure water ice grains via grain boundary sliding at the equatorial surface temperature of 155 K under a typical shear stress of $\sim 1 \mathrm{MPa}$ (Goldsby and Kohlstedt, 2001; Hayne and Aharonson, 2015). To simulate stronger crustal compositions corresponding to heavy contamination by hydrate and silicate phases, we test surface viscosity values up to $10^{28}$ Pas. Higher viscosities do not permit the decay of topography on solar system timescales even in the case of a very low viscosity interior.

We assume that viscosity decays exponentially with depth, corresponding to a composition with an Arrhenius law rheology in a conductive temperature gradient (See Supplementary Data for a discussion of adopted rheologies and uncertainties). The viscosity gradient is varied over a wide range between one logarithmic unit per 6 km and one logarithmic unit per 37.5 km (Table 1). For the activation energies and stress exponents of water ice and hydrates (Table 2), this range of viscosity gradients corresponds to thermal gradients of between $\sim 0.2$ and $12 \mathrm{Kkm}^{-1}$, which bracket the values of $0.5-1.0 \mathrm{~K} \mathrm{~km}^{-1}$ expected from thermal evolution modeling
(Castillo-Rogez and McCord, 2010). We clarify that, because the goal of the FE simulations is to identify a family of plausible raw viscosity profiles, no physical flow law is directly incorporated in the FE computations. However, once the plausible viscosity profiles are derived, we compare them to full non-Newtonian flow laws of different substances (Table 2) to identify the most plausible material composition.

The rheology of the deep interior within the silicate-rich mantle depends strongly on the assumed degree of lithification, which includes the processes of dewatering and cementation. Ordinary chondrite and other anhydrous, lithified rocks have compressive strengths on the order of $\geq 100 \mathrm{MPa}$ (Kimberley and Ramesh, 2011; Petrovic, 2001) and are not expected to deform at the shear stresses in the deep interior of Ceres ( $\leq 1 \mathrm{MPa}$ ) at temperatures below $\sim 600^{\circ} \mathrm{C}$ (see Section 4.2). On the other hand, given the high water content of Ceres, large regions of its deep interior may remain unlithified (e.g., Travis and Feldman (2016)). In these cases, deformation may be possible on geologic timescales via frictional sliding facilitated by pore fluids, which, assuming our adopted density structure and Earth ocean-like density of $1025 \mathrm{~kg} \mathrm{~m}^{-3}$ for interstitial water, lower the effective confining pressure by a factor of 2.2 to 3.8 in the uppermost 100 km of Ceres. Furthermore, fluids may affect grain boundaries, resulting in up to an order of magnitude decrease in the coefficient of friction (Bos and Spiers, 2002; Moore and Lockner, 2007). Laboratory studies and field studies of unconsolidated sediments such as glacial tills and sedimentary basins have found effective viscosities in the range of $10^{15}$ to $8 \times$ $10^{22}$ Pas for shear stress of order 1 MPa at temperature between 0 and $150^{\circ} \mathrm{C}$ (Boulton and Hindmarsh, 1987; Schneider et al., 1996). These values are much lower than our best-fit near-surface viscosities (see Section 3), implying that the deep interior of Ceres is effectively strengthless if it remains unlithified. At the same time,

Table 1
Summary of FE simulations.

| Set number | Viscosity gradient (km per $10 \times$ decay) | Lower mantle depth (km) | $\begin{aligned} & \text { Misfit } \\ & \left(\mathrm{km}^{4}\right) \end{aligned}$ | Misfit uncertainty, $1 \sigma$ ( $\mathrm{km}^{4}$ ) | Best-fit surface viscosity (Pas) | $\tau$ at degree 4 (My) | $\tau$ at degree 10 (My) | $\tau$ at degree 16 (My) |
| :---: | :---: | :---: | :---: | :---: | :---: | :---: | :---: | :---: |
| 1 | 6.0 | None | 3.25 | 0.61 | 4.9E+27 | 213 | 271 | 846 |
| 2 | 7.5 | None | 2.28 | 0.31 | $2.3 \mathrm{E}+27$ | 409 | 632 | 1692 |
| 3 | 10.0 | None | 1.72 | 0.22 | $8.6 \mathrm{E}+26$ | 963 | 1804 | 5242 |
| 4 | 12.5 | None | 1.46 | 0.16 | $4.4 \mathrm{E}+26$ | 1688 | 3348 | 18,304 |
| 5 | 15.0 | None | 1.51 | 0.26 | $3.2 \mathrm{E}+26$ | 2017 | 6444 | 25,140 |
| 6 | 20.0 | None | 2.26 | 0.25 | $1.4 \mathrm{E}+26$ | 3885 | 19,756 | 53914 |
| 7 | 25.0 | None | 2.46 | 0.24 | $7.7 \mathrm{E}+25$ | 6284 | 46,441 | 157,743 |
| 8 | 30.0 | None | 2.85 | 0.28 | $4.5 \mathrm{E}+25$ | 9679 | 62,564 | 184,425 |
| 9 | 37.5 | None | 3.53 | 0.28 | $3.3 \mathrm{E}+25$ | 14,907 | 89,765 | 411,102 |
| 10 | 10.0 | 46.0 | 21.20 | 0.11 | $5.7 \mathrm{E}+25$ | 114,021 | 35,165 | 28,234 |
| 11 | 10.0 | 58.0 | 12.37 | 0.20 | $1.7 \mathrm{E}+26$ | 17,549 | 6024 | 16,153 |
| 12 | 10.0 | 75.0 | 3.33 | 0.22 | $4.4 \mathrm{E}+26$ | 2618 | 2722 | 10,539 |
| 13 | 10.0 | 100.0 | 1.62 | 0.21 | $9.5 \mathrm{E}+26$ | 891 | 1690 | 4875 |
| 14 | 10.0 | 125.0 | 1.25 | 0.20 | $1.1 \mathrm{E}+27$ | 705 | 1632 | 5870 |
| 15 | 10.0 | 150.0 | 1.58 | 0.18 | $9.6 \mathrm{E}+26$ | 825 | 1473 | 5728 |
| 16 | 15.0 | 46.0 | 17.84 | 0.02 | $9.0 \mathrm{E}+24$ | 338,932 | 226,311 | 152,568 |
| 17 | 15.0 | 58.0 | 16.57 | 0.06 | $2.7 \mathrm{E}+25$ | 137,959 | 48,802 | 68,304 |
| 18 | 15.0 | 75.0 | 6.06 | 0.20 | 9.0E+25 | 18,689 | 12,680 | 37,668 |
| 19 | 15.0 | 100.0 | 1.27 | 0.20 | $2.2 \mathrm{E}+26$ | 3126 | 10,072 | 35,789 |
| 20 | 15.0 | 125.0 | 1.34 | 0.20 | $3.1 \mathrm{E}+26$ | 1959 | 7726 | 31,425 |
| 21 | 15.0 | 150.0 | 1.32 | 0.23 | $2.9 \mathrm{E}+26$ | 2173 | 7570 | 19,413 |

Notes: The misfit between simulated and observed topography is calculated according to Eq. (7). Columns 7-9 give the e-folding relaxation timescale ( $\tau$ ) at the indicated spherical harmonic degrees.

Table 2
Laboratory rheological parameters used in this work.

| Material | Formula | $\begin{aligned} & \log _{10}(A) \\ & \left(\mathrm{MPa}^{-\mathrm{n}} \mathrm{~d}^{\mathrm{m}} \mathrm{~s}^{-1}\right) \end{aligned}$ | m | $n$ | $\begin{aligned} & Q \\ & \left(\mathrm{~kJ} \mathrm{~mol}^{-1}\right) \end{aligned}$ | Reference | Minimum experimental temperature (K) |
| :---: | :---: | :---: | :---: | :---: | :---: | :---: | :---: |
| Water Ice | $\mathrm{H}_{2} \mathrm{O}$ | -2.4 | 1.4 | 1.8 | 49 | Goldsby and Kohlstedt (2001) | 208 |
| Epsomite | $\mathrm{MgSO}_{2} \cdot 7 \mathrm{H}_{2} \mathrm{O}$ | 36.2 | 0 | 3.5 | 270 | Durham et al. (2005) | 284 |
| Meridianiite | $\mathrm{MgSO}_{2} \cdot 11 \mathrm{H}_{2} \mathrm{O}$ | 9.6 | 0 | 5.0 | 90.3 | McCarthy et al. (2011) | 230 |
| Mirabilite | $\mathrm{Na}_{2} \mathrm{SO}_{2} \cdot 10 \mathrm{H}_{2} \mathrm{O}$ | 12.1 | 0 | 5.4 | 128 | Durham et al. (2005) | 230 |
| Magnesite | $\mathrm{MgCO}_{3}$ | -47.1 | 0 | 1.0 | 233 | Holyoke et al. (2013) | 673 |
| Methane Clathrate Hydrate | $\mathrm{CH}_{4} \cdot 5.75 \mathrm{H}_{2} \mathrm{O}$ | 8.6 | 0 | 2.2 | 90 | Durham et al. (2003) | 260 |
| Wet Olivine | $(\mathrm{Mg}, \mathrm{Fe}) \mathrm{SiO}_{4}+\mathrm{H}_{2} \mathrm{O}$ | 6.7 | 0 | 3.5 | 515 | Hirth and Kohlstedt (1996) | 1523 |
| Calcite | $\mathrm{CaCO}_{3}$ | 1.05 | 1.9 | 1.7 | 190 | Walker et al. (1990) | 673 |

Notes: Column three gives the logarithm of the pre-exponential constant used in an Arrhenius formulation (e.g., Eq. (1) in Goldsby and Kohlstedt (2001)), columns four and five give the grain size and stress exponents, and column six gives the activation energy. To illustrate the degree of extrapolation used, the final column provides the lowest laboratory temperature used to derive the creep law. Deformation mechanism for water ice is grain-boundary sliding, due to the relatively low stresses and temperatures. The creep law for sodium chloride (van Keken et al., 1993; Wawersik and Zeuch, 1986) shown in Fig. 6A is not based on an Arrhenius formulation and is therefore not tabulated here. The minimum experimental temperature for sodium chloride is 293 K .
even in the case of a lithified silicate-rich mantle, deformation over geologic timescales may still proceed via pressure solution under low temperature and pressure conditions, although the low abundance of low temperature soluble species such as quartz and carbonates may limit the extent of pressure solution (Rutter, 1976; Shimizu, 1998; Tada and Siever, 1989).

Several evolutionary scenarios of Ceres may result in the partitioning of the silicate-rich mantle into rheologically strong and weak sub-regions. First, stronger metamorphic heating may have occurred in the deeper interior, resulting in dehydration and lithification (Castillo-Rogez and McCord, 2010). Alternatively, grain size sorting during an early convective phase of the mantle may have concentrated large, lithified particles such as chondrules in the deep interior (Travis and Feldman, 2016). Finally, if deformation in an unlithified mantle is accommodated predominantly through frictional sliding, the greater confining pressures at depth may suppress deformation.

In summary, plausible compositions and evolutionary histories of Ceres may result in strong, weak, or heterogeneous rheology for the silicate-rich mantle of Ceres. We therefore use our FE model to test a range of mantle viscosities and to constrain the existence and location of an internal rheological transition. If a lower mantle exists, we impose a viscosity of $3 \times 10^{26} \mathrm{~Pa}$, effectively simulating a non-deforming material at the timescales of our models. For the
rheologically weak upper mantle, we continue the viscosity gradient from the surface across the crust-mantle interface. Due to the steep near-surface viscosity gradients in the best-fit simulations, the viscosity at depths corresponding to the outer mantle are $10^{3}-10^{5}$ times lower than surface values. This treatment therefore effectively regards the upper mantle as strengthless on geologic timescales.

Our assumed radius of the lower mantle, if it exists, varies between 429 (i.e., the radius of the entire mantle) and 320 km (Table 1). In addition, we simulate the case of a fully unlithified interior with no lower mantle. Conductive thermal evolution models that account for heat from radiogenic isotopes, phase changes, and the increase in solar luminosity since the faint young sun period suggest that temperatures in the very near-surface ( $<50 \mathrm{~km}$ depth) have not changed by more than $\sim 20^{\circ} \mathrm{C}$ during the past 4 Gy while temperatures at 100 km depth have evolved by $<30{ }^{\circ} \mathrm{C}$ (Castillo-Rogez and McCord, 2010). The changes in rheology corresponding to such temperature variations do not alter our conclusions regarding the likely constituent phases. We therefore retain a single viscosity profile during the course of our simulations. Future thermal evolution studies that incorporate the effects of porous convection may be used to test further this assumption by refining our knowledge of both the thermal evolution and shape of the vertical temperature profile. Due to the large viscosity gradients
implied by such a temperature profile, we truncated the lowest input viscosities in the FE model at $10^{-5}$ times the maximum value in the domain to prevent ill-conditioned stiffness matrices. Previous testing showed that this approach is sufficient to simulate the effect of larger contrasts with an accuracy of a few percent in flow velocities (Fu et al., 2014).

Compared to the viscosity structure, the elasticity profile of Ceres has a secondary effect on the outcomes of the simulations. For the shear modulus, we adopt a clathrate-like value of $6 \times 10^{9} \mathrm{~Pa}$ for the crust, using methane clathrate as an example, (Helgerud et al., 2009) and a carbonaceous chondrite-like value of $7 \times 10^{9} \mathrm{~Pa}$ for the mantle (Ibrahim, 2012). Increasing both assumed values to $4 \times 10^{10}$ Pa corresponding to anhydrous rock (Matsuyama and Nimmo, 2011) increases the resulting mean relaxation rate in the spherical harmonic degrees 4 through 20 range by $31 \%$ for a viscosity gradient of one decade per 25 km . Variations in the assumed value for the shear modulus therefore do not significantly affect our inferred internal structures.

To evaluate the sensitivity of the computed relaxation rates on the plasticity model adopted, we also implemented a damagedependent yield criterion (Schultz, 1993). Adopting a low Rock Mass Rating (RMR) of 45 corresponding to heavily damaged rock and an intact mechanical strength based on ordinary chondrites (Kimberley and Ramesh, 2011), we find that, on average, relaxation rates become $44 \%$ slower for simulations with viscosity that decays by one logarithmic unit per 15 km , respectively. Furthermore, running identical FE simulations with plasticity suppressed typically decreases relaxation rates by a factor of $\sim 2$ and does not strongly alter the wavenumber dependence of relaxation rates. The assumed criterion for plastic failure therefore does not significantly affect the results of the work with regard to interior structure.

To minimize computation time, we take advantage of the inverse linear relationship between strain rate and viscosity in Stokes flow and run one set of 24 FE simulations corresponding to each viscosity gradient and lower mantle radius with a fiducial surface viscosity of $3 \times 10^{26}$ Pas. We then scale the resulting strain rates to derive the relaxation velocities for structures where the viscosity differs by only a global constant. Even in the case of a viscoelastoplastic model, such scaled results are identical to those obtained by direct computation of a scaled viscosity profile. This is because when viscosities and the model time step are scaled by a constant $c$ and $c^{-1}$, respectively, the viscoelastic parameters that enter into the FE computation remain identical except for the effective viscosity, which is also scaled by $c$ (Eq. (1) in Keller et al. (2013)). The output strain rates are then scaled by $c^{-1}$, while the stresses remain unchanged. As such, the extent and effect of plasticity remains constant, while overall deformation of the body proceeds identically at a rate $c^{-1}$ times the original. To verify this expectation, we ran full viscoelastoplastic models with viscosity gradients of 1 logarithmic unit per 10 km and no stiff lower mantle with surface viscosities of $3 \times 10^{26}$ and $9 \times 10^{27}$ Pas. We found that derived relaxation rates across all wavelengths are offset by the expected factor of 30 with deviations at the $<1 \%$ level, confirming the validity of our scaling approach.

We begin our FE simulations with the expected topography of Ceres at approximately 4.3 Ga , immediately after the expected short period of intense early bombardment (Fig. 3; Marchi et al., 2016). Due to ongoing bombardment and relaxation, the instantaneous topography of early Ceres is unknown, implying that no single initial FE mesh is appropriate. We therefore perform a Monte Carlo analysis whereby a set of 24 FE simulations is run for the same internal structure using randomly generated FE meshes with the same spectral properties as expected for early Ceres. The topography of each initial mesh is described by:


Fig. 3. The cumulative number of collisions with impactors larger than the indicated diameter as a function of time before present. Each curve is the average of 25 Monte Carlo simulations. Impactor sizes are drawn randomly assuming a main-belt like size frequency distribution. Colored " $x$ "s on the $x$-axis denote the expected age of the most recent impact of the given size. The time evolution of the impact flux is taken from O'Brien et al. (2014). For more details see Marchi et al. (2016). (For interpretation of the colors in this figure, the reader is referred to the web version of this article.)
$r(\theta)=r_{0}+\sum_{n=2,4 \ldots}^{100} a_{n} P_{n}[\cos (\theta)]$
where $r(\theta)$ is the radius as a function of co-latitude, $r_{0}$ is the mean radius of $470 \mathrm{~km}, a_{n}$ and $P_{n}$ are the spherical harmonic coefficient and Legendre polynomial for degree $n$. Due to the axisymmetric model, the spherical harmonic order $m$ is always zero. To generate the initial Monte Carlo mesh, we randomly assign a positive or negative sign for $a_{n}$ for each even degree between 2 and 100 . Meanwhile, the magnitude of $a_{n}$ is determined by a power-law to approximate an unrelaxed, impact-saturated surface (Fig. 1; see Section 2.3) except at degree 2 , which is assigned the same amplitude as degree 4 due to lack of constraints on the continuation of a power-law to this wavelength.

We assume no initial topography at the crust-mantle interface aside from the rotational bulge. Impactors with diameter comparable to the crustal thickness are expected to create uplift at the crust-mantle interface below the crater (Ivanov and Melosh, 2013). For such craters, the initial topography is partially compensated isostatically, which may result in slower relaxation rates than in our models. However, other topographic features created in large impacts, such as crater rims and ejecta, are much less compensated initially, if at all. Observationally, the latitudinal dependence in topographic relief (see Section 1; Ermakov et al., in preparation) strongly suggests significant viscous relaxation at long wavelengths, implying that craters were not fully compensated post-impact. Finally, because the rate of topographic relaxation is an approximately linear function of the uncompensated amplitude (e.g., Hager and O'Connell, 1979), partial initial compensation of topography would have a small effect on our modeled relaxation rates compared to the assumed rheological structure. We therefore argue that the assumption of no pre-existing topography at the crust-mantle interface is not expected to produce significant errors, although this may be quantified pending numerical simulations of large impacts on Ceres.

Topography in a purely viscous system is expected to decay following an exponential function (Hager and O'Connell, 1979; Takeuchi and Hasegawa, 1965). We observe this behavior in the output of our FE simulations for the majority of spherical harmonics degrees and viscosity structures. Due to plastic yielding, which
is most significant at early times when the high initial topography results in greater stresses, topographic decay at the longest wavelengths ( $n \leq 8$ ) for steep viscosity gradients $\leq 12.5 \mathrm{~km}$ per decade decay show significant non-exponential behavior (see Supplementary Data discussion). To evaluate the effect of this error, we fit the FE outputs to a power-law instead of exponential function, which improves the quality of fit for setups with high plastic yielding. Viscosity gradients and lower mantle depths inferred from the power-law fitting procedure remain unchanged compared to analysis using a single exponential function (Fig. S3). We therefore regard the non-exponential behavior of decay at long wavelengths as a negligible source of error and compute relaxation rates for even spherical harmonic degree between 2 and 100 by fitting the evolution of topography to an exponential function. This recasting of the FE simulations into a set of wavelength-dependent topographic decay timescales is necessary for combining the effects of topographic growth and attenuation (Section 2.4).

We perform benchmarking runs of our FE model using simplified model parameters to verify the model's consistency with analytic theory. We follow Hager and Clayton (1989) and use the propagator matrix technique to compute the purely viscous relaxation timescale of topography for a finite-viscosity shell overlying a weak interior. Comparison to a relaxation sequence computed from the average of FE runs on 60 initial meshes shows close agreement (Supplemental Data; Fig. S4). At the same time, we use our FE model to compute the full viscoelastic evolution of martian topography, finding close agreement with analytical simulations (Fig. S4C; Zhong, 2002). Finally, we use the observed variations among individual mesh runs to compute uncertainty in the misfit between simulated and observed topographic PSDs for each internal structure.

### 2.3. Topographic growth model

We estimate the rate of topography creation on Ceres due to impact cratering since 4.3 Ga using an asteroid belt impactor flux model. Monte Carlo simulations of Ceres impacts based on the primordial depletion of the main belt (O'Brien et al., 2014) suggest that the surface of Ceres experienced 3-4 times more impacts during the first 250 My after formation than in the subsequent 4.3 Gy (Fig. 3; Marchi et al., 2016). As such, we choose 4.3 Ga as a starting time for our topographic evolution simulations and adopt a saturated, power-law topographic PSD as the initial condition. Based on time evolution models of cratered terrains where the topographic effect of each impact is modeled explicitly (Rosenburg et al., 2015), we assume that, in the absence of viscous relaxation, the topographic PSD approaches a saturation state exponentially:
$f_{k}(t)=f_{k, s a t}\left(1-e^{-t / \tau_{k}}\right)$
where $f_{k}(t)$ is the cumulative frequency of craters at wavenumber $k, f_{k, \text { sat }}$ is the saturation frequency of craters, and $\tau_{k}$ is the timescale of saturation. The observation of a power-law topographic PSD on Vesta supports this assumption. To estimate $\tau_{k}$, we adopt a modern-day main-belt impactor flux model (Bottke et al., 2005; Marchi et al., 2012) and compute the total crater size-frequency distribution (SFD) over a timespan of approximately 3.4 Gy (Fig. 3; Fig. 1 in Marchi et al. (2016)). We then adopt the observed crater SFD on Vesta as a proxy for an impact-saturated surface. By observing the difference between the modeled crater SFD at 3.4 Gy (i.e., $f_{k}(t)$ where $t$ is 3.4 Gy ) and saturation value inferred from Vesta [i.e., $f_{k, \text { sat }}$ ], we can fit for the value of $\tau_{k}$ in the viscously relaxed range of spherical harmonics (Fig. 4A; Table 3), finding that in general the timescale of cratering saturation increases for longer wavelengths as expected from the topographic evolution models of Rosenburg et al. (2015).


Fig. 4. Sensitivity analysis for the assumed cratering rate. (A) Fits to the e-folding cratering saturation timescale based on the computed timescales for Vesta, which are represented by black points (Table 3; Eq. (2)). Black line denotes the leastsquares fit while the blue curves represent the $95 \%$ confidence interval. Secondary $x$-axis labels refer to spherical harmonic degrees ( n ) for Ceres. (B) The misfits between observed and modeled Ceres topographic PSD for a range of assumed impactor fluxes. Baseline model is identical to that presented in Fig. 5D. Color coding of lines is consistent between panels (A) and (B). Differences in the PSD evolution in the three cases occur predominantly at the longest wavelengths, where the high flux end-member results in slower overall decay. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

To evaluate the robustness of our estimated impact saturation timescale, we vary the assumed impact flux between the extreme cases of no impacts after 4.3 Ga and the highest flux within the $95 \%$ confidence interval derived from our fit (Fig. 4A). We then use these lower and upper bound impact fluxes to find the best-fit rheological structure of Ceres and compare them to our baseline results. We find that varying the post-4.3 Ga impactor flux between zero and our baseline model leads to no appreciable change in the best-fit viscosity gradient, while the highest impactor flux endmember weakly favors shallower viscosity gradients (Fig. 4B). As such, varying our adopted impact saturation model does not strongly affect the resulting constraints on Ceres internal structure.

### 2.4. Combining the effects of topographic growth and decay

In the final step of our model, we combine the rates of topographic relaxation and growth derived above to compute the time evolution of the topographic PSD. We use an analytical approach where both topographic relaxation and growth are described by exponential functions. Within this framework, the viscous relaxation of topography in the absence of impacts corresponds to the rate law:

$$
\begin{equation*}
\left(\frac{d a_{n}(t)}{d t}\right)_{\text {relaxation }}=-A_{n} a_{n}(t) \tag{3}
\end{equation*}
$$

Table 3
Cumulative crater frequency observed on Vesta and in 3.4 Gy of modeled evolution with their derived saturation timescales.

| Crater diameter <br> $(\mathrm{km})$ | Cumulative Vesta frequency <br> $\left(\mathrm{km}^{-2}\right)$ | Cumulative modeled frequency at 3.4 Gy <br> $\left(\mathrm{~km}^{-2}\right)$ | Saturation timescale <br> $(\mathrm{Gy})$ |
| :--- | :--- | :--- | :--- |
| 170 | $6.21 \mathrm{E}-6$ | $3.23 \mathrm{E}-6$ | 4.6 |
| 200 | $4.93 \mathrm{E}-6$ | $2.27 \mathrm{E}-6$ | 5.5 |
| 250 | $3.70 \mathrm{E}-6$ | $1.78 \mathrm{E}-6$ | 5.2 |
| 350 | $2.45 \mathrm{E}-6$ | $7.40 \mathrm{E}-7$ | 9.5 |
| 500 | $1.22 \mathrm{E}-6$ | $4.19 \mathrm{E}-7$ | 8.1 |

Notes: Cumulative crater frequencies are from Marchi et al. (2016). Saturation timescales are computed from fits to Eq. (2).
where $a_{n}(t)$ is the topographic amplitude at each spherical harmonic degree $n$ and $A_{n}$ (equal to $\tau_{n}^{-1}$ where $\tau_{n}$ is the $e$-folding timescale of decay) is a rate constant describing the relaxation of topography derived from our FE simulations. Meanwhile, the growth of $a_{n}(t)$ towards the saturation value $\left(a_{\text {sat }, n}\right)$ is given by:
$\left(\frac{d\left[a_{s a t, n}-a_{n}(t)\right]}{d t}\right)_{\text {cratering }}=-B_{n}\left[a_{s a t, n}-a_{n}(t)\right]$
where $B_{n}$ is a constant that reflects the assumed impactor flux. Combining Eqs. (3) and (4) results in an expression for the evolution of topography:
$\left(\frac{d a_{n}(t)}{d t}\right)_{\text {total }}=-\left(A_{n}+B_{n}\right) a_{n}(t)+B_{n} a_{s a t, n}$
which yields the solution:
$a_{n}(t)=a_{0, n} e^{-\left(A_{n}+B_{n}\right) t}+\frac{B_{n} a_{\text {sat }, n}}{A_{n}+B_{n}}\left(1-e^{-\left(A_{n}+B_{n}\right) t}\right)$
where $a_{0, n}$ is the initial topographic amplitude at spherical harmonic degree $n$. The quantity $B_{n} a_{s a t, n} /\left(A_{n}+B_{n}\right)$ represents the equilibrium amplitude of topography at spherical harmonic degree $n$ for a given rate of topographic relaxation and growth.

We can then use Eq. (6) to compute the time evolution of the topographic PSD given the relaxation and growth rate constants $A_{n}$ and $B_{n}$ as computed from the preceding sections, a value for the topographic PSD at saturation, and an assumed initial topographic PSD at time zero. After computing the predicted present-day topographic PSD for each internal structure, we quantify the misfit between the modeled and observed topography using the simple mismatch parameter ( $\Delta^{2}$; Table 1):
$\Delta^{2}=\sum_{n=4}^{20}\left[\log \left(a_{F E, n}^{2}\right)-\log \left(a_{o b s, n}^{2}\right)\right]^{2}$
where $a_{F E, n}^{2}$ is the present-day topographic PSD at spherical harmonic degree $n$ according to our simulation and $a_{o b s, n}^{2}$ is the observed value. We do not consider the mismatch at degrees 2 and 3 due to the absence of equally long physical wavelengths on Vesta, which removes empirical justification of a power-law to describe the initial topography. Furthermore, degree 2 is strongly affected by possible changes in Ceres' rotation rate (Mao and McKinnon, 2016), the extent of which is currently poorly known. We truncate the summation in Eq. (7) at degree 20 as it represents the approximate upper bound to spherical harmonic degrees that show significant viscous relaxation (Fig. 1).

We note that an alternative approach to model topographic evolution would consist of an FE simulation where the surface of the FE mesh is updated with new craters at each time-step based on an impactor flux model. However, a reliable, quantitative description of the initial morphology of large craters is unavailable for Ceres, whose observed craters exhibit transitional behavior between smaller icy and larger rocky bodies (Schenk et al., 2016). Because the largest basins on Ceres appear to exhibit degradation
of topography due to viscous relaxation and possibly other effects (Marchi et al., 2016), observations of these structures do not provide the initial post-impact topography. We therefore do not place new craters into our FE model and use the analytical approach described above to describe the time evolution of the topographic PSD.

## 3. Results

In the case of a water ice-dominated crust with surface viscosity $8 \times 10^{19}$ Pas, the resulting topographic PSD shows highly efficient relaxation of topography at a wide range of wavelengths (Fig. 5A), which agrees with finite element simulations of crater relaxation on Ceres (Bland, 2013; Bland et al., 2016). Importantly, the loss of topography in $<10^{6}$ yr for $\lambda<150 \mathrm{~km}$ is inconsistent with the observed topographic PSD. The presence of a stiff lower mantle does not substantially alter this behavior.

In contrast, model simulations with high surface viscosities, which correspond to less water ice-rich compositions were able to reproduce successfully the observed topography on Ceres (Fig. 5B). We find that the closest matches occur for surface viscosities that decay by one decade per $10-15 \mathrm{~km}$ (Fig. 5D; Table 1). Plastic yielding occurs in the uppermost $20-30 \mathrm{~km}$ of the crust, resulting in maximum effective viscosities of order $10^{25}$ Pas (Fig. 2C-D). In these simulations, the lowest spherical harmonic degrees exhibited the most rapid decay rate, consistent with the expectation that long wavelength features are sensitive to the weaker deep interior of the body.

Model viscosity structures with no lower mantle or a lower mantle with $\geq 100 \mathrm{~km}$ depth from the surface showed the closest match to the observed topographic PSD (Fig. 5D; Table 1). Shallower lower mantles begin to impede significantly the relaxation of topography at the longest wavelengths (Fig. 5C,D), which is inconsistent with the observation that the longest wavelength topography on Ceres undergoes the most rapid relaxation.

## 4. Discussion

### 4.1. Crustal composition

Our inferences for the rheological structure of Ceres may be used to constrain the composition of the crust. We rule out a global shell volumetrically dominated by water ice, which is $10^{3}-10^{4}$ times less viscous than our best-fit rheologies (Fig. 6A). Rock mechanical experiments show that an intimate mixture of ice and stronger materials could provide the required viscosity enhancement as long as the combined fraction of ice and pore space remains below approximately 28 to 38 vol.\%, the upper bound of which we round conservatively to $35 \%$ (Durham et al., 2009; Mangold et al., 2002). Assuming approximately 10\% void porosity (Supplementary Data), the upper bound on the water ice content of Ceres' crust is $25 \%$.

To constrain the composition of the remaining volume of the crust, we compare our inferred rheology to those derived from laboratory creep experiments. We first estimate the temperatures


Fig. 5. Results of FE models showing the evolution of the Ceres topographic PSD assuming (A) a 60 km thick water ice layer with to-scale sketch of viscosity profile versus depth, (B) a strong 46 km thick crust and a range of viscosity gradients, and (C) a strong crust with viscosity decaying by one decade per 10 km depth and a range of depths for a stiff lower mantle. Simulations in (B) and (C) adopt a main belt impactor flux as the mechanism of ongoing topography generation. (D) The mismatch between the modeled and observed topographic PSDs for a range of viscosity gradients (black) and lower mantle depths (red) expressed as the sum of the squares of residuals. The red curve plotted corresponds to a crustal viscosity gradient of 15 km per decade decay, which closely parallels the results for the 10 km per decade decay case (Table 1 ). Only the residuals between degrees 4 through 20 are considered to compute the mismatch due to the poorly known topographic growth rate at degrees $\leq 3$ (see text). Error bars in (D) represent $1 \sigma$ and are derived from the scatter among individual FE model runs for a given viscosity profile. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
in the crust assuming an equatorial surface temperature of 155 K (Hayne and Aharonson, 2015) and a vertical temperature gradient of $0.5-1 \mathrm{~K}$ per km (Fig. 6A; Castillo-Rogez and McCord, 2010; Neveu and Desch, 2015). Carbonate minerals such as magnesite and calcite have been detected remotely on the surface of Ceres (De Sanctis et al., 2015). However, these materials, as well as anhydrous silicates such as olivine, have much higher viscosities than our inferred values (Table 2; Hirth and Kohlstedt, 1996). On the other hand, some salt species, which are observed on Ceres by VIR, are consistent with our inferred rheologies (Fig. 6A, Table 2; see Supplementary Data for discussion of rheological uncertainties).

We can identify the plausible abundances of these candidate crustal components by considering their densities. After accounting for bulk crustal porosity, which is estimated to be $10 \%$ (Supplementary Data), we estimate a grain density of $1430-1554 \mathrm{~kg} \mathrm{~m}^{-3}$ for the crust ( $2 \sigma$ range). Given this value, we first test the simplest hypothesis of a crust dominated by a single component plus void porosity. In isolated form, only hydrated salt phases such as mirabilite and meridianiite provide an adequate match to both the rheology and density. However, because such phases are not observed by the VIR instrument (De Sanctis et al., 2016), they are unlikely to be the predominant component of the crust. Moving on to two-component crustal compositions, a mixture of carbonates or phyllosilicates and $\leq 25$ vol.\% water ice (see above) may be rheologically permissible while containing phases detected by the VIR instrument. However, the high densities of such mixtures ( $>1750 \mathrm{~kg} \mathrm{~m}^{-3}$ ) are likely inconsistent with the grain densities inferred above (Fig. 6B).

Instead, assuming a void porosity of 10 vol.\% (Supplementary Data) and a crustal density of $1287 \mathrm{~kg} \mathrm{~m}^{-3}$ (Ermakov et al., in preparation), we prefer a three-component mixture of $\leq 29$ vol.\% carbonates and/or phyllosilicates, $\leq 25$ vol.\% water ice, and $\geq 36$ vol.\% low density, high strength phases, such as a hydrated salt and/or clathrate hydrate. Adopting instead a $2 \sigma$ upper bound of $1399 \mathrm{~kg} \mathrm{~m}^{-3}$ for the crustal density (Ermakov et al., in press), we compute a lower bound of 29 vol.\% hydrated salt or clathrate hydrate phases. Such compositions can simultaneously satisfy the rheological and density constraints on the crust. In addition, both carbonates and phyllosilicates are observed spectroscopically on the cerean surface. Meanwhile, the lack of remote detection of widespread hydrated salts or clathrate may be due to low abundance, contamination by infall, or the instability of these phases in the low pressure environment of the near surface where the VIR instrument is sensitive (Chastain and Chevrier, 2007). Even so, localized areas of hydrated sodium carbonate has been detected by VIR (Carrozzo et al., in review), although the rheological properties of this material is currently unknown.

We note that we cannot uniquely identify the composition of the low density, high strength component due to the lack of rheological data on the full range of chemically plausible phases. The high abundance of residual solute phases such as hydrated salts relative to water ice is compatible with the efficient impact-driven sublimation of ice (Castillo-Rogez et al., 2016). Alternatively, if the uppermost several kilometers of Ceres are enriched in water ice relative to the bulk crust, the overall volume fraction of water ice may exceed 25 vol.\% while the deeper portions of the crust that


Fig. 6. Rheological and density constraints on the crustal composition of Ceres. (A) Inferred viscosities of Ceres' crust compared to available laboratory creep experiments. Gray area denotes the approximate range of viscosities derived from topographic relaxation models assuming viscosity decay of one decade per $10-15 \mathrm{~km}$ depth between latitudes $-60^{\circ}$ and $+60^{\circ}$. Phases that overlap the gray area are consistent with, but not required by, the inferred Ceres viscosities. Abbreviations in (A) are: $\mathrm{MS} 7=\mathrm{MgSO}_{4} \cdot 7 \mathrm{H}_{2} \mathrm{O}, \mathrm{MS} 11=\mathrm{MgSO}_{4} \cdot 11 \mathrm{H}_{2} \mathrm{O}, \mathrm{NS} 10=\mathrm{Na}_{2} \mathrm{SO}_{4} \cdot 10 \mathrm{H}_{2} \mathrm{O}$. Viscosities for magnesite $\left(\mathrm{MgCO}_{3}\right)$ and calcite $\left(\mathrm{CaCO}_{3}\right)$ plot above the field shown. We adopt a strain rate of $3 \times 10^{-18} \mathrm{~s}^{-1}$ consistent with our FE simulation output to compute viscosities from experimental creep laws (see Table 2 for references). Grain size is 1.0 mm . (B) Density and rheology of three component mixtures. Each point in the plotted space represents a composition with phyllosilicate and hydrate volume fractions indicated by the $x$ and $y$ axes, respectively. The remainder consists of water ice and $10 \%$ void porosity. The compositions shaded blue are consistent with the rheological constraint of $<35$ vol.\% weak phases (i.e., water ice plus void). Orange and green lines denote compositions with indicated densities assuming that the hydrate phase is methane clathrate or mirabilite $\left(\mathrm{Na}_{2} \mathrm{SO}_{4} \cdot 10 \mathrm{H}_{2} \mathrm{O}\right)$, respectively. Reference densities for serpentine, methane clathrate hydrate, mirabilite, and water ice are $2500 \mathrm{~kg} \mathrm{~m}^{-3}, 920 \mathrm{~kg} \mathrm{~m}^{-3}, 1490 \mathrm{~kg} \mathrm{~m}^{-3}$, and $930 \mathrm{~kg} \mathrm{~m}^{-3}$, respectively (Forrisdahl, 2010; Fortes et al., 2008). Values of 1209, 1287, and $1399 \mathrm{~kg} \mathrm{~m}^{-3}$ refer to the bulk crustal density. Because methane clathrate and mirabilite represent end-members in density among the hydrated phases, the area between the green and orange lines in the blue shaded region is simultaneously consistent with both rheological and density constraints for a given bulk crustal density. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
controls long wavelength topography remains rheologically strong with $\leq 25$ vol.\% water ice. In this scenario, lower bulk concentrations of hydrated salts and clathrate hydrates would be required.

Our inferred crustal composition of water ice, carbonates, phyllosilicates, and salt and/or clathrate hydrate phases is most consistent with models of an ancient ocean layer that underwent progressive freezing, leading to the concentration of anhydrous and hydrated salts (Neveu and Desch, 2015). Based on the lack of definitive regional variations in the extent of viscous relaxation
apart from latitudinal trends, the composition of Ceres's crust as inferred from rheology is broadly uniform. This observation suggests that any ancient ocean layer likely covered large fractions of the surface and was potentially globally continuous. This result points to significant global aqueous fluid flow in early Ceres, which led to the alteration of accreted silicate phases and the open system delivery of aqueous fluid and solutes to a surface ocean layer, a scenario also supported by spectroscopic observations of Occator crater (De Sanctis et al., 2016) and measurements of elemental abundances (Prettyman et al., 2017). Ceres therefore suggests that global scale porous flow was an important process on planetesimal bodies despite low localized permeabilities measured from carbonaceous chondrites matrices (Bland et al., 2009).

### 4.2. Deep interior evolution

Our simulations indicate that a mechanically strong lower mantle in Ceres, if it exists, must be deeper than 100 km from the surface (Fig. 5C). Because the best-fit thickness of the crust is 41 km , this depth implies that at least the uppermost $\sim 59 \mathrm{~km}$ of the silicate-rich mantle is rheologically weak with effective viscosities below $\sim 10^{21}$ Pas. The low present-day temperatures below $100^{\circ} \mathrm{C}$ (Castillo-Rogez and McCord, 2010) and low shear stresses less than 1 MPa (Fig. 2C) expected for the uppermost region of the mantle limit the variety of mechanisms that can accommodate deformation. As discussed in Section 2.2, anhydrous silicates are not expected to deform under such conditions. Even unconsolidated dry phyllosilicates, which have frictional coefficient $>0.23$, would not yield under the $>3 \times 10^{6} \mathrm{~Pa}$ lithostatic pressures of the upper mantle (Procter and Hirth, 2016). On the other hand, pore fluids enhance deformation by mediating viscous creep via pressure solution and by lowering the effective normal stress, thereby facilitating frictional sliding even under the relevant low temperature and stress conditions (see discussion in Section 2.2). We therefore conclude that the mechanically weak upper mantle likely contains liquid pore fluids, implying that temperatures at the crust-mantle interface of Ceres during its evolution since 4.3 Ga , including the present-day, has exceeded the eutectic point of the aqueous solution.

At the same time, our result implies that peak metamorphism in the upper mantle during Ceres' early history must have been sufficiently low to permit the retention of interstitial water. Ordinary chondrites with petrologic grade $\geq 3.7$ generally contain no significant water (Huss et al., 2006; Jarosewich, 1990). However, textural, chemical, and isotopic evidence suggest that these meteorites experienced aqueous alteration before metamorphism (Grossman et al., 2000), implying that interstitial water in ordinary chondrites was lost upon heating to higher than $525-600^{\circ} \mathrm{C}$. Similarly, metamorphosed CV carbonaceous chondrites, such as Allende (petrologic type >3.6; Bonal et al., 2006), are also hypothesized to have undergone aqueous alteration followed by dehydration during metamorphism to $325-600^{\circ} \mathrm{C}$ (Cody et al., 2008; Krot et al., 1998). Finally, simple theoretical considerations suggest that devolatilization via porous flow in planetesimals is efficient once the silicate matrix is sufficiently hot to eliminate pore spaces via viscous compaction (Fu and Elkins-Tanton, 2014). Using the rheology of wet olivine (Hirth and Kohlstedt, 1996), which is produced by metamorphic dehydration in carbonaceous chondrites (Akai, 1992), lithostatic pressures in the upper mantle of Ceres can induce compaction on $<1 \mathrm{~Gy}$ timescales at temperatures $>630^{\circ} \mathrm{C}$. Taken together, these observations and calculations suggest that the upper mantle of Ceres has not experienced peak metamorphic temperatures in excess of $\sim 600^{\circ} \mathrm{C}$.

Given that the lower mantle is, at minimum, 100 km below the surface, these maximum metamorphic temperatures suggest that the accretion of Ceres occurred at least $\sim 3$ My after CAIs as-
suming a conductive thermal history (Castillo-Rogez and McCord, 2010). Meanwhile, an extended accretion process lasting over several million years may also produce the required low temperatures at 100 km depth (Neumann et al., 2015). An earlier accretion of Ceres may be permitted if convection in the deep interior, through porous flow or soft sediment deformation (Bland et al., 2013; Neveu et al., 2015; Travis and Feldman, 2016; Young et al., 2003), was able to transport heat efficiently to the surface. The past occurrence of global porous flow delivering aqueous solution to form the modern crust (see Section 4.1) is consistent with the observed presence of water ice and depletion of Fe in Ceres' regolith (Prettyman et al., 2017) and favors the hypothesis that convection of aqueous fluids played a significant role in the thermal evolution of Ceres' interior.

The low metamorphic temperatures and weak rheology of the upper mantle suggests a composition dominated by phyllosilicates instead of anhydrous rock. Such a composition is consistent with the correspondence between possible large impact basins and stronger spectral signatures of ammoniated phyllosilicates, which may have been excavated from the deep interior (Marchi et al., 2016). At the same time, the inferred high degree of hydration observed on the global scale supports the past occurrence of extensive fluid-rock interactions (Ammannito et al., 2016; De Sanctis et al., 2016; Neveu et al., 2015).

Our inferred composition for the upper mantle may be compared to the bulk density of the silicate-rich mantle derived from admittance analysis (Ermakov et al., in preparation), which yielded a best-fit value $2434 \mathrm{~kg} \mathrm{~m}^{-3}$. Unmetamorphosed carbonaceous chondrites in the CI and CM groups have grain densities in a range between 2400 and $2800 \mathrm{~kg} \mathrm{~m}^{-3}$ (Consolmagno et al., 2008). Because up to $\sim 10$ vol. \% interstitial water may be retained in fine-grained sediments at the pressures corresponding to the deep interior of Ceres (Mondol et al., 2007), the corresponding bulk densities of such compositions are between 2260 and $2620 \mathrm{~kg} \mathrm{~m}^{-3}$. These values are closely compatible with the admittance-derived bulk density above. Due to the range of possible densities for a water-saturated chondrite composition, the presence of a denser lower mantle is not required but cannot be ruled out.

Finally, the result that the uppermost $\geq 59 \mathrm{~km}$ of the silicaterich mantle is weakly lithified holds implications for the fate of impact ejecta and the potential for meteorites to sample Ceres. Mechanical properties are unlikely to affect strongly the impact ejection process itself given that fragmentation on Ceres occurs in the gravity regime (Ryan and Melosh, 1998). However, as discussed above and in Section 2.2, a low strength upper mantle most likely implies an unlithified, water-saturated soft sediment composition. Such materials, once ejected, may disintegrate due to the evaporation of interstitial fluids under zero pressure conditions. The resulting particles, which may be sub-micrometer in size assuming a chondrite-like size frequency distribution (Bland et al., 2009), would be susceptible to efficient removal due to solar radiation pressure (de Pater and Lissauer, 2010). At the same time, ejecta from the volatile-rich crust may be subject to a similar process of volatilization and removal (Rivkin et al., 2014). Together, these studies may explain the absence of an asteroid family associated with Ceres, even given large, mantle-excavating impacts that occurred in the past (Marchi et al., 2016). Alternatively, the absence of a modern Ceres family in the central main belt may be due to dynamical effects that deplete fragments ejected from Ceres (Carruba et al., 2016).

Even if some ejected fragments of the mantle avoided disintegration and were delivered to Earth-crossing orbits, they are unlikely to have survived the stresses associated with atmospheric passage as macroscopic meteorites (Sears, 1998). As such, an unlithified upper mantle potentially composing a majority of Ceres' mass represents a large reservoir of early solar system material
that may be sampled by interplanetary dust particles (IDPs), but are unlikely to be present in our meteorite collection (Vernazza et al., 2015). On the other hand, impacts may cause localized lithification of the near-surface (Ashworth and Barber, 1976; Bischoff et al., 1983), leading to the possibility that meteorites may sample Ceres' upper crust.

## 5. Conclusions

Using a viscoelastoplastic FE model and assuming that impacts are the dominant source of topography on Ceres at spherical harmonic degree $\geq 4$, we have identified a range of possible rheological profiles that can replicate the observed topographic spectrum of Ceres. The preferential attenuation of topography at wavelengths longer than 150 km is most consistent with high effective surface viscosity of order $10^{25}$ Pas, which decays rapidly with depth at a rate of 1 order of magnitude per 10 to 15 km . Furthermore, the extent of long wavelength relaxation inferred on Ceres is possible only if viscosities do exceed approximately $10^{21}$ Pas within the uppermost 100 km , below which our models are no longer sensitive. These and other results of this work are only weakly sensitive to the assumed rate of impact-induced topography creation.

These inferred surfaces viscosities are at least $10^{3}$ times higher than that of pure water ice, implying that water ice composes $<25$ vol.\% of Ceres' crust. A rock-ice mixture in such proportions is significantly denser than the $1287 \mathrm{~kg} \mathrm{~m}^{-3}$ bulk density of the crust derived from admittance analysis (Ermakov et al., in preparation). Instead, a three component mixture of $<25$ vol.\% water ice, $<36$ vol.\% phyllosilicates or carbonates, and $>29$ vol.\% low density, high-strength phases, such as hydrated salts or clathrate hydrates, is able to satisfy simultaneously the rheological and density constraints (Fig. 6B). The high inferred abundance of water ice and hydrated salt or clathrate phases strongly suggests the past presence of a global ocean or at least globally significant areas of solute-rich fluids at the surface, the progressive freezing of which led to the concentration of these species.

We also use our models to infer the strength of the silicaterich mantle underlying the volatile-rich crust. Adopting 41 km as the depth of the density discontinuity between the crust and mantle (Ermakov et al., in preparation), the dense region between 41 and at least 100 km must have low effective viscosities of $<10^{21}$ Pas despite its silicate-rich composition. Given the low temperature and stress regime, such low viscosities are indicative of deformation via pressure solution or frictional sliding in a fluid-saturated medium. Because these deformation mechanisms require the persistence of interstitial fluids, we infer that the outermost $\geq 59 \mathrm{~km}$ of the silicate-rich mantle ( $\geq 36 \%$ by volume) did not experience metamorphic dewatering, implying peak metamorphic temperatures below $\sim 600^{\circ} \mathrm{C}$ throughout the interior of Ceres.

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## Appendix A. Supplementary material

Supplementary material related to this article can be found online at http://dx.doi.org/10.1016/j.epsl.2017.07.053.

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