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1 Consequences of large impacts on Enceladus' core shape

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8 **Abstract**

9 The intense activity on Enceladus suggests a differentiated interior consisting
10 of a rocky core, an internal ocean and an icy mantle. However, topography and
11 gravity data suggests large heterogeneity in the interior, possibly including sig-
12 nificant core topography. In the present study, we investigated the consequences
13 of collisions with large impactors on the core shape. We performed impact simu-
14 lations using the code iSALE2D considering large differentiated impactors with
15 radius ranging between 25 and 100 km and impact velocities ranging between
16 0.24 to 2.4 km/s. Our simulations showed that the main controlling parame-
17 ters for the post-impact shape of Enceladus' rock core are the impactor radius
18 and velocity and to a lesser extent the presence of an internal water ocean and
19 the porosity and strength of the rock core. For low energy impacts, the im-
20 pactors do not pass completely through the icy mantle. Subsequent sinking and
21 spreading of the impactor rock core lead to a positive core topographic anomaly.
22 For moderately energetic impacts, the impactors completely penetrate through
23 the icy mantle, inducing a negative core topography surrounded by a positive
24 anomaly of smaller amplitude. The depth and lateral extent of the excavated
25 area is mostly determined by the impactor radius and velocity. For highly en-
26 ergetic impacts, the rocky core is strongly deformed, and the full body is likely
27 to be disrupted. Explaining the long-wavelength irregular shape of Enceladus'

28 core by impacts would imply multiple low velocity (< 2.4 km/s) collisions with
29 deca-kilometric differentiated impactors, which is possible only after the LHB
30 period.

31 *Keywords:* Enceladus, Impact processes, Cratering, Interiors, Accretion

32 1. Introduction

33 Despite its small size ($R = 252$ km), Saturn's moon Enceladus is one of
34 the most geologically active body of the Solar System. Its surprising endogenic
35 activity is characterized by a very active province at the South Pole, from which
36 eruptions of water vapor and ice grains emanating from warm tectonic ridges
37 have been observed by the Cassini spacecraft (*Porco et al., 2006; Hansen et al.,*
38 *2006; Waite et al., 2006; Spencer et al., 2006*). This activity is associated with
39 a huge heat power estimated between 5 and 15 GW from thermal emission
40 (*Spencer and Nimmo, 2013*), which implies a warm interior, consistent with a
41 liquid water layer underneath the ice shell and a differentiated interior (*Nimmo*
42 *et al., 2007; Schubert et al., 2007*). Models of tidal dissipation may explain why
43 the activity is concentrated at the poles, where dissipation is predicted to be
44 maximal (*Tobie et al., 2008; Behouňková et al., 2010*). However, there is still no
45 satisfactory explanation for why this activity is located only in the south, and
46 not in the north.

47

48 Based on the global shape data which show a depression at the south pole
49 (*Thomas et al., 2007*), it has been proposed that the ocean may be located only
50 in the southern hemisphere (*Collins and Goodman, 2007*), thus explaining why
51 the activity would be concentrated at the south (*Tobie et al., 2008*). Grav-
52 ity and shape data indicate that such an ocean would be at depths of about
53 30 to 40 kilometers and extend up to south latitudes of about 50° (*Iess et al.,*
54 *2014*). It has been proposed that the dichotomy between the north and south

55 hemispheres may be the result of asymmetry in core shape (*McKinnon, 2013*).
56 Due to the low pressure and moderate temperature expected in Enceladus' core,
57 large topography anomalies may indeed be retained on very long periods of time
58 (*McKinnon, 2013*) and may explain why convection-driven activities in the ice
59 shell is confined only to the south polar terrain (*Showman et al., 2013*). Besides
60 the south polar depression, core topography anomalies could explain, at least
61 partly, the existence of other big depressions observed at moderate latitudes (be-
62 tween 15°S and 50°N) and uncorrelated with any geological boundaries (*Schenk*
63 *and McKinnon, 2009*).

64

65 *McKinnon (2013)* proposed three hypotheses to explain the possible irreg-
66 ularity of Enceladus' rocky core: accretional melting of the outer region of the
67 icy moon associated with a degree-one instability; accretion of icy protomoons
68 around irregular rock chunks; and collisional merger of two previously differ-
69 entiated protomoons. Here we test the latter hypothesis by investigating the
70 consequences of the collision of a large differentiated impactor on the shape of
71 Enceladus' core. Collisions with large differentiated bodies were likely at the
72 end of satellite accretion, during the final assemblage phase (e.g. *Asphaug and*
73 *Reufer, 2013*). Large impact basins on other saturnian moons (e.g. Iapetus
74 (*Giese et al., 2008*), Mimas (*Schenk, 2011*), Titan (*Neish and Lorenz, 2012*))
75 and other solar system bodies (e.g. Vesta (*Schenk et al., 2012*)) could represent
76 remnant evidences of such collisions. Large impacts occurring at the end of
77 the accretion and after, during the rest of the satellite's evolution, likely influ-
78 enced the internal structure and especially the shape of its rocky core. It is also
79 important to determine the conditions under which Enceladus would have sur-
80 vived disruption by collisions with deca-kilometric objects, which would place
81 constraints on its accretion and the subsequent impact history.

82

83 To constrain the consequences of large-scale impacts on Enceladus, we sim-
84 ulated head-on collisions of differentiated impactors with diameter ranging be-
85 tween 50 and 200 km using the iSALE2D shock physics code (*Wünnemann*
86 *et al.*, 2006; *Collins et al.*, 2004; *Davison et al.*, 2010). From these simula-
87 tions, we tracked the evolution of rock fragments coming from the impactor and
88 the impact-induced modification of Enceladus’s core shape. In particular, we
89 quantified the sensitivity in these outcomes to key model parameters, such as
90 impactor velocity and radius, as well as structure and mechanical properties
91 of Enceladus’ interior (porosity, strength, temperature profile, core size, pres-
92 ence of an internal ocean). In section 2, we describe our numerical modelling
93 approach; in section 3 we present our results. We discuss our results in the con-
94 text of the presence of a water ocean in section 4. Conclusions are highlighted
95 in section 5.

96

97 **2. Impact modeling**

98 To model the thermo-mechanical evolution of material during an impact be-
99 tween two differentiated icy bodies, we use iSALE2D (*Wünnemann et al.*, 2006;
100 *Collins et al.*, 2004). This numerical model is a multi-rheology, multi-material
101 shock physics code based on the SALE hydrocode (*Amsden et al.*, 1980) that
102 has been extended and modified specifically to model planetary-scale impact
103 crater formation (e.g., *Amsden et al.*, 1980; *Melosh et al.*, 1992; *Ivanov et al.*,
104 1997; *Collins et al.*, 2004; *Wünnemann et al.*, 2006; *Davison et al.*, 2010). In
105 our simulations, the target structure and the impactor were simplified to two-
106 or three- layer spherical bodies consisting of a rocky core, an icy mantle and for
107 the three-layer case an internal ocean. Interpretation of gravity data collected

108 by the Cassini spacecraft indicates that the core density could be as low as 2400
109 kg m^{-3} , corresponding to a core radius of about 200 km (*Iess et al.*, 2014).
110 However, as Enceladus appears to be relatively far from hydrostatic equilibrium
111 (*Iess et al.*, 2014), there are still significant uncertainties on the core radius
112 and density. The low core density inferred from gravity data suggests that the
113 rocky core might be significantly porous, with pores filled by water ice and/or
114 liquid water, and that a significant fraction of the core may consist of hydrated
115 silicate minerals. Currently, iSALE2D does not have provision to describe the
116 behavior of an ice-rock or water-rock mixture. In our simulations, for simplicity,
117 we assume complete segregation of the rock and ice-water phase into discrete
118 layers and we consider dunite as representative of the rock phase (with density
119 $\rho_s = 3330 \text{ kg m}^{-3}$). We reduce the density of the core by including some ini-
120 tial porosity ϕ (defined as the ratio of pore volume to total volume) within it,
121 varying from 0 to 50%, corresponding to radius varying between typically 160
122 km and 200 km. Assuming a core made of pure dunite, a radius as large as 200
123 km is consistent with a core porosity of about 50%, which is at the upper end
124 of the estimated porosity in large asteroids (*Lindsay et al.*, 2015). A significant
125 fraction of the core may also consist of hydrated minerals such as serpentine.
126 In this case a 200 km core radius would imply a lower porosity. For simplicity,
127 we consider only dunite as core materials and vary the porosity up to values of
128 50%. We also test the possible effect of porosity in the ice shell by considering
129 values up to 20% as suggested by *Besserer et al.* (2013).

130

131 In our models, we consider the extreme case where the pores of both ice
132 and rocks consist of voids, and are not filled with secondary materials (i.e. wa-
133 ter or ice in rock pores). The difference between saturated porosity (with ice
134 or liquid water) and voids may lead to differences in terms of mechanical and

135 thermal properties. This aspect will be discussed in the last section. The effect
 136 of both rock and ice porosity is treated using the $\epsilon - \alpha$ porosity compaction
 137 model (*Wünnemann et al.*, 2006; *Collins et al.*, 2013), which accounts for the
 138 collapse of pore space by assuming that the compaction function depends upon
 139 volumetric strain. For sake of simplicity, we assume that the impactor material
 140 has an identical composition and porosity to those of the target.

141

142 The impact velocity v_{imp} can be decomposed into two contributions:

$$v_{imp} = \sqrt{v_{esc}^2 + v_{\infty}^2} \quad (1)$$

143 where v_{esc} is the escape velocity of the impacted planet and v_{∞} is the velocity of
 144 the impactor at a distance much greater than that over which the gravitational
 145 attraction of the impacted planet is important. The escape velocity of Ence-
 146 ladus is $v_{esc} = 240$ m/s. As we consider collisions with relatively large objects
 147 ($R_{imp} = 25 - 100$ km), we limit our analysis to moderate relative velocities,
 148 varying between v_{esc} and $10 \times v_{esc}$, in order to limit the impact-induced defor-
 149 mation of the satellite and avoid full disruption (*Benz and Asphaug*, 1999; *As-*
 150 *phaug*, 2010). Moreover, this low-velocity impact regime is representative of the
 151 collisional environment at the end of the accretion. Indeed, N-body simulations
 152 from *Dwyer et al.* (2013) show that random impact velocity of proto-satellites
 153 mostly ranges between v_{esc} and $5v_{esc}$.

154

155 We approximated the thermodynamic response of the icy material using the
 156 Tillotson EoS for Ice as in *Bray et al.* (2008) and of the rocky material using the
 157 ANEOS EoS for dunite material as in *Benz et al.* (1989); *Davison et al.* (2010)
 158 (see Tab. 1 for parameter values). Standard strength parameters for dunite were
 159 used to form the static strength model for the rocky core (*Collins et al.*, 2004;

160 *Davison et al.*, 2010). The static strength model for ice used in iSALE was de-
161 rived from low temperature, high pressure laboratory data and accounts for the
162 material strength dependence on pressure, damage and thermal softening (*Bray*
163 *et al.*, 2008). We also explored the effect on our results of the cohesion of the
164 damaged material (referred to here as Y_i for ice and Y_s rocks), which represents
165 the minimum zero-pressure shear strength of cold material (strength is reduced
166 to zero at the melt temperature). The minimum strength values considered in
167 our models range between 10 – 500 kPa for ice and 100 – 10⁴ kPa for silicate
168 material. The Tillotson EoS for ice is severely limited in its applicability for hy-
169 pervelocity impact; it includes no solid state or liquid phase changes. However,
170 as we limit here our analysis to low velocity encounters ($240 < v_{imp} < 2400$
171 m s^{-1}), thought to be dominant at the end of the accretion, as shown in our
172 simulations, no significant ice melting occurs and the use of Tillotson EoS is a
173 reasonable approximation. We also used the Tillotson EoS for the liquid water.

174

175 Material weakening during impact may also be achieved by acoustic fluidiza-
176 tion and/or thermal softening (*Melosh and Ivanov*, 1999), the latter of which is
177 especially efficient for large-scale events (*Potter et al.*, 2012). Our simulations
178 including acoustic fluidization that assumed typical block-model parameters fa-
179 vored in other works showed no significant effect on simulation results (see also
180 discussion section). Hence, for simplicity and to reduce the number of free pa-
181 rameters, we chose to neglect acoustic fluidization. We do, however, include the
182 effect of temperature on shear strength using the temperature-strength relation-
183 ship proposed by *Ohnaka* (1995) and described by *Collins et al.* (2004) and we
184 set the thermal softening coefficient in this expression to 1.2 as suggested by
185 *Bray et al.* (2008). Since we consider the thermal softening during the impact,
186 the thermal structure of Enceladus before the impact is probably a key parame-

187 ter governing the post-impact state. However, the early temperature profile for
188 such a small body is poorly constrained. Accretionary models seem to favour
189 a cold accretion with inner temperatures close to the equilibrium temperature
190 (*Schubert et al.*, 1981; *Monteux et al.*, 2014). To test the influence of the initial
191 thermal conditions, we consider three different pre-impact temperature profiles
192 for the impacted moon: constant temperature, conductive profile, two-layered
193 advective profile. The impactor is assumed to have a constant temperature with
194 $T = 100$ K.

195

196 Owing to the axisymmetric geometry of iSALE2D, we consider only head-on
197 collisions (impact angle of 90° to the target tangent plane). The role of impact
198 angle is left to future studies. To limit computation time, a 1-to-2 km spatial
199 resolution is used, which is sufficient to describe the deflection of the rock core
200 surface. The gravity is calculated from the density structure. For the largest
201 and fastest impacts, we use iSALE2D’s self-gravity gravity model (*Collins et al.*,
202 2011) to correctly assess the gravity field as the body is strongly deformed and
203 the center of mass of the target moves upon the collision. As this self-gravity
204 model is expensive in terms of computational time, we limit our post impact
205 monitoring to the time needed to deform the rocky core (i.e. we consider that
206 the fall-back of icy material and the icy-mantle slumping has only a very minor
207 effect on the morphology of the rocky core). For all the impacts characterized
208 here, this corresponds to the first hour after the impact.

209

210 3. Numerical results

211 3.1. Non-porous models

212 Fig. 1 shows three characteristic simulations: ($v_{imp} = 10v_{esc}$, $R_{imp} = 25$
213 km), ($v_{imp} = 10v_{esc}$, $R_{imp} = 75$ km) and ($v_{imp} = v_{esc}$, $R_{imp} = 75$ km). After
214 such events, a large volume of Enceladus' mantle is displaced or escapes the
215 orbit of the icy moon. To get a quantitative measure of deformation induced
216 by the impact event, we monitor the plastic strain experienced by the impacted
217 material. In particular, we calculate the total plastic strain which is the accumu-
218 lated sum of plastic shear deformation, regardless of the sense of shear (*Collins*
219 *et al.*, 2004). As represented in Fig. 1, the icy material is highly disturbed
220 by the impact and most of the plastic deformation occurs in this layer. For
221 the largest impact velocities (Fig. 1, left and middle), deformation also occurs
222 at the top of the rocky core and leads to the formation of a depression. The
223 material removed from the depression is displaced in a very small uplift of the
224 core, surrounding the depression.

225
226 For small impact velocities (Fig. 1, right), the icy mantle is also highly
227 deformed but the impactor's rocky core is trapped within the ice layer. In
228 this low-velocity case, the deformation of the target's core and the impact melt
229 production are minor but the surrounding ice is warmed up. Hence, over a longer
230 time scale governed by a Stokes' flow, the impactor's core gently spreads over
231 the surface of the pre-existing rocky core favoring the formation of successive
232 fragmented silicate layers (*Roberts*, 2015). Depending on the impactor size and
233 impact velocities, our simulations show that core merging occurs into three
234 distinct regimes (Fig. 2):

235 (1) For small impactors and impact velocities close to $\sim v_{esc}$, the impactor's
236 core is simply buried within Enceladus' icy mantle at a depth that scales with

237 the penetration depth p (Orphal *et al.*, 1980; Murr *et al.*, 1998) :

$$p/R_{imp} = Av_{imp}^{2/3} \quad (2)$$

238 where A is a function of the bulk sound velocity, the geometry and density
239 difference between the impactor and the target.

240 (2) For higher impact velocities or larger impactors, the kinetic energy in-
241 creases and hence penetration of the impactor’s core through the target ice man-
242 tle is facilitated. When the impactor penetration depth, p (Eq.2), exceeds the
243 ice-mantle thickness, δ_m , the impactor induces a deflection of the core bound-
244 ary (Fig. 2), the amplitude of which depends on the impactor energy remaining
245 after crossing the ice mantle. For $p \sim \delta_m$ or slightly larger, the impactor core
246 spreads above the target’s core (leading to a positive core-topography anomaly
247 defined as the difference of post- and pre-impact core radii below the impact
248 site). (3) However, if more energy is available, $p > \delta_m$ and the core is strongly
249 deformed, possibly leading to severe deformation of the satellite, as illustrated
250 in Fig. 2 for impactors larger than 75 km and/or impact velocities $\geq 10v_{esc}$. It
251 has to be noted that, as we limit our post impact monitoring to one hour, for
252 the most energetic impact cases with large impact velocities (≥ 6 km/s) and
253 large impactor radii (≥ 75 km) the rocky material excavated from Enceladus’
254 core and orbiting around the moon is still moving with significant velocity at
255 the end of the simulation.

256

257 The thermal softening is an efficient process for large-scale events (Potter
258 *et al.*, 2012). This process is strongly dependent on the pre-impact temper-
259 ature field that is unfortunately poorly constrained. To test the influence of
260 the pre-impact thermal state, we consider three different pre-impact temper-
261 ature profiles for the impacted moon (Fig. 3): constant temperature (with

262 $T \sim 100K$), conductive profile (with a temperature gradient value of 1 K/km),
263 two-layered convective profile (with a core temperature of 450 K and a mantle
264 temperature of 250 K). As illustrated in Fig. 3, a hotter temperature profile
265 in the icy shell strongly enhances the ice flow back and the refill of the core
266 depression. One hour after the impact, a large cavity remains open in the icy
267 mantle for the constant and cold temperature case. For the two-layered convec-
268 tive case where the mantle temperature is close to the melting temperature of
269 ice, the icy mantle rapidly flows back leading to a huge jet of ice at the impact
270 site. However, even if considering three pre-impact thermal states significantly
271 modifies the post-impact dynamics of the icy mantle, this only weakly affects
272 the depth of the depression within the rocky core that ranges between 12 and
273 15 km (Fig. 3). Hence, in the following, we consider models with a constant
274 pre-impact temperature field.

275 3.2. Influence of ice and rock porosity

276 The porosity of the material involved during the impact is known to be a
277 key factor in both the fragmentation and disruption of the impactor and the
278 target (*Jutzi et al.*, 2008, 2009), and therefore it may play a role in our results.
279 Enceladus is believed to contain a high degree of porosity, as are many other
280 small bodies in the different populations of asteroids and comets (e.g. *Lindsay*
281 *et al.*, 2015). To explain the long-wavelength topography of Enceladus, recent
282 models also invoke porosity values ranging between 20 to 30 % within the icy
283 mantle of Enceladus (*Besserer et al.*, 2013). We monitored the rocky core defor-
284 mation as a function of the icy mantle porosity with porosities ranging between
285 0 and 20%. Similar to the simulations with different initial thermal conditions
286 (Fig. 3), the dynamics of post-impact ice flow in the the deep cavity depends
287 significantly on the porosity, as it affects the ice mechanical properties (Fig. 4).
288 When the ice porosity equals 20% and because the compacted ice is thermally

289 softened, the icy material (which is heated by impact to temperatures up to 250
290 K) re-fills the impact induced cavity in less than one hour.

291

292 Nevertheless, as illustrated in Fig. 5, the effect of the icy mantle porosity on
293 the post-impact core morphology is rather small, at least for initial porosities
294 ranging from 0 to 20% and for impact parameters leading to moderate core
295 deformation ($v_{imp} = 10v_{esc}$ and $R_{imp} = 25$ km). Fig. 6 shows the depth of the
296 impact-induced core depression as a function of the mantle porosity. According
297 to this figure, the depth of the depression ranges between 8 and 13 km. As
298 mentioned earlier (see Fig. 4), the major influence of the mantle porosity is
299 its ability to flow back and refill the core depression. As the impacted ice is
300 severely deformed and compacted during the shockwave propagation, the im-
301 pact will increase locally the porosity and the temperature of the icy mantle
302 below the impact site.

303

304 Fig. 6 and Fig. 7 show that the influence of core porosity on core defor-
305 mation is larger than the corresponding influence of mantle porosity. Indeed,
306 increasing the porosity of the core from 0 to 50 % (and thus increasing its radius
307 from 160 to 200) increases the maximum depth of the depression caused by the
308 impact from ~ 13 km to ~ 31.5 km. To explain this feature, two effects shall
309 be invoked. The first one is that increasing the rocky core porosity increases
310 its size to maintain its mass. Hence, the top of the rocky core is closer to the
311 surface and the impactor penetration depth needed to deform the rocky core
312 is reduced accordingly. The second one is that porosity can enhance the rocky
313 core deformation because the core material is less dense and easier to compact.
314 To decipher between these two effects we ran a non-consistent model with a
315 non-porous 200 km rocky core radius surrounded by a 50 km thick icy mantle

316 (Fig. 8, first column). At the end of this model, the depression depth is 18.5
317 km (compared to 31.5 km when the rocky core porosity is 50% and to 13 km
318 when the rocky core has a radius of 160 km) meaning that both increasing the
319 core size and the porosity favour deeper impact-induced depressions. This also
320 suggests that density/compaction has a greater influence than core radius on
321 the depth of the impact-induced core depression. We also ran a model with a
322 50% porosity 160 km rocky core radius (Fig. 6) where the obtained depression
323 depth is 15 km (close to value obtained in the non-porous case). In this non-
324 consistent case, a 8 km-thick ice block is trapped between the impactor and
325 the target's core that prevents the formation of a deeper cavity. We should,
326 however, keep in mind that in our simulations, we consider void porosity, while
327 in reality pores should be filled by liquid water or water ice, which would affect
328 compaction. The results presented here should be considered as an estimation
329 of the maximal effect associated to impact-induced porosity compaction.

330

331 3.3. Influence of minimum strength values and water ocean

332 In all the models described above, the minimum strength values were set to
333 $Y_i = 500$ kPa for ice and $Y_s = 10$ MPa for silicate material. These values repre-
334 sent the upper range of the plausible values since recent estimates of the strength
335 of the surface of comet Tempel-1 obtained minima strength values in the order
336 of 1-10 kPa (*Richardson and Melosh, 2013*). For the minimum strength of the
337 rocky mantle, this value is also likely to range between the strength of the lunar
338 soil (1 kPa) to the strength of the terrestrial soil (< 100 kPa) (*Mitchell et al.,*
339 *1972; Lambe and Whitman, 1979*). We have tested the influence of these two
340 parameters using lower values, $Y_i = 10$ kPa and $Y_s = 100$ kPa. As illustrated
341 in Fig. 9 (second column) (called "highly deformable"), decreasing the min-
342 imum strength of both the ice and the rocky materials tends to increase the

343 deformability of the rock core leading to both a deeper and wider depression.
344 Ultimately, for a 200 km radius rocky core with 50% porosity (Fig. 9, second
345 columns), the depth of the depression can reach 54.5 km. Here again, the con-
346 ditions in term of porosity and strength are rather extreme, and the objectives
347 of this simulation are to illustrate the maximal depression depth that could be
348 generated by a large impact on Enceladus.

349

350 Fig. 8 (third and fourth columns) and 9 (third column) shows that the pres-
351 ence of a deep water ocean (considered as an inviscid fluid with a density of 910
352 kg/m³) above the rocky core tends to reduce the impact-induced deflection of
353 the core surface. Liquid water and water ice have comparable compressibility,
354 water being slightly more compressible. The main difference concerns their re-
355 sistance to shear. Liquid water has no strength (and is considered a completely
356 inviscid material in the simulation), while ice has some strength. In the presence
357 of liquid water, there is complete mechanical decoupling of shear deformation
358 between the water and the core, whereas in the latter case shear stresses exist at
359 the ice-core boundary. In the presence of the water ocean, the lateral extent of
360 the morphology anomaly as well as its depth are smaller than without an ocean.
361 Indeed, for $R_{core} = 160$ km, the depth of the impact induced cavity decreases
362 from 13 km without an ocean to 3.5 km with an ocean. For $R_{core} = 200$ km
363 and $\phi = 50\%$, the depth of the impact induced cavity decreases from 31.5 km
364 without an ocean to 22.5 km with an ocean. This tends to illustrate that it is
365 easier to enhance post-impact negative topography anomalies in the absence of
366 a water ocean. Including a thick subsurface water ocean has the opposite effect
367 of increasing the impact velocity or the impactor size, because it concentrates
368 deformation in the ice mantle above, decoupling it from the rocky core below.
369 On the other hand, the presence of the ocean seems to enhance the plastic strain

370 in the deepest part of the core (Fig. 8, third and fourth columns). In parallel to
371 compaction, impact-induced fracturing is likely to generate a porosity increase
372 (via the dilatancy process) (*Collins, 2014*) that could in return favour fluid cir-
373 culation within the deformed rocky core.

374

375 4. Discussion and Conclusion

376 In order to investigate the morphological consequences of collisions between
377 differentiated impactors and Enceladus, we performed numerical impact simula-
378 tions for impactor radii and velocities ranging between 10% to 40% Enceladus'
379 radius and 1 to 10 times Enceladus' escape velocity (0.24 to 2.4 km/s), and for
380 various assumptions for the structure and mechanical properties of Enceladus'
381 interior. Our results showed that the dynamical response of the icy mantle to
382 the impact is strongly dependent on the assumed thermo-mechanical properties
383 for the ice. However, the icy mantle response has minor effects on the impact-
384 induced deformation of the rock core. Only the presence of an internal ocean
385 between the icy mantle and the rock core can significantly limit the rock core
386 deformation.

387

388 Our simulations showed that the main controlling parameters for the post-
389 impact shape of Enceladus' rock core are the impactor radius and velocity. We
390 have identified three regimes: (1) For low energy impacts ($\leq 1.5 - 2 \times 10^{23}$ J),
391 the impactors do not pass completely through the icy mantle and the core sur-
392 face remains unmodified. The rock core of the impactors are deformed by the
393 impact events, but remains trapped within the icy mantle. The impactor core
394 embedded in the icy mantle would then progressively sink and spread, leading
395 to a positive core topographic anomaly. (2) For more energetic impacts, the

396 impactors completely penetrate though the icy mantle and hit the core sur-
397 face. The impact leads to a negative core topography surrounded by a positive
398 anomaly of smaller amplitude. The depth and lateral extent of the excavated
399 area is mostly determined by the impactor radius and velocity. The shape of
400 the excavated area can be significantly enhanced for high core porosity and very
401 low material strengths, but its amplitude and extent remain primarily deter-
402 mined by the impactor parameters. In this regime, accounting for the acoustic
403 fluidization does not change the final core morphology (not shown here). (3)
404 For even more energetic impacts, the core is very strongly deformed, which does
405 not appear to be compatible with Enceladus' core morphology (see Fig. 2).
406 Our simulations of these events do not follow the full evolution of the impact
407 scenario so we cannot predict the final core structure; however, it is likely that
408 some of these events lead to full body disruption and that, in non-disruptive
409 impacts, acoustic fluidization may contribute to the final shape of the rocky
410 core and would therefore need to be included to analyze possible outcomes.

411

412 For impact velocities higher than $2.4 \text{ km.s}^{-1}(10 \times v_{esc})$, moderate deforma-
413 tion of the core is possible only for impactors smaller than 25 km. During the
414 Late Heavy Bombardment, high-velocity collisions with impactors exceeding 20
415 km is likely and therefore, as recently highlighted by *Movshovitz et al.* (2015),
416 full disruption and re-accretion of the satellite may have occurred possibly sev-
417 eral times during this period. This implies that any large impact leaving a
418 long-wavelength signature on the core shape should have taken place after the
419 Late Heavy Bombardement. This also requires relatively low velocity impacts,
420 and therefore encounter with planetocentric bodies rather than with heliocen-
421 tric bodies. Alternatively, as proposed by *Charnoz et al.* (2011), Enceladus may
422 have formed late during the history of the Saturn system, thus limiting the risk

423 of full disruption. Following the model of *Charnoz et al.* (2011), Enceladus may
424 have accreted from a swarm of differentiated embryos emerging from the outer
425 edge of a massive ring system. In such a model, multiple low velocity colli-
426 sions between decametric differentiated impactors and a growing Enceladus are
427 expected. The irregular core shape of Enceladus, as constrained from Cassini
428 gravity and topography data (*McKinnon, 2013; Lefèvre et al., 2015*), may con-
429 stitute a record of this accretional process.

430

431 Various processes will probably alter the core topography after an impact
432 event, so that the amplitude of core deflection predicted in our simulations
433 should be considered as an upper limit. Rock fragments would be likely trans-
434 ported by the ice flow back to the impact cavity, filling partly the impact-induced
435 depression. Even if the core is relatively cold, topography relaxation may occur
436 to some extent, especially for low-strength rock material. Prolonged water inter-
437 actions may also partly erode the topography and again reduce the topography
438 anomaly. Detailed modelling of the subsequent topography evolution is beyond
439 the scope of the present study, and will require future modeling effort. The
440 2D nature of our simulations also optimizes the amplitude of impact-induced
441 core deflection as only head-on collisions can be considered. It is known that
442 impact angle affects the strength and distribution of the shock wave generated
443 in the impact and therefore the perturbed region (e.g. *Pierazzo and Melosh,*
444 *2000*). For more oblique impacts, the impactor kinetic energy will be more ef-
445 ficiently transferred to the icy mantle, leading to a more efficient deformation
446 of the icy mantle and a larger amount of escaping materials (e.g. *Korycansky*
447 *and Zahnle, 2011*). The volume of icy mantle affected by the impact, which
448 is already large for head-on collisions as shown with our 2D simulations, will
449 be further increased. Another limitation of our modelling approach is the as-

450 assumption regarding the mechanical properties of the rock core. We considered
451 dunite with various degree of void porosity as representative of the rock core
452 composition, since a relatively well-defined equation of state exists for this ma-
453 terial (*Davison et al.*, 2010). Based on the interpretation of the Cassini gravity
454 data, which suggest a low density core (2400 kg.m^{-3} , *Iess et al.* (2014)), the
455 rock core may contain a significant fraction of highly hydrated minerals, as well
456 as free water or/and ice in rock pores. Currently, we are not able to consider
457 a mixture of ice and rocks for both the impactor’s core and the target’s core.
458 However, to estimate an upper limit of the deformation, we have performed a
459 run corresponding to our classical impact model ($v_{imp} = 10v_{esc}$ and $R_{imp} = 25$
460 km) with 100% ice-filled pores (i.e. a core made of pure ice). In this unrealis-
461 tic case (not shown here), the impactor’s core is eventually buried at a depth
462 of ~ 170 km (i.e. 80 km below the core-mantle boundary) which is far larger
463 than the depth of the depression (~ 30 km) obtained with a 50% porous rocky
464 core. This limitation also stands for the structure of the impactor’s core that
465 is likely to have remained undifferentiated in the context of an early formation.
466 To estimate the influence of the impactor’s degree of differentiation, we have
467 also considered the $v_{imp} = 10v_{esc}$ case with a 25 km radius impactor made of
468 pure ice and an impactor made of pure dunite. In the first case, the impact
469 induces a flattening of ~ 0.4 km at the core’s surface below the impact site
470 (see Fig. 6). In the second case, the impact induces a flattening of ~ 23.2 km.
471 This result, even if performed for an unrealistic water ice content, suggests the
472 ice/rock ratio in the core may play a strong influence on the response of the
473 core to large impacts. This suggests that the results presented here should be
474 considered valid only for differentiated interior models with rock-dominated core
475 and a relatively small porosity content ($<10\text{-}20\%$). Future works are required
476 to constrain more precisely the effect of hydrated minerals and mixture with high

477 ice-water/rock ratio in the interior.

478

479 Large impacts are likely to modify the ice/rock ratio by eroding significantly
480 the shallower part of the impacted moon. Our results show that vertical im-
481 pacts with $v_{imp} > 6v_{esc}$ and $R_{imp} > 75$ km, can erode up to half the ice volume
482 from the impacted body (Fig. 1, second column). Several factors such as a hot,
483 porous pre-impact mantle and the presence of a deep water ocean increase the
484 ability of the icy mantle to deform. Hence, these parameters are also likely to
485 influence the post-impact ice/rock ratio by decreasing the fraction of ice in the
486 post-impact moon. The impact angle is another key parameter that governs
487 the fraction of escaped material (e.g. *Korycansky and Zahnle, 2011*). However,
488 to limit the computational time and as we have restricted our study to vertical
489 impacts, monitoring the long-term evolution of the ice/rock ratio is beyond the
490 scope of our study.

491

492 Despite the limitations, the simulations we performed highlight the crucial
493 role played by impacts on the evolution of Enceladus. Besides explaining the
494 irregular shape of the core, impacts also provide efficient mechanisms to en-
495 hance thermo-chemical exchanges between the deep interior and the surface.
496 For models with an internal water ocean, we can see that a large volume of the
497 ocean is temporarily exposed to the surface, thus potentially releasing a large
498 fraction of volatile initially stored dissolved in the ocean. Large impacts cause
499 a strong damage of the ice on a very large portion of the icy mantle, which
500 will likely have consequences on the subsequent convective mantle dynamics
501 and interaction with the fractured surface. These also lead to a large plastic
502 strain in the rock core underneath the impact site, which may enhance macro-
503 porosity. This would promote fluid circulation throughout a large fraction of

504 the core, favoring serpentinization processes (*Malamud and Prialnik, 2013*) and
505 hydrothermal activities (e.g. *Hsu et al., 2015*). Further modeling efforts will be
506 needed to understand the consequences of such impact events on the long-term
507 evolution of Enceladus. Lastly, the effects of large impacts are not confined to
508 Enceladus. Similar effects are very likely on the other moons of Saturn as well
509 as on other planetary objects, such as Ceres (*Davison et al., 2015; Ivanov, 2015,*
510 e.g.) and Pluto (*Bray and Schenk, 2015, e.g.*) for which impact bombardment
511 has probably played a key role in their evolution.

512

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Table 1: Typical parameter values for numerical models

Enceladus radius	R	250 km
Rocky core radius	R_{core}	160-200 km
Icy mantle thickness	δ_m	50-90 km
Surface gravity field	g_0	0.113 m.s ⁻²
Escape velocity	v_{esc}	240 m/s
Impactor radius	R_{imp}	25-100 km
Impact velocity	v_{imp}	240-2400 m/s
<i>Mantle properties (Ice)</i>		
Initial density	ρ_i	820 kg.m ⁻³
Equation of state type		Tillotson
Poisson		0.33
Porosity		0-20%
Minimum strength	Y_i	10-500 kPa
<i>Core properties (Dunite)</i>		
Rocky core density	ρ_s	3330 kg.m ⁻³
Equation of state type		ANEOS
Poisson		0.25
Porosity		0-50%
Minimum strength	Y_s	100 kPa-10 MPa

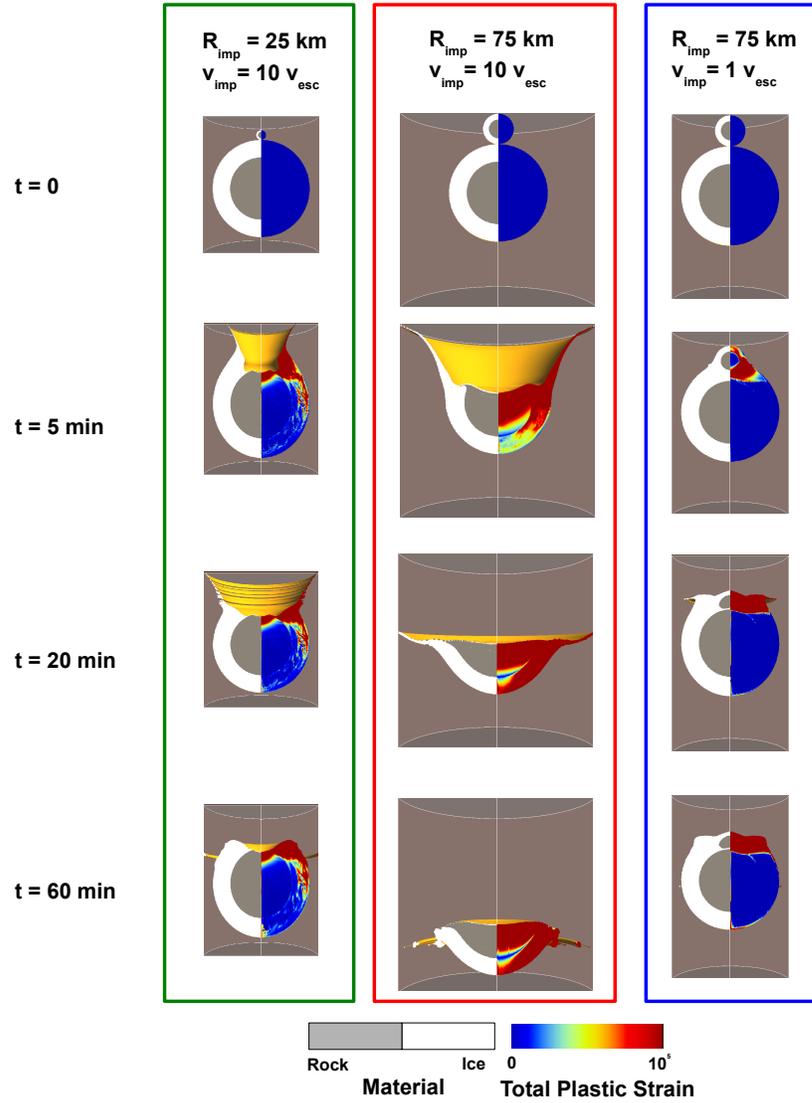


Figure 1: Material repartition (left column) and total plastic deformation (right column) as a function of time (from top to bottom) on Enceladus for 3 impact cases: ($v_{imp} = 10v_{esc}$, $R_{imp} = 25 \text{ km}$) (left), ($v_{imp} = 10v_{esc}$, $R_{imp} = 75 \text{ km}$) (centre) and ($v_{imp} = v_{esc}$, $R_{imp} = 75 \text{ km}$) (right). In these models, the grid resolution is 1 km in all directions. Here both the rocky core and the icy material are considered as nonporous materials.

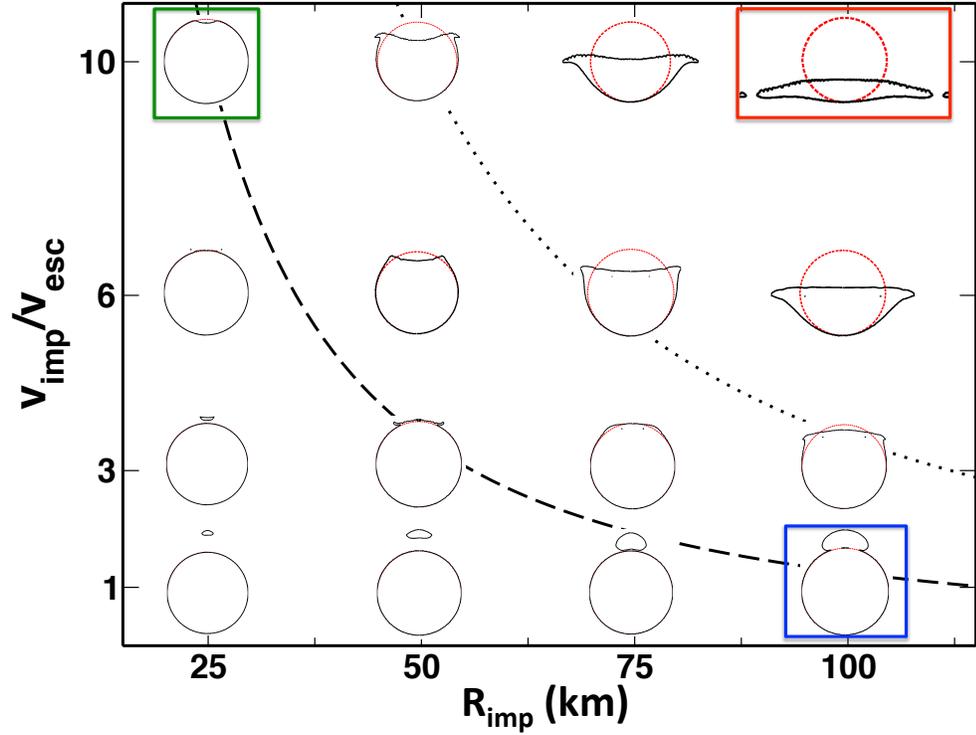


Figure 2: Rocky core morphology as a function of the impactor size and the impact velocity ($v_{esc} = 240$ m/s). In these models the porosity of the icy material is zero. For each morphology, the red circle represents the pre-impact spherical shape of the impacted core. The dashed black line represents Eq.2 with $A = 2$. Above this critical theoretical line, the impact induced topography is negative. Below this critical theoretical line, the impact induced topography is positive. The dotted black line represents Eq.2 with $A = 1$. Above this critical theoretical line, very highly deformed cores are formed and acoustic fluidization may contribute to their final shape. However the deformation is too large and probably not compatible with the Enceladus morphology. We limit our post impact monitoring to one hour which means that for large impact velocities (≥ 6 km/s) and large impactor radii (≥ 75 km) the rocky material excavated from Enceladus' core and orbiting around the moon is still moving with significant velocity at the end of the simulation.

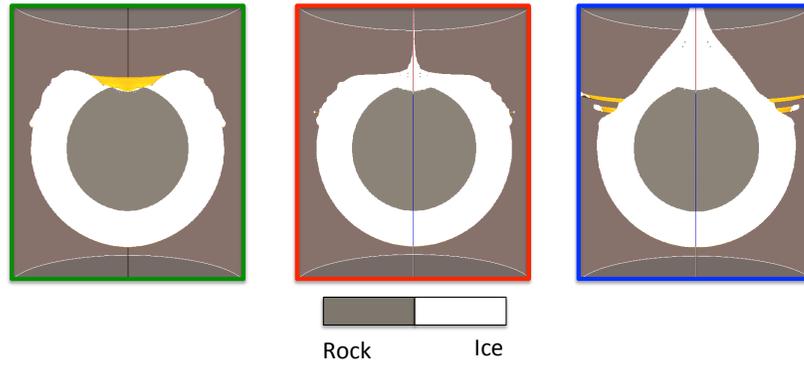
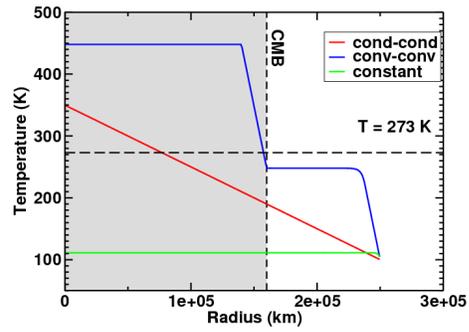


Figure 3: Material repartition one hour after the impact (bottom) for three different pre-impact temperature profiles (top) (with $v_{imp} = 10v_{esc}$, $R_{core} = 160$ km and $R_{imp} = 25$ km). The color of the temperature profile corresponds to the color of the rectangle surrounding the material repartition snapshot.

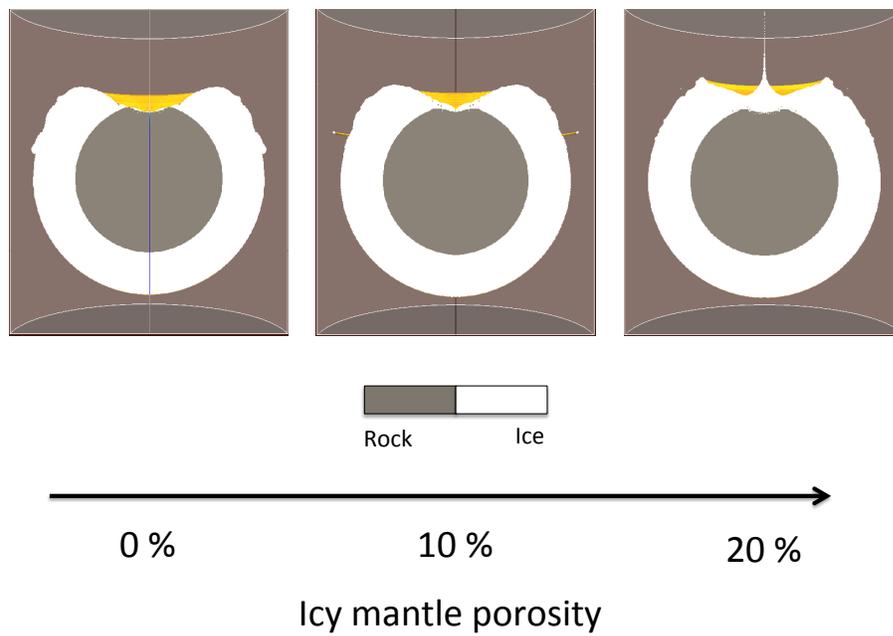


Figure 4: Material repartition as a function of the icy mantle porosity one hour after the impact ($v_{imp} = 10v_{esc}$, $R_{imp} = 25$ km). The rocky core is represented in grey while the icy material is represented in white. In these models, the grid resolution is 1 km in all directions.

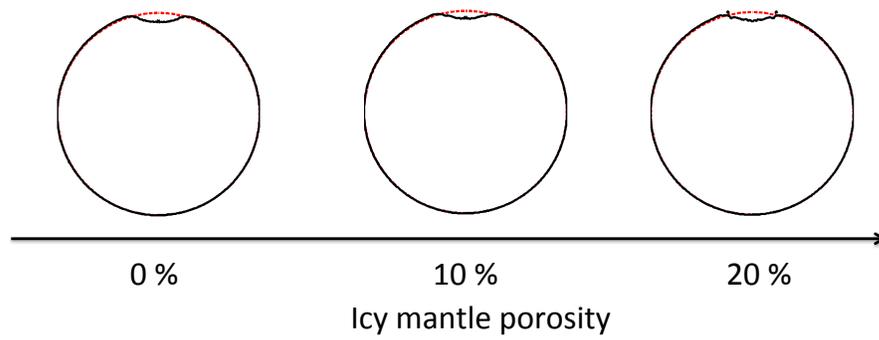


Figure 5: Rocky core morphology as a function of the icy mantle porosity (with $R_{core} = 160$ km). For each morphology, the red circle represents the pre-impact spherical shape of the impacted core.

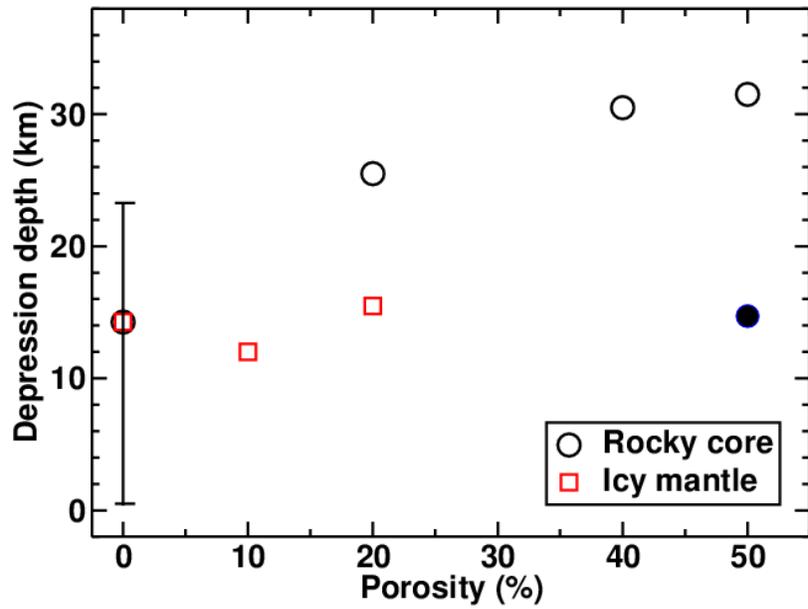


Figure 6: Depth of the impact induced depression as a function of the rocky core porosity (black circles) and as a function of the icy mantle porosity (red squares) ($v_{imp} = 10v_{esc}$ and $R_{imp} = 25$ km). The vertical line for 0% porosity represents the range of depression depths obtained when considering a 100% icy (lower value) and a 100% rocky (upper value) impactor. The black filled circle at 50% porosity represents the unrealistic case with a core radius of 160 km (while in the other cases the core radius increases with porosity).

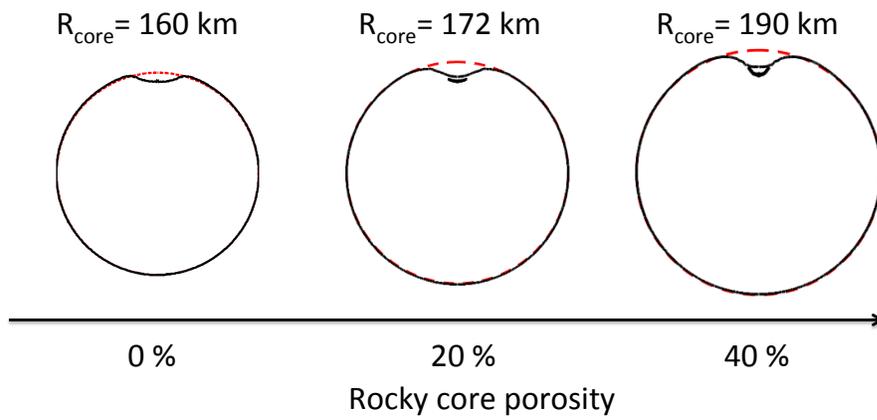


Figure 7: Rocky core morphology as a function of the rocky core porosity. For each morphology, the red circle represents the pre-impact spherical shape of the impacted core.

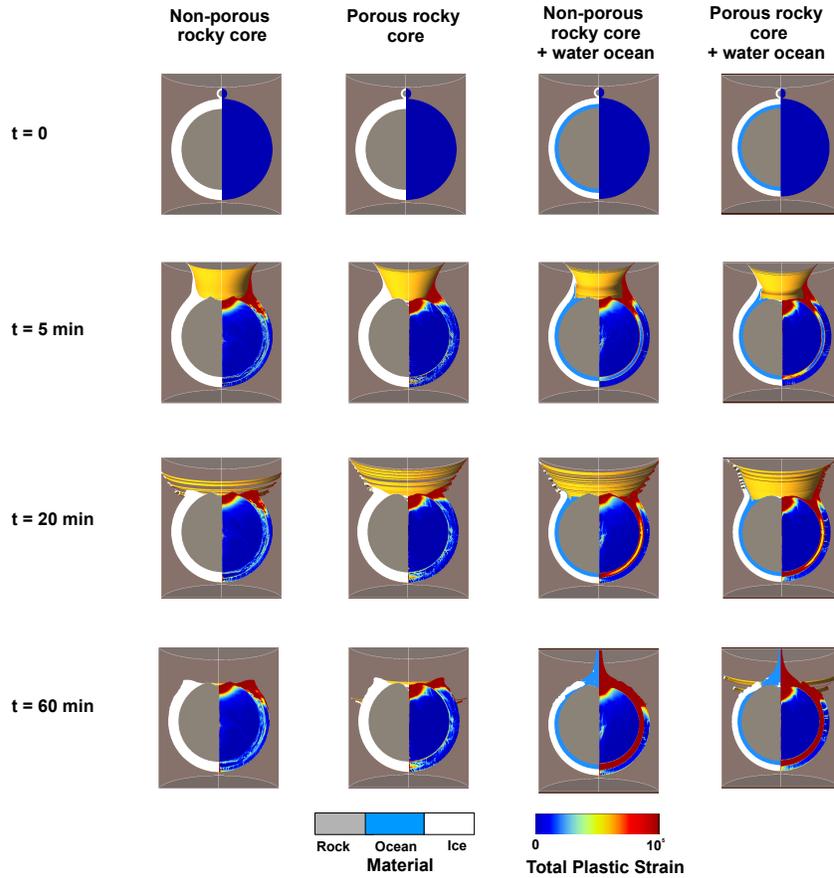


Figure 8: Material repartition (left column) and total plastic deformation (right column) as a function of time (from top to bottom) on Enceladus for $R_{core} = 200 \text{ km}$, ($v_{imp} = 10v_{esc}$ and $R_{imp} = 25 \text{ km}$). We consider 4 models: a non-consistent non-porous rocky core (first column), a porous rocky core with a porosity of 50 % (second column), a non-consistent non-porous rocky core overlaid by a 20 km thick water ocean (third column) and a porous rocky core with a porosity of 50 % overlaid by a 20 km thick water ocean (fourth column). In these models, the grid resolution is 1 km in all directions.

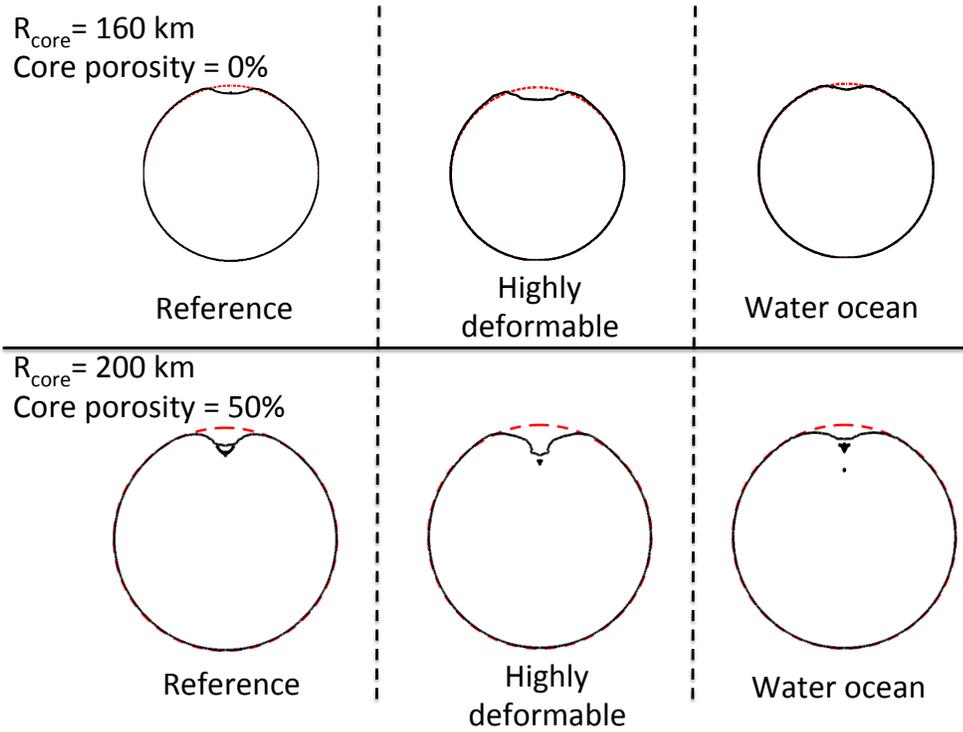


Figure 9: Rocky core morphology for different pre-impact core radii ($R_{\text{core}} = 160 \text{ km}$ (top) and 200 km (bottom)). First and third columns: $Y_i = 500 \text{ kPa}$ and $Y_s = 10 \text{ MPa}$, second column ("highly deformable") $Y_i = 10 \text{ kPa}$ and $Y_s = 100 \text{ kPa}$. In the third column we consider a water ocean (with a thickness of 20 km) above the rocky core. For each morphology, the red circle represents the pre-impact spherical shape of the impacted core.