

1 **Macroporosity and Grain Density of Rubble Pile**
2 **Asteroid (162173) Ryugu**

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16 **Key Points:**

- 17 • Ryugu's large bulk porosity is distributed between intrinsic boulder microporos-
- 18 ity and macroporosity due to void spaces in-between boulders.
- 19 • We use the boulder size-frequency distribution as observed on the surface together
- 20 with mixing models to estimate Ryugu's macroporosity.
- 21 • We find that macroporosity is 16 ± 3 %, indicating that Ryugu's large bulk poros-
- 22 ity of close to 50 % is governed by microporosity.

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23 **Abstract**

24 Rubble pile asteroids such as (162173) Ryugu have large bulk porosities, which are be-
 25 lieved to result from void spaces in between the constituent boulders (macroporosity)
 26 as well as void spaces within the boulders themselves (microporosity). In general, both
 27 macroporosity and microporosity are estimated based on comparisons between the as-
 28 teroid bulk density and both the bulk and grain density of meteorite analogues, and rel-
 29 atively large macroporosities are usually obtained. Here we use semi-empirical models
 30 for the macroporosity of multi-component mixtures to determine Ryugu’s macroporos-
 31 ity based on the observed size-frequency distribution of boulders on the surface. We find
 32 that Ryugu’s macroporosity can be significantly smaller than usually assumed, as the
 33 observed size-frequency distribution allows for an efficient packing of boulders, result-
 34 ing in a macroporosity of 16 ± 3 %. Therefore, we confirm that Ryugu’s high bulk poros-
 35 ity is a direct consequence of a very large boulder microporosity. Furthermore, using es-
 36 timates of boulder microporosity of around 50 % as derived from in-situ measurements,
 37 the average grain density in boulders is 2848 ± 152 kg m⁻³, similar to values obtained
 38 for CM and the Tagish lake meteorites. Ryugu’s bulk porosity corresponding to the above
 39 values is 58 %. Thus, the macroporosity of rubble pile asteroids may have been sys-
 40 tematically overestimated in the past.

41 **Plain Language Summary**

42 The carbonaceous asteroid (162173) Ryugu formed from fragments which re-accreted
 43 after its parent body was disrupted by a catastrophic collision. Asteroids of this type
 44 are also known as rubble piles and the re-accumulation process is thought to be one of
 45 the causes for their large bulk porosity. We have applied mixing models to determine the
 46 amount of inter-boulder porosity taking the observed abundance of large and small boul-
 47 ders on the surface into account. We find that the relative abundances of differently sized
 48 boulders allow for a very efficient packing, such that inter-boulder porosity in Ryugu is
 49 rather small and only 16 ± 3 %. This implies that a large part of Ryugu’s total poros-
 50 ity must reside inside the boulders themselves. Using estimates of boulder intrinsic poros-
 51 ity, we furthermore constrain the average density of the boulder’s constituent minerals
 52 to 2848 ± 152 kg m⁻³, which is consistent with values measured for carbonaceous me-
 53 teorites as collected on Earth. Thus, inter-boulder porosity of rubble pile asteroids may
 54 have been systematically overestimated in the past.

55 1 Introduction

56 Upon arrival of the Hayabusa2 spacecraft the C-complex asteroid (162173) Ryugu
 57 was found to be a spinning top-shaped rubble pile (Watanabe et al., 2019) with Cb-type
 58 spectrum and very low albedo around 0.045, consistent with thermally metamorphosed
 59 CM/CI meteorites (Sugita et al., 2019). Observations further show that a weak 2.7 μm -
 60 absorption is present, suggesting a small amount of hydrated minerals exist on the sur-
 61 face (Kitazato et al., 2019). Furthermore, the surface was found to be dominated by blocks
 62 and boulders (Sugita et al., 2019; Michikami et al., 2019), and 50% of the surface is cov-
 63 ered by boulders with diameters exceeding 0.5 m. A bulk density of $1190 \pm 20 \text{ kg m}^{-3}$
 64 was determined using the SFM20180804 shape model (Watanabe et al., 2019), which al-
 65 lowed for an estimate of asteroid porosity. Assuming typical grain densities for carbona-
 66 ceous chondrites (Britt & Consolmagno S.J., 2001; Macke et al., 2011; Flynn et al., 2018),
 67 bulk porosity estimates close to 50 % were obtained (Watanabe et al., 2019), which is
 68 consistent with the bulk porosity estimates for C-complex asteroids.

69 The bulk porosity inside rubble pile asteroids can be separated into two contribu-
 70 tions: the first one stems from the intrinsic porosity of rocks and boulders and is termed
 71 microporosity, while the second contribution refers to voids in-between particles and is
 72 termed macroporosity (Britt et al., 2002). The latter is directly related to the geomet-
 73 rical arrangement of the constituent blocks, also known as the packing state, which qual-
 74 itatively describes the arrangement of particles and can vary between random loose and
 75 random close packings. Macroporosity of average C-complex asteroids was estimated to
 76 be 25-30 % (Britt et al., 2002), which is generally consistent with numerical models of
 77 the reassembly of blocks after a catastrophic disruption, which result in macroporosi-
 78 ties of 20-40 % (Wilson et al., 1999). However, simulations suffer from unrealistically large
 79 lower cutoff sizes for the considered boulder population, such that rubble pile asteroids
 80 may still exhibit lower macroporosities.

81 Here we investigate the macroporosity of asteroid Ryugu using semi-empirical mod-
 82 els for the porosity of multi-component mixtures of non-spherical, cohesive particles (Zou
 83 et al., 2011). Such models predict the macroporosity of granular material given the par-
 84 ticle size as well as the particle shape distributions applying linear mixing and using the
 85 concept of controlling mixtures (Yu & Standish, 1991) to calculate the packing state. In
 86 general, polydisperse particle mixtures can have a macroporosity which is considerably

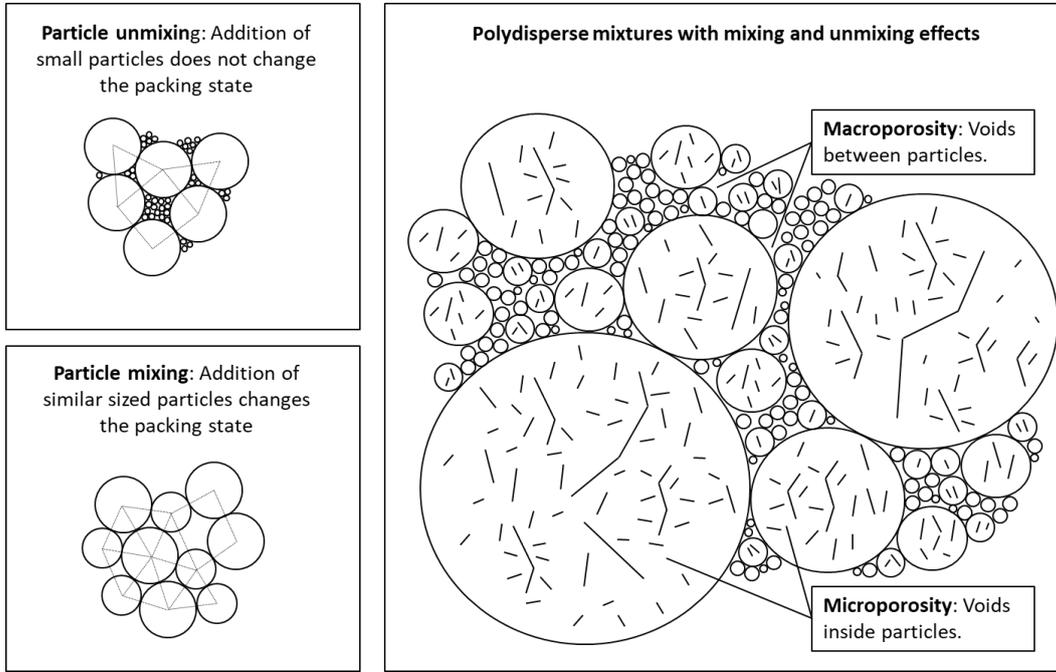


Figure 1. Top Left: Illustration of particle unmixing for particles with strongly disparate diameters. As small particles are added to a system of larger particles, the larger particles resist being displaced and the packing state does not change. A similar effect occurs for the addition of very large particles. Bottom Left: Illustration of particle mixing for particles with similar diameters. As similar sized particles are added to a system, particles can be displaced thus changing the packing state. Right: Two dimensional illustration of the random packing structure of strongly polydisperse spheres. As compared to monodispersed configurations, porosity is reduced by the filling of void spaces. Macroporosity refers to the porosity generated by the void spaces between particles, while microporosity is caused by void spaces and cracks that formed inside individual particles. Figure adapted from Yu & Zou (1998).

87 smaller than the canonical $\sim 36\%$ for a random close packing or $\sim 42\%$ for a random
88 loose packing of spherical, monosized particles (Scott, 1960), and values down to 10%
89 can be reached (Dullien, 1991).

90 In binary mixtures, the way particles interact depends on their size ratio, here de-
91 fined as the ratio of the respective particles' volume-equivalent diameters. If this ratio
92 is less than 0.154 (Graton & Fraser, 1935), small particles will not affect the packing state
93 and simply fill the gaps between larger ones. In contrast to this unmixing of particles,
94 similar sized particles will mix, creating a new packing structure (also see Yu & Zou (1998)
95 for a discussion of mixing and unmixing effects). Applying these concepts to polydisperse
96 mixtures, particle unmixing will take place for very small and very large particles, as smaller
97 particles start filling the gaps and larger particles completely fill some regions with solid
98 material. The component controlling the porosity of the mixture is then defined by in-
99 termediate sized particles, which do not change their packing state by the addition of
100 unmixing components (Yu & Zou, 1998). An illustration of particle mixing and unmix-
101 ing is shown in Fig. 1. The semi-empirical models by Yu & Zou (1998) and Zou et al.
102 (2011) can be applied to particle mixtures in loose and dense packing states. They have
103 been shown to reproduce the porosity of mixtures created using the funnel method, in
104 which particles are gently poured into a container, as well as the porosity of mixtures
105 tapped many times to reach maximum compaction.

106 It is important to note that packing is determined by the interplay of the differ-
107 ent grain sizes present, and it can be misleading to consider individual grain sizes only.
108 For example, while the addition of a single large block to the mixture can reduce poros-
109 ity by displacing smaller particles and filling void spaces, the addition of many large blocks
110 can increase porosity by creating large voids. Similarly, addition of some small particles
111 may reduce porosity, while many small particles can create a large number of small voids,
112 again increasing porosity. Therefore, the porosity finally attained by the mixture depends
113 on the details of the size-frequency distribution of the particles present.

114 In order to apply the theory of multi-component mixtures, the size and shape dis-
115 tributions of boulders need to be known. Here we use the boulder size and shape dis-
116 tributions determined by Michikami et al. (2019), who extend the analysis in Sugita et
117 al. (2019) using images from the Hayabusa2 optical navigation camera (ONC) (Kameda
118 et al., 2017; Suzuki et al., 2018; Tatsumi et al., 2019) which have near global coverage

119 and were acquired at altitudes between 20 km and 6.5 km. These have spatial resolu-
 120 tions down to 0.65 m/pixel, and global counts were performed for boulders with diam-
 121 eters > 2 m and a completeness limit of 5 m. In addition, smaller boulders, cobbles and
 122 pebbles with sizes of 0.02 to 9.1 m were studied using close-up images of the sampling
 123 areas, where images taken at altitudes from 67 m to 620 m with resolutions down to $<$
 124 0.01 m/pixel are available (Michikami et al., 2019). Overall, size-frequency and shape
 125 distributions were determined in the 0.02 to 140 m size range.

126 By applying the multi-component mixing model to the size distribution of boul-
 127 ders as observed on the surface, we assume that the same distribution holds in the in-
 128 terior. This assumption is supported by laboratory experiments on the disruption of mono-
 129 liths (Michikami et al., 2016), which suggest that boulders on bodies such as Itokawa,
 130 Bennu, and Ryugu are relicts of the direct formation of those asteroids by gravitational
 131 reaccumulation following the disruption of their parent bodies (Michel & Richardson,
 132 2013; Michel et al., 2020) rather than the result of impact events after formation has been
 133 completed. Impacts could reshape the size distribution by the production of smaller par-
 134 ticles after reaccretion has been completed, but the importance of this process may be
 135 limited. This is due to the so-called armoring effect (Sugita et al., 2019), by which a large
 136 fraction of the impact energy is lost when the projectile contacts the first large boulder,
 137 thus producing only few fragments. Another mechanism that could be responsible for
 138 a difference between the size-frequency distributions observed on the surface and present
 139 in the interior is seismic shaking, and the Brazil Nut Effect could lead to an overrepre-
 140 sentation of large boulders on the surface (Tancredi et al., 2015; Maurel et al., 2017). How-
 141 ever, the seismic efficiency of impacts in granular material appears to be low (Yasui et
 142 al., 2019; Nishiyama et al., 2020), such that surface modifications are likely localized. Nev-
 143 ertheless, seismic shaking could have an impact on the global boulder size-frequency dis-
 144 tribution over geological timescales. Finally, it has been argued that particle size sort-
 145 ing may take place during rubble pile reaccretion, with larger blocks accreting first and
 146 thus in the center (Britt & Consolmagno S.J., 2001). These caveats need to be kept in
 147 mind when interpreting the results presented below.

148 A second important input parameter for the multi-component mixing model is the
 149 material’s packing state, which can vary between a random loose and random close pack-
 150 ing. In general, little is known about the packing state of rubble pile asteroids follow-
 151 ing reaccretion, which depends on many parameters such as the distribution of angu-

lar momentum in the reaccreting system as well as the size distribution and shape of reac-
 creting fragments. While impact experiments indicate that shattered, elongated parti-
 cles with large deviations from a spherical shape can be produced (Nakamura & Fuji-
 wara, 1991; Durda et al., 2015; Michikami et al., 2016) the results of disruption exper-
 iments need to be interpreted with caution, as the high strain rates imposed during the
 experiment may not be representative for the destruction of larger blocks. Further, long
 term seismic shaking could lead to the reduction of pore spaces. Given these unknowns,
 we will systematically vary the packing state in the analysis below.

In the following, we will first introduce the theory of determining asteroid macro-
 porosity from observed size and shape distributions for the rocks and boulders. We will
 then derive a simple equation relating grain density to macro- and microporosity. Re-
 sults of the macroporosity calculation and relevant uncertainties will then be used to es-
 timate grain density of Ryugu’s constituent material given estimates of boulder micro-
 porosity (Grott et al., 2019; Hamm et al., 2020; Okada et al., 2020). Finally, results, as-
 sumptions, and implications will be discussed.

2 Methods

2.1 Particle Size and Shape Distributions

To estimate Ryugu’s macroporosity, the constituent boulder’s size and shape dis-
 tributions need to be known. These were determined by Michikami et al. (2019) who fit-
 ted size-frequency data using power laws. Power law exponents between 1.65 and 2.65
 were obtained, with 2.65 being the best fit for the global dataset. Furthermore, parti-
 cles were generally found to be elongated, and axis ratios for boulders > 2 m are close
 to 0.7 on average. The size-frequency distribution of boulders on small bodies may bet-
 ter be described by a Weibull distribution than a power law (Schröder et al., 2020), and
 we have used a cumulative Weibull (Rosin-Rammler) distribution (Rosin, 1933; Weibull,
 1951; Wingo, 1989; Brown & Wohletz, 1995) to represent the data provided by Michikami
 et al. (2019). The cumulative size-frequency distribution $N(D)$ is then given by

$$N(D) = N_T e^{-3(D/\lambda)^\beta/\beta} \quad (1)$$

where D is the mean horizontal diameter, and we determined the fit parameters $\beta =$
 0.09495 , $\lambda = 33.78$ m, and $N_T = 5.28 \cdot 10^{14}$ km $^{-2}$ by a weighted least-squares ap-
 proach as a practical means to obtain a good representation of the data. The resulting

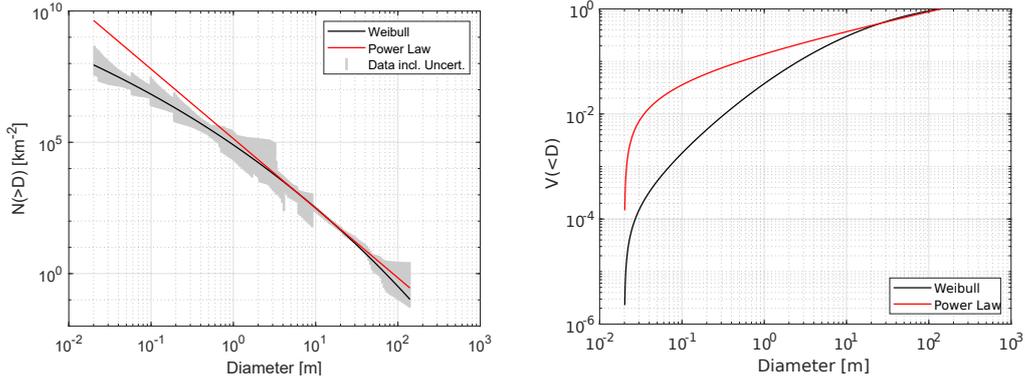


Figure 2. Left: Cumulative particle size frequency distribution (SFD) as derived for Ryugu by Michikami et al. (2019). Our Weibull fit to the data (black) is shown along with a power law fit with exponent $p = 2.65$ (red). Right: Cumulative volume fraction distribution for the SFDs on the left hand side of the figure.

182 distribution $N(D)$ is shown together with the uncertainty of the data in Fig. 2, where
 183 uncertainty comprises the Poisson uncertainty as well as the uncertainty of particle di-
 184 ameters introduced by the limited image resolution. It is worth noting that represent-
 185 ing the data using a single power law for the entire size range does not adequately rep-
 186 resent the data.

187 Given the size-frequency distribution $N(D)$ as determined from surface counts of
 188 boulders, the normalized cumulative volume distribution $V(D)$ can be calculated by nu-
 189 merical integration. It is given by

$$V(D) = c \left(N_{\text{tot}} D_{\text{min}}^3 - N(D) D^3 + \int_{D_{\text{min}}}^D N(D') 3D'^2 dD' \right) \quad (2)$$

190 where N_{tot} is the total number of particles counted per unit area, D_{min} and D_{max} are
 191 the minimum and maximum particle sizes of the particle size distribution $N(D)$, respec-
 192 tively, and c [m^{-1}] is a normalization factor chosen such that $V(D_{\text{max}}) = 1$.

193 In addition to the Weibull distribution fit to the data represented by Eq. 1, we will
 194 also consider a simple power law to systematically study the influence of the particle size
 195 distribution's power law exponent p on the obtained results. The distribution can then
 196 be expressed as

$$N(D) = N_{\text{tot}} (D/D_{\text{min}})^{-p} \quad (3)$$

197 where $p = 2.65$ represents the best fit to the global dataset (Michikami et al., 2019).
 198 For the power law defined by Eq. 3, Eq. 2 can be integrated analytically and the vol-
 199 ume size distribution is then given by

$$V(D) = \frac{D^{3-p} - D_{\min}^{3-p}}{D_{\max}^{3-p} - D_{\min}^{3-p}} \quad (4)$$

200 for $D_{\min} \leq D \leq D_{\max}$. For $D \geq D_{\max}$, $V(D) = 1$, whereas for $D \leq D_{\min}$ $V(D) = 0$.

201 Michikami et al. (2019) give the shape of boulders in terms of the maximum dimen-
 202 sions in three mutually orthogonal planes ($a \geq b \geq c$). Here we primarily regard the
 203 horizontal axis ratio b/a , with a being the maximum and b the intermediate dimension.
 204 As reported by Michikami et al. (2019), shape of particles on Ryugu appears to be largely
 205 independent of geographical longitude, whereas some dependence on latitude may in-
 206 dicate boulder migration. Nevertheless, average b/a is only weakly size-dependent and
 207 close to 0.7.

208 In general, particle sphericity is defined as the ratio of the surface area of a sphere
 209 (with the same volume as the particle) to the surface area of the particle (Wadell, 1932).
 210 However, this is difficult to evaluate in practice, and the Krumbein (Krumbein, 1941)
 211 or Riley (Riley, 1941) simplifications are usually applied. Working with two dimensional
 212 (image) data, we define sphericity Ψ as

$$\Psi = \sqrt{\frac{D_i}{D_c}} \quad (5)$$

213 where D_i is the diameter of the largest inscribed circle and D_c is the diameter of the small-
 214 est circumscribing circle for a given particle (Riley, 1941). Using Eq. 5, the shape pa-
 215 rameter b/a derived by Michikami et al. (2019) then translates into an average spheric-
 216 ity of $\Psi = 0.83$. In addition, Michikami et al. (2019) also estimated the third axis, c/a ,
 217 of 121 arbitrarily selected boulders. The mean axes ratio c/a was found to be 0.44, and
 218 the sphericity of a parallelepiped with axis ratios a:b:c of 1:0.71:0.44 is 0.796. On the other
 219 hand, sphericity of a triaxial ellipsoid with the same axis ratios is 0.913. Therefore, spheric-
 220 ity depends not only on axis ratios, but also on particle shape, and we will use $\Psi = 0.85 \pm$
 221 0.06 as an average sphericity rather than the average sphericity derived from the shape
 222 data in Michikami et al. (2019) when calculating interparticle forces and initial porosi-
 223 ties below.

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2.2 Macroporosity

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The macroporosity of Ryugu can be calculated from the volume size-frequency distribution (Eq. 2) assuming linear mixing models (Yu & Zou, 1998; Zou et al., 2011). In the mixing theory, the macroporosity achieved for a given size distribution will be a function of the volume fractions X_i , the initial porosity ϕ_i , as well as the nominal equivalent volume diameter d_i of particles in each bin. The latter represents the diameter of a volume-equivalent sphere. Further, $i = 1, \dots, n$ is the number of size bins used and $d_1 > d_2 > \dots > d_n$ for convenience. Then, the macroporosity ϕ_{Macro} can be expressed as

$$\phi_{Macro} = f(X_1, \dots, X_n; d_1, \dots, d_n; \phi_1, \dots, \phi_n). \quad (6)$$

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Note that the equivalent volume diameter d_i of particles is not strictly identical to the mean horizontal diameter as defined by Michikami et al. (2019), but as the observed boulder axis ratios on Ryugu change only little as a function of horizontal diameter, the shape factor relating horizontal diameter to the equivalent volume diameter d_i is close to constant. It can thus be factored out for the mixing model below and has a negligible effect on the Bond number.

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The above formulation holds if particle sphericity is independent of particle size, which is the assumption made in the following. However, we note for completeness that the method to estimate macroporosity used here can be generalized to arbitrary sphericity-size relations $\Psi(d)$ by introducing the equivalent packing diameter d_p , which then accounts for particle shape effects, i.e., mixing of particles that have different sphericities at different sizes. Then, the equivalent volume diameter d in Eq. 6 needs to be replaced by the equivalent packing diameter d_p , which is related to the observed equivalent volume diameter d through sphericity $\Psi(d)$ by (Yu & Zou, 1998) by the empirical relation

$$\frac{d}{d_p} = \Psi(d)^{2.785} e^{2.946(1-\Psi(d))} \quad (7)$$

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The dimensionless specific volume describing the packing state for each bin is defined as (Zou et al., 2011)

$$V_j = \frac{1}{1 - \phi_j} \quad (8)$$

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and the macroporosity finally attained by the mixture will be governed by the interaction of all differently sized particles. However, there will be one intermediate-sized bin

251 i that controls the packing structure (see Yu & Zou (1998), also compare Fig. 1). While
 252 the size-bin number i of the controlling component is not known a priori, the specific vol-
 253 ume \tilde{V}_i of a particular packing can in general be expressed as

$$\tilde{V}_i = \sum_{j=1}^{i-1} [V_j - (V_j - 1)g(d_i, d_j)]X_j + V_i X_i + \sum_{j=i+1}^n [V_j(1 - f(d_i, d_j))]X_j \quad (9)$$

254 where small particles have indices $j = 1 \dots i - 1$ and large particles have indices $j =$
 255 $i+1 \dots n$. The functions $f(d_i, d_j)$ and $g(d_i, d_j)$ are referred to as interaction functions
 256 between components i and j and were derived experimentally (Yu et al. (1997), Zou et
 257 al. (2011)). They are given by

$$f(d_i, d_j) = f(r_{ij}) = (1 - r_{ij})^{3.33} + 2.81r_{ij}(1 - r_{ij})^{2.77} \quad \text{and} \quad (10)$$

$$g(d_i, d_j) = g(r_{ij}) = (1 - r_{ij})^{1.97} + 0.36r_{ij}(1 - r_{ij})^{3.67} \quad (11)$$

258 and depend on the equivalent packing diameter size ratios r_{ij} between small and large
 259 particles of the two components. Parameters r_{ij} can be expressed as (Zou et al., 2011)

$$r_{ij} = (1 - x_{ij})R_{ij}^k + x_{ij}R_{ij} \quad (12)$$

260 where $R_{ij} = d_j/d_i$ is the small-to-large size ratio and $i < j$. The empirical parameter
 261 k is 0.451 (Zou & Yu, 1996), and x_{ij} depends on the type of particle-particle interaction
 262 (Zou et al., 2011). It is given by

$$x_{ij} = \begin{cases} 1 & d_j > d_{cri} \\ 0 & d_i < d_{cri} \\ 1 - 1.543 \cdot e^{-0.697d_i/d_{cri}} & d_j \leq d_{cri} \leq d_i \end{cases} \quad (13)$$

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 264 In the above equation, the critical particle diameter d_{cri} divides fine and coarse par-
 265 ticles, i.e., it is the particle diameter below which cohesion between particles starts to
 266 influence particle interactions. Under Earth gravity conditions, d_{cri} is close to 150 μm
 267 (Zou et al., 2011), but under micro-gravity conditions, cohesion can be relevant even for
 268 decimeter-sizes boulders (Scheeres et al., 2010; Kiuchi & Nakamura, 2015; Zou et al., 2011).
 269 Here, we define the critical diameter based on the Bond number B , i.e., the ratio between
 270 interparticle forces and the weight of a particle (Scheeres et al., 2010).

271 We define the Bond number assuming a cleanliness factor equal to unity and a par-
 272 ticle separation of $1.5 \cdot 10^{-10}$ m (Scheeres et al., 2010). Furthermore, we calculate the co-

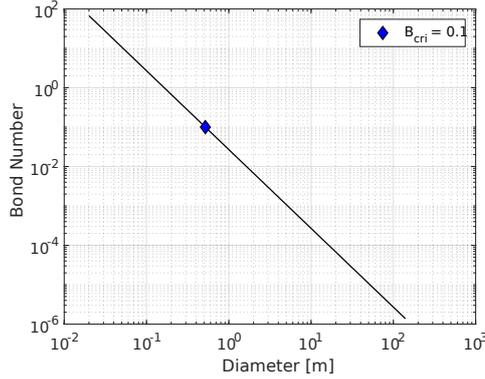


Figure 3. Bond number, i.e., the ratio between interparticle forces and particle weight, as a function of particle diameter, assuming parameters as appropriate for Ryugu. The diameter corresponding to a critical Bond number of $B_{crit} = 0.1$ is indicated.

hesive force for equally sized particles and include effects of particle sphericity Ψ and roundness Ω (Powers, 1953) by adding these as multiplicative factors (Wood, 2020). The Bond number is then given by

$$B(d) = \frac{1.1 \cdot 10^{17} A \Psi \Omega}{\rho g (d/2)^2} \quad (14)$$

where d is particle diameter, $g = 0.9825 \cdot 10^{-4} \text{ m s}^{-2}$ is volume averaged gravity of Ryugu (Yamamoto et al., 2020), and $A = 4.1 \cdot 10^{-20} \text{ J}$ is the Hamaker constant for olivine in high vacuum (Perko et al., 2001). While olivine is certainly not the most common mineral in carbonaceous material, we consider its Hamaker constant to be a more appropriate choice than, e.g., the widely used Hamaker constant for amorphous SiO_2 . In any case, the Hamaker constant needs to be regarded as highly uncertain. This also implies that the exact choice of parameters like boulder density, sphericity, and roundness has little influence on the results presented below. We choose boulder bulk density $\rho = 1420 \text{ kg m}^{-3}$ to match a macroporosity of 16 % and a bulk density of 1190 kg m^{-3} (Watanabe et al., 2019) for consistency, where ρ was determined using an iterative approach. Furthermore, we choose a particle roundness Ω of 0.24, as appropriate for angular to subangular particles (Powers, 1953).

The resulting Bond number for parameters appropriate for Ryugu is shown in Fig. 3 as a function of particle diameter. The critical diameter d_{crit} corresponding to a critical Bond number $B_{crit} = 0.1$ is indicated in blue and has been calculated using Eq. 14. We use $B_{crit} = 0.1$ as a baseline, i.e., we assume that cohesion starts to have a notice-

292 able effect on porosity once the interparticle forces exceed 10 % of the particle weight.
 293 For Ryugu, $B_{cri} = 0.1$ corresponds to $d_{cri} = 0.52$ m, but the influence of varying B_{cri}
 294 over a large range will also be discussed.

295 To evaluate Eq. 9, we first discretize the size range between $D_{\min} = 0.02$ m and
 296 $D_{\max} = 140$ m into $\log(D_{\max}/D_{\min})/\log(q)$ logarithmically spaced bins. We use a size
 297 factor of $q = 1.05$ from one bin to the next, resulting in a total of 182 size bins, which
 298 turned out to be sufficient. Volume fractions X_i in each size-bin were calculated accord-
 299 ing to the Weibull or power law representation of the size-frequency distribution as needed.
 300 Furthermore, initial specific volumes V_i and therefore initial porosities ϕ_i need to be pre-
 301 scribed. While initial porosities of coarse monosized spherical particles generally vary
 302 between 0.42 for loose random packing and 0.36 for dense random packing (Scott, 1960),
 303 cohesive forces between small particles can considerably increase porosities (Scheeres et
 304 al., 2010; Kiuchi & Nakamura, 2015). We use the empirical relation (Kiuchi & Nakamura,
 305 2015; Kiuchi & Nakamura, 2015b)

$$\phi_i = \phi_0 + (1 - \phi_0)e^{-\alpha B(d_i)^{-\gamma}} \quad (15)$$

306 to determine initial porosity, where ϕ_0 is the porosity of the non-cohesive particles and
 307 describes the packing state. Note that we here implicitly assume initial porosities as ap-
 308 propriate for spherical particles, as for the relevant range of observed sphericities the in-
 309 fluence of deviations from an ideal spherical shape on initial porosity is negligible (Zou
 310 & Yu, 1996). Particle shape enters Eq. 15 in the Bond number $B(d_i)$ only, and it is a
 311 secondary effect in the analysis presented for Ryugu below. The constants $\alpha = 2.414$
 312 and $\gamma = 0.1985$ have been derived from a new fit to the data of Kiuchi & Nakamura
 313 (2015). Finally, the specific volume occupied by the mixture is obtained by calculating
 314 the maximum of all specific volumes for the different controlling mixture sizes and

$$V = \max\{\tilde{V}_1, \dots, \tilde{V}_n\} \quad (16)$$

315 Mixture macroporosity is then given by $\phi_{Macro} = 1 - 1/V$.

316 In summary, the following steps need to be performed to determine the macropo-
 317 rosity of a granular mixture using the model above: First, volume fractions in the in-
 318 dividual size-bins need to be calculated from the given size-frequency distribution (Eq.
 319 8, 9). Then, initial porosity in each size-bin needs to be determined. This will primar-
 320 ily depend on the packing state. Further, it also depends on particle roundness and shape,

321 which influence cohesion (Eq. 14, 15) as well as the geometrical packing properties (not
 322 considered here). Finally, the macroporosity is determined by examining all possible par-
 323 ticle interactions (Eq. 16).

324 **2.3 Average Grain Density**

325 While the main goal of the present paper is a determination of the macro-porosity
 326 of rubble-pile asteroid Ryugu, additional information on the asteroid’s average grain den-
 327 sity can be derived. As macroporosity ϕ_{Macro} , microporosity ϕ_{Micro} , and bulk density
 328 ρ_{Bulk} are related by

$$\phi_{Macro} = 1 - \frac{1 - \phi_{Bulk}}{1 - \phi_{Micro}} \quad (17)$$

329 information on grain density ρ_{Grain} can be extracted from

$$\phi_{Bulk} = 1 - \frac{\rho_{Bulk}}{\rho_{Grain}} \quad (18)$$

330 Eq. 18 requires the macroporosity, microporosity, and bulk density to be known.
 331 While the bulk density of Ryugu was estimated to be $1190 \pm 20 \text{ kg m}^{-3}$ (Watanabe et
 332 al., 2019), the boulders’ microporosity cannot currently be unambiguously constrained
 333 due to the difficulties associated with extrapolating meteorite thermal conductivities to
 334 porosities in excess of 20 % (Grott et al., 2019; Macke et al., 2011). However, end-member
 335 models (Flynn et al., 2018; Henke et al., 2016) suggest microporosities ϕ_{Micro} of either
 336 $32 \pm 2 \%$ or $50 \pm 2 \%$ for Ryugu’s dark and rugged boulders (Hamm et al., 2020) which
 337 comprise the vast majority of all boulders observed on the surface (Sugita et al., 2019;
 338 Okada et al., 2020). We will use Monte-Carlo simulations to propagate these uncertain-
 339 ties to the determination of Ryugu’s grain density, while simultaneously taking the un-
 340 certainty associated with Ryugu’s macroporosity as derived from the linear mixing the-
 341 ory (Sec. 2.2) into account.

342 **3 Results**

343 Given the parameterization of the size-frequency distribution (Eq. 1) for the boul-
 344 ders observed on the surface of Ryugu, and assuming the distribution also applies to the
 345 interior, we have first calculated the corresponding volume frequency distribution using
 346 Eq. 2. Given roundness Ω , Hamaker constant A , particle bulk density ρ , and volume av-
 347 erage gravity g (see Eq. 14), we then varied the initial porosity ϕ_i in each size bin (Eq.

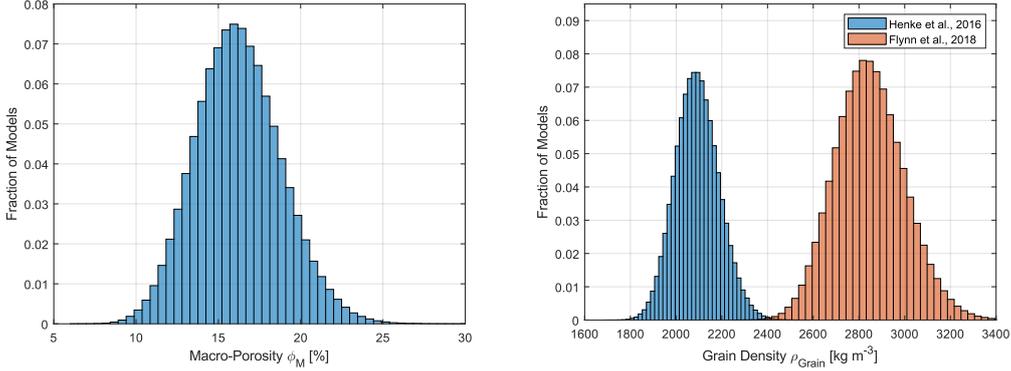


Figure 4. Left: Histogram of macroporosities ϕ_M obtained using Monte-Carlo simulations. Right: Ryugu grain densities derived from a second set of Monte-Carlo calculations (see text for details). The two distinct distributions result from the uncertainty of microporosity for Ryugu’s boulders (Grott et al., 2019; Hamm et al., 2020), and two end-member models for the microporosity have been assumed.

348 15) using a Gaussian distribution for ϕ_0 centered around 39.5 % with standard devi-
 349 ation of 3 %. In addition, particle sphericity was varied using a Gaussian distribution
 350 centered around 0.85 with standard deviation of 0.06, and 10^6 draws from these distri-
 351 butions were used in a Monte-Carlo simulation to calculate the resulting macroporos-
 352 ity according to Eq. 16.

353 Results of the calculation are shown in the left hand panel of Fig. 4, where a his-
 354 togram of the obtained macroporosities ϕ_M is shown. The range of macroporosities ob-
 355 tained in the calculations is $\phi_M = 16.2 \pm 2.6$ % (1-sigma), and thus considerably smaller
 356 than porosities of monodisperse packings. This is not surprising given the broad parti-
 357 cle size distribution observed on the surface of Ryugu.

358 Given the range of macroporosities derived above as well as estimates for the boul-
 359 der microporosities derived from in-situ thermal inertia measurements (Grott et al., 2017,
 360 2019; Hamm et al., 2020), we calculated the range of grain densities compatible with the
 361 observed bulk porosity of Ryugu (Watanabe et al., 2019) using Eq. 17 and 18. We ap-
 362 plied two endmember models for the microporosity ϕ_{Micro} : for the first model (Flynn
 363 et al., 2018) we use $\phi_{Micro} = 50 \pm 2$ %, while for the second model (Henke et al., 2016)
 364 $\phi_{Micro} = 32 \pm 2$ % (Hamm et al., 2020). In the 10^6 Monte-Carlo simulations performed,
 365 we varied microporosity using Gaussian distributions centered around 50 % and 32 %

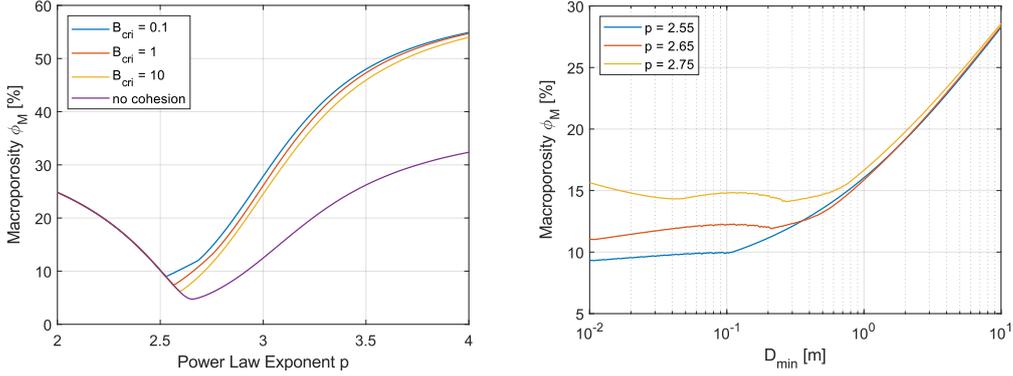


Figure 5. Left: Macroporosity ϕ_{Macro} (porosity caused by void spaces in-between particles) as a function of the power law exponent of the underlying size-frequency distribution and critical Bond number B_{crit} . For comparison, results obtained neglecting cohesion between particles are also shown. Right: Macroporosity ϕ_{Macro} as a function of lower cutoff size D_{min} for three different power law exponents p . For reference, the power law exponent for Ryugu as derived from the observed surface boulder size-frequency distribution is $p = 2.65$ on average (Michikami et al., 2019).

366 with standard deviations of 2 %, respectively. Furthermore, we varied bulk density using
 367 a Gaussian distribution centered around 1190 kg m^{-3} with a standard deviation of
 368 20 kg m^{-3} (Watanabe et al., 2019) and macroporosity using a Gaussian distribution cen-
 369 tered around 16.2% with a standard deviation of 2.6 %.

370 Results of the calculation are shown in the right hand panel of Fig. 4, where the
 371 resulting histograms for the grain densities ρ_{Grain} are shown for the two endmember mod-
 372 els. Owing to the two different models used to estimate boulder microporosity, two sep-
 373 arate peaks are obtained for the distribution of grain densities. For the model of Flynn
 374 et al. (2018), we find grain densities of $\rho_{Grain} = 2848 \pm 152 \text{ kg m}^{-3}$, whereas the model
 375 of Henke et al. (2016) results in $\rho_{Grain} = 2093 \pm 96 \text{ kg m}^{-3}$. As expected, higher mi-
 376 croporosities (Flynn et al., 2018) yield larger grain densities and vice versa to satisfy the
 377 constraint posed by Ryugu’s bulk density.

378 Results of a systematic study of the influence of critical Bond number B_{crit} and lower
 379 diameter cutoff sizes D_{min} on the obtained macroporosities ϕ_{Macro} are shown in Fig. 5.
 380 Here, the size-frequency distribution of boulders has been approximated by a power law
 381 with exponent p to facilitate a comparison of Ryugu with other rubble pile asteroids. For

382 smaller power law exponents, the size-frequency distribution is shallower as compared
 383 to distributions with larger p , and as a result, such distributions represent surfaces with
 384 a higher ratio of large particles.

385 In general, the macroporosities ϕ_{Macro} obtained using the above mixing theory show
 386 a distinct minimum at intermediate power law exponents p , whereas distributions which
 387 have too many small or too many large particles result in unfavorable mixing and larger
 388 ϕ_{Macro} are obtained. This minimum around $p = 2.5$ is known as the Fuller parabola
 389 in the engineering literature and has long been known as the optimum packing size dis-
 390 tribution for spherical particles (Fuller & Thompson, 1907). Results obtained varying
 391 the critical Bond number are shown in the left panel of Fig. 5, where the critical Bond
 392 number parametrizes the particle size below which interparticle forces result in signif-
 393 icant cohesion. As expected, low critical Bond numbers, corresponding to larger contri-
 394 butions from cohesive particles, result in larger macroporosities. However, the overall ef-
 395 fect is small and in the few percent range. The low critical Bond number of 0.1 adopted
 396 above therefore results in a conservative upper limit on macroporosity. It is also worth
 397 noting that results obtained using a power law distribution with $p = 2.65$, which over-
 398 estimates the fraction of small particles, are lower than those obtained using the Weibull
 399 representation of the data by 4-5 %, such that results obtained using global power law
 400 fits must be interpreted with caution. For comparison, results obtained neglecting co-
 401hesion are also shown in the left panel of Fig. 5, and macroporosity approaches a limit
 402 of 39.5 % (compare Eq. 15) for large p (not shown).

403 The influence of varying the lower cutoff diameter D_{min} of the size-frequency dis-
 404 tribution on the obtained macroporosity ϕ_{Macro} is shown in the right panel of Fig. 5,
 405 where ϕ_{Macro} is shown as a function of D_{min} for three power law exponents p . In the
 406 calculations, a critical Bond number of $B_{cri} = 0.1$ has been assumed. While the min-
 407 imum macroporosity that can be achieved by the packing is close to constant for small
 408 D_{min} , predicted macroporosity drastically increases for cutoff diameters larger than a
 409 few decimeters. In this case, unfavorable mixing is a result of the sparsity of smaller rocks
 410 to fill the gaps between larger blocks. These results indicate that image data with cen-
 411 timeter resolution are necessary to properly characterize the packing state of rubble pile
 412 asteroids, and that results presented above are largely independent of the cutoff size of
 413 $D_{min} = 0.02$ m imposed by the image data available for Ryugu.

4 Discussion and Conclusions

In the present paper, we have used semi-empirical models for the porosity of multi-component mixtures to estimate the macroporosity of Cb-type asteroid (162173) Ryugu based on the observed size-frequency distribution of boulders on the asteroid's surface and the assumption that the surface distribution of boulders is representative for the bulk asteroid. Using the concept of controlling mixtures (Yu & Standish, 1991; Yu & Zou, 1998; Zou et al., 2011), we estimated the macroporosity of Ryugu to be $\phi_M = 16.2 \pm 2.6$ %. Based on estimates of boulder microporosity, we furthermore constrained the average grain density of Ryugu's boulders to $\rho_{Grain} = 2848 \pm 152$ kg m⁻³ or $\rho_{Grain} = 2093 \pm 96$ kg m⁻³, depending on the microporosity model used.

Boulder shape can affect the above mixing model by changing interparticle cohesion, by changing the geometrical arrangement between different particle sizes, and by changing the initial porosity in each individual size-bin. In the modeling, we have taken the influence of shape on particle cohesion explicitly into account, while we neglected its influence on geometrical interactions and initial porosity. This is justified because for the case of Ryugu the majority of particles has axis ratios b/a in excess of 0.5 (Michikami et al., 2019), corresponding to sphericities larger than 0.7. For such particles, initial porosity is nearly independent of shape and equal to the value appropriate for spherical particles (Zou & Yu, 1996). It is also worth noting that for Ryugu all of the above are secondary effects when compared to the unknown packing state, which we address by considering the entire range stretching from a random loose to a random close packing.

For the case of Ryugu, the primary factor determining macroporosity is the boulder size-frequency distribution, and while the applied model takes cohesion between particles into account, disregarding cohesion results in only a slight modification of the obtained macroporosity ϕ_M . Switching off cohesion in the model by assuming a critical bond number of 10^5 results in a macroporosity of 16.1 %, only 0.1 % smaller than the value presented above. This is a direct consequence of the low volume fraction of small cohesive particles on Ryugu, which directly follows from the given boulder size-frequency distribution. This also implies that the results presented here are robust with respect to the exact choice of parameters like boulder density, roundness, and sphericity, which enter the calculation of the Bond number. It is also worth noting that the power law representation of the data significantly overestimates the influence of cohesion on macro-

446 porosity when compared to the Weibull fit by overestimating the volume fraction of small
447 particles.

448 The cratering experiment performed by Hayabusa2's small carry-on impactor re-
449 sulted in the formation of a crater in the gravity-dominated regime (Arakawa et al., 2020),
450 indicating that particle cohesion played a minor role in the crater formation process. On
451 the other hand, particles with diameters of 0.2 m were observed in the SCI crater wall,
452 which, according to Eq. 14, have Bond numbers close to unity and should therefore in-
453 teract cohesively. This apparent discrepancy is resolved by the fact that small particles
454 appear not to be volumetrically dominant inside Ryugu. This is indicated by the shal-
455 low particle size distribution for particles smaller than 1 m on the surface (Sugita et al.,
456 2019; Michikami et al., 2019) and inside the artificial crater (Arakawa et al. (2020), Fig.
457 S5), where the particle size distribution shows a power law exponent $p \sim 2$ (also com-
458 pare the volume-size distribution on the right hand side of Fig. 2). Therefore, results of
459 the cratering experiment confirm that cohesion has a small influence on Ryugu's pack-
460 ing state. However, cohesion may become significant for rubble pile asteroids with a steep
461 particle size distribution, e.g., power-laws with $p > 3$, where - in contrast to Ryugu -
462 the mixture is dominated by a high volume fraction of very small particles.

463 Although a full analysis using empirical fits of the cumulative boulder size-frequency
464 distribution of other small bodies has not been performed here, macroporosity results
465 can be qualitatively compared by considering the power law exponents of their respec-
466 tive size distributions and assuming similar size cutoffs D_{\min} and D_{\max} . The former have
467 been widely used to describe size distributions in the literature, and values of $p = 2.9 \pm$
468 0.3 and $p = 3.52 \pm 0.20$ have been obtained for Bennu (Lauretta et al., 2019) and Itokawa
469 (Michikami et al., 2008; Mazrouei et al., 2014; Michikami et al., 2019), respectively. As-
470 suming $B_{cri} = 0.1$ as above, these correspond to macroporosities between 10 and 38 %
471 for Bennu and 43 to 52 % for Itokawa. Assuming average grain densities of 2600 kg m^{-3} ,
472 Bennu's low bulk density of $1190 \pm 13 \text{ kg m}^{-3}$ (Lauretta et al., 2019) implies a bulk poros-
473 ity of 54 %, indicating significant microporosity. For Itokawa, average grain density has
474 been estimated based on the modal abundance of minerals in the returned samples, and
475 densities of 3400 kg m^{-3} have been obtained (Tsuchiyama et al., 2011, 2014). This im-
476 plies a bulk porosity of $39 \pm 6 \%$ (Abe et al., 2006; Fujiwara et al., 2006; Tsuchiyama
477 et al., 2011), consistent with the results obtained from the mixing theory. For Eros, the
478 power law exponent of $p = 3.31 \pm 0.06$ (Thomas et al., 2001) implies macroporosities

479 of 40 - 45 %, which is larger than the inferred bulk porosity of Eros. The latter is es-
 480 timated to be 21-33 % (Yeomans et al., 2000; Wilkison et al., 2002), indicating that macro-
 481 porosities derived using the presented mixing model are incompatible with the observa-
 482 tions. However, although Eros is a heavily fractured body, there is little evidence that
 483 it was ever catastrophically disrupted and later reaccumulated into a rubble pile (Wilk-
 484 ison et al., 2002), such that the theory presented here can probably not be applied.

485 Results for Ryugu have been obtained assuming minimum and maximum particle
 486 sizes of 0.02 m and 140 m, respectively, and these results are robust with respect to the
 487 cut-off at small particle sizes D_{\min} . Only shifting the cut-off D_{\min} to values larger than
 488 0.30 m has a noticeable effect on the macroporosity. The upper cut-off size D_{\max} was
 489 chosen to correspond to the Otohime boulder, which is the largest boulder observed on
 490 Ryugu's surface. However, boulders larger than Otohime could potentially reside in Ryugu's
 491 interior, which would decrease the obtained macroporosity through a filling of void spaces.
 492 Reasonable upper limits on monolith sizes are 200 m, as derived from observations of fast
 493 rotators in the asteroid population (Pravec & Harris, 2000) and the catastrophic disrup-
 494 tion threshold (Benz & Asphaug, 1999; Jutzi et al., 2010). Assuming $D_{\max} = 200$ m
 495 reduces ϕ_{Macro} to 15 %.

496 One way to increase macroporosity in the above models would be an increased ini-
 497 tial porosity in each size bin, which may for example be caused by mechanical interlock-
 498 ing of particles due to particle angularity. For a random loose packing, non-cohesive ini-
 499 tial porosity can increase from ~ 42 % for smooth frictionless particles to ~ 44 % for very
 500 rough particles (Onoda & Liniger, 1990; Jerkins et al., 2008). In the frame of the applied
 501 mixing model, this effect is taken into account in the chosen initial porosity (Eq. 15),
 502 and shifting the applied Gaussian distribution in the performed Monte-Carlo simulations
 503 by 2 % results in slightly increased macroporosities of 18.0 ± 3 %. Therefore, while rough-
 504 ness and particle interlocking can increase macroporosity, this is likely not a significant
 505 effect.

506 While the obtained macroporosity may appear to be relatively low, a significant
 507 reduction with respect to the porosity of random close packings of monodisperse spheres
 508 can be expected. Even binary mixtures of particles can be arranged in packing states
 509 with porosities of 15-20 % (Yu & Standish, 1991; Yu et al., 1992), such that it should
 510 not be surprising to achieve similar packing densities with the broad size distributions

511 used here. Ternary mixtures can achieve $\phi_{Macro} < 10\%$ (Yu & Standish, 1991), and
 512 while most common loose or compact granular materials have macroporosities between
 513 30% and 50%, almost any degree of macroporosity between 10 and 90% can be ob-
 514 tained for polydisperse angular particles (Dullien, 1991). Experimentally, macroporosi-
 515 ties down to 10% have been produced in the lab (Latham et al., 2002). Therefore, the
 516 macroporosity of Ryugu obtained here falls within a reasonable range, and Ryugu’s high
 517 bulk porosity is a direct consequence of the very large microporosity of Ryugu’s boul-
 518 ders.

519 The average grain densities obtained here are much lower than typical grain den-
 520 sities of ordinary chondrites, which range from 3520 to 3710 kg m^{-3} (Flynn et al., 2018),
 521 and also lower than those of most carbonaceous chondrites, which typically have grain
 522 densities in excess of 3360 kg m^{-3} (Flynn et al., 2018). Only the CM and CI sub-classes
 523 show lower grain densities, and $\rho_{CM,Grain} = 2960 \pm 40 \text{ kg m}^{-3}$ while $\rho_{CI,Grain} = 2420$
 524 kg m^{-3} (Consolmagno et al., 2008; Macke et al., 2011; Flynn et al., 2018). The Tagish
 525 Lake meteorite, an ungrouped carbonaceous chondrite, exhibits similar grain densities
 526 in the range between 2430 and 2840 kg m^{-3} (Ralchenko et al., 2014). While the larger
 527 grain densities of $2848 \pm 152 \text{ kg m}^{-3}$ are consistent with the CM and Tagish Lake re-
 528 sults, the lower densities of $2093 \pm 95 \text{ kg m}^{-3}$ are inconsistent with those of known me-
 529 teorite samples.

530 Estimates of grain densities discussed above indicate that extrapolating boulder
 531 porosities as a function of thermal conductivity using the model by Flynn et al. (2018)
 532 is preferred to extrapolations using the model by Henke et al. (2016). In addition, lab-
 533 oratory measurements of thermal conductivity (Hamm et al., 2019) using the UTPS Tag-
 534 ish Lake meteorite simulant (Miyamoto et al., 2018) provide further evidence of high boul-
 535 der microporosity. The UTPS simulant has a grain density of 2813 kg m^{-3} and a poros-
 536 ity of 47.5%, while thermal conductivity was determined to be similar to that of Ryugu’s
 537 rugged boulders (Hamm et al., 2019). It therefore seems likely that boulder porosity on
 538 Ryugu falls within the high range determined by Grott et al. (2019), but more labora-
 539 tory measurements of thermal conductivity at high porosity are needed to confirm these
 540 results and reduce uncertainties. If grain densities are indeed of the order of 2850 kg
 541 m^{-3} , Ryugu’s bulk porosity is estimated to be 58% (cf. Eq. 18).

542 It is noted that close-up images have revealed that many boulders on Ryugu and
543 Bennu exhibit morphologic properties consistent with a brecciated structure (Sugita et
544 al., 2019; Walsh et al., 2019). Breccia would have much larger microporosities than pris-
545 tine rocks, consistent with the large microporosities preferred here. Furthermore, the pres-
546 ence of breccia on Ryugu and Bennu is consistent with the fact that many carbonaceous
547 chondrites and, in particular, all CM and CI meteorites found on Earth are known to
548 be brecciated (Bischoff et al., 2006). However, it remains to be investigated if breccia-
549 tion is the main mechanism providing microporosity, or whether the boulder’s highly porous
550 structure is a result of the formation mechanisms acting in Ryugu’s parent body (Neu-
551 mann et al., 2014, 2015).

552 If microporosity in typical carbonaceous asteroids is as high as predicted here for
553 Ryugu, macroporosities of rubble pile asteroids may have been systematically overesti-
554 mated in the past (e.g., Consolmagno et al., 2008). Macroporosities have been estimated
555 based on measurements of asteroid bulk density and porosities of meteorite samples, the
556 latter of which could have been underestimated compared to values for actual carbona-
557 ceous material on asteroids derived from in-situ measurements (Grott et al., 2017, 2019).
558 This bias could be the result of filtering by the Earth’s atmosphere, as only the strongest,
559 densest carbonaceous meteoroids would survive atmospheric entry, while weaker sam-
560 ples would break up (Popova et al., 2011). This could explain the absence of high poros-
561 ity samples in our meteorite collections, where the most porous sample reported to date
562 is the Tagish Lake meteorite, which shows porosities in the range from 26 to 36 % (Ralchenko
563 et al., 2014). The samples to be returned from Ryugu by the Hayabusa2 mission will pro-
564 vide crucial information on this issue.

565 Results presented here assume that the size-frequency distribution observed on the
566 surface of Ryugu is representative for the entire asteroid, but as discussed in Sec. 1, the
567 reaccumulation process itself as well as post accretion surface modifications could influence
568 the observed size-frequency distribution. For example, meteorite impacts could increase
569 the number of small boulders on the surface and the observed size-frequency distribu-
570 tion would be steeper than the distribution in the interior. Therefore, macroporosity would
571 have been overestimated in the presented model, as the interior distribution would move
572 closer to the Fuller minimum (Fuller & Thompson, 1907). Conversely, the Brazil Nut
573 Effect could bias the slope of the surface size-frequency distribution towards smaller val-
574 ues, implying that macroporosity would have been underestimated. This topic can be

575 addressed once average grain density and possibly also microporosity have been deter-
 576 mined from the returned samples, as has been done for Itokawa (Tsuchiyama et al., 2011).
 577 Then, Ryugu’s macroporosity can be derived given the measured bulk density (Watan-
 578 abe et al., 2019). Any significant deviation from the macroporosity value calculated here
 579 will indicate a non-homogeneous boulder size distribution in the bulk volume of the as-
 580 teroid.

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