Length of main text: 3430 words, methods: 2421 words, captions: 812 words. Number of references: 65 (> 15 unique to the Methods). Number of figures: 6. Number of tables: 0. Extended data: 1 figure, 1 table.

DRAFT VERSION JULY 21, 2026
Typeset using LaTeX **manuscript** style in AASTeX63

The Role of Disruptive Impacts on Ocean Generation and Longevity in Icy Moons

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ABSTRACT

Among icy moons of the outer Solar System, subsurface water oceans and large collisions both appear common, but the impact of the latter on the presence and persistence of the former is unclear and has rarely been investigated. Here we interface a smoothed-particle hydrodynamics (SPH) model to simulate collisions and a thermal-structural evolution model to simulate the evolution of moons pre-, post-, and without a collision. Overall, even such large-scale collisions only affect ocean thickness or longevity, and ocean presence or absence, only for part of a moon's history. In reaccreted moons, the ocean survives the impact and becomes much thicker inside larger moons with radius near 1000 km, whereas an ocean that would otherwise arise inside moons with radius near 500 km is absent because the collision promotes ice-rock differentiation. Our simulations have not yielded an ocean developed post-impact –whether directly via collisional or reaccretional heating, or indirectly through tidal heating due to collision-induced orbital changes– in a moon that would otherwise have

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remained frozen. The ocean-enhancing effect is pronounced only for late disruptive impacts onto large, 1000 km-class targets, which are unlikely in recent solar system history.

MAIN TEXT

Disruption and reassembly of moons may be common in the outer Solar System. Within the Saturn system, multiple studies suggest that a precursor moon or system of moons was disrupted by an external impactor or through moon-moon collisions after satellite interactions destabilized the system. The present-day moon (or moons) are then assumed to have assembled from the collisional debris (1; 2; 3; 4; 5; 6; 7). At Uranus, a giant impact has been invoked to explain the planet's extreme axial tilt (e.g., 8; 9; 10; 11; 12). The tilting is thought to have destabilized the orbits of the primordial satellites, leading to catastrophic collisions, with the current satellite system reassembled from the resulting debris disk (e.g., 13; 14; 15) or a disk of material ejected by the impact onto proto-Uranus (9; 10; 11; 16). At Neptune, Triton, the largest moon, is thought to have been captured from a heliocentric orbit (17; 18). This capture likely disrupted Neptune's primordial satellite system, due to mutual collisions (19) and collisions onto the newly captured Triton (20).

Curiously, some of the moons associated with disruptive collisions and reassembly are also candidates for recent or present-day oceans, such as Mimas (21; 22; 23), Enceladus (24), Dione (25), Miranda (26), and Ariel (27). While the larger of these moons could develop oceans from early sources like radiogenic decay, heat of accretion, or tidal heating that predates rapid orbit circularization or despinning (e.g., 28; 29), these heat sources may not be effective after disruption and reassembly of a moon. Furthermore, smaller moons may not develop oceans even in the presence of these primordial heat sources (e.g., 29). Whether disruption and reassembly can provide additional heat sufficient to produce oceans is not currently known. Here, we address the evolution of the interiors of mid-sized moons, with and without oceans, in the aftermath of large collisions to determine whether such impacts can be ocean-promoting or ocean-limiting events.

To this end, we perform computer simulations of the thermal and structural evolution of ice-rock moons, pre- and post-collision, linking these two stages with smoothed-particle hydrodynamics (SPH) simulations of a disruptive collision (Fig. 1). Differences in the basic assumptions within these two numerical models preclude end-to-end simulations of moons with internal structures and temperatures that remain selfconsistent across thermal evolution and impact simulations. In addition, because simulations of the collision are significantly more computationally intensive than those of the thermal evolution, we sought to maximize use of a few disruption/reassembly simulations with many more thermal evolution simulations pre- and post-collision.

First, we identify the suite of plausible outcomes of the initial assembly and pre-impact interior evolution of mid-sized moons assuming primordial accretion in a circumplanetary disk (Methods). We then generate target moons as inputs to the impact simulations that have properties within the range of plausible moon structures. We conduct a suite of head-on impact simulations using different velocities and impactor properties (Extended Data Table 1). We then build post-impact moons consistent with the outcomes of collisions and evolve the interiors and orbits to track ocean generation and longevity. We also track the long-term evolution of moons without any collisions as a baseline to determine how the impact affects ocean development.

We focus on objects in two size classes: radius $R = 500$ km and R = 1000 km, differentiated at the time of impact, having collided head-on with objects respectively 250 (undifferentiated) and 500 km (differentiated) in radius. These sizes bracket those of Uranian and Saturnian moons (30; 31): Ariel (R = 579 km), Umbriel (R = 585 km), Titania (R = 789 km), Oberon (R = 761 km), Tethys (R = 531 km), Dione (R = 561 km), and Rhea (R = 763 km). The smaller impactor has a size close to those of Miranda (R = 236 km), Mimas (R = 198 km), and Enceladus (R = 252 km).

We perform impact simulations of two moons in the gravitational potential of a Uranus-like planet at 10 uranian radii (R_U). The collision between objects 500 and 250 km in radius is simulated using an impact velocity of 1, 2, or 3 times the mutual escape velocity, $V_{esc} = 0.4 \text{ km s}^{-1}$ (32). For the larger objects, 1000 and 500 km, disruption is achieved only at a higher velocity (33), hence we assume an impact at $3\, V_{esc} \approx 2.1 \text{ km s}^{-1}$.

RESULTS

The simulated collisions can strip material from the larger object or add material to it; i.e., the initial and final moons do not have the same size or mass. Thus, we assess the effect of collisions on post- and no- impact moons with radii and densities consistent with the final post-collision moon, which differ from assumed initial SPH radii, R = 500 and 1000 km, and densities (Methods). These sizes (when fully compacted) are R = 527, 421, 366, and 901 km, respectively. The respective compacted densities are 1370, 1550, 1480, and 1370 kg m^{-3}. Results are shown respectively in Fig. 2, 3, 4, and 5. In these figures, we also present results on the effect of collisions occurring later in Solar System history (i.e., with a lower radiogenic-to-impact heat ratio), using the same set of SPH simulations as for early collisions. For the collisions with 500-km targets, we choose a late collision time at which the cooling target moon's thermal profile roughly approximates that assumed for the SPH simulation (Supplementary Figure 2). The 1000-km target does not cool down enough by the present day, but bracket simulations with radionuclide contents corresponding to 4.5 and 50 Gyr of decay yield nearly identical results (Supplementary Figure 9). At these later times, the smaller impactor's internal temperatures (comprising $(1/2)^3 = 1/8$ of particles in the SPH simulation) are likely overestimated, because the impactor cools faster than the target moon (Supplementary Figure 2).

EFFECT OF COLLISIONAL HEATING AND RESTRUCTURING ON SUBSURFACE OCEANS

The collision between $R_1 = 500$ km and $R_2 = 250$ moons at 1 V_{esc} adds material to the target. The impacts at 2 and 3 V_{esc} remove mass, predominantly from the target's icy mantle; thus the bulk rock/ice ratio increases. Collisions involving the 500 km-class moon appear to limit a subsurface ocean by promoting internal differentiation.

Without a collision, the central regions of this moon warm throughout its first billion years as radiogenic heating exceeds cooling by conduction (Fig. 2-4a,b). In the larger two moons (R = 421 and 527 km), temperatures in the deepest ice exceed the melting point of water and an ocean develops around 0.25-0.3 Gyr after formation, initiating ice-rock differentiation and persisting for about 1 Gyr. Subsequently, as radiogenic heating decreases, the ocean refreezes and the now frozen, partially differentiated moon remains in this frigid state until the present. In the smallest moon (R = 366 km), neither melting nor differentiation are initiated.

Differences in the interior evolution of the moons can be traced to differences in mixing of ice and rock in the immediate aftermath of the impact. Collisions at 2 and 3 V_{esc} mix rock and ice in the outer layers (Fig. 6), but generate enough heating in these layers ($T \geq 250$ K beyond 300 km from the center; Extended

Data Figure 1b) to promote differentiation (Fig. 3-4c,d). The effects of this differentiation are to (1) bring water closer to the surface and (2) remove rock from overlying layers. Both effects substantially decrease the thermal insulation of subsurface water, preventing sustenance of an ocean post-collision. The result is a fully frozen moon, albeit warmer than in the no-collision case. In contrast, a slower collision at $1 \times V_{esc}$ results in outer layers (beyond 300 km from the center) that are colder (< 200 K; see Extended Data Fig. 1b) and rapidly cool (Fig. 2c), allowing ice and rock to remain mixed not in their primordial proportions, but in those set during the collision, i.e., ice-rich (Fig. 2d). If the collision occurs later, 1.5 Gyr after formation, the outcomes are the same (Fig. 2-4e,f). Irrespective of impact speed, no lasting ocean occurs post-collision.

In contrast to the smaller moons, the collision involving the 1000 km-class moon appears to promote the thickness and longevity of an ocean. Without a collision, interior warming over the first 1.5 Gyr leads first to full ice-rock differentiation 0.3 Gyr after formation, then to the onset of a subsurface ocean around 0.5 Gyr after formation (Fig. 5a,b). The ocean thickness increases to about 200 km for 2 Gyr, before decreasing to near-zero at present. A collision at 0.5 Gyr greatly enhances the ocean thickness for the two billion years that follow (Fig. 5c,d). During that time, much of the moon's hydrosphere is melted below a thin layer of conductive ice. After the ocean rapidly refreezes, the balance of radiogenic heating and heat transport through the ice shell shifts back to heating the shell, leading the lower ≈ 100 km of the ice shell to melt again until the present day. With negligible radiogenic heating post-impact, the ice shell still nearly fully melts for ≥ 0.8 Gyr (Fig. 5e,f).

These outcomes are insensitive to the initial accretionary temperature profile of the progenitors, corrections to their near-surface temperature at the time of collision, and post-impact surface temperature (Supplementary Sections 1, 6, and 7), with one exception. For the 1000 km-class moon, a warmer surface (80 K instead of 55 K) after a 0.5-Gyr impact delays freezing of the thick ocean from 2 to 2.3 Gyr after reaccretion, but impedes subsequent extensive remelting (Supplementary Figure 8). A present-day ocean is inhibited by a combination of extra heat lost due to delayed freezing of the whole hydrosphere, then mild solid-state convection in the warmer, less viscous ice shell, which transports heat more effectively than conduction prevalent in Fig. 5c-d.

EFFECT OF CHANGES IN ORBITAL PARAMETERS

In the above simulations, the collided moon's orbital semi-major axis a and eccentricity e are assumed to be unaffected by the collision, and are allowed to evolve owing to gravitational interactions with Uranus. In all cases, the large $a > 10\ R_U$ and small $e < 0.02$ lead to essentially negligible contributions of solid tidal dissipation, even in the post-impact warm ice shells.

In principle, orbital changes arising from a collision could lead to the onset of an ocean, if the moon is made more susceptible to tidal dissipation not only due to the heating of its interior ice, but also by bringing it closer to its host planet, on an eccentric orbit, or by imparting obliquity or non-synchronous spin. We did not evaluate changes in obliquity or spin period, but sought to quantify maximum changes in semi-major axis a and eccentricity e, and evaluate their effect on interior evolution.

Assuming collisions are instantaneous and fully conserve kinetic energy (neglecting, e.g., heating or ejection of material), the post-impact orbital speed of a moon colliding along its velocity vector can be expressed as:

$$v_{post-collision} = \sqrt{\frac{1/2\ m_{target}\ v^2_{pre-collision} - 1/2\ m_{impactor}\ v^2_{col}}{1/2\ m_{post-collision}}} \tag{1}$$

Adopting masses based on pre-collision radii and densities (see Methods), and setting v_{col} to $(1-3) \times V_{esc}$, we find that the orbital velocity decreases only by 0.1 to 2%. This translates to a decrease in semi-major axis of ≈ 500 to 10000 km, and a corresponding increase in eccentricity by $0.002 - 0.04$ from an orbit assumed circular before the collision.

Furthermore, the most violent collisions are also the ones that produce the most ejecta – some of which is likely to re-accrete (see Discussion). This extra material is almost certain to decrease the orbital eccentricity of the moon during the reaccretion process, because the ejecta itself would collisionally grind and dynamically cool.

The negligible additional tidal dissipation associated with these orbital changes does not noticeably affect interior evolution. Even if the orbital parameters remain fixed (no eccentricity decay, e.g., due to equilibrium dissipation in a mean-motion resonance), setting e to 0.05 is insufficient to cause tidally-induced melting in the ice shell at $a = 250000$ km $\approx 9.86\ R_U$, a lower-bound distance for the $R \sim 500$ km moons starting at 10 R_U. Thus, a collision-induced orbital shift alone cannot cause the onset of ice melting inside a moon due to increased solid tidal dissipation, unless perhaps the target moon is orbiting significantly closer to the planet at the time of impact.

DISCUSSION

Our simulations suggest that disruptive moon-moon collisions can be either ocean-promoting or ocean-limiting, depending on the target moon size. A collision onto a $R \sim 1000$ km moon thickens the ocean for ~ 2 Gyrs. Collisions onto smaller moons ($R \sim 500$ km) do not promote ocean formation, but the heat produced by the impact can promote differentiation which, unlike in the $R \sim 1000$ km moon, may not otherwise happen, or not completely (Fig. 2-4). This promotion of differentiation is not a consequence of starting with differentiated SPH targets: due to the impact heat, ice and rock mixed in the disruptive collisions (Fig. 6; Supplementary Figures 3-4) separate again fully (Fig. 3-4) or out to a larger radius than without an impact (Fig. 2). In Fig. 2-3, differentiation impedes ocean formation by bringing ice closer to the cold surface. Another consequence of disruptive impacts is to decrease the size of the target moon. Thus, collisions offer a potential explanation for how a moon could be differentiated despite its current size suggesting it should be undifferentiated. Had Mimas been too cold to ever be susceptible to significant tidal dissipation, collision-induced differentiation might thus have explained its layered structure despite its small size (34).

The effect of even such large-scale collisions appears generally small, affecting ocean thickness or longevity, and influencing ocean presence or absence only for part of a moon's history. Post-collision, the heat of reaccretion is comparable to radiogenic heat produced over < 10 Myr ($R \sim 500$ km moon) to 30 Myr ($R \sim 1000$ km moon) for an early collision. The ocean-enhancing effect may be much more pronounced only for a late collision involving large ($R \sim 1000$ km) moons, for the ≈ 1 Gyr that follow. Collisional heating corresponds to 200 Myr of radiogenic heating at the present day. However, barring episodes of dynamical instability in moon systems, later collisions are much less likely than collisions in the first billion years of Solar System history (35).

These simulations assumed a Uranian heliocentric distance ($\approx$20 AU). At Saturn, moon surfaces are $\approx$20% warmer for a given albedo, and a post-impact simulation with a 45% warmer surface shows a lesser propensity for present-day ocean retention in a 1000 km-class moon than for a 55 K surface (Supplementary

Figure 8c-d). This could suggest that at lesser heliocentric distances, collisions inhibit oceans in the long term.

Although other ocean-enhancing effects could hinge on changes to orbital or spin parameters, for the moon sizes simulated, reaccretion and radiogenic heating always greatly exceed eccentricity solid tidal heating rates by at least an order of magnitude at the orbital semi-major axes investigated ($\approx$ 10 R_U). Orbital changes (e.g., eccentricity increase) due to a disruptive collision at velocities no greater than 3 times the escape velocity might thus only play a role in a system more compact (relative to the planet radius) than the present-day Uranian system, such as among the mid-sized moons of Saturn, and most likely only for higher kinetic energies of impact (impactor mass or velocity). The effects of a change in obliquity or spin remain to be investigated. Tidal despinning tends to occur much faster and to produce more dissipation heating than orbital circularization (36), but fades within $\sim$10 Myr (37).

From an eccentricity-raising standpoint, the only favorable collision outcomes would be those in which most of the disrupted material rapidly coalesces back onto a moon, as in Fig. 6. Otherwise, collided moons could end up as ring debris, in a circular orbit less prone to tidal dissipation, albeit more prone to entering mean-motion resonances (e.g., 38). Yet, the chance of a post-collision moon orbit being near a mean-motion resonance seems low, at least at late times when presumably the moon system had expanded to near its current orbital footprint, for the Uranian system spanning $\gtrsim$ 20 R_U out to Oberon. Depending on rates of orbital expansion throughout Solar System history, e.g., whether they have reflected the effect of solid tides raised on Uranus or resonance locking (39), this likelihood may have been higher at epochs of high heat flow as recorded on the surfaces of icy moons such as Ariel (40; 41; 42) and Miranda (43; 44).

In moons smaller than $R \sim 1000$ km, the lesser ability to retain internal heat could be offset by the effect of antifreeze compounds such as ammonia in promoting the onset and persistence of subsurface oceans. However, simulations with the R = 421 km moon suggest that the presence of 2% ammonia does not appreciably enhance the longevity of an ocean, irrespective of whether the rocky core is consolidated or porous (i.e., allows hydrothermal circulation to better transfer core heat into the hydrosphere). In the latter case, much of the pore water in the core is liquid, irrespective of the occurrence of a disruptive collision (Supplementary Figure 10).

The results presented above hinge on major simplifying assumptions warranted by the coupling of SPH and thermal evolution models (see Methods). In SPH, undifferentiated rock-ice structures used for the 250-km impactor are approximated as pure serpentine, which has a higher bulk density than Saturnian and Uranian moons (equivalent to 30-35% more dry rock) and different mechanical properties. Since the impactor comprises 1/8th of particles, this can lead to a ~4% overestimated rock content in the reaccreted 500 km-class body. Temperature profiles at collision onset are truncated to 150 K and are necessary approximations representative of classes of moon sizes and collision times, although post-impact simulations initiated with temperature profiles that compensate for the pre-impact 150 K truncation show no major differences in outcomes (Supplementary Section 6). Material strength, ignored here, would likely result in less disruption (45), possibly yielding cooler reaccreted moons (Supplementary Section 5). However, the presence of a long-lived ocean appears much less (if at all) sensitive to the temperature at reaccretion than to longer-lived heat sources such as radioactive decay. Indeed, additional simulations performed with arbitrary variation of post-impact temperatures of the $R \sim 500$ km moon –up to 570 K– yielded no long-lived ocean, because higher temperatures generate more melt immediately post-impact and therefore more vigorous convection, faster heat loss, resulting in fast freezing in every case (Supplementary Figure 11). The manual rebinning of SPH outputs onto a radial grid for thermal evolution modeling warrants arbitrary choices, such as SPH simulation cutoff time and threshold porosity above which material compaction is forced. Finally, this handoff neglects any potential re-accretion of ejected debris. In the most violent collisions presented here, a few tens of percent of the impacting mass are ejected from the moon but still bound to the system. The ejecta will collide with each other and also the largest moon on timescales similar to their orbital period (tens of days).

Reaccretion of debris can result in 0.1 to 8 km of surface layering. The additional heat provided by the ejecta is likely to be negligible. First, the debris is collisionally evolving with itself and likely to grind to sizes small enough to effectively radiate heat before reaccreting. Second, it was found in test cases that even artificially increased surface temperatures at reaccretion did not alter outcomes due to effective convective transport through short-lived oceans.

The cases investigated here may be relevant to Saturn's moon Rhea, whose size ($R \approx 763$ km) is in between the small and large target moons simulated. Rhea's relaxed large craters indicate past heat fluxes of tens of mW m^{-2} (46), and its orbital inclination is an order of magnitude lower than it should be after its orbit expanded through the evection resonance (where Rhea's pericenter precession period matches that of Saturn's orbit around the Sun) at about 8.3 R_S. The current explanation for Rhea's low inclination is that a precursor moon had its inclination and eccentricity raised through the resonance, and was put on an orbital path toward a disruptive collision (4). Collisional debris dynamically evolved to a low inclination and reassembled into the Rhea we see today (5; 7). The framework of this study could test whether a reassembled Rhea would undergo enough heating to relax craters, differentiate, or possess an ocean. While computed near-surface heat fluxes do not exceed 5 mW m^{-2} for the post-impact 500-km class moons of Fig. 2-4 (panels c-d), they do reach 20 mW m^{-2} for the 1000-km class moon (Fig. 5c-d) at the time of ocean freezing, in the absence of significant tidal dissipation.

Future work could focus on modeling the effect of such collisions on smaller moons on closer-in orbits, analogous to Mimas, Enceladus, and Miranda, all of which display evidence of a subsurface ocean and/or high geologic activity, past or present (e.g., 47; 48; 23; 49). The framework presented in this paper could elucidate whether such activity could have been kick-started, following a disruptive collision, by reaccretion of a warm ice-rich shell susceptible to generating high levels of tidal dissipation.

METHODS

We investigate the extent to which a moon's interior evolves differently due to a large impact than it would have without the impact. This process involves modeling the pre-impact interior evolution of progenitors from initial assembly, 2) the impact and reassembly, 3) the post-impact interior evolution of the moon given the extra energy imposed by the impact and reassembly, and 4) the interior evolution of a moon from initial assembly to the present-day structure without an impact, which serves as the control case. We employ different modeling toolkits for the disruption/reassembly and the interior thermal and physical evolution, leading to a series of discrete modeling steps (Fig. 1).

NOTIONAL PRE-IMPACT INTERIOR EVOLUTION OF THE TARGET AND IMPACTOR

To inform the thermal state just before the impact, for use as initial condition of SPH simulations, the thermal and structural evolution of icy moon impact progenitors with radii 250, 500, and 1000 km is computed using the `IcyDwarf` software (34; 28).

During the pre-impact evolution, we ignore accretion of additional material on top of the moon (i.e., we assume it is fully formed at the start of a simulation). We initially assume a uniform interior temperature of $T_0 = 100$ K due to leftover accretional heating. Adopting nonuniform temperature profiles more consistent with accretional heating seldom affects the thermal evolution (Supplementary Section 1; Supplementary Figure 1). Given a bulk density of 1500 kg m^{-3}, ice and "rock", which we define as refractory material assumed to be mainly silicate and metal minerals as well as organic material, are initially distributed homogeneously on a one-dimensional, spherically symmetric grid. Proportions of rock and ice in each grid zone are calculated based on respective material densities of 3800 kg m^{-3} and 935 kg m^{-3}. Unlike post- and no-collision simulations, here ice is assumed to contain 2% ammonia.

Temperature changes are computed following the balance of i) heating by decay of long-lived radionuclides (abundances and decay parameters provided in Supplementary Table 1), tidal dissipation, and to a lesser extent, ice-rock differentiation and rock hydration, and ii) cooling due to conduction or convection. Convection can occur in solid ice, liquid water, and porous rocky core layers if the relevant critical Rayleigh number is exceeded. The surface temperature is assumed to remain fixed at 80 K, corresponding to a bolometric albedo of ≈0.1 in a Uranian system that formed at a closer heliocentric distance (10 AU) than today (50) with the early Sun at 70% of its current luminosity. This surface temperature is higher than for post- and no-collision simulations (see *Post-impact interior evolution* section and Extended Data Table 1); the difference can represent both an increase in albedo induced by the collision, which typically enriches the surface in bright ice and depletes it of darker rock or dust infall, and an increase in heliocentric distances during periods of gravitational interactions between giant planet systems and remnant planetesimals that made disruptive collisions more likely (13; 3). A detailed description of the model and implementation of each heating and cooling process is provided in ref. (34). The temperature profile after 200 Myr is used as the SPH input temperature profile.

We perform the same analysis on the impacting body, which is assumed to be the smaller of the two colliding bodies. The likely differentiation state of the impactor at the time of collision depends on its size. In an undifferentiated body, rock and ice are assumed to separate if a grid zone i reaches the temperature at which a liquid water-ammonia mixture first melts ($T_i \geq 176$ K). If the body is already differentiated out to R/2 (providing a reasonably flat interface at km scales) differentiation can take place by Rayleigh-Taylor instability in grid zones warmer than $T_i \geq 140$ K, i.e., where ice is sufficiently ductile (51). These assumptions result in the modeled $R = 250$ km pre-impact body remaining homogeneous, whereas the $R = 500$ km body differentiates within 100 Myr. Focusing on differentiated R = 500 km targets enables higher-fidelity SPH modeling of interior structures than undifferentiated bodies approximated as serpentine (see Discussion). However, full differentiation hinges on the early presence of ammonia and an 80 K surface, since otherwise a 500-km body with a 55 K surface would not be more than partially differentiated at any time in its history (Fig. 2b-3b).

COLLISION

Based on the above `IcyDwarf` pre-impact simulations, the $R \geq 500$ km objects are assumed to be differentiated at the time of impact t_{coll} and the R = 250 km object is assumed undifferentiated.

The radial profiles of temperature T_i from `IcyDwarf`, which comprise 100 grid zones, are converted to an SPH input as shown in Extended Data Figure 1a. The SPH code's equation of state (EOS) forces a minimum temperature of 150 K to ensure a non-negative pressure for the particles on the surface, which have fewer neighboring particles (hence artificially low densities). This minimum temperature is above that of the outer grid zones at t_{coll}. However, this difference in temperatures does not change the dynamical outcome of the impact, and correcting for it post-collision seldom affects the subsequent thermal evolution (Supplementary Section 6).

We use a modified version of the `Gadget2` SPH code (52) for the impact simulations. SPH represents material as spherically symmetric particles, with the spatial distribution of each particle defined by a spline density weighting function, known as the kernel, and a characteristic radius, known as the smoothing length. The kinematic state of each particle evolves due to gravity, compressional heating/expansional cooling, and shock dissipation. Material strength is not considered; the consequences of this simplification are discussed in Supplementary Section 5. This version uses a tabulated EOS (53; 54; 55) to represent the impacting material: 1) the semi-analytical EOS, M-ANEOS for forsterite (56) to represent the dry cores, 2) the H_2O five-phase EOS for ice (57) to represent the icy mantles, and 3) the serpentine EOS (58) to represent a hydrated rock (59) that did not undergo differentiation. The differentiated bodies contain 33% rock and 67% ice by mass, which replicates Rhea's structure (60). The lesser compositional detail, relative to `IcyDwarf`, mandated by the need for detailed material EOS in SPH modeling results in bulk densities (2500 kg m^{-3} for undifferentiated and 1300 kg m^{-3} for differentiated bodies) which are necessarily different from those of the pre-impact interior evolution simulations. Hence, we consider the thermal profiles and differentiation states from these simulations to be notional; the corresponding detailed interior structures are not carried over to SPH.

We use the temperature profiles provided by `IcyDwarf` to construct bodies, then simulate them in isolation within SPH for 10 hours to ensure a stable structure prior to impact. The temperature is corrected and/or re-equilibrated, if needed.

The two impacting bodies are placed on an orbit around a Uranian mass planet such that their combined central mass has a coplanar circular orbit with a representative semi-major axis of $10\ R_U$ (61) (the major Uranian moons presently orbit at about 5 to 20 R_U). We place the bodies at a mutual distance of $2(R_1 + R_2)$, where R_1 and R_2 are their respective radii, to ensure that the tidal distortion occurring as the two bodies approach each other is captured by the SPH code. The tidal distortion modeled by this approach is overestimated compared to what it would be if material strength was accounted for (45); however, for these fast impacts, the bodies do not have time to fully adjust their shape to each other's gravitational potential. To determine the initial positions and velocities that yield a desired impact angle and relative velocity v_{col} at contact, we integrate the three-body gravitational interaction backward in time, starting from the moment of contact and continuing until the bodies reach a separation of $2(R_1 + R_2)$. To maximize the thermal increase and disruption of the body caused by collisions, we restrict the impact configurations to head-on collisions.

The collisions onto the 500-km target are all subsonic; that onto the 1000-km target is a supersonic collision in water. The collision velocities are sufficient to disrupt the target moons, yet within the bounds of previously investigated circumplanetary impacts (mostly < 4 km s^{-1} for impacts within a putative primordial system after Triton's capture; 20), noting that such velocities depend sensitively on the perturbation that destabilizes the satellite system.

The planet is represented by a small number of SPH particles, sufficient to describe its gravitational potential. We use a total of $\sim 5 \times 10^5$ SPH particles, and the simulations span 24 or 48 hours after impact, depending on the time required to obtain a stable final shape.

In these head-on impacts, only one large fragment remains, which we hypothesize to be the newly reformed moon. It is warmer than the pre-impact moon (Extended Data Figure 1b), and generally less massive, except for the case of impact at $1 \times V_{esc}$ (R = 500-km target moon). The spatial distribution of mass m_i, temperature T_i, H_2O mass fraction $f_{\mathrm{H_2O,i}}$, and hydrated rock (serpentine) mass fraction $f_{\mathrm{serp,i}}$ of material (the rest being assumed to be dry rock) at that time, are binned in spherical shells with a grid step of 30 km, comparable to the typical smoothing lengths of the SPH particles. This 30 km spacing ensures reasonably smooth thermal and compositional profiles by averaging the temperatures and compositions of a representative number of particles in each bin.

POST-IMPACT INTERIOR EVOLUTION

The `IcyDwarf` software is used again to simulate the post-impact thermal and structural evolution of moons. Its approximation of spherical symmetry is justified because the SPH-simulated impacts are very disruptive, and lead to reaccretion of most of the target into a well-mixed moon with little to no temperature asymmetries (Fig. 6).

To convert the output of the SPH code into an input that can be read by `IcyDwarf` (Extended Data Figure 1c), first, the ice and rock masses in each radial grid zone, $M_{\mathrm{H_2O(s),i}}$ and $M_{rock,i}$, as well as an index of hydration $X_{hydr,i}$ (volume fraction of the rock that is hydrated; 62), are calculated from SPH simulation outputs as:

$$M_{\mathrm{H_2O(s),i}} = m_i \, f_{\mathrm{H_2O,i}} \tag{2}$$

$$M_{rock,i} = m_i \, (1 - f_{\mathrm{H_2O,i}}) \tag{3}$$

$$X_{hydr,i} = \frac{f_{serp,i}/\rho_{hydr}}{(1 - f_{\mathrm{H_2O,i}} - f_{serp,i})/\rho_{dry} \; + \; f_{serp,i}/\rho_{hydr}} \tag{4}$$

Unlike in the simulations used to obtain the notional pre-impact temperature profiles, here the ice is assumed to be pure H_2O (requiring $T \geq 273$ K for melting and onset of ice-rock separation in grid zones at $< R/2$), and the densities of dry and hydrated rock are chosen to match those of the EOS used in the SPH simulation; respectively, 3320 kg m^{-3} and 2500 kg m^{-3}. In addition to maximizing fidelity to the SPH output, assuming pure water ice separates out the effect of ammonia on ocean promotion (see Discussion section).

Next, a porosity in each grid zone based on the SPH code binning, $\phi_{SPH,i}$, is obtained from the ratio of the volume of material in each grid shell to that of the grid shell itself:

$$\phi_{SPH,i} = 1 - \frac{M_{rock,i}/[X_{hydr,i}\ \rho_{hydr} + (1 - X_{hydr,i})\ \rho_{dry}]\ +\ M_{\mathrm{H_2O(s),i}}/\rho_{ice}}{(4/3)\ \pi\ (r_i^3 - r_{i-1}^3)} \tag{5}$$

In the innermost grid zones, $\phi_{SPH,i}$ is on the order of a few percent, but in the outer grid zones, $\phi_{SPH,i}$ can be greater than 99%. Such high values reflect the lack of full consolidation of material at the end of the SPH simulation (see inset of Extended Data Figure 1b); i.e., grains of material at such porosities are very unlikely to be adjoined. This occurs because the SPH technique adjusts the smoothing length of each particle such that the number of neighboring particles is maintained $\sim$ constant, thus the surface particles have artificially low densities. In addition to being likely aphysical for a solid planetary body, such porosities are also too high for `IcyDwarf` to handle, as they result in exceedingly low thermal conductivities and, therefore, spuriously high temperatures in the overly insulated grid zones below. At such high porosities, heat transfer should instead occur primarily by radiation (63), leading to material temperatures closer to T_{surf}.

To remedy this issue, we re-bin the SPH radial mass distribution to ensure that the grid zone porosity $\phi_i \leq \phi_{SPH,i}$ does not exceed 0.7, the threshold above which we assume grains are no longer adjoined. If $\phi_{SPH,i} \geq 0.7$ in a grid zone, its outer radius r_i is recalculated as:

$$r_i = \left[r_{i-1}^3\ +\ (r_i^3 - r_{i-1}^3)\ \frac{1 - \phi_{SPH,i}}{1 - 0.7} \right]^{1/3} \tag{6}$$

This necessitates starting at the center and going toward the surface when checking and, as needed, changing grid zone radii. This compaction step results in an overall lower moon radius than implied by the SPH output (Extended Data Figure 1c). Subsequent porosity compaction is computed with `IcyDwarf` and depends on temperature and assumed material rheological flow laws (34). This limitation of porosity to 70% effectively speeds up reaccretion, without changing the predictable outcome of these gravitationally bound particles ending up contributing to object growth and final size.

For the `IcyDwarf` profile, the grid is made finer than the 30 km spacing of the SPH output in order to satisfy numerical convergence. For the majority of bins, starting at the center, we interpolate 8 to 10 grid zones (spherical shells) within each bin, holding grid zone temperature and porosity at the bin's values, and distributing the masses of rock and water so that the composition of each set of grid zones matches that of the corresponding SPH bin. For the outermost bins, whose high porosity results in very thin spherical shells after compaction, we interpolated only 1 to 5 grid zones, and in some cases merged 2 to 4 bins into a single grid zone. In those merger cases, we assigned to the grid zone the sum of the rock and water masses of the merged bins (thereby conserving mass), and the mass-weighted average of the porosity and temperature of the comprising bins (thereby nearly conserving energy, with insignificant differences due to compositional heterogeneity across the merged shells). The surface temperature is set to T_{surf} = 55 K, the equilibrium temperature for spherical objects with albedo 0.4 at 20 AU, the present-day heliocentric distance of the Uranian system. With this surface temperature, moon ice shells experience less ice-rock separation ($R \sim 500$ km moons) and shorter post-impact, whole-hydrosphere melting ($R \sim 1000$ km moon) than if the surface stays at 80 K (Supplementary Section 7, Supplementary Figure 8).

To obtain a control on the effect of the collision, we run no-collision simulations of moons with the same masses and densities as the reaccreted moons. For these simulations, we assume formation 10 Myr after CAIs, an initial homogeneous structure, void porosity of 0.3, and temperature of 100 K; we assume other (e.g., orbital) parameters as for the post-collision simulations (Extended Data Table 1). In particular, the orbit is assumed to have a low initial eccentricity of ≤ 0.02, found to be a representative result of gravitational interactions with a primordial circumplanetary disk (64; 65), and unlikely to be substantially modified by the collision (see Discussion).

DATA AVAILABILITY

Simulation inputs and outputs underlying the results presented in this article are available (66) along with data underlying all figures, except for intermediate panels of Figure 6 and Supplementary Figures 3-5.

CODE AVAILABILITY

`IcyDwarf` simulations were carried out with v24.7.2, available at github.com/MarcNeveu/IcyDwarf. This specific version is also archived together with simulation data (66). The modified version of the `Gadget2` SPH model and EOS tables are available in the supporting information of ref. (55).

CORRESPONDENCE

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ACKNOWLEDGMENTS

This manuscript was substantially improved thanks to comments from three anonymous reviewers.

AUTHOR CONTRIBUTIONS

AR, MN, RR, and KW planned the project. RR and YS performed SPH simulations and data analysis. MN performed thermal evolution simulations and data analysis. AR, MN, RR, and KW interpreted the results. MN drafted the manuscript, which all authors edited.

COMPETING INTERESTS

The authors declare no competing interests.

FUNDING STATEMENT

The authors disclose support for the research and publication of this work from NASA's Habitable Worlds program [grant number 80NSSC22K0403]. R.R. was additionally supported by a NASA Hubble Fellowship [grant number HST-HF2-51491]. M.N. was also supported by NASA's Goddard Space Flight Center [grant number 80GSFC24M0006].

FIGURE LEGENDS

Figure 1. Schematic of the processes investigated and overall modeling approach adopted in this work. We first model the pre-impact interior of the moon (see Methods). Using the thermal profile and ice-rock differentiation state (here ice, rock, ice-rock mix, and ocean are shown respectively in sky blue, brown, gray, and darker blue), we then model a disruptive collision onto the moon and its immediate reassembly. We then use the post-collision temperature and compositional profiles as inputs to model the long-term interior evolution. We compare the resulting interior to a moon of the same size as the reassembled moon, but that has not experienced a collision.

Figure 2. Thermal and structural evolution of a moon of radius R = 527 km and bulk density ρ = 1373 kg m^{-3}. This body results from the head-on collision of a differentiated $R = 500$ km moon with a homogeneous $R = 250$ km moon at $1 \times V_{esc}$. **(a)** Temperatures and **(b)** interior structure as a function of radius and time since Solar System formation, if the moon does not experience a disruptive collision. **(c,d)** Post-collision thermal and structural evolution for a collision 0.5 Gyr into Solar System history. **(e,f)** Same as in (c,d), but for a late collision 1.5 Gyr after Solar System formation. The insets in panels (e,f) show details of the thermal and structural evolution during the first 10 Myr post-reaccretion. That detail is not shown in panels (c,d); the evolution in those 10 Myr is indistinguishable from that shown in panels (e,f).

Figure 3. Thermal and structural evolution of a moon of radius R = 421 km and bulk density ρ = 1553 kg m^{-3}. This body results from the head-on collision of a $R = 500$ km moon with a $R = 250$ km moon at $2 \times V_{esc}$. **(a)** Temperatures and **(b)** interior structure as a function of radius and time since Solar System formation, if the moon does not experience a disruptive collision. **(c,d)** Post-collision thermal and structural evolution for a collision 0.5 Gyr into Solar System history. **(e,f)** Same as in (c,d), but for a late collision 1.5 Gyr after Solar System formation. The insets in panels (e,f) show details of the thermal and structural evolution during the first 10 Myr post-reaccretion. That detail is not shown in panels (c,d); the evolution in those 10 Myr is indistinguishable from that shown in panels (e,f). The impact strips part of the mantle, resulting in a $R = 421$ km moon. In comparison, a collision-free moon of the same size (panel a) does not fully differentiate. This suggests that moons which are currently too small to undergo differentiation may have formed from the collision of a larger, previously differentiated body.

Figure 4. Thermal and structural evolution of a moon of radius R = 364 km and bulk density ρ = 1481 kg m^{-3}. This body results from the head-on collision of a $R = 500$ km moon with a $R = 250$ km moon at $3 \times V_{esc}$. **(a)** Temperatures and **(b)** interior structure as a function of radius and time since Solar System formation, if the moon does not experience a disruptive collision. **(c,d)** Post-collision thermal and structural evolution for a collision 0.5 Gyr into Solar System history. **(e,f)** Same as in (c,d), but for a late collision 1.5 Gyr after Solar System formation. The insets in panels (e,f) show details of the thermal and structural evolution during the first 10 Myr post-reaccretion. That detail is not shown in panels (c,d); the evolution in those 10 Myr is indistinguishable from that shown in panels (e,f).

Figure 5. Thermal and structural evolution of a moon of radius R = 902 km and bulk density ρ = 1368 kg m^{-3}. This body results from the head-on collision of a $R = 1000$ km moon with a $R = 500$ km moon at $3 \times V_{esc}$. **(a)** Temperatures and **(b)** interior structure as a function of radius and time since Solar System formation, if the moon does not experience a disruptive collision. **(c,d)** Post-collision thermal and structural evolution for a collision 0.5 Gyr into Solar System history. **(e,f)** Same as in (c,d), but a low radionuclide content is assumed, corresponding to 4.5 Gyr of decay. A radionuclide content corresponding to 50 Gyr of decay yields nearly identical results. The insets in panels (e,f) show details of the thermal and structural evolution during the first 10 Myr post-reaccretion. That detail is not shown in panels (c,d); the evolution in those 10 Myr is indistinguishable from that shown in panels (e,f).

Figure 6. SPH simulation of a head-on collision between a differentiated object of radius 500 km and a homogeneous object of radius 250 km at $2 \times V_{esc}$. Panels show the 2-dimensional spatial distribution (scale bar: 1000 km) and temperature of water ice at 0.4h, 1.4h, 2.3h, 5.8h, 29h, and 48h after simulation onset. Dry and wet rock is shown in gray. All projections are on the equatorial plane with one hemisphere removed.

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