Modeling global impact effects on middle-sized icy bodies: applications to Saturn’s moons

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Abstract

Disrupted terrains that form as a consequence of giant impacts may help constrain the internal structures of planets, asteroids, comets and satellites. As shock waves and powerful seismic stress waves propagate through a body, they interact with the internal structure in ways that may leave a characteristic impression upon the surface. Variations in peak surface velocity and tensile stress, related to landform degradation and surface rupture, may be controlled by variations in core size, shape and density. Caloris Basin on Mercury and Imbrium Basin on the Moon have disturbed terrain at their antipodes, where focusing is most intense for an approximately symmetric spheroid. Although, the icy saturnian satellites Tethys, Mimas, and Rhea possess giant impact structures, it is not clear whether these structures have correlated disrupted terrains, antipodal or elsewhere. In anticipation of high-resolution imagery from Cassini, we investigate antipodal focusing during giant impacts using a 3D SPH impact model. We first investigate giant impacts into a fiducial 1000 km diameter icy satellite with a variety of core radii and compositions. We find that antipodal disruption depends more on core radius than on core density, suggesting that core geometry may express a surface signature in global impacts on partially differentiated targets. We model Tethys, Mimas, and Rhea according to their image-derived shapes (triaxial for Tethys and Mimas and spherical for Rhea), varying core radii and densities to give the proper bulk densities. Tethys shows greater antipodal values of peak surface velocity and peak surface tensile stress, indicating more surface damage, than either Mimas or Rhea. Results for antipodal and global fragmentation and terrain rupture are inconclusive, with the hydrocode itself producing global disruption at the limits of model resolution but with peak fracture stresses never exceeding the strength of laboratory ice.

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1. Introduction

The formation of our Solar System has been, and continues to be, one of the pinnacle problems in the field of planetary science. Because regular satellite systems around gas giant planets may originate and evolve in a manner closely analogous to planet formation, some researchers have sought information from these systems to improve understanding of our own Solar System origin (e.g., Stevenson et al., 1986; Canup and Ward, 2002; Mosqueira and Estrada, 2003). A key step in understanding the origin of satellite systems is to understand the internal structures of the satellites.

The interiors of satellites hold insights into the process of differentiation. Longer formation times produce less accretional heat from impacts which leads to less significant differentiation (Canup and Ward, 2002). Therefore, variations in the extent of differentiation can speak to conditions at the time of formation. For instance, according to moment of inertia data, Callisto is the only galilean satellite not completely differentiated (Anderson et al., 2001; Sohl et al., 2002). Some speculate that this is due to differences in the formation time of each galilean satellite (Canup and Ward, 2002), while others believe differences in tidal heating are responsible (Anderson et al., 1998).

Most of the current knowledge about the internal structures of the icy satellites of the outer Solar System is based on moment of inertia data and mass estimates for each satellite (cf. Dermott and Thomas, 1988; Thomas and Dermott, 1991; Anderson et al., 2003). These quantities, for all but the galilean satellites, are only loosely constrained until spacecraft flybys allow better parameterization. However, even with more accurate values for mass and moment of inertia, we still must make assumptions about the internal structure such as the size and density of the core. Investigating a satel-
litho’s cratering record may provide independent clues to its interior.

The internal structure of Earth was not known until the advent of seismology. The extensive global seismic network on Earth has led to many discoveries such as the fluid outer core (Oldham, 1906; Jeffreys, 1926; Lehmann, 1936) and discontinuities within the mantle (Engdahl and Flinn, 1969). The internal structure of the Moon was only realized following the deployment of seismometers during Apollo missions from 1969 to 1972 (Goins et al., 1981). Seismic networks have even been contemplated for Mars (Lognonne et al., 2000). However, for most planetary bodies, especially small satellites in the outer Solar System, it is not yet feasible to use seismology to determine internal structure, and we must consider other means of studying their interior structures. The method considered here is to examine global effects from large cratering events, beginning with a general examination of how internal structure affects the outcome (such as seismic shaking of the antipode, or the point opposite to an impact) and concluding with a specific analysis of three of the middle-sized icy satellites of Saturn.

Large impacts are relatively well-understood phenomena that have occurred on all satellites in the Solar System. For a number of planetary bodies, large impacts have been invoked to explain observed antipodal terrain disruption (e.g., Schultz and Gault, 1975; Moore and Ahern, 1983). Because an impact generates both internal and surface seismic waves that are capable of leaving widespread surface expressions, the modeling of large impact events and the analysis of their global effects may be used as a proxy for seismic data to determine the internal structures of impacted bodies (Chapman and McKinnon, 1986; Watts et al., 1991; Asphaug and Melosh, 1993; Asphaug et al., 2002).

In this paper we focus on impact events that may be large enough to cause observable damage at the antipode. On Earth, large impacts have been hypothesized to cause flood basalt and hotspot activity (Rampino and Stothers, 1988; Hagstrum and Turin, 1991; Rampino and Caldeira, 1992) through the focusing of seismic energy (Boslough and Chael, 1994). For a large terrestrial planet like Earth where distances are great, this process may only apply in the largest impacts, despite the low attenuation of surface wave energy (Boslough and Chael, 1994). For smaller bodies, antipodal disruption appears to be a part of the geologic record. Disrupted antipodal terrains have been identified opposite to Caloris Basin on Mercury (Murray et al., 1974), Imbrium Basin on the Moon (Wilhems and McCauley, 1971) and Stickney on Phobos (Fujiwara, 1991). Because seismic waves reflect and refract differently in different media, we may be able to use the degree of surface damage at the antipode to determine the size and density of a body’s core, if one exists. The first step in understanding this process is to accurately model the generation of impact shock energy, its conversion to seismic energy (surface and body waves) and the propagation of this energy globally, including focusing effects at the antipode.

Saturn’s satellites Tethys (DT = 1048 km), Mimas (DM = 394 km) and Rhea (DR = 1528 km) have giant impact structures Odysseus (D = 400 km; DT/DT = 0.38), Herschel (D = 135 km, D/DM = 0.34) and provisionally named Tirawa (D = 350 km, D/DR = 0.23), as shown in Fig. 1. The antipode of Odysseus on Tethys is a relatively smooth planar region. Ithaca Chasma, a trough system that extends some 270° around Tethys, is hypothesized to be the result of the impact event that formed Odysseus because they are of similar age (Moore and Ahern, 1983) and because Ithaca conforms closely to the impact equator (Moore et al., submitted for publication). That is, instead of being the result of antipodal disruption, Ithaca Chasma may be the result of whole body strain circumferential to the impact axis (Moore and Ahern, 1983) or reflection of stress channeling from a relatively shallow core-mantle boundary.

On Mimas, antipodal disruption from Herschel is not detectable due to low resolution in the Voyager images (Schenk, 1989). There is, however, some evidence for global scale fracturing (McKinnon, 1985). The antipode of Tirawa on Rhea is a region of rolling hills, which is consistent with impact degraded terrain (e.g., Schultz and Gault, 1975).

The highest quality images of both Tethys and Mimas have resolutions of 2 km/line-pair from the Voyager spacecrafts, resulting in fewer pixels across the much smaller Mimas. Although the highest quality images of Rhea have resolutions of ~1 km/line-pair, from the Voyager 1 spacecraft, they do not depict Tirawa clearly (Smith et al., 1981, 1982).

Building on the work of Hughes et al. (1977) and Schultz and Gault (1975), Watts et al. (1991) modeled antipodal pressures and surface accelerations caused by large impacts into several medium-sized bodies in the Solar System, including Tethys, Mimas, and Rhea. Using the Los Alamos Simplified Arbitrary Lagrangian Eulerian (SALE) 2D hydrocode (Amsden et al., 1980), Watts et al. (1991) found that the modeled surface accelerations and pressures were greater than the assumed surface material strength in most cases; therefore, giant impacts were sufficient to account for the observed disrupted antipodal terrain. However, many have been skeptical of these conclusions because a 2D axisym-
metry brings all wave energy to a perfect focus at the antipode (Boslough and Chael, 1994). Fujiwara (1991) found evidence for very different kinds of focusing in elastic ellipsoids. In this study, he solved the exact elastic equations and derived surface stress states, which he correlated with expressions of antipodal fracture manifest on Phobos from the Stickney impact. In particular, he observed stress trajectories consistent with the locus of fracture grooves on Phobos, for a particular value of the Poisson coefficient. Even for nearly spherical bodies such as Mimas and Tethys, whose major and minor diameters differ by only a few tens of kilometers, wave arrivals for a ∼ 10 km impactor are typically going to be out of phase at the antipode of a given impact and in phase where axisymmetric modeling would not predict. It is therefore critical to model giant impacts and their effects using a 3D code capable of simulating non-spherical bodies and non-symmetric interior structures.

2. Method

We modeled giant impact events on a fiducial middle-sized icy test satellite, as well as on Tethys, Mimas, and Rhea, using a 3D smooth particle hydrodynamics (SPH) code (Benz and Asphaug, 1995). We first explore how core size and density affect the focusing of seismic energy from an impact. Ultimately of interest is whether, and to what degree, large-scale features like those described above for Tethys, Mimas, and Rhea may be correlated to giant impact events and what such correlation implies for each satellite’s interior.

2.1. SPH

SPH is a three-dimensional gridless Lagrangian algorithm capable of incorporating realistic shape models (cf. Asphaug et al., 1998). (For a detailed review of the theory behind hydrodynamics codes, see Benz (1991); and for a detailed review of the SPH method, see Benz, 1990; Monaghan, 1992.) Our version of SPH has a unique fracture implementation (Benz and Asphaug, 1994, 1995), making it ideal for modeling geologic surface damage in solid body collisions. Although SPH is capable of using various equations of state, only a limited number of these are available for realistic icy satellite material. Appropriate liquid water models exist (cf. Gisler et al., 2003) but for Solar System ices, including water ice, good shock equations of state, including associated fracture and fragmentation and expansion of shock-generated vapor. (See Section 2.4 for a description of how SPH handles shock.)

Our models consisted of a target body and an impactor, as described below, undergoing hypervelocity collision. To investigate the global impact effects in our simulations, we calculated the peak surface velocity (as a proxy for mechanical devastation of a landscape) and peak surface tensile stress (as a proxy for surface rupture) for a particular location during an entire simulation. Direct calculation of fracture damage (as per Melosh et al., 1992; Asphaug et al., 1996) is possible but not particularly meaningful when shock resolution is > 10 kilometers at best and when fragmentation depends on extrapolation of the ice fracture constants, which are poorly established even for laboratory ices (see Table I in Asphaug et al., 2002), resulting in fracture stress extrapolations (size dependent strengths) that are not meaningfully constrained. For this reason we implement the fragmentation algorithm of Benz and Asphaug (1995) but note that for the chosen fracture constants, all of the impacts modeled here result in total damage of the target lithosphere by the end of the calculation. Effectively this means that cracks permeate the lithosphere at the model resolution (∼ 10 km). Because of the uncertainty in the ice fracture constants, when extrapolated from laboratory scales to satellite scales, and to the much lower strain rates associated with large-scale impact fragmentation, the degree to which the target lithosphere actually disrupts is better answered by examining the magnitude of the peak tensile stress and peak surface velocity during the course of the calculation.

Peak surface velocity indicates the magnitude of seismic shaking and the degree of terrain degradation. For degradation of terrains held together by gravity (deeply faulted mountains, ejecta deposits) the magnitude of degradation (expressed in terms of distance thrown, normalized to target radius) scales with $v_{\text{ej}}/v_{\text{esc}}$, where $v_{\text{ej}}$ is the peak surface velocity and $v_{\text{esc}}$ is the escape velocity at the surface.

Peak surface tensile stress is a measure of material failure, faulting and spallation. Given the evidence (e.g., Hoppa et al., 1999), however controversial, that ice lithospheres are very weak in their response to diurnal strains, we can only qualitatively apply our model results at present as an indicator of the propensity for surface rupture. In most cases, our peak tensile stresses are greater than the very low (∼ $10^4$ dyne cm$^{-2}$) stresses required by Hoppa et al. (1999) but strictly lower (though not always by much) than the stresses required to fracture laboratory ice (∼ $10^7$ dyne cm$^{-2}$; Hawkes and Mellor, 1972; Lange and Ahrens, 1983). The SPH code shows fracture damage at the modeled resolution owing to the rate- and size-dependence of strength (Housen and Holsapple, 1999), which remains poorly quantified for geologic ices. The use of a damage rheology in the simulations provides a more realistic response to impact than allowing the bodies to ring like elastic bells, although future models in the post-Cassini era will use what is learned about the detailed effects of these impacts in order to derive damage models that coincide with observations.
2.2. Input parameters for test satellite impact scenario

Our fiducial target is a 1000 km diameter differentiated sphere, with a mantle composed of pure water ice ($\rho = 0.917 \text{ g cm}^{-3}$) overlying a silicate core. The size and density of the core are free parameters within this model. We ran nine simulations, permuting the size and density of the test satellite’s core. Three simulations use a baseline core density of basalt ($\rho = 2.7 \text{ g cm}^{-3}$), three simulations use a low-density ($\rho = 2.0 \text{ g cm}^{-3}$) silicate core and three simulations use a high-density ($\rho = 3.97 \text{ g cm}^{-3}$) silicate core. For high-density silicate core, we use the equation of state for gabbroic anorthosite as computed from laboratory experiments by Ahrens and O’Keefe (1977). For the low-density silicate core, we use the equation of state of basalt (Ahrens and O’Keefe, 1977) but substituting 2.0 g cm$^{-3}$ as the zero-pressure reference density. For all three silicate types, rheology is equivalent to that used by Benz and Asphaug (1995) in their modeling of laboratory disruption experiments by Nakamura and Fujimura (1991). As mentioned in the preceding section, the materials we have chosen to use for these simulations are limited by the available equations of state.

The impactor is a 13 km diameter undifferentiated spheroid, representing a comet composed of pure water ice. In all cases the impact velocity is $20 \text{ km s}^{-1}$ and the impact angle is $0^\circ$.

2.3. Input parameters for saturnian satellite impact scenarios

Tethys, Mimas and Rhea have normalized dimensionless moments of inertia ($\gamma = I/(MR^2)$) less than 0.4 (Dermott and Thomas, 1988; Thomas and Dermott, 1991; Anderson et al., 2003), indicating they are differentiated. All of our model explorations of these satellites thus also consist of a differentiated target body composed of a pure water ice mantle and a silicate core. We represent Tethys and Mimas with triaxial ellipsoidal shapes. For Tethys, principal axes are $a = 535.9 \text{ km}$, $b = 527.5 \text{ km}$, and $c = 526.3 \text{ km}$ (Thomas and Dermott, 1991). For Mimas, principal axes are $a = 210.3 \text{ km}$, $b = 197.4 \text{ km}$, and $c = 192.6 \text{ km}$ (Dermott and Thomas, 1988). We represent Rhea as a spheroid with radius $r = 764 \text{ km}$ (Davies and Katayama, 1983). (See Table 1 for more properties of these satellites.)

The core size and density and impact angle varied over nine simulations for each satellite. The variety of core material is the same as that for the test satellite runs. However, the core size in each of the satellites’ simulations is determined by holding the bulk density of each satellite constant for all runs. All impacts are modeled in the true geodetic frame in order to facilitate application of the satellite 3D shape and for future use in evolving the ultimate fate of the Saturn-bound ejecta.

The diameter of the cometary impactor in each case ($d_i = 12.7 \text{ km to form Odysseus}, d_i = 2.9 \text{ km to form Herschel}, d_i = 10.9 \text{ km to form Tirawa}$) was calculated using gravity scaling for water ice (cf. Melosh, 1989) and assumed transient crater diameters $d_c = 246, 95, and 198 \text{ km}$, respectively (Moore et al., submitted for publication; W.B. McKinnon, private communication, 2003). Impact velocity in all cases was $20 \text{ km s}^{-1}$ and impact angle was allowed to vary between $0^\circ$ and $80^\circ$, measured from the surface normal, primarily in order to break any axisymmetry that might result from an unlikely vertical impact. We introduce the impactors in true geometry relative to the satellite geoid; on Tethys the impact occurred at the location of Odysseus, $30^\circ N$, $130^\circ W$; on Mimas the impact occurred at the location of Herschel, $11^\circ N$, $115^\circ W$; and on Rhea the impact occurred at the location of Tirawa, $36^\circ N$, $150^\circ W$ (Burns and Matthews, 1986, Map Section).

2.4. Model resolution

In all our model runs, we used $\sim 141,300$ particles for the target body and $700$ particles for the impactor, far beyond the requirement for resolution convergence established for catastrophic disruption of spherical targets (Benz and Asphaug, 1995) or for the impact origin of the Moon (Canup and Asphaug, 2001) but not sufficient to meaningfully resolve discrete fragments in the satellite lithosphere. While fracture damage is an important part of the satellite response to impact, including how it propagates and limits stress energy, at the modeled resolution about six geographical degrees (three particles) would define the smallest resolved discrete fragment plus its boundary, even if artificial viscosity (the means of modeling shocks in a hydrocode; Von Neumann and Richtmyer, 1950) did not smooth out the shocks, which lead to the powerful elastic stresses whose reverberations are ultimately responsible for fragmentation. Moreover, it is not very useful at present to extrapolate ice fracture statistics out to hundred-kilometer scales, when the strength scaling exponent $m$ remains poorly characterized (Table I in Asphaug et al., 2002).

<table>
<thead>
<tr>
<th>Satellite properties</th>
<th>Tethys</th>
<th>Mimas</th>
<th>Rhea</th>
</tr>
</thead>
<tbody>
<tr>
<td>$v_{esc}$ (m s$^{-1}$)$^a$</td>
<td>18.5</td>
<td>7.9</td>
<td>28.5</td>
</tr>
<tr>
<td>$D$ (km)</td>
<td>1048</td>
<td>394</td>
<td>1528</td>
</tr>
<tr>
<td>$D_{crater}$ (km)$^c$</td>
<td>400</td>
<td>135</td>
<td>350</td>
</tr>
</tbody>
</table>

$^a$ From Satellites (1986).
$^b$ Calculated from values found in Satellites (1986).
$^c$ Final crater diameters.
In short, stress propagation and its attenuation due to brittle fissuring of the lithosphere is poorly resolved. (Plastic yielding is included using the simple von Mises treatment which also reduces shear stresses to zero at the melting temperature; see Benz and Asphaug, 1995.) We thus focus on two model outcomes that are particularly robust: the peak particle velocity achieved at any time during the simulation (for particles within 5% of the target’s radius from the surface) and the peak tensile stress responsible for rupture. Future calculations on faster computers, with the availability of Cassini data, can usefully explore more detailed outcomes such as flexure, opening of grabens, sliding along thrust faults and peak acceleration. At present, maximum surface velocity is the most useful expression of time-integrated acceleration and of the maximum distance that surface debris can be expected to be thrown in a global-scale impact (cf. Asphaug and Melosh, 1993). Until fracture properties of these ices are better understood, we view maximum tensile stress as a good qualitative measure of the local degree of surface rupture.

For the Tethys impact scenario, and for the similarly sized test satellite, the SPH smoothing length of each particle in the target is ∼10 km. (Smoothing length is the width of the interpolation kernel used to express properties of the material, such as density, in terms of values at a set of disordered points or particles. For more information on smoothing lengths and kernels, see reviews by Benz, 1990, and Monaghan, 1992.) For the Mimas and Rhea impact scenario, the smoothing lengths of each particle in the target are ∼4 and ∼14 km, respectively. Particles belonging to the impactors have much smaller smoothing lengths, although the calculations are not sensitive to this choice since the impactor is mainly used to introduce impact energy and momentum. The principal limitation to these runs is the comparative lack of resolution in the target bodies. Runs utilizing a modified parallel version of the code would require ∼1000 times as many particles and almost 10,000 times as many cpu hours to attain a 10-fold enhancement in 3D model resolution.

3. Results

3.1. Test satellite

Peak velocity is highest at the impact site in all cases and decreases with distance from the impact with a slight increase at the antipode in two cases (see Fig. 2a, simulations IV and VIII). In these cases, the degree of focusing is small (no more than a factor of ∼2 increase at the antipode). The minimum and maximum peak velocities are similar in all cases. However, the average global peak velocity is highest for the cases with the largest core radius (see Fig. 2a, simulations III, VI, and IX). Varying the core density makes little difference in the peak velocity. For all test satellite simulations, the maximum value of peak surface velocity at the antipode is between 91 and 160 cm s⁻¹ (see Fig. 2a, simulation III).

Peak tensile stress is also highest at the impact site and decreases with distance from the impact with some antipodal increase in all cases. The degree of antipodal focusing in peak tensile stress is greater than for peak velocity. Peak tensile stress is enhanced by at least an order of magnitude in simulations using the largest core size (see Fig. 3a, simulations III, VI, and IX), while there is much less antipodal focusing for smaller cores. The dependence of global patterns in peak tensile stress on core radius is strong.

For all test satellite simulations, the maximum value of peak surface tensile stress is between 7 × 10⁵ and 1 × 10⁶ dyne cm⁻² (see Fig. 3a, e.g., simulations VI and VII). Although the global peak tensile stress is higher for smaller core sizes, the peak surface tensile stress never exceeds the rupture strength measured for ice in a laboratory, 1 × 10⁷ dyne cm⁻² (Hawkes and Mellor, 1972; Lange and Ahrens, 1983) except at the impact site itself. These stresses are, however, greater than the rupture strength (40 kPa or 4 × 10² dynes cm⁻²) required for Hoppa et al.’s (1999) Europa model of cycloidal ridge propagation by diurnal tides.

3.2. Rhea

Antipodal focusing is weak or non-existent in peak velocity (see Fig. 2c, e.g., simulations I and IV). Antipodal focusing in peak tensile stress is present in two cases and is enhanced by one order of magnitude at most (see Fig. 3c, simulations V and VIII). There are no clear trends of either peak velocity or peak tensile stress on impact angle or core properties.

For all Rhea simulations, the maximum value of peak surface velocity at the antipode is between 0.23 and 0.75 cm s⁻¹ (see Fig. 2c, simulations IV and VI). A velocity of 0.75 cm s⁻¹ is sufficient to throw material on Rhea a distance of only ∼0.1 mm, given g = 28.5 cm s⁻¹.

For all Rhea simulations, the maximum value of peak surface tensile stress at the antipode is between 6 × 10³ and 2 × 10⁴ dyne cm⁻² (see Fig. 3c, e.g., simulation V). These values of peak tensile stress do not exceed the rupture strength of laboratory ice and are less than the Hoppa et al. (1999) fracture threshold.

3.3. Mimas

There is slight peak velocity focusing at the antipode in some cases (see Fig. 2b, simulations II, III, and IV). Changing the core density and radius does not make a significant change in the global pattern of peak velocity. Antipodal peak velocity values decrease as the impact angle becomes more grazing. An asymmetric gradation is observed in nearly all cases (see Fig. 2b, e.g., simulation V). This asymmetry is likely due to the oblate nature of the shape model used for Mimas. For all Mimas simulations, the maximum value of peak surface velocity at the antipode is between 6 and
Fig. 2. Peak surface velocity for (a) test satellite, (b) Mimas, (c) Rhea, and (d) Tethys. Shown are cylindrical projections of each satellite’s surface. In all cases the impact occurs in the upper right-hand quadrant, with the antipode at the lower-left. Because the near-field target response is identical in all simulations of a given series, the area of impact in each series looks almost identical while the antipodal hemispheres vary with target structure and associated antipodal response. The white region in the center of each crater is empty space where particles were ejected from the crater and are, therefore, not included in our cylindrical projections. Mottled texture in these figures indicates either a boundary of two color-scale values or overprinting of one value above another (since several particles deep are being plotted as a surface). Both values could be present in these locations.

18 cm s$^{-1}$ (see Fig. 2a, simulation II, III, and IV). A surface velocity of 18 cm s$^{-1}$ is sufficient to throw material on Mimas a distance of only $\sim$ 20 cm, given $g = 7.9$ cm s$^{-1}$.

Antipodal focusing in peak tensile stress is weak or nonexistent in all cases. No clear trends are apparent for peak tensile stress dependence on varying core properties. As the impact angle increases, antipodal values of peak tensile stress decrease. For all Mimas simulations the maximum value for peak surface tensile stress at the antipode is between $4 \times 10^4$ and $1 \times 10^5$ dyne cm$^{-2}$ (see Fig. 3b, simulation I and III). This peak tensile stress does not exceed the rupture strength of laboratory ice and is lower than the tensile stress required by Hoppa et al. (1999) as a fracture threshold on Europa.

3.4. Tethys

Odysseus is 0.38 times the diameter of Tethys—larger, relative to the satellite diameter, than any other crater in the Saturn system. Antipodal focusing is observed in peak velocity for all cases. No clear trends exist for peak velocity dependence on either core properties or impact angle. The maximum value for peak surface velocity at the antipode is between 50 and 110 cm s$^{-1}$ (see Fig. 2d, e.g., simulations II and VI). With a surface gravity of 18.5 cm s$^{-1}$, a velocity of 110 cm s$^{-1}$ would throw loose materials up to $\sim 3.3$ m and severely degrade loosely consolidated terrains (such as crater rims) at scales of tens to hundreds of meters.

There is antipodal peak tensile stress focusing in some cases, with peak tensile stress enhancing by more than an order of magnitude in the strongest cases (see Fig. 3d, simulations I and IV). Varying core properties changes peak tensile stress more than varying the impact angle. The maximum value for peak surface tensile stress at the antipode is between $1 \times 10^6$ and $3 \times 10^6$ dyne cm$^{-2}$ (see Fig. 2d, simulation I). This would almost certainly result in ice rupture (see discussion section). The manifestation of this rupture awaits
Fig. 3. Peak surface tensile stress for (a) test satellite, (b) Mimas, (c) Rhea, and (d) Tethys. See caption of Fig. 2 for detailed explanation of plots.

higher resolution runs that can realistically express the fault geometry, linear extent and depth.

Although simulations of all the satellites we modeled indicate relatively weak antipodal focusing, Tethys shows both more antipodal focusing and greater values of peak surface velocity and peak surface tensile stress in the antipodal hemisphere than either Mimas or Rhea, as one might expect given its larger crater diameter relative to target diameter. An assessment of the observed surface geology coupled to the giant impacts on Tethys, Mimas and Rhea is forthcoming in Moore et al. (submitted for publication).

3.5. Ejecta

We estimate ejecta loss by considering how much mass reaches or exceeds escape velocity from each satellite surface and how much mass reaches or exceeds escape velocity for the Saturn system. Ejecta from all impact events on the saturnian satellites modeled stays in Saturn-bound orbit. Some ejecta from the Odysseus impact event on Tethys and the Herschel impact event on Mimas reaches each satellite’s respective escape velocity; however, no ejecta from the Tiritawa impact event on Rhea reaches escape velocity. More rigorous tracking of ejecta will be studied in a later paper.

3.6. Model time

All test satellite simulations were run for 800 s or three to four wave-crossing times. Simulations of the saturnian satellites were run for a minimum of two wave-crossing times or ~600 s for Tethys, ~250 s for Mimas and ~650 s for Rhea (see Fig. 4). In two wave-crossings the initial compressive (pressure) wave travels from the point of impact to the antipode and is reflected as a tensile wave, traveling back to the point of impact. Most of the surface damage occurs within two wave-crossing times, as the wave energy begins to attenuate after this; certainly the peak surface velocities have been achieved by this time.

4. Discussion

The results of modeling in 3D demonstrate relatively weak antipodal focusing of the peak surface velocity and
Our model results indicate a possible difference between the antipodal terrains of Caloris (Mercury) and Imbrium (the Moon) and those of Odysseus (Tethys) and Herschel (Mimas). Although the ratio of impact crater diameter to parent body diameter is less for Caloris and Imbrium than for Odysseus and Herschel, the antipodal terrains of Caloris and Imbrium have been plausibly correlated to giant impact events. Odysseus and its antipodal terrain have not been absolutely correlated on Tethys and, in fact, disturbed terrain antipodal to Herschel has not yet been identified on Mimas.

One possible reason for this discrepancy is the difference in material involved in the impact. Mercury and the Moon have silicate mantles, while Tethys, Mimas, and Rhea have icy mantles. The difference in mantle material may dictate how efficient an impact is at damaging the antipode. The composition of the impactors in all of these cases is unknown. In the saturnian system, the impactors are assumed to be comets; while for Mercury and the Moon the impactors could be either comets or asteroids. Perhaps the difference in impactor composition and velocity makes a difference in energy transmission to the target body during an impact. No researchers to date have modeled the formations of Caloris and Imbrium basins and the resulting global effects in 3D; therefore, the formation of their antipodal terrains is not well understood at present. In particular, the cause of the observed antipodal terrains (e.g., seismic shaking, tensile fracturing, etc.) is unknown. We plan to investigate this further in a future paper.

Besides the impact-induced shock wave, the reimpact of ejected materials (both prompt secondaries and those that go into temporary orbit about Saturn) may also explain the formation of disrupted antipodal terrain. For this reason we want to further investigate the trajectories of impact ejecta. Schultz and Gault (1975) showed that for large impact events on the Moon, surface (Rayleigh and Love) waves would not arrive at the antipode prior to the reimpact of secondaries. Although surface waves do more damage (due to their relatively low attenuation with distance and the constructive interference of the waves at the antipode), the damage due to the surface waves could not be distinguished from the damage due to reimpact of secondaries. Future simulations will include a ballistics calculation for ejecta, and we hope to investigate the arrival times of surface waves versus ejecta at the antipode.

Though we have concentrated on antipodal damage, it is possible for impact-induced terrains to appear elsewhere on a body. In fact, the most interesting feature on Tethys is Ithaca Chasma, which is not at the antipode to the Odysseus impact structure. In none of our simulations do the peak surface tensile stresses exceed the rupture strength of laboratory ice. However, fracture damage as computed by the Benz and Asphaug (1995) model, which uses the Weibull (1951) fracture coefficients to extrapolate to size dependent strengths, is global throughout the lithosphere since static strength decreases with roughly the cube root of size for masses of pure ice. Strictly speaking, a 1000 km diameter piece of ice should be hundreds of times weaker than a laboratory sample. It is obviously important to study in more detail any genetic correlation between Odysseus and Ithaca Chasma. To resolve geometry of fissuring around the impact’s equator, one needs to resolve surface elements to the extent that strain localization can become manifest—that is to say, so that the widespread lithospheric rupture seen in the present simulations is resolved into discrete crack propagation events that release stress in neighboring zones. At that point, one can explore which combination of ice fracture co-

Fig. 4. This time sequence follows a pressure wave traveling through the interior of our baseline (basalt core) Tethys simulation. The model sequence starts at \( t = 25 \) s and ends at \( t = 300 \) s, with snapshots every 25 s of model time. The Tethys basalt core case was chosen as an illustrative example because the ratio of core to mantle material and physical properties was such that this case had the longest wave-crossing time on any of the Tethys runs. Here we verify that 600 s is adequate time to follow two or more wave-crossings and the accompanying wave attenuation for this case. Because the simulations of Mimas and Rhea use identical materials with identical wave propagation speeds, 250 and 650 s, respectively, should be adequate time to follow at least two wave-crossings through these bodies.
efficients and interior structural models best fits the genesis of Ithaca Chasma.

Core shape may also affect the focusing of seismic energy. It is possible that this could not only explain the presence of Ithaca Chasma but its location, offset from the true antipode of Odysseus. In all of our simulations, each target body had a spherical core. However, Dermott and Thomas (1988) postulated that while Mimas has components of both ice and silicate, it may not be totally differentiated. It may be that Mimas has pockets of rock scattered throughout its interior rather than a single rocky core. Although the same has not been said for Tethys, a fragmented core is something we are able to model. Similarly, it may be possible to explain Ithaca Chasma by a variable porosity mantle, something that is also readily incorporated into hydrocodes.

5. Conclusion

The main result of this study is that, in multiple simulations, 3D modeling finds only weak evidence for antipodal focusing of seismic energy, in contrast to previous 2D calculations. In addition, for the test satellite simulations, antipodal damage is more dependent on core geometry (radius) than core density, indicating that incomplete or asymmetric differentiation can result in a unique response to large impact events. The simulations of Tethys, Mimas, and Rhea show that Tethys has larger values of peak surface velocity and peak surface tensile stress than either Mimas or Rhea, as one would expect for the larger crater diameter relative to target diameter. This indicates that Tethys is more damaged during its giant impact event than either Mimas or Rhea. Results for antipodal and global fragmentation and terrain rupture are inconclusive, with the hydrocode itself (using size- and rate-dependent strength scaling) producing global disruption at the limits of model resolution (tens of km) but with peak fracture stresses never exceeding the strength of laboratory ice. This problem awaits better characterization of the fracture properties of ice and much higher (10^7 particles or more) model resolution in 3D.

Antipodal focusing is not simple or straightforward. We need to better understand what magnitude of seismic shaking can account for a given degree of terrain degradation. We also need better constraints for the behavior of ice at the temperatures and pressures experienced during giant impact events on icy satellites in the outer Solar System. Detailed answers to these issues must await better images from Cassini, laboratory studies and future modeling.

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References


Global impact effects


