Impact Crater Morphology and the Structure of Europa’s Ice Shell

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Abstract

We performed numerical simulations of impact crater formation on Europa to infer the thickness and structure of its ice shell. The simulations were performed using iSALE to test both the conductive ice shell over ocean and the conductive lid over warm convective ice scenarios for a variety of conditions. The modeled crater depth-diameter is strongly dependent on the thermal gradient and temperature of the warm convective ice. Our results indicate that both a fully conductive (thin) shell and a conductive-convective (thick) shell can reproduce the observed crater depth-diameter and morphologies. For the conductive ice shell over ocean, the best fit is an approximately 8 km thick conductive ice shell. Depending on the temperature (255–265 K) and therefore strength of warm convective ice, the thickness of the conductive ice lid is estimated at 5–7 km. If central features within the crater, such as pits and domes, form during crater collapse, our simulations are in better agreement with the fully conductive shell (thin shell). If central features form well after the impact, however, our simulations suggest that a conductive-convective shell (thick shell) is more likely. Although our study does not provide a firm conclusion regarding the thickness of Europa’s ice shell, our work indicates that Valhalla class multiring basins on Europa may provide robust constraints on the thickness of Europa’s ice shell.

Plain Language Summary

Jupiter’s moon Europa has an ocean beneath its ice shell and is one of the targets for a search for life in the solar system. Impact craters on Europa exhibit unusually shallow depths at larger diameters and show interesting features such as central pits and domes. These characteristics are intrinsic indicators of conditions at depth, such as the thickness and structure of the ice shell. Despite recent developments in observational and theoretical techniques, the question whether the ice shell covering Europa’s ocean is thin or thick remains open. The implications of thin versus thick ice shell are extremely pertinent to the search for life on Europa and future space mission planning. We performed computer simulations to model the formation of impact craters on Europa to probe the thickness and structure of its ice shell. Our study explored whether Europa’s ice shell is thin or thick by simultaneously matching our simulation results to the observed crater dimensions and features. Our results suggest that both thin and thick ice shell scenarios can reproduce the observed crater depths, diameters, and morphologies. In future work, modeling even larger impact craters known as multiring basins might better constrain Europa’s ice shell thickness and structure.

1. Introduction

Geophysical observations of Jupiter’s moon Europa suggest that this icy world hosts a global subsurface ocean (e.g., Khurana et al., 1998; Kivelson, 2000; Kivelson et al., 1999). Europa’s ice shell has been estimated to be a few kilometers (e.g., Greenberg et al., 2000; Hoppa, 1999) to a few tens of kilometers thick (e.g., Moore et al., 2001; Pappalardo et al., 1999; Prockter & Pappalardo, 2000). The thickness and structure of the ice shell bear immense relevance to astrobiological, geophysical, and planetary body evolution considerations, as well as future space mission planning. On one hand, a thin shell over ocean would enable impact penetration (e.g., Cox et al., 2008), formation of chaos regions, heat exchange via conduction, and nutrient exchange via direct cracks (e.g., Greenberg et al., 2000). On the other hand, a thick shell, composed of a conductive stagnant lid and warm convecting ice underneath, would enable tidal heating dissipation (e.g., McKinnon, 1999; Nimmo & Manga, 2009), while diapirs initiated by thermal convection could be a mode for nutrient exchange (e.g., Ruiz et al., 2007) and formation of chaos regions (e.g., Schmidt et al., 2011). Despite recent developments in remote sensing (e.g., Schenk, 2002) and numerical modeling (e.g., Bray et al., 2014; Cox & Bauer, 2015), large uncertainties regarding the thermal structure and thickness of Europa’s ice shell remain.
Impact craters produced by hypervelocity collisions are among the most ubiquitous geological features on solid planetary surfaces. The final size and morphology of an impact crater depend on both the projectile (e.g., size and velocity) and target properties (e.g., rock, ice, and thermal gradient) (e.g., Melosh, 1989). Impact craters exhibit a progression in morphologies as a function of increasing diameter. On rocky bodies, simple, bowl-shaped crater morphologies transition to complex craters with flat floors and central peaks, then peak ring basins, and finally multiring basins (e.g., Melosh, 1989; Pike, 1980). During an impact, a bowl-shaped transient crater is excavated and subsequently succumbs to gravitational collapse or modification. Crater modification, and therefore the resulting crater dimensions and morphology, is highly sensitive to target properties, such as heat flow, layering, and the presence of underlying ocean (e.g., Collins et al., 2004; Ivanov et al., 2010; Johnson, Blair, et al., 2016; Senft & Stewart, 2011). Hypervelocity impacts generate tremendous amounts of energy, melting and vaporizing both the projectile and target. Ice is brittle and behaves like typical terrestrial rock at Europa’s surface temperature of 100 K (Ojakangas & Stevenson, 1989). However, considering that ice melts and deforms at temperatures substantially lower than that for rock, icy bodies will be even more susceptible to expressing the conditions at depth through crater morphologies (e.g., Schenk, 2002). Among tools used to estimate the thickness of Europa’s ice shell, analysis of impact craters remains the most cost-effective probe of ice shell thickness and likely conditions at depth (e.g., Bray et al., 2014, 2008; Cox & Bauer, 2015; Moore et al., 1998; Schenk, 2002).

Observations of impact craters, characterized as fresh, unmodified, and unrelaxed by Schenk (2002), on the icy Galilean moons reveal the depth-diameter (d-D) relationship that exhibits three distinct transitions. On Europa, these transitions are as follows: simple-to-complex transition (I), anomalous crater dimensions and morphologies (e.g., central pits and domes) with crater depths that decrease as crater diameter increases (II), and abrupt transition from modified central peak to shallow craters with multiring morphologies (III) (Schenk, 2002) (Figure 1). Decreasing crater depths with increasing diameters, Transitions II and III, are unique to the icy Galilean moons, occurring on Europa at D ~4 km and ~8 km, respectively, and larger diameters on Ganymede and Callisto. Schenk (2002) argued that these transitions on Europa are indicative of warm convecting ice at depths of 7–8 km and a liquid ocean at 19–25 km, respectively (Schenk, 2002). Figure 2 shows various crater morphologies of Europan craters ranging from 1.3 km to 23 km in diameter.

Numerical modeling by Turtle and Pierazzo (2001) demonstrated that Europa’s ice shell must be thicker than 3–4 km; otherwise the zone of material melted by the impact would intersect the underlying ocean and thus inhibit formation of craters with central peaks, which require significant material strength. Recent developments and improvements in numerical models have opened the door for more sophisticated studies. Bray et al. (2014) estimated the ice shell thickness at 7 km, while simultaneously matching the observed crater morphologies (e.g., summit pit and dome craters) and depth-diameter relationships (Schenk, 2002) (Figure 1). Cox and Bauer (2015) placed Europa’s icy shell thickness at 10 km by numerically probing the effects of crust breaching and matching the observed crater depth-diameter relationships (Figure 1), regardless of observed morphology. Both studies (Bray et al., 2014; Cox & Bauer, 2015) considered only a conductive ice shell over ocean scenario.

We modeled formation of impact craters to probe its internal structure. Our study differs from previous works in that we consider both a fully conductive ice shell over ocean and a conductive ice shell over warm convective ice to attempt to discern boundary conditions at the interface between the ice and the underlying ocean (Figure 3). In the latter scenario, it is assumed that a conductive ice lid and warm convective ice overlie the global ocean. We aim to elucidate the following questions regarding the anomalous morphology of Europan craters: Is the conductive ice shell over ocean the only plausible scenario? Could the conductive lid over warm convective ice (thick shell) scenario reproduce the observed depth-diameter and crater
Figure 2. (left column) The Europan crater profiles alongside (right column) the crater images. The crater rims are shown with arrows. All panels, except Figure 2a, have the same vertical scale. The horizontal scale varies across panels. The crater profiles were digitized (a–d) from Figure 6 in Bray et al. (2014), and (e) from Figure 9c in Schenk and Turtle (2009). Crater images are from the Galileo spacecraft. All craters, except that in Figure 2a, are complex craters with anomalous morphologies (Transition II) (Schenk, 2002).
morpheologies on Europa as well as the fully conductive ice shell (thin shell)? If yes, what are some probable parameters associated with the structure of the shell, specifically thickness of the conductive lid and the temperature of warm convective ice?

Our paper is organized into several sections. The methodology and model setups are described in section 2. Section 3 outlines our results and is further divided into subsections that discuss the modeled crater depth-diameter for conductive (thin) ice shell (section 3.1.1) and conductive-convective (thick) ice shell (section 3.1.2), as well as the comparison to observed crater morphologies (section 3.2) in the context of both simple-to-complex transition (section 3.2.1) and anomalous morphologies (section 3.2.2). Section 4 presents the discussion about the numerical modeling considerations relevant to formation of Europan crater morphologies through hydrocode simulations (section 4.1) and the effect of strength of ice on crater morphology (section 4.2). In this section we further discuss our findings in the context of pits and domes formation (section 4.3) and the thin versus thick ice shell, followed by implications for future studies (section 4.4). The conclusions are listed in section 5.

2. Methods

We modeled the formation of impact craters on Europa using iSALE-2D, a multimaterial, multi rheology shock physics code (Collins et al., 2004; Ivanov et al., 1997; Melosh et al., 1992; Wünemann et al., 2006), which is based on the SALE hydrocode solution algorithm (Amsden et al., 1980). Due to the axial symmetry of our models, only vertical impacts were considered. Although the average impact velocity ($v_i$) for Europa is 26 km/s (Zahnle et al., 2003), we use $v_i = 15$ km/s to reduce the simulation time and to maintain the consistency with previous work (Bray et al., 2014; Cox & Bauer, 2015). The ocean layer was represented by the Analytic Equation of State (ANEOS) (Thompson & Lauson, 1972) and ice Ih by Tillotson equations of state (EOS) (Ivanov et al., 2002; Tillotson, 1962).

The strength and damage models for ice shell and the parameters for the Block Model of acoustic fluidization (Melosh, 1979) correspond to Bray et al. (2014) and are listed in Tables 1 and 2, respectively. In the Block Model, the parameters controlling the degree to which the target is weakened during cratering are decay time ($T_{\text{dec}}$) and limiting viscosity ($\nu_{\text{lim}}$) of the fluidized target. Bray et al. (2014) estimated these empirically by performing a broad parameter study for uniform nonlayered ice to match the morphometry of impact craters on Ganymede. Since large craters ($D > 25$ km) are more susceptible to rheological changes that occur with depth (Schenk, 2002), the parameters were determined for small impact craters small ($D = 4–25$ km).

The grid resolution was set to 10 cells per projectile radius, as that offers optimal computing time, and a reasonably small error, while retaining the temporal evolution and morphology of the resulting crater (e.g., Elbeshausen et al., 2009; Silber et al., 2017, and references therein). The impactor radii ($R_i$) also correspond to Bray et al. (2014) and range from 26 m to 405 m (Table 2). Similar impactor sizes were also employed by Cox and Bauer (2015).

We considered a full viscoelastic-plastic ice rheology to account for any viscous contribution to material deformation. A viscoelastic-plastic rheology for rocky mantle material was first introduced in iSALE by Dirk Elbeshausen and is available as part of the iSALE-Dellen release (https://dx.doi.org/10.6084/m9.fgsharet.3473690).

Parameters appropriate for ice were subsequently implemented by

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**Figure 3.** The diagram depicting the two setups modeled in our study: (a) a fully conductive shell over ocean and (b) a conductive ice lid overlying warm convective ice. Schenk (2002) estimated the depth of the ocean at 19–25 km and noted that craters at Transition III would exhibit features indicative of the rheological changes (e.g., multiring structures). While it is assumed that there is ocean at some depth in the conductive-convective scenario (Figure 3b), it is not explicitly modeled. This is because at crater sizes investigated in this study ($D < 23$ km), the presence of the ocean at depths proposed by Schenk (2002) is not expected to make an appreciable contribution.

<table>
<thead>
<tr>
<th>Table 1</th>
<th>The Summary of Input Parameters for Ice</th>
</tr>
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<tbody>
<tr>
<td>Description</td>
<td>Values</td>
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<tr>
<td>Surface temperature (K)</td>
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</tr>
<tr>
<td>Melt temperature at zero pressure (K)</td>
<td>273</td>
</tr>
<tr>
<td>Thermal softening parameter</td>
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<tr>
<td>Cohesion, intact (kPa)</td>
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</tr>
<tr>
<td>Limiting strength at high pressure, intact (GPa)</td>
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</tr>
<tr>
<td>Cohesion, damaged (kPa)</td>
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</tr>
<tr>
<td>Coefficient of internal friction, damaged</td>
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</tr>
<tr>
<td>Limiting strength at high pressure, damaged (Gpa)</td>
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</tr>
<tr>
<td>Equation of state (EOS)</td>
<td>Tillotson, H$_2$O</td>
</tr>
<tr>
<td>Impact velocity (km/s)</td>
<td>15</td>
</tr>
</tbody>
</table>

Note: These quantities correspond to Bray et al. (2014), with coefficient of internal friction for damaged material taken from Bray’s (2008) fits to laboratory data (Beeman et al., 1988). The ocean layer (H$_2$O) is strengthless and thus not listed in the table. The ocean layer is represented by the ANEOS (Thompson & Lauson, 1972) and ice Ih by Tillotson (Ivanov et al., 2002; Tillotson, 1962) equations of state (EOS).
Thermal Gradients for Various Scenarios Modeled in This Study

Table 2
<table>
<thead>
<tr>
<th>Impactor radius (m)</th>
<th>Kinematic viscosity (10^3) (m² s⁻¹)</th>
<th>Decay time (s)</th>
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<tbody>
<tr>
<td>26</td>
<td>4.09</td>
<td>22</td>
</tr>
<tr>
<td>70</td>
<td>15</td>
<td>35</td>
</tr>
<tr>
<td>107</td>
<td>26.5</td>
<td>42</td>
</tr>
<tr>
<td>150</td>
<td>40</td>
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<td>230</td>
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<td>320</td>
<td>110</td>
<td>70</td>
</tr>
<tr>
<td>405</td>
<td>150</td>
<td>80</td>
</tr>
</tbody>
</table>

Table 3
<table>
<thead>
<tr>
<th>Ice shell thickness (km)</th>
<th>Thermal gradient (K/km)</th>
<th>Rollover depth (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>255 K</td>
<td>265 K</td>
</tr>
<tr>
<td>4</td>
<td>43.25</td>
<td>3.59</td>
</tr>
<tr>
<td>5</td>
<td>34.60</td>
<td>4.49</td>
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<tr>
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<tr>
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<td>21.63</td>
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</tr>
<tr>
<td>9</td>
<td>19.22</td>
<td>n/a</td>
</tr>
</tbody>
</table>

Note. These parameters control the degree to which the target is weakened during cratering. Bray et al. (2014) found these by performing a broad parameter study for uniform nonlayered ice to match the morphometry of impact craters on Ganymede using the impact velocity of 15 km/s. The parameters are described by the following equations: \( T_{\text{dec}} = 4.86R_i^{1.46} \) and \( \nu_{\text{lim}} = 55.48R_i^{0.32} \) (Bray et al., 2014). Since large craters \( (D > 25 \text{ km}) \) are more susceptible to rheological changes that occur with depth (Schenk, 2002), the parameters were determined for small impact craters \( (D = 4–25 \text{ km}) \).

Second, we modeled conductive-convective layering, where conductive ice (a conductive lid) with a steep thermal gradient (Table 3) overlays a region of convective warm ice that has an adiabat of ~0.05 K/km. A purely conductive layer is produced assuming a linear thermal gradient \((dT/dz)\) corresponding to the surface temperature of 100 K (Ojakangas & Stevenson, 1989) and temperature of 273 K for the underlying ocean. Thus, for a 7 km conductive ice shell, the heat flow \((q)\) is ~74 mW m⁻² and thermal conductivity \((k)\) is ~3 J/m/s (Bray et al., 2014). Values of thermal gradient for various ice shell thicknesses are given in Table 3.

Our study consists of two components. First, the aim was to model the fully conductive ice shell over the liquid ocean (Figure 3a) (as also done by Bray et al., 2014; Cox & Bauer, 2015) for comparison to our simulation that includes warm convecting ice and to investigate the influence of a viscoelastic-plastic ice rheology. To this end, we ran several suites of simulations. We modeled the fully conductive shell with thickness 7, 8, and 9 km using the viscoelastic-plastic ice rheology implementation. For comparison purposes, we also ran one simulation suite for the 7 km ice shell without the viscoelastic-plastic ice rheology implementation. Estimates of Europa’s surface heat flow range from approximately 50 mW/m² (Melosh et al., 2004; Ojakangas & Stevenson, 1989) up to ~100 mW/m² (Ruiz & Tejero, 2000). A purely conductive layer is produced assuming a linear thermal gradient \((dT/dz)\) corresponding to the surface temperature of 100 K (Ojakangas & Stevenson, 1989) and temperature of 273 K for the underlying ocean. Thus, for a 7 km conductive ice shell, the heat flow \((q)\) is ~74 mW m⁻² and thermal conductivity \((k)\) is ~3 J/m/s (Bray et al., 2014). Values of thermal gradient for various ice shell thicknesses are given in Table 3.

Note. We also list the depth at which the conductive regime “rolls” over to the convective regime with adiabat of ~0.05 K/km. For example, for the 5 km thick purely ice shell, the thermal gradient \((dT/dz = 34.6 \text{ K/km})\) corresponds to the difference between the surface temperature \((T = 100 \text{ K}; \text{Ojakangas & Stevenson, 1989})\) and the ocean temperature \((T = 273 \text{ K})\). Applying that same thermal gradient to a conductive shell over warm convective ice will result in the rollover occurring at a shallower depth because the temperature of the warm convective ice is lower \((T = 255–265 \text{ K})\) than that of water. Thus, for the warm convective ice at 255 K, the rollover depth will be 4.49 km, and for warm convective ice at 265 K, the rollover depth will be 4.78 km. We note that the rollover depth is listed only for convective ice scenarios that were investigated in our study.
(\(R_{\text{wall}}\)) of the crater wall and the preimpact surface (Figure 5). The uncertainty in crater radius is expressed as \(R_{\text{err}} = 0.25(R - R_{\text{wall}})\) and in overall diameter as \(D_{\text{err}} = 2R_{\text{err}}\). The uncertainty in crater depth is given as one grid cell in each direction (\(+/-\)), corresponding to the physical size of \(R/10\) and thus ranging from 2.6 m (\(R_i = 26\) m) to 40.05 (\(R_i = 405\) m).

Finally, since iSALE is not suitable for studying postimpact relaxation or small-scale deformations over long time scales (e.g., after the crater collapse is over), caution should be given when determining simulation run times. Extended simulation run times result in reflections, which coupled with numerical diffusion may give the appearance of shallowing craters. A couple of approaches to determine when a simulation is over can be applied. One is to qualitatively establish when all the motion stops from plotted time series. The other, more robust, is to plot up crater profiles (depth versus diameter) across several time frames before and after crater collapse is deemed to have finished and find the point at which crater depth, diameter, and morphology no longer undergo a notable change. Examples of this are shown in Figure 6. In Figure 6a, the reported simulation end time (\(t_{\text{end}}\)) is 215 s. The three subsequent time frames (\(t = 220, 225,\) and 230 s) show that crater depth, diameter, and morphology no longer undergo a notable change. The “steps” (e.g., on crater floor and crater rim), represent discrete cells in the simulation grid. Similarly, in Figure 6b, the crater profile at \(t = 400\) s is nearly identical to that at \(t_{\text{end}}\). The top of the crater rim is narrowed laterally by only one cell on each end, and the central peak is also reduced by one cell laterally. However, this makes no difference in either the measured crater depth or diameter, because the methodology to measure the crater dimensions described earlier in this section is robust. For example, both \(R\) and \(R_{\text{err}}\) remain unchanged from \(t_{\text{end}}\) to \(t = 400\) s. The uncertainty in crater depth, corresponding to a single cell in each direction, also accounts for any minute variations.

3. Results

Figure 7 shows the formation of anomalously shallow craters by a 320 m radius impactor for two possible ice shell structures. The left panel represents a 6 km thick conductive lid over warm convective ice at 265 K, and the right panel is for an 8 km thick ice shell over ocean. As we will show later, these conditions represent our best fit for reproducing observed crater depths and diameters. The collapse of the transient crater results in a large central uplift (Figure 7a), which subsequently collapses (Figure 7b). The central uplift is much larger and reaches more than twice as high for the convective ice case as compared to the 8 km conductive ice shell.
In the 8 km shell, the presence of the underlying ocean leads to pronounced crater floor uplift seen at 240 s (Figure 7b), which is retained in the final crater morphology (Figure 7c). While there is no significant crater floor uplift retention in the convective ice scenario, the final crater is anomalously shallow (Figure 7c). Regardless of the differences in the final morphology at the center of the crater, both simulations yield a crater that is 0.4 km deep and the diameters are similar (D = 18 km for the 6 km conductive lid over convective ice and D = 17 km for the 8 km conductive shell over ocean). We will discuss these differences in section 3.2.

3.1. Modeled Crater Depth-Diameter

In Figure 8 we compare the observed crater depth-diameter (d-D) (Schenk, 2002) to our simulation results. The conductive ice shell over ocean (Figure 8a) and the 255 K (Figure 8b) and 265 K (Figure 8c) warm convective ice scenarios are plotted in separate panels for clarity. The modeled crater diameters and depths, along with their uncertainties in measurement, are listed in Tables S1–S3 in the supporting information.

While crater diameters produced by impactors of a given size are relatively insensitive to changes in preimpact thermal structure considered here, the depths are significantly different and strongly dependent on the temperature gradient. The implementation of viscoelastic-plastic rheology produces relatively small differences in crater depth and diameters for the crater size range investigated here (D < 23 km) and no notable differences in crater morphology.

3.1.1. Conductive Ice Shell

In the ice shell over ocean scenario (Figure 8a), for crater sizes up to ~8 km in diameter (R1 = 150 m), all models produce very similar d-D, which is also consistent with observations. At smaller crater sizes (R1 = 107 m, D < 5.7 km), the crater depths and diameters for all three shell thicknesses converge, implying that around and below this threshold, impact craters are insensitive to the shell thickness and thermal gradient. At diameters larger than ~8 km, the 7 km ice shell results in craters that are on the shallow end of the observed d-D, with significant divergence and a steep dropoff in crater depth beyond D < 15 km. For the simulations with a 9 km shell, the 11 km crater (R1 = 230 m) is too deep, while the craters in larger size range are within the observed d-D bounds. The 8 km ice shell produces the best fit to the d-D of observed craters. An animation of the best fit simulation for R1 = 230 m is shown in the supporting information (Animation S1). This result is broadly consistent with the estimates derived from earlier numerical studies (Bray et al., 2014; Cox & Bauer, 2015).

3.1.2. Conductive Ice Lid Over Warm Convective Ice

In the 255 K warm convective ice scenario (Figure 8b), the 7 km conductive lid results in a flat d-D, with the roll-off nearly absent and with craters that are too deep. Both the 5 km and 6 km lid over warm convective ice...
The 4 km conductive lid results in very shallow craters, especially beyond \( D \sim 8 \) km. The conductive shell case, the presence of the underlying ocean at depths \( \geq 7 \) km does not have an influence on the simple-to-complex transition, also noted by Bray et al. (2014). However, the simple-to-complex transition does seem to be a function of thermal gradient. For example, a crater formed by the impactor with \( R_l = 107 \) m into an 8 km or a 9 km conductive shell over ocean exhibits a flat floor (Figure 9b). Conversely, the same impact into a warmer target (4 km conductive lid over warm convective ice at 255 K) leads to a complex crater with a well-defined central peak. The three best fits from our simulations are consistent with the morphological simple-to-complex transition (Transition I).

The modeled craters produced by impactors with \( R_l = 150 \) m (\( D < \sim 8 \) km) for all scenarios exhibit complex morphologies, including a central peak. The smallest craters (\( R_l = 26 \) m, \( D < 2 \) km) show the characteristics consistent with simple craters across all simulations. Beyond \( D \sim 4 \) km (\( R_l = 70 \) m), the modeled craters transition into complex structures; however, this transition is more gradual (flat floors and then central peaks) for some cases and more abrupt for others (e.g., Figures 9b and 9c). In the conductive shell case, the presence of the underlying ocean at depths \( \geq 7 \) km does not have an influence on the simple-to-complex transition, also noted by Bray et al. (2014). However, the simple-to-complex transition does seem to be a function of thermal gradient. For example, a crater formed by the impactor with \( R_l = 107 \) m into an 8 km or a 9 km conductive shell over ocean exhibits a flat floor (Figure 9b). Conversely, the same impact into a warmer target (4 km conductive lid over warm convective ice at 255 K) leads to a complex crater with a well-defined central peak. The three best fits from our simulations are consistent with the morphological simple-to-complex transition (Transition I).

3.2. Comparison to Observed Crater Morphologies

Figure 9 shows the modeled crater profiles for the three best fit scenarios: the conductive (8 km shell over ocean) and conductive-convective scenarios (5 km ice lid over warm convective ice at 255 K and 6 km ice lid over warm convective ice at 265 K). Figure 1 shows the European crater topographic profiles and images for five craters ranging from 1.3 km to 23 km in diameter. The crater profiles for Figures 9a–9d were digitized from Figure 6 in Bray et al. (2014) and Figure 9e from Figure 9c in Schenk and Turtle (2009). Although there are \( \sim 150 \) cataloged craters over 1 km in diameter on Europa, there is a relatively small number of craters larger than 10 km in diameter (Schenk, 2002; Schenk & Turtle, 2009). Excluding the multiring basins, only 11 of those have been assigned the morphological type (Schenk & Turtle, 2009). These craters are listed in Table S4. Since only a handful of topographical profiles are available for comparison to the modeled craters, in this section we perform a qualitative assessment and comparison of modeled and observed craters.

3.2.1. Simple-to-Complex Transition

The onset of complex structures occurs at crater diameters of approximately 4 km, which is consistent with observations (Schenk, 2002). The modeled craters produced by impactors with \( R_l = 150 \) m (\( D < \sim 8 \) km) for all scenarios exhibit complex morphologies, including a central peak. The smallest craters (\( R_l = 26 \) m, \( D < 2 \) km) show the characteristics consistent with simple craters across all simulations. Beyond \( D \sim 4 \) km (\( R_l = 70 \) m), the modeled craters transition into complex structures; however, this transition is more gradual (flat floors and then central peaks) for some cases and more abrupt for others (e.g., Figures 9b and 9c). In the conductive shell case, the presence of the underlying ocean at depths \( \geq 7 \) km does not have an influence on the simple-to-complex transition, also noted by Bray et al. (2014). However, the simple-to-complex transition does seem to be a function of thermal gradient. For example, a crater formed by the impactor with \( R_l = 107 \) m into an 8 km or a 9 km conductive shell over ocean exhibits a flat floor (Figure 9b). Conversely, the same impact into a warmer target (4 km conductive lid over warm convective ice at 255 K) leads to a complex crater with a well-defined central peak. The three best fits from our simulations are consistent with the morphological simple-to-complex transition (Transition I).
3.2.2. Anomalous Morphologies

Compared to craters on Ganyamede and Callisto, Europan craters transition to anomalously shallow depths and anomalous morphologies at smaller diameters. This trend is linked to the presence of underlying global ocean at depths of a few tens of kilometers (Kivelson, 2000; Pappalardo et al., 1999). However, complex Europan craters exhibit various morphologies at comparable diameters, making the validation of numerical models to observations challenging.

For example, Grainne ($D \sim 14$ km) and Amergin ($D \sim 19$ km) are both classified as disrupted central peak craters. Within this size range, there are also the flat floored craters Math ($D \sim 15$ km) and Rhiannon ($D \sim 16$ km) and the central peak craters Eochaird ($D \sim 17$ km) and Cilix ($D \sim 19$ km, Figure 1c). At larger sizes, despite having nearly the same diameters, Maeve ($D \sim 22$ km, Figure 2d) and Mannanán ($D \sim 23$ km, Figure 2e) have strikingly different morphologies. Maeve is a classic example of a central peak crater, while Mannanán is a disrupted central peak crater with a central pit (Moore et al., 2001).

The modeled craters with $D \sim 11–12$ km ($R_i = 230$ m) are consistent with the observed classic central peak craters of similar size, such as Avagddu ($D \sim 11$ km).

The morphology of larger modeled craters ($D > 15$ km), however, varies depending on the setup (conductive versus conductive-convective). For all shell thicknesses modeled in this study, a simple conductive shell over ocean leads to craters with prominent peaks and central pits (Figures 9e and 9f), consistent with some of the observed craters, such as Maeve and Cilix (Figures 2c and 2d). However, the formation of central pits might be a by-product of vertical impact setup and axial symmetry, rather a real effect (see section 3.3). The disrupted central peaks and flat floors are not well reproduced with the fully convective ice shell.

In the conductive ice lid over warm convective ice setup, the modeled craters with $D > 15$ km exhibit a wider range of morphologies but generally lack large peaks (Figures 9e and 9f). The 6 km conductive lid over warm convective ice at 265 K scenario leads to anomalously shallow craters (Figures 9e and 9f) and craters with irregular crater floor (Figure 9e), reminiscent of a disrupted central peak crater (e.g., Mannanán). The largest modeled crater ($D \sim 22$ km, $R_i = 405$ m) is broadly consistent with the largest shallowest observed crater with the disrupted central peak (Manannán) (Figure 9c). The 5 km conductive lid over warm convective ice at 255 K scenario produces anomalously shallow craters and anomalous morphologies (e.g., flat floors, absence of well-defined crater rims, and irregular crater floors, reminiscent of disrupted central peak). However, the largest modeled crater ($D \sim 21$ km, $R_i = 405$ m) is significantly deeper than the two observed shallowest large craters on Europa (Figure 9b). It is also deeper than its modeled counterparts using the 8 km conductive shell and the 6 km conductive lid over warm convective ice at 265 K (Figure 9f).

3.3. Material Splashing and Overflow

At larger crater diameters, during crater collapse, a central uplift will form. This central uplift is unstable and will also succumb to gravitational collapse. If the thermal gradient of the target is sufficiently high, as it is the...
case in several scenarios examined in this study (conductive ice lid over warm convective ice), collapse of the central uplift may cause warm material to splash out of the crater center and spread over the crater rim (Johnson, Blair, et al., 2016; Johnson, Bowling, et al., 2016). This occurs for two reasons. First, the combination of a thin conductive lid and a higher thermal gradient contributes to the presence of warm material at a shallow depth. This warm material is also much weaker than the cold material, because the strength of material depends on temperature (Collins et al., 2004), and deforms more readily than cold brittle ice. Second, the axial symmetry setup in hydrocode simulations leads to exaggerated central uplift as opposed to three-dimensional models of oblique impacts (Elbeshausen et al., 2009). Specific simulations affected with the material over flow are those for $R_i = 320$ m and $405$ m impacting a 4, 5, and 6 km conductive lid over warm convective ice at $255$ K, and $R_i = 405$ m impacting a 5 km conductive lid over warm convective ice at $265$ K.

An example of this is shown in Figure 10 for a 320 m projectile impacting a 4 km thick ice lid overlying warm convective ice at 255 K. The preimpact surface and transient crater are shown in Figures 10a and 10b. During the crater collapse, central uplift forms, as shown in Figure 10c at 110 s into the simulation. However, due to axial symmetry, the central uplift is exaggerated, as shown in the subsequent time segment in Figure 10d, at 180 s. Considering that the strength of material depends on temperature (Collins et al., 2004), this warm, weaker material will readily relax (Figure 10e) and free-fall into the center of the crater (Figure 10g), thereby pushing even more warm material out. The crater rim becomes disguised by a layer of warm material, which

Figure 9. The modeled crater profiles for best fit results from our simulations for conductive scenario (8 km shell over ocean (black)) and conductive-convective scenarios (5 km ice lid over warm convective ice at 255 K, and 6 km ice lid over warm convective ice at 265 K (red)). The legend shown in Figure 9a applies to all panels.
will obscure the true dimensions of the crater. Consequently, the new, apparent rim will not correspond to the actual crater rim, which remains concealed. Figures 10g and 10h show the intermediate step shortly before the final shape of the crater takes place at 685 s (Figure 10i).

The apparent rim produced by the outflow is broad and inconsistent with the sharp scarpas that define observed crater rims (Figure 2). Moreover, the final crater depth will also be affected due to the artificially raised rims such that the craters appear unrealistically deep. On a d-D plot, these craters exhibit depths that increase with an increasing crater diameter, which is not consistent with the observed craters on Europa (Schenk, 2002). The most extreme example of this is the scenario with the 4 km lid over warm convective ice at 255 K. For example, the crater depth for $R_i = 230$ m is 0.32 km (not affected by overflow). For comparison, for larger impacts where the crater depth is expected to further decrease, the overflow results in the apparent depth of 0.35 km ($R_i = 320$ m) and 0.40 km ($R_i = 405$ m). Animation S4 shows the overflow for

Figure 10. Time series showing the crater formation for the projectile with $R_i = 320$ m impacting a 4 km thick ice lid over warm convective ice at 255 K.
$R_i = 405 \text{ m}$ impacting the 5 km thick conductive ice lid over warm convective ice at 265 K. Data points for simulations that suffer from the apparent warm material overflow are represented by triangles in Figure 8. Finally, we note that the larger the impact, the more overflow effect there will be.

Thus, on one hand, if the warm material overflow is dominated by the axial symmetry setup rather than the behavior imposed by the conditions of the modeled ice shell, then the real depth and diameter dimensions will remain poorly constrained and such results cannot be used to infer the ice shell conditions on Europa. On the other hand, if the axial symmetry plays only a minor role, then none of the scenarios producing skewed $d/D$ (e.g., 4 km ice shell over warm conductive ice at 255 K, see Figure 8b) would be an appropriate consideration for the modeled ice shell conditions on Europa. Alternatively, there could be a more complex interplay between axial symmetry and modeled ice shell; however, more comprehensive studies are needed before a more definitive assertion can be made.

For other simulations, where the overflow is of lesser extent, we performed the measurement at the point at which the rim fully formed but before it was flooded over by the warm material (and thus before the entire crater fully formed). The measurement was performed following the same methodology as described earlier in this section. Therefore, we are able to report and plot (Figures 8b and 8c) crater dimensions. These are considered special cases and, as such, are denoted with half-filled squares in Figures 8b and 8c.
An example of the crater profile at two time steps used to measure the dimensions is shown in Figure 11. The profiles represent the results for $R_i = 320$ m into a 5 km thick ice lid over warm convective ice at 255 K. Since the final crater depth is shallow, and to better show the intermediate and final crater shapes, the profiles are plotted with the exaggerated vertical scale (Figure 11a). The crater rim was measured 140 s into the simulation, as this is when it fully formed, although the crater continues to collapse. The crater depth, on the other hand, was measured at 500 s, when the crater reached its final dimensions. Although minor oscillations (a remnant of splashing) localized at the very narrow central region of the crater floor took longer to completely cease (see Animation S4), they did not affect the final crater morphology or apparent depth and diameter. By $t = 500$ s point, the rim is no longer raised but appears flat (Figure 11c), with only the inner “edge” present. However, it is evident that the inner edge at 500 s overlaps the location of the rim at 140 s, thus confirming that our approach to measure the crater diameter at earlier time is robust.

4. Discussion

Recent numerical studies (Bray et al., 2014; Cox & Bauer, 2015) reproduced the observed crater $d/D$ and inflection points using a fully conductive icy shell over ocean. Bray et al. (2014) also matched crater morphology (e.g., summit pits and central domes) to discern the thickness of Europa’s ice shell. While numerical modeling using layering can produce domes and summit pits in larger craters ($D \geq 15$ km) (see Bray et al., 2014 for more details), there are certain considerations pertaining to hydrocode simulations that need to be examined.

4.1. Numerical Modeling Considerations

In our simulations, the impactor composition considered was ice, but if it was rocky or metallic, for example, impact-generated heating would be greater (e.g., Barr & Citron, 2011). In many hydrocodes, including iSALE-2D, the impact angle is set at 90°. In reality, however, the majority of impacts occur at an angle of 45° (Gilbert, 1893; Shoemaker, 1962). Thus, in some cases, axial symmetry in modeling might not accurately represent the real conditions involved in cratering process. Some of the factors that may be sensitive to angle of impact are the cratering efficiency (Elbeshhausen et al., 2009), impact-generated heating and melting (Pierazzo & Melosh, 2000), exaggerated central uplift, and formation of central features (e.g., pits, domes and summit pits). Additionally, the effect of impact velocity cannot be neglected, as it affects the extent of melting (Pierazzo et al., 1997) and final crater morphology (Silber et al., 2017). We will briefly touch on the effect of impact angle and velocity in this section.

4.1.1. The Effect of Impact Angle

Catering efficiency decreases as a function of impact angle, and transient crater depth, diameter, and volume all depend on impact angle (Elbeshhausen et al., 2009). Vertical impacts also produce a symmetrical isobaric core and the melting region around the point of impact (Pierazzo & Melosh, 2000). As the angle of impact decreases from the vertical (90°), the strength of the shock and the shape of the heated region changes accordingly and becomes more and more asymmetrical. Vertical impacts generate substantially more heating and melt as opposed to very oblique impacts (< 30°), where the melting efficiency is greatly diminished (Pierazzo & Melosh, 2000). These varying conditions might influence how central features, specifically pits and domes, form (e.g., formation of melt pools and conditions conducive to postimpact drainage). Moreover, axial symmetry often leads to exaggerated central uplift when compared to full 3-D simulations of oblique impacts (see section 3). Subsequent relaxation of the uplifted material could alter the final morphology (Moore et al., 2017) of the crater center by producing certain central features which might not form in oblique impacts (e.g., in a 3-D simulation). Hence, the formation of central features, such as pits, domes, and summit pits in a crater might be sensitive to the angle of impact and form mainly through the axial symmetry, as opposed to using full three-dimensional modeling of oblique impacts (Elbeshhausen et al., 2009). However, since there are no studies that specifically focused on this problem, it is not possible to make a conclusive determination as to what extent these features might or might not be affected by the impact angle. While the height of a central peak might be affected by axial symmetry, a study on Venusian craters has shown that the central peak diameter and its position relative to crater center has no correlation to the impact angle, at least for impacts on rocky bodies (Ekholm & Melosh, 2001).

4.1.2. The Effect of Impact Velocity

Impact velocity also plays an important factor in cratering process. For example, impacts at high velocities generate substantially more melt than impacts at low velocities (Pierazzo et al., 1997), which is important
for creating conditions conducive to formation of melt pools. The average impact velocity on Europa is 26 km/s (Zahnle et al., 2003), higher than that implemented in this study and other numerical studies (e.g., Bray et al., 2014; Cox & Bauer, 2015). Moreover, the morphology (e.g., onset of flat floors and central peaks) of impact craters, especially in the simple-to-complex transition regime, is highly sensitive to impact velocity, as shown by recent study on lunar craters (Silber et al., 2017). Silber et al. (2017) modeled lunar craters (D = 10–26 km) forming as a result of impacts at velocities 6–15 km/s. Modeling at very low impact velocities (2–10 km/s), representative of conditions at Pluto, also shows variations in crater depth and simple-to-complex transition as a function of impact velocity (Bray & Schenk, 2015). Given the small statistics of cataloged European craters with D > 10 km and their apparent diverse morphologies even at similar diameters (see Table S4), it is not possible to deduce which factors outlined above might dominate the observed morphologies. Moreover, at larger crater sizes, in addition to variations in morphology at the same crater sizes (see section 3 and Table S4), there is large ambiguity in whether central features form during crater collapse or well after.

If central pits and domes form post impact, then depending on the extent of the melt pool size, volume, and its emplacement within ice, the postimpact modification could result in a variety of possible crater morphologies. The conductive-convective scenario (thick ice) would be most consistent with such hypothesis. However, if central pits and domes form during the cratering collapse, then the purely conductive ice shell over ocean (thin shell) would be the most appropriate choice.

### 4.2. The Effect of Strength of Ice on Crater Size and Morphology

The strength of ice (Y) as a function of depth is an important consideration, and as demonstrated through our simulations, multiple strength profiles as a function of depth are possible (Figure 4a). Ultimately, it is the parameters for strength of ice that dictate what combination of thermal gradient, rollover temperature, and conductive lid thickness will be required to produce a given strength profile. The strength parameters for ice are derived (Bray et al., 2014, and references therein) from experimental data (Beeman et al., 1988), and the strength model used in iSALE is described in detail in Ivanov et al. (1997) and Collins et al. (2004). Therefore, we will only discuss the outcome in the context of our results. The strength-depth profiles for our best fits, represented with solid lines in Figure 4a, illuminate the linkage between crater morphology and the strength of ice. For example, the 8 km conductive ice shell over the ocean is notably stronger down to a depth of ~7 km than either of the best fits for the conductive lid over warm convective ice. However, below 8 km, it shifts to the strengthless regime. The strength profile for the conductive ice lid over warm convective ice is highly sensitive to the temperature of convecting ice. The 6 km conductive ice over warm convective ice at 265 K is stronger than the 5 km conductive ice lid over warm convective ice at 255 K down to depths of ~6 km, when the trend reverses and the strength becomes notably stronger for the cooler ice (Figure 4a). This might be the reason why a conductive layer needs to be thinner in order to produce the same depth crater when the lower, ductile ice layer is cooler and therefore stronger. Moreover, this may also explain why the conductive ice lid over warm convective ice scenario cannot produce the central peaks as large as the purely conductive ice shell over ocean. Although our results are robust for the best fit strength-depth profiles, they strongly depend on previously established parameters for ice based on laboratory experiments (Beeman et al., 1988). Any future improvements to ice parameters would also invoke an adjustment to thermal gradient, rollover temperature, and possibly conductive lid thickness.

### 4.3. Formation of Pits and Domes: A Postimpact Process?

Moore et al. (2017) reinforced the hypothesis that central pits and domes develop after the impact, suggesting that the morphology of the mature impact structure is strongly dependent on the melt pool size, volume, and its emplacement within ice. In principle, the water lens will undergo upwelling as it freezes out, altering the final crater morphology. Hence, the central pit and dome formation during the crater collapse does not have to be invoked if the crater d-D agree with observations. As noted by Moore et al. (2017), impactor composition and velocity will also affect the extent of postimpact formation of domes and pits. Schenk (2002) reported that all cataloged craters on Europa are unmodified and unrelaxed suggesting that postimpact formation of pits and domes proposed by Moore et al. (2017) may not affect crater depths.

Although the 8 km conductive ice shell over ocean and the 6 km conductive ice lid over warm convective ice scenarios lead to the very similar d-D trend, the latter is more conducive to formation of pits and domes.
4.4. Implications for Future Work

To evaluate and test the validity of the hypothesis that central pits and domes form post impact, more focused studies at higher resolution (to evaluate production of melt), with different impactor compositions and representative impact velocities at Europa, are recommended. While modeling at high impact velocities is computationally more demanding and expensive, future studies should also investigate morphological differences arising from impacts at varying velocities into icy bodies, such as Europa.

Test runs with larger impactors ($R_i > 405$ m) confirm relationships shown by Cox and Bauer (2015) that there is either a breach (ice shell over ocean) or the final crater becomes extremely shallow (nearly flat) and thus neither depth or diameter can be confidently discerned. While the resolutions used in our study are not sufficient to resolve any faulting indicative of rings formation, this might correspond to the progression to Transition III (Schenk, 2002). The ring structures should be very sensitive to the difference in rheology between warm convecting ice and liquid water (Singer et al., 2013). Thus, modeling the formation of multitiring basin on Europa may be provide important information about the structure of the ice shell and could be sensitive to ocean thickness even in the case of a conductive lid overlying warm convecting ice. The aim of future work will be to model these largest craters.

5. Conclusions

In our study, we modeled formation of impact craters on Europa to investigate thickness and internal structure of its ice shell. Our modeling results suggest that both a fully conductive shell (thin shell) and conductive ice over warm convective ice (thick shell) are capable of reproducing the observed crater morphologies and depth-diameter on Europa. If the ice shell is indeed thick, our study places an estimate on the conductive ice shell thickness between 7 and 10 km (Bray et al., 2014; Cox & Bauer, 2015). If central features within the crater, such as pits and domes, form during crater collapse, our simulations are in better agreement with a conductive-convective shell (thick shell). If central features form well after the impact, however, our simulations are more consistent with a conductive-convective shell (thick shell).


