Basin Crustal Structure at the Multiring Basin Transition

Evan Bjonnes1,2, Brandon C. Johnson3,4, and Jeffrey C. Andrews-Hanna5

Abstract Two impact basins on the Moon—Freundlich-Sharonov and Hertzsprung—are nearly the same size but exhibit different surface morphologies and subsurface structures. Gravity data reveal a bench-like transitional structure in the crust-mantle interface between the outer ring and the inner basin cavity in Hertzsprung, unlike that beneath both Freundlich-Sharonov and larger multi-ring basins. We use iSALE-2D to model the formation of impact basins into a 40-km thick pre-impact crust with a range of thermal conditions to understand the divergent development of these basins and gain insight into the factors affecting whether a basin forms with a peak-ring or multiring basin structure. We find that thermal gradients of at least 30 K/km result in Freundlich-Sharonov-type basins, in agreement with previous work. Cooler thermal gradients of approximately 15–20 K/km are needed to develop Hertzsprung-like multiring basins with observed bench-like structures in the crust-mantle topography. We find that for cooler models, the bench structure develops early in the cratering process as a rotated inner normal fault cutting the crust-mantle interface, whereas models with higher thermal gradients instead develop diffuse deformation zones instead of a discrete inner fault. The peak rings of both basins develop later in the cratering process as the collapsed central uplift. These results highlight the complex interplay between a strong lithosphere needed to develop ring faults and accessible ductile rocks that facilitate multiring basin formation. The varying thermal conditions giving rise to Hertzsprung and Freundlich-Sharonov impact basins may be a possible constraint on the lunar cooling rate and chronology.

Plain Language Summary Impact cratering is a complex process that depends on several factors describing a planetary body. We studied two large impact craters on the farside of the Moon, Freundlich-Sharonov and Hertzsprung basins, to understand how the temperatures of the lunar subsurface may have affected their development. We focus on recreating observed crater dimensions and subsurface features to evaluate our models. Specifically, we model the formation of a distinct “bench” feature between the crust and mantle layers that exists beneath Hertzsprung but is not present beneath Freundlich-Sharonov basin. We found that the temperatures at the time of impact play an important role in the nature of the structure that develops after a large impact. We successfully modeled basins that look like Freundlich-Sharonov using relatively warm temperature conditions and we modeled Hertzsprung-like basins using cooler temperature conditions. These varying thermal conditions directly relate to the character and location of faulting associated with each basin. These models underscore the effects derived from changing thermal conditions as a planetary surface cools and show that the impact craters that form on a planet or Moon are sensitive to these effects in different ways.

1. Introduction

The Moon, with its proximity to Earth and lack of erosional processes other than impacts, is an excellent natural laboratory to investigate planetary evolution. Developing a deeper understanding of lunar evolution not only teaches us about the Moon but also about the geologic evolution of terrestrial planetary bodies and broader Solar System evolution. Of particular importance are impact craters; impacts are a dynamic, ubiquitous process occurring on all solid surfaces, which makes them invaluable for teasing apart the nuances of how different planetary surfaces respond when subjected to extreme physical processes. When considering an individual planetary body, the range of thermal conditions that were present throughout its past could result in the prevalence of specific crater morphologies in different time periods, perhaps alluding to a relative chronologic sequence based on impact crater structure. To further investigate this possibility, we take the approach of comparing similarly
sized impact basins on the lunar farside and use their morphological differences to better understand the thermal evolution of the Moon.

Although both form as a result of planetary collisions, impact basins are morphologically distinct from impact craters by the development of two or more concentric topographic rings (Hartmann & Kuiper, 1962) and are subdivided into two categories: peak-ring basins and multiring basins. Lunar peak-ring basins have an annular topographic high that develops on the basin floor after the collapse of an unstable central uplift (Morgan et al., 2016) and are between 200 and 582 km in diameter (Neumann et al., 2015). Lunar multiring basins, on the other hand, are between 570 and at least 1,000 km in diameter (Neumann et al., 2015) and are surrounded by lithospheric-cutting faults that may comprise the outer rings of these basins (Andrews-Hanna et al., 2018). Ring tectonic theory is the primary hypothesis of multiring basin formation and describes the deformation caused by the flow of less viscous material at depth during the collapse of the transient crater (McKinnon, 1981; Melosh & McKinnon, 1978). In the case of a very large impact event, the transient crater penetrates to depths comparable to the deeper, ductile rocks below the brittle-ductile transition. When this happens, this deeper, less viscous material becomes involved during the crater collapse stage and its inward motion exerts a drag force on the base of the overlying brittle rocks, forming circumferential normal faults as it flows back toward the basin center (McKinnon, 1981; Melosh & McKinnon, 1978). The depth of the brittle-ductile transition is defined as the cross-over region of the brittle regime, where the increasing confining pressure increases the yield strength of the rocks, and the ductile regime, where elevated temperatures facilitate viscous deformation and decrease the strength of the rocks. The transition from peak ring to multiring basins denotes a shift in the dynamics of the cratering process from occurring primarily within the elastic lithosphere of the target body to one where deformation in the underlying ductile layer extends the affected area beyond the canonical impact basin. Consequently, the thermal state of the planetary body and its cooling history are highly connected to the depth of this transition zone and, accordingly, the scale of an impact that would be affected by the ductile layer at depth.

Recent modeling work supports the framework of ring tectonic theory and its description of the relationships between the pre-impact temperature and yield strength profiles in the target body and final basin morphology. Ivanov et al. (2010) described the effect of changing the thermal conditions on basins formed with projectiles greater than 150 km in diameter, conditions which would produce very large impact basins such as South Pole Aitken Basin on the Moon and Hellas Basin on Mars. Additionally, numerical models are able to resolve basin rings, demonstrating that hydrocode models are able to simulate multiring basin development (Bjonnes et al., 2021; Johnson et al., 2016, 2018); the deformation processes simulated in these multiring basin hydrocode models show that ring faults form in general accordance with ring tectonic theory. Further, ring fault spacing and location are highly sensitive to the thermomechanical state of the target, supporting the hypothesis that basin morphology is a function of both the size of the projectile as well as the thermomechanical state of the target body (Bjonnes et al., 2021; Johnson et al., 2016, 2018).

The Moon's geology is dominated by impact processes on both its nearside and farside, reflecting a rich history of impact cratering events through time. Impact basins on the nearside are generally larger than those on the farside but comparing structures of nearside and farside basins are complicated by the infilling of many nearside basins with lunar maria (Miljković et al., 2013). Moreover, the nearside is known to have a thinner crust compared to the farside (Wieczorek et al., 2013), further complicating drawing comparisons between impact structures located on different hemispheres. It is possible, however, to exploit these differences to gain a deeper understanding of lunar evolution; Miljković et al. (2013) determined that for given impactor size and velocity the basin produced on the relatively warm lunar nearside will be larger than a basin form on the farside. Here we explore the effect of temporal evolution of the Moon's thermal structure by comparing basins located on the same hemisphere, thus eliminating the effects of variable crustal thickness or heat flow between the two regions. When constrained in this way, it is possible to understand the variability of basin development within the broader context of lunar thermal evolution because impact basin development is so closely tied to the preimpact thermal state of the target body (Ivanov et al., 2010; Johnson et al., 2018; Miljković et al., 2013).

For this study, we focused on simulating the formation of two basins on the lunar farside—Hertzsprung Basin (Figure 1a) and Freundlich-Sharonov Basin (Figure 1b)—which are nearly equal in size but of different ages, and developed with different surface and subsurface morphologies. Freundlich-Sharonov is categorized as a peak-ring basin, has a diameter of 582 km, and has a peak ring encircling an inner basin with a diameter of 318 km. Hertzsprung is categorized as a multiring basin with a diameter of 571 km, a subtle inner ring fault at 408 km,
and a peak ring at 256 km diameter (Neumann et al., 2015; Wilhelms et al., 1987). Analysis of the Gravity Recovery and Interior Laboratory (GRAIL) gravity field (Zuber et al., 2013) allows for an interpretation of the subsurface characteristics. Inversions of the GRAIL Bouguer gravity field (e.g., Andrews-Hanna et al., 2018) reveal that these two basins show distinctly different patterns in the relief along the crust-mantle interface, with Hertzsprung showing a distinctive flattening of the interface (a “bench”) between the edge of the mantle plug and the deepest point of the interface. Although both basins are classified as pre-Nectarian/Nectarian in age, Freundlich-Sharonov has a higher density of 20 km-diameter impact craters superposed (N(20) of 140 ± 18, versus. 129 ± 22; Fassett et al., 2012) and has a markedly more degraded topographic signature. Thus, these two basins represent similar - sized structures formed in the crust of similar thickness, but presumably at different times and with different thermal conditions. Because both surface and subsurface structure are closely tied to the thermal conditions at the time of impact, we seek to determine if there is a genetic connection between the bench structure observed in the crust-mantle interface relief and the thermal structure at the time of impact and if there are any implications for lunar chronology.

By considering two similarly sized basins, which are both located on the lunar farside, we will directly compare how the lunar thermal conditions influenced their developments. Within this framework, we simulate impact crater formation using iSALE-2D focused on testing the effects of varying temperature profiles to constrain the thermal conditions in the crust and upper mantle and better understand how the thermal conditions affect the development of the observed features defining Freundlich-Sharonov and Hertzsprung. Because numerical impact simulations of Freundlich-Sharonov have been previously considered in the literature (e.g., Freed et al., 2014; Melosh et al., 2013), we primarily focus on determining the conditions which lead to a Hertzsprung-like impact structure. In this paper we begin by describing how we construct and evaluate our simulations (Section 2). We then present our model results, describing the basin-forming process and examining the effects of different parameters in detail (Section 3). Next, we discuss the implications of our work for lunar chronology and some details of basin-forming mechanics (Section 4). Finally, we conclude with a short summary (Section 5).

2. Methods
2.1. Gravity Analyses
The observed gravity and topography signatures can be inverted for a model of the relief along the crust-mantle interface and the resulting crustal thickness, based on assumptions of the densities of the crust and mantle and the mean thickness of the crust (Wieczorek, 2007; Wieczorek & Phillips, 1998). However, inversions of the gravity data for the relief along the crust-mantle interface become unstable at higher degrees as a result of the presence of small-scale density anomalies at shallow depths in the lunar crust (Jansen et al., 2017), which generate gravity anomalies that are unstably amplified when downward-continued to the crust-mantle interface. As a result, conventional crustal thickness models require strong filtering, with global crustal thickness models using a broad filter with an amplitude of 0.5 at spherical harmonic degree 80 (half-wavelength resolution of 68 km; Wieczorek

Figure 1. Topography of Hertzsprung (a) and Freundlich-Sharonov (b) basins. Dotted lines mark the approximate locations of the edge of topographic rings.
et al., 2013). However, for nearly circular impact basins, higher resolution gravity analyses can be accomplished by taking advantage of the symmetry of the basins (Andrews-Hanna, 2013; Kattoum & Andrews-Hanna, 2013). High resolution crustal thickness models of impact basins can be generated by smoothing more in the circumferential direction than in the radial direction, thereby diminishing the random small-scale gravity anomalies while preserving the radial basin structure (Andrews-Hanna et al., 2018). This smoothing can be accomplished in either the spatial domain or spherical harmonic domain, the latter by rotating the basin to the pole and using a degree- and order-dependent filter.

We adopt the model of Andrews-Hanna et al. (2018), in which the basins are rotated to the pole in the spherical harmonic domain and the crustal thickness is modeled using a degree- and order-dependent filter. A low-pass cosine taper was applied from degrees 120–140, corresponding to a half-wavelength resolution of 42 km in the radial direction. The azimuthal smoothing was implemented with a degree- and order-dependent filter, applied to wavelengths shorter than 4–6° of arc, for an azimuthal resolution of ~150 km at the pole. The resulting model of the relief along the crust-mantle interface provides a ~60% improvement in radial resolution compared to previous global crustal thickness models, while still resolving large-scale azimuthal variations in structure. Importantly, with basin radii of ~280 km and ring spacings of ~160 km, this model fully resolves the structure of the basin rings of Freundlich-Sharonov and Hertzspring. Crust and mantle densities were assumed to be 2,500 kg/m³ and 3,300 kg/m³, respectively, with a mean crustal thickness of 34 km (Wieczorek et al., 2013). Both basins are largely symmetric, so radial profiles were calculated by averaging azimuthally over the entire basin.

2.2. Hydrocode Modeling

We model impact crater formation on the Moon using iSALE-2D, a hydrocode developed to model hypervelocity impacts using a variety of planetary materials (Amsden et al., 1980; Collins et al., 2004; Ivanov et al., 1997; Melosh et al., 1992; Wünnemann et al., 2006). We include several model components tied to specific impact-related deformation mechanisms to accurately approximate the deformation behavior of the rocks: (a) the Collins dilatancy model (Collins, 2014), whereby dilatancy (an increase in porosity due to rocks deformed by shear failure; Reynolds, 1885) is a function of pressure, temperature, and distension, (b) the Collins damage model (Collins et al., 2004), which relates the damage accumulated in a model cell to the brittle-ductile and brittle-plastic transition pressures, and (c) an exponentially dependent damage model, which enhances strain localization along previously fractured rocks (Johnson et al., 2016; Montési & Zuiber, 2002). We evaluate final basin morphologies based on the post-impact crust-mantle topology and the development and spacing of ring faults associated with multiring basins. Multiring basin development is tightly related to the patterns of deformation deep in the subsurface; to fully model this deformation, we include visco-elastic plastic rheology for dunite comprising warm mantle material (Elbeshausen & Melosh, 2020). We considered including acoustic fluidization, a process incorporated to simulate additional necessary weakening in crater formation, in our models. However, after comparing the results of models with and without acoustic fluidization, we found that our final crater morphologies were unaffected by including it in the model and we consequently omitted acoustic fluidization to save computational expense, e.g. Johnson et al. (2016, 2018).

We focus on simulating lunar basins approximately 550–600 km in diameter, spanning the diameter range which results in either a peak-ring or multiring basin morphology (Neumann et al., 2015). Because basins of this size are not significantly affected by planetary curvature (e.g., Johnson et al., 2016), we use the two-dimensional, axisymmetric implementation of iSALE with a flat target geometry. Table 1 documents the physical properties of the iSALE mesh. High resolution cells are spaced at 1 km, corresponding to 20 cells per projectile radius. Projectiles are 40 km in diameter, comprised of dunite and strike the target vertically at 15 km/s (Le Feuvre & Wieczorek, 2011). The lunar target surface is composed of a 40-km thick crust overlying dunite mantle. We use the analytical equation of state for both granite and dunite (Table 2) and gabbroic intact and damaged strength parameters for the crust. Although granite is compositionally distinct from anorthosite, we use the granitic equation of state given its similar density compared to the lunar crust (Melosh et al., 2013). Similarly, we employ

<table>
<thead>
<tr>
<th>Table 1</th>
<th>iSALE Model Setup Parameters</th>
</tr>
</thead>
<tbody>
<tr>
<td>Description</td>
<td>Value</td>
</tr>
<tr>
<td>Cell Size</td>
<td>1 km</td>
</tr>
<tr>
<td>Cells Per Projectile Radius</td>
<td>20</td>
</tr>
<tr>
<td>Number of horizontal high-resolution cells</td>
<td>350</td>
</tr>
<tr>
<td>Number of vertical high-resolution cells</td>
<td>400</td>
</tr>
<tr>
<td>Physical size of mesh, horizontal direction</td>
<td>1.646 km</td>
</tr>
<tr>
<td>Physical size of mesh, vertical direction</td>
<td>2.992 km</td>
</tr>
<tr>
<td>Impact velocity</td>
<td>15 km/s</td>
</tr>
<tr>
<td>Surface gravity</td>
<td>1.62 m/s²</td>
</tr>
</tbody>
</table>
the strength parameters appropriate for gabbroic anorthosite, which is representative of the lunar crust (Potter et al., 2012) and to maintain consistency with previous work on lunar impacts. Both crustal and mantle material deform in accordance with the Collins et al. (2004) damage model.

Basin ring structure and crater collapse are driven by the thermal structure of the crust and upper mantle (Bjønnes et al., 2021; Johnson et al., 2018). Following previous work (e.g., Freed et al., 2014; Johnson et al., 2016, 2018) all thermal profiles have a surface temperature of 300 K and transition to an adiabatic temperature gradient corresponding to a rollover temperature of 1300 K. The lithospheric thermal gradients in our models vary between 5 and 40 K/km in increments of 5 K/km. These thermal gradients correspond to surface heat flows ranging from 10 to 80 mW/m², assuming a thermal conductivity of 2 W/m K (Maurice et al., 2020). For comparison, a model of the thermal evolution of the Moon assuming the sequestration of the majority of the KREEP-rich material on the nearside predicts far side heat fluxes ranging from 45 to 9 mW/m² from 4.5 Ga to today (Laneuville et al., 2013). In the shallow subsurface, yield strength increases due to the increasing lithostatic pressure (Collins et al., 2004). The temperature also increases with depth, resulting in a competing effect that as the temperature approaches the melt temperature, the yield strength of the target material decreases (Ohnaka, 1995). Figure 2 shows how changing the lithospheric thermal gradient alters the resulting yield strength profile in the preimpact target subsurface.

Table 2
Summary of iSALE Material Input Parameters

<table>
<thead>
<tr>
<th>Description</th>
<th>Crustal value</th>
<th>Mantle value</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>Equation of state</td>
<td>Granite (ANEOS)</td>
<td>Dunite (ANEOS)</td>
<td>Benz et al. (1989), Pierazzo et al. (1997)</td>
</tr>
<tr>
<td>Surface Temperature</td>
<td>300 K</td>
<td>N/A</td>
<td></td>
</tr>
<tr>
<td>Melting temperature</td>
<td>1513 K</td>
<td>1373 K</td>
<td>Davison et al. (2010), Potter et al. (2012)</td>
</tr>
<tr>
<td>Thermal Softening parameter</td>
<td>1.2</td>
<td>1.1</td>
<td>Davison et al. (2010), Potter et al. (2012)</td>
</tr>
<tr>
<td>Simon A parameter</td>
<td>1840 MPa</td>
<td>1520 MPa</td>
<td>Davison et al. (2010), Potter et al. (2012)</td>
</tr>
<tr>
<td>Simon C parameter</td>
<td>7.27</td>
<td>4.05</td>
<td>Davison et al. (2010), Potter et al. (2012)</td>
</tr>
<tr>
<td>Poisson's ratio</td>
<td>0.25</td>
<td>0.25</td>
<td>Davison et al. (2010), Potter et al. (2012)</td>
</tr>
<tr>
<td>Frictional coefficient (undamaged)</td>
<td>1.1</td>
<td>1.58</td>
<td>Davison et al. (2010), Potter et al. (2012)</td>
</tr>
<tr>
<td>Frictional coefficient (damaged)</td>
<td>0.71</td>
<td>0.63</td>
<td>Davison et al. (2010), Potter et al. (2012)</td>
</tr>
<tr>
<td>Strength at infinite pressure</td>
<td>2.49 GPa</td>
<td>3.26 GPa</td>
<td>Davison et al. (2010), Potter et al. (2012)</td>
</tr>
<tr>
<td>Cohesion (undamaged)</td>
<td>31.9 MPa</td>
<td>5.07 MPa</td>
<td>Davison et al. (2010), Potter et al. (2012)</td>
</tr>
<tr>
<td>Cohesion (damaged)</td>
<td>0.01 MPa</td>
<td>0.01 MPa</td>
<td>Davison et al. (2010), Potter et al. (2012)</td>
</tr>
<tr>
<td>Brittle ductile transition pressure</td>
<td>1.23 GPa</td>
<td>1.23 GPa</td>
<td>Collins et al. (2004)</td>
</tr>
<tr>
<td>Brittle plastic transition pressure</td>
<td>2.35 GPa</td>
<td>2.35 GPa</td>
<td>Collins et al. (2004)</td>
</tr>
<tr>
<td>Initial tensile strength</td>
<td>10 MPa</td>
<td>10 MPa</td>
<td>Collins et al. (2004)</td>
</tr>
<tr>
<td>Maximum distension</td>
<td>1.2</td>
<td>1.2</td>
<td>Collins (2014)</td>
</tr>
<tr>
<td>Maximum dilatancy coefficient</td>
<td>0.045</td>
<td>0.045</td>
<td>Collins (2014)</td>
</tr>
<tr>
<td>Dilatancy pressure limit</td>
<td>200 MPa</td>
<td>200 MPa</td>
<td>Collins (2014)</td>
</tr>
<tr>
<td>Frictional coefficient (maximum distension)</td>
<td>0.4</td>
<td>0.4</td>
<td>Collins (2014)</td>
</tr>
</tbody>
</table>

*Parameters to describe a rock-like strength model (Collins et al., 2004). *Parameters to describe the pressure-dependent melt temperature of a material (Wünnemann et al., 2008). *Parameters to describe dilatancy in iSALE (Collins, 2014).
The variability of yield strength in the shallow subsurface is evident in Figure 2b, where lower lithospheric thermal gradients result in colder temperatures in the crust and upper mantle, shifting the brittle-ductile transition (defined here as the depth of maximum strength) down to approximately 100 km for a 5 K/km thermal gradient. Basin development is sensitive to a depth on the order of the transient crater depth (Melosh, 1989; Schenk, 2002), predicting that the formation of an impact crater from a 40km diameter projectile would be sensitive to the rock properties down to approximately 125 km subsurface. Consequently, we expect basin development to be highly sensitive to the near-surface thermal profiles considered.

3. Results

3.1. Crustal Thickness and Basin Structure

In both map view and profile (Figure 3), both basins show the characteristic pattern of a central zone of thinned crust corresponding to the positive Bouguer anomaly (Neumann et al., 2015) surrounded by a zone of crustal thickening from the basin ejecta (Wieczorek & Phillips, 1999). The central zone of thinned crust is well-resolved in our models, taking the form of a flat-roofed column of uplifted mantle whose top diameter corresponds approximately to the inner ring diameter in the topography. The crustal thickening is contained entirely within the outer rings of both basins, leading to the zone of thickest crust being in the region of low topography between the inner and outer rings. The transition between the surrounding background crustal thickness and the thickening by ejecta occurs approximately where the outer ring fault would be expected to intersect the crust-mantle interface (Andrews-Hanna et al., 2018). As a result, it is not possible to confirm the presence of ring-faults crossing the crust-mantle interface using gravity data because a change in crustal thickness at this radial distance could be due to deformation from a ring fault, transition to the background crustal thickness, or both. The mantle uplift beneath Hertzprung is significantly narrower than that beneath Freundlich-Sharonov (diameters of 230 and 300 km, respectively, measured where the crustal thickness first attains the background value), despite the similar diameters of the topographic outer rings (∼570 and 580 km, respectively; Neumann et al., 2015).

The structures of the basins differ significantly in the zone between the rings. In Freundlich-Sharonov, the maximum crustal thickness is attained immediately outside the central zone of thinned crust, as also observed for larger multi-ring basins and consistent with the expected for the distribution of basin ejecta (Andrews-Hanna et al., 2018; Wieczorek & Phillips, 1999). In contrast, Hertzsprung exhibits a pronounced bench-like flattening of the crust-mantle interface relief at a level intermediate between the uplifted crust-mantle interface beneath the basin floor and the depressed interface in the zone of crustal thickening. Models were tested with a range of filters at different resolutions, and we found the bench-like structure to be a robust prediction. Although we tested modes out to spherical harmonic degree 170, we find that the results for degree 130 are the best balance

Figure 2. Preimpact temperature profiles (left) and corresponding static yield strength profiles (right) for models with 40 km crustal thickness and lithospheric thermal gradients of 5, 10, 20, and 40 K/km.
of attaining a high-resolution representation of the bench structure without introducing artificial ringing artifacts in the data that present at higher harmonic degrees. This bench does not resemble the ring structures around the larger Oriental multi-ring basin, in which the successive concentric ring faults make a stair-step pattern of increasing crust-mantle interface depth toward the basin center (Andrews-Hanna et al., 2018). A similar bench is not observed at Freundlich-Sharonov. The bench is not simply an artifact of the model as evidenced by the observation that the Bouguer gravity anomaly in this zone is transitional between the central gravity high and the
gravity low further beneath the thickened ejecta, unlike Freundlich-Sharonov in which the minimum Bouguer anomaly is immediately outside the central Bouguer high. Although the bench is near the limit of the resolution of the crustal thickness model, a pair of positive and negative gravity gradient anomalies at the outer edge of the bench (Figure 3; Figure S1 in Supporting Information S1) support the presence of a discrete change in crustal thickness at this location. This bench represents a zone of transitional crustal thickness outside the basin excavation cavity that is not found around Freundlich-Sharonov or larger basins. Similar benches could occur around smaller peak-ring basins but are not resolved. When matching these models to the iSALE outputs in the next section, we focus on the averaged crustal thickness profiles and crust-mantle interface structure in Figures 3c and 3d, limiting against overinterpreting small-scale model artifacts which may be apparent in map-view at the limit of model resolution.

3.2. Hydrocode Modeling

Our models result in basins with a range of final basin diameters depending on the thermal gradient implemented, consistent with previous work considering the thermal effect on basin size (e.g., Ivanov et al., 2010; Miljković et al., 2013, 2016; Johnson et al., 2018). We have modeled the formation of a peak-ring basin resembling Freundlich-Sharonov and multiring basin resembling Hertzsprung. Our models show that the development of ring faults, the primary means of identifying a multiring basin, depends on the model’s thermal gradient. In this section, we first outline the key steps during impact basin formation and then describe the effects of incorporating different thermal gradients.

Our baseline model is a 40 km diameter projectile striking a Moon-like target with a 40 km crustal thickness and 20 K/km lithospheric thermal gradient; this model setup results in a basin that resembles Hertzsprung in size, morphology, and topography along the crust-mantle interface. The development of this impact basin is shown through a series of key time steps in Figure 4. Approximately 75 s following the initial contact, the transient crater reaches its maximum depth of 125 km and diameter of 200 km (Figure 4a). At this point, the transient crater is a gravitationally uncompensated cavity and begins to undergo crater collapse, a stage of the impact process characterized by uplift of the crater floor and widening of the crater diameter (Figure 4b). During the crater collapse stage, material from up to 200 km away is accelerated inward and causes fault development (Figure 4c). The volume of material moving inward forms a central uplift which then also collapses. After the subsequent collapse of the central uplift, crustal material is pushed up in a bulge approximately 100 km from the basin center (Figure 4d). The central uplift at this time is also unstable, and material collapses inward again (Figure 4e). The formation and collapse of this bulge form the basins peak ring. This mechanism for peak ring formation in large basins is consistent with the simulations of Johnson et al. (2016, 2018). Crustal material flowing inward eventually generates a crustal cap covering the uplifted mantle in the basin center (Figure 4f). We note that this model develops ring faults associated with multiring basins, although the model rings develop at slightly different spacings from the basin center compared to observations. For this Hertzsprung-like model, ring faults develop at 440 and 580 km diameter, slightly different from Hertzspring’s ring faults observed at 408 and 582 km diameter (Figure 5).
find that this model develops an inner fault at approximately 250 km diameter, which cuts across the crust-mantle interface and ultimately becomes the bench feature, but this feature is subsequently covered by the peak ring with a diameter of approximately 250 km. The inner ring forms as crustal material collapses toward the basin center (Figure 4c) but is subsequently covered by that crustal material as it is folded over and outward during the uplift and then collapse of the mantle within the transient cavity leading to the loss of the surface expression of the fault (Figure 4d). All ring faults initially form with typical normal fault dips of ∼60°, but the innermost ring fault which demarks the bench is rotated during basin collapse to a final dip of <30°. The initial fault angle is consistent with Anderson's theory of faulting and the amount of inner fault rotation is directly related to how much underlying material is drawn to the basin center, dragging the deeper faulted rocks inward and ultimately upward during crater collapse. The innermost ring fault also differs from the outer ring faults in that it intersects the steep edge of the uplifted mantle beneath the basin center. The combination of the low dip and the intersection of the fault with the edge of the mantle plug result in the observed bench-like morphology and transitional crustal thickness between this innermost ring and the basin center.

We note that some mantle material is also excavated onto the surface. However, because axisymmetric simulations often exaggerate central uplift heights leading to overestimations of the degree of splash-out, we interpret this as a model artifact and not a reliable result. Including high melt viscosities which limit this mantle splashing is one way to limit final basin distortions due to these exaggerated central uplifts (Johnson et al., 2016).

The thermal conditions set at the start of the simulation have a substantial effect on final basin morphology. The stages of basin development outlined above describe an impact into a target with a 20 K/km thermal gradient and a 40 km thick crust, resulting in a final basin with a peak ring at approximately 250 km diameter, similar to the diameter of Hertzsprung's observed peak ring (256 km diameter). Figure 6 shows final basin features, including ring faults, as a function of thermal gradient using initially horizontal tracer lines to show subsurface deformation. We use tracer lines instead of cell-properties such as accumulated strain because tracer locations are not affected by numerical diffusion, whereas modeling faults this long after impact using a cell-centered variable such as accumulated strain may be highly affected by numerical diffusion at long model run-times. We find that although the post-impact locations of the bench feature and surface peak ring correlate, they are unrelated features (Figure 6b). The bench structure initially develops as the innermost normal fault along the crust-mantle interface during the early stages of crater collapse as mantle material moves inward toward the basin center (Figure 7a). The bench structure persists through additional phases of deformation as the central uplift collapses and pushes

![Figure 5. Tracer plot showing the surface locations of ring faults associated with basin formation. Tracer lines are initially horizontal and spaced 2 km apart and are colored according to material type, where brown tracers are crustal composition and green tracers are mantle composition. Red arrows denote the surface locations of ring faults associated with the basin formation and the blue arrow approximates the location of the peak ring.](Image)

![Figure 6. Tracer plots showing the deformation patterns in the subsurface 5,000 s post-impact into 40 km thick crust. Crust material is brown and mantle material is green; tracers more than 25 km apart are not connected. Thermal gradient is displayed in the upper right corner of each panel. Linearly organized bends in the tracer lines are interpreted as faults.](Image)
material outward at depths between 50 and 100 km, but only below the crust-mantle interface (Movie S1). We find that simulations with lithospheric thermal gradients of 30 K/km and higher do not develop the bench structure in the crust-mantle interface because normal faults do not develop close to the basin due to the higher overall ductility of the crust and mantle material (Figure 6d). Instead, these models predict a single normal fault defining the outer basin ring. We also note that the diameter of the mantle uplift is narrower relative to the outer ring fault crossing at the crust-mantle interface for the lower thermal gradient model (100 vs. 250 km) than for the higher thermal gradient model (150 vs. 260 km).

We find that our models with higher thermal gradients generally develop thicker crustal caps compared to models with lower thermal gradients (Figure 6, Figure S2 in Supporting Information S1) except for models with 5 K/km. For this very cold case, the yield strength of the target is so high as to resist as much excavation and displacement observed in the simulations with warmer starting conditions. This stronger yield strength condition causes more crustal materials to essentially stay in their original positions during basin development, resulting in a thicker crustal cap in the basin center. However, because the final basin structure for this scenario does not match any of the other observational constraints such as basin diameter and ring fault locations, we do not consider it a match to either Freundlich-Sharonov or Hertzsprung. Figure 3 shows crustal thickness estimates using the Bouguer gravity anomaly from GRAIL and suggests that there is a difference between the crustal cap thicknesses of these two impact basins: approximately 18 km at Freundlich-Sharonov and 25 km at Hertzsprung. Although our

Figure 7. X- and Y-velocities of target material in the crust and upper mantle during two simulations, one with 20 K/km (a) thermal gradient and the other with 30 K/km thermal gradient (b), each taken at 700 s post-impact. Thick black line denotes the crust-mantle boundary. For the model with 20 K/km thermal gradient, distinct fault planes develop early in the crater collapse stage (marked with black arrows). Models with thermal gradients of 30 K/km and higher, however, developed diffuse fault zones in the cratering process (marked by a bracket).
hydrocode models predict that Hertzsprung would have a thinner crustal cap than Freundlich-Sharonov (Figure S3 in Supporting Information S1), we do not use crustal cap thickness to evaluate our models given the high sensitivity of crustal cap thickness to preimpact crustal thickness. This discrepancy could be due to a difference in local preimpact crustal thicknesses at each basin location or unreliability of basin-centered, axisymmetric models such as iSALE near the point of impact. Tests with crustal thicknesses between 35 and 45 km confirm that changing the preimpact crustal thickness affects the crustal layer along the final basin floor by approximately the same amount, that is, models with 5 km thicker preimpact crustal thickness have 5 km thicker crustal caps within the basin structure.

Crustal thickness is another metric for evaluating hydrocode models against observations, although these plots can be affected by the effects of preimpact crustal thickness, as described above. When plotted against the GRAIL-derived crustal thickness inversions, we find that the model with 15 K/km thermal gradient produces a bench-like feature in the crustal thickness distribution, which more closely matches the variations observed around Hertzsprung basin (Figure S4 in Supporting Information S1). However, the model with a slightly higher thermal gradient of 20 K/km better replicates the surface features associated with Hertzsprung. For this reason, we conclude that appropriate thermal conditions in the crust and upper mantle at the time Hertzsprung formed are in the range of 15–20 K/km.

The gravitational signature of farside lunar basins is characterized by a negative free air gravity anomaly in a sub-isostatic annulus and a central positive gravity anomaly or mascon, with the coupling between sub-isostatic annulus in the crustal collar causing flexural uplift of the basin center producing the mascon geometry (Andrews-Hanna et al., 2018; Freed et al., 2014; Melosh et al., 2013). We find that models with a lower thermal gradient are left in a more strongly sub-isostatic state at the end of the simulation while models with a higher thermal gradient are closer to, but still do not attain, isostasy (Figure S5 in Supporting Information S1). Similarly, Hertzsprung is surrounded by a more strongly negative free air anomaly than Freundlich-Sharonov.

4. Discussion
4.1. Development of the Bench Structure and Peak Ring

The surface observations that define the Hertzsprung and Freundlich-Sharonov basin morphologies are well-documented but the subsurface character has thus far remained largely unknown. This hydrocode model of Hertzsprung, however, highlights how the development of the bench structure arises from an interplay between the ductile flow of the underlying mantle and the relatively brittle behavior of the crust above it as well as how its formation is different from Freundlich-Sharonov. A detailed examination of material velocities throughout the cratering process illustrates the interplay between these variables nicely (Figure 7, Movie S1); in the development of both Hertzsprung-like and Freundlich-Sharonov-like basins, there is a period of time during the crater collapse stage where material slumps inward. For the Hertzsprung-like model with an approximately 20 K/km thermal gradient, this inward slumping occurs along a distinct inner fault plane that develops approximately 700 s post-impact and extends through the crust-mantle interface, creating the bench structure (Figure 7a). For Freundlich-Sharonov-like models with thermal gradients 30 K/km or higher, however, the strain is accommodated through ductile deformation instead of brittle faulting and consequently the basin develops a diffuse zone of deformation rather than discrete fault planes. This difference in crust-mantle topography relates directly to the preimpact thermal gradient, where Hertzsprung-like models with bench features develop when the crust-mantle interface temperature is approximately 1000 K and is within the conductive portion of the thermal profile. Basins which resemble Freundlich-Sharonov, however, correspond with preimpact thermal profiles whereby the upper mantle is in the convective thermal regime at or above 1300 K. The final crust-mantle signature of these basins is a smooth boundary along the base of the thickened crustal collar, similar to the crust-mantle character observed in previous studies (Figure 7b; e.g., Freed et al., 2014; Johnson et al., 2016, 2018). We note that a similar bench-like structure is predicted in some azimuths around the Orientale basin (Andrews-Hanna, 2013). Azimuthal variability in the basin structure may be related to azimuthal variations in the nature of the collapse of the transient cavity, which could be related to differences in heat flow or crustal thickness. Orientale is a younger basin and thus likely formed in a cooler crust, consistent with our conclusion that lower thermal gradients favor the formation of a bench-like structure at the edge of the uplifted mantle plug. However, the larger size and different crustal thicknesses at Orientale prevent a quantitative comparison to Hertzsprung.

Interestingly, peak ring development occurs later in the basin formation process and is fully independent of any features along the crust-mantle interface, including the bench structure associated with Hertzsprung Basin. For
our models which develop benches, these features develop in the earliest stages of crater collapse when the transient crater walls first begin to move inward. At this time, the brittle deformation along the crust-mantle interface develops and remains throughout the rest of the crater formation process even after motion along the initial fault plane has stopped. The peak ring, however, forms during the collapse of the central uplift that forms during crater collapse regardless of the thermal gradient in the crust and upper mantle. Large impact events resulting in both Freundlich-Sharonov and Hertzsprung-like basins generate such large central uplifts during crater collapse that they fall under the force of gravity and push some of the uplifted material outwards and above the existing basin floor. The peak rings of both Hertzsprung and Freundlich-Sharonov form the leading edge of this collapsing material flowing along the basin floor and up onto the surrounding crust. Because they are generated during the latter stages of crater collapse, these peak rings are unrelated of any structure in the crust-mantle interface acquired during the earlier moments of crater collapse. The diameter of the peak ring structures increases with increasing thermal gradient (Johnson et al., 2018). Our simulations with thermal gradients of 20, 30, and 40 K/km produce peak rings with diameters of 250, 300, and 320 km, respectively. Despite similar basin sizes, the peak ring of Hertzsprung is 256 km in diameter while the peak ring of Freundlich-Sharonov is 318 km in diameter. Thus, the peak ring location of our simulation also suggests that the Hertzsprung is best reproduced using a thermal gradient of 20 K/km, while Freundlich-Sharonov requires a thermal gradient of at least 30 K/km. These same conditions also best reproduce the crust-mantle topography of these basins. The self-consistency of these results improves our confidence in our simulations and also further demonstrate the importance of subsurface structure in constraining basin formation models.

Our model results agree with previous lunar studies examining the effects of temperature on large basin development. We find that for models with the same pre-impact crustal thickness, higher thermal gradients lead to thicker crustal caps and a wider crustal annulus, conclusions which are in agreement with previous work (e.g., Freed et al., 2014; Miljković et al., 2016). Transient craters formed in targets with higher thermal gradients are wider and thus displace more crustal material early in the cratering process while also facilitating a larger amount of crustal return flow over the exposed basin floor during crater collapse, an interpretation supported by Johnson et al. (2018).

The different basin morphologies of Freundlich-Sharonov and Hertzsprung highlight key facets of ring tectonic theory, the current theoretical framework of multiring basin development. The higher thermal gradient needed for Freundlich-Sharonov suppresses ring fault development relative to a cooler impact with the same projectile characteristics. Although cooler thermal conditions favor multiring basin development, there are tradeoffs between the thermal and mechanical effects. The elastic lithosphere must have a certain degree of yield strength to facilitate brittle deformation and fault development but cannot be so thick as to render the underlying ductile material uninvolved during the cratering process. If the deeper ductile rocks are underneath too thick of an overlying elastic lithosphere as to facilitate enough inward movement during crater collapse, transient crater generated during a large impact event will form a peak-ring basin with one ring fault as its crater wall.

### 4.2. Implications for Lunar Chronology

Understanding the details of how these similar-sized basin-forming impacts lead to observably different basin structures highlights the very important role of temperature in the development of an impact structure. The only difference between the basins modeled in Figures 6 and 7 is the preimpact lithospheric thermal gradient. We find that higher thermal gradients are needed to replicate Freundlich-Sharonov Basin, consistent with previous work (Freed et al., 2014). This conclusion is further supported by our current understanding of lunar chronology given that Freundlich-Sharonov is a pre-Nectarian Basin and Hertzsprung is a Nectarian Basin (Fassett et al., 2012; Orgel et al., 2018; Wilhelms et al., 1987), where secular cooling of the Moon will result in lower surface heat flows and, accordingly, lower lithospheric thermal gradients later in the Moon’s history (Laneuville et al., 2013). However, more recent work yields similar crater ages for these basins with overlapping uncertainty (Fassett et al., 2012). By selecting basins in similar geologic settings (the lunar farside), we have avoided complications like variability in heat-producing elements in the crust such that the conclusion that similarly sized impacts may develop differently depending on the preimpact thermal conditions is robust (e.g., Miljković et al., 2013). With this in mind, we find that using Freundlich-Sharonov and Hertzsprung Basins as case studies to understand how thermal conditions in the lunar crust and upper mantle affect basin development illustrates that the conditions under which different impact structures develop are ever-changing, and secular cooling is enough to fundamentally change the rock behavior down to the subsurface structure of the crust-mantle interface.
The highest thermal gradient for which the inner bench does not form (30 K/km) would correspond to a heat flow of ~60 mW/m². This heat flow is at the upper limit of radiogenic heat production predictions for the lunar farside, possible only during the first few 100 Myr in thermochemical evolution models (e.g., Laneuville et al., 2013; Maurice et al., 2020). However, much higher temperatures and surface heat flows would occur during the early cooling of a hot crust formed by flotation above a magma ocean and may persist for timescales on the order of 100 Myr (Elkins-Tanton et al., 2011), comparable to the thermal diffusion timescale of the crust. Thus, the high heat flows consistent with the morphology of Freundlich-Sharonov suggest a very early formed impact basin, shortly after magma ocean crystallization. This pushes the possible age of Freundlich-Sharonov Basin back to 4.5–4.2 Ga based on revised solar system evolution models and the timing of the lunar magma ocean crystallization (e.g., Elkins-Tanton et al., 2011; Morbidelli et al., 2018). Despite having similar crater densities (e.g., Fassett et al., 2012), the degraded topography, bench-like multi-ring structure, and weaker sub-isostatic annulus of Freundlich-Sharonov all support a significantly warmer crust at the time of impact than for Hertzsprung. This difference may imply either a rapid cooling of the crust or short time lag or a lower impact flux or longer time lag between these two impacts.

With enough time after the impact, viscous relaxation is expected to occur to impact basins which form on relatively warm target. Recent work by Ding and Zhu (2022) found that impact basins forming in crust with lithospheric thermal gradients of 40 K/km and higher will likely experience a high degree of viscous relaxation, predicting total erasure of the crustal collar around Freundlich-Sharonov. Freed et al. (2014), who used a stronger crustal rheology than Ding and Zhu (2022), found negligible relaxation of the crustal annulus for Freundlich-Sharonov formed in a target with a 30 K/km thermal gradient. This is also consistent with work by Kamata et al. (2015) who found that substantial basin relaxation is expected to occur for crust-mantle interface temperatures >1300 K. Our conclusion of a 30 K/km thermal gradient at the time Freundlich-Sharonov formed is in agreement with previous hydrocode modeling of Freundlich-Sharonov (e.g., Freed et al., 2014) and, although close to the threshold determined by Ding and Zhu (2022), is within the parameter space whereby crustal collar relaxation is inhibited.

5. Conclusions

The Moon offers ample opportunity to study impact basin formation and the various factors that can alter basin morphology. Two lunar farside basins, Freundlich-Sharonov, and Hertzsprung, are nearly the same size and yet, in addition to being recognized as having different basin morphologies, they differ in subsurface and crust-mantle structure, where Hertzsprung exhibits a bench morphology in the crust-mantle interface and Freundlich-Sharonov does not. In this work, we used iSALE-2D, a shock-physics modeling code, to determine what factors may have contributed to the different surface and subsurface features based on the crust-mantle structure beneath the two basins.

Our models tested the effect of lithospheric thickness on final basin structure by changing the near-surface temperature profiles. Because preimpact geologic conditions affect basin development (e.g., Johnson et al., 2018), we focused our efforts on two farside impact basins with similar pre-impact crustal thickness and composition as case studies. We find that higher thermal gradients are associated with wider basins and more ductile deformation such that the fault formation is suppressed near the point of impact. These higher subsurface temperatures also cause a more pronounced degree of crater collapse, generally resulting in thicker crustal caps along basin floors. However, we note the differences in modeled crustal cap thicknesses of Freundlich-Sharonov and Hertzsprung Basins could also be due to variations in the preimpact crustal thickness at each location. GRAIL-derived crustal thickness maps suggest that small-scale variations in crustal thickness where the crust around Freundlich-Sharonov is approximately 40 km thick and the crust around Hertzsprung is approximately 50 km thick (Wieczorek et al., 2013), and Freed et al. (2014) demonstrate the high sensitivity of crustal cap thickness to preimpact crustal thickness. Finally, ring fault spacing is sensitive to projectile size (Johnson et al., 2018) and we note that our choice of projectile size is slightly too large, resulting in modeled basins which are slightly larger than Hertzsprung and Freundlich-Sharonov. Even with these considerations, the connection between the initial lithospheric thermal gradient and the final peak-ring or multiring basin morphology is well-illustrated through this study.

Data Availability Statement

iSALE is not currently available to the public and is accessible to the impact community on a case-by-case basis for non-commercial use. Scientists interested in using or developing iSALE can reference https://github.com/isale-code/iSALE2D for a description and application requirements. Model input files and outputs are uploaded to the Harvard Dataverse (Bjonnes, 2022).
Acknowledgments
The authors thank Shigeru Wakita for his programming help and John Brunton for insightful visualizations. The authors gratefully acknowledge the developers of isALE-2D and pyALEplot, including Gareth Collins, Kai Winnenmann, Dirk Elbershausen, Tom Davison, Boris Ivanov, and Jay Melosh. The authors thank Peter James and an anonymous reviewer for their helpful and insightful comments. This work was supported by Grant 80NSSC17K0341 from the NASA Lunar Data Analysis Program.

References


References From the Supporting Information