

Formation of Phobos and Deimos via a giant impact



Robert I. Citron^{a,*}, Hidenori Genda^b, Shigeru Ida^b

^a Department of Earth and Planetary Science, University of California Berkeley, 307 McCone Hall, Berkeley, CA 94720, USA

^b Earth-Life Science Institute, Tokyo Institute of Technology, 2-12-1-IE-1 Ookayama, Meguro-ku, Tokyo 152-8550, Japan

ARTICLE INFO

Article history:

Received 5 January 2015

Revised 6 February 2015

Accepted 7 February 2015

Available online 17 February 2015

Keywords:

Mars

Mars, satellites

Impact processes

Satellites, formation

ABSTRACT

Although the two moons of Mars, Phobos and Deimos, have long been thought to be captured asteroids, recent observations of their compositions and orbits suggest that they may have formed from debris generated by one or more giant impacts of bodies with $\sim 0.01\times$ target mass. Recent studies have both analytically estimated debris produced by giant impacts on Mars and numerically examined the evolution of circum-Mars debris disks. We perform a numerical study (Smoothed Particle Hydrodynamics simulation) of debris retention from giant impacts onto Mars, particularly in relation to a Borealis-scale giant impact ($E \sim 3 \times 10^{29}$ J) capable of producing the Borealis basin. We find that a Borealis-scale impact is capable of producing a disk of mass $\sim 5 \times 10^{20}$ kg ($\sim 1\text{--}4\%$ of the impactor mass), sufficient debris to form at least one of the martian moons according to recent numerical studies of martian debris disk evolution. While a Borealis-scale impact may generate sufficient debris to form both Phobos and Deimos, further studies of the debris disk evolution are necessary. Our results can serve as inputs for future studies of martian debris disk evolution.

© 2015 Elsevier Inc. All rights reserved.

1. Introduction

The origin of the two moons of Mars, Phobos and Deimos, is widely debated. One prevailing theory is that the two moons are captured asteroids, as indicated by their size, shape, and prior observations of their composition (Murchie et al., 1991; Burns, 1978; Forget et al., 2008). However, recent observations (e.g., Giuranna et al., 2011; Witasse et al., 2014) suggest that Phobos and Deimos may have formed from debris generated from re-accretion after one or more giant impacts of bodies with $\sim 0.01\times$ target mass, a process that could more readily explain the composition, density, and orbits of the martian moons (Rosenblatt, 2011). Compositionally, observations from both the European Space Agency's Mars Express mission and NASA's Mars Global Surveyor mission show that Phobos and Deimos have thermal infrared spectra in poor agreement with the primitive carbon-rich materials normally associated with carbonaceous chondrites (Giuranna et al., 2011; Witasse et al., 2014). Additionally, minerals were detected on Phobos that are also present on the martian surface (Giuranna et al., 2011; Witasse et al., 2014), suggesting that the material that formed the martian satellites could have been ejected from Mars. Furthermore, the Mars Radio Science Experiment on board Mars Express mission measured the density

of Phobos to be 1.87 g cm^{-3} with a porosity of 25–35% (Andert et al., 2010; Pätzold et al., 2014; Rosenblatt et al., 2010) – such low density and high porosity are uncommon for most asteroids. An asteroid with such a composition would not be expected to remain coherent if it was dynamically captured (Craddock, 1994), although it could be expected to form by material re-accreted from a debris disk (Richardson et al., 2002). Re-accretion could also explain the circularity of the orbits of Phobos and Deimos; if the martian moons were captured objects, they would be expected to have more eccentric orbits and randomly oriented orbital planes – and although the orbit of Phobos could be circularized via tidal perturbations from Mars, Deimos is too far away to be strongly affected by such perturbations. Additionally, moons that formed from the same impact that produced the current martian spin would be expected to orbit near the equatorial plane, as Phobos and Deimos do (Rosenblatt and Charnoz, 2012).

The observations discussed above makes a strong case that Phobos and Deimos formed via re-accretion, but this process requires the ejection of sufficient satellite-forming material into circum-Mars orbit, most likely via a giant impact (Craddock, 2011). Giant impacts were a common process in the late stages of planetary formation, and several large basins between 3300 and 4500 km in diameter suggest impacts that could have ejected significant debris into martian orbit. The largest possible impact site is the 7700 km-diameter Borealis basin that encompasses almost the entire northern hemisphere. Although whether the Borealis basin formed from

* Corresponding author.

E-mail address: ricitron@berkeley.edu (R.I. Citron).

exogenic or endogenic processes is still debated, recent analysis of the martian crust suggests that a giant impact could explain the elliptical shape of the basin and the sharp crustal thickness gradient on the basin perimeter (Andrews-Hanna et al., 2008). In addition, two independent studies numerically simulated Borealis-scale giant impacts ($E \sim 1\text{--}3 \times 10^{29}$ J) and found that such an impact could have formed a structure similar to the basin (Marinova et al., 2008; Nimmo et al., 2008).

A Borealis-scale impactor would likely have had a mass of ~ 0.026 Mars masses (Marinova et al., 2008), allowing the ejection of substantial material into martian orbit. The total mass of ejected material necessary to form Phobos and Deimos is unclear, and is complicated by the large number of elongate craters on the surface of Mars that could have formed via impacts by additional short-lived martian satellites (Schultz et al., 1982; Strom et al., 1992). The total mass of past and present martian satellites, M_{sats} , is estimated to be between 8.3×10^{17} and 1.5×10^{19} kg ($\sim 1 \times 10^{-6}\text{--}2 \times 10^{-5}$ Mars masses), based on the application of the Gault and π -group crater scaling laws (Gault, 1974; Gault and Wedekind, 1977; Melosh, 1989) to various estimates of the number of elongate martian craters attributed to the impact of de-orbited past satellites (Craddock, 2011; Schultz et al., 1982; Bottke et al., 2000). Although such an estimate is limited because the elongated crater population could also include grazing asteroid impacts and may not account for all prior martian satellites, the estimated mass of past and present satellites is still considerably greater than the current masses of Phobos and Deimos, which are 1.07×10^{16} and 1.48×10^{15} kg, respectively.

Recently, the feasibility of forming the martian moons from giant impacts has been studied both analytically (Craddock, 2011) and numerically (Rosenblatt and Charnoz, 2012). Craddock (2011) estimated that the mass of a circum-Mars debris disk should be twice the mass contained in the present and past martian satellites ($M_{\text{sats}}/M_d \sim 0.5$), based on the assumption that half of the disk material migrates inward while the other half migrates outward from Mars to form satellites. Comparing this estimate to the impactor mass of $\sim 0.02 M_{\text{Mars}}$ that would be necessary to cause the current martian spin-rate, Craddock estimated that $\sim 0.01\text{--}0.4\%$ of the impactor mass must be ejected into orbit to form the martian moons, a circum-Mars accretion disk of $\sim 1 \times 10^{18}\text{--}5 \times 10^{19}$ kg ($M_d/M_{\text{imp}} \sim 1 \times 10^{-4}\text{--}4 \times 10^{-3}$).

To determine if such an inserted mass could produce the total estimated mass of prior and current martian satellites, Rosenblatt and Charnoz (2012) modeled the evolution of a 10^{18} kg circum-Mars debris disk, a problem significantly different from prior models of satellite formation in a circum-Earth debris disk (e.g., Ida et al., 1997) because of the low mass and density of the proposed martian disk. Rosenblatt and Charnoz (2012) modeled circum-Mars disk evolution in both a strong tidal regime, where accretion occurs at the Roche limit close to the planet, and a weak tidal regime, where accretion occurs further away from the planet. In the strong tidal regime, Rosenblatt and Charnoz (2012) found that to account for a moonlet population of $8.3 \times 10^{17}\text{--}1.5 \times 10^{19}$ kg, the disk must have had a mass of $8.3 \times 10^{19}\text{--}1.5 \times 10^{21}$ kg, 30–80 times the mass predicted by Craddock (2011). If the relation, $M_d \sim 100M_{\text{sats}}$, obtained by Rosenblatt and Charnoz (2012) is correct, in order to be consistent with $M_{\text{imp}} \sim 0.026M_{\text{Mars}}$ and $M_{\text{sats}} \sim 1 \times 10^{-6}\text{--}2 \times 10^{-5}M_{\text{Mars}}$, M_d/M_{imp} must be $(M_{\text{sats}}/M_{\text{imp}}) \times (M_d/M_{\text{sats}}) \sim 0.004\text{--}0.08$. So far, the only other direct calculation of M_d/M_{imp} is by Canup and Salmon (2014), which was reported at the DPS meeting in 2014.

To directly determine an estimate for M_d/M_{imp} , we performed numerical Smoothed Particle Hydrodynamics (SPH) simulations of giant impacts into Mars. We take our canonical giant impact

from Marinova et al. (2008), a 45° impact with an impact energy $E_{\text{imp}} = 3 \times 10^{29}$ J and $M_{\text{imp}} = 0.026 M_{\text{Mars}}$, which has been shown to produce a crustal dichotomy similar to that currently observed on Mars. To test the robustness of the result, we vary impactor angle and impact energy, particularly to account for other proposed mechanisms of dichotomy formation, such as an impact megadome requiring an impact with $E_{\text{imp}} \sim 1\text{--}3 \times 10^{30}$ J (Reese et al., 2011). Our numerical study of a range of potential impacts allows for improved inputs into circum-Mars disk evolution models.

2. SPH simulations

2.1. SPH method

We model giant impact scenarios onto Mars using the SPH method (Monaghan, 1992). SPH codes have been used to model several types of giant impacts, including the putative Moon-forming impact (Benz et al., 1989, 1986, 1987; Cameron, 1997; Cameron and Benz, 1991; Canup, 2004, 2008; Canup and Asphaug, 2001) and other types of large-scale planetary collisions (Agnor and Asphaug, 2004; Genda et al., 2012). An SPH code has also been used to model the Borealis impact on Mars (Marinova et al., 2008), but these simulations focused on the structure of the impact basin and did not examine the formation of a debris disk.

We use the SPH code described in Genda et al. (2012), which is based on the convention of Monaghan (1992). Our simulations use the Tillotson equation of state (EOS) (Tillotson, 1962), which is widely used in large-scale planetary impact simulations (e.g., Benz et al., 1987; Canup et al., 2001; Marinova et al., 2008). The Tillotson EOS parameters used in our study are identical to those used in prior simulations of a giant impact on Mars (Marinova et al., 2008). To represent Mars, we use a differentiated iron/mantle target body with a core to mantle mass ratio of 0.3 and a total weight of 6.0×10^{23} kg. The impactor is an undifferentiated basalt body with a mass of $0.026 \times$ target mass. Our canonical impact has an impact velocity of 6 km/s, corresponding to an impact energy of 3×10^{29} J, similar to the nominal simulation from Marinova et al. (2008). We vary impact velocity ($\alpha = v_{\text{imp}}/v_{\text{esc}}$) and impactor mass, M_{imp} , to obtain disk masses for several types of giant impacts into Mars.

After each SPH simulation is complete, we quantify the mass and angular momentum of the impact-generated disk using the method described in Cuk and Stewart (2012); the disk contains SPH particles that have a negative total energy and an equivalent circular orbital radius that is greater than an estimated planetary radius defined by a 1 g cm^{-3} density contour. This method or a similar calculation is generally used to estimate disk masses from large collisions (e.g., Canup, 2008; Cuk and Stewart, 2012), however, it is only an effective estimate for the initial mass of the disk shortly after impact. For timescales greater than several tens of hours after impact, artificially induced spreading in the SPH simulation leads to an ineffective disk mass calculation. Furthermore, the SPH simulation is limited in that the collisionless particle representation cannot accurately simulate accretion in a disk, a computation better estimated using numerical models of disk evolution (e.g., Ida et al., 1997; Rosenblatt and Charnoz, 2012). Therefore, our initial disk mass estimates are meant to be compared with or used as inputs for long-term disk evolution models, and because the damping of inclination is rapid in comparison to the disk's radial evolution (Ida et al., 1997), disk mass should not change significantly beyond the time limit of our simulations due to damping.

For our canonical impact, we conducted a resolution test (Runs 0–3, Table 1) to compare disk mass evolution versus time for several total SPH particle counts (Fig. 1). For the first 40 h after the

Table 1
Disk masses for several SPH simulations with various impact parameters.^a

Run	N	M_{imp} (kg)	α	θ ($^\circ$)	E_{imp} (J)	γ	M_d/M_{imp}^b	M_d (kg) ^b	$L_d^{*b,c}$
0	5×10^4	1.68×10^{22}	1.4	45	3×10^{29}	0.028	0.026	4.3×10^{20}	0.95
1	1×10^5	1.68×10^{22}	1.4	45	3×10^{29}	0.028	0.029	4.8×10^{20}	0.98
2	3×10^5	1.68×10^{22}	1.4	45	3×10^{29}	0.028	0.033	5.5×10^{20}	0.92
3	1×10^6	1.68×10^{22}	1.4	45	3×10^{29}	0.028	0.033	5.5×10^{20}	0.90
4	3×10^5	1.68×10^{22}	1.4	30	3×10^{29}	0.028	0.024	4.0×10^{20}	0.89
5	3×10^5	1.68×10^{22}	1.4	60	3×10^{29}	0.028	0.017	2.8×10^{20}	1.01
6	3×10^5	1.68×10^{22}	2	45	6.3×10^{29}	0.028	0.024	4.0×10^{20}	0.95
7	3×10^5	3.5×10^{22}	1.4	45	6×10^{29}	0.058	0.033	1.14×10^{21}	0.86
8	3×10^5	6.0×10^{22}	1.4	45	1×10^{30}	0.100	0.023	1.4×10^{21}	0.89
9	3×10^5	1.0×10^{23}	1.9	45	3×10^{30}	0.167	0.013	1.3×10^{21}	0.76

^a Here, we list our ten simulations according to run number, number of particles in the simulation (N), impactor mass (M_{imp}), relative impact velocity ($\alpha = v_{\text{imp}}/v_{\text{esc}}$), impact angle (θ), energy of the impact (E_{imp}), impactor-to-target mass ratio ($\gamma = M_{\text{imp}}/M_{\text{target}}$), disk-to-impactor mass ratio (M_d/M_{imp}), disk mass (M_d), and normalized angular momentum (L_d^*).

^b M_d/M_{tot} , M_d , and L_d^* are all reported at $t = 30$ h.

^c We normalize the angular momentum as $L_d^* = L_d/(M_d \sqrt{GM_{\text{Mars}} R_{\text{Roche}}})$.

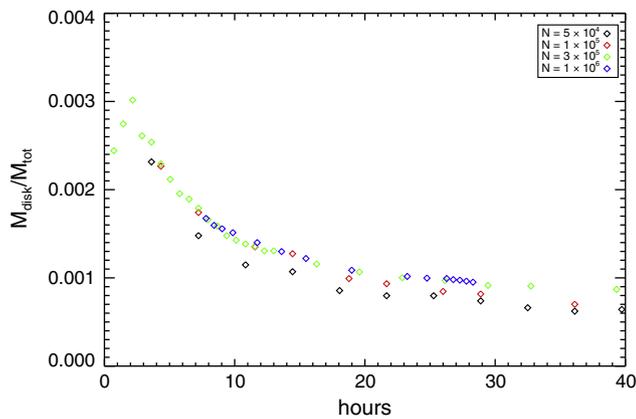


Fig. 1. Disk mass versus time for several total particle numbers

impact, the disk mass evolution of a simulation with 3×10^5 particles is similar to the simulation with 10^6 particles. Because of this, we use a resolution of $N = 3 \times 10^5$ particles for the remainder of our simulations.

2.2. SPH results

Because the proposed Borealis impact is lower energy than the putative Moon-forming collision, impacts studied in this investigation produced much smaller disks than previous studies of Earth–Moon formation. The disk masses for the collision scenarios we investigated are shown in Table 1. For our canonical impact (Run 2, Table 1), a disk of mass 5.55×10^{20} kg is produced 30 h post-impact, composed of ~ 280 particles.

An example of the evolution of a debris disk formed after our canonical Borealis-scale impact is shown in Fig. 2. In Fig. 2, the blue particles represent particles that are part of the planet or lack sufficient angular momentum to be considered part of the disk. The red particles are part of the disk, and black particles have been ejected from the system. The initial shock of the impact creates a large amount of ejecta that orbits the planet. While a degree of spreading of the ejecta occurs, a large amount of ejecta remains relatively close to the planet in a disk. An edge-on view of the disk is also shown in Fig. 3. The disk appears to undergo relatively quick damping, and Fig. 1 shows that the disk mass becomes more constant as simulation time increases.

To test the robustness of our result, we varied impact angle and impact energy, and compared the resulting debris disk masses. For impact angle, in addition to the nominal 45° impact angle, we also

tested impacts with angles of 30° and 60° . These changes in impact angle only slightly reduced the disk mass. For impact energy, we ran several simulations with higher impact energy by increasing the impactor mass and/or velocity (Runs 6–9, Table 1). While increasing the impact energy did result in larger mass debris disks, the highest energy impact we examined (Run 9) only resulted in approximately double the disk mass of the canonical impact (Run 2). In all cases, the amount of disk material produced was about 1–4% of the impactor mass. Therefore, we can conclude that $M_d/M_{\text{imp}} \sim (1-4) \times 10^{-2}$ is robust for Borealis-scale impacts. This appears to agree with the result of Canup and Salmon (2014), which found that an impactor with a few percent of Mars' mass produced a disk several orders of magnitude more massive than Phobos and Deimos.

We also examined the disk angular momentum, scaled as $L_d^* = L_d/(M_d \sqrt{GM_{\text{Mars}} R_{\text{Roche}}})$. If L_d^* is greater than 1, most of the mass is outside of the Roche limit, therefore this quantity represents the compactness of the disk compared to the Roche limit. Our simulations (Table 1) produce disks with $L_d^* < 1$ for all but one case. We find that L_d^* decreases when both the impactor mass and impact energy increase. The higher angle $\theta = 60^\circ$ impact (Run 5) produces the least compact disk. Our canonical simulation (Run 2) produces $L_d^* = 0.92$, representing a slightly compact disk where most of the mass is within the Roche limit, but moderate mass is also present outside of the Roche limit. The presence of mass outside the Roche limit is also reported in the Canup and Salmon (2014) abstract. Because a value of $L_d^* \sim 0.9$ should correspond to the strong tidal regime described by Rosenblatt and Charnoz (2012), our results indicate that accretion in the strong tidal regime is favored over accretion in the weak tidal regime.

3. Discussion and conclusion

Our examination of a Borealis-scale impact (Run 2, Table 1) shows that $\sim 5 \times 10^{20}$ kg of disk material is produced 30 h post-impact, well within the $8.3 \times 10^{19} - 1.5 \times 10^{21}$ kg disk mass Rosenblatt and Charnoz (2012) require in their strong regime to produce sufficient satellite mass to explain the elongated crater distribution on Mars. Modest adjustments to the impact angle or velocity produce similar disk masses, suggesting that sufficient material to produce Phobos and Deimos, or even the entire past/present satellite population, could have been ejected in a Borealis-scale impact. The larger impacts we examined, analogous to collisions required to form impact megadomes (e.g., Reese et al., 2011), produce over twice the disk mass as the smaller scale impacts. This is at the upper limit of the disk mass required to pro-

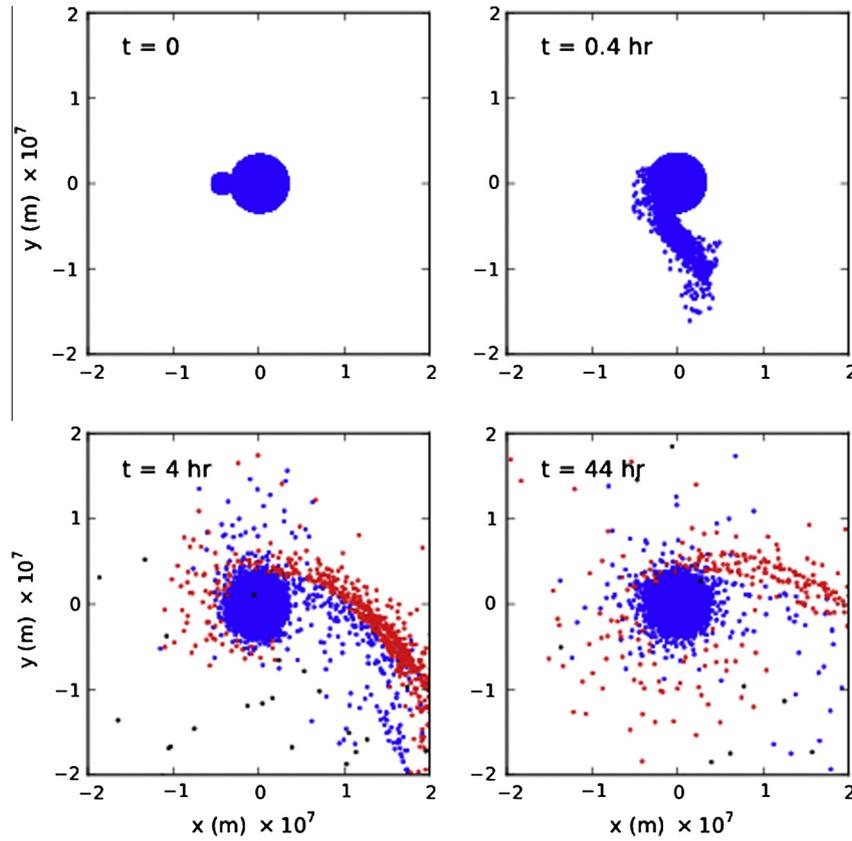


Fig. 2. Snapshots of an SPH simulation during the impact and disk formation. Ejected particles are shown in black and disk particles are shown in red. The blue particles are considered part of the planet. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

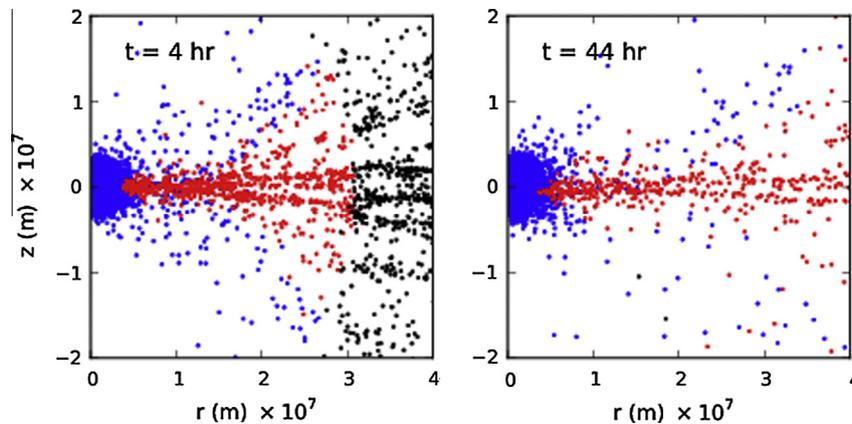


Fig. 3. An edge on view of the simulation from Fig. 1 at $t = 4$ and 44 h. Colors of particles are the same as in Fig. 1. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

duce the theoretical moonlet population and greater than the disk mass necessary to produce Deimos in the weak accretion regime.

Our finding that 1–4% of the impactor mass is inserted into an accretion disk is 20–100 times higher than the Craddock (2011) analytical estimate that 0.01–0.2% of the impactor is inserted into orbit. This is partly because Craddock (2011) estimates the ejected mass by assuming that half of a circum-Mars disk would migrate outward and accrete into satellites, and therefore the disk mass should be twice the mass of the past and present satellites. Additionally, the Gault and π -group scaling laws used to estimate the mass of past satellites are valid only for 90° impacts therefore should only provide a lower limit on the masses of the satellites

that could have formed elongate craters on Mars (Craddock, 2011). By numerically modeling the evolution of circum-Mars debris disks, Rosenblatt and Charnoz (2012) found that only 1% of the disk mass necessarily accreted into satellites. Therefore, although our simulations produce disks with higher mass than the Craddock (2011) estimates, if less of the disk mass accretes into moonlets and satellites, then similar satellite mass distributions can be achieved.

It should be noted that our study examines a narrow parameter space of Borealis-scale impacts that we find produce disks more massive than the 10^{18} kg disk examined by Rosenblatt and Charnoz (2012). Therefore, although we assume that 1% of the disk

mass forms satellites (Rosenblatt and Charnoz, 2012), it is not certain that this estimate holds at larger disk masses, and further simulations of circum-Mars disk evolution for larger disk masses are necessary. Additionally, accretion in more massive disks ($>10^{18}$ kg) could produce satellites larger in size than Phobos or Deimos (Rosenblatt and Charnoz, 2012); however, these satellites could eventually de-orbit and impact Mars, depending on where they form in the disk. Furthermore, in addition to Borealis-scale impacts, Craddock (2011) suggests that the Elysium and Utopia basins could have been formed by large enough impacts to produce the current martian spin rate, suggesting impacts as low as 0.015 Mars masses. A basic power law fit to our simulations with fixed $\alpha = 1.4$ and $\theta = 45$ (Runs 2, 7, and 8) yields $M_d = 8 \times 10^{21} (M_{\text{imp}})^{0.75}$, meaning an impactor with 0.015 target masses would produce a disk of mass $\sim 3 \times 10^{20}$ kg. Direct simulations of smaller impacts would be useful, but as the impact energy decreases, more total particles are needed in the simulation if the final disk is to contain at least the couple hundred particles necessary to accurately estimate the disk mass. Because our impact energies fall in the most-likely range reported by Craddock (2011), we leave the examination of less energetic impacts for future work.

Our simulations show that for Borealis-scale impacts, enough material is ejected into orbit to form accretion disks that could produce martian satellites. Our result is inconsistent with the formation of only Phobos and Deimos in the weak tidal regime considered by Rosenblatt and Charnoz (2012) because our disk is 100–1000 times more massive and the L_d^* values of our disks mostly indicate too much compactness for the weak tidal regime. Therefore, our results support accretion in the strong tidal regime. Although additional simulations of disk evolution with disks of initially larger mass are required to further investigate scenarios of martian moon formation, our results provide a numerical estimate of disk mass production in Borealis-scale impacts.

Acknowledgments

R. Citron would like to thank the NSF's East Asia and Pacific Summer Institute (Award 1107883) and the Japan Society for the Promotion of Science's Summer Science Program for their support of this research. We thank valuable and helpful comments by anonymous referees.

References

- Agnor, C., Asphaug, E., 2004. Accretion efficiency during planetary collisions. *Astrophys. J.* 613, L157–L160.
- Andert, T.P. et al., 2010. Precise mass determination and the nature of Phobos. *Geophys. Res. Lett.* 37, L09202.
- Andrews-Hanna, J.C., Zuber, M.T., Banerdt, W.B., 2008. The Borealis basin and the origin of the martian crustal dichotomy. *Nature* 453 (7199), 1212–1215.
- Benz, W., Slattery, W.L., Cameron, A.G.W., 1986. The origin of the Moon and the single-impact hypothesis. I. *Icarus* 66, 515–535.
- Benz, W., Slattery, W.L., Cameron, A.G.W., 1987. The origin of the Moon and the single-impact hypothesis. II. *Icarus* 71, 30–45.
- Benz, W., Cameron, A.G.W., Melosh, H.J., 1989. The origin of the Moon and the single impact hypothesis. III. *Icarus* 81, 113–131.
- Bottke, W.F. et al., 2000. Interpreting the elliptical crater populations on Mars, Venus, and the Moon. *Icarus* 145 (1), 108–121.
- Burns, J.A., 1978. On the orbital evolution and origin of the martian moons. *Vistas Astron.* 22, 193–210.
- Cameron, A.G.W., 1997. The origin of the Moon and the single impact hypothesis. V. *Icarus* 126 (1), 126–137.
- Cameron, A.G.W., Benz, W., 1991. The origin of the Moon and the single impact hypothesis. IV. *Icarus* 92, 204–216.
- Canup, R.M., 2004. Simulations of a late lunar-forming impact. *Icarus* 168 (2), 433–456.
- Canup, R.M., 2008. Lunar-forming collisions with pre-impact rotation. *Icarus* 196 (2), 518–538.
- Canup, R.M., Asphaug, E., 2001. Origin of the Moon in a giant impact near the end of the Earth's formation. *Nature* 412 (6), 708–712.
- Canup, R.M., Salmon, J., 2014. On an impact origin of Phobos-Deimos. In: *AAS/Division for Planetary Sciences Meeting Abstracts*, vol. 46 (abstract).
- Canup, R.M., Ward, W.R., Cameron, A.G.W., 2001. A scaling relationship for satellite-forming impacts. *Icarus* 150, 288–296.
- Craddock, R.A., 1994. The origin of Phobos and Deimos. *Lunar Planet. Sci.* 25, 293 (abstract).
- Craddock, R.A., 2011. Are Phobos and Deimos the result of a giant impact? *Icarus* 211, 1150–1161.
- Cuk, M., Stewart, S.T., 2012. Making the Moon from a fast-spinning Earth: A giant impact followed by resonant despinning. *Science* 338, 1047–1052.
- Forget, F., Costard, F., Lognonné, P., 2008. *Planet Mars: Story of Another World*. Praxis Publishing, Chichester.
- Gault, D.E., 1974. *Impact cratering. Primer Lunar Geol.*, 137–175.
- Gault, D.E., Wedekind, J.A., 1977. Experimental hypervelocity impact into quartz sand – II: Effects of gravitational acceleration. In: *Impact and Explosion Cratering: Planetary and Terrestrial Implications; Proceedings of the Symposium on Planetary Cratering Mechanics*, pp. 1231–1244.
- Genda, H., Kokubo, E., Ida, S., 2012. Merging criteria for giant impacts of protoplanets. *Astrophys. J.* 744, 137–144.
- Giuranna, M. et al., 2011. Compositional interpretation of PFS/MEX and TES/MGS thermal infrared spectra of Phobos. *Planet. Space Sci.* 59 (1), 1308–1325.
- Ida, S., Canup, R.M., Stewart, G.R., 1997. Lunar accretion from an impact-generated disk. *Nature* 389, 353–357.
- Marinova, M.M., Aharonson, O., Asphaug, E., 2008. Mega-impact formation of the Mars hemispheric dichotomy. *Nature* 453, 1216–1219.
- Melosh, H.J., 1989. *Impact Cratering: A Geologic Process*. Oxford University Press, New York, NY.
- Monaghan, J.J., 1992. Smoothed Particle Hydrodynamics. *Annu. Rev. Astron. Astrophys.*, vol. 30 (A93–25826 09–90) 30, 543–574.
- Murchie, S.L. et al., 1991. Color heterogeneity of the surface of PHOBOS – Relationships to geologic features and comparison to meteorite analogs. *J. Geophys. Res.* 96, 5925–5945, ISSN 0148-0227.
- Nimmo, F. et al., 2008. Implications of an impact origin for the martian hemispheric dichotomy. *Nature* 453, 1220–1223.
- Pätzold, M. et al., 2014. Phobos: Observed bulk properties. *Planet. Space Sci.* 102, 86–94.
- Reese, C.C., Orth, C.P., Solomatov, V.S., 2011. Impact megadomes and the origin of the martian crustal dichotomy. *Icarus* 213, 433–442.
- Richardson, D.C. et al., 2002. Gravitational aggregates: Evidence and evolution. *Asteroids III*, 501–515.
- Rosenblatt, P., 2011. The origin of the martian moons revisited. *Astron. Astrophys. Rev.* 19, 1–26.
- Rosenblatt, P., Charnoz, S., 2012. On the formation of the martian moons from a circum-martian accretion disk. *Icarus* 221 (2), 806–815.
- Rosenblatt, P., et al., 2010. Revisiting Phobos' origin issue from Mars Express Radio-Science observations. *EPSC Abstracts*, 1–2.
- Schultz, P.H., Schultz, R.A., Rogers, J., 1982. The structure and evolution of ancient impact basins on Mars. *Int. Colloq. Mars* 87, 9803–9820.
- Strom, R.G., Croft, S.K., Barlow, N.G., 1992. The martian impact cratering record. *Mars*, 383–423.
- Tillotson, 1962. *Metallic Equations of State for Hypervelocity Impact*. General Atomic Rept. GA-3216, 1–142.
- Witasse, O. et al., 2014. Mars express investigations of Phobos and Deimos. *Planet. Space Sci.* 102, 18–34.