Implantation of Martian Materials in the Inner Solar System by a Mega Impact on Mars

Ryuki Hyodo© and Hidenori Genda©
Earth-Life Science Institute/Tokyo Institute of Technology, 2-12-1 Tokyo, Japan; hyodo@elsi.jp
Received 2018 February 27; revised 2018 March 18; accepted 2018 March 19; published 2018 April 2

Abstract

Observations and meteorites indicate that the Martian materials are enigmatically distributed within the inner solar system. A mega impact on Mars creating a Martian hemispheric dichotomy and the Martian moons can potentially eject Martian materials. A recent work has shown that the mega-impact-induced debris is potentially captured as the Martian Trojans and implanted in the asteroid belt. However, the amount, distribution, and composition of the debris has not been studied. Here, using hydrodynamic simulations, we report that a large amount of debris (~1% of Mars’ mass), including Martian crust/mantle and the impactor’s materials (~20:80), are ejected by a dichotomy-forming impact, and distributed between ~0.5–3.0 au. Our result indicates that unmelted Martian mantle debris (~0.02% of Mars’ mass) can be the source of Martian Trojans, olivine-rich asteroids in the Hungarian region and the main asteroid belt, and some even hit the early Earth. The evidence of a mega impact on Mars would be recorded as a spike of 40Ar–39Ar ages in meteorites. A mega impact can naturally implant Martian mantle materials within the inner solar system.

Key words: meteorites, meteors, meteoroids – minor planets, asteroids: general – minor planets, asteroids: individual (A-type asteroids) – planets and satellites: composition – planets and satellites: formation

1. Introduction

Olivine is a major mineral of the Martian upper mantle with ~60 wt% (Bertka & Fei 1997; Zuber 2001). Also, at the surface of Martian grabens, such as Nili Fossae, an olivine-rich signature is detected (Hoefen et al. 2003; Mustard et al. 2009). Interestingly, seven out of the nine known Martian Trojans, called the Eureka family, show olivine-rich spectral features (Polishook et al. 2017). In addition, the rare A-type asteroids orbiting within the Hungarian region (~7% of the total mass) and within the main asteroid belt region (~0.4%) have olivine-rich spectral features (DeMeo & Crary 2013). The origin of A-type asteroids is still debated even though some could be mantle materials fragmented from differentiated parent objects (Sanchez et al. 2014).

More and more studies suggest that a mega impact occurs on Mars—a single giant impact could simultaneously explain the formation of the Borealis basin (Marinova et al. 2008), including its True Polar Wander (Hyodo et al. 2017b) spinning up Mars to the current orbital period of ~25 hr (Craddock 2011; Hyodo et al. 2017b), and the formation of Martian moons, Phobos and Deimos (Citron et al. 2015; Rosenblatt et al. 2016; Hesselbrock & Minton 2017; Hyodo et al. 2017a, 2017b). This giant impact could potentially distribute the impact-generated debris within the inner solar system, similarly to what happened after the giant impact on the Earth that may have formed the Moon (e.g., Bottke et al. 2015).

Recently, Polishook et al. (2017) suggested that the Martian Trojans may be captured from Martian mantle debris distributed near the orbits of Mars by a mega impact through an efficient capture during a sudden orbital change of Mars (Scholl et al. 2005). Such orbital change occurs due to the gravitational interactions with other planetesimals (Brasser et al. 2017). They also proposed that the rare A-type asteroids may have originated from the impact debris of the Martian mantle. However, it is not clear if a mega impact is indeed capable of producing a sufficient amount of ejecta with appropriate compositions and orbits that would explain the Martian Trojans and the rare A-type asteroids. Moreover, a giant impact is an energetic process that melts/vaporizes materials and thus the primordial Martian olivine-rich signature may be lost. Here we investigate the compositional and thermodynamic properties of the impact ejecta as well as their mass and heliocentric orbits, and discuss the possibility of forming Martian Trojans and rare A-type asteroids. In Section 2, we explain our giant impact simulations. In Section 3, we present our numerical results. Finally, our conclusions and discussions are presented in Section 4.

2. Numerical Methods and Models

We investigate the mass, composition, thermodynamic properties, and heliocentric orbits of the post-impact debris produced by the dichotomy-forming mega impact using the data obtained from smoothed particle hydrodynamics (SPH) simulations. We used the GADGET-2 code (Springel 2005) that includes tabulated equations of state (Cuk & Stewart 2012). The M-ANEOS equation of state (Melosh 2007) is used to model a differentiated object; its core is represented by pure iron and its mantle is represented by pure forsterite. This code includes self-gravity, but does not include material strength. For a huge impact (>100 km in radius), the effects of material strength are negligible (Jutzi et al. 2010; Genda et al. 2017). Note that M-ANEOS may underestimate the melting/vaporization fraction for a high pressure regime (>100 GPa where impact velocity of >8 km s−1; Kurosawa et al. 2012; Kraus et al. 2012), but our nominal case of 6 km s−1 impact (see below) would correctly predict the melting fraction.

Mars is modeled as a differentiated object with a core-to-mantle mass ratio of 0.3 and a total mass of 6.0 × 1023 kg and the impactor is modeled as an undifferentiated forsterite body with a mass of mimp = 1.68 × 1025 kg (Marinova et al. 2008). The total number of SPH particles in the simulation is N = 3 × 105 or 3 × 106. We placed the SPH particles in a three-dimensional lattice (face-centered cubic) for Mars and the impactor. Before the impact simulations, we carried out...
SPH simulations for gravitational relaxation for Mars and the impactor independently, so that the two bodies achieve their hydrostatic equilibrium. The initial entropy of Mars and the impactor is set to be 2000 J K\(^{-1}\) kg\(^{-1}\), which corresponds to 680 K at the surface of Mars. For our canonical dichotomy-forming impact (impact energy of \(E_{\text{imp}} = 3 \times 10^{39} \text{ J}\)), we apply an impact angle of 45° and an impact velocity of 1.4 times the mutual escape velocity (\(v_{\text{imp}} \sim 6 \text{ km s}^{-1}\)) (Marinova et al. 2008; Hyodo et al. 2017a, 2017b) with a total of \(3 \times 10^6\) SPH particles. The typical smoothing length in our SPH calculation is about \(50 \text{ km}\). We also carried out other impact simulations for the impact angle of 30° and 60° to see the effect of the impact angle. We also examined the dependence on mass ratio and impact velocity by considering the dichotomy-forming impact energy of \(E_{\text{imp}} = (3-6) \times 10^{39} \text{ J}\) (Marinova et al. 2008), where \(m_{\text{imp}} = 6.0 \times 10^{21} \text{ kg}\) and \(v_{\text{imp}} = 10 \text{ km s}^{-1}\) (\(E_{\text{imp}} = 3 \times 10^{39} \text{ J}\)) and \(m_{\text{imp}} = 3.36 \times 10^{22} \text{ kg}\) and \(v_{\text{imp}} = 6 \text{ km s}^{-1}\) (\(E_{\text{imp}} = 6 \times 10^{39} \text{ J}\)) with an impact angle of 45°. In these cases, we used \(3 \times 10^6\) SPH particles. Impact simulations are performed in an isolated system without the gravity of the Sun. We used snapshots obtained 20 hr after the impact. Note that we confirmed that the results do not significantly change as long as we use the snapshots obtained 5 hr after the impact.

In this paper, we are interested in the particles that are escaping from Mars’ gravity, and we categorize particles that are not gravitationally bound to Mars as ejected debris particles (Hyodo et al. 2017a, 2017b). In order to calculate the heliocentric orbits (semimajor axis, eccentricity, and inclination) of the ejected particles, we use positions and velocities of the ejected particles obtained from SPH simulations as input. If we know the direction of impact in the Sun-centered frame, we can uniquely define the orbits by considering the relative motions of the debris to Mars. However, the direction of impact that occurs in real life is very difficult to constrain. Giant impact is essentially isotropic (Kokubo & Genda 2010) and thus we decided to assume that the debris are isotopically distributed in the radial direction in the impact plane. Figure 2 also shows the original location of the debris derived from the Martian mantle completely avoids melting, but \(\sim 75\%\) of escaping debris derived from the Martian mantle completely avoids melting, but \(\sim 70\%\) of it undergoes complete melting (see Figure 3).

### 3. Debris from a Dichotomy-forming Impact

Figure 1 shows a snapshot of the mega impact simulations (\(N = 3 \times 10^7\), impact angle of 45° at 20 hr after the impact). A large amount of impact ejecta is produced (\(\sim 42,000\) ejected particles that are not gravitationally bound by Mars) and the ejecta contains not only impactor materials, but also Martian materials. Gravitational clumps are also formed and distributed randomly in the ejecta. We systematically sort particles and calculate their orbits at 20 hr after the impact. We also investigate the provenance of the ejected material—either from Mars or the impactor.

#### 3.1. Mass, Composition, and Thermodynamics of the Debris

Figure 2 shows the mass of the ejected debris and its composition in our nominal case. By changing the impact angles to a value between 30° and 60°, the debris mass differs between \(5 \times 10^{-3}\) and \(5 \times 10^{-2} M_{\text{Mars}}\) and the fraction of the Martian material differs between 5 and 40 wt%.

Statistically, an \(\sim 45°\) impact is the most likely (Shoemaker 1962) and is the sweet spot for the Mars hemispheric dichotomy formation (Marinova et al. 2008; Hyodo et al. 2017b). For the case of \(45°\)-impact high-resolution simulations (\(N = 3 \times 10^9\)), debris with a mass of \(\sim 10^{-3} M_{\text{Mars}}\) is produced and about \(\sim 20\%\) originates from Mars and the rest originates from the impactor. Figure 2 also shows the original location of the Martian material that is excavated by the impact. We found that about half of the Martian material originates from deep inside Mars, between 50 and 200 km in depth (the Martian mantle). Thus, our simulations indicate that about 10 wt% of the total debris is Martian mantle material. Note that the mass of pre-impact Mars is not significantly changed by this mega impact.

Here we focus on the thermal state of ejected debris. Figure 3 shows the temperature distribution of ejected escaping debris, which ranges between 1000 and 4000 K with a peak around 2000 K. Although this temperature distribution is similar to that of the debris disk orbiting Mars (Hyodo et al. 2017a), escaping debris has a somewhat wide temperature distribution because of the large variation of shock heating intensity during a mega impact. Since Mg\# (=Mg/(Mg+Fe) in mol) of bulk silicate Mars was estimated to be \(\sim 75\%\) (Elkins-Tanton et al. 2003), here we simply consider the (Mg\(_{0.75}\), Fe\(_{0.25}\)) SiO\(_2\) olivine solid solution as a major mineral of the Martian upper mantle. Its solidus and liquidus temperatures are about 1850 K and 2000 K (Boden & Schairer 1935), respectively. Therefore, \(\sim 10\%\) of escaping debris derived from the Martian mantle completely avoids melting, but \(\sim 70\%\) of it undergoes complete melting (see Figure 3). Taking into account partial melting, we can estimate that \(\sim 20\%\) of escaping debris derived from the Martian mantle avoids melting and preserves their primitive mineralogy. We also found that changing the impactor’s mass and impact velocity within the possible range (see Section 2) does not significantly change this unmelted fraction. Note also that in our simulations, we have arbitrarily set the initial surface temperature of Mars as 680 K since the epoch and surface condition of Mars are not well constrained when the mega impact occurs. However, we have confirmed that this unmelted fraction is \(> 10\%\) when the initial surface temperature is \(< 1100 \text{ K}\) and the fraction increases up to \(\sim 30\%\) as the surface temperature decreases to \(\sim 400 \text{ K}\). Further investigation is required to constrain the surface condition of ancient Mars. Thus, the unmelted Martian mantle material (olivine-rich material) is estimated to be about 2% of the total ejected mass (\(\sim 1.7 \times 10^{20}\) kg). This mass is much larger than those of A-type asteroids found in the Hungarian region (\(\sim 2.8 \times 10^{15}\) kg) and the main asteroid belt (\(\sim 8.9 \times 10^{18}\) kg; DeMeo & Carry 2013). Thus, our results imply that olivine-rich spectral features found among Martian Trojans (Polishook et al. 2017) and the rare A-type asteroids (DeMeo & Carry 2013) is attributed to this preserved olivine ejected from the Martian mantle during a mega impact.
3.2. Orbits and Size of the Debris

Figure 4 shows the heliocentric orbits of the debris at 20 hr after the impact, assuming the impact occurs at the current location of Mars. Here, we assumed that a mega impact happened on the Martian orbital plane. The debris is widely distributed within the inner solar system between about 0.5–3.0 au with an initial large eccentricity up to around $\sim 0.6$. The inclination from the impact plane is pumped up by the impact and is distributed up to about $\sim 0.3$ radian. The exact timing and radial location of Mars when the dichotomy-forming impact occurs is not well constrained. Mars may form at a greater heliocentric distance ($>2$ au) and is then scattered inward to the current location ($\sim 1.5$ au; Brasser et al. 2017). This can explain the fact that Mars has a distinct isotopic composition from the Earth (Tang & Dauphas 2014; Nugent et al. 2015). If the impact occurs at a larger semimajor axis before the inward scattering, the initial distribution of the debris also develops a larger semimajor axis, but the eccentricity and inclination distribution remain essentially the same.

The primordial size of the Martian Trojan Eureka is estimated to be a diameter of about $D \sim 2$ km (Nugent et al. 2015). The initial cumulative size frequency distribution (SFD) of large impact debris ($D > 100$ km) with the largest detected diameter of $D \sim 420$ km is directly derived from our impact simulations with sufficient resolution (Figure 5, see also Figure 1). We found that a steep slope of $q = 4.9$ for a power-law cumulative SFD fitted between 240 and 500 km (whose diameter is larger than $D$, $N_c(>D) = CD^{-q}$, where $C$ is constant). For smaller fragments ($D < 100$ km) where the simulation results are not directly applicable due to the resolution limit, we apply a single power-law cumulative SFD that satisfies the mass balance (total debris mass) for the rest of the debris, assuming a minimum size of $D = 0.1$ km to connect the simulation results of 240 km. By doing this, we derive a slope of $q \sim 2.1$ for this size range and predict the existence of the $\sim 10^6$ of $D \sim 2$ km or greater sized debris. Polishook et al. (2017) estimated the required number of debris of this size to explain Martian Trojans by the capture scenario as $\sim 10^5$ by assuming a largest fragment size of 114 km and a slope of $q = 2.85$ for the cumulative SFD. Our simulations...
shows that the total amount of ejected material by the dichotomy-forming impact is about 50 times larger than that estimated by Polishook et al. (2017). Then, 2% of the total ejected material is unmelted Martian mantle (Figure 3) and thus our results are consistent with the requirement derived from Polishook et al. (2017). Direct numerical simulations using our numerical results as input are required to understand the detailed process of capturing the debris. We will leave this matter to future work.

4. Discussion and Conclusions

In this Letter, we have shown that the dichotomy-forming impact produced a large amount of debris. The impact debris may be implanted within the asteroid belt. Figure 4 shows the post-impact distributions of eccentricity and inclination of the escaping debris, which are similar to those of the Moon-forming giant impact on Earth (Jackson & Wyatt 2012; Bottke et al. 2015). However, the heliocentric distance where a mega impact happened on Mars (~1.0 au or larger) should be larger than that for the Moon-forming giant impact (~1.0 au). Although some of the debris is expected to reaccrete onto Mars and the other terrestrial planets (Genda et al. 2017), a significant fraction of the debris can reach the asteroid belt region as a direct consequence of orbits resulting from a mega impact and/or through the successive planetary perturbations and resonances (Jackson & Wyatt 2012; Bottke et al. 2015). Our impact simulations demonstrated that the initial orbits of the debris can easily reach the asteroid belt region and thus unmelted Martian mantle material (a maximum of ~2% of the total debris mass) is potentially expected to settle into stable orbits as rare A-type asteroids found in the Hungarian and main asteroid belt regions. Investigation into the long-term evolution of the debris would be required to estimate the exact capture efficiency, which strongly depends on the initial orbits of the debris.

We also expect that the mega-impact-induced debris hit the pre-existing asteroids, which has been discussed in the context of the Moon-forming giant impact (Jackson & Wyatt 2012; Bottke et al. 2015). Although the amount of the debris produced by the dichotomy-forming impact is one order smaller than that produced by the canonical Moon-forming giant impact (a few percent of the Earth’s mass; Canup 2004), the location of a mega impact on Mars should be closer to the asteroid belt region. Therefore, we expect that the mega impact event that happened on Mars is also recorded in the asteroid belt. The impact velocity of the debris to pre-existing asteroids in the Hungarian region and in the main belt is expected to be
larger than 5 km s\(^{-1}\) (Figure 4), while the nominal collision velocity between asteroids is \(\sim 5\) km s\(^{-1}\) (Bottke et al. 1994). Such a high energetic impact can be recorded as a reset of \(^{40}\)Ar–\(^{39}\)Ar age and/or impact melts (Kurosawa & Genda 2018), which is also different from the case of the Moon-forming giant impact (Bottke et al. 2015). Since the timing of a mega impact on Mars is likely different from that of the Moon-forming giant impact (\(~100\) Myr after CAI condensation, Touboul et al. 2007), one of the multiple signatures for \(^{40}\)Ar–\(^{39}\)Ar resetting age in chondrites (Bottke et al. 2015) would be attributed to a mega impact on Mars.

Our simulations demonstrated that Martian materials are widely distributed within the inner solar system as a natural consequence of a mega impact and are eventually implanted as Martian Trojans and asteroids in the Hungarian and the main belt. Also, the impact ejecta can strike the pre-existing asteroid belt, and thus impact signatures that show a different epoch from those of the Moon-forming impact should be recorded on some meteorites. In addition, some of the debris should be transferred to early Earth. Since all ejected materials experienced a temperature above 2000 K in our simulations (Figure 3), this material transport does not directly support the panspermia hypothesis, in which life was born elsewhere in the universe (e.g., on Mars) and was transported to Earth. However, specific types of minerals or elements on Mars, such as borate or boron and molybdenum (Stephenson et al. 2013)—which are thought to play important roles in the origin of life (Ricardo et al. 2004), but would be lacking on early Earth—can be delivered to early Earth through this process. Future planetary explorations and the detailed data analysis of the meteorites will test our predictions.

This work was supported by JSPS Grants-in-Aid for JSPS Fellows (JP17J01269), MEXT KAKENHI grant (JP17H06457), and the Astrobiology Center Program of National Institutes of Natural Sciences (NINS) (AB291011).
ORCID iDs
Ryuki Hyodo https://orcid.org/0000-0003-4590-0988
Hidenori Genda https://orcid.org/0000-0001-6702-0872

References
Bertka, C. M., & Fei, Y. 1997, JGR, 102, 5251
Bottke, W. F., Nolan, M. C., Greenberg, R., & Krelowski, R. A. 1994, Icar, 107, 255
Bottke, W. F., Vokrouhlický, D., Marchi, S., et al. 2015, Sci, 348, 321
Bowen, N. L., & Schaefer, J. F. 1935, AmJS, 29, 151
Brasser, R., Mojzsis, S. J., Matsumura, S., & Ida, S. 2017, E&PSL, 468, 85
Canup, R. M. 2004, Icar, 168, 433
Citron, R. I., Genda, H., & Ida, S. 2015, Icar, 252, 334
Craddock, R. A. 2011, Icar, 211, 1150
Cuk, M., & Stewart, S. T. 2012, Sci, 338, 1047
DeMeo, F. E., & Carry, B. 2013, Icar, 226, 723
Elkins-Tanton, L. T., Parmentier, E. M., & Hess, P. C. 2003, M&PS, 38, 1753
Genda, H., Fujita, T., Kobayashi, H., et al. 2017, Icar, 294, 234
Genda, H., Iizuka, T., Kobayashi, H., et al. 2017, Icar, 294, 234
Genda, H., Iizuka, T., Kobayashi, H., et al. 2017, Icar, 294, 234
Hesselschroek, A., & Minton, D. A. 2017, NatGe, 10, 266
Hoefen, T. M., Clark, R. N., Bandfield, J. L., et al. 2003, Sci, 302, 627
Hyodo, R., Genda, H., Charnoz, S., & Rosenblatt, P. 2017a, ApJ, 845, 125
Hyodo, R., Rosenblatt, P., Genda, H., & Charnoz, S. 2017b, ApJ, 851, 122
Jackson, A. P., & Wyatt, M. C. 2012, MNRAS, 425, 657
Jutzi, M., Michel, P., Benz, W., & Richardson, D. C. 2010, Icar, 207, 54
Kokubu, E., & Genda, J. 2010, ApR, 714, L21
Kraus, R. G., Stewart, S. T., Swift, D. C., et al. 2012, JGRE, 117, E0009
Kurosawa, K., Kadono, T., Sugita, S., et al. 2012, JGRE, 117, E0007
Kurosawa, K., & Genda, H. 2018, GeoRL, 45, 620
Marinova, M. M., Aharonson, O., & Asphaug, E. 2008, Natur, 453, 1216
Melosh, H. 2007, M&PS, 42, 2079
Mustard, J. F., Ehlmann, B. L., Murchie, S. L., et al. 2009, Geophys. Res., 114, E00D12
Nugent, C. R., Mainzer, A., Masiero, J., et al. 2015, ApJ, 814, 117
Polishook, D., Jacobson, S. A., Morbidelli, A., & Aharonson, O. 2017, NatAs, 1, 0179
Ricardo, A., Carrigan, M. A., Olcott, A. N., & Benner, S. A. 2004, Sci, 303, 196
Rosenblatt, P., Charnoz, S., Dunseath, K. M., et al. 2016, NatGe, 9, 581
Sanchez, J. A., Reddy, V., Kelley, M. S., et al. 2014, Icar, 228, 288
Sanloup, C., Jambon, A., Gillet, P., et al. 1999, PEPL, 112, 43
Scholl, H., Marzari, F., & Tricarico, P. 2005, Icar, 175, 397
Shoemaker, E. M. 1962, in Physics and Astronomy of the Moon, ed. Z. Kopal (New York: Academic)
Springel, V 2005, MNRAS, 364, 1105
Stephenson, J. D., Hallis, L. J., Nagashima, K., & Freeland, S. J. 2013, Boron Enrichment in Martian Clay, ed. S. Maas, e64624 PLOS One 8
Tang, H., & Dauphas, N. 2014, E&PSL, 390, 264
Touboul, M., Kleine, T., Bourdon, B., Palme, H., & Wieler, R. 2007, Natur, 450, 1206
Zuber, M. T. 2001, Natur, 412, 220