Early bombardment of the moon: Connecting the lunar crater record to the terrestrial planet formation

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The lunar crater record features ~ 50 basins. The radiometric dating of Apollo samples indicates that the Imbrium basin formed relatively late — from the planet formation perspective — some \( \geq 3.9 \) Ga. Here we develop a dynamical model for impactors in the inner solar system to provide context for the interpretation of the lunar crater record. The contribution of cometary impactors is found to be insignificant. Asteroids produced most large impacts on the terrestrial worlds in the last \( \approx 3 \) Gyr. The great majority of early impactors were rocky planetesimals left behind at \( \sim 0.5-1.5 \) au after the terrestrial planet accretion. The population of terrestrial planetesimals was reduced by disruptive collisions in the first \( \tau \sim 20 \) Myr after the gas disk dispersal. We estimate that there were \( \sim 4 \times 10^5 \) diameter \( d > 10 \) km bodies when the Moon formed (total planetesimal mass \( \sim 0.015 \) M\(_{\text{Earth}}\) at \( \tau \sim 50 \) Myr). The early bombardment of the Moon was intense. To accommodate ~ 50 known basins, the lunar basins that formed before \( \geq 4.35-4.41 \) Ga must have been erased. The late formation of Imbrium occurs with a \( \sim 15\%-35\% \) probability in our model. About 20 \( d > 10 \) km bodies were expected to hit the Earth between 2.5 and 3.5 Ga, which is comparable to the number of known spherule beds in the late Archean. We discuss implications of our model for our lunar/Martian crater chronologies, Late Veneer, and noble gases in the Earth atmosphere.

1. Introduction

In the standard model of terrestrial planet formation (Wetherill, 1990), accretional collisions between 1 to 1000 km planetesimals lead to gradual build up of lunar- to Mars-size protoplanets that gravitationally interact and further grow during a late stage of giant impacts (Chambers and Wetherill, 1998). The hafnium–tungsten (Hf–W) isotopic system analysis indicates that the Moon-forming impact on proto-Earth happened relatively late, some \( \sim 30-150 \) Myr after the appearance of the first solar system solids (Kleine and Walker, 2017; Thiemens et al., 2019, 2021; Barboni et al., 2017; Maurice et al., 2020; Kruijer et al., 2021). The new-born Moon was molten, gradually cooled down, and was eventually able to support impact structures on its surface (Meyer et al., 2010; Elkins-Tanton et al., 2011; Miljković et al., 2021). This is time zero for the lunar crater record.

The early lunar bombardment was intense (Hartmann, 1966): at least 40 and up to 90 lunar basins (crater diameter \( D > 300 \) km) have been recognized or proposed (e.g., Wilhelms et al., 1987; Spudis, 1993; Fassett et al., 2012). Neumann et al. (2015) used gravity anomalies observed by GRAIL to report a complete list of all \( D > 200 \) km craters on the whole lunar surface (both the nearside and farside). In Table 1 of Neumann et al. (2015), there are 43 basins with the main ring diameter \( D > 300 \) km. From crater scaling laws appropriate for the lunar gravity (e.g., Johnson et al., 2016a), a diameter \( d \sim 20 \) km impactor is needed to excavate a \( D = 300 \) km crater.\(^1\) Miljković et al. (2016), modeling the GRAIL data from Neumann et al. (2015) with the iSALE-2D hydrocode, found that the lunar surface recorded \( \approx 50 \) impacts of \( d > 20 \) km bodies (the impact speed \( \sim 20 \) km s\(^{-1}\) is used here to convert the \( C \) parameter reported in their Table 2 to the impactor size). Here we use the results from Neumann et al. (2015) and Miljković et al. (2016) as an important constraint on our modeling efforts.

At least some of the lunar basins formed relatively late. The radiometric dating of impact melts found in the Apollo samples indicates that
the Imbrium basin formed \( \approx 3.92 \) Gyr ago (Ga; Zhang et al., 2019), that is \( \approx 650 \) Myr after the first solar system solids (Bouvier et al., 2007; Burkhardt et al., 2008). The Imbrium basin was excavated by a \( d \geq 100 \) km impactor (Miljković et al., 2013, 2016; Schultz and Crawford, 2016). Of all the known basins, only Orientale and Schrödinger have a lower accumulated density of superposed craters than Imbrium, and must therefore be (slightly) younger than Imbrium. The Orientale basin was produced by an \( d = 50 \) km (Miljković et al., 2016) or \( d = 64 \) km (Johnson et al., 2016b) impactor. A smaller, \( d \approx 20 \) km body impact is responsible for the Schrödinger basin \( (D = 312 \) km; Miljković et al., 2016).

Having (at least) three basin-forming impacts happening at \( t \geq 650 \) Myr after the first solar system solids is unexpected from the planet formation perspective. In the inner solar system, where the accretion processes have relatively short timescales \(< 100 \) Myr; Wetherill, 1990), the impact flux should have rapidly declined over time (Bottke et al., 2007). This has motivated the impact spike hypothesis where it was assumed that the Imbrium-era impacts mark an epoch of enhanced bombardment (Tera et al., 1974; Ryder, 1990, 2002; Cohen et al., 2000; Stöffler and Ryder, 2001; Kring and Cohen, 2002), and prompted search for possible causes (e.g., Lefèvre et al., 2001; Gomes et al., 2005). This epoch is often called the Late Heavy Bombardment (LHB). We avoid using this term here because LHB has too many strings attached to it (Hartmann, 2019).

The goal of this paper is to develop an accurate dynamical model of asteroid, comet and planetesimal impactors in the inner solar system. The model for asteroid and comet impactors is taken from previous publications (Nesvorný et al., 2017a,b). We take advantage of the recent simulations of terrestrial planet formation (Section 2) that faithfully match the masses, orbits and accretion histories of planets (Nesvorný et al., 2021a). Planetesimals remaining in these simulations after the Moon-forming impact (“leftovers”) are cloned and followed for additional 1 Gyr. All impacts on the terrestrial worlds are recorded in this interval. When combined with asteroids and comets, the new impact chronology model can be compared to the lunar crater record. We find that a steadily declining impact flux — in the spirit of Hartung (1974) — provides an adequate interpretation of available data.

Initial results from this study were reported in Nesvorný et al. (2022). Here we discuss the results in detail and, in addition: (1) provide a thorough description of the methodology, simulations and impact profiles obtained under different assumptions (e.g., as a function of the number and radial extension of terrestrial planetesimals), (2) explain how and why our results differ from those obtained in previous studies (e.g., Bottke et al., 2007; Brasser et al., 2020), (3) obtain the lunar and Martian impact chronologies, discuss their relationship and implications for ages of different terrains, (4) include constraints from highly siderophile elements for the Earth, Mars, and Moon, and (5) highlight comet impactors as the source of noble gases in the Earth atmosphere (Marty et al., 2016).

2. Terrestrial planet formation

We take advantage of our N-body simulations of the standard model (Nesvorný et al., 2021a). The simulations included, for consistency, the effects of radial migration and dynamical instability of the outer planets (Tsiganis et al., 2005). The instability was assumed to have happened early, within \( t \approx 10 \) Myr after the protoplanetary disk dispersal (Clement et al., 2018). In the specific case considered here, the outer planets started in a chain of mean motion resonances (3:2 and 2:1) and the instability happened at \( t = 5.8 \) Myr after the gas disk dispersal. We found that the details of outer planet instability do not matter with small Mars’ (potentially) forming even in the case where the outer planets are placed on their present orbits at the beginning of simulations.

\( ^2 \text{Mars represents only } \approx 11\% \text{ of Earth’s mass. Many terrestrial planet formation models suggest that a more massive planet should form at } 1.2-1.8 \text{ au. This is called the small Mars problem.} \)

The model with a radically extended disk of terrestrial planetesimals, however, failed to match the tight orbital spacing of Venus and Earth. To obtain the correct spacing, the terrestrial planetesimals must have started in a narrow annulus \( (0.7-1 \text{ au; Hansen, 2009}) \) and have (about) the Mars mass to begin with (Jacobson and Morbidelli, 2014). The tight radial spacing of Venus and Earth would be particularly easy to understand if terrestrial planetesimals convergently migrated toward \( \sim 0.7-1 \) au during the gas disk stage (Brož et al., 2021).

In the W11e/20M model from Nesvorný et al. (2021a), which we will utilize here, 20 Mars-mass protoplanets were initially \( (t = 0); \) marking the gas nebula dispersed) distributed in a narrow, dynamically cold annulus (orbital radius \( 0.7 < r < 1 \) au). In addition, 1000 planetesimals with the total mass of \( 2 \) \( M_{\text{Earth}} \) were placed in an extended disk \((0.3 < r < 4 \) au). The radial surface density profile of planetesimals was set to \( \Sigma(r) \propto r^{-1} \). We performed 100 simulations of the W11e/20M model where different seeds were used to generate slightly different initial conditions. This produced 100 growth histories of the terrestrial planets. The results were statistically analyzed using different criteria, including the number and mass of planets, their radial mass concentration, the angular momentum deficit (AMD) — a measure of their orbital excitation, etc.4

Here we adopt a reference case from the W11e/20M model (job #35) that produced a particularly good match to the terrestrial planets (Figs. 1 and 2). Venus, Earth and Mars formed in this simulation with nearly correct masses and semimajor axis: model mass \( 1.017 \) \( M_{\text{Earth}} \) and \( a = 0.727 \) au for Venus \((0.815 \) \( M_{\text{Earth}} \) and \( a = 0.723 \) au), \( 1.083 \) \( M_{\text{Earth}} \) and \( a = 1.034 \) au for Earth \((1 \) \( M_{\text{Earth}} \) and \( a = 1 \) au), \( 0.116 \) \( M_{\text{Earth}} \) and \( a = 1.525 \) au for Mars \((0.107 \) \( M_{\text{Earth}} \) and \( a = 1.524 \) au), where the numbers in parentheses list the real values for reference. The Mercury analog ended up to be too massive, which is related to the relatively large initial mass of protoplanets in W11e/20M and the assumption of inelastic mergers (Nesvorný et al., 2021a). The model orbits of Venus and Earth were only slightly less excited than in reality. The model orbit of Mars had a nearly perfect inclination \((\text{mean } i = 4.5^\circ) \) and somewhat lower eccentricity \((\text{mean } e = 0.028) \) vs. real mean \( i = 4.4^\circ \) and somewhat lower eccentricity \((\text{mean } e = 0.069) \). The orbital structure of the asteroid belt was reproduced.

An interesting feature of the case highlighted here is related to the Earth’s accretion history and the Moon-forming impact (Fig. 3). The initial growth of proto-Earth was fast. After accreting several Mars-class protoplanets, the proto-Earth reached mass \( \geq 0.52 \) \( M_{\text{Earth}} \) by \( t = 3 \) Myr after the gas disk removal. A prolonged stage of planetesimal accretion followed during which the proto-Earth modestly grew to \( \approx 0.57 \) \( M_{\text{Earth}} \) by \( t = 40 \) Myr. Then, at \( t = 41.3 \) Myr, an accretional collision between two roughly equal-mass protoplanets occurred, and the Earth mass shot up to \( \approx 1.05 \) \( M_{\text{Earth}} \). For comparison, Venus grew to 85% of its final mass in only \( \approx 3 \) Myr after the start of the simulation; the remaining 15% of mass was supplied to Venus by planetesimals over an extended timescale \((\approx 100 \) Myr). This supports the model of Jacobson et al. (2017), who argued that Venus has not developed a persistent magnetic field because it did not experience any late energetic impacts that would mechanically stir the core and create a long-lasting dynamo.

The accretional growth of the Earth shown in Fig. 3 would satisfy constraints from the Hf-W and U-Pb isotope systematics, assuming that the degree of metal-silicate equilibration was intermediate between the full and 40% equilibration considered in Kleine and Walker (2017); their Fig. 8). For example, Thiemens et al. (2019) found that the \( ^{182} \text{W} \) excess in lunar samples can be fully explained by the decay of now-extinct \( ^{182} \text{Hf} \), and suggested that the Moon formed \( \approx 40-60 \) Myr after

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3 Meaning that the initial planetary embryos must have been about the mass of Mars.

4 We found that in about 20% of cases, good Mars analogs formed (defined as planets with mass \( 0.5 < M < 2 \) \( M_{\text{Earth}} \) and \( 1.2 < a < 1.8 \) au) and the Earth and Venus ended with approximately correct masses, orbital excitation and separation (see Nesvorný et al., 2021a, for how this was quantified).
the first solar system solids.\(^5\) The low speed collision (the impact-to-escape speed ratio \(v_{\text{imp}}/v_{\text{esc}} = 1.01\)) between two nearly equal-mass protoplanets (the impactor-to-total mass ratio \(\Gamma = 0.46\)) falls into the preferred regime of Moon-forming impacts investigated in Canup (2012). In particular, the collision of two protoplanets with \(\Gamma > 0.4\) would lead to the Moon formation from a well mixed disk and would therefore imply matching oxygen-isotope compositions of the Earth and Moon (Williams and Cieza, 2011), and the simulation time \(\sim 3\) Myr (Williams and Cieza, 2011). The simulation time \(\sim 3\) Myr roughly corresponds to the formation of \(\sim 2\) km-scale lunar crater population (see (Bierhaus et al., 2018) for a review).

3. Chronology of lunar impacts

The radiometric ages, crater counts, and size distribution extrapolations are the basis of empirical models for impact cratering in the inner solar system (Neukum and Ivanov, 1994; Neukum et al., 2001; Hartmann and Neukum, 2001; Marchi et al., 2009; Hiesinger et al., 2012, 2020; Robbins, 2014; Marchi, 2021). On young lunar surfaces \((T \lesssim 100\) Ma; e.g., Cone, North/South Ray, Tycho; Hiesinger et al., 2016), where there is a good statistic only for very small, \(D \ll 1\) km craters, the crater counts can be extrapolated up with some assumed size distribution. The size distribution can be patched together from crater counts on different terrains, which implicitly assumes that the size distribution does not change with time, or related — via the crater scaling laws — to observations of near-Earth asteroids (NEAs; Marchi, 2012, 2020; Robbins, 2014; Marchi, 2021). The radiometric ages have been measured for Apollo, Luna and Chang'e samples, and for lunar meteorites (Stöffler and Ryder, 2001; Li et al., 2021). The main difficulty in interpreting these ages — in terms of the cratering and geological processes occurring on the Moon — is how to relate different samples to different events. For example, we know that major, basin-forming impacts were happening before 4 Ga (e.g., at \(T \approx 4.21\) and \(4.34\) Ga; (Pidgeon et al., 2010; Grange et al., 2013; Merle et al., 2013; Norman et al., 2015), but we do not know which specific basins formed at that time. We therefore cannot empirically determine \(N_1(T)\) for \(T > 4\) Ga, at least not with the existing data.

The Imbrium-basin formation has a strong presence in many collected samples. According to the most accurate U-Pb geochronological measurements, the Imbrium basin formed at \(T \approx 3.92\) Ga (Zhang et al., 2019). The radiometric ages of lunar maria range between \(3.1-3.7\) Ga (Stöffler and Ryder, 2001). The crater counts for Fra Mauro/Imbrium highlands establish that \(N_1(T)\) was about a factor of \(\sim 10\) higher at \(T \approx 3.92\) Ga than during the formation of most lunar maria \((T \approx 3.1-3.4\) Ga; Robbins, 2014). There clearly was a substantial increase in the lunar cratering rate at \(T \approx 3.5\) Ga.

The basic assumption of many empirical models for impact cratering is that the accumulated number of craters larger than \(D\) since time \(T\), \(N(D,T)\), is a separable function of \(D\) and \(T\); i.e., \(N(D,T) = N(D)N_1(T)\), where \(N(D)\) is the cumulative size distribution of craters (per surface...
around the Apollo and Luna landing sites, and performed new crater
and/or the size range of craters for which the counts were performed.
Arise from different choices of the terrains where craters were counted
\[ a = 2.2 \text{ au}, \quad q = a(1 - e) = 1.8 \text{ au}, \quad \text{and} \quad Q = a(1 + e) = 4 \text{ au} \]
in panel A, and \( a = 3.5 \text{ au} \) and \( i = 25^\circ \) in panel B. The solid lines in B are the \( v_6 \) and \( v_{16} \) secular resonances (plotted here for \( e \approx 0 \)). The mean orbits of the real terrestrial planets are indicated by planetary symbols. The simulation failed to reproduce the excited orbit of Mercury. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

Fig. 2. The mean orbital elements of planets (black dots) and asteroids (red dots) produced in the best simulation of the W11e/20M model (Nesvorný et al., 2021a). The lines approximately carve out the asteroid belt region. The dashed lines correspond to
\[ a = 2.2 \text{ au}, \quad q = a(1 - e) = 1.8 \text{ au}, \quad \text{and} \quad Q = a(1 + e) = 4 \text{ au} \]
in panel A, and \( a = 3.5 \text{ au} \) and \( i = 25^\circ \) in panel B. The solid lines in B are the \( v_6 \) and \( v_{16} \) secular resonances (plotted here for \( e \approx 0 \)). The mean orbits of the real terrestrial planets are indicated by planetary symbols. The simulation failed to reproduce the excited orbit of Mercury. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

area), the so-called production function, and \( N_1(T) \) is the chronology function. Neukum and Ivanov (1994) approximated the production function by a polynomial (Eq. (2) and Table 1 in Neukum et al., 2001). Based on extrapolations with this polynomial, the \( N_1 \) chronology function was given as
\[ N_1(T) = a \left( \exp(bT) - 1 \right) + cT \]
with \( a = 5.44 \times 10^{-15} \text{ km}^2 \), \( b = 6.93 \text{ Gyr}^{-1} \), and \( c = 8.38 \times 10^{-4} \text{ Gyr}^{-1} \text{ km}^{-2} \). Other published chronology functions used Eq. (1) with different coefficients. Marchi et al. (2009) constructed a production function based on the size distribution of NEAs and used it to determine \( a = 1.23 \times 10^{-15} \text{ km}^{-2} \), \( b = 7.85 \text{ Gyr}^{-1} \), and \( c = 1.30 \times 10^{-3} \text{ Gyr}^{-1} \text{ km}^{-2} \). Hartmann et al. (2007) added a quadratic term to Eq. (1) to express a slow (linear) decline of the impact rate over the last \( \sim 3 \text{ Gyr} \).

At least some of the differences between different works must arise from different choices of the terrains where craters were counted and/or the size range of craters for which the counts were performed. Robbins (2014), for example, carefully cared their terrains of interest around the Apollo and Luna landing sites, and performed new crater counts for \( D \geq 1 \text{ km} \) craters. This has a notable advantage over the previous results because \( N_1 \) is directly established from the counts themselves (no size extrapolation needed). For the oldest sites, however, the crater densities for \( D \sim 1 \text{ km} \) may be affected by crater saturation (Robbins, 2014).

Fassett et al. (2012) used topography from the Lunar Orbiter Laser Altimeter (LOLA) on the Lunar Reconnaissance Orbiter (LRO) to measure the superposed impact crater distributions for 30 lunar basins. They reported their counts in terms of \( N_{20}(T) \) — the same as \( N_1(T) \) but for \( D > 20 \text{ km} \) craters. For the Imbrium basin, Fassett et al. (2012) measured \( N_{20} = 30 \pm 5 \) (per \( 10^6 \text{ km}^2 \)) (later revised to \( N_{20} = 26 \pm 5 \) in Orgel et al., 2018). This is an important anchor of \( N_{20}(T) \) for a \( T \approx 3.92 \)-Gyr old lunar terrain (Zhang et al., 2019).

4. Modeling the impact flux in the inner solar system

There are at least three major source populations of impactors in the inner solar system: (i) leftover planetesimals in the terrestrial planet zone (0.3–1.75 au), (ii) main-belt asteroids (1.75–4 au; the range given here includes the Extended belt or E-belt for short, (Bottke et al., 2012)), and (iii) comets. Here we first discuss the results for (ii) and (iii) that were taken from previous work (Sections 4.1 and 4.2). The model for leftover planetesimals is described in Section 4.3.

4.1. Asteroids

In Nesvorný et al. (2017a), we published a dynamical model for asteroid impactors. The model used the same setup for the radial migration and instability of the outer planets as the terrestrial planet formation model described in Section 2 (the instability happens at \( t = 5.8 \text{ Myr} \) after the gas disk dispersal). To start with, the terrestrial planets were placed on the low-eccentricity and low-inclination orbits. The surface density profile of asteroids was assumed to follow \( \Sigma(r) \sim r^{-3} \).

The radial profile was smoothly extended from \( r > 2 \text{ au} \), where the model can be calibrated on observations of main belt asteroids (see below), to \( r = 1.75–2 \text{ au} \). This fixed the initial number of bodies in the now largely extinct E-belt (Bottke et al., 2012). The results had large statistics (50,000 model asteroids) and full temporal coverage (integration time \( t = 0–4.57 \text{ Gyr} \)).

The flux of asteroid impactors was calibrated from today’s asteroid belt. We showed that the model distribution at the simulated time \( t = 4.57 \text{ Gyr} \) (i.e., at the present epoch) was a good match to the orbital distribution of asteroids. The number of model \( d > 10 \text{ km} \) asteroids at \( t = 4.57 \text{ Gyr} \) was set to be equal to the number of \( d > 10 \text{ km} \) main-belt asteroids (\( \approx 8200 \)), as measured by Wide-field Infrared Survey Explorer (WISE; Mainzer et al., 2019). When propagated backward in time — using the simulation results — this provided the number of asteroids and asteroid impactors over the whole solar system history.

The size distribution of main-belt asteroids was used to set up the model size distribution. The size distribution is relatively steep for \( d \approx 10–15 \text{ km} \) (the cumulative power index \( y \approx -2 \)), flattens for \( d = 30–50 \text{ km} \) (\( y \approx -1 \)), and becomes much steeper for \( d > 100 \text{ km} \). The collisional evolution of asteroids during the early epochs was not accounted for in the model (see Nesvorný et al., 2017a, for a discussion); here we are mainly interested in the lunar crater record that likely does not reach back to these early stages (Section 8).

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\[^{8}\text{The E-belt, representing the inner extension of the main asteroid belt at 1.75–2 \text{ au}, was hypothesized by Bottke et al. (2012) to be an important source of terrestrial and lunar impactors. There are presently no asteroids in this region except for Hungarias with } i \approx 15^\circ–30^\circ.\text{ The E-belt population on low-inclination orbits was presumably cleared during the giant planet migration/instability.}\]
Nesvorný et al. (2017a) empirically approximated the impact profiles by a sum of three exponential functions, each with a different e-folding time. Here we find a simpler functional form with fewer parameters that approximates the asteroid impact flux equally well. In terms of the cumulative number of Earth impactors with diameters $d$ at times $t > t$, the new chronology function is

$$F(d, t) = F_1(d) \exp[-(t/\tau)^a] + F_2(d) T,$$

with $\tau = 65$ Myr, $a = 0.6$, $T = 4570 - t$ and $t$ in Myr. This expression should not be used for $t \sim 0$, because Eq. (2) does not take into account the effects of early collisional grinding.

The first term in Eq. (2) accounts for the dynamical decline of asteroid impactors during early epochs. The second term represents the constant impact flux in the last 3 Gyr. There are two size-dependent factors in Eq. (2), $F_1(d)$ follows the size distribution of main-belt asteroids (Bottke et al., 2005); the fit to simulation results gives $F_1(10\,\text{km}) = 225$. We use the factors mentioned above to scale the impact flux to any $10 \leq d \leq 100$ km. Also, from the main belt size distribution, $F_1(1\,\text{km}) = 3.0 \times 10^{20}$ (Section 10). $F_2(d)$ is calibrated on modern NEAs. Nesvorný et al. (2021a) estimated $\sim 3$ impacts of $d > 10$ km NEAs on the Earth per Gyr; we thus have $F_2(10\,\text{km}) = 3 \times 10^{-3}$ Myr$^{-1}$. Harris and D’Abramo (2015) and Morbidelli et al. (2020) estimated that $d > 1$ km NEAs impact the Earth on average every 0.75 Myr; we thus have $F_2(1\,\text{km}) = 1.3$ Myr$^{-1}$.

This fully defines the historical impact flux of asteroids on the Earth. We use the results of Nesvorný et al. (2017a) to obtain the impact flux for other terrestrial worlds as well. For the case considered here, Venus received $\sim 1.2$ times more impacts than the Earth, Mars received $\sim 2.9$ times fewer impacts, and the Moon received $\sim 20$ times fewer impacts (Mercury is not considered in this work). The overall Mars-to-Moon ratio in the number of asteroid impacts is $\approx 7$ (dominated by early impacts; see Section 13 for Mars). For reference, Nesvorný et al. (20000) reported the Mars-to-Moon ratio of $\sim 7.5$ from modeling impacts of large (modern) NEAs. We use these factors to re-scale the impact flux from the Earth to Venus, Mars and the Moon. This is an excellent approximation of the modeling results obtained in Nesvorný et al. (2017a). The mean impact speeds of asteroids on Venus, Earth, Mars and the Moon are 28.6, 23.5, 13.7, and 21.4 km s$^{-1}$, respectively (gravitational focusing included).

Fig. 4 shows the historical impact flux of asteroids on the terrestrial worlds. Most impacts happened early: for example, $\approx 90\%$ of impacts happened in the first 400 Myr. There were $\approx 239 \, d > 10$ km asteroid impacts on the Earth over the whole history of the solar system, $\approx 118 \, d > 10$ km impacts for $t > 42$ Myr (after the Moon formation in the case considered here), and $\approx 16 \, d > 10$ km impacts for $t > 650$ Myr (post-Imbrium). These fluxes imply, via the scaling factors mentioned above, only $118/3.5/20 \sim 1.7d > 20$ km (i.e., basin-scale) impacts on the Moon for $t > 42$ Myr. This implies that the asteroid contribution to lunar basin record was insignificant.\footnote{While this is true in the model presented here, where the asteroid depletion is relatively modest ($\lesssim 90\%$; Nesvorný et al., 2017a also see Roig and Nesvorný, 2015; Deienno et al., 2018), it has to be pointed out that some dynamical models imply much larger depletion ($> 90\%$, e.g., Clement et al., 2019). The asteroid contribution to the lunar basin record would presumably be more significant in these models.} We discuss this issue in detail in Section 8.

4.2. Comets

A model for cometary impactors was developed in Nesvorný et al. (2017b). To start with, a million cometesimals were distributed in a disk beyond Neptune, with Neptune on an initial orbit at 23 au. The bodies were given low orbital eccentricities, low inclinations, and the surface density $\Sigma(r) \propto r^{-1}$. The disk was truncated at 30 au to assure that Neptune stopped migrating near its current orbital radius at $\approx 30$ au (Gomes et al., 2004). The simulations were run from $t = 0$ (approximately the gas disk dispersal) to the present epoch ($t \approx 4.57$ Gyr). The effects of outer planet (early) migration/instability, galactic tides, and perturbations from passing stars were accounted for in the model. The results were shown to be consistent with the orbital distribution of modern comets, Centaurs and the Kuiper belt (Nesvorný et al., 2017b, 2019, 2020; Vokrouhlický et al., 2019).

The size distribution of outer disk cometesimals was calibrated from the number of large comets observed today, the number of $d > 10$ km Centaurs detected by OSSOS (Nesvorný et al., 2019), the size distributions of Jupiter Trojans and KBOs, and from the general condition that the initial setup leads to plausible migration/instability histories of the outer planets (see Nesvorný, 2018 for a review). The calibration gives $\sim 6 \times 10^9 \, d > 10$ km and $\sim 5 \times 10^7 \, d > 100$ km cometesimals in the original disk. The size distribution is expected to closely follow a power law with the cumulative index $\gamma \approx -2.1$ for $10 < d < 100$ km
of the scattered disk population, which is the main source of impactors during the early epochs, and the slower decay to the dispersal of the outer cometesimal disk, which was an important in the first ≃ 370 Myr. The wavy decline of cometary impacts is related to the initial number of comets in the original disk (see above).

The impact flux of comets on the terrestrial worlds was computed with the Öpik algorithm (Bottke et al., 1994; there were not enough planetary impacts recorded by the integrator to obtain the flux directly from the impact statistics). The results of the Öpik code were normalized to the initial number of comets in the original disk (see above). The calibrated model gives us the flux of cometary impactors over the whole history of the solar system. In total, over 4.57 Gyr, there were ≃ 10^7 impacts of d > 10 km comets on the Earth (Fig. 5). The great majority of cometary impacts happened early. For example, 90% of impacts happened in the first ≃ 55 Myr, and 99% of impacts happened in the first ≃ 370 Myr. The wavy decline of cometary impacts is related to the dispersal of the outer cometesimal disk, which was an important source of impactors during the early epochs, and the slower decay of the scattered disk population, which is the main source of modern comets.

An excellent approximation of the cumulative impact flux of comets on the Earth is

\[
F(d, t) = C_s(d) \left[ F_1 \exp[-(t/\tau_1)^{\alpha_1}] + F_2 \exp[-(t/\tau_2)^{\alpha_2}] + F_3(4570 - t) \right],
\]

(3)

with \(F_1 = F_2 = 6.5 \times 10^4\), \(\tau_1 = 7\) Myr, \(\alpha_1 = 1\), \(\tau_2 = 13\) Myr, \(\alpha_2 = 0.44\), \(F_3 = 4 \times 10^{-11}\) Myr^{-1}, \(C_s(d) = 1\) for \(d = 10\) km, and \(t\) in Myr (Fig. 5). The mean impact speeds of comets are higher than those of asteroid impactors: 28.2 km s^{-1} for the Earth,

\[\frac{\tau}{\tau^*} \approx 0.94\] times the number of Earth impacts,\(^{11}\) Mars \(\approx 5.0\) times fewer, and the Moon \(\approx 17\) times fewer. This gives the Mars-to-Moon ratio of cometary impacts \(\approx 3.4\). The mean impact speeds of comets are higher than those of asteroid impactors: 28.2 km s^{-1} for the Earth,

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\[\frac{\tau}{\tau^*} \approx 0.94\] times the number of Earth impacts,\(^{11}\) Mars \(\approx 5.0\) times fewer, and the Moon \(\approx 17\) times fewer. This gives the Mars-to-Moon ratio of cometary impacts \(\approx 3.4\). The mean impact speeds of comets are higher than those of asteroid impactors: 28.2 km s^{-1} for the Earth,
25.5 km s\(^{-1}\) for the Moon, 36.8 km s\(^{-1}\) for Venus, and 19.7 km s\(^{-1}\) for Mars (focusing included).

The physical lifetime of comets is shorter than their dynamical lifetime (Levison and Duncan, 1997; Levison et al., 2002), partly because comets become active and lose mass, but mainly because they spontaneously disrupt (Chen and Jewitt, 1994; Levison et al., 2006; Di Sisto et al., 2009). The exact mechanism behind the observed comet disruptions is unknown but rotational spin-up, thermal stresses and volatile-driven outbursts are all expected to contribute. In Nesvorný et al. (2017b), we tested several disruption laws and found that they produce comparable results. In the simplest of these models, the physical lifetime of comets is limited to a fixed number of perihelion passages \(N_p(q)\) below \(q\). To fit the orbital inclination distribution of observed \(d \sim 1\) km comets, we found \(N_p(q) \sim 500\) for \(q = 2.5\) au. The physical lifetime is expected to increase with comet size. To explain the observed number of \(d > 10\) km Jupiter-family comets, \(N_p(2.5) \sim 5,000\) is needed for \(d = 10\) km. This hints on a roughly linear dependence of \(N_p(2.5)\) on size, \(N_p(2.5) \sim 500 \times (d/1\) km), and \(N_p(2.5) \sim 50,000\) for \(d = 100\) km.

The impact flux of comets on the terrestrial worlds is reduced, relative to the expectations discussed above, if spontaneous comet disruptions are taken into account. We include comet disruption, using the approximate linear dependence described above, in the Ōpik calculation of the impact flux. The impact flux of comets on the Earth is reduced by a factor of \(\sim 59\) for \(d > 1\) km, \(\sim 6\) for \(d > 10\) km, and \(\sim 1.8\) for \(d > 100\) km. The reduction factors for Venus and the Moon are similar. The reduction factors for Mars are slightly lower, 36, 4.9, and 1.7, respectively. We account for these factors in Fig. 6. The total number of \(d > 10\) km comet impacts on the Earth is reduced to \(\sim 1,700\), of which \(\sim 200\) happen for \(t > 42\) Myr, and \(\sim 6\) happen for \(t > 650\) Myr. The Moon is expected to receive \(\sim 12\) \(d > 10\) km comet impacts after its formation (assumed here to happen at \(t \sim 42\) Myr with the instability at \(t < 10\) Myr). This is a factor of \(\sim 17\) below what would be needed to explain \(\sim 200\) \(D > 150\) km lunar craters (Bottke and Norman, 2017; a \(d = 10\) km impactor is assumed here to produce a \(D = 150\) km lunar crater). A more powerful source of lunar impactors is clearly needed.

4.3. Leftover planetesimals

4.3.1. N-body simulation setup

We are interested in the impact flux on the terrestrial worlds after the Moon-forming impact, \(t > 41.3\) Myr in the case described in Section 2. With this goal in mind we recorded the orbits of planets and planetesimals in the original simulation at \(t = 42\) Myr (i.e., shortly after the Moon-forming impact). The planetesimals in the asteroid belt region were ignored (see below and Section 4.1 for asteroids). Given that Mercury was too massive in the original simulation (Section 2), which could cause problems when examining the impact flux, we ignored Mercury in the follow-up simulations. All other planets, Venus to Neptune were included. We assumed that the terrestrial worlds were fully grown at \(t = 42\) Myr and their masses did not subsequently change. The orbits of planets \(t = 42\) Myr were taken from the original simulation (i.e., \(w = 0\) reset to the real orbits of the terrestrial planets).

To increase the model statistics, each planetesimal was cloned thousand times by slightly altering the velocity vector (< 10\(^{-6}\) fractional change). In total, we thus had 128,000 planetesimal clones at \(t = 42\) Myr. The \(N\)-body integrator known as \texttt{swift}\_\texttt{rmp4} (Levison and Duncan, 1994), which is an efficient implementation of the Wisdom-Holman map (Wisdom and Holman, 1991), was used to follow the system of planets and (massless) planetesimals over 1 Gyr. The simulation was split over 1280 Ivy-Bridge cores of the NASA Pleiades Supercomputer. All impacts of planetesimals on planets were recorded by the \(N\)-body integrator.

In the model considered here, planetesimals were originally distributed with a flat radial profile, \(\Sigma(r) \propto r^{-1}\), from 0.3 to 4 au (Section 2). The inner part of the planetesimal disk, \(r < 1.5\) au, contained the initial mass \(\approx 0.65 M_{\text{Earth}}\). This may or may not correctly approximate the physical conditions that existed in the inner solar system at \(t = 0\) (the time of gas disk dispersal). We developed a weighting scheme to test implications of different assumptions for the impact flux. Each planetesimal was assigned a weight, \(0 \leq w \leq 1\), depending on its starting orbit and the adopted surface density profile. For example, to test the effects of the planetesimal disk truncation at some outer radius, \(r_{\text{out}}\), the planetesimals with \(r < r_{\text{out}}\) at \(t = 0\) were given weights \(w = 1\), and the planetesimals with \(r > r_{\text{out}}\) were given weights \(w = 0\). For each assumption, the impact record was constructed by monitoring the weights of impactors recorded by the \(N\)-body integrator.

4.3.2. Collisional evolution model

We used the \textit{Boulder} code (Morbidelli et al., 2009a) to model the collisional evolution of planetesimals (Bottke et al., 2007). \textit{Boulder} employs a statistical particle-in-the-box algorithm that is capable of simulating collisional fragmentation of planetesimal populations. It was developed along the lines of other published codes (e.g., Weidenschilling et al., 1997; Kenyon and Bromley, 2001). In brief, for a given impact between a projectile and a target body, the algorithm computes the specific impact energy \(Q\), defined as the kinetic energy of the projectile divided by the total (projectile plus target) mass, and compares it with critical impact energy, \(Q_c\), defined as the energy per unit mass needed to disrupt and disperse 50% of the target. For each
collision, the mass of the largest remnant is computed from the scaling law for monolithic basalt (e.g., Benz and Asphaug, 1999).

The main input parameters of the Boulder code are: the (i) initial size distribution of the simulated populations, (ii) intrinsic collision probability \( p_i \), and (iii) mean impact speed \( v_i \). The initial size distribution is discussed in Section 5. The probabilities \( p_i(t) \) and velocities \( v_i(t) \) of mutual collisions between planetesimals were computed from the terrestrial planet simulation described in Section 2. We used the Opik algorithm (Botteke et al., 1994) and considered each pair of planetesimals at a time at each timestep. The results were averaged over all pairs (at each timestep) to give us a time-dependent description of the collisional environment. The leftover planetesimals were modeled as a single population in Boulder (e.g., we did not consider high-\( i \) or high-\( \iota \) orbits separately; simulations with the multi-annulus version of Boulder are left for future work). The collisional probabilities were initially high \( (p_i \sim 10^{-16} \text{ km}^{-2} \text{ yr}^{-1} \text{ for } r_{\text{cut}} = 1.5 \text{ au and } t < 3 \text{ Myr}) \), and quickly decreased as the planetesimal population dynamically and collisionally declined (Sections 5 and 6). The impact speeds between planetesimals increased from \( v_i \sim 10 \text{ km s}^{-1} \) at \( t \lesssim 5 \text{ Myr to } v_i \sim 20 \text{ km s}^{-1} \) later on. We implemented the impact speed dependence of \( Q_i \) from Leinhardt and Stewart (2012) (the disruption threshold shifts to higher specific energies when the impact speeds are higher).

5. Collisional evolution of planetesimals

The collisional evolution acted, along with the dynamical decay, to reduce the number of planetesimals in the terrestrial zone. We show below that the collisional grinding happened very early and produced divergence, where different (assumed) planetesimal populations at \( t = 0 \) evolved to similar populations by the time of the Moon-forming impact (both in the shape and overall normalization of the size distribution; Botteke et al., 2007). This result is used to approximately calibrate the number of leftover planetesimals and their impact profiles after Moon’s formation.

We defined the total mass of leftover planetesimals at \( t = 0 (M_0) \), and \( t = 42 \text{ Myr} (M_{42}) \), and performed collisional simulations with Boulder (Section 4.3.2) to understand the relationship between \( M_0 \) and \( M_{42} \) for different assumptions. Fig. 7 shows an example of Boulder run where we placed \( M_0 = 1 M_{\text{Earth}} \) in planetesimals at 0.5–1.5 au. The initial size distribution was assumed to be a broken power-law \( N(d) \propto d^{-\gamma} \) with \( \gamma = 1.5 \) for \( d < d^* \) and \( \gamma = 5 \) for \( d > d^* \), and \( d^* \sim 100 \text{ km} \). This setup was inspired by the streaming instability model of planetesimal formation (Youdin and Goodman, 2005), where the new-born planetesimals have the characteristic size \( d \sim 100 \text{ km} \), and the asteroid size distribution that shows a break at \( d \lesssim 100 \text{ km} \) (Morbidelli et al., 2009a). We do not consider very large, Ceres-class planetesimals that may have formed in the terrestrial planet zone, because the lunar impact record does not provide direct constraints on their population. According to our tests, however, energetic impacts on Ceres-class and larger objects produce fragments and this can influence the size distribution of planetesimals in the range relevant for the lunar impact record. Similarly, lunar impactors can be produced in collisions between protoplanets (Wishard et al., 2022). We leave these issues for future studies. The very large planetesimals may also have been important for the late delivery of highly siderophile elements (HSEs; Section 14) to the Earth (Bottke et al., 2010; Marchi et al., 2014) and Mars (Marchi et al., 2020).

The Boulder code was run from \( t = 0 \) to 1 Gyr. We found that the size distribution of planetesimals rapidly changed and reached an equilibrium shape by only \( t \sim 20 \text{ Myr} \). The subsequent collisional evolution was insignificant because the planetesimal population was reduced by a large factor \( \sim 100 \) by \( t = 42 \text{ Myr} \), collisional and dynamical removals combined. In this sense, the shape of the size distribution of leftover planetesimals at the time of the Moon-forming impact, and any time after that, is a fossilized imprint of the intense collisional grinding that happened in the first \( \sim 20 \text{ Myr} \) (Botteke et al., 2007).

Fig. 7. The collisional evolution of leftover planetesimals. Initially, at \( t = 0 \), the planetesimal disk at 0.5–1.5 au was assumed to have the mass \( M_0 = 1 M_{\text{Earth}} \), and a broken power-law distribution with a steeper slope for \( d > 140 \text{ km} \) and a shallower slope for \( d < 140 \text{ km} \) (dashed line). The black solid line shows the size distribution after 42 Myr of collisional grinding. The collisional evolution for \( t > 42 \text{ Myr} \) is negligible. For reference, we also show the size distribution of the main asteroid belt from WISE (green line; Mainzer et al., 2019) that was vertically shifted to plot near the black solid line. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

The equilibrium size distribution shows a break at \( d \approx 100 \text{ km} \), a shallow slope for \( d = 20–100 \text{ km} \), and a slightly steeper slope for \( d < 20 \text{ km} \). It is remarkably similar to that of the (scaled) asteroid belt (Fig. 7). This is an encouraging sign, because an asteroid-like size distribution is just what is needed to explain the size distribution of ancient lunar craters (Strom et al., 2005). Minton et al. (2015), however, pointed out that the main-belt-like size distribution of impactors would produce too many lunar basins. This could indicate that the size distribution of ancient impactors with \( d \gtrsim 10 \text{ km} \) was slightly steeper than that of today’s asteroid belt (Section 14).

The number of terrestrial planetesimals was strongly reduced by collisional grinding. With \( M_0 = 1 M_{\text{Earth}} \), and the initial size distribution defined above, there were \( \sim 1.9 \times 10^5 \text{ km} \), \( 6.7 \times 10^5 \text{ km} \), and \( 6.9 \times 10^5 \text{ km} \) planetesimals at \( t = 0 \). By \( t = 42 \text{ Myr} \), the planetesimal population was reduced to \( \sim 10^3 \text{ km} \), \( 2.6 \times 10^5 \text{ km} \), and \( 3.5 \times 10^5 \text{ km} \) planetesimals (\( \sim 19–26 \text{ reduction factors} \)). The total mass of the planetesimal population dropped by a factor of \( 19 \) to \( M_{42} = 0.053 M_{\text{Earth}} \). The numbers quoted above only account for the effects of collisional grinding. In the dynamical simulations, the number of planetesimals was reduced by a factor of \( \sim 3.1 \) from \( t = 0 \) to 42 Myr. Combining the dynamical and collisional removals we therefore estimate the overall reduction of \( \sim 59–81 \), and \( \sim 3.2 \times 10^5 \text{ km} \), \( 8.4 \times 10^5 \text{ km} \), and \( 1.1 \times 10^5 \text{ km} \) planetesimals at \( t = 42 \text{ Myr} \).12

12 An implicit assumption here is that the dynamical loss rate is independent of size.
The effects of collisional grinding depend on the initial mass $M_0$ (Bottke et al., 2007): stronger/weaker collisional grinding is expected if $M_0$ was higher/lower. We performed a large number of simulations with the Boulder code to characterize this dependence in detail. We found that the stronger grinding for larger initial masses leads to a situation where $M_{42}$, and the population leftover planetesimals at $t = 42$ Myr, do not sensitively depend on $M_0$, hence the aforementioned convergence. For $M_0 > 0.1 M_{\text{Earth}}$, we estimate $\sim (2.6-5.2) \times 10^5$ $d > 10$ km planetesimals at $t = 42$ Myr. The effects of collisional grinding are greatly reduced for $M_0 < 0.1 M_{\text{Earth}}$, but the number of planetesimals at $t = 42$ Myr ends up to be smaller in this case (because it was already small initially). For $M_0 = 0.03 M_{\text{Earth}}$, for example, we found $\sim 1.5 \times 10^5$ $d > 10$ km planetesimals at $t = 42$ Myr.

The results depend on other parameters as well, mainly on the shape of the initial size distribution and on the scaling law adopted in Boulder. Our choice of the initial size distribution described above puts most mass in $d \sim 100$ km planetesimals. It is possible that most mass was in smaller or larger planetesimals. If so, this would imply a smaller population of $d > 100$ km planetesimals at $t = 42$ Myr (Section 14). In the collisional simulations described above, we adopted the standard disruption law for monolithic basalt (Benz and Asphaug, 1999). Scaling laws that significantly differ from Benz and Asphaug (1999) (e.g., for porous bodies) would presumably produce different results (but note that most important collisions happen in the gravity regime where the material strength is not that important).

6. Impact flux of leftover planetesimals

In total, in the simulation described in Section 4.3.1, there were 14,091 impacts of leftover planetesimals on Venus, 9,418 impacts on the Earth, and 2,292 impacts on Mars in 1 Gyr. This represents 11%, 7.4%, and 1.7% of the population leftover planetesimals at the terrestrial planet zone at the time of the Moon-forming impact ($t = 42$ Myr). The number of planetesimal impacts on the Moon was obtained from the number of impacts on the Earth by re-scaling the results to the smaller cross-section and focusing factor of the Moon. This gives the Earth-to-Moon ratio of $\approx 20$, and the overall lunar impact probability of $\approx 0.4%$. The impact probabilities are substantially higher than the ones obtained for asteroids ($\approx 1%$ for Earth impacts from the inner main belt) and comets ($1.6 \times 10^{-6}$ for Earth impacts if comet disruption is ignored).

The raw impact profile — i.e., the (uncalibrated) impact profile where all planetesimals were given the same weight — is shown in Fig. 8. The initial impact flux decline is intermediate between those of asteroids and comets. Specifically, 90% of planetesimal impacts occur within 80 Myr (400 Myr for asteroids and 55 Myr for comets) and 99% of impacts occur within 300 Myr. The leftover planetesimals are therefore expected to produce a shorter tail of late impacts than asteroids. Compared to asteroids, however, the leftover planetesimals represented a much larger initial population and are therefore favored, by a large factor, to produce impacts for $t < 1$ Gyr (see below).

To a great accuracy, the profile shown in Fig. 8 can be approximated by a stretched exponential

$$F(d,t) = F(d) \exp\left[-\left(\frac{t}{\tau}\right)^\alpha\right]. \tag{4}$$

with $\tau = 12.2$ Myr and $\alpha = 0.5$. For comparison, Morbidelli et al. (2018) estimated $\tau = 10$ Myr and $\alpha = 0.5$. Our expression should not be used for $t < 10$ Myr, because Eq. (4) does not take into account the effects of early collisional grinding (Section 5). We assume that there were $4 \times 10^5 d > 10$ km planetesimals at $t = 42$ Myr (a factor of $\sim 2$ uncertainty; Section 5) and use this number to absolutely calibrate the impact flux for $t > 42$ Myr. To scale the results to larger impactors, given the results shown in Fig. 7, we adopt the asteroid belt size distribution as a reference. This gives $F(d) = 1.8 \times 10^7$ for $d = 10$ km, $F(d) = 5.1 \times 10^5$ for $d = 20$ km, and $F(d) = 6.9 \times 10^3$ for $d = 100$ km.

Similar results were obtained for Venus and Mars. The functional form in Eq. (4) provides an excellent approximation for Venus ($F(d) = 2.5 \times 10^5$ for $d = 10$ km — Venus receives $\approx 1.4$ times the number of Earth impacts). The same holds for Mars ($F(d) = 4.4 \times 10^4$ for $d = 10$ km — Mars receives 4.1 times fewer impacts than the Earth). For Mars, however, $\tau = 13.5$ Myr fits the model profile better than $\tau = 12.2$ Myr. This indicates that the decline of Mars impactors was slightly slower, presumably because the planetesimals with $r \approx 1.5$ au had longer dynamical lifetimes than the ones below 1 au (e.g., Deienno et al., 2019). For the Moon, we adopt the Earth impact profile and scale it down by a factor of $\approx 20$.

7. Dependence on the initial radial profile

The initial surface density profile, $\Sigma(r)$, of planetesimals in the inner solar system is unknown. The relatively low mass of the main asteroid belt ($\approx 4.5 \times 10^{-4} M_{\text{Earth}}$; DeMeo and Carry, 2013) probably suggests that the initial planetesimal mass at $r \approx 2-4$ au was low (see Raymond and Nesvorny, 2020 for a review). In the standard model of the terrestrial planet formation (Section 2), the initial planetesimal mass in the terrestrial planet region ($r < 1$ au) is usually assumed to be $M_p > 0.1 M_{\text{Earth}}$ (we adopted this assumption to calibrate the impact model in Section 6). How the profile changed from the high surface density at $r \sim 1$ au to the low surface density at $r > 2$ au is uncertain.

To fix the initial number of asteroids in the E-belt ($r = 1.75-2$ au), we adopted $\Sigma \propto r^{-1}$ for $r > 1.75$ au, and used the main belt
\( \tau \) was faster than in the case with \( r_{\text{out}} = 1 \) au. Note that having the parameterization of planetesimal impacts that we adopt in this work allowed to adjust as well, however, a slightly better fit was obtained \( \tau = 1.75 \text{ au} \) produced results that were very similar to those obtained partially compensate each other.

As for the impact profiles, the cases with \( r_{\text{out}} = 1 \) and \( r_{\text{out}} = 1.25 \text{ au} \) produced results that were very similar to those obtained for \( r_{\text{out}} = 1.5 \text{ au} \). The profiles can be approximated by Eq. (4) with \( r = 11.5 \text{ Myr}, a = 0.5 \), which is only a slightly shorter e-fold than \( r = 12.2 \text{ Myr} \) inferred from the raw profiles. When \( a \) was allowed to adjust as well, however, a slightly better fit was obtained with \( F = 3.8 \times 10^5, r = 6 \text{ Myr} \) and \( a = 0.45 \) (\( d > 10 \text{ km and Earth} \)). This is the parameterization of planetesimal impacts that we adopt in this work (Fig. 9). Note that having \( r = 6 \text{ Myr} \) does not mean that the decline was faster than in the case with \( r = 12.2 \text{ Myr} \) and \( a = 0.5 \), because \( a \) is now smaller (i.e., the exponential is more stretched). The half-life\(^{13} \) of impact flux decline steadily increases with time; it is \( \tau_{\text{half}} = 46 \text{ Myr} \) for \( t = 200 \text{ Myr} \) and \( t = 88 \text{ Myr} \) for \( t = 650 \text{ Myr} \). The case with \( r_{\text{out}} = 1 \) au could be interesting as well, but the planetesimal population declined more steeply in this case, and there are not as many late impacts as for \( r_{\text{out}} = 1.25–1.75 \text{ au} \).

The impact flux of leftover planetesimals on Venus closely follows the one obtained here for the Earth, but the overall impact rate is 1.4 times higher (for \( r_{\text{out}} = 1.5 \text{ au} \)). The lunar impact flux is \( \approx 20 \) times lower. The impact profile for Mars is flatter and can adequately be fit with \( F = 8.2 \times 10^4 \) for \( d > 10 \text{ km}, r = 6 \text{ Myr} \) and \( a = 0.42 \) (Fig. 9). This means that the lunar chronology should not strictly be applied to Mars. For example, the Imbrium-age Mars \( (t \approx 650 \text{ Myr or } T \approx 5.92 \text{ Ga}) \) would have accumulated \( \approx 18 \) times more planetesimal impacts than the same-age Moon, but a \( t = 42\text{-Myr old Mars} \) would have accumulated only \( \approx 4.1 \) times more planetesimal impacts than the Moon (values given here for the whole target surfaces, not per \( \text{km}^2 \)). The Mars-to-Moon impact ratio is expected to change for \( t > 1 \text{ Gyr} \), when the asteroid impacts start to dominate (the ratio \( \approx 7 \) is expected for asteroids; Section 4.1). The mean impact speeds of leftover planetesimals are \( 29.0, 23.2, 19.9 \) and \( 14.2 \text{ km s}^{-1} \) for Venus, Earth, the Moon, and Mars, respectively.

8. Lunar basins

We now turn our attention to lunar impacts. Fig. 10 shows the integrated history of lunar impacts including all major sources of impactors in the inner solar system: leftover planetesimals (Section 7), asteroids (Section 4.1), and comets (Section 4.2). We find that planetesimals dominated the early impact flux \( (t < 1.1 \text{ Gyr or } T > 3.5 \text{ Ga}) \). Asteroids took over and produced most impacts in the last \( \approx 3.5 \text{ Gyr} \). The cometary flux was never large enough, in the whole history of the inner solar system, to be competitive (the early instability at \( t \lesssim 10 \text{ Myr} \) is adopted here).\(^{14} \) In the first \( \approx 200 \text{ Myr}, \) comets produced more lunar impacts than asteroids, but the number of planetesimal impacts was far greater. Comets outpaced leftover planetesimals in the last \( \sim 3 \text{ Gyr}, \) but that was when the overall impact flux was ruled by asteroids. This would suggest that isotopic and other signatures of comets may be difficult to find on the lunar surface.

The overwhelming majority of impacts observed on the lunar surface must date back to \( T > 3.5 \text{ Ga} \), when the impact flux was orders of magnitude higher than it is today. The model predicts \( \approx 500 \text{ d} > 20 \text{ km} \) lunar impacts for \( t = 42 \text{ Myr} \) \((T < 4.53 \text{ Ga}) \). For comparison, modeling the lunar gravity anomalies detected by GRAIL, Miljković et al. (2016) found \( \approx 50 \) impacts of \( d > 20 \text{ km} \) bodies (Section 2). We therefore see that the number of impacts suggested by our model would be excessive, by a factor of \( \sim 10, \) if the Moon surface recorded all large impacts since its formation (Zhu et al., 2019a).\(^{15} \)

The Moon was fully molten when it accreted from the debris disk created by the giant impact on proto-Earth (see Canup et al., 2021, for a review). The subsequent evolution and solidification of the global lunar magma ocean (LMO) was controlled by a number of geophysical processes, including tidal heating, formation of an insulating flotation crust, etc. (Meyer et al., 2010; Elkins-Tanton et al., 2011). Radiogenic lunar crustal ages span from 4.47 to 4.31 Ga and suggest a prolonged stage of complete LMO solidification. (Shearer, 2006; Elkins-Tanton et al., 2011; Maurice et al., 2020). The lunar basins that formed while the LMO was still present would have been subject to

\[\text{Fig. 9. The impact flux of } d > 10 \text{ km leftover planetesimals the Moon (black line), Earth (blue), Venus (green), and Mars (red). The plot shows the accumulated number of impacts since time } t \text{ (i.e., on the surface with age } T), \text{ where } t = 0 \text{ (} T = 4.57 \text{ Ga}) \text{ is the birth of the solar system, and } t \approx 4.57 \text{ Gyr (} T = 0 \text{) is the present time. The profiles were normalized to having } 4 \times 10^5 \text{ d} > 10 \text{ km planetesimals at } t = 42 \text{ Myr}, \text{ and } r_{\text{out}} = 1.5 \text{ au}. \text{ The vertical dash-dotted lines show } t = 42 \text{ Myr and } t = 650 \text{ Myr for reference. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)} \]
were not as severe as we assumed here \(\approx 2\) \(\times 10^5\) \(\text{Ga}\), this could suggest that our nominal calibration of leftovers is a factor \(\approx 50\) to this time. If so, we would expect \(\approx 50\) \(d > 20\) km impacts to be recorded (Fig. 10), in a close agreement with the lunar basin record. Concerning the LMO lifespan, Morbidelli et al. (2018) and Zhu et al. (2019a) reached similar conclusions — they proposed \(T \approx 4.35\) \(\text{Ga}\), and transition from the planetesimal-dominated to asteroid-dominated impact stages (\(T \approx 4.38\) \(\text{Ga}\)).

It has to be noted that the LMO solidification is defined here in a specific context: the melt layer has to be thin enough for the topographic and crustal thickness signatures (Miljković et al., 2021). These early basins may be unidentifiable today.

The long-lived LMO would resolve the problem with the excess of planetesimal impacts on the young Moon. Suppose, for example, that the lunar surface started recording basin-scale impacts at \(T \approx 190\) Myr \((T \approx 4.38\) \(\text{Ga}\)), or roughly 150 Myr after Moon’s formation in our model — the oldest known basins (e.g., South Pole–Aitken) would date back to this time. If so, we would expect \(\approx 50\) \(d > 20\) km impacts to be recorded (Fig. 10), in a close agreement with the lunar basin record. Concerning the LMO lifespan, Morbidelli et al. (2018) and Zhu et al. (2019a) reached similar conclusions — they proposed \(T \approx 4.35\) \(\text{Ga}\) — based on the HSE constraint (Section 14). Scaling to smaller impactor sizes, we estimate \(\approx 180\) \(d > 10\) km impacts for \(T < 4.38\) \(\text{Ga}\), in a close agreement with \(\approx 200\) \(D > 150\) km lunar craters inferred in Bottke and Norman (2017) \((d = 10\) km impactor is assumed here to produce a \(D = 150\) km lunar crater; Johnson et al., 2016a).

The time of LMO solidification, \(T_{LMO}\), inferred here from the lunar basin record is uncertain. For example, for a slightly higher calibration with \(5.2 \times 10^5\) \(d > 10\) km planetesimals at \(T = 42\) Myr, we obtain \(T_{LMO} = 215\) Myr \((T_{LMO} \approx 4.35\) \(\text{Ga}\)). If, instead, there only were \(2.6 \times 10^5\) \(d > 10\) km planetesimals at \(T = 42\) Myr, then \(T_{LMO} = 160\) Myr \((T_{LMO} \approx 4.41\) \(\text{Ga}\)). Turning this argument around, the \(T = 4.31\) \(\text{Gyr}\) age for the youngest known crustal ages would imply \(T_{LMO} = 260\) Myr. This could suggest that our nominal calibration of leftovers is a factor of \(\approx 2\) too low, and that there were \(\approx 8 \times 10^5\) \(D > 10\) km planetesimals at \(T = 42\) Myr. It could indicate that the effects of collisional grinding were not as severe as we assumed here (e.g., stronger \(Q_D^\ast\) required).

9. Imbrium-era impacts

Having established that (nearly) all lunar basins formed from impacts of leftover planetesimals (only \(\approx 0.7\) and \(\approx 0.8\) \(d > 20\) km lunar

Fig. 10. Impacts of \(d > 20\) km bodies on the Moon (log time scale in the left panel, linear on the right). The plot shows the accumulated number of impacts since time \(t\), where \(t = 0\) is the birth of the solar system, and \(t \approx 4.57\) \(\text{Gyr}\) is the present time. The planetesimal, asteroid and comet profiles are shown by red, green and blue lines, respectively; the black line is the total impact flux. The vertical dash-dotted lines show the Moon formation in our model \((t = 42\) Myr), estimated start of the known lunar basin record \((t \approx 190\) or \(T \approx 4.38\) \(\text{Ga}\)), Imbrium formation \((t \approx 650\) Myr or \(T \approx 3.92\) \(\text{Ga}\)), and transition from the planetesimal-dominated to asteroid-dominated impact stages \((T \approx 1.1\) Ga or \(T \approx 3.5\) Ga). The numbers on the upper axis of the left panel indicate time \(T\) (in \(\text{Gyr}\)) measured looking back from today. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

extreme relaxation that would have reduced topographic and crustal thickness signatures (Miljković et al., 2021). These early basins may be unidentifiable today.

Concerning the LMO lifespan, Morbidelli et al. (2018) and Zhu et al. (2019a) reached similar conclusions — they proposed \(T \approx 4.35\) \(\text{Ga}\) — based on the HSE constraint (Section 14). Scaling to smaller impactor sizes, we estimate \(\approx 180\) \(d > 10\) km impacts for \(T < 4.38\) \(\text{Ga}\), in a close agreement with \(\approx 200\) \(D > 150\) km lunar craters inferred in Bottke and Norman (2017) \((d = 10\) km impactor is assumed here to produce a \(D = 150\) km lunar crater; Johnson et al., 2016a).

The time of LMO solidification, \(T_{LMO}\), inferred here from the lunar basin record is uncertain. For example, for a slightly higher calibration with \(5.2 \times 10^5\) \(d > 10\) km planetesimals at \(T = 42\) Myr, we obtain \(T_{LMO} = 215\) Myr \((T_{LMO} \approx 4.35\) \(\text{Ga}\)). If, instead, there only were \(2.6 \times 10^5\) \(d > 10\) km planetesimals at \(T = 42\) Myr, then \(T_{LMO} = 160\) Myr \((T_{LMO} \approx 4.41\) \(\text{Ga}\)). Turning this argument around, the \(T = 4.31\) \(\text{Gyr}\) age for the youngest known crustal ages would imply \(T_{LMO} = 260\) Myr. This could suggest that our nominal calibration of leftovers is a factor of \(\approx 2\) too low, and that there were \(\approx 8 \times 10^5\) \(D > 10\) km planetesimals at \(T = 42\) Myr. It could indicate that the effects of collisional grinding were not as severe as we assumed here (e.g., stronger \(Q_D^\ast\) required).

It has to be noted that the LMO solidification is defined here in a specific context: the melt layer has to be thin enough for the topographic and crustal thickness signatures of basins to survive and be identifiable today. Miljković et al. (2021) showed that impact basins experience extreme relaxation for melt layers as thin as \(\approx 10\) km (the melt layer needs to be close to the surface, Miljković et al. considered the melt layer in depths \(10\)–\(50\) km below the surface). Here we therefore provisionally interpret \(T_{LMO}\) as the time when the melt layer became thinner than \(10\) km. The youngest radiogenic crustal ages can post-date our \(T_{LMO}\) because they mark the time when the LMO solidification was brought to completion.

Oceanus Procellarum is the largest of the lunar maria, covering roughly \(10\)% of the total lunar surface. It is uncertain whether Procellarum is or is not an impact-generated basin (e.g., Andrews-Hanna, 2016). Here we estimate the probability of an impact of Ceres-class planetesimal on the Moon. Adopting the scaled asteroid-like distribution from Fig. 7 (green line), there would be \(\approx 100\) \(d > 1000\) km planetesimals for \(T = 42\) Myr, each having a \(0.4\)% of probability to impact on the Moon (Section 6). We can therefore roughly estimate that one \(d > 1000\) km impact on the Moon for \(T = 42\) Myr is a \(\approx 40\)% probability event (i.e., it can happen). This is consistent with the results reported in Zhu et al. (2019b), who considered impactor diameters \(\approx 500\)–\(1000\) km and found similarly large probabilities.
impacts of asteroids and comets are expected to happen for \( t > 190 \text{ Myr} \), respectively), we now consider the Imbrium-era impacts. What are our constraints? Evidence suggests that Imbrium formed at \( T \approx 3.92 \text{ Ga} \) (\( t \approx 650 \text{ Myr}; \) Zhang et al., 2019) by an impact of a \( d \gtrsim 100 \text{ km} \) object (Miljković et al., 2013, 2016; Schultz and Crawford, 2016). From the crater counts we know that the Orientale and Schrödinger basins formed after Imbrium (e.g., Orgel et al., 2018). These smaller basins were produced by \( d \approx 50-64 \text{ km} \) and \( d \approx 20 \text{ km} \) impactors, respectively (Miljković et al., 2016; Johnson et al., 2016b).

Having only two smaller basins with post-Imbrium formation ages is surprising. Adopting the asteroid belt size distribution, we find that there should be \( \sim 7.4 \) \( d > 20 \text{ km} \) impacts for every \( d > 100 \text{ km} \) impact. The expected number of \( d > 20 \text{ km} \) impacts would be even larger if we assumed a steeper size distribution (Minton et al., 2015). Our interpretation of this problem is that the Imbrium basin should have formed unusually late, by chance, relative to the expectation from the lunar impact chronology. In other words, statistically, the Imbrium basin formation needs to be an unusually late event.

The relatively late formation of some lunar basins is expected in our model. We find, for example, with the standard normalization considered here — \( 4 \times 10^6 \) \( d > 10 \text{ km} \) leftover planetesimals at \( t = 42 \text{ Myr} \), and the asteroid-like size distribution for \( d > 10 \text{ km} \) — that on average \( \approx 2.3 \) basins are expected to form for \( t > 600 \text{ Myr} \) (\( T < 3.97 \text{ Ga}; \) Fig. 10). Given the strong decline of planetesimal impacts for \( t > 600 \text{ Myr} \), any basin-scale impacts are expected to happen just after 600 Myr. As for Imbrium, we estimate that there should be \( \approx 0.31 \) \( d > 100 \text{ km} \) lunar impacts for \( t > 600 \text{ Myr} \), and that the Imbrium formation at \( t > 600 \text{ Myr} \) should therefore be \( \approx 23\% \) probability event (from the standard Poisson statistics), or 15%-35% considering the calibration uncertainty described in Section 5.

We now ask how likely it is for Imbrium to form late (\( t > 600 \text{ Myr} \), \( d > 100 \text{ km} \)) and have exactly two basins (Orientale and Schrödinger, \( d > 20 \text{ km} \) impactors) younger than Imbrium. This is a very restrictive conditions and the probability is therefore not expected to be high. But this is not the main point here. The main point is to understand whether a shorter or longer tail of impactors than the one obtained here, would fit data better. The condition is evaluated with our standard calibration of leftovers (see Nesvorný et al., 2022 for other calibrations). We fix \( \alpha = 0.45 \), vary \( r \) in Eq. (4), and generate a statistically large number of random impact sequences in each case.

We find that the probability has a broad maximum around \( r = 6 \text{ Myr} \) (Fig. 11), which was the best-fit e-fold that we obtained for leftover planetesimals. The probability drops for \( r < 5 \text{ Myr} \) because it becomes very difficult to obtain any late impacts in this case. It decreases for \( r > 6 \text{ Myr} \), because too many basins form after Imbrium if the decline of the impact flux is too slow. It thus cannot be argued that a very slow decline of impactors would help to explain the lunar basin record. Instead, the impact chronology obtained here (\( \alpha = 0.45 \) and \( r = 6 \text{ Myr} \)) is (nearly) optimal to satisfy the Imbrium-era constraints. A similar argument can be made about the overall normalization of leftover planetesimals. The results are not affected by small changes of the impact chronology. For example, the probability peaks at \( r \approx 9-12 \text{ Myr} \) when we use \( \alpha = 0.5 \) from Eq. (4), whereas \( r = 12.2 \text{ Myr} \) was our best fit with \( \alpha = 0.5 \) (Fig. 8).

The Imbrium-era impactors are often found to be stored on orbits with \( a \approx 1.3-1.7 \text{ au} \), \( e < 0.2 \), and in the mean motion resonances with Mars (e.g., 6:5, 7:6, 11:9; inside or outside of the Mars orbit). They either start with \( a \approx 1.3-1.7 \text{ au} \) at \( t = \) or are scattered to 1.3–1.7 au from \( \leq 1 \text{ au} \). The resonances provide a phase-protection mechanism against collisions with Mars. The orbits have low eccentricities such that they do not cross the orbit of Earth. The bodies stay in the resonances for hundreds of Myr, are eventually released to orbits with higher eccentricities, at which point they can impact. This is an example of the “storage places” hypothesized by Wetherill (1975). We are only able to see this because our model self-consistently followed the growth of the terrestrial planets and impact profile of leftover planetesimals (Sections 2 and 4.3).\(^{16}\)

10. Lunar \( N_{20} \) chronology

To compare our model results with lunar chronologies we need to compute the cratering flux for small impactors. We start discussing the \( N_{20} \) chronology (Fig. 12A), where we have more confidence in the model results (given that no wild extrapolations to very small sizes are needed). For the lunar impact speeds of leftover planetesimals and asteroids obtained here (\( \approx 20 \text{ km s}^{-1} \); Sections 4.1 and 7), we assume that a \( d = 1 \text{ km} \) projectile makes a \( D = 20 \text{ km} \) lunar crater (see Morbidelli et al., 2018 for a discussion of the scaling laws). The overall mean impact velocity of planetesimals on the Moon is lower (\( v_i \approx 17 \text{ km s}^{-1} \)), but the mean value is mainly contributed by the very early impactors that have, in general, lower impact speeds (because they are on orbits with \( a \approx 1 \text{ au} \)). The mean impact speeds of planetesimals for \( t > 200 \text{ Myr} \) are higher (\( v_i \approx 20 \text{ km s}^{-1} \)), and more similar to asteroids (\( v_i \approx 21 \text{ km s}^{-1} \); Section 4.1). This gives some justiﬁcation to using the same projectile size for both types of impactors.

To compute \( N_{20}(T) \) from our impact model, we assume that the size distribution of leftover planetesimals for \( 1 < d < 10 \text{ km} \) was similar to that of today’s asteroid belt (Section 5). There are roughly \( 1.1 \times 10^6 \)

\(^{16}\) If we would switch to the real orbits of planets at \( t = 42 \text{ Myr} \), or at any other time during the simulation, the resonant planetesimals would be prematurely released and this would (presumably) produce a shorter impact tail.
main-belt asteroids with $d > 1$ km (Bottke et al., 2020) and $\approx 8200$ with $d > 10$ km (Mainzer et al., 2019). We therefore scale the planetesimal flux profile from $d > 10$ km to $d > 1$ km by a factor $1.1 \times 10^7/8200 = 134$. This gives $F(1 \text{ km}) = 2.5 \times 10^2$ in Eq. (4) for the whole lunar surface ($r = 6 \text{ Myr}$ and $a = 0.45$ from Section 7). The same scaling factor is used for the early flux of asteroids (the first term in Eq. (2)). The constant term in Eq. (2) for $d > 1$ km is obtained from $d > 1$ km NEAs at the present epoch. Morbidelli et al. (2020) estimated that the average time interval between $d > 1$ km NEA impacts on Earth is $0.75 \text{ Myr}$ (also see Harris and D’Abramo, 2015). We thus have $F_d(1 \text{ km}) = 1.3 \text{ Myr}^{-1}$ in Eq. (2).

Neglecting comets, which should only contribute at the $< 10\%$ level, a simple $N_{20}(T)$ chronology function — the number of accumulated $D > 20$ km craters in area $10^6 \text{ km}^2$ of the lunar surface since $T$ — is given by

$$N_{20}(T) = a \exp[-(t/6 \text{ Myr})^{0.45}] + b \exp[-(t/65 \text{ Myr})^{0.3}] + cT$$ (5)

with $t = 4570 - T$ (t and $T$ given in Myr), $a = 6.7 \times 10^4$, $b = 40$ and $c = 1.7 \times 10^{-3} \text{ Myr}^{-1}$. The first term in Eq. (5) stands for leftover planetesimals, the second and third for asteroids. The second term is much less important than the other two and can be discarded for a rough estimate of $N_{20}$. Fig. 12A compares our model chronology with crater densities measured for different lunar terrains.

Fassett et al. (2012) reported $N_{20} = 30 \pm 5$ for the $\approx 3.92 \text{ Gyr}$ old Fra Mauro/Imbrium highlands. This value was slightly revised in Orgel et al. (2018), giving $N_{20} = 26 \pm 5$ (purple symbol in Fig. 12A), which nearly perfectly aligns with our model chronology obtained for the standard calibration of terrestrial planetesimals. We cannot plot $N_{20}(T)$ for other basins from Fassett et al. (2012) and Orgel et al. (2018) in Fig. 12A, because their radiometric age is unavailable or uncertain. For reference, we also show data inferred from Neukum et al. (2001). For that, we take the $N_1$ values reported in Neukum et al. (2001) and scale them to $N_{20}$ with the Neukum’s “old” production function ($N_1/N_{20} \approx 1000$). We believe this is a reasonable approach because the crater counts on old surfaces were typically done for relatively large craters, and were extrapolated down to $N_1$ with the same production function. Reporting them as $N_{20}$ should therefore be equally valid.

The $N_{20}$ value inferred from Neukum et al. (2001) for the Fra Mauro/Imbrium highlands is slightly larger than the more recent crater counts discussed above; it plots near the upper limit of our leftover calibration range. The results for young lunar maria are more discrepant. Here the $N_{20}$ values inferred from Neukum et al. (2001) plot below our chronology function, indicating lower crater densities, and the difference becomes larger for younger lunar maria (Oceanum Procellarum, Mare Imbrium, Mare Crisium, Mare Fecundatis) than for the older ones (Taurus Littrow, Mare Tranquilitatis).

This may mean one of several things. At the face value, inferred $N_{20}$ densities could indicate that our chronology function should more steeply drop for $T < 3.5 \text{ Ga}$ (the dashed line in Fig. 12A shows an example with $c = 0.85 \times 10^{-3} \text{ Myr}^{-1}$). For that to work, however, the Earth impact rate of $d > 1$ km asteroids at $T \approx 3.1-3.5 \text{ Ga}$ would have to be lower, by a factor of $\sim 2$, than the current impact rate (e.g., Morbidelli et al., 2020; the current impact rate of large NEAs is known to a better than $\sim 10\%$ precision), suggesting the number of impacts would have to increase at some point in the last 3 Gyr (Culler et al., 2000; Mazrouei et al., 2019).

Another possibility is that the $N_{20}$ values inferred from Neukum et al. (2001) for $T \approx 3.1-3.5 \text{ Ga}$ are systematically a factor of $\sim 2$
lower. Note that the density of \( d > 20 \) km craters on young lunar maria is very low and subject to small number statistics. We therefore extrapolated from smaller craters, which were actually counted with some confidence, to estimate \( N_{S_0} \). But perhaps the actual production function is flatter for \( D \geq 5 \) km than the one we adopted from Neukum et al. (2001), and would thus give a larger \( N_{S_0} \) value.17 Alternatively, the crater counts reported in Neukum et al. (2001) for young lunar maria may be lower than the actual values. The results of Robbins (2014) give some credit to this possibility, because their \( N_1 \) counts for young lunar maria (the actual counts of \( D > 1 \) km craters) are a factor of \( \sim 1.5-2 \) higher than \( N_1 \) from Neukum et al. (2001). If this difference propagates to \( N_{S_0} \), it could be fully responsible for the problem shown in Fig. 12A.

Our chronology can be used to estimate the formation ages of lunar basins from \( N_{S_0} \) counts. For example, the age of the Nectaris basin with \( N_{S_0} = 170 \) per \( 10^6 \) km\(^2\) (Orgel et al., 2018), is estimated to be \( T = 4.21-4.29 \) Gyr, where the uncertainty is dominated by the uncertain calibration of the lunar chronology function in Eq. (5) (here given for the two chronologies shown by solid lines in Fig. 12A). For comparison, Orgel et al. (2018) suggested a younger, \( T = 4.17 \)-Gyr age for the Nectaris basin from the “old” Neukum chronology function.

11. Lunar \( N_1 \) chronology

We have less confidence in extrapolating our model results to \( N_1(T) \), because very small bodies are subject to a host of dynamical (e.g., radiation effects; Vokrouhlický et al., 2015) and physical (e.g., rotational spin-up and mass loss) effects that were not modeled in this work. The small body populations such as the main asteroid belt are also not well characterized for \( d < 1 \) km. Some useful information can nevertheless be obtained by assuming that the impact flux of small asteroids was roughly constant in the last \( \sim 3 \) Gyr and equal to the impact flux of modern NEAs.

From the scaling laws we estimate that a \( d \approx 40 \) m asteroid impactor would produce a \( D \approx 1 \) km lunar crater.18 There are \( \sim 5 \times 10^5 \) \( d > 40 \) m NEAs (Rottke et al., 2020; debiased data from updates of Harris and D’Abramo, 2015). Assuming the usual impact probability \( p_1 = 1.5 \times 10^{-4} \) Myr\(^{-1}\) of NEAs with the Earth (e.g., Morbidelli et al., 2015), and scaling things to the Moon, we obtain \( \epsilon = 1.0 \) Myr\(^{-1}\) (per \( 10^6 \) km\(^2\) of lunar surface).19 This estimate is consistent with a recent analysis of planetary impacts from small NEAs (Nesvorný et al., 2000).

There are no useful constraints on the population of \( d > 40 \) m leftover planetesimals. To obtain a rough estimate of \( N_1(T) \) for \( T > 3.5 \) Ga, we adopt \( N_1/N_{S_0} = 1400 \) from Morbidelli et al. (2018) (also see Marchi et al., 2012). Note that this is just a reference ratio that turns out to give an acceptable fit to the \( N_1 \) data (Fig. 12B). This does not mean, however, that the adopted value is strictly correct. For example, a somewhat higher calibration of leftover planetesimals (e.g., 2.5 \( \times 10^5 \) \( d > 10 \) km planetesimals at \( t = 42 \) Myr) would still fit the data in Fig. 12A) quite well. If that calibration is adopted, \( N_1/N_{S_0} \sim 1000 \) — consistent with the “old” production function from Neukum et al. (2001) — would work well in Fig. 12B.

Neglecting comets, a simple \( N_1(T) \) chronology function — the number of accumulated \( D > 1 \) km craters in the area \( 10^6 \) km\(^2\) of lunar surface since \( T \) — is given by

\[
N_1(T) = a \exp[-(t/6 \text{ Myr})^{0.45}] + b \exp[-(t/65 \text{ Myr})^{0.41}] + cT
\]

with \( t = 4570 - T \) (and \( T \) in Myr), \( a = 9.4 \times 10^7 \), \( b = 5.6 \times 10^3 \) and \( c = 1.0 \) Myr\(^{-1}\) (again, the second term is much less important than the other two and can be neglected). The \( c \) coefficient is only \( \sim 16 \)% higher than the one in Eq. (1) (Neukum et al., 2001; Robbins et al., 2014), and would thus give a larger \( N_1 \) value obtained from counts of larger craters per Myr per \( 10^6 \) km\(^2\), which are the units used here). The main difference with respect to Eq. (1) in Neukum et al. (2001) is that the first term in Eq. (1) is an exact exponential function, whereas here we have a stretched exponential in Eq. (6).

Fig. 12B compares our \( N_1(T) \) chronology, with all the caveats mentioned above, to various crater counts and chronologies from Neukum et al. (2001) and Marchi et al. (2009). The main focus in Fig. 12B is the data from Robbins (2014) (blue dots in Fig. 12B), which are the actual counts of \( D \sim 1 \) km craters for Apollo and Luna landing sites; all other “data” points are the extrapolations from counts of larger craters.20

The \( N_1 \) crater counts from Robbins (2014) are slightly higher, in general, that the extrapolations from Neukum et al. (2001) and Marchi et al. (2009) (except for the old Mare Tranquillitatis age). They plot near the higher end of our chronology range (i.e., near the higher end of planetesimal calibration), but the agreement is reasonably good. While this gives some justification to the choice of parameters described above, we caution that other choices would potentially work as well. In particular, there is a degeneracy between the leftover planetesimal calibration and the \( N_1/N_{S_0} \) factor. If, for example, the NEA population was taken as a guide, and assuming that \( d = 40 \) m \((d = 1 \) km\)) impactors make \( D = 1 \) km (\( D = 20 \) km) craters, we would find \( N_1/N_{S_0} \sim 540 \), a factor of 2.6 below the \( N_1/N_{S_0} \) value adopted above. To obtain \( N_1/N_{S_0} = 1400 \) from the NEA population, one would have to assume that \( d = 25-30 \) m impactor makes \( D = 1 \) km lunar crater.

Comparing \( N_1 \) from Fassett et al. (2012) with \( N_1 \) from Neukum et al. (2001), both for Fra Mauro/Imbrium highlands, gives \( N_1/N_{S_0} \sim 1400 \). With \( N_1 \) from Fra Mauro/Imbrium highlands from Robbins (2014), we get \( N_1/N_{S_0} \sim 1800 \). Playing the same game for the young lunar maria, where the \( N_1 \) counts are secure (Robbins, 2014), but the \( N_2 \) counts require an extrapolation (Neukum et al., 2001), gives \( N_1/N_{S_0} \sim 700 \). Overall, these values may indicate that there was a trend of \( N_1/N_{S_0} \) with \( T \), with the younger terrains showing lower \( N_1/N_{S_0} \) values than the old terrains. If this is interpreted in terms of the size distribution of impactors, we would speculatively infer that the size distribution of leftover planetesimals below \( d \approx 1 \) km was (slightly) steeper than that of modern NEAs.

Additional \( N_1 \) calibration data point comes from the analysis of Chang’e-5 samples by Che et al. (2021), who found \( N_1(T) = 1200-1800 \) (per \( 10^6 \) km\(^2\)) for \( T = 1963 \pm 57 \) Ma old lunar terrain. Using the standard \( N_1(T) \) chronology from Eq. (6), we find \( N_1(T) \approx 1980 \) for \( T = 1963 \) Ma — only a slightly higher value than \( N_1 \) reported by Che et al. (2021). This is consistent with the observation made in Che et al. (2021) that the existing chronology curves, which are calibrated on very young lunar terrains \((T < 1 \text{ Ga})\), may indicate that the impact flux at \( T = 2 \) Ga may have been slightly lower (by \( \approx 20\% \)) than the one estimated here.

20 We do not show the chronology function from Robbins (2014), but note that their data-driven fit has a very different shape.
12. Archean spherule beds

When a large impactor strikes the Earth, it produces a vapor-rich ejecta plume containing numerous small melt droplets, most of which rise above the atmosphere. As the plume cools down, glassy spherules form and fall back, producing a global layer that can be several millimeters thick (for a Chicxulub-sized impact event) to many centimeters thick (Johnson and Melosh, 2012). The late Archean (2.5–3.5 Ga) spherule beds are thicker than those associated with the 66 million years old, 180 km wide Chicxulub crater — estimated $d > 10$ km impactor (Collins et al., 2020) — and should have therefore been produced by $d > 10$ km impactors (Bottke et al., 2012; Johnson et al., 2016a). Some $\sim 16$ spherule beds have been found in the late Archean (e.g., Marchi et al., 2021), although preservation biases and incomplete sampling may be an issue. At least some of these layers may have been produced by very large, $d \sim 50$ km impactors.

The impact flux of $d > 10$ km bodies on the Earth is shown in Fig. 13. We find $\sim 20$ of $d > 10$ km impacts on the Earth for $T = 2.5–3.5$ Ga. This is similar to the number of known spherule beds in the late Archean period. The leftover planetesimals and main-belt asteroids contribute equally to impacts in this time frame ($\sim 10$ impacts each).

Whereas the asteroid impacts were more uniformly spread over late Archean, nearly all planetesimal impacts should have happened for $T > 3$ Ga. The model predicts that $\sim 10$ and $\sim 2.2 \times 10^3$ km asteroids should have impacted the Earth in the last 2.5 and 0.6 Gyr, respectively. Assuming that the number ratio of $d > 10$ km to $d > 50$ km impactors is $8.9$, as inferred from the size distribution of main belt asteroids, the model implies $\sim 2 d > 50$ km impactors in late Archean.

13. Martian crater chronology

The impact flux of $d > 20$ km impactors on Mars is shown in Fig. 14. The Martian impact profile is different from that of the Moon, Earth and Venus. The impact profile is more extended in time and has a longer tail of late impacts. This is simply because leftover planetesimals at $r_o$ live longer and can impact later. The transition from the planetesimal-dominated to asteroid-dominated impact stages thus probably occurred later for Mars than for the Moon ($t \approx 1.4$ Gyr vs. $t \approx 1.1$ Gyr). The impact flux of leftover planetesimals on Mars is more sensitive to $r_o$ than that of any other terrestrial world. The results shown in Fig. 14 were obtained for our nominal case with $r_o = 1.5$ au (and $4 \times 10^5 \text{ km asteroids at } t = 42 \text{ Myr}$). The overall number of planetesimal impacts on Mars is $\approx 1.4$ times lower for $r_o = 1.25$ au, and $\approx 2$ lower for $r_o = 1$ au. This would reduce the planetesimal contribution and shift the transition back in time.

Following the method described in Section 11, the $N_i(T)$ chronology (per $10^6$ km$^2$; Fig. 16) for Mars can be given as

$$N_i(T) = a \exp\left(-\frac{c T}{6 \text{ Myr}}\right) + c T$$

with $a = 3.7 \times 10^7$ and $c = 0.41 \text{ Myr}^{-1}$. The second term in Eq. (6) — corresponding to the early impact flux of asteroids — in not included here for simplicity. Here we truncated the planetesimal disk at $r_o = 1.5$ au (at $t = 0$). The contribution of leftover planetesimals is $\approx 2$.
times lower for $r_{\text{out}} = 1 \text{ au}$; hence $a = 1.9 \times 10^7$ for $r_{\text{out}} = 1 \text{ au}$. The asteroid branch — factor $c$ in Eq. (7) — is rescaled from lunar $c = 1.0 \text{ Myr}^{-1}$ (Eq. (6)) using input from the dynamical models of modern NEAs (e.g., Granvik et al., 2018). The models indicate $R_b \approx 1.2$ for small NEAs (Nesvorný et al., 2000), where $R_b$ is the number ratio of Mars-over-Moon impacts normalized to a unit surface area; Hartmann and Neukum, 2001). Note that this value is much lower than the one adopted in Hartmann (2005) and Marchi (2021), $R_b \approx 2.6$.

We assume the impact velocities 14–15 km s$^{-1}$ for Mars and 19–20 km s$^{-1}$ for the Moon, fold in the effect of different surface gravities of the Moon and Mars, and estimate that Mars requires a ~1.4 times larger impactor than the Moon to create a $D = 1 \text{ km}$ crater (Johnson et al., 2016a). Our reference asteroid size distribution (Harris and D’Abramo, 2015; Bottke et al., 2020) has a steep slope for $20 < d < 50 \text{ m}$ with the cumulative power index $\approx 3.2$. The Martian $N_1(T)$ in Eq. (7) is therefore penalized by a factor of $1.4^{1.2} \approx 2.9$ relative to the lunar chronology. We use the same penalization for the planetesimal and asteroid branches.

Fig. 15 compares the Martian chronology obtained here with the chronologies from Hartmann (2005), Werner et al. (2014) and Marchi (2021). We plot two chronologies from Eq. (7), one for $r_{\text{out}} = 1 \text{ au}$ (thin black line) and one for $r_{\text{out}} = 15 \text{ au}$ (thick black line), to illustrate the dependence on the initial planetesimal profile. There is a significant difference between the two (the planetesimal branch of $N_1(T)$ is a factor of $\approx 2$ lower for $r_{\text{out}} = 1 \text{ au}$). This is bad news for the prospect of accurate dating: the age estimates derived from the two chronologies differ by up to ~200 Myr for $T > 2 \text{ Gyr}$. But this can be good news for the prospect of constraining the radial extension of the planetesimal disk from the Martian crater record — assuming that the radiometric age of an old Martian terrain with known $N_1$ will be measured in the near future (e.g., NASA Mars 2020).

The asteroid branches ($T < 2.5 \text{ Ga}$) of the Hartmann (2005) and Werner et al. (2014) chronologies are a factor of ~1.5 higher and lower, respectively, that the one derived here. We find a good agreement with the chronology of Marchi (2021) for $T < 2 \text{ Ga}$.

Looking back in time, the Hartmann, Werner, and Marchi chronologies continue relatively flat to $T > 3 \text{ Ga}$, where they show a sharp bend upward. Our chronologies instead connect to the planetesimal branch and start raising already at $T \sim 2.5 \text{ Ga}$.

Our best age estimate for the Noachian/Hesperian and Hesperian/Amazonian boundaries is 3.4–3.6 Ga and 2.6–2.9 Ga (the range given here for $r_{\text{out}} = 1–1.75 \text{ au}$ and the standard calibration of leftover planetesimals). Here we adopt the crater densities $N_1 = 4.8 \times 10^3$ and $N_1 = 1.6 \times 10^4$, both per 10$^6$ km$^2$, as defining the two boundaries (Tanaka, 1986; Hartmann, 2005). The Jezero crater — relevant to the NASA Mars 2020 mission — is estimated to be 2.2–2.5 Gyr old for $N_1 = 1.1 \times 10^3$ per 10$^6$ km$^2$ from Warner et al. (2020) or 2.4–2.7 Gyr old for $N_1 = 1.5 \times 10^2$ per 10$^6$ km$^2$ from Shahrzad et al. (2019).

For illustrative purposes, following the traditional approach (e.g., Hartmann and Neukum, 2001; Hartmann, 2005; Werner et al., 2014; Marchi, 2021), we also approximately rescale our lunar chronology from Eq. (6) to Mars. For that, we use $R_b = 1$ (roughly applicable for $r_{\text{out}} = 1.25–1.75 \text{ au}$) for the planetesimal branch and $R_b = 1.2$ for the asteroid branch (Nesvorný et al., 2000), and penalize Martian $N_1(T)$ by the factor of 2.9 to account for the impact velocity and surface gravity difference. For comparison, Marchi (2021) used $R_b = 0.5$ for planetesimals from Morbidelli et al. (2018), which is smaller than 0.81 reduction factor from Popova et al. (2003).
In Section 8, we estimated the Martian surface age. For comparison, we also show the Martian chronologies from Hartmann (2005, green line), Werner et al. (2014, blue), and Marchi (2021, red); early instability/NEA scaling. The horizontal dot-dashed lines are the crater counts for Noachian/Hesperian and Hesperian/Amazonian boundaries, and for the Jezero crater terrains. The vertical dashed lines are our age estimates for these units in the model with $r_{out} = 1.5$ au. The ages inferred with $r_{out} = 1$ au are ~ 200 Myr older. The numbers on the upper axis indicate time $T$ (in Gyr) measured looking back from today. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

Our value, presumably because their terrestrial formation model was effectively run with $r_{out} \approx 1$ au. Indeed we find $R_c = 0.63$ for $r_{out} \approx 1$ au (statistics based on all impacts). This is consistent with a smaller planetesimal contribution for disks truncated at $r_{out} \approx 1$ au (for Mars, not in general — the planetesimal branch for the Earth is only slightly steeper for $r_{out} = 1$ au than for $r_{out} = 1.5$ au, and the overall number of terrestrial impacts remains nearly the same — again adopting $4 \times 10^3 d > 10$ km planetesimals at $t = 42$ Myr, independently of $r_{out}$). Fig. 16 shows that the Mars chronology function is more extended in time than the lunar chronology, implying that $R_c$ is time dependent. Applying the lunar chronology function to Mars can lead to inaccurate age estimates that can differ, by up to ~ 500 Myr, from the age estimates obtained from the accurate Martian chronology.

Scaling from the most densely cratered surfaces of the Moon and Mars, Bottke and Norman (2017) estimated ~ 200 $D > 150$ km craters over the whole lunar surface, and ~ 500 $D > 150$ km craters over the whole Martian surface. From the scaling laws we find that a $d \approx 10$ km impactor is needed to produce a $D = 150$ km crater on the Moon, and a $d \approx 13$ km impactor is needed to make a $D = 150$ km crater on Mars (Holsapple and Housen, 2007; Johnson et al., 2016b; Morbidelli et al., 2018). In Section 8, we estimated ~ 180 $d > 10$ km lunar impacts for $T < 4.38$ Ga, which is consistent with Bottke and Norman (2017), assuming that the lunar surface recorded $D > 150$ km craters since $T \approx 4.38$ Ga (presumably the LMO solidification time). To obtain ~ 500 $D > 150$ km craters for Mars with our model-derived impact flux, we infer that the Martian surface would have to record $D > 150$ km craters since $T \approx 4.27$ Ga for $r_{out} = 1.5$ au, and $T \approx 4.35$ Ga for $r_{out} = 1$ au. Morbidelli et al. (2018) noted the same problem and suggested a global resurfacing event at $T \sim 4.4$ Ga, perhaps associated with the formation of the Borealis basin. There is currently no evidence for the late formation of Borealis. Robbins (2022) proposed that the formation of Borealis basin could have kept the surface warm enough for long enough to prevent large craters/basins from forming for an extended period of time. The volcanic and fluvial resurfacing of early Mars (Hartmann and Neukum, 2001) would have to be extremely powerful to globally erase $D > 150$ km craters for $T > 4.3$–4.4 Ga.

14. Discussion

We find that the terrestrial-zone planetesimals were the dominant source of lunar impactors for $T > 3.5$ Ga, and that asteroids were the dominant source of impactors for $T < 3.5$ Ga. This is in line with

Fig. 15. The $N_1(T)$ chronologies for Mars. The bold and thin black lines show our Martian chronologies for $r_{out} = 1.5$ au and $r_{out} = 1$ au, respectively. For comparison, we also show the Martian chronologies from Hartmann (2005, green line), Werner et al. (2014, blue), and Marchi (2021, red); early instability/NEA scaling. The horizontal dot-dashed lines are the crater counts for Noachian/Hesperian and Hesperian/Amazonian boundaries, and for the Jezero crater terrains. The vertical dashed lines are our age estimates for these units in the model with $r_{out} = 1.5$ au. The ages inferred with $r_{out} = 1$ au are ~ 200 Myr older. The numbers on the upper axis indicate time $T$ (in Gyr) measured looking back from today. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

Fig. 16. A comparison of lunar and Martian chronologies. The thin solid line is the lunar $N_1(T)$ chronology rescaled to Mars as described in Section 13. The bold solid line is the Martian chronology inferred in this work for the standard calibration of leftover planetesimals and the original planetesimal disk truncated at 1.5 au (Eq. (7)). The horizontal dot-dashed lines are the crater counts for Noachian/Hesperian and Hesperian/Amazonian boundaries, and for the Jezero crater terrains. The vertical dashed lines are our age estimates for these units. The ages inferred from the rescaled lunar chronology would be ~ 400–500 Myr older. The numbers on the upper axis indicate time $T$ (in Gyr) measured looking back from today.
the findings of Morbidelli et al. (2018).\textsuperscript{23} The leftovers are expected to evolve collisionally (Botte et al., 2007)\textsuperscript{24} and reach an asteroid-belt-like size distribution in only $\sim 20$ Myr after the first solar system solids. The size distribution of lunar impactors in the last $3.5$ Gyr should be similar to that of modern NEAs. The modern NEAs evolve from the asteroid belt by size-dependent radiation processes which favor mobility of very small bodies (Vokrouhlický et al., 2015). They therefore have slightly steeper size distribution for $d < 10$ km than the main belt. This explains why the crater size distribution on ancient lunar craters is related to the main belt asteroids, and why the modern impactors have a NEA-like size distribution (Strom et al., 2005). Head et al. (2010) suggested that the transition between the two populations of impactors happened $\sim 3.5$ Ga, which is what we find here from dynamical modeling.

Minton et al. (2015) pointed out that the asteroid-like size distribution of early lunar impactors would produce too many mega-basins ($D > 1200$ km) and suggested that the size distribution of impactors below $d = 100$ km was somewhat steeper than that of today’s asteroid belt. Johnson et al. (2016a) reiterated that point and proposed that the lunar basin data and Archean spherule beds could best be fit with a main-belt-like size distribution of impactors for $d \lesssim 50$ km, and a steeper slope for $d \gtrsim 50$ km. Here we find that the ancient impactors were leftover planetesimals from the terrestrial planet zone ($r \lesssim 1.5$ au), not asteroids. This makes it easier to understand any inferred differences. Note, however, that the size-distribution break at $d \sim 50$ km would imply fewer Imbrium impacts, and would diminish the probability of the late Imbrium formation in our model.

We showed that the lunar crater record is consistent with having $(2.6-5.2) \times 10^3$ $d > 10$ km planetesimals in the terrestrial planet zone ($\sim 0.5-1.5$ au) at $\sim 50$ Myr after the first solar system solids. The collisional evolution in the first $\sim 20$ Myr would have been stronger for higher initial planetesimal mass and weaker for lower initial mass. We therefore cannot predict, from the collisional modeling alone, the initial planetesimal mass. All that we can say is that there was at least $\sim 0.1$ $M_{\text{Earth}}$ in planetesimals to start with. This is significant, however, because it shows that there was a large population of planetesimals to start with. The formation models where the terrestrial planets grow from cm-size pebbles (e.g., Johansen et al., 2021) do not postulate any large planetesimal population in the terrestrial planet zone (they do not exclude it either).

Using the lunar basin record as a constraint on the LMO solidification (Section 8), we found that the LMO should have lasted to $T = 3.46-4.42$ Ga, roughly 160–210 Myr after the first solar system solids, and $\sim 110-160$ Myr after the Moon-forming impact in our model (Section 2). Morbidelli et al. (2018) reached similar conclusions based on considerations related to the HSEs in the lunar mantle. They considered the possibility that HSEs were sequestered from the mantle of the Earth during magma ocean crystallization, due to iron sulfide exsolution (Rubie et al., 2016), and showed that this likely affected the Moon as a well (if the lunar mantle overturn is taken into account; Elkins-Tanton et al., 2011). The HSE would accumulate in the lunar mantle only after the LMO crystallization, estimated to happen $\sim 100-150$ Myr after the Moon formation. This would correspond to $\sim 150-200$ Myr after the solar system solids if the Moon formed at $t \sim 50$ Myr.

Fig. 17A shows the HSE constraints for the Earth, Mars, and Moon. Assuming chondritic composition of impactors it has been inferred that these worlds accreted $3 \times 10^{23}$ g, $1.6 \times 10^{23}$ g, and $1.5 \times 10^{23}$ g during the Late Veneer (i.e., after their differentiation is fully over such that the accreted HSEs do not sink to the core). Assuming 30% retention of impactor mass for the Moon (Zhu et al., 2019a), we find that the LMO would have to solidify at $t = 200$ Myr ($T = 4.37$ Ga) to explain lunar HSEs. This is in an excellent agreement with the results of Morbidelli et al. (2018). The Earth differentiation should have ended within $\sim 10$ Myr after the Moon-forming impact (e.g., Elkins-Tanton et al., 2011). We find that the accreted mass for $t > 50$ Myr is a factor of $\sim 3$ too low to explain Earth’s HSEs (it would be a factor of $\sim 10$ too low if the Moon-forming impact happened at $t \sim 120$ Myr; Maurice et al., 2020; Kruijer et al., 2021). This shows the need for the accretion of very large planetesimals ($\sim 1000$–3000 km; Bottke et al., 2010; Marchi et al., 2014). The contribution of very large planetesimals to Earth’s HSEs was not accounted for in Fig. 17A, because we adopted a steep size distribution for $d > 100$ km (Fig. 7). The stochastic accretion of very large planetesimals could help to explain the large difference in the HSE content between the Moon and Earth ($\sim 2000$; Bottke et al., 2010).\textsuperscript{25}

If Mars accreted and differentiated early ($t \lesssim 10$ Myr, Dauphas and Pourmand, 2011; Marchi et al., 2020), it would have accreted $5 \times 10^{23}$ g in our model during the Late Veneer (in small planetesimals). This is a factor of $\sim 3$ higher than the late addition of chondritic material inferred from Martian meteorites (shergottite-nakhliite-chassigny, SNC; Marchi et al., 2020). It is possible that the average mantle abundance of HSEs is underestimated, perhaps because the average mantle abundance may be difficult to establish from (heterogeneous) SNC meteorites, or perhaps because our SNC collection is not fully representative of the Martian mantle. The retention of impactor mass could contribute as well. In Fig. 17A, we assumed a 100% retention factor for Mars but at least some work indicates that the retention factor may be lower (e.g., $\sim 60$% in Artemieva and Ivanov, 2004). Finally, as we already discussed, the impact flux of planetesimals on Mars is lower when the disk of terrestrial planetesimals is truncated at the lower orbital radius. For example, if $r_{\text{ret}} \approx 1$ au, instead of $r_{\text{ret}} = 1.5$ au that we used in Fig. 17A, the accreted mass would be reduced by a factor of $\approx 2$.

The cometary impact profile is more extended in time than previously thought (Morbidelli et al., 2018). This is a consequence of the much improved statistic of simulations reported in Section 4.2. With the standard comet calibration, accounting for spontaneous comet disruptions, and assuming that the Moon-forming impact happened $\sim 50$ Myr ($T = 4.52$ Ga) after the birth of the solar system, we find that the Earth would have accreted $\sim 1.5 \times 10^{22}$ g of cometary material for $T < 4.52$ Ga (Fig. 17B). This is consistent with comets being the source of noble gases in the Earth atmosphere (Marth et al., 2016). The mass accreted in cometary material would be larger if: (i) some comets fade instead of disrupting (Fig. 17B), (ii) the instability happened later, or (iii) the Moon formed earlier. In fact, the noble gas argument can be used to roughly constrain the delay between the instability and Moon formation, $\Delta t$. From Fig. 17B we infer $20 < \Delta t < 60$ Myr, with the exact value depending on the physical lifetime of comets (Section 4.2). This suggests, if the instability happened very early ($t < 10$ Myr; Clement et al., 2018; Liu et al., 2022), that the Moon must have formed early as well ($t < 70$ Myr or $T > 4.5$ Gy; Thiemens et al., 2019).

We reported the impact profiles of leftover planetesimals from the best terrestrial planet formation simulation from Nesvorný et al. (2021a). This falls short, given that only one case was tested, to understand the possible variability of the impact profiles when different assumptions are made. This should be the focus of future investigations. If the variability is large, it would perhaps be possible to rule out/confirm some specific setups based on the lunar crater constraints.

\textsuperscript{23} Brassé et al. (2020), instead, suggested that the early impact record was dominated by bodies from the E-belt.

\textsuperscript{24} Unlike Botke et al. (2007), we find that the effects of collisional grinding were not strong enough to prevent the leftover planetesimals from producing the dominant share of lunar basins.

\textsuperscript{25} A great share of this difference is explained by the smaller accretional cross-section of the Moon (factor of $\sim 20$, focused included), long-lived lunar LMO (factor of $\sim 10$), and lower retention factor of the Moon (factor of $\sim 3$; Zhu et al., 2019a).
Icarus 399 (2023) 115545

The originality of the present paper consists in explaining the lunar collisional evolution that we use in this work to calibrate the population of planetesimal impactors from new Boulder simulations.

15. Comparison with previous work

The results address several critical scientific issues, some of which date back to the Apollo program, including the late formation of Imbrium and Orientale, and provide context for the interpretation of classical/empirical impact chronologies. The model impact flux of planetesimals and asteroids on the Earth matches the number of spherule classes and provides context for the interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

For example, the initially very massive asteroid belt (Clement et al., 2019); $\sim 1 M_{\text{Earth}}$, would presumably lead to a much larger (a factor of $> 1000$) contribution of asteroid impactors to the early impact record. We tested the effects of the radial profile of the planetesimal disk in the terrestrial planet zone and found that the cases with the outer disk edge at $r \sim 1-1.75$ au work quite well to match different constraints (with the mass $> 0.1 M_{\text{Earth}}$, in planetesimals at the time of the gas disk dispersal). Whereas this is broadly consistent with at least some planetesimal formation models (e.g., Izidoro et al., 2021; Morbidelli et al., 2022), the planetesimal population needs to be better characterized (e.g., from the radiometrically calibrated Martian chronology) before a more specific inference can be made.

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find that the late lunar impactors were stored on Mars-crossing orbits at 1.2–1.7 au (Section 9).

Brasser et al. (2020), hereafter B20, developed a dynamical model for lunar impactors with four different components: E-belt, asteroid belt, comets and leftover planetesimals. The general methodology used in B20, is similar to that used in the present work (numerical integrations used to construct the impact model). The main methodology-related difference between our work and B20 is that the B20 model is uncalibrated and can (and must, see below, because it does not fit the lunar record) be adjusted to satisfy constraints. Our model is more rigid. All impactor populations in our model were calibrated from independent means (Sections 4 and 5). Our model can either fit or not fit the lunar record — there is no middle ground or room for data-driven adjustment.

Some of the main results in B20 were presented in terms of \( N_{20} \) in their Section 5.2 (their Figs. 5 and 8). Here, B20 adjusted their model to match the Neukum’s lunar chronology. This was done by assigning weights to leftover planetesimals and E-belt. The best-fit contribution of planetesimals was found to be zero; leftovers were therefore concluded to be only a minor source of lunar impacts. E-belt was assigned a weight equal to \( \sim 10 \) (to fit the Neukum chronology). This implies that the E-belt population would have to be increased by a factor of \( \sim 10 \) to fit the existing data. B20 proceeded by explaining why this choice may be reasonable.

This can be related to the present work where we found the dominant role of leftover planetesimals — with the standard calibration, no need for any (arbitrary) enhancement. We considered E-belt as an extension of the asteroid belt (this is what the letter E stands for; Bottke et al., 2012) and assumed that it is not justified to arbitrarily increase its initial population if things do not work (Section 4.1). There is no obvious reason for why the initial number density of asteroids immediately below 2 au should have discontinuously increased from that immediately above 2 au. B20 argued that the probability of Rhea/Callia basin formation on Vesta would be too low if the E-belt population were not enhanced, but this is simply not the case. In a work dedicated to Vesta’s cratering, Roig and Nesvorný (2020) determined a \( \sim 50\% \) probability to form Rhea/Callia over 4.5 Gyr (with standard assumptions). The Hungary population, which represents an important constraint on the E-belt (Bottke et al., 2012), was ignored in B20.

B20’s asteroid flux is \( \sim 10–20 \) times higher (as explicitly noted in their Section 5.2) than the one obtained here, where we calibrate asteroid impactors on the number of large main belt asteroids observed at the present time (and their orbital distribution; Nesvorný et al., 2017a). This represents an anchor. We use (forward) numerical simulations of asteroids, tie the results to the anchor at the present time, and compute the historical impact flux from that. This is a well-defined calibration method. B20 ‘calibrated’ the E-belt impactors on the lunar crater data, but that presumes that the E-belt was the dominant source of lunar impactors in the first place.

The contribution of asteroids to lunar impacts was simulated in B20 by assuming that all planets formed and remained on their current orbits. This ignored dynamical effects of the outer planet migration/instability. When the outer planet migration/instability is accounted for, the main asteroid belt is depleted by a much larger factor and the impact profile is different (more asteroid impactors early on). B20 therefore underestimated the contribution of the main belt to early lunar impacts (by a factor \( \sim 10 \) relative to the enhanced E-belt; compare Fig. 8 in B20 to Fig. 10 in Nesvorný et al. (2017a).

The outer planet migration/instability is needed to reduce the population of inner belt asteroids and match the orbital structure of the asteroid belt (Nesvorný et al., 2017a). It is not explained in B20 how

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26 Brasser et al. (2021) followed the implications of the model developed in B20 for HSEs. We do not comment on Brasser et al. (2021) here, because B20 is more closely related to our work.
that the Imbrium formation from leftover planetesimals was a ~ 15–35% probability event, and explain that many more younger/smaller lunar basins than Imbrium would be expected (only two are known: Orientale and Schrödinger), if the probability were much larger than that (Section 9).

B20 did not consider the terrestrial impact flux during the late Archean, and the general impact record in the inner solar system in the last 3.5 Gyr. It is therefore difficult to establish whether their results would satisfy the additional constraints considered in the present work.

16. Summary

We developed an accurate dynamical model that accounts for all major sources of impactors in the inner solar system (leftover planetesimals, asteroids and comets). Each impactor population was calibrated from independent means (not using crater records; Section 3). Here we summarize the main results:

- The leftover planetesimals produced most lunar impacts in the first 1.1 Gyr (t < 1.1 Gyr or T > 3.5 Ga). Asteroids were the main source of impacts in the last 3.5 Gyr. The transition from leftover planetesimals to asteroids has been imprinted in the crater size distributions (Strom et al., 2005; Head et al., 2010; Orgel et al., 2018). The comet contribution to the crater record is found to be insignificant (for the early instability case adopted here). This would suggest that isotopic and other signatures of comets may be difficult to find on the lunar surface.
- Some 500 d > 20 km planetesimals from the terrestrial planet zone (0.5–1.5 au) are expected to impact the Moon since its formation. The early crater record must have been erased plausibly because the lunar surface was unable to support basin-scale impact structures. The ~ 50 km lunar basins formed after t ≃ 160–215 Myr (T ≤ 4.36–4.41 Ga). This is consistent with the long-lived LMO (lunar magma ocean; (Elkins-Tanton et al., 2011; Morbidelli et al., 2018; Zhu et al., 2019a)). The Nectaris basin is estimated to be T = 4.21–4.29 Ga old (from crater counts). The South Pole–Aitken basin should date back to T ≃ 4.36–4.41 Ga.
- About two lunar basins are expected to form for t ≥ 650 Myr (T ≤ 3.92 Ga). The Imbrium basin formation (T ≃ 3.92 Ga, d > 100 km impactor) is estimated to happen with a ~ 23% probability in our model. Imbrium should have formed unusually late, relative to the expectations from the lunar impact chronology, to have only two smaller/younger basins than Imbrium (Orientale and Schrödinger); there would be many more younger/smaller basins otherwise (Section 9).
- The Imbrium-era impactors were leftover planetesimals that were stored on orbits with a ≃ 1.3–1.7 au, e < 0.2, and in the mean motion resonances with Mars (e.g., 6/5, 7/6, 11:9). The resonances provided a phase-protection mechanism against collisions with Mars. The orbits had low eccentricities such that they did not cross the orbit of Earth. The bodies stayed in the resonances for hundreds of Myr and were eventually released to orbits with higher eccentricities. This is an example of the "storage places" hypothesized by Wetherill (1975).
- The lunar and Martian chronologies can be given as a sum of two terms: the stretched exponential function (the leftover planetesimal branch) and a constant (the asteroid or NEA branch). This is similar to the classical (empirical) crater chronologies of Neukum et al. (2001) and Hartmann and Neukum (2001), except that the cratering rate profile in the first ~ 1 Gyr had a longer tail than the exact exponential. This can lead to modest, ~ 50 Myr differences in the estimates of lunar basin ages.

- The Martian chronology is found to have a slower decline at late times than the chronologies of the Moon, Earth and Venus. This is a consequence of long dynamical lifetimes of bodies at ~ 1.5 au — these bodies are more likely to produce late impacts on Mars. Applying the lunar chronology function to Mars can lead to age estimates that can differ, by up to ~ 500 Myr, from the age estimates obtained from the accurate Martian chronology (Fig. 16A).
- The Noachian/Hesperian and Hesperian/Amazonian boundaries are found to be T ≃ 3.4–3.6 Ga and T = 2.6–2.9 Ga, respectively (the range given here for r_{out} = 1–1.75 au and the standard calibration of leftover planetesimals). The Jezero crater is estimated to be 2.2–2.5 Gyr old for N_j = 1.1 x 10^3 per 10^6 km^2 and 2.4–2.7 Gyr old for N_j = 1.5 x 10^3 per 10^6 km^2. The asteroid branch of the Martian chronology (r <= 2 Ga) is found to be similar to the chronology developed in Marchi (2021).
- Our model predicts ≈ 20 d > 10 km impacts on the Earth for T = 2.5–3.5 Ga. This is similar to the number of known spherule beds in the late Archean (Bottke et al., 2012; Johnson et al., 2016a; Marchi et al., 2021). Both the leftover planetesimals and main-belt asteroids contribute to impacts in this time interval. Whereas the asteroid impacts were more uniformly spread over the late Archean, nearly all planetesimal impacts should have happened before 3 Ga.
- The cometary impact profile is more extended in time than thought previously. For example, assuming the instability at t < 10 Myr, 90% of impacts happened in the first ≃ 55 Myr, and 99% of impacts happened in the first ≃ 370 Myr. With the Moon-forming impact at t = 50 Myr (T = 4.52 Ga), we find that the Earth should have accreted ~ 1.5–2.5 x 10^{22} g of cometary material for T < 4.52 Ga. This is consistent with comets being the source of noble gasses in the Earth atmosphere (Marty et al., 2016).
- To explain lunar HSEs, we find — in agreement with Morbidelli et al. (2010) and our lunar basin results (Section 8) — that the LMO should have solidified at t ~ 200 Myr after the first solar system solids (T ~ 4.37 Ga). The mass accreted by the Earth for t > 50 Myr is a factor of ~ 3 too low to explain Earth’s HSEs. This shows the need for the stochastic late accretion of very large planetesimals (Bottke et al., 2010).
- Mars would have accreted a factor of ~ 3 more HSEs for t > 10 Myr in our nominal model (the standard calibration of planetesimals and r_{out} = 1.5 au) than what is inferred from the SNC meteorites (Marchi et al., 2020). It is possible that: (i) the average abundance of HSEs in the Martian mantle is underestimated (e.g., the SNC meteorites are not representative), (ii) the impactor-retention factor of Mars is smaller than 100%, and/or (iii) the accreted mass was lower because the ring of terrestrial planetesimals was originally confined to 0.5–1 au.

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Data availability

Data will be made available on request.

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27 Except that we used the crater record to infer at least ≥ 0.1 M_{Earth}, in planetesimals in the terrestrial zone at the time of the gas disk dispersal.