Survival times of meter-sized rock boulders on the surface of airless bodies

A.T. Basilevsky a,b, J.W. Head b, F. Horz c, K. Ramsley b

Abstract

Rock boulders are typical features of the surfaces of many airless bodies, so the possibility of estimating their potential survival times may provide insights into the rates of surface-modification processes. As an opening point of this study we employ estimates of the survival times of meter-sized boulders on the surface of the Moon based on analysis of the spatial density of boulders on the rims of small lunar craters of known absolute age (Basilevsky et al., 2013), and apply them, with necessary corrections, to boulders on other bodies. In this approach the major factor of rock destruction is considered to be impacts of meteorites. However another factor of the rock destruction, thermal fatigue due to day–night cycling, does exist and it was claimed by Delbo et al. (2014) as being more important than meteorite impacts. They concluded this on the basis of known presence of fine material on the surface of small asteroids, claiming that due to extremely low gravity on those bodies, the products of meteorite bombardment should leave these bodies, and thus their presence indicates that the process of thermal fatigue should be much more effective there. Delbo et al. (2014) made laboratory experiments on heating–cooling centimeter-sized samples of chondrites and, applying some assumptions and theoretical modeling concluded that, for example, at 1 AU distance from the Sun, the lifetime of 10 cm rock fragments on asteroids with period of rotation from 2.2 to 6 h should be only ~10^3 to 10^5 years (that is ~3.5 × 10^7 to 1.5 × 10^7 thermal cycles) and the larger the rock, the faster it should be destroyed.

In response to those conclusions we assessed the results of earlier laboratory experiments, which show that only a part of comminuted material produced by high-velocity impacts into solid rocks is ejected from the crater while another part is not ejected but stays exposed on the target surface and is present in its subsurface. This means that the presence of granulometrically fine material on the surface of small asteroids does not prove the predominance of thermal stresses over rupture by meteorite impacts as a factor in the comminution of the surface material. We then analyzed images of lunar rocks of decimeters–to–meters–size whose lunar surface exposure ages were radiometrically dated. This analysis shows that the presence of the fragment on the lunar surface for a time period 26–35 × 10^6 to 3 × 10^7 years (that is, ~3 × 10^9 to 5 × 10^10 day–night thermal cycles) did not lead to the formation of any features conclusively supporting rock destruction by thermal cycles. In turn, this means that on the lunar surface as well as on the surface of other bodies at 1 AU and further from the Sun, the destruction of rocks by thermal fatigue is secondary compared to rock rupture by the meteorite impacts. The possible implications of the difference in environments on fast spinning asteroids and on the Moon require additional analysis.

Then utilizing the entire catalog of inner solar system minor planet orbits out to Jupiter as a proxy for the distribution of potential impactors throughout the inner solar system, we calculated the meteorite flux and impact velocities for a number of airless bodies to use them for estimates of survival times of rock boulders on their surfaces (normalized to those for lunar boulders). We found that if the average survival time for meter-size rock boulders on the surface of the Moon is 1, then considering rupture by the meteorite impacts as the major process of rock destruction, for Phobos it is ~0.8, for Deimos ~0.7, for asteroid Itokawa ~1, for Eros ~0.3, for Vesta and Ceres ~0.03 and for the average of the first 150 Trojans discovered is ~12.5. Implications of these findings are that on the surfaces of Vesta and Ceres, compared to the Moon, the regolith layer should generally have a larger thickness and higher maturity, while small craters with rocky ejecta are rare. On the typical Trojans, where impact flux is closer to that...
1. Introduction

Rock fragments and boulders are typical features of the surfaces of many airless bodies and estimating their potential survival times may provide significant insights into the rates of surface-modification processes and thus the geologic histories of these bodies in general. Initial estimates of the survival times of hand specimen-sized rocks exposed to the lunar surface environment were made based on laboratory impact experiments into the collisional fragmentation of rocks combined with estimates of the lunar surface meteorite flux and stochastic modeling of the impact environment (e.g., Horz et al., 1975a, b; Cintala and Horz, 2008). The aim of the present study is to estimate the survival times of meter-sized rock boulders seen in the numerous orbital images of the surfaces of airless bodies (e.g., Thomas et al., 2001; Robinson et al., 2010; Mazrouei et al., 2014). As an opening point, we employ estimates of the survival times of meter-sized boulders on the surface of the Moon, as recently described by Basilevsky et al. (2013), and apply them, with necessary corrections, to boulders on other airless bodies. This new approach by Basilevsky et al. (2013) was based on the analysis of the spatial density of boulders on the rims of small lunar craters of known absolute formation age. Additionally, Basilevsky et al. (2013) assumed that the major factor in boulder destruction is catastrophic disruption by the meteorite impacts. The potential role of diurnal temperature cycling was mentioned there but not considered in detail.

In the present analysis we first briefly review the lunar results. Then we discuss the potential role of thermal cycling, and we will show, that it is probably less important than the collisional disruption by meteorite impacts. Finally, we consider the meteorite bombardment on a number of airless bodies in terms of projectile flux and velocities and, based on this, we will estimate the survival times of meter-sized boulders on various bodies.

2. Survival times of rock fragments on the surface of the Moon

As defined by Gault and Wedekind, (1969), a rock is deemed destroyed when its largest collisional fragment is < 0.5 the original target mass (M0); this condition obviously mandates some critical kinetic energy (Ecrit) of the impactor. This threshold energy can be delivered either by a single impact or—in cumulative fashion—by a number of modestly less energetic events (Horz et al., 1986). Single events > Ecrit will result in “overkill”, i.e. progressively more fine-grained fragment populations (Fujiwara, 1989; Cintala and Horz, 2008) compared to the Ekin = Ecrit case. Single impact events at < Ecrit will merely produce a crater in the target object, possibly some penetrative fractures, but will not result in the physical disintegration of the target. The specific energy to produce a crater may be expressed as ergs/g of displaced crater mass, and it is typically an order of magnitude smaller than Ecrit, the latter expressed as ergs/g of M0; (Fujiwara et al. 1989; Cintala and Horz, 2008). The generation of relatively few, penetrative fractures—such as those consuming modest amounts of energy—is typically insufficient for the collisional destruction of rocks and contrasts with the additional energy needed to contribute and physically eject the relatively fine-grained ejecta in a cratering event. This order of magnitude difference in the specific energy needed for the cratering and collisional destruction process also implies, that the catastrophic fragmentation of rocks on the Moon, but the impact velocities are by factor 4 lower, the situation should be the opposite: thinner layer of regolith, lower maturity and a larger percentage of small craters with rocky ejecta. These predictions and observations can be tested with future robotic and human exploration of the Moon and small bodies.
agreement with the results of Ghent et al. (2014) who on the basis of analysis of the data of Diviner instrument, Lunar Reconnaissance Orbiter, estimated the areal fraction occupied by exposed rocks of 1 m and larger associated with 9 craters of 19–97 km in diameter, and having ages from 4–1000 Ma.

Basilevsky et al. (2013) noted that these estimates of the survival times of meter-sized boulders are some factor of 5 shorter than those extrapolated from Horz et al. (1975a), probably because the larger targets are mechanically weaker due to the increased number and size of intrinsic flaws. This seems supported by appropriately scaled impact experiments of Housen and Holsapple (1999) who showed that the effective target strength of a 200 cm object decreases by a factor of 3–4 relative to a 20 cm object. Shorter survival times than those of Horz et al. (1975a) could also suggest that diurnal temperature cycling (Delbo et al., 2014, Molaro et al., 2015) may contribute to the destruction of lunar surface rocks.

A very important point (e.g., Horz et al., 1975a,b) is that the process of catastrophic rupture by meteorite impacts has a stochastic character; there will be some “unlucky” boulders destroyed in the very beginning of the time of exposure of the boulder population, and some “lucky” boulders that remain intact for very long times; the difference between the destruction time of the first 10% of rocks and the 99% destruction level is approximately a factor of 12. This disappearance of collisionally destroyed surface rocks over prolonged times seems a fundamental difference compared to temperature cycling, which should produce more uniform destruction times for rocks of a given mass, as all Sun-illuminated rocks are universally subjected to the same process; however specific illumination conditions and/or variable thermo-mechanical properties of rocks may also lead to widely variable survival times of similarly sized rocks. The relatively protracted history of boulder destruction illustrated in Fig. 2 seems to support collisional destruction as the major process.

3. Destruction of rocks by diurnal temperature cycling

High diurnal temperature variations on the lunar surface (up to 280 K, Vaniman et al., 1991) imply that thermal stresses may contribute to the destruction of rocks. However, neither the Surveyor Project Final Report (1968) nor Lunar Source Book (1991) mention destruction of lunar rocks due to thermal surface cycling as the potential process. Probably the first suggestion on the role of...
of thermal stress destruction of lunar rocks was made by Florensky et al. (1978) based on analysis of the morphology of the rock fragments seen in TV panoramas taken by Lunokhod 1. Evidence for the presence of that process included fractures on the surfaces of some rocks, small rock pieces adjacent to the main rock appearing to be spalls, and fillets of fine material in cases where they looked different than the adjacent regolith. The effectiveness of this process in comparison with meteorite bombardment was not considered in that analysis.

A quantitative approach in the analysis of the destruction of rocks on the surfaces of airless bodies was reported by Molaro and Byrne (2012). They used a simple, one-dimensional heat conduction model of a unit area surface on the Moon, Mercury and Vesta to calculate the maximum rates of temperature change (dT/dt) and explore optimal parameters for rapid temperature changes. They suggested that damage occurs in the form of microscopic cracks that result from a thermal cycle or a thermal shock. In the model runs for surfaces on the Moon and Mercury, they found the highest magnitude shocks are well below the canonical threshold of 2 K/min to initiate cracks, although on the surfaces of Vesta this heating rate may exceed. They also found that the magnitude of dT/dt values is primarily controlled by sunrise/sunset durations on fast-spinning bodies, such as Vesta, and by the distance to the Sun on slowly rotating objects, such as Mercury. The strongest temperature shocks are experienced by highly sloped sun-or antisun-pointing surfaces. The authors also concluded that they cannot say with any certainty whether or not the process competes effectively with other surface modification processes (e.g., from meteorite bombardment).

In subsequent work, Molaro and Byrne (2014) continued to consider this problem and came to different conclusions concerning the Moon and Vesta. They modeled grain-scale thermoelastic stresses produced on airless surfaces using a 2-D finite-element modeling program designed to simulate the behavior of microstructures. Their preliminary results indicate that thermoelastic stresses induced on lunar surfaces are likely to be high enough to cause rock breakdown. According to Molaro and Byrne (2014), the calculated diurnal maximum effective tensile stresses on the lunar surfaces may reach 529 MPa, while typical tensile strengths of rocks are on the order of 100 MPa. For the equatorial surface of Vesta, the calculated stress is 19 and 68 MPa, for undamaged and porous microstructures, respectively, thus not reaching the typical 100 MPa tensile strengths of rocks.

This direction of study was recently pursued further by Delbo et al. (2014), who interpreted the presence of fine debris on the surface of small asteroids as due to thermal surface cycling. They state that because of the very low gravity on the surface of these bodies, ejecta from meteorite impacts should escape the body, but space missions (Veverka et al., 2001; Yano et al., 2006) and thermal infrared observations (Gundlach and Blum, 2013) indicate that even small asteroids are covered by a layer of centimeter-sized and smaller particles. Delbo et al. (2014) suggest that formation of fine debris on these bodies is predominantly due to thermal cycling. In support for their hypothesis, they present a series of experiments on heating–cooling cycles of cm-scale pieces of ordinary and carbonaceous chondrites and undertake theoretical modeling of thermal stress induced crack formation and propagation. They find that thermal diurnal temperature variations break up centimeter-sized rocks more rapidly than micrometeoroid impacts. According to their calculations on asteroids with period of rotation from 2.2 to 6 h and which are located at 1 AU distance from the Sun, the lifetime of 10 cm rocks should be only ~10^3 to 10^4 years. They also calculate (see their Fig. 1) that the life times of “large” rocks are shorter than those of “small” specimen, a trend that is distinctly opposite to collissional destruction. The cause of this thermal cycling trend is due to the essentially isothermal destruction of small specimen, while “large” rocks sustain a thermal gradient and thus additional stress across the entire specimen.

Delbo et al. (2014) did not explicitly formulate testable evidence which could prove their conclusions on the dominant role of thermal fatigue in destruction of rock fragments of the surface of airless bodies. From what is written in their paper one can conclude such evidence on two pieces. One is the observed presence of fine debris on the surface of small asteroids and another one is the thermal-stress fracturing and thus fractured surfaces of the rocks undergoing this process and also geologically fast destruction rate suggesting absence of rocks with the surface exposure age larger i.e. ~10^7 to 10^8 years for the asteroids located at 1 AU distance from the Sun. In the following sections we analyze these pieces of evidence, first, taking issue with the basic assumption of Delbo et al. (2014) that impact is unable to produce a fine grained, unconsolidated regolith on very small, microgravity asteroids (Section 4) and then we present observations from in situ lunar rocks as well as the returned Apollo rock collection that suggest collisional fragmentation to dominate the lifetimes of lunar surface rocks (Section 5).

4. Impact cratering and rock comminution on very small asteroids

The basic motivation for the Delbo et al. (2014) study was the perceived inability of small asteroids to retain crushed and comminuted material, as “laboratory experiments (Housen et al., 1979) and impact models (Housen et al., 2011) show that crater ejecta velocities are typically greater than several tens of cm/s, which corresponds to the gravitational escape velocity of kilometer sized asteroids”. We explore this statement further and illustrate below that impact-derived comminution products as well as crater ejecta can indeed be retained on very small asteroids.

We will distinguish in this discussion between cratering processes in an infinite half space medium and collisional disruption processes of finite sized targets, i.e., surface rocks and boulders (see e.g. Holsapple et al., 2002). Both processes are of course transitional, depending on the relative size of both target and impactor at otherwise constant impact conditions. More specifically, the specific energy imparted into the target object will determine whether a crater will only form or whether—in addition—massive penetrative cracks will be generated that lead to the catastrophic disruption of the entire object.

4.1. Cratering

The cratering process is driven by a hemispherically expanding shock wave that accelerates specific target volumes and that sets up a macroscopic cratering flow field (e.g., Maxwell, 1977; Croft, 1980; Melosh, 1989). Only the high velocity part of the displaced material will form ballistic ejecta; much of the displaced volume is contained in subsurface flows that displace and comminute materials below the crater floors and rims. Indeed the major portion of topographically elevated crater rims is due to subsurface flows and associated structural uplift of displaced and thus fractured and brecciated materials (e.g. Sharpton, 2013). Also, gravity measurements and seismic studies attest to significant comminution and bulking below the bottom of many large craters (e.g., Pohl et al., 1977) or of the basaltic substrate from which lunar mare regolith evolved (e.g., Nakamura et al., 1975), the latter involving relatively modest crater sizes, typically < 100 m.

Significant weakening and production of fracture systems are also observed below the bottom of centimeter and decimeter-sized, experimental craters as illustrated in Fig. 3 (Horz, 1969): the major cracks visible in these images, especially in Fig. 3b, refer to
fracture system paralleling the crater pro(tensile) rarefaction of the shock wave. Also note the massive tensile failure of the target’s free surfaces upon reget’s free surfaces (b) as well as minor cracks parallel to the crater radial cracks around the crater (a), and the pronounced cracks paralleling the tar-
1969; 3 g Al-sphere, 7.3 km/s; 345 kg target mass). Note the presence of major, Experimental crater formed by high-velocity impact into granodiorite (Horz, Fig. 3. et al. (1977) speci
induced trajectories with a signi
fi
cally investigated the microcracks of this crater
or, and found crack density to
oor, and found crack density to
deadly downward component.
fragment speeds may be tens if not hundreds of meters/s for such
targets suspended by break-away wires, the vast majority of the disrupted mass will be accelerated downward. This implies that the fragmentation products of natural impacts will be predominantly driven into the planetary regolith substrate; only a small fraction of the disrupted mass assumes trajectories consistent with escape from a planetary object.
4.2. Collisional fragmentation

We now turn to the collisional disruption of surface rocks and limit our remarks to strength-dominated objects that are meter-sized or smaller, thus eliminating self-gravitational healing and re-accretion (e.g. Asphaug et al., 1998; Housen and Holsapple, 1999). As stated above, Gault and Wedekind (1969) were the first to experimentally explore the collisional disruption process and they deemed a target object “destroyed” when the largest fragment remaining was < 50% of the original target mass (Mₜ); this definition was adopted by all subsequent experimental efforts on the subject (e.g., Horz et al., 1975; Horz and Cintala, 1985, Fujiwara et al., 1989, Davis and Ryan, 1990, Flynn and Durda, 2004, Cintala and Horz, 2008). There is thus a considerable body of experimental studies addressing the collisional process, all suggesting the collisional disruption process to be significantly more efficient in generating unconsolidated debris, compared to the cratering process.

Importantly, high speed movies and videos, unfortunately unpublished, of such collisional laboratory experiments indicate that most of the fragment masses have velocity vectors along the impactor’s trajectory, i.e. into the planetary surface. Although fragment speeds may be tens if not hundreds of meters/s for such
height of impact structures in low gravity environments as seen, for example, on asteroid Itokawa images taken by the Hayabusa mission (Hirata et al., 2009). Thus, while it is of course correct to postulate that materials accelerated beyond the escape velocity of a parent object will be lost from the system, it is incorrect to postulate that this applies to all ejecta. There will always be a gravity-controlled tail end in the velocity distribution of these ejecta, as long as the cohesive strength of the target is significantly smaller than the gravitational forces.

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The granite target shown in Fig. 3 may serve as an example for such a planetary surface rock. The target obviously developed major spill surfaces at its antipodal side; the latter would most likely have dislodged during this event, if the entire target block were not encased in high-strength foam inside a stout plywood box (Horz, 1969). These spill plates would have had downward trajectories into the substrate, if the target-block were sitting on some unconsolidated surface. Also note the penetrative fractures on the vertical sides of the target block: a modest, subsequent impact could readily impart just enough energy to dislodge these materials at very low velocities and not necessarily along escape
trajectories. Also, the actual crater cavity in this experiment was finely pulverized and characterized by platy fragments, some loose, yet most barely attached to the host (see Fig. 3a); many of them could be handpicked, including one of the major spall plates surrounding the cavity. It is for this reason that the crater cavity was stabilized prior to cross-sectioning by filling it with epoxy (Fig. 3b). As illustrated in Fig. 3c, the vast majority of these spall plates surrounding the cavity are only loosely attached to the target and could readily become part of a local regolith layer if dislodged by subsequent impacts.

On the basis of these observations we suggest, that collisional fragmentation, substantially driven by the tensile rarefaction wave off the target's free surfaces, is perfectly capable of providing freshly comminuted material to the surfaces of microgravity objects. Much of the disrupted mass will be driven into the surface. Also, material loosely adhering to such targets may be readily displaced by subsequent impacts along trajectories that do not necessarily lead to escape from the parent object.

Fig. 4 shows that the floor material in the central part of this crater is finely crushed but not ejected. The impact direction was horizontal so the crater was formed on a vertically oriented surface and everything that did not strongly adhere to the floor left the crater. This is a good analogy with ejection of debris from impact craters in the environment of very low gravity: so a similar process of finely crushed but not ejected. The impact direction was horizontal so the crater was formed on a vertically oriented surface and everything that did not strongly adhere to the floor left the crater. This is a good analogy with ejection of debris from impact craters in the environment of very low gravity: so a similar process of freshly comminuted material to the surfaces of microgravity objects. Much of the disrupted mass will be driven into the surface. Also, material loosely adhering to such targets may be readily displaced by subsequent impacts along trajectories that do not necessarily lead to escape from the parent object.

5. Observations of lunar rocks of known age of exposure

In the following, we present lunar surface observations as well as those from returned Apollo rocks to suggest that the survival times deduced from lunar rocks seem inconsistent with the estimates of Delbo et al. (2014) for asteroidal objects that also reside at 1 AU. Obviously care is necessary when comparing the Moon with the fast spinning asteroids of Delbo et al., as an object's rotation rate will affect the rates by which surface rocks are heated and cooled; Delbo et al. asteroids were modeled to undergo thermal cycling once every 2.2 or 6 h, as opposed to the 708 h for the Moon. Also the max–min temperature difference at the low latitudes on the Moon is up to 290 K (Vaniman et al., 1991; Vasavada et al., 2012), somewhat higher than the 190 K excursions assumed by Delbo et al. (2014) for their model asteroids. Comparison of these two objects is thus difficult as detailed trade-offs between rotation period and absolute temperature excursion, which obviously govern a rock's heating and cooling rates and thus stress levels, are not well documented to our knowledge. We nevertheless undertake such a comparison, yet suggest that it can only be qualitative at best, as our basic approach merely relates the total number of thermal cycles, otherwise uncharacterized and thus identical, on either parent. In short: we normalize the absolute exposure times available for many lunar surface rocks into absolute number of lunar thermal cycles and compare the latter with the number of thermal cycles for the model-asteroid of Delbo et al. (2014) at 1 AU.

Considering the surface exposure ages of lunar rocks (e.g. Crozaz et al., 1974; Arvidson et al., 1975; Drozd et al., 1977; Eugster et al., 1993) one should keep in mind, however, that all spallogenic noble gases ages are normalized to direct surface exposure, although they can principally be acquired while buried anywhere as deep as 2 m in the subsurface. Thus, the rocks now observed on the surface or those collected by the astronauts could have spent some time in the shallow subsurface where they were protected from significant thermal cycling. These potential burial histories further complicate our comparisons with Delbo et al. (2014) on the basis of absolute number of thermal cycles. Gault et al. (1974) calculated that the regolith is being excavated at the 99% level to a depth of 7 cm over some 100 Ma, which is compared to 10 cm by Arnold (1975), the latter more similar to Bazilevskii (1975) who considered the regolith turnover based on size-frequency distribution of lunar crater populations. So to hopefully avoid rocks that resided for substantial times in the regolith subsurface, we consider in the following rocks which are either larger than ~10 cm across or that are chipped from 0.5–1 m and larger boulders. And we also consider those rocks that only have a simple surface history, the latter based on the distribution of omnipresent microcraters on the exposed and typically subrounded, because sand-blasted, surfaces, that contrast with the generally jagged and totally uncratered fresh fracture surfaces that were buried in the regolith (e.g. Sutton, 1981).

Tables 1 and 2 show lists of such samples found in the literature with excellent resource at http://curator.jsc.nasa.gov/lunar/lsc/index.cfm and Fig. 5 provides a visual view of the distribution of
their surface exposure ages measured by the spallogenic noble gases technique. Following Horz et al. (1975a) and Drozd et al. (1974, 1977) we selected age determination results preferring those taken by the $^{38}$Ar technique over the results taken by $^{38}$Ar method and then by $^{21}$Ne ones, because the first one largely eliminates sample shielding uncertainties and because the heavier noble gases are less vulnerable to the diffusion gas loss, and are among the available results preferring more recent ones over the older ones.

Fig. 5 It is seen from the data presented in the Tables 1 and 2 and Fig. 5 that the surface exposure ages of the lunar samples vary from 2 Ma (age of South Ray Crater) to tens and even several hundred of millions of years. Since the samples were biased towards those that most likely were indeed continuously exposed on the surface, most of them seem lucky survivors of the meteorite impact environment. They are, thus, promising objects to further search in their morphology for evidence of thermal fatigue. In the

### Table 1

<table>
<thead>
<tr>
<th>No</th>
<th>Site</th>
<th>Sample no</th>
<th>Lithology</th>
<th>Sample mass, g</th>
<th>Exposure age, Ma</th>
<th>Method</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>A16</td>
<td>60,015</td>
<td>Cataclastic anorthosite</td>
<td>5574</td>
<td>1.9 ± 0.1</td>
<td>Kr-Kr</td>
<td>Leich and Niemeyer (1975)</td>
</tr>
<tr>
<td>2</td>
<td>A16</td>
<td>60,016</td>
<td>Regolith breccia</td>
<td>4307</td>
<td>2.2 ± 0.6</td>
<td>Kr-Kr</td>
<td>Eugster (1999)</td>
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<tr>
<td>3</td>
<td>A16</td>
<td>60,025</td>
<td>Regolith breccia</td>
<td>1836</td>
<td>1.9 ± 0</td>
<td>Kr-Kr</td>
<td>K.Marti in Drozd et al. (1977)</td>
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<tr>
<td>4</td>
<td>A16</td>
<td>61,015</td>
<td>Dimict breccia</td>
<td>1803</td>
<td>2.4 ± 0.5</td>
<td>Ar</td>
<td>Eugster (1999)</td>
</tr>
<tr>
<td>5</td>
<td>A16</td>
<td>61,016</td>
<td>Impact melt rock</td>
<td>11729</td>
<td>1.84 ± 0.4</td>
<td>Ar</td>
<td>Niemeyer (1977)</td>
</tr>
<tr>
<td>6</td>
<td>A14</td>
<td>14,321</td>
<td>Crystalline matrix breccia</td>
<td>8996</td>
<td>23.8 ± 0</td>
<td>Kr-Kr</td>
<td>Lugmair and Marti (1972)</td>
</tr>
<tr>
<td>7</td>
<td>A14</td>
<td>14,305</td>
<td>Crystalline matrix breccia</td>
<td>2498</td>
<td>27.5 ± 1.5</td>
<td>Kr-Kr</td>
<td>Lugmair et al. (1984)</td>
</tr>
<tr>
<td>8</td>
<td>A16</td>
<td>67,016</td>
<td>Feldspathic breccia</td>
<td>4262</td>
<td>50 ± 0</td>
<td>Ar</td>
<td>Turner and Cadogan (1975)</td>
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<tr>
<td>9</td>
<td>A16</td>
<td>67,915</td>
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<td>2559</td>
<td>50.6 ± 1.5</td>
<td>Kr-Kr</td>
<td>Peepin et al. (1974)</td>
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<td>10</td>
<td>A15</td>
<td>15,515</td>
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<td>81</td>
<td>Kr-Kr</td>
<td>Marti and Lightner (1972)</td>
</tr>
<tr>
<td>11</td>
<td>A17</td>
<td>70,215</td>
<td>Mare basalt</td>
<td>8110</td>
<td>126 ± 3</td>
<td>Kr-Kr</td>
<td>Drozd et al. (1977)</td>
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<tr>
<td>12</td>
<td>A12</td>
<td>12,063</td>
<td>Olivine basalt</td>
<td>2426</td>
<td>140 ± 40</td>
<td>Kr-Kr</td>
<td>Burnett et al. (1972)</td>
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<tr>
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<td>2109</td>
<td>200</td>
<td>Ar</td>
<td>Hintenberger et al. (1971)</td>
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<td>70,017</td>
<td>Ilmenite basalt</td>
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<td>220 ± 20</td>
<td>Ar</td>
<td>Phinney et al. (1975)</td>
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<td>Olivine basalt</td>
<td>1806</td>
<td>230 ± 40</td>
<td>Kr-Kr</td>
<td>Burnett et al. (1972)</td>
</tr>
<tr>
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<td>A14</td>
<td>14,310</td>
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<td>~ 490</td>
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<td>Kr-Kr</td>
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### Table 2

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<th>No</th>
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<th>Exposure age, Ma</th>
<th>Method</th>
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<td>346</td>
<td>810 ± 60</td>
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following, we first consider detailed laboratory and lunar surface images of two examples for samples larger than \( \sim 10 \) cm, followed by similar imagery of boulder chips.

We first consider sample 65015 (Fig. 6) composed of poikilitic impact melt breccia which was collected from the lower slope of Stone Mountain at the Apollo 16 station 5. It is a very coherent, dense rock with little void space. It was about half buried in the soil, as judged by the obvious “soil line”, below which there are no micrometeorite pits (http://curator.jsc.nasa.gov/lunar/lsc/65015.pdf).

Surface exposure ages of this sample were measured by the \(^{38}\)Ar method by Kirsten et al. (1973) who found that it is \(365 \pm 20\) m.y. and by Jessberger et al. (1974) who found it to be \( \sim 490 \) Ma. The uncratered surface demonstrates a simple surface exposure history of approximately 400 Ma with the rock half buried in the soil. The number of thermal lunar cycles experienced by the exposed surface was \( \sim 5 \times 10^9 \) which is about 2–3 orders of magnitude larger than predictions of Delbo et al. (2014) to destroy a 10-cm rock by thermal fatigue. Meanwhile, no significant fracturing is seen on the surface exposed to space. Indeed, the presence of a brownish patina argues for small scale depositional processes of finely dispersed impact melts and vapors (McKay et al., 1991), and seems contrary to an omnipresent erosion process due to thermal fatigue. Every white dot in the above image—indicating removal of this patina—is caused by a microcrater and represents the very fine-grained comminution debris of such events.

The second example concerns sample 15555, composed of olivine normative basalt. It was found on the mare surface at Apollo 15 station 9 A, some 12 m from the edge of Hadley Rille and \( \sim 1800\) m west of the Lunar Module. Fig. 7 shows 4 sides of this sample.

Fig. 7 shows that all surfaces of the rock are covered with brownish soil, a common occurrence as it was impractical for the astronauts to pick up any sample without also depositing some loose soil material into the sample bags; this soil dusting is distinctly different from the patina described above, and can potentially be removed via vigorous gas-blast, yet the patina can not. Microcraters are more difficult to discern on such dusty surfaces, yet the center of South side contains a relatively large one, a few millimeters in diameter. The East and North sides display a long, penetrative fracture, the major reason for this illustration. Part of the top of the rock's N-face has a more rugged appearance than the rest of this surface and seems to represent the floor of an impact event in approximate cross section; the latter dislodged all of the more subdued and rounded “local” surface and produced a the major crack as well as a number of smaller ones. Examples of radial crack systems associated with impact events could be given, such as (undated) rock 73155 (Fig. 1 of Horz et al., 1975a). It can also be noted that neither the fracture in Fig. 7 nor many others observed on lunar rocks has any relation to the shape of the rock; while we would expect thermally induced stresses to be related to illumination geometry and thus to a rock's shape. Marti and Lightner (1972) determined the cosmic ray exposure age as 81 m.y.
by $^{81}$Kr method. Podosek et al. (1971) and York et al. (1972) determined exposure ages of 90 m.y. and 76 m.y., respectively, by the $^{38}$Ar technique. This age is not associated with any local crater (Arvidson et al., 1975). The impact that formed millimeter-sized microcrater on the South side would have caused the rock to jump or roll (http://curator.jsc.nasa.gov/lunar/lsc/15555r.pdf). In the given images the surface morphology looks more or less similar except that soil covers part of the West side. But that is rather small so it seems logical to conclude that for $\sim$80 m.y. the rock 15555 was not even partly buried but was present on the lunar surface probably changing its orientation but certainly undergoing long time for the day/night thermal variations. For this time period there happened $\sim 10^5$ diurnal thermal cycles that are about 2 orders of magnitude larger than the estimates of Delbo et al. (2014) needed to destroy the 10-cm rock by thermal fatigue, but no evidence of thermal stress destruction is seen in these images.

Now we consider in-situ images of lunar rocks as observed and photographed by the Apollo astronauts (Swann et al., 1971, 1972; Muehlberger et al., 1972, 1973). These illustrate examples of two types of rock fragments: (1) rocks $\sim$20 cm across, close to the sizes considered by Delbo et al. (2014) ("larger than a few centimeters"); thus, if the survival time estimates of these authors are correct, one may expect that when these values are reached, these rocks would be destroyed; (2) rock boulders of 3–5 m across; for these, one may expect that when the survival time estimated by Delbo et al. for the several-centimeter-size rocks is reached, the several-centimeter-thick surface layer on these larger blocks should be destroyed and form a fillet at the boulder base. Fig. 8 shows three examples of the rocks of the first type and in Fig. 9 three examples of the second type.

Parts a and d of Fig. 8 shows the in-situ and in-lab views of the “Big Bertha” rock. This rock was collected by Astronauts Shephard and Mitchell as a “football-size” sample 14,321 and consists of a crystalline matrix breccia. Its exposure age was determined as $\sim$24 Ma (Burnett et al., 1972; Lugmair and Marti, 1972; http://curator.jsc.nasa.gov/lunar/lsc/14321.pdf) and this agrees with the exposure ages of other samples taken on the rim of Cone crater whose age of formation is $\sim$26 Ma (Arvidson et al., 1975). In the in-situ image the rock looks somewhat knobby which is typical for breccias, as breccia clasts and matrix will erode differentially during micrometeoroid impact. The lab image shows details of the surface of the top side of the rock. The latter looks rather monolithic with almost no fracturing. Numerous microcraters 1–3 mm in diameter are seen here. They were studied and other rock samples were taken at Apollo 12 and 14 sites by Morrison et al. (1972).

Fig. 8b and e, shows the feldspatic fragmental breccia 67016, collected by Astronauts Young and Duke. Its exposure age was determined to be 50 Ma (Turner and Cadogen, 1975; http://curator.jsc.nasa.gov/lunar/lsc/67016.pdf), and also agrees with the exposure ages of other samples taken on the rim of North Ray Crater (Arvidson et al., 1975). In the in-situ image the rock looks very cohesive, with sharp edges. In the in-lab image the surface of its top part looks monolithic with almost no fractures and with numerous superposed microcraters.

Fig. 8c and f, shows ilmenite basalt 74275, chipped off the rock top of a much larger rock (arrow) by Astronauts Cernan and Schmitt. Its age of exposure on the lunar surface was determined as $\sim$30 Ma (Eugster et al., 1977; http://curator.jsc.nasa.gov/lunar/lsc/74275.pdf) and this agrees with the exposure ages of other samples taken on the rim of Shorty Crater (Arvidson et al., 1974, 1975). In the in-situ image, each of the four parts of the rock appear rather cohesive, with no fine-scale fracturing observed.

Fig. 8. Rock fragments on the rims of dated lunar craters: (a) the $\sim$20-cm rock Big Bertha on the rim of Cone Crater, Apollo 14 station C1, image AS14-64-9128; b) the $\sim$20 cm rock on the rim of North Ray Crater, Apollo 16 station 11, AS16-116-18658; c) the $\sim$20-cm rock on the rim of Shorty Crater (arrow), Apollo 17 station 4, image AS17-137-20990; (d, e, and f) – detailed in-lab images of the top surfaces of these rocks after they were returned to the Lunar Receiving Laboratory (images S-71-29183, S72-39231, S73-16018). The arrows point to a few of the numerous microcraters seen on these surfaces.
the lab image, sample 74,275 appears generally monolithic, but some fractures are seen, and microcraters are observed as well. Fig. 9 shows three rocks which are informative concerning the presence or absence of boulder-associated fillets. Fig. 9a shows Contact Rock. It was not sampled by the Apollo 14 Astronauts, but its position on the rim of Cone Crater among other relatively large boulders suggests that it is part of the ejecta from this crater, so its exposure age should be $\sim 26$ Ma (Arvidson et al., 1975). The distinctive albedo pattern on the camera-looking side of this boulder suggests that it is breccia, while its sharp edges indicate that it is a rather strong rock, not noticeably eroded by any process since its emplacement 26 Ma ago. Delbo et al. would predict a lifetime of some $10^{3}$–$10^{4}$ years for this meter-sized boulder, much shorter than the measured exposure age. Also, considerable fine-grained material produced by thermal cycling should have accumulated in a filament-like fashion around this rock over a period of 26 my, Ma which is not observed. No fracturing of this rock is seen on the camera-looking side of the rock.

Fig. 9b shows part of the 5-m long “Outhouse Rock” at the Apollo 16 site near the rim of North Ray Crater. Several samples including 67955 were chipped from this boulder by Astronauts Young and Duke. The boulder is composed of breccia, and sample 67955 represents a large clast of noritic anorthosite. Its exposure age is 50 Ma, and this agrees with exposure ages of other samples taken at the rim of the North Ray crater Drodz et al. (1974, 1977), http://curator.jsc.nasa.gov/lunar/lsc/67955b.pdf. Again, no noticeable filament accumulation is observed at the foot of Outhouse Rock.

Fig. 9c shows a group of 1–2 m size rocks on the southern rim of Camelot crater at the Apollo 17 site. These rocks were not sampled by Astronauts Cernan and Schmitt but sample 75035 was chipped off of a neighboring boulder. It is an ilmenite basalt and the exposure age determined for this sample is $\sim 90$ Ma (Turner and Cadogan, 1974; Crozaz et al. 1974; Arvidson et al., 1976; Bhandari, 1977; http://curator.jsc.nasa.gov/lunar/lsc/75035.pdf). The rock surfaces here look pitted and substantially rounded, consistent with a much longer surface exposure compared to the boulder in Fig. 9a. The absence of penetrating fractures in these boulders even after residing for 90 Ma on the lunar surface can be noted.

In summary, the in-situ and in-lab images of the $\sim 20$ cm rocks described here show no fracturing (samples 14,321 and 67,016) or minor fracturing (sample 74,275). Their exposure ages on the lunar surface are from 24 to 50 Ma; this corresponds to $\sim 3$–$6 \times 10^{8}$ lunar thermal cycles. The Delbo et al. (2014) estimates of the lifetime of such rocks are $10^{3}$–$10^{4}$ years (their Fig. 1), corresponding to $\sim 3.5 \times 10^{6}$ to $1.5 \times 10^{7}$ thermal cycles in their model. The time during which these 20 cm rocks spent on the lunar surface expressed in the number of thermal cycles is 1.5–2 orders of magnitude larger than the Delbo et al. (2014) lifetime estimates. So if the latter would be applicable to the 20 cm rocks analyzed, then these rocks should be destroyed many times over. Instead, they show almost no, or very minor, fracturing.

In all the three cases, no filament is seen at the foot of the rock. These rocks spent from $\sim 26$ to $\sim 90$ Ma on the lunar surface, corresponding to $\sim 3$ to $7.5 \times 10^{8}$ lunar thermal cycles. Following Fig. 1 of Delbo et al. (2014) and extrapolating to meter-sized boulders, the thermal-cycling lifetimes of the latter seem to be $< 10^{4}$ years, shorter than a 10 cm diameter rock; normalized to thermal cycles, the discrepancy between Delbo et al.’s chondritic asteroid at 1 AU and lunar observations exceeds 2–3 orders of magnitudes. This is again 1.5–2 orders of magnitude larger than the Delbo et al. (2014) estimates of the lifetime of the 10 cm rocks, and if the latter were applicable to the decimeter-scale surface layer of the rocks analyzed, the latter should be severely fractured, and fillets should have formed at their base. We also suggest that the fine-grained material removed by thermal cycling from a meter-sized parent boulder exposed for tens of millions of years should result in the universal accumulation of fillet-deposits around these boulders, which is not observed in our examples above.

However, in some localities such fillets are indeed observed around other lunar surface boulders, as illustrated in Fig. 10. Fig. 10a shows an irregularly-shaped, meter-sized boulder (seen behind the gnomon) with a well developed filament. This boulder was not sampled so its exposure age was not measured. But samples 15535 and 15536 were chipped off the smaller (0.75 m) gnomon-boulder by astronauts Scott and Irwin. These were found to be composed of olivine normative basalt and their ages of exposure were 110 Ma (Alexander et al., 1972 and Arvidson et al., 1975; http://curator.jsc.nasa.gov/lunar/lsc/1535.pdf). In close vicinity to this rock, sample 15555, a $30 \times 15$ cm$^{2}$ rock also composed of olivine normative basalt, was collected; its exposure age is $\sim 80$ Ma (Marti and Lightner, 1972; Podosek et al., 1972; York et al., 1972; http://curator.jsc.nasa.gov/lunar/lsc/1555.pdf). This filament (and others) can, however, not unambiguously be related to erosional debris of this local boulder, as the boulder may have been partially buried upon its emplacement as an ejecta block and/or regolith fines ejected from nearby craters may have accumulated at the boulder’s base.

The elongated field of meter-sized and smaller boulders marks the rim of the sinuous rille, Rima Hadley, and this field was interpreted to have been formed due to asymmetric impact
regarding: The near-rim impacts distributed material in all directions including into the rille but the narrow zone along the rille receives material only from the east because impacts that occur within the rille to the west do not eject enough material up to the rim (Swann et al., 1972). A similar phenomenon was described at the rim of Fossa Recta graben in the Lunokhod 2 study area (Fig. 10b; Basilevsky et al., 1977).

In summary, the meter-sized rock shown in Fig. 10a is probably one of the “lucky” long survivors, while the two smaller rocks are more recently formed fragments of “unlucky” boulders. This suggests that the larger boulder resided on the surface for a time longer than 110 Ma during which it underwent destruction of its surface layer and formation of a fillet at its base. One can surmise that during this destruction of the surface layer, both impacts by small meteorites and diurnal thermal stresses were involved. But even if the role of thermal stresses was not negligible, it produced a “visible” result only after the rock resided on the surface longer (probably much longer) than 110 Ma, that is ~10^9 thermal cycles. Fig. 10c shows the so-called Filleted Rock from the rim of Cone Crater at Apollo 14 site. This rock is about 1 m high and 2 m wide and it has prominent fillets as far as 3 m from the rock base; these characteristics and the rock name were given by astronauts Shepard and Mitchell (Swann et al., 1971). The boulder itself is composed of friable-appearing coarse breccia. Sample 14053 is an Al-rich mare basalt found perched on the side of the boulder. It is flat with one side freshly broken and the other side rounded and pitted by microcraters. Breccia material was found attached to the flat side, indicating that this basalt was a clast in the breccia boulder. Its exposure age was found to be 21 ± 5 m.y. (Husain et al., 1972; Stettler et al., 1973; Eugster et al., 1984; http://curator.jsc.nasa.gov/lunar/lsc/14053.pdf), and this agrees with the ~26 Ma age of Cone Crater (Arvidson et al., 1975). Note that this 1 m high boulder and its prominent, 3 m wide fillet contrasts dramatically with boulders of similar size and in the same area (e.g., Contact Rock, see Fig. 6a) that have no fillets. The distinctly jagged nature of the Contact Rock also contrasts with a much more rounded and subdued surfaces of Fillet Rock, suggesting dramatically different rheological properties for these boulders, which obviously affect any small scale erosion process. Thus part of the fillet in Fig. 10c could indeed be erosional debris, but the fillet seems too voluminous to be exclusively derived from the parent boulder, the reason why other contributors can not be excluded (such as partial burial upon boulder-emplacement and blocking of ejecta from nearby regolith craters).

In summary, the 20 cm sized rocks described here experienced 1.5–2 orders of magnitude, more thermal cycles than the number of thermal cycles which Delbo et al. (2014) interpreted should lead to their complete destruction; yet these rocks show no or little traces of thermal-stress destruction (Fig. 8). The meter-sized rock boulders described here experienced 1.5–2 orders of magnitude, more thermal cycles than the number of thermal cycles that are predicted to lead to complete destruction of rocks of about a decimeter in diameter, according to Delbo et al. (2014); however, these rocks show no accumulation of fillets at their base (Fig. 9). In the case of an even longer time period (≥100 Ma), and a correspondingly larger number of thermal cycles (Fig. 9a), some fillets did form, probably with involvement of thermal cycling. But according to the estimates of Basilevsky et al. (2013), at this time stage about 80–90% of the initial rock population should already have been destroyed by meteorite impacts (see Fig. 2). This indicates the relative effectiveness of the processes of meteorite impacts and thermal cycling in rock destruction.

If one accepts the logic of these analyses, then this means that the Delbo et al. (2014) estimates of rock survival times are not applicable, at least, to the case of lunar rocks. This might be due to a number of reasons.

1. The major reason may be the difference in the dynamics of the heating-and cooling rates as well as absolute temperature excursions on the lunar surface and on the model asteroids of Delbo et al. (2014) at 1 AU; our normalization of exposure ages to absolute number of thermal cycles may be a gross over-simplification. More detailed modeling of the thermo-elastic stresses produced on the Moon is needed.

2. The physical properties of planetary surface rocks can differ widely and also their susceptibility to small scale erosional processes, whatever be their nature. The experimental results of Delbo et al. (2014) on carbonaceous (CM2; Murchison) and ordinary (LL3.2; Sahara 97210) chondrites may not apply to lunar rocks, typically igneous basalts or impact breccias that range from competent, crystalline melt specimen to rather friable fragmental materials.

3. There is also an uncertainty whether the definition of “lifetime” is exactly equivalent between the thermal cycling process of Delbo et al. (2014) and the collisional destruction of rocks by meteorite impact (e.g. Gault and Wedekind (1969)). Delbo et al. considered a rock destroyed when “an initial 30-mm-long crack grows to a length equal to the rock diameter”. This definition is consistent with Gault and Wedekind’s suggestion, but it is not sufficient, as it does not necessarily lead to the physical disintegration of the rock. Many natural rocks are pervasively fractured, yet remain surprisingly competent and mandate for substantial additional energies to be physically disrupted in such a manner that the mass of the largest fragment remaining
is less than half of the initial rock mass. This, however, seems a minor effect considering the current order of magnitude differences in estimating the surface-lifetimes of lunar surface rocks.

In summary, current lunar observations and associated modeling of the collisional destruction process seem to suggest substantially longer surface residence times of lunar surface rocks than one would predict from thermal cycling and the liberal extrapolation of model-asteroids by Delbo et al. (2014). This is especially the case for meter sized boulders that may reside on the lunar surface for tens if not hundreds of millions years (Basilevsky et al., 2013), thus suggesting that thermal cycling may not be as prominent as suggested by Delbo et al. (2014).

In any case, our analysis of the morphologies of the lunar rocks of known exposure age showed that the role of meteorite impacts in rock destruction is dominant, at least at 1 AU distance from the Sun and at larger distances, while that of thermal cycling is secondary. We think that this situation is very certain for the destruction of rocks on the lunar surface and may be true for other airless bodies, although this latter conclusion requires further analyses.

6. Calculations of meteorite flux impact velocities and survival times of meter-sized boulders on different airless bodies

Here we calculate the meteor intersection rates and intersection velocities for a number of airless bodies to use them for estimates of survival times of rock boulders based on (1) estimates for the lunar boulders (see above Section 2) described above and (2) assuming that the thermal-stress destruction plays a minor role (see Section 5).

There are two aspects of the flux of meteors that intersect the solar system bodies: the intersection rate, and the average velocity of the intersections. To work out predictions of the rate and velocity of potential meteor intersections on inner solar system target bodies we utilize the entire Minor Planet Center “MPC Orbit (MPCORB)” database of ~400,000 inner solar system minor planet orbits out to the orbit of Jupiter (Minor Planet Center, 2015). In the absence of any known predicted intersections between members of the MPCORB database and potential target bodies in the present-day, we select minor planets orbits from the MPCORB database that cross target body orbits as proxies for the flux of unobservable smaller projectiles that are responsible for boulder degradation and destruction. Solar system fragments ≥300 μm are unaffected by solar photon flux and the solar wind, and therefore fragments that are ≥300 μm travel in normal gravitationally controlled Keplerian orbits (Dobrovolskis and Burns, 1980; Juhász et al., 1993; Hamilton and Krivov, 1996; Krivov et al., 1996). Consequently, the population of small unobserved projectile orbits that cross target body orbits should be proportional to the observed orbits of the MPCORB database of minor planets that cross the same target bodies, and the MPCORB database may therefore be used as a proxy to work out the flux of unobserved projectiles that degrade and destroy boulders on airless bodies.

There are two components for our model: The quantity of MPCORB orbits that cross the target orbit that we use to work out the intersection rate, and averaged populations of MPCORB orbits that are used to compute the average intersection velocity of target body impacts. If a MPCORB database orbit does not have a sufficient solar altitude (or excessive solar altitude) such that a proxy orbit does not cross the target body orbit at any solar altitude, we omit it from the population of potential MPCORB proxy orbits for this target body. Further, we assume that orbits with eccentricities that are ≥0.9 occupy unstable orbits that are likely to be removed from the solar system prior a target intersection, and our model omits these from the count of intersecting proxy orbits and from the average of intersecting velocities.

Intersection rate: To work out the rate of intersections, we count the number of MPCORB orbits that cross the target body orbit. This results in a relative rate in intersections. For example, if 1,000 MPCORB orbits cross the orbit of Target Body #1 and if 2000 MPCORB orbits cross the orbit of Target Body #2, we compute that Target Body #2 has an intersection rate that is 2 times greater than Target Body #1. In addition to factors that remove a proxy orbit from the count (no crossing orbit or an eccentricity > 0.9), the count is modified by factors that slightly increase or decrease the likelihood of a future target intersection. This includes the planetary gravitation focus that increases the effective target area in the vicinity of major planet and the largest minor planets, and also the effect of a moon’s orbital velocity which affects the length of time when intersections are possible.

Average intersection velocity: We assume that a population of unobserved projectiles populates orbits that intersect the target body from orbits that are the same in character as the MPCORB database of orbits, yet are aligned and timed in such a way that intersections take place. To work out the intersection velocity, an average of MPCORB orbital elements that cross the orbit of the target body are rotated into position. In order to account for the variety of orbital circumstances under which projectiles may intersect the target body, we work out 12 intersection scenarios and average of the 12 predictions (Fig. 11).
Comparison to published models: Our predictions for the intersection rates and the average intersection velocities of asteroidal projectiles for the Moon and in the vicinity of Mars are comparable with Ivanov (2001); Ivanov et al. (2002) and Feuvre and Wieczorek (2011), including predictions of the difference in the SFD of the lunar leading and trailing orbital apaxes (Le Feuvre and Wieczorek, 2011). Also, our model of intersection velocities of Ceres and Vesta are comparable with the predictions of Farinella and Davis (1992) for similarly located main Belt minor planets. Consequently, based on how our model is robust in reproducing comparable predictions of the intersection rates and the average intersection velocities of well-studied target bodies, and how projectiles that are \( \geq 300 \mu \text{m} \) travel around the Sun in Keplerian orbits, we are confident that our model is applicable for predicting the asteroidal meteor intersection rates and the average intersection velocities of asteroidal meteor impacts that degrade and destroy boulders on all airless target bodies of the inner solar system.

Lunar intersections as a standard rate reference: To relate the count of potential proxy intersections to a known standard of impact exposure that degrades and destroys boulders on airless bodies, we predict the interaction rate for the Earth's Moon where the Moon offers an accessible global impact record that has been studied extensively and a record of boulder destruction that is calibrated in years (Basilevsky et al., 2013), the predicted value of the intersection rate at a target body compared to the Moon may be used to calculate the rate of boulder destruction on any target body of the inner solar system in years. Our count of the MPCORB orbit intersection rate at the Moon works out to a lunar rating of “1” which corresponds to 672 MPCORB orbits that cross the orbit of the Moon. As an example of how we apply this rating, 2607 MPCORB orbits cross the orbit of Mars in the present-day, and therefore an intersection rate at Mars is \( \sim 3.88 \) greater than for the Moon (2707/672).

Applying the Moon as a standard intersection rate reference presents a challenge in three ways: (1) the Earth's gravitational focus increases the rate of intersections in the vicinity of the Moon, (2) the Moon's own gravitational focus increases the rate of intersections, and (3) the orbit of the Moon around the Earth is alternately sweeping a larger or a smaller volume of space that is added or subtracted from intersection rates depending on the time of the lunar month. In view of how the geological effects on lunar boulders are produced by impacts that are modified by the effects of gravitational focus and the cycle of the lunar orbit, the local gravitational environment of the Moon and the lunar orbital motion are included in our universal lunar standard intersection rate.

Intersection velocity modeling: To work out an average intersection velocity of MPCORB proxy orbits that intersect targets we compute scenarios that predict the intersection velocities of proxy projectiles at the target body perihelion, aphelion, and a distance from the Sun that is equal to the semimajor axis of the target body (Fig. 11a). We test the three orbital locations using two sub-populations of proxy orbits. The two sub-populations of proxy orbits are separated by semimajor axis. One group has a semimajor axis that is greater than the target semimajor axis and the other group has a semimajor axis that is less than the target body semimajor axis (Fig. 11b). Once we adjust the test projectile orbits to align them with the target orbit (by rotating the test orbit ascending nodes and augments of periapsis), we then rotate the ascending nodes by 180° in order to simulate the two possible orientations of intersection that take place, either on the rising or descending node of the orbit (Fig. 11c). The three orbital positions, the two populations of projectiles, and the two inclination models produce a total of 12 scenario predictions that are then averaged according to weighting that takes into account the number of projectiles in each scenario and the tendency of target bodies to occupy a greater portion of their orbital periods near aphelion.

Applying predicted velocities to boulder destruction rates: Using intersection velocity modeling, we predict the average velocity of proxy intersections at a target body. The predicted average velocities are used to compare the impact energy at the target body with the impact energy on the Moon. For example, if we predict an average meteor impact velocity on a target body that is 1/2 of the typical lunar meteor impact velocity, this means that 1/4 of the kinetic energy is being delivered to the target for a given projectile size compared to the Moon (\( E_k = 1/2 \, m \, v^2 \)). Therefore four times as many same-sized projectiles would be required to deliver the same extent of boulder destruction impact energy. If, however, the intersection rate at the target is four times greater than at the Moon, the extent of boulder destruction impact energy at the target body and on the Moon will be essentially the same because each impacts four times delivers four times less energy (\( 4 \times 1/4 \)).

A simplified model: Somewhat surprisingly, after extensive modeling that averages 12 testing scenarios, it turns out that every one of our velocity predictions to date can be approximated within 10% by the average of a simple zero-inclination circular test orbit that is applied at the target perihelion, aphelion, and at a distance from the Sun that is equal to the semimajor axis of the target body (Fig. 11a). Therefore, in view of the uncertainty of the population of potential meteor projectiles that intersect the target bodies, the simplified method of averaged circular meteor orbits may be sufficient to assess meteor intersection velocities to a first order. In this report, tabulated predictions that include target body intersection velocities at perihelion, aphelion, and the semimajor axis altitude are based on the 12-scenario method. Single predictions without orbital details are based on the simplified circular orbit method.

Conclusion: Based on the results of this numerical modeling, it was possible to estimate the meteorite fluxes and intersection velocities for a number of airless bodies, including Phobos, Deimos, asteroids Itokawa, Eros, Vesta, Ceres, and an average of the first 150 discovered Jupiter Trojans (http://www.minorplas...
and deduce the survival times of meter-sized rock boulders on their surfaces of (Table 3).

7. Discussion

In this work we use as the opening point our earlier estimates of the survival times of the meter-sized lunar rocks: For the time of a few millions of years only a small fraction of the initial population is destroyed. For the time of several tens of million years \( \sim 50\% \) is destroyed, and for times of \( 200–300 \text{ Ma} \sim 90–99\% \) of the original boulder population is destroyed (Basilevsky et al., 2013). Considering morphologies of the decimeter- and meter-sized lunar rocks with known age of their surface exposure (Swann et al., 1971, 1972; Muelhberger et al., 1972, 1973 and http://curator.jsc.nasa.gov/lunar/lsc/) we show that thermal-stress destruction due to diurnal temperature variations plays a secondary role, at least at 1 AU and at further distance from the Sun, and the major factor in rock destruction is catastrophic rupture by meteorite impacts. This is a solid conclusion for lunar rocks and may be essentially true for the rock destruction on other bodies considered.

Numerical modeling of orbital parameters was applied to estimate the meteorite flux and impact velocities for the satellites of Mars, Phobos and Deimos, asteroids Itokawa, Eros, Vesta, Ceres, and an average of the first 150 discovered Trojans. From these estimates, it was possible to deduce the survival times of meter-sized rock boulders on the surface of these bodies, compared to survival times on the lunar surface estimated earlier (see above Table 3). The calculated survival times can be used for analysis of histories and intensities of surface processes on these bodies. However, as it will be shown below, sometimes this is not a straightforward way.

It was found that survival times of meter-sized boulders at Phobos are, on average, slightly shorter compared to estimates for lunar rocks, with a significant (by a factor of 2) difference between the leading (noticeably shorter than on the Moon) and trailing (similar to that on the Moon) hemispheres. Survival times of such boulders at Deimos were found to be on average also shorter compared to the Moon and also with a significant (by a factor slightly smaller than 2) difference between the leading hemisphere (shorter than on the Moon) and the trailing (almost similar to that on the Moon). At first glance this difference in effectiveness of rock destruction on the leading and trailing hemispheres of Phobos and Deimos suggests that there may be differences in regolith maturity, including the presence and abundance of rock fragments on the leading and trailing hemispheres of these satellites. However, at the scale of these small and low-gravity bodies, even the low velocity part of the crater ejecta travels large distances and should probably homogenize the Phobos and Deimos regolith on the global scale, and may balance and/or mask the suggested differences.

The similarity of estimated average survival times of meter-sized boulders on Phobos and Deimos to those of the Moon suggests that the majority of meter-sized rock boulders formed in some event should be disrupted over a period of less than a few hundred of million years. However, typically observed relative rarity of meter-sized rock boulders associated with impact craters on the surface of Phobos and Deimos does not necessarily mean that all the craters considered are old. Rock fragments on the rims of fresh lunar craters are part of the crater ejecta deposited close to the crater. Crater ejecta on Phobos and Deimos should travel far, and even leave these bodies due to their very low gravity (Wilson and Head, 2015). And another factor is that thickness of regolith on these bodies may be much larger than that on the Moon (e.g. Basilevsky et al., 2014a,b, and references therein).

For Itokawa the average survival times were found also to be close to those of the Moon. But this is certainly a special case: on Itokawa there are areas with very abundant rock boulders and also are present smooth areas composed of relatively fine (centimeters) material (Fujiiwara et al., 2006). Analysis of observations and theoretical considerations show that the boulders on Itokawa destroyed by meteorite impacts may be replaced by newly-appearing boulders, for example, from the subsurface due to the Brazil Nut Effect (Sanchez and Scheeres, 2009) while the newly-formed fine debris may drain into gravitational lows represented by smooth areas (Miyamoto et al., 2007). So the high spatial density of boulders on Itokawa cannot be used for the terrain age estimates through direct comparisons with rock densities on the rims of lunar craters.

For Eros, survival times of meter-size boulders were found to be \( \sim 1/3 \) lunar values. This is because the meteorite flux on the surface of Eros is higher than on the Moon by a factor of 13 and impact velocity is about twice lower (compared to the Moon), only partly balancing the effect of the higher flux. So impact resurfacing on Eros is expected to be more effective than on the Moon: more intensive cratering, thicker regolith and faster destruction of rock boulders. However, the NEAR Shoemaker observations showed a different picture: the surface of Eros is characterized by a typical deficit of small (\(< 200 \text{ m in diameter} \) craters and a relatively high abundance of meter-sized rock boulders (Chapman et al., 2002). These authors suggest that this may be partly explained by a major depletion of meter-scale projectiles in the asteroid belt due to the Yarkovsky effect (Bell, 2001), so fewer number of small craters should be formed and boulders excavated in larger events will be destroyed less effectively. Additionally, the small size and low gravity of Eros may result in redistribution or loss of ejecta due to seismic shaking, thus preferentially destroying smaller craters formed in this regolith.

For asteroids Vesta and Ceres the survival times of meter-size boulders were found to be \( \sim 1/30 \) of lunar values. This is because of the very high meteorite flux (a factor of \( \sim 300 \) times that of the Moon); this high flux is only slightly balanced by the lower meteorite impact velocities (a factor of \( \sim 3 \) compared to the Moon). The available images of Vesta have insufficient resolution to see such boulders but predictions can be made. The Dawn mission showed that the surface of Vesta is heavily cratered, and thus is very old (e.g., Jaumann et al., 2012; Marchi et al., 2012). The shorter (compared to the Moon) survival times of the boulders are considered means a higher rate of destruction and implies a more intensive impact resurfacing that should lead to faster accumulation of regolith and its generally greater thickness. Thus, small craters with rocky ejecta on Vesta should be more rare than on the Moon. Intensive impact resurfacing should lead to a greater maturity of its regolith material.

However, the larger distance from the Sun implies weaker solar wind and the lower impact velocities that should make the regolith maturation on Vesta different from that typical for lunar regolith. This agrees with observations taken by the Dawn Visible and Infrared Spectrometer and Framing Camera: (1) no evidence was observed for accumulation of nanophase iron on regolith particles, (2) locally homogenized upper regolith generated through small-scale mixing of diverse surface components (Pieters et al., 2012) and (3) the presence of exogenous (meteoritic) hydrated carbonaceous material (McCord et al., 2012).

For Ceres, whose surface is also very old (http://www.nasa.gov/jpl/dawn/pia18920/) the predictions on the boulder/small crater relations as well as on regolith thickness appears to be generally similar to those for the surface of Vesta. However, the difference in composition of surface materials of Ceres and Vesta, carbonaceous chondrites versus. achondrites (Rivkin et al., 2006; Pieters et al.,...
2006, 2012), suggests that the issue of regolith maturation on Ceres may differ from that on Vesta.

For the average of the first 150 discovered Jupiter Trojans (Nicholson, 1961; Jewitt et al., 2004), the majority of which are kilometer-sized, the survival times of meter-sized boulders were found to be an order of magnitude larger than those for the Moon. This means that when emplaced on the surface by some event, for example by a crater-forming meteorite impact, the majority of the boulders will stay undestroyed for hundreds of millions of years; for complete or almost complete destruction of the boulder subpopulation, a few billion years will be needed. In parallel, general impact resurfacing will also be less effective than on the Moon; this should result in accumulation of thinner regolith layer on average and the maturity of the regolith material should be lower. Therefore, on these bodies the percentage of craters having rocky ejecta should be higher than that on the Moon.

8. Conclusions

The above consideration leads to the following conclusions:

1. Laboratory experiments show that high-velocity impacts into solid rocks produce comminuted material, part of which is ejected from the crater, while another part is not ejected but stays exposed on the target surface and is present in its subsurface.

2. This means that the presence of granulometrically fine material on the surface of small asteroids does not prove the predominance of thermal stresses over rupture by meteorite impacts as a factor in the comminution of surface material on these bodies.

3. Analysis of images, including microphotographs of several lunar rocks of decimeters-to-meters-size, whose exposure ages on the lunar surface was radiometrically dated, showed that their presence for a time period of ~26 – 400 Ma (equivalent to ~ 3 x 10^9 – 5 x 10^9 day–night thermal cycles), did not lead to formation of any features conclusively indicative of rock destruction by thermal cycles.

4. This means that on the lunar surface, as well on the surface of other bodies at 1 AU and further from the Sun, the destruction of rocks by thermal stresses due to day–night cycles is secondary compared to rock rupture by meteorite impacts.

5. Using the complete catalog of the Main Belt and innermost minor planets as a proxy for the distribution of potential impactors throughout the inner Solar System, we calculated the meteorite flux and impact velocities for a number of airless bodies to use them then for estimates of survival times of rock boulders on their surfaces normalized to those for the lunar boulders.

6. It was found that if the average survival time for meter-size rock boulders on the surface of the Moon is 1, then considering rupture by meteorite impacts as the major process of rock destruction, for Phobos it is ~0.8, for Deimos ~0.7, for asteroid Itokawa ~1, for Eros ~0.3, for Vesta and Ceres ~0.03 and, for the average of the first 150 discovered Trojans, ~12.5.

7. One may conclude, as implications of these estimates, that on the surfaces of Vesta and probably Ceres, compared to the Moon, the regolith layer should generally have a larger thickness and a higher maturity of its material, while small craters with rocky ejecta should be more rare. On the typical Trojans, the situation should be the opposite: a thinner layer of regolith, and a lower maturity of its material. These estimates consider meteorite impacts as the major process of rock destruction.

8. These predictions and observations can be tested with future robotic and human exploration of the Moon and small bodies.

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