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Research Paper

On the Survivability and Detectability of Terrestrial Meteorites on the Moon

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ABSTRACT

Materials blasted into space from the surface of early Earth may preserve a unique record of our planet's early surface environment. Armstrong *et al.* (2002) pointed out that such materials, in the form of terrestrial meteorites, may exist on the Moon and be of considerable astrobiological interest if biomarkers from early Earth are preserved within them. Here, we report results obtained via the AUTODYN hydrocode to calculate the peak pressures within terrestrial meteorites on the lunar surface to assess their likelihood of surviving the impact. Our results confirm the order-of-magnitude estimates of Armstrong *et al.* (2002) that substantial survivability is to be expected, especially in the case of relatively low velocity (*ca.* 2.5 km/s) or oblique (≤45°) impacts, or both. We outline possible mechanisms for locating such materials on the Moon and conclude that searching for them would be a scientifically valuable activity for future lunar exploration. Key Words: Meteorites—Moon—Biomarkers—Impacts. Astrobiology 8, 242–252.

INTRODUCTION

THE FIRST THOUSAND MILLION YEARS of Solar System history witnessed many events and processes of importance to astrobiology. These include the period of heavy bombardment and its destructive (*i.e.,* potentially extinction-causing) and constructive (*e.g.,* delivery of volatiles and organic molecules) aspects, and the origin and early evolution of life on Earth. Unfortunately, owing to its active geological and erosional history, Earth itself has preserved hardly any record of these early times (see Rollinson, 2007, for a recent review). However, Armstrong *et al.* (2002) suggested that the Moon will have collected materials, in the form of meteorites, blasted off Earth and other terrestrial planets throughout the history of the Solar System and that such samples may preserve a unique suite of evidence of the early surface environments of these planets. The recovery of such material would provide an important window into early planetary environments, including possible information on the nature and prevalence of early life, which is unlikely to be obtained in any

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other way. Moreover, if found on the Moon, this material, by its very nature, would provide a direct record of the rate at which material has been transferred between the terrestrial planets throughout Solar System history and thus help in constraining models of lithopanspermia (Mileikowsky *et al*., 2000; Burchell, 2004).

The greatest influx of terrestrial materials onto the lunar surface likely occurred during the so-called Late Heavy Bombardment (3.8–4.0 Gyr ago). This is a particularly interesting time in the history of life on Earth but one for which the geological record is extremely sparse. As reviewed by Westall and Southam (2006), there is now considerable evidence that life was well established on Earth by 3.5 Gyr ago. Microfossils [*e.g.*, Schopf and Walter, 1983; Schopf, 1993 (but see Brasier *et al*., 2002); Rasmussen, 2000; Furnes *et al*., 2004] and stromatolites/bacterial mats (Allwood *et al*., 2006; Westall *et al*., 2006) have been found to date from this time. Carbon isotope ratios may indicate the presence of life back to 3.8 Gyr (Mojzsis *et al*., 1996), and though this is more controversial (*e.g.*, Lepland *et al*., 2005), recent results support this interpretation (McKeegan *et al*., 2007). Molecular fossils only survive in less altered rocks but can provide information on organism type (Brocks and Summons, 2005). It would thus be of great value to try to push the record of life further back and to test the indications from phylogenetic analyses that life arose before 4.1 Gyr (Battistuzzi *et al*., 2004). For these reasons, identifying ancient terrestrial materials on the Moon constitutes a strong *astrobiological* argument for a vigorous robotic and human lunar exploration program (Crawford, 2006).

Armstrong *et al.* (2002) estimated that between 10^7 and several times 10^8 kg/km^2 of terrestrial materials may have landed on the Moon during the Late Heavy Bombardment (corresponding to an equivalent layer several mm to several cm thick; or between roughly 0.05% and 0.5% by mass for a 10 m thick regolith). The corresponding quantities for Mars and Venus are lower by approximately 2 and 3 orders of magnitude, respectively. To assess the potential value of the lunar surface as an archive of ancient terrestrial (and other terrestrial planet) material, it is necessary to demonstrate that some fraction of the material will have survived impact with the lunar surface and to consider how we might identify such exotic materials on the Moon. This is the purpose of the present paper.

SURVIVING THE IMPACT

As the Moon has no atmosphere to decelerate impacting meteorites, it is necessary to consider how much, if any, of this material survives impact with the lunar surface in any recognizable form. Armstrong *et al*. (2002) estimated that the maximum impact velocity of terrestrial meteorites landing on the Moon 3.9 Gyr ago would have been about 5 km/s (assuming that terrestrial debris was launched into near-Earth heliocentric orbits with which the Moon subsequently interacted). The minimum impact velocity would have been about 2.3 km/s (slightly lower than the Moon's escape velocity of 2.4 km/s owing to gravitational influence of Earth). In addition to its dependence on the impact velocity, the severity of the shock to which an impacting meteorite is exposed depends on the angle of impact, with oblique impactors being less severely shocked than those which strike the surface at high angles (Pierazzo and Melosh, 2000a, 2000b). In addition, the unconsolidated nature of the lunar regolith may further enhance the survivability of impacting meteorites. Based on these considerations, Armstrong *et al.* (2002) concluded that "the likelihood of terran ejecta surviving in some large aggregate sample is quite high." Here, we attempt to quantify this conclusion with the AUTODYN hydrocode (Century Dynamics, 2003; Hayhurst and Clegg, 1997), after first reviewing estimates obtained by simpler methods.

Calculating the peak pressures

Peak pressures are strongly dependent on the material densities involved in the impact, as well as the impact velocity and angle of impact (Melosh, 1989; Pierazzo and Melosh, 2000a, 2000b). Armstrong *et al*. (2002) assumed a basalt projectile impacting into a basalt target; for a vertical impact at 5 km/s, they estimated a peak pressure of 48 GPa by approximating the peak pressure to the force per unit area while the projectile decelerated over a distance equal to its own diameter within the target. However, shock-wave physics provides more sophisticated tools with which to calculate the peak pressures experienced within an impacting meteorite.

Shock wave equations of state (Hugoniots) are most often presented as relationships between shock velocity, U_s, and particle velocity, U_p, in the form:

$$U_s = C + SU_p$$

where C and S are material parameters from the normal linear shock wave speed equation of state [C is close to the speed of sound in the target and S, the slope of U_s versus U_p, is related to the Grüneisen parameter; see Melosh (1989, p. 39)]. The coefficients C and S are specific to the material involved, and vary according to the pressure regime being considered. Values of C and S for a range of shock regimes (*i.e.,* elastic shock, low pressure phase, mixed pressure region, and high pressure phase) were tabulated by Ahrens and Johnson (1995). We have combined the values given by Ahrens and Johnson (1995) into effective "low" and "high" pressure phases as described by Sharp and DeCarli (2006). This involves plotting the values tabulated by Ahrens and Johnson in a U_p versus U_s diagram and obtaining low and high pressure S and C values from the average gradients and intercepts of the best-fitting straight lines in the 2 regions (P. DeCarli, personal communication, 2007). The resulting values are listed in Table 1 for the basalt and sandstone projectiles considered in this paper.

We first estimated peak pressures in the projectile with the Planar Impact Approximation [PIA; see Melosh (1989, pp 54–57)], a one-dimensional technique adequate to assess the initial coupling of energy from the projectile to the target at the initial moment of the impact event. It is based on the shock Hugoniots of both the projectile and the target; the point at which the Hugoniots intersect represents the peak shock pressure attained in the impact. For the Armstrong *et al.* (2002) basalt-into-basalt impact at 5 km/s (and adopting the same basalt density, 2860 kg m^{-3}, for consistency), the PIA yields peak pressure estimates of 38 and 46 GPa for our low and high pressure choices of C and S, respectively.

A different approach to estimating the peak pressures, the so-called Late Stage Effective Energy (LE) approximation, was described by Mizutani *et al*. (1990). Here, the peak pressure, P_0, is given by

$$P_0 = {}^1\!/_2\xi\rho_{ot}\, C_t^2 \left(1 + {}^1\!/_2\, S\xi\, \frac{v}{C_t}\right)\left(\frac{v}{C_t}\right)$$

where C_t and S are the target material parameters described above, ρ_{ot} is the initial density of the target; v is the impact velocity; and ξ is a parameter related to the shock impedance matching (*i.e.*, the peak pressures which occur in the projectile and the target at the same time) defined by:

$$\xi \approx 2/\left\{1 + \left(\frac{\rho_{ot}C_t}{\rho_{op}C_p}\right)\right\}$$

where the subscripts t and p refer to the target and projectile parameters, respectively; if the target and the projectile are the same material, then ξ is equal to 1. Strictly, as noted by Mizutani *et al.* (1990), the LE approximation applies for pressures that are less than the bulk incompressibility, K, of the material (where $K = \rho_0 \times C^2$; ρ_0 is the density and C is the sound speed). For some of the highest pressures considered here, this approximation may break down; however, the broad agreement between the peak pressures obtained by the LE and the hydrocode simulations (discussed below) gives us confidence that the results have not been seriously affected.

For the choice of parameters given in Table 1, we obtained peak pressures of 38 GPa and 46 GPa for the low and high pressure values of C and S, identical to those obtained using the PIA. It was on the basis of similar values (*i.e.*, peak pressures $\leq$50 GPa) for basalt-into-basalt lunar impacts that Armstrong *et al.* (2002) predicted substantial survivability of terrestrial meteorites striking the Moon, as shock melting of basalt is expected to

TABLE 1. MATERIAL PARAMETERS IMPLEMENTED IN THE SIMULATIONS FOR THE PROJECTILES

		Shock equation-of-state parameters						*Mechanical properties*	
Material	ρ *(gcm^{-3})*	C_1 *(km s^{-1})*	S_1	V_b	V_e	C_2 *(km s^{-1})*	S_2	μ *(GPa)*	Y *(MPa)*
Basalt	2.860	5.22	0.04	0.630	0.610	2.4	1.6	25	135
Sandstone	1.993	2.19	1.08	0.752	0.724	1.67	1.78	10	104

Here, ρ is the density; C_1, S_1, C_2, and S_2 are material-specific equation-of-state parameters (see text); V_b and V_e are the relative volumes of the material before and after the change from low to high pressure phase; μ is the shear modulus and Y is the yield strength (Lama and Vutukuri, 1978).

occur in the pressure range 70 to 80 GPa (*e.g.*, Melosh, 1988; French, 1998; Stöffler *et al.*, 2006).

As shown by Pierazzo and Melosh (2000a, 2000b), lower pressures in the projectile are expected for oblique impacts. Specifically, Pierazzo and Melosh (2000a) found that the peak pressures obtained in oblique impacts scale approximately with the sine of the impact angle (*e.g.*, the peak pressure obtained in a 45° impact is approximately 70% of what it would be in a vertical impact, falling to 34% for a 20° impact, etc.). These lower pressures will in turn enhance the survivability of biomarkers in the projectile, as discussed by Pierazzo and Chyba (1999) in the context of amino acids delivered to planetary surfaces by comet impacts.

We consider now our AUTODYN results for non-vertical impacts and more realistic projectile and target materials than adopted by Armstrong *et al.* (2002).

AUTODYN model initialization

AUTODYN-2D and AUTODYN-3D (based at University College London and the Open University, respectively) were used to simulate vertical and oblique impacts of projectiles into the lunar surface [some background on our use of AUTODYN for impact simulations has been given previously by Baldwin *et al.* (2005)]. To determine the effect of the impact on different projectile materials, we performed separate simulations for basalt and sandstone projectiles, which represent igneous and sedimentary materials, respectively. The projectiles were modeled as cubes, 0.5 m on a side, striking the lunar surface at velocities in the range of 2.5 to 5.0 km/s and impact angles of 90°, 45°, and 20° to the horizontal. The Smooth Particle Hydrodynamics (SPH) solver—a gridless Lagrangian formulation—was adopted for the calculations, with a resolution of 20 SPH particles per projectile diameter; this number is a compromise between the requirements of adequate resolution (see below) and manageable computational time. The projectile was defined with 8000 particles and the target with 360,000 particles. A total of 500 gauge points (the maximum number available in AUTODYN-3D) were spaced at 50 mm intervals in regular arrays parallel to the *x*-axis of the projectile, but randomly distributed in the *y* and *z* directions to ensure coverage of a representative selection of gauges (Fig. 1); 450 gauge points were placed in the projectile, while 50 were distributed around the impact site in the target. For the 2D (90°, vertical) simulations, there were 200 gauge points available, which was sufficient to apply one gauge point to each SPH particle in the projectile. The gauge points were also Lagrangian; that is, they moved with the material location to which they had been assigned, and therefore always measured the state of the same volume of material.

To approximate the unconsolidated nature of

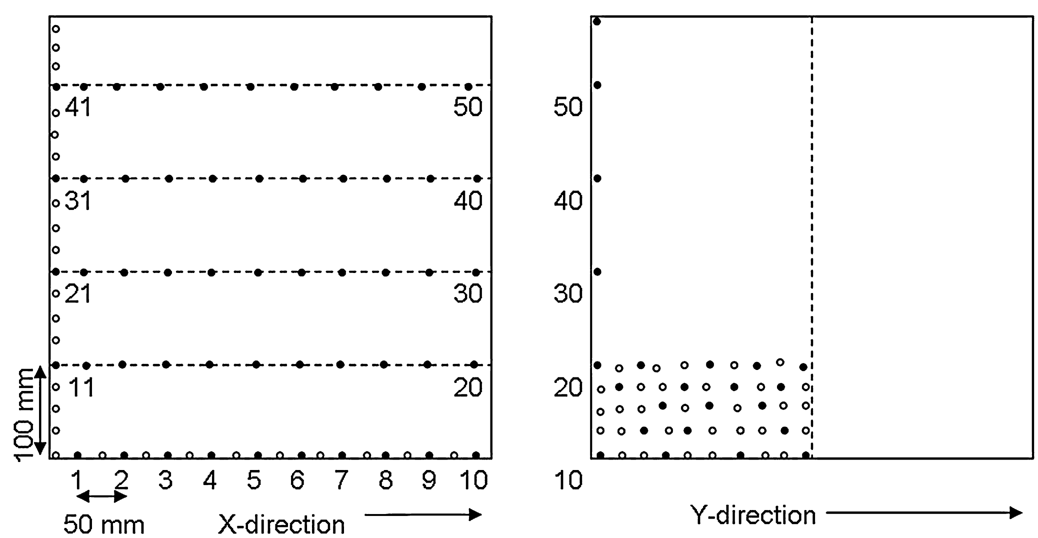

FIG. 1. Simplified diagram illustrating the distribution of gauge points (filled circles, with numbers assigned), and SPH particles (open circles), in the projectile. The left-hand plot shows 50 gauge points on the +XZ face; the gauges are spaced every 50 mm in the +X direction and spaced every 100 mm in the +Z direction. The right-hand plot shows the distribution of gauge points for a sample area of the projectile on the +YZ face.

the lunar regolith, we modeled the target as sand (the physical properties of which are incorporated into the AUTODYN material library using a compaction equation of state); in future efforts, we will attempt to refine these values to be more directly applicable to lunar regolith. The adopted shock equation-of-state input values are listed in Table 1. In addition to the equation of state, the materials are defined by various mechanical properties (*e.g.,* shear modulus, yield strength), which describe the way in which the material fails upon impact. We defined our basalt and sandstone projectiles using values taken from Lama and Vutukuri (1978). The adopted input values for the projectile materials are also presented in Table 1; the target material, sand, is pre-defined in the AUTODYN material library, based on work by Laine and Sandvik (2001).

For the purpose of our study, we were concerned with determining the peak pressures generated in the initial stages of impact. Therefore, we ran our models so that the duration of the shock state was included. The duration of the shock state is the time required for the shock wave to propagate through the projectile, reflect at the rear surface, and travel back through the projectile as a rarefaction wave that releases the material from its shocked state. This is approximately the time, t, for the projectile to traverse a depth equal to its own diameter, D, into the target (*e.g.,* Pierazzo and Melosh, 2000a);

$$t = D / v \sin \theta$$

where v is impact velocity and θ is the impact angle. For our choice of impact angles (90°, 45°, and 20°) and impact velocities (2.5 km/s and 5 km/s), the duration of the shocked state varies from 0.1–0.6 ms. Accordingly, all our models were run for several milliseconds, which represents between 500 and 1000 computational cycles.

Results

As an initial experiment and test of our models, we first used AUTODYN to duplicate the vertical 5 km/s impact of basalt into basalt considered by Armstrong *et al.* (2002). In this case, we obtained a peak pressure of 77 GPa, which is higher than the PIA and LE results. However, this pressure was reported by only one gauge point (located on the impacting face) out of a total of 500 gauge points that sampled the whole projectile; the median peak pressure recorded within the projectile was 22 GPa (with a minimum value of 2 GPa recorded on the trailing edge). Thus, as expected, peak pressures were higher close to the impacting face; if we consider the leading and trailing halves of the projectile separately (see below), we obtained median peak pressures of 28 and 14 GPa, respectively. Thus, while the PIA and LE methods described above appear to underestimate the peak pressures encountered close to the impacting edges of the projectile, they overestimate the peak pressures encountered within the body of the projectile.

Given that the lunar regolith (here modeled as sand) and terrestrial sedimentary rocks (here modeled as sandstone) have much lower densities than basalt, we would expect even lower peak shock pressures than those estimated by Armstrong *et al.* (2002) and, thus, a significantly increased probability of impact survival. We would also expect survivability to be further enhanced in the case of oblique impacts for both basalt and sandstone impactors, which also experience lower peak shock pressures (Pierazzo and Melosh, 2000a, 2000b).

To explore these effects quantitatively, peak pressures were computed for impact velocities of 2.5 and 5.0 km/s into sand and a range of impact angles. The results are given in Table 2. We present the median value of the peak pressure recorded by all gauge points and the range of peak pressures around this value. We also present the median peak pressure values for the leading and trailing halves of the projectile (defined in Fig. 2). Median values are listed because the maximum values are generally returned by just 1 or 2 gauge points, whereas the median represents the midpoint of the returned values and thus better represents the typical peak pressure experienced by the bulk of the projectile. Table 2 also lists the predictions made with the PIA and LE techniques for comparison. [For the latter calculations, the material parameters adopted for the projectile are those given in Table 1; those for sand are from Ahrens and Johnson (1995): ρ_{ot} = 1610 kg m^{-3} (very close to the measured bulk density of lunar regolith of 1660 kg m^{-3}; Carrier *et al.*, 1991); S_1 = 0.46; C_1 = 1.70 km s^{-1}; S_2 = 1.1; C_2 = 2.10 km s^{-1}, where the subscripts 1 and 2 refer to low and high pressure phases, respectively]. As would be expected, and as was confirmed by the earlier calculations of Pierazzo and Melosh (2000a, 2000b), peak pressures were

TABLE 2. RESULTS OF AUTODYN SIMULATIONS

Impact angle (degrees)	*Projectile material*	*Impact velocity (km/s)*	*Median peak pressure (GPa)* Total	Leading	Trailing	*PIA (GPa)*	*LE (GPa)*
90	Basalt	2.5	$4.5^{+14.2}_{-4.3}$	7.2	2.6	7–10	9–11
		5	$13.7^{+79.5}_{-12.9}$	18.9	7.3	19–28	25–31
	Sandstone	2.5	$3.4^{+18.0}_{-3.2}$	4.8	2.0	6–8	6–7
		5	$12.8^{+112.7}_{-11.8}$	17.9	7.7	16–24	15–19
45	Basalt	2.5	$3.7^{+15.3}_{-3.2}$	4.5	2.3	5–7	6–8
		5	$12.6^{+47.9}_{-11.1}$	15.2	7.6	14–20	18–22
	Sandstone	2.5	$2.8^{+10.9}_{-2.6}$	3.7	1.6	4–6	4–5
		5	$11.3^{+40.3}_{-10.1}$	13.1	7.4	11–17	11–14
20	Basalt	2.5	$1.9^{+11.6}_{-1.8}$	2.6	1.3	2–3	3–4
		5	$5.9^{+61.5}_{-4.9}$	7.9	4.1	7–10	9–11
	Sandstone	2.5	$1.1^{+9.5}_{-1.0}$	1.4	0.8	2–3	2–3
		5	$4.5^{+33.0}_{-4.0}$	5.9	3.5	5–8	5–7

Results are expressed as the median values of peak pressures recorded by the 500 gauge points sampling the projectile, and for the leading and trailing halves. For the whole projectile, the total range of peak pressures recorded around the median value is indicated. Results for the PIA and LE approximations are given for comparison, where the range corresponds to that obtained for low and high pressure values for the material properties; values for oblique impacts have been obtained by multiplying by the sine of the impact angle (see text for details).

higher in the leading halves of the projectile than in the trailing, which leads to an enhanced probability of projectile material surviving from the trailing portion of the projectile.

As discussed by Pierazzo *et al.* (1997), it is necessary to ensure that hydrocode calculations are performed at sufficient resolution to ensure that they have converged, and this is a particular concern for SPH codes (H.J. Melosh, personal communication, 2007). We do not have the computational resources to repeat all our calculations at higher resolution, but we have performed a resolution test by repeating the 90° basalt projectile impact with resolutions of 10, 20, 40, and 100 SPH particles across the projectile. The results are shown in Fig. 3, from which it is apparent that the pressures obtained in the 20 SPH particle case (*i.e.,* those listed in Table 2) are underestimates of the true values but probably only by about 30–50% in the case of the whole projectile. It should be noted that in most cases such a correction would bring the median hydrocode numerical values into better agreement with those estimated by the PIA and LE methods.

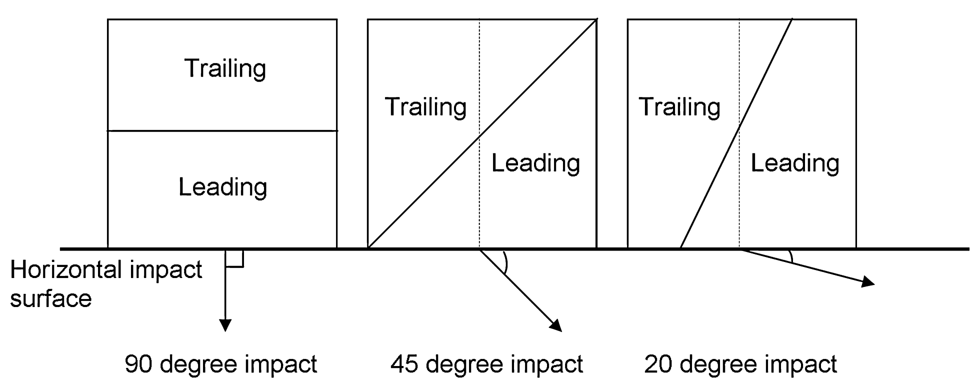

FIG. 2. Diagrams to show the definition of the leading and trailing halves of the projectile for impacts at angles of 90° (vertical), 45°, and 20° from the horizontal.

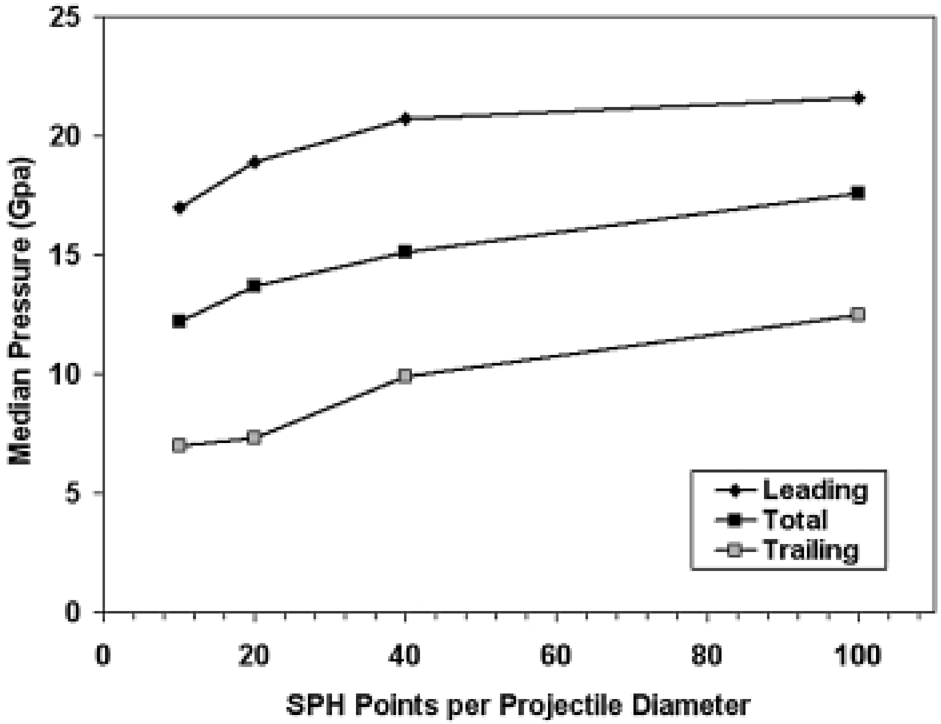

FIG. 3. The effect of SPH resolution on median peak pressures in the projectile for a basalt projectile and a 90° impact at 5 km/s. Results are plotted for the whole projectile and the leading and trailing halves, as indicated. It is clear that the results obtained with 20 SPH particles across the projectile (given in Table 2) will have underestimated the peak pressures to some extent (probably by 30–50% when considering the projectile as a whole), but not by so much as to materially affect the conclusions of the analysis.

SURVIVAL OF BIOMARKERS

The effects of impact-induced shock on geological materials were recently summarized by Stöffler *et al*. (2006) as follows: the transition between elastic to non-elastic behavior typically occurs in the range of 5 to 12 GPa; mechanical deformation and transition to high-pressure mineral phases occur between about 10 and 60 GPa; and finally, whole-rock melting begins at about 80 GPa for mafic rocks (such as our basalt projectile) and 40 GPa for porous siliceous rocks (such as our sandstone projectile). Post-shock temperatures would be expected to be of the order of 100°C at 10 GPa, rising to about 1000°C at 50 GPa (French, 1998).

The results shown in Table 2 indicate that, for the cases of a vertical impact at 5 km/s, some portion of the sandstone and basalt projectiles experienced pressures above that at which melting is expected to occur but that the bulk of the projectile, and especially the trailing half, was subjected to much lower pressures. Indeed, Table 2 shows that, in all cases (*i.e.,* even for 90° impacts at 5 km/s impacts), the median peak pressures attained in the trailing half of the projectile were all <10 GPa and generally less than 5 GPa. The projectile material should thus be essentially unaltered by impact-induced shock processes. At 2.5 km/s, no part of the projectile even approached a peak pressure at which melting would be expected. These conclusions are not substantially altered by the resolution issues discussed above, and they confirm the general conclusion reached by Armstrong *et al*. (2002), which was that substantial survivability of terrestrial meteorites on the Moon is to be expected. It should be noted that similar conclusions have recently been reported by Bland *et al.* (2007) in the context of "normal" (*i.e.*, non-terrestrial and therefore higher velocity) meteorite impacts on the lunar surface.

As Armstrong *et al*. (2002) noted with regard to survival in a projectile, biomarkers may range from those that are likely to be preserved in any intact meteorite (*e.g.,* isotope ratios and organic carbon) to those that may only be preserved in the least-shocked and, thus, rarest specimens (*e.g.*, complex molecular fossils or actual microfossils, or both). In this context, we note that for most of the impacts (and especially for the trailing halves) much of the projectile's volume was subjected to peak pressures below 5–12 GPa, at which the transition from elastic to plastic behavior is expected to occur, which led us to expect substantial survival of even quite fragile biomarkers in this type of material. On the other hand, the survival of at least some biomarkers (microfossils, for instance) will presumably depend on the extent to which an impacting meteorite remains physically intact, regardless of the peak pressures to which it is subjected. We might expect relatively weakly cemented sedimentary rocks to fare less well than crystalline igneous materials in this respect. It is important to note, however, that the search for evidence of life need not be restricted to relatively weak sedimentary rocks. Microbial communities are found in the glassy margins of ocean floor basalts (Thorseth *et al.,* 2001), and evidence of life in the form of mineralized tubes, organic carbon, and carbon isotope ratios has been found in Archean pillow lavas (Furnes *et al.,* 2004). It is also true that many rocks likely to contain ancient fossils (*e.g.,* chert) have a strength between that of sandstones and basalt. We hope to address some of these issues in future work.

METEORITE SURVIVABILITY ON THE MOON

Terrestrial meteorites that survive initial impact with the lunar surface will begin to be broken down by the same processes of micrometeorite bombardment responsible for the formation of the lunar regolith. This process is relatively rapid. For example, Hörz *et al.* (1991) estimated that a typical 1 kg rock is likely to survive for of the order of 10 Myr on the lunar surface. As noted above, the bulk of terrestrial materials on the Moon, and certainly that which is the most astrobiologically interesting, will have arrived during the heavy bombardment prior to 3.8 Gyr ago. While, on the one hand, the generation of large-scale basin ejecta blankets during the heavy bombardment and the later coverage of large areas by mare basalts would have buried most of this material to depths where it has been protected from later impact erosion, it is also true that such burial greatly complicates efforts to locate it.

Fortunately, as pointed out by Armstrong *et al.* (2002), the continual gardening of the lunar regolith by impacts of all sizes is continuously bringing buried materials to the surface, and this will presumably include some fraction of the total quantity of terrestrial materials that are sequestered at depth. Following this argument, Armstrong *et al.* (2002) estimated that at present the average concentration of terrestrial materials exposed at the lunar surface will be of the order of 7 ppm. However, the extent to which biomarkers within buried terrestrial material will survive the impact excavation, gardening, and comminution of the regolith is yet to be assessed, and it may be that the deliberate excavation of terrestrial meteorites preserved at depth [*e.g.,* in buried palaeoregolith deposits (Crawford, 2006; Crawford *et al.,* 2007)] will be required to obtain scientifically useful samples.

DETECTING EARTH ROCKS ON THE MOON

Although it appears that many terrestrial meteorites survive their impact with the lunar surface, locating them will be challenging. Fortunately, there are significant differences between terrestrial and lunar rocks, which will help with detection. First, there are differences in the bulk composition of the Moon. The Moon is devoid of water (*e.g.,* Papike *et al*., 1998), depleted in volatile elements, and enriched in refractory lithophile elements (Taylor, 2001; see Jolliff *et al*., 2006, for reviews). Second, the absence of water, and of plate tectonics, on the Moon means that many mineral-forming processes that operate on Earth cannot occur on the Moon. Thus many terrestrial rocks should be easily identifiable by their mineral composition. Infrared spectroscopy, particularly in the wavelength range from 1.4–3.0 μm, may be an effective tool for detecting terrestrial meteorites. Common lunar minerals such as pyroxene, olivine, and plagioclase have few spectral features in this wavelength range (Pieters *et al.*, 2006), though pyroxene has a very broad feature at $\sim$2 μm. However, hydrated silicates exhibit a strong absorption band at $\sim$3 μm and weaker bands at 1.4 and 1.9 μm, due to O–H stretching and H–O–H bending modes (Gaffey *et al*., 1993; Bibring *et al.,* 2005). Recent evidence from detrital zircon grains indicates that liquid water was probably present on Earth's surface as early as 4.4 Gyr ago (Wilde *et al*., 2001; Watson and Harrison 2005), so even the earliest Earth rocks may be expected to exhibit spectral evidence of hydration.

Similarly, carbonates exhibit a characteristic series of sharp, narrow absorption features in the spectral region between 2 and 3 μm (Gaffey, 1986; Gaffey *et al.*, 1993), which may also be diagnostic of a terrestrial origin. Other minerals that result from aqueous processes, such as clay minerals and sulfates, also have distinctive spectral features in this region.

Large meteorites (~1 m in size) might be detectable from orbit with the use of a hyperspectral imaging system based on the "push-broom" imaging approach (Mouroulis *et al.*, 2000). An example of such a system is the Moon Mineralogy Mapper (M3), which will fly on the Chandrayaan 1 mission. M3 (Green *et al.*, 2007; Pieters *et al.*, 2006) is a compact, low-mass instrument (8 kg) with a resolution of 10 nm over 430–3000 nm and has a spatial resolution of 70 m/pixel. A similar instrument, fed by a much larger telescope and perhaps operating in a lower orbit, could be capable of spatial resolutions of ~1 m or less. A rover equipped with a hyperspectral imaging system could search for smaller meteorites exposed at the surface. For such a system, an imager fed by an acousto-optic tunable filter (*e.g.*, Gupta 2005) would be an appropriate compact technology. For both the orbital and surface cases, the spectra of a very large number of image points would be collected. The system could be programmed to search for spectra with a particular spectral feature or to look for unusual spectra—*i.e.*, anything different from the typical spectra seen from normal lunar materials. We also note that the recent discovery of up to 5 meteorites on the surface of Mars by the Mars Exploration Rovers Spirit and Opportunity (*e.g.*, Rodionov *et al.*, 2005; Ashley *et al.*, 2007) demonstrates that the detection of meteorites on planetary surfaces by *in situ* robotic means is achievable (though, of course, the martian atmosphere would have decelerated these meteorites during infall, and they survived impact with the surface in a more pristine state than would be the case for lunar examples).

A parallel can be drawn with regard to the now-routine collection of meteorites in Antarctica: just as there are places in Antarctica where any rock sighted on the surface is likely to be a meteorite, we can imagine scanning the lunar surface with infrared instruments sensitive to hydrated silicates such that every specimen detected stands a good chance of being a terrestrial (or martian) meteorite deserving of more detailed analysis. It is certainly possible to design a suitable robotic infrared imaging system that could survey hundreds of square kilometers quite quickly and thus efficiently identify candidates despite their expected rarity. On the other hand, the collection of these candidate materials, their preliminary characterization *in situ*, and the return to Earth of selected specimens for more detailed analysis would be greatly facilitated by a human presence on the Moon (Crawford, 2004). Further, the search for buried materials (*e.g.*, in palaeoregolith deposits) may be impractical without a human presence.

CONCLUSIONS

Terrestrial meteorites on the Moon may preserve valuable information about Earth's early surface environment, including evidence for the conditions under which life arose on our planet. Here we have used the AUTODYN hydrodynamics code to calculate the peak pressures within terrestrial meteorites impacting the lunar surface. Our results confirm the order-of-magnitude estimates of Armstrong *et al*. (2002) that substantial survivability is to be expected, especially in the case of relatively low velocity (*ca.* 2.5 km/s) or oblique (≤45°) impacts, or both. We have considered how terrestrial materials might be located on the Moon and have identified a number of possible spectroscopic detection methods. While robotic searches for terrestrial (and other terrestrial planet) materials is possible in principle, for reasons set out by Crawford (2004) this is the kind of large-scale exploratory activity that would be greatly facilitated by a renewed human presence on the Moon.

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ABBREVIATIONS

LE, Late Stage Effective Energy; PIA, Planar Impact Approximation; SPH, Smooth Particle Hydrodynamics.

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