Shock compression and adiabatic release of lunar fines from Apollo 17

THOMAS J. AIHENES and DAVID M. COLE
California Institute of Technology, Pasadena, California 91109

Abstract—Shock compression and adiabatic release data on lunar fines (sample, 70027), with an initial density of 1.80 g/cm³, were obtained in the 26-125 kbar range. The sample contains approximately 60% pyroxenes, 30% plagioclase, and 10% brown glass and opaques, and has an estimated intrinsic density of 3.50 ± 0.10 g/cm³. Reversible compression, and presumably, indentation to a post-shock, zero-pressure density of 3.00 ± 0.05 g/cm³ occurs upon shock compression to 20 kbar (or lower) and subsequent adiabatic release. Above this pressure level the Hugoniot data are closely predicted, at a given volume, by using the thermal pressure calculated from the Max-Gruneisen equation (Gruneisen's ratio 0.80 to the pressure above the theoretical Hugoniot for intrinsic crystal density matrix. For shock compression to pressures of between 20 and 125 kbar, the calculated post-shock temperature varies from 2.15 ± 750 ± 155°C. When taken with the determined post-shock density variation from 3.1 to 2.6 g/cm³, this result implies that increasing quantities of glassy material is produced via solid-state reaction and possibly melting over this pressure range. The origins of post-shock thermal melting is calculated to occur upon impact of iron or stony meteoroids travelling at a minimum of 2.5 and 3 km/sec, respectively. This implies that appreciable solid-state transformation of plagioclase and melting of previously highly shocked surficial matrix will result only from primary impacts. However, the pressures required for irreversible compaction imply that liquefactions will form in the rigipith upon impact of secondary objects ejected at speeds on the order of only 1 km/sec.

INTRODUCTION

The continual bombardment of the upper several meters of the mixed and unconsolidated ejecta blanket covering the moon by meteoroids in the mass dominant 10⁻⁷ to 10⁻³ g range, gives rise to the continual (a) irreversible local compaction and agglutination of fines, (b) production of glasses within crystaline fragments by both shock-induced solid-state and thermal vitrification, and (c) selective volatilization. The details of these and other shock-induced processes (e.g. erosion of fusion tracks, argon loss or gain which modify individual mineral grains within the fines have yet to be extensively studied. The present equation of state experiments, the first to be carried out on lunar soil, emphasize the delineation of gross macroscopic and continuum properties resulting from shock loading.

Taken by themselves, the Hugoniot data for lunar fines may be used in conjunction with impedance match method (Rice et al., 1958), to determine the maximum shock pressures generated in the lunar regolith upon impact of stony or iron meteoroids at various velocities. In general, the mass of surface material

*Contribution No. 2425, Division of Geological and Planetary Sciences, California Institute of Technology, Pasadena, California 91109.
exposed to this peak pressure will be on the order of the mass of the impacting meteoroid. In the case where sufficient release adiabat and Hugoniot data are available, it is possible to construct a complete equation of state and apply it to the complete calculation of shock attenuation in the target (and projectile) material and hence determine the mass of material initially exposed to various dynamic pressures prior to being excavated and ejected from a given crater. Laboratory studies (Gault et al., 1968) have demonstrated that in unconsolidated media virtually all material exposed to even a few kilobars of dynamic pressure are subsequently ejected from the crater, or return as fall back, following passage of the initial shock wave.

Although no previous Hugoniot or release adiabat data are reported for basalt-like aggregates of mafic minerals and plagioclase, the dynamic properties of various silicic and alkali-rich tuffs and alluviums, in dry, wet and frozen condition have been studied by, for example, Lyne (1970), Murri and Grady (1973), Peterson et al. (1970), and Anderson (1968). In addition, a few very high pressure Hugoniot points for a pumice are reported by Jones et al. (1968). In general, the zero-pressure crystal densities of these materials varied from ~2.0 to 2.3 g/cm³, with the initial porous densities in the 1.5 to 1.8 g/cm³ range. Of special interest are the Hugoniot and release data obtained in vacuum by Peterson et al., for a 1.8 g/cm³ playa alluvium, which indicate that upon compression to ~10 kbar, irreversible compaction to a crystal density of ~2.2 g/cm³ is achieved. Also, for a dry Nevada tuff (intrinsic density 2.2 g/cm³), the release adiabats curves centered at a Hugoniot pressure of ~65 kbar as measured in air by Murri and Grady demonstrated that a post-shock density of only 2.0 g/cm³ was achieved.

The present experiments were carried out in order to place quantitative bounds, in terms of shock pressure and hence meteoroid impact velocity, on the conditions required to compact and lithify lunar fines (packed to an initial density typical of the regolith). Our objective was also to use the measured pressure-particle velocity release states to obtain approximate values of shock pressure associated with various post-shock volumes and temperatures concomitant with solid-state vitrification and thermal melting.

**Sample Material.**

Experiments were conducted with a series of aliquots of sample 70851. The sample consists of angular lithic and mineral fragments, ~5-500 μm in diameter, having the approximate composition: 60% pyroxenes, 30% plagioclase, and 10% brown glass and opaques. Although this sample was obtained from the same sample bag (BSL085) as rocks 70017, 70215, 70655, and 7035 collected near the base of the North Massif, its composition (Table 1) more closely matches fines 70161, 70181, 71041, 71061, and 71501 collected over the Taurus-Littrow region (LSPET, 1974). The estimated zero pressure intrinsic density of this sample, 3.1 ± 0.1 g/cm³ is substantially greater than that of the terrestrial tuffs and alluviums previously studied. Thus, there are no comparable data for terrestrial materials.
An initial packing density of 1.8 g/cm³ was used in the present study. This density was meant to be representative of the bulk density of the upper several meters of the regolith at the Apollo 12, 15, and 17 sites as determined on the basis of studies by Carrier et al. (1972, 1973), and Mitchell et al. (1973). In addition to the Hugoniot and release adiabat measurements on 70051, two experiments were carried out for comparison purposes using powdered Viscasil biscuit (Ahrens and Gregson, 1964), prepared to a similar starting density.

The Hugoniot measurements were made using the impedance-matched method (Rice et al., 1958). Initial sample density, sample shock velocity and projectile velocity are measured using previously described techniques (Ahrens et al., 1975). States along the release adiabats, centered at Hugoniot states, were determined using a variant of the buffer-material method described by Ahrens et al. (1969). Because of the low shock impedance of the samples, and our desire to carry out the entire experiment in a moderate vacuum (50-100 µHg), the use of organic fluids was unfeasible. Instead we chose a fine-grained (~0.1 mm) polystyrene foam with a zero-pressure bulk density of 0.0082 ± 0.0010 g/cm³ as the buffer material (Fig. 1). This is a very mechanically weak and compressible material. Although the Hugoniot data for styrofoam (McQueen et al., 1970) show considera-

Fig. 1. Diagram of release adiabat experiment for measurements on lunar regolith. (a) Cross section of release adiabat assembly along plane of sight. Impact psi, which triggers streak camera, rotated into plane of section. Piezoelectric crystal close to 90° from direction of the sonic impact of projectile bearing flyer plate, shock is driven into driven plate, specimen, and foam buffer material. Über three interaction of shock in sample with buffer material, adiabatic release wave propagating to the left in specimen. (b) Photograph of specimen showing position of Optex II of view (no streak operation). (c) Streak photograph, time 357.
ble scatter, the available results demonstrate that the shock velocity, $U_s$, depends very strongly on particle velocity, $u_0$, according to:

$$u_0 (\text{km/sec}) = 0.908 U_s - 0.051$$  \hspace{1cm} (1)$$

Moreover, Eq. (1) adequately describes the available data for polystyrene foams with initial densities varying from 0.032 to 0.174 g/cm$^3$. In the experiments, a shock is driven into the sample and is then transmitted to the overlying buffer material. As the shock propagates into the buffer material, a rarefaction wave is reflected back into the sample from the sample-buffer interface, releasing the material adiabatically from the Hugoniot state.

To determine the sample release state, we must determine the Hugoniot state in the buffer material and assume the pressure and particle velocity at the buffer-sample interface are continuous. The velocity of the shock in the buffer is measured and, using Eq. (1) and the Rankine-Hugoniot momentum equation, the particle velocity and pressure state along the release adiabat are determined.

**Experimental Details**

Preweighed aliquots of the sample material were cold pressed to an initial density of 1.80 g/cm$^3$ within rings of steel (16 mm internal diameter, 4 mm thick) and of lexan (8 mm internal diameter, 2 mm thick) for the Hugoniot and release adiabat experiments, respectively. By making the sample thickness large relative to the grain size, we hoped to achieve a dynamic equilibrium state behind the shock front. We doubt that true thermal equilibrium is achieved in the short duration of our experiments. One flat face of the sample was in direct contact with the W, Ta, or 2024 Al driver plate; the other surfaces were sealed with 0.01 mm thick, tightly stretched aluminized mylar. In order to test the sample loading procedure, we dissected a loaded 850 mg sample to recover the lunar material. Recovery of all of the sample, except for 0.6 mg, was demonstrated.

Projectile and shock velocities are measured by laser beam obscuration and streak camera techniques. As indicated in Fig. 1a, a 2.5 mm thick flyer plate impacted in the front surface of a lexan projectile impacts the driver plate on which the sample is mounted. The driver and flyer plates are of the same material. The arrival of the shock, resulting from flyer plate impact at the plane of the driver plate-specimen interface and the specimen free-surface is detected by the loss of specular reflectivity of the arrival mirrors and the aluminized mylar foil, respectively. In the release adiabat experiments, the loss of reflectivity of the mylar foil stretched across the surface of the 5 mm square, 1.1 mm thick polystyrene foam sample was used to detect the shock arrival at its free-surface. This shock arrival time was not recorded as clearly as the others (Fig. 1c), and the associated time and hence, the resulting uncertainty of the shock velocity in the foam represents the major source of error in the measurement of the release adiabat states, aside from uncertainties in the foam Hugoniot.
EXPERIMENTAL RESULTS

Above 50 kbar the Hugoniot data for sample 70051 (Fig. 2, Table 2) lie close to the extrapolated cold compression curve for the lunar fines examined by Stephens and Lilley (1970). The shock datum point at 20 kbar (shot 347) lies at a distinctly lower volume than the static data; it appears to lie along a compression curve which approximates the intrinsic pore-free material. Therefore, this point provides only an approximate upper bound to the shock pressure required to compress this type of material to a close to intrinsic density. We note also that the datum point (shot 304) for a powdered Vacaville basalt sample lies at least 25 kbar above the lunar data.

In order to separate the possible effects of mechanical resistance to shock compaction and shock heating, both of which tend to reduce the density at a given pressure, we constructed a theoretical intrinsic mineral Hugoniot (Figs. 2 and 3) from the mineralogic and chemical model of Table 1. Assuming Stephens and Lilley's compression curve for a material with intrinsic density of 3.13 g/cm³ and a Gruneisen's ratio, γ, of 0.8 (representing an average between 0.45 and 1.0 for plagioclase and pyroxene), a theoretical Hugoniot for an arbitrary dispersion ratio
Table 1. Assumed mineralogic and chemical models, lunar sample 78051.

<table>
<thead>
<tr>
<th>Volume (%)</th>
<th>Density (g/cm³)</th>
</tr>
</thead>
<tbody>
<tr>
<td>32</td>
<td>30</td>
</tr>
<tr>
<td>19</td>
<td>33</td>
</tr>
<tr>
<td>8.5</td>
<td>2.7</td>
</tr>
<tr>
<td>21</td>
<td>3.4</td>
</tr>
<tr>
<td>9.6</td>
<td>4.8</td>
</tr>
<tr>
<td>5.7</td>
<td>2.8</td>
</tr>
<tr>
<td>1.5</td>
<td>7.2</td>
</tr>
</tbody>
</table>

Weight (%)

<table>
<thead>
<tr>
<th>Element</th>
<th>Percentage</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>40</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>9</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>11</td>
</tr>
<tr>
<td>CaO</td>
<td>17</td>
</tr>
<tr>
<td>MgO</td>
<td>10</td>
</tr>
<tr>
<td>CaO</td>
<td>11</td>
</tr>
</tbody>
</table>

![Diagram](image)

Fig. 3. Theoretical Hugoniot for lunar tuff for various values of distortion (m). Data from present samples, m = 1.75, are plotted for comparison.

\[ \rho_0 = \rho_s \left[ \frac{1 - \frac{3}{2} \left( V_s / V \right) - 1}{1 - \frac{3}{2} \left( m V_s / V \right) - 1} \right] \]

where \( \rho_0 \) and \( \rho_s \) are the distorted and intrinsic density and \( p_s \) and \( p_n \) are the Hugoniot pressures along the intrinsic and distorted Hugoniot curves at a given compressed volume, \( V \). Equation (2) follows from the Mie-Grüneisen equation of state and the conservation of energy Runkle-Hugoniot equation (Ahrens and
<table>
<thead>
<tr>
<th>Shot No.</th>
<th>Initial density (g/cm^3)</th>
<th>Flyer plate velocity (km/sec)</th>
<th>Shock pressure (kb)</th>
<th>Shock density (g/cm^3)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jef</td>
<td>2.080 ± 0.002</td>
<td>0.010 ± 0.003</td>
<td>19.9 ± 0.4</td>
<td>3.16 ± 0.06</td>
</tr>
<tr>
<td>312</td>
<td>1.796 ± 0.002</td>
<td>1.047 ± 0.004</td>
<td>39.5 ± 0.2</td>
<td>3.16 ± 0.06</td>
</tr>
<tr>
<td>313</td>
<td>1.795 ± 0.002</td>
<td>1.444 ± 0.006</td>
<td>66.1 ± 0.2</td>
<td>3.41 ± 0.06</td>
</tr>
<tr>
<td>325</td>
<td>1.799 ± 0.002</td>
<td>1.573 ± 0.006</td>
<td>77.9 ± 0.2</td>
<td>3.49 ± 0.06</td>
</tr>
<tr>
<td>317</td>
<td>1.830 ± 0.005</td>
<td>2.105 ± 0.006</td>
<td>126.0 ± 0.2</td>
<td>3.68 ± 0.06</td>
</tr>
<tr>
<td>804</td>
<td>1.762 ± 0.005</td>
<td>1.181 ± 0.006</td>
<td>54 ± 0.2</td>
<td>3.97 ± 0.06</td>
</tr>
</tbody>
</table>

*2032 Air flyer plate.
*313 tantalum flyer plate.
*331 tantalum flyer plate.

O’Keefe, 1972). The present data, corresponding to m = 1.72, lie very close to the m = 1.7 curve. This result is consistent with the idea that shock compaction is essentially complete at 20 kb (or lower) whereas at higher pressures, the Hugoniot of the lunar fines is controlled by shock heating and compaction of the constituent minerals (i.e. not their mechanical compaction characteristics). Above ~150 kb this simple model will probably be invalid as the shock induced phase changes which are known to occur in plagioclase and pyroxene (Ahrens et al., 1969; Ahrens and Gaffney, 1971) undoubtedly occur.

The release adiabat results, although useful, are far from complete at this point as only a single point along each of the adiabats centered at ~42, ~76, and ~119 kb is explicitly determined in the pressure-particle velocity plane (Table 3, Fig. 4). Since the Hugoniot curve of styrofoam lies effectively at zero-pressure, as contrasted to the lunar fines’ Hugoniot, the release adiabat points determined are essentially at zero-pressure. (At a particle velocity of 3 km/sec, the Hugoniot pressure in the styrofoam is only ~4 kbar.)

To invert release adiabat data from the pressure-particle velocity plane to the pressure-volume plane, we must assume the existence of a simple release isentrope along which pressure is a single-valued function of the density. In this case, the Riemann integral formula applies:

\[ V = V_0 \left(1 + \int \frac{\partial n}{\partial V} \right) \] (3)
<table>
<thead>
<tr>
<th>Shot No.</th>
<th>Flyer plate velocity (km/sec)</th>
<th>Foam shock velocity (km/sec)</th>
<th>Inferred Hugoniot particle velocity (km/sec)</th>
<th>Inferred Hugoniot particle volume (cm³/g)</th>
<th>% release adjusted particle velocity (km/sec)</th>
<th>Post-shock volume (cm³/g)</th>
<th>Lower bound, post-shock temperature (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>345</td>
<td>0.810' ± 0.002</td>
<td>0.973 ± 0.013</td>
<td>19.0' ± 0.9</td>
<td>0.682' ± 0.002</td>
<td>0.316' ± 0.011</td>
<td>0.832 ± 0.001</td>
<td>0.327 ± 0.011</td>
</tr>
<tr>
<td>346</td>
<td>1.027 ± 0.004</td>
<td>1.255 ± 0.014</td>
<td>42.0 ± 0.9</td>
<td>0.985 ± 0.014</td>
<td>0.325 ± 0.004</td>
<td>1.343 ± 0.003</td>
<td>0.540 ± 0.003</td>
</tr>
<tr>
<td>336</td>
<td>1.051 ± 0.004</td>
<td>1.383 ± 0.008</td>
<td>42.0 ± 0.9</td>
<td>0.985 ± 0.014</td>
<td>0.325 ± 0.004</td>
<td>1.205 ± 0.006</td>
<td>0.536 ± 0.006</td>
</tr>
<tr>
<td>331'</td>
<td>1.097 ± 0.004</td>
<td>1.360 ± 0.013</td>
<td>44.8 ± 0.9</td>
<td>1.020 ± 0.014</td>
<td>0.323 ± 0.004</td>
<td>1.184 ± 0.002</td>
<td>0.529 ± 0.008</td>
</tr>
<tr>
<td>337</td>
<td>1.563 ± 0.006</td>
<td>2.311 ± 0.015</td>
<td>79.5 ± 0.9</td>
<td>1.435 ± 0.014</td>
<td>0.297 ± 0.014</td>
<td>2.047 ± 0.015</td>
<td>0.344 ± 0.015</td>
</tr>
<tr>
<td>340</td>
<td>1.947 ± 0.008</td>
<td>3.386 ± 0.074</td>
<td>117 ± 1.8</td>
<td>1.810 ± 0.275</td>
<td>0.2757 ± 0.014</td>
<td>3.023 ± 0.077</td>
<td>0.401 ± 0.015</td>
</tr>
<tr>
<td>339</td>
<td>1.935 ± 0.008</td>
<td>3.549 ± 0.16</td>
<td>119 ± 1.8</td>
<td>1.825 ± 0.275</td>
<td>0.2757 ± 0.014</td>
<td>2.990 ± 0.037</td>
<td>0.390 ± 0.037</td>
</tr>
</tbody>
</table>

1. 't' Aluminum flyer plate.
3. Xanthion flyer plate.
4. Sample, powdered Vacuval bolt. 1.800 g/cm³.
5. Tungsten flyer plate.
Here $V$ and $u$ are specific volume and particle velocity, respectively (the subscript $s$ refers to the Hugoniot shock state) and $c_s$ is the isentropic bulk sound speed along the release adiabat. A straight line adiabat was assumed in the pressure–particle velocity plane connecting states along the sample Hugoniot with the corresponding release states lying on the symmetric Hugoniot. Using Eq. (5) it can be shown that a straight line isentrope in the pressure–particle velocity plane maps into a straight line in the pressure–volume plane. With the above approximation and the further approximation that the pressure along the styrofoam is effectively zero, it follows from Eq. (5) that the zero-pressure, post-shock volume is given by

$$V_s' = V_s + (a - 10)p,$$  

(4)

where $p$ is the Hugoniot pressure in the sample and $a$ is the resulting styrofoam particle velocity. Consideration of the probable pressure–volume paths along isentropic release from successively higher pressures (Ahrens and O’Keefe, 1972) will only be a good approximation when $V_s'$ is less than or comparable to $V_s$, the zero-pressure crystal volume. The values of $V_s'$ calculated (Table 3, Fig. 5) imply that the present lunar fines, shock loaded to \( \approx 20 \text{kbar} \), the lowest pressures generated in our experiments are crushed to their intrinsic density and effectively remain near intrinsic density upon pressure release. Upon adiabatic release from higher pressures the limited data suggest that successively lower density material is produced. This result predicts that increasing quantities of glassy phases are to be expected over the present range of shock pressures, a density of only \( \approx 2.5 \text{ g/cm}^3 \) results from shock compression to \( \approx 120 \text{kbar} \) contrasting to a density of \( \approx 3.3 \text{ g/cm}^3 \), resulting from compression to \( \approx 20 \text{kbar} \). Because of the assumption regarding the pressure–volume release path, this variation in post-shock density should be viewed with some caution.
If the release adiabat paths discussed above are assumed, the post-shock temperature can be calculated from

\[ T'_s = T_i + \frac{1}{C_v} \left( \rho_i (V_m - V_i)/2 - \int p dV \right) \]  

(5)

Here \( T_i \) is the initial temperature (\(-25^\circ C\)) and \( C_v \) is the specific heat at constant pressure and the integral is taken along the release isentrope. Over the temperature range of interest, \( C_v \) will be in the range of 1.1 to 1.3 \times 10^7 \) ergs/\( ^\circ C \) g. Using these values for \( C_v \), the post-shock temperatures given in Table 5 and Fig. 6 are calculated. The temperature uncertainties indicated arise from uncertainties in the specific heat and uncertainties in the release adiabat data. In cases where detailed release adiabat measurements have been made, e.g. Lyne (1970) and Petersen et al. (1971), the actual release curves usually are convex downward in the pressure-volume plane and lie below the straight-line path (assumed for the integral term in Eq. 5). Hence, we may conclude that the present calculated post-shock temperatures represent a lower bound to the actual values. Also, since we are considering only continuum properties, the calculated temperatures are in a sense “bulk” property and we might expect that individual portions of the lunar fines will experience lower and higher temperatures. This phenomenon has been discussed by Kieffer (1971).

**Implications for Regolith Evolution**

The present data are pertinent to describing the processes of continual reworking of the initially lightly shocked and widespread lithic fragment ejecta.
Fig. 6. Theoretical minimum post-shock temperatures versus shock pressure, sample 70051.

blankets produced by the many large-scale impacts which dominated early lunar history. Although we have examined the properties of only a single (but probably typical) sample of the present lunar regolith, our results show that appreciable irreversible compaction occurs at shock pressure levels at, or below, ~20 kbar, and the measured shock impedance implies that hydrated breccia formation can take place as a result of rather mild impacts. The data of Fig. 3 imply that compaction will result from impact of objects having the shock impedance of basalt, or stony meteorites, at speeds of ~1 km/sec. That such low impact velocities are required for indentation suggests that the bombardment by secondary objects, that is, ejecta from other impacts, can play an important, and possibly dominant, role in producing the high proportion of agglutinates found in many of the lunar fines. The previous cold compression measurements on fines (Stevens and Lilley, 1970) imply that static pressures considerably in excess of 40 kbar are required to remove the porosity of these materials under quasistatic conditions. Christie et al. (1973) observed that lunar fines are indurated by 50 kbar shocks. The calculated "equilibrium" shock and post-shock temperatures in our experiments were higher than those in the Christie et al. experiment. The present experiments utilized single shocks whereas in the Christie et al. experiment, carried out within a copper container, the peak pressure is achieved via a series of wave reverberations and hence the thermodynamic path followed by the sample is more nearly isentropic.

The present calculated minimum post-shock temperatures and accompanying high post-shock specific volume (densities of ~2.5 g/cm³) achieved after compression to 100-120 kbar suggests that most of the crystalline plagioclase may have been transformed to a glassy state (quenched or diopside glass), presumably, via solid-state reaction. Such a range of pressure would be achieved in regolith fines as a result of the impact of an object with the shock impedance of a stony or iron meteoroid traveling at a speed of ~2.5 and 3 km/sec, respectively.
Because this velocity is greater than lunar escape velocity it is unlikely that substantial quantities of glass can result from impacts of ejecta. Thus glass production is likely to result only from direct meteorite impact.

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References
Abrams T. J. and O'Keefe J. D. (1972) Shock melting and vaporization of lunar rocks and minerals. The Moon 4, 244–249.
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